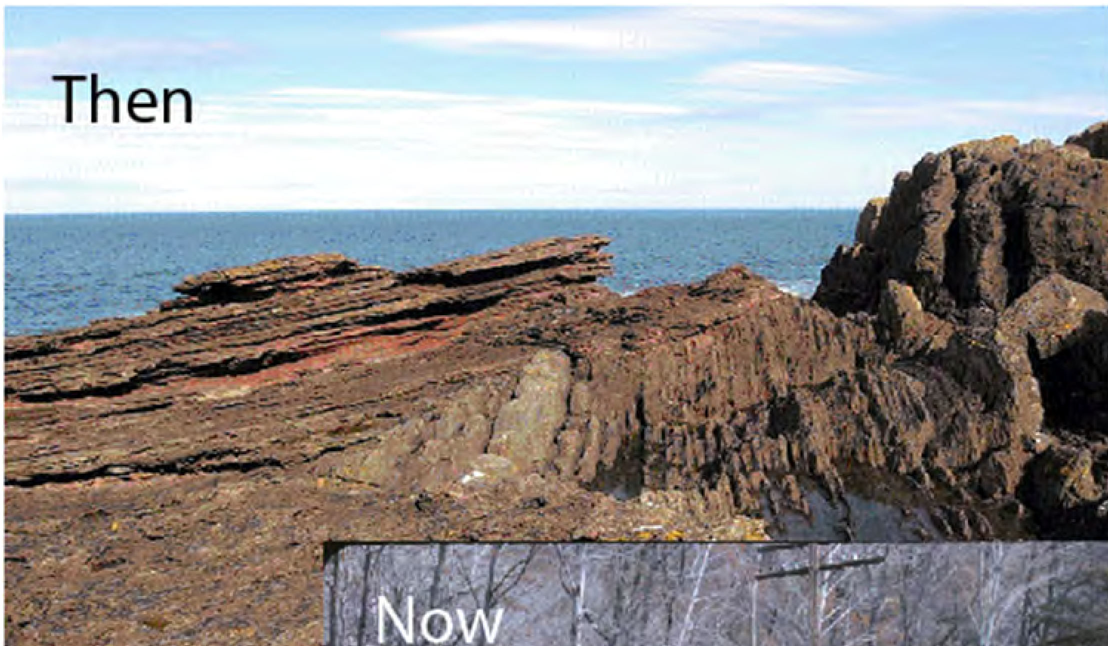
77th ANNUAL FIELD CONFERENCE OF PENNSYLVANIA EOLOGISTS

Journey Along The Taconic Unconformity, Northeastern Pennsylvania, New Jersey, and Southeastern New York



October 18-20, 2012
Shawnee-on-Delaware, PA

Hosts:
Pennsylvania Geological Survey
U.S. Geological Survey
New Jersey Geological Survey

Guidebook for the
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Journey Along The Taconic Unconformity, Northeastern Pennsylvania, New Jersey, and Southeastern New York

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Cover: Angular unconformities in geologic history: Then—James Hutton discovered this classic unconformity, where the Devonian Old Red Sandstone lies atop a Silurian graywacke at Siccar Point in Scotland, in 1787. This was his proof "that we find no vestige of a beginning, no prospect of an end." Now—The Taconic unconformity, with the Silurian Shawangunk Formation lying atop the Ordovician Martinsburg Formation, in a railroad cut in New York State. This outcrop has "gone the way of good old outcrops" (Epstein, 2012).

Cartoons: John A. Harper

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TABLE OF CONTENTS

Page

IN MEMORIAM

Thomas A. McElroy, 1949—2012

On February 8, 2012, we lost one of our former Survey colleagues, Thomas A. McElroy, to cancer at age 62. Tom passed away at home that evening after a battle with the disease that lasted for several years. He is survived by his wife, Deirdre (Dede).

Tom was born in Olean, N.Y. He served in the U.S. Air Force, including a tour of duty in Vietnam. After returning home with an honorable discharge from the military, Tom obtained a B.S. degree in geology from Harpur College of the State University of New York at Binghamton and an M.S. degree in geology from the University of Massachusetts at Amherst.

Tom came to the Survey in 1980 and worked for most of the time until his retirement in October 2010 as a hydrogeologist. He completed a number of major groundwater investigations and reports in the Allegheny Mountain section of the Appalachian Plateaus physiographic province. His countywide summaries for Fayette, Cambria, and Somerset Counties were published by the bureau. Tom also coauthored the county summaries of Indiana County, which were published by the U.S. Geological Survey. In all of these reports, Tom not only characterized the hydrogeology of the counties, but he also compiled and updated the mapping, with others, of the bedrock geology. The Somerset County report included a completely new geologic map. For this project, Tom was responsible for mapping the Mississippian Period rocks.

Tom's last hydrogeologic project was to help produce a statistical compilation of the hydrogeologic and well-construction characteristics of the geologic units of the state geologic map, and he was a coauthor of the resulting report, Water Resource Report 69. For that project, Tom reviewed each of the approximately 50,000 well records used to determine the appropriate geologic formation and physiographic section in preparation for the statistical analysis, probably his earliest foray into the use of GIS as an analytical tool.

After having done bedrock geologic mapping as part of his hydrogeological studies, it was not a major change when in his last few years with the Survey, he transferred to the Mapping Division. He first mapped the Great Bend 7.5-minute quadrangle in northeastern Pennsylvania. Then he began mapping the complexly folded and faulted bedrock in the Ridge and Valley province, in collabo-ration with retired State Geologist Don Hoskins. Together, with Tom as principal geologist, they prepared new detailed bedrock geologic maps for a number of topographic quadrangles in the Ridge and Valley province of central Pennsylvania—Lewistown , Belleville, Allensville, Newton Hamilton, McVeytown, and McCoysville. Each was published as a digital report and full-scale map.

As a result of their mapping work, Tom and Don organized and led the 72nd Annual Field Conference of Pennsylvania Geologists in 2007, *Geologic Mapping—"Walkabouts" in central Pennsylvania—1st-to-5th-Order Appalachian Mountain Folds; Folded Thrusts; Ordovician and Silurian Carbonates; Silurian Quartzites and Sandstones*.

Tom's mapping in both the Plateaus and the Ridge and Valley provinces resulted in articles in *Pennsylvania Geology* highlighting interesting and/or rare features that he discovered. Notable of these was Tom's article on recently exposed rocks on the west side of Lewistown, whose complicated geology led him to denote the area as "Oz".

Tom was also a long-time participant in the annual Field Conference of Pennsylvania Geologists. In addition to being a co-leader at the 2007 conference, he also co-led trips in 1993 (Somerset County) and 2002 (Tunkhannock). As a volunteer as well as a participant, he helped with the logistics of the trips.

Tom retired in October 2010, partly to concentrate on his health issues. He continued to participate in geologic activities until his health no longer permitted it. All of us who knew and worked with Tom miss his presence in our ranks.

Gary M. Fleeger and Donald M. Hoskins

IN MEMORIAM

John H. Way, 1943—2012

John H. Way, a former staff geologist at the Pennsylvania Geological Survey and later a professor of geology at Lock Haven University, died on February 21, 2012, in Williamsport following a brief illness. He joined our staff in 1971 and stayed for 15 years, making lasting contributions both as a field geologist and as a geological editor, before accepting a position on the faculty at Lock Haven University (LHU) in 1986.

John was born in Philadelphia in 1943 and grew up in the nearby community of Yeadon. His love of nature and geology was aroused early during visits to the Delaware County Institute of Science in Media, PA. There, under the mentorship of curator Harold W. Arndt, John developed an interest in mineralogy and geology. In addition to unique specimens that John was able to view there, John was inspired by Harold's stories of his field experiences with Sam W. Gordon's mineral collecting excursions in Pennsylvania, on which Harold had been the unofficial photographer. John started his own collection, and some of his favorite mineral-collecting areas are believed to have been Bancroft, Ontario; Herkimer, NY; and the Keystone Trappe rock quarry at Cornog, Chester County, PA.

John majored in geology at Franklin and Marshall College and earned an A.B. degree in 1965, followed by an M.S. degree in 1967 from the University of Pennsylvania and a Ph.D. in 1972 from Rensselaer Polytechnic Institute in Troy, NY. His master's degree research was a study of the sedimentary rocks that preserve a Carboniferous fossil forest that is exposed in cliffs along the Bay of Fundy near Jog-gins, Nova Scotia. His doctoral research was a study of the depositional environment and the potassium, uranium, and thorium content of Middle and Upper Devonian rocks in the Catskill Mountains of New York. While studying in Troy he met Roberta (Bobbie) Seibert, who became his wife of over 40 years. John and Bobbie raised a daughter, Mary.

During his years at the Survey, John was responsible for completing several major publications, most notably a major study of the geology of the Altoona area, *Geology and Mineral Resources of the Blandburg, Tipton, Altoona, and Bellwood Quadrangles, Blair, Cambria, Clearfield, and Centre Counties, Pennsylvania,* published with Rodger T. Faill and Albert D. Glover as Atlas Report 86 in 1989, and another Atlas Report, 154cd, *Geology and Mineral Resources of the Washingtonville and Millville Quadrangles, Montour, Columbia, and*

Northumberland Counties, Pennsylvania, published in 1993. But these reports barely tell the story of John's contributions here. He published many articles in *Pennsylvania Geology* and articles and abstracts in such outside publications as the Geological Society of America Abstracts with Programs and the guidebooks of the Field Conference of Pennsylvania Geologists, many coauthored with Survey colleagues. Research topics included the Devonian Tioga Ash Beds and Bald Hill Bentonites (with Robert C. Smith, II, Samuel W. Berkheiser, and Mary K. Roden), 19th century iron-making at Pine Grove Furnace, and the geology of South Mountain, Cumberland County. He also was a coauthor with Thomas M. Berg, Michael K. McInerney, and David B. MacLachlan of the Survey's *Stratigraphic Correlation Chart of Pennsylvania* (the "strat chart").

John's contributions to understanding the geology of the Ridge and Valley physiographic province and the economic geology of Pennsylvania were, and continue to be, significant. For example, the "strat chart" helps to define the framework of the Marcellus and Utica shale gas horizons. The Survey's Tioga Ash Bed study provided information on the base of the Marcellus play zone, but also delineated the correct direction of time transgression relative to lithologic facies. His careful work on volcanic ash beds in the Middle and Upper Ordovician Union Furnace section helped establish that exposure as Pennsylvania's de facto type section for those beds and the basis of the bentonite nomenclature used in Pennsylvania. His work on the lowermost Devonian Bald Hill Bentonites arose from the hypothesis that termination of extended periods of carbonate deposition would be marked by volcanic ash beds. The ashes marking the end of the Silurian Wills Creek, Tonoloway, Keyser, Coeymans, and New Scotland carbonates were found within an hour of searching in the transition to the lowermost Devonian Mandata black shale at Bald Hill, Blair County. Weekend trips with family eventually extended the known range of surface exposures of the Bald Hill Bentonites from near the Adirondack Mountains in New York to McDowell, WV. In the process, the resulting detailed stratigraphic sections disproved a then-current assumption that punctuated aggradational cycles (PACs) were time-stratigraphic surfaces.

Despite John's contributions that helped set the stage for the Marcellus gas development, he was concerned about its potential impact on the environment. John never profited from the Marcellus but, true to his principles, he volunteered countless hours to conservancies in central Pennsylvania seeking to protect watersheds.

John's ability to write in a clear and interesting way for the nongeologist was amply demonstrated in his publication, *Your Guide to the Geology of the Kings Gap Area, Cumberland County, Pennsylvania,* published by the Survey in 1986. John also used his excellent communication and writing skills to advantage during a three-year stint as a geologic editor at the Survey, from 1974 through 1977, when he helped convert a number of Survey publications from rough manuscripts to finished products. He performed a similar task as a volunteer, spending many hours of his personal time as the editor of the 300-page book, *The Mineralogy of Pennsylvania, 1966-1975,* by Robert C. Smith, II, published by the Pennsylvania Chapter of Friends of Mineralogy in 1978. To this same work, John contributed more than 50 finely executed pen-and-ink sketches. From alloclasite to the back piece geologic time scale, all of the drawings were drafted by John.

At Lock Haven, John proved to be a very effective teacher, conveying to his students not only his technical expertise, but his love of nature and excitement about the geological processes that shape the earth. John received the Teaching and Learning Center's Peers Choice

Award at LHU in 2004. Through papers published while he was there, he made significant contributions to understanding the regional and environmental geology of the Lock Haven and Williamsport areas. Among his contributions were four field guidebooks covering the geology of the Erie, Gettysburg, South Mountain, and Johnstown areas. He also made one additional contribution to the Survey while teaching at LHU, authoring the chapter on the physiography of the Appalachian Mountain section for the Survey's *Geology of Pennsylvania* compendium.

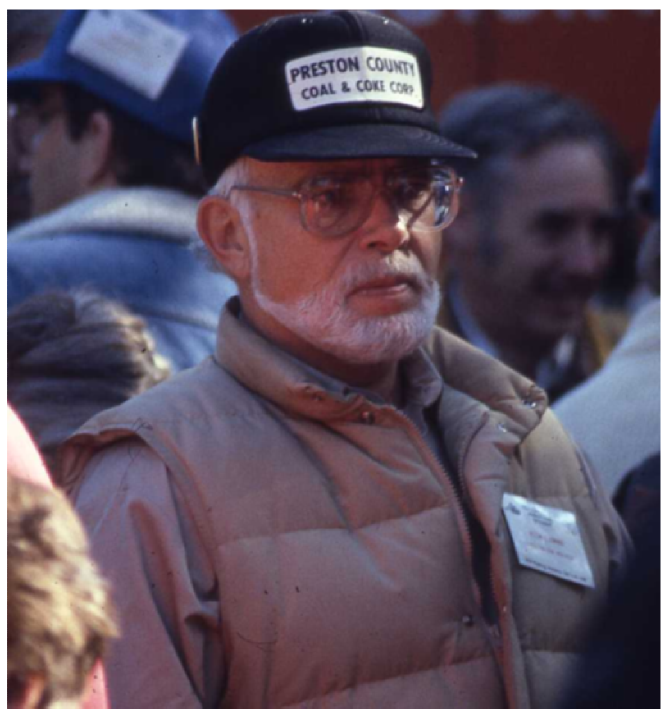
John's enthusiasm and energy were infectious. Anyone who came into contact with him was uplifted and energized by John's positive outlook and the joy that he took in everything that he did. This extended well beyond his work to include his community, his church, and his family. Those of us who were privileged to know and work with John are the better for it, and we will miss him.

John H. Barnes and Robert C. Smith, II

IN MEMORIAM

William E. Edmunds, 1932—2010

Another long-time participant of the Field Conference died on August 21, 2010. Bill Edmunds was largely a fixture at the annual conference for many years. He also was a co-leader of 6 Field Conferences during the 1980s and 1990s: 1981- Wellsboro; 1986- Huntingdon; 1988- Hazelton; 1993- Somerset; 1996- Chambersburg; 1997- Scranton



Bill worked for the Survey from 1960 to 1977, when he left for the world of consulting. He was an expert on the stratigraphy and sedimentology of the Pennsylvanian and Mississippian of the central Appalachians. In fact, there was hardly anything in geology that he couldn't explain. Likewise, there were few non-geological things that he didn't know anything about. He understood astronomy, physics, chemistry, botany and zoology, philosophy, Greek mythology, American history, photography, and, as the saying goes, "The list goes on and on."

Bill was a 'mentor' to many Survey (and other) geologists. He was a very intelligent person who was always willing to help someone understand rather complex stratigraphy, and he had a wonderful sense of humor. Bill was easy to learn from because he had the ability to take a complex concept or discussion, synthesize the main points, ignore unimportant details and rephrase the concept in plain language.

He was also a premier coal geologist who was responsible for following up what George Ashley had started. His work at the Survey during 1960s and early 1970s was first class in terms of both quantity and quality. The coal stratigraphy of Clearfield County is complex, and is further complicated by numerous strike slip faults. He, Tom Berg, Gary Glass, and Al Glover mapped most of the county, all under Bill's supervision as Chief of the Coal Section from 1967 to 1977. His contributions to the Survey far overshadowed many others.

Bill was extremely productive. For example, he single-handedly produced the last statewide accounting of coal reserves with the 1972 publication of Information Circular 72, *Coal Reserves of Pennsylvania: Total, Recoverable and Strippable (January 1, 1970).* In 2009, an analysis of the cost of re-evaluating the coal reserves for the entire Commonwealth resulted in an estimate somewhere in the neighborhood of $20 million. Bill was also responsible in 1966 for the initiation of TASIC (Temporarily Available Stratigraphic Information Collection). That program acquired large amounts of data from temporarily exposed strip-mine highwalls, and has since been expanded to include any temporary exposures.

He was the leader in defining the stratigraphy of the Mississippian in Pennsylvania. His last major work is a Magnum Opus of the Mauch Chuck Formation, and is currently being prepared for publication by the Survey.

William A. Bragonier, Gary M. Fleeger, and Tom Berg

Group Photo of the 2011 Field Conference of Pennsylvania Geologists at Phil Myers' Farm in West Virginia (photo by Yuriy Neboga).

A JOURNEY ALONG THE TACONIC UNCONFORMITY: INTERPRETATIONS, PERPLEXITIES, AND WONDERMENTS NORTHEASTERN PENNSYLVANIA, NORTHERN NEW JERSEY, AND SOUTHEASTERNMOST NEW YORK

Jack B. Epstein and Peter T. Lyttle
US Geological Survey
Reston, VA

Introduction

The "transitional" contact between Silurian and Ordovician rocks in central Pennsylvania becomes unconformable in eastern Pennsylvania to southeastern New York as the hiatus widens (Figure 1). Following the northeastward decrease in intensity of deformation in the Ridge and Valley through New Jersey, this trip will begin with the high-angle contact between the Tuscarora and Hamburg sequence at the Schuylkill River and proceed for 120 mi (193 km)

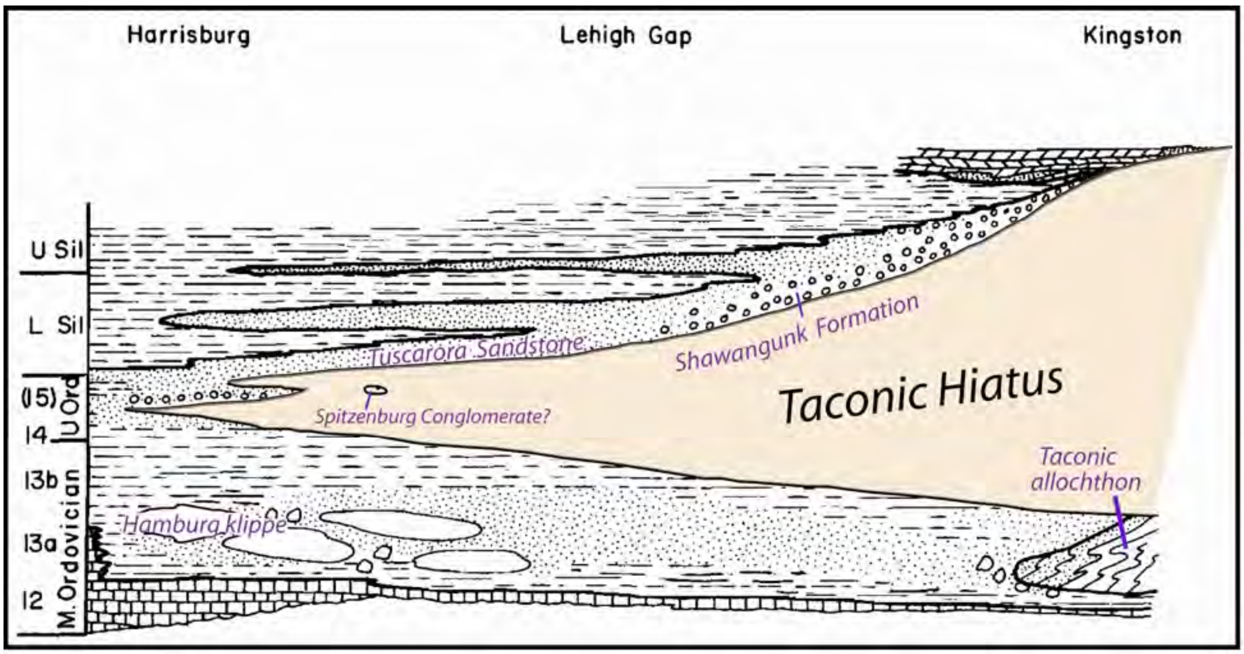


Figure 1. Schematic diagram showing the increasing west-to-east hiatus between Ordovician and Silurian rocks from south-central Pennsylvania to southeastern New York (modified from Rodgers, 1971, p. 1145). The continuous but sporadic sedimentation of Ordovician through Silurian rocks west of Harrisburg, PA, is represented by Martinsburg (or Reedsville) shales overlain by Bald Eagle (or Oswego) sandstones, then Juniata red beds, and finally sandstone of the Tuscarora. At the east end of the area in the diagram the Shawangunk pinches out and Late Silurian rocks overlie Ordovician rocks of the Taconic allochthon. This implies that the base of the Shawangunk and Tuscarora transgresses eastward from the Lower Silurian to Upper Silurian, a fact not in evidence as yet. Note that no Middle Silurian is recognized in untainted stratigraphic understanding, despite the inclusion of same elsewhere in this guidebook.

Epstein, J.B. and Lytle, P.T., 2012, A journey along the Taconic unconformity: Interpretations, perplexities, and wonderments, northeastern Pennsylvania, northern New Jersey, and southeasternmost New York, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 1-36.

along the very low-angle unconformable contact between Lehigh Gap, PA and Ellenville, NY. Past geologists have compared and rated Taconic and Alleghanian tectonism based on perceived geologic relations on either side of the contact, leading to continuing debate that has persisted for more than 170 years. The age and stratigraphic framework of the Ordovician rocks also has been disputed. The goal of this trip is to examine many localities along the contact, so that individual interpretations will benefit from the detailed mapping by the New Jersey, Pennsylvania, and U.S. Geological Surveys. By discussing our interpretation of the relative intensities of deformation, we will suggest predominant Alleghanian deformation along the contact and propose zones of increasing southeastward Taconic deformation away from the contact. The perplexing story of events during the Taconic hiatus, lasting perhaps 10-30 million years, will be illuminated by an unusual diamictite in southeastern New York. Several glaciations have profoundly affected the landscape of this area, especially the latest Wisconsinan. We cannot limit ourselves to bedrock geology on this trip, so the effects of glaciation also will be highlighted.

The Field Conference of Pennsylvania Geologists has visited many sites along the Ordovician-Silurian boundary in the past; several of those are unabashedly plagiarized on this trip, although new thoughts will be entered into the record. It offers an opportunity for younger geologists to see these exposures for the first time. At the time of preparation for this trip, access to a couple of the stops was in jeopardy, so alternative stops have been prepared. As a sad note, access to many instructive geologic localities has declined over the past many years due to safety, litigation, and terrorist-security concerns. We lament this loss and hope that the documentation in this field-trip will record some of these sites for posterity.

THE STOPS

Ten stops are planned for this field conference (see map on p. 166). They include:

1. Schuylkill Gap, PA—right-angle contact between Ordovician rocks of the Hamburg klippe and Tuscarora Sandstone; “Appalachian” structure in post-Ordovician rocks; greywacke.
2. Penn Big Bed Slate Quarry, PA—Stratigraphy and structure of the upper Martinsburg Formation.
3. Lehigh Gap, PA—Unconformity between the Shawangunk (Silurian) and Martinsburg (Ordovician) formations; folds and faults; northwest translation on bedding faults; rockfall mitigation.
4. Delaware Water Gap, PA—Silurian stratigraphy; slaty cleavage and age of deformation.

4A. Resort Point Overlook, Delaware River, PA—Bloomsburg wedges; defining a red bed formational boundary; history of tourism (*time permitting*).

5. Yards Creek Pump-Storage Generating Station, NJ—Martinsburg slaty cleavage affected by unconformably overlying Silurian rocks.
6. Home Depot parking lot, Newton, NJ—Retrodeforming Martinsburg deformation.
7. Lusscroft Farm and Beemerville syenite, NJ—An early Silurian intrusion into the Martinsburg Formation and unconformably overlain by the Silurian Shawangunk Formation.

8. High Point State Park, NJ—Summary of the Taconic unconformity in northern New Jersey; the Shawangunk Formation elucidated; regional glaciations.
9. Otisville railroad cut, NY—Two classic unconformities, Holocene and Taconic; enigmatic diamictite at the Taconic boundary that requires explanation.
10. Ellenville arch, NY—Retro-deformation of the Shawangunk and Martinsburg Formations; deformation zones in the Martinsburg Formation.

Stratigraphy of Rocks above and below the Taconic Unconformity, Northeastern Pennsylvania, New Jersey, and southeastern New York

A variety of rocks, in many structural configurations, are exposed for more than 110 mi (177 km) above and below the unconformity separating Ordovician and Silurian rocks in northeastern Pennsylvania, New Jersey, and southeastern New York (Figure 2). The basal Silurian clastic units, the Shawangunk Formation transitioning into the Tuscarora Sandstone in the southwest, is succeeded by a variety of sandstones, shales, and some red beds of the Clinton Formation, which is overlain by sandstone and fine clastics of the Bloomsburg Red Beds. In central New Jersey and New York the Shawangunk, also termed the Green Pond Conglomerate, bevel across Ordovician-through-Precambrian rocks in a narrow faulted syncline about 15 mi (25 km) southeast of the main ridge of Silurian rocks in New Jersey and New York. That ridge is named Blue, Kittatinny, and Shawangunk Mountain from southwest to northeast in the trip area.

The Ordovician rocks are dominated by several members in the Martinsburg Formation (Figure 2), having suffered (enjoyed?) a history of nomenclatural variation, related to discussions of its age, stratigraphic variations, and structural framework. These discussions have gone on for more than 150 years, and it is not over yet. Likewise, the Silurian rocks have had an interesting nomenclatural history, culminating in the interesting facies relations shown in Figure 3. Summaries of the stratigraphic units within the field trip area are given in Appendices 1 and 2.

Ordovician Rocks

The Martinsburg Debate

The southwest section of the field trip area is dominated by rocks of the far-travelled Hamburg klippe and by sandstones and shales comprising a deep syncline in Shochary Ridge (see Appendix 1). This is part of an area full of puzzlements, discussed at Stop 1, under the designation “Hamburg Triangle”.

The subdivision of the Martinsburg Formation in eastern Pennsylvania and New Jersey has been debated for nearly 100 years, and has been the discussed on several of our previous Field Conferences (1967, 1982, 1984, 2001). To plagiarize from Epstein (2001, p. 6-8), 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

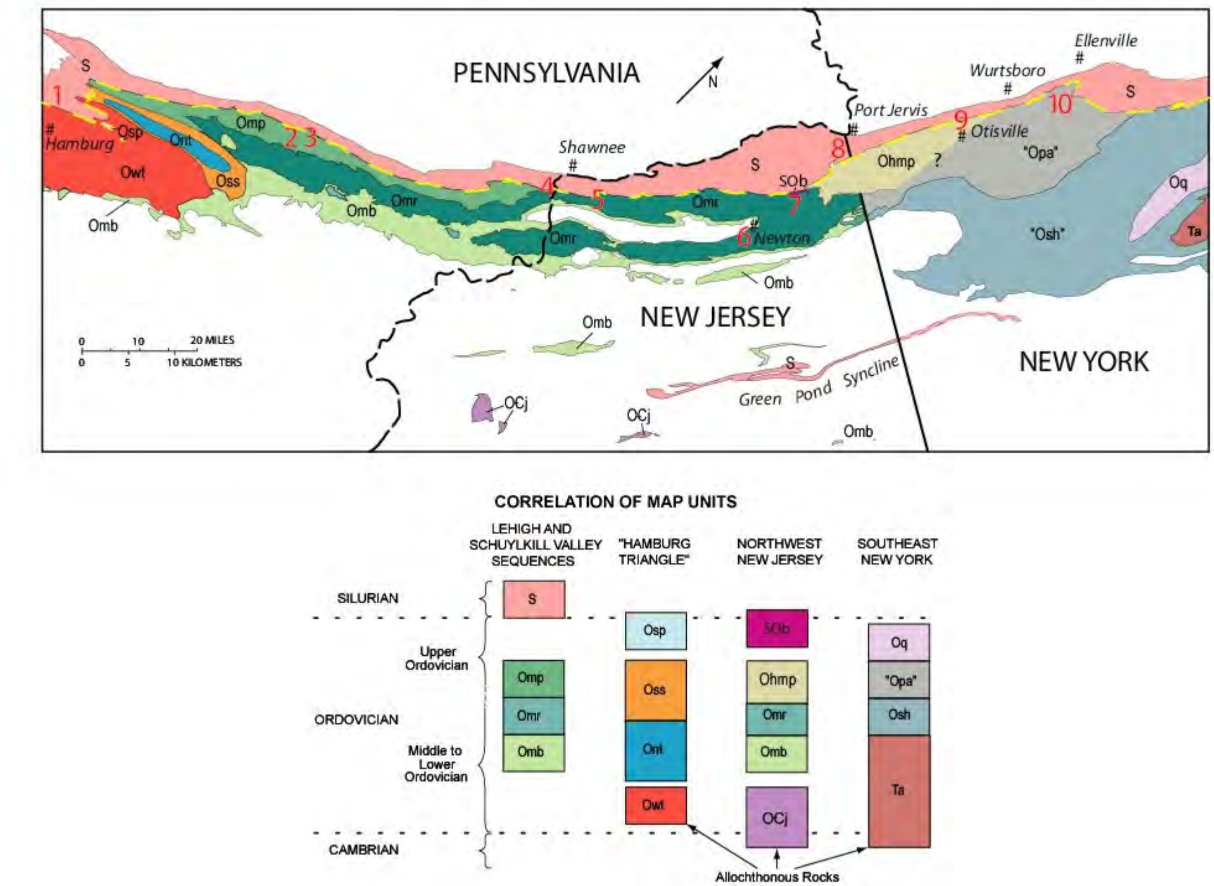


Figure 2. Geologic map showing stratigraphic units straddling the Ordovician-Silurian contact in eastern Pennsylvania, northern New Jersey, and southeastern New York. The Taconic unconformity is shown by the dashed yellow line. See Appendix 1 for description of stratigraphic units. S—Silurian rocks, Bloomsburg Red Beds, Clinton Formation, Tuscarora Sandstone, and Shawangunk Formation; Omp—Pen Argyl Member of the Martinsburg Formation; Omr—Ramseyburg Member of the Martinsburg Formation; Omb—Bushlkill Member of the Martinsburg Formation; Osp—Spitzenburg Conglomerate and sandstones at Sharps Mountain (the lovely hill, Spitzenburg [see Day 1 Road Log, Figure 13], is the easternmost of the two exposures); Oss—Shochary Sandstone of the Shochary Ridge sequence; Ont—New Tripoli Formation of the Shochary Ridge sequence; Owt—Windsor Township Formation in the Hamburg klippe; Sob—Beemervillesyenite intrusion; Ohp/Omk—High Point Member of the Martinsburg Formation/shale and greywacke at Mamakating and Sandstone at Pine Bush of Epstein and Lyttle (1987); OCj—rocks of the Jutland and Peapackklippen; "Opa"—Pen Argyl shale of Rogers et al. (1990), and Austin Glen Formation of Fisher et al. (1971); Osh—Snake Hill of Rogers et al. (1990), and Normanskill Shale of Fisher et al. (1971); Ta—rocks of the Taconic allochthon. Modified from Lyttle and Epstein (1987), Drake et al. (1996), Fisher et al. (1970), and Rogers et al. (1990).The "Hamburg Triangle" is discussed at Stop 1.

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 Appendix 1) in almost the same way as

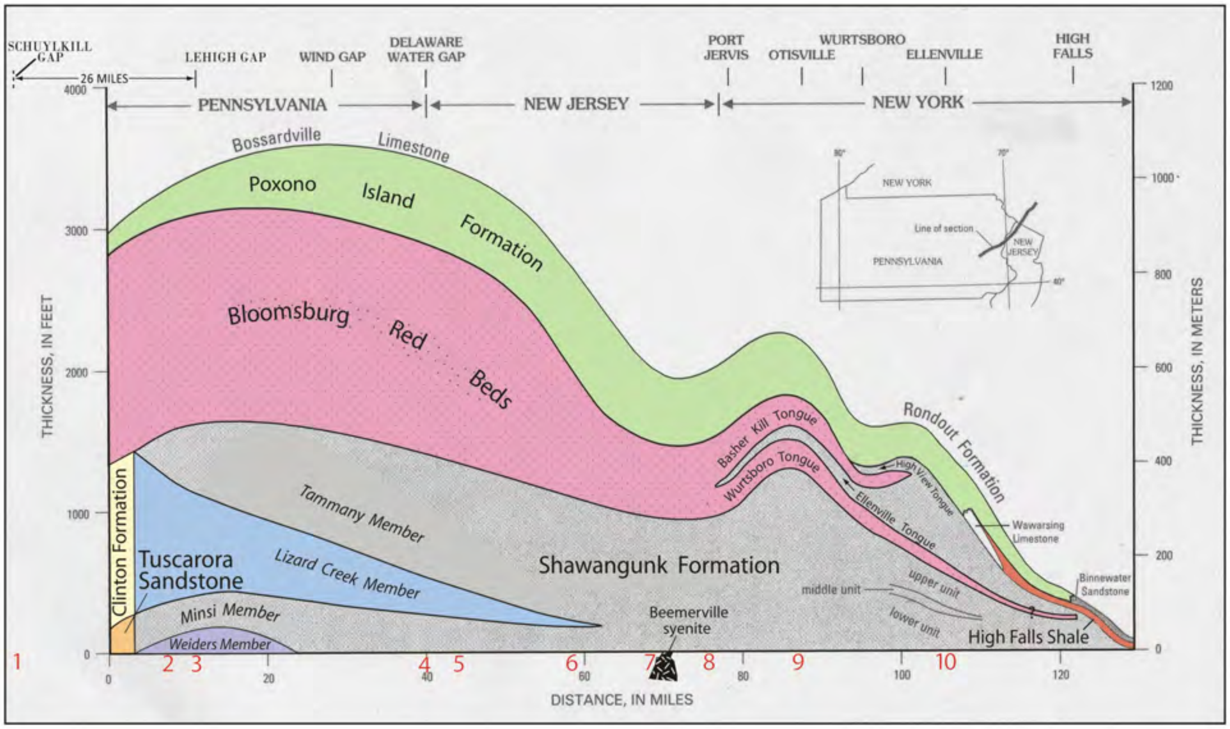


Figure 3. Generalized stratigraphic section of Silurian rocks between Schuylkill Gap in eastern PA to High Falls, NY, showing facies relations between the Bloomsburg Red Beds, High Falls Shale, Shawangunk Formation, Clinton Formation, and Tuscarora Sandstone. Red numbers are position of field stops perpendicular to the strike of the Ordovician-Silurian contact. Modified from Epstein (1993).

defined by Behre (1933). This should not be surprising because the best geologists 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 graywacke is interbedded with the slate. Those who accepted a two-fold interpretation have estimated that the Martinsburg is as thin as 3,000 ft (915 m) (Stose, 1930), whereas those who support the idea of three members have estimated thicknesses of more than 10,000 ft (3,050 m) (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 River 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., Epstein, 1980, fig. 6). 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 in (15 cm) thick and are generally less than 2 in (5 cm) thick. This laminated to thin-bedded characteristic is present everywhere in the member over an outcrop width of nearly 5 mi (8 km) in places and an outcrop length of more than 30 mi (48 km) in Pennsylvania. The Bushkill extends for more than 30 mi (48 km) in the Great Valley of New Jersey to and beyond the New York border. The overlying Ramseyburg Member comprises about 20 percent graywacke. The slates in the graywacke are thin bedded at the base and become thicker bedded upwards. The first 1,000 ft (300 m) 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 ft (6 m) thick. Most of these quarries are now inactive and flooded, but we will visit one of the few active ones on the first day of the field trip at Stop 2. The thick clear slate beds are the source for many pooltables. The thick-bedded material is not repeated south of the Ramseyburg outcrop belt, a fact long known to the slate quarrymen of the area.

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; Finney, 1985). A study of graptolites in the Delaware Water Gap area by Parris and Cruikshank (1992) supports the three -fold interpretation. 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.

Ordovician Rocks in the Wallkill Valley, Southeastern New York

The following discussion is unabashedly adapted from Peter Lyttle (*in* Epstein and Lyttle, 1987).

There has been, and continues to be, a great deal of confusion when dealing with the Ordovician clastic sediments of southeastern New York State. A brief (and by no means exhaustive) review highlighting some of the earlier work in these rocks is helpful to emphasize the complex history of names, and establish proper correlation of units. All the rocks in the area from south of Rosendale, New York, to the New Jersey border in the Wallkill Valley are considered to be units of the Martinsburg Formation (see Epstein and Lyttle, 1987, fig. 1.

In the Wallkill Valley, a thick section of glacial deposits covers much of the bedrock. Insofar as this makes structural analysis of the rocks in places difficult, if not impossible, this has a definite influence on resolving the stratigraphy. The tracing of faults and sometimes folds in the Ordovician rocks is extremely difficult, which explains why I and other geologists such as Vollmer (1981) and Kalaka and Waines (1986) have chosen to divide map areas into structural domains that can be defined in general descriptive terms. Added to this is the century-old problem regarding which Ordovician clastics in the Hudson Valley region are part of the far-traveled Taconic allochthon and which are part of the parautochthonous flysch that rests conformably on Middle Ordovician carbonates of the North American shelf. This is a problem that seems to be satisfactorily resolved to the southwest in New Jersey and Pennsylvania (Perissoratis et al., 1979; Lash and Drake, 1984; Lash, 1985; Lash et al., 1984) but remains a critical problem in parts of southern New York State, particularly along the Hudson River. There is an irony to this, since the existence of the far-traveled rocks was recognized much later in Pennsylvania (Stose, 1946) than in New York.

Mather (1840) first proposed the name "Hudson River slate group" for rocks that he had previously referred to as "transition argillite" (Mather, 1839). Subsequently, this name has seen at least seven variants including Hudson River Series, Hudson River Group, and Hudson River Formation. In addition to the variations in the name, the use of these names has been extended north into Canada and as far west as Wisconsin. Holzwasser (1926) gives a useful account of the tortured history of Hudson River nomenclature; unfortunately, she also decided to use the name Hudson River formation for the shales and graywackes of the Newburgh Quadrangle. Although the name has generally been abandoned since Ruedemann's work in the beginning of this century, it is still loosely used and misused in a variety of publications to this day. Ruedemann (1901, p. 561) first used the name Normanskill Shale for rocks in the gorge of the Normans Kill near Kenwood, New York. This type locality turns out to be one of the more spectacular exposures of melange in the northern Appalachians, as attested to by the extremely detailed mapping of the structures by Vollmer (l981). It should not be too surprising, therefore, that there has also been considerable confusion in the use of this geologic name. Later, when mapping the Catskill quadrangle, Ruedemann (1942) recognized two belts of rock that he included in the Normanskill Shale. The western "grit belt" was named the Austin Glen Member and the eastern "chert belt" was named the Mount Merino Member. Most geologists today would agree that both of the type localities for these members are within the Taconic allochthon (*sensu stricto*); that is, the rocks are part of the far-traveled slope-rise sequence. The name Snake Hill Shale was first used by Ulrich (1911), although he based his discussion on the work of Ruedemann who later published a number of papers using this name (Ruedemann, 1912, for example). The type locality for this unit is on the east side of Saratoga Lake. Berry (1963) suggested abandoning the name because restudy of this region showed that what was mapped as Snake Hill contains three different lithic units, all of which contain elements of the distinctive fauna that Ruedemann used as the unit's diagnostic feature. This points to yet another problem in the nomenclature of the clastics of the Hudson Valley region. Ruedemann and others often failed to discriminate between biostratigraphic and lithostratigraphic units, making it extremely difficult for later workers to fully appreciate the problems inherent in using a particular name.

More recently, Fisher (1962) and Offield (1967) have used the names Mount Merino Shale, Austin Glen Graywacke, and Snake Hill Shale for the lower, middle and upper units of the parautochthonous Middle and Upper Ordovician shales and graywackes that are found west of the Hudson River in the Wallkill Valley. Later, Fisher (1969, 1977; and Fisher et al., 1970) made a number of modifications to the mapping and naming of Ordovician clastic units in the vicinity of the Hudson River at the latitude of our field trip, but very little new work closer to the unconformable contact with the Silurian Shawangunk to the west has been published. For a summary of the most recent work near the Hudson River, particularly in the region underlying Marlboro and Illinois Mountains, see Waines (1986).

Offield (1967) produced a wealth of new stratigraphic and structural information in the Goshen 15-minute quadrangle (Middletown, Goshen, Warwick, and Pine Island 7.5-minute quadrangles) and recognized that his units might correlate with Behre's (1933) tripartite subdivision of the Martinsburg Formation in Pennsylvania. This subdivision was later refined by Drake and Epstein (1967) who recognized a lower thin-bedded slate unit called the Bushkill Member, a middle graywacke-rich unit called the Ramseyburg Member, and an upper thick-bedded slate unit called the Pen Argyl Member. Berry (1970) was one of several people to recognize significant similarities between the Delaware Valley sequence of Drake and Epstein and the sequence of rocks in the Wallkill Valley (Fisher, 1962; Offield, 1967). I feel that all of the names that Offield (1967) chose for the units of what he refers to as "the shale sequence" in the Wallkill Valley should be discontinued. One reason to do this is to avoid unnecessary confusion with the Normanskill Shale and its members that are clearly part of the far-traveled Taconic allochthon. Another reason, which is even more important, is that a better correlation can be made with units mapped in Pennsylvania and New Jersey. I have not done all of the detailed mapping that is necessary to establish these correlations in detail, but I feel confident that the correlations proposed herein are correct overall.

I believe that it is appropriate to refer to all of the parautochthonous Ordovician clastics in the Wallkill Valley as the Martinsburg Formation of Middle to Upper Ordovician age. All of the Ordovician clastics in the Great Valley from eastern Pennsylvania through northern New Jersey to the New York State border have been mapped in detail (1:24,000 scale), and summarized in Drake et al., 1996. The parautochthonous sequence west and southwest of Albany, New York, is not contiguous with the rocks of the Wallkill Valley; nor has the stratigraphy of the rocks near Albany been done in sufficient detail to warrant using the names established for that area in the Wallkill Valley.

There has been debate in eastern Pennsylvania over whether the Martinsburg is a tripartite sequence with a lower slate member, a middle graywacke member, and an upper slate member, or a bipartite sequence with an upper graywacke member and a lower slate member (see Lash et al, 1984, for a summary of this debate). I feel strongly that the published detailed mapping, which ultimately must answer all questions of this sort, supports the tripartite subdivision first discussed by Behre (1933) and later named along the Delaware Valley by Drake and Epstein (1967). The question that must now be answered is, how far away from the Delaware Valley can the three members of the Martinsburg be mapped? The upper Pen Argyl Member, which contains thick-bedded slates (up to 25 ft [8 m] thick), has been extensively quarried in Pennsylvania from the New Ringgold 7.5-minute quadrangle in the west, a few mi (km) east of the Schuylkill River, to the Stroudsburg 7.5-minute quadrangle in the east where it disappears beneath the Silurian Shawangunk Formation by structural overlap (Epstein, 1973). Based on my mapping and that of other geologists, it is not found in northern New Jersey and southern New York. However, the Pen Argyl correlates in part with rocks that I have mapped in the western Wallkill Valley unconformably beneath the Shawangunk, and that I am herein informally naming the shale and graywacke at Mamakating, subsequently called the Mamakating (Epstein and Lyttle, 1987, fig. 2). The Mamakating represents the upper part of the Martinsburg in the western Wallkill Valley and is named for the excellent exposure seen at Stop 7 along Route 17 (just east of Wurtsboro exit) in the eastern part of the Mamakating Township. The Mamakating first appears from beneath the Shawangunk in the Otisville 7.5-minute quadrangle, New York and extends northeastward. All of the Martinsburg we shall be seeing on this field trip is within the Mamakating. The Ramseyburg Member extends from the New Tripoli, 7.5-minute quadrangle, Pennsylvania, about 12 mi (7.5 km) east of the Schuylkill River, to the Middletown 7.5-minute quadrangle, New York. To the northeast it correlates for the most part with a unit that we are herein informally calling the sandstone at Pine Bush. The sandstone at Pine Bush extends from the High Point area of New Jersey through the Middletown and Pine Bush 7.5-minute quadrangles, New York, where it is thickest, and appears to die out somewhere in the vicinity of the southwest corner of the Gardiner 7.5-minute quadrangle. There are excellent exposures of the Pine Bush along Route 17K that underlie the unnamed hills 1.6 mi (1 km) west of Montgomery, New York, in the Pine

Bush 7.5-minute quadrangle. Since the details of the facies changes in the middle and upper Martinsburg have not been sufficiently mapped in southern New York State, it is safest to say that the combined Ramseyburg and Pen Argyl correlates with the combined Pine Bush and Mamakating. It may eventually be determined that the Pen Argyl correlates with all of the Mamakating and the uppermost Pine Bush. The Mamakating is everywhere unconformably overlain by the Shawangunk Formation. It grades conformably downward and laterally into the Pine Bush, and the contact is arbitrarily put where beds of medium-grained, clean protoquartzite make up more than 5 percent and are thicker than 2 in (5 cm). In most places in the Wallkill Valley the Pine Bush grades upward into the Mamakating, but to the southwest near High Point, New Jersey, it is unconformably overlain by the Shawangunk Formation. We have not done enough detailed mapping in the Pine Bush, Walden, and Gardiner 7.5-minute quadrangles, New York, to resolve what happens to the Pine Bush to the northeast. From reconnaissance, it would appear to pinch out and grade laterally into the Mamakating somewhere near the northeast corner of the Pine Bush quadrangle. The lower Bushkill Member of the Martinsburg has, by far, the greatest areal extent of the three members of Drake and Epstein (1967). It extends as far southwest as Reading, Pennsylvania (and probably considerably farther) and northeast at least as far as the Newburgh, New York area.

Several very general points can be made about the Martinsburg Formation. From eastern Pennsylvania to the field trip area in southern New York, the composite thickness of the Martinsburg appears to remain fairly constant with ranges estimating from about 8,000 to 12,800 ft (2,440 to 3,900 m). It is possible that the thickness decreases going towards the northeast, perhaps by as much as 3,000 ft (915 m). All thickness estimates may be on the generous side, because of the large number of thrust faults that duplicate portions of the unit, particularly the lower Bushkill Member.

The sedimentology of the lower part, or Bushkill Member, of the Martinsburg remains remarkably constant along strike from eastern Pennsylvania through southern New York. However, the middle part of the Martinsburg shows considerable facies variation along strike. To the southwest in Pennsylvania, the Ramseyburg Member rarely contains more than 20 percent medium- to very thick-bedded graywacke beds. Also, going up-section in the Ramseyburg, the thickness of slate beds increases dramatically near the contact with the Pen Argyl. From High Point, New Jersey northward, the Pine Bush (~ lower part of the High Point Member of Drake (1990) commonly contains up to 50 percent clean, medium- to very thick-bedded sandstone, and as best as we can tell from reconnaissance, the thickness of shale beds does not increase going up in section. Both of these factors would appear to suggest that the middle part of the Martinsburg is becoming more proximal to the northeast. The upper part of the Martinsburg also shows dramatic facies changes along strike. Although this part of the section is dominated by shales or slate everywhere, in eastern Pennsylvania, slate beds in the Pen Argyl Member are commonly 12 ft (3.7 m) thick, and can be as thick as 25 ft (7.6 m). In New York, the shale beds in the Mamakating (~ High Point Member) rarely exceed 3 in (7.6 cm) in thickness.

What IS the Martinsburg Formation?

The Martinsburg Formation extends for more than 100 mi (160 km) from southeastern New York to where it dives under the Tuscarora Sandstone at the west end of the field trip area. These rocks do not reappear in outcrop to the west. Here, it is about 115 mi (185 km) northeast of the type area near Martinsburg, West Virginia, where the Martinsburg was named by Geiger and Keith (1891). It apparently was first adequately described by Keith (1894) who characterized the formation as partly calcareous shale. In the Shenandoah Valley of Virginia, the Martinsburg is possibly about 3,000 ft (915 m) thick, predominantly calcareous shale, but containing greywacke sandstone higher in the section (Butts, 1973). The continuity of the Martinsburg belt is blocked by the rocks of the structurally overlying Hamburg klippe, which extends from near the Susquehanna River to just east of Stop 1 of this field trip. The Martinsburg in the Great Valley of Virginia, West Virginia, and Maryland contains fossiliferous clastic rocks, termed "Reedsville" in places, as well as the characteristic shale and greywacke (Diecchio, 1985, McBride, 1962). In Pennsylvania north of Blue Mountain, the name Reedsville Shale is applied to the finer-grained Ordovician clastic rocks. These fossiliferous rocks are more proximal (= nearshore) than the turbidite-bearing Martinsburg of eastern Pennsylvania. They may be analogous to isolated rocks in the Shochary Ridge sequence (Figures 2 and 4). Similar fossiliferous shales can be found in southeastern New York (Feldman et al., 2009). Clearly, there are significant differences in the 200+ mi (320+ km) length of the Ordovician clastic belt to warrant reexamination of the nomenclature of these rocks. The Martinsburg of eastern Pennsylvania is significantly different than the type Martinsburg to ask, "should it even be called Martinsburg"? The rocks identified as Martinsburg in the Harrisburg area (Wise et al., 2010, Stop 12) is a very fine-grained shale that is very different from any of the Martinsburg east of the klippe. It most closely resembles the New Tripoli Formation of the Shochary Ridge sequence. Perhaps a better understanding of the regional distribution of the various Ordovician clastic lithofacies is warranted at this time.

Silurian Rocks

The Tuscarora/Clinton-Shawangunk Relationship

Basal Silurian sandstones, quartzites, and conglomerates hold up a ridge system that extends throughout the length of the Appalachians of the Eastern United States. In the area of this field trip the ridge is made up of Blue, Kittatinny, and Shawangunk Mountains. In Pennsylvania, the Tuscarora Sandstone and overlying Clinton Formation (Clinton Group is some places) merges eastward into the Shawangunk Formation while the superjacent Bloomsburg Red Beds is happy to remain on top of all of these units (Figure 3). The change takes place about one mi (1.6 km) east of the Schuylkill River. As explained at Stop 4, Delaware Water Gap, the Shawangunk is subdivided into four members; two quartzite-conglomerate members (Weiders and Minsi) at the base and top (Tammany), separating sandstone-shale sequence in the middle (Lizard Creek). The Tammany and Weiders Members pinch out westward, leaving the Lizard Creek and Minsi, which according to stratigrpahic rules, become the Clinton and Tuscarora.

Silurian Clastic Facies in New Jersey

In the east section of the field trip area, the Silurian sequence seen in Delaware Water Gap thins dramatically across New Jersey towards southeastern New York (Figure 3). In New Jersey, the shales of the Lizard Creek Member become less abundant and more scattered

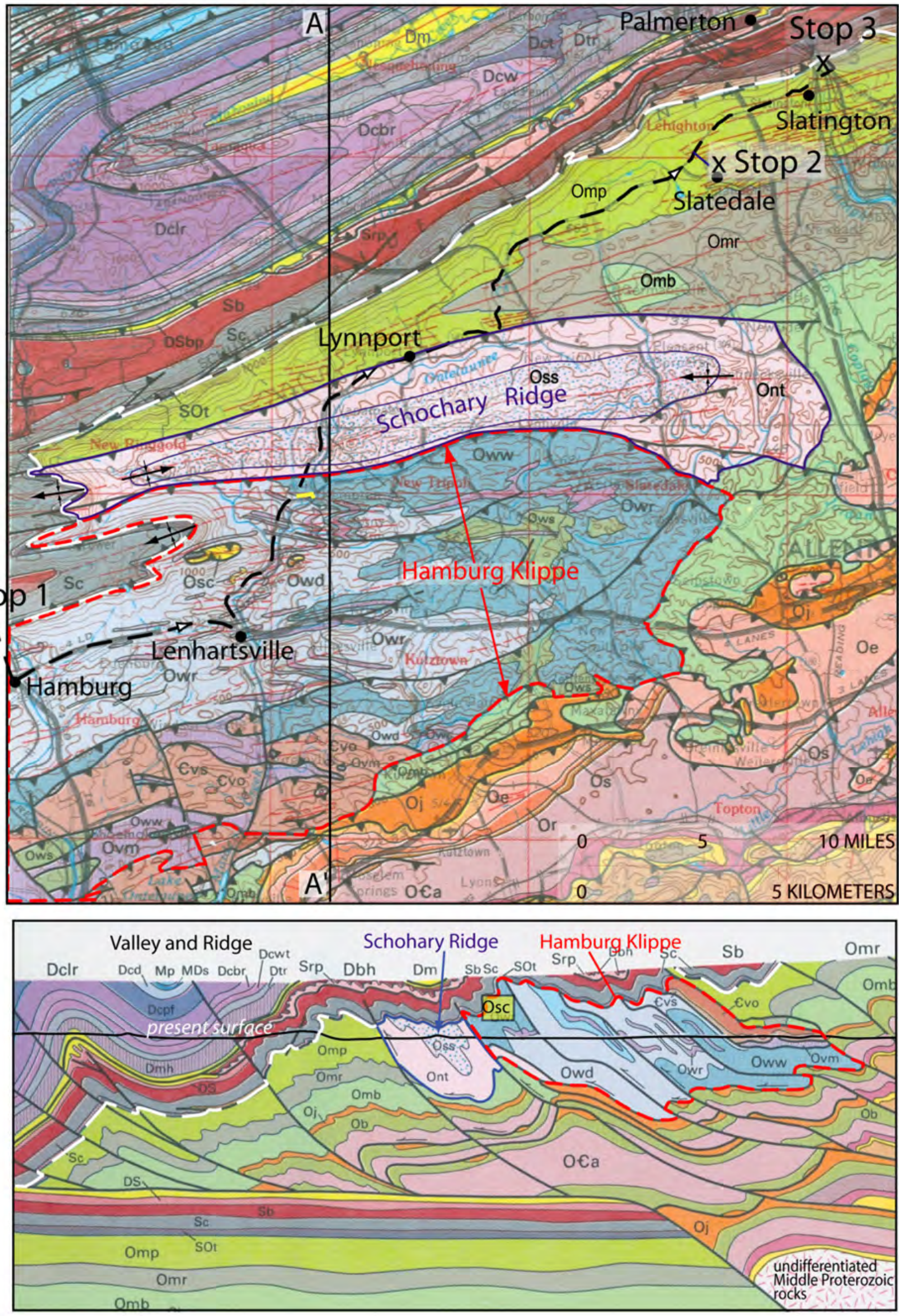


Figure 4. Geologic Map and cross section showing travel route (black dashed line) between Stops 1 and 3 and tectonic relations of the Hamburg klippe (red dashed line), Schochary sequence (purple dashed line) , and position of Osc. The white dashed line is the angular unconformity between Silurian and Ordovician rocks. Standard structure symbol shows direction of selected synclinal plunge. Significant formation symbols are: Dm—Marcellus Shale; Dsbp—Middle Devonian through Upper Silurian rocks, undivided; Sb—Bloomsburg Red Beds; Sc—Clinton Formation; Sot—Tuscarora Sandstone; Osc—Spitzenberg and Sharps Mountain; Oss—Schohary Sandstone; Ont—New Tripoli Formation; Owr—Windsor Township Formation; Omp—Pen Argyl member of the Martisnburg Formation; Omr—Ramsyeburg Member of the Martinsburg Formation; Omb—Bushkill Member of the Martinsburg Formation. Modified after Lyttle and Epstein, 1987.

throughout the section and the unit can no longer be mapped in the middle of the state. Thus, a unified member-less Shawangunk underlies red beds (Bloomsburg) northeastward. Earlier New Jersey State geologic maps (e.g., Lewis and Kummel, 1910-1912) identified the red beds above the Shawangunk as High Falls Shale, a name derived from the type section of that formation at High Falls, NY, about 40 miles northeast of the state border. However, as shown by Epstein (1993) the Shawangunk thins northeastward, two of its sandstone/conglomerate bodies intertongue with two red bed tongues of the Bloomsburg, all of which can be traced across New Jersey without complication to southeastern New York. These tongues terminate and do not intersect the rocks at High Falls, NY. These relations required the use of the name Bloomsburg on the most recent State geologic map (Drake and others, 1996). Some of the rocks of the Shawangunk tongues contain polymictic conglomerates, some of which are similar to those in the Green Pond Conglomerate of the Green Pond outlier, about 15 miles to the southeast in New Jersey. One can easily envision a stratigraphic section between the main outcrop belt and the outlier that shows the Shawangunk tongue becoming thicker and encompassing more of the lower part of the section going eastward. In the outlier most of the Green Pond Conglomerate would be included in the tongue.

Structural Geology

The Provinces

The area of the 77th Annual Field Conference of Pennsylvania Geologists will pass through several stratigraphic packages of differing structural style, summarized in Appendix 3. Figure 4 is a geologic map and section of the western section of the trip, seen during day 1.

The unconformable Ordovician-Silurian contact represents a hiatus of at least 10 m.y. and could be much greater depending on the age of the Beemerville syenite body (see Stop 7) and whether it is agreed that the Shawangunk lies nonconformably on the syenite. The divergence in dip at the unconformity is less than 15^{o} with the Martinsburg Formation east of the Schuylkill River through eastern Pennsylvania, northern New Jersey, and into southeastern New York (Epstein and Lyttle, 1986, 1987). To the west, a more pronounced angular unconformity exists between the same Silurian rocks and the complexly deformed Middle and Lower Ordovician rocks of the far-travelled Hamburg klippe.

In general, going from Pennsylvania to New York, structures become simpler (Figure 4), from highly faulted and folded along the Schuylkill River (Stop 1, Figure 1), where the Tuscarora Formation rests on both the Martinsburg Formation and rocks of the Hamburg klippe, to overturned and faulted rocks at Lehigh Gap (Stop 3, Figure 6), to oversteepened folds at Delaware Water Gap (Stop 4, Figure 4), and upright to slightly overturned folds at High Point, New Jersey (Stop 8, Figure 7), and finally into a fairly simple arch at Ellenville, New York (Stop 10, Figure 4). Slaty cleavage in both Ordovician and younger rocks is common, particularly in the southwestern part of the study area. A second-generation crenulation cleavage is also found in all rocks, more so in the southwest and generally absent in the northeast. Timing and degree of deformation of these rocks has been the subject of considerable long-standing debate. The three most important issues 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 (and what is the age of the folds,

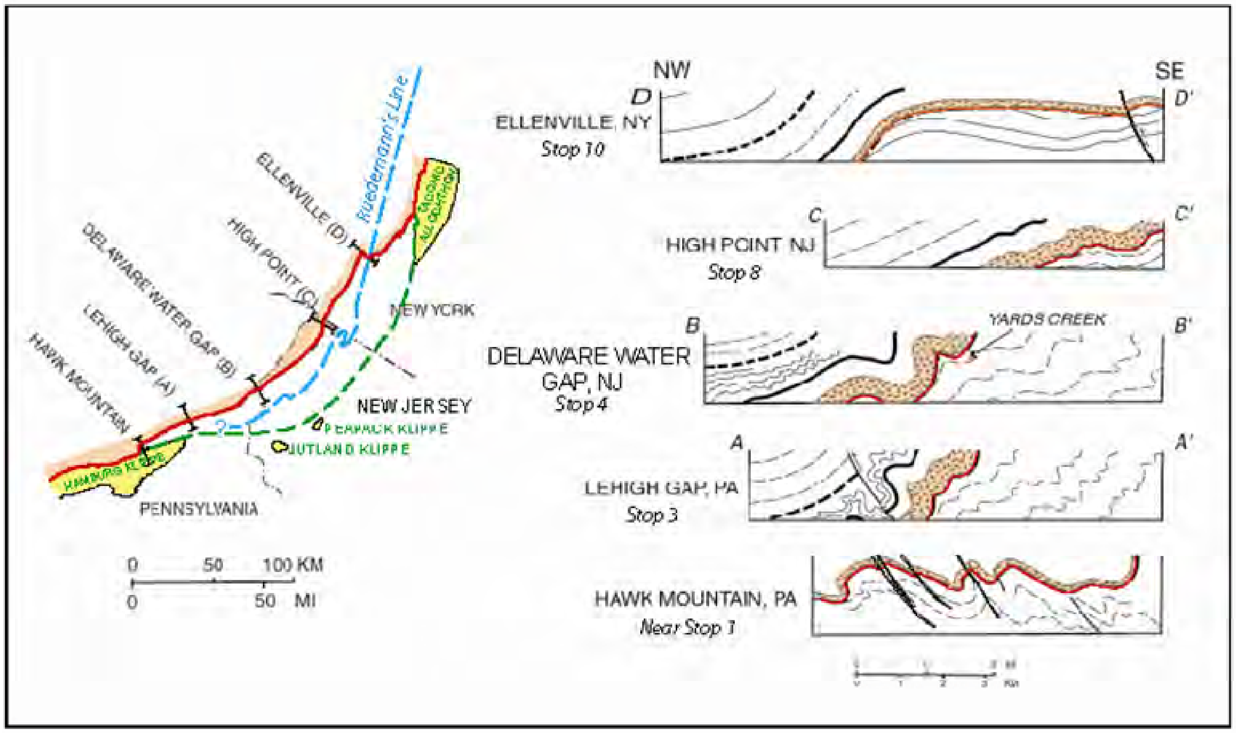


Figure 5. Generalized tectonic map and cross sections along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and New York. Red line is the Taconic unconformity separating Silurian clastic rocks from the underlying Ordovician rocks. Solid heavy line separates the lithotectonic unit containing relatively thin Upper Silurian limestones, shales, and sandstones from the thicker Bloomsburg, Clinton, Tuscarora, and Shawangunk rocks below. The heavy dashed line separates those rocks from the Devonian sandstones and shales of the moderately dipping to flat lying rocks of the Pocono and Catskill Plateaus. Ruedemann's line, in blue, separates broad open Taconic folds to the north and west from more intense structures to the south and east. Orange-shaded, dotted unit in the cross sections is the Tuscarora Sandstone – Shawangunk Formation. Modified from Epstein and Lyttle, 1987.

faults, and cleavage in these rocks); and 3) is the post-Taconic deformation Acadian or Alleghanian, or both?

On this field trip we will suggest that: 1) with only a few exceptions, the Shawangunk and equivalent Tuscarora Formation overlie the Martinsburg Formation with an angular unconformity that ranges between an angle that is barely discernible, to about 15^{o}; 2) the dominant regional folding in all rocks along the contact is Alleghanian in age, 3) the regional slaty cleavage in the Martinsburg formation, and possibly in the Hamburg klippe in the trip area is Alleghanian in age, 4) Taconic folds in the Martinsburg Formation below the unconformity are mostly broad and open along the entire 120 mi (193 km) length of the contact that we have studied; 5) southeast of Ellenville, New Yor, the structures in the Martinsburg become more intense and the angular disparity between beds above and below the unconformity is greater; and 6) the strike of Taconic structures trend a bit more northerly (by about 3-20^{o}) than later structures in the Ellenville area, and possibly elsewhere in the trip area.

