Geology of the Mount Rogers area, revisited, Blue Ridge, VA–NC–TN

Carolina Geological Society
Annual Field Trip

October 29–30, 2016

Guidebook Editor:

Arthur J. Merschat

Field Trip Leaders:

Christopher S. Holm-Denoma, Jamie Levine, Ryan J. McAleer, Arthur J. Merschat, Scott Southworth, Crystal G. Wilson

ACKNOWLEDGMENTS

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Reviewers

Mark W. Carter (USGS), Bart L. Cattanach (NCGS), Loren A. Raymond (Appalachian State University emeritus), Scott Southworth (USGS), Christopher S. Swezey (USGS), and Jonathon Tso (Radford University) provided thorough and detailed reviews of the manuscripts.

Cover photo: Spring time view from the pinnacles on Haw Orchard Mountain across Grayson Highlands State Park and Wilburn Ridge to Mount Rogers, elevation 5729 feet.

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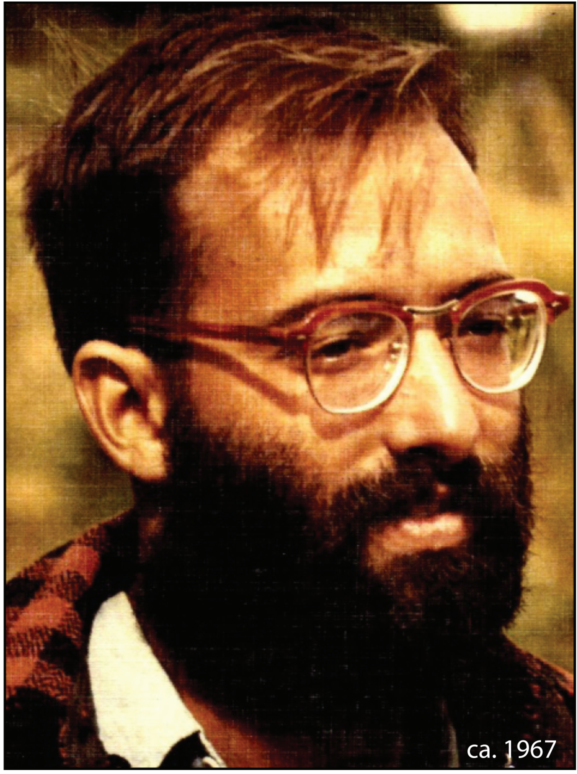
Dedication:
Douglas W. Rankin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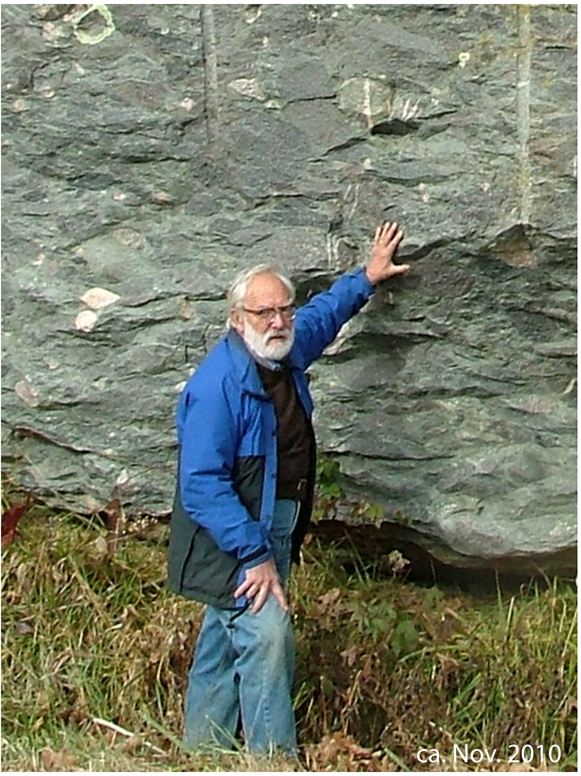
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For his 33 years of service to the U.S. Geological Survey and over 50 years of research into the geology of the Appalachian orogen and beyond.

CONTENTS

Memories of Doug Rankin

Bob Hatcher
University of Tennessee–Knoxville

Most people knew Doug Rankin for his long-time service to the USGS, for his geologic mapping—beginning with his dissertation in the Mt. Katahdin area in Maine, followed by his work in the Piermont allochthon in western New Hampshire and Vermont. He then focused his attention on the southern Appalachians—and developed his ideas about the tectonic history of the southern and central Appalachians.

I first knew Doug as my professor for undergraduate mineralogy, optical mineralogy, and petrology at Vanderbilt University during the late 1950s. He did not stay there very long and moved to the USGS to begin a long and successful career.

I do not think that Doug was ever comfortable in front of even the small classes that we had in those days, but a few of us were drawn to the subjects that he taught, and the way he taught them. [Throughout his career Doug frequently read his presentations at professional meetings—in the USGS tradition as he put it.] Mineralogy was a year-long course in those days, and we finished that with a thorough understanding of the subject. Optical mineralogy and petrology were not required courses, so most Vanderbilt undergrads took neither. There was something about the subject and perhaps Doug, however, that drew me into taking both optical mineralogy and petrology. Maybe it was his youth and willingness to talk with students one-on-one, not to mention that he would engage in snowball fights with us—with him as the target—whereas the other two (much older) faculty remained far aloof from this kind of activity. Nevertheless, when I signed up for optical mineralogy, only one other student signed up with me. This was my first experience in a class this small, but Doug chose to meet class in the usual fashion and deliver lectures, so the two of us got lots of attention both in lecture and lab.

The following semester I signed up for petrology and thought that there would be a similar arrangement, but it turned out that I was the only student in the class. Fortunately, class sizes were not subject to the kind of scrutiny experienced today or I might not have gotten to take petrology. Doug chose to meet the class in his office and, instead of lecturing, he gave me parts of a current igneous petrology book supplemented by some of the early 1900s papers on phase equilibria where each step in the crystallization process of some suite of minerals in a phase diagram was described in detail. Labs were great: I was taught to identify rock-forming minerals in thin sections one-on-one, and was fascinated by the variety and complexity of igneous rocks. (It also helped that I had had a year of physical chemistry by the time I began the course.) Doug's teaching method required me to prepare for several hours before every class and bring in questions about things I did not understand, or try to answer questions he would ask—again something to which I was not accustomed. This method worked well through igneous petrology; I learned a lot, and became fascinated with the subject.

We finished igneous petrology and I walked into his office to begin learning about metamorphic rocks but, instead of assigning me readings from standard texts (there were few), Doug handed me the Harvard notes from his graduate metamorphic petrology course under Jim Thompson. He then told me to begin studying the notes and to come to his office at class time prepared to discuss the contents as we systematically worked our way through them. Needless to say, this turned out to be a daunting challenge that required me to climb to a higher plateau where I had never been before as an undergrad or first-year grad student. The reason to go through all of this is that the experience of taking this course the way Doug chose to teach it, coupled with the subject, was one of the high points of my Vanderbilt education. This helped lay the foundation for other happenings (e.g., being taught field geology by two Tennessee Division of Geology geologists) that occurred in those early years that became other (serendipitous) blocks in the foundation that strongly influenced how my own career developed. In summary, I owe a great debt of thanks to Doug and hope that I adequately communicated that to him before he passed away two plus years ago.

Doug Rankin's Mt. Rogers work, together with reconnaissance mapping of the Winston-Salem 1° x 2° sheet (Rankin et al., 1972; Espenshade et al., 1975), provided the basis for several synthesis papers that cemented his long-term contributions to southern Appalachian tectonics (Rankin, 1970, 1976; Rankin et al., 1973). His work that resolved the different components of the Mt. Rogers Formation, the Konnarock Formation, and the resolution of the Ashe and Alligator Back Formations, are the products of careful geologic mapping and synthesis of his observations.

My first field trip to the Mt. Rogers area was in 1971; it was led by Doug (Rankin, 1971). He had been mapping there as a USGS geologist for several years, and organized the trip for the Southeastern Section meeting of the Geological Society of America. I had been out of college for ~6 years and received a very good dose of regional tectonics on the trip, along with Doug's new ideas about possible Neoproterozoic glacial deposits (now the Konnarock Formation; Rankin, 1993), the bimodal volcanic nature of the Mt. Rogers Formation (as it was known then), and the rifted-margin sedimentary rocks of the western Blue Ridge (Rankin, 1970;

Figure 1. Doug Rankin making a point on a Northeastern Section of GSA field trip in March, 2013, which he led to some of Marland Billings' original Devonian formations type localities in northern New Hampshire (Rankin and Rankin, 2014). Wife Mary Rankin is in the dark windbreaker to the right of Doug. Snowflakes for scale.

Rankin et al., 1969; Aleinikoff et al., 1995). Doug had a firm idea at that time of the tectonic implications of both the bimodal nature of the Mt. Rogers Formation, the Konnarock Formation, and their relationships to the Neoproterozoic rifting of the Laurentian margin (Rankin, 1975, 1976).

Doug subsequently mapped and did the petrology of the volcanic rocks in part of the U.S. Virgin Islands (Rankin, 2002), and did some mapping and petrologic work on volcanic rocks in the Absaroka Mountains in Wyoming. This pursued a theme of interest in volcanic rocks throughout his career beginning with his work on the Traveler Rhyolite in the Mt. Katahdin area. In addition, Doug coordinated a study of the area affected by the 1886 Charleston, SC, (actually Summerville) earthquake that resulted in publication of USGS Professional Paper 1028 (Rankin, 1977). He also played a major role in the Geological Society of America Decade of North American Geology project by publishing several syntheses in DNAG volumes (Rankin, 1994; Rankin et al., 1989, 1993), and being the lead author of one of the continent-ocean transects (Rankin et al., 1991). Throughout this time Doug never lost his interest in or ties to New England geology (e.g., Rankin et al., 2007). He was a contributor to the 2006 lithotectonic map of the Appalachians (Hibbard et al., 2006). One of his last contributions was leading a GSA Northeastern Section Meeting field trip on a bright but cool day (high in the mid-20s) in March to Marland Billings' localities in northern New Hampshire where many of the Devonian units were first described (Rankin and Rankin, 2014). He and wife Mary spent much of a day before the trip clearing snow from of several of the exposures.

It is appropriate to honor Douglas W. Rankin for his contributions to our understanding of the Mt. Rogers area and southern Appalachian tectonics as part of the 2016 Carolina Geological Society field trip. It is clear from his papers that many of his ideas on southern Appalachian tectonics had their roots in the Blue Ridge of northwestern North Carolina and adjacent southwestern Virginia and northeasternmost Tennessee. Additional details of Doug's professional life can be found in Tucker and Robinson (2015).

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Rankin Revisited

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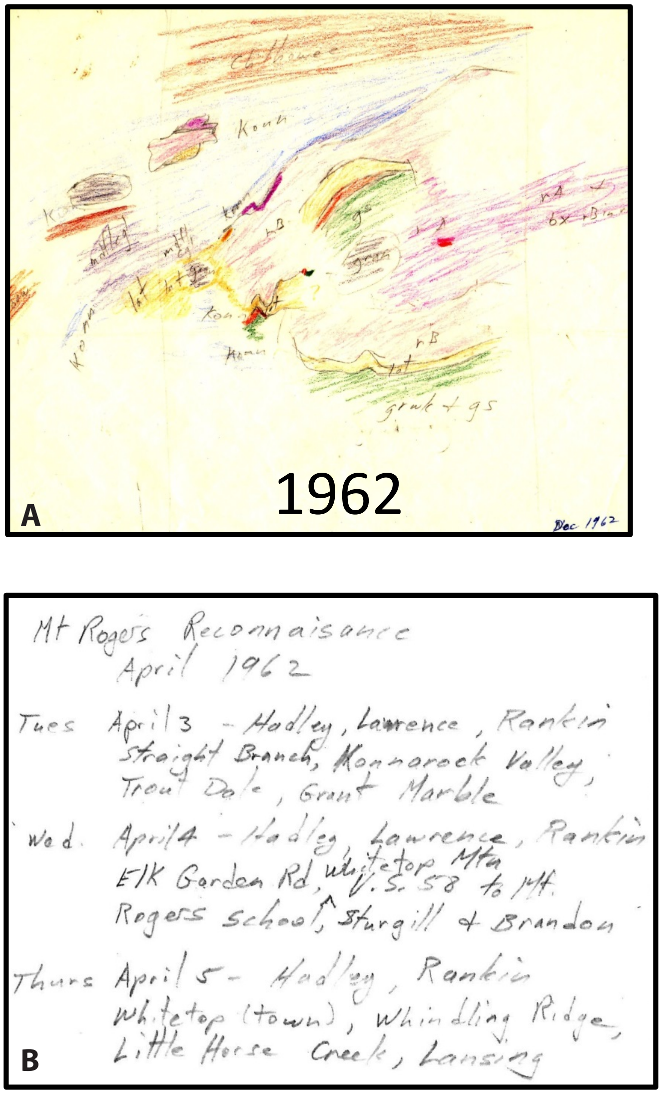
Douglas Whiting Rankin, (Doug), was born, raised, and lived on metamorphosed Ordovician rocks overlain by Tertiary gravel, first above an island arc complex (Newark, DE) and later above diamictite intruded by plutons (Washington, D.C.). Although born a flatlander, he was attracted to the Appalachian Highlands after summer vacations as a child in VT&NH which led to leading a trail crew for the Appalachian Mountain Club. He migrated to the highlands of northern Maine as a Ph.D. candidate at Harvard University, advised by Marland Billings. Doug loved a challenge so he pursued remote and difficult terrane for his dissertation: "Bedrock geology of the Katahdin-Traveler area, Maine" (1961). He was an Assistant Professor at Vanderbilt University (Nashville, TN) from 1958-61. "This was no place for a 'New Englander'. The faculty consisted of a southern gentlemen Professor and his son-in-law."

He was hired by USAID in Washington, D.C., and was in training to work in Chile when the budget was cut. His superior called the USGS and he was hired by Jerry Hadley. Jerry had been a Billings' student and had studied the bedrock where Doug had vacationed at Lake Fairlee, VT.

Jerry took Doug on a week-long field excursion in April 1962 to determine where and what he was best suited to study (Fig. 1). The plutonic-volcanic succession of Maine and New Hampshire attracted Doug to chose the bi-modal volcanic rocks around the Mount Rogers area of VA, NC, and TN. This logical area and topic were based on recently completed and on-going studies by Anna Jonas Stose, George Stose, Phil King, Herman Ferguson, Bruce Bryant, and Jack Reed. He started field work in May, 1962, driving south on Rt. 11 in a Wiley Jeep with his dog. He conducted extremely detailed field work until 1965. Prominent geologists visited to see the rocks and provide feedback, and he was most influenced by Billings and King. The nearly completed Konnarock and Whitetop Mountain 7.5-minute quadrangles were soon suspended and his new assignment was to map the west half of the Winston-Salem 1:250,000-scale quadrangle. It is not easy to shift from 1 or 2 maps to ***64*** 1:24,000-scale quadrangle maps. "I considered this assignment to be a dubious honor…I was scared to death. I did not know how it was going to pan out." His militaristic field work from 1965-69 resulted in the legacy map published in 1972.

In his spare time, Doug prepared a Guidebook and lead the Carolina Geological Society Field Trip to the Mount Rogers Area, October 14-15, 1967 (Fig. 2). It was his first formal presentation of his findings: Precambrian Cranberry Gneiss, mafic and felsic plutons and dikes that intruded it, the unconformably overlying



Mt Rogers Reconnaisance
April 1962
Tues April 3 - Hadley, Lawrence, Rankin
Straight Branch, Konnarock Valley,
Trout Dale, Grant Marble
Wed. April 4 - Hadley, Lawrence, Rankin
Elk Garden Rd, Whitetop Mtn, U.S. 58 to Mt.
Rogers School, Sturgill & Brandon
Thurs April 5 - Hadley, Rankin
Whitetop (town), Whindling Ridge,
Little Horse Creek, Lansing

Figure 1. Excerpts from Doug Rankin's field notebook from reconnaisance in the Mount Rogers area in 1962. (A) Field sketch map of the geology of the Mount Rogers area, and (B) descriptions of areas covered during reconnaisance.

Southworth, S., 2016, Rankin Revisited, *in* Merschat, A.J., ed., Geology of the Mount Rogers area, revisited, Blue Ridge, VA–NC–TN: Carolina Geological Society Annual Field Trip Guidebook, p. 1–2.

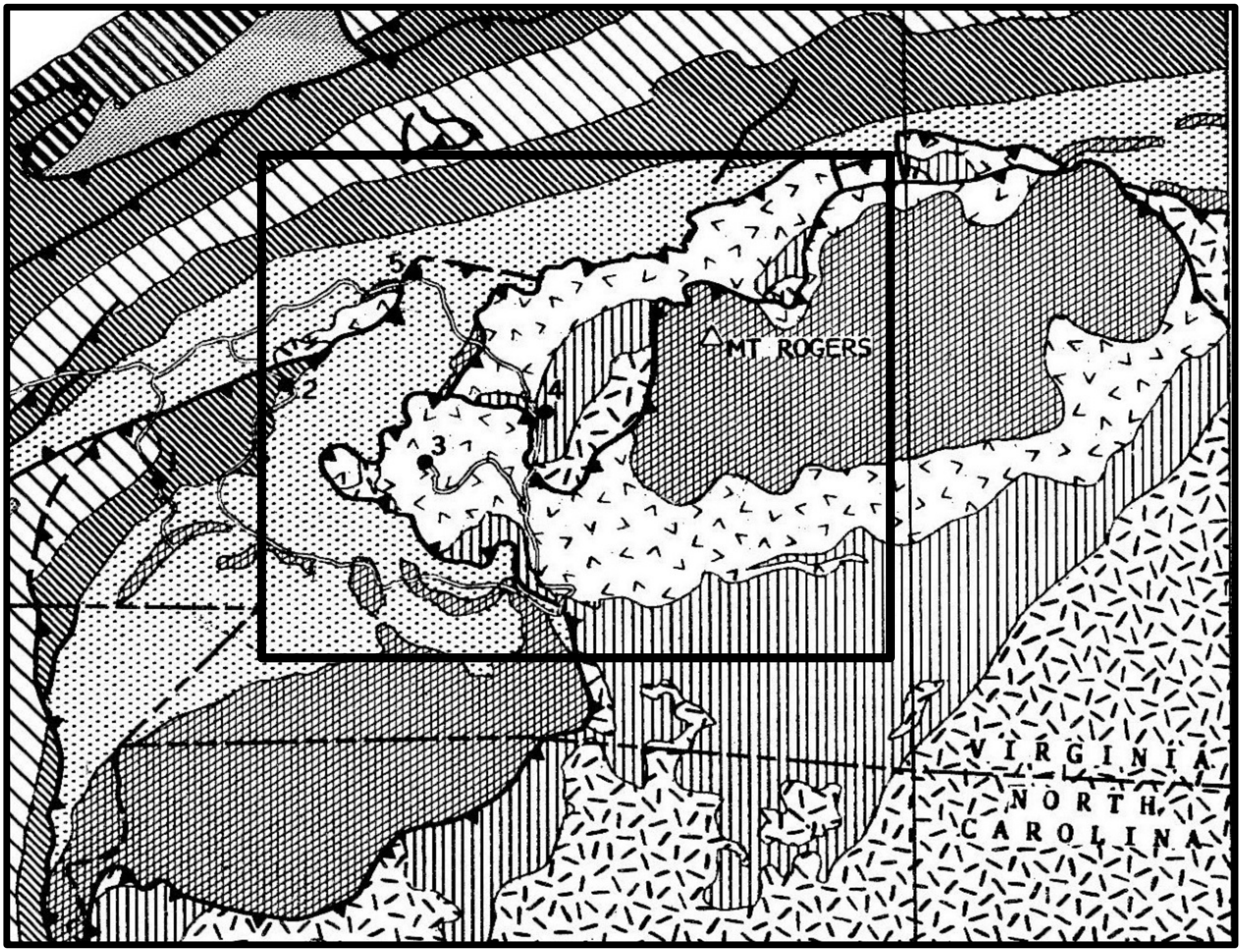


Figure 2. Geologic map of the Mount Rogers area, VA-NC-TN by Doug Rankin for the 1967 Carolina Geological Society Field Trip.

late Precambrian Mount Rogers Volcanic Group (MRVG), the "conformable Early Cambrian (?)" Unicoi Formation and basalts; amphibolite and mica gneisses and schists, of unknown age, were considered to be unconformable above Cranberry Gneiss on the "east limb of the anticlinorium". He recognized that granitoids and rhyolites comprised a magmatic cycle (to become the Crossnore Suite), based in part on Billings' influence on the White Mountain Series, NH, and Doug's work on the Katahdin Granite and Traveler Rhyolite. He recognized that the sedimentary and bi-modal volcanic rocks in the lower part of the MRVG were overlain by rhyolite in the middle part, and overlain by sedimentary rocks in the upper part. Three rhyolite units in the middle part constituted the approximate site of a volcanic center. Rhythmite and tillite in the upper part contained clasts that were "rafted into place by ice", thus were of glacial origin.

He formally named the rocks in 1993. The majority of the geologic story remained the same, including the delineation of thrust sheets and tectonic windows. With Tom Stern, Jack Reed, and Marci Newell, he published one of the first U-Pb studies of igneous zircon ages from the rhyolites at Mount Rogers (Science, 1969). He soon placed the rocks into the context of plate tectonics and continued studies of the rocks from 1973-80, 1984-89, 1991, 1995, and 2005, to refine his observations and interpretations and document the results. The bulk of his knowledge was well documented in the GSA books and maps on the Geology of North America (1989-90). Most remarkable was his ability to write and publish significant papers on the rocks 30-40 years later.

Doug's special talent was compiling and synthesizing bedrock geology. "Digital compilations" were with steady hands on green line maps and mylar using rapidograph pens and ink, standing on a low bench so he could hover over the large maps, wearing magnified lens. Precise line work was photographically reduced in scale, further compiled (1:100,000-, 1:250,000-, and 1:2,500,000-scales), and retained the accuracy. In retirement he taught himself Adobe Illustrator.

Doug led a field trip here in 1992 which planted the seed for our current studies. He demonstrated remarkable teaching skills on the outcrop. He persisted on field reviews to evaluate new findings. Doug's geologic map data will be published at 1:50,000-scale and database will include a robust suite of new geochronologic isotopic data. We are only embellishing the geology that he had figured out so very well when he lead the trip here in 1967. Doug's demeanor, skills, and mentoring are sorely missed.

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Geology of the Mount Rogers area, Revisited: Evidence of Neoproterozoic Continental Rifting, Glaciation, and the Opening and Closing of the Iapetus Ocean, Blue Ridge, VA–NC–TN

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ABSTRACT

Recent field and geochronological studies in eight 7.5-minute quadrangles near Mount Rogers in Virginia, North Carolina and Tennessee recognize important stratigraphic and structural relationships for the Neoproterozoic Mount Rogers and Konnarock formations, the northeast end of the Mountain City window, the separation of Mesoproterozoic rocks of the Blue Ridge into three age groups, and timing and emplacement of the Blue Ridge thrust sheet. The study area includes folded and faulted Paleozoic strata of the Valley and Ridge to metamorphic and igneous rocks of the Blue Ridge. In the Valley and Ridge, Cambrian to Middle Ordovician carbonate and clastic rocks are exposed in a syncline on the Pulaski thrust sheet; these rocks are overridden by the Blue Ridge thrust sheet. The northeast end of the Mountain City window is interpreted as a simple window; the Stone Mountain fault is folded and continues as the Iron Mountain fault on the NW-side of the window. The Stone Mountain fault does not exist to the NE near the Razor Ridge volcanic center. Instead a continuous section of Proterozoic gneisses, Mount Rogers Formation, Konnarock Formation and Chilhowee Group is now recognized.

Rhyolites of the Mount Rogers Formation range from 760–749Ma, with detrital zircon age populations from associated volcaniclastic rocks indicating magmatism and rifting began by ~780 Ma. Rhyolite blocks in the Konnarock Formation and a change from rift-related clastic rocks of the Mount Rogers Formation transitioning to maroon laminites and laminites with dropstones, suggest that the Konnarock Formation may be as old as ~749 Ma.

Mesoproterozoic crystalline rocks of the Blue Ridge, previously referred to as the Cranberry Gneiss, are separated based on field relationships and SHRIMP U–Pb geochronology: (1) pre-Grenvillian crust,1.33 Ga; (2) 1190–1140 Ma granitoids; and (3) 1075–1030 Ma granitoids.

Multiple greenschist-facies high-strain zones, including the 2–11 km wide Fries high-strain zone, occur in the Blue Ridge thrust sheet. Fabrics across the Fries and Gossan Lead faults have similar orientations and NW–directed contractional deformation. ^{40}Ar/^{39}Ar hornblende, muscovite, and K-feldspar ages indicate the western and eastern Blue Ridge had different thermal histories. The eastern Blue Ridge (Gossan Lead thrust sheet) experienced a 360–340 Ma amphibolite facies event prior to juxtaposition with the western Blue Ridge. ^{40}Ar/^{39}Ar muscovite ages in western Blue Ridge rocks document greenschist facies metamorphism and deformation and emplacement of the Blue Ridge thrust sheet at ~340 Ma; the Catface and Fries faults are tentatively interpreted to be contemporaneous. After initial emplacement of the Blue Ridge thrust sheet at ~340 Ma, shortening was accommodated by westward translation along the basal decollement, which carried the Blue Ridge thrust sheet to its final position.

Merschat, A.J., Southworth, S., Holm-Denoma, C.S., and McAleer, R.J., 2016, Geology of the Mount Rogers area, revisited: Evidence of Neoproterozoic Continental Rifting, Glaciation, and the Opening and Closing of the Iapetus Ocean, Blue Ridge, VA–NC–TN, *in* Merschat, A.J., ed., Geology of the Mount Rogers area, revisited, Blue Ridge, VA–NC–TN: Carolina Geological Society Annual Field Trip Guidebook, p. 3–28.

INTRODUCTION

Forty-nine years ago Douglas W. Rankin led the Carolina Geological Society on its first visit to the Mount Rogers area, VA–NC–TN (Rankin, 1967), and introduced the society to the spectacular and interesting geology. His mapping and research into geology of the Mount Rogers area became the foundation for many subsequent papers and maps (Rankin, 1970, 1975, 1993; Rankin et al., 1972, 1973, 1993) and field trips (Rankin, 1967, 1971; Rankin et al., 1994) on the tectonic history of the southern Appalachians, and recent USGS mapping and associated studies (Fig. 1). The goal of the 2016 Carolina Geological Society fieldtrip is to provide an updated overview of the geology in the Mount Rogers area from new information gleaned from recent field studies, modern geochronology and geochemistry. We will discuss the diversity of Mesoproterozoic rocks formerly included as the Cranberry Gneiss, the age and nature of Neoproterozoic volcanism associated with the Mount Rogers Formation, the age and significance of the glaciogenic Konnarock Formation, the rifting and development of the Laurentian Early Paleozoic margin, the structure of the northeast end of the Mountain City window, various high-strain zones (Fries and Gossan Lead faults), and the basement-cover contact of the eastern Blue Ridge.

Fieldwork in the Mount Rogers area began about 70 years ago with geologic investigations by Anna Jonas (e.g., Jonas and Stose, 1939; Stose and Stose, 1957). This work was followed by Doug Rankin in the early 1960's as part of detailed geologic mapping of the volcanic rocks on Mount Rogers, and reconnaissance work for the west-half of the Winston-Salem 1° x 2° sheet (Rankin et al., 1972). Recent field work and associated studies have focused on eight 7.5-minute quadrangles in Virginia, North Carolina and Tennessee (Fig. 2). Doug Rankin's detailed mapping, unpublished field sheets, and detailed descriptions have been the foundation on which present USGS research and academic partners build upon.

On Day 1 we will examine an overview of the geology of the Mount Rogers area: beginning in the oldest rocks, Mesoproterozoic basement gneisses, and progressing stratigraphically upwards through Neoproterozoic to Cambrian cover sequences (Mount Rogers, Konnarock and Unicoi formations), ending with evidence of the opening of the Iapetus ocean. During Day 1 we will discuss the evidence for different Mesoproterozoic orogenies, Neoproterozoic rifting and glaciations, the allochthonous nature of the Blue Ridge thrust sheet, its relationship to the Valley and Ridge, and the Late Neoproterozoic to Early Cambrian rifting of Laurentia and development of a passive margin. Day 2 will focus primarily on the eastern Blue Ridge: its relationship to the western Blue Ridge, and the lithostratigraphy and units within the Ashe Formation[1] (Ashe Metamorphic Suite of Abbott and Raymond, 1984).

GEOLOGIC SETTING

The Blue Ridge is allochthonous; a large crystalline thrust sheet—the Blue Ridge thrust sheet—stacked on top of Paleozoic sedimentary rocks of the Valley and Ridge during the late Paleozoic Alleghanian orogeny (Fig. 1). The composite Blue Ridge thrust sheet is composed of internal crystalline thrust sheets of Mesoproterozoic to Paleozoic rocks that are more penetratively deformed and higher metamorphic grade toward the hinterland or geographic east (i.e., Hatcher, 1989; Hatcher and Goldberg, 1991; Rankin, 1993). The thrust sheets contain rocks with different protoliths and tectonic histories: the western Blue Ridge represents part of the Laurentian margin (e.g., Rankin, 1975; Hatcher et al., 2007a), whereas the tectonic setting of eastern Blue Ridge rocks are still debated, but are generally considered to be related to closing of the Iapetus ocean (Fig. 1)(e.g., Rankin, 1975; Abbott and Raymond 1984; Horton et al., 1989; Bream et al., 2004; Hatcher et al. 2007a; Merschat et al., 2010; Carter and Merschat, 2014, this guidebook).

The composite Blue Ridge thrust sheet is commonly described as an antiformal thrust stack (e.g., Hatcher and Goldberg, 1991). In northwestern North Carolina and southwest Virginia, the composite Blue Ridge thrust sheet is comprised of several individual thrust sheets (from northwest to southeast): Shady Valley–Stone Mountain, Catface, Fries, and Gossan Lead thrust sheets (Figs. 1 and 2). Except for the Gossan Lead thrust sheet, all of these thrust sheets involve western Blue Ridge rocks. The Gossan Lead thrust sheet contains higher grade and more intensely deformed rocks of the eastern Blue Ridge.

From southeastern Tennessee into Virginia the base of the Blue Ridge thrust sheet is the Great Smoky–Iron Mountain–Holston Mountain faults, part of the Blue Ridge thrust system (Fig. 1) (Hatcher, 1989; Woodward, 1989; Rankin, 1993). The western edge of the Blue Ridge thrust sheet is commonly imbricated, and several windows occur along its leading edge. Prominent windows include the Mountain City window, Hot Springs window, and the Grandfather Mountain window (Fig. 1). These structures sole into a master fault beneath the Blue Ridge thrust sheet that extends eastward beneath the terranes of the Blue Ridge and Piedmont, and eventually terminates into a root zone beneath the Piedmont and Coastal Plain (Harris et al., 1981; Hatcher, 1989; Rankin et al., 1991; Hatcher and Hooper, 1992).

The Mountain City window is framed by the Stone Mountain and Catface faults on the southeast and the Iron Mountain fault at the base of the Shady Valley thrust sheet to the northwest (King et al., 1944; King and Ferguson, 1960). Duplexes within the window repeat sections of the Chilhowee Group, Shady Dolomite, and Rome Formation (King et al., 1944; King and Ferguson, 1960). The roof thrust of the window is the base of the Blue Ridge thrust sheet; it reemerges to the southeast to frame the Grandfather Mountain window (Bryant and Reed, 1970; Boyer and Elliot, 1982).

The western Blue Ridge consists of Neoproterozoic to Early Cambrian sedimentary rocks deposited unconformably on Mesoproterozoic rocks of the French Broad massif. The study area is within the northern part of the French Broad massif before it is overridden by the Fries fault (Fig. 1). The Cranberry Gneiss was the name previously assigned to much of the granitic rocks

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[1] * U.S. Geological Survey usage for this unit is formation, as defined by Rankin (1970) and Rankin et al. (1973). The Ashe Formation was elevated in rank to Metamorphic Suite by Abbott and Raymond (1984) and redefined by Raymond (2015).

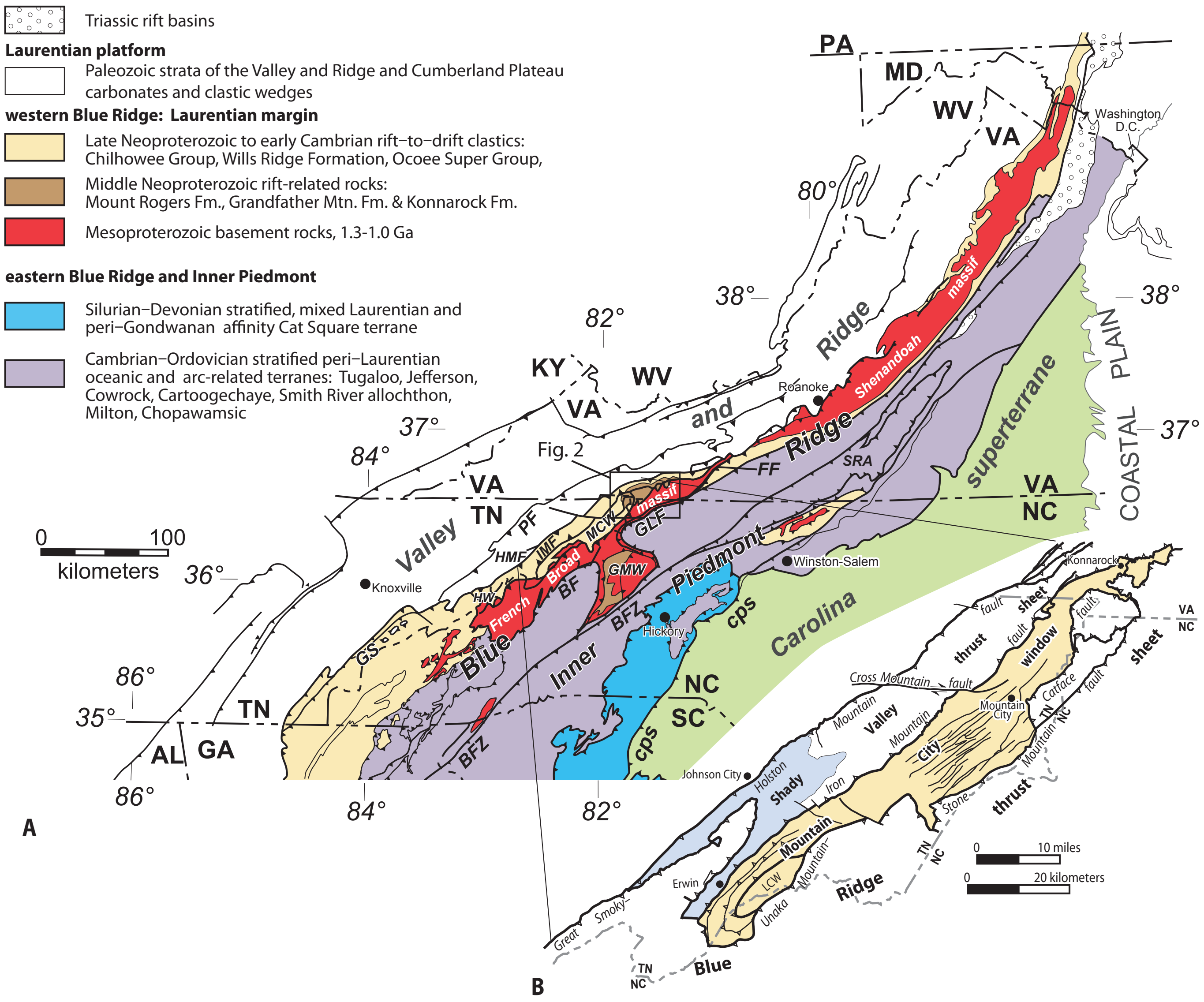

Figure 1. (A) Tectonic map of the southern and central Appalachians modified from Rankin (1993) and Hatcher et al. (2007a). (B) Inset map of the Mountain City window modified from King et al. (1944), King and Ferguson (1960), Rankin (1993), and Hatcher et al. (2006). Light blue indicates the distribution of Cambrian to Ordovician carbonates in the core of the Stony Creek syncline in the Shady Valley thrust sheet. BF–Burnsville fault, BFZ–Brevard fault zone, CPS–central Piedmont suture, FF–Fries fault, GMW–Grandfather Mountain window, GLF–Gossan Lead fault, HMF–Holston Mountain fault, HW–Hot Springs window, IMF–Iron Mountain fault, LCW–Limestone Cove window, MCW–Mountain City window, PF–Pulaski fault, SRA–Smith River allochthon.

exposed in the northern end of the French Broad massif (Keith, 1903; Rankin et al., 1972), although recent work has demonstrated the complexities and variations in the Mesoproterozoic basement (Carrigan et al., 2003; Merschat and Cattanach, 2008; Tollo et al., 2010, 2012). In southwestern Virginia and into North Carolina and Tennessee, the Mesoproterozoic basement is overlain by Neoproterozoic clastics and felsic volcanics of the 760–749 Ma Mount Rogers Formation and glaciogenic sediments of the Konnarock Formation (Rankin, 1993; Tollo et al., 2012, Merschat et al., 2014). In northwestern North Carolina, the Grandfather Mountain Formation is a similar sequence of Neoproterozoic clastics but contains considerably less felsic volcanic rocks (Bryant and Reed, 1970). The Mount Rogers and Grandfather Mountain formations, including both sedimentary and volcanic components, are part of the Crossnore Volcanic-Plutonic complex, which also includes the anorogenic plutonic rocks of the Beech Mountain Granite, Brown Mountain Granite, Striped Rock Pluton and Bakersville Gabbro and associated dikes (Rankin et al., 1993). Collectively, these units represent a Neoproterozoic (Cryogenian) episode of rifting in the Blue Ridge (Rankin, 1993; Aleinikoff et al., 1995). The Neoproterozoic to Cambrian Chilhowee Group unconformably overlies both Mesoproterozoic basement rocks and rift-related Neoproterozoic rocks. The Chilhowee Group was deposited during the opening of the Iapetus ocean and development of an early Paleozoic passive margin.

The structurally higher eastern Blue Ridge (Fig. 1) reflects a different depositional and tectonic history. Eastern Blue Ridge rocks are polydeformed, amphibolite-facies (garnet to kyanite zones), Neoproterozoic to Paleozoic siliciclastic metasedimentary rocks that contain mafic and ultramafic rocks, and are intruded by Paleozoic granitoids (Rankin et al., 1972; Abbott and Raymond, 1984; Rankin, 1993; Hatcher et al.,

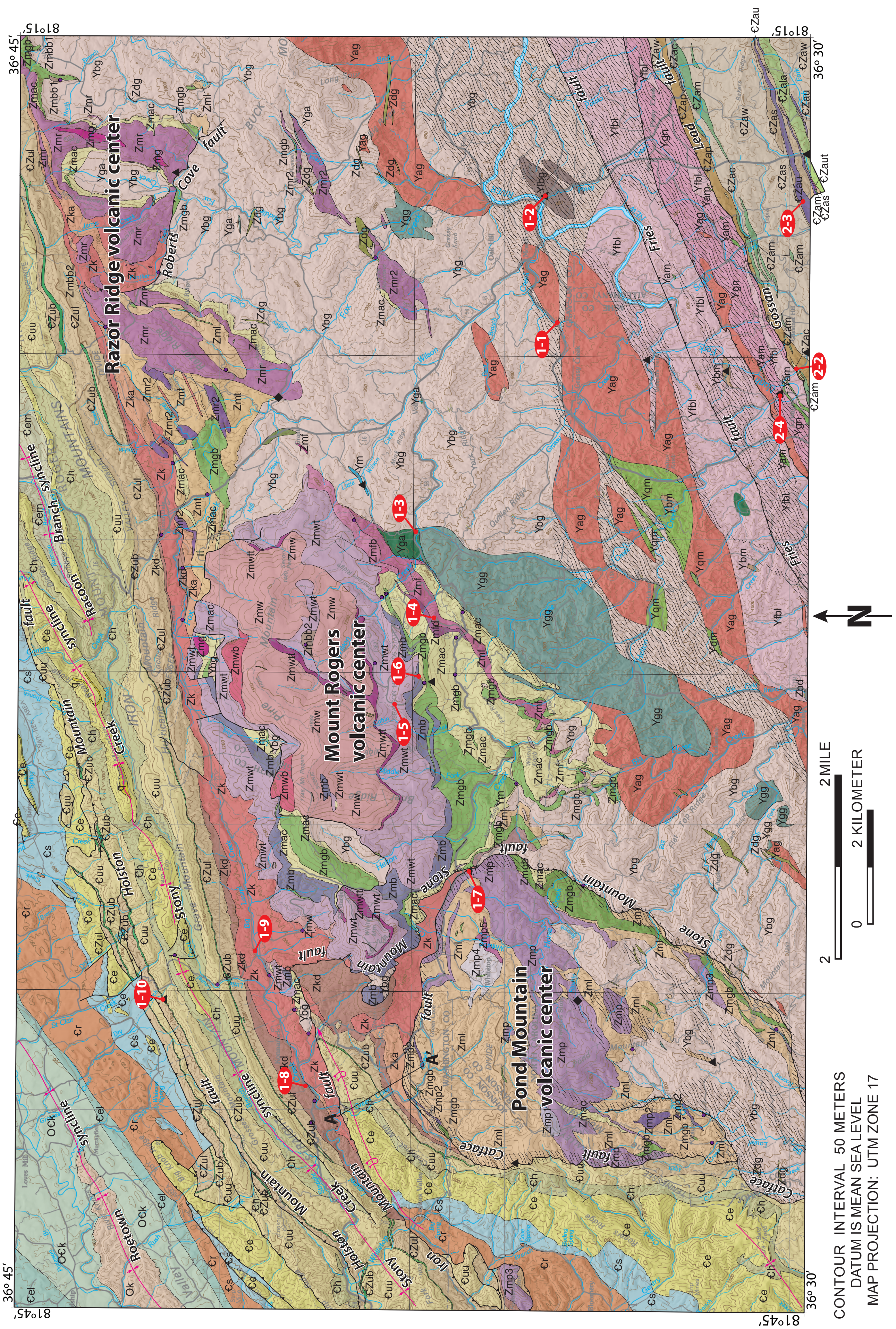

Figure 2. (A) Geologic map of the Mount Rogers area showing 2016 Carolina Geological Society Field trip stops. *(Continued on facing page).*

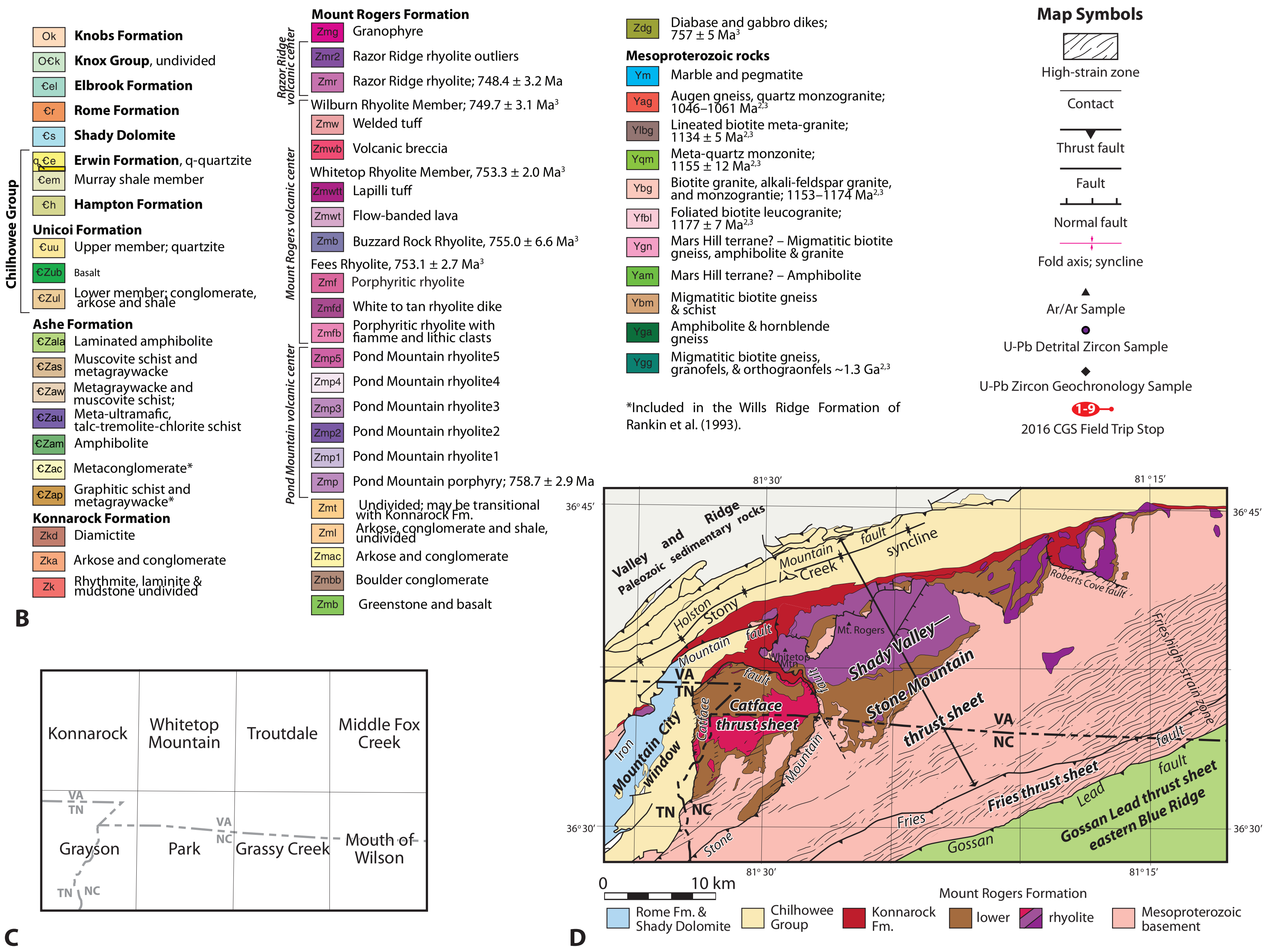

Figure 2 *(Continued)*. (B) Explanation of map units. (C) Eight 7.5–minute quadrangles that are the focus of study by the USGS and this field trip. (D) Simplified tectonic map of the Mount Rogers area showing the various thrust sheets and distribution of the Mount Rogers and Konnarock formations Tectonic map displays a larger area than part (A).

2007a). Rock units include the Wills Ridge, Ashe, Alligator Back, and Lynchburg formations (Stose and Stose, 1957; Rankin et al., 1972; Rankin, 1975; Rankin et al., 1993), or the Ashe Metamorphic Suite (Abbott and Raymond, 1984; Raymond, 2015) and Alligator Back Metamorphic Suite (Raymond et al., 1989; Raymond, 2015). These rocks have been interpreted as offshore equivalent deposits of the western Blue Ridge (Rankin, 1975), accretionary wedges (Abbott and Raymond, 1984), and as part of tectonic terranes of oceanic and exotic craton affinities (Williams and Hatcher, 1982; Abbott and Raymond, 1984; Horton et al., 1989; Hatcher and Goldberg, 1991; Hatcher et al., 2007a). A more in depth review and discussion of eastern Blue Ridge lithostratigraphy and tectonogenesis can be found in Carter and Merschat (this guidebook).

REGIONAL STRUCTURES

Mountain City window and Blue Ridge thrust sheet

The Mountain City window is more than 100 km (62 mi) long, from near Konnarock, Virginia, to south of Erwin, Tennessee, and is over 15 km (9.5 mi) wide (Fig. 1B). The internal structure consists of two oppositely verging duplexes: the southern end is the foreland-dipping duplex Limestone Cove culmination; the Doe Ridge culmination is a hinterland-dipping duplex in the central part that ends near Mountain City, Tennessee (Diegel, 1986). The imbricate thrusts within the Mountain City window merge into the Stone Mountain–Iron Mountain–Holston Mountain fault, the roof thrust and base of the Blue Ridge thrust sheet, which reemerges around the Grandfather Mountain window (Bryant and Reed, 1970; Boyer and Elliot, 1982). The floor thrust beneath the window is likely the Pulaski thrust (Boyer and Elliot, 1982; Woodward, 1989), which crops out 15–25 km to the northwest.

The northeast termination of the Mountain City window has been portrayed differently by various workers (Fig. 3). King and Ferguson (1960) connected the Catface and Iron Mountain faults. Stratigraphic and structural relationships led Rankin (1967, 1993) to map the Catface fault around the Pond Mountain area to the east, where it was overridden by the Stone Mountain fault. Bailey and Rose (1998) illustrate a simplified version of Rankin's (1993) eyelid window geometry, with the Stone Mountain fault located further to the southeast, but their position of the Catface fault is not consistent with field relations (e.g., Rankin, 1967, Rankin, 1993).

