The Piedmont: Old Rocks, New Understandings

2017 CONFERENCE PROCEEDINGS FOR

THE 34TH ANNUAL MEETING OF THE

OF THE GEOLOGICAL ASSOCIATION OF NEW JERSEY

Edited by

Howell Bosbyshell

West Chester University of Pennsylvania

Friday October 13, 2017 annual meeting hosted by West Chester University

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SCHEDULE

Friday, October 13th

Teachers' Workshops – Merion Science Center

11:00-12:00 What Every Earth Science Teacher Should Have In Their Rock Box
Timothy Macaluso, Brookdale Community College and GANJ President-Elect

Astronomy Education in the K-12 Classroom
Karen Schwarz, Ph.D., West Chester University

Poster Session – Sykes Student Union

11:00-5:00 Authors present 11:00-12:30

Oral Presentations – Sykes Student Union

12:00	Registration Opens
12:50-1:00	Dr. Michael J. Hozik, Stockton University and GANJ President - Opening Remarks
1:00-1:40	Dr. Howell Bosbyshell, West Chester University - Constraints on the Tectonic Evolution of the Central Appalachian Piedmont from Monazite and Detrital Zircon Ages
1:40-2:20	Dr. Lynn Marquez, Millersville University - The Baltimore Mafic/Ultramafic Complex
2:20-3:00	Dr. Ryan J. Kerrigan, University of Pittsburgh at Johnstown - The Bell's Mill Ultramafic Body
3:00-3:10	Break
3:10-3:50	Dr. LeeAnn Srogi, West Chester University - The 201-Ma Magma Plumbing System in the Western Newark Basin: How (Maybe) Why It's Different from the East
3:50-4:30	Dr. Donald U. Wise, University of Massachusetts at Amherst - Northern and Southern Appalachian Wilson Cycle Junction in Pennsylvania
4:30-5:00	GANJ Business Meeting
5:00-6:00	Keynote Presentation – Dr. Gale C. Blackmer, State Geologist/Director, Pennsylvania Bureau of Topographic and Geologic Survey – Coming Together to Divide the Wissahickon: A Case in Teamwork and the Role of Government Surveys in Regional Geologic Studies
6:30-8:00	Dinner and Conversation – Iron Hill Brewery
8:30-9:00	Planetarium Program – Dr. Sandra F. Prichard-Mather Planetarium Dr. Karen Schwarz, West Chester University - Looking UP: A Tour of the Fall Sky

Saturday, October 14th

Field Trip – Leaves from Lot M, West Chester University 8:00 am

8:45 to 10:15	Stop 1 – Bell's Mill Road and Wissahickon Creek, Philadelphia, PA - Classic Wissahickon Formation
11:00 to 1:00	Stop 2 – Ridley Creek State Park, Media, PA – Avondale Massif gneiss and lunch
1:15 to 2:15	Stop 3 – Novotni Brothers Construction-Mount Road and Rt. 452, Aston, PA – Wilmington Complex
2:30 to 3:30	Stop 4 – Chester Park, Chester, PA – Chester Park Gneiss
4:15 to 4:50	Stop 5 – Waltz Road, West Chester, PA – Embreeville Thrust/Shear Zone
4:50	Return to West Chester University

Contents

Dedication

James Edward "Bud" Alcock

1946 - 2017

It is very fitting that this geological field guidebook, focused on the bedrock geology of the southeastern Pennsylvania Piedmont, be dedicated to Bud Alcock. Bud grew up in the Piedmont (Landenberg, PA), was educated here (BA Haverford College, MAT Temple University, and PhD University of Pennsylvania), and devoted much of his professional career to as a geologist and educator to the study of these rocks (Professor, Pennsylvania State University, Abington, PA and 40-year resident of the Mount Airy neighborhood of Philadelphia).

Bud's 1989 PhD at Penn, *"Tectonic Units in the Pennsylvania-Delaware Piedmont: Evidence from Regional Metamorphism and Structure",* formed the basis of his expertise in field mapping, metamorphic petrology, structure, and tectonics. Despite moving on to petrologic and geochronologic research in the Adirondack Highlands of northeastern New York and the Ruby Mountains of southwestern Montana, remained strongly interested in the geology of the southeastern Pennsylvania Piedmont.

Beyond academics, Bud was eager to share his enthusiasm for geology with school groups, teachers and the general public. After completing his PhD, he collaborated in the development of an earth science exhibit at the Academy of Natural Sciences, and often led field trips in the Wissahickon Valley. Following retirement he continued his research activities, delighting in finding new questions and different ways of looking at data.

Generally quiet and unassuming, Bud is widely remembered by his family and friends for his kindness and generosity, his commitment to social justice and his determination to make the world better. He is also remembered for his wry humor (and awful jokes), love of sports, and many skills as a handyman, mechanic, and guitar player. Bud was a dedicated teacher and student mentor, supportive and trusted research colleague, and loyal friend.

Peter Muller, 9/22/17

Snaking Through the Pennsylvania Piedmont: Serpentinites of Southeastern Pennsylvania

Samuel Louderback and Ryan Kerrigan
Department of Energy and Earth Resources, University of Pittsburgh at Johnstown, Johnstown, PA 15904, U.S.A.

A petrographic and geochemical assessment of a suite of serpentine rocks from the Piedmont region of southeastern Pennsylvania was conducted to delineate their likely source and determine if any trends could be noted that could suggest spatial differences in tectonic history. The area is host to rocks that are metamorphosed to amphibolite facies accreted onto the continental margin during the Taconic orogeny (450-470 Ma) that overlay a basement composed of higher grade granulite facies metamorphic rocks of Grenville age (1.0-1.2 Ga). Serpentinites were collected from several pod-shaped bodies in southeastern Pennsylvania with some of these bodies adjacent to NE-SW trending shear zones.

Distinct differences were seen in field relationships and petrography whereas geochemical signatures remained uniform. Geochemical trace element data suggests that the sampled serpentinites were affected by subduction and refertilization of ultramafic protoliths during tectonic emplacement. The geochemistry of relatively immobile trace elements plotted on petrogenetic discrimination diagrams suggest an arc setting for the protolith. Examined bodies towards the east exhibited a zoned morphology with multiple rock types (talc, anthophyllite, chlorite, including serpentine) whereas bodies towards the west exhibited near complete serpentinization. Evidence of shear is present in all serpentinite samples and is particularly pronounced in the samples collected in the west.

Feasibility of Drone Use to Model Diabase Features

Daniel M. Bochicchio, LeeAnn Srogi, Tim Lutz, Howell Bosbyshell, Joby Hilliker
Department of Earth and Space Sciences, West Chester University, West Chester, PA 19383, U.S.A.

Collecting aerial imagery with drones/UAVs in order to create orthomosaic images and Digital Elevation Models (DEM) is becoming routine throughout industry and academia. Sensors that capture this data are constantly evolving; increasing in quality, availability, and lowering in prices rapidly. The boundaries of innovation are also expanding - this study focuses on the feasibility of using drones, photogrammetry, and 3D modeling to survey specific features of value within a diabase quarry. This method is highly valuable to hard rock geologists who seek to predict the cost of retrieval and payout of a targeted resource with higher accuracy than previous methods. We are working in a dimension-stone quarry where the patterns of mineral layering within the diabase determine the monetary value for monuments, countertops, and facing stone. The layering also is of interest to igneous petrologists seeking to understand how magma crystallizes. Each mineral layer may extend laterally for several meters but is thin, no more than 1-5cm thick. The primary goal of this project is to determine whether images of these thin layers can be captured, modeled, and interpolated in three dimensions across adjacent, orthogonal walls of the quarry. Because of the cost and multi-use functions of drone surveys, this is likely to be the predominant practice of non-invasive and non-destructive analysis in the near future.

Origin of erosional landforms along and near the Schuylkill River-East Branch Brandywine Creek drainage divide segment of the Schuylkill River-Delaware River drainage divide, Chester County, PA

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ABSTRACT

Analysis of erosional landform evidence as observed on detailed topographic maps was used to test a recently proposed hypothesis that present day southeast Pennsylvania Piedmont region drainage routes originated when deep valleys eroded headward across massive and prolonged southwest oriented floods. Map evidence for the Schuylkill River-East Branch Brandywine Creek drainage divide, which is also the Schuylkill-Delaware River drainage divide, and nearby drainage divides was analyzed to determine whether the hypothesis explained the origin of the erosional landforms such as drainage divides, valley orientations, barbed tributaries, water gaps, wind gaps, and through valleys including the Chester Valley segment between the Schuylkill River and East Branch Brandywine Creek. The hypothesis was found to be extremely productive in terms of its ability to explain previously unexplained origins of valley and drainage route orientations, barbed tributaries, water gaps, and drainage divide crossings such as through valleys and wind gaps. The origin of these erosional landforms can be explained if headward erosion of a deep southeast oriented Schuylkill River valley and its tributary valleys captured massive and prolonged southwest oriented floods that were moving immense quantities of water to what at that time was a deep and actively eroding Brandywine Creek-East Branch Brandywine Creek valley, which was also eroding headward from the southwest oriented Delaware River valley across the massive southeast oriented flood flow. Further the hypothesis explains how and why major regional drainage routes, such as East Branch Brandywine Creek flow across geologic structures and do not flow along zones of more easily eroded bedrock. The hypothesized immense and prolonged southwest oriented flood source was not determined, but a large and melting North American continental ice sheet could have produced the required water volumes.

Key Words: *Chester Valley, drainage basin erosion, Piedmont Province, southeast Pennsylvania*

INTRODUCTION

Philadelphia (PA) is located just north of the confluence of the southeast oriented Schuylkill River and the southwest oriented Delaware River. Upstream from Philadelphia the Delaware River flows in a southeast direction before turning in a southwest direction (see figure 1). The result is a Delaware-Schuylkill River drainage divide extends in roughly a north direction from Philadelphia and a Schuylkill-Delaware River drainage divide extends in roughly a west direction from Philadelphia. The Delaware-Schuylkill divide extends across the Piedmont Province in a north and northwest direction into the Appalachian Section of the Ridge and Valley Province, however the Schuylkill-Delaware divide ends west of the Brandywine Creek drainage basin and becomes the Schuylkill-Susquehanna River drainage divide.

Evidence of drainage routes that crossed the southeast Pennsylvania Piedmont Province prior to present day drainage basin formation can be found along these major divides and also along other regional drainage divides. Recently such evidence has been used to propose a hypothesis that massive and prolonged southeast oriented floods eroded three north oriented Montgomery County water gaps (Clausen, 2016a), the Tookany(Tacony) Creek drainage basin (Clausen, 2016b), and the Pennypack Creek drainage basin (Clausen 2017). The flood erosion hypothesis was extremely productive in terms of its ability to explain origins of erosional landforms such as drainage divide and valley orientations, barbed tributaries, water gaps, wind gaps, through valleys, and similar features. While study of one of the Montgomery County water gaps used evidence along the Schuylkill-Delaware River divide, study of the Tookany(Tacony) and Pennypack Creek drainage basins used evidence near the Delaware-Schuylkill River drainage divide. The study described here looks in detail at the Schuylkill River-East Branch Brandywine Creek drainage divide area to determine whether or not the massive and prolonged southwest oriented floods hypothesis is equally productive in terms of being able to explain erosional landform origins.

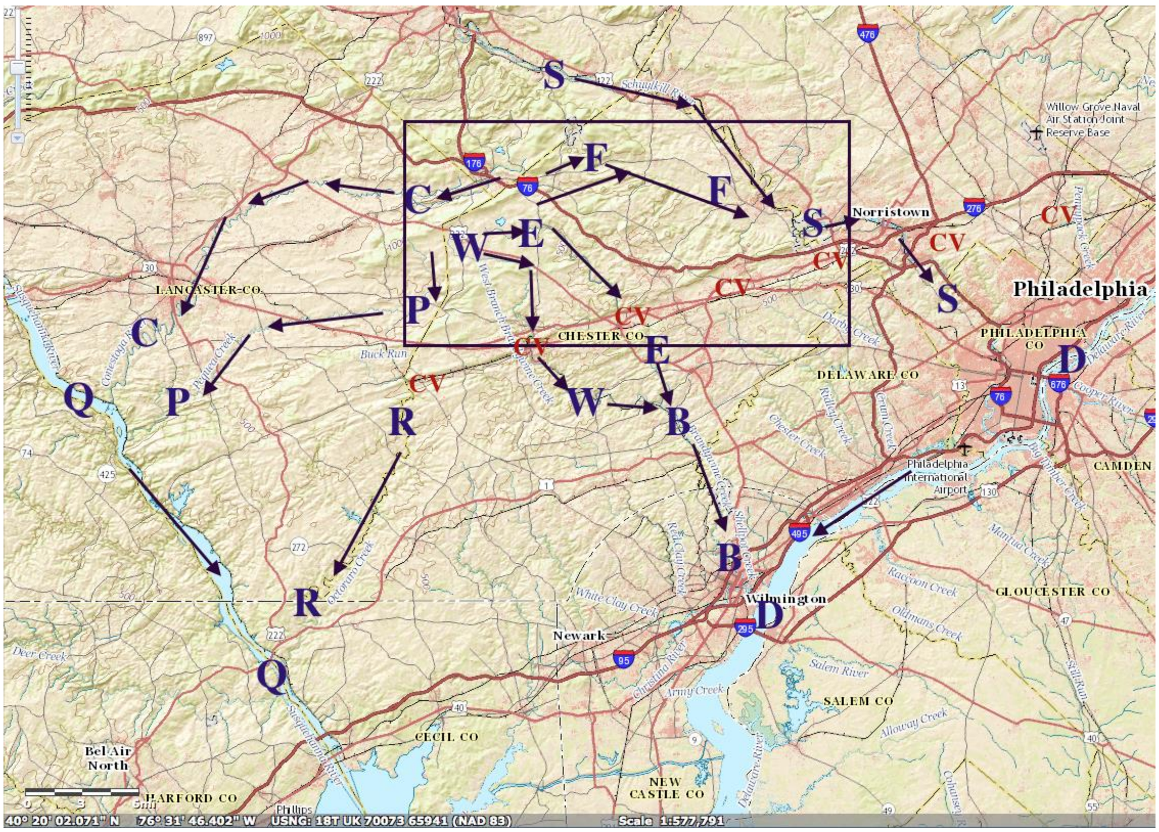

Figure 1: Modified map from USGS National Map website showing this paper's study area (inside the rectangle) in relation to Philadelphia (PA) and regional drainage routes. Arrows show present day flow directions and letters identify drainage routes discussed in the text. Scale bar in southwest corner represents 6 miles.

Figure 1 provides a location map for the Schuylkill-East Branch Brandywine Creek segment of the Schuylkill-Delaware River drainage divide investigated here. The rectangle identifies this paper's study area with letters identifying streams and rivers discussed in this paper and arrows indicating their flow directions. Rivers shown in figure 1 are the southwest oriented Delaware River (D) flowing to Delaware Bay, the south-southeast oriented Susquehanna River (Q) flowing to Chesapeake Bay, the southeast oriented Schuylkill River (S) flowing to the southwest oriented Delaware River, and the west and southwest oriented Conestoga River (C) flowing through Lancaster to the south-southeast oriented Susquehanna River. Streams and creeks discussed in this paper include Brandywine Creek (B) flowing to the southwest oriented Delaware River, East Branch Brandywine Creek (E), West Branch Brandywine Creek (W), southwest oriented Pequea

Creek (P) flowing to the Susquehanna River, south-southwest oriented Octoraro Creek (R) flowing to the Susquehanna River, and French Creek (F) flowing to the Schuylkill River. The letters “CV” identify the carbonate floored Chester Valley, which extends across much of the region from north of Philadelphia in a west-southwest direction and is crossed by multiple south or southeast oriented streams including Wissahickon Creek north of Philadelphia), the Schuylkill River, East Branch Brandywine Creek, West Branch Brandywine Creek, a major West Branch Brandywine Creek tributary, and several Octoraro Creek tributaries.

Two mid twentieth century United States Geological Survey (USGS) publications discuss study region landscape features, but do not address the Schuylkill River-East Branch Brandywine Creek drainage divide origin. Bascom and Stose in their 1932 USGS Atlas folio 223, which focuses on bedrock geology of the Coatesville and West Chester quadrangles, comment without explaining why that streams in the area “pursue southeasterly, southerly, or southwesterly courses, transverse to the trend of the geologic formations, and thus pass from soft to hard rocks, through which alike they have cut their channels… they occupy gorges which are sunken in wide U-shaped valleys which are in turn cut in still wider shallow valleys.” Six years later in USGS Bulletin 891 (Bascom and Stose, 1938) describing the northern study region note the East and West Branches of Brandywine Creek are “contesting the divide with tributaries of the Schuylkill on the northeast and of the Susquehanna on the west. They are strong streams, whose courses transverse to the strike of the formations were established in consequence of the slope of an uplifted peneplain.” Beyond that brief comment the drainage divide origins are not discussed. While the Brandywine Creek channel has been extensively studied for other reasons (e. g. Wolman, 1955; Pizzuto & Meckeinberg, 1989; Cinotto, 2003; and Pizzuto, Oneal, & Stotts, 2010) other literature describing study region erosional landforms origins was not found.

The immense southwest oriented floods discussed by Clausen (2016a, 2016b, 2017) flowed across the Delaware-Schuylkill River drainage divide north of Philadelphia and prior to being captured by headward erosion of the southeast oriented Neshaminy Creek and Delaware River valleys flowed to the southeast oriented Schuylkill River valley just north of this paper’s study region. Questions needing answers in this study are did the massive and prolonged southwest oriented floodwaters cross the present day Schuylkill River valley and if so did the floods also cross the Schuylkill River-East Branch Brandywine Creek segment of the Schuylkill-Delaware River drainage divide west of Philadelphia and continue to the Susquehanna River drainage basin? And is the flood hypothesis as productive in terms of being able to explain erosional landform origins west of Philadelphia as it was in terms of being able to explain erosional landform origins in the Delaware-Schuylkill River drainage divide area north of Philadelphia?

METHOD

Drainage divide erosion history is best studied using the most detailed topographic maps available where a small contour interval is critically important. For this reason USGS topographic maps used in this study were obtained from the Pennsylvania Department of Conservation and Natural Resources (DCNR) Interactive Map Resources website and from the USGS National Map Viewer and the USGS Historical Map Collection website. In addition an interactive digital geologic map available on the DCNR Interactive Map Resources website and the Honeybrook (Brown, 2006) and Wagontown (Marquez, 2005) quadrangle geologic maps were used to identify bedrock units. Topographic map interpretation of the study region erosional landform evidence began by inspecting the Schuylkill River-East Branch Brandywine Creek and other regional drainage divides to identify divide crossings such as through valleys, wind gaps, saddles, or other low points along the divide. These divide crossings were considered evidence of former drainage channels that once crossed each drainage divide. Valley orientations close to each of the identified divide crossing were (unless there was good reason to believe otherwise) used to determine the former channel orientations. Closely spaced divide crossings suggested the possibility of diverging and converging channels such as might be found in a large flood formed anastomosing channel complex, although other possibilities such as converging channels in a former dendritic drainage pattern were also considered.

Drainage routes on either side of each drainage divide were studied to determine why the earlier drainage route had been dismembered, which usually consisted of looking for stream capture evidence such as asymmetric drainage divides and barbed tributaries. In the case of opposing drainage routes (on either side of a divide crossing) the capture evidence also meant the divide had been created when one of the two opposing valleys had been beheaded and reversed. Long reversed flow valleys suggested the original channel had an extremely low gradient at the time of dismemberment and water captured from adjacent drainage routes was considered to have been responsible for eroding all deep reversed flow valleys. Shallow diverging and converging channels in a flood formed anastomosing channel complex were considered the simplest situation by which a reversed flow channel could obtain large volumes of water from adjacent channels during a stream capture and flow reversal event. Headward erosion of a deep valley across such a diverging and converging channel complex would behead and reverse flow sequentially with yet to be beheaded and reversed channels still moving large quantities of water that could be captured by a newly reversed flow channel which then moved the captured water to a lower base level.

Bedrock geology was checked to determine how geologic structures influenced drainage route locations. The most significant influence appeared to be in the carbonate-floored Chester Valley, which as described by Fenneman (1938) and Thornbury (1965) is a 55-mile long limestone-floored and remarkably straight trench. Potter (1999) includes it in the Piedmont Lowland Section (of the Piedmont Province) while uplands on either side are considered in the Piedmont Upland Section. Bascom et al in their 1909 United States Geological Survey Folio (p. 18) report finding no evidence that a stream ever flowed along this valley and suggest differential weathering was responsible for the valley formation. The Schuylkill-Delaware River drainage divide crosses the Chester Valley in the study region and the decision was made to consider the Chester Valley drainage divide origin independent of the drainage divide origin further to the north.

THE SCHUYLKILL-DELAWARE DRAINAGE DIVIDE IN THE CHESTER VALLEY

The Chester Valley segment between the southeast-oriented Schuylkill River and south oriented East Branch Brandywine Creek offers important clues when deciphering southeast Pennsylvania drainage history. Figure 2 provides a regional map of the Chester Valley segment between the Schuylkill River and East Branch Brandywine Creek. Arrows indicate stream flow directions and letters identify the Chester Valley (CV), North Valley Hills (NH), South Valley Hills (SH), the Schuylkill River (S), East Branch Brandywine Creek (E) flowing to the Delaware River, Valley Creek East (X) flowing to the Schuylkill River, Valley Creek West (Z) flowing to the East Branch Brandywine Creek, Pickering Creek (P), and French Creek (F). Carbonate rocks (limestone and dolomite) underlie the Chester Valley while the North and South Valley Hills are composed of much more erosion resistant gneiss, schist, quartzite, and similar material.

The Schuylkill-Delaware River drainage divide crosses the Chester Valley at location 1 where two north oriented streams, Valley Creek East (X) and Valley Creek West (Z) diverge with Valley Creek East flowing in an east-northeast direction along the Chester Valley floor to eventually reach the Schuylkill River at location 2 and Valley Creek West flowing in a west-southwest direction to eventually reach the East Branch Brandywine Creek at location 3. South of location 1 the Schuylkill-Delaware River drainage divide is located between the north oriented

Valley Creek East and Valley Creek West headwaters and then turns and follows the South Valley Hills crest in an east-northeast direction. North of location 1 the Schuylkill-Delaware River drainage divide follows the North Valley Hills crest in a west-southwest direction to slightly beyond location 4 where it turns in a north-northwest direction between south oriented East Branch Brandywine Creek tributaries and the headwaters of northeast oriented Pickering and French Creek tributaries. Locations 4 and 5 identify deep gaps in the North Valley Hills, which are drained in a northeast direction by Pickering Creek tributaries. Two of several questions raised by figure 2 evidence are why does the Schuylkill-Delaware River drainage divide cross the Chester Valley floor and how were the gaps at locations 4 and 5 eroded?

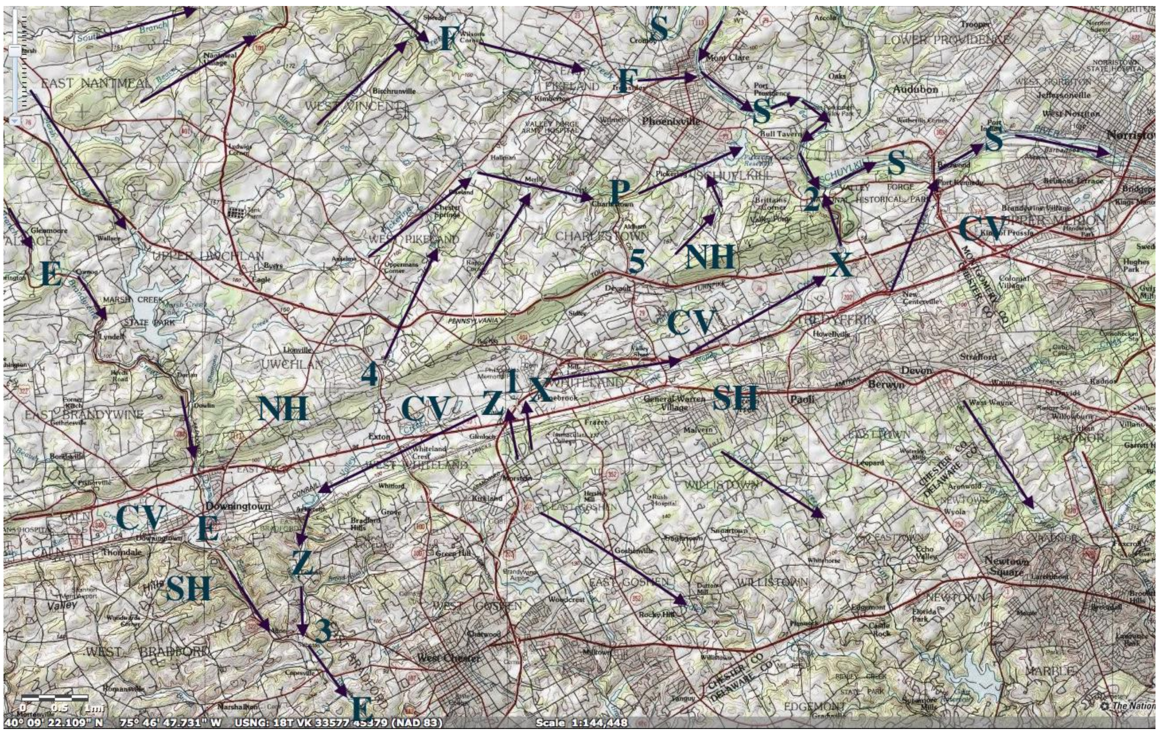

Figure 2: Modified USGS topographic map showing the Chester Valley segment between the Schuylkill River and East Branch Brandywine Creek. Arrows indicate stream flow directions while letters identify drainage routes and number identify locations as indicated in the text. Scale bar in southwest corner represents 1 mile.

Figure 3 shows a detailed topographic map of the Chester Valley midway between the Schuylkill River and East Branch Brandywine Creek (at figure 2 location 1) where the two opposing Valley Creeks originate. Both Valley Creeks originate almost in the same place on the Chester Valley south wall and flow parallel to each other in a north direction onto the Chester Valley floor where Valley Creek East turns in an east-northeast direction to flow along the Chester Valley floor (toward the Schuylkill River) while the Valley Creek West turns in a west-southwest direction to flow along the Chester Valley floor (toward East Branch Brandywine Creek). The small letters "d" identify the Schuylkill-Delaware River drainage divide location as it crosses the Chester Valley from the South Valley Hills crest to the North Valley Hills crest. A shallow through valley on the Chester Valley floor links the two opposing creeks and provides evidence that water once flowed in one direction or the other across the present day drainage divide.

The short through valley linking the two opposing Valley Creeks (seen in figure 3) is strong evidence one of the two opposing streams has been reversed so as to create the present day

Schuylkill River-Delaware River (East Branch Brandywine Creek) drainage divide. The question then becomes which of the two opposing streams was reversed and why was it reversed? The answer can be found by following each of the opposing Valley Creeks downstream in a search for other evidence that flow in it had been reversed. Valley Creek West flows in a west-southwest direction along the Chester Valley floor toward south oriented East Branch Brandywine Creek, but as seen in figure 2 makes an abrupt turn to leave the Chester Valley floor and to flow in a deep south oriented valley through the much higher South Valley Hills to eventually reach the deep south-southeast oriented East Branch Brandywine Creek valley. Northward flow from the south oriented Delaware River through the Brandywine and East Branch Brandywine Creek valleys and then along the south oriented Valley Creek West valley carved in the South Valley Hills seems extremely unlikely so the hypothesis that Valley Creek West once flowed in a north and then east-northeast direction toward the Schuylkill River is rejected.

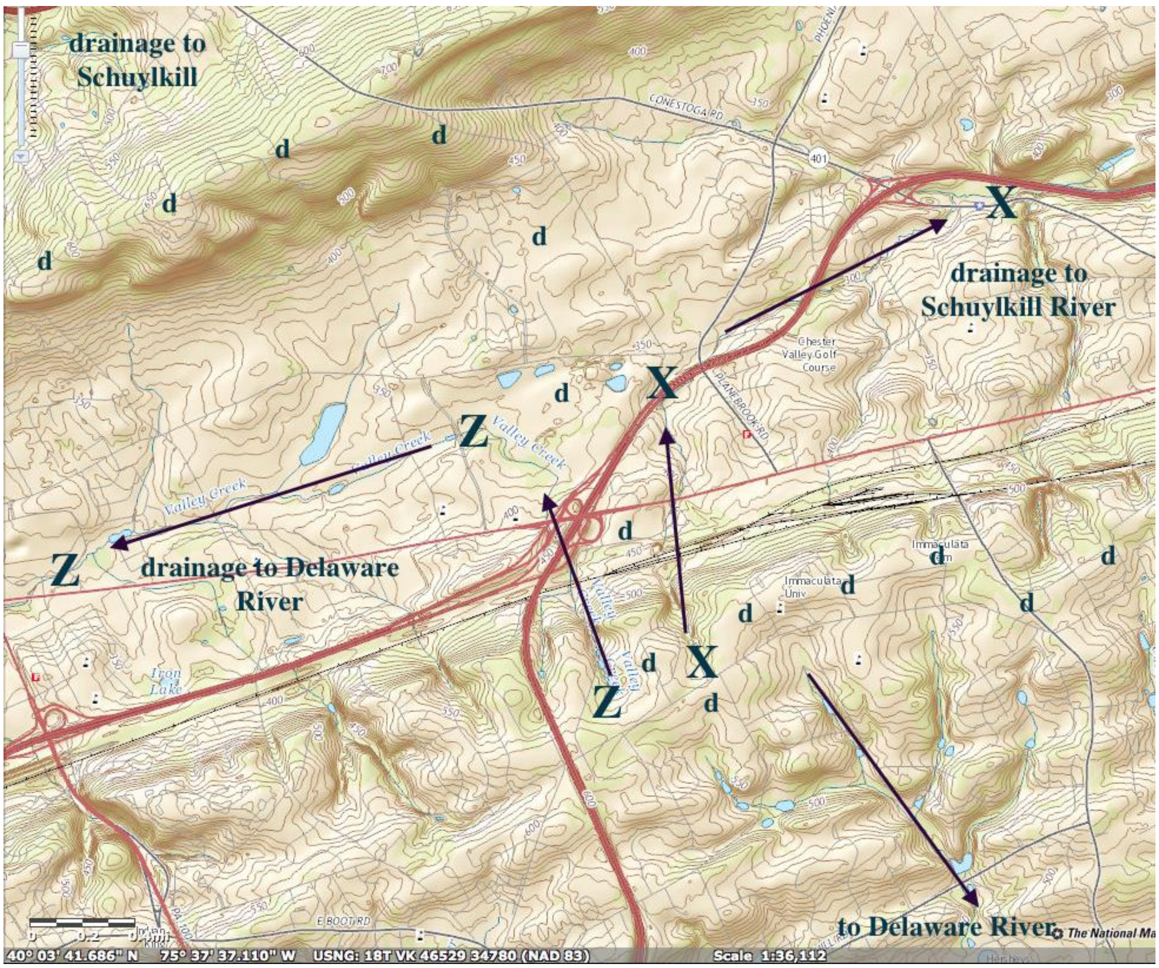

Figure 3: Modified map from the USGS National Map website showing where the Schuylkill-Delaware River drainage divide crosses the Chester Valley floor. Arrows indicate stream flow directions and letters drainage routes as described in the text. Scale bar represents .4 miles.

Figure 4 provides a detailed topographic map of the region near the figure 2 location 2 where Valley Creek East turns from flowing in an east-northeast direction on the Chester Valley floor to flow in a north direction through a deep water gap carved in erosion resistant quartzite bedrock across the North Valley Hills to reach the Schuylkill River as a barbed tributary. Overall the Schuylkill River is a southeast oriented river, but immediately north of the Valley Creek East mouth the Schuylkill River flows in S-shaped incised meanders (as seen in figure 2) and immediately upstream from the Valley Creek East mouth it is flowing in a southeast direction and then turns in a northeast direction away from the Valley Creek East mouth. At least some of the Valley Creek East water gap segments are eroded along what is probably a fault line, however the southeast-northeast Schuylkill River turn at the Valley Creek East mouth suggests south oriented water once flowed through the water gap to enter the Chester Valley.

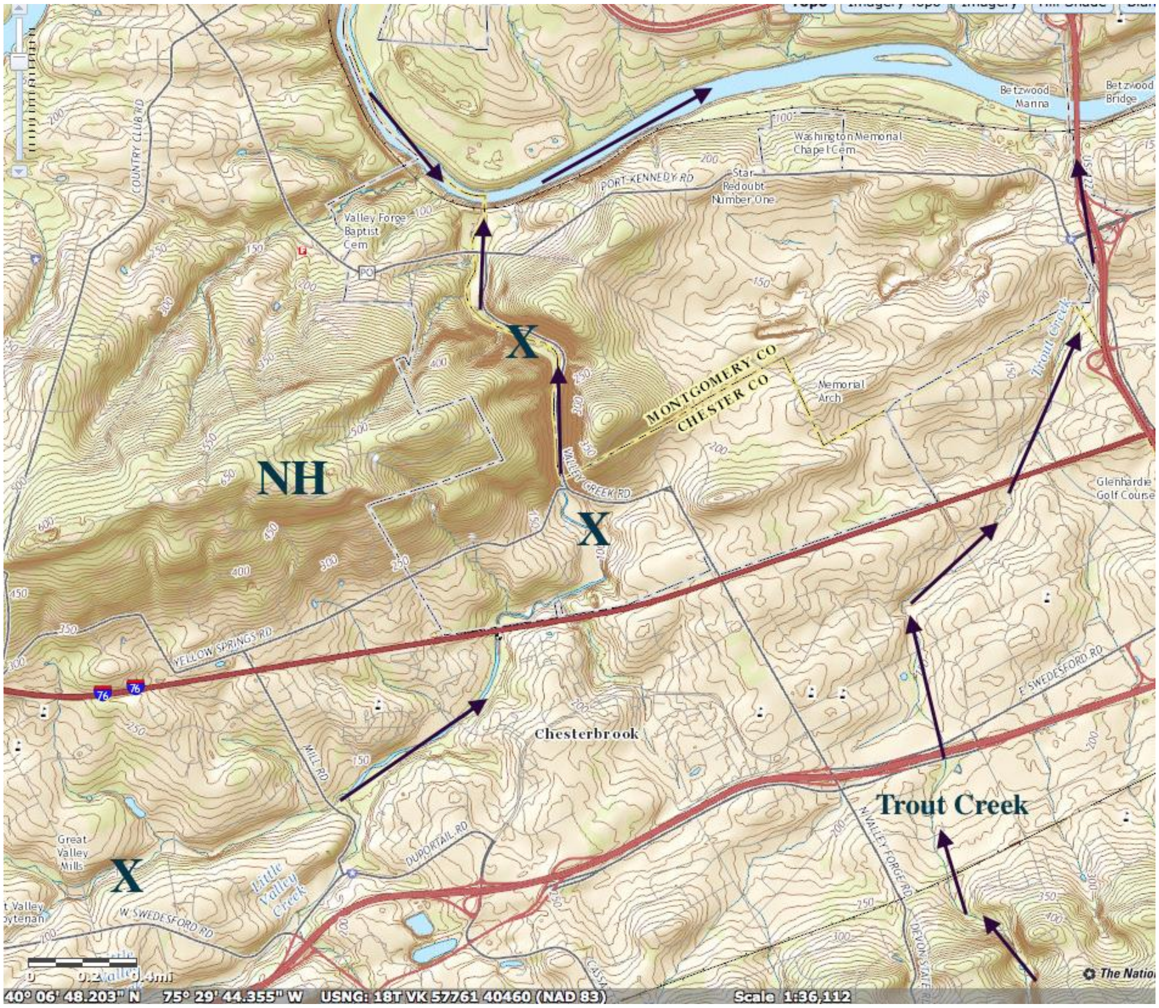

Figure 4: Modified topographic map from the USGS National Map website showing where Valley Creek East flows in a north direction through a deep water gap at Valley Forge to join the southeast and northeast oriented Schuylkill River. Arrows indicate stream flow directions and letters identify features as indicated in previous figures. Scale bar represents .4 miles.

The north oriented water gap at Valley Forge attracted the attention of previous researchers and Bascom and Stose (1938) determined the water gap is too large to have been carved in the erosion resistant quartzite bedrock by so small a stream and suggested the Schuylkill River once flowed south through the water gap and then turned in an east direction to flow across the Chester Valley (including across the north oriented Trout Creek valley which they do not mention). Gravel consisting of rounded quartzite pebbles up to 2 inches in size enclosed in finer grained sand at several locations was mapped to identify what they considered to be the former Schuylkill River channel about 140 feet higher than the current day Schuylkill River channel to the north. They also suggested erosion resistant bedrock in the water gap area retarded Schuylkill River down cutting and the Schuylkill River was diverted to its present channel when a tributary cutting into easier to erode bedrock further to the north captured it and Valley Creek East, which they suggested had always been an east-northeast oriented stream, was then captured by the deeper northern Schuylkill River valley, causing it to flow through the water gap as reversed flow on what had been the earlier south oriented Schuylkill River route.

An alternate and significantly less complex interpretation of the Valley Forge water gap origin was recently proposed by Clausen (2016a) who suggested the water gap was initially used by south oriented flood flow moving to the East Branch Brandywine Creek valley (and then to the Delaware River) prior to Schuylkill River development and Schuylkill River valley erosion. After flowing through the water gap the south oriented floodwater turned in a west-southwest direction along the Chester Valley floor to reach what at that time was a newly eroded and much deeper East Branch Brandywine Creek valley. The present day Schuylkill-Delaware River drainage divide on the Chester Valley floor was created when headward erosion of the deep Schuylkill River valley beheaded and reversed flow on what had been a very low gradient south and west-southwest oriented flow channel to create what is today the Valley Creek East drainage route.

According to this recently proposed hypothesis the south oriented Valley Forge water gap and the south oriented Valley Creek West valley through the South Valley Hills were initiated before Schuylkill River valley headward erosion and at a time when the entire region, including Chester Valley, was as high as the present day crests of the North and South Valley Hills. Massive and prolonged floods initially moved across this high level surface with water concentrated in shallow diverging and converging flow channels carved into easier to erode bedrock units (such as those in the Chester Valley), although no deep valleys existed until headward erosion of the deep East Brandywine Creek valley and its south oriented Valley Creek West tributary valley significantly lowered base level and enabled floodwaters to begin to lower the Chester Valley floor.

Clausen's 2016(a) paper does not address how reversed flow was able to erode what is today the east-northeast and north oriented Valley Creek East drainage route that has an elevation of about 370 feet near the Schuylkill-Delaware River drainage divide and an elevation of less than 80 feet where it enters the Schuylkill River. Figure 2 shows gaps in the North Valley Hills at locations 4 and 5. Northeast oriented Pickering Creek tributaries originate at those gaps and prior to being reversed by Pickering Creek valley headward erosion those northeast oriented valleys were used by southwest oriented flood flow moving into the Chester Valley. The beheading and reversal of flow in the Valley Forge water gap occurred before the actively eroding and deep Schuylkill River valley head had progressed far enough north and west to behead southwest oriented flow still moving water through the gaps at the figure 2 locations 4 and 5. Southwest oriented flow entering the Chester Valley at location 5 was captured by the reversed flow and made a U-turn to flow in an east-northeast and north direction to reach the much deeper Schuylkill River valley and in the process eroded what is today the deep Valley Creek East valley.

SCHUYLKILL RIVER-EAST BRANCH BRANDYWINE CREEK DIVIDE

North of the Chester Valley the Schuylkill River-East Branch Brandywine Creek drainage divide extends in a north-northwest and northwest direction and is crossed by closely spaced northeast to southwest oriented through valleys. Figure 5 provides a topographic map of the Schuylkill River-East Branch Brandywine Creek drainage divide area north of the Chester Valley. Arrows indicate stream flow directions and letters identify drainage routes as follows: East Branch Brandywine Creek (E) flowing to Brandywine Creek and the Delaware River, Marsh Creek (Em) flowing to East Branch Brandywine Creek, West Branch Brandywine Creek (W) flowing to Brandywine Creek and the Delaware River, Pickering Creek (P) and French Creek (F) flowing to the

Schuylkill River, South Branch French Creek (Fs), Pine Creek (Fp) and Beaver Run (Fb) flowing to French Creek, Conestoga River (C) flowing to the Susquehanna River, East Branch Conestoga River (Ce), West Branch Conestoga River (Cw), and Pequea Creek (P) all flowing to the Susquehanna River. The letters "CV" indicate the Chester Valley location and numbers identify locations discussed in the text with number 1 identifying Welsh Mountain, number 2 Thomas Hill, and number 3 the Baron Hills.

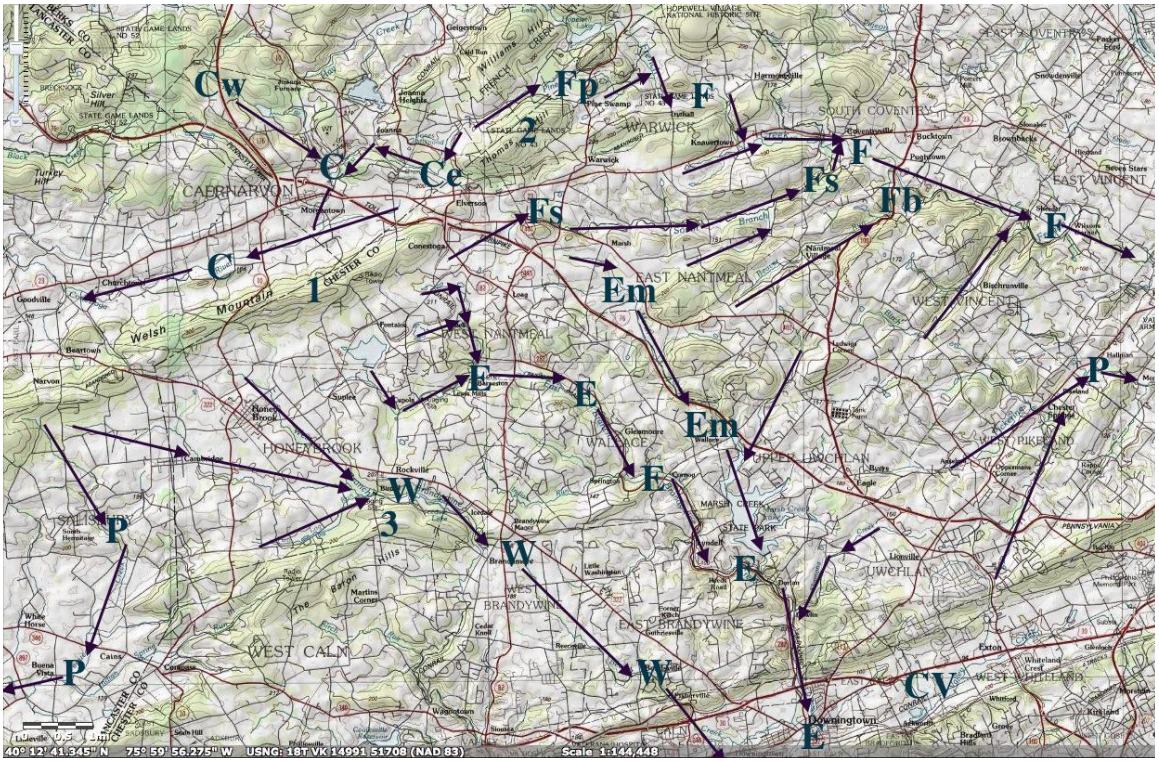

Figure 5: Modified USGS topographic map showing the Schuylkill River-East Branch Brandywine Creek drainage divide area north of the Chester Valley. Arrows indicate present day stream flow directions and letters identify drainage routes as indicated in the text while numbers give locations of features discussed in the text. Scale bar represents 1 mile.

North of the East Branch Brandywine Creek headwaters is a southwest to northeast oriented ridge known as Welsh Mountain (location 1 in figure 5). According to the digital geologic map at the DCNR Interactive Map Resources website Welsh Mountain is a Cambrian quartzite ridge with Precambrian gneiss and anorthosite located between it and the Baron Hills (another quartzite ridge at location 3) with Cambrian age carbonate rocks in the Conestoga River valley to the northwest. South of Welsh Mountain most drainage is to the East and West Branches Brandywine Creek with water flowing in a south and southeast direction to reach the Delaware River. North of Welsh Mountain drainage is to the Conestoga River with the water flowing in a west and eventually southwest direction to reach the Susquehanna River. West of the West Branch Brandywine Creek headwaters in the area south of Welsh Mountain Pequea Creek begins with the water eventually reaching the Susquehanna River. Near the Welsh Mountain northeast end drainage to the South Branch of French Creek is located between Welsh Mountain and the East Branch Brandywine Creek headwaters. Water in the South Branch of French Creek eventually reaches the Schuylkill River. Why would streams flowing to three different rivers originate in the small region near the northeast end of Welsh Mountain seen in figure 5?

