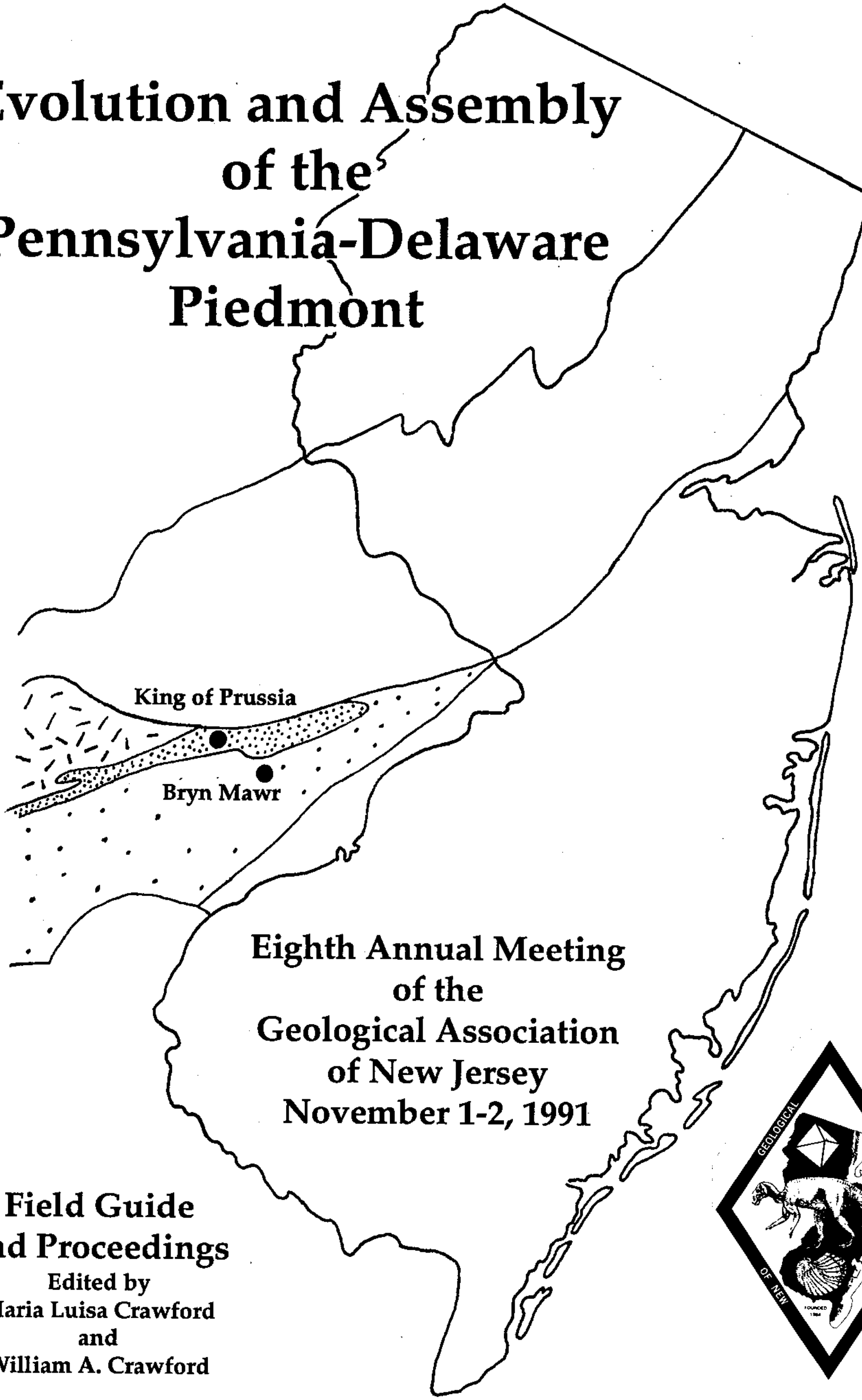
Evolution and Assembly of the Pennsylvania-Delaware Piedmont



Eighth Annual Meeting of the Geological Association of New Jersey
November 1-2, 1991

Field Guide and Proceedings

Edited by
Maria Luisa Crawford
and
William A. Crawford

EVOLUTION AND ASSEMBLY OF THE PENNSYLVANIA AND DELAWARE PIEDMONT

Edited by:
M. L. Crawford
W. A. Crawford
Department of Geology
Bryn Mawr College

EIGHTH ANNUAL MEETING
of the
GEOLOGICAL ASSOCIATION OF NEW JERSEY
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Geological Overview
M. L. Crawford
Department of Geology
Bryn Mawr College

Field Trip Leaders

STOP 1: Mary Emma Wagner
Department of Geology
University of Pennsylvania

STOP 2: W. A. Crawford
Department of Geology
Bryn Mawr College

STOPS 3 & 5:
C. Gil Wiswall
Department of Geology and Astronomy
West Chester University

STOPS 4 & 6:
J. Alcock
College of Earth and Mineral Science
Pennsylvania State University at Ogantz

STOP 7: LeeAnn Srogi
Department of Geology and Astronomy
West Chester University

K. D. Woodruff
M. O. Plank
Delaware Geological Survey

STOP 8: M. L. Hill
Department of Geology
Temple University

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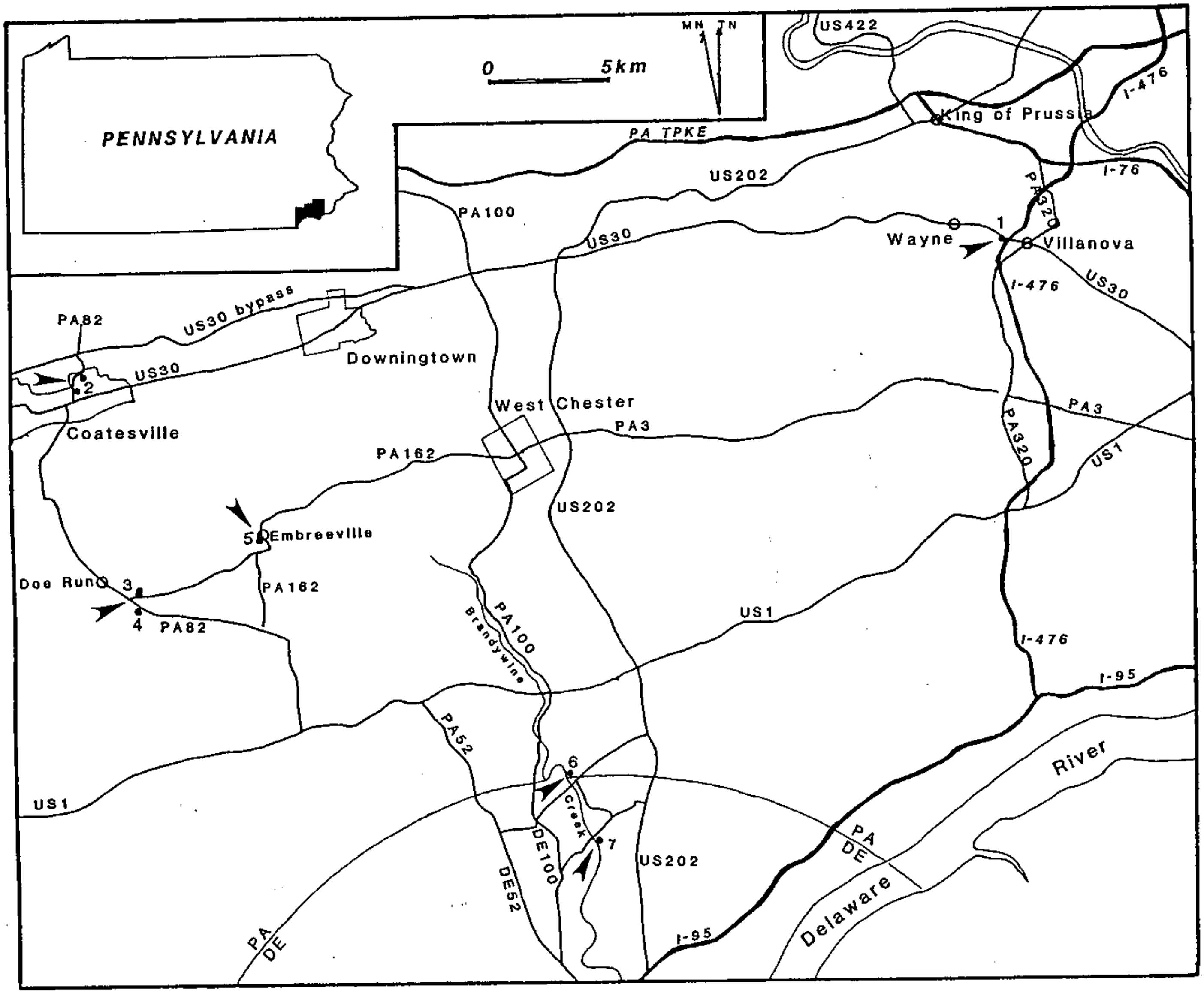


FIG. 0-1 Road map with STOP locations.

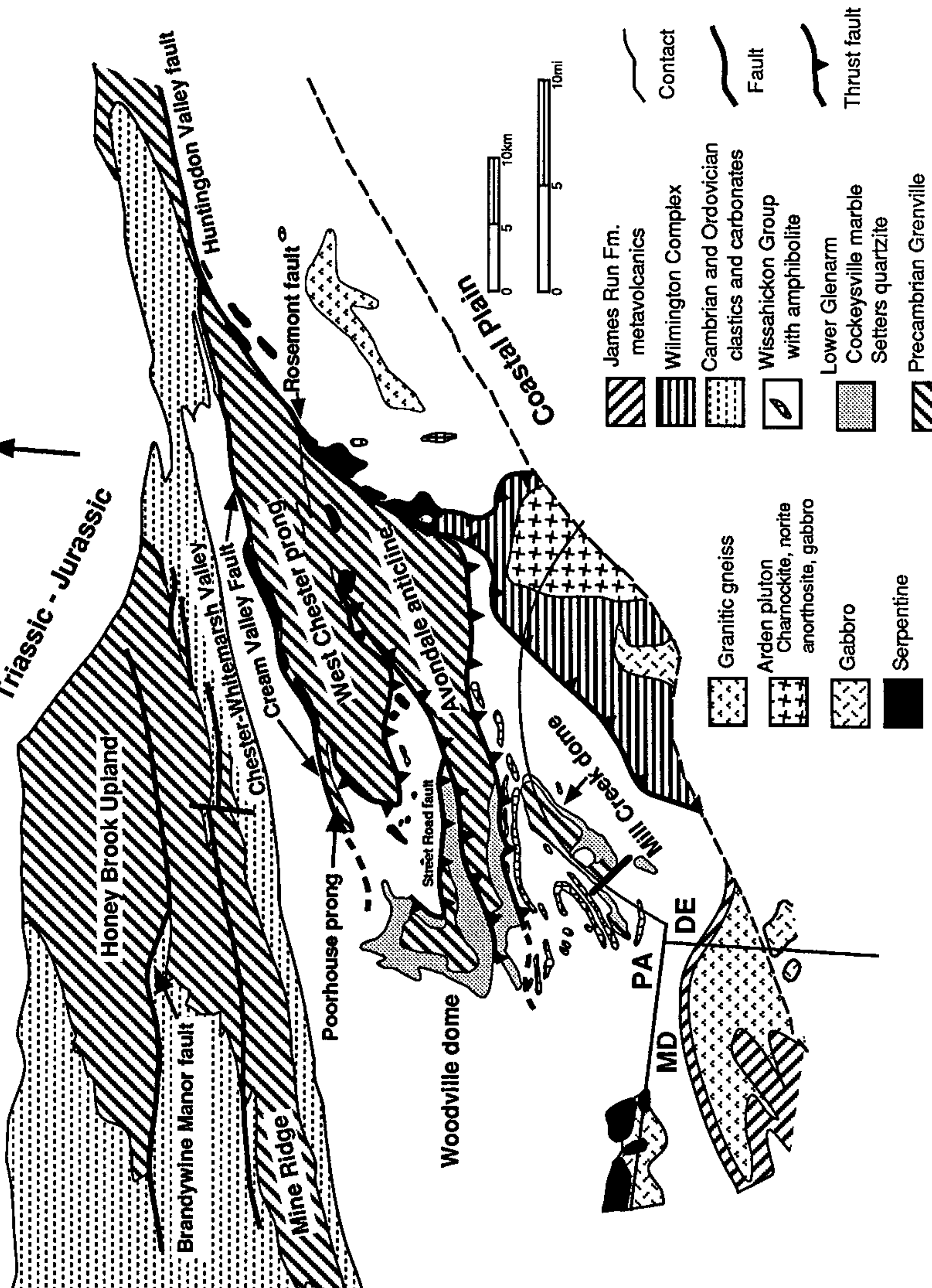

N
Triassic - Jurassic
Honey Brook Upland
Brandywine Manor fault
Mine Ridge
Chester-Whitemarsh Valley
Cream Valley Fault
West Chester prong
Avondale anticline
Huntingdon Valley fault
Rosemont fault
Coastal Plain
Poorhouse prong
Street Road fault
Woodville dome
Mill Creek dome
PA
DE
MD
0 5 10km
0 5 10mi
James Run Fm. metavolcanics
Wilmington Complex
Cambrian and Ordovician clastics and carbonates
Wissahickon Group with amphibolite
Lower Glenarm Cockeysville marble Setters quartzite
Precambrian Grenville gneiss
Contact
Fault
Thrust fault
Granitic gneiss
Arden pluton Charnockite, norite anorthosite, gabbro
Gabbro
Serpentine

Figure 1-1

GEOLOGIC OVERVIEW

Maria Luisa Crawford
Department of Geology
Bryn Mawr College

INTRODUCTION

The Pennsylvania and Delaware Piedmont has been studied for a number of years, starting with the account by Heilprin, 1885, and map of Rand, 1887, followed by the detailed mapping of Bascom (1905, 1909; Bascom and Miller, 1920; Bascom and Stose, 1932; 1938), continued by her students (Bliss and Jonas, 1916; 1922; 1923), and subsequently by many of the faculty and students at the Geology Departments of Bryn Mawr College, University of Delaware, LaSalle University, University of Pennsylvania, Temple University and West Chester University. The early mapping defined the main rock units and established the basic map pattern shown in Figure 1-1. Outcrops are sparse in much of the area, commonly limited to stream valleys, road and railroad cuts, and quarries. Because of population growth in the cities of Wilmington and Philadelphia and their suburbs, there is less access to the rocks in the field today than there was early in the century. For these reasons much recent work has concentrated on revisiting well known sites armed with tools and insights that permit us to interpret the data in different and perhaps better ways. Much of the information summarized in this guidebook relies on advances in the study and interpretation of metamorphic rocks initiated in the mid 1960's. Other insights have been obtained by applying techniques of structural investigation to rocks showing penetrative ductile deformation that have been developed during the last 10 years. The aim of the trip is to discuss the kinds of observations that can be made at the outcrop and, using additional information obtained from detailed petrographic and geochemical studies of the rocks, to present our current best interpretation of the evolution of the area between the Mine Ridge and the Wilmington Complex (Fig. 1-1). The stratigraphic sequence in the area is presented in Table 1-1.

GEOLOGIC SETTING

The north to south transect traced by this trip is in the center of a thick section of stacked thrust slices preserved in the Pennsylvania-Delaware Piedmont. The oldest rocks in the area are dated as Grenvillian (1050-980 Ma) based on U/Pb ages from zircons (Tilton and others, 1960; Grauert and others, 1973; Grauert and others, 1974). Bascom (1909; Bascom and Stose, 1938) recognized that many of these rocks have igneous precursors; a lesser volume are metasediments. Though she recognized that all the basement rocks had been metamorphosed, Bascom did not realize the temperatures were high enough that the original igneous mineral assemblages have largely been replaced by metamorphic minerals (Wagner and Crawford, 1975; Crawford and Huntsman, 1976; Crawford and Hoersch, 1984). In the northern part of the area to be visited, the Grenville-age gneisses are exposed in the Honey Brook Upland and Mine Ridge (Fig. 1-1). Uplift and exhumation of the Grenville gneiss to shallow crustal levels probably occurred by the late Precambrian. After uplift and exhumation the gneisses were covered by a clastic sedimentary sequence. At some time during this interval, but prior to deposition of the unconformably overlying sedimentary units, the gneiss was intruded by diabase dikes. It is tempting to speculate that these dikes are related to the late Proterozoic Catoctin mafic igneous rocks of the Blue Ridge and that they formed during a period of crustal extension and rifting.

Unconformably overlying the Grenville-age gneisses are metamorphosed Cambrian clastic sedimentary units: Chickies quartzite, Harpers phyllite, and Antietam quartzite (STOPS 2A and 2B). These are overlain in turn by a sequence of metamorphosed Cambrian and Ordovician carbonates. The carbonates occupy the Chester-Whitemarsh Valley, south of the Honey Brook Upland and Mine Ridge (Fig. 1-1) and continue westward into the Lancaster Valley. The only fossils in the area are found in these formations. *Scolithus linearis*, common in the Chickies quartzite, dates that formation as Cambrian. *Michelia* sp. has been reported from the Ordovician Conestoga Formation. South of the Chester-Whitemarsh Valley, and separated from the carbonate rocks by the Martic Line (Wyckoff, 1990), are the rocks of the Glenarm Supergroup which also unconformably overlie Grenville-age gneisses intruded by diabase dikes. These basement gneisses south of the Martic Line, which Bascom named Baltimore gneiss and which will be visited at STOP 1, form the cores of several structures: the West Chester prong, Poorhouse prong, Avondale anticline, Woodville dome and Mill Creek dome (Fig. 1-1).

In the field trip area the Glenarm Supergroup consists of the Setters Formation, the Cockeysville Marble, the Wissahickon Group and the Peters Creek Formation. Although the Grenville gneisses north and south of the Martic Line are different in composition and metamorphic grade (Crawford and Hoersch, 1984; Hoersch and Crawford, 1988; Wagner and Crawford, 1975), there has been a tendency to consider that the two sequences of metasediments overlying the Grenville gneisses north and south of the Martic Line can be correlated and and are probably of the same age. However, there is no proof that this is the case. The model presented here assumes the Cockeysville marble and the underlying Setters quartzite, along with the clastic and carbonate units that overlie the Mine Ridge and Honey Brook Upland, form part of a continental margin rock sequence which can be traced the length of the Appalachians.

In addition to the Cambrian and early Ordovician continental margin assemblage, the early Paleozoic geologic features of the area include an offshore magmatic arc separated by an unknown distance from the edge of the North American continent (Crawford, 1976; Crawford and Crawford, 1980; Wagner and Srogi, 1987) and a back arc (Crawford and Mark, 1982) or a fore-arc (Wagner and Srogi, 1987) basin between the arc and the continent that was the site of accumulation of sediments eroded from the arc and of a clastic wedge built out from the continental margin. The magmatic arc is represented by the Wilmington Complex gneisses and associated plutons; in Maryland the James Run Formation (Crowley, 1976) and the Baltimore Mafic complex are considered part of this magmatic arc. The Wilmington Complex is assumed to be Cambrian based on an age of 502 ± 20 Ma for the felsic igneous rocks of the Arden pluton that intrude the complex (Fig. 1-1) (Foland and Muessig, 1978).

The sedimentary rocks that accumulated between the magmatic arc and the continent are lumped together as the Wissahickon Group. They probably include units formed in a variety of settings. The age of these sediments could range from late Proterozoic to Cambrian. Crowley (1976) subdivided the Wissahickon Formation in Maryland into various formations thereby raising the Maryland Wissahickon to group status. This has not been carried out explicitly in Pennsylvania although most workers would agree that distinct lithologic units are mapped together as Wissahickon Schist. In particular, the phyllite that lies just south of the Chester-Whitemarsh Valley, mapped by Bascom as Octoraro Phyllite, is a lithologically as well as a metamorphically distinct formation within the Wissahickon Group. The high grade gneisses northwest of the Wilmington Complex (Fig. 1-1) may also be a distinct formation. There has been considerable discussion by geologists working both in Maryland and Pennsylvania concerning the age and origin of the Wissahickon Group. A summary of the discussions is presented by Higgins (1972), Higgins and others (1977) and Seiders and others (1975). Higgins and others (1977) suggested that the Glenarm Supergroup is younger than 650 Ma. The current interpretation for the rocks assigned to the Wissahickon Group in Pennsylvania and Delaware is that they represent deep water sediments, possibly distal turbidites, associated with the late Proterozoic and early Paleozoic continental margin and with an offshore magmatic arc. Metamorphosed mafic igneous rocks which occur as apparently conformable layers in the Wissahickon (Wier, 1962) may represent basaltic lava flows or shallow intrusions. Ultramafic rocks occur in linear arrays along or near contacts between the Wissahickon Group schists and the Baltimore gneiss in the area north and west of Philadelphia and around the West Chester prong (Fig. 1-1). These may be tectonically transported pieces of the ocean floor that originally underlay the sediments that form the protolith of this part of the Wissahickon Group.

FAULTS

One of the more significant emphases of recent work in this part of the Piedmont has been the role played by large-scale faulting, not only in the distribution of rock types but also for understanding the factors that control the pattern of metamorphism. Early workers identified major high-angle faults in the Pennsylvania and Delaware Piedmont that truncate units and juxtapose rocks of dissimilar ages, metamorphic grade and lithology (Fig. 1-1, Fig. 1-2). The Brandywine Manor fault bisects the Honey Brook Upland; Crawford and Hoersch (1984) suggest dip slip movement dropped the block on the south side relative to that on the north. The Cream Valley and Rosemont faults lie along the north and south sides of a basement block of Grenville-age gneisses northwest of Philadelphia. The Cream

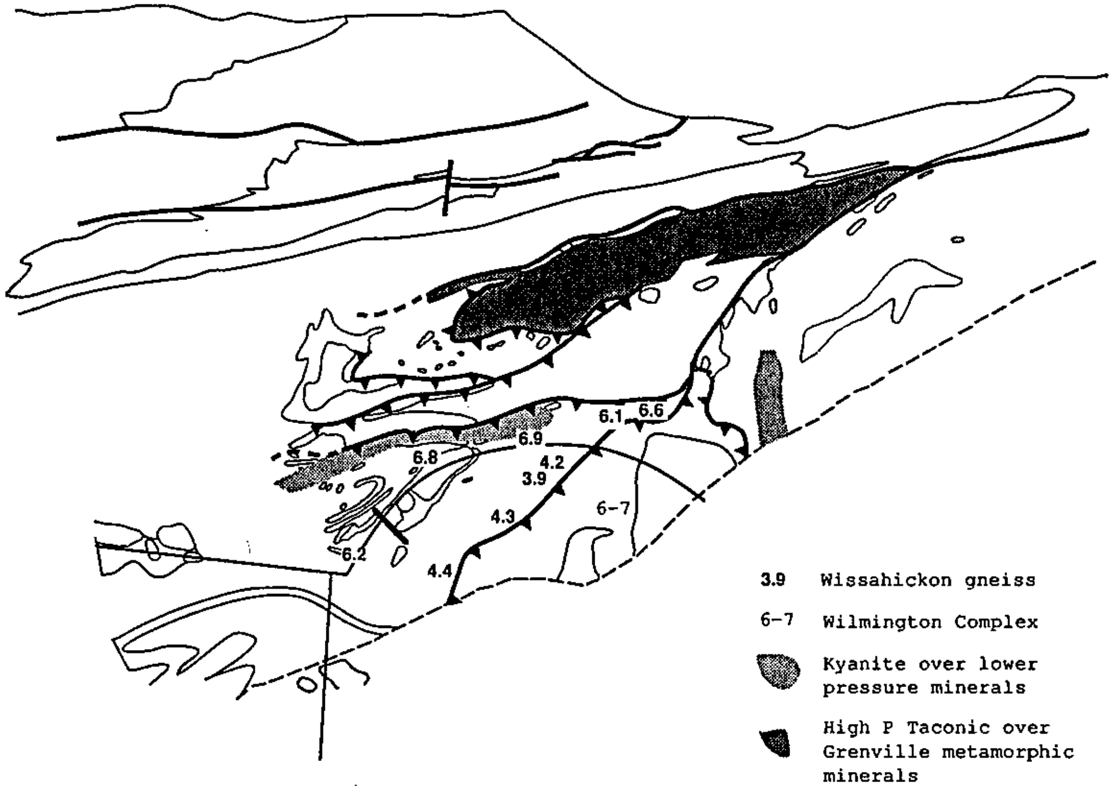


Fig. 1-2a. Pressures (in kilobars) inferred for the metamorphic rocks in the field trip area. Data from Crawford and Mark (1982), Plank (1989), Srogi, (1988).

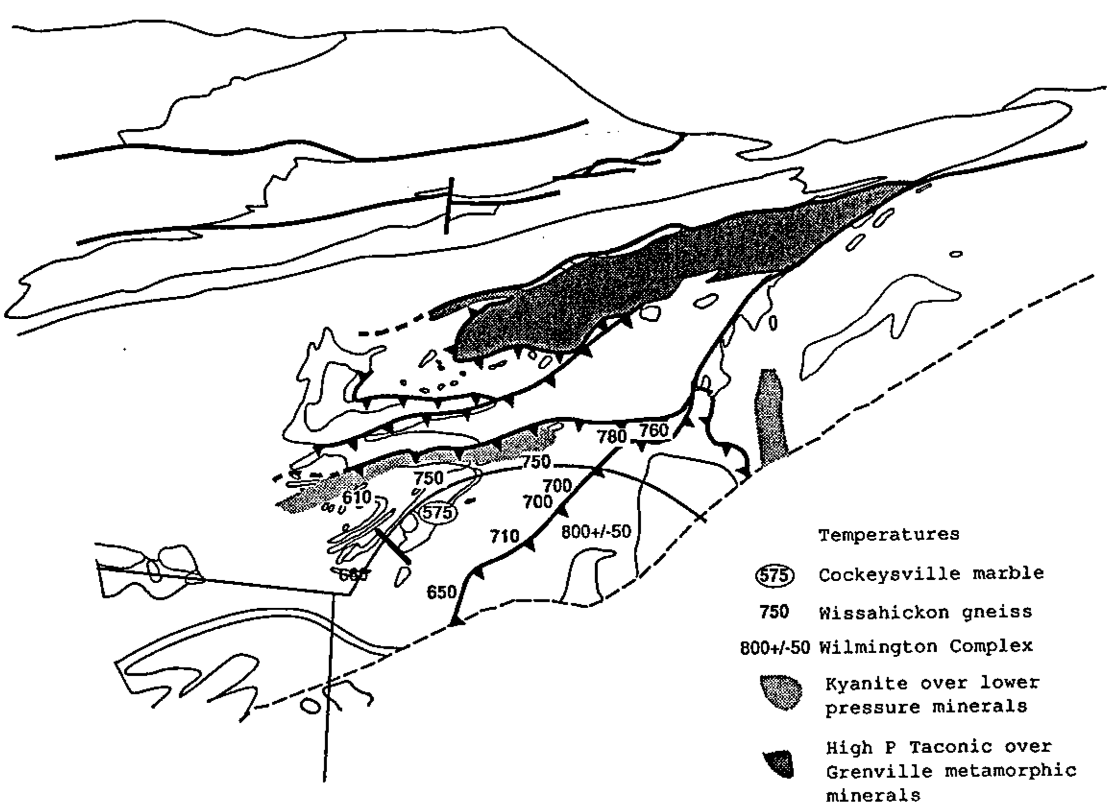


Fig. 1-2b. Metamorphic temperatures inferred for the field trip area. Data from Crawford and Mark (1982), Plank (1989), Srogi, (1988).

Valley fault can be traced from Philadelphia to west of West Chester (Fig. 1-1); the history of movement along this fault and the deformation of rocks on either side of the fault is discussed by Wiswall in connection with the structural observations at STOPS 3 and 5. Most workers have agreed that there was an early component of dip slip movement on these steep faults (Wagner and Crawford, 1975; Crawford and Hoersch, 1984; Howard, 1988). A later right-lateral offset was proposed for the Cream Valley Fault by McKinstry (1961) and Howard (1988), and a late left-lateral offset along the Brandywine Manor Fault by Valentino (1990). Hill (this guidebook) discusses the evidence for strike slip movement along many of the steep faults and describes some of the field evidence for right-lateral shearing associated with the Huntingdon Valley fault (STOP 8).

Low angle thrusts form another major group of faults. These have been controversial as they are more difficult to identify and locate. The first to be described was the Doe Run overthrust (Bliss and Jonas, 1916) which was proposed to explain the occurrence of Wissahickon schists (then thought to be Precambrian) above the quartzite and marble of the Glenarm Supergroup (then considered Paleozoic). Their field evidence for this thrust included observing the Wissahickon above marble in quarries and the fact that Wissahickon schist locally occurs at higher elevations than adjacent marble outcrops. The Woodville dome and Avondale anticline (Fig. 1-1) were interpreted as windows thrugh the thust in areas where the fault had been arched and eroded. At STOP 4 Alcock will discuss additional evidence for thrust transport of a slab of Wissahickon schist over the basement gneisses and associated Setters and Cockeysville Formations in the Woodville dome in the fashion proposed by Bliss and Jonas. In contrast to the upright sequence identified by Bliss and Jonas (1916), he suggests the Wissahickon overlies an inverted sequence of Baltimore gneiss, Setters quartzite and Cockeysville marble. This apparently inverted sequence of lower Glenarm units and Baltimore gneiss is possibly due to recumbent folding. Figure 5-1 shows Alcock's interpretation of the structural relations around the Woodville dome, which differs from that presented in Figure 1-1. If the Doe Run fault exists it does not extend as far as the edge of Chester Valley; the thrust plane must lie between the relatively high metamorphic grade Wissahickon shists of STOP 5 and the low grade Wissahickon at STOPS 3 and 4. Either the thrust thins out, or it is cut off by the westward extension of the Cream Valley fault.

