metamorphosed clastic sedimentary lithology, belonging to the Laurentian shelf sequence. These interlayers are distinct from foliation, but, the difference in orientation is only a few degrees, indicating that the bedding was transposed during shear-zone deformation. Station 51 has an excellent example of this structural relationship.

Along the southern boundary, the phyllitic unit is interlayered with the high-grade metamorphic rocks of the southern terrane. In this case, this unit is a phyllonite which belongs to the southern terrane. This feature lacks the indication of bedding that is seen in interlayering along the northern boundary. No distinct bedding is observed to the south, either along the shear zone boundary, where interlayering is observed, or anywhere else south of the shear zone. Also, cyclicity, which is the hallmark of sedimentary stratigraphy, is not observed in these layers. The chemical compositions of the Wissahickon schist and the phyllonite are very similar, indicating that the layering is not a result of compositional differences **(Table 5)**.

The development of phyllonite along the southern shear-zone boundary is thought to be the result of heterogeneous syntectonic retrograde metamorphism of the Wissahickon schist (*e.g.* Baker, 1987). Strain imposed by shearing caused discrete phyllonitization of some layers while others retained their high-grade metamorphic character.

Microstructures

In the center of the shear zone the phyllite is not distinguishable from the phyllonite owing to the intensity of shear-zone overprinting. In this region, on the microscopic scale, the shear zone foliation is very strong and is defined by ribbon quartz with muscovite and chlorite interlayering. Quartz shows subgrain formation and very strong crystal lattice preferred orientation (**Figure 25**). At some localities, larger-grained muscovite porphyroclasts are also present, some of which occur as mica fish (**Figure 26**). The grain size of minerals in the core of the shear zone is much smaller than that of minerals along the shear zone boundaries and beyond. This indicates grain-size reduction due to shear-zone deformation. Many minerals are grain-size reduced to the point that identification is not possible.

South of the terrane boundary, metamorphic oligoclase porphyroclasts of up to 1.5 mm occur. Some of these porphyroclasts contain inclusion trails which comprise chlorite, quartz, opaques, amphibole, epidote and tourmaline (**Figure 27**). Inclusion trails represent the fabric present in a rock at the time the host mineral grew (Shelley, 1993). Inclusion trails therefore indicate a pre-existing metamorphic fabric for this lithology.

There are two categories of oligoclase porphyroclasts, which are recognized based on inclusion trail morphology and crystal lattice deformation. Some of the inclusion trails in oligoclase porphyroclasts are crenulated (**Figure 28**) while others are straight (**Figure 29**) (Cushing, 1988). Some of the feldspar porphyroclasts show patchy or sweeping extinction (**Figure 30**), indicating some intracrystalline deformation at no less than 500° C (Tullis and Yund, 1992), while some appear unstrained. Whenever evidence of intracrystalline deformation is observed, the inclusion trails are either straight or missing. Porphyroclasts containing crenulated inclusion trails do not show signs of plastic deformation.

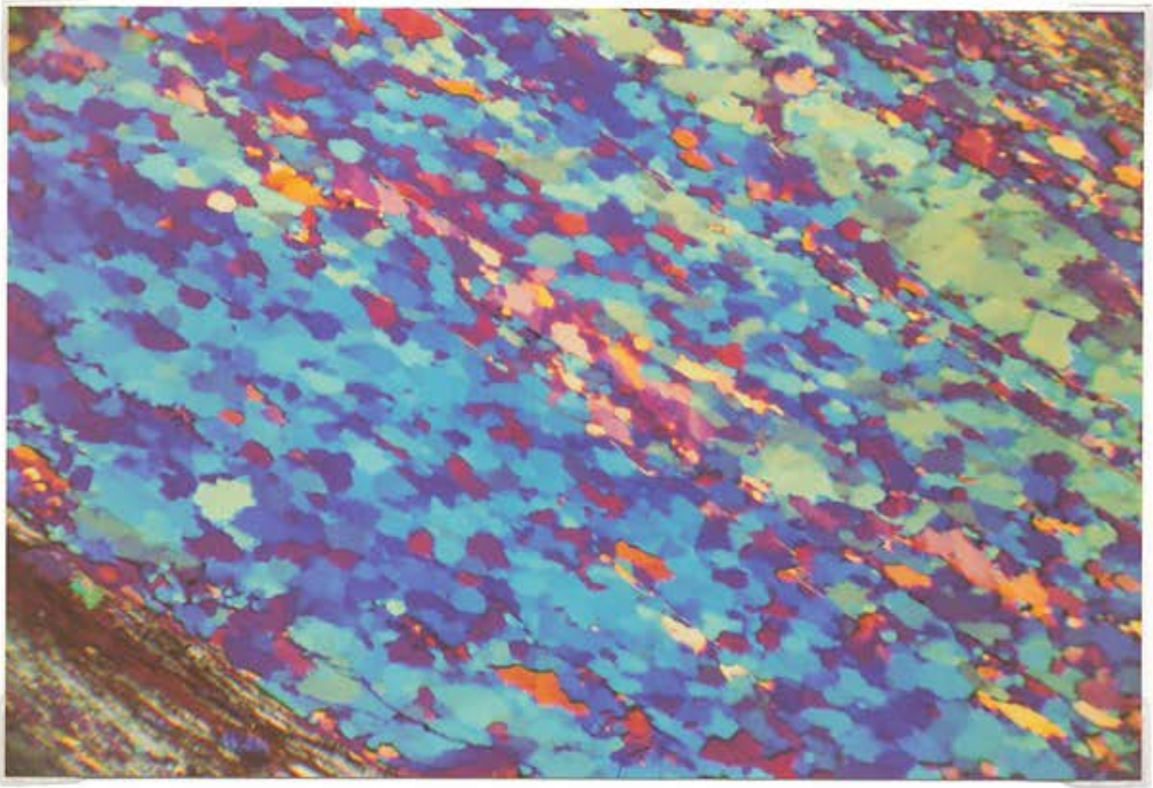

Figure 25. Quartz subgrains and crystal lattice preferred orientation. Uniformly colored areas are subgrains. Uniform blue color with insertion of the first-order plate indicates crystal lattice alignment. Both of these microstructures indicate crystal lattice deformation of quartz. Sample 45, cross-polarized light with first-order plate, Field of view ≈ 4.2 mm x 3 mm.

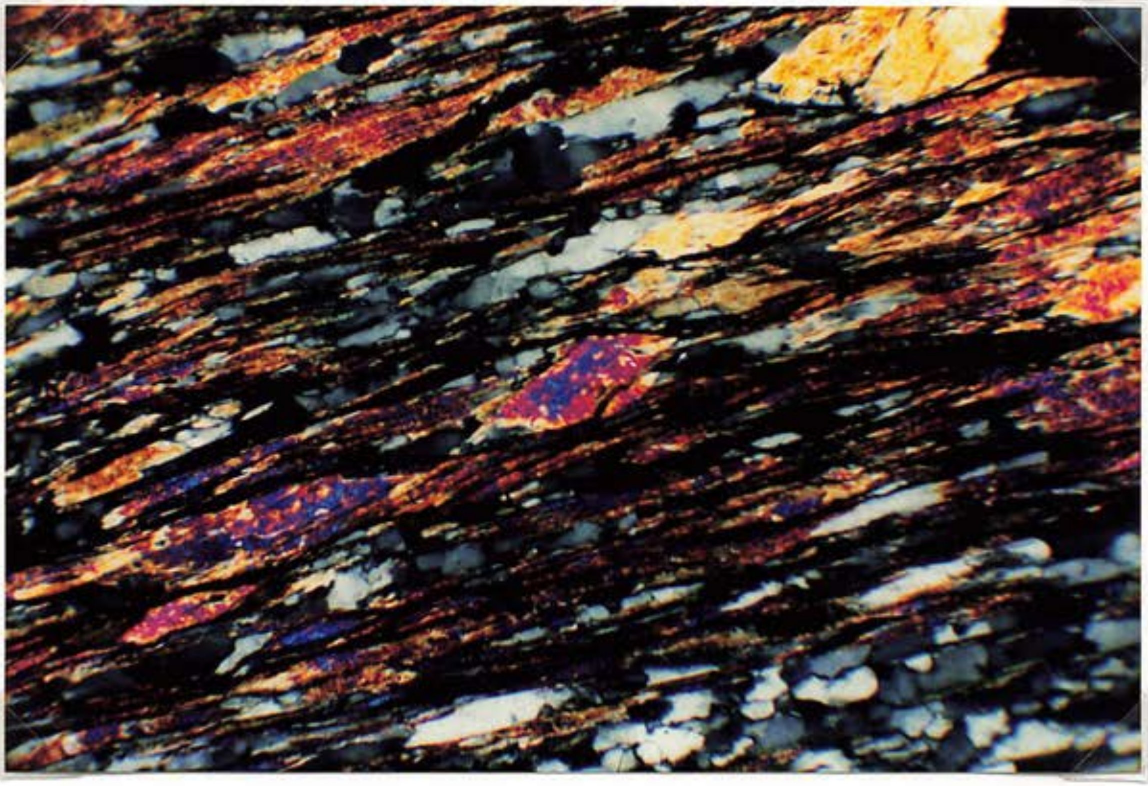

Figure 26. Muscovite porphyroclast (mica fish) in shear zone. Sample 2, cross-polarized light, Field of view ≈ 1.51 mm x .94 mm.

