CRYSTALLINE BEDROCK GEOLOGY OF THE
LOWER SUSQUEHANNA GORGE: CONOWINGO TO HAVRE DE GRACE,
MARYLAND

by
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The crystalline bedrock of the lowermost Susquehanna River Gorge, Conowingo to Havre de Grace, Maryland, consists of two discrete structural blocks, each with its own unique history prior to juxtaposition. The southern Havre de Grace Block is a Cambrian magmatic arc association (James Run Formation and Port Deposit Intrusive Complex), possibly developed on a rift fragment of Laurentia (Canal Road Formation). The northern Conowingo Block represents a precursory mélange (Conowingo Dam Formation) intruded during the Ordovician by a layered mafic complex (State Line Mafic Complex) and associated plutons (Basin Run Tonalite). Many olistoliths within the Conowingo Dam Formation could have originated in the Havre de Grace Block. The Havre de Grace Block was metamorphosed and deformed under amphibolite facies conditions during Middle Ordovician to Early Silurian time. It was subsequently thrust upon the Conowingo Block along the Elbow Branch Thrust, probably during the Late Ordovician or Early Silurian. From as early as the Middle Devonian, through the Pennsylvanian, dextral strike-slip shearing took place under greenschist facies conditions along the Rock Run Shear Zone, within the northern part of the Havre de Grace Block. Analysis of SC fabrics in the zone yields a minimum offset of 2 to 6 kilometers. From Late Paleozoic to Mesozoic time, strike-slip shearing gave way to dip-slip shearing across very thin, low grade, ductile shear zones.
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This thesis is dedicated to the three women in the author's life: his wife Zenah, and his two daughters, Travertine Juniata and Naomi Arcadia.

"Me, I'll sit and write this love song, as I all too seldom do. Build a little fire at midnight, it's good to be back home with you."

Ian Anderson
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Chapter One

Crystalline Bedrock Geology of the Lower Susquehanna Gorge: Conowingo to Havre de Grace, Maryland

1. Introduction
Rocks of the most southeastern part of the northeast-trending Central Appalachians crop out in spectacular fashion along the lower Susquehanna Gorge, providing an exceptional natural laboratory in which to study the geologic history of the crystalline core of the central part of the Appalachian orogenic belt.

1.1. Regional Geologic Setting
Figure 1.1 is a generalized geologic map of the northern Central Appalachian Piedmont, distributed symmetrically for approximately 100 kilometers along strike on either side of the Susquehanna River. The map pattern is characterized by discrete fault-bounded structural blocks, each with its own internal tectonic lithofacies association and dynamometamorphic history. Blocks northwest of the Pleasant Grove - Huntington Valley Shear Zone form the Westminster Terrane, and are clearly autochthonous to North America, exhibiting a Late Proterozoic/Early Paleozoic rift to drift metasedimentary cover sequence deposited on Grenville basement (Valentino and others, 1994). The James Run/Chopawamsic Terrane, volcanic and plutonic rocks along the extreme southeast margin of the Piedmont, clearly represent a Cambrian (Sinha and others, 1989) magmatic arc. The geologic history of the intervening rocks - those southeast of the Pleasant Grove/Huntington Valley Shear Zone and northwest of the paleo-magmatic arc - is hotly debated.
Figure 1.1- Regional Geologic Setting

Although in a general sense most authors agree that the Piedmont's history is one of Late Cambrian/Early Ordovician accretion and orogeny (Taconic) followed by Late Paleozoic (Acadian? and Alleghanian Orogeny) dextral transcurrent tectonics, they strongly disagree as to both the details of either event and the relative impact of the Paleozoic orogenies on producing the geologic structures and map distribution observed.
today. Three major tectonic models exist: the **magmatic evolution model**, the **accretionary model**, and the *Taconic collision-Alleghanian transpression* model.

In the **magmatic evolution model**, the geologic history is discussed in terms of three discrete thermal events, each accompanied by plutonism, metamorphism, and deformation (Sinha and others, 1989; Sinha and others, 1979). The **oldest event** is the Cambrian (520-490 Ma) magmatic arc plutonism and volcanism of the Chopawamsic Terrane. It is considered to be the result of a subduction zone of debatable polarity, developed on the present-day eastern margin of Laurentia, the ancestral North American continent characterized by Grenville (~1.1 Ga) Age basement rock (Sinha and others, 1979). The Middle Ordovician Taconic Orogeny represents a **second event** (460 to 440 Ma) signifying the collapse of the continental margin arc and its obduction onto Laurentia. The **final thermal event** is Perm-Carboniferous in age (330-270), and is associated with the Alleghanian Orogeny- collision of Laurentia with Africa.

The **accretionary model** (e.g. Horton and others, 1989) interprets the Central Appalachian Piedmont as having evolved primarily by Late Cambrian through Middle Ordovician accretion of allochthonous tectonostratigraphic terranes or tectonic motifs, culminating in the Taconic Orogeny. Each motif is separated by a major thrust fault and characterized by a precursory syntectonic sedimentary mélangé derived from the obducted hanging wall thrust sheet (Drake, 1985). In this model, the James Run/Chopawamsic Terrane is the structurally highest, most outboard tectonic allochthon. Subduction polarity during arc magmatism is uncertain. Pavlides (1981) considered subduction in Virginia to be east-facing (i.e. - west dipping) based on distribution of
volcanic rocks within the terrane. However, basin closure and accretion of the Chopawamsic Terrane to Laurentia is more consistent with a west-facing (i.e. -east-dipping) subduction zone, and, at least in the Central Appalachian Piedmont north of Virginia, is modeled as such by some authors (Gates and others, 1991; Muller and others, 1989). Depending on subduction polarity, the Chopawamsic Terrane was separated from Laurentia by either an oceanic (Iapetus) or marginal basin when the arc was active. Recently, proponents of the accretionary model have acknowledged that Alleghanian dextral strike-slip shear zones both cross-cut the accreted terranes and reactivate the sutures between them. This dextral strike-slip shearing is relegated a relatively minor role in producing the modern map pattern (e.g. Drake, 1995).

Conversely, the **Taconic collision-Alleghanian transpression** model (Valentino and others, 1994; Gates and others, 1991) treats Alleghanian dynamometamorphism as at least equal in importance to Cambro-Ordovician accretion and subsequent orogeny (Taconic). In this model, the Central Appalachian Piedmont is a collection of structural blocks juxtaposed across dextral strike-slip shear zones. Minimum offset estimates across individual shear zones range from a typical 25 kilometers on the Rosemont Zone (Valentino,1988) to over 150 kilometers on the Pleasant Grove-Huntingdon Valley Shear Zone (Valentino and others, 1994). Furthermore, although the structural blocks internally preserve earlier, presumably Taconic, metamorphic assemblages and structural surfaces, the metamorphic assemblages are variably retrograded and the structural surfaces sheared and folded during the Alleghanian Orogeny. Within individual blocks, many map scale structures, considered dominantly Taconic in origin under the accretionary model, are
considered to be more a result of Alleghanian tectonism. Major structures of this type include the Tucquan Antiform of the Westminster Terrane and the Baltimore Gneiss Domes of the Baltimore Terrane (Valentino and others, 1994; Gates and others, 1991; Gates, 1996). Under this model, the record of Cambro-Ordovician orogeny (Taconic, culminating in the Middle Ordovician) is preserved within discrete blocks as distinctive lithofacies assemblages, older structural surfaces, and generally peak metamorphic assemblages. However, instead of multiple terranes as in the accretionary model, this model allows for only one allochthonous terrane - the Cambrian James Run/Chopawamsic Terrane. The single terrane boundary - the Taconic Suture - is marked by the Liberty Complex, a polygenetic combination of sedimentary and tectonic mélange (Muller and others, 1989). The presence of multiple outcrop belts of mélange is attributed to structural duplication of a single mélange complex.

Although detailed geochemical and isotopic studies are certainly useful, testing the models described above requires integrating the results of such studies with a variety of very fundamental geologic data: field relations, petrography, lithofacies association, geologic structures, and metamorphic assemblages. Prior to this study, such data was ambiguous for rocks of the Susquehanna Gorge southeast of the State Line Complex.

1.2. Local Geology

Preexisting published geology maps of the lower Susquehanna Gorge are summarized in Figure 1.2. The radically different mapped distribution of bedrock on either side of the river, particularly from Conowingo Dam south (study area, boxed in Figure 1.2), underscores the difficulty in integrating the geology of the area into the
regional framework. The geology of the Susquehanna Gorge southeast of Conowingo consists primarily of variably foliated and sheared sodic granitoid, various proportions of which have been called Port Deposit Gneiss (Higgins, 1990; Southwick, 1969; Hershey, 1937). Southwick mapped a part of the Port Deposit Gneiss just south of the central supracrustal belt (described below) as Sheared Port Deposit Gneiss.

Samples from the traditional type section of Port Deposit Gneiss, an abandoned quarry along the Susquehanna River just northwest of the town of Port Deposit, yield an isotopic age of 515 Ma (U-Pb; Sinha, 1989), which is interpreted as an age of igneous crystallization. An Rb-Sr whole rock age of 467±21 Ma is interpreted to represent the timing of regional (Taconic) metamorphism (Lesser, 1982). Lesser (1982) calculated two ages (316 Ma and 380 Ma) based on Rb-Sr analyses of biotite separates. Whole rock Rb-Sr on Southwick’s Sheared Port Deposit Gneiss yielded an age of 360 ± 65 Ma (Lesser, 1982). Kohn and others (1993) report an Rb-Sr biotite age of 191 Ma for the Port Deposit Gneiss.

Three different lithostratigraphic associations of metamorphosed supracrustal rock occur as irregular northeast-trending belts surrounded by the granitoid. The most southeastern belt consists of metamorphosed island arc volcanics, mapped as James Run Formation (Higgins, 1990) and as Cecil County Volcanics (Southwick, 1969). The James Run Formation is Cambrian in age (516 Ma, U-Pb; Lesser, 1982). An Rb-Sr whole rock age of 430±21 Ma is interpreted as the age of peak metamorphism (Lesser, 1982).
Figure 1.2 - Summary of Pre-existing Geologic Maps of Lower Susquehanna Gorge

The central belt consists of complexly deformed metaturbidites, and has been mapped as Wissahickon Formation (Southwick, 1968) and as metagraywacke with amphibolite (no formation name; Higgins, 1986).

The northwestern belt consists of massive granofels with internally foliated inclusions of schist, graywacke, amphibolite, calcisilicate gneiss, and ultramafic rock, and is mapped as Boulder Gneiss Facies of the Wissahickon Formation by Southwick (1968) and as Conowingo Diamicite by Higgins (1986). It is generally agreed to be metamorphosed submarine debris flow deposits (Hopson, 1964). However, the close resemblance of this rock to a xenolith-rich granodiorite has caused considerable confusion (see Figure 1.2) over the extent of the mélange (Southwick, 1968; Higgins,
The olistolith assemblage has led most authors to conclude that, tectonically, these rocks were associated with terrane accretion, and represents a precursory mélange associated with the obduction of the James Run Terrane with either the Bel Air-Rising Sun Terrane (Higgins, 1990; Horton and others, 1989) or the Westminster Terrane (Gates and others, 1991).

Whole rock Rb-Sr analysis of Higgins’ Conowingo Diamictite produced ages of 473±38 Ma (Lesser, 1982). All but one of the analyzed samples were from outcrops previously mapped as Port Deposit Gneiss.

Mafic intrusive rocks are present as well. Quartz gabbro of the Ordovician State Line Mafic Complex (490±20 Ma; Nd-Sm; Shaw and Wasserburg, 1984) crops out in the northeastern portion of the study area, just below and northwest of Conowingo Dam (Southwick, 1968). Fine- to medium- grained, intermediate mafic dikes are common in the southeastern portion of the field area, (Higgins, 1986). The Aberdeen Metagabbro crops out in the extreme southwest portion of the study area. Diabasic dikes are present locally. Tongues of Cretaceous and Tertiary sands and gravels, not show on Figure 1.2, cover the crystalline bedrock in many upland areas (see Figure 1.3).

The nature of the contacts drawn between units varies radically across the river, as shown in Figure 1.2. Higgins (1986) in Cecil County considered contacts to be dominantly structural, characterized by a stack of thrust sheets. Higgins (1990) considered the Port Deposit Gneiss a hypabyssal pluton intruding its own volcanic ejecta, the James Run Formation. Conversely, Southwick (1968) considered nearly all contacts intrusive in nature, and the Port Deposit Gneiss a deep-seated pluton.
These discrepancies have hampered the integration of the rocks of the lower Susquehanna Gorge into tectonic models of the evolution of this portion of the Central Appalachian Piedmont. The objective of this study was to resolve this conflict by collecting and analyzing the very basic geologic data discussed above, and evaluate the results in the context of the regional tectonic models described above.

2. **Methods**

Detailed mapping of the Lower Susquehanna River Gorge was conducted on a 1:24,000 USGS topographical base. Ninety days were spent in the field, mapping, measuring structures, and collecting oriented samples for petrographic analysis. Approximately 150 thin sections were subjected to petrographic and microstructural analysis. Modal abundances of minerals were based on visual estimation. Stereographic analyses of field fabric measurements were integrated with thin section and outcrop petrographic observations to relate specific measurements to specific dynamometamorphic episodes. Plagioclase compositional estimates were based on standard flat stage Michel-Levy optical technique.

3. **Data**

Plate 1 is the new geologic map of the lower Susquehanna River Gorge. The map is summarized in Figure 1.3. Lithologic description of map units is presented first (Section 3.1), followed by structure and metamorphism (Section 3.2). Petrographic data is tabulated in Appendix A. Appendix B is a list of outcrop locations for photos and sketches shown in the text. Appendix C shows maps to outcrop locations.
Figure 1.3 - Generalized Geologic Map of Lower Susquehanna Gorge
3.1. Description of Units

3.1.1. Canal Road Formation

The pre-Cambrian to Cambrian Canal Road Formation (CZ_{CR}) consists of interlayered garnet + staurolite schist, fine-grained, greenish gray metaturbidites, and, locally, fine-grained, dark gray to black amphibolite. It crops out as a northeast trending wedge shaped body separating the highly foliated Port Deposit Gneiss (C_{PD}) to the southeast from the weakly foliated, xenolith rich Basin Run Granodiorite (O_{BR}) to the northwest (see Figure 1.3 or Plate 1). The Canal Road Formation is assigned a Cambrian or older age based on the isotopic age (515Ma, Sinha, 1989) of the Port Deposit Gneiss, which intrudes it. To the southwest of the study area, the Canal Road Formation pinches out along the Rock Run Shear Zone near the contact between the Port Deposit Gneiss and the Baltimore Gabbro-State Line Mafic Complex (Southwick, 1969). To the north and northeast of the study area, outcrop becomes sparse along a drainage divide.

The southeast contact of the Canal Road Formation with the Port Deposit Gneiss is complex (see Figure 1.4), and lies within the Rock Run Shear Zone. Abundant pelitic and psammitic xenoliths, possibly of Canal Road Formation, occur within the Port Deposit Gneiss. Irregular dikes of granodiorite intrude Canal Road Formation near the contact. Garnet is much more common within where the Port Deposit Gneiss at or very near the contact with the Canal Road Formation, where the Port Deposit Gneiss becomes progressively finer-grained as well. One could argue that these last two features may be due, in part, to shearing along the Rock Run Shear Zone (RRSZ) (Chapter 2, this thesis), within which the contact lies. However, retrograde assemblages are commonly
associated with RRSZ fabrics in both the Port Deposit Gneiss and the Canal Road Formation. Furthermore, petrographic examination of microstructures reveals that $S_2C$ angles within the Port Deposit Gneiss do not decrease near the contact, nor does the rock exhibit an overall more sheared appearance. Thus, both the fine grain size and increase in abundance of garnet probably mark the original chilled margin of the protolith pluton. As the contact is followed along strike to the southwest, it enters a high strain part of the RRSZ and most of the original igneous contact features disappear.

