Geology and Structural History of the Blue Ridge Basement Complex, Albemarle County, Virginia

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GEOLGY AND STRUCTURAL HISTORY OF THE BLUE RIDGE BASEMENT COMPLEX, ALBEMARLE COUNTY, VIRGINIA

A thesis submitted in partial fulfillment of the requirement for the degree of Bachelors of Science in Geology from The College of William and Mary

by

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Accepted for High Honors
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Abstract

The Mesoproterozoic (1180-1030 Ma) basement complex in the central Virginia Blue Ridge consists of several mineralogically and texturally heterogeneous units. Near Crozet, Virginia, five compositionally distinct rock types occur and include: 1) layered gneiss, 2) charnockite and charnockitic gneiss, 3) biotite granitoid gneiss, 4) leucogranite, and 5) megacrystic granitoid. Two distinct foliations, including a high-temperature amphibolite to granulite facies and a low-temperature greenschist facies fabric, are variably developed in rocks with different deformation intensities. Several anastomosing northeast-southwest trending high-strain zones, including the connected Rockfish Valley and White Hall high-strain zones, cut through the basement complex. Lenses of relatively less deformed rock occur within the 1-3 km wide high-strain zones. Kinematic analysis indicates general shear with possible triclinic symmetry and apparent flattening strain. Palinspastic reconstruction of a cross-section to its undeformed state reveals 35% total shortening and less than 3 km of vertical displacement.
Introduction

Basement rocks record the tectonic evolution of continents and provide insight to processes that occur deep within the crust. Magmatism, metamorphism, and deformation associated with continental collisions create diverse assemblages of basement lithologies. Orogenic events and erosion expose mid-crustal rocks at the surface, allowing for study at regional to micro-scales. Examination of ancient mountain belts helps in understanding the processes occurring during modern mountain building events.

The crystalline basement of the Blue Ridge geologic province in Albemarle County, Virginia records over a billion years of geologic history in the Appalachian Mountains. Mesoproterozoic granitoids and gneisses provide evidence for multiple pulses of magmatism during the Grenville orogeny, collisional tectonics in the Mesoproterozoic and Paleozoic, and crustal extension in the Neoproterozoic and Mesozoic. Heterogeneous lithologic and structural relations indicate the complicated evolution of the basement complex. A variety of hypothesis for lithologic variability have been proposed, involving different magmatic episodes, major tectonic contacts, and retrograde metamorphism, but the relationships between contrasting basement units remain uncertain.

Much progress has been made toward the understanding of basement geology and Blue Ridge tectonics in recent years. Recent detailed mapping by students from the College of William & Mary provides the basis for this project to better understand the basement complex of Albemarle County (Gattuso and others, 2007). The study area focuses on the Crozet 7.5’ quadrangle of north central Virginia, where the basement core and western limb of the Blue Ridge anticlinorium are exposed (Plate 1). This study aims
to fully describe each of the basement units, determine the nature of basement contacts, and quantify the kinematics of basement mylonites in high-strain zones.
Geologic Setting

The Blue Ridge province extends from southern Pennsylvania to Georgia and separates the fold and thrust belt of the Valley and Ridge in the west from the suspect terranes of the Piedmont in the east. In central and northern Virginia, the Blue Ridge forms a northeast-striking anticlinorium over 300 km long and approximately 25 to 30 km wide (Fig. 1). During Paleozoic contractional deformation, the allochthonous Blue Ridge thrust sheet moved northwestward along a major tectonic ramp on top of younger Valley and Ridge sedimentary rocks in the footwall (Evans, 1989).