The Taconic Unconformity: A Short Geologic History

Precambrian rocks more than one billion years old underlie parts of Pennsylvania. Following the breakup of the early Precambrian continent, Rondinia, about 725 million years ago, a shallow-water carbonate bank existed along the east coast of the North American continent during the Cambrian to mid-Ordovician. A volcanic arc in the middle of the ocean (*Iapetus* and its friends) subsequently collided with North America (Taconic Orogeny) breaking up the bank, forming highlands to the east which shed thick muds and sands into forearc basins. Uplift of the shales and turbidite sandstones (Martinsburg), beveling by erosion, and deposition of coarse clastics Shawangunk/Tuscarora) from the ancient Taconic Mountains during the Silurian Period, resulted in the Silurian-Ordovician unconformity. More intense hinterland Taconic deformation resulted in significant thrusting and formation of the Hamburg klippe, nappes, and all the fine stuff visited on the 2010 Field Conference of Pennsylvania Geologists (Wise and Fleeger, 2010; and in the References). The age of the rocks missing along the Taconic unconformity increases from central Pennsylvania (where Ordovician rocks appear to be transitional into Silurian rocks) through eastern Pennsylvania and New Jersey, into New York (Rodgers, 1971, for example; see Figure 2). Continued tectonic movement, not of concern here, resulted in deposition of all manner of sedimentary rocks during the remainder of the 200 million years of the Paleozoic, culminating in the Alleghany orogeny near the end of the era (of concern here), with the docking of the African plate with North American, and the birth of the super-continent *Pangea*. Pangea stayed intact for more than 50 million years, and near the end of the Permian, did some nasty things that killed off the largest number of species ever. But its days were numbered beginning in the late Triassic, when the docked African and North American blocks started their rifting separation during seafloor spreading, which is still going on today. Glaciers invaded eastern United States several times during the Pleistocene, upon which Ron Witte will elucidate at several stops. The geomorphic evolution of the Appalachians following the termination of mountain building and Mesozoic rifting has encouraged many differing thoughts on such subjects as peneplains, superposition, antecedent streams, and headward erosion. The mountains have also produced some of the finest breweries in the world. ENJOY!

Historical Perspective

The contact between the Martinsburg Formation of Ordovician age and the Silurian Shawangunk Formation in eastern Pennsylvania, New Jersey, and southeastern New York has attracted the attention of geologists every since Rogers (1838) recognized that it was an unconformity and later proclaimed that the orogeny was the "...most omentous...revolution" in North America (Rogers, 1858, p. 785). White (1882) described the contact as unconformable at Lehigh Gap, Pa., and Otisville, N.Y., but Chance (<u>in</u> White, 1882) and Lesley (1883) maintained that the angular relations were due to faulting. Later, Clark (1921) and Keith (1923), among others, maintained that the angular unconformity seen between Ordovician and Silurian rocks to the northeast is not to be seen in Pennsylvania.

Miller (1926) disagreed. He believed that an angular unconformity is present in Pennsylvania and based his conclusions on the following reasons: 1) the disconformable relations seen in exposures; 2) sericitized slate pebbles, apparently derived from the underlying Martinsburg, in the basal beds of the Shawangunk Formation; 3) omission of beds along strike; 4) the Martinsburg Formation is more highly metamorphosed than Devonian shales a few miles (kilometers) away; 5) structures in Ordovician and Cambrian rocks are more complex than those in Devonian and Silurian rocks; and 6) the cleavage in the Martinsburg, which was formed during the Taconic orogeny, is itself deformed into folds and was faulted during the Appalachian orogeny.

Behre (1924, 1933) argued that the Taconic orogeny produced slaty cleavage, close overturned folds, and thrust faults and was more intense than later Appalachian deformation which merely distorted the slaty cleavage. He (Behre, 1927, 1933) divided the Martinsburg into three members, a lower and upper slate separated by a sandstone unit. Stose (1930), however, maintained that the upper slate member of Behre is the lower member repeated by folding; hence, the Taconic orogeny must have been intense, for the Shawangunk Formation rests on the lower member of the Martinsburg. Stose's interpretation has had a profound influence on understanding the stratigraphy of the Martinsburg, as well as interpreting the intensity of Taconic deformation in the Martinsburg. Detailed mapping of the Martinsburg in the critical areas in the Delaware Valley clearly showed that it contained three mappable units rather than two (Davis et al., 1967; Epstein, 1973, 1990), which Drake and Epstein (1967) reestablished the threefold subdivision of the Martinsburg as proposed by Behre (1933), named the three members, and concluded that Stose's interpretations were wrong.

Willard and Cleaves (1939) showed that the angular unconformity extends as far southwest as Susquehanna Gap in Pennsylvania, where the Bald Eagle Conglomerate rests conformably on top of the Martinsburg Formation. Willard (1938) previously presented a cross section of the unconformity at Delaware Water Gap, but his interpretation does not agree with the interpretation presented in at stop 4.

Hess (1955) believed that the Taconic orogeny was so intense that it was not only the cause of folding of the sediments in the Appalachian "geosyncline", but rather the cause of the geosyncline itself. Woodward (1957) maintained that the slate belt of the Martinsburg is the result of the superposition of three periods of folding (Taconic, Acadian, Alleghanian), each having a different trend. However, no field evidence is recognized in Pennsylvania and New Jersey to support Woodward's Views.

Detailed mapping in the Delaware Valley by different geologists mapping on either side of the unconformity have led to different views on the intensity of Taconic and Alleghanian deformation. Drake et al. (1960), working in rocks older than Silurian, led to the interpretation that the Taconic orogeny was more severe than the Alleghanian orogeny in that area, resulting in regional fold nappes, and that the regional slaty cleavage in the Martinsburg is Taconic in age. At the same time, Arndt and Wood (1960), working in rocks generally younger than Ordovician, concluded that the Appalachian orogeny was by far the stronger. Additionally, Wood et al. (1963, p. 78) suggested that the discordant contact between the Martinsburg and Shawangunk might be largely the result of faulting. Lowry (1957) suggested that the regional cleavage in all rocks in eastern Pennsylvania formed at the same time—post-Ordovician.

Maxwell (1962) concluded that the flow cleavage in the Martinsburg Formation in the Delaware Water Gap area was produced by relatively minor deformation during the Taconic orogeny, believing that the regional slaty cleavage was the product of diagenesis, rather than metamorphism, and that the slaty cleavage does not migrate up into Silurian and younger rocks. He maintained that a later fracture cleavage cuts all rocks across the Taconic unconformity. His comments about cleavage is discussed at Stop 4.

The tacit acceptance of two orogenies led Broughton (1946) to believe that the slaty cleavage in the Martinsburg of New Jersey was a product of Taconic tectonism and that a cross-cutting slip cleavage was formed during the later Appalachian orogeny, although he suggested that "These structures might well be explained as the result of two peaks in the stress cycle of one period..." (Broughton, 1946**, p. 17**). However, both slaty cleavage and slip cleavage are found in all rocks in the Delaware Water Gap area, they are believed to have formed during the same continuing period of deformation (Epstein and Epstein, 1969). This same interpretation was accepted in the Harrisburg area by Root (1970). The idea that the Martinsburg was severely cleaved during the Taconic orogeny led Stevens (1966) to conclude that it was Taconic metamorphism that produced anthraxolite in the Martins-burg, although he noted (Stevens, 1966, p. 111) that, "It is a curious coincidence that the westward diminution of Taconic metamorphism in the Martinsburg conforms to the pattern of later metamorphism for the Pennsylvanian coals."

The interpretation that a large nappe underlies the Great Valley in easternmost Pennsylvania (Drake et al., 1961; Drake, 1967a, 1967b; Drake and Epstein,, 1967; Davis et al., 1967) with no structural counterpart in rocks younger than Ordovician argued for an intense Taconic orogeny. Nappes have been mapped in other parts of the Great Valley (Stose, 1950; Gray, 1954; Field Conference of Pennsylvania Geologists., 1966). While intense Taconic deformation is well documented in central Pennsylvania, such as in the Harrisburg area, the potential overriding effects of Alleghanian deformation has been recognized (Wise and Fleeger, 2010). Bird and Dewey (1970) explained the Taconic nappes in Pennsylvania in their plate tectonic model for evolution of the Appalachian orogen, although they were not certain to what extent Alleghenian deformation affected pre-Silurian rocks.

Based on detailed field mapping between Lehigh and Delaware Water Gaps, Epstein and Epstein (1967;1969; Epstein et al., 1974) concluded that the folding in Ordovician and Silurian and younger rocks in that area are dominantly Alleghanian in age. Rodgers (1971, p. 1164), on the other hand, based on the interpretation of regional nappes of Ordovician age with their associated cleavage, and which are cut by later (Alleghanian) cleavage, believed that the dominant structure in the Delaware Valley is pre-Silurian in age. Interpretation of regional nappes has changed over time. The present geologic map of New Jersey (Drake et al., 1996) interprets the structure of the Great Valley in that state as one of imbricate foreland thrusting; no regional nappes are shown. Finally, an Alleghanian age for the regional slaty cleavage is supported by 40Ar/39Ar whole-rock analysis from the Martinsburg Formation at Lehigh Gap (Wintsch et al.., 1996).

Structural Conclusions

In this report, I and my co-leaders conclude the following about the Ordovician-Silurian boundary issue in the field-trip area:

1. There is an angular unconformity between the Martinsburg and Shawangunk and Windsor Township formation and Tuscarora.
2. The Taconic orogeny was a period of mountain building, indicated by the thick coarse Silurian clastic wedge overlying the Martinsburg.
3. 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.
4. Folds of proven Taconic age in the Martinsburg Formation below the unconformity are mostly broad and open along the entire 120 mi (193 km) length of the contact between the area east of the Hamburg klippe in Pennsylvania to Ellenville, NY. Southeast of this zone of gentle folding the structures become tighter and faulting increases.
5. The regional slaty cleavage formed after the rocks were indurated at, or just below, conditions of low-grade metamorphism.

6. Arching of cleavage in the Martinsburg Formation as the contact with Silurian rocks is approached, which has been ascribed to folding of Taconic-age cleavage by later Alleghanian folding (Drake et al., 1960; Maxwell, 1962), is attributed to a pressure-shadow mechanism, to be discussed at Stop 3, 4, and 5, possibly 7.
7. 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 (see Day 1 Road Log, Figure 19). No unoriented slate fragments have been found in basal Shawangunk/Tuscarora beds. Argillite intraclasts in the basal Shawangunk, especially at Lehigh Gap, have cleavage that is parallel to one of two cleavages present in the beds above.
8. In the area east of the Schuylkill River Taconic deformation immediately adjacent to the unconformity is limited to gentle folding; Alleghanian deformation is much more intense.
9. The contact at Schuylkill Gap presents a puzzlement – rotating the near-vertical Silurian rocks there back to horizontal makes for a weird picture. *More field mapping is needed here!* Please see the section "The Role Of Geologic Mapping: A Call For Future Mappers" below.

The Hamburg Triangle: Figure 6

After we leave Stop 1 of the field trip, we will travel up the valley of Maiden Creek with the Tuscarora Sandstone holding up Blue Mountain on our left, passing thorough the Hamburg klippe, Sharps Mountain on our left and Spitzenburg on our right, then passing through rocks of the Shochary sequence and onto the fault-separated Martinsburg Formation. All this will happen over 11 mi (18 km) of travel. Many of the geologic features are a source of wonderment and have been discussed in geologic literature, not to be summarized here. Here they are. Few conclusions will be drawn. They are offered for future investigatory enlightenment.

1. Typical "Appalachian" folds, Pennsylvania style, lie to the east. The repetition of rock units over and over again as the folds are traversed to the north, indicate a shallow level of folding, before the rocks plunge deeper into the Alleghany Plateau. These folds don't last too long towards the east as they merge into the Pocono Plateau. Then there is the peculiar banana-shaped Lackawanna Syncline. And the thermal maturity line that separates Marcellus gas production for the "dead zone" to the south. Decollements at depth? Effects of Salina salt? Just asking.
2. Why does the Hamburg klippe end wher6e it does just as the folds in the Tuscarora skoot to the north east of the city of Hamburg?
3. What the heck is the Shochary syncline? A deep tight fairly uncomplicated structure in the midst of rocks with tight folds of all scales? How do the Shochary sequence rocks relate to those in the Hamburg klippe, to the Martinsburg Formation to the east, and to correlative rocks south of Harrisburg? Note that the axis of the syncline in those Ordovician rocks appears to pass into a syncline in Blue Mountain in the Tuscarora to the west (see Faill, 2011). *But*. . . whereas the fold pattern in the Tuscarora in Blue Mountain shows the folds there plunge to the southwest, the Shochary syncline, based on the contact relations between the Schocharie sandstone and New Tripoli Formation clearly show the syncline to plunge to the east Lyttle and Epstein, 1987). Just asking.
4. Sharps Mountain is weird. It is held up by a thin sliver of Tuscarora Sandstone. Its

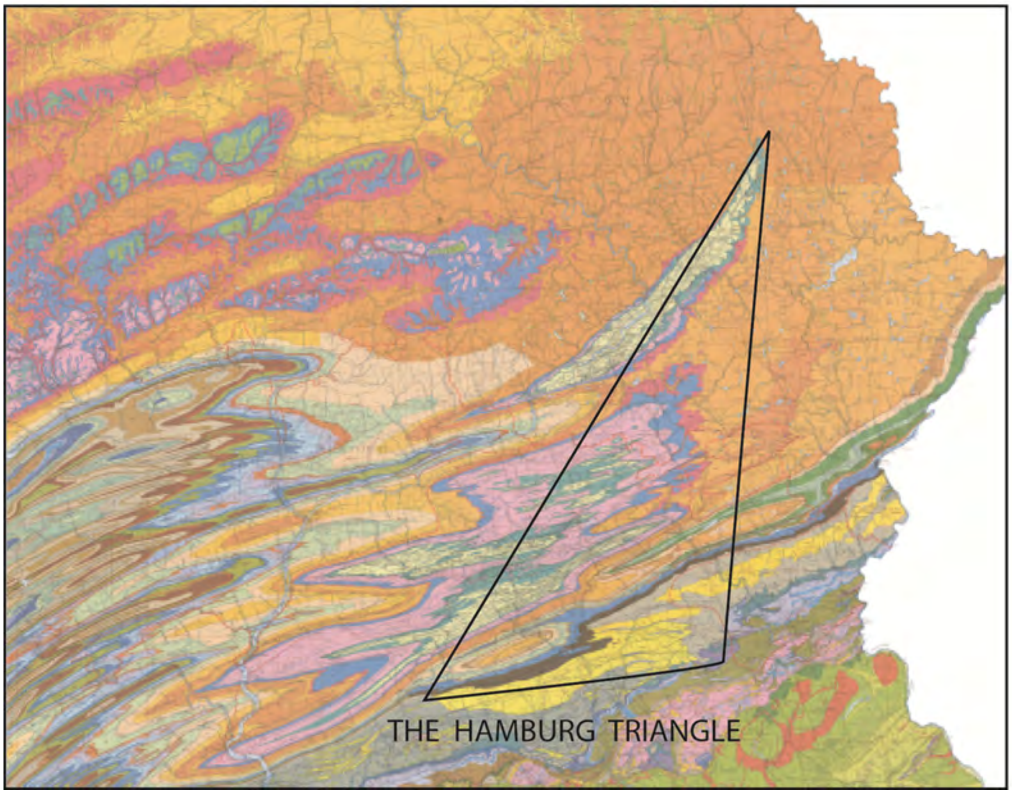


Figure 6. The Hamburg Triangle. The geology of eastern Pennsylvania changes relatively abruptly, almost along the straight red line, from the typical Pennsylvania/ Appalachian folds on the west to the Pocono plateau to the east, and where the Martinsburg slate belt disappears under the Silurian and the Hamburg klippe and Schochary Ridge terminate.

moderate northwest dip does not match the projected dips from the main outcrop belt a mi (1.6 km) to the west (see Lash, 1987). A trip a bunch of years ago suggested to me that the pile of Tuscarora there is a landslide.

5. Spitzenburg Conglomerate. Two small erosional outliers, containing unique crossbedded conglomaerates and sandstone with limestone cobbles, apparently derived from the Hamburg klippe and Great Valley rocks. They were discussed in a previous Field Conference (Lash et al., 1984), and concluded to have lay unconformably beneath the Tuscarora. *But* the same rocks are not seen at Sharps Mountain (graywackes are present there under the Tuscarora overburden, nor are similar conglomerates seen under the Tuscarora in the main outcrop belt immediately to the west. What gives? It is a bit of a stretch to correct these conglomerates with the Juniata or Bald Eagle many tens of miles (kilometers) to the west.

The Role Of Geologic Mapping: A Call For Future Mappers

Much of the information presented in this guidebook is the result of detailed field mapping by the co-leaders. Much of the discussion about the Hamburg klippe and Martinsburg Formation is the result of efforts by USGS geologists Peter Lyttle (now coordinator of the USGS National Cooperatiove Geologic Mapping Program) and Gary Lash (now at SUNY in Fredonia, NY, and involved in studies of the Marcellus Shale). Without detailed mapping by

both state and federal geologists and only by walking and visiting all outcrops, tracing rock units, and recording structural data, could the structural and stratigraphic conclusions many of the conclusions presented here could not have been possible. The sparse data available for Stop 1 cries out for the arrival of young detailed geologic mappers to fill in our geologic knowledge between the Auburn quadrangle on the Schuylkill River, to the Indiantown Gap Quadrangle several quads to the west where the Pennsylvania Survey is presently mapping.
In 1990 the US Congress realized that only 20 percent of the U.S. has detailed geologic map coverage. In 1992 Congress passed National Cooperative Geologic Mapping Act and in 1995 added the EDMAP is added as a component of the Program. The NCGMP is authorized until 2019. See **http://pubs.usgs.gov/fs/2010/3088/support/fs2010-3088.pdf** (accessed 9/11/2012)

The EDMAP program was established for university students to gain experience and knowledge in geologic mapping whilecontributing to national efforts to map the geology of the United States. It is a matching-funds grant program with universities. Geology professors whose specialty is geologic mapping request EDMAP funding to support upper-level undergraduate and graduate students at their colleges or universities in a 1-year mentor-guided geologic mapping project that focuses on a specifi c geographic area. Every Federal dollar that is awarded is matched with university funds. EDMAP is invaluable not only because it contributes to national geologic mapping efforts but also because it helps fund academic research, thoroughly prepares students for realworld careers in the geoscience field, and gives participants a competitive edge in the job market:

- Students participating in the EDMAP Program receive training and first-hand field experience in geologic mapping and thus acquire skills useful in many geoscience fields.
- EDMAP geology professors and their students frequently work closely with nearby State geological surveys and U.S. Geological Survey geologists.
- Student work contributes to geologic mapping of the United States.
- So far, EDMAP has benefi ted 144 universities and more than 850 students from geoscience departments across the Nation.
- Results from a recent EDMAP participant survey show that 95 percent of the respondents either went on to take jobs in the geoscience fi eld or pursued further degrees in geosciences.
- Surveyed participants considered the program to have been a great opportunity and one that was enjoyable and highly valuable to their career.

How To Apply

Grants are awarded through an annual, competitive, and matched grant process. Per-project funding that was available in fiscal year 2011, for example, for graduate projects is $17,500 and for undergraduate projects is $10,000. A peer-review panel consisting of university faculty, State Geologists, and USGS representatives determines awards.

Applications for professors and students interested in participating in this program are solicited via the online EDMAP Program Announcement, where you can find detailed instructions for submitting EDMAP proposals (U.S. Geological Survey, 2011). Visit **http://www.grants.gov** and click on "Find Grant".

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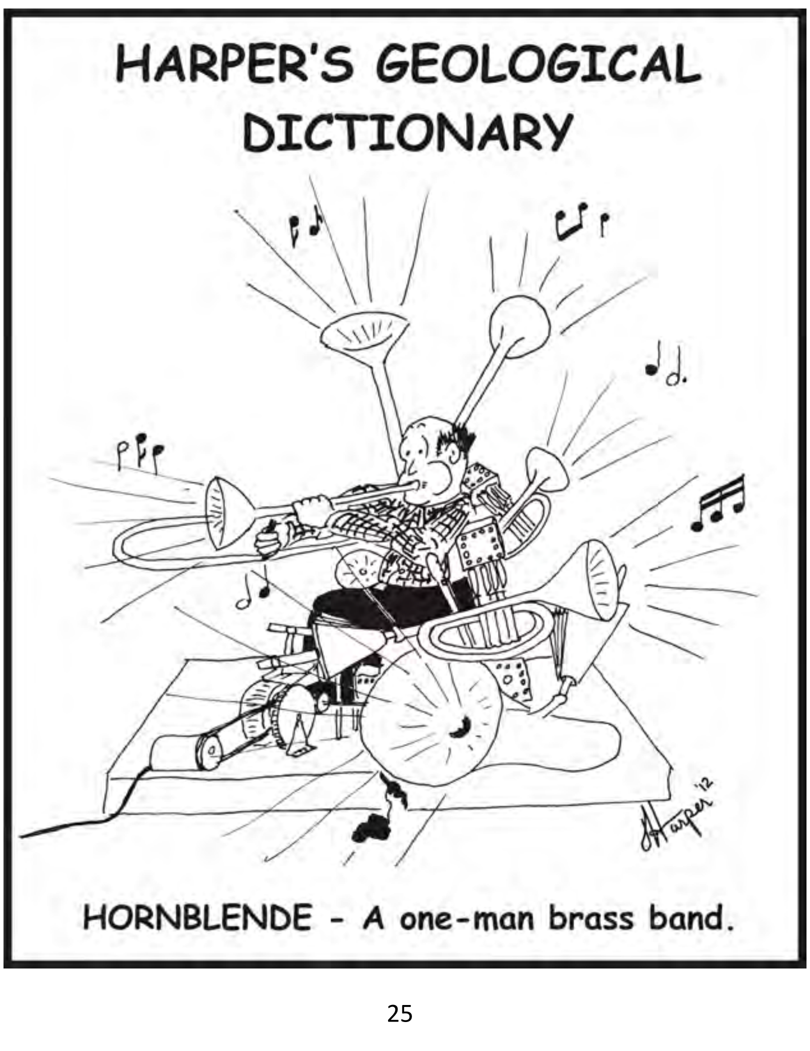

APPENDIX 1: Generalized stratigraphy of the rock units that we may encounter in the area of the first day of the field trip or are shown on figure 1 in Pennsylvania and New Jersey. Modified from Epstein and Lyttle (1987).

MAHANTANGO FORMATION (Middle Devonian, 1,200-2,500 ft [366-762 m] thick): Medium-dark-to dark-gray, poorly bedded and laminated to thin-bedded, bioturbated, shaly siltstone and silty shale. Fossils are diverse and scarce to abundant in biostromes different parts of the formation.

MARCELLUS SHALE (Middle Devonian, about 250-1,000 ft [76-305 m] thick): Medium-dark-gray to grayish-black, laminated to poorly bedded, sparingly fossiliferous shale and silty shale. Lower part, 70-200 ft (21-61 m) thick, consists of laminated to thin-bedded, calcareous, shaly siltstone and argillaceous limestone (Stony Hollow Member) above, and medium-dark-gray to grayish-black shale (Union Springs Member) below. The basal rocks of the Marcellus is deformed wherever exposed.

ROCKS OF ONESQUETHAW AGE (Middle to Lower Devonian, <50-about 700 ft [<15 - 213 m] thick): **Onondaga Limestone**—gray, fossiliferous, cherty limestone and argillaceous limestone; deeply leached in western exposures grades to the southwest into the **Palmerton Sandstone**—medium- to very coarse grained, generally massive sandstone and conglomeratic sandstone with quartz pebbles as much as 0.75 in (2 cm) long; **Schoharie Formation**—laminated to very thick bedded, slightly cherty, fossiliferous, calcareous siltstone with Taonurus and other burrows; **Esopus Formation**—Medium- to dark-gray, well-cleaved, generally laminated to poorly bedded siltstone with abundant *Taonurus.*

ORISKANY GROUP THROUGH HELDERBERG GROUP (Lower Devonian, <50-~400 ft [<15–~120 m] thick): **Ridgeley Sandstone,** Oriskany Group—fine- to very coarse grained, fossiliferous (spiriferid brachiopods), calcareous sandstone and conglomerate with quartz pebbles as much as 0.75 in (2 cm) long, with minor siltstone, arenaceous limestone, and chert; **Shriver Chert**, Oriskany Group—calcareous, fossiliferous shale and siltstone and minor calcareous sandstone; **Port Ewen Shale**, Helderberg Group—fossiliferous, calcareous shale and siltstone; **Minisink Limestone**, Helderberg Group—fossiliferous, argillaceous, limestone and calcareous shale; **New Scotland Formation**, Helderberg Group-- cherty, fossiliferous, calcareous and silty shale and argillaceous limestone, deeply leached in western exposures; **Coeymans Formation**, Helderberg Group—argillaceous and arenaceous, partly cherty, fossiliferous, burrowed, partly biohermal limestone, and medium-gray, fine- to coarse-grained and pebbly, fossiliferous, crossbedded and planar bedded, calcareous sandstone and conglomerate with quartz pebbles as much as 1 in(2.5 cm) long;

UPPER SILURIAN ROCKS (about 50-1,000 ft [15-300 m] thick): **Rondout Formation**—fossiliferous, laminated to thin-bedded limestone; mudcracked, calcareous shale; and dolomite; **Decker Formation**—calcareous, crossbedded to planar-bedded to flaser-bedded, fossiliferous, burrowed partly conglomeratic sandstone, calcareous siltstone and shale, arenaceous limestone, and mudcracked dolomite; **Andreas Red Beds**—grayish-red and light- to greenish-gray, fine- to coarse-grained and conglomeratic sandstone and burrowed shale and siltstone. May correlate with either the Coeymans Formation or upper part of the Decker Formation to the east; **Bossardville Limestone**—laminated to thin-bedded, fossiliferous, argillaceous, and partly dolomitic limestone; and calcareous shale with deep mudcrack polygons; **Poxono Island Formation**—Green to gray, partly mudcracked, laminated to thick-bedded, calcareous shale; very fine to fine-grained dolomite; and argillaceous limestone with dessication breccia locally at the top and pale-red, interbedded shale near the bottom. Grades into the High Falls Shale in New York (see Figure 3 on p. 5**).**

BLOOMSBURG RED BEDS (Upper Silurian, 800-1,800 ft [244-549 m] thick): Red, crossbedded and planar-bedded, laminated to thick-bedded shale, siltstone, very fine to coarse-grained sandstone, and minor conglomeratic sandstone with cut-and-fill structures, mudcracks, in fining-upward cycles in the east. Lower contact gradational, very irregular in places, and placed at base of lowest red bed above gray rocks of the Shawangunk or Clinton Formation (see mileage 142.9 of Day 1 Road Log).

SHAWANGUNK FORMATION (Middle Silurian, 1,600 ft [488 m] in the west, thins to about 400 in the east end of the field trip area): Comprises four members in western New Jersey and eastern Pennsylvania, from youngest to oldest: **Tammany Member** (0-800 ft [0-244 m] thick**—**Medium- to medium-dark-gray, planar-bedded and crossbedded, thin- to thick-bedded, fine- to coarse-grained and conglomeratic quartzite with quartz and argillite pebbles as much as 2 in(5 cm) long, and minor beds of dark-gray argillite. Lower contact gradational into the **Lizard Creek Member (**0-1,400 ft [0-427 m] thick)**—**Light- to dark-gray and light-olive- to dark-greenish-gray, very fine to coarse-grained, laminated to very thick bedded, planar-bedded, crossbedded, flaser-bedded, rippled, partly channeled, burrowed quartzite with flattened argillite cobbles as much as 4 in(10 cm) in diameter; minor dark-grayish-red-purple, fine-grained, burrowed, hematitic quartzite, interbedded with medium-light- to dark-gray and olive-to greenish-gray, laminated to thin-bedded, flaser-bedded, burrowed siltstone and shale; scattered thin beds contain collophane, siderite, and chlorite nodules, quartz pebbles as much as 0.25 in (0.6 cm) long, and *Lingula* and eurypterid fragments. Lower contact transitional and placed at base of lowest argillite in sequence containing abundant argillite above quartzites of underlying member; **Minsi Member** (100-350 ft [31-107 m] thick)**—**Very light- to dark-gray and light-olive- to greenish-gray, partly burrowed, planar-bedded and crossbedded, thin- to thick-bedded, fine- to coarse-grained, partly conglomeratic quartzite with pebbles of quartz and chert not more than 2 in(5 cm) in diameter and cobbles of shale as much as 7 in(18 cm) in diameter; minor medium-dark- to dark-gray and greenish-gray, laminated to thin-bedded, locally mudcracked siltstone and shale. Local *Arthrophycus*. Lower contact placed at top of uppermost bed of the Weiders Member containing quartz pebbles more than 2 in (5 cm) long in sharp unconformable contact with the Martinsburg Formation. 200-350 ft(61-107 m) thick; **Weiders Member** (0-220 ft [0-67 m] thick)**—**Medium-light- to medium-dark-gray and greenish-gray, planar-bedded and crossbedded, thin- to thick-bedded, medium- to very coarse grained quartzite; conglomerate with quartz and chert cobbles as much as 6.5 in(17 cm) in diameter and shale cobbles as much as 8 in(20 cm) in diameter; and rare, greenish-gray, laminated to thin-bedded argillite. Lower contact with Martinsburg Formation unconformable.0-220 ft(0-67 m) thick. The Minsi and Weiders Members merge into the Tuscarora Sandstone 4.5 mi (7.2 km) west of Lehigh Gap as the Lizard Creek Member merges into the Clinton Formation as the Tammany Member pinches out.

CLINTON FORMATION (Middle Silurian; 1,400 ft [426 m] thick): Gray, green, and red sandstone, siltstone and shale laterally continuous with and lithically similar to the Lizard Creek Member of the Shawangunk Formation, except that it contains more red beds.

TUSCARORA SANDSTONE (Lower or Middle Silurian; about 100 ft [31 m] thick at Schuylkill Gap): Gray quartzite laterally continuous with and similar to the Minsi Member of the Shawangunk Formation.

Many rocks of Paleozoic age are present in the Green Pond syncline in New Jersey and New York. It is a narrow, northeast-trending, faulted syncline containing a thin, but fairly complete section of Paleozoic sediments, that sit unconformably on the Proterozoic basement. Many of the Paleozoic rocks correlate with thicker units in the Valley and Ridge Province showing that these rocks once were present between the two areas, a distance of more than 15 mi (25 km). Units of interest include:

HIGH FALLS FORMATION (probably the Bloomsburg Red Beds; Upper Silurian; 300 ft [91 m] thick): Grayish-red, nonfossiliferous, silty shale with thin sandstone common in lower half.

GREEN POND CONGLOMERATE (Upper Silurian; 1,000-1,400 ft [304-427 m] thick): Gray and reddish-gray sandstone and conglomerate with predominantly white quartz and minor gray, green, red, and yellow chert, red shale, and red sandstone cobbles as much as 3 in(7.6 cm) long. Lower contact unconformable (see article by Herman in this guidebook).Also termed Shawangunk Conglomerate.

MARTINSBURG FORMATION (Upper and Middle Ordovician; probably more than 15,000 ft [4,572 m] thick): In New York the identifications of correlative rocks have been given various names (see Figure 2 on p. 4 and discussion by Lyttle below). In Pennsylvania and most of New Jersey, consist of three distinctive members: **Pen Argyl Member** (Upper Ordovician, *Climacograptus spiniferus* Zone, 3,000-7,000 ft [914-2,134 m] thick)**—**Dark-gray to grayish-black, thin- to thick-bedded, evenly bedded slate, commonly more than 12 ft(3 m) and in places 20 ft(6 m) thick; rhythmically interlayered with carbonaceous slate, sandy slate, and very fine to medium-grained graywacke with parallel lamination, lenticular bedding, convolute bedding, and sole marks. Units in fining upward sequences (turbidite-flysch sequence).Quarried extensively for slate ("soft slate" of Pennsylvania quarrymen). Upper contact is unconformable and site of a regional decollement. Lower contact is gradational and is placed where graywacke is in excess of about 5 percent of local sequences and supplies abundant float. In Pennsylvania the Pen Argyl is overlapped by the Shawangunk Formation just west of the Delaware River and is absent in northern New Jersey; **Ramseyburg Member** (Upper and Middle Ordovician, *Orthograptus ruedemanni* to lower *Climacograptus spiniferus* Zone, 2,000-3,500 ft [610-1,067 m] thick)**—**Medium- to dark-gray slate that alternates, in part cyclically, with light- to medium-gray, thin- to very thick bedded graywacke and graywacke siltstone (turbidites). Graywacke comprises about 20 percent of member, but may make up more than 50 percent of some thick parts of the section, and less than 5 percent in others. Slates are generally thick bedded at the top and thin- to medium-bedded at the bottom of the member. Lower contact is placed at the base of lowest conspicuous graywacke bed, generally recognized by abundant float, but contact may be transitional through several hundred ft (scores of m), where discontinuous and lenticular graywacke beds are present in the underlying Bushkill Member); **Bushkill Member** (Middle Ordovician, C*limacograptus bicornis* to *Diplograptus multidens* Zone, 1,500-5,000 ft [457-1,524 m] thick)**—**Medium- to dark-gray, laminated to thin-bedded slate containing thin beds of quartzose slate, graywacke siltstone, and carbonaceous slate in fining upward sequences. Bed thickness does not exceed 6 in(15 cm) throughout member, and is generally less than 2 in(5 cm), except for graywacke beds that probably are less than 12 in(30 cm) thick in discontinuous units near the top of the member. Lower contact transitional through 3 ft(1 m). Formerly quarried for slate ("hard slate" belt of Pennsylvania quarrymen). In northernmost New Jersey and adjoining New York, rocks of slightly different composition than the upper part of the Ramseyburg are included in the **High Point Member** by Drake (1990), Upper and Middle Ordovician, 4,500 ft [1,372 m] thick)**—**thick-bedded greywacke and shale. These include rocks that were informally named Mamakating, above, and Pine Bush, below by Lyttle (*in* Epstein and Lyttle, 1987); **Shale And Graywacke At Mamakating—** Thick sequences of thin- to medium- bedded, medium dark gray shale interbedded with very thin to thick-bedded graywacke (as much as 6 ft [1.8 m] thick) alternating with thinner sequences of medium-bedded graywacke interbedded with less thin- to medium- bedded shale. Grades downward and laterally into the sandstone at Pine Bush; **Sandstone at Pine Bush—** Medium-grained, medium- to thick-bedded, medium-gray, speckled light-olive-gray- and light-olive-brown-weathering quartzitic sandstone interbedded with, and containing rip-ups of, thin- to medium-bedded, medium-dark-gray, greenish gray-weathering shale and fine-grained siltstone. Lower contact with Bushkill Member is interpreted to be conformable, but in many places it is marked by a thrust fault.

JACKSONBURG LIMESTONE (Middle Ordovician, 65-800 ft [20-244 m] thick):Dark-gray argillaceous limestone at top ("cement rock facies) grading down into gray, medium- to coarse-grained, calcarenite and high-calcium limestone. In New Jersey, the lower contact which is conformable in

Pennsylvania, is marked by beds of dolomite pebble- to boulder-conglomerate.

BEEKMANTOWN GROUP (Middle and Lower Ordovician): **Ontelaunee Formation—**Fine- to coarse-grained dolomite, cherty at the base, grading into dolomite that contains beds of medium-grained calcilutite at the top in some places. Generally absent in New Jersey Thickness ranges from a feather edge to 650 ft(200 m); **Epler Formation** (Lower Ordovician, 900 ft [274 m] thick)**—**Very fine grained to cryptogranular, limestone and dolomite. Upper part is absent in much of New Jersey; **Rickenbach Dolomite** (Lower Ordovician, 700 ft [213 m] thick)**—**Fine- to coarse-grained dololutite, dolarenite, and dolorudite. Upper part generally thin-bedded and laminated; lower part characteristically thick-bedded; **Stonehenge Limestone** (Lower Ordovician,)**—**Fine-grained, thin-bedded limestone marked by silty or sandy laminae with subordinate beds of orange to buff-weathering, irregularly laminated and mottled dolomite.

ALLENTOWN DOLOMITE (Lower Ordovician and Upper Cambrian, 575 ft [175 m] thick): Alternating light- and dark-gray weathering, rhythmically bedded dolomite; with abundant flat-pebble conglomerate, calcilutite, calcarenite, ooliticcalcarenite, calcirudite, algal stromatolite, dolomicrite, and scattered beds and lenses of orthoquartzite.

LEITHSVILLE FORMATION (Middle and Lower Cambrian, 1,000 ft [305 m] thick): Interbedded dolomite and calcitic dolomite, light-gray to tan phyllite, and very thin beds and stringers of quartz and dolomitic sandstone.

HARDYSTON QUARTZITE (Lower Cambrian, 0-900 ft [0-974 m] thick): Fine- to medium-grained, fine-to medium-grained, massive, but in some places thinly laminated and crossbedded, *Scolithus*-bearing, feldspathic quartzite interbedded with arkose, quartz-pebble conglomerate, and silty shale or phyllite.

SPITZENBERG AND SHARPS MOUNTAIN OUTLIERS: These two small erosional remnants in the Great Valley north of Hamburg, PA contain reworked sediments of both the Hamburg klippe and the Great Valley sequence. These rocks rest unconformably, and perhaps with structural discontinuity, on rocks of the Hamburg klippe and unconformably beneath the Tuscarora Sandstone. Although difficult to prove, these rocks may have been deposited after the rocks of the Hamburg klippe were folded during the Taconic orogeny and may represent the youngest Ordovician clastics in the area.

SPITZENBURG CONGLOMERATE (Late Ordovician, 0-200 ft [0-61 m] thick): At Spitzenberg (see Figure 2 on p. 4), the unit consists of medium- to coarse-grained, medium- to thick-bedded, poorly to well-sorted, crossbedded, red and green weathering, conglomeratic sandstone, interbedded with a thick-bedded, polymict conglomerate. The clasts include green chert, milky-white calcilutite and laminated to cross-laminated calcisiltite, red and maroon shale, brown sandstone and siltstone, and clasts of the same red sandstone that is interbedded with the conglomerate. Conodont biostratigraphy shows that the clasts are youngest at the bottom of the unit and oldest at the top. At Sharps Mountain, the sandstone weathers differently to a greenish-white and the conglomerate is absent. These rocks are unconformable, and possibly in thrust contact, with the underlying rocks of the Hamburg klippe. The underlying klippe rocks are the source for this unit. At Sharps Mountain, the unit is unconformably overlain by the Tuscarora Sandstone.

ROCKS OF THE SHOCHARY RIDGE AREA: This group of Ordovician clastic rocks crops out over a fairly small area in the Great Valley of eastern Pennsylvania (Figure 2 on p. 4). It is entirely fault bounded, and sits structurally on top of all three members of the Martinsburg Formation (Pen Argyl, Ramseyburg, and Bushkill), and structurally beneath the Greenwich slice of the Hamburg klippe. The eastern end of the Shochary Ridge outcrop belt coincides almost exactly with the eastern end of the

Hamburg klippe. It is very likely that the rocks of Shochary Ridge are derived from the Hamburg klippe and were transported in Taconic time as a thrust slice beneath the two major slices of the klippe. The two faults that mark the present limits of the Shochary Ridge area, are probably Alleghanian in age. The rocks of the Shochary Ridge area probably formed as a local, northward-prograding fan from the advancing accretionary prism of the Greenwich slice of the Hamburg klippe. These rocks are roughly the same age as the slightly deeper water clastics of the Martinsburg Formation, which were deposited by longitudinal (dominantly northwest or southeast) currents within a very long northeast-trending basin.

SHOCHARY SANDSTONE (Upper and Middle Ordovician, about 5,000 ft [1,524 m] thick): Medium dark-gray, thin- to thick-bedded calcareous, pyrite-rich, graywacke turbidites interbedded with light-olive-brown weathering slate, calcisiltite, and minor thin beds of conglomerate. Graywacke generally comprises 10 to 20 percent with rare instances of 50 percent, particularly near the top of the exposed section, and beds become thick-bedded with rare parallel laminations. Graywacke beds contain abundant faunal debris. Sedimentary structures in some turbidites obliterated in places by bioturbation. Rusty-weathering channels filled with coarse-grained sandstone, abundant shelly fauna, and pyrite are common in light-olive-brown-weathering graywackes. Graywacke beds also contain rare clasts of rounded chert. Upper beds of this unit are not present due to faulting and erosion. The Shochary is structurally and unconformably overlain by the Tuscarora Sandstone. Lower contact transitional over 50 ft(15 m) with underlying New Tripoli Formation and is placed where sandstones commonly make up less than 10 percent and beds are less than 2-4 in(5-10 cm) thick.

NEW TRIPOLI FORMATION (Middle Ordovician, more than 4,900 ft [1,494 m] thick): Medium-dark-gray, light-olive-brown weathering, thin, evenly bedded, calcareous graywacke interbedded with fairly thick slate and calcisiltite beds as much as 20 in [50 cm] thick) The calcisiltite beds are most commonly 1-2 in(2-5 cm) thick and are more resistant to weathering, giving the rock a ribbed appearance. This contrasts markedly with the ribbon slate of the Bushkill Member of the Martinsburg Formation. Shelly fossil debris is common, especially near the upper contact of the unit, but not abundant. Lower contact is faulted.

HAMBURG KLIPPE: This structurally and stratigraphically complex group of far-travelled rocks is located in the Pennsylvania Great Valley and resembles the Taconic allochthon of New York State. The klippe is divided into two tectonic slices: 1) the Greenwich slice, an accretionary prism of sediments displaying scaly cleavage that composes the Windsor Township Formation, the only unit that will be seen on this field trip, and 2) the Richmond slice, a sequence of rise and slope deposits that composes the Virginville Formation. Although emplaced by thrust faults during the Taconic orogeny, these faults have been obscured by later Alleghanian folds and faults. In general, rocks of the Lehigh and Lebanon (Great)Valley sequences have been thrust on top of the rocks of the Hamburg klippe during Alleghanian time. The Weisenberg Member of the Windsor Township Formation only will be see on this trip.

WINDSOR TOWNSHIP FORmation (Middle and Lower Ordovician, about 12,000 ft [3,658 m] thick): **Dreibelis Member—**medium- to very thick bedded, partly graded, fine- to coarse-grained, locally conglomeratic, somewhat calcareous graywacke sandstone interbedded with dark-greenish- to light-olive-gray, fissile to poorly cleaved mudstone, siltstone, and shale; **Switzer Creek Member—** medium- to thick-bedded, massive, medium- to coarse-grained to conglomeratic graywacke interbedded with lesser amounts of dark- greenish-gray mudstone and shale. Graywacke is rich in carbonate grains and pebbles, which weather to very distinctive, rotten and porous, limonite-stained rocks; **Weisenberg Member—**At least 5,700 ft (1,737 m) of light-olive-gray to grayish-olive, fissile to poorly cleaved shale and mudstone to micaceous siltstone with minor amounts of medium-dark- to dark-greenish-gray, silicified shale, mudstone, and argillite. In some places, thin-bedded siltstone and graywacke sandstone, and debris flows of chert and silicified mudstone are interbedded with the shale and mudstone. Soft-sediment slump folds are common. Local channels contain a very distinctive conglomerate with chalky-white-weathering feldspar grains and rare volcanic rock fragments. Polymicitc conglomerates contain clasts as much as 10 ft (3 m) long of graywacke, carbonate, siltstone, shale, and chert. Scattered throughout are red beds of locally silicified mudstone and argillite, interbedded with very thin to thick-bedded siltstone, sandstone, black calcilutite to calcarenite, and chert. Carbonate is found locally as clasts and megaclasts within a matrix of varicolored shale and argillite.

VIRGINVILLE FORMATION (Middle Ordovician to Upper Cambrian, about 1,800 ft [549 m] thick): **Moselem Member—**mudstone and shale interbedded with silicified argillite; minor thin-bedded, ribbon limestone, black shale, dolostone and minor carbonate-clast conglomerate; **Onyx Cave Member—**calcarenite, peloidal limestone, quartzose limestone, and calcareous quartzite, thick-bedded carbonate-clast conglomerate, and laminated black shale and dolostone; **Sacony Member—**thick-bedded, structureless, micaceous siltstone and sandstone interbedded with grayish- to pale-blue-green micaceous shale and mudstone.

JUTLAND KLIPPE (Ordovician and Cambrian, 600 ft [183 m] thick): These rocks resemble some of the rocks found within the Greenwich slice of the Hamburg klippe in Pennsylvania. Unlike the Taconic allochthon and Hamburg klippe, which occur on the foreland side of the external massifs of the Berkshires and Reading Prong, the rocks of the Jutland klippe occur on the hinterland side of the Reading Prong. Consist of interbedded red and green shale; silty sandstone; and micritic limestone. Upper contact is placed where sandstone, siltstone, and sparry and micritic carbonate rocks (found throughout the Jutland klippe, but more rarely in this lower unit) become much more common. Upper part contains manganese-bearing shale and interbedded red and green shale, dolomite, limestone conglomerate, laminated limestone and shale, fine-grained sandstone, siltstone and shale, and yellow, red, green, and gray shales interbedded with chert.

BEEMERVILLE SYENITE (Late Ordovician to Early Silurian): Medium to coarse-grained nepheline syenite intruded into the Martinsburg Formation and unconformably overlain by the Shawangunk Formation (see comments in Stop 7). Surrounded by numerous xenolith-bearing ouachitite breccias and lamprophyre, phonolite, bostonite, and malignite dikes and sills.

References

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Epstein, J.B. and Lyttle, P.T., 1987, Structure and stratigraphy above, below, and within the Taconic unconformity, southeastern New York, *in* Waines, R.H., ed., Field Trip Guidebook. 59th Annual Meeting, New York State Geological Association, Kingston, NY, p. C1-C78.

APPENDIX 2 – Generalized stratigraphy of the rock units that are present in the area of the second day of the field trip in New York, from Epstein and Lyttle (1987).

The Mount Marion Formation wll be seen along US 209 between Otisville and Ellenville. The rocks below, to and including the Rondout Formation, will not be seen but are included to complete the listing. The rocks older than the Binnewater Sandstone are shown in Figure 3 on p. 5. The terminology for stratigraphic units above the Onondaga Limestone is not well established in this area (see Rickard, 1964).

MOUNT MARION FORMATION, Hamilton Group (Middle Devonian, 1,000+ ft [305+ m] thick): Olive-gray to dark-gray, platy, very fine- to medium-grained, sandstone, siltstone, and shale.

BAKOVEN SHALE, Hamilton Group (Middle Devonian, 200-300 ft [61-91 m] thick): Dark-gray shale. Zone of faulting at base in several places.

ONONDAGA LIMESTONE (Middle Devonian, 100 ft [30 m] thick): Cherty fossiliferous limestone.

SCHOHARIE FORMATION (Lower Devonian, 180-215 ft [55-66 m] thick): Thin- to medium-bedded, calcareous mudstone and limestone; more calcareous upwards.

ESOPUS FORMATION (Lower Devonian, 200 ft [61 m] thick; thickens to southwest): Dark, laminated and massive, non-calcareous, siliceous, argillaceous siltstone and silty shale.

GLENERIE LIMESTONE (Lower Devonian, 50-80 ft [15-24 m] thick): Siliceous limestone, chert, and shale, thin- to medium-bedded.

CONNELLY CONGLOMERATE (Lower Devonian, 0-20 ft [0-6 m] thick): Dark, thin- to thick-bedded pebble conglomerate, quartz arenite, shale, and chert.

PORT EWEN FORMATION (Lower Devonian, 70-180 ft [21-55 m] thick): Dark, fine- to medium-grained, sparsely fossiliferous, calcareous, partly cherty, irregularly bedded mudstone and limestone.

ALSEN LIMESTONE (Lower Devonian, 20 ft [6 m] thick): Fine- to coarse-grained, irregularly bedded, thin- to medium-bedded, argillaceous and partly cherty limestone.

BECRAFT LIMESTONE (Lower Devonian, 3-50 ft [0.9-15 m] thick): Massive, very light- to dark-gray and pink, coarse-grained, crinoidal limestone, with thin-bedded limestone with shaly partings near the bottom in places.

NEW SCOTLAND FORMATION (Lower Devonian, 100 ft [31 m] thick): Calcareous mudstone and silty, fine- to medium-grained, thin- to medium-bedded limestone. May contain some chert.
KALKBERG LIMESTONE (Lower Devonian, 70 ft [21 m] thick): Thin- to medium-bedded, moderately irregularly bedded limestone, finer grained than Coeymans Formation below, with abundant beds and nodules of chert and interbedded calcareous and argillaceous shales.

RAVENA MEMBER OF THE COEYMANS LIMESTONE (Lower Devonian, 15-20 ft [5-6 m] thick): Wavy bedded, fine- to medium-grained and occasionally coarse-grained, limestone with abundant thin shaly partings.

THACHER MEMBER OF THE MANLIUS LIMESTONE (Lower Devonian, 40-55 ft [12-17 m] thick): Laminated to thin-bedded, fine-grained, cross-laminated, graded, microchanneled, mudcracked, locally biostromal limestone with shale partings.

RONDOUT FORMATION (Lower Devonian and Upper Silurian, 30-50 ft [9-15 m] thick): Fossiliferous, fine- to coarse-grained, thin- to thick-bedded limestone and barren, laminated, argillaceous dolomite. Limestone lentils come and go, and the more persistent ones have been named (from top to bottom): Whiteport Dolomite, Glasco Limestone, and Rosendale Members.

BINNEWATER SANDSTONE (Upper Silurian, 0-35 ft [0-11 m] thick): Fine-grained, thin- to thick-bedded, crossbedded and planar-bedded, rippled quartz arenite , with gray shale and shaly carbonate. Probably grades southwestwardly into the Poxono Island Formation.

POXONO ISLAND FORMATION (Upper Silurian, 0-500 ft [0-152 m] thick): Poorly exposed gray and greenish dolomite and shale, possibly with red shales in the lower part.

HIGH FALLS SHALE (Upper Silurian, 0-80 ft [0-24 m] thick): Red and green, laminated to massive, calcareous shale and siltstone, occasional thin argillaceous limestone and dolostone. Ripple marks, dessication cracks.

BLOOMSBURG RED BEDS (Upper Silurian, 0-700 ft [0-213 m] thick)--Grayish-red and gray shale, siltstone, and sandstone. In southeastern New York the Bloomsburg intertongues with rocks of the Shawangunk Formation, forming two mappable units: The **Bascher Kill Tongue** above (0-300 ft [0-91 m] thick)**—**Grayish-red siltstone and shale and slightly conglomeratic, partly crossbeddedandstone with pebbles of milky quartz, jasper, and rock fragments, and gray sandstone; and the **Wurtsboro Tongue** below (Late Silurian, 0-330 ft [0-101 m] thick)**—**Interbedded red, green, and gray, cross bedded, polymictic conglomerate, sandstone, siltstone, and shale. Lithologies occur in fining-upward cycles as much as 12 ft (3.7 m) thick.

SHAWANGUNK FORMATION (Middle Silurian, 0-1,400 ft [0-427 m] thick): Crossbedded and planar-bedded, channeled, quartz-pebble conglomerate (rose quartz conspicuous in upper part), quartzite, minor gray, shale and siltstone, and lesser red to green shale. In southeastern New York the Shawangunk intertongues with rocks of the Bloomsburg Red Beds, forming two mappable units: The **High View Tongue** above (Middle Silurian, 0-100? ft [0-31? m] thick)**—**gray, thin-bedded quartz sandstone with well-sorted and rounded quartz grains, and some green shale; and the **Ellenville Tongue** below (Middle Silurian, 0-500+ ft [0-152+ m] thick)**—**medium-dark-gray to light-gray, medium- to thick-bedded, planar-bedded and cross-bedded, fine- to medium-grained quartzite, conglomerate, and minor red and greenish-gray shaly siltstone. Conglomerates are polymictic, with quartz, jasper, chert, and argillite pebbles. The Ellenville Tongue is similar to some of the rocks of the Greenpond Conglomerate exposed in central New Jersey.

DIAMICTITE (Lower Silurian or Upper Ordovician, <1 ft [<0.3 m] thick): Lying between the Shawangunk and Martinsburg Formations is a dark yellowish orange, compact sand-silt mixture with abundant clasts of fragments of the underlying Martinsburg Formation, quartz pebbles (similar to those found in the overlying Shawangunk), sheared vein quartz, clay gouge, and a variety of gray and reddish siltstone and sandstone pebbles, some of which are rounded.

MARTINSBURG FORMATION (Upper and Middle Ordovician, 10,000+ ft [3,048+ m] thick): Greater than 10,000 ft (3,050 m) thick. See Appendix 1 for descriptions of the **Ramseyburg, High**

Point, and Bushkill Members. The subdivision of Ordovician clastic rocks in southeastern New York is in a state of flux due to incomplete detailed mapping, as summarized in the section above.

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APPENDIX 3. Structural-stratigraphic packages within the field trip area. Modified from Lyttle and Epstein (1987)

VALLEY AND RIDGE PROVINCE: The Valley and Ridge province can be divided into lithotectonic units, each comprising a group of individual formations with their own deformation and erosion characteristics (Epstein and Lyttle, 1993; Epstein and Epstein, 1969). Incompetent rocks, such as the Marcellus Formation of Devonian age, the Poxono Island Formation of Silurian age, and the top of the Ordovician Martinsburg Formation, are zones of detachment or zones of rapid change in structural style, separating the lithotectonic packages. Alleghanian deformation, dominated by a thrust system of imbricate splays, has produced a series of northeast-trending, northwest verging upright to overturned folds. Many of the anticlines may be directly related to blind thrusts at depths.

GREEN POND SYNCLINE: The Green Pond syncline is a narrow, northeast-trending, faulted syncline containing a thin, but fairly complete section of Paleozoic sediments, located in the middle of the external massif of the Reading Prong, and sitting unconformably upon, or in fault contact with the Proterozoic basement. The Paleozoic section is cut by a number of thrust faults, some possibly having a strike-slip component. Many of the Paleozoic rocks correlate with thicker units in the Valley and Ridge Province showing that these rocks once were present between the two areas, a distance of more than 16 mi (25 km).

LEHIGH VALLEY SEQUENCE OF THE GREAT VALLEY: The Great Valley of Pennsylvania and New Jersey is a northeast-trending lowland bounded by Blue Mountain-Kittatinny Mountain on the northwest and the highlands of the Reading Prong on the southeast. The Great Valley comprises a thick section of Ordovician clastics, a thinner section of Ordovician and Cambrian carbonates, and a very thin Cambrian quartzite at its base. All of these rocks have been multiply deformed by a continuum of deformation between Taconic and Alleghanian time. These rocks contain numerous thrust faults, generally concentrated at several stratigraphic levels, such as the entire lower thin-bedded slate member of the Martinsburg Formation, the incompetent Jacksonburg Limestone which sits stratigraphically on top of the much more competent, dolomite-dominated, Beekmantown Group, and the base of the Cambrian Allentown Dolomite or Leithsville Formation. At the southwest end of the field trip area, the Lehigh Valley sequence rocks are structurally overlain by the Lebanon Valley sequence, which appear, at least in part, to be slightly deeper water, and perhaps farther travelled, equivalents of the Lehigh Valley sequence.

SPITZENBERG AND SHARPS MOUNTAIN OUTLIERS: These two small erosional remnants in the Great Valley north of Hamburg, PA contain reworked sediments of both the Hamburg klippe and the Lehigh Valley sequence (Lash et al., 1984, p. 74-78). These rocks rest unconformably, and perhaps with structural discontinuity, on rocks of the Hamburg klippe and unconformably beneath the Tuscarora Sandstone (Stephens, 1969). Although difficult to prove, these rocks may have been deposited after the rocks of the Hamburg klippe were folded during the Taconic orogeny and may represent the youngest Ordovician clastics in the area

ROCKS OF THE SHOCHARY RIDGE AREA:This group of Ordovician clastic rocks crops out over a fairly small area north of Lenhartsville, Pennsylvania. It is entirely fault bounded, and sits structurally on top of all three members of the Martinsburg Formation, and structurally beneath the Greenwich slice of the Hamburg klippe (Figure 4 on p. 11). It is very likely that the rocks of Shochary Ridge are derived from the Greenwich slice of the Hamburg klippe and were transported in Taconic time as a thrust slice beneath the two major slices of the klippe. The two faults that mark most of the present limits of the Shochary Ridge sequence of rocks are probably Alleghanian in age. The Shochary sequence probably formed as a local, northward-prograding fan from the advancing accretionary prism of the Greenwich slice of the Hamburg klippe. These rocks are roughly the same age as the slightly deeper water clastics of the Martinsburg Formation which were deposited by dominantly northwestward currents within a very long northeast-trending basin.

HAMBURG KLIPPE: This structurally and stratigraphically complex group of far-travelled rocks is located in the Pennsylvania Great Valley at the west end of the field trip area and resembles the Taconic allochthon of New York State. The klippe is divided into two tectonic slices: 1) the Greenwich slice, an accretionary prism of sediments displaying scaly cleavage that composes the Windsor Township Formation, and 2) the Richmond slice, a sequence of rise and slope deposits that composes the Virginville Formation (Lash et al., 1984). Although emplaced by thrust faults during the Taconic orogeny, these faults have been obscured by later Alleghanian folds and faults. In general, rocks of the Lehigh Valley sequence has been thrust on top of the rocks of the Hamburg klippe during Alleghanian time

JUTLAND KLIPPE: These rocks resemble some of the rocks found within the Greenwich slice of the Hamburg klippe in Pennsylvania. Unlike the Taconic allochthon and Hamburg klippe, which occur on the foreland side of the external massifs of the Berkshires and Reading Prong, the rocks of the Jutland klippe occur on the hinterland side of the Reading Prong.

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SOME TACONIC UNCONFORMITIES IN SOUTHEASTERN NEW YORK

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This Field Conference will have visited or discussed 12 locations where the Ordovician-Silurian contact is exposed or nearly so: six in Pennsylvania - Schuylkill Gap (Stop 1), Sharps Mountain, Spitzenburg Hill, northeast extension of the Pennsylvania Turnpike, Lehigh Gap (Stop 3), and Delaware water Gap (Stop 4); four in New Jersey - Yards Creek (Stop 5), Sunrise Mountain (mileage 54.5 of the Day 2 Road Log), the Beemerville intrusion (Stop 7), and High Point State Park (Stop 8); and two in New York - Otisville (Stop 9) and the Ellenville arch (Stop 10). Besides the contact exposed in the now-deeply flooded High View Railroad tunnel described by Inners and others in this guidebook, which is a mere 700 ft (213 m) south of the exposure along NY 17 (#3 of this list), there are an additional four excellent occurrences of the unconformity (Figure 1). They are described below. They add further insight into the faulted nature of the contact, adding to the puzzlement of how much of the contact is unconformable and how much is faulted.

1. Interstate 84

Coordinates: 41°22'24.64"N latitude; 74°37'32.26"W longitude; elevation 1,267 ft (386 m)

This exposure is located on the south side of the east-bound lane of I-84, 3 mi (5 km) east of Port Jervis, NY. The exposure is more than 1,500 ft (457 m) long, with gently folded Martinsburg (Figure 2, top) in contact with the Shawangunk, and with an angular discordance of 5° (Figure 2, bottom). Of particular interest is the occurrence of a 3-in (8-cm) layer of clay-fault gouge at the contact. The rocks were well exposed in 1966, soon after construction of the highway.

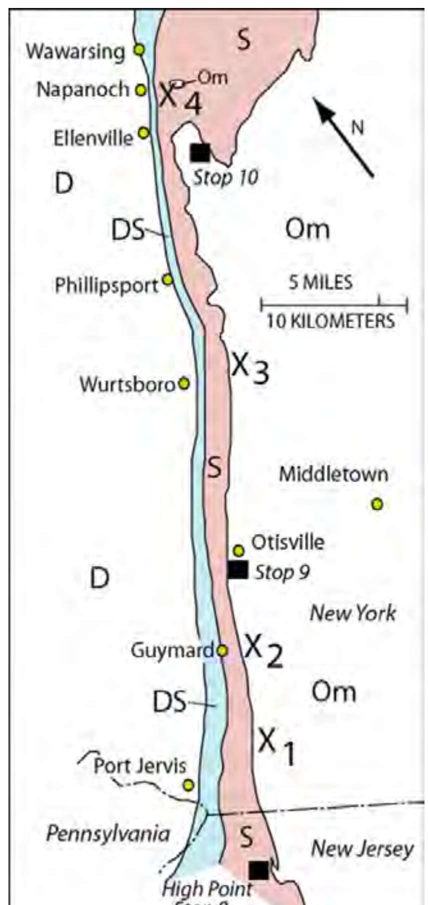

Figure 1. Generalized geologic map of parts of New Jersey and New York showing where exposures of the Taconic unconformity not seen on this Field Conference are located (X), and Stops 8, 9, and 10 (squares). *D*, Rocks of the Hamilton Group (Plattekill, Ashokan, Mount Marion, and Bakoven Formations and younger. *DS*, Rocks between the Onondaga Limestone and Binnewater Sandstone-Poxono Island Formation. *S*, High Falls Shale, Shawangunk Formation, Bloomsburg Red Beds, and tongues of the Shawangunk and Bloomsburg; Om, Martinsburg Formation. 1. I-84; 2. Guymard prospect; 3. NY 17 at Wurtsboro; 4. Napanoch Prison.

Epstein, J.B. and Lyttle, P.T., 2012, Some Taconic unconformities in southeastern New York, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 37-41.

Table 1. Description of units in Figure 2, bottom, I-84 (from base upwards).

Unit	Thickness (feet)	
Martinsburg Formation		
1a	1	Dark-gray argillite with interbedded argillite and greywacke below. Upper contact sharp.
1b	3.1	Medium-gray to medium-dark-gray, massive, fine- to medium-grianed, light-olive-gray to moderate-brown weathering, non-calcareouos greywacke; upper contact sharp. Much hematite in weathered rock.
1c	1.8	Interbedded, fine-grained greywacke with beds as much as 0.6 thick, interbedded with argillite in beds as much as 0.6 thick. Upper contact sharp.
1d	3	Same as in 1b.
2	3	Argillite, like unit 1a, partly concealed.
3	2.2	Graywacke, medium-dark-gray, non-calcareous, very fine- to fine-grained, feldspathic, with minor thin interbedded argillite. Upper contact gradatioal.
4	4.3	Medium-gray to medium-dark gray, very fine-grained to silty laminated graywacke interbedded with dark-gray to medium-dark-gray siltstone and dark-gray argillite. Upper contact gradational.
5	1.4	Medium-dark-gray, fine- to medium-grained, feldspathic greywacke with thin (< ¼ inch) dark-gray argillite. Upper contact sharp.
6a	0.8	Grayish black, to dark-gray, dark-yellowish-orange-weathering argillite. Upper contact gradational.
6b	0.8	Leached and sheared, medium-light-gray argillite. Shearing more intense at top with abundant moderate-reddish-brown iron staining.
7	0.5	Gouge, contains grayish black to dark-gray argillite fragments in a moderate-reddish-brown to dark-yellowish-orange clay matrix. Fragments generally up to one inch long, sheared parallel to overlying contact. Unit thins to about three inches where the Shawangunk along the contact overlies greywacke, and contain greywacke fragments along with argillite.
Shawangunk Formation		
8	9.4	Lower contact abrupt and disconformable. Medium-gray to medium-light -gray, massive quart- and chert-pebble conglomerate, quartz predominates. Pebbles rounded to well-rounded and as much as three inches long. Chert pebbles generally about one inch long. Pebbles slightly larger near base of unit. Matrix is medium- to coarse-grained sandstone. Upper contact gradational.
9	1.8	Greenish-gray, feldspathic, coarse- to very coarse-grained conglomeratic quartzite with rounded quartz pebbles at base as much as one ich long. Upper contact sharp.
10	8	Interbedded quartz-pebble conglomerate with pebbles as much as 1.5 inches long. Beds as much as two feet thick, interbedded with conglomeratic feldspathic sandstone between 0.5-2 feet thick. Quartz pebbles are iron-stained in upper part. Upper contact gradational.
11	2	Quartz-pebble conglomerate. Rounded pebbles as much as one inch log, with dark-yellowish-orange to moderate-reddish-brown iron staining on the pebbles. Unit is not exposed in valley to the north.