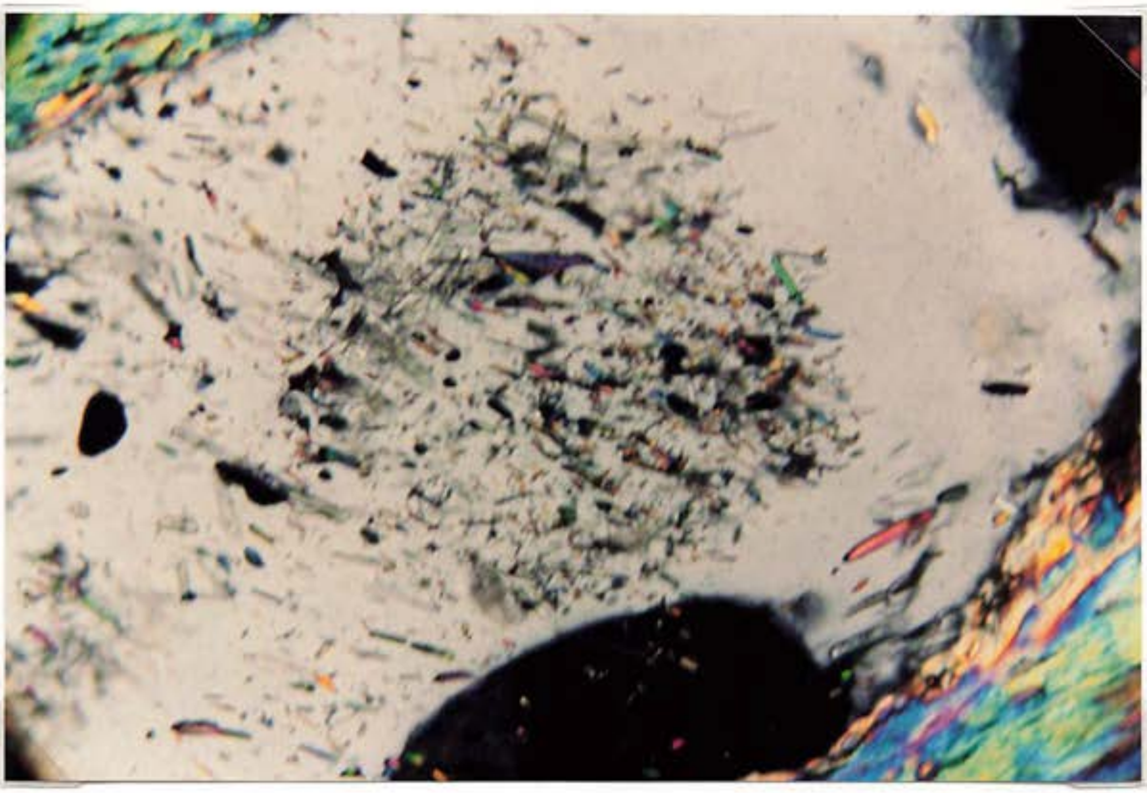

Figure 27. Mineralogy of inclusions in oligoclase. Inclusions comprise chlorite, quartz, opaques, amphibole, epidote and tourmaline. Sample 45, cross-polarized light, Field of view ≈ .63 mm x .4 mm.

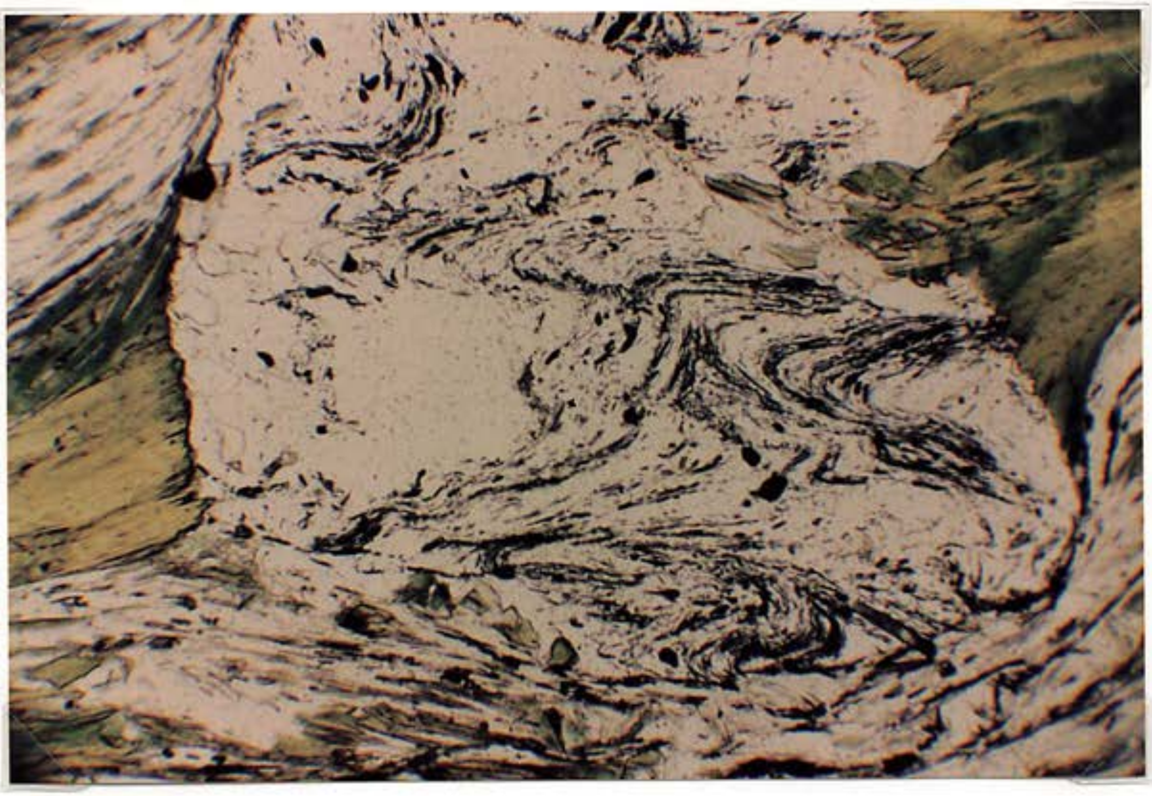

Figure 28. Crenulated inclusion trails in oligoclase porphyroclasts. The crenulated pattern represents the trace of a fabric in the phyllonite at the time this grain grew. Sample 45, plane-polarized light, Field of view ≈ 1.51 mm x .94 mm.

