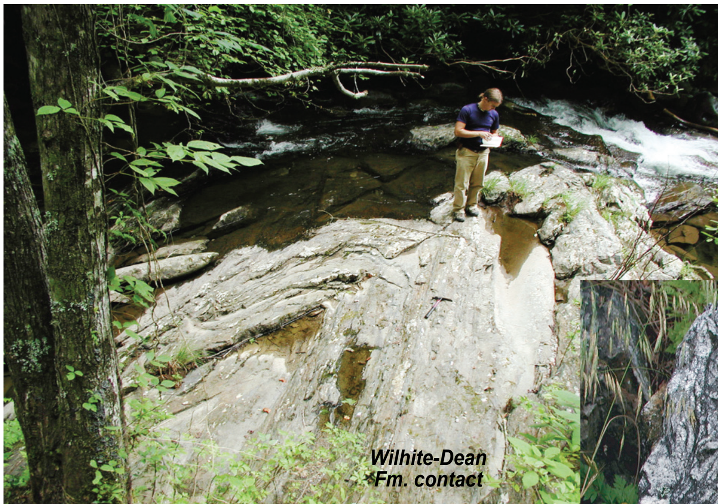
Blue Ridge Geology Geotraverse East of the Great Smoky Mountains National Park, Western North Carolina



Wilhite-Dean Fm. contact

Carolina Geological Society Annual Field Trip

November 5-6, 2005

Cartoogechaye terrane migmatite

Otto Fm.

Guidebook Editors:

Robert D. Hatcher, Jr. and Arthur J. Merschat

Ashe-Tallulah Falls Fm. migmatite

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(alphabetical)

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Rebecca Latham	Mathew A. Massey
Arthur J. Merschat	David P. Moecher
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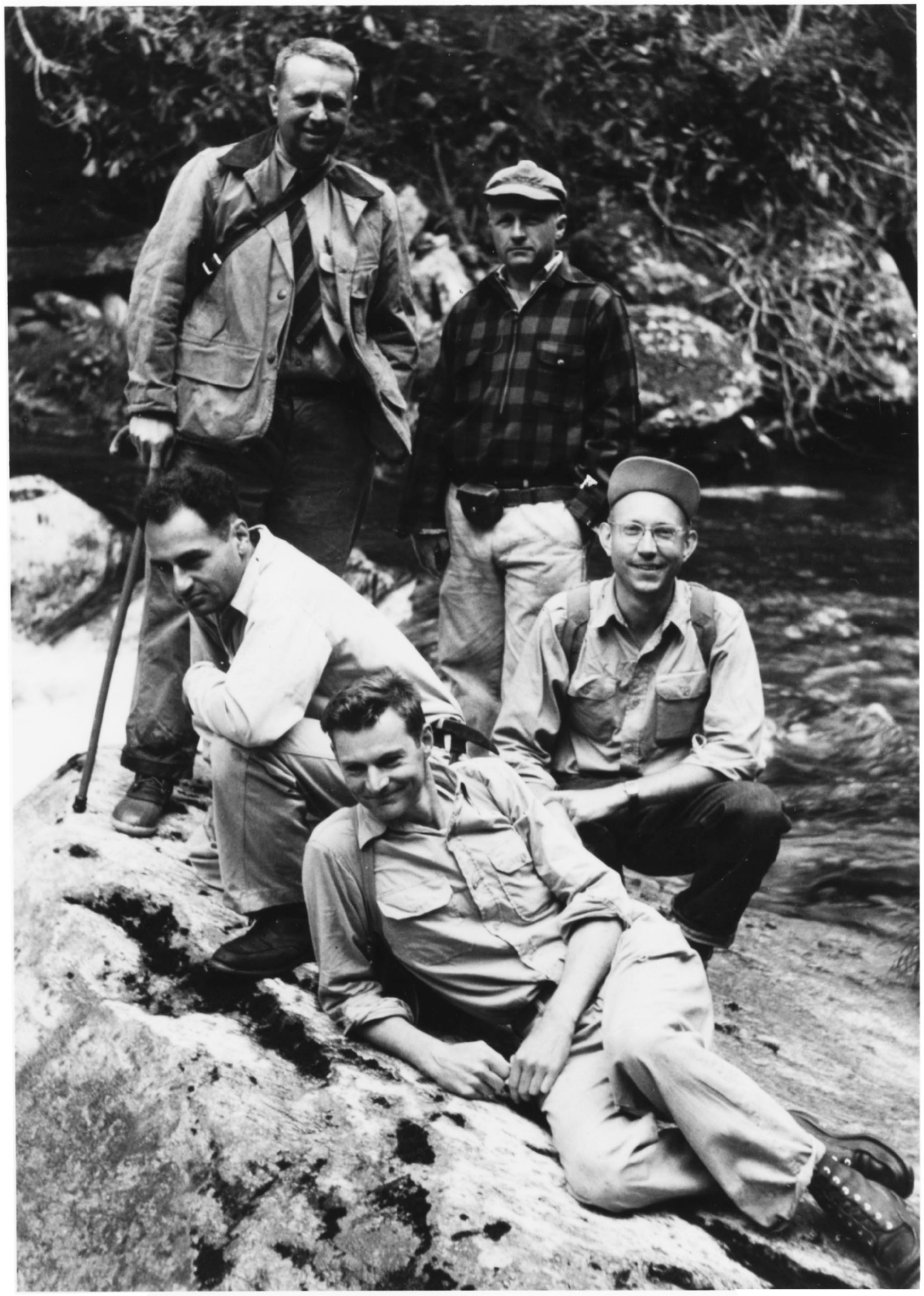

The threshold of modern Blue Ridge geology

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Carolina Geological Society
Annual Field Trip
November 5-6, 2005

Blue Ridge Primer

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INTRODUCTION

The western North Carolina Blue Ridge is the focus of the 2005 Carolina Geological Society field trip. Numerous previous CGS field trips have examined Blue Ridge geology, beginning with the trip by King et al. (1952) in the Great Smoky Mountains National Park (GSMNP) and vicinity. Subsequent field trips have been conducted in the Carolinas and northeast Georgia Blue Ridge by Bryant and Reed (1960), Power and Forrest (1971), Hatcher (1976), Kish et al. (1975), Wiener and Merschat (1988), and Kish (1991), as well as several Georgia Geological Society field trips by Hatcher (1974), Hartley (1974), Fritz and La Tour (1988), Fritz et al. (1989), and Costello et al. (2002). The value of these field trips has been to review the results of mostly detailed geologic mapping in different parts of the Carolinas and northeastern Georgia Blue Ridge. This mapping built islands of knowledge that have been joined and expanded to the present day. This field trip will attempt to review detailed geologic mapping to the south and southwest (e.g., Hatcher, 1980; Eckert, 1984; Eckert et al., 1989; Quinn, 1991), with that to the west (e.g., Southworth et al., 2005; Thigpen, 2005), which adjoins detailed mapping to the north and northeast (Wiener and Merschat, 1988; Edelman, unpublished; Cattanach and Merschat, this guidebook) (Fig. 1). The results summarized in the early CGS trip by King et al. (1952) were subsequently published by King et al. (1958), Hamilton (1961), Hadley and Goldsmith (1963), King (1964), and Neuman and Nelson (1965). This has to be considered the beginning of modern Blue Ridge geology, because of the accomplishments of this group, most of which remain unchallenged today because of the quality of their field work. USGS geologists have recently completed another cycle of work in the GSMNP and vicinity (Schultz, 1998, 1999; Schultz and Southworth, 1999; Southworth, 1995; Southworth et al., 1999; Southworth et al., 2005; Southworth et al., this guidebook). The work currently being completed north of I-40 by North Carolina Geological Survey geologists (e.g., Wiener and Merschat, 1988; Merschat et al., 2005 unpublished; Cattanach et al., 2005 unpublished; Cattanach and Merschat, this guidebook) has also provided ties farther north into the work by King and Ferguson (1960), Ferguson and Jewell (1951), and other work to the southwest in the mountains along the Tennessee-North Carolina state line and to the northeast and east (e.g., Oriel, 1950; Nelson, 1972; Keller, 1980; Dabbagh, 1975; Burr, 1997, unpublished; Merschat, 1977, 1990, 2003 unpublished; Morrow, 1978; Merschat and Wiener, 1988, 1991 unpublished; Carter and Wiener, 1999; Burr and Peterson, 2002, unpublished; Carter et al., 2002, unpublished; Merschat et al., 2002a, 2002b, unpublished; Peterson and Burr, 2002, unpublished; Merschat and Cattanach, 2004, unpublished; Berquist, 2003, unpublished), as well as the work by Bryant and Reed (1970) in the Grandfather Mountain window area. In addition, mapping by Livingston (1966) and McNiff (1967) in the southeastern North Carolina Blue Ridge provided new information and raised issues that could be addressed later; detailed maps by Horton (1982), Mohr (1975), and Ausburn et al. (1998) provide greater insight into Blue Ridge geology.

Our purpose here is to summarize some of the results and problems that have arisen based on current knowledge of bedrock and surficial geology of the Carolinas-northeast Georgia Blue Ridge, and the state-of-the-art geochronologic data that have been collected over the past 8-10 years that are not discussed in other articles in this guidebook (Tables 1 through 5). The current cycle of detailed geologic mapping began in the late 1960s by RDH and Clemson University undergraduates Louis L. Acker, James E. Wright, Jr., Hoke D. Petrie, Steven C. Godfrey, H. Brad Hubbard, and Gregory M. Bryan, and Donald T. Phillips, who assisted RDH in mapping the southern half of the Toxaway dome (Hatcher, 1977; Hatcher et al., 2004). During that time most of the South Carolina Blue Ridge and several quadrangles in the Georgia and North Carolina Blue Ridge were mapped. In the late 1970s RDH began work with graduate students and thus expanded the completion of detailed mapping in northeastern Georgia and southwestern North Carolina (e.g., Petty, 1982; Ausburn, 1983; Eckert, 1984; Farrell, 1986; McClellan, 1988; Stieve, 1989; Quinn, 1991; Hopson, 1994; Lamb, 2001). In addition, R. David Dallmeyer (University of Georgia) directed several masters theses involving detailed geologic mapping

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Hatcher, R. D., Jr., Merschat, A. J., and Thigpen, J. R.,, 2005, Blue Ridge Primer, *in* Hatcher, R. D., Jr., and Merschat, A. J., eds., Blue Ridge Geology Geotraverse East of the Great Smoky Mountains National Park, Western North Carolina: North Carolina Geological Survey, Carolina Geological Society Annual Field Trip Guidebook, p. 1-24.

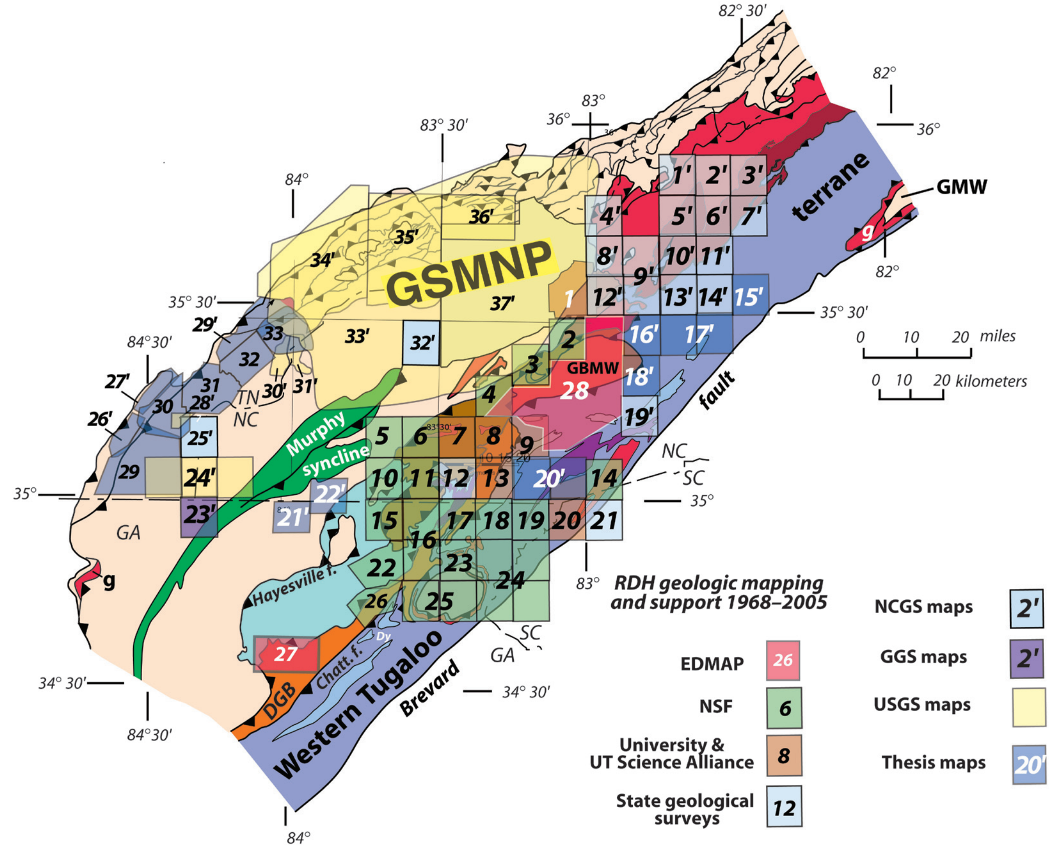

Figure 1. Part of the Carolinas, Georgia, and Tennessee Blue Ridge tectonic map and index of completed mapping in the area of the 2005 Carolina Geological Society field trip. GSMNP–Maps of the Great Smoky Mountains National Park and vicinity by Hamilton (1961), Hadley and Goldsmith (1963), King (1964), Neuman and Nelson (1965), King et al. (1958), Mohr (1973), and Southworth et al. (2005). Geologic maps completed by RDH or under his direction are indicated by unornamented numbers (with apologies for the uneven number system). Geologic maps produced by others are indicated by a number prime (X'). 1–Montes (1997). 2–Hazelwood quad, NC, S. H. Edelman unpublished. 3–Quinn (1991). 4–Greens Creek quad, NC, A. Burns, unpublished. 5–Ausburn et al. (1998). 6–Eckert (1984). 7–Franklin, NC, quad J. O. Eckert, Jr., 2005 unpublished; F. G. Lesure and Force (1993). 11–Rainbow springs, NC, quad; R.D. Hatcher, Jr., 2005 unpublished. 12–Hatcher (1980). 13–Lamb (2001). 14–Reid, NC-SC quad; partially mapped by Mark Hartford, unpublished, and Merschat et al., 2003. 8–Corbin Knob, NC, quad J.O. Eckert, Jr., and S. P. Yurkovich, 2005 unpublished. 9–Glenville, NC, quad; partially mapped by D. W. Stahr, II, proposed for completion with EDMAP support. 10–Shooting Creek, NC, quad; R.D. Hatcher, Jr., 2005 unpublished. 15–E. A. McClellan (1988). 16–Hopson (1994). 17–Dillard, GA-NC; L.L. Acker and R. D. Hatcher, Jr., 2002 unpublished. 18– Rabun Bald, GA-NC; L.L. Acker and R. D. Hatcher, Jr., 2003 unpublished. 19–Satolah, GA-NC quad; R. D. Hatcher, Jr., unpub. map. 20–Tamassee, SC-GA quad; R. D. Hatcher, Jr., L.L. Acker, and D. T. Phillips, II, unpub. map. 21–Hatcher and Acker (1984). 22–Farrell (1986). 23–Tiger, GA, quad R.D. Hatcher, Jr., unpub. map. 24–Hatcher et al. (in prep.). 25–Stieve (1989). 26–Settles (2002). 27–EDMAP-supported mapping projects by Shawna Cyphers, Arthur J. Merschat, and Donald W. Stahr, in progress.. 29–Costello (1993). 30–Carter (1994). 31–Martin (1997). 32–Geddes (1995). 34–Thigpen (2005).
1'–Carter et al. (2002 unpublished). 2'–Merschat et al. (2002a, 2002b). 3'–Merschat (2003). 4'–Merschat and Carter (2002). 5'–Peterson and Burr (2000 unpublished) and Berquist (2003). 6'–Merschat (1977). 7'–Merschat (1993). 8'–Carter and Wiener (2002). 9'–Merschat and Wiener (1988). 10'–Merschat and Cattanach (2004 unpublished). 11'–Burr and Peterson (2003). 12'–Merschat and Wiener (1991). 13'–Burr (1997 unpublished). 14'–Miller and Fryer (2002). 15'– Nelson (1972). 16'–Morrow (1978). 17'–Dabbagh (1975). 18'–Acker (1982). 19'–Horton (1982). 20'–Livingston (1966) and McKniff (1967). 22'–Wooten (1980). 21'–Shellebarger (1980). 23'–Hurst (1955). 24'. Hernon (1964). 25'–Merschat and Hale (1973). 26'–Rackley (1951). 27'–Phillips (1951). 28'–Hale (1974). 29'–Wiener unpublished map. 30'– Slack et al., (1984). 31'–Lesure et al., (1977). 32'–Mohr (1975). 33'–Southworth et al. (2005). 34'–Neuman and Nelson (1965). 35'–King (1964). 36'–Hamilton (1961). 37'–Hadley and Goldsmith (1963).

along the North Carolina-Georgia line west of the area where we were working (Shellebarger, 1980; Wooten, 1980; Gillon, 1982), and these data increased the size of the island that we had begun to build. Hadley (1945) earlier mapped the Buck Creek ultramafic body, and Hartley (1973) also mapped the Lake Chatuge mafic-ultramafic complex along the Georgia-North Carolina border in detail. More recently, RDH directed graduate students research involving detailed geologic mapping in the western Blue Ridge of southeastern Tennessee and northeastern Georgia (Costello, 1984, 1993; Carter, 1994; Geddes, 1995; Martin, 1997; Thigpen, 2005). The U.S. Geological Survey has now also completed a second cycle of mapping in the GSMNP and vicinity (Southworth et al., this guidebook), revising the work that King et al. completed in the 1960s. Consequently, today's knowledge of the field relationships and geometries of rock units and tectonic boundaries in this region is very good because of detailed geologic mapping over the last three decades in the eastern, central, and western Blue Ridge of the Carolinas, northeast Georgia, and Tennessee. In parallel and more recently, several geochemical studies have been completed, particularly of mafic and ultramafic rocks, some of which are presented elsewhere, and in this guidebook (McElhaney and McSween,

Table 1. Modern geochronologic data for Blue Ridge external massifs

Terrane	Massif	Rock Unit	Lat. & Long.	Age (Ma)	Inheritance	Source
Mars Hill terrane	(?)	Carvers Gap RM-1 (NC)	36º 06.2 N 82º 06.7W	1765 ± 140		Carrigan et al., 2003
		NC-49	36º 06.1 N 82º 07.5 W	1209±22		
		NC-81	36º 06.3 N 82º 07.6 W	> 1.6 Ga		
	(?)	Cloudland gneiss* RM-CLG (NC)	36º 06.2 N 82º 08.0 W		1.35-1.85 Ga	Carrigan et al., 2003
		NC-28	36º 06.2 N 82º 08.0W		0.8(?)-1.4	
	(?)	Felsic orthogneiss (Carvers Gap gneiss) RM 21 (NC)	(?)	1198±26	—	Ownby et al., 2004
		RM 38 (NC)		1200±26	(1) 1511±69 Ma	
		MBCL4 (NC)		1257±26	—	
		RM30 (NC)		1190±19	(6) 1320-1672 Ma	
		RM30B (TN)		1207±32	(4) 1273-1500, 1919, & 2653 Ma	
	(7)	Mafix opx gneiss RM30C (NC)	(?)	1209±35		Ownby et al., 2004
	(?)	Bakersville metagabbro gar-am. (NC)	(?)	728±16		Ownby et al., 2004
Laurentian Margin/ western Blue Ridge	Elk Park	Cranberry Gneiss (NC)	36º 10.7 N 81º 55.4 W	1192±11		Carrigan et al., 2003
	Watauga	Watauga River Gneiss (NC)	36º 18.8N 81º 51.1 W	1158±9		Carrigan et al., 2003
	Globe	Blowing Rock Gneiss (NC)	36º 11.1 N 81º 39.1 W	1181±14	5 cores, mean ~1084 Ma	Carrigan et al., 2003
	French Broad	Max Patch granite (NC)	(?)	1020±30		Berquist, 2005; Berquist et al., this guidebook
		Spring Creek granitoid (NC)	(?)	1170±10		Berquist, 2005; Berquist et al., this guidebook
		Indian Grave protomylonite (NC)	(?)	1050±20		Berquist, 2005; Berquist et al., this guidebook
		Spillcorn protomylonite (NC)	(?)	1050±10	(1) ~1250 Ma	Berquist, 2005; Berquist et al., this guidebook
		Sam Gap thrust sheet P25C (NC) P25 D (NC)	(?)	1060±20 1070±20		Berquist, 2005; Berquist et al., this guidebook
		Fork Ridge thrust sheet P32A (NC)	(?)	1380±60		Berquist, 2005; Berquist et al., this guidebook
		Stone Mountain thrust sheet P28 (TN) P34 (TN)	(?)	1170±20 1130±40		Berquist, 2005; Berquist et al., this guidebook
		Pardee Point thrust sheet PP3A (TN)	(?)	1070±20		Berquist, 2005; Berquist et al., this guidebook
Georgia Blue Ridge		Corbin Granite		1106 ± 13		Heatherington et al., 1996

1983; Hatcher et al., 1984; Swanson et al., this guidebook; Ryan et al., this guidebook; Warner and Hepler, this guidebook). Southern Appalachian mafic and ultramafic rocks geology was also summarized in Misra and Keller (1978) and Misra and McSween (1984). Studies of eastern Blue Ridge granitic rocks geochemistry by Miller et al. (1997, 1998, 2000), Bream et al. (2004), and others has broadened our understanding of the origin of these plutons. A large amount of state-of-the-art geochronologic data have appeared during the last 8 to 10 years (Tables 1–5) that have had a profound impact on our understanding of Blue Ridge tectonic history (e.g., Miller et al., 2000; Thomas, 2001; Mapes, 2002; McDowell et al., 2002; Bream, 2003; Carrigan et al., 2003).

Table 2. Modern geochronologic data for Blue Ridge internal massifs.

Terrane	Massif	Rock Unit	Lat. & Long.	Age (Ma)	Inheritance	Source
Tugaloo	Tallulah Falls dome	Wiley Gneiss (GA)	34º 48.2 N 83º 25.3 W	1159±19	—	Carrigan et al., 2003
	Tallulah Falls dome	Sutton Creek Gneiss (GA)	34º 44.2 N 83º 34.5 W	1156±23	—	Carrigan et al., 2003
	Toxaway dome	Toxaway Gneiss TOX-1 (SC) TOX-1B (SC) G-MA-2 (SC) B-TB-3 (SC) TXFL (NC)	 35º 0026 N 83º 04.29 W 35º 00.4 N 82º 59.6 W 35º 07.2 N 82º 55.5 W	 1151±17 1149±56 1157±36 1141±148 1149±32	—	Carrigan et al., 2003
	Tallulah Falls dome	Wolf Creek Gneiss	34.8332° N 83.4099° W	1129 ± 23		Bream et al., 2004
Cartoogechaye	Trimont Ridge	Trimont Ridge felsic gneiss (NC)	(?)	1103±69 (207/206)	—	Bream, 2003; Hatcher et al., 2004

Table 3. Modern geochronologic data for Blue Ridge detrital zircons.

Terrane	Sample	Location	Ages (Ga)	Source
Laurentian Margin/ western Blue Ridge	Mineral Bluff Fm. metass & phyllite (NC)	35º 05.715N 84º 1.058W	0.85, 1.0, 1.1, 1.3, 1.4, 1.5, 1.6, 1.7, 1.8, 1.9, 2.1, & 2.7	Merschat, unpub. data
	Walden Creek Grp. Sandsuck Fm. cgl. (TN)	35.548ºN 84.050ºW	1.1 ,1.2, 1.3,& 1.4	Bream et al., 2004
	Walden Creek Grp. Sandsuck Fm. ss. (TN)	35.248ºN 84.444ºW	1.1, 1.2, 1.3, & 1.4	Bream et al., 2004
	Great Smoky Grp. Deam Fm. ss (TN)	35.336ºN 84.296ºW	1.1, 1.2, 1.3, & 1.4	Bream et al., 2004
	Snowbird Grp. Longarm Quartzite (NC)	35.702ºN 83.040ºW	1.1, 1.2, 1.3, 1.4 7 1.5	Bream et al., 2004
Cowrock	Coleman River Fm. (NC)	34.908 83.612	1.1, 1.2, 1.3, 1.4, 1.6, & 2.7	Bream et al., 2004
Cartoogechaye	Winding Stair Gap (NC)	35.1224ºN 83.5435ºW	0.8, 1.0, 1.2	Miller et al., 1998
	Chunky Gal Mountain (NC)	35.0588ºN 83.6203ºW	0.7, 0.9, 1.0. 1.1, 1.2, & 1.3	Merschat, unpub. data
	Wolfpen migmatite (NC)	35.4709ºN 82.8850ºw	0.95, 1.1, 1.2, 1.7, & 1.8	Merschat, unpub. data
Dahlonega Gold Belt	Otto Fm. (GA)	34.560 84.074	1.1, 1.2, 1.3, 1.4, & 2.7	Bream et al., 2004
	Otto Fm. (GA)	34.533 84.018	1.1, 1.2, 1.3, 1.4, 1.8, 1.9, 2.0, 2.1,& 2.9	Bream et al., 2004
	Otto Fm. (NC)	35.060 83.453	1.1, 1.2, 1.3, 1.4, 1.5, & 1.6	Bream et al., 2004
	Otto Fm. (NC)	35º 25.85' N 83º 05.58'W	1.0, 1.1, 1.2, & 1.3	Merschat, unpub. data
Tugaloo	Tallulah Falls Quartzite (GA)	34.817ºN 83.424ºW	1.1, 1.2, & 1.3	Bream et al., 2004
	Tallulah Falls Quartzite (GA)	34.736ºN 83.391ºW	1.1, 1.2, & 1.3	Bream et al., 2004
	Tallulah Falls mgw (GA)	34.838ºN 83.422ºW	1.1, 1.2, & 1.3	Bream et al., 2004
	Tallulah Falls mgw (NC)	35.454ºN 82.661ºW	1.1, 1.2, & 1.3	Bream et al., 2004
	Tallulah Falls mgw (NC)	35.600ºN 82.538ºW	1.1, 1.2, &1.3	Bream et al., 2004

Table 4. Modern geochronologic data for Blue Ridge plutons.

Pluton	Lat. & Long.	Terrane	Age (Ma)	Inheritance	Comments	Source
Spruce Pine Pegmatite (NC)	35º 53' 56" N 82º 4' 49" W	**Tugaloo** Tallulah Falls/Ashe Fm.	361±2	None	Zr had accicular fan-like form	Mapes, 2002
Spruce Pine Pegmatite (NC)	35º 59' 20"N 82º 14' 57" W	**Tugaloo** Tallulah Falls/Ashe Fm.	392±2	None		Mapes, 2002
Trondhj. dike (NC)	35º 40' 03" N 82º 35' 24" W	??????	417±9	abun., 1.0-1.15 Ga		Miller et al., 2000
Trondhj. dike (NC)	35º 24' 51" N 83º 02' 08"W	**Dahlonega gold belt** Otto Fm.	402±4	abun., 1.0-1.2 Ga		Mapes, 2002
Rabun Granodiorite porphyritic (GA)	34º 58' 41" N 83º 21' 28" W	**Tugaloo** Tallulah Falls/Ashe Fm.	374±4	abun. 1.0-1.2 Ga; one 1.41 Ga & two 2.6-2.7	15-20% Zr are accicular and lack core	Miller et al., 1998, 2000
equigranular (NC)		**Tugaloo** Tallulah Falls	336±2	Abun.		D.Stahr, unpub. data
Walnut Creek granodiorite (NC)		**Tugaloo** Tallulah Falls	333±2	Abun.		D. Stahr, unpub. data
Yonah Mtn tonalite (GA)	34º 38' 22" N 83º 42' 53" W	**Tugaloo** Tallulah Falls/Ashe Fm.	346±17	abun. 0.9-1.5 Ga, five core between 571-760 Ma		Mapes 2002
Mount Airy granodiotrite (NC)		**Tugaloo** Alligator Back Fm.	352±6			Sinha pers. comm. in Mapes, 2002
Stone Mountain granodiorite (NC)	36º 23' 20"N 81º 04' 10" W	**Tugaloo** Alligator Back Fm.	353±6	abun. 0.9-1.2 Ga		Mapes, 2002
Pink Beds trondhjemite (NC)	35º 22' 57" N 82º 44' 25" W	**Tugaloo** Tallulah Falls/Ashe Fm.	388±5	abun., 1.0-1.25 Ga, single cores of 460, 650, 720 Ma		Miller et al., 2000
Looking Glass granodiorite (NC)	35º 18' 23" N 82º 21' 28" W	**Tugaloo** Tallulah Falls/Ashe Fm.	380±3	abun., 1.0-1.2 Ga with 1 1.42 Ga and 2 discordant 2.85 Ga.	3.2 Ga upper intercept of discord. core	Miller et al., 2000
Whiteside Trondhj–granodi (NC)		**Tugaloo** Tallulah Falls/Ashe	466±10	Common, 1.2-1.28; one 2.6 Ga		Miller et al., 1998, 2000
Persimmon Creek gneiss (NC-GA)		**Cowrock** Coleman R./Ridgepole Fm.	468±3	Common, 1.09-1.15Ga; two 1.4 Ga cores		McDowell et al., 2002
Cane Creek gneiss (GA)	34.576 84.017	**Dahlonega gold belt** Otto Fm.	482±4	common		Settles, 2002; Bream, 2003
Hillabee metadacite (AL)	34º 14' 32" N 86º 02' 34" W	**Talladega Belt** overlies Talladega group	470±4			Thomas, 2001; McClellan et al., 2005
Lake Burton felsic metatuff (GA)	34º 51' 45" N 83º 33' 17"W	**Dahlonega gold belt** Otto Fm.	468±6	common 0,975-1.3 Ga & 1.2-1.3 Ga		Thomas, 2001
Barlow gneiss	34º 29' 50" N 84º 01' 04" W	**Dahlonega gold belt** **????**	466±6	None		Thomas, 2001
Galts Ferry gneiss (GA)	34º 07' 42" N 84º 37' 33" W	**Dahlonega gold belt** **????**	463±3	None		Thomas, 2001
Villa Rica gneiss (GA)	33º 44' 00" N 84º 58' 24" W	**Dahlonega gold belt** **????**	458±3	None		Thomas, 2001

Table 5. Modern geochronologic data for Blue Ridge metamorphism.

Terrane	Rock Unit	Lat. & Long.	Age (Ma)	Type	Source
Mars Hill	Carvers Gap gneiss RM-1 (NC) NC-49 (NC) NC-81 (NC)	 36º 6.2 N 82º 6.7W 36º 6.1 N 82º 7.5 W 36º 06.3 N 82º 07.6 W	 1007 ± 150 966±51 (Pb loss?) 1022±40	Zircon rim	Carrigan et al., 2003
Mars Hill	Cloudland gneiss RM-CLG (NC) NC-28 (NC)	 36º 06.2 N 82 08.0W 36º 06.2N 82º 08.0W	 1035±19 963±64 (Pb loss?)	Zircon rim	Carrigan et al., 2003
Mars Hill	Felsic orthogneiss (Carvers Gap gneiss) RM 21 (NC)	(?)	1026±19	Zircon rim	Ownby et al., 2004
Mars Hill	Cloudland gneiss	(?)	957±31	Zircon rim	Ownby et al., 2004
Mars Hill	Bakersville metagabbro (gar.-am.) (NC)	(?)	~475	Zircon rim	Ownby et al., 2004
	Max Patch granite (NC)	(?)	~1000	Zircon rim	Berquist, 2005; Berquist et al., this guidebook
	Spring Creek granitoid (NC)	(?)	~1000	Zircon rim	Berquist, 2005; Berquist et al., this guidebook
	Indian Grave protomylonite (NC)	(?)	~1000	Zircon rim	Berquist, 2005; Berquist et al., this guidebook
Laurentian Margin	Great Smoky Gr. (NC)	(?)	480±10	Monazite	Moecher et al., 2003
Cartoogechaye	Winding Stair Gap, (NC)	35.1224ºN 83.5435ºW	460±12	Zircon rim	Moecher et al., 2004
Cartoogechaye	Chunky Gal Mtn, (NC)	35.0588ºN 83.6203ºW	463±6	Zircon rim	A. J. Merschat, unpub data
Cartoogechaye	Willets-Ochre Hill am., P22A (NC)	(?)	~450 Ma	Zircon rim	Berquist, 2005; Berquist et al., this guidebook
Dahlonega gold belt	Otto Fm. (NC)	35º 25.85' N 83º 05.58'W	~455	Zircon rim	A. J. Merschat, unpub, data
Tugaloo	Wiley Gneiss (GA)	34º 48.2 N 83º 25.3 W	1056±93	Zircon rim	Carrigan et al., 2003
Tugaloo	Sutton Creek gneiss, (GA)	34º 44.2 N 83º 34.5 W	1030±19	Zircon rim	Carrigan et al., 2003
Tugaloo	Toxaway Gneiss TOX-1 (SC) TOX-1B (SC) G-MA-2 (SC) B-TB-3 (SC) TXFL (NC)	 35º 0026 N 83º 0429 W 35º 00.4 N 82º 59.6 W 35º 07.2 N 82º 55.5 W	**~350** & 1027±27 1037±94 1039±82 921±45(Pb loss?) 1028±14	Zircon rim	Carrigan et al., 2003
Tugaloo	Tallulah Falls mig. am. P77D (NC)	(?)	~1000 & 450	Zircon rim	Berquist, 2005; Berquist et al., this guidebook

BLUE RIDGE TECTONIC ELEMENTS

Today we know that the Blue Ridge is a composite of several tectonostratigraphic terranes, which can be separated into the western (Laurentian margin), central, and eastern Blue Ridge, and western Inner Piedmont components. The western Blue Ridge extends from the Great Smoky-Holston Mountain to the Hayesville fault, and consists of Early to Middle Proterozoic basement rocks and the rifted margin and rift-to-drift assemblage (Rast and Kohles, 1986) (Fig. 2). The central Blue Ridge consists of the Cowrock and Cartoogechaye terranes (Hatcher, 2002), and Dahlonega gold belt, which, together with the Laurentian margin, comprise the southern Appalachian Taconian metamorphic core. The eastern Blue Ridge consists of the western part of the Tugaloo terrane, which was overthrust onto central Blue Ridge terranes along the Alleghanian Chattahoochee-Holland Mountain fault.

Western Blue Ridge

The western Blue Ridge consists of Grenville basement that includes the Mars Hill terrane, which consists of ~1.8 Ga Early Proterozoic rocks (Gulley, 1985; Ownby et al., 2004) accreted during the Grenville orogeny. It contains the oldest rocks in the Appalachians (Table 1). The remainder of the western and eastern Blue Ridge basement rocks are predominantly orthogneisses with ages in the range of 1.0 to 1.2 Ga, with a metamorphic overprint at ~980 Ma (Heatherington et al., 1996; Carrigan et al., 2003; Hatcher et al., 2004; Berquist et al., this guidebook; Cattanach and Merschat, this guidebook) (Tables 1 and 2). The most frequent age of basement orthogneisses is ~1.15 Ga. Basement rocks are nonconformably overlain by the Ocoee Supergroup and then by the rift-to-drift-to-passive-margin transition sequence consisting of the Chilhowee Group, Shady Dolomite, and Rome Formation.

Ocoee Supergroup rocks filled a basin that developed along the Laurentian margin wherein the thickness increases from zero just southwest of Elizabethton, Tennessee, to 15 km or more in the GSMNP to less than 10 km to the southwest (Fig. 3). The Ocoee Supergroup consists of the lower Snowbird Group overlain by the Great Smoky Group, which is overlain by the Walden Creek Group. The Great Smoky Group accounts for the greatest thickness of Ocoee Supergroup rocks.

The Snowbird Group consists of the basal impure clastic sedimentary rocks of the Wading Branch Formation, but the sequence cleans up into the Longarm Quartzite, an arkosic sandstone to quartz arenite (Montes and Hatcher, 1999), which is overlain by Roaring Fork Sandstone and the more pelitic Pigeon Siltstone (Hadley and Goldsmith, 1963). The Great Smoky Group, however, is separated from the Snowbird Group by the Greenbrier fault in the GSMNP. The bulk of Great Smoky Group rocks were apparently deposited farther southeast and then thrust along the Greenbrier fault (as a bedding thrust) over Snowbird Group rocks prior to the Taconian metamorphic peak. Great Smoky Group rocks consist predominantly of impure sandstone, graywacke, shale, and conglomerate with minor amounts of limestone and dolomite (King, 1964). Units within the Great Smoky consist of the Copperhill Formation (~Thunderhead and Elkmont Sandstone equivalent) of thick-bedded, coarse, graded-bedded sandstone and shale; the Wehutty Formation (~Anakeesta Formation equivalent) of graphitic, sulfidic slate interbedded with coarse sandstone; the Ammons Formation (~Hothouse equivalent) of medium-bedded, medium-grained sandstone and slate that locally contains the Grassy Branch Member (Mohr, 1975) of graphitic slate and sandstone; and the Dean Formation of thick-bedded, graded-bedded sandstone, slate and coarse polymictic, albeit milky quartz-dominated, conglomerate. The Dean Formation conformably underlies Wilhite Slate in southeastern Tennessee and southwestern North Carolina and the Nantahala Slate on the flanks of the Murphy syncline. Interestingly, the Walden Creek Group, consisting of the Wilhite and Sandsuck Formations, along with the Shields Formation, conformably overlies both the Snowbird Group and the Great Smoky Group (Hamilton, 1961; Costello, 1993; Carter, 1994; Geddes; 1995; Martin; 1997; Thigpen, 2005; Merschat et al., 2005 unpublished; Cattanach and Merschat, this guidebook). The Walden Creek Group consists of a mixed, rapidly deposited sandstone and conglomerate sequence (Shields Formation) overlain by the slate, minor sandstone, and limestone of the Wilhite Formation. The Wilhite is overlain by the Sandsuck Formation of shale, sandstone, conglomerate, and limestone. These units are succeeded by the Chilhowee Group of mostly clean sandstone and shale, with the exception of immature sandstone and shale and some basalt flows in the lower Unicoi Formation in northeastern Tennessee and southwestern Virginia (King and Ferguson, 1960). Except for late Neoproterozoic achritarchs reported by Knoll and Keller (1979) in the Wilhite and some unidentifiable soft-bodied organisms in the Sandsuck Formation (Broadhead et al., 1991) (Fig. 4), the Chilhowee Group contains the oldest fossils in the southern Blue Ridge (e.g., Walker and Driese, 1991). Consequently, King et al. (1958) chose to draw the Cambrian-Precambrian boundary beneath the Chilhowee. The unidentifiable soft-bodied organisms in the Sandsuck Formation raise the possibility that this unit could represent the missing lower Tommotian interval—the Lower Cambrian section that is so prominently developed on the western North American margin.

Great Smoky Group rocks of the central-western Blue Ridge are overlain by Murphy syncline rocks, a sequence that has been correlated with the Cambro-Ordovician section in the Valley and Ridge and westernmost Blue Ridge (e.g., Power and Forrest, 1973). Tull and Groszos (1988) suggested an unconformity exists beneath the uppermost unit in the Murphy syncline, the Mineral Bluff Formation, and that the Mineral Bluff Formation has a mid-Paleozoic age (Tull and Groszos, 1990; Tull, et al., 1993). Tull et al. (1993) reported an echinoid stem from the Murphy Marble beneath the unconformity, indicating the Murphy Marble cannot be older than Ordovician. The Mineral Bluff Formation could also represent the Middle Ordovician clastic wedge, and the unconformity beneath the Knox unconformity (Fig. 5). Our most recent detrital zircon SHRIMP geochronologic data from a sandstone in the Mineral Bluff Formation contains no Paleozoic zircons (Table 3, Fig. 6), suggesting that the zircon suite is derived from Grenville and older Laurentian basement rocks, and therefore probably is not younger than Ordovician.

Central Blue Ridge Terranes

Although Hadley and Goldsmith (1963) and Hadley and Nelson (1971) recognized a major boundary in the central Blue Ridge, the Hayesville fault was first recognized as a pre-

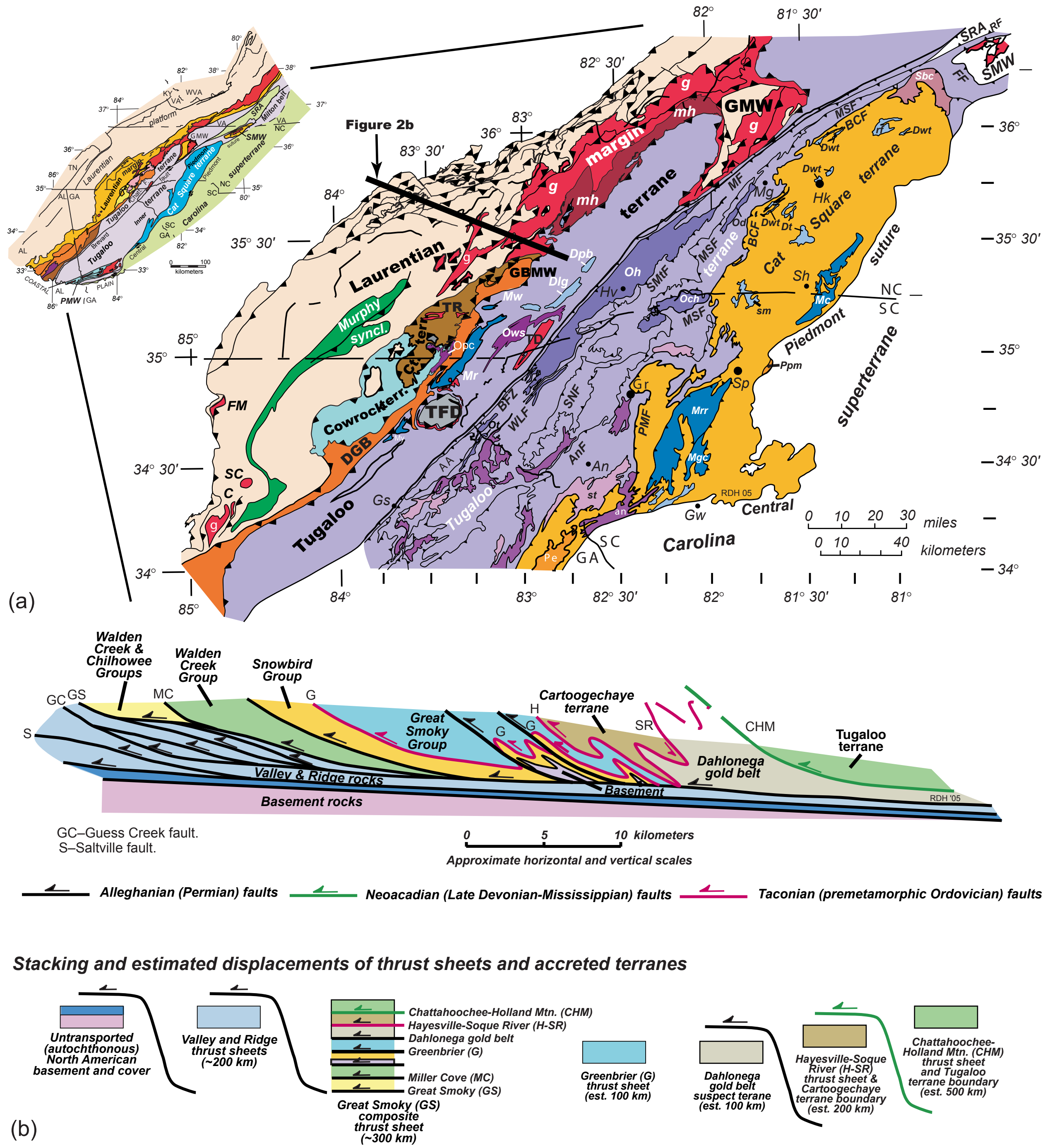

Figure 2. (a) Upper left map: major tectonic elements of the southern Appalachians. SMW–Sauratown Mountains window. GMW–Grandfather Mountain window. PMW–Pine Mountain window. SRA–Smith River allochthon. Larger map: tectonic elements of the southern Appalachian Blue Ridge and Inner Piedmont from northern Georgia to northwestern North Carolina. G–mostly Grenvillian orthogneisses. Mh–Mars Hill terrane. DGB–Dahlonega gold belt. GMW–Grandfather Mountain window. FM–Fort Mountain massif. SC–Salem Church massif. C–Corbin massif. TR–Trimont Ridge massif. TFD–Tallulah Falls dome. TD–Toxaway dome. Ct. terr.–Cartoogechaye terrane. AA–Alto allochthon. AnF–Anderson. BCF–Brindle Creek. BFZ–Brevard. FF–Forbush. MF–Marion. MSF–Mill Spring. PMF–Paris Mountain. RF–Ridgeway. SNF–Six Mile. WLF–Walhalla. Named Ordovician (purple) and Ordovician(?) (lighter purple) granitoids: Och–Caesars Crossroads. Devonian and Devonian(?) plutons (light blue): Drf–Rocky Face. Dt–Toluca. Dwt–Walker Top. sm–Sandy Mush. Mississippian granitoids: Mc–Cherryville. Mgc–Gray Court. Mr–Rabun. Mrr–Reedy River. Mw–Walnut Creek. My–Yonah (Mapes, 2002) Pennsylvanian granitoids: Pe–Elberton. Ppm–Pacolet Mills. (Carolina terrane). Towns: Gr–Greenville. Gs–Gainesville. Gw–Greenwood. Hk–Hickory. Hv–Hendersonville. Mg–Morganton. Sh–Shelby. (b) Schematic cross section through the Blue Ridge in the vicinity of the 2005 Carolina Geological Society field trip showing different tectonic units and how they may be assembled. See Figure 2(a) for location of section. (c) Generalized geologic map of part of the western, central, and eastern Blue Ridge in the area of the 2005 Carolina Geological Society field trip. Purple–1.8 Ga gneisses of the Mars Hill terrane (oldest rocks in the Appalachians). Lavender–1.15 Ga mostly granitic gneisses. Yellow–Snowbird Group clean sandstone and shale (meta.). Blue green–Great Smoky Group dirty sandstones and shales (meta.). Burnt orange–Murphy belt Cambrian and Ordovician(?) metasedimentary rocks (sandstone, shale, limestone). Light brown–Cartoogechaye terrane metasedimentary and metavolcanic rocks. Pink–Amphibolite. Dark green–Ultramafic rocks. Tan–Dahlonega gold belt metasandstone and schist. Green–Ashe-Tallulah Falls Formation dark metasandstone, schist, and metavolcanic rocks of the Tugaloo terrane. Black dots indicate field trip stops, accompanied by red stop numbers. Highways are darker red.