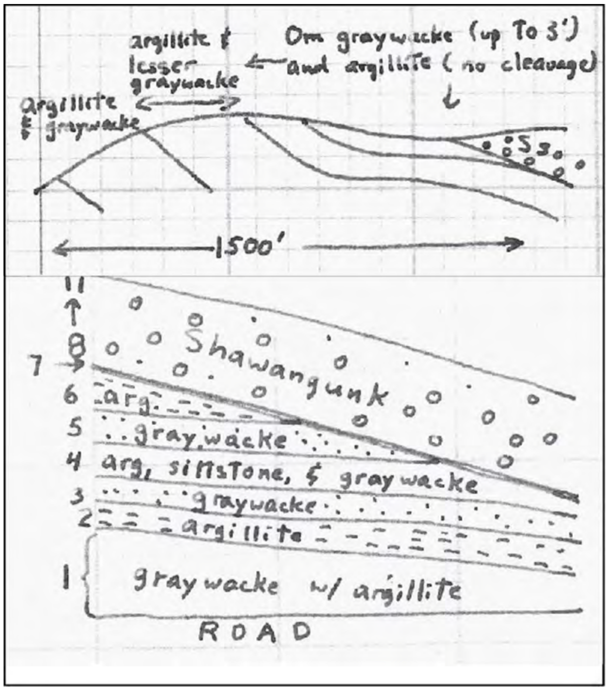

Figure 2. Field sketches of the Taconic unconformity made in 1966 along I-84, east of Port Jervis, NY. The units are described in Table 1.

Figure 3. Contact between the Shawangunk (above) and Martinsburg Formations along I-84. The angular difference between bedding in the Shawangunk (long dashes) and Martinsburg (short dashes) is 5°. Picture taken June 1966.

The outcrop has deteriorated since then. Figure 2 is a field sketch at that time, and the information gathered is given in Table 1. The pebbles here in the Shawangunk break out more readily than to the southwest at Delaware water Gap, possibly because the quartz "cement" is not as intensely diagenetically altered (= not as well "metamorphosed"). This correlates with the less intense folding noted from Pennsylvania through New Jersey, into southeastern New York. Figure 3 is a photograph of the exposure in 1966. It has deteriorated somewhat since then.

2. Guymard Prospect

Coordinates: 41°25'32.13"N latitude; 74°34'59.94"W longitude; elevation 1,180 ft (360 m)

This contact is located in a small wooded prospect hole about 25 ft (8 m) wide and extends for at least 100 ft (31 m) underneath the dip of the basal bed of the Shawangunk Formation. The Shawangunk here is a typical conglomerate with rounded quartz pebbles as much as 1.5 in (4 cm) long. Bedding in the Shawangunk is N15°E, 12°NW, whereas it is N30°E, 27°NW in the Martinsburg, a dip difference of 15°. There is a poorly developed spaced cleavage in the Martinsburg (N48°E, 57°SE). Very well-developed downward-projecting mullions grace the base of the Shawangunk (Figure 4). These trend N49°E. Immediately below the mullions is a shear zone made up of milky quartz veins,

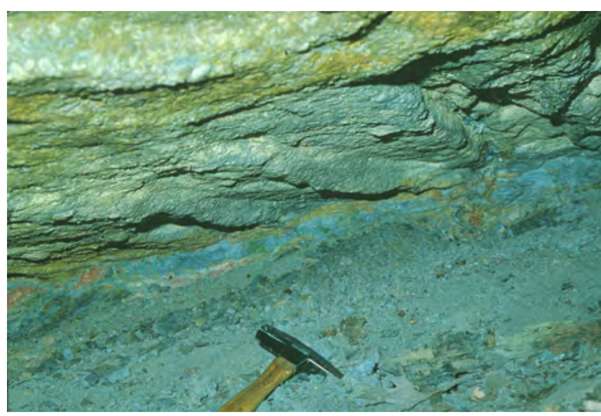

Figure 4. Assymetrical mullions that are 1 to 2 in (2.5 to 5 cm) deep in the Guymard prospect overlie about 6 in (15 cm) of clay gouge. They vary from straight (right side of the picture) to irregular (left side).

fragmented and slickensided, ireon-stained fragmented Martinsburg shale fragments in a medium-light-gray clay which is extremely contorted, similar to the contortions seen at Otisville (Stop 9, Figure 5A). Alternative to the fault interpretation, Gray (1953 and 1961), who examined many of the zinc-lead mines between Ellenville and Guymard, NY, believed that the plastic clay and yellow-brown sandy clay at the Martinsburg-Shawangunk contact at Guymard was a rock flour, representing a zone of weathering on the pre-Silurian surface. He believed that the ores were deposited from hypogene solutions that reached temperatures of about 350°F (176 °C) during the "Appalachian Revolution."

3. NY 17; Wurtsboro
Coordinates: 41°33'55.73"N latitude; 74°27'48.59"W longitude; elevation 960 ft (293 m)

This is another fine exposure of the Taconic unconformity in southeastern New York (Figure 5). The rocks of the Martinsburg and Shawangunk are similar to those we have seen elsewhere. The basal foot (0.3 m)of the Shawangunk contains much pyrite. At eye level the Martinsburg at the contact dips 16° to 31° NW, whereas the Shawangunk dips 25° NW. The angular discordance ranges up to 15° in spots. No secondary cleavage is seen in the Martinsburg at this spot. The lowest 10 ft (3 m) of the Shawangunk consist of massive, medium-gray to medium-light gray, planar-bedded, quartz- and chert-pebble conglomerate in a medium- to coarse-grained sandstone matrix. Quartz pebbles are rounded to well rounded and as much as 3 in (8 cm) long. The chert is generally not more than 1 in (2.5 cm) long. The overlying 12 ft (4 m) consist of finer, slightly feldspathic, quartz-pebble conglomerate and cross bedded conglomeratic sandstone with pebbles that are less than 15 in (38 cm) long. Graptolites of Zone 13 age (late Middle to early Late Ordovician) were identified by Berry in the Martinsburg within 135 ft (41 m) of the overlying Shawangunk (Offield, 1967, p. 53; Berry 1970). The character of the zone between the Martinsburg and Shawangunk varies from place to place. Near road level, the zone contains rotated shale fragments with some quartz veins, indicating appreciable shearing. Also included in this zone are disrupted shale fragments and a medium dark-gray sticky clay, a fault gouge.

Figure 5. Angular unconformity between the Martinsburg and overlying Shawangunk along NY 17, near the off-ramp, 1.3 mi (2.1 km) east of Wurtsboro, NY. Inset in lower left shows the sheared shale at the contact; specimen is 2 in (5 cm) long.

4. Napanoch-Maximum Security Prison
Coordinates: 41°43'55.38" N latitude; 74° 21'25.70"W longitude; elevation 630 ft (192 m)

This is an excellent exposure in the gulley behind the Eastern New York Correction Facility, a maximum security prison, servicing many gentlemen from the New York City area. The Martinsburg-Shawangunk contact is about 20 ft (6 m) long. The angular discordance

Figure 6. Contact between the Martisnburg and Shawangunk formation (at arrow) up the steep gully behind the prison at Napanoch, NY.

between the two formations is 4^o (Figure 6). The Shawangunk dips 22^o NW in the northwest limb of the Ellenville arch (Stop 10). Mullions are prominent on the basal Shawangunk surface, and there is a shear fabric in both the Martinsburg and Shawangunk. The mullions trend N58^{o}E, and bow down the slightly sheared bedding in the upper few inches (centimeters) of the Martinsburg. They are spaced about 6 to 12 in (15 to 31 cm) apart with amplitudes of about 2 in (5 cm). Within the basal 3 in (8 cm) of the Shawangunk there is a 1 to 2 mm spaced "shear cleavage" oriented N34^{o}E, 24^{o}NW. That is undoubtedly related to movement along the contact. The Shawangunk is medium to thick bedded (up to 2.5 ft [0.8 m] thick), planar bedded, with quartz pebbles up to 2 in (5 cm) long, with a medium- to coarse-grained sand matrix. Sedimentary structures include low-amplitude channels and cross-beds with a northwest current direction. Care should be taken if visiting this outcrop. Definitely seek permission from the prison before entering. Towers have a clean sight-line to the gulley.

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THE SILURIAN UNCONFORMITY WEST OF SCHUYLKILL GAP

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The westernmost exposure of the Silurian unconformity that we will visit on this Field Conference is the angular unconformity at Schuylkill Gap near Hamburg, Pennsylvania. However, the unconformity continues westward, north of Harrisburg and across the Susquehanna River (Figure 1). At its only two exposures through this stretch – Swatara Gap and Susquehanna Gap – the Martinsburg-Tuscarora contact is a disconformity, a significant time gap but with bedding generally parallel across the unconformity. The change from angular unconformity to disconformity through an area still profoundly affected by Taconic deformation is something of a puzzle.

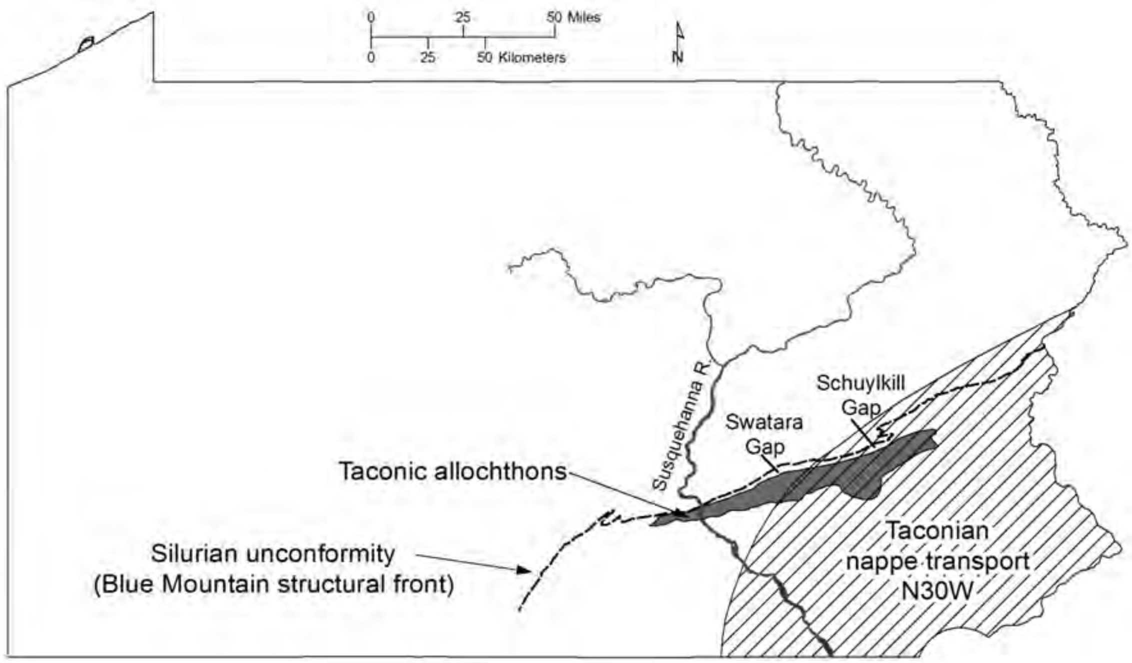

Figure 1. Model for changing character of the Silurian unconformity from the Susquehanna River to the Delaware River. White area between the Taconic allochthons and the Silurian unconformity is occupied by Martinsburg Formation. North of the Blue Mountain structural front, there was only minor deformation of the Martinsburg-Reedsville foreland in the Taconic Orogeny. Hatched area shows the general limit of major Taconian nappes. Taconian structures cut deeper and extended farther to the NNW nearer the Delaware. Alleghanian thrust sheets turned up the Silurian edge with more ENE trends, resulting in exposures having deeper pre-Silurian erosion and more intense angular unconformity toward the Delaware.

Through most of its length, the northern part of the Great Valley is occupied by the Martinsburg Formation. However, from west of the Susquehanna River in Cumberland County, through Dauphin, Lebanon, and Berks counties, the Martinsburg is interrupted by the Taconic

Ganis, G.R., Blackmer, G.C., and Wise, D.U., 2012, The Silurian unconformity west of Schuylkill Gap, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 42-45.

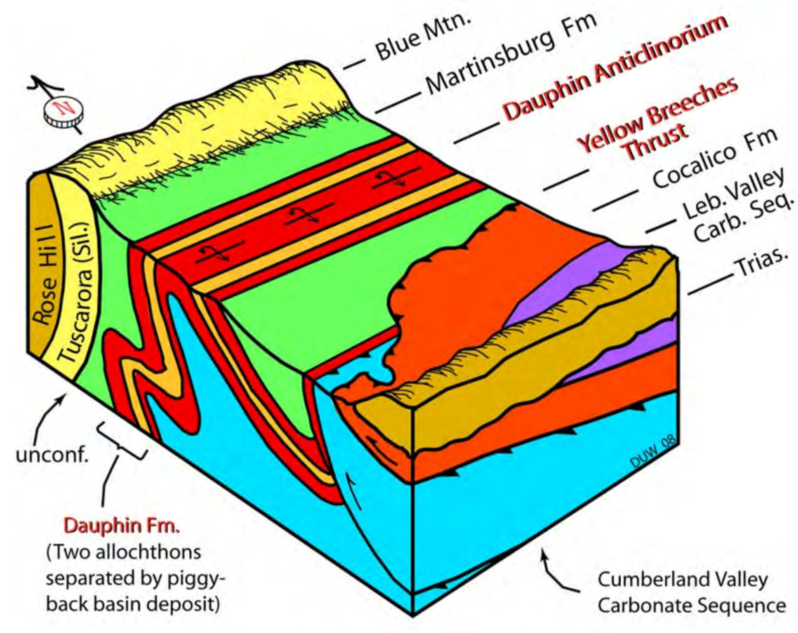

Figure 2. Cartoon block diagram showing basic structure of the Great Valley in Dauphin County. The western edge of the block is close to the east bank of the Susquehanna River.

allochthons (Figure 1). Stose (1946) called these Cambrian through Middle Ordovician rocks the Hamburg klippe, interpreting them as thrust over the Late Ordovician Martinsburg Formation. This interpretation was challenged but not overcome by Platt et al. (1972). More recent work has shown through faunal evidence (Ganis et al., 2001) and field mapping (Blackmer and Ganis, unpublished 1:24,000 maps) that the allochthons were emplaced prior to Martinsburg deposition and are therefore below the shale, rather than thrust over it (Ganis et al., 2001; Wise and Ganis, 2009; Ganis and Wise, 2008). Through Dauphin, Lebanon, and western Berks County, the allochthons and overlying Martinsburg are folded into the large overturned Dauphin anticlinorium (Figure 2; Ganis and Blackmer, 2010). The trend of the axial trace of this fold is generally parallel to the Blue Mountain structural front. The fold core exposes the "Hamburg" allochthons of the Dauphin Formation, so named by Ganis et al. (2001). The Martinsburg in the southern limb is cut off by the Yellow Breeches thrust of Alleghanian age (MacLachlan, 1967). The Martinsburg in the northern limb is overlain by the Silurian unconformity.

The overturned northern limb is marked on the ground by a belt of Martinsburg Formation. At Swatara Gap, the steeply overturned Martinsburg section has an outcrop width of about 2 mi (3 km), extending from the gap south about to the Lickdale interchange of I-81. This substantial thickness of Martinsburg thins toward Susquehanna Gap (about 18 mi [29 km] to the west) where, on the west side of the river, it is only a thin band. How much of this thickness change is due to structural causes versus original depositional conditions is difficult to decipher.

At Swatara Gap, late Martinsburg Edenian shelly fauna (Willard, 1943; Platt, 1972) and graptolites of equivalent age (Ganis ID) are directly below the Tuscarora. Moving south, through a progressively older inverted section toward Lickdale, there are additional fossiliferous exposures of late to middle Martinsburg age in the *Diplacanthograptus spiniferus* Zone and the

Diplacanthograptus caudatus Zone; and basal Martinsburg in the *Climacograptus bicornis* Zone in depositional contact above the Dauphin Formation (for details, see Ganis, 2004). Ganis (1972, unpublished data) identified a thin bed of Edenian shelly fauna, similar to the Swatara Gap fauna, at the east side of the Susquehanna Gap. Stose (1930; further discussed in Willard, 1943) found Edenian fossils, also similar to the Swatara Gap fauna, at the top of the Martinsburg on the west side of the Susquehanna. Both of these collecting locations have been destroyed by road construction. So, despite significant thickening of the Martinsburg from west to east, the contact of the Martinsburg and Tuscarora between Susquehanna Gap and Swatara Gap is consistently high in the Martinsburg. Thus it appears that the unconformity does not cut significantly downsection along this stretch.

The predominant expression of Taconic deformation in Pennsylvania is large overturned to recumbent nappe structures. It is easy to imagine creating an angular unconformity by laying down the Silurian sediments atop the Ordovician nappes, then folding the entire system during the Alleghanian orogeny to give the unconformity the steep dip it has today. It is more difficult to imagine a setting where Silurian sediments are laid down subparallel to bedding in the Martinsburg in either the upper or lower limb of a nappe fold, and still allow for exposure the core of the original nappe in the relatively straightforward map pattern observed in the Great Valley today.

The direction of tectonic transport of the Taconic nappes, as derived from a compilation of data from many small areas (Wise and Werner, 2004), was about N30^{o}W. The Taconic front was therefore perpendicular to that direction, or about N60^{o}E. This direction is slightly offset counter-clockwise from the later N70^{o}E trend of Alleghanian frontal structures of the Reading Prong – Musconetcong thrust. As a result, in proceeding westward along Blue Mountain, the Taconic deformation front migrates away from the Silurian unconformity, leaving the basal Silurian against the less deformed parts of the Taconic foreland (Figure 1). To the east along Blue Mountain, the unconformity migrates deeper into the Taconic frontal deformed zone. Predictably, missing stratigraphy and increasingly angular relationships become more prominent in that direction.

Inherited local structures may also have contribute to the lack of Taconic nappe folds west of Schuylkill Gap. Ganis and Wise (2008) and Wise and Ganis (2009) describe the thin-skinned nature of Taconic deformation in this part of the Great Valley. In the earliest stages of the orogeny, the great allochthonous slices which now constitute the Dauphin Formation were thrust from offshore sedimentary basins onto the Cambro-Ordovician carbonate platform. As orogeny progressed, the allochthons moved inboard across the platform by massive gravity-driven flow and were covered by Martinsburg flysch. As loading continued, the carbonates weakened sufficiently to flow and overturn, forming the Lebanon Valley nappes. The allochthons, however, may have accommodated the continued loading by sliding along the embedded fault planes rather than by buckling into folds (think of straightening a pile of disordered papers – when you push on the outside of the pile with your hands, the odd sheet might buckle but the vast majority respond by simply sliding over each other). As a result, the Martinsburg Formation above the allochthons may have been slightly eroded but otherwise little disturbed before its unconformable burial by the Silurian sediments.

If the Dauphin Formation allochthons and their Martinsburg cover had adjusted along existing low-angle faults rather than folding during the Taconic Orogeny, the Martinsburg still could have been relatively parallel to the depositional surface during Tuscarora deposition. The whole system then would have undergone Alleghanian folding, giving the disconformity its present near-vertical attitude. In fact, Alleghanian rather than Taconic folding of the Dauphin Formation fits better with field observations in the Harrisburg to Fredericksburg area (detailed mapping has not yet progressed east of Fredericksburg). The structure in this area appears to be a relatively straightforward train of asymmetric-to-overturned anticlines verging toward the Blue Mountain structural front without the complications expected from refolding. Outcrop-scale folds are commonly observed in regions affected by nappe folding. In two quadrangles of detailed field mapping, only one outcrop-scale fold was observed, in a limestone outcrop at Harper Tavern – an asymmetric anticline verging toward the Blue Mountain structural front. The Schuylkill Gap lies near the east end of the Dauphin Formation, where a thinner stack of allochthons may have allowed Taconic fold development.

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THE NATURE OF SILURIAN MOLASSE AND THE TACONIC UNCONFORMITY IN THE GREEN POND SYNCLINE, NEW JERSEY-NEW YORK, USA

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Introduction

This paper summarizes the outcrop and subsurface expressions of the Ordovician Taconic unconformity and the overlying Silurian molasse in the Green Pond syncline (GPS), an elongate belt of down-faulted Lower and Middle Paleozoic rocks in the New Jersey (NJ) Highlands and bordering the New York (NY) Hudson Highlands on the west (Figure 1). The Taconic unconformity is the erosion surface resulting from tectonic uplift during the Taconic orogeny of the present-day New England region southwestward through the NY recess and into the Pennsylvania Salient (Figure 1). The Taconic was among the first of a series of Paleozoic mountain building episodes affecting the eastern continental margin of ancestral North America (Drake et al., 1989). It is widely thought to have

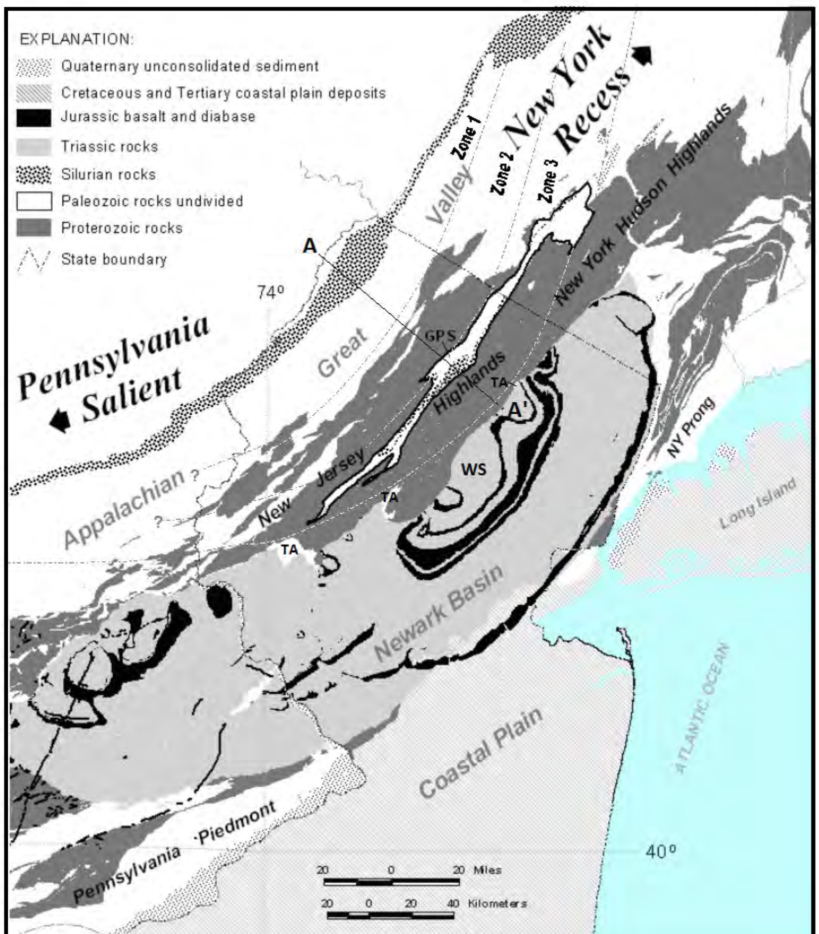

Figure 1. Generalized geology of the north mid-Atlantic region USA, showing named provinces and Appalachian terrain relative to the Green Pond Syncline (GPS). Cross section trace A-A' is coincident with parts of section A-A' of Drake et al. (1996) and Herman et al. (1997 - see Figure 2). The trace of Zones 1-3 are extrapolated southwestward into PA from New York State where Epstein and Lyttle (1987) reported increasing Taconic orogenic strains progressing southeastward toward the location of the Taconic allocthons, noted with a 'TA' along the northwest border of the Newark Basin. Fold and faults strains in each zone are detailed in the text. WS – Watchung syncline.

Herman, G.C., 2012, The nature of Silurian molasses and the Taconic unconformity in the Green Pond syncline, New Jersey-New York, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 46-84.

resulted from the collision and obduction of an island arc onto the continental margin beginning in the Early to Middle Ordovician period and ending in the Late Ordovician (~443 Ma), spanning a time period of about 15-20 million years (Wise and Ganis, 2009). Evidence for the unconformity now occurs where basal Silurian quartzite and conglomerate rest on Middle Proterozoic to Middle Ordovician bedrock exposed during orogenesis. The Taconic unconformity is an angular unconformity in renowned areas along the base of Hawk Mountain, Pennsylvania (PA), Kittatinny Mountain, NJ, and Shawangunk Mountain, NY at the front of the Appalachian Great Valley where dip angles across the unconformity between Early to Middle Ordovician foredeep sedimentary rocks and the Silurian molasse vary upward to as much as 15^{o} (Epstein and Lyttle, 1987). These relationships indicate that the current area of the Great Valley was a Taconic foreland region that was mildly to moderately strained with open-upright folds and tilting during the Taconic Orogeny (Epstein and Lyttle, 1987). Evidence of rock strain of Taconic age increasing southeastward within the Ordovician flysch of the Hudson Valley led Epstein and Lyttle (1987) to define three deformation zones before encountering Taconic allocthons (Figure 1). Zone 1 is mildly strained with open, upright folding; zone 2 contains tighter, steeper folds and localized thrust faulting, and zone 3 consists of thrust faults, overturned folds, and tectonic mélange. A recent effort of constructing palinspastic cross sections through the NJ region (Drake et al., 1996) is based on balanced foreland structures within zone 1 (Herman et al., 1997) that includes a depiction of a Taconic foreland-fold sequence of Lower Paleozoic rocks cored by Proterozoic basement (Figure 2). The restored

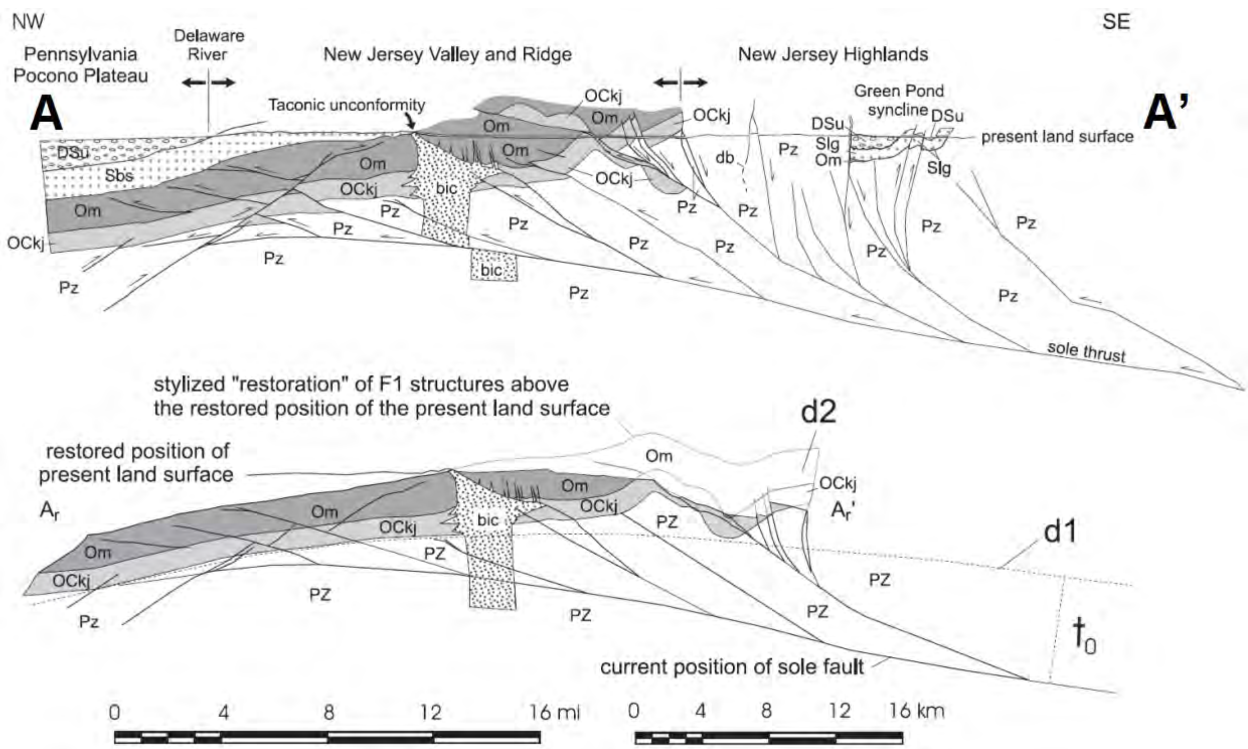

Figure 2. Details of the current (top) and palinspastic (bottom) restoration of a Taconic foreland sequence adapted from Herman et al. (1997). The section corresponds to the northwest 2/3 of A-A' of Figure 1. Abbreviations: bic—Beemerville intrusive complex; Om – Martinsburg Fm.; OCKj – Cambrian Ordvician Kittatinny Dolomite and Limestone; Pz – undivided Middle Proterzoic gneiss and granite; Sbs – Silurian Shawangunk and Bloomsburg Formations; DSu – undivided Silurian-Devonian section,; Slg Silurian Green Pond Conglomerate and Longwood Shale; d2 – restored Taconic alignment of the Cambrian-Ordovician (CO) cover sequence; d1 – restored CO base of the carbonate platform prior to sole-fault development.

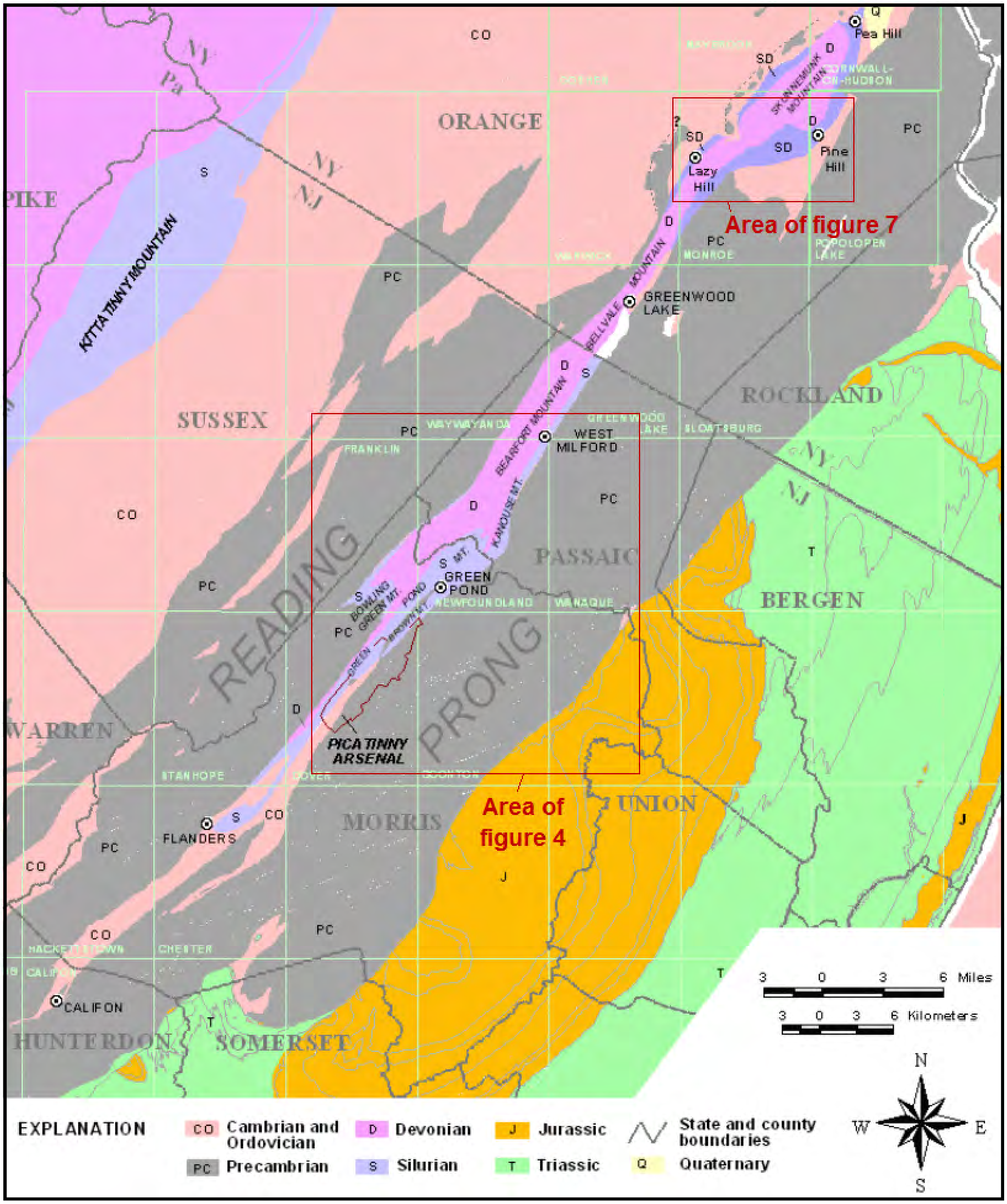

Figure 3. Generalized bedrock geology of the Green Pond Syncline. Adapted from Rickard et al. (1970) and Drake et al. (1996). Note the areas for Figures 4 and 7.

cover folds in NJ assumed a pre-Taconic, broadly arched shelf sequence for the geographic extent of the current southwest NJ Highlands, Kittatinny Valley, and Kittatinny Mountain. But the reconstructed Taconic foreland sequence doesn’t reach into the northern highlands region owing to the removal of Lower Paleozoic rocks from the structural culminations of the Reading Prong (Figure 3), and thereby lacks the basis for palinspastic, balanced cross sections in those areas.

Stratigraphic and structural details of the Taconic unconformity and overlying Silurian molasse reported from previous work are reviewed as a basis to discuss their geologic nature in the GPS. Early work in NY by Darton (1894 a, b) was followed closely in NJ by Kummel and Weller (1902), who mapped and reported detailed lithological and structural relationships throughout the Green Pond Mountain region where basal Silurian sedimentary rocks rest on older rocks from Precambrian to Ordovician age. In NY however, details surrounding the nature of the unconformity are sketchier as the Green Pond Formation is restricted in its distribution and strike length at three different locations around the northeastern end of the regional syncline. Moreover, the Green Pond Formation in NY is mapped using different names in different locations, and there is a lack of detailed, uniform mapping that could otherwise shed some light on the stratigraphic variations and structural complications seen in that area. Currently, Green Pond Conglomerate is mapped on the east side of the syncline (Dodd, 1965) whereas Shawangunk Formation is mapped on the west side (Jaffe and Jaffe, 1973). These stratigraphic units are of the same age, and part of the same depositional sequence, but are mapped as separate units because of contrasting matrix color and clast composition when moving from one side of the syncline to the other. These differences are briefly noted as part of a discussion on the formation nomenclature and why the name 'Green Pond Formation' is generally used herein.

A recent trip to Orange County, NY, was made in order to visit the two locations mentioned above on different sides of the syncline where the nature of the Silurian molasse differs in outcrop. The lithological differences seen in this area are compared and contrasted with those occurring elsewhere in the GPS. A new subsurface photo of the unconformity stemming from a borehole televiewer survey of a water well from the central part of the NJ Green Pond Mountain region is also documented. This is the first subsurface record of the stratigraphic contact in NJ, and the only known record of the unconformity in the GPS besides that mentioned by Kummel and Weller (1902), which has not been corroborated. Aspects of the unconformity and overlying molasse are added from other referenced sources to help gain perspective for a palinspastic stylization of the NJ Highlands. This stylization builds on the northernmost palinspastic section from Herman et al. (1997) showing a mildly strained Cambrian-Ordovician carbonate platform and overlying Middle Ordovician flysch, both of which are arranged in open, upright folds of Taconic age. The section is extended southeast to depict the strain profile of a Taconic fold and thrust belt, an erosional unconformity, and covering molasse.

Geologic Setting

The GPS is the largest Paleozoic Valley in the Reading Prong and lies at the juncture between the central and northern Appalachian region (Figure 2). It’s about 75 mi (121 km) long and up to 6 mi (10 km) wide, with about two-third’s lying in NJ and one-third in NY (Figure 2). The Paleozoic sedimentary rocks in the GPS border the NY Hudson Highlands on the west (Figure 1), but are faulted and folded within the Reading Prong in NJ and PA (Figures 1 and 2). In a structural sense, the "syncline" is more formally a regional, northeast-plunging synclinorium with folded and faulted Lower Paleozoic (Cambrian-Ordovician) at the surface in the southwest and Middle Devonian rocks coring large, closed synclines at the surface in northeastern NJ and into NY (Figure 4). It's unique among Reading Prong Paleozoic valleys because it contains Middle Paleozoic (Silurian-Devonian) units of stratigraphic affinity with those cropping out northwest of the Great Valley, beginning with Kittatinny and Shawangunk Mountains over 15 mi (24 km) distance to the northwest (Figures 2 and 3).

Paleozoic rocks in the GPS range in age from Early Cambrian to Middle Devonian. As many as two to three marine cycles shape the stratigraphic column, including a Lower Paleozoic (Cambrian-Ordovician) marine transgression followed by a series of Middle Paleozoic events, including: an Early Silurian regression, a Middle- to Late Silurian transgression, and a Middle-Devonian regression (Herman and Mitchell, 1991). It is a widely held view that these Lower Paleozoic rocks sustained at least three tectonic phases of uplift and erosion including: (1) early Ordovician; (2) Late Ordovician Taconic orogeny; and (3) Late Paleozoic Alleghanian orogeny (alternatively known as the Allegheny or Appalachian orogeny).

Prior and New Work on the Nature of the Unconformity and the Silurian Molasse

The focus here is on the unconformity and nature of rocks bracketing the erosional surface resulting from the Early Silurian regression and the Middle to Late Silurian transgression in the GPS. The outcrop expression of basal Silurian strata defines the nature of Taconic unconformity, which is mapped in the NJ Green Pond Mountain region (Figures 4, 5, and 6) and three locations in Orange County, NY, near the northeast end of the GPS (Figure 3).

A Note on the Nomenclature of the Green Pond Formation

The basal, Lower Silurian unit in the GPS consists of a mixture of pebble to cobble conglomerate, quartzose and subgraywacke sandstone, quartz siltstone, and shale. Its thickness, composition, color, and texture vary by location in the syncline. Accordingly, it has been called many different names. Rogers (1836) first described the 'Green-pond-mountain conglomerate' but miscorrelated it with Triassic border-fault conglomerate (Thomson, 1957). Early miscorrelation with Devonian Skunnemunk Conglomerate by many others in the region was resolved by Darton (1894a) who found Devonian Helderberg limestone between them. Ries (1895) called it the Medina formation (Lower Silurian) in NY. Kummel and Weller (1902) and Southard (1960) preferred Green Pond Formation, but more recent workers have used Green Pond Conglomerate (Thomson, 1957; Herman and Mitchell, 1991; Drake et al., 1996) or Shawangunk Formation (Barnett, 1976; Jaffe and Jaffe, 1973). In this report, the unit is referred to as the Green Pond Formation. The decision to use "Formation" here rather than "Conglomerate" is based on the consideration that conglomerate locally constitutes less than half of the formation gross lithology, with one comprehensive measure indicating 49% sandstone, 42% conglomerate, 6% shale, and 3% siltstone (Thomson, 1959). The term "conglomerate", or "conglomerate of the Green pond Formation", or more informally "Green Pond conglomerate", is sometimes useful when describing the distribution of conglomerate with respect to specific locations in the syncline. The use of "Formation" is consistent with the

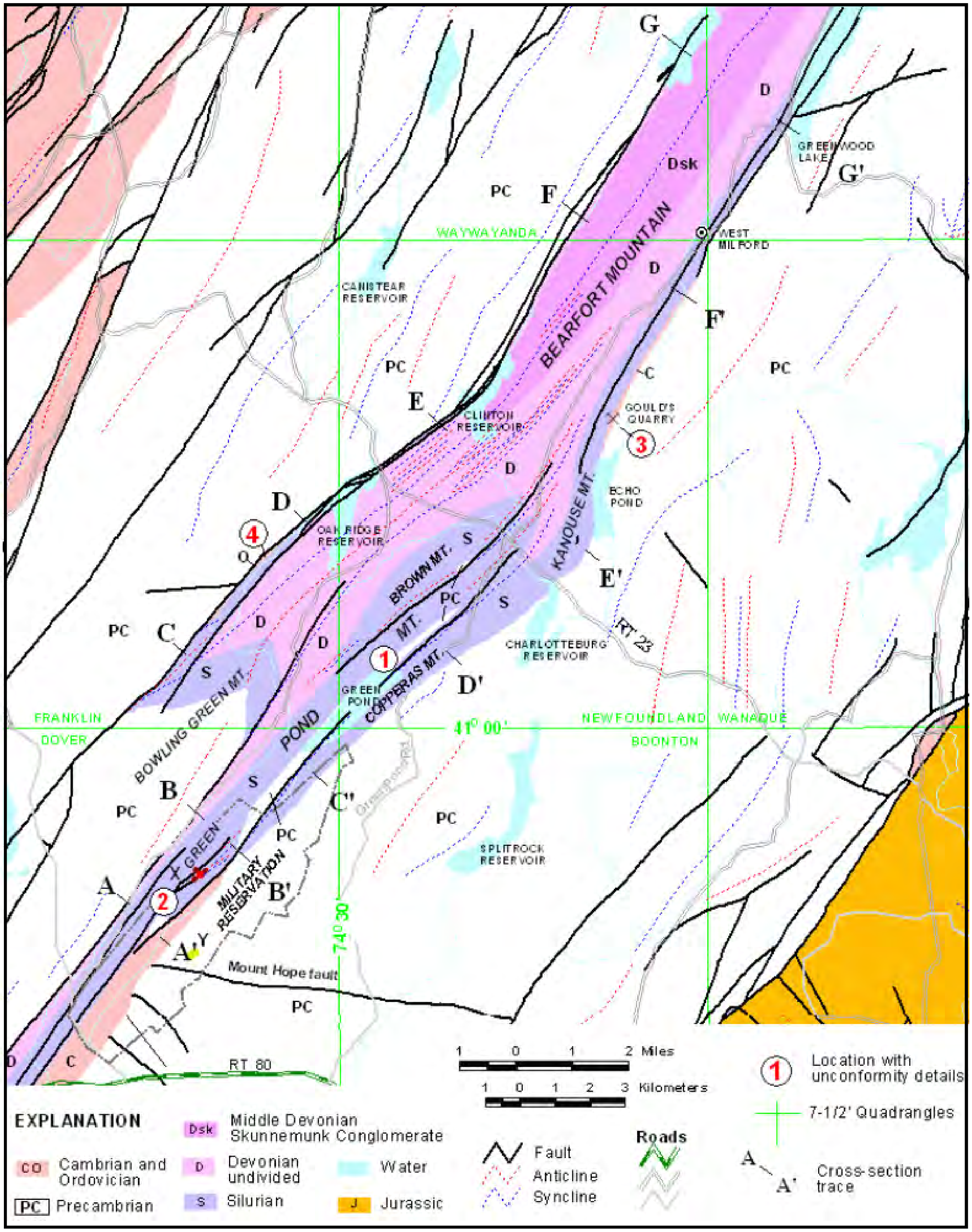

Figure 4. Generalized bedrock geology of the Green Pond Mountain region. Adapted from Herman and Mitchell (1991) and Drake et al. (1996). Note the location of section traces A through G (shown in Figure 5B). Details of the Taconic unconformity are discussed in the text for locations 1 to 4.

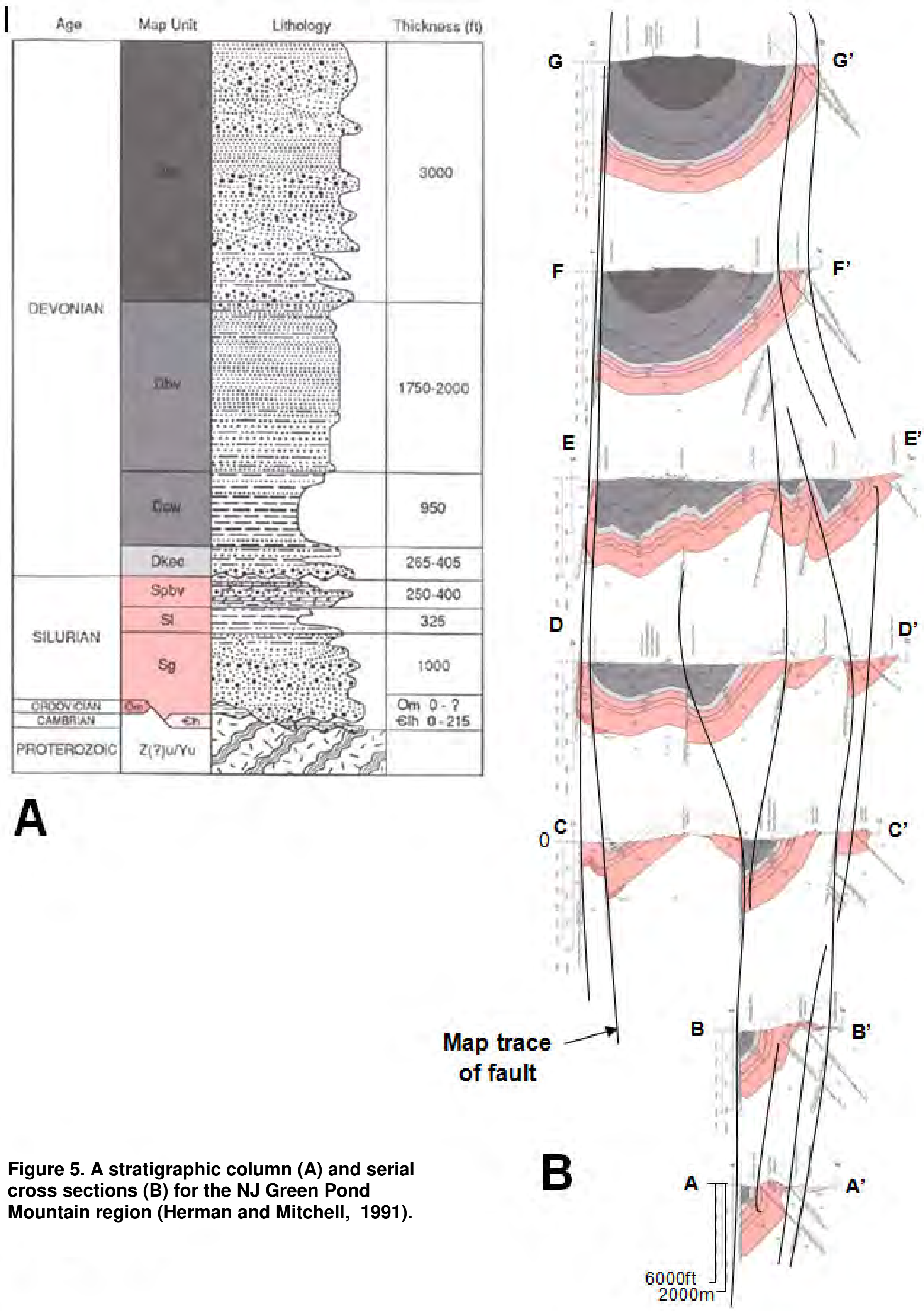

Figure 5. A stratigraphic column (A) and serial cross sections (B) for the NJ Green Pond Mountain region (Herman and Mitchell, 1991).

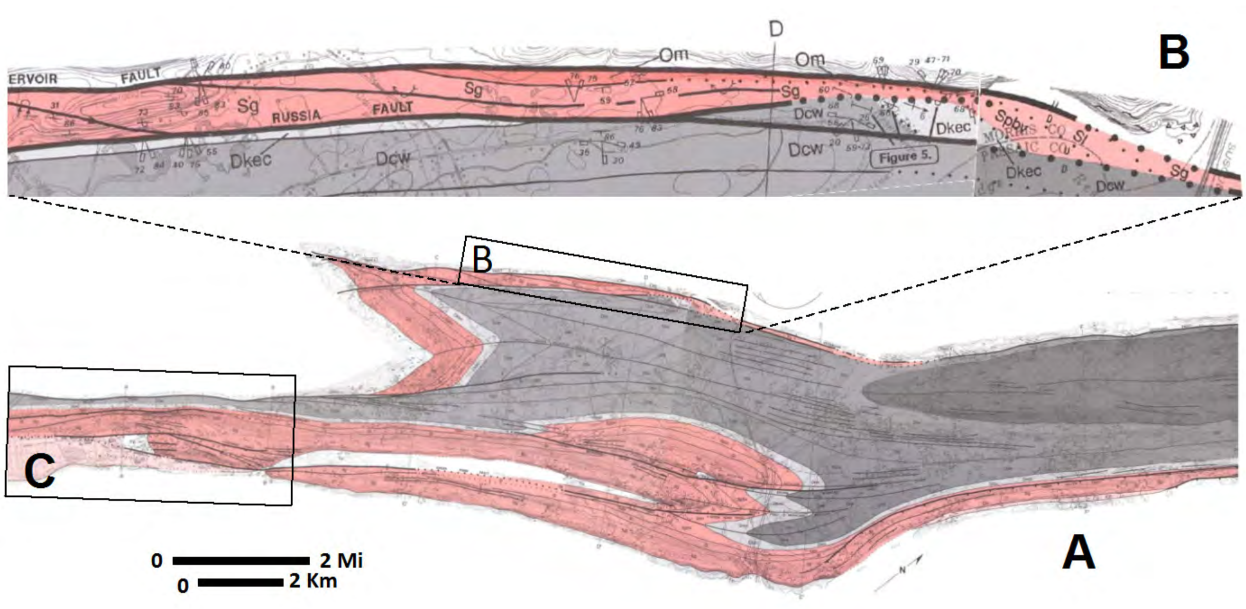

Figure 6. A. Geologic map of the central part of the Green Pond Mountain region adapted from Herman and Mitchell (1991). B. Inset map showing details along part of the northwestern boundary of the GPS where the Green Pond and Martinsburg Formations are mapped in a steeply northwest-dipping fold limb between two high-angle faults in the Reservoir fault system. The steeply-dipping limb contains is overturned steeply southeast just a short distance along strike to the west. The two shale occurrences mapped as Martinsburg are indicated by Om labels and leaders (top center of B.) C - Area C is shown in detail later in this report for the area of the US ARMY Picatinny Arsenal.

naming of the Shawangunk Formation, its stratigraphic equivalent in the Ridge and Valley Province that also contains abundant pebble conglomerate and varied alluvial-fluvial facies.

The Green Pond Mountain Region of New Jersey

The NJ part of the GPS historically has been referred to the Green Pond Mountain region and has been mapped over three time periods including the turn of the twentieth century (Kummel and Weller, 1902), around 1970 (Barnett, 1976), and again in the mid 1980s (Herman and Mitchell, 1991). The pioneering work of Kummel and Weller (1902) provides abundant stratigraphic and paleontological details, including observations of local strains such as fault attitude and slip motion.

The following excerpts are cited directly from Kummel and Weller's (1902) report on the geology of the Green Pond Mountain region:

> "The Green Pond Formation . . . consists of coarse, siliceous conglomerate, interbedded with and grading upward into quartzite and sandstones. The pebbles of the conglomerate range from one-half to three inches in diameter, and are almost entirely white quartz, but some pink quartz, black, white, yellow and red chert, red and purple quartzite and a very few red shale and pink jasper pebbles occur. The white quartz pebbles have frequently a pink tinge on their outer portion.
>
> "The quartzite . . . is interbedded in the upper portion of the conglomerate and rests upon it. It is in general a purple-red color, but presents various shades of pink,

yellow, brown and gray. Some of these beds are massive and show no laminae, but in others the thin stratification planes can be readily made out. The conglomerate beds are often very thick, with but slight trace of any bedding. In the southwestern part of the area, in the isolated hills southwest of the Rockaway River, the rock is much softer than farther north, and is friable sandstone rather than a quartzite. So completely disintegrated are some of these beds that they have been dug for sand and gravel for many years. This friable sandstone phase is well shown in the white rock cut on the D. L. & W. Railroad west of Port Oram and at the sand-pits in the vicinity of Flanders . . . The relationships of the conglomerate and quartzite to the older formations are not exposed in the isolated hills in the southwestern portion of the area."

"The relation of this formation to the overlying beds is simple. It passes upward somewhat abruptly into a soft red shale. Nowhere in NJ have the two been seen in actual contact, but they are frequently exposed in such close relationship as to render this conclusion a safe one."

"Various estimates have been made of the thickness of this formation. These range from 400 to 650 feet. All of these estimates, however, are believed to come far short of the actual thickness. In some cases they manifestly take into account only that part of the formation exposed in the steep eastward facing cliffs which characterize these ridges, and take no account of the higher beds which outcrop with steep dips on the back slopes of the mountain, and which add greatly to the thickness. In some cases, too, the small estimates may be due to an assumption that the ridges are formed by closely compressed folds. Our own estimates, measured on numerous section lines across the ridges, where at least the approximate portion of the enclosing formations were determined, and based on frequent observations of the angle of dip, indicate that the thickness of this formation is probably not less than 1,200 feet and locally it may be 1,500 feet. It is not asserted, however, that this entire thickness is exposed at any one locality, but we believe that these figures represent the thickness of the formation as developed along the greater portion of Green Pond and Copperas mountains. Toward the northern end of Kanouse mountain the thickness apparently diminishes somewhat; yet owing to the thick deposits of drift which conceal both the basal and upper portions, estimates of the thickness there may be somewhat in error. In the previous discussions of this region published in the Reports of the Survey, the conglomerate which occurs along Bearfort mountain was assumed to be the same as of Green Pond and Copperas mountains. Mr. Darton was the first to point out that this was an error, and the same conclusion was announced about the same time by Mr. Walcott. Our work corroborates completely the conclusions of these investigators in this respect. Although the conglomerates of Bearfort and Green Pond mountain resemble each other somewhat closely, yet critical examination of the two discloses at once marked lithological differences. These will be pointed out in connection with the description of the Bearfort conglomerate. Since the Green Pond formation rests unconformably in places upon the Lower Cambrian limestone, and perhaps upon Hudson River slate, and is overlain conformably by a red shale, which, as will be shown later, passes upward into a siliceous limestone containing Niagaran fossils, the correlation of the Green Pond formation with the Oneida conglomerate exposed in

Kittatinny mountain is probably correct. The lithological differences between the two are not so great as has been assumed by some observers. The lower beds of the Green Pond conglomerate are not infrequently of the same grey color and in almost every way identical with the conglomerate of Kittatinny mountain. Moreover, reddish conglomerates so common in the Green Pond rocks are not infrequent in the Kittatinny mountain beds. Although lithological resemblances and differences are not always safe guides for correlation, particularly in a formation which is so subject to variation as a conglomerate and sandstone, yet the structural position of the two is practically the stone, and there can be no question as to the correctness of this correlation, which was first announced by Merrill. He, however, included the conglomerate of Bearfort mountain as a part of the Green Pond formation."

Specific occurrences where the Green Pond Formation crops out in close proximity to older rocks in NJ are discussed below with respect to locations identified in Figure 4. The following details mostly stem from Kummel and Weller (1902) but are supplemented by work of Herman and Mitchell (1991).

Green Pond Formation over Middle Proterozoic crystalline basement (Figure 4, locations 1 and 2 for Copperas, Green Pond, Brown, and Bowling Green Mountains). – From Kummel and Weller (1902):

"The relation of this formation to the underlying rocks is readily determined, although only in one place has the actual contact been seen. Throughout the entire extent of Copperas mountain it rests unconformably upon the eroded surface of the crystallines, which form the lower part of the southeastern face of the mountain. At the mines opposite Green Pond the two formations are frequently exposed within twenty five or thirty feet of each other, although not in actual contact. Here the lowest conglomerate bed is rather gray in color and resembles closely the conglomerate of the Kittatinny and Shawangunk mountains. Along this face of the mountain the contact can be located definitely at an elevation of about 1,100 feet, or 225 to 250 feet below the crest"

"At Middle Forge, in the quarry west of the road (near location 2, fig. 4), the conglomerate of the Green Pond mountain apparently rests upon the Kittatinny limestone, but northward a quarter of a mile the conglomerate and gneiss are apparently in contact. West of the pond a fault evidently separates the high cliff of conglomerate from the Kittatinny limestone exposed in the quarry on the shore, both formations showing strong evidence of shearing and drag at outcrops nearest the hidden contact. Elsewhere along this mountain the conglomerate apparently rests upon the gneiss, and, although this contact is nowhere exposed, yet the two are shown in close proximity to each other at many places along the wild and narrow gorge of Green Pond brook, up which the gneiss can be traced continuously to about one-quarter of a mile southwest of the end of the pond where it is lost in the swamp. However, it reappears again a mile and a half east of the upper end of Green Pond forming a narrow bench fifty yards in width and several hundred yards long, immediately below the prominent summit of Green Pond mountain. Toward the lake the ledge becomes buried beneath the drift and to the northeast it disappears beneath the great blocks of talus, the dip of its contact with the overlying conglomerate and quartzite being such as to carry it beneath the surface within a short distance."

"The conglomerate is also seen to rest upon the gneiss in the offset of Green Pond mountain, southwest of Newfoundland, which is known locally as Brown's mountain. In Bowling Green mountain the conglomerate is wrapped around the northward end of a ridge of gneiss, and probably rests directly upon it; but the contact has not been seen, and nowhere have the rocks been found in such close proximity to each other as to eliminate beyond a doubt the possibility of a narrow strip of older sedimentary rocks between them."

Herman and Mitchell (1991) mapped Middle Proterozoic crystalline rocks with metamorphic layering dipping moderately to steeply southeast along the unconformity in the southeast-central and northeast part of the Green Pond Mountain region, whereas the Green Pond Formation dips moderately to steeply northwesterly. They also note that the contact of the conglomerate with the underlying basement rocks is not observed and is locally obscured by as little as 3 ft (0.9 m) of cover. The lack of pervasive tectonic strain fabric along the southeast syncline limb precludes a major structural contact, although limited shear strain associated with cover-layer folding is expected. These regional relations indicate that the Taconic unconformity is most pronounced in this area where the entire Lower Paleozoic section was eroded.

The basal conglomerate is coarsest to the east at Green Pond, Brown, and Kanouse Mountains where angular cobbles of shale and quartz are common, with some shale clasts measured up to 18 in (46 cm). To the west, at Bowling Green Mountain, the basal conglomerate contains mostly subangular to subrounded quartz-pebbles and is interbedded with quartzitic arkose and orthoquartzite.

Green Pond Formation over Cambrian-Ordovician Middle Proterozoic crystalline basement Gould's Quarry (Figure 4, Location 3). – From Kummel and Weller (1902):

"At Gould's quarry, large masses of the underlying limestone are included in a conglomerate, which is believed to be the basal layers of this formation. The matrix is comprised of quartz sand, is vitreous in texture and generally of a dull red color, but white, gray and greenish strata frequently occur, particularly in the basal portion, so that the formation is not so exclusively red as implied in most of the earlier reports. The beds are almost uniformly quartzitic in texture, and, on account of their hardness, form the long, narrow, steep-sided mountain ridges characterizing this region. Locally, however, the basal portion of the conglomerate is apparently quite friable and disintegrates readily, due probably to a greater or less amount of calcareous material derived from the limestone on which it rests in places. A good instance of this was found about 2 mi (3 km) north of Macopin lake, where the basal beds are so disintegrated that they have been dug for gravel".

Herman and Mitchell (1991) show bed strike in both the Lower and Middle Paleozoic units here are the same, but bedding dips more steeply (60^o to 70^o northwest) in units below the unconformity in comparison to those above (42^o to 56^o northwest).

Green Pond Formation over Middle Ordovician Martinsburg Formation along the Reservoir Fault (Figure 4, Location 4). – From Kummel and Weller (1902):

> "The outcrops of this formation southwest of Oak Ridge reservoir apparently rest upon a black shale, which may belong to the Hudson River formation, but no positive assertions can be made. Farther to the southwest they apparently abut against the crystallines and in the fault plane."

Barnett (1976) reports Ordovician brachiopods in this shale, resulting in his mapping them as Middle to Upper Ordovician Martinsburg Formation. There are two shale outcrops that are bounded on the southeast by fault slices of Green Pond pebble conglomerate (Figure 6). Worthington (1953) reported another occurrence of the Martinsburg Formation along the Reservoir fault farther southwest where the Green Pond Formation pinches out between Holland and Bowling Green Mountains. However, the black phyllite he described differs from the tectonized shales at Oak Ridge, and occurs with other anomalous rocks of uncertain affinity. Other tectonized sedimentary rocks that crop out along the trace of the Reservoir fault directly west of the Green Pond conglomerate show abundant stretched quartz grains included within a dark greenish-gray to dark reddish-brown phyllonitic matrix. Immediately to the southwest, and southeast across the trace of a subsidiary fault, the Green Pond Formation unconformably overlies a patchy strip of very low-grade metamorphic arkose and quartzite unlike other Middle Proterozoic basement rocks in the region. These rocks are similar to the Chestnut Hill Formation of Late Proterozoic (Z) age reported in the southwest NJ Highlands (Drake, 1984).

The Green Pond Formation in Orange County, New York

The Green Pond Formation Is mapped in three areas in Orange County, NY, covered by the Monroe, Lake Popolopen, and Cornwall-on-Hudson 7-½ minute quadrangles (Figures 2, 3 and 7). These locations are specifically discussed below with respect to Lazy Hill, Pine Hill, and two bedrock ridges near Pea Hill, respectively (Figure 2). Lazy Hill lies about 2 mi (3 km) west of Monroe on the western side of Bellvale Mountain, and the western limb of the syncline (Figures 2 and 7). Pine Hill lies immediately east of Highlands Mills, NY on the east side of the syncline where one long, northeast-striking, thin bedrock ridge rises about 200 ft (61 m) above base elevations along Skyline Drive (Figure 7). The ridges near Pea Hill are more than 1 mi (1.6 km) west of Cornwall, NY within the Cornwall-on-Hudson quadrangle and at the very northeast tip of the GPS. These are a little less conspicuous, rising only about 120 ft (37 m) above base elevations.