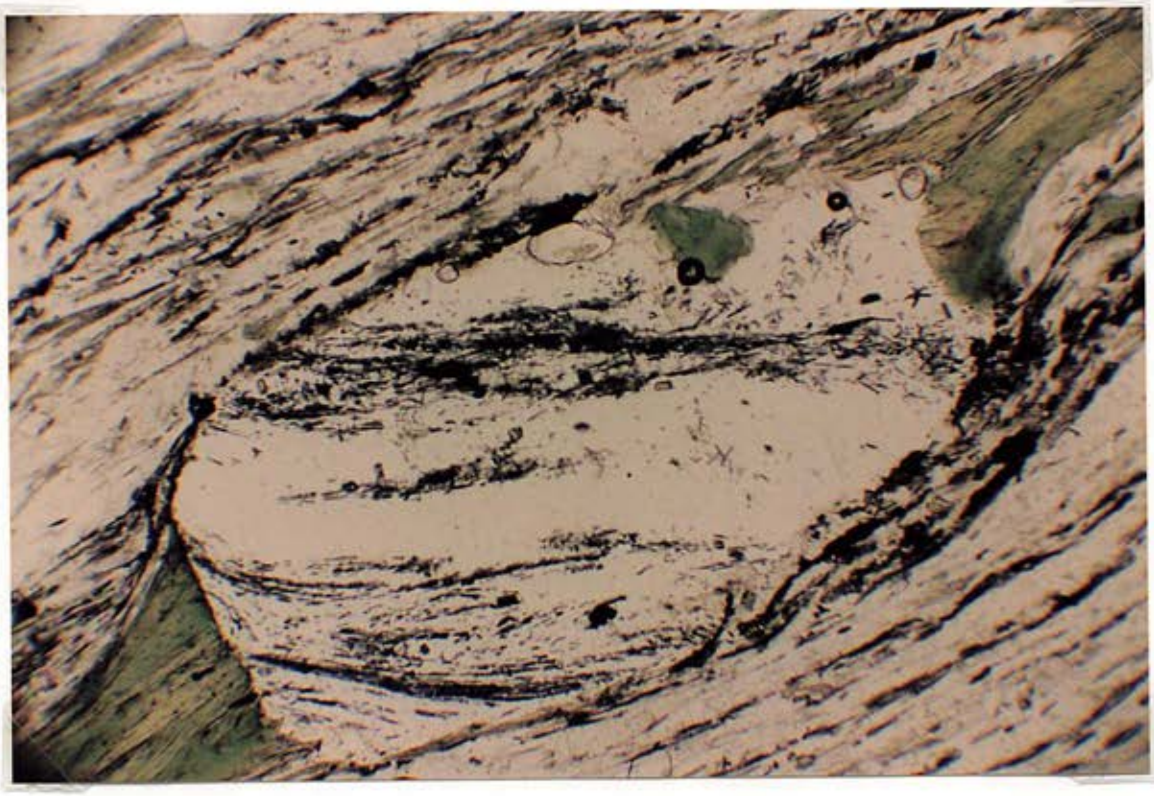

Figure 29. Straight inclusion trails in oligoclase porphyroclasts in phyllonite. Sample 45, plane-polarized light, Field of view ≈ 1.51 mm x .94 mm.

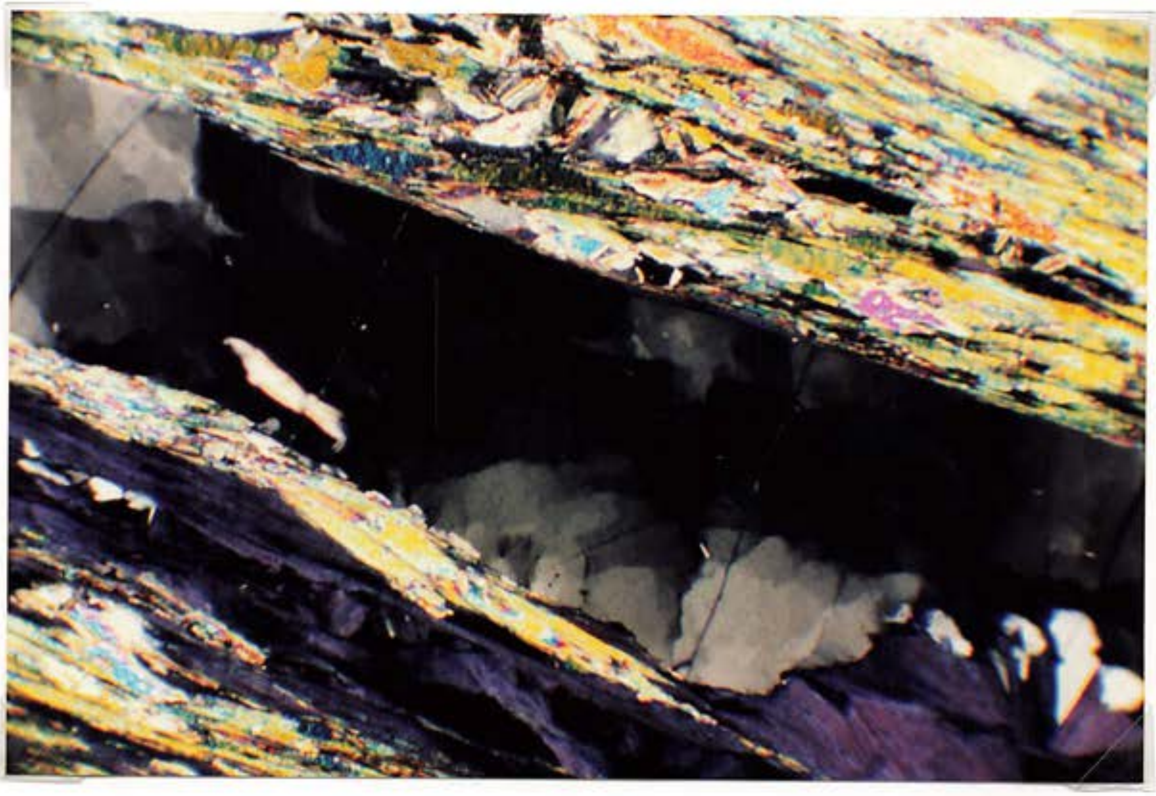

Figure 30. Patchy extinction in feldspar porphyroclast in phyllonite. Patchy extinction and crystal shape indicates crystal lattice deformation of this mineral. The temperature at which this deformation occurred was at least 500°C. Sample 45, cross-polarized light, Field of view ≈ 1.51 mm x .94 mm.

The differences in inclusion trail morphology and intracrystalline deformation are believed to be influenced by the porphyroclast orientation.

The inclusion trails in oligoclase, whether straight or crenulated, are not in alignment with the matrix fabric. Additionally, strongly crenulated trails contrast with the relatively straight matrix fabric. Because none of the oligoclase inclusion trails is coincident with the present matrix-fabric orientation, they are interpreted to represent a fabric which existed in this rock prior to the production of the present fabric.

Isolated crenulations of quartz and mica-defined foliation also occur in the southern portion of the shear zone (**Figure 31**). These micro-structures are not consistent with the planar shear-zone fabric and are truncated by the shear-zone foliation. These crenulations are interpreted as remnants of the same fabric which appears as inclusion trails in some of the oligoclase porphyroclasts. Truncation and inclusion in porphyroclasts of the crenulations indicate that these structures are the result a deformational event which predates shear-zone deformation.

Euhedral pyrite porphyroclasts similar in size to the oligoclase porphyroclasts are also present. Both pyrite and feldspar porphyroclasts have recrystallized chlorite tails (**Figure 32**).

Fabric Orientation

Foliations

The dominant fabric within the field area is the strong and penetrative foliation which strikes about N70°E and recurs throughout the field area. The consistency of foliation orientation and parallelism to the overall shear zone