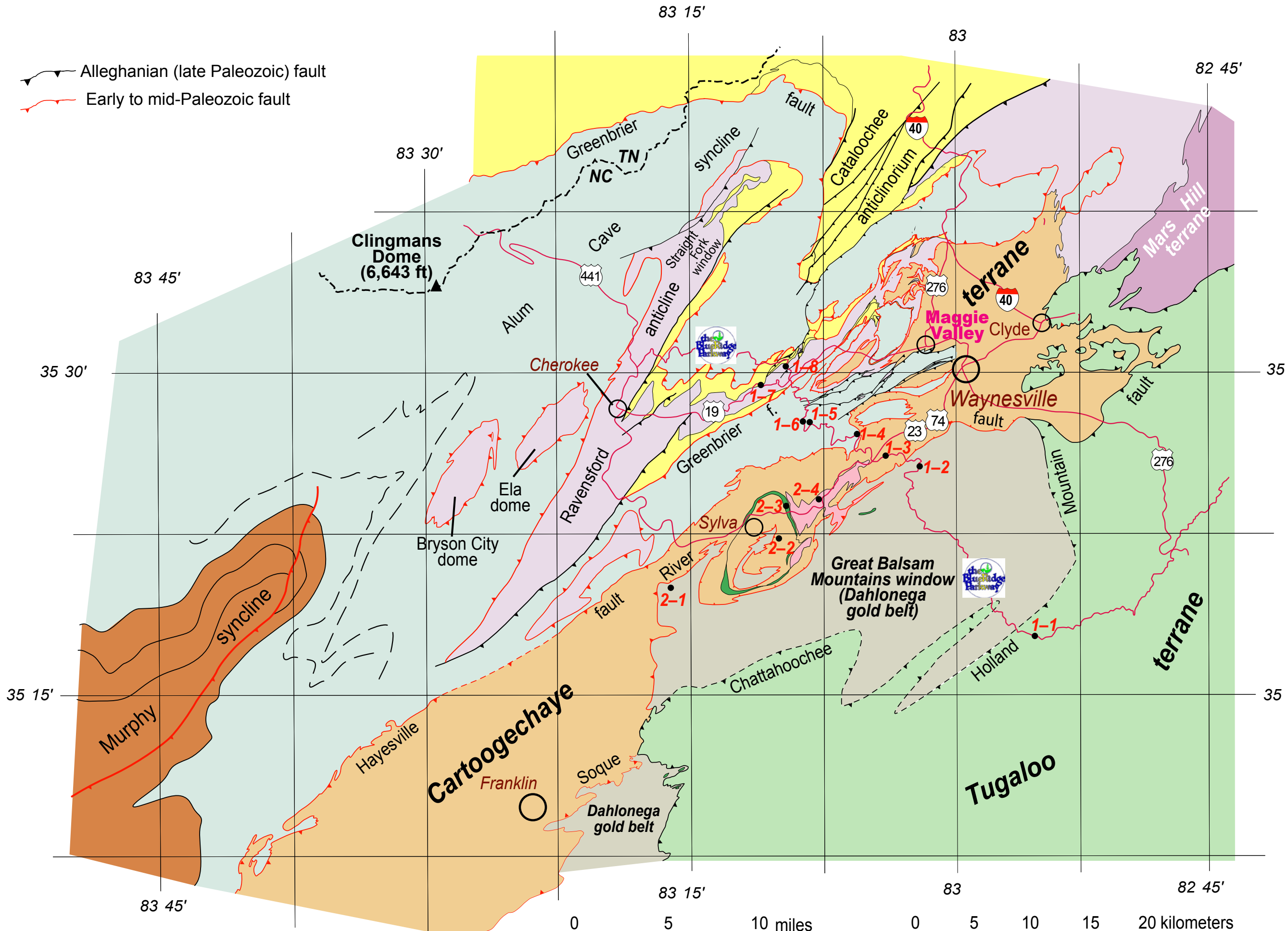

Figure 2. (continued)

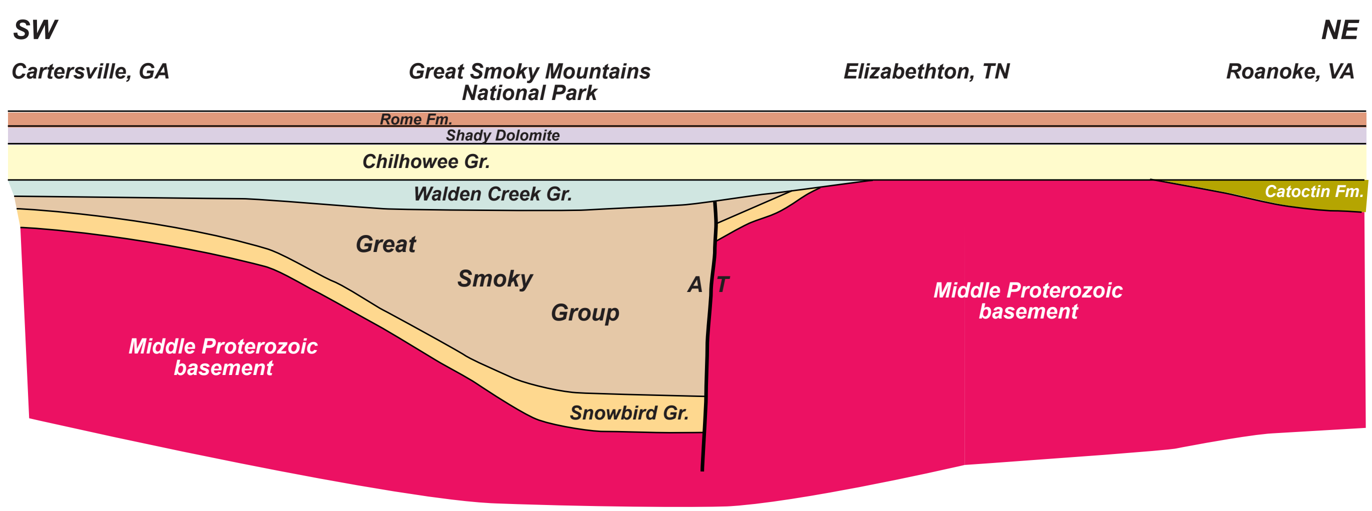

Figure 3. Along-strike reconstruction of the Ocoee basin. Note the abrupt termination of the basin to the northeast, gradual termination to the southwest, and the appearance of the Catoctin volcanic-sedimentary sequence to the northeast in Virginia.

metamorphic fault separating Great Smoky Group rocks of the western Blue Ridge from high-grade central Blue Ridge gneiss, schist, and amphibolite (Hatcher, 1978). Rock units on both sides of the fault are truncated (Eckert, 1984). Part of the basis for recognition of this fault, however, is the occurrence of abundant mafic and ultramafic rocks to the east (Hatcher, 1978), whereas to the west there are few or no igneous rocks (Hurst, 1955; Hadley and Goldsmith, 1963). Moreover, metasandstones to the east are dominated by quartz-biotite-plagioclase, whereas those to the west are two-mica, two-feldspar metasandstones. Completion of detailed mapping in the Hayesville thrust sheet has revealed two separate tectonostratigraphic terranes in the central Blue Ridge. The southernmost is the Cowrock terrane—located mostly in Georgia—truncated to the north by the Chunky Gal Mountain and Shope Fork faults (Hatcher et al., 2004) (Fig. 2). The Cowrock terrane sequence, the Coweeta Group (Hatcher, 1988),

Figure 4. Soft-bodied organism from the Sandsuck Formation in southeastern Tennessee. These organisms were first discovered by Phillips (1951) and Rackley (1951), and Phillips (1951) provided an exact location where they were found, so we were able to relocate the site and find another specimen during the late 1980s.

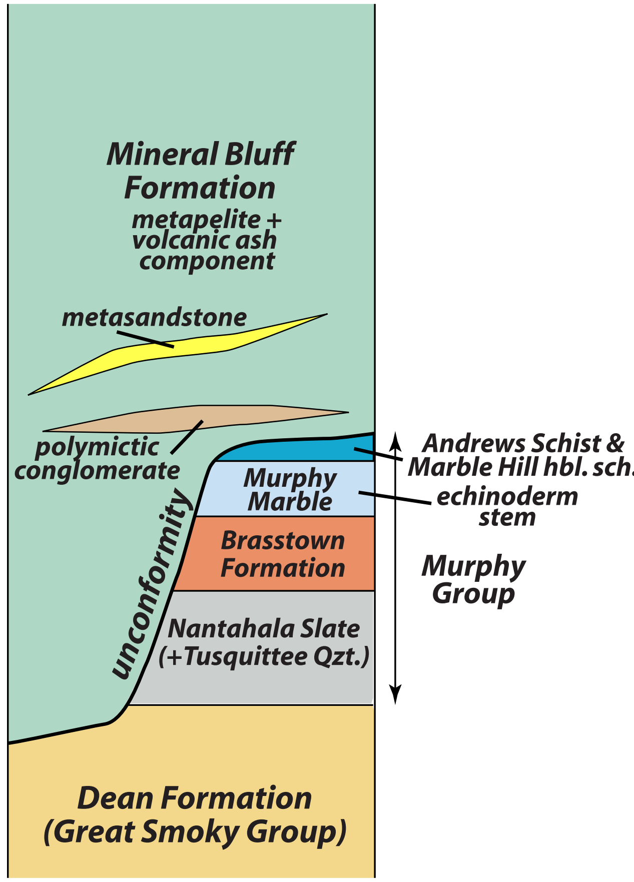

Figure 5. Murphy syncline stratigraphy from Tull and Groszos (1988).

consists of metasandstone, schist, and some mafic and ultramafic rocks in the Coleman River Formation, particularly around the Shooting Creek and Brasstown Bald windows (Hartley, 1973), along with additional mafic rocks in Georgia near the southwest end of the terrane (Nelson and Gillon, 1985). The Coleman River Formation of metasandstone and pelitic schist is overlain by the more diverse Ridgepole Mountain Formation, which consists of biotite schist, metasandstone, metagraywacke, and quartzite (Hatcher, 1979, 1988). Coleman River metasandstone contains Grenvillian and older detrital zircons (Bream et al., 2004), indicating these rocks are not basement, but are more likely a more distal Laurentian margin sequence. The Coweeta Group was intruded by the 466 Ma Persimmon Creek Gneiss (McDowell et al., 2002) in the southeastern part of the terrane (Hatcher, 1980). The eastern boundary of the Cowrock terrane is the Soque River fault, which truncates the Shope Fork fault, and continues northward into North Carolina forming the western boundary of the Dahlonega gold belt and the eastern flank of both the Cowrock and Cartoogechaye terranes (Fig. 2). Settles (2002) demonstrated that the Hayesville and Soque River faults join around the southwest end of the Cowrock terrane just west of Dahlonega, Georgia, where they truncate the Allatoona fault (Fig. 2).

The Cartoogechaye terrane lies to the northeast of and atop the Cowrock terrane. It consists of metasandstone, pelitic schist, and an assemblage of mafic and ultramafic rocks, along with some Grenville basement (Hatcher et al., 2004). These were metamorphosed to assemblages as high as granulite facies (Force, 1976; Hatcher and Butler, 1979; Absher and McSween, 1985; Eckert et al., 1989) during the Ordovician (Moecher and Miller, 2000; Moecher et al., 2004, this guidebook). Metasandstone from Winding Stair Gap west of Franklin and another metasandstone from near Waynesville yielded Grenvillian and older detrital zircons (Miller et al., 2000; A. J. Merschat, unpublished data) (Table 3), indicating that this terrane cannot consist predominantly of Grenville basement, but the Trimont Ridge gneiss appears to be a basement unit (Hatcher et al., 2004). The Cartoogechaye terrane is traceable northward from the Georgia-North Carolina line through Franklin and Sylva to the region just south of I-40 where it terminates in the Clyde, Cruso, or Dellwood quadrangles.

Dahlonega gold belt

The Dahlonega gold belt is traceable from the area of the CGS 2005 field trip southward through Dahlonega, Georgia, and beyond (German, 1985; Settles, 2002) (Fig. 2). Rocks of this terrane consist of the Otto Formation of pelitic schist and metasandstone (Hatcher, 1988), along with several mafic and ultramafic complexes. All of the mafic–ultramafic complexes dated thus far have yielded Ordovician 460 and 480 Ma ages (Thomas, 2001; Bream, 2003) (Table 4). Metasedimentary rocks of the Dahlonega gold belt consist of metasandstone containing both muscovite and biotite along with two feldspars, which no doubt led Hadley (1970) to correlate them with the Great Smoky Group. Detrital zircons from the Dahlonega gold belt metasandstones reflect a North American source in four samples (Table 3, Fig. 6), but one sample contains Gondwanan (2.0 Ga) zircons in addition to Grenvillian and older North American detrital zircons (Bream, 2003). The Gondwanan ages have not been reproduced the other samples that have been analyzed (Bream, 2003; A. J. Merschat, unpublished data) (Table 3).

The Dahlonega gold belt is bounded to the east by the southeast-dipping, and tightly folded, Chattahoochee–Holland Mountain fault. Northeast of Franklin (Corbin Knob quadrangle Eckert and Yurkovich, 2005 unpublished) Chattahoochee–Holland Mountain thrust sheet rocks have been mapped across

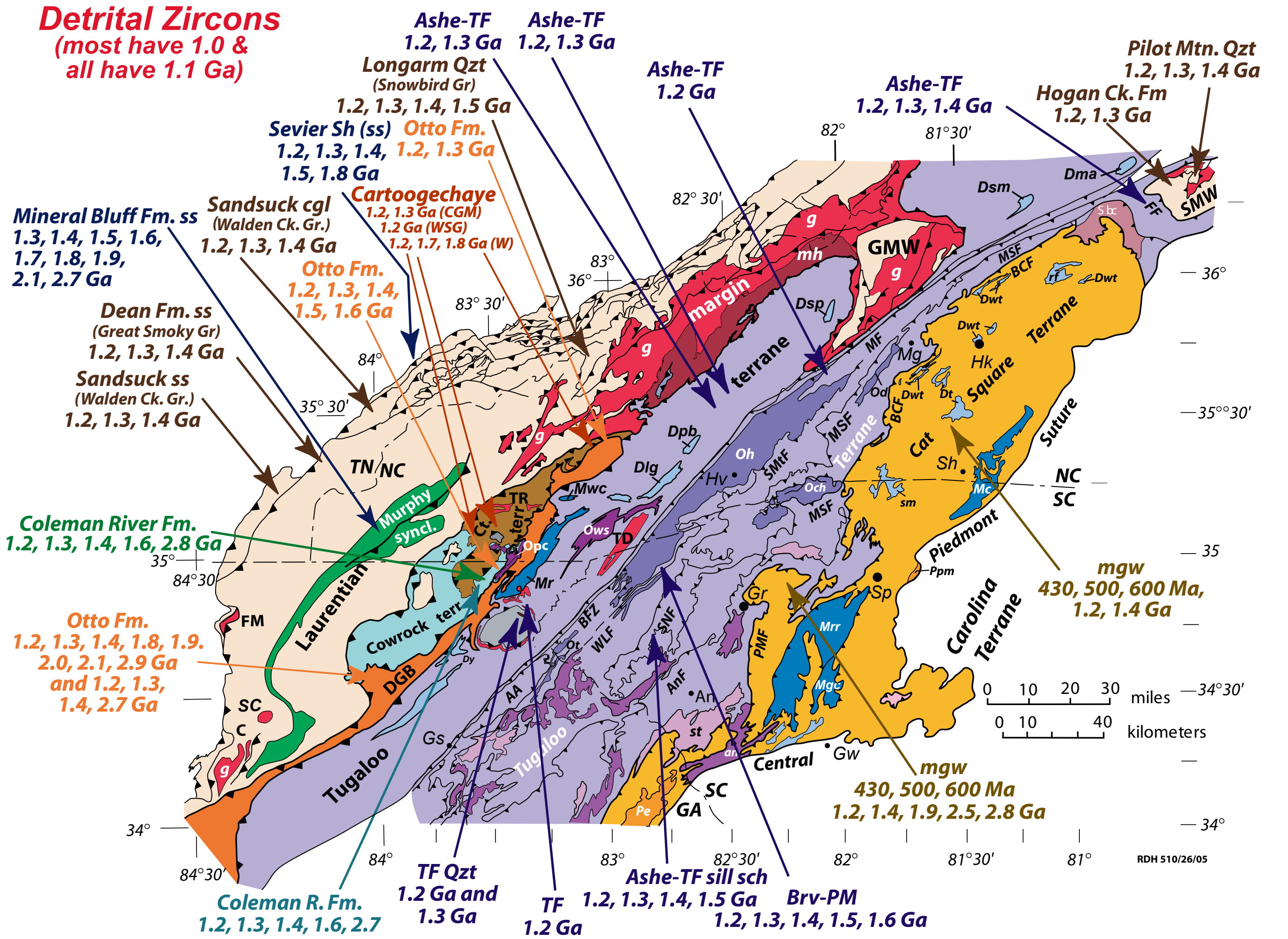

Figure 6. Detrital zircon ages from the southern Appalachians west of the Carolina superterrane (from Bream, 2003 and A. J. Merschat, unpublished). SHRIMP ages of detrital zircons from mostly sandstones and metasandstones from the Valley and Ridge, Laurentian margin, and Tugaloo terrane clearly indicate a Laurentian provenance from Grenville and granite-rhyolite province sources in the Laurentian basement to the west. The Dahlonega gold belt contains possible Gondwanan zircons, in addition to a dominant Laurentian zircon population. The Cat Square terrane contains Archean, Penokian (?), granite-rhyolite province, Grenvillian, Avalonian (Carolina), and Silurian zircons, requiring mid- to late Paleozoic docking of the Carolina superterrane.

the Dahlonega gold belt onto rocks of the Cartoogechaye terrane (Eckert and Yurkovich, 2005 unpublished). This fault promontory is preserved in a synform that separates the northern from the southern parts of the Dahlonega gold belt (Fig. 2). The northern part of the Dahlonega gold belt is exposed in a window, here called the Great Balsam Mountains window, in the Sylva–Waynesville–Shining Rock area. The southern part extends from just southwest of Sylva to south of Canton, Georgia (Fig. 2).

Tugaloo Terrane

East of the Chattahoochee-Holland Mountain fault lies the Tugaloo terrane, which consists of Ashe-Tallulah Falls Formation rocks, and small amounts of Grenville basement in the Toxaway and Tallulah Falls domes. In addition, there are several large granitic plutons that were intruded during either the middle or late Paleozoic, long after intrusion of the Middle Ordovician Whiteside pluton (Miller et al., 2000) (Table 4, Fig. 7). Most of these plutons consist of granodiorite, with some tonalite (Miller et al., 1997; McDowell et al., 2002; Stahr et al., 2005). In addition, numerous small ultramafic bodies and several mafic-ultramafic complexes occur in the Tugaloo terrane. The Tugaloo terrane extends from the Chattahoochee-Holland Mountain fault eastward across the Brevard fault zone to the Brindle Creek thrust, which separates the eastern Inner Piedmont Cat Square terrane from western Inner Piedmont Tugaloo terrane rocks (Hatcher, 2002) (Fig. 2).

The Tallulah Falls-Ashe Formation consists of a lower unit of metasandstone, pelitic schist, and amphibolite, which is overlain by an aluminous, white mica-dominated schist unit containing either kyanite or sillimanite, abundant garnet, and minor amphibolite. The aluminous schist unit is overlain by a metasandstone and pelitic schist unit that contains small amounts of amphibolite—or none (Hatcher, 1971, 1974). This sequence is readily mappable in detail throughout the eastern Blue Ridge and western Inner Piedmont of northeast Georgia and western North Carolina. These subdivisions have been recognized wherever detailed mapping has been conducted in the Carolinas and northeastern Georgia.

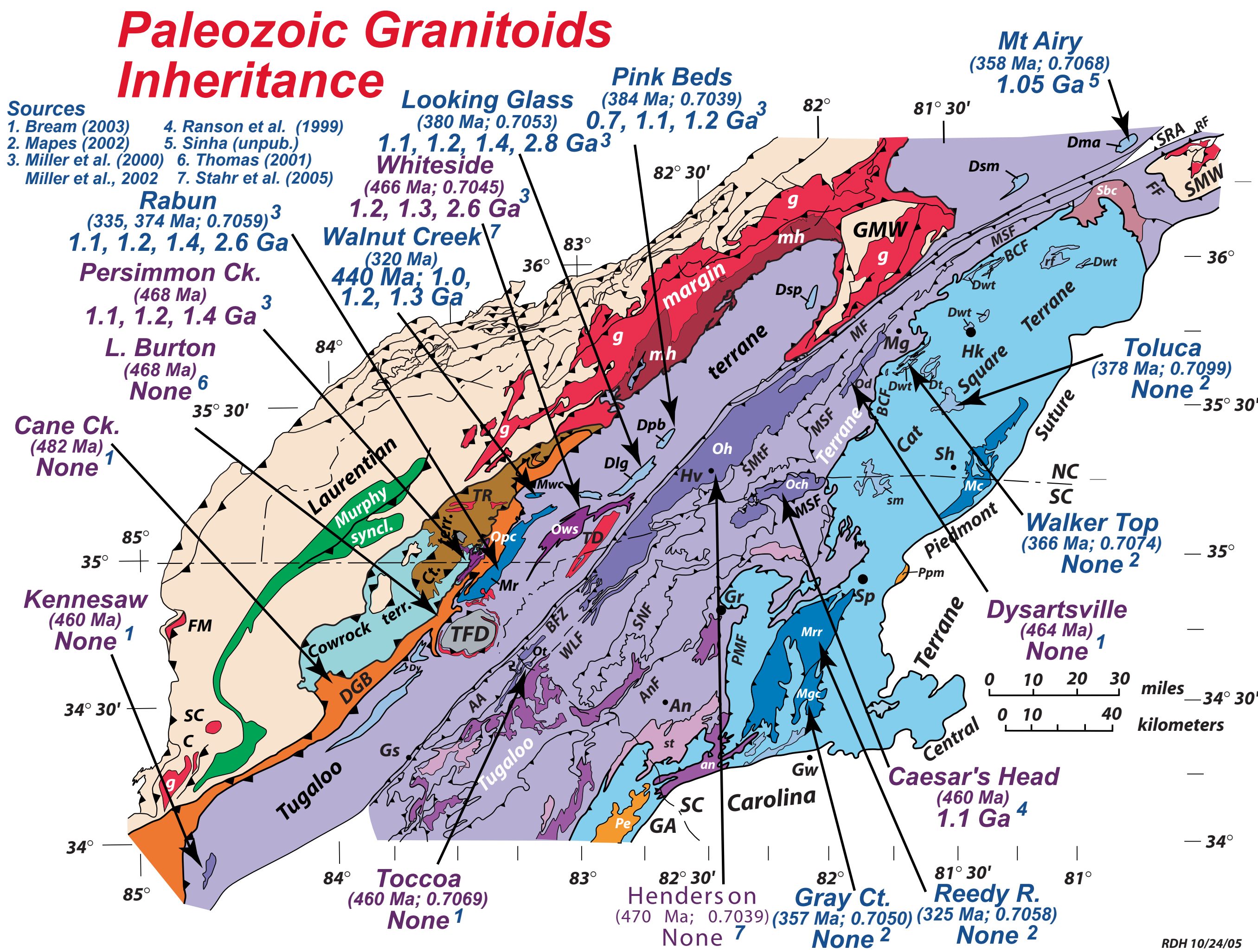

Figure 7. SHRIMP and ion microprobe pluton ages, $^{87}Sr/^{86}Sr$ initial ratios, and inheritance in the western half of the southern Appalachians. Ordovician plutons dominate in the Tugaloo terrane, with the exception of Devonian and Mississipian plutonic suites in the western Tugaloo terrane. The new Mississippian ages on the Rabun and Walnut Creek plutons identify the westernmost Carboniferous (Alleghanian?) plutons in the Appalachian orogen. Cat Square plutons are dominantly Acadian and Neoacadian (380, 366-350 Ma), with some Carboniferous and Permian plutons present. All of these appear to be anatectic. Remarkably little inheritance in the Inner Piedmont contrasts with varied and abundant inheritance in the western Tugaloo terrane (west of the Brevard fault zone). This break likely represents different crustal types encountered on either side of the Brevard fault zone during pluton emplacement.

Blue Ridge Paleozoic Plutons

A suite of Paleozoic plutons exists in the eastern Blue Ridge and western Inner Piedmont from near the Virginia-North Carolina state line to Alabama (Fig. 2). Most of these plutons are Middle Ordovician granitoids (Miller et al., 2000; McDowell et al., 2002; Bream, 2003), but a Devonian and Mississippian suite also occurs here (Miller et al., 2000; Stahr et al., 2005) (Table 4, Fig. 7). Most of these plutons are medium- to coarse-grained, equigranular granodiorite and tonalite, with the exception of the megacrystic phase of the Rabun pluton, which contains K-spar megacrysts up to 4 cm (Miller et al., 1997; Stahr et al., 2005). All of these plutons are foliated and, where enough layering exists to record strain, two foliations may be observed, one possibly related to flow and the other to deformation. Two foliations are frequently observed in the megacrystic Rabun Granodiorite. The Mississippian Rabun pluton (Miller and Stewart, 2002; Stahr et al., 2005) also is truncated by the Chattahoochee-Holland Mountain fault.

Mafic metavolcanic rocks (now amphibolite), and minor felsic volcanic rocks, also occur here. Mafic and ultramafic rocks are frequently spatially and temporally associated (e.g., McElhaney and McSween, 1983; Hatcher et al., 1984), but there are also numerous isolated small to large ultramafic bodies in the central and eastern Blue Ridge, and Dahlonega gold belt (Ryan et al., this guidebook; Swanson et al., this guidebook; Warner and Hepler, this guidebook).

LOCAL-REGIONAL PROBLEMS IN AND NEAR FIELD TRIP AREA

Several interesting problems have arisen as a result of detailed geologic mapping by the state geological surveys, the USGS, and faculty and graduate students from several universities (Fig. 1). The topics addressed are not intended to comprise an exhaustive list. Hopefully, they will define new problems and evoke new solutions among guidebook users that could lead to more productive research.

Nature of the Ocoee Basin and extent of early Paleozoic faulting

The Ocoee basin formed along the irregular margin of southeastern Laurentia during initial rifting of Rodinia, beginning ~700 Ma (Southworth et al., this guidebook). This was similar to the timing of breakup of Rodinia on the western North American margin (e.g., Sears and Price, 2003). The basin ends abruptly to the northeast, suggesting that basin could be bounded by a transform fault, but ends gradually toward the southwest (Fig. 3). The Ocoee basin is marked by its great thickness of sediment accumulation, and the absence of volcanic rocks. The Mt. Rogers-Grandfather Mountain Formation was deposited in another basin farther northeast (Bryant and Reed, 1970; Rankin, 1993), and probably farther outboard. The nearly coeval to slightly younger Catoctin Formation was deposited still farther northeast along the margin in Virginia (Aleinikoff et al., 1995).

A perplexing element of the Ocoee basin is that the oldest unit in the basin, the Snowbird Group is separated from the Great Smoky Group by the Greenbrier fault, but the Snowbird is conformably overlain by Great Smoky Group to the northeast (Cattanach and Merschat, this guidebook). Another enigma is the Great Smoky Group is largely faulted against the Walden Creek Group in the GSMNP, but is in conformable contact immediately south of the Park (Geddes, 1995; Martin, 1997; Thigpen, 2005), and around the Murphy syncline that contains Walden Creek equivalent sequences (Fig. 8).

Within the younger sequence is another controversy about the age of the Wilhite Formation. Mid-Paleozoic fossils were reported from samples of Wilhite and Sandsuck Formation clastic and carbonate rocks (Unrug and Unrug, 1990; Unrug and Palmes, 1991; Unrug et al., 2000). Tull and Groszos (1990) also reinterpreted the rocks of the frontal Blue Ridge and the upper part of the section in the Murphy syncline as having been deposited in successor basins. The enigma arises from the following: (1) Early progressive metamorphism in the western Blue Ridge is Ordovician (e.g., Moecher et al., 2004; this guidebook), before the reported time of deposition of these rocks; (2) The frontal block where several of the Unrug samples were collected consists of Sandsuck Formation that can be followed up into demonstrably Cambrian Chilhowee Group rocks; (3) Attempts to reproduce fossils from most of the same localities south of the GSMNP have been unsuccessful to date, except for a possible high conodont alteration index conodont fragment from the limestone exposed at Springtown, Tennessee (John Repetski, pers. comm., 2005). Clastic sedimentary rocks from within a km of this locality contain only a very weak pressure-solution cleavage, quartz contains undulatory extinction with no evidence of recovery (maximum temperatures <300° C), and there is no microscopic evidence of recrystallization of fine micas. The carbonate rocks contain a nearly pristine microtexture of micrite and sparry calcite containing only a few stylolites, undeformed rounded quartz sand grains, and late calcite-filled tension veins; and (4) Stable isotope ($\delta^{13}C$, $\delta^{18}O$) data and $^{87}Sr/^{86}Sr$ ratios favor a Neoproterozoic age for these rocks (Thigpen, 2005).

The Greenbrier fault, which has been carefully mapped at the east end of the Great Smoky Mountains (Hadley and Goldsmith, 1963; King, 1964; Southworth et al., 2005), has been traced southwestward out of the Park where it terminates beneath the Salt Spring Mountain fault (Geddes, 1995; Thigpen, 2005). Displacement on the Greenbrier has to be large based on stratigraphic separation across the fault and its relatively long trace length. Although Costello (1993), Carter (1994), Geddes (1995), Martin (1997), and Thigpen (2005) provided convincing evidence that the Greenbrier fault does not represent the Great Smoky-Walden Creek contact in southeastern Tennessee, there is still speculation that this fault may reappear as far south as the Fort Mountain massif in Georgia (Tull, 2005 in press). There, Great Smoky Group rocks lie in fault contact with footwall basement rocks. Only a short segment of this fault occurs here because it is truncated to both the northeast and southwest by the Miller Cove-Cartersville fault. Late Paleozoic dissection of the early Greenbrier trace has been well documented in the GSMNP (Hadley and Goldsmith, 1963; King, 1964; Southworth et al., 2005) and in southeast Tennessee (Thigpen, 2005). If the Greenbrier fault is present in Georgia and possibly farther south, the stratigraphic and tectonic implications of a fault with this magnitude of extent may previously have been underestimated.

Although the Greenbrier fault consumes much of the focus of the early Paleozoic western Blue Ridge history, the nature and emplacement timing of the Rabbit Creek and Dunn

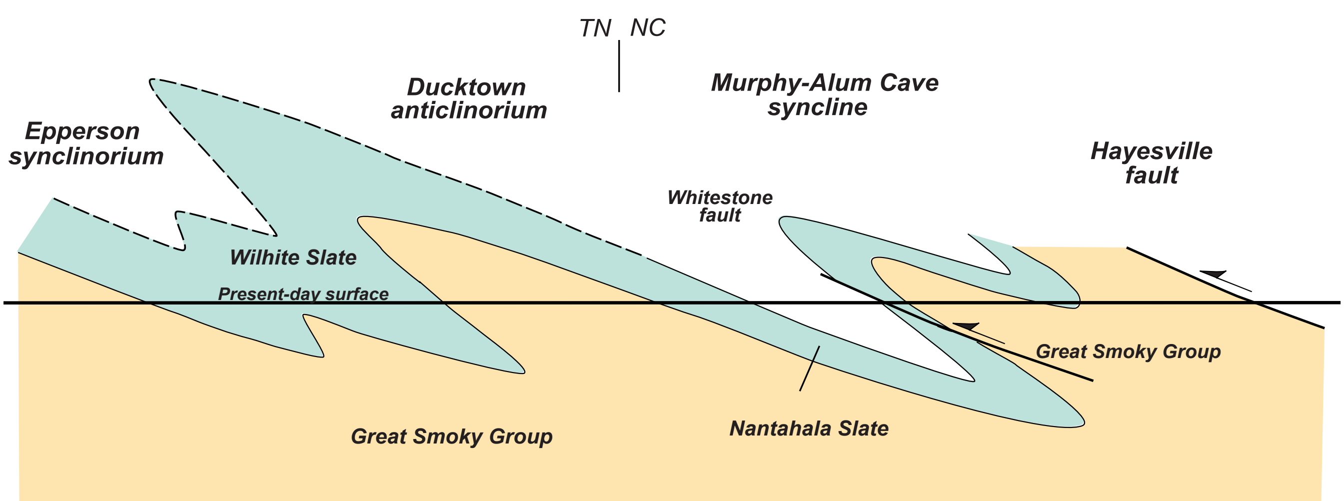

Figure 8. Schematic cross section from the Epperson syncline through the Murphy syncline showing the justification for correlating Walden Creek Group units (e.g., Wilhite Slate) with rocks in the Murphy syncline (e.g., Nantahala Slate)

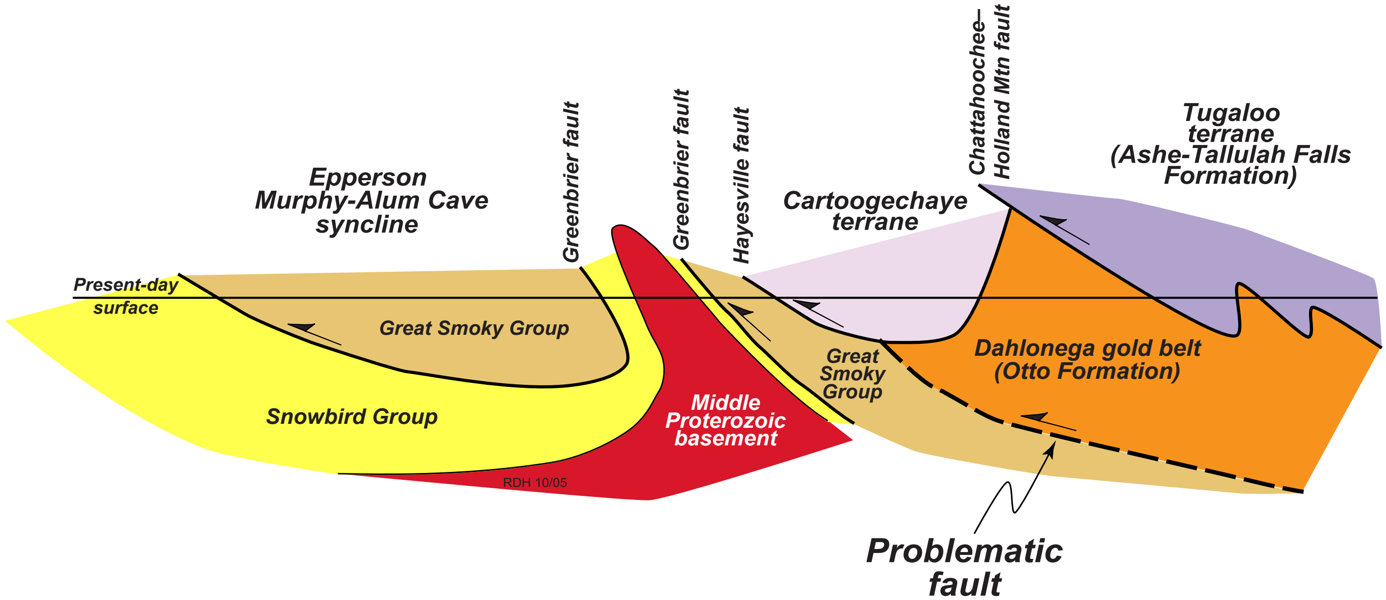

Figure 9. Schematic cross section through the area of the 2005 Carolina Geological Society field trip showing the relationships between the major tectonic units and the need for an additional fault if the Otto Formation is *not* a more easterly facies of the Great Smoky Group.

Creek-Line Springs faults and their relationships to each other and the Greenbrier fault remain somewhat enigmatic. Field relationships in the GSMNP and immediately southwest indicate that, although all three faults formed during early Paleozoic (Taconic) orogenesis, both the Greenbrier and Dunn Creek systems were subsequently folded and imbricated during late Paleozoic orogeny. The Rabbit Creek fault, despite revealing ductile fabrics and undisturbed pervasive foliation near the fault surface in the western Great Smoky Mountains (Neuman and Nelson, 1965), shows abundant evidence for brittle reactivation in the southeast Tennessee Foothills (Geddes, 1995; Thigpen, 2005). Why the Greenbrier and Dunn Creek were folded rather than reactivated like the Rabbit Creek is certainly related to post-emplacement geometry and orientation of these fault surfaces, although the implications of these observations are still not fully understood.

Northern Terminus of Cartoogechaye Terrane

The northern end of the Cartoogechaye terrane likely occurs somewhere between the Blue Ridge Parkway and I-40 in the Clyde, Cruso, or Dellwood quadrangles. Cartoogechaye terrane metasandstones contain Grenvillian detrital zircons (Table 4) precluding these metasandstones from being Grenville basement. The high metamorphic grade of these rocks has been demonstrated to be Ordovician, using modern geochronology (Miller et al., 2000; Moecher and Miller, 2000; Moecher et al., 2004). Therefore, basement rocks in the core of the western Blue Ridge should be truncated beneath the Hayesville fault near I–40 (e.g., Montes, 1997) and against the Chattahoochee-Holland Mountain fault farther east (Wiener and Merschat, 1988).

Dahlonega Gold Belt and Tallulah Falls Dome

Another perplexing problem is the relationships between the Dahlonega gold belt and the Cowrock and Cartoogechaye terranes, along with its relationships to the Laurentian margin rocks to the west. Hadley (1970) considered the Dahlonega gold belt rocks to be correlative with the Ocoee Supergroup, no doubt because of the similarities between Dahlonega gold belt metasandstone and schist to those of the Great Smoky Group. Both metasandstones are two-mica, two-feldspar sandstones that resemble those of the more fine-grained parts of the Great Smoky Group, like the Ammons Formation (Mohr, 1975; Ausburn et al., 1998), but could be distal facies of any part of the Great Smoky Group. One difficulty of making this correlation, however, lies in the presence of Gondwanan detrital zircons (1.8, 2.0 Ga) in one sample (Table 4, Fig. 6) (Bream, 2003). In the very narrow area separating the Dahlonega gold belt from the western Blue Ridge is a belt of Cartoogechaye terrane rocks in the Hazelwood quadrangle (Edelman, unpublished) (Fig. 9). If the Dahlonega gold belt rocks (Otto Formation, Hatcher, 1988) are a more eastern facies of the Great Smoky Group, there would be no problem to connect these rocks to western Blue Ridge Ocoee Supergroup rocks beneath the Cartoogechaye terrane. If the Dahlonega gold belt rocks are not correlative with the Great Smoky Group, a hidden fault is required that remains beneath the Cartoogechaye terrane throughout the extent of the Dahlonega gold belt, even beneath the narrow portion of the Cartoogechaye terrane in the Hazelwood quadrangle (Fig. 9). A candidate fault exists at the southwest end of the Cowrock terrane west of Dahlonega, Georgia, where the Allatoona fault is truncated by the Hayesville–Soque River fault (Settles, 2002), but it is difficult to perceive that this fault continues beneath the entire length of the Cowrock and Cartoogechaye terranes and does not emerge even where the Cartoogechaye terrane narrows.

A similar problem exists in the interior of the Tallulah Falls dome (Fig. 10) where Tallulah Falls Quartzite overlies metasandstone and schist originally correlated with the Tallulah Falls Formation (Hatcher, 1971). Later it became clear that the metasandstone beneath the quartzite is more like the western Blue Ridge or Otto metasandstone than Tallulah Falls metagraywacke—both the quartzite and metasandstone are two-mica, two-feldspar metasandstones (Hopson et al., 1989; Stieve, 1989). We had assumed the rocks in the interior of the Tallulah Falls dome are Otto Formation until the Gondwanan

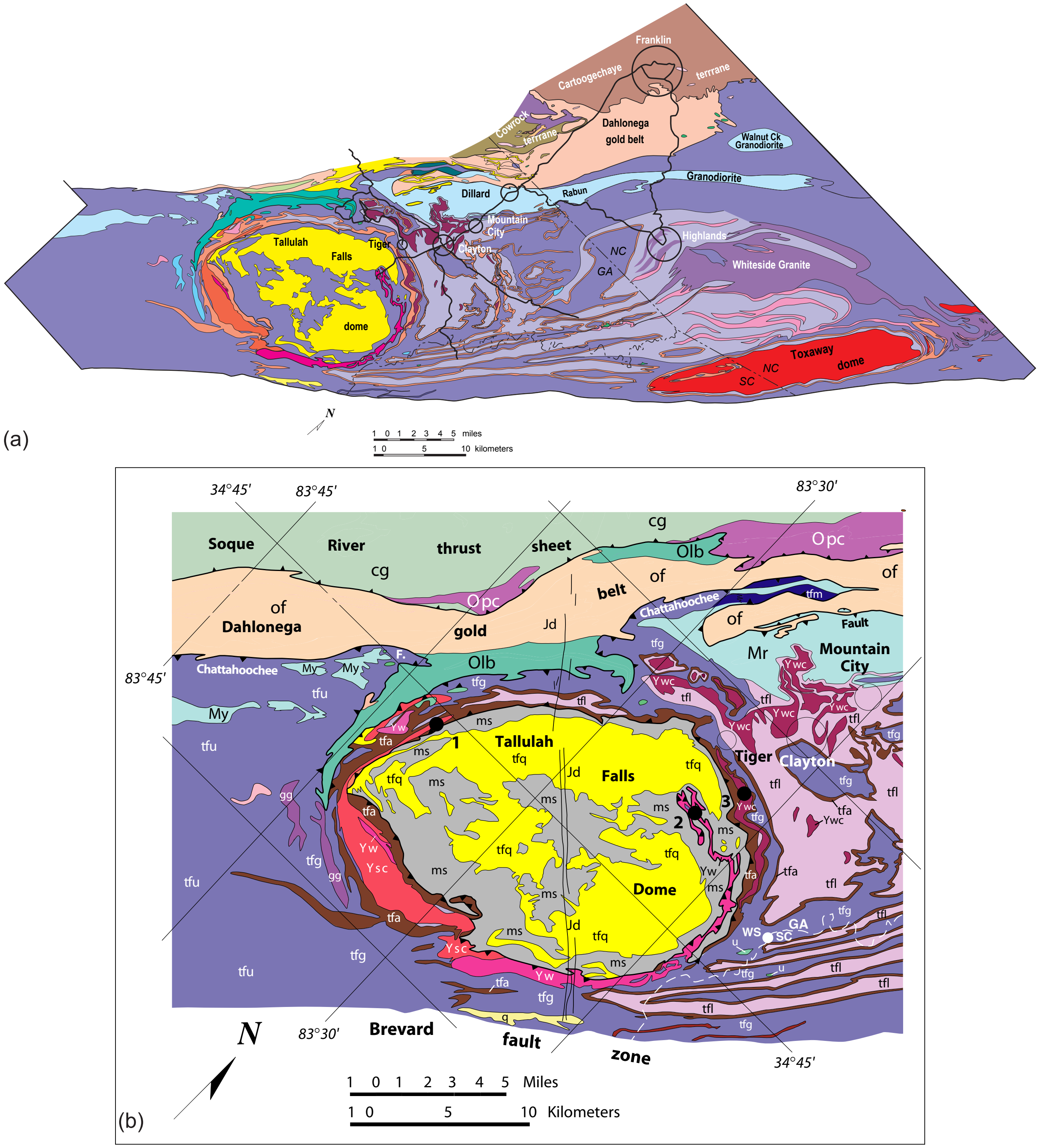

Figure 10. (a) Regional setting of the Tallulah Falls dome. Chattahoochee thrust sheet units: 1.15 Ga basement orthogneisses are shown in shades of red. Subdivisions of the Tallulah Falls Formation are shown in shades of lavender (upper and lower units), and in orange (aluminous schist unit). Amphibolite bodies are pink. Tallulah Falls Quartzite is yellow, and the unit surrounding it inside the Tallulah Falls dome is probably Otto Formation. Ultramafic bodies are medium green. Migmatite is dark blue. Ordovician Whiteside Granite and Persimmon Creek Tonalite Gneiss (Cowrock terrane) are colored reddish purple. Mississippian granitoids are colored light blue. (b) Geologic map of the Tallulah Falls dome and vicinity. Chattahoochee thrust sheet: Ysc– Sutton Creek Gneiss. Yw– Wiley Gneiss. Ywc– Wolf Creek Gneiss. Tallulah Falls Formation members: tfl–Graywacke-schist amphibolite member. tfa–Garnet-aluminous schist member (kyanite- or sillimanite-bearing). tfg–Graywacke-schist member. tfq–Tallulah Falls Quartzite. ms–metasandstone and muscovite-biotite schist beneath the Tallulah Falls Quartzite. tfu–undivided. tfm–migmatite derived from Tallulah Falls Formation metagraywacke. gg–granitoid gneiss, age unknown. q–quartzite and quartzose metasandstone. u–ultramafic rocks. Mr–Rabun granodiorite. My–Yonah granitoid. WS–Woodall Shoals (white dot on GA-SC line). Dahlonega gold belt: of–Otto Formation, undivided. Olb–Lake Burton mafic arc complex. Soque River thrust sheet: cg–Coleman River Formation metasandstone and schist. Opc–Persimmon Creek tonalite. Cutting all thrust sheets and the Brevard fault zone: Jd–diabase. Black dots and bold numbers represent the locations of Wiley and Sutton Creek Gneiss geochronology and geochemistry samples of Carrigan et al. (2003) and the Wolf Creek Gneiss sample (Hatcher et al., 2004). 1–Sutton Creek Gneiss along Raper Creek (sample SC–RC). 2–Wiley Gneiss, U.S. 23-441 Wiley, Georgia. 3–Wolf Creek Gneiss, Stekoa Creek Road near Stekoa Creek (sample WC). Modified from Hatcher and Hopson (1989).

Figure 11. Folded block-in-matrix structures at Savannah Church that were cut by late pegmatite veins.

detrital zircons were identified (Bream, 2003). The Tallulah Falls Quartzite zircon suite, however, is the same as that in the Tallulah Falls Formation and Ocoee Supergroup (Figure 6). If the interior Tallulah Falls Quartzite and metasandstone are different, another "unique" fault is required beneath the even narrower strip of hanging wall separating the Dahlonega gold belt Otto Formation and Lake Burton mafic-ultramafic complex from the interior of the Tallulah Falls dome (Fig. 10).

The bottom line may be the Dahlonega gold belt and interior Tallulah Falls dome rocks are both Otto Formation, and all of these rocks are really distal Great Smoky Group rocks. If so, this solves a major problem in the emplacement kinematics of Gondwanan rocks beneath Laurentian hanging wall (Cartoogechaye, Cowrock, and Tugaloo terranes) and Lauretian margin footwall rocks. Hadley's (1970) initial conclusion may therefore be correct after all.

Block-in-Matrix Structures

Block-in-matrix structures have been described at several localities in the central Blue Ridge, including several in the area of the 2005 CGS field trip (e.g., Stops 2–1 and 2–4) (Raymond et al., 1989) (Fig. 11). They consist of a diverse assemblage of rock types dominated by amphibolite in a matrix of biotite gneiss and migmatitic leucosome. They have been interpreted as mélanges (Lacazette and Rast, 1989; Raymond et al., 1989), but the strain state and metamorphic grade are so high that they could be dismembered amphibolite layers (boudinage) in a matrix of partial melt. The exposure at Savannah Church locality (Stop 2–1), however, contains blocks of calc-silicate, hornblende quartzite, and metasandstone (biotite-quartz-feldspar gneiss) in addition to the dominant amphibolite (Raymond et al., 1989), so these bodies may have a more complex origin than boudinaged amphibolite layers plus melt.

Mafic-Ultramafic Complexes

Larrabee (1966) compiled the known ultramafic bodies in the Appalachians, and revealed that they are distributed throughout the Cartoogechaye, Cowrock, and Tugaloo terranes, and the Dahlonega gold belt, in the present tectonic framework (Ryan et al., this guidebook; Swanson et al., this guidebook). Detailed geologic mapping during the past several decades has determined that many (most?) ultramafic bodies are associated with various kinds of mafic rocks and should be considered mafic-ultramafic complexes (Hatcher et al., 1984). Many of the ultramafic bodies in the region thus have an origin that is tied to the mafic rocks with which they are associated, but several possibilities exist for their origin (Fig. 12) (Hatcher et al., 1984). Early geochemical work during the 1980s (e.g., McElhaney and McSween, 1983; Hatcher et al., 1984; Hopson, 1989) suggested the mafic rocks have a noncontinental origin, and recent work

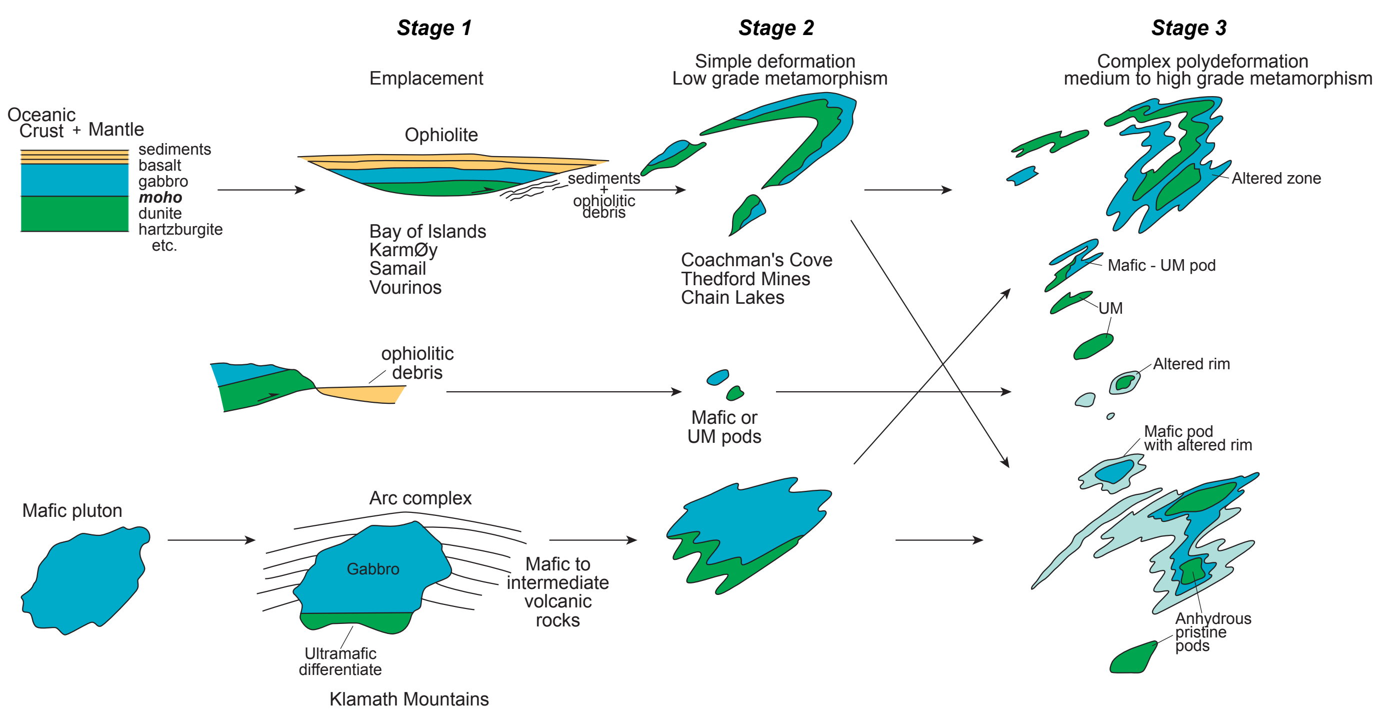

Figure 12. Alternative possibilities for the origin of deformed mafic-ultramafic complexes in orogenic belts. (After Hatcher et al., 1984.)

has corroborated these early conclusions (Berger et al., 2001; Raymond et al., 2001; Ryan et al., this guidebook; Swanson et al., this guidebook; Warner, this guidebook; and references cited therein). While most of these complexes may be ophiolite fragments, the possibility remains that some of them may be parts of differentiated plutons, or, in the case of the large numbers of small bodies of both amphibolite and ultramafic rocks, ophiolitic debris (Hatcher et al., 1984) (Fig. 12).

Detrital Zircons, Ages of Plutons, and Timing of Metamorphism

Modern geochronologic data in some cases have reconfirmed earlier age dates, but they also have provided new insights into the age of Blue Ridge plutons, timing of metamorphism, and provenance of rifted-margin, Cartoogechaye terrane, Dahlonega gold belt, and Tugaloo terrane rocks, and brought new problems to light (Tables 1–5). A major conclusion of detrital zircon studies is no (meta)sedimentary rocks younger than Ordovician exist at the latitude of the Carolinas and northern Georgia from the western Blue Ridge to the western margin of the Cat Square terrane (eastern Inner Piedmont), and all have a Laurentian source (Bream, 2003) (Fig. 6). Detrital zircon data again confirm that the Brevard fault is not a suture.