Lazy Hill and Fault Blocks in the Monroe Quadrangle

The bedrock geology map of the Monroe quadrangle provides the only record of detailed structural readings of strata bracketing the unconformity in the NY part of the GPS (Jaffe and Jaffe, 1973 and Figure 7). Basal Silurian conglomerate and quartzite are mapped just west of Monroe in a series of fault blocks that are referred to here as south, central, and north (Figure 7). The south and central blocks flank Bellvale Mountain, whereas the north block flanks Schunemunk Mountain (Figure 3). The unit is mapped as Shawangunk Formation, and is described as green-gray to white and buff-colored orthoquartzite (25%) and conglomerate

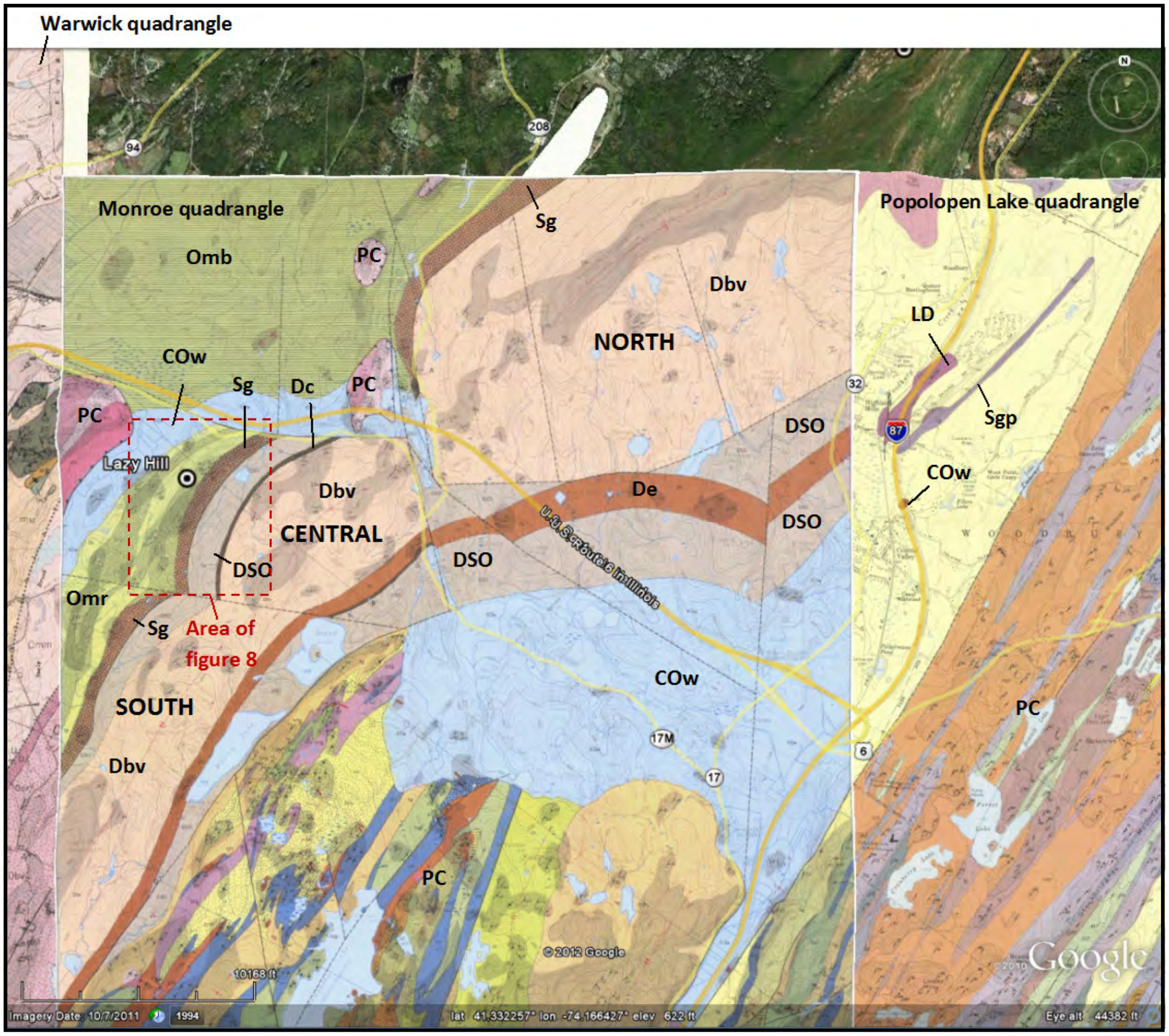

Figure 7. Composite display of three bedrock geology maps (Offield, 1967—left; Jaffe and Jaffe, 1973—center; and Dodd, 1965— right) registered in Google Earth that cover parts of the Green Pond Syncline in New York State. The Shawangunk Formation (Sg) is mapped to the west whereas the Green Pond Formation (Sgp) is mapped to the east. The Shawangunk is extended north of the Monroe quadrangle boundary based on the topographic expression of the associated hill. The relationship of the Taconic unconformity on the west side of the syncline is discussed in the text with respect to south, central, and north fault blocks in the Monroe quadrangle. Abbreviations: Dbv—Devonian Bellvale Formation; COw – Cambrian-Ordovician Wappinger Group (dolomite); Dc – Connelly Conglomerate; De – Esopus Formation; DSO - concealed Silurian, Devonian, and Ordovician strata; LD – Lower Devonian sedimentary rocks; Omr – Ordovician Martinsburg calcareous shale and quartzite; PC – Precambrian rocks.

(75%) consisting of coarse white pebbles of milky vein quartz in a matrix of fine pebbles and grains of rounded quartz.

According to Jaffe and Jaffe (1973):

> "A 10-meter section measured across the top of Lazy Hill, shows from west to east; about 1 meter of gray-buff orthoquartzite, 2.5 meters of coarse white pebble conglomerate, 4.5 meters of white orthoquartzite with ripple marks on bedding planes, and 1.25 meters of finer white quartz pebble conglomerate . . . Toward the

eastern edge of Lazy Hill scarp, the conglomerate carries some coarse orthoclase pebbles and unit grades into a red arkosic conglomerate below the ridge top . . . Small slabs of red argillaceous sandstone were found at localities near the base of the eastern edge of lazy Hill scarp and suggest a gradation to the Longwood Shale."

Pebbles in the conglomerate are about 0.5 to 4 in (1.3 to10 cm) long, and are strongly elongated (stretched) parallel to the syncline fold axes. The rocks are reported as being shattered, sliced, and veined. They map Shawangunk Formation overlying calcareous shale and quartzite of the Martinsburg Formation, thereby having the same stratigraphic relationships as seen about the Taconic unconformity at Kittatinny and Shawangunk Mountains about 15 mi (24 km) across the Great Valley to the northwest. But the Martinsburg at Lazy Hill is comparatively more deformed than its counterpart to the northwest, with local, southeast-verging asymmetric folds and steep, northwest-dipping faults (Jaffe and Jaffe, 1973). The Shawangunk designation was used for these rocks because they closely resemble those in Shawangunk Mountain where they are white- and buff-colored rather than the grayish-purple to grayish-red color of Green Pond Formation on the eastern side of the syncline at Pine Hill, NY (Figure 7) and in the central Green Pond Mountain region of NJ. The unit thickness in the Lazy Hill area is portrayed by Jaffe and Jaffe (1973) as about 400 to 600 ft (122 to 183 m) thick based on their cross-section interpretations and calculations based on their outcrop widths and dip angles. The Shawangunk Formation of Jaffe and Jaffe (1973) will be referred to as the Green Pond Formation for the remainder of this report in accordance with the nomenclature discussion above.

In the southern and central fault blocks, the Green Pond Formation forms ridge scarp peaking at 700 to 800 ft (213 to 244 m) elevations in comparison to 900 ft (274) crest elevations of nearby Bellvale Mountain to the east. The scarps are shown as being cut by concealed cross faults striking about N95^{o} E, having little offset of cross-cut Ordovician through Devonian strata, and are thus not included on the regional maps compiled here (Figures 1 and 2). In the southern fault block, the Green Pond Formation is mapped as overlying Martinsburg Formation calcareous shale dipping north-northeast to south-southeast at 20^{o} to 50^{o}, but there are no structural readings mapped close to the unconformity, and the angular relationship between the two units here is unknown. However In the central fault block, the Green Pond is mapped near outcrops of Ramseyburg calcareous quartzite (Figure 8), and the angular unconformity is characterized along two traverses across the crest of Lazy Hill (Figure 8). Both formations strike parallel (northeast-southwest), but the Martinsburg dips gently eastward 25 to 16^{o} beneath the Shawangunk, that is mapped having moderate eastward dips of 52^{o} to 40^{o}. Along the northern traverse, there is about a 10^{o} difference in northeast-southwest strike, but the formation dips are the same (60^{o} southeast).

The Green Pond Formation crops out sparingly in the North block on the westward-facing hillslope at about 600 ft (183 m) elevation, with peak elevations of Schunemunk Mountain in the 1,300 to 1,400 ft (396 to 427 m) range. In all three blocks, it's mapped as being fault-bounded on the eastern side. For the south and central blocks, undivided and concealed rocks of Ordovician to Devonian age are mapped directly east of the bounding fault. The Lower Devonian Connelly Conglomerate is mapped about 1,300 ft (396 m) to the east of the Green Pond Formation, and adjacent to the concealed unit in the central block. It is also mapped as a fault sliver at the southern end of the ridge scarp in the north fault block (Figure 7) where Middle Devonian Bellvale Sandstone is otherwise mapped directly east of the bounding fault.

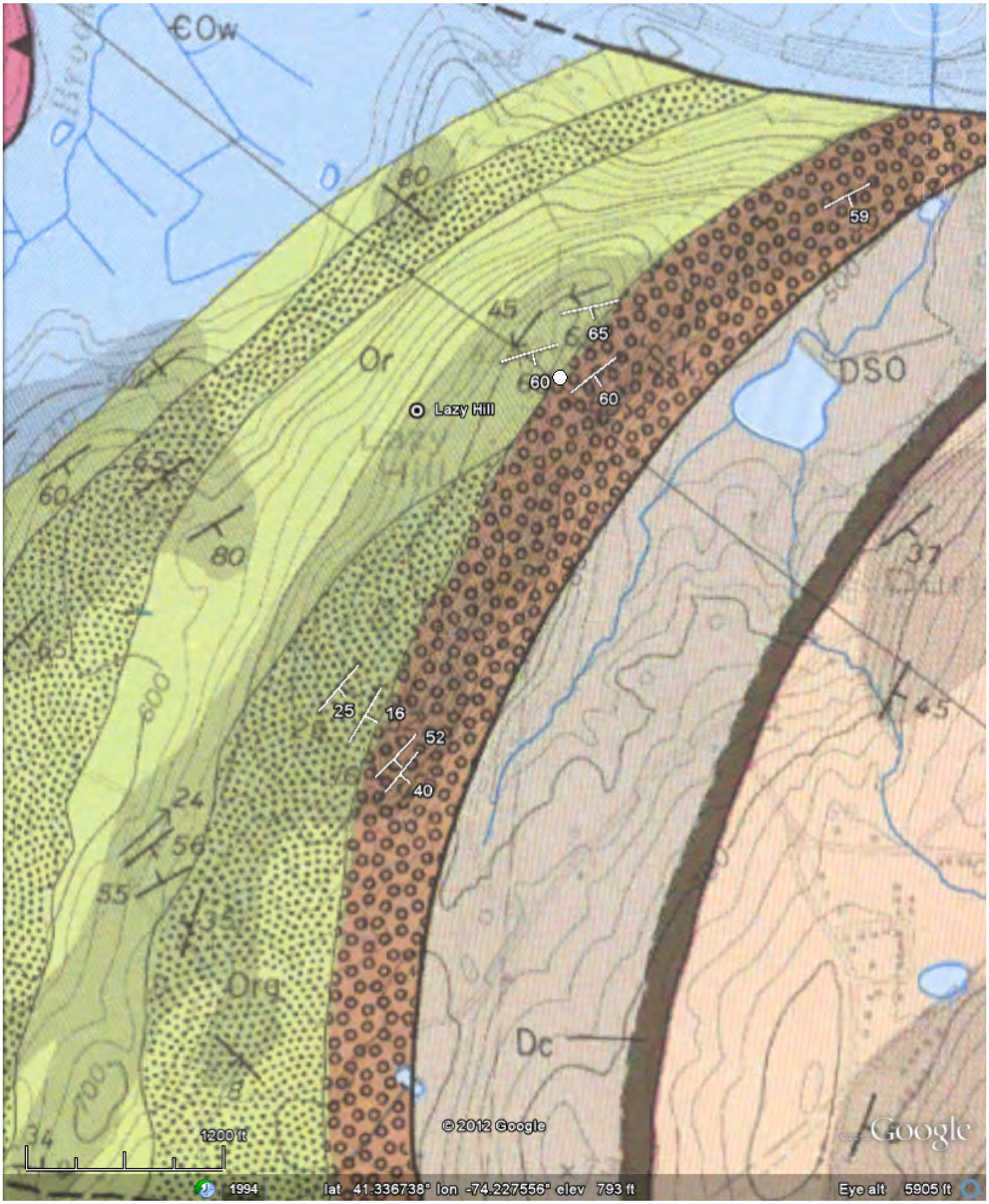

Figure 8. Structural details in the central fault block on the western limb of the GPS where the Taconic unconformity is mapped by Jaffee and Jafee (1973). Map location shown in Figure 7. Bedding strikes and dips for the two traverses across Lazy Hill are emphasized. SSk - Shawangunk Formation (referred to in the text as the Green Pond Formation), COw – Cambrian-Ordovician Wappinger Group (dolomite), Dbv – Devonian Bellvale Formation, Dc – Connelly Conglomerate, De – Esopus Formation, DSO – concealed Devonian, Silurian, and Ordovician sedimentary rocks, LD – Lower Devonian sedimentary rocks undivided, Omr – Ordovician Martinsburg calcareous shale and quartzite, PC – Precambrian rocks. Location of pictures on Figure 8 indicated by the solid white dot.

This distinction is interesting because there is enough distance between the Green Pond Formation here and the superjacent Middle Devonian Bellvale Formation to accommodate the sequence of Middle Silurian strata mapped above the Green Pond Formation elsewhere in the GPS. In NJ, the combined stratigraphic thickness of the Middle and Upper Silurian units occurring between the Green Pond Formation and the Connelly Conglomerate is about 800 ft (244 m) (Herman and Mitchell, 1991). The concealed interval in the central fault block can accommodate an 800-ft. section dipping at about 40^{o}. The closest Green Pond outcrop dips 59^{o} and the closest Connelly outcrop dips 32^{o} for an average of about 45^{o}. It is therefore possible that the Green Pond Formation is not fault-bounded on its eastern side, and for that matter, may not be fault-bounded along its entire length west of Schunnemunk Mountain where outcrops are seemingly scarce and the questionable interval is concealed by thick surficial deposits (Jaffe and Jaffe, 1973).

Don Monteverde, Jack Epstein, and I travelled to the central fault block on the west limb of Bellvale Mountain on June 28, 2012 with the hope of documenting the unconformity based on the mapping of Jaffe and Jaffe (1973). We targeted the northernmost of two traverses that they mapped across Lazy Hill in the central fault block where a power-transmission line provides access off NY Rt 17M at the base of Lazy Hill up to its crest (Figure 8). We hiked up approaching from the north, passing over pavement outcrops showing rhythmic cycles of shale, siltstone, and greywacke sandstone of the Martinsburg formation. At the location of the northernmost traverse of Jaffe and Jaffe (1973; Figure 8), there's a prominent ridge of white pebble conglomerate cropping out immediately east of the transmission line that overlies a quartzite that is about 3.3 ft (1 m) thick along the western base and southern tip of the ridge (Figure 9). Here, white pebble conglomerate sits atop siliceous, light-brown to gray, medium- to coarse-grained uartzite that is lcally thin-bedded. It was nonreactive with dilute hydrochloric acid on fresh surfaces, and it was difficult to tell if we we're looking at or Martinsburg well-cemented subarkosic sandstone or Green Pond quartzite, because both units are penetratively strained and cut by slickensided shear planes that locally offset and complicate their contact (Figure 10). But underneath a small overhang at the southern tip of the

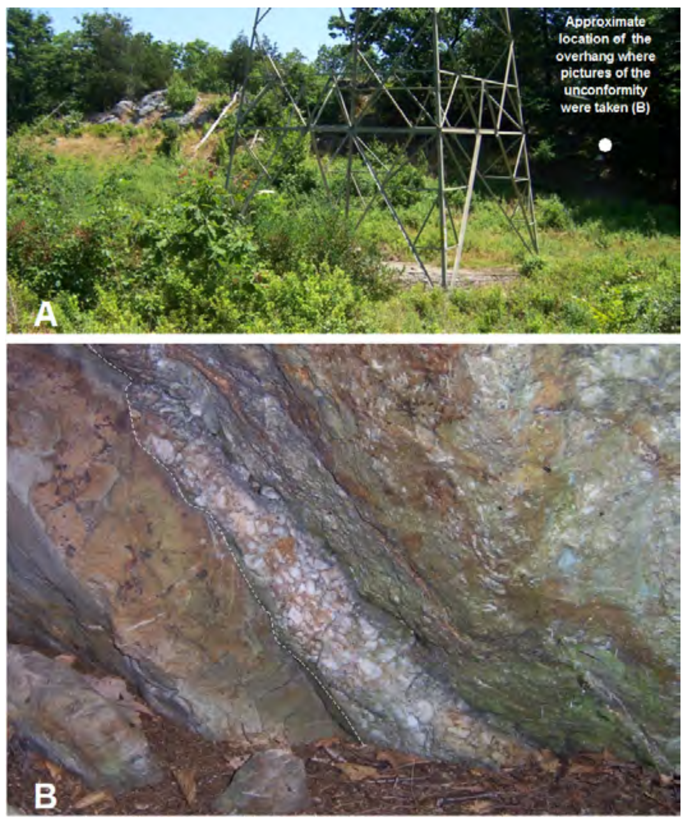

Figure 9. View of the Shawangunk Formation cropping out on a ridge atop Lazy Hill. See Figure 8 for the picture location. Photograph A is a southeast view from the trail towards the spine of the Hill. The solid white dot on the right side is the approximate location of the overhang where the unconformity crops out. Photograph B shows the overhang outcrop where white-pebble conglomerate of the Shawangunk Formation lies atop a subarkose sandstone that may be the Martinsburg Formation. Some of the strain features associated with these rocks are detailed in Figure 10, and discussed in the text. Photographs by Jack Epstein.

ridge, white pebble conglomerate sits directly on the quartzite, with very little divergence in strike and dip between the two units; the average strike/dip for the sandstone was N60°E/18°S compared to N73°E/24°S in the superjacent conglomerate. We walked the contact between the quartzite and conglomerate along strike for about 20 ft (6m) at the base of the ridge heading northeast from the overhang. The conglomerate closely resembles Green Pond conglomerate mapped along the Reservoir fault in NJ (Herman and Mitchell, 1991). In both places it contains milky-white, subangular to subrounded, vein-quartz pebbles ranging from an averaging diameter of about 1 to 2 in (2.5 to 5 cm), up to about 6 in (15 cm). The rocks along the Reservoir fault are more highly stretched and fractured in comparison to those here, but the abundance of mineral veins and stratigraphic slip and wedging at this location indicates significant Alleghanian strains here as well. The Martinsburg Formation appeared to coarsen upward towards the upper contact, but there was a significant covered interval between the Martinsburg outcrops near the trail and the base of the ridge. It is likely that the lower quartzite here is the 3-ft (0.9-m) thick quartzite at the base of the Green Pond (Shawangunk) Formation mentioned by Jaffe and Jaffe (1973). The search for the unconformity at the more southerly traverse (Figure 8) was not attempted due to time restrictions.

Pine Hill, Lake Popolopen Quadrangle

After hiking down Lazy Hill, we drove to the base of the ridge east of Highlands Mills, NY, along the base of Pine Hill in the Lake Popolopen quadrangle where the Green Pond Formation was mapped by Dodd (1965 and Figure 7). We drove into a new housing tract being built along the east side of the ridge base, and we stopped to inspect nearby boulders. The Green Pond here is the same polymictic, grayish purple and grayish red conglomerate that crops out in the eastern and central parts of the NJ Green Pond Mountain region. It has varicolored sedimentary gravel and cobbles along with abundant white, vein-quartz pebbles. The nature of the lower contact here is unknown, and it's difficult to tell whether this ridge is bounded by a fault on the east, as the ridge is separated from outcrops of Cambrian-Ordovician Wappinger Dolomite to the southeast by a large expanse of alluvium (Dodd, 1965). No description of the Green Pond is provided by Dodd (1965) as their primary focus was on the Precambrian geology. But this sequence was studied by Southard (1960) as part of a senior thesis that provides details of the lithological facies here and in the ridges near Pea Hill (Figures 3 and 11). He reports that the Green Pond Formation is nonfossiliferous and therefore of uncertain age, but probably the stratigraphic equivalent of the Shawangunk conglomerate to the west. He divided the Green Pond Formation into five subunits, four of which are quartzite that occur mostly in the eastern section. At Pine Hill he recognized an upward sequence of conglomerate, quartzite, and sandstone that is about 400 ft (122 m) thick. The conglomerate is about 300 ft (91 m) thick and is coarsest at its base where the largest pebbles (4.5 in [11 cm] in diameter) fine upward to white-quartz pebbles 1 to 2 in (2.5 to 5 cm) in diameter. The conglomerate matrix is coarse to very coarse sand that is red and yellow in color. Conglomerate bed thickness ranges from 0.5 to 15 ft (0.15 to 5 m). Conglomerate beds contain sandstone intervals that are less continuous along bed strike where they pinch and swell. The quartzite units are medium- to thick bedded, with varieties ranging in color from light-red to white, and totaling about 55 ft (17 m) thick. The upper sandstone units are more poorly exposed, range in thickness from about 5 to 15 ft (1.5 to 5 m), and show a variety of lithological textures and color variations. Red sandstone units contain minor red shale and siltstone, cross-bedded, and laminated varieties that range in

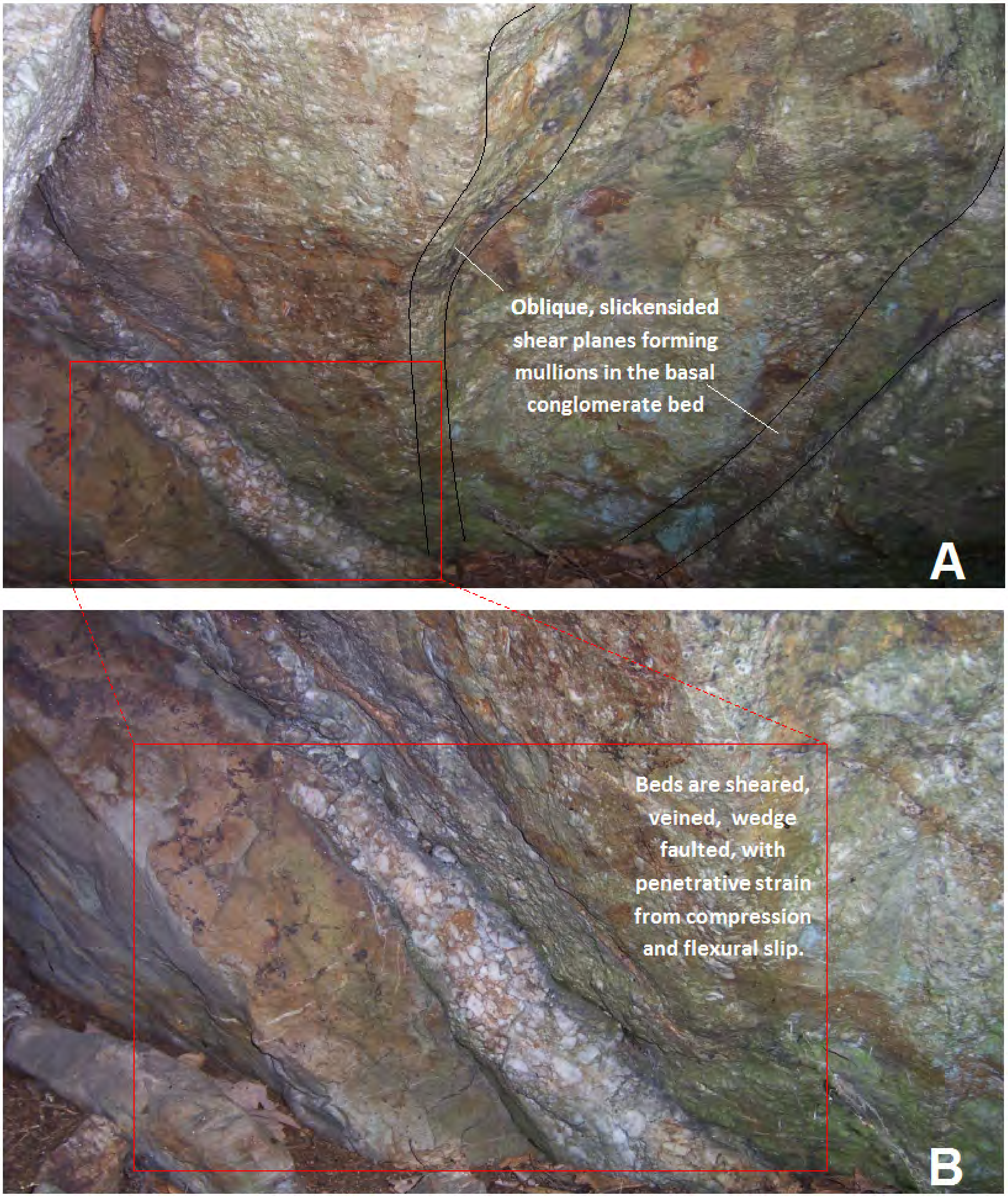

Figure 10. Detailed views of the rocks near the Taconic unconformity atop Lazy Hill. Photograph A is a northeast view of beds beneath an overhang. Beds dip left-to-right at about 20°-25° E. Slickensided shear planes form mullions (Epstein and Lyttle, 1987) at the base of a conglomerate-bed. Photograph B shows details about a contact between gray quartzite (left) and the suprajacent white-pebble conglomerate. Mineral-vein arrays and small faults locally offset the contact. The beds are penetratively strained and sheared along contacts. Photographs by Jack Epstein.

color from purplish white to purplish red and locally contain rip-up clasts, and ripple marks. This sequence at Pine Hill was also mentioned in earlier work of Darton (1894a) and Ries (1895). At that time, Ries (1895) referred to the Green Pond Formation as the Medina formation, including Oneida conglomerate at its base, with the quartzite and sandstone as upper members of the Medina formation.

The Ridges near Pea Hill, Cornwall-on-Hudson Quadrangle

At the northern end of the GPS just west of Cornwall, NY, Darton (1894a) reported conglomerate overlying Hudson Shale (Martinsburg Formation). He depicts their arrangement in a moderate- to steeply-northwest-dipping succession of Silurian to Devonian strata with Oneida conglomerate along its eastern face. Southard (1960) also studied this sequence where he reports basal conglomerate and quartzite units like those at Pine Hill, but lacking the upper sandstone unit. Ries (1895) described the east side of the eastern ridge as being formed by "coarse-grained red siliceous conglomerate (Oneida), with red sandstones and shales of Medina age". Darton (1894a) reported about 25 ft (8 m) of conglomerate and sandstone here.

The ridge locations near Pea Hill are tentatively mapped in Figure 11 based on Darton's (1894a) mapping as reported by Ries (1895). The Green Pond Formation was sketched in Google Earth as a pair of polygons corresponding to conglomerate noted on Darton's map (Figure 11B). Registration of the image proved challenging because the map is old, and contains sketched roads, railways, and a stream. The proposed alignment primarily uses two, linear topographic ridges and Pea Hill as reference features in Google Earth (Figure 11). Darton (1894a) depicted the structure and stratigraphy associated with the Green Pond (Medina formation) in the eastern ridge, but no details surrounding the western ridge were given. It is possible that the ridges may define a southwest plunging syncline axis of the GPS near its northeastern tip, but more detailed mapping is needed to verify it.

A New Subsurface Record of the Unconformity in the Green Pond Mountain Region from Picatinny Arsenal

A subsurface stratigraphic contact interpreted to be the Taconic unconformity was recently photographed as a digital borehole televiewer (BTV) optical image collected within the U.S. Army military reservation in Morris County, NJ, known as Picatinny Arsenal (Figures 3, 4, 12, and 13). In 2010 and 2011, the NJ Geological Survey was asked by Mr. Joseph Marchesani of the NJ Department of Environmental Protection Site Remediation Program to review optical BTV data collected by geophysical companies contracted by the US Army as part of on-going site characterization and remediation work at two groundwater-investigation (GWI) sites within the reserve (Figure 12). The 600 Area GWI site is underlain by the Green Pond Formation, whereas the Mid-Valley GWI site is underlain by Precambrian gneiss and granite lying immediately southeast of the GPS (Figures 4 and 13). The BTV surveys were provided as paper reports and used primarily used to determine the orientation of permeable stratigraphic layering and secondary brittle structures (fractures and faults) penetrated by the wells. The details of the BTV study of the Proterozoic rocks is beyond the scope of this work, but the orientation of metamorphic layering and folding geometry determined from a structural analysis of Mid-Valley area is incorporated into a cross-section interpretation below.

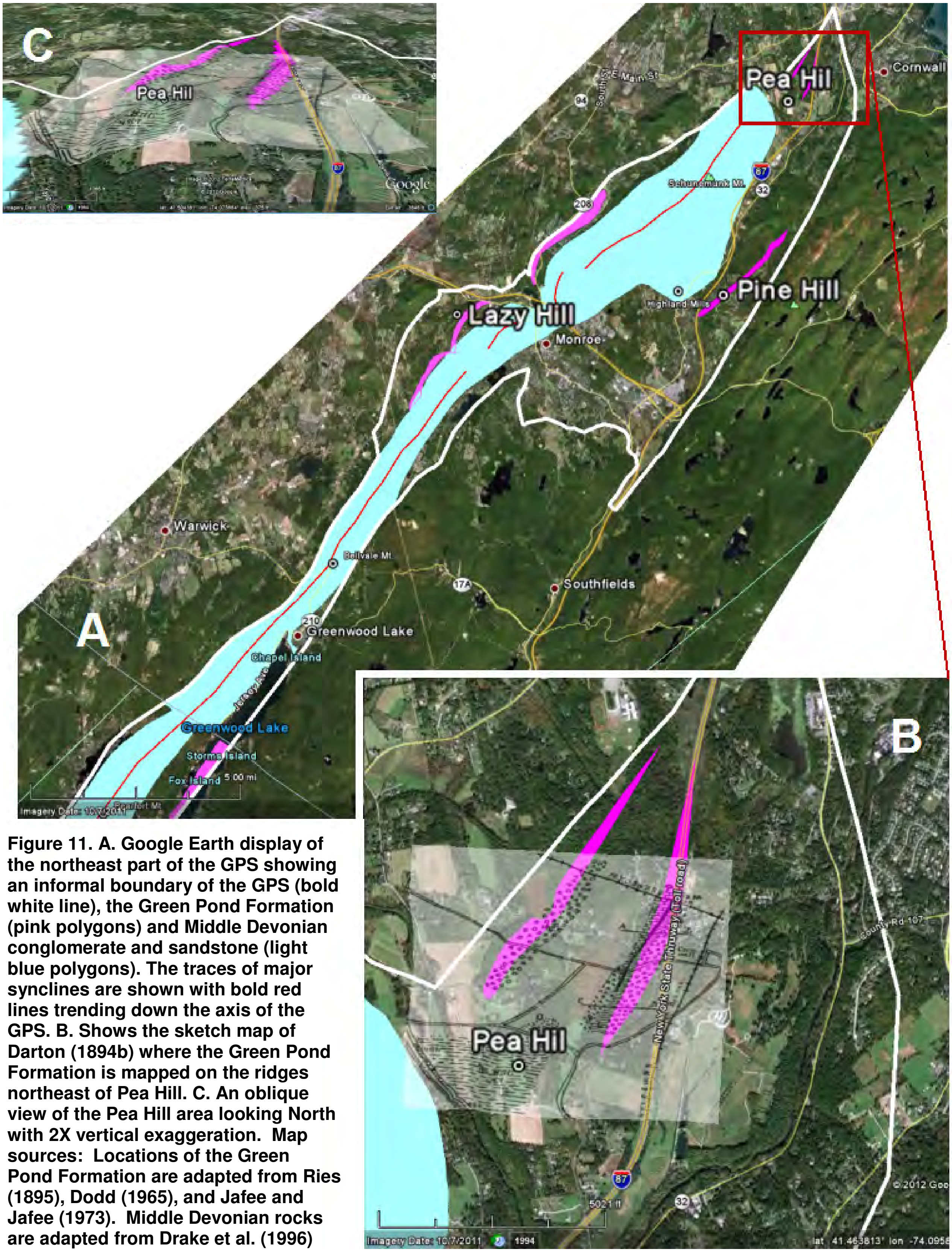

Figure 11. A. Google Earth display of the northeast part of the GPS showing an informal boundary of the GPS (bold white line), the Green Pond Formation (pink polygons) and Middle Devonian conglomerate and sandstone (light blue polygons). The traces of major synclines are shown with bold red lines trending down the axis of the GPS. B. Shows the sketch map of Darton (1894b) where the Green Pond Formation is mapped on the ridges northeast of Pea Hill. C. An oblique view of the Pea Hill area looking North with 2X vertical exaggeration. Map sources: Locations of the Green Pond Formation are adapted from Ries (1895), Dodd (1965), and Jafee and Jafee (1973). Middle Devonian rocks are adapted from Drake et al. (1996) and Rickard et al. (1970)

The unconformity was penetrated by the AWDF well within the 600 Area (Figure 14) at about a depth of about 419 ft (128 m) after passing through Green Pond conglomerate, sandstone, siltstone and shale. Below the unconformity, higher reflective, indurated, and fractured beds are interpreted to be the Cambrian Hardyston Quartzite (Figure 14). The contact between the two units is unremarkable and somewhat diffuse, but is picked at a depth of 418.65 ft (127.6 m) below ground surface (Figure 14). Figure 15 shows a statistical analysis of beds

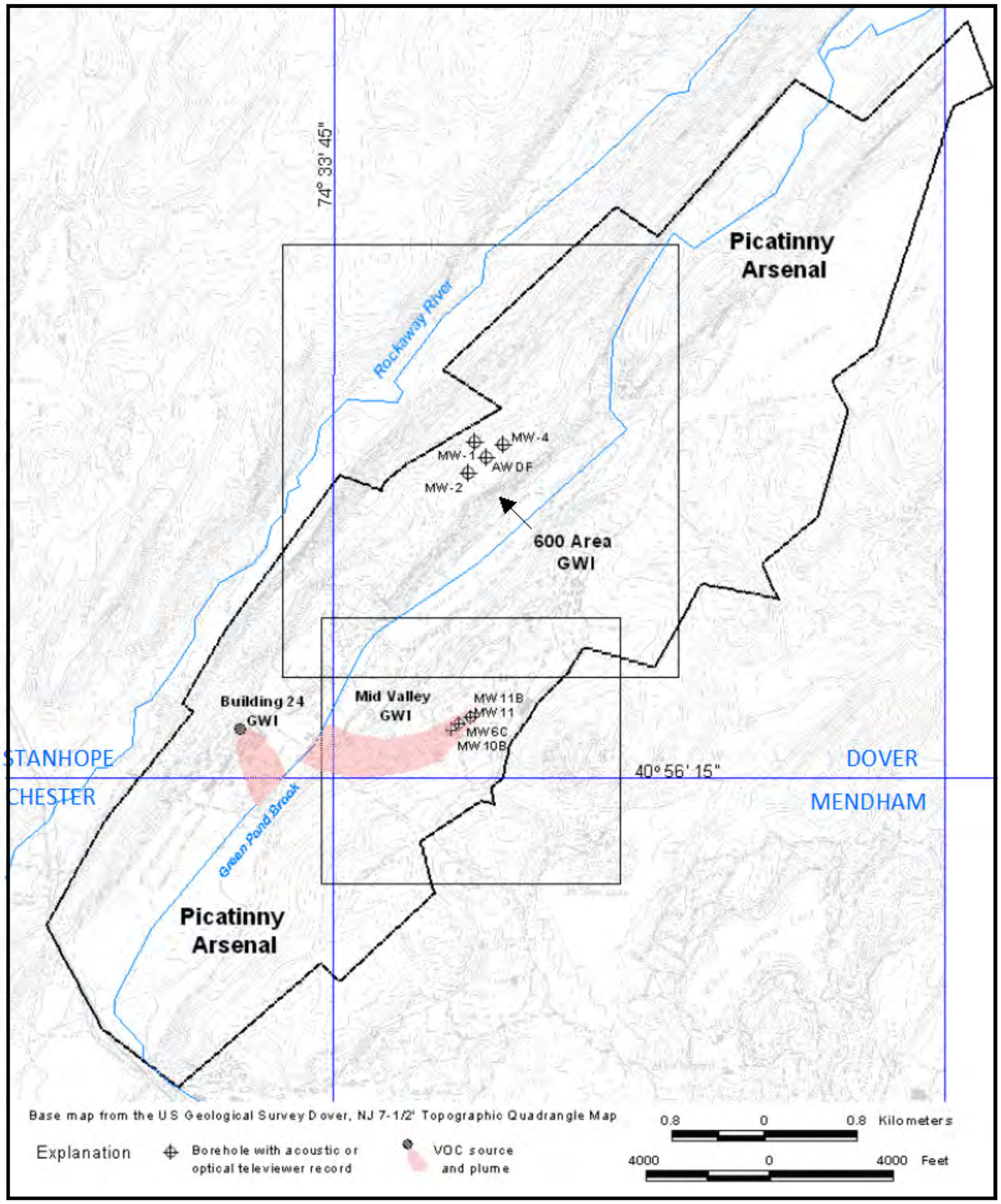

Figure 12. Location map of Picatinny Arsenal showing the locations of three groundwater investigation (GWI) sites and four USGS 7-1/2' topographic quadrangles. This paper focuses on the 600 Area GWI. The boxes correspond to the coordinate limits of digital elevation model (DEM) grids that were generated for two groundwater investigations. The unconformity was penetrated at a depth of about 419 ft (128 m) in well ADWF.

and fractures bracketing the unconformity. Figure 16 summarizes the well depths and average bed dips determined for each well from the BTV records.

The BTV images show the Green Pond Formation as a heterogeneous mix of olive green and grayish brown pebble conglomerate, sandstone, siltstone, and shale. The rock colors are described as they appear in the paper report and should be regarded with caution as they probably don’t capture true colors, and the type of borehole imaging tool and data acquisition and processing parameters are unknown. But compositions and textures seen in the borehole photos facilitate distinction between conglomerate, sandstone, siltstone, and shale. The reports were only available for a brief time during the file review, which was more focused on the structural interpretation than on the stratigraphic characterization. Consequently, only a general stratigraphic review of the record was conducted and not a detailed lithological summary. As noted in the bureau review conducted then, the coarsest conglomerate beds reach 4 to 6 ft (1.6 to 1.8 m) thick, whereas the shale beds are less than 1 ft (0.3 m) thick. Sandstone beds commonly range in thickness from less than 1 ft (0.3 m) to about 2 ft (0.6 m). Beds in the basal part of the Green Pond Formation are generally finer grained than beds higher in the section (Figure 14). The Hardyston is composed of tan, brown and bluish-white, thin-to medium bedded, fine to coarse sandstone or quartzite. This unit is not folded and lacks the metamorphic and igneous textures that are seen elsewhere in Middle Proterozoic rocks. However, this lower unit may have an affinity with other anomalous metasedimentary rocks of probable Late Proterozoic age like those mentioned earlier along the Reservoir fault (Herman, and Mitchell, 1991).

Some beds in the Green Pond Formation are highly fractured whereas others are not. Bed-parallel fractures are common, as are steeply-dipping extension fractures that locally show complex banding, and therefore multiple generations of tensile opening and mineralization. Some of the steeply-dipping extension fractures show normal dip-slip movement. Most of the fractures show alteration rinds and adjacent staining of the unit matrix by groundwater infiltration and movement. The chemistry of the fracture and matrix alteration and staining is unknown. Shear fractures seem to be more common in the finer-grained beds where local wedges in the strata probably reflect bed-parallel shear strain.

The BTV data were also used to refine the bedrock geology map (Figure 16) and to construct a new cross section through the area (Figure 17). The central part of 600 Area lies between the Green Pond and Picatinny reverse faults (Figures 13, 16 and 17). Herman and Mitchell (1991) show the Picatinny fault dipping steeply southeast and the Green Pond fault dipping steeply northwest. The BTV analyses confirm that the beds near wells MW-2 and AWDF straddle a southwest gently plunging, upright, and open anticline with limbs that dip gently northwest and southeast (Figures 15 to 17). The stratigraphic interpretation of the unconformity agrees with the existing cross section interpretations that depict a thin veneer of Hardyston Quartzite overlying Precambrian basement and underlying the Green Pond Formation (Herman and Mitchell, 1991).

3D diagrams of well-field components and bedding and fracture planes for the 600 Area wells were also produced as part of the file review. Some of these diagrams are included here to further convey the subsurface BTV results (Figures 18 and 19). The maps and 3D diagrams were generated and displayed using ESRI ArcView 3.2a software. The ArcView Spatial and 3D Analyst extensions were used for clipping digital elevation models (DEMs) and the 3D Analyst extension was used with the NJGS 3D well field visualization extension to generate and

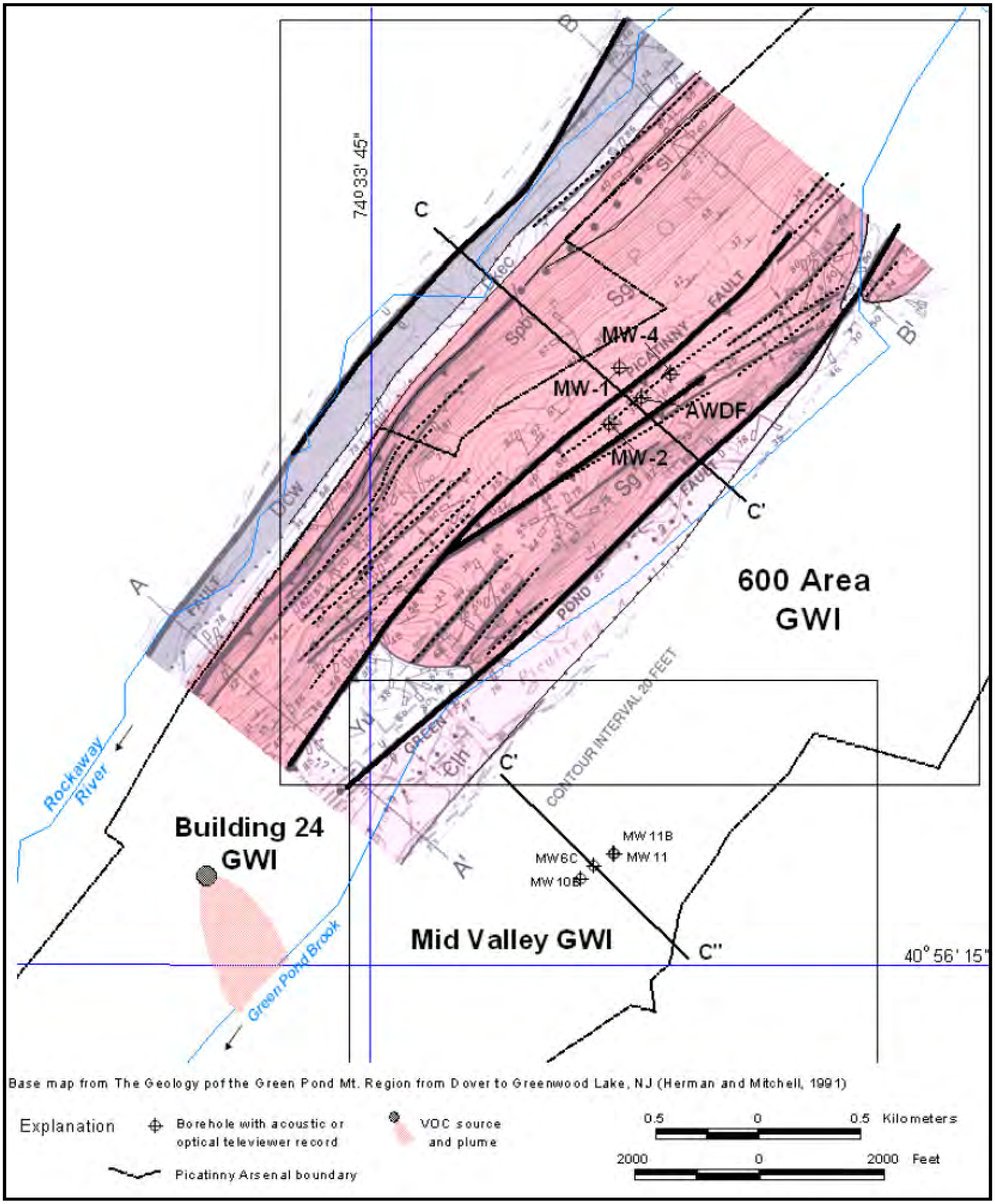

Figure 13. Part of a bedrock geological map covering the 600 Area GWI (Herman and Mitchell, 1991) was scanned and georegistered in ArcView GIS in order to refine the structural interpretation near the 600 Area based on the BTV data. The cross section interpretation C-C" is shown in Figure 17. The northwest part (C -C') lies between sections A-A' and B-B' " and edge matches to the section through the Mid Valley are (C'-C"). Note that the fault locations remain the same, but folds axes in the 600 Area have been slightly modified from those of Herman and Mitchell (1991) based on the BTV data.

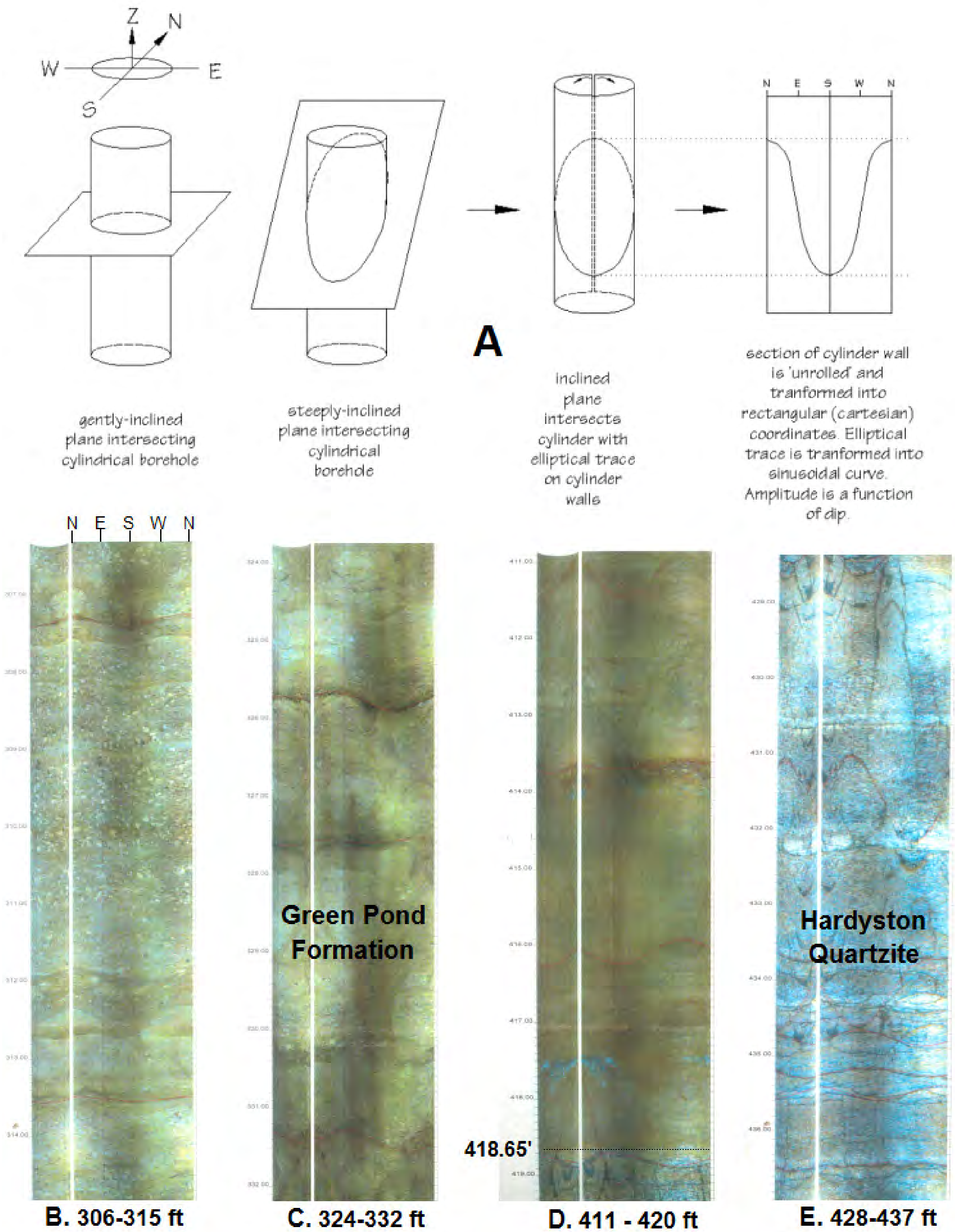

Figure 14. A. Schematic diagram illustrating how BTV images are "unrolled" and flattened for display and interpretation. B through E are optical BTV records of specific depth intervals in the well that were digitally scanned from a report by Mid-Atlantic Geoscience, LLC. Each BTV record includes a wrapped virtual core on the left showing just a sector of the well from a specific viewpoint, and a flattened, unwrapped image on the right showing the unwrapped borehole wall. Images B and C show pebble conglomerate and coarse to medium sandstone in the Green Pond Formation. Image D shows a stratigraphic contact between the Green Pond Conglomerate and the subjacent Hardyston Quartzite at about 419' depth. Image E shows a section of the Hardyston Quartzite starting about 10 ft (3 m) below the contact.

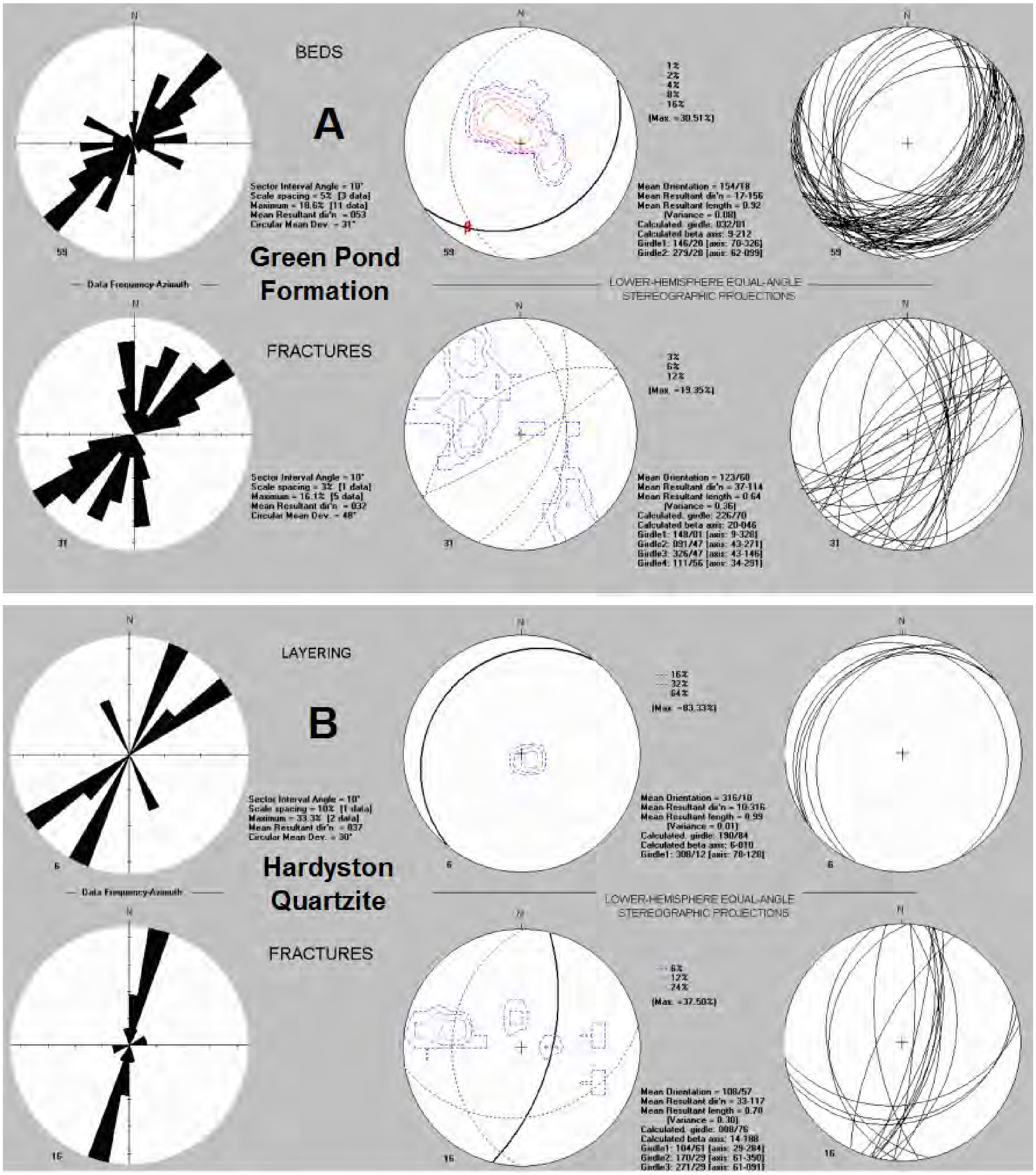

Figure 15. Structural analyses of interpreted BTV record of the AWDF well showing the orientation of beds and fractures bracketing the unconformity. A. Structures in the Green Pond Formation. B. Structures in the Hardyston Quartzite.

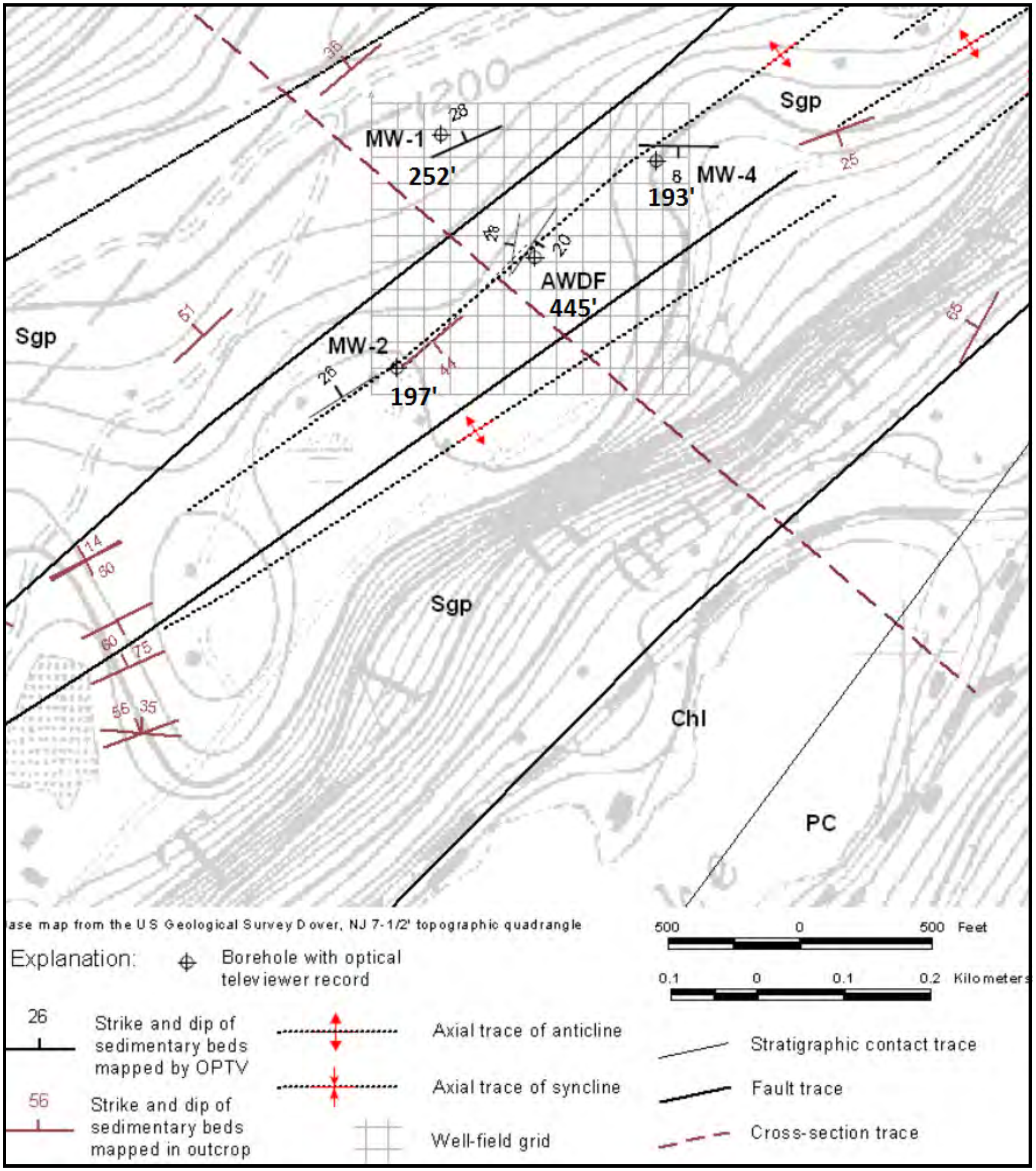

Figure 16. Geologic map near the 600 Area GWI showing a well-field grid. The AWDF well penetrates two limbs of an open, upright anticline in the Silurian Green Pond Formation before encountering the Taconic unconformity and Cambrian Hardyston Quartzite at 419 ft (128 m).

visualize 3D shapes of the boreholes, oriented structural planes, and well-field grid based on the BTV survey (N.J. Geological Survey, 2001). The borehole shapes used well-location and construction parameters taken from the consultant's report. The well and plane shapes were generated using a vertical borehole alignment because no borehole-deviation information was included in the reports.

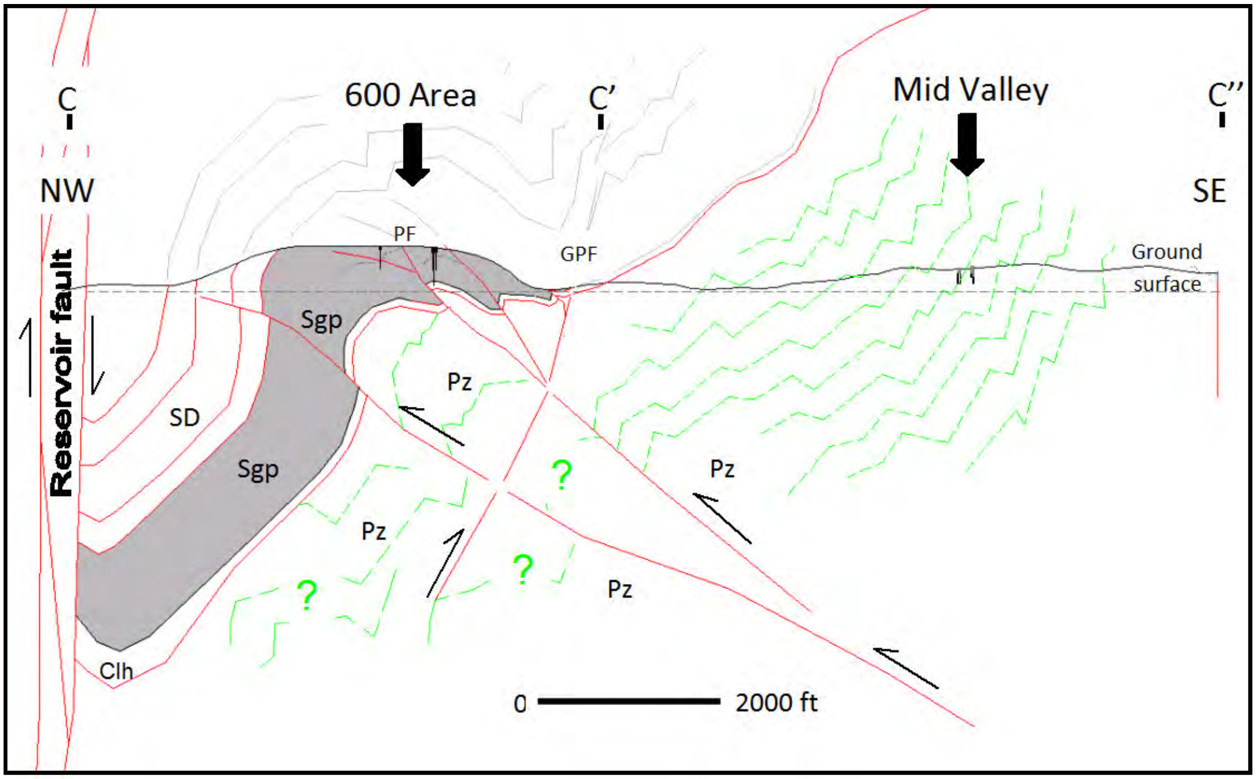

Figure 17. Cross section C-C-C" is based on earlier sections of Herman and Mitchell (1991) and the BTV analyses at the 600 Area and Mid-Valley sites (Figures 12 and 13). The section traces are shown on Figure 12. Abbreviations: Clh – Cambrian Leithsville and/or Hardyston Quartzite, Pz –Proterozoic gneiss and granite, Sgp – Silurian Green Pond Formation, SD – Undivided Silurian and Devonian rocks. GPF – Green Pond fault, PF – Picatinny fault.

Figure 18 shows that the upper beds of the Green Pond Formation in well AWDF dip gently to moderately southeast, whereas the deepest Green Pond beds dip gently (< 30^{o}) northwest , and a little more steeply than those of the underlying Hardyston Quartzite (~10^{o} northwest; Figure 15B). Beds close to the unconformity strike about the same. The histogram plot of dip azimuth for the lower set of Green Pond is interpreted here to reflect cross bedding (Figure 18C) that shows western and southern sedimentary dispersion.

A fault zone seen in the BTV record of MW-1 occurs at a depth of about 235 ft (72 m) (Figure 18A). The zone is a gently-dipping mineralized interval consisting of about 6 in (15 cm) of anastomosing light and dark vein-fill material. This horizon is interpreted as a reverse fault that is a subsidiary splay fault to the Picatinny reverse fault (Figure 17), and that separates beds of varying strike and dip. Upper beds in the Green Pond Formation in MW-1 dip gently northwest, and then become steep northwest deeper in the well and closer to the fault. Beneath the fault, bedding returns to dipping gently northwest.

The resulting well-field features were displayed relative to a DEM and a digital version of the cross section to examine the spatial relationships of the measured features from different viewpoints. Figure 19 shows the bedrock structures based on the BTV data relative to a DEM covering the 600 Area (Figure 10A) and a 3D display of the cross section C-C'-C".

DISCUSSION

A stylized, cross section across the Reading Prong in NJ (Figure 20) was constructed to help summarize and integrate different aspects of we know about the Taconic unconformity in the GPS region, and to facilitate discussion on some of the more speculative aspects with respect to its stratigraphic variability and probable connection with the Shawangunk Formation across the Great Valley (Figures 1 and 19). The section is pinned in the foreland by the Late Ordovician Beemerville intrusive complex (Ghatge et al., 1992) and uses the restored and balanced Lower Paleozoic cover sequence depicted in section A-A' of Herman et al. (1997). Their A-A' is extended southeast toward the Taconic hinterland roots to depict a "restored" Taconic foreland cover-fold sequence that arches over the current crystalline roots of the Reading Prong. The cover is folded above crystalline basement, with northwest verging folds

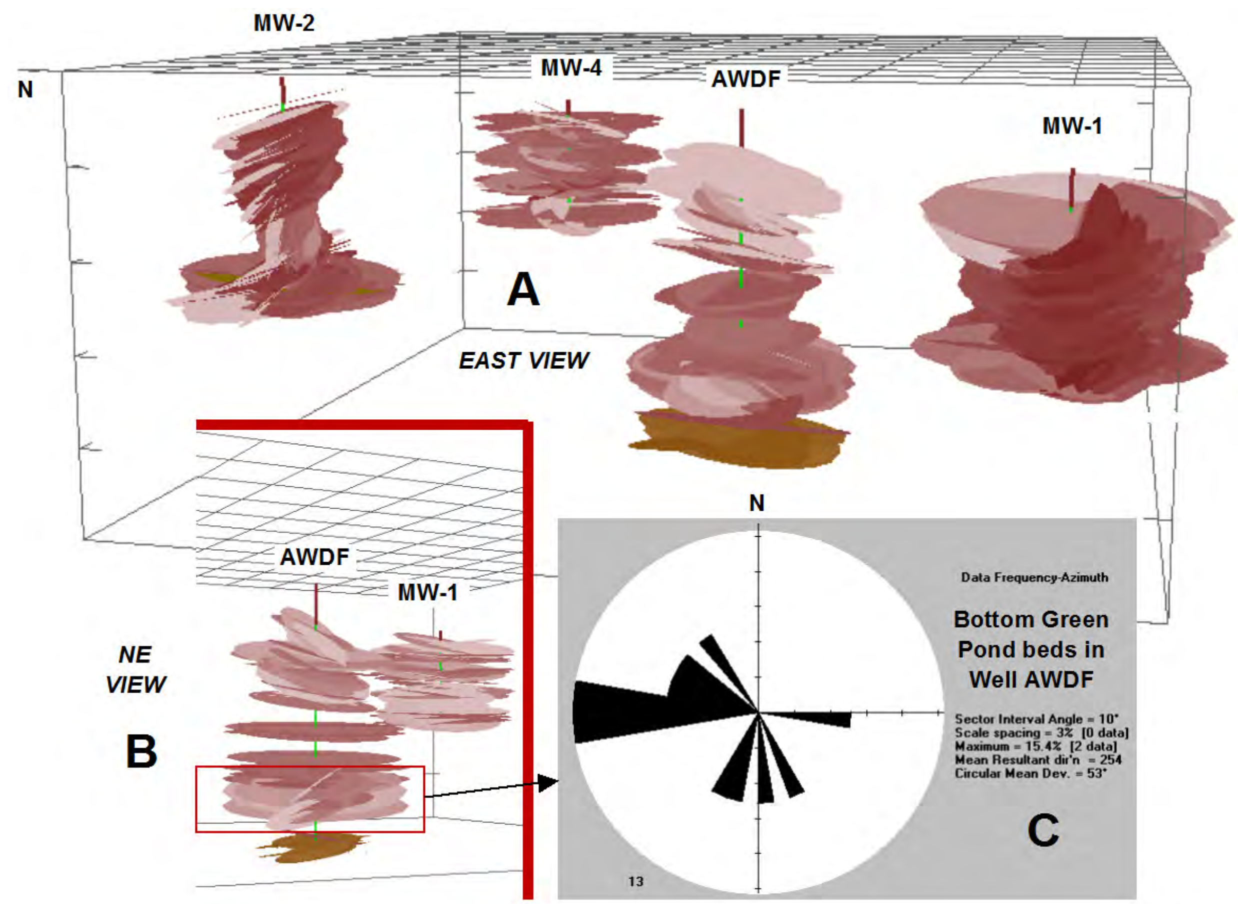

Figure 18. 3D graphic displays of the 600 Area well field and a circular histogram analysis of bed dip azimuth in the lower part of the Green Pond Formation in well AWDF. Image A is an east view showing a 3D well-field grid with 50-ft (15-m) divisions, ellipses oriented parallel to measured beds (pink polygons), and well parts including cased (upper, dark red) and open (lower, light green) well parts. Bed ellipses were generated at BTV record depths using 300 ft (91 m) major axes and 150 ft (46 m) minor axes. The pink planes are Green Pond Formation whereas the deepest beds in well AWDF (brass colored) are Hardyston Quartzite. Image B is a northeast view of wells AWDF and MW-1 showing the variability in bedding dip in well AWDF, and cross beds in the lowest part of the Green Pond Formation. Image C is a circular histogram analysis of bed dip azimuth of bottom beds in the Green pond formation from the AWDF well interval 331 to 373 ft (114 m) below land surface. The histogram and 3D display shows that the bottom beds are cross-stratified with S and W sedimentary dispersion.

that are progressively tighter with more gentle axial surfaces to the southeast, but lacking significant, emergent thrust faults that would have otherwise stack and repeat stratigraphic sections. In other words, it's one of the simplest geometric solutions that agrees with what we see at the surface, and that provides a minimal finite strain estimate based on having an eroded, folded cover sequence. As portrayed, the GPS syncline was shortened by over 50% from ensuing Alleghanian compression (Figure 20). One measurement of the penetrative, finite-shortening strain in the Devonian Bellvale Formation in the Green Pond Mountain region just from regional cleavage development is 44% (Herman, 1987).

