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Dating Deformation: Thermochronology and Chemical and Mineralogical Changes in the Blue Ridge Basement Complex in Virginia

Chelsea Jenkins
College of William and Mary

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Dating Deformation: Thermochronology and chemical and mineralogical changes in the Blue Ridge basement complex in 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

Chelsea Ellen Jenkins

Accepted for _______________________________
(Honors)

Dr. Christopher Bailey

Dr. Brent Owens

Dr. Deborah Bebout

Williamsburg, VA

2011
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Abstract

The Blue Ridge province in central and northern Virginia is underlain by Mesoproterozoic basement rocks and a Neoproterozoic to early Cambrian cover sequence. Basement rocks include granitoid gneiss, charnockitic gneiss, and younger charnockite and granite. Rift-related metasedimentary and metavolcanic rocks of Swift Run and Catoctin formations unconformably overlie the basement complex. The Blue Ridge basement and cover sequence experienced greenschist facies deformation during mid to late Paleozoic. $^{40}\text{Ar}/^{39}\text{Ar}$ analysis indicates this deformation occurred between the Acadian (420-380 Ma) and Alleghanian (325-260 Ma) orogenies in the late Devonian-early Carboniferous (Mississippian) during the Neoacadian orogeny (380-340 Ma). In places, charnockitic gneiss near the basement/cover contact has been converted to unakite. Consistency of these changes to a depth of 200m below the contact suggests alteration was not a result of Neoproterozoic contact metamorphism, but rather a product of Paleozoic regional metamorphism.
Introduction

Tectonics and erosion can produce striking geologic features, including volcanic arcs, canyons, and mountain ranges. The majestic Blue Ridge Mountains are one example of the magnificent scenery that these geologic processes can create. But these mountains were not always as we see them today. The granitic rocks that form the core of the mountains formed over a billion years ago and have been deformed several times throughout their history. Although rocks in eastern North America have experienced five major orogenies, or mountain building events, and three major rifting events (one failed and two that successfully opened an ocean). The exact timing of these deformation events remains enigmatic. Researchers have pieced together a timeline for these major geologic events in the Blue Ridge; however there are details about timing that have not been thoroughly agreed upon. My research focuses on the timing of Paleozoic deformation in the Blue Ridge province of central and northern Virginia. I also consider timing of observed mineralogical alterations within the lowermost crystalline rocks, or basement, near the contact with the overlying metasedimentary and metavolcanic rocks that cover the basement.

The Blue Ridge province is a major part of the southern and central Appalachians orogen that extends from Georgia to Pennsylvania (Fig. 1). Blue Ridge igneous basement rocks intruded over a billion years ago during the Grenville orogeny (Tollo and others, 2004a). By the end of the Neoproterozoic, approximately 570 Ma, uplift and erosion of the country rock had exposed the basement, and the Swift Run Formation was deposited directly on the older igneous rocks (Southworth and others, 2009). The Catoctin Formation extruded during rifting associated with the break up of Rodinia and the
opening of the Iapetus Ocean. Due to the basement’s uneven surfaces and the discontinuous nature of the Swift Run, the Catoctin sometimes directly overlies the crystalline basement. As the environment transitioned from an active rift zone to a passive continental margin, the overlying Chilhowee Group and Cambrian-Ordovician carbonates were deposited (Fig. 2).

Plate tectonics, the same mechanism that broke apart Rodinia, later reunited the continents into another super continent, Pangaea. As the components of the former super continent began to converge, the Iapetus Ocean between them gradually closed (Fig. 3). As the oceanic crust of the Iapetus Ocean subducted, continental crust riding atop the subducting basin collided with and accreted to Laurentia. The collisions that formed mountain ranges are called orogenies, and those that affected the central Appalachians are generally recognized in three different groups of collision events. The timing and intensities of these collisions can vary due to the shape of the paleo coastline and the nature of the collision (if the coastline protrudes further out in one area compared to another, the collision will have happened sooner and more intensely than in the more sheltered areas). The Taconic orogeny, the first of the Paleozoic central Appalachian orogenies, occurred in the early Paleozoic as the Taconic volcanic island arc collided with Laurentia (Drake and others, 1989). Next, the Acadian orogeny occurred in the middle Paleozoic, as the continent Baltica collided with Laurentia (Osberg and others, 1989). Traditionally considered the third and final orogeny, the Alleghanian orogeny occurred during the late Paleozoic when Gondwana collided with Laurentia (Hatcher and others, 1989). It was during these collisions that the rocks of the Blue Ridge were deformed and transported westward via thrust faults and high strain zones over the
younger rocks in the Valley and Ridge (Fig. 4). In the last decade researchers have begun to recognize an orogenic event between the traditional Acadian and Alleghanian orogenies known as the Neoacadian orogeny, wherein the Carolina Superterrane collided with the Laurentian margin (Merschat and Hatcher, 2007).

