A Taconic Backarc and Arc Terrane in the Southern Appalachians: Correlating Geologic Units of the Blue Ridge and Western Inner Piedmont of Georgia and Alabama

Edited by: Clinton Barineau and James Tull

51st Annual Field Trip of the Georgia Geological Society
Carrollton, Georgia, October 6-8, 2017
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Enjoy the trip!
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The publication of a field trip guidebook provides a record of the Society’s activities and allows us to make a permanent contribution to the advancement of our science. It is always a struggle to get everything pulled together in time for the trip. This year has been no exception, and we are particularly grateful to Randy Kath and Publications and Printing at the University of West Georgia for their efforts in producing the guidebook at the last possible minute.

The Saturday night social is an important part of the fieldtrip weekend. This year financial support was generously provided by Petrologic Solutions, Inc.

We hope everyone enjoys the trip. Several generations of geologists have contributed to the conclusions presented in this guidebook. Many of them are attending this meeting to discuss various interpretations. We can all look forward to a stimulating and fun weekend.
Taconic Backarc-Arc Terranes in the Southern Appalachians: Correlating Geologic Units of the Blue Ridge and Western Inner Piedmont of Georgia and Alabama

Clinton I. Barineau1, James F. Tull2, and Christopher S. Holm-Denoma3

1Columbus State University, Dept. of Earth and Space Sciences, Columbus GA 31907
2Florida State University, Dept. of Earth, Ocean, and Atmospheric Sciences Tallahassee, FL 32306
3U.S. Geological Survey, Central Minerals and Environmental Resources Science Center, Denver, Colorado 80225

Abstract

Decades of geologic mapping, geochronology, and geochemical analyses of rocks from the southern Appalachian orogenic belt of Alabama (AL), Georgia (GA) and North Carolina (NC) have led to a better understanding of its geologic history. In recent years, a synthesis of available geologic data from units within the Talladega belt (AL-GA), Opelika Group (AL), Eastern Blue Ridge (AL-GA) and Dahlonega gold belt (GA-NC), as well as units of the western Inner Piedmont Dadeville Complex (AL-GA), has resulted in a major change in our understanding of the Ordovician Taconic orogeny in this region. Unlike earlier models of Taconic collisional orogenesis for the Appalachians, which were largely developed in the Northern Appalachians and applied along the entire length of the orogen, data from Ordovician geologic units in the southern Appalachians indicate that the southern Appalachian Taconic orogeny developed in an accretionary orogen similar to that of the Sea of Japan back-arc and Japan arc in the western Pacific. This model of accretionary orogenesis for the southern Appalachians is primarily based on the presence of rocks representing a paired backarc-arc system in the western Inner Piedmont and Blue Ridge of AL and GA – both of which contain evidence that they directly sourced or were built atop Laurentian continental crust. This backarc-arc system, which faced the open Iapetus Ocean from the Early to Late Ordovician, remained largely undeformed until the Acadian/NeoAcadian orogeny in the latest Devonian-Mississippian, and was emplaced atop units of the Laurentian shelf during the Carboniferous-Permian Alleghanian orogeny as a number of distinct thrust sheets. Recognition of this Laurentian plate Ordovician suprasubduction system has important implications for our understanding of the southern Appalachian Taconic orogeny as an accretionary orogen, rather than the classic collisional orogenic model of the northern Appalachians. In this guidebook, we present a synthesis of geochemistry, geochronology, and mapping efforts from numerous workers in the region, correlate and refine stratigraphic units from the eastern Blue Ridge and western Inner Piedmont of GA-AL, along with their bounding faults, and present an updated Paleozoic tectonic history for this portion of the orogenic belt.

Geologic Overview

The southern Appalachians of AL, GA, and NC (Fig. 1.1) can generally be divided into different lithotectonic terranes bounded by significant faults, each with its own genetic origin. From northwest to southeast (Fig. 1.2), these include: the foreland fold and thrust belt (AL-GA); Talladega slate belt (AL-GA and referred to herein as the
Talladega belt) and equivalent western Blue Ridge (GA, NC); eastern Blue Ridge (AL-GA) and equivalent Dahlonega gold belt (GA-NC); western Inner Piedmont Dadeville Complex (AL-GA); and Opelika Group (AL, GA). Although some of these terranes include rocks that were once thought to be exotic to Paleozoic Laurentia, a growing body of evidence gathered from geologic mapping, geochronological and geochemical (including isotopic) work suggests all of these units formed on the southern margin (southeastern margin using a modern coordinate reference frame) of the Laurentian plate, and were telescoped by thrusting during the Acadian/neoAcadian and Alleghanian orogenies. Here, we provide a general overview of early Paleozoic stratigraphic units and bounding/internal faults (Fig. 1.2) for each of these southern Appalachian lithotectonic terranes as a reference frame for understanding the tectonic history of the region. Bounding faults are described in more detail in Tull, Holm-Denoma and Almuntashe (this guidebook).

**Figure 1.1.** Geologic map of the southern Appalachians depicting major terranes and bounding faults, Mesoproterozoic (Grenville) basement, and rocks of the Dadeville Complex (arc) and Wedowee-Emuckfaw-Dahlonega basin (back-arc). Adapted from Barineau et al., 2015.

**Foreland Fold and Thrust Belt**

The southern Appalachian foreland includes Paleozoic rocks that span the Cambrian to Pennsylvanian systems. These rocks are contained in a series of thrust sheets between the Appalachian Plateau on the northwest and the Talladega-Cartersville fault (AL-GA) and equivalent Great Smoky fault (GA) to the southeast (Fig. 1.2), separating them from the structurally overlying Talladega belt-western Blue Ridge lithotectonic terrane. The oldest of these foreland units, the lower Cambrian (time scale from Cohen et al., 2013; Fig. 1.3) Chilhowee Group, represents drift-facies shelf deposits which formed on the Iapetus-facing margin of Laurentia following its latest Neoproterozoic breakout from the Rodinian supercontinent (King et al., 1960; Simpson and Sundberg, 1987; Simpson and Eriksson, 1989). Atop rocks of the Chilhowee Group, a sequence of carbonates and subordinate mixed siliciclastic units comprise the Shady Dolomite, Rome Formation, Conasauga Formation, and Knox
Figure 1.2. Geologic map of the southernmost Appalachians of Alabama and Georgia depicting lithotectonic terranes, major faults and stratigraphic units. Modified from Barineau et al., 2017.
Group, all of which formed on the Laurentian passive margin during progressive opening and widening of the Iapetus Ocean (Rodgers, 1953; Thomas and Drahovzal, 1973; Rankin et al., 1989; Thomas, 1991; Glumac and Walker, 2000). These Cambrian to Lower Ordovician carbonate platform sequences are truncated by the post-Knox unconformity, separating them from overlying Middle Ordovician strata of the Lenoir Limestone (AL-GA), Little Rock Limestone (AL) and temporal equivalent Rockmart Slate (GA), and Athens Shale (AL-GA). The post-Knox unconformity has traditionally been assumed to mark uplift of the Cambrian-Ordovician carbonate platform as a response to tectonic loading by advancing thrust sheets during the Taconic orogeny (Keller, 1977; Shanmugam and Walker, 1978; Shanmugam and Lash, 1982; Quinlan and Beaumont, 1984; Beaumont et al., 1988; Hatcher, 1989; Diecchio, 1993; Finney et al., 1996; Ettensohn, 2004; Bayona and Thomas, 2006). This foreland basin-style model for the origin of these Middle Ordovician sedimentary sequences (Blount clastic wedge), however, has presented a number of problems for southern Appalachian geologists. These include observations that detritus in this subsiding Blount basin was completely derived from Laurentian units (i.e. no exotic components) and the Ordovician (i.e. Taconic) thrust sheets from which they were purportedly derived have never been identified. Timing of destabilization and subsidence of the passive margin is marked by deposition of black shale during the Darriwilian stage (Middle Ordovician of Cohen et al., 2013) along the Alabama promontory (Bayona and Thomas, 2003, 2006). These Middle Ordovician sedimentary sequences are intercalated with altered volcanic ash deposits (k-bentonites) interspersed from near the base (Darriwilian stage) to the top of the sequence (Katian stage). The most significant of these, the 448 ± 2 Ma Millbrig and 449 ± 2.3 Ma Deicke (Min et al., 2001) palinspastically thicken towards the Alabama promontory and Tennessee embayment, suggesting a source for these ignimbrite eruptions in this region (Kolata et al., 1998). Both units have geochemical and isotopic signatures that suggest they incorporated evolved continental crust during their genesis (Huff et al., 1992; Coakley and Gurnis, 1995; Kolata et al., 1996, 1998; Haynes et al., 2011; Samson et al., 1989).
Talladega Belt-Western Blue Ridge

Southeast of the foreland fold and thrust belt, in the hanging wall of the Talladega-Cartersville and Great Smoky faults (Fig. 1.2), the Talladega belt (AL-GA) and western Blue Ridge (GA) consist of metamorphosed Paleozoic units built atop Grenville-age basement and its cover of late Neoproterozoic rift sequences that developed outboard (seaward) of the palinspastic position of the foreland fold and thrust belt on the Laurentian plate. Lower Paleozoic units in the Talladega belt (Fig. 1.4) include the Cambrian-Ordovician Kahatchee Mountain Group, Sylacauga Marble Group and Hillabee Greenstone. The Kahatchee Mountain and Sylacauga Marble Groups, a sequence of lower greenschist facies metapelites (e.g. locally carbonaceous chlorite sericite phyllites), metagraywackes, quartzites, and marbles represent outboard equivalents of the lower Cambrian Chilhowee Group (Kahatchee Mountain Group) and Cambrian-Ordovician Shady Dolomite through Knox Group units (Sylacauga Marble Group) of the foreland fold and thrust belt (Johnson and Tull, 2002; Tull et al., 2010; Tull and Barineau, 2012). Paleontologic data indicate that these rocks are of Laurentian provinciality, and palinspastic restoration of the Talladega belt indicates these units formed at the edge of the Laurentian margin. As such, rocks of the Kahatchee Mountain and Sylacauga Marble Groups represent the most distally preserved portion of the lower Paleozoic shelf along the palinspastic location of the Alabama promontory (Ferrill and Thomas, 1988; Thomas, 2004, Tull et al., 2010; Tull and Barineau, 2012).

At the structural top of the Talladega belt (Fig. 1.4), a Middle Ordovician bimodal volcanic suite, the ca. 469 Ma Hillabee Greenstone (Tull and Stow, 1980; Tull et al., 2007; McClellan et al., 2007), was emplaced atop Upper Devonian-Mississippian (?) metasedimentary rocks of the upper Talladega Group (i.e. Jemison Chert-Erin Slate) along the pre-metamorphic Hillabee thrust (Tull et al., 2007; Barineau, 2009; Tull and Barineau, 2012). Because both the Hillabee Greenstone and structurally underlying metasedimentary rocks (i.e. Kahatchee Mountain, Sylacauga Marble and Talladega Groups) share the same metamorphic fabrics and facies, as well as the same deformational elements, emplacement of the Hillabee Greenstone atop Talladega Group shelf units must have occurred prior to Late
Mississippian metamorphism (ca. 330 Ma; McClellan et al., 2007; Hames et al., 2007). Additionally, because the Hillabee Greenstone did not experience significant deformation during tectonic emplacement along the Hillabee thrust, and thus was not likely to have been transported across one or more significant thrust ramps during translation, this metavolcanic sequence must have formed in a palinspastic location relatively proximal to the Paleozoic Laurentian shelf (Tull et al., 2007; Barineau, 2009; Tull and Barineau, 2012). Trace element geochemical analysis of the predominant greenstone (e.g. actinolite epidote chlorite phyllite and massive greenstone) lithologies suggest they represent tholeiitic basaltic magma erupted in a back-arc setting, whereas $\varepsilon$Hf values of zircon from subordinate metadacite (metaignimbrite) units indicate these magmas were partial melts derived Grenville-aged (ca 1.0-1.1 Ga) continental crust (Tull et al., 2007, 2014).

Lower Paleozoic rocks of the western Blue Ridge (Fig. 1.2), including the Nantahala and Brasstown Formations and Murphy Marble of the Hiawassee River Group, represent along-strike equivalents of the Talladega belt Kahatchee Mountain and Sylacauga Marble Groups and most workers similarly interpret these western Blue Ridge units of GA and NC as stratigraphic equivalents of the Chilhowee Group siliciclastic sequences and Shady Dolomite carbonate platform rocks (Groszos and Tull, 2007; Tull et al., 2010). Talladega belt lithologies, between the Talladega-Cartersville fault on the northwest and Hollins Line and Allatoona faults on the southeast, significantly narrow as they strike from AL and western GA into northern GA and equivalent rocks of the western Blue Ridge. Although more than 25 km wide in eastern AL along the GA state line, in the vicinity of Cartersville GA, the entire belt narrows to less than 3 km across the location of the Cartersville transverse zone (Tull and Holm, 2005). Interpreted as the location of a subsurface oblique thrust ramp (Fig. 1.5), lithologies of the Talladega Group from the southwest strike into those of the equivalent Mineral Bluff Group to the northeast in the western Blue Ridge (Tull and Holm, 2005; Barineau et al., 2017). Northeast of the Cartersville transverse zone, rocks of the western Blue Ridge include Talladega Group-equivalent passive margin,
drift facies units of the Nantahala and Brasstown Formations and Murphy Marble. However, northeast of the Cartersville transverse zone, the western Blue Ridge also contains an up to 10 km-thick underlying package of rift-related rocks, the Ocoee Supergroup, which has led many workers to interpret the Cartersville transverse zone as a first order structure in the rift architecture of the Laurentian margin – a significant transform fault separating the Alabama promontory from the Tennessee embayment (Rankin, 1975; Thomas, 1977, 1991; Tull and Holm, 2005). Transition from an upper plate to lower plate rifted margin (Fig. 1.5) at the AL promontory-TN embayment (Wernicke, 1985; Lister et al., 1986, 1991; Tull and Holm, 2005) coincides with the appearance of Grenville-aged basement massifs (e.g. Corbin Gneiss) unconformably overlain by basement-fault bounded continental rift-facies units of the Ocoee Supergroup in the western Blue Ridge (Li, 1996, Li and Tull, 1998). In contrast, no basement massifs or rift-facies rocks have been identified in the Talladega belt, consistent with interpretation of the AL promontory as an upper plate rifted margin.

Eastern Blue Ridge
Ashland-Wedowee-Emuckfaw Belt (AL)

The middle-upper amphibolite facies eastern Blue Ridge of AL (i.e. Ashland-Wedowee-Emuckfaw belt) lies in the hanging wall of the post-metamorphic Hollins Line fault (Fig. 1.2), a dextral transpressional footwall thrust duplex system on the southeastern flank of the Talladega belt (Tull et al., 2007; Barineau, 2009; Tull and Barineau, 2012). In GA, equivalent eastern Blue Ridge rocks lie in the hanging wall of the Allatoona fault, which post-dates and truncates the Hollins Line fault immediately west of the AL-GA state line (Fig. 1.2). Southwest of Hightower, displacement along the Allatoona fault gradually decreases into its tip point southwest of Goodwater, AL.

At the structural and stratigraphic base of the eastern Blue Ridge in AL (Fig. 1.6), the Higgins Ferry and equivalent Poe Bridge Mountain Groups (i.e. lower Ashland Supergroup; Neathery, 1975; Tull, 1978; Barineau et al., 2015) consists predominantly of biotite-muscovite schist of varying garnet, feldspar, quartz, and graphite content, locally containing staurolite, kyanite and sillimanite. Subordinate rocks include fine-grained paragneiss, gametiferous graphitic quartzite, pegmatites and amphibolite. Migmatitic units in the Ashland Supergroup, which when coupled
with the presence of high pressure-temperature (P-T) aluminosilicates, support conditions up to 775 °C and 8.6 kb (i.e. middle to upper amphibolite facies metamorphism; Allison and Morisani, 2002). Amphibolite in the lower Ashland Supergroup (Fig. 1.6), which makes up <10% of the stratigraphy (Tull et al., 2007) occurs as discontinuous lenses, but also as thick, laterally extensive layers (Fig. 1.2) >30 km along strike (e.g. Mitchell Dam and Ketchapedrakee amphibolites). Geochemical, petrographic and field relationships indicate igneous protoliths for these amphibolite bodies, which are interpreted to have developed as mafic flows and/or sills in a continental rift setting (Stow et al., 1984; Tull et al., 2014; Barineau et al., 2015).

Stratigraphically above the Higgins Ferry-Poe Bridge Mountain Groups, rocks of the overlying Hatchett Creek and equivalent Mad Indian Groups (Fig. 1.6) consist of paragneiss, micaceous quartzite, calc-silicate, and graphitic schist (Tull, 1978; Drummond et al., 1988; Allison, 1992). Like units in the lower Ashland Supergroup, rocks of the Hatchett Creek-Mad Indian Groups include pegmatite, migmatitic gneiss, and high P-T aluminosilicates (i.e. kyanite-sillimanite), attesting to their middle-upper amphibolite facies metamorphic conditions, with schists and paragneiss units interpreted as metamorphosed turbidites. When coupled with the presence of interlayered calc-silicates and orthoamphibolites with geochemical signatures (Fig. 1.7) ranging from mid-ocean ridge basalt (MORB) to within plate basalt (Tull et al., 2012), these interlayered amphibolites, metaturbidites, and calc-silicates strongly suggest that the Ashland Supergroup originated in a continental rift setting succeeded by deposition along the continental slope-rise, outboard of a continental margin (Tull, 1978; Thomas et al., 1980; Drummond et al., 1988; Allison, 1992; Tull et al., 2007, 2012, 2014; Barineau et al., 2015). Whole rock Nd model ages of 1055 and 964 Ma from metasedimentary units within the Ashland Supergroup suggest derivation of sediment from Grenvillian source rocks (Das, 2006). The structural position of Ashland Supergroup units above shelf rocks of the Talladage belt, across the Hollins Line fault (Figs. 1.1 and 1.2), indicates that these rocks originated as the basal rift and slope-rise prism units along the Neoproterozoic-lowest Paleozoic Laurentian margin.

**Figure 1.7.** Trace element tectonic discrimination diagrams for othoamphibolites from the Ashland-Wedowee-Emuckfaw belt of Alabama (eastern Blue Ridge). Adapted from Tull et al., 2012.
Rocks of the lower Ashland Supergroup would have formed coevally with deposition of Ocoee Supergroup rocks in the Tennessee embayment to the northeast, while upper Ashland Supergroup rocks may have been deposited coevally with units of the Kahatchee Mountain Group and equivalent Nantahala and Brasstown Formations in the Talladega belt-western Blue Ridge, as well as Chilhowee Group units in the foreland fold and thrust belt. Rocks equivalent to the Ashland Supergroup are not recognized in the eastern Blue Ridge of GA.

Structurally and stratigraphically above the Ashland Supergroup, the Wedowee Group (Fig. 1.6) consists of variably garnetiferous and graphitic two-mica schists interlayered with subordinate quartzite, biotite paragneiss, and amphibolite. These are interpreted as metagraywacke and uncommon orthoamphibolite (Neathery, 1975; Allison, 1992; Tull et al., 2014; Barineau et al., 2015). From an area near Hightower, AL, southwest towards Goodwater, AL, Wedowee Group rocks are separated from those of the underlying Ashland Supergroup by the Allatoona fault and potentially equivalent Goodwater-Enitachopco fault (see Tull et al., this guidebook). Southwest of Goodwater, AL, however, displacement on this fault ends at a tip point (Figs. 1.2 and 1.8), beyond which the lowermost Wedowee Group is in gradational contact with the underlying upper Ashland Supergroup, indicative of a stratigraphic boundary between the two units (Allison, 1992). The Wedowee Group was long considered to represent continued deposition of slope-rise turbiditic sediments above Neoproterozoic-lower Paleozoic slope-rise units of the upper Ashland Supergroup (Tull, 1978; Thomas et al., 1980; Stow et al., 1984; Drummond et al., 1988, 1994, 1997, Tull et al., 2007). However, U-Pb detrital zircon ages from the upper Wedowee Group and overlying Emuckfaw Group indicate a maximum deposition age of Early to Middle Ordovician for units stratigraphically above the Ashland Supergroup (Tull et al., 2012, 2014; Barineau et al., 2015).

