Tectonics of the Susquehanna Piedmont in Lancaster, Dauphin, and York Counties, Pa.

Proceedings of a symposium associated with the 75th Field Conference of Pennsylvania Geologists

Lancaster, Pa.
September 23, 2010

Organized and hosted by the Pennsylvania Geological Survey
The problem: How can tectonic transport diverge in the Piedmont without excessive stretching in the foreland fold belt?

Great Valley structure east of Harrisburg with Dauphin Anticlinorium exposing Hamburg Sequence allochthons in its core, directly over the carbonates.
Tectonics of the Susquehanna Piedmont in Lancaster, Dauphin and York Counties, Pa.

Proceedings of a 2010 symposium held in conjunction with the 75th annual field trip:

Field Conference of Pennsylvania Geologists
Chair: George E. W. Love  Vice-Chair: Tom G. Whitfield  Secretary: Jamie Kostelnik (Secretary-Treasurer - Editor Emeritus: Gary M. Fleeger)

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Howell Bosbyshell  Robert C. Smith II
Carol de Wet  Scott Southworth
Rodger T. Faill  John F. Taylor
G. Robert Ganis  David W. Valentino
Alexander E. Gates  Donald U. Wise

September 23, 2010
Lancaster, Pa.

Organized and hosted by the Pennsylvania Geological Survey
SOME TECTONIC AND STRUCTURAL PROBLEMS OF THE APPALACHIAN PIEDMONT ALONG THE SUSQUEHANNA RIVER

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25th FIELD CONFERENCE OF PENNSYLVANIA GEOLOGISTS
October 22-23, 1960

Hosts: Department of Geology
Franklin and Marshall College
Lancaster, Pennsylvania

1960
This guidebook is available online at the FCOPG web-site
(As are other past field trip guidebooks)
Dedication of this 2010 Tectonics Symposium and Guidebook

Geologic mapping of this region by George Stose, Florence Bascom, Anna Jonas (Stose), and Elenora Bliss (Knopf) in the 1900s through the 1930s allowed them to develop tectonic syntheses and interpretations of the Susquehanna Piedmont far ahead of their times. Their tectonic syntheses and maps were radical. In an era when even the existence of great overthrusts was hotly debated, many geologists found such models beyond belief. The idea that the Martic Zone involved a great overthrust was bad enough; making the entire Reading Prong a completely allochthonous thrust sheet bordered on geologic insanity. Since then revolutions in geologic thinking and methods have elevated those early outrageous models to the status of (almost) proven facts.

Had Stose not been the map editor of the USGS, most of these tectonic syntheses probably would never have been published. He started as a field assistant in the last days of Powell’s tenure and was honored in the 1950s for 75 continuous years of service with the USGS, maturing his thinking during those years. Today we should recognize the pioneering efforts of those early field geologists. Their work, quadrangle by quadrangle, stratigraphic unit by unit, is still easily recognizable within today’s stratigraphic names and modern geologic maps. Their ideas have also withstood the test of time. Not only did they see and interpret far more geology of the overall region than their predecessors ever did nor their successors likely ever will; they also integrated those myriad details into imaginative tectonic sketches, texts and maps still well-worth detailed study. In putting together some of the data and ideas in this volume, I have again read much of their work and have been repeatedly surprised at the advanced level of much of their thinking. There is far more in those works than first meets the eye or enters the mind.

When this conference had its first of 75 trips in 1931 (four years were missed during WWII) such tectonic models really were near the lunatic fringe. When Ernst Cloos led a Field Conference trip in 1949 to the famous New Providence Railroad cut where Wissahickon Schist rests on Conestoga Limestone, even he was unwilling to apply either “thrust” or "unconformity" to that contact despite his superb mapping that showed five underlying stratigraphic repetitions that he was willing to call "thrusts." However by the 1960 field trip many of those previously "outrageous” tectonic ideas and models of the early workers were beginning to be more widely accepted. Now as this conference, a half century later, tries to make sense of a host of new field observations, radiometric and fossil dates, petrology, geochemical signatures, geophysics, structural fabrics and plate models, many of those early "outrageous" ideas appear prophetic. The problems now focus on how plate tectonic processes created the complex Susquehanna orogenic salient by jamming and overprinting a succession of mountain systems into a rifted corner of the North American craton.

At the diamond anniversary field trip of this conference and the golden anniversary of the “Tectonic Cross-section of the Piedmont” 1960 field trip, it seems appropriate to dedicate a tectonically-oriented guidebook to those pioneers who laid the map, stratigraphic, and tectonic foundations upon which we build our modern and slightly "outrageous" tectonic syntheses.

Don Wise, September, 2010
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TECTONICS IN THE PENNSYLVANIA SALIENT'S ELBOW: OVERVIEW

Donald U. Wise, Dept. of Geosciences, Univ. of Massachusetts at Amherst

Tectonic Jig-saw Puzzle

Tectonic overviews of the Susquehanna Piedmont through the 20th and 21st centuries have resembled table-top assembly of a giant jig-saw puzzle. In the first third of the 20th century, tectonic studies resembled looking at the box in which the puzzle arrived. Pioneers like George Stose and the women of Bryn Mawr, Florence Bascom, Eleanora Bliss (Knopf), and Anna Jonas (Stose), among many others developed the outlines of the surface geology of the region and its basic stratigraphy. This was done at a reasonably detailed but still reconnaissance level mostly for USGS, Pennsylvania Survey and Maryland Survey quadrangle maps. Refinements of these and works of many other geologists continue with a still expanding and diversifying new generation of geologists. The history of the Field Conference of Pennsylvania Geologists, as described in a following article by Potter and others, has been intertwined through the years with assembling this tectonic jig-saw puzzle. Several of its field trips have coincided with major turning points in the assembly process.

By the 1930's when trips by this Field Conference had just begun, enough pieces of the puzzle were recognized to begin the next tectonic steps of wild-eyed arm-waving, commonly by the original mappers. Many of these tectonic models were "outrageous" for their day but far ahead of their time: the contact of the schist belt with the limestones was proposed as a giant thrust extending at least from the Susquehanna to the Delaware Rivers (Knopf and Jonas, 1929) while the entire Reading Prong was proposed as the rootless remnant of a giant overthrust (Stose, 1935). Many of these ideas earned strong condemnation during those early years as well as many constraints through detailed studies, most important being detailed mapping and petrofabric of the Martic area by Ernst Cloos (1941). In 1949, when Cloos ran part of the 15th Anniversary Field trip to the Martic Area, many details had been sorted out even though the complex schists and higher grade rocks of the Appalachian core zone were still considered to be "Precambrian." Even though the term "tectonics" was rarely used at the mid-point of the 20th century, the original box-cover picture of the jig-saw puzzle had developed into a real table-top tectonic puzzle with arguments of how certain pieces should be colored, orange, green, or gray, and how these various pieces might fit together. The next ten years saw remarkable advances.

Tectonic environment of 1960

By the time of the 25th Anniversary field trip in 1960, enough of the packages and controversies had been sorted out that the term "tectonic" could be included in the title of that trip. The trip revisited many of the former stops-- same rocks, new stories. By then, the new geologic map of Pennsylvania had just been completed, much of the crystalline core had been rejuvenated and become Paleozoic in age, multiple deformations and widespread structural superposition was beginning to be recognized more clearly, and the alpine-like nature of the Lebanon Valley nappes was beginning to unfold through work by Carlyle Gray and members of the Pennsylvania Survey, including Alan Geyer, Dave MacLachlan, and Sam Root.

By 1960, the first glimmers of the coming revolutions in tectonics had begun to appear: published data on modern sea floors would soon explode into plate tectonics, increasingly precise radiometric dates were about to place real numbers on the geologic time scale, while newly developing structural, sedimentological, geochemical, and petrological signatures were beginning to relate given rock units to specific tectonic environments. The original jig-saw puzzle was now beginning to be

recognized as several superposed tectonic puzzles. The task was just starting of sorting the pieces into their appropriate piles on several table-tops and actually beginning to assemble them.

It is difficult to overestimate the magnitude of change in the entire field of tectonics that occurred about 1960. A personal vignette during the compilation of the 1960 guidebook might be an illustration. That trip revisited Cloos's classic New Providence railroad cut that then provided excellent exposure of the Martic contact. I had convinced him to lead the discussion for that stop. In our pre-trip talks he continued to maintain his 1941 position of neutrality on the on the nature and origin of the actual Martic contact: possibly an unconformity, possibly a thrust. Even though his 1941 map identified and his text described the five-fold repetition of stratigraphy just north of the contact as thrusts, for the actual contact he continued to maintain that position of neutrality. He insisted the 1960 field trip notes leave his name off as co-author for that stop when I interpreted the contact as one more thrust in an imbricate stack. Equally bad, the field trip title included the word "tectonic." At that time, "Der Erdmeister" gave some fatherly advice to a young geologist. "Now, Don, you should give up this emphasis on tectonics. Tectonics is like fashion in ladies' skirts. Whether they go up or down means nothing in the long run. It is the descriptions of the rock outcrops that really last and are truly important." In a sense, he was quite correct in summarizing how the ephemeral tectonic arguments of the previous half century contrasted so sharply with the lasting contributions made by his careful geologic mapping and outcrop descriptions. Most tectonic ideas of those early times are in the dustbin of history but his outcrop observations are as valuable today as they were in 1941. He was quite correct about the past but at that moment we were standing on a hinge-pint in tectonic thought and his predictions of its future left a bit to be desired. Nevertheless they contained some advice, too often forgotten. The basic foot-on-the-rock contact with the outcrops should never be lost even as monumental changes of several tectonic revolutions continue to sweep through our science. In part I have taken his advice: larger tectonic questions of mountain-building mechanics, global sea level controls, lunar origin and the like have received much of my attention but the field geologist's detailed approach to outcrop examination has always been extremely important for me and my students.

**Figure 1. Potential geologic factors associated with the original rifted corner of the craton.** Rift edges (Rankin, 1976, Thomas, 1977) tend to have warm mantle under them and sink slowly over a broad area whereas transform edge should have a narrower coastal zone and less subsidence. Shoreward migration of island arcs and micro-continents is likely to partially close off a seaway, one possible explanation for the complex lithologies in the York County Winchester Domain. When the great Alleghanian continental collision occurred, the protruding corner was likely to suffer the greatest effect, being detached as the Reading Prong chip of basement bounded on the SW by the old transform fault. The potential sharp fault-bounded corner (white circle) is proposed as the origin of the corner of the Lancaster

**Piedmont Tectonics as of 2010**

Now, on this organization's 75th or diamond jubilee anniversary field trip and its 50th or golden anniversary trip, it seems appropriate to summarize the past half century's results. Over this time "tectonics" has gone from Cloos's category of tectonics being barely respectable to having the term
"plate tectonics" be part of everyday American vocabulary. Cloos even changed his mind and in later years began to make tectonic interpretations of the greater Blue Ridge. These post-1960 geologic revolutions have been so profound that younger readers may find it hard to imagine how convoluted and naïve were our pre-1960 views of the world; older readers may remember how difficult it was to keep track of all those apparently unrelated facts while trying to fit them into one of several poorly defined tectonic models.

In the half-century separating the two tectonic maps in the front of this volume, the tectonic jigsaw puzzle has changed from being a single puzzle into several superposed puzzles. The primary piece begins with Rankin (1976) and Thomas's (1977) transform fault that created an inside corner of the craton as a template (Figure 1). Almost all subsequent tectonic episodes have used that template to build the Pennsylvania Salient or its more restricted "elbow" geology. Now with the jigsaw puzzle of 2010, colors as well as patterns on many of the pieces finally seem to be established and many data sets are now organized into reasonably coherent local tectonic pictures and syntheses as indicated on Figure 2.

In effect, clusters of puzzle-pieces are beginning to be assembled on the "table top" even if their positions in the total picture remain somewhat ambiguous. The following list includes some high points the following speakers will make.

- Radiometric dating coupled with good mapping, petrological and structural analysis have begun to recognize the Paleozoic additions to the North American craton within the Potomac and Delaware cross-sections of the inner Piedmont. These additions take the form of a complex array of island arcs, micro-continents and intervening basin fills as will be described by Scott Southworth and Hal Bosbyshell respectively. Unfortunately, the lower Susquehanna elbow remains a bit of a no-man's-land across and into which projections of current data and tectonic zones from these adjacent areas seem to fit poorly.
- Radiometric dates with precisions approaching a million years have become abundant in the region as summarized in Bob Smith's paper but as yet are comparatively rare in the elbow.
- Closer to the Susquehanna, Gale Blackmer describes the original "Wissahickon Schist" as one of at least four separate formations that occur between the Martic region and the Delaware River. She describes how the early Paleozoic Octoraro Seaway was filled with the Octoraro and possibly Peters Creek schists that were later thrust up onto the craton and its carbonate platform.
- John Taylor describes the Cambro-Ordovician platform that developed on the rifted edge of the old craton and shed its debris down over submarine slopes into the adjacent deep seas.
- Bob Ganis summarizes the results of his seminal graptolite studies in the Taconic foreland basin, with evidence of far travelled deep sea allochthons that seem to invalidate former "Hamburg Klippe" interpretations.
- Wise uses recent radiometric dates of the Martic hinterland along with the new Ganis dates of foreland deposits to paint a picture of the Taconic Orogeny as a 15 - 20 my Ordovician event driven by roll-back tectonics of an offshore island arc to invert basin contents into largely gravity-driven mass that moved across the cratonic foreland.
- Mid-and late Paleozoic events are summarized by Gates and Valentino who point out evidence for widespread strike-slip strain across the region and emplacement of the rifted edge of the old craton across the entire Piedmont as Alleghanian mega-thrusts.
- Faill presents evidence and arguments questioning the traditional view of the Mesozoic basins as one-sided grabens that filled as they were tilted. Instead he presents evidence for a two stage model of a broad regional sag that filled with Mesozoic sediment and was intruded by large sills before a younger tectonic event tilted them into their present geometry.
Figure 2. Successive tectonic elements associated with an old transform fault and cratonic corner. The underlying base map on all panels is part of the frontispiece tectonic map. The upper left shows initial geometry with a corner near the city of York. In the upper right, the Cambrian facies of the formations onlapped by the Conestoga limestone slope deposit show the old corner was located in about the same place as the elbow from Figure 3. The next three panels show stages in the Taconic Orogeny. The lower right shows the major late Paleozoic basement-related thrusts along with previous panel transport arrows, a mixed age of transport directions but always being sub-perpendicular to their half of the elbow.
Pazzaglia opens an entirely new window on Piedmont tectonics, describing cosmo-chemical dating of terraces and other features of the lower Susquehanna. With these data, the rates of uplift and incision of different areas can be determined and potentially separated from those of glacial origin.

Parrish will unveil results of a brand new seismic line across the region. These appear to show the Piedmont mega-thrust sheet is much thinner than previously interpreted. As a result, the Survey is preparing to drill a test hole for velocity data and verification of the new thin-skin seismic interpretation.

**Susquehanna Elbow and a junction between N and S Appalachians**

Most of the above questions and problems are of relatively local interest but one overarching tectonic question seems emerge from this symposium and its trips: "How did these various processes interact to create a narrow orogenic belt across the orogen as the link between the Northern and Southern Appalachians?" The answer begins with the late-Precambrian rift corner (upper left, Figure 1) but the broader answer must lie somewhere in understanding the nature and evolution of the Pennsylvania Salient. More precisely, this junction must occur in the Susquehanna "elbow" that forms the axial zone of the salient. The following paragraphs and figures attempt to define the geometry of that elbow and link it with a few key tectonic elements.

The seemingly simple curve of the Pennsylvania Salient has a much more complicated geometry than is immediately obvious in map view (Figure 3.) From the south, the relatively linear Blue Ridge segment of the Southern Appalachians meets the equally linear Reading Prong of New England as a relatively sharp corner or elbow. The modern Susquehanna River follows that zone across most of the orogen about 40 km east of the central axis. Unlike any other east-flowing river of the Appalachians, the Susquehanna reaches the sea without having to pass through an external basement massif, an advantage that has allowed it to develop into the largest of the east-flowing drainage basins.

Figure 3. Contours of azimuths of dominant structural grain around the Susquehanna Elbow. The contours are not those of a simple smooth curve. Instead they show a sharp bend propagated from the Piedmont across the orogen. A double curve on either side (the pink and tan areas with reversed sense of curvature) show an outward bulge indicative of greater cumulative tectonic transport in the axis than on either limb, a "knobby elbow."

A contoured map of numerical values of azimuths of local tectonic grain (Figure 1) shows the corner elbow quite clearly. Only a few contours appear on the straight segments on either side of the elbow; most are concentrated as a parallel array down the center of the salient. Further, fold axes and
other tectonic elements traced along strike pass through the elbow with a double bend, a feature most easily seen on the figure as the outward extension of the Allegheny Front into the Plateau but also as zones of repeated azimuth values on either side of the elbow. Those contours show the double curvature extends NW across most of the orogen from at least the Piedmont's Martic Zone, across the rest of the Piedmont and through the entire foreland fold belt before transforming into more simple radial geometry of contours beyond the Allegheny Front.

In the Piedmont, the elbow axis is closely associated with tectonic transport. A plot of these vectors, each from small, individual study areas scattered throughout the region, shows tectonic transport falls into two distinct populations, each normal to the trend of its half of the elbow (Figure 4). Even though each half of the elbow had very large amounts of tectonic transport that diverged by about 30 degrees, the central axial zone in both Piedmont and foreland shows little sign of the expectable excessive stretching of fold axes or fabric. Instead, the two sets of Piedmont tectonic transport arrows are intermingled and overprinted in a 20-30 km wide zone west of the Susquehanna, a location coinciding with the parallel array of clustered contours that define the elbow.

Fig. 4. Tectonic transport vectors within the Susquehanna elbow. The fogged base map is the same as Figure 3. Each arrow is from an equal area net plot of data collected in some small field area. (Details are in Wise and Werner, 2004). In some areas the deformation includes both Taconic and Alleghanian structures but the same direction always seemed to be reactivated, perpendicular to that half of the elbow. Separating the two domains is a zone of overprinting, about 40 km west of the Susquehanna River and lying approximately on the center of the zone of maximum curvature of the elbow (green band).
The 19th century speculations on the origin of the salient have evolved greatly. The late-Precambrian transform fault that offset the rifting edge of the Laurentian Craton (Rankin, 1976, Thomas, 1977) and separated the two future halves of the US part of the orogen is now a part of the established tectonic picture. A number of workers has considered mechanisms by which successive tectonic events have propagated that early corner geometry across or around the cratonic edge without excessive stretching. Some use largely foreland motion vectors and geometry for mechanisms based on outward and radial flow of a foreland mass (Gray and others). Another model based on the two directions of Piedmont transport suggests multiple directions of displacement of the entire foreland and Piedmont (Wise, 2004).

**The Alleghanian megathrusts**

It has been long known, that the amount of shortening in the Appalachian fold belt requires massive displacement of the Piedmont. Great transport was at least proposed in Knopf and Jonas's (1929) Martic thrust and Stose's (1935) interpretation of the entire Reading Prong as a massively overthrust basement sheet. All of us know the entire Pennsylvania Piedmont is allochthonous but for the most part we tend to ignore it and concentrate on the nearer surface geology. With the release of the Pennsylvania Survey's new seismic line, this mega-thrust feature can become much better integrated into the regional tectonic picture. Existing cross sections of the Piedmont (Figure 5, after Wise and Ganis, 2008) showed the basement slab and some of its details but show the sheet as far thicker than the present seismic data would suggest.

![Figure 5. Schematic cross section of the Piedmont in the vicinity of the Susquehanna River (2008 version). The entire Piedmont is interpreted as being a vast mega-thrust that bulldozed the Appalachian Fold Belt ahead of it. The basement sheet of Mine-Ridge-Honeybrook upland is probably much thinner than shown (See Parrish discussion, this volume).](image)

The new interpretation of the Lancaster Seismic Zone (Figure 6 and discussion in notes for Stop # 4) suggest the Reading Prong-Honeybrook-Mine Ridge (RPHMR) thin sheet of basement can be traced to a sharp but shallowly buried corner near the Susquehanna River. As such this would represent the end point of New England basement in the shallow surface. Associated non-basement rocks may continue farther west toward the York Junction. The presence of this slab so far west, further narrows the gap in external basement massifs through which the Susquehanna flows. Its endpoint is probably the reason the RPHMR) arm of the Susquehanna elbow also terminates near the Susquehanna.
With the thin basement slab model of the RPHMR, the effects of its anisotropy on overlying structures need to be examined more closely. As suggested by dotted lines on Figure 6, the edge of preserved Conestoga Formation coincides with the Mine Ridge-Honeybrook suture, significant edge effects of the Mesozoic Hammer Creek alluvial fan coincide with an extension of the Mesozoic basement boundary, and the Oregon thrust probably represents a boundary between Reading Prong and Honeybrook basement areas. Where Mesozoic rocks were on the slab, subsequent tilting was relatively minor. Once beyond its edge and underlying strong support, the tilting became much greater to form the "narrow neck" of Mesozoic rocks. Not shown here, is Stose's (1924) "new type of structure in the Appalachians." This is a series of faults that extend westward into the area N and W of Lancaster, turning southward to end in that region. The most obvious large-scale effect is to drop the Ephrata graben, the extended yellow areas of Mesozoic on the frontispiece map. Some of these effects are discussed in the notes for Stop #7. One model would invoke later Mesozoic, left-lateral stain carrying through the slab and splaying into a series of normal faults of the Fruitville area near its western end.

![Figure 6. The Lancaster Seismic Zone as an edge effect of the shallowly buried Reading Prong-Honeybrook-Mine Ridge basement slab.](image)

A potential origin for this zone as the ending of an Alleghanian-age basement thrust against an old transform edge of the continent is illustrated with the tip in a white circle on Figure 1. Old boundaries between lithologic blocks within the sheet at shallow depths seem to have had strong hereditary control of younger structures. (data from Wise and Faill, 2002).

**Ongoing and future Piedmont Tectonic studies.**

In the jig-saw puzzle analogy, 2010 sees the beginning of a new phase. The edge pieces of the puzzle are now largely assembled into the bordering edges of several puzzles; within these bounds
several clusters of many pieces are now assembled even though their locations and interrelationships within the larger picture remain uncertain. However, various piles of different colored pieces are still lying all over the tabletops awaiting interpretation. And the bad news is: we probably are still missing many of the critical pieces. Some of the key points follow.

- A remaining open question is just how and if features interpreted from adjacent cross-sections make the transition into or through the Susquehanna Piedmont region. Structures and domain boundaries from Delaware and Potomac cross-sections are now being expanded into the Susquehanna elbow with emphasis on much more intense radiometric dating (Southworth, personal communication).
- The magnitude, dating and extent of strike-slip faulting proposed by Gates and Valentino remain controversial but are certainly far greater than commonly recognized. In particular these elements as well as directions of displacement along or near the Pleasant Grove-Huntingdon Valley shear zone are relatively unconstrained.
- The "Spechty Kopf" source problem of basement cobbles in an Upper Devonian formation in the middle of the Ridge and Valley folds continues to loom as a great imponderable. Dennis (2007) argues that these cobbles have glacial striations on them and are matched in lithology with sources in the Cat Creek Terrane of North Carolina. If so, alpine mountains hosted glaciers that extended far into the foreland basin but subsequently have been displaced by hundreds of km of right lateral offset along or near the Huntingdon Valley-Pleasant grove Shear Zone (frontispiece map). Alternatively, Smith (personal communication) argues that the cobbles can be matched with sources in the Maryland Piedmont. In either case, if the striations are really glacial, the mid-Paleozoic Piedmont Alps were high enough for glaciers to spread far into the foreland basin.
- The nature, extent, and geometry of island arcs, offshore micro-continents, and local basins is poorly known. Did the tectonic mélanges go over micro-continents or pass through gaps between them?
- Ultramafics are now spread along many of the faults. From whence these came and how they were emplaced at these locations remains a mystery.
- The thin nature of the Piedmont mega-thrust needs to be verified, along with how and if it thickens toward root zones to the SE. The geometry and character of the sedimentary units under the thrust sheet are completely unknown but are certain to get renewed attention.
- The two-stage Mesozoic Basin model of Faill seems reasonable but needs more testing. A deeper seated problem is the origin of the East Coast magnetic anomaly. This regional feature of the East Coast seems to follow the Mesozoic basins at depth as some kind of a transition from continental crust toward a denser semi-oceanic crust. How or even if the anomaly is related to the Mesozoic Basins is unknown.
- The new model of the Lancaster Seismic zone, as a shallowly buried basement slab bounded by the intersection of the Reading Prong Thrust and the old transform fault, was proposed above and is discussed further in the notes for Stop #4. As yet that model is untested but if valid, it further constricts the gap in external basement massifs through which the Susquehanna flows and provides a reason for the abrupt ending of the east half of the Pennsylvania Salient or "elbow." It also may explain by tectonic hereditary, many of the structures in its thinly buried cover.

**Conclusion.**

Even though tectonic understanding of the Susquehanna Piedmont has advanced greatly in the past half century, its complex, overprinted geology renders it one of the least understood areas of the Appalachians. Unfortunately, it also contains one of the most critical boundaries of the chain, the junction between the two US halves. The boundary occurs within a ~30 km wide band centered
about 40 km west of the Susquehanna River and appears to be a much sharper break than most would imagine. It passes in or near the city of York and hence might be called the York Junction. A major task of the coming generation of geologists should be to unravel the complex geology of this zone. This symposium and its field trips represent a step in that direction, helping define the locus more clearly and providing new information and tectonic models for areas and problems in and around it. It would make an excellent candidate for an Earthscope array.

References


TECTONICS OF THE MARYLAND PIEDMONT ALONG THE POTOMAC RIVER: INSIGHT SINCE 1960 AND POTENTIAL TRANSFER TO THE PENNSYLVANIA PIEDMONT

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Introduction

This is a summary of a half century of research in the Maryland Piedmont and how it may or may not have implications for the Piedmont of Pennsylvania. Much of the field mapping and all of the isotopic analyses of rocks and minerals of the Maryland Piedmont have been conducted since the 1960 Field Conference of Pennsylvania Geologists “Some tectonic and structural problems of the Appalachian Piedmont along the Susquehanna River”. The Piedmont rocks of Maryland and Pennsylvania occur in a critical place within the central Appalachian Pennsylvania embayment (Thomas, 1977), which likely contributed to the distribution of lithologies and structures.

In the last 50 years plate tectonics and accretionary structures (Drake, 1989) have superceded the geosynclinal model for the stratigraphy (Hopson, 1964). Isotopic techniques largely developed and applied in the last 20 years have significantly improved our understanding of the tectonic history. The complex depositional setting of the rise and slope prism of the Laurentian margin, represented by rocks of the Westminster terrane, and the accreted rocks of the Potomac terrane, coupled with multiple metamorphic and deformation events with many out-of-sequence faults, make working in the Piedmont a continuing challenge. Lessons learned along the Potomac River hopefully can be applied to rocks along the Susquehanna River to see how different the geology is across the Mason-Dixon Line.

The geology here is presented by elements and by terranes. The Potomac terrane is juxtaposed against the Westminster terrane along the Pleasant Grove fault (Drake, 1989; Horton and others, 1989; 1991). Our field mapping (Southworth and others, 2007), the systematic collection of samples for \(^{40}\text{Ar}/^{39}\text{Ar}\) analyses (Kunk and others, 2004; Wintsch and others, 2010), and new U-Pb ages of detrital zircons (Martin and others, 2010) provide a better understanding of the protoliths and the time and conditions of deformation and metamorphism. These data suggest some important relationships: (1) a change in the provenance of detrital zircon ages occurs within the rocks of the Westminster terrane and extends eastward across the Potomac terrane rocks, (2) regional foliations across the Piedmont formed over a period of ~175 million years, (3) previously recognized fault strands were reactivated to form broad high strain zones, (4) faulting and shearing across the Piedmont were not in sequence, and (5) the dominant deformation events preserved across the Piedmont were Ordovician in the middle, Silurian in the extreme west, with the majority being Devonian to Carboniferous in age.