Crystalline granitoids and gneisses are exposed in the core of the northwest-verging anticlinorium and are flanked by metasedimentary and metavolcanic rocks in the limbs (Fig. 1). A significant unconformity exists between the Mesoproterozoic basement complex and overlying Neoproterozoic to early Paleozoic cover sequence (Fig. 2). On the western limb of the Blue Ridge, the variably thick (0-350 m) Swift Run and metabasaltic Catoctin formations nonconformably overlie the basement (Nelson, 1962). Neoproterozoic metasedimentary rocks, including the Swift Run and Mechum River formations, occur as fault-bounded structural inliers in contact with the basement core, while the thick (~5 km) Lynchburg Group overlies the eastern limb. The Catoctin Formation has been radiometrically dated to approximately 575 Ma and can be traced on both limbs and around the nose of the anticlinorium in Maryland and southern Pennsylvania, where it gently plunges into the subsurface (Badger and Sinha, 1989; Aleinikoff and others, 1995). The early Cambrian Chilhowee and Evington groups conformably overlie the Catoctin in the western and eastern limbs respectively.
Figure 1: Regional geologic map encompassing the Charlottesville 1:100,000 quadrangle (38º to 38º 30’ N, 77º to 78º W). Mesoproterozoic basement rocks (Y) form the core of the Blue Ridge anticlinorium, flanked on both limbs by Neoproterozoic metasediments and metavolcanics. Inset shows location in Virginia. C = Crozet 7.5’ quadrangle, the study area.
Figure 2: Schematic diagram illustrating intrusive relationships between basement units. All other units intrude Layered gneiss (Ygn), the oldest unit in the study area. Biotite granitoid gneiss (Ybg) surrounds and divides pods of charnockite and charnockitic gneiss (Ycg) in the southeast. Dikes of Neoproterozoic (Zd and Zhg) as well as Jurassic (Jd) age cut through the older units. Neoproterozoic metasedimentary and metavolcanic rocks of the (Zcs, undivided) unconformably overlie Mesoproterozoic basement.
Mesozoic basins occupy structural depressions in the eastern margin of the Blue Ridge province and diabase dikes cut the basement complex and cover sequence.

The cover sequence of the Blue Ridge records the transition from rifting of the supercontinent Rodinia and creation of the Iapetus Ocean in the Neoproterozoic through the development of the Laurentian passive margin in the early Paleozoic. The terrestrial sediments of the Swift Run Formation and the flood basalts of the Catoctin Formation formed during Iapetan rifting, while the marine siliciclastics of the Chilhowee and Evington group and the overlying Cambro-Ordovician carbonates were deposited on the passive margin. Paleozoic deformation occurred as a result of collisional tectonics and the formation of Pangea, creating the Appalachian Mountains during a series of orogenies including the Taconic, Acadian, and Alleghanian. Mesozoic rifting caused Pangea to split apart, injection of diabase dikes into the basement and cover sequence, and formation of basins along the eastern margin of the Blue Ridge.
Previous Work

Regional Geology

Geologists in Virginia have long recognized the heterogeneity of the Mesoproterozoic basement complex. Jonas (1928, 1935) distinguished the biotite-bearing Lovingston and Marshall formations from hypersthene-bearing granodiorites. Bloomer and Werner (1955) named the predominantly granodioritic rocks the Pedlar Formation. Nelson (1962) mapped the geology of Albemarle County, distinguishing between the Lovingston Gneiss Formation and the Virginia Blue Ridge complex. Lovingston formation consists of coarse-grained quartz monzonite, although elliptical areas of igneous rocks with different compositions have been injected into Lovingston gneiss (Nelson, 1962). The Virginia Blue Ridge complex includes granodiorite, hypersthene granodiorite, and the Marshall and Crozet granites.

Gathright and others (1977) recognized a zone of cataclastic rocks more than 5 km wide in the adjacent Waynesboro East 7.5’ quadrangle, and correlated it to the Rockfish Valley fault of Bartholomew (1977) separating the Lovingston and Pedlar formations. Bartholomew and others (1981) renamed this feature the Rockfish Valley ductile deformation zone and developed a tectonic model involving the juxtaposition of two basement massifs along a major northwest-directed thrust in the Paleozoic. This model requires significant movement of the allochthonus lower-grade Lovingston massif in the hanging wall over the parautochthonus higher-grade Pedlar massif in the foot wall.

Alternatively, Evans (1984, 1991) proposed a model in which the contrast between the Lovingston and Pedlar terranes resulted from rift related metamorphism and crustal extension during the Neoproterozoic, without significant differences in crustal
depth between the two terranes. According to this hypothesis, the Lovingston terrane, overlain by a much thicker sequence of hydrous metasediments, experienced hydration metamorphism with crustal heating from the intrusion of Neoproterozoic plutons. Meanwhile, the Pedlar terrane, overlain by a much thinner sedimentary sequence and further west toward the craton interior, remained largely unchanged.