Figure 1.8. Regional geologic map near Goodwater, Alabama, depicting the relationship between the Goodwater-Enitachopco fault, Ashland Supergroup, and Wedowee Group. Displacement along this faulted contact tips out at a point southwest of Goodwater, Alabama, beyond which it is a polydeformed stratigraphic boundary. Adapted from Barineau et al., 2017. See Figure 1.2 for unit symbology.
Rocks of the overlying Emuckfaw Group (Fig. 1.6), like the Wedowee Group, consist predominantly of variably garnetiferous and graphitic metapelites (i.e. two-mica schist) and subordinate quartzite, and biotite paragneiss. In contrast, metasedimentary units of the Emuckfaw Group are interlayered with thick and laterally continuous orthoamphibolite layers (e.g. Beaverdam amphibolite), which along with the intercalated metasedimentary rocks, are intruded by a number of metagranitoids ranging in composition from alkali granites to tonalities. U-Pb ages from zircons extracted from the two largest of these orthogneiss bodies (i.e. Zana Granite and Kowaliga Gneiss) yield ages ranging from 477 Ma to 455 Ma (Early to Middle Ordovician; Sagul, 2016; Sagul et al., 2017). Coupled with detrital zircon ages from the upper Wedowee and lower Emuckfaw Groups, these magmatic ages indicate that the >3 km of sedimentary and mafic protoliths of the Emuckfaw and upper Wedowee Groups were deposited/emplaced over the course of a few million years during the Early-Middle Ordovician.

Eastern Blue Ridge (GA)

In AL, the structural top of the Emuckfaw Group is truncated by the late kinematic Abanda fault along the northwestern boundary of the Brevard fault zone (Fig. 1.2). To the northeast, across the AL-GA state line, the Abanda fault cuts up section in its footwall and the outcrop width of the Emuckfaw Group significantly widens from <5 km near Roanoke, AL, to nearly 20 km southwest of Roopville, GA (Fig. 1.2). In addition, the lower bounding fault of the eastern Blue Ridge in eastern AL and GA – the Allatoona fault – truncates the Hollins Line and rocks of the Ashland Supergroup near Hightower, AL, such that equivalent units are not exposed in the eastern Blue Ridge of GA. The Allatoona fault continues to cut upsection into the Wedowee Group northeast of the AL-GA state line, such that the >20 km wide section of Wedowee Group rocks exposed in the vicinity of Wedowee, AL, shrinks to <4 km southwest of the Mulberry Rock Recess (see Tull et al., this guidebook).

In GA, rocks along strike of those mapped as Emuckfaw Group in AL were originally named the Heard Group by Bentley and Neathery (1970) for exposures of interbedded schist and paragneiss (metagraywacke) exposed in Heard County, GA. Originally divided into the Roopville, Glenloch, and Centralhatchee Formations, Neathery and Reynolds (1975) were unable to identify this tripartite stratigraphic sequence in AL, and formally reassigned these rocks to the Emuckfaw Formation (later upgraded to group status) in AL (named for exposures along Emuckfaw Creek in Tallapoosa County). In GA, subsequent workers (e.g., McConnell and Abrams, 1984; German, 1989) assigned units along strike of the Emuckfaw, immediately east of the AL-GA state line to the “western belt” of the Sandy Springs Group, correlating them with rocks of the Sandy Springs Group in the hanging wall (eastern side) of the Chattahoochee fault (Higgins and McConnell, 1978). Farther to the northeast, McConnell and Abrams (1984) assigned predominantly bimodal metavolcanic rocks interlayered with minor metapelites to the New Georgia Group, which stretches from Carroll County, east of the Alabama state line, northeast to Cobb and Paulding Counties, Georgia. German (1985, 1988, 1989) correlated New Georgia Group units with those of the Pumpkinvine Creek Formation, Canton Formation, and Helen Group in the vicinity and southwest of Dahlonega, GA. Age constraints and recent mapping
TACONIC BACKARC-ARC TERRANES IN THE SOUTHERN APPALACHIANS

(Holm-Denoma, 2006; Gilmer and Barineau, 2012; Tull et al., 2012, 2014; Smith and Tull, 2014; Barineau et al., 2015), however, indicate that rocks of the Emuckfaw Group and New Georgia Group are correlative, whereas rocks of the Pumpkinvne Creek Formation, Canton Formation and Helen Group are contained in a separate thrust sheet (i.e. Dahlonega gold belt; see Tull et al., this guidebook). Some workers (e.g. Merschat et al., 2010), have considered rocks of the Dahlonega gold belt to be part of the central Blue Ridge terrane in north Georgia, due to the belt’s intermediate position between thrust sheets mapped as western and eastern (western Tugaloo terrane of Hatcher et al., 2007) Blue Ridge in north Georgia and North Carolina. Because, however, rocks of the Dahlonega gold belt share a genetic origin with those of the Emuckfaw Group and correlative New Georgia Group, we consider them to be part of a composite eastern Blue Ridge terrane (Tull et al., 2014; Barineau et al., 2015).

Although German (1985, 1988, 1989) and McConnell and Abrams (1984) proposed a stratigraphic sequence for the New Georgia Group and surrounding units, we have subsumed many of those units into previously established, correlative stratigraphic units identified and named earlier in Alabama (Tull et al., 2012, 2014; Barineau et al., 2015), and reinterpret proposed structural relationships in this segment of the eastern Blue Ridge. In AL, variably garnetiferous graphitic muscovite-biotite mica schists of the Wedowee Group pass into non-graphitic (or graphite-poor) muscovite-biotite schist and orthoamphibolite of the Emuckfaw Group (Fig. 1.6) across a gradational contact. This contact can be readily mapped (Figs. 1.2, 1.8 and 1.9) from southwest of Alexander City, AL, along the southeastern flank of the Elkahatchee batholith, northeast to the AL-GA state line near Ephesus, GA. At this location, the contact steps northwest in a series of isoclinal folds and is eventually truncated by the Burnt Hickory Ridge fault on the western side of the Mulberry Rock recess (Figs. 1.2 and 1.9; Tull et al., this guidebook). In contrast, McConnell and Abrams (1984) and German (1989) interpreted the overall structure of the New Georgia Group and Dahlonega gold belt in Carroll and Paulding counties as an antiformal nappe with >35

Figure 1.9. Geologic map depicting the Wedowee-Emuckfaw stratigraphic contact crossing from Alabama into Georgia in the vicinity of Ephesus, Georgia. Adapted from Barineau et al., 2017. See Figure 1.2 for unit symbology.
km half-wavelength, overturned to the northwest. In their structural model, rocks of the New Georgia Group defined the core of the structure whereas rocks of the “western Sandy Springs Group” defined the northwestern and southeastern “limbs” of their structure. German (1989) traced stratigraphy of the “western Sandy Springs Group” to the GA-AL state line, inferring that they could be traced into the Ashland-Wedowee-Emuckfaw belt of AL. Rocks of the “western Sandy Springs Group” were initially proposed by McConnell and Abrams (1984) to be stratigraphic equivalents of the Sandy Springs Group (Higgins and McConnell, 1978) east of the Chattahoochee fault in the Chattahoochee thrust sheet (western Tugaloo terrane of Hatcher et al., 2007). Following the example of Holm-Denoma (2006), Tull et al. (2012, 2014), and Barineau et al. (2015), we suggest the term “western Sandy Springs Group” should be abandoned due to the difficulty of correlating non-unique stratigraphy across a major structural/metamorphic/tectonic boundary (Chattahoochee fault). In addition, because the overturned antiformal nappe structure interpreted for the eastern Blue Ridge of western GA (Carroll County gold belt of German, 1989) is not present in rocks of the Wedowee and Emuckfaw Groups immediately west of the GA-AL state line, it is difficult to adopt the structural and stratigraphic model proposed for the eastern Blue Ridge of GA with that of AL. It is possible, however, to reconcile the structural and stratigraphic constraints of the GA eastern Blue Ridge with those already established for the Ashland-Wedowee-Emuckfaw belt of eastern AL. Thus, we reassign rocks of the “western Sandy Springs Group” to those of the Wedowee and Emuckfaw Groups in eastern AL and correlate rocks of the New Georgia Group with those of the Emuckfaw Group (Holm-Denoma, 2006; Tull et al., 2014; Barineau et al., 2015).

In contrast to German’s (1989) structural interpretation, a number of authors (e.g., Hatcher et al., 2007; Merschat et al., 2010) have suggested that rocks of the New Georgia Group lie in a separate thrust sheet than those of the Emuckfaw and Wedowee Groups (or German’s “western Sandy Springs sequence”), interpreting the Ashland-Wedowee-Emuckfaw belt as a structurally higher thrust sheet than that containing the New Georgia Group. Detailed mapping of the Emuckfaw-Wedowee contact from eastern Alabama to northwestern GA (Fig. 1.2), however, indicates that stratigraphic sequences from the AL eastern Blue Ridge can be mapped across these proposed faults, although rocks of the Wedowee Group and correlative Emuckfaw-New Georgia Groups lie in a separate thrust sheet from those of the structurally lower Pumpkinvine Creek and Canton Formations of the Dahlonega gold belt. We consider all rock units in the Ashland-Wedowee-Emuckfaw belt, Dahlonega gold belt and Chattahoochee thrust sheet (western Tugaloo terrane of Hatcher et al., 2007), to be part of a composite eastern Blue Ridge.

**Dahlonega Gold Belt (GA, NC)**

The Dahlonega and Carroll County gold belts have been the subject of intense study due to the gold-bearing rocks they contain (Lindgren, 1906; Abrams and McConnell, 1984; Albino, 1988; German, 1985, 1988, 1989). Located in the Blue Ridge of northwest GA and western NC, the Dahlonega gold belt extends for more than 300 km northeast of the Mulberry Rock recess in the hanging wall of the Allatoona fault. Although rocks of the central Blue Ridge—framed by the Hayesville and Soque River faults—separate the western Blue Ridge and Dahlonega gold belts in
northeastern GA, a number of researchers (e.g., Nelson and Gillon, 1985; Bream et al., 2004; Hatcher et al., 2007; Merschat et al., 2010) interpret rocks of the central Blue Ridge as a structurally higher klippe above both the Dahlonega gold belt and western Blue Ridge. This structural interpretation suggests that the central Blue Ridge must palinspastically restore to a position southeast of the original location of the Dahlonega gold belt. In this interpretation, rocks of the Dahlonega gold belt extend beneath the central Blue Ridge and reemerge to the northeast in the Great Balsam Mountains window west of Asheville, NC. Beneath the central Blue Ridge thrust sheet (Cowrock and Cartoogechaye terranes of Hatcher, 2007), rocks of the Dahlonega gold belt are either in stratigraphic or faulted contact with rocks of the western Blue Ridge (Hatcher et al., 2007). Like those in the Ashland-Wedowee-Emuckfaw belt, we assign rocks of the Dahlonega gold belt to the eastern Blue Ridge because of their structural position above rocks of the western Blue Ridge in Georgia and North Carolina. Other workers (e.g., Merschat et al., 2010), however, assign rocks of the Dahlonega gold belt to the central Blue Ridge of north Georgia due to their intermediate position between thrust sheets mapped as western and eastern (western Tugaloo terrane of Hatcher et al., 2007) Blue Ridge in north GA and NC. The structural, and potentially stratigraphic base of the Dahlonega gold belt beneath the central Blue Ridge is not exposed, but the southeastern and structural top of the belt in GA and NC is the Chattahoochee and Holland Mountain faults, which separate it from the western Tugaloo terrane to the southeast (Hatcher et al., 2007; Merschat et al., 2010).

Structural separation of rocks in the Dahlonega gold belt beneath the central Blue Ridge klippe, as well as along-strike variation in volcanic versus sedimentary components, have led to a myriad stratigraphic names for rock units in this thrust sheet. From northeast to southwest, rocks in the Dahlonega gold belt include the Otto Formation in the Great Balsam Mountains window and immediately southeast of the central Blue Ridge klippe, the Helen Group and correlative Pumpkinvine Creek and Canton Formations in North Carolina and north-northwestern Georgia, respectively, and the Sally Free mafic complex west of Dahlonega, Georgia (Gillon, 1982; Nelson and Gillon, 1985; Hatcher, 1988; German, 1989; Settles, 2002). In the Carroll County gold belt, which lies in a separate thrust sheet structurally above the Dahlonega gold belt (see Tull et al., this guidebook), rocks of the eastern Blue Ridge in GA, immediately east of the Alabama state line, have been assigned to the “western belt of the Sandy Springs Group” by a number of workers (e.g., McConnell and Abrams, 1984; German, 1989). We follow the example of Holm-Denoma (2006) and Tull et al. (2012, 2014) and assign these rocks to the Wedowee and Emuckfaw Groups, abandoning the term “western Sandy Springs Group” for rocks between the Allatoona fault and Brevard fault zone in western central Georgia (see discussion above).

Otto Formation

The Otto Formation in the Great Balsam Mountains window and immediately southeast of the central Blue Ridge in NC, between the Chattahoochee–Holland Mountain and Shope River faults, is commonly divided into two units: a dominant metasandstone and subordinate interlayered schist and metasandstone (Lamb, 2001; Merschat, 2009). Otto Formation metasandstones are commonly micaceous, garnetiferous, and feldspathic; biotite-muscovite schists are commonly garnetiferous
and quartzose; and all units are locally sillimanite-bearing, sulfidic, and migmatitic. Schist-metasandstone sequences also contain subordinate micaceous, garnetiferous, feldspathic quartzose gneiss, and calc-silicate bodies, in addition to isolated lenses of amphibolite and altered ultramafic bodies (Merschat, 2009). Otto Formation detrital zircons yield abundant Mesoproterozoic (1.3–1.0 Ga) and Neoproterozoic (900–700 Ma) ages (Bream et al., 2004; Merschat et al., 2010), as well as a minor Archean–Paleoproterozoic population (2.7–1.6 Ga, Bream et al., 2004). Farther to the southwest, rocks of the Helen Group are considered correlatives of the Otto Formation (Hatcher, 1988).

**Helen Group**

The Helen Group in North Carolina and northern Georgia is typically divided into three subunits, from northwest to southeast: the Robertstown, Horton, and Nacoochee Formations (Nelson and Gillon, 1985). The southeastern and structurally lowest unit, the Nacoochee Formation, consists of thick amphibolites interlayered with variably graphitic mica schist, micaceous quartzites, and biotite feldspathic paragneiss. Rocks of the Nacoochee Formation grade into those of the structurally overlying Horton Formation, with Nacoochee graphitic schists and amphibolites grading into micaceous, feldspathic quartzites, and locally sulfidic metasiltstones, interlayered with subordinate amphibolite and aluminous mica schist of the Horton Formation. The structurally highest unit, the Robertstown Formation is conformable with the underlying Horton Formation, and is composed predominantly of muscovite-biotite feldspathic quartzites interlayered with micaceous schist, micaceous feldspathic paragneiss (metagraywacke?), and minor metaconglomerate and amphibolite (Gillon, 1982; Nelson and Gillon, 1985; German, 1985).

**Sally Free Mafic Complex**

Structurally below rocks mapped as the Helen Group (Nelson and Gillon, 1985) and New Georgia Group (German, 1985), west-northwest of Dahlonega, Georgia, rocks of the Sally Free mafic complex make up the structurally lowest stratigraphy of the Dahlonega gold belt (Settles, 2002), or alternatively lie in a separate thrust sheet (German, 1985). German (1985) mapped the Allatoona fault at the tectonic base of the Dahlonega gold belt along a generally northeast trace into northeast Lumpkin County (plate 1 in German, 1985), northeast of the city of Dahlonega, where he interpreted it as truncated along the Shope Fork fault. Settles (2002) mapped felsic-intermediate orthogneiss bodies, orthoamphibolites, and metagabbros as part of a large (>34 km²) bimodal metaigneous complex in the region between German’s (1985) Allatoona and Shope Fork faults, naming it the Sally Free mafic complex. Based on lithologic, geochemical, and metamorphic similarities between rocks of the western Blue Ridge, central Blue Ridge, and Dahlonega gold belt, Settles (2002) interpreted the Sally Free mafic complex to lie, at least in part, in contact with the Otto Formation (Helen Group of Nelson and Gillon, 1985; New Georgia Group of German, 1985). Significantly, the Cane Creek felsic gneiss, interpreted as a metamorphosed shallow intrusive or volcanic sequence within the Sally Free mafic complex, yielded a U-Pb age of 482 ± 7 Ma (Settles, 2002; Bream, 2003), suggesting an Early Ordovician age for this metaigneous complex. In the interpretation of Settles (2002), this would place
the Lower Ordovician Sally Free mafic complex structurally and stratigraphically below rocks mapped as the Middle Ordovician Pumpkinvine Creek Formation (466.1 ± 5.2 Ma and 461.8 ± 3.3 Ma, Holm-Denoma and Das, 2010; 463 ± 3 Ma and 466 ± 5 Ma, Thomas, 2001), and make it the oldest known igneous unit in the Dahlonega gold belt (Tull et al., 2014; Barineau et al., 2015). In the alternative interpretation of German (1985), the Sally Free mafic complex would belong within the central Blue Ridge thrust sheet.

**Inner Piedmont**

The Dadeville Complex and Opelika Group of the western Inner Piedmont (eastern Tugaloo terrane of Hatcher et al., 2007) have structural characteristics reminiscent of those between the central Blue Ridge and Dahlonega gold belt farther to the northeast. Immediately southeast of the eastern Blue Ridge, rocks of the Dadeville Complex were emplaced as a thrust sheet in the hanging wall of the Stonewall Line/Katy Creek fault system (Fig. 1.2). Subsequently, these rocks were deformed into a >4,300 km², ~5 km thick synformal structure (Tallassee synform) and erosionally breached to form a klippe atop rocks of the underlying Opelika Group. Like the central Blue Ridge klippe, which must be palinspastically restored southeast of the Dahlonega gold belt, rocks of the Dadeville Complex must root southeast of the Opelika Group (see Farris et al., this guidebook). Rocks of both the Opelika Group and overlying Dadeville Complex have lithologic, geochronologic, and geochemical characteristics that suggest a genetic link with a suprasubduction system that formed on the seaward margin of the Laurentian plate, beyond the continental hinge zone.

**Opelika Group**

The Opelika Group (Bentley and Neathery, 1970; Sears et al., 1981), southeast of and structurally below the Dadeville Complex, includes a sequence of rocks lying between the Stonewall Line shear zone on the northwest and Towaliga fault on the southeast. Divided into two stratigraphic packages, rocks of the Opelika Group include the Auburn Formation at the structural base, overlain by rocks of the Loachapoka Formation (Bentley and Neathery, 1970; Sears et al., 1981). The Auburn Formation consists of migmatitic, variably garnetiferous biotite gneiss interlayered with kyanite-staurolite-bearing, variably garnetiferous two-mica schist and rare calc-silicate pods (Bentley and Neathery, 1970; Goldberg and Steltenpohl, 1990). These units have been interpreted as metamorphosed, interlayered greywacke and pelitic strata (Bentley and Neathery, 1970), or alternatively as metasomatized paragneiss associated with intrusion of the Farmville Metagranite (Colberg and Chalokwu, 1990). The structurally higher Loachapoka Formation consists of a thick package (>1 km) of interlayered kyanite and sillimanite, variably graphitic and garnetiferous schists, interlayered with quartzite and minor amphibolite (Bentley and Neathery, 1970; Goldberg and Steltenpohl, 1990). Discontinuous lenses of locally conglomeratic, sillimanite-bearing, garnetiferous, micaceous, feldspathic quartzite as much as 80 m thick (Grimes et al., 1993) are mapped as the Saugahatchee quartzite or Saugahatchee quartzite member. These readily mappable quartzite layers strike for >180 km into the central Georgia Piedmont (Bentley and Neathery, 1970; Sears et al., 1981; Goldberg and Steltenpohl, 1990). Both the Auburn and Loachapoka Formations...
are intruded by the Farmville Metagranite (Bottle Granite of Bentley and Neathery, 1970), a garnetiferous, muscovite, biotite, feldspathic orthogneiss interpreted as a peraluminous, S-type granite formed during crustal anatexis (Goldberg and Steltenpohl, 1990). Steltenpohl et al. (2005) reported SHRIMP U/Pb zircon analyses with complex U-Pb systematics from of a sample of Farmville orthogneiss. The eight youngest grains plot along concordia, with $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 477 ± 20 Ma to 425 ± 18 Ma, but also include inherited grains with ages of 1476 ± 34 Ma ($^{207}\text{Pb}/^{206}\text{Pb}$) and 564 ± 24 Ma ($^{206}\text{Pb}/^{238}\text{U}$). Hawkins et al. (2013) reported 36 additional zircon analyses of the Farmville orthogneiss, assigning it a crystallization age of 440 ± 7 Ma ($^{206}\text{Pb}/^{238}\text{Pb}$). The complex age systematics, along with a whole-rock Nd depleted-mantle model age of 1.12 Ga suggest a magmatic source for the Farmville consisting, at least in part, of Grenville crustal components.