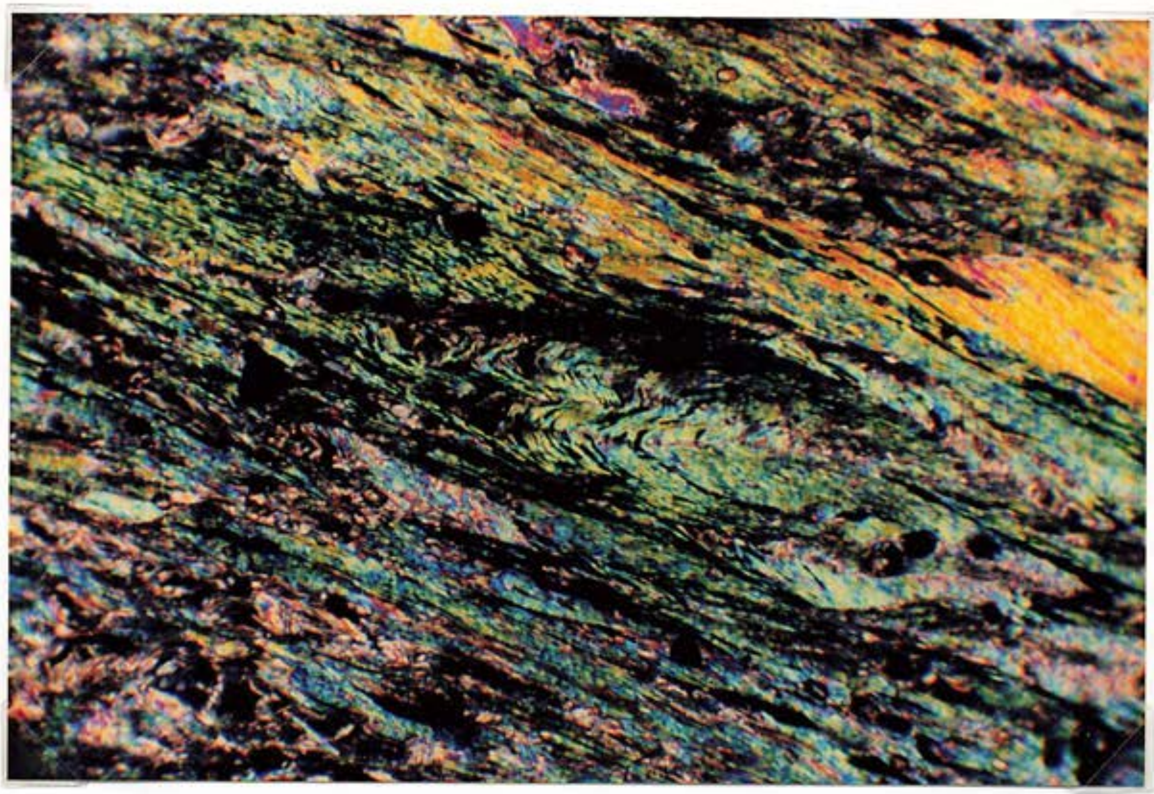

Figure 31. Isolated crenulation within shear-zone fabric. Crenulations are truncated by the dominant straight foliation of the shear zone. These structures are believed to represent an older fabric which was present in this lithology prior to shear-zone deformation. Sample 47, cross-polarized light, Field of view ≈ 1.51 mm x .94 mm.

Figure 32. Pyrite porphyroclast with chlorite tails in phyllonite. Sample 45, plane-polarized light, Field of view ≈ 1.51 mm x .94 mm.

orientation **(Figure 33)** indicate that this fabric formed synchronously with the shear zone. The mean foliation orientation, as derived from eigenvector E3, is 069°/74° **(Figure 34)**. Eigenvalues support the cluster pattern in the data set (E3>>E2≈E1) **(Table 6)**. Woodcock ratios for this data set are K = 2.894 and C = 3.271.

Table 6. Statistical Data for Shear Zone Foliation Orientations

Vector	Trend	Plunge	Value	Max/Min Angle (95% cone)
E1	245°	14°	7.4878	12.44°/1.78°
E2	116°	69°	17.345	12.44°/2.72°
E3	339°	16°	197.17	2.72°/1.79°

Lineations

Lineations are abundant in the phyllitic units within the shear zone. Lineations occur as crenulation hinge axes which form conjugate sets. The angle between conjugate crenulations is too low to discriminate two distinct clusters in stereographic analysis. Mineral stretching lineations were not recognized. The trend of lineations within the shear zone is close to the overall strike of the shear zone **(Figure 35)**. The mean lineation orientation, derived from eigenvector E3, is 32°/236° **(Figure 36)**. A moderately developed cluster is consistent with the eigenvalues for this data (E3>>E2≈E1) **(Table 7)**. Woodcock ratios for this data set are K=2.246 and C=2.628.

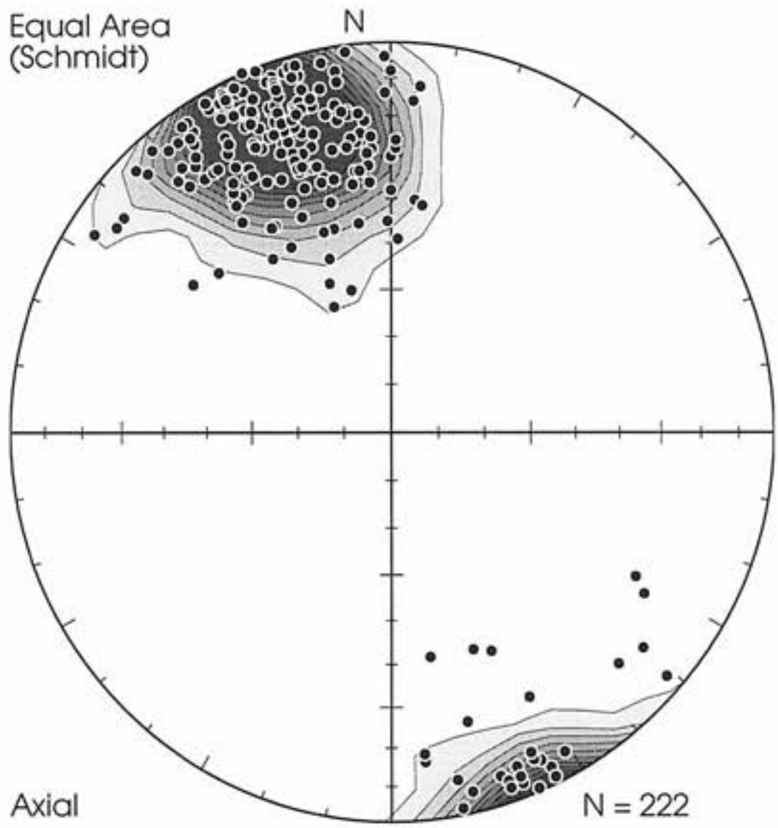

Figure 33. Poles to foliation and density contours within shear zone. Cluster pattern demonstrates consistency of fabric orientation caused by shear-zone deformation. Contour lines are at 2% intervals from 0-16%.

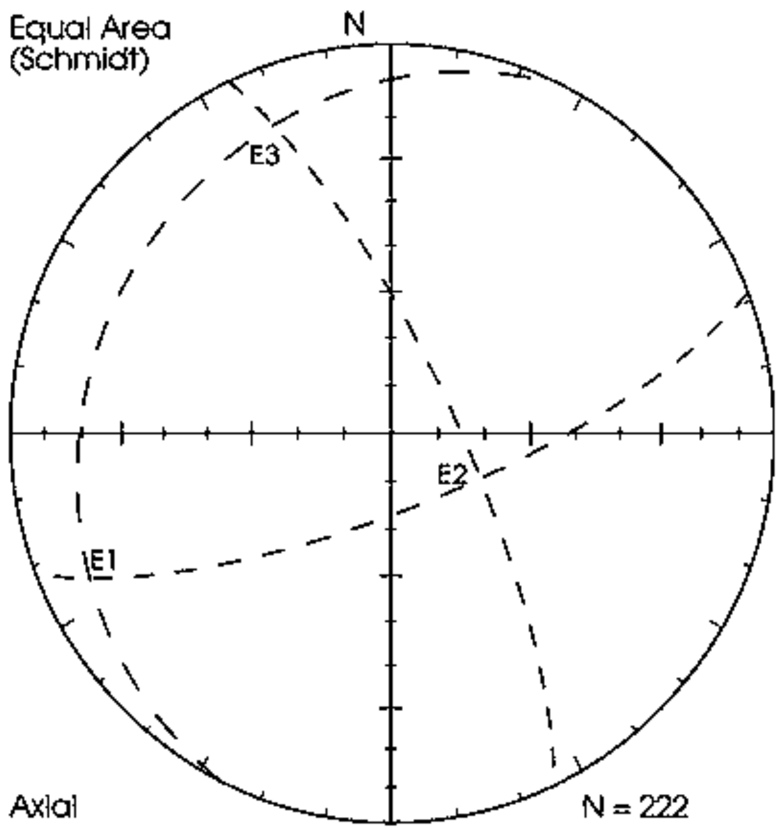

Figure 34. Eigenvectors and planes - shear zone foliation. E1 is normal to the best fit plane through the data (girdle). E3 indicates the statistically greatest clustering. The plane normal to E3 represents the mean fabric orientation (069°/74°).