Mapping in the Razor Ridge area (Trout Dale and Middle Fox Creek quadrangles; Fig. 2), supports a continuous flat-lying to northwest-dipping stratigraphic section from basement to the Chilhowee Group, with no compelling evidence of the Stone Mountain and Trout Dale faults (Fig. 2; Merschat and Southworth, 2011). Thus the Stone Mountain-Catface fault merge with the Iron Mountain fault, and the Shady Valley thrust sheet is part of the Blue Ridge thrust sheet. Ordovician rocks in the southwestern core of the Stony Creek syncline in the Shady Valley thrust sheet are part of the Blue Ridge thrust sheet (Figs. 1B).

The rocks in the northeasternmost end of the Mountain City window terminate in an overturned, antiformal syncline (060/20). The overturned west limb of the syncline is overridden by the Iron Mountain fault, which frames the window on the west (Fig. 4). Rocks of the Cambrian Rome Formation and Shady Dolomite plunge northeast under older rocks of the Chilhowee Group and Konnarock Formation in the hinge of the fold. Reclined mesoscopic folds have a similar geometry and orientation as map-scale folds (Fig. 4A). King and Ferguson (1960) recognized similar overturned folds and faults near Forge Mountain, east of Mountain City, Tennessee, where quartzites of the Erwin Formation overlie Shady Dolomite beneath the Stone Mountain fault (Fig. 4C).

A problem not yet resolved is the relationship of the Stone Mountain and Catface faults. At their type locations, both faults occupy the same structural position and frame the southeast side of the Mountain City window (King et al., 1944; King and Ferguson, 1960). King and Ferguson's (1960) Stone Mountain fault is structurally above the Catface fault, similar to that portrayed by Rankin (1993) (Fig. 3). Bailey and Rose (1998) portrayed the Stone Mountain fault on the southeast side of the southwestern most occurrence of Mount Rogers Formation in the Park and Grayson quadrangles (Fig. 3). Both basement and Mount Rogers Formation rocks are penetratively deformed and cut by numerous high-strain zones, an intensity of deformation not characteristic of the internal parts of the Mountain City window. At this point it is not clear if this deformation is related to the Stone Mountain fault or high-strain zones similar to the Fries high-strain zone. However, the location of the Stone Mountain fault in Figure 3 is similar to that of Rankin (1993) and Bailey and Rose (1998).

Fries fault

Stose and Stose (1957) defined the Fries fault to the east of our study area, where several discrete fault strands of basement and cover rocks are juxtaposed. Recent mapping has shown that high-strain zones occur throughout the western Blue Ridge from the Cat Face and Stone Mountain faults to the western edge of the eastern Blue Ridge, Gossan Lead fault (Fig. 2; Merschat, 2011; Tollo et al., 2012, Merschat et al., 2014). Multiple high-strain zones as much as 11 km wide comprise a major crustal-scale shear zone that may sole into the base of the Blue Ridge thrust sheet. These high-strain zones are characterized by greenschist facies deformation, down-dip lineations, and top-to-NW kinematic indicators (Merschat, 2011; Tollo et al., 2012; Merschat et al., 2014).

WESTERN BLUE RIDGE LITHOSTRATIGRAPHY

Mesoproterozoic Rocks

The Cranberry Gneiss and variations thereof (e.g., Cranberry Granite of Keith, 1903; Cranberry Suite of Bartholomew and Lewis, 1984) have been used to describe most of the Proterozoic felsic granitoids and gneisses of the northern end of the French Broad massif (Rankin et al., 1972). Recent mapping and geochronologic studies in the Mount Rogers area (Tollo et al., 2010, 2012) and to the southwest (Carrigan et al., 2003; Merschat and Cattanach, 2008) have documented that the Cranberry Gneiss contains various lithologies that range in age from 1.33–1.0 Ga. Tollo et al. (2010, 2012) distinguished three

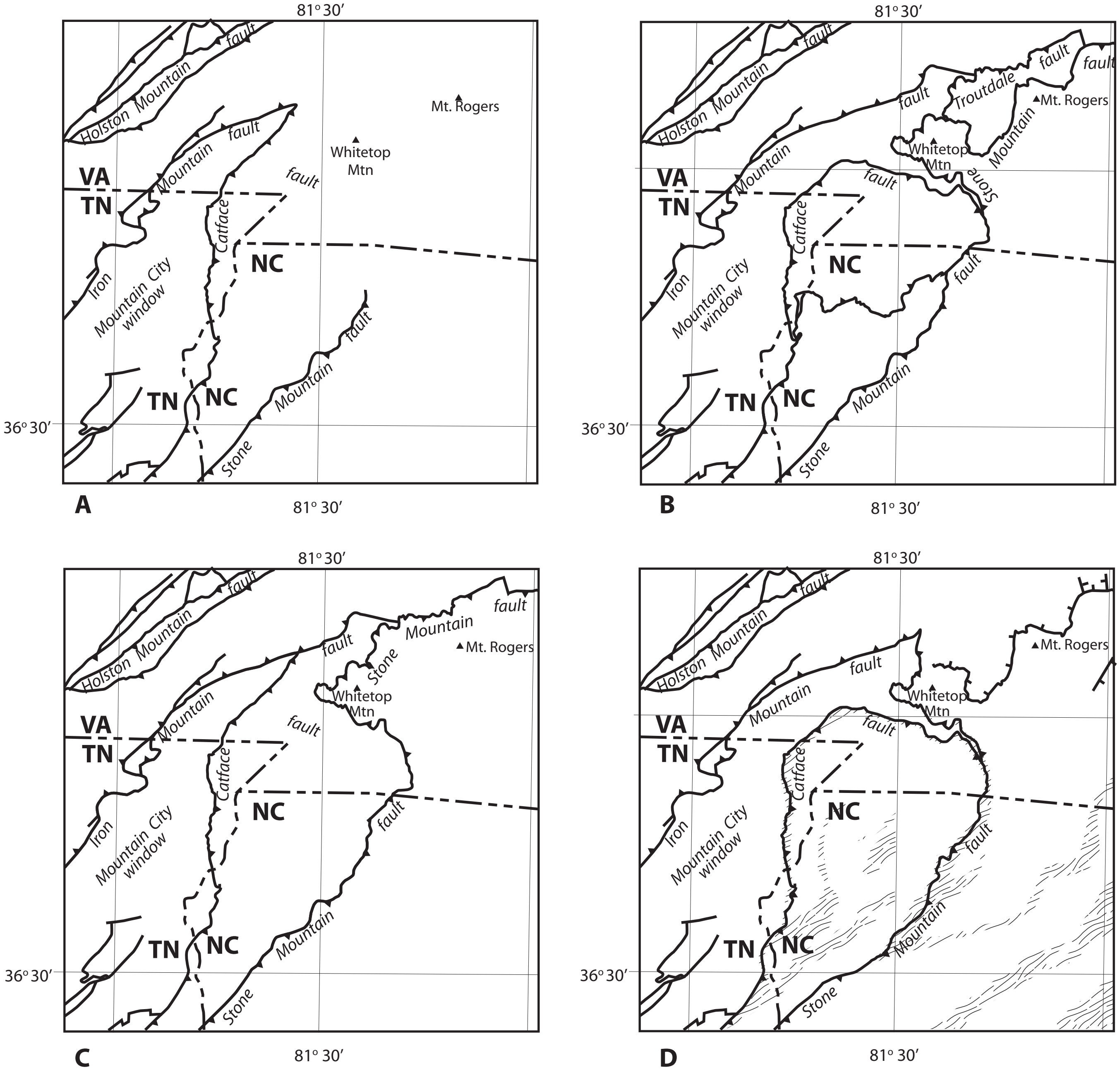

Figure 3. Comparison of different fault geometries of the northeast end of the Mountain City window. (A) King and Ferguson (1960), (B) Rankin (1993), (C) Bailey and Rose (1998), and (D) this study. Pattern of wavy lines shows distribution of high-strain zones, see Figure 2.

age groups of Mesoproterozoic rocks based on field relationships and SHRIMP U-Pb geochronology: (1) pre–Grenvillian crust, ~1.32 Ga migmatitic orthogranofels and amphibolite; (2) 1.19–1.14 Ga granitoids (early magmatic suite); and (3) 1.07–1.03 Ga granitoids (late magmatic suite) (Fig. 2; Table 1). Similar ages of rocks are recognized in the Shenandoah massif (Southworth et al., 2010; Carter et al., 2012, 2013) and southern part of the French Broad massif (Carrigan et al., 2003). These age relationships are temporally correlative with Shawinigan and Ottawan phases of the Grenville orogeny, respectively (McLelland et al., 2010, 2013; Tollo et al., 2010, 2012).

The oldest rocks are ~1.33 Ga orthogranofels and migmatitic amphibolites (Tollo et al., 2010, 2012). The orthogranofels have layers of hornblende–actinolite–chlorite–epidote and quartz–K-feldspar (Tollo et al., 2010; 2012). Amphibolite gneiss occurs as smaller bodies completely enclosed by early and late magmatic rocks, xenoliths. Although no protolith ages have been obtained for the amphibolites, their occurrence as xenoliths permits correlation with the ~1.3 Ga orthogranofels (Tollo et al., 2012). The migmatitic granofels consist of a large NW-trending body in the Grassy Creek quadrangle and small NW-trending bodies in the southern part of the Park quadrangle (Fig. 2). Although locally containing retrograde epidote and K-feldspar, the migmatitic granofels preserve a gneissic foliation defined quartz–plagioclase–sillimanite–biotite assemblage (Fig. 5). Collectively, these rocks represent pre-Grenville crust intruded by the 1.18–1.14 and 1.07–1.03 Ga magmatic suites.

The majority of basement rocks in the Mount Rogers area yields magmatic ages from 1.19–1.14 Ga, and are grouped as the early magmatic suite (Tollo et al., 2010, 2012). The

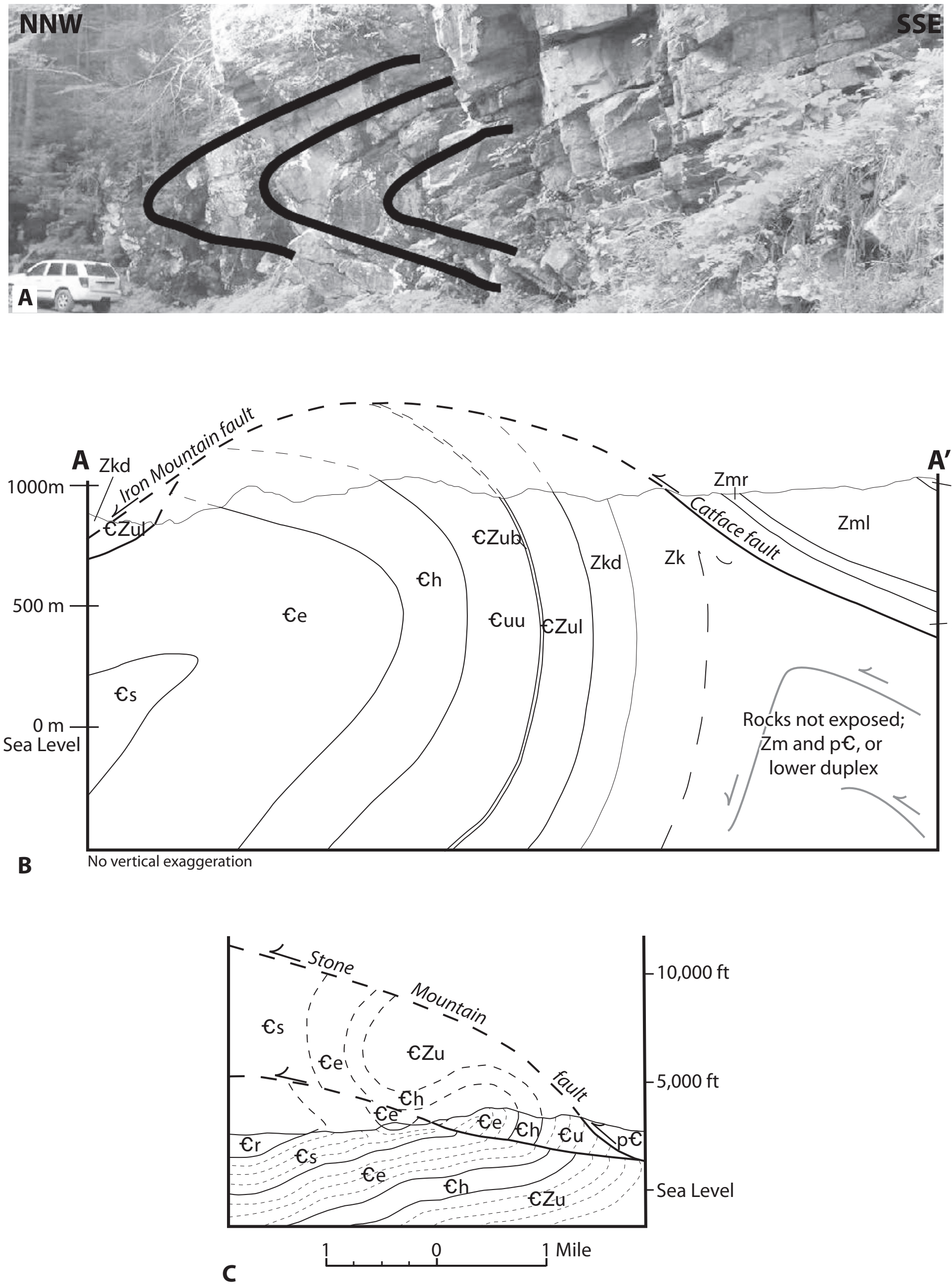

Figure 4. Mesoscale and map–scale examples of overturned fold from the northeastern end of the Mountain City window. (A) Reclined fold in quartzites of the upper Unicoi Formation. (B) Cross section through the northeast end of the Mountain City window following VA Route 859. Location of cross section is shown on Figure 2. (C) Structural interpretation by King and Ferguson (1960) of recumbent folds in the Forge Mountain, near Mountain City, TN. Abbreviations of map units are pꞒ—Mesoproterozoic granitoids; Zml—lower Mount Rogers Formation; Zm—Mount Rogers Formation rhyolite; Zk—Konnarock Formation undivided; Zkd–Konnarock Formation diamictite; ꞒZu—Unicoi Formation undivided; ꞒZul—lower Unicoi Formation; ꞒZub—Unicoi Formation basalt; ꞒZuu—upper Unicoi Formation; Ꞓh—Hampton Formation; Ꞓe—Erwin Formation; Ꞓs—Shady Dolomite.

rocks generally have a bulk composition of granite but have considerable lithologic and textural variation (Tollo et al., 2012). The dominant lithology is a variably foliated coarse-grained, equigranular to porphyritic, biotite monzogranite. Pegmatitic and aplitic dikes, and ~750 Ma mafic and rhyolite dikes (Tollo et al., 2012) intruded the early magmatic suite.

The late magmatic suite, 1.07–1.03 Ga, is bimodal and ranges from syenogranite to monzodiorite (Tollo et al., 2010, 2012). The syenogranite is commonly megacrystic to porphyroclastic with blocky to rounded, centimeter–sized alkali feldspar and 3–5 mm blue quartz phenocrysts in a strongly foliated matrix of biotite, muscovite (sericite) and minor chlorite (see Stop 1–1). They are moderately to strongly foliated, and commonly mylonitic; the result of Paleozoic deformation.

Mount Rogers Formation

The 760–749 Ma Mount Rogers Formation consists of a lower clastic sedimentary sequence interbedded with basalt, and an upper sequence dominated by rhyolites. The rocks were first described by Jonas and Stose (1939) but carefully studied and defined by Rankin (1993). The Mount Rogers Formation

Table 1. Compilation of U-Pb zircon ages of Mesoproterozoic rocks from the northern French Broad massif.

Name or lithologic description	Label	Crystallization Age (Ma)	Metamorphic Age (Ma)	Ref.
Late Magmatic				
Biotite meta-quartz monzonite and quartz monzodiorite	Yag	1046 ± 14	1055 ± 11	2
biotite meta-quartz monzodiorite	Yag	1055 ± 5	~990	2
prophyroclastic metagranite	Yag	1061 ± 5	1053 ± 5	2
Blowing Rock Gneiss		1081 ± 14		1
Early Magmatic				
lineated biotite metagranite	Ylbg	1140 ± 9	~1050 & 997 ± 12	2
porphyroclastic biotite metagranite	Ylbg	1134 ± 5	~1050 & 984 ± 18	3
meta-alkali feldspar granite	Ybg	1153 ± 8		3
meta-alkali feldspar granite	Ybg	1162 ± 4	1157 ± 6 & ~1050	2
meta-monzograntie	Ybg	1161 ± 7	1156 ± 10 1117 ± 6 & 989 ± 9	2
mata-quartz monzonite	Yqm	1155 ± 12	1.0-1.1 Ga	3
Watauga River Gneiss		1158 ± 9		1
alkali feldspar meta-leucogranite	Ybg	1166 ± 9	~1.1 Ga	3
meta-monzogranite (med. Grained)	Ybg	1168 ± 10	0.9-1.0 Ga	3
meta-monzogranite (porphyritic)	Ybg	1172 ± 10		3
meta-leucogranite	Ybg	1174 ± 7	1158 ± 7 & 1031 ± 12	2
foliated meta-leucogranite	Yfbl	1177 ± 7	~1160 & ~1035	3
amphibole-bearing granitic gneiss		1177 ± 15	1180-1100	3
nonfoliated amphibole meta-leucogranite		1183 ± 18	1.18 Ga & 1.1-1.05	3
Cranberry Gneiss (type locality)		1192 ± 11		1
pre-Grenville				
layered amphibolite	Yga		1145 ± 5 & ~950	3
medium-grained amphibolite	Yga		1164 ± 6 & 1148 ± 9 & 1040	2
orthogranofels	Ygg	1327 ± 7	1140-1060	2

References cited: 1—Carrigan et al., (2003); 2—Tollo et al., (2010); 3—Tollo et al., (2012)

unconformably overlies Mesoproterozoic basement, and is overlain by the Konnarock Formation. The formation reaches a thickness of about 3000 m (9,843 ft), and occurs in three separate areas (northeast to southwest): Razor Ridge, Mount Rogers and Pond Mountain volcanic centers of Rankin (1993) (Fig. 2). Of the named volcanic centers, only the Mount Rogers volcanic center has been studied and described in detail (Rankin, 1993; Novak and Rankin, 2004).

The lower part of the formation consists of conglomerate, arkose, shale, basalt, and rhyolite. Regionally there are spatial variations in the thickness and lithologic character. Detrital zircon studies show that clastic rocks of the Mount Rogers Formation are dominated by 1160, 1050 and 760 Ma sources, with minor components of 1.4–1.2 Ga (Holm-Denoma et al., 2013, 2014; see discussion below). These studies show that the provenance of the Mount Rogers Formation was local, with sediment derived from Blue Ridge Mesoproterozoic rocks and cannibalized volcanic deposits. Basaltic rocks vary texturally from amygdaloidal, porphyritic to aphanitic. The upper part is dominated by ~610 m (2000 ft) of characteristically purple to maroon porphyritic rhyolites (Rankin, 1993). Phenocrysts usually consist of quartz, alkali feldspar, and plagioclase, but the size, abundance, and assemblage vary. Rare accessory minerals include aegirine, sodic amphibole (riebeckite), biotite, and fluorite (Rankin 1975, 1993; Novak and Rankin, 2004). Chemically the rocks are ferroan, peraluminous to metaluminous, high silica (74–78 wt. %) rhyolites (Novak and Rankin, 2004; Tollo et al., 2012). Trace-element data indicate

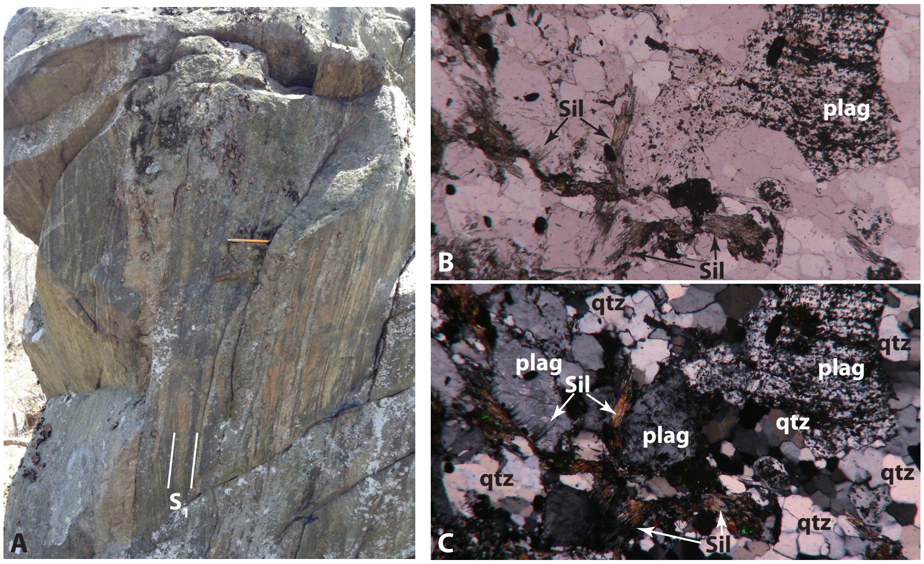

Figure 5. (A) Outcrop of migmatitic granofels on Zach Ridge, Park quadrangle, displaying S_1 gneissic foliation trending NW–SE. Pencil for scale is 14.5 cm. Photomicrographs of migmatitic granofels in plane polarized (B) and cross-polarized (C) light with the prograde assemblage of quartz-plagioclase-sillimanite-biotite. Plagioclase contains inclusions of epidote giving it a clouded appearance. Field of veiw is 2.5 mm. Abbreviations: qtz—quartz; plag—plagioclase; sil—sillimanite.

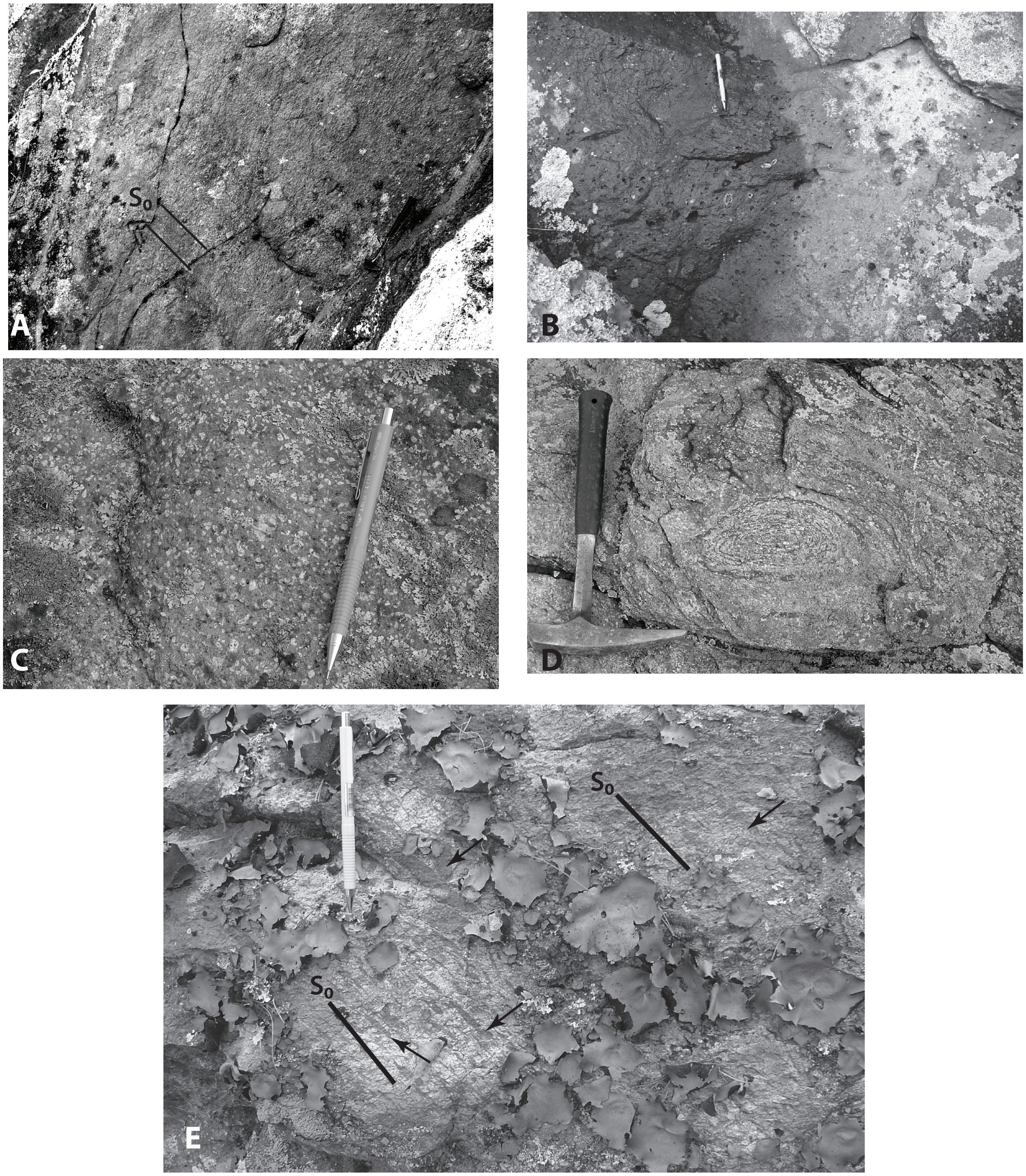

Figure 6. (A) Massive coarse-grained, conglomeratic arkose of the lower Mount Rodgers Formation from Mikes Ridge (Middle Fox Creek quadrangle). Clasts are sub-angular to rounded, pebble– to coble–sized quartz, microcline and some coarse–grained, biotite granite. Bedding is faint, graded and dips moderately to the northwest. (B) Vesicular, amygdaloidal basalt flow in the lower Mount Rogers Formation rest directly on basement and is about 60 m (~200 ft) thick. Porphyritic rhyolite with 1.5 cm long alkali feldspar phenocrysts (C) and concentric jasperoid rings (D) from Razor Ridge rhyolite on Bald Rock. (E) Abundant fiamme, 1-10 cm long, define volcanic compaction layering in porphyritic rhyolite on Newland Ridge. Fiamme are darker than surrounding rhyolite and several are marked with the small black arrows. Pencil is 14.5 cm long; hammer is 32.5 cm long.

the Mount Rogers rhyolites are derived from A-type, intraplate magmas, like the granitoids of the Crossnore Complex (Rankin, 1975; Novak and Rankin, 2004; Tollo et al., 2012). McClellan and Gazel (2014) noted that the greenstone, basalts and rhyolites of the Mount Rogers Formation have a geochemical signature of mantle-derived melts. Thin beds of lithic wacke, arkose, and conglomerate are interbedded with the rhyolites. Foliated diabase dikes cut the youngest rhyolites (Rankin, 1967).

Field Relationships

Mount Rogers volcanic center

The field and stratigraphic relationships described herein are from Rankin (1993). Rhyolite units include (1) porphyritic, phenocryst-rich Fees Rhyolite in the lower Mount Rogers; (2) the low-silica, porphyritic Buzzard Rock Rhyolite Member; (3) phenocryst-poor Whitetop Rhyolite Member; and (4) the uppermost Wilburn Rhyolite Member, a porphyritic welded tuff (Rankin, 1993). Additional work has focused on the southeastern boundary of the volcanic center and the lower Mount Rogers, Fees Rhyolite and basement contacts (James, 1999; Yonts et al., 2011; Jessee et al., 2012; McClellan et al., 2012; Tollo et al., 2012). The Mesoproterozoic basement-lower Mount Rogers Formation contact is inclined to the southeast, with strongly foliated, mylonitic basement, structurally overlying younger Mount Rogers rocks.

Razor Ridge volcanic center

The northeasternmost occurrence of the Mount Rogers Formation, the Razor Ridge volcanic center, consists of clastic rocks, rhyolites and lesser basalt. Rankin (1993) interpreted the rhyolites of this volcanic center to be structurally lower and older than the Mount Rogers volcanic area to the southwest. The Razor Ridge area consists of several separate flat-lying to gently dipping panels of rhyolite and underlying clastic rocks, as well as small outliers of rhyolite interbedded with laminites typical of the Konnarock Formation (Fig. 2) (Merschat and Southworth, 2011). The prominent ridges (Razor, Mike, Newland and Flat) are capped by rhyolite that dips gently northwest and north beneath the Konnarock Formation and Chilhowee Group rocks of the southeast limb of the Stony Creek syncline. Between the panels of rhyolite is a synform of interbedded maroon sandstones, conglomerate and rhythmically laminated mudstone, which are likely the Konnarock Formation; there is no evidence of the Stone Mountain fault as previously mapped (Rankin et al., 1972; Rankin, 1993) (Figs. 2A and 2D).

The basal part of the Mount Rogers Formation varies from 0–300 m (0-984 ft) thick, and consists of arkose, conglomerate (Fig 6A), and purple to gray shale/mudstone. Amygdaloidal and vesicular basalt flows (now greenstones) (Fig. 6B) are common near the base of the formation, including a 60 meter-thick flow with entrained basement boulders near Cinnamon Ridge in the northeast part of the Middle Fox Creek quadrangle (Fig. 2). In other places the nonconformable base of the formation above Mesoproterozoic basement locally is a boulder conglomerate composed almost exclusively of granitoid clasts, coarse-grained arkose, and locally a purple to maroon shale. Paleo-relief on the basement was as much as 100 m, locally cutting out most of the lower clastic part of the formation except locally ~1–2 m of red flinty rock that is likely a hydrothermally altered paleosol beneath rhyolite. Purple to maroon rhyolites are characteristically phenocryst-rich (30–40 percent). Phenocrysts, generally 3–10 mm across, are embayed quartz, alkali feldspar and plagioclase (Fig. 6C–E). Zones of fiamme-rich rhyolite occur locally (Fig. 6E). Petrographic and geochemical studies have distinguished two units that are texturally similar, but one is less chemically evolved and displays variation in silica content (Merschat et al., 2014). Zircon from a sample of porphyritic Razor Ridge rhyolite was analyzed by Sensitive High Resolution Ion MicroProbe (SHRIMP). Ten concordant analyses resulted in a U-Pb Concordia age of 748.8 ± 3.2 Ma (2 sigma-error) (Fig 7), one of the youngest known rhyolite ages from the Mount Rogers Formation.

Pond Mountain volcanic center

Located in the Catface thrust sheet, the Pond Mountain volcanic center is overridden by the Mount Rogers volcanic center along the Stone Mountain fault (Rankin, 1993; Bailey and Rose, 1998). The rocks are penetratively deformed and locally strongly tectonized (Fig. 8). The Catface thrust bounds these rocks and juxtaposes lower clastic rocks and rhyolites against the Chilhowee Group and Konnarock Formation in the Mountain City window (Fig. 2). Although upper contact relationships are excised by faulting and erosion, portions of the basement–cover contact are preserved. Lower clastic rocks nonconformably on basement are overlain by rhyolite. A shallow NW-dipping fault along the basement-Mount Rogers Formation contact, as shown by Rankin (1993), could not be identified (Fig. 2D). The thickest accumulation of Neoproterozoic rocks in this volcanic center is at Pond Mountain. Here, the lower Mount Rogers Formation consists of arkose, pebble to boulder lithic conglomerate, shale (slate), and minor basalt. Distinct purple shale with gray layers, chips, and reduction spots occurs at the base locally. Bailey and Rose (1998) separated three different rhyolitic units and suggested that the Pond Mountain volcanic center may have partially collapsed into its magma chamber. This study as also recognized multiple rhyolites in the Pond Mountain volcanic center (Fig. 2). Rhyolites exposed on Pond Mountain and near Farmers Store, NC (the main body

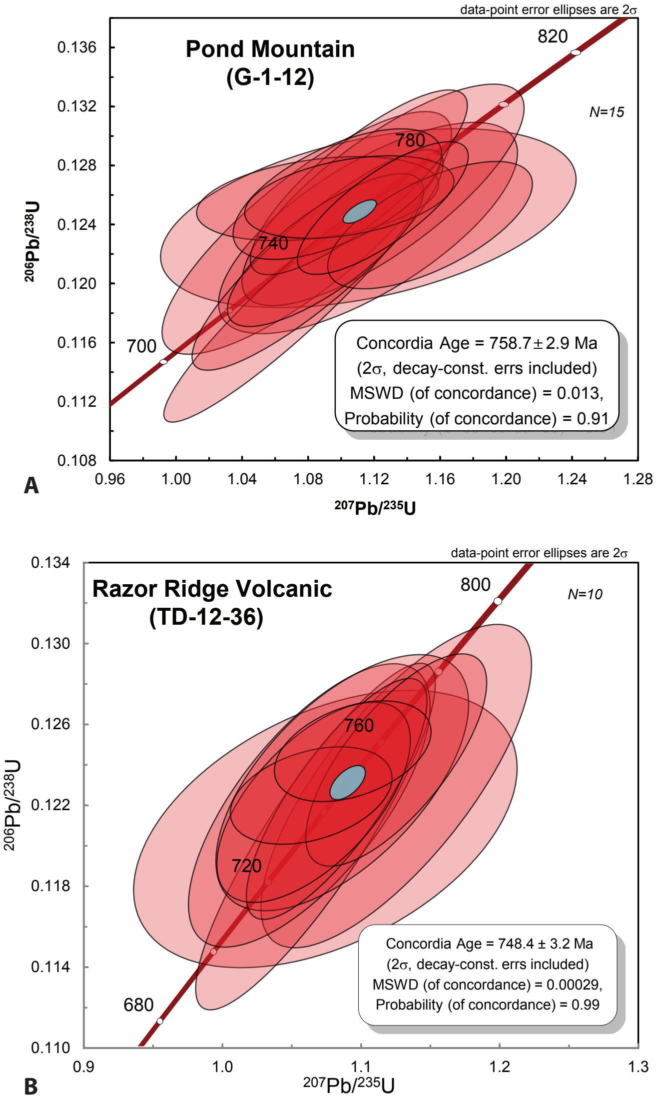

Figure 7. U-Pb Concordia diagrams of laser ablation-inductively coupled plasma-mass spectrometry analyses of zircon from a sample of the Pond Mountain rhyolite porphyry near Framers Store, North Carolina (A), and porphyritic Razor Ridge rhyolite near Grant, Virginia (B).

of rhyolite in the Pond Mountain volcanic center) are purple to maroon, and distinctly porphyritic to megacrystic. Phenocrysts comprise 30–40 percent of the rhyolite and consist of euhedral and embayed quartz, and alkali feldspar as long as 4 cm (Fig. 8A). Plagioclase phenocrysts are rare. Linear bodies of reddish to maroon rhyolite have similar phenocryst assemblages, but are only 0.5–1 cm across, and locally contain lapilli and fiamme (Fig. 8C). The large phenocrysts and mascular cavities suggest that this rock might be a hypabassal intrusive rock (Rankin, 1993) but the deformed lapilli, embayed quartz phenocrysts, and fractured zircon grains (Fig. 8D) all indicate an extrusive origin. The distinctly porphyritic rhyolite from south of Farmers Store, NC was sampled for U-Pb geochronology of zircon by SHRIMP. Fifteen concordant analyses yielded a U-Pb Concordia age of 758.7 ± 2.9 Ma (2 sigma-error) (Fig. 7). The porphyritic, and locally strongly deformed, Pond Mountain rhyolite is the oldest rhyolite of the Mount Rogers Formation; however, detrital zircon data suggest rhyolites as old as ~780 Ma may have existed and were reworked into the clastic deposits (Holm-Denoma et al., 20014; McClellan et al., this guidebook).

Geochronology

Improved analytical techniques have resulted in more precise age constraints of the Mount Rogers Formation (Table 2). Aleinikoff et al. (1995) obtained an age of 758 ± 12 Ma from the Whitetop Member using thermal ionization mass spectrometry. Chemical abrasion thermal ionization mass spectrometry single zircon ages from the defined rhyolite members of the Mount Rogers volcanic center span 6–10 m.y., and the weighted averages with associated errors overlap in age (Tollo et al., 2012; Table 2). New SHRIMP U-Pb zircon ages from the Pond Mountain rhyolite date the oldest rhyolite deposit in the Mount Rogers Formation, 758.7 ± 2.9 Ma (2 sigma-error), while rhyolite from the Razor Ridge, 748.8 ± 3.2 Ma (2 sigma-error), is the youngest rhyolite (Fig. 7; Table 2). Based on the rhyolites ages, the Mount Rogers volcanic center magmatism likely persisted for at least10 m.y. or longer (Tollo et al., 2012), and eruption of the Mount Rogers rhyolites occurred between 760–749 Ma. The sub-volcanic magma chamber of the Mount Rogers Formation, however, may have existed as early as ~772 Ma (oldest zircon analyzed; Tollo et al., 2012). Additionally, SHRIMP and laser ablation-inductively coupled plasma-mass spectrometry U-Pb detrital zircon geochronology of arkoses and conglomerates from the Mount Rogers identified zircons

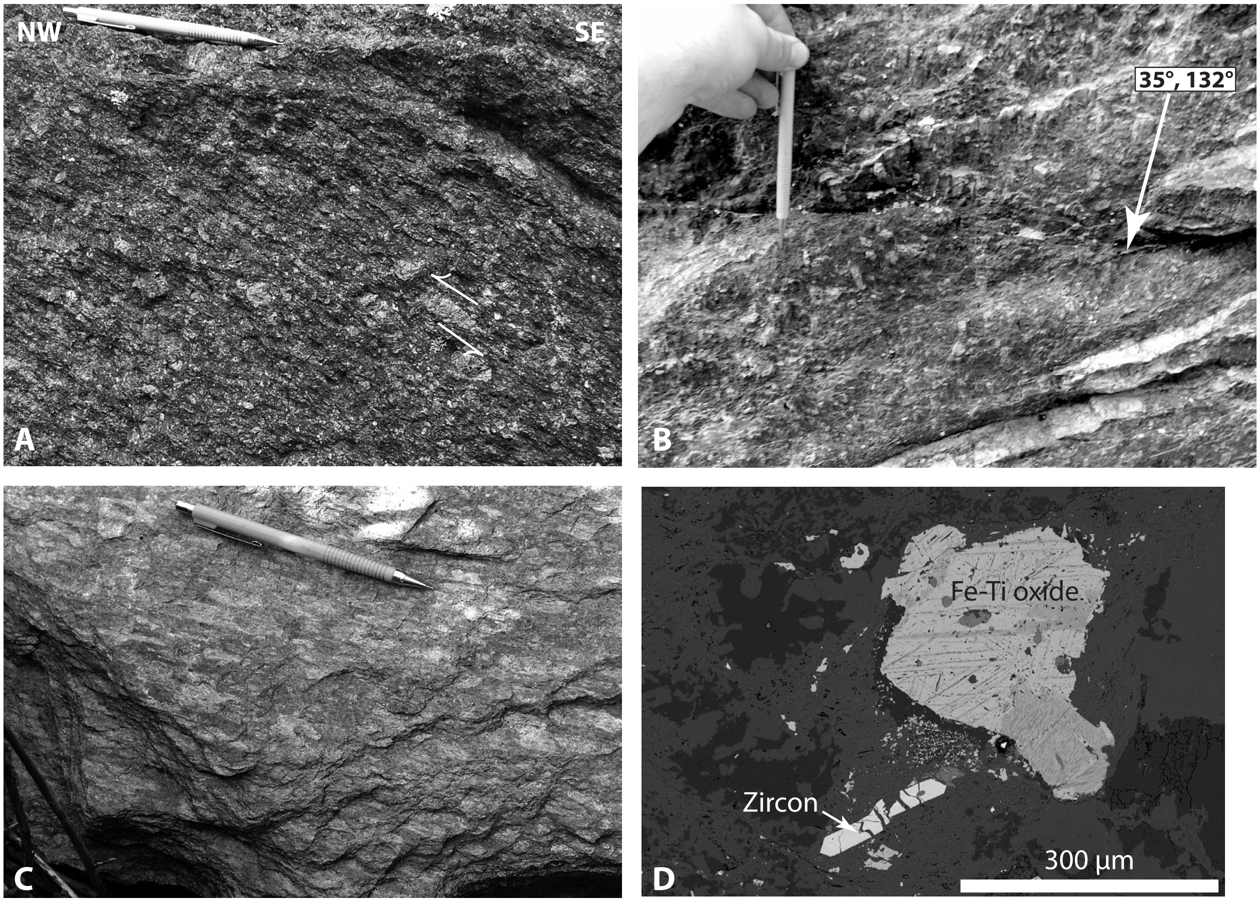

Figure 8. (A) Foliated Pond Mountain metarhyolite porphyry with centimeter–size alkali feldspar phenocrysts. Dark matrix is coarsely recrystallized quartz, sericite and chlorite. Asymmetric tails yield top-to-NW shear sense. (B) Alkali feldspar porphyroclasts in porphyritic Pond Mountain metarhyolite are fractured and elongated down-dip to the SE with top-to-the-NW sense of motion (toward the reader). (C) Mylonitic metarhyolite with sheared lapilli (gray spots) now flattened and elongated. Pencil is 14.5 cm long and parallel to mineral lineation and elongation direction of lapilli in (B) and (C). (D) Backscattered electron image of fractured prismatic zircon and Fe-oxide with Ti-oxide exsolution lamellae of the Pond Mountain rhyolite porphyry (MR-354). Fractured zircon crystal is indicative of air fall deposits. Matrix phases include muscovite, K-feldspar, quartz, albite, Fe-Ti oxide, and apatite.

Table 2. Crystallization ages of the Crossnore Plutonic-Volcanic Suite

Unit	Crystallization age (Ma)	Method	Notes	References
Mount Rogers Formation				
Razor Ridge rhyolite	748.4 ± 3.2	SHRIMP		—
Wilburn Rhyolite	749.7 ± 3.1	CA-TIMS	754.5 ± 1.2 (wtd. mean)	3
Whitetop Rhyolite	753.3 ± 2.0	CA-TIMS	757.1 ± 3.6 (wtd. mean)	3
Burrard Rock Rhyolite	755.0 ± 6.6	CA-TIMS	754.7 ± 4.7 (wtd. mean)	3
Fees Rhyolite	753.1 ± 2.7	CA-TIMS	759.8 ± 2.6 (wtd. mean)	3
Pond Mountain rhyolite	758.7 ± 2.9	SHRIMP		—
Metagabbro	757 ± 5	SHRIMP		3
Crossnore Plutonic Suite				
Beech Pluton	745 ± 6	TIMS		2
Beech Pluton	757 ± 6	TIMS	1424 ± 29 upper intercept	2
Lansing pluton	739 ± 4	TIMS		2
Warrensville pluton	740 ± 8	TIMS		2
Crossnore pluton	754 ± 5	TIMS		2
Bakersville Metagabbro	734 ± 26	Rb-Sr isochron		1
	759 ± 7	SHRIMP		4

References: 1—Goldberg and Fullagar (1986); 2—Su et al. (1994); 3—Tollo et al. (2012); and 4—J.N. Aleinikoff U.S. Geological Survey (2016, oral commun.)

wtd.—weighted; CA-TIMS—chemical abrasion-thermal ionization mass spectrometry;

SHRIMP—sensitive high-resolution ion microprobe; TIMS—thermal ionization mass spectrometry

that range from 800–710 Ma (Holm-Denoma et al., 2013, 2014, 2015) demonstrating that older volcanic sources existed (Fig. 9). Several separate clasts from Mount Rogers Formation conglomerates ranged from 780–752 Ma (McClellan and Holm-Denoma, 2014; McClellan et al., this guidebook). The oldest three rhyolite clasts and a granitoid clasts, 780–772 Ma, further support the long lived duration of magmatism and initiation of rifting by ~780 Ma (Fig. 9) (McClellan et al., this guidebook). Additionally, these ages overlap with published ages from plutonic components of the Crossnore Complex (Table 2) (e.g., Rankin, 1975; Tollo et al., 2012).

KONNAROCK FORMATION

The Konnarock Formation crops out for ~60 km in the Blue Ridge thrust sheet, and Mountain City window (King and Ferguson, 1960; Hardeman, 1966; Rankin, 1993; Miller, 2004). Rankin (1993) recognized the glaciogenic origin of the unit and defined it as separate formation from the Mount Rogers Formation. The Konnarock Formation locally unconformably overlies Mesoproterozoic basement and Mount Rogers Formation, although the contact is rarely exposed and/or is faulted (Rankin, 1993). The formation is in excess of 1000 m thick (Rankin, 1993; Miller, 2004). Maroon polylithic boulder to cobble diamictite and varved rhythmite containing dropstones are the most notable and characteristic lithologies (Rankin, 1993) (Fig. 10). These are interbedded with maroon to pink, massive sandstone/arkose, polylithic pebble conglomerate, maroon laminite, massive mudstone and other lithologies (see Miller, 2004).

The Konnarock Formation represents a glaciolacustrine depositional environment in a continental rift setting (Rankin, 1993; Miller, 2004). The Konnarock Formation was deposited in a thermally stratified lake basin that was actively subsiding (Miller, 2004). Varves indicate a seasonal control of the lake (melting and deposition during the warm periods) and dropstone stones indicate occurrence of ice-rafted material (Fig. 10) (Miller, 2004). Many of the sandstones, and interlayered sandstones and laminites represent subaqueous delta deposits (Miller, 2004; Moyer et al., 2016). The massive to laminated, locally cross bedded, well-sorted sandstones represent proximal delta deposits. Interlayered laminated mudstones and graded sandstones represent turbidites or gravity flows deposited in the delta fans. Diamictite, occurring at the top of formation, was deposited proximal to the glacier and likely indicates advance of a glacier across the paleo-Lake Konnarock. The diamictite may have been deposited as a lodgment till on top of a frozen lake (Rankin, 1993), or a subaqueous lacustrine deposit below an ice cliff or ramp, which explains interbedded sandstone within the diamictite (Miller, 2004). Clasts in the diamictite are dominantly Mesoproterozoic granitoids, while conglomerates in the Mount Rogers include both volcanic rocks and Mesoproterozoic granitoids (Rankin, 1993; Miller, 2004). Detrital zircons from the Konnarock Formation are dominantly Mesoproterozoic and mimic this provenance shift (Holm-Denoma et al., 2012, 2013). The source area was dominated by Mesoproterozoic granitoids, indicating the glacier likely advanced from the west (current geographic coordinates) because no Mesoproterozoic rocks were cropping out to the east at the time of deposition; alpine glaciers on volcanic edifices as portrayed by Miller (2004) were not the setting because of the lack of volcanic clasts and detrital zircons of the appropriate age within the Konnarock.