The pattern of drainage routes seen in figure 5 supports the massive and prolonged southwest oriented flood hypothesis. North of Welsh Mountain and Thomas Hill and of the Brandywine Creek drainage basin is the drainage divide between the southwest oriented Conestoga River (flowing to the Susquehanna River) and northeast oriented Pine Creek, which flows to southeast oriented French Creek and then to the southeast oriented Schuylkill River. South of Welsh Mountain and Thomas Hill the south oriented Brandywine Creek drainage basin is between west and southwest oriented Pequea Creek and northeast oriented French Creek tributaries. Drainage divides between the northeast oriented French Creek tributaries and the Conestoga River headwaters and also the south oriented East Branch Brandywine Creek drainage basin cross through valleys (just like in the Chester Valley to the south). This pattern of northeast and southwest oriented drainage routes suggests the southeast oriented West Branch Brandywine Creek valley and southeast and south oriented East Branch Brandywine Creek valleys eroded headward across multiple southwest oriented channels delivering large quantities of water to the Pequea Creek drainage basin (and then to the Susquehanna River) with some reversals of flow creating the present day northeast oriented West and East Branch Brandywine Creek tributary drainage routes. Southwest oriented flood flow across the figure 5 map area ended when headward erosion of the deeper southeast oriented French Creek valley beheaded and reversed the flow to create the northeast oriented French Creek tributaries seen today.

Bascom (1921 and 1938) when interpreting the physiography of the figure 5 map area considered tops of the erosion resistant Cambrian quartzite ridges making up Welsh Mountain, Thomas Hill, and the Baron Hills to represent a higher level erosion surface (Bascom's Schooley surface ranging from about 900 feet in the Baron Hills to as much as 1080 feet on Welsh Mountain) and the "quartz monzonite and granodiorite" bedrock underlying the region between Welsh Mountain and the Baron Hills to represent a lower level erosion surface (Bascom's Harrisburg surface with elevations ranging between 700 and 800 feet). Bascom further notes these and other regional erosion surfaces "do not conform to the underlying rock formations, which have highly complex structure, nor do the master streams" and interprets the different erosion surfaces to have been developed in response to successive uplifts with each new erosion surface developed as the previous erosion surface(s) was (were) partially dissected. While this hypothesis does not explain how the modern day figure 5 map area drainage divides originated or why those drainage divides cross multiple through valleys or even why drainage routes such as the East and West Branches of Brandywine Creek and French Creek cross regional geologic structures the hypothesis in some respects appears to be consistent with the massive and prolonged southwest oriented flood hypothesis, although such an interpretation requires that the present day southeast oriented Schuylkill River valley did not exist at the time floodwaters crossed the French Creek-East Branch Brandywine Creek drainage divide and also implies that floodwaters lowered much of southeast Pennsylvania Piedmont surface by several hundred feet or more.

Perhaps the best place to understand the Schuylkill River-East Branch Brandywine Creek drainage divide origin is at the north end of the south oriented East Branch Brandywine Creek drainage basin. Figure 6 shows a detailed topographic map of the region where drainage routes to three different rivers diverge. The letter "E" identifies the East Branch Brandywine Creek and the letters "Em" identify Marsh Creek, which flows to the East Branch Brandywine Creek (in the Brandywine Creek drainage basin flowing to the Delaware River). The letters "F" identifies French Creek, "Fs" the South Branch French Creek, "Fp" Pine Creek and "Fb" Beaver Run (in the French Creek drainage basin flowing to the Schuylkill River). The letters "C" identify the Conestoga River and "Ce" the East Branch Conestoga River (in the Susquehanna drainage basin). The small letters "d" identify the Schuylkill River-East Branch Brandywine Creek drainage divide location, which ends on the southeast flank of the Welsh Mountain northeast end. The small letters "x" identify the location of the Schuylkill-Susquehanna River drainage divide, which begins north of the East Brandywine Creek drainage basin northern limit.

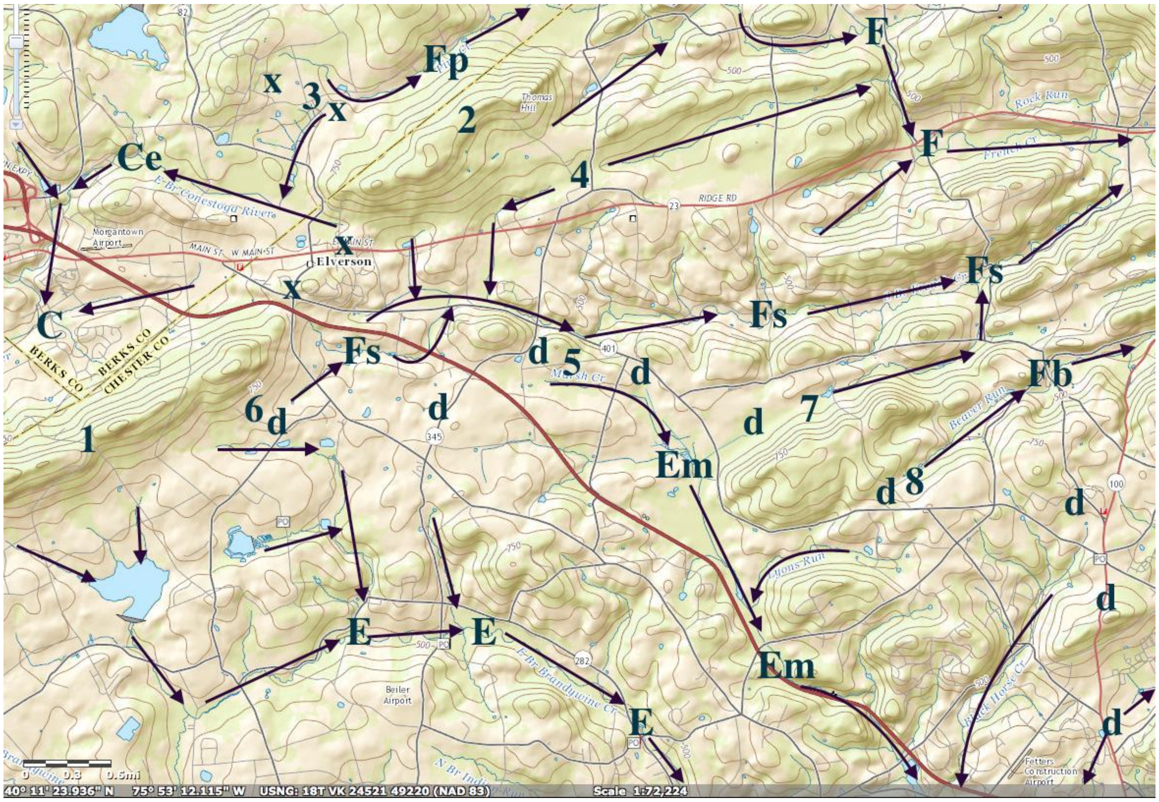

Figure 6: Modified map taken from the USGS National Map website showing drainage divides between streams flowing to three different rivers. Arrows indicate flow directions and the letters identify drainage divide location as indicated in the text. Scale bar represents .6 miles.

Locations 1 and 2 on figure 6 are the same as on figure 5 and identify Welsh Mountain and Thomas Hill respectively. Number 3 identifies a through valley linking the northeast oriented Pine Creek valley (draining to French Creek and the Schuylkill River) with a southwest oriented East Branch Conestoga River tributary (flowing to the Susquehanna River). This valley was eroded as a southwest oriented flood flow channel prior to headward erosion of the deeper French Creek valley, which beheaded and reversed the flow to create the present day drainage divide. Location 4 is interesting because it is in a through valley linking a northeast oriented French Creek tributary with a southwest and south oriented South Branch French Creek tributary. This through valley was also eroded by southwest oriented flood flow, which was captured by the reversal of flow that occurred when headward erosion of the French Creek valley beheaded and reversed flow in the present day South Fork French Creek valley. This capture occurred before French Creek valley headward erosion had progressed far enough north to behead and reverse the southwest oriented flow moving past location 4 and illustrates how reversed flow valleys captured water from yet to be beheaded and reversed flood flow channels.

Numbers 5, 6, 7, and 8 (on figure 6) are all located on the French Creek-East Branch Brandywine Creek drainage divide. Number 5 is located on a low ridge separating the South Branch French Creek valley from the Marsh Creek headwaters (with Marsh Creek being an East Branch Brandywine Creek tributary). Note how at its headwaters Marsh Creek flows in almost the same direction as the South Branch French Creek to the north before turning in a southeast direction indicating that headward erosion of the southeast oriented Marsh Creek valley beheaded and reversed west oriented flood flow probably moving to the what at that time were newly eroded East Branch Brandywine Creek tributary valleys south and southwest of the number 6. The number 6 illustrates how the South Branch French Creek headwaters were created by the reversal of southwest oriented flow that was being channeled along the Welsh Mountain southeast flank. Prior to that flow reversal the southwest oriented flow along the Welsh Mountain southeast flank was beheaded and reversed by headward erosion of south oriented East Branch Brandywine Creek tributary valleys. The grid of northeast-southwest and south oriented East Branch Brandywine Creek tributary valleys south of number 6 was formed by this process with number 7 being located in a valley on the divide between a present day northeast and north oriented South Branch French Creek tributary and Marsh Creek which is located in a former flow channel that once continued in a west-southwest direction along the present day alignment of the two lakes south of number 6. Number 8 identifies the divide between northeast oriented Beaver Creek (flowing to French Creek) and southeast oriented Marsh Creek.

DISCUSSION AND CONCLUSIONS

Probably the best test of any hypothesis is its ability to explain previously unexplained evidence. The hypothesis that the Brandywine-East Branch Brandywine Creek valley captured massive and prolonged southwest oriented floods prior to headward erosion of the southeast oriented Schuylkill River valley explains not only the creation of the Schuylkill River-East Branch Brandywine Creek drainage divide, but also the erosion of what are today multiple through valleys crossing that divide along with the orientations of many, if not all, of the opposing East Branch Brandywine Creek and Schuylkill River tributaries heading along or near the divide. In addition the hypothesis also explains the origin of previously unexplained barbed tributaries, water gaps, and wind gaps. There can be no question that the hypothesis explains considerable previously unexplained evidence, however the hypothesis also raises several important questions.

One of the questions raised relates to sediments the flood waters may have deposited. Scattered deposits of poorly explained Cenozoic sediment are found covering the older southeast Pennsylvania Piedmont Province bedrock surface and while the southwest oriented floodwaters may have deposited those sediments the question arises as to why those deposits are so sparse. The only geologic event known to this author capable of generating immense and prolonged floods of the type described in this paper is the melting of a 1 to 3 mile thick or thicker continental ice sheet. The southeast Pennsylvania Piedmont region was not glaciated although there is good evidence what may have been the margin of such a thick continental ice sheet was nearby and extended across northern New Jersey. If such a thick continental ice sheet existed and rapidly melted massive southwest oriented melt water floods almost certainly would have flowed across the southeast Pennsylvania Piedmont Province. Warm climates would be required to initiate and continue the rapid continental ice sheet melting with mild climate flora and fauna probably flourishing in ice marginal regions. Further, most melt water from such a 1 to 3 mile thick or thicker continental ice sheet would contain little or no sediment as it flowed from the ice sheet upper layers. Such melt water would deeply erode regions over which it flowed and leave little or no sediment evidence to indicate the water source. Logically massive floods of such melt water could have crossed the Schuylkill River-East Brandywine Creek drainage divide and represent the most likely erosion agent responsible for erosion events this paper describes.

Another important unanswered question relates to the upstream Schuylkill and Delaware River valleys. Floodwaters described in this paper crossed not only the Schuylkill River-East Branch Brandywine Creek drainage divide but also the Delaware-Schuylkill River drainage divide, which means the southeast oriented Schuylkill River valley and perhaps the southeast oriented Delaware

River valley (upstream from the Trenton, NJ area) eroded headward across the immense southwest oriented flood flow. Unlike the Brandywine Creek drainage basin discussed in this paper and the Tookany(Tacony) and Pennypack Creek drainage basins discussed in Clausen's 2016 and 2017 papers the Schuylkill and Delaware River drainage basins extend into the Valley and Ridge Province and then into the Pocono and Catskill Mountain regions respectively. The unanswered question arises did massive and prolonged southwest oriented floods erode the upstream Schuylkill and Delaware River drainage basins or did headward erosion of southeast oriented Schuylkill and Delaware River valley segments now crossing the southeast Pennsylvania Piedmont region capture flow from previously developed upstream drainage basins? Based on today's topography it may seem unlikely that massive southwest oriented melt water floodwaters flowed across and eroded what are today the entire upstream Schuylkill River and Delaware River drainage basins. However, the concept is not impossible. In fact continental ice sheets are known to have once covered what are today upstream Schuylkill River and Delaware River drainage basin areas and if one envisages a thick and rapidly melting continental ice sheet occupying those areas it is easy to also envisage massive southwest oriented melt water floods flowing across the Valley and Ridge Province. While further work is needed to verify that massive southwest oriented floods eroded the upstream Schuylkill and Delaware River basins such a hypothesis is not impossible and is consistent with the hypothesis tested here.

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PETROLOGY AND GEOCHEMISTRY OF THE BELLS MILL ROAD ULTRAMAFIC BODY, PHILADELPHIA PA

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Abstract

A petrographic and geochemical assessment of the Bells Mill Road ultramafic body was conducted to examine its alteration history and to test hypotheses on the source of the ultramafic protolith. The Bells Mill Road ultramafic body is located in the Piedmont region of southeastern Pennsylvania within Wissahickon Valley Park at the intersection of Bells Mill Road and Wissahickon Creek. The Bells Mill Road ultramafic body is one of several altered ultramafic bodies scattered throughout the Pennsylvanian Piedmont province. The local area is host to rocks that are metamorphosed from greenschist to amphibolite facies during accretion onto the continental margin during the Taconic orogeny (450-470 Ma) that overlay a basement composed of higher grade metamorphic rocks of Grenville age (1.0-1.2 Ga).

The geologic history of the Pennsylvanian Piedmont has been thoroughly studied, but the origin of the ultramafic bodies remains a source of contention. There are several competing hypotheses to explain the protolith of these ultramafics, namely: oceanic crust (ophiolite), diapiric mantle, and arc-related magmatic differentiates. Geologic mapping of the body has shown the following lithologic zones: serpentine-talc rock, talc-tremolite schist, anthophyllite-chlorite schist, chlorite schist, and talc-serpentine schist. Spinel chemistry was obtained using EDS and whole-rock chemistry was obtained using ICP-MS and XRF. Elemental concentrations of spinel minerals reveal significant metamorphism (greenschist to amphibolite facies), altering the inherited protolith signals. However, plotting whole-rock trace element concentrations on multiple petrogenetic discrimination diagrams confirms an island arc origin. A minority of samples exhibit a mid-ocean ridge basalt affinity on petrogenetic diagrams, however, these rocks are rich in Al-phases (garnet, kyanite, and corundum), which may suggest a high level of recrystallization/metasomatism altering inherited signals. The ultramafic body is most likely the ultramafic differentiate of an arc-related magma system.

Introduction

The Bells Mill Road ultramafic body is part of a set of exotic rocks located in the Pennsylvanian Piedmont that have evaded proper characterization and scientists remain uncertain about their exact origin. Published geologic maps that contain coverage of the ultramafic bodies (Bascom et al., 1909; Weiss, 1949; Amenta, 1974; Amenta et al., 1974; Berg and Dodge, 1980; Bosbyshell, 2006; among others), contain contradictory information pertaining to the precise locations and contact relationships between the ultramafic bodies and adjacent rocks. Furthermore, all of the geologic maps for this region lump the ultramafic bodies as single units and do not differentiate between the various lithologies within the ultramafic bodies. Field work on the Bells Mill Road body revealed complex zonation that has not been adequately characterized.

Limited studies have been conducted on the ultramafic bodies in the Pennsylvanian Piedmont (Busé and Watson, 1960; Carnes, 1990; DeSantis, 1978; Roberts, 1969; Zarnowsky, 1995). Furthermore, these studies present contradictory information related to the lithologies and alteration zones within the bodies as well as varied contacts for the ultramafic bodies with respect to adjacent rocks. Previous studies on the ultramafic bodies included limited structural data for the ultramafic bodies. Addressing these discrepancies and omissions for the ultramafic bodies is essential for understanding the Paleozoic tectonic evolution of the Central Appalachian Mountain Belt.

Regional Geology

The basement rocks in the region are Precambrian gneisses (Baltimore Gneiss) that were metamorphosed during the Grenville orogeny approximately 1000 Ma (Wagner and Crawford, 1975). The Baltimore Gneiss can be observed approximately 1 km northwest of the Bells Mill Road ultramafic body, once crossing over the Rosemont shear zone (Figure 1 and 2). Regionally, the Rosemont fault separates the Baltimore Gneiss from the Wissahickon Formation, a pelitic to quartzofeldspathic schist with interspersed amphibolite layers. Ultramafic pods are found within the Baltimore Gneiss and Wissahickon Formation throughout the Piedmont Province and the pods are elongated, which trend parallel to major structural features trending generally northeast.

The Baltimore Gneiss is interpreted to have sedimentary and igneous origins and experienced its peak metamorphism at granulite facies during the Grenville orogeny approximately 1000 Ma (Wagner and Crawford, 1975). During rifting of supercontinent Rodinia in the late Proterozoic and the opening of an ocean basin, the Baltimore Gneiss served as the bedrock of the Laurentian continental margin as carbonate and siliclastic sediments were unconformably deposited. Shortly after, an island arc developed under an eastward dipping subduction zone on the eastern margin of the Laurentian continent (Crawford and Mark, 1982;

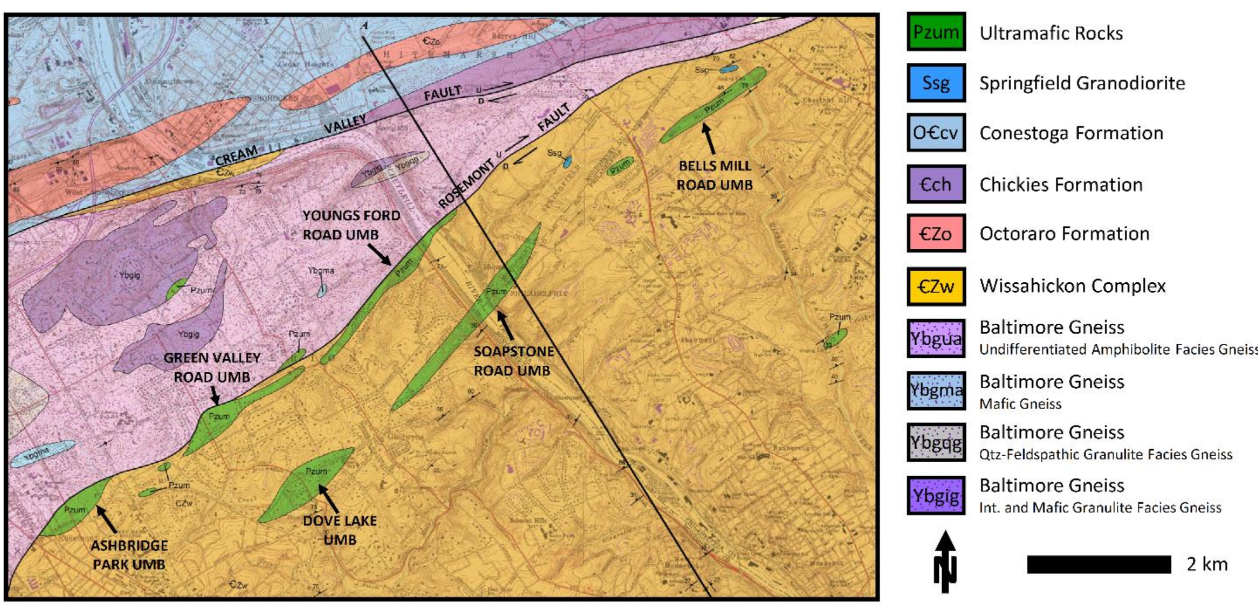

Figure 1. Map of the Bells Mill Road vicinity with regional ultramafic bodies labeled (modified after Bosbyshell, 2006).

Wagner and Srogi, 1987). The Wissahickon Formation was originally deposited in the marginal basin enclosed by the Laurentian continent and the offshore island-arc (Crawford and Crawford, 1980; Wagner and Srogi, 1987). Before or during the late Ordovician Taconic orogeny, a series of ductile nappes and thrust sheets were transported west as the island arc was accreted onto Laurentia (Crawford et al., 1999). New interpretations suggest the Taconic arc may have been accreted near the North Appalachians (i.e., New England) and transported by transcurrent faulting to its present location by the Late Paleozoic (Bosbyshell et al., 2016). The crustal thickening during the Taconic orogeny and high heat flow from the accreted island arc complex deformed the region into the Devonian (~385 Ma) from greenschist to amphibolite facies with grades increasing from the northwest to southeast toward the proposed core of the arc complex, the Wilmington Complex (Wagner and Srogi, 1987). After or during peak metamorphic conditions, the thick crustal orogen began to isostatically uplift. The later Alleghenian orogeny during the late Paleozoic appears to have had little effect on the rocks of the Piedmont Province; however, shortly after, erosion began to unroof the buried rock to expose the Piedmont units.

Ultramafic pods are scattered throughout the Baltimore Gneiss and the Wissahickon Formation ranging in size from 0.5 to 2.5 km long and 0.2 to 0.7 km wide. Geologists have speculated on the origins of the ultramafic bodies and offered a number of possible hypotheses: dikes of metapyroxenite and metaperidotites (Bascom et al., 1909; Busé and Watson, 1960); diapiric mantle intrusions (Weiss, 1949; Amenta, 1974; Amenta et al., 1974); and oceanic crust ophiolite sequences or ophiolite fragments within an accretionary prism (DeSantis, 1978; Wagner and Srogi, 1987; Carnes, 1990; Faill, 1997). Research conducted on the State Line serpentinites, ultramafic bodies approximately 80 km southwest of the Bells Mill Road ultramafic body, show a close association with the Baltimore Mafic Complex and have been interpreted to be part of a large layered mafic intrusion (McKague, 1964; Hanan and Sinha, 1989). The possibility of the ultramafic bodies in the Philadelphia region being related to a large layered mafic intrusion has not been ruled out. The age and emplacement history of the ultramafic bodies are unknown, however, they are believed to have been emplaced prior to or during metamorphism of the country rock based on seemingly conformable trends of the contacts, foliations, and lineations. Additionally, the lack of hornfels aureoles around the ultramafic bodies would support the pre- or syn-metamorphic age of the ultramafic bodies.

Field Relationships and Lithologies

Field work, petrography, and whole-rock geochemistry has been completed to understand the nature of the alteration zones, deformation history, and the potential protolith of the Bells Mill Road ultramafic body. The Bells Mill Road ultramafic body is an elongate unit (~0.25 by 1 mile in dimensions) emplaced within the Wissahickon Formation of southeastern PA and is oriented parallel to regional foliations (NE-SW trending). Foliations, textures, and contacts within the body all follow a general northeast-southwest trend with steep dips from ~70° to vertical and varying in dip direction (northwest to southeast). Direct contact between the body and the country rock is not observable.

The body can be separated into five distinct lithologic units: serpentine-talc rock, talc-tremolite schist, anthophyllite-chlorite schist, chlorite schist, and talc-serpentine schist. The units are layered within the body and the best exposures reveal an interweaving/duplication of units (Figure 2). Some shear is present in the rocks, however, only on microscale and large shear indicators are absent on outcrop scale. Duplication of units and shear indicators may suggest the body was thrusted and stacked during emplacement.

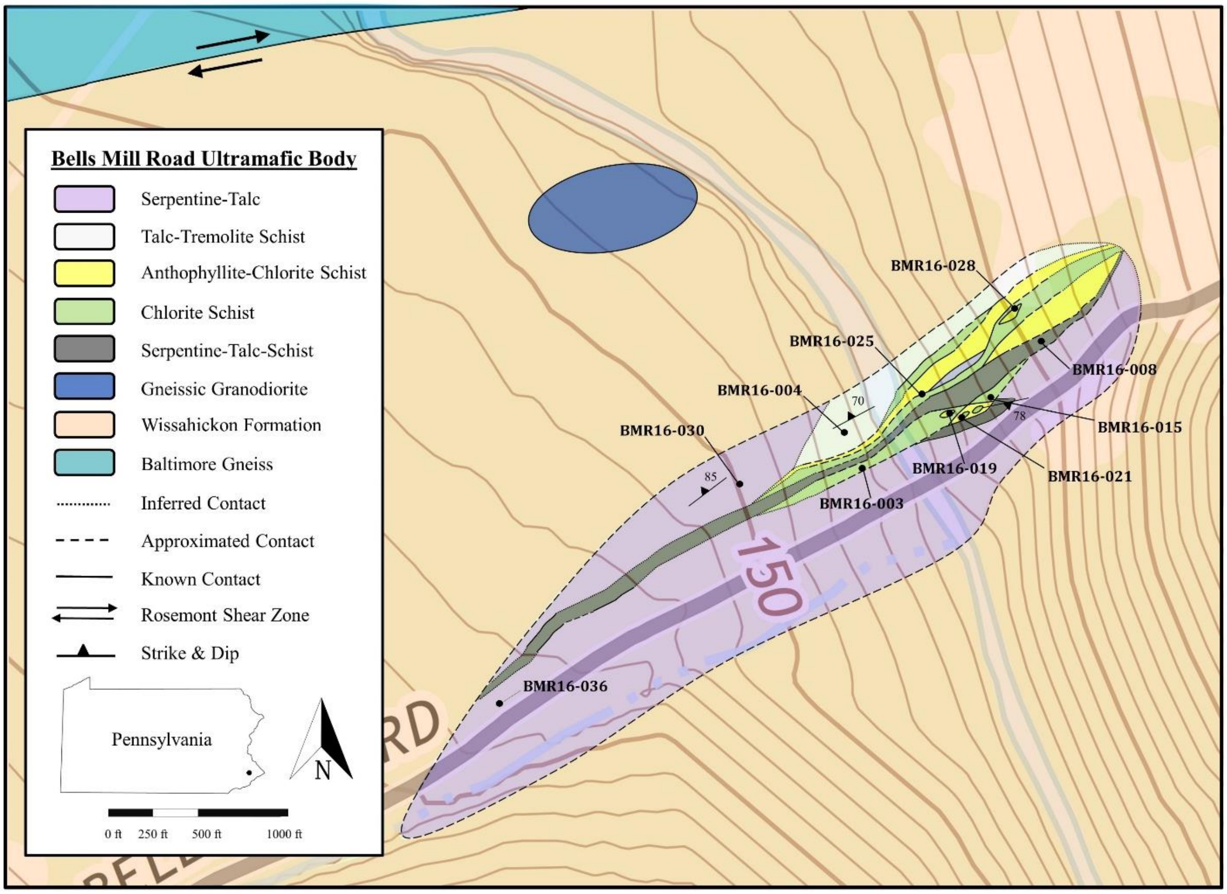

Figure 2: Geologic map of the Bells Mill Road ultramafic body. The lithologic zones of alteration are defined as: serpentine-talc, talc-tremolite, anthophyllite-chlorite schist, chlorite schist, and talc-serpentine schist (Simboli et al., 2017)

Serpentine-talc rock

The serpentine-talc rock is most abundant unit within the body and encompasses the outside edge of nearly the entire body. The serpentine-talc rock is characterized by large inclusions of serpentine surrounded by a matrix of talc, anthophyllite, and magnesite. The serpentine inclusions range in size from 1 to 10 cm and provide hand samples a spotted appearance (Figure 3). Most of the serpentine inclusions are oblong and seemingly randomly orientated. Most inclusions are rounded but, increase their angularity with increasing size. In thin section, the serpentine inclusions have mesh textures serpentine suggesting the replacement of olivine (O'Hanley, 1996). Other serpentine textures are present (i.e., chysotile

Figure 3. Serpentine-talc rock textures seen in outcrop. The black is serpentine inclusions surrounded by a matrix of talc, anthophyllite and magnesite.

veining, bastite), however, they are minor. Accessory phases within the serpentine inclusions include: magnetite, hematite, chromite, awarite, and millerite (DeSantis, 1978).

Overall, the rock is matrix-supported with most serpentine inclusions surrounded by the matrix. The serpentine inclusions can range from 30% to 60% of the rock, but most samples maintain an approximate 50% ratio of inclusion to matrix. The grey matrix is a complex network of talc, anthophyllite, magnesite with minor amounts of chlorite and opaques, likely magnetite, chromite, and ferritchromite. Talc is the dominant mineral in the matrix and defines the schistosity with dispersed acicular anthophyllite. Magnesite porphyroblasts are common but have ragged grain boundaries with clear centers.

Talc-tremolite schist

The talc-tremolite schist is a massive, fine-to-medium grained, grey rock located mainly in the northerly portions of body. In hand sample the rock's schistosity is revealed by white stringers of tremolite (Figure 4). The talc-tremolite schist appears only in the northern half the body and pinches out with many units near the center of the exposed body.

Figure 4. Talc tremolite schist hand sample view.

Anthophyllite-chlorite schist

The anthophyllite-chlorite schist consists of anthophyllite, chlorite, talc, and occasional relic orthopyroxene. The chlorite is fine grained making up the matrix with large acicular anthophyllite throughout the sample. The anthophyllite is often kinked or distorted exhibiting brittle deformation. At the contact with the chlorite schist unit, the anthophyllite becomes coarser grained splaying blades into the chlorite schist (Figure 5). The presence of orthopyroxene may be relic from the protolith.

Figure 5. The contact between chlorite schist (upper green) and anthophyllite-chlorite schist (lower beige)

Chlorite schist

The chlorite schist is a fine to medium grained green rock with several diverse layers in several locations. The majority of this lithologic unit is overwhelmingly chlorite in composition, however, some locations have scattered layers of magnesite (Figure 6), magnetite (see Spinel Mineralogy section), and Al-rich phases (Figure 7) all restricted to specific field locations and not representative of the lithology. Magnesite present

Figure 6. Thin section photomicrograph in XPL of the chlorite schist with a large magnesite grain. FOV = 1 mm

as porphyroblast up to 1 mm and occasionally exhibiting deformation twinning (Figure 6). Magnetite in the chlorite schist occurs as layers up to 10 cm thick of magnetite porphyroblasts from 0.01 to 1 cm. In one unique location there is a layer of chlorite rock containing porphyroblasts of garnet, kyanite, and corundum (Figure 7). The presence of such high-Al phases in an ultramafic body suggests a large amount of component exchange and metasomatism. The Al-rich layer occurred with a magnetite rich layer as well.

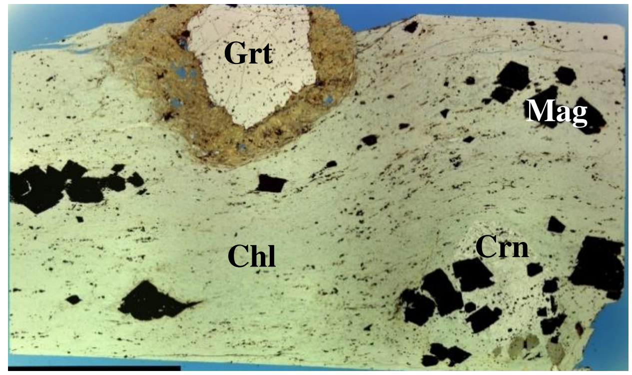

Figure 7. Thin section photomicrograph in PPL of sample BMR17-015 showing Al-rich phases within the chlorite schist. FOV = 4 cm

Talc-serpentine schist

The talc-serpentine schist is a fine to medium grained rock with magnesite scattered throughout specific locations as well as anthophyllite. The talc, serpentine and anthophyllite define the foliation while the magnesite is porphyroblastic. Most samples containing magnesite develop a strong 1 to 2 cm weathering rind that bleached the silicates a dirty bright white and a rusty red-brown replacing the magnesite. The weathered rock becomes extremely soft and crumbles with the strike of a hammer.

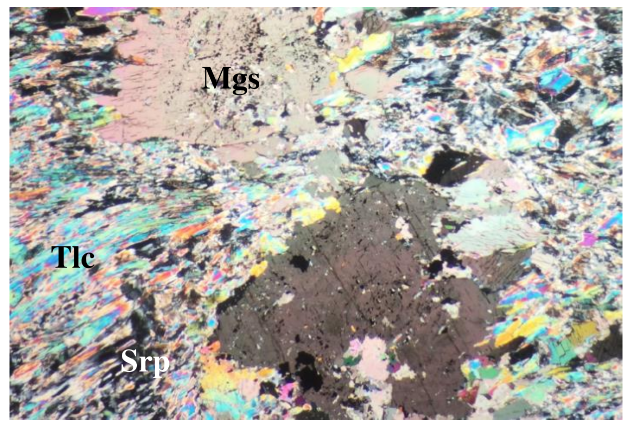

Figure 8. Thin section photomicrograph in XPL of sample BMR17-008. FOV = 1 mm

Spinel Mineralogy:

Spinel minerals, particularly magnetite and chromite, are common in mafic and ultramafic rocks over a wide range of conditions. This group of oxides, specifically chromites, was one of the first mineral systems to be used as petrogenetic indicators as reported by Irvine (1965). Chromites can be one of the first minerals to crystallize in a mafic or ultramafic magma including unfractionalized signature of the magma. Chromites are refractory and relatively resistant to alteration especially when compared to the silicate minerals within these systems. However, pervasive alteration can change inherited geochemical signatures and metamorphism must always be assessed prior to applying spinel compositions on petrogenetic diagrams.

Large Cr-rich spinels from samples showing the least apparent alteration were analyzed for geochemical evidence of tectonic origin. Cr-rich spinel were analyzed using energy dispersive spectroscopy electron probe microanalysis (EDS-EPMA) on the microprobe at the National Institute of Standards and Technology (NIST) following the methods of Mengason et al. (2017). Quantitative values were arrived at using standards-based multiple linear least squares fitting with DTSA-II software from NIST. Individual spinel grains were selected for anlaysis based on size and position in the sample (either thin section or epoxy mount). Transects of the grain in at least one direction were performed and apparent 'plateau' values were averaged for plotting in discrimination diagrams.

Spinel minerals in samples from Bells Mill Road examined were from the following lithologic units: talc-serpentine schist (BMR16-008), serpentine-talc rock (BMR16-030 and BMR16-036), and chlorite schist. All examined spinels from the chlorite schist were found to be nearly pure magnetite and therefore not plotted on the below diagrams (Figures 10, 11, and 12) as this study focused on Cr-rich spinels. Along with Bells Mill Road samples (BMR) plotted on the diagrams below were samples from other regional ultramafic bodies. Sample SRP16-011A is serpentinite from an ultramafic body near Newtown Square, PA. Sample SRP16-024 is serpentinite from an ultramafic body at Strode's Mill in southern West Chester, PA. In addition, Cr-rich spinel compositions of Smith and Barnes (2008) from the Goat Hill Serpentine Barrens in southwestern Chester county, part of the State Line Serpentinite, are plotted for comparison.

Spinels from Mid Ocean Ridges (MOR) show a more limited range in Cr:Al ratio compared to those from volcanic arcs (Arc) (Figure 9 for reference). The core compositions of spinel from nearby serpentinite units and from Goat Hill Serpentine Barrens plot largely outside of the MOR field and within the Arc field (Figure 10). However, the BMR samples are significantly Al-depleted and plot outside of all fields. This correlates well with the reported petrogenesis of the State Line deposits being closely related the Baltimore Mafic Complex (McKague, 1964; Hanan and Sinha, 1989; Smith and Barnes, 2009).

Two core compositions from BMR16-008 are more highly oxidized and depleted in Al by alteration. They can be described as 'ferritchromite' with rim compositions plotting as metamorphic magnetite overgrowth (Figure 10). The core compositions from BMR16-030 and -036 are just slightly Cr-enriched compared to their rims and have homogenized with their magnetite overgrowth (Figure 10). All of these spinels have undergone significant alteration and no-longer reflect initial compositions. However, it is possible to use their elemental concentrations to assess metamorphic grade and confirm the syn-metamorphic nature with the adjacent Wissahickon schist.

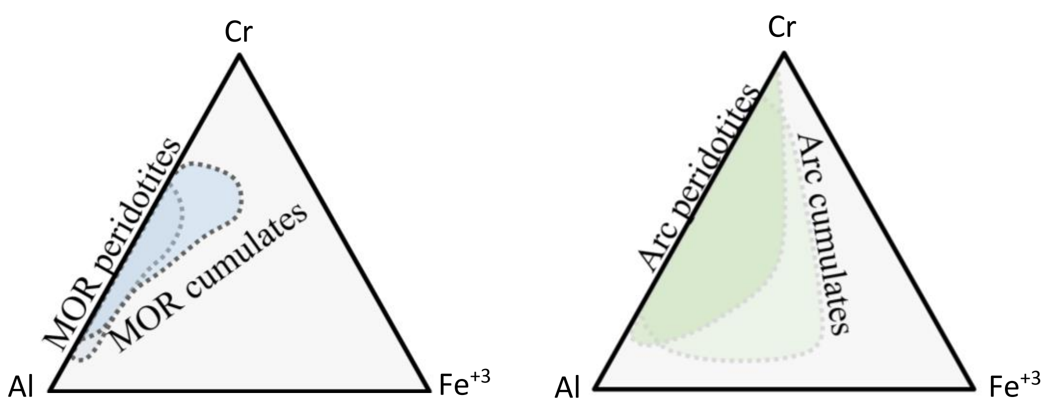

Figure 9. Fe^{3+}-Cr-Al ternary diagram in atomic % (as below). Fields encapsulate data from Arai et al. (2011)

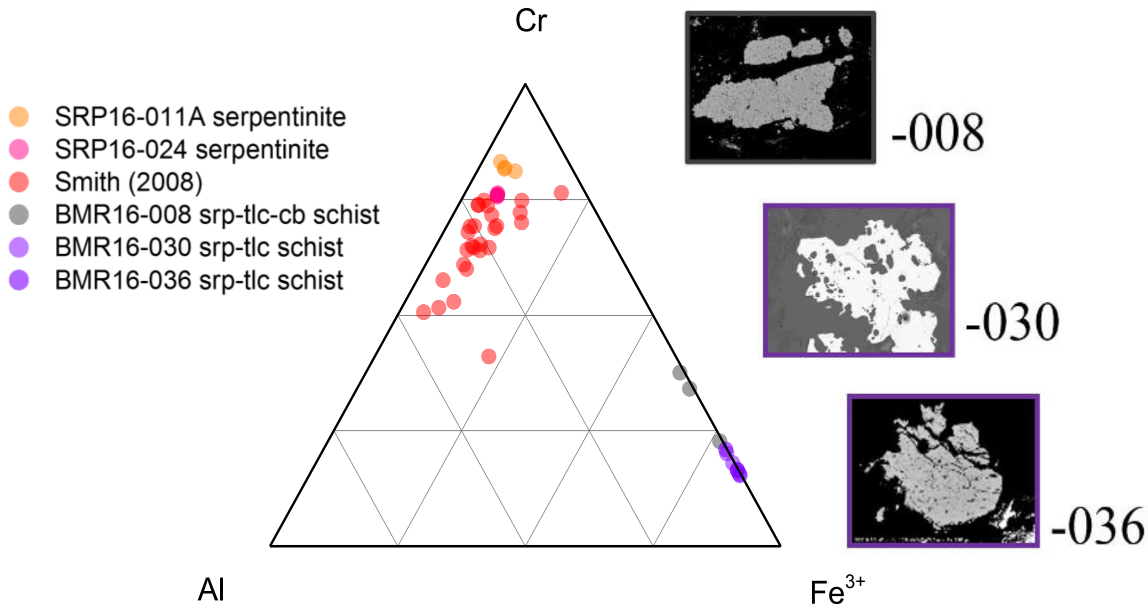

Figure 10. Core compositions taken from cross-grain transects of Cr-rich spinels. Fe^{3+} calculated from total Fe assuming charge balance and full site occupancy. Backscatter images of the spinels are shown on the right.

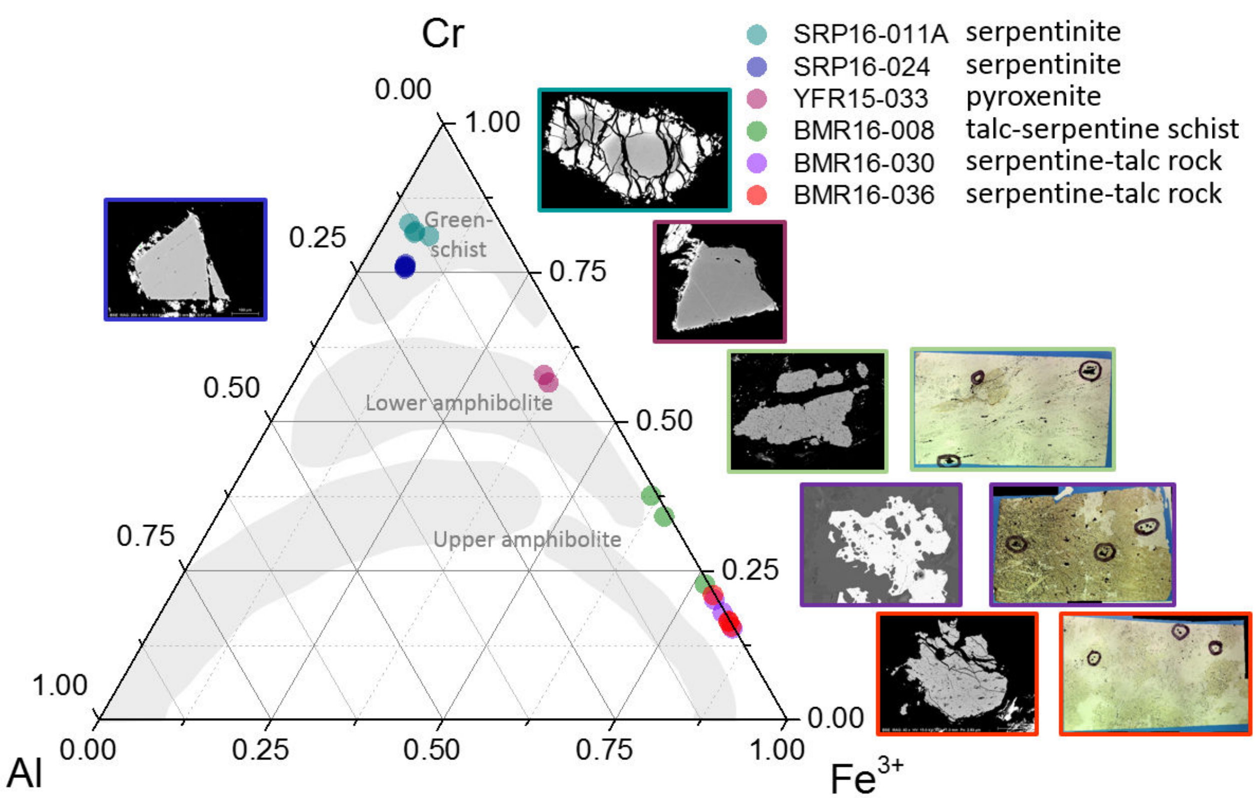

Figure 11. Chromite composition from chlorite schists of this study and related units. Data from chlorite schists plot in the field of ferritchromite and Cr-magnetite. Fields of composition for different metamorphic facies after Evans and Frost (1975) and Suita and Streider (1996).