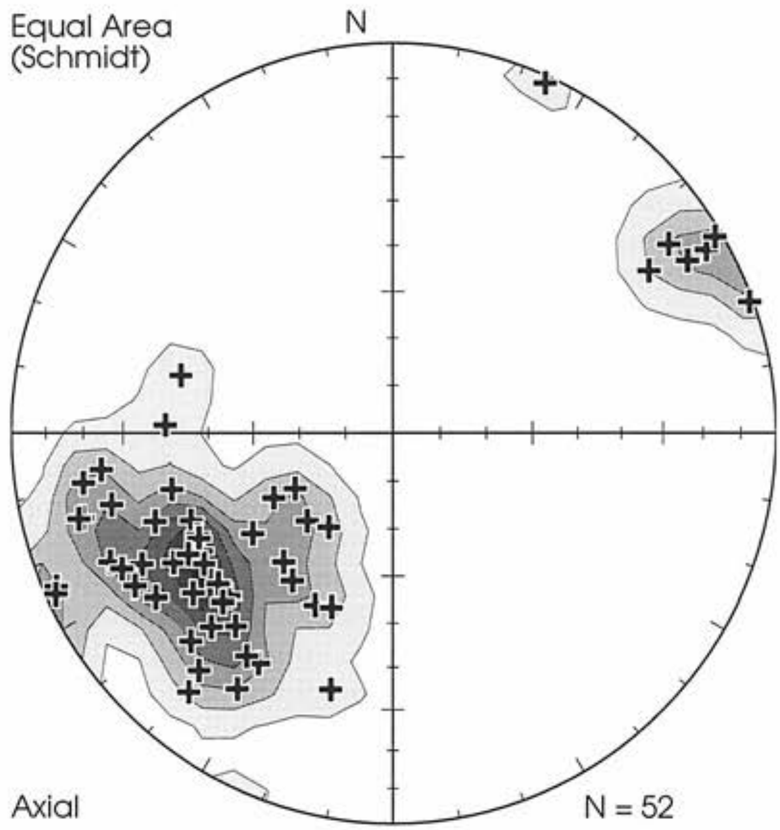

Figure 35. Lineations and density contours within shear zone. Moderately developed cluster pattern indicates rotation of lineation into the direction of the long axis of the shear-zone strain ellipsoid. Contours are at 2% intervals from 0-12%.

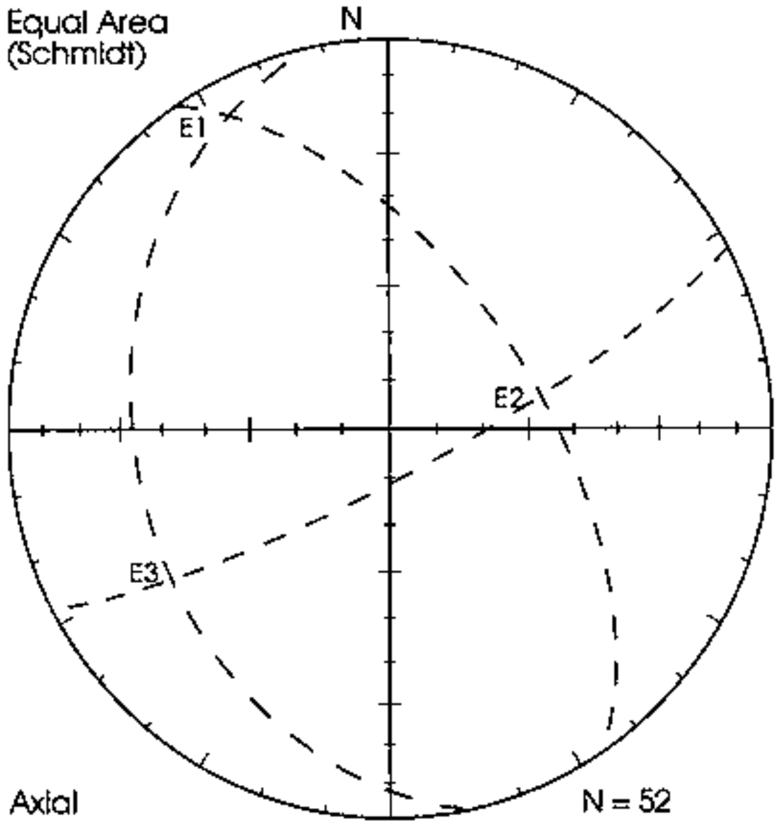

Figure 36. Eigenvectors and planes - shear zone lineations. E1 is normal to the best-fit plane through the data. The best-fit plane represents the average orientation of the plane which contains the lineations. E3 represents the average orientation of the lineations (32°/236°).

Table 7. Statistical Data for Shear Zone Lineation Orientations

Vector	Trend	Plunge	Value	Max/Min Angle (95% cone)
E1	335°	14°	3.0420	16.95°/5.26°
E2	86°	55°	6.8356	17.04°/7.83°
E3	236°	32°	42.122	8.05°/5.26°

The moderate cluster in orientation of these structures is thought to be the result of rotation toward the long axis of the strain ellipsoid during shear-zone deformation. This finding is consistent with the works of Hill (1987), Baker (1987), Cushing (1988) and Ketterer (1995), all of whom reported subhorizontal mineral lineations which indicate the displacement direction along the shear zone.

Structural Interpretation

Microstructural evidence suggests that the unit mapped as the chlorite-albite schist facies of the Wissahickon Formation is actually a phyllite in the northern exposures and a phyllonite to the south. The similarity in appearance of these two units is attributed to the most recent deformational event which both units experienced: the activation of the Cream Valley shear zone.

In the northern terrane, shearing was accompanied by metamorphic conditions which transformed Laurentian pelitic sediments to phyllite. Along the northern boundary the only signs of deformation occurred at the time of displacement along the shear zone. These features include grain-size reduction, bedding transposition and lineation rotation. Evidence of pre-shear-zone deformation, which is abundant in the southern regions of this structure, does not exist in the northern regions.

In the southern terrane, the units had been exposed to high-grade metamorphism prior to formation of the Cream Valley shear zone. Therefore, conditions which accompanied shearing caused retrograde metamorphism and grain-size reduction in the southern terrane.

The rocks in the southern exposures of the shear zone appear to have experienced a more complex deformational history than those in the northern exposures. Inclusion trails and truncated structures indicate that this unit experienced deformation prior to Cream Valley shear zone activation. Interlayering of this unit with high-grade rocks of the southern terrane indicates that the phyllonite in the southern portion of the shear zone is retrograded Wissahickon schist.

Deformation of feldspar porphyroclasts may have occurred with the onset of slip along the Cream Valley shear zone. The earliest stages of deformation along the shear zone could have occurred at lower amphibolite-facies conditions. These conditions would have been sufficient to produce crystal-lattice deformation in feldspar.

Parallelism of fabric orientations throughout the shear zone is attributable to foliation development and alignment during shear-zone deformation. In the southern region, pre-existing fabrics were overprinted by shear-zone deformation. Shear-zone overprinting explains the findings of this study that distinct deformational events occurred in the southern portion of the shear zone, despite the overall consistency of fabric orientations within the shear zone and throughout the rest of the field area. Although the relative effects of Cream Valley shear-zone deformation were different in each of the terranes, this deformational event created a similarity in appearance of the adjacent rocks on the northern and southern terrane.

Metamorphism

Within the center of the shear zone, extreme grain-size reduction makes mineral identification quite challenging. The only minerals that are positively identified are quartz, muscovite, chlorite and feldspars (**Figure 37**). The presence of biotite is questionable. Tourmaline, titanite and opaques are present as accessories.

Along the northern shear zone boundary, calcite occurs. Primary calcite is unique to the northern region of the shear zone. As with the Laurentian sequence, quartz and carbonates coexist as a stable assemblage in the northern portions of the shear zone (**Figure 38**). Peak metamorphism of phyllite in the northern shear zone occurred at greenschist-facies conditions. These are the same conditions at which late-stage Cream Valley shear-zone deformation occurred, indicating that the only metamorphism to occur north of the terrane boundary took place synchronously with shear-zone deformation.

