THE CREAM VALLEY SHEAR ZONE:

A PROPOSED TERRANE BOUNDARY IN

SOUTHEASTERN PENNSYLVANIA

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MASTER OF ARTS

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by

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ABSTRACT

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Master of Arts
Temple University, 1998
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The Cream Valley Shear Zone, in southeastern Pennsylvania, contains a proposed terrane boundary between Laurentia (north of the shear zone) and a displaced terrane (south of the shear zone). The Cream Valley shear zone is part of a Paleozoic dextral strike-slip shear zone system which extends from New Jersey through southeastern Pennsylvania and into Maryland. Within the field area, in southeastern Pennsylvania, the Cream Valley Shear Zone varies in width from 3 to 5 km and is characterized by a strong, penetrative, steeply dipping foliation with an overall strike of N70°E. The proposed terrane boundary lies between retrograde greenschist-facies phyllonite and prograde greenschist-facies phyllite near the southern edge of the shear zone.

The terrane boundary divides Laurentian shelf sequence lithologies, including Cambrian Chickies quartzite and Cambro-Ordovician carbonates, from high-grade metamorphic rocks of the southern terrane. Rocks of the Laurentian cover sequence have experienced peak metamorphism of greenschist facies and a simple deformation history; at some localities primary structures are preserved.

The southern terrane consists of schists and gneisses with some ultra-mafic pods. South of the boundary, Laurentian shelf-sequence rocks do not occur. Southern lithologies have experienced amphibolite to granulite facies

metamorphism. Structures in these rocks indicate a complex deformation history not observed in the Laurentian rocks to the north.

To the east, the Cream Valley shear zone is continuous with the Huntingdon Valley Shear Zone, which also separates distinct lithologies and metamorphic grades. A gravity anomaly, which is interpreted to result from major crustal offset, coincides with the Cream Valley shear zone. Dissimilarities in lithology, metamorphic grade and deformation history north and south of this structure indicate that these rocks are not correlative. Paleozoic orogen-parallel displacement on a terrane boundary may explain the juxtaposition of these distinctly different metamorphic suites.

ACKNOWLEDGMENTS

Thanks to my thesis advisor, Dr. Mary Louise Hill, for encouraging me to undertake the Master's program, providing this project and helping me complete it. Thanks also to Dr. Peter Goodwin for inspiring an appreciation in a sub-discipline other than the one I came to Temple to study and for being a patient listener and enthusiastic pool coach (despite a not-so-gifted pupil). Dr. David Grandstaff, who is a Great Teacher, has provided much technical assistance on my thesis and other projects throughout my residence in this program. He deserves special mention and a Guiness for his efforts. Dr. George Myer helped with mineral identification. The faculty and staff of the Temple University geology department as a whole have assisted in the completion of this project.

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My fellow graduate students have become both good friends and valuable sources of geologic discussion which has served me well in my time at Temple. Thanks to all who have lent their ears, their opinions or their assistance. They have kept me somewhat sane.

My parents have always encouraged me in my educational pursuits and provided stability in my life, thus helping me to achieve my goals. My sister has been my voice of reason on occasion.

Brett inspired me.

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CHAPTER 1. INTRODUCTION

Hypothesis

The focus of this research is to test the hypothesis that the Cream Valley shear zone in southeastern Pennsylvania marks a terrane boundary between the edge of ancient Laurentia, to the north, and a displaced terrane, to the south. Testing the terrane boundary hypothesis involves comparison of the rocks on either side of the proposed terrane boundary, based on mapping of the surface geology. Evidence for distinct geologic histories on opposite sides of the proposed terrane boundary would support this hypothesis. The specific observations and evidence used to demonstrate distinct geologic histories are differences in metamorphism, deformation and lithologies.

Setting

Location

The Cream Valley shear zone is part of the Pleasant Grove-Huntingdon Valley shear zone system, a major through-going shear zone system that extends from New Jersey through southeastern Pennsylvania and into Maryland. The field area for this study is in southeastern Pennsylvania, just west of the Schuylkill River (**Figure 1**). The northern and southern boundaries of this study are at 40°07'30" and 40°00'00", respectively. Eastern and western boundaries are at 75°15'00" and 75°37'30", respectively. This is a heavily developed suburban area within the western suburbs of Philadelphia known as the Main Line. Field work was conducted along an approximately 12 km long section of the shear zone in an

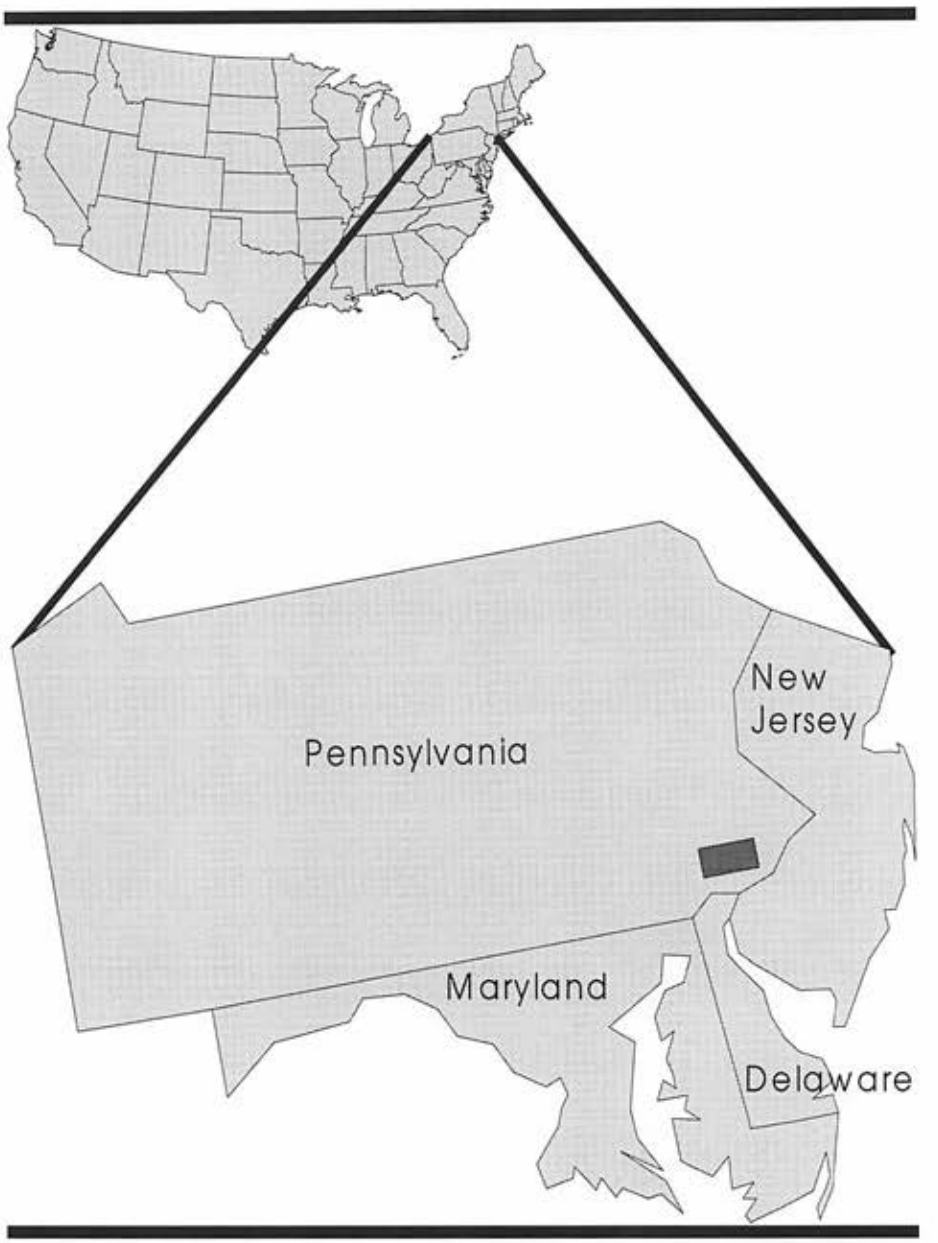

Figure 1. Field area location map.

area of sparse outcrop. Most of the outcrop exists as road cuts, railway cuts and construction sites. Occasionally, stream valleys provide outcrop as well.

Geologic Setting

This field area is located within the central Appalachian Piedmont geologic province along the Paleozoic edge of Laurentia. The Triassic Newark Basin is north of the field area and the Schuylkill River is to the east. Lithologically, this area is composed of metamorphic rock of unknown age as well as Paleozoic shelf metasediments (**Figure 2**). South of the shear zone, schists, gneisses and ultramafics are the dominant lithologies. North of the shear zone the lithologies include Cambro-Ordovician carbonates and quartzites as well as the Precambrian Honey Brook Upland suite of igneous and metamorphic basement rock.

Within the field area the Cream Valley shear zone is characterized as a 3 to 5 km wide zone of ductile deformation with evidence of dextral displacement. The fabric observed within the shear zone is typically a steeply southeast dipping (60-90°) penetrative foliation which strikes about N70°E. The dominant lithologies within the shear zone are phyllite and phyllonite. The proposed terrane boundary lies within the shear zone, near its southern boundary (**Figure 3**), and may be approximately coincident with the Cream Valley fault as it is mapped on the current version of the state geologic map (Berg, 1980).

The Cream Valley fault was proposed to explain the contact between the Precambrian gneisses (to the south) and the Wissahickon schist (to the north)

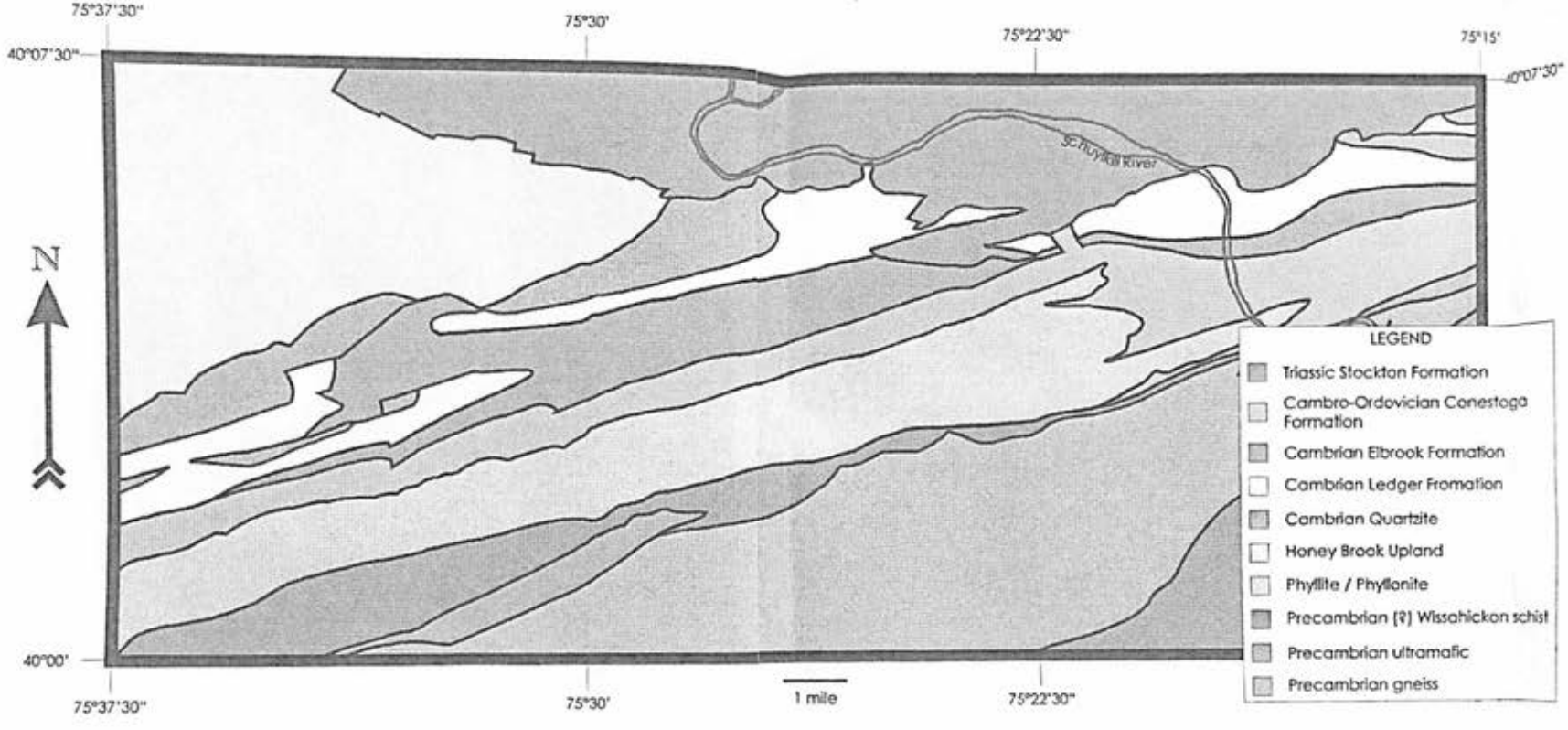

Figure 2. Lithologic map of field area. After Berg (1980).