An overthrust relation between rocks of the Wissahickon Group and the Grenville-age Baltimore gneiss basement of the the West Chester prong is proposed by Wagner and Srogi (1987). They cite the abundance of ultramafic rocks in the Wissahickon near the contact and the absence of the Setters Formation and the Cockeysville marble overlying the gneiss as evidence for a fault at this location. They do not think this thrust is the same as the Doe Run thrust of Bliss and Jonas (1916) and Alcock (Fig. 5-1) but prefer the Bailey and Mackin model (Bailey and

Mackin, 1937; Mackin, 1962). In this alternative interpretation the Woodville dome and the Avondale anticline gneiss massifs represent the cores of nappes with gently dipping upright southern limbs and inverted northern limbs (Mackin, 1962). Woodruff and Plank (this guidebook) present drill core evidence (Fig. 6-3) that the Mill Creek dome, a structure along the Pennsylvania-Delaware state line first described by Higgins and others (1983), is a northward overturned nappe of this type. In these nappes the Setters and Cockeysville Formations are present in the upright limbs; on the inverted limbs they have been cut by low angle thrusts, such as the Street Road thrust (Fig. 1-1) on the north side of the Avondale anticline, along which the nappe cores overrode the units in the lower limb. Alcock suggests the Street Road fault is later than and cuts the Doe Run Thrust. Detailed arguments setting forth the alternative views of Alcock and Wagner and Srogi are presented in Wagner and others (1991).

Penetrative deformation in the schists and phyllites north and east of the Woodville dome is most intense along the westward projection of the Cream Valley fault (Wiswall, 1990). This may mark the northern edge of the overthrust units debated by Alcock and Wagner and Srogi.

The Honey Brook Upland and Mine Ridge are thought to be allochthonous thin crustal slabs thrust over the lower Paleozoic carbonates of the continental shelf (Crawford and Hoersch, 1984; Hoersch and Crawford, 1988). This interpretation is based mainly on modelling of magnetic and gravity data.

Another thrust fault proposed by early workers lies between the lower Paleozoic carbonates of the Chester-Whitemarsh Valley and the Octoraro phyllite of the Wissahickon Group. This is the Martic thrust, discussed in detail in numerous reports and papers published by G. W. and A. J. Stose and by E. B. Knopf after 1929. Evidence for this fault is most controversial. There is no metamorphic break across the proposed fault and structural evidence for a thrust fault is poor (Wise, 1970). The history of the controversy surrounding this feature is summarized in Wyckoff (1990).

METAMORPHIC EVIDENCE

Metamorphic discontinuities continue to be used to identify thrust faults associated with the Taconic orogeny, even though other field evidence for the faults may be difficult to find. Interpretation of the metamorphic conditions in the Piedmont suggests that significant crustal thickening in this area resulted from tectonic stacking of crustal slices by thrust faults and nappes such as those just described. The complex metamorphic patterns document: 1) initial differences in temperature and pressure of the rocks in different thrust slabs (reflected in the metamorphic grade); 2) superimposed prograde metamorphism that reflects the response of previously metamorphosed units to new temperatures and pressures as a consequence of tectonic loading or exhumation; and (3) local late retrograde changes in the metamorphic minerals in response to conditions during late episodes of deformation, cooling and uplift.

The basement gneisses show evidence of metamorphism in the deep crust during the Grenville between 1050 and 980 Ma. At STOPS 1 and 2A evidence is presented of a second, and thus younger, metamorphism superimposed on these gneisses. The age of this younger event is constrained by the fact that it also affects the overlying lower Paleozoic rocks. North of the Martic Line this younger metamorphism is of relatively low grade (mid-greenschist facies) and no diagnostic assemblages for inferring burial depth are available (Crawford and Hoersch, 1984; Hoersch and Crawford, 1988). In contrast, upper amphibolite grade metamorphic assemblages formed during the second metamorphism in the Grenville-age basement rocks of the West Chester prong. Data provided Wagner and Crawford (1975) indicate these rocks were buried to depths of 33-38 km during this younger metamorphism.

Throughout the Piedmont evidence for high-pressure metamorphic minerals replacing an earlier metamorphic mineral assemblage is used as evidence of tectonic crustal thickening. Crawford and Mark (1982) report overprinting of relatively low pressure metamorphic assemblages including andalusite, representing burial of 14-15 km, by higher pressure minerals, including kyanite, that must have formed at depths of 25-30 km. They suggested that the sequence of events recorded by the metamorphic minerals reflects crustal loading by a thick crustal slab, possibly the Wilmington Complex and associated high grade Wissahickon gneiss which outcrop nearby. Figure 1-2 shows locations throughout the Pennsylvania Piedmont where similar high pressure overprinting relations are seen in the Wissahickon.

Alcock interprets a temperature discontinuity between the sillimanite-bearing Wissahickon gneiss (STOP 6) and the underlying Cockeysville marble south of the Woodville dome as evidence for the Doe Run thrust. The presence of abundant muscovite in the Setters Formation rather than the alternate metamorphic mineral assemblage sillimanite + potassium feldspar which occurs in the adjacent Wissahickon gneiss, also records a temperature discontinuity between the Wissahickon gneiss south of the Woodville dome and the rocks in the dome interpreted to represent the ancient margin of North America.

CRUSTAL ASSEMBLAGE

The Wilmington Complex, lies at the top of the crustal section as presently exposed. At STOP 7 we will discuss the relations between the Wilmington Complex rocks and the adjacent Wissahickon gneisses. The lower contact of the Wilmington Complex is a low angle surface that dips gently southeast (Fig. 7-1). According to Wagner and Srogi (1987) this contact is a thrust along which the Wilmington complex was transported over the underlying Wissahickon. Rocks of the Wilmington complex and the adjacent sillimanite-bearing gneisses of the the Wissahickon Group that outcrop northwest, north and east of the Wilmington Complex in Pennsylvania (Stop 6) and Delaware reached high metamorphic temperatures (Wilmington Complex: 800 + 50^{o}C, Srogi, 1988; Wissahickon: 650 + 50^{o}C, Plank, 1989). Pressures vary: 1) intermediate pressures of 6-7 Kb (or 22-25 km depth) for the Wilmington Complex (Wagner and Srogi, 1987; Srogi, 1988); 2) a band of low pressure (4-4.5 kb or 14-16 km depth) sillimanite-bearing gneisses close to the Wilmington Complex (Plank, 1989); and 3) moderate pressure gneisses (6-7 kb or 22-25 km depth) north of the low pressure Wissahickon gneiss and parallel to the Avondale anticline (Plank, 1989) (Fig. 1-2).

Mineral textures and mineral assemblages in the Wilmington Complex and the Wissahickon gneiss suggest both the Complex and the gneiss were displaced upward from deeper to shallower crustal levels while they were hot. Wagner and Srogi (1987) suggest that hot Wilmington Complex magmatic arc rocks were thrust over adjacent sedimentary units. This caused the high temperature metamorphism of the Wissahickon in this area and the mineral textures in the Wilmington Complex that record a pressure decrease. Crawford and Mark (1982) note that the low pressure metamorphism in the Wissahickon units east of the Wilmington Complex could be earlier and be due to high thermal gradients associated with a tectonic setting near the magmatic arc in a back-arc basin. Subsequent uplift and thrust transport of the combined Wilmington Complex-Wissahickon gneiss over the rocks presently exposed in the Mill Creek dome and the Avondale anticline would explain the low pressures recorded by the high temperature gneisses and possible decompression melting in Wissahickon gneiss adjacent to the Wilmington Complex (Calem, 1987).

The base of this overthrust slab of hot gneisses coincides with the temperature discontinuity documented by Alcock between the Wissahickon gneiss and the underlying lower Glenarm formations. Emplacement of this overthrust slab may also be reponsible for the higher pressure mineral assemblages developed in the Wissahickon immediately south of the Avondale anticline and east of the Wilmington Complex (Fig. 1-2).

North of the Avondale anticline the picture is less clear. Wagner and Srogi (1987) propose that the Baltimore gneiss in the Woodville dome is the core of a nappe. They suggest, following reconstructions based on down structure projections (Bailey and Mackin, 1937; Mackin, 1962), that the gneiss, together with the quartite and marble that envelope the gneissic core of that nappe, were thrust over the schists that in turn overlie the West Chester prong. This would explain the high pressure metamorphism superimposed on the earlier Grenville metamorphic mineral assemblages of the West Chester prong. However, they have not documented similar high pressure metamorphism in the schists between the Avondale anticline and Woodville dome and the West Chester prong, which would be required by their model. Alcock suggests the thrust fault that transported the low pressure Wissahickon gneiss over the Glenarm rocks south of the Avondale anticline extends northward over the Avondale anticline, the Woodville dome, and the West Chester prong. This would appear to require that the Wissahickon north of the Avondale anticline also be a low pressure and high temperature gneiss similar to the Wissahickon gneiss south of the Avondale anticline, which is not the case, particularly south of the Cream Valley fault and north of the West Chester prong.

Following the episode of crustal telescoping by thrusting, the West Chester prong was uplifted along the Cream Valley fault. This fault juxtaposes higher temperature and higher pressure schists on the south against lower temperature schists on the north. Pressures for the rocks north of the Cream Valley fault are not known. These units north of the Cream Valley fault, including the Mine Ridge and Honey Brook Upland allochthonous crustal slabs, thus represent higher portions of the thrust pile or crustal segments that were more distant from the high temperature conditions associated with the magmatic arc, or both.

AGE CONSTRAINTS

The time of thrusting and associated metamorphism is constrained to be younger than the ages of the rocks involved, and therefore must be younger than mid-Ordovician. These event must also be older than 360 MA, the cooling age of biotite in the rocks along the Susquehanna obtain by Lapham and Bassett (1964). Our best estimate for the timing of thrusting is a date of 440 Ma on the metamorphic rocks of the Wilmington Complex (Grauert and Wagner, 1975) and $^{40}Ar/^{39}Ar$ ages of 413-360 Ma on hornblende in Wissahickon amphibolite (Sutter and others, 1980). Stratigraphic and structural evidence recorded to the north and west of the Piedmont, in rocks of the Great Valley and Valley and Ridge provinces, also suggest a significant compressional tectonic event, the Taconic orogeny, occurred close to the end of the Ordovician. It is impossible, to ascertain whether the thrusting events were concentrated during a short interval or continued intermittently over a relatively lengthy time period.

SUBSEQUENT DEFORMATION

Late strike-slip movements on many of the steep faults discussed above (Wiswall, this guidebook; Hill, this guidebook) produced the present map pattern. The amount of strike-slip transport is hard to document; in some cases it appears to be at least on the order of 10's of kilometers. The timing is also unknown; it may be as late as Alleghenian.

TABLE 1-1

STRATIGRAPHIC CORRELATION CHART
(after Berg, and others, 1986)

Honey Brook Upland	West Chester Prong/ Woodville Dome/ Mill Creek Dome	Wilmington Complex
	Probably Lower Ordovician	
	Peters Creek Schist Wissahickon Schist	
	Cambrian	
	Cockeysville Marble	Wissahickon Gneiss Wilmington Complex
	Eocambrian	
Harpers Formation Chickies Formation	Setters Formation	
	Precambrian	
Felsic Amphibolite Grade Gneiss	Baltimore Gneiss	

BALTIMORE GNEISS OF THE WEST CHESTER PRONG

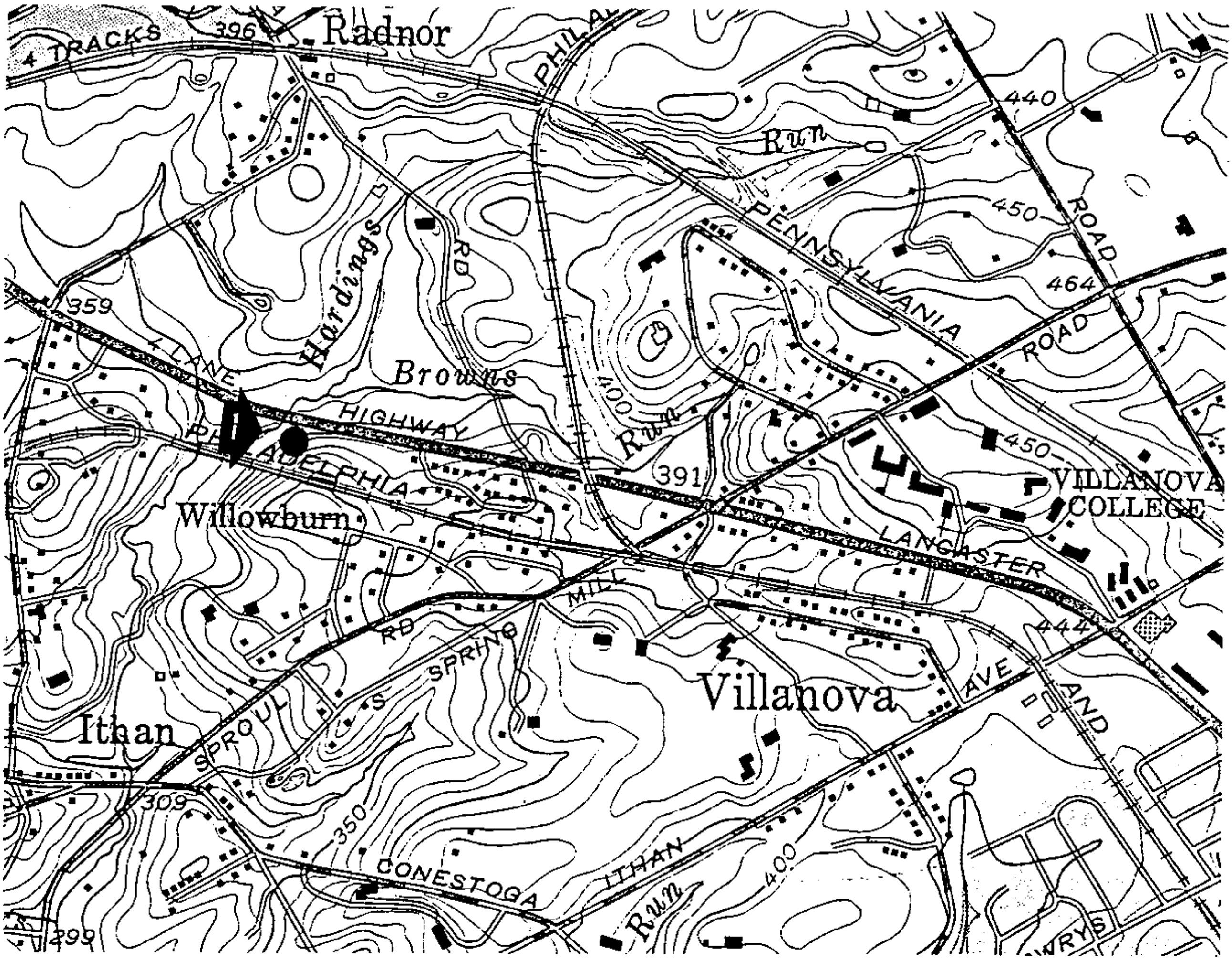
Mary Emma Wagner
Department of Geology
University of Pennaylvania

INTRODUCTION

Wagner and Srogi (1987) have proposed a model suggesting collision between a magmatic arc and the North American continent during the Ordovician Taconic orogeny, based in large part on differences in metamorphic grade among the various units exposed in the Piedmont, as well as on present-day outcrop patterns. The Wilmington complex is the highest structural level presently exposed and is considered to be the infrastructure of a magmatic arc that grew above an east-dipping subduction zone. Collision occurred when the North American continent entered the subduction zone, effectively bringing subduction to an end. In this interpretation, the continental crust of the West Chester prong was carried on the downgoing plate to beneath the arc and is the deepest level exposed in the Pennsylvania-Delaware Piedmont. Basement-cored nappes and highly deformed metasediments of the Wissahickon Formation, that probably originated in a forearc basin and/or accretionary prism, lie above the West Chester prong and below the Wilmington complex.

STOP 1

The felsic gneisses exposed in these large road cuts are near the center of the West Chester prong, a block made up of Precambrian rocks with a long and complex history. They were completely recrystallized under granulite facies conditions between 1000 and 1100 Ma during the Grenville orogeny. This



Norristown 7.5' Q. Lat 40°02'12" Long 75°21'36"

high-grade metamorphism was so pervasive that it is difficult to decipher the origin of the gneisses, although it is probable that the massive rocks exposed at this stop were originally part of an igneous pluton.

Wagner (1972) and Wagner and Crawford (1975) described an area of the West Chester prong a few miles west of this stop, where the rocks are heterogeneous gneisses that include massive felsic gneisses similar to these, as well as mafic gneisses and aluminous quartzites. The Grenville granulite-facies assemblage in felsic rocks is typically hypersthene + garnet + mesoperthite (or plagioclase + orthoclase in some areas) + quartz + ilmenite or magnetite biotite; in mafic rocks, hypersthene + diopside + plagioclase + ilmenite ± garnet ± hornblende; in aluminous quartzites, sillimanite + quartz ± spinel; or garnet + quartz; or garnet + sillimanite + quartz.

When the Wilmington Complex arc collided with the North American continent (Crawford and Mark, 1982; Wagner and Srogi, 1987), the portion of crust that now makes up the West Chester prong was buried to a depth of 35 km or more. As a result, all of the rocks of the West Chester prong were subjected to a second metamorphism at high pressure (1.0-1.1 GPa) but moderate temperature (650-700oC). In the least deformed rocks, the reactions that took place during this second metamorphism did not proceed to completion, so that many of the gneisses of the West Chester prong retain minerals from the Grenville granulite-facies event, overprinted by high-pressure assemblages of the Taconic metamorphism. Nearly all of the gneisses that still retain Grenville minerals and textures have garnet coronas that grew during Taconic metamorphism. The coronas are found around mafic minerals where they had been in contact with plagioclase. In addition to garnet, the coronas often contain fine-grained clinopyroxene and quartz between the garnet and the mafic minerals, especially orthopyroxene. The plagioclase adjacent to the garnet coronas is commonly zoned to a more sodic composition, suggesting a reaction such as:

$$\underset{\text{hypersthene}}{4(Mg,Fe)SiO_3} + \underset{\text{anorthite}}{CaAl_2Si_2O_8} = \underset{\text{clinopyroxene}}{Ca(Mg,Fe)Si_2O_6} + \underset{\text{garnet}}{(Mg,Fe)_3Al_2Si_3O_{12}} + \underset{\text{qtz}}{SiO_2}$$

a reaction that proceeds to the right with increasing pressure or decreasing temperature (or both). The aluminous quartzites contain pseudomorphs of kyanite after sillimanite, also evidence of either increasing pressure or decreasing temperature.

Gneisses that were penetratively deformed were more thoroughly recrystallized. The assemblages in these rocks record upper amphibolite-facies conditions. The felsic rocks contain biotite, garnet, plagioclase, orthoclase or microcline, quartz, and magnetite. These felsic rocks contain two generations of garnet (Grenville and Taconic), evidence that the amphibolite-facies rocks were originally Grenville gneisses. The mafic amphibolites consist of hornblende, plagioclase, sphene, $\pm$ epidote $\pm$ quartz. No aluminous quartzites have been found in areas that were penetratively deformed.

Evidence that the present metamorphic mineralogy of the Grenville basement gneiss is a result of two separate

Grenville basement gneiss is a result of two separate metamorphic episodes, and not of isobaric cooling from high temperatures at depth following Grenville metamorphism, is given by diabase dikes that cut the Grenville gneiss. The diabase dikes are fine- to medium-grained, have chilled borders against the gneiss, and many have igneous ophitic texture. The fine grain size and chilled borders suggest the dikes were intruded while the gneiss was at a shallow depth, probably during late Precambrian rifting. The dikes have been metamorphosed and have garnet coronas similar to those in the gneiss. Igneous clinopyroxene is partially replaced by fine-grained metamorphic clinopyroxene or amphibole; calcic plagioclase is often replaced by sodic plagioclase and garnet; rare olivine is partially replaced by orthopyroxene, clinopyroxene and garnet. Taken together, the evidence of a metamorphic episode that recrystallized metamorphic assemblages in the Grenville-age gneiss and recrystallized igneous minerals in the dikes suggests that the whole West Chester prong was buried and subjected to high-pressure metamorphism during Taconic collision (Wagner and Srogi, 1987; Wagner and others, 1991).

The felsic gneisses at this stop differ from those farther west described by Wagner and Crawford (1975) in that the mafic minerals inside the coronas have been largely replaced by intergrowths of randomly oriented biotite laths, fibrous amphibole, quartz, apatite, plagioclase and scapolite. The garnets of the coronas are euhedral to subhedral (in contrast to the larger anhedral Grenville garnets) and contain small inclusions of feldspar and quartz that are typical of the Taconic garnets throughout the West Chester prong.

In one of the cuts at this stop, an orange-beige colored granitic dike cuts the gneiss. This granite contains euhedral garnets similar to those in the coronas of the gneisses and is therefore most likely pre-Taconic. In two of the exposures, post-Grenville, pre-Taconic diabase dikes cut across the felsic gneiss. The dikes are undeformed and have igneous ophitic texture that is visible in thin section, overprinted by Taconic garnet and amphibole.

A few subvertical shear zones less than 1 cm wide are present. Pseudotachylite has been described associated with similar shear zones elsewhere in the West Chester prong (Armstrong, 1941; Howard, 1988).

THE SOUTHERN MARGIN OF THE HONEY BROOK UPLAND

William A. Crawford
Department of Geology
Bryn Mawr College

OVERVIEW

Crystalline rocks, now charnockites, are the oldest in the Honey Brook Upland. They were overlain by a calc-alkaline volcanic suite and clastic sediments derived from both the ancient crystalline suite and the volcanic rocks. Carbonate rocks were precipitated in regions of diminished clastic deposition. This sequence underwent severe metamorphism during the Grenville orogeny (Sutter, and others, 1980). The ancient crystalline suite was transformed to become charnockite of lower granulite grade. More deeply buried volcanic and sedimentary rocks also reached low-pressure granulite conditions. The less deeply buried portions of the sequence became gneisses of the upper amphibolite grade. Some time prior to or during the Grenville Orogeny the charnockites were intruded by rocks of the anorthosite suite ranging in composition from anorthosite to gabbro. The anorthosite suite now contains minerals characteristic of the low-pressure granulite grade. In the Mine Ridge a mixture of crystalline basement rocks, calc-alkaline volcanic rocks, and clastic sedimentary rocks also underwent upper amphibolite grade metamorphism during the Grenville Orogeny.

All these rocks were exhumed in the late Precambrian and then overlain by clastic and carbonate rocks ranging in age from lower Cambrian to early-Ordovician. Once again, the entire sequence was buried and metamorphosed during the Taconic Orogeny (Sutter, and others, 1980). Retrograde greenschist minerals grew in the old high grade crystalline rocks and prograde greenschist minerals formed in sedimentary rocks of the appropriate compostion.

The Honey Brook Upland and Mine Ridge crystalline rocks with their overlying Cambrian to Ordovician meta-sedimentary cover rocks comprise thin slices of crust thrust upon the edge of the North American Continent sometime during or following the late Ordovician greenschist metamorphic event (Crawford and Hoersch, 1984, and Hoersch and Crawford, 1988). Whether these two crystalline regions combined to form one thrust sheet or arrived separately in their current juxtaposition cannot be resolved. The presence of the Gap Overthrust (Bascom and Stose, 1938) suggests, but does not prove, the former.

An extensive set of outcrops are found along Pennsylvania Highway 82 south of the intersection with U.S. Highway 30-bypass and north of the railroad overpass at Coatesville. Greenschist grade Cambrian clastic meta-sediments of the Chickies quartzite and Harpers phyllite formations overlies amphibolite grade Precambrian gneisses of intermediate color index. This exposure is one of the best places to observe the rocks at the southern edge of the Honey Brook Upland.

PRECAMBRIAN GNEISS

Huntsman, 1975, applied modern metamorphic terminology to the crystalline gneisses of the Honey Brook Upland (Table 2).

TABLE 2. AMPHIBOLITE FACIES ROCKS OF THE HONEY BROOK UPLAND

ROCK TYPE	BANDED	COLOR	GRAIN SIZE
Mafic banded amphibolite gneiss	YES	Dark bands predominate over light bands	medium-grained
Intermediate gneiss	YES	Light bands predominate over dark bands	fine-to medium-grained
Leucocratic gneiss	NO	Light color only	fine-to medium-grained

He suggested and Crawford and Hoersch, 1984, confirmed that the protoliths of those gneisses lying south of the Brandywine Manor fault were a calc-alkaline suite of volcanic and volcani-clastic origin subsequently metamorphosed to the amphibolite facies of regional metamorphism. The Highway 82 outcrop contains intermediate gneiss with the mineral assemblage:

quartz-plagioclase(An27-39)-biotite-epidote-+/-(muscovite-hornblende-chlorite-garnet)-opaque minerals-accessory minerals (abundant apatite, less abundant zircon and sphene, and occasional carbonate grains).

Quartz grains are granulated and broken plagioclase laths containing inclusions of sericite and clinozoisite predominate over randomly oriented, interstitial biotite in the felsic bands. Pleochroic light yellow-green/dark green-brown biotite flakes and yellow-green/green/blue-green columnar hornblende grains define the foliation of the dark bands.

Anatectic pegmatites composed of quartz and plagioclase are a promient feature of this outcrop. Calcite-graphite thermometry from amphibolite facies graphite-bearing gneisses of the Upland yield average temperatures of 680 $\pm$ 40 C for the metamorphism commensurate with upper amphibolite facies conditions. Though mineral assemblages suitable for geobarometery are scarce in the Upland pressure estimates of 5 to 6 kb are certainly reasonable. Such pressures coupled with temperatures on the order of 680 C certainly provide appropriate conditions for the generation of anatectic pegmatites.

Using 40Ar/39Ar cooling ages, Sutter, et al, 1980, demonstrated the high grade metamorphic event in the Upland is of Grenville age while a greenschist grade event affected both the gneisses and the Cambrian meta-sediments during the Taconic orogeny.

CHICKIES QUARTZITE AND HARPERS PHYLLITE

Huntsman, 1975, described the very well-sorted, clean quartz sandstones that overlie the Grenville age gneisses. A thin, basal unit, the medium-bedded, coarse-grained Hellam quartzite conglomerate, contains pebbles of light-bluish quartz and pink feldspar. Conformable above the basal unit lies a thin-bedded, pure, vitreous quartzite containing thin lenses of brown, sericitic quartzite with biotite and chlorite. Abundant black tourmaline crystals occur on the bedding planes of all these units.

In thin section, quartz, feldspar, sericite, chlorite, biotite, and tourmaline comprise all of the minerals found in the quartzites. The quartz grains commonly show undulatory extinction and well-developed sutures, and vary from 1 mm in diameter in the conglomerates to less than 0.2 mm in the fine-grained, greatly sheared, sericitic quartzites. Micaceous minerals make up to 25% of the rock. They wrap around individual quartz grains in the sericitic layers and are interstitial to quartz in the conglomerate beds. Two types of tourmaline crystals occur in bedding planes of quartzites (Trautwein, 1983). Stretched and segmented prismatic crystals, up to 13 cm long, plunge at low angles to the southwest, parallel to fold axes (see Structure Section) and are apparently syntectonic. Needle-like tourmaline crystals are common and are randomly oriented within bedding planes. This second type is probably post-tectonic.

A moderately well-defined layering is visible in the Chickies quartzite in the field. There is good evidence that the layering observed is relict bedding. Cross-bedding is clear and abundant, and Scolithus linearis, reported by Bascom and Stose, 1938, have burrows normal to sub-normal to the bedding. Field observation and thin section samples show a general upwards fining of grain size normal to the layering. The Harpers Phyllite conformably overlies the Chickies Quartzite. The phyllite is fine grained with a general gray-green color and cosists of quartz, muscovite, chlorite, albite, and biotite. Bascom and Stose, 1932, noted that a schistosity was the only discernible structural feature; field observations confirm this and suggest that recrystallization of the original sediments destroyed bedding and formed a foliation whose cleavage planes sparkle with fine mica.

STRUCTURES

All rocks display the dominent northeast to east-northeast structural grain of the Honey Brook Upland, Mine Ridge, and Chester Valley portions of the Pennsylvania Piedmont (Fig. 3-1). Pervasive foliation in the Precambrian gneisses striking N75E and dipping steeply to the south. This same trend characterizes the axial plane orientation of isoclinal folds in the Cambrian meta-sedimentary rocks of the North Valley Hills and the lower Paleozoic carbonate rocks of the Chester Valley.