Figure 1. Map of eastern North America depicting the Appalachian Mountains and the extent of the Blue Ridge province. Inset shows the Blue Ridge province within the state of Virginia relative to the other geologic provinces.
Figure 2. General stratigraphic column of Blue Ridge in Shenandoah National Park, Virginia (from Bailey, 2010.) Included within the Mesoproterozoic (Y) basement are the three temporally distinct igneous intrusions (Y$_1$:1183-1144 Ma, Y$_2$:1120-1111 Ma, Y$_3$: 1078-1020 Ma; Tollo and others, 2004a)

Figure 3. (next page.) A reconstruction of paleogeography from the late Neoproterozoic through the Paleozoic depicting the opening and closing of the Iapetus Ocean. (Maps modified from Blakely, 2010)
Fig 3.a. 600 Million years ago in the Neoproterozoic, the majority of Earth’s continents were in the super continent, Rodinia.

b. By 560 Ma, Rodinia had begun to drift apart. The Iapetus Ocean was opening up between the new continents.

c. By 500 Ma, in the Cambrian, the Chilhowee groups was depositing along the southern coast of Laurentia.

d. By 470 Ma, during the Ordovician, the Taconic Arc was colliding with Laurentia in the Taconic Orogeny.
e. By 450 Ma, in the Silurian, the Taconic Arc had already accreted onto Laurentia. Now Baltica and Laurentia are on a collision course.

f. By 430 Ma, Laurentia and Baltica were converging, starting the Acadian Orogeny.

g. By 400 Ma, in the Devonian, Laurentia was right in the middle of the Acadian Orogeny.

h. By 370, the Acadian orogeny had finished and the Iapetus ocean was restricted to the basin between Laurentia and Gondwana.
I. By 340 Ma, in the Carboniferous, Gondwana was on a collision course with Laurentia.

j. By 300 Ma, in the beginning of the Permian, the Alleghanian Orogeny was well underway and the Iapetus Ocean was no more. The mountain range between the two continents is the precursor to the modern Appalachians.

k. By 260 Ma, towards the end of the Permian, Laurentia and Gondwana had finished colliding to form Pangaea.
Proposed timing for Paleozoic greenschist facies deformation in the Virginia Blue Ridge includes the Taconic orogeny in the Ordovician (Evans, 1991), the Acadian orogeny in the late Ordovician to Devonian (Bartholomew and Lewis, 1984), the Neoacadian orogeny in the late Devonian to Missippian (Wooton and others, 2005, Bailey and others, 2007), and the Alleghanian orogeny in the Pennsylvanian (Mitra and Elliot, 1980). To investigate timing of deformation, I will compile published thermochron data from the Blue Ridge, as well as some new data. Ar/Ar analyses quantify the time elapsed since particular minerals in a rock cooled through a certain threshold, different for each mineral, known as the closure temperature (~480°C for hornblende, ~350°C for muscovite and white mica) (McDougall and Harrison, 1999). The results indicate a cooling age, or a minimum age for the most recent episode of deformation and regional metamorphism.