![Figure 1.4 - Canal Road Formation - Port Deposit Gneiss Contact](image)

The northwest contact with the Basin Run Tonalite is highly sheared. The lack of fine-grained plutonic rock along this contact is an important difference between it and the southeastern contact with the Port Deposit Gneiss.

Higgins (1986) postulated a thrust fault contact to the northeast separating the Canal Road Formation from pelitic schist underlying a poorly exposed drainage divide.
Four types of outcrops are present outside of the Rock Run Shear Zone: 1) interlayered graded argillaceous feldspathic quartzite (5-30 cm thick) and garnet schist (generally 3-5 cm thick, locally up to 30 cm), 2) highly aluminous staurolite garnet schist in which all primary structures are obliterated, and 3) alternating layers of quartzite (5-20 cm thick) and garnet schist (3-10 cm thick) that contain some amphibolite layers (.75 m through 2.5 meters thick), and 4) outcrops of massive amphibolite. Because of poor exposure, the relationship between these rock types is undetermined. Both pelitic and psammitic layers locally pinch out within one another. A sharp contact between rock types 3 & 4 above is locally exposed along the margin of the Rock Run Shear Zone, and may represent an initial depositional contact modified by later shearing.

Despite metamorphism to lower amphibolite facies and multiple episodes of deformation, many primary sedimentary structures can be seen in the Canal Road Formation. These structures include bedding ($S_0$), erosional bed boundaries, relict graded bedding, and dewatering cones. Such structures are generally best preserved within the Canal Road Formation along regional strike to the northeast, and across strike to the northwest. The best place to observe these is along Basin Run northeast of Liberty Grove.

Figure 1.5 illustrates two of the primary sedimentary features preserved in the Canal Road Formation - flame structures and dewatering cones. Flame structures, representing remnant surfaces of erosion by produced by the turbidity currents which deposited the overlying graded bed, are common throughout the Canal Road Formation. Curvilinear surfaces, superficially similar to cross-beds, within the psammitic layers are
consistently concave in a single direction within the outcrop, and resemble dewatering cones observed in modern sediments. Where such structures are observed, the upper surface of psammitic layers is disrupted, locally causing individual layers to look simultaneously upright and overturned.

Graded beds are present only cryptically, at outcrops where internally graded 10-20cm thick psammitic layers are separated by pelitic layers 3-6 cm thick. Metamorphism and deformation has produced an inverse grading within psammitic layers at these locations (See Figure 1.6). Psammitic layers are aggregates of polygonally intergrown, fine-grained (d<1mm) quartz and plagioclase with various proportions of biotite defining the foliation. Within individual psammitic layers, both the proportion of biotite and the foliation intensity increase, and the aluminous phases chlorite, garnet and muscovite appear, each increasing in abundance toward the top of a layer. The thin, aluminum rich pelitic layers show no internal grading.

Near their bases, psammitic layers are dominantly quartz with up to 15% plagioclase, minor garnet, and generally less than 15% phyllosilicates. Neither alkali feldspars nor lithic fragments were observed. Grain size is small (d<1mm), and the quartz and plagioclase grains are intergrown in a mosaic texture. The increase in abundance of aluminous phases within a layer, described above, represents original
compositional grading. Near their tops, the graded layers are comprised of up to approximately 50% aluminous phases.

Figure 1.6 - Grade Bedding in the Canal Road Formation

Most schist samples contain a 18% to 50% quartz, and are generally semi-pelites.

Garnet is the highest grade index mineral present at all outcrops, while staurolite is found only in the most aluminous samples from outcrops of massive schist. Muscovite, chlorite, and biotite are present in varying amounts, and collectively constitute 40% - 70% of a sample. Opaque phases generally constitute less than 5% of the rock.

Although amphibolite layers where present are concordant with psammitic and pelitic layers, no primary structures are preserved within them or along their boundaries. Their occurrence and orientation suggest interlayered basalt flows. However, the possibility of a post-depositional intrusive origin cannot be ruled out. Petrographically, they are generally quartz-bearing hornblende plagioclase amphibolite containing
accessory titanite overgrowing an opaque phase, most likely ilmenite. The plagioclase composition is An$_{35-40}$.

The high degree of deformation, metamorphism, and recrystallization hampers attempts at detailed protolith determination for the Canal Road Formation. Boundaries between individual clasts within the protolith turbidites were completely recrystallized during dynamometamorphism. Flattening of layers probably accompanied recrystallization and may have differentially reduced layer thicknesses, thinning the less competent schist layers relative to the stronger graded psammitic layers. However, preserved primary structures and association of rock types makes an attempt at protolith estimation possible.

The association of psammitic layers - exhibiting flame structures, graded beds, dewatering cones, and possible cross-beds - separated by thin layers of pelitic schist and interlayered with massive pelitic schist and amphibolite is consistent with a protolith that was a sequence of thin, distal turbidites (graded psammites) separated by mud drapes (pelitic and semi-pelitic schist), and interbedded with basalt flows (interlayered amphibolite). The mineral assemblage of the psammitic layers - dominantly quartz (> 70%) and plagioclase (<20%) - points to a moderately mature sedimentary protolith. However, such a general lithofacies association is not by itself definitive, particularly when it comes to assigning an upper limit for the age of the Canal Road Formation or determining the tectonic setting of deposition.

The pervasive recrystallization of the psammitic layers prevents proper application of the provenance discriminant diagrams of Dickinson (1985), which are
based on the detrital modes of sandstones. Although a quantitative application of these diagrams is unreasonable because of the lack of preserved discrete sedimentary grains, a qualitative estimate based on the mineralogy of the recrystallized psammites places all analyzed samples of the Canal Road Formation along the quartz - total feldspar join of the quartz - total feldspar - lithic fragment ternary diagram. While primary structures are quite common, no relict lithic fragments were observed. This suggests that there were no lithic fragments in the protolith, and that the qualitative position of the metamorphic rock on the ternary plot may be a reasonable approximation of the protolith's position. The provenance fields corresponding to the position of the Canal Road Formation are transitional continental and craton interior (Dickinson, 1985). Turbidites of such provenance, particularly when associated with interlayered basalt flows, are quite commonly associated in the geologic record with a continental rift tectonic setting (Dickinson et al, 1983). Therefore, the Canal Road Formation may represent metamorphosed continental rift turbidites.

Correlation of the Canal Road Formation to other metasedimentary rocks of the Maryland Piedmont is problematic. Rocks which compositionally resemble the Canal Road Formation, such as the Oella Formation north of Baltimore (Avery Drake, personal communication), lack primary sedimentary structures indicative of deposition by turbidity currents. One the other hand, the Peters Creek Formation, thick metaturbidites of demonstrated continental rift affinity, interlayered with basalt, differs significantly from the Canal Road Formation in both bulk and clast composition, containing an abundance of potassium feldspar clasts and lithic fragments (Valentino, 1993).
Furthermore, the Peters Creek Formation and Canal Road Formation are separated by
major thrust and possibly strike-slip shear zones. Until such time as a correlation to other
formations is demonstrated, the Canal Road Formation should be considered
independently as a sequence of compositionally mature, distal metaturbites interlayered
with pelitic schist and basalt, probably deposited in a continental rift tectonic setting.

3.1.2. **Sheared Ultramafic Rock**

Well-foliated, fine-grained serpentinite intensely veined with asbestos crops out at a
small quarry near the entrance to Susquehanna State Park Campground, Harford County.
It occurs as a small lensoidal body along the contact between the Canal Road Formation
and the Port Deposit Gneiss, within the Rock Run Shear Zone. Protolith determination
is impossible due to both the limited exposure and the high degree of shearing. It
probably represents one of three possible protoliths: an ultramafic layer or sill in the
Canal Road Formation, a large xenolith in the Port Deposit Gneiss, or a fragment of a
larger ultramafic body cut by the Rock Run Shear Zone along strike. The mineral
assemblage is serpentine, calcite, chlorite, relict hornblende, and talc, with some
unidentified opaque phases.

3.1.3. **Fine-grained Amphibolite**

Fine-grained, variably sheared amphibolite lenses up to one thousand feet long and three
hundred feet wide crop out within and adjacent to the Rock Run Shear Zone. They
consist of an aggregate of hornblende, plagioclase, epidote, quartz, and opaques with
fibrous actinolite bridging fractured amphiboles and chlorite along discrete shear
surfaces. No primary features providing clues to protolith were observed. This unit is locally associated with sheared ultramafic rock (IIIB above), and may be genetically related. However, all fine-grained amphibolite within and adjacent to the Rock Run Shear Zone may not be of the same origin. This unit may be polygenetic, with some exposures related to the ultramafic rock above, and others related to the amphibolite dikes which cut the Port Deposit Gneiss and James Run Formation between Port Deposit and Havre de Grace, Maryland. Only the larger occurrences of this formation are plotted on Plate 1.

3.1.4. James Run Formation
The Cambrian James Run Formation crops out in an irregular 2-3 km wide, northeast striking belt (see Figure i.3 or Plate 1) of aphanitic to porphyritic, interlayered dacitic, basaltic, and, to a lesser extent, andesitic gneiss. The Cambrian age is based on a U-Pb zircon upper intercept age of 516 Ma for the James Run Formation calculated by Lesser (1982) applying the Steiger and Jager (1977) decay constants to the data of Tilton (1970), from analysis of samples from the Gatch Quarry in Churchville, Maryland.

Thickness of individual layers within the James Run Formation ranges from a few millimeters to greater than a meter. A 0.5 to 1.5 km wide belt of dominantly fine-grained, locally pillowed amphibolite crops out near the northwest margin (Gilpins Falls Member of Higgins (1990)). Numerous thin lenses of coarse amphibolite ($C_{CA}$) intrude the James Run Formation, usually within mafic layers, xenoliths of which are common. The James Run Formation is intruded to both the northwest and southeast by the Cambrian Port Deposit Gneiss. The contact is highly sheared. However, xenoliths of
James Run in the Port Deposit Gneiss and dikes of Port Deposit Gneiss within the James Run Formation near the contact suggest the contact was originally intrusive. The southeast contact is less sheared, locally marked by an igneous intrusion breccia, and other places very sharp (see Figure 1.7) Along strike to the northeast, the James Run is clearly intruded by the Tom's Creek Trondhjemite. Igneous intrusion breccias are preserved along that contact as well, and are weakly foliated. To the southwest, the Aberdeen Metagabbro has intruded the James Run Formation (Southwick, 1968).

Figure 1.7 - The Southeastern James Run Formation - Port Deposit Gneiss Contact
(Arundel Corporation Quarry, Havre de Grace, MD)

Compositional layering is well preserved in the James Run Formation, and represents original layering. Primary volcanic structures are common. Large (up to 1 cm) euhedral to subhedral quartz (beta), plagioclase, and hornblende crystals are present.
and appear to be relict phenocrysts. Flattened, relict amygdules are abundant in the mafic James Run. Epidote filled spherulites as much as a few centimeters in diameter are common as well. Volcanic flow breccias are locally preserved in the mafic James Run. Higgins (1990) described well-preserved pillow structures outside of the present study area along strike to the northeast.

Felsic specimens of the James Run Formation are aphanitic to porphyritic with a variably foliated fine-grained quartz rich matrix containing up to forty percent porphyroclasts. The matrix consists of quartz (35-80%) ± plagioclase (An_{28}, 0 - 35%) ± hornblende (0 - 10%) ± opaques (mainly magnetite, 9-10%) ± alkali feldspar (perthite, 0 - 10%) ± epidote (0-35%) + biotite (0-10%) ± muscovite (0-10%) ± chlorite (0-9%) ± garnet(0-1%) ± actinolite (trace amounts). Porphyroclast (relict phenocryst) phases are dominantly quartz, plagioclase, and hornblende. In most samples, quartz and plagioclase are strongly recrystallized, and what appear as single porphyroclasts in the field are actually aggregates of polygonally intergrown grains. In the least deformed samples, both quartz (beta form) and plagioclase occur as euhedral to subhedral crystals and are interpreted as relict volcanic phenocrysts. Blue-green hornblende occurs in elongate crystals as much as a centimeter long, which are randomly oriented to strongly aligned. Blocky, equant, anhedral hornblende crystals are present locally, and may be relict cores of primary amphibole phenocrysts. Small amounts of anorthoclase, fluorite, and magnetite and are present locally, and appear to be relict phenocryst phases as well. Garnet and anhedral epidote occur as metamorphic porphyroblasts. Chlorite locally rims garnet.
Mafic rocks within the James Run Formation consist of fine- to medium-grained, generally well-foliated and lineated amphibolite, locally containing as much as ten percent porphyroclasts. The matrix consists of 50-75% anhedral amphibole (dominantly blue-green hornblende with as much as 10% actinolite), 5-10% anhedral plagioclase (An$_{33}$), 0 -10% opaques, 0-13% epidote, 0-15% biotite, and 0 - 5% chlorite. Hornblende and plagioclase appear euhedral in less deformed specimens, and may be relict volcanic phenocrysts. Common elliptical cavities filled with coarse crystalline epidote, plagioclase, calcite, and amphibole are flattened amygdules.

Less common andesitic layers consist of 30-50% hornblende + quartz + plagioclase ± opaques ± epidote ± biotite ± chlorite ± garnet ± actinolite, and have plagioclase, hornblende, and/or quartz porphyroclasts, which may be relict volcanic phenocrysts.

Previous authors agree that the James Run Formation represents a metamorphosed volcanic arc sequence(e.g. Higgins, 1990; Lesser, 1982). Observations made in this study fully support the volcanic arc protolith interpretation. Higgins (1971) suggests subaqueous deposition of the James Run volcanioclastics. Handy-Barringer (1983) suggests that this deposition was in a forearc because of the James Run Formation’s boninitic chemical affinities.

3.1.5. Old Mill Tectonite
The Old Mill Tectonite is quartzo-feldspathic blastomylonite which crops out along Principio Creek between the Tom’s Creek Trondhjemite to the southeast and the Port
Deposit Gneiss to the northwest. Contacts are not exposed, but are probably highly sheared, based on the character of this unit.

Petrographically, the Old Mill Tectonite is characterized by millimeter scale porphyroclasts of rounded to locally euhedral plagioclase, rounded quartz, radial sphene with ilmenite cores, and, locally, potassium feldspar. These porphyroclasts lie in a fine-grained, polygonal matrix of quartz, plagioclase, biotite, epidote, and apatite with opaques present locally. Very small inclusions of schist are common, as are small lensoidal mafic. Although the protolith is unknown, bulk composition and map distribution of units support the notion that the Old Mill Tectonite is most likely either sheared James Run Formation or sheared Tom’s Creek Trondhjemite, and is thus Cambrian in age.

3.1.6. The Port Deposit Intrusive Suite

3.1.7. Aberdeen Metagabbro of the Port Deposit Intrusive Suite

Medium to coarse-grained amphibolite of the Cambrian Aberdeen Metagabbro crops out in the southwestern corner of the study area. The age is based on the fact that it both intruded the Cambrian James Run Formation and was itself intruded by the Cambrian Port Deposit Gneiss. It was described by Southwick (1969) as weakly to strongly lineated epidote amphibolite and “epidiorite” with locally well-preserved gabbroic textures. Southwick (1969) lists the characteristic mineral assemblage hornblende, plagioclase, epidote, and opaques with or without quartz, chlorite, sphene, rutile, apatite, and clay minerals (presumably a recent, near surface alteration product).
The Aberdeen Metagabbro was not investigated in detail during this study. Observation of outcrops near Havre de Grace, where xenoliths of the James Run Formation are common within the Aberdeen Metagabbro and dikes of Port Deposit Gneiss cross-cut the it, confirmed Southwick's (1969) contact relations. Coarse-grained lenses of cummingtonite-bearing amphibolite which intruded the James Run Formation in Cecil County may or may not be Aberdeen Metagabbro.