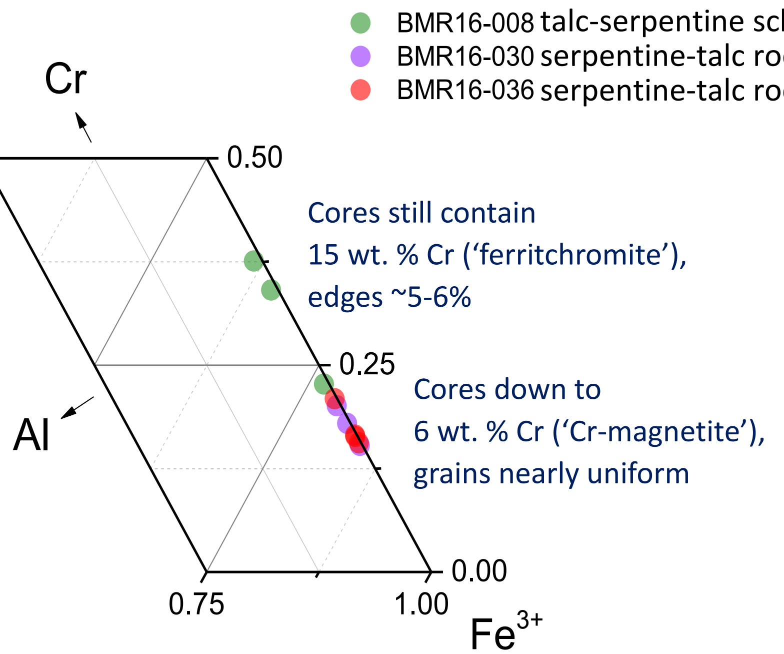

Figure 12. Cr-rich spinel in samples plotted approaches magnetite in composition. Elevated Cr concentration in cores indicates initial igneous origin and lack of full homogenization. Possibly represents initial conversion to 'ferritchromite' and later ongoing replacement by magnetite.

Spinels from Bells Mill Road and spatially related metamorphosed mafic/ultramafic rocks can give context to the metamorphic grade and history of these altered samples. Figures 11 and 12 show spinels from samples plotted with respect to reported metamorphic grades associated with spinel compositions (Evans and Frost, 1975; and Suita and Streider, 1996)

Based on plots shown above and interpretations detailed by Barnes (2000), Barnes and Roeder (2001), and Gonzalez-Jimenez et al. (2009), the generalized chromite alteration sequence is as follows:

- Low-temperature serpentinization sees some overgrowth of 'ferritchromite' or magnetite and minimal alteration of core.
- Lower Greenschist facies sees Mg loss during Fe^{2+}↔Mg^{2+} exchange with olivine, growth of "ferritchromite" rim and gradual loss of Ti, Cr, and Mg from the core to the rim progressing inward from the contact.
- Upper Greenschist facies sees effective loss of Al by fluid-mediated reaction to form chlorite along with an increase in Fe^{3+}, reflecting increasing fO_2 and the Mg^{2+}/Fe^{2+} ratio no longer buffered by olivine.
- Lower Amphibolite facies sees continued Cr, Mg, and Al loss and increase in Fe^{3+} with a shrinking miscibility gap between chromite and magnetite diminishing above 500°C and closing above ~550°C.

Geochemistry

Major and trace abundances have been measured by ICP-MS and XRF on all of the lithologies present at the Bells Mill Road ultramafic body (geochemistry of representative samples are present in Table 1). Mineral modal abundances were found by point counting >600 points per slide and are compatible with major element geochemistry. Major minerals present are represented in Table 1. All samples exhibit elevated concentrations of chromium, cobalt, and nickel, which is typical for ultramafic rocks.

Plotting the trace elements on petrogenetic discrimination diagrams is common practice in igneous petrology where inherited signals are sustained due to minimal disturbance from alteration. The rocks of the Bells Mill ultramafic body have been significantly altered through possibly multiple stages of deformation, making it difficult to plot on discrimination diagrams, particularly if the elements plotted are mobile/incompatible. Application of discrimination diagrams to metamorphic systems requires the use of elements that are relatively immobile/compatible within rock. Although large fluxes of fluid significantly alter the parent rocks, some trace element concentrations can remain unaltered retaining geochemical signatures that are inherited from their initial crystallization (Deschamps et al., 2013).

A combination of igneous and metamorphic petrogenetic discrimination diagrams have been used to plot the trace elements of relatively immobile elements for mafic to ultramafic systems to determine the potential origin of the Bell Mill Road ultramafic body. Two mafic igneous systems designed for basaltic systems (Figure 14: i.e., Shervais, 1981 and Pearce and Cann, 1971) and two metamorphic systems for serpentinites (Figure 15: i.e., Deschamps et al., 2013) have been used to test hypotheses of possible protolith origins. Geochemical discrimination diagrams can be used to identify tectonic origin. These diagrams are not specifically designed for ultramafics, however, consistent results were observed. It can be inferred that the source of magma originated from a subducted island arc setting. Figure 14 shows several samples plotting in the MORB field suggesting an ophiolite setting for the

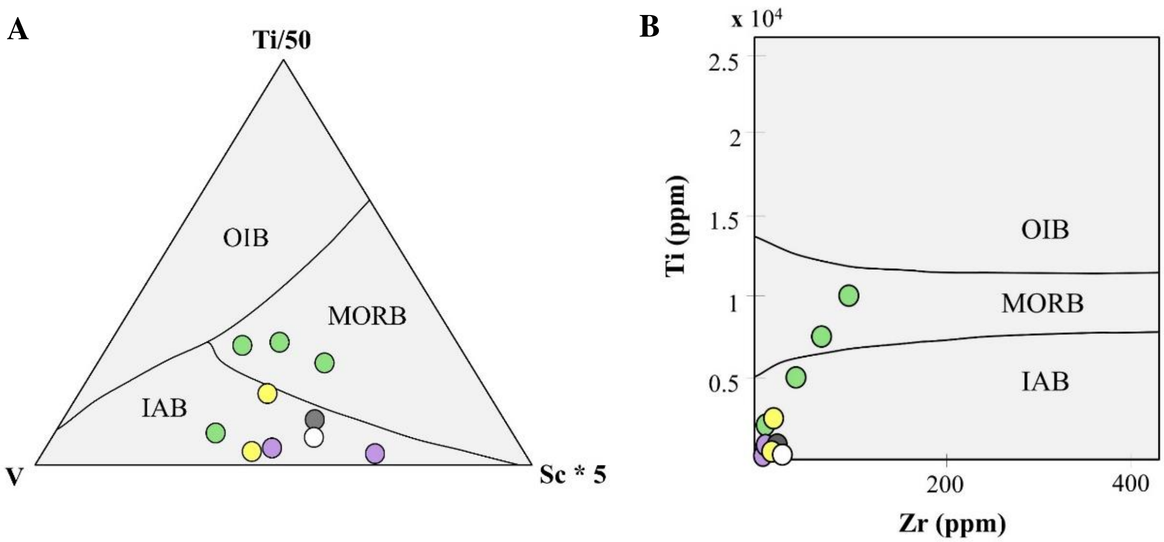

Figure 14: Petrogenetic discrimination diagrams for basaltic igneous systems (**A**) after Shervais, 1981; (**B**) after Pearce and Cann, 1971. IAB – Island Arc Basalt (arc related magmatism), OIB – Ocean Island Basalt (mantle derived hot spot activity), and MORB – Mid-Oceanic Ridge Basalt (oceanic floor derived ophiolite). Colors correspond to mapped lithologies in Figure 2: purple – serpentine-talc rock, white – talc-tremolite schist, yellow – anthophyllite-chlorite schist, green – chlorite schist, and grey – talc-serpentine schist.

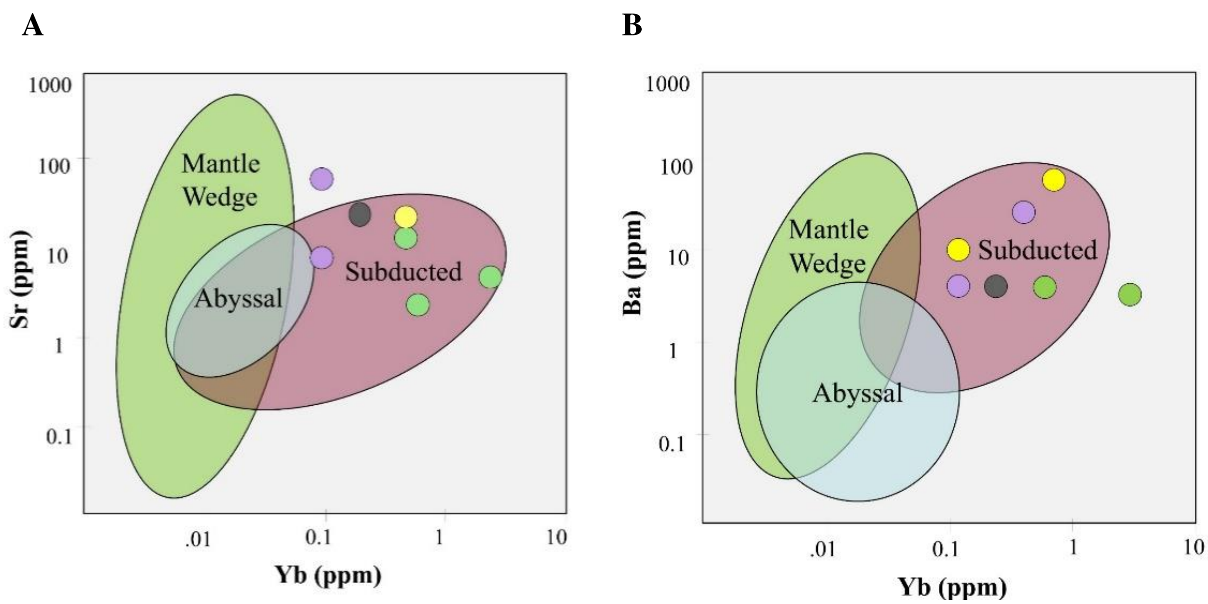

Figure 15: Petrogenetic discrimination diagrams for serpentinite systems (**A & B)** after Deschamps et al., 2013. Abyssal serpentinites represent oceanic floor derived ophiolite, Mantle wedge represent diapiric mantle derived material, and subducted represent arc related magmatism. Colors are described in the figure above.

protolith. However, all samples plotting in the MORB field were from the chlorite schist, which display significant metasomatism/component exchange as evidenced by the presence of garnet, corundum, and kyanite in some samples. The presence of these minerals in limited zones suggests localized significant component exchange with fluids derived from the Al-rich country rock. It is unclear why this would drive compositions toward the MORB field but they did have a consistent deviation.

Conclusions

The Bells Mill Road ultramafic body shows several distinct alteration zones with the following lithologies: serpentine-talc; serpentine-talc-carbonate; talc-tremolite; chlorite schist; and anthophyllite-chlorite schist. Shearing appears present exhibited by duplication of units and shear-induced microstructures (shear banding, pressure shadows, deformation twins, etc.). Chromites have undergone significant alteration and no-longer reflect initial compositions. A small minority of samples show a mid-ocean ridge basalt affinity, but these rocks are rich in Al-phases (garnet and corundum) which may suggest a high level of recrystallization/metasomatism altering inherited signals. Interpretations of petrogenetic discrimination diagrams suggests that the ultramafic body is most likely a differentiate of an arc-related magma chamber.

	BMR16-003 Chlorite schist	BMR16-004 Talc-Trem schist	BMR16-008 Talc-Serpentine schist	BMR16-028 Anthophyllite Chlorite schist	BMR16-036 Serpentine-Talc rock
Major Elements in Wt%					
SiO_2	23.22	53.45	39.48	42.87	41.41
Al_2O_3	12	3.13	1.22	7.34	1.21
Fe_2O_3(T)	29.77	6.5	6.92	8.7	7.11
MnO	0.046	0.089	0.053	0.131	0.055
MgO	24.27	25.09	32.1	29.25	34.97
CaO	0.01	6.38	1.22	1.28	0.35
Na_2O	< 0.01	0.05	0.02	0.03	0.02
K_2O	< 0.01	< 0.01	< 0.01	< 0.01	< 0.01
TiO_2	0.297	0.048	0.074	0.366	0.017
P_2O_5	< 0.01	< 0.01	< 0.01	0.02	< 0.01
LOI	9.04	4.67	18.08	8.73	13.54
Total	98.65	99.42	99.16	98.72	98.69
Trace Elements in ppm					
Cr	20	2220	2200	1620	2060
Co	100	68	80	79	96
Ni	150	1280	1700	1540	2150
Sc	36	12	9	20	7
V	343	49	36	122	40
Ba	< 2	5	5	29	< 2
Sr	< 2	9	32	31	10
Y	< 1	15	2	5	< 1
Zr	5	15	3	10	< 2
Zn	70	60	40	60	30
Ga	10	4	2	7	1
Nd	0.6	3.3	0.5	1.3	0.2
Sm	0.2	1.5	0.2	0.4	< 0.2
W	2	2	2	2	2
Mineral Modal Abundances in %					
Srp	-	6%	20%	-	53%
Chl	90%	-	1%	34%	4%
Tlc	-	49%	40%	17%	25%
Ath	-	15%	18%	42%	13%
Tr	-	27%	-	-	-
Opx	-	3%	-	5%	-
Mgs	-	-	19%	1%	4%
Opaques	10%	-	2%	1%	1%

Table 1: Geochemistry and modal mineral abundances of representative samples from each of the lithologic zones at the Bells Mill Road ultramafic body.

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The 201-Ma Magma Plumbing System in the Western Newark Basin: How and (maybe) Why it's Different from the East

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ABSTRACT

The Newark basin (PA-NJ-NY) is part of the Central Atlantic Magmatic Province and contains Triassic-Jurassic sediments and mafic igneous rocks that formed at about 201 Ma during Mesozoic extension and rifting of Pangaea. Compared with the more well-known eastern Newark basin, the western basin is narrower; contains heterogeneous, coarse-grained alluvial fan sediments; and intrusions are less laterally-extensive and cut further up-section. Different underlying Paleozoic structures and orientations of the border fault influenced Triassic sedimentation and later deformation resulting in greater tilting, folding, and deeper erosion that exposes natural cross-sections through the upper-crustal plumbing system. The Morgantown "sheet" is actually a set of interconnected sills, inclined sheets, and dikes similar to but larger than the Jacksonwald intrusive complex; estimates of intrusion emplacement depth range from <0.5 km to >5.7 km. The variety of sizes and orientations indicates that intrusion structures were controlled more by magma volume, overpressure, and crustal heterogeneities than by extensional tectonic stresses. Sites of phenocryst accumulations are used to identify three feeder structures that supplied very similar but distinct pulses of magma into the upper-crustal complexes. Chill margin and lava flow samples are virtually identical in whole-rock rare-earth element and phenocryst composition, suggesting they formed from the same magma. This, along with the distribution of cumulates in deeper intrusions and more evolved and altered rocks in shallower intrusions, indicates a high degree of connectivity and that these sill-sheet-dike complexes operated as a system rather than as isolated intrusions.

INTRODUCTION

The architecture of magmatic plumbing systems is fundamental to understanding earth processes and volcanic hazard assessment (e.g., Putirka, 2017). Seismic imaging beneath active volcanic areas (e.g., Huang et al., 2015; Greenfield et al., 2016) is interpreted to show magma chambers at different depths with low percentages ($\leq$ 10%) magmatic liquid, which must operate as interconnected transport systems during times of eruption. Field studies remain critical in providing a more complete picture of sub-volcanic intrusive networks because seismic methods typically cannot image igneous intrusions that are small or thin or oriented close to vertical (dikes). As reviewed by Magee et al. (2016), sills and sill complexes are significant conduits for lateral magma transport, especially in upper-crustal plumbing systems. Well-exposed mafic intrusive complexes such as the Ferrar dolerites in Antarctica and the Karoo basin, South Africa, provide models for intrusion geometry and magma transport (e.g., Muirhead et al., 2012; Polteau et al., 2007) and crystallization processes (e.g., Marsh, 2004; Bédard et al., 2007; Galerne et al., 2010; Neumann et al., 2011). Petrologic and geochemical data that constrain magma emplacement, transport, and crystallization can provide an essential link between the volcano and its intrusive network, and between active and fossil magmatic systems (e.g., Cooper, 2017; Puffer et al., 2009).

This paper focuses on the sub-volcanic magmatic network in the Newark basin, Pennsylvania, part of the Central Atlantic Magmatic Province (CAMP) that formed at about 201 Ma associated with the rifting of Pangaea and the opening of the modern Atlantic Ocean (McHone, 2000). The CAMP has the greatest area of any Large Igneous Province on Earth and an estimated volume of 2-3 x 10^6 km^3 of basaltic magmas (Marzoli et al., 1999; McHone, 1996). Basaltic dikes are numerous in the Precambrian and Paleozoic basement rocks of the Eastern North America (ENA) portion of the CAMP, but other intrusive structures are almost exclusively found within the late Triassic-Jurassic rift basins (Froelich and Gottfried, 1985). The Newark basin (Figure 1) is one of the most comprehensively-studied of the ENA continental rift basins, but the western basin has had less research than the east. According to Withjack et al. (2013), the Newark basin opened in the Late Triassic as a narrow, asymmetric half-graben and progressively widened over time into a continuous basin extending from Virginia to Connecticut by the beginning of CAMP magmatism at ca. 201.5 Ma (Blackburn et al., 2013; Marzoli et al., 2011). Deformation before, during, and after CAMP magmatism tilted the sedimentary and igneous rocks towards the northwest border fault and produced transverse plunging synclines and anticlines (Schlische, 1992; Withjack et al., 2013), including the Jacksonwald syncline in the western Newark basin. Subsequent erosion removed several kilometers of sedimentary and igneous rocks (Withjack et al., 2013, and references therein). Froelich and Gottfried (1985, p. 84) speculated that "a map view of the tilted, asymmetrical, deeply-eroded early Mesozoic basins of the Eastern United States offers a natural tangential cross-section."

By taking into account the structural effects of post-magmatic folding and northwesterly tilting we show that the Morgantown "sheet" and Jacksonwald syncline in the western Newark basin are in fact interconnected sill-sheet-dike complexes that functioned as magmatic systems (Srogi et al., 2014; Martinson et al., 2015), similar to the Palisades intrusive complex in the eastern Newark basin (Husch, 1992; Puffer et al., 2009; Block et al., 2015). The problem with continuing to use the term "sheet" to describe an intrusive that is "composed of multiple, interconnected and irregular bodies with ring-like outcrop patterns" (Daniels, 1985, p. 128) is that a "sheet" implies a single intrusion with one floor, one roof, both lateral and vertical continuity, and a single magmatic history. If instead there are several smaller intrusive bodies, some sill-like and some inclined sheets and dikes, then there are multiple floors, roofs, and sides at different depths. The connectivity and continuity of magma flow could be intermittent and variable over time and space, and each intrusion could have a distinct history of cooling, replenishment, melt flow and crystallization. In short, geometry matters. We quantify structural relief and emplacement depths in the western Newark basin for the first time and construct a more accurate model of the diabase intrusive structures including feeder dikes. We synthesize a variety of published and new information to investigate factors that control intrusion geometry and the transport of material within the intrusive network.

GEOLOGIC SETTING AND PREVIOUS WORK

The Mesozoic rift basins at their present erosional level include the Gettysburg basin (MD-PA) and the Newark basin (PA-NJ-NY), connected by the Narrow Neck (Figure 1). These boundaries have been defined differently (e.g., Robinson and Froelich, 1985; Schlische, 1992), but the basins were continuous during the time of CAMP magmatism (Withjack et al., 2013). The southwestern margin of the Newark basin narrows in two abrupt steps to the Narrow Neck (Figure 1), partly the result of greater erosion of an asymmetric basin that narrows with depth (Faill, 2003; Withjack et al., 2013). The central and eastern Newark basin contains laterally-extensive lacustrine sediments of the Lockatong and Passaic Formations (Olsen et al., 1996), but at the western end the Lockatong Fm pinches out and the Passaic Fm contains inter-digitating coarse-grained sandstones and conglomerates deposited in an alluvial fan environment (Figure 2). Pennsylvania maps (Figures 2 and 3) use Brunswick Fm (TRb) for lacustrine Passaic Fm and Hammer Creek Fm (TRh, TRhc) for alluvial fan sediments in the Passaic Fm, (Lyttle and Epstein, 1987).

In the western Newark basin (Figure 2), at least five intrusions and the Jacksonwald basalt are folded by and exposed along the well-known Jacksonwald syncline (Olsen et al., 1988; Schlische, 1992). The Morgantown "sheet" has the 250-meter-wide Birdsboro dike (Figure 2) forming its northeast side (Smith, 1973; Smith et al., 1975; Froelich and Gottfried, 1999) and a series of irregularly-shaped segments that extend laterally to the west and up-section across the entire basin. Diabase intrudes the basement contact along the southeast margin (St Peters sill, Figure 2), while intrusions near the border fault cut Triassic sediments corresponding to the middle of the Passaic Fm (Faill, 2003; Olsen et al., 1996). A discontinuity in the sedimentary and igneous units on either side of the Birdsboro dike results from a fault

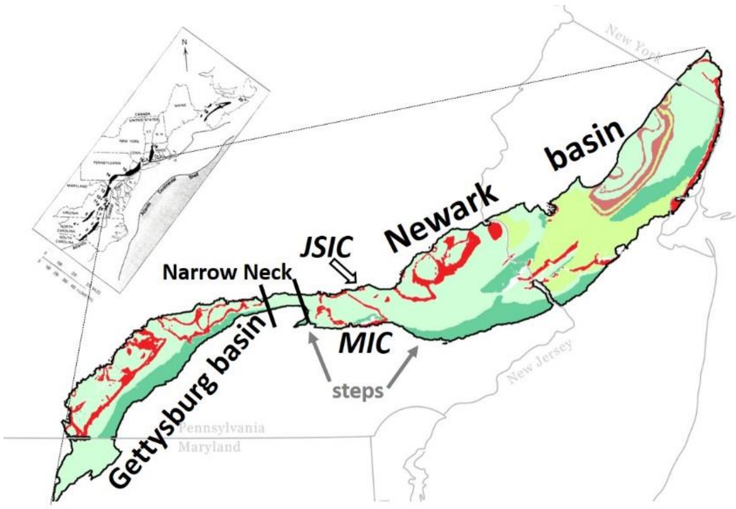

Figure 1. Generalized geologic map of the igneous rocks (red and brown colors) and Mesozoic sedimentary rocks (shades of green) in the Newark and Gettysburg basins showing locations of the Morgantown intrusive complex (MIC) and Jacksonwald Syncline intrusive complex (JSIC). Steps in the southeastern margin of the western Newark basin are indicated by gray arrows. Modified from Withjack et al. (2013) and Dickin et al. (2005). Inset: Mesozoic rift basins of the Eastern North American portion of the CAMP, modified from Robinson and Froelich (1985).

along the northeast side of the dike (MacLachlan, 1981; Schlische, 1992, Figure 3, cross-section D-D´) that dropped the Jacksonwald syncline rocks down on the northeast side and uplifted the Morgantown rocks on the southwest side. Wise (2014) regards this as a reactivated Paleozoic zone of crustal weakness trending NW-SE, the Schuylkill Tectonic Zone, where two major Paleozoic allochthons come together in the subsurface. This produced the intra-basin high and alluvial fan sedimentation during the Triassic (Schlische and Withjack, 2005), the NW-SE-striking fault along the Birdsboro dike, and the steps along the southern margin of the western Newark basin (Wise, 2014). The fault displacement, along with the northwest-plunging fold, resulted in less erosion of the northeast side and preservation of the Jacksonwald lava flow and overlying Jurassic Feltville Fm. On the southwest side, all units were tilted towards the border fault and sedimentary units younger than the middle-to-late Passaic Fm (Hammer Creek Fm) were eroded (Olsen et al., 1996; Faill, 2003), exposing diabase intruding basement rocks and Stockton Fm at deep structural levels (Srogi et al., 2010, 2012, 2014).

Diabase intrusions in the Mesozoic rift basins from Virginia to New Jersey have been described as “sheets,” but this interpretation does not hold in the western Newark basin. The interpretation of a lens- to saucer-shaped intrusive form is based on roughly elliptical outcrop patterns, limited drill-core and geophysical data, and by analogy to more well-exposed examples such as the Karoo basin (Hotz, 1952; Smith et al., 1975; Froelich and Gottfried, 1999). Residual Bouguer gravity anomaly ***highs*** forming concentric closed-loop patterns centered over ring-shaped outcrop are consistent with buried, saucer- or lens-shaped sheets (Hersey, 1944; Sumner, 1977; Daniels, 1985). However, concentric, closed-loop gravity anomaly ***lows*** led Daniels (1985, p. 128) to conclude that the “Morgantown body is an exception…” and that “diabase is absent or very thin in the subsurface beneath the center of the ring;” thus, it is not a buried saucer-shaped sheet. The Birdsboro dike that dips 80°SW along the east side of the Morgantown “sheet” is not the same as the gently- to moderately-dipping inclined sides of saucer-shaped sheets (e.g., Polteau et al., 2007). Whereas individual sheets in well-studied sill complexes rarely cut up-section more than two kilometers vertically (e.g., Magee et al., 2016; Muirhead et al., 2012; Cartwright and Hansen, 2006), the Morgantown “sheet” traverses the entire width of the western basin stratigraphy from the basement to middle-to-late Passaic Fm. At least five intrusions are exposed along the axis of the plunging Jacksonwald syncline at different depths beneath the basalt lava flow nearest the border fault (Figure 2); individually they may be sheet-like, but they are probably interconnected as a sill complex not a single sheet. We use the term “intrusive complex” to refer to the entire set of interconnected intrusions; i.e., the Morgantown intrusive complex and the Jacksonwald intrusive complex.

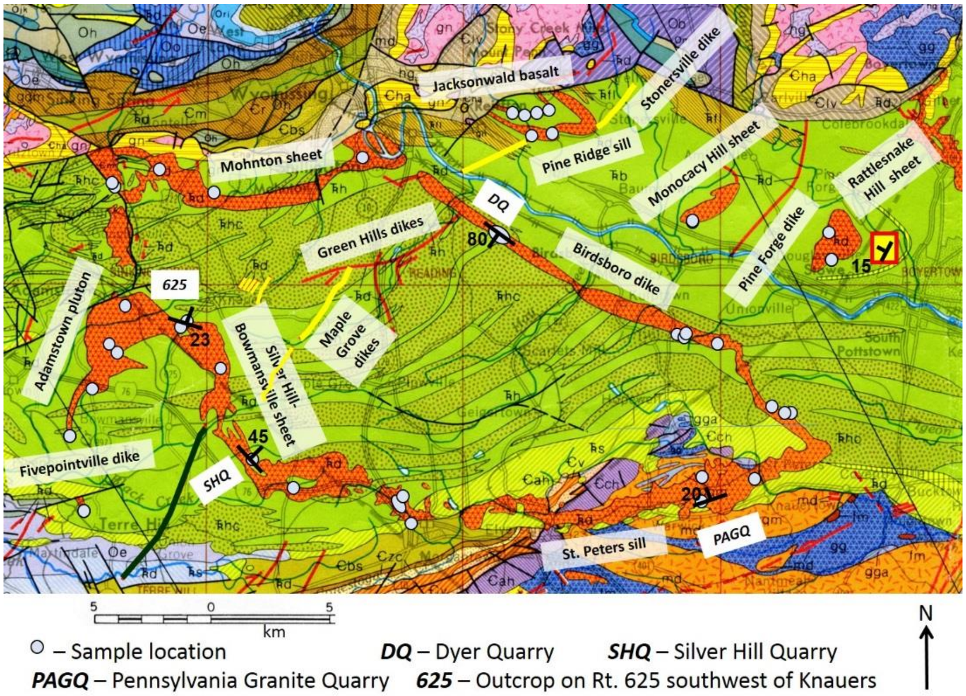

Figure 2. Geologic map of the Morgantown and Jacksonwald areas (Berg et al., 1980) showing igneous unit names, locations, and orientations discussed in the text. All HTQ (York Haven-type) tholeiite is red; LTQ (Rossville-type) tholeiite dikes are yellow; olivine-normative (Quarryville type) dike is dark green (after Smith et al., 1975).

METHODS

All sample locations to date are shown on Figure 2 (light blue symbols). Care was taken to sample diabase with the least possible weathering at each locality; very fresh samples were obtained from quarries and roadcuts. Thin sections from outcrop samples are oriented.

Geochemical analyses were made at Duke University by direct current plasma emission spectrometry (DCP, Fisons SpecterSpan 7) or The College of Wooster by XRF (4-kW Rh-target Rigaku ZSX Primus II). Low-abundance trace elements for DCP samples were measured by inductively-coupled plasma mass spectroscopy (ICP-MS, VG PlasmaQuad 3) at Duke University. Analytical procedures are described in Pollock et al. (2014). Rare-earth element concentrations are shown in Table 1.

Semi-quantitative mineral analyses by energy-dispersive spectrometry (EDS) and backscattered electron (BSE) images were obtained on polished, carbon-coated thin sections using the FEI Quanta 400 Environmental Scanning Electron Microscope (SEM) at West Chester University. Analyses made from 2004 to April 2017 used Oxford Inca EDS with SiLi x-ray detector; subsequent analyses were made with Oxford Aztec EDS with a silicon-drift detector (SDD). The accuracy, precision, and reproducibility of EDS results are judged by repeat analyses and by comparing with wavelength- dispersive spectrometry (WDS) analyses of the same samples using a Cameca SX-50 electron probe micro-analyzer (EPMA) at the University of Massachusetts-Amherst. Operating conditions for the Oxford Inca EDS and Cameca SX-50 WDS analyses can be found in Willis et al. (in press). For the Aztec analyses by SDD, a spot size of 5.5-5.6 was used to generate x-ray counts of 35-40 kcps with dead time of ~35%, as recommended by Oxford; count rates and dead time were monitored and did not change significantly during analysis. Pyroxene analyses were made by setting the total x-ray counts to 500,000; dwell time was on the order of 5-15 sec. Magnification was 400x or higher and the scanned area dimension was on the order of a few µm for point analyses. For pyroxenes with exsolution, the box tool was used to define a scan area a few µm on each side;

Table 1. Rare-earth element concentrations in basalt and chill margin samples. RHS: Rattlesnake Hill sheet, Douglassville Quarry. DQ: Birdsboro dike, Dyer Quarry. Normalizing values from Sun and McDonough (1989).

Sample		**La**	**Ce**	**Pr**	**Nd**	**Sm**	**Eu**	**Tb**	**Gd**	**Dy**	**Ho**	**Er**	**Yb**	**Lu**
BiQ-151	basalt near top	9.5	21.9	3.0	12.7	3.3	1.1	0.7	3.9	3.9	0.8	2.2	2.0	0.3
BiQ-123	basalt, middle	10.0	22.1	3.0	13.2	3.4	1.1	0.7	4.0	3.9	0.8	2.2	2.1	0.3
BiQ-129	basalt, base	8.6	19.5	2.7	11.6	3.0	1.0	0.6	3.5	3.6	0.7	2.0	1.9	0.3
BoQ-10	chill, RHS	9.7	21.4	2.9	12.8	3.3	1.1	0.6	3.7	3.8	0.8	2.2	2.1	0.3
BoQ-11	chill, RHS	10.1	22.3	3.1	13.1	3.4	1.1	0.7	3.9	3.9	0.8	2.3	2.1	0.3
BiQ-b6-40	chill, DQ	10.3	22.9	3.1	13.5	3.5	1.1	0.7	4.0	4.1	0.9	2.4	2.2	0.3
BiQ-b6-47A	chill, DQ	10.5	21.9	3.1	13.6	3.4	1.1	0.7	4.0	4.0	0.8	2.3	2.1	0.3
	normalizing value	*0.237*	*0.612*	*0.095*	*0.467*	*0.153*	*0.058*	*0.0374*	*0.2055*	*0.254*	*0.0566*	*0.1655*	*0.17*	*0.0254*
		La_N	**Ce_N**	**Pr_N**	**Nd_N**	**Sm_N**	**Eu_N**	**Tb_N**	**Gd_N**	**Dy_N**	**Ho_N**	**Er_N**	**Yb_N**	**Lu_N**
		42.2	36.1	31.9	28.2	22.2	19.2	17.7	19.2	15.5	14.5	13.4	12.4	12.6
		36.4	31.9	28.4	24.8	19.9	17.5	16.2	17.3	14.2	13.2	12.1	11.1	11.4
		40.1	35.7	31.4	27.3	21.8	19.3	17.5	18.9	15.4	14.0	13.4	11.8	12.5
		41.1	35.0	30.7	27.3	21.7	18.7	17.2	18.1	15.2	14.1	13.4	12.2	12.5
		42.5	36.5	32.2	28.1	22.0	18.9	17.5	18.9	15.5	14.4	13.7	12.4	12.5
		43.7	37.4	33.0	28.8	22.8	19.5	18.0	19.3	16.3	15.2	14.4	12.7	13.2
		44.4	35.8	32.7	29.1	22.1	19.3	18.0	19.4	15.8	14.8	13.9	12.5	12.9

this is analogous to using a defocused probe beam on an EPMA. The level of x-ray spectra processing (unitless “process time”) was set to 4 or 5 as recommended by Oxford. Routine calibration and beam current monitoring was performed by analyzing a Co metal standard at the beginning of the analyses and about every two hours. Element abundances in weight percent were calculated by the software with oxygen by stoichiometry and normalized to 100%. The detection limit for any individual element is 0.1 wt.%. Pyroxene formulas were recalculated to 4 cations. We use the cation ratio [Mg/(Mg+Fe+Ca)] to compare pyroxene compositions in different samples. Cation ratios are more robust and have less error than absolute cation concentrations, and our EDS and WDS molar cation ratios are the same within ≤1 mol.%. Average core analyses of pyroxenes in the Jacksonwald basalt and intrusions chill margins are in Table 2. Orthopyroxene (opx) occurs only as inclusions (i) inside augite (aug) phenocrysts in most samples, but opx phenocrysts (p) were analyzed where present (Table 2).

RESULTS

Within each intrusive complex we use the term “sheet” for a tabular intrusion inclined at an angle between 10-70°; “sill” for a sub-horizontal, tabular intrusion even if locally discordant to bedding; “dike” for a steeply-dipping (>70°) tabular intrusion; and “pluton” for an intrusion with a more complex shape where individual sill-like and sheet-like portions can be recognized but not separated (terminology similar to Muirhead et al., 2012). On Figure 2, existing names are used for the Rattlesnake Hill sheet, Monocacy Hill sheet, Pine Ridge sill (sheet), and Jacksonwald basalt (Gottfried et al., 1991), and the Birdsboro dike (Smith, 1973; Smith et al., 1975). Additional informal names are assigned as needed.

Table 2. Average pyroxene core compositions analyzed by EDS (*n=no. of pts averaged*). **BiQ**: Jacksonwald basalt. **Chill margins**: **MH**: Monocacy Hill. **BoQ**: Rattlesnake Hill sheet. **MoQ**: Silver Hill Quarry. **EQ**: St Peters sill. **THQ-295**: Fivepointville dike. **THQ-260**: 625 locality. **i**: opx inclusion; **p**: opx phenocryst.

	ORTHOPYROXENE									AUGITE											
	i	**i**	**i**	**p**	**i**	**i**	**p**	**p**	**i**												
Element (wt. %)	**BiQ-123**	**BiQ-b6-40**	**MH-1**	**BoQ-10**	**BoQ-10**	**BoQ-11**	**MoQ-227**	**EQ-329**	**THQ-295**	**BiQ-123**	**BiQ-129**	**BiQ-151**	**BiQ-b6-40**	**BiQ-b6-47A**	**BiQ-b5-137**	**MH-1**	**BoQ-10**	**BoQ-11**	**MoQ-227**	**THQ-295**	**THQ-260**
Na												0.1			0.2		0.1		0.2	0.2	0.2
Mg	16.8	18.0	18.4	16.9	16.8	16.7	18.1	17.8	17.8	10.6	10.6	10.4	11.2	10.2	10.9	11.3	10.8	10.6	11.3	11.2	11.0
Al	0.6	1.2	0.6	0.9	1.1	0.9	1.1	0.8	1.3	1.5	1.4	1.2	1.2	1.5	1.6	1.6	1.4	1.3	1.1	1.4	1.5
Si	26.8	26.1	26.4	26.4	26.6	26.7	25.7	26.3	25.8	25.1	25.0	25.4	25.0	25.0	24.6	24.7	25.4	25.5	24.6	24.7	24.7
Ca	1.8	1.9	1.8	1.9	1.8	1.9	1.8	1.8	1.9	12.8	12.9	13.2	13.4	12.9	13.7	13.1	12.3	12.5	12.9	13.1	13.6
Ti		0.1	0.1	0.1			0.1	0.1	0.1	0.2	0.3	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
Cr	0.3	0.4	0.2	0.4	0.4	0.3	0.4	0.2	0.5	0.8	0.7	0.7	0.5	0.6	0.7	0.6	0.5	0.6	0.5	0.5	0.6
Mn		0.2	0.2	0.2		0.3	0.2	0.2	0.2	0.1			0.2	0.1	0.1	0.1	0.2	0.1	0.1	0.2	0.1
Fe	8.4	8.3	8.1	7.8	7.9	8.0	7.8	9.2	8.2	5.1	5.1	4.8	5.6	5.6	5.0	5.3	5.1	5.1	5.3	5.5	5.2
O	45.4	43.7	44.3	45.3	45.5	45.4	44.7	43.6	44.3	43.9	43.9	44.0	42.7	43.8	43.1	43.0	44.1	44.1	44.1	43.2	42.9
Total	100.0	100.0	100.0	100.0	100.0	100.2	100.0	100.0	100.0	100.1	100.0	99.9	100.0	100.0	100.2	100.0	100.0	100.1	100.2	100.2	100.1
moles																					
Si	2.04	1.93	1.96	2.01	2.02	2.04	1.93	1.96	1.93	1.97	1.97	1.99	1.93	1.97	1.91	1.91	1.99	2.00	1.93	1.92	1.91
Al	0.05	0.09	0.05	0.07	0.09	0.07	0.09	0.06	0.10	0.12	0.11	0.09	0.10	0.12	0.13	0.13	0.12	0.11	0.09	0.11	0.12
Ti	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Fe	0.32	0.31	0.30	0.30	0.30	0.31	0.29	0.34	0.31	0.20	0.20	0.19	0.22	0.22	0.19	0.21	0.20	0.20	0.21	0.22	0.20
Mn	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.01	0.01	0.00	0.00		0.01	0.01	0.00	0.01	0.01	0.00	0.00	0.01	0.01
Cr	0.01	0.02	0.01	0.02	0.02	0.01	0.02	0.01	0.02	0.03	0.03	0.03	0.02	0.02	0.03	0.03	0.02	0.03	0.02	0.02	0.03
Mg	1.48	1.54	1.58	1.49	1.48	1.47	1.57	1.52	1.54	0.96	0.96	0.95	1.00	0.93	0.98	1.01	0.98	0.96	1.03	1.01	0.99
Ca	0.09	0.10	0.09	0.10	0.10	0.10	0.09	0.10	0.10	0.70	0.71	0.72	0.72	0.71	0.75	0.71	0.67	0.69	0.71	0.71	0.74
Na										0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00
O	6.08	5.68	5.78	6.06	6.07	6.08	5.89	5.68	5.80	6.05	6.05	6.06	5.78	6.05	5.87	5.83	6.06	6.08	6.08	5.89	5.83
	n=2	*n=3*	*n=3*	*n-3*	*n-3*	*n=4*	*n=8*	*n=8*	*n=5*	*n=5*	*n=1*	*n=5*	*n=6*	*n=3*	*n=10*	*n=10*	*n=3*	*n=7*	*n=9*	*n=16*	*n=8*
Mg/(Mg+Fe)	**0.82**	**0.83**	**0.84**	**0.83**	**0.83**	**0.83**	**0.84**	**0.82**	**0.83**	**0.83**	**0.83**	**0.83**	**0.82**	**0.81**	**0.83**	**0.83**	**0.83**	**0.83**	**0.83**	**0.82**	**0.83**
Mg	78.1	78.9	79.9	78.9	78.9	78.3	80.2	77.6	79.0	51.5	51.5	50.9	51.5	49.9	51.0	52.4	52.8	52.1	52.7	52.1	51.2
Fe	17.0	16.0	15.2	15.8	16.1	16.4	14.9	17.5	15.9	10.8	10.7	10.2	11.2	12.0	10.1	10.7	10.9	10.8	10.8	11.2	10.4
Ca	5.0	5.1	4.8	5.3	5.1	5.3	4.8	4.9	5.1	37.7	37.8	38.9	37.3	38.1	38.9	37.0	36.3	37.2	36.5	36.8	38.4

For the Jacksonwald syncline we measured the distance of each intrusion in Google Earth™ along the fold axis (bearing 292 or N68W) to estimate the original depth below the paleo-surface marked by the base of the Jacksonwald basalt assuming a 15° plunge (Olsen et al., 1988). Results are shown in Figure 3. We also estimated the thickness of each intrusion taking the slope of the earth's surface into account: the Rattlesnake Hill sheet, Monocacy Hill sheet, and Pine Ridge sill are all about 300 m thick. If the Pine Forge dike is tilted northwest by the fold plunge, it likely intersects the Rattlesnake Hill sheet below the present-day surface and the Monocacy Hill sheet above the surface, resulting in a sill-dike-sill geometry stepping up through the Triassic sediments. Connections to the Pine Ridge sill or Jacksonwald basalt are not exposed.

In the Morgantown intrusive complex, orientations were measured at contacts exposed in the Dyer quarry (DQ) northwest of Birdsboro, and in a roadcut along Rt. 625 southwest of Knauers (Figure 2). The contact in the Silver Hill quarry (SHQ, Figure 2) was estimated from outcrop and cross-sections verified with drill-core data provided by H & K Group, Inc., (personal communication, 2012-13). The orientation of the St Peters sill and the best estimate for the tilting of the Morgantown intrusive complex comes from magmatic structures exposed in the Pennsylvania Granite Quarry (PAGQ; Srogi et al., 2010). The walls of this dimension-stone quarry are cut parallel and perpendicular to the strike of the St Peters sill and reveal magmatic layering consisting of cm-thick, plagioclase-rich, light-gray layers and darker layers richer in orthopyroxene phenocrysts in an overall mafic diabase (Srogi et al., 2010, Figure 1). Gimson et al. (2009) conducted a statistical analysis of the layers digitized on photographs. The average orientation of the magmatic layering is close to horizontal (0° ± 15°) on a wall cut along strike, and dips 18-22° on a wall cut perpendicular to strike, along a bearing of 345 or N15W (Srogi et al., 2010, Figure 7). Furthermore, the magmatic layering is transected by channels of mafic diabase that are roughly orthogonal and vertical on walls cut parallel to strike (Srogi et al., 2010, Figure 4A), but that also dip to the northwest on walls cut perpendicular to strike by about 67° ± 12° (n = 38). Our conclusion is that the St Peters sill was tilted about 20° NNW after solidification, presumably by continued subsidence and rotation along the border fault (Faill, 2003; Withjack et al., 2013). The outcrop pattern of the St Peters sill is consistent with this interpretation and shows no significant folding (Figure 2). Taking the slope of the land as well as the 20° tilt into account, we estimate the thickness of the St Peters sill to range from about 200 m at its thin, southwestern end to more than 650 m just east of the PAGQ. Magmatic layering also dips approximately 20° N-NE in the roadcut along old I-176 north of Morgantown (Smith, 1973; Srogi et al., 2010) and in the Rt. 625 roadcut (Figure 2).

For the Morgantown complex, minimum values for the depths of intrusions were estimated assuming the entire complex was tilted by 20° and measuring along bearings of 345 (N15W) to the youngest Triassic sediment of the Passaic Fm adjacent to the border fault (since there is no lava flow to mark the paleo-surface). Despite uncertainty due to assumptions, measurements, and the effects of intrabasin faulting, the resulting depths (Figure 3) are consistent with thickness estimates for Triassic sediments in the western basin by Faill (2003) and Withjack et al. (2013). We also back-rotate the intrusions at 625, SHQ, PAGQ, and DQ by 20° about a horizontal axis oriented 75 (N75E) to estimate their orientations at the time of emplacement (Figure 3). The St Peters sill at PAGQ is approximately horizontal and the Birdsboro dike at DQ is essentially vertical, while diabase at both the 625 outcrop and the SHQ still dip northeast (Figure 3). Our results cannot rule out some gentle synclinal folding along the western side of the Morgantown complex which could have helped to expose the igneous structures.