Additionally, I will be examining basement rocks near the contact between the basement and the overlying Swift Run or Catoctin formations. In Shenandoah National Park, researchers report chemical and mineralogical changes within the basement, apparently related to the unconformity (Tollo and others, 2004b). Where observed,
recrystallized feldspars in bleached charnockite basement below the contact suggest contact metamorphism (Tollo and others, 2004b). Around fractures near the contact, epidote replaces plagioclase, and hematite replaces some alkali feldspar, converting charnockitic (orthopyroxene bearing granite) basement rocks to unakite (altered granite characteristically containing epidote, alkali feldspar, and quartz). This could be a result of hydrothermal fluids either during the extrusion of the Neoproterozoic Catoctin Formation or during Paleozoic deformation (Tollo and others, 2004b). I will investigate chemical and mineralogical changes in the basement rocks exposed in the northeast area of the Elkton East 7.5’ quadrangle (Fig. 5). The relationship between these changes and the unconformity will help to determine both the cause and the timing of the alteration of the charnockite (Fig. 6).
Figure 5. Area locator- samples collected in the northeast Elkton East 7.5’ quadrangle. Inset shows Virginia, with this map excerpt shown by a red square.
Figure 6. Schematic representations of two possible hypotheses for observed alterations in the basement. The pink hashes denote where alterations occurred. In a regional metamorphism scenario, alterations would be observed pervasively or through depths within the rock due to widespread heat and pressure. In a contact metamorphism scenario, alterations would be limited to the paleo-surface and around fractures where volcanic fluids would have flowed and made direct contact.
Precise timing of Paleozoic deformation in the Blue Ridge has proven elusive. Workers have attributed the deformation to three different Appalachian orogenic events. Evans (1991) noted that clasts of metamorphosed Cambrian rocks occur in the Fincastle conglomerate at the base of the mid-Ordovician Martinsburg Formation, indicating that deformation occurred during the Ordovician Taconic orogeny (460-440 Ma). Bartholomew and Lewis (1984) argued that the greenschist facies fabric associated with Paleozoic deformation developed synchronously with movement along the Fries fault system during the Acadian orogeny (420-380 Ma). Bartholomew and Lewis argue that water introduced during thrusting allowed the development of observed greenschist facies mineral assemblages. Mitra and Elliot (1980) use the South Mountain cleavage observed in Devonian rocks within the Massanutton syncline, but not in Triassic basins east of the Blue Ridge anticlinorium to bound the age of Paleozoic deformation to the Alleghanian orogeny (320-280 Ma). Wooton and others (2005) use isotopic dating of hornblende and white mica to suggest that deformation occurred sometime during the late Acadian to early Alleghanian orogenies. Interestingly, reported hornblende cooling ages vary by about 545 million years from the western limb (~900 Ma) to the eastern limb (~400 Ma) of the anticlinorium (Bailey and others 2007a), suggesting a complex deformation or cooling history.

Wooton and others (2005) investigated quartz microstructures and mylonite mineralogy, and suggested that rocks in the western Blue Ridge experienced temperatures of approximately 300-350°C during deformation, whereas rocks in the eastern Blue Ridge experienced temperatures up to 400°C. Isotherm dating on muscovite in western
Blue Ridge rocks suggests that these rocks had cooled through 350 °C during the early Alleghanian (Wooton and others, 2005). Burton and others’ (2000) work in the eastern limb of the Blue Ridge anticlinorium indicate that deformation occurred at peak temperatures above 700 °C at 6kb. Work on timing of this deformation in the Piedmont Carolina terrane and the Chopawasmic terrane bounds Paleozoic greenschist and amphibolite facies metamorphism to pre-305 Ma, given the presence of unmetamorphosed granodiorite dikes of that age. Analysis of $^{40}\text{Ar}^{39}\text{Ar}$ ages of hornblende, biotite, and muscovite from the greenschist and amphibolite facies country rocks suggest these rocks cooled through 500°C approximately 284 Ma, and through 350°C by 260Ma (Burton and others, 2000). It is clear that further work must be done to better understand the timing of Paleozoic deformation in the Blue Ridge.

**Geologic Setting**

The Blue Ridge anticlinorium is a series of folds that extends from southwest Pennsylvania through Maryland and Virginia. The western limb of the anticlinorium lies within the Blue Ridge province and the eastern limb lies within the neighboring Piedmont province (Espenshade, 1970). The anticlinorium is interpreted to be a hanging wall anticline at the hinterland edge of a late Paleozoic thrust complex that moved the Blue Ridge rocks over the Valley and Ridge rocks to the west (Evans, 1989). Exposed at the core of the anticlinorium is the Mesoproterozoic basement complex, and exposed in the limbs are the siliciclastic metasedimentary and metavolcanic cover sequence (Espenshade, 1970).
The igneous basement rocks intruded over three temporally distinct episodes during the Mesoproterozoic (Tollo and others, 2004a). The oldest suite consists of granitoid gneisses that date to 1.16-1.14 billion years ago. The second suite consists of orthopyroxene bearing granites that date to 1.1 billion years ago and is not as expansive as either other intrusions. The third suite consists of charnockite, leucogranites, and other granitoids that date to 1.08-1.05 billion years ago (Tollo and others, 2004a). The oldest granitoid intrusion in the Blue Ridge is interpreted to both intruded and been deformed during the long-lived Grenville orogeny (Tollo and others, 2004a), a mountain building event associated with the collision of continents to form the supercontinent Rodinia.