Bailey and Simpson (1993) provide evidence for multiple stages of deformation in basement high-strain zones, including extensional and contractional movement. The older, extension related event involved top-to-the-southeast movement and occurred under amphibolite facies conditions with temperatures greater than 450 °C. The younger top-to-the-northwest thrust related deformation occurred under greenschist facies. They also suggest that extensional fabrics coincide with the emplacement of Neoproterozoic plutons to the north of the study area (Bailey and Simpson, 1993).

Burton and Southworth (2004) point out problems with both the Bartholomew and Evans hypotheses. They cite a lack of evidence for retrograde metamorphism caused by a thermal peak in the Neoproterozoic and suggest that lithologic differences cannot be explained by Paleozoic movement alone. Instead, they argue that the difference between the terranes results from a Grenville age discontinuity that was later offset by Neoproterozoic normal faulting (Burton and Southworth, 2004).

Recent geochronology studies indicate three groups of intrusive events related to the Grenville orogeny: Group 1 at 1180-1160 Ma, Group 2 at 1120-1080 Ma, and group 3 at 1080-1030 Ma. (Aleinikoff and others, 2000, 2005, 2008; Burton and Southworth, 2004; Tollo and others, 2004a, 2004b, 2006). In addition to the pluton emplacement ages
for orthogneisses, detrital zircon ages for paragneiss units yield 1260-1020 Ma, indicative of sedimentation during Grenvillian orogenesis (Southworth and others, 2008a).

Two major zones of high-strain cut through the study area (Fig. 3). The anastomosing Rockfish Valley high-strain zone strikes generally northeast-southwest and intersects the north-south striking White Hall high-strain zone. Previous workers have noted top-to-the-northwest sense of shear from meso- and micro-scale kinematic indicators (Gathright and others, 1977; Bailey and Simpson, 1993; Bailey, 1995). However, the amount of displacement across these zones is debated (Bartholomew, 1981; Evans, 1991; Berquist, 2000). Quantitative estimates of the kinematics of high-strain will elucidate basement contacts and verify proposed models.

**Kinematic Analysis**

The deformation of naturally occurring rocks in high-strain zones has been the subject of recent study. High-strain zone kinematics can be resolved into three main components: finite strain, volume change, and vorticity. Finite strain is a three-dimensional measure of the geometric change in shape resulting from stresses applied to materials experiencing ductile deformation. Volume change results from the removal of material through dissolution by hydrothermal fluids. Kinematic vorticity ($W_k$) indicates the amount of rotation associated with shearing.

Finite strain of occurs along three mutually perpendicular axes that define an ellipsoid (Fig. 4). The $X$, $Y$, and $Z$-axes correspond to the maximum, intermediate, and minimum finite strain axes, respectively. Deformation of natural rocks results in alignment of minerals or aggregates of minerals such a phyllosilicates, with their long
Figure 3: Generalized geology and sample location map of the study area. Units include layered gneiss (Ygn), charnockite and charnockitic gneiss (Ycg), leucogranite (Ylg), megacrystic granitoid (Ymg), biotite granitoid gneiss (Ybg), and undifferentiated Neoproterozoic rocks (Zcs). Contacts are thin lines, faults are thicker black lines, and high-strain zone boundaries are dashed red lines. A cross section reference line from X to X’ is also shown.
Figure 4: Schematic diagram of three-dimensional strain. (a) Strain ellipsoid showing the orientation of long X-axis, the intermediate Y-axis, and the short Z-axis. (b) Box diagram showing the strain ellipse on each plane as a cube is deformed to a rectangular prism with monoclinic symmetry. (c) Diagram of a naturally deformed mylonite with foliation developed in the XY plane and a mineral elongation lineation parallel to the X-axis. Notched rectangles show position of oriented thin sections.
axes parallel to the $X$-axis defining a foliation in the $XY$ plane. Mineral elongation lineations develop parallel to the $X$-axis.

Kinematic vorticity defines the amount of rotation associated with deformation relative to stretching and indicates the type of shear. Pure shear ($W_k=0$) occurs with no rotation, only change in length of the principal axes. Simple shear ($W_k=1$) occurs only with rotation, and no change in shape of the principal axes. Intermediate values of vorticity ($0<W_k<1$) give rise to general shear, a combination of rotation and change in shape of the principal axes.