We combine all existing field-based data into a composite, schematic cross-section of the Morgantown and Jacksonwald intrusive complexes looking northwest and with as little vertical exaggeration as possible (Figure 4). Overall, the network of sills and inclined sheets on the western side

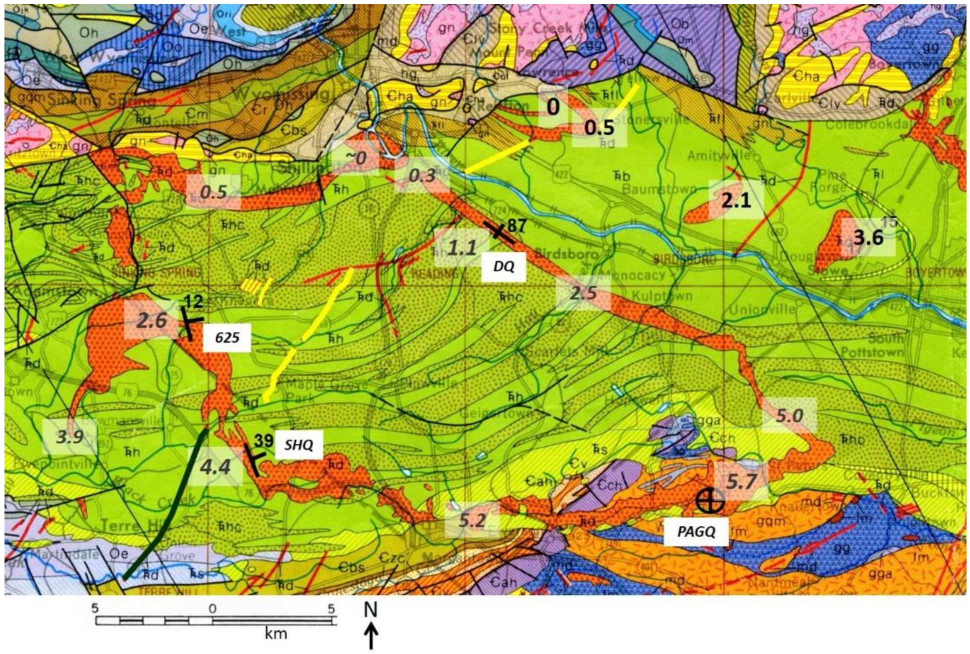

Figure 3. Geologic map as in Figure 2 showing estimates of paleo-depths and orientations after back-rotation about axis of tilting, as discussed in the text.

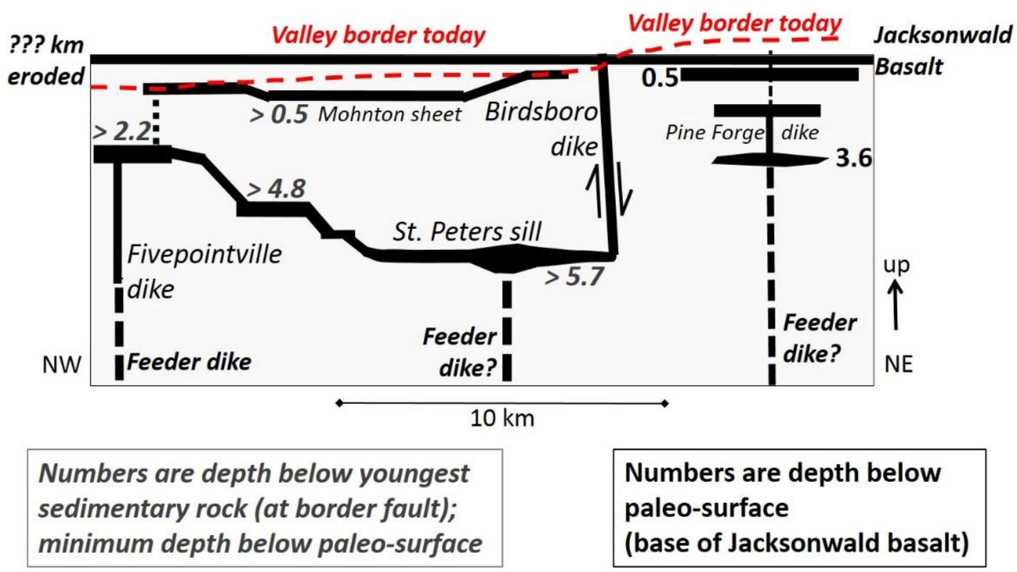

Figure 4. Schematic cross-section of the Morgantown and Jacksonwald intrusive complexes, viewed looking to the north-northwest and with minimal vertical exaggeration. See text for discussion.

cuts vertically up section by roughly 3.5 km and no individual sheet cuts up section more than 1.5-2 km (Figures 3 and 4), consistent with studies of other sill complexes world-wide (e.g., Magee et al., 2016; Muirhead et al., 2012; Cartwright and Hansen, 2006). The SPS and other sill-like intrusions are roughly parallel to bedding in the Triassic sediments and present-day topographic contours. The outcrop pattern of the western end of the SPS suggests short segments forming a ramp and upper sill like a saucer-shaped intrusion (e.g., Polteau et al., 2007). Although the Mohnton sheet looks like a sill in this cross-section, it must be an inclined sheet with moderate to near-vertical dips based on estimated dips of contacts in this area (MacLachlan et al., 1975; Smith, 1973) and the positive gravity anomaly low (Figure 5a) that indicates no diabase in the sub-surface (Daniels, 1985). The Fivepointville dike is narrow and strongly discordant with a small piece offset by faulting to the south (Figure 2). The Silver Hill-Bowmansville intrusion seems to be a discordant sheet that crosscuts bedding and topographic contours and varies from moderately- to steeply-dipping; the central portion resembles peripheral apophyses that extend from sills and sheets (Muirhead et al., 2012), but this may be an artifact of exposure. Triassic sedimentary rock south of the 625 locality (Figures 2 and 3) must be an inlier within diabase because elevation decreases continuously from south to north; this strongly resembles a bridge structure where a block of country rock becomes sandwiched between two sheets stepping up from a deeper to shallower level (e.g., Magee et al., 2016). Intrusions between the Adamstown pluton and Mohnton sheet in the northwest (Figure 2) are so disrupted by faulting and small folds (MacLachlan et al., 1975) they are not shown on the cross section (Figure 4). The Jacksonwald basalt extends westward above the Morgantown intrusive complex on Figure 4, consistent with the Withjack et al. (2013) model for a continuous basin at 201.5 Ma, the age of the lava flow. We show the Birdsboro dike feeding the lava flow, which is reasonable given how close the dike is to the paleo-surface (Figure 3), but the connection is not exposed and remains speculative.

The mineralogy of the diabase is consistent with the northwesterly tilt and exposure of deeper intrusion levels to the southeast, as noted by previous workers (e.g., Froelich and Gottfried, 1999; Woodruff et al., 1995). Shallower intrusions contain more coarse-grained diabase with intergrowths of quartz with K-feldspar and albitic plagioclase (granophyre); more iron- and titanium-rich diabase with magnetite; more abundant hydrous minerals such as biotite and amphiboles; and have greater hydrothermal mineralization (Smith, 1973; MacLachlan et al., 1975). Moreover, positive aeromagnetic anomalies (Figure 5b) with amplitudes of 500-1000 gammas, indicating more Fe-rich diabase containing more magnetite (Socolow, 1974), are centered over shallower intrusions, such as the Silver Hill-Bowmansville sheet and Pine Ridge sill, while positive anomalies of lower amplitude (100-200 gammas) are centered over deeper intrusions that have less evolved diabase, such as the Rattlesnake Hill and Monocacy Hill sheets. (In the St Peters and Morgantown areas very large, positive, concentric anomalies overlie magnetite skarn deposits in calcareous country rocks). Orthopyroxene (opx) phenocrysts are abundant (~15% to ~40% by volume) in deeper intrusions and absent at higher structural levels (Figures 6 and 7). The trends in diabase mineralogy have been attributed to lateral and up-dip migration of evolved, hydrous melt and hydrothermal fluids within single large sheets (Smith, 1973; Smith et al., 1975; Mangan et al., 1993; Woodruff et al., 1995; Froelich and Gottfried, 1999). Husch (1992) drew a schematic cross-section of the western parts of the Palisades sill showing this up-dip progression. In future work, we will combine the mineralogic data with paleo-depth estimates to quantify melt and fluid transport within the sill-dike complex and make more quantitative comparisons with other intrusive systems.

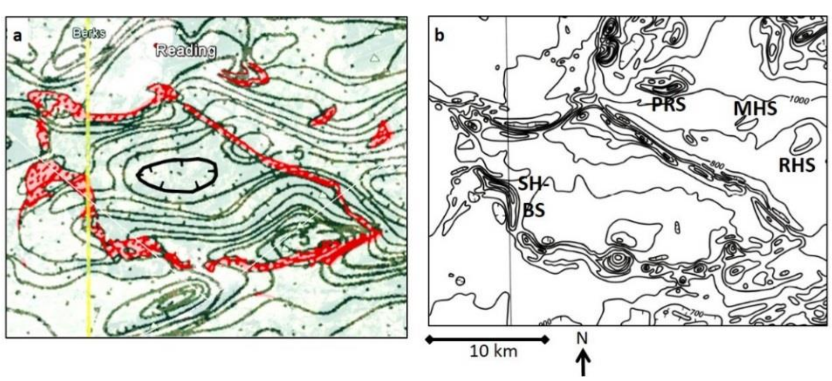

Figure 5. Geophysical maps of Morgantown-Jacksonwald area. Vertical lines are 76°W longitude. a) Residual Bouguer gravity anomaly map after Sumner (1977) with diabase in red and gravity anomaly low highlighted. b) Aeromagnetic anomaly map (1978). **SH-BS**: Silver Hill-Bowmansville sheet. **PRS**: Pine Ridge sill. **MHS**: Monocacy Hill sheet. **RHS**: Rattlesnake Hill sheet.

DISCUSSION

Feeder structures and magma replenishments

The dikes or other structures bringing basaltic magma into the Newark basin are generally not well-exposed and their locations must be inferred from field, mineralogic, and geochemical evidence. However, the Fivepointville dike on the west side of the Morgantown intrusive complex is connected to and apparently fed into the Adamstown pluton (Figures 2, 3, 4). Coarse-grained diabase from the tip of the dike (Figure 6b), its faulted southern extension (Figure 6c), and the base of the Adamstown pluton (Figure 6a) are virtually identical and contain abundant large, euhedral to subhedral orthopyroxene (opx) and small olivine (ol) phenocrysts in a matrix of clinopyroxene, plagioclase, and minor granophyric intergrowths. Fine-grained diabase from the dike is similar to other chill margins as discussed below.

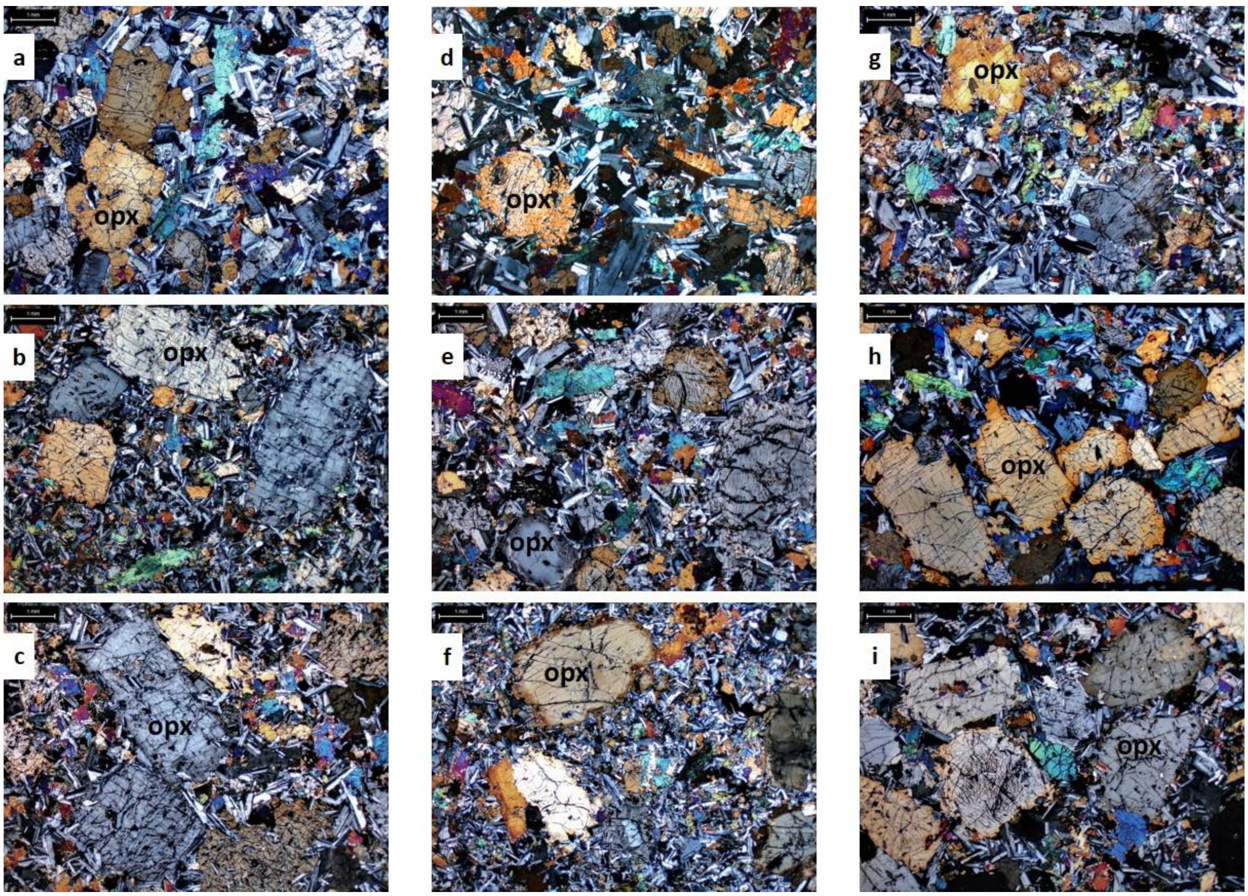

Figure 6. Photomicrographs of diabase samples containing large orthopyroxene phenocrysts, organized from west (left) to east (right). Scale bar in upper left of each image is 1 mm. a) THQ-306B – base of Adamstown pluton, no olivine (ol). b) THQ-305 – Fivepointville dike, ol present. c) THQ-298 – Fivepointville dike, faulted southern segment, ol present. d) EQ-97E-6a – PAGQ, no ol. e) MoQ-212 – New Morgan sheet near upper contact, no ol. f) MoQ-374 – New Morgan sheet near lower contact, no ol. g) PQ-341 – Birdsboro dike, Coventry Woods Park, no ol. h) BoQ-294b – Rattlesnake Hill sheet, cumulate closer to upper contact, ol present. i) BoQ-3 – Rattlesnake Hill sheet, cumulate near lower contact, ol present.

The evidence from the Fivepointville dike suggests that concentrations of large opx phenocrysts near the base of the intrusions can be used to identify approximate locations of feeder structures, consistent with previous studies of the Palisades and similar intrusive systems (e.g., Husch, 1992; Woodruff et al., 1995; Marsh, 2004). We propose an additional feeder structure in the Morgantown intrusive complex located beneath the thickest part of the St Peters sill (Figure 2). The cut walls of the dimension-stone quarry at PAGQ reveal spectacular layering of opx and plagioclase crystals forming lens- to wok-shaped packets averaging about 0.3 m thick (Srogi et al., 2010), that we interpret as shear-induced flow sorting within a crystal slurry (e.g., Petford, 2009) near its source feeder: the "fine structure" within the opx cumulate that "records the emplacement dynamics," (Marsh, 2004; Bedard et al., 2007). Magma from this feeder (Figure 6d) spread laterally through the St Peters sill, to the east and up-dip into the lower part of the Birdsboro dike (Figure 6g), and to the west and up-dip (Figures 6e and 6f), forming rocks with opx phenocrysts and no ol (Figure 7). Alternatively, a dike-like feeder could extend along the length of the St Peters sill. Any feeder structures tilted to the north-northwest that were truncated at the base of the intrusion would dip southeast under the basement rocks and therefore not be exposed today. The Rattlesnake Hill and Monocacy Hill sheets contain small ol phenocrysts not observed in the St Peters sill (Figures 6h, 6i, and 7), so we propose a third feeder into the Jacksonwald intrusive complex. The Pine Forge dike is probably tilted to the northwest and connects the Rattlesnake Hill and Monocacy Hill sheets; if it extends beneath the Rattlesnake Hill sheet to the southeast then it may have functioned as a feeder into the system, as well.

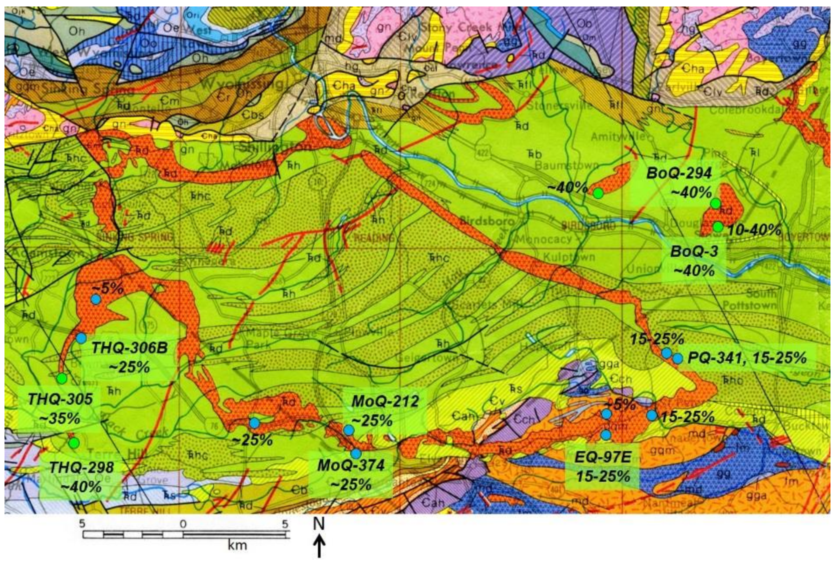

Figure 7. Geologic map (same as Figure 2) showing locations of, and approximate volume percentages of opx phenocrysts in, samples containing only opx and aug (blue-filled circles) and opx, aug, and ol phenocrysts (green filled circles). Sample numbers refer to photomicrographs shown in Figure 6.

The 250-meter-wide Birdsboro dike, possibly the widest dike in the Eastern North American CAMP, seems an obvious choice for a feeder structure, but several factors suggest this is not the case. Unlike the extensive dike swarms in this part of the ENA, it does not continue southeast beyond the St Peters sill into the basement rocks and its NW-SE strike is roughly orthogonal to the orientation expected for extension-related magma propagation (Schlische et al., 2003, and references therein). If the Birdsboro is an accommodation dike (Schlische, 1992) it narrows downwards and is unlikely to connect to a deep magma source. The Birdsboro dike connects with the St Peters sill and contains similar phenocrysts (Figures 6d, 6g) suggesting that it was fed by and propagated from the St Peters sill, similar to sill-dike relationships in the well-exposed Franklin sill complex, Canada (Bedard et al., 2012, Figure 10). The Birdsboro dike lies within and is parallel to the Schuylkill Tectonic Zone (Wise, 2014) at the intersection of major Paleozoic allochthons that may have functioned as a crustal weakness that guided the propagation of the Birdsboro dike from the sill. The Birdsboro dike was not a feeder from depth to the upper-crustal intrusions although it may have transferred magma from the intrusions to the lava flow.

There are numerous dikes in the Newark basin and in basement rock southeast of the basin, but most can be ruled out as feeders for the intrusions based on their geochemical composition (Smith, 1973; Smith et al., 1975, Figure 1) or their distance and geometry. The diabase intrusions and basalt comprising the Morgantown and Jacksonwald complexes are all high-Ti, quartz-normative (HTQ) tholeiites as defined by Weigand and Ragland (1970), also called the York Haven-type diabase by Smith (1973) and Smith et al. (1975). Laterally-extensive dikes about 25 km southeast of the intrusions are HTQ tholeiites, but their orientations are not known and they may not have intersected the now-eroded intrusions. The low-Ti, quartz-normative (LTQ) tholeiite dikes (Weigand and Ragland, 1970; also called Rossville-type diabase by Smith), such as the Maple Grove dikes in the Morgantown and Stonersville dike in the Jacksonwald (Figure 2) and the long dike that extends from Doe Run to East Greenville, PA, cannot be feeders to the HTQ intrusions. Similarly, the Quarryville-type olivine-normative dikes just southwest of the Morgantown intrusive complex (Figure 2) cannot be feeders. The Green Hills dikes are HTQ tholeiite and could connect to the Birdsboro dike and/or the Mohnton sheet to the north, but they cannot connect the Mohnton sheet and St Peters sill because the gravity anomaly low precludes any significant diabase in the sub-surface in this area (Figure 5a; Daniels, 1985).

There is field and petrographic evidence for emplacement of multiple magmas within parts of the Jacksonwald and Morgantown intrusive complexes. Different phenocryst populations in cumulate rocks of the Morgantown and Jacksonwald complexes (Figure 7) and in chill margins, suggest the presence of slightly different parental magma batches. In the Rattlesnake Hill sheet all chill margin samples within 50 cm of the lower contact of contain variable proportions of similar-sized aug, ol (Fo_{80}), and opx phenocrysts, consistent with small pulses of very similar magma carrying slightly different phenocryst populations, possibly mixed shortly before emplacement (e.g., Marzoli et al., 2014). Additional variations in grain size and phenocryst abundance over the basal 10 m of the sheet may indicate meter-scale replenishments. Walls of the Pennsylvania Granite Quarry in the St Peters sill show 0.3-1 m thick, sill-like lobes of magma, some with scalloped lower margins (see Srogi et al., 2010, Figure 1C), that appear to be small-scale replenishments intruded into crystal mush (e.g., Wiebe, 1993). Extremely narrow (~1-3 mm), internal chills are observed in thin sections of the chill margin of the Birdsboro dike, and in outcrop and in thin sections from the 625 locality. Mingling textures between slightly more felsic and more mafic magmas are observed in outcrop along the southwest side of the old I-176 highway north of Morgantown. We have not yet sampled to identify larger-scale magmatic units within the intrusions or Jacksonwald basalt comparable to those recognized in the Palisades sill and Orange Mtn basalt (Puffer et al., 2009).

Comparison of lava flow and chill margin compositions

A comparison of the intrusion chill margins with the Jacksonwald basalt is important to judge the connectivity of the plumbing system. Puffer et al. (2009) demonstrated that there are portions of the Palisades sill that are geochemically equivalent to the Orange Mtn and Preakness basalts and discussed a model for repeated intrusion, mixing, and eruption in this long-lived system. In the Morgantown and Jacksonwald complexes all rocks are the oldest, HTQ tholeiite type (e.g., Smith et al., 1975; Froelich and Gottfried, 1999) corresponding to the Orange Mtn basalt (Puffer et al., 2009). Figure 8 compares published (Merle et al., 2013) and new (Table 1) normalized rare-earth element concentrations in the Orange Mtn and Jacksonwald basalts and chill margins from the Rattlesnake Hill sheet and Birdsboro dike. REE elements were chosen for comparison because they are little affected by alteration or assimilation of country rock. Our samples of two Jacksonwald basalts and four chill margins are essentially identical, and very similar to a third Jacksonwald basalt and published data for the Jacksonwald and Orange Mtn basalts (Merle et al., 2013). The range in concentration results from a combination of inter-laboratory comparison and analytical error and variations in phenocryst abundance through the flow. The important point is that the same magma

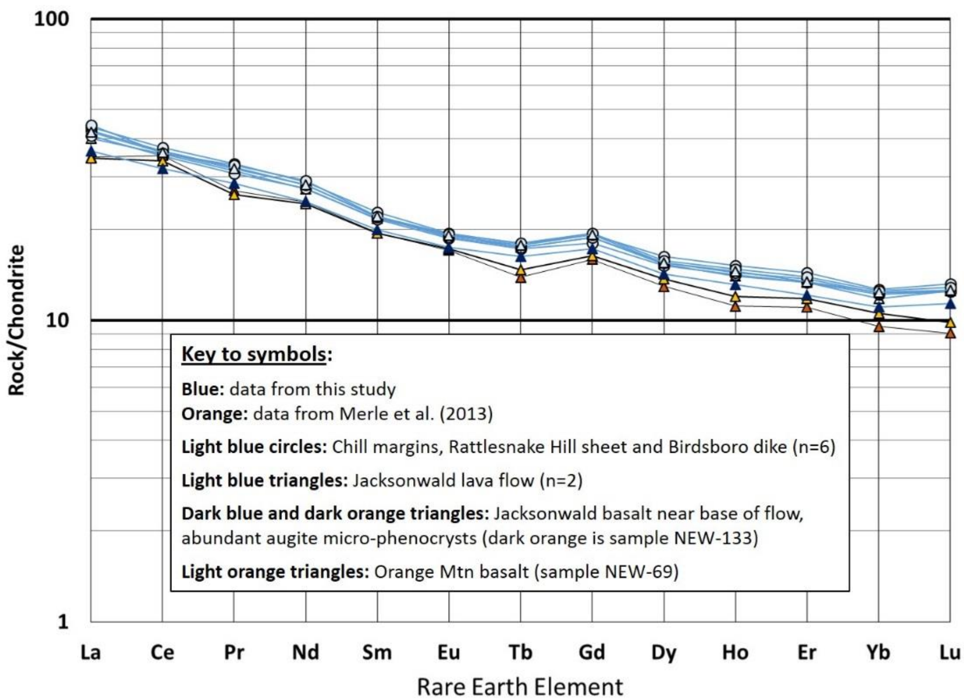

Figure 8. Chondrite-normalized rare-earth element patterns of Jacksonwald basalt, Orange Mtn basalt, and chill margins, from Morgantown and Jacksonwald intrusive complexes. See Table 1 for data from this study. Normalization values from Sun and McDonough (1989).

was chilled at the base of the Rattlesnake pluton (~3.6 km depth) in the Jacksonwald complex, and at the edges of the Birdsboro dike (>1.1 km depth) in the Morgantown complex, and erupted at the surface, suggesting there was connectivity at least in these parts of the plumbing system. The lower chill margin of the Palisades sill shows comparable similarity to the Orange Mtn basalt (Puffer et al., 2009). Thirteen additional chill margin samples are being analyzed for REE elements to establish the similarity of chill margins across the intrusions and better evaluate connectivity in the system.

Additional evidence for the formation of the basalt and chill margins from a common magma comes from phenocryst compositions. Opx occurs as inclusions within aug phenocrysts in almost all lava flows and chill margins (Figure 9, inset images), along with (usually relict and altered) ol phenocrysts. Some chill margins also contain opx phenocrysts (Figure 9, inset images). The similarities in the back-scattered electron images are striking. Figure 9 also summarizes the pyroxene compositions in terms of Mg/(Mg+Fe) ratio, as analyzed by SEM-EDS at West Chester University, for three basalt and ten diabase chill margin samples (representative analyses are in Table 2). The Mg/(Mg+Fe) ratio has the same narrow range of values between 0.80 – 0.85 for the interiors of both opx and aug phenocrysts; rims and matrix grains were not included in Figure 9. Differences in the ratio between samples are so small that they are likely due to analytical error and sampling phenocryst margins rather than cores (cut effect). The similarity in phenocryst appearance and composition between three lava flow samples and all chill margins analyzed to date is strong evidence for the same magma being present at an early stage throughout the plumbing system, from the Jacksonwald syncline in the east to the Fivepointville dike and upper contact at the 625 locality in the west. This supports the conclusion of previous workers, based on whole-rock geochemistry and isotopic compositions, that a large volume of magma with little compositional heterogeneity established the upper-crustal magmatic system at 201.5 Ma at least in the Newark basin, probably in the ENA, and perhaps even throughout the CAMP (Marzoli et al., 1999; McHone, 1996, 2000).

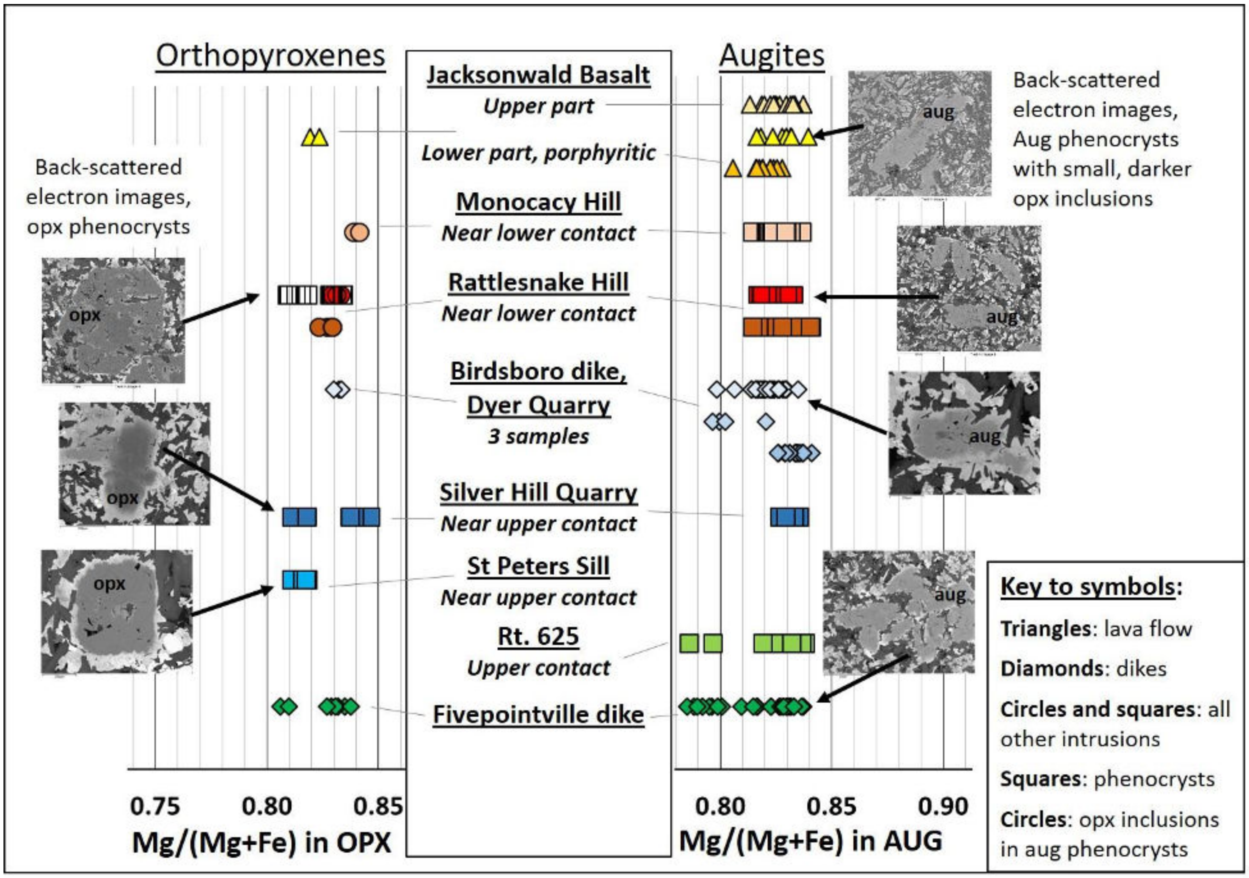

Figure 9. Pyroxene compositions in terms of molar Mg/(Mg+Fe). Similar composition of opx and aug and similar appearance in back-scattered electron images should be noted. See text for discussion.

Factors controlling intrusive geometry

The geometry of the western intrusions is different from the folded sheets and ring structures and the laterally-extensive Palisades sill to the east (Figure 1). While this may be due partly to deeper erosional level in the Morgantown intrusive complex, it is likely that small-scale heterogeneities in the alluvial fan sediments in the western Newark basin helped deflect the intruding magma into smaller segments that stepped up through the stratigraphy (e.g., Magee et al., 2016). The St Peters sill intrudes basement rocks in the southeast corner of the Morgantown complex and there is a fold in the overlying Triassic sediments (Figure 2) also visible on Lidar images. Daniels (1985) proposed these rocks were uplifted by intrusion of the magma, a type of forced folding produced by roof uplift (Magee et al., 2016; Muirhead et al., 2012). If correct, this may have been influenced by sill inflation above the proposed feeder structure and flow from the St Peters sill into the Birdsboro dike. A small intrabasin fault north of the St Peters sill and west of the basement rocks could have formed by the inflation and deflation of the St Peters sill during magma transport and crystallization. The gentle synclinal fold observed in the Triassic sediments but not in the St Peters sill (Figure 2) might also result from forced folding and roof uplift accompanying sill inflation.

Dike swarms typically reflect far-field tectonic stresses (e.g., Schlische et al., 2003), whereas shallowly-dipping sheets are the product of an over-pressured magmatic environment and/or deformation of country rock ahead of an advancing sill (Muirhead et al., 2012). The variety of intrusion shapes and orientations in the Morgantown intrusive complex, and especially the NW-SE orientation of the Birdsboro dike orthogonal to rift-related tectonic stresses, indicates that the orientation of the stress regime was not the major factor controlling magma intrusion in the upper crust. While it is also possible that parts of the plumbing system with different orientations formed at different times, the nearly identical mineralogy and chemistry of the basalt and chill margins suggests that the entire plumbing system was connected and formed together over a relatively short time. Thus, intrusion architecture was likely controlled by an abundant magma supply, overpressure and local heterogeneities in the rocks within and surrounding the Newark basin, similar to conclusions based on other sill-dike complexes (e.g., Magee et al., 2016).

Distribution of rock types and implications for transport processes in the plumbing system

Figure 10 summarizes the general distribution of diabase mineralogy in the Morgantown and Jacksonwald intrusive complexes. The concentration of phenocrysts near the floors and in the deepest intrusions, and the occurrence of evolved and hydrothermally-altered diabase in the shallowest intrusions, is not surprising. Previous workers note that studied intrusions in the Newark and Gettysburg basin are not in mass balance (e.g., Smith, 1973; Smith et al., 1975; Froelich and Gottfried, 1999); i.e., the average composition of an intrusion does not match the chill margin "parental magma," containing either too much pyroxene cumulate or too much evolved, Fe-rich diabase with granophyre. However, most diagrams like Figure 10 and models involving differentiation and migration of evolved melt and fluids are developed for *a single, saucer-shaped sheet* such as the York Haven sheet (Mangan et al., 1993; Woodruff et al., 1995), Palisades sill (Husch, 1992; Steiner et al., 1992; Puffer et al., 2009; Block et al., 2015), and Golden Valley sill, Karoo basin (Galerne et al., 2010). By contrast, we find similar rock distribution patterns over an interconnected sub-volcanic system extending ***vertically*** across almost six kilometers in the upper crust. This easily explains the lack of mass balance within an individual intrusion: the intrusions were connected and operated as a system open to magma coming in from below or flowing laterally and to magma and hydrothermal fluids leaving for other parts of the network. An outstanding question in volcanology today is whether gas content or the rate of gas release controls volcanic eruptions and eruption intensity (Putirka and Cooper, 2017). Subvolcanic, interconnected sill-dike complexes influence the evolution and transport of melt and volatiles and so are a critical part of understanding the eruptive behavior of volcanic systems.

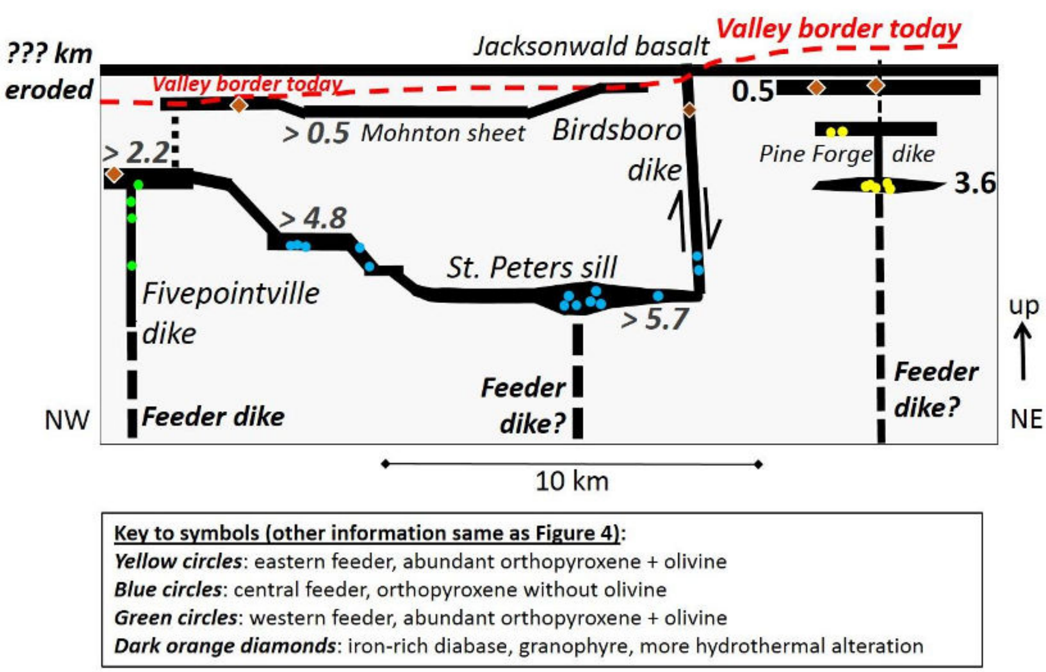

Figure 10. Schematic cross-section showing distribution of opx and ol phenocrysts and evolved, Fe-rich diabase. See text for discussion.

CONCLUSIONS AND FUTURE WORK

The western Newark basin differs from the central and eastern basin primarily in having a deeper level of erosion and greater heterogeneity in sedimentary lithology. The result is an apparently narrower basin exposing interconnected sill-sheet-dike networks in an upper-crustal cross-section of more than five kilometers beneath the Jacksonwald basalt (Orange Mtn equivalent). There are at least three feeders into the intrusive complexes, with numerous features that indicate small-scale magma pulses or replenishments. To better define intrusive structures and connectivity, we are starting a pilot project using magnetic susceptibility to identify magmatic flow fabrics. A drone-based photogrammetry project is underway to image the complex magmatic layering in the PAGQ and understand better the flow sorting in crystal slurries. We also continue work on volcanic structures in the Jacksonwald basalt (Leflar et al., 2015).

Whole-rock and phenocryst compositions support previous conclusions that a large volume of relatively homogeneous magma built the entire upper-crustal and volcanic system in the Newark basin. We continue quantitative testing of this idea with new whole-rock and mineral analyses of the lava flow and chill margins sampled at different emplacement levels. Pyroxene phenocrysts brought up from depth hold information about crystallization and transport in the deeper plumbing system that can be investigated using thermodynamic modeling and trace element profiles. Data presented in this paper are consistent with models of differentiation and migration of evolved melt and volatiles developed for single diabase sheets; we show that these processes also operate in interconnected intrusive networks that span kilometers of the upper crust. This is our most important finding because it directly relates to the operation of active sub-volcanic systems. We continue mineral analyses and thermodynamic modeling to trace the evolution of the magma and volatiles within the entire system.

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Northern and Southern Appalachian Wilson Cycle Junction in Pennsylvania

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ABSTRACT.

Overlapping tectonic patterns of multiple orogenies at the Susquehanna Piedmont's Northern and Southern Appalachian junction are the Devil's Workshop, designed to induce insanity in any geologist stupid enough to attempt an explanation. This challenge is presented as a four-act mystery play that may make it an open and shut case for a Wilson Cycle.

BACKGROUND.

The Appalachian orogen can be divided into tectonic zones that extend almost continuously from Newfoundland to Alabama (tectonic maps by Williams (1958), Hibbard (2006), and Hatcher (2007). The western zones record opening and mostly later Paleozoic closing of the paleo-Atlantic Ocean. The closing collision detached multiple megathrust sheets of basement, now exposed almost continuously along the backbones of both the Northern Appalachians (N-Appl) and Southern Appalachians (S-Appl). Continued closure overlapped the ends of two megathrust sheets, the Blue Ridge and Reading Prong Megathrusts, and joined them as the Susquehanna Piedmont Superthrust. That combined sheet drove into the foreland stratigraphy for ~ 150 km(+) to form the Ridge and Valley fold belt (Figure 1). At the presently exposed erosion level, the Reading Prong sheet's sedimentary cover rests on basement of the Blue Ridge sheet. The result was a ~100 km-wide gap that separates basement exposures of the U.S. part of the chain into two halves separated by the Susquehanna Piedmont Superthrust with the Susquehanna River flowing along its azis.

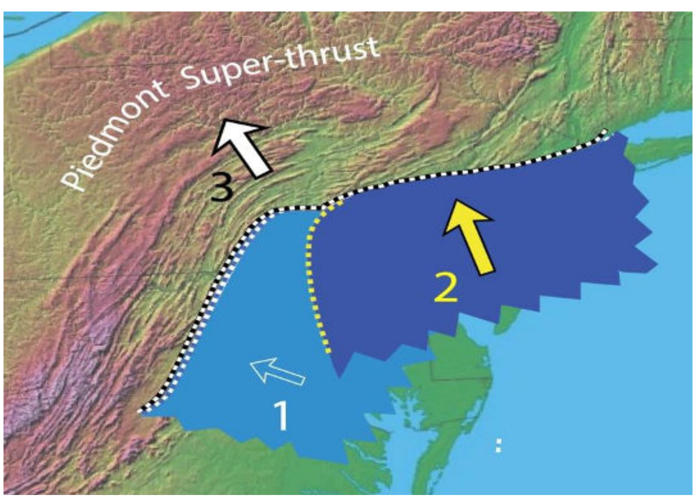


Fig. 1. Junction of the Northern and Southern Appalachians. *Continental colliion in the later Paleozoic Alleghaninan Orogeny drove the Reading Prong Megathrust's end (sheet 2) through and over the end of the slightly older Blue Ridge Megathrust (sheet 1) to form the Susquehanna Piedmont Superthrust (sheet 3). Continuing NNW oriented collision followed by WNW collision drove that sheet into the foreland to form the Ridge and Valley Province.*

The Susquehanna Piedmont includes that gap in the more accessible N-Appl sheet (Figure 2). That sheet's history of multiple orogenies of different styles, trends, and phases makes it tectonic museum that should be viewed through the lens of tectonic heredity's reactivation of older structures. Generations of geologists mostly in the 20th century produced Figure 2's record of that history. Two versions of the state geologic map, classics in both detail and color selection, compiled and organized much of that work. Excellent snapshots of data and tectonic thinking at the end of the 20th century are in compilations by many authors in *Geology of Pennsylvania* (Schultz, ed. 1999) and in Faill's (1997a, 1997b, 1998) summary papers of separate orogenic events. Despite that rich background, the overall tectonic syntheses of of this region are among the least advanced of any part of the chain.