Lithologies

Westminster terrane

The Westminster terrane is comprised of a heterogeneous suite of sedimentary and volcanic rocks that were metamorphosed at greenschist-facies conditions (fig. 1). Formational names and traditional stratigraphic principles are difficult to apply to the complex depositional setting of the rise and slope here and past practices of applying them in that mode has resulted in confusion. The use and misuse of formational names led to erroneous interpretations based on stratigraphic constructs (Scotford, 1951; Stose and Stose, 1951; Cleaves and others, 1968; Edwards, 1984; Edwards, 1994, for example). The mapping of lithologic units is preferred, such as metagraywacke, slate, phyllite,
conglomerate, greenstone, etc., as well as tectonites, such as phyllonite (Knopf, 1931) and migmatite. Some lithologies are not restricted to specific formations. For example, greenstone (likely metabasalt), recognized as the Sams Creek Formation, also occurs within Ijamsville Phyllite and Marburg Formation. Additionally, Wakefield Marble stratigraphically overlies Sams Creek metabasalt but marble is also found within several different phyllite units. The lithologic map (fig.1) reveals the complex nature of stratified sediments of the rise and slope, which includes lenses (channels) of clastic debris and sedimentary blocks of rock (olistostromes).

**Potomac terrane**

Rocks along the Potomac River were mapped as three parts of the Wissahickon Formation (Boulder gneiss, Metagraywacke, and Upper Pelitic schist) (Fisher, 1970). The Boulder gneiss was later named the Sykesville Formation (Crowley, 1976). The Wissahickon rocks were reclassified as the Peters Creek Formation (Jonas and Knopf, 1921; Drake and Morgan, 1981) and subsequently renamed the Mather Gorge Formation (Drake, 1994). Zones of migmatite and phyllonite were recognized within the formation (Fisher, 1970; Drake and Froelich, 1997). New research (Kunk and others, 2004; 2005; Southworth and others, 2006; Wintsch and others, 2010) demonstrates that
disparate metamorphic and structural histories of the rocks require the use of “domains” rather than stratigraphic units.

![Figure 2.](image)

**Pennsylvania**

Most of the Westminster terrane here was classified as schist (including gneiss, quartzite, and phyllite) (Berg and others, 1984). Layers of greenstone mapped in Maryland pinch out toward the Susquehanna River and one closed depression was mapped as Wakefield Marble (Knopf and Jonas, 1929). Most of the rocks on the north side of the Susquehanna River were called Octoraro Formation (Valentino and others, 1994; Octoraro Phyllite of Lyttle and Epstein (1987) and Octoraro Schist of Knopf and Jonas (1929)), yet the rocks southwest of the river were called Marburg Schist (Jonas and Stose, 1938a, b) and renamed the Marburg Formation by Drake (1994). Rocks formerly assigned to the Peters Creek Formation were redefined into two members of the Peters Creek Formation and three new members of the Scott Creek Formation, all of which are interpreted to be stratigraphically above the Octoraro Formation (Fail and Smith, 2010). Rocks of the Peters Creek Formation were considered to be Neoproterozoic turbidites of the Laurentian margin, correlative to Lynchburg and Fauquier formation rocks of the eastern Blue Ridge (Valentino and Gates, 1995).

**Ages of the protoliths**

**Westminster terrane**

No fossils have been reported from rocks in this terrane. Samples of Silver Run Limestone and Wakefield Marble collected at several localities were dissolved in acid and were barren of conodonts and other fossils, suggesting that they are older than Upper Cambrian (D.K. Brezinski, Maryland Geological Survey, oral comm., 2009, and J. Repetski, USGS, oral comm., 2009). Metabasalt of the Sams Creek Formation is chemically indistinguishable from metabasalt of the
Catoctin Formation that occurs to the west in the Blue Ridge-South Mountain anticlinorium. Metarhyolite near the base of the Catoctin in northern Virginia and at the top of the formation on South Mountain, PA, provide zircon U-Pb TIMS ages of 571 and 563 Ma, respectively (Southworth and others, 2009). Therefore, the Sams Creek Formation is likely Neoproterozoic in age. Rocks classified as Libertytown Metarhyolite (Stose and Stose, 1946) are phyllites that do not contain igneous zircons (J.N. Aleinikoff, USGS, oral comm., 2010). Rocks of the Sugarloaf Mountain Quartzite and Urbana Formation that underlie the Sugarloaf Mountain anticlinorium are exposed in a tectonic window through the Martic thrust sheet. They are interpreted to be equivalent in age to the Early Cambrian Weverton and Harpers Formations that under the Blue Ridge-South Mountain anticlinorium to the west (Southworth and others, 2007). Nickelsen (1956) reported fragments of trilobite in the Harpers Formation near Harpers Ferry, W.Va., but locally elsewhere the Harpers contains only the trace fossil *skolithus linearis*.

*Potomac terrane*

No fossils have been reported from rocks of this terrane either. The minimum age for sedimentary rocks of the Sykesville Formation is constrained by muscovite trondhjemite of the Dalecarlia Intrusive Suite that yielded a zircon U-Pb SHRIMP age of 478±6 Ma (Aleinikoff and others, 2002). The minimum age for sedimentary rocks of the Mather Gorge Formation is constrained by bodies of amphibolite that yielded a 40Ar/39Ar amphibole cooling age of 475 Ma (Kunk and others, 2005). Metagraywacke of the Mather Gorge Formation contains a U-Pb detrital zircon ~580 Ma and diamictite of the Sykesville Formation contains several detrital zircons ~550 Ma (Martin and others, 2010). Thus, rocks of the Potomac terrane are poorly bracketed as Neoproterozoic to Ordovician. These ages support the interpretation that the Sykesville Formation diamictite was deposited unconformably on Mather Gorge Formation (Morgan Run Formation of Muller and others, 1989).

*Pennsylvania*

U-Pb ages of detrital zircons from rocks assigned to the Cambrian Harpers and Antietam Formations and rocks of the Octoraro and Peters Creek Formations suggest that they are all younger than Early Cambrian, at ~530 Ma (Blackmer and Bosbyshell, 2010).

*Structures*

*Foliations and folds*

The primary foliations in the rocks include bedding, cleavage, and schistosity. These are overprinted by phyllonitic and mylonitic foliations, shear band cleavage, and spaced crenulation cleavage. Characteristically, the rocks contain composite foliations that mostly strike to the northeast. The foliations in the Westminster rocks dip steeply to the southeast. High strain zones resulted in transposition and formation of phyllonitic and mylonitic foliations (Fleming and Drake, 1998; Southworth and others, 2007). Transposition foliation containing folded vein quartz is structurally aligned parallel to faults and shear-band cleavage formed in high strain zones. Isoclinal folds were refolded by north-trending upright to steep folds (Jonas, 1937; Fisher, 1978; Drake, 1989) with S2 axial- planar cleavage. Adjacent to faults and high strain zones locally are steep, tubular sheath folds and antiforms of foliation that have vertical crenulation cleavage (Southworth and others, 2007). Late contraction folded the schistosity and cleavage into antiforms and synforms. The largest antiform is along Parrs Ridge in the Westminster terrane and was likely related to formation of the Tucquan arch (Stose and Stose, 1946; Valentino and others, 1994) in Pennsylvania. The dominant foliation in the Potomac terrane rocks is schistosity which forms a pseudo fan structure along the Potomac River (Drake, 1989); schistosity in the west dips gently eastward and in the east it dips steeply westward, but they are not of the same generation (Kunk and others, 2004).
**Faults and high strain zones**

**Westminster terrane**

For many years the controversial Martic Line was the only fault recognized in the Piedmont (Cleaves and others, 1968). The statement by Cleaves and others (1968) that “the Martic Line was interpreted as a fault; however, stratigraphic relationships in Maryland may not require a major fault along this line”, was likely influenced by Weaver’s (1954) research in Pennsylvania. Drake first portrayed the Martic fault (MF) around the Sugarloaf Mountain anticlinorium as framing a tectonic window (Rankin and others, 1990). This window and the smaller Bush Creek window were verified by field mapping (Southworth, 1996; Southworth and others, 2007) and $^{40}\text{Ar}^{39}\text{Ar}$ analyses of minerals in cleavage (Wintsch and others, 2010; Southworth and others, 2010). Erosion of the folded thrust sheet formed klippen; the westernmost belt of Ijamsville Phyllite occupies a long narrow klippe. The MF experienced dextral strike slip motion in the Alleghanian orogeny, as suggested by Horton and others (1989), but displacement is negligible (Southworth and others, 2006; 2010).

Several regional thrust faults (Southworth and others, 2007) were pre- to syn-metamorphic as there is no metamorphic age discontinuity across them (Wintsch and others, 2010). Elsewhere, broad high strain zones were recognized by $^{40}\text{Ar}^{39}\text{Ar}$ metamorphic discontinuities (Wintsch and others, 2010), but not in outcrop (Southworth and others, 2007).

**Potomac terrane**

Major faults here separate distinct bedrock units and formations (Drake, 1989). Retrograde shear zones (Fisher, 1970) were mapped as the Pleasant Grove fault (PGF) and Plummers Island fault (PIF), respectively (Drake, 1989) (fig. 2). The southeast-dipping PGF was interpreted to be the Taconian suture, which thrust Mather Gorge Formation, rocks of the Potomac terrane onto rocks of the Westminster terrane. The northwest-dipping PIF was interpreted to be the sole of a Cambrian thrust sheet of Mather Gorge Formation whose debris formed the underlying diamictite of the Sykesville Formation (Drake, 1989). The reactivation of all of these fault strands created high strain zones and shear zones that are as much as 13 km wide (fig. 2). For example, the Rock Creek shear zone was initially a sinistral strike slip fault during emplacement of the Ordovician Kensington Tonalite, later reactivated as a dextral strike slip fault (Fleming and Drake, 1998; Fleming and others, 1994) at ~350 Ma with brittle faulting at ~306 Ma (Kunk and others, 2010). As suggested by Horton and others (1989) and Krol and others (1999), the PFG experienced dextral strike slip motion in the Alleghanian orogeny (Kunk and others, 2004; Wintsch and others, 2010) as evidenced by kinematics of shear band cleavage (Southworth and others, 2006), but the amount of displacement is unknown. The surface trace of the PGF (Drake, 1989) was expanded to the 1-2 km wide Carboniferous Pleasant Grove shear zone (Krol and others, 1989, and further expanded to be at least 5 km wide (Wintsch and others, 2010).

**Pennsylvania**

The MF is clearly folded (Wise, 1970; Wise and Ganis, 2009). There is evidence of pervasive strike slip strain associated with it and to the north of it in Lancaster County (Valentino and Gates, 1995), but no single large fault zone has been identified with certainty. The PFG has recently been called the Delta Duplex and a schematic cross section (Faill and Smith, 2010) implies minor displacement of stratigraphic units across it.

**Metamorphism**

The Piedmont rocks were deformed and metamorphosed at different times, at different grades, and under different conditions, and subsequently were juxtaposed and transported westward during Paleozoic orogenesis. The Paleozoic metamorphic grade is highest (sillimanite) in the west.
central part of the Potomac terrane, and lowest (chlorite-sericite) in the western Westminster terrane, but it is not a simple Barrovian sequence related to one event.

**Westminster terrane**

These rocks experienced a complex metamorphic history under greenschist-facies conditions (Wintsch and others, 2010). They contain varying proportions of muscovite, chlorite, paragonite, quartz, magnetite, calcite, actinolite, epidote, stiplnomelane, and sericite. Retrograde phyllonitic rocks contain multiple generations of white mica. Chloritoid and albite crystals overgrew foliation in the eastern area, while rocks in the western area (Sugarloaf Mountain anticlinorium beneath the MF), contain only sparse sericite and chlorite in cleavage. Muscovite grew first in the lower thrust sheets (western area) and later in the higher thrust sheets (eastern area) during a Late Devonian reactivation of the Pleasant Grove shear zone. Early Silurian muscovite ages (439-437 Ma) (Wintsch and others, 2010) across the MF suggest that it was folded at that time but emplaced earlier (Southworth and others, 2010).

**Potomac terrane**

These rocks experienced at least two prograde amphibolite-facies events and several retrograde and secondary prograde greenschist-facies events (Fleming and Drake, 1998). In the sillimanite-grade zone (Fisher, 1970; Drake, 1989), migmatization occurred in the Ordovician (Becker and others, 1993; Kunk and others, 2005) as indicated by amphibole cooling ages of ~475 to ~455 Ma (Jäeger, 1979). Metamorphism was complete by the Late Devonian as indicated by the 363±3 Ma SHRIMP crystallization age of the nonmetamorphosed and undeformed Guilford Granite (Aleinikoff and others, 2002) and the cooling ages of two biotite samples from undeformed lamprophyre dikes (363 ± 13 Ma and 360 ± 13 Ma, 2σ) (Reed and others, 1970). The ages and closure temperatures of amphibole, S1 and S2 muscovite, and zircon fission-tracks show discontinuities across faults (see Kunk and others, 2004; Southworth and others, 2006; Wintsch and others, 2010). Cooling curves converge between 300 and 280 Ma, when the rocks were uplifted and cooled through the zircon closure temperature of ~235 °C at ca. 280 Ma (Naeser and others, 2004; Kunk and others, 2004; Wintsch and others, 2010) during westward transport to higher crustal levels above the North Mountain thrust fault in the Permian (Southworth and others, 2006). Apatite ages decrease from ~200–180 Ma at the eastern Piedmont to ~115 Ma in the western Blue Ridge, as the loading of crust by Coastal Plain strata caused uplift to the west (Naeser and others, 2004). The Piedmont rocks were at temperatures in excess of 90–100 °C before the latest cooling through the apatite closure temperature in Early Jurassic to Early Cretaceous time (Naeser and others, 2004).

**Pennsylvania**

Blackmer and others (2007) reported ~400 Ma white mica and ~365 Ma cooling ages of white mica across the MF and WT and PT rocks northeast of the Susquehanna River. Wise and Ganis (2009) demonstrated that Taconian metamorphism and deformation across the MF near the Susquehanna River was ~461 to 443 Ma. About 100 km to the southwest, deformation and metamorphism was at ~443 Ma and ~365 Ma. However, in the WT, the earlier event was restricted to the westernmost rocks and the later event was restricted to the easternmost rocks (Wintsch and others, 2010).

**Baltimore Gneiss Domes and Wilmington Complex**

Structurally and tectonically distinct from the Potomac terrane rocks are the Baltimore Gneiss Domes and Wilmington Complex, which extend from Maryland north to Pennsylvania and Delaware. SHRIMP U-Pb titanite metamorphic ages of 374 ± 8, 336 ± 8, and 301 ± 12 Ma (Aleinikoff and others, 2004) define thermal events recorded in the rocks of the Baltimore Gneiss Domes. Amphibolite- to granulite-facies metamorphism of rocks of the Wilmington Complex is
recorded by ages of metamorphic zircon (432 ± 6 and 428 ± 4 Ma) and monazite (429 ± 2 and 426 ± 3 Ma), and at least three other thermal episodes are recorded by monazite growth at 447 ± 4, 411 ± 3, and 398 ± 3 Ma (Aleinikoff and others, 2006).

Tectonic domains

Westminster terrane

The Parrs Ridge fault (PRF) was determined to be a metamorphic discontinuity (Wintsch and others, 2010). Rocks to the west and beneath the PRF were emplaced, and later folded with axial planar cleavage ~430 Ma, whereas rocks of the upper plate were reactivated ~370 Ma during motion along the PGF. U-Pb detrital zircon ages (Martin and others, 2010) suggest another important boundary is along the Hyattstown fault (HF; Southworth and others, 2007) that is situated between the PRF and MF. The HF placed Marburg Formation on Sams Creek Formation pre- or syn-430 Ma (Wintsch and others, 2010; Southworth and others, 2010). Of note, populations of detrital zircons in the Marburg Formation resemble those from rocks of the Potomac terrane (Martin and others, 2010).

Potomac terrane

40Ar/39Ar data allowed the rock formations to be further subdivided into domains (Kunk and others, 2004; Wintsch and others, 2010). The Mather Gorge Formation (Drake and Froelich, 1997) was subdivided into three domains. Only the middle domain retains the Taconian age metamorphic signature, as the adjacent domains were reactivated in the Early (east) and Late (west) Devonian. Populations of detrital zircons change across the Blockhouse Point fault (BPF), and suggest a significant boundary prior to Devonian deformation (Martin and others, 2010). Specifically, rocks in the western domain may not be Mather Gorge Formation, as they received detritus from different source areas.

Acknowledgements

The author gratefully acknowledges Avery A. Drake, Jr., for providing the opportunity to work in the Piedmont. His mantra “geology is an iterative process” resulted in our enhanced understanding of Piedmont tectonics. I also thank Don Wise for his continued interest and enthusiasm on Piedmont geology and for inviting me to participate in his second 50th anniversary. Reviews by Don Wise and Mark Carter were helpful. Funding and support by the USGS National Cooperative Geologic Mapping Program is gratefully appreciated.

References Cited


SOME THOUGHTS AND SPECULATIONS ON RADIOMETRIC DATES RELEVANT TO PENNSYLVANIA

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Observations related to the compiled tables.

1) The age of the classic Grenvillian Orogeny has stood the test of time better than most early (double meaning intended) dates. McLelland et al. (2001), among others, have done a fine job teasing out the distinct ~ 1.25 Ga Elzeverian and ~ 1.05 Ottawan stages within the Grenvillian. Mainly through observation in the Reading Prong, Smith suspects an additional post-Grenvillian extensional stage seen as U, Mo, and REE-bearing dikes. Joe Pyle (2006) may have extended this post-Grenvillian activity to the Honey Brook Upland of Chester County. The Grenvillian also seems to have been a period when many magmas were first separated from the mantle to later continue their saga during long-term storage beneath the Laurentian crust until it was attenuated by later, post-compressional extensions.

2) The circa 735 Ma Neoproterozoic extension seems to have had far more widespread effects than once envisioned. For example, it is the intrusion/extrusion age of the older members of the Robertson River Igneous Suite of Dick Tollo and the mantle separation age for the Baltimore Mafic Complex and Sword Mountain Olivine Mellilitite. Vestiges of the ~ 735 event may eventually be confirmed in places such as the Morgan Hill shear zone, Reading Prong, of Avery Drake, the informal mafic “C.K. Williams Quarry metabasalt” of Smith and Barnes, and the felsic Marble Mountain volcanics of Avery Drake, Rich Volkert, and others. Various other faults now assumed to be post-Neoproterozoic based on present geometry may also eventually be found to be reactivated, early rifting faults. These seem to be especially possible in the Honey Brook Upland where Joe Pyle has already detected granulite facies metamorphism of approximately this age suggesting the possibility of associated crustal thinning.

3) Ediacaran extension yielded the Catoctin Metabasalt and Catoctin Metarhyolite. These volcanics are associated with what is probably the most underrated event in the central Appalachians. Smith maintains that the initial rifting basalts gradually evolved over some millions of years to more oceanic drifting metabasalts. These provide an interesting tool to help establish terrain boundaries and interpret associated sedimentary sequences. Some Catoctin derivatives, such as the Jonestown Volcanic Complex, with an interesting gouge zone at the base, are far traveled. At present, Smith cannot even rule out highly metamorphosed Catoctin volcanics in the Honey Brook Upland.

4) Nice work, primarily by John N. Aleinikoff, has done a fine job establishing an age of ~ 480 Ma for the Wilmington Complex. This is slightly younger than dates provided for the Baltimore Mafic Complex. Although both their igneous ages appear to be too old to have been directly part of the Taconic orogeny, they might still have been Taconian impactors.

5) The Taconic orogeny has been nicely established at 450 Ma and the James Run Island Arcs, dated at 459 Ma, are likely one of the significant impactors, which also left their marks as bentonites, observed in the Ridge and Valley. As such they may both provide décollement surfaces in the SE portion of the Ridge and Valley and mark the end of a 100 Ma carbonate shelf. For the James Run Formation, consider high elevation stratovolcanoes.

6) Like most good compressional orogenies, the Taconic was followed by a very widespread extensional magmatic and thermal event (STM) at 432.6 Ma. Smith et al. (2004) couldn’t even rule out that this extension initiated the Appalachian basin.

7) Except for what the wind may have blown in, discrete evidence for the Acadian orogeny at the latitude of Pennsylvania remains difficult to find despite evidence to the NE and SW.

8) It has been difficult to date the Alleghanian orogeny so far, but Ar/Ar of micas in faults should help. We know it exists, but need a strategy to radiometrically date the various recognized stages. It seems likely that there is also a dateable late extensional phase. Or perhaps the elephant is already in the room and we can’t see it because it is too obvious. Those apparent Mesozoic faults cutting the Bald Friar Metabasalt along Brandywine Creek, Chester County may be older, Late Alleghanian faults.

8) Because it is so well recorded in titanite, zircon, and apatite FTA and other data, the extremely widespread Mesozoic Thermal Pulse (MTP) is both impressive and provides a nice picture of post-Alleghanian erosion. There appears to have been a continuum of cooling rather than a series of discrete heating events.

9) Finally, if there is a gross cyclicity (to think pretoriously, as in Desmond Aubrey Pretorius of South Africa) to extension in the mid-Atlantic region, where might the Eocene volcanics in Pennsylvania be located?

Bibliography for Annotated Table of selected dates relevant to southeastern Pennsylvania


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<td>Early Neoproterozoic extension and hydrothermal (?)</td>
<td>Uraninite-Thorianite, Reading Prong, Northampton County, Pa.</td>
<td>$^{207}\text{Pb}/^{206}\text{Pb}$ TIMS</td>
<td>948 +/- 5</td>
<td>Est. of age of U-Th mineralization in Chestnut Hill portion of Reading Prong. Gray and Zeitler (1997) found 948 in detrital zircons in Pottsville.</td>
<td>R. I. Grauch and K. R. Ludwig (1979) on 2nd, deeper 3.5 m channel sample cut by RCSII.</td>
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<td><strong>Cryogenian Extension and Magmatism</strong></td>
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<td>Separation of ultramafic component of BMC island arc from mantle</td>
<td>OIrRuFeNi (52.3% Os) Lancaster County, Pa.</td>
<td>$^{187}$Os/$^{188}$Os TIMS</td>
<td>735 Ma</td>
<td>Similar to igneous age RRS volcanism of Tollo and Aleinikoff (1996). Also see Sword Mountain O.M.</td>
<td>Analyses by Ryan Mathur at University of Arizona on micronugget collected R.C.S.II.</td>
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<tr>
<td>Separation of ultramafic component of Clear Spring Olivine Melilitite</td>
<td>Olivine Melilitite, Washington County, Md.</td>
<td>Isotope dilution for elemental Nd and Sm. Isotopes by TIMS.</td>
<td>735 Ma</td>
<td>$T_{Nd}$ model mantle separation of ultramafic Sword Mountain Olivine Melilitite</td>
<td>K. A. Foland in Smith, Foland, and Nickelsen (2004).</td>
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<tr>
<td><strong>Ediacaran Extension and Magmatism</strong></td>
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<td>Peralkaline Rittenhouse Gap Felsite dikes. (Associated with ~ similar age Tunnel Mountain Metadiabase dikes of mantle origin.)</td>
<td>Zircon, Berks County, Pa.</td>
<td>Avg. of concordant $^{206}$Pb/$^{238}$U, $^{207}$Pb/$^{235}$U TIMS.</td>
<td>602.3 +/- 2 Ma. Less precise 628 Sr/Rb whole rock isochron</td>
<td>Geochemically, this fits the peralkaline Battle Mountain of the Robertson River Igneous Suite, but the age is closer to the Catoctin. Therefore, stages of Neoproterozoic extension might not be as distinct as once assumed.</td>
<td>Jahandar Ramezani, M. I. T., in Smith (2003). Latter includes Sr/Rb and Nd/Sm isotopes by K. A. Foland, O.S.U.</td>
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<td>Catoctin Metarhyolite</td>
<td>Zircon, ~8 km SSE Fickels Hill, Adams County, Pa.</td>
<td>$^{206}\text{Pb}/^{238}\text{U}$ $^{207}\text{Pb}/^{235}\text{U}$ TIMS “concordia”</td>
<td>564.3 +/- 9.3 Ma</td>
<td>Points widely scattered. Four fractions having lowest $^{207}\text{Pb}/^{206}\text{Pb}$ used.</td>
<td>Aleinikoff et al. (1995)</td>
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<td><strong>Ordovician Island Arcs</strong></td>
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<td>Nd/Sm Three concordant isochrons.</td>
<td>TIMS for mineral separates from 2 gabbros and a norite.</td>
<td>490 +/- 20 Ma</td>
<td>Large error, but consistent with Sinha (1997) zircon discordia.</td>
<td>Shaw and Wasserburg (1964)</td>
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<td>Zircons, highly zoned, New Castle County, Delaware</td>
<td>SHRIMP U-Pb</td>
<td>476 +/- 4 to 483 +/- 7 Ma</td>
<td>Range of ages of igneous crystallization for four units. Also metamorphic monazite: 447 +/- 4, 411 +/- 3, and 398 +/- 3.</td>
<td>Aleinikoff et al. (2006)</td>
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<td><strong>Taconic Orogeny</strong></td>
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<td>Taconic metamorphism of BMC</td>
<td>Lower intercept for zircon discordia</td>
<td>TIMS for 4 zircons from 3 pyroxene gabbros and 1 norite.</td>
<td>453 +/- 11 Ma</td>
<td>Taconian metamorphic zircons formed from release of Zr from alteration of cpx. to amphibole.</td>
<td>A. K. Sinha et al. (1997)</td>
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<td>Taconic Orogeny</td>
<td>Monazite Kline Quarry, Wrightsville, York County, Pa.</td>
<td>Th-U-Total Pb by Ultracehron electron microprobe</td>
<td>NNNN? In progress.</td>
<td>Monazite is in an assemblage similar to that at Pequea, but nil evidence for Taconic in York County.</td>
<td>Michael J. Jercinovic, U. of Mass., (p.c. to Don U. Wise and R.C. Smith, II, N/NN/10)</td>
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<td>Wilmington Complex Faulkland Gneiss metamorphic monazite</td>
<td>Monazite, New Castle County, Delaware</td>
<td>$^{206}$Pb/$^{238}$U SHRIMP</td>
<td>449.3 +/- 5.0 and 447.8 +/- 2.5 and Ma</td>
<td>Permissive of correlation with Taconic Orogeny and above noted bentonites.</td>
<td>Aleinikoff et al. (2006)</td>
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<td>Early Silurian Thermal and Magmatic Event</td>
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<td>Late Taconian stitching of obducted BMC onto Laurentia</td>
<td>Baddeleyite in “iron ore” highly enriched in CrTiZ, Lancaster County, Pa.</td>
<td>TIMS determination of U-Pb isotopes.</td>
<td>441.7 +/- 7.3 Ma from isochron upper intercept (6 points typically having 1% discordance)</td>
<td>Late Taconian, but large analytical error so can’t rule out Silurian Thermal and Magmatic event.</td>
<td>Analyses by Jahandar Ramezani, MIT, in Smith and Barnes (2008).</td>
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<tr>
<td>Late Taconian stitching of obducted BMC onto Laurentia</td>
<td>Monazite, Faulkland Gneiss of Wilmington Complex, New Castle County, Delaware</td>
<td>$^{206}$Pb/$^{238}$U SHRIMP</td>
<td>441.7 +/- 6.9 Ma for core. Rim is 414.8 +/- 6.7 Ma</td>
<td>Late Taconian, but large analytical error so can’t rule out Silurian Thermal and Magmatic event for core.</td>
<td>Aleinikoff et al. (2006). They placed it into a statistically generated 447 +/- 4 Ma grouping.</td>
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<td>Wilmington Complex prograde amphibolite to granulite facies meta-morphism.</td>
<td>Zircon, metamorphic, New Castle County, Delaware</td>
<td>SHRIMP</td>
<td>432 +/- 6 Ma</td>
<td>Part of STM. Metamorphic monazite 429 +/- 2 for Faulkland Gneiss and 428 +/-4 Ma Brandywine Blue Gneiss.</td>
<td>Aleinikoff et al. (2006)</td>
</tr>
<tr>
<td>Wissahickon metamorphic monazite</td>
<td>Monazite, near Yorklyn, Newcastle County, Delaware</td>
<td>U-Pb concordia by TIMS</td>
<td>424.86 +/- 0.50 Ma. SHRIMP yielded 426 +/-3 Ma.</td>
<td>Cooling from STM in lower portion of Wissahickon Formation. Detrital zircons include peaks at ~ 1050 and ~ 950 Ma. Youngest 735 Ma.</td>
<td>Aleinikoff et al. (2006)</td>
</tr>
<tr>
<td>Bald Hill Bentonite C, type locality. BHB B and BHB A are older.</td>
<td>Zircon, Blair County, Pa.</td>
<td>$^{206}\text{Pb}/^{238}\text{U}$ by TIMS</td>
<td>~ 417 Ma. Close to Sil.-Dev. boundary.</td>
<td>Geochemistry suggest from an island arc. For strat. position at end of Silurian carbonates in Pennsylvania see Smith et al. (1988).</td>
<td>Jahandar Ramezani, M.I.T. Additional work in progress. See Aleinikoff et al. (2010) for 419.3 +/- 0.4 diorite, Vt.</td>
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<td>Devonian Island Arcs and Acadian Orogeny (?)</td>
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<td>Acadian Orogeny ???</td>
<td>Zircon, metamorphic Brandywine Blue Gneiss and Faulkland Gneiss, New Castle County, Delaware</td>
<td>$^{206}\text{Pb}/^{238}\text{U}$ by SHRIMP</td>
<td>411 +/- 3 and 399 +/-3</td>
<td>Metamorphism, possibly in the poorly constrained Acadian Orogeny.</td>
<td>Aleinikoff et al. (2006)</td>
</tr>
<tr>
<td>Tioga Ash Bed B</td>
<td>Monazite, Zeigler Pit, Union County, Pa. (Zircons showed recent Pb loss.)</td>
<td>$^{207}\text{Pb}/^{205}\text{U}$, but said to be concordant or v. slightly reverse discordant TIMS.</td>
<td>390.0 +/- 0.5 Ma</td>
<td>Geochemistry suggests from a peraluminous island arc. Sample from Way et al. (1986) which also gives strat. position. Slight excess $^{206}\text{Pb}$, so didn’t use $^{206}\text{Pb}/^{207}\text{Pb}$.</td>
<td>M. K. Roden, R. R. Parrish, and D. S. Miller (1989)</td>
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<td><strong>Wilmington Complex Faulkland Gneiss metamorphic monazite</strong></td>
<td>Monazite, Newcastle County, Delaware</td>
<td>$^{206}\text{Pb}/^{238}\text{U}$ SHRIMP</td>
<td>387.6 ± 4.3</td>
<td>Docking island arc???</td>
<td>Aleinikoff et al. (2006)</td>
</tr>
<tr>
<td><strong>Ellicott City Granodiorite calcalkaline peraluminous I-type</strong></td>
<td>Zircon (?), Ellicott City, Howard County, Md.</td>
<td>?</td>
<td>375 Ma</td>
<td>Acadian Orogeny. Other radiometric ages reported: 450 +/- 20 and 458 Ma may be Taconian metamorphism. However, cannot rule out relation to Springfield Granodiorite.</td>
<td>A.K. Sinha (p.c. to J. N. Aleinikoff, 1999) in Aleinikoff et al. (2002)</td>
</tr>
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<td><strong>Guilford Granite, calcalkaline peraluminous S-type monzo-granite having two micas</strong></td>
<td>Zircons, Guilford, Howard County, Md.</td>
<td>$^{206}\text{Pb}/^{238}\text{U}$ SHRIMP weighted average from 16 of 18 points.</td>
<td>362 +/- 3 Ma</td>
<td>Late Acadian Orogeny?</td>
<td>Aleinikoff et al. (2002)</td>
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<td><strong>Alleghanian Orogeny</strong></td>
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<td>Early or main phase of Alleghanian orogeny</td>
<td>Baddeleyite BMC, Lancaster County, Pa.</td>
<td>TIMS</td>
<td>~310 Ma from lower intercept</td>
<td>Apparent baddeleyite rim growth. R. T. Faill (per. comm., 6/8/07) suspects this corresponds to Alleghanian duplex tectonics that advanced the BMC farther onto Laurentia from its Taconic docking position.</td>
<td>Jahandar Ramezani, M.I.T., analyst.</td>
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<td>Late Alleghanian Orogeny lateral shear</td>
<td>Muscovite from mylonite zone, Peach Bottom Slate, Lancaster County</td>
<td>$^{40}\text{Ar}/^{39}\text{Ar}$ by TIMS</td>
<td>276 +/- 6 Ma</td>
<td>Muscovite crystallization at &gt; 300°C in fault.</td>
<td>Kruger Enterprises, (8/19/1994) in Faill and Smith (1994). [Modern $^{40}\text{Ar}/^{39}\text{Ar}$ plateau analyses needed.]</td>
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<td>Mesozoic Extension and Cooling from Mesozoic Thermal Pulse</td>
<td>Zircons Palisades sill, Bergen County, New Jersey</td>
<td>$^{206}\text{Pb}/^{238}\text{U}$ by TIMS</td>
<td>201.2 +/- 1.3 Ma</td>
<td>Two fractions of best, clear fragments of zircon from Palisades sill.</td>
<td>Dunning and Hodych (1990)</td>
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<td>York Haven Diabase, a.k.a. HTQ</td>
<td>Whole rock diabase, Gainesville East 7 ½’ quadrangle, Culpeper Basin, Fairfax County, Va.</td>
<td>$^{40}\text{Ar}/^{39}\text{Ar}$ weight average plateau</td>
<td>201.2 +/- 1.3 Ma</td>
<td>Median of 3 Ar/Ar for lateral equivalents each from a different area.</td>
<td>Sutter (1988)</td>
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<tr>
<td>Rossville Diabase, a.k.a. LTQ</td>
<td>Zircon, near Wellsville York County, Pa.</td>
<td>$^{206}\text{Pb}/^{238}\text{U}$ by TIMS</td>
<td>201.1 +/- 1.0 Ma</td>
<td>Median for 5 clear zircon fractions.</td>
<td>Dunning and Hodych (1990)</td>
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<td>Zircon</td>
<td>FTA</td>
<td>184 Ma</td>
<td>Cooling to ~ 225°C from Mesozoic Thermal Pulse</td>
<td>Kohn et al. (1993)</td>
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<td>Cooling from MTP</td>
<td>K-spar from granophyre and contact zones, Newark and Culpeper Basins.</td>
<td>$^{40}\text{Ar}$ closure</td>
<td>175 Ma</td>
<td>Cooling to 200°C from Mesozoic Thermal Pulse (Smith and Faill 2000).</td>
<td>Sutter (1988)</td>
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<tr>
<td>Cooling from MTP ???</td>
<td>Apatite, but any data from SE Pa.?</td>
<td>U-Th/He</td>
<td>Would be ~125 Ma if linear cooling from MTP???</td>
<td>Cooling to ~ 65°C.</td>
<td>Frank Pazzaglia and Peter Zeitler (in press?) [Watching for ref.]</td>
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K by ICP-AES.  
Ar TIMS. | 6 +/- 0.5, 12.5 +/- 1.5, 15 +/- 1, 24 +/- 2, 57 +/- 12, and 58 +/- 5 Ma. | Supergene mineral.  
5 of 6 formed during dissolution of Cambrian dolomite.  
THE DELAWARE TRANSECT: PROGRESS AND PROBLEMS.