The geometric relationship between fabric elements, high-strain zone boundaries, and finite strain depend upon the type of symmetry (Simpson and De Paor, 1993). Orthorhombic symmetry may result from pure shear deformation, and the foliation plane develops parallel to the high-strain zone boundary. Simple and general shear may cause monoclinic deformation, in which the foliation develops at an angle ($\theta$) oblique to high-strain zone boundaries. This angle decreases at higher values of finite strain. In triclinic shear, the $X$-axis of the finite strain ellipsoid does not parallel the mineral elongation lineation. Triclinic shear represents continuously superimposed instantaneous monoclinic deformations, so that the mineral elongation lineation only records the most recent instantaneous strain, rather than the accumulation of all of them (Forte and Bailey, 2007).
Methods

This study integrates both field and laboratory methods, including geologic mapping, field observations, structural measurements, petrographic analysis and strain analysis. Geologic mapping of the Crozet 7.5’ quadrangle at a scale of 1:24000 provided the initial framework for this study, delineating the spatial relationships between lithologically distinct basement units and the extent of high-strain zones (Fig. 3, Plate 1). Field data such as rock type, rock structures and deformation intensity were observed while mapping. Rock types were distinguished by mineralogy, grain size, texture, color and weathering. Structural measurements include the orientation of foliation and lineation, as well as shear-sense indicators.

Deformation intensity was qualitatively estimated based on the degree of mylonitization, according to a modified version of the classification scheme developed by Olney and others (2007). Rocks were assigned a deformation intensity number on a scale ranging from 0 to 5 based on the following criteria: (0) massive rock, no visible foliation, (1) weakly deformed rock, difficult to measure foliation, (2) moderately deformed rock, easy to measure foliation, (3) protomylonite with visible porphyroclasts and well developed foliation (4) mylonite with significant grain size reduction, abundant porphyroclasts and strong foliation, (5) ultramylonite with intense grain size reduction, few porphyroclasts and laminar foliation. Plotting deformation intensity data on a map allowed for delineation of high-strain zone boundaries (Fig. 5).

Several samples of each rock type were collected, including oriented samples of varying deformation intensities. A single thin section chip was prepared from samples with isotropic fabrics, while two perpendicular thin section chips were prepared
Deformation Intensity

- 0. massive - no foliation
- 1. very weakly foliated
- 2. weakly to moderately developed foliation, easy to measure
- 3. moderately developed foliation, protomylonite
- 4. strong foliation, with abundant porphyroclasts, mylonite
- 5. strong foliation, matrix-dominated, mylonite to ultramylonite

Figure 5: Deformation intensity map of the Crozet 7.5' quadrangle delineating the Rockfish Valley and White Hall high-strain zone boundaries in red. Colors correspond to a qualitative measure of ductile deformation, with warmer colors indicating higher strain and cooler colors indicating lower strain. Deformation is heterogeneous throughout the anastomosing high-strain zones.
from foliated rocks. Both sections were cut perpendicular to foliation, with one section parallel to lineation (XZ plane), and one section normal to lineation (YZ plane). Thin sections chips were notched on one side, and the position of the notch was indicated on the hand sample. Sample orientation marks in the field allowed the strike and dip of each thin section plane to be determined in the lab.

Thin sections were analyzed using a petrographic microscope to determine the mineralogy and texture of each sample. Minerals were listed in order of decreasing abundance and characteristics such as grain shape, crystal texture, alteration, and habit were noted. The rock texture was also examined, including segregation banding, microstructures, porphyroclasts, recrystallization and deformation.

Strain analysis of quartz grains involved the use of the $R/\phi$ method. The quartz grains in each thin section, including both the XZ and YZ planes, were traced by hand and the ratio of the long axis relative to the short axis of each grain was calculated. The angle between the long axis and the foliation plane, $\phi$, was measured for samples with low deformation intensities. In mylonitic samples, all quartz grains were aligned nearly parallel to the foliation plane, so measuring the $\phi$ angle proved unnecessary.