New geochronologic data from the western Tugaloo terrane Rabun and Walnut Creek plutons (D. W. Stahr, 2005, unpublished; Table 5) revise the published 374 Ma age date on the Rabun pluton (Miller et al., 2000) to ~335 Ma (dating both the megacrystic and equigranular phases), with a similar age on the Walnut Creek pluton. These ages reconfirm the 335 Ma TIMS age of Miller and Stewart (2002). Assuming the age of both the Rabun and Walnut Creek plutons is late Mississippian, this raises new questions about their origin and identifies them as the westernmost Alleghanian plutons in the Appalachians. The origin of the undeformed Carboniferous plutons in the Atlanta area (e.g., Stone Mountain, Ben Hill, etc.; Higgins et al., 2003) is difficult to understand, although Sinha and Zietz (1982) suggested that they are part of an Alleghanian plutonic arc. The deformed Rabun and Walnut Creek plutons are even more difficult to understand not only because they are located in the westernmost part of the Tugaloo terrane, but also because they were so strongly deformed during and immediately after emplacement (Figs. 2b and 10a). The Rabun is truncated by the Chattahoochee–Holland Mountain fault, and this fault was also tightly folded after emplacement. These relationships confirm that: the crust was still hot when the Rabun and Walnut Creek plutons were emplaced, and the Chattahoochee–Holland Mountain fault formed during the Alleghanian.

Timing of metamorphism in the western and central Blue Ridge has been confirmed to be Ordovician (Moecher et al., 2004), but that in the Tugaloo terrane appears to be Neoacadian (Table 5, Fig. 10a) (Carrigan et al., 2001; D. W. Stahr, unpublished data; A. J. Merschat, unpublished data). Both monazite and zircon ages (Moecher and Miller, 2000; Moecher et al., 2004) have documented the Ordovician event. The 360-350 Ma Neoacadian event is well confirmed as the dominant event in the Tugaloo and Cat Square terranes (Bream, 2003), suggesting that subduction of the Inner Piedmont and eastern Blue Ridge beneath the docking Carolina superterrane is responsible for the mid-Paleozoic metamorphic event and the contrasts in deformational style in the eastern Blue Ridge and Inner Piedmont (Merschat et al., 2005; Hatcher and Merschat, in press). The intriguing aspect of the 360-350 Ma event is that it may have overprinted the western Blue Ridge in the Carolinas and northeastern Georgia. Connelly and Dallmeyer (1993), Naeser et al. (this guidebook), and Southworth et al. (this guidebook) have reported several 360 Ma ages using a variety of age dating techniques in the western North Carolina Blue Ridge. Docking of Carolina superterrane accompanied by subduction of all of Tugaloo and Cat Square terranes to depths sufficient to produce upper amphibolite to lower granulite facies metamorphism in these terranes. This could easily produce more far-field effects such as the 360 Ma thermal effects, and possible faulting or fault reactivation, in the western Blue Ridge.

Post-Paleozoic History

The Blue Ridge is a remnant highlands of the Appalachian Mountains formed 320-270 Ma by the collision of Africa with North America producing a chain comparable in height to today's North American Cordillera, the Alps, or Himalayas. Erosion during the Mesozoic might have initially reduced the mountainous Blue Ridge to a low plateau like the Piedmont, but renewed uplift during the Tertiary following Cretaceous stress reversal from tensional to compressional in eastern North America (Prowell, 1988) has produced the Blue Ridge and other highlands topography in the western Piedmont and Valley and Ridge-Plateau that we see today.

The North Carolina-Tennessee Blue Ridge contains the highest mountains in the Appalachians from Alabama to Newfoundland. Despite this, few outside of this region have heard of most of the high mountains in this area, but, ironically, most have heard of Mt. Washington, New Hampshire. There are 20 mountains in North Carolina and Tennessee higher than Mt. Washington, which may be the only named mountain higher than 6,000 ft in New England, while there are 56 mountains in North Carolina-Tennessee higher than 6,000 ft (Table 6). The mountains in New England, however, involve consistently higher relief than those in the southern Appalachians, because the valleys there are mostly closer to sea level. In the southern Appalachians, however, many valleys in the Blue Ridge are above 3,000 ft so the relief to the higher summits is much less, yielding moderate relief in the North Carolina-Tennessee Blue Ridge. The vertical distance from Waynesville (2,644 ft) to the top of Waterrock Knob (6,292 ft) is >3,600 ft over a distance of little more than 7 mi. From Gatlinburg, Tennessee, (~1,400 ft) to the top of Mt. LeConte (6,593 ft), the relief is ~5,200 ft in a distance of <5 mi. In the east-west valley along U.S. 321 east of Gatlinburg at ~2,000 ft, the relief is >4,600 ft between the valley and the top of Mt. Guyot (6,621 ft) ~ 6 mi south. An interesting comparison is from Denver, Colorado, (~5,000 ft) to the top of Mt. Evans (14,148 ft) some 40 mi west, the relief is a little greater than 9,000 ft.

The higher parts of the southern Blue Ridge today comprise a temperate rainforest, with rainfall estimated to range from 65 to just under 100 in per year in the Smokies, depending on elevation (http://www.nps.gov/grsm/gsmsite/natureinfo.html). Continuous rainfall records from 1934 at the U.S. Forest Service Coweeta Hydrologic Laboratory, located south of Franklin, North Carolina, reveal an average rainfall of ~70 in per year at the headquarters (2,249 ft elev.) to ~90 in per year on Albert Mountain (5,223 ft), the highest point in the laboratory located

Table 6. Elevations of ≥6000' mountains in North Carolina and Tennessee

Name	Elevation	Name	Elevation	Name	Elevation
Mt. Mitchell	6684	Roan High Knob	6285	Mt. Love	6098
Mt. Craig	6647	Roan Mountain	6285	Reinhart Knob	6095
Clingmans Dome	6643	Mt. Lyn Lowry	6280	Craggy Dome	6090
Mt. Guyot	6621	Browning Knob	6250	Thermo Knob	6090
Balsam Cone	6600	Percys Peak	6240	Plott Balsam	6088
Cattail Peak	6600	Luftee Knob	6234	Tennet Mountain	6060
Mt. LeConte	6593	Gibbs Mountain	6220	Grassy Cove Top	6055
Mt. Buckley	6535	Mt. Kephart	6217	Sam Knob	6055
Big Tom	6526	Black Balsam Knob	6214	Patton Knob	6045
Mt. Gibbes	6520	Winter Star Mountain	6203	Yellow Face	6032
Potato Hill	6440	Grassy Ridge Bald	6189	Cold Mountain	6030
Potato Knob	6420	Mt. Collins	6188	Chestnut Bald	6025
Mt. Chapman	6417	Tricomer Knob	6180	Shining Rock	6010
Richland Balsam	6410	Marks Knob	6169	Big Butt	6000
Old Black	6370	Deer Mountain	6160	Balsam Corner	6000
Hallback	6360	Big Cataloochee Mtn	6155		
Blackstock Knob	6330	Mt. Yonaquska	6135		
Celo Knob	6327	Mt. Hardison	6134		
Clingmans Peak	6320	Mt. Ambler	6130		
Waterrock Knob	6292	Horse Rock	6120		
Mt. Washington, New Hampshire	***6288***	Mt. Hardy	6110		

* From http://ww.americasroof.com/highest/nc.shtml

<3 mi from the headquarters. The maximum rainfall measured in one year at Albert Mountain is 140 in (http://cwt33.ecology.uga.edu/catalog_monthly_noaa.html). The huge measured variations in rainfall over short distances may be representative, but we really do not know how much rainfall occurs on the tops of the highest mountains, because there are few permanent or seasonal meteorological stations on the high mountains in this region.

Timberline in Colorado is ~10,500 ft, and in northern New England is ~5,000 ft. Block fields and pattern ground near the tops of the highest mountains in North Carolina and Tennessee indicate timberline was possibly as low as 5,000 ft during Pleistocene glacial maxima, and may locally have been as low as 3,500 ft (Hadley and Goldsmith, 1963; King, 1964).

Today's Appalachian topography may not have developed by uniform static erosion of the end-of-Paleozoic mountains through the varied climates of the Mesozoic and Cenozoic, as has recently been suggested (Naeser et al., 2001; Matmon et al., 2003). Instead, it may be the product of a crust that was under tension during the initial stages of rifting of Pangaea that formed the Mesozoic basins, the Cretaceous-Tertiary trailing margin, the present Atlantic and the North American Plate. The early unroofing history is revealed by the zircon fission-track data of Naeser et al. (2005; this guidebook), where they noted that zircon fission tracks from Valley and Ridge sedimentary rocks retain the ages of their original sources. Blue Ridge zircons, however, were reset ~360 Ma, and those in the Piedmont were reset ~280 Ma, reflecting a Paleozoic west-to-east uplift history. The mid-Mesozoic tensional stress regime was inverted during the Late Cretaceous yielding a compressional field that remains today throughout eastern North America (Prowell, 1988; Zoback and Zoback, 1991). Blue Ridge elevations today decrease uniformly to the southwest until all that remains of the mountainous topography is a few outliers, like Talladega and Cheaha Mountains in Alabama, and the entire Blue Ridge geologic province passes beneath the Coastal Plain a short distance to the southwest (Osborne et al., 1989) (Fig. 13). The present major river drainages that cross the mountains to the west into the Ohio, like the Tennessee and its tributaries, or the New River (North Carolina, Virginia, and West Virginia), head in the eastern Blue Ridge and have more gentle gradients. Streams that drain eastward across the Piedmont into the Atlantic have steeper gradients and over time have beheaded the more perched upper reaches of the Tennessee, and even the Chattahoochee that drains into the Gulf of Mexico (Acker and Hatcher, 1970; Hack, 1982). Prowell and Christopher (2000) concluded the truncation of the entire Appalachians by the Coastal Plain in Alabama is evidence that the mountain chain had been eroded flat and uplifted again during the Tertiary northeast of the Coastal Plain overlap.

Topographic expression of bedrock structure and lithology is very clear in the Valley and Ridge where tilted rocks with markedly different resistance to weathering and erosion occur and surface processes have taken advantage of these contrasting processes (Figs. 13 and 14). Subtle differences between ridge-forming rock units can be distinguished with a little background on Valley and Ridge rock units, e.g., the sharp, "cockscomb" ridges held up by the Rome (Lower Cambrian) Formation and Rockwood–Clinch (Lower Silurian) sandstones, and the broad rolling ridges underlain by different units of the Knox Group. In contrast, the Blue Ridge and western Piedmont are composed largely of metasandstone, schist, metavolcanic rocks, and granitoids, with lesser amounts of ultramafic and carbonate rocks. While ultramafic bodies host a unique potassium-starved flora (Pittillo et al., 1998), they are commonly not large enough to influence the topography. Blue Ridge topography is thus more random, although detailed geologic mapping frequently reveals subtle relationships of bedrock to topography. Less subtle features include the Murphy syncline, which is evident on any scale of observation (Fig. 14); the frontal Blue Ridge in Georgia, Tennessee, Virginia; the Grandfather Mountain window; and the Brevard fault zone with its carbonate and highly deformed fine-grained rocks. All contain greater diversity of rock types.

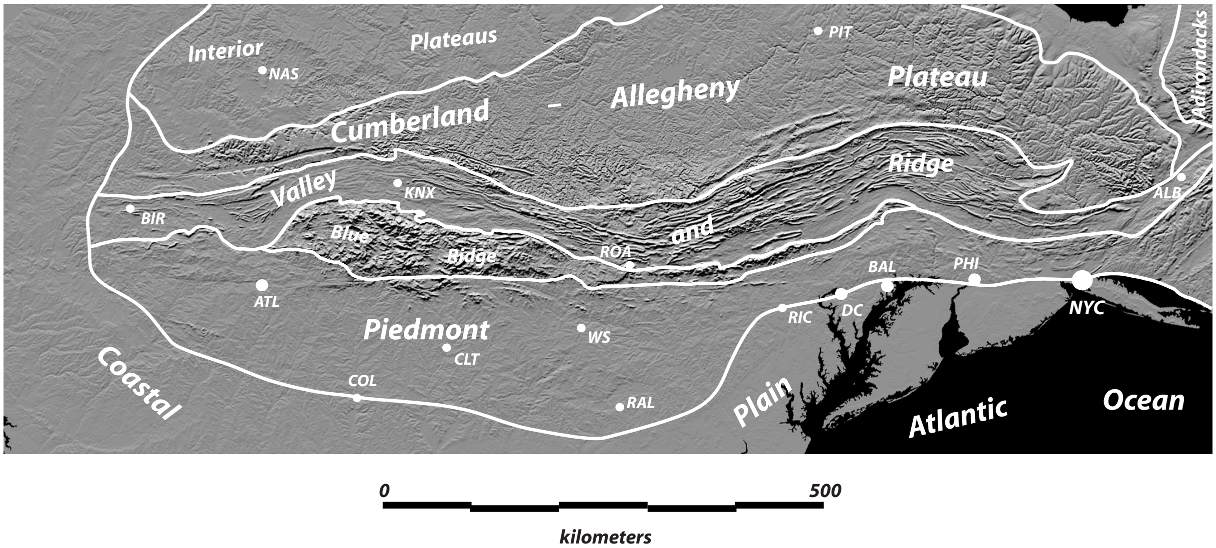

Figure 13. Southern and central Appalachians digital elevation model (USGS).

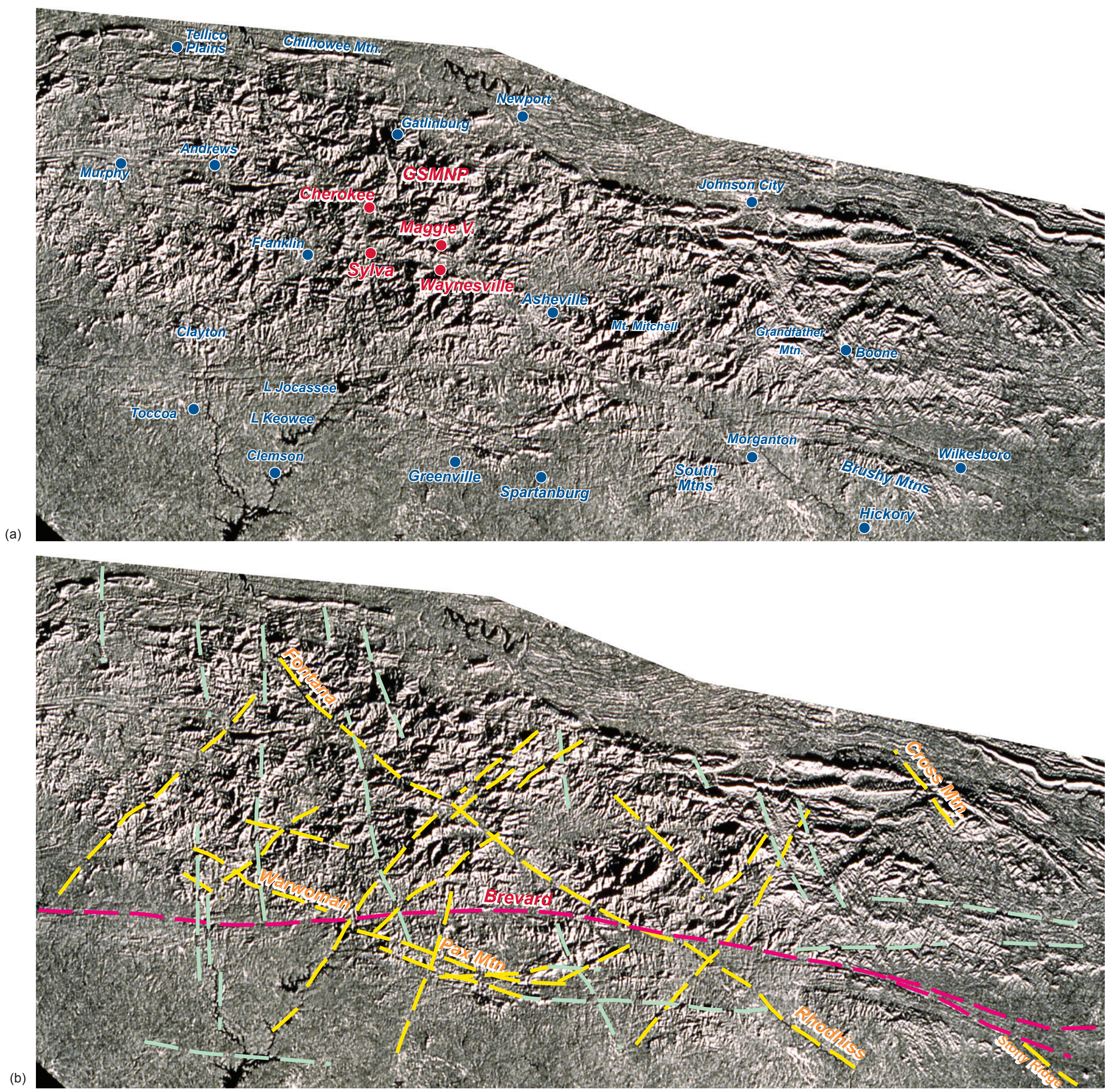

Figure 14. Side-looking radar image parts of the southern Blue Ridge and adjacent parts of the Valley and Ridge and Piedmont with geographic labels (a) and interpretation of lineaments and other geologic features (b).

Here also are several lineaments, in addition to the Brevard fault zone, which were interpreted from topographic maps (earlier work by Acker and Hatcher, 1970), digital elevation models, and side-looking radar images (Fig. 14). Such interpretations are made with the pitfalls described by Wise (1982) in mind. These lineaments are mostly related to fractures and post-Paleozoic faults. Secondary drainages in the western Piedmont and Blue Ridge are partially controlled by prominent N45°W, N45°E, N10°W, and N70-80°E and N70-80°W joint sets (Acker and Hatcher, 1970; Schaeffer et al., 1979). These fractures probably formed during the late Paleozoic (NW and NE sets) and Mesozoic (~N-S and E-W sets). In addition, there are several prominent nearly E-W (to ENE) fault-related lineaments in northwestern Georgia, the Carolinas and eastern Tennessee: the Warwoman (and smaller parallel Lake Rabun) in northeastern Georgia (Hatcher, 1974) that projects into the Marietta-Tryon fault system (Garihan et al., 1993), which consists of several named faults, with the Cross Plains fault being the first recognized (Birkhead, 1973; Snipes et al., 1979); the Stony Ridge fault system in the Sauratown Mountains window (Stirewalt, 1969); and others (see Garihan et al., 1993, their Fig. 1). Despite speculation by Dennison and Stewart (2001) that these are Tertiary structures, they probably are coupled with Late Triassic-Jurassic extension and dike emplacement. Garihan et al. (1993, p. 55-56) noted that Mesozoic diabase dikes cut one or more of these faults, but in places diabase dikes are also displaced along these faults, strongly indicating they are Mesozoic structures. The Fontana–Rhodhiss lineament is one of the longer fault systems in this group. The almost E-W Cross Mountain fault and another subparallel to it in the frontal Blue Ridge and Valley and ridge of northeastern Tennessee (King and Ferguson, 1960) are probably also part of this system (Fig. 14).

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An overview of the Mesoproterozoic basement complex on the west half of the Asheville 1:100,000 scale geologic map

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ABSTRACT

Bedrock geology of all 7.5 minute quadrangles on the west half of the Asheville 1:100,000-scale map has now been mapped in detail by the North Carolina Geological Survey (NCGS) and other mappers with financial assistance from the STATEMAP program. A preliminary map by the NCGS is presented here as part of an ongoing compilation project. Mesoproterozoic basement rocks in the compilation area consist of five major units: four highly deformed metamorphic units and the relatively undeformed Max Patch Granite. Contacts between the four metamorphic units are still uncertain but may be either faults, intrusive, or gradual changes of dominant rock type. A newly dated Neoacadian mylonitic shear zone separates Max Patch granite from the highly deformed metamorphic basement rocks. To the southeast, basement rocks are in fault contact with the Ashe Metamorphic Suite along the Chattahoochee-Holland Mountain Fault. To the northwest, Ocoee Supergroup sedimentary rocks are separated from the basement along unconformities and Paleozoic fault zones. Paleozoic metamorphic overprinting intensifies from the northwest to southeast but was not complete in all areas. Granulite-facies rocks are present in localized bodies of the metamorphic basement units and likely preserve Grenvillian foliations.

INTRODUCTION

This paper focuses on the history and relationships of the igneous-metamorphic basement rocks in the western half of the Asheville 1:100,000-scale (100k) geologic map. Crystalline basement of western North Carolina comprises the Mesoproterozoic metamorphic core of the southern Appalachian Mountains. Geographically, it separates the Neoproterozoic Ashe Metamorphic Suite/Tallulah Falls Formation from Neoproterozoic rocks of the Ocoee Supergroup (Fig. 1).

Each of the sixteen 1:24,000-scale quadrangles that make up the western half of the 100k sheet has been mapped in detail by the North Carolina Geologic Survey (NCGS) and contract geologists. Most of these maps were completed under the auspices of the National Cooperative Geologic Mapping Program, STATEMAP component.

Surprisingly, very little detailed mapping had previously been conducted in this area. Arthur Keith (1904) produced a 1:125,000 scale map of the region, subdividing basement rocks into four broad units: Carolina Gneiss, Roan Gneiss, Cranberry Granite, and the Max Patch Granite. The 1958 North Carolina Geologic State Map (NCGS, 1958) was published at a 1:500,000 scale and recognized four major basement units: Mica gneiss, Hornblende gneiss, Cranberry granite gneiss, and Max Patch granite gneiss. Hadley and Nelson (1971) mapped the area at 1:250,000 scale and divded the basement into two major units: Layered gneiss/migmatite and the Max Patch Granite/Cranberry Gneiss. The 1985 North Carolina Geologic State Map (NCGS, 1985) named four basement units: Migmatitic Biotite Hornblende Gneisses, Biotite Granitic Gneiss, Granodioritic Gneiss, and Max Patch Granite. In addition, the 1985 map includes a Biotite Gneiss unit of questionable Mesoproterozoic age.

Cattanach, B. L., and Merschat, C. E., 2005, An overview of Middle Proterozoic basement complex on the west half of the Asheville 1:100,000 scale geologic map, *in* Hatcher, R. D., Jr., and Merschat, A. J., eds., Blue Ridge Geology Geotraverse East of the Great Smoky Mountains National Park, Western North Carolina: North Carolina Geological Survey, Carolina Geological Society Annual Field Trip Guidebook, p. 25-32.

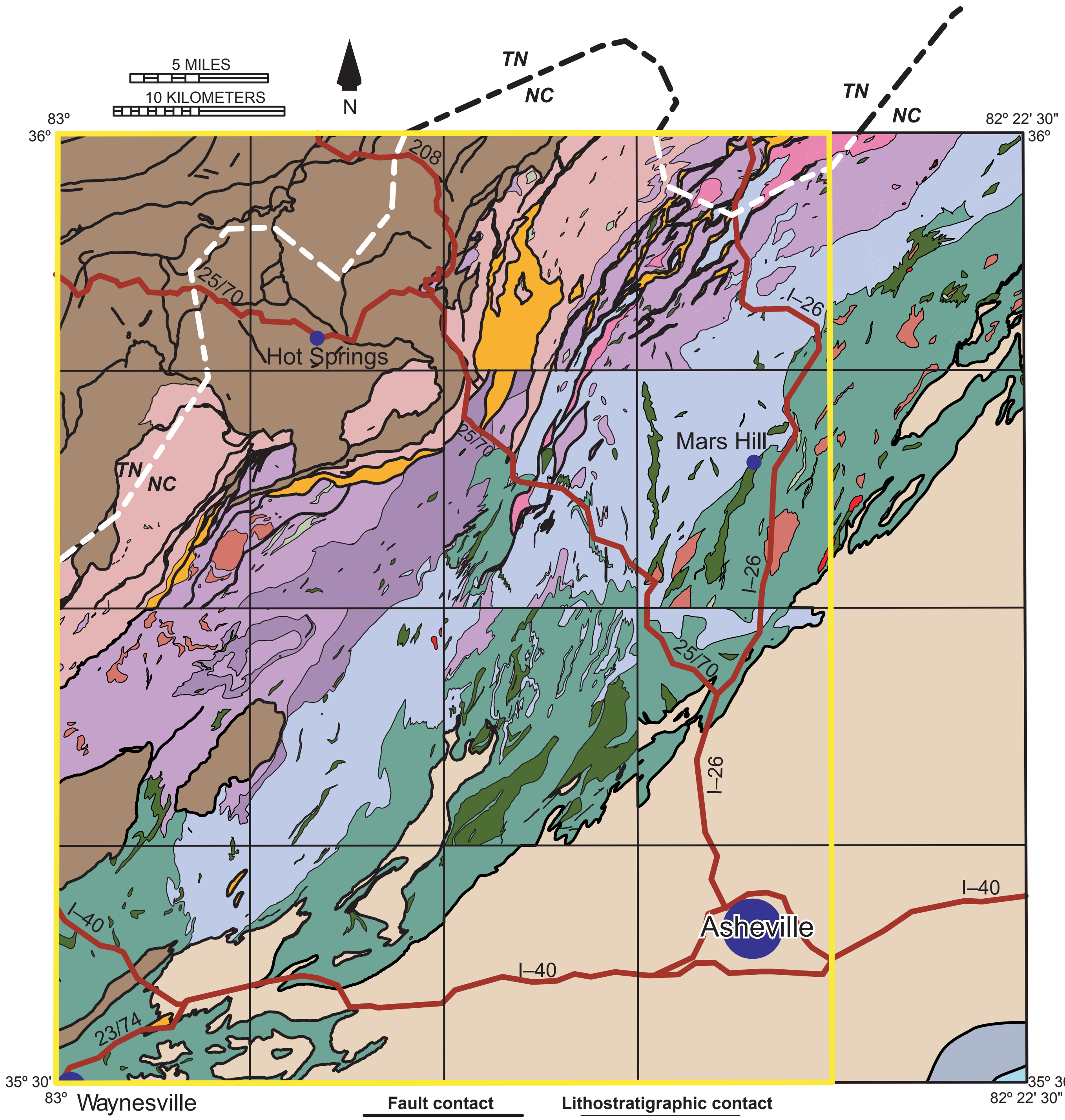

Figure 1. Preliminary compilation of the western portion of the Asheville 1:100,000 scale geologic map (west half shown by yellow box). Original digital cartography by Mark W. Carter.

Various other workers have completed mapping projects contributing to the geologic interpretation of the Asheville sheet (Finlayson, 1957; Brewer, 1986; Burton, 1996) but none has compiled such a wide area of detailed geologic mapping. Although the NCGS compilation is preliminary, we here outline the map units, their ages as currently known, and their structural and metamorphic histories.

Several units on the separate 1:24,000-scale quadrangles are possibly equivalent but have different formal names. In order to simplify the initial map, formal unit names are not used except for the Max Patch Granite. We instead name these units for the major rock types that define the map patterns. All of the various units mapped in each quadrangle have been correlated in Table 1 and correspond to map units in the 1:100,000-scale Asheville compilation.

MAP UNITS

The basement rocks consist of five major units (Fig. 1): migmatitic biotite-hornblende gneiss, layered biotite granitic gneiss, protomylonitic granitoid gneiss, biotite granitoid gneiss, and the Max Patch Granite and equivalents. These five units

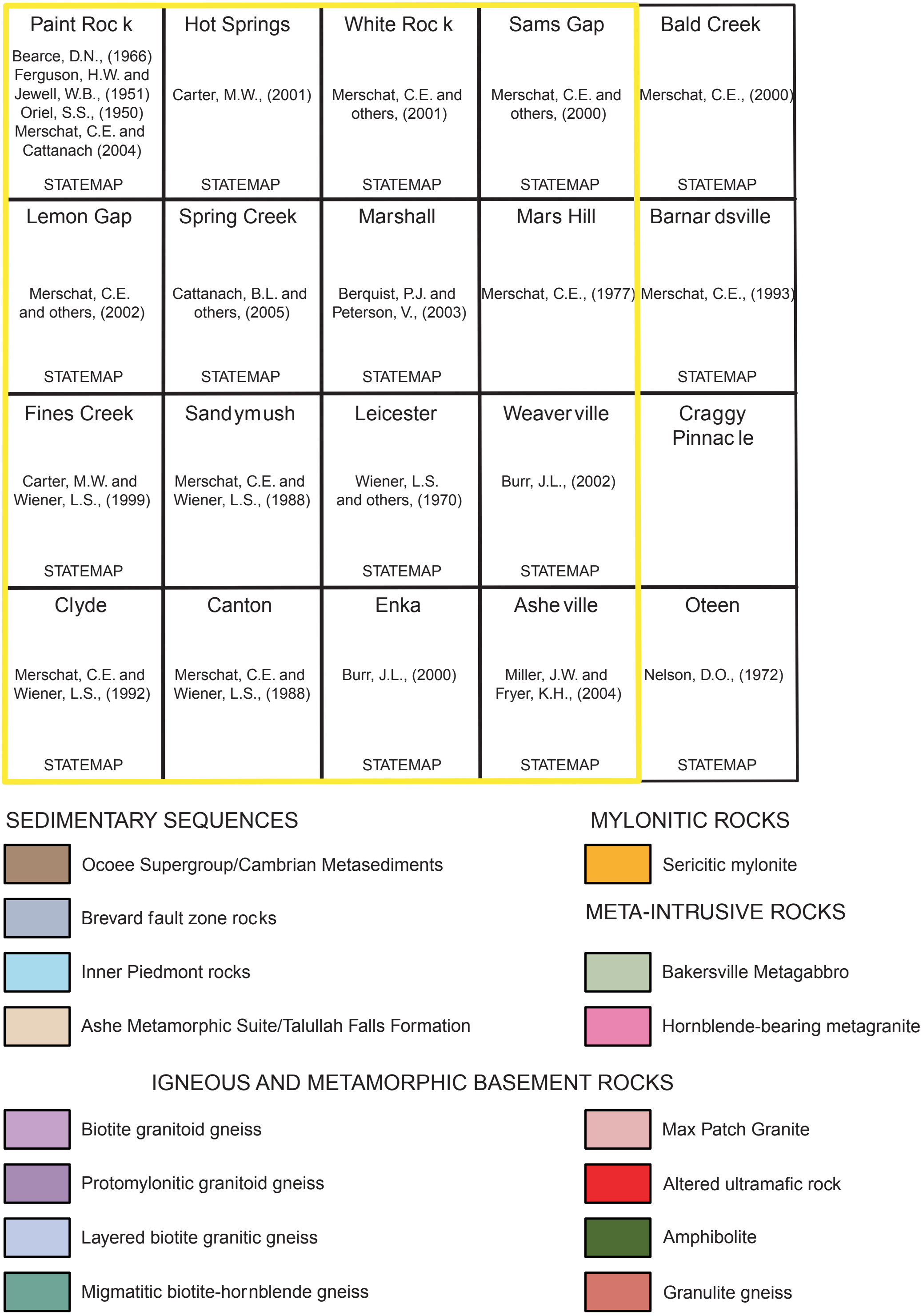

Figure 1. (continued)

can be grouped into two broad categories based on age and deformational style: 1) the younger and relatively undeformed Max Patch Granite, and 2) the older but highly deformed rocks of the migmatitic biotite hornblende gneiss, layered biotite granitic gneiss, protomylonitc granitic gneiss, and biotite granitic gneiss. Minor components of the older units include amphibolite, altered ultramafic rocks, hornblende-bearing granite, calc-silicate bodies and granulite gneisses.

The common rock types associated with the older basement gneisses are found in several different map units, thus making it difficult to compile an easily recognizable "stratigraphy" that can be applied to the area. Detailed field mapping at 1:24,000-scale does, however, make it possible to recognize proportional differences between rock types and mineralogy within the various map units. For example, although it is possible to find biotite-bearing granitic gneisses in all of the major basement map units, there is much less of this rock type in the migmatitic biotite-hornblende gneiss unit than in the biotite granitoid gneiss unit.

Migmatitic Biotite-Hornblende Gneiss: 1210 Ma (Fullagar et al., 1979). This is a strongly metamorphosed and deformed gneiss that contains a higher percentage of amphibolite and hornblende-bearing gneisses than the layered biotite granitic

Table 1. List of all 7.5' quadrangle formal and informal unit names grouped within map units shown on the west half of the Asheville 100k initial compilation (Fig. 1).

Basement Rock Summary

Migmatitic Bio-Hbl Gneiss	Biotite Granitoid Gneiss	Granulites
Bg- Biotite gneiss	Hbgn- Hornblende-biotite gniess	Bhgb- Hypersthene metagabbro
Bgn- Biotite gneiss	Pzc- cataclasite	Hygg- Hypersthene granitic gneiss
Bhg- Biotite-horn gneiss	Pzfgr- Leucocratic Granite	Hyp- Hypersthene plagioclase rock
Bhm- Bio/horn migmatite	Ybg- Biotite granitoid gneiss	Ygbg- Granoblasitc biotite granitic gneiss
Hbgn- Hornblende-biotite gniess	Ybga- Amphibolite	Yggg- Garnet-rich granulite gneiss
Mbg- Magnetite granitic gneiss	Ybggf- Aluminous granofels	Ygqm- Quartz Monzodioritic Granulite
Mss- Metasandstone	Ybgh- Hornblende granitoid gneiss	Ygt- Tonalitic granulite
Yam- Amphibolite	Ybgm- Mylonitic bio granitoid gneiss	Yhyp- Pyroxene granulite
Ybhg- Bio/horn gneiss	Ybgmy- Mylonite gneiss	Ymg- Mafic granulite
Ybhm- Migmatitic biotite-hornblende gneiss	Ybgmy- Mylonitic biotite granitic gneiss	
Ybn- biotite gneiss	Ymb- Marble	
Ycs- Calc-silicate granofels	Ybgn-Feldspar granitoid gneiss	**Ultramafic Rocks**
Ydn-Biotite dioritic gneiss	Yggg- Granoblastic granitic gneiss	Zud- Dunite body
Ye- Earlies Gap	Ylg- Leucocratic granitoid gniess	Zua- Altered ultramafic
Yea- Earlies Gap amphibolite	Ypzpbg- Biotite granitoid gneiss	Yut- Talc body
Yfg- felsic gneiss	Ysc- Spring Creek granitoid gniess	Um- Ultramafic rocks
Ymbcs- Marble + calc-silicate	Ysca- Interlayers of amphibolite	Yum- Ultramafic rocks
Ymg- Magnetite gneiss	Yscm- Protomylonitic Spring Creek granitic gneiss	Du- Dunite body
Ypzpghg- Granitic and bio/horn gneiss	Yscmy- Mylonitic Spring Creek granitic gneiss	
Yr- Richard Russell Bio Gneiss	Yscs- Calc-silicate granofels	
Yra- Richard Russell Amphibolite	Yscu- Mylonitic Spring Creek granitic gneiss	**Miscellaneous**
Yrg- Richard Russell Metagraywacke		Trondhjemite
Yrm- Richard Russell mus-bio gneiss		Pzm- Mylonite/protomylonite
	Max Patch and equivalents	Pzc- Cataclasite
	Pzcpmy- Protomylonite	Pzcf- Ferruginous cataclasite
Layered Bio Granitic Gneiss	Ym- Max Patch granite	Pzmy- Paleozoic Mylonite
bgg- biotite granitic gneiss	Ymfl- Leucocratic granite	
Gg- granite gneiss	Ymmy- Mylonitic granite	
Ybgg- layered biotite granitic gneiss	Ymp- Protomylonitic granite	**Bakersville Metagabbro**
Ybggmy- mylonitic layered biotite granitic gneiss	Ypzmmp- Mylonitic max patch	Zbg- Bakersville metagabbro
Yea- Earlies Gap amphibolite	Zgg- Max Patch	
Ys- Sandymush felsic gneiss	Zggmy- Mylonitic Max Patch	
Ysa- Sandymush amphibolite	Zggp- Protomylonitic Max Patch	
Yscs - Sandymush calc-silicate	Zm- Max Patch granite	
Ysm- muscovite bearing Sandymush felsic gneiss	Zymgm- Protomylonitic monzogranitic gneiss	
Ysp- Sandymush Protomylonite	Zymgu- Mylonitic monzogranitic gneiss	
Protomylonitic Granitoid Gneiss		
Ydg- Doggett Gap protomylonite		
Ypzpg- Protomylonitic granitic gneiss		
Yzcbgg- Cranberry granitic gneiss		
Yzkag- K-spar granite/granitic gneiss		
Ydgc- Coarse-grained Doggett Gap protomylonite		

gneiss to the northwest. This unit is commonly migmatized and contains lenses of calc-silicate, altered ultramafic rocks, and felsic and mafic granulite gneisses.

It is characteristically a light-brownish-gray to light-pinkish-gray, massive to well-foliated, locally migmatitic gneiss. It consists mostly of interlayered and intergraded biotite-hornblende gneiss and granitic gneiss. The biotite-hornblende layers predominate and may represent thin metamorphosed sills of gabbro or diabase. Locally, biotite-hornblende layers exhibit a chaotic block-in-matrix structure. The granitic layers range from granite to quartz diorite with most of the layers approximating quartz diorite. Migmatization is more prevalent near large bodies of mafic rock.

Layered biotite granitic gneiss: 1270 Ma (Fullagar, 1983). This is a thick, repetitive sequence of layered rocks dominated by biotite granitic gneiss to quartz dioritic gneiss interlayered with biotite gneiss, biotite schist and amphibolite. Layers range from millimeters to meters thick. It is composed of quartz, plagioclase, potassium feldspar, biotite, epidote group minerals, muscovite, opaques, garnet, titanite, apatite, and zircon and is locally migmatitic and mylonitic. Isoclinal folds are prominent. This unit is interpreted to be a highly deformed sequence of predominantly felsic metavolcanic rocks with a minor mafic component.

Protomylonitic granitoid gneiss: This unit consists of medium- to coarse-grained granitoid gneiss exhibiting a protomylonitic fabric consisting of feldspar porphyroclasts in a bioite-rich matrix. It has been mapped on the basis of this fabric and outcrops in a belt 25 km long and 6 km wide. It is unclear if the protomylonitic fabric is the result of Paleozoic or Grenville deformation. Close association and compositional similarities with the biotite granitoid gneiss raise the possibility that the two map units are the same lithologically, differing only in their deformational fabric.

Biotite granitoid gneiss: 1174-1360 Ma (Officer et al., 2003; Berquist, 2005). Highly metamorphosed and deformed gneiss that is typically more massive than the layered biotite granitic gneiss. It is well foliated and locally layered. Compositions of layering vary widely from granitic/granodioritic gneiss to biotite schist. Layer thickness ranges from decimeters to meters. Granitic/granodioritic gneiss is composed of potassium feldspar, quartz, plagioclase, biotite, muscovite, epidote group minerals, garnet, and opaques. The biotite granitic gneiss exhibits a strong mylonitic fabric in Middle to Late Paleozoic shear zones along its northwestern margin.

Max Patch Granite and equivalents: 1000-1100 Ma (Officer et al., 2003; Berquist, 2005). This is a large granitic body with an outcrop exposure averaging over 7 km wide and 35 km long on the 100k map. Composition ranges from granite to alkali-feldspar granite to granodiorite. It is typically massive to weakly foliated, but is strongly mylonitic along in Middle to Late Paleozoic shear zones. Amphibolite xenoliths are present locally in the unit. The unit is in fault contact with the biotite grantitic gneiss to the southeast and is unconformably overlain by Ocoee Supergroup metasedimentary rocks to the northwest.

MINOR MAP UNITS

These units include amphibolite, garnet-hornblende granofels, altered ultramafic rocks, calc-silicate bodies, marble, and granulite gneisses.

STRUCTURE

All basement map units follow the NE-SW trend prominent in this part of the Appalachian Orogen. To the southeast, the basement is in fault contact with the Ashe Metamorphic Suite/Tallulah Falls Formation. To the northwest it is both in fault contact with and unconformably overlain by the Ocoee Supergroup. A major shear zone separates the older, more deformed basement units from the Max Patch Granite (Fig. 2).

The vast majority of foliation measurements in the Asheville 100k align with outcrop patterns and are oriented NE-SW. Within the granulite bodies, however, most foliations trend NW-SE and are only locally overprinted by NE-SW-trending Paleozoic deformation. Merschat and Wiener (1990) suggested that these granulite facies NW-SE-trending foliations were formed during the Grenville orogeny. Recent geochronologic studies by Berquist (2005) support this interpretation. NW-SE foliations found in non-granulite assemblages elsewhere in the basement may also preserve this relict Grenville fabric. Such foliations are more common to the northwest because higher-grade Paleozoic metamorphism has largely overprinted much of the Grenville fabrics to the southeast (Cattanach et al., 2004; Merschat and Carter, 2004).

Southeastern boundary

The Ashe Metamorphic Suite/Tallulah Falls Formation cover rocks are separated from the migmatitic biotite hornblende gneiss and layered biotite granitic gneiss by the Holland Mountain fault. The Holland Mountain fault is likely a pre- to syn-metamorphic Taconic thrust fault that has subsequently been folded, thus producing isolated klippes of Ashe Metamorphic Suite/Tallulah Falls Formation within the basement suite (Merschat and Wiener, 1988). Hatcher (2004) correlated the Holland Mountain fault with the Chattahoochee fault and interpreted this contact to be a Neoacadian or Alleghanian thrust fault. Farther to the northeast, Stewart et al. (1997) interpreted the Burnsville shear zone to be an Acadian reactivation of the Holland Mountain fault with dextral strike-slip offset. Compilation of the west half of the Asheville 100k map will re-examine the Holland Mountain fault to determine its extent and geologic history.

Intra-basement unit boundaries

Internal boundaries between the highly deformed basement units are much less distinct and have been delineated based on the relative percentages of the major rock types. The overall map pattern exhibits a progressive trend from the high-grade mafic-rich biotite-hornblende migmatitic gneiss to the more felsic-rich biotite granitoid gneiss. It is unclear if this progression is representative of stacked thrust sheets or a heterogeneous nonconformable sequence.

The contact separating the protomylonitic granitic gneiss and biotite granitic gneiss from the Max Patch Granite is more distinct. Middle to Late Paleozoic mylonitic thrust zones place highly deformed basement rocks on the relatively undeformed and weakly metamorphosed Max Patch Granite. Shear-sense indicators and mineral stretching lineations indicate northwest directed thrusting. New Ar-Ar dating on sericitic muscovite in these shear zones yields ages of 355 to 336 Ma, suggesting that the deformation is Neoacadian (Southworth et al., this guidebook). An interesting feature of this mylonitization is that it affects a much wider area in the north-central part of the 100k map than farther to the southwest. To the southwest, most of the deformation associated with Neoacadian thrusting is confined to the sub-kilometer wide Meadow Fork fault zone and the discrete Cold Springs fault that terminates in the Max Patch Granite. The Meadow Fork fault zone increases in width to the northeast and joins an E-W trending fault zone along a lateral ramp bordering the Hot Springs window. Farther northwest, the mylonite zone is over 10 km wide in the White Rock and Sams Gap quadrangles. Mylonitic effects have also been mapped within the basement suite away from unit contacts. These mylonites form an anastomosing network of ductile deformation that locally overprints earlier foliations, but leaves nearby rocks unaffected.

The boundaries of the Mars Hill terrane, traditionally viewed as the extent of the biotite-hornblende migmatitic gneiss, will be examined to further define this important geological feature that is the focus of much new research. Preliminary findings suggest that outcroppings of Mars Hill terrane rocks in the Asheville 100k are discontinuous and more widespread than previously thought.

Northwestern boundary

The contact between the Max Patch Granite and the Neoproterozoic Ocoee Supergroup in the western portion of the Asheville 100k sheet is a nonconformity that is locally overprinted by Neoacadian mylonitization. To the northeast, this contact is mapped as a series of Paleozoic thrust faults.

AGE

U-Pb zircon ages have been reported by a number of workers that reinforce the two-fold division of the basement rocks (Fig. 2). The Max Patch crystallization age of 1019 Ma is slightly younger than those of the highly deformed basement gneisses and granulite bodies (1057-1375 Ma) (Berquist,

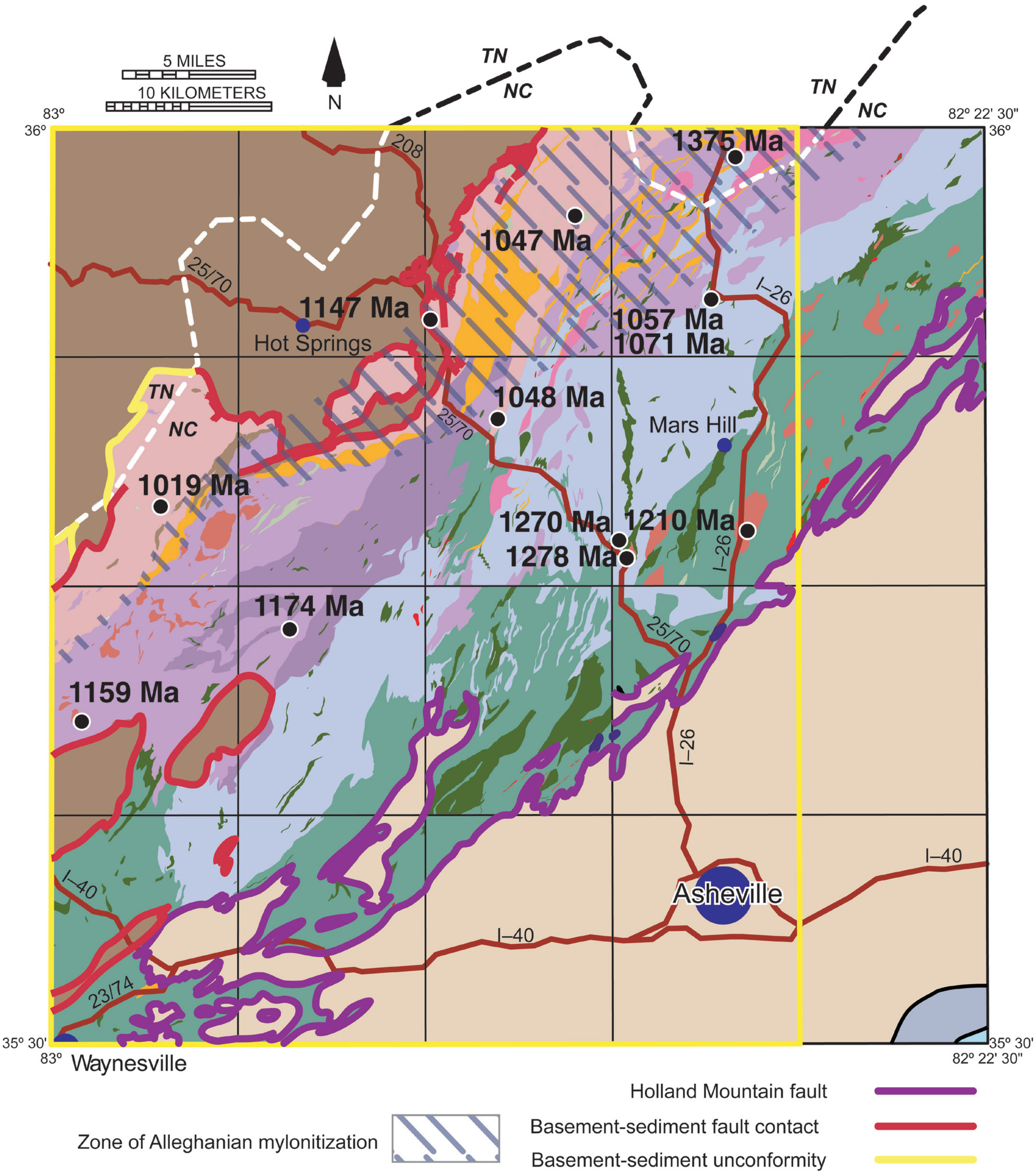

Figure 2. Map showing major structural boundaries and age dates for basement rocks of the west half of the Asheville 1:100,000 scale geologic map.

2005). We believe the Max Patch Granite was intruded after peak Grenville metamorphism and/or in a location distant from the Mesoproterozoic deformation affecting the other basement gneisses. In this scenario, the highly deformed and metamorphosed basement rocks were only juxtaposed with the pristine Max Patch during the Neoacadian. Alternatively, it is possible that some of the relatively undeformed granite bodies in the biotite granitic gneiss could be related to the slightly post-Grenville intrusion of Max Patch Granite or even younger intrusives.

METAMORPHISM

The basement rocks have a complex metamorphic history spanning at least 1.38 b.y. Unfortunately, it is difficult to interpret this history for many of the map units because they do not contain reliable metamorphic index minerals.

Three major periods of metamorphism have been identified. The earliest is Grenville granulite-facies metamorphism that produced garnet-hypersthene gneisses. These are present along the northwest margin of the deformed gneiss/Max Patch contact and in scattered bodies in the migmatitic biotite hornblende gneiss near the eastern margin of the basement rocks. We believe this grade of metamorphism was widespread in the basement gneisses, but few bodies persist because of later metamorphic overprinting. It is also possible, however, that the basement gneisses and amphibolites did not reach granulite facies conditions in all areas.

The second period of metamorphism is interpreted to be either Acadian or Taconic. Metamorphic intensity drops from sillimanite grade to the southeast (Ashe Metamorphic Suite) to biotite grade to the northwest (Max Patch Granite). Locally, metamorphic conditions were not sufficient to significantly retrograde the Grenvillian granulite assemblages. This could be related to lower temperatures and pressures or a lack of metamorphic fluids in dry granulite bodies.

The final metamorphic event reached chlorite grade and was associated with Neoacadian mylonitization. Effects of this event include sericitization and saussuritization and are limited to the areas of ductile shearing.