A Neoproterozoic to early Cambrian cover sequence overlies the basement. This includes the rift related metasedimentary and metavolcanic Mechum River, Swift Run and Catoctin formations, and the nonmarine to marine siliciclastic rocks of the Chilhowee Group. The Mechum River Formation contains metasedimentary rocks that crop out in a narrow band in the central Blue Ridge province and are the product of a failed rifting event around 730 Ma (Bailey and others, 2007b). The Swift Run includes arkosic quartzite, metasandstone, metaconglomerate, arkose, phyllite, and volcanoclastic metasediments (King, 1950; Gathright, 1976). The Catoctin Formation, extruded 562± 5 Ma (Southworth and Aleinikoff, 2007), is primarily massive to schistose metabasalt with interlayered metasedimentary rocks including phyllite, meta-arkose, and metaconglomerate (Badger and Sinha, 2004). Both the Swift Run and Catoctin formations are interpreted to represent the second major rifting event in the region- this one successfully opening an ocean basin (Southworth and Aleinikoff, 2007). In north-central Virginia, the Chilhowee Group includes the Weverton, Harpers, and Antietam
formations (Fig. 2). The Weverton contains pebbly meta-conglomerate interbedded with phyllite. The Harpers contains drab metasandstones, metasiltstones and phyllite. The Antietam Formation is known for clean sandstones with characteristic *Skolithos linearis* trace fossils (Gathright, 1976). Together the Chilhowee Group represents the subsequent transition from terrestrial to marine environments associated with the opening of the Iapetus Ocean and the transgressive and regressive sequences that occurred thereafter (Southworth and others, 2009). Cambro-Ordovician carbonates cap this sequence, deposited along the resultant Laurentian passive continental margin.

The region became tectonically active again later in the Paleozoic. The Blue Ridge province experienced four major collision events, at least one of which could have deformed the Blue Ridge rocks to the greenschist facies (Bailey and Simpson, 1993). First was the Taconic orogeny (480-450 Ma) during which a volcanic island arc collided with the margin of Laurentia (Hatcher, 2008). Next was the Acadian orogeny (420-380 Ma) and Neoacadian orogeny (380-340 Ma) (Merschat and Hatcher, 2007), during which small continents collided with Laurentia. Last was the Alleghanian orogeny (325-260 Ma), during which Gondwanaland collided with Laurentia, producing the supercontinent Pangaea (Hatcher, 2008). It was approximately 280 Ma, during Alleghanian contraction, that the Blue Ridge province was thrust westward over the Valley and Ridge province (Meyer and others, 2005).

Mesozoic rifting later broke apart Pangaea, creating the modern Atlantic Ocean. This rifting is evidenced in the Blue Ridge by a suite of northwest- and northeast-trending Jurassic diabase dikes (Southworth and others, 1993). Since then, eastern North America has been comparatively tectonically inactive, and the thick sediments of the
Coastal Plain province have accumulated to the east. The modern mountainous topography expressed today formed by differential erosion in the Cenozoic (Poag and Sevon, 1989).

Methods

Timing of Deformation

Isotopic dating can be used to determine the age of deformation in rocks. In this study, the \(^{40}\text{Ar}/^{39}\text{Ar}\) method will be used to estimate the time elapsed since the mineral has been at or below its closure temperature - the temperature at which the mineral’s crystal structure no longer allows particular elements to diffuse through the lattice, trapping all constituents inside. The mineral becomes solid when it first crystallizes, but can reopen with sufficient heat during deformation. The mineral closes again while cooling down after deformation. When the crystal closes, elements that undergo radioactive decay are trapped inside the crystal along with the daughter elements they produce. During analysis, the time elapsed since closure is determined using both the ratios of a parent element to its radioactive decay product and the decay constant. This method gives a lower limit to the age of deformation (cooling occurs after a deformation event), or to the growth of the crystals (if the deformation never reached temperatures high enough to reopen the crystals) and is useful when piecing together a region’s geologic history.

In \(^{40}\text{Ar}/^{39}\text{Ar}\) analyses, rocks are crushed and potassium-bearing minerals are isolated from the rest of the sample. These minerals are irradiated, converting the \(^{39}\text{K}\) and
to $^{39}\text{Ar}$, and then fused through an ultrahigh vacuum system. After purification, they are run through a mass spectrometer to measure abundances of $^{40}\text{Ar}$, $^{39}\text{Ar}$, $^{37}\text{Ar}$, and $^{36}\text{Ar}$ (McDougall and Harrison, 1999). The $^{37}\text{Ar}$ and $^{36}\text{Ar}$ are used to correct for certain interferences in the process, such as the presence of non-radiogenic argon. The samples are heated in stepwise increments, with the ratio of $^{40}\text{Ar}/^{39}\text{Ar}$ analyzed at each increment. The result or plateau age, comes from three or more of these increments that coincide within a 2$\sigma$ error range, while obtaining at least 50% of the argon released (Berger and York, 1981; Albarede 1982).