Approximately 50-100 grains per section were measured, and the median value was taken to represent the average strain ratio. From this ratio, an average value for each deformation number was calculated (Fig. 6). Also, three-dimensional strain was determined by combining data from the XZ and YZ planes, assuming a stretch of 1 for the Y-axis, and solving for $X$ and $Z$. These data were then plotted on a logarithmic Flinn diagram to indicate the three-dimensional strain geometry (Fig. 7).
Figure 6: The correlation between deformation intensity and averages of measured strain ratios. Strain ellipses represent the XZ plane and colors correspond to deformation intensity.

<table>
<thead>
<tr>
<th>DI</th>
<th>$R_s$</th>
<th>Symbol</th>
</tr>
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<tbody>
<tr>
<td>0</td>
<td>1.0</td>
<td>![Blue Ellipse]</td>
</tr>
<tr>
<td>1</td>
<td>1.4</td>
<td>![Blue Ellipse]</td>
</tr>
<tr>
<td>2</td>
<td>2.0</td>
<td>![Green Ellipse]</td>
</tr>
<tr>
<td>3</td>
<td>2.7</td>
<td>![Yellow Ellipse]</td>
</tr>
<tr>
<td>4</td>
<td>5.2</td>
<td>![Orange Ellipse]</td>
</tr>
<tr>
<td>5</td>
<td>11.0</td>
<td>![Red Ellipse]</td>
</tr>
</tbody>
</table>
Figure 7: Logarithmic Flinn diagram for samples of this study denoted by Xs combined with samples from Bailey (1995) denoted by circles. All samples plot in the field of apparent flattening, indicating true tectonic flattening, volume change, or a combination of both.
After correlating deformation intensity with measured strain ratios, strain ellipses were plotted on a cross section (Fig. 8). Foliation measurements were projected from the surface into the subsurface. Palinspastic reconstruction of the cross section involved transforming each ellipse back into a circle using a combination of simple shear and pure shear resulting in general shear (Fig. 8c). Reference points on the modern topographic surface were used to measure the total displacement between the restored and deformed state cross sections.
Figure 8: Three cross sections from X to X’ on the 1:24000 scale map. (a) Top cross section shows bedrock geology, including hypothesized basement unit and cover sequence contact geometry in the subsurface. (b) Middle cross section shows the current, deformed state geometry along with observed deformation intensities projected into the subsurface along strike of foliation. (c) Bottom cross section shows modern topography prior to Paleozoic deformation. Note that all ellipses from the above cross section have been restored to unit circles. Magenta labels serve as reference points.
Description of Basement Units

The heterogeneous basement complex consists of five map-scale units (Fig. 3) distinguished by mineralogy, grain size, texture, color and weathering. Intrusive contacts, where observed, help delineate the age relations between different units (Fig. 2). Petrographic analysis accompanied with field observations are reported for each basement unit.

Layered gneiss (Ygn)

The oldest mapped unit in the study area, layered gneiss, consists of orthogneiss and paragneiss (Fig. 9). Orthogneiss displays variably thick 5 mm to 20 cm bands that segregate quartz and feldspar from biotite, chlorite and opaque minerals like ilmenite. Fabrics include both high-temperature foliation and low-temperature foliation. High-temperature foliation, where present, is delineated by amphibolite to granulite facies minerals. Low-temperature foliation is pervasive throughout the study area, and is especially well developed in high-strain zones.

Paragneiss crops out in the Old Trail housing development approximately 1 km south of Crozet and contains both quartzitic and garnetiferous lithologies. The garnetiferous paragneiss contains up to 35% garnet, as well as quartz, potassium feldspar, plagioclase, white mica, and biotite. The quartzitic paragneiss consists almost entirely of quartz with some interstitial white mica, and likely has a quartz arenite protolith. Detrital zircon geochronology of paragneiss elsewhere in the Blue Ridge yields an age range of 1260-1020 Ma, which overlaps with and postdates ages for Mesoproterozoic orthogneiss
Figure 9: (a) Photomicrograph of layered orthogneiss (Ygn) with polars crossed. Note distinct segregation banding between quartz and feldspar and biotite, muscovite and epidote which represents a relict high-temperature fabric. (b) Photomicrograph of garnetiferous paragneiss (Ygn) in plane-polarized light (PPL). This rock contains 25-30% garnet.
(Southworth and others, 2008a). The sedimentary protoliths of the paragneiss were probably deposited unconformably on the basement complex.