(Bascom, 1909). Bascom (1909) originally described this contact as a thrust fault, an interpretation which has been used repeatedly as a description of the northern boundary of the Precambrian gneisses in the field area (e.g. Armstrong, 1941; Wagner, 1966 and 1972; Crawford and Hoersch, 1984). The current version of the state geologic map (Berg, 1980) represents the Cream Valley Fault as part of a continuous fault plane which stretches from Morrisville, PA to near Mortonville, PA, striking approximately N70°E.

Previous Work

The Cream Valley shear zone is part of a regional shear zone system the Pleasant Grove-Huntingdon Valley shear system, which stretches from New Jersey through southeastern Pennsylvania and into Maryland (Valentino, 1993). Various segments of this system have been studied and mapped in detail by others (e.g. Song, 1987; Valentino, 1993; Solar, 1994; Ketterer, 1995; Duan, 1997).

There is a steep regional gradient in Bouguer gravity anomalies in the region of the Cream Valley shear zone. Surface factors, such as differences in geophysical properties of rock types along the gradient, cannot wholly account for the strong gradient (Song, 1987; Duan, 1997). The probable cause of the Bouguer gravity anomaly was interpreted as a deep, nearly vertical crustal structure, which offsets the Moho by about 8 km at 25 to 33 km depth (Duan, 1997). This interpreted structure coincides spatially with surficial mapping of the Cream Valley shear zone; its strike and dip also match the orientation of the Cream Valley shear zone foliation. These coincidences, which support the

terrane boundary model, initiated the investigation of the terrane boundary hypothesis for this study.

As part of a regional through-going shear zone system, the Cream Valley shear zone is contiguous with other shear zones at both ends of the field area. To the east, the Cream Valley shear zone is essentially strike-parallel and continuous with the Huntingdon Valley Shear Zone, interpreted by Ketterer (1995) as a terrane boundary based on a mylonite zone which is a contact between two distinct rock suites (the Laurentian assemblage and the Wissahickon assemblage). West of the current field area, the Pleasant Grove shear zone is the continuation of the shear system in the Susquehanna River region. Mapping in this region revealed discordant metamorphic and structural histories across the shear zone (Valentino, 1993). To explain the juxtaposition of rocks which originated in geologically different regions, Valentino (1993) estimated a displacement of 150 km along the Pleasant Grove-Huntingdon Valley shear zone system.

The area of the current study lies between sections of the Pleasant Grove-Huntingdon Valley system which have been mapped in detail. Within this area, Solar (1994) identified the Cream Valley shear zone as the southernmost shear zone in a system comprising four individual shear zones. Detailed mapping of these shear zones has shown that the Cream Valley shear zone is the most significant shear zone in this system in terms of regional extent and displacement (Solar, 1994). Thus, the Cream Valley shear zone stands out within this system as the most likely candidate to coincide with a terrane boundary. Mapping in the

current area is necessary to constrain both the location of the shear zone system in this area and to further test the terrane boundary hypothesis.

Methods

Field Work

Field work was conducted primarily in the summer of 1996. The main goal of field work was to identify and map the exact location and lateral extent of the Cream Valley shear zone within the field area and to gather data that could be used to test the terrane boundary hypothesis. These goals were accomplished by mapping the surface geology using the Norristown, Malvern and Valley Forge USGS 7.5 minute topographic series quadrangles as base maps. Outcrop in this area is quite sparse, consisting mostly of road cuts, railroad cuts, and an occasional stream valley.

At each outcrop, a station number was assigned, its location was marked on the base map (**Figure 4**) and relevant data were gathered. Qualitative descriptions of each outcrop included color, crystallinity, hardness, mineralogy, weathering and grain size. Presence and strength of structural fabrics were also recorded. All fabrics observed were measured. Strike and dip were measured for bedding and foliation planes; trend and plunge were measured for mineral lineations and crenulation cleavage hinge axes. Oriented samples were taken wherever possible, following the methods of Prior, et al. (1987).

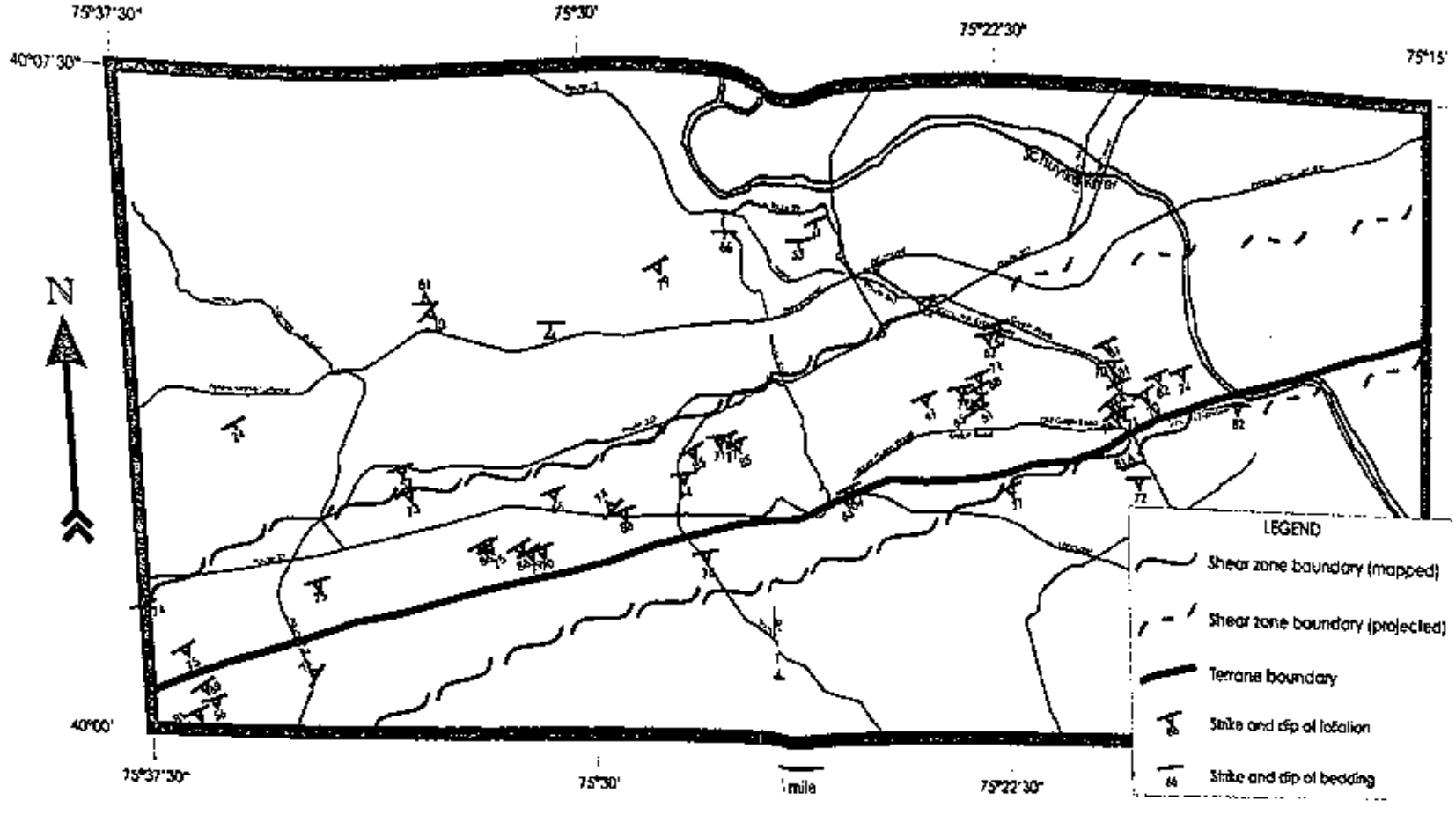

Figure 3. Structural map of field area.

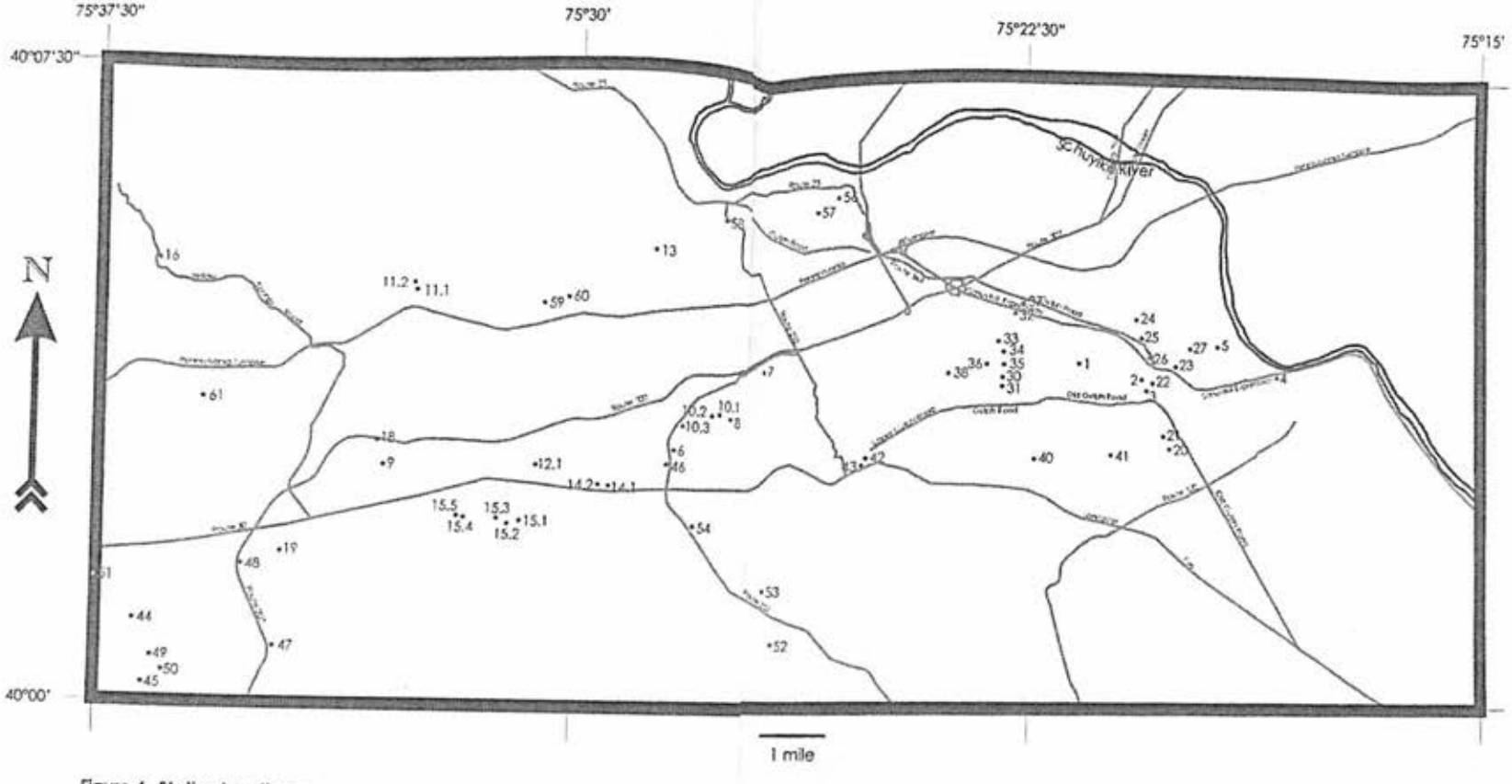

Figure 4. Station location map

Lab Work

Lab work for this project was conducted in the fall of 1996 and spring and summer of 1997. This work focused primarily on petrographic analysis of thin sections supported by some geochemical analyses.

Samples were selected for thin sectioning based on lithology and location. Thin sections of each of the major lithologies in the field area were made. Several thin sections of samples taken from within the shear zone were made to allow petrographic analysis of a north-south transect through the shear zone. This analysis was necessary to study micro-scale evidence of distinctive characteristics at the margins of the shear zone. For consistency of analysis, and in keeping with convention, all samples were prepared so that the cut plane was perpendicular to foliation and parallel to lineation (where lineations existed). This orientation allowed viewing of the XZ plane of the strain ellipsoid. Chips were glued to allow down-dip viewing.

Thin sections were analyzed by use of a petrographic microscope. Information on rock textures and mineralogy was gathered in order to characterize the rocks throughout the field area and to provide a basis for comparing and contrasting rocks that lie on opposite sides of the shear zone.

Whole rock geochemical analyses were performed using x-ray fluorescence spectrometry at Franklin and Marshall College with samples that had been partially prepared at Temple University. Eight samples were analyzed. All but one of these samples were taken from the phyllitic rock within the shear zone. The samples were chosen to represent northern and southern exposures

within the shear zone for the purpose of comparison of the two suites. The eighth sample was of Wissahickon schist south of the shear zone. This analysis was done to study potential chemical differences on opposite sides of the proposed terrane boundary.

Rock samples were crushed in a shatterbox to an 80 mesh size powder. Each sample powder was mixed with lithium tetraborate in a ratio of 0.4000 ± 0.0001 g : 3.6000 ± 0.0002 g, then heated to its melting point and poured into a mold, producing a glass disk for use in the x-ray fluorescence spectrometer, the Phillips 2400 XRF Vacuum Spectrometer with a rhodium target x-ray tube. The results of x-ray fluorescence analysis were reported as compositions in weight percentage of ten major element oxides (K_2O, CaO, Na_2O, MgO, P_2O_5, TiO_2, Al_2O_3, SiO_2, Fe_2O_3, and MnO) and parts per million of two trace elements (V and Cr). Loss on ignition was also determined by heating the powdered samples in porcelain crucibles and subtracting the weight after heating from the weight before heating.