SUMMARY

The Mesoproterozoic basement rocks of the west-half of the Asheville 100k map consist of a series of highly deformed and metamorphosed gneisses and the relatively undeformed Max Patch Granite. These map units were juxtaposed along Neoacadian mylonitic fault zones. The northwestern boundary of the basement rocks is a nonconformity in some places and a Middle to Late Paleozoic fault in others. To the southeast, basement rocks are in fault contact with the Ashe Metamorphic Suite/Tallulah Falls Formation along the Chattahoochee fault/ Holland Mountain fault/Burnsville shear zone.

Grenvillian granulite-facies deformation is preserved in localized domains and commonly displays northwest-southeast-trending foliations. These fabrics were overprinted during lower-grade Taconic or Neoacadian metamorphism. Finally, chlorite-grade metamorphism occurred in areas affected by Middle to Late Paleozoic shearing.

Although the basic foundation is in place, many questions remain unanswered in the Blue Ridge basement rocks of the western half of the Asheville 100k sheet. Once synthesis is completed, the compilation will provide a detailed geologic framework on which further work and interpretation can be based.

ACKNOWLEDGEMENTS

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Geochemistry and U-Pb zircon geochronology of Blue Ridge basement, western North Carolina and eastern Tennessee: Implications for tectonic assembly

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ABSTRACT

New and recent geochronologic and isotopic data for Proterozoic basement of the western, central, and eastern Blue Ridge (WBR, CBR, EBR) place important constraints upon the origin, evolution, and assembly of ancient Appalachian crust. U-Pb (SHRIMP) geochronology document three major pulses of Proterozoic magmatism: 1.02-1.08, 1.13-1.18 Ga, and 1.22-1.27 Ga. The first of these is recorded only in the WBR, the second is common in all three areas, and the last is characteristic of the Mars Hill terrane (MHT) portion of the CBR. Ages on important regional units of the WBR include 1.02 Ga for Max Patch Granite and 1.17 Ga for Spring Creek granitoid gneiss. Three samples of basement from elsewhere in the CBR and EBR also yield ages in the older group. In addition, single MHT and WBR samples have magmatic ages of ~1.77 and 1.38 Ga, respectively. Early Proterozoic detrital and inherited ages are common in the MHT near Roan Mountain but absent elsewhere. Metamorphic overgrowths from 1.00-1.05 Ga are common throughout the Blue Ridge; older metamorphic rims ($\geq$~1.1Ga) for some WBR and CBR samples are rarer and imprecise. Except for 5 samples from the WBR with model ages of 1.3-1.5 Ga, all samples have Nd depleted mantle model ages >1.5 Ga. Most are in the range 1.5-1.7 Ga, except for the MHT, where ages near 2.0 Ga are typical. A well-defined $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ trend with elevated $^{207}Pb/^{204}Pb$, initially recognized by Sinha (1996), characterizes the southern and central Appalachian. The MHT at Roan Mountain has a steeper Pb/Pb trend and higher $^{207}Pb/^{204}Pb$.

These data suggest that the southern Appalachian basement comprises distinct zones that were assembled during Grenvillian and Appalachian orogenies. Furthermore, all of this basement appears to be out of place with respect to exposed or inferred basement to the northwest, which appears to fit simply into the southeastward younging trend of Laurentia. In comparison to the juvenile Grenville and mid-continent granite-rhyolite terranes, southern Appalachian basement has distinctly older Nd model ages and higher $^{207}Pb/^{204}Pb$ (at a given $^{206}Pb/^{204}Pb$). The apparent greater antiquity of southern Appalachian crust formation suggests that the southern Appalachian basement may be related to distant, older Laurentian and/or Gondwanan terranes, and that the southeastern Laurentian margin may be best characterized as a composite of accreted terranes.

Berquist, P. J., Fisher, C. M., Miller, C. F., Wooden, J. L., Fullagar, P. D., and Loewy, S. L., 2005, Geochemistry and U-Pb zircon geochronology of Blue Ridge basement, western North Carolina and eastern Tennessee: Implications for tectonic assembly, *in* Hatcher, R. D., Jr., and Merschat, A. J., eds., Blue Ridge Geology Geotraverse East of the Great Smoky Mountains National Park, Western North Carolina: North Carolina Geological Survey, Carolina Geological Society Annual Field Trip Guidebook, p. 33-44.

INTRODUCTION

Purpose

Basement rocks – defined here as those with protoliths of Mesoproterozoic or older age (>1 Ga) – provide the only direct record of the history of continental crust that predated the intense orogenies that affected the southern Appalachians during the Paleozoic. Recent studies of the southern Appalachian basement, especially when taken in the context of global interpretations of the Grenville orogeny and the construction of Rodinia, suggest that this crust may have a much more complicated history than previously suspected (e.g., Sinha et al., 1996; Loewy et al., 2003; Carrigan et al., 2003; Tohver et al., 2004; Ownby et al., 2004; see Tollo et al., 2004). Geochronologic and isotopic evidence call into question the notion that southern Appalachian basement is simply an extension of the crust of the North American/Laurentian interior. Furthermore, these data suggest the presence of distinct basement provinces in the southern Appalachians.

We present new geochemistry, Sm-Nd-isotope, and U-Pb zircon geochronology data from basement samples throughout the western, central, and eastern Blue Ridge provinces (WBR, CBR, and EBR, respectively) of western North Carolina and eastern Tennessee. (We herein define basement as rock older than 1.0 Ga). Samples were collected to address two broad objectives: (1) to characterize the age and crustal affinity of WBR basement, and (2) investigate the nature and distribution of "anomalous basement" that appears lithologically distinct from and potentially much older than the remainder of southern Appalachian crust. We combine our data with those of Carrigan et al. (2003) and Ownby et al. (2004) to better elucidate the age, origin, and distribution of southern Appalachian basement and to constrain its tectonic evolution with respect to adjacent terranes.

Geologic Setting

In the past, the core of the southern Appalachians has been broadly divided into two provinces: the eastern and western Blue Ridge (EBR and WBR; e.g., Hatcher, 1978) (Fig. 1). The EBR is separated from the Inner Piedmont to the southeast by the Brevard fault zone and the WBR from the Valley and Ridge province by the Blue Ridge fault system to the west. The EBR includes sparse Mesoproterozoic-aged basement and its origin is controversial; plausible interpretations suggest that it is rifted Laurentian basement (Hatcher, 1989) or accreted suspect terranes (of possible Gondwanan-affinity?) (Stewart et al.,1997). The WBR contains abundant Mesoproterozoic basement and is commonly considered to be of native Laurentian origin. Carrigan et al. (2003) characterized the basement of the EBR and WBR as generally granitic in composition with U-Pb zircon data that record a major magmatic pulse at 1.15-1.19 Ga, less abundant magmatism at ~1.08 Ga, and widespread Grenvillian metamorphism at ~1.03 Ga. More recently, the complex zone that had been labeled EBR has been subdivided into four distinct terranes: Cowrock, Cartoogechaye, Dahlonega gold belt, and Tugaloo (e.g., Hatcher et al., 2004). The Tugaloo terrane extends across the Brevard fault to encompass the western portion of the Inner Piedmont. In the scheme of Hatcher et al. (2004), the WBR is referred to as Laurentian margin.

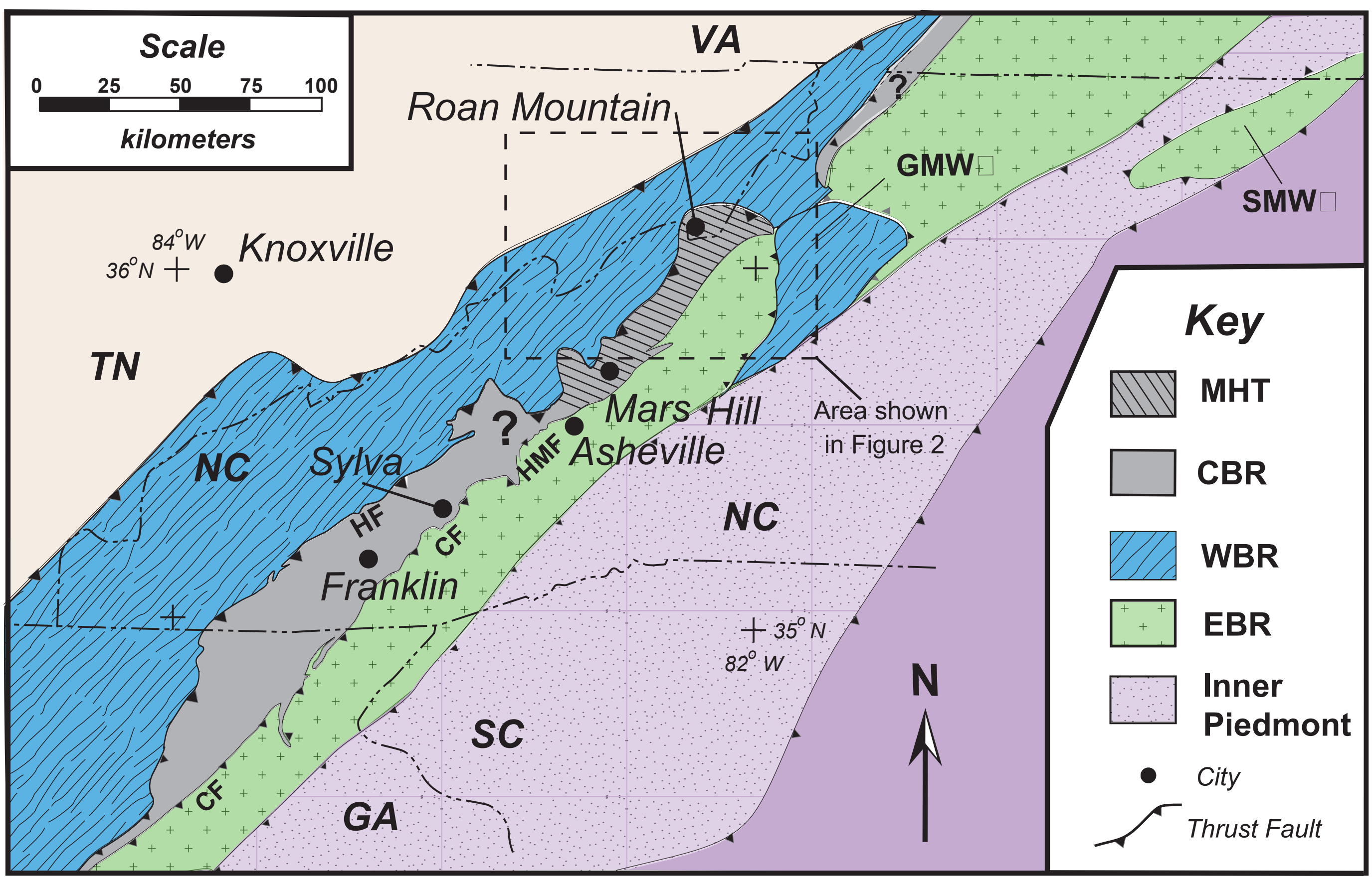

Figure 1: Generalized tectonic map of the southern Appalachians. HF = Hayesville fault, HMF = Holland Mountain fault, CF = Chattahoochee fault, GMW = Grandfather Mountain window, SMW = Sauratown Mountain window. Modified from Hatcher et al. (1990).

For simplicity, we retain the EBR and WBR terminology in this paper, but we add a central Blue Ridge that lies in between. The CBR comprises the Mars Hill, Cowrock, and Cartoogechaye terranes and the northern end of the Dahlonega gold belt as mapped by Hatcher et al. (2004); it is also equivalent to the central Blue Ridge plus the Mars Hill terrane and northern Dahlonega gold belt of Bream et al. (2004). Basement exposed within the CBR is generally more lithologically diverse than that of the EBR and CBR, and it is in part older than the EBR and WBR basement (Berquist et al., 2003; Ownby et al., 2004; this paper). In our tables and maps we include locality designations both as WBR, CBR, or EBR and within the terrane classification of Hatcher et al. (2004). The northernmost portion of this belt, in the vicinity of Roan Mtn, NC-TN, includes the oldest magmatic age, to our knowledge, in the Appalachians (whole rock Rb-Sr and U-Pb SHRIMP zircon ages of 1.8 Ga [Monrad and Gulley, 1983; Carrigan et al., 2003; Ownby et al., 2004]), ancient Sm-Nd and Pb-isotopic signatures (Sinha, 1996; Carrigan et al. 2003; Ownby et al. 2004), and diverse lithologies that experienced granulite-grade metamorphism during the Mesoproterozoic. Previous workers have identified similar lithologic characteristics within the CBR farther south to the Asheville area, and commonly referred to the Roan Mountain to Asheville portion of the CBR as the Mars Hill terrane (MHT). This belt is recognized by other names with varying boundaries (Merschat, 1977; Bartholomew and Lewis, 1988, 1992; Brewer and Woodward, 1988; Raymond et al., 1989; Merschat and Weiner, 1990; Johnson, 1994, Raymond and Johnson, 1994; Adams et al., 1995; Stewart et al., 1997, Carrigan et al., 2003; Ownby et al., 2004; Trupe et al., 2004). The southern extent of the MHT remains poorly constrained. In portions of the CBR south of Asheville field observations document high-grade, mafic to felsic orthogneiss basement, at least in part of Mesoproterozoic age (Hadley and Nelson, 1971; Brown et al., 1985; Raymond et al., 1989; Quinn and Wright, 1993; Berquist et al., 2003), which are characteristics reminiscent of the MHT (as defined by Ownby et al. 2004).

Samples

Representative samples of WBR rocks were collected along a NW-SE transect from newly exposed outcrops along Interstate 26, NW of the TN-NC state line (Fig. 2). This portion of the WBR is characterized by SE-dipping, NW-vergent imbricate thrust sheets decreasing in Paleozoic metamorphic grade toward the NW. Samples for petrographic analysis, geochemistry, Sm-Nd isotopes, and zircon geochronology were collected from each thrust sheet: the Sams Gap, Fork Ridge, Stone Mountain, and Pardee Point (Stewart et al., 1997; Trupe et al., 2004). Regionally extensive units were sampled elsewhere in the WBR basement, including the Max Patch Granite and granitoid gneiss and possible correlative units (Spillcorn and Indian Grave protomylonitic granitoid gneiss), and the Spring Creek granitoid gneiss. Additional samples were collected in the WBR for petrographic and geochemical analysis and combined with WBR basement data of Carrigan et al. (2003) (see Berquist (2005) for expanded sample descriptions).

Some basement exposures from the WBR, CBR, and EBR share some of the distinguishing characteristics of the MHT. We refer to these samples as "anomalous basement," because they contrast with the typical, broadly uniform basement of EBR and WBR (as described by Carrigan et al., 2003). The salient characteristics of the MHT that differentiate it from EBR and WBR basement, as defined by Ownby et al. (2004), and thus the characteristics that we use to define "anomalous basement," include: (1) highly variable geochemistry, (2) mafic to felsic orthogneiss interspersed on a range of scales, including amphibolitic (mafic) migmatites; (3) evidence for Mesoproterozoic granulite-grade metamorphism ; (4) Paleoproterozoic magmatic, inherited, or detrital zircon ages, and (5) Paleoproterozoic Nd depleted mantle model ages (T_{DM} ≥1.8 Ga). Migmatized mafic rocks, although present locally within the EBR cover sequence, are absent in typical EBR and WBR basement. Granulite grade metamorphism of Proterozoic age has not been documented in the EBR basement, but our data demonstrate it in isolated pods within the WBR and evidence for it is widespread in the MHT portion of the CBR. While all of these attributes are common in MHT lithologies near Roan Mountain, the MHT elsewhere and our "anomalous basement" exhibit only one or a few of the above discriminating characteristics.

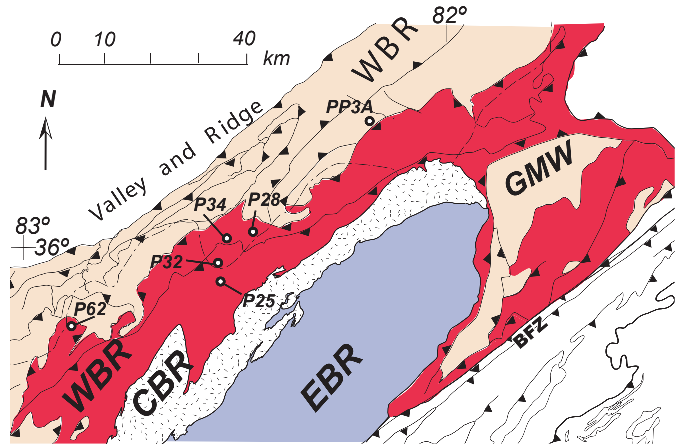

Figure 2: General location map of Western Blue Ridge thrust sheets. Open black circles denote sample locations. BFZ= Brevard Fault zone, CBR= central Blue Ridge, EBR= eastern Blue Ridge, GMW= Grandfather Mountain Window, WBR= western Blue Ridge. Map modified from Hatcher et al. (2004).

Three "anomalous basement" samples were collected for geochronology, Sm-Nd isotope geochemistry, and major- and trace-element geochemistry. These include granulitic gneiss of the CBR (MHT near Marshall, NC) and WBR (Fines Creek Granulite gneiss) and layered migmatitic amphibolite gneiss from within the EBR (Blue Ridge Parkway milepost 418) and the northernmost Dahlonega gold belt as defined by Hatcher et al. (2004) (northeast of Sylva, NC). Numerous additional samples were collected throughout the southern portion of the CBR for petrographic and geochemical analyses.

METHODS

The following methods (and numbers of analyses) specifically apply to this study. Methods of Carrigan et al. (2003) and Ownby et al. (2004) were essentially identical.

Approximately 5 kg of sample for U-Pb geochronology and 1 kg of sample for geochemical analyses were collected. Powders for elemental and isotopic geochemistry were prepared in an alumina-ceramic Shatterbox™. Fourteen samples were analyzed for elemental geochemistry by Activation Laboratories

Ltd. of Ontario, Canada, using inductively coupled plasma mass spectrometry and instrumental neutron activation analysis. An additional 34 samples were analyzed for elemental geochemistry on an Oxford Instruments MDX $^{1080+}$ multi-dispersive X-ray fluorescence spectrometer at Middle Tennessee State University. Sm-Nd isotope samples were analyzed by P. D. Fullagar at the University of North Carolina-Chapel Hill using a VG Micromass Sector 54 multicollector thermal ionization mass spectrometer, following the procedures of Fullagar et al. (1997) and Fullagar and Butler (1979).

Zircons from selected samples were separated using standard procedures. Approximately 60 zircons per sample were mounted in epoxy, polished to expose mineral interiors, and imaged by cathodoluminescence on the JEOL JSM 5600 scanning electron microscope at the Stanford-USGS Micro Analysis Center (SUMAC). Individual spots, measuring approximately 40-microns in diameter, were analyzed on the Stanford Sensitive High-resolution Ion Microprobe-Reverse Geometry (SHRIMP-RG) following the procedures of Bacon et al. (2000). Zircon standards for U-Pb age and U-concentration, R33 (419Ma) and CZ3 (550 ppm) respectively, were provided by the SUMAC. All SHRIMP data were reduced with SQUID (version 1.03, Ludwig, 2003) and plotted with Isoplot (v. 3.00, Ludwig, 2003).

GEOCHEMISTRY

Elemental Chemistry

CBR and WBR samples exhibit a similar wide range in SiO_2 concentration (47 – 82 wt % SiO_2) (Fig. 3). In contrast to both the WBR and CBR, Carrigan et al.'s (2003) EBR basement data define narrower fields and only range between 63 and 72 wt % SiO_2. The chemical diversity of the MHT and WBR data reflects the varied lithologies of each province. The scale of variability, however, is considerably different: the MHT, especially near Roan Mountain, is extremely variable on less than 1 m to 100's of m scale and contains abundant mafic rocks. The WBR is mostly granitic, fairly homogeneous over much larger tracts, and lacks mafic migmatites.

Trace element concentrations for almost all samples with >60 wt% SiO_2 plot dominantly in the volcanic arc granite field of Pearce et al. (1984) tectonic discrimination plots (Fig. 4). These plots are appropriate for evaluating the tectonic environment of relatively felsic samples only, as defined by Pearce et at. (1984), thus more mafic samples were not included. A single mafic sample of migmatitic layered amphibolitic gneiss from near Sylva, NC is characterized by a flat trace element (TE) pattern and resembles enriched-mid-ocean ridge basalt values (Sun and McDonough, 1989).

Protolith interpretation is based in large part on major element geochemistry, including the degree of alumina-saturation (see Ownby et al., 2004, for more detailed protolith interpretation criteria). All analyzed samples are metaluminous to weakly peraluminous, which we interpret to reflect igneous protoliths. Three samples are slightly more peraluminous; these samples are characterized by saussuritization of feldspars and large analytical uncertainties and were difficult to interpret. Thus, we infer an igneous protolith for these three samples.

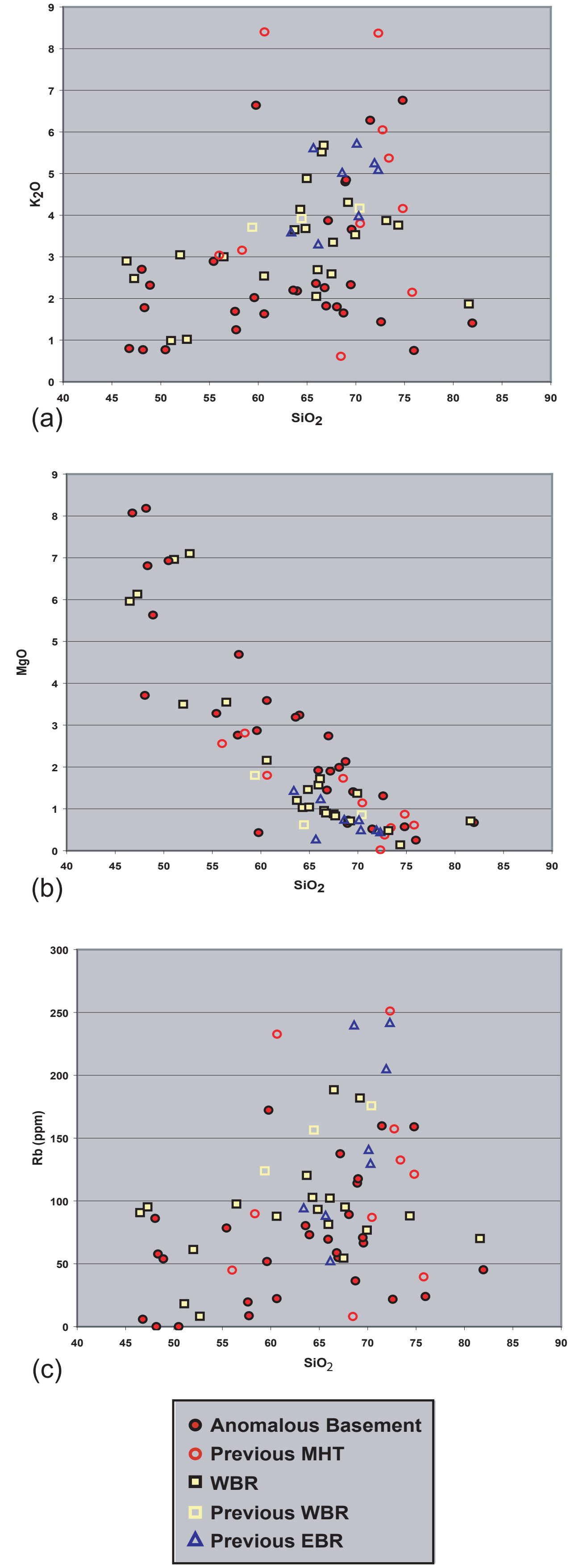

Figure 3: Harker plots of major element oxides and selected trace elements plotted versus SiO2. Filled symbols denote new samples from this study. Open symbols denote samples from previous studies; EBR and WBR from Carrigan et al. (2003); MHT from Carrigan et al. (2003) and Ownby et al. (2004). a) K_2O-SiO^2 b) MgO-SiO2 c) Rb-SiO2.

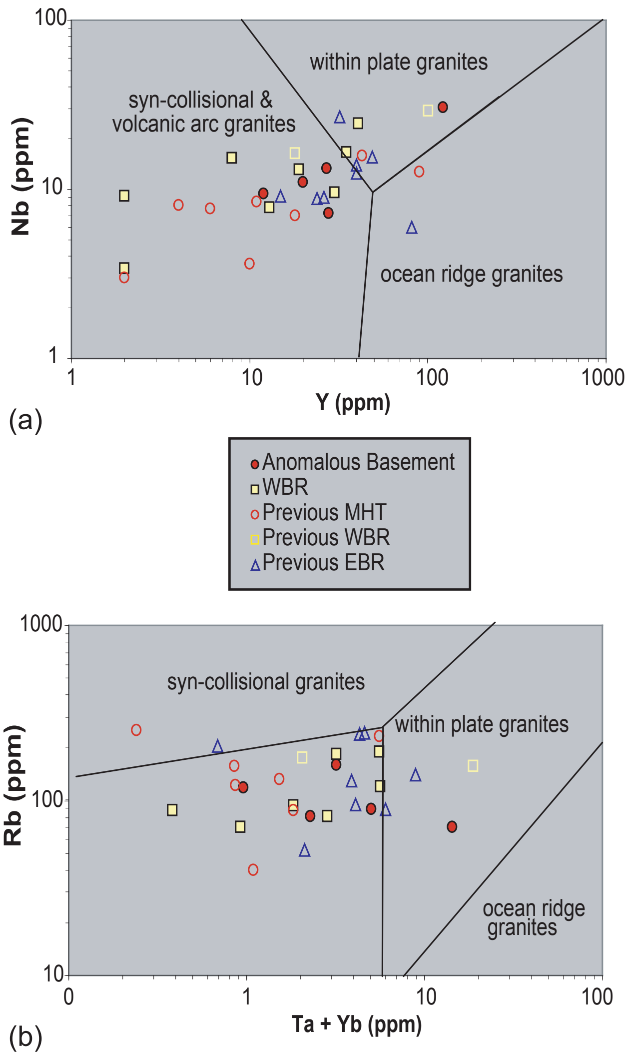

Figure 4. Tectonic discrimination diagrams, after Pearce et al (1984). (a) Nb versus Y ; (b) ; Rb versus Ta + Yb. Previous EBR and WBR data from Carrigan et al(2003) . Previous MHT data from Carrigan et al. (2003) and Ownby et al. (2004).

Isotope Chemistry

ε_{Nd} values are calculated at two ages: (1) initial values based on available magmatic ages from zircon geochronology, and (2) values at 1100 Ma to directly compare all samples near the time of the Grenville (Fig. 5). "Anomalous basement" samples exhibit ε_{Nd} values at 1100 Ma ranging from -4.1 to +1.2, almost identical to WBR samples, which range from -3.5 to 0.5 (Table 1). EBR samples have values near 0 (Carrigan et al., 2003). Depleted mantle model ages (T_{DM}, DePaolo, 1981) can be used to estimate the age at which the source for a crustal rock was extracted from a depleted mantle reservoir and may serve as a means to compare sources of crustal rocks. Anomalous basement T_{DM} ages (1.66 - 1.88 Ga, 4 samples) overlap with but range to older values than those of the EBR (1.5-1.6 Ga; Carrigan et al., 2003). WBR model ages range from 1.34 to 1.71 Ga, with a single exception at 1.82 Ga. Five of the 12 WBR samples have ages ≤ 1.50 Ga. Younger T_{DM} ages are restricted to the WBR; their possible significance is discussed in Conclusions. The two samples from the southern MHT, near Mars Hill, have T_{DM} of 1.53 and 1.65 Ga. It appears that the oldest model ages are restricted to the Roan Mountain area of the CBR (1.8-2.3 Ga for 6 of 7 samples; Carrigan et al., 2003; Ownby et al., 2004).

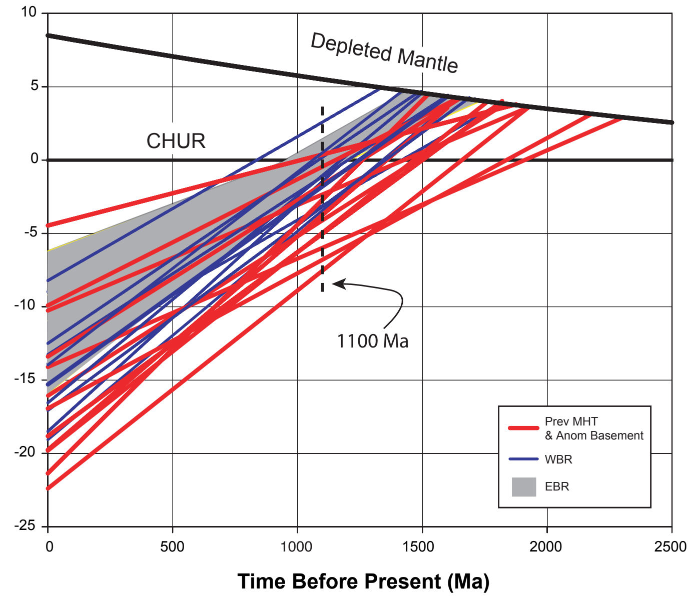

Figure 5. Evolution of εNd over time. Data for EBR and WBR from Carrigan et al. (2003) and Previous MHT from Carrigan et al. (2003) and Ownby et al. (2004). Anon Basement = anomalous basement (this study). Depleted mantle model from DePaolo (1981).

GEOCHRONOLOGY

Summary

Zircons extracted from southern Appalachian basement samples exhibit complex magmatic and metamorphic zoning that is and evident in cathodoluminescence (CL-) and back-scatter-electron (BSE-) imaging, attesting to their multi-stage history. We employ SHRIMP, versus conventional TIMS, geochronology because this technique is more appropriate for dating complex zircons by permitting discrimination of multiple, discrete ages from within single crystals and in turn minimizing mixed ages. All sample data were corrected for common Pb using a ^{204}Pb correction (assuming concordance). CL- and

Table 1. Sm-Nd isotope geochemistry.

sample	($^{147}Sm/^{144}Nd$)	ε_{Nd} at 1.1 Ga	Preferred Mag Age[1]	T_{DM} (Ga)
"Anomalous Basement"				
P22A	0.1121	-4.1	1268	1.84
P22B	0.2092	1.2	1268	(11.18)[2]
P61	0.1137	-1.7	1159	1.66
P77D	0.1625	0.4	1261	1.88
southern MHT				
MBCL-4	0.0645	-2.8	1260*	1.53
P54	0.1297	-0.5	1278	1.65
WBR				
P27A	0.1205	-3.2	~1100	1.82
P29	0.0759	-1.5	~1100	1.50
P25 C	0.0860	-3.5	1056	1.65
P 28	0.1108	-1.2	1171	1.60
P32 A	0.0984	-3.2	1375	1.69
P34	0.0869	-1.1	1130	1.50
P62A	0.1068	0.2	1019	1.49
PP3A	0.0941	0.5	1073	1.43

[1] Preferred magmatic ages (* from Ownby et al., 2004)
[2] Sm/Nd > chondritic - T_{DM} meaningless (mafic sample)

BSE-imaging were used to distinguish zones and thus guide the location of the ion-beam for individual analyses. We interpret euhedral, oscillatory, and sector zoning to reflect magmatic growth, whereas metamorphic zones are generally structureless and surround magmatic interiors (Hanchar and Miller, 1993). In some grains, multiple generations of metamorphic overgrowths are distinguished by varying brightness in CL-imaging. Most samples experienced ancient or recent Pb loss, reflected in discordance, and therefore magmatic ages were primarily determined from upper-intercept ages of best-fit discordia lines and $^{207}Pb/^{206}Pb$ weighted averages of core analyses. For some samples, magmatic ages were determined by pooling the oldest, coherent population of $^{206}Pb/^{238}U$ core ages.

Metamorphic ages were determined by pooling concordant rim analyses. In many samples, especially those with magmatic ages >1.1 Ga, discordant analyses do not define one discrete discordia line. Rather, smaller populations of magmatic core analyses may be fit to multiple discordant arrays, whose lower-intercept ages generally coincide with either metamorphic zircon growth or Pb loss events (Mesoproterozoic or early Paleozoic) more clearly recognized in other samples from this study and elsewhere throughout the basement. Therefore, we interpret this distribution of discordant analyses that cannot reliably fit to one discrete discordia line to demonstrate a multi-stage Pb loss and/or metamorphic history. Many discordant analyses fit best along a discordia line with the lower intercept forced through zero, however, representing recent Pb loss. Discordia lines through the origin yield an upper intercept age that is equivalent to averaging $^{207}Pb/^{206}Pb$ ages of discordant points.

Metamorphic rims ~ 1.0 Ga are nearly ubiquitous in samples throughout the southern Appalachian basement. These rims often have exceptionally low Pb concentrations, and therefore yield less precise SHRIMP ages (Carrigan et al., 2003). For many samples, we have omitted these analyses in determining ages due to their exceedingly large errors. We speculate that this event could be better defined by other techniques (e.g., SHRIMP or electron probe monazite analyses).

WBR samples generally fall within two ranges of magmatic ages, from 1.02-1.07 Ga and 1.13-1.17 Ga. A single sample (P32A) yields an upper intercept age of 1.38 Ga from core analyses, which we interpret as the magmatic age. Older than the Grenville, but slightly younger than most rocks of the granite-rhyolite province of the mid-continent to the west, a 1.4 Ga magmatic event of this age has, to our knowledge, not previously been documented within the southern Appalachians. Metamorphic rims from WBR samples define a complicated metamorphic history. A dominant, but poorly constrained, rim growth event evident in many samples occurred between 1.00 and 1.05 Ga. Five samples also preserve an older, ~1.1 Ga old generation of metamorphic rims. Pb loss events throughout the Grenville (~1.1 Ga to 1.0 Ga), early Paleozoic (~450 Ma), and in recent times are suggested by lower intercept ages of discordia lines fit to discordant analyses.

Two of the three samples of "anomalous basement" – those from amphibolitic migmatites - preserve magmatic ages of 1.26-1.28 Ga, as does the new sample from the MHT near Mars Hill. The remaining "anomalous basement" sample (P61), Fines Creek granulite, yields an age of 1.16 Ga. All of these samples except the amphibolitic leucosome near Sylva sample (P22A) exhibit the same imprecisely defined 1.00-1.05 Ga event. Zircons from P22A do not appear to record the 1.0 Ga event; they do have numerous early Paleozoic rims (~450 Ma).

Overall, inherited grains are rare within the WBR and "anomalous basement" samples; four samples include single inherited ages of 1.15 Ga, 1.20 Ga, 1.25 Ga, and 1.40 Ga (samples P25D, PP3A, SP, and P54, respectively).

All age uncertainties are reported to 2σ-confidence unless otherwise stated.

Refer to Table 2 for interpreted magmatic and metamorphic ages and Berquist (2005) for detailed descriptions of individual samples.

DISCUSSION AND CONCLUSIONS

Characteristics of Southern Appalachian Basement

WBR samples define magmatic events at 1.02 – 1.08 Ga, 1.13 – 1.19 Ga, and 1.38 Ga (Carrigan et al., 2003; Ownby et al., 2004; this work). EBR samples from Carrigan et al. (2003) establish a magmatic episode between 1.14 – 1.17 Ga. The CBR (including the MHT) is typified by generally older magmatic ages of ~1.2-1.3 Ga, and the oldest magmatic age documented in the southeast Appalachians of 1.8 Ga (Ownby et al., 2004, and this work). Throughout the southern Appalachian basement, zircon age analyses of metamorphic rims document a widespread metamorphic rim growth event ~ 1.0 Ga; metamorphic rim ages >1.1 Ga are documented in four WBR samples and one "anomalous basement" sample from the EBR (Carrigan et al., 2003, and this work) (Fig. 6).

New data from extensive, previously undated WBR units document their magmatic ages and complex metamorphic and Pb loss history. The Spring Creek granitoid gneiss is characterized by a magmatic age of 1.17 Ga. Max Patch Granite (sampled near Max Patch Mountain) yields a crystallization age of 1.02 Ga. One sample (sample MP) collected NE of Max Patch Mountain, from a unit thought to be correlative with the Max Patch, yields a magmatic age of 1.15 Ga, clearly demonstrating that it is older and not correlative with Max Patch Granite. Samples from the Indian Grave and Spillcorn protomylonites exhibit magmatic ages of ~1.05 Ga, supporting geochemical and field evidence suggesting that these tectonites share a similar igneous protolith. The uncertainties associated with the magmatic ages from the Spillcorn, Indian Grave, and Max Patch samples permit the possibility that the protomylonites were derived from a Max Patch protolith. These samples document the regional extent of the Max Patch and other ~1.02-1.05 Ga units.

Two "anomalous basement" samples from the EBR and CBR have magmatic zircon ages of 1.26 and 1.27 Ga, older than the dominant EBR and WBR basement and similar to MHT ages of 1.26 and 1.28 Ga for granulitic bodies near the town of Mars Hill. These "anomalous" samples, felsic gneisses associated with amphibolitic migmatites (CBR, north of Sylva, and EBR along the Blue Ridge Parkway), appear to be consistent with extension of the MHT well beyond its previously recognized extent (cf. Raymond et al., 1989). However, the full suite of MHT characteristics that clearly indicate distinctive antiquity appears to be restricted to the vicinity of Roan Mountain - neither MHT samples near Mars Hill nor the "anomalous" EBR and CBR samples contain the inherited Paleoproterozoic zircons that are common in samples from Roan Mountain, and only the

Table 2. U-Pb zircon geochronology summary, new analyses.

Sample	Location, unit	Preferred Magmatic age, ± 2σ	Method[1]	Possible Pb loss events[2]	Metamorphic age
	WBR				
P25C	Sams Gap thrust sheet	1.06 ± 0.02 Ga	int	~1.0 Ga, recent	~ 1.0 Ga
P25D	Sams Gap thrust sheet	1.07 ± 0.02 Ga	Pb-Pb	~1.0 Ga	~1.0 Ga
P32A	Fork Ridge thrust sheet	1.38 ± 0.06 Ga	int	> 1.2 Ga, ~1.0 Ga	~1.2 Ga, ~1.0 Ga
P28	Stone Mountain thrust sheet	1.17 ± 0.02 Ga	Pb-Pb	~1.0 Ga, ~early Paleozoic	~1.0 Ga
P34	Stone Mountain thrust sheet	1.13 ± 0.04 Ga	int	>1.1 Ga, ~1.0 Ga, recent	No rim analyses
PP3A	Pardee Point thrust sheet	1.07 ± 0.02 Ga	int	~early Paleozoic	No rim analyses
IG	Indian Grave protomylonite	1.05 ± 0.02 Ga	int	~1.0 Ga, recent	~1.0 Ga
SP	Spillcorn protomylonite	1.05 ± 0.01 Ga	Pb-Pb	~1.0 Ga, recent	~1.0 Ga
P62	Max Patch Granite	1.02 ± 0.03 Ga	Pb-Pb	~1.0 Ga, recent	~1.0 Ga
SC	Spring Creek granitoid gneiss	1.17 ± 0.01 Ga	Pb-Pb	> 1.1 Ga, ~1.0 Ga, early Paleozoic, recent	~1.1 Ga, ~1.0 Ga
MP	granitoid gneiss	1.14 ± 0.01 Ga	Pb-Pb	~1.0 Ga, early Paleozoic	~1.0 Ga
	southern MHT				
P54	Granulitic gneiss (MHT)	1.28 ± 0.01 Ga	conc	multiple, ~1.0 Ga	~1.0 Ga
	"Anomalous Basement"				
P22A	Leucosome of amphibolitic migmatitic gneiss (CBR)	1.27 ± 0.02 Ga	int	early Paleozoic	~450 Ma*
P77D	Leucosome of amphibolitic migmatitic gneiss (EBR)	1.26 ± 0.08 Ga	Pb-Pb	~1.0 Ga, early Paleozoic	~1.0 Ga*, ~450 Ma
P61	Granulitic gneiss (WBR)	1.16 ± 0.01 Ga	Pb-Pb	~1.0 Ga, early Paleozoic, recent	~1.0 Ga, ~450 Ma

[1]How ages were calculated: int - discordia upper intercept age; Pb-Pb - weighted mean of $^{207}Pb/^{206}Pb$ ages; conc – weighted mean of concordant analyses.
[2] Possible Pb loss events do not imply growth of zircon, rather episodes in which the U-Pb isotopic system was upset. Ages given in billions of years (Ga), except for early Paleozoic time ~ 450 Ma; recent implies Pb loss within last 50 Ma. Calculated based on discordant populations and/or discordia intercept ages of core and rim analyses.
[3] Approximate metamorphic ages reflect periods of actual zircon growth and are based on concordant populations or intercept ages of rim analyses only.
* Denotes possible age of anatectic melting, defined by coincident concordant core and rim analyses

Blue Ridge Parkway sample at 1.88 Ga has T_{DM} approaching 2 Ga.

Some samples from the WBR have T_{DM} ages of 1.3-1.5 Ga, setting them apart from the remainder of the Blue Ridge basement. This may suggest a distinct crustal heritage, perhaps linking them with the adjacent Laurentian craton to the west (see below).

Comparison to Adjacent Laurentia: Tectonic Implications

Traditional models for Laurentian growth suggest a progressive younging of crust from an Archean nucleus through Early Proterozoic mid-continent to a Middle Proterozoic margin (Fig. 7) (e.g. Karlstrom et al., 1999). This model is consistent with observed Nd-model ages >2.5 Ga in the interior Superior province, with progressive younging to the 1.4 – 1.5 Ga model ages of the mid-continent granite-rhyolite province (Van Schmus et al., 1996), and 1.3 – 1.4 Ga model ages of juvenile Grenville crust exposed in the Adirondacks and central Texas Llano uplift (Patchett and Ruiz, 1989; Marcantonio et al., 1990; Daly and McLelland, 1991; Aleinikoff et al., 1990; Ratcliff et al., 1991; Davidson, 1995; McLelland et al., 1996; Rivers, 1997; Mosher, 1989; Karlstrom et al., 1999; Bickford et al., 2000). This interpretation implies the presence of juvenile, 1.4 Ga (or younger) crustal ages in the Appalachian basement. The few WBR samples with model ages < 1.5 Ga are consistent with this trend, but the remainder of the available isotopic data for the southern Appalachians appear to refute its generality. With the exception of five WBR samples, all southern Appalachian basement T_{DM} ages are >1.5 Ga. The presence of ca. 1.8 Ga ages in the Roan Mountain area of the CBR is also at odds with

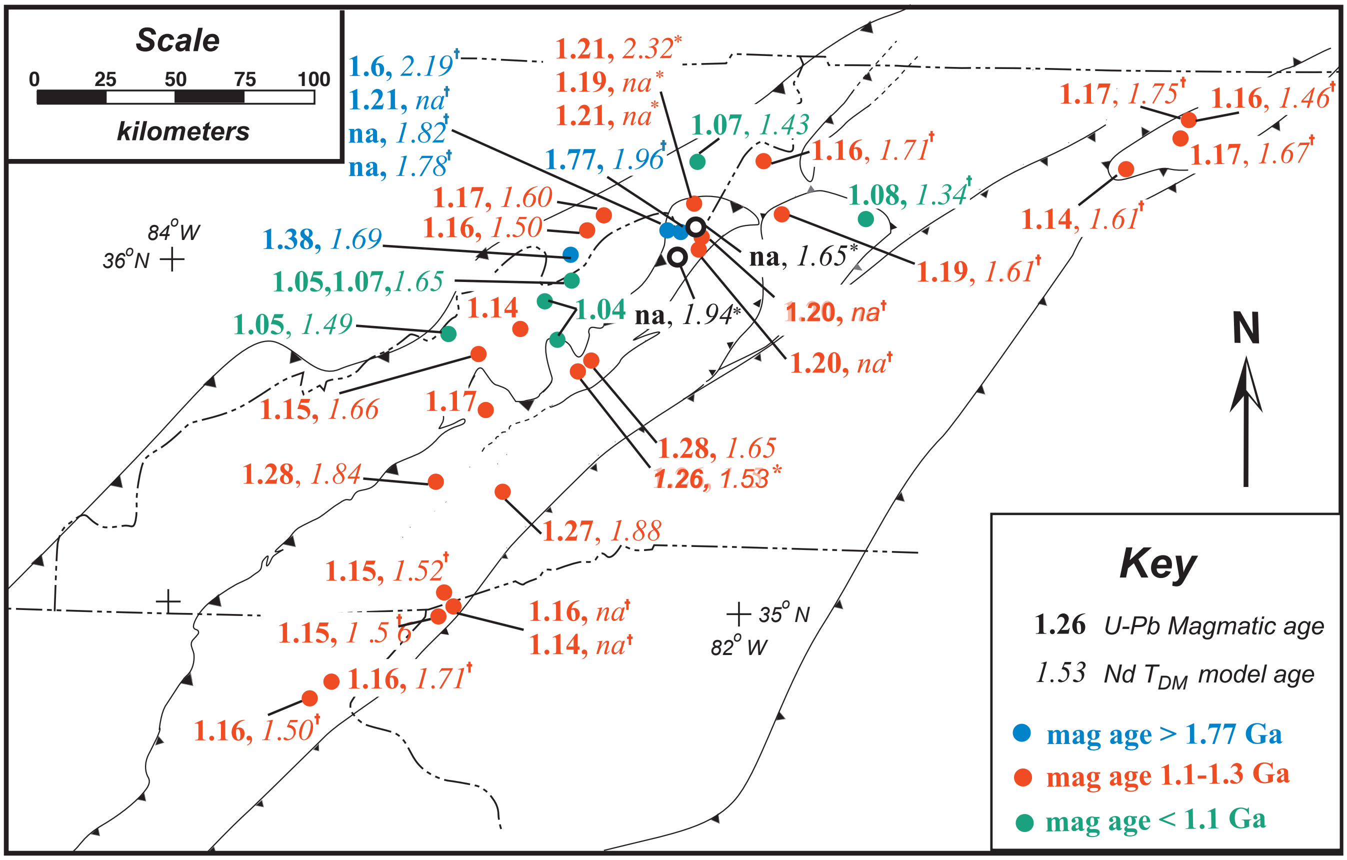

Figure 6. Distribution of U-Pb zircon magmatic ages and Nd-depleted mantle model ages from throughout the southern Appalachian basement. Samples with an asterisks analyzed by Ownby et al. (2004); samples with a cross analyzed by Carrigan et al. (2003).

progressive southeastward growth. The Blue Ridge is outboard of the ca. 1.5-1.4 Ga granite-rhyolite province. Thus, one would expect to find only Late Mesoproterzoic and younger rocks. Therefore both Nd and U/Pb data suggest that most or all of the southern Appalachian basement is exotic with respect to the adjacent Laurentian mid-continent and the North American Grenville belt.

Equally compelling evidence for an exotic origin of the southern Appalachian basement comes from regional Pb isotope studies. Whole-rock Pb isotopic data from Appalachian basement (Sinha et al., 1996, plus our new data), the Adirondacks (Sinha and McLelland, 1997), West Texas (Cameron and Ward, 1998), the Llano uplift (Smith, 1997), and subsurface data from southern Missouri (Doe, 1983) can be used to characterize the Pb signature of Proterozoic eastern North American crustal blocks (Fig. 8). Data from the juvenile Laurentian crust found in the Adirondacks, Texas, and the mid-continent define a coherent trend in $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ space that is distinct from the compositions of the southern Appalachian basement rocks. Compositions of southern and central Appalachian basement rocks define a parallel trend with a higher $^{207}Pb/^{204}Pb$ for a given $^{206}Pb/^{204}Pb$. The fact that the two trends do not overlap or cross indicates that the southern Appalachian basement rocks were derived from a different source than the juvenile Laurentian rocks. Based on this observation Sinha and McLelland (1997) concluded that the southern and central Appalachian basement may be exotic. Data from the Roan Mountain area of the MHT (Carvers Gap of Sinha et al., 1996) are distinct from the rest of the southern Appalachian basement in that they define a steeper trend, and they have an even more elevated $^{207}Pb/^{204}Pb$ signature. These elevated $^{207}Pb/^{204}Pb$ signatures in the southern Appalachian basement and Roan Mountain region of the MHT are similar to those of the Mojave (Hammond and Wooden, 1990), Wyoming (Wooden and Mueller, 1988) and Slave cratons (Davis et al., 1996) of western Laurentia and the Amazon (Tohver et al., 2004) and Kalahari cratons (Luais and Hawkesworth, 2000) of Gondwana. Incorporation of material from these regions could

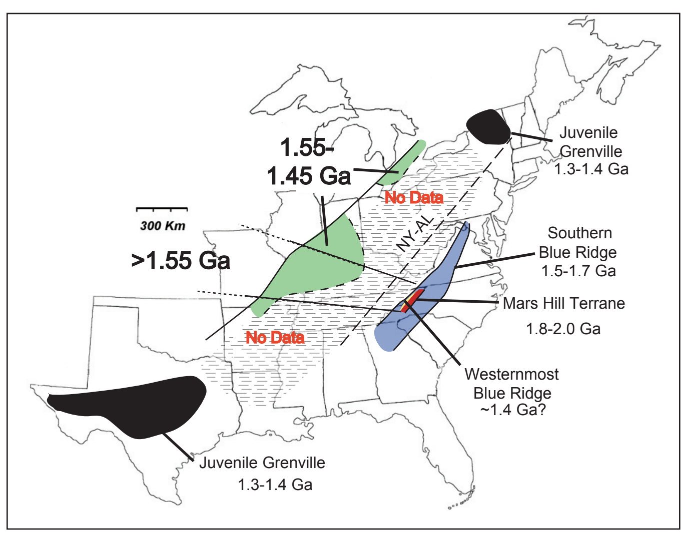

Figure 7. Revised Nd-depleted mantle model age map of eastern Laurentia based on available data. Mapand data modified from Van Schmus et al. (1996). NY-AL - New York Alabama magnetic lineament.