**Dadeville Complex**

The Dadeville Complex is a thick (~15 km of lithostratigraphy), extensive metagneous complex with subordinate metasedimentary rocks located in the core of the map-scale Tallassee synform. The complex is separated from the Jacksons Gap Group of the Brevard fault zone by the Katy Creek fault on its northwestern flank, whereas on its southeastern flank it is separated from the structurally lower Opelika Complex by the Stonewall Line (Bentley and Neathery, 1970). The structurally lowest and spatially most extensive unit (~70% of the complex’s nonintrusive rocks with a structural thickness >9 km) is the Ropes Creek Amphibolite and its northwestern equivalent, the Waresville Formation and/or Amphibolite (Bentley and Neathery, 1970; Stow et al., 1984). The bulk of the Ropes Creek–Waresville unit consists of metamorphosed mafic to intermediate tuffs (Bentley and Neathery, 1970; Sears et al., 1981) and tholeiitic basalts (Neilson and Stow, 1986). The basement to these units has never been identified, but these metamorphic rocks are intercalated with andesitic and/or dacitic gneisses (Waverly Gneiss), minor pelitic metasedimentary rocks, metagraywacke (metaturbidite?), spessartine garnet and/or magnetite quartzite (metachert?), and calc-silicate gneiss (Clark, 1973; Brown and Cook, 1981; Sears et al., 1981). Locally, the unit contains rocks interpreted as mafic and ultramafic olistostromes, potentially representing an ophiolitic mélangé (Neathery, 1968; Sears et al., 1981). In the core of the Tallassee synform, the Ropes Creek Formation is structurally overlain by the uppermost unit of the Dadeville Complex, the Agricola Schist (Sears et al., 1981), a >1-km-thick sequence of interbedded pelitic schist, thin metagraywacke, and amphibolite. Intruding Ropes Creek and Agricola strata, several suites of felsic (metatonalite and metagranite) and mafic/ultramafic (metanorite, metagabbro, and metaorthopyroxenite) plutonic rocks make up >50% of the Dadeville Complex (Bentley and Neathery, 1970; Sears et al., 1981; Stow et al., 1984; Neilson and Stow, 1986; Spell and Norrell, 1990; Steltenpohl et al., 1990). Neilson and Stow (1986) identified both mafic and ultramafic plutonic rocks of the Dadeville Complex, metanorite and metaorthopyroxenite cumulates of the Doss Mountain suite and metagabbro of the Slaughters suite, which they interpreted as forming within a volcanic arc based on their collective geochemical characteristics (Stow et al., 1984; Neilson and Stow, 1986). Tholeiitic E-MORB geochemical signatures for the thick pile of metabasalt and associated lesser intermediate metavolcanic rocks suggests
formation within either an oceanic or back-arc rift, sediment-starved basin (Stow et al., 1984) with very little sediment input. Spell and Norrell (1990) report detailed geochemistry for rocks they identified as belonging to the Ropes Creek Amphibolite and/or Waresville Formation. However, many of the samples they analyzed were sampled from amphibolites in different thrust sheets that were erroneously correlated with those of the Ropes Creek Amphibolite in the Dadeville Complex. Their commixture of amphibolites originating in different lithotectonic regimes, and potentially with different ages of formation, calls into question some of their conclusions. Recent trace element geochemistry from rocks of the Dadeville Complex indicate these units have arc-like characteristics, including large negative Nb-Ta anomalies and large ion lithophile element (LILE’s) enrichment. Zircons from felsic plutons intruding the more dominant amphibolite lithologies yield Taconic U-Pb ages of 461 ± 4 Ma to 447+/-3 Ma. Zircons from the Agricola Schist and Ropes Creek Amphibolite yield a significant fraction of Grenville and Granite-Rhyolite Province ages, but also grains as young as 440 Ma (see Farris et al., this guidebook).

TECTORIC SYNTHESIS

Because of the potential for translation of large thrust sheets over significant distances, Appalachian geologists have historically struggled with interpretations of genetic origin for rocks in different lithotectonic regimes. In recent decades, fossil discoveries, improvements in age resolution via radiometric dating of single mineral grains, and tectonic discrimination of rocks using trace element geochemistry, have provided valuable insight into interpretations of origin for rocks along the entire orogen. In the southern Appalachians, this work has revealed the presence of an Ordovician back-arc (Wedowee-Emuckfaw-Dahlonega back-arc basin of Tull et al., 2014) and corresponding arc (Dadeville Complex, this guidebook) built on the seaward margin of the Laurentian plate. Here, we present a model for the genetic origin of latest Neoproterozoic and lower Paleozoic rocks in the western Blue Ridge, eastern Blue Ridge, and western Inner Piedmont based on a synthesis of ours and others' work in the region.

Neoproterozoic to earliest Cambrian Rifting

Following the Grenville orogeny and Mesoproterozoic assembly of the supercontinent Rodinia, rift-related volcanic rocks from the Canadian Maritimes in the northern Appalachians south to Tennessee indicate two phases of continental rifting between Laurentia and its Gondwana conjugate margin. The first, an aborted rift event between 760 and 680 Ma (Su et al., 1994; Aleinikoff et al., 1995; Fetter and Goldberg, 1995; Ownby et al. 2004; Burton and Southworth, 2010), was followed by initial phases (615-550 Ma) of the successful rifting event that led to development of the Iapetus Ocean (Aleinikoff et al., 1995; Cawood et al., 2001; Southworth et al., 2009; O’Brien and van der Pluijm, 2012). Asymmetric rifting of Laurentia during the breakup of Rodinia produced a series of promontories and embayments in the Iapetus-facing southern (eastern in modern coordinates) margin of the continent (Thomas, 1976, 1977, 1991) due to polarity reversals of lithospheric-scale detachment faults (Fig. 1.5). In the southern Appalachians, this resulted in significant differences in thickness of rift-related sediment deposited across the margin.
of the AL promontory and adjacent TN embayment due to the underlying upper plate-
lower plate architecture (Lister et al., 1986; Thomas, 1991). Rift sedimentation within
the TN embayment (i.e. Ocoee Supergroup) reached thicknesses of as much as 15
km, while equivalent units deposited seaward of the AL promontory (i.e. lower Ashland
Supergroup) probably reached thicknesses of no more than 5-7 km.

In the southern Appalachians, variably metamorphosed, interbedded
conglomerates, sandstones, siltstones, and pelitic rocks of the Ocoee Supergroup in
the western Blue Ridge (GA, TN), and middle-upper amphibolite-facies quartzite,
paragneiss, schist, and orthoamphibolite of the Ashland Supergroup in the eastern
Blue Ridge (AL) are interpreted to have formed in these evolving rift basins along the
TN embayment and AL promontory, respectively. Despite its allochthonous nature, in
the western Blue Ridge, rocks of the Ocoee Supergroup are stratigraphically overlain
by those of the Laurentian affinity, Cambrian-aged Chilhowee Group, attesting to their
deposition in rift basins along the Neoproterozoic to earliest Cambrian margin of
Laurentia. In the eastern Blue Ridge of AL, however, the palinspastic location of rocks
of the Ashland Supergroup are more equivocal. We argue, however, that the following
characteristics strongly suggest the lower Ashland Supergroup (i.e. Higgins Ferry–
Poe Bridge Mountain Groups) formed in a rift setting along the Laurentian margin.
First, rocks of the lower Ashland Supergroup consist predominantly of
metamorphosed pelitic units (e.g. garnet biotite muscovite schist) interlayered with
subordinate greywacke (fine-grained biotite paragneiss and quartzofeldspathic
schist), calc-silicate and mafic flows and/or sills (orthoamphibolite). Alternating layers
of metapelite and metagraywacke are suggestive of turbidite deposits, while
interlayered calc-silicate and orthoamphibolite are consistent with eruption of mafic
lava in a submarine setting. Secondly, tectonic discrimination of orthoamphibolite (e.g.
Mitchell Dam amphibolite) using trace element geochemistry (Tull et al., 2012), places
these rocks in MORB to within-plate basalt fields on tectonic discrimination diagrams
(Fig. 1.7), including the high field-strength element tectonic discrimination diagrams
(e.g. Ti-V tectonic discrimination diagram of Shervais (1982) and Zr/Y-Y basalt
discrimination diagram of Pearce and Norry (1979). Third, metasedimentary units
within the Ashland Supergroup yield whole-rock Nd model ages of 1055 and 964 Ma,
suggesting derivation of sediment from a Grenvillian source (Das, 2006). Fourth, rocks
of the Ashland Supergroup lie structurally and stratigraphically below units of the
Lower Ordovician Wedowee Group, providing an upper depositional age for these
interlayered metasedimentary and metavolcanic units. They have not, however,
experienced the granulite facies metamorphism typical of Grenville-age basement
units (e.g. Corbin gneiss) in the region, suggesting they are post-Grenville in age.
Finally, rocks of the Ashland Supergroup lie in the hanging wall of the Hollins Line fault
system (Fig. 1.2), placing them structurally directly atop Laurentian shelf units of the
Talladega belt (i.e. Kahatchee Mountain, Sylacauga Marble, and Talladega Groups).
Thus, lithologic, geochemical and geochronologic constraints on rocks of the lower
Ashland Supergroup indicate they formed in a post-Grenville to Lower Ordovician
submarine basin simultaneously receiving coarse- to fine-grained siliciclastics via
turbidity currents from Grenville-aged source rocks, as well as mafic lava flows from
nearby eruptive centers. Additionally, rocks of the Ashland Supergroup must have
lain in a palinspastic position which would allow them to be thrust directly atop
Paleozoic shelf units of the Talladega belt. Because there is no evidence that the bounding fault (i.e. Hollins Line roof thrust) between these two thrust sheets (i.e. Talladega belt and AL eastern Blue Ridge) marks a suture between terranes on opposing plates, it is difficult to model rocks of the Ashland Supergroup as originating from the rifted margin of a continent other than Laurentia. Collectively, these data are consistent with deposition of the lower Ashland Supergroup on the rifted margin of Laurentia during the breakup of Rodinia in the latest Neoproterozoic, with Alleghanian thrusting emplacing these rocks atop the Laurentian shelf (Drummond et al., 1988; Allison, 1992; Tull et al., 2012, 2014; Barineau et al., 2015). It is possible, however, to interpret the base of the Ashland Supergroup as latest Cambrian to earliest Ordovician in age, in which case it would represent the initial phases of back-arc rifting in the Wedowee-Emuckfaw-Dahlonega back-arc basin (WEDB of Tull et al., 2014). An unequivocal interpretation awaits geochronological analyses of rocks from this portion of the stratigraphic sequence.

**Cambrian to Early Ordovician Shelf**

Lower Paleozoic rocks of the southern Appalachian foreland and Talladega belt-western Blue Ridge (Fig. 1.10) record stabilization of the rifted Laurentian margin and deposition of shallow water siliciclastics (Chilhowee Group, Kahatchee Mountain Group) during a cratonward transgression (Sauk sequence of Sloss, 1963). The precise age of the rift-to-drift transition in this segment of Laurentia is difficult to establish. Trilobite trace fossils reported in the upper Unicoi Formation (basal unit of the Chilhowee Group) place the beginning of passive margin sedimentation in the “Early” Cambrian (Tommotian or younger; Simpson and Eriksson, 1989). Revisions to the Cambrian time scale (Babcock and Peng, 2007; Gradstein et al., 2004; Cohen et al., 2013), however, make correlations between the original tripartite Cambrian subdivisions (i.e. Early, Middle, Late) and more recent series/epochs (e.g. Terreneuvian, Furongian) and stages/ages (e.g. Fortunian, Drumian) subdivisions challenging. Additionally, newly proposed revisions to the Cambrian time scale place the Tommotian within the Terreneuvian series – before the first appearance of trilobites (Ogg et al., 2016). Recent paleontological work on the Chilhowee places the Neoproterozoic-Cambrian boundary in the upper Unicoi Formation, suggesting the rift-drift transition developed in the latest Neoproterozoic to earliest Cambrian (Hageman and Miller, 2014). If so, movement on basement faults (Thomas and Astini, 1999) on the shallow Laurentian shelf (e.g. Birmingham graben) continued well into the Cambrian (Cambrian series/epoch 2 of Cohen et al., 2013) after a passive margin had been established. Regardless of the exact timing of the rift-drift transition, by Cambrian series/epoch 2 (ca. 515 Ma), as Laurentia drifted into tropical waters south of the lower Paleozoic equator, an extensive carbonate platform developed along the extent of the iapetus-facing margin, marked by deposition of shallow water carbonates preserved in the foreland (Shady Dolomite through Knox Group), the Talladega belt (Sylacauga Marble Group) and western Blue Ridge (Murphy Marble) of the southern Appalachians. Carbonate deposition, punctuated by the influx of intermittent siliciclastics from the continental margin, continued from early Cambrian through the Early Ordovician on the distal continental shelf.
Cambrian to Early Ordovician Slope-Rise

Assuming that rocks of the lower Ashland Supergroup represent Neoproterozoic to earliest Cambrian sedimentary deposits and mafic volcanic rocks on attenuated continental crust beyond the distal Talladega belt shelf and continental hinge zone (Fig. 1.10), stratigraphy of the upper Ashland Supergroup (i.e. Hatchett Creek-Mad Indian Groups) ideally represents slope-rise deposition following stabilization of the Laurentian margin. Dominated by schist (metapelite) and paragneiss (metawacke), units of the upper Ashland Supergroup are reasonably interpreted as metaturbidites. Detrital zircon from the overlying upper Wedowee Group (Tull et al., 2012, 2014; Barineau et al., 2015) place a maximum depositional age for the upper Ashland Supergroup and lower Wedowee Group of Early Ordovician or older, consistent with interpretation of the upper Ashland Supergroup and lower Wedowee(?) as slope-rise sediment deposited on a Cambrian-Lower Ordovician continental margin outboard of the continental shelf. By the earliest Ordovician (ca. 480 Ma), however, rocks of the Wedowee Group in the eastern Blue Ridge (AL), and Sally Free mafic complex and Pumpkinvine Creek Formation in the Dahlonega gold belt (GA-NC) record volcanic activity associated with the Wedowee-Emuckfaw-Dahlonega back-arc basin (see below). Because stratigraphy of the Taconic foreland, Talladega belt, and western Blue Ridge contains no evidence for Early Ordovician plutonism or effusive volcanism in coeval rocks of the shelf, the locus of back-arc rifting must have been beyond the continental hinge zone, on attenuated continental crust or transitional crust at the continent-ocean boundary. The palinspastic distance between rocks of the Laurentian shelf and slope-rise explains the seemingly contradictory nature of a passive margin shelf and active margin slope-rise during the Early Ordovician, and indicates that igneous activity occurring beyond the continental hinge zone did not affect carbonate sedimentation cratonward of the developing back-arc basin.

Early to Middle Ordovician Back-Arc

Rocks of the Talladega belt (AL), eastern Blue Ridge (AL, GA), Dahlonega gold belt (GA, NC), Opelika Complex (AL, GA) and Dadeville Complex (AL, GA) all contain interlayered, metamorphosed Ordovician volcanic and siliciclastic units. Many of these sequences are, additionally, intruded by Ordovician silicic plutons. Recent application of modern, single-grain radiometric dating techniques to these sequences (e.g. Thomas, 2001; Settles, 2002; Bream, 2003; Tull et al., 2007; McClellan et al., 2007; Holm-Denoma, 2006; Holm-Denoma and Das, 2010; Ingram, 2012; Tull et al., 2014; Barineau et al., 2015; Sagul, 2016) yield U-Pb zircon ages spanning the entire Ordovician (ca. 480-443 Ma). Although many of these lithotectonic terranes have traditionally been interpreted as portions of an exotic or peri-Laurentian arc obducted onto the Laurentian margin during the Taconic orogeny (Shanmugam and Lash, 1982; Hatcher, 1987, 2005; Higgins et al., 1988; Drake et al., 1989; Hatcher et al., 2007; McClellan et al., 2007), a number of complicating factors make this collisional orogenic model untenable for rocks of the lower Paleozoic, southern Appalachian margin of Laurentia.

In the southern Appalachians, the geometry of assembled lithotectonic terranes and constraints on timing contrasts with many aspects of collisional orogenic models.
Figure 1.10. Reconstruction of the late Cambrian Laurentian margin in the vicinity of the AL promontory and TN embayment. Upper plate-lower plate geometries adapted from Sutra et al., 2013.
The Ordovician history of the southern Appalachians includes a number of critical elements that contrast sharply with well-studied Cenozoic collisional orogens, including the subducting Australian margin beneath Timor and western Papua New Guinea (Harris et al., 1998; Audley-Charles, 2004; Cloos, 2005; Harris, 2006; Keep and Haig, 2010) and the subducting Asian margin beneath Taiwan (Huang et al., 2006). First and foremost, in those collisional orogens (A-type subduction) the obducting arc is separated from the foreland by a forearc and accretionary prism – which is uplifted during initial subduction of the continental margin. In the southern Appalachians, however, back-arc terranes were built directly on or structurally emplaced atop rocks of the Laurentian shelf-slope-rise with no evidence of an intervening forearc or accretionary prism (Tull et al., 2007, 2014). Additionally, the southern Appalachians lack Taconic-age thrust faults/allochthons, in significant contrast to the earliest phase of Taconic orogenesis in the northern Appalachians (Chapple, 1973; Robinson and Hall, 1980; Rowley and Kidd, 1981; Stanley and Ratcliffe, 1983, 1985; Waldron and van Staal, 2001; van Staal et al., 2007). Critically, rocks of these southern Appalachian terranes suggest the presence of significant extensional basins sourcing sediment from both the Laurentian margin and an adjacent suprasubduction system, while at the same time receiving significant input of volcanic material.

The collective lithologic, structural and geochemical aspects of rocks in the Talladega belt (AL), eastern Blue Ridge (AL, GA), Dahlonega gold belt (GA, NC) and Opelika Complex (AL, GA) suggest they formed in an Early-Middle Ordovician marginal (back-arc) Laurentian basin (Fig. 1.11) above thinned continental, transitional and/or oceanic crust outboard of the continental margin hinge zone (Tull et al., 2007, 2014; Holm-Denoma, 2006; Barineau, 2009; Holm-Denoma and Das, 2010; Barineau et al., 2015). Geochronological analyses of detrital zircon from these interlayered metasedimentary-metavolcanic sequences indicate both a Grenville basement source as well as one or more proximal and coeval Ordovician volcanic/plutonic bodies (Tull et al., 2012, 2014). Field evidence and major/trace-element chemistry of mafic and felsic metavolcanic units indicate these rocks originated in a suprasubduction setting – specifically a back-arc (Tull and Stow, 1980; McConnell, 1980; McConnell and Abrams, 1984, 1986; Gillon 1989; German, 1989; Hopson, 1989; Spell and Norrell, 1990; Durham, 1993; Davis, 1993; Yanagihara, 1994; Tull et al., 1998; Thomas, 2001; Settles, 2002; Bream, 2003; Kalbas, 2003; Holm-Denoma, 2006; Holm-Denoma and Das, 2010; McClellan et al., 2007; Tull et al., 2007, 2012, 2014; Barineau, 2009; Tull and Barineau, 2012; Barineau et al., 2015). Volcanic units within the Wedowee-Emuckfaw-Dahlonega (WEDB) marginal basin display a number of characteristics consistent with development in back-arc setting (Tull et al., 2014; Barineau et al., 2015).