The depositional age of the Konnarock Formation is constrained by the youngest rhyolite (~749 Ma) and the disconformably overlying Unicoi Formation (Rankin, 1993). This broad age range for the Konnarock overlaps with several recognized global Neoproterozoic glaciations: Kaigas, ~750 Ma; Sturtian, 716.5–700 Ma; Marinoan, 660–631 Ma; and

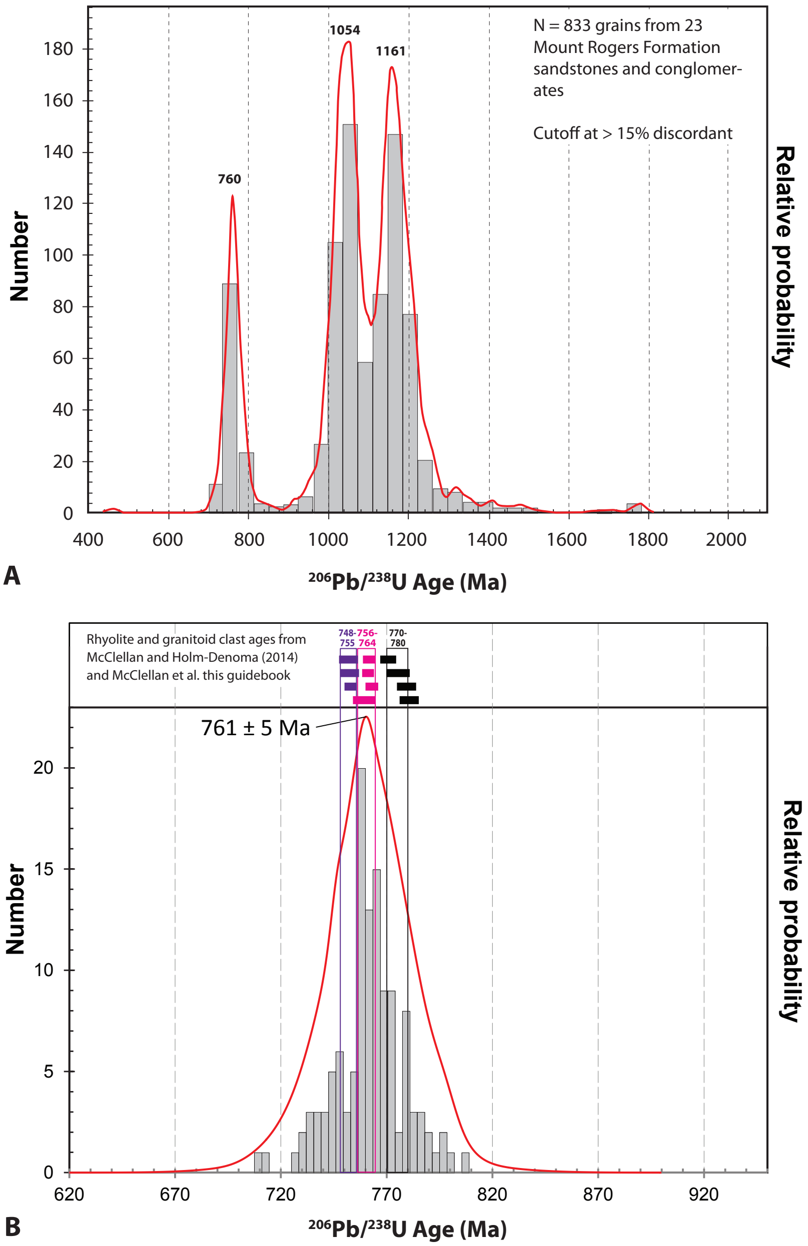

Figure 9. (A) Relative probability plot of laser ablation-inductively coupled plasma-mass spectrometry of twenty-three detrital zircon samples from the Mount Rogers Formation. Ages are less than 15 percent discordant and define peaks at ~1161, ~1054, and ~760 Ma, which agrees with existing geochronology from Mesoproterozoic basement and Neoproterozoic rocks (see Tables 1 and 2). (B) Relative probability plot of ages between 800–700 Ma and rhyolite and granitoid clast ages of McClellan and Holm-Denoma (2014) and McClellan et al. (this guidebook). Detrital zircon and rhyolite clast ages suggest rifting and magmatism initiated by ~780 Ma

Gaskiers 583–582 Ma (Hoffman and Schrag, 2002; Macdonald et al., 2010; Fig. 11). The supercontinent Rodinia was situated at low latitudes (4°–5°; 20°–30°) at this time (e.g., Hoffman and Schrag, 2002; Evans and Raub, 2011), so the Konnarock glaciation may be related to a global event. Rhyolite bodies within the glaciolacustrine rocks are small (1 to 2.4 km long) isolated bodies of rhyolite that are structurally separated from the main masses of rhyolite in the Razor Ridge area (Merschat and Southworth, 2011; Merschat et al. 2014) (Fig. 2). Polylithic pebble conglomerate above a body of rhyolite contains clasts of rhyolite, thus rhyolites are interlayered with glaciogenic rocks and not a klippe (Fig. 2) (e.g., Rankin et al., 1972; Rankin, 1993). Texturally, these purple to maroon, porphyritic rhyolites resemble those of the ~749 Ma Mount Rogers Formation. Until new U-Pb ages are obtained for these outliers, the tentative age of ~749 Ma is applied to the Konnarock Formation and glaciation in this area may be related to a pre–Sturtian event (Fig. 11).

Chilhowee Group

Unconformably overlying basement and the Konnarock Formation is the late Neoproterozoic to Cambrian Chilhowee Group succession of conglomerate, shale, and sandstone (Stose and Stose, 1957; King and Ferguson, 1960; Rankin et al., 1972; Rankin, 1993). The Unicoi Formation, the lowest unit, consists of a lower part of conglomerate, arkose, shale and basalt flows. The upper part is cross-bedded quartz arenite, which marks the transition from fluvial to marine environments (Simpson and Eriksson, 1989; Smoot and Southworth, 2014). The overlying

Hampton Formation is composed of black shales, some sulfidic, and gray sandstones that are turbidite successions deposited in deeper marine water (Smoot and Southworth, 2014). White vitreous quartz arenites and light greenish gray shales of the Erwin Formation represent a return to shallow marine conditions; the top is mostly calcareous sandstones and shales of the Helenmode Member, which commonly host manganese deposits (King et al., 1944; King and Ferguson, 1960). In a detailed study of the ichnotaxa of the Chilhowee Group, Hageman and Miller (2016) identified a nevadiid trilobite specimen from the Erwin Formation (Murray Shale) on Chilhowee Mountain in southeast Tennessee. The nevadiid trilobite constrains the Erwin Formation to Cambrian Stage 3 (521–517 Ma) (Hageman and Miller, 2016). The Chilhowee Group rocks are succeeded by blue-gray Shady Dolomite and maroon shale, sandstone and minor carbonates of the Rome Formation (King et al., 1944; King and Ferguson, 1960; Rankin, 1967).

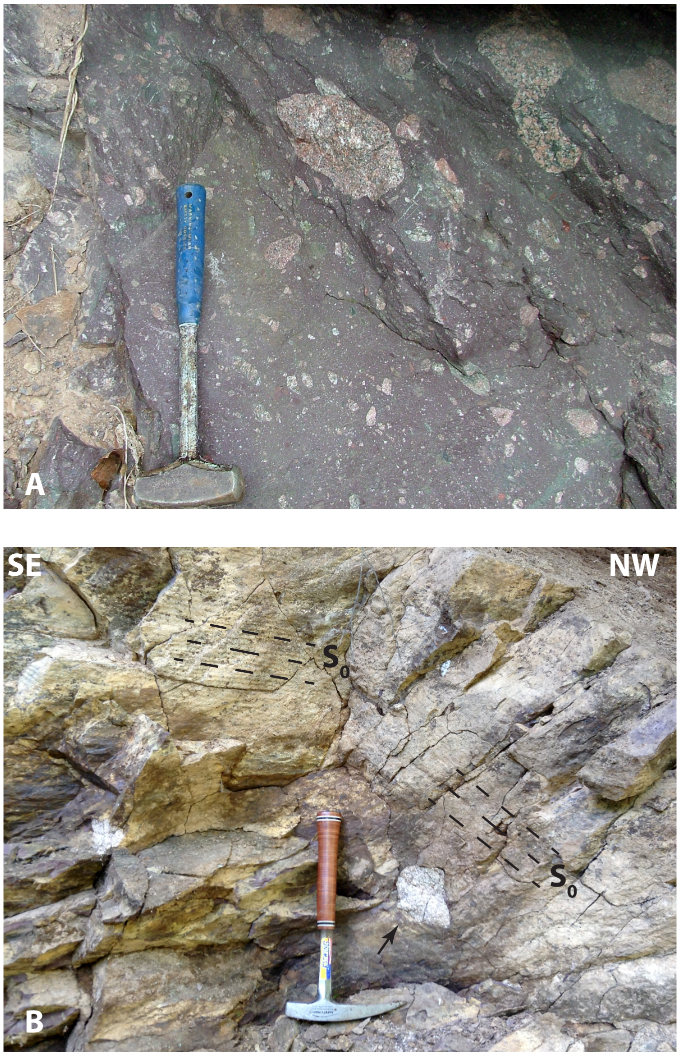

Figure 10. Diamictite and rhythmite are distinct glaciogenic lithofacies of the Konnarock Formation. (A) Maroon diamictite with various granitoid cobbles and boulders in a matrix of clay, silt, and sand occurs in the upper part of the formation. Outcrop is located on US 58 near intersection with VA 728, see Stop 1–9. Sledge is 36 cm long. (B) Cobble-sized granitoid dropstone (arrow points to dropstone) in rhythmite from small quarry on VA 859. Beds, S_0, (dashed lines) dip to the NW and distinct couplets with sharp boundaries indicate these are likely varves. Hammer is 32.5 cm long.

DETRITAL ZIRCON GEOCHRONOLOGY OF NEOPROTEROZOIC ROCKS AND BASIN DYNAMICS

Several samples (N=23) from the Mount Rogers, Konnarock, and Unicoi formations from each of the above mentioned volcanic centers/thrust sheets were sampled and analyzed for detrital zircon U-Pb ages by SHRIMP and LA-ICPMS. A pooled probability density plot of all detrital zircon analyzed (N=~1300 analyses <15% discordant, 833 from the Mount Rogers Formation as shown in Figure 9) for this study shows three main age peaks, ~1160 Ma, ~1040 Ma, and ~760 Ma, which coincide with the Grenville early and late magmatic suites (Tollo et al., 2010, 2012) and Mount Rogers volcanism, respectively. Within the Neoproterozoic (~760 Ma) peak, 243 analyses between 800-700 Ma were evaluated and treated statistically to deconvolute and identify multiple zircon age populations under the ~760 Ma curve. The "Unmix" function of Isoplot v. 3.75 (Ludwig, 2012) uses the relative misfit parameter of the Sambridge and Compston (1994) method for deconvolution of Gaussian data. More than 90% of the data falls into the age grouping of ~761 Ma (Figure 9b); however, two other minor age populations were identified at approximately 785 Ma and 747 Ma. This is significant as three rhyolite and a granitoid clast from the lower Mount Rogers Formation yielded ages between ~780-772 Ma as well as several rhyolite clasts that are ~760 Ma and ~753 Ma (see McClellan and Holm-Denoma, this guidebook; Holm-Denoma and McClellan, 2014, 2015; McClellan and Holm-Denoma, 2014) (Figure 9b). This suggests that magmatism in this region started by at least 780 Ma and persisted for at least 30 m.y.

A Mount Rogers area composite stratigraphic section (including samples from each volcanic center/thrust sheet) based on the detrital zircons collected for this study was created to show detrital zircon provenance trends through time (Figure 12). Several observations are noted from the comparison of individual samples in this data set. (1) The lower Mount Rogers Formation has variable amounts of Mesoproterozoic and Neoproterozoic detrital zircon populations. However, the early magmatic suite Mesoproterozoic zircons are much more abundant than late magmatic suite derived zircon. This suggests that during initial rifting and basin formation the Early Magmatic suite granitoids were volumetrically more exposed than late magmatic suite granitoids. (2) The upper Mount Rogers Formation detrital zircon population is almost entirely derived from volcanic rocks erupted during ~760 Ma rifting of the region. (3) Between the lower and upper Konnarock Formation, ~760 Ma detrital zircon is no longer being sourced. This suggests that (a) volcanism has ceased in the region, and (b) that detrital zircon of that age is no longer accessible/exposed for erosion and deposition into the Konnarock basin. (4) Detrital zircon in the upper Konnarock is dominantly derived from late magmatic suite rocks. This suggests that either the source of the Konnarock has switched due to shifting drainage patterns or, if the sediment is still being derived locally, that the early magmatic suite dominated the upper portion of the basement rocks, and that progressive unroofing of the basement exposed lower crustal assemblages

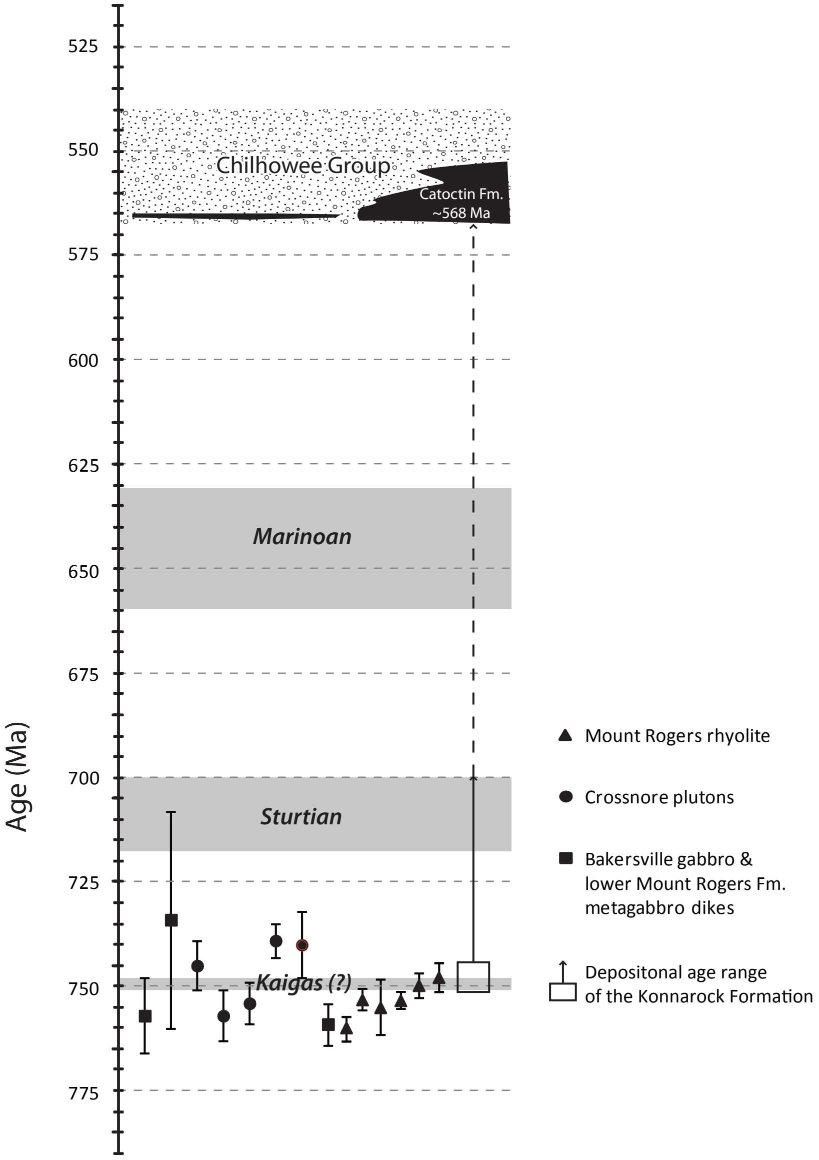

Figure 11. Comparison of ages of the Mount Rogers Formation rhyolite members, plutonic rocks of the Crossnore Complex, Konnarock Formation, and Neoproterozoic global glaciations. Sources of ages are listed in Table 2. Neoproterozoic glaciations are compiled from Hoffman and Schrag (2002) and MacDonald et al. (2010).

dominated by the late magmatic suite—essentially exposing a stratified basement architecture.

The detrital zircon age population variations within samples of the lower Mount Rogers Formation coincide with their structural/volcanic center relations (Fig. 12). Samples from the Mount Rogers volcanic center/thrust sheet have detrital zircon that are dominantly ~760 Ma with relatively minor Mesoproterozoic grains. This suggests that the basin was dominantly filled by volcanic and volcaniclastic rocks. The Razor Ridge volcanic center/thrust sheet is apparently devoid of ~760 Ma volcanic-derived detrital zircon. This suggests that the basin was isolated from volcanic rocks during deposition. Interestingly, the Razor Ridge rhyolite dated at ~749 Ma (Figure 7) is one of the youngest in the area. We suggest that the Razor Ridge basin was an isolated basin or is relatively young compared to the Mount Rogers and Pond Mountain basins. It is possible that the Razor Ridge basin was filled quickly prior to eruption of the ~749 Ma volcanism. The Pond Mountain volcanic center has a relatively bimodal detrital zircon age distribution including both basement (Mesoproterozoic) and volcanic (~760 Ma)-derived detritus. This suggests that the Pond Mountain basin was receiving both volcanic and volcaniclastic detritus as the basement was extended and eroded. The three volcanic centers have unique detrital zircon age populations, but each was formed during the active extension and uplift of the area in which the lower Mount Rogers formation was being deposited into basins. The basement paleotopography was quite variable and had a profound impact on the facies preserved in the basins.

EASTERN BLUE RIDGE — GOSSAN LEAD THRUST SHEET

The eastern Blue Ridge consists of an amphibolite-grade assemblage of dominantly siliciclastic rocks with mafic and ultramafic rocks that record the closing of the Iapetus ocean. Rankin (1970) defined two formations in the Blue Ridge of northwestern North Carolina, the Ashe and overlying Alligator Back formations. Tectonic aspects of the rocks led Abbott and Raymond (1984) to call these rocks the Ashe Metamorphic Suite within the Gossan Lead thrust sheet. The Wills Ridge Formation, as defined by Rankin et al. (1993), consists of graphitic schists, arkoses, and conglomerate deposited nonconformably on basement and underlying the Ashe along a pre–metamorphic fault. Eastern Blue Ridge lithostratigraphy and structures are

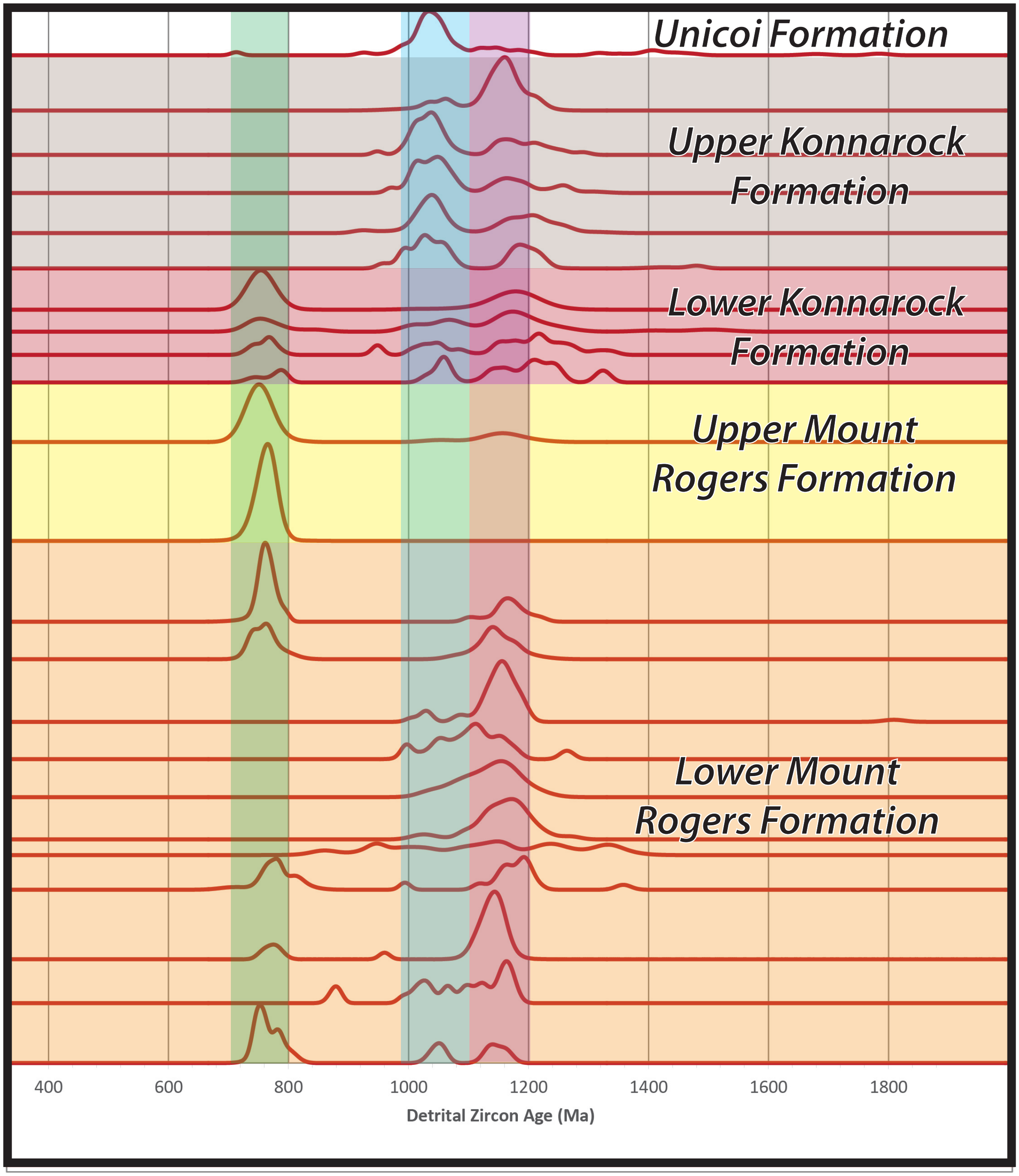

Figure 12. A pooled probability density plot of SHRIMP and LA-ICPMS detrital zircon U-Pb ages from twenty-three samples from the Mount Rogers, Konnarock, and Unicoi formations (N=~1300 analyses <15% discordant). Samples are stacked in stratigraphic order. Age range of the early magmatic suite (purple), late magmatic suite (blue), and Crossnore Volcanic-Plutonic Complex (green) are indicated. Probability density plot constructed using Isoplot v. 3.75 (Ludwig, 2012).

further discussed in Carter and Merschat (this guidebook). To the southwest the Ashe Formation is intruded by the Devonian Spruce Pine pegmatites (392–361 Ma; Mapes, 2002; Miller et al., 2006) and to the southeast the Mississippian Stone Mountain and Mount Airy granites intrude the Alligator Back Formation (Rankin et al., 1972; Miller et al., 2006). The age of the Ashe Formation remains loosely constrained by the youngest detrital zircons and oldest intrusion, 750–466 Ma (e.g., Bream et al., 2004; Merschat et al., 2010). Interpretation of the Ashe Formation as being deposited after the opening of the Iapetus ocean, suggests that it likely ranges from Late Neoproterozoic to Ordovician, 570–466 Ma (Rankin et al., 1975; Rankin, 1993, Hatcher, 2007a; Merschat et al., 2010).

In the southeastern corner of the field trip area, the Ashe Formation consists of metagraywacke (two-mica gneiss) and interlayered pelitic schist, graphitic schist, amphibolite, chlorite-talc-amphibole schist, and an important locality of metaconglomerate near the contact with Mesoproterozoic basement (Rankin, 1971; Rankin et al., 1972). Graphitic schist, metagraywacke and metaconglomerate along the western part of the Gossan Lead thrust sheet (Fig. 2) may be correlated with the

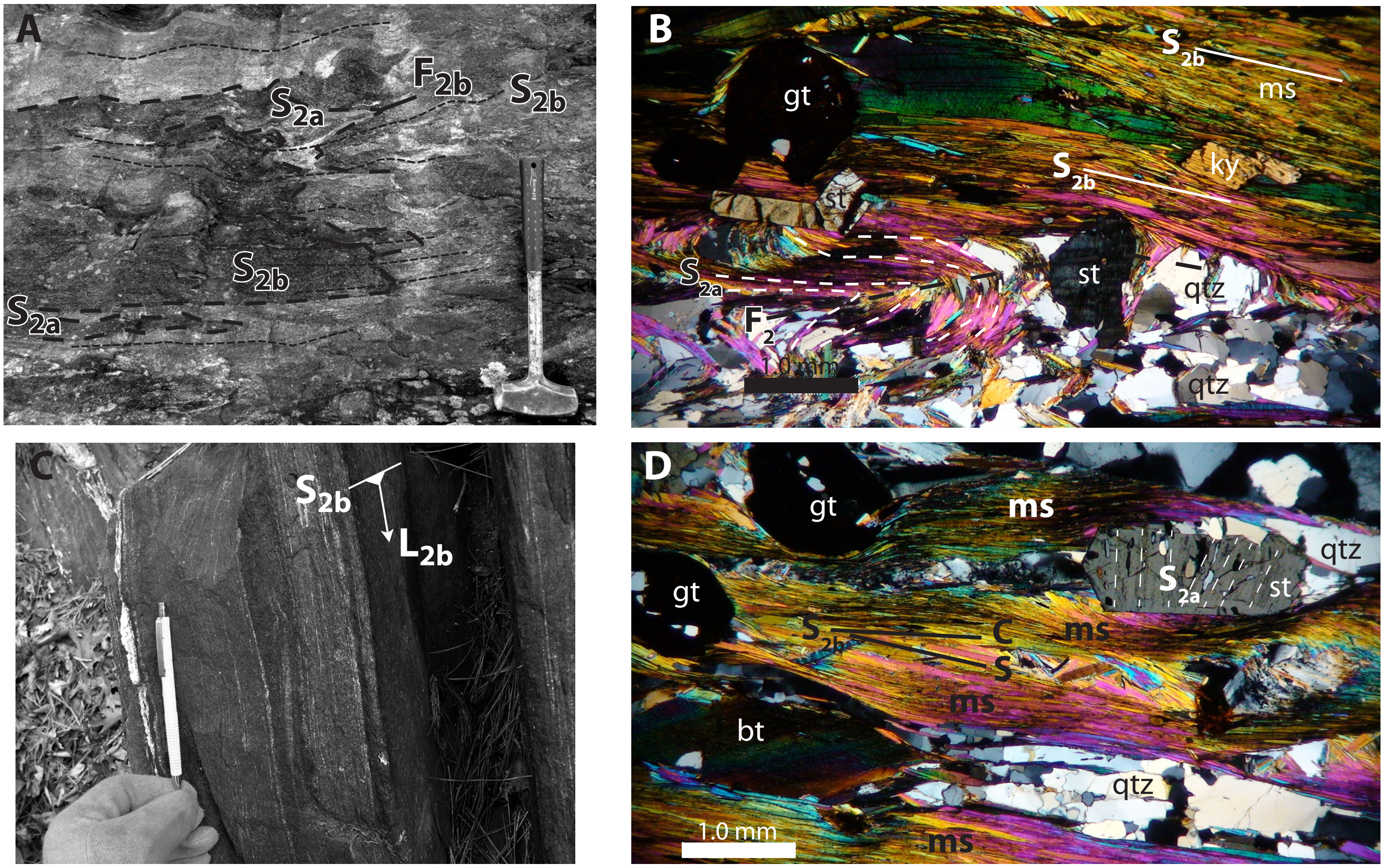

Figure 13. (A) Isoclinal F_{2b} folds in schist and metagraywacke. S_{2a} foliation (long-dashed thick lines) is deformed into tight to isoclinal F2b folds (solid line) with axial-planar S_{2b} foliation (short-dashed thin lines). Sledge is ~36 cm long. (B) Kyanite-garnet-staurolite-muscovite schist with an earlier foliation (S_{2a}) folded by F_{2b} and axial planar S_{2b}. Linear inclusion tail, S_{2a}, in dark bluish gray staurolite is also folded. (C) Laminated amphibolite comprised of quartz and hornblende exposed at the abandoned quarry on NC Route 113. A preferred $^{40}Ar/^{39}Ar$ hornblende cooling age of 340 Ma was obtained from this location. S_{2b} foliation and L_{2b} lineation are defined by hornblende. Pencil is ~14.5 cm long. (D) Porphyroclastic garnet-staurolite-muscovite schist with recrystallized quartz ribbons, and S-C fabric. Porphyroclasts are deformed and have pressure shadows of quartz, muscovite, and biotite. Staurolite porphyroclasts contains S_{2a} linear inclusion trails that are nearly orthogonal to the S_{2b} foliation. Abbreviations: ms–muscovite, qtz–quartz, st–staurolite.

Wills Ridge Formation (Rankin et al., 1993) or the Lynchburg Formation (Stose and Stose, 1957; Carter and Merschat, 2014, this guidebook). Metamorphic grade is kyanite-staurolite zone locally overprinted by a retrograde greenschist-facies assemblage. Interlayered quartzo-feldspathic and pelitic layers may be beds, but sedimentary structures are transposed into the strong regional foliation. Linear inclusion trails in garnet and staurolite porphyroblasts, intrafolial folds, and oblique foliations preserved in amphibolite boudins indicate the regional foliation is at least S_2 (Fig. 13). The regional foliation is superposed by a later F_3 crenulation that folds the earlier foliations and layering.

The origin of rocks in the eastern Blue Ridge remains controversial. Rankin (1970; 1975) suggested that the Ashe and overlying Alligator Back formations were deposited unconformably on basement, but later restricted only the Wills Ridge Formation to a nonconformable position above basement. Abbott and Raymond (1984) suggested that the Ashe Formation was deposited on oceanic crust due to the large volume of amphibolite and small ultramafic bodies, and transported onto basement along the Gossan Lead fault of Stose and Stose (1957). Whisonant and Tso (1992) suggested that the rocks were deposited in slope and rise setting, based on quartz, feldspar, and clay modal proportions of metasandstones. Detrital zircon suites from along strike segments of the eastern Blue Ridge have yielded 1.4, 1.3, 1.2, 1.1, 1.0 and minor 0.7 Ga detrital zircons spectra suggesting a peri-Laurentian provenance (Bream et al., 2004; Carter et al., 2006; Merschat et al., 2010). Tectonic compilations of the orogen have included the eastern Blue Ridge in various terranes: Piedmont terrane (Williams and Hatcher, 1982), Jefferson (Horton et al., 1991), Toe (Raymond et al., 1989), Piedmont zone (Hibbard et al., 2006), and the Tugaloo (Hatcher et al., 2007a).

STRUCTURE

The Blue Ridge is polydeformed by multiple Proterozoic to Paleozoic orogenies. Structures across the Blue Ridge vary in style and age. This study recognizes the following deformational events: D_1, Grenville orogeny, 1.2–1.0 Ga; D_2, middle Paleozoic Taconic and Acadian-Neoacadian orogenies, 470–440 Ma and 395–340 Ma; D_3 Alleghanian orogeny, 340–260 Ma.

The oldest structures occur in the basement gneisses and are the result the Grenville orogeny, which actually consists of three to four different phases or events. The Grenville orogeny resulted in the formation of the Proterozoic supercontinent Rodinia. As noted previously, associated tectonomagmatic events include the Elzevirian, Shawinigan, Ottawan and Rigolet (McClelland et al., 2010, 2013). Structures that formed during these events are difficult to recognize, largely due to the intense overprinting Late Paleozoic deformation, D_3. In areas of less intense late Paleozoic deformation, high-grade D_1 structures have been recognized (Tollo et al., 2010, 2012). D_1 structures are primarily defined by high-grade (amphibolite to granulite facies) gneissic foliation S_1 and roughly parallel migmatite layers in the orthogranofels, granofels, and basement amphibolites. In some granitoid lithologies, S_1 may be defined by the alignment of alkali feldspar. Although the orientation of S_1 may vary, it generally strikes NW and dips steeply (mean foliation 322/83, Fig. 14). Folds and mineral lineations are also associated with D_1, but are rare. Tollo et al. (2010, 2012) delimited the formation of these structures using U-Pb geochronology and field relationships to either the Shawinigan, or Ottawan orogenic phases.

The middle Paleozoic orogenies are combined into one event, D_2, because the different structures or age of the structures cannot be clearly resolved by this study (i.e., Taconic versus Acadian). Middle Paleozoic deformation is limited to the eastern Blue Ridge and possibly into the immediate footwall of the Gossan Lead fault (Fig. 14). No evidence for middle Paleozoic deformation, D_2, has been recognized in the western Blue Ridge, Stony Creek syncline, Mountain City window, and Valley and Ridge. The eastern Blue Ridge, Ashe Formation rocks, in the southeast corner of the study area are polydeformed amphibolite facies rocks that reached kyanite and staurolite zones. The different amphibolite facies deformations in the eastern Blue Ridge are designated with "a, b, c . . ." to indicate different generations of the structures. The dominant foliation, S_{2b}, is defined by biotite + muscovite + quartz + garnet + plagioclase + staurolite + kyanite in pelitic and psammitic rocks, and hornblende + quartz + plagioclase in basic rocks. Linear inclusion trails in garnet and staurolite porphyroclasts define an earlier middle Paleozoic foliation S_{2a} that is oblique to S_{2b} (Fig. 13). The L_{2b} mineral lineation is defined by mineral aggregates of quartz + muscovite, and aligned minerals including amphiboles, and, in some schists, staurolite. Abbott and Raymond (1984) described five deformations and three metamorphic events that affected the rocks of the Ashe Metamorphic Suite during the Taconic and Acadian orogenies. Our D_{2a} and D_{2b} deformations likely correspond to their D_3, D_4, and D_5 events (Taconic and Acadian deformation). To the southwest, however, multiple studies have separated Taconic and Acadian-Neoacadian structures in the eastern Blue Ridge of northwestern North Carolina (Goldberg and Dallmeyer, 1997; Trupe et al., 2003; Miller et al., 2010). Based on the results of the ^{40}Ar/^{39}Ar thermochronology, discussed below, the D_2 event occurred before ~340 Ma.

Most of the structure of the Mount Rogers area is related to the Alleghanian orogeny, D_3. These structures include a penetrative greenschist facies foliation and cleavage (S_3), regional SE-plunging mineral stretching lineation (L_3), northwest-verging folds and crenulations (F_3) (Fig. 14). Alleghanian fabrics decrease in grade and intensity from SE to NW. The Valley and Ridge, Mountain City window and Shady Valley thrust sheet are characterized by a spaced cleavage that is axial-planar to Alleghanian folds (Fig. 14).

The S_3 foliation is defined by greenschist facies minerals muscovite (sericite) + quartz + chlorite + magnetite ± epidote ± K-feldspar ± albite. Alleghanian deformation is more intense in high-strain zones. Largely granitic to rhyolitic protoliths are altered to phyllonites dominated by fine-grained muscovite (sericite) formed through the replacement of feldspar. The L_3 mineral lineation defines a regional pattern across the western and eastern Blue Ridge (Fig. 14). The lineation is defined by fractured and stretched feldspar, and mineral aggregates of muscovite + quartz + chlorite (Fig. 8B and 8C). Top-to-NW kinematics associated with the mineral lineation suggests NW-directed transport of the Blue Ridge thrust sheet. Folds are not common in many of the deformed basement rocks, but are common in thin bedded, fine-grained sedimentary rocks in the Mountain City window and Stony Creek syncline. These folds plunge gently (3–20°) to the ENE (35–060), with a steep to subvertical axial planar cleavage (Fig 14).

^{40}AR/^{39}AR THERMOCHRONOLOGY

Samples containing hornblende, muscovite, and K-feldspar were collected from the western edge of the Blue Ridge thrust sheet to the eastern Blue Ridge for ^{40}Ar/^{39}Ar thermochronology (Stokes et al., 2010; Stokes, 2013; M. Kunk, unpublished data; Tollo et al., 2012; McAleer et al., 2015). Samples were collected from different units—basement gneisses, marbles, muscovite-rich phyllonites, foliated rhyolites of the Mount Rogers Formation, and Ashe Formation schists and amphibolites—and across different structures—the Catface fault, greenschist facies high-strain zones related to the Fries fault, and the Gossan Lead fault. A summary of the results is shown in Figure 15.

Two patterns are evident in the ^{40}Ar/^{39}Ar data. First, from the Catface fault SE across the Gossan Lead fault, muscovite yields plateau or near plateau ages of ~340 Ma for all samples (Fig 15). Second, the ^{40}Ar/^{39}Ar ages of amphibole and K-feldspar are markedly different on either side of the Gossan Lead fault. Amphibole in the basement of the western Blue Ridge records an age of ~950 Ma, but amphibole immediately east of the Gossan Lead fault is ~340 Ma. K-feldspar age spectra from rocks west of the Gossan Lead fault climb to ages > 340 Ma (older than muscovite ages) in all cases (n = 5), but only climb in age to ~300 Ma east of the Gossan Lead fault. These patterns suggest the western and eastern Blue Ridge had different thermal histories (Fig. 16).

The preservation of Grenvillian amphibole ages west of the Gossan Lead fault indicates these rocks were never heated above ~500° C during Paleozoic orogenesis. Similarly, because some K-feldspar age spectra climb to >470 Ma, Paleozoic metamorphic conditions likely never exceeded the lower greenschist-facies conditions west of the Gossan Lead fault. The fact that all (n = 5) K-feldspar spectra climb to ages that exceed the age of foliation-forming muscovite separated from the same rocks further confirms that Paleozoic deformation occurred in the lower greenschist-facies conditions. This low grade of Paleozoic metamorphism in the western Blue Ridge is consistent with all field and petrographic observations.

In light of these data we interpret the ~340 Ma age from foliation-forming muscovite in the western Blue Ridge to record the time of foliation formation rather than the time of cooling

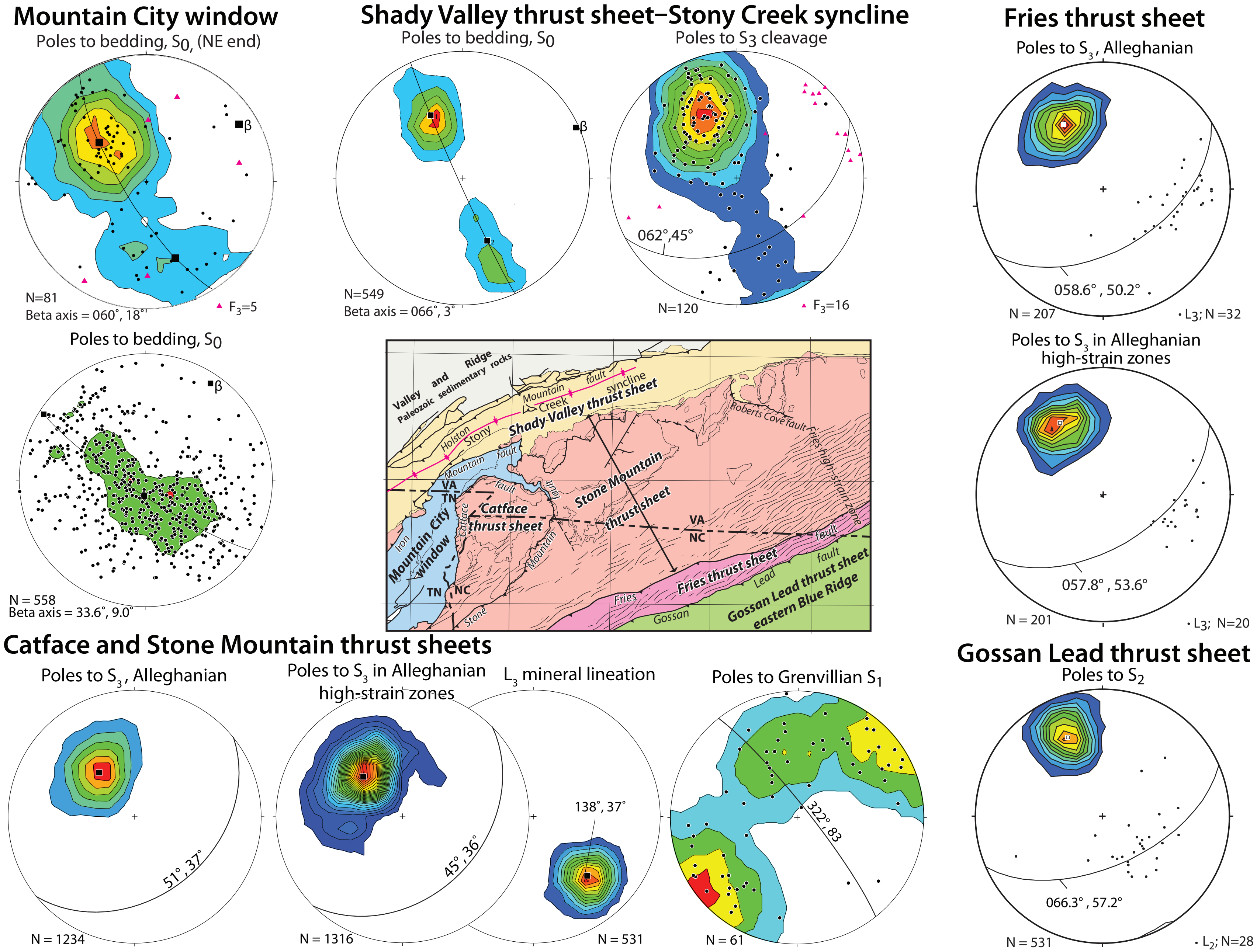

Figure 14. Equal area, lower hemisphere stereonets of structural data from the Blue Ridge of southwestern Virginia and parts of North Carolina and Tennessee separated according to thrust sheet: Mountain City window, Shady Valley thrust sheet/Stony Creek syncline, Catface-Stone Mountain, Fries and Gossan Lead thrust sheets. Contour interval is 2% per 1% area; mean vector of planes is indicated by great circle. All plots were created using Stereonet v. 9.3.2 (Allmendinger et al., 2012). Inset tectonic map is modified from Fig. 1D.

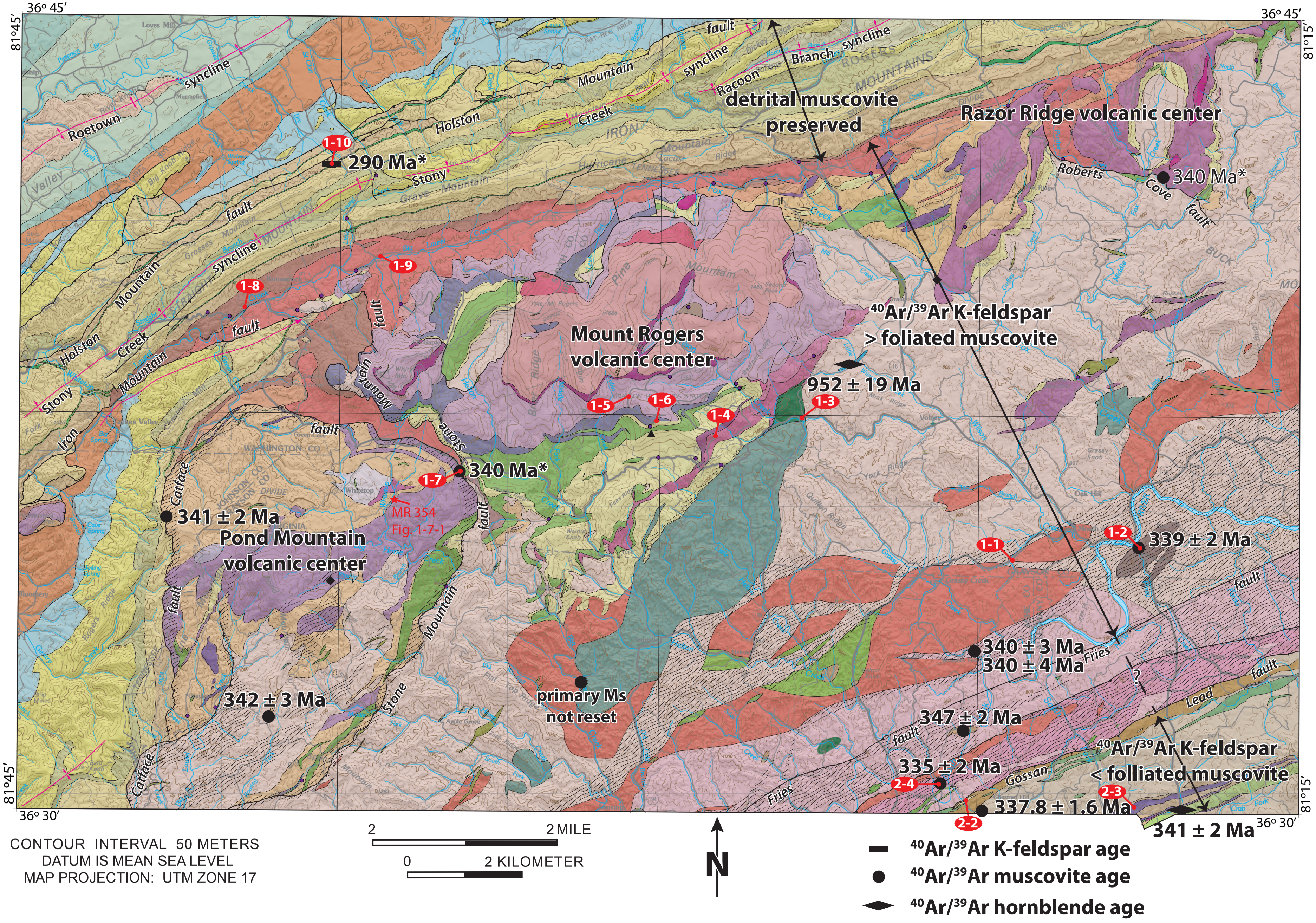

Figure 15. Compiled $^{40}Ar/^{39}Ar$ hornblende, muscovite, and K-feldspar ages from the Mount Rogers area, VA–NC–TN. Ages with errors are plateau ages; preferred ages are indicated with an asterisk. Data are from Stokes (2012), Tollo et al. (2012), M.J. Kunk, USGS (oral comm., 2012), and this study. Geologic map is the same as Figure 2.

through muscovite closure. In other words, the penetrative Paleozoic deformation (see Stops 1–2 and 1–7, Fig 1–7–1) observed in the Fries fault and Blue Ridge thrust (Catface and Stone Mountain faults) occurred at ~340 Ma. These samples yield nearly ideal plateau ages because the folia are in rocks with no primary muscovite (e.g. rhyolite) and so only one generation of muscovite is present. Where foliation-forming muscovite overprints primary basement muscovite, a complex, older age spectrum results, which confirms the below-closure growth of muscovite in the western Blue Ridge (Fig. 16; Stokes, 2012).

In the eastern Blue Ridge amphibole and muscovite ages are as young as ~340 and ~335 Ma. Because the microchemistry of the amphibole shows that it grew at 700° C (Stokes et al., 2012), and muscovite occurs in folia along with biotite (indicating minimum growth temperatures of >400° C), these muscovite and K-feldspar ages are interpreted to reflect the time of cooling through 350° C following peak metamorphism.

SUMMARY OF TECTONIC EVENTS

The tectonic history of the Mount Rogers area (Fig. 16) begins with ~1.33 Ga, pre–Grenville crust. These rocks were intruded by the early magmatic suite plutonic rocks at 1190–1140 Ma, which corresponds with high-grade deformational structures (D_1, S_1) in the pre-Grenville rocks and crystallization of metamorphic zircon rims (Tollo et al, 2010, 2012). The early magmatic suite constitutes the majority of Mesoproterozoic rocks in the Mount Rogers area (Tollo et al., 2010, 2012), the French Broad massif, smaller internal massifs in the southern Appalachians (Carrigan et al., 2003), and the Shenandoah massif (Southworth et al., 2010; Carter et al., 2012, 2013). This event is defined as the Shawinigan orogeny, 1190–1140 Ma, and represents the initial collision of Amazonia and Laurentia (McClelland et al., 2010, 2014). The late magmatic suite, 1075–1030 Ma correlates with the Ottawan Phase, 1090–1020 Ma, of the Grenville orogeny and continued collision between Amazonia and Laurentia (Tollo et al., 2010). Whole-rock Pb isotopic studies indicate that Mesoproterozoic rocks of the southern and central Appalachians, *including the rocks of the Mount Rogers area*, were part of Amazonia (Sinha et al., 1996; Loewy et al., 2003; Fisher et al., 2010). The continuity of magmatic and deformational ages from the central and southern Appalachian massifs to the Adirondacks and the Canadian Grenville Provence and Pb isotopic signatures demonstrate the collision of Amazonia and Laurentia began during the Shawinigan orogeny (Carrigan et al., 2003; McClelland et al., 2010, 2013, Tollo et al., 2010), and may have continued episodically through 0.98 Ga (Aleinikoff et al., 2013).

During the Cryogenian there was a period of continental rifting, volcanism, and glaciation. Deep crustal basement rocks were exhumed to the surface (Fig. 16) and intruded by mafic dikes (757 ± 9 Ma; Tollo et al., 2012) that locally breeched the surface and formed basalt flows now in the lower Mount Rogers Formation. Intertonguing arkoses, conglomerates, shales and mudstones represent the variation and close proximity of high to low energy sedimentary environments of alluvial fan and shallow basin deposits that developed in the rifts associated with the Mount Rogers Formation. Rhyolites from the Mount Rogers Formation span at least from 760–749 Ma (Tollo et al., 2012)—rhyolite clast ages suggest volcanism and rifting initiated as early as ~780 Ma (McClellan and Holm-Denoma, 2014; McClellan et al., this guidebook). Geochemically the rhyolites are A-type, rift-related, within-plate magmas, similar to granitoid plutons of the Crossnore Complex in northwestern North Carolina (Rankin, 1975; Novak and Rankin, 2004; Tollo et al., 2012). Likewise, basalts (greenstones) from the Mount Rogers Formation and mafic dikes from the underlying basement are similar to Bakersville Gabbro near the type area (Tollo et al., 2012). Ultimately, the underwelling of mantle-derived mafic magma resulted in the generation of siliceous magma of the 760–740 Ma granitoid plutons and metagabbro bodies of the Crossnore Complex and the 760–749 Ma Mount Rogers Formation rhyolites (McClellan and Gazel, 2014). Detrital zircons and rhyolite clasts suggests rifting and magmatism began by ~780 Ma (McClellan and Holm-Denoma, 2014; Holm-Denoma et al., 2014, 2015; McClellan et al., this guidebook). Globally, ca. 780 Ma is recognized as a mantle plume/super plume event (Li et al., 2008). McClellan and Gazel (2014) suggested that the Mount Rogers Formation represented results of a mantle plume, but exactly how the plume is manifested in the geology of the Blue Ridge remains unclear. Regionally, Neoproterozoic magmatic rocks in the Blue Ridge of northwest North Carolina to central Virginia young to the northeast (Burton and Southworth, 2010; McClellan and Gazel, 2014). Analogues for middle Neoproterozoic tectonics of the Blue Ridge are the East African rift (Rankin, 1975; Novak and Rankin, 2004), Yellowstone hot spot (Tollo et al., 2012), or mantle plume (McClellan and Gazel, 2014).