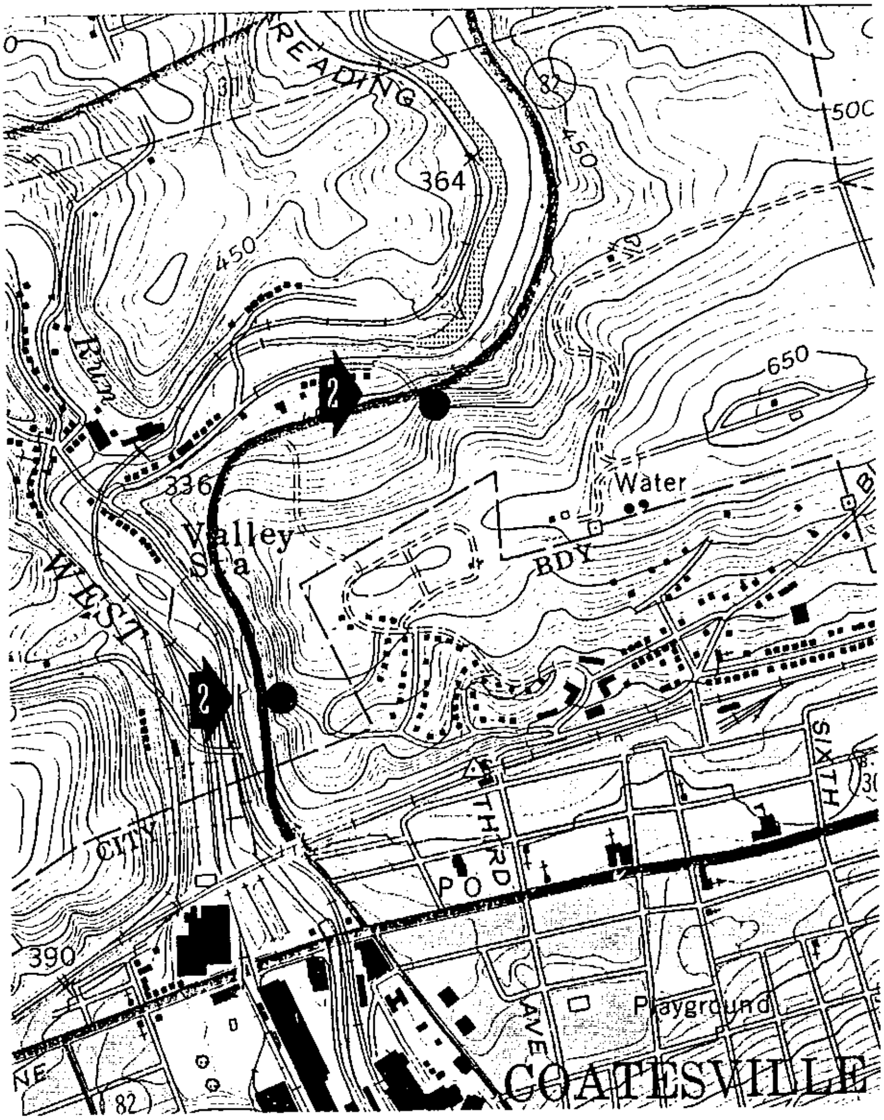
Note the poles to foliation cluster together in the meta-sedimentary rocks and form a more diffuse pattern in the basement rocks. this diffuse pattern may be the result of a Taconic refolding of these basement rocks (Hoersch and Crawford, 1988).

STOP 2

Stose provides a section measured in the 1920's in the gorge of the West Branch of Brandywine Creek (Bascom and Stose, 1932). He noted that a complete measurement could not be made because of the repetition of beds, difficulty in determining dips, and obscure outcrops, but the following relations were observed (Table 1).

TABLE 1. SECTION OF CAMBRIAN ARENACEOUS ROCKS AT COATESVILLE, PA.

Harpers phyllite and probably Antietam quartzite:	
Silvery micaceous quartzose schist and mica schist	278 feet
Chickies quartzite:	
Thin quartzite and mica schist	270 feet
Vitreous quartzite and quartz schist	150 feet
Sheared quartzite and quartz schist (the lower part is conglomeratic west of the creek - Hellam conglomerate member)	137 feet
	557 feet
Gabbro, Precambrian	



Coatesville 7.5' Q. Lat 39°59'07" Long 75°49'34"
39°59'33" 75°49'22"

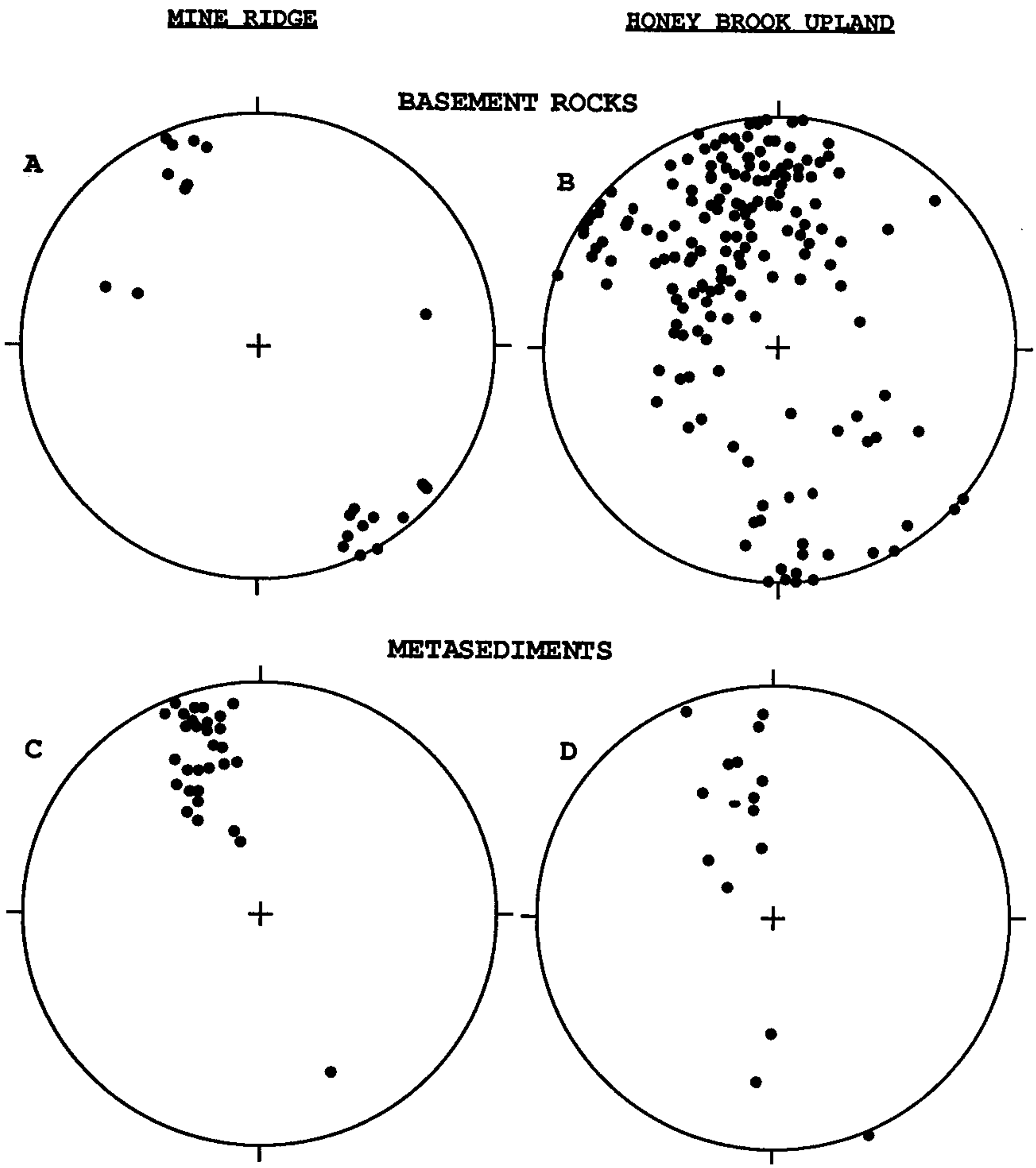


Figure 3-1
Equal area projections of foliation data from the crystalline rocks (A,B) of the Mine Ridge and Honey Brook Upland and surrounding Cambro-Ordovician metasediments (C,D). Data compiled from Postel (1951), Wise (1970), Huntsman (1975), Demmon (1977), Thomann (1977), Trautwein (1983), and Hoersch andCrawford (1988).

STRUCTURAL EVOLUTION OF THE CREAM VALLEY FAULT ZONE

C. Gil Wiswall
Department of Geology and Astronomy
West Chester University

INTRODUCTION

The formation, movement history, and tectonic significance of the Cream Valley fault zone (Fig. 1-1) remain poorly understood. The first geologic maps of the region show the Cream Valley fault zone as a steeply dipping thrust fault (Bascom, 1909). In the few exposures of the structure near its juncture with the Rosemont fault at Conshohocken, PA (Fig. 1-1), it is a near vertical zone of mylonitization. This orientation and the juxtaposition of Grenville age gneisses in the West Chester Prong against the lower grade and younger Octoraro phyllite across the fault zone led to the suggestion of reverse (south side up) motion (Armstrong, 1941). More recently, detailed fabric studies in the vicinity of the Cream Valley fault zone (Hill, 1987; Howard, 1988; Valentino, 1990; Valentino and Wiswall, 1991) and the recognition of dextral strike-slip along ductile faults in similar tectonic positions elsewhere in the Piedmont (e.g. Drake, 1989; Hatcher and others, 1988; Horton and others, 1988) have led to the suggestion that the Cream Valley fault zone shows significant dextral strike-slip movement. This interpretation would explain the remarkably straight map trace and the presence of Grenville age gneisses across the fault zone (north of the Huntington Valley fault) and northeast of the West Chester Prong gneisses of similar age.

This portion of the trip will examine exposures along the west Cream Valley fault zone in Chester County, PA. This segment of the fault zone is distinctive in that basement gneisses are not present in the hanging wall and the structure exhibits a shallow southeast dip. This is in contrast to the near vertical orientation of the foliation in mylonitized gneisses found to the east. These stops will demonstrate that structures associated with the Cream Valley fault zone occur in a band of variable width and extend along strike further than previously recognized. In addition, the deformation associated with this zone documents a complex kinematic history that probably spans much of the Paleozoic.

AREAL EXTENT OF THE CREAM VALLEY FAULT ZONE

Recently, several workers have identified with Cream Valley fault zone as a major through-going terrane boundary (Drake, 1989; Faill and MacLachlan, 1989; Gates and Sun, 1989; Hatcher, 1989; Sheridan, 1989; Wagner, Srogi, and Alcock, 1989). However, current geologic maps show the fault zone terminating at the southwestern end of the Poorhouse nappe (Berg and others, 1980). If the Cream Valley fault zone is a major terrane boundary, then it must continue to the southwest beyond its presently mapped extent. Data from the vicinity of the Poorhouse nappe and to the southwest suggest that this is the case.

Around the Poorhouse nappe, the mapped trace of the Cream Valley fault zone corresponds to a distinct structural discontinuity and steep metamorphic gradient. Analysis of dominant foliation orientation shows that this area can be

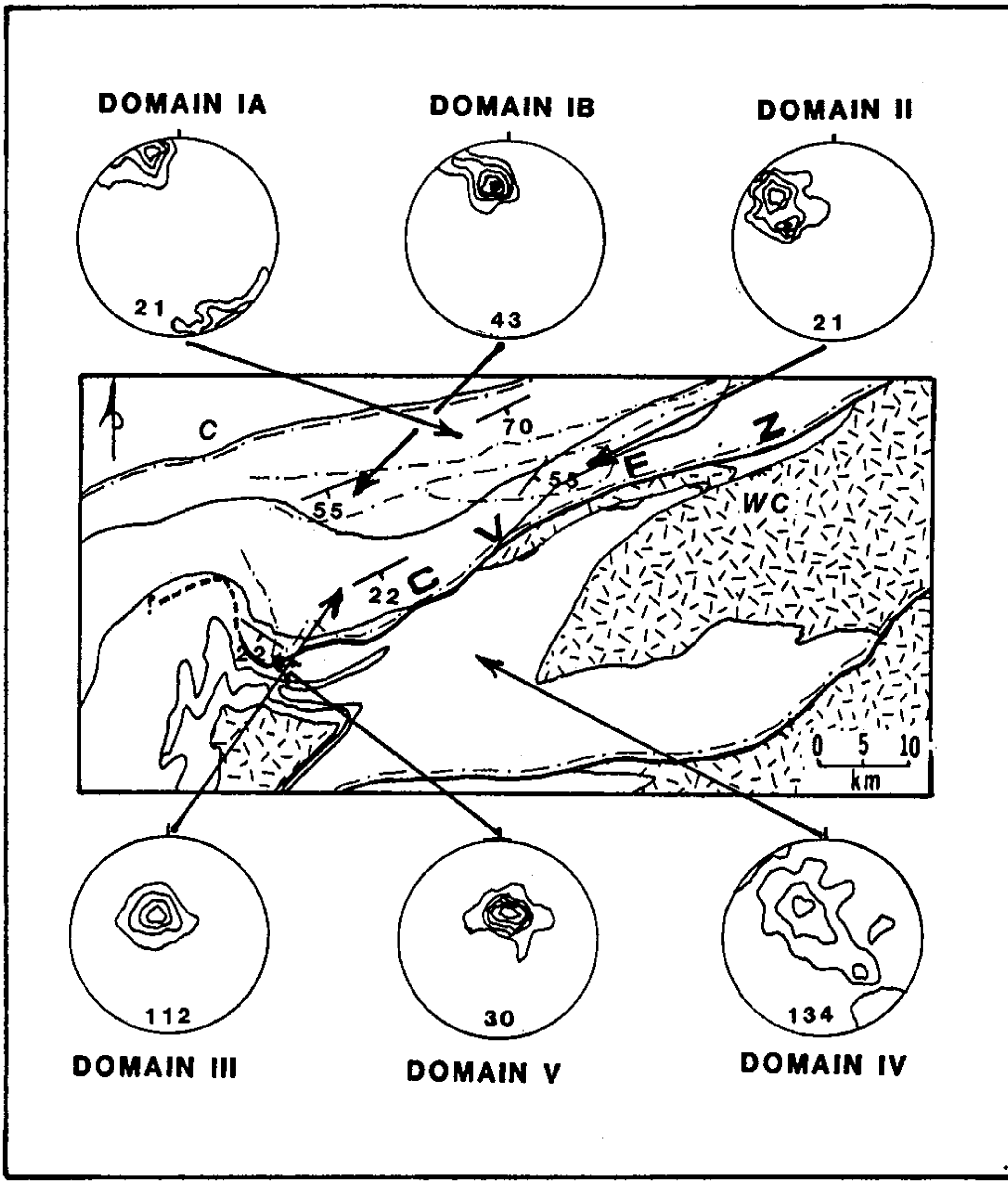


Figure 4-1. Domain map of dominant foliation orientations in the vicinity of the western Cream Valley fault zone. Orientation data plotted on lower hemisphere Schmidt nets with contours equal to 5% per 1% area. Dash-dot lines show domain boundaries; strike and dip symbols show average foliation orientations in the domain. Grenville age gneisses are patterned. C = Coastesville; WC = West Chester.

divided into several domains (Fig. 4-1). The trace of the Cream Valley fault zone corresponds to a boundary separating foliation orientations which define a point maximum in stereoprojection on the north from those defining a girdle distribution on the south. Hanging wall rocks on the south contain coarse-grained mineral assemblages in the middle/upper amphibolite facies. In contrast, rocks in the footwall adjacent to the map trace are fine-grained schists in the garnet zone of the greenschist facies. Texturally, rocks in the footwall adjacent to the Cream Valley fault zone trace are S-C mylonites whereas hanging wall rock fabrics are best described as refolded folds.

The structural, textural, and metamorphic relationships which characterize the Cream Valley fault zone in the vicinity of the Poorhouse nappe can be traced along strike to the southwest to at least the eastern edge of the Woodville nappe. Reconnaissance mapping to the north and west of the Woodville nappe suggests the fault zone continues to coincide with the foliation domain boundary and parallels formational contacts around the northern end of the nappe. Thus it has been folded. The Cream Valley fault zone may traverse the entire southeast Pennsylvania Piedmont.

DEFORMATIONAL HISTORY

Cross cutting fabrics in rocks bordering the western Cream Valley fault zone suggest three deformational events. Similar structural sequences occur in both hanging wall and footwall. The earlier deformations occurred at middle/upper amphibolite facies conditions in the hanging wall, but upper greenschist facies in the footwall. Later deformation in both blocks occurred under greenschist facies conditions. Structural analysis, detailed below, suggests that the Cream Valley fault zone formed as a thrust associated with the emplacement of basement-cored nappes probably during the Taconic orogeny. Subsequent transpressional deformation appears to have reactivated the fault zone with the nature of recurrent motion controlled by the orientation of the inherited fabric. Along the western Cream Valley fault zone in the area examined on this trip, the structure maintains a shallow southeast dip and the later deformation resulted in folding and the formation of a non-penetrative, oblique slip crenulation cleavage.

Since the Cream Valley fault zone juxtaposes rocks of different metamorphic grade along its length, and has experienced a protracted movement history, it may not be reasonable to correlate structures across the fault zone. This is particularly true of structures formed in the early stages of deformation. As a consequence, fabric element notations used in the following discussion are suffixed with "f" or "h" to denote relative age of that element in the footwall or hanging wall, respectively.

Descriptive Analysis

Lithology. The hanging wall of the Cream Valley fault zone lies south of the map trace (Fig. 1-1). It is composed of the West Chester, Poorhouse, and Woodville nappes. Each nappe consists of a core of Grenville age gneiss with overlying metasediments including Wissahickon schist and locally, lower Glenarm Series Setters Formation and Cockeysville Marble. Where the lower Glenarm Series rocks are present, they lie between the Wissahickon and the gneisses. Where these rocks are absent, the Wissahickon is in structural contact with the gneisses.

The footwall, north of the map trace, contains the Peters Creek schist and Wissahickon schist. Rocks mapped as Wissahickon schist in the footwall are mineralogically and texturally quite distinct from rocks bearing the same name south of the fault zone. The Wissahickon schist in the hanging wall is coarse grained with quartz, plagioclase, muscovite +\- biotite, garnet, staurolite, and kyanite. In the footwall, both the Wissahickon and Peters Creek schists are pelitic to psammitic, fine grained, albite-chlorite-muscovite phyllites and schists with variable biotite and garnet. In general, the Peters Creek tends to be more psammitic than the Wissahickon.

Planar Fabrics. Each lithology bordering the Cream Valley fault zone exhibits several generations of planar fabrics. Psammitic units within the Peters Creek schist preserve original compositional layering (SO) defined by centimeter scale alternation of quartz/feldspar and phyllosilicate layers. Individual compositional layers may be traced on the order of meters through subsequent folds. These units record structures which are not commonly observed elsewhere due to a lack of marker horizons. Most notable among these are isoclinal folds of bedding with axial planar metamorphic schistosity (S1) or with axial planes parallel to the regional metamorphic schistosity (S1). The alignment of undeformed metamorphic minerals to form S1 demonstrates that D1 deformation coincided with regional

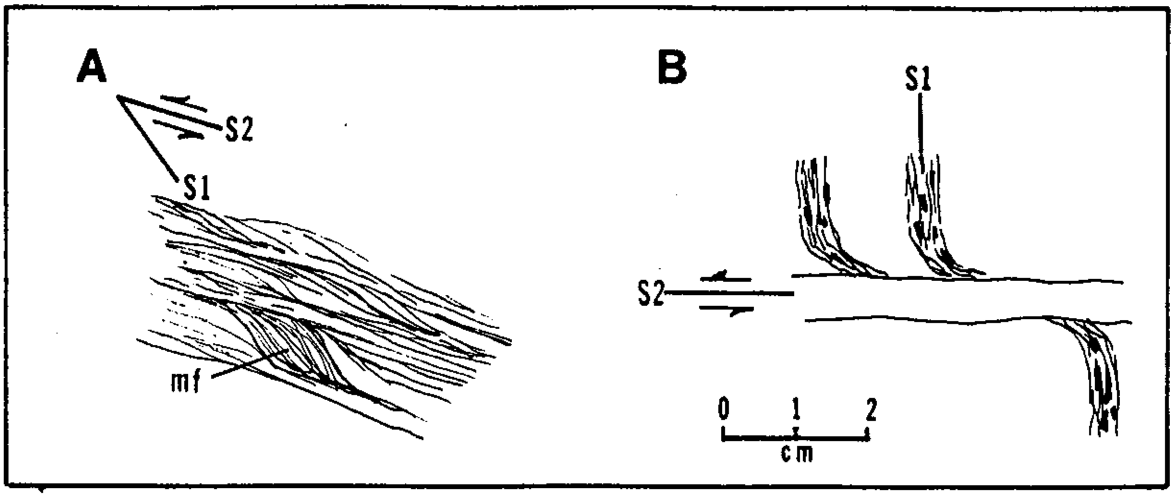


Figure 4-2. Sketches showing S1/S2 relationships in the hanging wall. A. Typical fabric in the Wissahickon schist. S1 is observed within muscovite fish (mf) and between S2 slip surfaces. B. Gneissic banding (S1) is offset by S2 ductile deformation zones.

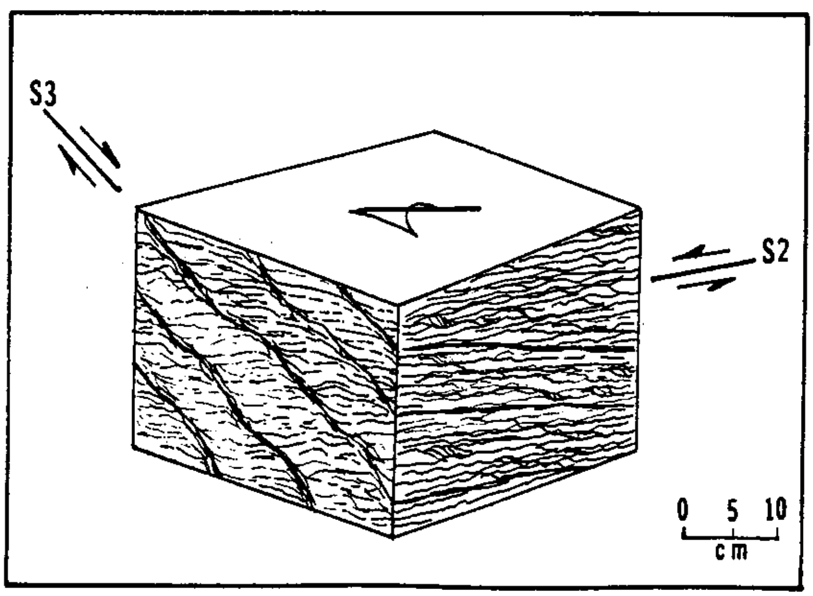


Figure 4-3. Generalized block diagram showing S2/S3 relationships in the footwall.

prograde metamorphism in both the hanging wall and footwall.

A metamorphic foliation (S1) is the oldest recognizable fabric in rocks where S0 is not preserved. In the hanging wall gneisses the foliation is a gneissic layering; in the Wissahickon it is a schistosity (S1h) defined by aligned middle/upper amphibolite facies minerals usch as muscovite, kyanite, and/or staurolite. In the footwall, S1f schistosity is defined by muscovite, chlorite, and/or biotite.

In both the hanging wall and footwall, S1 is cut by later planar fabrics. Deformation of S1h in the Wissahickon occurred along shear surfaces (S2h). Deflection and crenulation of S1h by slip along S2h produced asymmetric muscovite porphyroclasts (mica fish); porphyroclasts of garnet, kyanite, staurolite, and feldspar with asymmetric recrystallized tails; zones of grain size reduction; and quartz ribbons. The shear fabric forms a small acute angle with S1h (Fig. 4-2A). Consequently, S1h is commonly difficult to detect mesoscopically. In the hanging wall gneisses, S2h is defined by thin (0.5-2 cm thick) ductile deformation zones that crosscut gneissic layering, commonly at high angles. The older fabric is deflected adjacent to these zones but largely unaffected between them (Fig. 4-2B).

Several lines of evidence suggest that the shearing in the hanging wall resulting in S2h occurred under amphibolite facies conditions. In the gneisses, the presence of epidote and sphene in S2h zones indicates temperatures in the greenschist/ amphibolite transition to lower amphibolite facies (400-450o C). Dynamic recrystallization of feldspar in both lithologies suggests a similar temperature range (Tullis and Yund, 1985). Finally, garnet, hornblende, and kyanite show no significant retrograde reactions associated with the S2h fabric. Thus, although D2 occurred after peak metamorphism, the rocks were (still?) warm. Strong retrogression has occurred locally in hanging wall rocks close to and within the fault zone. The degree of retrogression decreases sharply to the south away from the fault zone. The association of retrograde reactions with the fault zone indicates post-D2 deformation at lower temperatures in the presence of fluids.

In the footwall, two planar fabrics cut S1f. These rocks record high shear strains which, in some areas, obliterates S1f. Outcrops of pelitic units in the footwall exhibit a wavy schistosity. In thin section, the dominant penetrative schistosity (S2f), defined by oriented muscovite, chlorite, and a strong grain-shape fabric in quartzofeldspathic layers, is seen to cross cut an older foliation defined by chlorite and muscovite. The older foliation (S1f), is deflected or dragged into a new orientation that resulted from shear deformation of the previously penetratively foliated rocks (Fig. 4-3). The waviness seen mesoscopically is the result of this shear strain. Asymmetry of S1f/S2f fabrics records a top to the northwest sense of motion (Fig. 4-3).

The second cross-cutting planar fabric (S3f) in the footwall consists of generally south-dipping shear surfaces spaced at intervals of several centimeters (Fig. 4-3). These exhibit clear S-relationships (Lister and Snoke, 1984) with respect to S2f.

S2f (s-surfaces) are deformed and crenulated by S3f (c-surfaces) (Fig. 4-3). S3f is best developed adjacent to the map trace of the Cream Valley fault zone and, although it varies in intensity along stroke, is locally sufficiently penetrative to produce button schists (Lister and Snoke, 1984). S3f shows oblique normal slip with a dextral component, or top down and to the southwest. This younger fabric gradually disappears northward farther into the footwall until only S2f is present.

Each of the planar fabrics in the footwall is defined by minerals (chlorite, muscovite, +/- biotite) of the garnet zone of the greenschist facies. However, several features suggest a polymetamorphic history. Two types of garnets occur. Millimeter scale grains with euhedral rims around anhedral cores indicate two periods of garnet growth. Since the rims overgrow S2f, the second period of garnet growth must either coincide with or post-date D2. Larger, anhedral to subhedral grains deflect S2f and often show some retrograde reaction to chlorite. These are interpreted to represent garnets which grew during prograde metamorphism and were later retrograded and resorbed either during D2 or D3. Chlorite grains may be found which both parellel S2f and appear to overgrow it. Particularly in quartzo-feldspathic layers, euhedral chlorite grains with random orientation have acted to pin grain boundaries during annealing. These features suggest either two metamorphic events both at upper greenschist conditions, or that the rocks remained at greenschist facies conditions throughout the later (D3) deformation history.

Folds. Fold style and cross cutting relationships show that multiple fold sets have formed about similarly oriented axes in both the hanging wall and footwall. In the hanging wall, at least two generations of folds deform the S1h foliation and thus formed after the metamorphic peak. The older set developed in the ductile regime, the younger near the brittle-ductile transition. The older folds are similar style, northwest vergent, overturned to recumbent, nappe-like structures and are best observed in the Wissahickon schist. Since these structures commonly fold but are also locally cut by S2h, they are labeled F2h and are interpreted to have occurred late in D2. Younger folds in the Wissahickon are tight to open, upright to slightly northwest vergent structures of a more brittle style than F2h. Since these structures fold F2h, they are labeled F3h. However, because the later folds formed about axes with similar orientations to F2h (20 S60W), the two generations are often difficult to distinguish in the field.

In the hanging wall gneisses, fold relationships are less clear. The dominant structures visible in outcrop are similar to the F2h folds in the Wissahickon in terms of orientation and their nappe-like geometry. However, they are parallel rather than similar in style suggesting less ductile behavior. The difference in foldstyle and apparent ductility may be associated with the difference in competence expected in gneisses and schists under similar conditions of temperature and pressure. In addition to these, there are buckle folds that are open and upright and commonly associated with brittle features such as offset lithologic contacts and S2h zones. No crosscutting relationships have been observed to determine if these structures represent two different fold sets. In the absence of any evidence to the contrary, the folds exhibiting nappe geometry are correlated with F2h; the buckle folds associated with brittle features are considered synchronous with F3h.

Two post-F1 fold sets also occur in the footwall. The older are northwest vergent, close to tight, asymmetric folds (F2f). These are the most common folds observed in outcrop. An incipient axial plane schistosity defined by biotite or chlorite is present in some F2f folds suggesting that this deformation occurred while the rocks were warm but not undergoing prograde metamorphism. These structures are cut by asymmetric, northwest verging, sharp-crested, kink-like folds. They most commonly occur on the shallow limbs of asymmetric F2f. Planar, 1-5 cm thick, southeast dipping shear zones that are axial planar to F3f commonly attenuate the lower limb of F3f suggesting that these folds formed by drag. The resulting asymmetry indicates transport toward the northwest.