Howell Bosbyshell, West Chester University of PA

Introduction

High grade metamorphic rock which underlies the Delaware River Basin of Southeastern Pennsylvania has been studied for more than one hundred years. Understanding the relationships between different lithotectonic units in complexly deformed, poorly exposed, high grade metamorphic terrains such as this requires that the full range of geologic tools be brought to bear. The last three decades have seen a dramatic increase in the sophistication of the tools that are available to interpret metamorphic rocks. The use of internally consistent thermodynamic databases and powerful computational techniques enable accurate determination of the pressures and temperatures of metamorphism. New databases of the geochemical composition of modern igneous rocks provide the ability to infer the ancient tectonic settings for igneous precursors of metamorphic rocks. Finally, advances in in situ radiometric dating allow us to determine the absolute age of metamorphic minerals and thus constrain the age of cross-cutting fabric elements and deformation.

This contribution summarizes the current understanding of the bedrock geology which underlies Philadelphia, Chester, and Delaware counties in Southeastern Pennsylvania, with particular emphasis on knowledge that has been gained since the 2004 Field Conference of Pennsylvania Geologists (Blackmer and Srogi, 2004). One important purpose of that conference was to present a provisional subdivision of the Wissahickon Formation prior to publication of a new compilation map (Blackmer, 2005). As defined at that time, the Wissahickon Formation (referred to here as the Type-Wissahickon Formation) was restricted to all pelitic and psammitic gneiss and schist and interlayered amphibolite from the Wilmington Complex east to the Coastal Plain, excluding a narrow band of rock immediately adjacent to the Wilmington Complex. This narrow band, along with all meta-sedimentary rock and interlayered amphibolite south of the Avondale massif, was assigned to the Mt Cuba Wissahickon formation. The Glenarm Wissahickon comprised rock formerly mapped as Wissahickon Formation, which underlies the area from the Avondale massif north to the Embreeville Thrust and underlying Peters Creek Formation.

While the above configuration appears on Blackmer’s (2005) compilation, uncertainty in the location of the Mt. Cuba/Type Wissahickon boundary was discussed (Plank and Schenk, 2004; Blackmer, 2004b) and an alternative contact, corresponding to the western limit of arc-magmatism, was also shown on the guidebook map. This contact is adopted in the map presented here. Uncertainty in the origins of the West Chester and Avondale massif and the associated Glenarm and Mt. Cuba Wissahickon cover rocks prompted Bosbyshell et al. (2007a) to divide the area into three broad regions corresponding to rocks of Laurentian affinity, arc affinity and uncertain affinity (inset on map). These divisions serve as the framework for the outline presented below.

Since the 2004 Field Conference we have gained insight from new geochronological and geochemical data. We have learned from monazite dating that peak metamorphism in the Glenarm Wissahickon at ~410 Ma (Bosbyshell et al., 2007a) is significantly younger than the 425-Ma peak in the Mt. Cuba Wissahickon (Aleinikoff et al., 2006). Additional new geochronological data include a detrital zircon probability spectrum for the Windy Hills Gneiss (Aleinikoff et al., 2006) that may indicate a peri-Gondwanan origin for the Wilmington Complex. New Ar-Ar ages (Blackmer et al., 2007) indicate cooling through Ar closure in muscovite by ~365 Ma for all major units in the area, providing constraints on displacements along major shear zones. New geochemical data demonstrate that amphibolite in the type-section of the Wissahickon Formation is arc-related (Bosbyshell et al., 2009). Collectively, these data lead to a new interpretation of the tectonic history of the region.
involving protracted oblique convergence spanning the upper Ordovician through at least the upper Devonian and likely beyond. The rocks were affected by tectonism described in the Appalachian orogen as the Taconic, Salinic, Acadian, and Alleghanian orogenies.

I. Current state of knowledge

A. “Delaware transect” includes geology underlying Chester, Delaware and Philadelphia counties. Geologic map (fig. 1) – inset shows three broad divisions: Laurentian affinity, uncertain affinity, and arc affinity (Bosbyshell et al., 2007a).

1. Laurentian affinity: includes Mine Ridge, Honey Brook Upland – amphibolite and granulite facies Precambrian gneiss (Bascom and Stose, 1938; Crawford and Hoersch, 1984; Pyle et al., 2006); unconformable Paleozoic cover, including Chickies Fm, Antietam-Harpers and Cambro-Ordovician carbonates; and Octoraro and Peters Creek Formations which Blackmer will discuss.

   a. Mine Ridge and Honey Brook Upland – Neoproterozoic cooling ages for hornblende and biotite (Sutter et al., 1980)

   b. Chickies and Conestoga Formations carbonates contain Acadian- and Alleghanian-aged monazite (Pyle, 2006; Pyle et al., 2006)

2. Rocks of uncertain affinity include:

   a. the Glenarm Group, the Setters Formation and Cockeysville Marble (Drake, 1993); probably Neoproterozoic to Cambrian in age

   b. the Glenarm and Mt. Cuba Wissahickon, probably Neoproterozoic to Cambrian in age (Blackmer, 2004a, 2004c; Plank and Schenk, 2004);

   c. the West Chester massif, Precambrian Gneiss, Grenville-aged granulite facies metamorphism (Grauert et al., 1973, 1974; Wagner and Crawford, 1975) rifted microcontinent (Faill, 1997) or deep Laurentian crust (Wagner and Srogi, 1987);

   d. Avondale massif, historically considered Precambrian Baltimore Gneiss (Bascom et al., 1909; Crawford and Crawford, 1980), but age and genetic relationships are uncertain (Grauert et al., 1974; Bosbyshell et al., 2006; Pyle et al., 2006); amphibolite facies metamorphism.

3. Rocks of arc affinity include:

   a. The Wilmington Complex, Ordovician-aged magmatic arc rocks metamorphosed to upper amphibolite and granulite facies, and Silurian-aged intrusions (Ward, 1959; Wagner and Srogi, 1987; Plank et al., 2000a; Aleinikoff et al., 2006);

   b. The type-Wissahickon Formation, probably lower Paleozoic meta-sedimentary rock, greywacke and shale protolith; generally amphibolite to upper amphibolite facies metamorphism (Bascom et al., 1909; Amenta, 1974; Crawford, 1977; Crawford and Mark, 1982; Bosbyshell et al., 1999a).

B. Major structures

1. Pleasant Grove/Huntington Valley shear zone – right lateral transpressive shear zone (Valentino et al., 1994) which separates Octoraro and Peters Creek formations (west) and Type Wissahickon Formation from Precambrian gneiss of the Trenton Prong (Volkert and Drake, 1993) (east)
2. Embreeville Thrust – base of shallowly SE dipping nappe which places Glenarm Wissahickon and West Chester massif on top of Peters Creek Formation (Wiswall, 1990, 2005; Bosbyshell et al., 2006); boundary between Laurentian and “uncertain rocks”

3. Street Road Fault: shallowly SE dipping ductile thrust shear zone which places Avondale massif on West Chester massif; possible window in Ridley Creek State Park, (Bosbyshell et al., 2006).

4. Rosemont Fault: right-lateral transpressive shear zone (Valentino et al., 1995, 1999; Bosbyshell, 2001) which separates arc-related Wilmington Complex and type-Wissahickon from Avondale massif; trace of this structure is uncertain south of Avondale massif.

C. Geologic history of major units

1. Rocks with arc affinity: Wilmington Complex
   a. geochemistry: suprasubduction zone setting for mafic igneous rock (Plank et al., 2000b)
   b. arc magmatism 475 to 485 Ma; SHRIMP zircon (Aleinikoff et al., 2006)
   c. granulite grade metamorphism and mafic and composite plutonism (Srogi and Lutz, 1997); ~430 Ma SHRIMP zircon (Aleinikoff et al., 2006), also electron probe micro-analyser (EPMA) total Pb monazite and TIMS monazite ages (Bosbyshell et al., 1998; 1999b).
   d. Array of Neoproterozoic detrital zircon ages in meta-volcanic Windy Hills Gneiss (Aleinikoff et al., 2006) could indicate proximity of Gondwanan block. Major peaks in zircon probability spectrum are Grenvillian.
   e. Type-Wissahickon Formation is likely arc-substrate: boninitic dikes intrude Wissahickon Formation near contact (Bosbyshell et al., 1999a; Bosbyshell, 2004) and possible screens of Wissahickon Formation are present along contact in Delaware (could also be tectonic inter-leaving).

2. Rocks with arc affinity: Type-Wissahickon Formation
   a. depositional environment: subduction related basin (Wagner and Srogi, 1987), probably back-arc, based on amphibolite geochemistry – boninitic near Wilmington Complex and BABB interlayers to east, including type-section along Wissahickon Creek (Bosbyshell et al., 2000a, 2000b, 2009)
      i. D2 – isoclinal folding of an older metamorphic foliation (cryptic D1) synchronous with high T metamorphism locally
      ii. D3 – widespread folding; locally developed axial planar foliation; timing suggests that this is onset of deformation related to crustal thickening which led to M3 metamorphism
      iii. D4 – most recent ductile deformation in Rosemont shear zone, localized shearing, synchronous with dominant metamorphism in much of Wissahickon Formation; over printing of older high T metamorphism locally
      iv. D5 – widespread crenulation cleavage development and associated folding
c. metamorphism
   i. M1 – cryptic contact metamorphism associated with Wilmington Complex magmatism; main evidence is 480 Ma monazite growth (Bosbyshell et al., 1999a; Bosbyshell, 2001)
   ii. M2: early high T – low P (0.4 to 0.5 GPa); gradient in M2 metamorphism from very high T (~700°C) nearest Wilmington Complex to apparently very low grade along Wissahickon Creek (Bosbyshell et al, 1999b, 2007b)
   iii. overprinting by M3 higher P, ~ 0.7 GPa (Crawford and Mark, 1982; Bosbyshell et al., 1999). Barrovian sequence along Wissahickon Creek is M3.

d. geochronology: preliminary EPMA total Pb monazite ages M1: 480 Ma, same as Wilmington Complex magmatism; M2: Silurian, 440 – 430 Ma; D3 deformation ~416 Ma; M3: Devonian, 380 – 370 Ma (Bosbyshell 2001; Pyle et al., 2006; Bosbyshell 2008)

3. Uncertain affinity: the Glenarm Wissahickon
   a. interpreted to be in thrust contact (Alcock, 1994; Alcock and Wagner, 1995) with Glenarm Group, the Setters Formation and Cockeysville Marble (Drake, 1993), but Blackmer (2004a) removed thrust and interpreted this contact as depositional
   b. locally complex deformation (Alcock, 1994; Blackmer, 2004a, 2004c), but dominant foliation is shallow to moderately dipping S2, thrust related fabric.
      i. Given the wide geographical extent of this fabric its formation is likely diachronous, rather than the result of an “event.”
   c. metamorphism: peak metamorphic conditions exceeded staurolite stability, local partial melting, followed by isobaric or slight increase in pressure during cooling, based on presence of texturally late kyanite and thermobarometry using rim compositions of garnet and plagioclase (Bukeavich et al., 2006; Srogi et al., 2007)
      i. staurolite breakdown/garnet growth is latest syn- to post-S2 formation (Bukeavich et al., 2006; Srogi et al., 2007)
   d. geochronology:
      i. excellent control on timing of peak metamorphism using EPMA total Pb monazite analyses and detailed examination of metamorphic textures (Bosbyshell et al., 2007a) max T at 410 Ma;
      ii. Glenarm samples also yield monazite ages of approximately 454 Ma (Bosbyshell, unpublished data) Taconic metamorphism? Monazite of this age has not been identified in the Type-Wissahickon or Mt. Cuba.

4. Uncertain affinity: Mt. Cuba Wissahickon: West of Wilmington Complex; south of Avondale massif
   i. depositional setting: detrital zircon ages indicate Grenvillian source (Aleinikoff et al, 2006); generally more psammitic than Glenarm Wissahickon; geochemistry of interlayered White Clay Creek amphibolite suggests continental rift setting (Smith and Barnes, 2004); though Plank et al. (2000b) interpreted this geochemistry to indicate an ocean island origin for the basaltic precursor.
ii. deformation; dominated by southeast dipping S2, but detailed structural analysis is needed

iii. metamorphism: upper amphibolite facies; partial melting

iv. geochronology: monazite growth at 425 Ma, likely the time of high T metamorphism (Aleinikoff et al., 2006)

II. Tectonic timeline

A. Late Mesoproterozoic to early Neoproterozoic: Granulite facies metamorphism of West Chester massif in Grenville orogeny

B. Neoproterozoic through Cambrian: Deposition of Glenarm and Mt. Cuba Wissahickon, probably along rifted margin of Laurentia (Plank and Schenk, 2004; Blackmer, 2004)

C. Early Ordovician: formation of Wilmington Complex arc ~480 Ma;
   1. Minor 480 Ma monazite in Wissahickon Formation near the Wilmington Complex contact indicates some degree of metamorphism of Wissahickon Formation at this time (M1), though relic silicate porphyroblasts of this generation have not been identified.

D. Late Ordovician: 450 Ma monazite cores in Glenarm Wissahickon samples (Bosbyshell et al., 2007) indicate some burial and heating during Taconic Orogeny.

Silurian through Devonian is a period of protracted oblique convergence, which successively buried blocks or sheets containing the Wilmington Complex and adjacent Type-Wissahickon; Mt. Cuba Wissahickon and Avondale massif; Glenarm Wissahickon and West Chester massif; and ultimately reburied the Type Wissahickon and Wilmington Complex.

E. Early Silurian: Burial of Early Ordovician volcanic rocks in the Wilmington Complex must be complete by this time. Some of this burial could be through sedimentation and younger arc volcanism, but pressure of Silurian metamorphism, ~ 0.5 GPa or depth of 15-20 km, indicates tectonic burial. Is this the Taconic or Salinic orogeny (see Problems and future research, below)?

F. 435 – 425 Ma (Bosbyshell et al., 2001; Aleinikoff et al, 2006): High T, low to moderate P metamorphism at of Wilmington Complex and adjacent Type-Wissahickon.
   1. Metamorphic conditions require high heat flow; mafic magmatism suggests mantle involvement; geochemistry of mafic rock suggests back-arc setting (Plank et al., 2000b).
   2. Metamorphism is the result of an extensional environment – back-arc rifting or slab delamination (Bosbyshell 2001; Srogi, 2004)
   3. Emplacement of still-hot arc terrane on Mt. Cuba Wissahickon helping to drive metamorphism in those rocks at that time?

G. Early Devonian – the onset of Acadian deformation and metamorphism
   1. 416 Ma: D3 deformation in Type-Wissahickon Formation; onset of convergence that ultimately resulted in crustal thickening and D3 metamorphism in Type-Wissahickon
   2. 410 Ma: Formation of pre- to syn-metamorphic S2 fabric (Bukeavich et al., 2006) and attainment of maximum temperature in Glenarm Wissahickon (Bosbyshell et al., 2007a)
   3. Are deformation and metamorphism in the Glenarm Wissahickon a response to burial by still warm Mt. Cuba Wissahickon on Street Road Fault?
H. Late Devonian – 380 Ma-375Ma: M3 metamorphism in Type-Wissahickon, ~7 kb pressure,
25 – 30 km of crust on top of present exposure level – the Acadian mountains in
Pennsylvania.

1. Fabrics in Rosemont shear zone indicate some right-lateral translation at this time, but
extent of movement is limited because Catskill Delta isopachs and paleocurrent
indicators suggest that source of these sediments is approximately at the latitude of the
Type-Wissahickon (Harper, 1999).

I. All rocks described above cooled through Ar closure in muscovite (350C) at 365 Ma
(Blackmer et al., 2007).

1. This result precludes significant relative vertical displacement of different crustal
blocks subsequent to 365 Ma. However, Alleghanian strike-slip deformation is possible.
2. Relatively rapid cooling from Acadian metamorphism for some rocks, notably the
Type-Wissahickon.

III. Problems and future research

A. Origins of basement rocks.

1. West Chester massif – almost certainly Laurentian, but is this rifted micro-continent
(Faill, 1997) or the deep edge of the craton (Wagner and Srogi, 1987)?
2. Avondale massif – Grauert et al. (1974) thought this was distinct from West Chester
massif; recent preliminary work (Bosbyshell et al., 2006; Pyle et al., 2006) found no
Precambrian monazite, in fact most robust monazite population is same age as
Wilmington Complex magmatism (~480 Ma) – is the Avondale Massif Wilmington
Complex arc basement?

B. Origins of Mt. Cuba Wissahickon Formation, Glenarm Wissahickon, and Glenarm Group.

1. Both Mt Cuba and Type Wissahickon appear to be in depositional contact with
Glenarm Group, but seem to have distinctly different times of peak metamorphism.
Planned detrital zircon provenance study, structural analysis in Mt. Cuba Wissahickon
and additional monazite geochronology will help to resolve this problem.

C. Latest Ordovician and Silurian tectonics – is burial of the Wilmington Complex Salinic or
Taconic?

1. If Wilmington Complex is peri-Gondwanan (as suggested by detrital zircon data) this
tectonism is probably not the Taconic Orogeny; the arc would have been on the wrong
side of Iapetus Ocean at that time.
2. Possible sequence of events and tectonic mechanisms in the Salinic orogeny:
   a. burial of arc as a result of collision with Laurentia ahead of Peri-Gondwanan
      (Ganderia or Carolinia?) block (Hibbard et al., in press)
   b. magmatism and metamorphism as a result of back-arc rifting above west-dipping
      subduction zone as Avalon approaches
   c. metamorphism of Mt. Cuba Wissahickon as a result of emplacement of still warm
      arc.
      i. lack of intrusive rocks in the Mt. Cuba suggests that arc terrane arrived after
         cessation of magmatism
D. Metamorphic history of basement rocks. Metamorphic textures in the Avondale massif, plagioclase haloes around garnet, indicate rapid isothermal decompression (Johnson and Bosbyshell, 2010), consistent with cooling history indicated by Ar-Ar results (Blackmer et al., 2007). Interestingly, rocks in the West Chester massif exhibit the opposite texture, garnet haloes on plagioclase, indicating a sharp increase in pressure (Wagner and Crawford, 1975; Wagner and Srogi, 1987). Solution is in the timing, which is uncertain.

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References


**SUSQUEHANNA TRANSECT**

**Gale C. Blackmer,** Pennsylvania Geological Survey

It is my task to introduce you to the Susquehanna Transect through the Piedmont, share some current thoughts about this area, and attempt to tie together the three transects that have been presented to you today. For the purposes of this presentation, “Susquehanna Transect” refers to southern Lancaster County and parts of Chester County.

Wise (1970) and others (see summary in Wise, this volume) first addressed the geologic structures of this area in a modern sense. Recent mapping by the Pennsylvania Geological Survey and cooperators, largely under the STATEMAP portion of the USGS National Cooperative Geologic Mapping Program, essentially supports their big structural picture, so I will not dwell on that in this presentation. You can view recent work in the area on these maps: Blackmer, 2007; Bosbyshell, 2007; Hill, 2007; Hill, 2008; Blackmer and others, 2010; Hill and others, in prep.

Many interesting advances over the last 50 years have come in our understanding of the origin and assembly of the various tectonic elements that constitute this part of the Piedmont. That is what I will concentrate on here.

I. Tectonic Elements of the Susquehanna Transect (figure 1)


B. Laurentian shelf: basal clastic rocks (Cambrian) – Chickies, Antietam-Harpers; carbonates (Cambrian into Ordovician) – Vintage through Conestoga.

C. Octoraro Formation (Westminster Terrane) – quartzose schist (Cambrian).

D. Peters Creek Schist (Potomac Terrane) – metasandstone, metaconglomerate, and schist (Cambrian).

1. Peach Bottom Slate and Cardiff Conglomerate

E. Various kinds of Wissahickon and Baltimore Gneiss internal basement massifs – see Bosbyshell (this volume). These are geographically off of the transect but still play a part in understanding the tectonics.

F. Baltimore Mafic Complex – arc-related stratified mafic complex, which includes mafic, ultramafic, and felsic rocks (early Ordovician).

G. Martic Fault – traditionally, the boundary between Laurentian shelf and Octoraro (we’ll talk more about this).

H. Pleasant Grove-Huntingdon Valley shear zone – boundary between Octoraro and Peters Creek; significant high-strain zone.

II. The “old” stories

A. Knopf and Jonas (1929) defined the Martic overthrust as a fault that carried Precambrian schists over Paleozoic sediments. In large part, age of the rocks was determined by degree of metamorphism – rocks south of the Martic fault are metamorphic, rocks north of the fault are not. In 1929, metamorphic rock was equivalent to Precambrian rock.

B. The Baltimore Gneiss massifs east of this transect represent a microcontinent that was rifted from Laurentia during the opening of Iapetus in the Late Proterozoic. A narrow ocean basin called the Octoraro Sea lay between the microcontinent and Laurentia. Sediments that now constitute the Peters Creek Schist were shed from the microcontinent into the east side of the Octoraro Sea. The Octoraro Formation represents deep water deposits in the sea. This depositional basin was active from the early Cambrian until it was closed and overrun during the Ordovician Taconic Orogeny (Faill, 1997).

Figure 1. Major features of the Susquehanna transect in Pennsylvania. The area north of the Martic Fault is known to be Laurentian platform (unlabeled unit shown in blue is undivided Cambrian carbonates). Extending the Antietam-Harpers Formations south of the Martic Fault also extends the platform and changes the significance of the Martic Fault. All the contacts on the map have some level of uncertainty. The contact of the Antietam-Harpers and Octoraro is particularly uncertain. Red lines are faults. The Westminster Terrane lies between the Martic Fault and the PGHV shear zone. The Potomac is roughly the extent of the Peters Creek Schist. PGHV = Pleasant Grove-Huntingdon Valley.