The location of the current GPS is placed into a restored position as a Taconic synclinorium (Figure 20). The restored GPS contains broad, open, asymmetric, northwest-verging folds with localized, post-Taconic erosion of a basement-cored anticline. The anticline is a subsidiary fold in the greater synformal structure that was breeched by early erosion to expose the Proterozoic core during Taconic uplift and before Silurian deposition. Basal Silurian molasse now overlies crystalline rocks in the central area of the GPS near Green Pond, NJ, but Lower Paleozoic rocks elsewhere in all directions along strike and toward the foreland. Both Silurian and Middle Devonian molasse now occupies valleys in the GPS between crystalline blocks of the Reading Prong (Figures 3 and 4). This implies that the GPS was a structural depression then, and remains depressed today. This geometry also suggests that regional culmination during the Taconic was near the Green Pond Mountain region, where the maximum amount of erosion occurred within a great structural depression. This led Finks (1968) to conclude that there were “Taconic Islands” in this area stemming from Ordovician orogenesis.

Thomson (1957) conducted a comprehensive, petrologic and petrographic comparative study of the Shawangunk and Green Pond Formations in the NJ region. He found many sedimentary and stratigraphic relationships that bear on the depositional setting, continuity, and contemporaneity of the Silurian molasse in this region. He concluded that the two formations were probably continuous at one time from analyses of lithologies, cross-beds, and heavy-mineral fractions. Both formations contain an abundance of white quartz pebbles of probable igneous, metamorphic, and vein origin. Both formations show west to northwest cross bedding and contain white and pink zircons indicating two distinct sedimentary-source areas east of the current Great Valley and GPS. One nearby source includes the Lower Paleozoic cover and Precambrian gneiss of the Reading Prong as the source of pink zircons. The second, easternmost terrain includes argillaceous rocks of the allocthons and a “Taconia” root where other sedimentary, metasedimentary, and acidic granitic and/or metaigneous rocks provided euhedral, white zircons that reportedly are rare in the Reading Prong. He also noted that the basal Shawangunk contains considerable feldspar and the clear, euhedral zircon not found in basal sections of the Green Pond, even though is lies furthest northwest from the easternmost source. The basal Green Pond contains more jasper, chert, and flint, although Kummel and Weller (1902) noted significant plagioclase clasts in conglomerate on the east side of the GPS, and Jafee and Jafee (1973) reported coarse orthoclase pebbles at Lazy Hill, NY. The common occurrence of varicolored chert in Green Pond conglomerate led Emery (1952) to include Ordovician flysch as source rocks. Thomson (1957) also found that zircons are more rounded in the Green Pond in comparison to more elongate ones in the Shawangunk. He noted that the euhedral, white zircons first occur about 500 ft (152 m) above the base of the Green Pond Formation. He therefore proposed that the lower 500 ft (152 m) of basal Shawangunk is older than basal Green Pond because erosion began earlier, progressed more rapidly, and continued

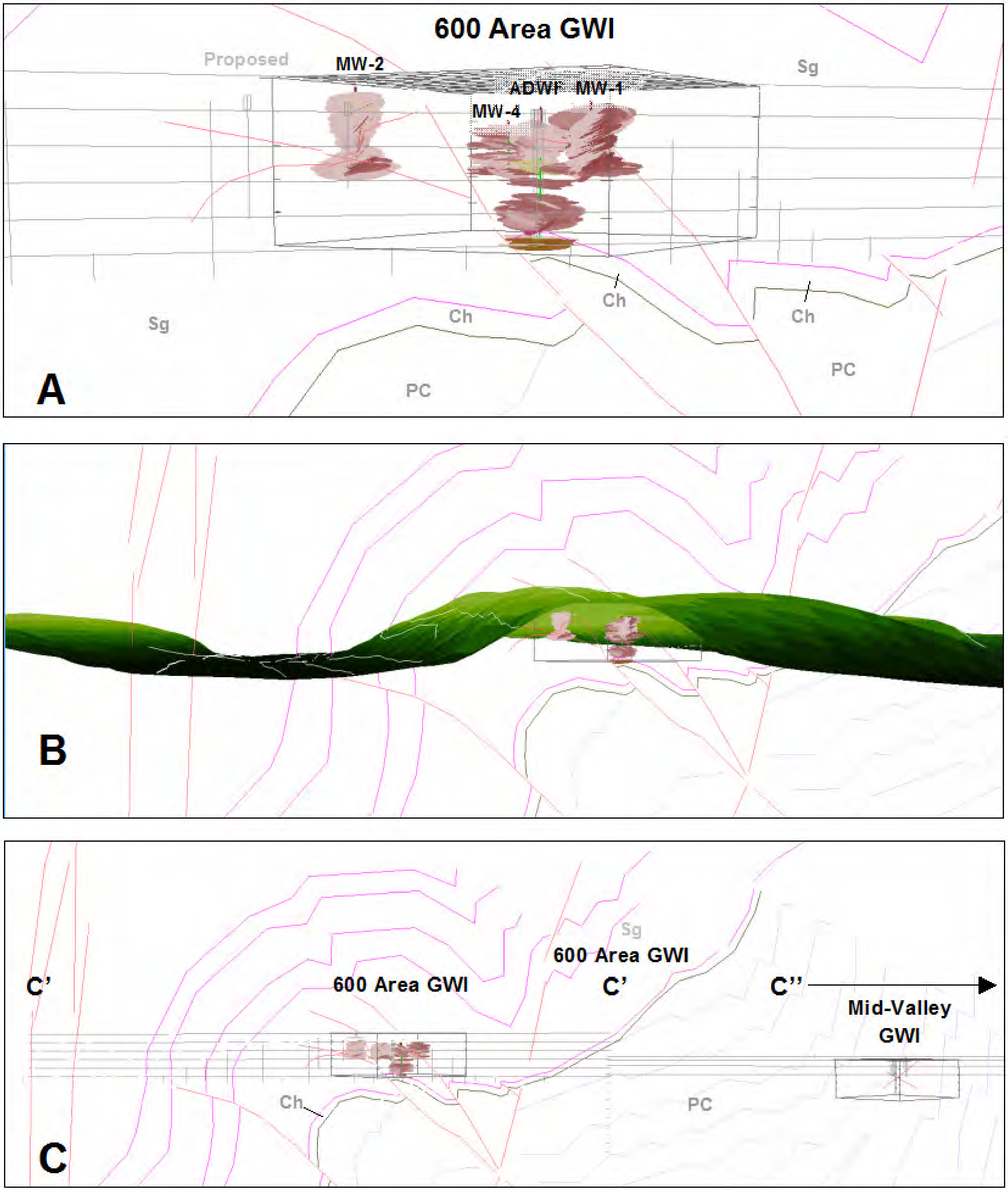

Figure 19. 3D graphic displays looking northeast of geologic and topographic details near 600 Area well field. Image A is shows the well field relative to cross section C-C', a grid with 50-ft (15-m) cells, and oriented bedding ellipses (pink polygons) that were generated at BTV record depths using 300 ft (91 m) major axes and 150 ft (46 m) minor axes. The deepest planes in well AWDF correspond to the Hardyston Quartzite (Ch in the section). Image B is shows the well field relative to ground surface represented by a digital elevation model (N.J. Geological Survey, 1999). Image C is shows the 600 and Mid-Valley groundwater investigation areas relative to cross-section C-C'-C". Profile locations mapped on Figure 13.

later in the Green Pond region than further northwest, where Early Silurian deposition began under "an eastward transgressing sea". But an alternate explanation for the apparent delay of the white zircons in the Green Pond hinges on the accumulation of early, thick, locally sourced materials without the need to invoke west-to-east stratigraphic deposition of the regional Silurian molasse. That is, initial alluvial facies of the Silurian molasse were probably deposited in localized structural depressions first, and then were overlapped by a thick, mature, alluvial cover that grew into the foreland as the hinterland source area culminated over time. In this manner, the basal Green Pond is older, and the Shawangunk becomes coeval with higher sections of the Green Pond.

A more recent description of a Taconic source area includes rift facies, pre-margin material and slope-and-rise rocks of Laurentia (Herman and Mitchell, 1991). The lower, conglomeratic part of Green Pond Formation is probably consolidated piedmont alluvium derived from nearby, deeply-weathered source area rising to the east, but its age is unknown owing to the lack of fossil or radiometric-age data. According to Thomson (1957), the Green Pond conglomerate is composed principally of white quartz and chert pebbles in a matrix of quartz sand and silt containing considerable amounts of hematitic clay, resulting in the deep purplish-red coloration. Sandstones in the lower sections of the formation grade laterally and vertically into conglomerates. The red shale intercalated with red, coarse alluvium strongly suggests a terrestrial origin; initially red sand grains washed by waves on beaches, and by streams on alluvial plains tend to lose their red color (Raymond, 1942). Having brown to light-brown orthoquartzite-conglomerate with only white vein-quartz pebbles on the west side of the syncline, and a red polymictic quartzite-conglomerate on the east limb points to the probability that the east-side, basal conglomerates were deposited within thick alluvial channels near an upland source area with subaerial exposure, deep weathering and soil development. At the same time, marginal-marine sedimentation may have occurred simultaneously along the west margin of the restored GPS where the red-purple rocks first give way to the typical brown, buff, and white colors of the Shawangunk Formation. And so it seems that alluvial and shallow marine deposits of Early to Middle Silurian age in the GPS resulted from the differential erosion of Middle and Upper Proterozoic and Lower Paleozoic rocks that supplied the sedimentary load for distributary channels draining localized uplands. Eventually, a thick blanketing molasse accumulated on the northwest regional slope in a complex transitional marine-continental environment (Epstein and Epstein, 1972) with oscillating marine shorelines in equatorial climates. Finks (1968) argued that the middle-to-upper parts of the Green Pond Formation and the lower parts of the Longwood Shale represent piedmont alluvium and marine sand deposited from a westward-transgressing shoreline.

It seems more plausible that the Green Pond was deposited first closer to a hinterland source, then was later blanketed by more regional, alluvial, and marginal-marine deposits. The Shawangunk Formation is reported as Early to Late Silurian (Epstein, 1993), and possibly Late Ordovician (Epstein and Epstein, 1972), whereas the current published age of the Green Pond Formation is Early to Middle Silurian (Drake et al., 1996, but they query the "Early" designation). The well rounded nature of the zircons toward the source area higher in the section is explained by repeated abrasion in a transitional marine, beach environment near the GPS, and reflects the tendency for Green Pond quartzite to be comparatively more mature than Shawangunk quartzite. It's more likely that erosion and deposition began in the east near the GPS where basal, immature, and thick Silurian alluvium filled local, elongate, structural troughs

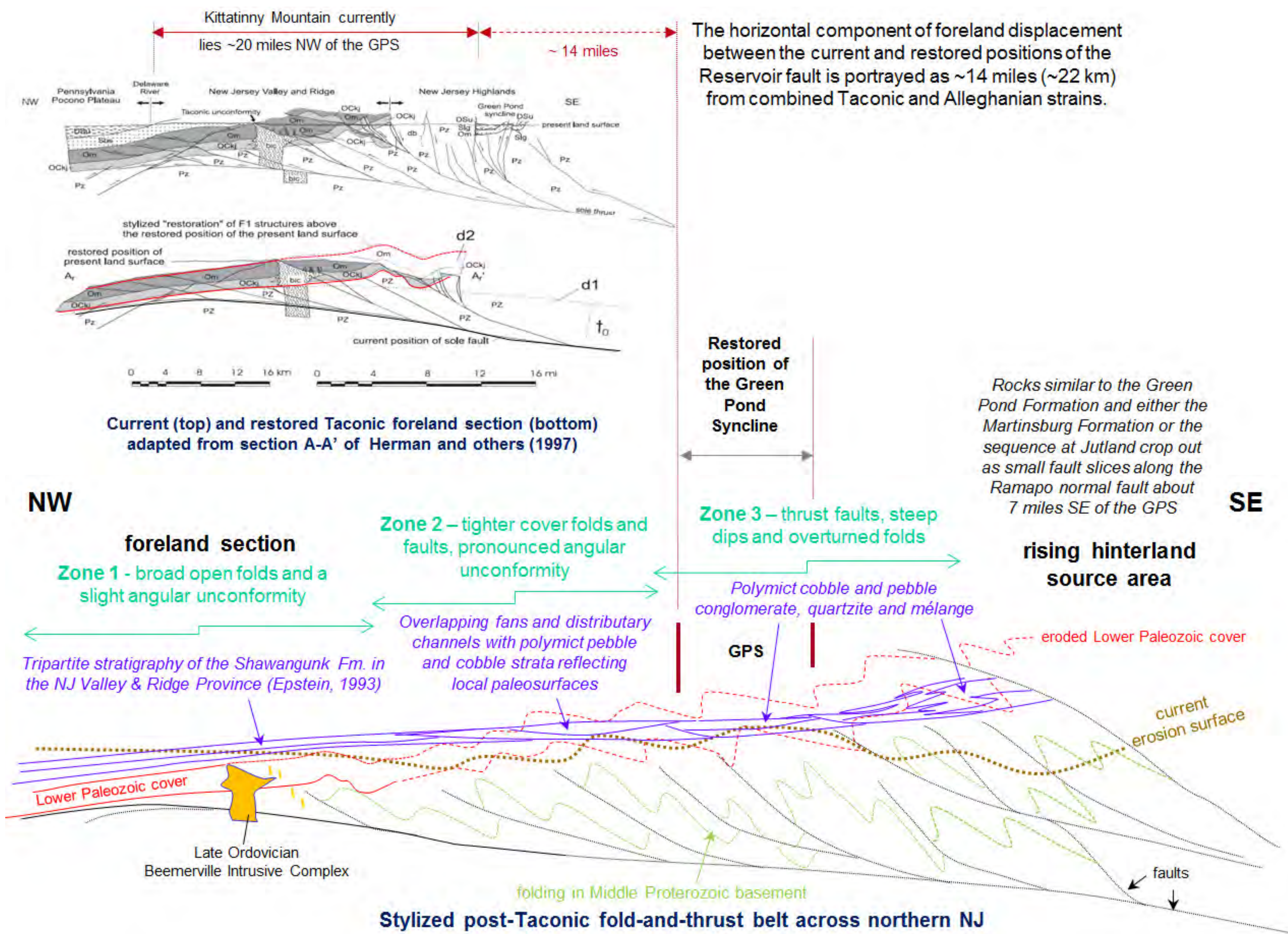
Kittatinny Mountain currently
lies ~20 miles NW of the GPS
~ 14 miles
The horizontal component of foreland displacement between the current and restored positions of the Reservoir fault is portrayed as ~14 miles (~22 km) from combined Taconic and Alleghanian strains.
NW
Pennsylvania Pocono Plateau
Delaware River
New Jersey Valley and Ridge
Taconic unconformity
New Jersey Highlands
Green Pond syncline
SE
present land surface
sole thrust
stylized "restoration" of F1 structures above the restored position of the present land surface
restored position of present land surface
d2
d1
current position of sole fault
Current (top) and restored Taconic foreland section (bottom) adapted from section A-A' of Herman and others (1997)
Restored position of the Green Pond Syncline
Rocks similar to the Green Pond Formation and either the Martinsburg Formation or the sequence at Jutland crop out as small fault slices along the Ramapo normal fault about 7 miles SE of the GPS
NW
SE
foreland section
Zone 1 - broad open folds and a slight angular unconformity
Zone 2 – tighter cover folds and faults, pronounced angular unconformity
Zone 3 – thrust faults, steep dips and overturned folds
rising hinterland source area
Polymict cobble and pebble conglomerate, quartzite and mélange
Overlapping fans and distributary channels with polymict pebble and cobble strata reflecting local paleosurfaces
GPS
eroded Lower Paleozoic cover
Tripartite stratigraphy of the Shawangunk Fm. in the NJ Valley & Ridge Province (Epstein, 1993)
current erosion surface
Lower Paleozoic cover
Late Ordovician Beemerville Intrusive Complex
faults
folding in Middle Proterozoic basement
Stylized post-Taconic fold-and-thrust belt across northern NJ

Fig
ure

before being covered and buried by more mature quartzite and sandstone in a subsiding basin that extended well past the Great Valley to the west.

The stylized nature of the Green Pond Formation in Figure 20 uses a tripartite stratigraphy for the Shawangunk Formation based on the work of Drake and Epstein (1967), Epstein and Epstein (1972), and Epstein and Lyttle (1982). Moving eastward over the Reading Prong, the molasse is shown as having overlapping and interfingered sedimentary lenses of restricted down-slope range meant to portray the infilling of restricted structural troughs and depressions that are aligned with regional strike, and lie about normal to the line of section. Moving farther southeast, the molasse is characterized as overlapping alluvial fans and mélange near the footwall of major Taconic over thrust raising a hinterland source area. However this thrust is probably not Thomson's (1957) "Taconia" root lying further southeast and hinterland of the Huntingdon Valley-Cream Valley fault in eastern PA and Cameron’s line in NY (Figure 20).

Figure 20 also depicts a normal fault having Lower Paleozoic strata dropped down along the ancestral Reservoir fault. This interpretation is consistent with having repeat episodes of slip along older, deeply rooted faults within the Proterozoic blocks, some of which may have been active during Late Proterozoic and Early Mesozoic rifting (Ratcliffe, 1980; Hull et al., 1986). For example, the Reservoir fault has a long, complicated history of slip activity (Hull et al., 1986; Herman and Mitchell, 1991; Malizzi and Gates, 1989; Gates and Costa, 1998). Early normal slip on this fault may help explain the spotty occurrence of Martinsburg Formation on the western limb of the syncline in the Green Pond Mountain region, that must have at one time tied into the Martinsburg Formation in the southern part of the Hudson Valley (Figures 7 and 8). Therefore, in the center of the GPS, it appears that that Martinsburg flysch may have been deposited atop Middle Proterozoic basement because of the lack of other, nearby Lower Paleozoic rock (Herman and Mitchell, 1991; Figure5), although A. A. Drake, Jr., upon review of Herman and Mitchell’s (1991) map, pointed out that this would be a unique occurrence in the Appalachians. It is more probable that the current cross sections underrepresent the presence and thickness of the Paleozoic cover beneath Martinsburg shale in the western, faulted limb of the syncline in NJ, but there is simply no way to know solely based on surface mapping. Where the Martinsburg is first mapped bordering the northeast end of GPS on its northwest side, there are Lower Paleozoic carbonates mapped between the Martinsburg Formation and Middle Proterozoic basement (Offield, 1967; Jafee and Jafee, 1973). But pre-Taconic, arching and open folding of the Lower Paleozoic sequence is well established in the region (Offield, 1967; Epstein and Lyttle, 1987; Herman and Monteverde, 1989; Herman et al., 1997), where early folds in the carbonate platform resulted in restricted, local stratigraphic variations of Middle Ordovician limestone and Upper Ordovician flysch (Monteverde and Herman, 1989).

More work is needed to resolve some issues arising from this work. A more comprehensive assessment of regional subsurface records, like the BTV record shown here (Figure 14), would be very useful for helping in characterizing the lateral stratigraphic heterogeneity and thickness of the Silurian molasse, and the angular discordance about the Taconic unconformity. The AWDF BTV record shows that basal parts of the Green Pond Formation can be locally fine grained near areas where it is also the thickest and coarsest. This area may have been subject to a brief period of post-Taconic relaxation with basement block faulting and local horst and graben structures involving Lower Paleozoic cover rocks, similar to Silurian Connecticut Valley–Gaspé Trough to the northeast (Rankin et al., 1997). In this case, the coarse grained clastics may have originated near fault scarps while more distal, finer alluvium filled more distal parts of local basins.

The Green Pond Formation may be the thickest near Green Pond Mountain NJ (~1,399 ft [426 m]), where the level of Taconic erosion within the GPS was the greatest (Figure 20). The unit thins to the northeast, down to as little as 25 ft (8 m) near Pea Hill, NY,= (Darton, 1894; Ries, 1895), before apparently pinching out. The thickness in the southwest part of the GPS is difficult to tell because of the friable nature of the unit and thick cover, but it may mimic the outcrop expression of the Shawangunk Formation, which generally increases in thickness from the northeast to the southwest across the NY recess (Figure 21). Darton (1894b) reports about 3 ft (0.9 m) of Shawangunk grit at the northeast end of the outcrop belt near the fourth Lake at Binnewater, NY (Figure 21). Darton (1894 a, b) also notes about 290 ft (88 m) at Ellenville, NY, and up to 2,000 ft (610 m) in NJ. Epstein and Epstein (1972) report about 1,800 ft (549 m) near Delaware Water Gap, and it may reach a maximum thickness near Pine Ridge, PA (Figure 21), where the map unit changes from the Shawangunk Formation to the Clinton Formation continuing into the Pennsylvania Salient along strike (Swartz and Swartz, 1930; Epstein and Epstein, 1967). However there are structural complications in the sections from Delaware Water Gap and further southwest that makes an accurate accounting of the unit thickness difficult and imprecise.

The angular unconformity within GPS has only been directly observed in one place, in the subsurface where Green Pond overlies the unit interpreted as Hardyston Quartzite (Figure 14). Both units strike subparallel (northeast-southwest) but at slightly different dip angles (~10^{o} below the unconformity compared to ~30^{o} above). In areas of NJ where the Green Pond overlies Lower Paleozoic rocks, the contact has not been directly observed with certainty, but nearby dip angles vary, from negligible to as much as 18^{o} (see Gould's quarry note above). Where Green Pond Formation overlies Precambrian basement, gneissic layering immediately southeast of the GPS dips steeply southeast in northwest-verging, overturned folds only near the central and northeastern parts of the NJ Green Pond Mountain; further along strike in both directions, basement fold limbs become upright, with layering dipping northwest beneath the northwest-dipping limb of the Paleozoic syncline (Figure 17; and Jafee and Jafee, 1973). It's probably that a main phase of basement folding occurred in the Reading Prong from Taconic orogenesis, and the GPS reflects area-wide, semi-ductile infolding of this age. More work is needed in comparing the geometry of basement folding with that found in the Lower Paleozoic sequence in order to test this hypothesis.

Another aspect arising from this work is the need for more detailed geological maps of Orange County, NY. Only the 1:250,000 scale geological maps for New York are available in digital form, and there are some errors in this coverage in the form of mislabeled (miscoded) polygons for at least the Lower Hudson bedrock coverage. These errors were only detected and corrected in the personal themes needed for generating this report, but they persist in the source themes. Notice of these errors is being sent to the source agency, but users of these data should beware. Also, there are mismatched map units and misaligned unit contacts along the NY-NJ state border that need better definition, as well as those occurring along the quadrangle boundaries at the 1:24,000 scale (for example, see Figure 7). More detailed mapping is also needed in both NJ and NY in and around the GPS for assessing the nature of the late-stage cross faults that cut and offset the Lower Paleozoic rocks. The age and movement on these faults is intriguing and could include Late Paleozoic through Cenozoic movements. The strike of these faults in NY is similar to some cross-faults mapped in the NJ Highlands that may also cut and offset the GPS there, but they are currently mapped as not doing so (see the Mount Hope fault in Figure 4).

In conclusion, the bulk evidence points to the probability that a prominent, Taconic-age structural culmination occurred immediately hinterland of what is now the New York recess, with Taconic root structures now buried beneath Mesozoic cover of the Newark basin and Cenozoic coastal plain sediment laid down on the ensuing passive margin (Figure 1 and 20). It is also likely that these Taconic roots include both 'external' and "internal" basement massifs (Drake et al., 1989), the former including rocks of the Reading Prong and miogeoclinal cover,

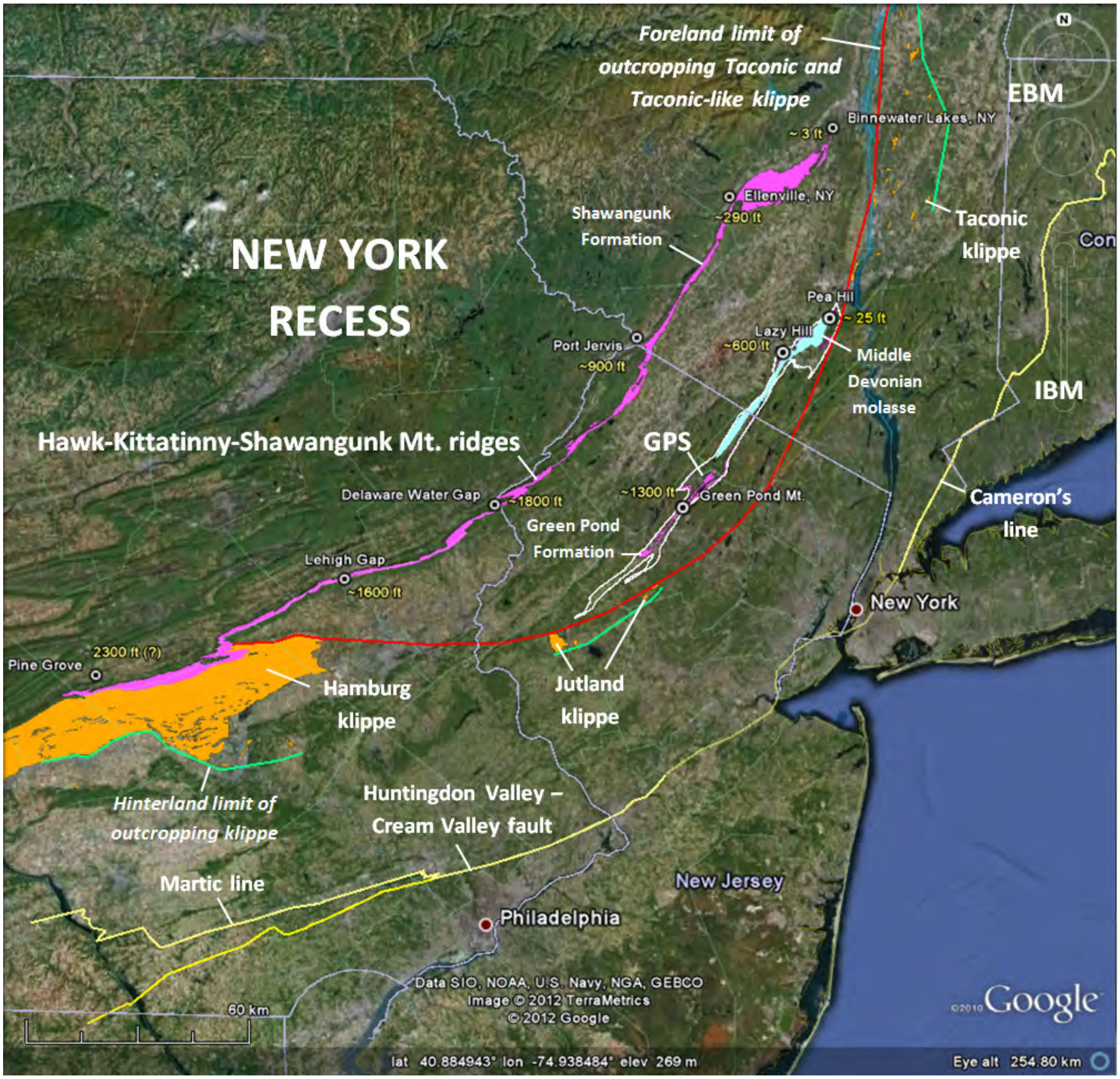

Figure 21. Aspects of the Taconic orogeny around the New York recess summarized using Google Earth. Geological shapes adapted from the US Geological Survey Online Spatial Data; Geological maps of US States (http://mrdata.usgs.gov/geology/state/). Klippe and other terminology after Drake et al. (1989). EBM – Terrain including "external basement massifs" and the foreland basin on the Laurentian margin. IBM – Terrain including "internal basement massifs" and their migeoclinal cover poking through thrust sheets containing eugeosynclinal rocks. The best estimate for the thickness of the Silurian molasse is compiled at a number of places for the Green Pond Formation in the GPS, and for the Shawangunk Formation in the Hawk-Kittatinny-Shawangunk Mountain ridges. See the discussion section for further explanation.

and the latter including abundant mafic rocks, granite, and highly strained eugeosynclinal rocks with “gneiss domes” containing Proterozoic basement like those mapped in New England and the PA Piedmont. These “internides” (Hatcher, 1987) may be the source of Thomson’s (1957) euhedral, white zircons in the heavy mineral fraction of the Silurian molasse. It’s probably more than just coincidence that the Watchung syncline (Figure 1), which contains the thickest accumulation of Early Mesozoic strata in the Newark basin, lies directly southeast of the area of maximum erosion of Paleozoic cover in the GPS. This spatial relationship suggests that the location of maximum structural relief on the New York recess stemming from Paleozoic orogenesis is coaxial with the location of maximum relaxation and normal slip on ensuing Mesozoic faults, including, but not limited to, the Ramapo fault.

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THE BEEMERVILLE ALKALINE COMPLEX, NORTHERN NEW JERSEY

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Introduction

The Beemerville complex is located at the western end (Figure 1) of the Cortlandt-Beemerville magmatic belt (Ratcliffe, 1981). The belt cross-cuts the structural grain of the Appalachians. There is a trend from calc-alkalic and alkalic rocks at the eastern end of the belt to strongly alkalic rocks at the western end. Mafic and intermediate dikes that intrude the magmatic belt show a similar trend in rock composition. Ratcliffe (1968, 1981) concluded that, relative to Taconic dynamo-thermal metamorphism, the Cortlandt complex was emplaced syntectonically while the Beemerville complex was post tectonic. Thin section examination shows no evidence of a metamorphic or tectonic fabric supporting the conclusion that the Beemerville complex is post tectonic.

In this paper I will focus on the geology, mineralogy, and petrography of the Beemerville complex. For a more complete description of the complex in terms of mineral and rock chemistry please consult Eby (2004). A pdf of Eby (2004) is available on request from the author (**nelson_eby@uml.edu**).

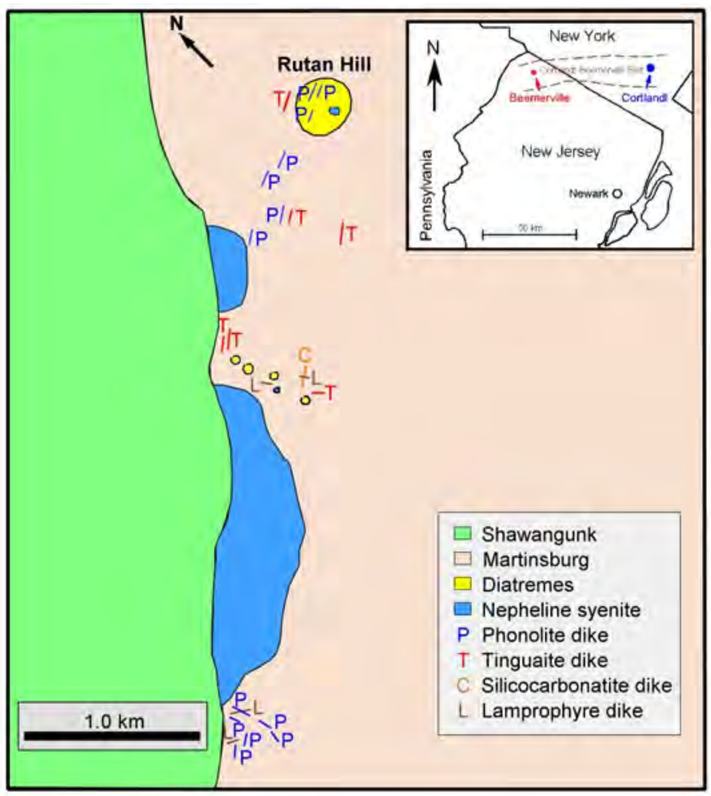

Figure 1. Location and geologic map for the Beemerville complex. Modified from Maxey (1976) and Eby (2004).

Geology of the Beemerville Complex

The following geologic description of the Beemerville complex (Figure 1) is summarized from Maxey (1976) and taken from Eby (2004). The nepheline syenite occurs in two large bodies and as several small bodies, one located in the Shawangunk conglomerate and the other associated with the diatreme of Rutan Hill. The large bodies intrude the Ordovician Martinsburg shale and to the west are overlain by the Silurian Shawangunk conglomerate. The nepheline syenite that crops out in the

Eby, G.N., 2012, The Beemerville alkaline complex, northern New Jersey, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 85-91.

Shawangunk is stratigraphically higher than the basal Shawangunk and may represent an intrusion of nepheline syenite from the southern large nepheline syenite body into the Shawangunk. The Martinsburg shale has been metamorphosed to a hornfels near the boundary with the nepheline syenite plutons, but actual contacts between the nepheline syenite and the shale are not observed. The nepheline syenite is medium- to coarse-grained and consists of nepheline, orthoclase, clinopyroxene, biotite, titanite, ±melanite, magnetite, apatite, and trace amounts of pyrite and zircon.

Several diatremes are found in the Martinsburg shale, and the largest comprises Rutan Hill. The diatremes consist of a variety of angular to subangular xenoliths and autoliths in a dark, dense matrix. Xenoliths noted by Maxey (1976) include shale and graywacke (Martinsburg Formation), fine-grained pale-blue dolomite (Kittatinny Formation?), cream-colored, fine-grained limestone (Jacksonburg Formation), and gneiss (similar to that observed in the Reading Prong). Autolithic inclusions are nepheline syenite, micromelteigite, and carbonatite. The matrix consists of an extremely fine-grained groundmass with fine-to coarse-grained megacrysts of biotite, diopside, aegirine-augite, orthoclase, magnetite, apatite, and nepheline. A nepheline syenite plug, approximately 30 m (98 ft) in diameter, intrudes the Rutan Hill diatreme. Fenitization of the greywacke in the Martinsburg is noted in a fracture zone associated with the small diatremes. The fenitized greywacke consists of albite, aegirine, sodic amphibole, and trace amounts of biotite and calcite.

Phonolite, tinguaite and lamprophyre dikes occur in the Beemerville area. Phonolite dikes occur in the southern nepheline syenite body, crosscut diatremes, and intrude the Martinsburg Formation. Dikes vary from a few cm to up to 50 m (164 ft) in width and both fine-grained and porphyritic varieties are noted. The phonolites associated with the Rutan Hill diatreme are extensively altered. Tinguaite dikes are found in both nepheline syenite bodies and the Martinsburg formation. The dikes vary in width from 0.25 to 50 m (10 in to 164 ft). Most are porphyritic and the common phenocrysts are nepheline, orthoclase, clinopyroxene, and titanite. The lamprophyre dikes occur throughout the complex and range in width from 10 cm to 7 m (4 in to 23 ft) and most trend northwest. Many of the dikes are extensively altered and the primary minerals are largely replaced. Where observable, the primary minerals are diopside, aegirine-augite, bioite, nepheline, orthoclase, titanite, melanite, magnetite, and apatite. Ocelli are common in some of the dikes.

Based on gravity data, plus supporting aeromagnetic data, Ghatge et al. (1992) modeled the Beemerville complex as a thin body near the surface that broadened and elongated with depth. The Beemerville body extends to the southwest and plunges at high angle to the southeast. From their data they also inferred the existence of a second, subsurface, intrusive body approximately 9 km (6 mi) southeast of the Beemerville complex.

Petrography

(from Eby, 2004)

Phonolite and Tinguaite

Mineralogically phonolite and tinguaite are the same rock. Historically the distinction was based on the presence of phenocrysts (tinguaite). The term tinguaite is now considered archaic. As used on the geologic map (Figure 1), phonolite indicates a fine-grained dike rock while

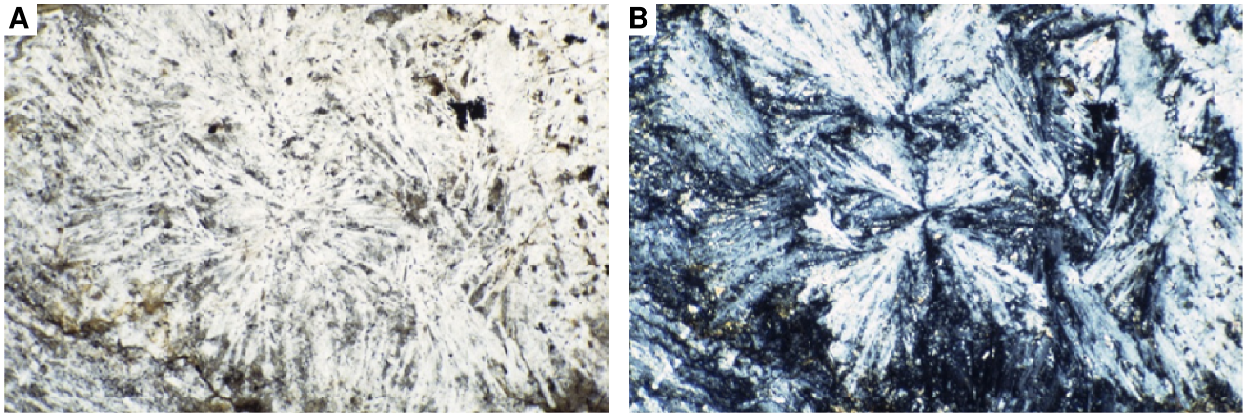

Figure 2. Phonolite dike at base of Rutan Hill (BEM16). Major minerals are nepheline and orthoclase with minor aegirine-augite. Plumose texture indicative of rapid cooling. Width of field, 2 mm. A - Plane light; B – Crossed polars.

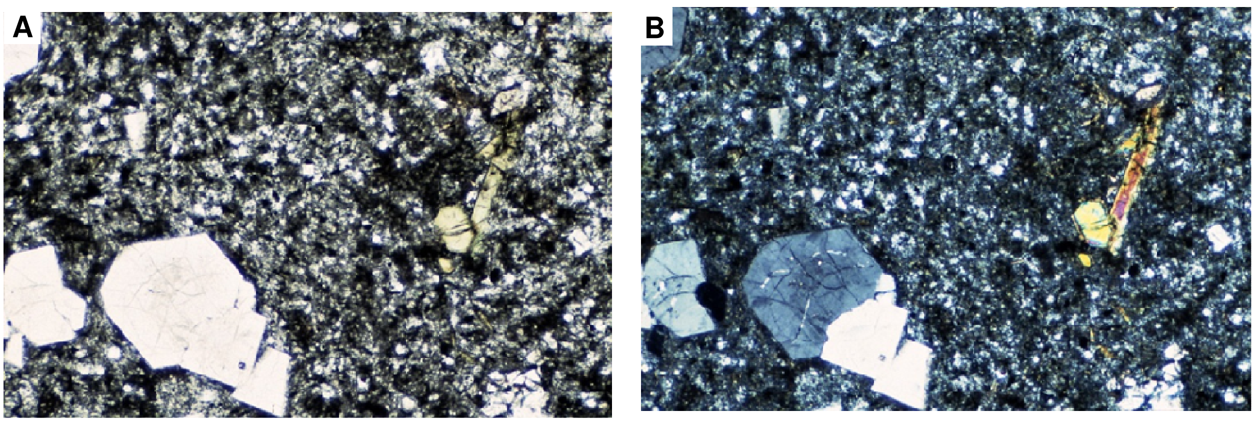

Figure 3. Phonolite (*tinguaite*) dike (BEM19). Nepheline, aegirine, and titanite phenocrysts in a fine-grained matrix of nepheline, orthoclase, and aegirine. Width of field, 5 mm. A - Plane light; B –Crossed polars.

tinguaite indicates the presence of phenocrysts in the dike rock. The major minerals in the phonolites are orthoclase, nepheline, and clinopyroxene. Minor minerals are biotite, titanite, and opaque minerals. A fine-grained phonolite, with a texture indicative of rapid cooling, is illustrated in Figure 2. In the porphyritic varieties, orthoclase, nepheline, clinopyroxene, and titanite are the common phenocryst phases (Figure 3).

Nepheline Syenite

The nepheline syenite is medium- to coarse-grained, equigranular to subporphyritic. Major minerals are orthoclase, nepheline, and clinopyroxene. Minor minerals are biotite, titanite, ±melanite, fluorite, cancrinite, sodalite, calcite, magnetite, and apatite. Based on probe data, the feldspars are orthoclase ($\geq Or_{80}$). Optically, the feldspars do not show unmixing textures. The pyroxenes vary in composition from diopside to aegirine. Zoned clinopyroxenes have diopside cores and aegirine-augite rims or aegirine-augite cores and aegirine rims. Biotite occurs as a minor mineral either replacing clinopyroxene or as discrete grains. Cancrinite and sodalite replace nepheline. Fluorite is a common trace mineral. Figures 4 – 7 illustrate the various types and textures of nepheline syenite.

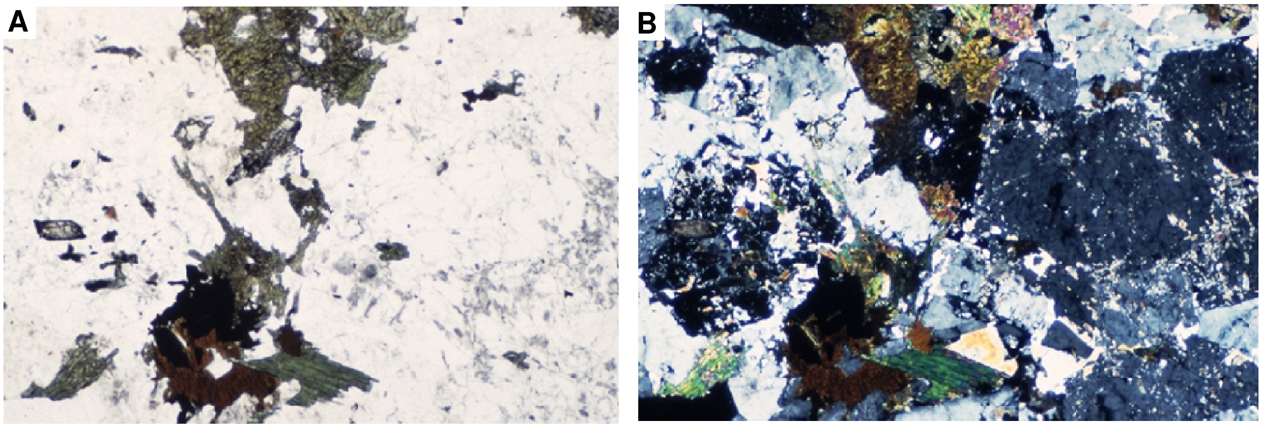

Figure 4. Coarse-grained nepheline syenite (BEM4). Nepheline, orthoclase, aegirine-augite, and minor biotite. Width of field, 5 mm. A - Plane light; B – Crossed polars. Sodalite replaces nepheline (isotropic mineral).

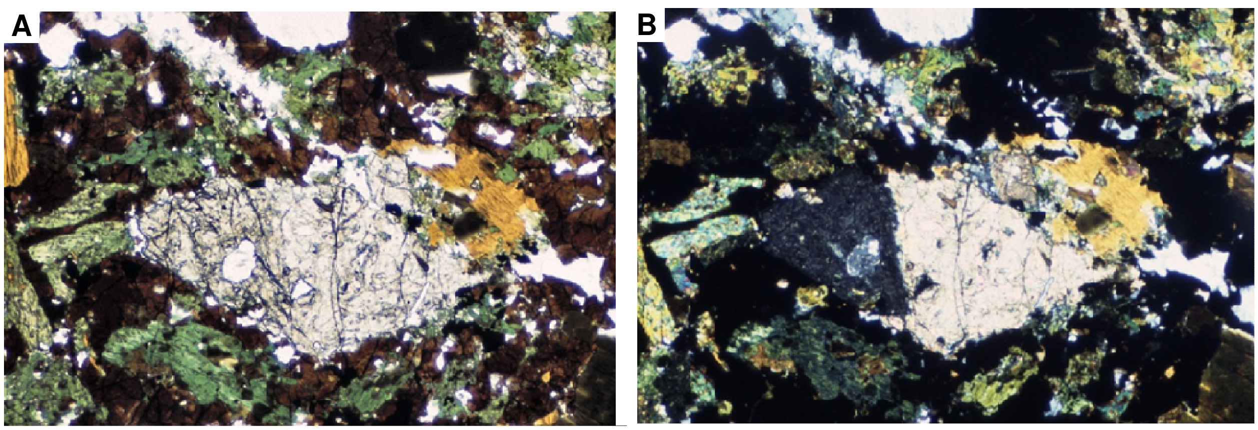

Figure 5. Nepheline syenite (BEM29). Intergrown melanite and aegirine-augite, minor biotite, feldspar, titanite. Width of field, 5 mm. A - Plane light; B – Crossed polars. The high birefringence grain in the center of the field of view is calcite.

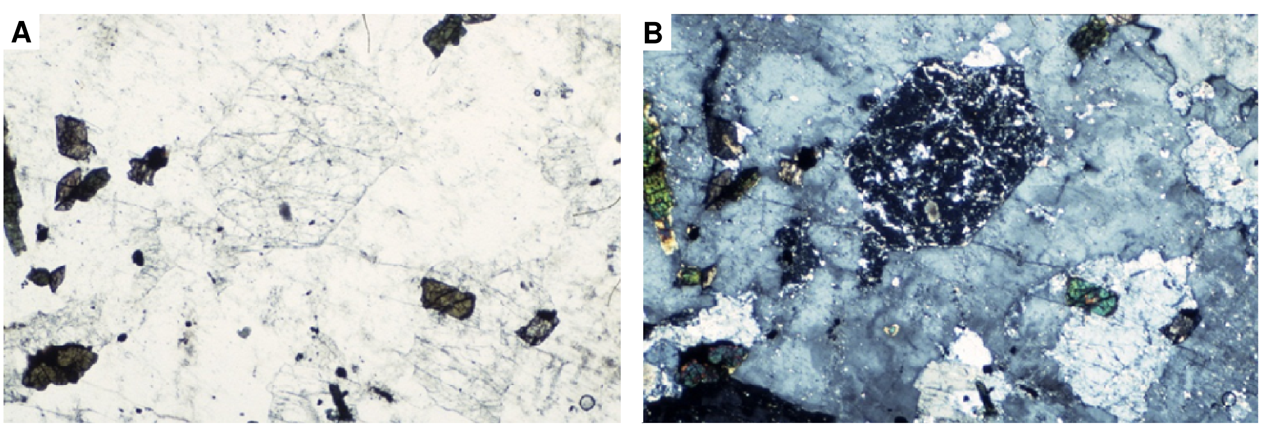

Figure 6. Nepheline syenite (BEM33). Euhedral to subhedral nepheline, aegirine-augite, and titanite in a large orthoclase. Width of field, 5 mm. A - Plane light; B - Crossed polars. Note that the mineral inclusions in the orthoclase are the same as the phenocryst minerals in the phonolite (Figure 3A and B).

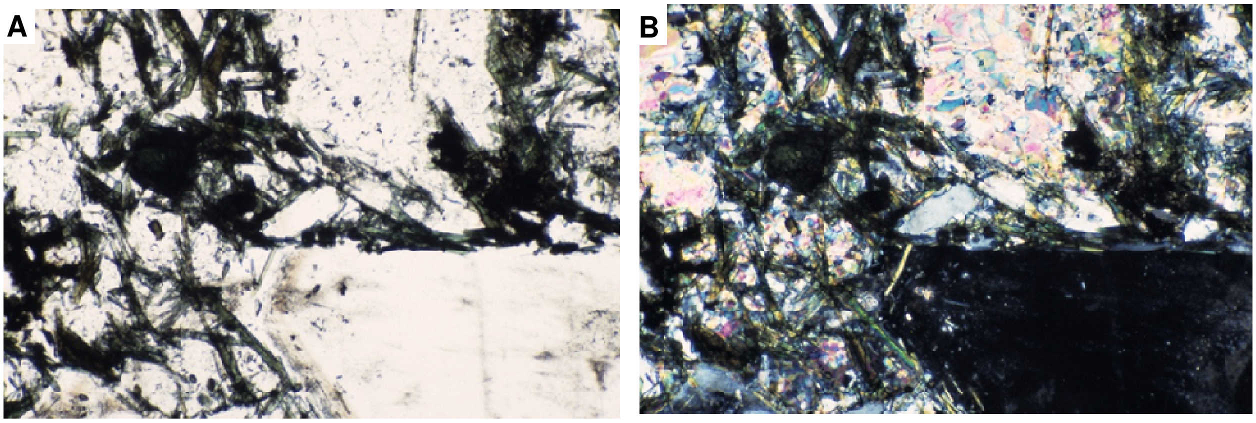

Figure 7. Border phase of northern nepheline syenite body (BEM2). Phenocrysts of nepheline and orthoclase in a fine-grained matrix of feldspar and aegirine. The nepheline has been extensively altered by cancrinite. Width of field, 2 mm. A - Plane light; B - Crossed polars. The high birefringence mineral is cancrinite.

Diatreme Breccia

The diatreme breccias are heterogeneous at all scales and show a wide range of xenoliths and autoliths. Locally calcite is an important component. An example of this carbonate breccia is shown in Figure 8.

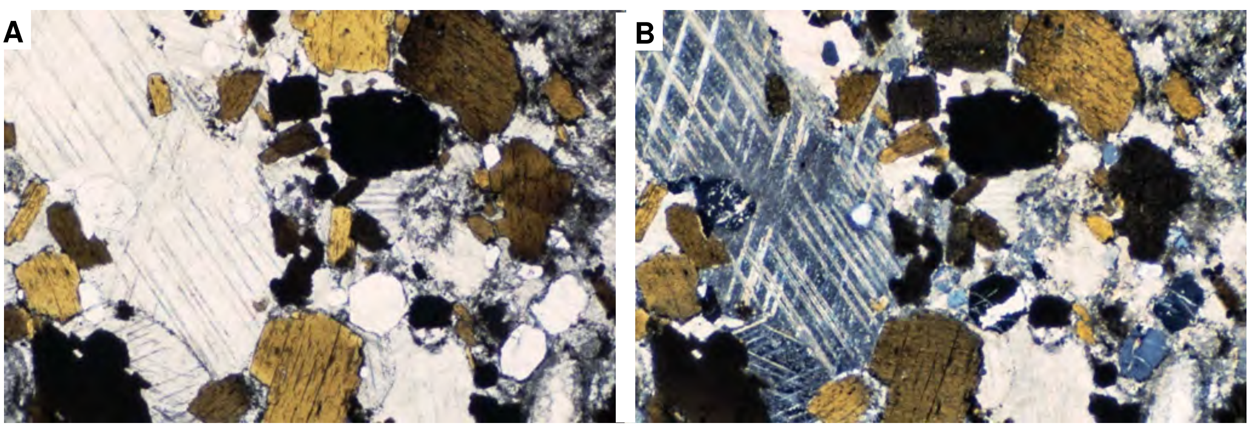

Figure 8. Carbonate breccia from Rutan Hill (BEM11). Biotite, apatite, and opaque minerals in a coarse-grained matrix of calcite. Width of field, 5 mm. A - Plane light; B - Crossed polars.

Lamprophyre Dikes

Many of the lamprophyre dikes are intensely altered obscuring the primary mineralogy. For relatively unaltered specimens the major minerals are diopside or aegirine-augite, and/or biotite phenocrysts in a fine-grained matrix of clinopyroxene, biotite, nepheline, orthoclase, titanite, magnetite, calcite, and apatite. Some of the dikes contain ocelli (Figure 9).

Discussion

The Beemerville complex is of interest because it is the only occurrence of strongly silica-undersaturated rocks of Ordovician age in the northeastern United States. Hydrous minerals

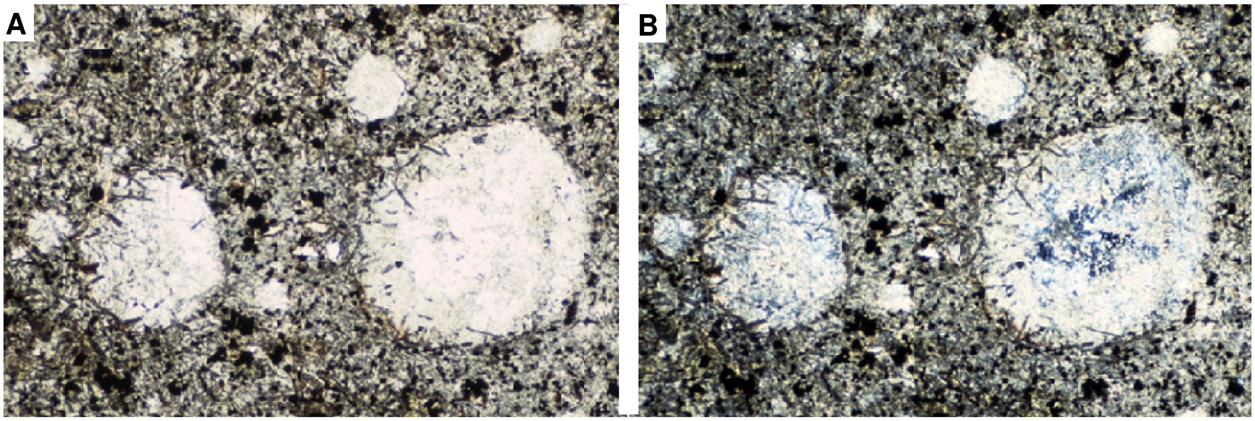

Figure 9. Ocelli in lamprophrye dike (BEM20). Width of field, 5 mm. A - Plane light; B - Crossed polars.

(with the exception of minor biotite) are absent in the nepheline syenites and fluorite occurs as a minor mineral. This suggests that the nepheline syenites crystallized from an anhydrous F-rich magma. On the basis of texture and rock chemistry, it is inferred that the various syenites are largely composed of cumulate minerals (alkali feldspar, nepheline, and clinopyroxene). The only potential solidified liquids, although their chemistry has been modified by hydrothermal alteration, are the phonolites and lamprophyre dikes. The carbonate-rich matrix of the diatreme breccia has chemical characteristics typical of carbonatites.

A number of radiometric ages have been determined for the Beemerville complex. Eby (2004) reported a mean apatite fission-track age of 156 ± 4 Ma and a mean titanite fission-track age of 420 ± 6 Ma. Zartman et al. (1967) reported a K-Ar biotite age of 443 ± 22 Ma (corrected to currently accepted decay constants) for a syenite. Ratcliffe et al. (2011) reported a titanite U-Pb TIMS age of 447 ± 2 Ma for a nepheline syenite. Apatite retains fission-tracks at temperatures less than 120°C, while for titanite fission-track retention starts at temperatures less than ~275°C. Thus the fission-track ages reflect the cooling history of the pluton. The biotite K-Ar and titanite U-Pb ages may represent the age of the intrusion, but it should be noted that the closure temperature for biotite is ~350 °C and the closure temperature for the U-Pb system in titanite is ~450 °C. The older ages can be compared to the currently accepted age of 443.7 ± 1.5 Ma (Gradstein and Ogg, 2004) for the Ordovician-Silurian boundary. Thus one can conclude from the existing geochronological data that the Beemerville complex was emplaced in the late Ordovician. By 420 Ma the intrusion had cooled to ~275 °C, but ambient temperatures of ~100 °C were not reached until 156 Ma. The 27-million-year difference in titanite fission-track and titanite U-Pb ages, coupled with phase equilibria considerations (Eby, 2004), lead to the inference that at least 3 km (1.9 mi), and as much as 6 km (3.7 mi), of rock overlay the Beemerville complex, relative to its present level of exposure, at the time of intrusion. The young apatite fission-track age is most likely related to the unroofing of the Appalachian orogenic belt.

Isotope data are lacking for the Beemerville complex, but the mineralogy and chemistry of the rocks of the complex suggest that the magma originated in the upper mantle at depths at which garnet was a stable phase. The most likely source is enriched lithospheric mantle. The presence of a carbonatitic matrix in the Rutan Hill diatreme breccia also points to a deep lithospheric origin for the magma(s). The emplacement of alkaline silica undersaturated rocks and carbonatites is invariably associated with an extensional (or at least non-compressive) tectonic environment. The Beemerville pluton was intruded during a time of medium to high-grade metamorphism and closure of the Iapetus Ocean. Thus, the intrusion of the Beemerville complex represents a variance from the overall compressive regime for this region during the Ordovician. Ratcliffe et al. (2011) have suggested that the intrusion of the rocks of the Cortlandt – Beemerville belt was controlled by "post-collisonal sublithospheric tears and mantle upwelling oblique to the suture" formed during closing of the Iapetus Ocean.

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GLACIATIONS OF WESTERN NEW JERSEY AND EASTERN PENNSYLVANIA: A VIEW ACROSS THE DELAWARE RIVER

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Preface

The 2012 Pennsylvania Field Conference from Port Clinton, PA to Ellenville, NY will take us through an area that has experienced multiple glaciations over the last two million years. Based on the distribution and degree of preservation and weathering of glacial drift, at least three glaciations have been defined for New Jersey and eastern Pennsylvania. With the exception of eastern New Jersey, each succeeding glaciation stopped short of the older one, preserving a terrestrial record that has kept geologists (at least the good kind) busy for more than a hundred years. The last two million years also included multiple cycles of periglacial weathering where large volumes of colluvium, formed by fragmental disintegration of rock, was shed from uplands, and cycles of temperate to sub-tropical interglacial weathering that was dominated by the formation of saprolite and decomposition residuum.

Starting just outside the glacial border near Port Clinton, PA, the field trip will traverse areas that were glaciated during the pre-Illinoian, Illinoian, and Late Wisconsinan glaciations (Figure 1). The Late Wisconsinan border, in many places marked by a terminal moraine, divides the field trip area into two contrasting landscapes: 1) a southern area marked by extensive colluvium, weathered bedrock and in places patchy, weathered glacial drift; and 2) a northern area marked by extensive fresh to lightly weathered glacial drift and numerous bedrock outcrops.

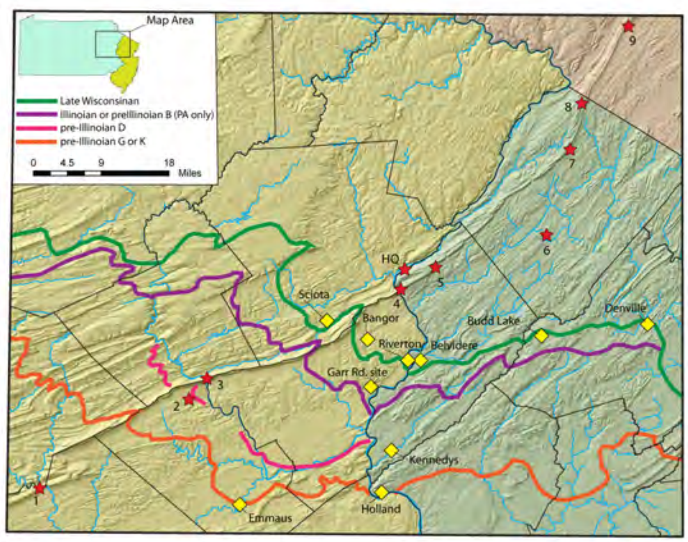

Figure 1. Map showing the extent of glaciations in eastern Pennsylvania and western New Jersey, and location of field trip stops and headquarters for the 2012 Field Conference of Pennsylvania Geologists. Glacial limits from Braun (1996a), Sevon and Braun (1997), and Stone et al. (2002). List of field stops: 1—Port Clinton; 2—Slatedale; 3—Lehigh Gap; 4—Delaware Water Gap; 5—Yards Creek Pumped Storage Facility; 6—Newton; 7—Beemerville; 8—High Point; 9—Otisville; and HQ—Shawnee-on-Delaware.

This is a summary discussion about the glacial history of western New Jersey and eastern Pennsylvania. It could not have been written without

Witte, RW. and Germanoski, Dru, 2012, Glaciations of western New Jersey and eastern Pennsylvania: A view across the Delaware River, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 92-117.

the contributions of so many that have labored in the woods and fields since the late 1800s. The earliest investigators, Lewis, Chamberlin, Salisbury, White, Williams and later Leverett and MacClintock wrote detailed reports that laid the foundation for modern studies. What follows is the most current understanding about glacial limits, ages of glaciations, and character of glacial drift. Besides, it's always interesting to discuss geology across state borders.

Introduction

New Jersey's terrestrial glacial record shows that the Laurentide ice sheet reached New Jersey at least three times (Stone et al., 2002) over the last two million years. These glaciations (Figure 1), following the terminology of Richmond and Fullerton (1986), are from youngest to oldest the Late Wisconsinan (Marine Isotope Stage (MIS) 2), Late Illinoian (MIS 6), and pre-Illinoian G (MIS 22) or older and possibly more than one glaciation. Braun (2004) cited evidence of four glaciations in Pennsylvania: Late Wisconsinan, Late Illinoian or pre-Illinoian B (MIS 12), pre-Illinoian D (MIS 16), and pre-Illinoian G (MIS 22). Similar to New Jersey's oldest glacial deposits, those in Pennsylvania may represent more than one glaciation. There is some disagreement concerning the age of the older glaciations and number of pre-Illinoian glaciations, but there is a remarkable congruency between the glacial limits mapped on either side of the Delaware River. The youngest glacial deposits laid down during the Late Wisconsinan substage provide the clearest record of glaciation. The glacial record, indicated by the Illinoian and especially the pre-Illinoian deposits, is much less clear due to an extensive and complex periglacial and weathering history.