Kinematic Interpretation

D1 is characterized by isoclinal folding and was accompanied by prograde regional metamorphism in both the hanging wall and footwall. Age data bearing on this metamorphic event suggest that it is associated with the Taconic orogeny. U/Pb analyses of zircons from the Grenville age gneisses yield a lower intercept age of 450 Ma (Grauert and others, 1973, 1974). Lapham and Bassett (1964) obtained K/Ar ages from micas in the Wissahickon

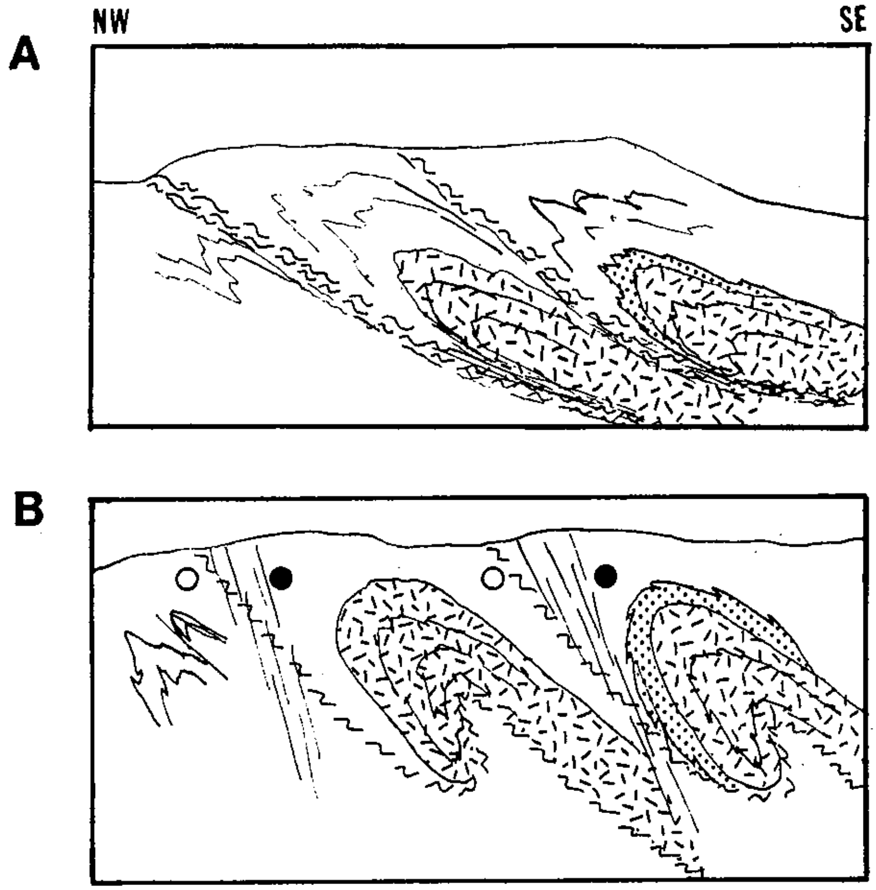


Figure 4-4. Generalized and schematic cross sectional cartoons. Random dash pattern = Grenville age gneiss; dots = lower Glenarm Series; unpatterned = schists and phyllites. A. Nappe-forming stage of D2 deformation. B. Strike-slip phase of D3. Blocks have been internally folded into antiforms with S3 foliations formed between blocks. Slip is occurring on S3 zones (open circle moving away from the observer, solid circle moving toward the observer).

ranging from 320 to 350 Ma and suggested that these represent Taconic cooling ages. 40Ar/39Ar ages from a variety of lithologies range from 465 to 360 Ma (Sutter and others, 1980).

D1 structures are rare. Isoclinal folds in compositional layering with axial planar schistosity (S1f) are observed primarily in psammitic lithologies in the footwall. Variably developed gneissic layering in the basement massifs of the hanging wall probably also developed at this time (as opposed to Grenville) based on the amphibolite facies mineral assemblage which defines it.

Northwest-directed shearing (D2) occurred after the metamorphic peak on shallow southeast-dipping surfaces. In both the hanging wall and footwall schists, movement was initially controlled by the metamorphic foliation with slip resulting in S-C fabrics. In the gneisses, strain was highly partitioned into discrete ductile deformation zones. Based on mutually cross-cutting relationships between F2h and S2h, slip gave way to folding as the dominant deformation mechanism late in D2. Second generation folds in the footwall and hanging wall (F2f and F2h) have similar orientation and geometry. The consistent northwest vergence, geometry, and the presence of overturned sections beneath the gneisses of the Woodville structure (Alcock, 1989) and Poorhouse nappe suggest that this folding event records nappe formation and emplacement (Fig. 4-4A). The change in style of deformation from shearing to nappe formation presumably occurred as the rocks were transported to shallower crustal levels where cooling resulted in an increase in competence. However, it is

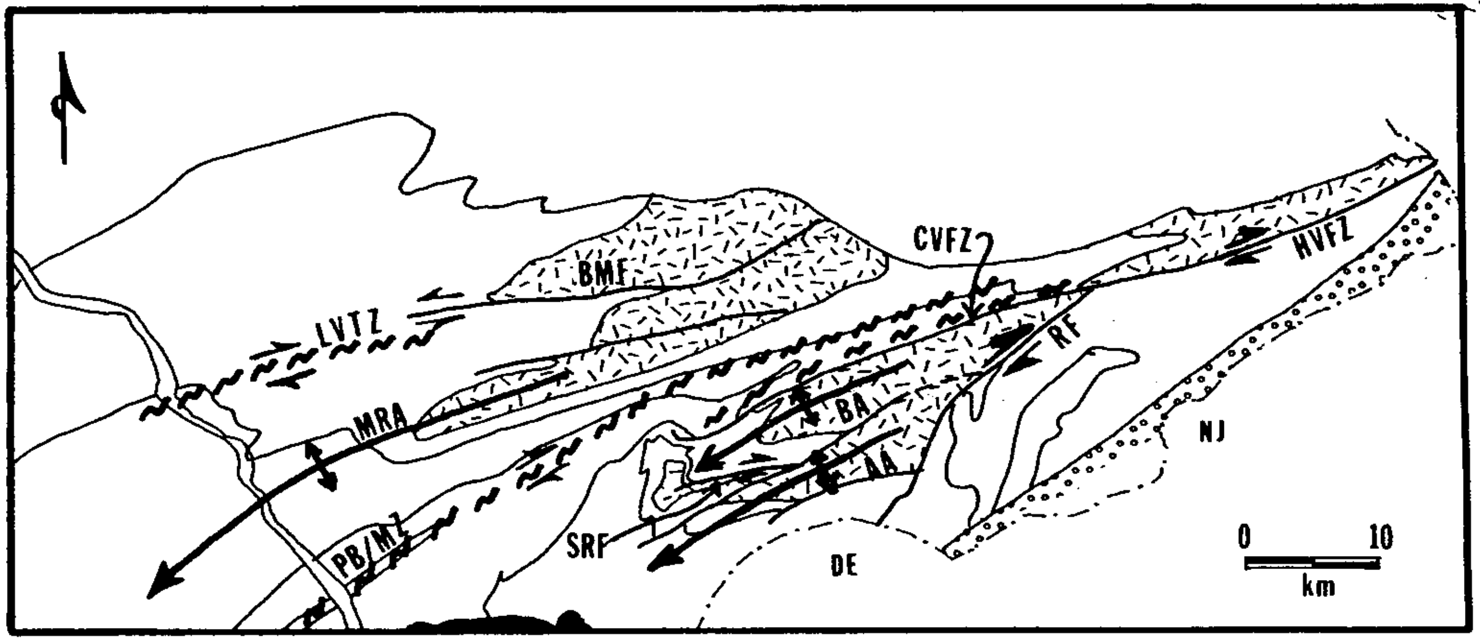
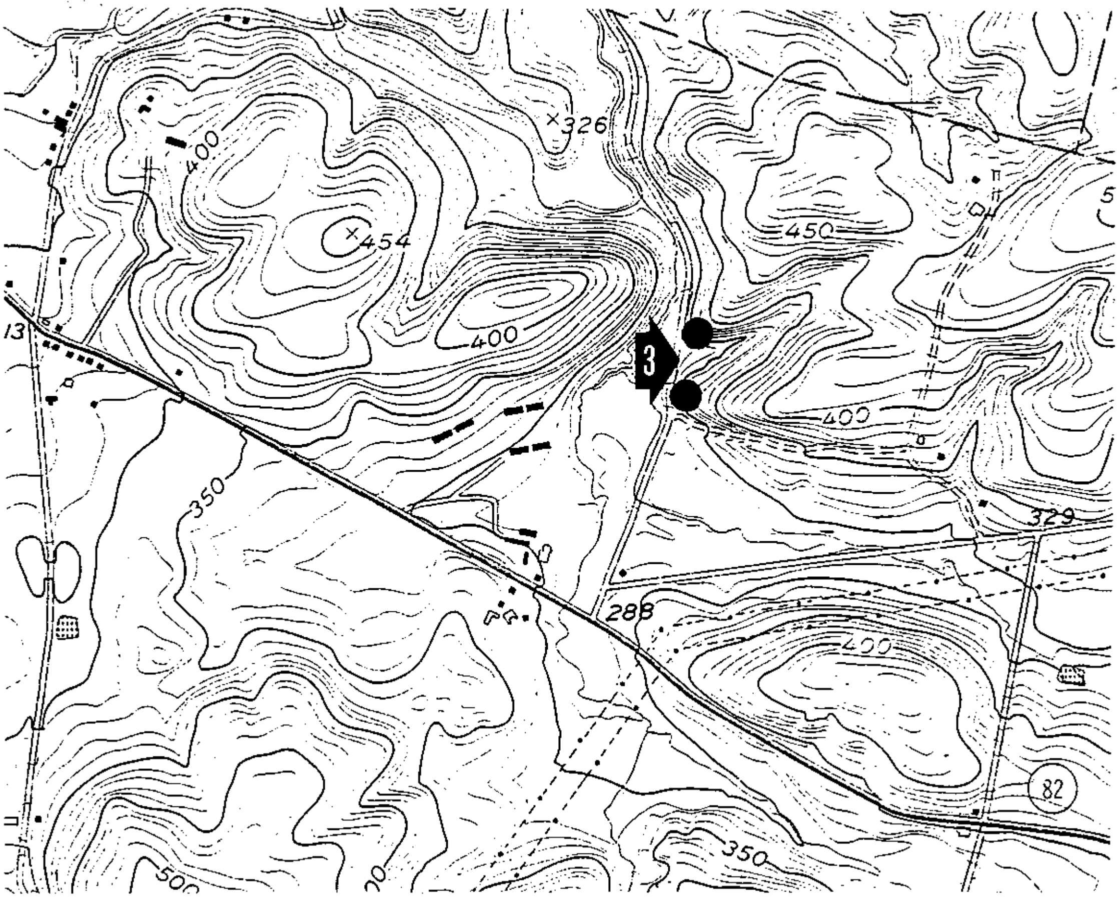


Figure 4-5. Map of structures formed or active during D3. LVTZ = Lancaster Valley Tectonite Zone, BMF = Brandywine Manor Fault, MRA = Mine Ridge antiform, PB/MZ = Peach Bottom/Martic Zone, CVFZ = Cream Valley fault zone, BA = Brandywine Antiform, AA = Avondale Antiform, SRF = Street Road Fault, RF = Rosemont Fault, and JVFZ = Huntington Valley Fault Zone. Modified from Valentino and Wiswall, 1990.

important to note that the hanging wall remained at temperatures within the amphibolite facies during D2 while the footwall rocks were cooler. Thus, the hanging wall may not have reached its present structural position relative to the footwall during D2. It seems likely that D1 and D2 represent a progressive deformation involving imbrication of the North American margin during Taconian mountain building (Wagner and others, 1991).

D3 deformation is recorded by folds (F3h,f) and shear fabrics (S3f) which deform earlier structures. In general, D3 folds are restricted to the hanging wall and psammitic lithologies in the footwall. At the macroscopic scale, the West Chester nappe was refolded during this event resulting in the present map pattern that indicates a southwest-plungng antiform (Fig. 4-4B). In the vicinity of the Cream Valley fault zone, D3 shearing was restricted primarily to the fault zone but also occurred along discrete zones in the pelitic lithologies within the footwall.

Similar relationships between early and later overprinting structures have been recognized throughout the Pennsylvania Piedmont (Valentino and Wiswall, 1991). The youngest folds and planar fabrics can be explained by a single compressive event directed at a high angle to the Taconic structural grain producing a transpressional environment. At the regional scale, post-Taconic deformation presents a picture of relative horizontal movement between crystalline-cored crustal blocks internally deformed by folding (Fig. 4-5). Between blocks, early folding produced a penetrative axial planar cleavage. Once formed, this fabric was favorably oriented with respect to the strike-parallel component of stress. Slip along this foliation produced shear fabrics. Within blocks, lithologies were not prone to slip and folding resulted. The boundaries of individual blocks appears to have been established during Taconian mountain building.



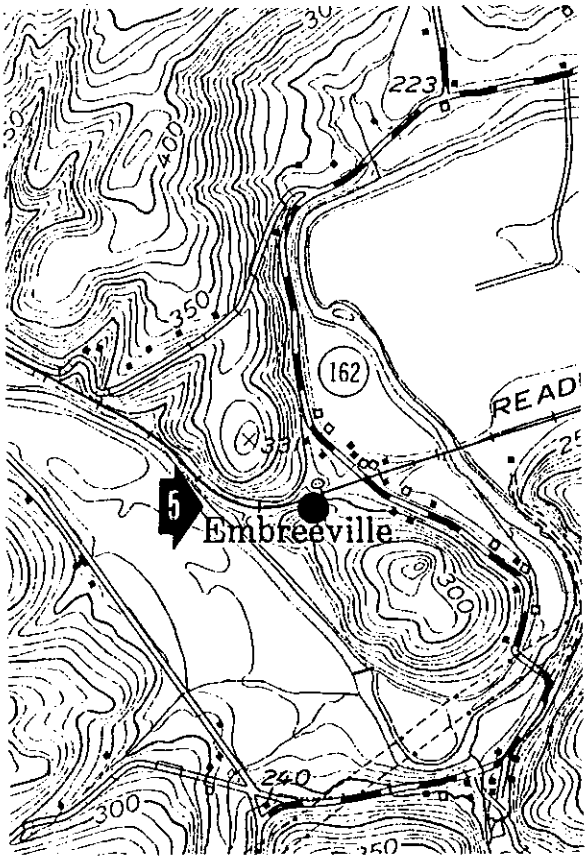
Coatesville 7.5' Q. Lat 39°54'56" Long 75°47'51"
39°54'40" 75°47'52"

STOP 3: BUCK AND DOE RUN FARM

This location provides the opportunity to compare rock fabrics in the hanging wall with those in the footwall. A small outcrop occurs in the roadbed and east bank approximately 460 m north of the entrance gate off Apple Grove Road. Although not a good exposure, this rock is similar to that which will be seen at Stop 5. In thin section, the rock is highly sheared and contains strongly retrograded staurolite.

The next outcrop occurs approximately 150 m farther to the north. Previous mapping makes no distinction between the rock here and the schists in the hanging wall; both are called Wissahickon. However, the two lithologies are quite different in terms of metamorphic grade and fabric. This rock is a fine grained schist with garnet, chlorite,muscovite, plagioclase, and quartz.

Three foliations make up the fabric (refer to figure 4-3). The dominant foliation, dipping gently to the southeast, is S2f which has almost completely overprinted S1f at this location. Careful examination may reveal small, isolated domains of S1f showing asymmetry indicating top to the northwest. S3f surfaces are observed on the north face or the outcrop. They rake across the face toward the southwest. Deflection of S2f by these surfaces indicates oblique slip with the top moving down dip and to the southwest.



Unionville 7.5' Q. Lat 39°55'47" Long 75°43'10"

STOP 5: EMBREEVILLE RAILROAD CUT

This stop is located approximately 150 meters north of PA Route 162 along the Octoraro Railway tracks in the village of Embreeville. The southwestern end of the Poorhouse nappe is approximately 1 km to the northeast along strike. The trace of the Cream Valley fault zone crosses the railroad tracks approximately 0.3 km north of this location. Stop 3 lies approximately 7 km along strike to the southwest.

The outcrop is typical of the Wissahickon schist in the hanging wall close to the fault zone. It is a coarse-grained, muscovite-quartz-plagioclase-garnet schist. Garnets are sufficiently abundant that this rock was quarried in the past. Staurolite is present and has been retrograded to varying degrees ranging from minor sericitization to complete replacement by chlorite and sericite.

The dominant foliation is S2h. Careful examination of the fabric reveals shear features such as those shown in Figure 4-2A. In addition, recrystallized tails are commonly developed around the garnets. No D3 structures have been documented at this location.

Figure 1

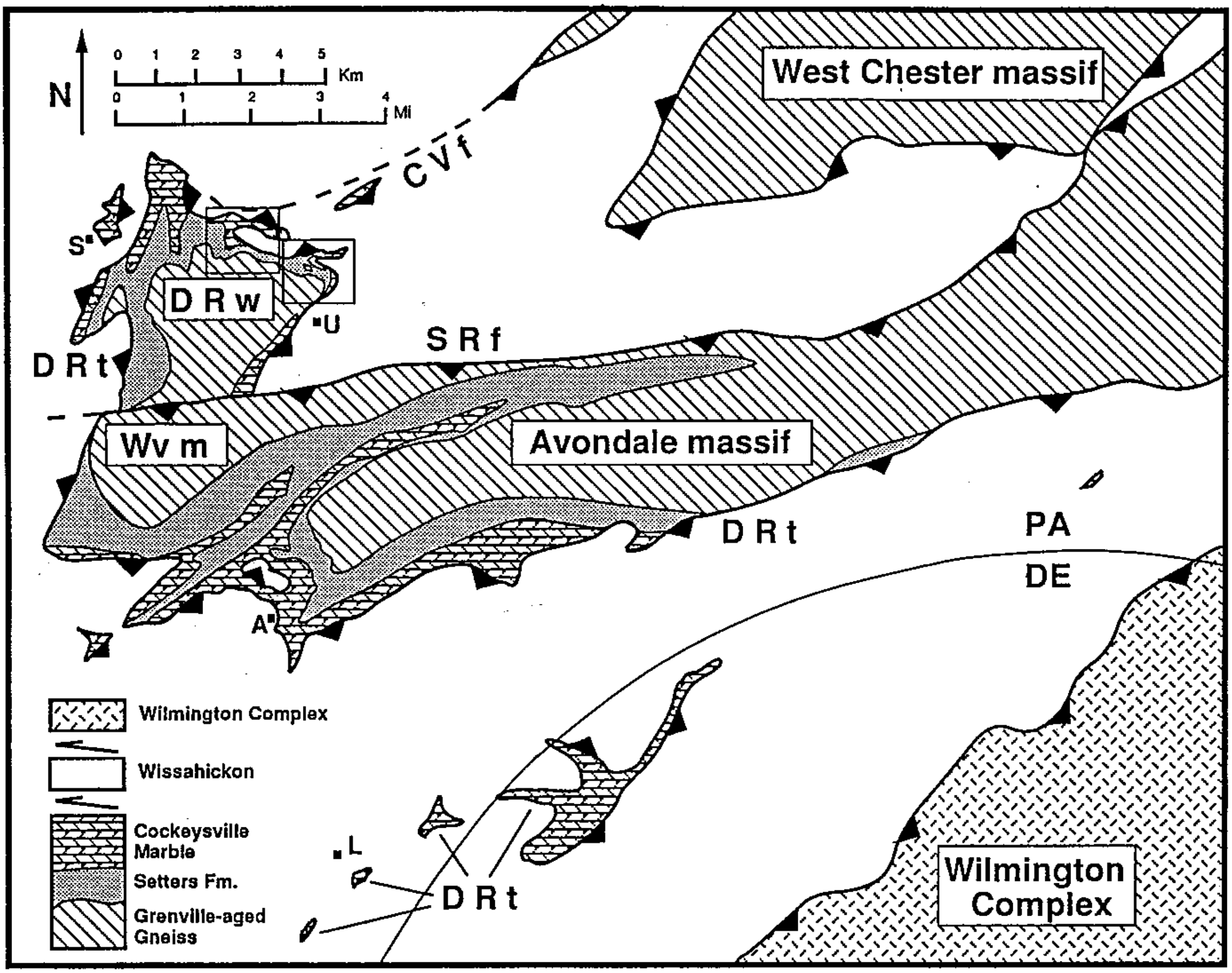


Figure 5-1: Geologic map of the Pennsylvania-Delaware Piedmont. Faults are indicated by heavy contact, hanging wall by solid triangles. Boxes at Doe Run window locate Fig. 5-5, 5-6. Abbreviations: A, Avondale; L, Landenberg; S, Springdell; U, Upland; Bq, Buck and Doe Run Farm quarries; Lq, Logan quarry; CVf, Cream Valley fault zone (dashed line is extension of fault proposed by Wiswall (Wagner et al, 1991)); DRt, Doe Run thrust; DRw, Doe Run window; SRf, Street Road fault; Wvm, Woodville massif.

STRUCTURAL DEVELOPMENT OF THE PENNSYLVANIA PIEDMONT, WEST GROVE AND COATESVILLE QUADRANGLES.

J. Alcock
Penn State University, Ogontz Campus
Abington, Pa, 19001

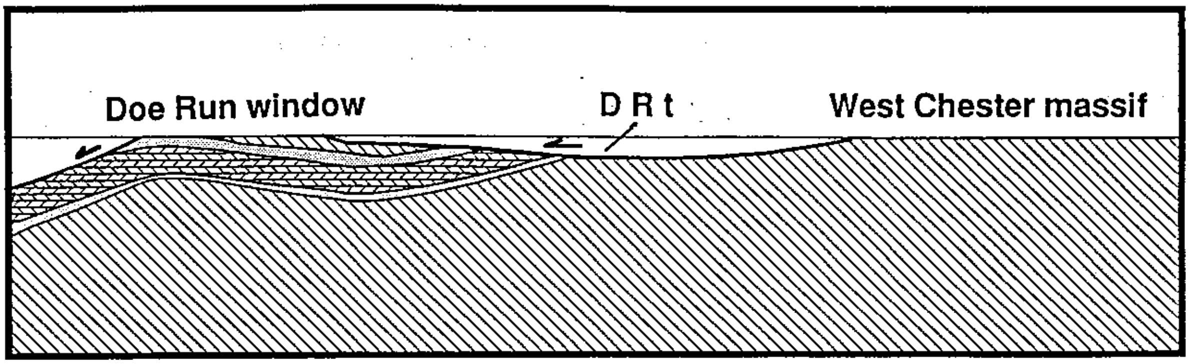
INTRODUCTION AND GEOLOGIC SETTING

Rocks now exposed at the surface in the Pennsylvania-Delaware Piedmont were deformed and metamorphosed under greenschist to granulite-facies conditions at the boundary of two colliding lithospheric plates (Crawford and Crawford, 1980; Wagner and Srogi, 1987). These rocks belong to units from the pre-Taconic North American continental margin, which lay on the subducting plate, and from the interior of the arc terrain that overrode the margin and that were accreted to the North American plate during the collision.

At least three distinct tectonic units, juxtaposed during the Taconic orogeny, can be identified in the Pennsylvania-Delaware Piedmont (Fig. 5-1). One unit consists of the pre-Taconic North American continental margin (Grenville-aged gneiss and its Late Precambrian to Early Paleozoic meta-sedimentary cover, the Setters Formation and Cockeysville Marble) . The Wilmington Complex, the infrastucture of the overriding arc, is a second unit, and the pelitic to semi-pelitic schist and gneiss of the Wissahickon Formation and its associated meta-igneous rocks are the third (Crawford and Crawford, 1980; Crawford and Mark, 1982; Wagner and Srogi, 1987; among others). (In this paper the first unit will be referred to collectively as NA margin rocks and the meta-sediments and meta-igneous rocks of the third, as the Wissahickon.)

The stratigraphic and structural relation of the Wissahickon Formation to the other units has been a source of debate for the past 75 years. Bliss and Jonas (1916) originally mapped the base of the Wissahickon Formation as the Doe Run thrust (Fig. 5-1) on the basis of an observed structural discordance and a presumed age difference between the Wissahickon and the underlying units.

SW Figure 2a NE



SE Figure 2b NW

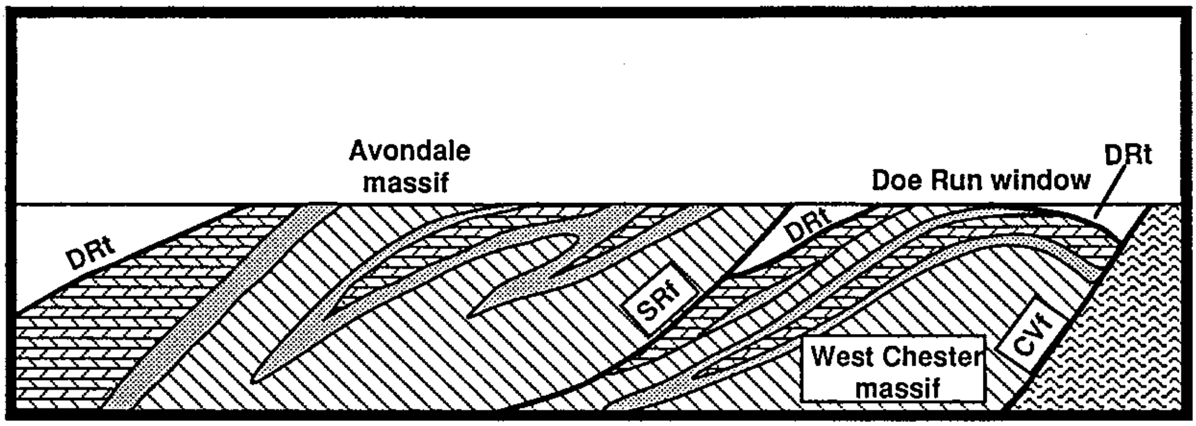


Figure 5-2: Cartoons presenting structural model. 5-2a is parallel to regional strike. 5-2b is perpendicular to strike running. Lines intersect near Upland (Fig. 5-1). In 5-2b rocks north of the Cream Valley fault zone are undifferentiated schists. Other patterns as in Fig. 5-1.

Later they re-interpreted the contact to be conformable (for example, Knopf and Jonas, 1929). Published geologic maps of the area (The West Chester Folio, Bascom and Stose, 1932 and the Pennsylvania State Geologic Map, Socolow, 1980) show the contact as conformable; and most structural models of the region accept this interpretation (for example, Bailey and Mackin, 1937; McKinstry, 1961).

More recent work has again suggested that the Wissahickon Formation may be in thrust contact with the units beneath it (Crawford and Crawford, 1980; Wagner and Srogi, 1987; Alcock, 1989). In this paper, I summarize evidence for the thrust emplacement of the Wissahickon* -- a metamorphic inversion across its contact with the Cockeysville Marble south of the Avondale massif and structural discontinuities at its base, especially in the area of the Doe Run window (Fig. 5-1). Cross-cutting relationships at the margins of the Doe Run Window constrain the timing of thrusting to post-date the formation of recumbent folds in the NA margin rocks exposed in the Avondale and Woodville massifs and in the Doe Run window. Post-thrust deformation includes additional thrusting that cuts the Doe Run thrust and two sets of relatively open folds.

Recognition of thrusting and its relative timing requires a reinterpretation of the structure of the Pennsylvania Piedmont in southern Chester County (Fig. 5-2). The Wissahickon should be treated as a relatively thin blanket that truncates and in places conceals pre-thrust structures in the units beneath it. Post-thrust deformation has resulted in uplift and erosion to expose the Grenville-aged gneiss, Setters Formation, and Cockeysville Marble in complex windows like the Avondale massif and Doe Run window.

*Note: The name, Doe Run thrust (Jonas and Bliss, 1916), is reassigned.

EVIDENCE FOR THRUSTING AT THE BASE OF THE WISSHICKON

South of the Avondale massif a metamorphic discontinuity coincides with the Wissahickon-Cockeysville Marble lithologic contact. Higher grade schist and gneiss of the Wissahickon lie

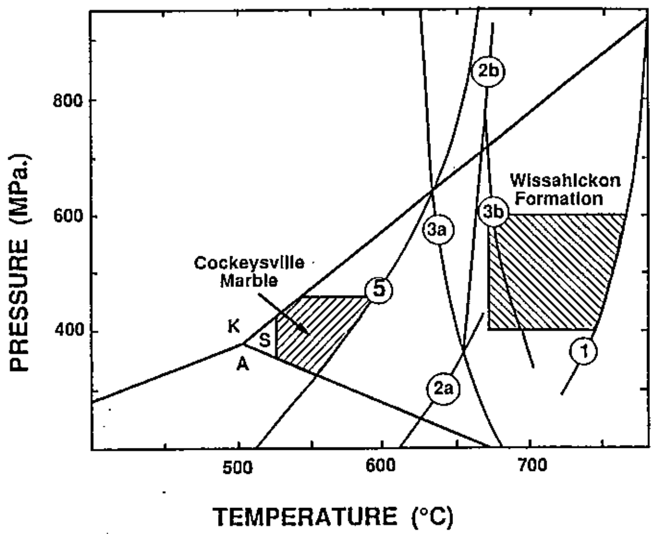


Figure 3

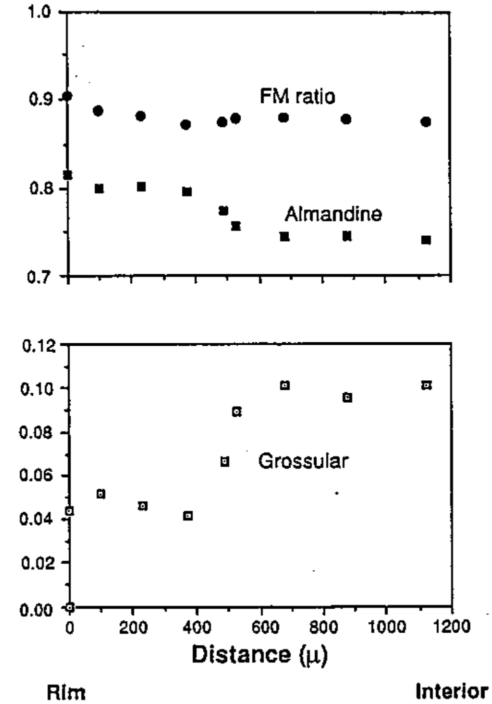


Figure 4

Figure 5-3: Presure-temeprature diagram illustrating metamorphic conditions consistent with mineral assemblages and compositions in the Wissahickon and Cockeysville Marble south of the Avondale massif.