III. New thoughts – Antietam-Harpers Formations, Octoraro Formation, Martic Fault
A. At their type localities near Harpers Ferry, West Virginia, the Harpers and Antietam Formations are separable into primarily meta-siltstone and primarily meta-sandstone, respectively. In the Susquehanna area, the rocks occupying the same interval above the Cambrian quartzites are fairly uniform interlayered more and less quartzose schists. These have traditionally been mapped as Antietam-Harpers undivided.
B. In the Knopf and Jonas (1929) picture, the rocks north of the Martic Fault were not metamorphosed and therefore differentiated from the Precambrian rocks south of the Martic Fault. But there is garnet in Antietam-Harpers at the west end of Mine Ridge.
C. On the east side of the Susquehanna River, the lithologies in the Antietam-Harpers north of the Martic Fault are indistinguishable from the Octoraro south of the Martic Fault (including...
the presence of albite porphyroblasts, which are the hallmark of the Octoraro Formation). On my map of the Conestoga and Quarryville quadrangles (Blackmer, 2007), I mapped the rocks on both sides of the Martic Fault as the same unit and called them Antietam-Harpers. I did draw a contact, which left a band of Octoraro against the Peters Creek, but that was done on fairly weak grounds and it may not be correct.

D. If the rocks on both sides of the Martic Fault really are the same unit, then the fault is “demoted” from a major suture to just another fault in the complex of duplex structures and imbricate fans that dominate this area (see Blackmer, 2007).

E. Detrital zircon data support this conclusion. Four samples – one from Antietam-Harpers north of the Martic Fault and three from Antietam-Harpers and Octoraro south of the Martic Fault – have similar zircon age distributions: a large peak of Grenvillian-age zircons, a few Archean (ca. 2650 Ma) zircons, and a few 530 Ma zircons.

1. Similar age distributions suggest that all the sediments had a similar source, and the simplest assumption is that they were deposited in the same basin.

2. The maximum age of the rocks is early Cambrian.

F. The Octoraro lithology can be followed east along its outcrop belt, although it becomes increasingly phyllonitic and exhibits textures indicative of increasing strain. This is especially true after crossing into Chester County, in the narrowest part of the belt.

F. The Octoraro Formation and/or Antietam-Harpers in this transect, which occupy the same space between the Martic Fault and the Pleasant Grove shear zone as the Westminster Terrane, do not correlate lithologically or geochronologically with the Westminster Terrane in Maryland (see Southworth, this volume). Different detrital zircon populations suggest different sources for the Maryland and Pennsylvania sediments (Martin and others, 2010).

IV. New thoughts – Peters Creek Schist

A. The Peters Creek is full of detrital quartz and, to a lesser extent, detrital feldspar. It clearly has a nearby continental source and is a more sandy than muddy sediment. Faill (1997) has it deposited against a microcontinent, across a narrow ocean basin from Laurentia, over a time range extending from late Cambrian to middle Ordovician. In contrast, I think it could be the oldest sediment on the Laurentian margin, deposited in the Iapetan rift.

1. One of the lithologies in the Peters Creek is a biotite-epidote schist. Epidote is locally common enough to make the rock nearly black. A possible protolith for this rock would be volcaniclastic material falling into the sandy sediments.

2. Thin greenstone layers scattered throughout the Peters Creek have the chemistry of ocean-floor basalts (Bald Friar metabasalt; Smith and Barnes, 1994, 2004). A rift environment would accommodate coarse, sandy sediment and bimodal volcanism, represented now by greenstone layers and more acidic volcaniclastic material.

3. Gates and Valentino (1991) and Valentino and Gates (1995) classified the Peters Creek as a rift sediment based on sandstone petrography. They postulated a Late Proterozoic age based on the lack of passive margin detritus (e.g., carbonate clasts).

B. Detrital zircon ages from one sample of Peters Creek metasandstone, collected in eastern Lancaster County near the northern boundary of the map extent, are similar to those of the Antietam-Harpers-Octoraro samples – a major Grenvillian age peak, an Archean peak, and one 530 Ma zircon. The Archean peak is larger in the Peters Creek, comprising 10% of the sample. This means that the Peters Creek sediment had greater input from an Archean source.

1. The maximum age of the Peters Creek is 530 Ma (early Cambrian).

2. Does the detrital zircon age distribution mean that the Peters Creek was deposited in the same basin as the Antietam-Harpers-Octoraro, and therefore did not move far along the Pleasant Grove-Huntingdon Valley shear zone; OR
3. Does the greater Archean zircon signature indicate that the Peters Creek was farther north, closer to Archean sources in the interior of the continent, and moved significantly southward on the Pleasant Grove-Huntingdon Valley shear zone.

C. Another interesting stratigraphic issue related to the Peters Creek is the Peach Bottom Slate-Cardiff Conglomerate interval. Although there is currently general agreement that this interval should not be interpreted as the core of a syncline, there is disagreement about what it does represent.

1. My mapping in the Kirkwood quadrangle revealed a belt of coarse, pebbly metasandstone in the northern part of the Peters Creek that encompasses the Cardiff Conglomerate and extends at least all the way across the quadrangle. This makes me suspect that the Cardiff is really a conglomeratic facies of the Peters Creek. Although the Peach Bottom undoubtedly started as a clay-rich mud, microtextures show that it has also absorbed significant strain. It is to be expected that strain would partition into the micaceous rock rather than the coarse metasandstone, creating an intraformational shear zone. I envision a depositional relationship between the quartz sand and mud, perhaps a near-shore delta-like environment where channels alternate with quiet, lagoonal areas.

2. For a different interpretation of this area, see Faill and Smith (2010).

D. A number of workers have recognized diamictite within the area between the Pleasant Grove-Huntingdon Valley shear zone. This has led some to correlate the Peters Creek, or parts of it, with the Sykesville Formation of Maryland. Recently, Hill and others (in prep) mapped a relatively small area of diamictite in contact with the Baltimore Mafic Complex (figure 1). They postulate that the diamictite was either carried with the Complex on its basal fault or is related to its movement and emplacement. In contrast, Faill and Smith (2010) mapped a large part of the area identified as Peters Creek in figure 1 as diamictite and called it Sykesville.

D. Moving east from the Susquehanna transect into Chester County, the outcrop belt of Peters Creek narrows. The pebbly, conglomeratic lithology disappears. The biotite-epidote schist becomes more prevalent, interlayered with quartzose schist containing detrital grains and rock fragments.

E. The Peters Creek in this transect occupies the same position south of the Pleasant Grove shear zone as the Potomac Terrane in Maryland. There is some lithologic similarity between the Peters Creek and the Potomac Terrane, but there is not a direct correlation. The Potomac Terrane in Maryland is divided into several fault-bounded blocks. Similar faults have not (yet?) been identified in Pennsylvania. The detrital zircon population from the northwestern part of the Peters Creek outcrop belt is similar to detrital zircon populations from the southeastern Potomac Terrane. The northwestern Potomac Terrane is different (Martin and others, 2010).

V. What we still don’t know

A. Are there faults or contacts within what is currently mapped as Peters Creek?

B. Do we have the Antietam-Harpers/Octoraro division correctly mapped?

C. What is the source of the Archean zircons in the clastic rocks in the Susquehanna transect?

Archean zircons might come from several sources: 1) an interior continental source (must account for absence of Archean zircons in Laurentian margin sediments in Maryland); 2) a Gondwanan source across a narrow rift basin; 3) Gondwana-sourced sediments deposited on Laurentia when the continents were together, left behind at rifting, and subsequently eroded and deposited in the new basin.

D. What is the source of the 530 Ma zircons? (Note: these are about 30 m.y. younger than the Catoctin volcanics).

E. Many other questions of correlation of units along strike, depositional settings, and timing of assembly. We have just enough data to raise more questions than it answers.
VI. What we know (but maybe don’t yet understand) about origins and assembly

A. None of the clastic rocks that have been sampled for detrital zircons along the Susquehanna transect received material from an Ordovician arc, which means they were buried prior to arc development (more likely) or they were receiving sediment through the Ordovician but were not in proximity to an arc (less likely).

1. We know the detrital zircon signature of one Ordovician arc, the Wilmington Complex, from a sample of the Faulkland Gneiss, which consists of interlayered metavolcanic and metasedimentary rocks. It includes Ordovician zircons, ca. 474 Ma (Aleinikoff and others, 2006).

2. Given that there are Ordovician bentonites of similar age throughout the Valley and Ridge and further into the interior of the continent, and bentonites in Ordovician carbonates in the Great Valley, it is fair to assume that the Antietam-Harpers, Octoraro, and Peters Creek would show some evidence of Ordovician input had they been receiving sediment at that time.

3. Therefore, it appears that these metasediments are older than late Early Ordovician.

4. This permits, but does not confirm, a rift-early Laurentian margin origin for the clastic units. It makes more difficult, but does not preclude, Faill’s (1997) Octoraro Sea model.

B. There is garnet in the Peters Creek only close to the contact with the Glenarm Wissahickon, along the entire exposed length of that contact (Embreeville fault). This suggests that the Wissahickon was warmer than the Peters Creek when the two were juxtaposed, and the extra heat from the hanging wall raised the temperature of the footwall sufficiently to grow garnet. I’m not sure what that means in terms of the story, but it’s all I can say right now about the relationship between the Wissahickon and Peters Creek.

C. Monazite ages from the Pequea Silver Mine (Conestoga Formation, north of the Martic Line; Wise and Ganis, 2009) and from the Glenarm Wissahickon (Bosbyshell and others, 2007) indicate metamorphism at about 450 Ma. Wise and Ganis (2009) interpret this age as representing a mature mountain system built on what had been the shelf edge. The Glenarm Wissahickon monazites show additional metamorphism and fabric development at about 400-410 Ma (Devonian). The Pequea monazites do not show this overprint. Our attempts to find monazite in the Octoraro Formation were unsuccessful. We haven’t looked yet in the Peters Creek.

D. The map pattern of the Baltimore Mafic Complex (BMC) in Pennsylvania is incompatible with all of the other structures in the Piedmont. Recent mapping (Hill and others, in prep) shows that the contact of the northwestern “finger” of the BMC with the Peters Creek Schist is subhorizontal. The main body of the BMC has a higher-dip contact with the schist units, but cuts close to perpendicular across the Embreeville fault. All of this suggests that the Peters Creek and Glenarm Wissahickon were assembled prior to emplacement of the BMC over them on a thrust fault. That would make the emplacement of the BMC Devonian at the oldest (Bosbyshell, this volume).

E. The detrital zircon signatures of known Laurentian margin and Westminster Terrane rocks in Maryland are different from [what we assume to be] equivalent rocks in Lancaster County, Pennsylvania.

F. $^{40}\text{Ar}/^{39}\text{Ar}$ muscovite ages for the Westminster and Potomac terranes in Maryland and Virginia show faults between and within the terranes with a long history of motion from the Devonian through the Pennsylvanian (Kunk and others, 2004). In Pennsylvania, the pattern is quite different. Muscovite cooling ages from across the entire Piedmont in Delaware and Chester Counties, including all subdivisions of the Wissahickon Formation, Avondale Massif, Peters Creek Schist, and Octoraro Formation have cooling ages of about 365 Ma (Devonian) (Blackmer and others, 2007). Although these ages are comparable with the Potomac Terrane
in Maryland and Virginia, there is no evidence of jostling of blocks within the terrane as we see in the south.

VII. Wild speculations

A. There appears to be a fundamental break in sedimentation and tectonism on the Laurentian margin somewhere in York County, between Maryland and the east side of the Susquehanna River. Might there be a transform fault somewhere near the Susquehanna?

B. The evidence we have suggests that the Peters Creek and Octoraro Formations are Cambrian deposits, attached in some way to the Laurentian margin, that were already buried when Ordovician tectonism began.

C. Although the Baltimore Gneiss massifs may represent a microcontinent, the Octoraro Sea model for the deposition of the Peters Creek and Octoraro/Antietam-Harpers is not supported by geochronologic data.

D. The metamorphic and structural history of the Glenarm Wissahickon suggest that was not juxtaposed against the Peters Creek until Devonian time. Although the Glenarm Wissahickon shows an Ordovician metamorphism appropriate to roots of “Taconic alps,” perhaps those alps were not our Lancaster Platform alps.

References


Introduction

The purpose of this paper, prepared for a symposium held in advance of the 2010 Field Conference of Pennsylvania Geologists, is to provide a concise summary of progress made over the past two decades in unraveling the complex depositional history recorded in lower Paleozoic outer platform and deep-water carbonates of the Conestoga Valley and adjacent areas (Figure 1). This year's conference marks the 20th anniversary of the 1990 Field Conference, where a series of stops in the York and Lancaster areas allowed the trip leaders to introduce a new, highly refined lithostratigraphy (Ganis and Hopkins, 1990) and a derivative depositional model (Taylor and Durika, 1990) for Cambrian carbonates in the Conestoga Valley. One goal of the 2010 symposium and field trip is to review those stratigraphic interpretations and see which have stood the test of time and critical scrutiny with new data. Much of what is presented here has been distilled from a comprehensive paper on the lower Paleozoic strata of the central Appalachians (Brezinski et al., In Press) soon to be published as part of an Association of Petroleum Geologists (AAPG) Memoir on carbonate deposits of the "Great American Carbonate Bank" preserved throughout Laurentian North America.

Figure 1. Areas of exposed lower Paleozoic carbonates in the central Appalachians referred to in the text. Nittany Arch and Great Valley successions represent inner shelf and outer platform environments, respectively; the Conestoga and Frederick Valley strata, exposed within the Piedmont, also include shelfbreak and off-platform, deep marine facies.
Figure 2. Stratigraphic model proposed by Taylor and Durika (1990), depicting the Emigsville (EM) and Greenmount (GM) members of the Kinzers Formation as laterally continuous transgressive tongues of basinal facies, a diachronous Kinzers-Ledger formation contact attributed to progradation of the shelf margin, and a Willis Run (WR) limestone Member in the Ledger Formation separating informal upper and lower dolomite members (LD and UD).

The 1990 Stratigraphic Model

The refined, member-level lithostratigraphy and age correlations proposed in 1990 (Figure 2) were based on extensive new lithologic and biostratigraphic information recovered from outcrop and cores in the Kinzers and Ledger formations, primarily in the York area, where these units are nearly an order of magnitude thicker than they are in and around Lancaster. The new data supported the interpretation made earlier by Rodgers (1968), Gohn (1976), and Read (1989) that the Kinzers and Conestoga Formations formed in deep, off-platform environments seaward of a shelf rimmed by carbonate sand shoals that produced the dolomitized grainstone of the Ledger Formation. Recognition of what appeared to be lithologically distinctive and laterally persistent intervals at the top of the Kinzers (the Greenmount Member = GM) and in the bottom third of the Ledger Formation (the Willis Run Member = WR) improved correlation along strike between quarries within the West York Block, north of the Gnatstown Fault (Ganis and Hopkins, 1990). Trilobites recovered from both of these new members in a continuously exposed succession in the quarry (then owned by Delta Carbonate) along Roosevelt Avenue proved to be Early Cambrian in age. This was a surprising development given an earlier report (Campbell, 1977) of Middle Cambrian trilobites in the highest
strata of the Kinzers in the Lancaster area. A simple Waltherian model, involving shelf-margin progradation through the latest Early Cambrian and into earliest Middle Cambrian time, was proposed to reconcile the age difference of basal Ledger strata in the York and Lancaster areas (Taylor and Durika, 1990).

1) The Greenmount Member (GM), like the Emigsville Member (EM) at the base of the formation, was interpreted as a laterally continuous transgressive tongue of basinal facies deposited very late in the Early Cambrian. The onlap of more distal, clastic-rich facies onto the toe-of-slope apron of pure carbonate (the York Member = YM) was attributed to a rise in sea level that inhibited downslope transport of carbonate sediment from the shelf margin. The Greenmount Member, whose high siliciclastic silt content makes it slightly more resistant than the adjoining pure carbonates, apparently was the same unit mapped by Stose and Jonas (1939) as an "Upper Kinzers Sandstone", which created a series of low ridges in the York area.

2) The Willis Run Member (WR) was established when an interval of limestone was found low within the "Ledger Dolomite" both in the quarry on Roosevelt Avenue, and in the subsurface during core drilling adjacent to the Baker dolomite quarry. The Lower Cambrian trilobites and brachiopods recovered from the member at the former locality were the first fossils ever recovered from this formation. The fossil-bearing Willis Run in the Roosevelt Avenue Quarry was also unusual in consisting primarily of burrowed lime mudstone to wackestone, whose mud-supported fabric reflected conditions far less energetic than those that created the typical, well-winnowed oolitic grainstone that was dolomitized to produce the bulk of the Ledger Formation. The lithologies observed in the cores and a few excavated blocks from the Willis Run along Berlin Road near the Baker quarry, however, displayed grain-supported fabrics more typical of the Ledger. They also contained some very large marine-cement-lined cavities that prompted us to predict on the field trip that continued excavation of the area underlain by limestone would produce evidence of shelf-edge microbial reefs.

3) The Conestoga Formation, which overlies the Ledger Formation in both the York and Lancaster areas, comprises limestone-shale rhythmite and limestone conglomerate/breccia that clearly originated by down-slope transport of carbonate sands and gravels from a high-relief shelf margin. The emplacement of the deep-water carbonate turbidites, grainflows, and debris flows directly atop the shelfbreak grainstone of the Ledger must have involved down-faulting and a major back-step of the shelf margin. Trilobites from very low in the Conestoga near Baker Road (the first mega-fossils recovered from strata unquestionably assignable to this formation) dated that event as occurring during the Middle Cambrian.

Subsequent advances in stratigraphic insight (delusion?)

Our understanding of the environments, processes, and events recorded in the Piedmont carbonate succession has improved for a number of reasons. New lithologic and faunal data, recovered as quarrying operations continued in the Conestoga Valley and adjacent areas, identified some critical errors in age assignment and miscorrelations of some key exposures and units. Detailed geologic mapping and refinement of the lithostratigraphy of age-equivalent strata in the Ridge and Valley Province also have yielded valuable insight to stratigraphic and structural relationships essential to an accurate interpretation of the Piedmont carbonates. Figure 3 is a revised stratigraphic model for the Lower and Middle Cambrian carbonates of the Conestoga Valley, with the equivalent formations in the eastern Great Valley added on to the cross-section. The specific changes (e.g. the fate of the "Greenmount" and "Willis Run" members) are described briefly in the sections that follow.
On a broader scale, progress in subdivision and correlation of Cambrian-Ordovician units throughout North America has paid dividends in identifying events of inter-regional scale that left their mark in the central Appalachian stratigraphic succession, as elsewhere. We have also benefited immensely from the emergence of new and powerful tools such as carbonate sequence stratigraphic analysis and stable isotope stratigraphy. All the new data and techniques have produced a more refined chronostratigraphic and lithostratigraphic framework for the lower Paleozoic stratigraphy of the central Appalachians (Figure 4). In some cases, the information archived in central Appalachian stratigraphic units played a pivotal role in the recognition of these widespread events and provided the key to accurate interpretation of the sedimentary record of other sedimentary basins. For example, the Stonehenge Transgression (Taylor et al., 1992) is the name now widely used (Miller et al., 2004; Runkel et al., 2008; Brezinski et al., In Press) for a major transgression that occurred in the earliest Ordovician, affecting deposition in sedimentary basins throughout North America and on other paleo continents (Nielsen, 2004).

**Figure 3. A new stratigraphic model for Lower and Middle Cambrian carbonates in the eastern Great Valley (Hagerstown Valley) and Conestoga Valley in southern Pennsylvania.** Carbonate ramp deposits shown in green. Cyclic shelf-margin facies in orange, shelfbreak facies in blue, and proximal and distal off-shelf deposits in light and dark purple, respectively. Shale-dominated units in black. (Modified from Brezinski et al., In Press)

**Piedmont carbonates in the central Appalachians: the 2010 perspective:**

The next two sections summarize what we presently know about the depositional history of Cambrian and Lower Ordovician (Sauk Megasequence), passive margin carbonates preserved in the central Appalachians Piedmont. The active margin, Middle and Upper Ordovician (Tippecanoe Megasequence) deposits are covered elsewhere in this volume. The first section below details the differences between the 1990 and 2010 stratigraphic models of the Conestoga Valley (Figures 2 and 3), and the rationale for those changes. The second section is an outline summary of the depositional
history of the Sauk Megasequence carbonates in this region. This includes a summary of recent discoveries in the Frederick Valley succession in Maryland, where considerable progress in linking the events recorded in the shallow marine carbonates to their off-platform equivalents is described to illustrate how new data and techniques are improving our understanding of even these poorly exposed, sparsely fossiliferous, and deformed successions. The hope is that similar progress can be made in the near future in one of the most intriguing, yet poorly known units in the Piedmont, the Conestoga Formation.

Figure 4. Chronostratigraphic units (systems, series, stages) and supersequences currently recognized in the lower Paleozoic of Laurentian North America, along with the lithostratigraphic units (groups and formations) currently in use in areas of the central Appalachians. Vertically ruled areas represent stratigraphic gaps. (Modified from Brezinski et al., In Press.)
The new Conestoga Valley stratigraphic model differs (aside from being much more colorful) from its predecessor in a number of significant respects, two of the more conspicuous being the member stratigraphy of the Kinzers and Ledger formations, and the less profound diachroneity of the contact between these formations. The addition of the eastern Great Valley succession is also noteworthy, and is discussed piecemeal (unit by unit) in the outline summary of events below.

A. The "Greenmount Member" - Additional trilobite collections recovered in the early 1990's from exposures of the impure, siliciclastic-rich strata mapped as "Upper Kinzers Sandstone" (Stose and Jonas, 1936) or Greenmount Member (Ganis and Hopkins, 1990) at various locations in the York area established that these outcrops do not represent a single, Early Cambrian transgressive event (Taylor et al., 1997). The fauna recovered from at least one locality assigns those strata to the basal Middle Cambrian Poliella-Plagiura Zone (Figure 5). This age variation refuted the posited existence of a single, impure carbonate package at the very top of the Kinzers Formation. Consequently, the Greenmount Member was abandoned and the revised stratigraphic model (Figure 3) shows multiple incursions of impure, upper slope "Greenmount" facies onto the pure York Member carbonate fan during the latest Early and earliest Middle Cambrian.

B. The "Willis Run Member" - The discovery of unquestionably younger (Middle Cambrian) strata high in the Kinzers Formation in the York area prompted reevaluation of the interval assigned to the Ledger Formation in the Roosevelt Avenue Quarry. The limestone package with Lower Cambrian fossils (including the trilobite Zacanthopsis, as shown in Figure 5), which was correlated with the limestone interval in the Ledger along Baker Road to form the basis of a laterally continuous "Willis Run Member", is now recognized to be part of the Kinzers Formation. The Roosevelt Avenue quarry succession, it turns out, was not unusual in preserving an interval of limestone within the Ledger Formation; rather, it is odd in containing an interval of Ledger-like dolomite in the upper part of the Kinzers Formation.

C. The "Willis Run facies" - Continued excavation of the limestone interval within the Ledger Formation near the Baker quarry did yield the anticipated evidence of shelfbreak microbial reefs. In fact, what emerged was a spectacular, previously undescribed shelf-margin facies created at least in part by microbially-induced syndeposition al cementation of portions of the shelf-edge carbonate sands. This is the interval shown as "fenestral grainstone" low in the Ledger in the central column of Figure 3. First reported (a bit tongue-in-cheek) as "Pennsylvania's Great Barrier Reef" (Taylor and Kraewic, 1993), this remarkable shelfbreak facies has been described much more thoroughly, and its extraordinary attributes detailed, in a series of excellent papers by the faculty and students at Franklin and Marshall College (de Wet et al., 1999, 2004, In Press). Among the many intriguing discoveries reported in those papers was evidence from the geochemistry of the syndepositional cements that the deposit is indeed Middle Cambrian in age. Subsequent recovery of a few trilobites (Taylor et al., 1997), including Kootenia cf havasuensis and Glossopleura sp. allow assignment to the lower Middle Cambrian Glossopleura Zone, confirming that age determination. Consequently, there is little (if any) evidence of diachroneity for the base of the Ledger Formation across the Conestoga Valley.
Figure 5. Diagram showing the stratigraphic distribution and zonal assignment of trilobite collections and key genera recovered from the Conestoga Valley carbonate succession through the 1990's, necessitating revision of the lithostratigraphy.

2) The historical sequence of events in the evolution of the central Appalachian carbonate platform is provided in outline form below. It began with the development of a carbonate ramp (lower Tomstown and Vintage Formations), which evolved into a highrelief, narrow rimmed shelf (upper Tomstown and Waynesboro Formation; western Kinzers and Ledger Formations), and finally a rimmed shelf that was bordered on its seaward margin with a broad complex of peritidal environments from the Middle Cambrian through Early Ordovician (highest Tomstown Formation; Elbrook and Conococheague Formations; Beekmantown Group). The deep-water, off-platform deposits that accumulated seaward of the shelf now compose the shale-rich members of the eastern Kinzers Formation, the Conestoga Formation, and (in Maryland) the Frederick Formation.

Phase #1 - Initiation of the carbonate platform:

A. Base of the Tomstown, Vintage, and Frederick Formations, all of which directly overlie the youngest transgressive clastics at the top of the Chilhowee Group, represent the initiation of the carbonate platform in their respective areas in the Early Cambrian (Read, 1989; Brezinski, 2004; Brezinski et al, In Press).
B. However, this formation contact, at least in the Great Valley and Conestoga Valley, bears a strong structural overprint in the form of a major mylonite that records decoupling of the carbonate stack from the underlying clastics - e.g. the Keedysville marble bed at the base of the Tomstown Formation (Brezinski, 1992; Brezinski et al., 1996; Campbell and Anderson, 1996).

C. Lithofacies of both basal Tomstown (basal three members) and Vintage indicate initial carbonate ramp deposition, although pure dolomite of Benevola Member of Tomstown and basal carbonate component of Emigsville Member of Kinzers record widespread winnowed shoal deposition.

Phase #2 - High-relief, narrow rimmed shelf deposition
A. Later in Early Cambrian carbonate deposition became restricted to very edge (few 10's of km) of shelf; remainder of shelf behind the rim shale-dominated, forming Waynesboro Formation (at least shale-dominated portions) and Rome Formation farther to the west.
B. Shelfbreak grainstone of Ledger formed in that narrow rim, feeding periplatform, pure carbonates of the upper (York) member of Kinzers by downslope gravity flow.
C. At times during narrow-rim deposition, distal "basin margin" facies experience very slow, "starved basin" deposition (e.g. upper, shale-rich member of eastern Kinzers.
D. Relationship/timing of Conestoga Formation deposition relative to these other Conestoga Valley units very poorly established, although overlap appears to have been minimal.
E. Frederick Valley succession also appears to record starved basin deposition then.

Phase #3 - Broad-rimmed (peritidal complex) shelf deposition (Figure 6).
A. Development of a broad complex of tidal flats at the distal edge of the shelf confirmed as latest Early Cambrian by cyclic peritidal carbonates of the Dargan Member at the top of the Tomstown Formation (Brezinski, 2004).
B. Peritidal cyclic facies of the Middle Cambrian Zooks Corner Formation (Meisler and Becher, 1971) in the Conestoga Valley (equivalent of lower Elbrook Formation in Great Valley) provide earliest record of the broad rim in the CV.
C. Conestoga Formation basin-margin facies accumulated in deep water seaward of the rimmed shelf in the Middle Cambrian. No fossils yet recovered to confirm that this any part of this formation was deposited in Upper Cambrian, more or less in the Ordovician.
D. Sauk II-Sauk III boundary drawdown left its mark in all areas: unconformity and condensed succession at base of Conococheague Formation in Great Valley, and Sitz Creek Formation of Conococheague Group in CV (; LST (low-stand systems track) package of deep marine fan deposits (lower Frederick Formation - Rocky Springs Station Member) of Frederick Valley (typical "reciprocal sedimentation" of condensed on non-deposition on platform coinciding with accumulation of thick LST in deep water).
E. Subjacent strata of upper Elbrook Formation represents TST on platform during sea level rise; reciprocal sedimentation yielded starved basin facies at top of Monocacy Member at base of Frederick Formation during Crepicepha/us Zone time.
F. Third- and fourth-order sequences deposited during sea level oscillations in the late Cambrian being resolved with increasing precision in the Conococheague and ageequivalent portions of the Frederick Formation. For example, "Adamstown Submergence Event" in middle Late Cambrian created reef-dominated TST in Conococheague (Thrombolite III), and simultaneously stalled down-slope transport to Frederick Formation sites, producing base of Adamstown Member of Frederick Formation (Taylor et al., 2009).
**Figure 6 - Cross section of Upper Cambrian and basal Ordovician units of the central Appalachians region**, showing advances in linking shallow inner shelf, outer shelf (peritidal rim) and deep marine (basin margin) successions by event stratigraphy with biostratigraphic control.