3.1.8. The Port Deposit Gneiss of the Port Deposit Intrusive Suite
The Cambrian Port Deposit Gneiss consists dominantly of foliated tonalite and granodiorite (see Figure 1.8). Minor amounts of quartz-rich metatonalite is present within intensely foliated zones, and quartz diorite was observed locally. The Cambrian age is based on a U-Pb upper intercept age calculated from analyses of zircons from the abandoned Port Deposit Quarry along the Susquehanna River just northwest of the town of Port Deposit (Sinha, 1989). The northwest contact of the Port Deposit Gneiss with the Canal Road Formation appears to be an original igneous contact, subsequently strongly sheared (see section 3.1.1 above). The contact lies entirely within the Rock Run Shear Zone, and is offset across centimeter spaced subvertical northeast-striking microscopic shear zones (see Chapter 2, this thesis). The Port Deposit Gneiss clearly intrudes the James Run Formation. Xenoliths of amphibolite and hornblende quarzite are common. Igneous contact breccia is present locally and is rich in partially digested xenoliths. In many places, however, the contact is intensely foliated (e.g. - the northwest boundary of James Run Formation in vicinity of the Susquehanna River), and original contact relations are unclear. Overall, the contact appears to be an original intrusive contact
variably modified by subsequent shearing. Along strike to the southwest, the Port Deposit Gneiss intrudes the Aberdeen Metagabbro (Southwick, 1969). Schist, volcaniclastic gneiss, and amphibolite are common as xenoliths in the Port Deposit Gneiss, and probably stem from the Canal Road Formation, the James Run Formation, and the Aberdeen Metagabbro, respectively.

![Diagram of IUGS Classification of felsic phases of the Port Deposit Intrusive Suite](image)

Figure 1.8 - IUGS Classification of felsic phases of the Port Deposit Intrusive Suite, Fields after Streckheisen (1979)

The characteristic mineral assemblage of the Port Deposit Gneiss is plagioclase, quartz, biotite, and clinopyroxene, with or without muscovite, microcline, garnet, chlorite, hornblende, zircon, apatite, and opaques. Plagioclase occurs dominantly in centimeter scale aggregates, equant to ribbon in shape, of polygonally intergrown, anhedral millimeter and smaller plagioclase crystals(An$_{25-37}$) exhibiting albite twinning (see Figure 1.9). Large (centimeter scale), zoned subhedral plagioclase crystals are present, but very rare, and exhibit Carlsbad and albite twinning. Quartz occurs in ribbons of fine-grained (millimeter and smaller) anhedral polygonal crystals lying within foliation domains. In
the less foliated rocks, biotite occurs in clumps (relict books?) surrounding clinozoisite. Generally, however, it lies within and helps define the foliation. Muscovite occurs within foliation planes. Chlorite occurs only in or near subvertical shear zones which cross-cut the pervasive foliation. Small, euhedral garnets are present locally, especially in the more foliated portions. Microcline occurs in small, anhedral masses interstitial to plagioclase and quartz, and is most common near subhedral plagioclase crystals, to which it is sometimes connected by perthitic intergrowths. Hornblende occurs locally in association with biotite. Zircon and apatite are present as accessory phases and occur as inclusions in biotite and plagioclase, respectively. Sericite and epidote occur as alteration products of plagioclase (saussurite). The inferred probable primary igneous assemblage was, in the order of crystallization, apatite and zircon, followed by plagioclase and clinozoisite, then biotite, quartz, and finally microcline. Biotite, muscovite, garnet, and chlorite are metamorphic phases. Note that biotite formed both during igneous crystallization and metamorphism.

The northeastern third of the Port Deposit Gneiss lies within the Rock Run Shear Zone, and has a well-developed, pervasive SC fabric (see Chapter 2, this thesis). Chlorite is more abundant in the Port Deposit Gneiss within the shear zone, as is magnetite. Traditionally, the Port Deposit Quarry in Cecil County has served as the de facto type section for the Port Deposit Gneiss. The quarry is located just northwest of
the town of Port Deposit, in the middle of the Rock Run Shear Zone. Because of the strongly polydeformational history of the Port Deposit Gneiss within the shear zone, the Port Deposit Quarry is probably one of the worst places in the relict pluton for sample collection for chemical analysis. Fluids associated with movement of the Rock Run Shear Zone could have significantly altered bulk and possibly trace element chemistry, at least locally. The high number of quartz veins within the Port Deposit Gneiss in the shear zone is consistent with the presence of abundant shear zone fluid. Furthermore, while southeast of the Rock Run Shear Zone plagioclase within the Port Deposit Gneiss has a composition of An_{25-37}, within the zone plagioclase is dominantly An_{05}, with a few less deformed samples having the composition An_{28}. Nevertheless, the majority of samples analyzed in geochemical studies of the Port Deposit Gneiss have been collected from the Port Deposit Quarry (e.g. Kohn and others, 1993; Davis, 1992; Sinha and others, 1989; Lesser, 1982).

3.1.9. **Tom's Creek Trondhjemite of the Port Deposit Intrusive Suite**

Tom’s Creek Trondhjemite is a weakly foliated, medium-grained, idiomorphic to hypidiomorphic granular tonalite with less than ten percent mafic phases, thus meeting the definition of trondhjemite (Streckheisen, 1979) (see Figure 1.8). It crops out in Cecil County as two small plutons, one along the Susquehanna River and the other along Principio Creek. Xenoliths of hornblende-bearing felsite of the James Run Formation are common, especially in igneous intrusion breccia, which is well exposed along the southeast contact of the pluton exposed on Principio Creek. The pluton on Principio
Creek is bound to the northwest by the Old Mill Tectonite of the Rock Run Shear Zone. This contact is not exposed.

The smaller pluton of Tom’s Creek Trondhjemite along the Susquehanna River is bound to the northwest, northeast, and southeast by mafic and felsic James Run Formation, which it intruded. Across the river along strike to the southwest crops out Port Deposit Gneiss. The contact between the Tom’s Creek Trondhjemite and the Port Deposit Gneiss is not exposed.

The age of the Tom’s Creek Trondhjemite is inferred to be Cambrian based on cross-cutting relationships, it intruded the Cambrian James Run Formation and was itself intruded by Cambro-Ordovician amphibolite dikes (described below), and petrographic similarity to the Cambrian Port Deposit Gneiss.

Petrographic analysis of the Tom’s Creek Trondhjemite revealed the mineral assemblage plagioclase, quartz, chlorite, biotite, and clinzozeite with or without muscovite, microcline, chlorite, allanite, opaques, zircon, and. Plagioclase (An_{27}) crystals (see Figure 1.10) up to a millimeter or more in long dimension are anhedral to subhedral, highly saussuritized, sometimes exhibit Carlsbad and/or albite twinning, and frequently enclose clinzozeite inclusions. Clinzozeite is present as both anhedral crystals in plagioclase and as anhedral to subhedral, locally zoned crystals associated with biotite and chlorite. Quartz occurs both
as recrystallized phenocrysts a few millimeters in size and as interstitial matrix to plagioclase and other quartz. Anhedral muscovite is found within and helps to define foliation planes. Anhedral chlorite is common both within and outside of foliation domains. Trace amounts of microcline is present in small, anhedral masses interstitial to plagioclase and quartz. Hornblende is locally found with biotite. Zircon and apatite are present as accessory phases and occur as inclusions in biotite and plagioclase, respectively. Sericite and epidote occur as alteration products of plagioclase (saussurite). The observed textures suggest the primary igneous assemblage was, in order of crystallization, apatite, zircon, clinzoizesite, and allanite, followed by plagioclase, then quartz and biotite, and finally microcline. Biotite, muscovite, epidote, and chlorite are metamorphic phases.

3.1.10. Amphibolite Dikes of the Port Deposit Intrusive Suite

A swarm of anastomosing, mafic to andesitic dikes of amphibolite generally less than three meters thick intruded the Port Deposit Gneiss, the James Run Formation, and the Tom’s Creek Trondhjemite. They may have intruded the Canal Road Formation as well. The dikes dip steeply southeast (see Figure 1.11). They contain hornblende phenocrysts in a matrix of plagioclase, quartz, potassium feldspar, epidote, sphene, and opaques. Within the Rock Run Shear Zone, they are finer grained and have chlorite plus actinolite in the matrix. South of the Rock Run Shear Zone, they are ubiquitous. They have both an internal foliation, parallel at some outcrops to that in their country rock and at other outcrops parallel to their margins, and locally show a steeply east plunging ductile mineral lineation. Their contacts cross-cut the foliation at low angles.
Figure 1.11 - Typical Dike of Amphibolite intruding the Port Deposit Gneiss (looking northeast)

A Cambro-Ordovician age is inferred for the amphibolite dikes, because they intruded Cambrian rocks and were deformed and metamorphosed along with their country rock during the Ordovician. The age of metamorphism is based on whole rock Rb-Sr ages determined by Lesser (1982) from samples of the James Run Formation (430±21 Ma) and Port Deposit Gneiss (467 ±21 Ma; 456 ± 21 Ma).

3.1.11. Conowingo Dam Formation

The Conowingo Dam Formation consists of medium-grained granofels with abundant inclusions of bedded psammitic schist, pelitic schist, metadacite, laminated calcsilicates, epidote amphibolite, and ultramafic fragments (see Figure 1.12). It crops out as a thin wedge (600m wide) striking northeast along the intrusive contact between quartz gabbro of the State Line Mafic Complex and the Basin Run Tonalite, both of which intruded it. The Conowingo Dam Formation was formerly mapped as
Wissahickon Boulder Gneiss in Cecil County by Southwick (1969) and as the Conowingo Diamictite by Higgins (1986).

![Image of block in matrix structures in the Conowingo Dam Formation](image)

- **Figure 1.12 - Block in matrix structures in the Conowingo Dam Formation**

Parts of the Conowingo Dam Formation closely resemble a granitoid rich in xenoliths. The following observations suggest the sedimentary nature of its protolith, and that the inclusions represent olistoliths in a protolith mélange deposit. Inclusions show no digestion rind as one would expect of xenoliths in a pluton, nor do they exhibit hornfelsic textures. The matrix itself contains only fractured, anhedral quartz and feldspar, and never the euhedral crystals characteristic of the plutons (described below) which intruded it. Furthermore, the appearance of the matrix is strongly heterogeneous compared to typical intrusive textures.

However, much of the bedrock previously mapped in Cecil County (Higgins, 1986) as Conowingo Diamictite has proven in this study to indeed have an igneous origin (see Basin Run Tonalite, Section 3.1.9), with euhedral igneous plagioclase, igneous
textures such as interstitial microcline, and xenoliths exhibiting well-developed digestion rinds. Because of this discrepancy between the previously mapped distribution of metamorphosed mélange and that determined in this study, the name Conowingo Diamictite is not used. Furthermore, the name has never been wholly appropriate because of its obvious glaciogenic connotations, and the overall failure of the rock to even meet the descriptive criteria for a diamictite.

Internally, olistoliths retain features which predate their incorporation as olistoliths into the Conowigo Dam Formation. Disruptive structures associated with soft sediment deformation are common within psammitic clasts. Most clasts bear an internal tectonic foliation.

A detailed analysis of this formation was beyond the scope of this study. However, several thin sections were examined to determine composition and metamorphic grade. The granofels matrix consists dominantly of anhedral quartz, locally rutile, anhedral plagioclase, and anhedral opaques. Biotite, epidote, and chlorite are present as secondary phases, with biotite defining the weak foliation. Specimens of the psammitic olistoliths exhibit the same mineral assemblage as the granofelsic matrix, but are much finer grained. Olistoliths of both fine- and coarse-grained epidote amphibolite were analyzed, and have the characteristic assemblage hornblende, epidote, plagioclase, quartz, and opaques with actinolite and chlorite. Metadacitic olistoliths consisted of finely interlayered, fine-grained epidote bearing quartzites.

The Conowingo Dam Formation probably correlates to the Sykesville Formation as described by Muller (1989), which crops out to the northwest, structurally below the
State Line Mafic Complex. An important similarity between the Conowingo Dam Formation and the Sykesville Formation is that both the granofelsic matrix and the psammitic olistoliths within each of them contain very little to no alkali feldspar. The granofels matrix in each is dominantly quartz and plagioclase. Gates and others (1991) interpret the Sykesville Formation as a metamorphosed precursory mélangé deposit associated with the collision between Baltimore Terrane and the Westminster Terrane.

State Line Mafic Complex - Upper Part (3.1.12 through 3.1.13)

3.1.12. Quartz Gabbro and Quartz Diorite of the State Line Mafic Complex

Quartz gabbro and quartz diorite of the Early Ordovician State Line Mafic Complex, as mapped by Southwick (1968), crops out just below Conowingo Dam in both Cecil and Harford Counties. These rocks intruded the Conowingo Dam Formation and are themselves intruded by the Basin Run Tonalite. To the northwest, the quartz gabbro and quartz diorite grades into metagabbro. The Ordovician age is based on a Sm-Nd of 490±20 Ma (Shaw and Wasserburg, 1984). Upper intercept zircon ages of 1,200 Ma (Shaw and Wasserburg, 1984) are interpreted to represent incorporation of country rock zircons into the State Line Mafic Complex magma.
The quartz gabbro and quartz diorite was not studied in detail in this investigation. However, the field relations described above were verified, and thin section petrography of several samples was performed. Although locally sheared, much of the quartz diorite has a nearly pristine igneous appearance in the field (see Figure 1.13). The results of petrography concurred with the assemblage of Southwick (1969): plagioclase (strongly zone, An₈₅₋₄₅) and quartz, with or without clinopyroxene, orthopyroxene, green hornblende, and biotite. Fibrous pale-green amphibole and epidote are ubiquitous as secondary phases. Locally, rims of nearly pure, unaltered albite are present on plagioclase.

3.1.13. Basin Run Tonalite of the State Line Mafic Complex

The Ordovician Basin Run Tonalite consists of nonfoliated to weakly foliated tonalite and granodiorite, and is rich in xenoliths of rocks from the Conowingo Dam Formation (see Figure 1.14), which it intruded. The abundance of xenoliths is presumably one reason why Higgins (1986) mapped much of the Basin Run Tonalite as Conowingo Diamictite, metamorphosed mélange. Another reason may be the presence of shear zones, described in Section 3.2.4 below, which drastically alter the appearance of the tonalite or granodiorite protolith.
Within the area mapped as Basin Run Tonalite in this study, the xenolithic nature of the inclusions is clear. They have well-developed digestion rinds and are commonly cross-cut by dikelets of Basin Run Tonalite. Furthermore, the matrix to the inclusions is clearly plutonic in nature. It has a generally uniform igneous texture with abundant igneous microtextures, described below. Locally, hanging wall pegmatite is observed, clearly indicating an igneous origin, and suggesting that the Basin Run Tonalite formed by multiple magma injections. The Basin Run Tonalite intruded the quartz gabbro and quartz diorite of the State Line Mafic Complex as well.

To the southeast, the contact of the Basin Run Tonalite with the Canal Road Formation is highly sheared. A well-developed dextral strike-slip SC fabric virtually identical to that of the Port Deposit Gneiss is present along this contact and extends a few hundred meters into the Basin Run pluton. This boundary, however, probably has a polydeformational history, because it juxtaposes pervasively deformed amphibolite facies rocks to the south with relatively pristine, middle greenschist facies rocks to the north. Because of the orientation of the $S_1$ surface south of this contact (see Section 3.2.2 below), and because of regional considerations, this contact is inferred to have originated as a southeast-dipping thrust surface, probably Taconic in age.
The Ordovician age of the Basin Run Tonalite is based on an Rb-Sr whole rock age of 473±38 Ma (Lesser, 1982) calculated for the Conowingo Diamictite of Higgins (1990). All but one of these samples come from outcrops mapped in this study as Basin Run Tonalite. Because the Basin Run Tonalite lacks of a pervasive deformation, and contains a mineral assemblage (described below) indicative of metamorphism to only middle greenschist facies, the Rb-Sr date is interpreted here to correspond to the age of igneous crystallization.