Glaciation was locally penecontemporaneous with late felsic volcanism (Merschat and Southworth, 2011; Merschat et al., 2014). The dominance of Mesoproterozoic basement clasts in Konnarock diamictites suggest that glaciers advanced from the west (present geographic coordinates; Rankin, 1993). Paleolatitude estimates for Laurentia place the Konnarock Formation at low to middle latitudes (4°–5°; 20°–30°) at ~740 Ma and the existence of glaciers at this latitude may reflect a global event (Hoffman and Schrag, 2002; Evans and Raub, 2011). Rhyolite bodies in the Konnarock Formation provide a possible age constraint, ~750 Ma and suggest a correlation with a pre-Sturtian glacial event. Field relationships, however, are not conclusive. The rhyolites may represent erosional rhyolite domes that were topographic highs in the paleolake Konnarock basin. Further, global evidence of a pre-Sturtian glacial event is in doubt based on isotopic evidence and field relationships (Rooney et al., 2015). The Sturtian event (717–700 Ma) is well documented in the Laurentian craton in northwestern Canada (e.g., Macdonald, 2010), and may be an equally reasonable time of deposition of glacial rocks represented by the Konnarock Formation.

Basalt flows in the lower Unicoi Formation are likely related to continental basalts of the 571–565 Ma Catoctin Formation and the break-up of Rodinia (i.e., Aleinikoff et al., 1995; Burton and Southworth, 2010; Smoot and Southworth, 2014). Fluvial environments represented by conglomerates, arkose, and shales of the lower Unicoi Formation transgressed to marine and passive margin sediment in the upper Unicoi Formation (Simpson and Eriksson, 1989; Walker et al., 1994; Smoot and Southworth, 2014). Eventually, clastic sedimentation ceased and an Early

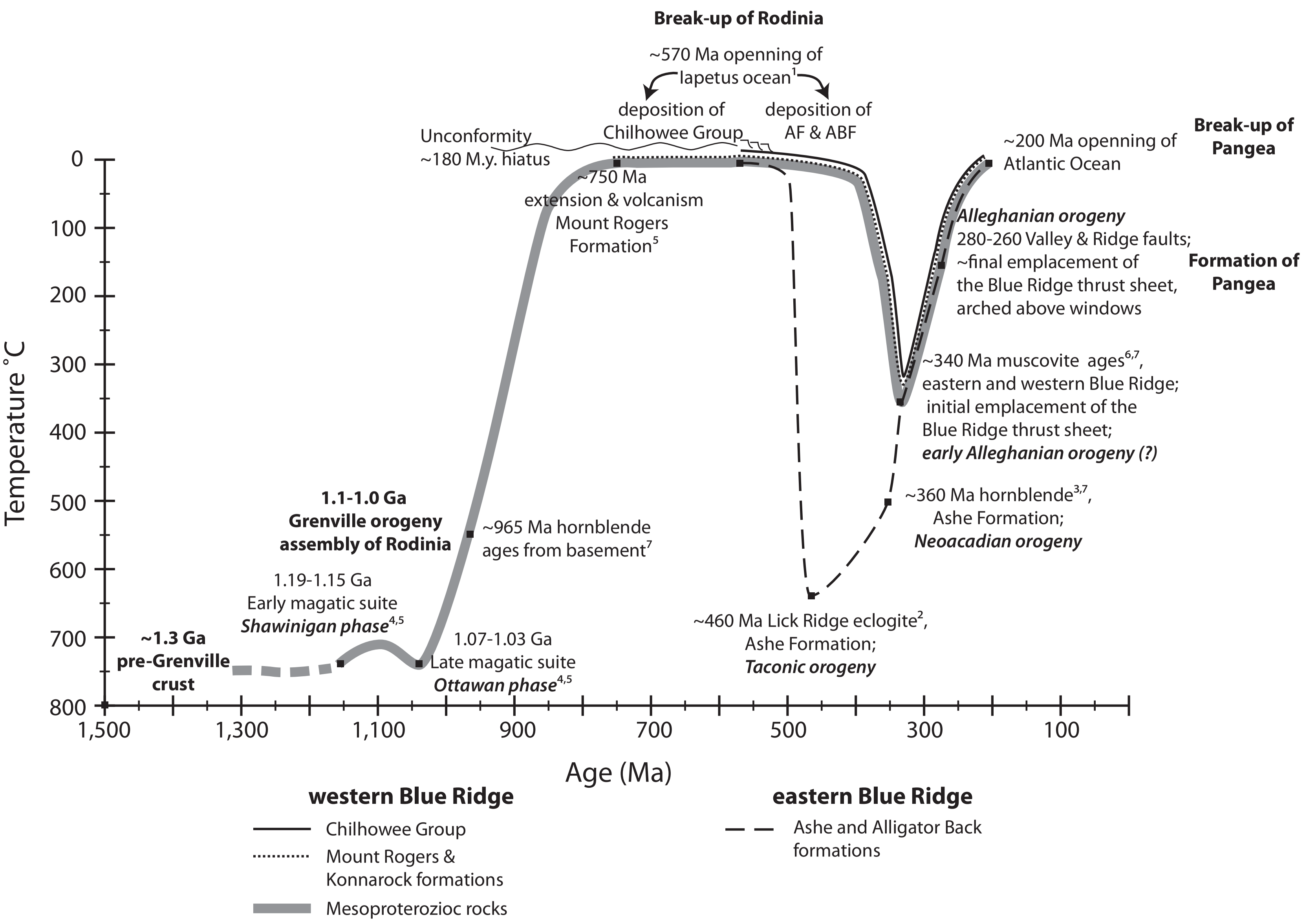

Figure 16. Tectonic summary of the Mount Rogers area plotting time verses temperature or depth for the different geologic units of the Blue Ridge. Data sources: 1—Aleinikoff et al. (1995); 2—Miller et al. (2010); 3—Stokes et al. (2010); 4—Tollo et al. (2010); 5—Tollo et al. (2012); 6—M.J. Kunk, USGS (oral comm., 2012); and 7—Stokes (2013).

Cambrian to Ordovician carbonate shelf developed. During this same period, eastern Blue Ridge rocks (Ashe Formation) were deposited in the opening Iapetus ocean basin. Provenance studies support a Laurentian source likely in a slope and rise setting proximal to Laurentia (Bream et al., 2004; Carter et al., 2006; Merschat et al., 2010).

The eastern Blue Ridge was affected by multiple Middle Paleozoic orogenies resulting in the closing of the Iapetus ocean and amphibolite-facies metamorphism (Fig. 15). Evidence for both a 465–450 Ma Taconian and a 390–345 Ma Acadian-Neoacadian event has been recognized in the eastern Blue Ridge elsewhere, however, only Devonian to Mississippian deformation and metamorphism are recorded in northwesternmost North Carolina by ^{40}Ar/^{39}Ar hornblende cooling ages 360–340 Ma (Stokes et al., 2010; Stokes, 2013), which likely reflects Acadian-Neoacadian orogenesis and collision of the Carolina superterrane with Laurentian (Merschat et al., 2005; Hatcher et al., 2007a; Merschat and Hatcher, 2007). South of the Grandfather Mountain window, the Ashe Formation contains eclogite dated at 459.0 +1.5/-0.6 Ma (Miller et al., 2010), and 480–440 Ma Sm-Nd and Rb-Sr garnet and hornblende ages (Goldberg and Dallmeyer, 1997). These isotopic ages represent subduction and closure of the Iapetus ocean and regional metamorphism associated with the Taconic orogeny, respectively (Fig. 16).

The Alleghanian orogeny was the culminating, late Paleozoic collisional event and resulted in the formation of the supercontinent Pangea (e.g., Hatcher, 1989). Based on restored cross sections across the orogen, crustal shortening in the southern and central Appalachians was ~50% (Hatcher, 1989; Hatcher et al., 2007b); the Blue Ridge thrust sheet was displaced as much as 350–400 km to the northwest. Preservation of detrital ages in Chilhowee Group in the Mountain City window, and metamorphic ages in basement not reset by Paleozoic metamorphism suggests the Blue Ridge thrust was a thin thrust sheet, or at least the western edge was thin. ^{40}Ar/^{39}Ar muscovite growth ages ~340 Ma delimit greenschist-facies metamorphism and deformation in Blue Ridge thrust sheet associated with the westward emplacement of the Blue Ridge thrust sheet onto the then actively deforming foreland fold-thrust belt. Further, we tentatively suggest that the Fries fault and Catface fault/ Blue Ridge thrust were contemporaneous, but the Fries fault was a slightly deeper ductile structure. This is supported by the wide deformation zone associated with Fries Fault (2–11 km wide), the intensity of mylonitic fabric decreases to the west, and coarse-grained muscovite and biotite are stable in the S_3 mylonitic foliation within the Fries fault zone. The rapid cooling from amphibole (500° C at ~340 Ma) through muscovite closure (350° C at ~335 Ma) for Ashe Formation rocks of the Gossan Lead thrust sheet is a direct consequence of motion on these ~340 Ma high strain zones. However, to explain the ~340 Ma amphibole ages east of Gossan Lead fault requires greater exhumation moving eastward following mylonitization—explanations may be east side up tilting (or doming) (Stokes, 2013) or unrecognized faults within the Ashe and Alligator Back Formations. Finally, ca. 340 Ma deformation is much older than the generally accepted ages for Alleghanian orogeny, 325–260 Ma (Secor et al., 1986; Hatcher et al., 1989), and fault gouge ages in the Valley and Ridge (280–260 Ma; Hnat and van der Pluijm, 2014). We suggest that the ca. 340 Ma metamorphism and deformation in the Blue Ridge represent the initial emplacement of the Blue Ridge thrust sheet during the Alleghanian orogeny, which closely followed the Acadian-Neoacadian orogeny. After initial emplacement of the Blue Ridge thrust sheet at ~340 Ma, shortening was accommodated by westward translation along the basal decollement, which carried the Blue Ridge to its final position. Additionally, we speculate that periodic changes between partitioning of motion onto orogen-parallel structures like the Brevard fault zone and more inboard structure are due to far-field changes in plate motion (e.g. zippering and rotation, Hatcher, 2002). Development of several large antiformal duplexes beneath the thrust sheet arched the thrust sheet and formed the Mountain City and Grandfather Mountain windows (Bryant and Reed, 1970; Hatcher et al., 2007b).

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Provenance Analysis of Conglomerates in the Neoproterozoic Mount Rogers Formation, SW Virginia

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ABSTRACT

Early stages of breakup of the Rodinian supercontinent along the eastern Laurentian margin are preserved in the Blue Ridge of SW Virginia as ca. 760–750 Ma bimodal volcanic rocks and clastic sedimentary deposits of the Neoproterozoic Mount Rogers Formation (MRF). The stratigraphy and inferred eruptive sequence of the MRF were described in detail by Rankin (1993), who divided the formation into an "upper" section comprising rhyolite lavas and ash-flow sheets, and a "lower" section consisting of bimodal basalt/rhyolite volcanics interlayered with sedimentary rocks. Based on field relationships, rocks of the lower MRF were inferred to be older. However, recent U-Pb zircon ages reported for rhyolites in the upper and lower MRF (Tollo et al., 2012) show that their ages largely overlap within error of the analyses; therefore, the sequence of eruption is uncertain.

Coarse clastic sedimentary deposits in the lower MRF are dominated by arkosic sandstone and conglomerate. The poorly sorted, compositionally immature sediments, rapid changes in lateral and vertical facies, and generally lenticular geometry suggest deposition as alluvial fan deposits. Polymict conglomerates are dominated by boulder- to cobble-sized clasts of rhyolite and granitoids, assumed to be locally derived. Zircon U-Pb geochronology of ten rhyolite clasts and one granitoid from the lower MRF conglomerates yields ages (within error of each group) of 775–780 Ma, ~760 Ma, and ~752 Ma. The 760 Ma and 752 Ma ages are consistent with MRF rhyolites (dated by SHRIMP and TIMS). The youngest of these clast ages demonstrates that the conglomerates of the lower MRF

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must be younger than at least parts of the upper MRF, and thus calls for reassessment of the stratigraphy and structural relationships within the MRF. In addition, the older (>760 Ma) rhyolite clasts appear to signal geological processes that occurred prior to deposition of the MRF, and suggest that magmatic activity related to the initial intracontinental rifting began 20 m.y. earlier than previously known.

INTRODUCTION

Earth history is punctuated by supercontinent cycles, in which major continental masses amalgamate during convergent plate tectonics, only to break apart again and drift separately as plate motions change over time. The most recent and by far the best-studied supercontinent episode was the formation and eventual breakup of the supercontinent Pangaea between approximately 300 million years ago to 200 million years ago, when the modern-day Atlantic Ocean opened and the continental fragments drifted to their present position. The Blue Ridge in Virginia, however, contains vestiges of an older cycle, spanning ~1.2-1.0 billion years ago with assembly of the supercontinent Rodinia, and ending with rifting and breakup about 565 million years ago (Badger and Sinha, 1988; Aleinikoff et al., 1995). Breakup of a continent is accomplished by intracontinental rifting, followed by initial formation of small ocean basins (e.g., the modern East African Rift and Red Sea), and finally continued spreading that leads to an open ocean separating major continental masses. Geological processes characteristic of this environment include eruption of distinctive 'bimodal' volcanics (intermixed silica-rich and silica-poor lavas), and deposition of coarse-grained clastic sediment in down-dropped basins along rift-related faults. Each of these settings forms deposits that may be preserved in ancient rocks, yielding evidence of ancient rifting episodes.

The Mount Rogers Formation (MRF) in SW Virginia and NW North Carolina records volcanism and sedimentation during initial rifting of Rodinia. The upper part of the Mount Rogers Formation constitutes an eruptive center dominated by rhyolite lavas and ash-flow sheets (Rankin, 1993). U-Pb zircon dating of the rhyolites yields ages of ~750-760 million years (Aleinikoff et al., 1995; Tollo et al., 2012). The lower Mount Rogers Formation, as defined by Rankin (1993), comprises bimodal volcanic rocks (basalt and rhyolite) intermixed with clastic sedimentary rocks, predominantly conglomerate and arkose. The contact between the basement intrusive rocks, which range in age from ca. 1.3 to 1.0 billion years (Tollo et al., 2010, 2012), and the lower MRF is generally interpreted as a nonconformity (e.g., Rankin, 1993). However, the presence of ductile fabrics in highly strained rocks at the basement-cover contact led Bailey and Rose (1998) to propose that the contact is a fault, and similar fabrics are prominent along the contact throughout the region (James, 1999; McClellan et al., 2012a). It is possible that a nonconformable contact became a localized zone of high strain during Paleozoic deformation, perhaps due to competency contrast between the more massive granitoids and the weaker volcanic or sedimentary lithologies (McClellan et al., 2012a).

The Neoproterozoic history of the southern Appalachian mountain chain is not well understood, in part because of the limited rock record of this period of time. Although the absolute ages of rocks in the lower MRF have been uncertain until recently, field relationships have led to the assumption that these rocks are older than the massive rhyolites of the upper MRF, and therefore represent the beginning stages of continental rifting. This assumption has been challenged by new U-Pb zircon ages reported for rhyolites in the upper and lower MRF (Tollo et al., 2012). These data show that the ages largely overlap within the errors of the analyses, leading the authors to suggest that all the rhyolites formed during a geologically brief period of volcanic activity, perhaps as short as ~1 million years (Tollo et al., 2012). Significant questions remain, however, some of which may be answered by provenance studies of MRF sedimentary rocks. In particular, is there a record of an earlier Neoproterozoic history, perhaps of rocks that have mostly eroded away? And, if the lower MRF rhyolite is essentially contemporaneous with rhyolites in the upper MRF, as the new dates suggest, what are the stratigraphic and age relationships between the sedimentary and volcanic rocks, and what do they tell us about the sequence and processes of rifting? In this study, we investigate these questions through geochronological and petrological analysis of rhyolite and granitoid lithic clasts in conglomerates of the lower MRF.

GEOLOGIC SETTING OF THE MOUNT ROGERS FORMATION

Previous Work

Jonas and Stose (1939) and Stose and Stose (1944) first described the Mount Rogers 'series' based on exposures of lightly metamorphosed volcanic and sedimentary rock on and in the vicinity of Mount Rogers, the highest mountain in the state of Virginia. Rankin (1970, 1993) subsequently formalized the rock unit as the Mount Rogers Formation. In the latter work, Rankin (1993) named three separate volcanic centers, and described in detail rhyolites of the upper part of the Mt. Rogers volcanic center, differentiating them based on field distribution as well as contrasting composition and mineral content of the rhyolite units. The lower Mount Rogers Formation remained largely undifferentiated on his map, and was described as a complexly deformed mixture of clastic sedimentary rocks, basalt, and rhyolite. Only the felsic volcanic rock was given a formal name, the Fees Rhyolite, for its prominent exposure on Fees Ridge. Subsequent work in the Mount Rogers area concentrated on the distribution, chemistry, and geochronology of the volcanic rocks from the main eruptive centers in the upper part of the formation (e.g., Aleinikoff et al., 1995; Fetter and Goldberg, 1995; Novak and Rankin, 2004; Tollo et al., 2012), the stratigraphy and geochemistry of rocks in the lower MRF (e.g., McClellan et al., 2011; 2012a; 2012b; McClellan and Gazel, 2014), and the Razor Ridge volcanic center to the east (Merschat and Southworth, 2011; Rubin and Tollo, 2012).

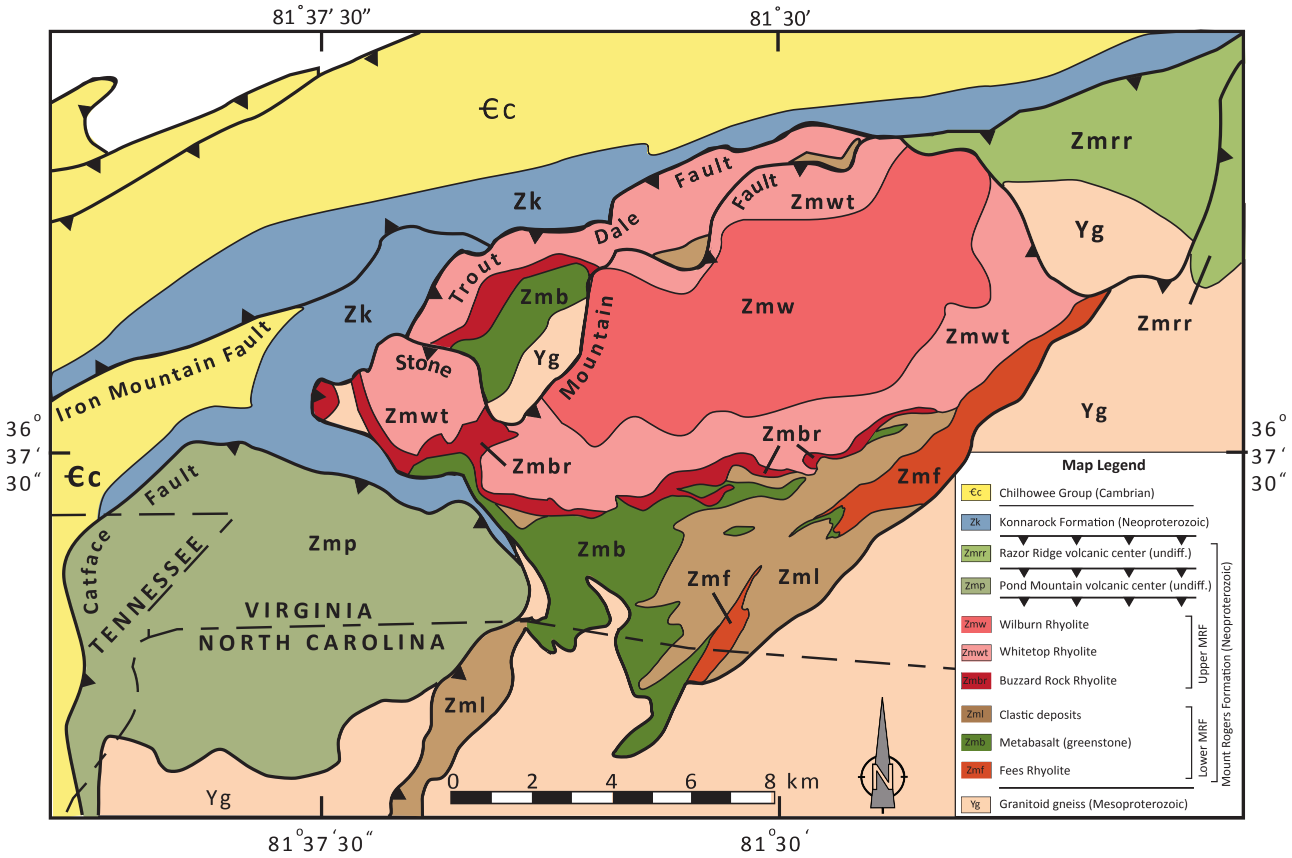

Figure 1. Generalized geologic map of the Mount Rogers area, southwestern Virginia, northwestern North Carolina, and eastern Tennessee (after Rankin, 1993). Contacts in the lower Mount Rogers Formation from Jessee et al. (2011), McClellan et al. (2012a), Brett et al. (2014). Placement of the Stone Mountain fault in North Carolina from Bailey and Rose (1998). Box is approximate location of Fig. 2.

Stratigraphy of the Mount Rogers Formation

Volcanic and clastic sedimentary rocks of the MRF were metamorphosed to lower greenschist facies during Paleozoic tectonism, likely during the earliest stages of the Alleghanian orogeny (Hames et al., 2014). Identification of protoliths is generally straightforward despite the metamorphic overprint and locally intense foliation development. Therefore, the prefix ‘meta-‘ will be omitted in the descriptions below.

Lower Mount Rogers Formation

The lower MRF comprises a diverse assemblage of rock types. The Fees Rhyolite Member (Fig. 1, Fig. 2) is dominated by porphyritic rhyolite with prominent phenocrysts of perthitic alkali feldspar and quartz and lesser plagioclase, which distinguishes this unit from similar rhyolites in the upper Mount Rogers Formation (Rankin, 1993; McClellan and Gazel, 2014). Locally, lithic clasts of both rhyolite and granite are present, as well as fiamme (compressed pumice clasts). The presence of ignimbrite textures, lithic clasts, and fiamme indicate the Fees Rhyolite is dominantly pyroclastic in origin. Basalt flows dominate the lower Mount Rogers Formation in the western part of its outcrop area (Fig. 2), and range from relatively undeformed basalt to greenstone or foliated greenschist, all having the typical chlorite and epidote assemblages characteristic of the low-grade metamorphism. Several different varieties occur, including vesicular and amygdaloidal basalt, homogeneous aphyric greenstone or greenschist, and plagioclase-phyric basalt porphyry (the “turkey track” greenstone of Rankin, 1993). The latter rock may represent a hypabyssal intrusive, but relationships in outcrop show interlayering between vesicular flows and the porphyry, suggesting an extrusive origin. Lindsey (2010) calculated a thickness of approximately 1600 m for lower MRF basalts exposed in the Trout Dale thrust sheet north of our study area (Fig. 1).

Volcanic rocks of the lower MRF are overlain by a sequence of coarse-grained clastic rocks, largely composed of conglomerate, pebbly arkose and lithic wacke (Fig. 2). Conglomerates range from grain- to matrix-supported, and typically contain cobble-sized to boulder-sized clasts of rhyolite and granitoid, with lesser basalt and other lithologies (Fig. 3). In some outcrops, original bedding in the conglomerate can be observed, defined by layers of pebbly sandstone within the conglomerate. Over the outcrop area, conglomerate grades laterally into coarse arkosic sandstone that contain granules and pebbles of vein quartz, granitoid, and rhyolite.

Upper Mount Rogers Formation

Detailed descriptions of the upper Mount Rogers Formation can be found in Rankin (1993). The formation is divided into three rhyolite units (Fig. 1), with an estimated total thickness of approximately 1850 m (Rankin, 1993). The uppermost Wilburn Rhyolite is a well-preserved ash-flow sheet (ignimbrite).

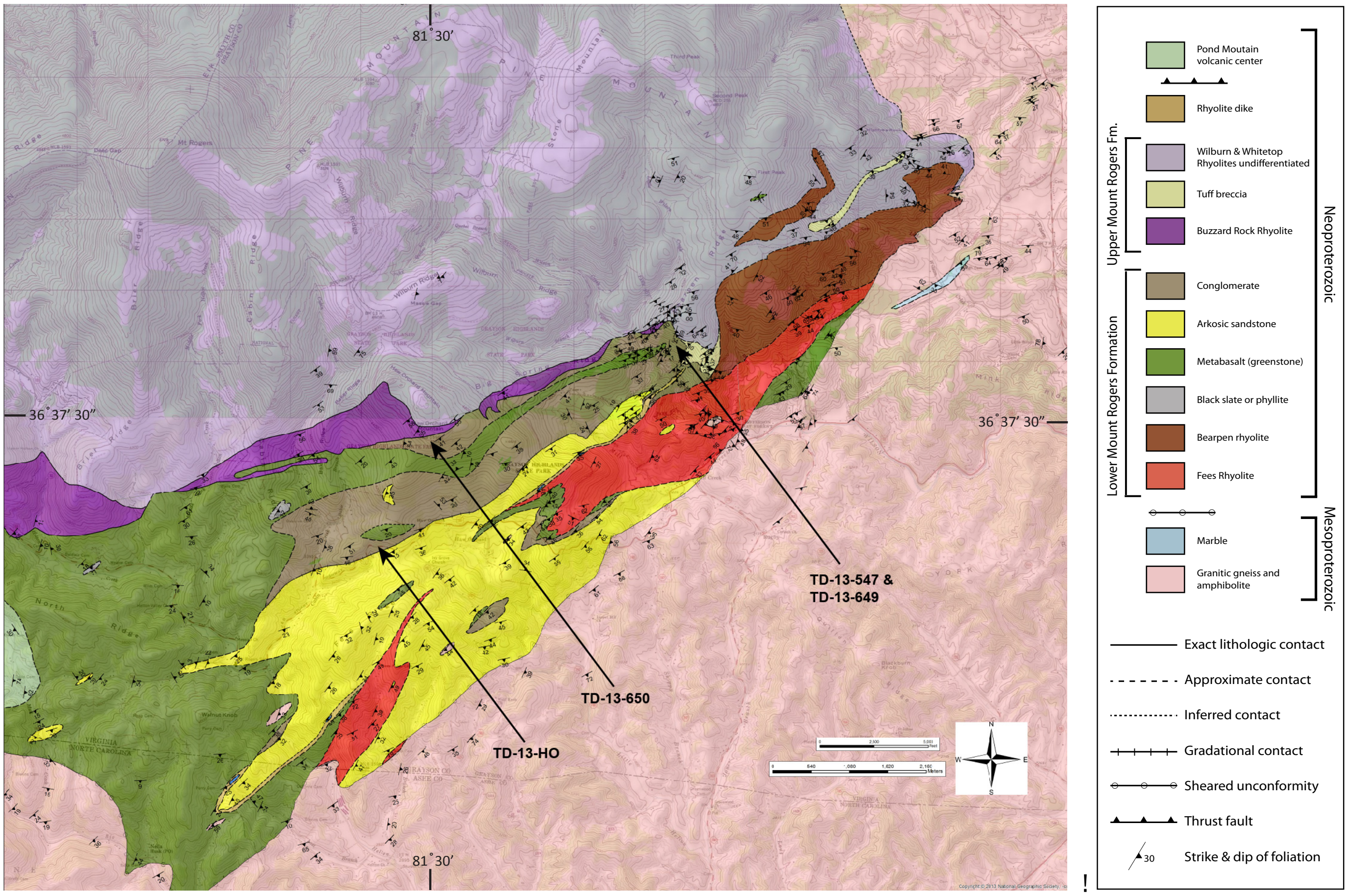

Figure 2. Detailed map of the lithologies of the lower Mount Rogers Formation based on Radford University research 2010-2015. Outcrop pattern of the Buzzard Rock Rhyolite from Rankin (1993). USGS topographic quadrangles. Clast sample locations shown by arrows.

Geochemical interpretations suggest that the magma chamber was compositionally zoned from metaluminous to peralkaline, and this relationship was inverted during eruption (Novak and Rankin, 2004). Fewer published geochemical analyses exist for the other units, but existing data show that the Whitetop Rhyolite is geochemically similar to the Wilburn, whereas the Buzzard Rock Rhyolite is lower in silica and less chemically evolved (e.g., higher in FeOt, TiO2, Sc, V) than the other units (Tollo et al., 2012). Whole rock major and trace-element data indicate that the Mount Rogers rhyolites were formed in a within-plate setting, and are interpreted as part of an A-type suite of plutonic and volcanic rocks related to intracontinental rifting (Novak and Rankin, 2004; Tollo et al., 2004, 2012).

Age of Mount Rogers Formation Rhyolite Units

Stose and Stose (1944) interpreted the Mount Rogers volcanics to be correlative with basalts of the Catoctin Formation, which crop out from southeastern Pennsylvania to central Virginia. This correlation was accepted for decades (e.g., Rankin, 1975). Aleinikoff et al. (1995) applied zircon U-Pb isotopic dating techniques to both units and showed that the upper Mount Rogers volcanic rocks are distinctly older [758 (± 12) million years] than the Catoctin volcanic rocks [564 (± 9) million years]. The Catoctin age agrees with data from similar volcanic sequences that crop out from central Virginia to Maritime Canada, and is now accepted as the age of final rifting of the North American margin of Rodinia and opening of the Iapetus Ocean. In contrast, the Mount Rogers is interpreted to represent an older initial stage of rifting (Aleinikoff et al., 1995; Burton and Southworth, 2010). More recently, Tollo et al. (2012) reported CA (chemical abrasion)-TIMS single zircon ages for the three rhyolite units in the upper MRF, as well as the Fees Rhyolite in the lower MRF. The youngest zircons from each of the units, 755-750 Ma, overlap in age within the analytical uncertainties (Fig. 4), leading the authors to suggest that rhyolites in both the upper and lower MRF were produced during a short-lived episode of eruptive activity.

ANALYSIS OF CLASTIC ROCKS IN THE LOWER MOUNT ROGERS FORMATION

Until recently, the sedimentary rocks of the lower Mount Rogers Formation were not studied in detail. Nonetheless, the immature polymict conglomerates, in particular, possess a remarkable potential to provide information applicable to studies of stratigraphy, sediment provenance, and depositional environment. Below we summarize the results of our recent investigations of these rocks, and explore the implications for understanding timing and processes associated with the Neoproterozoic/Cryogenian episode of intracontinental rifting.

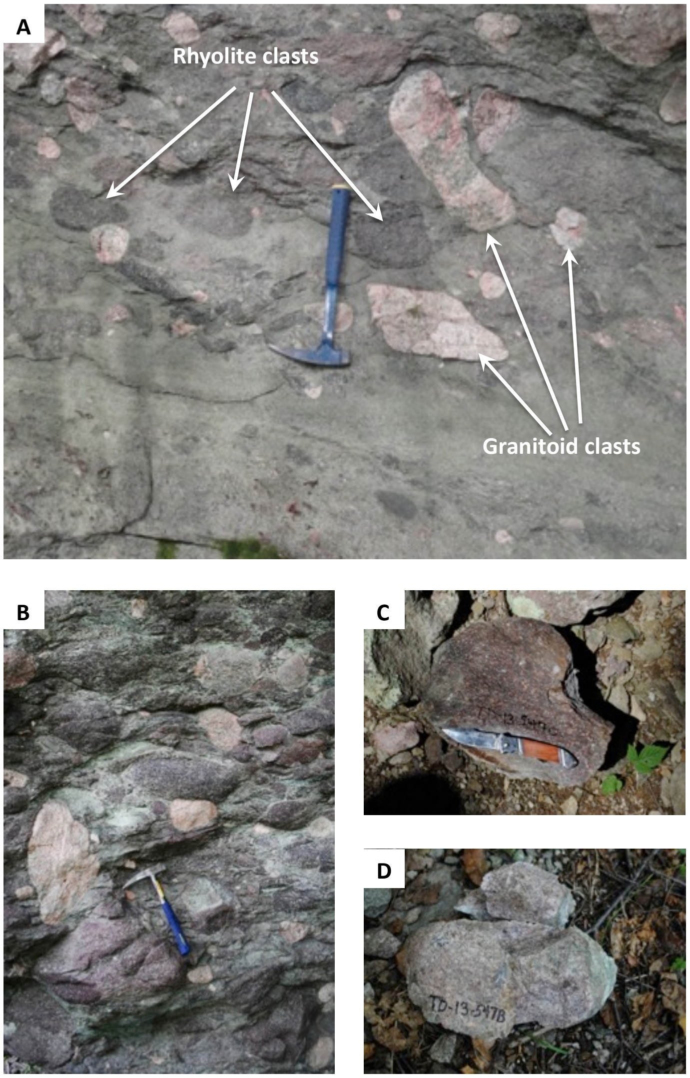

Figure 3. Mount Rogers conglomerate outcrops and clasts. (A) Typical outcrop of Mount Rogers Formation conglomerate, dominated by clasts of rhyolite and pink granite. Original bedding defined by interlayered arkosic sandstone. (B) Representative outcrop of conglomerate from which clasts were collected (TD-13-547; see Fig. 2 for location). Light-colored clasts are granite, dark maroon clasts are rhyolite. (C) Porphyritic rhyolite clast (TD-13-547A). (D) Pink granite clast (TD-13-547B).

Depositional Setting

Detailed mapping and outcrop studies have shown that conglomerates and arkosic sandstones in the lower MRF likely represent a progradational sequence formed during synsedimentary faulting associated with intra-continental rifting of Rodina, in which sediment was eroded off the rift flanks and deposited into subsiding rift basins (e.g., McClellan et al., 2010, 2012b; Yonts et al., 2011; Jessee et al., 2012). In such an environment, sedimentation occurs rapidly by the action of intermittent, high-volume stream flows, debris flows, and sheet floods, and results in fan-shaped accumulations of poorly sorted sediment, generally of locally derived material. The features of the Mount Rogers deposits conform to many of the 23 numbered criteria for alluvial fan deposits as described by Nilsen (1982), as described below. Note that the numbers in parentheses refer to Nilsen's original criteria:

- The majority of clasts in the conglomerates derive from felsic volcanic rocks similar to MRF rhyolites or plutonic rocks similar to Mesoproterozoic granitoids in the region, indicating sediments were deposited relatively close to their source area (1).
- Cobble and boulder-sized clasts, some up to a meter in diameter, are common in MRF conglomerates, requiring transport and deposition by high-energy flows (3).
- MRF sedimentary rocks are compositionally immature (6), containing abundant feldspar and lithic clasts. In addition, they are extremely poorly sorted (4) and have moderate to poor rounding (5) reflecting short distances of transport.
- MRF conglomerates are unstratified or weakly stratified (16), suggestive of deposition by debris flows. Inverse grading, which typically occurs in basal parts of debris flow deposits (Naylor, 1980) is observed in a few localities. The conglomerates are interlayered with, and grade into, better stratified arkosic sandstones that may represent streamflow and sheetflood deposits (14, 16).
- MRF sedimentary structures are limited to normal to inverse grading, planar stratification and occasional cross-stratification, as typical of alluvial fan sediments (12).

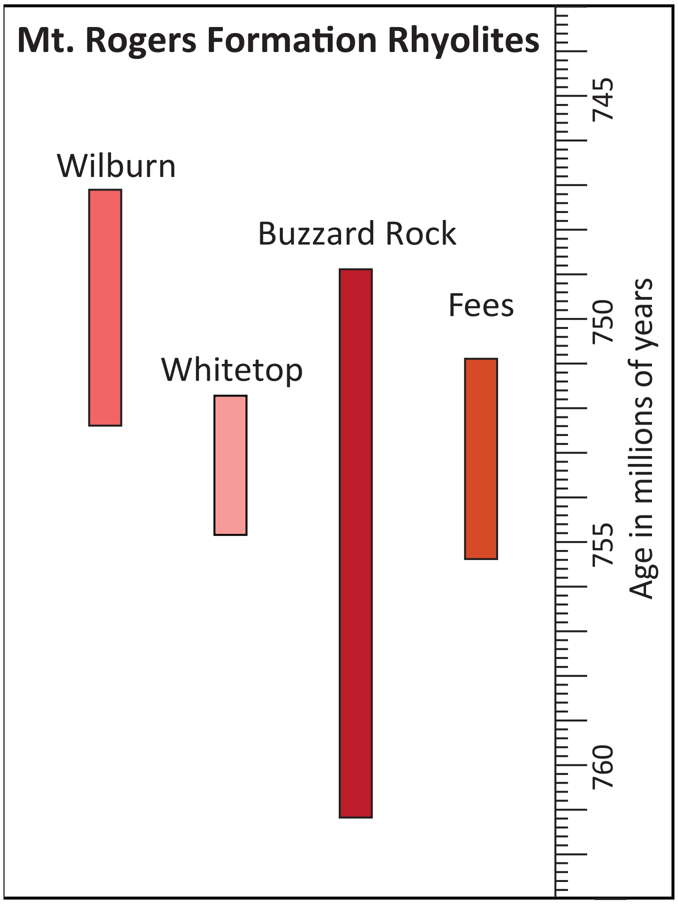

Figure. 4. Uranium-lead zircon CA-TIMS ages of rhyolites from the Mount Rogers Formation (Tollo et al., 2012). Bars represent preferred ages with 2-sigma-error: Fees Rhyolite— 753.1 ±2.7 Ma; Buzzard Rock Rhyolite— 755.0 ± 6.6 Ma; Whitetop Rhyolite— 753.3 ± 2.0 Ma; Wilburn Rhyolite— 749.7 ± 3.1 Ma.

- Although the original geometry of the Mount Rogers clastic deposits has been highly modified by later tectonic deformation, major changes in lateral and vertical facies, and generally wedge-shaped geometry are still evident (Jessee et al., 2012), as discussed below. Alluvial fans are characterized by lenticular or wedge-shaped geometry (13). Sediments are typically distinguished by major changes in lateral and vertical facies, especially in the downfan direction (7), including downfan decreases in average and maximum clast size (8).

We used spatial analysis of the sedimentary rocks to determine the aerial extent, facies variation, and geometry of the interpreted alluvial fan deposits (Jessee et al., 2012). Data collection involved detailed field mapping, and point-counting clasts in outcrops using a 1 m^2 grid divided into 4 cm^2 sections. Data were interpreted using spatial statistics, and three distinct subsystems were defined. Subsystem 1 is composed of dominantly clast-supported conglomerates with matrix/clast ratio of < 1, and clast size commonly > 10 cm in diameter. Subsystem 1 conglomerates are polymict, but rhyolite clasts are most abundant and make up > 50 percent of the total rock volume. Subsystem 2 comprises polymict matrix-supported conglomerates with matrix/clast ratio > 1, and clasts size typically < 10 cm in diameter; rhyolite is again the dominant clast type, but makes up < 50 percent of the total volume. Subsystem 3 is composed of pebbly arkosic sandstone with scattered granule- to pebble-sized lithic fragments and vein quartz clasts. The subsystems are interpreted to represent proximal, midfan, and distal facies respectively (e.g., Larsen and Steel, 1978), and their spatial distribution, both along and across strike, aids in reconstructing the original fan geometry (Fig. 5). We interpret the complex present-day map pattern to arise from the considerable paleotopography that persisted on a broader scale as the sediments were being deposited, and subsequent modification of Neoproterozoic syndepositional basin-bounding faults by Paleozoic folding, faulting, and top-to-NW high strain zones (Jessee et al., 2012; McClellan et al., 2012b).

Variation in Clast Types

Recognizing that several dissimilar rhyolite types occur as clasts, we subsequently undertook a detailed clast count study to distinguish them based on texture (porphyritic vs. non-porphyritic), phenocryst abundance, and matrix color (Stokes et al., 2015). The use of the latter as a basis is complicated by weathering and/or degree of oxidation, but the common correlation between matrix color and phenocryst abundance

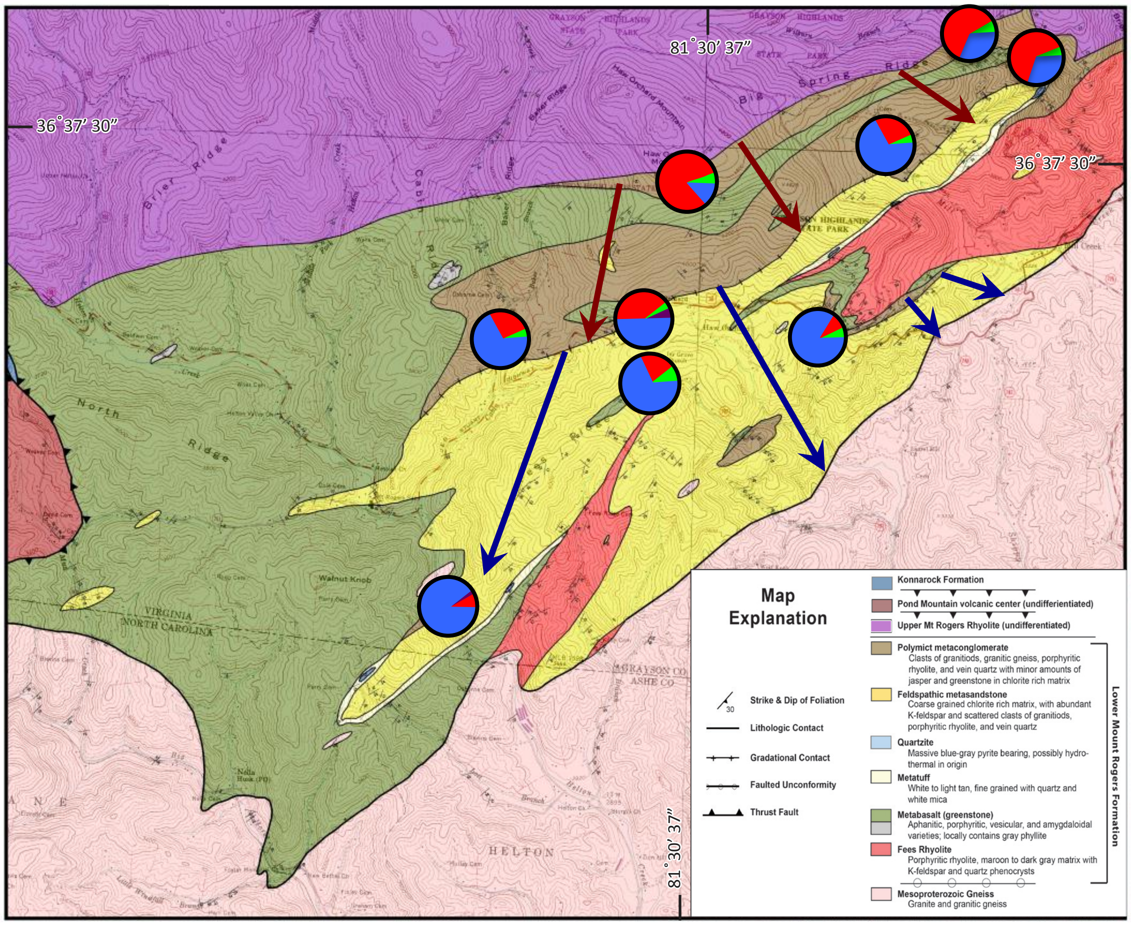

Figure 5. Spatial distribution of facies and clast types in the lower Mount Rogers Formation sedimentary deposits. Pie charts represent the clast-to-matrix ratio and clast type: Blue – matrix; Red – rhyolite clasts; Green – granitoid clasts; Purple – total of other clast types. Arrows represent interpretation of sediment dispersal patterns: Red arrows – proximal fan facies; Blue arrows – distal fan facies.

seems to support its inclusion as a parameter. Following the 'area method' of Howard (1993), a 1 m^2 area was defined on each conglomerate outcrop face, and every clast >2 cm (coarse pebble-sized) within the area was counted. A total of ~130-160 clasts were counted in each of eight separate outcrops. In all outcrops analyzed, rhyolite is the most abundant clast type, composing 40–70% of the clast population. The proportion of rhyolite clasts increases toward the NE in the outcrop belt, consistent with the results of Jessee, et al. (2012). The most common rhyolite clast type contains 3–7 mm alkali feldspar and quartz phenocrysts and lesser plagioclase in a dark gray to maroon-gray aphanitic matrix. Other less abundant types have red, light gray, or green matrix, and some contain alkali feldspar and plagioclase phenocrysts with lesser quartz. Plutonic rocks typically make up ~30% of the clast types. Pink non-foliated granite clasts are found throughout all of the conglomerates counted, whereas white granitoids, rarely with gneissic banding, make up less than a third of the total plutonic clasts. Lithic fragments of sandstone, black slate, and vein quartz occur as a minor component in most outcrops.

AGE AND PROVENANCE OF CONGLOMERATES

The marked variation in rhyolite clast types suggests there may have been multiple sources, and perhaps different ages, of igneous-sourced sediment entrained in the fan deposits. To test provenance interpretations, we visited several conglomerate outcrops (Fig. 2) and collected large (decimeter to sub-meter scale) rhyolite clasts. One granite clast was also collected. Rock saws were used to trim any matrix surrounding the clast and carefully isolate only the clast core. Samples were processed at the U.S. Geological Survey in Denver, Colo., for preparation (crushing, grinding, sieving, water table, magnetic separation, and heavy liquids), imaging (reflected light and electron microscopy), and analysis (laser ablation coupled with high-resolution inductively coupled plasma-mass spectrometry-(LA-ICPMS)-and data reduction).

Rhyolite Clast Petrography

Table 1 summarizes the petrographic characteristics of the analyzed samples. All the rhyolite samples exhibit significant alteration to sericite, and pervasive fracturing at the grain scale. Most are characterized by a felsitic groundmass, composed of a fine-grained to microcrystalline mixture of nearly indistinguishable quartz and feldspar; spherulitic outlines are visible in some at the thin section scale (Table 1). These textures provide compelling evidence of devitrification of an originally glassy groundmass (e.g., Ewart, 1971). In thin section, none of the samples resemble the pyroclastic Fees Rhyolite of the lower Mount Rogers Formation, which is distinguished by a very fine cryptocrystalline groundmass, strongly eutaxitic texture, fiamme (pumice clasts, now recrystallized), and the presence of lithic fragments. Three of the samples do exhibit a moderate to strongly developed foliation (Table 1), however, which may indicate a relict eutaxitic or flow banding.

Despite the overall similarity in groundmass textures, the rhyolite clasts are distinguished by their contrasting phenocryst abundances. Alkali feldspar perthite is dominant in most samples, but the relative proportions of quartz and plagioclase vary (Table 1). Only one sample (TD-13-649A) is devoid of plagioclase, whereas two samples (TD-ad-547E, TD-13-650D) have very little quartz as a phenocryst phase. The significance of the phenocryst assemblages in relation to clast ages will be explored below.

Zircon Analysis

Several clasts from the lower Mount Rogers Formation were analyzed for U-Pb zircon geochronology. Eleven clasts of a dozen sampled yielded sufficient zircon to be dated—ten rhyolite clasts and one granitoid clast. Each clast sample was crushed and ground down to individual mineral grains. The splits then underwent standard mineral separation techniques including water density separation (Wilfley table), magnetic separation (Frantz Isodynamic Magnetic Separator), and heavy liquid separation (Methylene Iodide). Zircon concentrates were then hand-picked and mounted in epoxy and imaged by SEM-cathodoluminescence to identify internal characteristics of the grains. The zircons were analyzed using the Laser Ablation-Inductively Coupled Plasma Mass Spectrometry (LA-ICPMS) U-Pb method. All analyses were completed at the Southwest Isotope Research Laboratory at the U.S. Geological Survey in Denver, Colo. using a Photon Machines Excite 193 nm excimer laser and Nu Instruments ATTOM high-resolution magnetic sector ICPMS.

Zircons from the rhyolite and granitoid clasts in general have textures and U/Th ratios typical of igneous origin. The zircon morphologies range from blocky to euhedral and in some cases display some resorption textures. Oscillatory zoning is the most common internal feature in the zircons; however, sector zoning is relatively common as well (Fig. 6). Another common feature in zircon from rhyolite clasts is an abundance of melt inclusions. These seem to be common in zircon derived from extrusive/hypabyssal felsic rocks of the MRF. All the zircon was imaged to identify xenocrystic cores as an indicator of inheritance. While xenocrystic textures are noted in the zircons, few if any yielded ages that indicated inheritance from an older zircon population, with the exception of the granitoid clast, sample TD-547B (see discussion below).

The U-Pb zircon ages of the clasts fall into three age groupings: ~780-770 Ma, ~764-756 Ma, and ~752 Ma (Fig. 7; Fig. 8). The youngest age group agrees well with the rhyolites of the Fees, Buzzard Rock, Whitetop, and Wilburn rhyolites of the MRF (Tollo et al., 2012) (Fig. 8) and rhyolite from the Razor Ridge area (see Merschat et al., this guidebook). A rhyolite from the Pond Mountain area (759 Ma, see Merschat et al., this guidebook) falls within the 764-756 Ma clast age grouping, but the oldest age group is older than any reported U-Pb zircon ages on rhyolite exposed in the area. Included in the older age group is the single granitoid clast that yielded sufficient zircon to be dated. We interpret the granitoid clast to be ~780 Ma. It contains xenocrystic and inherited zircon components at 1.77 Ga, 1.5-1.4 Ga, and 1.3-1.16 Ga.