**#4 - Early Ordovician sea level oscillations**

A. Stonehenge Transgression at the very start of the Ordovician briefly terminated deposition of peritidal cyclic facies in the Appalachians (base of Beekmantown Group), and in other depositional basins throughout Laurentian North America.

B. Rest of Beekmantown Group records third-order alternations of sea level rise and fall with former producing restricted, cyclic peritidal deposits, and latter (at times) suspending cyclic facies deposition.

C. Great caution needed to sort out the effects of later-stage dolomitization that masks or overwhelsms that depositional signal in the Beekmantown Group (e.g. the Rickenbach and Larke Dolomites).

**REFERENCES CITED**


Edwards, J. Jr., 1988, Geologic map of the Woodsboro Quadrangle, Carroll and Frederick Counties, Maryland: Maryland Geological Survey Geologic Map, scale 1:24,000.


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TACONIC FORELAND BASIN AND ALLOCHTHONS

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Prologue:
Somewhere out in the Iapetus Ocean (s.l.), and we wish we could be more specific, deepwater sequences containing Pacific province graptolites and cold water (Baltic) conodonts accumulated from Late Cambrian through early Middle Ordovician. The graptolites are non-Gondwanan but are permissive of Laurentia. The conodonts have no Laurentian affinity. Taken together, the fossils indicate that the original sediments were deposited across a significant oceanic barrier from Laurentia, closer to a land mass that was not part of Gondwana. These rocks were pushed into a peri-Laurentian trench during the Middle Ordovician. We now find these umetamorphosed rocks as allochthons in the Late Ordovician Martinsburg foreland.

The discovery:
In 1943, Brad Willard found Lower Ordovician graptolite beds within the outcrop belt of the Late Ordovician Martinsburg Formation in Susquehanna Gap. To the east, in the Martinsburg belt of Dauphin, Lebanon, and Berks Counties, Willard (and others) found an assortment of odd non-Martinsburg-like rocks including red shales and cherts, lime-turbidites and limestone conglomerates, black shales, hemipelagic shales/siltstones, and even basalts and diabase dikes. In addition to the obviously older graptolites, more graptolite beds were found that might be older than Martinsburg.

The conclusion jumped:
It was well known that rocks of similar description and age formed the “Great Taconic Allochthon” of Vermont and New York. In 1946, George Stose proposed a Taconic allochthon for Pennsylvania incorporating these older odd rocks as the “Hamburg klippe,” emplaced on an overthrust above the Martinsburg Formation. Stose (1946) also concluded that the metamorphic Cocalico Formation in Lancaster County was part of the allochthonous series.

The fight begins:
The proposed Hamburg klippe met with varying degrees of acceptance – some pro, some con. The conspicuous massive overthrusting of the Taconic ranges in Vermont and New York over the carbonate lowlands of the Champlain valley made for an imperfect analogy to the subdued Hamburg klippe resting on the clastic Martinsburg. To many observers, the odd and older rocks seemed “mixed” with perfectly acceptable-looking Martinsburg. It was also possible that some of the “older” graptolite fossils might be lower Martinsburg age. Perhaps these allochthonous rocks were within the Martinsburg, rather than above it. Two camps emerged and slugged it out, paper by paper, field trip after field trip, for the rest of the 20th century (see Ganis and others (2001) for an account). Without age control, mapping seemed impossible and the rocks appeared chaotic.

Find me a fossil:
It was generally accepted during this warfare that the allochthons-mixed-with-Martinsburg model required undisputed Martinsburg age fossils somewhere in the mix, else all the rock was older than Martinsburg and supported the Hamburg klippe idea. In 2001, Ganis and others announced that several unequivocal Martinsburg age graptolite locations had been found within the delimits of the proposed overthrust, dealing a crushing blow to the Hamburg klippe model. Also revealed in this paper was a comprehensive account of additional older-than-Martinsburg graptolite localities,

offering great potential for organizing a stratigraphy within the allochthonous sequence, which they
named the Dauphin Formation with several distinct members. Clearly, only diligent work would be
required to separate the allochthonous rocks from the Martinsburg.

**Trench warfare:**

The contents of the allochthon say much about its history. Everything older than Middle Ordovician
occurs as olistoliths (dated with graptolites and conodonts) contained within a Middle Ordovician
trench fill olistostrome complex (*stop 10*). This trench fill contains complex interfingering facies (all
dated with graptolites and/or conodonts) of hemipelagites, turbidites (*stop 12b*), and pelagites (*stop
11a*). The formation of the trench fractured the subducting plate just enough to allow localized
outpouring of basaltic magma, the Jonestown volcanics (Lash, 1984), which cooked some of the
carbonate olistoliths containing Lower Ordovician conodonts.

The allochthon is actually a series of slices, the earlier of which has syntectonic deposition on top
(piggyback basin; *stop 11b*) containing graptolite fossils that date the emplacement. Figure 1 shows
the contents of the allochthon, its age, and an explanation (suggestion) of how it was assembled.

The allochthon (Dauphin Formation) and its Martinsburg cover once extended south and east over a
much broader area than the current extent, covering at least the area underlain by the Lebanon
Valley carbonates. They apparently also extended over parts of the Piedmont. Remnants of the latter
are preserved as the Cocalico Formation, now metamorphosed to phyllite. The correlation is
suggested by the lithologic similarity of the Cocalico protolith to Martinsburg shale and to the red
shales of the Dauphin Formation, as well as by the discovery of conodonts and deformed graptolites
in the Cocalico, which point to a similar deep water origin.

**Maps!**

As the number of dated localities grew, both of Martinsburg age (*stop 12a*) and older allochthonous
rock, the preliminary stratigraphy and lithotectonic emplacement scheme proposed by Ganis and
others (2001) for the Great Valley appeared to hold up and hinted of large-scale structure. Although
tentative, a large Alpine-scale overturned anticlinorium cored by the allochthons and flanked by the
Martinsburg, appeared across Dauphin and Lebanon Counties, supporting the idea that Taconic
nappes overran the Great Valley (see Ganis and Wise, 2008). In 2009, Gale Blackmer teamed with
Ganis under the StateMap program and began a serious mapping program in these counties. The
emplacement of the allochthons as an early event later covered by Taconic foreland fill (Martinsburg
Formation) has been mapped out. The Alleghanian transport of the “meta-Martinsburg/allochthon”
terrane of the Cocalico (*stop 9*) on the Yellow Breeches fault has been traced out as well. We are
pleased to provide a copy of the new preliminary map for Dauphin and Lebanon Counties for this
Field Conference (see guidebook).
Figure 1. Fossil-dated relationships among phases of the Taconic orogeny in the Martinsburg/Dauphin foreland. The arrows indicate packages that have been moved as allochthons to be embedded into younger units with fossil-dated matrices. Stratigraphic time gaps of few m.y. commonly separate the two events. References to fossil dates are at the bottom of the figure. Not 1 on figure, the Jonestown volcanics were emplaced into the olistostrome. Note 2 on figure, presumably, this occurred as the Westminster terrane obducted onto the Laurentian margin. Note 3 on figure, in this area the foreland basin formed about one graptolite zone or perhaps 1 to 1.5 m.y. earlier than area to the east (Ganis and Wise, 2008)

The following series of diagrams illustrates the events involved in the emplacement of the allochthon(s) and the changes in geology that resulted, culminating with the formation of the Taconic Alps.

References


The allochthons, some with piggyback basins on board, have been pushed out of the trench and are moving towards a newly forming foreland. Obduction of the allochthons onto the Laurentian margin causes subsidence of the carbonate platform, which is balanced by a peripheral bulge (topographic high) farther back on the continent. A restricted, euxinic, carbonate/clastic mud filled trough (Myerstown basin) forms on the subsided platform. The dirty limestone of the Myerstown Formation (stop 8), deposited within this basin, marks the earliest part of the platform to foreland transition in Pennsylvania. The top of the Myerstown loses carbonate and transitions to black shale. The trajectory of the allochthons and obduction/subsidence in general is causing only marginal effects to the Chambersburg Formation to the southwest.

**Impinging “Hamburg” Allochthons**
(L. Cambrian to M. Ordovician Dauphin Formation)  
With Piggyback Deposition  
(Linglestown Formation, informal)
The allochthons have been emplaced over the Myerstown basin, perhaps with a thin layer of black shale in between. The peripheral bulge to the northeast has eroded away (the Blackriveran unconformity), in the process providing Beekmantown carbonate clasts to the Hershey facies of the Myerstown. The transitional basin has migrated to the northeast to become the Jacksonburg basin. To the southwest, the Chambersburg Formation is gradually losing its carbonate and is transitioning to a foreland as well. Ash beds are being deposited in the upper Chambersburg and in the Jacksonburg. Ash may have fallen on the allochthons as well, but it was not preserved in that tectonically active environment.
The "foreland fill" phase has begun, with deposition of Martinsburg muds over the emplaced allochthons (stop 12) and the Chambersburg Formation. To the northeast, the Jacksonburg basin is slowly losing its carbonate content and it, too, is transitioning to foreland deposition. The lower part of the Martinsburg covering the allochthons contains graptolites that are the same age as the very top of the Jacksonburg. Deposition of the Martinsburg cover over the allochthons must have begun just after the giant Millbrig ash fall, which is present in the Jacksonburg but not in the Martinsburg.

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**Early Martinsburg Formation**
(Covers Allochthons)
"Foreland Fill"
Foreland deposition continues as Taconic fill. The hinterland continues to imbricate and rise to the formative Taconic range (source of the Martinsburg sediment). The foreland migrates diachronously to the northwest, continuing to replace the drowning carbonate platform.
The Martinsburg Formation and the Dauphin Formation allochthons are folded together into the giant nappes of the Taconic Alps (the Taconic Orogeny!). That was a progressive event (~ 450-455 Ma) moving from southeast to northwest, and more intense in the southeast. The equivalent units in the Piedmont portion are called the Cocalico Formation (stops 7 and 9).
Epilogue: During the Alleghanian orogeny, some of the Cocalico Formation (stop 9) was transported as a very large (multi-county-sized) overturned nappe on the Yellow Breeches fault and juxtaposed against its non-metamorphic equivalents (Martinsburg and Dauphin Formations) in the Great Valley.

Mixed Terrane of Martinsburg and Allochthons (greenschist grade)

Belt of Martinsburg Foreland in the Great Valley

Alleghanian Thrusting

(includes the Yellow Breeches fault)

Moves “Cocalico” (mixed terrane of lower greenschist grade Martinsburg and allochthons) northward into the Great Valley.
TACONIC NAPPE STRUCTURES OF THE SUSQUEHANNA PIEDMONT

Donald U. Wise, University of Massachusetts at Amherst

Slab Roll-back Tectonic Model

This talk deals with structures associated with the "Taconic Orogeny," an event generally considered to be of later Ordovician age. In reality the Ordovician date represents the culmination of Iapetan Ocean processes that had been moving island arcs, micro-continents, and their associated basins ever cratonward through most of Cambrian and Ordovician times. Only in later Ordovician time did these processes finally impinge on the Laurentian craton to leave a relatively easily read, sedimentary and structural record of those final events. "The Taconic Orogeny" represents only the final 15-20 my of a 100 my oceanic migration that finally arrived at the cratonic edge to pile its accumulated debris against the craton, to flow onto a peripheral carbonate platform, and as its dying act, to deform that allochthonous mass and some of the underlying carbonate platform into a great nappes complex (Ganis and Wise, 2008, Wise and Ganis, 2009). Driving this migration was the long-term landward roll-back of the hinge zone in front of an island arc's steadily sinking oceanic slab (Figure 1). As the slab sank its frontal hinge zone migrated landward dragging the arc with it, a mechanism that ended when the slab passed beneath the thick old craton, was no longer able to peel off the slab, detached and sank into the mantle's depths.

This "Taconic Orogeny" has many similarities to the modern Italian Apennines with their exotic "argille scagliosa" masses of seafloor material piled onto a carbonate platform. (Parotto and Praturlon, 2004; Butler, et al., 2006). As the underlying Tyrrhennian plate continues to sink beneath Italy, its hinge zone is migrating eastward, with its fast-moving asthenospheric wedge dragging the overlying crust with it, folding and telescoping it into the Apennine mountains (Figure 2). This advancing load is depressing and moving a foreland basin ahead of it with contents that include some of the former Tyrrhenian sea floor, the famous "argille scagliosa." In the Apennines the carbonate platform is being intruded and locally covered by the great Latian volcanoes. The Taconic nappes on the Pennsylvania platform show now evidence of this local igneous activity. In Italy roll-back has been able to continue because the basement craton is young Hercynian crust offering only minor resistance to sub-crustal peel-off. The Taconic slab encountered the billion-year-old Laurentian crust when it made landfall. It broke off and sank, terminating the volcanic phase before the carbonate platform could become involved.

Specific Piedmont Features Related to the Roll-back Model

This section attempts to fit specific details of the Pennsylvania - Maryland Piedmont into the proposed roll-back model. In Cambrian and early Ordovician times, roll-back tectonics acting far offshore steadily closed the Iapetan Ocean to pull the James River and possibly other island arcs shoreward toward Baltimoria, that was then separated from the main Laurentian craton by the Octoraro Seaway.

As the Iapetan Ocean closed, parts of its floor and associated deep sea sediments were pushed onto and over Baltimoria. Some of that micro-continent's upper crust and sedimentary cover was detached and displaced cratonward into the Octoraro basin to become future mantled gneiss domes. Other masses of Iapetan deep-sea materials were transported across those micro-continent roots into the Octoraro basin to be deposited largely on its seaward side. These allochthons included deep sea pelagic sediments, volcanics, sea floor slabs, and olistostromic masses with a variety of exotic

fragments, the future Dauphin Formation, Jonestown Volcanics, mafic and ultramafic Piedmont masses, Sykesville and related diamicites.

Figure 1. Slab roll-back model for the Taconic Orogeny in the Susquehanna Piedmont. Lithospheric slabs consisting of oceanic crust and upper mantle mostly sink rather than descend obliquely as most text books show. A rolling hinge zone of sharper curvature occurs where the flat-lying ocean plate bends to begin its downward trip. As the slab sinks, the hinge rolls away taking the island arc with it. This model explains how the most distal deep-water sediments can be the first allochthons to be deposited so far inland directly onto the carbonate floor of the foreland basin. Further descriptions are given in the text.
As roll-back continued to push the arc(s) and micro-continental roots cratonward, partly lithified contents of the Octoraro Basin were thrust, folded, and thickened into a weak mass that finally reached elevations to flow and thrust onto the slightly lower edge of the craton's bordering carbonate platform. As part of the overriding process, the mass, picked up pieces of submarine slope deposits of the platform and imbricated them into thrust repetitions now exposed in the Martic Zone. Once on the craton, isostatic processes helped lower the surface and develop an advancing foreland basin into which the newly arrived materials flowed in a generally N30W. Distal allochthons from Iapetus that had ridden high on the mass but continued to move forward in caterpillar tread fashion finally reached the leading edge to be deposited onto the flooded carbonate floor of the basin and then be covered by thick flysch of the Martinsburg and Cocalico Formations. The basin was a migrating one, the rear edges being thrust and folded into mountains while advancing nappes continued to move the depo-centers farther inland as part of the migrating foreland basin (Ganis and Wise, 2008).

Meanwhile in the hinterland, metamorphic and igneous process continued to rework the complex mixture of two marine basin fills and island arc-micro-continent remnants into the evolving Taconic Alps. Thermal effects and flowage formed hinterland s-surfaces and schists with orientations parallel to flow planes in the still-moving masses. In the Martic Zone, the earlier Martic contact and associated imbricate slices were deformed into "enigmatic folds" discussed below and developed foliation parallel to their axial surfaces. Time-wise, by ~ 450 m.a. (Wise, Smith, Jercinovic, and others, 2007) these hinterland mountains were well developed when quartz vein mineralization at Pequea Mine (Figure 4) was localized along the axial surfaces of the main Taconic folds. At the same 450 ma date, graptolites show the distal edge of the still active foreland basin was receiving the last of the Martinsburg flysch (Ganis and Wise, 2008). During the next 5 my, the outward advancing nappe structures engulfed these final deposits and erosion reduced the distal parts to levels where deposition of the Silurian unconformity began a new cycle of foreland accumulation. The Laurentian craton had accreted a Taconic package to its eastern edge.
Multiple Folding:

Subsequent discussions require recognition of multiple folding and strong later Paleozoic (Alleghanian) warping and telescoping of the Taconic fabric. By the end of Ordovician time, the regional flow and axial plane pattern resembled the cartoon lower panel of Figure 3 with a steeply dipping root zone that flattened into relatively horizontal axial surfaces across the craton. Later, the Alleghanian Orogeny (upper panel) warped the Taconic flow planes and axial surfaces across the Tucquan Antiform to have steeply north-dipping orientations in the Martic Zone, a tilt far beyond original relatively horizontal or recumbent orientation.

This two-fold deformational history had been recognized as early as 1944 by Stose and Stose who noted that the main schistosity had been deformed across a younger Tucquan Arch. Details of the associated multiple deformations and fold systems were defined by Freedman, et al., (1964) and extended through the Martic area by Wise (1970) as shown in Figure 4 with terminology explained in its caption.

The Martic Problem:

The Martic problem began when Knopf and Jonas (1929) noted that on the projection of the west-plunging nose of Mine Ridge's crystalline basement (Figure 4), schists then regarded as Precambrian rested on the Conestoga Limestone, then regarded to be Ordovician in age. In an era when even the existence of such structures was debatable, they proposed a regional-scale "Martic Thrust" to explain this age anomaly. Their model included several areas of Conestoga Limestone south of the thrust line (Figure 3), interpreted as "windows" or fensters through the thin edge of the thrust sheet.
Others, including Miller of Lehigh University (1935), claimed the schists were only the metamorphosed equivalent of Martinsburg clastics that rest on Ordovician carbonates of the Great Valley. Mackin (1935) took a more structural down-plunge approach to support the thrust model and so the fight began. Stose (1935) added fuel to the controversy by proposing that the entire Reading Prong was the erosional remnant of an even larger thrust sheet.

Subsequent careful mapping by Cloos (Cloos and Hietanen, 1941) showed that blob-like outcrop patterns on Knopf and Jonas's geologic map actually interconnected as a series of five folded thrust sheets repeating the sequence Antietam schist, Vintage Dolomite and Conestoga Limestone (Figure 3). Cloos (1941) reports having spent 400 days in the field producing that map, a model of good basic geology. In the text he notes the five-fold repetition of the stratigraphy and concludes that this must result by thrusting from the south with subsequent folding and overprinting of the thrust sheets by younger schistosities and lineations. However, he leaves open the nature of the actual Martic Contact, noting it follows bedding and that the "Precambrian" schist above the contact does not look more metamorphosed than the underlying schists. At one location near Quarryville (Q on Figure 4) where an Antietam Ridge meets the "Wissahickon Schist" at the Martic contact the lithologies are too similar to tell apart. (He left a blank spot on the map at that location.)

Cloos retained this cautious, neutral interpretation through the 1949 field trip to the key stops and through the writing of the 1960 field trip guidebook. Even though Cloos's caution seems unwarranted with today's interpretations, his mapping has withstood the tests of time. During the 1960s, Franklin and Marshall College students remapped the area as a series of small senior theses. Many details were added as the multiple folds were unscrambled but even with new outcrops and improved maps, Cloos's contacts could not be shifted by more than about 100 meters.

During the 1950s and 60s as great overthrusts became widely recognized elsewhere, Hubbert and Rubey's (1959) excess fluid pressure model for thrust mechanics appeared, and multiple fold analyses began to unscramble deformations of the Martic imbricate sheets, the Martic thrust model has been generally accepted. Knopf and Jonas's Wissahickon Schist has been split into several components (Blackmer, 2004), the Martic areas becoming Octoraro Schist of the Octoraro Seaway (See also Blackmer this volume) and the Antietam and Conestoga Formations recognized as Cambrian slope deposits from the Lancaster-York carbonate platform into that basin (See Taylor this volume).

The original "Martic Problem" seems to have been largely solved with recognition of the contact as one of many Taconic-age thrust repetitions of Cambrian slope deposits off the carbonate platform. Knopf and Jonas' "fensters" now appear to be a thin cover of Conestoga Formation riding on the back of yet another imbricate thrust slice (Figure 4). Cloos's stated inability to distinguish between "Antietam" and "Wissahickon" where the two meet at Quarryville turns out to have been quite valid. These are essentially the same unit brought into juxtaposition by a thrust. Even though the schists riding just above the Martic contact could be called "Antietam" as hinted by the coloring on Figure 4, there is little to be gained by shifting the official names and contact upward by one thrust sheet. Recent work by Blackmer (this volume) on zircon provenance supports the same interpretations. As noted above, the main folding and schistosity that deform these thrusts are also Taconic as shown by the recent 450 m.y. monazite date from the Pequea Mine (Figure 4) where quartz mineralization is localized along an F\textsubscript{0x2} fold hinge.
Figure 4. Multiple folding of imbricate thrust sheets in the Martic Zone. (Mapping after Cloos and Hietenan, 1941; interpretations after Wise, 1970, and Wise and Ganis, 2009.) This terminology in the Martic Zone (Figure 3) utilizes a double subscript numbering system, the first number being the s-surface being folded, the second is the number in the deformation sequence. For example, the fold axis of bedding ($S_0$) in the Antietam sheets by the main Taconic deformation ($D_2$) is marked $F_{0x2}$ whereas the axis of the Tucquan Antiform, formed by folding of the main cleavage ($S_2$) by the Mine Ridge ($D_3$) basement events is marked $F_{2x3}$, etc.

Enigmatic Folds at Safe Harbor and Turkey Hill.

At the west end of the Martic Zone, the Martic contact is scalloped into a series of folds, the largest and most pronounced at Turkey Hill (Figure 3). Schamel's (1963) mapping included many measurements of bedding and schistosity related to several overprinted fold patterns. Figure 5's map and plots, extracted from data in his honors thesis at Franklin and Marshall College, show several apparently conflicting structural details about these folds.

1) Equal area plots of bedding poles on Figure 5 form a girdle indicative of folds plunging $S_{85W}$ at small angles. These are rolled leading edges of the early-Taconic thrust sheets. The Vintage-covered nose of one of these appears just east of Stop #2.
2) Poles to axial planes and main schistosity show a $N_{63E}$ strike, axial planar to the large-scale, west-plunging folds of the Martic contact and essentially normal to the regional Taconic $N_{30W}$ transport.

In nearby areas, the Martic contact changes from its zig-zag pattern in passing through these folds to $N_{70-75E}$ trends on the projection of the north edge of the Mine Ridge basement uplift. (Figure 3 and Wise, 1970) The oblique relationship between trends of the older schistosity and folds with respect to the Martic contact and its younger, Mine Ridge-related trends at that location is evident at the south edge of Figure 5.
Figure 5. Enigmatic folds of the Safe Harbor Area. The cross section shows they are synformal anticlinal lobes but unlike typical anticlines have younger rocks in their cores. Similar folds occur to the north, the most prominent at Turkey Hill (Figure 3).

At first glance, folds in the Safe Harbor cross-section of Figure 5 seem to show all the thrust sheets, including the Martic contact, folded as a single package into typical synclinal folds. Appropriately, the youngest sheet of Octoraro Schist from the top of the pile appears as in the synclinal cores. However fold asymmetry and north dipping cleavage suggest these folding resulted from south-directed transport, contrary to regional tectonic patterns. Cloos proposed a second mountain-building phase of south-directed transport produced these anomalous geometric relationships. A mistaken "Aha!" occurs when one realizes these folds are noses of Taconic anticlinal lobes, once recumbent but now rotated to north dipping axial surfaces by the Tucquan Antiform (Figure 3). Tectonic transport on the folds was actually downward and northward into the ground. With this, the apparent synclinal fold pattern, cleavage dip, and tectonic transport direction all make geometric sense --- but then the enigma appears. If these really are anticlinal lobes how can they have youngest, highest thrust sheets in their cores?

One model to explain this enigma is presented in the cartoon sequence of Figure 6 and its caption. When the first Octoraro Basin materials arrived, the carbonates in the thrust sheets and platform edge were cold, thinly buried and relatively strong. As the overriding mass continued to thicken, the increasing load raised fluid pressures in what is now the Harpers Phyllite beneath the Antietam schists. Hubbert and Rubey (1959) excess fluid pressure mechanisms caused the Antietam and its overlying Vintage and Conestoga carbonates to detach across a broad area and move forward easily as part of the overriding mass to pile up as the Martic imbricate thrust zone.
As known from generations of rock mechanics experiments (Robertson, 1960), limestone becomes very weak and ductile when burial depths exceed about 2 km. As the mass of Octoraro clastic sediments and future schists continued to thicken and add to the total load, underlying carbonates became increasingly ductile. This change was further enhanced by thermally driven recrystallization as greenschist facies conditions migrated into the pile. What had once been a strong platform slope and edge, changed strength characteristics to become largely similar to those of its overriding, northward-flowing clastic mass. Upper parts of the Cambrian platform and its slopes that had been the strong floor for the overriding clastics and thrust sheets, now began to flow northward along with the thrust sheets and the moving clastics. Within the Martic zone, these flow planes cut down-section through the contact and the imbricate sheets to form anticlinal noses with younger emplaced Octoraro Schist in their cores (Figure 6, center panel). With Alleghanian tilting across the Tucquan Antiform, and subsequent erosion these became the observed enigmatic synformal anticlines with age reversal in their cores.

**Figure 6. Model to explain the enigmatic Safe Harbor and Turkey Hill folds.** Following early-stage Taconic imbricate thrusting of the Martic Zone, the overriding schist mass continued to increase in thickness. Deeper burial caused underlying carbonates to become more ductile allowing some flow lobes to deflect downward and pass through the imbricate stratigraphy. These formed outward-advancing anticlinal lobes with younger cores. Later Alleghanian deformation rotated these past recumbency for subsequent erosion to leave as synformal anticlines with anomalous age relationships.
Nappe transport of distal allochthons to become basal units of the foreland basin

Interpretations of the basal contact of the Martinsburg-Dauphin-Cocalico over the carbonates have changed with time. Early workers noted the Cocalico and Martinsburg Formations rest on a variety of formations of the upper Beekmantown group as well as a few post-Beekmantown basins of Myerstown cement rock and Annville pure limestone. As a result most stratigraphies from Jonas and Stose in the Lancaster Quadrangle (1930) to Hobson in Berks County (1937), to Stose and Stose in York County (1944) to Meisler and Becher in the Lancaster Area (1971) have interpreted an erosional unconformity at the base of these flysch units.

Such interpretations were plausible so long as the red units at the base of these formations were believed to be volcanogenic sediments deposited on an erosional surface. (above authors). As usual Gray (1952) was ahead of his time with a tectonic interpretation. Recognition of these red units as allochthons from distant deep sea deposits, many with ages older than the units upon which they rest (Ganis et al., 2001, Ganis and Wise, 2008), invalidated all previous models based on simple deposition of the flysch above an unconformity. Some kind of far-travelled transport mechanism had to move this most-distant-sourced material across most of the carbonate platform. Furthermore, the fact that there is no sign of these allochthons in the Martic Zone requires the mechanism to have moved them across that early and actively evolving platform edge leaving no sign of their passage. Finally they must have arrived in the middle and distal parts of the foreland basin to become the initial Dauphin deposits on that carbonate platform.

The middle panel of Figure 6 suggests a solution. The future allochthons from Iapetus were emplaced across Baltimoria from temporary emplacement on the seaward edge of the Octoraro Basin. As basin inversion began, these rear locations became the upper parts of that tectonic mass, moving forward in caterpillar-tread fashion. In the early stages, Octoraro Seaway materials from the shoreward part of that basin were in contact with the carbonates. They formed the basal contact relationship leaving no record of the deep-sea allochthons about to pass over at higher levels. As the mass continued travel to the north, the caterpillar tread motion slowly shifted these "passengers" to the "front of the bus" to slide or be rolled onto the relatively pristine carbonate floor of the foreland basin. Whether most slid off right-side-up or rolled over to have upside-down orientations remains to be determined. As sinking of the foreland basin continued, Cocalico and Martinsburg flysch was deposited across the newly arrived allochthons.