Multiple glacial cycles have greatly modified the Garden State and Penn's Woods. Valleys were deeply scoured, and bedrock ridges, hills, and slopes were worn down by glacial erosion. In places, glacial ice and drift dammed valleys, rerouting streams and establishing new drainage ways. Most of the eroded debris entrained by the ice sheets was deposited as till and meltwater sediment. Numerous ice-marginal lakes formed in valleys dammed by moraine, ice-contact deltaic deposits and ice. These lakes and their associated deposits, and recessional moraines provide a detailed record of deglaciation. The many unweathered and lightly weathered bedrock outcrops north of the terminal moraine show that most of the pre-existing weathered bedrock and surficial material had been removed by glacial erosion. This area lies in stark contrast to that south of the Late Wisconsinan Glacial Maximum (LWGM) where outcrops are far fewer (Figure 2). The LWGM also divides the largely glacial landscape to the north and the largely colluvial landscape to the south.

Although the effects of glaciation in modifying the landscape are pronounced, these modifications in the older glacial landscapes have been largely masked or removed by periglacial weathering. Based on the sawtooth record of the marine isotope record, Braun (1989) indicated that there may have been as many as ten glaciations of a magnitude sufficient to glaciate or introduce a periglacial climate to Pennsylvania and New Jersey. Glacial/ periglacial periods in New Jersey were short-lived and marked by intense physical weathering. Colluvium, a major weathering product of periglacial climate, was shed off uplands onto the lower parts of hillslopes and onto the floor of narrow valleys and heads of drainage basins. It is chiefly a monolithic diamicton derived from weathered bedrock (chiefly by fragmental disintegration of outcrop and regolith by frost shattering) and transported downslope largely by creep. Over time it accumulated at the base of slopes, forming an apron of thick material, and it also collected on the floors of narrow valleys and in first-order drainage basins. In places it is

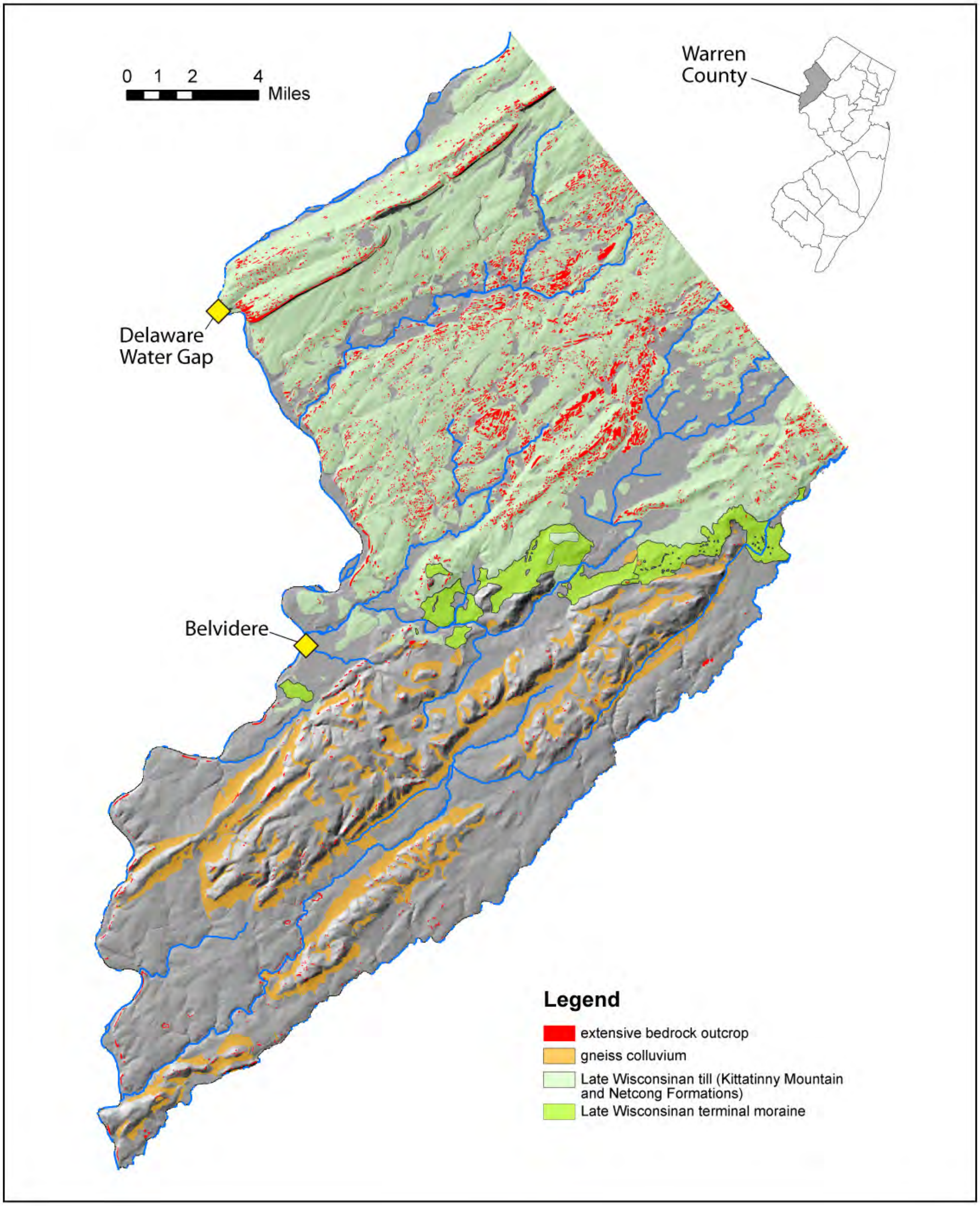

Figure 2. The Late Wisconsinan terminal moraine divides the largely colluvial landscape to the south where rock outcrops are sparse from the glacial landscape to the north where rock outcrops are numerous. Data from Witte and Stanford (1995) and Stone et al. (2002).

greater than 50 ft (15 m) thick and it covers large parts of the landscape. In contrast during the warm interglacials, the relative rate of chemical weathering increased and an extensive cover of deeper-rooted vegetation helped reduce the rate of mass wasting. During these periods thick soils were formed and bedrock was deeply weathered forming saprolite and decomposition residuum. Braun (1999) suggested that sub-till dissolution of carbonate bedrock in the Great Valley has been sufficient to overprint the primary glacial topography, particularly on the oldest till (≥ 850,000 yr BP). Constructional knob and kettle glacial topography, subsequently altered by periglacial processes and bedrock dissolution produced a composite topography that Braun (1999) refers to as "pseudo-moraine".

Previous Investigations

New Jersey

Cook (1877, 1878, and 1880) discussed the geology of New Jersey's glacial deposits in a series of Annual Reports to the State Geologist. He included detailed observations on the terminal moraine, recessional moraines, distribution and kinds of drift, and evidence of glacial lakes. Deposits of "older" weathered drift were discussed by Cook (1880) who noted the distribution of quartzose boulders and scattered patches of thin gravely drift in Pohatcong and Musconectcong Valleys in western New Jersey. Most of this material was thought to be "modified glacial drift", possibly deposited by meltwater and reworked later by weathering and fluvial erosion. On greater inspection (Salisbury, 1893) this "modified glacial drift" was determined to be of glacial origin and called extra-morainic drift because of its distribution south of the terminal moraine.

A voluminous report by Salisbury (1902) detailed the glacial geology of New Jersey region by region. The terminal moraine and all surficial deposits north of it were interpreted to be products of a single glaciation of Wisconsinan age. Salisbury also noted that "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." South of the terminal moraine Salisbury (1902, Plate XXVIII) shows two deposits of extra-morainic glacial drift. The first, forming a narrow belt just outside the terminal moraine, consisted of glacial drift of late glacial age mixed with material that was older than the terminal moraine. Salisbury indicated that the drift was deposited during a temporary advance of ice beyond the terminal moraine, or was carried out by running water. The second body of extra-morainic drift is largely glacial and much older than the terminal moraine based on its deep weathering and patchy distribution. It lies as much as 20 mi (32 km) beyond the terminal moraine. Salisbury (1902) assigned a Kansan age to the older drift because its deeply weathered appearance suggested it was the product of a much older glaciation than the Wisconsin. Chamberlin and Salisbury (1906) correlated the oldest drift with the sub-Aftonian glacial stage of Iowa, using the term "Jerseyan" as an equivalent stage for the older glacial deposits in Pennsylvania and New Jersey. Bayley et al. (1914) divided the extra-moranic drift into "early glacial drift" that was largely till deposited during the Jerseyan stage and "extra-morainic drift" that consisted of a mix of Wisconsin and early drift.

MacClintock (1940) concluded that there were also three ages of glacal drift in New Jersey, the youngest of Wisconsinan age and two pre-Wisconsian drifts of Illinoian and Kansan age. He largely based his conclusions on the degree of weathering of medium to coarse grained gneiss and pegmatite clasts. Ridge (1983) and Cotter et al. (1986) indicated the youngest glacial deposits in Pennsylvania and New Jersey are Late Wisconsinan age, and are correlative with the Olean drift in Pennsylvania, and Ridge et al. (1990) showed that older and weathered drift in the Delaware Valley north of Marble Mountain is Late Illinoian age and not early Wisconsinan. The only early Wisconsinan deposits were colluvium and fluvial deposits that were observed in the Delaware Valley near Brainards, New Jersey about 4 mi (6 km) downstream from the Late Wisconsinan terminal moraine. Stone et al. (2002) also indicated that the youngest glacial deposits in New Jersey are Late Wisconsinan age, and that the two older drifts are Illinoian, and pre-Illinoian age.

Stanford (1997) suggested that the oldest glacial deposits in New Jersey may be pre-Illinoian K age (2.14 – 2.01 Ma), based on similarity of weathering characteristics, topographic position, and erosional preservation to that of the Pennsauken Formation, a Pliocene age braided stream deposit. Additional evidence for antiquity of these deposits includes the identification of pre-Pleistocene pollen identified in the lower part of a 60-ft (18-m) core from Budd Lake, NJ by Harmon (1968). Sediment sampled (below 45 ft [14 m]) may be laminated clays of pre-Illinoian proglacial lake deposits. It is overlain by late Wisconsinan/Illinoian proglacial lake sediment. Harmon (1968) dismissed the finding of the older pollen because the exotic taxa appeared to have been rebedded prohibiting their use as a trusted indication of Pliocene age.

Ridge (1983) and Witte (1988, 1997, 2001a) detailed the Late Wisconsinan deglaciation for northwestern New Jersey and a small part of northeastern Pennsylvania. Accordingly, deglaciation was characterized by the systematic northeastward retreat of the Kittatinny Valley and Minisink Valley ice lobes (Figure 3). This interpretation was based on the distribution of ice-marginal meltwater deposits (morphosequences) and moraines, and correlative relationships between elevations of delta topset-foreset contacts, former glacial-lake-water plains, and lake spillways. The identification of ice-retreatal positions by mapping morphosequences was first introduced in New England by Jahns (1941) and later refined by Koteff (1974) and Koteff and Pessl (1981).

Pennsylvania

White (1882) described the glacial geology of Pike and Monroe Counties, PA. The terminal moraine of the youngest glaciation was assigned a Late Wisconsinan age by Chamberlin (1883) and mapped in Pennsylvania by Lewis (1884). Williams (1893 and 1894) mapped glacial deposits of pre-Wisconsinan age in the Lehigh Valley. Leverett (1934) also assigned a Late Wisconsinan age to the terminal moraine and the glacial drift north of it, and revised its terminal limit. Crowl and Sevon (1980) remapped the terminal moraine in northeastern Pennsylvania, further refining its distribution and extent of the late glacial limit. They also concluded that the glacial deposits in eastern Pennsylvania consisted of the late Wisconsinan Olean drift, and two older deposits represented by the Warrensville drift of early Wisconsinan age and the Muncy drift of Illinoian age. Braun (2004) indicated that the terrestrial glacial record in northeastern Pennsylvania records at least four glacial advances. The youngest of Late Wisconsinan age (Marine Isotope Stage (MIS) 2 after Richmond and Fullerton (1986), the next oldest Late Illinoian (MIS 6) or pre-Illinoian B age (MIS 12) and two older drifts, the oldest pre-Illinoian G (MIS 22) based on reversed magnetic signal of proglacial lake deposits in the West Branch of the Susquehanna River Valley (Gardner et al., 1994) and a

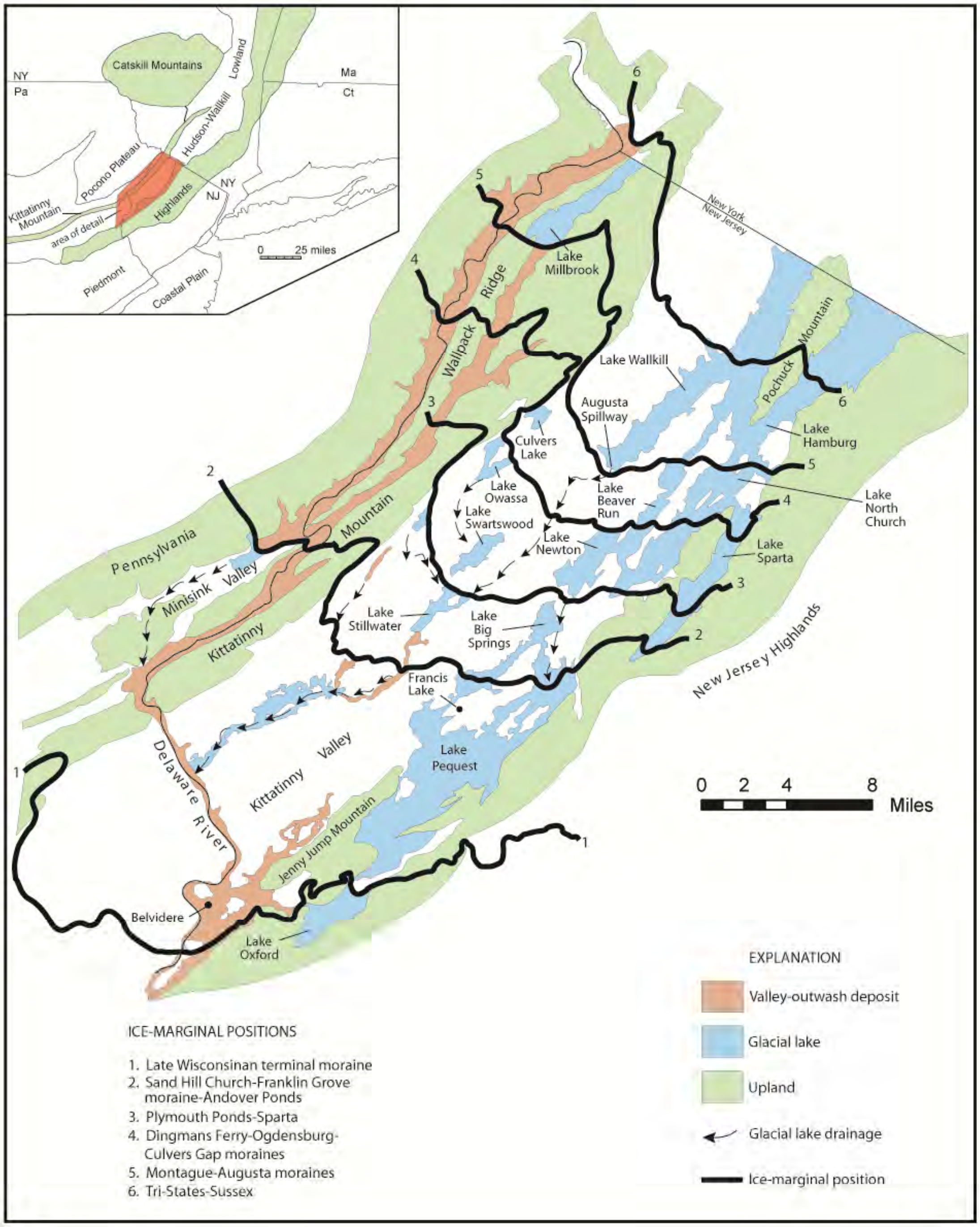

Figure 3. Major Late Wisconsinan ice margins of the Kittatinny and Minisink Valley ice lobes, location of large glacial lakes, and extensive valley-outwash deposits. Figure modified from Witte (1997).

younger pre-Illinoian D (MIS 16) based on a normal magnetic polarity of proglacial lake deposits found between the pre-Illinoian G (maximal glacial limit) and Late Illinoian or pre-Illinoian B limit (Sasowsky, 1994).

Glacial Limits and Deposits

Late Wisconsinan Glaciation

The LWGM is represented in most places by a terminal moraine that forms a nearly continuous, west-trending, low, uneven ridge of boulders and soil that sweeps across the northern part of the state from Perth Amboy to Foul Rift, NJ (Figure 3). In northwestern New Jersey the deposits of this glaciation are represented by meltwater deposits of the Rockaway Formation and till of the Kittatinny Mountain, and Netcong Formations (Stone et al., 2002). These deposits are lightly weathered, generally lie on nonweathered rock, are found in the modern valleys, and glacial landforms exhibit well-preserved constructional topography.

The terminal moraine's course divides New Jersey into two contrasting landscapes (Figure 2). North of the moraine there are many fresh to lightly weathered rock outcrops, thick stony soils, valleys filled with thick deposits of stratified sand and gravel, silt, and clay, and numerous wetlands and lakes. South of the moraine rock outcrops are sparse and weathered, soils are typically more clayey, and wetlands are sparse. Because the terminal moraine was a readily distinguishable feature of New Jersey's landscape, it was well studied around the turn of the 20^{th} century. R. D. Salisbury, in his magnum opus, *The Glacial Geology of New Jersey*, devoted 38 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. Only fifty 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, 1910-1912). This surely caused consternation among the day's geologic elitists, who viewed the study of surficial deposits as a lowly endeavor and not the proper field of study for serious scientists.

The terminal moraine in western New Jersey follows a nearly continuous looping course through Warren County (Figure 2). It consists of non-compact, bouldery, silty-sandy to sandy till with minor beds and lenses of water-laid sand, silt, and gravel (Figure 4 and 5). 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."

The age of the terminal moraine and precise chronology of deglaciation are uncertain due to a lack of appropriate organic material for radiocarbon dating, inadequacies of dating bog bottom organic material and concretions, and use of sedimentation rates to extrapolate bog bottom radiocarbon dates. Also, varved lake bottom exposures that can be used for chronology are scarce. However, thick deposits of these annual silt-clay couplets are found beneath swamp and bog deposits in the many glacial lake basins in northern New Jersey (Figure 3), would likely provide information on deglaciation history if they can be sampled.

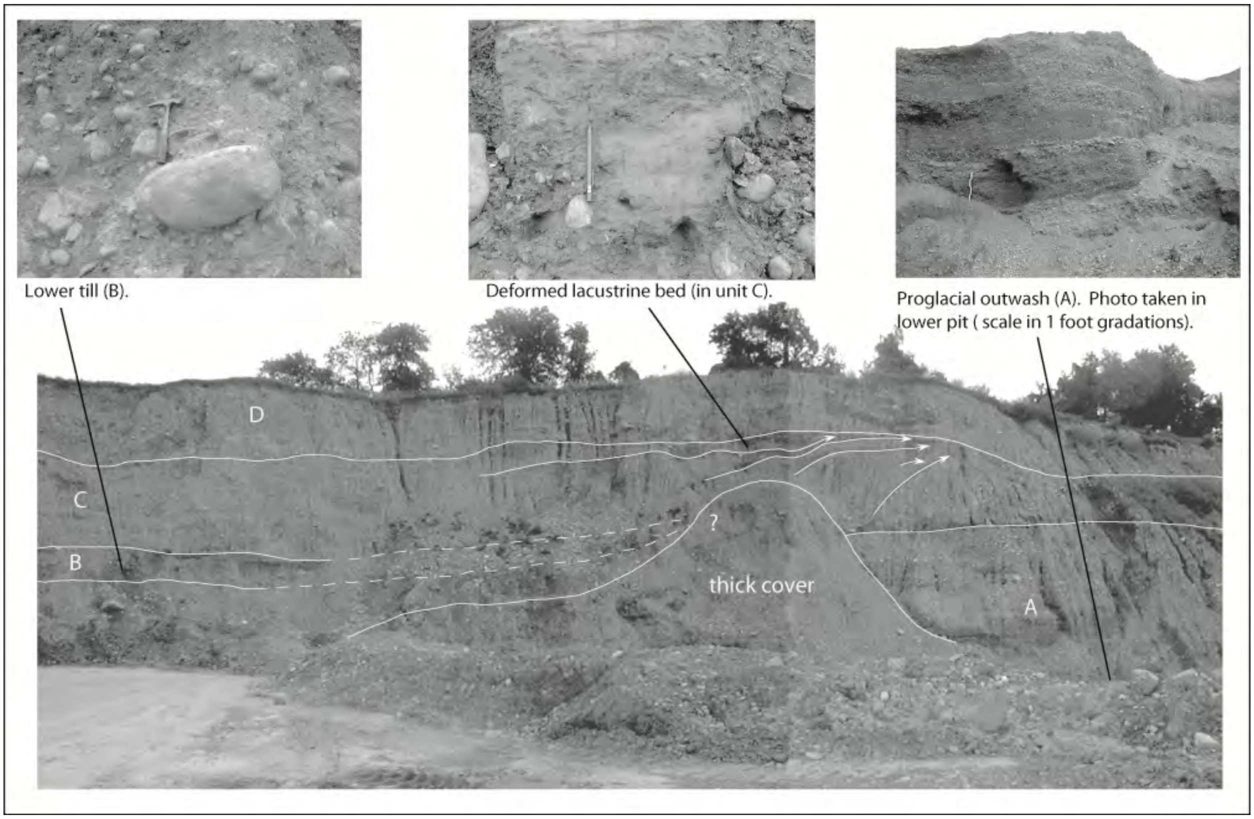

Figure 4. Composite section of the east wall, Foul Rift pit. Units: A—Proglacial outwash; B—Lower till (basal); C—Deformed outwash, arrows denote thrusts; and D—Upper till (Foul Rift moraine). Thrusts typically marked by deformed beds of silt and clay. Figure from Witte (2001b).

There are a few radiocarbon dates that bracket the age of the terminal moraine. Basal organic material cored from Budd Lake by Harmon (1968) yielded an age of 22,890 +/ 720 yr BP (I 2845), and a concretion sampled from sediments of Lake Passaic by Reimer (1984) that yielded an age of 20,180 +/ 500 yr BP (QC 1304) suggests that the age of the terminal moraine dates to 22,000 to 20,000 yr BP. Basal organic material cored from a bog on the side of Jenny Jump Mountain approximately 3 mi (4.8 km) north of the terminal moraine by D. H. Cadwell, New York State Geological Survey (written comm., 1997) indicated a minimum age of deglaciation at 19,340 +/- 695 yr BP (GX-4279). Similarly, basal organic material from Francis Lake in Kittatinny Valley, which lies approximately 8 mi (13 km) north of the terminal moraine indicates a minimum age of deglaciation at 18,570 +/ 250 yr BP (SI 5273) (Cotter, 1983). Based on the assumed age of the LWGM (22 – 21 ka) and deglaciation chronology outlined for northwestern New Jersey, the terminal moraine was deposited during an interval of about 1,000 to 1,500 years. It represents a time when the ice sheet's margin remained in a nearly constant position, neither retreating nor moving forward, except within the narrow zone marked by the moraine.

The Late Wisconsinan recessional 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), and Witte (1997 and 2001a) showed that deglaciation was characterized by the systematic northeastward retreat of the Kittatinny Valley and Minisink Valley ice lobes. This interpretation is based on the distribution of ice-marginal meltwater deposits and moraines, and correlative relationships between elevations of delta topset-foreset contacts, former glacial-

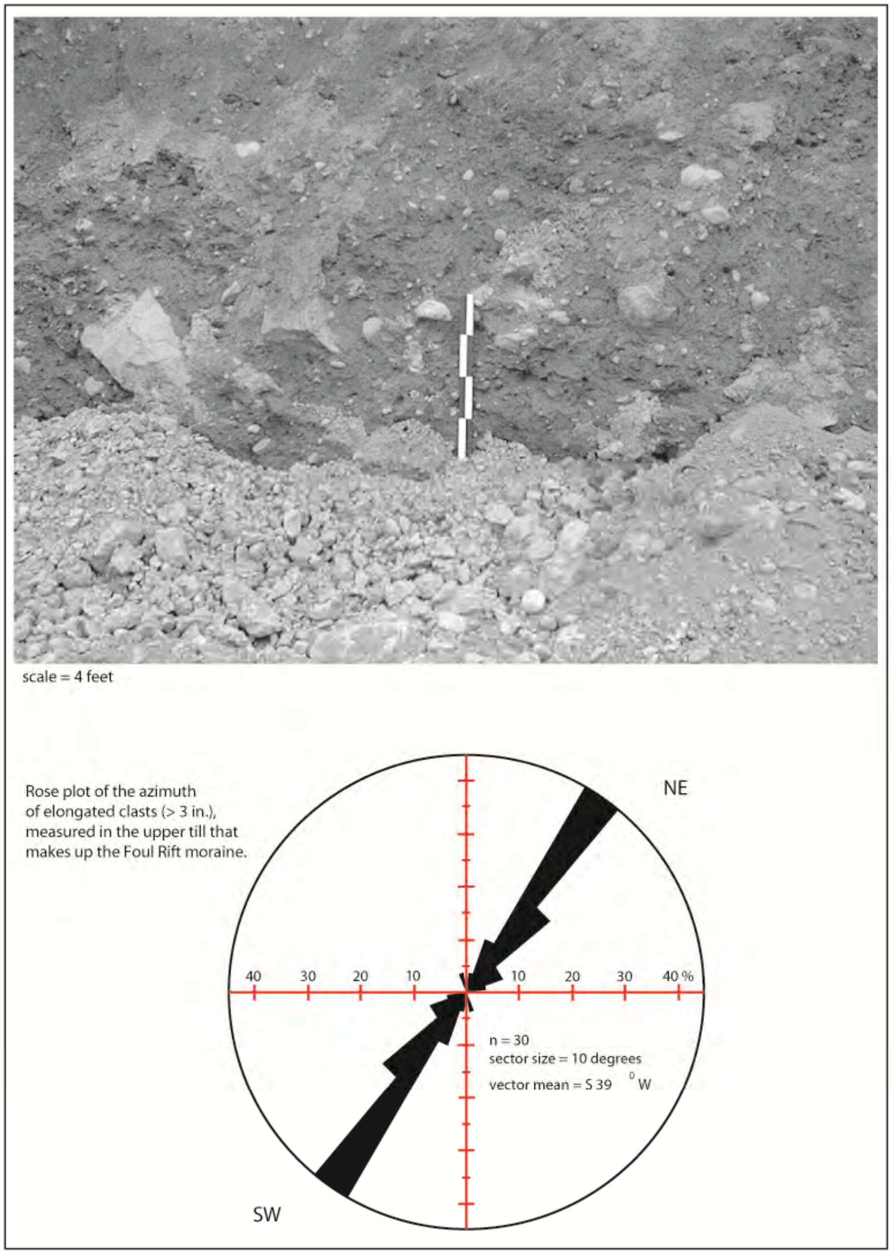

Figure 5. Upper till exposed along the north rim of the Foul Rift pit. Compact, fissile, sandy-silty matrix containing by volume 7 to 10 percent subangular to subrounded stones. Many stones are striated and elongated clasts have a pronounced downvalley fabric. Although the till forms part of the Foul Rift moraine (Late Wisconsinan terminal moraine), it has characteristics of a basal till and it may represent a subglacial till facies associated with a push moraine. Figure modified from Witte (2001b).

Outwash - Planar-bedded gravel and sand (unit 3).

Basal Till - Compact, stony, polylithic diamicton (unit 1).

lake-

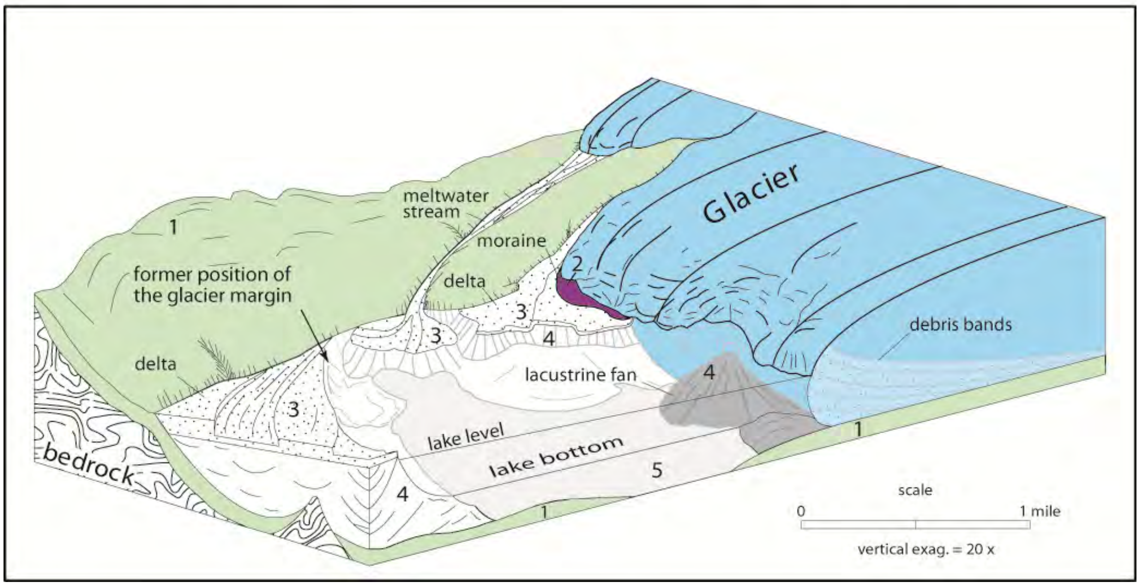

Figure 6. Composite diagram showing the depositional setting of glacial sediments. 1—Basal till; 2—Flowtill; 3—Planar beds of gravel and sand; 4—Inclined beds of sand beds of sand and some gravel; and 5—Laminated beds of silt, clay, and very fine sand.

water plains, and lake spillways. During glacial retreat, meltwater sediment was chiefly laid down in glacial lakes (Figure 6) that occupied valleys now drained by the Pequest River, Paulins Kill, and Wallkill River, and to a lesser extent in small upland basins and valleys (Figure 3). These former lake basins had been dammed by stratified drift, moraine, and stagnant blocks of ice, or by the glacier's margin.

Five ice margins (Figure 3), 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 (Witte, 1997). 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.

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. 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 ft/mi (660 and 1,531 m/km) within the first few miles (kilometers) 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 ft/mi (422 to 2,138 m/km) with a "best measure" estimated at 225 ft/mi (1,188 m/km).

Illinoian Glaciation

The Illinoian glacial maximum (IGM) lies as much as 5 mi (8 km) south of the LWGM, and follows a similar parallel course across western New Jersey (Figure 1). This glaciation is represented by the Lamington (meltwater deposits) and Flanders Formations (till) of late Illinoian age (Stone et al., 2002). These deposits (Figure 7) are moderately weathered, lie on weathered bedrock, and are found in modern valleys at slightly higher topographic positions than Late Wisconsinan glacial drift. Constructional topography is preserved in many places with moraine, delta, valley-outwash plain, and meltwater terrace land forms recognizable. Illinoian deposits in many areas are preserved on low to moderate slopes, whereas the drift on steep slopes has been stripped by erosion. Near the Late Wisconsinan border, Illinoian deposits have been found beneath Late Wisconsinan drift and in the core of some drumlins (Stone et al., 2002). Older soils have been largely stripped, although where buried by younger glacial drift and colluvium, a truncated, red Bt soil (Figure 8) presumably of Sangamon age (MIS 5) may be preserved (Ridge et al., 1990; Stone et al., 2002). Depth of carbonate leaching is as much as 12 ft (4 m), crystalline clasts (mostly gneiss) have moderately thick weathering rinds (Figure 9), and manganese staining is common, especially along the faces of soil joints. Exposed quartzite boulders commonly have a surface coating of ferro-manganese oxide. Late Wisconsinan drift near the LWGM also contains weathered clasts, but there is always fresh material present to distinguish it from Illinoian drift. The "extra-morainic drift" of Bayley et al. (1914) consists of Late Wisconsinan till that extended beyond the terminal moraine, and Illinoian till. Most of this material has been remapped as Illinoian till (Stone et al., 2002).

The southern edge of the Illinoian drift sheet in many places is poorly defined, although in places a terminal moraine and heads-of-outwash mark its southern limit. East of Denville NJ (Figure1), Illinoian deposits have not been recognized, the border covered by Late Wisconsinan deposits laid down by the Passaic ice lobe. In western New Jersey, heads-of-outwash and remnants of a terminal moraine mark the Illinoian border in the larger valleys. Elsewhere, patches of till and ice-contact stratified deposits (largely deltaic) define the border.

The similarity in the course of the Illinoian and late Wisconsinan drift borders shows that the Illinoian deglaciation proceeded similarly as the late Wisconsinan with glacial lakes forming in the valleys. An end moraine in Pequest Valley below Townsbury, NJ is the only Illinoian recessional moraine identified in western New Jersey. Bayley et al. (1914), and Ridge (1983) indicated this feature was Late Wisconsinan age, based on their finding a few lightly weathered dolostone boulders on its surface. However, closer inspection has shown that the dolostone clasts were only found near dolostone outcrops on the western side of the ridge. The dolostone clasts may have also been derived from ice-rafted debris laid down in Lake Oxford in Late

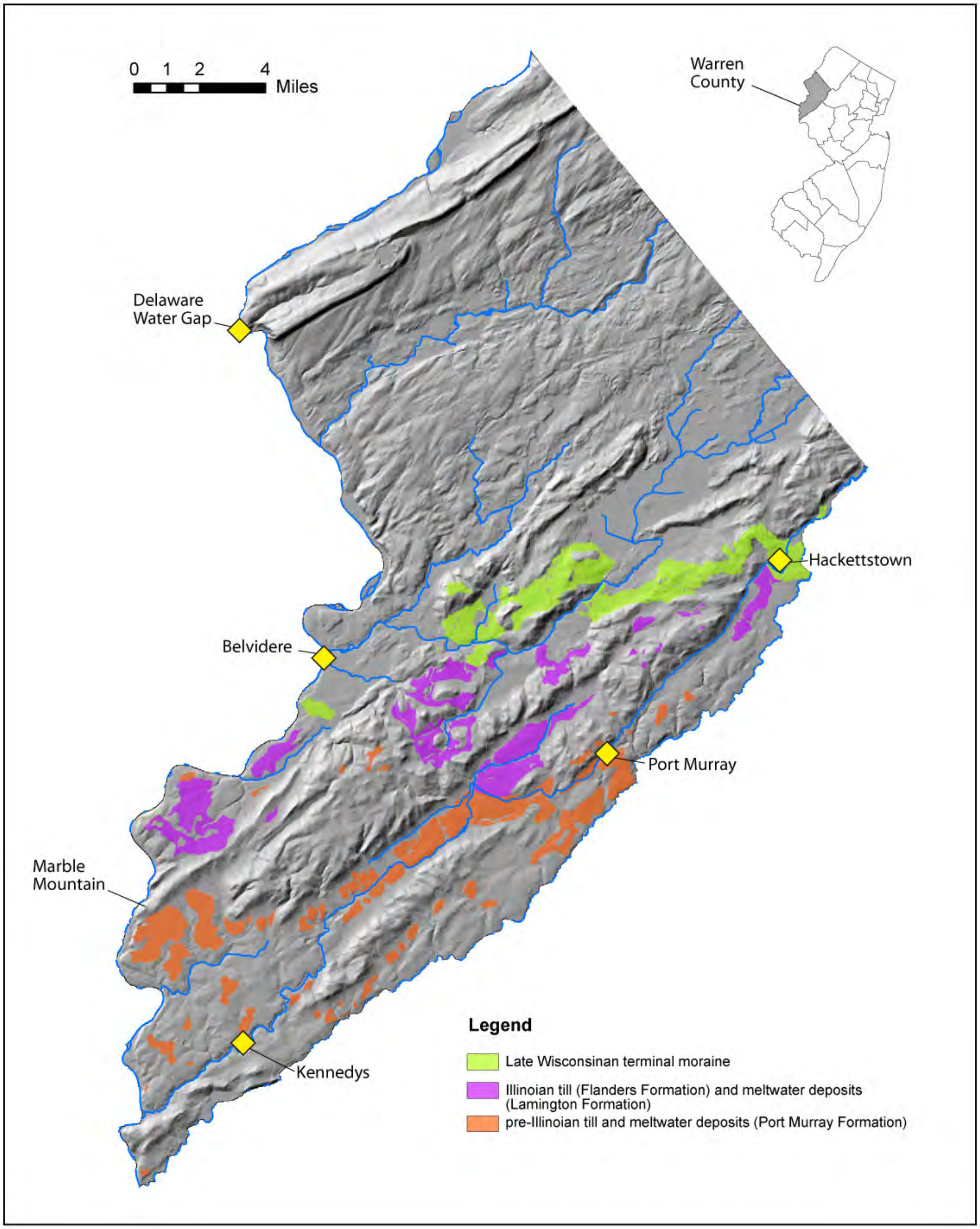

Figure 7. Distribution of Illinoian and pre-Illinoian glacial drift in Warren County, NJ. Data from Witte and Stanford (1995) and Stone et al. (2002).

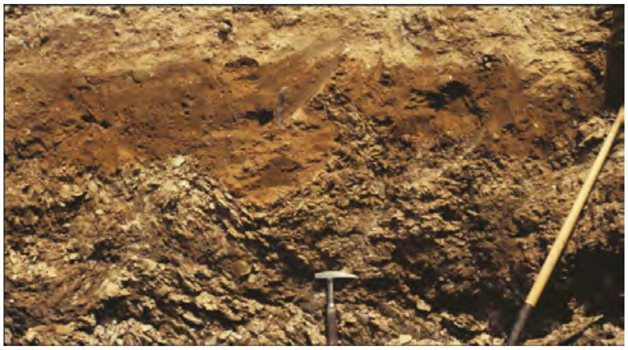

Figure 8. Truncated red Bt horizon developed in Illinoian till (Flanders Formation) overlying frost involution surface developed on weathered and colluviated Martinsburg Formation. Pit location is approximately 2 mi (3 km) outside the Late Wisconsinan limit, near Hackettstown, NJ.

Figure 9. Medium to medium-thick weathering rinds developed in gneiss cobbles (except for chert pebble in upper left). The clasts were collected from Illinoian till (Flanders Formation) that makes up a small recessional moraine in the Pequest Valley, approximately 0.5 mi (1 km) outside off the Late Wisconsinan glacial limit.

Wisconsinan time. Additionally, a 12-ft (4-m) deep trench dug at the crest of the moraine showed moderately-weathered till devoid of carbonate clasts. Based on the lack of carbonate material, the advanced degree of weathering of crystalline clasts, and pervasive ferro-manganese staining, this material compares favorably with other till in the area that has been mapped as Illinoian. An exploratory boring located near the crest of the moraine shows that the drift there is 52 ft (16 m) thick and provides additional evidence that this cross-valley ridge is a moraine rather than thin till perched on carbonate bedrock.

Pre-Illinoian Glaciation

New Jersey's oldest glaciation is represented by the Port Murray Formation (Stone et al., 2002) (Figure 1), which replaces the term "Jerseyan". The formation consists of till, till-stone lag, and meltwater deposits. It is deeply weathered, thin and patchy, and it lies on weathered bedrock (Figure 10). Deposits are generally only preserved in valleys in areas of low relief protected from erosion and more rarely in uplands where deposits have been trapped on low topographic saddles and broad low-relief surfaces. In places, these older deposits have been found beneath colluvium. In western New Jersey, most of the Port Murray Formation is preserved on weathered dolomite and slate in the Pohatcong, Musconectong, and Delaware Valleys (Figure 8). It is typically less than 15 ft (5 m) thick. Constructional topography is not preserved. The drift is not found in the modern valleys, more often lying as much as 100 ft (31 m) above

Figure 10. Till of the pre-Illinoian Port Murray Formation overlying deeply weathered Martinsburg Formation (below dashed line) exposed in an old clay pit near Port Murray, NJ. Fully decomposed gneiss cobble by arrow. The inclined contact may represent relief on the Martinsburg surface. The underlying unit may also be glaciotectonized, frozen substrate formed when the pre-Illinoian glacier overrode deeply weathered Martinsburg more than 800,000 years ago.

modern valley floors in areas that are protected from mass wasting and fluvial erosion. In most places these older deposits appear to be till. However, after a long and complex weathering history most of these polymictic materials become clayey, slightly stony diamictons. Original characteristics are often difficult to discern. However, the strong scientific consensus is that these materials are of glacial origin, based on clast provenance and position on the landscape

The southern limit of the pre-Illinoian glaciation (pIGM) is based on the most southerly occurrence of thin, deeply weathered patchy till, till-stone lag, and a few stratified deposits and erratics. In most places it is poorly defined, its trace a "best guess" between patches of the Port Murray Formation and erratics. The pIGM lies as much as 15 mi (24 km) south of the IGM, following a westward trending course from Denville to Holland, NJ (Figure 1). Similar to the Illinoian deposits, pre-Illinoian glacial deposits have not been recognized east of Denville.

Figure 11. Weathered silty clay bed of the Port Murray Formation exposed in a small gravel pit near Kennedys, NJ. The 10-in (25-cm) thick bed was shown to have a reversed magnetic polarity indicating that it is older than 788 ka (J. C. Ridge, Tufts University, written comm., 1998). The planar-bedded material here may also be of non-glacial origin possibly deposited by the Delaware River when its course was farther east than its present configuration. The small normal fault (center of photo) may have formed because of glacial collapse or it is neotectonic in origin.

The age of this oldest and most extensive glaciation is uncertain. The Port Murray Formation is correlative with the oldest glacial deposits in eastern Pennsylvania, which are older than 788 ka based on the reversed magnetic polarity of lake-bottom deposits cited in Braun (2004). As previously discussed, Braun (2004) assumes a pre-Illinoian G (850 ka) age for these deposits whereas Stanford (1997) favors a pre-Illinoian K (2.1 Ma) age. Samples collected from a silty-clay bed in a deeply weathered glaciofluvial deposit in the Pohatcong Creek valley near Kennedys (Figure 1 and Figure 11), show that these sediments were laid down during a period of reversed magnetic polarity (J.C. Ridge, Tufts University, written comm., 1998). This places the age of the deposits at older than 788 ka. Based on the correlation of continental glaciations in the northern hemisphere with the offshore oxygen isotope record this deposit may have been laid down during Late Pliocene (2.1 Ma) or during the middle part of the Pleistocene (1.1 or .85 Ma). If the deposit is not outwash, but rather alluvium laid down by Pohatcong Creek or the Delaware River, its position within the belt of older glacial drift shows that the older drift is the same age or older than the suspected outwash.

Due to the scant distribution and poor preservation of the Port Murrayan deposits, and lack of recognizable recessional deposits, the history of the pre-Illinoian deglaciation is problematic. If deglaciation proceeded as it did in the late Wisconsinan, then ice-recessional positions may have been marked by the heads-of-outwash of glaciofluvial deposits laid down in the Musconetcong and Pohatcong Valleys, and ice-contact deltas laid down in glacial lakes in Kittatinny Valley. Except for a few high-standing remnants in the lower parts of the Musconetcong and Pohatcong Valleys (Witte and Stanford, 1995) all pre-Illinoian stratified deposits have been removed by erosion.

In eastern Pennsylvania thick belts of older glacial material and a few ice-contact stratified deposits near Emmaus and Hellertown define the terminal edge of the pre-Illinoian ice sheet.

Lake clays observed by Williams (1893) near Emmaus and Allentown, PA confirmed the existence of an ice-dammed lake named glacial Lake Packer in the lowland now drained by Little Lehigh Creek. A lake would have also existed in Saucon Valley up until the time Saucon Gap was uncovered by ice and the lake drained into the Lehigh Valley. Because glacial lakes provide much of the basis for deglaciation chronologies in New Jersey and Pennsylvania there may be an opportunity to define part of the pre-Illinoian deglaciation in eastern Pennsylvania, an opportunity that's lacking in western New Jersey.

Tracing Glacial Limits into and across Eastern Pennsylvania

Glacial limits in western New Jersey (Stone et al., 2002) are largely accordant with limits in Pennsylvania (Braun, 2004) with the exception of the third oldest (pre-Illinoian D) glaciation, which is not recognized in New Jersey (Figure 1).

Late Wisconsinan

In New Jersey the outer edge of the terminal moraine marks the LWGM in most places. The moraine follows a nearly continuous looping course through western New Jersey. Morainal topography is typically 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 some places a narrow belt of Late Wisconsinan till extends as much as 3,000 ft (914 m) beyond the terminal moraine. Tracing the LWGM into eastern Pennsylvania, there is a very close accordance with the limit showed in Braun (2004).

Similar to the Late Wisconsinan terminal moraine in New Jersey, the terminal moraine traces a lobate westward and northwestward course through Pennsylvania. In most places it marks the LWGM, sharply separating the lightly weathered, drab-looking glacial drift to the north of it with the weathered, brighter-colored glacial drift to the south. The moraine is generally characterized by constructional hummocky topography throughout most of eastern Pennsylvania (Figure 12) except where ice crossed Kittatinny Mountain, Godfrey Ridge, and other significant bedrock landforms (Epstein, 1969). From Riverton (located across the Delaware River from Belvidere, NJ), the moraine trends northwest near the Villages of Ackermanville, Bangor and Rosetto (Braun and Ridge, 1997). The

Figure 12. Hummocky constructional moraine topography, Late Wisconsinan terminal moraine, Riverton, PA.

Figure 13. Stone row south of Kittatinny Ridge dominated by Shawangunk cobbles and boulders (note Bloomsburg Formation cobbles in center of photo).

Figure 14. Poorly sorted Late Wisconsinan till near Bangor, PA (rock hammer in center of photo for scale).

Figure 15. Fresh to lightly weathered shales and siltstones in kame pit north of Saylorsburg, PA.

moraine forms a reentrant northwest of the Big and Little Offsets in Kittatinny Mountain and crosses the ridge near Fox Gap (Epstein, 1969). The moraine follows the north flank of Kittatinny Mountain southwest to Saylorsburg (Epstein, 1969), where constructional knob and kettle topography is also well developed. The LWGM curves north to Sciota and then northwest towards Tamaqua (Braun, 1996a). Glacial till varies in thickness from being too thin to map over the crest of Kittatinny Ridge at Fox Gap (Epstein, 1969), to thicknesses greater than 100 ft (31 m) in valleys (Braun and Ridge, 1997). South of Kittatinny Ridge, cobbles and boulders in till are dominated by the Shawangunk quartzite, with scattered clasts derived from the Bloomsburg Formation present in most localities (Figure 13). The moraine is marked by higher concentrations of cobbles and boulders, but is characteristically poorly sorted (Figure 14). Till and meltwater deposits deposited north of Kittatinny Ridge are dominated by locally derived Devonian lithotypes including weathered shales, siltstone, sandstones, and carbonates (Figure 15).

Outwash terraces and valley trains have been mapped in many of the major drainages extending away from the ice margin including Oughoughton and Martins Creek (Ridge, 1983; Ridge et al., 1992; Braun and Ridge, 1997) and the Delaware River (Crowl, 1971; Ridge, 1983; Witte, 2001a; Medford et al., 2011). Stratified fluvioglacial deposits associated with deglaciation are present south of Kittatinny Mountain in the Jacoby Creek watershed northwest of Portland, PA (Ward, 1938; Epstein, 1969) and north of Kittatinny Mountain in Cherry Valley between Kittatinny Mountain and Godfrey Ridge (Epstein, 1969). Ice disintegration kames are common both as large mounds in the center of the valley (Figure 16) and also as kame terraces deposited between stagnant ice and the valley sides (Figure 17). Kames consist of stratified sand and gravel (Figure 17) and have been mined for aggregate by individual landowners

Figure 16. Kame (low hill, photo's center) in Cherry Valley, PA.

Figure 17. Stratified kame sand and gravel exposed in a kame terrace deposited against the south flank of Godfrey Ridge in Cherry Valley, PA.

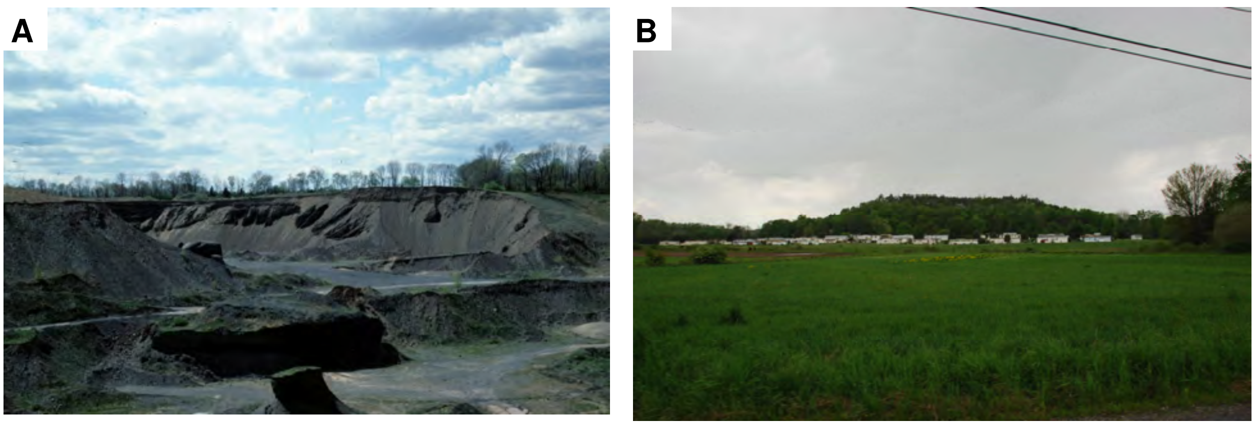

Figure 18. A—Ice-contact ("kame") delta near Sciota, PA; B—Esker 'upstream' of kame delta shown in A, Sciota, PA.

(Cherry Valley) and commercially (Jacoby Creek). Ice flow was clearly influenced by these prominent ridges during the LWGM, and the topographic influence likely became more pronounced as the ice sheet thinned. As the ice sheet wasted towards the northeast during deglaciation, the ridges may have stranded ice south of Kittatinny Mountain in the Jacoby Creek Valley and also in Cherry Valley between Kittatinny Mountain and Godfrey Ridge. Whether the kames represent heads of outwash associated with temporary ice margin positions, or whether there were multiple ice blocks is not entirely clear, but the elongate nature of several of the kames suggests that ice wasted towards the northeast as a solid mass (Epstein, 1969).

The most impressive stratified deposits in the area consist of a large ice-contact ("kame") delta and esker near Sciota (Epstein, 1969; Epstein and Epstein, 1969; Figures 18A and B). The esker is discontinuous and most pronounced immediately upstream of the delta (Figure 18B). The delta (originally a lacustrine fan) prograded from a subglacial meltwater tunnel into glacial lake Sciota, an ice-marginal lake formed between the terminal moraine and the retreating ice margin (Epstein, 1969; Epstein and Epstein, 1969). Although the delta-esker pair shown in Figures 18A and 18B is the largest and most prominent, Epstein (1969) mapped and described multiple ice-contact ("kame") deltas that prograded into a dynamic Lake Sciota whose level decreased as lower outlets were uncovered by ice retreat.

Illinoian

The Illinoian ice sheet (Figure 1) reached its most southerly limit on the north side of Marble Mountain during the late Illinoian stage. In western New Jersey Illinoian till is represented by the Flanders Formation. (Qit) is preserved on many hillslopes, and it lies in the modern river valleys. In some places its surface is marked by a truncated red soil of presumably Sangamon age. Elsewhere, the soil has been stripped by colluviation. Generally clasts and matrix material are moderately weathered, and dolostone and limestone clasts are leached to depths of at least 12 ft (4 m).

Well preserved, moderately weathered till and ice-contact outwash delineate the IGM in the Delaware Valley and the northwestern edge of the New Jersey Highlands. The location of high-standing meltwater deposits underlain by thick clay and silt north of Marble Mountain and Chestnut Hill show that the Marble Mountain Gap was probably blocked by a tongue of ice during the Illinoian maximum. This resulted in the formation of a high-standing glacial lake during the early phase of glacial retreat from the Illinoian terminal position. Illinoian deposits in the Delaware Valley in the vicinity of Marble Mountain appear significantly more weathered due to rubification of matrix material and staining of clasts than deposits farther east in the New Jersey Highlands. These differences result from differences in parent material with the more carbonate-rich drift in the Delaware Valley appearing to be more weathered than the crystalline-rich (gneiss and granite) drift found in the Highlands. Also, insitu weathering of drift in the Highlands was highly attenuated due to colluviation in areas of higher relief, s compared to the moderately flat valley floors.

Tracing the IGM into Pennsylvania, there exists a close accordance with the limit showed by Braun (1996a, 2004), based on Illinoian (MIS 6) deposits (or pre-Illinoian B (MIS 12)) mapped near Easton, PA. The Illinoian limit in eastern Pennsylvania is defined by heads-of-outwash in valleys and patchy till on uplands, the limit generally paralleling the LWGM and as much as 7 mi (11 km) beyond. Deposits are deeply weathered (10 m or more) and constructional topography has been removed or greatly subdued by erosion Braun (2004). Overall these deposits appear to be weathered greater than equivalent deposits in western PA (west of the Salamanca re-entrant) (Braun, 2004) and western NJ (personal commun, 1987). Based on degree of weathering and preservation Braun doubts that only 150,000 yrs. have passed since deposition and favors a pre-Illinoian B age (440 ka). Nonetheless, our examination of till mapped in Eastern Pennsylvania by Braun (1996a) as possibly pre-Illinoian B, appears to be too fresh to be 440,000 yrs. old. Weathering differences of the Illinoian till are largely due to lithologic variation and that much of the Illinoian in eastern Pennsylvania (Lehigh Valley) was deposited on the Martinsburg Formation where it was more easily eroded by colluviation and congeliturbation.

The Illinoian drift in Pennsylvania is well represented just south of Mud Run in Forks Township, PA, where large concentrations of boulders from the till have been assembled in large continuous stone rows. The cobbles and boulders in these stone rows are clearly more weathered than those found in stone rows developed from tills of Late Wisconsinan age, however they are remarkably unweathered if one were to assume they were pre-Illinoian B age (Figures 19A and B). The Garr Road site in Forks Township overlies carbonate rocks situated approximately 700 to 1,000 ft (213 to 305 m) south of the contact with the Martinsburg escarpment. This is the exact same geologic setting as exists near Riverton, PA where the

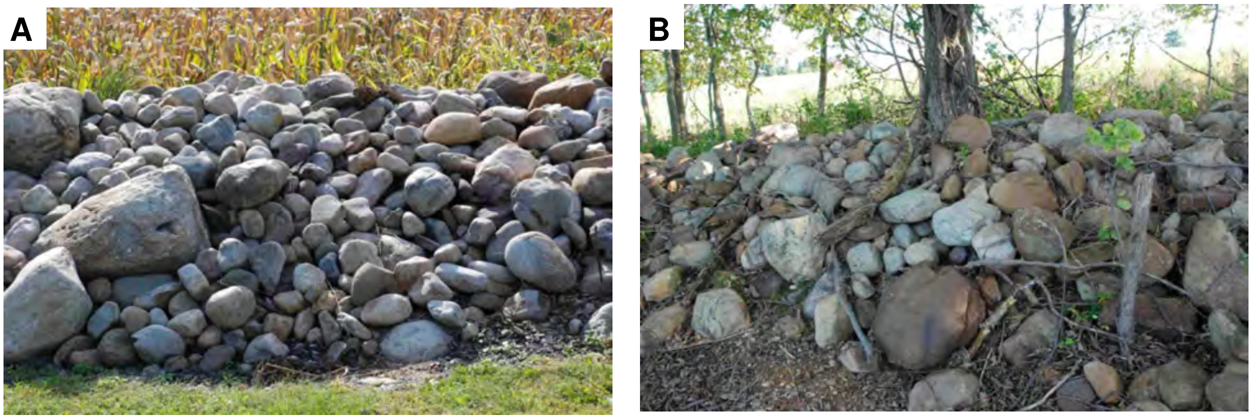

Figure 19. A—Typical stone row composed of Late Wisconsinan boulders (west of Riverton, PA on Martins Creek-Belvidere Highway); B—Stone row composed of Late Ilinoian boulders (Forks Township, PA on Garr Rd.). Note ferromagnesian weathering rind on many boulders.

LWGM deposits are extensive (overlying carbonate rocks situated approximately 700 to 1,000 ft (213 to 305 m) south of the contact with the Martinsburg escarpment). Because the Wisconsinan and Illinoian ice sheets would have overrun the same lithologies in each location, these sites provide a good opportunity to compare the degree to which weathering has altered the rocks without the typical complication of comparing the weathering of disparate lithotypes. We sampled 93 rocks in stone rows at each of these locations selected at 10-in (25-cm) intervals in four 43-ft^2 (1-m^2) grids (some grid intersections landed on vein quartz, shattered Martinsburg slate, or two grid intersections covered the same rock). The majority of the lithologies sampled at each site were Shawangunk or other sandstone cobbles, and four clasts of the Bloomsburg Formation were included in each sample. Weathering rinds were measured at the mm scale on all specimens that had clear oxidation rinds. Cobbles that were weathered through were arbitrarily assigned a rind thickness of 35 mm to separate them from the rocks having discrete rinds. Figure 20 left shows the frequency distribution of rind thicknesses at the two sites and Figure 20 right, shows a representative sample of specimens from each of the two sites included in the rind data along with a set of cobbles collected in till in Bethlehem Township Pa, well south of the Late Illinoian border. Although the samples collected at the Mud Run site are clearly more weathered than those from the LGWM site in Riverton, they are far less weathered than those collected further south in Bethlehem Township.

Figure 20. Left—Weathering rind thickness on clasts collected from stone rows near Riverton, PA (LWGM deposits) and near Mud Run (LIGM deposits). Clasts labeled 30-35 mm are actually weathered throughout the interior; Right—Examples of samples represented in Figure 20a. Note the relatively fresh nature of specimens labeled as late Illinoian relative to those labeled pre-Illinoian. The labels are situated on a Bloomsburg clast for comparison for each group.

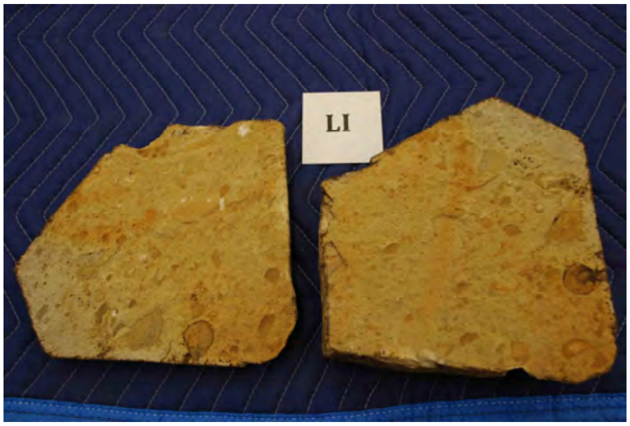

Figure 21. Fossiliferous specimen of the Middle Devonian Hamilton group from Illinoian till near Mud Run. Fossils are very well preserved in this sample.

Morris Metz, the landowner at the Mud Run site, collected a partially weathered fossiliferous cobble in the field adjacent to the stone row that contains a fossil assemblage characteristic of the Middle Devonian Hamilton that only crops out north of Kittatinny Ridge (Andy Bush and David Sunderlin, Pers. Comm., 2012). The rock is weathered, but too fresh, we believe, to represent a till stone of pre-Illinoian B age (Figure 21). Therefore, based on the high concentration of relatively unweathered cobbles and boulders near the glacial border in the Mud Run drainage basin, the similarity in weathering between these cobbles and those mapped in New Jersey as Late Illinoian, and the disparity in weathering compared to rocks picked from till approximately 6 mi (10 km) south in Bethlehem Township, we believe the border mapped as Illinoian or pre-Illinoian in Figure 1 is most likely late Illinoian in age rather than pre-Illinoian B.

Pre-Illinoian

The pre-Illinoian glacial maximum (pIGM) in New Jersey is based on the most southerly occurrence of thin, deeply weathered patchy till, till-stone lag, and a few stratified deposits collectively known as the Port Murray Formation (Stone et al., 2002) and erratics. In western New Jersey the limit lies just south of Musconectong Mountain crossing the Delaware River near Holland, NJ (Figure 1). Weathered deposits of Triassic fanglomerate make the identification of pre-Illinoian glacial drift in this area difficult. Alternatively, patches of weathered till just north on Musconetcong Mountain (New Jersey Highlands) show a maximum glacial position nearby.

Tracing the pIGM into Pennsylvania, there exists a close accordance with pre-Illinoian boulder and cobble lags mapped by Braun (1996a), near Monroe, PA. Similar to New Jersey, the identification of glacial drift in areas of fanglomerate is problematic. Alternatively, some of these lags may be weathered pre-Illinoian outwash given their position along and above the Delaware River. Farther west the limit is traced to the pseudo-moraine areas in Saucon Valley and the Lehigh Valley (Braun, 2004).

Braun (2004) also defines a second pre-Illinoian limit in the Lehigh Valley based on a thick belt of drift about 6 mi (10 km) north of the glacial maximum. Similar to the older deposits farther south they are deeply weathered and poorly preserved. However, in a few places pseudo-moraine topography is persevered. Tracing the limit eastward it wraps around the north end of Morgan Hill, PA crossing the Delaware Valley just south of Phillipsburg, NJ. Continuing an eastward trend and adjusting for topography, the limit would fall near the reversed polarity site (discussed previously) near Kennedys, NJ, too close to determine its location inside or outside of the limit. In New Jersey, thick deposits of pre-Illinoian glacial drift and/or pseudo-moraine topography have not been observed in areas underlain by dolomite and limestone. This limit is untraceable in New Jersey. For now it's an unresolved issue of whether there existed a pre-Illinoian D glaciation.

In Pennsylvania evidence of a glacial event that extended to the position of Braun's pre-Illinoian G margin was recognized as early as 1893 by Williams. Williams also postulated the existence of proglacial Lake Packer, an ice dammed lake formed between the ice margin on the northeast in Lehigh Valley and the Reading Prong hills to the southwest in the Little Lehigh drainage basin. He based Lake Packer on lacustrine clays that he called the Packer Clay and topographic enclosure related to ice damming. Williams (1893) and Leverett (1934) mapped a moraine that, more or less, mirrors Braun's (1999) pre-Illinoian G margin, based on tills and stratified drift. The Saucon Valley drainage basin is bounded by Reading Prong basement blocks such as South Mountain on the north and other Reading Prong blocks on the west, south and east, and drains to the northeast through a narrow gap between South Mountain and Green Hill, into the Lehigh River. Early workers *(*Leverett, 1934; Miller et al., 1939) mapped a moraine looping across the south central portion of Saucon Valley through the south end of Hellertown and Bingen. Based on topographic and drainage relationships, and clays mapped in the Saucon Valley, Braun (1999) suggested that a proglacial Lake bounded by the ice margin to the north and the reading Prong Hills to the south and west, formed in the Saucon Valley similar to Lake Packer to the west. One of us (Germanoski) encountered clean clays beneath the colluvial wedge prograding from the south side of South Mountain in a heat pump well drilled near Summit Lawn that that may be representative of these lacustrine clays. This ice-marginal lake would have formed as ice advanced into the Saucon Valley, blocking the South Mountain-Green Hill gap. The lake sediments were probably overlain by till as the ice advanced to its maximum southernly extent (pre-Illinoian G), with a spillway over the southern drainage divide into Cooks Creek (Braun 1996b). As ice retreated out of the Saucon Valley, the lake would have persisted until the drainage outlet was breached at the South Mountain-Green Hill gap. This corresponds with Braun's pre-Illinoian D boundary (Figure 1). Therefore, as Braun (1999) acknowledges, his pre-Illinoian D boundary may represent a recessional position of the pre-Illinoian ice sheet. This interpretation may also explain the lack of a well-defined counterpart to this ice margin in western New Jersey where pre-Illinoian proglacial lakes have not been identified.