Reactions: 1. Si + Bi + Q + Pl = Gt + more calcic Pl + Melt (LeBreton and Thompson, 1988); 2a. Mu + Q = Si + Kf + H2O (Chatterjee and Johannes, 1974); 2b. Mu + Q + Pl = Si + more calcic Pl + Melt (Kerrick, 1972); 3a. and 3b. Minimum melting of granite composition at P(H2O) = P(lith) and .5 P(lith) respectively (Kerrick, 1972); 5. Ph + Cc + Q = Tr + Kf + fluid (Hoschek, 1973). Alumino-silicate triple point after Holdaway (1971).

Figure 5-4: Profile of chemcial zoning in garnet from the Wissahickon Formation at its contact with the Cockeysville Marble at Landenberg, south of the Avondale massif. Grossular and almandine are presented mole fraction of total. Note vertical scales of graphs are different above lower grade marble. The metamorphic discontinuity cannot be documented north of the Avondale massif in the Doe Run area; however, structural evidence for thrusting at the base of the Wissahickon there suggests that the absence of the discontinuity should not be interpreted as evidence for a conformable contact in that area.

Metamorphism: The Wissahickon

Uppermost amphibolite-facies gneiss of the Wissahickon contains the assemblage biotite-sillimanite-garnet-K-feldspar-plagioclase-quartz-ilmenite $\pm$ secondary muscovite, indicating that it was metamorphosed above the second sillimanite isograd (Plank, 1988; Alcock, 1989). Garnets in migmatitic gneiss appear to have grown into zones of biotite and sillimanite suggesting a reaction such as:

$$\text{Si} + \text{Bi} + \text{Q} + \text{Na-Pl} = \text{Gt} + \text{Ca-Pl} + \text{melt}. \qquad 1$$

Textures, for example embayed biotite and plagioclase as inclusions within garnet, the separation of biotite-sillimanite-garnet restite from leucosomes with mostly plagioclase-K-feldspar-quartz, and the occurrence of euhedral microcline grains partially enclosed by garnet, document partial melting associated with garnet growth (reaction 1 going to the right). The presence K-feldspar and sillimanite, but not muscovite, in the leucosome indicates that crystallization of the melt occurred outside the stability field of muscovite (above reaction 2a or 2b, Fig. 5-3).

Experimental investigation of the second sillimanite isograd (Chatterjee and Johannes, 1974) and melting in pelites (for example, Kerrick, 1972 and LeBreton and Thompson, 1988) indicate that the temperature in the Wissahickon was greater than 650°C and probably 700-750°C (Fig. 5-3).

Garnets from the Wissahickon Formation exhibit textural and chemical zoning (Fig. 5-4) that result from changing metamorphic conditions and mineral reactions during garnet growth. Typically garnets found in uppermost amphibolite facies rocks show only a limited zoning at the rim since diffusion at high temperature can create a chemically homgeneous garnet. Rim zoning results from

The Gt-Al2SiO5-Pl-Q barometer (GASP, Newton and Hazelton, 1981; Ganguly and Saxena, 1984 and Koziol and Newton, 1988) is based on the exchange of calcium between garnet and plagioclase through the reaction

$$Gr + 2\ Al_2SiO_5 + Q = 3\ An. \qquad 4$$

Growth zoning of calcium is preserved in the interiors of garnets in the Wissahickon and calcium is also commonly exchanged between garnet rims and plagioclase during cooling. Therefore, the best estimates of pressure for peak metamorphism should come from garnet compositions that do not show late modification of the FM ratio (also caused by post metamorphic peak reactions) but which are outside the high-grossular zones of the garnet interior (Fig. 5-4).

Chemical variation of plagioclase is also pronounced. The presence of rare albite grains, often within K-feldspar, and normal zoning of plagioclase adjacent to K-feldspar indicate that the anorthite content of plagioclase was reduced by the release of Na from K-feldspar, during cooling and retrograde reactions that produced secondary muscovite. For this reason, the highest measured per cent anorthite within a thin section is considered to be most representative of plagioclase composition at peak metamorphic conditions and has been used in the GASP barometer. A pressure estimate, 500 $\pm$ 100 MPa (5 $\pm$1 Kb.), is based on a variety of points from different grains and thin sections.

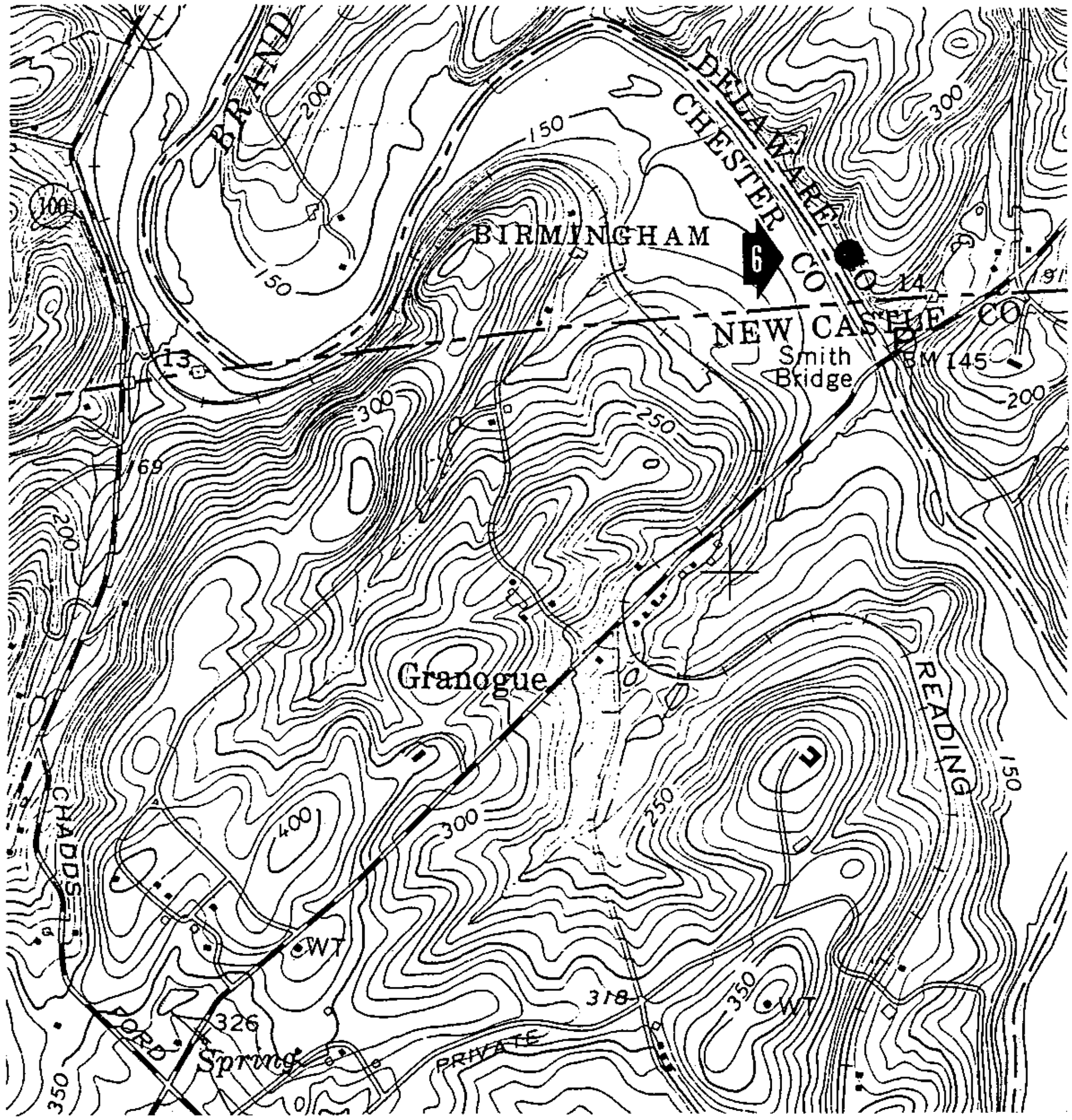
Stop description, Stop # 6 Smith Bridge, Brandywine Creek

High grade Wissahickon gneiss showing partial melting

The Cockeysville Marble

The Cockeysville Marble is thought to be derived from an impure dolomite-calcite limestone. Metamorphic reactions formed phlogopite, tremolite, diopside, additional calcite and K-feldspar. Layers richer in aluminum and perhaps sodium contain plagioclase, scapolite, and/or epidote minerals in addition to those listed above.

Of particular interest are two low variance assemblages belonging to the $MgO-CaO-SiO_2-KAlO_2-H_2O-CO_2$ system. The



Wilmington North 7.5' Q. Lat 39°50'13" Long 75°34'45"

assemblages, dolomite-calcite-quartz-phlogopite-tremolite and calcite-quartz-K-feldspar-phlogopite-tremolite, are found in the Cockeysville Marble in a single outcrop south of Landenberg (Fig. 5-1) within 15 m. of high-grade Wissahickon gneiss and at several other localities in close proximity to high-grade Wissahickon south of the Avondale massif (Alcock, 1989). In the model system, these assemblages define an isobaric invariant point at the intersection of the reactions

$$\text{Ph} + \text{Cc} + \text{Q} = \text{Tr} + \text{Kf} + \text{Fluid} \qquad 5$$

$$\text{Do} + \text{Q} + H_2O = \text{Tr} + \text{Cc} + CO_2 \qquad 6$$

The pressure is not constrained by the mineral assemblages in the Marble but is estimated to be ≈ 400 MPa based on the continued presence of sillimanite in the overlying Wissahickon and the results of geobarometry discussed above. Experiments place the isobaric invariant point at T ≈ 575°C (Eggert and Kerrick, 1981; Hoschek, 1973, Alcock, 1989). The maximum temperature for reaction 5 is similar (Fig. 5-3).

Minor impurities in the minerals of the marbles (for example, <1 weight% Fe and Al in tremolite) are assumed to have not affected reaction temperatures significantly. Modelling the effect of the impurities on the reactions using thermodynamic calculations with an ideal solution model (Ferry, 1976) indicates a shift in the invariant point to only slightly higher temperature (≈ 585°C).

The differences in metamorphic grade of the Wissahickon (peak T ≈ 700°C) and Cockeysville Marble (peak T ≈ 575°C) indicate a metamorphic inversion across their contact. The discontinuity is interpreted to be evidence that the Wissahickon in this area was metamorphosed under uppermost amphibolite facies conditions and subsequently thrust over the Cockeysville Marble.

Structural Discontinuity

The metamorphic inversion across the Wissahickon-Cockeysville Marble contact cannot be documented in area of the Doe Run window;although field evidence of structural discontinuities at the base of the Wissahickon support the interpretation that it was emplaced along a region-wide thrust (Alcock, 1989). The

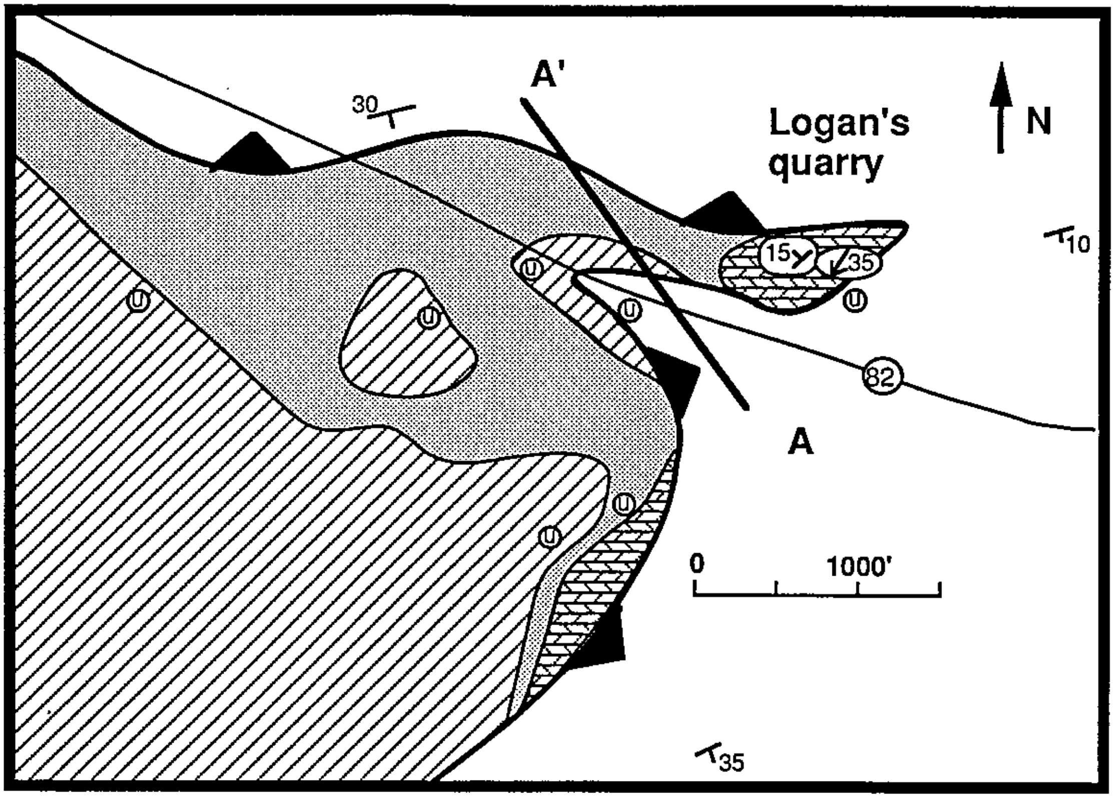


Figure 5a

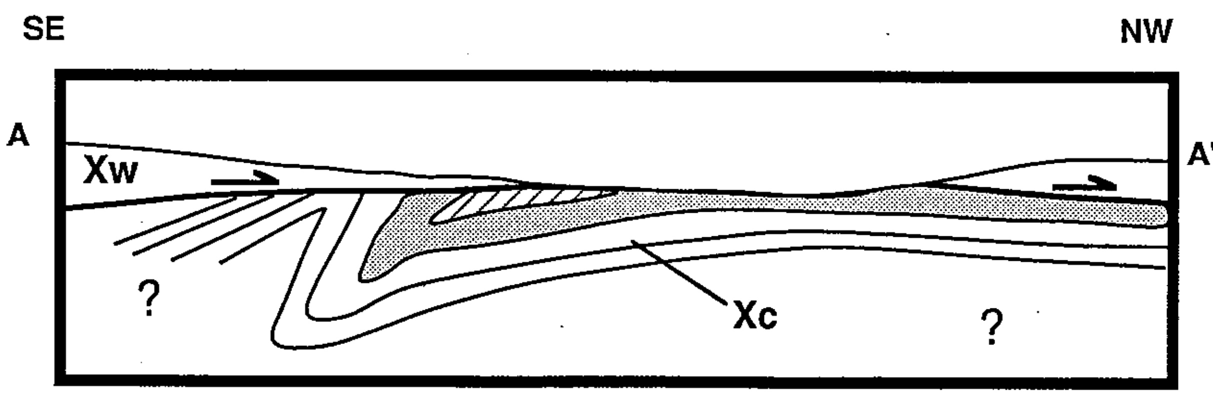


Figure 5b

Figure 5-5: Detail of geologic map in area of Logan's quarry with schematic cross-section. Legend is same as in Fig. 5-1 except that Cockeysville Marble is identified as Xc, Wissahickon as Xw, and "u" indicates structurally higher unit where contact relations are clearly defined.

Wissahickon in this area is at staurolite-kyanite grade indicating peak temperature between 500 and 650°C, a temperature range consistent with the mineral assemblages observed in the Marble, for example dolomite-calcite-quartz-phologpite ±K-feldspar. However, the absence of an observed metamorphic inversion does not preclude a thrust contact. The broad range of temperature consistent with the mineral assemblages found in the two units might prevent recognition of an existing discontinuity.

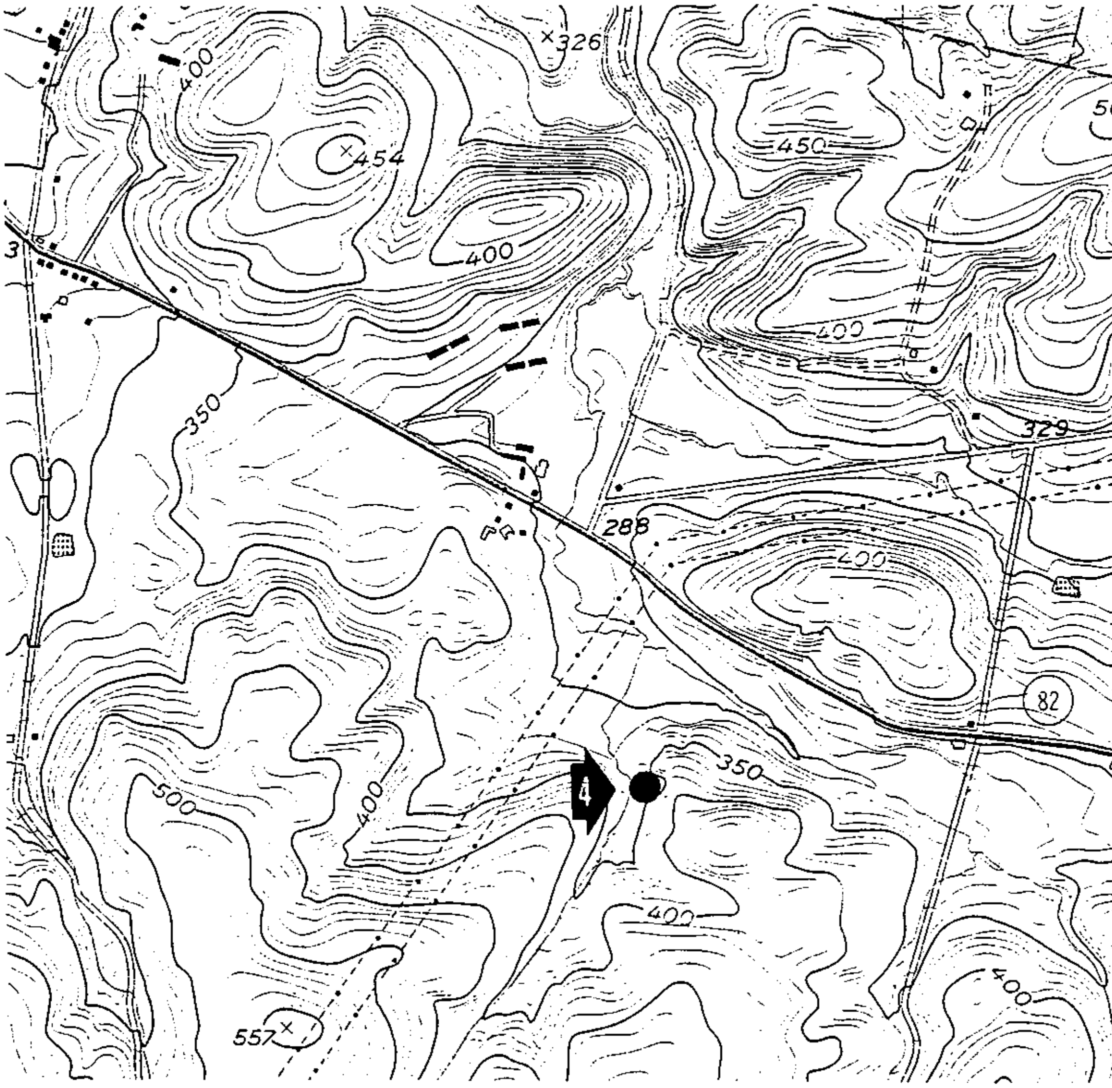
Structurally, the Wissahickon appears to lie above and crosscut an overturned stratigraphic sequence in the NA margin rocks at a number of localities around the window. This relationship has been observed south of Springdell (Fig. 5-1) where the Wissahickon is above overturned Grenville-aged gneiss, Setters Formations and Cockeysville Marble. At Logan's quarry large recumbent folds, exposed in Cockeysville Marble, are truncated by the Wissahickon that has apparently cut across the axial planes of the folds (Fig. 5-1, 5-5). Preliminary analysis of ground magnetic data supports this interpretation, suggesting that the Wissahickon lies above the NA margin rocks both north and southeast of the quarry (Alcock, unpublished data). Structural models assuming a conformable Wissahickon have it passing under the marble on the north side of the quarry (McKinstry, 1961; Mackin, 1962).

Stop Description, Stop # 4, Buck and Doe Run Farm quarry (Fig. 5-6).

Approx. .3 mi east of Apple Grove Rd on Rte. 82. Wissahickon is exposed in the road banks along Rte. 82. An old farm road leads south from Rte. 82, across creek and into woods. Grenville-aged gneiss is found in the woods. Cockeysville Marble and Setters Formation are exposed in 3 quarries along the old road.

This locality has been chosen as an example of the basic structural features common to the Doe Run window, the overturned stratigraphy in the NA margin rocks and discordance of the Wissahickon to structures in those rocks, and to illustrate the difficulty of defining structural relations in this area.

Float of mylonitic Grenville-aged gneiss, is common in the woods, south of and structurally above the rocks in the quarries.



Coatesville 7.5' Q. Lat 39°54'09" Long 75°47'55"

In the south and east walls of the southern quarry, Setters Formation is exposed above the Cockeysville Marble. (The contact can be reached in the southeast corner.) The inverted stratigraphic sequence of Cockeysville Marble beneath the Setters Formation and Grenville-aged gneiss suggests that the quarry exposes the lower limb of a recumbent fold.

Small, open folds in the Setters Formation-Cockeysville Marble contact with approximate E-W axes can be seen in the SE wall of the quarry. These may be associated with the dominant regional N60-70E trend. Turning and looking to the north, one can see that the marble forms the core of antiform trending N10°W with a nearly horizontal axis. This and similar folds create the irregular map pattern around the north margin of the Doe Run window (Fig. 5-1). The presence of marble beneath Setters where the creek crosses Rte. 82 (Fig. 5-6), directly along strike of the axis of the antiform seen in the quarry, suggests that the creek crosses Rte. 82 at the axis of the antiform. The location of the antiform's axis and the NE dip of the Setters on the east side of the antiform seems to indicate that the rocks exposed in the quarry lie beneath Wissahickon schist exposed in the ridge along Rte. 82 north and east of the quarry. If the Wissahickon is conformable, then it should lie beneath the Cockeysville Marble not above the Setters Formation.

Along Rte 82 a mylonitic schist to augen gneiss and a coarser grained muscovite-garnet-staurolite schist are found as float and badly weathered outcrop(?). (Please be careful on Rte 82. Large trucks are common.) The coarser-grained garnet-rich schist belongs to the Wissahickon Formation. It is difficult to identify the formation of the mylonitic schist in the field. However, in thin section, microcline which the author has not found in the Wissahickon at any other locality in the Doe Run area, can be seen to be a significant constituent. Microcline is a common and important mineral in both the Grenville-aged gneiss and Setters Formation. For this reason the mylonites are thought to belong to one of these units.

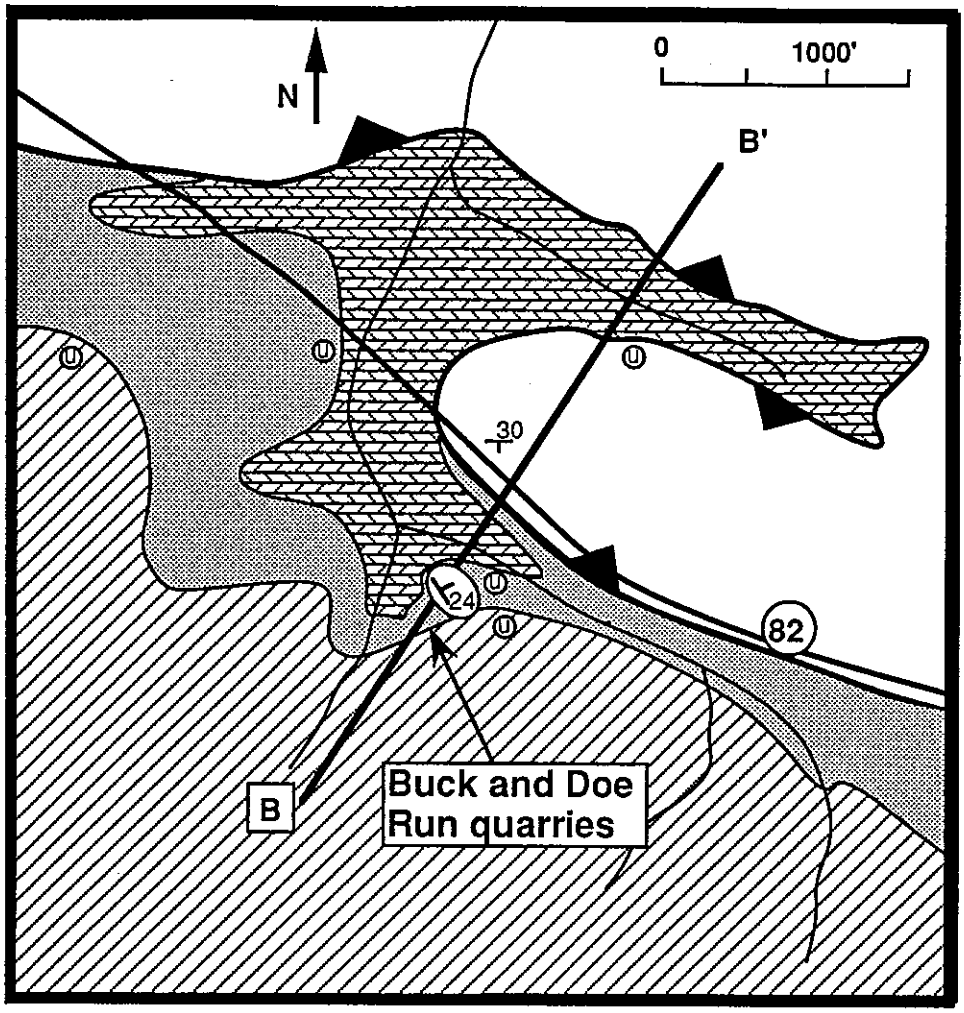


Figure 6a

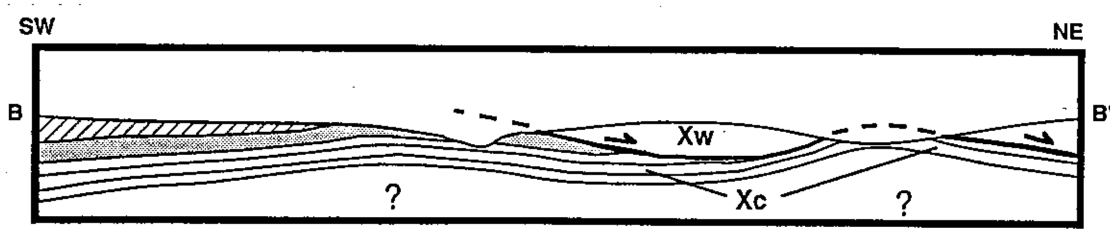


Figure 6b

Figure 5-6: Detail of geologic map in area of Buck and Doe Run quarries (Stop #4) and schematic cross-section. Legend is same as in Figure 1 except that Cockeysville Marble is identified as Xc, Wissahickon as Xw, and "u" indicates structurally higher unit where contact relations are clearly defined.

The foliation in weathered Wissahickon schist found in the road bank beneath high-tension power line strikes N90°E, dipping 30°S which is highly discordant to structures in the quarry. While the strike and dip at this location may be altered by downhill slumping, it is similar to the most common orientation of foliation in the Wissahickon (N60-70°E, 30°SE) in this area.