TABLE 1. GEOCHRONOLOGY AND PETROGRAPHY OF CLASTS FROM THE MOUNT ROGERS FORMATION CONGLOMERATES

Sample #	Rock Type	U-Pb Zircon Age	% Phenocrysts	Phenocryst Petrography	Groundmass Textures
P-13-HOX	Rhyolite	752.3 +/-2.8 Ma	~25%	**Phenocrysts: Perthite > plag = qtz**. Alkali feldspar perthite grains up to 4 mm. Plag as single grains from ~1.5 mm, or in glomeroporphyritic clusters 5 mm. Quartz ranges from ~0.5-2 mm, and is commonly embayed.	**Groundmass:** Felsitic, patchy. Quartz and perthite phenocrysts commonly surrounded by cryptocrystalline 'halos'. Moderate alteration to sericite and Fe-oxides.
TD-13-547A	Rhyolite	752 +/-4.5 Ma	25-30%	**Phenocrysts: Perthite > qtz > plag.** Alkali feldspar perthite ranges from 2-5 mm. Quartz phenocrysts (0.5-2 mm) are commonly brittely fractured, some embayed. Plag less abundant, grains small (~0.4 mm), some in glomeroporphyritic clusters.	**Groundmass:** Felsitic, with vague spherulitic structures. Cryptocrystalline 'halos' around quartz phenocrysts appear syntaxial to host grain, also contain fine opaques. Moderate alteration to secondary sericite.
TD-13-650D	Rhyolite	753.3 +/-4.1	30-35%	**Phenocrysts: Perthite = plag >> qtz.** Alkali feldspar perthite up to 7 mm. Abundant plag phenocrysts, up to 5 mm. Some glomeroporphyritic clusters of plag and perthite. Very little quartz, most grains are 1-3 mm.	**Groundmass:** Relatively coarse-grained felsitic, some spherulitic textures evident, including around perthite and quartz phenocrysts. Moderate alteration to sericite and Fe-oxides.
TD-13-547C	Rhyolite	759.8 +/- 5.2 Ma	20-25%	**Phenocrysts: Perthite > qtz >> plag**. Alkali feldspar perthite phenocrysts up to 5 mm, highly fractured. Quartz typically 3-4 mm, only a few grains embayed, many fractured. Minor plag - ~0.5 mm. Possible lithic clast. Large opaques.	**Groundmass:** Generally felsitic, but displays strong foliation (flow layering or eutaxitic texture?). Minor alteration to sericite, epidote.
TD-13-547E	Rhyolite	759.3 +/- 3.3 Ma	35-40%	**Phenocrysts: Perthite > plag >> qtz.** Alkali feldspar perthite phenocrysts 3-4 mm. Plag occurs in glomeroporphyritic clusters (+/- perthite) up to 5 mm. Very little quartz, ~1.5-2.5 mm.	**Groundmass:** Felsitic with visible spherulitic structures, especially around perthite phenocrysts. Minor alteration to sericite, opaques. Large opaque grains pseudomorph unknown orthosilicate.
TD-13-649A	Rhyolite	761.4 +/-2.8 Ma	~25%	**Phenocrysts: Perthite > qtz. Plag absent.** Alkali feldspar perthite phenocrysts up to 6 mm, highly fractured and altered to sericite. Abundant rounded quartz phenocrysts range from 1 to 4 mm, larger ones embayed.	**Groundmass:** Felsitic, with patches of much finer grained, microcrystalline quartz (devitrified pumice?), and polycrystalline quartz "clasts" - perhaps amygdaloidal structures? Significant alteration to sericite.
TD-13-649B	Rhyolite	765.3 +/-3.3 Ma	~40%	**Phenocrysts: Perthite = plag > qtz.** Alkali feldspar perthite up to 1 cm in length, highly fractured. Plag, up to 3 mm, some in glomeroporphyritic clusters, some grain display bent twins. Qtz phenocrysts up to 4 mm, are fractured and micro-faulted.	**Groundmass:** Generally felsitic, with obvious radiating to random spherulitic structures, including around perthite and quartz phenocrysts. Weak banding. Moderate alteration to sericite.
TD-13-650A	Rhyolite	780.3 +/-4.6 Ma	~35%	**Phenocrysts: Perthite > qtz = plag.** Alkali feldspar perthite typically 4-5 mm, up to ~ 1 cm. Quartz 0.5-2.5 mm, commonly embayed. Plag fairly abundant but small grains (0.5-1.5 mm).	**Groundmass:** Felsitic, highly altered to sericite and abundant small (< 0.5 mm) opaques.
TD-13-650B	Rhyolite	779.8 +/-3.1 Ma	25-30=%	**Phenocrysts: Perthite = qtz > plag**. Alkali feldspar perthite fractured, up to 6 mm. Quartz up to 4 mm, typically embayed. Most plag grains <1 mm - 2.5 mm, but some large glomeroporphytitic clusters.	**Groundmass:** Felsitic, with moderate development of layering (flow layering or foliation?). Cryptocrystalline 'halos' around quartz phenocrysts. Abundant sericite, calcite, and epidote alteration.
TD-13-650C	Rhyolite	775.7 +/-5.4 Ma	~20%	**Phenocrysts: Qtz > perthite > plag.** Quartz phenocrysts abundant; range from 1 to 4 mm. Alkali feldspar perthite (3-4 mm) fractured, altered to sericite. Fairly abundant plag, up to 4 mm .	**Groundmass:** Felsitic, contains patches of much finer grained, microcrystalline quartz (devitrified pumice?). Highly altered to sericite, minor epidote
TD-13-547B	Granite	779.5 +/-11 Ma	N/A	Dominated by quartz and microcline or microcline perthite; plag 10-15% [classifies as granite, close to alkali feldspar granite]. Overall equigranular, a few larger feldspars up to 9 mm but typical grain size ~2-4 mm. Primary(?) muscovite grains highly oxidized.	**Grain textures**: Quartz commonly undulose. No alignment of grains. Moderate alteration of feldspars to sericite.

SIGNIFICANCE OF CLAST ANALYSIS

Insight into several questions may be gained through this study, including: (1) can the clasts be matched with potential source rocks within the MRF; (2) what are the stratigraphic and age relationships between the sedimentary and volcanic rocks of the MRF, and what do they tell us about the sequence and processes of rifting; and (3) do the conglomerate clasts hold a record of an earlier Neoproterozoic history, perhaps of rocks that have mostly eroded away?

In answer to the first of these questions, the 760-752 Ma clast ages are consistent with ages of upper and lower MRF rhyolites, although based on petrography, correlations with specific MRF units of similar age are not clear. The majority of clast samples of the ~760 Ma and ~752 Ma populations contain plagioclase as a significant component of the phenocryst assemblage. The ~760 Ma clast population appears to have two contrasting rhyolitic sources (Table 1). One is characterized by a moderate percentage of phenocrysts (20-25%), little to no plagioclase, and groundmass textures that may indicate a pyroclastic origin (relict eutaxitic texture, possible devitrified pumice clasts), whereas the other is especially phenocryst-rich (up to 40%), contains abundant plagioclase, and exhibits a felsitic groundmass with obvious relict spherulites. Petrographic characteristics of the oldest clast population (770-780 Ma) vary to some degree (Table 1), but all the samples are relatively rich in quartz phenocrysts.

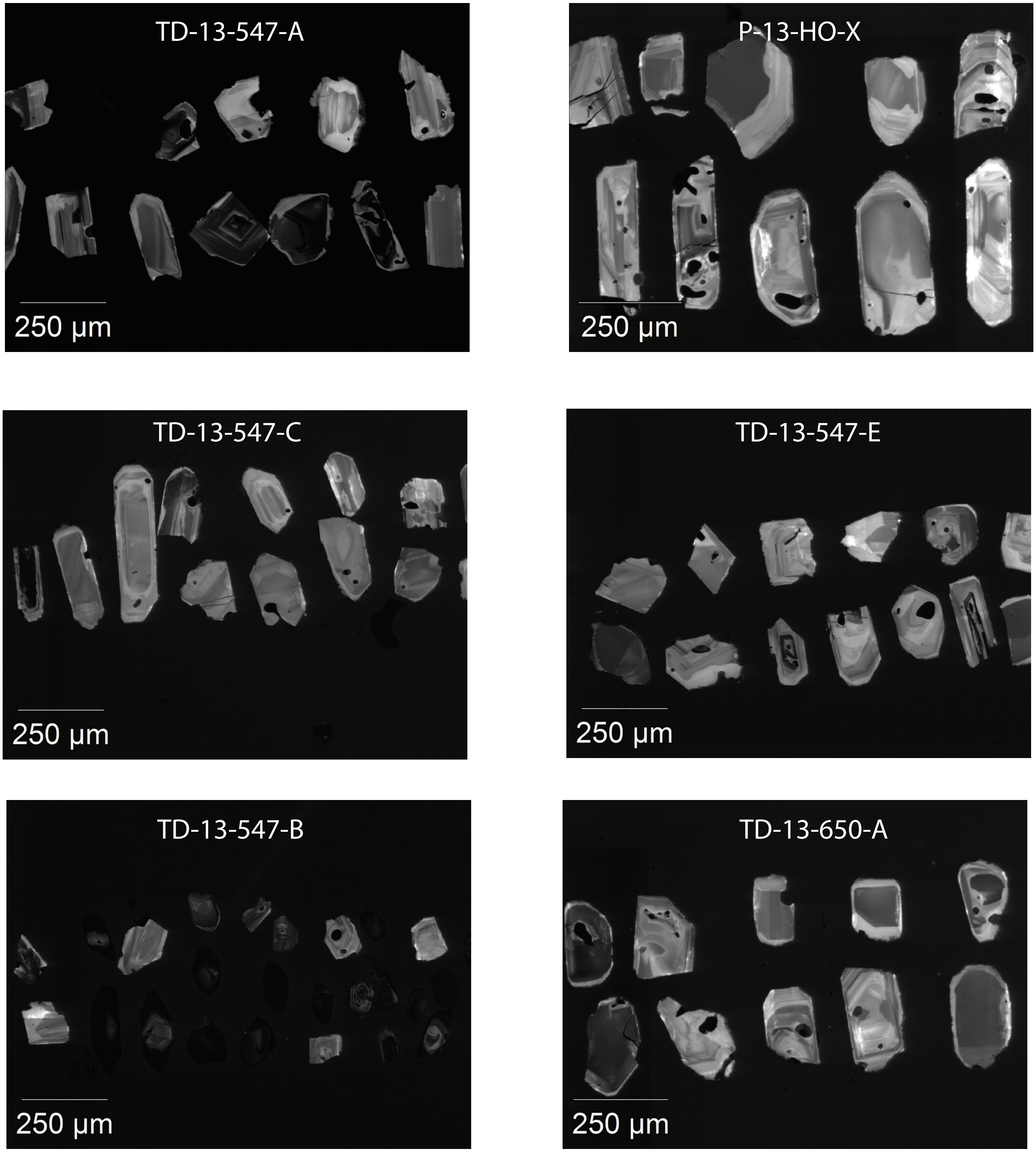

Figure 6. Examples of scanning electron microscope-cathodoluminescence (SEM-CL) images of zircon from clasts from the lower Mount Rogers Formation. Most zircons show typical magmatic zoning; however, complex zonation is exhibited in several grains. The complex domains do not show evidence of inheritance in the individual grain except for sample TD-13-547-B, which is a granitoid clast. In this sample, the dark grains have high uranium concentrations and are generally xenocrystic in nature.

In contrast to the ~752 Ma clast samples, similarly aged Wilburn and Whitetop rhyolites of the upper MRF are dominated by perthite and quartz phenocrysts with little to no plagioclase (Rankin, 1993). Lavas of the Whitetop Rhyolite are phenocryst-poor, and commonly display distinctive flow banding. Texturally, the Wilburn Rhyolite displays diagnostic features of a welded ash-flow tuff, including fiamme, devitrified glass shards, eutaxitic layering, and locally abundant lithic clasts of the underlying Whitetop Rhyolite. Pyroclastic rocks of the Fees Rhyolite Member of the lower MRF contains all three phenocryst phases, although plagioclase makes up a small percentage of the assemblage (Rankin, 1993). Only the

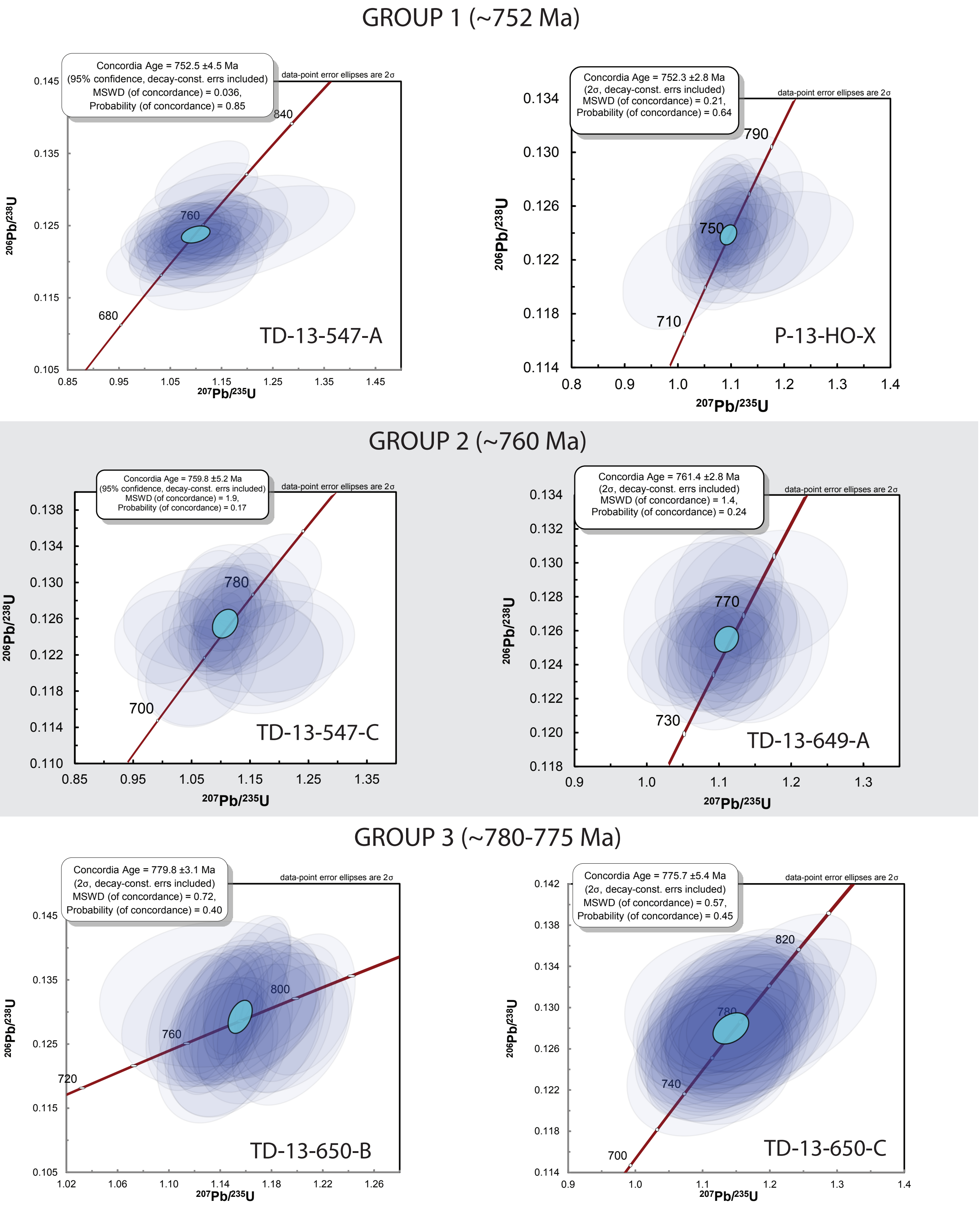

Figure 7. Representative zircon U-Pb concordia diagrams from the three rhyolite clast age groups determined by laser ablation ICP-MS U-Pb zircon analyses. Thirty analyses were conducted for each sample. Highly discordant analyses and statistical outliers were removed for the determination of concordia ages (Ludwig, 1998). All ages reported are at 2-sigma error or at 95% confidence.

Buzzard Rock Rhyolite, which may be as old as ~760 Ma (Fig. 4, Table 1) contains abundant plagioclase, commonly in glomeroporphyritic aggregates with or without perthite (Rankin, 1993). Rhyolite of the Razor Ridge volcanic center to the east (Fig. 1) is petrographically similar to the Buzzard Rock

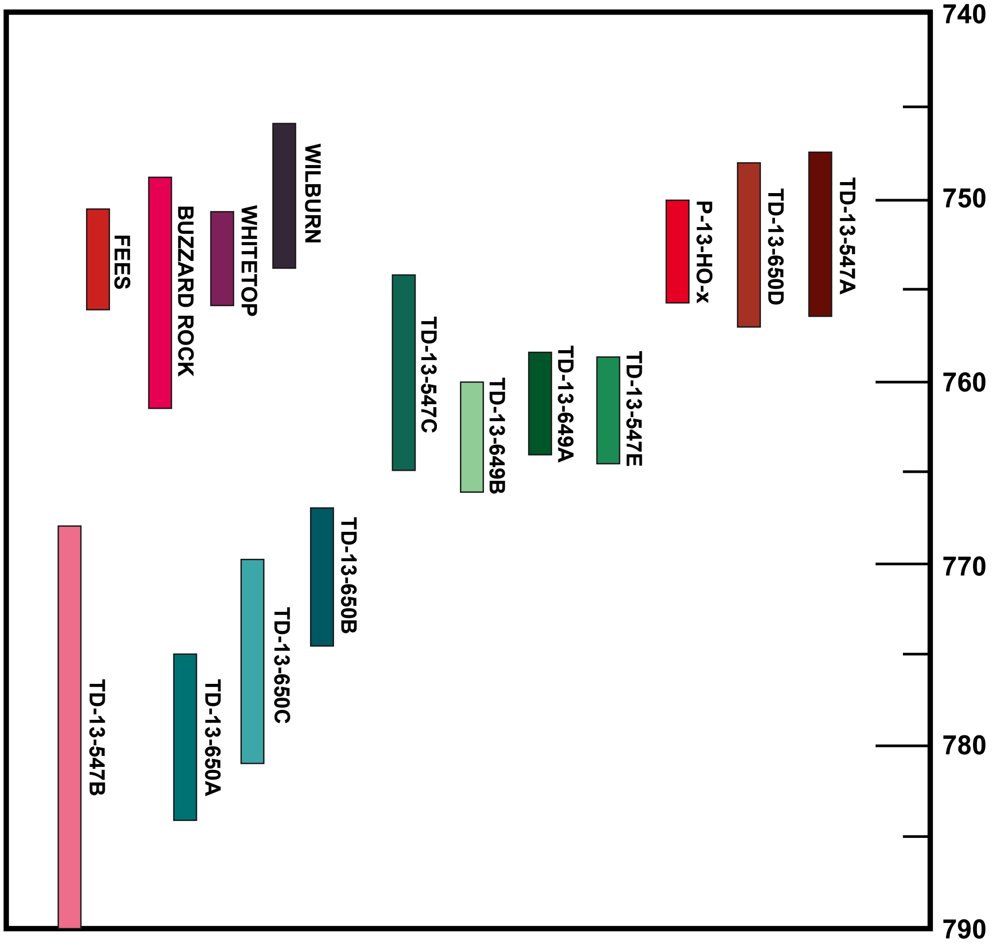

Figure 8. Ages of rhyolite and granite lithic clasts from Mount Rogers Formation conglomerate, determined by laser ablation ICP-MS U-Pb zircon analyses. Bars represent preferred ages within error; see Table 1. Ages of rhyolite units of the Mount Rogers Formation for comparison are from Tollo et al. (2012).

(Merschat et al, 2014), and thus a potential source outside of the main MRF eruptive center.

In regard to stratigraphic and age relationships within the MRF, the data presented here suggest that the assumed ‘lower’ vs. ‘upper’ stratigraphy of the MRF must be reassessed. Following Charles Lyell’s classic principle of inclusions, conglomerates of the lower MRF must be younger than the ~752 Ma clasts they contain. Considering that the age of the Fees Rhyolite in the lower MRF overlaps with rhyolites of the upper MRF (Tollo et al., 2012), and that the conglomerates must be even younger, it is evident that the lower MRF must be, at least in part, younger than most or all of the upper MRF. Work is currently underway (e.g., McClellan et al., 2016) to assess the influence of paleotopography and ancient block faulting on the present-day distribution of lithologies.

A significant find of this project is that the oldest age group of clasts is older than any reported U-Pb zircon ages on volcanic rocks exposed in the area. Rhyolite clasts older than 760 Ma demonstrate that a Neoproterozoic volcanic field existed ~20 m.y. earlier than previously thought. Additionally, not all granitoid clasts were sourced from the Mesoproterozoic basement rocks as formerly assumed, but a significant proportion may represent part of the 780 Ma magma plumbing system. Two observations can be made from the incorporation of these ~780 Ma plutonic rock clasts into the lower Mount Rogers Formation: (1) the granitoid intruded the crust ~780 Ma, cooled, and was rapidly uplifted and exposed to erosion, suggesting very rapid basin formation; and (2) the melt likely incorporated a sedimentary rock that contained recycled materials from the Laurentian continental craton (1.77 Ga Trans-Hudson orogeny/Yavapai-Mazatzal terrane), Granite-rhyolite province (1.5-1.4 Ga), and pre- to early phases of the Grenville orogeny (1.3-1.16 Ga). The entire volcano-plutonic field must have been exhumed and rapidly eroded prior to deposition of the conglomerate, suggesting dynamic processes such as doming which may occur above a mantle plume (e.g., McClellan and Gazel, 2014).

CONCLUSION

Age dating of lithic clasts in conglomerates can provide significant information about cryptic geological processes and events. Determining the age of lithic clasts in conglomerates can go beyond the bounds of traditional detrital zircon studies, in that not only does this technique yield information about the age of sediment sources, but also about specific rock types that were incorporated into the sedimentary deposit. In the MRF, clasts older than 760 Ma demonstrate the existence of a Neoproterozoic magmatic (volcano-plutonic) system of ~20 m.y. longer duration than previously thought, whereas clasts as young as ~752 Ma indicate that the lower MRF must be, at least in part, younger than most or all of the upper MRF.

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Linking the Central and Southern Appalachian Blue Ridge: What We Know and Don't Know about Stratigraphy, Structure, Tectonism, and Regional Correlation Between the Eastern Limb of the Blue Ridge in Central Virginia and the Eastern Blue Ridge in Southern Virginia

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ABSTRACT

The transition from Neoproterozoic Lynchburg Group rocks on the eastern limb of the para-autochthonous Blue Ridge anticlinorium in central Virginia to the fault-bounded Ashe Formation and Alligator Back Formation in southern Virginia has been a source of intense debate and speculation for decades. There are fundamental differences in the tectonogenetic interpretation for these rock packages, despite many similarities in lithology. This problem is compounded by insufficient detailed mapping (1:24,000-scale) in critical key areas of this transition. Lynchburg Group rocks unconformably overlie Mesoproterozoic meta-igneous rocks and underlie Catoctin greenstone on the east limb of the anticlinorium in central Virginia. In southern Virginia, similar metasedimentary lithologies – metasandstone (meta-feldspathic greywacke, meta-quartz arenite, meta-quartz wacke), graphitic schist, and pebble metaconglomerate – and mafic to ultramafic rocks of the Ashe Formation and Alligator Back Formation are faulted onto Mesoproterozoic meta-igneous rocks along the Gossan Lead and Red Valley faults. Internal to the eastern Blue Ridge of southern Virginia, the Rock Castle Creek fault separates Ashe Formation, consisting of Lynchburg-like lithologies, from polydeformed Alligator Back Formation rocks, which crop out in the core of the Ararat River synclinorium. Regional reconnaissance suggests Ashe Formation rocks re-emerge on the eastern limb of the synclinorium. Fundamental and conflicting differences in tectonogenetic models for these rocks compound the problem. The Neoproterozoic Lynchburg Group has long been thought to represent Laurentian margin rift-related rocks, with intrusive mantle-derived dikes and sills of mafic and ultramafic character. In contrast, the Ashe Formation and Alligator Back Formation in southern Virginia and northwestern North Carolina are possibly in part younger, and interpreted to be part of a distal margin accretionary wedge with entrained and tectonically emplaced dismembered ophiolite fragments of mafic and ultramafic rocks. Only detailed mapping in critical areas, coupled with new and emerging geochemical and geochronologic analyses will solve the persistent questions about the various units.

Carter, M.W., and Merschat, A.J., 2016, Linking the Central and Southern Appalachian Blue Ridge: What We Know and Don't Know about Stratigraphy, Structure, Tectonism, and Regional Correlation Between the Eastern Limb of the Blue Ridge in Central Virginia and the Eastern Blue Ridge in Southern Virginia, *in* Merschat, A.J., ed., Geology of the Mount Rogers area, revisited, Blue Ridge, VA–NC–TN: Carolina Geological Society Annual Field Trip Guidebook, p. 43–57.

INTRODUCTION

It has long been recognized that a major transition in the architecture of the Blue Ridge occurs in the vicinity of the James River in central Virginia (Fig. 1). North of the James River, the Blue Ridge consists of a para-autochthonous, north-plunging anclinorium that extends well into Maryland (Cloos, 1947; Southworth and Brezinski, 1996). The core of the Blue Ridge anticlinorium is composed of Mesoproterozoic meta-igneous rocks (basement), whereas Neoproterozoic to Cambrian metasedimentary and metavolcanic cover strata comprise the limbs. The Neoproterozoic Catoctin Formation is the only stratigraphic unit recognized on both flanks (Fig. 1).

South of the James River, the Blue Ridge is an allochthonous composite megathrust sheet that extends southward into North Carolina, South Carolina, Georgia and Alabama (Hatcher, 1978). Low angle thrust faults repetitively stack basement rocks and overlying metasedimentary and metavolcanic cover. The Grandfather Mountain window in northwestern North Carolina demonstrates minimum displacement to the west of up to 80 km (Bryant and Reed, 1970; Boyer and Elliot, 1982), but maximum estimates from seismic reflection profiling suggest more than 350 km of total translation (Cook et al., 1979; Hatcher and Zietz, 1980; Hatcher, 1989). The Cambrian Chilhowee Group is the only unit that may be correlated from the northern anticlinorium to the southern composite megathrust sheet. Strata of the Chilhowee Group occur both on the western limb of the Blue Ridge anticlinorium north of the James River and in the frontal thrust sheets of the western Blue Ridge to the south (Fig. 1).

The Geologic Map of Virginia (VDMR, 1993) suggests a smooth transition from rocks of the eastern limb of the Blue Ridge anticlinorium to those in the fault-bounded eastern Blue Ridge in southern Virginia, but correlating the units is problematic, particularly in the absence of detailed mapping in critical areas south of Lynchburg, Virginia (Fig. 1). Greenstone of the Catoctin Formation on the east limb of the Blue Ridge anticlinorium extends as far south as Lynchburg, where it stratigraphically pinches out or is cut out by faults. Beneath the Catoctin Formation on the east limb of the Blue Ridge anticlinorium is a variably thick package of metasedimentary and metavolcanic rocks of the Lynchburg Group, which rests on an unconformity above basement (e.g., Nelson, 1932; Bailey, 2014). In southern Virginia, however, rocks of the Ashe Formation[1*] (usage of Rankin, 1970) and Alligator Back Formation[2*] (usage of Rankin et al., 1973) comprise the fault-bounded eastern Blue Ridge, which is structurally emplaced above basement of the western Blue Ridge. Fundamental differences between orogenic models developed over decades of independent work in these areas only compound what should be a simple exercise in correlation.

This paper outlines the few similarities and many differences in the stratigraphy, the structure, and the tectonic models between the northern and southern portions of the Blue Ridge in Virginia on the basis of detailed geologic mapping north of Lynchburg in central Virginia, and recent work, both along the Blue Ridge Parkway and in the Mount Rogers area of southern Virginia. Because answers are few, and many are speculative at best, this paper also highlights areas where future detailed work must be conducted to resolve lingering questions in this intriguing region.

SIMILAR LITHOLOGIES – DIFFERENT NAMES

The term "Lynchburg" has been used, since the early work of Jonas (1927), for metasedimentary and metavolcanic strata between basement meta-igneous rocks and the Catoctin Formation on the eastern limb of the Blue Ridge anticlinorium in Virginia from the James River northward to Charlottesville. Wehr (1985) formally defined an internal lithostratigraphic nomenclature for these metasedimentary strata (Fig. 2), building primarily upon earlier models of Nelson (1962) and Furcron (1969). Lynchburg Group stratigraphy of Wehr (1985) is still in use (e.g., Carter, 2008), although not all workers accept this stratigraphy (e.g., Wang, 1991; Johnson et al., 2014), or significantly modify his model (Driggers, 2016). Along the Rockfish River in central Virginia, Wehr (1983, 1985; following Nelson, 1932) recognized at the base of the Lynchburg Group a conglomerate (Rockfish Conglomerate) that lies unconformably on basement meta-igneous rocks. The Rockfish Conglomerate is variably thick and stratigraphically discontinuous, as it is not recognized everywhere at the base of the Lynchburg Group (Bailey, 2014; Driggers, 2016). Above the Rockfish Conglomerate is a unit composed of medium-grained to pebbly, poorly sorted meta-feldspathic greywacke (usage of Pettijohn et al., 1987) with minor metaconglomerate, metasiltstone, and graphitic slate that is mapped as the Thorofare Mountain Formation. This unit is overlain by a unit consisting of coarse-grained to pebbly meta-quartz wackes and cross-bedded meta-quartz arenite that is mapped as the Ball Mountain Formation. Within the Ball Mountain Formation, a distinctive, thick package of graphitic schist is mapped as the Johnson Mill Member. The graphitic schist is overlain by a unit of greenschist-facies metasiltstone and phyllite with isolated beds of medium- to coarse-grained meta-feldspathic greywacke that is mapped collectively as the Charlottesville Formation. Southward from Charlottesville toward Lynchburg, where the metamorphic grade increases from lower greenschist-facies to upper greenschist- to lower amphibolite-facies, the Charlottesville Formation consists of garnetiferous muscovite-rich phyllite (Carter, 2008). The upper part of the Lynchburg Group is meta-volcanic arenite and phyllite near Lynchburg (Carter, 2008). Throughout the Lynchburg Group, there are concordant, tabular, dike- and sill-like bodies of mafic and altered ultramafic rocks, but the larger and more abundant bodies of these rocks occur in the Charlottesville Formation.

During the late 1950's, the term "Lynchburg" was extended into southern Virginia to include greenschist- to amphibolite-facies metasandstone (meta-feldspathic greywacke, meta-quartz arenite) graphitic schist, marble, and mafic and ultramafic rocks occurring in the eastern Blue Ridge (Stose and Stose, 1957; Dietrich, 1959). During the early 1970's, however, Rankin and colleagues named a unit of sulfidic biotite gneiss interlayered with mica schist and amphibolite as the Ashe Formation, for

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[1*] U.S. Geological Survey usage for these units is formations, as defined by Rankin (1970) and Rankin et al. (1973). The Ashe Formation was elevated to Metamorphic Suite by Abbott and Raymond (1984) and the Alligator Back Formation to Metamorphic Suite by Raymond et al. (1989). Raymond (2015) redefined both units.

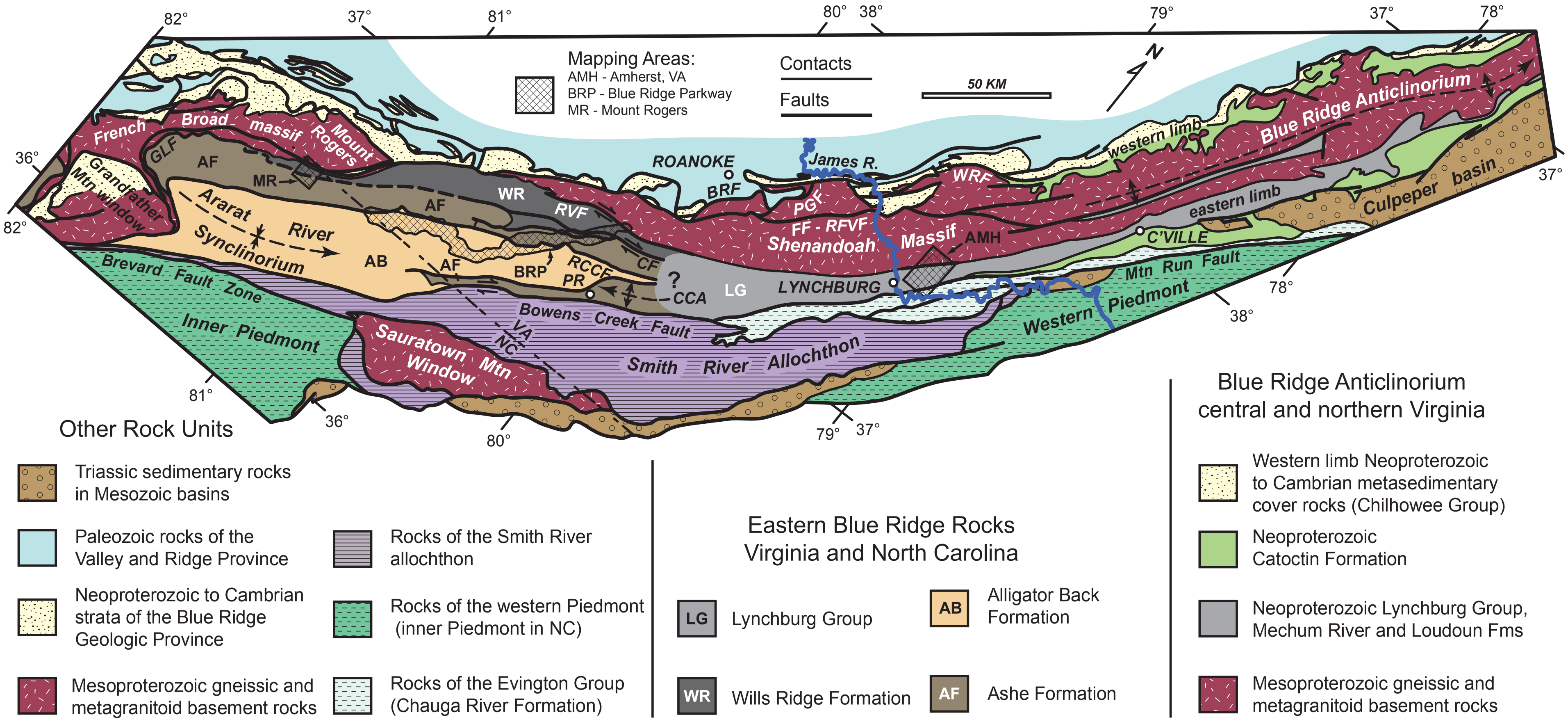

Figure 1. Tectonic map of a part of the Blue Ridge and adjacent regions in Virginia and North Carolina. The transition of lithologies and commensurate rocks names in North Carolina and Virginia (i.e., Wills Ridge Formation, Ashe Formation, Alligator Back Formation, and Lynchburg Group) are the focus of this paper. Detailed mapping in this region includes the Amherst area (AMH) and the Blue Ridge Parkway corridor (BRP) by Carter and Mount Rogers area (MR) by Merschat. The Ararat River synclinorium is labeled on the map. The Cooper Creek anticline (CCA) is cored by rocks of the Ashe Formation. Fault abbreviations: BRF – Blue Ridge fault; CF – Callaway fault; FF-RFVF – Fries-Rockfish Valley fault; GLF – Gossan Lead fault; PGF – Powell Gap fault; RCCF – Rock Castle Creek fault; RVF – Red Valley fault. Other Abbreviations: C'VILLE – Charlottesville; PR – Philpott Reservoir. Map compiled from numerous sources.

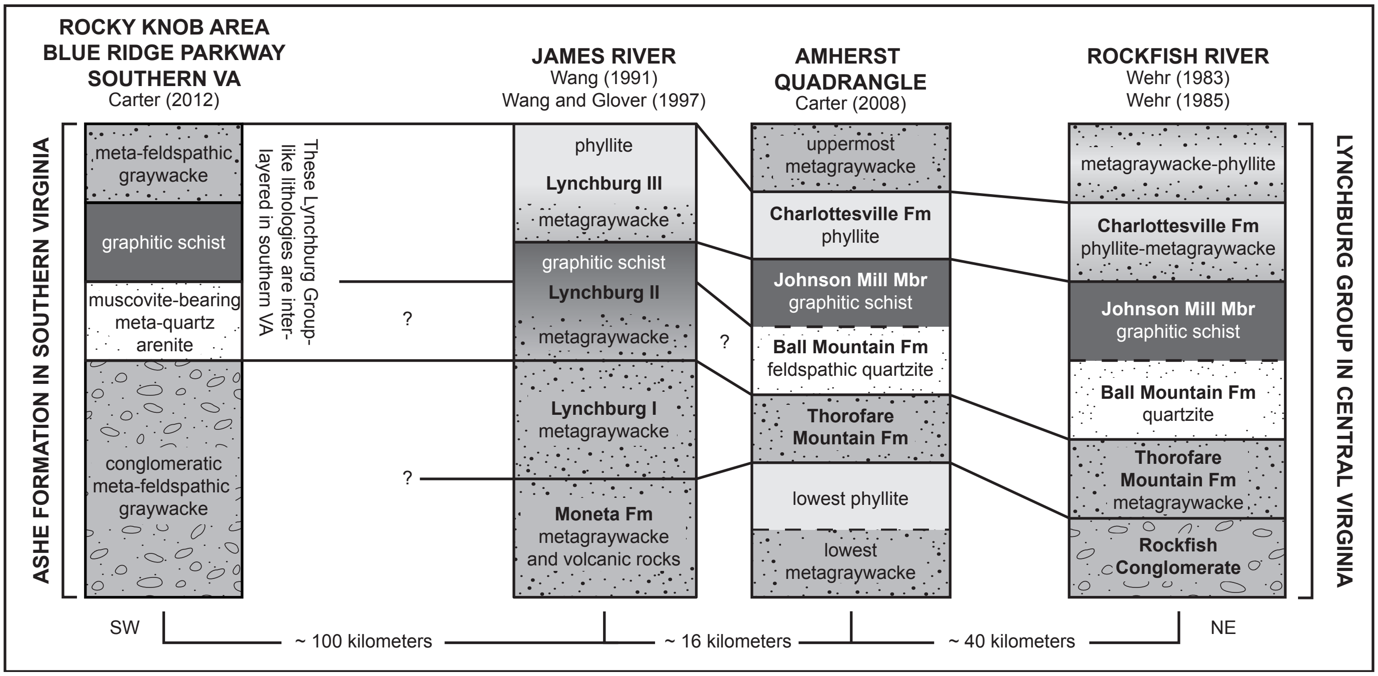

Figure 2. A schematic correlation table for rocks of the Lynchburg Group and regionally associated lithologies of the Ashe Formation, along ~160 km of strike, from the Rockfish River in central Virginia to the Blue Ridge Parkway in southern Virginia, on the basis of detailed geologic mapping and field relations (modified from Carter, 2008; 2010; 2012; Carter and Merschat, 2014).

exposures along the South Fork of the New River in northeastern Ashe County, North Carolina (Rankin, 1970; Rankin et al., 1973). Rocks formerly assigned to the Lynchburg Formation in Virginia by Stose and Stose (1957), Brown (1958), and Dietrich (1959) were included in the Ashe Formation by both Rankin et al. (1973) and Espenshade et al. (1975). Rankin et al. (1973) designated a unit of mica gneiss, "pelite," and amphibolite, with minor amounts of "quartzite" and marble, in the core of the Ararat River synclinorium in southern Virginia and northwestern North Carolina (Fig. 1), as the Alligator Back Formation, and noted prominent pin-striped laminae as a distinguishing feature. Abbott and Raymond (1984) proposed that the Ashe Formation constituted a Metamorphic Suite, with metabasite and ultramafic rocks as principle lithologies interlayered within metasedimentary strata. Raymond et al. (1989) also classified the Alligator Back Formation as a Metamorphic Suite, and Raymond (2015) re-defined the two units.

The recognition of far travelled lithotectonic terranes in the Cordilleran (e.g., Coney et al., 1980) created a paradigm shift in the evaluation of different tectonic and stratigraphic elements of an orogen. Analysis of the similarities and differences of stratigraphic, magmatic, metamorphic and deformational history can be used to recognize internally homogenous, fault-bounded terranes and help resolve the accretionary history of an orogen. Williams and Hatcher (1982) applied this concept to the Appalachian orogen, and combined most of the eastern Blue Ridge and Piedmont into the Piedmont terrane, a suspect terrane accreted to Laurentia. Horton et al. (1989) also recognized that much of the eastern Blue Ridge contained mafic and ultramafic rocks of oceanic affinity and defined the Jefferson terrane, which included the Ashe Formation and Alligator Back Formation in northwestern North Carolina and southern Virginia. Similarly, Raymond et al. (1989) proposed the Toe terrane to recognize the accretionary and mélange-related nature of the eastern Blue Ridge. Rankin et al. (1993) defined a new formation to the northwest of the Jefferson terrane, the Wills Ridge Formation, which contains graphitic schist, phyllite, metasiltstone, metagraywacke, metagrit and metaconglomerate, but devoid of ultramafic rocks. Thus the Wills Ridge was interpreted to be deposited on the Laurentian margin and the contact with the Jefferson terrane represents a pre-metamorphic Taconic suture (Rankin et al., 1993). The Fauquier Group, redefined by Rankin et al. (1993), represented part of the rifted Laurentian margin north of the James River. Hatcher (2002) and Hatcher et al., (2007) included the peri-Laurentian, metasedimentary, magmatic and oceanic rocks of the eastern Blue Ridge and much of the Inner Piedmont in the Tugaloo terrane. Hibbard et al. (2006) included the eastern Blue Ridge in the Piedmont domain, a grouping of Late Neoproterozoic to middle Paleozoic lithotectonic elements that represented the opening and closing of the Iapetus ocean.

THE GROUND TRUTH – CONTACT RELATIONS

It is well established that strata of the Lynchburg Group of Wehr (1985) on the east limb of the Blue Ridge anticlinorium in central Virginia rest on an unconformity above Mesoproterozoic basement meta-igneous rocks (Jonas and Stose, 1939; Bloomer, 1950; Brown, 1953; Bloomer and Werner, 1955; Brown, 1958; Nelson, 1962; Furcron, 1969; Wehr, 1983, 1985; Wang, 1991; Tollo and Arav, 1992; Bailey et al., 2007; Johnson et al., 2014; Driggers, 2016). In the vicinity of Charlottesville, Virginia, and northward, it is also clear that these rocks of the Lynchburg Group are present beneath the greenstone of the Catoctin Formation (Fig. 1). Thus, the age of the Lynchburg Group is constrained by ca. 1.0 Ga basement (below) and the ca. 571-564 Ma age of the overlying Catoctin Formation (Southworth et al., 1994; Aleinikoff et al., 1995). The unit has a maximum age of about 730-680 Ma on the basis of granitoid clasts within the Rockfish Conglomerate (Bailey et al., 2007). These relations become less clear southward toward Lynchburg, as some workers have mapped both the upper and lower contacts as faults (e.g., Espenshade, 1954; Carter, 2008).

The transition from the para-autochthonous east limb rocks of the Blue Ridge anticlinorium in central and northern Virginia to the allochthonous, thrust faulted, eastern Blue Ridge rocks in southern Virginia is significantly complicated by faults that bound discrete domains of different lithologies, structural style, deformational history, and metamorphic conditions. The Gossan Lead fault was defined by Stose and Stose (1957) as a major overthrust that juxtaposed Lynchburg gneiss above mostly mylonitic basement metagranitoid rocks. They reported beds of shiny, slate-like phyllonite and quartzose metasandstone that dip gently SE and folds with NW vergence. This contact was reinterpreted by Rankin (1970, 1971) and Rankin et al. (1972) as an unconformity on the interpretation that a metaconglomerate present locally in the Ashe Formation is a basal conglomerate containing clasts of basement meta-igneous rocks. The amphibolite grade (kyanite zone) Ashe Formation rocks are mylonitic at the contact, however, and in noting those facts, Abbott and Raymond (1984) suggested that Ashe Metamorphic Suite rocks were juxtaposed on lower grade basement meta-igneous rocks along the Gossan Lead fault of Stose and Stose (1957). Rankin et al. (1993) and Rankin (1993) re-assigned the conglomerate and interbedded graphitic schist to their Wills Ridge Formation and placed it immediately above an unconformity on Mesoproterozoic basement. Ashe Formation rocks, which include metasedimentary rock types associated with the Lynchburg Group to the north (e.g., "quartzite" of the Ball Mountain Formation, graphitic schist of the Johnson Mill Member of the Ball Mountain Formation, and phyllite of the Charlottesville Formation), as well as ultramafic rocks were considered by them to structurally overlie the Wills Ridge Formation (which does not contain ultramafic rocks) above a pre-metamorphic fault (Rankin, 1988).

COMPLICATING STRUCTURES

Numerous regional studies from southern Virginia into western North Carolina have concluded that the Ashe Formation and correlative rocks (e.g., Tallulah Falls Formation) are in fault contact with basement meta-igneous rocks and (or) metamorphic rocks of the western Blue Ridge (Raymond and Abbott, 1984; Mies, 1988; Merschat and Wiener, 1988; McSween et al., 1989; Whisonant and Tso, 1992; Trupe et al., 2003; Waters–Tormey and Stewart, 2010; Merschat et al., 2012). The metamorphic rocks west of the Grandfather Mountain window along the contact with basement meta-igneous rocks have dextral strike-slip kinematic indicators of pre-ca. 360 Ma motion (Adams et al., 1995; Trupe et al., 2003). Yet, ^{40}Ar/^{39}Ar muscovite cooling ages across the contact in southern Virginia suggest cooling through 350 °C at approximately 340 Ma and top-to-the-NW motion, with emplacement before 340 Ma (Carter and Merschat, 2014). Northward toward Roanoke, Virginia, the Gossan Lead fault has been correlated with both the dextral transpressive Red Valley shear zone that juxtaposes basement metasedimentary rocks of the Wills Ridge Formation rocks against basement meta-igneous rocks (Henika, 1997; Henika, 2000; Trupe et al., 2003). The Gossan Lead fault has also been correlated with the Callaway fault, which places metamorphic rocks of the Ashe Formation on metamorphic rocks of Wills Ridge Formation and Mesoproterozoic basement meta-igneous rocks (Henika, 2011a; 2011b; 2011c). The Callaway fault thus would be the major suture that juxtaposes the ultramafic rock-bearing metamorphic rocks of the Ashe Formation over ultramafic rock-free metamorphic rocks of the Wills Ridge Formation (Rankin, 1988; Rankin et al., 1993; Horton et al., 1994; Hibbard et al., 2006). Recent mapping across this proposed major structural boundary east of the Mount Rogers area, however, did not reveal any structural variations (Merschat, 2011; Carter and Merschat, 2014). Similarly, there are no significant differences in lithology across this proposed boundary in the vicinity of the Blue Ridge Parkway (e.g., altered ultramafic rocks occur in both the footwall and hanging wall – Carter, 2012; Carter et al., 2016).

The contact between the Alligator Back Formation and the underlying Ashe Formation in North Carolina was originally defined as conformable by Rankin et al., (1973). In contrast, recent mapping demonstrates that the contact is a fault in southern Virginia (Carter, 2012). Along the Blue Ridge Parkway in the vicinity of Rock Castle Creek (Fig. 3), the fault contact is marked by a zone of tectonite with northwest-vergent fabric in which tight to isoclinal folds of mm-thick laminations in pin-striped gneiss and quartz-feldspar layered amphibolite are transposed into penetrative S_2 foliation, with down-dip mineral stretching lineations (e.g., see Carter, 2012; Carter and Merschat, 2014; Carter et al., 2016). This high-strain zone can be followed along strike on the basis of a number of easily recognized lithologic, deformational, and metamorphic features. The fault marks a significant change in both lithology and structural style. Beneath the fault, bedding (S_0 layering) in pebble metaconglomerate, abundant graphic schist, and meta-feldspathic greywacke is present locally, with penetrative axial-planar S_1 foliation. Above the fault, pin-striped compositional layers of the Alligator Back Formation constitute an S_1 foliation (likely transposed S_0; S_0 is not observed in hanging wall rocks), and both an S_2 axial-planar foliation that is locally penetrative and an S_3 cleavage that is locally common, are present. Significant lithologic differences between rocks in the footwall (Ashe Formation) and rocks in the hanging wall (Alligator Back Formation) also exist across the structure (Fig. 4). These include the following. 1) Volumetrically, there are far more ultramafic rock bodies and more graphitic schist in the footwall (Ashe Formation) than in the hanging wall (Alligator Back Formation), although Cattanach et al., (2016) assign a belt of graphitic schist that occurs along a thrust fault to the Alligator Back Formation. 2) Pin-striped, two-mica gneiss of the hanging wall (Alligator Back Formation) is interlayered throughout the unit with garnet-mica schist, which is absent in the footwall rocks (Ashe Formation).