The Pre-Illinoian G ice margin is reasonably well marked in Pennsylvania by a number of ice-contact stratified deposits that may be ice-contact ("kame") deltas in the Hellertown and Allentown East quadrangles (Braun, 1996b, 1996c, 1999). One of the more striking deposits is exposed between Emmaus and the northwest flank of South Mountain (Braun, 1999). This delta would be a pre-Illinoian G analog to the Late Wisconsinan Sciota proglacial delta described earlier, except that this proglacial lake was trapped between the ice margin and bedrock topography rather than an ice margin and a terminal moraine. Material in the Emmaus deposit (Figure 22) is deeply weathered, far more so than the Late Illinoian material and other pre-Illinoian material described in this summary. Individual clasts, lack physical strength and matrix was largely derived from the weathering of primary

Figure 22. Deeply weathered material in a kame delta between South Mountain and Emmaus (photo courtesy of Frank Pazzaglia, Lehigh University).

materials (Braun, 1999). These deposits are heavily weathered in part because they were originally coarse, permeable sand and gravel.

Summary

There is a remarkable consistency between glacial margins mapped on either side of the Delaware River given that the scope of work here spans more than 130 years and includes the work of many geologists. Dates of glaciations have changed as more has become known of continental glaciations in North America. Current thinking indicates that eastern Pennsylvania and New Jersey have been glaciated at least three times over the last 2.1 million years. These are the Late Wisconsinan (21 ka), Illinoian (150 ka), and pre-Illinoian (855 ka or older). Comparison of Illinoian till and outwash across the river from New Jersey to Pennsylvania shows highly variable weathering, an indication that composition, topographic position, and permeability greatly influence apparent age. In studies where these differences were kept to a minimum (Garrs Road site), measured weathering rinds suggest a more likely Late Illinoian age (150 ka) rather than the pre-Illinoian B age (440 ka) for the glaciation that lies just outside of the LWGM. However, it must be said that without absolute dates it comes down to "in the Delaware Valley, Illinoian clasts look too fresh to be sitting around for 440,000 years and they nearly occupy the same topographic position as Late Wisconsinan glacial drift."

The lack of a pre-Ilinoian D margin in New Jersey suggests that the margin mapped in Pennsylvania behind the glacial maximum limit may be a pre-Illinoian G recessional position. A study on the magnetic polarity of proglacial lake sediments in the Lehigh Valley area could resolve the legitimacy of the p-ID margin and age of the oldest glacial deposits (p-IG or older). Similarly, a re-investigation of the proglacial lake sediments beneath Budd Lake in New Jersey may establish the age of the oldest glacial deposits

Lastly, the terrestial record in terms of glacial landforms has been consistent from glacial episode to glacial episode. The strongly developed northeast-southwest topographic grain of eastern Pennsylvania and western New Jersey has shaped ice flow dynamics and the nature of the glacial deposition by channeling ice into lobes, forming moraines along lobate glacial margins, stranding ice during deglaciation to varying degrees, and in particular facilitating the development of proglacial lakes and deposition of ice-contact (kame) deltas. Except for a few moraines, these lake deposits provide the clearest record of terminal limits and deglaciation and although this record is greatly obscured in areas of older glacial episodes, parts of it remain. What little we do know suggests that glaciation proceeded in the same manner regardless of age and that long interglacial periods of weathering and erosion have made the history of pre-Late Wisconsinan glaciations a challenge to understand.

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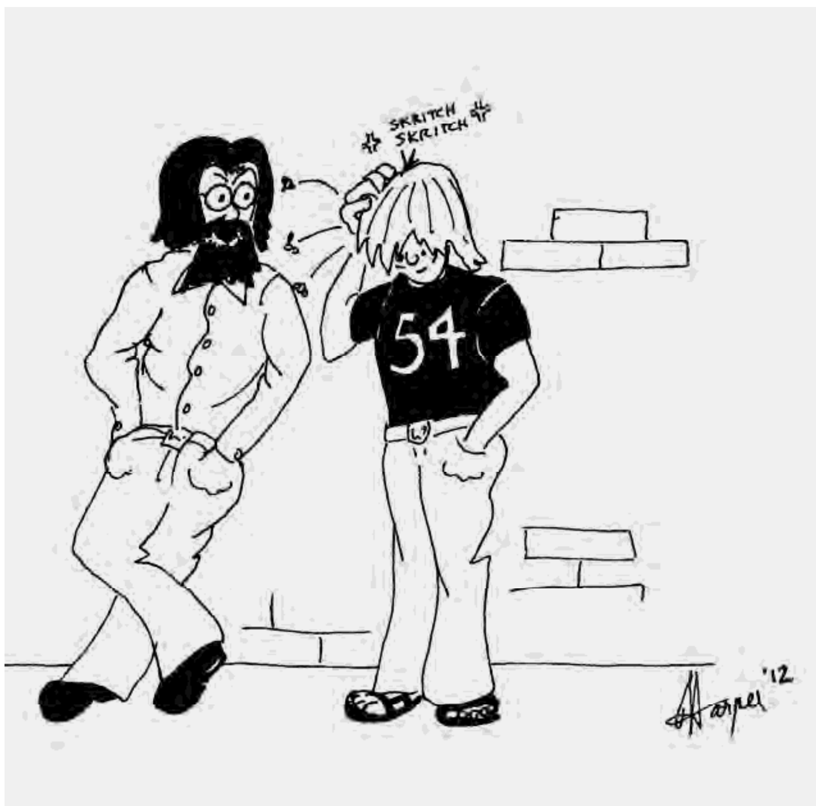

Hey, man, quit buggin' me!

BREACHING THE MOUNTAIN BARRIER: HISTORY AND ENGINEERING GEOLOGY OF THE SHAWANGUNK MOUNTAIN RAILROAD TUNNELS, ORANGE AND SULLIVAN COUNTIES, NEW YORK

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Introduction

Shawangunk Mountain in New York constitutes the northern 44 mi (71 km) of a remarkable topographic ridge that bounds the northwestern margin of the Appalachian Great Valley for 244 mi (388 km) from Franklin County, PA, ~10 mi (~15 km) north of the Maryland State Line, to Rosendale, Ulster County, NY. The 144 mi- (232 km-) long ridge in Pennsylvania, breached by numerous water gaps (see Stops 1, 3, and 4 of this guidebook), is called Blue Mountain. The 34 mi- (55 km-) long ridge in New Jersey, known by the very appropriate Lenni Lenape Indian name of Kittatinny ("Endless") Mountain, extends northwest from Delaware Gap into New York State, deeply notched only at Culvers Gap, a prominent preglacial wind gap in Sussex County (see Witte, 2001). Shawangunk Mountain is an uninterrupted homoclinal ridge (with a few wind gaps) northeast from the New Jersey State Line for 30 mi (48 km) to near Ellenville, Ulster County, but from there to its termination broadens out as a result of several Alleghanian folds. (Shawangunk is also apparently of Lenni Lenape derivation, meaning "there is smoky air" and perhaps referring to the Dutch burning of a Munsee Indian fort in 1663 [Wikipedia, 2012c; but see Fried, 2005].) Throughout its entire length, the Blue-Kittatinny-Shawangunk Mountain ridge is underlain by Silurian Tuscarora-Shawangunk Formation quartzite and conglomerate on its summit and northwest slope and Ordovician Martinsburg Formation shale and sandstone on its southeast slope. (The above statement must be modified slightly by pointing out that at one point at least, in the Town of Mount Hope, Orange County, NY, the Martinsburg lies at the crest of the mountain.)

Over the last two centuries a number of proposals have been made to bore transportation tunnels through Shawangunk Mountain in New York, the mountain ridge being a significant barrier between the Hudson River valley and New York City on the east and the economically important Catskill foothills and Neversink-Basher Kill-Mamakating valley on the west. However, in all this time only two such tunnels have been dug (Figure 1). The High View Tunnel was among the earliest successfully completed in the Northeast. The Otisville Tunnel was bored nearly 50 years later when tunnel technology was considerably more advanced.

Early Proposed Tunnels

In the 1820's, 10 years before railroads entered the transportation picture, a group of Orange County businessmen lobbied for the planned Delaware and Hudson (D&H) Canal to

Inners, J., Skye, S., Epstein, J., Bierly, A., and O'Hara, A., 2012, Breaching the mountain barrier: History and engineering geology of the Shaqwangunk Mountain railroad tunnels, Orange and Sullivan Counties, New York,, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 118-129.

HISTORICAL CHRONOLOGY

(Wakefield, 1970; Inners et al., 2002; Houck, 2006; Skye, 2009; Wikipedia, [undated], 2012a, 2012b, 2012d, 2012e.)

1828	**Delaware and Hudson (D&H) Canal completed between Honesdale, PA, and Kingston, NY. (Construction of the canal was initiated in 1825.)**
1832	**(April 24) The New York and Erie Rail Road is chartered in New York, the line to connect Piermont, NY, north of New York City and west of the Hudson, with Dunkirk, NY, on Lake Erie.**
1836	**Construction of the Erie Rail Road begins.**
1847	**The New York and Erie Rail Road cuts through Shawangunk Mountain at Otisville, NY. The line opens to Port Jervis in January 1848.**
1861	**The New York and Erie Rail Road is reorganized as the Erie Railway.**
1868	**The New York and Oswego Midland Rail Road is organized. The mainline runs from Weehawken, NJ, to Oswego, NY.**
1871	**The New York and Oswego Midland Rail Road completes the High View Tunnel through Shawangunk Mountain between Wurtsboro on the west and Bloomingburg on the east.**
1878	**Brick tunnel-liner installed in parts of the High View Tunnel.**
1880	**Due in part to the earlier financial machinations of Jay Gould, the Erie Railway goes bankrupt and is sold off, becoming the New York, Lake Erie and Western Railroad. The New York, Ontario and Western (Ontario and Western, O&W) Railway takes over the mainline of the New York and Oswego Midland Rail Road.**
1895	**The New York, Lake Erie and Western Railroad goes into bankruptcy, and then emerges as the Erie Railroad.**
1906-08	**The Erie Railroad bores the Otisville Tunnel through Shawangunk Mountain.**
1947-48	**Last passage of steam trains through the High View Tunnel.**
1953	**(September 10) Last O&W passenger train passes through the High View Tunnel, bound for Roscoe to the west.**
1957	**Last trains run through the High View Tunnel. The tunnel is abandoned. (Sometime later the tracks are removed.)**
1960	**(March 29) The O&W Railway is liquidated, and all assets are auctioned off. (October 17) The Erie Railroad merges with the Delaware, Lackawanna and Western Railroad to form the Erie-Lackawanna.**
1976	**The Erie-Lackawanna becomes part of Conrail.**
1983	**The Metro-North Railroad is formed to take over commuter operations of Conrail in New York State, operating the Port Jervis Line through the Otisville Tunnel.**
1999	**Conrail system is split up, with the Norfolk Southern (NS) Railroad taking over the old Erie line through Otisville. Metro-North leases the NS tracks for commuter operations down to the present day.**
2003	**Metro-North leases the NS tracks for commuter operations, with the possibility of outright purchase in the future.**
2005-2006	**New York Highway Department undertakes attempt to dewater the High View Tunnel because of concerns about subsidence beneath new NY Route 17 being constructed over the mountain above the tunnel.**

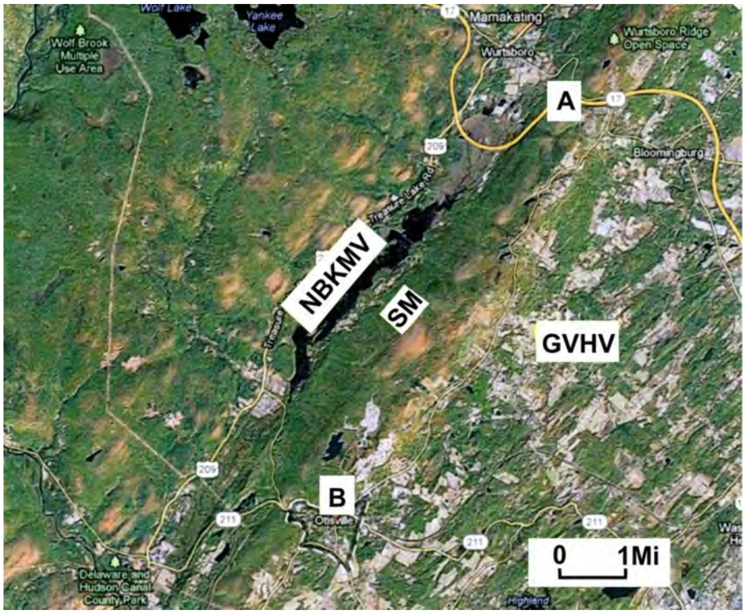

Figure 1. Map showing the locations of the High View (A) and Otisville (B) Tunnels in relation to the Neversink-Basher Kill-Mamakating Valley (NBKMV), Shawangunk Mountain (SM), and the Great Valley-Hudson Valley (GVHV) in Orange and Sullivan Counties, NY.

cross Orange County. In 1825 these gentlemen proposed that a tunnel be dug through the Shawangunk ridge and that a canal be built through Orange County to Newburgh where it would join the Hudson River. Since one of the strong backers of the D&H Canal was George Duncan Wickham, a prominent citizen of Orange County and a member of the D&H Board, the D&H Board of Managers had to treat the proposal seriously. Wickham made a motion to the Board to explore alternatives to the planned route up the valley west of Shawangunk Mountain to Kingston on the Hudson and the board approved (Skye, 2009). Benjamin Wright, the nation's foremost canal engineer was asked to explore alternatives to the Kingston route. Wright evaluated the proposed Orange County route and determined that a tunnel two miles long would be needed and that the additional cost would be prohibitive. It is worth noting that the black powder blasting technology available at the time would surely have delayed the completion of the canal well beyond the actual completion date of 1828 when the canal was opened to Kingston (Skye, 2009).

Ten years after Wright rejected the ideal of a D&H Canal tunnel, he had to consider the idea of a Shawangunk tunnel again. Wright had become the chief engineer for the Erie Railroad and had to decide whether the Erie should cross the Shawangunks at Otisville by going over the top or through a tunnel. He opted for a deep cut through the Deerfield Gap at Otisville in the route plan he completed in 1835. He did not support the idea of a tunnel at the time since the amount of traffic expected could not justify the expense of a tunnel he estimated would have to be over half a mile (>0.8 km) long. He did state in his report to the New York Secretary of State that in 20 years time the increase in the railroad's business would demand that such a tunnel be built. In 1847 the railroad finally accepted Wright's recommendation and built the line through Deerfield Gap (Figure 2; see Stop 9). In 1873 the Erie reconsidered its decision on building a tunnel, but nothing came of that effort (Skye, 2009).

Figure 2. Abandoned Erie Railroad cut through Shawangunk Mountain in the wind gap at Otisville, looking east (GPS 41°28'32.4"N/ 74°33'05.9"W). The cut is 0.15 mi (0.24 km) south of the later tunnel ~100 ft (~30 m) higher in elevation. The beds exposed are northwest-dipping (~30°) quartzitic conglomerate and sandstone of the Shawangunk Formation, the uncon-formable contact with the Martinsburg Formation being exposed at the east end of the cut (STOP – of this guidebook, p. ----).

Engineering Geology of the Shawangunk Tunnels

The High View Tunnel

In 1868, more than 20 years after the Erie rejected the idea of a tunnel, Clinton Stephens designed a tunnel through the Shawangunk ridge between Wurtsboro on the west and Bloomingburg on the east for the New York and Oswego Midland Rail Road (later the Ontario and Western [O&W] Railway). Stephens had previously done considerable work on the Erie Canal and had also contracted with the Erie Railroad. Construction of the so-called High View Tunnel began in 1868 and was completed in 1871. Work was started at both ends of the tunnel simultaneously. When both teams met in the middle, they were only a few feet off. This was quite an engineering feat at the time, especially since the tunnel is curved near the east portal (Skye, 2009).

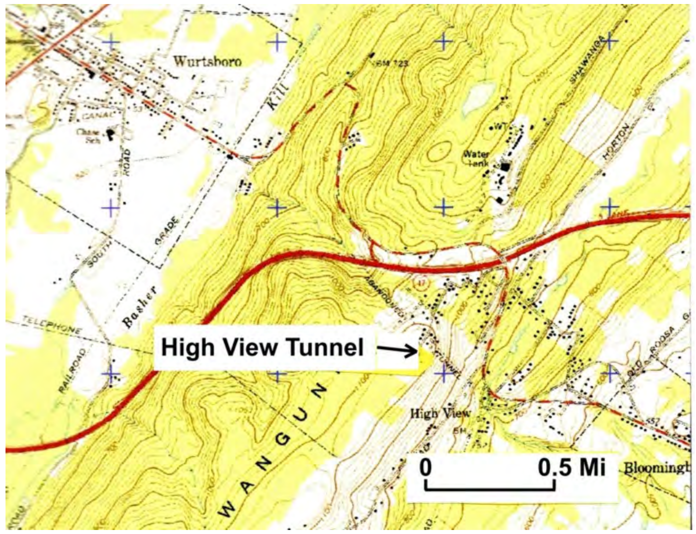

Figure 3. Location map of the O&W Railroad's tunnel through Shawangunk Mountain between High View and Bloomingburg,

The now abandoned High View Tunnel cuts through Shawangunk Mountain just south of a high wind gap ~1 mi (~1.5 km) southwest of Wurtsboro, Sullivan County (Figure 3). Elevation of the floor of the gap is about 1000 ft, the mountain rising to knobs ~1200 ft (~365 m) immediately to the northeast and southwest. Just to the southeast, and also on the line of the tunnel is a narrow spur-ridge ~100 ft (~30 m) higher than the floor of the gap. The elevation of both portals is ~840 ft (~255 m), and therefore ~250 ft (~75 m) of rock lie above the deepest part of the tunnel. Original length of the tunnel was 3,857 ft (1,176 m), cut through solid rock. The rock was excavated using steam-powered drills and black powder.

Figure 4. Southeast portal of the abandoned O&W tunnel at High View (GPS ~41°33'30.3'N/-74°27'30.6"W). Bedding in the Martinsburg shale and sandstone in the immediate vicinity of the portal is very gently west dipping.

The southeast portal of the tunnel (Figure 4) is situated several hundred feet (scores of meters) northeast of the former High View Station (Figure 5). The Martinsburg Formation exposed at the portal and in cuts to the south is predominantly evenly interstratified, thin-bedded gray shale and siltstone, exhibiting only slight internal deformation. Bedding strikes N35-65E and dips rather uniformly 10-15NW. Attitude of cleavage in the shales is ~N75E/20SE. Attitudes of the most prominent joints are N75E/90, N28E/87SE, and N2W/83E.

The present northwest portal is on the

Figure 5. High View Station, now an elegant residence, along the abandoned railroad grade a few hundred feet (scores of meters) from the east portal of the O&W tunnel. In the late 1960's, the station "brood[ed] hauntingly over a field-like thicket, a vandalized shell, nothing more than a shelter for a flock of chickens" (Wakefield, 1970, p. 38). Things have since decidedly improved.

Figure 6. Northwest portal of the abandoned O&W tunnel, about 0.8 mi (1.3 km) east-southeast of Wurtsboro (GPS 41°34'00.0"N/74°28'05.7"W). Bedding in the Shawangunk conglomerate here strikes N33E and dips 26NW. Note the brick lining (with stone facing at entrance) and water flowing from pipe in foreground.

mountainside directly downslope from the westbound lanes of recently constructed new NY Route 17 (Figure 6). Well exposed at the portal and in the rock cut to the northwest is medium - to thick-bedded, light-gray to white Shawangunk quartzitic sandstone and conglomerate. Bedding is relatively uniform, striking N33E and dipping 26 NW. Numerous large, angular quartzite blocks fill the cut just beyond the existing portal, a result of a "botched attempt" to block the original portal that shortened the tunnel ~ 20 ft (~6 m) (Houck, 2006).

The High View Tunnel was originally "jerry-built" with little regard to safety and geologic conditions (Wakefield, 1970, p. 45). Particularly troublesome were a pocket of clay encountered in the course of construction (probably related to deep weathering in the Martinsburg at the northwest-dipping contact with the Shawangunk), the fractured (jointed) nature of the Shawangunk quartzite, and constant water problems. In 1878 the Midland installed brick lining in parts of the tunnel (Figure 7), but leaving large rooms of solely rock bore construction between three lined segments (Wakefield, 1970, p. 45; Houck, 2006). In 1897, seventeen years after taking over the tunnel from the Midland, the O&W Railway attempted to further strengthen the arching of the tunnel. (The local press reported, however, that the men performing the work were exposed to almost unbearable gas and smoke conditions [Wakefield, 1970, p. 47].)

Bearing testimony to the tunnel's constant water problems are the large pool of

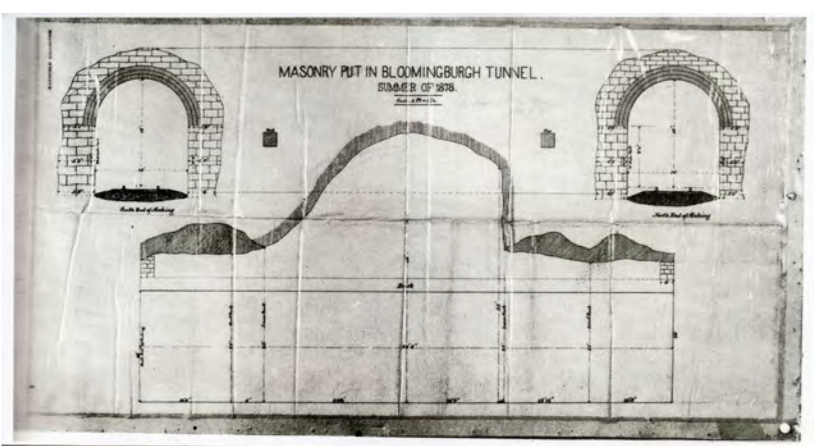

Figure 7. Engineering department drawing showing details of brick lining built to protect trains from persistent roof rock-falls in the O&W's High View-Wurtsville ("Bloomingburgh") tunnel. The August 30, 1878, Liberty (NY) Register reported "the Midland is doing a good job in the tunnel, laying a solid brick arch which will make it save for all trains to come" (Wakefield, 1970, p. 47).

Figure 8. Close-up of the southeast portal of the O&W tunnel's east portal showing gently west-dipping bedding, water in tunnel (a constant problem during railroad operations), and lack of lining at portal.

water at the southeast portal (Figure 8) and the water flowing from a similar pool into a pipe at the northwest portal (see Figure 6). The pipe was emplaced about a half-dozen years ago during construction of new NY Route 17, when New York Highway Department engineers and geologists became concerned that "the standing water in the tunnel would hasten the further deterioration of the rock bore [and thereby increase the likelihood of additional rock falls and subsidence above the line of the tunnel bore]" (Houck, 2006).

Though suffering periodic safety, engineering and geologic problems, the High View Tunnel continued to operate effectively throughout the first half of the 20th century until the tunnel was abandoned and the O&W Railway was liquidated in 1957.

The Otisville Tunnel

The idea of an Erie Railroad tunnel at Otisville lay dormant until 1906 when work finally began on the long awaited tunnel. Troubles beset the tunnel builders as they worked their way through the mountain in 1906. In August fighting broke out within the African-American crew that was digging the tunnel. The local Justice of the Peace and his constables had to be summoned to restore order. In September a blast of the explosives being used to excavate the tunnel caused the tunnel roof to collapse, killing one workman and trapping a number of others. That same month it was reported that the new tunnel was causing local water "veins" to dry up. This resulted in farmers' wells running dry and in large amounts of water spilling into the tunnel (Skye, 2009). As inscribed on the east portal, the tunnel was completed in 1908.

The Otisville Tunnel trends WNW-ESE through Shawangunk Mountain at Deerfield Gap, a short distance north of the original Erie grade over the ridge just west of Otisville, Orange County (Figure 9). The elevation of the gap is ~850 ft (~260 m) at the east end, rising gently to ~900 ft (~275 m). At the west end a mountain spur juts into the gap from the northeast, sloping off from an elevation of ~1,090 ft (~330 m) to ~1,025 ft (~310 m) at an imposing quartzite cliff high

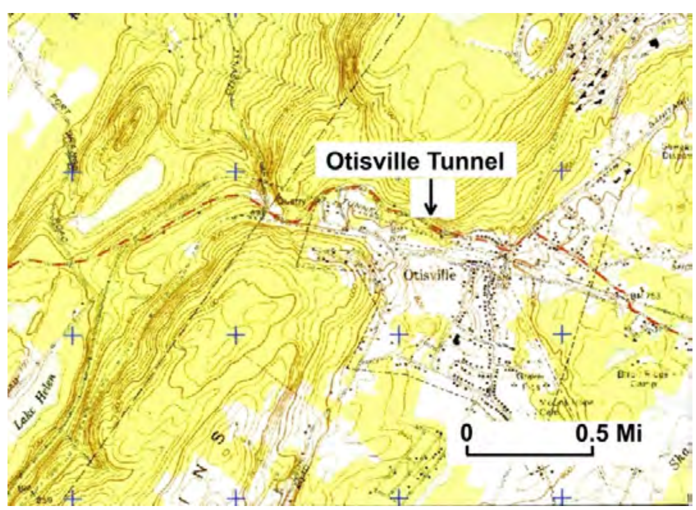

Figure 9. Location map of the Erie Railroad's tunnel through Shawangunk Mountain at Otisville, Orange Co., NY (Otisville, 7.5' quadrangle).

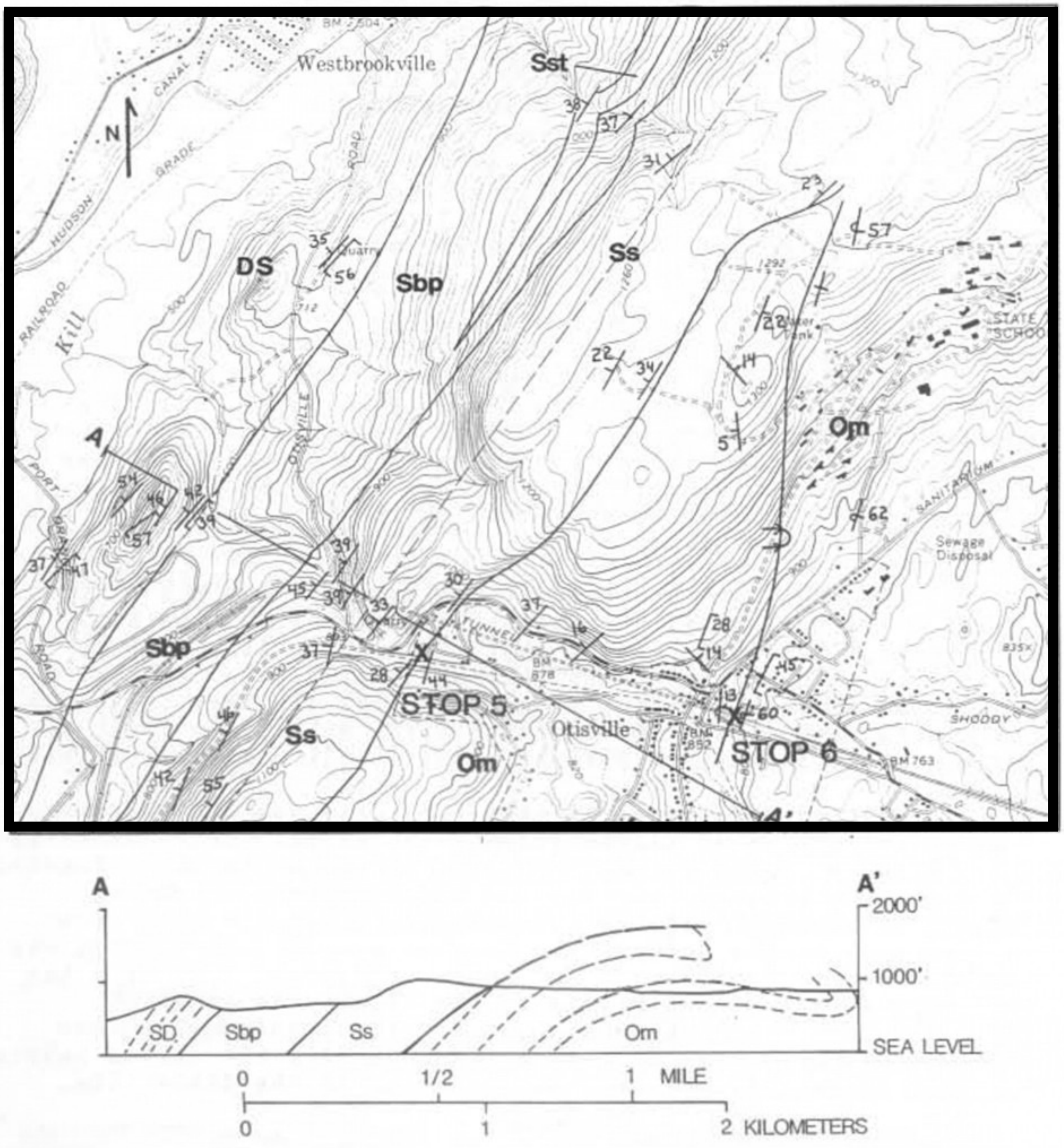

Figure 10. Preliminary geologic map and section of the Otisville, NY, area, showing the angular unconformity between the Shawangunk and Martinsburg Formations, the overturned syncline overlapped by the Taconic unconformity, and location of STOPS 5 and 6 (1987 NYSGA). Standard structure symbols used for bedding, cleavage, and axial tract of syncline. DS = Schoharie Formation through Bossardville Limestone; Sbp = Poxono Island Formation and Bloomsburg Red Beds; Ss = Shawangunk Formation; Om = Martinsburg Formation. Base from Otisville 7.5' topographic quadrangle, 1969. (Epstein and Lyttle, 1987, fig. 16)

Figure 11. East portal of the Erie Railroad tunnel at Otisville (GPS 41°28'25.1"N/74°32'13.2'W). Intensely deformed Martinsburg shale and sandstone are exposed on both side of the cut

over the tunnel. Elevation of both portals is ~780 ft (~240 m). Therefore, the eastern 0.75 mi (1.2 km) of the tunnel is only ~100 -120 ft (~30-35 m) beneath the floor of the gap, but under the spur ridge at the west end nearly 250 ft (76 m) of rock lies overhead. The tunnel is 5,314 ft (1,620 m) long (Wikipedia, 2011). At the west portal the 1908 grade curves to the south and joins the 1847 grade ~2 mi (~3 km) to the southwest.

Figure 10 is a preliminary geologic map of the Otisville area, showing the areal geology in the vicinity of the old Erie Railroad Tunnel and the structure at the portals (Epstein and Lyttle, 1987).

The Martinsburg Formation at the east portal of the Otisville Tunnel (Figure 11) is spectacular exposed in cuts on both sides of the tracks leading into the portal. While generally within the open-fold Taconic frontal zone in the Otisville area, the Martinsburg at the east portal is complicated by a faulted overturned fold (Figures 12 to 14). The formation here consists of shale and interbedded thin- to thick-bedded graywackes. Sole marks (grooves, flutes, and loads) are prominent on the undersurfaces of bedding in the overturned limb (Figures 15 to 18). Cleavage is well developed and is axial-planar to the fold. The axis of the fold trends about N10E and is overlapped by the Shawangunk ~1.3 mi (~2.1 km) to the north. Because the Shawangunk does not appear to be folded at the unconformity, the fold, faults, and cleavage in the Martinsburg here must be Taconic in age. Outside the area of this fold, cleavage is generally not developed (Epstein and Lyttle, 1987).

The west portal of the Otisville Tunnel is quite picturesque, being cut into the Shawangunk ridge beneath wooded cliffs high above and a deep roadcut on NY Route 61 just

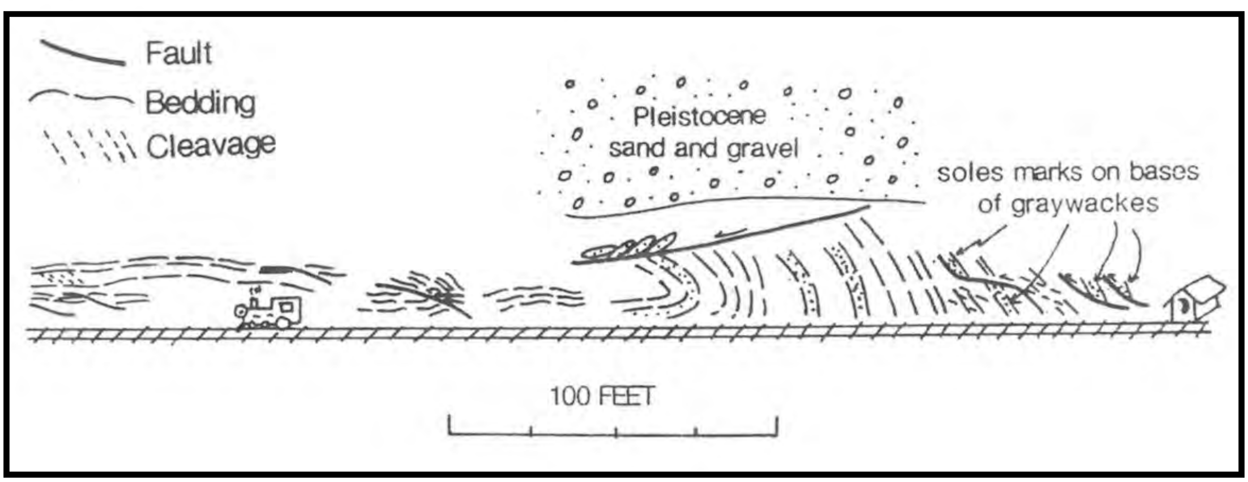

Figure 12. Field sketch (looking to north) of faulted overturned syncline along old Erie (now Norfolk Southern) Railroad at the east portal of Otisville Tunnel (Epstein and Lyttle, 1987, fig. 17).

Figure 13. West-verging, overturned (recumbent) syncline predominantly in sandstone on the north side of the cut at the east portal of the Otisville Tunnel.

Figure 14. The west-verging overturned syncline as expressed in shaly Martinsburg strata on the south side of the cut. Note the overlying west-dipping fault, movement on which is presumably to the west (see Figure 12).

Figure 15. Steeply dipping, overturned Martinsburg beds with sole markings on the north side of the tracks near the east end of the cut. Attitude of bedding is N25E/62 SE (overturned). Jeff Chiarenzelli for scale.

Figure 16. Groove and flute casts on the base of overturned Martinsburg sandstone bed at

Figure 17. Shallow load casts on the base of overturned Martinsburg sandstone bed at Figure 15 locality.

Figure 18. Large load casts on thick overturned Martinsburg sandstone bed at Figure 15 locality. Attitude of bedding is N25E/62 SE (overturned).

Figure 19. Cliffs of Shawangunk conglomerate at crest of Shawangunk Mountain high above the west portal of the Erie Railroad tunnel at Otisville and just north of the earlier rock cut through the wind gap.

Figure 20. West portal of Erie Railroad tunnel at Otisville (GPS 41°28'41.6"N/74°33'19.9"W). Note outcrop of Shawangunk conglomerate to left of portal. The inscription reads "19—Otisville—08," the year of completion of the tunnel.

overhead (Figure 19). Bedding in the Shawangunk quartzite on the north side of the tracks at the portal (Figure 20) strikes N47E and dips 45 NW. The tunnel is lined for ~100 ft directly in from the concrete portal (Figure 21), then unlined for ~750 ft (~230 m). The unsupported rock bore apparently extends to some point just west of the Shawangunk-Martinsburg contact, with concrete and corrugated metal lining from there through the Martinsburg to the east portal.

The Otisville Tunnel remained in the hands of the Erie Railroad until 1960, when the financially strapped Erie merged with the Delaware, Lackawanna and Western Railroad to form the Erie-Lackawanna Railroad. In 1976 The Erie Lackawanna became part of Conrail; and finally in 1999, on the split up of Conrail, the Norfolk Southern Railroad took over the line through the tunnel (Wikipedia, 2011, 2012a, 2012e). In 2003 the Metro-North Railroad began leasing this "Port Jervis Line" for commuter service from Suffern to Port Jervis, NY.

Figure 21. Unlined west part of the Erie Railroad tunnel at Otisville. The Shawangunk conglomerate here forms a ragged, but solid, arch that does not require lining.

NOTE WELL: Metro North runs frequent New York-bound and Port Jervis-bound trains every weekday, as well as on weekends. This is in addition to the Norfolk Southern freight trains operating on the tracks (Wikipedia, 2012d).

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Acknowledgments

Inners and O'Hara gratefully thank Dave Valentino and Jeff Chiarenzelli for granting us a few hours off from our rigorous responsibilities with the SUNY Oswego Delaware Valley Geologic Field Camp to examine the Martinsburg cuts at the east portal of the Otisville Tunnel in late May 2012.

Appendix—Structural Data from Cut in Martinsburg Formation at East Portal of Otisville Tunnel (measured by A. O'Hara)

Overturned Bedding

Strike	Dip	Dip direction
040	47	E
018	42	E
046	52	E
022	54	E
010	67	E
022	62	E
027	71	E
023	68	E

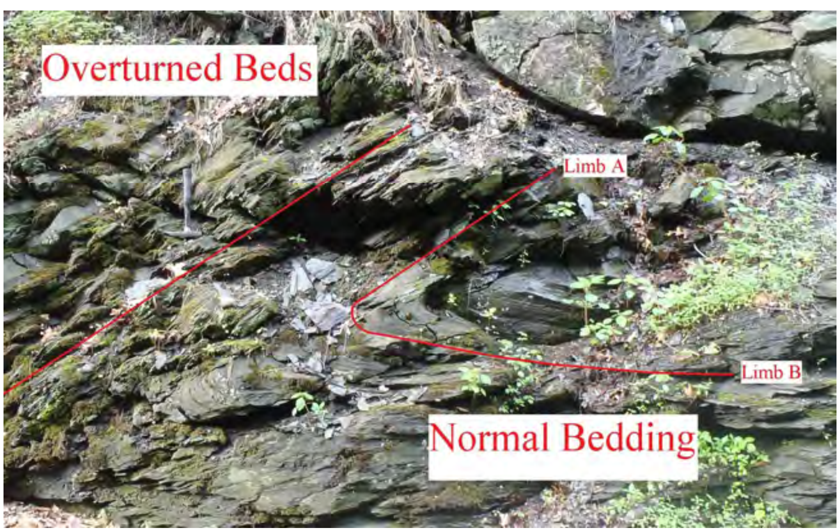
Fold

Limb A

Strike	Dip	Dip direction
036	36	E

Limb B

Strike	Dip	Dip direction
030	02	E



BLUE MOUNTAIN BOULDER COLLUVIUM

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Introduction

Blue Mountain is a unique feature in the geology of PA. The mountain forms a Ridge and Valley Province physiographic boundary that separates the Great Valley Section from the Appalachian Mountain and Blue Mountain Sections to its north. Blue Mountain (BKM), called Kittatinny Mountain east of Wind Gap, extends southwestward from the Delaware River to the Maryland state line and is broken only by several water gaps that cut through the mountain. This trip is concerned with the mountain only from the Schuylkill River northeastward and this discussion will focus mainly on the mountain a few miles southwest and northeast of the Lehigh River.

The crest of BKM is underlain by the Silurian age Shawangunk Formation, a unit of mixed rock types, but of particular importance are white sandstones and conglomerates that are the most erosion-resistant rocks in PA. Southwest of the Schuylkill River, and particularly in the Appalachian Mountain Section, these rocks are called the Tuscarora Formation (TF), or more commonly, the Tuscarora quartzite. Considerable information on the SF is presented in this guidebook and an excellent reference is Epstein and others (1974) and the several Epstein and Epstein references cited therein. A simplified geologic map of PA shows the Shawangunk and Tuscarora (S-T) as one continuous unit from the Delaware River to Maryland (Figure 1).

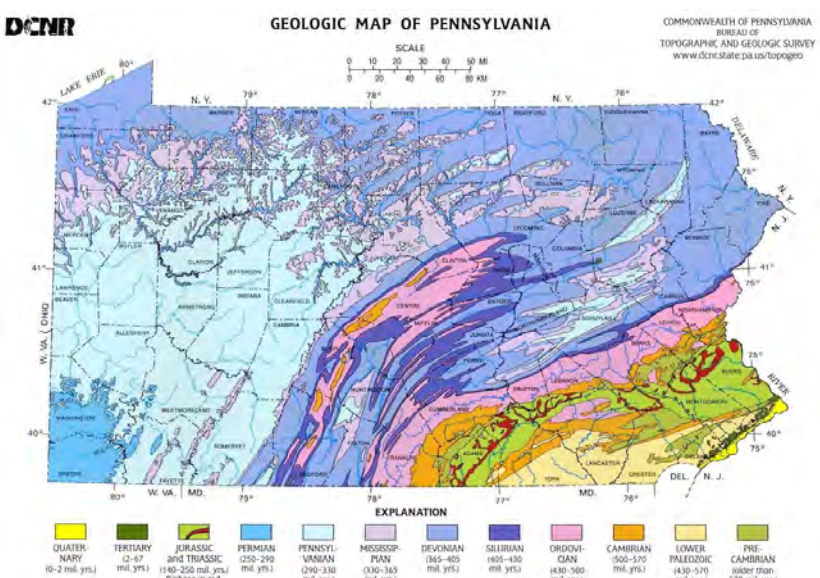

Figure 1. Simplified geologic map of PA. Copied from PA Geological Survey post card.

Sevon, W.D., 2012, Blue Mountain boulder colluvium, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 130-138.

East of the Susquehanna River, the S-T is exposed only on BKM because folding during the Alleghany orogeny put the S-T at considerable depth and it does not outcrop again in PA. In the Appalachian Mountain Section of the Ridge and Valley Province SW of the Susquehanna River, the TF forms the crests of several ridges as the result of anticlinal folding that brings the rock unit to the surface.

The Rocks

The rocks that comprise the south-facing slope of BKM above the unconformable contact with the underlying shales of the Penn Argyl Member of the Martinsburg Fm. (MF) are those of the Shawangunk Fm. (SF). The contact of the SF and MF is topographically defined by a marked change in slope from steep (SF) to gentle (MF) (Figure 2). The rocks of the SF comprise four members that are, in descending order, the Tammany, Lizard Creek, Minsi, and Weiders. All of the members contribute to the south-slope colluvial deposits, but the lower three members, and possibly mainly the lower two members, contribute the most. The Lizard Creek is the thickest member at 1,225 ft (373 m) at the Lehigh River water gap and frequently the crest and/or the upper south slopes of the mountain. The unit consists of shale, siltstone, sandstone, conglomerate, calcareous sandstone, and scattered red beds. Bedding thickness is variable and the various lithologies are intermixed.

Figure 2. Blue Mountain steeper BKM slope underlain by SF contrasting with gentler slopes developed on the MF, here defined by greener leafing-out trees.

The Minsi Member consists dominantly of white, planar-bedded to cross-bedded conglomeratic quartzite with variable bedding thickness. Some scattered, 0.5-3 in. (1.3-7.5 cm) thick, argillite beds occur. The unit is 225 ft (68.6 m) thick at Lehigh River water gap.

The Weider Mbr. consists dominantly of white, cross-bedded, planar-bedded, and massive conglomerate and quartzite. The conglomerates are up to 3 ft (0.9 m) thick. The quartzite occurs in 1-6 in. (2.5-15 cm) thick beds. About 3 percent of the member consists of thin argillite beds. The member is 220 ft (67.1 m) thick.

All of the members have abundant bedding-normal joint planes that were formed during the Alleghany orogeny. Figure 3 shows a large outcrop of SF on the east side of Lehigh River water gap. Both bedding and joint planes are apparent. The spacing of the planes is suggestive of the size of the boulders and blocks that result from breakdown of the outcrop and become the surface debris. The crest of the mountain and much of the outcrop is Lizard Creek Mbr. The lower right part of the outcrop has a prominent, laterally continuous, white, massive unit that is

Figure 3. SF sandstones and conglomerates on the east side of the Lehigh River water gap. BKM crest and much of slope is Lizard Creek Mbr. Lower white ledge is Minsi Mbr.

Figure 4. Remnants of SF outcrop at crest of BKM now broken into boulders and blocks.

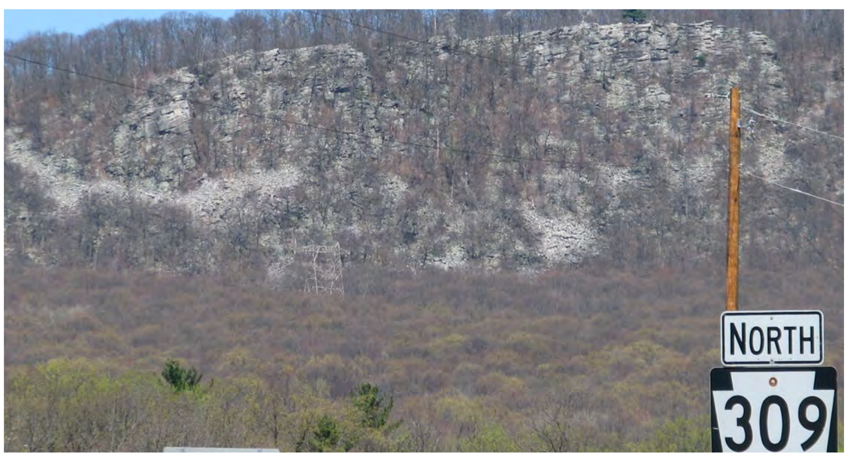

Figure 5. Large area of SF outcrop just east of where BKM is crossed by US Rte 309. Best exposed large area showing features. Note bedding and fracture planes and the large amount of BC on slope.

Figure 6. BC on the upper steeper slope of BKM where no GLs occur.

the Minsi Mbr. These rocks plus the underlying Weiders Mbr. are the source of much of the rock that covers the south slope of BKM.

Boulder Colluvium

The south-facing slopes of BKM are covered with abundant rock debris called boulder colluvium (BC). This BC results from the physical breakdown of rocks of the SF into boulders (rounded to subrounded, >10 in. (256 mm) diam.) and blocks (angular to subangular, > 10 in. (256 mm), and generally larger than nearby boulders) of various size controlled by the variable spacing of the bedding and joint planes. Figure 4 shows an area at the crest of BKM where an outcrop has been totally broken down at the surface by physical processes. Figure 5 shows a very large area of mixed outcrop and BC lower on the BKM slope. Much of the BKM slope is covered with too much vegetation for the large extent of BC to be readily apparent.

Physical breakdown of the SF was intense during the Pleistocene when periglacial activity dominated the non-glaciated areas of PA. Such activity was definitely associated with the pre-Illinoian, Illinoian, and Late Wisconsin glaciers that crossed NE PA and crossed BKM in different places NE of the Schuylkill River. The southwesternmost crossing was that of the pre-Illinoian glaciation that crossed BKM southwest of the Lehigh River and deposited glacial materials in the New Tripoli quadrangle (Braun, 1996). Subsequent BKM glacial crossings were farther NE. The closeness of glacial ice intensified the periglacial activity in nearby areas and resulted in the abundance of BC in areas close to but SW of the ice crossing, e.g., Figure 4. In areas where the ice crossed BKM, much material was removed by the ice and the resultant amount of BC is less.

On the steeper upper BKM slopes the BC forms a relatively uniform covering of boulders and blocks with essentially no surface form other than the irregularity caused by variations in size of the rocks (Figure 6). The lack of surface morphology results from downslope movement caused primarily by gravity. This all changes when the contact between the SF and MF is reached. The shallower slopes associated with the MF impeded gravity caused flow and periglacially aided down-slope flow took over. The periglacial flow involved water and ice as well as gravity.

This periglacial flow, working through freeze and thaw action, gradually moved debris down the lower gradient slopes. As it did so, it created a lower angle slope that varied from very low angle to nearly flat to slightly depressed in places. The debris accumulated as an elongated mound at the front of the moved mass. The frontal mass had either a relatively straight or lobate front. The front of the lobate mass was steeper than the surface in front or behind. The individual elongate mass is called a 'gelufluction lobe' (GL).

The fronts of the GLs are only a few feet high and can be missed as significant geologic features, particularly in wooded areas, by one who does not know their significance. Figure 7 is a LIDAR image of part of the New Tripoli quadrangle showing the fronts and backs of a number of GLs, both moderately to very lobate. The dark area at the north side of the GLs is the steep slope of BKM underlain by the SF. The GLs occur on the slopes of the MF.

Figure 8 shows the subtle front and top of a GL with almost no surface boulders. The angle of the frontal slope of the GL shows best near the left margin of the photo. A similar GL front is shown in Figure 9. The boulders and blocks in these GLs by the finer grained matrix material. In contrast, Figure 10 shows the front of a lobe composed entirely of boulders and

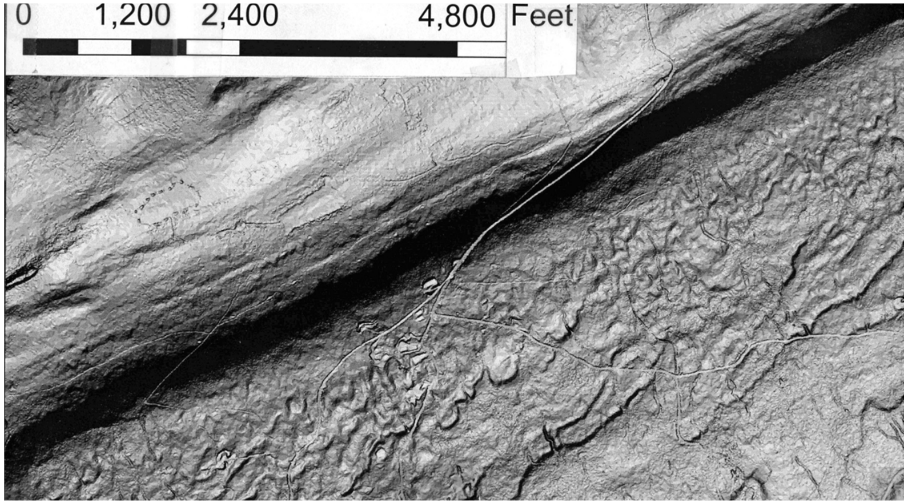

Figure 7. LIDAR image of part of the New Tripoli quad. Showing moderately to very curved GLs. Note the well defined GL fronts. Dark area north of GLs is the BKM south slope.

Figure 8. Front and top of a GL with almost no exposed boulders or blocks. Slope of GL front best shown on the left side of the photo.

Figure 9. Front end and laterally continuous top of a GL with no exposed boulders or blocks.

Figure 10. Front end and crest of a GL composed totally of boulders and blocks. Front slope both on the left and towards the viewer.

Figure 11. Typical surface of the north-sloping side of BKM showing minimal surface debris and no BC similar to that on the south-facing side of BKM.

blocks. Note the lack of boulders on the surface in front of the GL. That area presumably contributed its rocks to the next down-slope GL.

Hopefully this discussion, the LIDAR image, and the photographs will make these topographic forms more clear and recognizable when encountered in the field.

Mention should be made of the north-facing slopes of BKM. These slopes are dip slopes that parallel the slope of the SF as it disappears underground. The slopes are underlain in part by the uppermost part of the SF and for the most part by the overlying Bloomsburg Fm. (BF). Because of the dip slope and the general softness of the BF red sandstones, siltstones, and shales, the amount of debris on the north-facing slopes is minimal and the debris is much finer grained than that on the south-facing slopes of BKM and BC is absent. Figure 11 shows a fairly typical example of the nature of that surface.

Devil's Potato Patch

The flat area of the floor of Little Gap in BKM NE of the Lehigh River is a boulder field called the Devil's Potato Patch. Braun (1997) calls this an area of boulder colluvium, which it is in sense, but, because of the flatness and breadth of surface, I prefer to call it a boulder field. Figure 12 shows well the flatness, width, and partial length of the accumulation of boulders. Figure 13 shows the angular character of the blocks and boulders as well as the almost total lack of rounding of any of the rocks. The rocks are all derived from the slopes on the sides of the gap.

The boulders and blocks were physically broken away from the SF and moved into their present position periglacially. This had to have occurred after the gap itself was eroded and after the pre-Illinoian glaciation had occurred. The Illinoian and Late Wisconsin glaciations were sufficiently close to have had considerable periglacial affect on the gap area.

Noticeable in Figure 13 is the vertical to sub-vertical orientation of many of the elongate boulders/blocks. Why? I don't really know, but I suggest that when an elongate rock mass reached the bottom of the side slope through gravity processes, it became encased in ice and was moved onto the boulder field in that orientation. The same would hold true for rocks in

Figure 12. View of the Devil's Potato Patch showing flatness, length, and general nature of the boulders and blocks.

Figure 13. Close-up view of the boulders and blocks forming the Devil's Potato Patch. Note the vertical orientation and angularity of many of the blocks.

other orientations. Note in both Figures 12 and 13 the lack of rounding of any of the rocks and the lack of any visible matrix material. This angularity is a total contrast to the Hickory Run Boulder Field (Hickory Run State Park, Carbon Co., PA, about 18 mi. (29 km) north of Little Gap) where rounding of boulders is extensive (Sevon, 1990). Obviously, the two boulder fields have considerably different histories.

Origin of Blue Mountain Wind and Water Gaps in the Field Trip Area

In 1979 I pointed out the direct correlation between the position of water gaps in BKM and the former position of Paleozoic input centers through which came the sediment that filled the Appalachian basin (Sevon, 1979). Despite the temptation to suggest a simple drainage reversal following the Alleghany orogeny, the whole history of the Mesozoic discounts that idea. Instead, the best explanation is Mesozoic and post-Mesozoic headward erosion by streams that originated both on the north side of the Birdsboro basin (Faill, 2003) and on the new coastal margin of North America following its separation from Africa.

The first headward-eroding stream was the Triassic precursor to the Schuylkill River that eroded into the Anthracite basin that was once up to 5.9 mi. (9 km) higher as the result of Alleghany orogeny overthrusting (McLachlan, 1985; Faill, 1998). This headward erosion brought sand and gravel southward into the north part of the Birdsboro basin and formed the Hammer Creek Fm. (Glaesser, 1966; Faill, 2003). Although basically unstudied for correlation to this concept, late Triassic Birdsboro basin conglomerates marginal to both the Delaware and Susquehanna Rivers suggest that headward erosion by those streams may also have been underway at an early date.

Why have the headward-eroding streams chosen to cut through BKM at sites correlative with the former sediment-input centers where the hardest and coarsest rocks, sandstone and conglomerate occur? This is a particularly relevant question because most if not all, of these coarser, harder rocks, e.g., the Pottsville on the Schuylkill River, the (in descending order) Duncannon, Clarks Ferry, Berry Run, and Packerton Mbrs. of the Catskill Fm. on the Lehigh River, and the SF on the Delaware River, are central to these former input centers and change laterally through facies change to finer grained, more erodible rocks. To me, the only logical explanation is the erosion-susceptibility of the combination of bedding planes and orogenically produced fracture planes that have greatest concentration in the coarser and harder rocks that occur in the former sediment-input centers.

Wind gaps in the field trip area occur only at Little Gap and Wind Gap, the site from which the geologic term originated. There seems to be little doubt that streams once flowed through each of these gaps and that these streams were long ago beheaded by tributaries to larger streams on the north side: Aquashicola Creek, tributary to the Lehigh River in the case of Little Gap, and Cherry Creek, tributary to the Delaware River, in the case of Wind Gap.

Final Words

I hope that this commentary and particularly some of the illustrations will be an illumination of some of the geology that occurs not only here in this field trip area, but also in other parts of PA. The gelufluction lobes are widespread and particularly noticeable one you know what to look for. All the slopes of BKM and those of South Mountain in Franklin and New Cumberland Counties are very demonstrative of these features.

The trip does not stop at the Devil's Potato Patch, but it does pass through Little Gap and the boulder field can be seen in passing. There's a good parking area on the SW side if you want to come back.

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KARST SUBSIDENCE PROBLEMS ALONG THE BUSHKILL CREEK, NORTHAMPTON COUNTY, PENNSYLVANIA

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Introduction

During the period 1999 through 2006, both banks and the floodplain of the Bushkill Creek in the Stockertown area (Figure 1) were affected by an inordinate amount of sinkhole activity. Consequently, existing highways and nearby residential areas were impacted

The story is a complex weave of karst geology, surface mining, hydrogeology, stream geomorphology, engineering, land development and the weather. Combine all these variables with a handful of geo-technical firms, state and federal government agencies, municipal government officials, legislators, the general public and the story takes on serial proportions.

Although not a stop in this year's Field Conference, the events that took place during this time period provide an interesting aside into the geotechnical problems associated with karstic subsidence. It is suggested that what has been observed along the Bushkill can serve as a model case study for the geologist and engineer.

Figure 1. The Stockertown area; Nazareth 7.5-minute quadrangle. Coordinates for the former SR 2017 bridge at the Bushkill Creek, 40° 44'55.51", 75°15'44.43". BC = Bushkill Creek, BW = Brookwood Community, ST = Stockertown, LBC = Little Bushkill Creek. North is up.

Regional Karst Geology

The Bushkill Creek has its headwaters to the north at Blue Mountain, a ridge of siliciclastic Lower Silurian-aged bedrock. As it flows south of the ridge, the creek crosses Ordovician slates and shales of the Martinsburg Formation then across mixed shales, shaley limestone and limestone of the Jacksonburg Formation, and finally across interbedded

Kochanov, W.E., Karst subsidence problems along the Bushkill Creek, Northampton County, Pennsylvania, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 139-152.

limestone and dolostone of the Epler Formation (Aaron, 1971; Drake, draft; and Epstein, 1990). A generalized geologic map is shown in Figure 2.

Bedrock structure is complex. Bedding generally strikes in an east-west direction with variable dips that reflect low to high-angle folding. High-angle joints strike north south with variance of 20^{o} to 30^{o} towards the east. One feature worth noting is the "Stockertown Fault," or fault complex (Epstein, 1990). It is interpreted as one of a series of imbricate folded thrusts in this general area (Epstein, 1990). Based on the east-west strike orientation of faulting observed in the nearby Hercules limestone quarry, the general straight-lined nature of the Bushkill Creek, and borehole data (SAIC, 2002), it is suggested that the "Stockertown Fault" or fault complex may extend along the Bushkill Creek from the Hercules Quarry to the Little Bushkill.

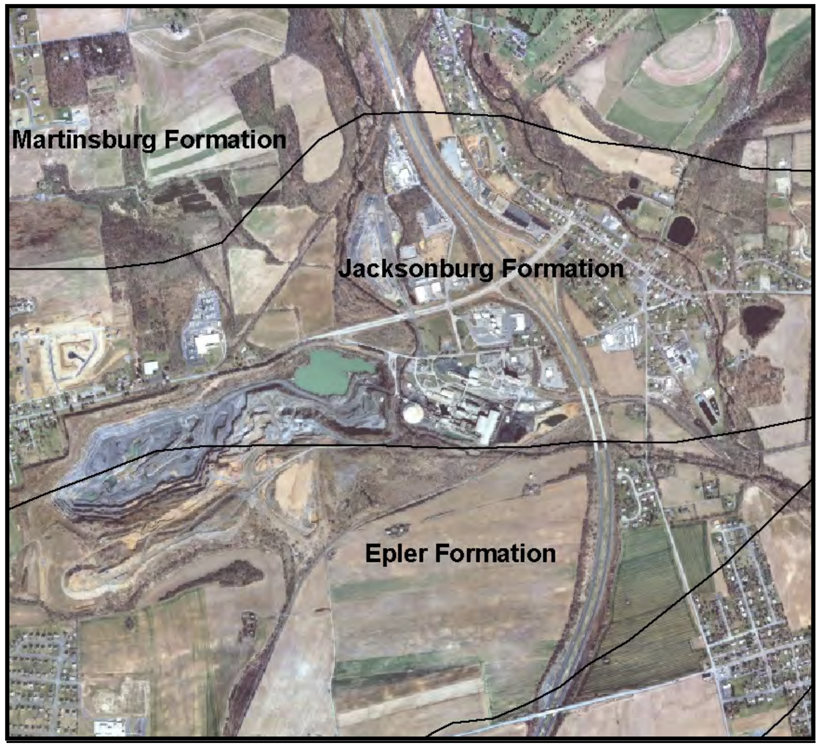

Figure 2. Generalized geologic map of the Stockertown area. Geology from Aaron (1971) and Drake (unpublished).

Surficial geologic mapping by Braun (1996) shows pre-Illinoian glacial till and lag deposits over much of Northampton County. The tills tend to lie in the often straight-lined drainages that are oriented with regional bedding and joint patterns (Kochanov, 2005). Till is commonly exposed within the channel and in sinkholes along the Bushkill Creek between the Hercules Quarry and the Little Bushkill Creek (Figure 3).

Surface Features

The topographic surface of the limestone belt in eastern Pennsylvania exhibits a relatively low-relief karstic surface. Surface depressions are the dominant features ranging upwards of 328 ft (100 m) in diameter and 3 to 10 ft (1 to 3 m) in depth. This low, undulating surface is the result of being covered by a variable thickness of alluvial, colluvial, and glacial sediments making it appear "flat" in many areas. Coupled with land disturbances from urban and rural activities, surficial evidence for karst subsidence features can be difficult to discern from the ground level.

Statistically the Ordovician Epler Formation, along with the Cambrian Allentown Formation, account for 85 percent of recorded sinkholes (Wilshusen and Kochanov, 1999) and have the highest number of karst surface features per unit area in the eastern third of the Great

Valley Section (Kochanov, 1993). The Ordovician Jacksonburg Formation, on the other hand, has the lowest rate per unit area.

Sinkholes on average can range in size from 3 to 22 ft (1 to 7 m) in diameter and the same in depth. Bedrock is rarely observed within sinkholes inferring that the sink structures are the result of voids propagating upward through the regolith.

Figure 3. Sinkhole along the north bank of the Bushkill Creek showing coarse glacial till (T) overlain by finer alluvial sediment (A)

Although the Epler Formation is highly karstic, the degree of karstification can vary regionally. Density patterns on recent mapping shows that the Epler can exhibit areas with a high density pattern yet have low-density areas or areas lacking observable karst surface features within the same outcrop belt (Kochanov and Reese, 2003; Reese and Kochanov, 2003). Even though lithologic composition is an important variable in karst development, structural deformation is the key component in determining the pathways for groundwater movement. It is the orientation of bedrock discontinuities that directs the groundwater and thus determines areas of preferential dissolution. Additionally, dolostones are more brittle than limestone and would be more fractured as a result of structural deformation, allowing more groundwater to move through those discontinuities. This can make some dolostones more karstic than limestones even though the limestone may be more soluble. The interbedded nature of the Epler and Allentown, where more soluble limestone is in contact with more fractured dolostone, would also help to direct and enhance the carbonate dissolution process. The geographic distribution of these limestone-dolostone sequences could account for the distribution pattern of sinkhole occurrences within the Epler and Allentown Formations throughout the Lehigh Valley.

In this section of the Lehigh Valley, the carbonate belt straddles the mapped glacial border. This raises the question as to whether identified surface depressions are karstic or glacial/periglacial features. Braun (1994, 1996) discusses surface depressions occurring within the slate and shale belt of Lehigh County both inside and outside the glacial limit inferring that they are periglacial features. Smaller-scale periglacial depressions have also been observed superimposed atop larger-scale karstic dissolution features in the carbonate belt of the Lehigh Valley (Braun, 1996). These periglacial depressions however, commonly contain wetlands and perennial ponds and differentiate them from the surrounding karst landscape.

Records based on marine oxygen isotopes and radiometric dating of terrestrial volcanic and glacial deposits suggest that as many as ten glaciations may have approached the late Wisconsinan terminus and four probably reached beyond that limit (Braun, 1999); the late Wisconsinan advance destroyed nearly all traces of previous advances right up to the late Wisconsinan terminal margin (Braun, 1999).