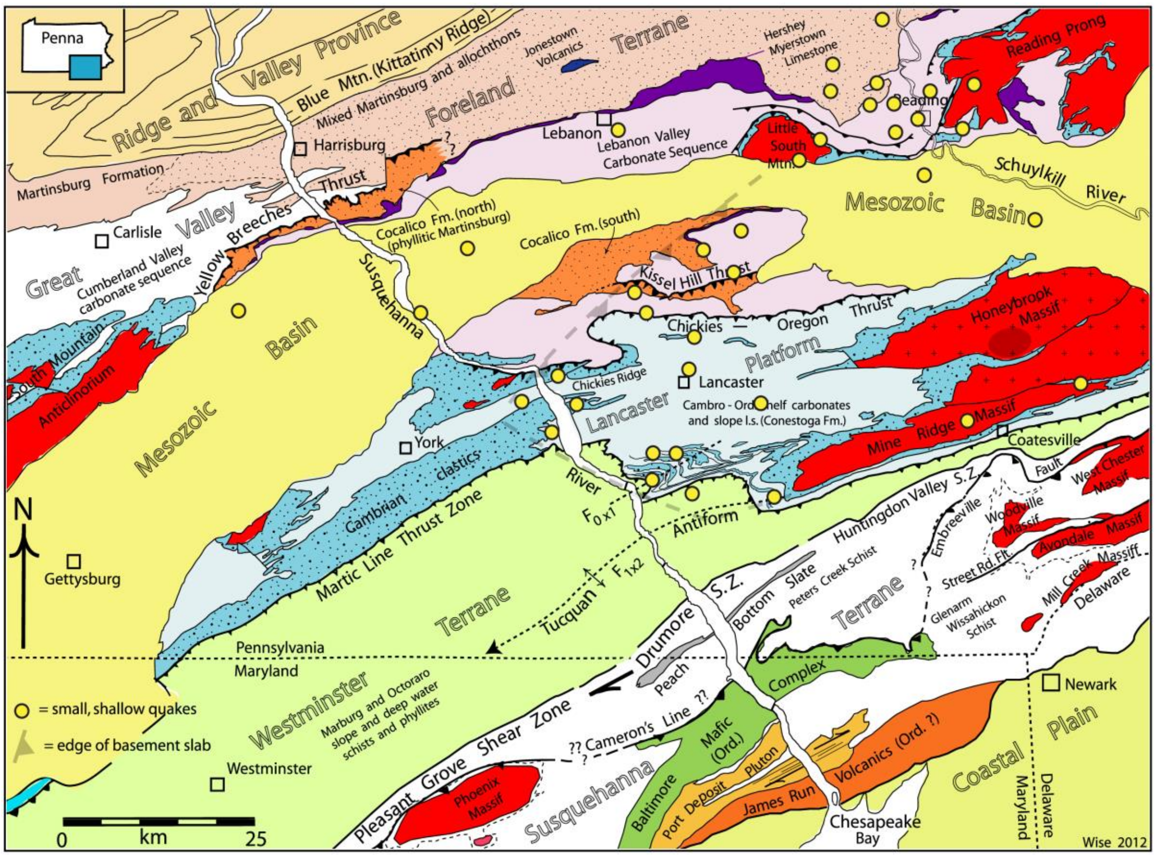
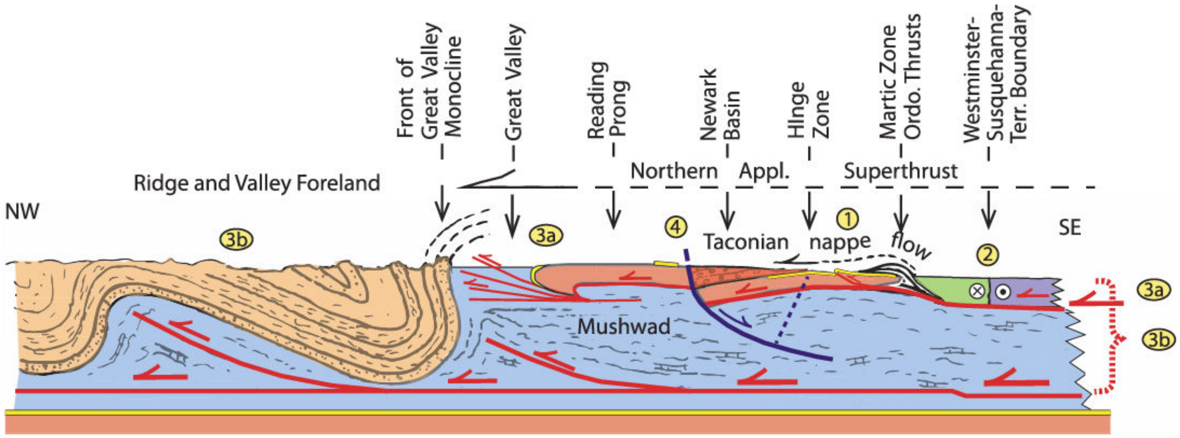


Fig 2. Tectonic map of the Susquehanna Piedmont. The Northern Appalachin Megathrust comprises most of this area beneath the Mesozoic cover. The Precambrain basement massifs at the east and their buried extensions outlined by seismic activity (yellow dots) are the core slab of the megathrust. Its Cambo-Ordovician cover it rests on the Precambrian core of the Southern Appalachian Megathrust as the Yellow Breeches Thrust. The Martic Line separates North American terranes from offshore additions.

Most of what follows is lifted directly from evidence long-available on the state map but not interpretable without more recent age control, newer detailed studies, and newer tectonic ideas. In this century a new cycle of detailed mapping combined with more abundant ~m.y. precision dates began to unscramble the tectonic record and allow more regional views of individual orogenies and processes. This paper organizes some old and new data and ideas and adds a few missing pieces to the puzzle as the basis of an orogeny-by-orogeny tectonic history. Some parts are speculative, some are outdated, a few are outrageous, and all are tentative. Some processes or styles are unique to this transform-protected corner. Others may be applicable as models for other segments of the chain or beyond. It is presented here in preliminar form in the hope that with testing and correction it may be a framework for improved understanding of this critical area.



<u>Figure 3. Schematic cross section of Figure 2.</u> The numbers indicate the sequential sequence of features created by the several orogenies. The text describes its evolution as a Wilson Cycle four act play. Act 1 created the Martic edge of the craton. Act 2 (number 1 above) pushed the adjacent sea floor over that edge in the Ordovician (2) and welded the Susquehanna Terrane onto it in the Mid-Paleozoic (3). Act 3 in the late Paleozoic detached the basement slab and joined it with the S. Appl. Megathrust to form the Piedmont superthrust and drive it into foreland stratigraphy (3a) while raising it on a giant mushwad and continuing to push it forward (3b). Act 4 collapsed the mushwad to form the Mesozoic basin (4) when when seafloor opening withdrew lateral support.

The tectonic map of Figure 2 covers the critical junction area. It exposes the Piedmont from Coastal Plain to Ridge and Valley Province. The Martic line separates the inner and the outer Piedmont. At the type Martic locality just east of the Susquehanna River this line preserves the edge of the Cambrian carbonate platform, its transition into offshore slope deposits, and deep water clastics of the Westminster Terrane. This edge of the North American continent originated in latest Precambrian time by breakup of the Supercontinent of Rodinia and those structures have influenced its tectonics ever since. This rarely preserved boundary separates the original continent from all that has been added or modified. The tectonic maps of the chain listed above trace the Martic boundary for the length of the chain, albeit with different names. The Inner Piedmont SE of the line has two components, the Westminster Terrane of offshore deposits, now mostly schists, and the Susquehanna Terrane, an allochthonous terrane typical of southern New York and Connecticut, that docked with this area in the Mid Paleozoic.

Most of the near-surface rocks between the Martic Line and the edge of the Ridge and Valley Province on are parts of a thin skinned Taconian nappe that flowed across the region in later Ordovician times. This along with the initial offset of the content must be described and its effects mentally added to all that followed, a story attempted in the format described below.

*****

Prelude to a four-act play.

A Wilson Cycle is the underlying theme for Figure 2's complex series of events overprinted orogenies. The schematic sequence of events illustrated in Figure 3 can be told as a four-act play wherein the same actor can play different roles in successive acts and scenes. The following prelude to that play introduces the principal actors, sets the superthrust stage upon which they act, and suggests the origin mechanism for the basement slab that became their stage. The four acts, several with multiple scenes, are:

Act 1: *Cambrian edge of eastern North America formed by breakup of Supercontinent Rodinia.*

Act 2: *Continental margin greatly changed by interactions with oceanic tectonics.*

Act 3: *Closing the ocean basin to form the Appalachian Chain within the Gondwana supercontinent.*

Act 4: *New cycle begun by collapse of the Appalachian edge to form the modern Atlantic* Ocean..

Principal actors. Much of Figure 2 resulted from later Paleozoic or Alleghanian-age activity on the greater basement sheet of the N-Appl shown in the Lidar image of Figure 4. The buried west end of the N-Appl basement block does not show clearly on the Lidar image but the western edge of the Lancaster-Reading Seismic Zone (yellow dots on Figure 2) defines it's approximate loctation (Wise and Faill, 2002).

On that image, the Reading Prong forms the north boundary of the N-Appl greater basement slab. Two forks separated by the Oley Valley with Little South Mountain as a fault-displaced end-fragment of the western prong. Ridges of steeply dipping Hardyston Quartzite resting unconformably on basement define west faces of each bloc that make them stand out as orientational anomalies in the overall structural grain. They are interpreted below as abandoned klippes that lay beyond the terminating tear fault of the S-Appl Megathrust but were caught up by the N-Appl Megathrust.

The Lidar image also shows a remarkably linear, ~120 km-long N80E-trending topographic and geologic feature that bounds the basement slab's trailing edge. It includes Valley Forge, a location George Washington probably chose as his headquarters partly for the forge area's capabilities and infrastructure but also for its location at the intersection of two major transport arteries, the lineament and the Schuylkill River. Segments of the lineament have different local names but none describe it as a single feature. One or another of these difficult-to-find, difficult-to-remember location names could be elevated as a name for the lineament's full length but the nationally recognized and easily remembered name of Valley Forge Lineament seems more appropriate. The lineament is a late-stage Alleghanian-age rolldown of the rear edge of the 60 km-wide basement sheet that forms the core of the Susquehanna Megatrhust. At its north edge, the Great Valley Cambro-Ordovician carbonates and Martinsburg Flysch separate the fold belt's first or Silurian ridge from the Reading Prong basement slab.

The NE corner of the Lidar imafe shows the boundary between the older,N75E-trending Susquehanna basement slab and the younger N45E-trending NJ Highland/ Shawangunk basement domain. The younger sheet's end-loading and shear of the Reading Prong, N edge of the sheet produced the 15 km-wide, sinistral Hawk Mountain kink that transects the entire valley from basement front to Hawk Mountain folds. A second kink at the end of the Reading Prong produced the Schuylkill Tectonic Zone. This controls much of the locl Triassic-Jurassic tectonic activity and its axis is now host to a linear stretch of the Schuylkill River.

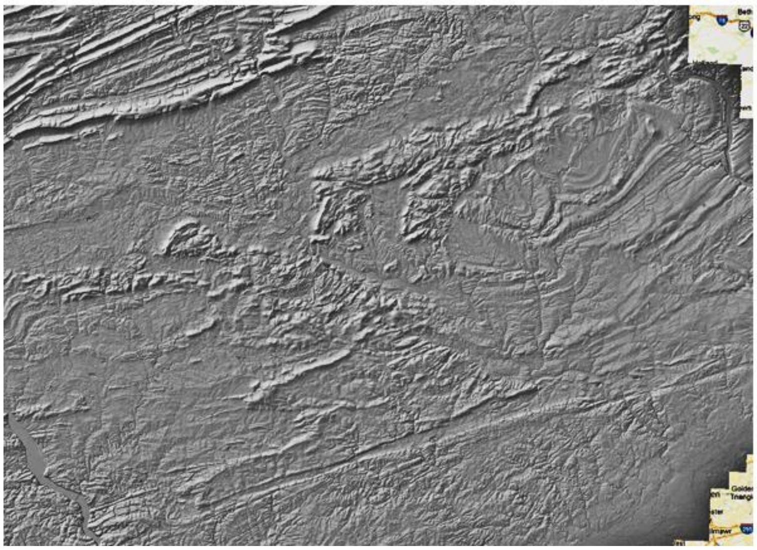


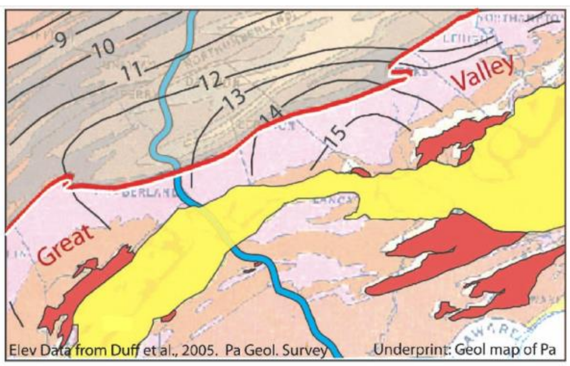
Figure 4. Lidar image of terminal structures of the greater N-Appl basement slab. Major actors in the Wilson Cycle text are identifiable by their topographic signatures.

Parallel to and N of the Reading Prong's front, a linear contact truncates the Hamburg Allochthon's fabric. That line, sub-parallel to the Valley Forge Lineament, is probably the true northern edge of the Susquehanna Megathust's greater basement core.

Two megathrusts make a Piedmont Superthrust. Act 3 involved final closing of the Paleo-Atlantic Ocean during the later Paleozoic Alleghanian Orogeny. Figure 5 shows the joining had to occur as a two-stage process. Cambrian clastics on edges, plunging noses, and locally tops of both basement sheets are at a common level of ~ 0.5 km above sea level. The N-Appl's basement extends ~ 60 km to the rear and behind that the basement in mantled gneiss domes is at about the same elevation. The same surface on the underlying autochthonus craton is ~15 km below sea level and slopes gently to the SE to pass without change beneath the edge of the Superthrust's Great Valley foreland. The broad red line on the figure is the vertical Silurian ridge, frontal boundary zone of the Piedmont Superthrust. Despite an Alleghanian orogenic cycle that detached two distnctly different slabs of that cratonic surface, juxtaposed them, and then lifted them by ~ 15 km, some remarkable process either preserved or restored that surface to it's original horizontality.

These geometric relationships drive several tectonic conclusions. **1)** Nearly identical elevation of the tops of both megathrust sheets shows their juxtaposition predated superthrust uplift**. 2)** Minimal differences in elevation among the tops of the overlapping sheets indicates the edge of at least the upper sheet is very thin. **3)** The geometry of first stage superthrust assembly at depth requires a second stage accumulation to raise the sheet to its present level. **4)** Uplift was by a 12 -15 km thick mushwad that accumulated beneath the superthrust's forward-moving leading edge. A "mushwad" is a Malleable, Unctuous SHale, Weak-layer Accretion in a Ductile duplex (Thomas, 2001) **5).** At this scale and burial depth, the mushwad was so weak that it's almost hydrostaticbehavior is the probable mechanism that maintained the superthrust's original horizontality during uplift.

Origin mechanism of basement slab. This thin, extremely large slab of N-Appl basement is not an unusual feature but is origin is enigmatic. This Piedmont example provides an unusually well-exposed, well-preserved set of constraints on the detachment mechanism. The initial detachment cut through the craton's Martic edge at a depth of several km. That plane propagated inland for 40+ km at essentially constant depth, through a complex Grenville basement that included an anorthosite pluton. It finally climbed through the slab to create its basement leading edge and continued in the Cambrian stratigraphy as the megathrust's thin leading edge. The process was not unique to the this area's Alleghanian orogeny; similar features appear in the S-Appl (Hatcher, 2002), basement sheets occur in the Taconian-age mantled gneiss domes along the chain,and other mountain chains such as the Eastern Alps with theTauern Fenster have similar large sheets of detached basement.



<u>Figure 5.</u> The tops of all the Piedmont Megathrust basement exposures are ~ 0.5 km above sea level while the same surface of the autochthonous craton is 15 km below them as it passes seaward. About 10 km of weak Cambro-Ordovician lithologies piled up as a giant mushwad to raise this leading edge of the superthrust. Sub-sea contours in km are from Alexander et al, 2005.

This basement sheet's geology precludes ordinary stratigraphic control. The fault tip had to begin at the cratonic margin and propagate through complex Grenville basement for at least 40 km at essentially constant depth below the Cambrian unconformity. The sheet is too thing to be driven from the rear; the driving mechanism had to be the downward drag of 8-10 km of overriding stratigraphy. Without lithologic control, the only remaining independent processes are pressure, temperature, and perhaps a seismic wave guide. Hatcher (personal communication) notes this detachment problem for the S-Appl. Wise (2000.) invokes it for the Beartooth and Bighorn Ranges of Wyoming-Montana. There the sedimentary load of only 3-4 km depressed the detachment planes cut at about the same depth as the Appalachian examples to form 6-8 km thick basement edges. As those detachments broke through the basement surface, their leading edges rolled into a frontal brow-fold with ~ 2 km radius After that the slab plowed forward and upward with the fold as a passive passenger. Some of the frontal folds of the Appalachian slabs may be of similar origin.

If the temperature was the control mechanism for quartz weakening to begin, that isotherm in a stable craton would be far deeper than a few km in basement. The more probable candidate is some form of pressure-dependent weakening at depths of 10 -12 km. The general seismic pattern of localization of most small quakes in the upper ~ 10 km is probably related mechanism.

*** ACT 1: The Continental edge ***

ACT 1, scene 1: Cambro-Ordovician carbonate platform and the Susquehanna Tranform Fault.

Many of Figure 2's basic tectonic elements were inherited from a transform fault associated with latest Precambrian break-up of the Supercontinent of Rodina (Thomas, 2005). The Susquehanna Transform Fault (Figure 6) is an example. The fault on one side of the Latest Precambrian rift flattened to become the dominant slip plane (Figure 7). The exposed deeper levels collapsed on its relatively flat surface to become semi-oceanic floor. The rift's opposite or passive side was part of the cratonic platform and remained little changed. In rifts, this early stage dominance in slip behavior commonly switches from side to side with a transform fault required to allow adjacent active floors to move in opposite directions. Eventually, those along strike-changes in a single coastline become jogs between recessive, actively sinking margins and protruding passive margin extension of the platform.

In this segment of the Appalachians, the Susquehanna Transform fault separated the passive Lancaster Platform from the active York Basin and its floor of rapidly sinking semi-oceanic crust above a cooling but still hot mantle (Figure 7). The red ellipses on Figure 7 suggest changing stratigraphy on the transform's opposite sides. On the Lancaster Platform, the basal Cambrian Mine Ridge clastics are only a few 100 m thick and transition upward into a relatively thin cover of Cambrian carbonates. In the York basin, west of the river, clastics and carbonates increase in thickness by factors of x5 to x10 and show equally dramatic facies changes. The Kinzers Shale on Figure 6 is an example.

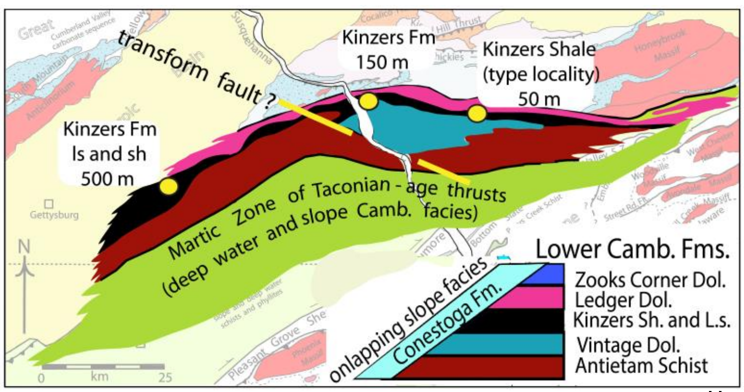


Figure 6. Approximate location of the shallowly buried Susquehanna Transform Fault below a linear stretch of that river. The color areas are sub-crop exposures of successive carbonate formations beneath climbing Conestoga Formation slope deposits that grew vertically in concert with them.

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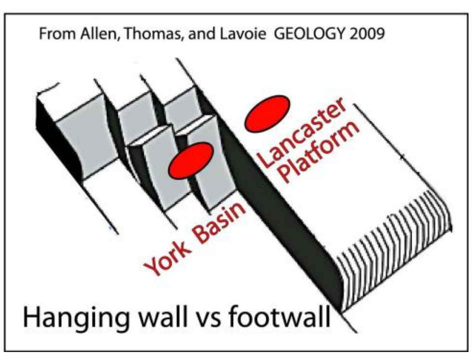


Figure 7. Differing subsidence rates across the Susquehanna Transform Fault. The hot active side of the original rift cooled and sank rapidly while the passive side sank very slowly. The text describes thickness changes of an order of magnitude between the two areas indicated by the red ellipses.

The stratigraphy on Figure 8 includes the Conestoga Formation, mapped in dark blue. This slope unit grew just downslope of and in concert with the edge of vertically growing, Cambrian platform carbonates (Figure 8C). Olistostromes with large fragments of the underlying and adjacent carbonate formation are commonly interbedded. The photo of 8B, located at the yellow dot locations on A and C, shows the Conestoga slope deposits resting on a the Vintage Dolomite, the same formation that cropped out just upslope of it as the source of the Conestoga's interbedded conglomerate lens The edge of the growing platform was just upslope of this location. As the Cambrian carbonate bank grew vertically the Conestoga kept pace as a paleo-shoreline indicator.

The white lines on the map are approximate paleo-shoreline/ platform edges for each formation on panel C. They are drawn just outboard of the outermost outcrops of each successive formation. Clustering of these lines at both edges of the map (Figures 8A and D) shows locations where the shoreline remained relatively fixed while the carbonate units accumulated. Stability of location was independent of platform growth rates that differed by ~ an order of magnitude between the two areas.

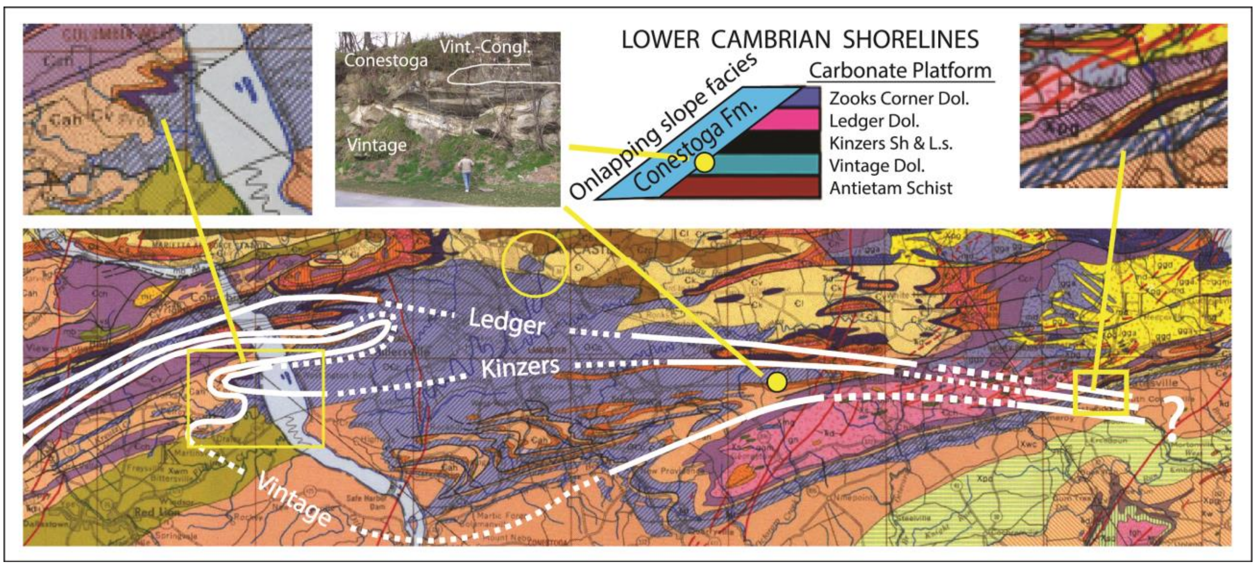


Figure 8. Changing Lower Cambrian shorelines across the Susquehanna Transform boundary. These limits of accumulating formations in subcrop exposures beneath the Conestonga Fm slope deposts (dark blue) record a collapse of the extant platform edge (circle), probably over the Paradise Fault. The fault should lie beneath the yellow-to-blue contact. The platform and fault originally extended to the SE but the Valley Forge Lineament at the S edge of the Mine Ridge massif truncated it.

The shoreline migrated very rapidly between those two stable locations. For the oldest unit the shoreline was nearly parallel with the linear segment of the river. About then, a subsidiary fault began to develop to the east as recorded by progressive migrations of the Kinzers and Ledger lines. Migrations finally began to stabilize along the Ledger shoreline. During Ledger accumulation, the Conestoga slope deposits suffered a large landslide collapse (circle). Now inaccessible quarries at the north edge of Lancaster exposed the Conestoga Formation with angular meter-scale blocks of Ledger, fallen from the headwall scarp as debris in Conestoga olistostromes. The new shoreline became stable as the yellow-to-blue boundary on the map, evidence of a new, small fault in the Cambrian platform. A quake along that fault probably triggered the slide. The fault, inboard of and at least sub-parallel to the transform edge, was probably part of larger collapse of the platform edge into the transform-bounded edge.

The fault needs a name and the town of Paradise located in its middle stretch seems more appropriate than its well-known tourist-attracting neighbor. Figure 8 also shows another aspect of its location at a major change between Mine Ridge and Honeybrook basement lithologies, an indication of hereditary control. At the map's edge, the Valley Forge Lineament truncates Paradise shorelines. It also truncates the Susquehanna Transform fault, located approximately by the facies data descried above and by epicenters of Lancaster-Reading Seismic Zone (Figure 2). These quakes, all of < 3 magnitude, < 5 km depth, and located as yellow dots on the figure, appear to be modern crackling of the shallowly buried edge of the N-Appl basement slab, a continuation of the westward plunging Mine Ridge and Honeybrook massifs (Wise and Faill, 2000). The SW end of that swarm turns SE approximately under the linear stretch of the river (Figure 2) as probable locus of Thomas's (2005) buried Susquehanna Transform Fault. The fault's NW projection just beyond the plunging end of South Mountain Anticlinorium suggests it was reactivated as the N boundary of the S-Appl basement sheet.

*** ACT 2: Interactions with the Oceanic realm ***

<u>Act 2, Scene 1.</u> The Taconian Orogeny as first major orogenic event to impact the coast.

Through the Cambrian and Mid-Ordovician, this edge of North America accumulated 1 – 2 km of platform carbonates across a broad area that extended from NY state through much of the S-Appl foreland to VA and and beyond. It was centered in a in a NNE-trending trough (Figure 9). Toward the end of Mid-Ordovician times, an island arc in the highly active, young Iaptian Ocean, rolled into the craton's New York Promotory. This initiated debris-laden turbidite flows that deposited the Martinsburg flysch from NY to WVA with isopacs along a new ~N55E trend.

Southwest of the NY Promontory, the Susquehanna Transform Fault protected the coastal deposition from the main impact. Like much of the Cambrian coastline, a ribbon continent's back-arc basin separated the coast from the main ocean basin. Arrival of the island arc began to close the basin and push its debris apron across the ribbon onto the intervening Octoraro Seaway, and compress its fill (Figure 10). That laterally shortenning mass soon developed enough height to flow onto the platform. The arc"s angular convergence geometry caused that flow to climb obliquely upward across the transform margin (Figure 11). This flow first detached large thrust sheets of the Octorato Formation at the rear and then acquired new sheets at the base to form a growing duplex. As advance along the transform edge continued, successive sheets picked up Conestoga Fm on Antietam schist, then Conestoga on Vintage, etc.,. Eventually the overflow dragged that duplex up over the Lancaster Platform edge and for abandonment as Cloos's famous Martic duplex (1941).

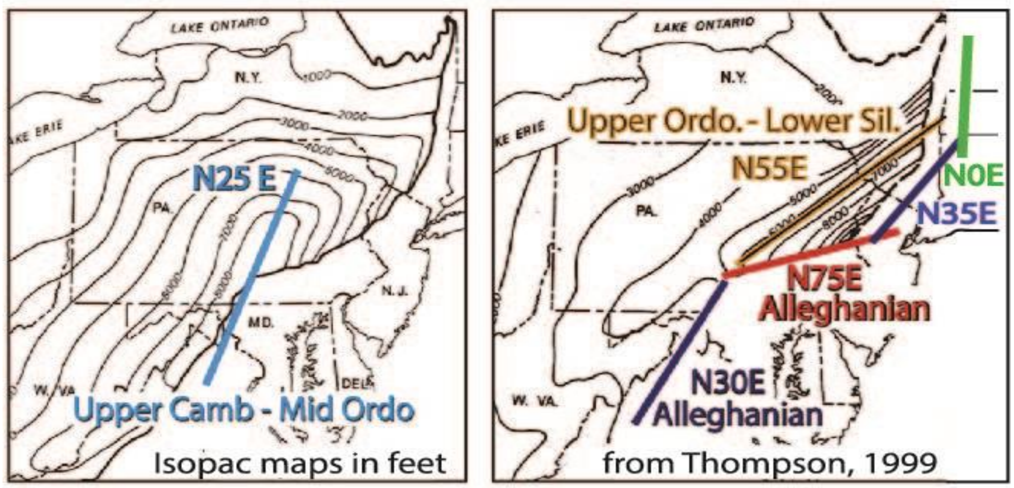


Figure 9. Changing orogenic trends with time along the continetal margin. Base map from Thompson (Geology of PA, 1999).

The "Martic Line" in its type localith is merely the the first offshore sheet that did not include Conestoga slope carbonates. About four sheets to its south, a more oceanic sheet includes enough magnetic material to define a "magnetic Martic Line" on aeromagnetic maps (Wise,1970). In its type locality, the Martic "Line" is atypical, a special creation of the Susquehanna Transfrom Fault, combined wiith chance direction of Taconian closure, and ideal exposure elevation by the Alleghanian Orogeny and modern erosion. Beyond the Martic type locality takes many forms but few are the actual Ordovician edge of the continent, and even fewer owe their origin to a transform fault.

The additional tectonic load depressed the Lancaster Platform edge and allowed easier flow onto and across it. Once onto the undeformed platform, gravity took over to drive the Lebanon Valley Nappes of quadrangle after quadrangle of inverted and recumbently folded, mosly Ordovician carbonate units and overlying Martinsburg/Cocalico shales to phyllites. Evidenee of stretching of oveturned limbs is widespread; non-stretching dolostone beds sanwiched between ductile carbonate beds appear as boudins of all scales (Figure 11). Exposutes of this nappe system extend from the Chickies-Oregon thrust on the south to the north edge of the Great Valley, interrupted only by the Mesozoic basin cover. The gravity flow process detached the upper units of the carbonate platform along with their overlying and barely lithified Martinsburg flysch to roll them over into inverted limbs of single folds with axial surfaces traceable for kms across strike. Tectonic transport/flow diffection lineations, boudin stretch directions, extension veins, etc., show a remarkably consistent N30W flow across the region (Wise and Werner, 2004).

The tectonic edge loading and flexural rigidity of the lithosphere produced an inland bulge. That uplift exposed some of the platform's upper beds for eerosion and redepositon as identifiable clasts in the resulting trough developed just ahead of the bulge. The flowing mass included the Hamburg Allochton as a piggy-back passenger on the caterpillar-like overflow and advance (Figure 10) for eventual deposition in the foreland trough on top of the ~ 1 km-thick Marinsburg flysch (Figure 8).

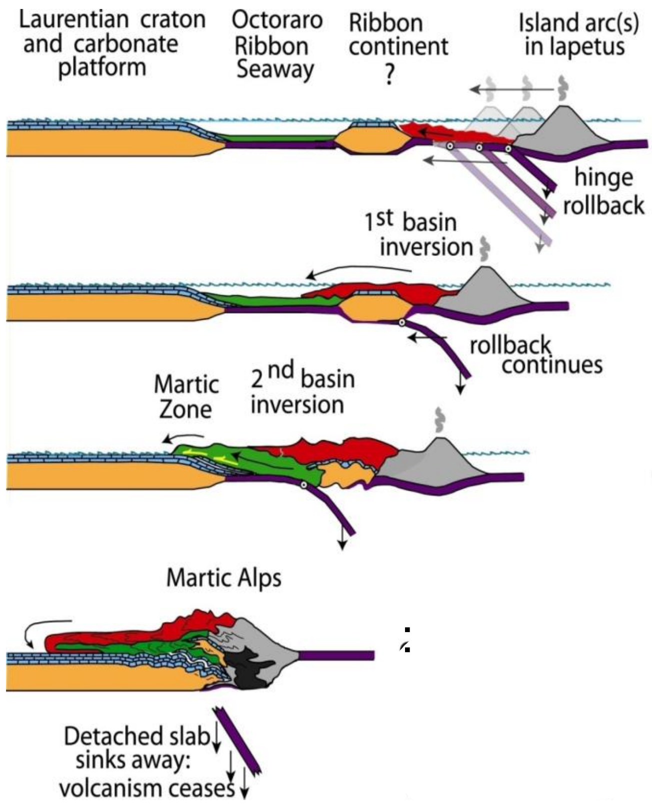


Figure 10.. Evolution of the thin-skinned Taconic Orogeny in the Susquehanna Region. As described in the text, the Susquehanna Transform Fault protected and preserved this region from modifications and complexities typical of other regions of the chain. Landward roll-back of an island arc pushed a thick mass of offshore deposits and its own debris up the Martic edge to become a gravity-driven mass that flowed for 50 km across the platform, overturning and stretching its underlying weak units as the Lebanon Valley nappes.

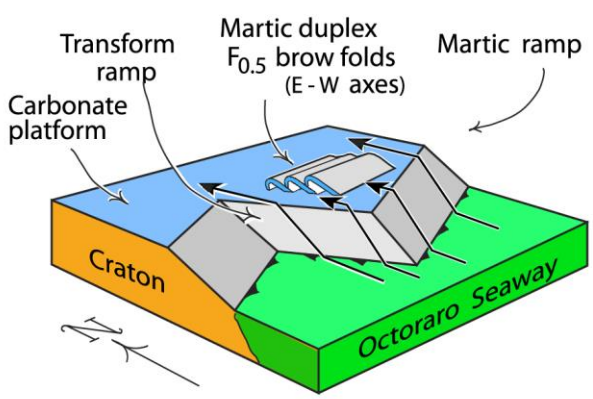


Figure 11. The Martic thrust duplex as a product of oblique upthrust of Octoraro deep-sea material up and

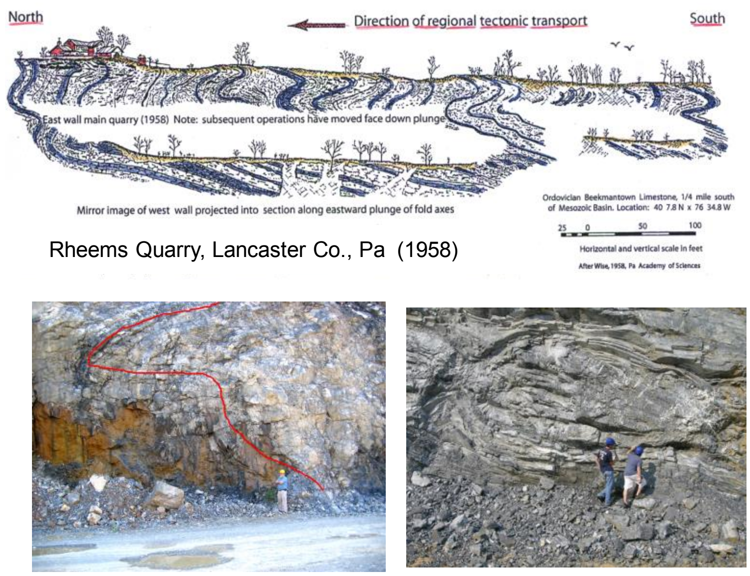
over the Susquehanna Transform Fault. The early-stage, small E-W trending F0.5 detachment brow thrusts at the west end of the duplex should not be confused with the more prominent N60E trending map scale F1 folds and S1 cleavage superposed by later sstages of the Taconian Orogeny. (This old sketch should to be changed so the folds are closer to the transform edge and it should continue much farther to the right.) (Brow fold mapping data from S. Schamel, honors thesis, F&M College, 1963).

This nappe complex, with preservation and exposure from its root zone south of the Martic Line to the mass's final resting place in Great Valley, is a superb example of large-scale gravity flow tectonics. A unique history created and presrved it at just the right elevotion for modern erosion to slice through it at multiple levels. The Martic area exposes both its early detachment of slope deposits and their thrusting advance onto the craton as well as contiuing flow and folding in the root while metamorphism, schistosity, and mineral isograds developed. The Lancaster Platform exposes cleavage and isoclinal folds with many types of N30W flow indictors of deeper level, downward effects of that overflowing mass. At that platform's north edge, the Chickies-Oregon Thrust dropped the overlying level of inverted and recumbently folded Lebanon Valley Nappes as an example of incorporation of the upper platfrom straatigraphic units into the overflowing mass. Finally, the Great Valley's Hamburg Allochthon preserves a sample of the oceanic mass with its own history of multiple Cambrian and Ordovician offshore assembly of pieces of the oceanic floor (Ganis, 2001, 2008). It includes the town-sized area of Jonestown volcanics (Figure 2) as a former seafloor package with pillow lavas.,volcanic breccias and sedimentary fans, intermingled with former deep sea oozes (Lash, 1984).

The famous argille scagliosa (scaly clay) of Italy (Speeranza, 2002, Parotto, 2004) is a similar but much larger island-arc-driven overflow mass. Miocene arc rollback drove large masses of the Tyrrhenian seafloor onto the Italian craton. Part of it was onto the 5 km thick, Central Apennine carbonate platform. Unlike this example, that arc did not reverse or have its down-going slab drop away to limit further evolution. Instead that roll-back continues eastward beneath Italy today. The deep slab generates the Latian volcanics of Rome and Naples. Contiued eastward roll-back produced a series of rising Apennine fault blocks that shifted the initial trough progrressively and more complicated example version of this simpler and smaller example of the same process.

Act 2, Scene 2. Mid Paleozoic terrane. accretion

In the Mid-Paleozoic, a piece of mostly Taconian hinterland, characteristic of the Southern Berkshires of southwestern CT and southern NY docked with this area along the Pleasant Valley- Huntingdon Valley Shear Zone (Figure 2). This area's geology has frustrated generations of geologist but more rcent work, including that by the convenors of this session, has begun to unravel it's complexities. Better petrology, new field work, and more abundant precision dates that include fingerprint scans of zircon populations are beginning to expose its secrets. Much is described and revealed by many authors in the 2014 NE GSA Field Trip and is likely to be expanded upon at this meeting. The Alleghanian discussion below includes some discussion of these events but there is neither space nor talent for further disussion here.

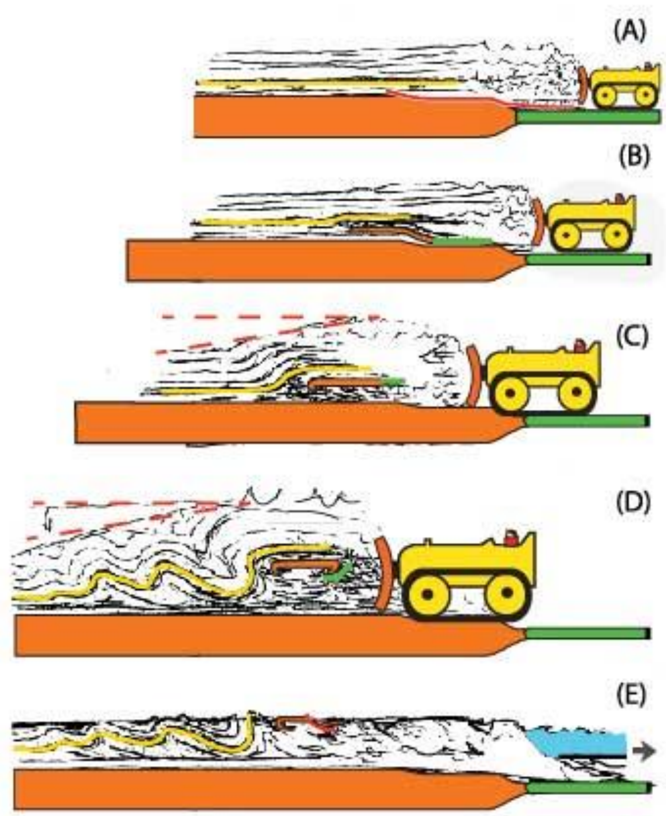


<u>Figure 12</u>. Recumbent folds of the Lebanon Valley nappe system. Completely overturned beds are in the Beekmantown Formation. Contiued flow and stretching did not include the right-side-up limbs that created widespread boudingage development by breaking non-flowing dolostone beds and sucking of the ductile limestones into the voids. The Rheems stratigraphy has thinner dolomites and hence smaller boudins than the pictured thicker bed from a quarry two km to the south. (Wise, 1960).

*** ACT 3: Superthrust emplacement ***

<u>Act 3, scene 1.</u> Superthrust advance, Great Valley Monocline, and roll-under tectonics.

This paper's prelude section described the origin of the N-Appl and S-Appl basement slabs and their juxtaposition at depth to form the Susquehanna Piedmont Superthrust during the initial stage of the Alleghanian Orogeny. This section continues with a description of the Piedmont Superthrust's advancing uplift into the foreland. Figure 13A-D suggests the sequence of events: 1) early bedding-parallel shortening; 2) bedding plane thrusting and doubling of some stronger units, especialy the strong Silurian clastic unit that overlies the Great Valley's weak Ordovician section; 3) initial draping of strong Silurian units across the thickening Great Valley stack of weak units as an early-stage development of the Great Valley Monocline; and 4) continued roll-up and increase in dip until the monocline is fully developed with its vertical limb as the superthrust's frontal structure.



<u>Figure 13</u>. Alleghanian evolution of the Great Valley Monocline (A to D), followed by Triassic collapse of its underlying mushwad (E).

A mini-version of this sequence occurs a 500 m-long road cut through the frontal Silurian units at Schuylkill Gap, N of Reading (Figure 14). That cut, fresh in 1958 but now severly degraded, exposed 175 m of doubled Silurian strtigraphy cut by two "normal faults." The one through strongly hematite-banded Clinton Fm has unmistakably clear normal fault drag. This displacement sense might be no problem in other environments but are anmalous in this domanantly compressional regime. If the beds are returned to their deppositonal orientation, the "normal" faults become forward directed underthrusts as suggested in the caption. The underthusting is a middle part of the 1, 2, 3 advance sequence described above.

The unfortunate but almost universal habit of calling such structues "back thrusts" is a reversal of tectonic transport, at least in the mind of the interpreter. A common result is mis-interpretaton of a new and non-existent orogenic phase. Underthrusting followed by bedding plane doubling of the section, followed by progressive rotationthese eds draped over the superthrust's advancing front to become the bulldozer blade for its continued advance..

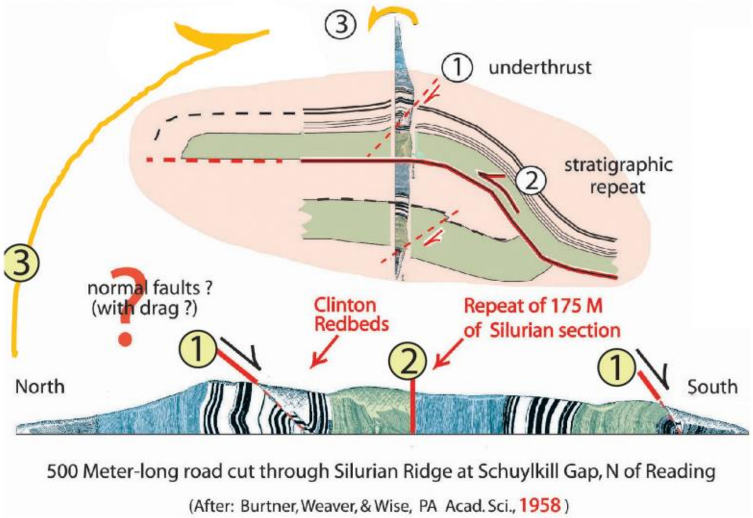


Figure 14. *Vertical limb of the Great Valley Monocline over the Piedmont Superthrust's frontal zone. The now degraded road cut was well-exposed when mapped in detail in 1958 (Burtner, 1958.). The large "normal faults" are underthrusts developed in early stages of the Great Valley Monocline's origin.*

The foreland immediately ahead of the advancing vertical limb remained relatively undistrubed, it's early-stage advance was concentated in the weak, still-horizontal C-O units. The frontal limb, once established as the bulldozer blade, began to plow into those units and, like modern snow plows, the resulting pile ahead of it developed a constant critical taper angle at it's top ((Figure 13C and D). As advance continued, the pile grew in width and height while the blade accommodated this by progressive roll-up of the flat Silurian unit in the forelalnd to became the bottom of the rising blade (Figure 15).

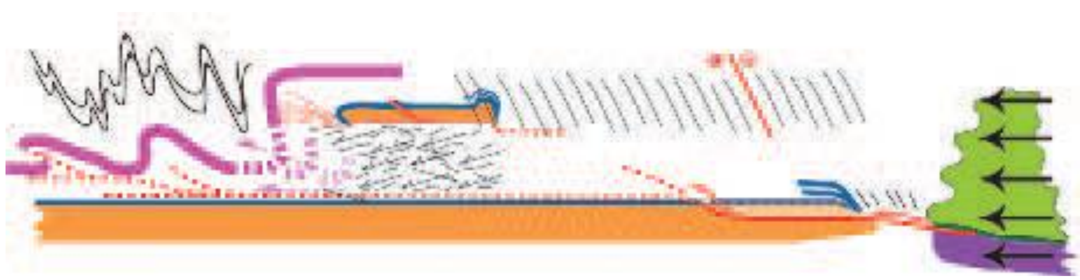


Figure 15. A schematic history of the Alleghanian Orogeny as described in the text.

The Great Valley and its basement slab were lifted by new additions of weak, overrun units at its base.. These grew into the mushwad that raised the Piedmont Superthrust by 15 km and draped the Great Valley Monocline over its front (Figures 3, 5, 13, and Thomas, 2001). The only publically available data on this mushwad that I have ever seen is a seismic traverse across the entire Piedmont just east of the Susquehanna River (Parrish, 2010). Figurre 16 is a sample of the data from under the Great Valley near Lebanon. The processing is only the contractor's first -cut but even at that level it shows the predicted package-after-package of semi-horizontal units, all kinked by later horizontal compression and offset by numerous- small displcement, steeply dipping faults.