Regional reconnaissance suggests that the Alligator Back Formation structurally occupies the core of the Ararat River synclinorium (Fig. 1). To the west of the synclinorium axis, rocks of the Ashe Formation, which includes rock types like those in both the Wills Ridge Formation and Lynchburg Group, extend to the Gossan Lead fault separating the French Broad massif and the Ashe Formation rocks (Whisonant and Tso, 1992). In the vicinity of the Blue Ridge Parkway, coarse-grained rocks

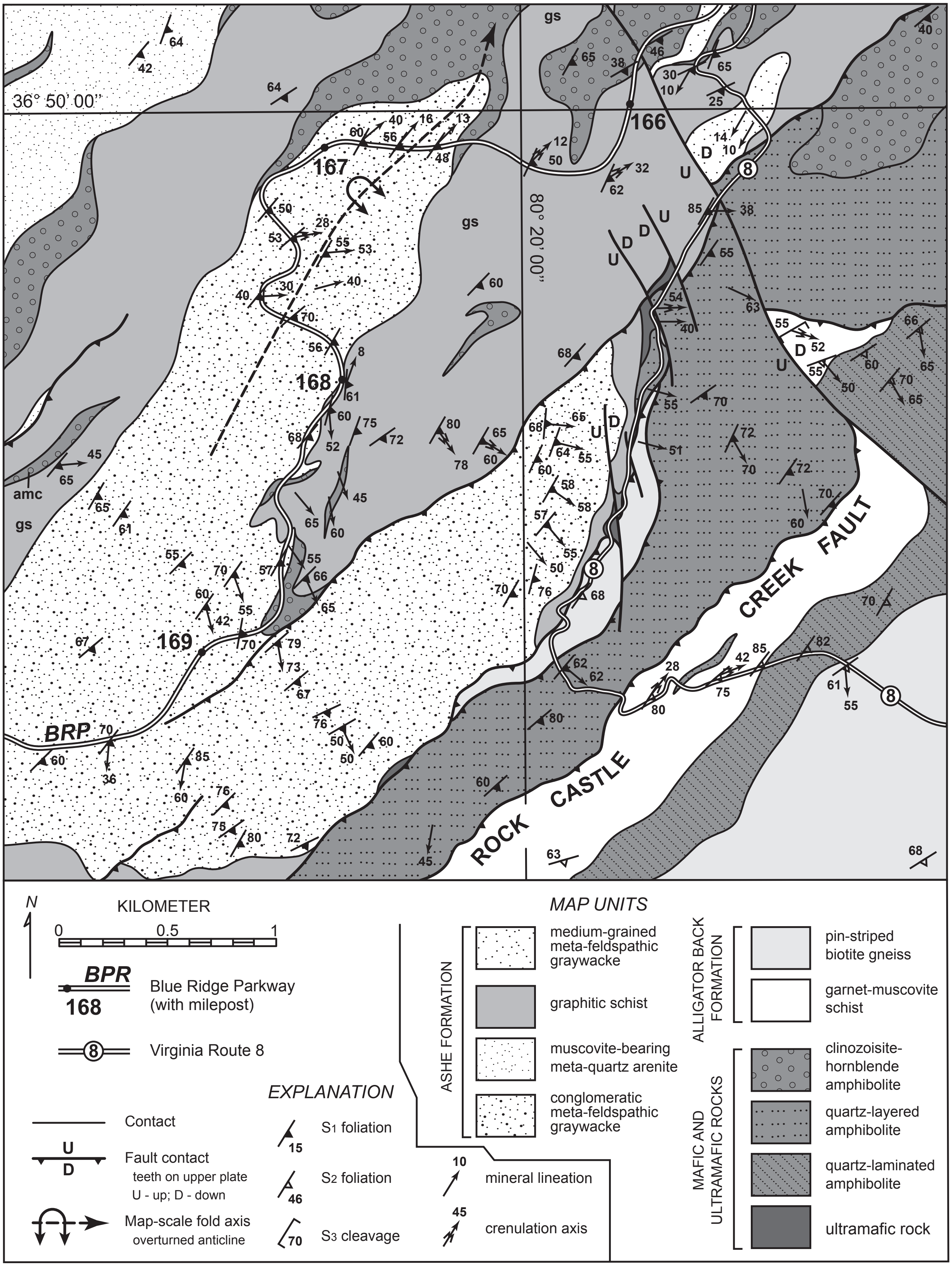

Figure 3: Geologic map in the vicinity of the Rock Castle Creek fault, south and east of the Blue Ridge Parkway, on the Woolwine 7.5-minute quadrangle, Patrick County, Virginia. Pin-striped gneiss, schist, and amphibolite of the Alligator Back Formation occur structurally above conglomeratic meta-feldspathic greywacke, graphitic schist, and ultramafic rocks of the Ashe Formation containing lithologies very similar to those in the Lynchburg Group to the north. Modified from Carter (2012), Carter and Merschat (2014), and Carter et al. (2016).

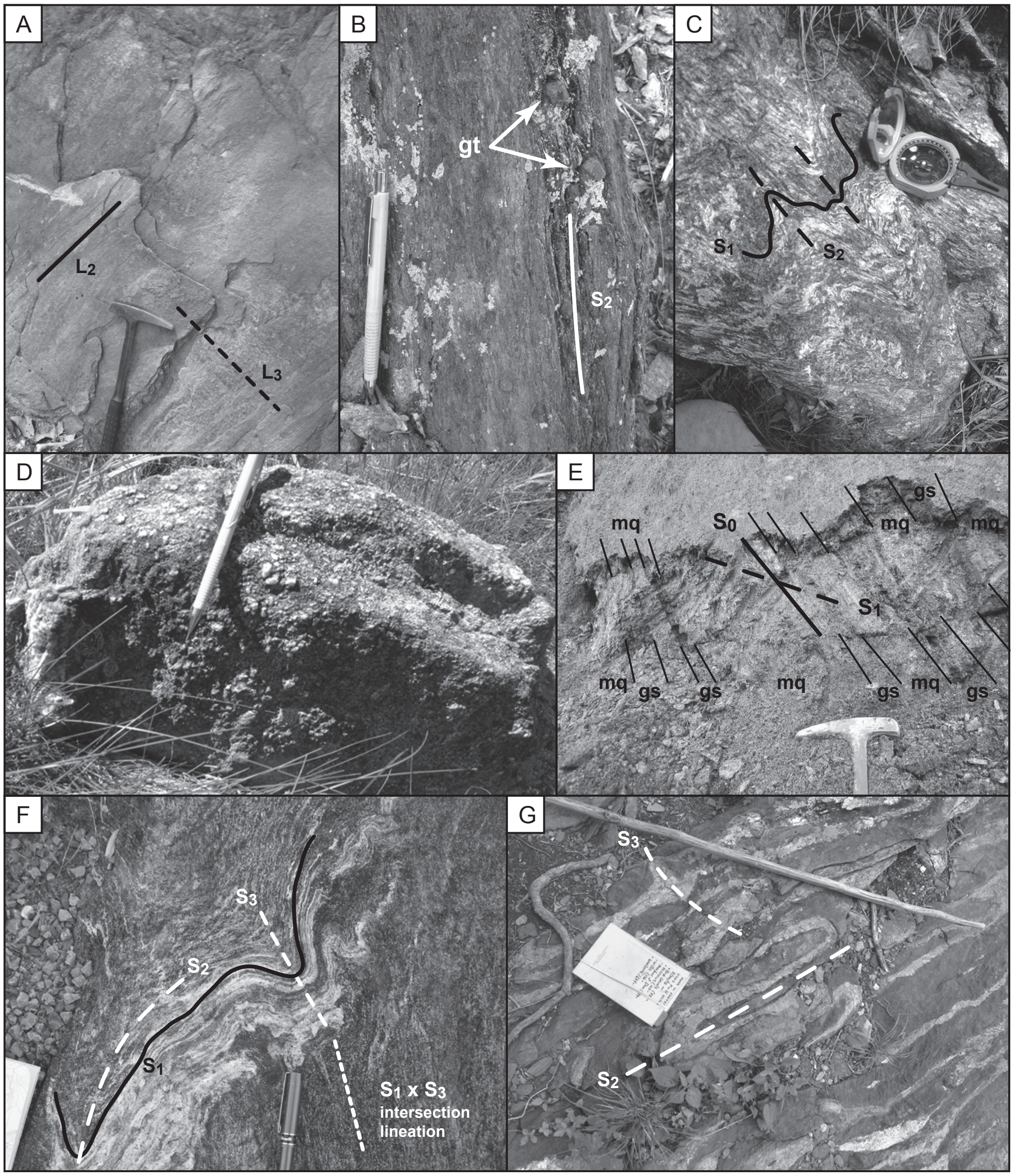

Figure 4. Photographs of rocks and features observed in the vicinity of the Rock Castle Creek fault, Patrick County, Virginia. (A) Strongly foliated and lineated schistose gneiss within the Rock Castle Creek fault zone. The surface in the photograph is penetrative S_2 foliation, with well-developed down-dip L_2 mineral stretching lineation (marked by a solid black line). Crenulation axes (L_3 – marked by a dashed black line) formed perpendicular to L_2 lineation (S_3 crenulation cleavage dips perpendicular into the rock). Exposed part of hammer is approximately 30 cm long. (B) Garnet-mica schist ~ 0.2 km southeast of the trace of the fault, where it crosses VA Route 8. The garnets are up to several cm in diameter (see arrows) in the S_2 foliation (white line). Some may be rotated. Pencil is approximately 14 cm long. (C) Pin-striped biotite-muscovite gneiss of the Alligator Back Formation. S_1 compositional layers comprise the pin-stripes and are characteristically folded into tight to isoclinal F_2 folds, with axial-planar S_2 foliation. Brunton compass is approximately 18 cm long. (D) Granule metaconglomerate in the parking lot across the Blue Ridge Parkway from the Rocky Knob Visitor Center near Mile Post 169 on the Blue Ridge Parkway. This rock is in the footwall of the Rock Castle Creek fault. Note the absence of intense penetrative deformation in the conglomerate. Bedding is evident by gradation in grain-size in some layers. Exposed part of pencil is approximately 13 cm long. (E) Interbedded graphitic schist (gs) and meta-quartz arenite (mq) approximately 10 km northeast of the Rocky Knob Visitor Center, east of Floyd, Virginia (see Fig. 3 for distribution of graphitic schist in the vicinity of the Rocky Knob Visitor Center, in the footwall of the Rock Castle Creek fault). Hammerhead is approximately 16 cm long. (F) Quartz-laminated amphibolite of the Alligator Back Formation, in the hanging wall of the Rock Castle Creek fault, exposed south of the Rock Castle Creek fault, near Meadows of Dan, approximately 10 km southwest of the Rocky Knob Visitors Center. S_1 pin-striped compositional layering (solid black line), composed mostly of quartz, is deformed into tight to isoclinal F_2 folds, with axial-planar S_2 foliation (long-dashed white line); both S_1 and S_2 are deformed by F_3 cross-folds (short-dashed white line). S_1 X S_3 intersection lineation (very short-dashed white line) parallels mineral stretching lineations on the S_1 foliation plane. This outcrop is ~0.7 km west of quartz-laminated amphibolite exposed at Stop 6 of Carter and Merschat (2014). Pen cap is approximately 5 cm long. (G) Quartz-layered amphibolite in the footwall of the Rock Castle Creek fault. This outcrop is on the trail to the crest of Buffalo Mountain, Floyd County, Virginia, ~11 km west of the Rocky Knob Visitors Center. F_2 isoclinal folds deform cm-thick S_1 quartz layers, but axial-planar S_2 foliation is weak to absent (compare to quartz-laminated amphibolite in the hanging wall of the Rock Castle Creek fault). F_2 isoclinal folds are deformed by F_3 cross folds locally. Field book is approximately 20 cm long.

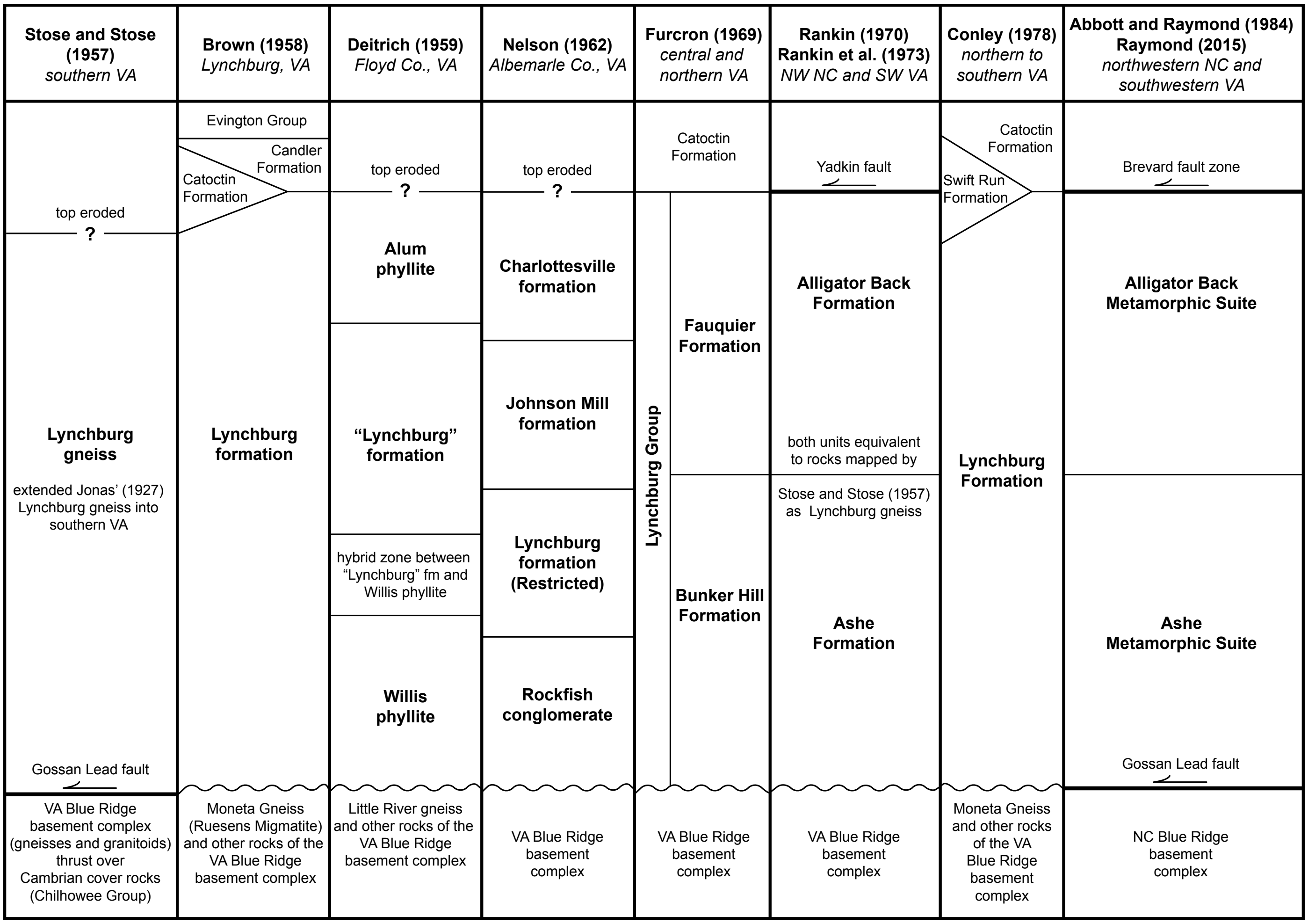

Figure 5. Chart showing the use of geologic nomenclature of eastern Blue Ridge and eastern limb rock units in southern and central Virginia from Stose and Stose (1957) to the present (modified from Carter and Merschat, 2014). Note that the stratigraphic column of Abbott and Raymond (1984) and Raymond (2015) is duplicated on both pages for reference.

core map-scale anticlines (Carter, 2012; Carter et al., 2016), and are flanked by finer-grained graphitic schist and fine-to medium-grained meta-feldspathic greywacke (Carter, 2012; Carter et al., 2016 – Fig. 3). To the east of the synclinorium axis, lower-grade graphitic schist and medium- to coarse-grained meta-feldspathic greywacke re-appear from beneath higher-grade and polydeformed Alligator Back rocks west of Philpott Reservoir (Fig. 1); graphitic schist flanks meta-feldspathic greywacke comprising the core of the Cooper Creek anticline (Conley and Henika, 1970; McCollum, 1989). Farther east, the top of the eastern Blue Ridge sequence is truncated by the Bowens Creek fault and rocks of the Smith River allochthon (Fig. 1).

CONFLICTING MODELS

Since the 1970's, the accepted interpretation of the Lynchburg Group has been one of fluvial-deltaic-deep water submarine fan turbidite deposition, typical for a rifted continental margin basin (Brown, 1970; Conley, 1981; Conley, 1985; Wehr and Glover, 1985; Swab, 1986; Kline, 1991; Bailey and Peters, 1998). Detrital zircon ages of ca. 1.4-0.9 Ga for Lynchburg Group rocks are consistent with Laurentian margin sources (Carter et al., 2006). In contrast, rocks of the Ashe Formation and Alligator Back Formation of the eastern Blue Ridge in North Carolina are viewed by many as an accretionary unit or mélange (Hatcher, 1978; Abbott and Raymond, 1984; Conley, 1985; Horton et al., 1989; Hatcher, 1989; Raymond et al., 1989; Adams et al., 1995; Miller et al., 2006), despite many lithologic similarities of rock types with rocks of the Lynchburg Group to the north. Abbott and Raymond (1984) cite the following evidence for mélange from intercalated ultramafic rocks, noting that: 1) these rocks preserve an earlier deformational history before incorporation and prior to high-grade Paleozoic metamorphism; and 2) the linear distribution of ultramafic bodies suggests that they were emplaced along pre- to syn-metamorphic faults within the accretionary wedge. Interestingly, similar evidence is used in central Virginia to interpret linear mafic to ultramafic rocks within the Lynchburg Group as dikes and sills formed through fractional crystallization of a picritic-basaltic magma related to Neoproterozoic Iapetan rifting (Wang and Glover, 1997; Jensen, 2012). Petrologic and geochemical studies indicate mafic and ultramafic rocks of the Ashe Formation and Alligator Back Formation formed in an oceanic setting (Misra and Conte, 1991; Raymond et al., 2001, 2003, 2016). Misra and Conte (1991) concluded that amphibolites of these formations have a MORB-like geochemical signature. Investigations into the petrology and geochemistry of ultramafic bodies from northwestern North Carolina indicate the bodies represent dismembered ophiolitic fragments (Raymond et al., 2001, 2016). Further, chromium spinel compositions from many of the ultramafic bodies suggest

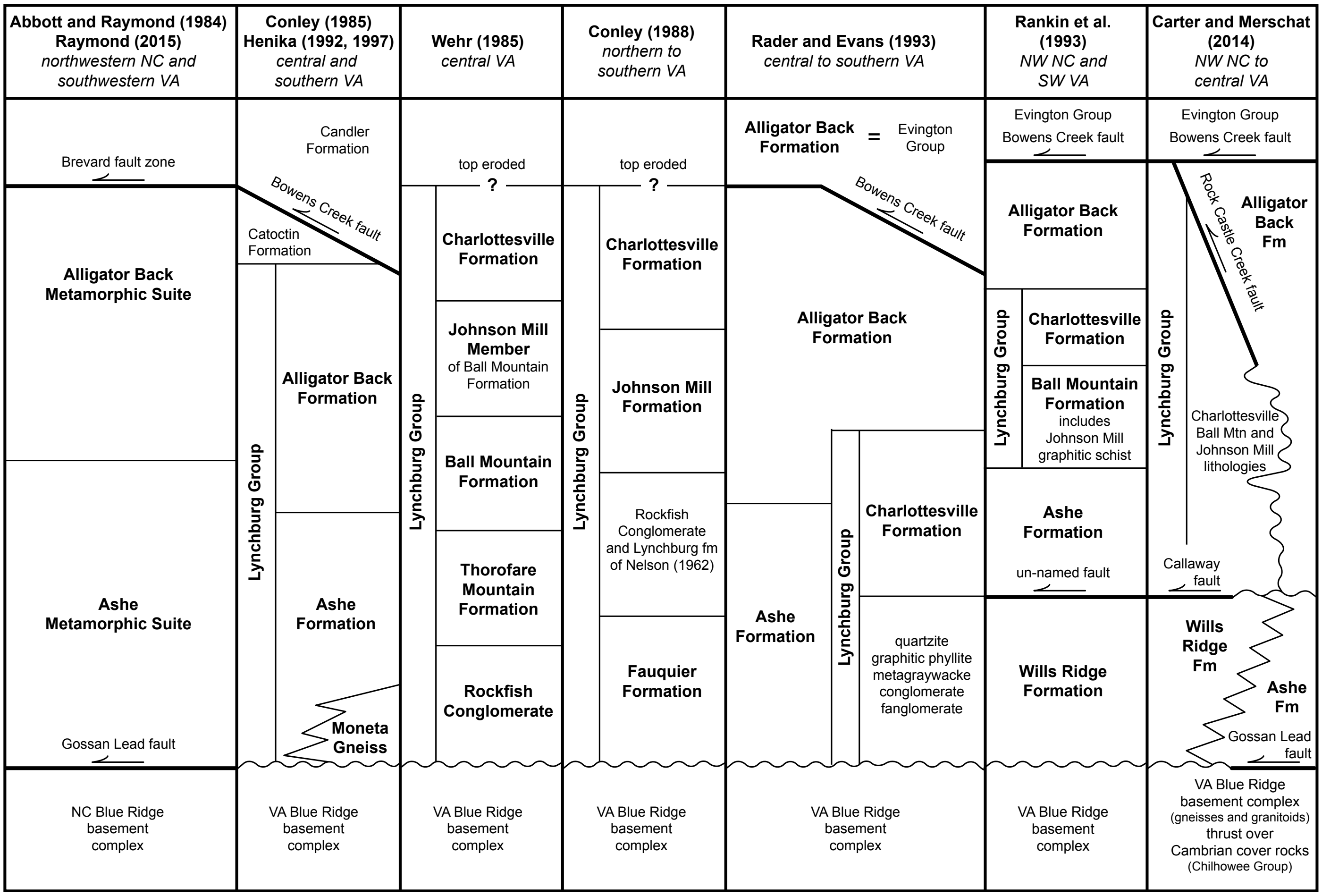

Figure 5. Continued.

a supra-subduction zone setting (Raymond et al., 2003).

Despite these conflicts, eastern Blue Ridge stratigraphy of North Carolina has historically been applied to central Virginia, as the lower (Ashe) and upper (Alligator Back) formations of the Lynchburg Group (Rankin et al., 1972; 1973; Conley, 1985; Henika, 1992; Rankin et al., 1993; Fig. 5). Scheible (1975) and Henika (1991) suggest that the Alligator Back Formation of Rankin et al. (1973) represents the upper part of the Lynchburg Group type section along the James River. Moreover, strata previously assigned to the Evington Group of Espenshade (1954) east of the Bowens Creek fault (Slippery Creek Greenstone, Mount Athos Formation, Pelier Schist, Arch Marble, and Joshua Schist, as designated by Brown, 1951; 1958), were considered to be equivalent to the Alligator Back Formation, and thus the upper Lynchburg Group (Henika, 1992; Rader and Evans, 1993). Given the rock types, this correlation suggests that graphitic schist is common in the Alligator Back Formation, but is a minor component of the Ashe Formation, where it marks the tops of thick, fining-upward graded sequences (Rader and Evans, 1993, pp. 29-30). In southern Virginia, graphitic schist is a major lithology in both the Ashe Formation and Wills Ridge Formation, and from southern Virginia southward, graphitic schist is rare in rocks assigned to the Alligator Back Formation, except along faults (e.g., Cattanach et al., 2016).

Regional correlations remain a problem. A tectonic model on the basis of the distribution of ultramafic rocks was devised by Rankin et al. (1993). According to his model, the Wills Ridge Formation was deposited on Laurentian continental crust and therefore should contain no ultramafic rocks. The Lynchburg Group (Charlottesville and Ball Mountain formations, including graphite schist of the Johnson Mill Member) was deposited on oceanic crust, and therefore contains numerous ultramafic rocks. An association of ultramafic rocks of oceanic affinity and Lynchburg Group sediments only makes sense, however, if the Lynchburg rocks were deposited on that crust, in which case, the mafic and ultramafic rocks would not likely be dikes cutting the stratigraphic section. In this scenario, the Alligator Back Formation, which generally lacks ultramafic rocks, occurs stratigraphically above the Ashe Formation, which is rich in ultramafic rocks. The Evington Group (Brown, 1953) is a separate sequence of rocks exposed east of the Bowens Creek fault and has no clear relationship to the preceding two units. Stratigraphic similarities and structural position east of the Bowens Creek fault suggest that the Evington Group in Virginia, however, may be the northern equivalent of a unit of quartz schist, mica schist, and graphitic schist in South Carolina and North Carolina named the Chauga River Formation (Hatcher (1972).

WHAT ELSE DON'T WE KNOW?

With the publication of the geologic map of east Tennessee, Rodgers (1953) provided a list and short description of outstanding projects for future geologic mapping, which "should provide structural information of more than merely local importance, such as information on the interrelation of major structural features or on the mechanics of the deformation." This paper follows in the footsteps of Rodgers (1953) with the following shorter list and discussion of future research directions in the Blue Ridge of Virginia, to resolve the many lingering questions concerning correlations between the eastern limb of the Blue Ridge anticlinorium in central Virginia and the eastern Blue Ridge to the south:

1) We do not know the full extent of the Red Valley ductile deformation zone. The Red Valley ductile deformation zone needs to be traced to the northeast and mapped in detail. Henika (1997) showed that the dextral transpressive Red Valley shear zone terminates on the west limb of a SW-plunging map-scale fold with a core comprised of Mesoproterozoic basement meta-igneous rocks, approximately 7 km west of Lynchburg and 3 km southwest of the type section of the Lynchburg Group. Carter (2008), however, characterizes the contact between the basement meta-igneous rocks and the overlying metamorphic rocks of the Lynchburg Group approximately 15 km north of Lynchburg as a top-to-northwest thrust fault. Yet most have argued that the contact between the basement meta-igneous rocks and the overlying metamorphic rocks north of the James River is stratigraphic rather than structural (e.g., Nelson, 1932; Jonas and Stose, 1939; Bloomer, 1950; Brown, 1953; Bloomer and Werner, 1955; Brown, 1958; Nelson, 1962; Furcron, 1969; Wehr, 1983, 1985; Wang, 1991; Tollo and Arav, 1992; Bailey et al., 2007; Bailey, 2014; Johnson et al., 2014; Driggers, 2016). More than 50 km of the trace of the Red Valley fault through southern Virginia has not been mapped at a scale larger than 1:100,000, including the region at the fault terminus west of Lynchburg. Detailed geologic mapping on seven 7.5-minute quadrangles, coupled with structural analysis, high-resolution geophysical surveys, and ^{40}Ar/^{39}Ar geochronology, may pin the fault location, determine kinematic variation along strike, and provide much needed data for regional fault and terrane reconstructions.

2) We do not know the details of the relations between rocks of the Lynchburg Group and Ashe Formation and Alligator Back Formation rocks. Detailed mapping of approximately nine 7.5-minute quadrangles southwest of Lynchburg, at a scale of 1:24,000, is needed to resolve correlation issues between the Lynchburg Group rocks to the north and rocks to the south. Graphitic rocks of the Johnson Mill Member (formation of Nelson, 1962) can be traced without break from north of Charlottesville to the type section of the Lynchburg Group along the James River in Lynchburg. Four tiers of 7.5-minute quadrangles occupy the area between outcrops of graphitic schist on the James River, as well as outcrops of abundant graphitic schist on the Rocky Mount 7.5-minute quadrangle (McCollum, 1989). To the west on the Callaway 7.5-minute quadrangle, similar rocks occur (Carter, 2012; Carter et al., 2016). Are these belts of graphitic schist the same, indicating that graphitic schist in the Lynchburg Group is a marker bed that can be traced into southern Virginia and can define major structural folds and faults or were conditions in the depositional basin at that time conducive to repeated deposition of carbon-rich sediments?

3) We do not fully understand the relations between the Ashe Formation and the Alligator Back Formation. The largest-scale mapping currently available that crosses the entire width of the eastern Blue Ridge in southern Virginia is the 1:250,000-scale map of Espenshade et al. (1975). Recent detailed mapping along the Blue Ridge Parkway (Carter, 2012; 2014; Carter et al., 2016) and regional reconnaissance (Carter and Merschat, 2014) raise the following tantalizing new questions that can only be addressed with new detailed 1:24,000-scale mapping across seven 7.5-minute quadrangles north of latitude 36°45′ and west of Austinville, Virginia. Important questions include: a) Is the contact between the Ashe Formation and Alligator Back Formation along the length of the west limb of the Ararat River synclinorium a structural contact, as first proposed by Raymond et al., (1989), and confirmed in southern Virginia by Carter (2012); b) How far east do high-grade and polydeformed rocks of the Alligator Back Formation extend and is the east limb contact in the Ararat River synclinorium structural – e.g., what is the contact relation between the Alligator Back Formation and lower-grade graphitic schist and meta-feldspathic greywacke at the nose and west flank of the Cooper Creek anticline (Conley and Henika, 1970; McCollum, 1989)? and c) Would detailed 1:24,000-scale mapping along the New River in northwestern North Carolina, where lower metamorphic grades reveal a blurred lithologic distinction and contact relations between the Ashe Formation and Alligator Back Formation, as noted by Carter and Merschat (2014), force re-definition of these units?

4) We do not understand regional variations and tectonic significance of ultramafic rocks of the Blue Ridge Geologic Province in Virginia and North Carolina. Some work has been done independently on ultramafic rocks of the Lynchburg Group in central-western Virginia (e.g., Burfoot, 1930; Hess, 1933; Conley, 1986; 1987; Wang and Glover, 1991; 1997; Weiss, 1999; Jensen, 2012), and considerable work has been done on ultramafic rocks of the Ashe Formation in the eastern Blue Ridge in North Carolina and southernmost Virginia (e.g., Misra and Keller, 1978; Raymond and Swanson, 1981; Swanson, 1981; Scotford and Williams, 1983; Hatcher et al., 1984; Misra and McSween, 1984; McSween and Hatcher, 1985; Hopson, 1989; Misra and Conte, 1991; Tenthory et al., 1996; Abbott and Raymond, 1997; Berger at al., 2001; Raymond et al., 2001; 2003; Warner, 2001; Peterson et al., 2004; Swanson et al., 2005; Swanson and Raymond, 2010; Warner and Swanson, 2010; Raymond and Merschat, 2011; Raymond et al., 2016). It's now time to study these rocks as a package in order to answer the following basic questions: Chemically, are these rocks similar or different? Raymond et al. (2016) suggest that there are regional variations. If the Lynchburg ultramafic rocks are dikes and sills, why then did they not create easily observed thick contact metamorphic areoles in the wet sediments into which they intruded? If they are ophiolites, then can the faults along which they were emplaced within the metasedimentary strata be identified, and reconciled with regional kinematics? If Lynchburg and

Ashe ultramafic rocks are different, then where do these rocks make a transition from mantle-derived dikes and sills to ophiolites, as this transition site will mark the major hinge zone (Wehr and Glover, 1985) or suture between the Neoproterozoic paracontinental rift-related rocks and likely Paleozoic distal marine sediments and oceanic crust?

5) We do not have the needed constraints on ages of rock units. Age relations of the various units are still loosely constrained on the basis of field relations and they range from Neoproterozoic to early Paleozoic. Detrital zircon studies (Bream et al., 2004; Carter et al., 2006; Bailey et al., 2007; Merschat et al. 2010) have provided important information regarding provenance, but have not significantly refined the depositional age of these units. Intercalated felsic volcanic rocks need to be identified and described through detailed field studies, and dated by U-Pb geochronology techniques.

Despite decades of attempted correlation of units from the eastern limb of the Blue Ridge anticlinorium in central Virginia to the eastern Blue Ridge in southern Virginia and western North Carolina (Fig. 1), fundamental differences in origin and evolution remain problematic. These correlation problems are a significant hindrance to our full understanding of the Blue Ridge in the central and southern Appalachians. Much work remains to be done in the decades to come.

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Field Trip Stop Descriptions

INTRODUCTION

This field trip guide contains 14 stops in the North Carolina and Virginia Blue Ridge that examines the geology of Mount Rogers, Virginia and surrounding area (see Fig. 2, Merschat et al., this guidebook). Some of the stop descriptions in this field guide are modified from previous field trip stops, and additional information may be found in the following guides (Rankin, 1971; Tollo et al., 2012; Carter and Merschat, 2014; Merschat et al., 2014). On Day 1 we will examine the geology of the western Blue Ridge, beginning in the oldest rocks, ~1.33–1.05 Ga basement gneisses, and continuing through Late Neoproterozoic (780–750 Ma) felsic volcanic, clastic, and glaciogenic rocks, and finishing with Late Neoproterozoic to Early Cambrian rift-to-drift sequence preserved in the Chilhowee Group. Day 2 will focus on the eastern Blue Ridge rocks and key contacts with basement rocks of the western Blue Ridge. Most of the stops are along public roads or on public land, but a few are on private property. Please honor property owners' requests when entering private property. Please note that Stops 1–4 through 1–6 are located within Grayson Highlands State Park and no collecting or rock hammers are allowed.

SAFETY

As always, please be very cautious and mindful of yourself and others at every stop. Board and leave the buses carefully and quickly, watch for traffic and warn others, be careful crossing highways, and be mindful of others when hammering on an outcrop. Do not go close to overhanging ledges, particularly where there is loose rock, and do not attempt to climb road cuts or other rock faces at any of the stops. Each stop presents different kinds of hazards, so please look around and assess your surroundings before examining the geology.

We wish everyone a safe, educational, and enjoyable 2016 Carolina Geological Society field trip.

DAY 1. OVERVIEW OF THE GEOLOGY OF THE MOUNT ROGERS AREA

Stop 1–1. ~1061 Ma Porphyroclastic granite gneiss (Yag) with blue quartz on VA 16 near Mouth of Wilson Baptist Church (Mouth of Wilson 7.5–minute quadrangle; 36.579872° N, 81.362313° W)

Exposed in the roadcut along VA 16 is a porphyroclastic granite gneiss (augen gneiss) with blue quartz. The gneiss contains 0.5–5 cm long blocky to subrounded alkali feldspar (microcline) megacrysts to porphyroclasts in a dark matrix of biotite, blue quartz (2–5 mm diameter), muscovite, epidote and chlorite. The foliation is well developed, characterized by greenschist facies assemblage: quartz + biotite + muscovite + chlorite + epidote +magnetite, and is similar to Alleghanian foliations observed in many of the Mesoproterozoic rocks around the Mount Rogers area. Feldspar porphyroclasts may contain tails or pressure shadows of quartz and muscovite that define a lineation (140/25) and often display top-to-the-NW shear sense. Geochemically, the gneiss varies from syenogranite to monzogranite (Tollo et al. 2010). Rankin et al. (1972) mapped this as augen gneiss and classified it as part of the Cranberry Gneiss. Geochronologic and petrologic investigations in conjunction with geologic mapping have shown that the porphyroclastic gneiss is one of the younger Mesoproterozoic units (Tollo et al., 2010, 2012). A SHRIMP U-Pb zircon age of 1061 ± 5 Ma was obtained from a sample of porphyroclastic biotite gneiss located to the southwest along Big Horse Branch. The Blowing Rock Gneiss, located in the Grandfather Mountain window further to the southwest, is texturally and compositionally similar. Carrigan et al., (2003) obtained a SHRIMP U-Pb zircon age of 1081 ± 14 Ma for the Blowing Rock Gneiss, which suggests it may be temporally related to the porphyroclastic granite gneiss here.

Stop 1–2. ~1140 Ma Lineated biotite granite (Ylbg) and foliated mafic dikes (now greenstone; Zdg) VA 93 south of bridge over New River (Mouth of Wilson 7.5–minute quadrangle; 36.584315° N, 81.313263° W)

The large roadcut on VA 93 just south of the bridge over the New River represents a microcosm of the northern end of the French Broad massif: various Mesoproterozoic lithologies intruded by mafic dikes and locally strongly overprinted with a mylonitic, greenschist facies, Paleozoic fabric. The roadcut contains migmatitic layered mafic gneiss (Ygg), coarse-grained to porphyritic biotite granite gneiss (Ylbg), porphyritic granite gneiss (Yag), and several mafic dikes, which are now chlorite-albite schists or greenstone. A strain gradient occurs in the roadcut. The least deformed rocks occur near the bridge in the northern part

Figure 1–1–1. Megacrysts/porphyroclasts of alkali feldspar in the ~1061 Ma porphyroclastic granite (augen gneiss) are blocky to subrounded, and rotated into the Alleghanian S_3 foliation defined by matrix biotite-muscovite-quartz-feldspar. Some of the alkali feldspar porphyroclasts indicate top-to-the-NW shear sense. Small felsic dike is transected by S_3.

of the roadcut, in front of a ~2-m-thick mafic dike transposed into the regional foliation (052/61) (Fig. 1–2). The rest of the roadcut is protomylonitic to mylonitic. The coarse-grained porphyritic to porphyroclastic granite gneiss contains 0.5–1.2 cm long alkali feldspar, plagioclase, quartz (often blueish gray), and biotite, overprinted with a metamorphic assemblage of muscovite, chlorite, and epidote. Biotite-rich schlieren 1 x 20 cm across occur in the granite but are only obvious in the low strained granite at the north end of the roadcut. The long-axis of the schlieren are parallel to S_1 (316/45), which is likely a Grenvillian foliation (Ottawan phase). The protomylonitic to mylonitic granite gneiss contains porphyroclasts of alkali feldspar and sometimes quartz, whereas muscovite + chlorite + epidote + clinozoisite define the S_3 Paleozoic foliation. In places where the protomylonitic granite gneiss is more massive, epidote veins are common. A sample of lineated biotite granite from a quarry located ~ 1 km to the east yielded a SHRIMP U-Pb age of 1140 ± 9 Ma and a second mylonitic sample from this roadcut yielded an age of 1134 ± 5 Ma (Tollo et al., 2012). The other Mesoproterozoic lithologies include migmatitic mafic gneiss (Ygg) and a porphyritic to porphyroclastic granite gneiss (augen gneiss; Yag). The overprinting mylonitic Paleozoic foliation makes these two lithologies difficult to distinguish from the mylonitic biotite granite. In the middle of the roadcut, ~5 m south of the road sign, the rock is characterized by centimeter to decimeter layers of quartz-feldspar leucosomes and amphibolite of the migmatitic mafic gneiss (Ygg). The alternating amphibolite and leucosome layers represent a Grenville age foliation S_1 transposed in the regional mylonitic Paleozoic foliation S_3. Near the southern end of the roadcut, the feldspar porphyroclasts are noticeably larger. Although this could be a coarser-grained phase of the biotite granite (Ylbg), it is interpreted to be the ~1061 Ma porphyroclastic granite gneiss seen at Stop 1–1.

At least seven bluish green to dark greenish gray mafic dikes (Zdg) in the roadcut are transposed into the regional Paleozoic foliation and locally strongly sheared (Fig. 1–2). The mafic dikes vary from 7 m to 0.2 m thick, although most are 2.5 to 1 m thick, and are characterized by greenschist facies minerals, chlorite + albite + magnetite + epidote. In areas where the Paleozoic overprint is not strong, the mafic dikes vary from diabase to gabbro, often may be ophitic to subophitic, and truncate S_1 foliation in the country rocks (See Stop 1–3). Zircon separated from a gabbro dike located 10 km east of Lansing, NC, yielded a SHRIMP U-Pb age of 757 ± 5 Ma (Tollo et al, 2012), within error of a SHRIMP U-Pb zircon age of 759 ± 9 Ma from a Bakersville Gabbro dike on Roan Mountain, Tennessee (J. Aleinikoff, USGS, written commun., 2016). Thus, the mafic dikes are part of the Bakersville dike swarm and are related to an early stage of rifting and underplating of mafic, mantle-derived material beneath the Blue Ridge (Rankin, 1975; Burton and Southworth, 2010; McClellan et al., 2014) and likely were feeders to the basalt in the lower Mount Rogers Formation.

The Paleozoic fabric in this roadcut is defined by a greenschist facies assemblage of muscovite + quartz + chlorite + epidote. Quartz ribbons in the mylonitic granitic gneiss display polygonal annealed textures, while other quartz grains are undulose and contain subgrains. Plagioclase porphyroclasts are sericitized and wrapped by the mylonitic foliation. An asymmetric foliation tail on a 1 m thick mafic dike indicates top-to-the-NW kinematics. Samples were collected from this outcrop for $^{40}Ar/^{39}Ar$ hornblende, muscovite, and K-feldspar geochronology. Muscovite from this roadcut yielded an $^{40}Ar/^{39}Ar$ plateau age of 339 ± 2 Ma, but hornblende and biotite yielded strongly disturbed $^{40}Ar/^{39}Ar$ spectra (Tollo et al., 2012). K-feldspar ages are disturbed as well, but yielded ages greater than the foliation-forming muscovite $^{40}Ar/^{39}Ar$ age suggesting that the latter is a growth age.

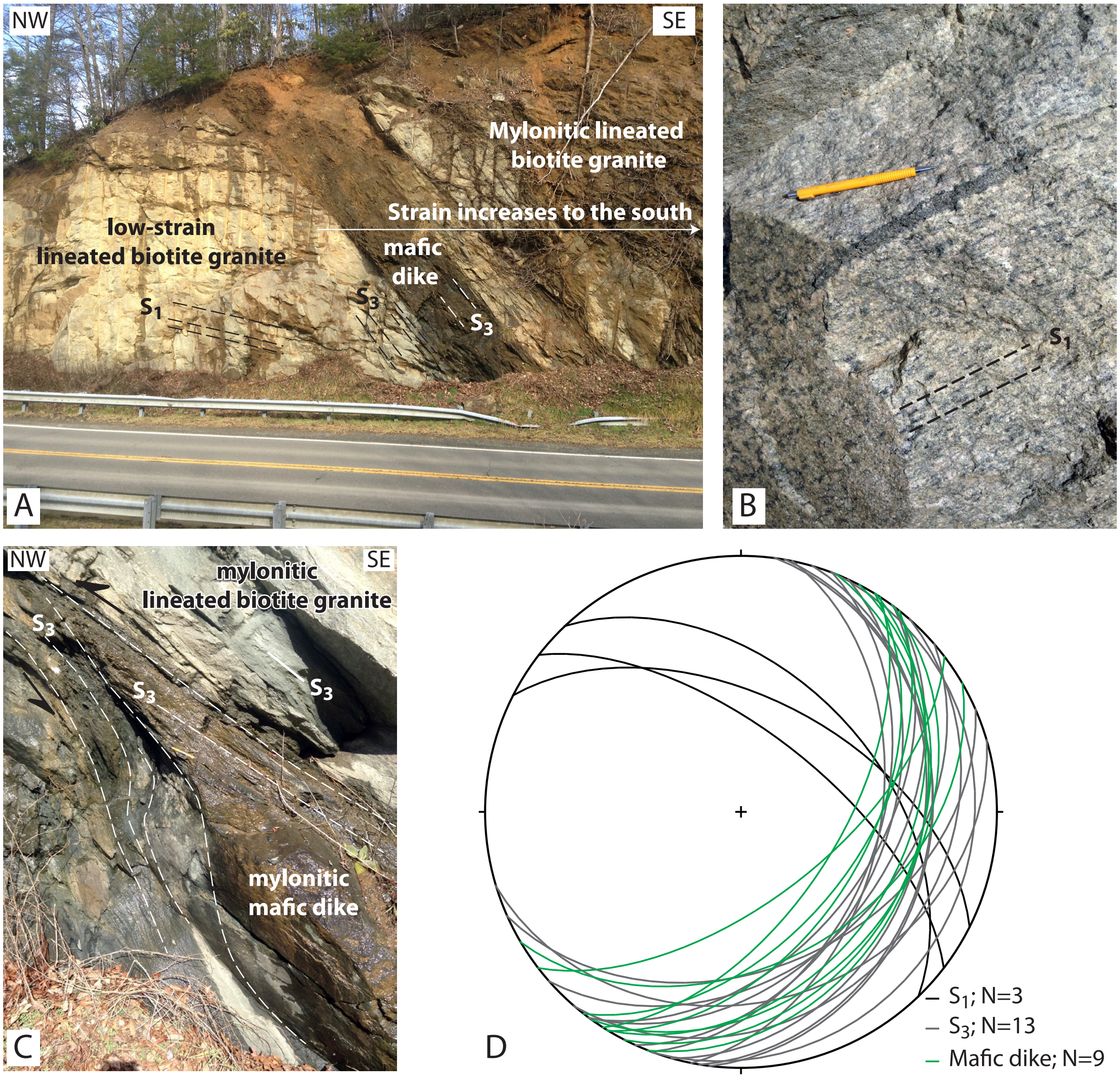


Figure 1–2–1. (A) Low-strained lineated biotite granite is located to the north of a ~2m dike mafic dike, now greenstone. The low-strained granite contains an earlier foliation, S_1, and biotite-rich schlieren. (B). Coarse-grained, lineated biotite granite with biotite-rich schlieren in NW-striking S_1 foliation (310/50). (C) Asymmetric S_3 foliation developed in transposed and sheared mafic dike suggests top-to-the-NW kinematic shear sense. (D) Equal area lower hemisphere stereonet showing the orientation of S_1, mafic dikes, and mylonitic foliation, S_3, in high-strained granite and mafic dikes (chlorite schists or greenstone). Fabric diagram created using Stereonet v. 9.3.2 (Allmendinger et al., 2012).

Stop 1–3. Migmatitic amphibolite gneiss (Ygg) cut by a mafic dike (Zdg) (Grassy Creek quadrangle; 36.624138° N, 81.444673° W)

This is also a snack break.

This roadcut is located on the inside of a blind curve on U.S. 58, so please be careful when crossing the road and while examining the roadcut. Furthermore, it is not necessary to cross the road to see the important structural relationships; everything can be observed from the gravel pull-off and bridge on Rugby Road, VA 743. Fresh exposures of the migmatitic amphibolite gneiss can be seen below the bridge in the creek (but be careful walking down the steep bank).