Two interpretations may be made of observations: 1. The Wissahickon lies above NA margin rocks. This would seem to be consistent with the observed structures in the quarry and the apparent continuation of the antiform observed in the quarry across Rte 82. Or, 2. The Wissahickon passes under Cockeysville Marble as would seem to be indicated by the strike and dip of its foliation. I prefer the former interpretation for the following reasons: 1. Digging into the road bank it is possible to find the garnet schist overlying the augen gneiss, but separated from it by a 5-10 cm clay layer. 2. Marble passes under the Wissahickon on the north slope of the ridge along Rte. 82 (Fig 5-6a). The simplest model joining the marble on both sides of the ridge is a gentle synform dropping the marble below a structurally higher Wissahickon (Fig. 5-6b). 3. The N-S trending antiform is thought to have formed during the latest period of folding in this area. This suggests that the axis of the antiform should be straight which would place the Wissahickon to its east and, therefore, above the marble and Setters exposed along the trend of the axis where the creek crosses Rte. 82. 4. The contact between the Setters Formation and Grenville-aged gneiss both east and west of the quarry (Fig. 5-6) is suggestive of synforms. For the Setters and Cockeysville to pass over the Wissahickon these would have to be an antiforms.

Additional observations of structural relations at Upland and Springdell near the Doe Run window (Fig. 5-1) allow a further test of the discordant thrust model presented here and previous models that assumed a concordant Wissahickon (McKinstry, 1961; Mackin, 1962). Field observations and well logs indicate that the Wissahickon lies above the Cockeysville Marble at Upland, consistent with its emplacement along a thrust. However, in the

Mackin model, the Wissahickon at Upland is the lower limb of a recumbent fold and should pass beneath the NA margin rocks as it plunges to the SW. At Springdell Cockeysville Marble is exposed in two moderate sized quarries between exposures of Wissahickon. The model presented here predicts that the Wissahickon would lie above the Marble on both sides of a N-S trending antiform. In contrast, both McKinstry (1961) and Mackin (1962) have the Wissahickon passing below the Marble on the east side of the quarries in the lower limb of a tight recumbent synform, verging to the west. A single contact with Wissahickon above Cockeysville Marble is exposed in the west wall of the quarries. Field observations including a gentle easterly dip in the Marble in the east wall of the quarries, lenses of Wissahickon in the uppermost layers of the Marble at the NE end of the quarries, and ground magnetic data consistent with a shallow, east-dipping schist-marble contact, all suggest that the Wissahickon lies above the Marble on the east as well.

POST-THRUST METAMORPHISM AND DEFORMATION

At Avondale on the south margin of the Avondale massif (Fig. 5-1), the Wissahickon directly above the Cockeysville Marble preserves evidence of a post-thrust high pressure metamorphism (Alcock, 1989). Kyanite and staurolite grow in the rocks that were originally at sillimanite grade (though probably not above the stablility of muscovite); and garnets show a marked increase in grossular content at their rims. GASP geobarometry (Koziol and Newton, 1988) estimates a pressure increase from 450 MPa (4.5 Kb.) to 650 MPa (6.5 Kb.) based on near rim and rim compositions of garnet. The increase is consistent with the replacement of sillimanite by kyanite and the growth of staurolite. Evidence of a similar relatively late pressure increase is reported for the Wissahickon at Garnet Mine Rd., also south of the Avondale massif (just to the east of Fig. 5-1) (Crawford and Mark, 1982).

Mineral assemblages in the staurolite-kyanite grade schists at the Doe Run window do not show a similar overprint. However, garnets from this area also show a distinct increase in grossular content at or near the rim suggestive of a significant pressure increase (Alcock, 1989) . GASP barometry estimates a pressure increase from 4.5 to 8.5 MPa. (4.5-8.5 Kb.). During the Taconic orogeny high-pressure metamorphism (to 1000 MPa.) also affected Grenville-aged gneiss of the West Chester massif now exposed at surface approximate 20 km. along strike from the Doe Run window (Wagner and Crawford, 1975; Wagner and Srogi, 1987). It seems likely that the same high pressure event affected the both units, and that the high pressure metamorphism affecting the Doe Run area, therefore, was a Taconic event.

The late pressure increase is interpreted to result from the stacking of blocks containing both NA margin rocks and the Wissahickon along NW directed thrusts. The Street Rd. fault (Fig. 5-1), for example, places the NA margin rocks of the Avondale massif and the Wissahickon to its south above the Wissahickon to the north of the fault (Fig. 5-2b). An additional fault not identified in the field may lie between Landenberg and Avondale (Fig. 5-1) since there is no evidence of late higher pressure metamorphism affecting the Wissahickon at Landenberg.

As can be seen in the regional map pattern the Doe Run thrust is also deformed by two sets of folds. The axes of the dominant folds are oriented at approximately S60W, most commonly plunging at about 10-20°. These folds bring the West Chester and Avondale massifs to the surface. The axes of the second set are more nearly horizontal and trend N-S, like the fold seen at Stop .... As map scale folds these are limited to the Doe Run area and appear to create the irregular northern margin of the window through interference with S60W trending folds.

CONCLUSION

Evidence for metamorphic inversion at the base of the Wissahickon in the area south of the Avondale massif and for structural discontinuities at the base of the Wissahickon around the Doe Run window are interpreted to reflect the thrust emplacement of the Wissahickon onto NA margin rocks during the Taconic orogeny. Crosscutting relationships in the Doe Run window show the thrust is discordant to recumbent folds in the NA margin rocks and thus thrusting post-dates a period of recumbent folding that affected these rocks.

Recognition of the Doe Run thrust and its discordance to earlier recumbent folding, requires that regional structural models based on
down-plunge projections be abandoned. The Wissahickon is considered to lie as a relatively thin blanket in the hanging wall of the thrust above NA margin rocks (Fig. 5-2a). Presently, the thrust dips gently to the SW with local irregularities created by two sets of post-thrust folds with axes that trend ≈ N60E and ≈ N-S. Originally, however, the Wissahickon cut up-section from east to west since it lies directly above Grenville-aged gneiss in the east and above the Setters Formation and Cockeysville Marble in the west.

Later thrusts and folds deformed the Doe Run thrust. Faults like the Street Road fault broke the Doe Run thrust and emplaced the NA margin rocks of the Avondale massif above the Wissahickon (Fig. 5-2b). This second period of thrusting resulted in high-pressure metamorphism of the footwall rocks. Folds along S60W and N-S axes, uplift and erosion resulted in the formation of windows. Most significant uplift has occurred along an antiform with a S60W axis through the West Chester massif and Doe Run window.

GEOLOGY OF THE MILL CREEK DOME

Kenneth D. Woodruff and Margaret O. Plank
Delaware Geological Survey
Newark, DE

INTRODUCTION

The effect of rapid population and housing growth in the Delaware Piedmont with accompanying loss of recharge to ground-water supplies has become a concern to both State and County planning officials. In particular, the Hockessin-Yorklyn and Pleasant Hill valleys have experienced intense development in the past five years. These valleys are underlain by marble of the Cockeysville Formation which is a major source of ground water for both public and private wells.

To address these concerns the Water Resources Agency for New Castle County and the Delaware Department of Natural Resources and Environmental Control requested that the Delaware Geological Survey (DGS) undertake a two year study of the Cockeysville Formation. Specific goals were to (1) refine the existing geologic mapping of the Cockeysville, (2) determine the amount of ground water available from the unit on a sustained basis, and (3) determine the theoretical loss of additional recharge on the ground-water yields. The U.S. Geological Survey carried out the hydrologic investigations under a joint-funded program while the DGS provided the geologic studies and program coordination.

As a result of the mapping project new information on the geology of the Mill Creek Dome has been obtained. A

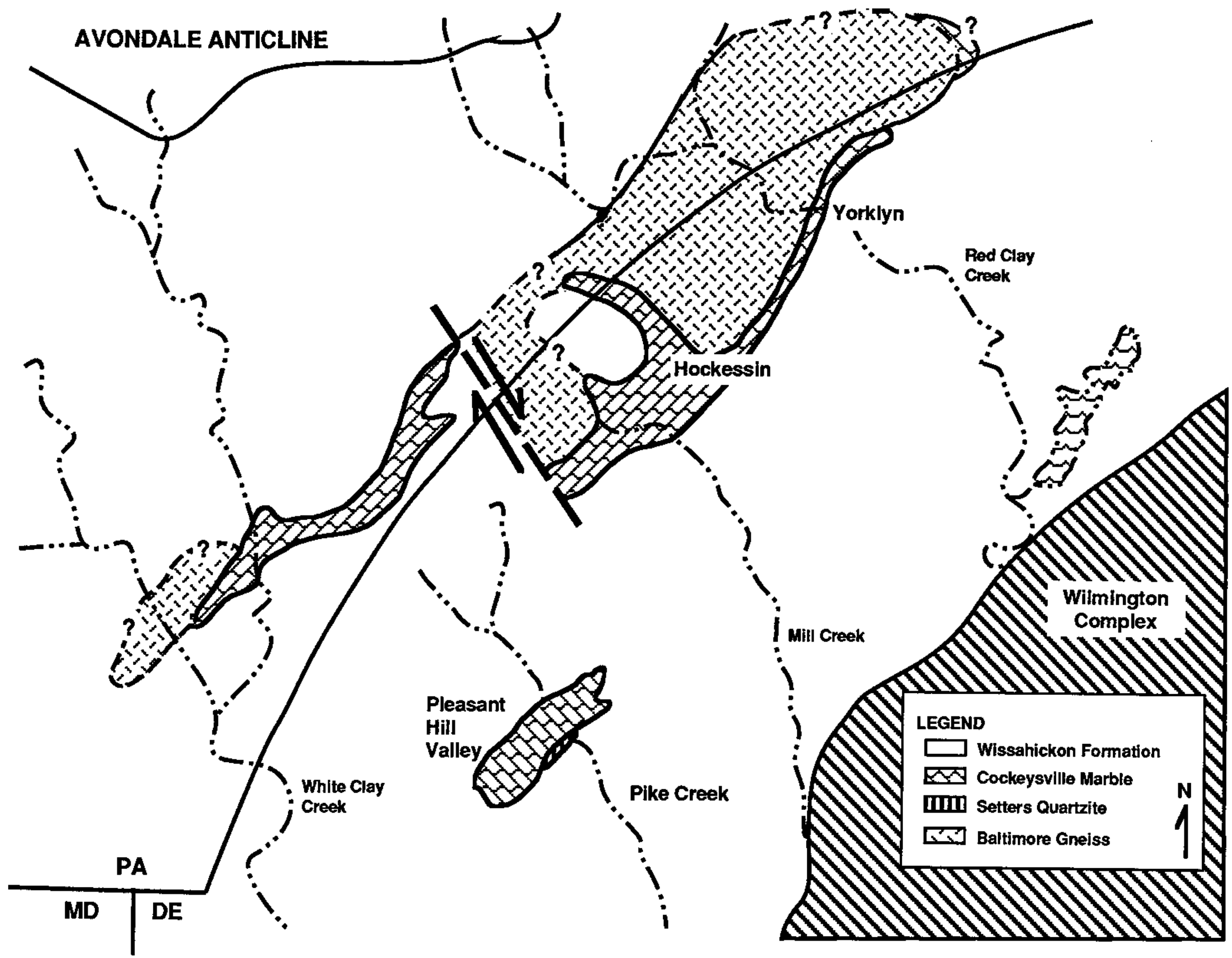


Figure 6-1. Geology of the Mill Creek Dome

northwest-striking fault southwest of Hockessin displaces the western end of the dome to the northwest into Pennsylvania (Fig. 6-1).

MILL CREEK DOME

The Mill Creek Dome is one of several anticlinal structures or nappes in southeastern Pennsylvania cored by a basement complex of Baltimore Gneiss of presumed Grenville age (~1,100 M.a.). Unlike the other anticlinal structures in the Pennsylvania Piedmont: the West Chester Prong, the Poorhouse Prong, the Avondale Anticline, and the Woodville Dome, the Mill Creek Dome was not identified until its presence was postulated by Higgins and others (1973) based on the deep magnetic low shown on the aeromagnetic map of Henderson and others (1963). After a reconnaissance field survey to the area Higgins and others named the antiform defined by the magnetic low the"Mill Creek Dome". The delay in recognition and mapping of the dome was due to poor exposures and high grade metamorphism of the Wissahickon especially in Delaware. Highly metamorphosed psammitic gneisses of the Wissahickon are almost impossible to distinguish from the Baltimore Gneiss.

The main features that distinguish the Baltimore gneiss from the Wissahickon gneiss of the Delaware Piedmont are: (1) The Baltimore Gneiss is less aluminous than the Wissahickon. Sillimanite and primary muscovite are abundant in the Wissahickon Formation, but absent from the Baltimore Gneiss. (2) Orthopyroxene, possibly a relic of Grenville granulite facies metamorphism, is found in the Baltimore gneiss but not in the Wissahickon.

Three lithologies have been identified within the Baltimore Gneiss: biotite-quartz-feldspar gneiss; biotite-hornblende gneiss; and amphibolite. Many of the biotite gneisses are light colored with discontinuous, thin, biotite-rich layers. The fabric in the gneisses grades from weakly to strongly layered and the gneiss is commonly migmatitic. Strikes and dips of the foliations are variable even within individual outcrops, except along the southeastern boundary where foliation attitudes parallel those in the Cockeysville and Wissahickon. Textures of the Baltimore Gneiss are crystalloblastic. No original sedimentary or igneous textures were recognized.

The Baltimore Gneiss cover sequence, the Glenarm Supergroup, consists from the basal unit upward of Setters Quartzite, Cockeysville Marble and the Wissahickon Formation. The Setters outcrops at one locality in the Delaware Piedmont (Fig. 6-1). At that site, Eastburn's Quarry in Pleasant Hill Valley, fine-grained gneiss that petrographically resembles descriptions of the Setters Formation in Pennsylvania and Maryland, except for the absence of muscovite, overlies the Cockeysville Marble in inverted order.

Exposures of the Cockeysville Marble are rare in Delaware because it erodes to form broad, flat valleys. As a result most of our knowledge of the marble comes from well data and from several abandoned quarries.

Indeed, the lack of surface exposure and float is usually evidence for the presence of marble. The contact with the adjacent rocks is usually marked by a topographic break that can be readily identified in the field. No internal stratigraphy has been recognized in the Cockeysville, but three lithologies have been identified from well cores; dolomite marble, calcite marble and calc-schist. The dolomite marble is the dominant lithology and makes up about 90% of the Cockeysville in Delaware. Many dolomites show minor retrograding to serpentine.

Calcite marble occurs as thin irregular layers within the dolomite marble and consists chiefly of calcite and diopside with coarse phlogopite concentrated at the contacts between the two rock types. The absence of dolomite in these layers may be due to the breakdown of dolomite in the presence of water-rich fluids. Alcock (1989) suggests the following dedolomitization reactions:

3 dolomite + K-feldspar + H_2O = phlogopite + calcite + 3 CO_2

dolomite + 2 quartz = diopside + CO_2

5 dolomite + 8 quartz + H_2O = tremolite + 3 calcite + 7 CO_2

The calc-schists are also dolomite free and consist of more than 50% silicate minerals. The calc-schists commonly occur at the contact between the marble and Baltimore Gneiss, Setters, and Wissahickon, and are probably a result of metasomatism along the contacts (Crowley, 1976; Alcock, 1989) .

An interesting feature of the marble is its association with large deposits of kaolin. During the 1800's kaolin was mined in several pits in the Hockessin valley. The kaolin appears to have formed by weathering of pegmatites. These pegmatites have no surface expression, however they appear to parallel the foliations in the marble.

In Delaware the Wissahickon Formation is an areally extensive sequence of pelitic and psammitic metasedimentary gneiss with interlayered amphibolite and lenses of serpentinite. No internal stratigraphy has been identified in the complexly folded and faulted lithologies that comprise the Wissahickon, and the large-scale structures are poorly understood.

The pelitic gneiss contains biotite, quartz, plagioclase (oligoclase to andesine) and Fe-Ti oxides. Staurolite, sillimanite, orthoclase, muscovite and garnet vary with increasing metamorphic grade. Cordierite appears in a few places adjacent to the Wilmington Complex. Metamorphic grade, as recorded by the minerals in the pelitic rocks increases eastward from amphibolite facies to above the second sillimanite isograd as documented by the presence of sillimanite and orthoclase and the absence of primary muscovite. Peak temperatures increase from 650°C west of Newark to 750°C east of Concord Pike. Pressure estimates vary from 4 to 7 kilobars (Calem, 1987; Plank 1988). The highest-grade assemblages cluster east of the Red Clay Creek and around the Mill Creek Dome.

Isograds based on the progress of discontinuous reactions show a complex pattern reflecting (1) the regional increase in grade from northwest to southeast, (2) heat from a local source in the east, possibly the Wilmington Complex as suggested by Wagner and Srogi (1987), and (3) post-metamorphic emplacement of the Mill Creek Dome (Plank, 1989).

GRAVITY

Gravity measurements were made in the study area using a Worden Prospector Gravity meter. Gravity modeling of the marble underlying the Hockessin Valley indicates the marble is approximately 600 feet thick and dips to the southeast.

TEST DRILLING

Nine test holes were drilled to obtain data on lithology, contact location, and depth of weathering of the marble. One hole was continuously cored and eight holes were cored at selected depths.

Cores from three holes will be displayed at Brandywine Park on the field trip. One core (Cb12-10) is dolomite marble with thin irregular layers of calcite marble from a depth of 400-419' from Pleasant Hill Valley.

Core Bb44-30 was drilled along Mill Creek south of the Hockessin marble valley. The hole was drilled through 20 feet of overburden, 60 feet of Wissahickon, and bottomed at 100 feet in marble. In this core the Wissahickon is a garnet-rich pelitic gneiss, and the marble is a calc-schist with cataclastic textures. Adjacent to this hole the DGS maintains an observation well drilled to 300 feet. A 10 foot core was recovered from the bottom of this well and is a typical Wissahickon pelitic gneiss. Gamma-ray logs from this well show that the marble here is approximately 40 feet thick. Dips of the foliation in the Wissahickon vary between 90 and 30°, and in the marble between 25 and 45°. Although the cores are not oriented the dips of the foliations are probably to the southeast, if in agreement with dips in the surrounding rocks.

A test hole drilled on the south side of the marble valley between Yorklyn and Hockessin (Bb35-16) traversed calc-schist retrieved in a core at 48 feet followed by 400 feet of white dolomite marble to a depth of 450 feet. At this depth the drilling suggested a change in rock type, and subsequent cores at 450 and 520 feet contained amphibolite and biotite gneiss respectively. Changes in lithology and the contact depths can be verified on geophysical logs. Textures and thin section analyses suggest this rock underlying the marble is Baltimore Gneiss. Dips in both the marble and the gneiss are between 30 and 45°, presumably to the southeast.

Three additional test holes, 75 feet apart in a north-south line, were located so as to straddle the mapped contact between the Wissahickon and the Cockeysville south of Hockessin. The northernmost hole Bb34-55 penetrated pegmatite and marble. Hole Bb34-54, immediately to the south, penetrated highly weathered Wissahickon underlain by a moderately weathered calc-schist. A gamma-ray log indicated the top of the calc-schist at a depth of 50 feet. The southernmost hole (Bb34-53) penetrated weathered Wissahickon schist to about 65 feet, and pegmatite from 65 to 81 feet.

GEOLOGIC STRUCTURE

The test holes show the marble in the Hockessin-Yorklyn area overlies the Baltimore Gneiss and underlies the Wissahickon in a normal stratigraphic sequence. Foliations in all units strike N40-45°E, parallel to the regional strike, and dip to the southeast. In the Wissahickon immediately south of the Mill Creek Dome foliations dip southeast between 23 and 60°, and then steepen to 70-90°. North of the Dome the Wissahickon foliations dip southeast under the Baltimore Gneiss at a shallow angle, 20-35°SE. The structures and map pattern suggest that the Mill Creek Dome is a nappe overturned to the northwest. The absence of the Cockeysville and Setters along the north side of the dome may be due to shearing or thinning out of these units on the overturned limb of a nappe, or to removal by faulting. At the contact of the Baltimore Gneiss and Wissahickon on the northern side of the dome the rocks are riddled with pegmatites.

Cross sections B-B', C-C' and D-D' (Fig. 6-2, 6-3) illustrate this interpretation. The structure is more complicated in the Pleasant Hill Valley where interpretations must account for the "upside-down" sequence in which Setters overlies the Cockeysville and Wissahickon overlies the Setters. Our interpretation, presented in section A-A' (Fig. 6-3), shows the Wissahickon faulted over an overturned section of Setters-Cockeysville.

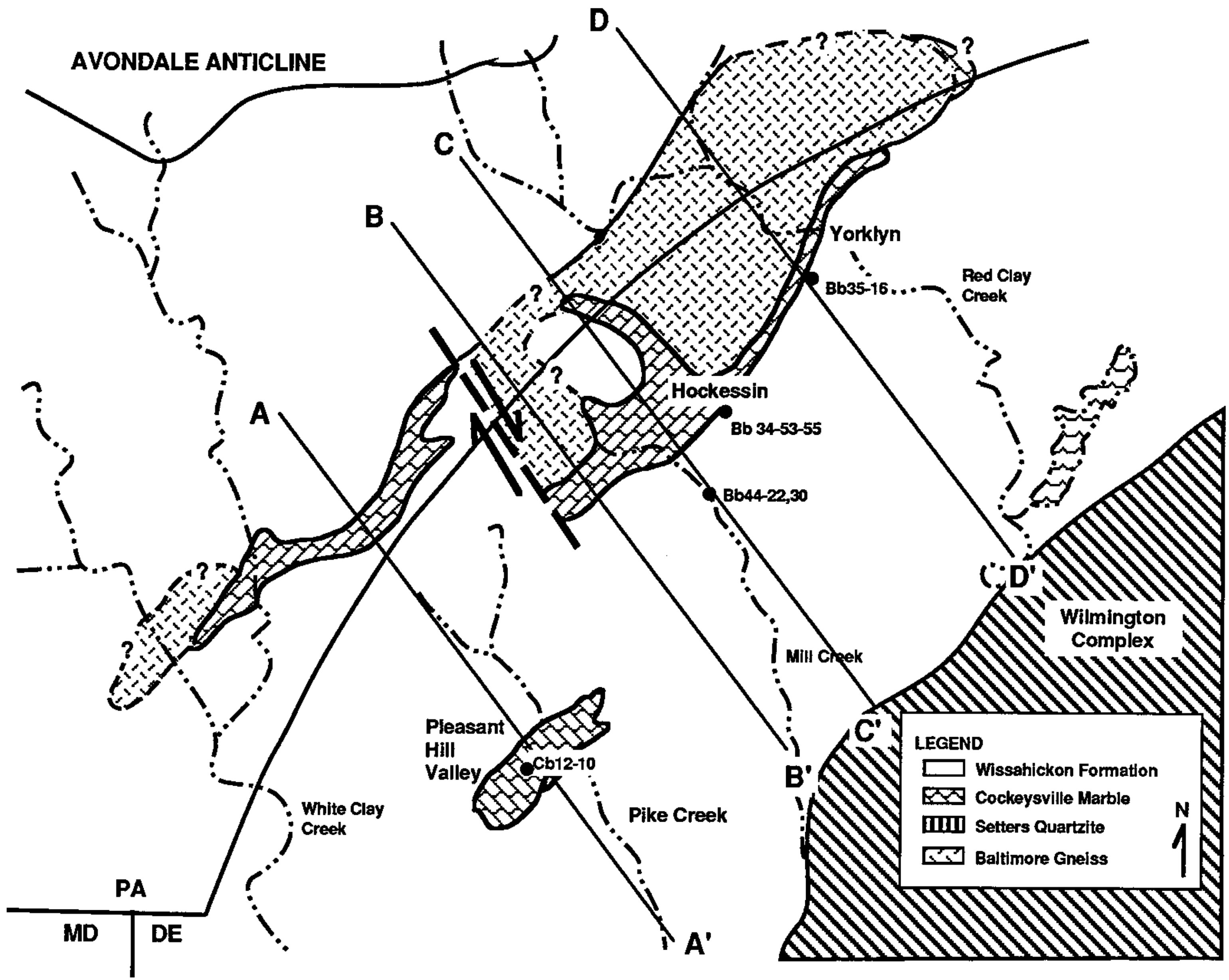


Figure 6-2. Location of drill holes and cross section lines, Mill Creek Dome

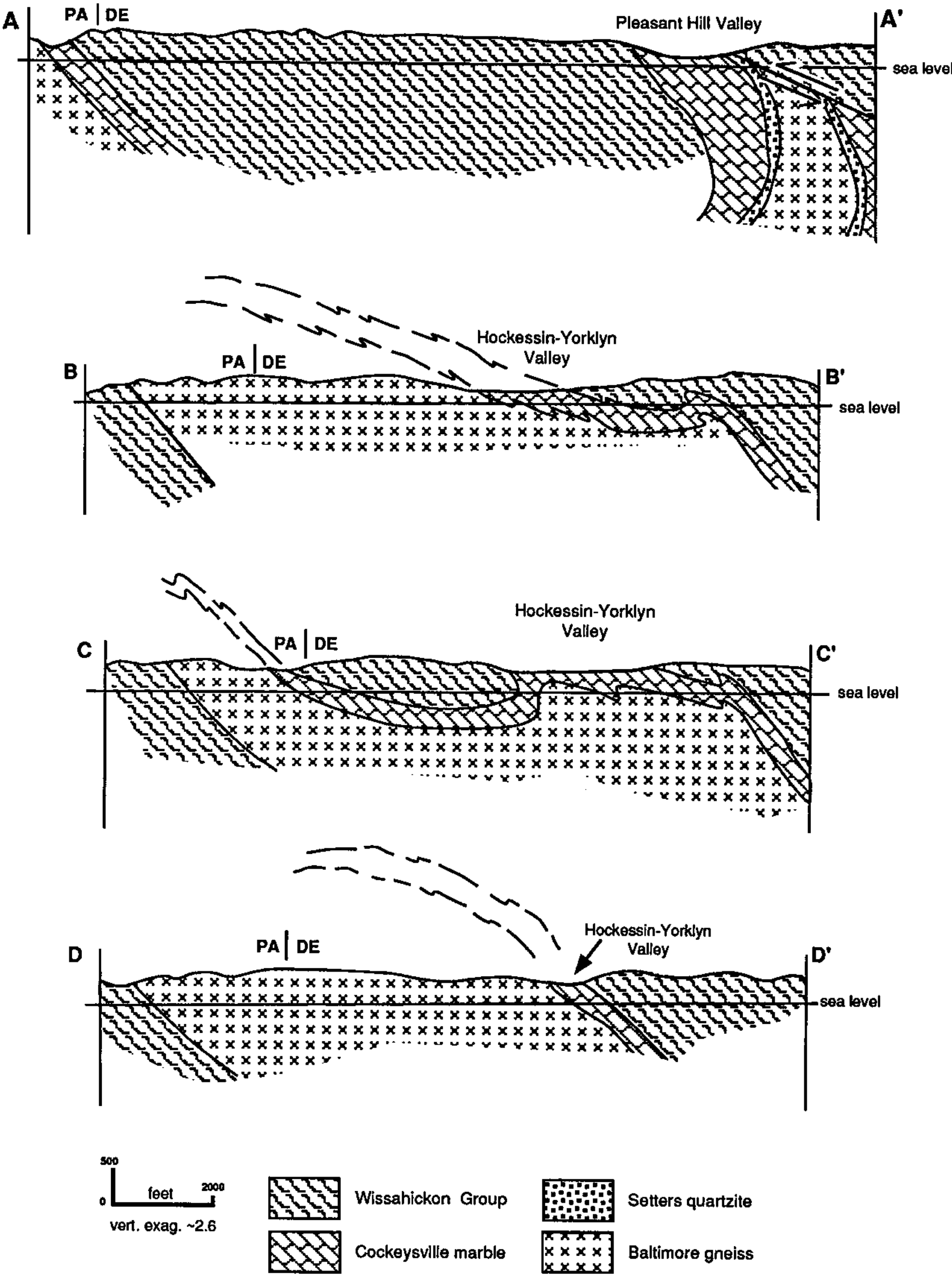


Figure 6-3. Cross sections of the Mill Creek Dome.

TABLE 6-1: LITHOLOGIC CHARACTER OF SETTERS QUARTZITE

Minerals	P431	P431	495-2	Kuhlman (1975)		Hopson (1964)			Southwick (1969)	
	Dk layer	Lt layer		344A	344D	351	H36-1	H106-7	2	3
Quartz	30	28	64	14	34	29	58	45	36	40
Plagioclase	2	2	1	1	x	0	2	8	28	x
Microcline	50	64	16	55	57	46	19	32	14	42
Biotite	16	6	10	18	4	13	12	11	16	10
Muscovite		x		11	2	11	6	2	3	5
Garnet			7.6							
Opaques	x	x	x	x	x	x	1		1	3
Accessories										
Zircon	x	x	x				x	x	x	x
Sphene	x	x	x				x	x	x	x
Apatite	x	x	2.4	x	x		x		x	x
Chlorite/Bio	x			x	x	x				
Clay/Pla	x							x	x	

P431 Southeast Side of Eastburn's Quarry, gneiss overlying marble, Delaware
405-2 Southeast side of Eastburn's Quarry, gneiss overlying marble, garnet rich layer, Delaware
344A Quarry at Avondale, Pennsylvania
351 Quarry NE of Avondale, Pennsylvania
H36-1 Clarksville dome, 1-1/2 miles NW of Pine Orchard, Maryland
H106-7 Woodstock dome, at Marriotsville, Maryland
2 Lower Bynum Run about 0.5 mile north of Hookers Mill Road, Maryland
3 Lower Winters Run about 0.9 mile upstream from Interstate Highway 95, Maryland

WILMINGTON COMPLEX

LeeAnn Srogi
Department of Geology and Astronomy
West Chester University

INTRODUCTION

The Wilmington Complex is composed of granulite-facies gneisses intruded by numerous small plutons. The tectonic models of Crawford and Crawford (1980) and Wagner and Srogi (1987) identify the Wilmington Complex as a fragment of the deep infrastructure of a magmatic arc, tectonically emplaced onto North America during the Taconic orogeny. Most of the igneous rocks in the Wilmington Complex are undeformed, and the Complex as a whole has not been significantly overprinted by post-Taconic deformation or re-metamorphism. The Wilmington Complex therefore is the place to examine the arc basement and compare it with North American Grenville basement and other terranes of the Appalachian-Caledonian belt.