Along the southern boundary the mineral assemblage consists of quartz, feldspar, chlorite, muscovite and biotite (**Figure 39**). The same accessory minerals occur here as in the center of the shear zone. Metamorphic porphyroclasts which are unique to southern exposures within the shear zone include oligoclase, pyrite and apatite (**Figure 40**). Epidote is also identified in this region (Cushing, 1988). Apatite and epidote serve as index minerals of upper greenschist or

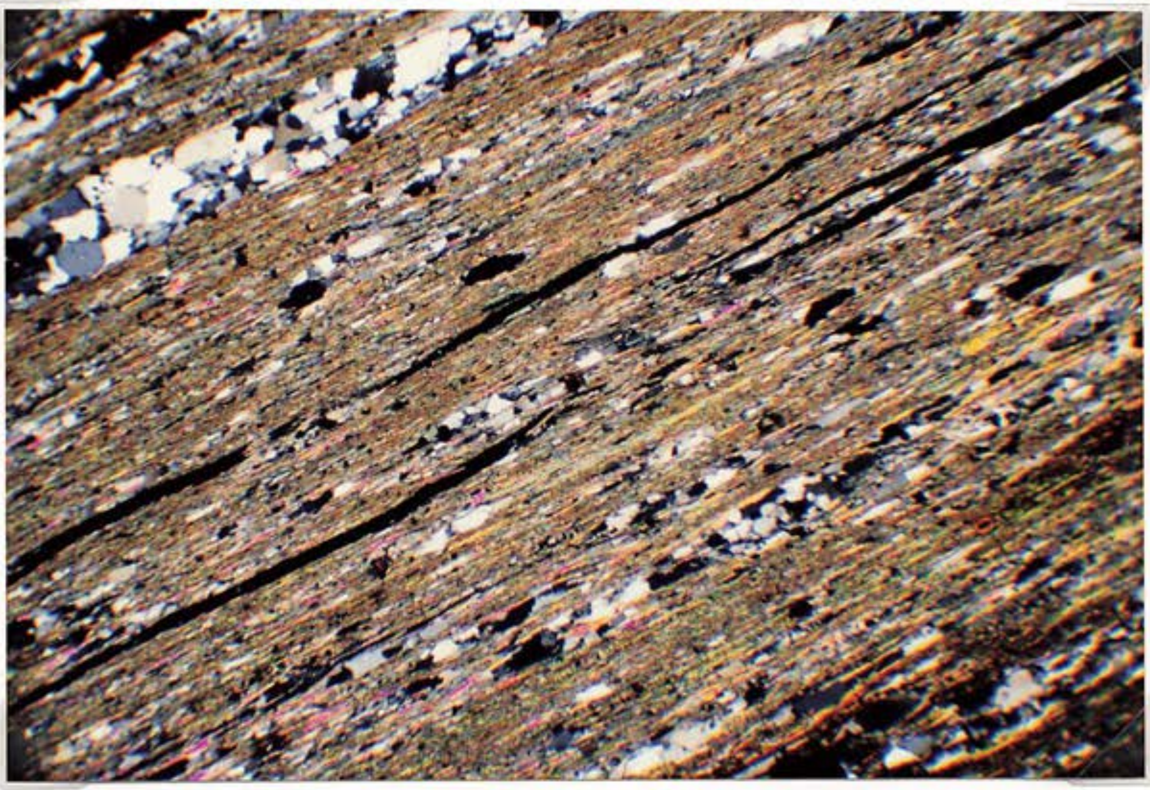

Figure 37. Shear zone mineral assemblage. Identifiable minerals in this assemblage include quartz, muscovite and feldspar as primary minerals and tourmaline, titanite and opaques as accessories. Sample 6, cross-polarized light, Field of view ≈ 4.2 mm x 3 mm.

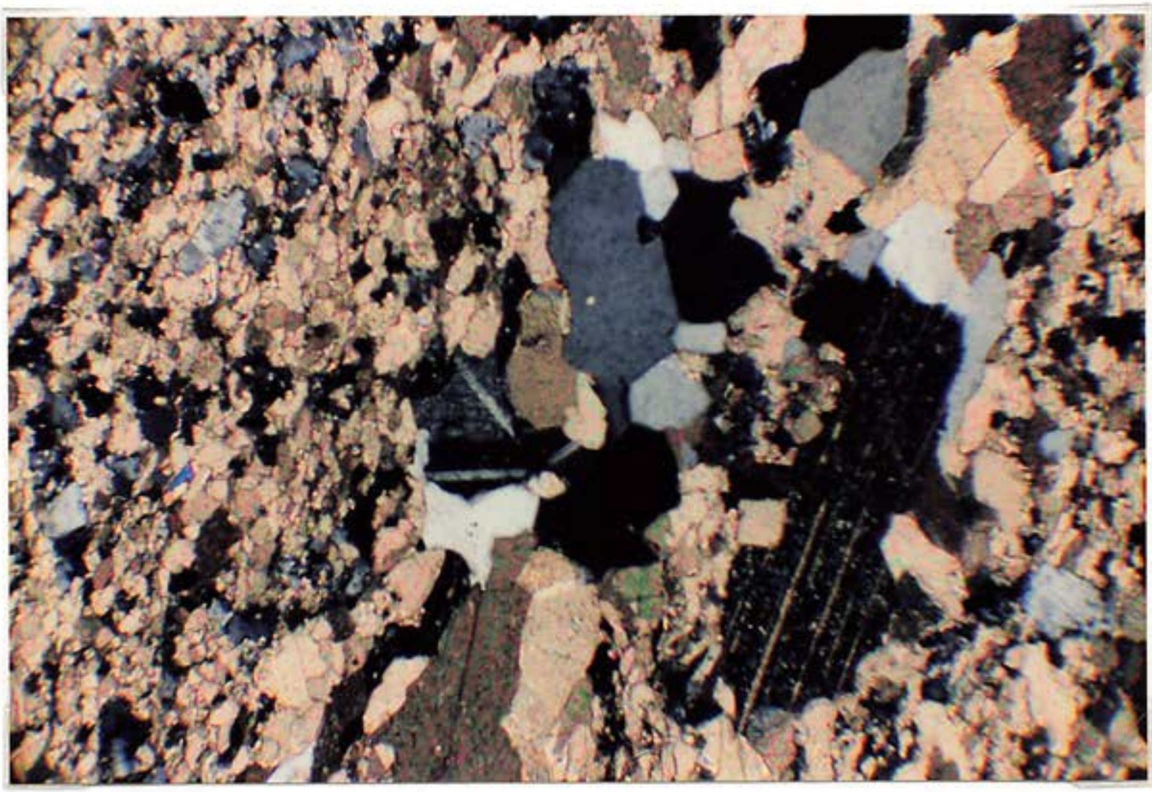

Figure 38. Quartz and calcite along northern shear zone boundary. There is no reaction along contacts between these minerals, indicating that this is a stable assemblage. This assemblage constrains the metamorphic conditions of this lithology to greenschist facies. Sample 9, cross-polarized light, Field of view ≈ 1.51 mm x .94 mm.