Amphiboles are common in intermediate to felsic igneous rocks and amphibolites facies metamorphic rocks (Perkins, 2002). Amphiboles are identifiable by their dark color and characteristic amphibolic cleavage ($56^\circ/124^\circ$) (Fig. 7). Hornblende (K,Na)$_{0.1}$(Ca,Na,Fe,Mg)$_2$(Mg,Fe,Al)$_5$(Si,Al)$_8$O$_{22}$(OH)$_2$ is an amphibole that commonly has potassium or sodium filling in the otherwise vacant 10-12 coordination site. While not extremely abundant in Blue Ridge rocks, hornblende is common enough to find enough samples for this study. The closure temperature of hornblende for Ar is approximately 480$^\circ$C (McDougall and Harrison, 1999), which is towards the upper end of the greenschist (350$^\circ$C -500$^\circ$C) metamorphic facies. Muscovite is convenient due to its abundance, and is used in this study in addition to the hornblende data. Muscovite’s closure temperature is $\sim$350$^\circ$C (McDougall and Harrison, 1999) and using cooling ages from both minerals provides more information about the rate of cooling after deformation.
11 samples were collected, some from both the east and west side of the Blue Ridge (Fig. 8). Analysis of the metamorphic minerals in the samples was done at the United States Geological Survey labs in Reston, Virginia. These results were then compiled with previously measured Ar/Ar age spectra analyses and plotted spatially and according to which mineral was analyzed using ArcGIS. Plotting the ages spatially allows for comparison in ages across the Blue Ridge, both to verify the previous observation that deformation ages from hornblende vary from east to west (Bailey and others, 2007a), and to understand how the deformation and cooling history varies regionally in the Blue Ridge.

Figure 7. Charnockite in thin section (5x), pictured are hornblende (H) and white mica (WM), the colorless minerals are quartz (Q) and plagioclase (P).
Figure 8. Collection sites for Ar/Ar dating from May, 2010, in orange. Inset shows where this map is in Virginia. Background is a DEM, with lighter colors representing higher elevations and darker colors represent lower elevations.
Chemical and Mineralogical Changes at the Basement/Swift Run Unconformity

5 samples were collected from the basement in the Elkton East 7.5’ quadrangle in the Virginia Blue Ridge (Fig. 5). A structure contour map of the surface of the basement was made for the area so that each sample could be located relative to the contact between the basement and the overlying Swift Run Formation (Fig. 9a). In places where the information available was insufficient for constructing basement structure contours, a combination of cross sections (Fig. 9b) and contours of the bottom of the Weverton were used instead. Estimated thicknesses of the intermediary units were used to adjust the contours to the contact. Thin sections were made from the collected rocks and the samples were sent to Acme Analytical Labs for chemical analysis, where abundances of the major oxides and some minor elements were analysed by ICP-emission spectrometry following a Lithium metaborate/tetraborate fusion and dilute nitric digestion. The geochemistry and normative mineralogy of the new samples were compared to published data on other Blue Ridge charnockites. The actual mineralogy of the rock was observed in thin section and then compared to the expected mineralogy determined by performing CIPW norms based on the results of the chemical analyses. The depths at which differences between the observed and expected mineralogy of a sample are observed will suggest a timing and nature for the alteration.
Figure 9a. Sample collection locations 19, 20, and 22 are pictured (the location for sample 18 is to the northwest of this picture). Red lines represent structure contours of the top of the basement, or the elevation of the contact. Rock type is also pictured, the purple unit represents the basement, the pale orange the Swift Run, the green the Catoctin, and the brown the Weverton. Distance from the contact is determined by subtracting the current topographic elevation of the sample from where the contact would be if not for erosion (the structure contour). Contour Interval for the structure contours as well as the topographic contours are in feet.

Figure 9b. An example of the same concept illustrated in figure 9a., but in cross section. Y is the basement complex, Zsr the Swift Run, Zc the Catoctin, Cw the Weverton, Ch the Harpers, and Ca the Antietam. Dashed lines represent hypothesized projections if there were no erosion. Red number and lines show the distance between the surface (where samples are collected now) and the paleo-surface (depth from contact).
Results