First, primitive WEDB metabasalts have compositions intermediate between normal mid-ocean ridge basalt (N-MORB) and volcanic arc basalt, which suggests thinned crust in an extensional (riifting) suprasubduction setting. The degree of crustal thinning in the WEDB is difficult to estimate, but must have been substantial enough
Figure 1.11. Reconstruction of the Ordovician Wedowee-Emuckfaw-Dahlonega back-arc basin and Dadeville Complex arc (ca. 480-440 Ma), depicting major stratigraphic units forming in this time interval.
to accommodate a basin up to 10 km thick and >115 km wide (Tull et al., 2014). The WEDB does not appear to have evolved beyond a back-arc basin into one producing oceanic crust at a true mid-ocean ridge. Since spreading in back-arc basins is commonly temporally episodic (Taylor and Karner, 1983) and partitioned into numerous sub-basins (Yoon et al., 2014), the fact that multiple units in the overall WEDB stratigraphic section contain metavolcanic rocks may suggest that multiple episodes of extension (480-460 Ma) resulted in back-arc magmatism. Second, bimodal volcanism with suprasubduction signatures, which some workers have suggested accompanies early stages of back-arc rifting (Marsaglia, 1995), is consistent with metavolcanic sequences of the Talladega belt, eastern Blue Ridge, and Dahlonega gold belt that are dominated by stratified greenstones or amphibolites and subordinate metarhyolite/metadacite units. Third, metalliferous hydrothermal deposits (i.e. Cu-Zn dominated base-metal deposits) and chemical sediment (banded-iron formation) within the WEDB are consistent with similar deposits recognized in failed or intra-crustal rifts and extensional back-arc regions of volcanic arcs (Ohmoto and Skinner, 1983). Fourth, Ordovician subvolcanic felsic plutons (e.g., Zana Granite, Kowaliga Gneiss, Farmville Granite, Villa Rica Gneiss) intruding rocks of the WEDB are geochemically and petrologically similar to those found in back-arc units of the Lachlan orogen in southeastern Australia (Gray and Foster, 2004). Because the volcanic components of Ordovician WEDB rocks make up <20% of the total stratigraphy, and numerous detrital zircon studies (Hatcher et al., 2007; Holm-Denoma and Das, 2010; Merschat, et al., 2010; Tull et al., 2012, 2014) have failed to yield evidence for significant input of arc detritus (e.g. rare Ordovician zircons), it is difficult to interpret these units as forming within an arc sensu stricto or forearc setting. The observation that underlying, late Neoproterozoic-earliest Ordovician (?) eastern Blue Ridge strata (e.g., the Ashland Supergroup in Alabama) and strata of the WEDB do not appear to have been intruded by significant subduction-generated magmas is also inconsistent with an arc setting for WEDB units. Finally, structural constraints on emplacement of the Hillabee Greenstone, at the structural top of the Talladega Group, makes it difficult to envision an exotic origin for this unit. The ca. 470 Ma Hillabee Greenstone is the structurally highest unit of the Talladega belt, and is in tectonic contact along the pre-metamorphic Hillabee thrust with the late Early Devonian to earliest Mississippian (?) Jemison Chert/Erin Slate. Age constraints on these sequences suggest emplacement of the Hillabee along a pre-metamorphic fault during the latest Devonian or earliest Mississippian (Tull et al., 2007; McClellan et al., 2007; Barineau, 2009; Tull and Barineau, 2012). Little or no deformation of hanging or footwall units could have affected the Hillabee prior to or during thrust emplacement, which post-dates its formation by >90 m.y. The minimal deformation associated with emplacement of the Hillabee suggests a palinspastic position near the Laurentian shelf (Tull et al., 2007; Barineau, 2009; Tull and Barineau, 2012). In addition, to structural constraints, isotopic signatures (e.g., εHf) of zircons from Hillabee metadacite units suggest these magmas incorporated Grenville-age continental crust during their genesis (Tull et al., 2007), which contrasts sharply with the arc-continent collisional model proposed by some workers to explain the Hillabee’s tectonic position atop rocks of the Laurentian shelf (e.g., McClellan et al., 2007). In this arc obduction model of emplacement, back-arc sequences of the Hillabee would
need to be thrust over the arc, fore-arc, and accretionary prism components as it was translated from an opposing plate. To accomplish this without significantly deforming the Hillabee prior to its final tectonic emplacement on the Laurentian plate is unlikely (Tull et al., 2007; Barineau, 2009; Tull and Barineau, 2012; Tull et al., 2014; Barineau et al., 2015).

Lateral distribution of metavolcanic sequences in the Wedowee, Emuckfaw, New Georgia, and Helen Groups is highly variable, suggesting kilometer-scale, along-strike heterogeneities that are typical of back-arc basins (Gamble and Wright, 1995; Gray and Foster, 2004; Yoon et al., 2014). An abundance of Grenville-age detrital zircons, typical of Laurentian crustal materials, indicates a nearby cratonic source instead of a juvenile exotic arc source. Although most WEDB rocks are structurally emplaced at their base, thus complicating our understanding of the basement on which they formed, the common presence of Grenvillian, mid-continent (Granite/Rhyolite), Yavapai/Mazatzal (Central Plains), and trans-Hudson/Penokean detrital zircons within the basin’s metasedimentary rocks suggests the presence of rifted Laurentian basement and/or cover beneath the basin and/or along its flanks. Isotopic evidence for involvement of older crust (most likely Mesoproterozoic) is also found in rocks of the Hillabee Greenstone and Pumpkinvine Creek Formation metadacites (Tull et al., 2007; Holm-Denoma and Das, 2010).

Although inconsistent with typical collisional orogenic models and modern collisional orogenic belts, the geologic history of the southern Appalachians is consistent with the development of extensional and contractional phases of an accretionary orogen similar to those associated with opening of the Sea of Japan (East Sea) during the Cenozoic (Yoon et al., 2014) and the Lachlan orogen during the Paleozoic (Gray and Foster, 2004). Orogenic elements that develop on the overriding plate during B-type subduction of oceanic lithosphere (accretionary orogenesis) include a number of commonalities. First, rocks identified as fragments of oceanic crust commonly have back-arc basin geochemical signatures ranging from N-MORB to volcanic arc basalts (e.g., Lachlan orogen, Gray and Foster, 2004). Second, accretionary orogens commonly contain extensive turbidite deposits that are generally devoid of arc-derived lithic fragments (Cawood et al., 2009). Third, emplacement of back-arc basement and cover sequences atop adjacent pre-rift terranes is typical of contractional orogenic phases (e.g., New Caledonia, Cluzel et al., 2001). Fourth, intrusion by syn- and post-tectonic granitoids into back-arc basement and sedimentary cover is typical of some accretionary orogens (e.g., Lachlan orogen, Gray and Foster, 2004; Yukon-Tanana terrane, Piercey et al., 2001). Finally, the assembly of terranes within an accretionary orogen along a continental margin generally consists of back-arc rocks structurally emplaced between rocks of the adjacent continent and an outboard arc-forearc-accretionary prism—although intrabasinal subduction of back-arc oceanic crust can complicate this geometry (Cluzel et al., 2001; Gray and Foster, 2004; Cawood et al., 2009).

**Middle-Late Ordovician Shelf**

During the Middle Ordovician (Darriwilian stage, Cohen et al., 2013), the Cambrian–Ordovician carbonate platform experienced uplift and erosion (post–Knox unconformity), and seaward deposition of deep-water black shales (D. murchisoni
graptolite zone; Finney et al., 1996; Bayona and Thomas, 2006) along the Alabama promontory. Subsequent sedimentation progressed cratonward and northeast toward the Tennessee embayment through the early Late Ordovician (Bayona and Thomas, 2003, 2006). This advancing clastic wedge, the Blount basin, consists of a >2.5-km-thick sequence of mud, silt, and sand in its thickest preserved parts, but also includes volcanic ash sequences (k-bentonites) from silicic eruptive centers. The oldest of these, near the base of the sequence (Shanumugam and Walker, 1980; Kolata et al., 1996; Bayona and Thomas, 2006), is Darriwillian in age (Fig. 1.3), similar to the youngest WEDB bimodal metavolcanics (e.g., Pumpkinvine Creek Formation). Two of these foreland k-bentonites, the 448 ± 2 Ma Millbrig and 449 ± 2.3 Ma Deicke (Min et al., 2001), thicken toward a palinspastic location between Alabama and Pennsylvania (Kolata et al., 1998), and may have been sourced from volcanic centers located in the WEDB or Dadeville Complex. Isotopic evidence for the Deicke and Millbrig k-bentonites suggests they formed in a volcanic arc influenced by continental crust (Huff et al., 1992; Coakley and Gurnis, 1995; Kolata et al., 1996, 1998; Haynes et al., 2011; Samson et al., 1989). Apart from these k-bentonites, rocks of the Blount basin consist exclusively of Laurentian margin, shallow water sedimentary rocks that must have been built atop Grenville basement (Kellberg and Grant, 1956; Cressler, 1970; Bayona and Thomas, 2003). Blount basin sedimentary rocks lack evidence for detrital contributions from volcanic or deep-water sedimentary sources (Mack, 1985), or Taconic-age detrital zircons (Merschat et al., 2010). Modeled as a classic foreland basin, certain characteristics of the Blount basin contrast sharply with both modern and ancient foreland basins, especially its lack of coeval (i.e. Ordovician) faults and allochthons that would have not only generated the lithostatic load necessary for subsidence, but would have also provided the sedimentary detritus for the developing clastic wedge. These and other inconsistencies indicate the Blount basin may be better modeled as a retro-arc basin on the continental margin of a Laurentian plate suprasubduction system rather than a traditional foreland basin cratonward of a collisional orogenic belt and advancing thrust sheets (Tull and Barineau, 2014; Barineau et al., 2015; Tull et al., 2017).

Middle-Late Ordovician Arc

The Dadeville complex in the western Inner Piedmont of AL and GA consists of >15 km of predominantly metagneous rocks, unlike the sediment-dominated units of the WEDB, with lithologic and geochemical characteristics indicative of their formation in an arc setting (see Farris et al., this guidebook). Mafic and felsic igneous rocks in the Dadeville have geochemical characteristics typical of arc magmas, including enrichment in large ion lithophile elements (LILE), large negative Nb-Ta and large positive Pb anomalies, and moderate heavy rare earth element (HREE) concentrations. U-Pb geochronology on zircon extracted from metavolcanic and metaplutonic components of the Dadeville indicates crystallization of these igneous bodies between 460 and 444 Ma (Middle-Late Ordovician), while metasedimentary units contain detrital zircons ranging in age from 460 to 1300 Ma, with significant age populations at 530 Ma and 1030 Ma. The dominant Precambrian ages are consistent with Grenville and Granite-Rhyolite Province sources, while the younger, Neoproterozoic to early Cambrian populations probably represent Iapetus sources.
ranging from more proximal Iapetus rift volcanics and plutonics (760-680 Ma and 615-550 Ma) of the “Blue Ridge” rift system (Thomas, 1991) to the more distal failed rift system of the Southern Oklahoma Aulocogen (e.g. ca. 536-539 Carlton Rhyolite Group, ca. 533 Ma Roosevelt Gabbros, ca. 530 Glen Mountains Layered Complex; Brueseke et al., 2016). U-Pb zircon ages from metaigneous and metasedimentary rocks, coupled with trace element geochemical characteristics typical of volcanic arc rocks, indicate Dadeville Complex rocks represent volcanic and plutonic elements of an Ordovician arc with intercalated sedimentary sequences which sourced not only the local arc terrane, but also rocks typical of the Laurentian interior (e.g. Granite-Rhyolite Province, Grenville basement, Southern Oklahoma Aulocogen) and Appalachian Iapetus margin. These characteristics are consistent with development of Dadeville Complex rocks as a Laurentian plate volcanic arc. Additionally, its palinspastic location – southeast of the Talladega belt, eastern Blue Ridge, Dahlonega gold belt and Opelika Complex – places it in a position that should have been occupied by the corresponding arc to the WEDB back-arc.

CONCLUSIONS

The collective geologic data set from these southern Appalachian lithotectonic terranes indicates they formed in a paired backarc-arc system on the seaward edge of the Laurentian plate, outboard of the continental hinge zone (Fig. 1.11). This includes observations that portions of these back-arc sequences (WEDB) were built directly atop portions of the Neoproterozoic–Cambrian Laurentian continental margin (Ashland Supergroup); that Early and Middle Ordovician WEDB units formed in an extensional environment capable of producing significant volumes of mafic volcanic material through decompression melting (e.g., Hillabee Greenstone, Pumpkinvine Creek Formation, Sally Free mafic complex, New Georgia Group), while similarly aged Ordovician granitic magmas incorporated Grenville-aged crustal components (e.g., Hillabee metadacites, Galts Ferry gneiss, Cane Creek gneiss); that sediment deposited in this extensional setting sourced both Ordovician volcanic units and Grenville crustal sources (Wedowee, Emuckfaw, and New Georgia Groups); that these WEDB rocks apparently formed in a tectonic setting proximal to the Paleozoic Laurentian margin such that the Hillabee Greenstone could be emplaced on top of younger shelf rocks with little evidence for synchronous or prior deformation; that both the WEDB and Dadeville Complex formed in a setting that allowed them to be emplaced atop Laurentian marginal sequences without the presence of an intervening accretionary prism or forearc; and that rocks of the Dadeville Complex were emplaced atop those of the WEDB during Alleghanian orogenesis. The only tectonic model capable of accommodating all of the observed characteristics of the WEDB and Dadeville Complex is one in which an Ordovician back-arc basin and arc system developed along the seaward margin of the Laurentian plate immediately outboard of the continental hinge zone (Holm-Denoma, 2006; Barineau, 2009; Holm-Denoma and Das, 2010; Tull et al., 2007, 2014; Barineau et al., 2015), and was subsequently telescoped by Alleghanian thrusting during later continental collisional events. This Ordovician accretionary orogenic setting contrasts with models of Ordovician collisional orogenesis in the northern and central Appalachians, and suggests that one or more significant transforms (Fig. 1.12) between the two ends of the orogen
accommodated a reversal of subduction polarity, not unlike that observed along the Alpine fault of New Zealand (Tull et al., 2014; Barineau et al., 2015).

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IS THE DADEVILLE COMPLEX PART OF THE “MISSING” SOUTHERN APPALACHIAN TACONIC ARC?

David W. Farris¹, 2, James Tull¹, Paul Mueller³, and Ben Davis¹

¹Florida State University, Dept. of Earth, Ocean, and Atmospheric Sciences Tallahassee, FL 32306
²Present address: Dept. of Geology, Washington and Lee University Lexington, Virginia 24450
³University of Florida, Dept. of Geological Sciences, Gainesville, Florida 32611

ABSTRACT

The Taconic orogeny in the northern Appalachians is characterized by obduction of Ordovician arc terranes onto Laurentian crust, however exotic obducted arc terranes have not been identified in the southern Appalachians. The Dadeville complex of the Inner Piedmont comprises the most likely Taconic arc assemblage in the southern Appalachians because it contains mafic/ultramafic intrusions, batholith-scale granitic bodies, and volcanic sequences with a combined tectonostratigraphic thickness of ~15 km. It forms a broad klippe within a regional synform that was thrust from southeast of the Mesoproterozoic Pine Mountain basement massif over a proposed eastern Blue Ridge/Inner Piedmont Ordovician back-arc assemblage. The Ropes Creek Amphibolite, derived from ~9 km of tholeiitic basalt, metasedimentary rocks, and minor felsic tuff, forms the base of the Dadeville thrust sheet. It is overlain by the Agricola Schist, a >1 km thick sequence of pelitic schist, metagreywacke, and amphibolite. Both units were intruded by felsic and mafic-ultramafic plutonic rocks that comprise a majority of the complex. All magmatic units exhibit arc-like geochemistry with large negative Nb-Ta anomalies and enriched LILE’s. Zircons from the felsic plutons yield Taconic U-Pb ages of 461 ± 4 Ma to 448 ± 2 Ma. Zircons from the Agricola Schist and Ropes Creek Amphibolite metasediments yield a significant fraction of Grenville and Granite-Rhyolite Province ages, but also grains as young as 440 Ma. Although it has been previously proposed that the Dadeville is an accreted, peri-Gondwanan arc fragment obducted onto the Laurentian (or peri-Laurentian) Inner Piedmont and eastern Blue Ridge/Brevard zone, we interpret 440-480 Ma U-Pb ages of zircons from Dadeville complex metagneous rocks, Paleozoic and Precambrian detrital zircons from Dadeville complex metasedimentary rocks, initial ⁸⁷Sr/⁸⁶Sr of igneous protoliths (basalt-trondhjemite-granite), initial ⁴⁰Ar/³⁹Ar values of igneous and detrital zircons, and normalized trace element abundances of mafic rocks to indicate that the Dadeville complex formed in an Ordovician suprasubduction environment on Laurentian crust.

INTRODUCTION

Although the Taconic-Caledonian orogen was one of the first to be placed within a plate tectonic context, there remain fundamental questions with regard to its expression in the Appalachians of eastern North America. The Taconic in the northern
Appalachians records a collisional orogeny triggered by initial subduction of Laurentian (Grenville) crust beneath lithosphere of peri-Laurentian or oceanic (i.e. Iapetus) origin, leading to subsequent obduction of arc crust onto Laurentia (Karabinos et al., 1998; Zagorevski et al., 2006; van Staal et al., 2007). An important, but enigmatic aspect of this model is the lack of a similar record in the southern Appalachian orogen. This enigma may have several explanations, including erosion of obducted arc crust or burial of arc crust beneath younger (i.e. Alleghanian) thrust sheets. One of the more challenging issues with modeling the southern Appalachian Taconic orogeny in a collisional setting is that Early Paleozoic southern Appalachian passive margin strata rarely, if ever, contain Taconic-age detrital zircons (Park et al., 2010). Such zircons are found in more recent strata, but their provenance could be from outside the southern Appalachians (Eriksson et al., 2004; Mueller et al., 2008; Park et al., 2010). Additionally, the southern Appalachians lack coeval (i.e. Ordovician) faults and allochthons (Tull et al., 2014), as well as a paired arc/accretionary prism terrane structurally above the subducted continental margin typical of collisional (A-type) margins (van Staal et al., 2007). Consequently, alternatives to a Taconic subduction model with a Laurentian lower plate must be considered for the southern Appalachians. Here we use structural, geochemical, and geochronologic evidence to argue that the Dadeville Complex of the Appalachian Inner Piedmont (Figs. 1.2 and 2.1) is a fragment of the missing Taconic arc.

**Geologic Setting**

The Dadeville complex sits structurally above metamorphosed Early to Middle Ordovician supracrustal rocks of the Wedowee and Emuckfaw Groups, and rocks of the Dahlonega Gold Belt. These latter units are composed of bimodal metavolcanic rocks intercalated with thick sequences of flysch-like siliciclastic rocks (Fig. 2.2A). Tull et al. (2014) refer to this sequence as the Wedowee-Emuckfaw-Dahlonega basin (WEDB) and propose that they formed in a Taconic, back-arc environment along the former post-Rodinia, Laurentian passive margin. The Dadeville complex is currently nested in a large (>4,300 km²), synformal klippe (Tallassee synform) with a structural thickness of ~5 km along the Inner Piedmont's western flank (Fig. 2.2). Unfolding the synform and restoring the complex southeast of the underlying Opelika Group (part of the WEDB) yields a displacement of the Dadeville Complex along the basal thrust of >50 km, i.e., the Dadeville complex originated outboard of the WEDB sequence. The basal thrust (Katy Creek/Stonewall Line fault) also separates the Dadeville complex from the Jacksons Gap Group of the Brevard Zone. To the south, the complex is bounded by the Loachapoka Formation of the Opelika Group, which correlates with the Jackson's Gap Group around the hinge of the Tallassee synform (Bentley and Neathery, 1970; Grimes, 1993; Tull et al., 2014; Barineau et al., 2015) (Fig. 2.2A). Along its southeast trace the fault is concordant to lithologic units and planar fabrics in both hanging and footwalls, but in the synform’s hinge, Dadeville unit contacts and internal planar fabrics are highly oblique to the boundary (Fig. 2.2A), indicating that the thrust is post-metamorphic, and has the geometry of a hanging wall (Dadeville complex) ramp on a footwall (WEDB) flat (Fig. 2.2B).
Figure 2.1. Geologic map showing the major lithotectonic belts in the southern Appalachians, the distributions of the Dadeville complex (red), the Wedowee-Emuckfaw-Dahlonaga Basin (orange), Grenville basement massifs (black), and the location of Figure 2.2.
Drummond et al. (1997) report peak metamorphic conditions of ~5-8 kb and ~600-700°C in the Agricola Schist at the top of the Dadeville and pressures of ~10 kb and ~750-800°C in the Ropes Creek Amphibolite at its base. In contrast, the P-T conditions in the underlying Opelika Group are ~540-580°C at ~6.8-8.2 kb (Goldberg and Steltenpohl, 1990), indicating an inverted metamorphic gradient across the basal thrust. Temporal constraints on the thrusting are minimal. The Farmville metagranite (440 ± 7 Ma; Hawkins et al., 2013) intrudes the underlying Opelika complex immediately beneath the basal fault, but does not cross it. In turn, no Dadeville plutonic units occur below the fault. Thermochronologic data from the Dadeville include 40Ar/39Ar hornblende (347.1 ± 1.8 to 320.1 ± 1.6 Ma) and conventional K-Ar muscovite (336 to 331 Ma) ages, suggesting regional cooling at that time (Stonebraker, 1973; Steltenpohl and Kunk, 1993). These observations suggest the Dadeville did not reach its current structural configuration until at least the early Alleghanian.