With this many periods of glacial advance and retreat the karst surfaces that had developed during warmer, interglacial periods probably had been eroded away. However, with the rise and fall of local and regional base levels associated with these glacial/interglacial periods, *subsurface* conduits were probably preserved and continually modified with regard to connectivity, extent, and the amount of sediment moving through the conduit network.

These conduits are static features as chemical and mechanical weathering and erosional processes continually deepen and widen these zones over long periods of time. Water pathways through the regolith may be initially small, filling the interstitial spaces between soil granules and within macropores. Over the course of time they continually change course and evolve into more complex water insurgence systems linking with other pathways that eventually lead to the conduits developed in the soluble bedrock. These insurgences mimic the function of fractures and bedding partings in the carbonate bedrock in that they provide a means for water to flow (Kochanov, 2005).

Ponors, or swallow holes, can be considered the end result of this linkage process between surface drainage and the water table. Ponors are dynamic fixtures in the karst plumbing system and vary in size, shape and location within the regolith, changing with the seasonal fluctuations of groundwater and insurgent water. One of the most common forms is the alluvial ponor as discussed by Milanovic (1981) where water is gradually lost along the length of a surface stream through the unconsolidated sediment lining the streambed.

The Bushkill Creek is an alluvial ponor. A significant percentage of the stream waters is lost through alluvial and glacial sediments that cover a well-developed karstic bedrock surface (Kochanov, 2005). Monitoring data concluded that the Bushkill Creek was losing approximately 70 percent of its water between the SR 33 bridges and the SR 2017 bridge (D. Zeveney, U.S. Army Corps of Engineers, pers. comm.).

Sinkhole History

Historical evidence for sinkholes along the Bushkill Creek can be traced from aerial photographs as far back as the late 1930s. Perlow (1985) documented sinkholes along the SR 33 corridor during its construction, as well as surveys for the Lehigh Valley by Kochanov (1987).

Sinkholes began to be reported by local residents along the stream and within the floodplain of the Bushkill Creek in 1999. The main areas of occurrence were at the SR 2017 bridge, within the channel of the Bushkill Creek and along the south bank of the creek adjacent to the Brookwood Community.

In the fall of 2000, one residential property within the Brookwood Community had a large sinkhole open in the backyard with over $20,000 in remediation costs. Continued subsidence eventually forced the residents to abandon the house. Over the course of the next few years, sinkholes continued to open within the Creek and along the banks of the stream eventually compromising the SR 2017 bridge and undermining the approach roadway to the bridge (Figure 4). Individual sinkholes around the 2017 bridge coalesced to make one large subsidence area (Figure 5). Sinkhole activity was not limited to the 2017 area but also upstream and in the medial area of SR 33. In April 2001, large sinkholes opened along the south bank of the Bushkill with a portion of the Norfolk Southern railroad bridge being damaged.

During January of 2004, sinkhole activity focused around the abutments of the northbound span of SR 33 bridge. It was during this time that the SR 33 bridge suffered serious structural

Figure 4. A—A large sinkhole on the south bank of the Bushkill Creek in 2001; B—The SR 2017 bridge in October 2002.

damage, was razed and a new span constructed. In a proactive measure, the southbound span was also replaced. Replacement cost exceeded $30 million.

Figure 5. Coalescing sinkholes along the SR 2017 bridge in 2002. Photo courtesy of the Brookwood Group.

Concurrent with the damages to the bridges, a small drama was unfolding along the north bank of the Bushkill Creek near SR 33. Two sinkholes opened, coalesced, and created a nick point along the bank allowing the Creek to develop a small meander bend (Figure 6).

The formation of the small meander loop was intriguing and resulted in much speculation as to the end result of the stream breach. It was hypothesized that the stream might be attempting to revert back to a previous channel configuration. This goes back to the construction of SR 33 (late 1960s through early 1970s). At that time the channel of the Bushkill Creek was split around a mid-stream gravel bar (Figure 7A). As part of the construction activity, the southern channel of the Bushkill was filled and the entire stream flow was directed into the northern channel (Figure 7B). The hypothesis was that the stream would play out a game of connect-the-dots, where the the stream would link

Figure 6. View from the SR33 bridge in January, 2004 showing the location of sinkhole pools (P) in the bed of the Bushkill Creek. A sinkhole (SH) had opened along the bank and has started to divert water from the stream. This diversion helped to form a small meander loop. Photo courtesy of the Brookwood Group.

Figure 7. A—Aerial photograph from 1939 on the left shows the island bar on the Bushkill Creek with the channel split circumventing the island; B—The 1971 photo on the right shows the location of SR 33 and the modified channel of the Bushkill Creek. Flow of the Bushkill is from left to right, north is up.

to new and existing sinkholes, starting with those in the field and finally connecting to sinkholes at the SR 2017 bridge. By doing so it would create a new island and have the split channel reform. What really happened, however, was that the stream did connect up with sinkholes but the stream ended up going in a different direction (Kochanov, 2005). During the summer of 2004 the remnants of Hurricane Ivan dumped approximately 7 in (18 cm) of rain in the Northampton County area. The resulting flood wiped out the remaining meander “neck” creating a new recessed bank edge (Figure 8A). This recession brought the stream closer to an existing sinkhole that was located in the field (Kochanov, 2005).

In December of 2004, a breach developed connecting the sinkhole in the field and the Bushkill channel (Figure 8B). In the following week, additional sinkholes opened in advance of the prograding stream as the stream was pirated across the field. The stream had nearly advanced across the field and threatened to link up with one recently activated sinkhole near the SR 2017 highway and potentially affect the approach to the SR 2017 Bridge (Figure 8C).

A thin finger of land was serving as a partial barrier during these events (Figure 8D). It was plain that if this neck of land failed then the major flow of the Bushkill would have followed the route across the field. This prompted a rapid response to temporarily fill in the new stream segment (Figure 9) to prevent further deterioration of private property and potential damages to SR 2017 (Hill, 2005). This action by the State Department of Environmental Protection put an end to the progradation of the Bushkill Creek across the field at this time and put a stopper in this interesting case of sinkhole piracy.

Discussion

The overriding thought throughout this case history is what was the primary cause for the dramatic sinkhole occurrences to occur in this particular area? A number of variables come into play, each adding their own weight at any one time.

One simple version is that:

1. The area is underlain by carbonate bedrock. Sinkholes occur in areas underlain by carbonate

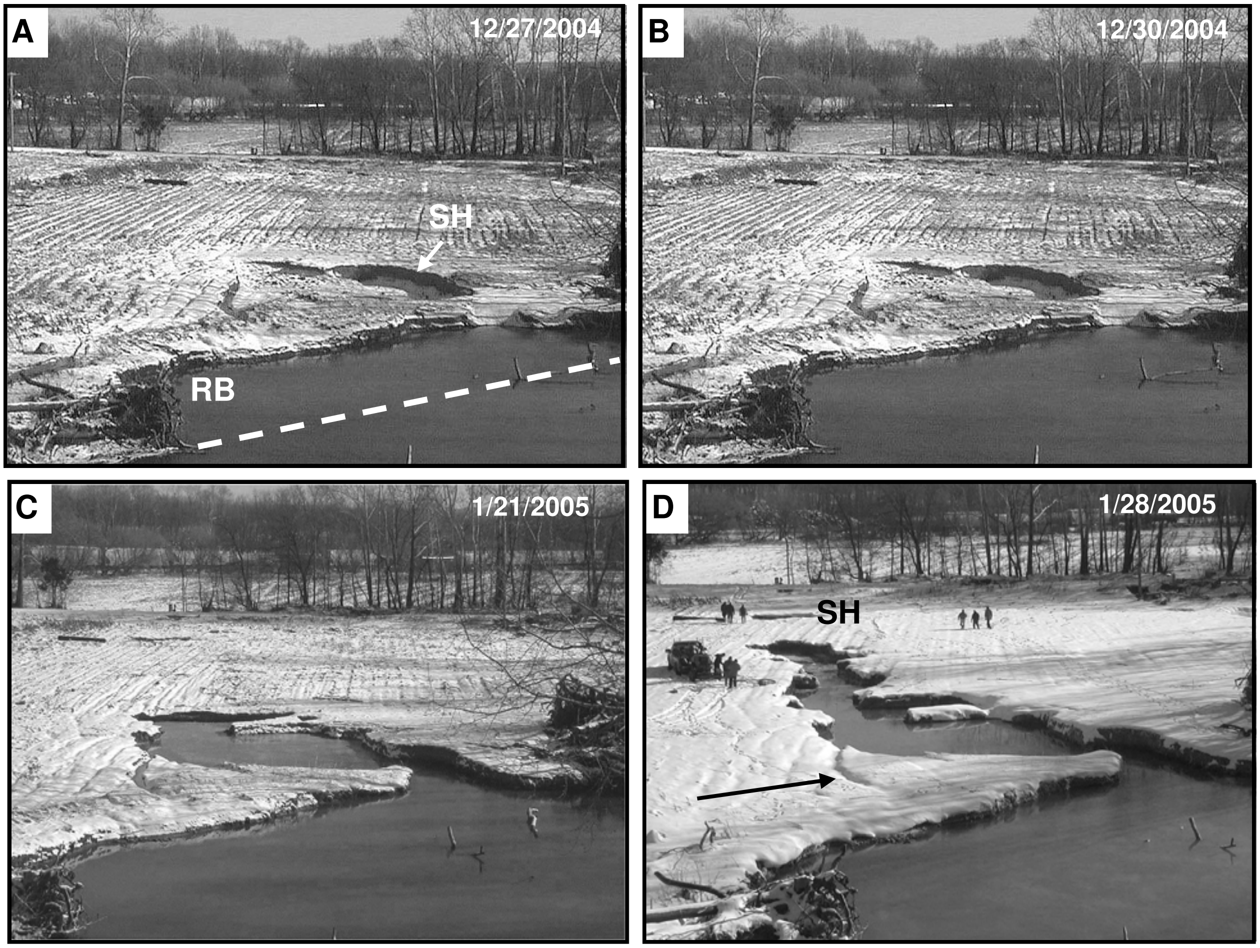

Figure 8. Photographic sequence showing the progradation of the Bushkill Creek across the field. A—Note the recessed bank (RB) and the nearby sinkhole (SH) in the field; B—Shows the breach in the bank (arrow) allowing stream water to enter the sinkhole; C—Shows the beginning of stream progradation across the field; D—Shows the stream almost to another sinkhole (SH) and SR 2017 (line at the top of SH letters). Also note in D the finger of stream bank (arrow) serving as a tenuous barrier. The outline of curving tension cracks can also be seen (arrow). All views are looking east.

Figure 9. Drained segment of the "new" Bushkill channel. It was subsequently filled with soil and shot rock as part of the rapid response plan.

bedrock. It was just this areas turn on the big wheel.

Evidence from borings and geophysical surveys indicate that a deeply weathered zone exists in the vicinity of the SR 33 bridge. In addition, borehole data along the Bushkill Creek show depths to bedrock ranging from 7 to 50 ft (2 to 15 m) indicating a pinnacled bedrock surface. For example, on the northbound segment of the SR 33 bridge, the depth to bedrock ranges from approximately 80 ft (25 m) on the south side of the creek to over 330 ft (100 m) on the north side of the creek (K. Petrasic, PDOT, pers. comm.). Sinkhole distribution is reflected in the regional attitude of bedding and joints.

Borehole camera views filmed by PA DEP showed a cave invertebrate in monitoring well 3 located on the south side of the Bushkill along SR 33 at depths of approximately 100 ft (30 m) below stream level. Core from a borehole on the north side of SR 33 depths of over 425 ft (130 m) showed rounded limestone gravels similar to pebbles observed in cave streams (pers. obs.).

Add in the fact that the Bushkill is a losing stream between SR 33 and SR 2017, it is quite apparent that the area is karstic. But what was the trigger?

Much of Pennsylvania was in the grips of drought during the period September 1995 through November 2002. The drought years were followed by a period of above-normal precipitation. What is noteworthy was that the drought lasted beyond the onset of sinkhole activity along the Bushkill. Additionally, high precipitation events due to Tropical Storms Dennis (2 in [5.5 cm]) and Floyd (6 in [16 cm]) in 1999 were also coincident with the onset of sinkhole activity.

Drought conditions can exacerbate sinkhole development. As the soil dries out, clays will shrink with desiccation and cracking of the soil would result in the development of more pathways for surficial water to enter the subsurface. In addition, drought conditions would have depressed the water table.

Depending on the cohesive properties of the regolith, infiltrating surface water can play an important role in promoting instability. Residual sediment typically has minimal interstitial cement holding the grains together and would have a low liquid limit. As water comes into contact with such loosely cemented material, cohesion is lost and liquefaction occurs. This is commonly observed during drilling where fluids are often lost at the soil-bedrock interface where the residual rind from the dissolution process is typically encountered. In another instance, the rise and fall of a potentiometric surface could also increase and decrease soil pore pressure and the effective stress between soil particles causing liquefaction and create sinkholes.

Over time, dewatering and the lowering of the potentiometric surface (i.e., during a drought) creates a temporary base level as equilibrium is reached. Sinkhole activity is generally low during equilibrium phases. Sudden changes in this equilibrium, such as what occurs during extreme swings in precipitation amounts (i.e., dry to wet times), can be the trigger that affects the hydraulic gradient to such a degree that sinkhole activity is high until the next plateau of equilibrium is reached.

2. Mining activity was concurrent with the onset of sinkhole activity. The affected area was compromised by the cone of depression developed by pumping during the mining process.

Sinkholes and limestone quarries are as acid mine drainage and coal mining; one often occurs with the other. Connections between quarry operations and sinkhole occurrence has been discussed in the literature (Foose 1953; Knight, 1970; Foose and Humphreville, 1979; Newton, 1987; Kochanov, 1999; Langer, 2001).

Dewatering, or the removal of water from sediment, commonly occurs during mining. The high pumping rates involved with removing groundwater from a mine can lower the water table and increase the zone of influence for miles (kilometers). Changes in the potentiometric surface from pumping can affect hydrostatic pressure and in turn cause gradual or sudden removal of support for the land surface. As this support is removed, the land surface sags, creating a depression and increasing potential for collapse.

A nearby quarry (approximately 6,070 ft [1850 m] west of southbound SR 33) has been in operation since 1919 primarily mining the Jacksonburg limestone and argillaceous limestone. On an average they pump 20-25 million gallons per day (mgd) (76-95 million liters per day [mld]) ranging to 32 mgd (121 mld) during periods of high precipitation out of the active pit and return it to the Bushkill Creek (DEP, 2000). Hydrographs of the pumpage rates during the period of increased sinkhole activity indicate that there has been a steady increase in pumping over time (DEP, 2000). However, one would think that sediment-laden water would be observed in the water being pumped out of the quarry with each sinkhole collapse. This was not always the case and leaves room to speculate that sediment could be traveling in some other direction or simply that the finer sediment never made its way back to the quarry. The observation of cave invertebrates, cave-type of sediment observed in deep core returns, and significant water loss between SR 33/SR 2017, leaves the investigator with some degree of certainty that a subsurface conduit system of some undefined extent is present in the Bushkill-Brookwood-SR 33 area.

Two other quarries (~1.7 mi [~2.7 km] SW of the SR 33 bridge) were also operating during the same time period. It was felt that the cone of depression for the three quarries overlapped at some point but the precise location of the combined cone of depression was indeterminate (S. Hill, DEP, pers. comm.).

Although no direct hydrologic connection has been determined quantitatively by means of a dye trace, it is generally assumed that there has been significant impact from the nearby mining activity and that the water table has been lowered significantly in the vicinity of the SR 33/SR 2017 stretch.

Dewatering and the lowering of the potentiometric surface with an increase (or decrease) in pumping rates could create a temporary base level and somewhat artificial groundwater equilibrium. A disruption of this equilibrium by increased pumping could remove the hydraulic support of the land surface and cause sinkholes to occur.

3. The construction of SR 33 set the stage for sinkhole development at the SR 33 and SR 2017 bridges as well as the sinkholes between SR 33 and SR 2017.

It is interesting to note that the onset of sinkhole activity began in the area where the Bushkill channel had been modified through the construction of SR 33. It would appear that the change of the Bushkill channel had some influence on stream processes.

Straight banks are not the norm for any significant distance but can contain many of the channel features common to meandering streams (Ritter, 1986; Leopold and others, 1964).

Typically, straight reaches contain sediment that accumulate along alternating sides of the stream as alternate bars with the thalweg, or deepest part of the channel, migrating back and forth (Ritter, 1986). The sediment within the Bushkill channel is generally a poorly sorted mixture of sand and cobbles and the channel is reflective of the dynamics of the stream flow rate and volume. The base of the channel alternates with a series of shallow riffles and deep pools basically directing the flow of water from side to side of the channel as it flows towards the Delaware River (Kochanov, 2005).

Pools in the Bushkill appear to be directly linked to sinkhole development within the stream channel. Keller (1971) observed that that as discharge increases, the velocity in the pool approaches that of the riffle and from this he suggests that in bankfull discharge conditions, the velocity in the pool will exceed that of the riffle. During periods of high flow, the pools are scoured, with sediment being deposited on “high” reaches of the stream. These topographic highs correspond to places where bedrock is closer to the surface. As stream energy dissipates during the waning of a flood event, the coarser sediment falls out while the finer sediment is transported and deposited to the next pool.

Kochanov (2005) suggested that sinkhole pools in the Bushkill Creek have had a major impact on determining flow direction. Within the main streambed, a sinkhole can serve as a deflector, forcing the water laterally towards the banks as well as directing water downward within the sinkhole pool (Figure 10). Once deflected, erosional processes along the banks would be more focused. Changes in sediment pore sizes, such as what would be encountered with the poorly sorted glacial sediment, can result in more turbulent groundwater flow and enhance erosion. As the banks are undercut, connections are made to voids and other subsurface drainageways and promote new sinkhole development ahead of the prograding stream (Figure 11).

Within the channel pool, the downward deflection of water may increase the size of the sinkhole or shift its location through erosion. In the case of the Bushkill reach between SR 33 and SR 2017, sinkholes within the channel appeared to approach a certain maximum size, roughly 10 ft (3 m). Sinkholes within the stream channel went through cycles of opening and filling in with sediment over time (Kochanov, 2005).

Summary

Since the fall of 1999, increased sinkhole activity south of the Borough of Stockertown had resulted in significant damage to public and private property. To date, three houses, three highway bridges, one railroad support, hundreds of work hours plus materials for other related sinkhole remediation, stream monitoring, borehole drilling, geophysical testing, along with dozens of cookies courtesy of Aunt Sylvia, contributed to making this case truly unique.

In karst areas, it may take a relatively long period of time to set the stage for a subsidence event, and when it does occur, it can proceed rapidly. Aerial photography established that the catastrophic sinkhole that occurred in the Borough of Macungie in 1987 (Dougherty and Perlow, 1987; Kochanov, 1987b) was related to a large sinkhole that had been filled some 30 years earlier. Sinkholes can recur in the same location due to variables such as precipitation, water table fluctuations, surface drainage alteration, and repair methods (if applicable). In that instance, the large sinkhole noted in the 1964 aerial photographs was filled and subsequently farmed. It is unknown whether the historical account for this occurrence was recorded. It took

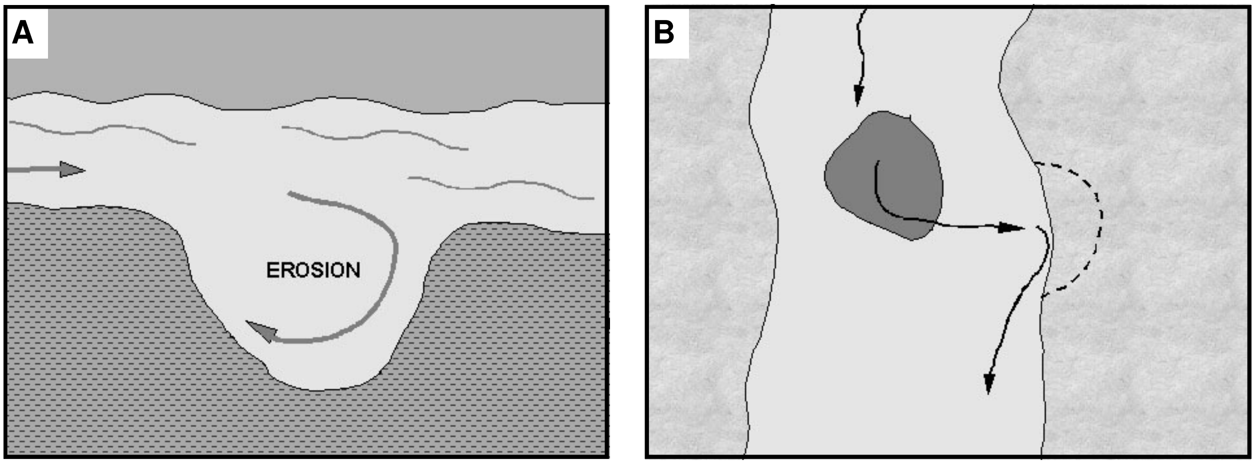

Figure 10. A sinkhole can serve as a deflector, forcing the water downward within the sinkhole pool (A) or directing water laterally towards the banks (B).

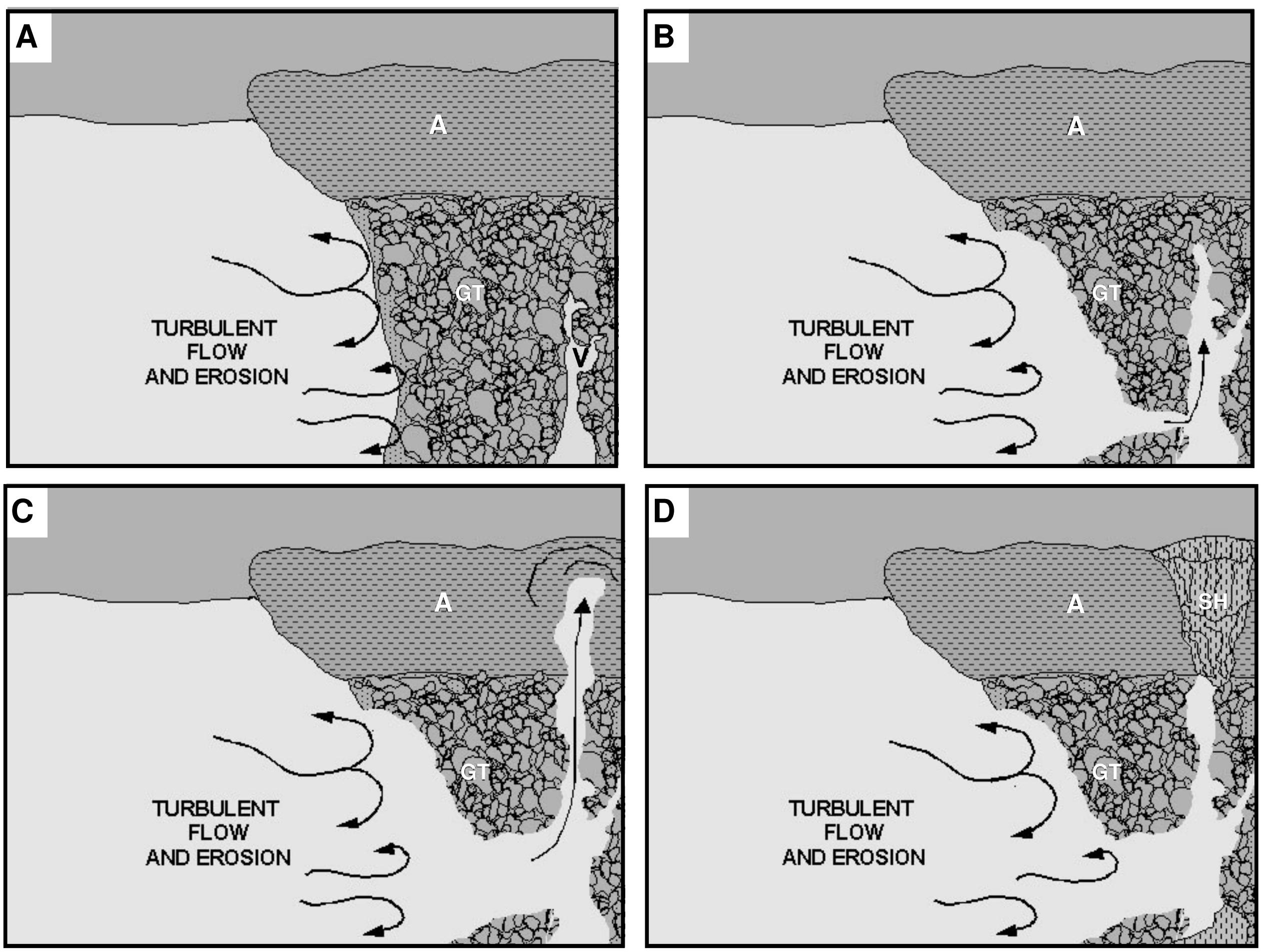

Figure 11. As stream water is deflected towards the bank and encounters poorly sorted glacial sediment (GT), more turbulent groundwater flow and enhanced erosion can occur. As the banks are undercut, connections are made to voids (V). Increased hydrostatic pressures (arrows) within pore spaces and within the voids can flush out sediment more easily and promote new sinkhole (SH) development.

decades for the sinkhole to be rejuvenated but when it did, it opened in a residential area and not rural farmland. Parallels could be made with the sinkhole activity at Stockertown. The construction of SR 33 and the modifications to the Bushkill channel occurred 25 years before the onset of the more recent activity. Mining surely had its influence as well as the impact of swings in precipitation amounts. It was probably a combination of factors that contributed to the Stockertown events, each adding their own weight at any one given time.

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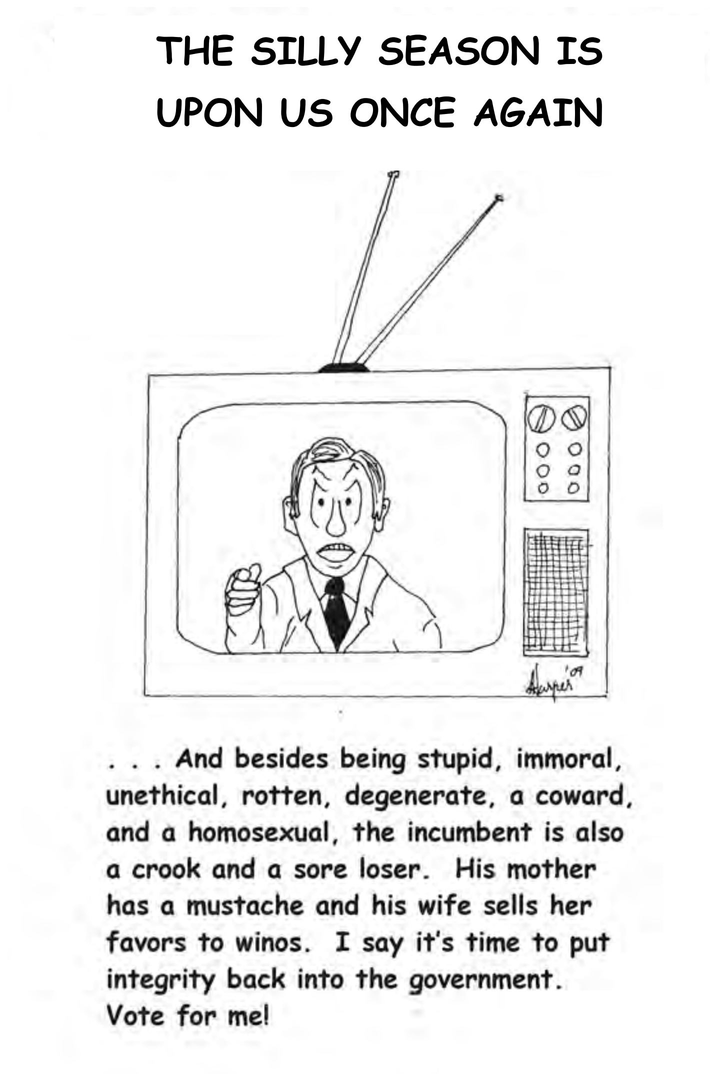

A HISTORY OF DELAWARE WATER GAP AND ITS RESORTS*

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** Published online by the Antoine Dutot Museum and Gallery, Main Street, Delaware Water Gap, Martin Wilson, East Stroudsburg University Pa. (http://www.dutotmuseum.com/history.htm), taken from his 1984 East Stroudsburg University Master's thesis.*

As a sunflower seed needs fertile earth, an adequate supply of water, mild temperatures, and plenty of sunlight to grow, so too, a resort community, in order to flower, requires specific conditions. For Delaware Water Gap, those conditions existed during the last half of the nineteenth, and the first third of the twentieth centuries. During that period, America's vacation habits and the limitations of transportation, coupled with the scenic beauty of the area and the entrepreneurial spirit of some local residents, conspired to transform the tiny borough into the heart of one of the most popular inland resort areas in the eastern United States. Each summer during that period its year-round population of about 400 was augmented by approximately 2500 visitors, many of whom stayed the entire season.

Life before indoor plumbing, super highways, and air-conditioning is hard to imagine for those of us who did not experience it. Summer, for city dwellers especially, must have been unpleasant and even unhealthful. Depending on individual economic circumstances, urbanites responded to unbearable summer heat in a number of ways. The wealthiest escaped for the entire season to Bar Harbor, Newport, or to other playgrounds of the rich. Working class and lower middle class Philadelphians traveled to Atlantic City to enjoy the cool sea breezes and the ever-present holiday atmosphere. For New Yorkers, Coney Island served as the destination of choice every year. Not everyone,

Wilson, M.W., 2012, A history of Delaware Water Gap and its resorts, *in* Harper, J. A., ed., Journey along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and southeastern New York: Guidebook, 77th Annual Field Conference of Pennsylvania Geologists, Shawnee on Delaware, PA, p. 153-165.

though, preferred the excitement and noise of these two seaside playgrounds. Many more prosperous middle-class city dwellers opted for the refreshing mountain air and the scenic beauty of America's inland resort areas, one of which was Delaware Water Gap, Pennsylvania.

The Settlement of Delaware Water Gap

In 1793, when Antoine Dutot arrived in the area with the intention of founding a city, the vicinity just north of the geological formation known as the Delaware Water Gap had been the site of human habitation for thousands of years. Known as the Minisink by the Lenni-Lenapes, it is estimated that the area was first inhabited by the Paleo-Indians as early as 10,000 to 12,000 B.C. When the first white men reached the region in 1614, they encountered the Minsi tribe of the Wolf Clan of the Lenni-Lenape Nation (the Lenni-Lenape were commonly referred to as the Delaware Indians because they ranged from the headwaters of the Delaware River to the shores of the Delaware Bay).

The Minisink was first explored by Europeans in the early seventeenth century by three travelers from New Amsterdam who entered the area from the Hudson River. One theory, recently contested by historians, is that the Dutch then mined copper from the mines on the eastern side of the river and built a road connecting the mines to Esopus (today Kingston, New York) on the Hudson. The first settler on the west bank of the Delaware River in the Minisink was Nicholas Depui who, in 1727, moved his family from the Hudson Valley to present day Shawnee.

Due to the difficulty of travel through the Gap (the mountains reached right down to the river leaving no room for a road or path), settlers in the Minisink knew little or nothing of settlements to the south. In 1730, Thomas Penn, son of William, sent Nicholas Scull on an expedition from Philadelphia to the Minisink to investigate rumors of settlements there. As a result of Scull's visit, Depui was required to repurchase land from William Allen (who had obtained it from Penn) that he had previously bought from the Indians. After Scull's sojourn, settlers from south of the mountains began to travel into the area. (Northern-bound settlers reached the area via Wind Gap.) It was not until the end of the eighteenth century, however, that the flow from the south eclipsed that of the north.

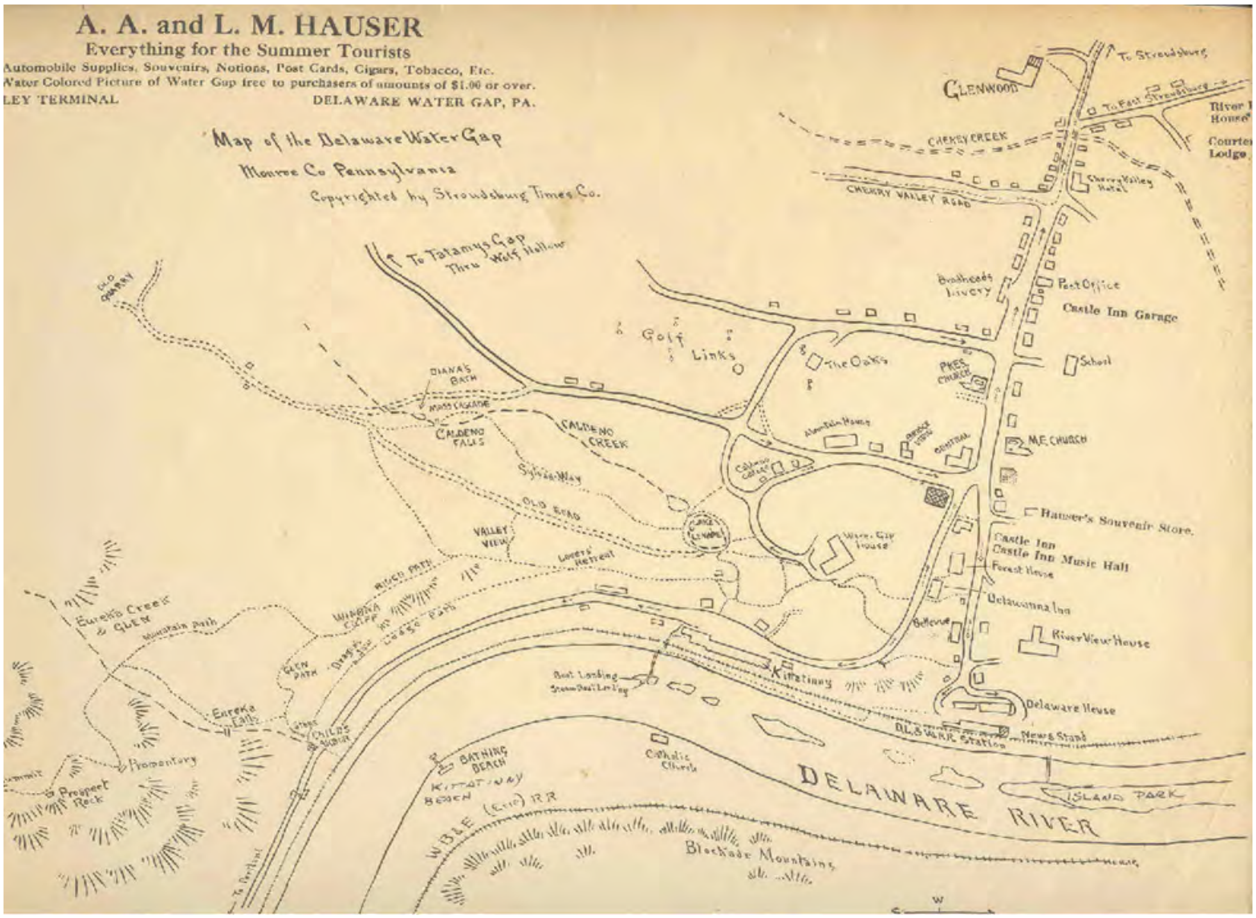
Dutotsburg

A settler from present-day Albany, Daniel Brodhead, moved his family to the area in 1737. Settling in present-day East Stroudsburg, Brodhead lent his name to the new town of Dansbury. The Indian wars of mid-eighteenth century led to a thinning of settlers as many moved away to avoid hostilities. By the time another settler, Jacob Stroud, returned to the area after the Revolutionary War, the Indian threat had been eliminated. Stroud was able to acquired several abandoned farms at very little cost. By 1806, he owned so much land that the area in which he lived began to be called Stroudsburg.

Delaware Water Gap remained unsettled long after settlements nearby had grown. In 1793, Antoine Dutot, a French plantation owner in Santa Domingo, fled the slave uprising there and headed toward Philadelphia. Upon arriving in the Quaker city, Dutot was advised to travel up the Delaware River to the Gap, where he purchased a large tract of land and began to lay out an inland city. He erected a dozen or more wooden buildings, designated a triangular piece of ground for a market, and named the new town after himself. Dutotsburg never became the bustling city its founder had envisioned, however. People moving into the tiny borough built their own houses and Dutot's structures fell into disrepair. Eventually Dutotsburg became known as the borough of Delaware Water Gap, probably in order to benefit from the inherent advertising benefits associated with the well-known geological formation.

Early Growth of the Resorts

This map was given (or possibly sold) at Hauser’s Trolley Terminal & Souvenir Store. It



shows where all the hotels and guest houses were located.

The natural beauty of the Delaware Water Gap proved to be an attraction to people traveling through the area. As early as 1820, visitors began staying in the small town where they roomed with local families in order to enjoy the scenery. Conscious of the possibilities, Dutot began constructing a small hotel overlooking the Delaware River in 1829. By 1832, however, he had run out of money and sold the incomplete building to Samuel Snyder. Snyder enlarged and completed the hotel which he named the Kittatinny. The new structure could accommodate twenty-five people and was filled the first season it opened. William A. Brodhead rented the Kittatinny from 1841 to 1851, when he bought it and increased its capacity to sixty. Over the next fifteen years the Kittatinny's size was increased on four separate occasions, first under William Brodhead, and, after 1857, under its new manager, Luke W. Brodhead. By 1860, the hotel could accommodate two hundred and fifty guests.

The success of the Kittatinny led to the establishment of other hotels. In addition, families opened their homes to visitors as a means of augmenting their income. At least one private home gradually grew into a full-fledged resort (the River Farm). By the Civil War, Delaware Water Gap's popularity as a resort area was becoming well-known throughout the northeastern United States. The strained economy of the war years led to a decline in the budding

Kittatinny House

Water Gap House

resort industry, but the reconstruction period found city dwellers once again traveling to the Gap. By 1867, the Brainerd, the Lenape, the Glenwood, the River Farm, and the Arlington, had joined the Kittatinny in offering accommodations to visitors. On June 20th, 1872, a new hotel that rivaled the Kittatinny in size and splendor, the Water Gap House, opened its doors.

Water Gap's Popularity

"Delaware Water Gap was the second largest inland resort town in the United States after the Civil War (ranking behind Saratoga Springs, N.Y.), and its clientele were the upper classes of Philadelphia and New York." So says one writer about the area. Although such rankings are hard to quantify, it is clear that the Gap enjoyed a national reputation for its resorts and drew prominent financiers, politicians, and society people from the time of the Civil War until World War I. Even a United States President visited the town (Theodore Roosevelt visited the Water Gap House on August 2, 1910). A publisher of world famous guide books in the nineteenth century included Delaware

Teddy Roosevelt

Water Gap among the fifteen scenic marvels of the United States. In 1906, an advertising pamphlet estimated that over one-half million people visited the Gap annually.

Unlike today's vacationer who may stay at a hotel for only one night or perhaps a week, Victorian Americans would often spend an entire season at their favorite resort—no doubt as a means of escaping the insufferable summer heat in the city. It was the custom among those families who could afford it to pack mom and the kids off to a hotel in the country for the entire summer where the father would join them on weekends. Summer visitors returned to the same resort year after year, calling it their second home.

What did the Gap have that attracted city visitors? According to Luke W. Brodhead, one of the managers of the Kittatinny and author of a book about the history and legends of the Gap:

> *The principal sources of amusement and recreation are the rambles over miles of mountain paths with vistas of great beauty opening at frequent intervals; carriage drives in many directions over a picturesque and interesting country; steamboat and rowboat service, and good bass fishing on the river in season and trout fishing in the adjacent streams."*

"Perhaps the featuring asset of the Gap, aside from its beautiful gorge, through which flows the placid Delaware, is its health-giving atmosphere, which permeates everywhere and which in itself has given the region much of its charm and popularity." This claim was made by an author extolling the beauty of the area in a book published in 1897. Whether the "atmosphere" in the region is any more healthful than anywhere else is, of course, open to debate. Nevertheless, that theme was played repeatedly in advertisements of the late nineteenth and early twentieth centuries. "The atmosphere is pure and dry, always cool evenings, and even at mid-day seldom so warm as to be uncomfortable. The whole region is free from mosquitoes or malaria." (This from an 1895 book.) As early as 1866, the local newspaper, The Jeffersonian Republican, ran a story reporting that the hotels and boarding houses were full; thus city people were escaping the danger of cholera, it said. In 1873, Doctor F. Wilson Hurd decided that Monroe County would be an ideal spot for his Wesley Water Cure. The Water Cure of Experiment Mills (later the Water Gap Sanitarium) was located just off the current Marshall's Creek exit of Rt. 80, and was instrumental in increasing the influx of visitors to the area.

For the last quarter of the nineteenth century the Gap's popularity earned it repeated mention in The New York Times. During the summer season, four to five articles a month appeared in that paper written by a correspondent in the town.

In order for families to take advantage of Delaware Water Gap as a vacation spot, good transportation was needed to insure that the patriarch could travel back to the city for the week's labor. In the late nineteenth and early twentieth centuries, good transportation (inland) meant railroads.

Transportation to the Gap

Roads

As we have seen, the natural barrier of the Blue Mountains led to early settlement of the area by people moving south from the Hudson River valley instead of north from Philadelphia. Prior to 1800, when Abram B. Giles constructed a wagon road through it, the Delaware Water Gap was not considered a practical passage north or south. Only rough Indian trails wound round the base of the mountains on both sides of the river. (A main Indian trail, upon which a road was later built by colonists, wound through what is now called the Wind Gap as it passed over the mountains.) Shortly after Giles completed his road, a visitor traveled the route and described it as a:

> *wagon road leading between the mountain's edge & the river & which all the labour of the inhabitants have been ineffectual to make more than about 8 feet wide or to clear from excessive roughness as it leads over one rough hillock to another the whole distance.*

Around 1799, in anticipation of the completion of the road, Benjamin Bonham constructed a small inn along it—the first in a town later to become famous for its hotels.

Antoine Dutot built a road in 1798 from his saw mill, below where the Kittatinny once stood, to the site of his planned city. A few years later he obtained a charter for a toll-road and extended his existing road to the River Farm where it connected with one running from Shawnee to Tatamy Gap. Although he set up a toll-gate along the way, he had trouble collecting tolls. In 1823, his road was superseded by one built by the state.

In order to meet the needs of the growing county, roads were widen and improved, and stagecoach lines began to operate. By 1846, a passenger and mail stagecoach stopped in Stroudsburg on the way to Milford from Easton three times a week. By that time, the road through the Gap was sufficiently improved to carry stagecoach travel.

In the early nineteenth century, Henry Drinker, owner of large tracts of land in northeastern Pennsylvania, dreamed of a rail line between the coal fields of Lackawanna County and the Delaware Water Gap. Drinker hoped to connect his line with one into New York, thus improving the marketability of the anthracite coal that had been discovered in the valley. It was not until March 11, 1853, however, that the Delaware, Lackawanna and Western Railroad was formed from the consolidation of two smaller lines. On January 21, 1856, the first train ran from Scranton to the Delaware River five miles below the Gap. It could go no further because the Warren Railroad in New Jersey was not yet open. By May 13 of that year, though, trains could travel from Great Bend (north of Scranton) to New York (actually the route terminated at Elizabethport, New Jersey, opposite the northwest tip of Staten Island). The Southern Division of the Delaware, Lackawanna and Western Railroad was officially opened on May 27, 1856. A train leaving New York at 7:30 in the morning arrived in Delaware Water Gap at 1:15 that afternoon, a trip of almost six hours.

With the intention of gaining access to a terminal closer to Manhattan, the D.L.& W. signed a lease with the Morris & Essex Railroad on December 10, 1868. The lease provided that the D.L.& W. would take over the Morris and Essex on December 31, 1868; thus Hoboken, right across the Hudson from New York City, became the D.L.& W.'s New York station. A ferry ran from the Hoboken terminal to the foot of Christopher Street, directly across the river in Manhattan, and to the foot of Barclay Street which is further downtown. The changes cut over an hour from the trip to the Water Gap.

Railroads

A common ingredient in the success of the towns of Delaware Water Gap, Atlantic City, and Coney Island as resorts was the existence of railroads. The introduction of rail service to these areas resulted in their increased popularity (in fact, Atlantic City did not exist until a rail line was built to the New Jersey shore).

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In 1900, William Truesdale, president of the D.L.& W., perceived that a new route was needed across New Jersey to forestall competitors from gaining the upper hand in passenger traffic. During 1906 and 1907, three studies were conducted to examine the feasibility of shortening the trip from New York to the Gap. It was decided to build a new route from Lake

Hopatcong to Slateford, Pennsylvania. The following account, published in a history of the D.L.& W., illustrates the enormity of the new line (commonly called the New Jersey Cut-Off):

> *The country to be crossed was anything but level. Valleys and roads ran north and south; the railroad ran east and west. There were to be no grade crossings. The new route would require 28.5 miles of new track, two large viaducts, and a fill three miles long and from 75 to 140 feet high. West of the Pequest fill, as it was named, were six miles of continuous cuts and fills. There were thirteen fills, most of which were about fifty feet high, and with fifteen cuts with the big Cut west of Johnsonburg being a maximum of one hundred feet deep and a mile long.*

Truesdale staked the future of his railroad on the success of the new line. Finished on December 24, 1911, at a cost of $11,065,511.43, the new route was a fast and smooth downhill run of twenty-eight miles. It cut eleven miles and twenty-seven minutes off the trip from New York.

In 1895, it cost $2.55 for a ticket from New York to the Gap. Ten years later, it cost twenty cents less. By 1933, the price was up to $2.82. With faster trains and more efficient scheduling, the time it took the train to reach Water Gap from Barclay Street gradually decreased. In 1959, it took just under three hours. Passenger service on the D.L.& W. ended on January 5, 1970.

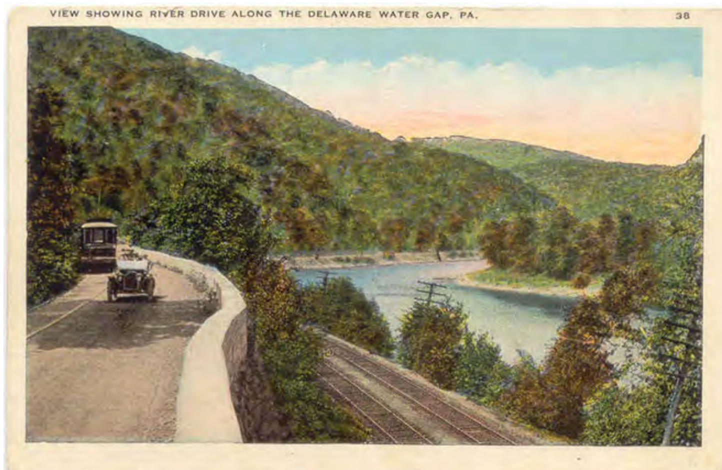
Another railroad company, the New York, Susquehanna & Western, provided passenger service to the area. Starting on October 24, 1882, the N.Y.,S.& W. ran from Weehawken, New Jersey and stopped in North Water Gap (Minisink Hills), and in Stroudsburg (near the present V.F.W.). The line crossed the Delaware just north of the Route 80 toll bridge (its stone supports can still be seen in the river). N.Y.,S.& W. service to the Poconos ended in 1940.

Passenger service from Philadelphia to the Gap was available on the Belvidere-Delaware Railroad (Trenton to Belvidere). Sometime around 1850, the Belvidere-Delaware extended its track to Manuka Chunk where it connected with the Warren Railroad. Passenger service was provided until October 4, 1947. (The line had earlier been absorbed by the Pennsylvania Railroad.)

Trolleys

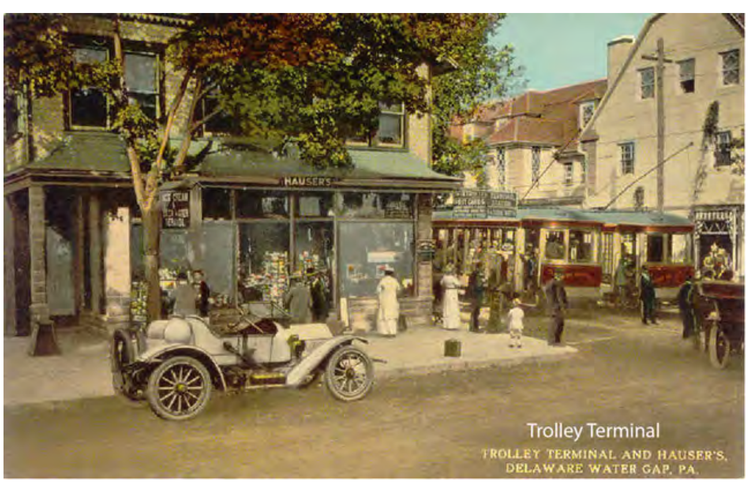
On July 10, 1907, The Mountain View Line, connecting Delaware Water Gap with existing trolley lines in Stroudsburg, began operations. During the school year, the trolley served as a school bus, charging students fifteen cents each way.

Meanwhile, trackage was being laid south of the Blue Mountain by the Lehigh Valley Traction Company that would eventually reach the Water Gap resorts. In connection with that company, on August 28, 1905, the Bangor and Portland Traction Company entered Portland from the west, having underpassed the Delaware, Lackawanna and Delaware tracks after a three year conflict. Railroad companies were reluctant to allow trolleys, their competitors, to cross rail lines. The plan was to continue the line into Stroudsburg, but the Lehigh and New England Railroad Company refused permission for trolley tracks to be laid across their rails, and the extension to the resorts was abandoned. Tourists from Philadelphia could travel north on the trolley to Nazareth where they had to change cars. From Nazareth they traveled on the Slate Belt Electric Railway Company's cars to Bangor where they switched cars again to those of the Bangor and Portland Traction Company. At Portland, passengers could ride a bus into Water Gap, or they could take the D.L.& W. The first "Delaware Water Gap Limited" left Chestnut Hill at 9:30 on the morning of July 17, 1908, and reached the Gap six hours and forty minutes later.



Wanting to gain access to the resorts at Water Gap for their "Liberty Bell" route, the Lehigh Valley Traction Company invested $50,000 in the Water Gap and Portland Street Railway Company. On February 21, 1911, portions of the mountain at the narrowest part of the Gap were dynamited to permit space for the tracks. By October, trolleys were running between Stroudsburg and Portland on the newly created Stroudsburg, Water Gap and Portland Railway Company. Open, screen-sided double truck cars painted lemon-yellow were in service in the summer and enclosed cars were used the rest of the year.

On April 1, 1910, the Lehigh Valley Traction Company announced an arrangement with the Philadelphia and Western Railway Company to use part of its line. The use of this track with its terminal at the 69th Street Station in Upper Darby was part of a larger upgrading of the entire rail system. By 1912, passengers could make the entire trip from Upper Darby to Portland without changing cars. Passengers dined during scheduled dinner stops at hotels in either Allentown, Rittersville, Bethlehem, or Nazareth. Alterations made to the cars on the Water Gap route for the comfort of passengers on the long ride included black leather seats with arm rests; baggage racks; carpeted floors; iced drinking water facilities; a



Trolley Terminal

uniformed "tour guide" who pointed out points of interest along the way; and a flashy, newly painted Liberty Bell Limited sign. At Portland, where the Lehigh and New England still refused a right-of-way to the trolley, passengers had to pick up their bags, get off one trolley and walk across the L.N.& E. tracks, and then board another trolley for the ride into Delaware Water Gap.

Direct service to Portland was short-lived. Before the 1913 vacation season opened, continuous service on the Water Gap route was canceled. Passengers had to change cars in Allentown.

In addition to the Liberty Bell Route, the Delaware Valley Route of the Philadelphia and Easton Transit company ran a trolley from Philadelphia to the Gap between 1908 to 1915. The journey took six hours and cost $2.40 round-trip. North of Easton the line was called the Blue Mountain Route and continued in service until November 25, 1926. From Bangor to Portland the route shared L.V.T. Company's tracks.

In 1917, the Stroudsburg, Water Gap and Portland Railway Company became the Stroudsburg Traction Company. The growing popularity of the automobile, however, rang the death-knell of the trolleys. On March 20, 1926, the Bangor-Portland was abandoned and the right-of-way was sold to Northhampton County for construction of a new highway between Portland and Mount Bethel. In November of the same year, the lease of the right-of-way between Portland and Water Gap, which was owned by the D.L.& W., was canceled thus ending service between the two towns. Stroudsburg Traction Company ceased operations in 1928 after trying unsuccessfully to compete with growing bus lines. The last trolley in Stroudsburg ran on September 8. In commemoration several hundred people turned out to witness the end of an era. A local band played "The Old Grey Mare Ain't What She Used To Be."

The Mountain Echo

For a time, beginning in 1879, Delaware Water Gap had its own newspaper. Called The Mountain Echo, the small, seasonal paper focused on activities at the hotels and on local places of interest. The editor was local photographer Jesse A. Graves. One of the services dutifully carried out by the periodical was the listing of all the guests staying at the various resorts.

The Hotels

A 1909 guide to summer resorts in the area had this to say about Delaware Water Gap:

> *Its quota of hotels is second to none in the Unites States. They compare favorably with those in any other section of the country in size and attractiveness and are comparable only to the very finest in the matter or cuisine.*

It is difficult to accurately determine how many hotels operated in the Gap. A search in surviving pamphlets and newspapers for advertisements reveal evidence of only the larger establishments. In addition, as some hotels changed owners, they also changed names, further clouding the issue. Nevertheless, it is estimated that the town of 400 permanent residents could accommodate over 2500 people. Long-time Water Gap resident Casey Drake remembers that, as a boy, the town was so crowded in the summer that it was often difficult to walk down the street.

The two largest and perhaps best known of the hotels were the Kittatinny and the Water Gap House (see photos above). The Kittatinny was located at the present site of the overlook along Rt. 611 just south of the borough. Part of its foundation still stands beneath the spot from which visitors look out at the Delaware River and the Rt. 80 bridge. The same view was enjoyed by guests of the Kittatinny as they stood on the hotel's large veranda. In 1874, the Brodhead brothers increased the hotel's capacity to 275. Then, in 1892, the building was razed to make room for a larger, more elegant New Kittatinny. Able to accommodate 500 guests, the hotel boasted, in addition to spectacular views and cool breezes, the following:

> *Electric lights, elevators, steam heat, running mountain spring water in rooms [and a mountain stream running under the kitchen—which can still be seen from the Rt. 80 bridge], private baths, etc. Noted for its cuisine and service, and the hotel's farm gives to the table products par Excellence."...Bell phone 92; telegraph office in hotel, orchestra, social diversions.*

A 1908 advertisement lists G. Frank Cope as proprietor. Similarly, one from 1917 lists John Purdy Cope as owner.

The Water Gap House was located above the Kittatinny on Sunset Hill (so named because when one stands facing east on the hill one can see the shadows on the mountain across the Delaware slowly rise as the sun sets in the West). Opened by Luke W. Brodhead on June 20, 1872, the Water Gap House had first and second story piazzas twelve to fifteen feet wide and 650 feet long looking out over one of the finest views in the area. In keeping with the mores of the times, Brodhead built the hotel with no bar.

In 1908, the Water Gap House was completely rebuilt at a cost of over $100,000. John Purdy Cope, its new owner, advertised its attractions in the June 14, 1908 edition of The New York Times :

> *Capacity, 300. A MOUNTAIN PARADISE; highest altitude, coolest location, always a breeze, no humidity . . . Commanding views for 30 miles in every direction of the grandest scenery east of the Rockies. Hotel is surrounded by its magnificent park of Old Shades, Rhododendron, Wild Flowers, Rare Plants, and Fine Lawns. . . . entertaining refined, high-class patronage. Running mountain spring water and stationery stands in all rooms. Fifty private tile bats, also public baths. . . . Telephones and telegraphs. Solariums and balconies on all floors. Steam heat, open log fireplaces. Electric lights. Hydraulic elevator. Most modern sanitary arrangements. . . . Hotel supplied from own greenhouse and farm with early vegetables and poultry. Milk from our own dairy of registered cows. Every outdoor sport and indoor amusement. Orchestra and frequent social functions. Private riding academy with high-class saddle horses and instructors; nine-hole golf links; garage and livery—all within the grounds. Coaches meet all trains.*

The Glenwood House opened its doors to summer visitors in 1862 after serving for a while as a boy's academy. In 1897, it was catering to 200 guests, was opened from May to November, and could boast private balconies on the second floor. A 1909 advertisement claimed a capacity of 400. The Glenwood also supplied its tables with fresh fruits and vegetables from its own farm. Of the old resort-hotels, the Glenwood is the only one still operating as a resort today. (The Central House, now the Deer Head Inn, still functions as a rooming house and its bar enjoys a reputation as something of a jazz mecca.)

The Castle Inn opened for business in 1909, and was the last of the great hotels built in the Gap. When it opened it had 112 guest rooms, a ball room, recreation rooms, its own power plant, and its own freezing plant.

The Bellevue was known by two other names over the years. First it was the Juniper Grove House, and later it was called the Arlington. As the Bellevue, it could sleep 150 guests and claimed to be the popular hotel for young people. A big selling point for this and some of the other hotels was their proximity to the train station.

The hotel located closest to the station was the Delaware House, which was situated just across the street. Open all year, the Delaware House could accommodate 50 people and offered, in addition to the normal activities such as fishing, boating, and bathing, also bowling, pool, and billiards.

The Riverview, also located near the station, had a capacity of 250. The Mountain House could hold eighty guests, and the Forest House could hold 100.

These are just some of the hotels located in the Gap. Many hotels, while not located in Delaware Water Gap, nevertheless maintained an address in town in hopes of benefiting from the Gap's popularity. The Karamac, for instance, was located across the river in New Jersey, and yet advertised its Delaware Water Gap address.

The End of an Era

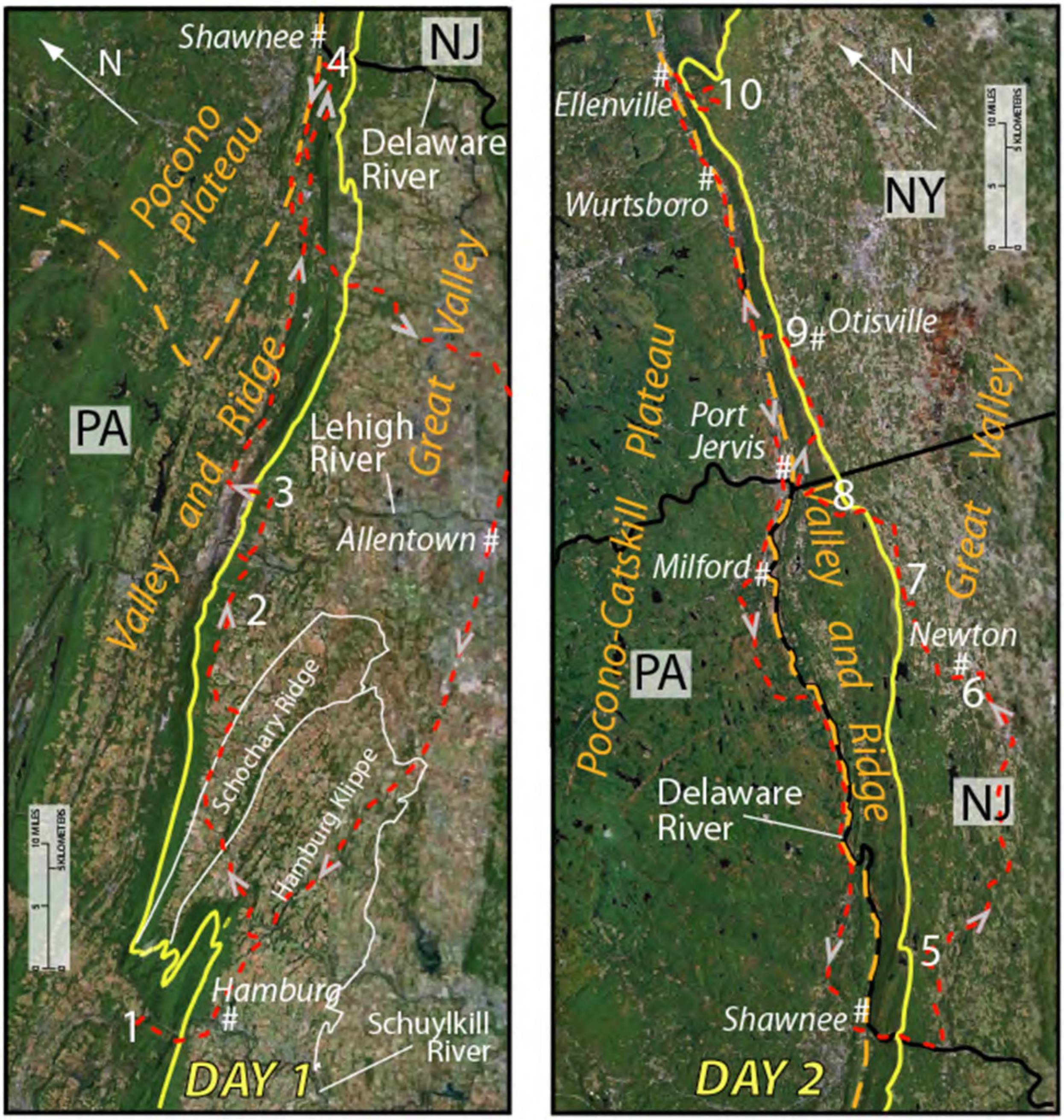
At five o'clock in the afternoon of Thursday, November 11, 1915, workmen, helping to close the Water Gap House for the winter, discovered a fire which had broken out in one of the guest rooms of the hotel. An alarm was sounded and several fire companies responded; but their efforts were in vain. Though a light rain was falling at the time, the entire structure was leveled in only a matter of hours. The loss was estimated at between $150,000 and $200,000. Four days after the fire, it was announced that a new hotel, as large as the Water Gap House, would be built on the same site. The planned hotel was to be fire-proof and, hopefully, would be open for some of the 1916 season. The hotel was never built.

Cope experienced another disaster in 1931, when the Kittatinny burned to the ground. He and his family were awakened at four o'clock on the morning of October 30, by a passing motorist who had seen flames coming from the Kittatinny. By six o'clock, the entire structure was engulfed—a loss of between $500,000 and $750,000.

Why was neither hotel rebuilt? Over the years the Poconos have continued to be a major resort region. Delaware Water Gap, however, has steadily declined as a resort community. Part of the answer for the Gap's decline as a resort lay with changing transportation trends; there was a clear symbiotic relationship between the resort and transportation industries in the town and surrounding area. The large hotels were in an ideal location to benefit from the easy access that the rail lines and trolleys provided. The hotels also furnished the varied transportation companies with a "draw" or need for transportation which the various companies were eager to fulfill. As the business of travel matured into the automobile oriented industry of today, however, the demand for the large hotels located on rail lines diminished. The popularity of the automobile after World War I, in part, changed the way people took vacations. No longer tied to the rail system for transportation, a whole new concept of vacationing developed. In 1909, a story in The *New York Times* anticipated this trend when it reported that a weekend outing with the entire family, stopping for a night's lodging at some comfortable but not too expensive hotel, was superseding the summer-long separation of the father from his family.

The automobile was only part of the answer though. Tough economic times of the 1930's erected a hurdle that, in combination with other factors mentioned, proved too high for Water Gap's resorts to overcome. When the resort industry began to expand after World War II, Delaware Water Gap seemed, for the most part, content to let the resurgence pass the town by. Many of the small boarding houses were converted into private residences. Most of the old hotels were either destroyed by fire, were closed, or continued to operate as best they could under changed conditions. Water Gap's heyday as a resort had come to an end.

Postcards courtesy Lucy Kosmerl.



Map of the 77th Annual Field Conference of Pennsylvania Geologists travel route (red dashed line) showing locations of the ten stops, physiographic provinces, rocks of Shochary Ridge and the Hamburg klippe, and the Taconic unconformity (yellow line) (constructed from Google Maps). Note the rapid decrease in width of the Valley and Ridge east of the Lehigh River.