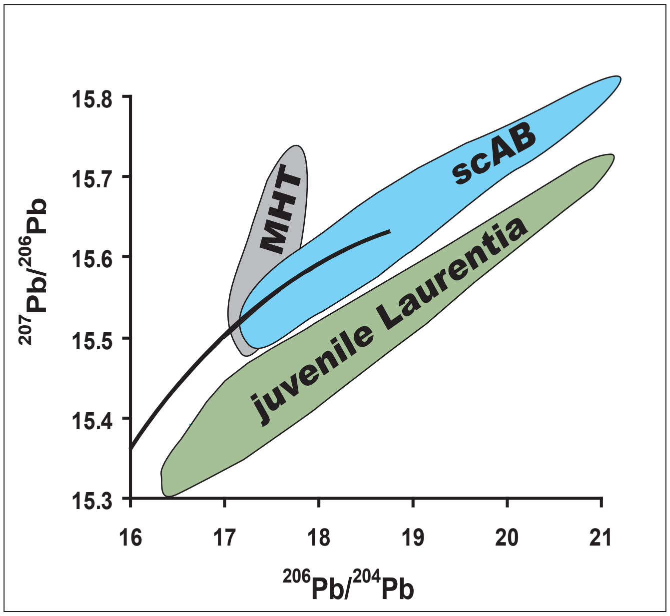

Figure 8. Whole-rock Pb isotopic data from Proterozoic basement of eastern North America. MHT= Mars Hill terrane, scAB= southern and central Appalachian Basement. Juvenile Laurentia includes Adirondacks, eastern granite-rhyolite terrane, Llano uplift and west Texas. (See text for references). The black line is the average crustal evolution curve of Stacey and Kramers (1975).

easily generate the observed signatures. Based on similarities in Pb isotopic composition Loewy et al. (2003) proposed that the southern and central Appalachian basement may have been a piece of Amazonia transferred to Laurentia after a ca. 1 Ga Amazonia-Laurentia collision. This idea was supported by Tohver et al. (2004), who took into account the similarities in both Pb and Nd isotopic signatures.

Conclusions

1. The WBR basement is characterized by three main pulses of Proterozoic magmatism at 1.02 – 1.08 Ga, 1.13 – 1.17 Ga, and 1.38 Ga.

2. Regionally extensive, previously undated units within the WBR yield magmatic ages of 1.17 Ga for the Spring Creek granitoid gneiss; 1.02 Ga for the Max Patch Granite (Nd-model age of 1.5 Ga); 1.15 Ga for a unit previously thought to be a correlative of the Max Patch Granite; and 1.05 Ga for the Spillcorn and Indian Grave protomylonites.

3. Magmatic ages and T_{DM} ages from thrust sheets of the northwestern WBR include (from structurally highest to lowest): Sams Gap (ages 1.06 and 1.07 Ga; T_{DM} ages of 1.50, 1.65, and 1.82); Fork Ridge (magmatic age of 1.38 Ga, T_{DM} of 1.69 Ga); Stone Mountain (age of 1.17 and ~ 1.13 Ga, T_{DM} of 1.50 and 1.60 Ga); and Pardee Point (age of 1.07 Ga, T_{DM} of 1.43 Ga).

4. The MHT lies within the CBR. It is most distinct near Roan Mountain, where a magmatic age of 1.8 Ga is recorded, T_{DM}s are ~2.0 Ga, and Paleoproterozoic inherited and detrital zircons are common. It is less distinct elsewhere and its southern extent is uncertain.

5. Local "anomalous basement" within the WBR, CBR, and EBR has some of the characteristics of the MHT - either granulite facies metamorphism of Mesoproterozoic age, magmatic ages of 1.2-1.3 Ga, or exposures of migmatitic amphibolite. Such rocks occur beyond the established boundaries of the MHT.

6. Nd T_{DM} ages <1.5 Ga for some samples from the WBR suggest that part of the WBR may be linked with the eastern Laurentian craton.

7. U-Pb magmatic and Nd-model ages, and Pb-isotopic signatures of southern and central Appalachian basement are, on the whole, distinct with respect to the North American Grenville basement and the adjacent Laurentian mid-continent. These data demonstrate that the southern Appalachian basement may be more similar to distant terranes in western Laurentia, South America, or Africa.

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Geochronology of the Great Smoky Mountains National Park region, TN/NC, with correlation to the rocks and orogenic events of the Appalachian Blue Ridge

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ABSTRACT

New geochronologic data in the Great Smoky Mountains National Park (GSMNP) area, western Blue Ridge provides new insight into the complex tectonic history of the region: 1) ~1195 to 1030 Ma plutonism and tectonism of the Mesoproterozoic Grenvillian orogeny; 2) uplift, cooling, erosion, extension, sedimentation, and minor volcanism from ~990 to 670 Ma; 3) metamorphism and(or) deformation events pre-~440 Ma, ~440 Ma, ~365 to ~350 Ma, and ~280 Ma; and 4) ~220 Ma of unroofing, uplift, extension, and erosion at rates of between 28 and 18 m/m.y. Ages of deformation and metamorphism are a composite of Paleozoic events: 1) pre ~440 and ~440 Ma in the Mesoproterozoic rocks beneath the Greenbrier fault; 2) ~365 Ma within the Dunn Creek thrust sheet; 3) ~350 Ma within the Greenbrier thrust sheet; 4) ~280 Ma en masse Alleghanian transport of the Miller Cove, Great Smoky, and Guess Creek thrust sheets; and 5) post ~280 Ma block faulting within the highlands. Uplift of the central highlands after the Alleghanian formed a "Great Smoky Mountains dome." As suggested by previous workers, peak Paleozoic metamorphism was Taconian with weaker overprints in the Acadian and Alleghanian orogenies.

Geochronologic data also indicate that the rocks of the Great Smoky Group were deposited between ~990 and 670 Ma, and that the Ocoee Supergroup is entirely Neoproterozoic. A metamorphic discontinuity occurs along the Greenbrier fault where Great Smoky Group rocks (muscovite cooling ages of ~350 Ma) were juxtaposed onto Mesoproterozoic rocks of the Bryson City and Ela domes (~440 Ma sphene and hornblende and biotite cooling ages of ~430 Ma). Motion on the Greenbrier fault in this area occurred between ~350 Ma and ~340 Ma (muscovite cooling ages of mylonite).

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INTRODUCTION

The geology of the GSMNP region of Tennessee and North Carolina (Keith, 1895; 1907; King et al., 1968; Hadley and Nelson, 1971) was further studied from 1993 to 2003 as part of a cooperative investigation of the FEDMAP Program of the U.S. Geological Survey (USGS) with the National Park Service (NPS) (Southworth et al. 2005). In addition to our work in the Park, we investigated the geochronology of: 1) Mesoproterozoic rocks; 2) Neoproterozoic Ocoee Supergroup (Aleinikoff et al., 2004); 3) deformation, metamorphism, uplift, and denudation (Naeser et al., 2004); and 4) erosion rates using cosmogenic ^{10}Be (Matmon et al., 2003a, 2003b).

Modern geochronology defines four orogenic events that affected rocks of the southern Appalachians (summarized by Rodgers, 1970; Butler, 1991; Hatcher and Goldberg, 1991; Thomas et al., 2004; Hatcher et al., 2005): 1) the Grenvillian orogeny (~1200 to 1030 Ma plutons); 2) the Taconian orogeny (490 to 440 Ma plutons, 480 to 460 Ma metamorphism, and 454 to 453 Ma K-bentonite beds in Middle Ordovician strata); 3) the Acadian orogeny (420 to 350 Ma plutons and 400 to 350 Ma metamorphism); and 4) the Alleghanian orogeny (330 to 270 Ma plutons, and 316 Ma and 311 Ma Pennsylvanian volcanic ashes. Hadley (1964) first related isotopic age data to crustal heating and sedimentation in the Appalachian basin. The timing of thermal peaks that he noted at 440 Ma, 360 to 340 Ma, and 260 Ma, using Rb-Sr and K-Ar techniques, correspond remarkably well with modern isotopic data. The lack of correlation between the time of foreland sedimentation and the hinterland metamorphic cooling ages (Hadley, 1964), as well as the ages and provenance of detrital zircons (Thomas et al., 2004), suggest involvement of complex processes during construction and destruction of mountain ranges. The break-back sequence of orogenesis and metamorphism (Hatcher and Odom, 1980; Higgins et al., 1988; Butler, 1991; Thomas et al., 2004) is further complicated by break-forward (piggy back) sequence of metamorphism and related late Paleozoic thrusting (Trupe et al., 2004), as well as out-of-sequence faulting. Uncertainty remains in the correlation of metamorphic and deformation events with the three classic Paleozoic Appalachian orogenies (Abbott and Raymond, 1984), the ages and stratigraphic correlations among the clastic sediments in the Appalachian basin (Hadley, 1964), and the faulting sequence.

This paper is a progress report on investigations of the geochronology of the rocks and orogenic events of the GSMNP area, and how they correlate to rocks elsewhere in the Appalachian Blue Ridge. The ages of the periods cited are from the International Geologic Time Scale (Gradstein et al., 2004).

MESOPROTEROZOIC ROCKS

Mesoproterozoic rocks in the GSMNP area consist of a polymetamorphic complex of paragneiss, migmatite, and orthogneiss that were locally partially melted and mylonitized during the Mesoproterozoic and Paleozoic (Cameron, 1951; Hadley and Goldsmith, 1963; Kish et al., 1975). Paleozoic deformation, uplift, and erosion have exposed these rocks in six antiforms and along several late Paleozoic thrust faults (Fig. 1). The rocks have been subdivided based on petrology, foliation, and Sensitive High Resolution Ion Microprobe (SHRIMP) U-Pb zircon geochronology (Southworth et al., 2005). Paragneiss lithologies include amphibolite, ultramafic rocks, hornblende-biotite gneiss, migmatitic biotite gneiss, and isolated pods of calc-silicate granofels. Orthogneiss lithologies include leucocratic metagranite, granitic gneiss, granodiorite, biotite augen gneiss, and porphyritic granitic gneiss.

The Mesoproterozoic rocks of the GSMNP area can be correlated with rocks recognized elsewhere in the North Carolina Blue Ridge (Quinn and Wright, 1993; Carrigan et al., 2003; Ownby et al., 2004) and in the Virginia Blue Ridge (Aleinikoff et al., 2000; Tollo et al., 2004a, 2004b) that comprise three groups, as defined by Aleinikoff et al., (2000) and Tollo et al. (2004a, 2004b). The oldest Mesoproterozoic rocks are migmatitic paragneisses that are intruded by orthogneisses of Group 1 age. Group 1 rocks contain both the oldest dated granitoids (1194-1192 Ma) and a younger suite that crystallized between 1168 and 1163 Ma. The younger suite contains inherited zircons (1190 and 1180 Ma) from the earlier suite of Group 1 rocks. Group 2 rocks crystallized at ~1117 Ma. Group 3 orthogneisses crystallized between ~1056 and 1029 Ma. A major deformation event occurred between the crystallization of Group 2 and Group 3 rocks, or between ~1117 and 1081 Ma. The lithologic and age similarities between the migmatitic gneiss in the Ela dome (Fig. 1) with the Cranberry-Mine Layered Gneiss of the Elk River massif near Grandfather Mountain window (Carrigan et al., 2003), as well as felsic gneiss of the Mars Hill terrane above the Fries fault also near Grandfather Mountain window (Ownby et al., 2004), suggest that rocks in the western Blue Ridge are similar to some of those in the Mars Hill terrane.

NEOPROTEROZOIC OCOEE SUPERGROUP

King et al., (1958) subdivided the Ocoee Supergroup rocks into the Snowbird, Great Smoky, and Walden Creek Groups. Stratigraphic relations between the three Groups in the GSMNP are not clear (Rast and Kohles, 1986; Costello and Hatcher, 1991; Walker and Rast, 1991). Great Smoky Group rocks may be conformable above the Snowbird Group (Montes and Hatcher, 1999); Snowbird and Walden Creek Groups rocks may transition into one another across the Dunn Creek fault (Hamilton, 1961); Walden Creek Group rocks may be conformable above the Snowbird Group (Oriel, 1950); and Walden Creek Group rocks are unquestionably conformable above Great Smoky Group (Costello and Hatcher, 1991; Thigpen and Hatcher, 2004). Since the three groups are in fault contact in the GSMNP region, geochronologic techniques offer the best means to determine relative ages of deposition and diagenesis (Aleinikoff et al., 2004).

The Snowbird Group consists of, in ascending order, the Wading Branch Formation, Longarm Quartzite, Roaring Fork Sandstone, Pigeon Siltstone, Metcalf Phyllite, and Rich Butt Sandstone (King et al., 1958). These rocks are lithologically similar, intergradational, current-bedded metaclastic rocks. Metcalf Phyllite (Zm) found west of the Pigeon Forge fault is tectonized Pigeon Siltstone (King et al., 1968). The entire belt of Metacalf Phyllite is fault- bounded, has pervasive shear-band cleavage, and locally has crenulation cleavage.

The Great Smoky Group consists of, in ascending order, the Elkmont Sandstone, Thunderhead Sandstone, Anakeesta Formation (King et al., 1958), Copperhill Formation, Wehutty

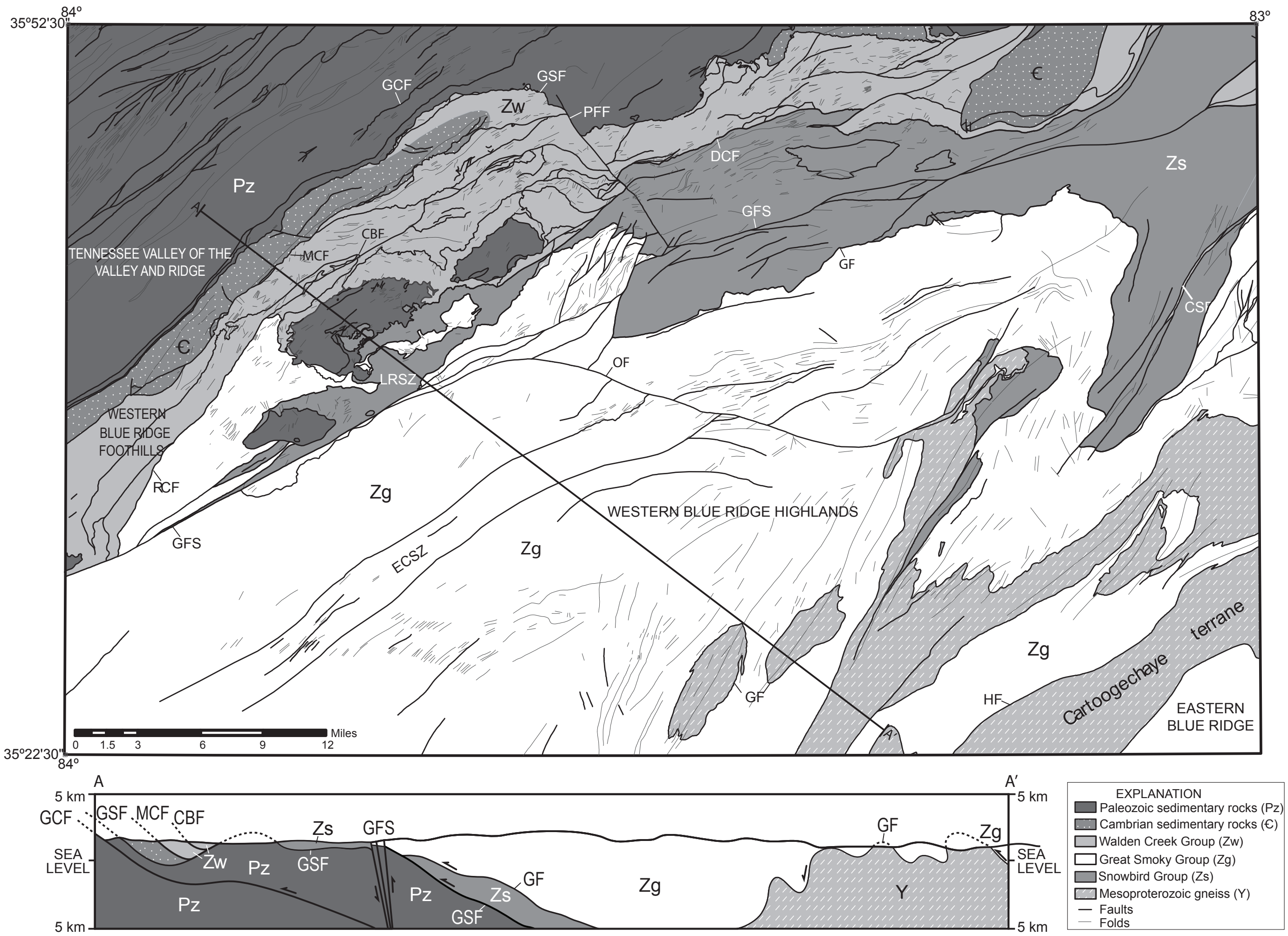

Fig. 1. Generalized geologic map and cross section A-A' of the GSMNP region, Tenn./N.C., showing major faults, fold axes, and isograd metamorphic zone markers (modified after Southworth et al., 2005). Capshaw Branch fault (CBF); Cold Spring fault (CSF); Dunn Creek fault (DCF); Eagle Creek shear zone (ECSZ); Guess Creek fault (GCF); Greenbrier fault (GF); Gatlinburg fault system (GFS), Great Smoky fault (GSF), Hayesville fault (HF); Little River shear zone (LRSZ); Miller Cove fault (MCF); Oconaluftee fault (OF); Pigeon Forge fault (PFF); Rabbit Creek fault (RCF).

Formation, Grassy Branch Member of the Ammons Formation, Dean Formation, and Nantahala Formation (Wiener and Mershat, 1992). The formational subdivisions are problematic and somewhat arbitrary because of complex interbedding of coarse- and fine-grained rocks and the structural and metamorphic history. Rocks of the Great Smoky Group in North Carolina are stratigraphically above rocks in Tennessee (Wiener and Merschat, 1992), and are lithologically similar although they have different names.

The Walden Creek Group consists of, in ascending order, Licklog, Shields, Wilhite, and Sandsuck Formations (King et al., 1958). Walden Creek Formation rocks are stratigraphically overlain by Chilhowee Group rocks along the length of Chilhowee Mountain (King, 1964; Thigpen and Hatcher, 2004).

PALEOZOIC ROCKS

Paleozoic rocks crop out in the Foothills along the Great Smoky fault, and within the Calderwood, Cades Cove, Tuckaleechee Cove, and Wear Cove windows. Lower Cambrian rocks of the Chilhowee Group, Shady Dolomite, and Rome Formation are in the hangingwall of the Great Smoky fault; Ordovician Jonesboro Limestone, Lenoir Limestone, and Blockhouse Shale are exposed in the Calderwood, Cades Cove, Big Spring Cove, Whiteoak Sink, Tuckaleechee Cove, and Wear Cove windows, and in the footwall of the Great Smoky fault.

NEOPROTEROZOIC AND PALEOZOIC IGNEOUS ROCKS

Metadiabase, metadiorite, pegmatite, and trondhjemite intrude rocks of the Snowbird Group and Great Smoky Group (Laney, 1907; Espenshade, 1963; Hadley and Goldsmith, 1963; King, 1964; Southworth, 1995). Chemical analyses and some modal data of metadiabase and metadiorite indicate they are subalkalic basalt or diorite (Southworth, 1995; Tull et al., 1998). There are multiple dikes and sills, and most are altered to greenstone and carbonate-chlorite schist. In one locality, fine-grained tuff—crystal tuff containing plagioclase phenocrysts, lapilli tuff containing angular fragments of crystal tuff, and tuff breccia containing sub-rounded blocks of crystal tuff as much as 10 in (25 cm) long—occur near massive greenstone. These rocks may be correlative with the Ducktown assemblage of

Higgins et al., (1988), which consists of amphibolite dikes and sills, felsic metatuffs, coarse tuff breccias, and coarse, poorly sorted volcanic-epiclastic conglomerates within the Copperhill Formation in northern Georgia (Higgins et al., 1988). Although volumetrically small, these dikes and sills may be associated with pre-Iapetus rifting and formation of the Great Smoky Group basin.

Tabular dikes and sills of pegmatite and trondhjemite intruded the Mesoproterozoic rocks and the staurolite- to kyanite- grade rocks of the Snowbird and Great Smoky Groups. Pegmatite is concentrated along the western margin of the Bryson City window (Cameron, 1951). Most pegmatites are not foliated, and some intruded mylonitic rocks of the Greenbrier fault. However, some pegmatites are pre- to syn-kinematic, because they exhibit foliation paralle to that in the country rocks and are folded. Trondhjemite dikes mostly occur east and southeast of Dellwood and the Qualla-Dellwood belt, and they post-date the foliated rocks (R.D. Hatcher, Jr., 2005, written comm.).

AGE OF THE OCOEE SUPERGROUP

Great Smoky Group

Overgrowths of xenotime (YPO_4) and monazite ($CePO_4$) in Thunderhead Sandstone-Cades Sandstone were dated using SHRIMP U-Pb microanalysis (Aleinikoff et al., 2004). Thin (<20 µm), delicate, irregular overgrowths of xenotime, formed epitaxially on isostructural detrital zircon, and monazite overgrowing detrital monazite are interpreted as diagenetic and/or metamorphic. Both minerals were dated *in situ* from polished thin sections. Because overgrowths can form during multiple events, the oldest overgrowths represent the best estimate of the minimum age of deposition of the sediment. Conversely, the youngest detrital zircon and monazite ages provide the best estimates of the maximum age of deposition. Chlorite-grade metasandstone along the Little River, opposite Meigs Falls, TN., yielded a monazite overgrowth age of 670 ± 21 Ma (9 analyses on 5 grains), and a xenotime overgrowth age of 560 ± 16 Ma (11 analyses on 7 grains), plus one additional analysis that is about 780 Ma. Detrital zircon and monazite range from 1200-990 Ma and 1080-1030 Ma, respectively, which correspond well to the dated Mesoproterozoic rocks in the area. These data indicate that the Thunderhead Sandstone-Cades Sandstone was derived from Mesoproterozoic source rocks and was deposited sometime between ~990 and 670 Ma. However, uplift, erosion, and deposition after zircon crystallization and hornblende cooling, probably requires that the maximum age is less than 900 Ma. The monazite age either records the time of diagenesis during shallow burial, or the ages of both monazite and xenotime overgrowths record subsequent thermal events. Interestingly, the xenotime overgrowths formed during the Catoctin Formation volcanism interval in central Virginia to southern Pennsylvania (575 to 560 Ma) (Aleinikoff et al., 1995).

Rocks of similar age in the Blue Ridge province include the 810 Ma metavolcanic rocks of the Grandfather Mountain Formation (Rankin et al., 1969) (U/Pb Zr technique), the 734 ± 26 Ma Bakersville metadiabase dike swarm (Goldberg et al., 1986) (whole rock Rb-Sr technique), the 760 to 740 Ma Crossnore Complex (Su et al., 1994) (U/Pb Zr technique), the 758 ± 12 Ma metarhyolite of the Mount Rogers Formation (Aleinikoff et al., 1995) (U/Pb Zr technique), and the 730 to 700 Ma Robertson River Igneous Suite (Tollo and Aleinikoff, 1996) (U/Pb Zr technique). In addition, there are 707 Ma metavolcanic rocks beneath the Neoproterozoic Fauquier Group (U/Pb Zr technique), 705-702 Ma volcanic and volcaniclastic rocks in northern Virginia (Tollo and Hutson, 1996) (U/Pb Zr technique), which may be intercalated with rocks of the Mechum River Formation (Bailey and Peters, 1998), and the 607 ± 7 Ma Reusens migmatite of the Moneta Gneiss in the lower part of the Lynchburg Group in central Virginia (Wang and Glover, 1991) (U/Pb Zr technique).

Walden Creek Group

Unrug and Unrug (1990) reported that shale enclosing limestone olistoliths contains microcrinoids (post-Ordovician?), fenestrate bryozoans (Ordovician and younger?), ostracodes, trilobites (post-Ordovician), and agglutinated foraminifera. These fossils suggest a Cambrian to Ordovician or younger age. Bryozoans, foraminifera, ostracods, pelmatozoans, and trilobites in seven different localities of rocks of the Walden Creek Group were interpreted as Silurian or younger (Unrug et al., 2000). Although Unrug et al., (2000) reported the shale to be of the Wilhite Formation, it is part of the Sandsuck Formation (R.D. Hatcher, Jr., 2005, written comm.). We sampled, dissolved, and analyzed limestone at five localities of Unrug et al., (2000) and they are devoid of fossils (J. Repetski, U.S. Geological Survey, 2003, oral comm.). Additional samples analyzed from limestone southwest of the Park may contain fragments of conodonts, trilobites, and ostracodes (J. Repetski, U.S. Geological Survey, 2005, oral comm.).

Carbon and oxygen isotope compositions and ratios measured from Wilhite Formation limestone along Chilhowee Lake, TN., have $\delta^{13}C_{carb}$ values that are lower than typical Silurian carbonates but are very characteristic of carbonates deposited during several intervals of the Neoproterozoic (Herbert, 2003 and written comm., 2004).

C-shaped soft-bodied metazoan fossils in the Sandsuck Formation (upper part of the Walden Creek Group) suggest an Early Cambrian age at the youngest but are probably Neoproterozoic (Broadhead et al., 1991). Acritarchs recovered from shale of the Wilhite Formation and conglomerate of the Shields Formation near Kinzel Springs, shale of the Sandsuck Formation northeast of Chilhowee Mountain, and limestone of the upper part of the Yellow Breeches Member of the Wilhite Formation near English Mountain also suggest a Neoproterozoic age (Knoll and Keller, 1979). Some members of the acritarch assemblage [*Bavlinella faveolata* (Shepleva), *Sphaerocongregus* (=*Bavlinella*) and leiosphaerid] range into the Cambrian (Knoll and Swett, 1985). The acritarch assemblage is a facies-specific biota widely associated with fine-grained basinal sediments of Neoproterozoic age, but ranges into the Phanerozoic. Until proven otherwise, the Walden Creek Group has a Neoproterozoic age.

Chilhowee Group

The Neoproterozoic-Cambrian boundary (~545 Ma) (Isachsen et al., 1994) probably occurs within the middle or upper part of the basal Cochran Formation (Walker and Driese, 1991). The underlying Sandsuck, Wilhite, and Shields Formations, Walden Creek Group, contain Neoproterozoic

(Vendian) acritarchs (Knoll and Keller, 1979). A SHRIMP U-Pb Zr age of ~560 Ma from metarhyolite in the Neoproterozoic Catoctin Formation on South Mountain, PA, is approximately 1 m beneath the Weverton Formation (Cochran equivalent).

Trace fossils in the Chilhowee Group include *Skolithus* and *Planolites* in the Nichols Shale, *Skolithus, Rusophycus*, and *Cruziana* in the Nebo Quartzite, *Planolites*, *Rusophycus*, and *Cruziana* in the Murray Shale, and *Skolithus* in the Hesse Quartzite (Walker and Driese, 1991). The Murray Shale contains lingulellid brachiopods and trilobites (Keith, 1895) and the archaeocopid ostracode *Indiana tennesseenis* (Laurence and Palmer, 1963). The Helenmode Formation contains lingulellid brachiopods, the trilobites *Olenellus*, *Isoxys chilhoweana*, the ostracode *Indiana tennesseenis* and *Hyolithes* (Walcott, 1891; Resser, 1938; Neuman and Nelson, 1965). All of these indicate a Lower (?) Cambrian age for the Chilhowee Group.

STRUCTURE OF THE GSMNP REGION

The deformation and metamorphism of the rocks of the GSMNP region occurred tens of kilometers deep within the crust and hundreds of kilometers to the southeast of their present position (Hatcher, 1978). Several phases of faulting, folding, and cleavage development occurred throughout the Paleozoic, creating a complex structural framework (Fig. 1). There are four major fault systems in the study area: 1) the Greenbrier fault, Dunn Creek fault, and related faults and folds; 2) the Great Smoky fault, Miller Cove fault, and related faults and folds in the foothills; 3) the Gatlinburg fault system and Pigeon Forge fault; and 4) the thrust sheets of the Tennessee Valley and Cumberland Plateau bounded by the Sequatchie Valley and Pine Mountain thrust faults (floor) and the Great Smoky fault (roof).

Premetamorphic faults (Hamilton, 1961; Hadley and Goldsmith, 1963; Milton, 1983) include the Greenbrier fault (Hadley and Goldsmith, 1963), the Rabbit Creek fault (Neuman and Nelson, 1965), and the Dunn Creek fault (Hamilton, 1961). Pre-metamorphic east-northeast-trending folds (F1) in the region are the Copeland Creek anticline, the Alum Cave synclinorium, and the Murphy synclinorium.

Postmetamorphic faults include the Eagle Creek shear zone (Southworth, 1995), Mingus fault (King, 1964), Little River shear zone (Southworth et al., 2005), Capshaw Branch klippe (Neuman and Nelson, 1965), Cold Spring fault (Hadley and Goldsmith, 1963), Caldwell Fork fault (Hadley and Goldsmith, 1963), Great Smoky fault (Keith, 1895, 1927; Gordon, 1920; Wilson, 1935), Miller Cove fault (Neuman and Nelson, 1965), and Walden Creek fault (King, 1964). The latest faults are the Pigeon Forge fault (King, 1964), Bogle Spring fault (King, 1964), Oconaluftee fault (King, 1964), and the Gatlinburg fault system (Hadley and Goldsmith, 1963; King, 1964).

The F_1 folds in the eastern half of the highlands have been overprinted by the north-northeast-trending F_2 folds and faults. The domes and anticlinoria that expose Mesoproterozoic rocks are F_2 folds. A more northerly trending cleavage (S_3) transected these structures throughout the Park and post-dates the F_1/F_2 folding events (Hadley and Goldsmith, 1963; Connelly and Woodward, 1992). Early tectonic transport was northward and later transport, probably in two different phases, was north-northwestward and later northwestward.

METAMORPHISM AND POSTMETAMORPHIC COOLING OF THE GSMNP REGION

Mesoproterozoic

Some of the Mesoproterozoic rocks in the GSMNP area contain the mineral assemblage hornblende-orthopyroxene-microcline in a textural equilibrium that indicates granulite-facies metamorphism. U-Pb geochronology suggests that the migmatitic gneiss formed during a partial melting event (~1195 Ma) that preceded the intrusion of granitoids between ~1170 to 1030 Ma (Southworth et al., 2005). A thermal event at ~1045 Ma roughly coincided with the ~1040 Ma crystallization of granodiorite exposed on Cove Mountain. The biotite augen gneiss (~1030 Ma) crystallized during the Ottawan phase peak Grenvillian metamorphism, at about 1028 ± 9 Ma (Ownby et al., 2004).

Paleozoic

The Neoproterozoic rocks in the western Blue Ridge were metamorphosed from sub-chlorite grade in the far northwestern Foothills, to kyanite grade in the highlands near Cherokee (Hadley and Goldsmith, 1963; Connelly and Dallmeyer, 1993). Locations of the metamorphic zone markers that approximate isograds, as shown by Hamilton (1961), Hadley and Goldsmith (1963), King (1964), Neuman and Nelson (1965), Carpenter (1970), Lesure et al., (1977), and Mohr and Newton (1983), were generalized based on exposure, bulk rock composition, samples collected, and thin sections studied, and they apparently are not offset by faults (Fig. 2).

Metamorphic mineral assemblages in the respective zones are as follows. Sub-chlorite zone rocks have very weak pressure solution cleavage (R.D. Hatcher, Jr., 2005, written comm.) but no visible metamorphic minerals. Chlorite zone rocks have chlorite porphyroblasts and minute crystals of sericite. Biotite zone rocks seldom contain visible porphyroblasts of biotite. Garnet zone rocks contain visible porphyroblasts of garnet, biotite, muscovite, and chloritoid. Rocks in the staurolite and kyanite zones contain albite, garnet, biotite, muscovite, staurolite, and kyanite, visible in hand sample. Some porphyroblasts are confined to the bedding (compositional layering), and peak thermal conditions produced randomly oriented porphyroblasts that grew statically. Kyanite zone rocks were locally partially melted during peak metamorphic conditions, producing migmatite and pegmatite segregations.

Three staurolite-kyanite grade schists in the Great Smoky Group in the Murphy synclinorium, and schist along the Blue Ridge Parkway near Soco Gap, yield P-T conditions that range from 700 to 580^{o} C and 8 to 6.6 kbar (Nesbitt and Essene, 1982; Mohr and Newton, 1983; Kohn and Malloy, 2004; Anderson et al., 2005), at depths of 27 to 23 km (Dallmeyer, 1975).

Abundant evidence of retrograde metamorphism occurs in the southern and southeastern part of the highlands (Mohr, 1973; Southworth, 1995; Montes and Hatcher, 1999). Axial planar pressure solution cleavage cuts across chlorite and sericite in retrograde garnet and biotite porphyroblasts.

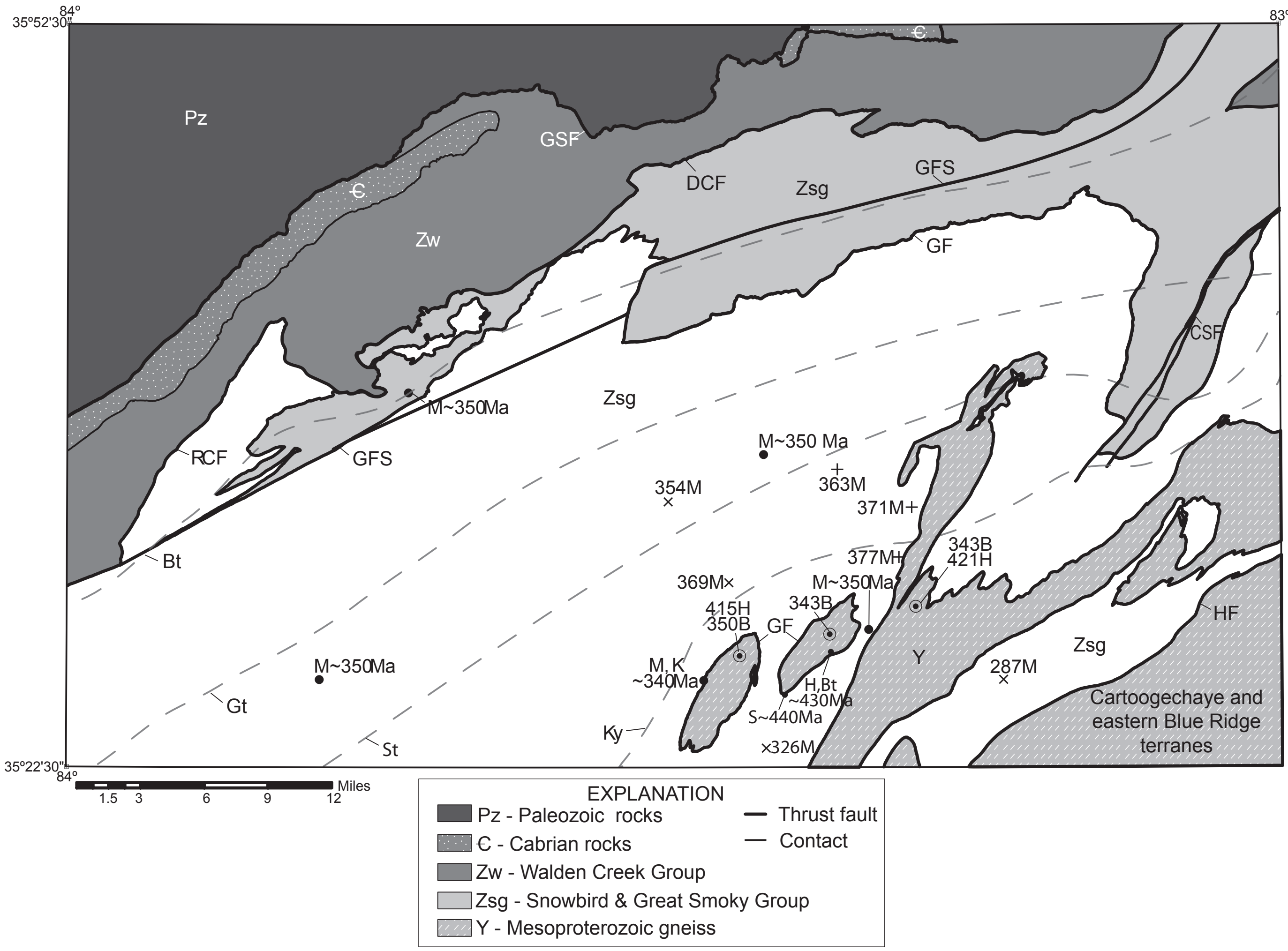

Fig. 2. Generalized geologic map showing isograds, sample locations, and mineral ages. Biotite (B), hornblende (H), muscovite (M), and sphene (S) ages from Dallmeyer (1975) (circled dot), Kish (1991) (x), Connelly and Dallmeyer (1993) (+), and this study (dot). The monazite samples of Kohn and Malloy (2004) are near the samples of Connelly and Dallmeyer (1993). Cold Spring fault (CSF); Dunn Creek fault (DCF); Greenbrier fault (GF); Gatlinburg fault system (GFS); Great Smoky fault (GSF), Hayesville fault (HF); Rabbit Creek fault (RCF). Bt, biotite; Gt, garnet; St, staurolite; Ky, kyanite.

Previously Published Isotopic Data

K-Ar and $^{40}Ar/^{39}Ar$ ages of biotite, hornblende, muscovite, and whole-rock specimens in this area suggest a complex metamorphic history (Kish, 1991; Fig. 2). $^{40}Ar/^{39}Ar$ hornblende cooling ages for Mesoproterozoic rocks are 421 ± 15 Ma in the Qualla-Dellwood belt, and 415 ± 15 Ma in the Bryson City dome (Dallmeyer, 1975). $^{40}Ar/^{39}Ar$ biotite incremental gas-release ages from Mesoproterozoic rocks are 343 ± 15 Ma in the Ela dome and Qualla-Dellwood anticlinorium, and 350 ± 15 Ma in the Bryson City dome (Dallmeyer, 1975). $^{40}Ar/^{39}Ar$ analysis of rocks within the Murphy belt to the southwest of our study area suggests 440 to 425 Ma cooling of hornblende with a 350 to 325 Ma thermal overprint with 330 Ma muscovite (Dallmeyer, 1988). Across the Murphy synclinorium near Bryson City, K-Ar and $^{40}Ar/^{39}Ar$ analyses of muscovite from Great Smoky Group rocks yield ages of 369, 354, 347, and 326 Ma, and 287 Ma from rocks just west of the Hayesville fault (Kish, 1991). Three samples of schist of the Copperhill Formation north of Cherokee have $^{40}Ar/^{39}Ar$ muscovite growth ages that range from 377 to 362 Ma (Connelly and Dallmeyer, 1993).

U-Pb TIMS analyses of monazite and U-Th-Pb monazite chemical ages suggest that metamorphism occurred between 480 to 440 Ma (Moecher et al., 2003; 2004) near Soco Gap, NC, and between 437 Ma and 357 Ma (~400 Ma interpreted age using electron microprobe and ion microprobe techniques) (Kohn and Malloy, 2004) near Bryson City, NC. A pegmatite from the Cox Number 1 mine near the "border gneiss" mylonite of the Bryson City window had a Rb-Sr whole rock age of 435±28 Ma (2σ) (Kish et al., 1975; Kish, 1991).

New Isotopic Data

A U-Pb sphene age of ~440 Ma, an $^{40}Ar/^{39}Ar$ biotite cooling age of ~430 Ma, and an $^{40}Ar/^{39}Ar$ hornblende cooling age of ~430 Ma were obtained from Mesoproterozoic rocks in the Ela dome (Fig. 2). An $^{40}Ar/^{39}Ar$ muscovite cooling age of ~350 Ma was obtained from schist of the Copperhill Formation that surrounds the Bryson City dome. An $^{40}Ar/^{39}Ar$ muscovite cooling age of ~340 Ma and a potassium feldspar age of ~340 Ma were obtained from mylonite of the Greenbrier fault of the Bryson City dome. Pegmatite that intrudes rocks of the Copperhill Formation just east of the Ela dome has a $^{40}Ar/^{39}Ar$ muscovite plateau age of ~350 Ma. $^{40}Ar/^{39}Ar$ muscovite cooling ages of ~350 Ma were obtained from rocks of the Eagle Creek shear zone and Little River shear zone, 15 km and 35 km to the northwest of the Bryson City dome, respectively. In addition, preliminary $^{40}Ar/^{39}Ar$ muscovite cooling ages of mylonitic Mesoproterozoic rocks in the Asheville 30 by 60-minute quadrangle range from 355 Ma to 336 Ma (Cattanach and Merschat, this guidebook).

Zircon fission-track ages in the central highlands are statistically indistinguishable at about 280 Ma over >1.25 km vertical relief. The zircon ages in surrounding rocks are significantly older. Apatite fission-track ages southeast of the Gatlinburg fault system generally decrease with decreasing elevation, from ~185 Ma near the TN/NC line from Clingmans Dome to Newfound Gap to ~ 85 Ma at the base (Naeser et al., this guidebook).

DISCUSSION

Chlorite- to kyanite-grade rocks interpreted as a typical Barrovian metamorphic sequence (Hadley and Goldsmith, 1963) now appear to be a composite of several metamorphic events (Connelly and Dallmeyer, 1993), in part, telescoped along thrust faults (Fig. 2). Chlorite-grade rocks are restricted to the Miller Cove and Dunn Creek thrust sheets in the Foothills. The biotite isograd is along the Gatlinburg fault system that separates the Foothills and Highlands. Our preliminary U-Pb and $^{40}Ar/^{39}Ar$ isotopic data from Mesoproterozoic rocks in the Ela dome support early Silurian (~440 Ma) metamorphism (sphene growth), middle Silurian (~430 Ma) cooling of hornblende from an early Silurian or Ordovician (?) metamorphism, and ~430 Ma cooling of biotite due to rapid uplift. $^{40}Ar/^{39}Ar$ cooling ages of ~340 Ma for muscovite and potassium feldspar from mylonite of the Greenbrier fault framing the Bryson City window, and muscovite from pegmatite that cut the bounding faults of the Bryson City and Ela domes, suggest that the Greenbrier fault was emplaced between ~350 and ~340 Ma (Fig. 3).

The staurolite- to kyanite-grade cover rocks near Bryson City and Cherokee contain 437 to 357 Ma monazite (Kohn and Malloy, 2004) and muscovite grew between 377 and 347 Ma (Connelly and Dallmeyer, 1993). Regional cooling and muscovite growth in shear zones occurred ~350 Ma.

The cooling ages of hornblende (~430, 421± 15, and 415 ± 15 Ma) could be from peak metamorphism in the Ordovician (Dallmeyer, 1975; Moecher, 2005), although the ~440 Ma sphene suggests a thermal event. Biotite cooling at the same time as hornblende (~430 Ma) suggests very rapid uplift and cooling (~ 1 km/m.y.). Regionally consistent muscovite cooling ages of ~340 Ma in mylonite of the basement-cover detachment and ~350 Ma in two different shear zones suggests a deformation event between ~350 and ~340 Ma.

A major Alleghanian fault may exist between the Qualla-Dellwood and Cherokee-Ravens Fork anticlinoriums. Hornblende in Mesoproterozoic rocks in the Qualla-Dellwood anticlinorium cooled through the 500^{o} C closure temperature at 421 ± 15 Ma (Dallmeyer, 1975), monazite ages in cover rocks are 480 to 440 Ma, 430 to 415 Ma, and 410 to 360 Ma (Moecher et al., 2003), and muscovite ages are 287 and 281 Ma (Kish, 1991). Zircon fission-track ages of ~280 Ma in the central highlands indicate rapid cooling of rocks in the Early Permian, most likely related to emplacement of Alleghanian thrust sheets. Older zircon ages from the surrounding biotite- and higher-grade rocks most likely reflect regional cooling from pre-Alleghanian metamorphism. The pattern of zircon ages suggests doming in the area of the central Highlands, during or after emplacement of the Great Smoky thrust sheet (Naeser et al., this guidebook). The Gatlinburg fault system and the Cold Springs-Caldwell Fork fault system may mark the northwestern and southeastern boundaries of the dome.

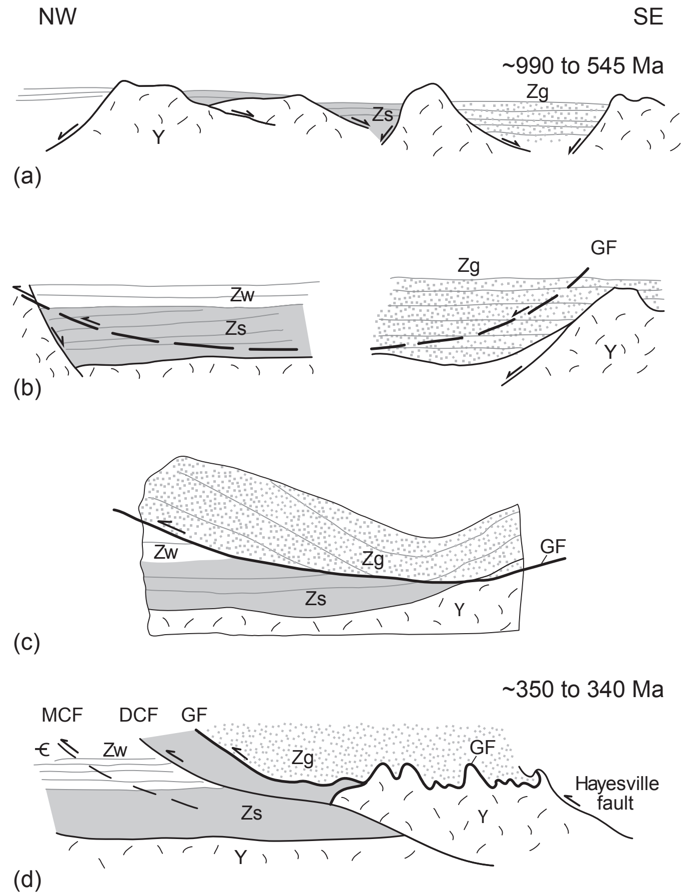

Fig. 3. Schematic diagram of the evolution of the Greenbrier fault (GF). *A*) Neoproterozoic crustal extension of Mesoproterozoic rocks (Y) resulted in grabens and half grabens. The grabens developed into depositional basins for the Snowbird Group (Zs), Great Smoky Group (Zg), and Walden Creek Group (Zw). *B*) The dashed line across the basins is the projected Greenbrier fault that had a listric geometry as it cut downsection in the Great Smoky basin and upsection in the Snowbird basin. *C*) The resulting geometry of the hangingwall and footwall rocks of the Greenbrier fault. Along strike from southwest to northeast, the fault cuts upsection in both the hangingwall and foothill rocks. *D*) The Greenbrier fault was reactivated between ~440 Ma and ~340 Ma, folded, and transported westward above the Dunn Creek (DCF) and Miller Cove faults (MCF), with later transport along the Great Smoky and Guess Creek faults ~280 Ma. C, Cambrian rocks.

In summary, our data support these events (Figs. 4 and 5):

1) plutonism, granulite-facies metamorphism and ductile deformation in the ~1 Ga Mesoproterozoic rocks.

2) pre-metamorphic faulting and folding (Fig. 4A).

3) ductile deformation and peak amphibolite-facies metamorphism before or about 440 Ma. Hornblende in Mesoproterozoic rocks cooled from ~430 to 415 Ma.

4) rocks of the Great Smoky Group were juxtaposed onto Mesoproterozoic rocks along the Greenbrier fault between ~350 and ~340 Ma.

5) non-coaxial F_2 folding and faulting under retrograde conditions ~365 to 340 Ma (Fig. 4B).