**Charnockite and charnockitic gneiss (Ycg)**

Charnockite refers to a group of igneous or metamorphic rocks containing orthopyroxene, quartz, and perthite (Le Maitre, 1989). This unit has also been referred to in the literature as hypersthene granodiorite, granodiorite, and the Pedlar Formation (Jonas, 1928, 1935; Bloomer and Werner 1955; Gathright and others, 1977; Bartholomew and others 1981). Charnockite varies from coarse-grained and massive to well-foliated and porphyroclastic (Fig. 10). In the vicinity of Mt. Moriah church, the high-temperature fabric is defined by 2 to 5 mm bands of hypersthene and ilmenite between quartz and feldspar. Recent U-Pb zircon geochronology of this unit from the Free Union 7.5’ quadrangle yields an emplacement age of 1177 ± 11 Ma (Southworth, 2008b).

Massive to lightly deformed charnockite spheroidally weathers to rounded boulders with thick weathering rinds of light orange to white feldspars. Fresh surfaces show blue and green feldspars with a waxy appearance. Charnockitic gneiss predominates in the northwestern portion of the study area while multiple smaller bodies of charnockite crop out on and around Taylors Mountain southeast of the Rockfish Valley high-strain zone and transitions into biotite granitoid gneiss across narrow (50-100 m) zones.
Figure 10: (a) Photomicrograph of charnockite (Ycg) from Taylors Mt. with polars crossed (XPL). Orthopyroxene (Opx) is hypersthene and K-feldspar is a microperthite. (b) Field photograph of weakly deformed (DI = 1) leuchocharnockitic gneiss (Ycg). Foliation is too poorly developed to measure accurately.
Figure 11: (a) Photomicrograph of leucogranite (Ylg) with crossed polars.  Quartz and feldspar compose over 95% of this rock.  (b) Outcrop photograph of a leucogranite (Ylg) boudin in protomylonitic biotite granitoid gneiss (Ybg).
Leucogranite (Ylg)

Leucogranite is distinguished based on its very low mafic mineral content (<5%) and pale white color in fresh and weathered surfaces (Fig. 11). Leucogranite is typically medium-grained to coarse-grained with white potassium feldspar and clear, smoky, or blue quartz. Trace amounts of biotite, chlorite, epidote, and ilmenite occur. Leucogranite dikes intrude charnockite on Taylors Mountain and on Broad Axe Road, indicating the leucogranite crystallized later. Where leucogranite dikes are present in high-strain zones, they form boudins of more competent, less deformed rock (Fig. 11b). Foliations are less well defined in this unit because of the lower mafic mineral content.

Megacrystic granitoid (Ymg)

Formerly named the Crozet granite (Nelson, 1962), megacrystic granitoid contains very coarse-grained feldspars and quartz with variable proportions of biotite, uralite and pyroxene (Fig. 12). Feldspars weather white to light grey, and quartz is usually smoky to dark grey. Pyroxene includes orthopyroxene, such as hypersthene, and clinopyroxene, such as augite, which is altered in some places to uralite and biotite. An abundance of ilmenite and deep red colored biotite grains suggest high titanium content. Fabric is usually massive to weakly deformed, although significant grain size reduction occurs in the White Hall high-strain zone resulting in well-foliated mylonites. Near the basement-cover contact, megacrystic granitoid shows extensive alteration to greenschist facies, sometimes forming unakite with pink potassium feldspar and green chlorite. Recent U-Pb zircon geochronology from the Browns Cove 7.5’ quadrangle yields an
Figure 12: (a) Photomicrograph of megacrystic granitoid (Ymg) with polars crossed. Clinopyroxene (Cpx) is augite and K-feldspar is a microperthite. (b) Field photograph of megacrystic granitoid (Ymg) exposed one kilometer southwest of White Hall. Feldspars are white and exceed 5 cm in size, while quartz is smoky grey.
emplacement age of 1058 ± 6 Ma for megacrystic granitoid just north of the study area (Southworth and others, 2008b).

*Biotite granitoid gneiss (Ybg)*

Biotite granitoid gneiss is the most prevalent rock type in the study area (Fig. 3). Mineralogy includes quartz, potassium feldspar, plagioclase, and 5-25% mafic minerals including biotite, chlorite, epidote, and ilmenite (Fig. 13). Fabric ranges from massive to mylonitic, with aligned biotite and chlorite defining the greenschist facies, low-temperature foliation. Grain size ranges from medium- to coarse-grained and textures include equigranular, porphyritic and megacrystic.