This roadcut, which was further excavated in the winter of 2016, consists of migmatitic amphibolite gneiss (Ygg) intruded by a mafic dike (Zdg) that is more than 7 m thick. The migmatitic amphibolite gneiss is one of the lithologies that comprise the older group of gneisses in the northern end the French Broad massif, and may be regarded as pre-Grenville crust into which voluminous amounts of 1150 Ma and 1050 Ma magmas intruded. These vestiges of pre-Grenville crust may be best considered as xenoliths or screens. This outcrop is near the northeastern end of a larger body of migmatitic layered gneiss referred to as orthogranofels (Tollo et al., 2010; 2012). The ~1.33 Ga orthogranofels (Ygg is the oldest rock in the Mesoproterozoic basement around Mount Rogers and possibly the southern Appalachians (Tollo et al., 2010). Sub-vertical gneissic layering defined by hornblende + plagioclase +

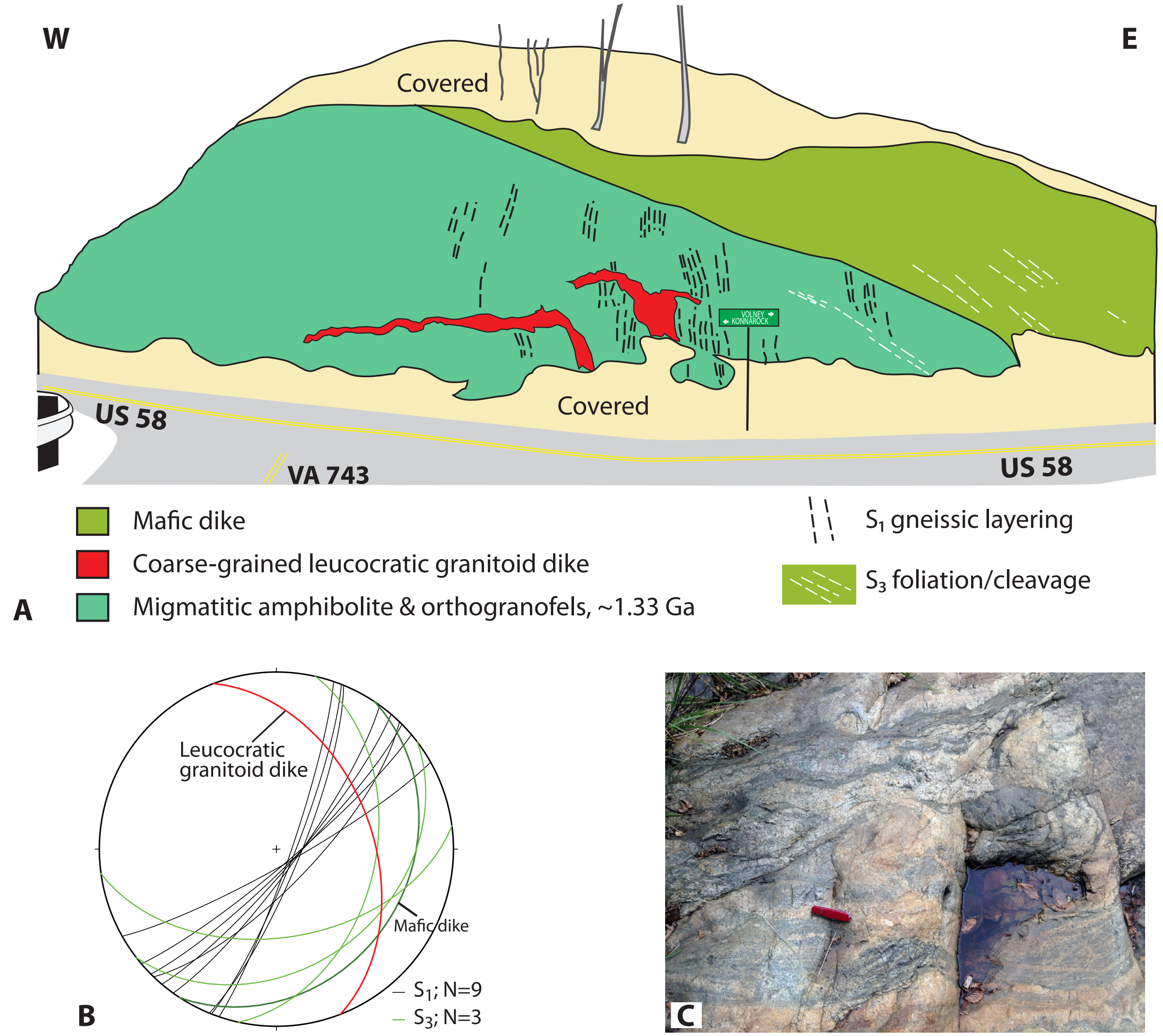


Figure 1–3–1. Sketch of roadcut of migmatitic amphibolite and orthogranofels with foliation, S$_1$, cut by mafic dike. Mafic dike is transposed into the regional D$_3$ Paleozoic foliation. (B) Equal area lower hemisphere stereonet showing the relationship between S$_1$, leucocratic granitoid dike, mafic dike, and S$_3$. The contact of the dike is roughly parallel with S$_3$. Stereonet v. 9.3.2 (Allmendinger et al., 2012) was used to create the plot. (C) Migmatitic amphibolite exposed in the Wilson Creek. Knife is 9 cm long.

quartz and migmatite leucosome layers define S$_1$ (049/82), and are truncated by a ~0.3 m leucocratic dike (339/45) and the mafic dike (033/22). Foliation within the mafic dike (S$_3$) is 032/33; the dike is transposed into the regional penetrative Paleozoic foliation. Although this dike contains a greenschist-facies Alleghanian foliation, the deformation is not as intense or pervasive as at Stop1–2. Similar to the dikes observed at Stop 1–2, these dikes are interpreted to be part of the ~757 Ma Bakersville dike swarm and likely represent feeder dikes to the basalt flows in the lower part of the Mount Rogers Formation.

Stops 1–4 through 1–6 are located within the Grayson Highlands State Park and collecting is prohibited without a permit. No hammers please.

Stop 1–4. Fees Rhyolite Member (Zmf), Grayson Highlands State Park (Grassy Creek 7.5–minute quadrangle: 36.619116° N, 81.477144° W)

The dark grayish purple, weathering light purplish gray, rhyolite in the roadcut in the curve is the Fees Rhyolite Member (753.1 ± 2.7 Ma; Zmf) of the Mount Rogers Formation. The Fees Rhyolite Member is a porphyritic rhyolite with 10–25 percent phenocrysts (up to 40 percent; Rankin, 1993) of pink perthite alkali feldspar, quartz, and plagioclase (Fig. 1–4–1). Pink perthite

Figure 1–4–1. The Fees Rhyolite Member is a dark grayish purple, porphyritic lithic crystal tuff, which contains inclusions of porphyritic rhyolite recognized by close clusters of phenocrysts (white dashed ellipse), and inclusions of granite pebbles (white arrows). Some rhyolite inclusions in this roadcut may be dominantly aphanitic. Pencil is 14.5 cm long.

Figure 1–5–1. Contorted flow bands in the Whitetop Rhyolite Member. Pencil is 14.5 cm long.

alkali feldspar is the most abundant phenocryst and ranges from 1-4mm across. Quartz is the second most abundant phase and is 1–3 mm across and frequently is embayed. Rankin (1993) noted that plagioclase phenocrysts constitute a small, variable amount of phenocrysts, but its presence is a distinguishing characteristic of the Fees Rhyolite. The phenocrysts often occur in circular dense clusters 5–20 cm long and flattened into the foliation. Near the southern end of the roadcut, the rhyolite contains rare lithic, pebble to cobble-sized inclusions of quartz, muscovite granite, and rhyolite (Fig. 1–4–1). A foliation, S_3, is moderately to weakly developed throughout the roadcut.

The Fees Rhyolite Member was interpreted to occur near the base of the Mount Rogers Formation as it is separated from the main section of rhyolites of the Mount Rogers volcanic center by arkose, conglomerate, and metabasalt, which are often associated with the lower part of the formation (Rankin, 1993). The Fees is often in close proximity with basement and locally rests nonconformably on basement (see Stop 9, Merschat et al., 2014). Yet ambiguity still exists regarding its field relationships, age and stratigraphic position (see McClellan et al., this guidebook). Structurally, the Fees overlies the rhyolites of the Mount Rogers volcanic center, either along a NW-verging ductile thrust, the overturned limb of a speculative NW-verging fold (although the core of the fold has not been found), or it is a younger rhyolite within a younger clastic sequence. Tollo et al. (2012) reported CA-TIMS U-Pb single zircon ages from the Fees rhyolite that range from ~769 to ~753 Ma (759.8 ± 2.6 weighted mean) and the youngest zircon, 753.1 ± 2.7 Ma, was interpreted as the age. If this age is correct the Fees is about the same age as the overlying Buzzard Rock and Whitetop Rhyolite members.

Stop 1–5. Flow bands in the Whitetop Rhyolite Member (Zmwt), Grayson Highlands State Park. (Whitetop Mountain 7.5–minute quadrangle; 36.624713° N, 81.506812° W)

The Whitetop Rhyolite Member (Zmwt) (753.3 ± 2.0 Ma; Tollo et al., 2012) is distinct among all of the rhyolites of the Mount Rogers Formation in that it generally lacks phenocrysts and is commonly flow banded. The light purplish gray flow banded rhyolites exposed along the road and in roadcuts are typical of the Whitetop Rhyolite. Especially well developed here are folds in the flow bands (contorted) and some of the tighter folds show evidence of auto-brecciation (Fig. 1–5–1). Common flow bands indicate that the Whitetop Rhyolite Member was a lava that flowed on the surface as compared to most of the other rhyolites, excluding parts of the Buzzard Rock Rhyolite (see Stop 1–6), that were erupted as ignimbrites or pyroclastic flows (Rankin, 1993). The sub-vertical, contorted flow bands observed here indicate it is near the core or center of a lava flow. Other lithologies of the Whitetop Rhyolite Member that are not seen here include lithic lapilli tuff, volcanic flow breccia, and multishelled lithophysae rhyolite (see Rankin, 1993).

Stop 1–6. Listening Rock Overlook and Virgil J. Cox Visitors Center, Grayson Highlands State Park (Park 7.5-minute quadrangle; 36.623129° N, 81.501452° W). Lunch and Restroom break.

This stop (~1.5 hours long) includes both a lunch and restroom break. There are picnic tables in the trees around the overlook and also scattered about the visitor center. Bathrooms are available at the Virgil J. Cox Visitors Center. The outcrops we will be examining are by the overlook, but there are lots of other exposures of the Buzzard Rock Rhyolite Member around the visitor center. There are also several hiking trails with spectacular geology, but we do not have time for a hike; please do not attempt them.

On the left side of the paved path to the overlook is an outcrop of the Buzzard Rock Rhyolite Member (Zmb)(755.0 ± 6.6 Ma), the lowest and oldest rhyolite of the Mount Rogers volcanic center. The Buzzard Rock Rhyolite is a dark bluish-purple porphyritic rhyolite with 10–15 percent phenocrysts of perthitic alkali feldspar (usually pink) and plagioclase (white to greenish-white), usually in equal proportions (Rankin, 1993). Phenocrysts are generally 1–4 mm across. Glomeroporphyritic plagioclase clusters, a texture too delicate to survive explosive airborne deposition, occur in the Buzzard Rock Rhyolite (Tollo et al., 2012). In the outcrops around the overlook, road, and visitor center, dark gray to dark purple centimeter-scale flow bands are well developed (Fig. 1–6–1), which suggests that parts of the Buzzard Rock Rhyolite were lava flows, if not entirely (Rankin, 1993; Tollo et al., 20102). Chemically the Buzzard Rock Rhyolite is distinct; it is lower in silica (69.1–70.3 %) and the least chemically evolved of all the other rhyolites from the Mount Rogers Formation (Tollo et al., 2012).

Polylithic boulder conglomerate (Zmac) crops out at the overlook, and makes a series of ledges and boulders along the south side of the road and slope (Fig. 1–6–1). (This is also a popular location for bouldering, a type of rock climbing, and hence the white, chalky residue on many surfaces of the outcrops.) The boulder conglomerates are generally massive, and contain rounded and

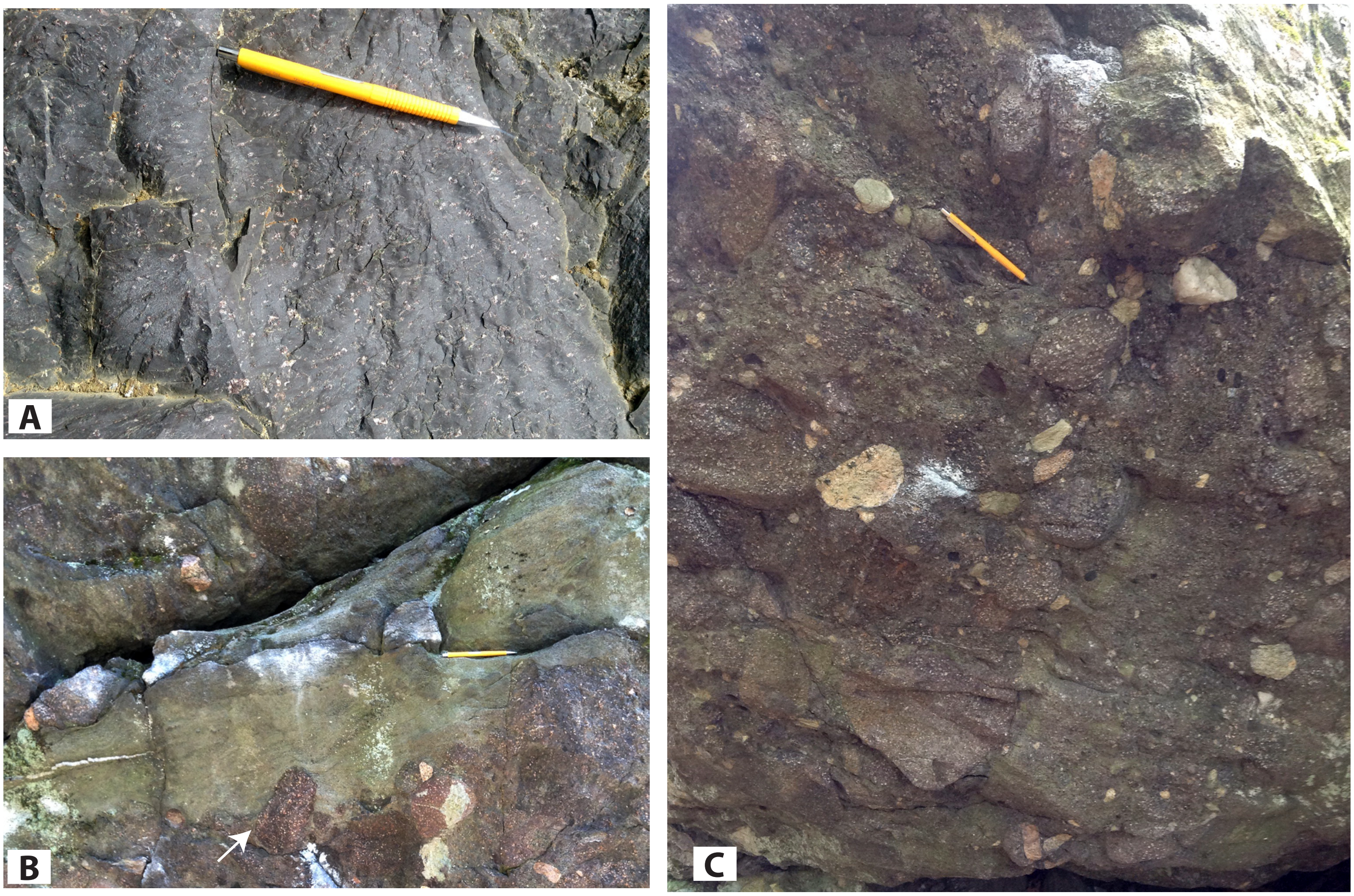


Figure 1–6–1. (A) Porphyritic Buzzard Rock Rhyolite Member with typical dark bluish gray color, 10-20 percent phenocrysts , and faint flow bands. (B) Boulder conglomerate preserving bedforms, with rhyolite clasts embedded into laminated and graded sands (black arrow). Embedded rhyolite clast suggests that the bedform is overturned. (C) Polylithic boulder conglomerate at Listening Rock Overlook. Most of the round clasts are porphyritic rhyolite; coarse-grained biotite granitoid, greenstone, arkose, quartz and feldspar comprise minor populations (see also McClellan et al., this guidebook). Dusty white areas in (b) and (c) are chalk. Pencil is 14.5 cm.

elongate clasts of porphyritic rhyolite, granite (pink, medium-grained, and coarse-grained biotite granite), sandstone, greenstone, mudstone, quartz, and alkali feldspar (see McClellan et al., this guidebook). Bedforms, although not common, consist of decimeter-scale laminated sand to silt layers. Bedforms around the overlook and seen along part of Listening Rock Trail display graded beds, and draped or embayed clasts that indicate the boulder conglomerate beds are upright and dipping to the east and south. The contact relationship between the conglomerate and rhyolite is either: (1) the rhyolite overlies the conglomerate and the porphyritic rhyolite is sourced elsewhere; or (2) the conglomerate overlies the rhyolite and was sourced from the rhyolite. The large amount of rhyolite clasts, and upright bedforms suggests the latter maybe possible, however a source for the granitoid clasts, presumably Mesoproterozoic basement, is needed. Detrital zircon analysis and U-Pb clast analysis (see McClellan et al., this guidebook) indicates the source was a larger volcanic field with 780–740 Ma volcanics, ~780 Ma granitoids, and Mesoproterozoic basement exposed.

Finally, the view from the overlook is to the south. North Carolina! The high mountains include Phoenix Mountain (several towers on the southwest end), Mount Jefferson (pyramidal mountain with a tower behind Phoenix), Three Top, and Elk Knob (both to the southwest from Phoenix Mountain); these mountains are all underlain by amphibolite of the Ashe Formation. On a clear day, Beech Mountain, NC and Grandfather Mountain, NC can be seen.

Stop 1–7. Catface fault and mylonitic Pond Mountain porphyry (Zmp), Mud Creek Road (Park 7.5-minute quadrangle; 36.608051° N, 81.578419° W)

One of the more interesting structures on the geologic map of Rankin (1993) is a narrow, hooked reentrant of the Mountain City window that extends to the southeast through the community of Whitetop, VA exposing the Konnarock Formation and bounded by shallow dipping segments of the Catface and Stone Mountain faults (see Fig. 2, Merschat et al., this guidebook). Mud Creek Road transects the middle of the structure and the roadcut at Stop 1–7 contains mylonitic Pond Mountain rhyolite porphyry (Zmp). A short distance further south on Mud Creek Road are ditch exposures of cleaved maroon laminites assigned to the Konnarock Formation.

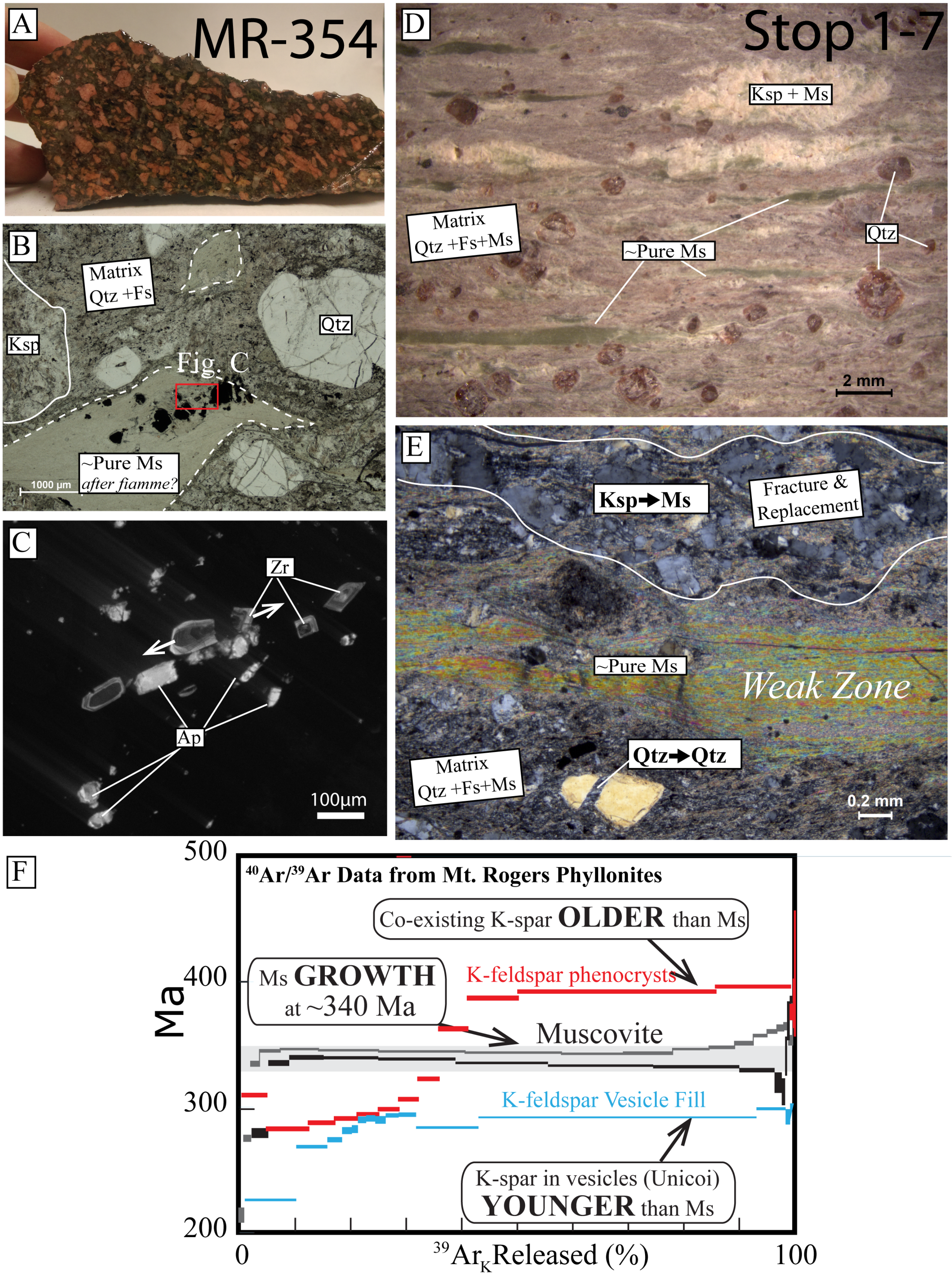
The rocks here are porphyroclastic phyllonites with quartz and alkali feldspar porphyroclasts in a matrix of muscovite and quartz. The rock may not look mylonitic, but these rocks can be mapped into coarse porphyritic rhyolites of the Pond Mountain porphyry containing alkali feldspar phenocrysts 1–4 cm across (Fig 1–7–1A). On many of the foliation surfaces are rounded bumps of quartz porphyroclasts/phenocrysts, elliptical patches of light green, very fine-grained muscovite (scales), and a strong lineation defined by streaked muscovite, quartz ribbons, and fractured and stretched feldspars. The mylonitic foliation, S_3, dips gently to the southeast, and a lineation, L_3, plunges gently to the southeast (15–20°), part of the regional lineation pattern related to the NW transport of the Blue Ridge thrust sheet.

In thin section K-feldspar and plagioclase porphyroclasts are fractured, pulled apart, and partially altered to fine-grained muscovite. The elongate light green scales on the foliation surfaces are also comprised of muscovite, however the muscovite in these scales is slightly coarser grained (Fig. 1–7–1) and these domains contain no relict feldspar. These textural differences suggest that the scales have a different origin than muscovite in the rest of the rock. We suggest the light green scales originated as glassy domains, and possibly as pyroclasts (fiamme or pumice lapilli). Examination of less deformed porphyritic rhyolites shows the presence of similar (though less elongate) homogenous domains of green muscovite despite no lower greenschist facies foliation and little to no replacement of K-feldspar by muscovite (Fig. 1–7–1). Zircon and apatite are anomalously abundant in these domains, and zircon is commonly fractured and brecciated (Fig. 1–7–1). However, muscovite in the fractures shows no preferred orientation. One possible explanation for the observed mineralogy and textures is that these domains were originally late stage liquids concentrated in incompatible and accessory phases. Flow of this highly viscous and cooling silicate liquid resulted in fracturing of the included accessory phases. Later, the metastable glass was replaced by randomly oriented muscovite during fluid infiltration and/or metamorphism, likely during the Alleghanian.

Another interesting textural aspect of these phyllonites is the conspicuous preservation of quartz phenocrysts. Most quartz phenocrysts are slightly rounded and some are fractured and offset in an orientation consistent with the regional sense of shear. However, in general the phenocrysts are well preserved and some phenocrysts even remain angular and preserve igneous zoning in cathodoluminescence (Fig. 1–7–1). All of these observations are consistent with the limited recrystallization and rigid behavior of quartz phenocrysts despite the location of these rocks in a major shear zone that was active at temperatures of ~300° C. Quartz is expected to deform ductilely at these temperatures.

Our observations from high and low strain rocks suggest that an initial aphanitic matrix of feldspar, quartz and volcanic glass was replaced by a weaker assemblage of muscovite + quartz (reaction softening), likely aided by microcracks (McAleer et al., 2015). Strain localization into zones dominated by this weaker assemblage led to alignment of muscovite into contiguous bands and further weakened the rock (textural softening). The rock—now characterized by these bands of muscovite—could accommodate the strain at a lower yield stress than that necessary to deform quartz into ribbons. The end result preserved nearly euhedral quartz phenocrysts within the Catface fault at the base of the Blue Ridge thrust sheet.

Because the muscovite apparently grew at low temperature and defines the foliation, we attempted to date the fabric-forming muscovite to determine the time of deformation on the Catface fault. $^{40}Ar/^{39}Ar$ ages of muscovite from this phyllonite as well as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⁴⁰Ar/³⁹Ar Data from Mt. Rogers Phyllonites
Co-existing K-spar OLDER than Ms
Ms GROWTH
at ~340 Ma
K-feldspar phenocrysts
Muscovite
K-feldspar Vesicle Fill
K-spar in vesicles (Unicoi)
YOUNGER than Ms
Ma
500
400
300
200
0
100
³⁹Ar_K Released (%)

several other nearby localities are all ~340 Ma and we suggest this is the time of emplacement of the Blue Ridge thrust sheet above what is now the Mountain City window (Fig. 1–7–1F). This is older than the age of faulting as determined in the fold and thrust belt. We suggest that shortening after ~340 Ma was accommodated along the basal decollement and linked with the westward propagation of faults in the Valley and Ridge that eventually carried the Blue Ridge thrust sheet to its present location by ~260 Ma. Post ~340 Ma deformation and fluid flow is indicated by (1) open folding of mylonitic fabrics along the Blue Ridge thrust; (2) folding of the Shady Valley thrust sheet and duplexing in the Mountain City window; (3)^{40}Ar/^{39}Ar ages of ~280 Ma from fault gouge to the west (Hnat et al., 2014); (4) ~280 Ma zircon fission track (Naeser et al., 2005); and (5) ~290 Ma K-feldspar filling vesicles in basalts of the Unicoi (Stop 1–10).

Stop 1–8. Boulder diamictite of the Konnarock Formation (Zkd), intersection of VA 859 and US 58 (Konnarock 7.5–minute quadrangle; 36.65914° N, 81.66052° W).

Maroon polylithic boulder diamictite (Zkd) is one of the characteristic glaciogenic rock types of the Konnarock Formation. Diamictite has been mapped both northwest and southeast of the valley in which the community of Konnarock is located, and occurs in both the Shady Valley-Blue Ridge thrust sheet and within the Mountain City window (Rankin, 1967; 1993) (see Fig. 2, Merschat et al., this guidebook). This exposure of diamictite is within the Shady Valley thrust sheet and is stratigraphically above the rhythmite at Stop 1–9. It is part of a nearly continuous belt of diamictite near the top of the formation that is 400–500 m thick (Rankin, 1993). The Unicoi Formation disconformably overlies the Konnarock Formation and part of the diamictite may have been removed by erosion.

The diamictite consists of randomly oriented, subrounded boulders of largely pink, coarse-grained, non-foliated biotite granite in a poorly sorted maroon matrix of clay and some sand. Rhyolite, greenstone and sedimentary clasts are not common, but do occur. Bedforms are not evident and cleavage is not well developed, although slickensides occur on the surfaces of some boulders. Local beds of arkose or medium- to fine-grained sandstone beds occur within the diamictite along strike (Rankin, 1993). The diamictite represents either a lodgment till (Rankin, 1993), or waterlain till deposited proximal to an ice cliff or ramp by subaqueous melting of the glacier as it advanced across "lake Konnarock" (Miller, 2004). The lack of striated clasts and common subaqueous facies of the Konnarock Formation (see Stop 1–9) suggests a waterlain till near an ice cliff or ramp is more likely (Miller, 2004). Non-foliated granitoid clasts are dominant (~80 percent basement granitoid and the rest of the clasts are volcanic and sedimentary clasts; Rankin, 1993) throughout the diamictite and suggest that it was not an alpine glacier on a volcanic edifice.

Stop 1–9. Maroon rhythmite with dropstones, Konnarock Formation (Zk), intersection of VA 600 and VA 603 (Konnarock 7.5–minute quadrangle; 36.674506° N, 81.608850° W). Park at gas station and walk along road to roadcut on VA 603.

The two better known lithologies of the Konnarock Formation are the diamictite (Stop 1–8) and maroon laminated rhythmite with dropstones (Zk). This roadcut displays the variation of rock types within the Konnarock Formation. Exposed here are pinkish medium-grained sandstone, maroon mudstone, laminated mudstone and rhythmite with dropstones; all of these lithologies reflect deposition in a glaciolacustrine environment. "Lake Konnarock" had a surface area of perhaps 1200 km ^{2} although these are only minimum estimates based on the current deformed distribution of the Konnarock Formation.

At the southwest end of the cut is a 1-2m thick, well sorted feldspathic quartz arenite (sandstone) bed that overlies maroon mudstone. The sandstone bed is internally featureless except for rip-up clasts of maroon mudstone and exhibits load casts along its base. The sandstone cuts downward through beds of maroon mudstone and probably represents subaqueous outwash or delta deposit (Miller, 2004)(Fig. 1–9–1A). The mudstone is laminated and bedding dips toward the west and northwest. The mudstone contains pellets of friable feldspar and a cobble-sized granitoid clast (dropstone) cut by a down-to-the-NW normal fault (Fig. 1–9–1B). Moving to the northeast along the roadcut, maroon rhythmite occurs beneath the mudstone and in the northeast part of the roadcut. The rhythmite consists of cm- to mm-scale varves; couplets with green-gray siltstone to fine-grained sandstone (base)

◄———

Figure 1–7–1. (A-C) Largely undeformed porphyritic rhyolite of the Mount Rogers Formation. (A) Slabbed hand sample. (B) The aphanitic matrix is dominantly K-feldspar and quartz. Quartz and feldspar phenocrysts are preserved and pure green muscovite "blebs" are present with inclusions of Fe-Ti oxides, zircon and apatite. (C) Apatite and zircon are abundant in the muscovite domains and are commonly fractured. (D) Mylonitic rhyolite from Stop 1–7. Pure green muscovite zones are present but are elongate. Muscovite is also common in the matrix and replacing K-feldspar. Red mineral is jasperoid (not garnet). (E) Photomicrograph shows contiguous muscovite in the green domains. A fractured and pulled apart K-feldspar is replaced by muscovite and some of this muscovite is parallel to foliation. Fine-grained metamorphic quartz fills fractures in quartz phenocrysts. (F) ^{40}Ar/^{39}Ar data from phyllonites developed in the Mount Rogers Formation. Muscovite from 5 localities (2 shown) produced a flat age spectra at ~340 Ma. K-feldspar phenocrysts from the same samples (1 shown) yield age spectra that climb to ages older than 340 Ma, supporting our conclusion that muscovite grew below its closure temperature and dates the time of phyllonitization. K-feldspar from vesicles in the basalts in the Unicoi (Stop 1–10) indicate K-feldspar precipitated there at ~290 Ma, likely related to fluid flow associated with transport of the composite thrust sheet to the west. Abbreviations: Fs—feldspar, Ksp—K-feldspar, Ms—muscovite, and Qtz—quartz.

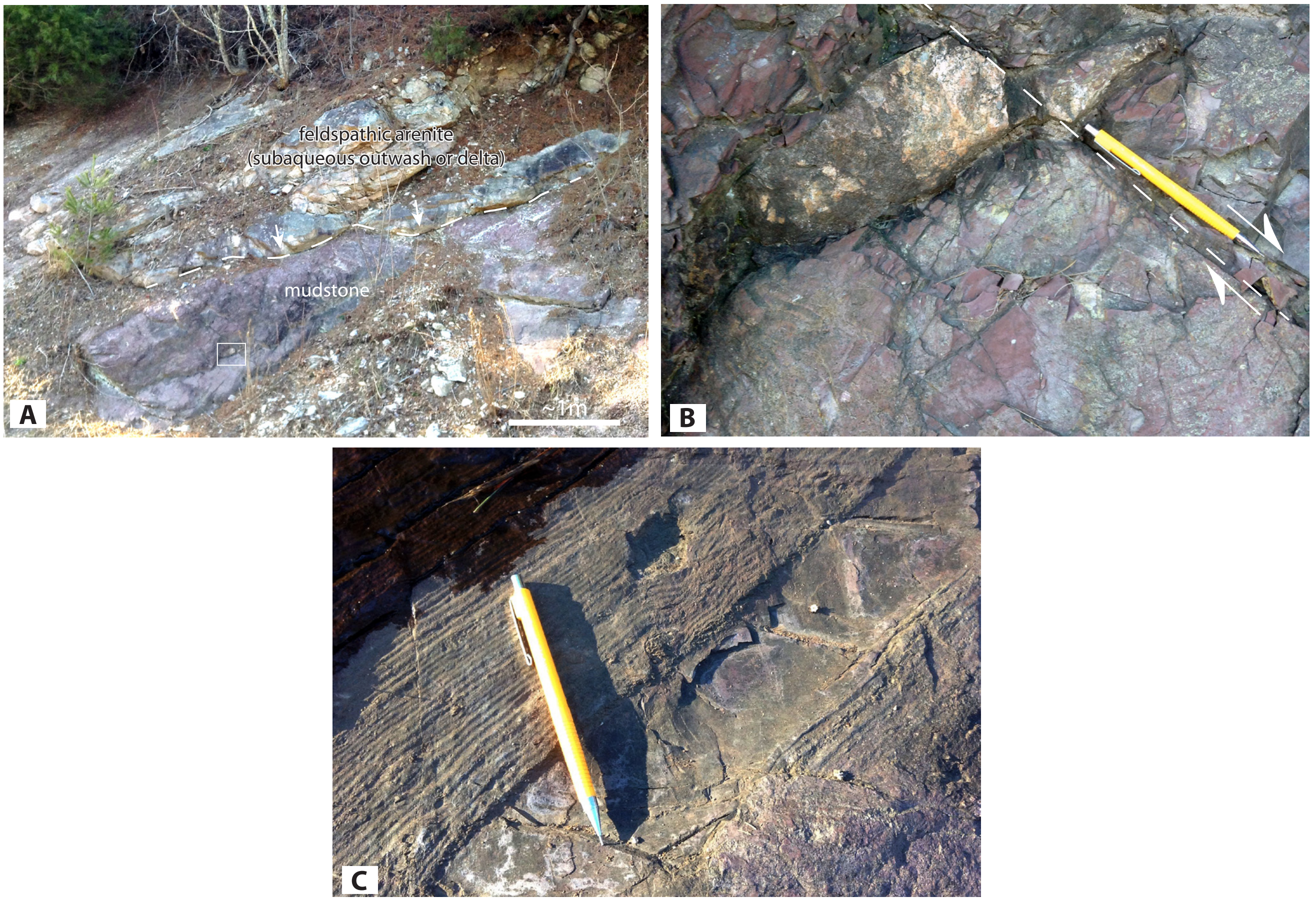

Figure 1–9–1. (A) An approximately 1 m thick sandstone bed cuts down into bedded maroon mudstone. The well-sorted feldspathic sandstone exhibits load casts (see white arrows and dashed lines) and some rip-up clasts of maroon mudstone. Load casts at the base of the sandstone, rip-up clasts, and down-cutting nature of the sandstone suggest rapid, subaqueous deposition in a glacial outwash or delta deposit. Small white box shows location of (B). (B) Coarse-grained granitoid dropstone (~20 cm long) cut by a small late Paleozoic normal fault, top-down-to-NW. (C) Granitoid dropstones embedded and truncating millimeter-scale rhythmically layered beds, varves (181/46). Pencil is 14.5 cm and points downward in (B) and (C).

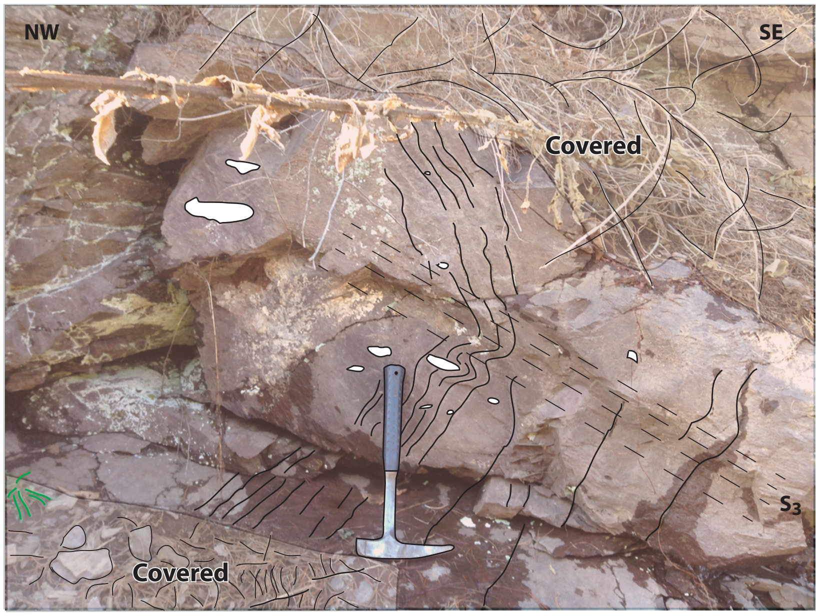

Figure 1–9–2. Folded maroon rhythmite with granitoid dropstones that is overturned to the NW. Note long-axis of the clasts embed downward and are nearly axial planar to the fold. A weak cleavage can be noted (23/75). Black lines trace light colored, coarser-grained beds, 5–10 mm thick, which represent seasonal melting of ice.

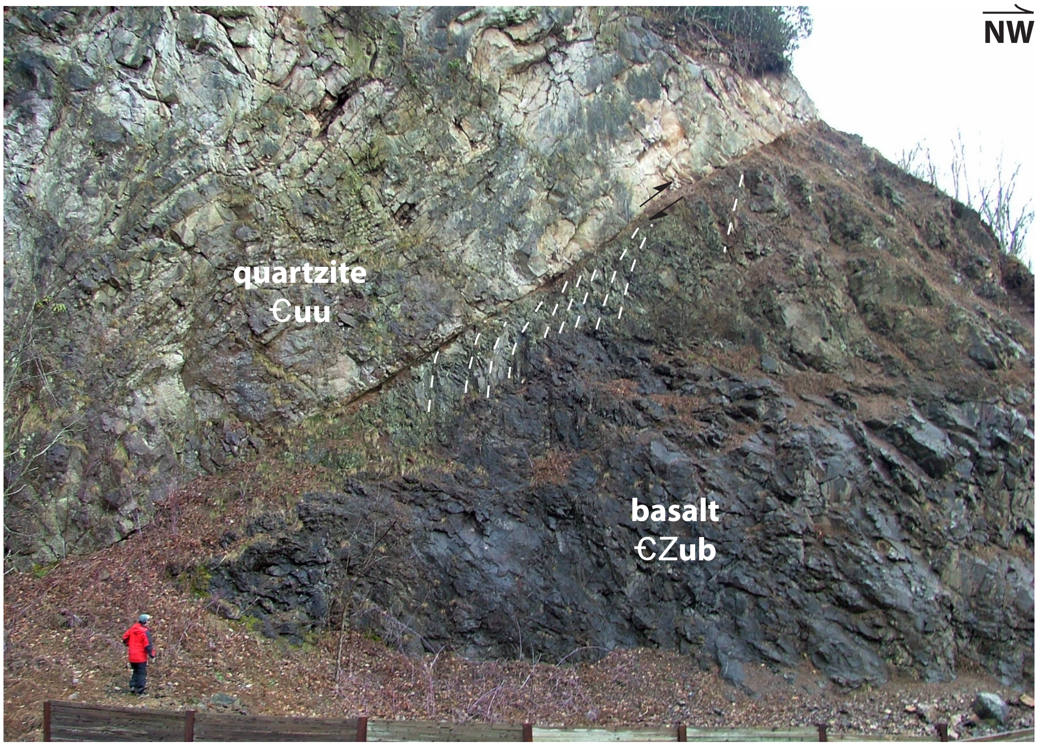

Figure 1–10–1. Large amygdaloidal basalt flow (ꞒZub) at the contact between the marine quartzites of the upper Unicoi Formation and fluvial rocks of the lower Unicoi Formation beneath the basalt (out of view). The lower Unicoi Formation is comprised of conglomerate, coarse-grained feldspathic wackes, drab and purple shale, and several basalt flows. Note the planar nature, recessed weathering, fine-grained material, and deflection of cleavage (white dashed lines) along the upper contact. The contact was re-activated as a NW-directed thrust fault along the anisotropic and mechanically different units. Person is approximate 1.8 m tall.

grade upward into maroon mudstone (top) (Fig. 1–9–1C) (Rankin, 1993). Varves form as the result of freeze/melt cycles within a glacial lake (Rankin, 1993; Miller, 2004). Dropstones occur in the northeast end of the roadcut and are primarily pebble- to cobble-sized granitoid clasts. Bedding in the northeast end of the outcrop, approximately 2 m above road level, is folded (Fig. 1–9–2). The folding is overturned to the NW and related to the emplacement of the Blue Ridge thrust sheet during the Alleghanian orogeny.

Stop 1–10. Amygdaloidal metabasalt flow in the lower Unicoi Formation, VA 600 (Whitetop Mountain 7.5–minute quadrangle; 36.70806°N, 81.61692°W).

This roadcut exposes a cross section through the uppermost basalt flow (ꞒZub) in the lower Unicoi Formation (ꞒZul) . It is located above the Holston Mountain fault in the northwest limb of the Stony Creek syncline in the Blue Ridge thrust sheet (Fig. 1). The flow is ~12 m thick and consists of bluish black to dark gray, amygdaloidal, vesicular metabasalt (Fig. 1–10–1). Amygdules are filled with white calcite and rimmed by pink K-feldspar. ^{40}Ar/^{39}Ar spectra indicate the K-feldspar precipitated at ~290 Ma (Fig. 1–7–1F), likely related to fluid flow corresponding with transport of the composite Blue Ridge thrust sheet to the west . The base of the basalt is above a 10–20 cm zone of drab to dark gray, shale to mudstone, ~1 m of dark gray laminated feldspathic wacke, and a 0.5–1 m thick bed of quartz pebble conglomerate and feldspathic wacke. The depositional environment of the lower Unicoi Formation is described as fluvial (Simpson and Eriksson, 1989; Walker et al., 1994; Smoot and Southworth, 2014). The upper contact of the basalt is a minor intraformational bedding-slip fault (058/32). Deflected and truncated cleavage in the metabasalt is consistent with top-to-the-NW reverse motion (Fig. 1–10–1). The fault (younger-over-older) likely developed along contrasting rheologies during contraction associated with the formation of the Stony Creek syncline and emplacement of the Blue Ridge thrust sheet during the Alleghanian orogeny. The basalt is overlain by light purplish gray to white, thick bedded, medium- to coarse-grained quartzites and feldspathic arenites of the upper Unicoi Formation (072/ 46). Sandstones are cross-bedded, wavy laminated, grade upwards and represent the transition from fluvial environments to nearshore marine environments upsection (Simpson and Eriksson, 1989; Walker et al., 1994; Smoot and Southworth, 2014).

Several basalt flows occur in the lower Unicoi Formation from central Tennessee to southwest Virginia (King et al., 1944; King and Ferguson, 1960; Rankin, 1993; Smoot and Southworth, 2014). This is one of the more extensive flows and has been traced for 20 km and also crops out on the southeast limb of the Stony Creek syncline (King and Ferguson, 1960; Rankin, unpub. data). These basalt flows are likely related to the 575–555 Ma continental flood basalts and bimodal volcanism of the Catoctin Formation (central Virginia to south-central Pennsylvania) (Aleinikoff et al., 1995; Burton and Southworth, 2010; Smoot and Southworth, 2014).

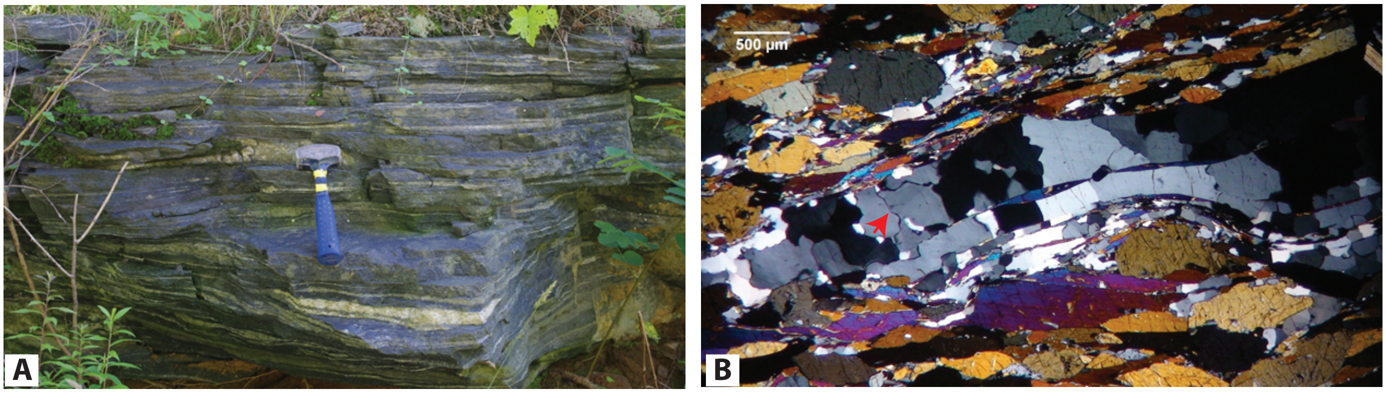

Figure 2–1–1. (A) Outcrop of amphibolite of the Ashe Metamorphic Suite. There is a well-developed foliation in this outcrop associated with the C-foliation. This outcrop contains abundant quartz veins and granitic dikes. (B) Photomicrograph of amphibolite containing quartz veins. Quartz grains are coarse and some grain boundaries are amoeboid (red arrow) indicating that this rock is on the boundary between subgrain rotation recrystallization and fast grain boundary migration. In the middle right of the image quartz ribbons appear to be wrapping around a more rigid amphibole grain.

Therefore, this age has been applied to the Unicoi Formation, and the age of rifting of Rodinia and opening of the Iapetus ocean (Aleinikoff et al., 1995; Hatcher et al., 2007). A recent study of ichnotaxa from the Chilhowee Group suggests that the middle and upper parts of the group are Middle Cambrian (Stages 2 and 3; Hageman and Miller, 2016). Most importantly, a newly discovered nevadiid trilobite specimen from the Erwin Formation (Murray Shale) on Chilhowee Mountain in southeastern Tennessee places faunal control on the age of the Erwin Formation (521–517 Ma) (Hageman and Miller et al., 2016). Thus the Chilhowee Group either spans perhaps 40 m.y. or rifting occurred later than interpreted (Walker and Driese, 1991; Smoot and Southworth, 2014; Hageman and Miller, 2016).

DAY 2. ASHE–BASEMENT CONTACT

Stop 2–1. Amphibolite of the Ashe Metamorphic Suite near Warrensville (Warrensville 7.5-minute quadrangle; 36.458525° N, 81.515036° W). (Jamie Levine)

At this outcrop we will see excellent examples of the amphibolite of the Ashe Metamorphic Suite (AMS) of Abbott and Raymond (1984). This location, just outside Warrensville, NC is very close to the Gossan-Lead fault, an Alleghanian-aged fault with thrust kinematics. The Gossan Lead fault, which is mapped to the northwest of the town of Warrensville, strikes northeast-southwest in this area and separates rocks of the AMS from basement gneisses and schists. Due to the proximity of the Gossan Lead fault, and further to the northwest the presence of the Fries fault, another Alleghanian-aged thrust fault, this outcrop displays a well-developed S-C fabric. Along both sides of the Gossan Lead fault the C-fabric is the dominant foliation, and it consistently strikes north-northeast and dips to the east-southeast. Amphibole needles form a strong down-dip mineral lineation that typically plunges moderately to shallowly to the east-southeast. Amphibolites of the AMS commonly contain foliation sub-parallel granitic dikes and quartz veins (Fig. 2–1–1A).

In thin section these amphibolites contain abundant amphibole, and quartz with well-developed quartz ribbons parallel to the dominant C-foliation, in addition to less abundant plagioclase garnet (Fig. 2–1–1B). Several samples display S-C and C' fabrics that are consistent with top-to-the-NW kinematic shear sense. There is some chlorite growth, which can be observed along C' fabrics. The quartz in these samples has undergone dominantly rotational recrystallization with some samples showing evidence for fast grain boundary migration, followed by some amount of static recrystallization as evidenced by polygonal grains. Rotational recrystallization in quartz is typical of deformation occurring at greenschist-facies to lower amphibolite-facies conditions, consistent with the presence of chlorite in some samples. Samples of metagraywacke adjacent to amphibolites commonly display more evidence for fast grain boundary migration in quartz, suggesting that temperatures may have been more definitively within the amphibolite-facies. Amphibolite samples may not provide evidence for these higher-temperatures because the amphibole, instead of quartz, may be the more highly strained phase.