Values reported from geochemical analyses were used to compare the abundance of elements in sample that were believed to come from opposite sides of the terrane boundary. This was done simply by comparing the raw weight percent of each major element and ppm of each trace element for all samples that were analyzed.

Structural Data Analysis

Stereonet and statistical analysis of structural data gathered in the field was done in the spring and fall of 1997. SpheriStat™ (Pangaea Scientific software,

Stesky, 1996) was used to produce stereonets and to graphically and statistically analyze similarities and differences between rock fabric orientations on opposite sides of the shear zone. Uniformity testing was used to indicate the presence of a preferred trend in the data.

Eigenvector analysis, also done in SpherlStat™, was used to determine the orientation of principal planes in different data sets. Eigenvector E3 corresponds to the statistically greatest clustering and indicates the mean of the poles to bedding. The minimum eigenvector, E1, represents a normal to the best-fit plane, or girdle, through the data. Eigenvector E1 therefore, represents an axis around which plane orientations rotate.

Woodcock ratios, K and C, determined based on eigenvector analysis, were used to interpret the distribution pattern and pattern strength for a given data set. K ratios (ln((E3/E2)/(E2/E1))) with values less than one indicated a girdle pattern; values greater than one indicated a cluster pattern. C ratios (ln(E3/E1)) with values less than C_{95} indicated a uniform distribution, and pattern strength increased with increasing C values (C_{95} is the critical C value for the 95% confidence level). All density distribution contouring was done using the Gaussian counting method (which weights the data) with K=100 (fractional counting area is 1% of the net area). All statistical tests were done at the 95% confidence level.

Objectives

The principal objective of this study is to test the hypothesis that the Cream Valley shear zone is a terrane boundary. If this hypothesis is correct, the bodies of rock on opposite sides of the shear zone will exhibit geologic histories which are

distinct from one another. This hypothesis is tested by interpreting the geologic histories of rocks on opposite sides of the shear zone and comparing them. Disparate geologic histories of rocks lying on opposite sides of the shear zone would support the hypothesis that the shear zone is a terrane boundary.

Determining the geologic history of the rocks in the field area involves several steps. First, field work is done to map the extent and location of these rocks and their relationship to the shear zone. Outcrop scale observations are used to determine the rock type and possible origin.

In the lab, petrographic and geochemical analyses supplement interpretations made in the field and add another dimension to the study. Interpretation of metamorphic history is aided by petrographic analysis. For each lithology, metamorphic facies is determined based on the mineral assemblage identified in thin section. The overall metamorphic facies, as well as the presence or absence of certain metamorphic reactions, is then used to constrain the temperatures and pressures to which each lithology has been exposed. Multiple metamorphic episodes may be identified by the presence of features such as reaction rims.

Geochemical analyses help to determine if the rock within the shear zone can be distinguished as two units based the chemical signature of each. The abundance of different elements in samples from different areas of the shear zone is compared to see if a pattern of chemical composition can be used to distinguish rocks from opposite sides of the terrane boundary.

Textures and structures observed on all scales aid in interpreting the deformational histories of these lithologies. Features such as inclusion trails indicate deformational history prior to shear-zone deformation. Preservation of primary structures indicates absence of penetrative deformation prior to shearing.

Once the geologic histories of the rocks on opposite sides of the shear zone are interpreted, the comparison of the two suites can be made. A sharp contrast in degree and extent of metamorphism and deformation on opposite sides of the shear zone, would constitute support for the terrane boundary hypothesis.

CHAPTER 2. GEOLOGY NORTH OF THE SHEAR ZONE

Overview

Lithologies north of the shear zone consist of igneous and metamorphic rocks of the Honey Brook Upland suite and Paleozoic metasediments. Metasediments lying unconformably on the Honey Brook Upland suite contain sedimentary structures that provide clues to the setting in which they formed. The metasediments include quartzites and carbonates which formed along the early Paleozoic Laurentian margin.

The metasediments appear to have been undeformed until activation of the Cream Valley shear zone to the south. Metamorphism and deformation of these rocks at greenschist-facies conditions accompanied shear zone formation.

Honey Brook Upland

The Honey Brook Upland, in the northwestern part of the field area, exposes Precambrian basement rock (Stose, 1937). This suite comprises several types of gneiss as well as an anorthosite body and two generations of dikes (Berg, 1980). The units of this suite mapped within the field area are: metadiabase, quartz monzonite and quartz monzonite gneiss, granodiorite and granodiorite gneiss, gabbroic gneiss and gabbro, graphitic gneiss and Franklin marble. Pegmatite and diabase dikes are also present (Berg, 1980).

Crawford and Hoersch (1984) subdivided the Honey Brook Upland suite into three lithologic categories: amphibolite facies gneiss, granulite facies gneiss and an anorthosite suite. The granulite facies gneisses comprise felsic and mafic units. Included in the amphibolite facies gneisses are felsic, intermediate and banded

amphibolite gneiss. The felsic units in the granulite facies gneiss and the felsic and intermediate units in the amphibolite gneiss are locally graphite-bearing. The anorthosite, lying to the west of the current field area, consists of 40% anorthosite, 40% leuco-hornblende gabbro and 20% hornblende gabbro (Crawford and Hoersch, 1984).

The gneisses of the Honey Brook Upland differ from those found south of the shear zone as they contain graphite in some places and are associated with an anorthosite body. The Honey Brook Upland gneisses are not correlative with the gneisses found in the southern terrane (Berg, 1980; Crawford and Hoersch, 1984).

Paleozoic Metasediments

Lithologies

Paleozoic shelf metasediments include carbonate and quartzite units. These units range in age from lower Cambrian to Cambro-Ordovician and unconformably overlie the Honey Brook Upland suite. All of the metasedimentary units display an essentially linear outcrop pattern which strikes parallel or subparallel to the strike of the Cream Valley shear zone (approximately N70°E).

The quartzite formations mapped in the field area are the Chickies, Antietam and Harper Formations. The Chickies is a Lower Cambrian tidal sand body and is the oldest of the metasediments in the field area. In most places it is a nearly pure quartzite with few accessory minerals. At its northern boundary it is in contact with the Honey Brook Upland. The Antietam and Harper Formations are undifferentiated in the field area and are described as Cambrian phyllite,

quartzite and quartz schist (Berg, 1980). These formations crop out just south of and lie stratigraphically above the older Chickies Formation. After Solar (1994), quartzites in this study are discussed as a single lithologic unit and referred to as Cambrian quartzites.

The Paleozoic carbonate units in the field area are the Ledger, Elbrook, Conestoga and Vintage Formations, collectively known as the Conestoga Valley carbonates. The Ledger Formation, consisting largely of peritidal carbonate facies, is the basal carbonate unit in the field area and crops out northwest of the other carbonate units (**Figure 2**). It is a Lower Cambrian, dolomite/dolomitic marble with accessory quartz and pyrite. The Elbrook Formation is a Middle Cambrian marble which stratigraphically overlies the Ledger. Mineralogically it is composed of calcite, dolomite, mica and quartz. Distinction between the Elbrook and Ledger Formations is based on the greater modal percent of dolomite in the Ledger Formation. The Conestoga Formation is a Cambro-Ordovician marble which overlies the Elbrook Formation. The Vintage Formation, a Cambrian marble and dolomite, is poorly exposed in the field area. The predominant minerals in the Paleozoic carbonates are calcite, dolomite and quartz with mica and pyrite in some localities. The Conestoga Valley carbonates are thought to mark a transition from carbonate platform to off-platform environments (Ganis and Taylor, 1990)

Structures

Structures in the quartzites are primarily of sedimentary origin. Structures present in the carbonate units are primary sedimentary structures, foliations,

lineations and folds. The orientations of both bedding planes and foliation planes were measured and statistically analyzed. Lineations were also measured but were not sufficiently abundant for statistical analysis.

Primary Structures

The Chickies Formation contains primary sedimentary structures of both biogenic and physical origin in undeformed localities. Biogenic structures are *Skolithus* tubes (Stose and Jonas, 1922; Knopf and Jonas, 1923, Goodwin and Anderson, 1974) with *Monocraterion* tops (Goodwin and Anderson, 1974). *Skolithus* tubes are long, vertical, circular to cylindrical tubes which served as dwelling burrows (**Figure 5**) (Goodwin and Anderson, 1974). *Monocraterion*, the funnel shaped structure occurring at the top of a *Skolithus* tube, is interpreted as the result of the feeding motion of the organism occupying the *Skolithus* tube (Goodwin and Anderson, 1974).

Physical structures include relict bedding with a variety of bedforms. These bedforms are graded horizontal stratification, large-scale planar cross-strata and large-scale curved cross-strata (**Figure 6**) (Goodwin and Anderson, 1974).

Based on the sedimentary structural assemblages, Goodwin and Anderson (1974) identified three distinct facies within the Chickies Formation. Of these facies, the intertidal flat facies, represented by burrowed sand, occurs in the area of this study.

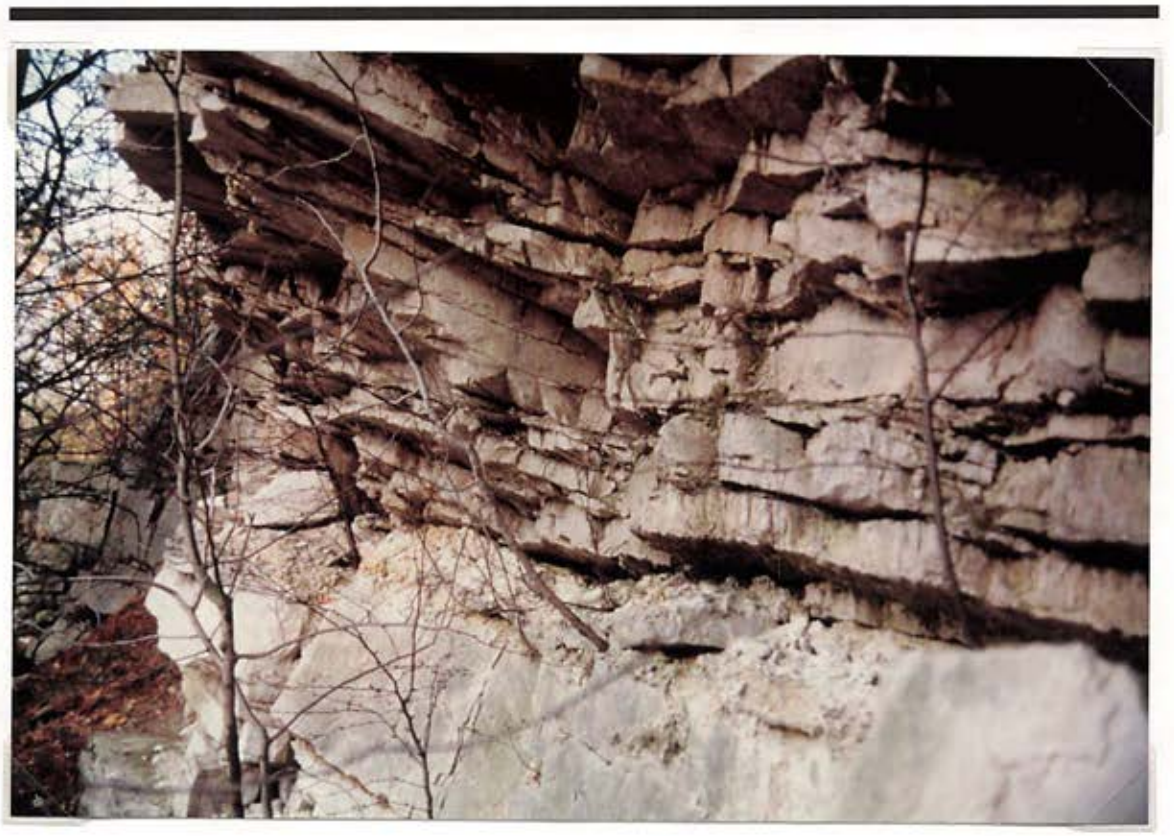

Figure 5. *Skolithus* tubes in Cambrian quartzite. These structures are thought to have been vertical dwelling burrows. Station 61.

Figure 6. Cross-beds in Cambrian quartzite. Preservation of these primary structures indicates low strain. Station 61.

Bedding

Outcrop-scale bedding is recognizable in the carbonate units at the northern margin of the shear zone near the contact between phyllite (in the shear zone) and the Conestoga Formation (station 51). Interlayers of phyllite and marble in this region are presumed to represent primary shale and limestone bedding. In addition to primary bedding, fine layering, interpreted as cryptalgal-laminites, is present in the Ledger Formation at Valley Forge National Park. The laminites are gently folded at this location (**Figure 7**).

Relict bedding orientations show a non-uniform distribution with a weakly developed girdle and cluster (**Figure 8**). The relative value of E3 is much higher than E1 or E2, indicating a cluster, or point density (**Table 1**). A higher value for E2 than E1 indicates a weakly developed girdle pattern. The Woodcock ratios confirm the weakly developed cluster and girdle pattern. The strike and dip of mean bedding orientation, derived from E3, are 084°/39° (**Figure 9**). Woodcock ratios for this data are K = .876, C = 5.903.