In the vicinity of Taylors Mountain, transitional mineral assemblages from pyroxene-bearing charnockite to biotite-bearing granite occur. Hypersthene in charnockite is rimmed by fine-grained fibrous amphibole (uralite), which in turn is surrounded by biotite. Orthopyroxene, clinopyroxene, uralite, and biotite occur together in transitional rock.
Figure 13: (a) Photomicrograph of biotite granitoid gneiss (Ybg). Some feldspars have been sausseritized and replaced by sericite and quartz. (b) Field photograph of protomylonitic (DI=3) biotite granitoid gneiss (Ybg), with feldspar porphyroclasts and ribbons of blue quartz.
Structural Analysis

Basement rocks are variably deformed throughout the study area, but frequently display a measurable foliation. Both a high-temperature foliation defined by amphibolite to granulite facies minerals, and a low-temperature foliation defined by greenschist facies minerals occur. High-temperature foliation appears in the oldest basement units, charnockitic gneiss and layered gneiss, and is sometimes overprinted by the younger low-temperature foliation. High-temperature foliations do not retain a consistent orientation throughout the study area (Fig. 14a).

Low-temperature foliation is defined by greenschist facies minerals including chlorite, biotite, muscovite, and aligned quartz grains. Foliation strikes northeast-southwest and dips moderately to the southeast (Fig. 14b-d). Where present, an associated mineral elongation lineation plunges moderately down-dip to the southeast (Fig. 14e).

Rocks in the study area show a complete transition from massive, or unfoliated, to mylonitic and ultramylonitic. Deformation is heterogeneous across the study area and is concentrated in high-strain zones. In general, deformation intensity increases to the southeast of the study area (Fig. 5). Charnockite and megacrystic granitoid are generally massive to slightly deformed. Biotite granitoid gneiss and layered gneiss are often moderately to well foliated, but range from slightly deformed to ultramylonitic. Leucogranite does not show strong foliation, but does occur as foliated layers and boudinaged dikes in other units.

Strain analysis of quartz grains in the XZ-plane indicates the maximum finite strain ellipse for variably deformed rocks (Fig. 6). The finite strain ratio may be
Figure 14: Stereograms showing the orientation of structural elements in the study area. (a) Layered gneiss (Ygn) and charnockitic gneiss (Ycf) display randomly oriented Grenville-age amphibolite-facies high-temperature fabric. (b-d) Greenschist-facies minerals define the low-temperature foliation present in all Proterozoic units. The Paleozoic greenschist-facies fabric commonly dips to the southeast, and is relatively consistent across the high-strain zones. (e) Mineral elongation lineations are approximately down-dip of foliation measurements.
correlated with deformation intensity. Strain ratios in the $XZ$-plane are moderate to high ($R_s > 5$) and strain ratios in the $YZ$-plane are moderate in mylonitic samples. Plotting these ratios on a Flinn diagram shows that all samples fall within the realm of apparent flattening (Fig. 7). These data are consistent with Bailey (1995) and may indicate volume loss as much as 50%.

Asymmetries of sigma-tailed porphyroclasts as well as cross-cutting shear bands display top-to-the-northwest movement, although opposing shear-sense indicators occur (Fig. 15). In the lineation normal sections, a few asymmetries are found, suggesting that the strain includes at least some component of triclinic symmetry. Accounting for the strain throughout the study region leads to the creation of the restored cross section (Figure 8). Overall vertical displacement is approximately 1 km within the high-strain zone with 60% horizontal shortening, compared to 35% horizontal shortening and less than 3 km of vertical displacement across the entire study area.
Figure 15: Photomicrographs of oriented biotite granitoid gneiss (Ybg) samples under crossed polars, showing a progression from (a) protomylonite (DI = 3), (b) mylonite (DI = 4) and (c) ultramylonite (DI = 5). Note the decrease in grain size and porphyroclast abundance. All sections show the XZ plane and asymmetries are top-to-the-northwest where present.
Discussion

Map patterns, field relations, petrography and kinematic analysis allow for the evaluation of existing models relating contrasting basement units. Field relations and map patterns provide evidence for distinct magmatic events, concurring with the results of recent geochronology (Aleinikoff and others, 2000, 2008; Burton and Southworth, 2004; Tollo and others, 2008). Layered gneiss and charnockitic gneiss likely formed during the group 1 pulse of magmatism (1180-1140 Ma) and display a high-temperature foliation developed earlier (1078-1064 Ma) during the Grenville orogeny (Tollo and others, 2004a).