6) transport west along the Great Smoky fault system ~280

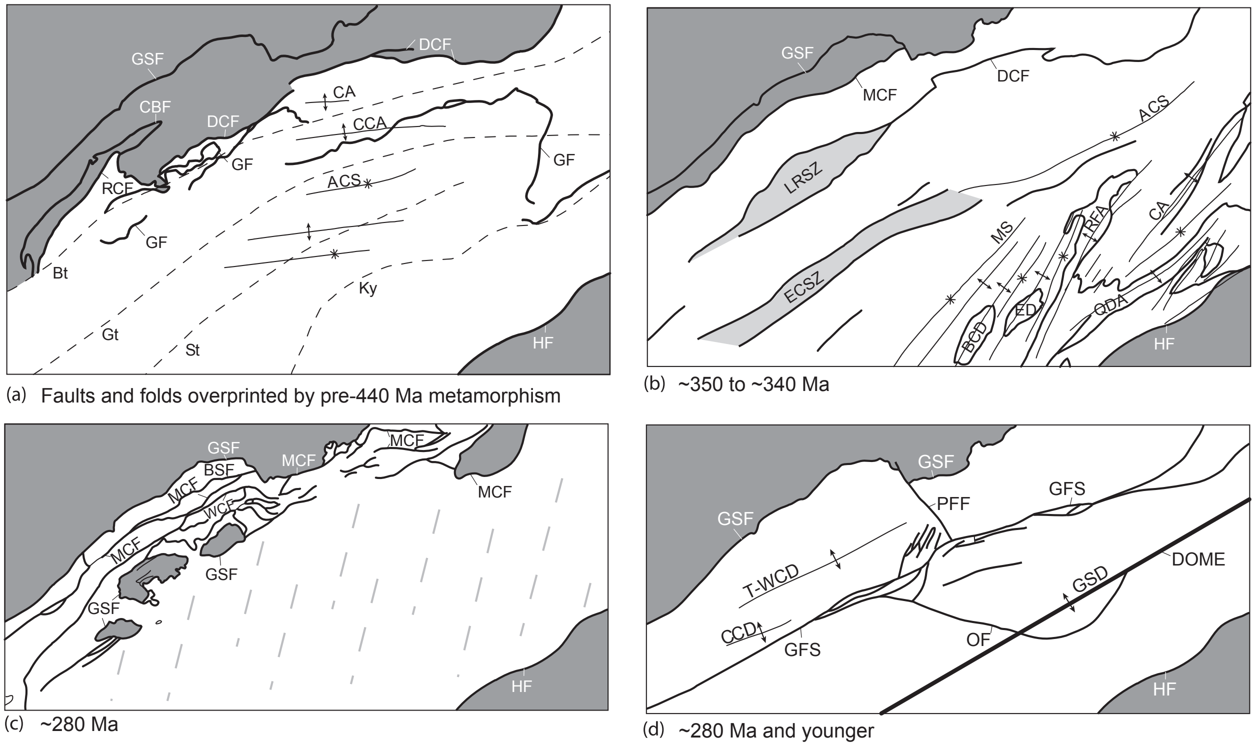

Fig. 4. Sequential structure maps of the GSMNP region. Surface trace of northwestern-most Great Smoky fault (GSF) and the Hayesville fault (HF) is shown for reference. Emphasis is on unshaded areas. *A*) Premetamorphic faults and folds and pre ~440 Ma amphibolite-facies metamorphic zone markers. Faults are the Greenbrier (GF), Dunn Creek (DCF), Rabbit Creek (RCF), and Capshaw Branch (CBF) faults; F_1 folds are the Alum Cave synclinorium (ACS), Copeland Creek anticline (CCA), and Carterstown anticline (CA). Metamorphism pre ~440 Ma; metamorphic zone markers are biotite (Bt), garnet (Gt), staurolite (St), and kyanite (Ky). *B)* The Greenbrier fault is reactivated in the southeastern area, and F_2 folds, faults and shear zones formed ~350 to 340 Ma. Folds are Murphy synclinorium (MS), Bryson City dome (BCD), Ela dome (ED), Ravens Fork anticlinorium (RFA), Cataloochee anticlinorium (CA), and Qualla-Dellwood anticlinorium (Q-DA). Faults are Eagle Creek shear zone (ECSZ) (shaded), Little River shear zone (LRSZ) (shaded), and Miller Cove fault (MCF). *C),* The Great Smoky fault (GSF) system includes the Walden Creek (WCF), Miller Cove (MCF), and Bogle Spring (BSF) faults, which were active ~280 Ma. North-northeast-trending cleavage transects all previous structures. *D*) Post-280 Ma faults and folds. Faults are Oconaluftee (OF), Gatlinburg fault system (GFS), and Pigeon Forge fault (PFF). Folds are the Tuckaleechee-Wear Cove dome (T-WCD), Cades Cove dome (CCD), and the Great Smoky dome (GSD).

Ma and north-northeast-trending cleavage transects all of the structures. (Fig. 4C).

7) ~280 Ma the Great Smoky fault system was cut by the Gatlinburg fault system, and the Oconaluftee and Pigeon Forge faults (Fig. 4D). Domes formed along Tuckaleechee Cove, Wear Cove, Cades Cove, and Green Mountain as the result of duplex beneath the Great Smoky fault as it moved forward.

8) A paleothermal dome formed in the central Highlands. Slow regional uplift and denudation in the Mesozoic and Cenozoic (Naeser et al., this guidebook).

METAMORPHIC AND STRUCTURAL EVENTS ELSEWHERE IN THE BLUE RIDGE PROVINCE

Paleozoic tectonic events that have been proposed for the rocks of the western Blue Ridge include:

1) 510 to 460 Ma (Middle Cambrian to Late Ordovician Taconian), with the continuous increase in metamorphic gradient to the southeast partly due to discontinuities along syn- and postmetamorphic faults (Butler, 1991).

2) 440 to 425 Ma (Taconian) overprinted by 350 to 325 Ma (Alleghanian) (Dallmeyer, 1988),

3) 480 to 440 Ma (Taconian) and ~390 Ma (Acadian) (Kish, 1991),

4) 460 to 440 Ma (Taconian) overprinted by 380 to 360 Ma (Neoacadian) (Connelly and Dallmeyer, 1993),

5) 480 to 450 Ma (Taconian), 330 to 300 Ma (Alleghanian) (Butler, 1991;Goldberg and Dallmeyer, 1997) and 325 to 265 Ma (Alleghanian) (Trupe et al., 2004),

6) 458 Ma (Taconian) (Moecher, 2005),

7) ~400 Ma (Acadian) (Kohn and Malloy, 2004),

8) 472 Ma (Taconic) and 327 to 320 Ma (Alleghanian) (Steltenpohl et al., 2001) (K-Ar and $^{40}Ar/^{39}Ar$ of hornblende and muscovite in Georgia).

9) 405.7 Ma (Acadian) ($^{40}Ar/^{39}Ar$ of phlogopite), and 350 to 300 Ma ($^{40}Ar/^{39}Ar$ muscovite) (Alleghanian) (Burton et al., 1992; Kunk and Burton, 1999), in northern Virginia,

10) 360 to 350 Ma (Neoacadian) and 330 to 325 Ma (Alleghanian) (R.D. Hatcher, Jr., 2005, written comm.),

11) 320 to 260 Ma (Alleghanian) ($^{40}Ar/^{39}Ar$, muscovite) (Wooten et al., 2005), in central Virginia.

12) Zircon fission-track data from Maryland south to North Carolina suggest regional cooling from temperatures less than 240°C at times ranging from 305 to 280 Ma (Alleghanian) (Naeser et al., 2004).

In summary, the rocks of the western Blue Ridge were metamorphosed at different conditions and at different times, and experienced differential uplift, cooling, and erosion, as different faults were active at different times. Locally the rocks were metamorphosed at amphibolite-facies from 472 to 458 Ma, ~440 Ma, and 400 Ma, and experienced greenschist-facies metamorphism and (or) cooling from 390 to 260 Ma

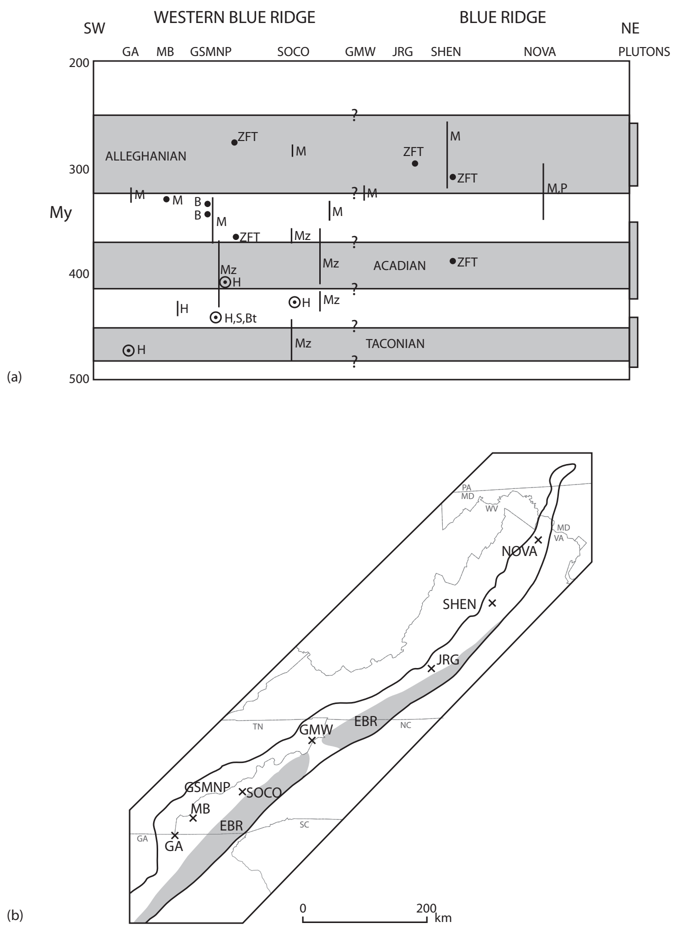

Fig. 5. Summary of Paleozoic geochronology of the Blue Ridge. *A)* Horizontal shaded bars show general timeframe of Taconian, Acadian, and Alleghanian orogenies; times of intrusions of plutonic rocks are shown as vertical bars on the right margin. B, biotite; H, hornblende; M, muscovite, Mz, monazite, P, phlogopite, S, sphene, ZFT, zircon fission track. *B)* Index map of Blue Ridge province showing locations and sources of data. GA (Georgia), $^{40}Ar/^{39}Ar$ H and M (Steltenpohl and others, 2001). MB (Murphy belt), $^{40}Ar/^{39}Ar$ H and M (Dallmeyer, 1988). GSMNP (Great Smoky Mountains National Park), $^{40}Ar/^{39}Ar$ H, Bt, and M (Dallmeyer, 1975; Kish, 1991); U-Pb S; ZFT (Naeser and others, this Guidebook); Electron microprobe and ion microprobe Mz (Kohn and Malloy, 2004). SOCO, U-Pb TIMS and U-Th-Pb chemical ages Mz (Moecher et al., 2003); K-Ar and $^{40}Ar/^{39}Ar$ M (Kish, 1991). GMW (Grandfather Mountain Window), $^{40}Ar/^{39}Ar$ M (Goldberg and Dallmeyer, 1997). JRG (James River Gorge), ZFT. SHEN (Shenandoah National Park), $^{40}Ar/^{39}Ar$ M (Wooten et al., 2005). NOVA (northern Virginia), $^{40}Ar/^{39}Ar$ M and P (Burton et al., 1992; Kunk and Burton, 1999). EBR, eastern Blue Ridge.

(Fig. 5). Regional cooling ranging from 305 to 280 Ma (Naeser et al., 2004, this guidebook), of western Blue Ridge rocks from North Carolina north to Maryland, and at about 280 Ma in the easternmost Piedmont Province (Kunk et al., 2004; 2005), is likely due to Alleghanian transport along the Pine Mountain-North Mountain thrust faults that tip out in the Valley and Ridge province. Recent isotopic studies across a 25-km- wide zone in the eastern Piedmont along the Potomac River near Washington, D.C., show that polyphase metamorphism and out-of-sequence deformation occurred at ~475 Ma, ~400 Ma, ~350 Ma, and ~300 Ma, prior to regional cooling at ~280 Ma (Kunk et al., 2004; 2005). It appears that a similar history of Paleozoic amalgamation is recorded in the rocks of the western Blue Ridge.

CONCLUSIONS

Suites of plutons (Aleinikoff et al., 2000; Carrigan et al., 2003; Tollo et al., 2004a, 2004b; Southworth et al., 2005) and at least two deformation events (Burton and Southworth, 2004; Tollo et al., 2004a, 2004b) characterize the 200 m.y. Grenvillian orogeny (Fig. 6). Following hornblende cooling ages of ~998 Ma (Kline et al., 1994) and 920 Ma (Kunk and Burton, 1999), crustal extension resulted in volcanic and plutonic activity over a ~100

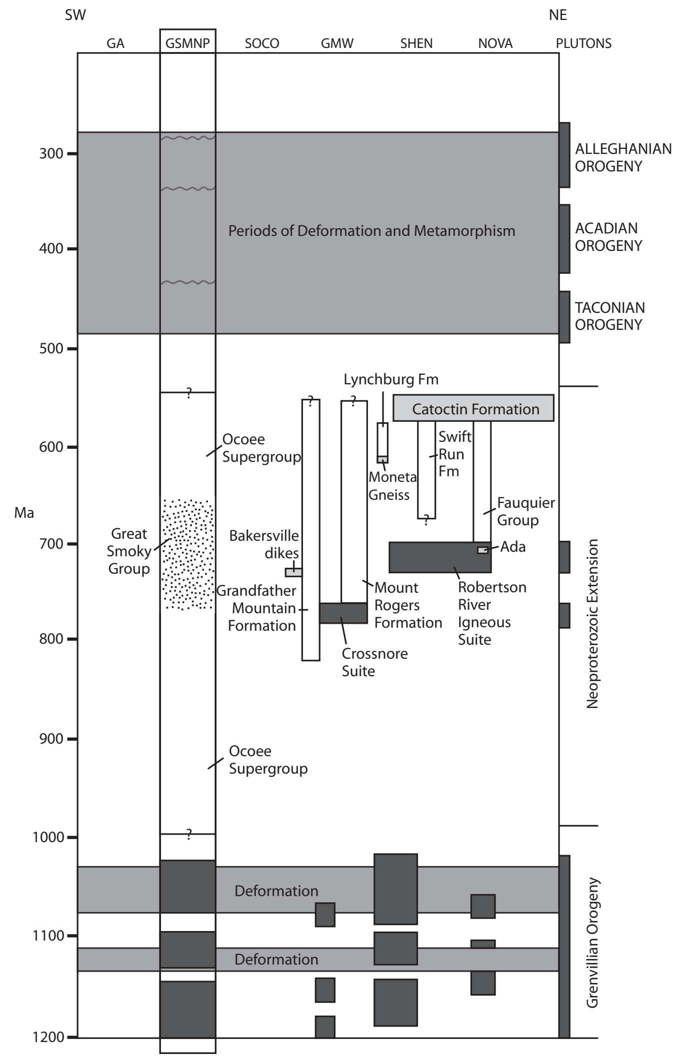

Fig. 6. Summary of geochronologic data of orogenies and tectonic events of the Blue Ridge province. The vertical shaded bar is the rocks of the GSMNP region. The times of intrusions of plutonic rocks are shown as vertical bars on the right margin. Suites of plutons (dark shade) (Aleinikoff et al., 2004; Carrigan et al., 2003; Southworth et al., 2005) and at least two deformation events (light shade) (Burton and Southworth, 2004; Tollo et al., 2004 a, 2004b) are associated with the ~1200 to 1029 Ma Grenvillian orogeny. Crustal extension resulted in volcanic and plutonic activity from ~810 Ma Grandfather Mountain Formation (Rankin et al., 1969) to the ~700 Ma Robertson River Igneous Suite (Tollo and Aleinikoff, 1996). Rocks of the Great Smoky Group have diagenetic mineral growth between 780 and 649 Ma (dark shaded), but the rocks of the Ocoee Supergroup may range from ~990 to 545 Ma (light shaded) (Aleinikoff et al., 2004). Metavolcanic events are at ~810 Ma (Grandfather Mountain Formation), 758 Ma (Mount Rogers Formation), 707 Ma (Ada metavolcanics), 607 Ma (Moneta Gneiss), ending with the 575 to 560 Ma Catoctin Formation. Although not regionally, spatially, or temporally consistent, plutonism, metamorphism, deformation, uplift, and cooling was happening somewhere in the Blue Ridge province from ~480 to 280 Ma.

m.y. range from ~810 Ma (Rankin et al., 1969) to ~700 Ma (Tollo and Aleinikoff, 1996) north of the GSMNP region. Great Smoky Group rocks were deposited during this time in geographically restricted basins with little local igneous activity. The duration of deposition of Ocoee Supergroup rocks, however, may be more than 350 m.y. (~990 to 550 Ma) (Aleinikoff et al., 2004). Correlative sedimentary rocks of the Mechum River Formation, Swift Run Formation, Fauquier Group, and Lynchburg Group (Hadley, 1970), were deposited on Mesoproterozoic rocks in the northern Blue Ridge over a period of 125 m.y. (~700 Ma to 575 Ma). Volcanism and intrusions of mafic dikes over a period of ~25 m.y. coincided with the opening of the Iapetus ocean in the northern Blue Ridge. The uppermost Catoctin metavolcanic rocks (~560 Ma) are overlain by the Chilhowee Group, with the 545 Ma Cambrian boundary in the lower part of the Chilhowee Group. The rocks were metamorphosed, deformed, and transported westward and to higher crustal levels over a period of ~150 m.y. They have seen ~300 m.y. of uplift, erosion, and cooling, to form the present landscape.

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Timing and pattern of metamorphism in the western and central Blue Ridge, TN and NC: Status and outstanding problems

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ABSTRACT

Details regarding the absolute age of regional metamorphism, porphyroblast-matrix microstructural relationships, deformation history, and 3-D orientation of isograds in the Western and Central Blue Ridge of the Great Smoky Mountains and adjacent terranes remain largely uncompiled on a regional scale. The Barrovian progression (chlorite through sillimanite) developed in Great Smoky and Snowbird Group metaclastic rocks of the Western Blue Ridge was shown to continue into migmatitic metapelites of the Central Blue Ridge, with the WBR-CBR boundary (Hayesville Fault) being folded and metamorphosed. Further regional along-strike comparisons of tectonometamorphic histories (e.g., into Blue Ridge Thrust Complex) remain to be elucidated. Staurolite through kyanite grade metapelites and migmatites of the Great Smoky Mountains Barrovian sequence in the eastern WBR are overprinted by a phase of deformation (tight to isoclinal folding) under retrograde conditions that produced recrystallization of matrix biotite, muscovite, and quartz (generally not plagioclase), folding and kinking of kyanite, and growth of biotite at expense of garnet. The foliation and microstructural features generated by this deformation event are widespread and could be used to correlate deformation histories throughout the WBR and CBR. Despite provocative proposals that orogenesis in the WBR and CBR is Acadian or Alleghanian, age data and the distribution of prograde isograds in the WBR and CBR are most consistent with Taconian orogenesis. Precise U-Pb zircon geochronometry on granulites and eclogites in the CBR yield an age of ~458 Ma, interpreted to be the time of peak thermal or baric conditions. This is supported by a dominant monazite Th-U-Pb chemical age mode of ~470 Ma obtained for staurolite and kyanite grade Great Smoky Group metapelites in the WBR. The age of the post-peak metamorphic deformation event is proposed to be Acadian and related to tectonometamorphism in the Eastern Blue Ridge and Inner Piedmont that propagated northwestward into the CBR and WBR. Much of the published geochronology for metapelites and metabasic rocks the Spruce Pine thrust sheet (Rb-Sr, Sm-Nd mineral isochrons and mineral-rock isochrons; U-Pb ID-TIMS analysis) is consistent with an Acadian overprinting of Taconian peak metamorphic assemblages. The meaning of much of the available K-Ar and ^{40}Ar-^{39}Ar geochronology is equivocal owing to complex deformation histories, lack of information on fabric relationships and mineral growth histories, disturbed isotopic systematics, or isotopic inheritance.

INTRODUCTION

The classic Barrovian metamorphic progression in the Ocoee Supergroup, eastern Great Smoky Mountains (TN-NC border), was first mapped by Hadley and Goldsmith (1963). Based on hand sample and thin section observations, metamorphic grade increased southward from chlorite grade in rocks immediately south of the Great Smoky fault to kyanite grade in the Cherokee-Maggie-Dellwood areas. Isograds strike ENE. The distribution of these isograds was further delineated to the south, southwest and northeast by Carpenter (1970) using heavy minerals separated from alluvial deposits. A sillimanite isograd was delineated, which approximately coincides with the presently inferred boundary between the western and central Blue Ridge terranes (Hayesville fault). Eckert et al. (1989) mapped

Moecher, D. P., Massey, M. A., and Tracy, R. J., 2005, Timing and pattern of metamorphism in the western and central Blue Ridge, TN and NC: Status and outstanding problems, *in* Hatcher, R. D., Jr., and Merschat, A. J., eds., Blue Ridge Geology Geotraverse East of the Great Smoky National Park, Western North Carolina: North Carolina Geological Survey, Carolina Geological Society Annual Field Trip Guidebook, p. 57-66.

additional isograds from Ocoee rocks of the western Blue Ridge (WBR) into the highest-grade region of the central Blue Ridge (CBR-Cartoogechaye terrane) where migmatites developed in pelitic gneisses and mafic rocks contain orthopyroxene. It has become generally accepted that this WBR-CBR prograde succession resulted from a single regional metamorphic event. Compilation of regional isograds in crystalline terranes of western North Carolina are also included on the state geologic map of North Carolina (Brown et al., 1985), and in Bream (2002). It is now clear, however, that these isograds must be regionally polychronic (i.e., they represent the effects of three phases of Appalachian tectonometamorphism partially overlapping in space and time). As a result, mineral assemblages and fabrics of younger tectonometamorphic events will likely overprint those of earlier events in broad zones parallel to the strike of isograds.

Although the WBR-CBR metamorphic progression in the Great Smoky Mountains has been well known for decades, we do not yet fully understand the details regarding timing of metamorphism, prograde reaction mechanisms, porphyroblast-matrix microstructural relationships throughout the P-T-D path, the 3-D orientation of isograds, and the relationship of minerals to the deformation history. The purpose of this paper is to review the status of these issues in the western and central Blue Ridge terranes in the area of Great Smoky Mountains National Park and adjacent terranes to the southwest and northeast, elucidate outstanding problems, and outline strategies for resolving these problems.

STATUS

Timing of metamorphism: Previous Work

Debate regarding the age of isograds in the WBR-CBR sequence is as old as the geochronologic methods applied to them. Remarkably, this debate continues today, due mainly to the interpretation of the meaning of dates obtained by different radiometric systems. Radiogenic isotope geochronology was in its infancy at the time of the early work of Hadley and Goldsmith (1963) (work carried out in the 1950s), and only broad age limits could be derived. The general absence of fossils in Ocoee strata was a further impediment to constraining timing of metamorphism.

Hadley and Goldsmith (1963, p. B2) stated that regional metamorphism was contemporaneous with deformation "…that began probably in Ordovician time and continued intermittently throughout most of the Paleozoic era." leading one to think they considered the isograds to be Taconian. However, perhaps influenced by exciting results in the nascent field of radiogenic isotope geochronology (Aldrich et al., 1958; Long et al., 1959), and assuming correlation of Bryson City pegmatites with those near Spruce Pine (dated at ~375 Ma based on Rb-Sr, U-Pb, and Pb-Pb ages on uraninite, muscovite, and feldspar, and ~340 Ma based on K-Ar ages on muscovite and biotite from Spruce Pine pegmatite and enclosing "Carolina gneiss": Aldrich et al., 1958), they state that: "Since the pegmatites of the Bryson City area probably accompanied the thermal maximum of post-Ocoee metamorphism, it follows that this metamorphism occurred 340 to 375 million years ago, in Middle or Late Devonian time…" (p. B107), implying the isograds are Acadian. This apparent contradiction by Hadley and Goldsmith neatly encompasses the issues being debated today. Based on a review of much of the same early geochronology, Carpenter (1970) also concluded that the isograds are mid-Paleozoic (375 to 325 Ma).

In a previous CGS Field Conference volume, Kish (1991) reviewed the methods, results and interpretation of K-Ar and ^{40}Ar-^{39}Ar ages for muscovite, biotite, and hornblende available at the time for WBR and CBR schists and gneisses between Great Smoky Mountains National Park and the NC-GA state line. Unfortunately, additional K-Ar dates for muscovite, biotite, and whole rocks did not resolve the issue of the timing of peak metamorphism (Taconian vs. Acadian). It is possible that there are some geologically meaningful dates in this dataset, particularly slates that have been assessed for muscovite crystallinity and that involved analysis of neocrystallized white mica (< 2 mm size fraction), or for coarsely crystalline muscovite in garnet grade or higher schists. However, the fact that many of the slate whole rock ages are older than the time of Taconian orogenesis demonstrates presence of a Precambrian detrital component in the slates. A general eastward younging (up to 350 Ma) of such ages occurs up to biotite grade, consistent with purging of inherited ^{40}Ar during prograde metamorphism. Muscovite K-Ar ages at garnet grade and higher scatter mostly within the range 340-370 Ma, with a preponderance of ages near 350 Ma, which agree with the whole rock ages. Kish (1991) also compiled new K-Ar and existing (Dallmeyer, 1975) ^{40}Ar-^{39}Ar ages for hornblende. Hornblende K-Ar ages in the sillimanite zone vary by 100 m.y. (378 to 490 Ma), but ^{40}Ar-^{39}Ar plateau ages for the staurolite-kyanite zone southeast of Great Smoky Mountains National Park and in the Murphy belt are more consistent (422-439 Ma).

Connelly and Dallmeyer (1993) obtained a similar pattern of ages as Kish (1991) in Ocoee slates and schists in the Great Smoky Mountains National Park, using ^{40}Ar-^{39}Ar incremental heating methods. Samples of the greenschist facies Pigeon metasiltstone (Chl-Ms-Qtz-Pl-KFs) contain white mica exhibiting illite crystallinity consistent with formation at ambient metamorphic grade. Three samples yield ^{40}Ar-^{39}Ar plateau ages of 460-440 Ma; four samples yield total gas ages of 420 Ma. In contrast, samples of Roaring Fork, Thunderhead, and Anakeesta Formation metasandstone and schist at biotite and garnet grade yield a single plateau age of 350 Ma and total gas ages of 345, 350, and 375 Ma, respectively. Samples of Thunderhead Fm. schist at staurolite and kyanite grade all yield muscovite plateau ages ranging from 360 to 370 Ma.

In interpreting solely the K-Ar and ^{40}Ar-^{39}Ar ages, we focus first on the highest grade rocks that yield straightforward Ar release patterns and define plateaus in the spectra: hornblende ages in basement gneisses and muscovite ages in staurolite and kyanite grade rocks in Great Smoky Mountains National Park (GSMNP) near Cherokee, NC. Kyanite-grade conditions in the southern GSMNP are 650-700 ^{o}C at 8-9 kbar (Anderson et al., 2005). Hornblende and muscovite plateau ages are reasonably interpreted as the time at which each mineral passed through their respective Ar blocking temperatures (~500 and 350 ^{o}C, respectively: McDougall and Harrison, 1988). One potential scenario is that regional metamorphism occurred at 460-440 Ma (plateau ages for greenschist facies Pigeon Fm. phyllites) and slow cooling (~5 ^{o}/m.y.) occurred through the Acadian. In contrast, Connelly and Dallmeyer (1993; p. 351) interpret the muscovite plateau ages (360-380 Ma) as the time of growth of biotite-grade through kyanite-grade porphyroblasts and attainment of peak thermal conditions, implying regional metamorphic isograds are Acadian. They further concluded that Acadian metamorphism overprinted an earlier Taconian metamorphism, indicateded by 460-440 Ma plateau ages in the phyllites, but apparently not preserved in higher-grade rocks. Clearly, if the plateau ages are cooling ages (the conventional interpretation) and muscovite ages are "Acadian", such metamorphism had to have occurred *before* 360-380 Ma (e.g., approx. 400 Ma). However, this interpretation does not explain the hornblende plateau ages for nearby basement rocks that are ~420 Ma. The hornblende and muscovite ^{40}Ar-^{39}Ar ages *together* are most consistent with slow post-Taconian cooling.

The above discussion assumes a single prograde metamorphic-deformation history and that K-Ar and ^{40}Ar-^{39}Ar analysis of *bulk* muscovite separates is valid for Ocoee metasediments in the Great Smoky Mountains region. We will discuss evidence for a more complex deformation and crystallization history in a subsequent section, and revisit the issue of interpretation of the Ar ages after a discussion of more recent developments in geochronology of these rocks.

Timing of Metamorphism: Recent Developments

Although ^{40}Ar-^{39}Ar is often the system of choice for dating greenschist to sub-greenschist facies metamorphism and deformation of rocks containing one or more generation of white mica (Dunlap et al., 1991; Markley et al., 1998; Sherlock et al., 2003), more robust approaches and systems for dating growth of index minerals need to be applied. This involves dating porphyroblast minerals, dating accessory phases that can be tied to specific isograd reactions, or dating fabrics that can be related to metamorphic mineral growth.

Two geochronometric systems were recently applied to obtain more precise and accurate limits on the timing of metamorphism in the WBR and CBR proper: zircon and monazite U-Pb geochronology employing isotope dilution-thermal ionization mass spectrometry (ID-TIMS), and microanalytical methods (ion microprobe U-Pb isotope analysis; electron microprobe Th-U-Pb chemical age dating). These methods are applicable in upper amphibolite to granulite (zircon) and lower amphibolite to granulite facies (monazite) metamorphic rocks.

The prograde succession of isograds mapped in the granulite facies high of the southern Blue Ridge in western North Carolina by Eckert et al. (1989) begins in kyanite grade rocks and extends through sillimanite I and II isograds. The latter isograds correlate with an increase in the degree of leucosomes present in migmatites. The progression overlaps the position of the Hayesville fault (CBR-WBR boundary) and corresponds structurally with the road cut at Stop 1–4 of the 2005 CGS field trip. The road cut at Winding Stair Gap on U.S. 64 southwest of Franklin, NC (Absher and McSween, 1985) exposes a variety of migmatite styles with leucosomes that probably formed by several melting reactions. One particular leucosome studied by Moecher et al. (2004) formed by extensive biotite dehydration melting of Pl-Grt-Sil-Bt-Ru gneisses, and generated Pl + Grt leucosomes containing an exceptionally large amount of zircon that was also a product of the melt-generating reaction. The leucosomes cross cut the youngest fabric in the enclosing gneisses, are undeformed, and equilibrated at ~850 ^{o}C and 8-9 kbar. These conditions agree with the thermobarometry of Eckert et al. (1989) and Tenthorey et al. (1996) for nearby granulites, and are consistent with conditions determined experimentally for formation of such melt reactions (Vielzeuf and Holdaway, 1988; Patiño Douce and Johnston, 1991). Zircon was analyzed by ID-TIMS and SHRIMP, yielding ages of 458 $\pm$ 1 Ma and 460 $\pm$ 12 Ma, respectively (Moecher et al. 2004). We interpret the age to correspond to the time of peak P-T conditions in the granulite facies zone of the CBR in North Carolina. If these high-grade rocks are a product of the same metamorphic event that generated the regional isograds, then the latter are Taconian.

Th-U-Pb chemical ages calculated from electron probe microanalysis of monazite were obtained for several samples of Great Smoky Group schist collected on the Blue Ridge Parkway along the CGS field trip route. Although this method has been tested for nearly a decade, and several studies demonstrate the power of the method when carefully applied (Williams and Jercinovic, 2002), many workers remain dubious that the system yields meaningful ages. These concerns are well founded. Chemical ages are very sensitive to the measured Pb content, which occurs in low concentrations (100-1000 ppm) in Paleozoic monazite; the measured Pb concentration is in turn sensitive to background spectral interferences and other analytical parameters (Pyle et al., 2002). Furthermore, individual monazites may possess complex compositional zoning (Foster et al., 2002; Gibson et al., 2004). Successive growth periods may be observed in single grains, with all growth zones not necessarily being concentrically arranged (Spear and Pyle, 2002). Phases of monazite growth observed by X-ray mapping in some grains may not all necessarily correspond to a period of regional metamorphism. In other cases, complex growth zones may correspond to the same phase of metamorphism, with zones corresponding to reactions among major and accessory minerals during the prograde path. Bearing these complexities in mind, a program of electron probe microanalysis of monazite was carried out to further constrain the timing of staurolite- and kyanite-grade metamorphism in the Great Smoky Mountains.

Samples of Grt+Ky$\pm$Str+Ms+Pl+Bt+Qtz+Ru+Ilm schist were collected along the Blue Ridge Parkway near Big Witch, Soco, and Balsam Gaps. X-ray mapping and electron probe microanalysis of monazite was carried out by Prof. Robert Tracy at VPI. Compositional domains are identified by X-ray mapping of each monazite grain, and then multipoint traverses are carried out across each zone. Individual grains may involve 50 to 100 point analyses that take 30 minutes each, in order to accumulate a statistically meaningful sample. An age is calculated for each grain, based on the measured concentration of Th, U, and Pb. Histograms are then constructed from the ages, and in some cases probability age distributions are calculated. Numerous studies carried out in the VPI microprobe lab have yielded ages that agree closely with ID-TIMS analysis of monazite or other geochronologic systems, where such comparisons are possible.

Two samples of Great Smoky Group schist are particularly useful (Moecher et al., 2004b). One is a Grt-St-Ky schist from the BRP near Big Witch Gap (sample GT02-2) that does not exhibit evidence of the late fabric overprint that is widespread in the field area (see the following section and Stop 1–4 description). The second sample is a Gr-St-Ky schist (MM22A) that is weakly overprinted by S_3 collected at Stop 1–4. Monazite in such rocks is commonly interpreted to grow during garnet-grade and higher

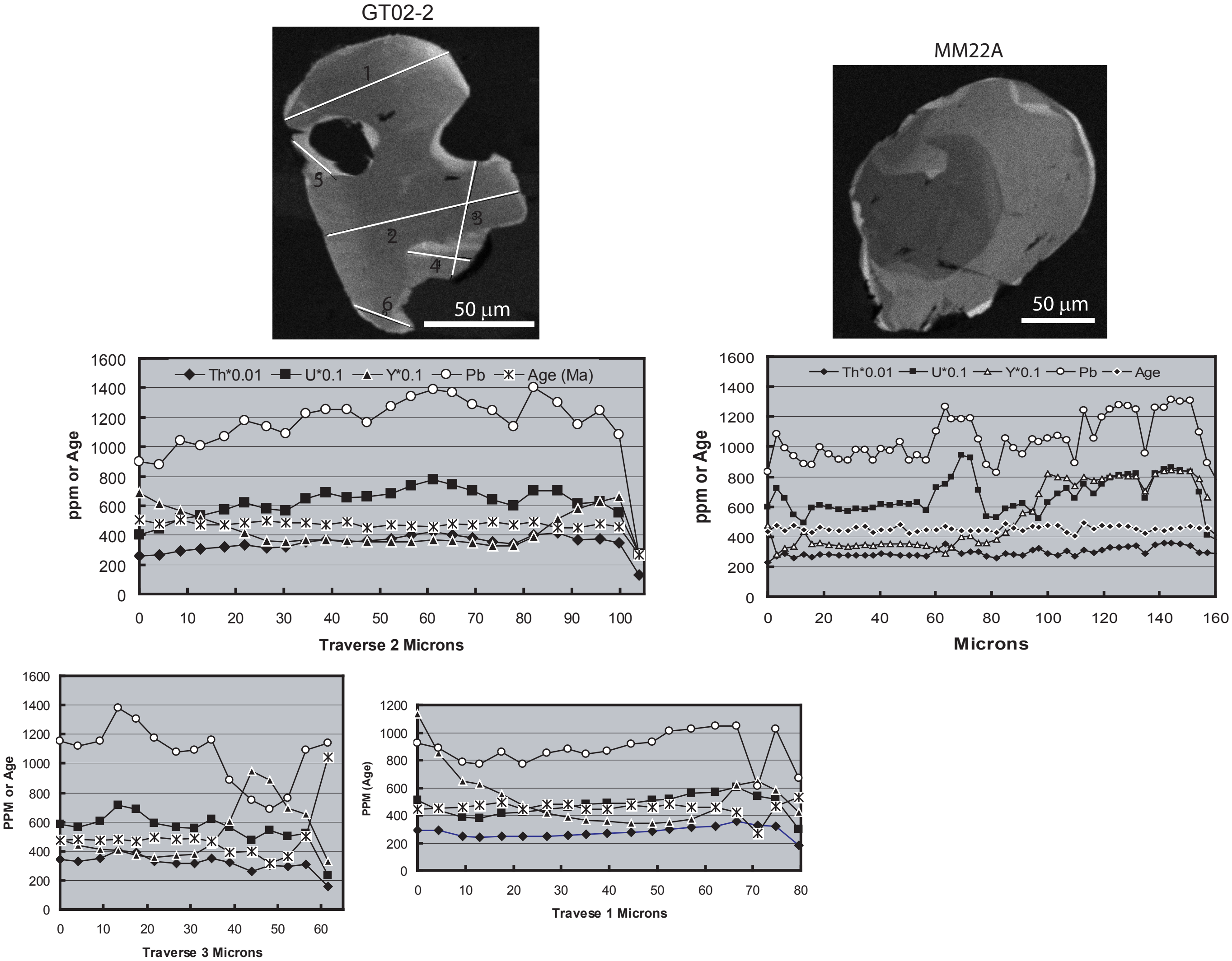

Figure 1. Yttrium X-ray maps (lighter contrast corresponds to higher Y concentration) and concentrations of Y, Th, U, and Pb from probe traverses (lines) for monazite from two samples of Great Smoky Group schist collected along the Blue Ridge Parkway. Sample GT02-2 exhibits a weak core to rim zoning with local patches of high Y content. MM22A is more distinctly zoned with a distinct core and rim (probably a growth feature). The zoning outlined by the Y contrast and apparent on the probe traverses do not necessarily correlate with Th or U concentrations. Although distinct compositional variations may exist across each grain, in these cases the calculated dates are generally internally consistent for each grain.

prograde metamorphic reactions (Smith and Barreiro, 1990; Spear and Pyle, 2002; Gibson et al., 2004; Kohn and Malloy, 2004). Therefore, the age of monazite in these samples should correspond closely to the time of regional metamorphism that formed the isograds. The monazites selected for analysis in these samples show weak or concentric yttrium zoning (Fig. 1). Both samples exhibit a dominant age mode of 450-480 Ma, and the vast majority of all ages obtained are in the range 450-500 Ma (Fig. 2). Although the monazite Th-U-Pb chemical age method does not permit a precise age determination, the ages in general are consistent with monazite growth during Taconian metamorphism of Ocoee rocks in the vicinity of GSMNP. The ages are also consistent with the zircon U-Pb analysis of the highest-grade rocks in the CBR to the southwest. We thus have strong evidence that the regional metamorphic isograds in the WBR-CBR area are Taconian. This does not preclude the possibility that Acadian or younger metamorphic effects are preserved in the area (discussed further below). In fact, Kohn and Malloy (2004) reported monazite chemical ages of ca. 400 Ma from Murphy belt and Great Smoky Group garnet-staurolite schists from the northeast end of the Murphy Syncline (southwest of the study area).

Deformation Phases and Effects

Although there appears to be a smooth, unbroken pattern of isograds in the GSMNP region, a punctuated deformation history was recognized by Hadley and Goldsmith (1963). Their deformation history involves an early premetamorphic folding of bedding (S_0: Hadley and Goldsmith did not use the S_n/F_n designations for foliation-forming fold events) to generate map scale F_1 open to tight folds (e.g., Alum Cave syncline), and an axial plane slaty cleavage (S_1) in chlorite-grade rocks of the Foothills belt and within shale interbeds in the Thunderhead sandstone below garnet-grade. Apparently, no index mineral growth is associated with this fold phase. A second generation of folding (F_2) was interpreted to produce a regional crenulation cleavage ("slip cleavage" of Hadley and Goldsmith, 1963) and foliation (S_2). Growth of isograd minerals is inferred to occur during a static phase of heating between the development of S_1 and S_2. Unfortunately, no other details exist that provide

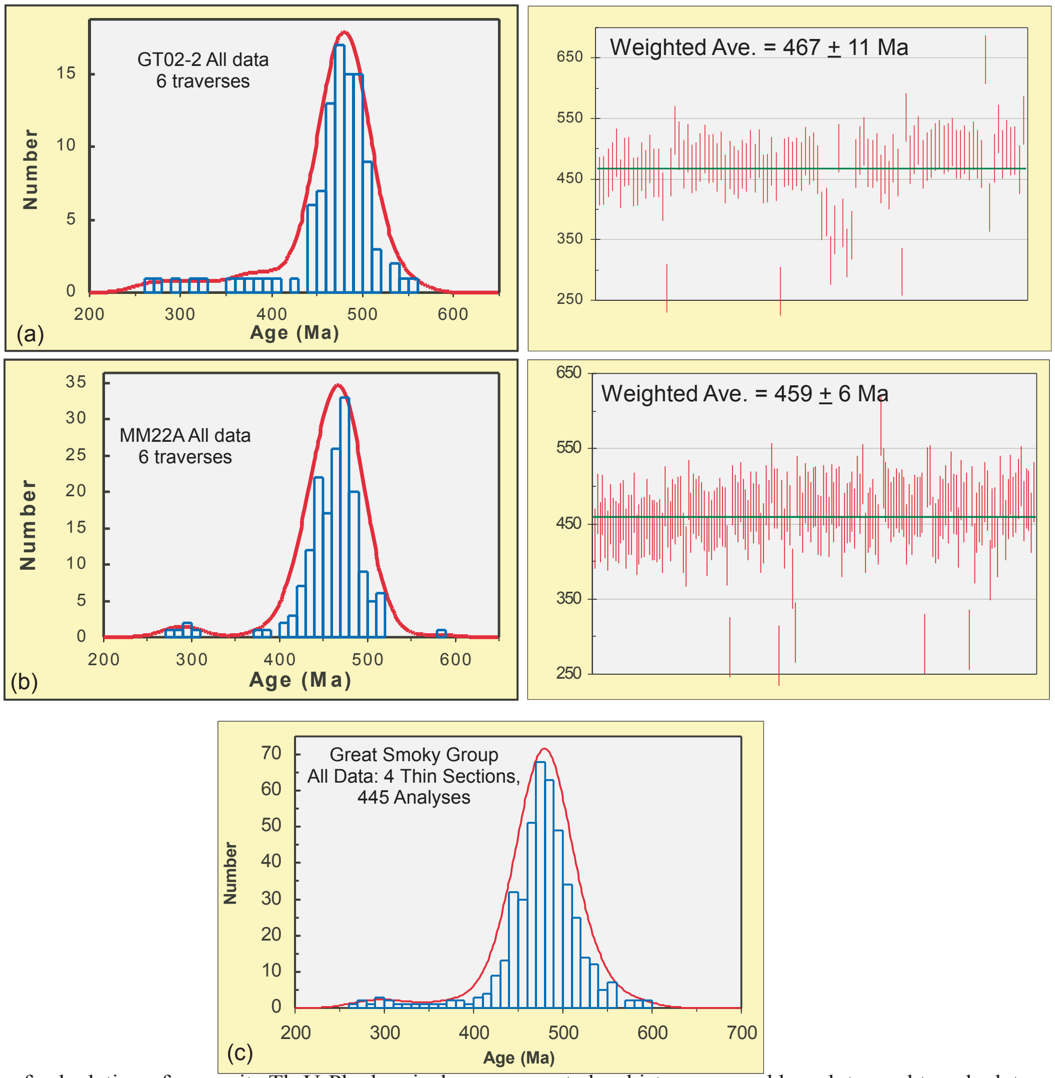

Figure 2. Results of calculation of monazite Th-U-Pb chemical ages presented as histograms and bar plots used to calculate a weighted average age for each monazite (Moecher et al., 2004b). Weighted average ages use only those values that fall within a narrow range near the mean. (a) Results for Great Smoky Group schist sample GT02-2 collected west of Big Witch Gap, Blue Ridge Parkway. (b) Results for Great Smoky Group gneiss collected at WBR-CBR boundary on Blue Ridge Parkway at Stop 1–4 (Woodfin Valley overlook). (c) All ages from four thin sections. Note that the number of older ages (525-575 Ma) is greater on this histogram. These ages could correspond to preservation in another sample from the Blue Ridge Parkway (MM9b: Moecher et al., 2004b) of detrital Neoproterozoic/Cambrian monazite that was overgrown by the Taconian generation, or a Neoproterozoic/Cambrian phase of tectonometamorphism (e.g., Hibbard et al., 2003). A weighted average age would not be appropriate to calculate for this sample.

greater insight into the metamorphic and fabric evolution of the Ocoee metasedimentary rocks in the GSMNP area during the isograd-forming tectonometamorphic event. The key questions are whether and how an S_1 schistosity (in coarse-grained muscovite-rich rocks, and temporally equivalent to S_1 slaty cleavage) developed in rocks at garnet grade and higher, what are the porphyroblast-matrix microstructural relationships for this event, and how was the earlier schistosity overprinted by the late crenulation cleavage?

The important feature present in the Ocoee metasedimentary rocks is the late crenulation cleavage developed during the open to tight folding event (corresponding to the F_3 event of Massey and Moecher, 2005; discussed further in the field trip guide). This event generated many of the outcrop and map scale folds in the Cherokee-Dellwood area (higher-grade regions) of the Great Smoky Mountains (Hadley and Goldsmith, 1963), including the exposures at Stop 1–4. The axial plane foliation (S_2 of Hadley and Goldsmith, 1963; S_3 of Massey and Moecher, 2005) associated with that event is characterized by numerous, relatively low temperature (≤ garnet grade) deformation microstructures and fabric elements, which are discussed in detail in the description of Stop **1–4**. This foliation is nearly pervasive throughout the area, but isolated occurrences exist where Great Smoky Group schists were not affected by the S_3-generating deformation phase (e.g., Grt-St-Ky schist near Big Witch Gap; Anderson et al., 2005; Fig. 3, this study). The dominant outcrop scale effect is open to tight folding of migmatitic layering in gneisses near the WBR-CBR boundary, and millimeter-scale crenulation of mica folia in fold hinges. However, such exposures reveal little regarding the degree of grain scale deformation effects, which are pervasive and striking in thin section. The deformation event that produced this foliation was accompanied by remarkably little retrogression of regional porphyroblasts, but nearly complete recrystallization of matrix micas and quartz. Most striking is the kinking and folding of individual kyanite porphyroblasts (Fig. 4a-c), which is ubiquitous in the study area and occurs as far

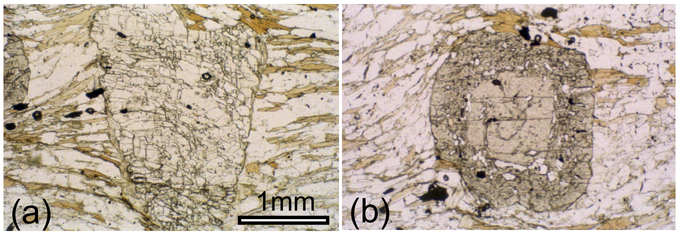

Figure 3. Textural features of porphyroblasts in Great Smoky Group schist collected near Big Witch Gap, Blue Ridge Parkway. This sample is only mildly affected by the late D_3 folding event that generated the "slip cleavage" of Hadley and Goldsmith (1963) (designated S_3 by Massey and Moecher, 2005; see Stop 1–4 description in accompanying field guide). (a) Photomicrograph (plane light) shows kyanite porphyroblast overgrowing matrix foliation that is only slightly deflected around the grain margin, suggesting minor post-kyanite-growth matrix recrystallization. Staurolite porphyroblasts (not shown) exhibit a similar relationship. (b) Garnet porphyroblast with randomly oriented inclusions (plagioclase, quartz, biotite, rutile); matrix is slightly more deflected around garnet compared to kyanite in A.

southwest as the Fontana Lake area studied by Mohr (1973). The description of Stop 1–4 provides documentation of these outcrop to grain scale deformation features. A final phase of mineral growth in schists and gneisses involves the nucleation of very fine-grained fibrous sillimanite along plagioclase grain boundaries (Fig. 4d). X-ray mapping reveals plagioclase grain boundaries are enriched in Ca relative to cores. The growth of grain boundary sillimanite has been interpreted to occur by a base cation leaching reaction such as:

$$2NaAlSi_3O_8 + 9H_2O + 2H^+ = Al_2SiO_5 + 2Na^+ + 5Si(OH)_4$$

(Vernon, 1979). The sodium depletion of plagioclase observed in X-ray maps, and absence of any other reaction or product phases, is consistent with this mechanism. Kerrick (1987) proposed that the source of acidic fluids for such reactions is crystallizing granitic plutons. Although this type of sillimanite is not related to regional metamorphic sillimanite developed at the highest grades in the WBR and CBR (Massey and Moecher, 2005), it is of regional extent, having been described by Goldberg et al. (1992) in the Ashe metamorphic suite of the Spruce Pine thrust sheet. This sillimanite occurrence is an important link between the two areas, as discussed further below.

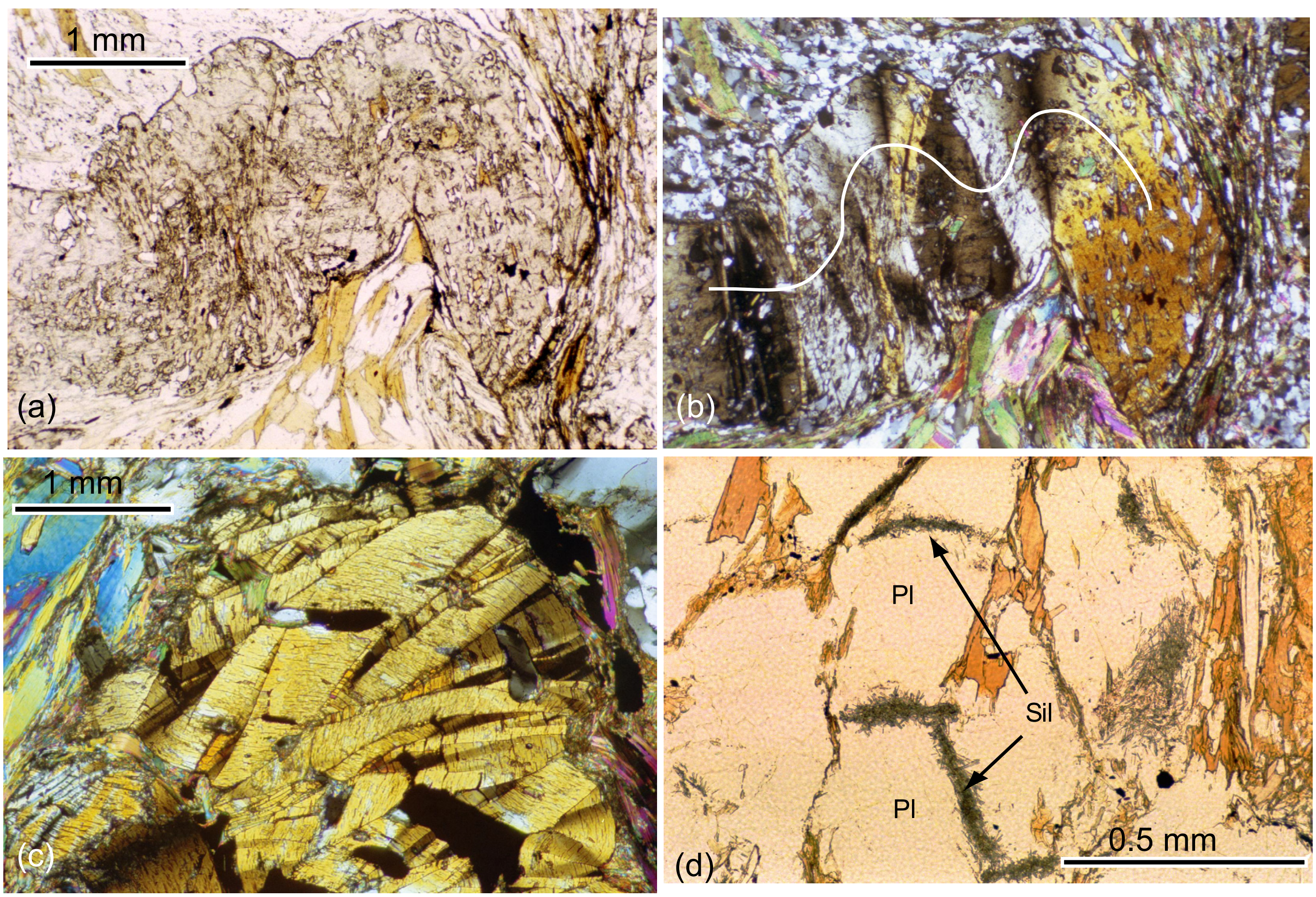

Figure 4. Textural relationships exhibited by Al_2SiO_5 polymorphs affected by D_3, post-peak metamorphic folding event in Great Smoky Mountains area. (a) and (b) Plane light and crossed polars view of folded kyanite porphyroblast in Anakeesta Fm. collected along NC Hwy 28 near Almond, NC. Folding is accommodated here primarily by bending of the kyanite lattice through a ~110° angle (white line outlines C axis of elongate kyanite blade). (c) Crossed-polars photomicrograph of folded kyanite in Great Smoky Group schist collected along Blue Ridge Parkway. Folding is accommodated primarily by kinking and twinning, with some bending of the lattice. (d) Post-peak metamorphic grain boundary sillimanite nucleating along plagioclase grain boundaries. These types of features (also see Stop 1–4 description) can be used to correlate deformation and metamorphic events along strike in the western and central Blue Ridge.