Regional inveresion of the Lebanon Valley nappes.

Another problem of local nappe mechanics is how the great allochthonous masses could have been transported at least 50 km or more to be deposited on the carbonate platform in so gentle a fashion that the contact could be mistaken for an unconformity. In some areas there was enough fault motion at their basal contact to cut out as much as 1000 feet of the then undeformed and apparently quite strong carbonate platform (Meisler and Becher's 1971 Manheim Valley area), a geometric omission formerly ascribed to pre-unconformity erosion. The problem is compounded by the fact that everything, including the carbonate platform, was engulfed in overturning and nappe flowage on a massive, regional scale following so gentle an initial emplacement.

The nappe mechanisms were essentially the same as those of the Martic zone except here there was no significant thermal pulse to aid the flowage. Once emplacement began, continued thickening by sedimentation and tectonic overriding increased lithostatic pressure on the carbonates to turn them from strong platform into "toothpaste." The basin floor became ductile with strength approaching that of the overlying fill. All began to flow forward together in a fashion similar to a cold bottom glacier, anchored at the bottom of the carbonate platform but moving along with the
basin fill at intermediate and shallow levels to form great flow lobes. Present erosion levels in the Lebanon Valley region are at about that intermediate level where most beds have been inverted and stretched into regionally overturned nappes.

Figure 7. Timing of events in the foreland basin. (After Ganis and Wise, 2008). Traditional "Taconic" platform events covered ~15 my. Allochthons include Late Cambrian deep-water components from Iapetus and matrix materials that predate the last carbonate platform deposits at ~ 458 ma. The Martinsburg flysch is in depositional contact over Allochthon #2. Foreland orogeny ended with erosion for Silurian unconformity at 443 ma but hinterland must have continued activity into the Silurian. Dates based largely on graptolites and conodonts from Ganis et al. (2001). A more stratigraphic version explaining this assembly process is in notes for Stop #7.

References:


LATE PALEOZOIC COMPRESSION AND DEXTRAL TRANSPRESSION TO TRANSTENSIONAL COLLAPSE IN THE PENNSYLVANIA PIEDMONT

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Introduction

The Alleghanian Orogeny has left the most conspicuous mark on the Appalachian orogen of any of the mountain building events. The deformation in the vast Valley and Ridge Province is overwhelmingly the result of this event (Hatcher et al., 1989). Research in the 1970’s showed that there was more than 225 km of Alleghanian overthrusting in the southern Appalachian Valley and Ridge (Hatcher, 1981). This overthrusting is thin-skinned west of the Blue Ridge but basement involved to the east. There is decreasing throw on the thrust faults into the central Appalachians (Hatcher, 2002) but recent studies suggest that overthrusting in Pennsylvania may still be between 100 and 200km (Faill, 1998).

In the 1980’s, a new aspect of the Alleghanian orogeny emerged in the southern Appalachians. Significant dextral strike slip faulting in the Piedmont was found to be a major aspect of the orogeny (Gates et al., 1986). Tectonic models of the Appalachians were adjusted to propose oblique convergence between North America and Africa or include a rigid indenter generating tectonic escape of eastern provinces to the south (Arthaud and Matte, 1977; Gates et al., 1988; Hatcher et al., 2002). This dextral offset is estimated to exceed 341 km in Virginia and North Carolina where it was best studied (Gates et al., 1988). The dextral strike slip component may therefore exceed the overthrusting.

The Pennsylvania Piedmont followed a similar path to the rest of the Appalachians, with identification of contractional deformation (Freedman et al., 1964; Wise, 1970) followed by dextral strike slip deformation (Gates, 1992; Valentino et al., 1994). The Pennsylvania Susquehanna River traverse is an especially good area to study the effects of the Alleghanian Orogeny in the Piedmont because it crosses from dominantly strike slip deformation in the southeast to mixed deformation in the central portion to contractional deformation in the northwest. This paper will first review the bedrock evidence for strike-slip deformation in the Piedmont, it will then speculate on how much lateral shuffling may have taken place and finally it will place strike-slip deformation into the context of the well-documented classic contractional Alleghanian orogeny (Faill, 1998).

Geologic Setting

The Pennsylvania Piedmont is in a unique position within the U.S. Appalachians. In contrast to the rest of the northeast-trending Appalachians, the Piedmont trends generally east-west from the New York Promontory into the Pennsylvania Reentrant (Figure 1). This unusual geometry was apparently created during Late Proterozoic rifting (Rankin, 1975; Thomas, 1977) as a transform fault in the opening Iapetus Ocean. The rift created a depocenter for sediments of the Peters Creek Formation at that time (Gates and Valentino, 1991; Valentino and Gates, 1995). Colliding continents with this unusual geometry during the Paleozoic orogenies developed a distinctive pattern of tectonism. The New York Promontory bore the brunt of collisions and the rocks were highly deformed and metamorphosed (Hall, 1968). Such is the case with the Manhattan Schist, Hartland Formation, Inwood Marble, Lowere Quartzite and Fordham Gneiss (Table 1) and their equivalents north of New York City and in Connecticut. In contrast, the time equivalent rocks in the western Pennsylvania Piedmont such as the Peters Creek Formation and Octararo Formation, and others are
Figure 1. Geologic map of the Pennsylvania Piedmont (with inset of the eastern U.S.) showing the main geologic features including the strike-slip shear zones (with arrows) that form the duplex (see Table 1 for stratigraphy). LVSZ = Lancaster Valley shear zone; T-MR = Tucquan-Mine Ridge antiform; PGSZ = Pleasant Grove shear zone; DSZ = Drumore shear zone; PBZ = Peach Bottom shear zone; CVSZ = Cream Valley shear zone; HVSZ = Huntington Valley shear zone; OSZ = Octararo shear zone; RRSZ = Rock Ridge shear zone; PDSZ = Port Deposit shear zone; RMSZ = Rosemont shear zone; CCSZ = Cream Valley shear zone.

Table 1. Lower Paleozoic shelf stratigraphy from the Manhattan Prong, Philadelphia-Maryland domes and the Pennsylvania Valley and Ridge.

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comparatively less deformed and of lower metamorphic grade. It is the region between these two areas that exhibits the more complex relations.

**Distribution Of Strike-Slip Faults**

The Late Paleozoic strike-slip faults form a left-stepping strike slip duplex across the southeastern and central Pennsylvania Piedmont (Figure 1). The duplex opens abruptly at the eastern Pennsylvania state line with the Huntington Valley – Cream Valley shear zone (HV-CVsz) which bounds the north side of the duplex and the faults that bound the several horses across the Piedmont progressively transfer strike-slip motion to the duplex bounding fault to the south, the Rock Run – Port Deposit shear zone (RR-PDsz) (Figure 1) (Gates et al., 1999; Orndorff, 1999). The duplex is asymmetrical in its abrupt eastern edge but diffuse distribution of faults and structures to the southwest.

By far the most impressive of the shear zones in the duplex is the HV-CVsz that stretches from Trenton, NJ westward through the Philadelphia area (Figure 1). The zone is marked by extensive near vertical mylonite to ultramylonite that exhibits greenschist retrograde metamorphic assemblages with consistent right lateral kinematic indicators. Most of the strike slip offset of the entire rest of the Piedmont is carried by the very high strain HV-CVsz in this area. The offset carried by the HV-CVsz on the north side of the duplex steps across to southern zone, RR-PDsz, throughout the Piedmont.

The first and easternmost left-stepping dextral fault in the duplex is the Rosemont shear zone near Philadelphia (Figure 1). Consistently dextral kinematic indicators are well developed in the high grade mylonite through the Wilmington Complex and high grade portions of the Wissahickon Formation. This zone has an apparent conjugate sinistral strike slip zone, the Crumb Creek shear zone (Valentino et al., 1995). The presence of the conjugate pair indicates a pure shear component to an otherwise simple shear system. This indication of pure shear occurs closest to the New York promontory (Valentino and Gates, 2001).

There are several other largely strike slip shear zones that cross the Piedmont in the transfer of offset from the HV-CVsz to the RR-PDsz. These zones divide the Piedmont into strike slip horses, many of which exhibit transpressional features. To the east, these fault blocks contain interpreted transpressional domes with cores of Grenville gneiss and covers of Setters quartzite and Cockeysville marble. Individual shear zones are relatively thin but related low strain transpressional features such as overprinting crenulation cleavages in metapelite to metapsammite schists are dispersed across the area (Freedman et al., 1964; Wise, 1970; Valentino and Gates, 2001). One shear zone, the Octararo shear zone (OSZ) crosses the western edge of the State-Line mafic complex through the serpentinite rim (Figure 1). This zone is composed of an anastomosing series of faults with approximately 10-12 km of total offset (Gates, 1992). The mylonite displays consistent dextral shear sense at greenschist facies conditions with apparent decreasing temperature to brittle conditions. The Octararo shear zone combines with the RR-PDsz in Maryland to form the Gunpowder Shear Zone (Gates et al., 1999), another large strike slip fault.

Several of the faults do not connect with the RR-PDsz within the Pennsylvania Piedmont if at all (Figure 1). The Pleasant Grove zone is a low grade extension of the HV-CVsz that continues into Maryland where it bounds the western Baltimore terrane, also within the strike slip duplex (Valentino et al., 1994; Gates et al., 1999; Valentino, 1999). It is well defined by the Drumore phyllonite where it crosses the Susquehanna River (Valentino, 1999). The Peach Bottom shear zone forms the famous slate of the same name which is really an ultraphyllonite (Valentino, 1999).

The areas between the dextral shear zones form strike slip horses but are internally deformed as well. In addition to the transpressional domes and anticlines near Philadelphia (Avondale,
Woodville, West Chester) and Baltimore (Towson, Phoenix, Woodstock, Texas, Chatauque, Clarksville), there are multiple crenulation cleavages (Freedman et al., 1964; Wise, 1970) and low angle fault surfaces as well. Many of these fault surfaces are thrust faults that accommodated crustal thickening during the Taconic and Acadian (?) Orogenies (Faill, 1998; Wise and Ganis, 2009). Several of these faults were reactivated during late Paleozoic transpression as orogenic float surfaces (Gates et al., 1999; Valentino et al., 2004). These shear zones contain mineral lineations that are parallel to the orogen (east-west in Pennsylvania) rather than perpendicular to it as would be expected with thrust faulting. They are typically low grade, shear banded silver metapelite schists that contain multiple crenulation cleavages (Table 2). Some orogenic float surfaces are not reactivated thrust faults but were produced during dextral transpression, such as the strike parallel displacement on the Martic thrust within the Tucquan-Mine Ridge antiform (Valentino et al., 2004).

**Transtensional Collapse**

Late in the strike slip event and farther to the west, the faults and associated structures exhibit transtensional collapse (Valentino and Gates, 2001). Features demonstrating the collapse include conjugate extensional cleavage (Table 2) in the Tucquan-Mine Ridge antiform (Figure 1), normal faults and a normal component on several of the strike slip faults and overprinting mineral relations showing dropping temperatures during shearing in many shear zones but especially the Rosemont-Crumb Creek system. The collapse appears to be asynchronous and with different amounts of offset across the Pennsylvania Piedmont.

The proposed model for the development of this transtensional collapse rests largely on the geometry of the New York promontory-Pennsylvania reentrant (Figure 1). As terranes entered the New York Promontory through dextral strike slip movement they entered a restraining bend and underwent transpression. As they moved into the Pennsylvania reentrant, they entered a releasing bend and underwent transtension. Because the rocks were so compressed in the promontory they were likely elevated with respect to surrounding areas (Figure 2). Therefore, there may have also been a component of gravitational collapse (Dewey, 1988) in the late extension (Valentino and Gates, 2001). Such gravitational collapse of Alleghanian transpressional structures has been documented in the eastern Virginia Piedmont (Gates and Glover, 1987).

There have been several geochronology/thermochronology dating studies in the Pennsylvania and Maryland Piedmont. Gates et al. (1999) compiled all of the published geochronology data and found that the peak of late Paleozoic using K/Ar on micas was from 360 to 260 Ma whereas for Rb/Sr on micas, it was 360 to 280 Ma. Krol et al. (1999) performed a detailed Ar/Ar analysis across the Pleasant Grove shear zone (PGsz) to the Gunpowder shear zone in Maryland on muscovite and biotite. They found that the rocks within a restraining bend of the PGsz exhibit younger ages of 308-313 Ma and rapid uplift. The rocks within the Baltimore terrane horse exhibit the same young ages and uplift (302-310 Ma). Pyle (2006) identified both Acadian and Alleghanian low grade signatures in Paleozoic quartzite of the central Piedmont (378 Ma and 272 Ma), and K-Ar whole rock ages of 279 and 276 Ma were obtained from the ultraphyllonite of the Peach Bottom shear zone (Faill and Smith, 1994). In contrast, rocks in releasing bend of the PGsz and to the west of the duplex in the Westminster terrane (Peters Creek –Octararo Formations) exhibit consistently older ages of 338-368 Ma. From these data, we interpret a general uplift on the whole area from about 365 to 335 Ma (Figure 2) but continued dextral movement with uplift and exhumation of rocks through transtensional collapse within the strike slip duplexes until about 280-260 Ma depending upon location.
Table 2. Fabric elements for rocks within the southwestern Pennsylvania Piedmont.

279 and 276 MA were obtained from the ultraphyllonite of the Peach Bottom shear zone (Faill and Smith, 1994). In contrast, rocks in releasing bend of the PGsz and to the west of the duplex in the Westminster terrane (Peters Creek–Octararo Formations) exhibit consistently older ages of 338-368 Ma. From these data, we interpret a general uplift on the whole area from about 365 to 335 Ma (Figure 2) but continued dextral movement with uplift and exhumation of rocks through transtensional collapse within the strike slip duplexes until about 280-260 Ma depending upon location.

Maguire et al. (2009) found even younger Ar/Ar ages of 244 Ma on biotite in metamorphic rocks under the New Jersey Coastal Plain to the east. They interpreted these ages to reflect continued
Alleghanian transpression east of the Pennsylvania Piedmont. Gates and Glover (1988) found similar ages and relations in eastern Virginia.

Figure 2. Distribution of Paleozoic metamorphism by index minerals in the Pennsylvania Piedmont.

Model For Transported Terranes

There was certainly synchronous contraction and lateral transport in the Pennsylvania Piedmont during the late Paleozoic. The question remains about the degree of lateral transport. If it is assumed that all of the high grade metamorphism was produced in the buttress of the New York Promontory while the rocks in the Pennsylvania Reentrant were protected, the amount of translation of high grade rocks from the Promontory into the Reentrant could be a measure of the offset (Figure 2). The transtensional collapse of the compressive structures like the Tucquan-Mine Ridge antiform could complicate this relationship but simple calculations argue for a dextral offset of at least 100 km primarily on the southeastern side of the Piedmont/duplex.

Valentino et al. (1994) attempted to correlate the stratigraphy of early Paleozoic of the Manhattan Prong with equivalent rocks from the Maryland Piedmont (Table 1). This correlation with the inclusion of the early Paleozoic shelf edge as defined by Rodgers (1968) allows a palinspastic reconstruction of the Piedmont that requires 150 km of dextral offset (Figure 3). The total offset occurring along the PG-HVSZ proposed by Valentino et al. (1994) is certainly an oversimplification and in disagreement with the more detailed results presented here but it provides a first attempt at reconstruction.

The total amount of dextral offset in the Appalachian Piedmont Province during the late Paleozoic is far greater than these estimates. Gates et al. (1988) estimated that the minimum Alleghanian dextral offset across the Virginia to North Carolina Piedmont is 341 km. Clearly, there cannot be such a radical change in offset along strike without major structural expression. There is no such expression in the intervening Piedmont. This means that either the offset estimates for the Pennsylvania are too low or more likely, there are other major strike slip faults outboard of the Piedmont and buried under the Atlantic Coastal Plain as proposed by Maguire et al. (2009) based upon their radiometric dating.
An even more intriguing possibility for large dextral offsets comes from the Spechty Kopf complex of Pennsylvania-Maryland. This Devonian age complex resides in the Valley and Ridge Province but contains clasts of basement material. Dennis (2005) proposed that the clasts could not have been derived from local units but instead found them to be more likely to have been derived from rocks of the Inner Piedmont belt of the Carolinas. If this correlation is true, dextral strike slip offset during the late Paleozoic may have been more than 500 km. This magnitude of offset rivals that produced in the southeast Asian strike slip faults produced by tectonic escape generated by the Himalayan collision (Tapponier et al., 1982). Such large offsets could only be on strike slip faults now buried under the Atlantic Coastal Plain rather than those in the Pennsylvania Piedmont.

**Partitioning Of Strain**

The Piedmont may have had significant Late Paleozoic strike slip deformation but the Valley and Ridge Province to the west and north underwent purely contractional deformation with well developed folds, cleavage and thrust faulting. Faill (1998) estimates overthrusting in the Valley and Ridge to be 100 to 200 km which is as large or larger than the total dextral strike-slip offset in the Piedmont. There is a transition from dominantly strike-slip to dominantly contractional deformation in the western Piedmont near the Tucquan antiform (Valentino, 1999; Wise and Ganis, 2009) (Figure 4). The transition is marked by complex mixed modes of deformation and overprinting relations. Some minor strike slip faulting even extends into the Valley and Ridge where they are overprinted by reverse motions in some cases (Nickelsen, 2009).
There are two possible explanations for the distribution of strike-slip faults and related deformation (DSZ, PBZ, OSZ, PDSZ) to the southeast and thrust faults (Yellow Breeches thrust, Chickies-Oregon thrust) and related deformation to the northwest in the Susquehanna River area (Figure 4): they could be asynchronous or there could be strain partitioning. The overprinting of fault types by Nickelsen (2009) demonstrates that they could be asynchronous with strike slip first followed by thrusting. On the other hand, the cooling ages found by Maguire et al. (2006) and similar studies in the southern Appalachians (Gates and Glover, 1987) show that there must be some temporal overlap of strike slip in the Piedmont hinterland with thrusting in the Valley and Ridge foreland (Gates, 1988; Hatcher, 2001). This overlap supports large scale partitioning of strain.

Large scale partitioning of strain in an oblique plate tectonic collision was first documented in Pacific island arcs (Fitch, 1972) where oblique convergence is partitioned into orthogonal subduction at the trench and parallel strike slip faulting on the arc. This means that the oblique convergent motion is spatially resolved into vector components of reverse (contraction) and strike slip movement and deformation. Employing this mechanism for the Susquehanna traverse, a single oblique décollement to the east would have resolved into the transpressional deformation in the Piedmont and the thrust/contractional deformation in the Valley and Ridge as shown by Wise and Ganis (2009) (Figure 4).

On the other hand, most radiometric ages in the transpressional Piedmont are 300 Ma or older with only a few areas with ages in the 280-300 Ma range. In comparison, much of the Alleghanian overthrusting in the Valley and Ridge was Pennsylvanian to Permian (Faill, 1998) which is much younger. Although transpression may have continued on the orogen scale, it is likely that the transpression in the Pennsylvania Piedmont ended or was waning by the time major overthrusting began. The Pennsylvania Piedmont strike slip duplex would have been carried passively westward on a thrust sheet during the later part of the Alleghanian Orogeny.

References
Crowley, W.P., 1976, The geology of the crystalline rocks near Baltimore and its bearing on the evolution of the eastern Maryland Piedmont, Rept. of Inv. no. 27, Md Geol. Surv., 40 pp.

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Rodgers, J., 1968, The eastern edge of the North Am. continent during the Cambrian and Early Ordovician, Studies in Appalachian Geology, Northern and Maritimes, p. 141-149.


TECTONICS OF THE MESOZOIC GETTYSBURG AND OTHER BASIN REMNANTS OF THE BIRDSBORO BASIN

Rodger T. Faill

Introduction

The Gettysburg “basin” is not a separate, independently-formed Mesozoic basin. It is one of four erosional remnants remaining of a much larger (500+/- by 75+/- km), elongate sedimentary trough the Birdsboro basin (Faill, 2003a). Deformation late in the early Jurassic tilted, folded, and faulted this basin. Subsequent erosion through the remainder of the Mesozoic, and through the Cenozoic, has left four erosional remnants from New York to Virginia, the Newark, Gettysburg, Culpeper, and Barboursville remnants (Fig. 1).

Figure 1. Basin remnants of the once continuous early Mesozoic Birdsboro basin, consisting of the Barboursville, Culpeper, Gettysburg, and Newark remnants. The principal formations are shown for each basin remnant. Fanglomerates along the non-faulted, southeast margin occur at: C—Conoy Creek; H—Hanover; N—Norristown; P—Potomac River; and T—Thomasville (see text for detailed discussions); along the faulted, northwest margin: F—Fairfield.

The four remnants share a number of features. The sedimentary rocks are fluvial to lacustrine, ranging from conglomeratic sandstone to fine-grained argillite. The basin accumulated as much as 7+ km of sediment in the basin center, with progressively less toward the basin margins. The sediments were deposited in subhorizontal layers, but presently, the dips are generally to the

northwest, ranging from 5 to 60 degrees. As has long been noted, the northwest down dip margin is faulted along much of its length, presumably by a normal, steeply southeastwardly dipping faults. These features have given rise to the widely held idea that the basin remnants formed as half-graben, in which downward movement on “border faults” provided depressions that concurrently filled with sediment.

The half-graben model may seem, in a casual view, to be a simple and appropriate interpretation for the early Mesozoic basin remnants. However, many detailed geologic features within and surrounding the remnants are not consistent with, and even contradict, that simple model. Such features include fanglomerates adjacent to overlaps, post-depositional age of the faults, basin floor inliers adjacent to the “border faults,” steep bedding dips in Jurassic lacustrine beds, absence of diabase in the faults, among others. In short, a half-graben model is not an adequate paradigm.

I instead present a two-phase model of a non-faulted sedimentary trough (Birdsboro basin) of late Triassic and early Jurassic age, that filled with 7+ km of sediment and was subsequently deformed (faulted, rotated, and folded) late in the early Jurassic. This two-phase paradigm is preferred because it presents a more comprehensive and consistent explanation and history for these Mesozoic “basin remnants.”

The interpretation presented here is a two-phase model: The Birdsboro basin formed over an initial, early late Triassic crustal downwarp that was subparallel to the Alleghanian tectonic trends. Throughout the late Triassic and into the early Jurassic, this linear trough (from New York to Virginia) progressively deepened and widened, accumulating the 7+ km of terrigenous sediment derived from all sides of the basin. Late in the early Jurassic, a second, post-depositional phase of deformation occurred, during which the basin was faulted, folded, and rotated, which imparted the monoclinal dip to bedding. Subsequent erosion removed much of the deformed Birdsboro basin, leaving today the four basin remnants.

This is not a new idea. A two-phase history was proposed in the late nineteenth century by Davis (1888, 1897; Kummel, 1898); and later by Wherry (1913). It was only later, early in the twentieth century, that the two phases became merged into a single half-graben model (e.g., Russell, 1892; Barrell, 1915; Longwell, 1922, 1937; Bain, 1932; Kay, 1951; and many others).

The simplicity of the half-graben paradigm is beguiling. Students entering geology soon learn of it, and, absent any further exposure to the basins, generally accept it without question and it becomes the established paradigm. Yet, many of the basin features contradict the half-graben model (Faill, 1973, 1988). They include:

1) As much as one-third of the sediment entered the basin from the southeast, across the non-faulted margin.
2) The margin fanglomerates occur along non-faulted overlaps as well as faulted margins—they are not sufficient evidence of coeval faulting.
3) One-third of the northwest margin has been mapped as an overlap, not a “border fault.”
4) At several locations along the faulted, down-dip margin, the basin floor is exposed (as inliers), rather than being at several kilometers depth.
5) The faulting within and at the basin northwest margin was post-depositional, offsetting the youngest, early Jurassic sediments.
6) The early Jurassic igneous rocks do not intrude the faults; many of them are offset by the faults.
7) The youngest Jurassic lacustrine beds contain folds, faults, and a monoclinal dip (up to 60 degrees in the Virginia Culpeper remnant), indicating a late (post-depositional), basin-wide deformation.

The two-phase model avoids all these contradictions.
The Birdsboro Basin

The Birdsboro basin commenced filling early in the late Triassic (Olsen, 1997). It progressively deepened and enlarged to more than 500 km in length and 75 km in width during the late Triassic and early Jurassic (Faill, 2003a), accumulating up to 7 km (or even more) of fluvial and lacustrine continental sediments from both sides of the basin (McLaughlin and Gerhard, 1953; Glaeser, 1966; Klein, 1969; Lindholm et al., 1979; and Smoot, 1991). Perhaps as much as one half of this total came through numerous streams from southeast of the basin, forming a bajada (coalesced alluvial fans) along the southeast side of the basin. Most of the remainder entered the basin from the northwest, through one or more large river systems from which regional alluvial fans developed. The finer-grained fractions (sands and silts) that entered and passed through the upper bajada and the fan apices reached the lower, distal portions of the bajada/fans; the finest-grained fractions (silts and muds) settled even farther into the playas and lakes of the interfan areas (Faill, 2003a).

This pattern of basin filling, with sediment equally from both sides, is not consistent with a half-graben model. In that model, the rising footwall adjacent to the descending hanging wall is presumed to be the source of sediment filling the half-graben. Nothing in that model provides for a significant sediment source beyond the southeastern overlap margin—that area is presumably descending (to a lesser extent) as part of the hanging wall. Indeed, the presence of cobbles up to 10 cm in size in the local fluvial conglomerates near the base of the New Oxford Formation (Gettysburg remnant) indicates intermittent high-energy sediment-input environments along the southeast margin.

The Margin Fanglomerates

The local fanglomerate deposits along the Birdsboro basin margins are exceptions to this sediment dispersal pattern. They are poorly sorted and poorly bedded, and of limited areal extent, in distinct contrast, lithically and texturally, to the moderately well sorted, well-bedded conglomerates and conglomeratic sandstones in the alluvial fans and bajadas. Volumetrically, they constitute much less than 1 percent of the total sediment volume—they were not the primary mechanism by which the basin filled. However, the inordinate role they are given in the half-graben models for the Mesozoic basin remnants warrants some discussion.

The fanglomerates within the Birdsboro fans are distinctive, generally being diamicitic deposits exhibiting poorly developed, thick to very thick bedding. They are poorly sorted, containing pebble, cobble, and even boulder clasts in a sandy, silty, and muddy matrix. Their poor sorting indicates that the finer-grained fraction was not winnowed and carried farther into the basin as occurred with the fluvial deposits, but rather remained in the deposit at the margin. The poor bedding, coarse-grained texture, and poor sorting suggest are characteristics of debris flows rather than fluvial deposits (Lorenz, 1988, p. 115ff; Smoot, 1991). They entrained whatever regolith, talus, colluvium, and alluvium that had accumulated in valleys near the basin margins.

The fanglomerates of the Birdsboro basin (and its remnants) occur along both the partially faulted northwest margin, and along the overlap southeastern margin. Their presence along non-faulted margins indicates that they are not sufficient evidence of coeval faulting. An example at Thomasville, Pennsylvania (west of York) is illustrative.

Thomasville, Pa. (southeast margin)

The Thomasville Stone and Lime Company operates an underground mine at Thomasville, where the lower Cambrian Kinzers Formation dips gently to the northwest under the upper Triassic unconformity. They conducted an exploration program in the late 1960's to evaluate the Kinzers Formation under the Triassic basin north of the then current workings (Cloos and Pettijohn, 1973).
The first hole (A in Fig. 2), 50 m north of the contact, encountered the Paleozoic carbonates at 20m (65 ft), confirming that the unconformable contact, the basin floor, dips northward at about the same dip as the overlying Triassic beds.

Cloos, unsure of the identity of that Paleozoic carbonate, continued drilling. Seventeen meters (55 ft) farther, the drill encountered the Salterella zone, the distinctive fossil zone at the base of the Kinzers Formation. Interestingly, at this depth, this zone lay exactly where expected if it were projected northward from the underground mine (Fig. 2). In other words, no vertical offset of the Salterella zone exists between the underground mine and #1 Rife, which would rule out the presence of a fault.