GEOLOGIC UNITS OF THE DADEVILLE COMPLEX
The Dadeville complex is dominated by meta-igneous units ranging from ultramafic to true granite in composition, but also contains metasedimentary rocks (Fig. 2.1). There are five primary units:

1. **Ropes Creek Amphibolite.** This structurally lowest unit is >9 km thick and is composed of strongly foliated amphibolite characterized by highly planar sub-cm to meter thick layers alternating from plagioclase- to hornblende-rich compositions. The protolith was dominantly tholeiitic basalt (Stow et al., 1984; Neilson and Stow, 1986; Neilson et al., 1997), but thin interlayers of dacitic meta-tuff (Fig. 2.3) and metasedimentary rocks are present. Several larger bodies of metasedimentary rock occur near the structural base of the Ropes Creek, but above a mafic and ultramafic olistostromal sequence (Sears et al., 1981) that may mark an internal unconformity (Fig. 2.2A). These metasedimentary bodies are up to 1.5 km thick and consist of metapelite, quartzofeldspathic mica schist and metagraywacke, with 1 cm to meter scale layering. Continuous meter-scale layers extend for >100 meters in the amphibolite. We interpret the Waresville Formation (Bentley and Neathery, 1970), the Waverly Gneiss, and the Zebulon Formation (Sears et al., 1981) to be part of the Ropes Creek Amphibolite rather than being designated as separate lithodemic units.

2. **Agricola Schist.** The Agricola Schist lies structurally above the Ropes Creek and has a typical thickness of ~1 km. It contains cm- to meter-scale layering of metapelitic and more quartzofeldspathic compositions, plus pods of migmatite and <1 m thick amphibolite layers. It is interpreted to have a meta-turbidite protolith composed of siltstone, pelite, and greywacke interbedded with rare basalt flows.

3. **Camp Hill Granite.** This batholith-scale (~8 km thick) metaigneous rock intruded and now largely obscures the contact between the Agricola and Ropes Creek (Fig. 2.2A). Local peraluminous zones contain muscovite and garnet, similar to the Chattasofka Creek Gneiss (see below). Based upon the lack of K-feldspar and the high silica content (70-76 wt. % SiO₂), the Camp Hill is classified as a trondhjemite
IS THE DADEVILLE COMPLEX PART OF THE “MISSING” SOUTHERN APPALACHIAN TACONIC ARC?

Figure 2.2. A) Geologic map of the southern Appalachians showing the Dadeville complex and the Wedowee-Emuckfaw-Dahlonega back-arc basin. Location of U-Pb geochronology samples shown. B) Cross-section from X-X’ across the Dadeville complex. Note the inversion of metamorphic grade across the basal Stonewall line/Katy Creek thrust.
(Neilson et al., 1996). It has an intermediate strength subsolidus fabric, and is moderately coarse-grained.

4. Chattasofka Creek Gneiss. This unit is exposed on the northwest side of the Tallassee synform, often in close spatial association with the mafic/ultramafic complex. To the northeast it extends into the Rock Mills and Franklin Gneisses (Seal and Kish, 1990) (Fig. 2.2A). It is a strongly peraluminous, S-Type granite with 70-78 wt. % SiO2. It is distinguished from the Camp Hill by the abundance of K-feldspar and muscovite. Both the Chattasofka Creek and Camp Hill contain xenoliths of the Ropes Creek Amphibolite and Agricola Schist (Neilson and Stow, 1986; Neilson et al., 1996; Davis, 2015). The Chattasofka intruded the Doss Mountain Suite described below.

5. Doss Mountain Suite. Neilson and Stow (1986) describe abundant gabbro, pyroxenite, and norite that occur mainly in the core of the Tallassee synform as sills, layered intrusions, and dikes, some of which intrude the Agricola schist and the Ropes Creek Amphibolite. The pyroxenite is coarse-grained, and contains orthopyroxene and clinopyroxene with minor spinel. Olivine is not a significant phase in either the norite or pyroxenite. The gabbroic units (e.g., Slaughters gabbro) contain significant plagioclase and are both olivine and non-olivine bearing. Although parts of the mafic complex have been recrystallized into mafic schists or amphibolites, substantial volumes retain primary igneous textures and mineralogy. Smaller mafic/ultramafic bodies intruded the lower Ropes Creek, structurally below the main masses of the mafic/ultramafic suite in the core of the Tallassee synform (Fig. 2.2A).

Figure 2.3. Outcrop of Ropes Creek Amphibolite on Ropes Creek (32°37′35.00″N, 85°41′10.48″W) showing thinly bedded layers of amphibolite (grey) and light colored layers of rhyodacite metatuff.
GEOCHEMISTRY OF THE DADEVILLE COMPLEX

In terms of major element chemistry, Dadeville complex igneous units define discrete fields separated by large compositional gaps (Fig. 2.4A). Members of the ultramafic Doss Mountain suite have high average MgO contents (19-22 wt. %). The Slaughters gabbro and the Ropes Creek Amphibolite are compositionally similar, with 5-11 wt. % MgO at 45-50 wt. % SiO₂. The Camp Hill and Chattasofka Creek have higher silica (70-78 wt. % SiO₂ at <1.5 wt. % MgO), but are best distinguished using SiO₂-K₂O relationships, with the former plotting mostly in the low-K series (<1.5 wt. %) and the latter plotting in the high-K series (>3.5 wt. % K₂O) (Fig. 2.4B). Overall, Dadeville mafic rocks and the Camp Hill gneiss follow tholeiitic trends, whereas the Chattasofka Creek gneiss is calc-alkaline (Fig. 2.4B).

In terms of trace element signatures, Dadeville mafic and felsic igneous rocks share significant commonalities, including: large ion lithophile (LILE) enrichment, a large negative Nb-Ta anomaly, a large positive Pb anomaly, and moderate heavy rare earth element (HREE) contents (Fig. 2.5A). Ultramafic rocks exhibit these characteristics to a lesser extent, with their trace element signatures dominated by pyroxene. As with major elements, the Slaughters gabbro and Ropes Creek Amphibolite have very similar trace element compositions, although the gabbroic rocks have somewhat lower HREE values (0.3x vs. 0.5x N-MORB, respectively). The Camp Hill and Chattasofka Creek gneisses have somewhat higher HREE abundances (1.0x N-MORB), and significantly higher LILE’s (50x vs. 5x N-MORB). The granitic rocks also show evidence of Fe-Ti oxide fractionation (large negative Ti anomaly), and plagioclase fractionation (moderate negative Sr anomaly). Overall, the trace element signature of Dadeville igneous units exhibits features characteristic of subduction-driven igneous activity. In particular, the basaltic and gabbroic rocks are very similar to compositions of basalt found in intra-oceanic arcs (Fig. 2.5B). However, as discussed below, detrital zircon ages clearly show that the Dadeville rocks formed in proximity to continental crust.

GEOCHRONOLOGY OF THE DADEVILLE COMPLEX

Previous geochronologic studies of Dadeville units are limited to the cooling ages described above, and a whole-rock Rb-Sr age of 462 ± 4 Ma interpreted as the crystallization age of the Franklin Gneiss (aka Chattasofka Creek gneiss; Seal and Kish, 1990). Here we report U-Pb zircon analyses by laser ablation inductively coupled plasma mass spectrometry (Mueller et al., 2008) for Dadeville igneous protoliths and metasedimentary units of the Agricola Schist and Ropes Creek Amphibolite. All ages plotted are either <5% discordant (206Pb/238U ages <900 Ma) or <10% discordant ages >900 Ma. All errors reported at 2x the standard error of the mean.

Among the felsic rocks, U-Pb ages were determined for the Camp Hill and Chattasofka Creek gneisses and a felsic layer from the Ropes Creek (Fig. 2.6A-E). Two samples from the Camp Hill Gneiss (CH-197 and CH-216) yielded ages of 448 ± 2, and 446 ± 2 Ma based on averaging the 20 and 11 most concordant grains, respectively (Figs. 2.6A and 2.6B). These results place the Camp Hill as a late Taconic intrusion and establish a minimum age for the Ropes Creek and Agricola units, which were intruded by the Camp Hill. The Chattasofka Creek Gneiss (CH-58, n=61) yielded
Figure 2.4. Dadeville Complex major and trace element geochemistry. A) MgO wt. % vs. SiO$_2$ wt. %, B) K$_2$O wt. % vs. SiO$_2$
Figure 2.5. A) N-MORB normalized trace element spider diagram (normalized to values in Klein, 2003), B) Oceanic Arc basalt normalized trace element spider diagram (normalization values are from Kelemen, et al., 2003). Both major (Fig. 2.4) and trace element data contain significant arc geochemical signatures.
a somewhat older crystallization age of 461± 4 Ma based on 31 grains with 0-2% discordance (Fig. 2.6C). That result is compatible with the lower ages obtained from the Camp Hill and may help explain the peraluminous parts of that unit as rafts of the older gneiss in the younger granitoid. Zircons from a felsic layer in the Ropes Creek (TH-183, n=48) range from 318 to 2570 Ma (Figs. 2.6D and 2.6E). Although arbitrary, the best estimate of crystallization age for this sample is the average of the ten oldest analyses at 444 ± 4 Ma, which, within error, is compatible with field relations. The Precambrian zircons are likely detrital, and suggest a volcaniclastic origin of the felsic layer.

Two samples from the Agricola Schist, yielding significant numbers of detrital zircons (CH-30 and CH-34, Fig. 2.7A-C, sample localities 2 & 3, Fig. 2.2A ), give ages from ~ 460 to 1300 Ma with prominent concentrations centered about 530 Ma (Iapetan rifting) and 1030 Ma (Grenville). Detrital zircons from two samples (LN-55 and 62) of metasedimentary rocks near the base of the Ropes Creek Amphibolite (Figs. 2.7D and 2.7E, sample localities 7 & 8, Fig. 2.2A) are characterized by significant discordance, with <50% of the analyses passing the discordance screens. For LN-55, 50 of 108 analyses pass the discordance screens and range from 286 to 1671 Ma (Fig. 2.7E) and only 43 of 102 analyses pass the discordance screens for LN-62 (300-1464 Ma) (Fig. 2.7D). Both samples yield mostly Precambrian ages with concentrations in the Mesoproterozoic (Grenville and Granite-Rhyolite Province age ranges). Even at this sampling density, it is clear that the Agricola schist and Ropes Creek Amphibolite metasedimentary protoliths in the Dadeville complex had diverse, pre-Ordovician, continental provenances, with a strong Mesoproterozoic signal that is difficult to reconcile with Gondwanan sources.

Viewed collectively, the Agricola and Ropes Creek metasedimentary samples do not exhibit aspects of solely Laurentian or Gondwanan sources, and do not compare favorably to any of the spectra characteristic of Early Paleozoic sandstones from the accreted Gondwanan Suwannee terrane (Florida sub-surface; Mueller et al., 1994; 2014). These differences include Grenville-age zircons in the Agricola and Ropes Creek, but not in the Suwannee sandstones. The Suwannee rocks also contain Birimian (Africa)/Trans-Amazonian (South America) grains (~2.1-2.3 Ga), but the Agricola and Ropes Creek samples do not. Moreover, the Dadeville rocks show evidence for Acadian metamorphism, for which there is no evidence in the Suwannee rocks. Overall, ages of zircons from rocks of the Dadeville complex rocks with both igneous and sedimentary protoliths suggest the complex is more likely to be a Laurentian or peri-Laurentian terrane rather than a Gondwanan one as suggested by VanDervoort et al. (2015), which implies B-type subduction beneath the Laurentian margin during the Taconic orogeny in the southern Appalachians. Paired with the Ordovician Wedowee-Emuckfaw-Dahlonega back-arc basin and the similarly aged Blount retro-arc basin, we propose that the Dadeville complex was part of a subduction system constructed on extended Laurentian margin crust during the accretionary Taconic orogeny (Fig. 2.8).
Figure 2.6. Kernel density estimator (KDE) plots (Vermeesch, 2012) of U-Pb ages of zircons from metaigneous rocks of the DVC. All ages plotted have passed the 5% and 10% concordance screens discussed in the text. Age concentrations (peaks) are labeled, but, should not be interpreted to indicate the times of specific events. A) Camp Hill gneiss sample CH-197, B) Camp Hill gneiss sample CH-216, C) Chattasofka gneiss sample CH-58, D) Ropes Creek felsite sample TH-183 with the full age range, and E) Inset of Ropes Creek felsite sample TH-183 from 300-500 Ma.
Figure 2.7. Kernel density estimator (KDE) plots (Vermeesch, 2012) of U-Pb ages of zircons from metasedimentary rocks of the DVC. All ages plotted have passed the 5% and 10% concordance screens discussed in the text. Age concentrations (peaks) are labeled, but, should not be interpreted to indicate the times of specific events. A) Agricola schist sample CH-34, B) Inset of Agricola schist sample CH-34 from 300-800 Ma, C) Agricola schist sample CH-30, D) Ropes Creek metasediment sample LN-62, and E) Ropes Creek metasediment sample LN-55
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CONCLUSIONS

The Ordovician Dadeville complex comprises an ~5 km thick, synformal klippe containing >15 km of mostly metaigneous stratigraphy in the western Inner Piedmont that was thrust over an extensive Laurentian back-arc basin sequence of similar age during late Paleozoic collisional events. The complex represents the largest and most coherent fragment of intact Taconic arc crust in the southern Appalachian orogen. Structural reconstructions of the Dadeville complex indicate a thickness of >15 km, consistent with the observed 5 kbar range in metamorphic pressure. Trace element signatures of Dadeville silicic and mafic/ultramafic rocks are broadly arc-like and best explained via formation in an extensional arc setting. However, the significant population of Grenville and Granite-Rhyolite Province-age detrital zircons in the Agricola Schist and Ropes Creek Amphibolite indicates that the complex most likely formed near or on the Laurentian margin. Additionally, rocks of the Dadeville complex were tectonically emplaced atop rocks of the WEDB, Laurentian-plate back-arc – without an intervening accretionary prism or forearc terrane. Therefore, we interpret the Dadeville complex as part of a fringing Laurentian arc outboard of the coeval Wedowee-Emuckfaw-Dahlonega back-arc basin, both of which formed due to subduction of Iapetus oceanic lithosphere beneath Laurentian Mesoproterozoic continental crust (Fig. 2.8).
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REFERENCES


CRYSTALLINE THRUST SHEETS OF THE FRONTAL SOUTHERN
APPALACHIAN OROGEN IN GEORGIA AND ALABAMA: INSIGHTS FROM THE
COMPOSITE MULBERRY ROCK RECESS HALF-WINDOW

James F. Tull¹, Christopher S. Holm-Denoma², and Nawwaf Almuntshry¹

¹Department of Earth, Ocean and Atmospheric Science, Florida State University, Tallahassee, FL 32306
²U.S. Geological Survey, Central Mineral and Environmental Resources Science Center, Denver Federal Center, Denver, CO 80225-0046

ABSTRACT

Breached regional folds that trend oblique to regional strike in a terrane occupied by thrust faults allow for a three-dimensional view of the thrusts’ hanging and foot wall geometries, and for an analysis of regional kinematic relationships. Several of these structures exist along the eastern-western Blue Ridge boundary in Georgia and Alabama. Among the most prominent is the Mulberry Rock recess, which indents that boundary 20 km, forming a half-window and allowing for geometric examination of four major southern Appalachian thrust sheets. The thrusts described here telescoped parts of the previously deformed and metamorphosed Laurentian outer margin stratigraphy, significantly shortening that margin and placing higher-grade rocks on lower grade rocks, cutting obliquely through earlier metamorphic fabrics and lithostratigraphy. Examination of this recess and those to the southwest in Alabama allow for minimal estimates of the displacements of all of the frontal crystalline thrusts of the southern Appalachians. It appears that with respect to much of the deformation in the external foreland thrust wedge, all of the frontal hinterland thrusts described here, with the exception of the Hillabee thrust, represent out-of-sequence faulting which propagated through previously deformed and metamorphosed lithostratigraphy and earlier emplaced thrust sheets. Importantly, these hinterland thrusts post-dated major thrusting and associated deformation in at least the southeastern part of the more cratonward foreland wedge.

INTRODUCTION

The Mulberry Rock recess is a 20 km north-south indentation into the eastern Blue Ridge frontal fault system in the southern Appalachian orogeny, which exposes four metamorphic thrust sheets within a breached, composite occupying an area of ~65 km² (Figs. 1.2, 3.1 and 3.2). The half-window corresponds with an open, upright, doubly-plunging, ~north-south-trending antiform (Mulberry Rock antiform). Analysis of the structural framework of this feature provides important insight into the structural development of this key part of the Appalachian orogen, and into the kinematics and trajectories of thrusts which cut previously deformed and metamorphosed sequences. The four thrust sheets exposed within parts of the recess include the footwall Talladega belt (western Blue Ridge), and the Mulberry Rock, Allatoona, and Burnt Hickory Ridge allochthons of the eastern Blue Ridge. The thrusts at the base of the latter two allochthons are of regional significance to the southern part of the Appalachian orogen.
Figure 3.1. Generalized geologic map of Talladega Belt and western and eastern Blue Ridge in Alabama and Georgia showing the locations of Figures 3.2, 3.3, and 3.5. Minimum horizontal components of net slip for the Talladega–Cartersville, Hillabee, and Hollins Line faults are shown by dashed arrows.
Figure 3.2. Geologic map of the region of the Mulberry Rock recess including portions of 5 thrust sheets: 1) Foreland Fold and Thrust belt, 2) Talladega belt, 3) Mulberry Rock, 4) Pumpkinvine Creek Formation, 5) Emuckfaw/New Georgia Group. Includes data from the following 7.5' quadrangles: Burnt Hickory Ridge, Dallas, Draketown, New Georgia, Rockmart South, Taylorsville, and Yorkville. Location of this figure in Figure 3.1.
COMPOSITION OF THE THRUST SHEETS

The geology (Fig. 3.2) of the four separate regionally developed thrust sheets exposed within the core and rim of the Mulberry Rock recess is described below in order of overlying structural succession. Although the sequences within the two intermediate thrust sheets are locally absent, the relative structural order is always preserved.

Talladega thrust sheet/Talladega-Cartersville fault (Sheet A)

The core and north flank of the recess is occupied by the greenschist facies Talladega belt, an extension of the western Blue Ridge and the frontal metamorphic allochthon in this segment of the Appalachians (Tull and Holm, 2005) (Figs. 1.2, 3.2 and 3.3). The rocks in the environs of the recess within this thrust sheet consist predominantly of biotite-bearing chlorite sericite phyllite, with lesser lenses (thickness up to ~100 m) of metagraywacke and metaconglomerate, belonging to the Silurian–Devonian Lay Dam Formation (Tull and Telle, 1988). The thrust at the base of this sequence is the Talladega-Cartersville fault, which extends northeastward as the Great Smoky fault (Fig. 1.2). This fault has a minimum horizontal component of net slip of 23 km (Tull, 1984).

Mulberry Rock thrust sheet/Mulberry Rock fault (Sheet B)

The Mulberry Rock Gneiss (McConnell and Abrams, 1984) occurs structurally above rocks of the Talladega belt along the Mulberry Rock thrust as a series of thin (up to ~300 m thick) horses discontinuously along the inner rim of the recess, occupying a structural position along the interface between the eastern and western Blue Ridge (Figs. 1.2, 3.2 and 3.3). The gneiss is a medium grained, slightly metaluminous, garnet-bearing two-mica granitic gneiss with subequal amounts of quartz, plagioclase, and K-feldspar (Holm-Denoma, 2006). It contains an amphibolite facies gneissosity, which is cut by the surrounding faults, and is the only lithology within this thrust sheet. Higgins et al. (1988) argued that the gneiss was correlative to the Corbin Gneiss to the northeast in the southwest corner of the western Blue Ridge (Fig. 1.2). The Corbin is a component of the Grenville basement (Odom et al., 1973; Heatherington et al., 1996), but later U/Pb studies of zircon from the Mulberry Rock Gneiss showed it to be significantly younger than Corbin basement. Laser ablation ICPMS and ion microprobe (SHRIMP) zircon U/Pb systematics of the gneiss are complicated, but the most concordant analyses fall between 432 and 450 Ma (P.E. Mueller and R.Das, written communication). The gneiss is similar in age and geochemical character to the ca. 440–430 Ma Austell Gneiss (Higgins et al., 1997) in the overlying thrust sheets containing the Wedowee and Emuckfaw/New Georgia Groups (Fig. 1.2), suggesting that it may share a genetic origin. Zircons from both plutonic bodies include a fraction of inherited Grenville aged (~1.1 Ga) zircons and zircon cores, suggesting that these formed while incorporating or intruding into Laurentian-affinity, Grenvillian crust.