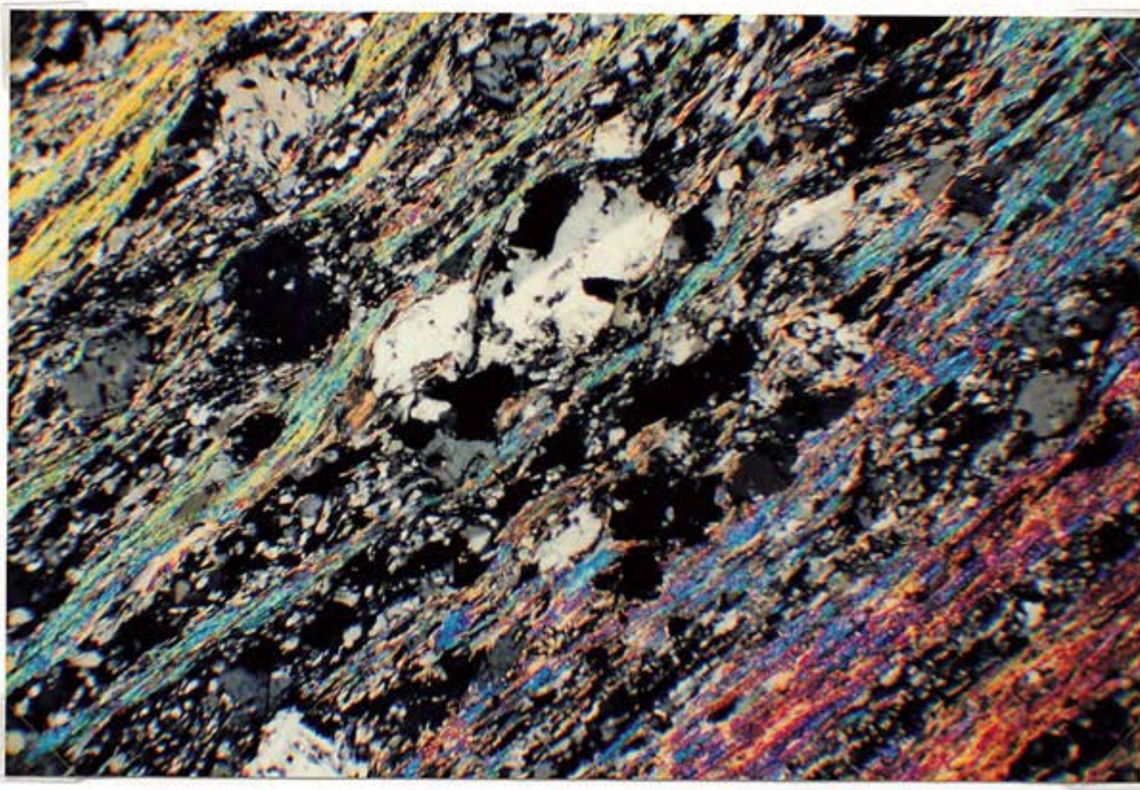

Figure 39. Phyllonite mineral assemblage, southern shear zone. The mineral assemblage here is similar to that in the center of the shear zone. Minerals present in this location are quartz, oligoclase, chlorite and muscovite. Sample 49, cross-polarized light, Field of view ≈ 4.2 mm x 3 mm.

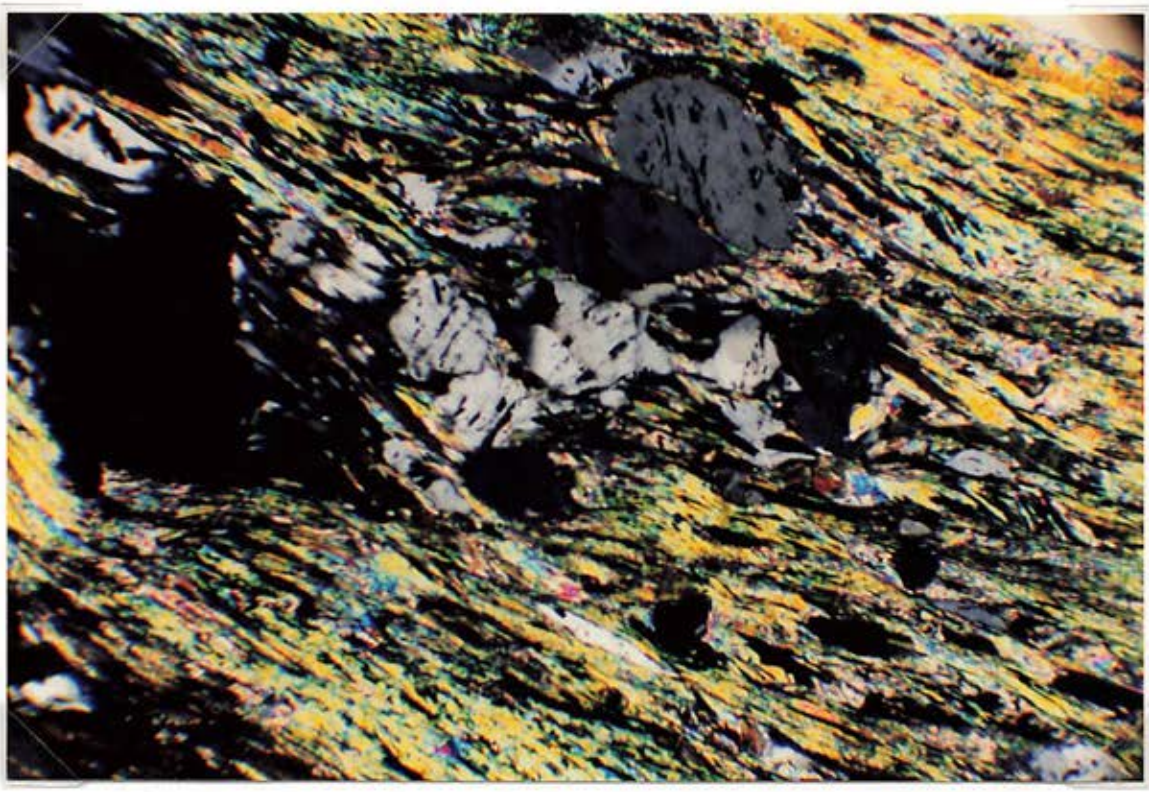

Figure 40. Porphyroclasts in phyllonite. The presence of porphyroclasts is unique to the southern region of the shear zone. Large grey porphyroclasts in the center are oligoclase with inclusion trails. Large black grain on left edge is pyrite. Two small black grains in lower right corner are apatite. Sample 47, cross-polarized light, Field of view ≈ 1.51 mm x .94 mm.

epidote-amphibolite facies conditions (Cushing, 1988). The preservation of large albite porphyroclasts reflects a prior coarse-grained metamorphic texture.

Epidote, apatite and oligoclase are distinctive to the southern phyllonite. These minerals establish the occurrence of at least one metamorphic event in the south for which there is no evidence to the north.

Plagioclase porphyroclasts record the first metamorphic event in the phyllonite. The deformation of these porphyroclasts indicated by strain of the crystal lattice, occurred next. The history established for the phyllonite in the southern portion of the shear zone resembles the metamorphic history of rocks in the southern terrane.

Conditions of deformation along the Cream Valley shear zone imposed a prograde metamorphism on the northern rocks and retrograde metamorphism on the southern rocks simultaneously. This evidence supports the hypothesis that rocks of the northern and southern portions of the shear zone have had distinct geologic histories.

Delineating the Terrane Boundary

Delineating the location of the terrane boundary within the shear zone is a complicated task in this field area. The strength of the shear-zone overprint on the phyllite and phyllonite within the structure effectively obliterated any distinguishing features of the protoliths for these rocks. The result is an apparently uniform lithology which even geochemical and petrographic analyses cannot distinguish. However, at the northern phyllite boundary and the southern phyllonite boundary clear distinctions exist. Unique features along these boundaries indicate that these are distinct units and that the contact between these units must therefore lie somewhere in between. The contact between the phyllite and phyllonite is the terrane boundary.

Although the exact location of the proposed terrane boundary is difficult to discern, this study suggests that the terrane boundary may lie near the southern shear zone boundary (**Figure 3**). The Cream Valley shear zone juxtaposes two sets of rocks that had experienced different histories, and as such reacted differently to the transpressional deformation that accompanied the interaction of the Laurentian plate margin and the allochthonous terrane.

Retrograde metamorphism in the southern terrane occurred simultaneously with prograde metamorphism in the hydrous Laurentian sequence during shear-zone deformation. Therefore, the Laurentian rocks to the north were more easily affected and shear-zone deformation was more far-reaching into these rocks than into the high-grade rocks to the south. This study suggests therefore that the terrane boundary is more likely to lie closer to the southern shear-zone boundary. This suggestion is consistent with field and petrographic observations.

Geologic History

The rocks in the southern portion of the shear zone formed in Precambrian time and have experienced a geologic history which is consistent with that of the rocks south of the shear zone. Metamorphic minerals such as epidote and oligoclase with inclusion trails support the interpretation of a pre-shear-zone metamorphic and deformational history for the southern rocks. At the time of shear-zone deformation, retrograde metamorphism and mylonitization occurred.

The combination of these effects produced a phyllonite in the southern portion of the shear zone.

Rocks in the northern portion of the shear zone belong to the Laurentian shelf sequence which formed during Cambrian and Ordovician time. These rocks experienced peak metamorphism and deformation at greenschist-facies conditions during late-stage shear-zone deformation. The primary evidence of deformation in this region is grain-size reduction and bedding transposition which accompanied shearing. Metamorphic porphyroclasts, indicative of an earlier metamorphism, are absent in this region, in contrast to the abundance of such evidence of pre-shear-zone metamorphism observed in the southern portion of the shear zone.