Resistancce to early stages of the basement sheet's push from the rear buckled its edges and broke it's center to form the Chickies-Oegon Thrust (Figure 2, 15, and 17A). These folds with N70E trends extended beyond the present basement exposures as suggested in Figure 17A. Some uplifted Taconian structures of different trends (Figure 18). As the mushwad grew upwards the effective compressional shortening became more concentrated in ever deeper levels. This rolled-under the adjacent terranes and fault lines to give the appearance of fault truncation by the newly created Valley Forge Lineament with a N75E trend, ~5° CW of the slightly older fold trends. Continuing compression imposed a south-dipping S2 cleavage on most rocks of the inner Piedmont and on the Lancaster Platform (Figure 15). At the Great Valley's buried leading edge it rolled-under the late-stage extremely linear structure identified above on the Lidar imaage. This structrue is parallel with the Valley Forge Lineament and the Reading Prong front. It cuts the Prong's frontal carbonate structures, truncates the internal grain of the the Hamburg Allochthon, and infaults part of the allochton along it's strike.

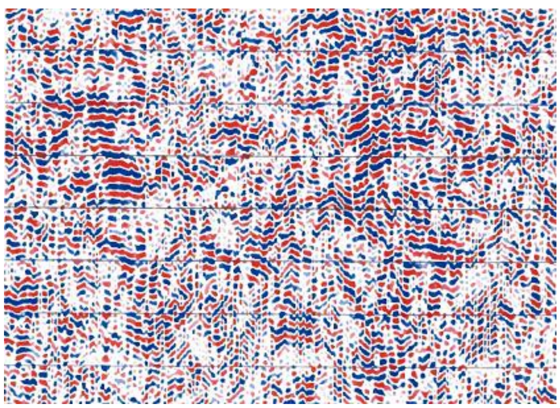
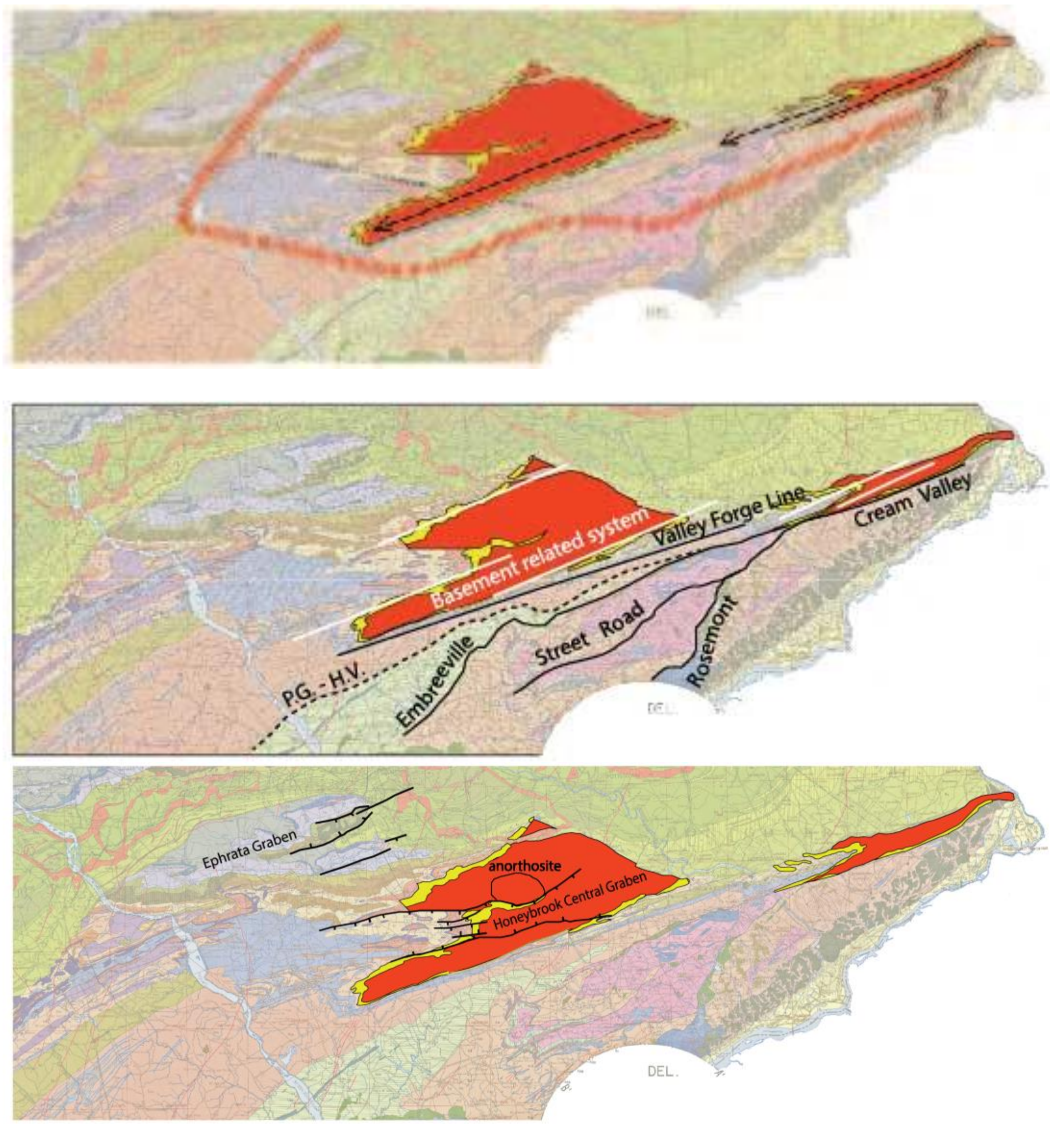


Figure 16. *Seismic image of the mushwad beneath the Great Valley. This minimally-processed sample of a seismic line across the Susquehanna Piedmont is from uncertain depth beneath the Lebanon Area of the Great Valley (Parrish, 2010) It shows a pile of probable C-O carbonates, kinked and deformed by later horizontal compresion and broken by small vertical faults.*

The multiple overprints can be organized into a somewhat coherent scenario by taking liberties with and merging data and interpretations Southworth, et al, (2010), Gates and Valentino, (2010), and Bosbyshell, et al., (2014)., may explain the origi of the Valley Forge Lineament's origin (Figrue 19). The Octoraro and Peters Creek deposits were initial deposits of the new ocean floor, the Octoraro off the Martic edge and the Peters Creek as a turbidity current deposits off the New York Promontary. Both became components of outer parts of the Taconian orogeny while the Brandywine Terrane developed somewhere to the north, closer to its orogenic core.



<u>Figure 17</u>. Evolution of the Valley Forge Lineament As described in the text early compression buckled the and folded the slab. ***An alternative second thought*** *and possibly more probable interpretation would eliminate the Honeybrook Graben on C and add its northern fault to A as part of the Chickies-Oregon thrust system, and the sourhern fault to the Valley Forge system.*

A new phase of major strike-slip dextral displacement brought the Wilmington Terrane into contact with these structures and splays may have helped separate them by that motion especially along the Huntingdon Valley and Embreeville Faults. A later change in tectonic orientation truncated all these structures along or rolled them into the transverse Cream Valley Fault.

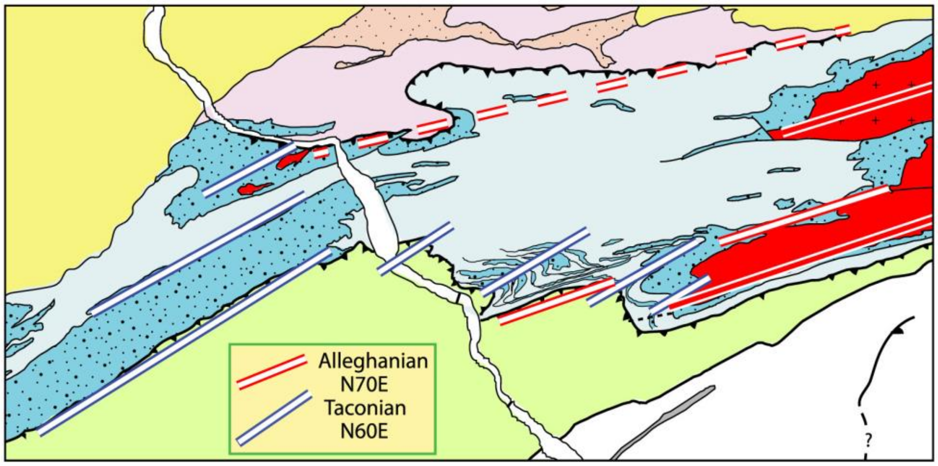


Figure 18. Aleghanian sructural trends superposed on Taconian ones. The north flank of Mine Ridge tilted the older structures ito a linear belt above it. See figure 9 for another scale view of changing tectonic trends. transport direction.

Subsequent early sages of Alleghanian compression detached all this as part of the basement sslab and folded it's edges into Mine Ridge and related folds. Continuing compression in a slighly more clockwise sense and rise of the slab on the growing mushwad, transferred the strain to progressively deeper levels. This rolled-under the Devonian Cream Valley cut-off edge and its along-strike extension to become the the Valley Forge Lineament with ~ 5° deviation in strike from the Mine Ridge system. Absence of displacement of the Martic Line on the nose of Mine Ridge (Figure 18) precludes any significant Mine Ridge or Valley Forge Lineament fault activity at that west end . Some faults at the lineament's east end may have been reactivated but deformation included significant roll-over and roll-under truncation.

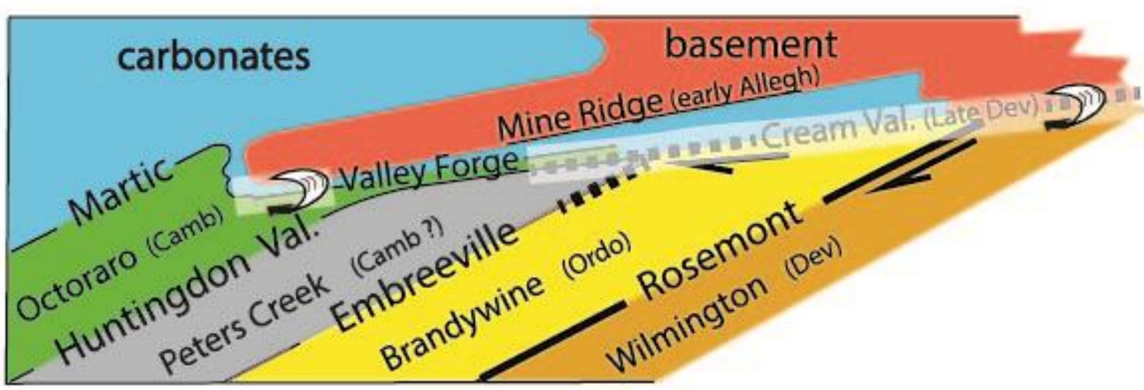


Figure 19. *Schematic model for origin of the Valley Forge Lineament. See text.*

Act 3, scene 2. New Jersey Highland's new dirrection of foreland advance .

In later stages of the orogeny, the closure direction shifted from NNW to WNW (Figure 9). The NY Promontory and VA took the brunt of the impact, while the Susquehanna region's recessive location excluded it from that intense deformation

As suggested by Figure 9, this new impact direction by the NJ Highlands and its leading edge of Shawangunk Silurian outcrops sheared and end-loaded the Susquehanna Piedmont Superthrust. Greater displacement on the north buckled the sheet horozontally into at least two ~ 15 km-wide kink zones (Figure 20). The strong basement slab broke into multiple small domains defined by on the state map by parallel outcrops of Hardyston Quartzite. That outcrop pattern breaks the Reading Prong basement into several lettered domains. Blocks F and I are the largest and most consistently oriented with tends that persist northwestward across the Great Valley to the Silurian Ridge. Segments G and H, between them, break that segment into smaller altenating kinks as a zone and pattern that trends NW across the valley to become the Hawk Mountain fold group.

Farther SW the Reading Prong ends with a similar but more complicted kink. The tear fault interpretation of Little South Mountain would make domains B and D displaced terminal structures of the S-Appl Megathrust. E becomes a fault block associated with their attachment to the N-Appl block. The dextral slrike-slip fault that separated B from D passes through the gap north of C. The details of these multiple intersections await clarification by separation of their special history from that of the Schuylkill Kink Zone that passed through them with further modification by Triassic-Jurassic tectonics as described in the next section.

In the bigger picture, that end-loading compression thickened weaker underlying weak rocks NE of the Hawk Mountain Kink Zone to uplift that segment of the Great Valley create the west-plunge that ends the Hamburg Allochthon. It also uplifted domain I commonly regarded as a vast sheet of overturned basement klippes exposed in synforms and Great Valley units exposed in t intervening antiforms (Drake and Epstein). Ths area outlined on Figure 21as the Allentown-Bethlehem-Easton area may owe its deeper erosional exposure to thickening of the underlying units by that s kink-related compression from NJ.

This younger phase of New Jersey Highland phase of compresion also caused broad folding of the foreland. The thick apron of Pocono Platean cover inhibited deformation of its immediate front with resulting concentration of forward displacement as the Lackawana Synclinorium. That structure postdates Reading Prong deformation: coal masures within it are folded and thrust again and again with Reading Prong trends and truncated by turn-up against the younger erosion-resistant, ridge-forming Pottsville Conglomerate's valley walls. Southward of that valley, the dominant axes of the other Anthracite coal baasins are parallel with the Valley Forge late-stage underthrust structures but regionally all plunge towards an synclinal axis with trend similar to the Lackawana axis (Figure 21). In an even broader sense, a thickened and uplifted region continues westward across the Ridge and Valley to the Susquehnna River .One dashed line on Figure 21 shows the nose-line of west-plunging folds in the Pocono Fm. The next line to the west is the approximate beginning of less elevated marine Devonian units on a line northward on the projection of the Blue Ridge.

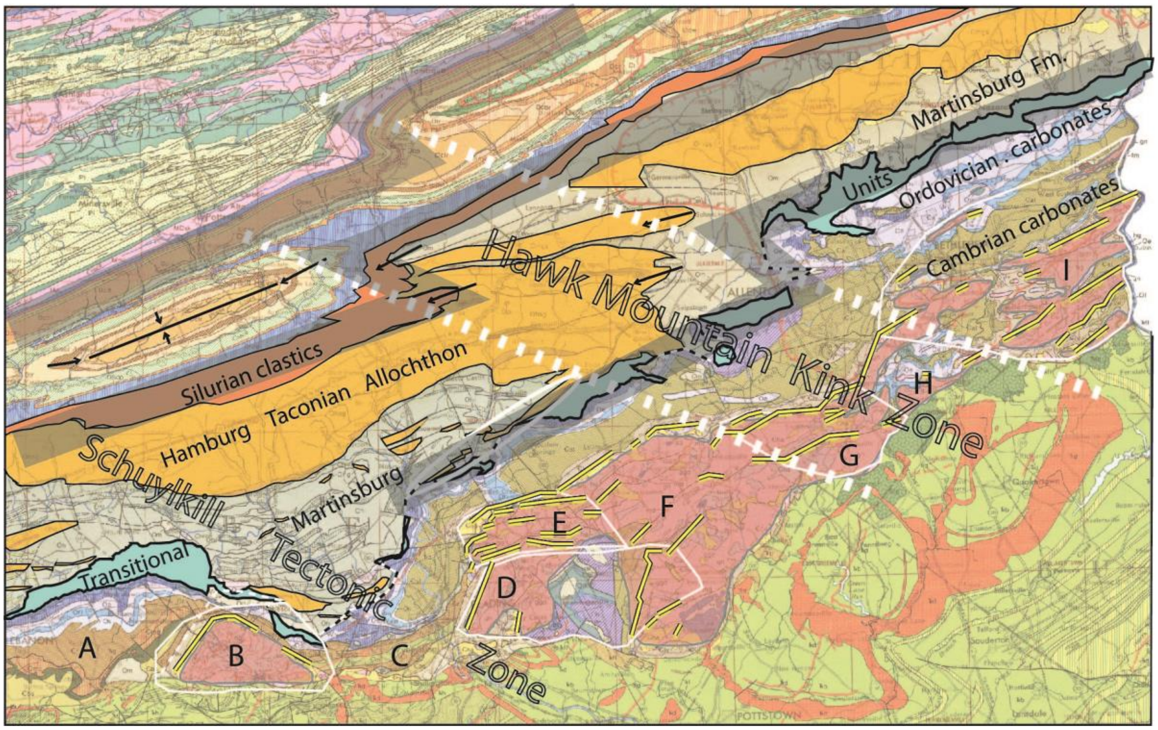


Figure 20. Mega-kink zones imposed on the Piedmont Superthrust sheet by the Shawangunk-NJ Highland orogenic phase.

Many debates and even heated arguments have swirled about the origin of the arc in Pennsylvania's Ridge and Valley Province. The other half of that problem is the origin of the angular junction of the Reading Prong arm of Piedmont Superthrust with the Shawangunk/NJ Highland segment. A similar angular junction occurs at the south end of the Blue Ridge.

The rule along the Appalachians is arcuate boundaries that alternate with angular internal ones. The Susquehanna Piedmont Superthrust's junctions suggest a relatively simple answer. Early rift and transform fault orign of the continental margin left a segmented cratonic edge of the Appalachians to be further segmented by younger orogenic overprints. In the Alleghanian collision, an early stage of WNW-directed motion detached edges of the of the craton to form the Blue Ridge half of thes uperthrust. A change to NNW contient-continent collision detached slabs of the cratonic margin from intervening segments. It rolled or broke them into massifs perpendicular to the transport direction and joined them with the older slabs to form the Appalachian Superthrust. A third change in closing direction drove the entire Appalachain

Superthrust in a WNW direction. These closing directions were not constant along the entire chain. The last one had increasing displacement southward with consequent pivoting and CW change in transport direction. In the Susquehanna region this reactivation transported the Blue Ridge segment in its third displacement.

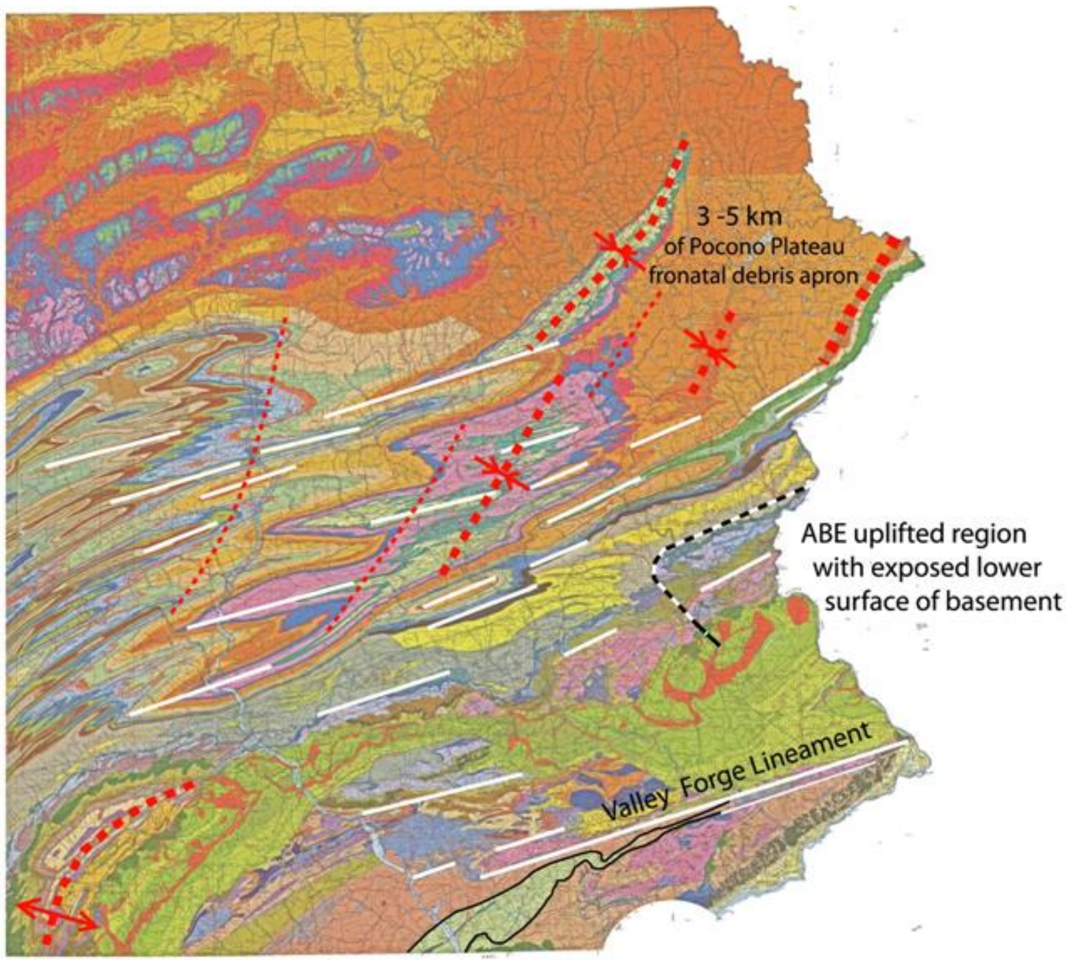


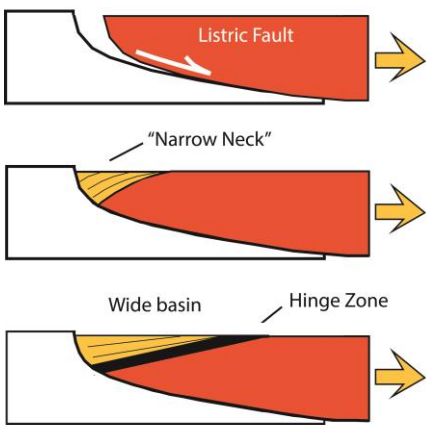
Figure 21. Reading Prong linear structures overprinted by Shawangunk-NJ Highland structures (in red).

Basement massifs at converging junctions became anguar overprint zones like that at the Delaware River or sheared and buckled large kink zones in the older sheet. In the foreland it overprinted the older fold systems with a new orientation. At diverging junctions the new faults new faults sliced across the rear parts of the the older systems terminal structures to incorporate them as part of the new sheet. In distal parts of diverging junctions displacement, the third displacement was sub-parallel with the first. It reactivated some of the older trends, superposed sub-parallel fonds on them and overprinted their new fold system on the older trends of the older system with no requirement of extensile strain. The map pattern became a giant drape fold with a vertical axis.

*** **ACT 4. Orogenic collapse** ***

The start of a new cycle in the Triassic ended the Paleozoic Wison cycle and began a new one. The previous three acts described the first cycle's cratonic origin, additions to the new craton's edges and finally the closing of it's ocean to create the supercontinent of Gondwana. It had ~ 30 m.y. reign of glory that involved isostatic uplift and erosional stripping to a rolling landscape at essentially modern levels before the end.

By Mid-Triassic, slow stretching began to open the modern Atlantic. Removal of lateral support caused the outboard half of the giant mushwad to begin slow collapse toward the SE as a giant landslide. It had a steep headwall fault that flattened downward in a teaspoon-shaped listric shape. It left a headwall void as it moved and an outward-migrating hinge-line void for gravity collapse too keep filled with rock and above that for surface processes to fill a new sedimentary basin. As stretching and sedimentation continued, that tilt-and-fill process tilted the contents into a narrow zone of steeply dipping sedimentary rocks, the 10-12 km-wide "narrow neck" that separates the Newark and Gettysburg Basins (Figure 22A -B, and Figure 2).



<u>Figure 22</u>. Differing types of collapse basins controlled by presence of a strong basement slab. Gravity collapse will fill the open space associated with a flattening etension fault. Presence of a strong slab will expand the distance between headwall and hinge zone and require a much shallower dip to fill the same volume.

Those collapse mechanics applied to to the relatively isotropic rock mass in the gap with no basement slab. Where the strong basement slab was present, as in Figue 22C, the hinge line was fixed at a much greater distance from the headwall. The volume to be filled was the same but a much more gentle tilt was needed to replace it. The Newark and Gettysburg Basins owe their width to their underlying basement slab. Their locatization on the outboard half of each basement massifs resulted from flow collapse of the outboard half of their underlying mushwads. Between them the Narrow Neck was inevitable. It's ~ E-W trend probably reflects a headwall reactivated along the Little South Mountiain's tear fault.

During detachment of the N-Appl basement slab, the migrating tip climbed westward to "daylight" through the slab. The resulting westward thinning controlled the westward changes in Newark Basin's tansition inot the Narrow Neck. Th thinner slab at the west of the broad Newark basin tited more steeply with less distatance between headwall and hinge zone. This shearing action made use of the Schuylkill tectonic zone to separate the different tilts and hinge locations (Figue 23). The white line separates untilted Honeybrook basement from the slowly tilting wide basin. The Honeybrook side's hinge line is farther north at the location later pirated by a Jurassic intrusive. Total vertical displacement increased northward along that line between the two basins that were tilting at different rates. The more rapidly sinking side became the Hammer Creek Basin, progressively filled by alluvial fans pouring through the Little South Mountan gap. Individual fans, lightly colred on the figure, show the earliest southern-most fans came across the low initial scarp with little while younger, larger ones to the north were deposited over the increasingly higher fault scarp.

That scarp, now occupied by a Jurasic dike, was part of the CAMP (Central Atlantac Magmatic Province) volcanism associated with the break-up of Gondwana. Dates of that system concentrate in a m.y. time-span at ~201 m.y., just after the beginning of Jurassic times. In the basins, most CAMP intrusives are sills and these fix an upper time limit on basin tilt. The sills in the Narrow Neck have esentially untilted horizontal geometry as they slice through the neck's steeply tilted Upper Triassic beds. The same pattern exists in less dramatic form for the more gently dipping basin fills of both adjacent broad basins.

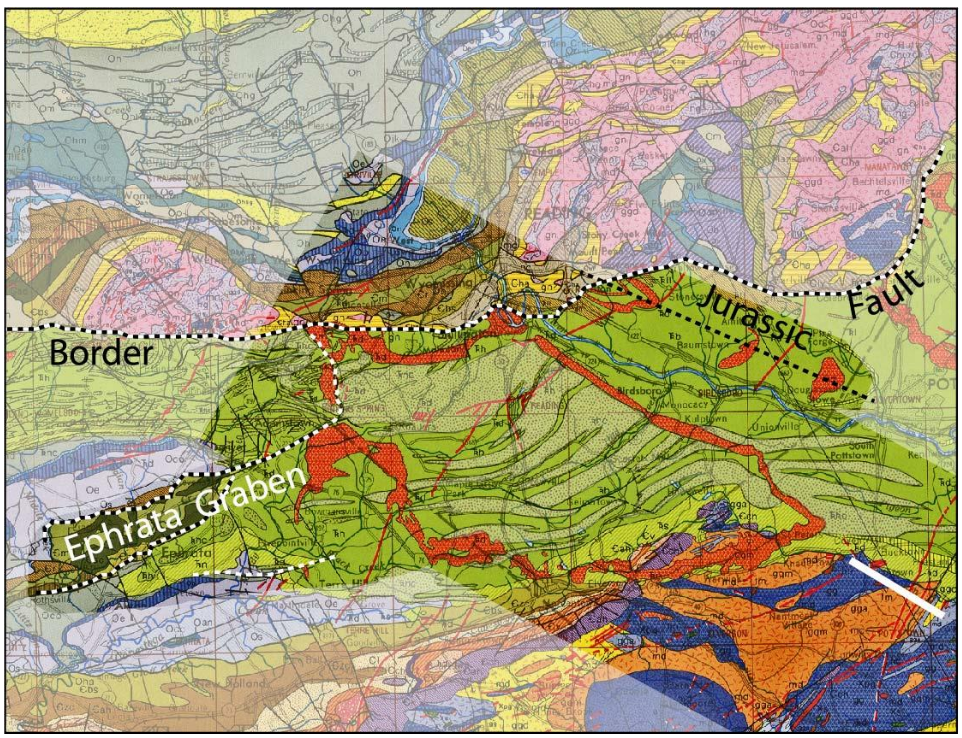


Figure 23. Mesozoic structures that reactivate the Schuylkill Tectonic Zone. The lighter greens are the Triasic Hammer Creek alluvial fans that fill the faulted basin. See text for other details.

The dike that followed the bounding scissors (Figure 23) became the feeder for the differentiated sill's magma as described by Scogi, et al. Two additional fault zones completed the sill's structural control into a rhomboidal outline. One follows the border fault and the other is part of a swarm of faults that separate the sill block from the Ephrata Graben (Figure 23), localized along in a zone of N-S trending faults that separate the sill-block from even thinner basement to the west. The graben's western terminus approximately the platform edge defined by the the Lancaster-Reading Seismic Zone.

***** The End: catclls, rotten fruit, etc. *****

Acknowledgements and apologies

This is assembly of ideas and data has been accumulating ever since my introduction to the Martic Problem in 1949. Even though I was an incoming freshman geology student at Franklin and Marshall College, Pete Foose suggested I might tag along with the Field Conference of Pennsylvania Geologists. As a result, I heard Ernst Cloos describe the Martic Problem standing at its classic exposure in the New Providence railroad cut. Since then, interactions with and contributions by several generations of colleagues, students, and friends, far too numerous to mention by name, have helped refine and improve this larger view of that and a host of related problems. For all those individuals and those that a failing memory has not brought up, I offer my acknowledgement and thanks. In recent years, conversations with Mike Williams have sharpened and added to the ideas. Several years' work on the Taconic and Martic problem with Bob Ganis have helped isolate this area's thin-skinned Taconic problem from basement-related Alleghanian tectonics. Discussions with Bob Hatcher have been part of the background for the slab detachment mechanism and those with Bill Thomas have refined the mushwad picture.

This is not a finished product but an agglomeration of data and ideas produced by generations of geologists, partially digested during 2/3 of century, and frantically assembled to meet editor Hal Bosbyshell's deadline only four weeks late. As such, I trust readers will forgive some redundancy, lack of referencing, a few un-finished illustrations, and minimal time for proof reading and proper editing. References are incomplete but credit is given to sources as best as I can remember. Some of the proposed tectonic mechanisms probably have application to other segments of the chain or beyond. Many are tentative, speculative, and potentially outrageous but perhaps the overall document can serve as a first-draft integration of the this critical region's tectonic evolution, something for the next generation to test refine, correct, and expand with new data and ideas.

DUW

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The Piedmont: Old Rocks, New Understandings

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ABSTRACT

The results of recent geochronological investigations in the central Appalachian Piedmont of southeastern Pennsylvania and northern Delaware have prompted a reevaluation of the tectonic history of rocks in the high grade metamorphic core of this portion of the orogen. Monazite results reveal that major metamorphism and deformation in the Piedmont occurred in the Silurian and Devonian and.is not related to the Taconic orogeny. The West Grove Metamorphic Suite and Wissahickon Formation lie on opposite sides of the Rosemont Shear Zone. Distinct differences in metamorphic history and detrital zircon provenance between these units demonstrate that the Rosemont Shear Zone is a significant tectonic boundary. Detrital zircon results further show that the Chester Park Gneiss likely originated in a peri-Gondwana basin and may be correlative with the Moretown Terrane in the New England Appalachians.

INTRODUCTION

This field trip focuses on the high-grade metamorphic rocks in the central Appalachian Piedmont of southeastern Pennsylvania and northern Delaware (Fig. 1). Metamorphism and deformation in the rocks of the area have long been interpreted to be products of the Taconic Orogeny: the Ordovician collision between a peri-Laurentian volcanic/magmatic arc and the Laurentian margin (Wagner and Srogi, 1987; Aleinikoff et al., 2006; Wise and Ganis, 2009; Sinha et al., 2012). The importance of younger transcurrent deformation, that likely translated the Taconic hinterland from its original location within the orogen, has also been recognized (Valentino et al., 1994; Wise and Ganis, 2009). Monazite results (Bosbyshell, 2001, 2004, 2008; Pyle et al., 2006; Bosbyshell et al., 2014, 2016) show that, while there is evidence for middle to late Ordovician, Taconic-aged, metamorphism in some rocks, the bulk of the metamorphism is Silurian through mid-Devonian in age. Thus, the Salinic and Acadian orogenies, which mark the accretion of Gondwanan terranes Gander and Avalonia, respectively, in the Northern Appalachians played a significant role in Piedmont tectonics. These results also underscore the importance of transcurrent deformation in the assembly of the Piedmont and demonstrate that today's configuration is the result of tectonism that spanned a significant portion of the Paleozoic Era and not merely the Ordovician.

In addition to monazite geochronology, detrital zircon analysis has also been important in reshaping our understanding of the bedrock geology and the provided key evidence in support of defining a new lithodeme in the lower Paleozoic metasedimentary rocks. Bosbyshell et al. (2014) cite differences in detrital zircon provenance, geochemistry of interlayered amphibolite, and metamorphic history between units that have historically been considered part of the Wissahickon Formation. They proposed a new lithodeme, the West Grove Metamorphic Suite (WGMS), for the portion of the Wissahickon formerly known informally as the "Glenarm Wissahickon." Detrital zircon results (Bosbyshell et al., 2014; 2015) further reveal that the Wissahickon Formation, *sensu stricto*, likely received sediment from a Gondwanan source in addition to a Laurentian (Grenville-aged) source. The Chester Park Gneiss, which outcrops along the Coastal Plain onlap (Fig. 1), contains relatively few Grenville-aged detrital zircon and

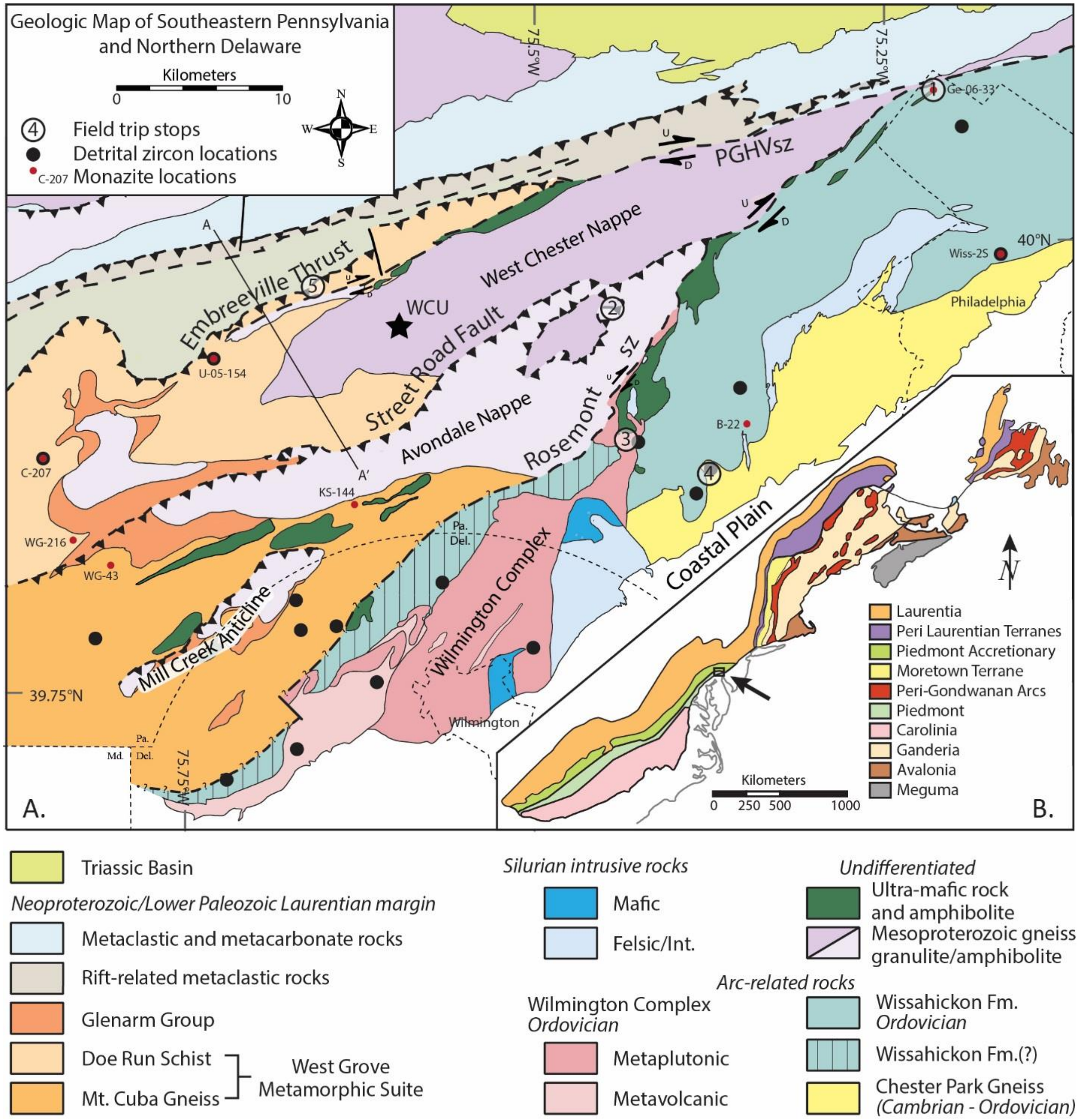


Figure 1. (a) Geologic map of southeastern Pennsylvania and northern Delaware, modified after Blackmer (2005) and Schenck et al. (2000). PGHVsz = Pleasant Grove and Huntingdon Valley shear zone. Cross-section A-A' is shown in Figure 2. (b) Map of Appalachian orogen after Hibbard et al. (2006) as modified by Macdonald et al. (2014), arrow indicates location of map (A).

was apparently deposited in a basin adjacent to a Gondwanan terrane. This unit may be correlative with, or is possibly a fragment of, the Moretown Terrane (Macdonald et al., 2014) in New England.

Geologic Setting

The bedrock geology of southeastern-most Pennsylvania and northern Delaware in the area of the field trip (Fig. 1) consists of two main lithotectonic elements: rocks of the Laurentian margin, including

Mesoproterozoic basement gneiss and a lower Paleozoic cover sequence, and an early Ordovician magmatic/volcanic arc and associated metasedimentary rocks, including the Wilmington Complex and Wissahickon Formation. Silurian-aged intrusive rocks are present in the arc terrane, but are not found in the Laurentian rocks. The Rosemont Shear Zone (RSZ) is the boundary between arc rocks and the Laurentia margin.

Laurentian margin rocks

Rocks of the Laurentian margin include the Mesoproterozic-aged Baltimore Gneiss, which occurs in the West Chester nappe, the Avondale nappe, the Woodville dome and the Mill Creek anticline, and cover rocks including the Glenarm Group and West Grove Metamorphic Suite (Fig. 1). The basement gneiss is among the least studied rock in the Appalachian orogen. Granulite facies metamorphism in the West Chester gneiss was investigated by Wagner and Crawford (1975) and Wagner and Srogi (1987). Limited geochronologic data for the gneiss (Grauert et al., 1973, 1974) is consistent with the Mesoproterozoic age obtained from zircon results in Baltimore gneiss domes (Aleinikoff et al., 2004). An age of 1075 +/- 9 Ma was from zircon in the Woodville Dome (John Aleinikoff, unpublished data). Zircon results for gneiss of the Avondale massif (Grauert et al., 1974) are quite discordant and indicate a large component of Paleozoic zircon growth.

The lower Paleozoic cover sequence which unconformably overlies the gneiss consists of the Glenarm Group: the Setters Formation and Cockeysville Marble; and the recently named West Grove Metamorphic Suite (WGMS) (Bosbyshell et al., 2014). The WGMS consists of rock historically known as Wissahickon Formation (Bascom, 1902, 1905; Bascom et al., 1909). Named for exposures along Wissahickon Creek in Philadelphia, Pennsylvania, the Wissahickon Formation once extended throughout the mid-Atlantic Piedmont, from New Jersey into Virginia (a full discussion of the evolution of Wissahickon nomenclature is given by Schenck, 1997). Differences between the Wissahickon Formation east of the Rosemont Shear Zone, including the type locality in Philadelphia, and the metasedimentary gneiss and schist west of the Rosemont (Fig. 1) have been recognized at least for 25 years, since Faill and MacLachlan (1989) defined the Philadelphia terrane (east) and Brandywine terrane (west). The informal nomenclature associated with lithologic units within the terranes evolved, but the name "Wissahickon" was retained for rocks in both terranes (e.g. Blackmer 2005). Bosbyshell et al. (2014) outlined differences in the detrital zircon populations, geochemistry of interlayered amphibolite, and metamorphic history between the western and eastern rocks and proposed the WGMS, to encompass amphibolites and metasedimentary rock west of the RSZ and above the Glenarm Group.

The West Grove Metamorphic Suite consists of rock formerly mapped as "Glenarm Wissahickon" (Blackmer 2004a, 2004b, 2005) and includes metasedimentary rock of the Mt. Cuba Gneiss, Doe Run Schist, and Laurels Schist as well as metavolcanic rock of the White Clay Creek Amphibolite and Kennett Square Amphibolite (Smith and Barnes, 1994, 2004; Plank et al., 2001). Detrital zircon results, discussed further below, suggest that the depositional age of the WGMS is most likely early to middle Cambrian.

Mt. Cuba Gneiss consists of interlayered pelitic gneiss and pelitic schist. Small bodies of granitic pegmatite appear to be locally-derived products of partial melting. Mt. Cuba Gneiss is the metasedimentary cover rock above the Street Road Fault in the Avondale nappe (Blackmer, 2004b). Mt. Cuba Gneiss also overlies Baltimore Gneiss in the Mill Creek anticline, although the bedrock geologic map of Delaware (Plank et al., 2000; Schenck et al., 2000) uses the older formal nomenclature and the unit is shown as Wissahickon Formation.

Doe Run Schist is silvery to dark gray pelitic schist dominated by medium-grained muscovite giving the rock a spangled appearance. Locally, quartz and feldspar content is sufficient to make the rock psammitic. The Doe Run Schist is in the hanging wall of the Embreeville Thrust (Fig. 1), in contact with Baltimore Gneiss of the West Chester massif and with the Glenarm Group and Baltimore Gneiss in the Woodville nappe (Blackmer, 2004a). The Laurels Schist (Blackmer, 2004a; Wiswall, 2005) is fine-grained, silvery to grayish-green, muscovite-chlorite phyllitic schist occurring in a narrow band adjacent to the trace of the Embreeville Thrust.

The geochemistry of amphibolites in the WGMS, the White Clay Creek Amphibolite (WCCA) and Kennett Square Amphibolite (KSA), resemble modern basalt. Smith and Barnes (2004) suggest that the WCCA resembles continental initial rift basalt and correlate the WCCA with the Catoctin metabasalt. The KSA is very similar to modern mid-ocean ridge basalt (MORB) (Smith and Barnes, 1996, 2004; Plank et al., 2001). The cover sequence including the Glenarm Group and WGMS is interpreted to have been deposited along the rifting Laurentian margin in the latest Neoproterozoic to Cambrian (Blackmer 2004c, 2005; Bosbyshell et al., 2014).

Arc-related rocks: the Wilmington Complex and Wissahickon Formation

The Wilmington Complex has been studied for well over a hundred years (Chester, 1890; Bascom et al., 1909), and was defined by Ward (1959) to include the igneous and metamorphic rocks across northern Delaware from Cecil County, Maryland, to Chester, Pennsylvania. Detailed geologic mapping (Plank et al., 2000; Schenck et al., 2000), geochemical analyses of mafic units (Plank et al., 2001), and ages determined by Sensitive High-Resolution Ion Microprobe (SHRIMP) U-Pb isotopic analysis of zircons from felsic gneisses (Aleinikoff et al., 2006), were used to re-define the lithodemic units of the Wilmington Complex and constrain the age and tectonic affinity. Geochemically, metavolcanic and metaplutonic units resemble modern subduction related igneous rock (Plank et al., 2001). Zircon result demonstrate that they are early Ordovician in age, 475 to 485 Ma; the relatively undeformed Arden pluton is Silurian, 434 ±5 Ma (Aleinikoff et al., 2006).

The Wissahickon Formation consists of fine- to coarse-grained pelitic schist interlayered with psammitic granofels and quartzite. Pelitic schist is composed of quartz, plagioclase, biotite, muscovite, garnet, and local sillimanite, kyanite, and/or staurolite. Psammitic layers are composed of quartz, plagioclase, biotite, and muscovite with rare garnet. Also present are thin layers of granofels, composed of plagioclase, quartz, hornblende and epidote, with or without garnet and/or dolomite. These may have originated as marl or other calcsilicate. This interlayered lithology alternates at the map scale with massive to layered, medium- to coarse-grained psammitic to semipelitic schist. Throughout the Wissahickon Formation, metasedimentary rock is interlayered with amphibolite, up to 20 m thick. The Wissahickon Formation contains detrital zircon as young as 475 Ma (Bosbyshell et al., 2012; 2014) and is host to the 427 ± 3 Ma Springfield granodiorite (Bosbyshell et al., 2005). Thus, the depositional age of the Wissahickon Formation is most likely middle to late Ordovician.