TACONIAN OR ACADIAN METAMORPHISM? OR BOTH?

Although the recent studies cited above are most consistent with regional isograds in the study area being Taconian, there is in all likelihood an Acadian metamorphic "overprint" that impacts fabrics and potentially isotopic systematics of geochronologically useful systems. Acadian metamorphism has been proposed to be the dominant event the neighboring eastern Blue Ridge and Inner Piedmont-Tugaloo terranes (Bream, 2002). Acadian deformation and thermal effects in the latter terranes would be expected to impinge to varying degrees on easternmost WBR and CBR rocks. In order to gain further insight into the potential nature of this overprint in the Great Smoky Mountains area, we must rely on the results of geochronologic studies in the Blue Ridge thrust complex to the northeast. In particular, the geochronology of Goldberg and Dallmeyer (1997) on the Ashe metapelites and amphibolites, and Miller et al. (2000) on the Spruce Pine eclogites, illustrate the nature of the solution to the problem of fully understanding polymetamorphism in the WBR and CBR near the Great Smoky Mountains.

The polymetamorphic nature and protracted tectonic history of crystalline rocks in the Blue Ridge thrust complex have been known for some time (Butler, 1973; McSween et al., 1989; Goldberg et al., 1992; Goldberg and Dallmeyer, 1997; Trupe et al., 2003). Thrust sheets consisting of amphibolite facies Ashe Formation Metamorphic Suite lithologies and Spruce Pine granitic plutons are bounded by thrust faults (e.g., Holland Mtn. fault) or strike slip shear zones (e.g., Burnsville fault; Trupe et al., 2003), and contain numerous internal shear zones that overprint regional metamorphic assemblages and folds. Goldberg and Dallmeyer (1997) used Rb-Sr and Nd-Sm mineral- (garnet, muscovite, hornblende) whole-rock isochrons to date mineral growth in Ashe pelitic schist and amphibolite in the Spruce Pine and Pumpkin Patch thrust sheets. The mineral assemblage in pelitic schists is Ms + Bt + Pl + Qtz + Gt $\pm$ Ky $\pm$ St $\pm$ Sil. Kyanite exhibits late kinking, and sillimanite is a fibrous grain-boundary variety (mainly plag-plag grain boundaries). Both such Al_2SiO_5 textures are observed in the Great Smoky Group rocks, as described above. Thermobarometry on Ashe metapelites in the Spruce Pine sheet yields 7-9 kbar at 650-700 ^oC (Goldberg et al., 1992); amphibolites tend to yield higher pressures. These too are similar to results obtained for Great Smoky Group schists and amphibolites (Anderson et al., 2005). Amphibolites analyzed by Goldberg and Dallmeyer (1997) are generally Hbl + Pl $\pm$ Ep $\pm$ Grt $\pm$ Ep assemblages, but relict higher-grade (Cpx-bearing, some eclogitic) metabasites are present in the Blue Ridge thrust complex north and south of the Grandfather Mountain window (Willard and Adams, 1994; Abbot and Greenwood, 2001).

Most of the garnet-whole rock Rb-Sr and Nd-Sm ages range between 450 and 460 Ma; one sample near the base of the Spruce Pine thrust sheet is younger (~390 Ma) (Goldberg and Dallmeyer, 1997). Another sample of Ashe metapelite from the Spruce Pine thrust sheet (their sample 24) yielded relatively precise garnet-whole rock Rb-Sr and Nd-Sm ages of 456 and 455 Ma, respectively, but the same sample yielded a Rb-Sr muscovite-whole rock age of 368 Ma. The hornblende-whole rock isochron ages are more evenly split, defining age modes of ~450-470 Ma and ~370-390 Ma. A garnet-bearing Bakersville metagabbro yielded Rb-Sr and Sm-Nd mineral isochron (defined by Hbl, Pl, Ep, and whole rock isotopic ratios) ages of 458 and 457 Ma, respectively. In contrast, an epidote-bearing amphibolite yields mineral isochron ages of 397 and 375 Ma. This pattern of ages is nearly identical to that obtained for the Spruce Pine eclogites. Eclogite facies metamorphism, based on U-Pb zircon ages (Miller et al. 2000), is dated at 459 Ma. The age of the widespread amphibolite facies overprinting (including deformation) of eclogite assemblages and fabrics is represented by the time of titanite formation (394 Ma).

All hornblende ^{40}Ar-^{39}Ar release spectra from the Spruce Pine and Pumpkin Patch thrust sheets are markedly discordant. This implies that either the crystal structure or chemical composition of hornblendes have been modified to an unknown degree since growth and initial decay of K to Ar. Calculated isotope correlation ages for non-mylonitic hornblendes in amphibolites are 385-400 Ma. All muscovite samples from Ashe metapelites exhibit release spectra that define plateaus with ages in the limited range of ~325-335 Ma.

Interpretation

A temperature-time plot summarizes the results of the above geochronology and provides a framework for comparing alternative scenarios to explain the age data (Fig. 5). Temperatures are based on thermobarometry or conventionally interpreted closure temperatures for Ar diffusion in hornblende and muscovite. When interpreting the Rb-Sr and Nd-Sm data, recall that a true isochron requires three or more points to define an age (Faure, 1986), although numerous other studies employing the two point isochron approach have yielded important insights into rates of metamorphism (Christensen et al., 1989; Vance and

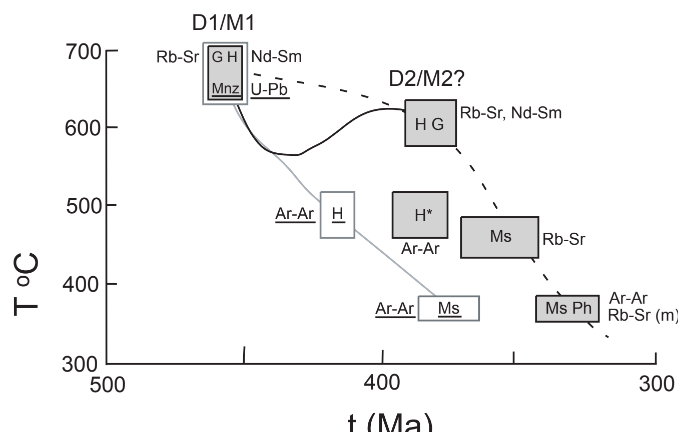

Figure 5. Schematic temperature-time-deformation (T-t-D) histories for Great Smoky Mountains (faint boxes and lines) and Blue Ridge thrust complex (black boxes and lines) based on thermochronology and thermobarometry discussed in the text. H* designates disturbed ^{40}Ar-^{39}Ar systematics for hornblende from Blue Ridge thrust complex amphibolites (Goldberg and Dallmeyer, 1997). Although data are sparse for the Great Smoky Mountains area of the western Blue Ridge, and the post-peak metamorphic S_3 folding event is widely distributed, the T-t-D history appears to be a simple static post-Taconian cooling, assuming hornblende and muscovite ^{40}Ar-^{39}Ar ages (Dallmeyer, 1975) were not affected by the S_3-forming event of Massey and Moecher (2005). Dashed and solid black lines portray potential alternative T-t-D histories for the Spruce Pine thrust sheet (Goldberg and Dallmeyer, 1997). The alternatives differ in the rate and degree of cooling from post-D_1/M_1 Taconian conditions. Mineral assemblages and thermobarometry suggest that D_2/M_2 was at lower grade than D_1/M_1 but do not permit assessment of the Taconian to Acadian T-t path.

O'Nions, 1990). However, the multimineral Rb-Sr and Sm-Nd isochron ages agree with several garnet- and hornblende-whole rock two point isochron ages, suggesting the latter are meaningful ages (particularly in view of their consistency). These ages, and zircon U-Pb TIMS ages for eclogites in the Spruce Pine thrust sheet (Miller et al., 2000), constitute internally consistent evidence that regional metamorphism was Taconian (M_1 on Fig. 5), and essentially the same age as that in the WBR-CBR region of the Great Smoky Mountains and granulite facies high to the southwest. However, the Ashe metapelites and particularly the amphibolites exhibit evidence of Acadian-age mineral growth, and resetting or disturbance of Taconian isotopic systematics. The apparently scattered distribution of isotopic effects suggests that the cause of Acadian (400-380 Ma) 'metamorphism' in the Great Smoky Mountains and Blue Ridge thrust sheets ("M_2?", Fig. 5) was non-penetrative, albeit regional in extent. An Acadian thermal perturbation would be conducted essentially equally throughout the terrane such that all rocks would experience the same thermal effects. Alternatively, deformation partitioning within zones of particularly reactive or weak lithologies also explains the pattern of overprinting. Deformation partitioning during the retrograde P-T-t path from Taconian high grade conditions (dashed line, Fig. 3) would assist in driving mineral reactions or (partial to complete) elemental and isotopic re-equilibration at whatever point along the post-Taconian P-T path where the rocks were deformed.

For example, sample 24 of Goldberg and Dallmeyer (1997) yielded a Taconian garnet-whole rock two point Rb-Sr isochron date, but an Acadian muscovite-whole Rb-Sr rock isochron age date. Clearly, the Rb-Sr isotopic systematics of the sample were modified since garnet growth as the three components do not define an isochron. The garnet represents equilibration with the system (whole rock) at a different time (earlier) than the muscovite. The relatively high closure temperature of garnet for Sr diffusion (Coughlan, 1990; Vance and O'Nions, 1990) renders garnet closed to re-equilibration with other phases in the rock upon crystallization during prograde metamorphism. A phase of deformation at ~600 ^{o}C that generates a new fabric via recrystallization of weak matrix phases would permit muscovite to re-equilibrate with the bulk of the rock (except garnet). Since so little of the Rb and Sr is contained in garnet (0.3 and 0.7 wt %, respectively), the whole rock values of $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ are changed only slightly by removing garnet from calculation of the whole rock composition. Doing so affects the calculated two-point muscovite whole-rock age by 1 (garnet mass fraction of 0.01) to 15 m.y. (garnet mass fraction of 0.10). The muscovite whole-rock ages are still broadly Acadian. A more meaningful age would be obtained by analyzing the isotopic composition of the schist with the garnet having been physically removed before dissolution rather than by calculation from a mineral mode as is done here.

Although the style of post-Taconian tectonism (specifically, Alleghanian thrusting) is distinctly different, several observations permit us to link the metamorphic history of the Spruce Pine thrust sheet to the WBR-CBR region studied here, and provide a hypothetical framework in which to consider WBR-CBR tectonometamorphism. First, peak metamorphism appears to be Taconian in both settings. Although many questions remain (e.g., How essentially contemporaneous eclogite and granulite facies metamorphism related?) the P-T-t histories for the Barrovian style assemblages start at nearly the same point for the eclogite assemblages. Second, peak kyanite-grade assemblages are overprinted by a enigmatic fibrous grain boundary sillimanite growth event, and a period of fabric overprinting in which kinking of the kyanite occurred, matrix phases were deformed around porphyroblasts, and a new axial plane foliation developed (S_3 of Massey and Moecher, this volume). We propose, therefore, that these late deformation-induced features in the WBR-CBR correlate with the overprinting in the Blue Ridge thrust sheets to the northeast, and are likewise Acadian.

PROPOSALS FOR TESTING HYPOTHESES AND RESOLVING UNCERTAINTIES

Although there exists a plethora of geochronometric dates for metamorphic rocks in the western and central Blue Ridge, few are readily interpretable, and unfortunately many are useless. Many samples exhibit 'disturbed' systematics (e.g., ^{40}Ar-^{39}Ar hornblende data for Blue Ridge thrust sheets), or, as with the K-Ar ages, possess uncertainties regarding unsupported Ar that cannot be assessed as to the argon source. Such situations result in scattered ages over even small areas where the T-t history should be the same for all samples. Many ages are 'ballpark' estimates (e.g., two point isochrons) that seem to cluster around what most of us believe to be the major Appalachian tectonometamorphic events. The ages are at best broad constraints in too many cases. One is left wondering what all the geochronology means. Unfortunately, we spend too much time arguing among ourselves about poorly constrained dates.

Part of the reason for the lack of meaningful ages is our lack of understanding of the details of the deformation, structural, and metamorphic history of our rocks, i.e., the context for the phases on which we perform geochronology. With certain exceptions, it is fair to say that large areas in the WBR (e.g., biotite through staurolite grade in the Ocoee Supergroup, Great Smoky Mountains) suffer from an incompletely documented megascopic deformation history, structural and fabric evolution, porphyroblast-matrix microstructural evolution, and foliation development history. Therefore, the first step in making meaningful advances in our understanding would be mapping and detailed structure and fabric analysis of rocks in quadrangles between the Blue Ridge thrust complex and the WBR and CBR in the Great Smoky Mountains proper. For example, Merschat and Wiener (1988) published or have available maps for the four quadrangles immediately east of GSMNP. Detailed petrologic and structural studies within these areas would permit extrapolation of deformation histories to the northeast, toward the Blue Ridge thrust complex. Similar petrologic studies moving southwestward out of the Blue Ridge thrust complex toward the Great Smoky Mountains region would also help resolve several longstanding questions and issues.

Detailed characterization of samples would be followed by geochronology using robust metamorphic phases that allow precise, geologically meaningful ages to be obtained. For example, U-Pb ID-TIMS analysis of titanite in amphibolite grade mafic to intermediate rocks is probably the best system to apply regionally. Titanite possesses a high closure T for Pb diffusion (Cherniak, 1993), it grows during prograde metamorphism, and relict igneous vs. neocrystallized metamorphic compositions can be discriminated via electron probe microanalysis. Zircon in high-grade rocks is also a useful phase, although there are diverging opinions on the how to obtain the most meaningful ages (ion beam microanalytical methods versus ID-TIMS). Some interpretation is involved when inferring which generation of zircon in a migmatitic gneiss is metamorphic or magmatic. Age precision is a limitation of the microanalytical approach, and averaging of distinct age modes via single grain or multigrain analysis is potentially a problem for the ID-TIMS approach unless samples are well characterized. However, careful assessment of zircon morphology and ID-TIMS analysis yields highly precise ages (e.g., Miller et al., 2005). The two approaches have their most appropriate applications and relevance, and a thorough program of zircon geochronology would utilize both.

Dating cross-cutting plutons remains a proven approach for bracketing the time of deformational events in the Blue Ridge thrust complex (e.g., Spruce pine granites cross-cutting mylonitic Ashe metapelites: Trupe et al., 2003). Mineral isochrons (garnet-whole rock Nd-Sm, Rb-Sr, U-Pb) remain a powerful approach for dating metamorphism (Christensen et al., 1989; Vance and O'Nions, 1990; Vance and Holland, 1993), but inclusions of light REE accessory phases in garnet present a potential complication (DeWolf et al., 1996; Thöni, 2002). Step leaching of garnets is an effective means of dealing with inclusion problems.

Bulk multigrain analysis of hornblende or muscovite via ^{40}Ar-^{39}Ar in samples exhibiting multiple deformation phases will not be useful for dating peak metamorphism or obtaining thermochronologic constraints on cooling histories. This method has great utility if the specific details of the structural state, composition, and growth history of the relevant phases is well documented and presented for others to critically assess. Laser-heating-based microanalytical ^{40}Ar-^{39}Ar analysis of various textural types and generations of muscovite from Great Smoky Group schists may be necessary to assess the accuracy of ages obtained by step heating of bulk samples of coarse-grained muscovite (Connelly and Dallmeyer, 1993).

Finally, open discussion in forums such as the CGS Field Trips and Southeastern Section meetings of the Geological Society of America, and formation of collaborations among working groups from a variety of institutions are critical for reaching consensus on Blue Ridge tectonics.

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Tracking across the southern Appalachians

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ABSTRACT

Fission-track ages have been determined on apatite and zircon separated from rocks along a transect extending from eastern Tennessee through western North Carolina into northern South Carolina, and from the Great Smoky Mountains northeast to Grandfather Mountain. These ages are part of a larger fission-track study of the Paleozoic to Recent thermochronology of the central and southern Appalachians.

In the central highlands of the Great Smoky Mountains, zircon fission-track data from the upper plate of the Great Smoky fault indicate that the rocks underwent rapid cooling through the zircon closure temperature (~235°C) at ~280 Ma (early Permian). This cooling most likely was related to rapid uplift and denudation of upper plate rocks associated with emplacement of the Great Smoky and related thrust sheets during the late Paleozoic Alleghanian orogeny.

Most rocks yield significantly older zircon fission-track ages in all directions away from the central highlands. These rocks were evidently at sufficiently shallow burial depths (low temperatures) by the late Paleozoic that thrust-related cooling is not recorded in the zircon fission-track ages. Some of these rocks were at high enough temperatures to totally anneal fission tracks in zircon (>~240°C) earlier in the Paleozoic, but cooled through the zircon closure temperature prior to the Alleghanian. On the far northwestern side of the Blue Ridge and in the Valley and Ridge, however, rocks have never been buried to temperatures sufficient to totally anneal fission tracks in zircon; in some rocks, zircon may have undergone little if any postdepositional annealing. The pattern of zircon fission-track ages suggests that the present central highlands of the Great Smoky Mountains is a paleothermal dome, formed during or after emplacement of the Great Smoky thrust sheet.

Fission-track data indicate that the entire study region, with the possible exception of the far northwestern Blue Ridge and the Valley and Ridge, was still at temperatures >100°C in the early Mesozoic. Beginning no later than Late Triassic or Early Jurassic, apatite fission-track data indicate that the region has undergone uplift and denudation at rates that, on average, are very slow. Apatite fission-track data record similar slow regional uplift in the northern Blue Ridge in Virginia. The one known exception in this regional pattern is the Grandfather Mountain window.

In the Great Smoky Mountains, the apatite data suggest that continuing uplift has been accompanied by Late Cretaceous or younger offset (southeast side up) on one or more faults in the upper plate of the Great Smoky fault, most likely including the Gatlinburg fault system. In contrast, neither the apatite nor the zircon ages show any detectable change across the Brevard fault zone, suggesting there has been no significant vertical displacement on the Brevard fault since the late Paleozoic.

INTRODUCTION

This paper is a progress report on part of a larger fission-track study in the central and southern Appalachians. The purpose of the study is to use fission-track analysis to gain a better understanding of the low-temperature thermotectonic history of the region. So far, apatite and zircon fission-track ages have been determined on 194 samples collected in West Virginia, Maryland, District of Columbia, Virginia, Tennessee, North Carolina, and South Carolina. We will discuss data from 61 of these samples that form a broad transect extending from the Valley and Ridge in eastern Tennessee through the Blue Ridge of Tennessee and western North Carolina into the Piedmont of North Carolina and northern South Carolina, and from the Great Smoky Mountains northeast to the Grandfather Mountain area (Fig. 1). Sampled Blue Ridge rocks in the Great Smoky Mountains area (Southworth et al., 2005) are from the upper plate of the Great Smoky thrust fault, part of the Blue Ridge fault

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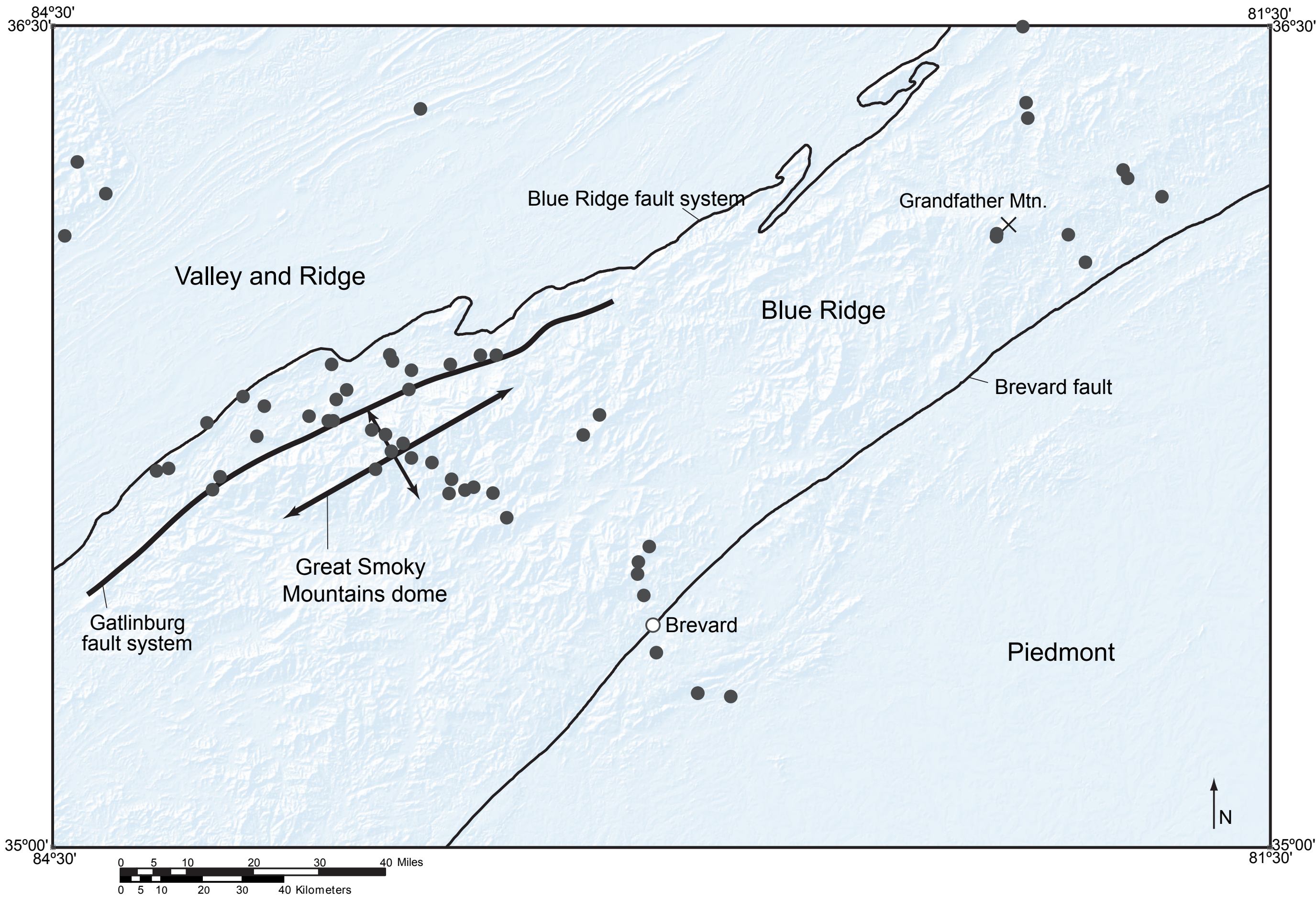

Figure 1. Location of 61 fission-track samples collected along a transect extending from the Valley and Ridge in eastern Tennessee southeast to the Piedmont of northern South Carolina, and from the Great Smoky Mountains northeast to the Grandfather Mountain area. The Blue Ridge fault system includes the Great Smoky fault in the southwestern part of the study area (Hatcher et al., 1989).

system, which carried Blue Ridge rocks over Valley and Ridge rocks during the late Paleozoic Alleghanian orogeny (Hatcher et al., 1989). To the northeast in the Grandfather Mountain area, samples were collected both in the Grandfather Mountain window and in the surrounding Blue Ridge thrust sheet (Bryant and Reed, 1970; Rankin et al., 1972).

FISSION-TRACK ANALYSIS

Fission tracks are zones of damage in a crystal that are caused by the passage of highly charged and energetic nuclei from the spontaneous fission of ^{238}U atoms. Fission-track analysis is particularly well suited to the study of the low-temperature history of rocks, below ~240°C. The number of tracks in a mineral increases predictably over time, and thus the age of a mineral can be calculated from the number of tracks and amount of uranium it contains, but fission tracks disappear at elevated temperatures through track annealing. Heating a mineral containing fission tracks allows ions displaced along the damage zone to move back into normal crystallographic positions in the mineral. This repair of the damage zone leads to shortening and ultimately to the total disappearance of the fission track, resulting in an anomalously young fission-track age.

Significant annealing occurs over a temperature range, which is referred to as the "partial annealing zone." If a mineral is heated to high enough temperatures for sufficiently long time, tracks are effectively totally annealed and the mineral yields a "zero" fission-track age. When the mineral cools, tracks again begin to accumulate, resetting the fission-track clock. Annealing temperatures in apatite are a function of both the duration of heating and the chemical composition of the apatite. In a relatively stable thermotectonic environment, apatites with compositions similar to the Durango fluorapatite age standard (including apatites analyzed in the present study) undergo significant annealing over an effective partial annealing zone that extends from ~60 to ~110°C (Fitzgerald et al., 1995). Modeling indicates that, for rocks cooling through the apatite partial annealing zone at rates comparable to those in the Blue Ridge and Piedmont, the effective apatite closure temperature (Dodson, 1979) in fluorapatite is ~90-100°C.

Annealing temperatures in zircon, which are a function of the effective duration of heating and the amount of α-radiation damage in the zircon crystal, are not as well known, but are significantly higher than apatite annealing temperatures. In natural α-radiation-damaged zircon, effective closure temperatures are probably in the range of ~210-260°C for cooling rates of 1 to 100°C/m.y., respectively (Brandon et al., 1998). For the present study, we have adopted a mean value of 235 ± 25°C for the zircon closure temperature, and 180 ± 25°C and 240 ± 25°C as the approximate low- and high-temperature boundaries of the effective partial annealing zone (based on Brandon et al., 1998, their Fig. 6).

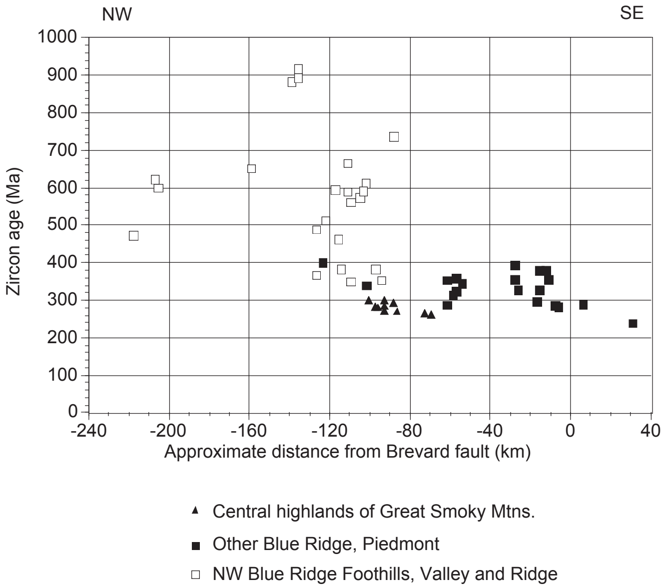

Figure 2. Zircon fission-track ages plotted as approximate distance northwest and southeast of the Brevard fault.

Apatite and zircon grains coarser than 74 μm (200 mesh; fine-sand size) were separated from crushed samples using standard heavy-mineral separation techniques (e.g., Naeser et al., 1989, their Fig. 10.7). Fission-track ages were determined using the external detector method (Naeser and McKee, 1970; Naeser et al., 1989), as follows. Apatite separates were mounted in epoxy, polished to expose internal grain surfaces, and etched in 7% nitric acid for 40 seconds at 22 to 23°C. Zircons were mounted in teflon, polished, and etched in a eutectic KOH-NaOH melt (Gleadow et al., 1976) for times between 23 hours and 64 hours at 207 to 220°C. Grain mounts were covered with low-uranium-content-muscovite external detectors and irradiated at the U.S. Geological Survey Reactor in Denver, Colorado. Grain mounts and external detectors were counted at 1250x magnification using a 100x oil immersion lens. Ages were calculated using the zeta calibration method (Hurford and Green, 1983), with zeta factors of 10,752 (apatite) and 319.6 (zircon). The central age (Galbraith and Laslett, 1993) is reported for samples that fail the χ^2 test ($P(\chi^2)<5\%$) (Galbraith, 1981; Green et al., 1989).

RESULTS

Fifty-three of the 61 samples yielded sufficient zircon for fission-track analysis. The zircon fission-track ages range from ~240 Ma in the northern South Carolina Piedmont to ~920 Ma in Blue Ridge rocks near the northwestern edge of the Great Smoky thrust sheet in eastern Tennessee (Fig. 2). High-elevation samples from the central highlands of the Great Smoky Mountains yielded statistically indistinguishable (at $\pm 2\sigma$) zircon ages that average ~280 Ma over a vertical relief of >1.25 km (from our highest sample, at 1,951 m above sea level) (Fig. 3). Rocks generally yield significantly older zircon fission-track ages in all directions away from the central highlands.

Fifty-one samples yielded apatite fission-track ages, ranging from less than 80 Ma in Proterozoic rocks in the Grandfather Mountain window to ~200 Ma in the lower Mississippian Grainger Formation just west of the Great Smoky fault along the base of Chilhowee Mountain (Fig. 4). Much of the scatter in apatite ages along the transect reflects the elevation dependence of the apatite ages (Fig. 5). In particular, ages from the central highlands of the Great Smoky Mountains and from other Blue Ridge and Piedmont rocks to the southeast and northeast define a trend of decreasing age with decreasing elevation, from ~185 Ma near the top of the section to ~85 Ma at the base. The exception to this trend is the Grandfather Mountain window where ages are statistically indistinguishable, at ~85 Ma, over a vertical relief of >1.1 km.

To the northwest, in the Foothills of the Blue Ridge and in the Valley and Ridge, the range of apatite fission-track ages is generally comparable to the range of ages to the southeast. With few exceptions, however, rocks of a given age occur at lower elevations in the Foothills and Valley and Ridge (Fig. 5).

DISCUSSION

The lack of change of zircon fission-track age with elevation in the central highlands of the Great Smoky Mountains (Fig. 3) indicates that rocks in this part of the Great Smoky thrust sheet were buried at temperatures high enough to totally anneal fission tracks in zircon prior to undergoing rapid cooling through the zircon fission-track closure temperature (~235°C) at ~280 Ma (early Permian). The time and required rate of cooling suggest

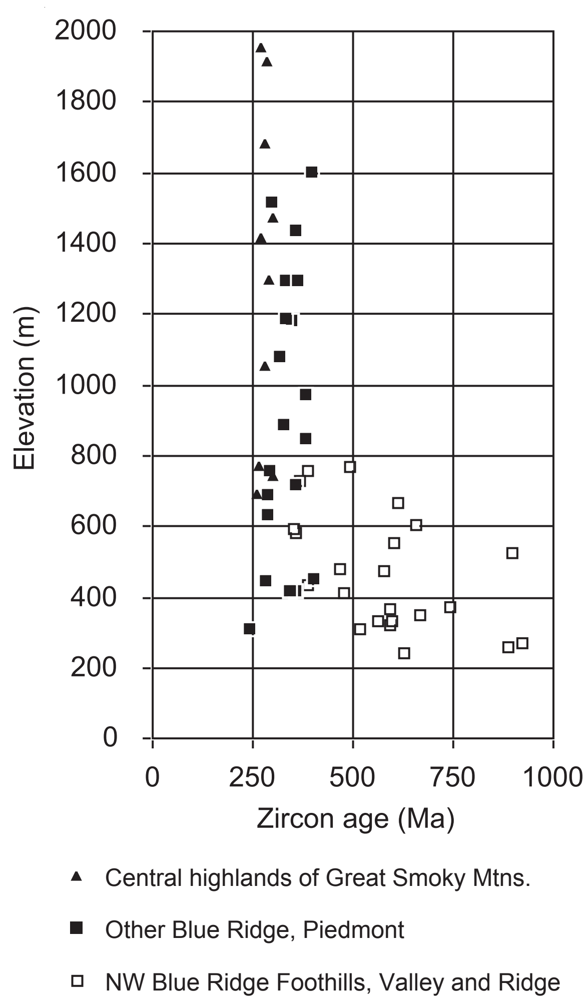

Figure 3. Zircon fission-track ages plotted against sample elevation.

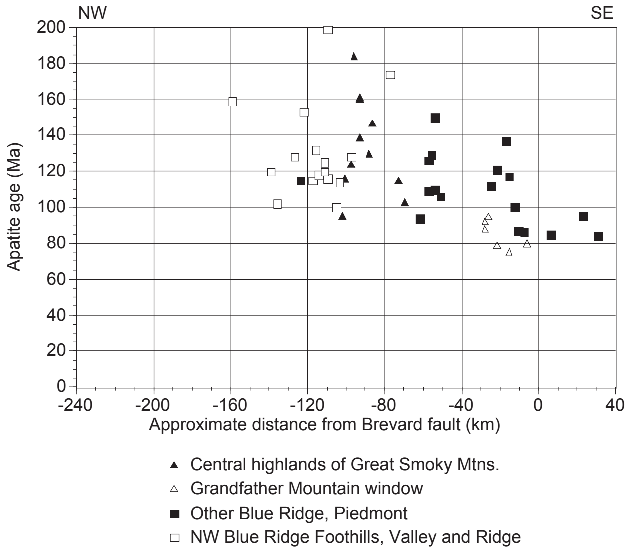

Figure 4. Apatite fission-track ages plotted as approximate distance northwest and southeast of the Brevard fault.

that the cooling was most likely related to rapid uplift and denudation of upper plate rocks associated with emplacement of the Great Smoky and related thrust sheets in the late Paleozoic Alleghanian orogeny (N.D. Naeser et al., 2004). In the northern Blue Ridge, zircon fission-track ages similarly record rapid cooling of Alleghanian thrust sheets, at ~295 Ma in the James River area in central Virginia, and at ~305 Ma in the Shenandoah National Park area of northern Virginia (N.D. Naeser et al., 2004).

Away from the central highlands in all directions—in rocks from the periphery of the highlands, the northwest Foothills, the Blue Ridge to the southeast, and the Grandfather Mountain area to the northeast—most rocks yield significantly older zircon fission-track ages (Naeser et al., 1999; C.W. Naeser et al., 2004) (Figs. 2, 3). These rocks were evidently at sufficiently shallow burial depths (low temperatures) by the time of Alleghanian thrusting that thrust-related cooling is not recorded in the zircon ages. Some of these rocks were heated to temperatures high enough to totally anneal fission tracks in zircon (>~240°C) in the Paleozoic, but cooled through the zircon closure temperature prior to the Alleghanian, probably reflecting regional cooling from pre-Alleghanian Paleozoic metamorphism. In contrast, zircon fission-track ages near the northwestern edge of the Great Smoky thrust sheet in the Chilhowee Mountain-Little Tennessee River area, and in the Valley and Ridge, are all equal to or older than the depositional age of the rocks. These rocks, which are not metamorphosed, have never been buried to temperatures high enough to totally anneal zircon. Some may contain zircon that has undergone little, if any, postdepositional annealing, restricting maximum burial temperatures to <~180°C. The contrast between zircon ages in the central highlands and surrounding rocks suggests that the central highlands of the Great Smoky Mountains is a paleothermal dome, formed during or after emplacement of the Great Smoky thrust sheet.

Apatite fission-track ages record the lower temperature, Mesozoic-Cenozoic history of the region. Preliminary modeling suggests that the rocks share a similar history of relatively continuous cooling, related to uplift and erosion, beginning no later than Late Triassic to Early Jurassic and continuing to the present at rates that, on average, are very slow (<2°C/m.y.). Average effective denudation rates for the time when the rocks were cooling through the apatite closure temperature (~90-100°C in these rocks) are on the order of 20 m/m.y. (Fig. 5), very similar to rates observed in the northern Blue Ridge in central and northern Virginia (N.D. Naeser et al., 2004), and to average denudation rates previously estimated for the Appalachians (summarized in Matmon et al., 2003a, 2003b).

The anomalous trend of apatite ages in the Grandfather Mountain window may indicate late tilting or late doming of the Grandfather Mountain window. Fission-track analysis of new samples collected on the northwest flank of Grandfather Mountain should resolve the Late Cretaceous and younger tectonism in this area.

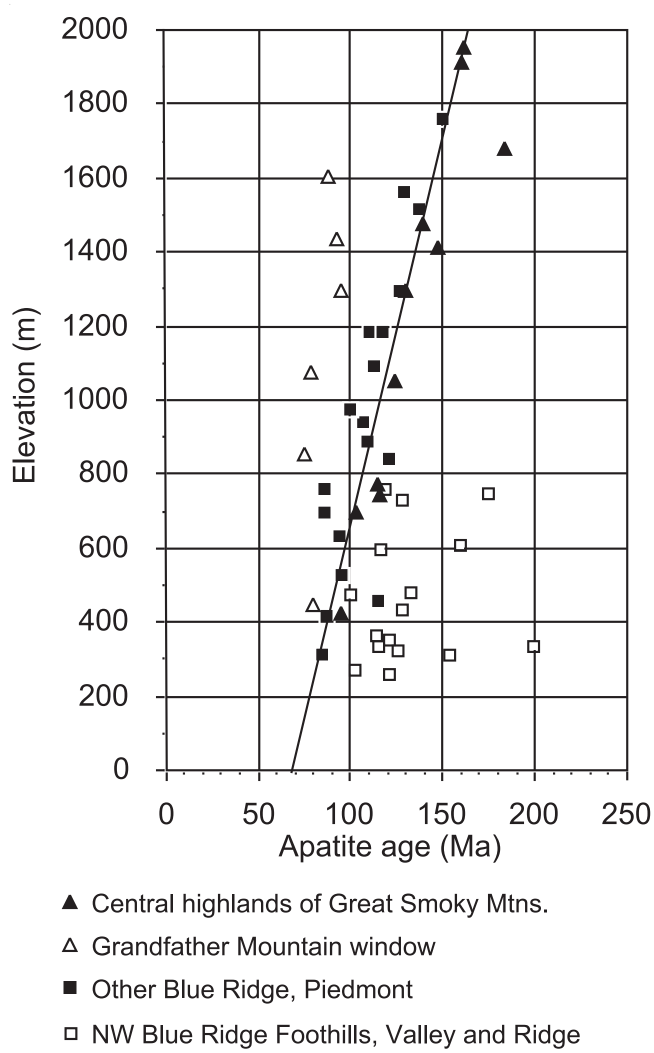

Figure 5. Apatite fission-track ages plotted against sample elevation. The best-fit line through data from the Blue Ridge (excluding ages from the Grandfather Mountain window and northwest Foothills) and Piedmont indicates an average effective denudation rate of ~20 m/m.y. during the time when the rocks were cooling through the apatite fission-track closure temperature (~90-100°C in these rocks).

The elevation offset in apatite ages between the central highlands and Foothills in the Great Smoky Mountains (Fig. 5) suggests that continuing Mesozoic to Cenozoic uplift has been accompanied by Late Cretaceous or younger movement (southeast side up) on one or more faults in the upper plate of the Great Smoky thrust fault, most likely including the Gatlinburg fault system (Naeser et al., 1999). In contrast, in the area where we have samples on both sides of the Brevard fault zone, near the town of Brevard, there is no detectable difference in either zircon or apatite ages across the fault (Figs. 1, 2, 4), suggesting there has been no significant vertical displacement on the Brevard fault since the late Paleozoic. Consistent with this interpretation, the Brevard fault is cut by Jurassic diabase dikes in northeastern Georgia (R.D. Hatcher, Jr., written communication, 2005).

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Petrotectonics of mafic and ultramafic rocks in Blue Ridge terranes of western North Carolina and northern Georgia

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INTRODUCTION

The purpose of this paper is to summarize the geochemical, petrologic and mineralogic data on the mafic and ultramafic rocks in the eastern Blue Ridge province of the southern Appalachian orogen and to relate these rocks to the various terranes postulated for this area. The recent use of high resolution ion microprobe geochronology (Carrigan et al., 2003; Bream et al., 2004; Hatcher et al., 2004; Ownby et al., 2004), coupled with detailed geologic mapping has resulted in an evolving concept of terranes in the this area (Hatcher et al., 1999, 2004, this volume, Hatcher, 2001).

Tectonostratigraphic terranes are recognized based on their faulted boundaries, crustal affinity, age, and tectonic history. For the purposes of this paper, we adopt the terminology and distribution of these terranes as depicted in this guidebook (Hatcher, this volume), but recognize that delineation of terranes is an evolving process and that terrane boundaries are subject to reinterpretation based on new data. With the goal of using ultramafic rocks to help unravel the character and history of these packages of rock, we present a review of the character of the ultramafic and associated mafic rocks.

ULTRAMAFIC ROCKS AND TECTONICS

Alpine ultramafic rocks provide critical information on the tectonic history of orogenic belts. In particular, ophiolites mark suture zones and their petrology can provide clues to details of tectonic settings. At the onset of the plate tectonics revolution, ophiolites were viewed as bits of exhumed oceanic crust, developed at spreading centers, that were incorporated into mountain belts during the closing of ocean basins (Hess, 1955; Moores, 1970). Not long thereafter, an origin associated with accretion of island arcs to continental margins was advocated (Miyashiro, 1973). Now, formation of ophiolites along convergent plate boundaries in a back-arc, suprasubduction zone setting, as well as in response to other subduction-related processes, is recognized as important (Pearce et al., 1984; Pearce, 2003). Thus, even within the same mountain belt, ophiolites may have different origins and may be quite different in terms of their origin and emplacement tectonics (Dilek et al., 2000; Dilek, 2003; Nicolas and Boudier, 2003;Wakabayashi and Dilek, 2003).

Attempts to understand what ophiolites represent in terms of tectonic setting and to then classify the ophiolites in

Swanson, S. E., Raymond, L. A., Warner, R. D., Ryan, J. G., Yurkovich, S. P., Peterson, V. L., Petrotectonics of mafic and ultramafic rocks in Blue Ridge terranes of western North Carolina and northern Georgia, *in* Hatcher, R. D., Jr., and Merschat, A. J., eds., Blue Ridge Geology Geotraverse East of the Great Smoky Mountains National Park, Western North Carolina: North Carolina Geological Survey, Carolina Geological Society Annual Field Trip Guidebook, p.73-90.

a genetically significant way have met with limited success. Miyashiro (1975) attempted to use geochemistry of mafic rocks as a prime indicator of tectonic setting and divided ophiolites into three types based on that chemistry. Moores (1982) broadened the description and definition of ophiolites to include associated rocks and recognized two categories of ophiolite based on emplacement tectonics: obducted ophiolites emplaced on recently passive crystalline continental margins (Tethyan type) and accretionary complex ophiolites emplaced at active margins (Cordilleran type). Ishiwatari (1985) divided ophiolites on the basis of the petrology of ophiolites into three types:the Liguria type with alkali basalt and lherzolites; the Yakuno type with high-alumina tholeiite and clinopyroxene (cpx) harzburgites; and Papua type with low-alumina tholeiite and cpx-free harzburgites. Simultaneously, Boudier and Nicolas (1985) divided ophiolites, on the basis of petrology, petrogenesis, and structure, into lherzolite ophiolite subtypes (LOT) and harzburgite ophiolite subtypes (HOT). Later, they added an intermediate type of ophiolite, the lherzolite-harzburgite ophiolite type (LHOT), with characteristics intermediate between those of the two end members (Nicolas and Boudier, 2003). Ishiwatari et al. (2003) adopted the lherzolite vs. harzburgite distinction and nomenclature, with the harzburgite type including normal and depleted harzburgite types. Hence, they, too recognize three types of ophiolites. Seven classes of ophiolite were recognized by Dilek (2003) on the basis of petrologic, structural, petrotectonic, and genetic characters. Although complete descriptions and exclusive sets of characters are not provided, his subdivision emphasized the complexity and variability of ophiolite formation. Attempts at classification during recent years, conducted against the backdrop of the mid-ocean ridge vs. suprasubduction zone petrogenetic controversy, in some cases, have drawn on knowledge of the structure of fast- and slow-spreading mid-ocean ridge crust formation processes and resulting structure in an attempt to link the physical-chemical features of the ophiolites to fast, normal, and slow spreading processes of crustal genesis. Because of the variations in mantle chemistry and the varying roles of water and temperature in controlling melt chemistry and rock petrogenesis, these attempts have not been altogether successful (Nicolas and Boudier, 2003). Nevertheless, the division of ophiolites into LOT bodies with a petrochemical relationship with enriched mantle rocks, LHOT bodies with a petrochemical relationship with partially depleted mantle rocks, and HOT bodies with a petrochemical relationship with depleted mantle rocks provides a sound petrochemical foundation to this classification scheme.

In an attempt to use observable features (chemistry, texture, and structure) rather than inferred genesis, we here adopt a three-fold classification of ophiolites (Table 1) modeled on those of Nicolas and Boudier (2003) and Ishiwatari et al. (2003). Features listed in Table 1 apply to all parts of the ophiolite section, from overlying sedimentary rocks to basaltic/gabbroic rocks , to cumulate ultramafic rocks and depleted mantle.

Rarely in orogenic belts is a complete ophiolite section preserved, and this is the case in the Blue Ridge. Typically, Blue Ridge ultramafic rocks occur as isolated pods surrounded by metasedimentary rocks, prompting some workers to refer to the rock bodies with isolated masses of mafic and/or ultramafic rocks as melanges (Abbott and Raymond, 1984; Lacazette and Rast, 1989; Adams and Trupe, 1997). Other workers have queried whether or not these rocks are broken or deformed ophiolites (McSween and Hatcher, 1985). Mafic rocks are sometimes closely associated with the ultramafic rocks, but a direct link between mafic and ultramafic rocks is rarely apparent. Modification of primary textural and/or structural features and the modification of protolith chemistry during regional metamorphism of the Blue Ridge assemblage (Swanson, 2001; Raymond et al., 2003) present challenges to reconstructing original igneous protoliths. Given these problems, it is easy to see why the petrotectonic history of the Blue Ridge ultramafic rocks remains unresolved. In this paper we pull together existing data (published and unpublished) on the Blue Ridge ultramafic rocks and associated mafic rocks and attempt to interpret this information in light of the ophiolite model outlined in Table 1.

DISTRIBUTION OF MAFIC/ULTRAMAFIC ROCKS IN THE GEORGIA AND NORTH CAROLINA BLUE RIDGE

Building on the earlier work of Pratt and Lewis (1905), Hess (1955) recognized the occurrence of two belts of ultramafic rocks in the Appalachians. Later work by Larrabee (1966) refined the distribution of these belts, and Misra and Keller (1978) summarized the considerable literature on the ultramafic rocks available at the time. In the Blue Ridge belt, ultramafic rocks are enclosed in moderate to high grade schists and gneisses and are, themselves, metamorphic rocks (Raymond et al., 2001; Swanson, 2001; Warner, 2001; Raymond et al., 2003). Enclosing mafic rocks are typically amphibolite (hornblende schist and gneiss). The ultramafic rocks occur as granoblastic metadunites and metaperidotites or as more hydrated ultramafic rocks such as serpentinites and amphibole-talc-chlorite schists. The rocks represent a range of metamorphic conditions, from high to low grades reflecting recrystallization following peak metamorphism at granulite or eclogite facies conditions. For our purposes, we will not dwell on the metamorphic character of these rocks. Instead we will try to unravel the protolioth compositions of the mafic and ultramafic metamorphic rocks and relate this original character to the geologic history of the host terranes. Our study is focused on Hess's western belt of ultramafic rocks, cropping out mostly in the eastern Blue (Fig. 1).

One of the peculiarities of the western belt of ultramafic rock bodies of Hess (1955) is that the belt crosses apparent terrane boundaries. In southern North Carolina, the belt, in fact, seems to bifurcate into a western and an eastern part. The eastern part occurs entirely within the Neoproterozoic (?) eastern Blue Ridge of Hatcher (1978), whereas the western part occurs in a belt earlier assigned to the eastern Blue Ridge, but now recognized as belonging to the Mesoproterozoic Mars Hill terrane (Bartholomew and Lewis, 1988; Hatcher, 2002; Carrigan et al., 2003).

Methods

Individual bodies and complexes of mafic and ultramafic rocks in the various terranes of the southern Blue Ridge were selected for discussion based on the availability of data either in the literature or in our files. The approach is to use mineralogy, petrology, geochemistry, rock assemblages, and structures as a

Table 1. Characteristics of ophiolite types.