The Wilmington Complex is covered by Coastal Plain sediments to the south and is bordered by the Wissahickon Formation along most of its margin to the northwest and northeast (Fig. 1-1). The Wilmington Complex was first mapped by Bascom and Miller (1920) and Bascom and Stose (1932). More recent maps and descriptions are by Ward (1959), Woodruff and Thompson (1975), Mark (1977), Brick (1980), Wagner, Srogi, and Brick (1987), Wagner and Srogi (1987), and Srogi (1988). Ward (1959) recognized that the Wilmington Complex rocks are distinct from the high-grade gneisses and meta-igneous rocks of Grenville age in southeastern Pennsylvania.

For example, the relatively monotonous felsic and mafic gneisses do not have the diversity of the Grenville gneisses in the West Chester prong (Wagner and Crawford, 1975) or the Reading Prong (Drake, 1984).

A number of plutons intrude the Wilmington Complex, most of which are gabbroic stocks and dikes too small to show at the scale of Fig. 1-1. Ward (1959) identified and mapped the two largest plutons, the Bringhurst and the Arden, (Fig. 1-1). The Bringhurst pluton contains only gabbroic lithologies, including gabbronorite, olivine gabbronorite, and olivine gabbro. An excellent exposure of the Bringhurst pluton is described in Wagner, Srogi, and Brick (1987). The Arden pluton, the largest of the intrusions, covers about 20 km^2 and is a composite pluton. In addition to gabbroic lithologies, the Arden pluton contains a suite of pyroxene-bearing quartz diorites, granodiorites, and granites with a Rb-Sr whole-rock crystallization age of 502 ± 20 Ma (Foland and Muessig, 1978). Field relationships in the Arden pluton indicate that the gabbroids are the same age or slightly older than the granodioritic suite.

Interpretation of the tectonic setting of the Wilmington Complex is based in part on the characteristics of the igneous rocks. The granodioritic suite in the Arden pluton is calc-alkaline and is chemically similar to other plutons from orogenic environments (Wagner and Srogi, 1987). Mineral assemblages and compositions of the Bringhurst gabbroids are similar to those in continental tholeiitic intrusions and orogenic layered mafic complexes which contain intermediate rather than anorthitic plagioclase (Srogi, Lutz, and Wagner, 1991). Preliminary Nd and Sr isotopic data for the Bringhurst gabbroids show evidence of crustal contamination (L. Srogi, unpub. data), and overlap the range of isotopic compositions for the 490-Ma Baltimore Mafic Complex (Shaw and Wasserburg, 1984; Hanan and Sinha, 1989). We interpret the tectonic setting of the igneous rocks of the Wilmington Complex to be a convergent plate boundary, consistent with the intrusion of gabbroids and calc-alkaline granitic rocks into continental or thick, arc-related crust. The Wilmington Complex may be a more deeply-exposed portion of the Cambro-Ordovician magmatic arc of the Virginia and Maryland Piedmont (Pavlides, 1981; Sinha et al., 1980).

The intrusive relationships of the igneous rocks into high-grade gneisses are well-exposed in the Wilmington Complex, and are not affected by later tectonism. Foliated, granulite-facies gneisses are intruded by plutonic rocks with undeformed igneous textures, indicating that the penetrative deformation in the gneisses is older than the magmatism. Migmatitic textures and disruption of metamorphic banding are observed in the gneisses around the margins of the Bringhurst pluton (Brick, 1980), the Arden pluton (Rdesinski, 1983), and some smaller intrusive plugs (Luborsky, 1983; L. Srogi, unpub. field mapping). Our conclusion is that the granulite-facies mineralogy and foliation in the gneisses are older than ≈ 500 Ma, and that intrusion of high-temperature magmas produced a thermal overprint on the gneisses. Mineral assemblages and compositions in felsic, mafic, and garnet-bearing gneisses constrain peak metamorphic conditions to about 800 ± 50°C and 0.6-0.7 GPa, (Wagner and Srogi, 1987; Srogi, 1988; Srogi, Wagner, and Lutz, ms. in review). In the gabbroic rocks, coronas surrounding olivine, separating it from plagioclase, contain spinel in symplectic intergrowth with low-Ca orthopyroxene and pargasitic amphibole. The lack of penetrative deformation of the gabbroids suggests that the coronas formed during subsolidus cooling to ambient granulite-facies conditions at ca. 21-28 km depth, consistent with the pressures estimated from the gneisses.

The timing and style of emplacement of the Wilmington Complex into its present position have important implications for tectonic models of the central Appalachian Piedmont. Zircons from one sample of felsic gneiss were analyzed for U-Pb isotopic composition by Grauert and Wagner (1975), who interpreted the nearly-concordant lower intercept of 441 Ma as the age of granulite-facies metamorphism during the Taconic orogeny. Wagner and Srogi (1987), however, suggested that the U-Pb age represents cooling of the Wilmington Complex from high temperatures following tectonic emplacement onto the North American continent. The presence of undeformed, ≈500-Ma igneous rocks in the Arden pluton clearly demonstrates that the Taconic orogeny did not cause pervasive deformation throughout the Wilmington Complex.

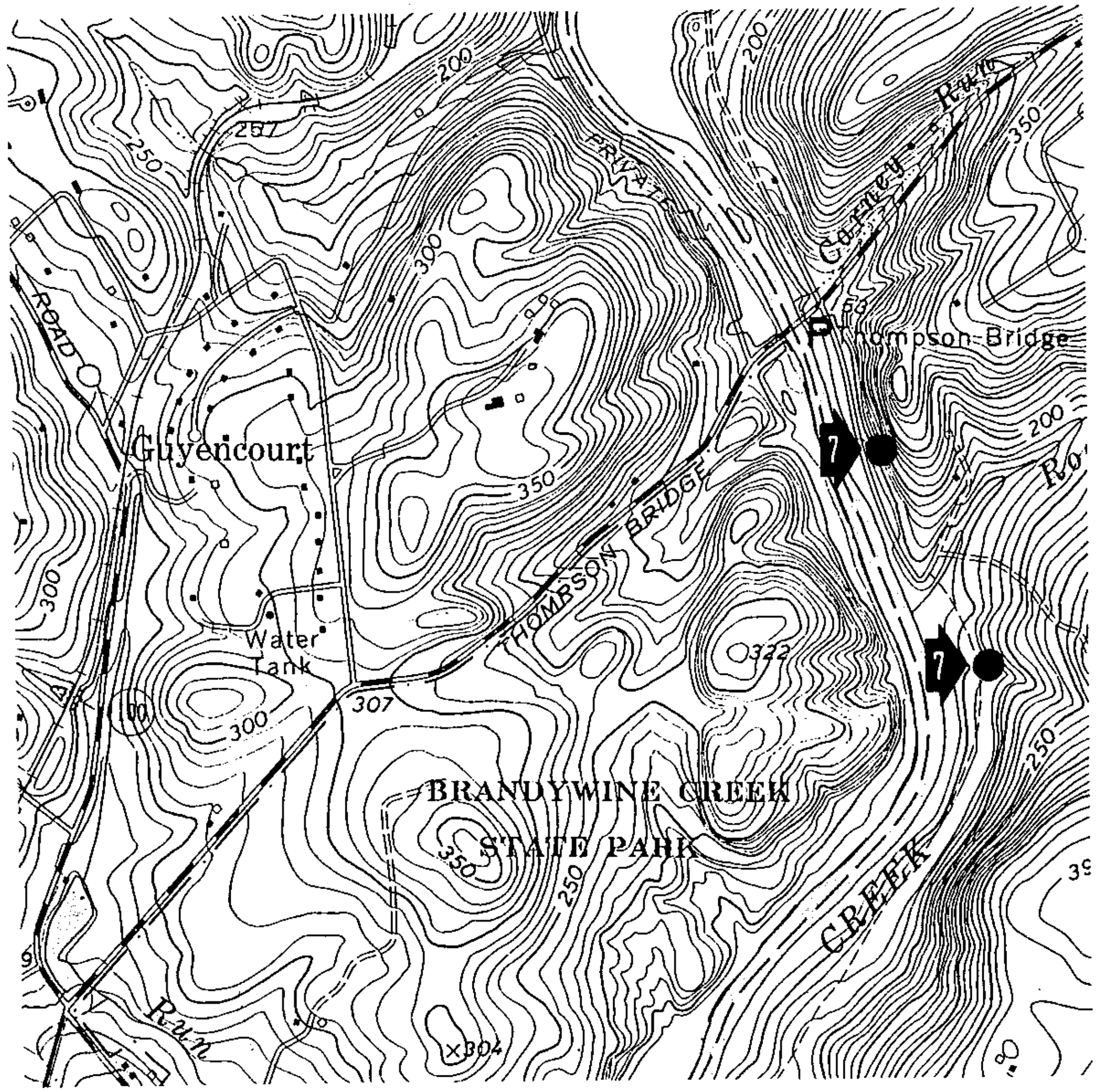
Wagner and Srogi (1987) consider the Wilmington Complex to be the highest structural level exposed in the northern Delaware-southeastern Pennsylvania Piedmont, although Horton, Drake, and Rankin (1989) think it is overthrust by the "Potomac terrane." At our field stop in Brandywine Creek State Park, we will discuss the contact relationships between the Wilmington Complex and the adjacent Wissahickon Formation based on recent mapping in this area and to the northeast. The foliation in both units dips steeply to the northwest, suggesting that the Wilmington Complex dips beneath the Wissahickon Formation and that the contact surface is steep. However, the outcrop pattern of the two units suggests that the Wilmington Complex **overlies** the Wissahickon Formation along a surface that dips gently to the southeast (Srogi, 1982; Wagner and Srogi, 1987). It is possible that post-Taconic tectonism has disrupted the original relationships to some extent and produced the steep foliations near the contact. The nature of the Wilmington-Wissahickon contact must be resolved in order to constrain models for the assembly of the orogen.

STOP 7 -- Wilmington-Wissahickon contact relationships, Brandywine Creek State Park

The parking lot for the Nature Preserve is on the southeast side of DE Rte. 92 (Thompson Bridge Rd.), just on the east side of Brandywine Creek. Walk south from the parking lot on a trail that runs parallel to the Creek.

At this stop we will review what is known of the relationship between the Wilmington Complex and the Wissahickon Formation. The actual contact is never exposed, and consideration of the outcrop distributions and the structures of the two units lead to contradictory interpretations of the contact relationships. These contradictions must be resolved before an improved tectonic model for the central Appalachian Piedmont can be achieved. Detailed field mapping and examination of microstructures is presently underway and our latest results will be discussed on this trip. Wagner et al., (1991) also describe this stop, including different outcrops which illustrate the same features.

In Brandywine Creek State Park and in the Nature Preserve on the east side of Brandywine Creek, are exposed the felsic and mafic gneisses of the Wilmington Complex and the pelitic to semi-pelitic



Wilmington North 7.5' Q. Lat 39°49'00" Long 75°34'12"

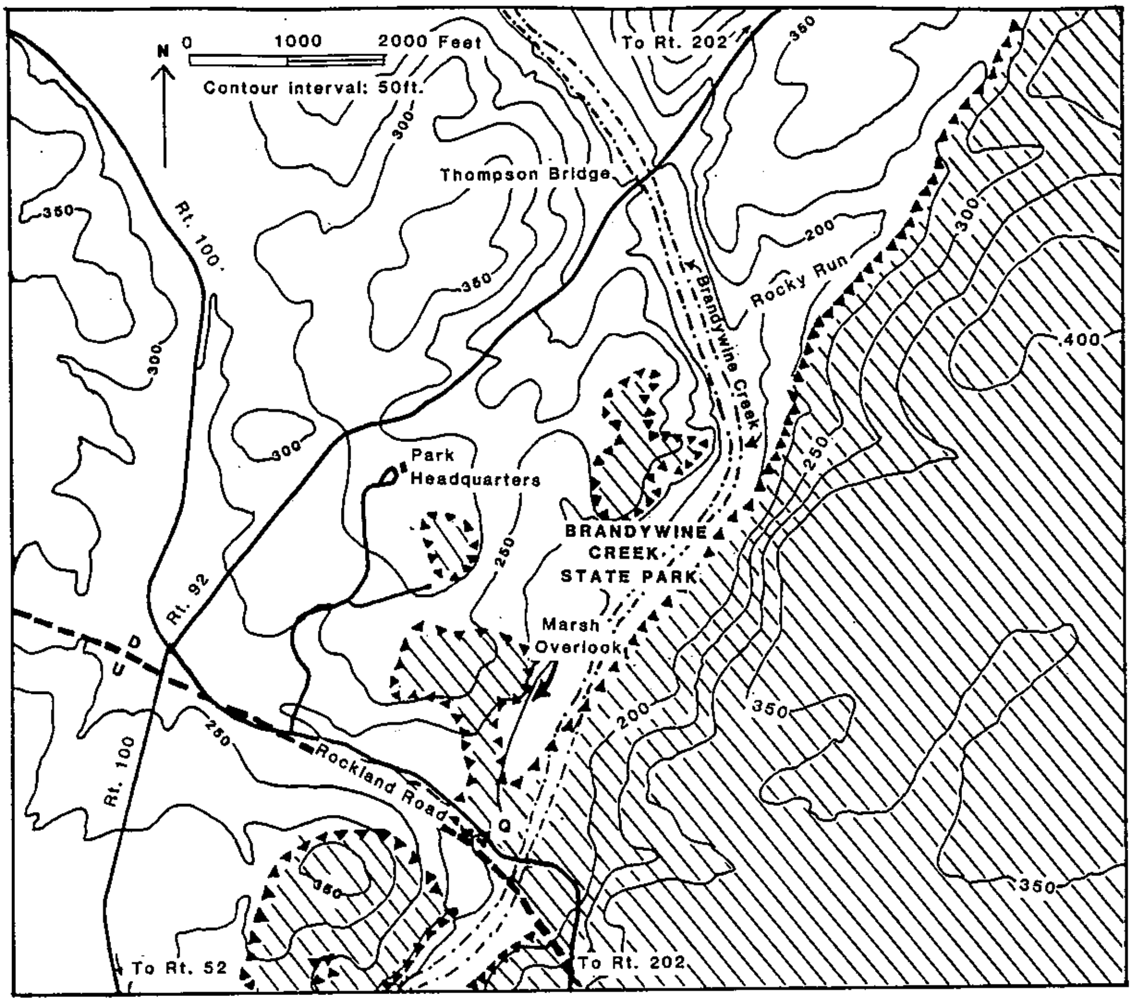


Figure 7-1. Interpretive geologic map of Brandywine Creek State Park. The contact shown is the intersection with the topography of a plane striking N45°E and dipping 5° southeast. It corresponds closely to the contact between the Wilmington Complex and the Wissahickon Group. Diagonal lines represent the Wilmington Complex; unpatterned, Wissahickon Group. Contour interval: 50 ft.

schists and gneisses of the Wissahickon Formation. The foliations in both the Wilmington Complex and the Wissahickon Formation have similar orientations, generally striking northeast and dipping moderately to steeply northwest. This led Ward (1959) to suggest a concordant stratigraphic contact between the two units, with the Wilmington Complex underneath the Wissahickon Formation. In more recent work beginning with Hager and Thompson (1975) and Woodruff and Thompson (1975), the contact is described as a fault. Woodruff and Thompson (1975) interpreted the contact to be a steep normal fault dipping to the southeast, nearly perpendicular to the attitude of foliation, with the Wilmington Complex on the hanging wall side. Ductile deformation of garnet in Wissahickon gneiss (Valentino, 1986 and 1988) and pyroxene in Wilmington Complex gneiss (Srogi, 1982) observed in samples collected near the contact indicates high-temperature deformation in both units.

Srogi (1982) suggested that the Wilmington Complex overlies the Wissahickon Formation along a contact surface that dips at a shallow angle to the southeast, based on the outcrop pattern of the two units. An attitude of about N46°E, 5° SE, (Fig. 7-1), was determined by a planar regression through nine contact points inferred from three cross-sections for the Brandywine Creek State Park area. This model surface predicts the occurrence of the two units very well for at least 2-3 km along strike to the northeast, while southwest of Brandywine Creek State Park, the contact surface has the same strike and dip but is about 100 feet higher in elevation (Hamre, 1983). Field evidence also suggests a shallowly-dipping contact surface farther to the northeast, in the

Marcus Hook and Media Quadrangles (Mark, 1977; M.L. Hill and D. Valentino, pers. comm., 1985; D. Valentino, pers. comm., 1990; L. Srogi, unpub. field data).

Although the model surface seems to predict the location of the contact between the two units, neither the Wilmington Complex nor the Wissahickon Formation contain structures parallel to this surface. The discordance between the contact surface and the steep foliations caused Horton, Drake, and Rankin (1989) in a recent review article to reject the contact surface model without, however, proposing an alternative model that explains the outcrop pattern. It is possible that the steep foliations developed after tectonic juxtaposition of the two units, and there is field evidence for offset of the contact surface by later shearing according to D. Valentino (pers. comm., 1990).

Boulders of Wilmington Complex gneiss can be examined next to Brandywine Creek in the parking area and along the trail approximately 250 meters southeast of the confluence of Rocky Run and Brandywine Creek (Fig. 7-1). Most typical are felsic gneisses with discontinuous layers or lenses of mafic gneiss. Some gneisses have strong mafic-felsic banding on a scale of centimeters, while others have layers on a scale of meters consisting of only felsic or mafic gneiss. Minerals present include plagioclase (andesine to labradorite), orthopyroxene, clinopyroxene, and magnetite, with variable amounts of quartz, brown-green hornblende, and biotite; K-feldspar and garnet are absent. A mafic unit containing abundant blue-green to brown-green hornblende, plagioclase, and less abundant pyroxene occurs close to the contact with the Wissahickon Formation in Brandywine Creek State Park and adjacent areas. At four localities elsewhere in the Wilmington Complex garnet-bearing gneisses, with more aluminous bulk compositions, occur as lenses or pods less than a meter in any dimension, interlayered with felsic and mafic gneisses. All of the gneisses are dark-colored on fresh surfaces and the foliation is most clearly visible on weathered surfaces.

South of the parking lot, the pelitic and semi-pelitic schists and gneisses of the Wissahickon Formation crop out along the hill on the east side of the trail. The rocks in the Wissahickon Formation which attained the highest metamorphic temperatures are found adjacent to the Wilmington Complex. Temperatures exceeded the second sillimanite isograd and the schists and gneisses contain garnet, biotite, sillimanite, K-feldspar, quartz, plagioclase, and ilmenite, ±cordierite and hercynitic spinel. Many of the rocks are extremely garnet-rich, or in some places, appear migmatitic. According to Calem (1987), cordierite and spinel formed in the Wissahickon as a result of incongruent melting of biotite-garnet-sillimanite assemblages.

Crawford and Mark (1982) suggested that cordierite-bearing assemblages in the Wissahickon formed during a low-pressure (0.4-0.5 GPa) metamorphic event which also affected the Wilmington Complex. However, cordierite-bearing assemblages in the Wilmington Complex record moderate pressures of 0.6-0.7 GPa, and grew during thermal metamorphism associated with igneous intrusions which have no direct counterpart in the Wissahickon Formation.

Wagner and Srogi (1987) suggested that emplacement of hot

Wilmington Complex along a mid-crustal level thrust fault over amphibolite-facies Wissahickon Formation caused heating of the Wissahickon beneath the thrust and relatively rapid cooling of the overlying Wilmington Complex. The mineral assemblages and compositions of zoned garnets in the Wissahickon Formation northwest of the Wilmington Complex (Plank, 1989; Alcock, 1989), can be explained by a model of heating during overthrusting by the Wilmington Complex followed by decompression due to uplift and northwest-verging thrust faulting within the Wissahickon Formation. Alcock (1989) envisions the metamorphism and thrusting taking place within the crust of the arc-forearc infrastructure.

The rocks near the contact between the Wilmington Complex and the Wissahickon Formation provide evidence for the juxtaposition of these two units during the Taconic orogeny, as well as evidence of later deformation that may have reoriented the contact and produced steep foliations.

POST-TACONIC TRANSPRESSION IN THE PENNSYLVANIA PIEDMONT

Mary Louise Hill
Dept. of Geology
Temple University

INTRODUCTION

The recognition and analysis of a system of steeply dipping ductile shear zones crosscutting the metamorphic rocks of the Pennsylvania Piedmont provides fundamental evidence for transpression in this part of the Central Appalachians sometime after the Taconic orogeny (Myer et al., 1985; Hill, 1987; Valentino, 1988; D. Valentino, 1989; Valentino, 1991). Most of these shear zones preserve evidence of dextral strike-parallel displacement, with evidence in some cases for synchronous compression across the zones (Valentino and Hill, 1991; Valentino and Wiswall, 1991).

Deformation on these discrete shear zones postdates peak metamorphism associated with the Taconic orogeny. The lateral displacements related to the shear zones occurred during post-Taconic cooling. If the Taconic orogeny was associated with crustal telescoping and/or the accretion of terranes by thrusting at the convergent plate margin of North America, then the post-Taconic shear zone deformation was responsible for disruption and redistribution of the Taconic assemblage. Since the system of shear zones was active for a considerable time and apparently produced significant displacements, it is important to understand this part of the tectonic history before modeling the details of the Taconic orogeny in the Central Appalachians.

EARLIEST POST-TACONIC SHEAR ZONES

The transcurrent ductile shear zone systems overprint metamorphic mineral assemblages and fabrics attributed to the Taconic orogeny. In southeastern Pennsylvania, the Taconic foliation is characterized by variable orientation due to folding, but dips gently to moderately to the northwest or southeast in most places.

The oldest major structures cutting this Taconic fabric are the Rosemont and Crum Creek shear zones (Valentino, 1988; R. Valentino, 1989). Both are near-vertical ductile shear zones with horizontal displacement, and both were active at amphibolite facies conditions. The Rosemont shear zone had dextral displacement; the Crum Creek shear zone had sinistral displacement. Evidence for contemporaneous displacement (mutual overprinting relationships and similar conditions of deformation) as well as the geometric relationship of these two shear zones suggest that the Rosemont and Crum Creek shear zones are a conjugate set of structures (R. Valentino, 1989; Valentino and Valentino, 1991).

However, the Rosemont shear zone outlasted the Crum Creek shear zone and was responsible for much larger displacements. Displacement along the Rosemont shear zone was great enough that different lithologies are juxtaposed across the zone along most of its length. The Crum Creek shear zone terminates at the Rosemont shear zone, indicating that the Rosemont shear zone is the major transcurrent structure from this phase of the deformation history.

These two shear zones disrupt Taconic isograds, although microstructural evidence indicates that the shear zone deformation occurred at amphibolite facies temperatures (Valentino and Valentino, 1991). Since there is no geochronologic evidence for significant reheating of the region after Taconic time, this deformation must have occurred early in the post-Taconic cooling history.

LATER POST-TACONIC SHEAR ZONES

In the Wissahickon Valley area, the Rosemont shear zone is truncated by the Huntingdon Valley shear zone. This shear zone, together with the Cream Valley, Martic (Myer et al., 1985) and Peach Bottom (Valentino, 1991) shear zones, is part of a system of major ductile shear zones with dextral transcurrent displacement that strike across the southeastern Pennsylvania Piedmont from west of the Susquehanna River to the Delaware River. The shear zone system has a general strike of 070o in most places, but both the strike and character of the structures change westward toward the Susquehanna River.

Although the shear zones are fairly broad, they consist of relatively narrow zones of high strain anastomosing around less strained areas. The earliest recognizable deformation on these shear zones occurred at amphibolite facies temperatures; fabrics and microstructures from the earliest high-temperature deformation are preserved in low strain areas marginal to the major loci of displacement. Deformation continued as the rocks cooled to greenschist facies conditions. In the zones of highest strain, amphiibolite facies assemblages were replaced

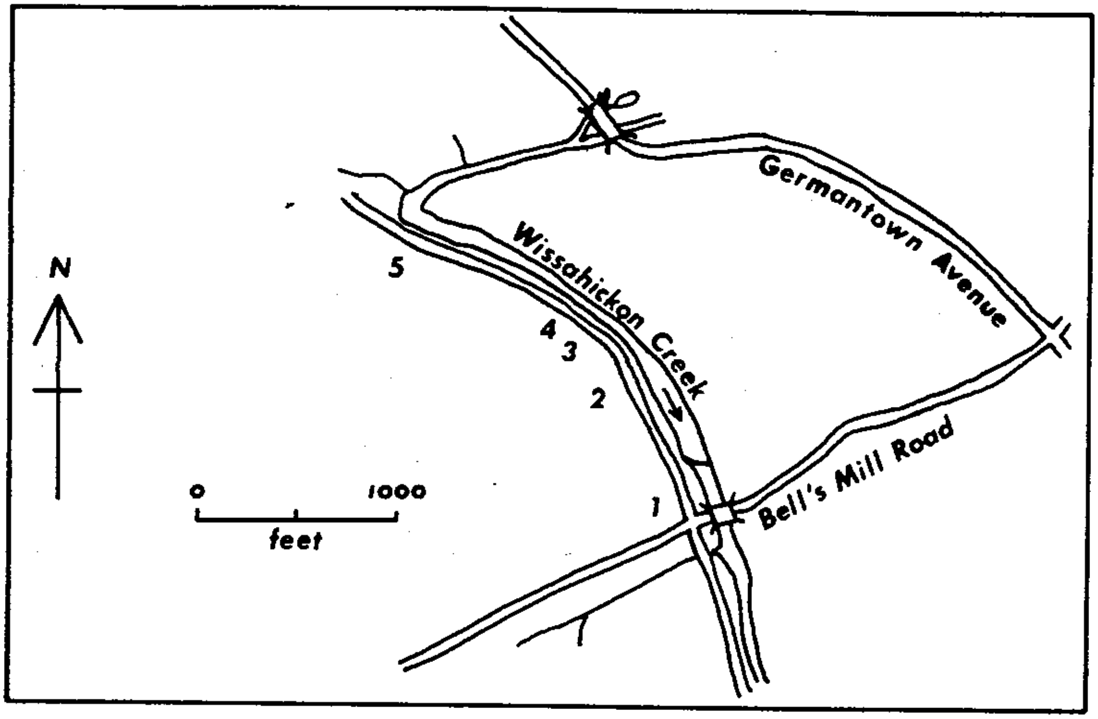


Figure 8-1. Location map for STOP 8

by greenschist facies assemblages. As temperatures decreased, the ductile deformation became increasingly restricted to narrow discrete shear zones.

The shear zones in this ENE-striking system are characterized by a prominent steeply dipping foliation which is extremely straight and consistent along strike. In most lithologies significant grain size reduction is apparent; the zones of highest strain are characterized by extreme mylonitization. A strong subhorizontal stretching or mineral lineation commonly lies within this foliation. A variety of asymmetric microstructures (S-C fabric, pressure shadows, recrystallized porphyrocast tails,etc.) indicate a dextral sense of displacement (Hill, 1989).

This system of latest ductile shear zones coincides with a steep gradient in Bouger gravity anomalies (the Appalachian gravity gradient), suggesting that it is a major crustal structure (Song, 1987). This gravity signature continues along strike across New Jersey where the structure is buried. A plausible explanation for the steep gravity gradient is a steeply dipping boundary between crustal blocks with different gravity structures, i.e. a steeply dipping terrane boundary. This interpretation is consistent with other evidence for significant transcurrent displacement on this shear zone system (Song and Hill, 1988).

STOP 8: WISSAHICKON CREEK, NORTH OF BELL'S MILL ROAD

Directions: Park in the parking lot on Bell's Mill Road just west of Wissahickon Creek. Walk north on Forbidden Drive (the gravel road along the west side of the creek). There are several outcrops beside the road and along the creek between Bell's Mill Road and the 140° bend in Wissahickon Creek (about 1/2 mile north of Bell's Mill Road).

Description: This stop exposes the southern transition zone of the Huntingdon Valley shear zone. The shear zone is exposed near the bend in the creek about 1/2 mile north of Bell's Mill Road. The earlier Rosemont shear zone is truncated by the Huntingdon Valley shear zone here. Much of the strain in the rocks at this stop occurred within the Rosemont shear zone, with later transposition by the Huntingdon Valley shear zone.

The Rosemont shear zone is a post-Taconic ductile shear zone that had dextral transcurrent displacement. South of the transition zone of the Huntingdon Valley shear zone, the Rosemont shear zone generally strikes about 030° to 040°. Deformation fabrics and petrologic relations indicate that this shear zone was active at amphibilite facies conditions (Valentino, 1988). At this stop, mylonitic foliation associated with the Rosemont shear zone strikes between 055° and 070 and is very steeply dipping.