Table 1. Statistical Data for Bedding Orientation North of the Shear Zone

Vector	Trend	Plunge	Value	Max/Min Angle (95% cone)
E1	109	19	0.0384	7.23°/2.04°
E2	212	33	0.8937	9.51°/7.16°
E3	354	51	14.068	9.46°/2.04°

Figure 7. Gently folded cryptalgal-laminites in Ledger dolomite. Preservation of bedding at this location indicates low strain. Location 57, Valley Forge National Park.

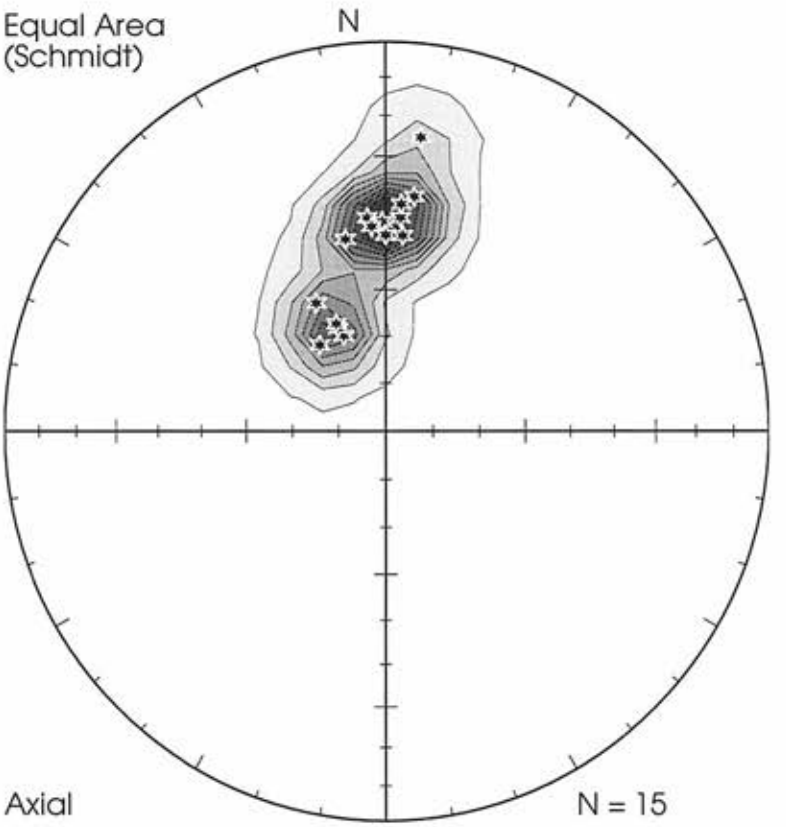

Figure 8. Poles to relict bedding and density contours north of shear zone. Consistency of bedding orientations is seen in the cluster pattern of this data set. Moderately developed girdle is due to transposition. Contours are at 2% intervals from 0-16%.

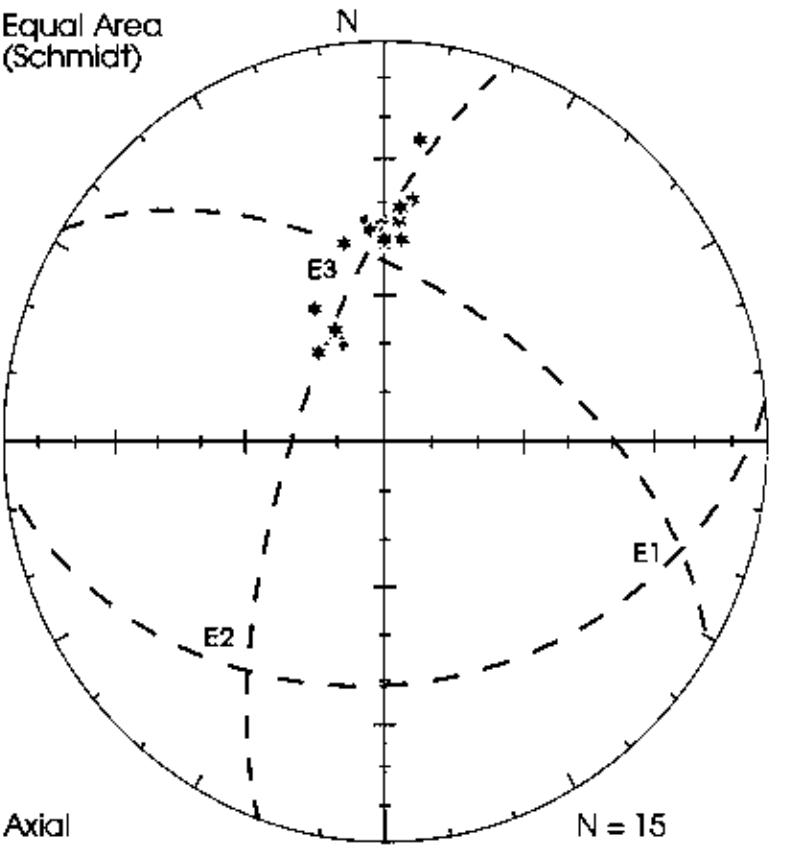

Figure 9. Eigenvectors and Planes - relict bedding north of shear zone. E1 is normal to the best fit plane through the data (girdle). E3 indicates the statistically greatest clustering. The plane normal to E3 represents the mean orientation of relict bedding (084°/39°).

The cluster pattern in this data set illustrates the fairly regular bedding orientations in the area. Variations in the bedding orientations arise as a result of gentle folding due to transposition, illustrated by the weak girdle pattern. All bedding in this domain has been transposed by shear-zone deformation. Evidence of transposition can be seen in the similar strike orientation of bedding and foliation.

Foliation

Foliation is most apparent in mixed units, such as the interlayered carbonates and phyllites. In this particular unit, foliation is manifest as a secondary planar fabric which is parallel to subparallel to, yet distinct from, bedding on outcrop scale. Parallelism of these structures is probably due to transposition of bedding during the same deformation which produced the foliation. In other locations foliation is not apparent, especially where outcrops consist of pure carbonate, such as the Valley Forge locations where Ledger Dolomite is exposed. Microscopic analysis of samples from these locations reveals no preferred orientation of mineral-grain shape or alignment that would indicate the presence of a metamorphic fabric.

Foliation orientations north of the shear zone show a preferred trend supported by statistical analysis (**Figure 10**). Eigenvalues of foliation orientations indicate a moderately developed girdle and cluster (E3>>E2>E1) (**Table 2**). Woodcock ratios agree with this pattern interpretation. The statistical mean of foliations north of the shear zone, as determined by E3, is 073°/77° (**Figure 11**). Woodcock ratios for this data are K = 1.245, C = 3.030.

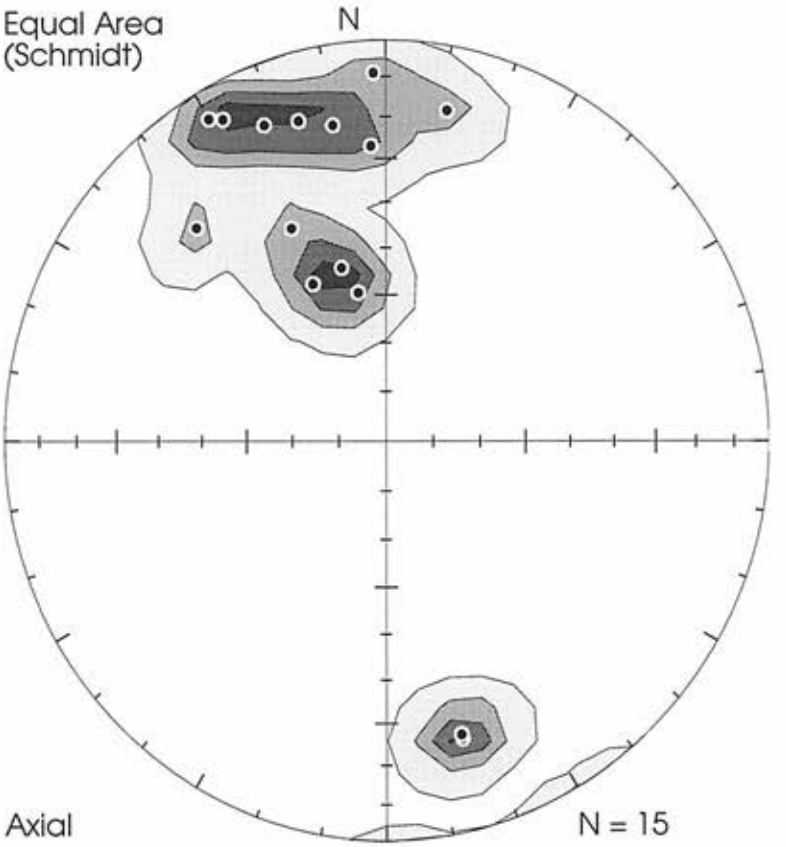

Figure 10. Poles to foliation and density contours north of shear zone. A moderately developed cluster and girdle pattern are seen here. Contours are at 0, 2, 4 and 6%.

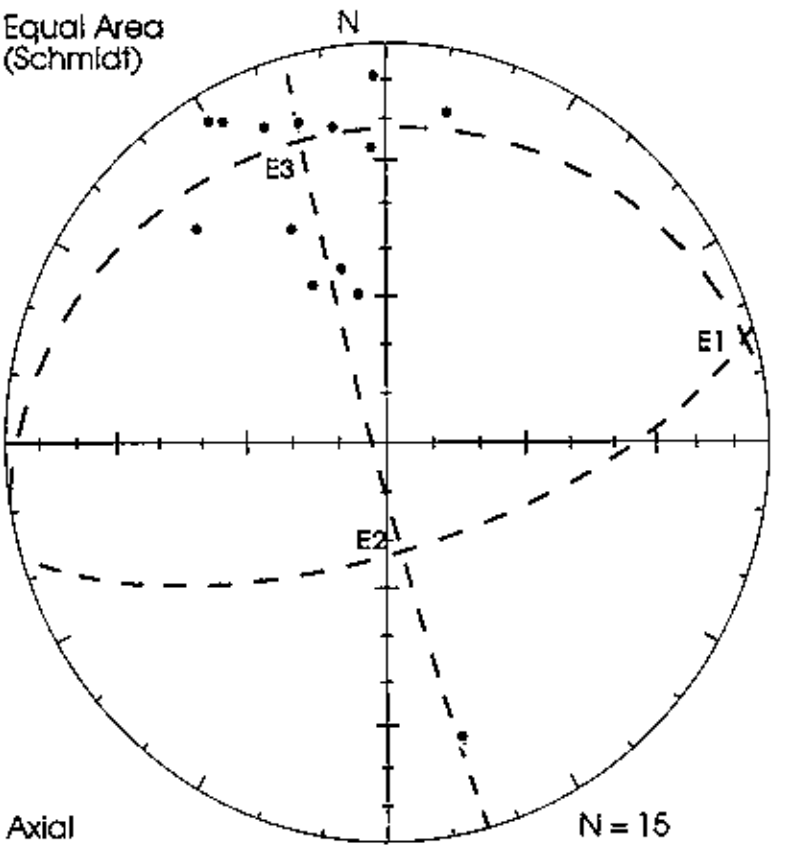

Figure 11. Eigenvectors and planes - foliation north of shear zone. E1 is normal to the best fit plane through the data (girdle). E3 indicates the statistically greatest clustering. The plane normal to E3 represents the mean orientation of foliations (073°/77°).

Table 2. Statistical Data for Foliation Orientations North of the Shear Zone

Vector	Trend	Plunge	Value	Max/Min Angle (95% cone)
E1	75	3	.5868	11.51°/7.74°
E2	172	67	2.2633	17.42°/11.36°
E3	343	22	12.150	17.33°/7.46°

Lineation

Lineations in the carbonate units north of the shear zone are sparse and when present appear to be the result of late stage brittle deformation (slickensides).

Structural Interpretation

Deformation of autochthonous Laurentian Paleozoic metasediments is indicated by structural orientations, outcrop scale folds and an overall fine grain size. Closer to the shear zone, the transposition of bedding into parallelism with metamorphic fabric is a further indication of deformation of these lithologies. However, the preservation of primary structures within these same units reveals that deformation must have been fairly gentle.

Metamorphism

Quartzite units of this sequence contain quartz and feldspar with minor mafic minerals. Minerals found in the carbonate units of the Laurentian sequence include calcite, quartz, muscovite and pyrite in some locations (**Figure 12**).