Samples of Conowingo Diamictite analyzed by Sinha and others (1971) probably came from areas mapped in this study as Basin Run Tonalite. The upper limit zircon age for these samples is 1,200 Ma, identical to that of the State Line Mafic Complex.

Petrographically, the Basin Run Tonalite is characterized by the assemblage plagioclase, quartz, biotite, microcline, and clinopyroxite with or without muscovite, chlorite, ilmenite, allanite, sphene, opaques, zircon, apatite, and garnet. Figure 1.15 shows typical igneous textures. Plagioclase (An$_{39}$) occurs as strongly zoned, saussuritized, euhedral to subhedral rectangular crystals which exhibit both carlsbad and albite twinning. Albite rims free of alteration are common, and may reflect minor reequilibration under greenschist facies metamorphic conditions. Quartz occurs as equant aggregates (up to a centimeter) of weakly recrystallized (millimeter-scale) anhedral polygonal crystals, and typically exhibits sweeping extinction and deformation bands. Biotite occurs in books surrounding clinopyroxite and titanite. Alignment of biotite locally defines a weak foliation. Muscovite occurs with biotite, and as an alteration product of plagioclase. Small amounts of chlorite are present locally in clumps
with biotite. Small, euhedral garnets are present at a single outcrop, in a sheared pegmatite vein. Microcline occurs in small, anhedral masses interstitial to plagioclase and quartz, and is most commonly found adjacent to plagioclase crystals, to which it is typically connected by perthitic intergrowths. Reddish brown, well-zoned, euhedral allanite crystals up to a few mm in size occur locally, and are always rimmed by thin radial growths of clinozoisite. Zircon and apatite are present as accessory phases and occur as inclusions in biotite and plagioclase, respectively. Sericite and epidote occur as alteration products of plagioclase. The order of igneous crystallization, based on nearly pristine igneous textures, was zircon and apatite, followed by plagioclase and allanite, sphene and clinozoisite, biotite, quartz and possible muscovite, and finally microcline. Biotite, muscovite, chlorite, and epidote are metamorphic phases.

![Photomicrographs showing igneous textures in the Basin Run Tonalite](image)

**Figure 1.15 - Photomicrographs showing igneous textures in the Basin Run Tonalite**

Based on contact relationships observed in the field, thin section petrography, and pre-existing geochronology, their is most likely a strong genetic relationship between the State Line Mafic Complex proper and the Basin Run Tonalite. The State Line Mafic
Complex intrudes the Conowingo Dam Formation and is itself intruded by the Basin Run Tonalite. The Basin Run Tonalite itself may represent either the final felsic differentiate of a common parent magma or the felsic top of a single body of relatively primitive magma, differentiated in situ. A detailed petrogenetic analysis of the relationship is beyond the scope of this study. However, the similarity in ages of igneous consolidation - 490±20 Ma for the State Line Mafic Complex compared with 473±38 Ma for the Basin Run Tonalite - is consistent with the above hypothesis. Furthermore, the overlapping ages of crystallization coupled with the fact that both units yield zircon upper intercept ages of 1,200 Ma indicates that both units most likely intruded through the same age crust in the same tectonic setting.

3.1.14. Diabase

A thin dike of fine-grained, Mesozoic diabase crops out along an unnamed stream in Susquehanna State Park. It is undeformed, with well-developed diabasic textures, and a plagioclase - pyroxene igneous assemblage. It is assigned a Mesozoic age based on similarity to other such dikes throughout the piedmont. Other such intrusions may be present in the project area, and were either missed or misidentified as amphibolite during mapping.

3.1.15. Cretaceous and Tertiary Gravels and Sands

Unconsolidated Cretaceous and Tertiary gravels and sands crop out in uplands near the Susquehanna River, and southeast in the coastal plain. These units were not part of this investigation. Map distribution is from Southwick and Owens (1968) and Higgins and Conant (1986).
3.1.16. *Quaternary Alluvium*

Quaternary alluvium deposits are present locally. Only large deposits of this kind are shown on the geologic map. Such deposits were not part of this investigation.

3.2. **Structure and Metamorphism**

Deformation and metamorphism are both complicated and broadly synchronous in the study area. Determination of the relative timing of dynamometamorphic events is critical for development of a tectonic model. For these reasons, structure and metamorphism of the above described rocks will be considered simultaneously in this report, in sequence from oldest to youngest foliation surface or metamorphic assemblage.

3.2.1. *Inclusions trails of quartz in garnet (S0.5)*

Garnets in both the Canal Road Formation and the James Run Formation contain abundant inclusions of quartz aligned in lines within the planes of thin sections (see Figure 1.16). Since this linear alignment is present regardless of thin section orientation, the inclusions are inferred to lie in planes. Within individual trails, the quartz grains are generally asymmetric, with their long axes lying parallel to the trails. These features reflect a preexisting foliation overgrown by the garnets. The inclusion trails are assigned the ambiguous label $S_{0.5}$ because of the difficulty in determining whether or not they correlated to a fabric in the matrix of the rock outside of the garnets. Within the James Run Formation, they lie parallel to $S_1$ (Section 3.2.2). Within the Canal Road Formation, they appear randomly oriented relative to $S_1$ (Section 3.2.1), and appear to define folds when trails from nearby garnets are projected through the matrix. However, the $S_1$
surface in the Canal Road Formation is highly deformed by movement along the Rock Run Shear Zone (Section 3.2.3 and Chapter 2), and the lack of $S_1 - S_{0.5}$ parallelism may be due to this subsequent deformation. Within the Canal Road Formation, staurolite overgrows both garnet and $S_1$, which is consistent with the inclusion trails representing $S_1$ overgrown by the peak $M_1$ metamorphic assemblage.

Within the Canal Road Formation immediately adjacent to the Rock Run Shear Zone, inclusion trails show slight curvature near the outer edge of garnets, indicating that some garnet growth may have occurred during early stages of movement along the Rock Run Shear Zone (Chapter 2).

![Image: Microstructures and garnet zone assemblage in the Canal Road Formation](image)

**Figure 1.16 - Microstructures and garnet zone assemblage in the Canal Road Formation**

3.2.2. - $D_1M_1$, *Amphibolite facies dynamometamorphism ($S_1M_1L_1$)*

All Cambrian and older rocks southeast of the Basin Run Tonalite exhibit a well-developed foliation associated with amphibolite facies metamorphism. Within both the Canal Road Formation and the James Run Formation, this foliation is parallel to
compositional layering. In addition, a well developed ductile lineation is present in the Port Deposit Gneiss, James Run Formation, and Aberdeen Metagabbro (Southwick, 1969).

Within the Canal Road Formation, schistosity and gneissosity parallel to bedding define S₁. Both alignment of phyllosilicates (biotite or muscovite) and segregation of minerals into quartzofeldspathic and phyllosilicate domains are present in all specimens. Possible folds of S₀ intrafolial to S₁ are present locally. Conversely, these structures may represent folded quartz veinlets. S₁ is associated with peak amphibolite facies metamorphic conditions. Stable assemblages are garnet - biotite- chlorite - muscovite in psammites, staurolite-garnet - chlorite - muscovite or garnet-chlorite-biotite-muscovite in pelitic schist, and hornblende - plagioclase(An₃₇) in amphibolite.

Within the more aluminous layers of the Canal Road Formation, garnets are overgrown by randomly oriented staurolite crystals up to 5 cm long which also overgrow S₁. Interlayered amphibolites are weakly to non-foliated. This indicates that the M₁ metamorphic peak postdates the formation of S₁. This is consistent with S₀₅ corresponding to an S₁ overgrown by metamorphic minerals. Similar relationships where the peak metamorphic assemblage overgrows the layer parallel (S₁) schistosity, are common regionally (Lang, 1986).

S₁ within the James Run Formation is defined by alignment of schistosity, alignment of amphiboles, or gneissosity, depending on composition of the individual layer. Boudins of individual layers and flattened amygdules are common, and lie in S₁. A strong lineation is present locally along foliation surfaces, defined both by preferred
alignment of hornblende crystals and by a mineral stretching lineation, and plunges steeply east. \( S_1 \) is associated with the peak metamorphic assemblages blue-green hornblende, plagioclase (\( \text{An}_{35} \)), and epidote within metabasalts and garnet, biotite, and possibly chlorite in the metadacites. These assemblages are indicative of peak metamorphic conditions in the lower amphibolite metamorphic facies (Spear, 1993).

Higgins (1990) mapped the sequence as isoclinally folded, with the northwest portion overturned. No evidence for this was observed in this investigation. However, Higgins may have had better way up indicators in the James Run Formation along strike to the northeast.

The peak \( M_1 \) metamorphic assemblage cummingtonite, hornblende, and plagioclase characterized the coarse-grained amphibolite which intrudes the James Run. Alignment of \( M_1 \) amphiboles defines a steeply east-plunging mineral lineation.

The \( S_1 \) fabric in the Port Deposit Gneiss is characterized in the field by a foliation defined both by alignment phyllosilicate phases and by ribbons of recrystallized quartz and plagioclase (see Figure 1.17). Substantial variation in fabric intensity is observed in both outcrop and thin section, with grain size reducing as foliation strength increases (see Figure 1.18). The fabric strikes northeast and dips moderately to steeply southeast. Fine-grained muscovite and biotite lie in and help define the foliation plane. More steeply
dipping and intensely foliated portions appear to have a type I SC fabric (Berthe and others, 1979) with S nearly wrapped into C, yielding an oblique reverse sense of shear. A well-developed mineral streaking lineation rakes steeply east along the S₁ foliation surface. The long axes of ellipsoidal shaped deformed xenoliths parallel this lineation.

The Tom's Creek Trondhjemite is characterized by a weak foliation defined by alignment of phyllosilicate phases. This foliation is approximately parallel to the more well-developed S₁ in other units south of the Rock Run Shear Zone.

The Old Mill Tectonite is dominated by structures related to the Rock Run Shear Zone (D₂M₂ below). However, S₁ and L₁ are preserved as highly flattened, elongate xenoliths whose long axes define a steeply plunging lineation.

Figure 1.19 presents orientation of S₁ and L₁ south of the Rock Run Shear Zone. Note the similarity in orientation of these features across different units. Presumably, S₁ within the Canal Road Formation had a similar orientation before it was tightly folded during the D₂M₂ event described below.

The S₁ surface north of the Canal Road Formation - Basin Run Tonalite contact is defined by the internal foliation of both olistoliths in the Conowingo Dam Formation and xenoliths in the State Line Mafic Complex and Basin Run Tonalite. Depending on the clast or xenolith, the grade of metamorphism ranges from greenschist to epidote amphibolite.
Figure 1.18 - Microtextures in the Port Deposit Gneiss

A. Low Intensity S1 (dipping to east) in Port Deposit Gneiss; note aggregate of plagioclase

B. Saussurite aggregate outlining original feldspar grain, rounded during deformation (same field of view as A)

C. High Intensity S1, Port Deposit Gneiss

D. Polygonally Intergrown aggregate of plagioclase

E. High Intensity S1, Port Deposit Gneiss; extremely fine-grained

F. Close-up of Highly Foliated Port Deposit Gneiss; some grains are as small as .01 mm.
A. Orientation of S, (419 poles, contoured) and L, (39 data) in Port Deposit Gneiss south of the Rock Run Shear Zone

B. Orientation of S, (149 poles, contoured) and L, (18 data) in James Run Formation

C. Orientation of S, (42 poles, contoured) and L, (14 data) in James Run Formation

D. Orientation of S, (20 planes)

Figure 1.19 - S, and L, southeast of Rock Run Shear Zone
3.2.3. - \textit{D}_{2}\textit{M}_{2}. Dextral strike-slip shearing under greenschist facies conditions and associated deformation (S\textsubscript{2}C and S\textsubscript{2}; M\textsubscript{2}; L\textsubscript{2})

Rocks within and northwest of the Rock Run Shear Zone are affected by the D\textsubscript{2}M\textsubscript{2} event. The lack of a significant D\textsubscript{2}M\textsubscript{2} overprint on rocks southeast of the Rock Run Shear Zone is probably related to the bulk composition and prior history of the James Run Formation and Port Deposit Gneiss.

The northwest third of the Port Deposit Gneiss lies within the Rock Run Shear Zone (Chapter 2) and is characterized by a well-developed dextral strike-slip, type 2 SC fabric (S\textsubscript{2}C) (Berthe and others, 1979) which cross-cuts and generally obliterates S\textsubscript{1} (see Figure 1.20. The S\textsubscript{2} surfaces strike north by northeast and dip steeply southeast. S\textsubscript{2} is offset in a dextral strike-slip sense across discrete, closely (centimeter) spaced ductile shear zones (C) which strike east by northeast and dip very steeply northwest and southeast. Metamorphic minerals helping define the S\textsubscript{2} and C surfaces are biotite, muscovite, and chlorite. One, two, or all three of these phases are found locally within both S and C bands, and no systematic variation in their distribution was documented.

Sheared Port Deposit Gneiss is the dominant rock type in the Rock Run Shear Zone (Chapter 2, this thesis).
An important difference between $S_1$ and $S_2$C in the Port Deposit Gneiss is that in $S_1$, quartz and plagioclase are about equally flattened, indicating similarly ductile behaviour during $D_1$. In $S_2$C, however, only quartz is highly flattened. This is inferred to reflect a higher temperature of deformation associated with $D_1$ than with $D_2$ (Tullis and Yund, 1985).

C planes of the Rock Run Shear Zone define the dominant foliation surface in the Old Mill Tectonite. This unit is highly sheared, and reflects the complex deformational history of the Havre de Grace Block. Abundant syn deformational quartz veins and porphyroblasts reflect a high degree of fluid activity associated with $D_2$ (see Chapter 2).

SC mylonite is developed in the southeastern part of the Basin Run Tonalite as well, and defines a 100-200 meter wide northwestern splay off of the Rock Run Shear Zone.

Within the Canal Road Formation, between the Rock Run Shear Zone and its northwestern splay, $S_1$ foliation surfaces are tightly crenulated and folded along with bedding, and $M_1$ metamorphic assemblages are slightly retrograded, with biotite and chlorite rimming and embaying garnets. A spaced foliation ($S_2$) is axial planar to the folds. $S_2$ is approximately parallel to the $S_2$ component of the $S_2$C mylonite of the Port Deposit Gneiss in the Rock Run Shear Zone.

Crenulations and folds plunge steeply northeast. Layers on limbs of folds all dip steeply, and are frequently overturned. Interlimb angles of folds and crenulations are highly variable both by location and by rock type, varying on the outcrop scale from $<30^\circ$ to $\sim 135^\circ$, and are typically on the order of $60^\circ$. Crenulation is most strongly developed
in the more pelitic layers where the pre-existing $S_1$ is stronger. Fold amplitude also varies with composition and thickness, with thinner, more pelitic layers more tightly folded, and ranges from the scale of decimeters to tens of meters. Order of magnitude of crenulation amplitude is cm to mm scale. $S_2$ axial planar cleavage frequently appears to obliterate $S_1$ within the pelitic layers, while $S_1$ is preserved, essentially undisturbed, within psammitic layers (refer back to Figure 1.16).