The Cream Valley shear zone became active during the Paleozoic. Activity along this structure probably began at amphibolite-facies conditions, based on deformation of feldspar which is believed to have occurred coincidentally with this event. Deformation along this system may have continued for 100 Ma or longer, based on the range of radiometric dates for biotite, which is believed to have crystallized during shear-zone deformation (Valentino *et al.*, 1994). Potassium-Argon ages on muscovite from the Susquehanna River region of this system place this event at 320 Ma (Lapham and Basset, 1964; Valentino *et al.*, 1994).

CHAPTER 5. DISCUSSION

Terrane Boundary Model

Terranes are generally defined as fault-bounded geologic units that possess an internally homogeneous geologic history which contrasts sharply with that of adjoining units (Coney, 1978; Jones *et al.*, 1983; Horton *et al.*, 1989). This model predicts that terrane boundaries will mark discontinuities that cannot be explained by metamorphic or depositional facies changes or unconformities (Coney *et al.*, 1980; Williams and Hatcher, 1983). The Cream Valley shear zone fits this description. This structure separates Laurentian passive-margin sediments to the north from Precambrian high-grade gneisses to the south. Sharp contrasts in lithology, deformation and metamorphism exist between these two shear-zone bounded suites. Additionally, this model accounts for the transcurrent shear zone at the contact between these disjunct units.

Terranes may be identified by contrasts in any or all of the following: stratigraphy, structure, metamorphic history, faunas, mineral deposits and paleomagnetic signatures (Coney, *et al.*, 1980; Williams and Hatcher, 1983). Genetic or tectonic implications are not necessary to recognize a terrane (Coney *et al.*, 1980) Lithologies lying on opposite sides of the Cream Valley shear zone show contrasts in at least three of these criteria: stratigraphy, structure and metamorphic history.

The geology of the field area agrees with the predictions of the terrane boundary model. Moreover, this model explains the geology of the field area more completely than previous models. Several prior models explained this structure as a thrust fault (Bascom, 1909; Armstrong, 1941; Wagner, 1966; Wagner, 1972). Such models fail to explain the vertical dip of the structural contact. Additionally, there is no correlation whatsoever between rocks of the "overthrust hanging wall" to the south and the "footwall rocks" to the north. Of particular note is the lack of correlation between the southern high-grade rocks, which were supposedly thrust upon the metasediments, and the Honey Brook Upland, which is the basement upon which the metasediments were deposited. The lack of correlation between basement rocks across the structure requires a terrane-boundary model.

Knopf and Jonas (1923) explained the change in metamorphic grade between the granulite/amphibolite facies rocks of the southern terrane and the greenschist-facies rocks of the Laurentian shelf as a metamorphic facies change. A gradual facies change fails to explain two important observations in the field area. First, the contact between the high and low-grade rocks is clearly structural. The Cream Valley shear zone is the structure which separates these distinct metamorphic assemblages. Second, greenschist and granulite-facies rocks are found one kilometer from each other in places, this distance is too small to be explained by a gradual facies change.

In addition to the surficial geologic evidence for a terrane boundary, geophysical evidence also supports the terrane boundary model. Bouguer gravity anomalies indicate that the Cream Valley shear zone is the surficial expression of a structure that offsets the Moho (Song, 1987; Duan, 1997).

The Cream Valley shear zone and surrounding geology support the terrane boundary model. This model explains all of the observations and data gathered in the course of this study. This model is compelling, not only because it fully explains the geology of the field area, but because it also accounts for aspects of the geology which previous models have failed to explain. Specifically, the vertical shear zone at the contact between the Laurentian margin and the southern terrane could not be explained by more traditional models which explained the contact as a thrust fault or metamorphic facies change. The terrane boundary model is a simple solution which explains more of the geology than any prior model.

CHAPTER 6. SUMMARY

Geologic Summary

Lithologies

Data collected for this study indicate that rocks lying on opposite sides of the shear zone experienced geologic histories which are distinct from one another. Rocks to the north of the shear zone are Cambro-Ordovician carbonates and quartzites, deposited on a basement of Precambrian crystalline rock. Rocks to the south of the shear zone are high grade metamorphic rocks of Precambrian age, uncorrelated with Precambrian basement to the north.

Structures

The presence of primary sedimentary structures within the metasediments north of the shear zone contrasts sharply with the inclusion trails and strong metamorphic fabric seen in rocks south of the shear zone. Rocks to the north experienced a less complex and lower temperature deformation than that to which the southern terrane was exposed. This contrast indicates separate deformational histories for rocks lying on opposite sides of the shear zone.

Metamorphism

Mineralogic evidence shows a sharp contrast in metamorphic grade of rocks on opposite sides of the shear zone. The Laurentian sequence, containing unreacted calcite and quartz is juxtaposed with the southern terrane, which contains minerals indicative of granulite and amphibolite facies metamorphism. Peak metamorphism of the autochthonous Laurentian shelf reached greenschist-facies conditions during the Paleozoic. Conversely, the southern terrane reached temperatures of up to 950° C during Precambrian time and temperatures between 650-700° C sometime during the Paleozoic (Wagner, 1966 and 1972). The disparity in metamorphic grade and timing of metamorphism of each package of rocks supports the idea that these suites have experienced distinct metamorphic histories.

Extent of Shear Zone

The Cream Valley shear zone is part of a shear zone system which is over 130 km long in Pennsylvania alone. The shear zone continues to the northeast as the Huntingdon Valley shear zone and southwest as the Pleasant Grove shear zone. Offset along this structure has been estimated as at least 150 km (Valentino, 1993). This structures ranges from 3-5 km in width within the field area, it narrows to the northeast (Ketterer, 1995) and may widen to the southwest (Valentino, 1993). This structure is vertical to near vertical and shows dextral strike-slip offset (Hill, 1987; Valentino, 1993; Solar, 1994; Ketterer, 1995).

Model Summary

The Cream Valley shear zone is consistent with the model which defines a terrane boundary as a structural contact between rocks of disparate lithologic, metamorphic and structural character. Lithologic, structural and metamorphic evidence in rocks around the Cream Valley shear zone supports the hypothesis that the Cream Valley shear zone marks a terrane boundary in southeastern Pennsylvania. This evidence indicates that the geologic history of rocks north of the shear zone differs radically from that of rocks south of the shear zone. Evidence in support of this interpretation is as follows:

Northern Suite

- Carbonate units and metapelites only occur north of the shear zone.
- Primary sedimentary structures such as laminites, crossbedding, *Skolithus* and *Monocraterion* are intact.
- Structural orientations support interpretation of a mild deformational history.
- Lack of tremolite or wollastonite in the carbonate units constrains the metamorphic temperatures to greenschist-facies conditions.

Southern Suite

- Granulite and amphibolite facies rocks occur only to the south of the shear zone (with the exception of the Precambrian Honey Brook Upland suite, which is not correlative with the southern gneisses).
- Structural orientations do not match the shear zone orientation, indicating a prior deformational history.
- Structural orientations are not consistent, indicating a complex deformational history.
- Mineral textures indicate deformational temperatures of greater than about 500°C.
- Mineral assemblages indicate metamorphism at granulite to amphibolite facies conditions.
- Metamorphic textures such as reaction rims indicate polyphase metamorphic history.

Shear Zone Structure

- The structural nature of the contact between the northern and southern suites is not explained by prior models.
- The vertical orientation of this structure is consistent with a transcurrent terrane boundary.
- This Cream Valley shear zone is the surface expression of a vertical structure which offsets the Moho.
- The identification of two distinct units within the shear zone helps in delineating the location of the terrane boundary within the shear zone.

Conclusion

In general, rocks in the northern terrane display many clues that metamorphism and deformation took place at low grades and were not geologically long-lived. Conversely, rocks in the southern terrane contain evidence of high-grade metamorphism and deformation and a long-lived geologic history. The juxtaposition of such radically different suites of rocks across a narrow zone, in addition to the geophysical and regional geologic evidence, suggests the presence of a major tectonic structure. A terrane boundary model fits the evidence more completely and more elegantly than previous models.

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