Samples for Argon Dating

CJ-01 (Fig. 10a) is a fine-grained, well-foliated amphibolite with abundant amphibole, plagioclase feldspar, and quartz. CJ-02 is a well-foliated amphibolite with slightly larger grains. The hornblende makes up the majority of the rock, with some quartz and plagioclase, and minor epidote grains. CJ-03 is an amphibolite with little to no foliation and more heterogeneously shaped and sized grains. Hornblende is somewhat more abundant than the quartz and plagioclase combined. CJ-04 is a metagranite with no apparent foliation, and is primarily white mica, quartz, and plagioclase. Biotite is also present and there are some occurrences of garnet, microcline, and epidote. CJ-05 and CJ-06 (Fig. 10b) are altered charnockites, with quartz, plagioclase, white mica, and biotite, as well as some hornblende, orthopyroxene, some garnet, and opaque minerals. CJ-08 is a metagranite with a very weak foliation, best expressed by alignment of biotite grains. CJ-08 (Fig. 10c) contains primarily larger, scattered grains of plagioclase within a finer biotite, white mica, and quartz fabric. CJ-09 is a medium grained amphibolite with no readily apparent foliation. CJ-09 (Fig. 10d) is mostly hornblende with some plagioclase and a very small amount of quartz, with a dark stain around some of the hornblende, and some scattered biotite. CJ-10 is a charnockite that primarily consists of large quartz and feldspar (plagioclase and microcline) grains. Orthopyroxene grains stand out among the mostly colorless grains and exhibit fantastic pleochroism. CJ-11 is a highly fractured charnockite gneiss with mostly medium grains. Quartz and plagioclase are well represented and hornblende and orthopyroxene occur less frequently.
Figure 10a. Micrograph (ppl, 10x) of amphibolite (CJ-01) pictured are: hornblende (A), quartz (Q), plagioclase (P), and opaque minerals.

10b. Micrograph (ppl, 10x) of charnockite (CJ-06), featuring hornblende (A), white mica (WM), quartz (Q), plagioclase (P), and orthopyroxene (opx).

10c. Micrograph (ppl, 10x) of metagranite (CJ-08), featuring white mica (WM), quartz (Q), and biotite (B).

10d. Micrograph (ppl, 10x) of another amphibolite (CJ-09), with larger plagioclase (P) grains, and hornblende (A) grains with no preferential alignment.
Age results for the samples collected in May are still pending (Table 1), so analysis will focus on previously collected data. Minerals analyzed include amphibole, white mica, and biotite. Data ranges geographically from central Virginia to northern Maryland and through both the Blue Ridge and Piedmont provinces.

Amphibole data provide information regarding the higher temperature major deformation (Fig. 11a). Some hornblende was not reset during Paleozoic deformation, with the furthest west and northern samples recording cooling ages older than 900 Ma. Two of the Piedmont samples had cooled through hornblende’s closure temperature during the Neoproterozoic, and the rest of the hornblende Ar/Ar data records cooling ages from the Paleozoic. Paleozoic amphibole cooling ages are concentrated in the eastern Blue Ridge to Piedmont, and primarily record deformation from the Taconic orogeny (cooling ages from 448-430 Ma), with one sample recording Acadian deformation (cooling age of 400 Ma) and one sample in the Piedmont recording an Alleghanian age (cooling age of 295 Ma).

Biotite (Fig. 11b) and white mica (Fig. 11c) data provide information regarding the lower temperature deformation and cooling rates after deformation. Data available are concentrated in central Virginia and Maryland. Biotite ages range from 329 Ma to 455 Ma, recording deformation as late as the Carboniferous (Mississippian). Biotite ages primarily fall during the Acadian orogeny and the Neoacadian orogeny (intervening time during the Devonian-Mississippian before the Alleghanian orogeny). White mica ages cover much of the Paleozoic and into the Neoproterozoic, ranging from 667 Ma to 250 Ma.
Cooling temperatures were also plotted by age across the Blue Ridge (Fig. 12) to illustrate cooling over time. Samples were sorted into age bins and plotted by closure temperature. The Blue Ridge appears to have been hottest during the Ordovician in association with the Taconic orogeny (4 of the 6 amphibole samples plot in the 480-420 Ma bin). This same time accounts for 14% of the white mica samples and 13% of the biotite samples. 1 amphibole sample, 4 white mica samples (9.5%), and 6 biotite samples (40%) plot within the Acadian orogeny (420-380 Ma). The late Devonian-Mississippian interval between the Acadian and Alleghanian orogenies (380-320 Ma), during which the Neoacadian orogeny occurs, contains the majority of the white mica and much of the biotite data. This interval includes 26 white mica samples (62%) and 7 biotite (47%) samples. The final time interval (320-280 Ma) represents spans the Alleghanian orogeny and contains 1 hornblende and 6 white mica samples (14%).