Photomicrographs of the pelitic layers show $S_1$ tightly folded (interlimb angle < 30°) with the axial planes of these folds defining $S_2$ (refer back to Figure 1.16). Pressure shadows around garnets are common and lie within $S_2$. Chlorite and biotite porphyroblasts commonly occur with basal planes aligned parallel to $S_2$, but locally randomly overgrow $S_2$. Retrograde biotite and chlorite on garnet and the relative low intensity of the $S_2$ fabric when compared to $S_1$ are consistent with $D_2$ occurring under greenschist facies (biotite grade) metamorphic conditions ($M_2$). Randomly oriented chlorite and biotite porphyroblasts locally overgrow $S_2$.

Where the Canal Road Formation lies within the boundaries of the Rock Run Shear Zone proper, $S_1$ is completely obliterated, and $M_1$ metamorphic assemblages are only cryptically preserved (see Chapter 2). The foliation lies between $S_2$ and $C$.

Sheared amphibolite and ultramafic rock within the Rock Run Shear Zone along the Canal Road Formation-Port Deposit Gneiss contact in Harford County show a fabric parallel to $C$, with SC structures present locally in the ultramafic rock.

Discrete northeast-striking dextral strike-slip shear zones cross-cut the Basin Run Tonalite, and are interpreted as splays off of the Rock Run Shear Zone. Within these
zones, the tonalite is converted to a mylonite with rounded quartz porphyroclasts in a
fine-grained green matrix of epidote and chlorite. Plagioclase is nearly obliterated.

Figure 1.21 presents stereographic plots of orientations of fabric elements
associated with the D2M2 event. Figure 1.21A presents the average $S_2C$ orientation of
the sheared Port Deposit Gneiss. Detailed analysis of all fabrics is presented in Chapter
2. Figure 1.21B shows the SC fabric in the Basin Run Tonalite in the northwest splay of
the Rock Run Shear Zone Figure 1.21C shows fabric elements in the Canal Road
Formation, and that folding of $S_1$ is consistent with deformation by the Rock Run Shear
Zone, as the resultant $S_2$ surfaces are roughly parallel to the $S_2$ component of $S_2C$. Figure
1.21D shows the weak foliation in the Basin Run Tonalite, correlated $S_2$ of $S_2C$, as well
as the orientation of discrete dextral strike-slip shear zones which anastomose across the
Conowingo Block.

3.2.4. - D3. Dip-slip faulting under greenschist and lower metamorphic conditions

Subvertical, dip-slip shear zones cross-cut bedrock both north and south of the
Rock Run Shear Zone. These zones are generally low grade, are associated with
abundant quartz and quartz plus epidote veins, and yield both reverse and normal senses
of shear, with the northwest side generally moving up relative to the southeast side.

Microstructural kinematic analyses of oriented thin sections show a strong dip-
slip normal sense of shear across S1 planes in more intensely foliated samples of the
James Run Formation. Retrogradation under greenschist facies conditions is documented
by chlorite rims on garnet and actinolite, biotite, and chlorite after hornblende. The
degree of retrogradation is much higher in rocks with good dip-slip kinematic indicators.
The dip-slip motion along $S_1$ is thus related to remobilization of $S_1$ and $S_0$ surfaces in the James Run Formation under retrograde metamorphic conditions. Kinematic indicators present include feldspar dominoes on plagioclase (Fig 1.22), sigma tails on relict quartz phenocrysts (Fig 1.22), sympathetic fracturing and extension of hornblende crystals, imbrication of hornblende crystals at a low angle to $S_1$, and intrafolial folds (Hanmer and Passchier, 1991) (Fig 1.22). A single sample with a steeply west-dipping foliation yielded a reverse sense of shear. Mesoscopically, asymmetric tails on quartz-epidote augen and pods of vein quartz and plagioclase yield a normal sense of shear as well.

Northeast-striking subvertical dip-slip shear zones cross-cut $S_1$ within the Port Deposit Gneiss (refer back to Figure 1.17). Chloritized fault surfaces and veins of quartz and epidote are associated with these zones. $S_1$ wraps into these zones yielding a consistent northwest up/southeast down sense of shear in both reverse and normal zones.

Similar dip-slip shear zones cross-cut the Basin Run Tonalite, and generally exhibit a reverse sense of shear with the northwest side moving up. Discrete subvertical shear zones in the Conowino Dam Formation and the State Line Mafic Complex may be related to the same event. Figure 1.23 summarizes orientations of $D_3$ fabric elements.
A. Contours of $S_2$ and $C$ surfaces in the Port Deposit Gneiss (~150 data, 16, 32, and 64 times uniform); average $S,C$ fabric and 4 lineations are shown as well.

B. Plots of 15 $S$ (gray) and $C$ (black) planes in the Basin Run Tonalite

C. Contour Plot of poles to $S_2$ in the Canal Road Formation (104 data); best fit great circle and pole as black plane and large gray circle; $S_2$ planes (16 data) as grey lines; $F_3$ folds (15 data) as circles; crenulation of $S_2$ shown as triangles

D. Contour Plot of poles to $S$ in Basin Run Tonalite (87 data; average $S$ shown as a black plane); discrete dextral strike-slip shear zones shown as gray planes

Figure 1.21 - Orientations of $D_2M_2$ Fabric Elements
A. Sigma Tails on Relict Quartz Phenocryst
B. Fractured and Extended Plagioclase Domino (note Quartz Fracture fill)
C. Intrafolial fold of opaque-rich layer, qtz-epidote mylonite

Figure 1.22 - D_3 Kinematic Indicators in the James Run Formation

A. D_3 Shear Zones in the Port Deposit Gneiss (25 data)
B. D_3 Shear Zones in the Basin Run Tonalite (9 data)

Figure 1.23 - Orientations of D_3 Shear Zones
4. Interpretations

4.1. The Havre de Grace and Conowingo Structural Blocks

The restriction of amphibolite facies metamorphism and an associated pervasive foliation (D1,M1) to units southeast of the Basin Run Tonalite indicates that the crystalline bedrock of the lower Susquehanna Gorge belongs to at least two structural blocks separated by the Basin Run Tonalite - Canal Road Formation contact. The structural block southeast of the contact is here defined as the Havre de Grace Block, and belongs to the Chopawamsic Terrane of Horton and others (1989). It includes, from oldest to youngest, the Canal Road Formation, the James Run Formation, the Aberdeen Metagabbro, the Port Deposit Gneiss, the Tom's Creek Trondhjemite, and dikes of amphibolite. The structural block north of the Basin Run Tonalite - Canal Road Formation contact is here named the Conowingo Block. It includes, from oldest to youngest, the Conowingo Dam Formation, the State Line Mafic Complex, and the Basin Run Tonalite. The Conowingo Block probably belongs to the Potomac Terrane of Horton and others (1989), based on the striking similarity of the Conowingo Dam Formation to the Sykesville Formation, which is the precursory mélange underlying the composite Potomac Terrane proper (Muller and others, 1989).
A key isotopic difference between the two blocks is that the intrusive rocks of the Conowingo Block have zircon system upper intercept ages of 1,200 to 1,300 Ma, much older than their Ordovician ages of crystallization (Shaw and Wasserburg, 1984; Sinha, 1971). Shaw and Wasserburg (1984) also report that the geochemical signature of the Baltimore Gabbro - State Line Complex indicates intrusion through continental crust. These upper intercept isotopic ages are strikingly similar to the 1,260 Ma age of the Baltimore Gneiss, which is interpreted to represent metamorphism associated with the Middle Proterozoic Grenville Orogeny (Shaw and Wasserburg, 1984). Grenville ages are characteristic of the autochthonous basement gneiss of eastern North America, so it can be reasonably inferred that the plutonic rocks of the Conowingo Block intruded through North American crust, while volcanic and plutonic rocks of the Havre de Grace did not.

The differences in geologic history, described above, between the two blocks is summarized in Figure 1.24 along with preexisting geochronology. When considered simultaneously, they point to a convincing model for the geologic evolution of the lower Susquehanna Gorge.
Figure 1.24 - Tectonic Evolution of the lower Susquehanna gorge
4.2. **Cambrian volcanic arc sequence metamorphosed to amphibolite facies - The Havre de Grace Block**

In the Havre de Grace block, the Pre-Cambrian to Cambrian Canal Road Formation was deposited first, probably in a continental rift setting. Later, the Cambrian James Run Formation Arc volcanics were deposited. The James Run Formation was intruded by the Port Deposit Intrusive Suite - Aberdeen Metagabbro and associated coarse amphibolite, Port Deposit Gneiss, Tom's Creek Trondhjemite, and dikes of amphibolite. Only the Port Deposit Gneiss phase intruded the Canal Road Formation.

The relationship of the Canal Road Formation to the James Run Formation is uncertain, because nowhere in the study area are they in direct contact. However, there is no difference in their respective geologic histories after deposition and prior to emplacement of the Port Deposit Intrusive Suite to indicate that the initial boundary between the two was structural. Instead, their dynamometamorphic histories are identical up until the D$_2$M$_2$ event discussed below. An initial depositional contact between the two is thus possible, most likely with the James Run Formation arc volcanics deposited on top of the Canal Road Formation. Southwick (1969) reports James Run Formation deposited on top of Wissahickon Formation south of the area considered by this study. If the Canal Road Formation consists of metamorphosed Late Proterozoic continental rift deposits related to the opening of Iapetus, and if the James Run Formation is deposited on top of it, then volcanic arc of the Chopawamsic Terrane may have been developed on a rift fragment of Laurentian crust.
After emplacement of the Port Deposit Intrusive Suite, all rocks of the Havre de Grace Block were subsequently metamorphosed to amphibolite facies and deformed during the Middle Ordovician (Taconic Orogeny) to Early Silurian, developing a steeply southeast-dipping foliation, which is layer parallel in the James Run and Conowingo Dam Formations. A strong dip slip lineation developed as well.

4.3. **Ordovician intrusions into mélange - The Conowingo Block**

Within the Conowingo Block, the Conowingo Dam Formation is deposited first, probably as a precursory mélange associated with the collision of the Chopawamsic Terrane with the Potomac Terrane. It is intruded during the Early to Middle Ordovician by the State Line Mafic Complex and Basin Run Tonalite, respectively. If the State Line Mafic Complex intrudes the Peters Creek Formation of the Westminster Terrane as proposed by Gates and others (1991), then emplacement of both the State Line Mafic Complex and Basin Run Tonalite postdates the collision between the Potomac and Westminster Terrane, and placing a lower age limit of Early Ordovician on this major tectonic boundary.

4.4. **Relationship between blocks prior to juxtaposition**

The Havre de Grace Block and Conowingo Block were probably related prior to structural juxtaposition. Olistoliths of metasedimentary and metavolcanic rocks in the Conowingo Dam Formation are low grade equivalents of the Canal Road Formation and James Run Formation, respectively. Thus, the Havre de Grace Block may have acted as a sediment source for the Conowingo Block during the deposition of the Conowingo Dam
Formation. The $S_1$ surface within olistoliths in the Conowingo Dam Formation could, then, be a more shallow crustal equivalent of $S_1$ in the Havre de Grace Block.

4.5. **Middle Ordovician to Early Silurian Thrusting - The Elbow Branch Thrust**

Juxtaposition of the Havre de Grace and Conowingo Blocks clearly postdates the metamorphism of the Havre de Grace Block, and is thus restricted to post-Middle Ordovician time. Several observations are consistent with juxtaposition of the two blocks across a major thrust fault during the Late Ordovician to Early Silurian. First of all, placement of high grade metamorphic rocks against previously unmetamorphosed rocks is consistent with faulting of the high grade block over the low grade block along a southeast-dipping thrust. However, a northwest-dipping normal fault could also place low grade hanging wall rocks against high grade footwall rocks. The lithofacies assemblage of the two blocks is more consistent with a thrust fault which transports the arc sequence of the Havre de Grace block northwest over the accretionary complex rocks of the Conowingo Dam Formation. Finally, the local dip-slip reverse indicators in the $S_1$ fabric and the strong $L_1$ dip-slip lineation within rocks of the Havre de Grace Block may reflect a thrusting component to the $D_1M_1$ (Taconic) dynamothermal event. As thrusting proceeded and the Havre de Grace Block moved to relatively shallow crustal levels, movement may have localized along a single surface, the Canal Road Formation - Basin Run Tonalite contact. Because of this contact's probable origin as a thrust fault, it is designated on the geologic map as the Elbow Branch Thrust. Conclusive kinematic evidence for its origin as a thrust fault is missing, because the contact was subsequently
reactivated during the Late Paleozoic as a dextral strike-slip splay off of the Rock Run Shear Zone, obliterating earlier structures.

4.6. **Late Paleozoic transpression - The Rock Run Shear Zone**

The Rock Run Shear Zone (D$_2$M$_2$) formed and was active from as early as the Middle Devonian through the Pennsylvanian. It began as a two kilometer wide zone of low strain SC mylonite in the Port Deposit Gneiss along and southeast of the Canal Road Formation and which evolved over time into an anastomosing network of thin high strain mylonite zones cutting the earlier S$_2$C fabric (Chapter 2). A major splay of the Rock Run Shear Zone formed along the Elbow Branch Thrust and within the Basin Run Tonalite to the north, obliterating evidence for earlier thrusting. An anastomosing network of discrete dextral strike-slip shear zones subsequently formed within the Basin Run Tonalite. Metamorphic conditions during D$_2$M$_2$ were retrograde relative to the peak amphibolite facies metamorphic assemblages of the Havre de Grace block, and reflect shearing under greenschist facies regional conditions. The deformation mechanisms in the shear zone are consistent with deformation under greenschist facies conditions as well (Chapter 2). Total offset across the Rock Run Shear Zone was a minimum of two to six kilometers, based on the degree of S$_2$C fabric development (Chapter 2). Considering the presence of anastomosing high strain zones cross-cutting the pervasive S$_2$C fabric, the existence of one major and numerous minor shear zone splays, and the fact the strain estimate in Chapter 2 does not incorporate offset across individual C bands, the total offset across the Rock Run Shear Zone and its splays may have been on the order of ten or more kilometers. Not knowing the magnitude of the offset limits the certainty with
which one can correlate earlier geologic features across the shear zone. The Rock Run Shear Zone fits well into the conceptual framework of Valentino and others (1994) in which the Central Appalachian Piedmont is modeled as an assemblage of structural blocks, each with its own internal history, bounded by dextral strike-slip shear zones (refer back to Figure 1.1).

The transpressional nature of the D2M2 event is indicated by the presence of a subvertical S2 foliation parallel to the S2 of the S2C fabric within blocks bounded by the Rock Run Shear Zone and its splayed. Within the Canal Road Formation, the S1 surface was tightly folded during D2, which S2 axial planar to these folds.

4.7. Late Paleozoic to Mesozoic dip-slip shearing

Dextral strike-slip shearing gradually evolved into oblique and then dip-slip shearing, concentrated along discrete, steeply dipping shear zones (D3). Thin subvertical dip-slip zones within the Basin Run Tonalite and the Port Deposit Gneiss moved in both reverse and normal senses, with the northwest side uplifted relative to the southeast. Layer boundaries within the James Run Formation were reactivated as normal faults. D3 was active from as early as Early Permian through as late as Middle Jurassic, and may reflect orogenic collapse associated with the Alleghanian Orogeny (Valentino, 1993), transtensional deformation (Gates, 1994), or Mesozoic rifting.
5. **Conclusions**
The crystalline bedrock of the lowermost Susquehanna River Gorge, Conowingo to Havre de Grace, Maryland, consists of two discrete structural blocks, each with its own unique history prior to juxtaposition. The southern, Havre de Grace, block originated as a Cambrian magmatic arc (James Run Formation and Port Deposit Intrusive Complex), possibly developed on a rift fragment of Laurentia (Canal Road Formation). The Havre de Grace Block belongs to the Chopawamsic Terrane of Horton and others (1989). The northern, Conowingo, block represents a precursory mélange (Conowingo Dam Formation) intruded by a layered mafic complex (State Line Mafic Complex) and associated plutons (Basin Run Tonalite). The Conowingo Block belongs most likely to the Potomac Terrane of Horton and others (1989). Many olistoliths within the Conowingo Dam Formation could have originated in the Havre de Grace Block. The Havre de Grace Block was metamorphosed and deformed under amphibolite facies conditions during Middle Ordovician to Early Silurian time, and subsequently thrust upon the Conowingo Block along the Elbow Branch Thrust, probably during the Late Ordovician or Early Silurian.