Pumpkinvine Creek thrust sheet/Allatoona fault (Sheet C)

Around the flanks of the Mulberry Rock recess, the Pumpkinvine Creek and Canton formations of the Dahlonega gold belt lie structurally above both the Talladega
Figure 3.3. Geologic map of the region of the Mulberry Rock recess differentiating the four thrust sheets described herein. Blue dashed arrows show the minimal horizontal components of the slip vectors of the three upper thrust sheets. Location of this figure in Figure 3.1.
belt and Mulberry Rock Gneiss along the Allatoona fault, forming a fault-bounded sequence 3–6 km wide, with a maximum structural thickness of ~4 km (Figs. 1.2, 3.2 and 3.3). These units are not exposed to the southwest beyond the west rim of the recess, but continue for >240 km to the northeast as the Helen Group in the Dahlonega gold belt in north Georgia (Figs. 1.2 and 3.1) and the correlative Otto Formation in North and South Carolina (Gillon, 1982; Hatcher, 1988). The Pumpkinvine Creek Formation is a bimodal metavolcanic unit consisting predominantly of orthoamphibolite interlayered with subordinate felsic gneiss (Galts Ferry gneiss member of McConnell, 1980, and Barlow gneiss of German, 1985; 1989), in addition to minor ferruginous quartzite and garnetiferous quartzose two-mica schist (Holm-Denoma, 2006; Holm-Denoma and Das, 2010). The informally designated Galts Ferry and correlative Barlow gneisses yield consistent Middle Ordovician U/Pb zircon ages (Thomas, 2001; Holm-Denoma and Das, 2010), whereas trace element and isotopic geochemistry from intercalated orthoamphibolites suggest development in a suprasubduction setting — specifically a back-arc (Holm-Denoma, 2006; Holm-Denoma and Das, 2010). In stratigraphic contact with the Pumpkinvine Creek Formation, rocks of the Canton Formation are dominated by variably graphitic, garnetiferous two-mica schist (metapelite), but also include subordinate biotite plagioclase quartz gneiss (metagraywacke) and amphibolite (McConnell and Abrams, 1984; German, 1985; Holm-Denoma, 2006; Holm-Denoma and Das, 2010). The sequence is at garnet grade, below the kyanite isograd (Fig. 3.3). A suite of detrital zircons separated from a metasandstone within the Canton Formation yielded 80% ages between 1,017 and 1,190 Ma (Holm-Denoma and Das, 2010). These ages are typical of Laurentian crustal materials formed during the Grenville orogeny and likely represent sediment derived from the Laurentian continental margin (Bream et al., 2004; Becker et al., 2005). Two analyses, one each at 513 Ma and 589 Ma, are potentially related to synrift (Iapetus rifting) igneous rocks, and older age peaks of 1,441 Ma and 1,582 Ma correspond to ages of the Eastern Granite-Rhyolite Province (Becker et al., 2005).

Wedowee and Emuckfaw/New Georgia Groups thrust sheet/Burnt Hickory Ridge fault (Sheet D)

Rocks of the Wedowee and Emuckfaw/New Georgia Groups are faulted above the Pumpkinvine Creek thrust sheet along the Burnt Hickory Ridge thrust (Figs. 1.2, 3.2 and 3.3). In Alabama, the Wedowee lies stratigraphically and structurally above rocks of the Ashland Supergroup, but is separated from the Ashland across much of the Alabama eastern Blue Ridge by the (Goodwater-Enitachopco) Allatoona fault (Tull, 1978) (see below). However, southwest of Goodwater, where displacement on the Allatoona fault ends (Fig. 3.1), the boundary between the two units is a polydeformed stratigraphic contact (Allison, 1992; Tull et al., 2014; Barineau et al., 2015). Rocks of the Wedowee Group are dominated by locally highly carbonaceous metapelite (variably garnetiferous graphitic schists) interlayered with less common quartzites and thin layers of highly feldspathic schist and/or fine-grained biotite gneiss interpreted as metagraywacke (Neathery and Reynolds, 1973; Allison, 1992). Accessory minerals include tourmaline, staurolite, kyanite, and sillimanite. The Emuckfaw Group (Heard Group of Bentley and Neathery, 1970) in Alabama and the equivalent New Georgia Group (McConnell and Abrams, 1984; German, 1985) in Georgia consists of a sequence of variably graphitic-garnetiferous two-mica schist interlayered with less common fine-grained variably garnetiferous, biotite paragneiss (metagraywacke), micaceous quartzite, metaconglomerate, metadiamicite, and rare calc-silicate rocks, in addition to variable amounts of orthoamphibolite. The lower contact of the Emuckfaw/New Georgia Group with
the underlying Wedowee Group is commonly marked by gradation of highly graphitic schists of the Wedowee Group into non-graphitic or graphite-poor schists of the lower Emuckfaw over an interval of 5–15 m, often coinciding with the appearance of orthoamphibolite in the lower Emuckfaw. Less than 20% of the Emuckfaw Group section in Alabama and west Georgia consists of amphibolite (metabasalt), with most of the section being composed of metasedimentary rocks dominated by mica schist (Fig. 1.2). However, northeastward in Carroll County, Georgia, in the stratigraphically equivalent New Georgia Group, the volume of metabasalt increases significantly in this section, becoming the predominant lithology (Figs. 1.2 and 3.2). This section is also marked by an abundance of banded iron formation and small metagabbro bodies. This massive sequence of dominantly mafic metavolcanic rocks occupies the type area of the New Georgia Group and appears to be centered on the Mulberry Rock recess.

U/Pb detrital zircon studies have been conducted on a number of metasedimentary units from this thrust sheet. All of the units contain an overwhelming population of Mesoproterozoic U/Pb zircon ages between 1.2 and ca. 0.9 Ga, which are typical of Laurentian crustal materials formed during the Grenville orogeny (Rodinian collision; Tollo et al., 2010) and most likely represent sediment derived from the adjacent Laurentian continental margin or exposed rifted-margin basement blocks (and/or their cover) along that margin (Tull et al., 2014). Older age peaks between 1.5 Ga and ca. 1.3 Ga are not uncommon and correspond to ages of Laurentia’s eastern Granite-Rhyolite Province (Becker et al., 2005). In addition to the Mesoproterozoic detrital zircons, those extracted from metasandstone and metagraywacke in the middle and upper Wedowee and lower Emuckfaw Groups (Figs. 3.2 and 3.6) also yield U-Pb ages between ca. 474 Ma and 460 Ma, suggesting a Middle Ordovician as well as a Mesoproterozoic (Grenville) source for much of the sediment within this thrust sheet (Tull et al., 2014). The presence of Middle Ordovician detrital zircons suggests proximity to Middle Ordovician igneous units of this predominantly sedimentary basin. Units of this age in the region include the metavolcanic rocks within the Pumpkinvine Creek Formation (see above). Ca. 460 Ma detrital zircons from the Wedowee Group provide an approximate maximum depositional age for the sequence, whereas this thrust sheet contains numerous pre-kinematic intrusive bodies of felsic gneiss that provide an upper age constraint on deposition. These include the 463 ± 4 Ma Acworth Gneiss (Holm-Denoma, 2014 personal communication; Barineau et al., 2015), the 458 ± 3 Ma Villa Rica Gneiss (Thomas, 2001), the 467–455 Ma Kowaliga Gneiss (Barineau et al., 2015; Sagul, 2016; Sagul et al., 2017), and the 477–457 Zana Granite (Sagul, 2016; Sagul et al., 2017), some of which are similar in age and composition to the Mulberry Rock Gneiss (Fig. 1.2).

STRUCTURES ASSOCIATED WITH THE MULBERRY ROCK RECESS

Mulberry Rock Antiform

The dominant structural feature of the recess is an open, upright, shallowly doubly-plunging antiform (Mulberry Rock antiform) with an axial trace trending ~ N 4° E (Fig. 3.3). Erosional breaching of this antiform resulted in exposure of the structural framework of the composite half-window at the current erosional level. The antiform folds the bounding thrusts, with those on the east limb dipping eastward and those on the west limb dipping westward. However, measuring the limb dips is difficult because it requires accurate knowledge of the dips of the surrounding thrusts, which are rarely exposed. Analysis of planar fabrics (compositional layering and cleavage-schistosity-
gneissosity), which predate thrusting within the surrounding thrust sheets, however, allows for some understanding of the antiform’s limb dips. These fabrics mostly strike north-northeast and dip moderately to steeply to the southeast, oblique to the approximately north-south trend of the antiform’s axis (Figs. 3.5-3.8). The analysis below shows that these fabrics are little affected (rotated) by the folding that produced the antiform.

To assess the geometry of the antiform, the region was divided into nine subareas for structural analysis, consisting of portions of thrust units outside, inside, on each limb, and at the southern closure of the antiform (Fig. 3.4). Although the sub-horizontal ~N 04° E (004°) axis of the antiform is oblique to the compositional layering and cleavage/schistosity/gneissosity within the thrust sheets, progressive tightening (increasing limb dips) of the antiform should rotate these planar data on the west limb westward, resulting in shallower southeast dips than the original regional orientation, and rotating planar data on the east limb eastward, resulting in steeper southeast dips than the regional orientation on that limb. For the Talladega belt (thrust sheet A), this analysis illustrates that there is essentially no difference in subparallel cleavage and bedding orientations both inside and outside the half-window or on either side of the antiform’s axial trace, i.e. cleavage is undeflected across the axial trace (Fig. 3.5). Cleavage and bedding outside the window have only a slightly steeper (~6°) southeast dip than that inside the window (Fig. 3.6 A and B). In addition, cleavage and bedding on each limb of the antiform within the window in thrust sheet A have essentially the same orientation and show no noticeable relative rotation as a result of formation of the antiform (Fig. 3.6, C-D vs. E-F). Within thrust sheets C (Fig. 3.7, Subareas 4 and 7) and D (Fig. 3.8, Subareas 3) subparallel cleavage and compositional layering data also show almost no variation in orientation between both limbs of the antiform (Fig. 3.7, A-F; Fig. 3.8, A-F). Gneissosity in thrust sheet B (Mulberry Rock Gneiss) is subdivided into two subareas (6 and 8), with Subarea 6 on the east limb of the antiform and Subarea 8 within the hinge of the antiform (Fig. 3.9). Like the other thrust sheets, sheet B shows little
evidence of the earlier fabrics being noticeably rotated by the later antiform. **We interpret these relationships to indicate that the Mulberry Rock antiform is a very gentle fold with a large interlimb angle of ≥140° that has arched the initially southeast-dipping thrusts into the current geometry of the recess.** A fold with increased tightness such as an open fold (interlimb angle 120° to 70°) or tighter would noticeably change the orientations of cleavage/schistosity/gneissosity on each limb of the antiform relative to the other, and this is not the case.

**Figure 3.5.** Geologic maps showing S1 slaty cleavage measurements in the Talladega slate belt on both sides of the Mulberry Rock antiform’s axial trace (dashed lines) north of (A.) and inside (B) the Mulberry Rock ½ window. See Figure 3.1 for locations.
Figure 3.6. Lower hemisphere equal area projections of poles to S1 slaty cleavage in subareas 1 and 2, thrust sheet A in Figure 3.4. Contoured poles in D and F.
Figure 3.7. Lower hemisphere equal area projections of poles to S1 schistosity in subareas 3 and 5, thrust sheet C in Figure 3.4. Contoured poles in B, D, and F.
Figure 3.8. Lower hemisphere equal area projections of poles to S1 schistosity and gneissosity. In subareas 4 and 7, thrust sheet D in Figure 3.4. Contoured poles in B, D, and F.
The oldest and structurally highest fault associated with the recess is the Burnt Hickory Ridge fault at the base of thrust sheet D. This fault is semi-concordant to its footwall stratigraphy (Pumpkinvine Creek and Canton Formations), especially on the eastern side of the Mulberry Rock recess. On the west side of the recess, however, the fault cuts down section to the northwest in the hanging wall obliquely through the metamorphic fabrics and, significantly, through >7 km of the lithostratigraphy of the Wedowee and Emuckfaw/New Georgia Groups (Fig. 3.2). This geometry is similar to that of the Hollins Line fault (Fig. 1.2) to the southwest in Alabama (discussed below). The geometry within the half-window indicates that the minimum horizontal component of net slip on the Burnt Hickory Ridge fault

Figure 3.9. Lower hemisphere equal area projections of poles to S1 gneissosity in subareas 6 and 8, Figure 3.4. Contoured poles in B and D.
is 14.8 km (Fig. 3.3), as it displaced the Wedowee and Emuckfaw/New Georgia Groups to the northwest over the Pumpkinvine Creek and Canton Formations. The structural position of thrust sheet B (Mulberry Rock Gneiss) is odd because neither of its bounding thrust sheets, A or C, contains any plutonic rocks. As noted above, sheet D, however, contains similar aged, sized, and composition plutonic bodies to the Mulberry Rock Gneiss. A likely possibility is that the Mulberry Rock Gneiss was originally a large (>65 km²) plutonic component of thrust sheet D, located on the northwest flank of that structure (Fig. 3.10A and B). If so, then the Burnt Hickory Ridge fault must have initially displaced the composite sheet B and D allochthon structurally over the location of the deformed and metamorphosed Pumpkinvine Creek basin rocks, upward onto the Talladega belt (Fig. 3.10B), because the Mulberry Rock Gneiss lies directly atop Talladega belt rocks along the approximately north-south trending Mulberry Rock fault on the inner flanks of the recess (Figs. 1.2 and 3.2). This scenario suggests that the
Mulberry Rock and the Burnt Hickory Ridge faults were originally the same structure (Fig. 3.10A and B) but were later cut by the younger Allatoona fault, with the Mulberry Rock Gneiss left as an isolated horse in the footwall of the latter fault, and thrust sheet D and the Burnt Hickory Ridge fault carried forward passively in a piggyback fashion along the Allatoona fault (Fig. 3.10 C and D). The minimum horizontal component of net slip on the Mulberry Rock fault is ~12 km in a northwest direction (Fig. 3.3), as it displaced the Mulberry Rock Gneiss over the Talladega belt. If the Mulberry Rock and the Burnt Hickory Ridge faults were originally the same fault, then the minimum horizontal component of net slip on it would be the sum of these two faults, or ~27 km. Fifty kilometers northeast of the recess, southeast of Canton, Georgia, the Burnt Hickory Ridge fault converges with the Chattahoochee fault (Figs. 1.2 and 3.1). The relative ages of these two faults are currently unknown.

Allatoona Fault

The youngest fault in the regional kinematic sequence is the Allatoona fault, an out-of-sequence thrust which cuts both the Mulberry Rock and the Burnt Hickory Ridge faults (Figs. 1.2, 3.2 and 3.3), but also cuts structures in its footwall associated with the underlying “Cartersville transverse zone,” including the Cartersville cross-antiform (Tull and Holm, 2005) (Fig. 3.1). This transverse zone is an alignment of thin-skinned transverse structures in the foreland and western Blue Ridge/Talladega belt which are, at least in part, the consequence of sinistral thrust motion across a large, oblique, contractual footwall ramp during the compression and telescoping of the margin associated with late Paleozoic Alleghanian collisional events (Tull and Holm, 2005). The transverse zone is outlined by a regional-scale shallowly-southeast plunging, gentle–to-open cross-antiform (Cartersville cross-antiform) that arches the western Blue Ridge/Talladega belt and underlying foreland thrust sheets. This antiform is decapitated by the Allatoona fault. Thus, the Allatoona fault is among the youngest structures in the kinematic sequence in the frontal part of the southern Appalachian orogen.

Cleavage and bedding in the Talladega belt within the Mulberry Rock recess have an approximate regional strike of N 28° E (028°) and dip of ~56° SE (Fig. 3.6 B, D, and F). Given the likely northwest-directed displacement on the Mulberry Rock and Allatoona thrusts, the trajectory of both of these faults cuts obliquely and stratigraphically down section in that direction through >11 km of the previously metamorphosed and deformed footwall Talladega belt stratigraphy, with footwall stratigraphic cutoffs along these faults trending parallel to the regional strike (Fig. 3.3). Around much of the recess, the immediate footwall of the Allatoona fault is thrust sheet B, but along the north and west flanks of the recess that fault cuts down through sheet B and sheet A becomes the immediate footwall.

At two locations in the southern part of the half-window the Allatoona fault also cuts across the Burnt Hickory Ridge fault and then structurally farther downward to the southeast into sheet B (Fig. 3.2), indicating that sheet C is eliminated down dip to the southeast between sheets B and D. Sheet C (Pumpkinvine Creek and Canton Formations) extends along strike from at least as far north as Dahlonega, Georgia for 120 km to the Mulberry Rock recess, but is not found to the southwest beyond the recess (Figs. 3.1 and 3.3). Southwest of these locations the Allatoona fault carries thrust sheet D directly over thrust sheet A, implying that southwest of the recess, the Allatoona fault has cut up through sheet C, eliminating it farther to the southwest, and has cut off the older Burnt Hickory Ridge fault, becoming the structural base of sheet D as it extends southwest toward Alabama. Within the recess, the Allatoona fault carries thrust sheet C over sheets B and A. These relationships show that, depending on the location around the recess, sheet D is in direct contact with all three underlying thrust
sheets, and at different locations the base of this sheet is either the older Burnt Hickory Ridge or the younger Allatoona fault. The minimum northwest horizontal component of net slip on the Allatoona fault measured from the southern point of the recess to the southeast flank of the Talladega belt outside of the recess is 14.3 km (Fig. 3.3).

REGIONAL FAULT RELATIONSHIPS

**Talladega-Cartersville Fault**

The post-metamorphic, low-angle Talladega-Cartersville fault forms the base of the Talladega slate belt and the upper boundary of thrust sheets of the structurally underlying foreland fold and thrust belt (Figs. 1.1 and 1.2). The minimum horizontal component of net slip of this far-traveled thrust sheet is 23 km in the southwest part of the belt (Tull, 1984) and 21 km in the Rockmart area (Tull and Hilton, 2008) (Fig. 3.1). The fault is oblique to cleavage, and stratigraphic units (as young as upper Mississippian), and cuts progressively structurally downward through several major foreland thrust sheets toward the southwest. It decapitates map-scale folds in both its hanging and footwalls (Tull, 1984; Tull and Hilton, 2008), and places greenschist facies rocks above sub-greenschist facies rocks. Palinspastic restoration of the Talladega belt (by unstacking of underlying thrust sheets) indicates it lay at least 190 km southeast of its present position prior to late Paleozoic, Alleghanian thrusting, placing it astride the approximate present position of the Mesoproterozoic to early Paleozoic Pine Mountain Window (Thomas and Bayona, 2005). Overlying thrust sheets must root farther to the southeast.

**Allatoona Fault**

As within the northern part of the Mulberry Rock recess, both to the northeast and southwest, outside the recess, the Allatoona fault marks the regional structural top of thrust sheet A, extending for >165 km from Dahlonega, Georgia, across the north flank of the recess to Hightower, Alabama (Figs. 1.1, 1.2 and 3.1). From the northwest tip of the recess to Dahlonega, the immediate hanging wall of the Allatoona fault is thrust sheet C, but to the southwest of the recess that fault cuts up through the overlying Burnt Hickory Ridge fault into thrust sheet D (see footwall cutoff by the Allatoona fault of the latter fault at the northwest tip of the recess, Fig. 3.3), and thrust sheet D lies in the immediate Allatoona hanging wall, with sheet A in the footwall. From the northwest tip of the recess, this relationship, with the Allatoona fault forming the boundary between sheets D and A continues for 64 km southwestward to Hightower, Alabama. At Hightower, the Allatoona cuts the older Hollins Line fault in its footwall (see below) and extends for another 106 km southwestward as an internal eastern Blue Ridge fault (Figs. 1.2 and 3.1). Southwest of Hightower, the Allatoona corresponds with a fault previously termed the Goodwater-Enitachopco fault (Figs. 1.2 and 1.8), interpreted here as a thrust fault (Fig. 3.1) (Tull, 1978). Since the term “Allatoona fault” was first used by Hurst (1973), that name takes precedent for this entire fault system. Southwest of Hightower, thrust sheet D continues to constitute the hanging wall of the Allatoona fault (previously the Goodwater-Enitachopco fault of Tull, 1978), but a thick sequence of Ashland Supergroup strata (Tull, 1978) emerges from beneath the Allatoona fault.
and overlying Hollins Line fault system at the top of thrust sheet A (Talladega belt), resulting in the Ashland Supergroup forming the Allatoona's footwall for 100 km farther to the southwest from Hightower to near Goodwater, Alabama (Fig. 1.8). Southwest of Goodwater, however, the Allatoona fault apparently tips out (Fig. 1.8) and the boundary between the Wedowee Group and the Ashland Supergroup is interpreted as a polydeformed stratigraphic contact (Allison, 1992, Tull et al., 2014; Barineau et al., 2015). Thus, the trace of the Allatoona fault apparently extends for at least 280 km as the youngest major thrust fault in the metamorphic core of the southern Appalachians. The distinctive sequence of the Ashland Supergroup (Tull, 1978), which forms the hanging wall of the Hollins Line fault (Fig. 3.1) (see below), is never exposed again to the northeast of Hightower. Field photos of the Allatoona fault are shown in Figures 3.11 and 3.12.