The apparent offset of the Paleozoic carbonate floor must have another explanation. The offset may simply be a steep (more than 70°) erosional escarpment, a 200+ m high cuesta on the Kinzers Formation, created solely by late Triassic erosion prior to the deposition of the basal Carnian-age beds (Faill, 1973, 1988; Cloos, 1971, personal communication). The fanglomerate material in the lower part of the core represents colluvial accumulation and/or fluvial deposition (debris flow) of mechanically weathered detritus from this steep erosional escarpment. In other words, the basin floor had significant topography from which local deposits were derived.

Northwest Margin

The northwest margin of the Gettysburg remnant is widely believed to be faulted, but faults can be demonstrated for only one-half of their length. West of the Susquehanna River, Root (1977) described an overlap seen in outcrop along the Pennsylvania Turnpike, and stated that nearby shallow drilling indicates that the contact dips south about 12°. East of the river, much of the northwest margin has also mapped as an unconformity (Geyer (1970), Geyer et al. (1958, 1963), Gray et al. (1958), and MacLachlan et al. (1975). Overlap northwest margins have been mapped.
farther to the east, to central New Jersey, and farther southward, in Maryland and Virginia. The presence of unconformable overlaps along the northwest margin is distinctly at odds with the half-graben paradigm.

**The Inliers (basin floor exposures)**

Inliers are surface exposures of the basin floor within the basin remnants. The most prominent and well-known example is the Buckingham Mountain inlier, which lies in the center of the Newark remnant where crossed by the Delaware River. Its presence there is a consequence of the post-depositional movement on the Chalfont-Furlong-Flemington fault system. The Ridge, in the southern end of the Culpeper remnant, is another striking example.

However, other inliers exposing basin floor rocks lie adjacent to the northwest margin. Four are in the Culpeper remnant in Virginia and Maryland. A few more exist in New Jersey. In Pennsylvania, the most prominent one occurs at Fairfield, Pa.

This large (12 sq. km.) inlier exposes Ordovician (St. Paul Group?) carbonates and lies in fault contact (“border fault”) with the Neoproterozoic rocks of South Mountain to the west (Fig. 3). Shales, siltstones, and carbonate fanglomerates of the Triassic Gettysburg Formation (Fig. 4) surround and unconformably overlie the inlier on the other three sides (Stose and Bascom, 1929). This inlier raises the question of how can the basin floor be at the surface if it had descended some 5 to 7 km to accommodate the basin fill. Some published explanations are quite tortuous.

Stose (1949), recognizing the difficulties the inliers presented for the half-graben hypothesis, attempted to resolve it by proposing that the inlier did not really belong to the descending hanging wall, but was actually part of the footwall, and that the main "border fault" lay east of the inlier (his Fig. 3). He hypothesized that, late in the basin filling, a new, minor fault
developed west of the inlier, and only the youngest sediments, the fanglomerate deposits, covered this "hanging-wall horse" (the inlier).

**Figure 4. Cross section, A—A’, of the Fairfield inlier.** See Inlier text, and Figure 3, for discussion.

However, no corroborating evidence for a buried "border fault" on the east side of the inlier has been presented. First, fanglomerates are not restricted to the “youngest sediment”—they occur throughout the entire stratigraphic section (Faill, 2003a). Stose (1949) also suggested that diabase intruded the fault, thereby masking the fault, but no continuation of the fault has been mapped north of the diabase, and the diabase does not align with the “border fault” farther north. It appears that this hidden "border" fault on the east was proposed solely to resolve the contradiction between the half-graben paradigm and the exposure of inlier rocks.

**The Age Of The Faults**

The central thesis of the half-graben hypothesis is that the “border fault” along the down dip margin was active during basin filling. That is, the continuous (or episodic) movement on the fault provided a deepening basin for the accumulating sediment. Or, phrased another way, were the faults syndepositional, or were they superimposed on the non-faulted Birdsboro trough after accumulation of the sediment? Again, geologic relations in the Gettysburg remnant are instructive.

As mentioned above, the northwest margin of the Gettysburg remnant lies against South Mountain (Fig. 3)—it has been mapped as a “border fault” (Stose and Bascom, 1929), even though its trace is somewhat irregular. South of Fairfield, the south-southeast-trending margin (a fault) truncates much of the late Triassic section, including the early Jurassic diabase. This suggests a late, post-depositional fault movement. Northward from Fairfield, several large fanglomerates in the Gettysburg Formation lie adjacent to the margin (“border fault”). These fanglomerates contain quartzite clasts characteristic of the Chilhowee Group, which overlies the late Neoproterozoic Catoctin metavolcanics in South Mountain. However, rather than Chilhowee quartzites, older Catoctin rhyolites lie adjacent to the fanglomerates. This juxtaposition suggests a post-fanglomerate displacement, of as much as 2 to 3 km (Faill, 2008). Some rhyolite clasts are present in the fanglomerates, but they do not resemble chemically or texturally any of the rhyolites presently
exposed in that part of South Mountain (Robert C. Smith, 2002, personal communication). Direct evidence of any earlier (late Triassic) displacement on this fault has not been found.

Similar examples of post-depositional faulting on the “border fault” are present elsewhere along the northwest margin of the Birdsboro basin. Steeply dipping early Jurassic lake deposits are truncated by the “border fault” on the west side of the Culpeper remnant in Virginia. The Ramapo “border fault” (Ratliffe, 1971) along the northwest margin of the Newark remnant in central New Jersey truncates beds of the Sinemurian Boonton Formation, indicating that at least some of the fault movement was post-depositional. Splays from this fault die out down-section in Jurassic beds (Drake et al., 1996), suggesting that the movement on those faults was late Sinemurian or younger.

Additionally, if any faults within the Birdsboro basin were active syndepositionally, they would have affected sediment distribution within the basin, providing breaks in the basin stratigraphy. A good example is the Buckingham inlier in the Newark remnant to the east (see Inliers above). This large fault system offsets two structural blocks by as much as 4 km (Schlische, 1992), yet the remarkable similarity in the lithic and stratigraphic sections between the two blocks strongly suggests that this fault system was active only post-depositionally (Faill, 2005).

A fine-scale lithic cyclicity in the Lockatong and Passaic Formations (in the Newark remnant) signifies how tectonically inactive the Birdsboro basin was during the late Triassic and early Jurassic. These formations are dominated by a pervasive, elevation-sensitive, nested cyclicity (Van Houten, 1962, 1964; Olsen, 1986) that was strongly influenced by a subtle astronomic orbital cyclicity (Olsen et al., 1996b). The repetition and lateral continuity of lithologies, lithic sequences, and specific cycles on both sides (Olsen, 1988; Olsen et al., 1996a) of the Buckingham fault system implies a widespread depositional continuity and tectonic quiescence that would have been easily disrupted by any small elevation changes (such as produced by syndepositional faulting). That, and the absence of any local lithic changes adjacent to the faults, indicates that the faulting in this part of the Birdsboro basin was post-depositional (Olsen et al, 1989, p. 91).

Finally, drill cores through a pair of faults of the intensely mineralized Furlong fault zone in western New Jersey reveal that the local fault zone dips to the southeast at 47 to 50° (Ratcliffe and Burton, 1985, 1988). The truncation of the diabase intrusion by these faults, the mineralization of the diabase in the fault zone, the absence of mineralization in the adjacent dolomite, and the presence of a horse of Jurassic diabase between the two faults clearly indicate that the faulting and mineralization were post-intrusion.

The Igneous Rocks

Four discreet episodes of igneous activity (Puffer, 1992; Tollo and Gottfried, 1992) occurred over a period of 600,000 years just after the beginning of the Jurassic period (Olsen and Fedosh, 1988; Olsen et al, 1996b; Hames et al, 2000). These continental tholeiites were emplaced in three different forms: basalt lava flows; intrusive, largely bed-concordant diabase plutons (sheets or sills); and subvertical diabase dikes. The narrow compositional range of the magmas in each episode indicates that all the forms taken by a particular magma were coeval (Smith et al., 1975).

The surprising aspect of the igneous intrusions is that the magma did not intrude any of the faults. Indeed, the very opposite is true—the faults truncate the plutons, dikes, and flows of all the magma episodes (e.g., Houghton et al., 1992), which clearly makes them post-igneous events. Even the Ramapo fault and its branches truncate the few small diabase bodies in western New Jersey (Drake et al., 1996). The absence of any intrusion into these faults strongly suggests that they were post-intrusion in age.
The Deformation

The deformation of the Birdsboro basin took several forms—the monocline, the folds, and the faults. All of these structures developed post-depositionally, probably late in the early Jurassic (Faill, 2003b).

The monocline

All of the beds throughout the Birdsboro basin dip gently to moderately to the northwest, north, or west. Dips range from gentle (a few degrees) to moderate (>40°) to the northwest and north. Bed dips do not decrease systematically across the basin, as would be expected if the tilting progressed with sedimentation (Faill, 1973, Plate 2). Indeed, the steepest dips (60 degrees) in the Culpeper remnant occur in the Jurassic lacustrine rocks along the northwest margin (Froelich et al, 1982). Their rotation from horizontal must have been post-depositional. Paleomagnetic data confirm that the beds were tilted after deposition, during the folding (Volk, 1977; Witte and Kent, 1991; Witte et al., 1991; Kodama et al., 1994).

The folds

Moderately open folds are present in the Birdsboro basin, especially along the northwest margin, superimposed on the basin-wide monoclinal structure. The most prominent folds lie adjacent to the northwest margin of the Newark remnant, plunging gently toward the margin at a large skew to the basin trend (Faill, 1973; Schlische and Olsen, 1988, Schlische, 1992; Drake et al., 1996). They occur in the youngest Triassic (Norian) and Jurassic rocks, and thus they formed after deposition in the basin.

The Jacksonwald syncline along the Schuylkill River south of Reading, Pa. (in the Newark remnant) is the most prominent fold in the basin. Several aspects of the fold indicate a post-depositional development. The lava flow high in the section is folded as much as the older sediments (MacLachlan, 1983); the folding rotated clastic dikes (Lucas et al, 1988); and the axial-planar pressure-solution cleavage does not fan about the fold (Lucas et al, 1988). In addition, the early (syndepositional) high-temperature (C-component) magnetization passes the fold test (thus predates the folding, Witte et al, 1991), whereas the post-monoclinal, lower-temperature (B-component) magnetization is associated with the folding (Witte and Kent, 1989, 1991; Kodama et al., 1994). Paleomagnetic data elsewhere in the basin (Volk, 1977) also indicates that the folds developed after the intrusive phase.

The faults

None of the faults in the Birdsboro basin were syndepositional, because no sedimentary effects of coeval faulting (such as abrupt lateral changes in lithology or thickness) have been identified. The absence of diabase in faults and the common truncation of igneous bodies by faults clearly indicate a post-intrusion age for the faults. The truncation of folds containing early Jurassic rocks suggests that the faults are post-depositional as well.

A Two Phase History For The Birdsboro Basin

Examination of many of the aspects of the Birdsboro basin and its present-day remnants demonstrate that a single-phase origin of these Mesozoic subbasins as half-grabens is not tenable in the least. The large volume of sediment from southeast of the basin, the presence of fanglomerates on both sides of the remnants, the overlaps along the northwest margin, and the basin floor as inliers along the “border faults” all contradict the half-graben model. The absence of sedimentary effects of coeval faulting, the presence and lateral dying-out of faults in the youngest, Jurassic beds, the absence of igneous rocks in any of the faults, and the involvement of all of the basin sediments, Norian to Sinemurian, in the monoclinal structure, faults, and folds Indicates activity that was
necessarily post-depositional. Post-depositional deformation is not part of the half-graben paradigm. For these reasons, I turn to a two-phase history for the Birdsboro basin and its remnants that was first proposed by Davis (1888).

Not much is known about the post-tectonic central Appalachian Alleghany orogen that lay within the Pangean continent. Presumably, the mountain system formed during the Permian Alleghany orogeny was gradually eroded during the early and middle Triassic (Slingerland and Furlong, 1989; MacLachlan, 1999). The paleosurface of where the Birdsboro basin would form (northwest of the orogenic core) probably sloped to the northwest away from the mountains. This area possessed a moderate relief and a drainage system for transporting the erosion products from the higher lands.

Sometime during the Triassic, perhaps late in the middle Triassic, mantle convection formed an upwelling plume beneath Gondwana/Laurentia suture that began a northwest/southeast crustal extension. The surface response was the development of several basins elongated parallel to the orogen. The Birdsboro was the northwesternmost in the (future) mid-Atlantic region. Sediment accumulation in the Birdsboro basin began with the capture of the formerly through-going drainage from the southeast in the early Carnian (~230 Ma, Olsen, 1997).

With time and basin subsidence, rising source areas northwest of the basin began contributing sediment, which accumulated in the regional alluvial fans and their lateral equivalents, the playas. This pattern continued throughout the remainder of the Triassic. Just after the beginning of the Jurassic (~200 Ma), a short period of tholeiitic magma intrusions and effusions occurred in the basin, and beyond its margins. However, sedimentation continued after this igneous episode. By the Sinemurian (~190 Ma), at least 7 kilometers of sediment had accumulated in the deepest parts of the Birdsboro basin over a period of 40 m.y.

It is not known how long sediment accumulation continued in the Birdsboro basin beyond the Sinemurian. In that the opening of the Atlantic Ocean was the most singular event at this time, it seems plausible to attribute the deformation to the crustal separation of Africa from North America. The presence of Aalenian (~175 Ma) sediments on the U.S. continental rise (Poag and Sevon, 1989) suggests that the basin deformation had occurred by then.

Continued crustal extension enhanced the basin subsidence and caused increasing brittle behavior in the crustal rocks, that finally lead to extensional faulting. However, it was during the actual splitting of Pangea (at 180 to 175 Ma?) some 300 km east of the Birdsboro basin that produced the deformation. Now released of extensional strain, the (by now) North American continental margin rebounded northwardly, augmented by the newly initiated ridge push in the Atlantic. This northwestward compression produced a crustal arching (?) which passively rotated entire basins, imparting to them the pervasive monoclinal structure, accompanied by the folding and faulting.

Subsequent erosion through the remainder of the Jurassic, and much of the Cretaceous, beveled the rotated Birdsboro basin sufficiently for the southeastern edge (in Pennsylvania) to be overlapped by Coastal Plain sediments during the Late Cretaceous. Today, we see merely the remnants of what was once a large, elongate sedimentary basin.

Implications

The Mesozoic tectonism has affected the underlying Paleozoic structures simply by being a later event in the same area. The tectonism of the Birdsboro basin was primarily a vertical movement, without any allochthony, so the geographic location of earlier structures was not changed. But their present depths and associated burial metamorphism were changed. The entire basin floor was at the surface at the beginning of the late Triassic. These areas descended with time,
to as much as 7 km, some of which remains at depth to the present day. Other areas, such as along the present overlap contacts, including the inliers, were buried and subsequently uplifted back to the surface. The specific movements of each part of the basin remnants and the areas outside the remnants are a direct function of the model for the Birdsboro basin. Which paradigm best fits the Mesozoic basins is important to the interpretation of the tectonics in the older rocks.

If the single-phase half-graben model applies, then the rocks under and southeast of the basins have been rotated northwestward. The rocks northwest of the basin have also been rotated northwestward with the rise of the footwall, being greatest at the “border fault”, and progressively less away from the fault. In other words, all of the pre-Triassic rocks under and next to the basins now have orientations different than at the end of the Permian. Those structures (thrust faults, folds, cleavages, nappes) that were northwest dipping are now more steeply northwest dipping. Those that were originally southeast dipping are now dipping less to the southeast, or, if they originally had a low dip, may now be dipping gently to the northwest.

On the other hand, if the two-phase basin filling/later deformation model applies, then only those rocks southeast of the basin and under the southeast part were rotated northwestward. Those northwest of the basin, and under the northwest part, were rotated southeastward. Whereas the two models had similar effects southeast of the basin, the rotations were opposite on the northwest side. This difference is particularly significant in the interpretations of the Reading Prong, the Taconic nappes, and even the South Mountain structures.

The early Jurassic structures (monocline, normal (mostly) faults, and open folds) may seem to be simple structures when compared with the underlying Paleozoic tectonism. Yet the size of the Birdsboro basin (500+/− by 75+/− km) indicates that these structures affected much of the underlying crust, especially the Paleozoic structures under and near the “basin remnants.” Very little attention has been devoted to how much the earlier Paleozoic structures were rotated and offset by the subsequent early Jurassic deformation.

References
TECTONICS AND TOPOGRAPHY OF THE CENOZOIC APPALACHIANS

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Introduction

The Appalachian Mountains in the eastern United States (Fig. 1) are an old orogen whose crustal root was grown and modified during a long period of Paleozoic collisional tectonics, and then thinned by erosion and stretching associated with the opening of the Atlantic Ocean. Evolution of the Appalachian mountain landscape has engendered influential tectonic (Williams, 1978; Boyer and Elliott, 1982; Rodgers, 1987; Faill, 1997a, b; 1998; Hatcher et al., 2007) and geomorphic (Davis, 1889; 1899; Hack, 1960) paradigms and continues to shape thinking as new data and analytical techniques are able to increasingly quantify processes and rates (Pazzaglia and Gardner, 1994; Pazzaglia and Brandon, 1996; Granger et al., 1997; 2001; Reusser et al., 2004, 2006). The Appalachians are one of the few decay-phase orogens where a wealth of erosional data has already been assembled from thermochronometric (Hulver, 1997), river incision (e.g. Mills, 2000), and most recently cosmogenic (e.g. Matmon et al., 2003) methods.

The wealth of geologic, geomorphic, and geophysical data accumulated for the Appalachians make them an excellent case study in long term erosion, persistence of mountainous topography, and coupling between surface and tectonic/ isostatic/ eperiogenic processes in an ancient, decaying orogen. The landscape of the Pennsylvania Piedmont harbors important geomorphic data that bears on these diverse processes.

For the Piedmont and the Appalachians in general there are six key observables that must be reconciled to explain its post-orogenic (postTriassic) evolution. First, the metamorphic core of the range currently lies at the lowest mean elevation, and much of it is below sea level, covered by a coastal plain. What is commonly referred to as "the Appalachians" are really what is left of the former Appalachian foreland, now topographically inverted. Second, a long-term record of unroofing exists in the form of siliciclastic sediments in Atlantic margin shelfslope basins. These sediments argue for unsteady exhumation driven by epeirogeny, climate, or both (Pazzaglia and Brandon, 1996) although there is some ambiguity regarding the precise provenance of the basin material. The most recent pulse of sediment delivered to the shelf-slope basins arrived in the middle Miocene and high sediment rates persist through the Quaternary. Third, the longitudinal profiles of Atlantic slope rivers increase in gradient towards the Atlantic, forming a zone of seemingly anomalous rapids (Fall Zone) near the coast. Furthermore, the rivers of the midAtlantic States possess knickpoints upstream of the Fall Zone of common elevation that have no apparent relation to rock-type or structure. Fourth, the divide between Atlantic slope and Ohio drainages is not a static feature; rather, it has a long-term history of unsteady translation westward. Fifth, Appalachian topography is locally rugged, with lofty, steep mountains and deep, narrow canyons. The presence of mountainous topography in a foreland that has experienced kilometers of erosion over 180 m.y. is enigmatic. Lastly, thermochronologic, river incision, and emerging cosmogenic data that are increasingly quantifying Appalachian erosion rates show surprising correspondence across long-time scales despite variable tectonic histories, eustasy, rock-type, and relief, but discordance over shorter times scales.

In this paper, we focus on geomorphic (river incision) preliminary thermo-chonometric, published cosmogenic, and published sediment yield data that quantify Cenozoic erosion of the Appalachians.
across a spectrum of time and space scales. These erosion and incision rates and their (un)steadiness constrain geodynamic models responsible for post-orogenic rock uplift and its interaction with surficial processes.

**Appalachian Thermochronology**

Thermochronology is the study of apparent or real changes in the temperature of the crust as recorded by different mineral phases, the presence or absence of crystal structure damage due to radiogenic alpha-decay, or the parent-daughter ratios of radiogenic elements bound within the mineral. Rocks cool as they move from deep in the crust to the surface. Radiogenic damage to crystals or decay products are annealed or lost, respectively, when the mineral is hot and buried deep in the crust. At cooler temperatures, crystal damage and decay products are retained and their accumulation in different mineral phases can be used to reconstruct the time-temperature path of the mineral approaching the land surface. The closure temperature of a thermochronometer is the temperature of the rock at the time corresponding to its apparent age.

Thermochronology data interpreted as rock uplift carries the implicit assumption that the uplift of rocks above regional results in erosion, and thus mineral cooling, at a closely balanced rate.

The most recent cooling due to erosion in the Appalachians may be realized by a low-temperature thermochronometer, such as the Apatite U-Th/He (AHe). The AHe system is based on the stepwise alpha decay of $^{238}\text{U}$, $^{235}\text{U}$, and $^{232}\text{Th}$ to stable Pb isotopes and the quantitative accumulation of the produced $^{4}\text{He}$ in the mineral gram.

**Results**

**Bedrock AHe cooling ages from Pennsylvania and New Jersey**

Pennsylvania and New Jersey U-Th/He cooling ages ($n=31$) range from $\sim60$ - $240$ Ma and show different pooled ages for the Ridge and Valley ($\sim198$ Ma), Blue Ridge ($\sim98$ Ma) and Piedmont ($\sim172$ Ma) provinces (Fig. 2). The Blue Ridge cooling ages are more tightly clustered and do not

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*Figure 1. Digital shaded relief of the Pennsylvania Piedmont and surrounding region. Inset map shows location with the high topography of the Appalachians (light shaded) and the shelf-slope basin (BCT, dark shaded). The dashed line in the inset map is the shelf-slope break at a bathymetry of $\sim200$ m. The rivers sampled in New England for AHe data are also shown in the inset map. Physiographic and geographic features on the colored map are: FZ = Fall Zone, CP = Coastal Plain, P = Piedmont, NB = Newark Basin, BR = Blue Ridge, GV = Great Valley, RV = Ridge and Valley, PP = Pocono Plateau, AP = Allegheny Plateau, PI = pre-Illinoian glacial margin, W = Wisconsinan glacial margin. The X-X’ cross section is illustrated in Figure 4 and 14.*
have any overlap with the Piedmont or Ridge and Valley samples that exhibit considerable overlap in their ages. Ridge and Valley and Piedmont AHe mean cooling ages are younger than corresponding AFT (Roden and Miller, 1989) and ZFT (Kohn et al., 1993) cooling ages respectively; however, there is overlap in the cooling age ranges for these different thermochronometers.

Assuming a uniform geothermal gradient of ~20°C/km for all of Pennsylvania and New Jersey over the cooling history of these samples, it appears that the Ridge and Valley and Piedmont provinces cooled through 70°C in the Early to Middle Jurassic and the erosion rate since that time has averaged ~16 mlm.y. (3000 m in 185 m.y.). For a similar geothermal gradient, the Blue Ridge has been eroded more rapidly at ~31 mlm.y. (3000 m in 98 m.y.).

**Apatite Single Grain Ages of New England Alluvial Sediments**

Single grain cooling ages (n = 120) from the Connecticut and Merrimac rivers are plotted as relative probabilities that include the analytical uncertainty and the error related to the alpha correction (Ludwig, 2003; Fig. 3). Detrital apatite (U-Th)/He ages of the Merrimac river basin range from the Paleozoic to Late Cretaceous (400-65 Ma) with a major peak of Cretaceous ages ranging from 130-80 Ma (average peak 104±9 Ma; n=46) and a small population of Middle to Late Jurassic ages (165-150 Ma; n=6). The Connecticut River shows a more complex distribution with ages ranging from the Late Paleozoic to Cenozoic (280-15 Ma). Similar to the Merrimac sample, the major age peak is Cretaceous with ages ranging from 130-80 Ma (average peak 107±14 Ma; n=36). There is good overall agreement between the bedrock AHe and alluvial AHe data in New England.

![Figure 3. AHe alluvial single grain cooling ages from New England.](image)
Assuming a uniform geothermal gradient of ~20°C/km for all of New England over the cooling history of these samples, it appears that New England mostly cooled through 70°C in the Late Cretaceous and the erosion rate since that time has averaged ~ 28 \text{ m}^3/\text{m}^2/\text{y}. (3000 m in 106 m.y.).

\textit{Shelf-slope basin sediment}

Actively subsiding shelf-slope basins, such as the Baltimore Canyon Trough (BCT), formed on the Atlantic margin and have been an effective trap for detritus transported by Atlantic slope rivers (Fig. 4a). The long-term depositional and subsidence history of the BCT, a syn- and post-rift lower-plate passive-margin basin 400 km long, 100 km wide, and up to 18 km deep has been particularly well studied (Karner and Watts, 1982; Poag, 1985, 1992; Poag and Sevon, 1989; Steckler et al., 1988, 1999). Collectively, the equivalent average of about 4 km of rock removed from an area spanning the modern central and New England Atlantic slope are preserved in the BCT, a thickness that compares favorably to Appalachian thermochronology data (Hulver, 1997). The sedimentation history of the BCT is summarized as rapid siliciclastic sediment fluxes in the Jurassic, several distinct pulses of siliciclastic sedimentation in the Cretaceous separated by periods of carbonate deposition, slow siliciclastic sediment fluxes throughout the Paleogene, and a large pulse of siliciclastic sediment flux beginning in the Miocene and continuing to the present (Fig. 4b). No less than 1.1 km of rock, spread evenly across the planimetric area of modern Atlantic drainage basins, is needed to account for the Miocene to present volume of sediment in these offshore basins (Braun, 1989).

\textbf{Figure 4. (a) Cross-section of the Baltimore Canyon Trough} (see location in Figure 1) showing major lithostratigraphic packages of sediment.

\begin{itemize}
  \item J = Jurassic,
  \item K = Cretaceous,
  \item eT = Early Tertiary,
  \item M = Miocene,
  \item P-Q = Pliocene-Quaternary.
\end{itemize}

\textbf{(b) Integrated flux of sediment to the BCT and erosion rate} considering a fixed contributing basin of area equal to the modern Atlantic slope catchment.

\textit{Cosmogenic erosion rates}

Terrestrial cosmogenic nuclides (TCN) such as \(^{10}\text{Be}\) have been used to measure the watershed-scale, average rate of erosion (Granger et al., 1996). The technique operates on the principle that rocks and soil accumulate in-situ production of \(^{10}\text{Be}\) as they lay exposed on hillslopes. These materials are delivered to channels where the quartz sand fraction contains an integrated inventory of \(^{10}\text{Be}\) that was produced within the upper 0.6 m of the soil profile throughout the drainage basin that is proportional to the residence time in the landscape. Rapid erosion of hillslope soils leads to small alluvial \(^{10}\text{Be}\) concentrations whereas slow erosion leads to high concentrations.
Recent results from TCN inventories in Appalachian alluvial sediment report mean, basin wide erosion rates of 27 m/m.y. for the Smoky Mountains (Matmon et al., 2003), 9 - 22 m/m.y. for the somewhat lower elevation and lower relief landscapes of the Pennsylvania Piedmont, Ridge and Valley, and Allegheny Plateau (Reuter et al., 2005), and 4-14 m/m.y. for steep slopes underlain by resistant rock types in the Blue Ridge (Duxbury et al., in press). Erosion rates of bedrock determined by exposure-age dating indicate even slower rates of erosion for uplands at or near drainage divides south of the glacial boundary (2.5-5 m/m.y., Reuter 2005; 2-8 mm/y, Hancock and Kirwan, 2007). In comparison, erosion of Appalachian ridge tops near the glacial boundary by periglacial processes, not determined by TCN methods, may be as high as -80-100 m/m.y. (Braun, 1989). Fluvial quartz cobbles embedded in Piedmont colluvium located in an upland, unincised setting at an elevation of 152 m at Black Baron, PA (see below) have an average exposure age date of -100 k.y. (Bierman, unpublished data).

**Sediment yield erosion rates**

Cenozoic erosion of the Appalachians has been reconstructed from suspended sediment loads, although it is important to point out that particularly in slowly eroding landscapes these data are known to be affected by relatively shortterm climate factors, human land use activities, and sediment storage (Wilkinson and McElroy, 2007). The suspended sediment yield is documented for the reservoirs of the lower Susquehanna that have sequestered decades of sediment, much of it the result of agricultural runoff and coal mining activity throughout the Susquehanna watershed. Calculated for a short 5 year period from 1985-89, the average annual suspended sediment yield of the Susquehanna River was 2,812,320,000 kg (3,100,000 short tons; Ott et al., 1991; Langland and Haney, 1997), which has a density-corrected equivalent volume of 0.001082 km" (1,081,661 m"). This volume distributed evenly throughout the 71,225 km2 watershed yields an average surface lowering rate of 0.015 mm/yr or 15 m/m.y. Other historic river suspended sediment yields suggest similar erosion rates between 20 - 40 m/m.y. (Sevon, 1989; Milliman and Syvitsky, 1992; Conrad and Saunderson, 1999). Carbonate dissolution rates typically range between -8 and 30 m/m.y. (White, 1984, 2000). Solute loads for the Piedmont region translate to a landscape lowering rate of -5-10 m/m.y. for mineral dissolution (Cleaves et al., 1970, 1974; Pavich et al., 1989).