Table 1. Samples for Ar/Ar analysis (pending)

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<th>Rock Type</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Minerals Analyzed</th>
<th>Apparent Age</th>
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<td>37.57696N</td>
<td>079.04347W</td>
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<td></td>
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<tr>
<td>CJ-02</td>
<td>amphibolite</td>
<td>37.50759N</td>
<td>079.12123W</td>
<td></td>
<td></td>
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<tr>
<td>CJ-03</td>
<td>amphibolite</td>
<td>37.54674N</td>
<td>079.07580W</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CJ-04</td>
<td>metagranite</td>
<td>37.60759N</td>
<td>079.02769W</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CJ-05</td>
<td>charnockite</td>
<td>37.70692N</td>
<td>079.05106W</td>
<td></td>
<td></td>
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<tr>
<td>CJ-06</td>
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Figure 11a. Ar/Ar cooling ages for amphibole at orange points, text denotes cooling age in Ma, green lines are state boundaries. (Data from Pavlides and others, 1994, Kunk and Burton, 1999, Wooton and others, 2005, Bailey and others, 2007, Kunk and McAleer, 2008).
Fig. 11b. Ar/Ar cooling ages for biotite at purple points, text denotes cooling age in Ma, green lines are state boundaries. (Data from Pavlides and others, 1994, Kunk and Burton, 1999, Wooton and others, 2005, Bailey and others, 2007, Kunk and McAleer, 2008). Abbreviations: VR is Valley and Ridge, BR is Blue Ridge, MzB is Mesozoic basins, and P is Piedmont.
Fig. 11c. Ar/Ar cooling ages for white mica at green points, text denotes cooling age in Ma, green lines are state boundaries. (Data from Pavlides and others, 1994, Kunk and Burton, 1999, Wooton and others, 2005, Bailey and others, 2007, Kunk and McAleer, 2008). Abbreviations: VR is Valley and Ridge, BR is Blue Ridge, MzB is Mesozoic basins, and P is Piedmont.
Figure 12. Ar/Ar cooling ages sorted into time bins through the Paleozoic, blue dots represent biotite (300°C), green dots represent muscovite (350°C), and orange dots represent amphibole (480°C). Abbreviations: VR is Valley and Ridge, BR is Blue Ridge, MzB is Mesozoic basins, and P is Piedmont.

320-280 Ma
Charnockite Samples

Samples collected for analysis were collected from basement outcrops in the northeastern corner of the Elkton East 7.5’ quadrangle. Locally, the Swift Run Formation is thin or absent, and in places the Catoctin Formation rests directly atop the basement. Three of the five samples collected are charnockitic gneiss (samples 18A, 18B, and 20) and two are altered charnockitic gneiss (samples 19 and 22 were uralitic). Each sample was collected from outcrop in or around the West Branch of Naked Creek. Sample 22 was the closest to the contact (12 m below), samples 20 (60 m) and 18 (180 m) were intermediate, and sample 19 (250 m) was the deepest sample collected.

Mineralogy and Chemistry

Quartz, feldspar, epidote, and opaque minerals commonly occur in each sample (Fig. 13). Garnet was observed in 18A, 19, and 20. Orthopyroxene occurred in samples 18A, 18B, and 20. Biotite occurred in sample 18B, and white mica in 19, and were likely constituents within the fine groundmass of 20 and 22. Hematite was apparent in samples 18B and 22. Each sample is moderately to highly fractured, samples 19 and 22 were the most fractured. Every sample featured sericite, a mineral that replaces feldspar during deformation. Uralite, a secondary hornblende that replaces pyroxenes and primary hornblende, is also present in each of the samples.

The CJ samples plot within the low silica range of the granite region of a QAP diagram (fig. 14). Normatively, the CJ samples contained a range from 17 to 25% quartz and 64-70% feldspar (Table 2). Each contained normative corundum, but the Al₂O₃ this represents is usually observed in the rock in other minerals, for example micas and
garnet. None of the newly collected samples contained clinopyroxene (normative diopside). All of the samples contained normative hypersthene (orthopyroxene), ranging from 4-11%. Other normative minerals include ilmenite (<2%), magnetite (<1%), and apatite (<.63%). None of the samples contained normative hematite.

A Mann-Whitney test was used to compare the mean percent of the normative minerals in the published samples (n=3) versus the CJ samples (n=5) (Fig. 15). The normative mineralogy of the CJ samples did not significantly differ from the published samples in quartz (p=.18), diopside (p=.051), or hypersthene (p=.101). The published samples contained more anorthite (p=.025), magnetite (p=.036), ilmenite (p=.025), and apatite (p=.025). The CJ samples contained more normative corundum (p=.024), orthoclase (p=.025), and albite (p=.025).
Figure 13a. CJ-18 in thin section (5x), with quartz (Q), plagioclase (P), sericite (S), and uralite (U).