Discussion of Ashe Metamorphic Suite Amphibolite Geochemistry (Crystal G. Wilson)

In addition to map-scale amphibolite bodies throughout the eastern Blue Ride, amphibolite can also be found as minor occurrences within pelitic units of the Ashe Metamorphic Suite (AMS). Recent detailed geologic mapping at Elk Knob State Park, ~15.3 km north of Boone, NC, prompted geochemical investigation of the origin and possible relationship of these smaller,

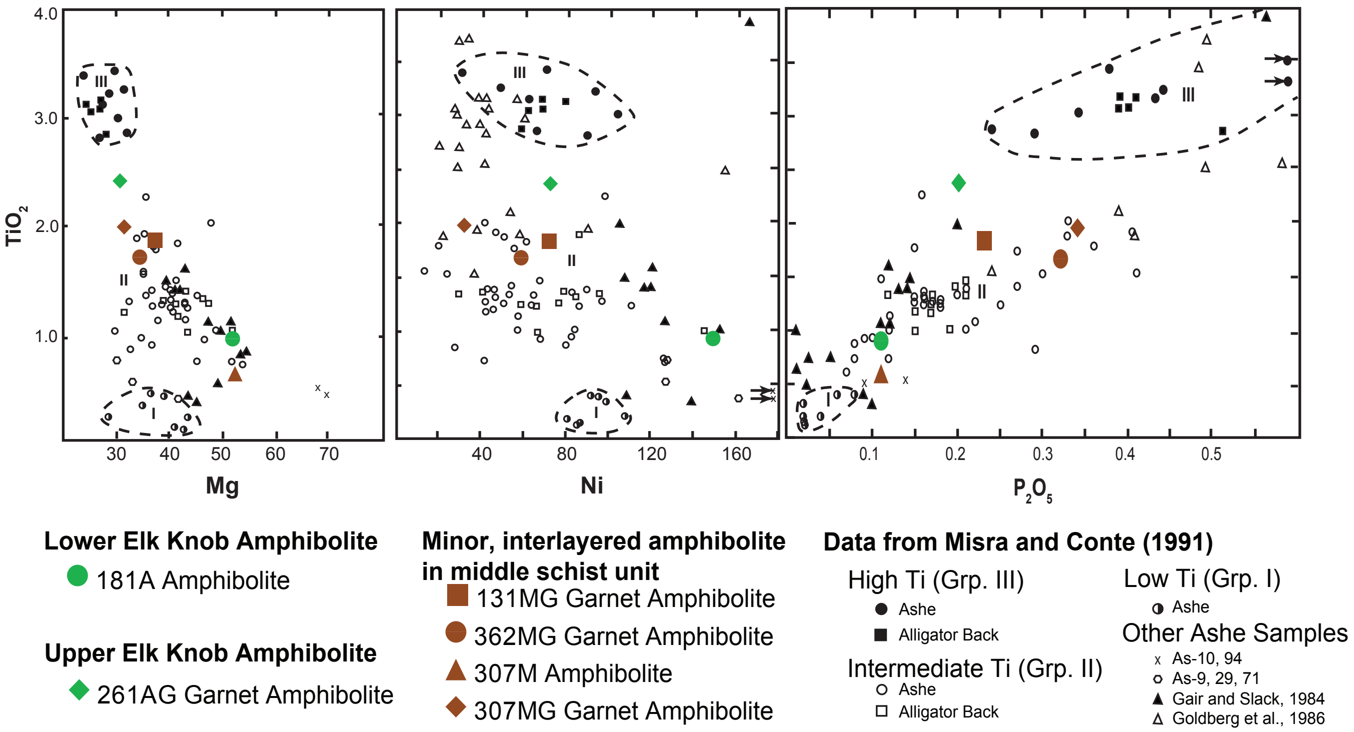

Figure 2–1–2 . Ti bivariate plots graph immobile and incompatible Ti against the fractionation index Mg* (100[Mg/(Mg+Fe+2)]), a compatible element Ni, and highly incompatible (P).

interlayered amphibolite bodies to the dominant map-scale bodies of amphibolite throughout the region. Four samples of interlayered amphibolite and one sample each from under- and overlying map-scale amphibolite bodies at Elk Knob State Park were crushed and powdered at Appalachian State University's rock prep lab and sent for analyses of major, trace, and REE geochemistry at Activation Labs in Ontario, Canada.

Comparison of Elk Knob amphibolite geochemistry to earlier major, minor, and trace element analyses of AMS amphibolites by Misra and Conte (1991), show both minor interlayered amphibolite and dominant map-scale amphibolite samples from Elk Knob to have geochemical traits consistent with Misra and Conte's (1991) compositional group II (of three compositional groups) of intermediate-Ti, which characterizes the majority of AMS amphibolites in northwestern NC (Fig. 2–1–2). Although there is scatter in several plots, major elements (Fe, Mg, Ca, Al) and compatible elements (Ni, Cr) exhibit discernable to strong fractionation trends when plotted against Zr (Fig. 2–1–3). Misra and Conte (1991) interpret decreasing MgO, Cr, and Ni with increasing Zr to indicate that compositional group II amphibolites are derived from low-pressure fractional crystallization of olivine + clinopyroxene. Misra and Conte (1991) also attribute the large decrease in V with increasing Zr content to indicate that magnetite was also a fractionating phase toward the later stage of crystallization.

At least three trends are apparent in chondrite-normalized REE patterns with samples varying from depleted to enriched in LREE ([La/Yb]N = 0.399 to 1.94), with one anomalous sample being highly enriched in LREE ([La/Yb]N = 2.91) (Fig. 2–1–4). Misra and Conte (1991) attributed similar variation in REE trends of other AMS amphibolites to dynamic melting, as LREE and HREE patterns controlled by fractionation processes alone are more consistent (not criss-crossing) and display a more narrow range of [La/Yb]N. Although there is a slight Eu anomaly in the four Elk Knob samples analyzed, plagioclase is likely not a *major* fractionating phase, which is also supported by inverse correlation of decreasing Ca and increasing Al trends in Zr plots.

Tectonic discrimination diagrams (Figs. 2–1–5A-F) show most amphibolites plot within distinct-to-overlapping N- and E-type MORB fields, as well as within-plate fields (Pearce & Cann, 1973; Floyd and Winchester, 1975; Wood, 1980; Pearce & Norry, 1979; Rollinson, 1993). The variation and overlap into E-type MORB fields and within-plate fields is due to their enrichment in LREEs, as is typical of basalts in these settings. Amphibolite samples may be enriched in LREEs due to either (1) enrichment of the source, (2) a small percent of partial melting, and/or (3) fractionation of garnet. If garnet was a fractionating phase, there should be a strong decrease in Cr ($D^{cr}_{gt/l}$ > 5) with increasing Zr/Y and decreasing Al_2O_3, as Al and Y are preferentially removed by the garnet (Misra and Conte, 1991; Pearce and Norry, 1979). Decreasing and constant Cr values with increasing Al_2O_3 and Zr/Y, respectively, coupled with inconsistent REE patterns, preclude garnet as a fractionating phase. Misra and Conte (1991) interpret the lack of LREE-HREE correlation in their amphibolite samples and the narrow range of $(La/Ce)_N$ ratios (0.81-1.3), traits which Elk Knob samples do share, to indicate that some amount of dynamic melting occurred. Conversely, mixing of a depleted asthenosphere with enriched, plume components is supported by variation in Y/Nb-Zr/Nb in Elk and other AMS amphibolites (Fig. 2–1–5G).

In conclusion, the large compositional variation in AMS amphibolites can be attributed to fractionation of an olivine + clinopyroxene +/- magnetite parental magma, a small percent of dynamic melting, and magma mixing/enrichment of depleted asthenosphere with plume-like components. Although it is not possible to distinguish whether AMS amphibolite originated at a mid-ocean ridge versus back-arc spreading center, it is clear that both map-scale bodies and interlayered amphibolite associated with pelitic rocks represent oceanic crust formed in a rift environment (Misra and Conte, 1991).

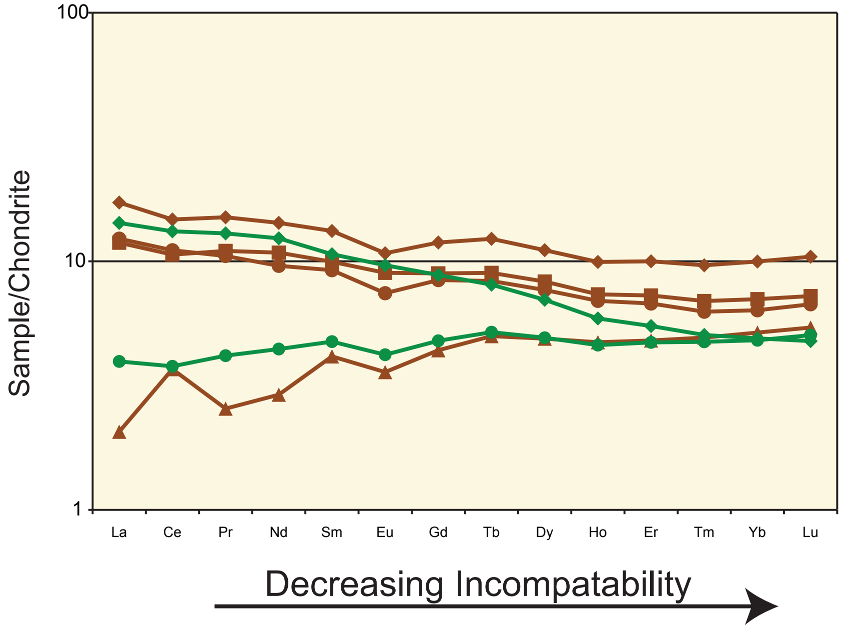

Figure 2–1–4. REE-plots normalized to chondrite values for Elk Knob amphibolite samples. See Fig.2–1–2 for explanation of symbols.

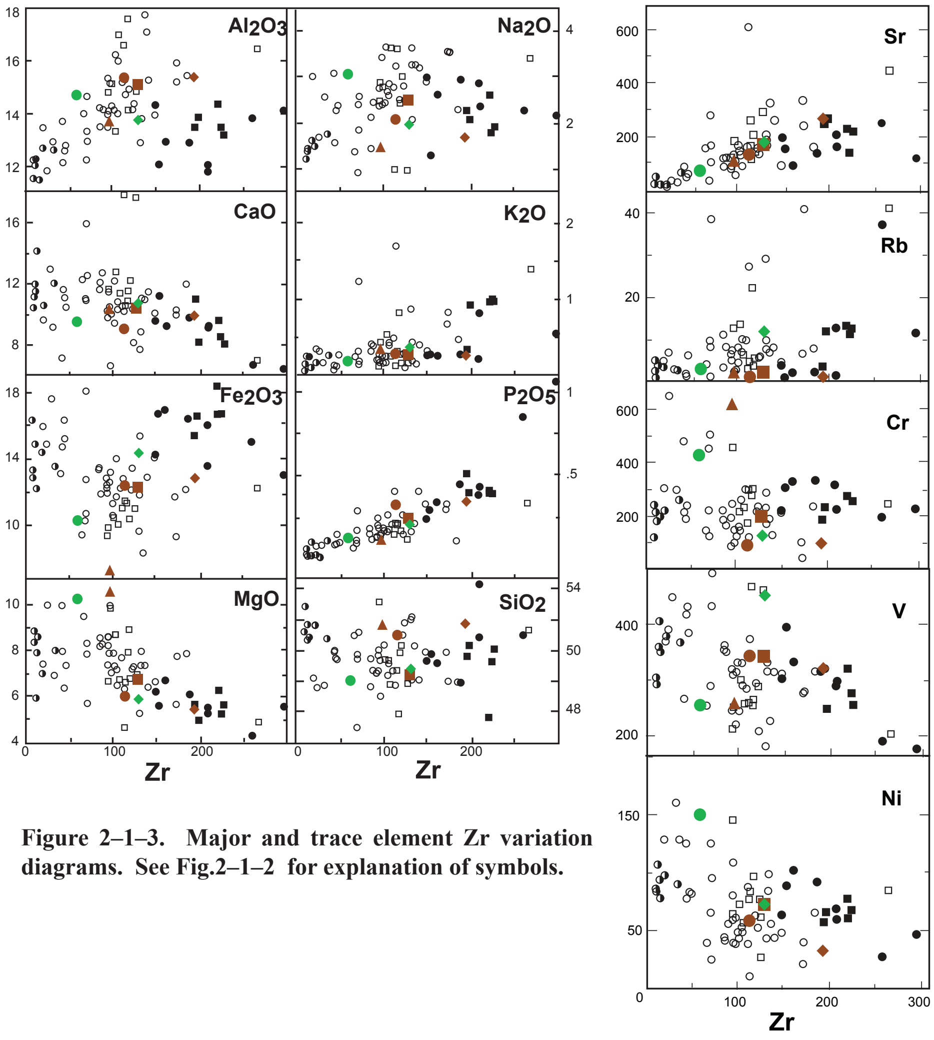

Figure 2–1–3. Major and trace element Zr variation diagrams. See Fig.2–1–2 for explanation of symbols.

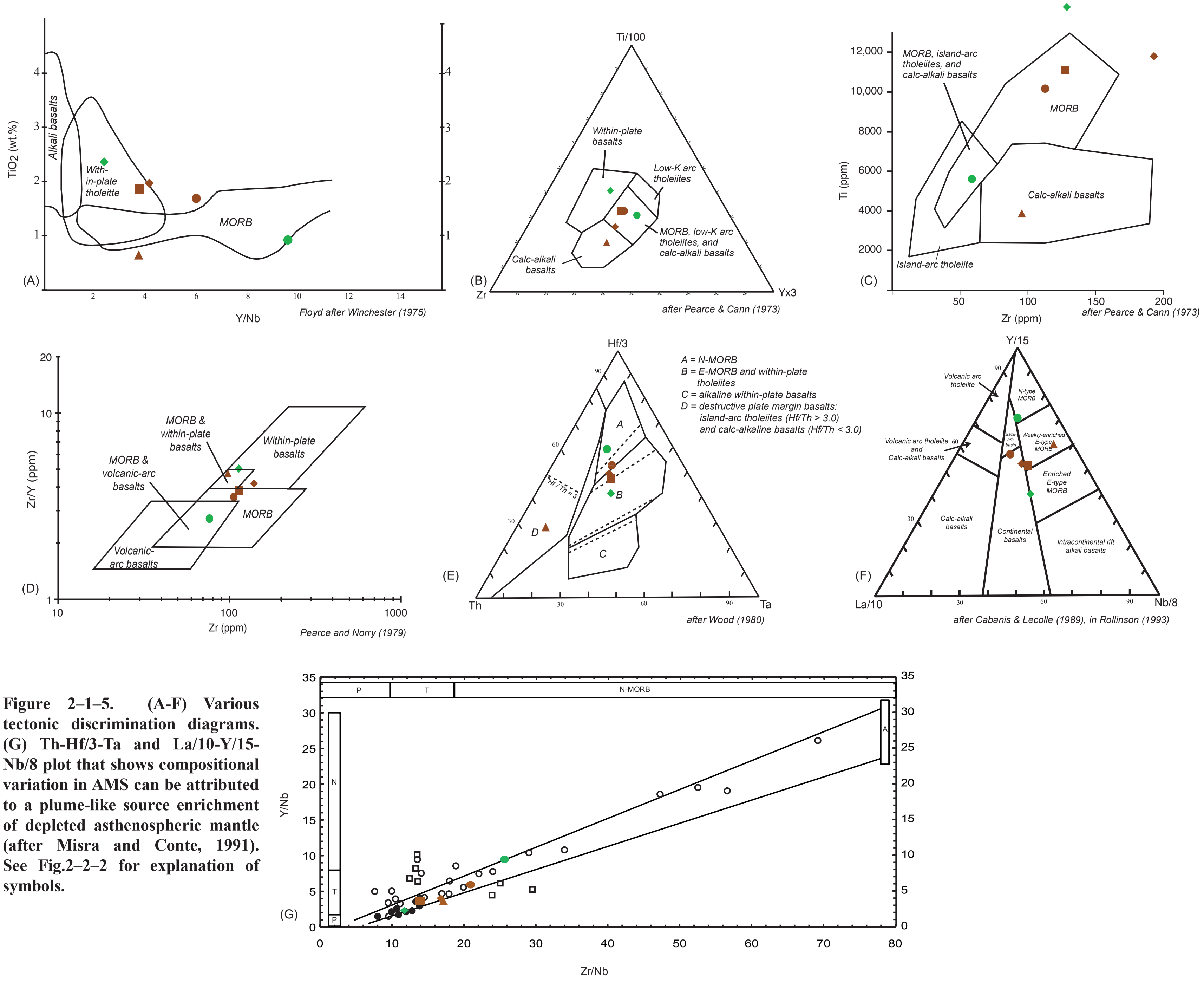

Figure 2–1–5. (A-F) Various tectonic discrimination diagrams. (G) Th-Hf/3-Ta and La/10-Y/15-Nb/8 plot that shows compositional variation in AMS can be attributed to a plume-like source enrichment of depleted asthenospheric mantle (after Misra and Conte, 1991). See Fig.2–2–2 for explanation of symbols.

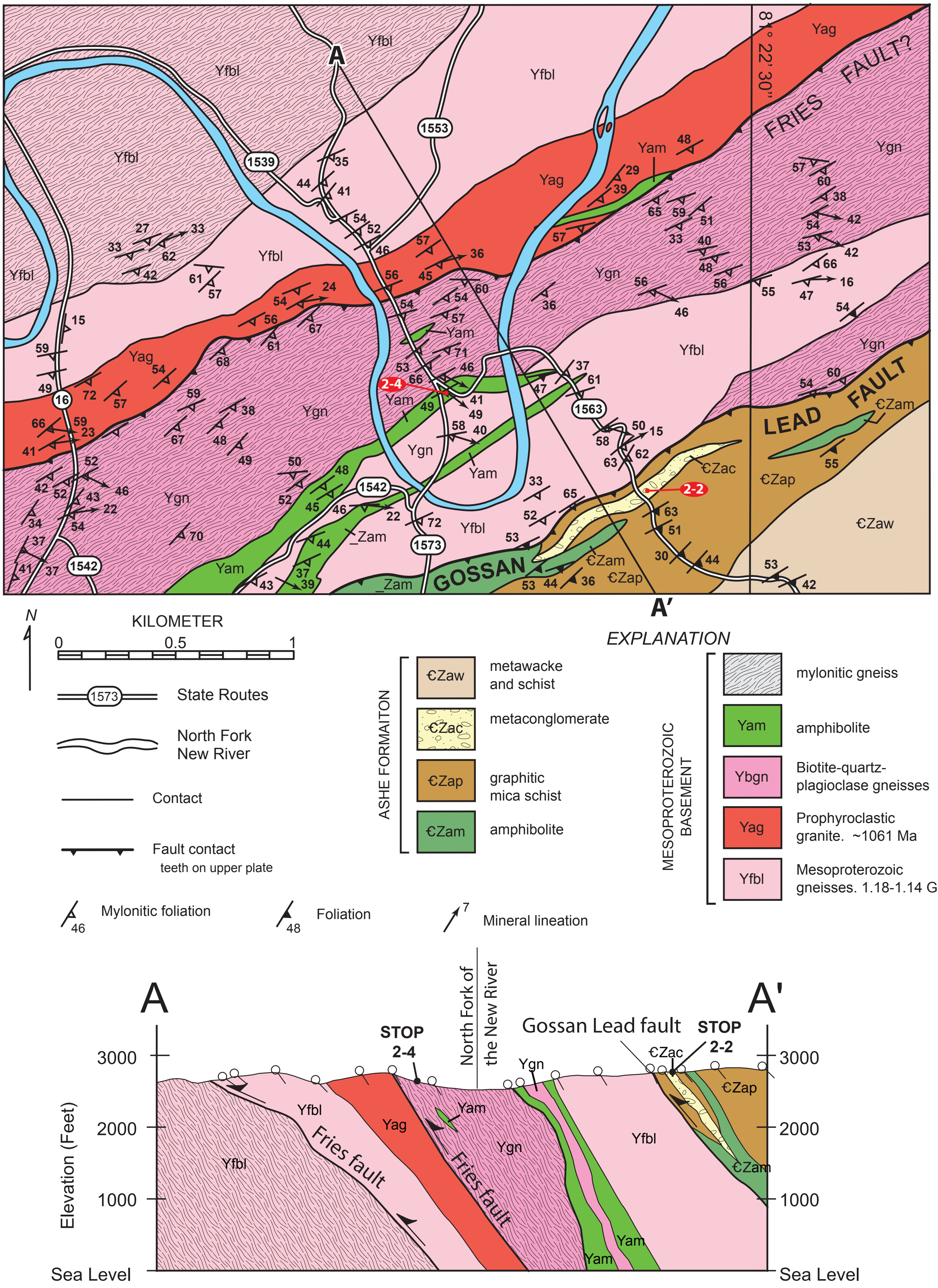

Figure 2–2–1. Geologic map with cross section of part of Grassy Creek and Mouth of Wilson 7.5–minute quadrangles, NC–VA indicating the location of Stops 2–2 and 2–4. Modified from Carter and Merschat (2014).

Stop 2–2. Metaconglomerate and graphitic schist in the Ashe Formation near Crumpler, NC (Grassy Creek 7.5–minute quadrangle; 36.50414° N, 81.37988° W). Snack break.

Rankin (1971) first described and visited this metaconglomerate locality, and it remains a pivotal outcrop that has been visited by many. We request NO HAMMERS at this outcrop to preserve it for future discussions (or a road widening project). The metaconglomerate outcrop is located ~ 500 m to the north along the road on private property (please respect the owners' property).

We will park the buses at a gravel parking lot and walk to the outcrop. On the way to the metaconglomerate outcrop are roadcuts of graphitic schist with interlayered metagraywacke layers, 5–20 cm thick. The schist has a silvery-gray sheen of fine-grained white mica and is slightly sulfidic. It was assigned to the Ashe Formation (ꞒZap) by Merschat (2011). Foliation dips moderately southeast. Biotite porphyroclasts, 1–5 mm across, occur throughout; garnet occurs locally. A subtle S-C foliation has top-to-the-NW shear sense. The metaconglomerate is within the graphitic schist.

Metaconglomerate. The outcrop is an ~2 m high ledge of metamorphosed pebble conglomerate with a pronounced foliation (055/60) and nearly down-dip mineral lineation (115/57) defined by quartz ribbons, muscovite and biotite streaks, and elongate pebbles 1–3 cm long; maximum is 14 cm (Fig. 2–2–2A). Most of the clasts are quartz, but a few clasts are quartz and feldspar, and leucocratic granitoid. Asymmetric tails on pebbles suggest top–to–NW kinematics (Fig. 2–2–2B).

The extent of the metaconglomerate (ꞒZac) is limited to a 1 km long lens within graphitic schist of the Ashe Formation (Fig. 2–2–1). As mapped, it extends from just east of the Crumpler Post Office and general store to just east of this outcrop. Included with the metaconglomerate are gray coarse-grained, metagraywacke similar to the matrix of the metaconglomerate. Located ~50 m to the northwest, structurally below the metaconglomerate, is mylonitic basement float and saprolite separated from the metaconglomerate by a thin septum of graphitic schist.

This locality was interpreted as the basal contact of the Ashe Formation on basement (Rankin, 1971, 1970; Rankin et al., 1972, 1973). Abbott and Raymond (1984) re-interpreted this outcrop to be mylonitic and in conjunction with a higher metamorphic grade, map-scale truncations, and various mafic and ultramafic rocks, proposed that the Ashe Metamorphic Suite was juxtaposed against basement along the Gossan Lead fault of Stose and Stose (1957). Later, the graphitic schist, metasandstone, and metaconglomerate were reclassified as the Wills Ridge Formation, a metaclastic unit deposited unconformably on Laurentian basement, and separated from the Ashe to the southeast by a pre-metamorphic fault (Rankin et al., 1993; Rankin, 1993). Merschat (2011) mapped graphitic schist to just southwest of Crumpler, NC, however, no structural discontinuities (faults) were recognized between the graphitic schist and other metaclastic rock units classified as part of the Wills Ridge Formation of Rankin et al. (1993), and typical schist, metagraywacke, amphibolite and ultramafic rocks of the Ashe Formation to the southeast (Figs 2–2–1). The graphitic schist can be traced from Crumpler, NC to central Virginia, where it is included with the Lynchburg Group (Carter and Merschat, 2014, this guidebook). Various studies to the southwest have concluded that the Ashe Formation and equivalents (e.g., Tallulah Falls Formation) are in fault contact with the basement or Late Neoproterozoic rocks of the western Blue Ridge (Merschat and Wiener, 1988; Trupe et al., 2003; Waters–Tormey and Stewart, 2010; Merschat et al., 2012), but documented different kinematics and time of deformation along this boundary.

^{40}Ar/^{39}Ar muscovite cooling age from the metaconglomerate did not yield a plateau, but suggests a minimum cooling age of ~334 Ma (M.J. Kunk USGS oral commun. 2012). Muscovite from a schist in the Ashe Formation ~800 m to the south yielded an ^{40}Ar/^{39}Ar cooling age of 337.8 ± 1.6 Ma and overlaps with muscovite ^{40}Ar/^{39}Ar ages from mylonitic basement to the northwest (see

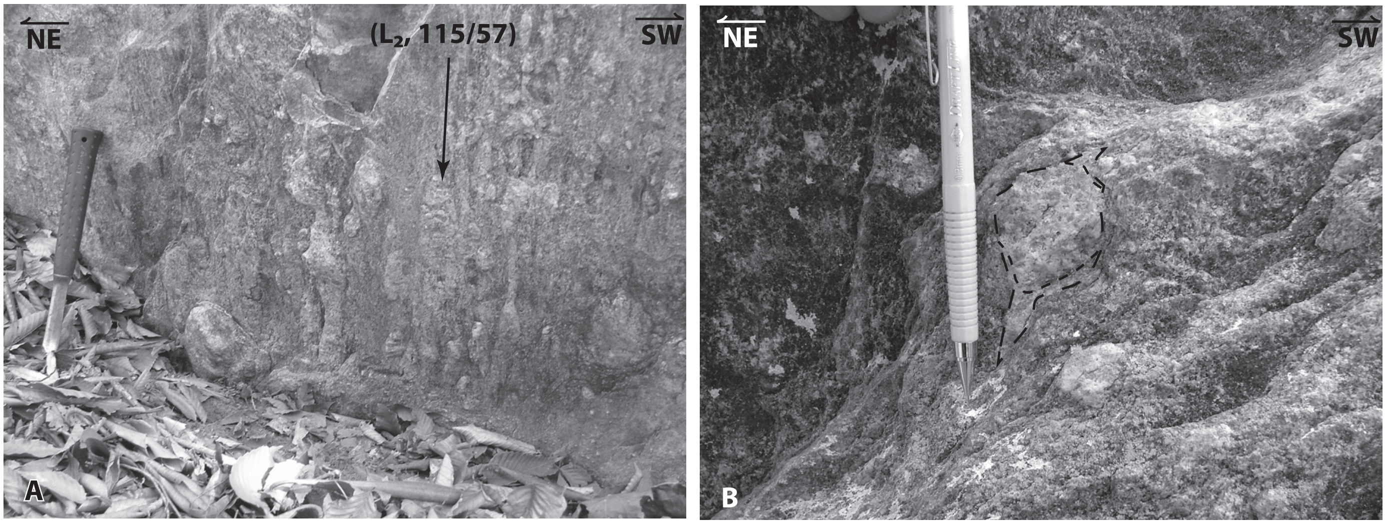

Figure 2–2–2. (A) Down-dip view of elongated quartz pebble and cobbles from the metaconglomerate near the base of the Ashe Formation. (B) Close up of quartz-feldspar-biotite pebble with asymmetric quartz tails; slightly oblique view suggests top–to–the–NW and–SW (dextral) shear sense. Pencil is 14.5 cm long; hammer is 32.5 cm long.

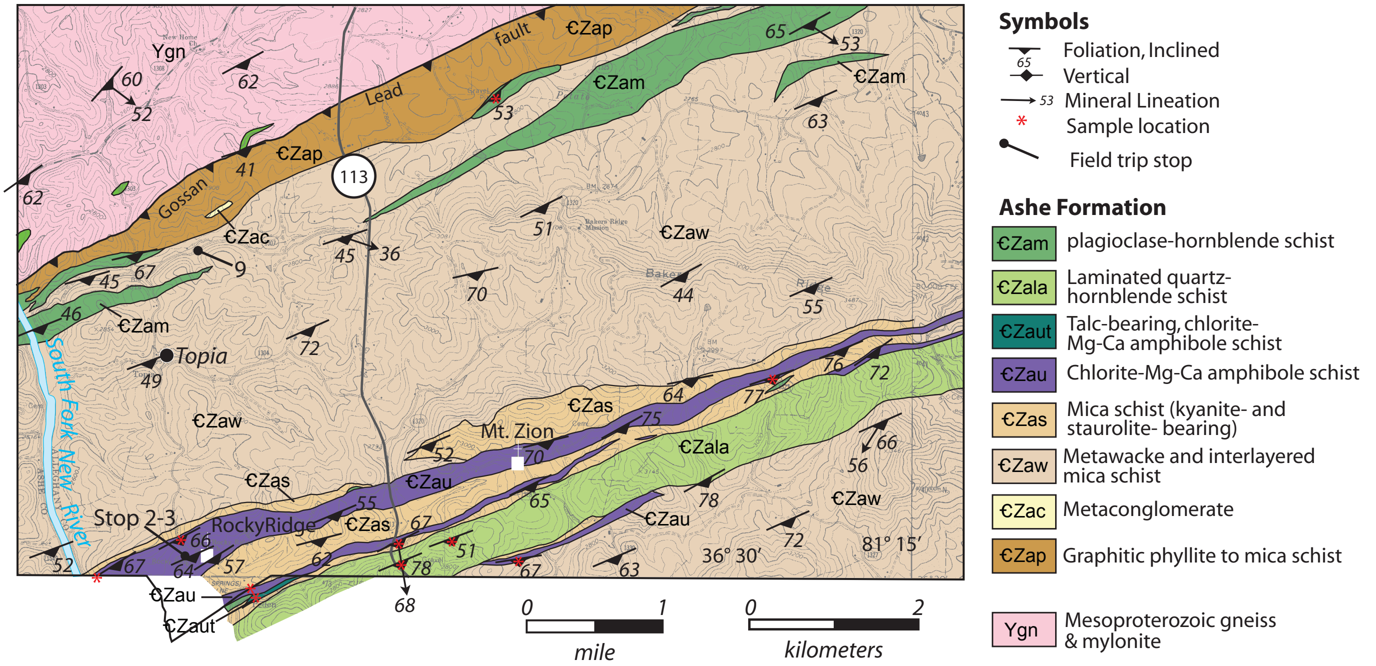

Figure 2–3–1. Simplified geologic map of part of the Mouth of Wilson, Laurel Springs, and Sparta West 7.5-minute quadrangles, showing the location of Stops 2-3. Map modified Raymond et al. (2016) and shows location of geochemical analyses presented therein.

Stops 1–2, 1–7, and 2–4). Collectively, the eastern and western Blue Ridge experienced the same deformational event related to the initial emplacement of the Blue Ridge thrust sheet at ~340 Ma. ^{40}Ar/^{39}Ar hornblende cooling ages of 360–340 Ma (Stokes et al., 2010, Stokes, 2012) indicate that the eastern Blue Ridge experienced a Devonian to Mississippian amphibolite facies metamorphic event that reached conditions in excess of 700° C (Stokes et al., 2012) prior to being juxtaposed with the basement at 340 Ma.

Stop 2–3. Rocky Ridge Ultramafic Body, Topia Road (Mouth of Wilson 7.5-minute quadrangle; 36.50198°N, 81.31300°W).

The Rocky Ridge ultramafic body (ЄZau) crops out in a linear trace for nearly 20 km (12 miles), and is concordant with the regional foliation in the Ashe Formation. It is separated from the Peden body (Raymond et al., 2016) another linear body to the south, by a thin septum of schist (Fig. 2–3–1). These ultramafic bodies contain a variety of rock types (e.g., Raymond et al., 2012, 2016). The Rocky Ridge body is largely a chlorite, Ca-Mg-amphibole schist, with minor talc, and rare pods of amphibolite. It has a porphyroclastic schistose texture containing amphibole porphyroclasts in a matrix of amphibole, chlorite, and talc. Amphibole grew in three events: early inclusion-rich Mg-hornblende cores rimmed with clear Mg-actinolite (Fig. 2–3–2). Some matrix amphiboles grew at the same time as rims on the porphyroclasts, and compositions do not vary significantly in the rims (Raymond et al., 2012, 2016). Minor late stage cummingtonite is the matrix, originally reported by Stapor et al. (2010), occurs in some rocks. A pod of amphibolite consists of amphibole–plagioclase–quartz–sphene–magnetite and is not as strongly foliated as the chlorite–amphibole

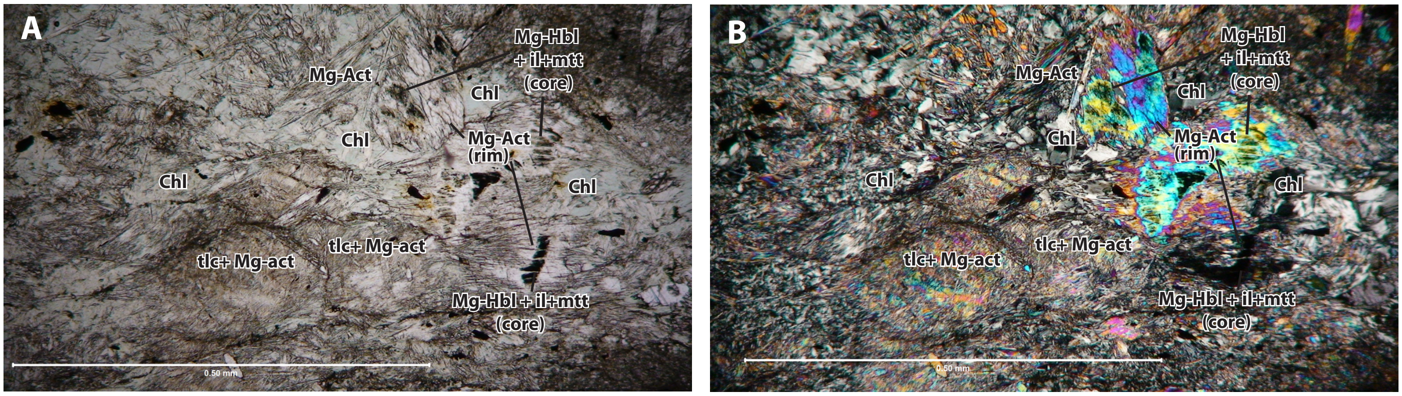

Figure 2–3–2. Photomicrograph of Rocky Ridge talc-chlorite-amphibole schist with multiple amphibole generations: early included Mg-hornblende (Mg-Hbl) cores, and late Mg-actinolite (Mg-act) rims and matrix amphiboles. Matrix also contains chlorite (chl), talc (tlc), magnetite (mtt), and illmenite (il). Photos are plane polarized light (A) and cross-polarized light (B).

schist. The Rocky Ridge and Peden ultramafic rocks have chemistry similar to other rocks in the Ashe Formation that are interpreted to be ophiolite fragments (Raymond et al., 2016).

Stop 2–4. Fries fault and high-strain zones, Old NC 16 (Grassy Creek 7.5–minute quadrangle; 36.50984° N, 81.39068° W).

The weathered rocks exposed in the poison ivy-covered outcrops (CAUTION!) consist of different protomylonitic and mylonitic basement lithologies (Fig. 2–2–1). These rocks are part of a broad high-strain zone of the Fries fault and have been interpreted to include rocks of the Mars Hill terrane and Pumpkin Patch Metamorphic Suite (Raymond, 2000; Hatcher et al., 2006). To the northeast, Stose and Stose (1957) portrayed the Fries fault as a fault zone and is mapped to the in the area around Mount Rogres as a series of anastomosing greenschist-facies high-strain zones that vary from meter to kilometers wide and occur throughout the northern end of the French Broad massif (Tollo et al., 2012; Southworth et al., 2012). Stop 2–4 is within one of the wider high-strain zones, which also includes Stop 1–2, and is in the immediate footwall of the Gossan Lead fault, located ~ 0.8 km (0.5 miles) to the southeast (Stop 2–2).

The rocks are greenish-gray to silvery greenish-gray porphyroclastic epidote–chlorite–muscovite–biotite–quartz–plagioclase gneiss and schist (Ygn), muscovite–quartz–plagioclase schist and minor amphibolite (Yam). Centimeter-thick layers of quartz and feldspar parallel to the mylonitic foliation pinch and swell, and are boudinaged. These layers likely were migmatite leucosomes from a high-grade gneiss or schist protolith, which are now mylonitic to protomylonitic (S–L tectonites) with a down–dip mineral stretching lineation, L_3, (trends 105°–120° and plunges 40°–50°) and top-to-the-NW kinematic shear sense. Mylonitic foliations (mean 58/54) in the basement is roughly parallel to S_2 Ashe Formation (Fig. 2–2–1; see also Merschat et al., this guidebook, Fig. 14). Biotite within the mylonitic foliation, S_3, annealed quartz ribbons with some subgrains, and slightly undulose feldspar porphyroclasts suggest deformation temperatures in the eastern part of the Fries high-strain zone were higher than further west, possibly related to being proximal to the Gossan Lead fault. Muscovite from these outcrops yielded an $^{40}Ar/^{39}Ar$ age 335 ± 2 Ma (M. Kunk USGS oral commun. 2012) and is within error of muscovite $^{40}Ar/^{39}Ar$ ages from the Ashe Formation/Gossan Lead thrust sheet. The rocks of the Fries high-strain zone and eastern Blue Ridge experienced the same deformation event at ~340 Ma related to the emplacement of the Blue Ridge thrust sheet.

To the southwest near Boone, NC, Bryant and Reed (1970) mapped a "mixed unit" of biotite–plagioclase schist to gneiss, amphibolite and calc-silicate located structurally beneath rocks now assigned to the Ashe Formation. This unit was correlated with the granulite-facies paragneisses and migmatitic hornblende gneisses of the Mars Hill terrane west and southwest of the Grandfather Mountain window (Fig. 1; Merschat et al., this guidebook) (Raymond and Johnson, 1994; Raymond, 2000; Trupe et al., 2004; Hatcher et al., 2006). Paragneiss and amphibolite are also found northeast of the North Carolina–Virginia state line along the boundary between basement and cover rocks, although Rankin et al., (1972) classified the rocks as basement above the Fries fault and beneath the unconformable cover rocks. We suggest that these paragneisses and associated rock types occupy a similar position may be a Mesoproterozoic sedimentary cover sequence and could potentially be correlative with rocks assigned to the Mars Hill terrane. Stop

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NOTES:

	Ok	**Knobs Formation**
	OꞒk	**Knox Group**, undivided
	Ꞓel	**Elbrook Formation**
	Ꞓr	**Rome Formation**
	Ꞓs	**Shady Dolomite**
Chilhowee Group	qꞒe	**Erwin Formation**, q-quartzite
Chilhowee Group	Ꞓem	Murray shale member
Chilhowee Group	Ꞓh	**Hampton Formation**
Chilhowee Group		**Unicoi Formation**
Chilhowee Group	Ꞓuu	Upper member; quartzite
Chilhowee Group	ꞒZub	Basalt
Chilhowee Group	ꞒZul	Lower member; conglomerate, arkose and shale
		Ashe Formation
	ꞒZala	Laminated amphibolite
	ꞒZas	Muscovite schist and metagraywacke
	ꞒZaw	Metagraywacke and muscovite schist;
	ꞒZau	Meta-ultramafic, talc-tremolite-chlorite schist
	ꞒZam	Amphibolite
	ꞒZac	Metaconglomerate*
	ꞒZap	Graphitic schist and metagraywacke*
		Konnarock Formation
	Zkd	Diamictite
	Zka	Arkose and conglomerate
	Zk	Rhythmite, laminite & mudstone undivided

B

		Mount Rogers Formation
Razor Ridge volcanic center	Zmg	Granophyre
Razor Ridge volcanic center	Zmr2	Razor Ridge rhyolite outliers
Razor Ridge volcanic center	Zmr	Razor Ridge rhyolite; 748.4 ± 3.2 Ma
Mount Rogers volcanic center		Wilburn Rhyolite Member; 749.7 ± 3.1 Ma^3
Mount Rogers volcanic center	Zmw	Welded tuff
Mount Rogers volcanic center	Zmwb	Volcanic breccia
Mount Rogers volcanic center		Whitetop Rhyolite Member, 753.3 ± 2.0 Ma^3
Mount Rogers volcanic center	Zmwtt	Lapilli tuff
Mount Rogers volcanic center	Zmwt	Flow-banded lava
Mount Rogers volcanic center	Zmb	Buzzard Rock Rhyolite, 755.0 ± 6.6 Ma^3
Mount Rogers volcanic center		Fees Rhyolite, 753.1 ± 2.7 Ma^3
Mount Rogers volcanic center	Zmf	Porphyritic rhyolite
Mount Rogers volcanic center	Zmfd	White to tan rhyolite dike
Mount Rogers volcanic center	Zmfb	Porphyritic rhyolite with fiamme and lithic clasts
Pond Mountain volcanic center	Zmp5	Pond Mountain rhyolite5
Pond Mountain volcanic center	Zmp4	Pond Mountain rhyolite4
Pond Mountain volcanic center	Zmp3	Pond Mountain rhyolite3
Pond Mountain volcanic center	Zmp2	Pond Mountain rhyolite2
Pond Mountain volcanic center	Zmp1	Pond Mountain rhyolite1
Pond Mountain volcanic center	Zmp	Pond Mountain porphyry; 758.7 ± 2.9 Ma
	Zmt	Undivided; may be transitional with Konnarock Fm.
	Zml	Arkose, conglomerate and shale, undivided
	Zmac	Arkose and conglomerate
	Zmbb	Boulder conglomerate
	Zmb	Greenstone and basalt

Zdg	Diabase and gabbro dikes; 757 ± 5 Ma^3
	Mesoproterozoic rocks
Ym	Marble and pegmatite
Yag	Augen gneiss, quartz monzogranite; 1046–1061 Ma^{2,3}
Ylbg	Lineated biotite meta-granite; 1134 ± 5 Ma^{2,3}
Yqm	Meta-quartz monzonite; 1155 ± 12 Ma^{2,3}
Ybg	Biotite granite, alkali-feldspar granite, and monzograntie; 1153–1174 Ma^{2,3}
Yfbl	Foliated biotite leucogranite; 1177 ± 7 Ma^{2,3}
Ygn	Mars Hill terrane? – Migmatitic biotite gneiss, amphibolite & granite
Yam	Mars Hill terrane? – Amphibolite
Ybm	Migmatitic biotite gneiss & schist
Yga	Amphibolite & hornblende gneiss
Ygg	Migmatitic biotite gneiss, granofels, & orthograonfels ~1.3 Ga^{2,3}

*Included in the Wills Ridge Formation of Rankin et al. (1993).

Map Symbols

▨ High-strain zone

Contact

Thrust fault

Fault

Normal fault

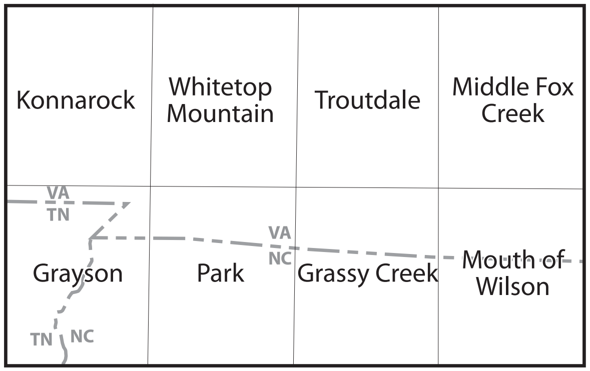
Fold axis; syncline

▲ Ar/Ar Sample

● U-Pb Detrital Zircon Sample

◆ U-Pb Zircon Geochronology Sample

1-9 2016 CGS Field Trip Stop



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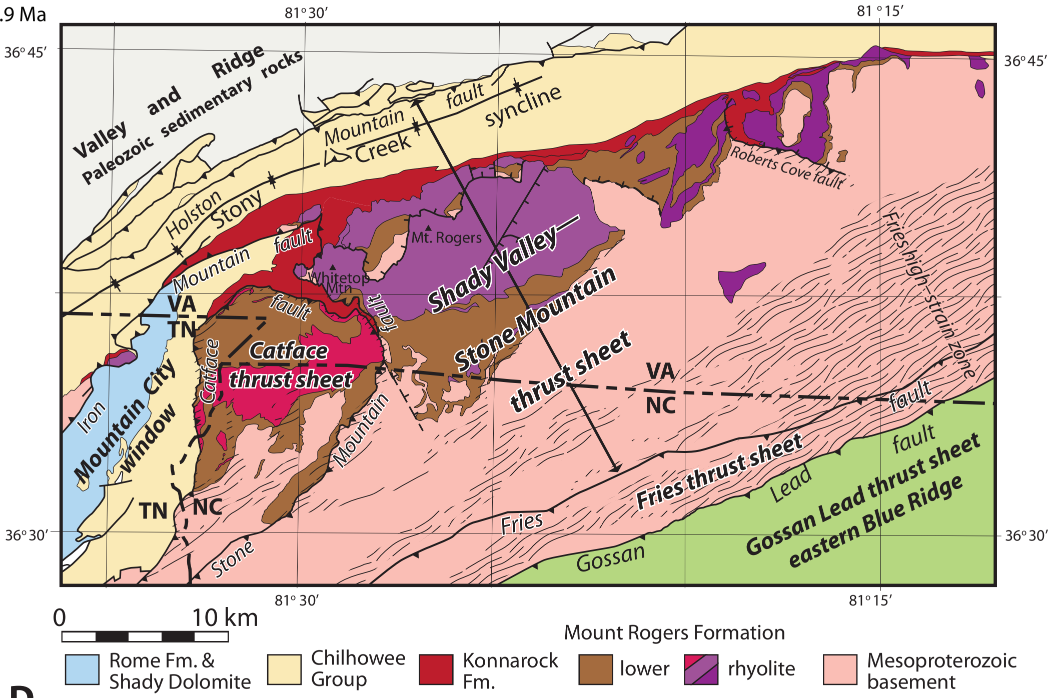

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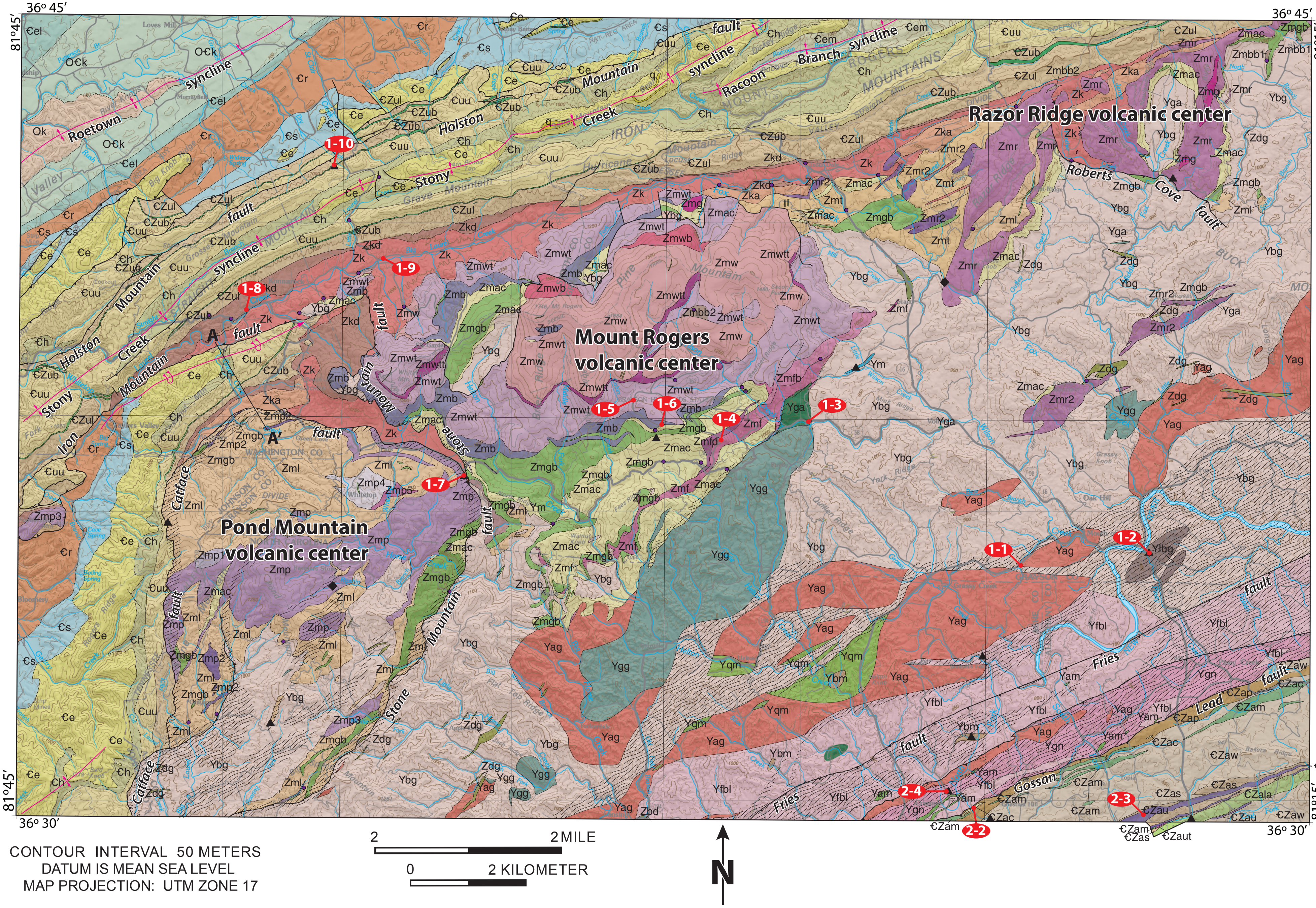

CONTOUR INTERVAL 50 METERS
DATUM IS MEAN SEA LEVEL
MAP PROJECTION: UTM ZONE 17