The carbonate units have a crystalline texture and fine grain size. Although all of the carbonates have a small grain size, there is an increase in size from the southernmost exposures to the northernmost exposures. Calcite

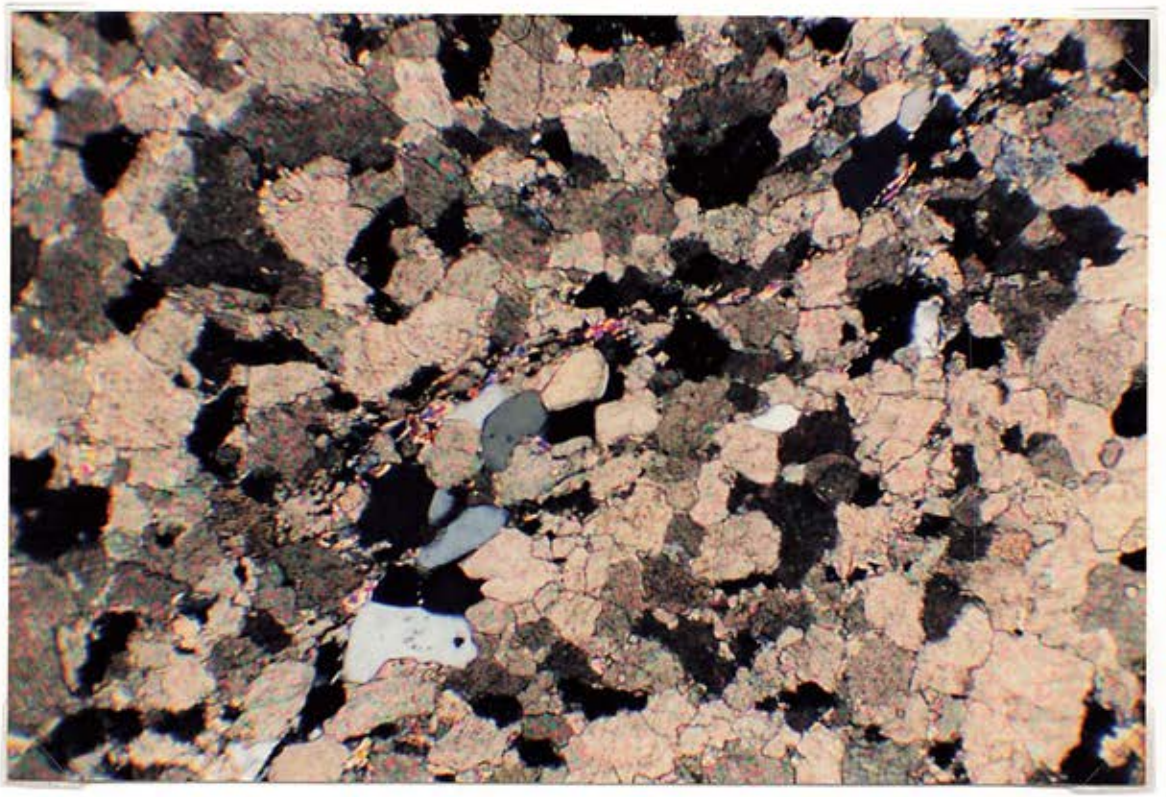

Figure 12. Mineral assemblage of Laurentian carbonate sequence. Dominant mineral is calcite. Large grey grains in upper right corner are quartz. Small bright interstitial material in center of photo is muscovite. Sample 18, cross-polarized light, Field of view ≈ 1.51 mm x .94 mm.

grain sizes from exposures of the Ledger Formation at Valley Forge National Park measure up to .73 mm, while the grain sizes of the Conestoga Formation are about an order of magnitude smaller (**Figure 13**).

The crystalline texture and development of foliation in carbonates north of the shear zone indicates that these units have been metamorphosed. The presence of phyllite associated with carbonate as interlayers along the southern carbonate boundary is further evidence of metamorphism in these units.

Based on the mineral assemblages of the formations in the Laurentian sequence, these rocks experienced peak metamorphism at greenschist-facies conditions. The presence of quartz and calcite in contact with each other, having not reacted to form wollastonite in the presence of a hydrous phase (muscovite), is indicative of greenschist-facies metamorphism. Constraining the exact temperature of peak metamorphism of these rocks is not possible because the fluid phase composition is unknown.

Geologic History

During the Cambrian and Ordovician periods, clastic and carbonate sediments were deposited in tidal settings along the Paleozoic Laurentian shelf. The Honey Brook Upland suite was the basement upon which these sediments were deposited. Structural and metamorphic evidence preserved in these rocks indicates that activation of the Cream Valley shear zone was synchronous with the single episode of low grade metamorphism and deformation that affected these rocks.

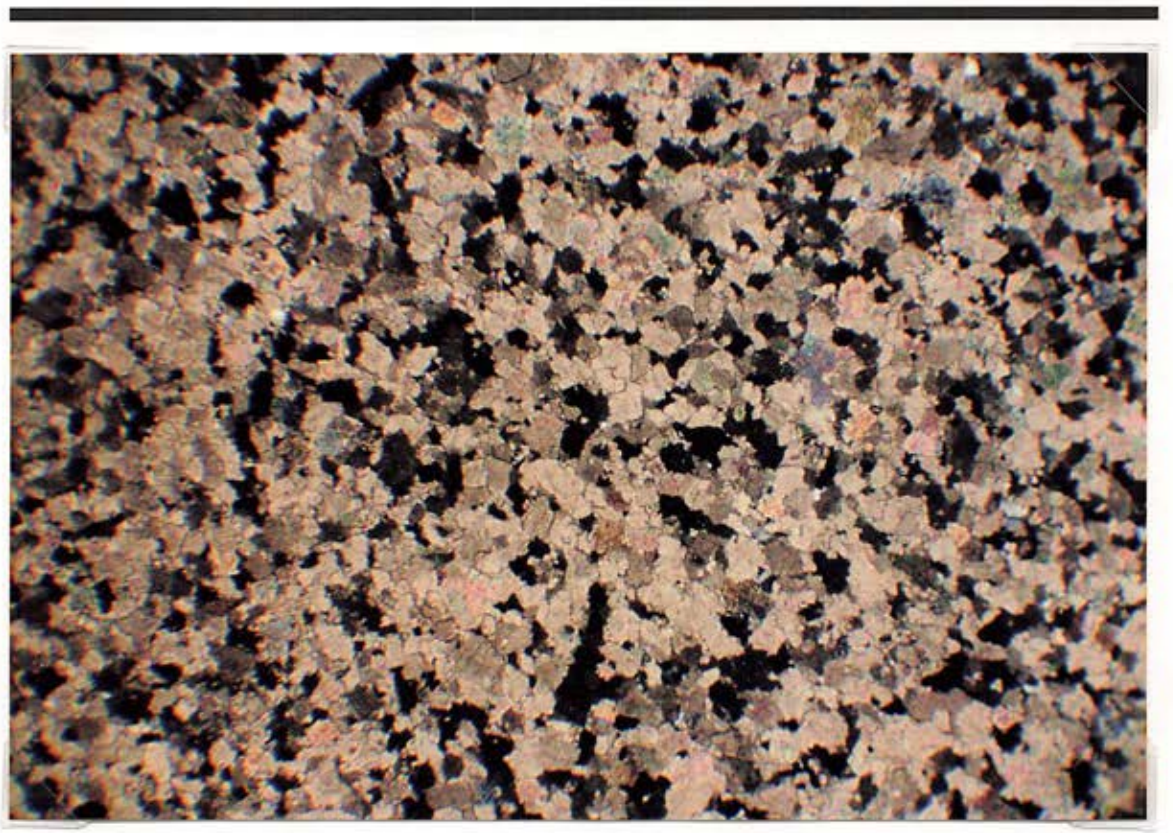

Figure 13. Texture of Laurentian carbonate units. Dominant mineral is calcite. Interlocking grains illustrate crystalline texture. Fine grain size is due to plastic deformation. Sample 18, cross-polarized light, Field of view ≈ 4.2 mm x 3 mm.

The Cream Valley shear zone was active at greenschist-facies conditions. Low-grade metamorphism in the Laurentian shelf sediments is indicated by crystalline textures and the presence of phyllite in these lithologies. The similarity of foliation orientation to shear zone orientation indicates that foliation in these rocks formed as a result of shearing. Near the northern shear zone boundary, the parallelism of bedding orientation with foliation indicates transposition by shearing. Despite metamorphism and deformation, primary structures such as *Skolithus, Monocraterion,* and bedding are preserved, indicating that these rocks have undergone only mild deformation and metamorphism at low grade.

CHAPTER 3. GEOLOGY SOUTH OF THE SHEAR ZONE

Overview

The units occurring south of the shear zone are high-grade metamorphic rocks of Precambrian age. These rocks include gneisses, ultramafic bodies and the Wissahickon Formation. Metadiabase, diabase and pegmatite dikes are also found in this region.

Petrographic and field observations indicate that these rocks have had a long, complex geologic history. Indications are that these rocks have undergone polyphase deformation and at least two periods of metamorphism at high temperatures and pressures (Wagner 1966 and 1972; Wagner and Crawford, 1973).

Lithologies

Precambrian Gneiss

The current Pennsylvania state geological map shows four distinct types of gneisses in this region: mafic hornblende-bearing, mafic pyroxene- bearing, felsic hornblende-bearing and felsic pyroxene-bearing units (Berg, 1980). These rocks were given the name Baltimore gneiss based on correlation with similar gneisses in Maryland (Bascom, 1909). They have also been referred to as Chester Prong gneiss (Wagner, 1972). The gneisses in this region are distinct from those of the Honey Brook upland (discussed in chapter two) and never crop out north of the shear zone (Berg, 1980).

Although the protolith of these gneisses is not known, many theories as to their origin have been proposed over the years. Bascom (1909) proposed that

there were two distinct facies within the gneiss: a massive granitic facies, and a gneissic facies. A sedimentary origin was proposed for the gneissic facies, based on texture (Bascom, 1909). Bascom (1909) proposed an igneous origin for the granitic facies. The gneisses of the southern terrane have also been called "hybrid rocks" and interpreted as plagioclase amphibolite of igneous origin which was injected by felsic material (Armstrong, 1941).

Depending on the isotopes used, radiometric dates of the Precambrian gneiss in the field area range from 380 to 550 Ma for biotite and 950 to 1120 Ma for zircon (Tilton *et al.*, 1960; Long *et al.*, 1959) (**Table 3**). Applying the discussion of Tilton *et al.* (1958) for work done on similar gneisses in Baltimore, dates for both zircon and biotite were interpreted to represent crystallization dates for the gneiss in the field area (Tilton *et al.*, 1960). The wide disparity between biotite-derived dates and zircon-derived dates suggests two periods of metamorphism of the Baltimore gneiss.

Table 3. Radiometric Dates for the Precambrian Gneiss in the Field Area

Location	Age (Ma)	Isotope	Mineral	Reference
Conshohocken, PA	1010	Pb^{206}/U^{238}	Zircon	Tilton *et al.*, 1960
Conshohocken, PA	1050	Pb^{207}/U^{235}	Zircon	Tilton *et al.*, 1960
Conshohocken, PA	1120	Pb^{207}/Pb^{206}	Zircon	Tilton *et al.*, 1960
Conshohocken, PA	950	Pb^{208}/Th^{232}	Zircon	Tilton *et al.*, 1960
Conshohocken, PA	380	Rb/Sr	Biotite	Tilton *et al.*, 1960
Conshohocken, PA	550	K/Ar	Biotite	Tilton *et al.*, 1960
Radnor, PA	396 ± 14	K/Ar	Biotite	Long *et al.*, 1959

Wissahickon Formation

Two members of the Wissahickon Formation are mapped within the field area (Berg, 1980). The difference in mineralogy and character of these members was explained as a metamorphic facies change by Knopf and Jonas (1923) who assigned the names oligoclase-mica schist (higher grade) and albite-chlorite schist (lower grade) of the Wissahickon Formation. The oligoclase-mica schist member, herein referred to as Wissahickon schist, is mapped north and south of the gneiss units (Berg, 1980) and occurs only south of the shear zone. Within the field area, this unit is an oligoclase, biotite, garnet, staurolite, quartz, muscovite schist.

The albite-chlorite schist member is mapped both south of and within the shear zone, and is the only unit to occur on both sides of the terrane boundary. The distribution of this unit makes it critical to this investigation, as the hypothesis being tested by this study predicts that lithologies to the north and south of the terrane boundary will be totally distinct from one another. The albite-chlorite schist member is mapped in a linear outcrop pattern which parallels the outcrop pattern of the carbonates north of the shear zone (discussed in chapter two) (Berg, 1980). The Octoraro phyllite (Bascom, 1909), which is a fine-grained phyllitic rock, is currently mapped as part of this member (Berg, 1980).

The results of this study indicate that the body of rock mapped as the albite-chlorite schist member of the Wissahickon may actually comprise two units which were deformed and metamorphosed simultaneously, during shear-zone deformation. These proposed units are a phyllonite, which formed by

deformation of the Wissahickon schist at retrograde metamorphic conditions, and a phyllite, which formed by prograde metamorphism of Paleozoic clastic sediments on the Laurentian margin.

Structures

Microstructures

Microstructures in the schists and gneisses south of the shear zone indicate that they have undergone polyphase deformation. The schist contains numerous microstructures including microfolds of interlayered quartz ribbons and muscovite. In general the Wissahickon schist is finer-grained than the gneisses.

Inclusion Trails

Garnet porphyroblasts in the Wissahickon schist contain inclusion trails. In larger porphyroblasts the inclusion trails reveal a tightly folded geometry (**Figure 14**). Folded inclusion trails form when a metamorphic mineral overgrows a pre-existing fabric, preserving the trace of the fabric in the new mineral. The tightly folded inclusion trails represent the trace of a pre-existing fabric in the schist.