Confusion has arisen as a result of the recognition of a late normal fault, potentially with an earlier right-slip, mylonitic foliation, along a railroad cut in downtown Goodwater, Alabama, also referred to as the Goodwater fault (Tull et al., 1985; Drummond, 1986; Allison, 1992; Drummond et al., 1994, 1997; Tull, 2011; Steltenpohl et al., 2013; Barineau et al., 2015). However, this fault is internal to the Allatoona fault allochthon and to the Wedowee Group, with the trace of the Allatoona fault occurring ~600 m northwest of that fault in Goodwater. Along its northeast trace, from just northwest of Goodwater, for ~5 km northeastward, the Allatoona fault trends almost perpendicular to, and abruptly truncates the south-southeast-trending prominent ridges in the Higgins Ferry Group of the Ashland Supergroup.

**Hillabee Thrust**

As the Allatoona fault is traced from the northwest corner of the recess 30 km to the southwest to near Tallapoosa, Georgia, the Hillabee Greenstone appears in the immediate

![Figure 3.11. Boulders showing fault zone features of the Allatoona fault in immediate Talladega belt footwall rocks. S-C fabric within fault zone shown by deformed quartz veins in graphitic phyllonite matrix. Location: 33° 53’ 09.40” N, 84° 57’ 06.55” W.](image)
footwall at the structural top of the greenschist facies Talladega belt (Fig. 3.1). The Hillabee is a >2.6-km-thick bimodal volcanic sequence of tholeiitic metabasalt and subordinate interstratified calc-alkaline metadacite/rhyolite (Tull and Stow, 1980; Tull et al., 2007). It extends discontinuously in the immediate footwall of the eastern Blue Ridge southwestward from Tallapoosa, Georgia, for 230 km to inliers within the Gulf Coastal Plain, but does not extend beyond 12 km to the northeast of Tallapoosa (Figs. 1.2 and 3.1). It is similar in age, thickness, and composition to the Pumpkinvine Creek Formation in the Allatoona fault's hanging wall on the east flank and to the northeast of the recess (McConnell, 1980; Tull et al., 2007), but is at lower greenschist facies as opposed to the lower amphibolite facies of the Pumpkinvine Creek. Their similar ages, geochemical characteristics, and inferred palinspastic positions, however, strongly suggest they formed in the same volcanic basin outboard of the Laurentian margin (Tull et al., 2014; Barineau et al., 2015). The Hillabee, however, was emplaced onto younger rocks (Middle Devonian to earliest Mississippian?) of the Talladega belt much earlier than the formation of the Allatoona fault. The Hillabee was emplaced along a basal cryptic pre-metamorphic (pre ~330 Ma) early Alleghanian fault (Hillabee thrust) (Tull et al., 2007; McClellan et al., 2007; Barineau, 2009; Tull and Barineau, 2012), which, unlike the Allatoona fault, is remarkably concordant to both hanging wall and footwall sequences across its entire extent (>230 km) (Tull et al., 2007; Barineau, 2009; Tull and Barineau, 2012). This fault has a minimum horizontal component of net slip of 18 km (Fig. 3.1).
Hollins Line Thrust Duplex

The Hollins Line fault system (Fig. 1.2) is a regional, post-metamorphic, dextral, transpressional, footwall duplex thrust system of Alleghanian age. It consists of a zone of post-metamorphic imbricate faults that bound large thrust horses of Talladega belt units beneath the high-grade eastern Blue Ridge allochthon (Moore and Tull, 1989; Tull, 1995; Barineau, 2009; Tull and Barineau, 2012) (Fig. 3.1). It consists of (1) a roof thrust (Hollins Line fault), (2) large (many 10’s of km$^2$ in outcrop area) dextrally offset imbricate slices containing the Hillabee Greenstone, the underlying pre-metamorphic Hillabee thrust, and the upper Talladega Group, and (3) a floor thrust (Fig. 3.1). The imbricate faults extend obliquely from the floor to the roof thrust in an en échelon-right sense. Below the floor thrust, the Talladega belt parautochthon also contains the Hillabee Greenstone and the structurally underlying Talladega Group. The roof thrust separates lower greenschist facies footwall rocks from the overlying upper amphibolite facies eastern Blue Ridge (Ashland Supergroup). The imbricate faults and the floor thrust cut the metamorphic fabrics but separate rocks of the same metamorphic grade. The roof thrust (Hollins Line fault) has a minimum horizontal component of net slip of 19 km (Fig. 3.1).

Importantly, the Hollins Line fault shares many characteristics with the Burnt Hickory Ridge fault. It carries the same stratigraphic sequence within its hanging wall (the Ashland/Wedowee/Emuckfaw belt); is in a similar position in the regional kinematic sequence (being the earliest regional thrust after the Hillabee thrust); displays a similar geometry relative to sequences in its hanging and footwalls; and has likely correlative stratigraphy (Hillabee Greenstone vs. Pumpkinvine Creek Formation) in its immediate footwall (Fig. 3.13). Like the Burnt Hickory Ridge fault, the trajectory of the Hollins Line roof thrust was at a much shallower angle than the regional lithostratigraphy and metamorphic planar fabrics, slicing down section in the direction of displacement through >12 km of hanging wall lithostratigraphy (Fig. 3.14). However, in its footwall the Hollins Line roof thrust, like the Burnt Hickory Ridge fault, was closely structurally controlled by an Ordovician metavolcanic sequence (the Hillabee Greenstone vs. Pumpkinvine Creek Formation), remaining subparallel to that unit over a strike distance of > 150 km.

DISCUSSION: “ANOMALOUS(?)” CHARACTERISTICS OF THE BLUE RIDGE CRYSTALLINE THRUST SHEETS

The final event of the Appalachian orogenic “revolution” in the southern and central parts of the orogen involved the latest Paleozoic (Alleghanian) collision of Laurentia with northwest Gondwana (Africa), and is believed to have completed the assembly of Pangea (Pangean collision) (Hatcher et al., 1989; Ziegler, 1989). It was during this continent-continent collision that structures within the Appalachian thrust belt developed, including those described here. The Alleghanian crystalline thrust sheets discussed herein behaved mechanically as “thin-skinned” thrust sheets. In contrast to the foreland thrust sheets and the earlier Hillabee allochthon, they were internally strong, brittle slabs of intact cover sequences that had been previously metamorphosed and more complexly polydeformed during earlier orogenic pulses. The low-angle trajectories of these thrusts were little affected by internal mechanical anisotropies like compositional layering or cleavage/schistosity/gneissosity, or by the presence of older thrusts boundaries.
The oldest thrust in the region, the pre-metamorphic, Lower to Middle Mississippian Hillabee thrust, involved emplacement of a thick (~2.6 km) intact slab of volcanic rocks onto Laurentian shelf rocks, but the overall laterally extensive flat-on-flat geometry of the thrust suggests that neither its hanging nor footwalls were extensively deformed prior to emplacement (Tull et al., 2007; Barineau, 2009; Tull and Barineau, 2012), unlike the other hinterland thrusts described here. Peak metamorphism of all of the crystalline thrust sheets described here was during the early Carboniferous (early Alleghanian). Based on paleontologic (Gastaldo, 1995) and 40Ar/39Ar white mica ages (McClellan et al., 2007; Hames et al., 2007) the Talladega and Hillabee thrust sheets were metamorphosed together ca. 334-321 Ma in the Middle to Late Mississippian (Visean). The youngest plutons that carry the peak metamorphic fabrics in the Ashland/Wedowee/Emuckfaw-New Georgia thrust sheet (sheet D) yield U/Pb zircon ages of 343.4 to 334.6 Ma, also Visean in age (Ingram, 2012). With the exception of the older Hillabee thrust, the timing of the thrusts described here was likely latest Alleghanian (post-Carboniferous: 286-266 Ma) (Hatcher et al., 1989). This means that these thrust sheets were metamorphosed and polydeformed 35 to 70 my. before they were emplaced. Although many workers in the southern Appalachians have traditionally
attributed metamorphism and deformation of the broad “Blue Ridge-Piedmont thrust sheet” to the Taconic and Acadian orogenies (Shanmugam and Lash, 1982; Hatcher, 1987; Higgins et al., 1988; Drake et al., 1989; Hatcher et al., 2007), geochronological data and relative age relationships in the Talladega belt and eastern Blue Ridge of GA and AL indicate that the Alleghanian orogeny was largely responsible for this dynamothermal metamorphism (Unrug and Unrug, 1990; Tull and Groszos, 1990; Tull, 1998; Repetski et al., 2006; McClellan et al., 2007; Tull et al., 2007, 2014), with the southern Appalachian Taconic and Acadian orogenies dominated by magmatic-volcanic activity associated with suprasubduction systems on the Laurentian margin (Ingram, 2012; Holm-Denoma and Das, 2010; Tull et al., 2014; Barineau et al., 2015). Earlier workers previously believed Alleghanian metamorphism to be restricted only to the southeastern-most flank of the orogen (Hatcher et al., 1989).

In general, thrusts propagate from the hinterland to the foreland in an orogenic belt with time (Armstrong and Oriel, 1965; Balley et al, 1966; Boyer and Elliott, 1982), and it has been suggested that the composite Blue Ridge-Piedmont crystalline thrust sheet (the frontal allochthons of which are addressed here), “drove the foreland deformation in front of it” (Hatcher, et al., 1989). In this interpretation, the Blue Ridge-Piedmont thrust sheet became detached at the brittle-ductile transition zone, and ramped upward to the northwest toward the craton along an extensive low-angle detachment at the base of the deforming orogen onto rift and then drift sedimentary rocks, and eventually propagated into higher and higher detachments onto the platform sequences of the foreland. The level of the detachment in southeasternmost foreland thrust sheets in the southern Appalachians is within the Lower Cambrian Chilhowee Group. To the southeast, within the frontal metamorphic allochthon, the detachment is, as expected, at a deeper stratigraphic level, within the Grenville basement and its Neoproterozoic cover (Ocoee Supergroup) in the western Blue Ridge along part of the Tennessee salient of Thomas (1977). Southwest of the Cartersville transverse zone, along the Alabama recess (Thomas, 1977) in the Talladega belt, however, the detachment is at a similar stratigraphic level as that of the eastern foreland thrusts, within Chilhowee-equivalent units (Kahatchee Mountain Group), and locally within the Silurian-Devonian Talladega Group.

Figure 3.14. Geologic map of the northern salient of the Ashland Supergroup showing locations of Allatoona fault footwall cutoffs of the Hollins Line fault, and the latter fault cutting obliquely down section to the northwest through >12 km of Ashland Supergroup lithostratigraphy.
The western Blue Ridge-Talladega belt allochthon and all thrust sheets to its northwest contain Laurentian rifted-margin and/or more cratonward trailing-margin shelf sequences formed on relatively thick continental crust of North American plate (Laurentian) affinity. Importantly however, once the Hillabee Greenstone and the thrust sheets of the eastern Blue Ridge (described in this study) appear in the thrust stack, the detachment had switched to generally a much higher stratigraphic level in Ordovician units. These thrust sheets containing deep-water sequences are interpreted to have made up an extensive Laurentian-margin fringing Middle Ordovician back-arc basin that formed in a palinspathic position southeast (outboard) of the continental margin hinge zone on thinned Laurentian continental crust, and thus at a lower elevation than the older sequences northwest of the hinge zone (Tull et al., 2014). These sequences are now incorporated into major thrust sheets that have undergone tens of kilometers of displacement along thrust planes with very low regional dips.

It appears that with respect to much (most?) of the deformation in the external foreland thrust wedge, all of the frontal hinterland thrusts described here, with the exception of the Hillabee thrust, represent out-of-sequence faulting which propagated through previously deformed and metamorphosed thrust sheets (Morley, 1988). Importantly, these hinterland thrusts post-dated major thrusting and associated deformation in at least the southeastern part of the more cratonward foreland wedge. There appears to be no evidence, particularly given their trajectories, that these crystalline thrusts involved reactivation of earlier thrusts. Nevertheless, such out-of-sequence thrusts are not uncommon in orogenic belts and can be considered part of the normal deformation sequence (Morley, 1986).

REFERENCES


51st Annual Field Trip: Road Log

Day 1, Saturday, October 7, 2017:
We will depart from the meeting headquarters at the Best Western Plus, 104 S Cottage Hill Rd, in Carrollton GA at 8:00am Eastern Time.

- From the parking lot of the Best Western Plus, turn right on South Cottage Hill Rd, then immediately right on Sue Alice Ln.
- Follow Sue Alice Ln ~0.1 mile to its intersection with U.S. Hwy 27 (S. Park St). Turn left (north) on U.S. Hwy 27.
- After ~17.5 miles, turn left (southwest) on Old Bush Mill Rd (~2.5 miles south of the town of Buchanan GA).
- After ~0.5 miles, turn right (west) on Old Ridgeway Rd.
- After ~0.5 miles, turn right (northwest) on Estavanko Loop.
- After ~1.5 miles, continue straight to stay on Macedonia Church Rd.

Figure 4.1. Georgia Geological Society Field Trip 2017. Regional geologic map showing field trip stops, select cities, major highways, major faults and lithotectonic terranes. For map symbology, see Figure 1.2. Adapted from Barineau et al., 2017.
• After ~1.0 miles, turn left (south) onto GA Hwy 120 W.
• The parking area for Stop 1 is ~0.1 miles from the intersection of Macedonia Church Rd and GA Hwy 120W at the site office trailer for the bridge project on GA Hwy 120W at Beach Creek.

STOP 1 (~1 hour): Hillabee Greenstone and Talladega Group of the Talladega belt (Latitude 33° 45' 48.11"N, Longitude 85° 13' 20.72"W).

Clint Barineau (Columbus State University) and Jim Tull (Florida State University)

Micaceous quartzite of the Tally Mountain Quartzite and metavolcanic units of the Hillabee Greenstone are exposed in road cuts on the north and south side of Highway 120 between Buchanan and Tallapoosa, GA (Figs. 4.1 and 4.2). Excavation associated with the Beach Creek bridge project exposed extensive outcrops on the south side of the highway, but retaining walls were being installed in late August/early September, 2017, and thus, these outcrops are not likely to still be exposed on Day 1 of the 2017 GGS field trip. Instead, we will visit one of the outcrops on the north side
of the highway. At this location, chlorite phyllites and micaceous quartzites are exposed in a nearly 100 meter-long outcrop along the road and adjacent hillside. Although detailed work on the Hillabee Greenstone and upper Talladega Group, has largely been conducted in AL, geologic mapping by Robert Weimer (1976), George Heuler (1993) and Martin Balinsky (1997) in Haralson County, GA, reveals similar stratigraphic units to those mapped to the west in neighboring Cleburne County (eastern AL) and farther to the southwest. As described in this guidebook (Tull, Holm-Denoma, and Almutseshy), rocks of the Talladega belt in AL lie in the foot wall of the Hollins Line fault system, whereas stratigraphic equivalents in western GA lie in the foot wall of the younger Allatoona fault – which truncates the Hollins Line near the community of Hightower, AL, ~6 km west of the AL-GA state line. Rocks in the immediate footwall of the Hollins Line in AL and the Allatoona fault immediately east of the AL-GA state line often consist of interlayered epidote-actinolite-chlorite phyllites, massive greenstones (metabasalts and/or metamafic tuffs), and subordinate foliated, micaceous hornblende-quartz metadacites of the Hillabee Greenstone (Fig. 4.3). Where these metavolcanic units are present, they are always found structurally above micaceous quartzites and quartzose sericite-chlorite phyllites of the late Early to early Late Devonian Jemison Chert (AL) and Tally Mountain Quartzite (GA).
The age of these volcanic units was the source of much debate in the 1970’s, 80’s, and 90’s, with field relationships suggesting a Devonian or younger age (e.g. Tull et al., 1978; 1998; Tull and Stow, 1980; Paris and Cook, 1989; Paris, 1990; German, 1990) and radiometric data suggestive of an Ordovician crystallization age for Hillabee metadacite units (Russell, 1978; Russell et al., 1984). Modern single-grain U-Pb isotopic analyses on zircon from metadacites in Clay County, AL, (McClellan and Miller, 2000; McClellan et al., 2005, 2007; Tull et al., 2007), however, confirmed the isotopic ages, indicating the ca. 470 Ma (lowermost Middle Ordovician) Hillabee Greenstone was tectonically emplaced atop the latest Devonian-earliest Mississippian Erin Slate-Jemison Chert of the upper Talladega Group. This same relationship can be observed at Stop 1, with micaceous and occasionally graphitic, fine- to coarse-grained quartzites and quartzose phyllites of the Tally Mountain Quartzite – nearly 300 meters of exposure in roadcuts on the west side of the Beach Creek bridge project – structurally below mafic phyllites and interlayered metadacites of the Hillabee Greenstone. The contact between these two units, the Hillabee thrust, is exposed in outcrops on both sides of the highway at Stop 1. Because the Hillabee Greenstone and Talladega Group, as well as the underlying Sylacauga Marble and Kahatchee Mountain groups (AL) share the same metamorphic grade and fabrics, this fault is
interpreted to be pre-metamorphic in nature. While any fault-related cataclastic/mylonitic fabrics that may have been associated with the Hillabee thrust were erased by Mississippian (ca. 330 Ma) greenschist-facies metamorphism, map relationships between metasedimentary units in the Talladega Group and internal layering in the Hillabee Greenstone (e.g. metadacites, massive sulfide zones) suggest the Hillabee back-arc allochthon was thrust atop rocks of the Talladega Group shelf along a hanging wall-footwall flat that spanned a lateral distance greater than 200 km.

- From the GDOT bridge project office parking lot, turn left (south) on GA Hwy 120W (Buchanan Hwy) towards the city of Tallapoosa, GA.
- After ~4.4 miles, turn left (south) on Head Ave.
- After ~0.8 miles, continue straight across GA Hwy 100S (E. Atlanta St), then immediately left (east) on E. Alabama St.
- After ~200 feet, turn right (south) on GA Hwy 100S (Alewine Ave) towards the town of Bowdon GA.
- After ~15.9 miles, continue straight to stay on Bowdon Tyus Rd.
- After ~5.7 miles, turn left (east) on GA Hwy 5 N.
- Stop 2 is ~2.0 miles east of the intersection of Bowdon Tyus Rd and GA Hwy 5 N, on the south side of the highway, approximately 1700 feet west of the intersection of GA 5 and Needmore Road.

Figure 4.5. Location and generalized geology for Stop 2. Adapted from Barineau et al., 2015.
STOP 2 (~30min): Amphibolite and Felsic Metavolcanic Rocks of the Emuckfaw Group, Josie Leg Formation (Latitude 33° 27' 46.17"N, Longitude 85° 10' 10.65"W).

**Clint Barineau (Columbus State University) and Jim Tull (Florida State University)**

Stop 2 for this field trip is at the same location as Stop 8 for a field trip associated with the 2015 Annual Meeting of the Southeastern Section of the Geological Society of America (Barineau et al., 2015).