Structural Geology

Here we present an overview of the deformation history of rocks. Different generations of foliation (S) and folding (F) are number sequentially, but these are not correlated across the Rosemont Shear Zone. A “W” or an “E” is appended to the subscript (e.g., S_{2W}) to indicate west (within the WGMS) or east (Wissahickon Formation) of the RMZ.

We adopt a structural framework that is based on relatively recent mapping by the Pennsylvania and Delaware geological surveys (Schenck et al., 2000; Blackmer, 2004a, 2004b, 2005; Wiswall, 2005;

Blackmer et al., 2010). To the west of the RSZ, the Embreeville Thrust (Fig. 1) is the lowest structure in a series of nappes composed of basement gneiss and metasedimentary cover. From structurally lowest to highest, these include the West Chester nappe, Avondale nappe, and Mill Creek (Hockessin-Yorklyn) anticline (Schenck et al., 2000). The dominant, S_{2W}, foliation in this area dips shallowly to moderately to the southeast (Blackmer 2004a, 2004b, Wiswall, 2005), and is axial planar to overturned to recumbent outcrop scale folds, which exhibit top to the northwest asymmetry (Alcock, 1994; Blackmer, 2004a). This foliation is also parallel to thrust-sense shear zones at the base of the nappes (Bosbyshell et al., 2006a). The S_{2W} foliation, therefore, likely formed as a result of thrust emplacement. The S_{2W} foliation is deformed by upright folds (Fig. 2), especially in the northwestern- and southeastern-most rocks (Alcock, 1994; Blackmer, 2004a; Wiswall, 2005). This upright folding is attributed to younger, transpressive deformation in the Pleasant Grove-Huntingdon Valley and Rosemont shear zones (Valentino et al., 1994, 1995).

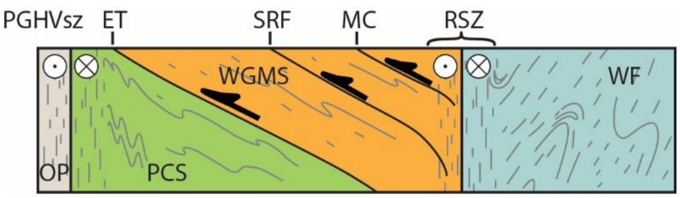


Figure 2. Schematic cross section of field trip area. PGHVsz = Pleasant Grove/Huntingdon Valley shear zone; ET = Embreeville Thrust; SRF = Street Road Fault; MC = base of Mill Creek nappe; OP = Octoraro Phyllite; PCS = Peters Creek Schist; WGMS = West Grove Metamorphic Suite; WF = Wissahickon Fm. Basement gneiss and Wilmington Complex not shown. Modified from Bosbyshell et al. (2016).

The nappes and associated southeast dipping fabrics are truncated to the southeast by the steeply dipping RSZ (Valentino et al., 1995; Bosbyshell 2005a, 2005b), the western boundary of the arc terrane, which includes the Wilmington Complex, Wissahickon Formation, and Chester Park gneiss (Figs. 1,2). Detailed structural analyses in the Wissahickon Formation (Amenta, 1974; Tearpock and Bischke, 1980; Bosbyshell 2001, 2008) describe similar deformational histories, involving five recognizable stages. Different generations of structures are preserved to varying degrees at the map or even outcrop scale, depending on local metamorphic history (Amenta, 1974; Bosbyshell, 2001). The oldest deformation is recognized as an early foliation present in the hinges of S_{2E} folds and as transposed F_{1E} hinges rarely preserved within the S_{2E} schistosity. The regional schistosity, the S_{2E} foliation, is axial planar to F_{2E} isoclinal folds and generally dips moderately to steeply to the northwest. S_{2E} and F_{2E} folds are in turn folded by F_{3E}, close to tight, upright to recumbent folds associated with a variably developed sub-vertical to moderately northwest-dipping axial planar foliation. S_{4W} fabrics are associated with the RSZ and cross cut F_{3E} folds (Amenta, 1974; Bosbyshell, 2001). The youngest ductile fabrics include sub-horizontal crenulation (S_{5E}) and associated outcrop scale open folds which are variably developed throughout the area (Amenta, 1974; Tearpock and Bischke, 1980; Valentino and Gates, 2001; Bosbyshell, 2008).

Gneissic fabrics in metaigneous rock of the Wilmington Complex are sub-vertical to steeply northwest dipping along the northwest margin of the Complex, approaching the RSZ, and dip moderately to the northwest elsewhere (Schenck et al., 2000). The pattern is similar to that in the Wissahickon

A note on terminology: Nappe is a structural geology term equivalent to thrust sheet. Historically, the West Chester and Avondale nappes have also been referred to as massifs. Here we use the term massif to refer to the body of gneiss within the thrust sheet. Massif has many definitions, but usually refers in some way to a fault-bounded mountain. The gneissic lithologies in the Piedmont tend to be more resistant to weathering than metasedimentary cover rocks and tend underlie higher elevations, hence our usage (though clearly they are not mountains!).

Formation to the northeast of the Wilmington Complex and contrasts with the shallow to moderate southeast dips in rocks to the northwest throughout the WGMS.

COMPARISON OF WEST GROVE METAMORPHIC SUITE AND WISSAHICKON FORMATION

Bosbyshell et al. (2014) provide a detailed comparison of the WGMS and Wissahickon Formation in support of separating the WGMS as a distinct unit. They describe distinct differences in detrital zircon population, amphibolite geochemistry, and metamorphic pressure-temperature-deformation-time (P-T-D-t) paths between these units. The P-T-D-t histories are further detailed by Bosbyshell et al. (2016). These differences are summarized below.

Detrital zircon geochronology

The results of recent detrital zircon analysis (Bosbyshell et al., 2012, 2014, 2015) demonstrate significant differences in the source regions for metasedimentary rocks of the WGMS and Wissahickon Formation. Most samples in the Doe Run Schist and Mt. Cuba Gneiss exhibit well-defined peaks at 960 Ma and 1020–1050 Ma, an array of smaller Mesoproterozoic peaks and latest Neoproterozoic peaks at 550 Ma (Fig. 3). Doe Run samples contain Archean zircon, which is absent in the Mt. Cuba. The youngest zircon in the Doe Run Schist yielded a concordant age of 528 Ma. The youngest zircon in the Mt. Cuba gneiss is 544 Ma. Thus, the WGMS can be no older than Cambrian, and the depositional age of the Doe Run Schist may be younger than that of the Mt. Cuba Gneiss.

Detrital zircon populations in samples from the Wissahickon Formation east of the Wilmington Complex (Fig. 1) contain Mesoproterozoic zircon, but the distribution of peaks is considerably different from those in the WGMS (Fig. 3). Two samples from the Wissahickon Formation, including one from a location intruded by arc-related magmatic rock near the contact with the Wilmington Complex, contain sizable populations of zircon with late Neoproterozoic ages which likely indicate a Gondwanan source. Most other Wissahickon samples, including those in a belt along the western side of the Wilmington Complex, yield zircon of this age although it is much less abundant. A meta-volcaniclastic unit adjacent to the Wilmington Complex also contains a small number of zircon grains of this age. These results suggest that the Wilmington Complex and the Wissahickon Formation may not have a peri-Laurentian origin, or that a peri-Gondwanan source was proximal to Laurentia in the early Ordovician when the Wilmington Complex arc was active.

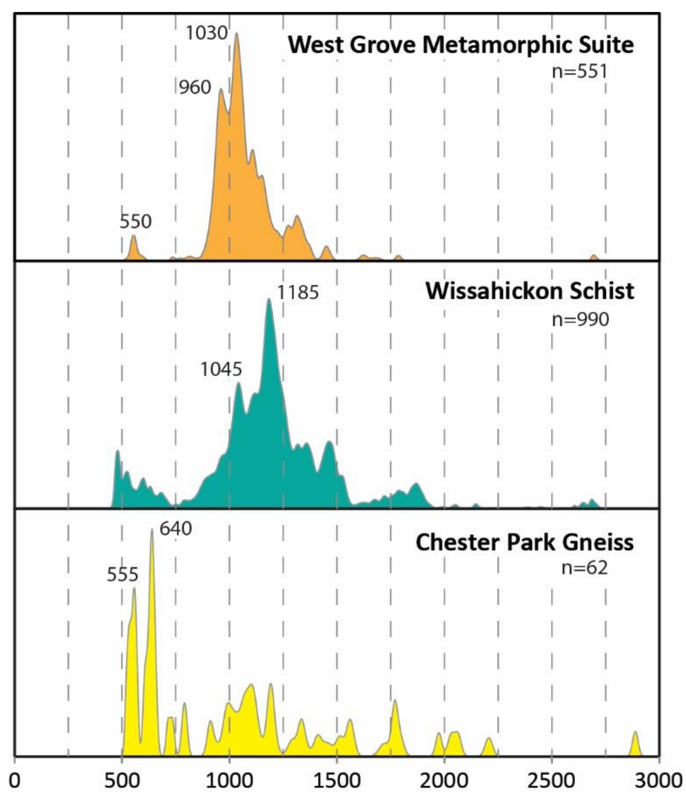


Figure 3. Detrital zircon results. The Wissahickon Schist results include the Windy Hills Gneiss (Aleinikoff et al., 2006), a volcanic clastic unit along the western margin of the Wilmington Complex, and two additional metaclastic samples within the Wilmington Complex. Note that the Chester Park Gneiss results from a single sample.

The detrital zircon population of the Chester Park Gneiss is dominated by

Neoproterozoic ages (peaks at 555 Ma and 640 Ma), that are likely derived from a peri-Gondwanan source, and a small number of grains of Mesoproterozoic, Paleoproterozoic and Archean age. Voluminous Neoproterozoic arc magmatism of is well known in peri-Gondwanan terranes (Nance et al., 2008) and similar detrital zircon signatures are present in sedimentary and metasedimentary rocks derived from these terranes in New England (Macdonald et al., 2014; Connard et al., 2015) and the Canadian maritime provinces (Fyffe et al., 2009). The Chester Park gneiss detrital zircon population is very similar to that of the Moretown terrane in New England (Macdonald et al., 2014).

Cawood et al. (2012) examined the detrital zircon signature of rocks deposited in different tectonic settings. They suggest that in rocks deposited in an extensional tectonic environment, the age of fewer than five percent of detrital zircons is within 150 Ma of the depositional age of the rock (cumulative probability curves will pass below the blue star in Fig. 4). In contrast, rocks deposited in a convergent tectonic setting contain a significant proportion of zircon grains (>30%) with ages within 100 million years of the depositional age of the rock (curves pass to the left of the red star). A "collisional" setting is inferred for rocks between these endmembers. Following this analysis (Fig. 4), an extensional depositional environment is inferred for the West Grove Metamorphic Suite, while Wissahickon Schist samples plot mainly in the collisional field with two samples nearly meeting the convergent criteria. The Chester Park Gneiss was likely deposited in a convergent tectonic setting.

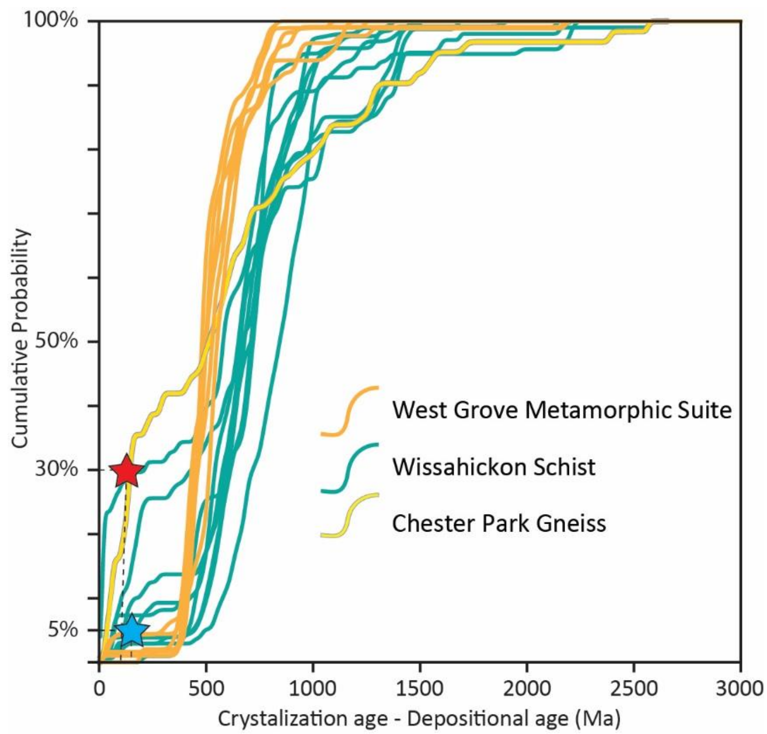


Figure 4. Cumulative probability plot of the difference between zircon crystallization ages and the depositional age of study area samples, after Cawood et al. (2012). The plot was prepared using estimated depositional ages of 520 Ma for the WGMS, 500 Ma for the Chester Park Gneiss and 480 Ma (or the age of the youngest zircon, whichever is younger) for the Wissahickon.

Amphibolite geochemistry

The West Grove Metamorphic Suite includes the Kennett Square Amphibolite (KSA) and White Clay Creek Amphibolite (WCCA). For detailed description and analysis of these units, the reader is referred to Smith and Barnes (1994, 2004) and Plank et al. (2001). The Kennett Square Amphibolite occurs within the Mt. Cuba Gneiss as map-scale bodies up to 10 km long in a belt along the southern portion of the Avondale nappe (Fig. 1). The KSA is characterized by flat REE patterns with slight LREE depletion or enrichment that is characteristic of mid-ocean ridge basalt (MORB) (Smith and Barnes, 1994; Plank et al., 2001).

The White Clay Creek Amphibolite is interlayered within the Mt. Cuba Gneiss and less commonly within Doe Run Schist (Blackmer, 2004b). WCCA is characterized by high total Fe, moderate to high TiO2, and high overall REE abundances. The REE patterns are characteristic of within-plate basalts from either a continental or oceanic setting. Plank et al. (2001) favored an oceanic setting, while Smith and Barnes (2004) proposed a continental initial rift environment and correlate the WCCA with the Catoctin metabasalt.

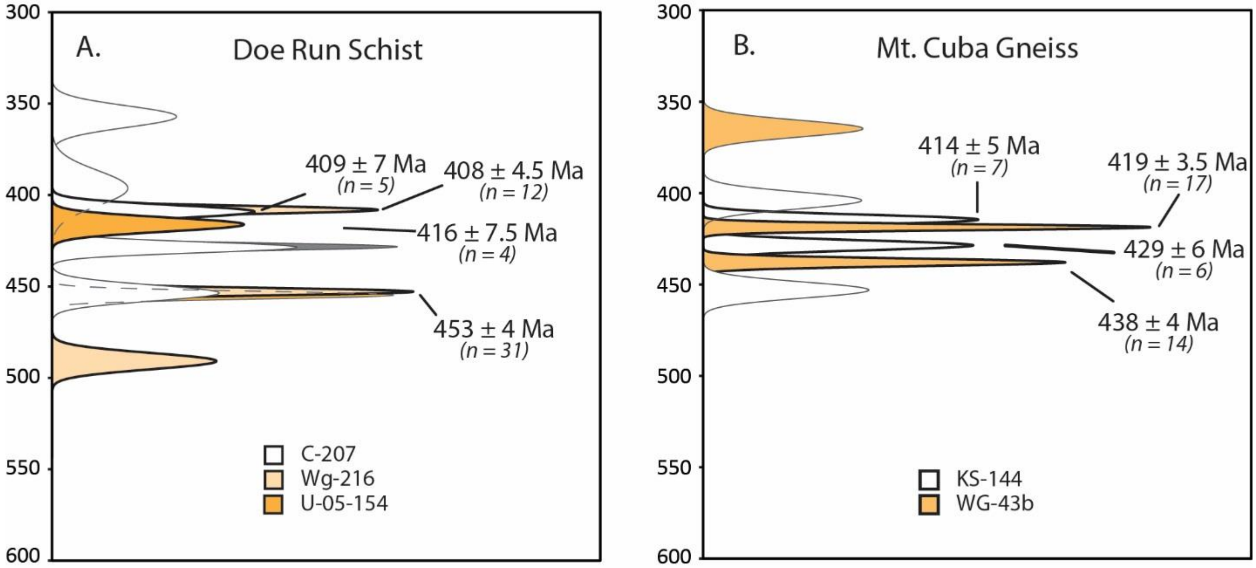
Three geochemically distinct amphibolites – the Confluence Dikes, Bridgewater Amphibolite, and Smedley Park Amphibolite – occur within the Wissahickon Formation in different geographic areas and possibly stratigraphic position (Bosbyshell, 2001; Bosbyshell et al., 2014, 2015). These amphibolites are described in detail by Bosbyshell et al. (2105). The Confluence Dikes occur in both the Wilmington Complex and Wissahickon Formation, near the contact between these units in the southern Media 7.5 minute quadrangle (Stop 3). The Bridgewater Amphibolite is found within and near the Chester Park gneiss (Bosbyshell, 2001, 2005a, 2005b) in a northeast-southwest trending belt along the Coastal Plain onlap. The Smedley Park Amphibolite is found within the Wissahickon Formation to the northwest of the Bridgewater Amphibolite.

The Confluence Dikes are nearly identical to mafic layers in the Rockford Park Gneiss of the Wilmington Complex (Bosbyshell et al., 2015), and both are very similar to modern boninities, distinctive volcanic rocks found in the forearc regions of modern western Pacific island arcs (Hickey and Frey, 1982; Bloomer, 1987; Stern et al., 1991). The Bridgewater Amphibolite occurs as 1 to 5 m thick layers that are concordant with foliation and compositional layering in surrounding rock. In many outcrops, the Bridgewater Amphibolite is distinctive for coarse (up to 0.5 cm diameter) magnetite grains that are surrounded by a halo of plagioclase. The chemical composition is similar to highly-evolved Fe-Ti basalts and more closely resembles within-plate basalts than MORB or arc tholeiites. The Smedley Park Amphibolite occurs in layers thicker greater than 20 m in many locations (Bosbyshell, 2001, 2004). Geochemically the Smedley Park Amphibolite is very similar to modern back-arc basin basalt (Bosbyshell et al., 2015).

Pressure-temperature history

The metamorphic and deformation histories of the WGMS and the Wissahickon Formation are distinct from each other, and within the WGMS, that of the Doe Run Schist differs somewhat from Mt. Cuba Gneiss. The Mt. Cuba Gneiss attained higher temperatures than the Doe Run Schist (Plank, 1989; Alcock and Wagner, 1995; Bosbyshell et al., 2016). The Doe Run contains evidence for an older period of metamorphism and widely-dispersed, locally-intense retrograde overprinting by hydrous fluids (Moore, 2007), that are absent from the Mt. Cuba.

The older period of metamorphism in the Doe Run Schist is present as an early fabric defined by aligned sillimanite preserved in microlithons within the dominant foliation. This fabric may be late Ordovician based on monazite core ages of ~450 Ma (Bosbyshell et al., 2016), but no direct textural evidence links the monazite cores to older fabrics. The dominant foliation, S_{2W}, wraps around garnet and staurolite porphyroblasts, but staurolite also occurs parallel to this foliation. In several instances staurolite is rimmed by fine sillimanite and small euhedral garnet which cross-cut foliation. These observations suggest that peak metamorphic conditions in the Doe Run Schist exceeded the high-T stability limit of staurolite and that peak temperatures were attained subsequent to formation of S_{2w} foliation. Bosbyshell et al. (2016) utilized garnet isopleth thermobarometry to estimate peak conditions of approximately 700 ± 50°C at 550 ± 100 MPa in a sample in the footwall of the Street Road Fault (WG-216, Fig. 1). Monazite geochronology demonstrates that maximum temperatures were attained at ~410 Ma (Fig. 5). Infiltration of hydrous fluids caused variable retrogression of higher-temperature assemblages, ranging from partial rimming and replacement of porphyroblasts by fine-grained muscovite or chlorite to complete



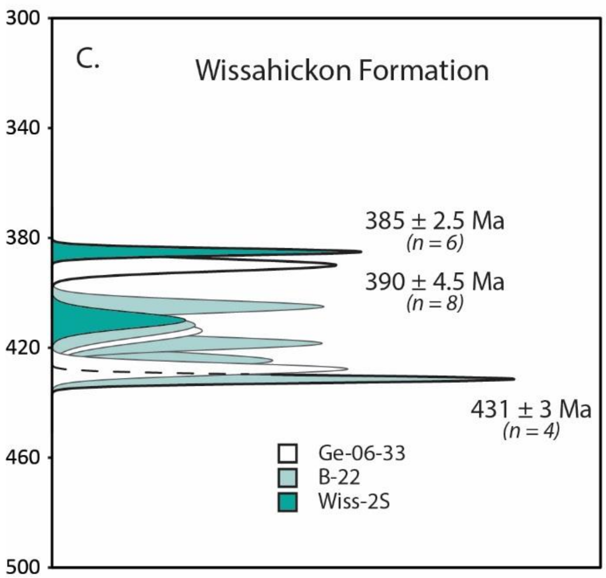


Figure 5. Summary of monazite results. Histograms show a Gaussian distribution calculated using the weighted mean and standard distribution of U-Th-total Pb age calculations. Distributions that are labeled and outlined in bold lines are the weighted mean of *n* compositional domains; curves without the bold outline are the results of individual compositional domains that are not used in the weighted averages. For sample DR-2, n is the number of individual spot analyses used. In all other samples, n is the number of dated domains used in the average age calculation. Data are from Bosbyshell (2008) and Bosbyshell et al. (2016).

replacement and recrystallization of foliated matrix biotite by mimetic chlorite. Retrograde metamorphism is most common near the Embreeville Fault (Moore et al., 2007).

In the Mt. Cuba Gneiss, Plank (1989) found that peak metamorphic conditions varied from 600 ± 50°C at 550 ± 100 MPa in southeastern-most Pennsylvania to 750 ± 50°C at 650 ± 100 MPa nearest the Wilmington Complex in Delaware; estimates which are supported by subsequent work (Alcock, 1989, 1994; Alcock and Wagner, 1995; Bosbyshell et al., 2016). In high-grade, migmatitic Mt. Cuba Gneiss, some leucosome contains biotite fish parallel to mesosome foliation while other leucosome contains randomly-oriented biotite. These high-temperature microstructures suggest syn- to post-kinematic partial melting with respect to S_{2w} foliation formation. The timing of maximum temperatures is constrained by monazite ages in two samples: monazite in the Mill Creek nappe (sample 44069) gives a Silurian age of 425 Ma (Aleinikoff et al., 2006); an age of 415 Ma for monazite inclusions in staurolite from the Avondale nappe provide a maximum age for peak metamorphism there (Fig. 5). Alcock (1989) describes evidence for higher pressure metamorphism following peak temperature in the Mt. Cuba, inferred to reflect crustal thickening following thrust stacking.

Two periods of metamorphism are well-documented in the Wissahickon Formation east of the Rosemont shear zone: early, high temperature, low- to moderate-pressure assemblages are variably overprinted by a second period of higher pressure metamorphism at 650 ± 50°C, 700 ± 100 MPa (Crawford and Mark, 1982; Bosbyshell et al., 1999; Bosbyshell, 2001). The temperature associated with the early metamorphism varies from west to east. Nearest the Wilmington Complex, peak conditions were likely in excess of 700°C at 500 ± 100 MPa while less than 10 km to the east, andalusite was part of the early assemblage, implying temperatures of approximately 500°C (Bosbyshell et al, 1999). In the type section of the Wissahickon Formation, the early metamorphism is present only near Silurian intrusions and the Barrovian-style moderate pressure metamorphism evident there reflects the second period of metamorphism (Bosbyshell et al., 2016). Monazite ages (Bosbyshell, 2001; Pyle et al., 2006; Bosbyshell, 2008; Bosbyshell et al., 2016) constrain the early metamorphism to the Silurian (~430 Ma) and the high-pressure overprint to the Devonian (~390 Ma) (Fig. 5). Metamorphism in the Wissahickon Formation is further described below (Stop 1.5).

TECTONIC IMPLICATIONS

Tectonism in the field trip area has long been interpreted to be the result of the Taconic Orogeny: the Ordovician collision between a peri-Laurentian volcanic/magmatic arc and the Laurentian margin (Wagner and Srogi, 1987; Aleinikoff et al., 2006; Wise and Ganis, 2009; Sinha et al., 2012). Constraints on the timing of metamorphism described above require revision of this model. Bosbyshell et al. (2016) propose that Silurian to earliest Devonian metamorphism in the WGMS is the result of the accretion of a peri-Gondwana terrane, Ganderia (Fig. 6) in the dominantly sinistral transpressive tectonic regime thought to be present at that time (Hibbard, 2000; Hibbard et al., 2007; 2010). Hibbard et al. (2007) suggest that the New York promontory acted as a restraining bend in this transpressive setting and was

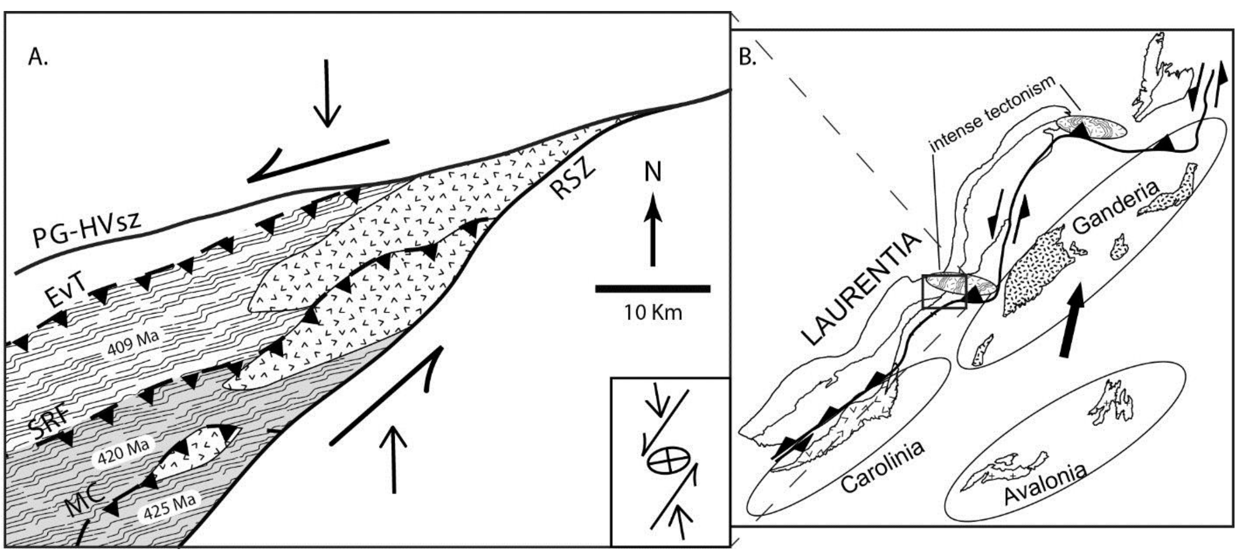


Figure 6. Schematic geologic map illustrating that the geometry thrust sheets in relation to the vertical shear zones is consistent with a sinistral transpressive regime. The age of metamorphism, with peak temperatures attained in the structurally lowest block subsequent to those in higher sheets, is interpreted to be the result of successive stacking of thrust sheets from southeast to northwest with the warmer overriding sheets contributing to the thermal budget of lower sheets. (b) Plate tectonic reconstruction for the Silurian modified from Hibbard et al. (2007) illustrating approach of Ganderia in sinistral transpression. Rectangle shows possible location of deformation shown in A. PGHVsz = Pleasant Grove and Huntingdon Valley shear zone; EvT = Embreeville Thrust; SRF = Street Road Fault; MC = unnamed fault below Mill Creek anticline; RSZ = Rosemont shear zone. From Bosbyshell et al. 2016.

the site of intense tectonism at this time. Bosbyshell et al note that in the Mill Creek anticline, the structurally highest nappe, peak metamorphism is older (425 Ma) than in the lower Avondale nappe (415 Ma). In turn, metamorphism in rock at the lowest structural level, the Doe Run Schist in the West Chester nappe, is even younger (409 Ma). They interpret this sequence to represent successive stacking of thrust sheets from southeast to northwest (present geography) with the warmer overriding sheets contributing to heating of the lower sheets. They suggest that the geometry of the thrust slices containing the Doe Run Schist and Mt. Cuba gneiss in relation to the steeply dipping shear Pleasant Grove-Huntingdon Valley shear zone is consistent with the sinistral restraining bend proposed by Hibbard et al. (2007).

The middle Devonian age of monazite suggests that metamorphism in the Wissahickon Formation is the result of crustal thickening during the Acadian orogeny, the accretion of Avalon in the northern Appalachians. Previously, the effects of the Acadian orogeny in southeastern Pennsylvania were the subject of reasoned inference (Amenta, 1974; Valentino et al., 1995) or were considered to be absent (Faill, 1997). Metamorphism of this age is well known in southern New England (Lanzirotti and Hanson, 1996; Robinson et al., 1998; Lancaster et al., 2008). Given the evidence for younger, dextral transcurrent motion regionally on the Pleasant Grove/Huntington Valley and Rosemont shear zones (Valentino et al., 1994; 1995) and throughout the Appalachians (e.g., Dennis, 2007; Hibbard and Waldron, 2009) Bosbyshell et al. (2015, 2016) proposed that the crustal block east of the Rosemont shear zone, which contains the Wissahickon Formation and Wilmington Complex, was originally located some distance to the north.

In the northern Appalachians, the Taconic Orogeny has also been modeled as the result of arc-continent collision involving a peri-Laurentian volcanic/magmatic arc (e.g. Stanley and Ratcliffe, 1985). Recent U-Pb detrital zircon geochronology (Macdonald et al., 2014) demonstrates that the Moretown terrane, upon which the bulk of the Taconic Shelburne Falls arc (Karabinos et al., 1998) is built, is Gondwana-derived. In a reevaluation of the Taconic orogeny, Macdonald et al. (2014) suggest that the Moretown terrane was the first of several rifted Gondwanan terranes – analogous to, but distinct from, Carolinia in the southern Appalachians and Ganderia in the northern Appalachians – to collide with the Laurentian margin. Metasedimentary rock structurally below the Moretown terrane, the Rowe belt, are Laurentian-derived, while rock above the Moretown, the Hawley belt (including the Cram Hill formation in Vermont), contains zircon from both Laurentian and Gondwanan sources (MacDonald et al., 2014; Connard et al., 2015) and is also host to arc-related metaplutonic and metavolcanic rocks (Kim and Jacobi, 1996).

There are many similarities between the rocks described here and arc-related rocks in New England. In the central Appalachian Piedmont, the Wilmington Complex at 476 to 483 Ma (Aleinikoff et al., 2006) is the same age as the Shelburne Falls arc, where the main magmatic pulse is 475 Ma (MacDonald et al., 2014). The Wissahickon Formation, like the Hawley belt in New England, is host to boninitic amphibolites (Kim and Jacobi, 1996); geochemical characteristics of the Smedley Park amphibolite are very similar to the Group IV, BABB-like amphibolite of Kim and Jacobi (1996). Detrital zircon age spectra of the Wissahickon Formation have similar peaks to the Hawley belt, including Gondwanan peaks (Connard et al., 2015) and Grenville-aged peaks that are distinctly different from those in adjacent Laurentian rift-related metasediments (the WGMS in the Central Appalachians, the Rowe belt in New England). The distinct lithological character of the Chester Park gneiss is very similar to the "pinstripe" granofels of the Moretown terrane (MacDonald et al., 2014). The detrital zircon age spectra obtained from the Chester Park gneiss is very similar to the Moretown terrane. As in the Moretown Terrane, deformed tonalitic rock is present in the Chester Park gneiss.

Bosbyshell et al. (2015) proposed that the arc-related rocks described above may have originally been part of the Taconic arc in New England that were translated by strike-slip deformation to their present location. In the Piedmont, middle to late Paleozoic transpressive deformation with a significant strike-slip component is associated with the Pleasant Grove – Huntingdon Valley shear zone (PGHV) (Valentino et al. 1994, 1995) and the RSZ (Valentino et al., 1995; Bosbyshell, 2001). As much as 150 km of dextral displacement has been proposed for the PGHV, based in part on similarities between rock in Manhattan, N.Y. and Baltimore, Md. (Valentino et al. 1994). A Devonian age (384 ± 6 Ma) was determined for syn-tectonic monazite in the RSZ (Bosbyshell, 2001) and fabrics related to deformation in the RSZ crosscut Devonian-aged (390 ± 4.5 Ma) mineral assemblages at stop one (Bosbyshell et al., 2016). These results suggest that deformation in the Rosemont is associated with the Acadian orogeny. Final emplacement is likely the result of younger movement in the PGHV shear which was active into the Pennsylvanian-Permian Alleghenian orogeny (Valentino and Gates, 2001; Blackmer et al., 2007).

FIELD TRIP OVERVIEW

This year's trip visits five locations which highlight recent research and new insights into the tectonic evolution of rock underlying Philadelphia, Delaware and Chester counties of southeastern Pennsylvania. At stop one we will examine the Wissahickon Schist, associated ultramafic rocks and highly sheared rock at the margin of the West Chester massif. Detrital zircon data reveal that the Wissahickon Formation may be derived in part from sediment eroded from a peri-Gondwanan terrane (Bosbyshell et al., 2015). The main period of metamorphism is now known to be Devonian (~390 Ma) in age (Bosbyshell et al., 2016). Kerrigan et al. (this volume) suggest that the ultramafic rock is most likely a differentiate derived from an arc-related magma system, supporting the inference that the depositional setting of the Wissahickon Formation is a basin adjacent to a volcanic arc. Stop two will afford us the opportunity to view Laurentian basement gneiss of the Avondale nappe. Recent research here has focused on an outcrop along the eastern edge of the nappe, adjacent to the Rosemont Shear Zone, which contains evidence for possible ultra-high temperature or pressure metamorphism (Trice et al., 2014; Noble et al., 2015). The Wilmington Complex, the remains of a volcanic/magmatic arc that formed along the Laurentian margin in the Ordovician will be seen at stop three before visiting the Chester Park Gneiss at stop four. Detrital zircon from this unit reveals that the sediment source is peri-Gondwanan (Bosbyshell et al., 2015) and that these rocks may be correlative with, or even a fragment of, the Moretown Terrane in New England (MacDonald et al., 2104). At stop five, we return to the West Chester area and Laurentia. Here we will view the Doe Run Schist and Laurels Schist, which are part of the early Paleozoic metasedimentary cover sequence, formerly known as "Glenarm Wissahickon," now the West Grove Metamorphic Suite (Bosbyshell et al., 2014).

ROAD LOG

The trip departs from M-lot, West Chester University.

Mileage	*Cum. mileage*	*Description*
0.0	0.0	Exit M-lot and turn right onto S. Matlack St.
0.6	0.6	Turn left at light onto US-202 N
17.1	17.1	Follow signs for I-76 E
18.6	18.6	Merge on I-76 (Schuylkill Expressway)
22.2	22.2	Take exit 331B to merge onto I-476N (toward Plymouth Meeting)
25.0	25.0	Take exit 18A toward Conshohocken
25.4	25.4	Merge onto Ridge Pike
29.1	29.1	Turn left onto Bells Mill Rd. (after Northwestern Ave and Ayrdale Rd)
30.0	30.1	Parking lot on right.

STOP ONE: BELLS MILL ROAD ULTRAMAFIC BODY AND WISSAHICKON SCHIST

This stop is within Wissahickon Valley Park, a 1,800 acre park in Philadelphia, PA that follows Wissahickon Creek approximately 1 km north of this location and 10 km south to the confluence of Wissahickon Creek and the Schuylkill River. The area offers excellent exposure of the Wissahickon Schist and Bells Mill Road ultramafic body. Exposures along Wissahickon Creek comprise the type section of the Wissahickon Formation, first described by Bascom (1905). Crawford (1987) provides details of the traverse from just north of our present location south to the mouth of Wissahickon Creek.

The walking/biking path along Wissahickon Creek, known as "Forbidden Drive (Fig. 7)," offers several outcrops that characterize that lithologies present within the ultramafic body and Wissahickon Schist. This stop is an excellent location for teaching exercises in igneous and metamorphic petrology due to the presence of the ultramafic body surrounded by the Wissahickon Schist and a small quarry within a granodioritic body where tectonized contacts with schist are visible. The northern most exposures along the creek provide a view of Mesoproterozoic gneiss which exhibits a mylonitic texture, evidence of intense deformation in the Rosemont Shear Zone (RSZ).

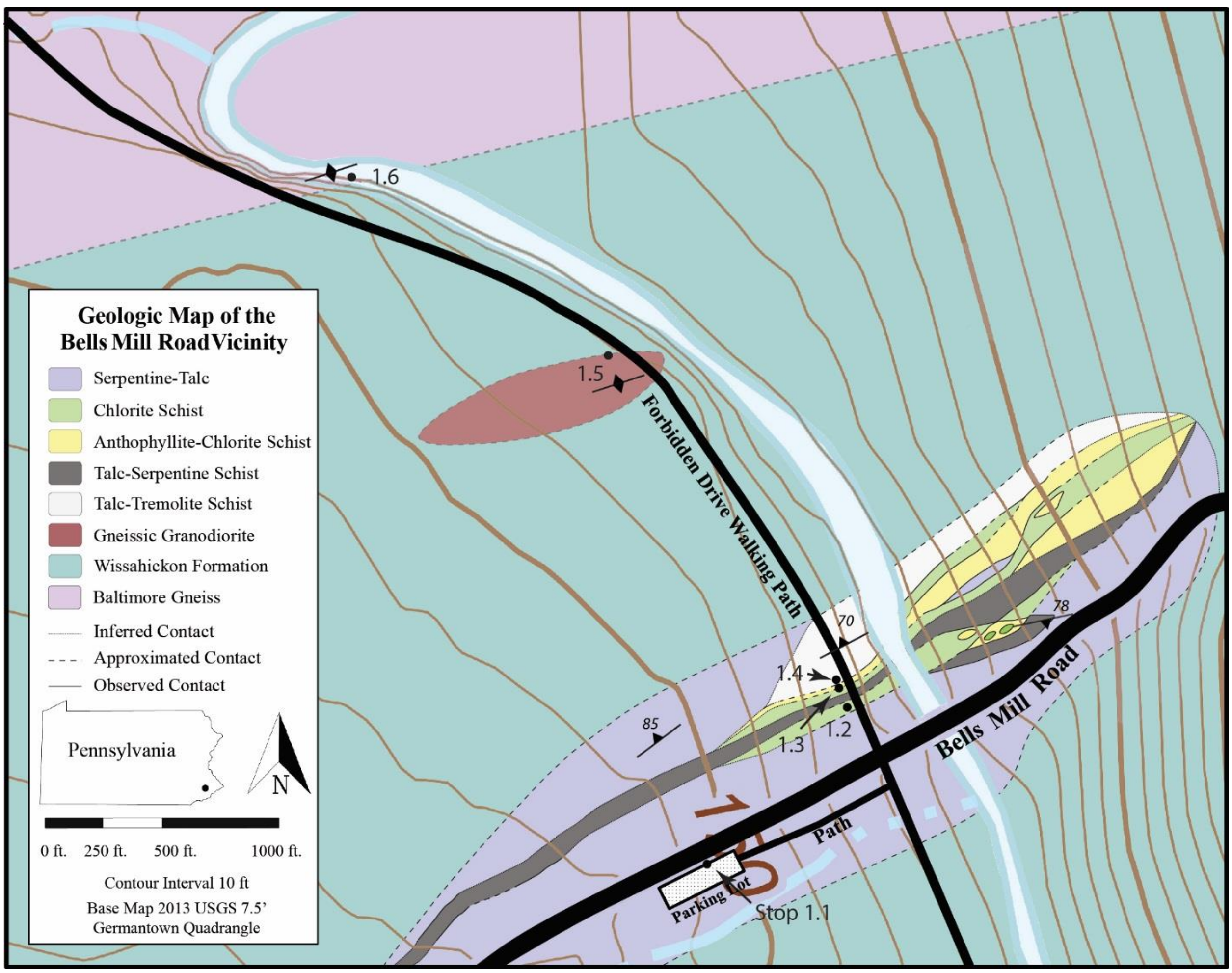


Figure 7: Geologic map of the Bells Mill Road ultramafic body showing the field trip stops. The lithologic zones of alteration are defined as: serpentine-talc, talc-tremolite, anthophyllite-chlorite schist, chlorite schist, and talc-serpentine schist (modified after Simboli et al., 2017)

STOP 1.1 - Serpentine Talc Rock

In the parking lot there is a large boulder of a serpentine-talc rock observed which encompasses the majority of the ultramafic body. The serpentine-talc rock is the largest of all lithologic units exposed at the Bells Mill Road ultramafic body. The rock has oblong, sub-rounded inclusions of serpentine surrounded by a talc-anthophyllite-magnesite matrix. The rock is matrix-supported and throughout the rock there are scattered oxides, mostly magnetite and chromite.

Walking from STOP 1.1 to STOP 1.2 follow a path leading from the parking lot to the main walking path of the Wissahickon Valley Park, Forbidden Drive. Along the short path from the parking lot there are numerous boulders scattered on the side of the path which show some of the diversity of the serpentine-talc rock. Some inclusions can reach up to 10 cm in diameter. Further discussion of the petrology and geochemistry each of these lithologies is contained in the accompanying paper in this volume.

STOP 1.2 - Chlorite Schist

Walking approximately 20 meters northwest from Bells Mill Road along the walking path, Forbidden Drive, there is an outcrop of the chlorite schist. The chlorite schist at this location is approximately 20 meters thick. The rock is dominated by chlorite (~90%) with scattered talc, anthophyllite, tremolite, magnesite, and magnetite. The northwestern end of the unit exhibit abundant magnetite up to 1 cm in diameter and some with nice octahedral crystal habit. Within this unit there are locations where the density of magnetites is so high that Brunton compasses are often thrown out of alignment by the magnetic pull of the minerals.

STOP 1.3 - Anthophyllite-Chlorite Schist

Approximately 20 meters northwest from STOP 1.2 there is an outcrop of the anthophyllite-chlorite schist. The anthophyllite-chlorite schist at this location is approximately 5 meters thick. Point counting analysis of the anthophyllite-chlorite schist reveals approximately 42% anthophyllite, 34% chlorite, 17% talc, 5% orthopyroxene, 1% magnesite, and 1% opaques. The orthopyroxene are observable only in thin section and may be relic from the protolith. Anthophyllite exhibits a bladed to acicular texture sometimes radially orientated.

STOP 1.4 - Talc-Tremolite Schist

Approximately 20 meters northwest from STOP 1.3 along the walking path, there is an outcrop of the talc-tremolite schist. The talc-tremolite schist at this location is approximately 20 meters thick and appears to pinch-out to the southwest and increase in width to the northeast. The rock is massive and mainly talc (~49%) with tremolite (27%), anthophyllite (15%), serpentine (6%) and orthopyroxene (3%). On fresh surfaces stringers of tremolite are visible helping to define the foliation. The outcrop is dense rock with wavy foliation.

Continue north along the trail, examining outcrops of Wissahickon Schist. A small abandoned quarry is found approximately 500 meters north of STOP 1.4.