Rotated Porphyroclasts

In the Wissahickon schist, pressure shadows and recrystallized tails have developed on garnet, staurolite and plagioclase porphyroclasts (**Figure 15**). The mineral composition of pressure shadows ranges from monomineralic quartz or mica to mixed quartz and mica (predominantly muscovite with small biotite and chlorite, which may be secondary in some cases); in some cases opaques (graphite) are included. Muscovite porphyroclasts form mica fish. The presence

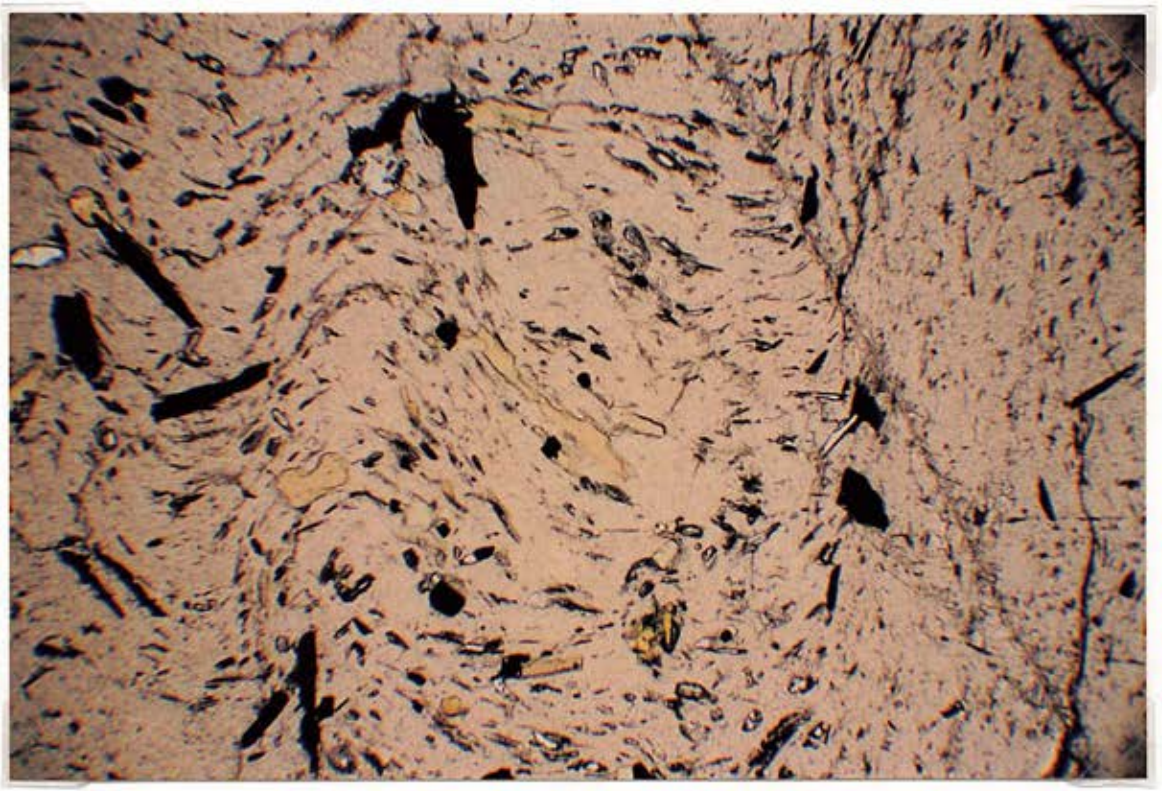

Figure 14. Tightly folded inclusion trails in garnet porphyroclast. Pattern of inclusion trails represents the trace of a fabric that was present in the rock at the time of garnet growth. Yellow inclusions are staurolite. Sample 40, plane-polarized light, Field of view≈ 4.2 mm x 3 mm.

Figure 15. Staurolite porphyroclasts with chlorite tails. Sample 40, plane-polarized light, Field of view≈ 4.2 mm x 3 mm.

of pressure shadows in combination with inclusion trails in garnet indicates a polyphase deformation of the schist.

Ductile Deformation of Feldspar

Bent deformation twins and patchy extinction of feldspar appear in the gneiss (**Figures 16 & 17**). In the Wissahickon, feldspar displays sweeping extinction. These features indicate intracrystalline deformation of both lithologies at temperatures above about 500°C (Tullis and Yund, 1992).

Fabric Orientation

If the Cream Valley shear zone had affected the gneisses south of the shear zone, an oriented fabric pattern matching that of the shear zone would have developed. No such fabric pattern exists in this area; instead, foliation orientations have a scattered distribution (**Figure 18**). Eigenvalues of data collected south of the shear zone have relatively similar values (**Table 4**) and also indicate that the fabric is non-directional. The statistical mean of foliations south of the shear zone, as determined by eigenvector E3, is 233°/84° (**Figure 19**). Woodcock ratios derived from the eigenvalues are K = 1.284 and C = 1.110.

Table 4. Statistical Data for Foliation Orientations South of the Shear Zone

Vector	Trend	Plunge	Value	Max/Min Angle (95% cone)
E1	52	73	3.0038	41.93°/24.73°
E2	255	16	4.8832	41.90°/18.42°
E3	163	6	9.1131	27.66°/18.49°

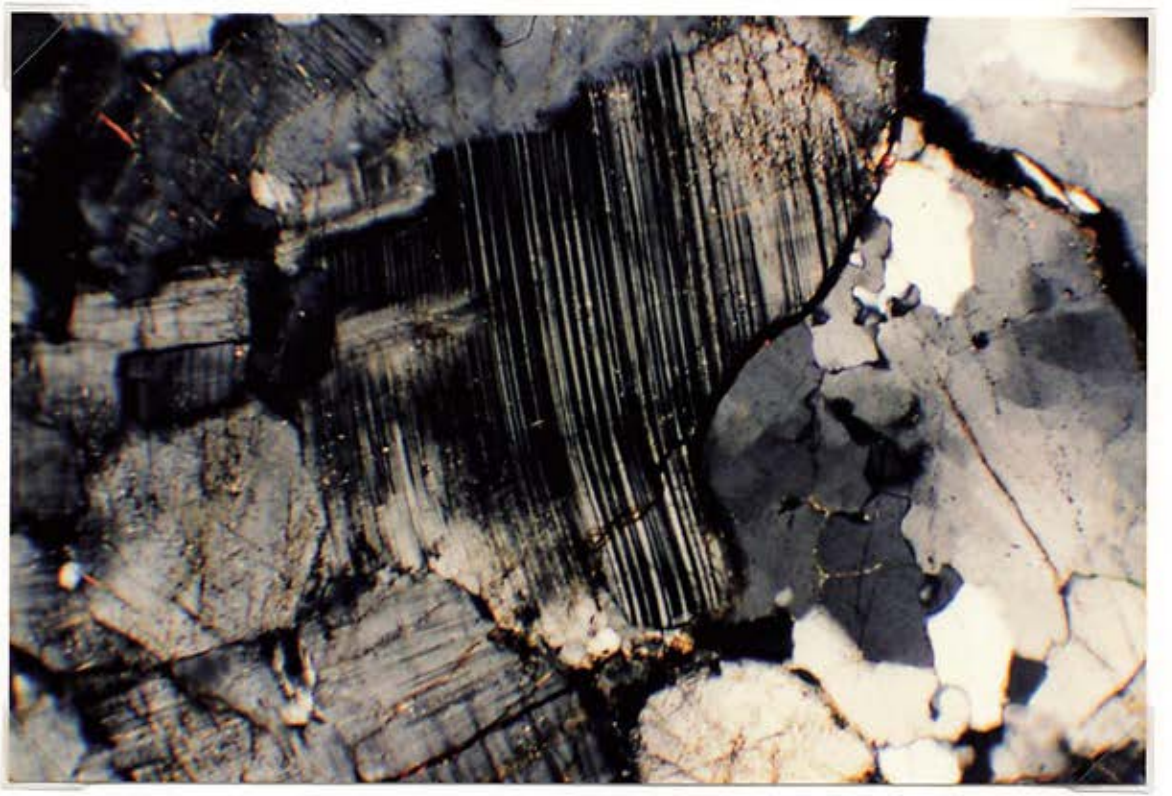

Figure 16. Plagioclase porphyroclast from Precambrian Gneiss. Bent deformation twins indicate intracrystalline deformation of this lithology. Sample 53, cross-polarized light, Field of view ≈ 1.51 mm x .94 mm.

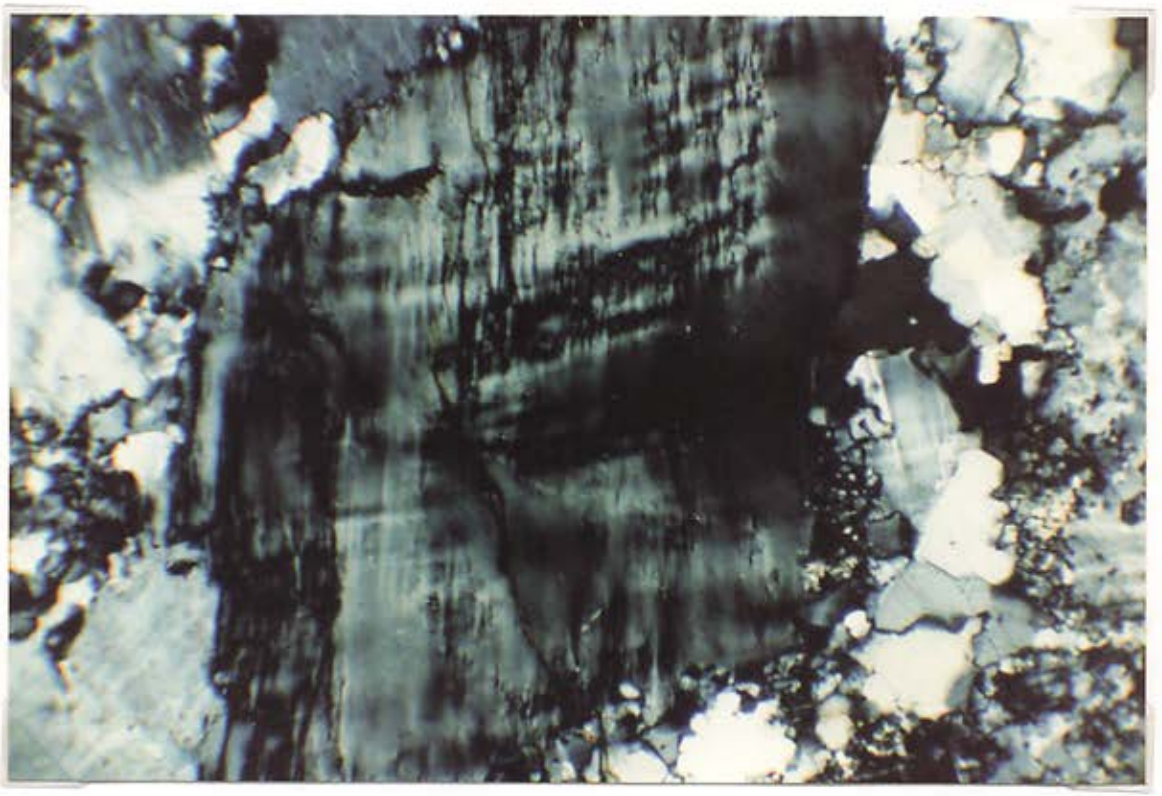

Figure 17. Orthoclase porphyroclast from Precambrian gneiss. Patchy extinction displayed by this grain indicates intracrystalline deformation at amphibolite-facies conditions. Sample 20, cross-polarized light, Field of view ≈ 1.51 mm x .94 mm.

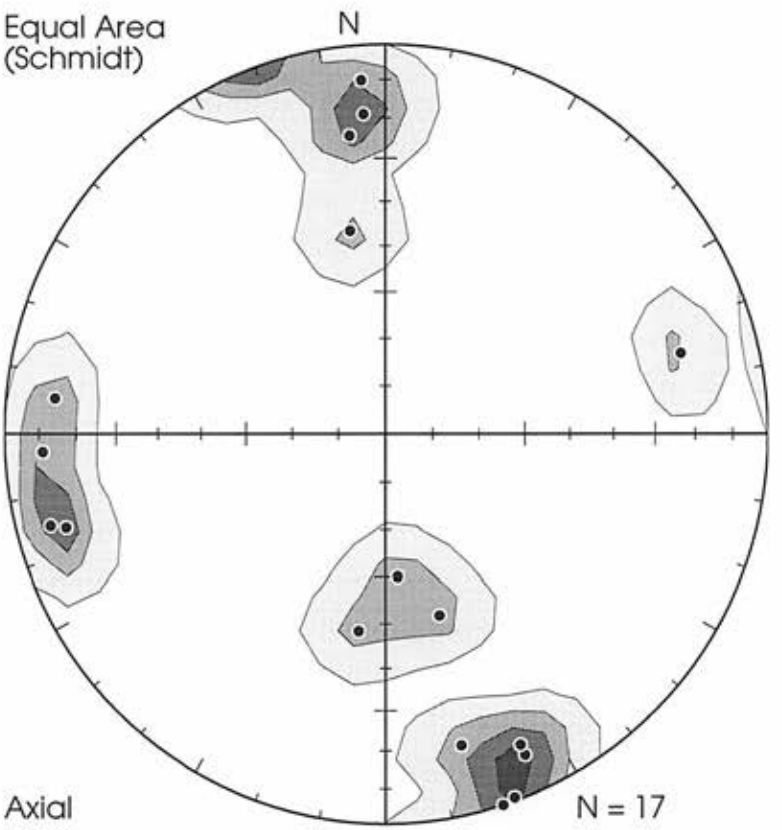

Figure 18. Poles to foliation and density contours south of shear zone. Lack of preferred orientation is demonstrated here by the scattered orientation data. Contours are at 0, 2, 4 and 6%.