During the Middle to Late Paleozoic, beginning as early as the Devonian and extending through the Pennsylvanian, dextral strike-slip shearing took place under greenschist facies conditions along the Rock Run Shear Zone (Chapter 2), part of a shear zone system active across the entire Central Appalachian Piedmont (Valentino and
others, 1994). This shearing shuffled blocks with earlier geologic history an unknown magnitude, making correlation across strike difficult.

During Late Paleozoic to Mesozoic time, strike-slip shearing gave way to dip-slip shearing across very thin, low grade ductile shear zones. This shearing may have been related to Mesozoic rifting, orogenic collapse (Valentino, 1993), or transtensional deformation (Gates, 1995).

These conclusions best match the Taconic collision-Alleghanian transpression model (Valentino and others, 1994; Gates and others, 1991). No evidence was found for multiple terrane boundaries, as would be expected in the accretionary model of Horton and others (1989). The magmatic model (Sinha and others, 1989) is not directly addressed by this study, because it approaches the problem from an entirely different perspective - magmatic versus the dynamometamorphic approach used here. However, geochronology obtained from studies incorporating the magmatic model proved invaluable in restricting the timing of events.
Chapter Two

The Rock Run Shear Zone: a Late Paleozoic dextral strike-slip shear zone in the Central Appalachian Piedmont

1. Introduction

Recent studies in the Piedmont of the north central Appalachians have resulted in the discovery of a network of crustal scale, large offset, generally dextral strike-slip, anastomosing ductile shear zones of Late Paleozoic age (Valentino, Gates, and Glover, 1994). The major shear zones which comprise this system are shown Figure 2.1, a generalized geologic and structure map of the North Central Appalachian Piedmont. The Rock Run Shear Zone is the southeastern most of these zones. Valentino (1994) shows the Rock Run Shear Zone as a southwestern extension of the Rosemont Shear Zone (Valentino, 1988). This correlation is likely based on the projection of the two shear zones using existing geologic maps. However, this connection has not yet been mapped in detail, and is shown in gray in Figure 2.1

2. Methods

Methods described in Chapter One, Section 2 of this thesis were incorporated in the investigation of the Rock Run Shear Zone. Emphasis was placed on microstructural analysis of oriented thin sections and stereographic analysis of fabrics measured in the field. Microstructural analysis emphasized the determination of the shear sense, the metamorphic conditions under which shearing occurred, and, where possible, the magnitude of shear strain.
Figure 2.1 - Bedrock Geology and Major Structures, North Central Appalachia Piedmont

(after Valentino and others, 1994; Williams, 1978; and Southwick, 1968) Structural Abbreviations:
LV=Lancaster Valley Tectonite Zone; PG-HV=Pleasant Grove, Huntington Valley Shear System;
RZ=Rosemont Shear Zone; CCZ=Crum Creek Shear Zone (sinistral conjugate shear zone); BWA=Baltimore-Washington Anticlinorium; BGD= Baltimore Gneiss Domes

3. Data

Figure 2.2 shows the Rock Run Shear Zone and surrounding rock in the vicinity of the Susquehanna River in Cecil and Harford Counties, Maryland. Two map units - the Port Deposit Gneiss (well-foliated metatonalite) and the Canal Road Formation (amphibolite facies meta-turbidites and amphibolite) - are cross-cut by the Rock Run...
Shear Zone. A third, the Old Mill Tectonite, is restricted to the zone, protolith uncertain. In addition, three other rock types are found in the shear zone in decreasing quantities: vein quartz, amphibolite, and coarse-grained, highly altered serpentinite. All of these rock types yield a consistent, dextral strike-slip sense of shear based on outcrop and thin section petrography structural analyses. Each rock type and its associated microstructures is discussed below.

Splays off of the Rock Run Shear Zone cross-cut rocks to the northwest. A major spay is localized in the footwall of the Elbow Branch Thrust (Chapter One, Section 4.4.5).

3.1. The Port Deposit Gneiss

3.1.1. Sheared Tonalite

Sheared tonalite comprises the overwhelming volume of rock at outcrops within the Rock Run Shear Zone in the study area. The Port Deposit Gneiss is characterized by a moderate to high strain type 1 SC mylonitic foliation (Lister and Snoke, 1984; Berthe and others, 1979). Previous workers interpreted the distinctive texture as the result of either the intersection of two independent cleavages (Higgins, 1990; Lesser, 1982) or as cataclastic shearing (Southwick, 1968; Lesser, 1982). This paper presents the first detailed description of SC mylonite of the Port Deposit Gneiss.

Figure 2.3 shows the mylonitic foliation at outcrop. Point of view is looking down into the earth, approximately perpendicular the SC intersection. Figure 2.7 shows several polished slabs from the same perspective. Note the variation in fabric intensity.
Grain size reduction due to shearing is not obvious at this scale. Instead, highly deformed grains of quartz and plagioclase appear, with quartz much more deformed than plagioclase. Microscopically, these grains are composed of aggregates of much finer grained quartz and plagioclase (See Figure 2.4).

Several kinematic indicators other than the SC fabric itself are present. Folded quartz ribbons with dextral vergence, imbricated aggregates of plagioclase, and dextrally offset xenoliths and quartz-rich domains are common. In addition, dextral strike slip structures in vein quartz and lamprophyre dikes which cross-cut the granodiorite are described below.

Figure 2.5 (a & b) presents contour plots of poles to field measurements of S (schistosity) and C (cissaillement [French for “shear”]) planes. Both plots show a strong bullseye pattern, reflecting normal distributions of both C and S surfaces about their respective maxima. The C surfaces are more tightly grouped than the S surfaces. This is probably due to two factors. First, in SC mylonites the S surfaces form and are then variably rotated toward C surfaces with increasing strain, while C surface orientations remain relatively constant. Thus the spread of S may reflect varying degrees of strain. Secondly, S surfaces are inherently more difficult to measure at most outcrops in the Rock Run Shear Zone, where C surfaces are generally continuous and truncate S surfaces. In either case, the peak represents the average S plane orientation, striking N34E and dipping 83 degrees to the southeast. The average C plane strikes N64E and dips 87 degrees north. These mean planes are plotted in Figure 2.5.
Figure 2.2 - Geologic Map of Rock Run Shear Zone and surrounding rocks
Figure 2.3 - Annotated outcrop photograph of SC mylonite in Port Deposit Gneiss

The intensity of the SC foliation is heterogeneous across the shear zone, and is observed varies substantially even within single outcrops. This non-systematic variation renders strain estimation (as attempted below) difficult, especially if a relationship between intensity of SC foliation and weathering of outcrop. In other words, some of the higher strain zones may be missed because they don’t crop out.
Figure 2.6 illustrates the variation in SC fabric style across RRSZ along the Susquehanna River in Cecil County, MD. In addition to the overall variation, particularly note the within outcrop variation at Outcrop 406.4. Figure 2.7 shows essentially the same thing in polished slabs.

A and B above are plane and polarized light photomicrographs of typical microstructures within the Port Deposit Gneiss. Several features are worthy of note. Biotite aligned in S wraps into C. Quartz and plagioclase both occur as aggregates of submillimeter, polygonal grains. These appear as single, deformed grains at the outcrop scale, and represent grain size reduced igneous crystals. These aggregates exhibit very different behaviour. Quartz aggregates are highly elongate and occur as ductile ribbons flattened into S, sheared across C planes, and sometimes folded between adjacent C planes. Plagioclase aggregates are subequant, and are the competent phase facilitating the development of type 2 SC mylonite. The grain size reduction of plagioclase is interpreted to predate the development of the Rock Run Shear Zone, and is probably a remnant of an earlier amphibolite facies deformational event during which both quartz and plagioclase were ductile, and which is well preserved southeast of the Rock Run Shear Zone.

Figure 2.4 - Microstructures in Sheared Post Deposit Gneiss
Figure 2.5 - Orientations of fabric elements in the sheared Port Deposit Gneiss

a. S surfaces \( (n=162) \)

K=38.00  
Sigma=1.000  
Peak=76.75

b. C Planes \( (n=140) \)

K=100.00  
Sigma=1.000  
Peak=63.71

c. Average SC Foliation in Port Deposit Gneiss

d. Mineral stretching lineations (o) and orientation of amphibolite lenses (solid planes) on SC (dotted planes)
Comments

Dextrally sheared Port Deposit Gneiss - Canal Road Formation Conduct; note discrete offset of contact along C surfaces.

Typical fabric trace over much of zone; both continuous and discontinuous C bands are present.

Ultramylonite from a 1 meter wide high strain zone; note dominance of C bands, near total obliteration of S, and the presence of East striking C' bands offsetting C. Closed polygons define quartz ribbons.

Within Outcrop Fabric Variation

The fabric traces of three thin sections from samples at the same, high strain outcrop illustrate the local variation in fabric intensity which places severe limitations on the accuracy of strain estimates based on sample traverses. Sample 406.4,2 has the highest apparent strain fabric, with closely spaced c planes (1 to 5 mm) and a low SC angle. Sample 406.4,1 has c planes spaced at regular 5 to 6 mm angles, and an SC angle similar to 406.4,2. Sample 406.4,5 has irregularly spaced, anastomosing c planes 1 to 10mm apart and the highest SC angle observed at the outcrop. Despite having the apparently lowest strain fabric observed at the outcrop, 406.4,5 is the only sample exhibiting possible c' bands. These observations may indicate that both SC intersection angle and C planes spacing may depend on other factors in addition to strain. These factors probably include initial texture, preexisting foliations, and initial grain size.

SC fabric with both discontinuous and continuous C surfaces spaced at ~1cm. Note overall low intensity of fabric and corresponding coarse-grained appearance of rock. SC angle approaches 60°.

Figure 2.6 - Sketch showing variation in SC fabric development across Rock Run Shear Zone
Figure 2.7 - Polished slabs showing variation in SC fabric across the Rock Run Shear Zone

3.1.2. Anastomosing High Strain Zones

Thin anastomosing zones of extremely fine-grained mylonite up to a meter thick crosscut the SC fabric in the sheared Port Deposit Gneiss. They are vertical, and strike from N20E to N60E. Discrete strike-slip zones in the Basin Run Tonalite to the northwest exhibit similar orientations (Chapter 1, Section 3.2.3), and are probably late
splays off of the main shear zone. The sense of shear on all of these zones is dextral strike-slip. The boundaries of the zones appear sharp in the field. However, analysis of thin sections and cut slabs reveals the boundaries to be gradational over a very distance, changing from the typical SC fabric (Figure 2.6B) to ultramylonite over a few centimeters. Within these ultramylonite zones, original SC structures are nearly obliterated. Although a few cryptic S surfaces are observed locally in thin section, the overwhelmingly dominant structure is the C surface.

Kinematic indicators and well-developed mylonitic microstructures are common in the ultramylonite. Figure 2.8 presents annotated photomicrographs of these structures. Figure 2.8A shows the typical mylonitic fabric, with anastomosing C surfaces spaced at sub-millimeter intervals. Mica fish with tails forming C surfaces are common and yield a consistent dextral strike slip sense of shear. Figure 2.8B shows an imbricated rounded plagioclase porphyroclast separating two adjacent C bands. Such porphyroclasts are interpreted as remnant igneous phenocrysts. Their presence may indicate that grain size reduction of plagioclase may be the limiting factor in the reduction of C plane spacing during mylonitization. Lenses of highly recrystallized quartz up to a few mm thick perpendicular to C and up to a few cm long parallel to C are common. Careful inspection reveals these to be fold trains of ribbon quartz (Figure 2.8C) showing dextral strikeslip vergence. In addition to folding, quartz ribbons are also sheared across C planes (Figure 2.8D). Internally, these quartz ribbons have a highly sheared texture with
a strong grain shape preferred orientation indicative of type 2 SC mylonite (Lister and Snoke, 1984). The sense of offset based on these type 2 structures is dextral strike slip.

Figure 2.8 - Microstructures within anastomosing high strain zones in the Port Deposit Gneiss

3.1.3. **Sheared Vein Quartz**

Strongly sheared pre- to syndeformational vein quartz commonly occurs within the Port Deposit Gneiss as thin (~1-20 cm) layers oriented either parallel to C or at a low angle to C between the S and C surfaces. Those veins not oriented parallel to C exhibit discrete dextral offset across C surfaces and intrafolial folds with dextral vergence. Type
2 SC mylonite (Lister and Snoke, 1984) is variably developed within these veins. Figure 9 shows variable development of the Type 2 SC fabric in the sheared vein quartz.

In general, the apparent intensity of type 2 SC fabric development bears no relation to either appearance of the sheared vein quartz in the field or the intensity of the SC fabric in the surrounding granodiorite. For instance, Sample 406.4.4 comes from a zone of rather high strain within the granodiorite, but does not show a particularly strong grain shape preferred orientation. Sample S8B comes from an outcrop where the type 1 SC mylonite in the granodiorite is less intense, yet exhibits a stronger grain shape preferred orientation. These observations are consistent with the theory (Means, 1981; Hanmer and Passchier, 1991) that type 2 SC mylonites are both strain insensitive and cyclic, repeatedly experiencing periods dynamic recrystallization producing a grain size reduced, grain shape preferred orientation fabric followed by periods of recovery or grain growth resulting in a coarser fabric comprised of more equant grains. The fabrics in Figure 9 probably represent different stages in this cycle, and not different degrees of bulk strain.
An exception to this is Sample 507B1 (Figures 9 and 10), which exhibits the most well-developed type 2 structures and which has the most highly sheared appearance at outcrop, including a subhorizontal mineral stretching lineation defined by quartz rods exposed along C surfaces. Such lineations are generally absent. This sample was collected within one of the anastomosing high strain zones which cross-cut the distributed SC fabric. The finer grain size and stronger fabric suggest that in these late zones, the rate of deformation exceeded the rate of recrystallization, resulting in the strongly sheared appearance. This could be due to lower temperature conditions of deformation, higher strain rates, or a combination of the two (Means, 1981).

Measurements of crystallographic preferred orientations (CPO) within these veins were not performed as part of this project. However, preliminary measurements of quartz c-axes using the universal stage microscope performed on samples collected in 1991 (Valentino and others) at the Port Deposit Quarry yielded both dextral strike-slip and subvertical

Figure 2.10 - Sheared vein quartz from an anastomosing high strain zone
Sample 506H1
Weakly developed type 2 SC mylonite in vein quartz. Grain size is millimeter scale with serrate grain boundaries. X:Z ratios typically 3:1. Dominant preferred orientation is long grain axis parallel to C with second mode approximately parallel to S.

Sample 406.4.4
Weakly to moderately developed type 2 SC mylonite in vein quartz. Grain millimeter to 1/10 millimeter scale. Serrate grain boundaries. X:Z ratios up to 4:1. Long axis of grains preferentially oriented in C with a secondary mode preferentially in S.

Figure 2.9 - Type 2 SC fabrics in sheared quartz veins in the Port Deposit Gneiss

Sample 406.4.4
Well developed type 2 SC mylonite in vein quartz. Grains millimeter to 1/10 millimeter scale. Serrate grain boundaries. X:Z ratios up to 4:1. Long axis of grains preferentially oriented in S. C bands spaced at intervals of a few mm; none obvious in this section.