**Susquehanna River incision**

The Appalachians are traversed by steep bedrock streams, many of which have convex longitudinal profiles set in deep gorges in their lower reaches. This is particularly true for Atlantic Slope rivers of the middle Atlantic states such as the Susquehanna. Recent advances in Quaternary geochronology and biostratigraphy of the Coastal Plain (Pazzaglia et al., 1997) have allowed for geomorphic markers such as river terraces to be dated and incision rates to be reconstructed. In summary river incision rates range from 5-40 m/m.y. (Pazzaglia and Gardner, 1994; Granger et al., 1997, 2001; Mills, 2000; Ward et al., 2005) with some arguments forwarded that incision rates have been increasing in the late Cenozoic (Mills, 2000).

Flat uplands flanking the river within 1-2 km from the gorge lip through the Piedmont reach are locally mantled by a rounded gravel lag representing the remnants of strath terraces (Fig. 5). Locally, the fluvial gravels are still in place, or nearly in place as they are cemented by hematite. Pits dug at gravel locations, such as at Black Baron, PA, showed that they are part of a matrixsupported colluvial deposit that is ubiquitous on the Pennsylvania Piedmont (Pazzaglia and Gardner, 1993). The gravel is found throughout at least 2-3 m of colluvium that lies atop saprolite and bedrock. There are three upland gravels Tg1, Tg2, and Tg3 that have similar texture with the possible exception of Tg1 which tends to be coarser grained and perhaps a bit more angular. Compositionally, the upland gravels are dominated by resistant rock types such as vein quartz, quartzite, and quartz sandstone; however, the amount of quartz sandstone increases with decreasing
stratigraphic age. Tg1, Tg2, and Tg3 are petrographically distinct enough to use clast counts as a means of correlation along the river. They are also much less heterolithic than QTg which is only locally preserved through the Piedmont reach, typically at the confluence of major tributaries.

Figure 5. Schematic cross-section illustrating Tertiary and Quaternary terraces preserved along the lower Susquehanna River.

Petrologic criteria are used to correlate the strath terrace remnants throughout the Piedmont and tie them into biostratigraphically-dated Coastal Plain fluvial and marine deposits (Pazzaglia and Gardner, 1993; 1994; Pazzaglia et al., 1997). The reconstructed terrace longitudinal profiles mimic the long profile of the Susquehanna River through the Piedmont, but tend to be steeper towards the Fall Zone and more gentle towards the Great Valley, an observation first noted by Campbell (1929, 1931). Terraces Tg1, Tg2, and Tg3 project downstream into Fall Zone upland gravel deposits intersecting the Bryn Mawr Formation that unconformably overlies Piedmont bedrock and the Cretaceous Potomac Group at an elevation of ~ 1 00 m (Fig. 6). The base of the Bryn Mawr Formation is thought to be ~ 12 my old based on preserved palynofloras (Pazzaglia et al, 1997) and petrographic down-dip correlation to the Choptank Formation that has a well known Serravallian marine fauna. The top of the formation is thought to be as young as 6 Ma based on downdip correlation to the upper part of the Chesapeake Group and along-strike correlation to the

Figure 6. Long profiles of the lower Susquehanna River and river terraces through the Pennsylvania Piedmont.
biostratigraphically-dated Brandywine Formation in Maryland (McCarten et al., 1990). Based on these and related data Tg1, Tg2, and Tg3 have been assigned ages of 12, 10, and 6 m.y. respectively. QTg can also be traced downstream all the way to the Coastal Plain where it projects both in elevation and in composition to the upper part of the Pliocene Pensauken Formation, which fixes its age to the late Pliocene - early Pleistocene (~2.5 - 1.0 Ma).

Pleistocene terraces are similarly correlative from tidewater to the glacial margin and well preserved along some reaches of the Susquehanna River (Peltier, 1949; Engle et al., 1996). Particularly rich flights of Pleistocene terraces are known from Washington Boro and Marietta where at least six Pleistocene terraces (Qt1-6) and QTg are preserved. Water well data and former gravel pits confirm that treads Qt1 through Qt6 are underlain by several meters of stratified sand and gravel. Terraces Qt1 through Qt6 are distinguished compositionally by containing granite and gneiss, rock-types not found in the upper Susquehanna basin that could only have been introduced by glaciation. For this reason, Qt1 through Qt6 are generally held to be genetically related to glacial outwash, which helps establish their ages. Pleistocene incision rate for the river as it enters the Piedmont reach is ~ 18 - 50 m/m.y.

The deepest, most narrow section of the Susquehanna Piedmont reach is the Holtwood gorge. Here, Tg1, Tg2, Tg3, and QTg lie 136, 104, 70, and 36 m respectively above the channel. Based on the ages of these terraces determined from their stratigraphic correlation to the Coastal Plain, the late Miocene-early Pleistocene incision rate was a relatively stable ~9 m/m.y. (Fig 7). The incision rate since the carving of the QTg strath is faster and in the range of ~15-36 m/m/y depending on the precise age of the strath.

The age of Pleistocene straths can be determined for the Holtwood gorge area where the terraces lack an alluvial mantle using TCN exposure age dating (Reusser et al., 2006). These straths are intimately related to the numerous islands that characterize the bedrock floor of the Susquehanna channel and it is generally assumed that the results from the Holtwood area apply elsewhere, including the islands near Safe Harbor. Exposure ages modeled from $^{10}$Be activities indicate that fluvially eroded bedrock surfaces within Holtwood gorge increase predictably in age with height above the channel floor, and that all are late Pleistocene features. The highest well preserved terrace (level 3) yields a mean exposure age of 36.1 ± 7.3 ka (n=14). The middle and lowest terraces, levels 2 and 1, yield mean exposure ages of 19.8 ± 2.7 ka (n=20) and 14.4 ± 1.2 ka (n=10), respectively. Two samples collected from heavily weathered and eroded topographic high points (LR-01 & LR-43), standing >20 meters above the channel floor, yield model ages of 97.1 ± 10.5 ka and 84.5 ± 9.1 ka respectively; the removal of rock and the associated cosmogenic nuclides by weathering and erosion means that these surfaces could be far older than their model exposure ages suggest.

The incision rates indicated by these data show that the Susquehanna River has increased its rate of down cutting at Holtwood in the past 100 6 ky to an average rate of ~255 m/m/y.

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**Figure 7. Susquehanna River incision rates at the Holtwood gorge.** Inset graph shows detail in the incision rates for the past 2.5 m.y
Incision rates in the Holocene have slowed considerably to 70 m/my, but that value is still nearly seven times faster than the long-term average suggested by the higher, older Tg and Qtg terraces.

**Susquehanna bedrock channel erosion and knickpoints**

The bedrock channel of the Susquehanna River features impressive erosional forms that are commonly associated with knickpoints and attest to the local ability of the river to lower its bed at a rapid rate. At low water, the Susquehanna channel is revealed to be a relatively flat wide strath generally lacking an alluvial cover, with not more than 1-2 m of local relief. At six locations tucked up against the eastern wall of the Piedmont gorge are narrow, elongate, intensely-potholed spoon-shaped "deeps" (Mathews, 1917). The floor of the deep at Holtwood is over 40 m below the water surface which places it at approximately 6 m below sea level. The deeps and smaller, related bedrock erosion forms like potholes have been used to argue for large, perhaps catastrophic floods in the lower Susquehanna River during the Pleistocene (Thompson, 1985; 1990; Shaw, 1989; Kochel et al, 2009) in part because these features scale with similar features associated with the scabland floods of the Columbia River. Alternatively, the potholes and the deeps are not relict, but rather modern features actively being carved by the river. The presence of similar features on rivers like the Potomac or tributaries to the Susquehanna River that never experienced glaciation in their headwaters argues against catastrophic floods. Furthermore, the Susquehanna River has generated large historic discharges such as hurricane Agnes (> 28,000 cms; \(10^6\) cfs), and spring rain-on-snow events in the past 20 years have generated ice-chocked discharges in excess of 25,000 cms (900,000 cfs).

The entire Piedmont reach of the Susquehanna River is a broad knickzone that attests to late Cenozoic base level fall (Figs. 6, 8). The timing of the initiation of that base level fall is best indicated by the middle-late Miocene age of the Bryn Mawr Formation and Tg terraces in the Piedmont. The base level fall signal is propagating upstream and into the Piedmont tributaries separating a lower, steep, bedrock channel gorge from an upper, gentle, wide valley.

Further upstream, there are other prominent knickpoints in the Susquehanna channel and its major tributaries, that argue for older base level falls still working their way towards the watershed divide (Fig. 8). The knickpoints are objectively identified using the stream-length index (Hack, 1973) that normalizes local channel gradient to the total length of the stream, measured from the divide.

The SL index identifies four knickpoints at ~140, ~160, ~300, and ~400 m preserved on the Main Branch, Sinnemahoning Branch, and Juniata River (Fig. 8). These knickpoints have no obvious relationship to a durable rock type, contact between a durable and soft rock, or structure. In fact, several of these knickpoints are located in soft rocks in otherwise flat strike valleys.

**Discussion**

**Steadiness of Appalachian Landscape Development**

The geologic, geomorphic, and thermochronologic data presented here indicate that postorogenic erosion and incision has been unsteady, but coincidentally, the long-term erosion rate averaged over 100 m.y. is similar to the short-term Quaternary rates averaged over < 2 m.y. (Fig. 9). The Appalachians have experienced slower than the long-term average rates of erosion, indicated by gentle graded segments of river long profiles and little sediment delivery to the BCT interspersed with pulses of more rapid than average erosion, indicated by river knickpoints and abundant sediment flux to the BCT (Fig. 10).
The pace of landscape change quickened in the late Cenozoic as evidenced by the large volume of sediment delivered to the BCT and the steepening of the lower segments of Atlantic slope river profiles. Increases in river incision rates in the Pleistocene is a global phenomena commonly associated with the onset of Northern Hemisphere glaciation (Pazzaglia, in press); however, the two- order of magnitude increases for Appalachian streams is so dramatic that a base level fall is very likely contributing to the incision. Some studies have suggested that several tens of meters of late Cenozoic base level fall could be the result of eustatic lowering since the middle Miocene (15 Ma; Pazzaglia and Gardner, 1994). More recent studies demonstrate that middle Miocene sea levels were indistinguishable from modern Holocene levels (Miller et al., 2005) which places all of the base level fall on rock uplift above regional. The actual Pleistocene average sea level is ~ 74 m below Holocene sea level, a result of long periods of glacio-eustatic drawdowns. But the knickpoints formed during those drawdowns have not progressed beyond the Fall Zone as evidenced by the buried Pleistocene river valleys in the Coastal Plain (Colman et al., 1990) and the estuarine mouths of all of the major mid-Atlantic rivers such as Chesapeake Bay and Delaware Bay. Gorges, convex river profiles, and incision rates of ~250 m/m.y. are here interpreted to indicate 100-200 m of post-Middle Miocene epierogeny, several models of which are described below. If the river incision rates are to be taken literally, most of this epeirogenic rock uplift is occurring now, in the Pleistocene, having accelerated from initially slower rates in the middle Miocene. Unfortunately, the resolution of AHe thermochronology cannot detect such a recent pulse of rock uplift. More detailed studies of suspended sediment or TCN-based erosion rate data may yet corroborate the river incision data and recent increase in the pace of landscape change.

Figure 8. Long profiles of the (a) Susquehanna River and its major tributaries, (b) the Main Branch above Westport, and (c) the Juniata River. The black profile are the raw data extracted from a 90-m DEM, the blue line is a lowest filtered profile to remove high frequency noise, the red line is the Hack SL-index (Hack, 1973). The shaded rectangles indicate knickpoints. Rock-types: MDhm=Huntley Mountain Fm, DSkm = Keyser through Mifflintown Fm undivided, Dtr = Trimmers Rock Fm, Pcg = Conemaugh Group, Pa = Allegheny Fm, Mb = Burgoon Fm, Desc = Sherman Creek Mbr, Catskill Fm, Dh = Hamilton Group, St-Obe = Tuscarora through Bald Eagle Fms, Swc = Wills Creek Fm, Sc = Clinton Group. An asterisk indicates possible rock type control on the knickpoint.
Figure 9. Comparison of erosion and incision rates over a range of time and space scales. 1 = AFT Ridge and Valley (Roden and Miller, 1989), 2 = AHe Ridge and Valley (this report), 3 = AHe PA-NJ Blue Ridge (this report), 4 = AHe PA Piedmont (this report), 5 = ZFT PA Piedmont (Kohn et al., 1993), 6 = Alluvial AHe New England (this report), 7 = Pazzaglia and Brandon (1996), 8 = Pazzaglia and Gardner (1993) and Reusser et al. (2006), 9 = Smokies (Matmon et al., 2003), 10 = PA Allegheny Plateau (Reuter, 2005), 11 = PA Ridge and Valley (Reuter, 2005), 12 = PA Piedmont (Reuter, 2005), 13 = Shenandoah N.P. (Duxbury et al., in press), 14 = Dolly Sods, WV (Hancock and Kirwan, 2007), 15 = PA Ridge and Valley (Braun, 1989), 16 = Ott et al., 1991, 17 = Conrad and Saunderson (1999), 18 = Sevon (1989), 19 = Carbonates (White, 1984, 2000), 20 = Piedmont schist (Cleaves et al., 1970, 1974; Pavich et al., 1989).

Flexural and Dynamic Models of Late Cenozoic Appalachian Epeirogeny

The non-eustatic base level fall signal propagating through Appalachian rivers and driving the unsteadiness in erosion and landscape change is direct evidence for unsteady epeirogenic rock-uplift. There are several possible causes of this epeirogeny of which two, flexural and dynamic, have been proposed and modeled. Both present plausible mechanisms that are supported by the geologic and geomorphic data. Not tested or considered here, but of equal possible importance are effects, both proximal and distal, to the now well-documented Chesapeake impact structure, the subsidence history of which is known to be both long-lived and unsteady (Hayden et al., 2008).

Flexural epeirogeny results from an elastic lithospheric response to redistributed surface loads in the crust. For the Appalachians, erosion onshore and deposition offshore in the BCT represents this load redistribution. Because the BCT has collected sediment from a broad part of the middle and northern Atlantic margin, but concentrated most of the deposited mass in its center, the lithospheric flexural response could be amplified for a more localized onshore region. The Fall Zone of the middle Atlantic margin and Pennsylvania Piedmont represents one such region where topography is unusually high-standing (Campbell, 1929) and river incision is particularly deep.
Figure 10. Comparison of BCT depositional record and AHe cooling from New England. (a) Cross-section oriented orthogonal to the New Jersey continental shelf showing the accumulation of siliclastic detritus eroded from the postorogenic Appalachians and preserved in the Baltimore Canyon trough (BCT) (Pazzaglia and Brandon, 1996). (b) Detrital AHe data from New England, representative of an Appalachian-wide data set that argues for broad cooling of the rocks at the surface at 100 Ma. (Pazzaglia et al., unpublished) (c) Unsteadiness in post-orogenic Appalachian erosion reconstructed from (a) and expressed as the flux of eroded rock (left axis) and erosion rate (right axis) for a contributing basin equal to the modern Atlantic Slope watershed of 300,000 km² (Pazzaglia and Brandon, 1996). The shaded region under the curve amounts to 2 km equivalent of rock removed from the Appalachians which represents all of Cenozoic, and a small portion of the Cretaceous section shown in the transparent window in (a). Accounting for dissolution of ~10m/m.y. over the past 100 m.y., 1 km of rock has been dissolved, added to 2 km of rock by erosion, sums to 3 km of rock removed in 100 m.y. Thus, the BCT and thermochronologic data agree in the total amount of postorogenic erosion; however, even the AHe data are insensitive to the nearly order of magnitude variation in erosion unsteadiness in the past 100 Ma.

A geodynamic model of flexural isostatic deformation of the middle Atlantic Coastal margin has been presented by Pazzaglia and Gardner (1994, 2000), following from general solutions presented in Turcotte and Schubert (2001). The interested reader is directed to these papers for full treatment of the methodology and results. In summary, four stratigraphic and geomorphic markers spanning a 15 m.y. time period from the middle Miocene to the present were established using Susquehanna River terraces, Fall Zone upland gravels, Coastal Plain deposits, and BCT lithostratigraphic packages. These markers were subsequently loaded by known volumes of sediment in the BCT, and unloaded...
by erosion in the Appalachians. Two models were considered: a one-dimensional line load oriented perpendicular to the margin and sub-parallel to the Susquehanna River and a two-dimensional distributed point load that encompassed the middle Atlantic states and the BCT. Both models treat the lithosphere as an infinite, perfectly elastic plate of finite thickness that response instantaneously to a fixed point load. Sediment accumulation offshore depresses the lithosphere and causes flexural uplift in the Appalachians at a distance determined by the rigidity of the plate. Erosion in the Appalachians amplifies the flexural uplift response. Both erosion and flexural rigidity were left as free parameters, the values of which were solved for by applying known BCT loads, and then applying a least squares regression fit to the four marker horizons.

The models are consistent in their prediction of an effective elastic thickness of 30 km for the Atlantic margin lithosphere and an average erosion rate of 10 m/m.y. (Fig. 11). More importantly, the models predict a steep, exposed flexural hinge at the Fall Zone which offers one explanation for that long-lived geomorphic feature that has separated the Piedmont from the Coastal Plain since the Cretaceous.

The models predict a maximum of about 80 m of rock uplift, above regional since the middle Miocene which is a good portion of, but not the total amount of uplift suggested by river incision in the Pennsylvania Piedmont. The absence of eustatic fall calls upon dynamic mantle forces, driven by continued subduction of the Farallon slab, to contribute to long-wavelength uplift and subsidence in the Appalachians (Forte et al., 2007; Moucha et al., 2008). In one model, dynamic mantle topography predicts 200 m of epeirogenic uplift of the middle Atlantic margin, centered on the Pennsylvania Piedmont, in the past 30 m.y. Some combination of these dynamic and flexural processes is in good general agreement with the geologic and geomorphic data. The emerging geochronology and river incision rate data would suggest that much of the flexural and dynamic uplift has occurred in the recent past, perhaps much of it in the Quaternary.

Figure 11. Results of (a) 1-D and (b) 2-D flexural isostatic models, see Pazzaglia and Gardner, 1994, 2000 for description of model details.
The coincidence of the Appalachian continental divide and predicted crest of the flexural forebulge suggests that the two features are related (Fig. 11). If the flexural forebulge is a dynamic feature that moves as surface loads are redistributed or in response to dynamic forces, it follows that the drainage divide would also be a dynamic feature, explaining the growth of the Atlantic slope basins (Harbor et al., 2005).

**Summary and Conclusions**

Appalachian landscape development, including the Pennsylvania Piedmont, has been unsteady in terms of base level fall, river incision, erosion and orogen unroofing, and delivery of sediment to the shelf-slope basin (BCT). This unsteadiness is likely driven by several factors including climate change, sediment storage in the landscape and the Coastal Plain, eustasy, and episodic jumps of the continental divide westward as the Atlantic slope basins continue to grow on the post-rift, decaying orogen flank. The ultimate and perhaps most important factor driving the unsteadiness has been variable rock uplift (base level fall) as the result of a lithospheric flexural response to surface loads and sub-lithospheric dynamic forces in the mantle.

The long term rate of Appalachian rock uplift and erosion is -20-30 m/my., a result consistent with AHe thermochronology and the long term average sediment yield to the BCT, but a bit misleading in terms of the range that has characterized post-orogenic landscape development over the past 180 m.y. The fact that several measures of modern erosion rates including TCN and sediment yields also return rates of -20-30 m/my. is interpreted as coincidental. These rates are faster than the earlier part of the Cenozoic, but still slower than contemporary rates of river incision downstream of the large Piedmont knickzone. The more rapid response of fluvial systems through incision in comparison to the slower processes eroding hillslopes argues that river gorges like the Susquehanna gorge in the Pennsylvania Piedmont are recent responses to recent base level fall that has not yet worked its way through the landscape.

**References**


Boettcher, S. S. and Milliken, K. L., 1994, MesozoicCenozoic unroofing of the souther Appalachian basin: apatite fission track evidence from Middle Pennsylvaniaan Sandstones: 1. of Geology, 102,655-663.


Duxbury, J., Bierman, P., Larsen, J., Pavich, Mj., Southworth, S., Miguens-Rodriguez, M., and Freeman, S., in press, Erosion rates in and around Shenandoah National Park, VA, determined using analysis of cosmogenic \( \text{l}^{7}\text{Be} \) and \( \text{l}^{26}\text{Al} \): American Journal of Science.


Granger, D. E., Kirchner, J. W., and Finkel, R.C., 1997, Quaternary downcutting rate of the New River, Virginia, measured from differential decay of cosmogenic \( \text{l}^{26}\text{Al} \) and \( \text{l}^{10}\text{Be} \) in cave-deposited alluvium: Geology, v. 25, p. 107-110.


Hancock, G. and Kirwan, M., 2007, Summit erosion rates deduced from \( \text{l}^{10}\text{Be} \) data: implications for relief production in the Central Appalachians, Geology v. 35, p. 89-92.


FIRST SEISMIC REFLECTION PROFILE FROM THE RIDGE AND VALLEY TO THE MARYLAND BORDER IN LEBANON AND LANCASTER COUNTIES, PA: PRELIMINARY RESULTS

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Introduction

Under a mandate to explore the Commonwealth for carbon sequestration potential, a 2D seismic reflection line which recorded in excess of 4 seconds of data was conducted. The project was named the Carbon Sequestration Technical Analysis (CSTA).

The route for the seismic acquisition was chosen by the author in conjunction with ARM Geophysics (ARM), taking into account logistics as well as specific geologic questions to be answered. Among these were: what is the character of faulting in the Great Valley? Is there a defined fault boundary of the Mesozoic basin? Can recent seismic events in the Lancaster area be identified as connected to proposed faults? What is the nature of the Piedmont south of the Martic Line and does the Martic Line show up in seismic reflection data?

To answer these questions, however, the seismic line must pass through some heavily populated regions. The rate of growth, particularly in Lancaster County, would prohibit a similar project in just a few years as there would be too much urban noise. The route was in three segments. The first segment from Ft. Indiantown Gap to well into the Mesozoic was conducted with an accelerated weight drop. The north edge of the Mesozoic to East Petersburg was collected using Vibroseis, as was the segment from Willow Street to Little Britain. The urban area of Lancaster was avoided due to logistical and noise problems. However the Mesozoic basin was captured twice.

Methods

ARM Geophysics of Hershey, PA, headed the project with AOA Geophysics of Houston, TX, doing the data collection. Reservoir Definition, Inc. of Houston provided processing expertise. Urban Seismic Specialists, Inc., of New Elm, TX provided peak particle velocity monitoring, and Flagger Force of York, PA, provided traffic control.

All data were collected along rights-of-way of public roads. Velocity monitoring was necessary. Shot points were located every 330 feet with receivers every 165 feet. The first segment was collected with 30-fold coverage while the vibrator data was at 90-fold. RDI conducted conventional seismic processing, with poor results due to the many dipping horizons and faults. By applying a specialized pre-stack process (Roscoe Sequence) it was possible to clean up the data and present a section with more coherent reflectors.

The northernmost line (Figure 2) avoided the Lebanon urban area and ended well in the Mesozoic basin. Heavily trafficked roads (RT 422, 322) left zones where data was unable to be collected and therefore the seismic reflection profile has a v-shaped data gap in the near surface. (Deeper horizons were picked up with offset shots).

Figure 1. AXIS accelerated weight source. (photo: Helen Delano).

Figure 2. Location of northernmost seismic line segment. (Tom Whitfield).
The second segment (Figure 3), using Vibroseis (Figure 4), started north of the Mesozoic and proceed to east Petersburg, PA, avoiding Manheim.

Figure 3. Location of second segment. An earthquake in December 2008 was located on the southern end of the line. (Tom Whitfield)
Results and Discussion

The Great Valley is marked by a large number of thrusts and faults (Fig. 6), making it difficult to trace a horizon any distance (Fig. 7). The region is marked by numerous listric faults and décollements (ARM report, 2010). The Mesozoic did not reveal a marked contrast as one might expect with a fault boundary.

The second segment (Line 301V), Figure 8, shows remarkable coherent reflectors that appear relatively undeformed contrary to what might be expected for the Cambro-Ordovician of central Pennsylvania.

South of the Mesozoic, where definitive evidence exists of a thrust fault in mapping by Wise and Ganis (2010, personal communication) and a corresponding earthquake epicenter (3.3 magnitude) there is no readily identifiable feature in the seismic data.

In 1974 a low level aeromagnetic map was made by Applied Geophysics of Salt Lake City, UT, for Chester Engineering as part of a study of the proposed Fulton Power Plant Site. The paper contour maps were shared by Don Wise with BTGS. The aeromagnetic data were digitized and compiled by GeographIT (a geospatial consulting Company in Lancaster, PA). The resulting raster image has a few artifacts resulting from mislabeled contours in the original paper map (e.g. the NW-SE linear feature evident in Figure 12).

South of Lancaster, where aeromagnetic data and topographic data show a textural change, there is a corresponding low angle fault (proposed as the Buck Thrust) to the south in the seismic data. This corresponds to the zone described by Blackmer as Antietam-Harpers, which was previously mapped as schist, and fills the zone between Buck and the Martic Line in the seismic section in Figure 11.
Figure 5. Third segment. (Tom Whitfield).
Figure 6. Northernmost line, acquired with accelerated weight drop source. (ARM)

Figure 7. Detail of character of seismic data. Located under Rt 422. Note extreme folding and numerous faults. (ARM)
Figure 8. Line 301V. Note persistent reflectors. (ARM)

Figure 9. Detail showing where the projection of the thrust mapped by Wise should be evident (CDP 950). (ARM)
Figure 10. Topographic map of Buck, PA, region. Note topographic change paralleling Route 372, from a higher frequency to lower amplitude and frequency to the south. Buck is at CDP 720, which corresponds to the surface projection of the major thrust from the south. (Tom Whitfield).

Figure 11. Southernmost line, from Willow Street to Little Britain. (ARM)
Seismic velocity profiles derived from stacking velocities used for processing show velocity highs in the southern portion of each seismic profile, consistent with a thrust bringing faster material nearer to the surface. This suggests that there may be at least three significant thrusts. While the Buck Thrust is readily apparent, the others appear to be lost in a series of other thrusts and folds.
Figure 14. Velocity profiles. Note high velocity material brought closer to surface in the southern portion of each line. (ARM).
Conclusions

In summary, the extensive décollement and faulting of the Great Valley blends into the Mesozoic with no great discontinuity, suggesting that, at least at this point the Mesozoic is shallow and probably underlain by carbonates. The mapped geology bears little resemblance to the large scale thrusts evident in the seismic data, with some distinctive geologic features not being evidenced in the seismic data to any great extent (e.g. the Martic Line). The presence of large structures at depth which seemingly have little connection to the surface geology indicate the need for more detailed geophysical studies.

The seismic data are extremely noisy with few good coherent reflectors. This could be corrected with a proposed deep hole to be drilled by BTGS on the northern seismic line, which could then be used to correct velocity estimations. Additional holes would be highly desirable in the southern region and a better velocity map could be constructed. Funding is in progress for a gravity survey to be conducted by Larry Malinconico of Lafayette College. The final interpretation and report on this portion of the CSTA project will be made available on-line by BTGS.

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References

Aeromagnetic Survey-Fulton Power Plant Site, 1974, Chester Engineering
Blackmer, Gale, 2010, personal communication.
Wise, D. and Ganis, R., 2010, personnel communication.
Schematic cross-section of the Piedmont just east of the Susquehanna River (2008 version)

Locations of field trip stops
Assembly of foreland basin sequence (After Ganis and Wise, 2008)
Lithotectonic Map of the Susquehanna Piedmont