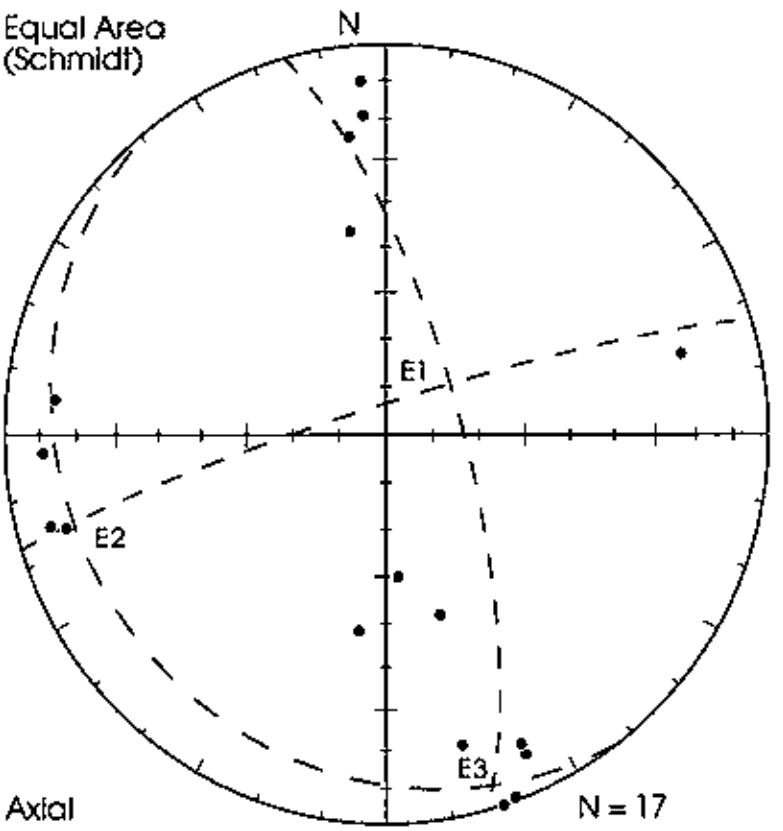

Figure 19. Eigenvectors and planes - foliation south of shear zone. E1 is normal to the best fit plane through the data (girdle). E3 indicates the statistically greatest clustering. The plane normal to E3 represents the mean orientation of foliations (233°/84°).

Parallel fabric orientations are expected in shear zones. The lack of parallelism in rocks south of the shear zone suggests a prior deformational history. This interpretation agrees with the microstructural evidence for a prior deformational history.

Metamorphism

The Wissahickon schist assemblage contains staurolite, garnet, plagioclase, muscovite and quartz as primary minerals. Opaques (graphite and hematite), biotite, chlorite and apatite are present as accessory minerals. Some chlorite may have formed by replacing biotite. Porphyroclasts of garnet (1.5 cm), muscovite (>2.15 mm), staurolite (2.15 mm), plagioclase (>1 mm) and apatite are present (**Figure 20**). Presence of staurolite in the Wissahickon schist places this unit in the Barrovian staurolite zone or above.

The felsic gneiss assemblage includes feldspar (plagioclase and orthoclase), quartz and amphibole (hornblende) as primary minerals (**Figure 21**). Epidote, titanite, tourmaline and biotite are present as accessories. Feldspars show alteration along twin planes.

Primary minerals in the mafic gneiss assemblage are quartz, garnet, orthopyroxene and plagioclase. Biotite and opaques are accessories. Sericitization has taken place on feldspar grain boundaries. Some orthopyroxene grains are partially to fully enclosed by garnet reaction rims which formed by reaction between the orthopyroxene and plagioclase (**Figure 22**) (Wagner, 1966; Wagner, 1972).

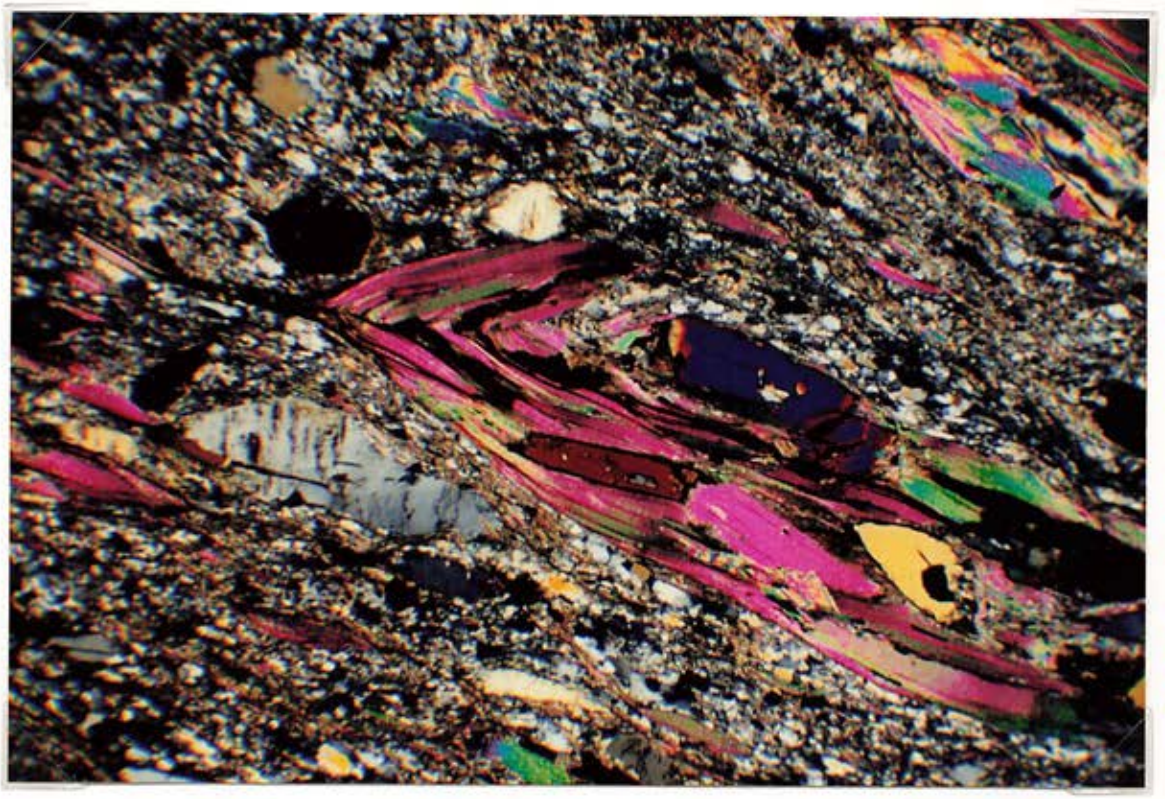

Figure 20. Porphyroclast assemblage in Wissahickon schist. Bright pink folded grain in the center is muscovite. Brownish-orange grain within muscovite, purplish-blue grain adjacent to muscovite to the upper right and pale yellow grain in lower right corner are all staurolite. Grey and white grain with twins on lower left edge of muscovite is plagioclase. Black round grain at the upper left tip of muscovite grain is garnet. Sample 40, cross-polarized light, Field of view ≈ 4.2 mm x 3 mm.

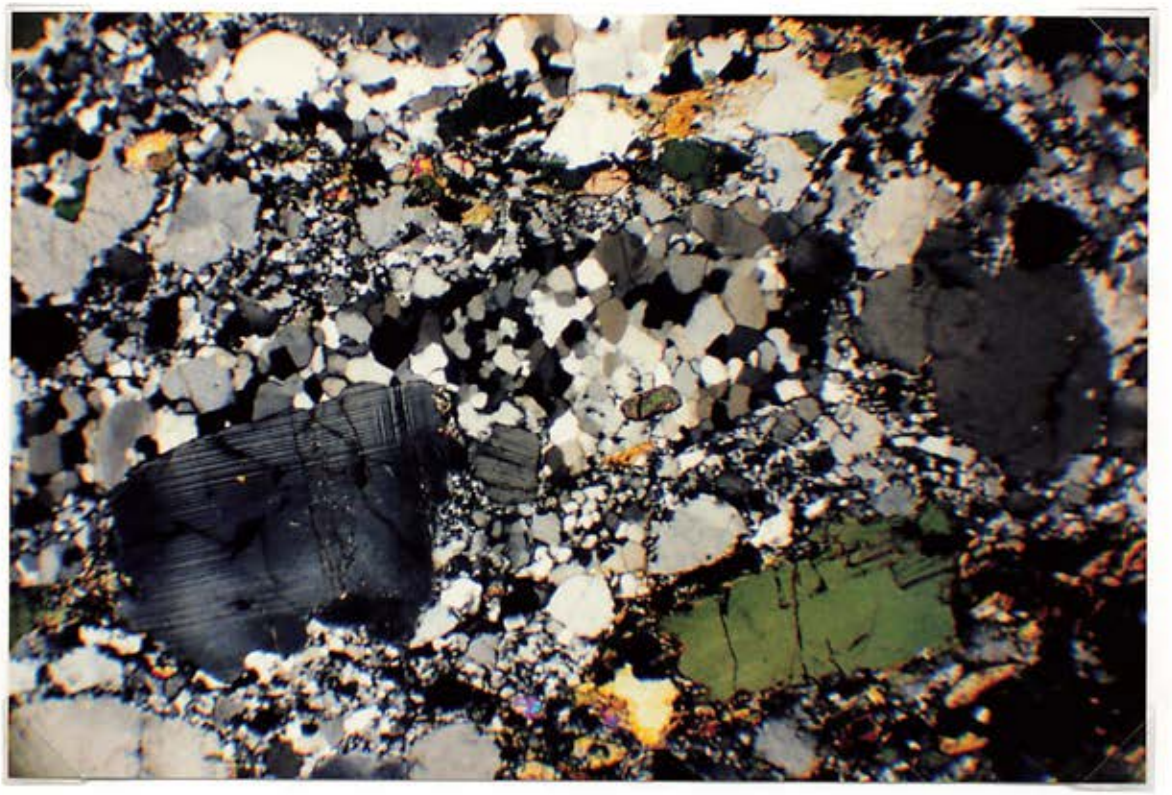

Figure 21. Mineral assemblage of felsic gneiss. Small grey grains in matrix are quartz. Large dark grey grain with twinning in lower left corner is plagioclase. Green grain in lower right corner is hornblende. Large grey grain on right edge is orthoclase. Small colored grain in center of photo is titanite. Sample 20, cross-polarized light, Field of view ≈ 4.2 mm x 3 mm.

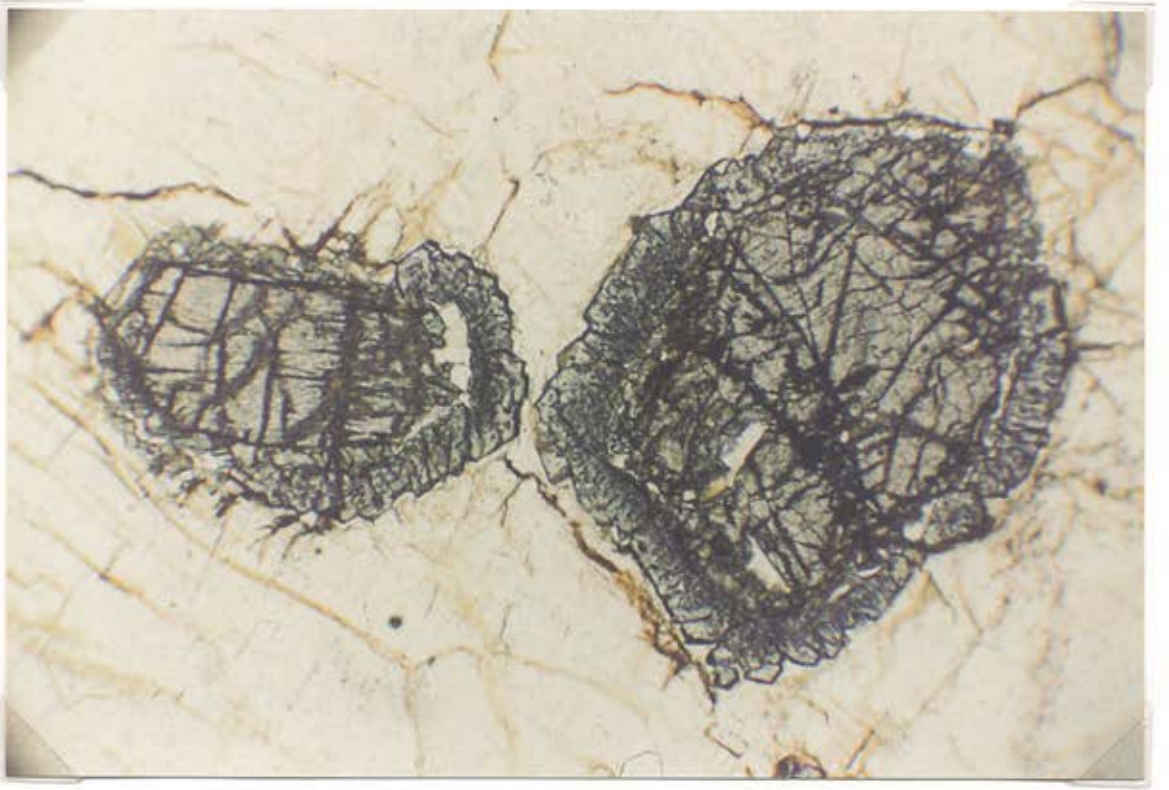

Figure 22. Garnet reaction rims in Precambrian gneiss. Garnet coronas in this figure are formed by reaction between orthopyroxene and feldspar. The pinkish-brown grains within the coronas are orthopyroxene. The material surrounding the coronas is feldspar. Sample 53, plane-polarized light, Field of view ≈ 1.51 mm x .94 mm.