Megacrystic granitoid intruded these older gneisses during a later pulse of Grenville-age magmatism, likely group 3 (~1050 Ma). Leucogranite also intrudes these older units as dikes, and forms boudinage in high-strain zones. The variety of scales at which leucogranite occurs makes it difficult to define as a unit, and may in fact include multiple leucogranite bodies that represent late stage crystallization of residual melts. Regardless, the map-scale body of leucogranite does not exhibit the high-temperature foliation, suggesting its emplacement postdates that of charnockite gneiss and layered gneiss.

A 50-100 m wide transitional contact exists between biotite-bearing granitoid and charnockite, especially in the vicinity of Taylors Mountain (Plate 1). A full range of hydrous metamorphic mineral assemblages are present, including orthopyroxene, clinopyroxene, uralitic hornblende, and biotite. These observations are consistent with the following reactions from Evans (1991):

\[
\text{qtz} + \text{plag} + \text{K-spar} + \text{opx} + \text{cpx} + \text{mgt} + \text{ilm} + \text{O}_2 + \text{H}_2\text{O}
\]
\[
\begin{align*}
= & \text{qtz + albite + K-spar + biotite + amph + musc + ep + mgt ± ilm +O}_2 + \text{H}_2\text{O} \\
= & \text{qtz + albite + K-spar + biotite + musc + ep + titanite}
\end{align*}
\]
In addition, biotite granitoid gneiss typically shows stronger deformation than charnockite, with transitional rock showing intermediate deformation. This suggests that greenschist metamorphism was synchronous with the development of the low-temperature fabric. Field data is consistent with the Evans hypothesis for this one particular contact, but many questions remain. For example, while the lithology of the charnockite southeast of the high-strain zone closely resembles that of the charnockitic gneiss to the north, the two bodies may not be genetically related. Geochronology of the Taylors Mountain charnockite and surrounding biotite granitoid gneiss may reveal the true nature of the transitional contact.

Strain estimates of mylonites from the Rockfish Valley and White Hall high-strain zones yield moderate values for vertical displacement with approximately 1 km of throw across the zone and less than 3 km of throw for the entire study area. Meanwhile, these rocks accommodated approximately 35% shortening at the regional scale and 60% across the high-strain zone. While these data provide minimum estimates, the total displacement is not enough to separate rocks of significantly different crustal depths. The Bartholomew and others (1981) hypothesis involving considerable displacement in order to juxtapose rocks of different crustal levels across the Rockfish Valley high-strain zone should therefore be reevaluated.

Kinematic analysis also reveals the a component of triclinic shear indicated by asymmetries in \(YZ\) plane in addition to monoclinic shear found elsewhere along the high-strain zone. This implies that previous, two-dimensional models for deformation across
high-strain zones do not account for the more complicated three-dimensional strain geometries observed in natural rocks. Also, the presence of triclinic shear implies the orientation of the shear direction changed as deformation progressed.
Conclusion

The Mesoproterozoic basement complex of the Blue Ridge experienced protracted events of magmatism, metamorphism and deformation. At least two pulses of Grenville-age magmatism formed the protoliths of the rocks in the study area. Deformation occurred during the Grenville, resulting in high-temperature foliations in the older basement units, followed by Paleozoic deformation at greenschist facies and development of a pervasive low-temperature foliation. Greenschist facies deformation was heterogeneously distributed at the micro-scale to map-scale, creating localized zones of high-strain. Strain analysis shows that basement rocks experienced approximately 35% total shortening and less than 3 km of vertical displacement, while rocks in the high-strain zone experienced approximately 60% shortening and less than 1 km of vertical displacement. These results suggest that the Rockfish Valley and White Hall high-strain zones did not accommodate enough displacement to juxtapose different grade metamorphic rocks.
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Appendix: Strain Data

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Note: Bold data is from this study and is combined with data from Bailey (1995).