Amphibolite is exposed over >300 feet in a road cut on the south side of GA 5 at this location (Figs. 4.4 and 4.5). Like similar amphibolites (e.g., Beaverdam Amphibolite) in the Alabama Ashland-Wedowee-Emuckfaw belt, this amphibolite body is assigned to the Emuckfaw Group (Josie Leg Formation). A sample collected for geochemical analysis plots in the basalt field on the total alkalis vs. silica diagram (Fig. 4.6) of Le Maitre et al. (1989), and exhibits the same geochemical characteristics of the majority of metamafic rocks found throughout the WEDB – moderate enrichment in large ion lithophile elements and a flat rare earth element (REE) pattern typical of suprasubduction-zone mafic rocks. A thin (<0.5 m), fine-grained quartzofeldspathic unit interlayered with the predominant amphibolite lithology is interpreted as a

![Figure 4.6. Total alkalis vs. silica diagram of Le Maitre et al. (1989) with greenstone-amphibolite lithologies of the Wedowee-Emuckfaw-Dahlonega basin (black x's), and the amphibolite and silicic metavolcanic units of lithologies at Stop 2 (red stars). Adapted from Tull et al., 2014.](image-url)
metamorphosed rhyolitic tuff, with a sample collected for geochemical analysis plotting in the rhyolite field on the total alkalis vs. silica diagram (Fig. 4.6) of Le Maitre et al. (1989). Zircon separated from this silicic metavolcanic unit is currently awaiting isotopic analysis at the U.S. Geological Survey in Denver, CO, for the purposes of radiometric dating.

- From Stop 2, drive east on GA Hwy 5 N.
- After ~2.8 miles, turn right (south) onto US Hwy 27.
- After ~10.5 miles, turn right (west) onto Franklin Pkwy.
- After ~1.0 miles, turn right (west) onto Old Field Rd.
- After ~1.1 miles, Stop 3 will be on the left (east) side of Old Field Rd, ~500 feet north of its intersection with GA Hwy 100, northwest of the community of Franklin, GA.

**Figure 4.7.** Location map for Stop 3, ~150 meters north of the intersection of GA Hwy 100 and Old Field Rd, northwest of Franklin, GA.
STOP 3 (~30min): Ropes Creek Amphibolite of the Dadeville complex (Latitude 33° 17' 48.72"N°, Longitude 85° 7' 9.18"W).

Ben Davis (Florida State University)

Figure 4.8. Ropes Creek Amphibolite outcrop at this stop displaying well developed compositional layering. Light-colored layers are interpreted as layers of dacitic meta-tuff.

The exposure at this stop (Fig. 4.7) is in a massive outcrop of amphibolite on the northwest flank of the Dadeville complex, sandwiched between thick lenses of the Ordovician Franklin gneiss, a high-silica granitic orthogneiss not seen at this exposure (Seal and Kish, 1990). Amphibolites in the region of this exposure, on the west limb of the Tallassee synform, and cropping out between the southeast flank of the Brevard Zone and the overlying Agricola Schist in the hinge of the synform in Alabama, have been referred to by previous authors as the Waresville Schist, formerly named the Waresville Amphibolite by Bentley and Neathery (1970). On the east limb of the Tallassee synform, in an identical structural/stratigraphic position beneath the Agricola Schist, are very similar lithologies which have been mapped by essentially all workers as Ropes Creek Amphibolite. In the hinge (closure) of the Tallassee synform just north of the Gulf Coastal Plain in Alabama, each of these lithologically very similar units (Waresville and Ropes Creek) can be mapped directly into one another. For these reasons we consider the Waresville to simply be the upper part of the Ropes Creek Amphibolite, and not distinguishable lithologically or stratigraphically as a separate
unit. This conclusion of the equivalency of these two units has also been made by previous authors (Stow et al., 1984; Seal and Kish, 1990). For these reasons use of "Waresville" as a lithodemic name in the Dadeville complex should be discontinued.

The Ropes Creek Amphibolite was derived from the metamorphism of tholeiitic (low-K) basalt, and major and trace element geochemical analyses suggest formation in a volcanic arc setting. The Ropes Creek Amphibolite is ~9-10 km thick and is the major stratified unit of the Dadeville complex. Ropes Creek units are delicately layered to massive amphibolite (Fig. 4.8) consisting of 0.1-3.0 mm thick alternating hornblende and plagioclase-rich layers, both exhibiting accessory epidote, quartz, garnet, and sphene. Interlayered with the layered and laminated amphibolites of the Ropes Creek are concordant layers of trondhjemite gneiss, ranging from a few centimeters to several meters thick (Fig. 4.8). These are common on the west flank of the Tallassee synform in the environs of this stop (see Seal and Kish, 1990), but have been observed throughout the Ropes Creek stratigraphy. We interpret these trondhjemites to be interlayered dacitic ash deposits within the dominant basalt and basaltic ash of the Ropes Creek Amphibolite. Zircon ages from a felsic layer in the type section of the Ropes Creek Amphibolite in Lee County, AL, range from 318 to 2570 Ma, which suggests that the protolith was volcaniclastic. The best estimate of crystallization age, 444 ± 4 Ma, is the average of the ten oldest 206Pb/238U grains (Farris et al., this guidebook). Although a minimum, this age is compatible with field relations that place the eruption of the Ropes Creek basaltic protoliths early in the Dadeville sequence, and is compatible with the 446-448 Ma ages of the Camp Hill gneiss, which intrudes the Ropes Creek Amphibolite.

Figure 4.9. Location map for Stop 4, ~300 meters southwest of Ridgeway Rd, northwest of the community of Texas, GA.
For further context of the geology at this location, see Stop 6 (p. VII124-126) in Seal and Kish (1990) in the Guidebook for Field Trip VII, Southeastern Section of the Geological Society of America, edited by Steltenpohl, M.G. Kish, S. A., and Neilson, M. J. Their Stop 6 is ~650 meters northeast of our Stop 3.

- From Stop 3 on Old Field Rd, drive south to GA Hwy 100 (~500 ft).
- Turn right (northwest) on GA Hwy 100.
- After ~1.0 miles, turn left (west) on Frolona Rd.
- After ~8.2 miles, turn left (south) on Ridgeway Rd.
- After ~2.5 miles park on the right of way, across from an unmarked dirt track on the left (west) side of Ridgeway Rd.
- Stop 4 is ~300 meters west of Ridgeway Rd on a recently timbered hillside.

**Stop 4 (~1 hour): Granite-bearing metadiamictite (pebble mudstone) in the Emuckfaw Group** *(Latitude 33°17’ 14.86” N, Longitude 085°14’ 08.62” W)*.

***Jim Tull and Nick Carpenter (Florida State University)***

This stop is accessible by walking ~200 m southward along the path from the paved road (Ridgeway Rd) and then turning westward and walking ~100 m along the property line to the large clear-cut area which contains the stop exposures (scattered over ~10,000 m²). At this stop (Fig. 4.9) we will observe a unique lithofacies within the Timbergut Formation of the Emuckfaw Group, a metadiamictite or olistostromal deposit (pebbly mudstone). Nearby units contain kyanite. This ~60 m thick unit can be traced along strike for at least 20 km from the Roanoke East quadrangle in Alabama, northeastward to the Frolona quadrangle in Georgia. It extends continuously from rocks mapped within the Emuckfaw Group in Alabama, into units in Georgia that have been previously mapped as the Bill Arp Formation of the Sandy Springs Group. This is an example of the inherent challenges associated with reconciling stratigraphic nomenclature when tracing...
geologic units across the state line between AL and GA, which we attempt to address in this guidebook. The Emuckfaw was formally named in 1975 (Neathery and Reynolds), whereas the Sandy Springs was formally named in 1978 (Higgins and McConnell), and the Bill Arp was formally named in 1984 (McConnell and Abrams). It is our practice to apply the formal name to a lithodemic unit based on the chronological order in which different names were proposed. In this instance, for example, we apply the term Emuckfaw to the same units in Georgia which were later named “Sandy Springs” and “Bill Arp,” thus eliminating the chronologically later terminology.

This metadiamicite tends to form large linear outcrops, some extending up to 7 m above the surrounding ground surface (Fig. 4.10 A). The lithofacies consists of a sandy quartz-feldspar rich muscovite-biotite-garnet schist matrix, hosting an array of gravel to cobble-sized granitoid and feldspar fragments (Fig. 4.10 B-E). These rocks are mostly unsorted, ungraded, and unbedded, with the clasts “floating” in the matrix, and were probably emplaced by a gravity-flow mechanism such as debris flow in a submarine fan-like environment. These types of gravity flows tend to move very rapidly, entraining large consolidated clasts within their mass. This facies is interlayered with other turbidite facies including metagraywacke, exposures of which can be seen along the trail leading into the stop from the highway.

One feature of interest here are the granitoid clasts (Fig. 4.10D), some of which contain a gneissic fabric that predates their deposition in the diamicite. All of the granitoid bodies in the Emuckfaw (and adjacent sequences) are intrusive into it and range from Middle Ordovician to Mississippian in age (see discussion by Barineau et al., this guidebook). Thus, it is highly likely that these pre-Ordovician clasts are fragments of Grenville basement, derived from relatively nearby basement exposures. This, combined with
the plethora of detrital zircons in this sequence and other isotopic evidence presented in Barineau et al. (this guidebook) supports the idea that the Wedowee-Emuckfaw-Dahlonega basin was deposited in a region underlain (at least partially) by Grenvillian basement, and was originally a part of the distal rifted Laurentian margin. Relatively

**Figure 4.10.** Above and previous two pages. A.) Typical “shark fin” exposure of diamicite unit. (Location: 33°14′07.48 N, 85°15′49.90 W) B & C) Exposure of diamicite facies containing granitoid and feldspar clasts at this stop. (Location: 33°17′14.86″ N, 85°14′08.62″ W) D) Gneissic granitoid pebble in diamicite facies. (Location: 33°15′07.17″ N, 85°17′04.13″ W) E) Upper part of photo (above pen) shows a small channel in diamicite facies with possible cross-beds at this stop. An S-C fabric can be observed in the underlying pebbly mudstone. (Location: 33°15′07.17″ N, 85°17′04.13″ W).

**Figure 4.11.** Location map for Stop 5, ~0.4 miles east of the AL-GA state line near the community of Mason, GA.
proximal basement uplifts (unseen at today’s erosional level) were likely present during deposition of this strata.

A second interesting feature that can be observed at this stop is the imprint of a shear fabric (shearbands) superimposed on the original metamorphic schistosity. This produces an S-C fabric (Fig. 4.10E) found throughout these outcrops, indicative of dextral shear. In most cases, the shear fabric is not intense enough to disrupt the primary sedimentary features.

- Return to Ridgeway Rd along the unmarked dirt road and drive south on Ridgeway Rd.
- After ~1.5 miles, turn right (west) on Awbreys Gin Rd.
- After ~1.6 miles (~0.3 miles west of the intersection of Awbreys Gin Rd and Mason Rd), park on the right of way.
- Stop 5, a series of metasandstone exposures in an unnamed branch of Cedar Creek, is ~80 meters southeast of Awbreys Gin Rd.

**Stop 5 (~30 minutes): Large-scale sigmoidal cross-stratification in an Emuckfaw Group metasandstone unit** *(Latitude 33°16’ 03.63 N, Longitude 85°15’23.27 W)*

Jim Tull (Florida State University)

![Figure 4.12. Large-scale (~65 cm thick) cross-bed set with sigmoidal (tangential) foreset beds in Emuckfaw Group metasandstone. The current direction is S 45°E (135°).](image-url)
This stop is located on a southeast-flowing branch of Cedar Creek (Fig. 4.11), accessible from a dirt road (Awbreys Gin Road) 0.3 miles to the northeast of the crossroads in the community of Mason (on the Alabama/Georgia line). Leave the road and walk down the creek ~80 m to where the stream turns southeastward and cuts downward through a small ravine. The exposures here are kyanite-grade metasandstone, metagraywacke, and garnet mica schist of the Emuckfaw Group’s Timbergut Formation. Contained in this section on an upstream-facing outcrop near water level is one of the very rare exposures anywhere in the Appalachian eastern Blue Ridge that contains well preserved cross stratification (Fig. 4.12). Here we can observe a large-scale (65 cm thick) cross-bed set with sigmoidal (tangential) foreset beds in metasandstone. The current direction (with beds rotated to horizontal along their strike line) is S 45°E (135°), or perpendicular to both regional strike and to the likely orientation of the basin margin (cratonal Laurentia), implying a source to the northwest.

The dominant depositional mechanisms for the metasedimentary rocks that we have studied over large areas of the eastern Blue Ridge appear to have involved suspended load gravity flow (e.g. turbidity currents, debris flows). This is a rare example in this region of a traction mechanism (possibly deep-water tidal currents in

Figure 4.13. Thinly layered metasandstone-metawacke units within the Emuckfaw at Stop 5. Note the sinistral, synmetamorphic, isoclinal flow fold (red dashed line, \( F_1 \)). Rock hammer for scale.
the lower flow regime) involving saltation related to subaqueous bedforms like megaripples.

Immediately downstream from the cross stratification exposure are thinly layered metasandstone/metagraywacke exposures which contain sinistral, mesoscopic, synmetamorphic, isoclinal (F1) flow (similar) folds (Fig. 4.13). A younger, more symmetric mesoscopic fold set affecting the schistosity can be observed along strike on the west bank of the stream. This exposure is ~300 m stratigraphically above the diamictite (olistrosomal) unit seen at Stop 4.

- Return to Awbreys Gin Rd.
- Stop 5 concludes Day 1 of the 2017 Georgia Geological Society Field Trip. We will return to the trip headquarters in Carrollton, GA (~45 minutes).

END OF DAY 1.
DAY 2, Sunday, October 9, 2016:

We will depart from the meeting headquarters at the Best Western Plus, 104 S Cottage Hill Rd, in Carrollton GA at 8:00am Eastern Time.

- From the parking lot of the Best Western Plus, turn left (west) on S. Cottage Hill Rd.
- After ~0.3 miles, turn left (west) on Cottage Hill Rd.
- After ~0.3 miles, turn right (north) on Tabernacle Dr.
- After ~0.2 miles, turn right (east) on GA Hwy 166.
- After ~4.2 miles, turn right (east) to stay on GA Hwy 166 (Bankhead Hwy).
- After ~2.9 miles, continue straight on GA Hwy 61 (Carrollton-Villa Rica Hwy).
- After ~8.0 miles, turn right (east) on S. Carroll Rd.
- After ~1.6 miles, turn left (northwest) on Main St.
- After ~0.4 miles, turn right (north) on GA Hwy 61 (Dallas Hwy).
- After ~14.5 miles, turn left (northwest) on Merchants Dr.
- After ~0.7 miles, continue straight to merge onto E. Memorial Dr.
- After ~0.3 miles, turn right (north) on GA Hwy 61 (Cartersville Hwy).
- After ~3.7 miles, turn left (west) on S. High Shoals Rd.
- After ~2.1 miles, at the intersection of High Shoals Rd and Raccoon Creek Rd, park on the right of way immediately west of High Shoals Baptist Church.
• Hike ~60 meters west along High Shoals Rd to the parking area for High Shoals Falls and the location of Stop 6. High Shoals Falls is ~200 meters downhill from the High Shoals Falls parking area.

Stop 6 (~1 hour): Pumpkinvine Creek and Canton formations at High Shoals Falls (Latitude 33° 59' 1.02"N, Longitude 84° 53' 36.90"W).

Clint Barineau (Columbus State University)

Located ~8 km northwest of Dallas, GA, High Shoals Falls (Figs. 4.14 and 4.15) lies on an unnamed branch of Raccoon Creek, ~15 meters southeast (upstream) of the contact between amphibolites of the Pumpkinvine Creek Formation and variably garnetiferous quartzofeldspathic schist/paragneiss of the Canton Formation (Fig. 4.16). The falls have carved out a small pool where they flow over a ~10 meter cliff of amphibolite (Fig. 4.14). The contact between the amphibolite and more easily weathered quartzofeldspathic schist/paragneiss is located a few meters northwest (downstream) of the pool (Fig. 4.14). Amphibolites of the Pumpkinvine Creek Formation are predominately composed of hornblende and plagioclase, but also include variable epidote and garnet as well as retrograde chlorite-actinolite. These
units are often compositionally layered and relict amygdules have been reported at other locations. Amphibolites of the Pumpkinvine Creek Formation plot overwhelmingly as basalts on a number of geochemical discriminant diagrams and have trace element compositional patterns intermediate between normal mid-ocean ridge basalt (N-MORB) and arc basalts – including modest enrichment of large-ion lithophile elements (LILE) and a negative Ta-Nb anomaly typical of backarc-basin basalts (Holm-Denoma and Das, 2010). U-Pb data from interlayered felsic metavolcanics (informal Galts Ferry gneiss) indicate a Middle Ordovician (ca. 460 Ma) crystallization age for the protoliths of this bimodal metavolcanic sequence.

The predominant foliation (S₁) at this location generally parallels compositional layering within and between the metavolcanic and metaclastic units (S₀), and is oriented ~N45E 35SE (045° 35). Amphibolites exposed in the cliff face, however, exhibit complex intrafolial deformation, which commonly disrupts the predominant foliation. From just below the pool, at the basal contact of the Pumpkinvine Creek Formation (Fig. 4.15), metaclastic units of the Canton Formation are exposed in the creek bed for another 15-20 meters downstream (northwest). Canton Formation

![Figure 4.16. Geologic map of the area in the vicinity of Stop 6. The contact between amphibolites of the Pumpkinvine Creek Formation and the underlying Canton Formation lies ~15 meters downstream (northwest) of the falls.](image-url)
metasedimentary units exposed here are quartzofeldspathic, with the feldspar component consisting of microcline, plagioclase and untwinned feldspar.

- Return to the parking lot of High Shoals Falls.
- Drive east on High Shoals Falls Rd.
- After ~1.4 miles, turn right (southwest) on Johnny Monk Rd.
- After ~2.7 miles, turn right (west) on McPherson Church Rd.
- Cross over the Silver Comet biking-hiking trail after ~0.9 miles.
- Park on the left (south) side of McPherson Church Rd on the immediate west side of the Silver Comet trail.
- Stop 7 is on the eastern side of the Silver Comet trail, immediately north of McPherson Church Rd.

**Figure 4.17.** Regional geologic map around the Mulberry Rock recess showing the locations of Stops 6 and 7, as well as major highways and cities.
Stop 7 (~1 hour): Mulberry Rock Gneiss at McPherson Church Rd and the Silver Comet Trail (Latitude 33° 56' 42.94"N, Longitude 84° 54' 59.43"W).

Jim Tull (Florida State University) and Clint Barineau (Columbus State University)

The Silver Comet trail, paved atop an abandoned CSX railway line, runs through Cobb, Paulding, and Polk Counties, GA, and is commonly used for hiking and biking. The 61.5 mile trail begins in Smyrna, GA and ends at Cedartown, GA near the AL-GA state line. In this section of the trail, the former railway cut through a number of outcrops of the Mulberry Rock Gneiss (Figs 4.17 and 4.18). The Mulberry Rock Gneiss (see Tull et al., this guidebook) lies in a series of distinct structural blocks (horses) between the eastern and western Blue Ridge terranes (Fig. 4.17). Horses containing the Mulberry Rock Gneiss, typically no more than 300 meters in thickness, lie between the Allatoona fault and Mulberry Rock fault (likely equivalent of the Burnt Hickory Ridge fault) at its structural top and base, respectively (Figs. 4.17 and 4.18).

The Mulberry Rock Gneiss is a foliated, garnet-bearing, biotite muscovite granitic gneiss (Figs. 4.19, 4.20 and 4.21) dominated by subequal amounts of quartz, plagioclase, and K-feldspar, with zircon, epidote, and allanite present as accessory phases (Holm-Denoma, 2006).
Mineralogically, the Mulberry Rock Gneiss plots as a granite (Fig. 4.22). At this location (Fig. 4.20), the Mulberry Rock Gneiss is a foliated (S₁ N14E 50SE/014° 50), medium to coarse-grained muscovite biotite orthogneiss (Figs. 4.19, 4.20 and 4.21). Originally interpreted as a Mesoproterozoic, Grenville pluton, laser ablation ICPMS and ion microprobe (SHRIMP) zircon U/Pb systematics of the gneiss suggest a crystallization age between 432 and 450 Ma (P.E. Mueller and R.Das, written communication).

Figure 4.20. Outcrop of Mulberry Rock Gneiss on the Silver Comet Trail, Stop 7.

Figure 4.21. Coarse-grained Mulberry Rock Gneiss at Stop 7. Camera lens cap for scale.

Figure 4.22. Albite (Ab), Anorthite (An), Orthoclase (Or) feldspar classification diagram of Barker, 1979. Mulberry Rock Gneiss samples (squares) plot overwhelmingly in the granite field. Adapted from Holm-Denoma, 2006.
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BARINEAU AND TULL


END OF DAY 2.-END OF FIELDTRIP
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