The formation of garnet reaction rims has been characterized as a retrograde reaction from granulite to amphibolite facies conditions, implying at least two episodes of metamorphism in the Precambrian gneiss (Wagner, 1972; Wagner and Crawford, 1973). As evidenced by presence of orthopyroxene, the first recognized metamorphic event reached granulite facies conditions with temperatures between 850-950°C and pressures of 9-11 kb (Wagner, 1972). The oldest radiometric dates for this area (**Table 3**) range from 950-1120 Ma on zircon (Tilton *et al.*, 1960) (**Table 3**). The zircons are believed to have formed at the time of crystallization of the gneiss based on the fact that their ages match the ages of non-clastic microcline grains from the same formation (Tilton *et al.*, 1958). If the zircons were detrital, they would be expected to give an older date than that for the metamorphic microcline. The age determined from these minerals is thought to coincide with the granulite-facies metamorphic event (Wagner, 1972).

Retrograde reaction rims of garnet around orthopyroxene are evidence of a lower grade event which occurred at amphibolite facies conditions with temperatures and pressures between 650-700°C and 7-8 kb (Wagner, 1972). Radiometric dates on unstrained biotite from gneiss in the field area give a range of ages from 380-550 Ma (Tilton *et al.*, 1960; Long *et al.*, 1959) (**Table 3**). The biotite ages are interpreted to indicate the date of a second metamorphism at which time biotite was recrystallized and feldspar was deformed (Tilton *et al.*, 1958). Wagner (1972) proposed that the retrograde, amphibolite-facies metamorphism occurred during the Paleozoic. If so, the reported biotite ages may reflect this

metamorphic event which may coincide with the onset of deformation along the Cream Valley shear zone.

Geologic History

Mineral assemblages and reactions that have occurred in these rocks indicate at least two periods of metamorphism. The maximum metamorphic grade for which there is evidence occurred around 1000 Ma (Tilton *et al.*, 1958 and 1960) at granulite-facies conditions (Wagner, 1972). During this episode, temperatures and pressures reached 850-950° C and 9-11 kb (Wagner, 1972). Pre-existing fabric, preserved as inclusion trails in garnet porphyroblasts in the Wissahickon schist, may also be a relict of this high-grade, Precambrian event.

A second metamorphism occurred between 380-550 Ma (Tilton *et al.*, 1960, Long *et al.*, 1959) at amphibolite-facies conditions (Wagner, 1972). Temperatures and pressures of this metamorphism were 650-700° C and 7-8 kb (Wagner, 1972). In the gneiss, this event is expressed as the existence of garnet coronas around orthopyroxene (Wagner, 1972), recrystallization of biotite and deformation of feldspar (Tilton *et al.*, 1958). In the Wissahickon, the growth of garnet and staurolite porphyroblasts may have coincided with this metamorphic event. The timing of this event may also coincide with the earliest stages of deformation along the Cream Valley shear zone.

The exact ages of the protoliths for schists and gneisses south of the shear zone is unknown, but they must be Precambrian, as radiometric dating gives a Precambrian age for early metamorphism. The Cream Valley shear zone is a

Paleozoic structure. Therefore, the earliest deformation and metamorphism of these suites predated activation of the Cream Valley shear zone.

CHAPTER 4. THE CREAM VALLEY SHEAR ZONE

Overview

The Cream Valley shear zone is a regional-scale structure which separates domains having distinct lithologies, metamorphic histories and deformational histories. Late-stage movement along this structure occurred during deformation and metamorphism of Cambro-Ordovician passive-margin metasediments to the north of the shear zone. Based on this field evidence, the Cream Valley shear zone is Ordovician or younger. Radiometric dating of mica within the Pleasant Grove-Huntingdon Valley shear system near the Susquehanna River (west of this study) gives an age of 320 Ma (Lapham and Basset, 1964). This age was interpreted to represent the age of deformation along this structure (Valentino *et al.*, 1994).

The main rock type within the shear zone has a phyllitic character and is currently mapped as a single unit. The proposed terrane boundary lies within this body of rock, the only unit that occurs both to the north and south of the terrane boundary. Detailed field and petrographic investigations were done to determine whether this body of rock is actually a single lithology or whether it includes more than one lithology that merely look the same on outcrop scale because of the effects of shear-zone deformation. These investigations revealed significant differences between the northern and southern exposures of this lithology.

Lithologies

Throughout the field area, the Cream Valley shear zone is lithologically dominated by a fine-grained, greenish-grey phyllitic rock with a strong, penetrative foliation. This study finds that there are actually two distinct units within this body. However, the current version of the state geologic map designates this entire body of rock as the albite-chlorite schist member of the Wissahickon Formation (Berg, 1980). Previously, this body has been mapped as the Octoraro phyllite (Bascom, 1909). Within the field area, this phyllitic unit is in contact with the Conestoga Formation to the north and with the Wissahickon schist to the south.

Variations between the extreme northern and southern exposures of the phyllitic rock are a key component of this study, as they serve to establish two distinct map units and distinguish the Laurentian sequence from the southern terrane. Evidence that these are actually two separate units with different origins includes the fact that along the northern shear zone boundary this unit is interlayered with the carbonates of the Laurentian shelf sequence but along the southern shear zone boundary this unit is interlayered with the high-grade gneiss of the southern terrane (**Figure 23**).

The proposed map units are a phyllite, which occurs north of the terrane boundary, and a phyllonite, which occurs south of the terrane boundary. Phyllite is the name given to a fine-grained micaceous rock that has undergone prograde metamorphism at low-grade conditions. The term phyllonite describes a mylonite that appears similar to a phyllite, but has gained its appearance

Figure 23. Interlayered phyllite and gneiss. Interlayering of these lithologies occurs near the southern shear zone boundary and indicates heterogeneous deformation of the southern terrane. Sample 45.

through ductile shear-zone deformation involving grain-size reduction and retrograde metamorphism rather than through prograde mica growth. Both of these rock types possess a fabric defined by mica alignment and both are generally fine-grained. Recognizing the differences in these similar-looking rocks is critical to delineating the exact location of the terrane boundary within the shear zone.

In the center of the shear zone, where deformation is most intense, the distinctions between phyllite and phyllonite are blurred. This region contains the terrane boundary (**Figure 24**). Because the two units are not distinguishable in this region, the term "phyllitic" is used when discussing the rocks from this particular region and when discussing both the phyllite and phyllonite.

Geochemical analyses done on samples from within the phyllitic unit reveal a chemical composition that is typical for metapelites, with the exception of a relatively high weight percentage of Al_2O_3 and relatively low weight percentage of SiO_2 (**Table 5**). This result suggests volume loss by solution during shear-zone deformation. Soluble phases where decreased while insoluble phases were concentrated. Geochemical analyses did not indicate any chemical distinctions between the samples which could be used in identifying from which terrane they came. The homogeneity of sample chemistry is probably due to the fact that both terranes contain metapelites and that fluid flow during deformation altered the original composition of the rocks on both terranes.

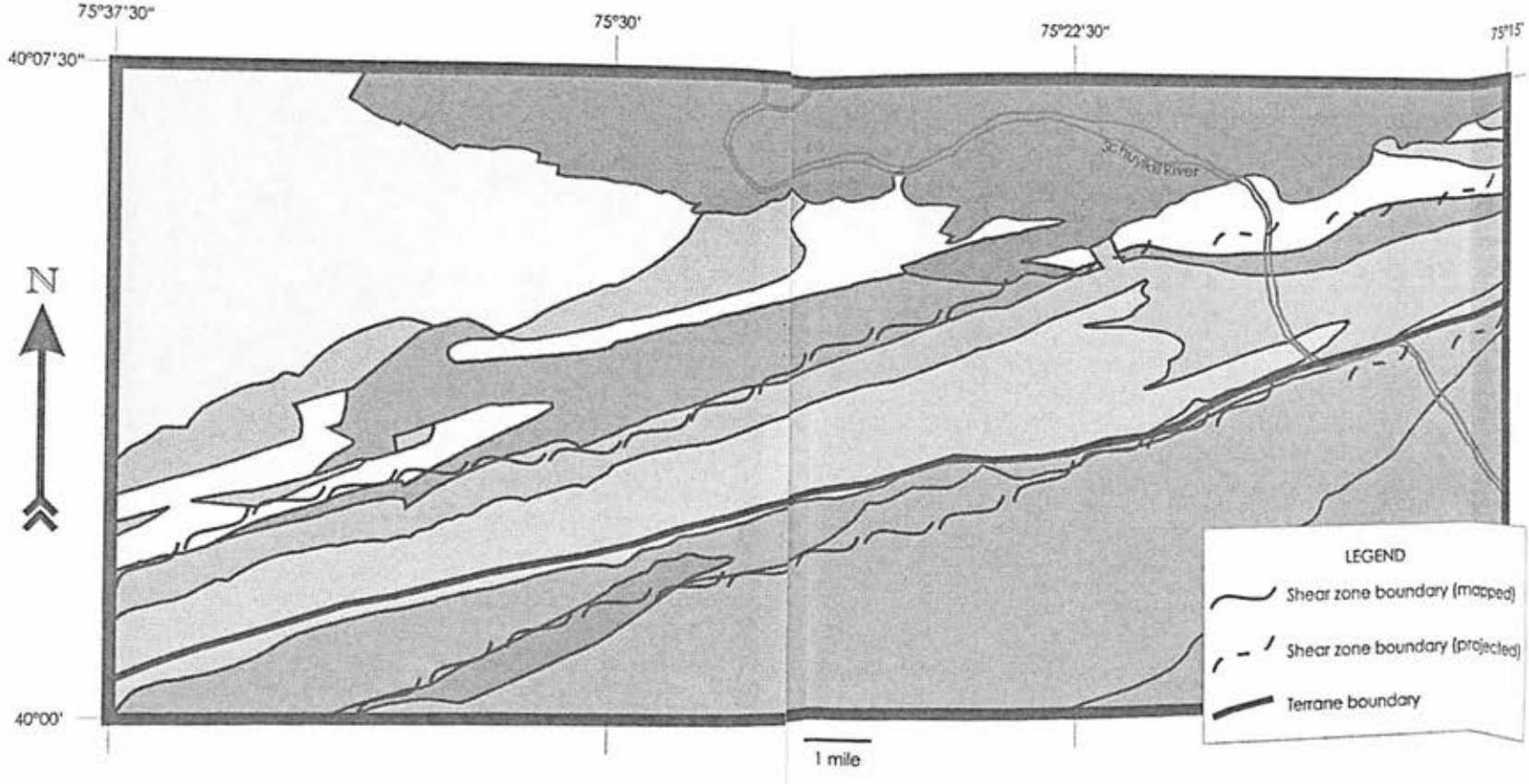

Figure 24. Relationship of lithologies to shear zone and terrane boundary. See figure 2 for legend to lithologies.

Table 5. Geochemical Analysis of Phyllitic Rocks

	1	2	6a	15c	40	45a	47a	50a
SiO_2(wt%)	53.858	44.755	47.119	49.831	59.821	44.961	51.343	36.757
TiO2(wt%)	0.939	1.228	1.516	1.007	1.108	1.658	1.234	1.760
Al_2O_3(wt%)	20.975	31.612	26.728	26.873	19.721	26.441	24.448	31.232
Fe_2O_3(wt%)	13.466	8.200	13.067	10.041	9.237	12.791	11.808	13.443
MgO(wt%)	2.194	1.033	1.934	1.527	1.884	2.067	1.268	2.426
MnO(wt%)	0.060	0.059	0.070	0.045	0.116	0.192	0.163	0.096
CaO(wt%)	0.058	0.097	0.076	0.163	1.107	0.718	0.507	0.269
Na_2O(wt%)	0.523	1.970	1.963	2.080	1.650	1.241	2.647	0.986
K_2O(wt%)	3.579	5.094	2.498	3.430	3.553	5.521	3.316	5.560
P_2O_5(wt%)	0.120	0.134	0.099	0.187	0.181	0.219	0.185	0.191
LOI(wt%)	4.813	6.470	5.603	4.884	2.100	4.395	3.492	6.501
Totals	100.585	100.652	100.673	100.068	100.478	100.204	100.411	99.221
V(ppm)	105.434	142.978	142.257	114.078	118.866	203.502	183.496	184.031
Cr(ppm)	91.147	110.390	94.311	80.080	74.106	100.445	91.068	184.844

Structures

Planar deformation fabric is the dominant structure within the shear zone. In the center of the shear zone, foliation is very strong and penetrative. Toward the northern and southern shear-zone boundaries, shear-zone deformation was less intense, and pre-existing differences in the rock influenced expression of the shear zone fabric.

Along the northern shear zone boundary, Laurentian carbonate rocks are interlayered with the phyllitic unit. This interlayering appears to be preserved primary bedding of mudstone and limestone. In this case, this unit is a phyllite, a