The Huntingdon Valley shear zone strikes about 070° and is continuous from here to Trenton, New Jersey. It is a major post-Taconic ductile shear zone that had dominantly dextral transcurrent displacement. A strong mylonitic foliation striking approximately 070° with very steep to vertical dip characterizes this shear zone. The temperature conditions of deformation apparently ranged from lower amphibolite facies to greenschist facies.

The following locations are shown on Figure 8-1:

Location 1: Ultramafic rocks with a strong deformation fabric parallel to the foliation in the enveloping Wissahickon schist.

Location 2: Fine-grained Wissahickon schist with a prominent and "straight" vertical foliation and subhorizontal mineral lineation. Location 3: Old quarry in deformed granitic gneiss with well-developed S-C fabric indicating dextral displacement.

Location 4: Wissahickon schist folded around boudins of granitic material.

Location 5: Mylonitized gneiss within the Huntingdon Valley shear zone.

A general description of lithologies exposed at this stop can also be found in the Geological Society of America Centennial Field Guide (Crawford, 1987).

REFERENCES

Alcock, J. A., 1989, Tectonic units in the Pennsylvania-Delaware Piedmont: evidence from regional metamorphism and structure, Ph.D thesis, University of Pennsylvania, 259 p.

Armstrong, E., 1941, Mylonization of hybrid rocks near Philadelphia, Pennsylvania: Geological Society of America Bulletin, v. 52, p. 667-694.

Bailey, E. B. and Mackin, H., 1937, Recumbent folding in the Pennsylvania Piedmont; preliminary statement: American Journal of Science, v. 33, p. 187-190.

Bascom, F., 1905, Piedmont district of Pennsylvania: Geological Society of America Bulletin, v. 16, p. 289-328.

Bascom, F., Clark, W. B., Darton, N. H., Kummel, H. B., Salisbury, R. D., Miller, B. L. and Knapp, G. N., 1909, Philadelphia Folio, United States Geological Survey Atlas, Folio 162, 37 p

Bascom, F. and Miller, B. L., 1920, Elkton-Wilmington Folio, United States Geological Survey Atlas, Folio 211, 15 p

Bascom, F. and Stose, G. W., 1932, Coatesville-West Chester Folio, Pennsylvania-Delaware. United States Geological Survey Atlas, Folio 223, 15 p

---- 1938, Geology and Mineral Resources of the Honeybrook and Phoenixville Quadrangles, Pennsylvania,United States Geological Survey Bulletin 891, United States Geological Survey, 145 p.

Berg, T. M., Edmunds, W. E., Geyer, A. R., Glover, A. D., Hoskins, D. M. and and others (1980): Geologic Map of Pennsylvania. Pennsylvania Geological Survey. 4th Series. Map 1. Scale 1:250,000.

Berg, T. M., McInerney, M. K., Way, J. H. and MacLachlan, D. B., 1986, Stratigraphic Correlation Chart of Pennsylvania, Pennsylvania Geological Survey General Geology Report 75, Pennsylvania Geological Survey, p.

Bliss, E. F. and Jonas, A. I., 1916, Relation of the Wissahickon mica gneiss to the Shenandoah limestone and Octoraro schist ~~of the~~ Doe Run-Avondale region, Chester Co., Pa: United States Geological Survey Professional Paper, v. 98, p. 9-34.

Brick, E., 1980, Field and petrographic study of the Bringhurst pluton, northern Delaware, A.B. thesis, University of Pennsylvania, 24 p.

Calem, J., 1987, A petrologic study of migmatites of Brandywine Creek State Park, northern Delaware, and implications for the mechanism of migmatization, M.S thesis, University of Pennsylvania, 111 p.

Chatterjee, N. D. and Johannes, W., 1974, Thermal stability and standard thermodynamic properties of synthetic 2M1-Muscovite, $KAl_2AlSi_3O_{10}((OH)_2)$: Contributions to Mineralogy and Petrology, v. 48, p. 89-114.

Chipera, S. J. and Perkins, D., 1988, Evaluation of biotite-garnet geothermometers: application to the English River subprovince, Ontario: Contributions to Mineralogy and Petrology, v. 98, p. 40-48.

Crawford, M. L. 1976, The Pennsylvania Piedmont - remnant of a back-arc basin? In: Proceedings of the First Northeastern Women's Geoscientists Conference. : St. Lawrence University, Canton, NY, 35-43.

--- 1987, The Wissahickon schist type section, Wissahickon Creek, Philadelphia, Pennsylvania. In: Northeastern Section of the Geological Society of America. Ed: Roy, D. C. Centennial Field Guide, 5. Geological Society of America, Boulder, CO, 77-80.

Crawford, M. L. and Crawford, W. A., 1980, Metamorphic and tectonic history of the Pennsylvania Piedmont: Journal of the Geological Society of London, v. 137, p. 311-320.

Crawford, M. L. and Mark, L. E., 1982, Evidence from metamorphic rocks for overthrusting, Pennsylvania Piedmont: Canadian Mineralogist, v. 20, p. 333-347.

Crawford, W. A. and Hoersch, A. L., 1984, The geology of the Honey Brook Upland, southeastern Pennsylvania: Geological Society America Special Paper, v. 194, p. 111-125.

Crowley, W. P., 1976, The geology of the crystalline rocks near Baltimore, and its bearing on the evolution of the eastern Maryland Piedmont,Maryland Geological Survey Report of Investigations 27, Maryland Geological Survey, 40 p.

Demmon, F. E., III, 1977, Investigations of the origins and metamorphic history of Precambrian gneisses, Downingtown 7.5 minute quadrangle, southeastern Pennsylvania, M.A. thesis, Bryn Mawr College, 71 p.

Drake, A. A., Jr., 1989, The Taconic suture in the central Appalachians: Geological Society of America Abstracts with Programs, v. 21, p. 11.

---- 1984, The Reading Prong of New Jersey and eastern Pennslvania: An apraisal of rock relations and chemistry of a major Proterozoic terrane in the Appalachians: Geological Society America Special Paper, v. 194, p. 75-110.

Eggert, R. G. and Kerrick, D. M., 1981, Metamorphic equilibria in the siliceous dolomite system: 6 Kbar experimental data and geologic implications: Geochimica et Cosmochimica Acta, v. 45, p. 1039-1049.

Faill, R. T. and MacLachlan, D. B., 1989, Tectonic terranes of southeastern Pennsylvania: Geological Society of America Abstracts with Programs, v. 21, p. 13.

Ferry, J. M., 1976, P, T, f_{CO2} and f_{H2O} during metamorphism of calcareous sediments in the Waterville-Vassalboro area, south-central Maine: Contributions to Mineralogy and Petrology, v. 57, p. 119-143.

Ferry, J. M. and Spear, F. S., 1978, Experimental calibration of the partitioning of Fe and Mg between garnet and biotite: Contributions to Mineralogy and Petrology, v. 66, p. 113-117.

Foland, K. A. and Muessig, K. W., 1978, A Paleozoic age for some charnockitic-anorthositic rocks: Geology, v. 6, p. 143-146.

Ganguly, J. and Saxena, S. L., 1984, Mixing properties of aluminosilicate garnets: constraints from natural and experimental data, and applications to geothermo-barometry: American Mineralogist, v. 69, p. 88-97.

Gates, A. E. and Sun, D., 1989, Geological and geophysical evidence for thrust faulting in the Taconic suture zone of the southeastern PA Piedmont: Geological Society of America Abstracts with Programs, v. 21, p. 17.

Grauert, B., Crawford, M. L. and Wagner, M. E., 1973, U-Pb isotopic analysis of zircons from granulite and amphibolite facies rocks of the West Chester prong and Avondale anticline, southeastern Pennsylvania: Carnegie Institution of Washington Yearbook, v. 72, p. 290-293.

Grauert, B. and Wagner, M. E., 1975, Age of the granulite facies metamorphism of the Wilmington Complex, Delaware-Pennsylvania Piedmont: American Journal of Science, v. 275, p. 683-691.

Grauert, B., Wagner, M. E. and Crawford, M. L., 1974, Age and origin of amphibolite-facies rocks of the Avondale anticline (southeastern Pennsylvania) as derived from U-Pb isotopic studies on zircon: Carnegie Institution of Washington Yearbook, v. 73, p. 1000-1003.

Hager, G. M. and Thompson, A. M., 1975, Fault origin of major lithologic boundary in the Delaware Piedmont: Geological Society of America Abstracts with Programs, v. 7, p. 68-69.

Hamre, J., 1983, The contact between the Wilmington Complex and the Wissahickon near Brandywine Creek State Park, A.B. thesis, University of Pennsylvania, 10 p.

Hanan, B. B. and Sinha, A. K., 1989, Petrology and tectonic affinity of the Baltimore Complex, Maryland: Geological Society America Special Paper, v. 231, p. 1-18.

Hatcher, R. D., 1989, Sutures in the Appalachian orogen: Geological Society of America Abstracts with Programs, v. 21, p. 21.

Hatcher, R. D., Osberg, P. H., Drake, A. A., Robinson, P. and Thomas, W. A., 1988, The Appalachian orogen in the United States: Geological Society of America Abstracts with Programs, v. 20, p. A132.

Henderson, J. R., Johnson, R. W. and Gilbert, F. P. (1963): Aeromagnetic map of the Wilmington, Delaware area and adjacent parts of Pennsylvania and Marland. United States Geological Survey. Geophysical Investigations. Map GP-363.

Higgins, M. W., Fisher, G. W. and Zietz, I., 1973, Aeromagnetic discovery of a Baltimore gneiss dome in the Piedmont of northwestern Delaware and southeastern Pennsylvania: Geology, v. 1, p. 41-43.

Higgins, M. W., Sinha, A. K., Zartman, R. E. and Kirk, W. S., 1977, U-Pb zircon dates from the central Appalachian Piedmont; a possible case of inherited radiogenic lead: Geological Society of America Bulletin, v. 88, p. 125-132.

Hill, M. L., 1987, Recent studies and new ideas in metamorphic geology of the S.E. Pennsylvania Piedmont: Geological Society of America Abstracts with Programs, v. 19, p. 19.

---- 1989, Structure of the Martic Zone: Geological Society of America Abstracts with Programs, v. 21, p. 22.

Hodges, K. V. and Spear, F. S., 1982, Geothermometry, geobarometry and the Al_2SiO_5 triple point at Mt. Moosilauke, New Hampshire: American Mineralogist, v. 67, p. 1118-1134.

Hoersch, A. L. and Crawford, W. A., 1988, The Mine Ridge of the SE Pennsylvania Piedmont: Northeastern Geology, v. 10, p. 181-194.

Hopson, C. A. 1964, The crystalline rocks of Howard and Montgomery Counties. In: The Geology of Howard and Montgomery Counties. : Maryland Geological Survey, Baltimore, MD, 27-215.

Horton, J. W., Drake, A. A., Rankin, D. W. and Dallmeyer, R., 1988, Preliminary tectonostratigraphic terrane map of the central and southern Appalachian orogen: Geological Society of America Abstracts with Programs, v. 20, p. A123.

Horton, J. W., Jr., Drake, A. A., Jr. and Rankin, D. W., 1989, Tectonostratigraphic terranes and their Paleozoic boundaries in the central and southern Appalachians: Geological Society America Special Paper, v. 230, p. 213-245.

Hoschek, G., 1973, Die reaktion phlogopit + calcit + quarz = tremolit + kalifedspar + H_2O + CO_2: Contributions to Mineralogy and Petrology, v. 39, p. 231-237.

Howard, C. T., 1988, The structural and metamorphic history of the Cream Valley fault zone near Conshohocken, PA, M.S. thesis, University of Pennsylvania, 281 p.

Huntsman, J. R., 1975, Crystalline rocks of the Wagontown 7.5 minute quadrangle, M.A. thesis, Bryn Mawr College, 69 p.

Kerrick, D. M., 1972, Experimental determination of muscovite + quartz stability with $P_{H_2O}<P_{total}$: American Journal of Science, v. 272, p. 946-958.

Knopf, E. B. and Jonas, A. I., 1923, Stratigraphy of the crystalline schists of Pennsylvania and Maryland: American Journal of Science, v. 5, p. 40-62.

---- 1929, Geology of the McCalls Ferry-Quarryville district, Pennsylvania,United States Geological Survey Bulletin 799, United States Geological Survey, 156 p.

Koziol, A. M. and Newton, R. C., 1988, Redetermination of the anorthite breakdown reaction and improvement of the plagioclase-garnet-Al_2SiO_5-quartz geobarometer: American Mineralogist, v. 73, p. 216-223.

Kuhlman, R., 1975, Stratigraphy and metamorphism of the Wissahickon Formation in southeastern Chester County, Pennsylvania, M.A. thesis, Bryn Mawr College, 47 p.

Lapham, D. M. and Bassett, W. A., 1964, K-Ar dating of rocks and tectonic events in the Piedmont of southeastern Pennsylvania: Geological Society of America Bulletin, v. 75, p. 661-668.

Lasaga, A. C. 1983, Geospeedometry: An extension of geothermometry. In: Kinetics and equilibrium in mineral reactions. Ed: Saxena, S. K. Springer-Verlag, New York, 81-114.

LeBreton, N. and Thompson, A. B., 1988, Fluid absent (dehydration) melting of biotite in metapelites in the early stages of crustal anatexis: Contributions to Mineralogy and Petrology, v. 99, p. 226-237.

Lister, G. S. and Snoke, A. W., 1984, S-C mylonites: Journal of Structural Geology, v. 6, p. 617-638.

Loomis, T. P., Ganguly, J. and Elphick, S. C., 1985, Experimental determination of cation diffusivities in aluminosilicate garnets: II. Multicomponent simulation and tracer diffusion coefficients: Contributions to Mineralogy and Petrology, v. 90, p. 45-51.

Luborsky, P., 1983, Petrology of route 141 outcrop between routes 52 and 48, Wilmington, Del. Unpublished report, University of Pennsylvania. 45 p

Mackin, J. H., 1962, Structure of the Glenarm Series in Chester County, Pennsylvania: Geological Society of America Bulletin, v. 73, p. 403-410.

Mark, L. E., 1977, Petrology and metamorphism in the Marcus Hook Quadrangle, southeastern Pennsylvania, M.A. thesis, Bryn Mawr College, 57 p.

McKinstry, H. E., 1961, Structure of the Glenarm Series in Chester County, Pennsylvania: Geological Society of America Bulletin, v. 72, p. 557-578.

Muller, P. D. and Chapin, D. A., 1984, Tectonic evolution of the Baltimore Gneiss anticlines, Maryland: Geological Society America Special Paper, v. 194, p. 127-148.

Myer, G. H., Baker, P. S., Cushing, B. R. and Hill, M. L., 1985, Metamorphism and ductile shear along the Martic "Zone" in southeastern Pennsylvania: Geological Society of America Abstracts with Programs, v. 17, p. 55.

Newton, R. C. and Haselton, H. T. 1981, Thermodynamics of the garnet-plagioclase-Al_2SiO_5-quartz geobarometer. In: Thermodynamics of Minerals and Melts. Eds: Newton, R. C., Navrotsky, S. and Wood, B. J. Springer Verlag, Berlin, 131-147.

Pavlides, L., 1981, The Central Virginia volcanic-plutonic belt: an island arc of Cambrian(?) age, United States Geological Survey Professional Paper 1231-A, United States Geological Survey, 34 p.

Perchuk, L. L. and Laverent'eva, I. V. 1983, Experimental investigation of exchange equilibria in the system cordierite-garnet-biotite. In: Kinetics and equilibrium in mineral reactions. Ed: Saxena, S. K. Springer-Verlag, New York, 199-239.

Plank, M. O., 1989, Metamorphism in the Wissahickon Formation of Delaware and adjacent areas of Maryland and Pennsylvania, M.S. thesis, University of Delaware, 126 p.

Postel, A. W., 1951, Problems of the Precambrian in the Phoenixville and Honey Brook quadrangles, Chester County, Pennsylvania: Pennsylvania Academy of Sciences Transactions, v. 25, p. 113-119.

Rand, T. D. (1887): A geological map of the serpentine ranges of Radnor Township, Delaware Co. and Lower Merion Township, Montgomery Co. to show their relation to the Philadelphia Belt. Geological Survey of Pennsylvania. .

Rdesinski, A., 1983, The contact between the Wilmington Complex gneisses and the Arden pluton, A.B. thesis, University of Pennsylvania, 18 p.

Shaw, H. F. and Wasserburg, G. J., 1984, Isotopic constraints on the origin of Appalachian mafic complexes: American Journal of Science, v. 284, p. 319-349.

Sheridan, R. E., 1989, Rootless Taconic suture under the New Jersey coastal plain: implications for accretion of Appalachian terranes: Geological Society of America Abstracts with Programs, v. 21, p. 65.

Sinha, A. K., Sans, J. R., Hanan, B. B. and Hall, S. T. 1980, Igneous rocks of the Maryland Piedmont. In: The Caledonides in the USA. Ed: Wones, D. R. Virginia Polytechnic Institute and State University Memoir, 2. Virginia Polytechnic Institute and State University, Blacksburg, VA, 131-136.

Song, T. S., 1987, The origin and implication of the steep gravity gradient in the vicinity of the Martic Zone, southeastern Pennsylvania, M.A. thesis, Temple University, 77 p.

Song, T. S. and Hill, M. L., 1988, The Martic Zone, southeastern Pennsylvania: a terrane boundary?: Geological Society of America Abstracts with Programs, v. 20, p. 71.

Southwick, D. L. 1969, Crystalline rocks of Harford County. In: The Geology of Harford County, Maryland. : Maryland Geological Survey, Baltimore, MD, 1-76.

Srogi, L., 1982, A new interpretation of contact relationships and early Paleozoic tectonic history of the Delaware Piedmont: Geological Society of America Abstracts with Programs, v. 14, p. 85.

---- 1988, The petrogenesis of the igneous and metamorphic rocks in the Wilmington Complex, Pennsylvania-Delaware Piedmont, Ph.D. thesis, University of Pennsylvania, 613 p.

Srogi, L., Lutz, T. M. and Wagner, M. E., 1991, The Wilmington Complex, PA-DE: Deep crust of a Cambrian arc?: Geological Society of America Abstracts with Programs, v. 23, p. 133.

Srogi, L., Wagner, M. E. and Lutz, T. M., in review, Disequilibrium during prograde dehydration partial melting: an example from the granulite facies Wilmington Complex, Pennsylvania-Delaware Piedmont: American Journal of Science, .

Sutter, J. F., Crawford, M. L. and Crawford, W. A., 1980, $^{40}Ar/^{39}Ar$ age spectra of coexisting hornblende and biotite from the Piedmont of Pennsylvania: their bearing on the metamorphic and tectonic history: Geological Society of America Abstracts with Programs, v. 12, p. 85.

Thomann, W. F., 1977, Igneous and metamorphic petrology of the Honey Brook Upland in the Elverson, Pottstown, and Phoenixville 7.5 minute quandrangles, southeastern Pennsylvania, M.A. thesis, Bryn Mawr College, 65 p.

Thompson, A. M., 1976, A summary of the geology of the Piedmont in Delaware: Transactions of the Delaware Academy of Science, v. 7, p. 115-134.

Tracy, R. J. 1982, Compositional zoning and inclusions in metamorphic minerals. In: Characterization of metamorphism through mineral equilibria. Ed: Ferry, J. M. Mineralogical Society of America Reviews in Mineralogy, 10. Mineralogical Society of America, Washington, 355-397.

Tracy, R. J., Robinson, P. and Thompson, A. B., 1976, Garnet composition and zoning in the determination of temperature and pressure of metamorphism, central Massachusetts: American Mineralogist, v. 61, p. 163-173.

Trautwein, J. W., 1983, Structural trends across the Precambrian/Cambrian boundary in the Honey Brook Upland of southeastern Pennsylvania, A.B. thesis, Bryn Mawr College, 38 p.

Tullis, J. and Yund, R. A., 1985, Dynamic recrystallization of feldspar: a mechanism for ductile shear zone formation: Geology, v. 13, p. 238-241.

Valentino, D. W., 1986, High-temperature deformation of gneiss in Delaware County, southeastern Pennsylvania: Geological Society of America Abstracts with Programs, v. 18, p. 73.

---- 1988a, The geometry and conditions of deformation of the Rosemont shear zone, southeastern Pennsylvania Piedmont, M.A. thesis, Temple University, 115 p.

---- 1988b, Rosemont shear zone, central Appalachian Piedmont: Dextral displacement of the Wilmington Complex: Geological Society of America Abstracts with Programs, v. 20, p. 77.

---- 1989, Evidence for dextral strike-slip shearing in the Pennsylvania Piedmont along the Susquehanna River: Geological Society of America Abstracts with Programs, v. 21, p. 72.

---- 1990, Post-Taconian structures of the western Piedmont province of Pennsylvania: The Tucquan antiform, the Lancaster Valley tectonite zone and the Peach Bottom structure. In: 55th Annual Field Conference of Pennsylvania Geologists. Ed: Scharnberger, C. K. Field Conference of Pennsylvania Geologists, Harrisburg, PA, 40-65.

---- 1991, The Peach Bottom shear zone: a new model for the Peach Bottom structure in southeastern Pennsylvania: Geological Society of America Abstracts with Programs, v. 23, p. 142.

Valentino, D. W. and Hill, M. L., 1991, Post-Taconic transpression in the Central Appalachian Piedmont of Pennsylvania, USA: Geological Association of Canada-Mineralogical Association of Canada Program with Abstracts, v. 16, p. A126.

Valentino, D. W. and Wiswall, C. G., 1991, Post-Taconian deformation in the Pennsylvania Piedmont and the tectonic implications: Geological Society of America Abstracts with Programs, v. 23, p. 143.

Valentino, R., 1989, Analysis of a shear zone in the southeastern Pennsylvania Piedmont: application of a conjugate pair model: Geological Society of America Abstracts with Programs, v. 21, p. 72.

Valentino, R. W. and Valentino, D. W., 1991, The Rosemont and Crum Creek conjugate shear zones of the Philadelphia area, Pennsylvania Piedmont: Geological Society of America Abstracts with Programs, v. 23, p. 143.

Wagner, M. E., 1972, Metamorphism of the Baltimore Gneiss in Southeastern Pennsylvania, Ph.D. thesis, Bryn Mawr College, 89 p.

GEOLOGICAL ASSOCIATION OF NEW JERSEY
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MINERAL IDENTIFICATION

BY: DR. FREDRIC R. GOLDSTEIN
DEPARTMENT OF PHYSICS
TRENTON STATE COLLEGE

In order to properly identify minerals, the following determinations should be made and recorded on the Mineral Identification Chart. **1. color 2. streak 3. luster 4. hardness 5. cleavage 6. density 7. miscellaneous determinations that you think might be applicable.**

If your mineral has a distinctly colored streak, use Table I. Further comparisons of your determinations with the information provided in Table I should help you to correctly identify the unknown mineral.

If your mineral does not have a distinctly colored streak, Use Table II, III, IV, V or VI. Relatively soft minerals (those that can be scratched by a penny) are listed in Table II. Minerals of intermediate hardness (those that can be scratched by a knife, but not a penny) are listed in Table III. Relatively hard minerals (those that cannot be scratched by a knife or a penny) are listed in Tables IV, V and VI. Further comparisons of your mineral's properties with those listed in Tables II through VI should help you to correctly identify the unknown mineral.

You will receive a number of unknown minerals to identify. Complete the Mineral Identification Chart on page 4. Your results will be checked at the end of the period.

GLOBAL TECTONIC INSIGHTS IN OCEANOGRAPHY

JOHN MARCHISIN

JERSEY CITY STATE COLLEGE
DEPARTMENT OF GEOSCIENCE
2039 KENNEDY MEMORIAL BOULEVARD
JERSEY CITY, NEW JERSEY 07305

INTRODUCTION

The following triad of exercises are suitable for middle school and high school Earth Science learners. The activities assume a unit in Oceanography containing material on ocean floor features and Global Tectonics has been or is concurrently being discussed in class. The exercises collectively involve the application of marine geology information and skills in map interpretation, graphing, and mathematics.

In Part A, Ocean Floor Topography, the learner must have access to a World Ocean Floor Map or facsimile. The questions involve the description of ocean floor features and the relationships between them.

Part B, Profile of the Ocean Floor, requires graph paper. Upon graphing the distance and water depth data, the student identifies the dominant basin features. After labelling the features, graphic analysis and math skills are used to develop gradient and water depth relationships.

Part C, Global Tectonics - Crustal Plates, uses a Physiographic Chart of the Sea Floor or world map in conjunction with the World Ocean Floor Map. The learner draws the plate boundaries and labels the crustal plate names. Using different colored arrows, the type and direction of motion are depicted.

While the presenter has field tested the activities, you the instructor should not take them for granite. Rather, use, change, rephrase, and adopt as you see fit.

References

Garrett, Gilbert M., ed., 1981, World Ocean Floor Map, National Geographic Society, Washington.

_______________, 1967. Physiographic Chart of the Sea Floor, Hubbard Scientific Company,Northbrook, Illinois.

PART A - **OCEAN FLOOR TOPOGRAPHY**

Survey the ocean floor map; note the configuration of all the ocean basins and the features associated with the ridge-rift system. Study the continental margins and their respective shelves and slopes throughout the world. Note the differences and changes in the various continental masses.

Now go on to the following specific questions. Answer in writing.

1. Define the following terms:

    a. continental shelf

    b. submarine canyon

    c. abyssal plain

    d. continental slope

    e. seamount

    f. submarine channel

    g. tablemount

    h. trench

2. What topographic features would suggest that continental crust is different from ocean floor crustal rock? Explain.

3. Where are most submarine canyons located? What is their probable origin?

4. What features are found on the floor beyond the mouths of all major estuaries?

5. How does the ridge-rift in the Pacific Ocean differ from that in the Atlantic?

6. On the basis of general ocean floor features, where would you expect the greatest number of earthquakes to occur?

PART A - continued

7. Where on the ocean floor would you expect to find:

   a. the oldest sediments

   b. the thickest accumulations of sediments

   c. the youngest rocks

   d. the area of greatest seismic activity

8. What feature is founded off the west coast of South and Central America?

9. Explain why Japan has numerous severe earthquakes while few occur in central Australia.

10. What types of submarine features exist immediately north and south of the Aleutian Islands? How were these features produced?

11. Trace the ridge-rift from the Indian Ocean to the Gulf of Aden. How does it seem to be related to the Red Sea and the Mediterranean?

12. How is the rift associated with the Sea of Cortez and California?

13. Find the Mariana Islands. What major feature is found there?

14. How is the rift valley in Africa related to ocean basis topography? Why?

PART B - **PROFILE OF THE OCEAN FLOOR**

Draw a profile of the Atlantic Ocean floor along the 39° N. parallel from Cape May, NJ to Cape Roca, Portugal.

Ocean depths from Cape May to Cape Roca in fathoms (1 fathom = 6 feet)

Distance	Depth	Distance	Depth
0 miles a	0 fathoms	2200 miles	1150 fathoms
100	100	2225	750
125	1000	2325	700
325	2000	2450	550
475	2500	2475 b	0
650	3000	2525	1000
900	2800	2700	2000
1125	2900	2800	2800
1300	3125	3150	2700
1425	2900	3300	2300
1475	2000	3400	1000
1600	1600	3425	500
1900	2500	3475	100
2000	2100	3500 c	0
2150	1800		

a = Cape May

b = Graciosa Island, Azores

c = Cape Roca

Part B - continued

Using 10 square per inch graph paper:

Suggested: Horizontal Scale of 1 inch = 500 miles (1 small square = 50 miles) and a Vertical Scale of 1 inch = 1000 fathoms (1 small square = 100 fathoms)

Questions:

1. The continental shelf break is 100 fathoms deep. How many feet is this?

2. How wide is the shelf at Cape May? At Cape Roca?

3. How does the shelf on the east coast of North America compare to the west coast of Europe?

4. Vertical Exaggeration = $\frac{\text{Feet per inch in horizontal scale}}{\text{Feet per inch in vertical scale}}$

   Compute the vertical exaggeration on your graph.

5. Accurately label the continental shelves, slopes, and rises on both sides of the profile.

6. Identify the position of the ridge-rift system and label the rift.

7. Calculate the gradient in feet per mile for both shelves and slopes.

8. Calculate the average depth of the profile.

9. Calculate and compare the percent profile depth below 2000 fathoms with the percent profile depth at or above 2000 fathoms.

10. How does the calculated average profile depth (#8) compare with the world "average" ocean depth of 2000 fathoms?

PART C - GLOBAL TECTONICS - CRUSTAL PLATES

Part A: On the World Map

1. Accurately draw in the Plate Boundaries such that the plates are clearly identified.

2. Name and label (name) each plate.

3. Using three different colored pencils, draw arrows showing direction of movement at the boundaries as evidence of Divergence (Red), Convergence (Blue) and Shear (Green).

   Do this for each Ocean.

Part B: Based on Part A above.

1. Name an ocean which is getting larger. Cite evidence.

2. Name two other bodies of water that are growing. Explain.

3. Cite an area where continental mountain building is happening. How is this depicted?

4. List three specific locations where the ocean floor is being destroyed.