Sample 507B1
Well developed type 2 SC mylonite in vein quartz. Grains millimeter to 1/100 mm scale, typically 1/10 mm scale. Serrate grain boundaries, strong undulose extinction. X:Z ratios typically 4:1. Long axis of grains preferentially oriented in S. S domains separated by thin C planes of very fine-grained mica and quartz.

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simple shear CPO fabrics (Eric Gardener, 1993, personal communication). The latter may be related to D₃ dip-slip shearing which affected rocks both within and outside of the RRSZ (see Chapter 1, Section 3.2.4).

3.1.4. Sheared Amphibolite

Thin lenses of highly sheared amphibolite typically less than 1 meter thick and up to tens of meters in length crop out in the Port Deposit Gneiss near the southeast margin of the Rock Run Shear Zone in Cecil County, and are generally oriented parallel or subparallel to C. They most likely represent the sheared equivalent of the ubiquitous lamprophyre dikes which intrude the Port Deposit Gneiss, Tom’s Creek Granodiorite, and James Run Formation southeast of the shear zone. These rocks experienced both grain size reduction and retrograde metamorphism during shearing within RRSZ. Type 1 SC structures are well developed (Figure 11A), and the alignment of the long axes of hornblende grains define a subhorizontal lineation along exposed C surfaces. Rounded, relatively equant hornblende porphyroclasts (Figure 11B) represent relict igneous hornblende phenocrysts, and are surrounded by finer-grained hornblende needles, biotite, chlorite, and colorless amphibole (tremolite-actinolite?) oriented parallel to S or C within the corresponding domain. In addition, thin discrete felsic ribbons comprised of a fine-grained polygonal mosaic of plagioclase and quartz are present.
3.1.5. *Sheared Ultramafic*

A lens of serpentinized ultramafic rock crops out along the contact between the Port Deposit Gneiss and the Canal Road Formation in a small quarry at the entrance to Susquehanna State Park Campground in Harford County, and was the only such outcrop documented in the present investigation. A well-developed, dextral SC structure is present on both outcrop and thin section scales. Also present is a distinctive system of thin anastomosing dextral shear zones dividing the serpentinite into relatively undeformed lenticular pods, similar to that described by Gates (1991).
3.2. Canal Road Formation

The contact between the Port Deposit Gneiss and the lower amphibolite facies interlayered metaturbidites of the Canal Road Formation (see Chapter 1, Sections 3.1.1 and 3.2) occurs within the Rock Run Shear Zone. The contact is probably not a simple fault contact, but rather an original igneous intrusive contact highly modified by later shearing (see Chapter 1, Section 3.1). The southeast few hundred meters of the Canal Road Formation (see Figure 2 or Plate 1) are intensely sheared and retrograded within the Rock Run Shear Zone. The degree of shearing of Canal Road Formation northeast of the river is uncertain due to poor exposure.

Abundant kinematic indicators, including Type 2 SC fabric, sigma tails on retrograded garnet porphyroclasts, chlorite and muscovite mica fish (Figure 12), shear bands, and crenulation vergence, all indicate a consistent dextral strike-slip sense of shear (Figure 13, all as described in Hanmer and Passchier, 1991). Thin amphibolite layers are common in the Canal Road Formation, and exhibit microstructures (Figure 13.7) similar to those observed in the sheared amphibolite of the Port Deposit Gneiss, and also suggest dextral strike slip movement in the shear zone.
The orientations of various structural elements of the Canal Road Formation within the Rock Run Shear Zone are shown in Figure 2.14. At the outcrop level, a single pervasive schistosity is measurable. This schistosity generally corresponds to the S component of the SC mylonite of the Port Deposit Gneiss, which also corresponds to the S2 axial planar cleavage within the Canal Road Formation northwest of the Rock Run Shear Zone (see Chapter 1, Section 3.2.3). At higher strain outcrops, particularly those in Harford County near the contact with the Port Deposit Gneiss, the schistosity measured in the field corresponds to C bands and, locally, C' extensional crenulation bands or shear bands (Simpson, 1984). Crenulations are generally very steeply plunging, and are oriented approximately the same as the few measurable fold axes measured within the shear zone. Their orientation corresponds to the lineation defined by the intersection of the S and C surfaces within the Port Deposit Gneiss.
Figure 2.14. Fabric element orientations in sheared Canai Road Formation
3.3. Old Mill Tectonite

Four kilometers east-northeast of the Susquehanna along the strike of the shear zone crops out the Old Mill Tectonite, a mylonite very different in appearance from those observed along the Susquehanna River. This rock is a quartzo-feldspathic blastomylonite comprised of 5-50% mm scale porphyroclasts in a very fine-grained (.01mm <d<.1mm) matrix of polygonal quartz and plagioclase. Two opaque phases, probably magnetite and ilmenite, are common in the matrix, the latter frequently overgrown by radial sphene aggregates forming a composite porphyroclast phase. Biotite is the dominant phyllosilicate, lying along closely spaced (<1mm) east by northeast striking, subvertical C planes (Figure 2.15), and comprises up to 15% of the rock. The porphyroclast population is 90% well-rounded plagioclase, up to 10% sphene-ilmenite composite porphyroclasts, and sparse, subhedral apatite porphyroclasts. Thin inclusions (relict xenoliths?) of pelitic material are common, are highly flattened with length to width ratios of up to 10:1, and lie along the C surfaces. Locally, small stringers (up to a 10 cm long) of highly sheared granodiorite are present as elongate lenses aligned with the C planes. Thin plagioclase bearing quartz veins are common throughout the zone. They cross-cut the mylonitic foliation, and are themselves in turn sheared. Elongate trails of plagioclase porphyroclasts can frequently be traced back to pods of coarse-grained quartz and plagioclase. Most of the porphyroclasts within the Old Mill Mylonite occur in planar trails oriented at a low angles to C, striking between N40E and N65E and dipping subvertically (See Figure 2.16A and D; See also Figure 2.15B and C). These trails are
interpreted as representing sheared syndeformational dikes oriented in the extensional strain field of a dextral strike slip mylonite. Independent evidence of the extension of these trails is the presence of porphyroclast and matrix phases between rounded plagioclase porphyroclasts which do not occur within the quartz-plagioclase dikes. Only a single vein was observed oriented within the compressional field (Figures 2.16 B,D, & G), and it clearly cross-cuts the extended dikes. This vein exhibited fold vergence sympathetic to dextral strike slip motion. Other mesoscopic kinematic indicators include well-rounded, ellipsoidal pods of vein quartz imbricated counterclockwise 30° to 40° from C(Figure 2.16E), and pressure shadows on scattered cm scale quartz-plagioclase composite porphyroclasts (Figure 2.16F). These indicators yield consistent evidence of dextral, strike-slip motion along the shear zone.

Multiple kinematic indicators are observed in thin section, including sigma tails of biotite on composite ilmenite-sphene porphyroclasts, antithetic feldspar domino structures (Figure 2.17B), sigma and delta tails on rounded plagioclase pclasts (Figure 2.17C), SC fabrics in areas with a high porphyroclast density (Figure 2.17D). These structures are consistently dextral strike slip in horizontally oriented thin sections.

The Old Mill Tectonite appears to have a well preserved, polydeformational history. While dextral strike slip kinematic indicators abound on horizontal surfaces in both thin section and outcrop, the dominant lineation measured is dip slip, raking steeply northeast along C planes (Figure 2.15C). Furthermore, the long axis of xenoliths are oriented down-dip, parallel to the lineation. While this lineation
may locally be an SC intersection lineation consistent with dextral strike-slip motion, in most places it is clearly a mineral stretching lineation suggesting dip slip movement along C planes.

![Image of diagrams]

**Figure 2.15. Orientation of fabric elements in the Old Mill Tectonite**

Investigation of lineation parallel, foliation perpendicular thin sections reveals the presence of dip slip microstructural kinematic indicators. Though not as well developed
as the strike-slip structures, these indicators reveal a late history of dip slip movement in which the northwest side of the zone was uplifted relative to the southeast side. Similar, subvertical dip-slip shear common throughout the study area (See Chapter 1, Section 3.2.4). These structures cross-cut and thus post-date the dextral strike-slip movement along the Rock Run Shear Zone.

As stated above, the long axes of the highly flattened pelitic inclusions are invariably oriented subparallel to the mineral stretching lineation. This dip-slip elongation coincides with that observed in xenoliths within the Port Deposit Gneiss southeast of the Rock Run Shear Zone, at outcrops clearly unaffected by either dextral strike slip shearing or the later dip slip shearing (see Chapter 1, Section 3.2.2). If this is so, then the shapes of the pelitic inclusions are not reliable strain markers within the Old Mill Tectonite, as they are the result of multiple deformations.

The protolith of the Old Mill Tectonite is elusive due to its intensely mylonitic nature. The similarity of the pelitic inclusions to xenoliths in the Port Deposit Gneiss suggests an igneous protolith. In addition, the bulk composition is granodioritic. Finally, the Old Mill Tectonite is bounded to the south by the Tom’s Creek Trondhjemite (Chapter 1, Section 3.1.9), and small stringers of granodiorite occur in the mylonite near the contact. These observations suggest that the Old Mill Tectonite may be highly sheared Tom’s Creek Trondhjemite or Port Deposit Gneiss. The bulk composition is also similar to that of metadacite within the James Run Formation, a protolith more consistent with the Old Mill Tectonite’s fine grain size.
A. Porphyroclast trails representing sheared syndetformational dikes in the extensional strain field. Scale is parallel to mylonitic foliation (C). Dikes rotating dextrally to C.

B. Evidence for syndetformational origin of pegmatite dikes. Pod of pegmatite in bottom of frame cross-cuts C, but itself in turn sheared along its margins. Area in box blown up in D below.

C. Along strike view of mylonitic foliation (C) cross-cut by and wrapping around a syndetformational pegmatite pod

D. Rare, thin folded dike in compressional field showing dextral vergence. Note that it cross-cuts the extended dikes (porphyroclast trails). Flattened xenoliths lie in C. Annotated sketch below.

E. Rounded felsic pod imbricated at low angle to C and flattened xenolith lying in C (pod is 15cm long)

F. Slightly assymmetric sigma tail on large plasst of plagioclase

G. Major features of D above:
- extended syndetformational dike
- xenolith
- folded dike or vein

Figure 2.16 - Field photographs showing mesoscopic structures in the Old Mill Mylonite
4. Interpretations

4.1. Conditions of deformation

Movement along the Rock Run Shear Zone occurred under greenschist facies metamorphic conditions. Evidence for this includes metamorphic mineral assemblages in the shear zone, inferred deformation mechanisms based on fabric style, and plagioclase compositions.
The amphibolite facies S1 foliation and M1 metamorphic assemblage within the Canal Road Formation are almost completely obliterated in the shear zone. The peak garnet + biotite + muscovite ± staurolite assemblage (Chapter 1, Section 3.2.2) becomes becomes biotite + muscovite + chlorite within the shear zone. The only vestiges of the M1 assemblage preserved are garnet fragments within clusters of biotite and chlorite. The biotite-muscovite-chlorite assemblage in the Port Deposit Gneiss also suggests shearing under middle greenschist facies conditions. Sheared amphibolites exhibit fibrous amphiboles characteristic of reequilibration under low grade condition.

The marked ductility contrast between weak quartz and much stronger plagioclase aggregates necessary for the development of SC fabric (Berthe and others, 1979) in the Port Deposit Gneiss is indicative of temperatures of deformation between 250° and 450° C (Tullis and Yund, 1985, 1987).

Within sheared Port Deposit Gneiss, the plagioclase composition is around An₄₅, while south of the shear zone plagioclase compositions range from An₂₅ to An₃₇. This suggests reequilibrations of plagioclase in the shear zone to lower-grade metamorphic conditions.

4.2. **The role of shearing in retrogradation**

All rocks caught up in the Rock Run Shear Zone exhibit lower grade metamorphic assemblages in the shear zone than elsewhere. This variation in assemblage from outside to inside the shear zone takes place over very short distances.
This implies that rocks both within and outside of the shear zone were subject to virtually identical pressures and temperatures. Therefore, some other factor related to the shear zone must have facilitated retrogradation.

Strain induced destabilization of the peak assemblage and the presence and nature of shear zone fluids are possible causes. Independent evidence of the former was not documented. However, there is abundant evidence of shear zone fluids. The presence of abundant, thin, syndeformational plagioclase bearing quartz veins in both sheared amphibolite layers of the Canal Road Formation and in the Old Mill Tectonite along strike to the northeast suggest an active, hydrous fluid component during shearing. In addition, many of the sheared quartz veins within the Port Deposit Gneiss may have formed during shearing.

The preexisting M1 assemblage indicates amphibolite facies peak metamorphic conditions outside the shear zone (see Chapter 1, Section 3.2.2). Retrogradation to greenschist facies involves many hydration reactions, which require water to proceed (Spear, 1993). These observations suggest that it was not the shearing itself which facilitated retrogradation, but rather the fluids associated with the shear zone.
4.3. Strain estimate

4.3.1. SC Mylonite in Port Deposit Gneiss

A simple estimate of the minimum offset across the Rock Run Shear Zone can be made using these average SC fabric orientations and Ramsay and Graham's (1970) equations:

$$\gamma = 2 \left( \tan 2\theta' \right)^{-1}$$

$$\text{displacement (S)} = \int_0^x \gamma \, dx$$

where $\gamma$ = shear strain, $\theta'$ = angle between S and C planes, and $x$ = width of shear zone. Simpson (1989) points out four characteristics of this approach which cause it to yield a significant underestimate of actual offset: 1) it ignores the curvature of S planes near C planes where most of the strain concentrates, 2) it assumes no net slip along C planes, 3) at high shear strains $\theta'$ is very small, and small errors in measurement can lead to large differences in $\gamma$ estimates, and 4) strain is generally heterogeneously distributed in large scale shear zones. Factors 1, 2, and 4 are the most important limitations which regards to the Port Deposit Gneiss, where $\alpha$ averages $30^\circ$. Another, more fundamental problem with this technique is discussed below.

The modified Ramsay and Graham (1970) approach (MRG) assumes S planes initiate at $\alpha=45^\circ-\theta'$ (the angle between c planes and the long axis of the strain ellipsoid after infinitesimal shear strain $\, d\gamma$). In the low strain portions of the Port Deposit Gneiss, however, measured $\alpha$ commonly exceeds $45^\circ$. The maximum observed value of $\alpha$ is $60^\circ$, and occurs in samples where the fabric intensity is qualitatively observed to be the
weakest. This suggests that S planes initiate at 60° rather than 45°. Thus α cannot be treated like θ' for the purposes of strain estimation.

Simpson (1989) suggests that type 1 SC mylonites are analogous to the ideal Ramsay and Graham's (1970) shear zones. Using a combination of theoretical derivation and real examples, Ramsay and Graham (1970) show that the angle between unrotated schistosity and shear planes should be approximately 45°. S surfaces form parallel to planes of maximum and intermediate stretch, and thus initiate perpendicular to σ1. The shear zones studied by Ramsay and Graham (1970) formed in metagabbro under dry, amphibolite facies conditions, and exhibit a maximum SC angle of 45°. This value agrees well with theoretical predictions for ductile shearing of a homogeneous crystalline solid (Ramsay and Graham, 1970). It does not necessarily apply to Type 1 SC mylonites.

Type 2 SC mylonites are not purely ductile structures, but composite ductile-brittle ones. It is this ductility contrast which causes them to form in the first place (Lister and Snek, 1984; Berthe and others, 1979). In addition, fluids seem to play a significant role in the development of a Type 1 SC fabric, at least in the Port Deposit Gneiss.

Consider instead the scenario shown in Figure 18. Suppose the S and C planes are forming independently in response to the regional stress field illustrated. Assuming the S planes form perpendicular to σ1, one can work backwards and calculate the angle between σ1 and C based on the maximum angle α, observed between S and C, which is
taken to represent the lowest simple shear strain fabric. For the Port Deposit, is $\sim 60^\circ$, are thus C is oriented at an angle of approximately $30^\circ$ to $\sigma_v$.