13b. CJ-19 in thin section (5x), with quartz (Q), sericite (S), uralite (U), and epidote (E) (outlined in orange).

13c. CJ-20 in thin section (5x) with quartz (Q), plagioclase (P), and sericite (S).

13d. CJ-22 in thin section (5x), with quartz (Q), white mica (WM), epidote (E), and sericite (S).
Figure 14. Charnockite samples of this study plotted (using normative proportions) on a QAP diagram (alkali feldspars here labeled as K-spar) with charnockite samples from the literature (Tollo and others, 2004a; Gilmer, 1999). Normative albite and anorthite were combined and plotted as plagioclase.
Table 2

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Yqfj (Tollo and others 2004a). F19(AG99) (Gilmer, 1999).
Figure 15. Mean normative abundance of Tollo and others’ (2004a) and Gilmer’s (1999) charnockites and the CJ samples.
Discussion

Geochronology

Amphiboles record primarily Paleozoic cooling ages in the east, and early Neoproterozoic ages to the west, suggesting the rocks in the east were more heavily deformed than rocks in the west. Collisions that drove the orogenies occurred along the Paleozoic coastline, to which the eastern rocks are closer than the western rocks. Rocks in the Piedmont and eastern Blue Ridge were closer to the actual continental collisions and experienced a higher grade of metamorphism.

Because white mica and biotite cooling temperatures are very close to the temperature at which deformation occurred, the observed cooling ages are likely very close to deformation ages. White mica and biotite data suggest that deformation occurred during each of the traditionally recognized Paleozoic Appalachian orogenies, as well as during the interval between the Acadian and Alleghanian orogenies, recently recognized as the Neoacadian orogeny. This time interval (380-320 Ma) accounts for the majority of the white mica cooling ages collected.

Chemical and Mineralogical Changes in Charnockite

All of the Elkton East sample rocks may have originally been charnockitic, as they contain normative orthopyroxene, but that pyroxene may have since been replaced by uralite. Samples were heavily fractured at all depths, and the most fractured at the least and greatest depths below the contact, indicating that depth did not control the amount of fracturing. Some of the observed alterations, including fracturing and presence of hematite, could have occurred after the rocks were exposed at the surface. This is
especially true of the occurrence of hematite for rocks exposed in the creek. However, sericitization and conversion of pyroxene to uralite requires deformation, not simply chemical or physical weathering. Accordingly, these rocks were likely altered by hydrothermal fluid flow during Paleozoic regional metamorphism.

Future Work

To investigate the geochronologic examination of Blue Ridge deformation further, more Argon data should be collected in central Virginia and further west in the Blue Ridge province. It would be helpful to have more hornblende ages to see if the observed patterns hold with more data. Continued work with the altered charnockites should begin with the collection and analysis of more samples. It would also be interesting to look at other basement exposures in the Blue Ridge to see if other locations are altered in the same way.

Conclusion

The Blue Ridge province in Virginia experienced four major Paleozoic orogenies including the Neoacadian orogeny. $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages from biotite, muscovite, and amphibole indicate that greenschist foliation in the northern and central Virginia Blue Ridge occurred during this recently recognized Neoacadian deformation event.

Physical and mineralogical changes in charnockite in the Elkton East area of the Blue Ridge at various depths below the contact between the charnockitic basement rocks and the metasedimentary and metavolcanic rocks indicate that these changes occurred during Paleozoic regional metamorphism, rather than during Neoproterozoic extrusion of the Catoctin formation.
Acknowledgements

Thank you to Dr. Chuck Bailey for guidance and considerable patience along the way. Thank you to the rest of the geology faculty for all the support and for making geology so exciting. Their enthusiasm makes every hour spent seem worthwhile. Thank you to the Elkton Easter Bunnies (Molly Cox and Grace Dawson). I simply haven’t got the words to explain how incredible they are, so I won’t try. Thank you to all the other Geology majors (past and present), but especially the class of ’11. I can’t think of a more amazing group of people. Thank you to my family and friends, I obviously would not have gotten anywhere without their support. Thank you to my honors committee for taking the time to consider my project.
References


Merschat, A.J., Hatcher, R.D., 2007, The Neoacadian orogeny in the Southern and Central Appalachians; a kinematic model linking Middle Devonian- Early Mississippian accretion of the Carolina Superterrane, orogenic channel flow, and
foreland sedimentation: Geological Society of America Abstracts with Programs, v. 39, p. 68


**Appendix 1: Geochemical Analysis (Tables and Graphs)**

Geochemical Analysis of Tollo and Gilmer samples

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