STOP 1.5 - Small quarry exposing granodioritic gneiss and Wissahickon Schist

In an investigation of metamorphism along the entire Wissahickon Creek transect, Bosbyshell et al. (2007; 2016) prepared x-ray composition maps of garnet from 12 samples. With the exception of one sample, from this quarry at the contact with the granodiorite, all garnet along the transect shows relatively simple zoning indicative of prograde metamorphism. Garnet from this outcrop exhibits distinct small very low-Ca, high-Mn cores indicative of two stages of metamorphism (Fig. 8). Pelitic schist here contains the

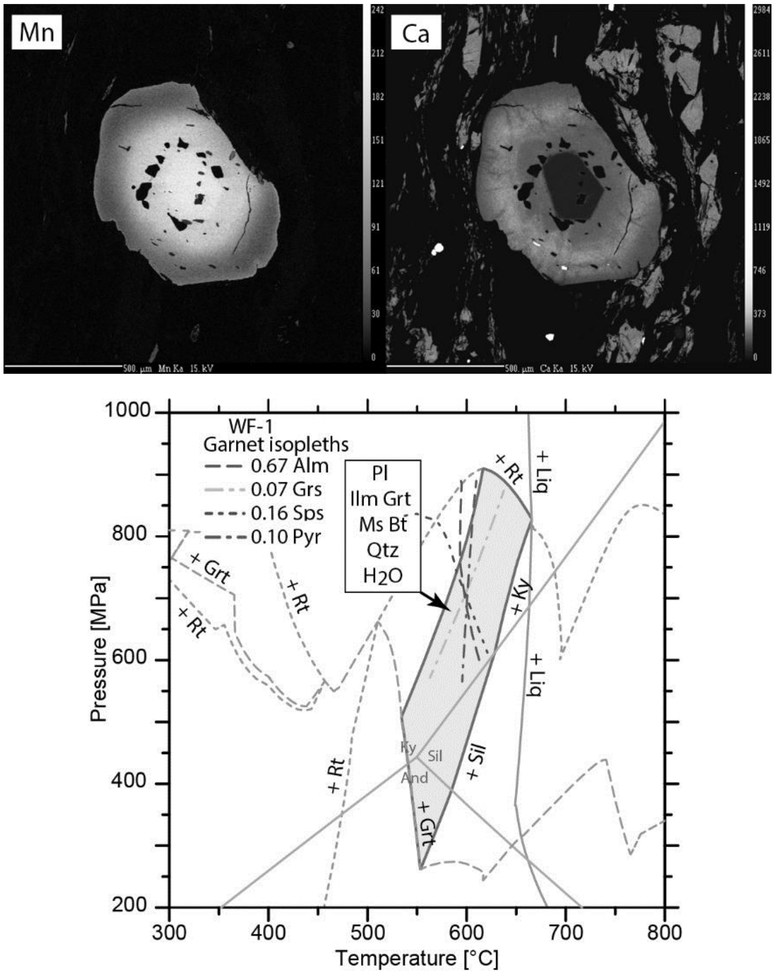


Figure 8. (a) X-ray composition maps of garnet from Stop 1.5, note well defined high-Mn, low-Ca core, which we infer is the result of contact metamorphism at the time of granodiorite intrusion. (b) Estimated metamorphic conditions calculated with outer garnet composition using Theriak-Domino (de Capitani and Petrakakis, 2010). From Bosbyshell et al. (2016)

assemblage muscovite + biotite + garnet + quartz + plagioclase + ilmenite with minor staurolite and kyanite. Bosbyshell et al. (2016) estimated metamorphic conditions of approximately 600°C and 700 MPa based on garnet isopleth thermobarometry calculated with Theriak-Domino (de Capitani and Petrakakis, 2010).

Bosbyshell et al. (2016) report EPMA U-Th-total Pb results of monazite from this location. The age of metamorphism, based on analysis of eight monazite grains, is 390 ± 4.5 Ma (Fig. 5). Devonian monazite ages have been reported in the Wissahickon Formation from the sillimanite zone in Philadelphia (Bosbyshell, 2008) and in rock along strike approximately 15 km to the southwest (Bosbyshell, 2001). Slightly discordant Devonian monazite ages from a nearby sample and from the sillimanite zone downstream were previously obtained using thermal ionization mass spectrometry (TIMS) (Bosbyshell et al., 1998). The age of the igneous protolith of the granodioritic gneiss here is unknown, but a larger intrusion of similar composition, the Springfield granodiorite (Fig. 1), yielded a U-Pb zircon age of 427 ± 3 Ma (Bosbyshell et al., 2005). One monazite grain in the Bosbyshell et al. (2016) study contained a core which yielded an indistinguishable age of 428 ± 4.5 Ma. They suggest that the monazite core and the small low-Ca garnet core likely formed during a period of contact metamorphism related to the granodiorite intrusion.

Nearest the Wilmington Complex (stop two), rock of the Wissahickon Formation contains Silurian-aged, high-T low-P mineral assemblages, with Devonian kyanite-bearing assemblages present only in shear zones, where fluids likely facilitated metamorphic recrystallization (Bosbyshell et al., 1999; Bosbyshell, 2001; Pyle et al., 2006). Bosbyshell et al. (1999) describe a metamorphic gradient in the early assemblage – from west to east over a distance of less than 10 km, sillimanite + k-feldspar assemblages give way to sillimanite + muscovite and andalusite-bearing assemblages. The younger kyanite overprinting becomes pervasive as the grade of the early metamorphism decreases. Here, and along the remainder of Wissahickon Creek, the early metamorphism is found only in rock adjacent to intrusions. Thus, the early metamorphic gradient identified by Bosbyshell et al. (1999) continues to the east. This suggests that the Wissahickon Formation in the type section is higher and relatively younger than the deeper and likely older portion of the Wissahickon Formation that is exposed near the Wilmington Complex.

Continue walking north along Forbidden Drive. Descend to creek level approximately 150 meters northwest of STOP 1.5 to examine outcrops in the bank of Wissahickon Creek.

STOP 1.6 – Mesomylonite in the Rosemont Shear Zone

The term mylonite is used for rock that has experienced a high degree of dynamic recrystallization in a ductile fault or shear zone, resulting in significant grain size reduction. The rock here is mesomylonite, so named because the grain size reduced matrix comprises 50 to 90% of the rock (if a rock consists of >90% matrix it is termed ultramylonite). The highly foliated rock in this outcrop consists of plagioclase feldspar, epidote, clinozoisite, and possibly alkali feldspar porphyroclasts which range from 0.25 – 0.5 mm diameter. Pyroxene porphyroclasts are also present in very low abundance. These are largely replaced by amphibole and epidote and can be as large as 1mm. The matrix consists of micron to tens of microns scale grains of plagioclase, biotite, quartz and epidote. Quartz also occurs as mm-scale polycrystalline ribbons. Texturally late epidote, which overgrows mylonitic foliation is also present. The presence of pre-, syn- and post-tectonic epidote suggests that deformation occurred under conditions corresponding to the epidote-amphibolite metamorphic facies, similar to, or at slightly lower temperature than, the conditions estimated for Devonian metamorphism in the nearby Wissahickon Formation. Fabrics in schist indicate that shear zone deformation occurred subsequent to the growth of garnet and staurolite porphyroblasts.

Return to vans and proceed to stop two.

Mileage	*Cum. mileage*	*Description*
0.0	30.1	Exit the parking lot turning left on Bells Mill Rd.
1.1	31.2	Turn right onto Ridge Ave.
3.8	33.9	Slight right onto I-476 S access ramp
4.0	34.1	Merge on I-476 S
13.8	43.9	Take exit 9 to merge onto PA-3 W
20.5	50.6	Turn left onto N. Sandy Flash Dr. (<0.5 mi from Providence Rd.)
21.4	51.5	Turn right into parking lot of Colonial Plantation

STOP TWO: THE AVONDALE MASSIF IN RIDLEY CREEK STATE PARK.

Most of the land which is now Ridley Creek State Park was the former site of the estate of Walter Morrison Jeffords, Sr., a Philadelphia businessman and prominent Thoroughbred racehorse owner and breeder in the first half of the twentieth century. The park encompasses some 2600 acres and is underlain by gneiss that has historically been mapped as the Mesoproterozoic Baltimore Gneiss of the Avondale nappe (Bascom et al., 1909), although mapping and preliminary geochronology cast doubt on this association (Bosbyshell et al., 2006b). Recent research (Trice et al., 2014; Noble et al., 2015) has focused on an outcrop some two miles east of our present location, along Bishop Hollow Road near the eastern entrance to the park, where a unit informally referred to as the Sycamore Mills formation is exposed (Fig. 9). However, given the small size of the outcrop, parking considerations, and the number of participants on this trip, we will view gneiss of the Avondale massif here, near the Colonial Plantation.

On the east side of the park, the Bishop Hollow Road outcrop exposes a rare pelitic lithology in the Avondale massif, informally designated the Sycamore Mills formation. The presence of garnet with crystallographically oriented inclusions of rutile needles (Trice et al., 2014; Noble et al., 2015) makes this

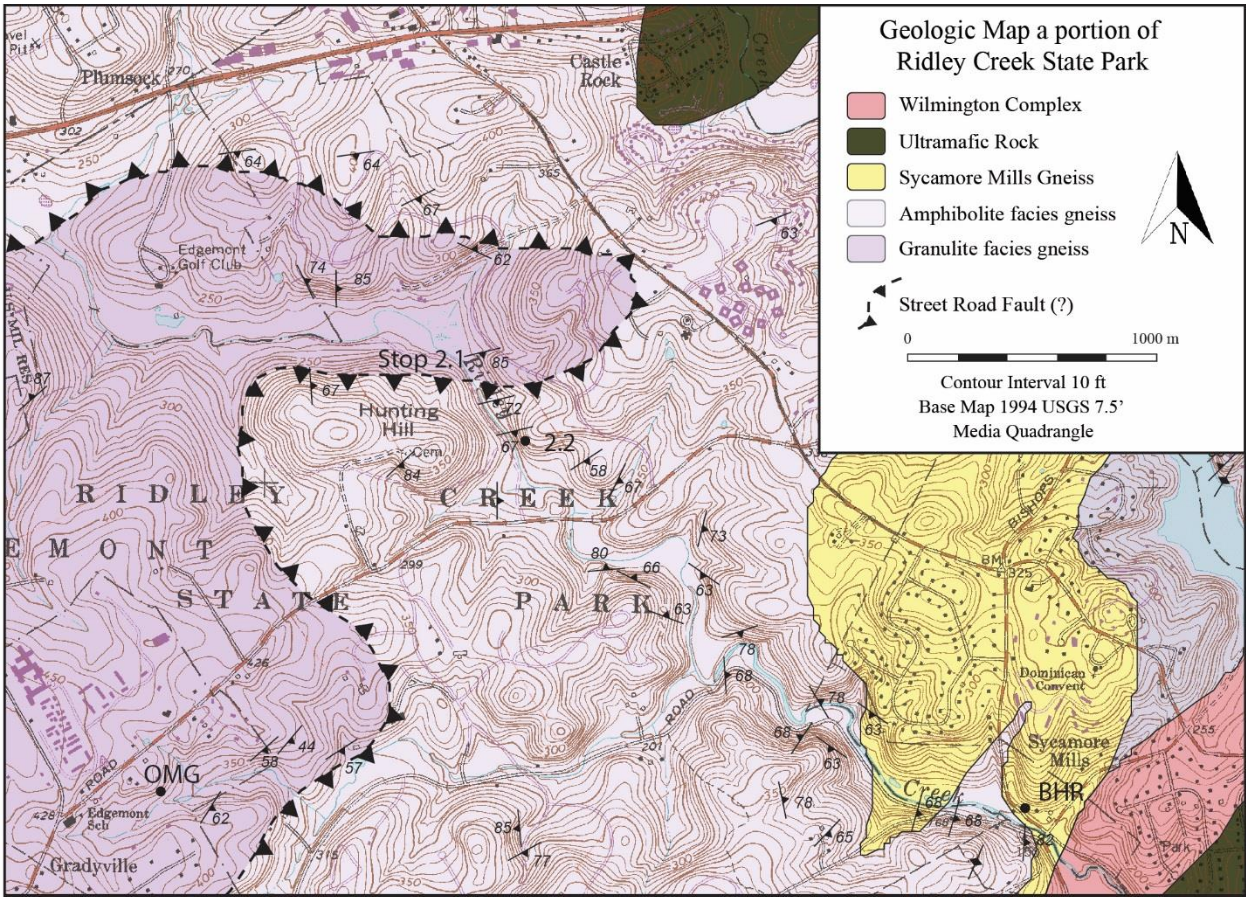


Figure 9. Geologic map of a portion of Ridley Creek State Park and the surrounding area, showing locations of Stops 2.1 and 2.2, Bishop Hollow Road (BHR) and olivine metagabbro (OMG) outcrops described in text.

rock especially interesting. The oriented rutile is possibly indicative of ultrahigh-temperature (UHT) and/or ultrahigh-pressure (UHP) metamorphism (e.g., Ague and Eckert, 2012). The rock is medium- to coarse-grained migmatitic gneiss composed of the middle to upper amphibolite facies assemblage garnet + biotite + sillimanite + quartz + plagioclase + alkali feldspar with accessory ilmenite. In addition to rutile needles, garnet cores contain inclusions of kyanite and coarsely exsolved antiperthitic plagioclase feldspar. An equilibrium assemblage diagram calculated using Theriak-Domino (de Capitani and Petrakakis, 2010) indicates minimum pressure of approximately 1100 MPa (or burial to a depth of ~35 km) for kyanite + rutile bearing assemblages. Using the Tomkins et al. (2007) calibration of the Zr in rutile thermometer, Trice et al. (2014) estimated metamorphic temperatures of 775 °C at this pressure. Noble et al. (2015) reintegrated the composition of antiperthitic plagioclase inclusions and estimated temperatures in excess of 1000 °C.

Outer portions of garnet also contain quartz inclusions which are surrounded and separated from garnet by plagioclase-Al_2SiO_5 intergrowths (Fig. 10). This texture suggests that plagioclase formed from garnet, indicating rapid decompression from this depth while still at high temperature (Whitney, 1991). The age of the early high T, high P metamorphism remains poorly constrained, but rapid decompression in the Sycamore Mills formation likely took place in the early Devonian (Bosbyshell, unpublished data). These plagioclase coronas against garnet contrast markedly with textures in the granulite facies rocks at Stop 1.2, described below, which are indicative of crustal thickening, or tectonic burial of the rocks.

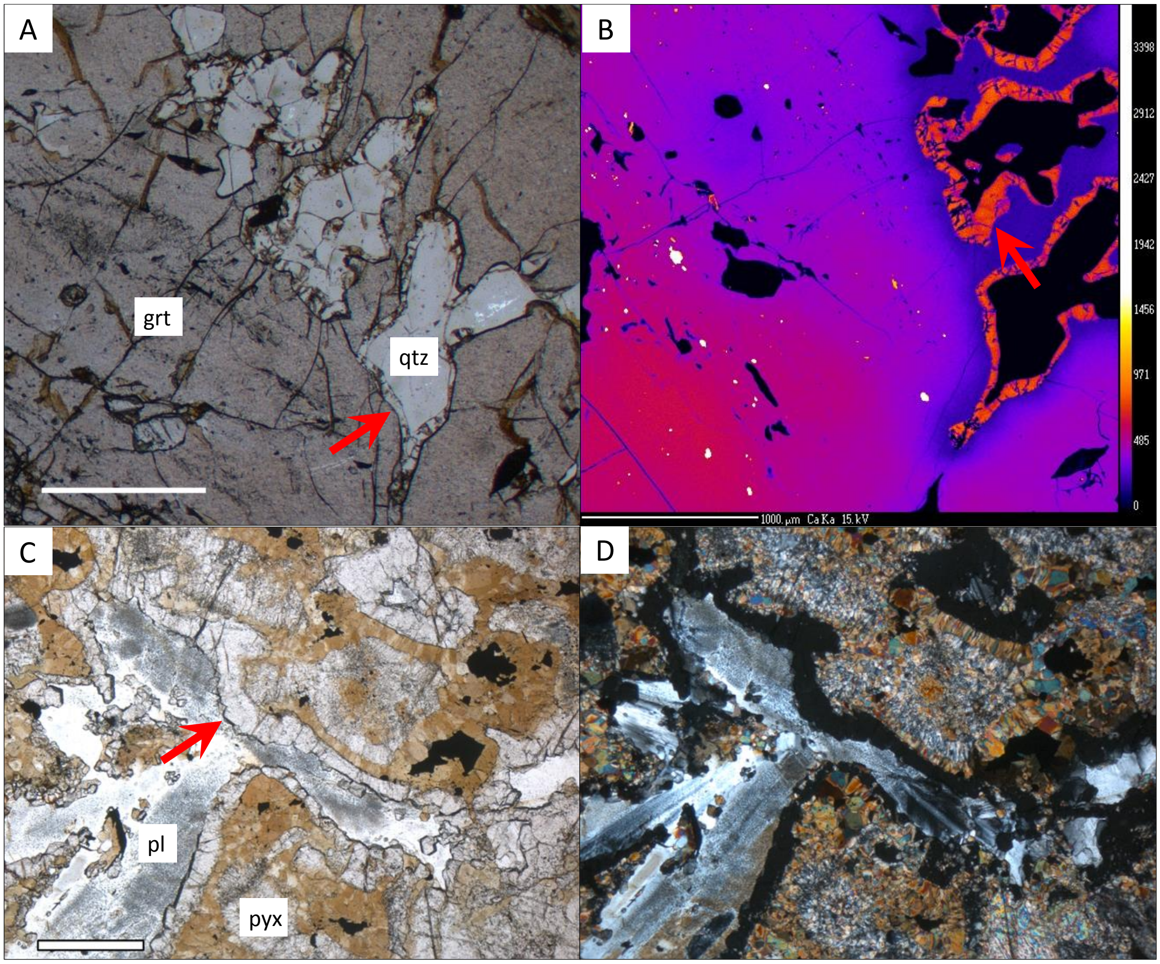


Figure 10. Textures in rocks of the Avondale massif indicating opposite P-T histories – but the time of their formation is unknown. A. Garnet (grt) from the Bishop Hollow Rd. outcrop (BHR, Fig. 9) in which quartz inclusions exhibit plagioclase coronas (arrow) against garnet, indicative of high temperature decompression. Fine inclusions in the garnet core are rutile. B. Ca-x-ray of garnet in C showing plagioclase coronas (orange). The highest-Ca inclusions (white spots) are apatite. C and D (crossed polarizers). Olivine metagabbro from the granulite zone (OMG, Fig. 9) exhibiting garnet coronas (arrow) on plagioclase (pl) against complex amphibole coronas on pyroxene (pyx), indicating an increase in pressure during metamorphism. Scale bars in A and C are 0.5 mm.

Walk down the trail to an outcrop approximately 10 m north of the trail.

Stop 2.1 – Granulite facies gneiss – a window through the Street Road Fault?

The rocks here are heterogeneous, massive to moderately foliated and lineated, medium to coarse grained, granulite facies gneiss. Felsic to intermediate compositions predominate in the Avondale massif, but cm- to m-thick mafic layers and larger metagabbro bodies are present (although not at Stop 2.1). Gneissic fabric is defined by cm- to m-scale compositional layering; some felsic rocks are characterized by lineated fabric defined by aligned mafic minerals and 2-5 cm quartz ribbons. These rocks are composed of plagioclase, quartz, orthopyroxene, clinopyroxene, garnet, hornblende and biotite. Coronas of garnet and amphibole or clinopyroxene are present between plagioclase and orthopyroxene in all granulites examined. Metagabbro is medium grained with a sub-ophitic texture and consists of olivine, orthopyroxene, clinopyroxene, and plagioclase. As in the granulite gneiss, garnet coronas are present between pyroxene and plagioclase. Massive to lineated amphibolite grade gneiss consisting of plagioclase, hornblende, garnet, and quartz is also present locally. These rocks are similar to the eastern granulites in the West Chester massif, described by Wagner and Crawford (1975) and are interpreted to be Proterozoic in age, though the metagabbro is clearly younger than the host gneiss.

Granulite facies gneiss in the Avondale massif contains evidence for two periods of metamorphism: an older high temperature episode, which produced the granulite facies assemblages; and a younger metamorphic episode that resulted in the formation of garnet coronas and complete re-equilibration to amphibolite facies assemblages in some rocks. Olivine metagabbro retains undeformed igneous subophitic textures, but does exhibit evidence for the second metamorphism in the form of garnet coronas. Coronal textures in granulite facies rocks have been interpreted to be the product of a single prograde metamorphic episode in some areas (Harley, 1989), including metagabbro in the Adirondack highlands, a Grenville gneiss terrane (Whitney and McLelland, 1973). However, the relationships described above are identical to those involving diabase dikes and host granulite gneiss of the West Chester massif previously described by Wagner and Crawford (1975) and Wagner and Srogi (1987). The authors reasoned that because cross-cutting, undeformed diabase dikes exhibit coronal textures, metamorphism of the diabase must be younger than that in rocks which are crosscut, and that coronal textures in granulite gneiss also formed during this younger, post-Grenville, metamorphic episode. Like the diabase dikes in the West Chester massif, gabbro likely intruded sometime subsequent to granulite facies metamorphism but prior to amphibolite facies metamorphic overprinting.

Garnet coronas likely formed by reactions of the general form:

orthopyroxene + anorthite → garnet + clinopyroxene + quartz (1)

or if fluid is present:

orthopyroxene + anorthite + H2O → garnet + amphibole (2).

Reaction (1) has a positive slope in pressure-temperature space (Green and Ringwood, 1976), with the garnet-bearing assemblage on the higher pressure, lower temperature side of the curve. The position and slope of reaction (2) depends on lithostatic pressure and on pH2O (Essene et al., 1970), so the significance of reaction (2) is difficult to assess without additional data. Wagner and Srogi (1987) estimated that corona formation in the West Chester massif reflected metamorphic conditions of ~650 – 700 °C at pressures of 9 – 11 kb.

While most of the rock in the West Chester massif is at the granulite facies, granulite facies gneiss is only present within a relatively small area of the Avondale massif (Figs. 1, 9). The distinct lithologic difference between the granulite facies gneiss, as seen here at Stop 2.1, and amphibolite facies gneiss throughout the remainder of the Avondale massif (Stop 2.2) led Bosbyshell et al. (2006a) to propose that the area of granulite facies rock is actually gneiss of the West Chester massif exposed in a window through the Street Road Fault.

Stop 2.2. Folded amphibolite facies gneiss.

Follow the trail along Ridley Creek until you come to a large boulder at the edge of the stream. Take a narrow trail up into the woods to an outcrop of folded amphibolite facies gneiss. The rock consists of hornblende, plagioclase, biotite and quartz. If the speculation of Bosbyshell et al. (2006a) is correct, we have crossed the Street Road Fault. The outcrop illustrates the complexity of folding in the Avondale massif. An outcrop-scale synformal hinge is visible on the uphill end of the exposure; this is likely a second generation fold. On the long side of the exposure up-right folds which arch the older synform are evident.

Return to buses, proceed to Pavilion 14 for lunch.

Mileage	*Cum. mileage*	*Description*
0.0	51.5	Exit Colonial Plantation parking lot and turn right on Sandy Flash Dr.
1.1	52.6	Turn right onto Gradyville Rd.
1.8	53.3	Turn left into the main entrance to the park, Sandy Flash Dr.
3.3	54.8	Turn left into parking area of Pavilion 14

Lunch. After lunch, return to buses and proceed to Stop 3.

Mileage	*Cum. mileage*	*Description*
0.0	54.8	Exit parking lot and turn right on Sandy Flash Dr.
1.5	56.3	Exit park turning left on Gradyville Rd.
2.8	57.6	Turn left at light onto PA-352 S/Middletown Rd.
5.3	60.1	Turn right onto PA-452 S/Pennell Rd.
7.1	61.9	Turn left onto Mount Rd., Stop 3, Novotni Bros., is on left

STOP THREE: CONFLUENCE GNEISS AT NOVOTNI BROTHERS PAVING COMPANY

This stop (Fig. 11) examines an old quarry in the Confluence Gneiss, a unit that is part of the Wilmington Complex. The Confluence Gneiss is heterogeneous gneiss, consisting of hornblende plagioclase quartz biotite granofels, interlayered mafic, felsic, and intermediate orthogneiss, and several large amphibolite bodies (Bosbyshell, 2001). This unit is contiguous with the Brandywine Blue gneiss of the Wilmington Complex (Schenck et al., 2000), but shares many characteristics with the Rockford Park Gneiss (also part of the Wilmington Complex), including the scale of layering and the boninitic affinity of some mafic rocks. Zircon in this unit yielded an age of 476 ± 4 Ma (Bosbyshell et al., 2015), identical to zircon ages of the Brandywine Blue Gneiss (476 ± 6 Ma), Rockford Park Gneiss (476 ± 4 Ma) (Aleinikoff et al., 2006) and Springton Tonalite (476 ± 4 Ma), which crops out along the trend of the Rosemont Fault approximately 7 km from Stop 3 (Bosbyshell et al., 2006b).

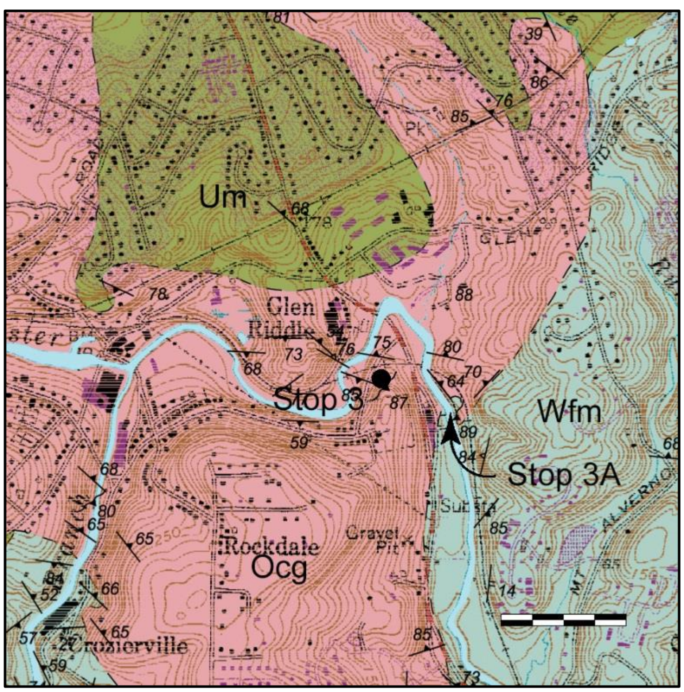


Figure 11. Geologic map of a portion of the southern Media quadrangle showing location of Stop 3. Contact between Wilmington Complex (Ocg = Confluence gneiss) and Wissahickon Fm. is exposed at Stop 3A. Scale bar is 400 m.

Geochemically, the Confluence Gneiss is similar to modern volcanic arc rocks. Intermediate gneiss resembles andesite or the intrusive equivalents, diorite, tonalite or granodiorite. Mafic layers are basaltic, and some are geochemically similar to modern boninites, rocks that occur almost exclusively in fore arcs. Boninitic amphibolite layers are also present in the Wissahickon Formation, at the contact with the Confluence Gneiss in an outcrop approximately 400m from the

Novotni Brothers quarry on the opposite side of Chester Creek. These boninitic amphibolites are collectively referred to as the Confluence Dikes.

Unlike much of the Wilmington Complex, which experienced granulite facies metamorphism, the Confluence Gneiss contains mineral assemblages characteristic of the amphibolite facies: hornblende + plagioclase ± quartz ± biotite ± garnet ± epidote. Orthopyroxene is present in some rocks to the west of Stop 3 in outcrops along the West Branch of Chester Creek, suggesting that granulite facies conditions were achieved at least locally in the Confluence Gneiss. Three zircon grains with equant morphology, indicative of a metamorphic rather than magmatic origin, yielded an age of 427 ± 7 Ma (Bosbyshell et al., 2015), indistinguishable from zircon and monazite ages from other Wilmington Complex rocks (~428 Ma; Aleinikoff et al., 2006).

Multiple episodes of deformation are evident here. A large synform is visible on the NE-SW trending quarry face (parallel to Mount Rd.). This fold and the many related minor folds deform an older metamorphic foliation and so are part of at least the second episode of deformation to have affected these rocks. The dominant foliation here strikes approximately east-west and is parallel to the axial planes of these folds. Close inspection of the folds indicates that some degree of refolding has occurred, with the axial surface of the younger folds approximately parallel to that of the first. Whether this represents progressive deformation during a single deformation event or discreet deformation episodes cannot be determined. On the horizontal surface at the north end of the quarry wall doubly closed folds suggest sheath fold geometry.

Several shear zones, which cross cut the fold set and dominant foliation described above, are visible in the other quarry wall (Fig. 12), though recent mass wasting has significantly degraded the exposure. Many similar shear zones occur along the Rosemont Shear Zone. The shear zones are characterized by a moderately to steeply plunging lineation defined by aligned hornblende crystals and sheath fold axes. Folded foliation at shear zone margins and fabric asymmetry indicate east-side down motion with a dextral strike-slip component (Bosbyshell, 2001; Kellogg et al., 2001).

At first glance, the shear zones may appear to be brittle structures, due to intense weathering of the well-defined, planar, shear zone foliation. However, petrographic analysis of mylonitic shear zone rock reveals microstructures which are characteristic of deformation under amphibolite facies conditions. Similar shear zones in metapelitic rock contain kyanite, indicating that deformation was synchronous with Devonian moderate pressure metamorphism.

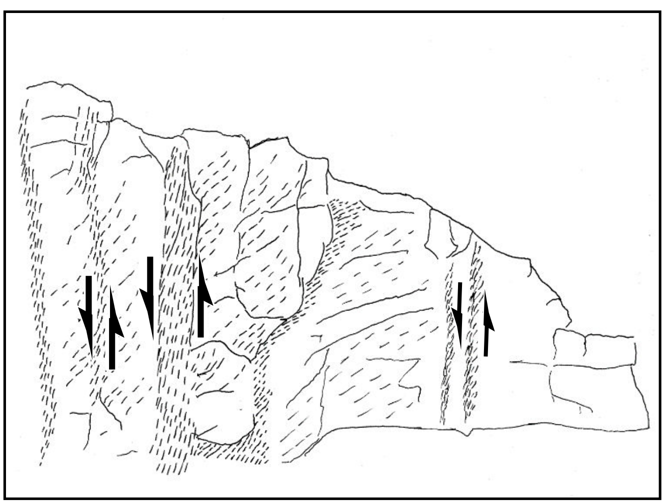


Figure 12. Sketch of shear zones in northeast facing quarry wall before recent mass wasting (Kellogg *et al.*, 2001).

As mentioned above, the contact between the Confluence Gneiss and the Wissahickon Formation, where boninitic layers/dikes are present in both units, is exposed just across Chester Creek, approximately 400m from the Novotni Brothers facility (Stop 3A in Fig.11). Unfortunately, these outcrops cannot accommodate a group of this size, so we will not visit them. The location is described in detail in the 2004 Field Conference of Pennsylvania Geologists guidebook (http://fcopg.org/wp-content/uploads/2014/06/69th2004.pdf) and more recently by Bosbyshell et al. (2015), in a guidebook prepared for the Geological Society of American Annual Meeting in Baltimore, Md. If you have interest in visiting these exposures please note it is private property; obtain permission from King's Mill banquet facility before visiting.

Return to buses.

Mileage	*Cum. mileage*	*Description*
0.0	61.9	Turn left onto PA-452 N
0.2	62.1	Turn right onto Glen Riddle Rd.
1.6	63.5	Turn right at light onto PA-352 S/Middletown Rd.
4.3	66.2	Turn left onto E Brookhaven Rd.
4.6	66.5	Turn right onto Waterville Rd.
5.2	67.1	Waterville Rd. becomes Chestnut Parkway
5.3	67.2	Turn right onto Chester Park Drive

STOP FOUR: CHESTER PARK GNEISS IN CHESTER PARK: A PERI-GONDWANAN TERRANE?

Chester Park, one of the oldest municipal parks in Delaware County, is owned and maintained by Chester City. It consists of 71 acres along the banks of Ridley Creek. The rock exposed in Chester Park is medium to coarse grained, quartzo-feldspathic biotite gneiss and schist designated the Chester Park Gneiss. New detrital zircon results (Fig. 3) from a sample at this location (Fig. 13) indicate that the Chester Park Gneiss is derived from a Gondwanan source, the first exposure of peri-Gondwanan rock to be recognized in this portion of the Appalachian orogen.

Some portions of the Chester Park Gneiss are sufficiently micaceous to be called schist. Garnet, fibrolitic sillimanite, replaced by kyanite blades at some locations, and cordierite, replaced to varying degrees by intergrowths of pale-green low-Ti biotite with either kyanite or sillimanite (Crawford and Mark, 1982; Bosbyshell et al., 2005), are also present in more aluminous domains. Accessory minerals include Fe-Ti oxides, apatite, monazite, and zircon. Alkali feldspar is present only in pegmatite pods and lenses. This lithology can be mapped from southern Darby Creek, where it is host rock to the Springfield granodiorite, south to Marcus Hook Creek, where it hosts the Arden pluton. It may be equivalent to the Fairmont member of the Wissahickon Formation in Philadelphia (Bosbyshell, 2008). Portions of the Chester Park Gneiss were originally mapped by Bascom et al., (1909) in the Philadelphia Folio as both Granitic Gneiss and Wissahickon Gneiss. Bascom et al.'s (1909) contacts were incorporated into the 1960 geologic map of Pennsylvania, but on the most recent state map (Berg et al., 1980) the area of granitic gneiss decreased and portions were mapped as gneiss of the Wilmington Complex.

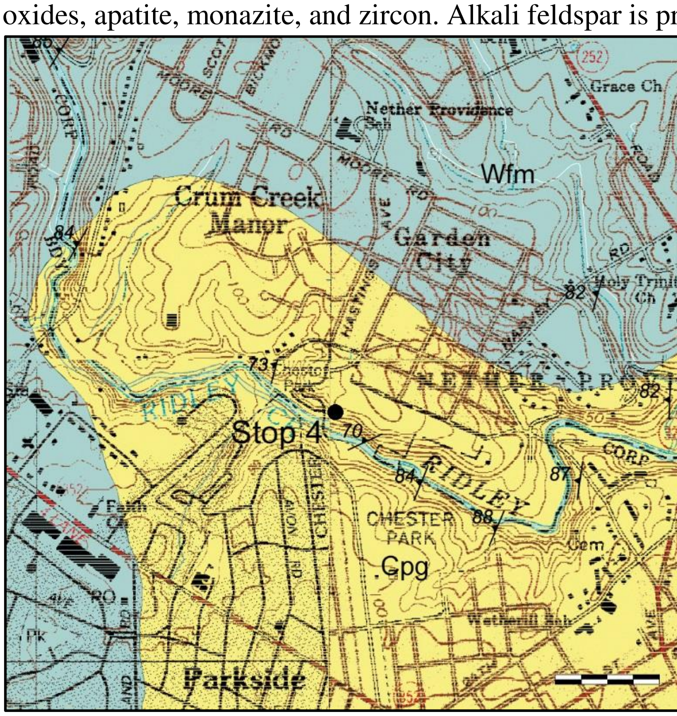


Figure 13. Geologic map of Stop 4, junction of Media, Bridgeport, Marcus Hook and Lansdowne 7.5 min quadrangles. Scale bar = 400m.

A sedimentary protolith is inferred for much of the Chester Park gneiss, based the quartz rich nature of some rock, the aluminous character of other rock and the rounded, detrital morphology of zircon grains (Bosbyshell et al., 2008). Compositional layering is present at some locations, but is generally not well developed. Other rock may have an igneous protolith. This rock frequently has a massive appearance and contains irregularly shaped and elongate, biotite-rich enclaves (xenoliths?), that vary in size from a few centimeters to several meters long, suggesting a possible intrusive igneous origin.

Mileage	*Cum. mileage*	*Description*
0.0	67.2	Turn left onto Chestnut Parkway/Waterville Rd.
0.7	67.9	Turn left onto Brookhaven Rd.
1.0	68.2	Turn right onto PA-352 N/Edgemont Ave.
11.7	78.9	Turn left onto PA-3 W/PA-352 N/West Chester Pike
14.9	82.1	Use the right lane to merge onto US-202 N/US-322 W
15.7	82.9	Take the US-322 W exit toward Downingtown
18.6	85.8	Turn right to stay on US-322 W
20.9	88.1	Turn left onto Sugars Bridge Rd.
		Turn left onto Waltz Rd.

STOP FIVE: DOE RUN AND LAURELS SCHIST; EMBREEVILLE FAULT: INVERTED METAMORPHISM

At this stop (Fig. 14) we will view the Doe Run and Laurels Schist, two units in the West Grove Metamorphic Suite which, together with the Mt. Cuba Gneiss, were formerly known as the "Glenarm Wissahickon." This area is within a zone of very complex deformation (Fig. 15) involving the Pleasant Grove-Huntingdon Valley Shear Zone, the Embreeville Thrust and the Cream Valley Fault (Wiswall,

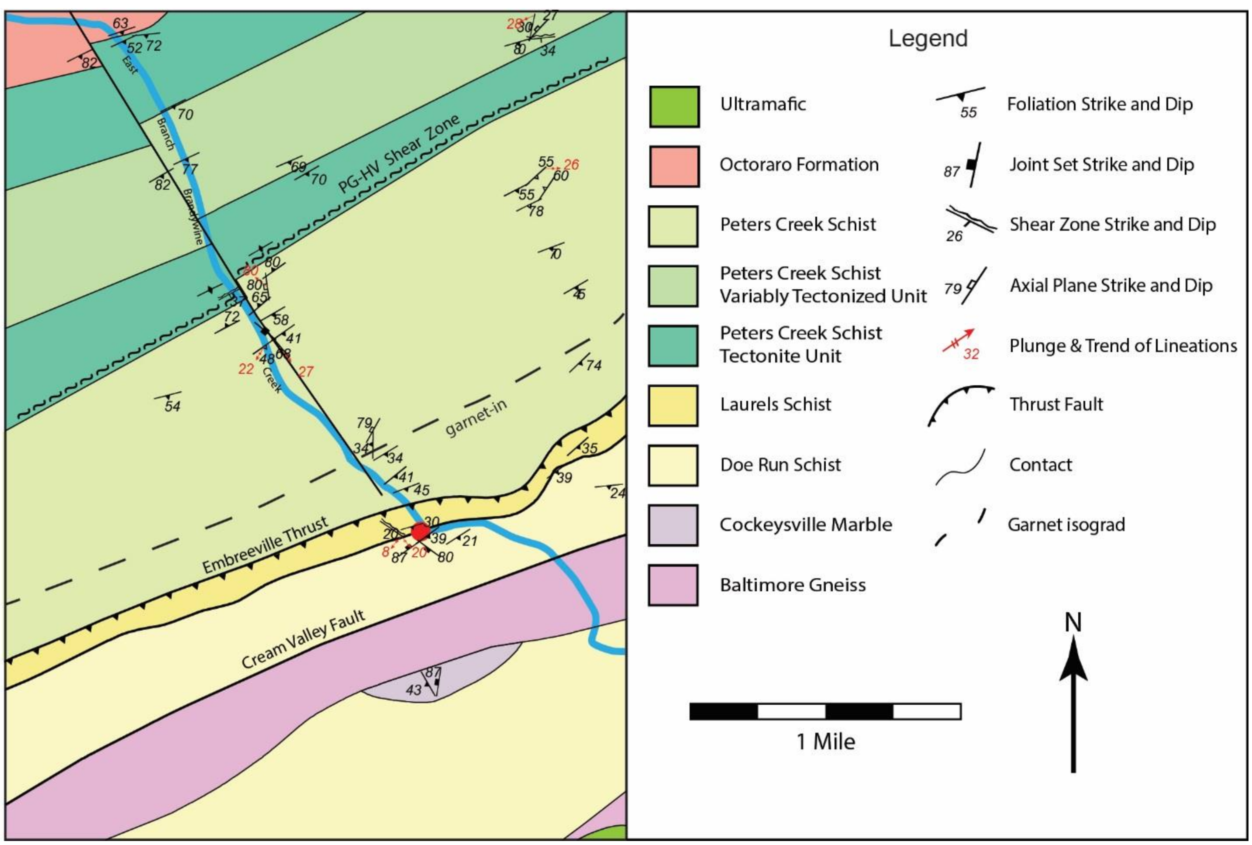


Figure 14. Geologic map of the vicinity of Stop 5 (red dot) modified from Wiswall (2005).

1990, 2005). Here the fabrics are an expression of the Embreeville Thrust, a ductile shear zone at the base of the West Chester nappe, which places amphibolite facies metamorphic rocks above rocks at the greenschist facies (Fig. 15).

Walk east to a small outcrop at the bend in Waltz Rd. This is the Doe Run Schist in the hanging wall of the Embreeville Thrust. The rock here contains garnet + biotite + muscovite + staurolite + quartz + plagioclase. The dominant foliation dips southeast.

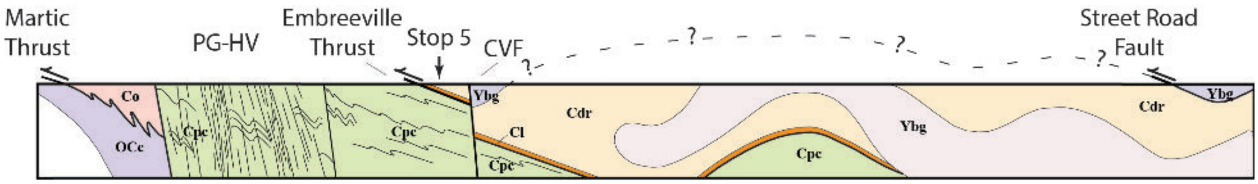


Figure 15. Cross-section of the Unionville 7.5 minute quadrangle modified from Wiswall (2005). PG-HV = Pleasant Grove-Huntingdon Valley shear zone, CVF = Cream Valley Fault. Section corresponds to A – A' in Figure 1; relative along-strike position of stop five is shown.

Return to the larger outcrops at the parking site. Here the rock is psammitic to semipelitic Laurels Schist, which contains the assemblage chlorite + muscovite + quartz + plagioclase ± garnet. The complex fabric in these rocks is an S/C mylonite. Three distinct surfaces can be discerned: an S surface formed by partitioning of the shortening component of strain, the dominant SE-dipping foliation is C, the shear surface, the third are more steeply dipping C' shear bands. The fabrics in these outcrops are an expression of the Embreeville Thrust, we are in the highest strain portion of the shear zone.

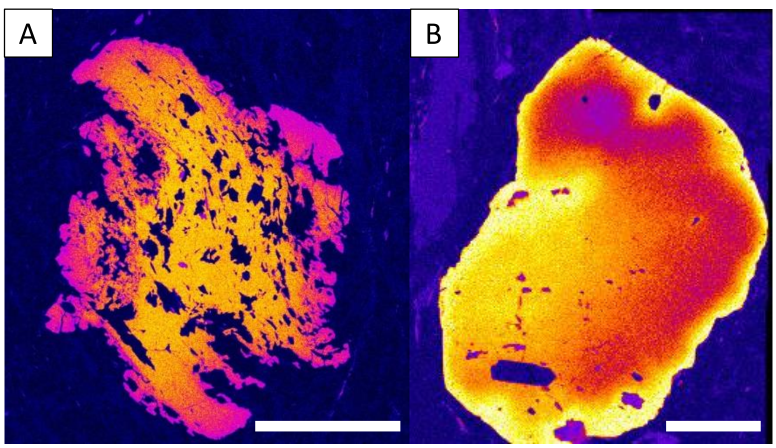


Figure 16. Mn x-ray maps of garnet. A. Syntectonic garnet from Peters Creek Schist in footwall of Embreeville Thrust. Core to rim decrease is indicative of prograde growth. B. Garnet from hanging wall (from outcrop at bend in Waltz Rd.). Sharp increase of Mn in rim indicates some degree of garnet resorbtion.

The garnet-in isograd occurs in the Peters Creek Schist approximately 300m northwest of our present location. Garnet in the Peters Creek Schist exhibits curved inclusion trails suggesting syn- to post-kinematic garnet growth (Ford and Bosbyshell, 2015). Garnet core compositional isopleths intersect at 520 C and 600 MPa, which likely indicates considerable over-stepping of the garnet-in reaction. Prograde garnet zoning (Fig. 16) suggests that maximum temperatures were somewhat higher. Peak metamorphic conditions in the Doe Run Schist in the hanging wall are estimated to have been in excess of 600 C at approximately 700 MPa based on the presence of staurolite and kyanite. Garnet in the hanging wall exhibits a sharp increase in Mn content at the rim (Fig. 16), indicating some degree of garnet dissolution which we interpret as the result of retrograde metamorphism driven by fluids released from the footwall during thrusting.

The timing of peak metamorphism in the Doe Run Schist is constrained by monazite ages, including monazite inclusions in garnet and staurolite, which range from 394 ±8.6 to 409 ±5.2 Ma

(Bosbyshell et al., 2016). Here at Stop five, monazite within relatively coarse-grained microlithons within the shear zone yields a somewhat older age, 424 ±10.6 Ma. Elongate, asymmetric, likely syntectonic monazite is younger at 387 ±6 Ma (Bosbyshell and Ford, 2017). Based on these observations, Bosbyshell and Ford (2017) suggest that emplacement of the Doe Run Schist drove prograde metamorphism in the structurally lower Peters Creek Schist.

Return to West Chester University. I hope you've enjoyed the trip!

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