![Diagram](image)

**Strain Marker Geometry Evolution with Progressive Simple Shear**

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<th>$\gamma$</th>
<th>$\beta$</th>
<th>$\alpha$</th>
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<td>0.00</td>
<td>150$^\circ$</td>
<td>60$^\circ$</td>
</tr>
<tr>
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<tr>
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<td>80</td>
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<td>9$^\circ$</td>
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- Conjugate shear zone (rarely developed at map scale)

- Preferred shear planes, paralleling orogen and preexisting structures

- Vicinity of Rock Run Shear Zone

**Figure 2.18 - Evolution of strain markers in a ductile shear zone during progressive simple shear**
Figure 19 shows the derivation of the strain equation for SC mylonites using initial SC intersection angle $\alpha_0 = 60^\circ$, which is

$$\gamma = 1/\tan \alpha - 1/\tan 60^\circ.$$ 

Thus, the expression for displacement becomes

$$\text{displacement (S)} = \int_0^x (1/\tan \alpha - 1/\tan 60^\circ) \, dx.$$

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<tr>
<th>Conditions (values sketched)</th>
<th>Geometry</th>
<th>Strain Equation</th>
</tr>
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<td>s plane</td>
<td>$\gamma = 0$</td>
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<tr>
<td>$\gamma = 0$</td>
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<td>$\theta' = 45^\circ$</td>
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<td></td>
</tr>
<tr>
<td>$0 &lt; \gamma &lt; 1$</td>
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</tr>
<tr>
<td>$45^\circ &gt; \theta &gt; 32^\circ$</td>
<td></td>
<td>$\gamma = x/y = (x-x_0)/y$</td>
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<tr>
<td>$60^\circ &gt; \alpha &gt; 32^\circ$</td>
<td></td>
<td>$= (y/\tan(\alpha) - y/\tan(\alpha_0))/y$</td>
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<td>($\psi = 15^\circ, \gamma = .27$, $\theta' = 41^\circ, \alpha = 50^\circ$)</td>
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<td>$\theta = \alpha$</td>
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<td>($\psi = 60, \gamma = 1.73$, $\theta' = 25^\circ, \alpha = 23^\circ$)</td>
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Figure 2.19 - Derivation of Strain Equation for SC mylonites
Figure 20 shows the differences in strain estimated by this approach and by the modified Ramsay and Graham (1970) approach as a function of SC angle $\alpha$. Note that this new approach retains the integration of strain over the entire shear zone component of Ramsay and Graham (1970), but recognizes the difference between $\alpha_o$ and $\theta'$. Three features from this graph are very important. First, for $60^\circ > \alpha > 45^\circ$ negative strain is predicted by MRG, a prediction totally incompatible with discrete positive offset observed across C planes in low strain SC mylonite where $\alpha > 45^\circ$. Secondly, for $45^\circ > \alpha > 30^\circ$, MRG predicts significantly lower strains than the method presented in this paper, the error increasing with increasing $\alpha$. Finally, MRG slightly overestimates $\gamma$ for $30^\circ > \alpha > 10^\circ$. For $\alpha < 10^\circ$, the difference in strain estimation between the two models is insignificant. The two models predict equivalent $\gamma$ at $30^\circ$.

![Strain Estimate Comparison](image)

**Figure 2.20 - Theoretical comparison of SC strain estimates used in this paper with the method of Ramsay and Graham (1970)**
Two southeast to northwest transects across the Port Deposit Gneiss, T1 along the Susquehanna River and Rock Run in Harford County and T2 along the Susquehanna River and Rock Run in Cecil County, were performed to take field measurements of SC intersection orientations for the purpose of strain estimation. Figure 21 exhibits the results for the two transects. Note the areas of negative strain predicted by the use of Ramsay and Graham (1970). For the Harford County transect, the new approach yields an offset estimate of 1818 meters compared to 1455 meter using the Ramsay and Graham (1970) method. The difference for the Cecil County transect is even larger, 2048 meters using the new approach versus 970 meters using MRG.

The revised technique developed here eliminates the problem of negative offset in low strain regions, yields a greater minimum offset across a type 1 SC shear zone, and appears to produce more consistent results. However, all of the limitations of the Ramsay and Graham approach noted by Simpson (1984) still apply.

4.3.2. Sheared syntectonic veins in Old Mill Tectonite

An independent minimum offset estimate for the Rock Run Shear Zone based on the orientation of sheared syntectonic veins within the Old Mill Tectonite can be attempted based on the equation derived in Figure 2.23. This assumes that the dikes orientation during intrusion was approximately 150° counterclockwise from C, as shown in Figure 2.18. Using only the width of exposure of the Old Mill Tectonite (~1.2 kilometers), assuming approximately homogeneous strain across the exposed transect, and using an average $\beta$ of 18° (from Figure 2.15), this technique yields a minimum offset
estimate of about six kilometers across the southeastern portion of the Rock Run Shear Zone.

**Comparison of New Method and MRG**
Transect 1, Harford County

*Offset = Area under curve*

- New method = 1818 meters
- MRG = 1455 meters

**Comparison of New Method and MRG**
Transect 2, Cecil County

*Offset = Area under curve*

- New method = 2048 meters
- MRG = 970 meters

*distance from SE margin of RRSZ(m)*

**Figure 2.21 - Comparison of the technique developed in this paper with that of Ramsay and Graham (1970)**
4.3.3. **Geochronology**

Very little geochronology exists for the Rock Run Shear Zone. Analyses by Lesser (1982) suggest the zone was active from as early as the middle Devonian through the Early Pennsylvanian. Two Rb-Sr analyses of biotite separates of Port Deposit Gneiss with the distributed SC fabric yielded ages of 380 Ma and 316 Ma, respectively (Lesser, 1982). Whole rock Rb-Sr on a sample of highly sheared Port Deposit Gneiss yielded an
age of $360 \pm 65$ Ma (Lesser, 1982). Biotite Rb-Sr by Kohn and others (1993) yielded an anomalous age of $180$ Ma. This sample was probably taken near one of the numerous $D_7$ shear zones (Chapter 1, Section 3.2.4) which cross-cut the shear zone fabric.

Based on cross-cutting relationships, Valentino and others (1994) interpret the entire Central Appalachian Piedmont transpressional system as post-Taconic, most likely associated with the Alleghanian Orogeny (345-245 Ma, Glover (1989)).

5. Conclusions
The Rock Run Shear Zone is a component of the dextral transpressional shear zone system described in Valentino and others (1994). Like most of the zones in the system, it cross-cuts and retrogrades pre-existing (Taconic) structures and metamorphic assemblages. The limited geochronology supports this conclusion.

Like many of these zones, offset determination is hampered by the lack of a map scale offset marker. However, offset across the zone was a minimum of two to six kilometers, based on the orientation of fabric elements - SC intersection angles and the rotation of syndeformational veins. Considering both the minimum nature of the strain offset calculations used in this study and the presence of an anastomosing network of high strain zones which cross-cut the pervasive SC fabric, the offset could have been many times the minimum estimate, perhaps a few tens of kilometers.

The proposed correlation of the Rock Run Shear Zone with the Rosemont Zone as drawn in Valentino and others (1994) (see Figure 2.1) is reasonable based on offset considerations. Valentino (1988) estimated a minimum offset of 28 kilometers across
the Rosemont Zone based on the correlation of a block of mafic gneiss within the shear zone to mafic gneiss of the Westchester basement massif which the shear zone intersects along strike. This order of magnitude of offset is not unreasonable for the Rock Run Shear Zone.

The moderate to high offset, Acadian (?) through Alleghanian dextral strike-slip shear zone system of the Central Appalachian Piedmont requires reconstruction in order for meaningful models of earlier tectonic events to be developed. Not only do blocks bound by the dextral strike-slip shear zones need to be restored to their pre-transpressional positions, but also the deformation internal to these blocks caused by transpression needs to be recognized and undone. For example, the Tucquan antiform, lying between the Lancaster Valley and Pleasant Grove - Huntingdon Valley Shear Zones has recently been shown to be a transpressional feature (Valentino, 1993).

Future studies in the Central Appalachian Piedmont are needed to further delineate the extent of Late Paleozoic shearing. Such studies should integrate geochronology and quantitative metamorphic petrology with structural petrology and field relations in order to tie down the timing, thermal conditions, and offset associated with Late Paleozoic transpressional deformation in the Central Appalachian Piedmont.
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Appendix A - Petrographic Tables
(modes based on visual estimation)

1. Canal Road Formation

   a. graded psammites (metaturbidites)

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<th>Afs</th>
<th>Lithic Fragments</th>
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   b. pelitic and semipelitic schist

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   c. amphibolite

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4. James Run Formation

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108
5. Old Mill Tectonite

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* - generally radial overgrowths on opaque (ilmenite?) inclusions

6. Aberdeen Metagabbro and Coarse Amphibolite
(only coarse amphibolite was analyzed in this investigation)

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109
7. Port Deposit Gneiss

| Sample  | PI  | Qtz | Afs | Bt | Hbl | Cz | Ms  | Aln | Tm  | Mg1 | Ilm | Opa | %An | Qz  | Mc | Pl  | Zrn | Ap  | Uk  | Ep | Ms | Chl | Bt | Grt | Location |
|---------|-----|-----|-----|----|-----|----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| 301.01  | 30  | 30  | 10  | 10 |     |    |     |     |     |     |     |     |     |     |     |     |    |    |     |     |     |     | D4-1 |
| 205.02.1| 40  | 40  | 1   | tr | 7   | tr |     | tr  |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D4-2 |
| 3152A   | 70  | 10  | 1   |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D4-3 |
| 3152B   | 40  | 40  | 5   |    | 7   |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D4-3 |
| 327.10.1| 25  | 60  |     |    | 5   |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D5-1 |
| 340.3.2 | 40  | 40  | 5   |    | tr  | 5   |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D5-1 |
| 341.11A | 37  | 40  | 4   |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | E4-1 |
| 341.11B | 37  | 40  | 4   |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | E4-1 |
| 405.8.1 | 36  | 36  | 8   |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D4-4 |
| 406.4.1 | 33  | 40  | 2   | 5  | 8   | tr |     | tr  |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C4-1 |
| 406.4.2 | 25  | 56  | 1   |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C4-1 |
| 406.7.2 | 39  | 40  | 1   | 5  | 5   | tr |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C4-2 |
| 407.7.3 | 35  | 35  |     |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C4-2 |
| 504F1   | 45  | 45  | 1   | 5  | 4   |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D3-2 |
| 506C1   | 35  | 35  | 4   | 5  | 6   |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C4-2 |
| 506F2   | 40  | 40  | 7   |    | tr  | 7   |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D3-3 |

Contact Brecia

| Sample  | PI  | Qtz | Afs | Bt | Hbl | Cz | Ms  | Aln | Tm  | Mg1 | Ilm | Opa | %An | Qz  | Mc | Pl  | Zrn | Ap  | Uk  | Ep | Ms | Chl | Bt | Grt | Location |
|---------|-----|-----|-----|----|-----|----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| 333.1.1 | 25  | 50  | 10  | 5  |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | E7-1 |
| 332.1.1 | 10  | 40  | 10  | 25 |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | E7-1 |

Xenoliths

| Sample  | PI  | Qtz | Afs | Bt | Hbl | Cz | Ms  | Aln | Tm  | Mg1 | Ilm | Opa | %An | Qz  | Mc | Pl  | Zrn | Ap  | Uk  | Ep | Ms | Chl | Bt | Grt | Location |
|---------|-----|-----|-----|----|-----|----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| 315.30  | 40  | 40  | 8   |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D4-3 |
| 323.07.1| 13  | 86  | 3   | 1  | 1   |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D4-3 |

8. Tom's Creek Trondhjemite

| Sample  | PI  | Qtz | Afs | Bt | Hbl | Cz | Ms  | Aln | Tm  | Mg1 | Ilm | Opa | %An | Qz  | Mc | Pl  | Zrn | Ap  | Uk  | Ep | Ms | Chl | Bt | Grt | Location |
|---------|-----|-----|-----|----|-----|----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| 317.14.1| 50  | 30  | 0   | 2  |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | E5-1 |
| 339.17.1| 50  | 34  | 1   | 3  |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C7-2 |
| 341.4.1 | 50  | 35  | 5   | 5  |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C7-2 |
| 341.5.4 | 55  | 30  | 5   |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C7-3 |
| 341.7.1 | 40  | 40  | 1   | 2  | 13  |     |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C7-3 |
| 341.9.1 | 60  | 30  | 1   |    |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | C7-4 |

Xenoliths

| Sample  | PI  | Qtz | Afs | Bt | Hbl | Cz | Ms  | Aln | Tm  | Mg1 | Ilm | Opa | %An | Qz  | Mc | Pl  | Zrn | Ap  | Uk  | Ep | Ms | Chl | Bt | Grt | Location |
|---------|-----|-----|-----|----|-----|----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| 339.15.2| 40  | 40  | 0   | 8  | 8   |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D7-1 |
| 339.17.2| 30  | 35  | 25  | 5  |     |    |     |     |     |     |     |     |     |     |     |     |    |    |    |    |    |    | D7-1 |
9. Amphibolite dikes

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10. Conowingo Dam Formation

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11. Basin Run Tonalite

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Sheared Samples

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Xenoliths

| Sample |                |                |                |          |

12. Quartz diorite and quartz gabbro of State Line Mafic Complex

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CD1

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111
13. Sheared quartz veins in Rock Run Shear Zone

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Appendix B - Index to Locations of Photos, Photomicrographs, and Sketches

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Appendix C. Locations of samples and photographs
Vita

I was born on January 9, 1967, just after one of the worst snowstorms in local history hit Strasburg, Virginia, my home town. My father had arranged for a helicopter to fly my mother to the hospital in the event that the roads were impassable. We ended up driving. I was raised on a two hundred odd acre farm in the Shenandoah Valley by varying configurations of my father, William H. Orndorff, my mother, Kay S. Fiocca, and my maternal grandmother, Lucy Spiker, in a house that had served as a hospital during the Civil War. As kids, my brother and I would find lead bullets with tooth marks on them in the cellar. Across the road from the farm was a large, abandoned quarry in Cambro-Ordovician limestone and dolostone. I spent many hours exploring the tailings piles for calcite samples. Beds of brick red clay cropped out on a hillside on the farm. Rain would wash the clay away and leave behind clear quartz crystals several inches long. I liked school, and graduated valedictorian of the Strasburg High School Class of 85. I matriculated to Johns Hopkins University, where I was graduated in 1990 with a bachelor of arts degree in Natural Science, and a newfound interest in geology. While I was at Hopkins, I became an avid cave explorer. This led me to the Department of Earth and Planetary Sciences at Hopkins near the end of my undergraduate years, where I got hooked on geology. A weird combination of cave exploration, choral music, science writing, and geology brought my future wife, Zenah Viola Wilson, and I together shortly thereafter. After graduation, I worked for a year and a half at the Smithsonian Environmental Research Center in Edgewater, Maryland, studying what happens to nitrates in wetlands near cornfields. In the evenings, I took classes in geology at the University of Maryland so I could qualify for graduate school. In 1992, Virginia Tech accepted me into the Geology Department and my fiancé into the Department of Crop and Soil Environmental Science. After a summer at field camp in South Dakota, we got married and drove to Blacksburg. Many people thought we would never leave, including my advisors. Surprise. Two beautiful daughters, twenty-odd pounds, and almost four-years later, the masters degree is completed, and I am still alive. The Klingon creed - “What does not kill me will only make me stronger!” - is appropriate. I hope to apply the strength I have gained from my time at Virginia Tech to all future endeavors, public and private, which I may choose to undertake. Peace.

[Signature]

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