Feature	Harzburgite Ophiolite Type (HOT)	Lherzolite/Harzburgite Ophiolite Type (LHOT)	Lherzolite Ophiolite Type (LOT)
crustal thickness	4-6 km	2-3 km	0-1 km
basalt chemistry	lo-Al tholeiite	hi-Al tholeiite	alkali basalt
dikes & sills in ophiolite	uncommon	X	X
sheeted dikes	X	in some bodies	generally absent
plutonic complex thickness	thick	thin to absent	generally absent
quartz diorite "plagiogranite"	X	X	X
gabbro types			
troctolite			X
ol-gabbro	X	X	uncommon
gabbronorite	X	dominate	uncommon
ferrogabbro	uncommon	X	uncommon
layered gabbros	X	X	thin
foliated gabbros	X	poor foliation	X
penetrative fabric	magmatic	magmatic to plastic	?
ultramafic rock types			
dunite	X	X	local dunite lenses
peridotite			
wehrlite	locally abundant	X	rare
lherzolite	tectonized	local, plag or spinel	plag-spinel
harzburgite	X	with cpx	X
pyroxenite			
clinopyroxenite	X	X	dikes
websterite	with ol	locally present	X
orthopyroxenite	X	rare	
chromitite		abundant	uncommon to absent
inferred setting & spreading rate	slow	normal	very slow

Source: modified from Nicolas and Boudier (2003)
X = present, blank indicates absent

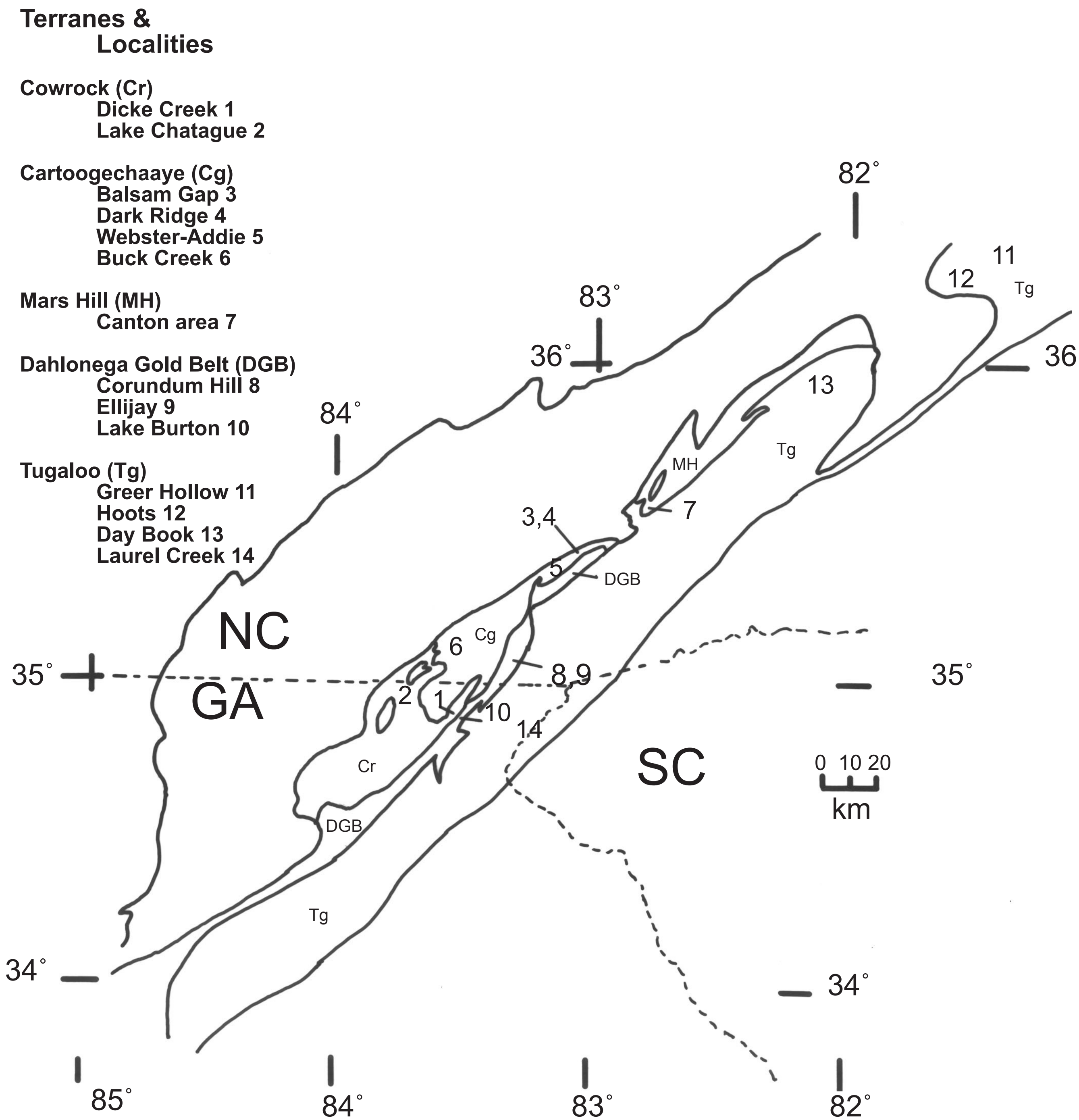


Figure 1. Locations of rocks in terranes of the Blue Ridge (see Hatcher et al., this guidebook).

guide to the tectonic setting of the magmatism that produced these rocks. This approach emphasizes the original formation of the mafic and ultramafic rocks, not post-magmatic emplacement of these mafic/ultramafic rocks into their present positions.

Much of the mineralogy in Blue Ridge ultramafic rocks has been affected by metamorphic recrystallization and metasomatic exchange, in particular hydration. Dunite and peridotite bulk compositions form olivine- and pyroxene-rich rocks at higher metamorphic grades, while basalt/gabbro compositions form amphibolites. The widespread outcrops of olivine-rich rocks in the Blue Ridge were interpreted by many geologists as primary igneous rocks, but we now recognize their metamorphic character. Most of the minerals, including olivine and pyroxene, have undergone recrystallization and metasomatic exchange, and their compositions thus reflect metamorphic conditions (e.g, Tenthorey et al., 1996; Swanson, 2001; Warner, 2001). The lack of zoning in pyroxenes and plagioclases reflects metamorphic homogenization. Some minerals show the effects of multiple periods of recrystallization. Amphibole sometimes shows overgrowths of cummingtonite on tremolite (Swanson, 2001) or tremolite on edenite (Warner, 2001), reflecting incomplete recrystallization during metamorphism. Symplectic textures developed between olivine and plagioclase in the metatroctolite at Buck Creek record recrystallization during high-grade metamorphism (Tenthorey et al., 1996; Lang et al., 2004). Compositions of the metamorphic minerals are not generally useful in characterizing igneous processes, although the bulk compositions of such rocks can be useful protolith indicators (see Ryan et al., this volume).

Chromian spinel (chromite and related CrFe oxides) shows a wide range of compositions in Blue Ridge ultramafic rocks (Lipin, 1984; Warner, 2001; Raymond et al., 2003; Swanson and Raymond, in preparation). Recrystallization of chromite at moderate metamorphic grades produces chromian clinochlore and a less aluminous spinel. Small amounts of chlorite mantle zoned chromite grains, while a thicker chlorite corona forms about more-recrystallized homogeneous chromite grains. Intergowths of chlorite in spinel along the {111} planes, called "lattice texture" by Lipin (1984), are typical of the latter. This texture probably formed by the exsolution of some Al-rich phase from chromite and the subsequent alteration of that phase to chlorite. At lower metamorphic grades, recrystallization of chromite produces a more Fe-rich spinel, initially as chromian magnetite (the so-called "ferritchromite" alteration of many workers) and finally magnetite. The least altered chromites may represent primary igneous compositions (Swanson and Raymond, 2004, in preparation), but identification of relict spinels requires careful work. In this paper we will use the composition of the least recrystallized chromite (typically Al-rich cores of zoned crystals or grains included in olivine) as a guide to igneous petrogenesis.

Mafic/ultramafic rock bulk compositions are readily recrystallized to a variety of mineral assemblages during metamorphism, especially in the presence of a fluid phase. However, the major element bulk compositions of these rocks often remain unchanged, except for the hydration, and reflect their igneous heritage. Mafic and ultramafic rocks recovered from the sea floor by drilling or dredging often show the effects of hydrothermal alteration (Manning and MacLeod, 1996; Fletcher et al., 1997), but the major element chemistry of these rocks is little altered. During incorporation into orogenic belts, regional metamorphism converts basalts and gabbros to amphibolite and more foliated varieties of metamorphic rock. Introduction of the smallest amount of fluid to the metaultramafic rocks can produce a variety of hydrated minerals including serpentine, amphibole, chlorite, and talc. It is common in ultramafic rocks of the eastern Blue Ridge to find a variety of rocks (metadunite, chlorite-talc schist, serpentinite, etc.) in the same small body, reflecting varying degrees of recrystallization related to interaction with fluids at a range of P-T conditions during cooling from peak metamorphic conditions. Metasomatic alteration, either along contacts between mafic/ultramafic rocks and more Si-rich rocks, or by interaction with fluids, can alter original igneous compositions, especially the abundances of fluid-sensitive species like the alkali metals. We attempt to use the olivine/pyroxene mineralogy of the metaultramafic rocks as a guide to the original protoliths and avoid hydrated lithologies, in which metasomatic alteration may be a problem. For the mafic rocks, we use bulk compositional data (Al-Fe-Mg after Pearce et al., 1977) of amphibolites to try and characterize their tectonic setting. The tectonic fields delineated by Pearce et al. (1977) were based on compositions of basaltic rocks representing magma compositions. Amphibolites derived from gabbros, especially from cumulate gabbros, do not represent magmatic compositions. Care must be exercised in identifying amphibolites formed via a cumulate process. Trace elements offer one way to distinguish cumulate and magmatic amphibolites (Berger et al., 2001; Peterson et al., 2004, Ryan et al., this guidebook), but trace element analyses are not available for all of the amphibolites. We use the major element diagrams of Pearce et al.(1977) because of the paucity of trace element and/or isotopic data for amphibolites in many areas of interest, realizing that some degree of scatter in these data may be related to nonmagmatic (cumulate) amphibolite compositions.

Few clearly primary structural features remain in the mafic/ultramafic rocks of the Blue Ridge. Emplacement of mafic/ultramafic rocks in the Blue Ridge predates the Ordovician Taconic orogeny and associated amphibolite to granulite grade regional metamorphism. Primary textural and structural features associated with these rocks were mostly removed during this and subsequent orogenic events. Compositional layering is locally preserved in some ultramafic rocks. Chromitite and pyroxenite layers preserved in metadunite and mineral layering in metagabbro may reflect relict igneous layering. More commonly, lenses and pods of pyroxenite and chromitite (disrupted layers or dikes?) are enclosed in metadunite. Most of the larger Blue Ridge ultramafic bodies are associated with mafic rocks, but the smaller bodeis are not and this may in part be a result of deformation and disaggregation during emplacement.

ULTRAMAFIC ROCKS AND TERRANES

Ultramafic rocks occur in several terranes in northeastern Georgia and western North Carolina (Fig. 1). Nomenclature and the geology of these terranes is described elsewhere in this guidebook, but some salient details are summarized here.

Cowrock Terrane

The Cowrock terrane is relatively small and contains only a few small ultramafic bodies. In the area of Lake Chatuge (GA/NC border), rocks of the Cowrock terrane frame (form the hanging wall of) two windows in the Hayesville thrust sheet and this zone is host to the rocks of the Lake Chatuge complex (Hartley, 1973). Elsewhere, minor mafic and ultramafic rocks occur among the dominant rocks of the terrane.

Amphibolite in the Lake Chatuge area encloses various metagabbros and apparently formed by the hydration of gabbroic protoliths (Hartley, 1973). Hartley identified various gabbroic rocks including troctolite and olivine gabbro, some with relict igneous layering, in the Lake Chatuge complex. Peterson et al. (2004) reported the amphibolites from the Lake Chatuge complex have compositions consistent with "oceanic gabbro/cumulate protoliths," and Berger et al. (2001) reported these rocks are light rare-earth element depleted, with ocean ridge-like signatures.

Dicks Creek Suite. The Dicks Creek suite includes two small bodies of ultramafic rock and amphibolite along Dicks Creek, west of Clayton, GA (Vincent et al., 1990). Nelson (1983) mapped these ultramafic bodies as inclusions along splays of the Soque Fault. Mapping by Hatcher and his colleagues (map included in Hopson et al, 1989) however, shows these bodies as inclusions in biotite gneiss and schist. One of the ultramafic bodies is serpentinite. The other body is a composite mass of amphibolite, serpentinite, and olivine clinopyroxenite (Vincent et al., 1990). The olivine clinopyroxenite is composed of olivine and augite along with anthophyllite, talc, and serpentine. A body of amphibolite occurs between the two ultramafic bodies. The ellipsoidal bodies of mafic and ultramafic rock are concordant with the foliation of the enclosing rocks.

The lack of chromitite and dunite in the Dicks Creek suite (Table 2) along with the small size of the bodies favors an assignment to the LOT type of Nicolas and Boudier (2003). The data, however, are limited.

Lake Chatuge Complex. The Lake Chatauge complex is a sheet of mafic and ultramafic rocks framing two windows through the Hayesville thrust sheet along the Georgia/North Carolina boundary. Outcrops of mafic and/or ultramafic rocks are continuous around the Brasstown Bald window and discontinuous around the Shooting Creek window. Metagabbros and amphibolites are the most common rock types, but some metadunite also occurs in the complex (Hartley, 1973). The petrology is similar to the Buck Creek body, with lenses of corona-bearing olivine troctolite enclosed in dunite. Amphibolites within the complex represent metasomatically altered gabbros (Hartley, 1973). Chlorite and talc-chlorite schists occur as isolated bodies and along the margins of some amphibolites. Corundum, associated with the chlorite schists, was mined in the late 1800s. Foliations within the complex are conformable with surrounding biotite gneiss and Hartley (1973) suggested that the complex may represent a differentiated sill.

Protoliths for the metagabbros are olivine gabbro and troctolite (Meen, 1988). Meen (1988) inferred that the amphibolites represent more hydrated metagabbros of various lithologies. Ultramafic rock protoliths include dunite and plagioclase-bearing olivine clinopyroxenite. Serpentinization of olivine is common and more widespread in the ultramafic rocks than in the metagabbros. Olivine and clinopyroxene in the metagabbros typically have metamorphic coronas of pyroxene and spinel. Metamorphic amphibole is also widespread within the metagabbros. The presence of plagioclase-bearing olivine clinopyroxenite reported by Hartley (1973) is unique in the Blue Ridge ultramafic rocks. Some pods of garnet clinopyroxene granulite represent recrystallization of clinopyroxenite or gabbro at high pressure (Dallmeyer, 1974; Meen, 1988).

REE patterns and abundances in the garnet-clinopyroxene granulite are similar to those of amphibolites of the Buck Creek complex (Berger et al., 2001) suggesting a similar magmatic origin for both. Meen (1988) suggested that fractionation of an alkali basalt magma produced the Lake Chatuge lithologies. The Lake Chatuge pyroxene granulites and most Buck Creek amphibolites are LREE-depleted (Berger et al., 2001) reflecting a depleted mantle source. Shaw and Wasserburg (1984) reported Nd isotopic data for both Lake Chatuge and Buck Creek mafic rocks consistent with an LREE-depleted source region.

The abundance of olivine gabbro (and troctolite), lack of chromitite, presence of dunite (Table 2), a somewhat depleted source area for the magmas, and a moderate size for the complex are most characteristic of the HOT type ultramafic body of Nicolas and Boudier (2003).

Amphibolite in the Lake Chatuge area encloses metagabbros and apparently formed by the hydration of gabbroic protoliths (Hartley, 1973). Hartley identified various gabbroic rocks in the complex, including troctolite and olivine gabbro, some showing relict igneous layering. Peterson et al. (2004) report the amphibolites from the Lake Chatuge complex have compositions consistent with "oceanic gabbro/cumulate protoliths." No data are available on chromite compositions.

Cartoogechaye Terrane

The Cartoogechaye terrane is an assemblage of high grade (upper amphibolite to granulite facies) gneisses and schists bounded on the west by the Hayesville fault. The Cowrock terrane lies in the Hayesville thrust sheet beneath and immediately south and the Mars Hill terrane is northeast of the Cartoogechaye terrane (Hatcher and others, this guidebook). Cartoogechaye terrane rocks are metasedimentary schists and gneisses interlayered with hornblende granulites and amphibolites (Absher and McSween, 1985; Eckert et al.,1989). The Cartoogechaye terrane is host to a number of large mafic/ultramafic complexes.

Balsam Gap. The Balsam Gap metadunite body is enclosed in gneiss (metagraywackes?) of the Tallulah Falls Formation of the Cartoogechaye terrane near the community of Balsam in western North Carolina. The body is dominated by metadunite exhibiting varying degrees of seprentinization (Honeycutt and Heimlich, 1980), and was mined for olivine. A few stringers (single grain thick) of chromite are found in the olivine-rich fabric. Rare pods of light-colored hornblende diorite and orthopyroxenite are found in the metadunite. Locally hydrated lithologies include tremolite-chlorite-talc schist and serpentinite.

The ultramafic body is included as an elongate, pod-like mass in the enclosing gneiss. The long axis of the ultramafic body is parallel to foliations in the enclosing gneisses (Honeycutt and Heimlich, 1980). Petrofabric studies of the metadunite reveal lattice preferred orientation (LPO) fabrics, but do not show any relationships between olivine petrofabrics and the shape of the body or the foliations of the enclosing gneisses (Astwood et al., 1972; Honeycutt and Heimlich,1980).

These characterisitics have elements of both the HOT and LHOT classes of ophiolite (Table 2), but the abundance of dunite is not consistent with a LOT body. The lack of chromitite and presence of orthopyroxenite favors a HOT classification for the Balsam Gap dunite, but the presence of the diorite argues for a HLOT assignment (Nicolas and Boudier, 2003). Compositions of the associated amphibolites (discussed below) are consistent with either a LHOT or HOT assignment for Balsam Gap.

Dark Ridge. The Dark Ridge ultramafic body, which measures approximately 310 m by 460 m (Schiering, 1979), is located about 2 km southwest of Balsam, and about 6 km east of Addie, NC. It is enclosed in high grade rocks (chiefly biotite gneiss) of the Cartoogechaye terrane. A thin layer of amphibolite occurs along one margin. The body is composed of both metadunite, serpentinized in part, and metaharzburgite, much of which has been steatitized (Warner and Helper, this guidebook). In places exceptionally coarse-grained olivine (up to 18 cm crystals) is present (Hunter, 1941). Hunter (1941) considered Dark Ridge to be one of the most outstanding olivine deposits in the western North Carolina olivine belt. Near Dark Ridge Creek, vein-like lenses of chromite up to 0.6 m in thickness occur near Dark Ridge Creek, from which chromite was mined in the early 1900s (Hunter et al., 1942). Possibly these chromitite layers represent primary igneous layering. A thin rind of chlorite-talc

Table 2. Characteristics of ophiolite types (from Table 1) compared to localities discussed in this paper.

Feature	HOT	LHOT	LOT	Cowrock Terrane		Cartoogechaye Terrane				Mars Hill Terrane	Dahlonega Gold Belt Terrane			Tugaloo Terrane			
				DC	LC	BG	DR	WA	BC	Canton	CH	EJ	LB	GH	H	DB	LC
associated amphibolites	X	X	X		X				X						X		X
adjacent amphibolite				X		X	X	X		X	X	X	X	X		X	
amphibolite chemistry	lo-Al tholeiite	hi-Al tholeiite	alkali basalt		oceanic	oceanic, subduction-related			oceanic island	subduction - related			low-K tholeiites	MORB			oceanic
gabbro types																	
troctolite			X		X				X								
ol-gabbro	X	X	X														
gabbroniorite	X	dominant	rare						X								
ferrogabbro	rare	X	rare														
um rock types																	
dunite	X	X	local lenses		X	X	X	X	X	X	X	X	X		X	X	X
peridotite																	
wehrlite	locally abundant	locally present	absent														
lherzolite		local plag or sp	plag, sp														
harzburgite	X	with cpx					X	X					X	X	X	X	
pyroxenite																	
cpxite				X	X				X								
websterite	lenses	local						X									
opxite						X			X			X			X		X
chromitite	rare	abundant	rare to absent	none	none	none	X	X	none	none	X	none	none			X	X
wt % Al_2O_3 in chromite cores				no data	no data	4 - 15	1 - 32	1 - 53	27 - 47	no data	3 - 8	16	6 - 17	no crsp	no data	1 - 16	4 - 11

schist separates the ultramafic rocks from surrounding gneiss and amphibolite.

A petrofabric study (Astwood et al., 1972) showed no preferred orientation of olivine in the Dark Ridge body. However, strain features (undulatory extinction, strain banding) preserved in relict olivine porphyroclasts led Dribus et al. (1976) to infer a diapiric emplacement origin. Post-emplacement recrystallization of olivine has produced extensive domains with granoblastic polygonal texture. Hydration coupled with upper amphibolite/lower granulite facies metamorphism gave rise to abundant talc and other secondary hydrous minerals (Warner and Helper, this guidebook). Additional hydration accompanying retrograde greenschist facies metamorphism resulted in formation of serpentine.

The presence of chromitite is consistent with a LHOT class of ophiolites, but the abundance of harzburgite characterisitics is better described by the HOT class (Table 2). The same package of amphibolite and gneiss is associated with both the Balsam Gap and Dark Ridge bodies.

Webster-Addie. The Webster-Addie complex is a sheet of discontinuous ultramafic rock domed and eroded to yield an outcrop ring diameter of almost 10 km. The rocks extend from Addie in the north to Webster in the south. Ultramafic rocks are conformable with the enclosing Cartoogechaye terrane gneisses and amphibolites and consist mainly of metadunite, with lenses of metawebsterite, metaorthopyroxenite, metaclinopyroxenite, and metaharzburgite enclosed in the olivine-rich rocks (Madison, 1968; Pike, 1968; Cronin, 1983; Swanson and Warner, 2001; McIlmoil et al., 2002; Ryan et al., this guidebook). A large lens of metawebsterite (the type locality for the rock *websterite*) is enclosed in metadunite just south of Webster. Small pods and stringers of chromite occur within the metadunite (Hunter et al., 1942). Some light-colored bodies of hornblende diorite are also enclosed in the metadunite. The pod-like character of these rocks is probably the result of disruption of original layer or dikes during deformation. Dikes of granitoid pegmatite intrude the ultramafic rocks and produce an extensive metasomatic reaction zone rich in anthophyllite, chlorite, talc, and vermiculite (Ryan et al., this guidebook). Rocks of the Webster-Addie complex were mined for olivine near the Addie community and near Chestnut Gap, to the south of Addie. Ryan et al. (this guidebook) provide detailed field and petrologic descriptions of Webster-Addie rocks near Addie.

Abundances of REE in pyroxenites from Webster-Addie are low and exhibit a slightly "U"-shaped REE pattern, as do pyroxenites from Balsam Gap and the Moores Knob dunite in the Ellijay area (see below) distribution patterns (Berger et al., 2001) reflecting a residual mantle source. Shaw and Wasserburg (1984) reported Nd isotopic data consistent with a slightly enriched source region and Berger et al. (2001), Soraruf et al.(2002b), and Ryan et al.(this guidebook) suggested possible metasomatic enrichment of LREE.

Protoliths of isolated amphibolite bodies near the Webster-

Addie complex are inferred to have a subduction-related origin (Peterson et al., 2004; Ryan et al, this guidebook), based on elevated LREE and silica contents. Similar amphibolite bodies occur to the north near the Balsam Gap and Dark Ridge ultramafic bodies.

These characteristics favor a LHOT classification for the Webster-Addie complex (Table 2), and the abundance of dunite is not consistent with LOT-type bodies. The presence of chromitite and websterite clearly favors a LHOT assignment. The presence of lenses of ortho- and clinopyroxenite (Swanson and Warner, 2001; McIlmoil et al., 2002) in the Webster-Addie dunites is consistent with either a HOT or LHOT assignment. Compositions of the associated amphibolite are also consistent with a LHOT assignment for the Webster-Addie bodies.

Buck Creek. The Buck Creek complex (termed the Chunky Gal Mountain complex by McElhaney and McSween, 1983) in Clay County, North Carolina is another large mafic/ultramafic complex in the Cartoogechaye terrane. The complex contains the largest body of metadunite in the Blue Ridge along with a much smaller body of metadunite, all surrounded by amphibolites and subordinate mica gneiss. Lenses of metatroctolite are enclosed in the large body of metadunite (Hadley, 1949; Kuntz, 1964; McElhaney and McSween, 1983; Berger et al., 2001; Warner, 2001). Buck Creek is located near the western margin of the Cartoogechaye terrane in a region of granulite facies regional metamorphism (Absher and McSween, 1985; Eckert et al., 1989), west of Franklin, North Carolina. Metadunite, some containing amphibole, is the only abundant ultramafic rock in the Buck Creek complex.Uncommon (but prominent) lenses/layers of metatroctolite are found within the metadunite. The lenses of metatroctolite grade into corundum-amphibole-feldspar-zoisite-margarite±kyanite rocks, called edenite-amphibolites by Hadley (1949) but more recently described as edenite-margarite schists (Tenthorey et al., 1996; Collins et al., 1998). Rare hornblende diorite is also included within the metadunite. Granulite facies metamorphism of the troctolite produced sympletic orthopyroxene-clinopyroxene-spinel and sapphirine-bearing coronas resulting from dry reactions between primary olivine and calcic plagioclase (Tenthorey et al., 1996; Lang et al., 2004).

The protolith for the Buck Creek ultramafic rocks and enclosed mafic rocks was probably a cumulate series consisting of interlayered olivine (metadunite) and olivine-calcic plagioclase (metatroctolite/corundum-bearing amphibolite) adcumulates, both containing accessory Al-rich chromite (Tenthorey et al., 1996; Warner, 2001). High-P-T led to the development of coronal (dry) and amphibole-corundum-zoisite bearing assemblages, and subsequent lower-grade metamorphism produced chlorite, antigorite, and tremolite-bearing rocks (Emilio, 1998; Warner, 2001). Other hydrated rocks include serpentinite located along fractures and shear zones and actinolite-chlorite-talc schist found in the northern extension of the ultramafic body (Emilio, 1998; Thomas et al., 1999).

Major and trace element systematics of Buck Creek mafic ultramafic rocks enclosed within the metadunite demonstrate that these units represent layers of cumulate troctolite, anorthosite, and gabbro (Tenthorey et al., 1996; Berger et al., 2001).

The presence of extensive dunite without pyroxene is consistent with either a HOT or LHOT assignment for the Buck Creek rocks. Chromitite is not known at Buck Creek (Table 2) despite the large exposure of dunite, and this suggests Buck Creek is a HOT body. Compositions of the associated amphibolite (discussed below) are also consistent with a HOT model for Buck Creek.

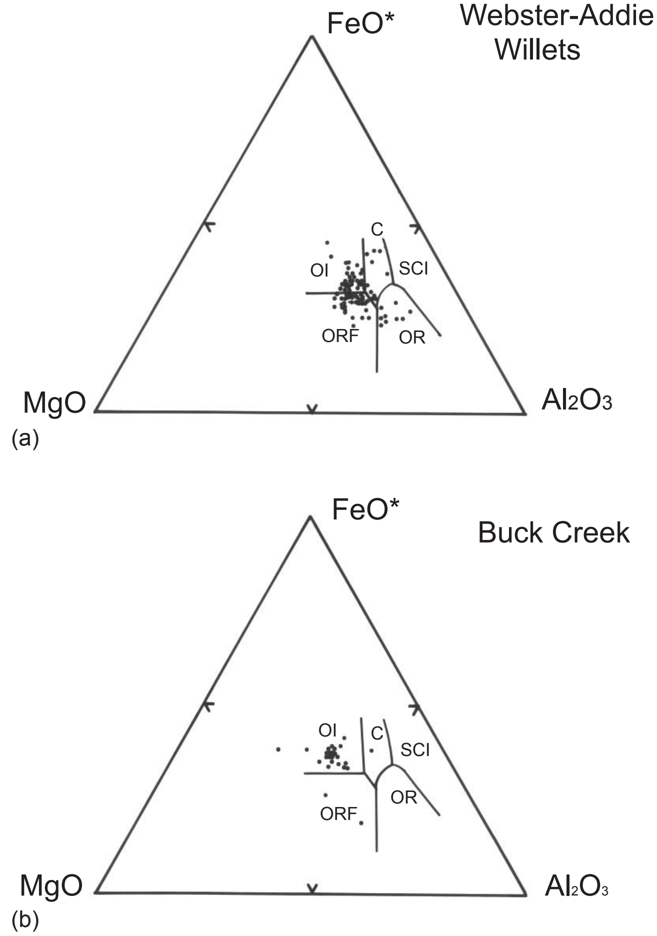


Figure 2. Amphibolite compositions from the Cartoogechaye terrane plotted on the discriminant diagram of Pearce et al. (1977) showing compositional fields for basaltic rocks from ocean floor and ridge ORF), oceanic islands (OI), continents (C) spreading center islands (SCI), and orogenic zones (OR). FeO* is total Fe expressed as FeO. Amphibolite data from the Webster-Addie/Willets area (a) from Quinn (1991) and for the Buck Creek (b) are from McElhaney and McSween, (1983) and Berger et al. (2001). Most opf the amphibolite compositions plot in the field for oceanic islands. The two analyses from Buck Creek (b) that plot in the ocean floor field (ORF) are from Berger et al. (2001) and are from cumulate metagabros included within the metadunite.

Amphibolite Compositions. Amphibolites form layers and pods that range in size from hand sample to map-scale bodies within the biotite gneiss of the Cartoogechaye terrane in the Webster-Addie area (Quinn, 1991). Quinn (1991) suggested coarser-grained amphibolites may represent diabase or gabbro protoliths. XRF analyses of 82 amphibolite samples yield a range of 42-57 % SiO_2 with most of the analyses in the range of 48-52 % (Quinn, 1991). Quinn (1991) concluded that most of the amphibolites represent “oceanic basalts (MORB’s).” Recasting Quinn’s data (Fig. 2a) reconfirms the oceanic character of the amphibolites. Other studies of amphibolites in the Webster-Addie area report higher silica values (up to 63 % SiO_2; Soraruf

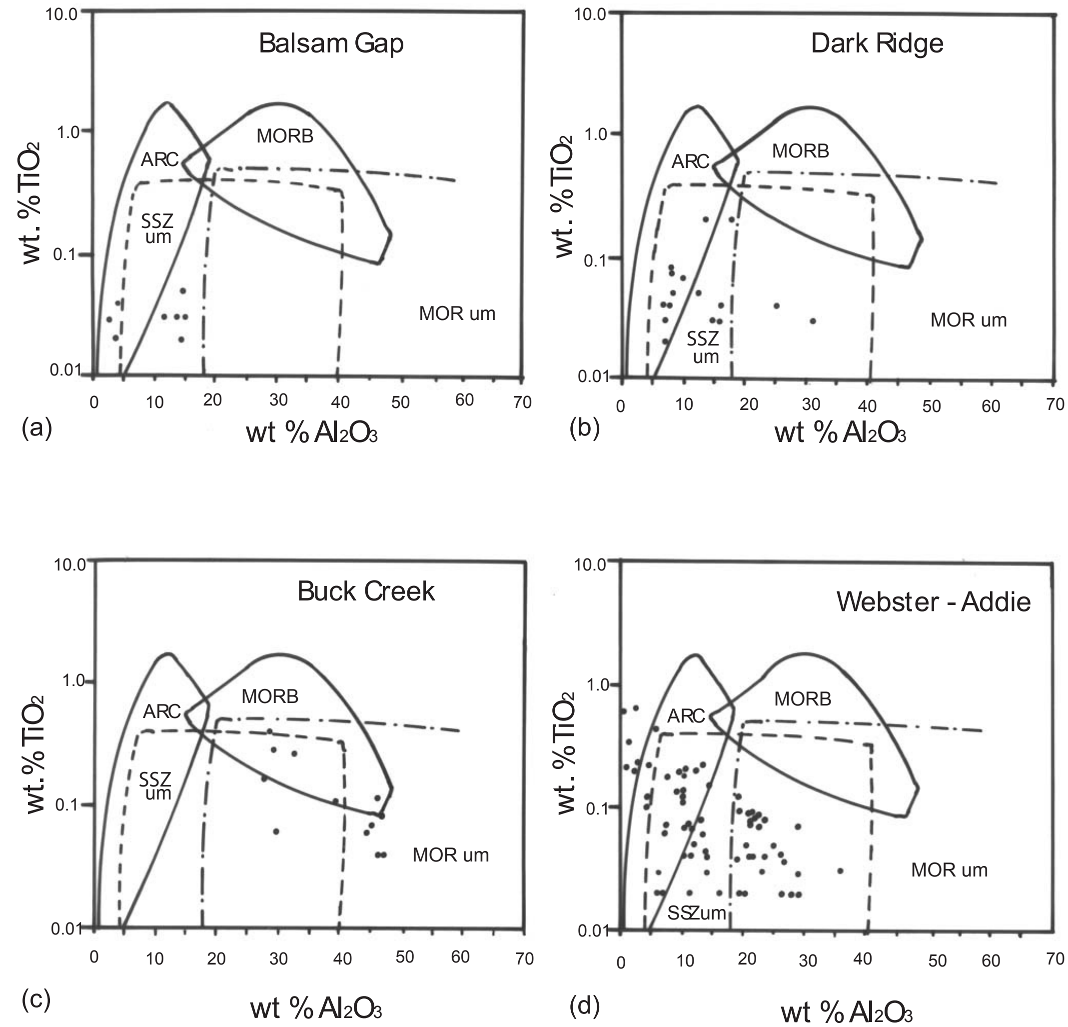


Figure 3. Chromite core compositions from metadunites in the Cartoogechaye terrane plotted on the discriminant diagram of Kamenetsky et al. (2001) showing compositional fields for chromite from subduction-related (ARC) and ocean floor (MOR) volcanic rocks and mantle peridotites from oceanic (MOR) and suprasubduction zone (SSZ) settings. Chromites from Balsam Gap (a), and Dark Ridge (b) and many of the Webster-Addie (d) plot in the SSZ field. A scattering of Webster-Addie analyses at low Al contents is related to recrystallization to lower Al contents, even in the cores of the grains. Chromite from Buck Creek (c) is largely recrystallized, but a few apparently unrecrystallized core compositions have high Al contents, overlapping the SSZ and MOR ultramafic fields.

et al., 2002a, b). Ryan et al. (this guidebook) found most of the amphibolites in the Addie area to be more siliceous than those reported by Quinn (1991), and suggested the likely protoliths were subduction-related andesitic lavas.

Amphibolites completely enclose the Buck Creek ultramafic body, and are petrogenetically related to it. McElhaney and McSween (1983) suggested an opholite origin for these rocks, a view supported by the later work of Tenthorey et al. (1996) and the results of the WCU-USF REU Site program (Collins et al., 1998; Morman et al., 1999; Berger et al., 2001). The rocks plot in the oceanic island field (Fig. 2b). Limited Nd and Sr isotopic compositions are consistent with a noncontinental setting for the magmatism (Shaw and Wasserburg, 1984). Trace element and REE compositions suggest the Buck Creek amphibolites formed in a "oceanic gabbro/cumulate" setting (Peterson et al., 2004).

Chromite Compositions. Chromite compositions in Dark Ridge and Balsam Gap metadunites have a wide range in compositions related to the loss of Al. Chromite with little or no chlorite shows modest zoning from Al-rich cores to Al-poor rims. Chromite with the "lattice texture" are mostly homogeneous and low in Al, reflecting loss of Al due to exsolution of an Al-rich phase. Even the most Al-rich cores from Balsam Gap have only modest Al contents (Fig. 3a) and suggest a suprasubduction zone setting. Chromite from Dark Ridge has a greater range of Al contents (Warner and Helper, this guidebook), but most of the data still plot in the suprasubduction zone field (Fig. 3b).

Chromites from the metadunites in the Buck Creek complex have a wide range of composition (Warner, 2001). The pattern of Al loss associated with chlorite formation and "lattice textures" is similar to other bodies. The most Al-rich chromite cores are higher in Al than other chromites (Fig. 3c) and suggest either a suprasubduction zone or MOR setting for formation of the Buck Creek complex.

Modest amounts of chlorite are associated with chromite in metadunites of the Webster-Addie complex. The chlorite occurs along the margins of chromite grains and as a pseudomorphic replacement of a mineral that exsolved from the chromite forming the "lattice" texture described by Lipin (1984). The lattice chromites are uniformly low in Al. Chromite rimmed by chlorite is zoned from Al-rich cores to Al-poor rims. There is a wide range of core compositions (Fig. 3d), probably reflecting varying degrees of Al loss associated with formation of chlorite. The most Al-rich chromites suggest either a suprasubduction zone or MOR setting for formation of the Webster-Addie complex.

Mars Hill Terrane

Rocks of the Mars Hill terrane include abundant granitic and pelitic gneisses, mafic rocks, and sparse ultramafic rocks and include the oldest currently known rocks in the Appalachians south of Virginia (Ownby et al., 2004). These rocks are typically migmatitic, reflecting the Precambrian granulite facies metamorphic grade of the rocks. Mafic rocks include amphibolite, various varieties of biotite hornblende gneiss and metagabbro, and two-pyroxene granoblastites. Geochronologic relations (Paleo- and Mesoproterozoic rocks, 1.0 Ga age for metamorphism) contrast with basement rocks of the Western and Eastern Blue Ridge and suggest the geologic history of the Mars Hill terrane had little in common with its now-adjacent neighbors (Ownby et al., 2004).

Canton Area. Several small bodies of metadunite are enclosed in biotite gneiss in the Mars Hills terrane near Canton, North Carolina (Merschat and Wiener, 1988) . The Newfound Gap metadunite (Palmer et al., 1977; Hirt et al., 1987) is the largest of these bodies. The metadunite is composed mostly of olivine with some chromite. Serpentine is the most common hydrous metamorphic mineral, but some serpentine-poor rocks contain abundant talc. Small amounts of chlorite, anthophyllite and tremolite also occur within the body (Hirt et al., 1987). Petrofabric studies of the olivine reveal a LPO fabric. (Palmer et al., 1977). Small bodies of chlorite-talc schist also crop out locally within the Mars Hill terrane (Merschat, 1977, 1993; Merschat and Wiener, 1988).

The abundance of dunite and the lack of chromitite (Table 2) suggest the Canton area metadunites are HOT ophiolites of Nicolas and Boudier (2003). However, the data are very limited.

Amphibolite Compositions. Amphibolites occur as relatively thin layers and pods within the feldspathic schists and gneisses of the Mars Hill terrane near Canton, North Carolina (Merschat and Wiener, 1988). Compositions of amphibolites in the Canton area west of the Holland Mountain fault are similar to amphibolites in the Webster-Addie area and suggest a intermediate protolith with a subduction-related origin (Peterson et al., 2004).

Chromite Compositions. No data are available on chromite compositions from the Mars Hill terrane.

Dahlonega gold belt

Rocks of the Dahlonega Gold belt are dominantly pelitic schists and gneisses. Hatcher (1971, 1974; and German, 1985) describe these rocks as metagraywacke. Regional metamorphism to amphibolite facies assemblages has not resulted in extensive partial melting of these rocks.

Mafic and ultramafic rocks in the Dahlonega gold belt are sparse. The Sally Free complex northwest of Dahlonega is a relatively large fault-bounded unit composed of amphibolite, metagabbro, hornblende gneiss and the Cane Creek gneiss. Settles et al. (2002) concluded these rocks were part of a Early Ordovician volcanic arc formed in a "volcanic arc and/or back-arc basin." Small bodies of metadunites and associated talc schist and serpentinite crop out near Dillard, GA in the Beavert Suite (Vincent et al., 1990). Minerals included within the ultramafic rocks include olivine, anthophyllite, talc, and serpentine (Vincent et al., 1990). In the Corbin Knob and Prentiss quadrangles of North Carolina small, unnamed bodies of ultramafic rocks are found in rocks of the Dahlonega Gold belt (Hatcher, 1980; Hatcher and Hopson, 1989)

Lake Burton. The Lake Burton mafic-metadunite complex crops out along Highway 76 to the west of Clayton, GA (Hopson, 1989). A package of amphibolite and metagabbro is interlayered with metasandstone and mica schist. Small bodies of metadunite, serpentinite and metapyroxenite are enclosed within the mafic rocks. The metadunite consists mainly of olivine with lesser amounts of chromite and orthopyroxene. Serpentine surrounds olivine in a mesh texture and veinlets of serpentine cut the olivine and pyroxene. Small coronas of chlorite surround chromite grains (Vincent et al., 1990).

The abundance of metabasalt/gabbro and dunite along with the absence of chromitite (Table 2) suggests the Lake Burton dunite is a harzburgite-type (HOT) ophiolite (Nicolas and Boudier, 2003).

Ellijay area. Several bodies of metadunite occur in the vicinity of Ellijay community in Macon County, NC. These bodies are along strike with the Corundum Hill body to the southwest. Two of these bodies, the Deposit No. 9 (Dribus et al., 1982) and Mincey Mine (Hahn and Heimlich, 1977), are described in the literature. Host rocks for these bodies are muscovite-biotite quatrzofeldspathic gneiss with some amphibolite. The ultramafic rocks have an elliptical outcrop pattern. The No. 9 body is elongate parallel to the regional foliation (Dribus et al., 1982). The elongation of the Mincey Mine body is almost perpendicular to the regional foliation (Hahn and Heimlich, 1977). Fractures, commonly filled with talc, tremolite, and anthophyllite within the Mincey Mine metadunite, parallel the regional foliation. A small dike of pegmatitic granite intrudes the metadunite along one of these fractures. A thin, metasomatic rind of talc-amphibole schist separates the metadunites from the enclosing country rocks.

The ultramafic rocks are dominantly metadunite, but minor metaharzburgite occurs at the Mincey Mine (Hahn and Heimlich, 1977). Small amounts of chromite and variable amounts of talc, serpentine, chlorite, tremolite, and anthophyllite complete the mineral assemblage. Strained porphyroclasts of olivine are enclosed in a matrix of fine-grained polygonal olivine. Internal structures in the metadunite (chromite lineations and olivine orientation) appear unrelated to the present shape of the bodies.

These characteristics (abundant dunite, harzburgite and lack of chromitite, (Table 2) best match the features of the harzburgite-type (HOT) ophiolites of Nicolas and Boudier (2003).

Corundum Hill. The best known ultramafic body in the Dahlonega gold belt is the Corundum Hill metadunite (Yurkovich, 1977). Mining of abrasive and gem corundum during the early 1900s gave Corundum Hill its name. Today, Corundum Hill is the name given to the subdivision built on the site. Host rocks are the same mica gneisses that enclose the Ellijay area metadunites. Contacts between the metadunite and gneiss were observed in several locations and suggest the body has a tabular, sill-like form (Yurkovich, 1977).

The ultramafic rock is mainly metadunite, but variable amounts of orthopyroxene in some rocks indicate metaharzburgite and meta-olivine pyroxenite (Yurkovich, 1977). Disseminated chromite occurs in the metadunite and small masses of chromitite occur locally (Hunter et al., 1942). A variety of hydrous minerals, including serpentine, talc, chlorite , and amphiboles (magnesiocummingtonite/anthophyllite, tremolite) attest to the metamorphic character of the rocks. Veins of talc (locally with magnesite) and chlorite (with some corundum and green spinel) cut the metadunite.

These characteristics (abundant dunite, presence of harzburgite and chromitite, Table 2) best match the features of the harzburgite-type (HOT) ophiolites of Nicolas and Boudier (2003).

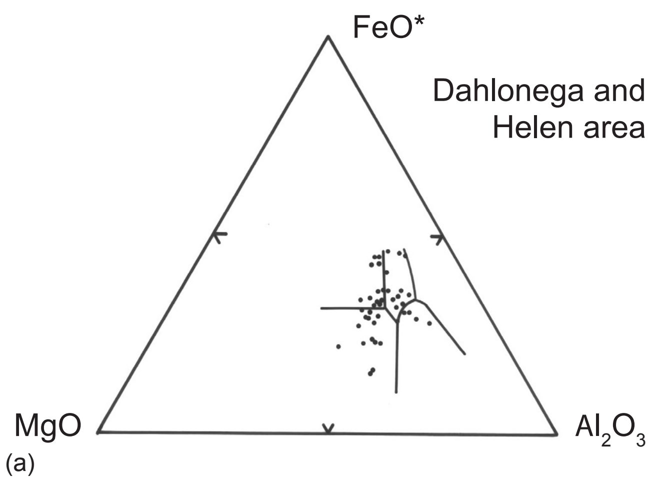
Amphibolite Compositions. Several studies provide data on the composition of the mafic rocks within the Dahlonega gold belt (Abrahms and McConnell, 1984; German, 1985; Hopson, 1989; Gillon, 1989; Settles, 2002). German (1985) and McConnell and Abrams (1984) favored a back-arc origin for the mafic rocks in the Dahlonega gold belt in Georgia, southwest of the area of this field trip. These rocks plot in the "continental" and "oceanic" fields of Figure 4a. Gillon (1989) and Hopson (1989) concluded the mafic rocks associated with the Lake Burton mafic-ultramafic complex and near Helen, Georgia represented low-K tholeiitic basalts as found in island arcs. Most of the analyses of Lake Burton amphibolites plot in the "arc" field of Figure 4b, but there is a scatter of analyses in other fields. Hopson (1989) concluded these protolith of the Lake Burton amphibolites was a low-K tholeiitic island arc basalt.

Chromite Compositions. Chromite occurs as small accessory grains, typically with a corona of chlorite, in metadunites from Lake Burton and Corundum Hill. The chlorite coronas are thicker at Lake Burton, often half as thick as the residual chromite inside the corona. Chlorite coronas on Corundum Hill chromites are not as thick; typically one flake of chlorite separates chromite from the other minerals. Development of chlorite coronas came at the expense of the chromite and chromite immediately adjacent to the chlorite is lower in Al than chromite in the cores. A vein of green spinel with chlorite and corundum cuts the metadunite at Corundum Hill.

Cores of chromite in all five of the analyzed samples from Corundum Hill are uniform in their Al content (Fig. 5a). The chromite corresponds to an "arc" or suprasubduction zone setting. Chromite analyses are available from one sample of metadunite from the small metadunite body along U.S. 76 west of Clayton, Georgia. The compositions of chromite cores in this one sample are more variable than all of the Corundum Hill samples (Fig. 5). Lake Burton chromite compositions plot in the "arc" and suprasubduction zone fields on Figure 5b. The higher Ti content and scatter of the Lake Burton chromite compositions may be related to higher degrees of recrystallization associated with the formation of thicker chlorite coronas. Raymond et al. (2003) noted an increase in chromite Ti with decreasing Al and related this to the formation of chlorite. Lipin (1984) reported one chromite analysis from the Ellijay metadunite, located near Corundum Hill. That Ellijay chromite composition plots in the suprasubduction zone field (Fig. 5b).

Tugaloo Terrane

The Tugaloo terrane is the most extensive terrane of the eastern Blue Ridge belt (and includes the western Inner Piedmont, as well) (Hatcher, 2002). This terrane, called the Spruce Pine thrust sheet, the Gossan Lead block, and the Toe terrane (see Raymond, 1998), is dominated by the Ashe and Alligator Back Metamorphic suites, and the equivalent Tallullah Falls Formation (Hatcher, 1973; Abbott and Raymond, 1984; Rankin et al., 1989). The dominant rock types are pelitic schists, mica-quartz-feldspar semischists; and hornblende schists and gneisses ("amphibolites"). Minor constituents of the terrane include the ultramafic rocks and rare marbles and metaconglomerates. Miller et al. (2000) reported these rocks were intruded by siliceous Ordovician and Devonian plutons, but recent data reveal Pennsylvanain ages, rather than Devonian (Hatcher, personal communication).



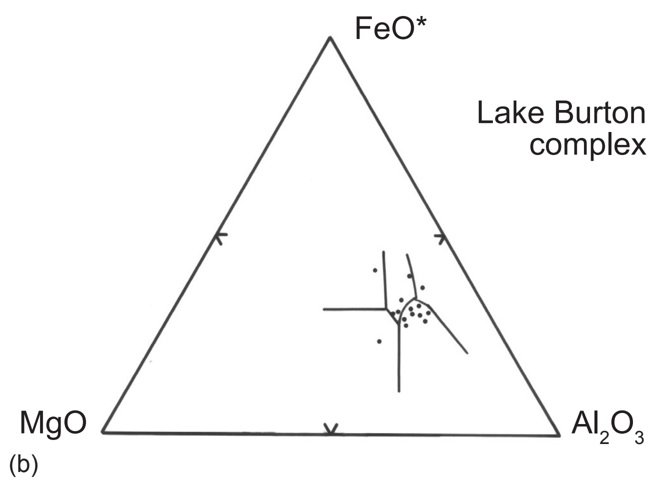


Figure 4. Amphibolite compositions from the Dahlonega gold belt plotted on the discriminant diagram of Pearce et al. (1977) showing compositional fields for basaltic rocks from ocean floor and ridge ORF), oceanic islands (OI), continents (C) spreading center islands (SCI), and orogenic zones (OR). FeO* is total Fe expressed as FeO. Amphibolite data from the Helen and Dahlonega area are from German (1985), Gillon (1989), and Settles (2002). Compositions of amphibolite for the Lake Burton complex (b) are from Hopson (1989).

The Tugaloo terrane contains a number of ultramafic bodies. Swanson (2001) described 15 bodies of ultramafic rock within the Tugaloo terrane near Spruce Pine, North Carolina. Raymond and coworkers (Raymond and Abbott, 1997; Raymond et al., 2001) described ultramafic outcrops in the Tugaloo terrane near Spruce Pine and north of Boone, North Carolina. Rather than review all of the ultramafic bodies in this terrane, several bodies along the strike of the Tugaloo terrane in western North Carolina are described below.

Greer Hollow. The Greer Hollow ultramafic body is exposed in the Ashe Metamorphic Suite in the Gossan Lead (thrust) bock of the Tugaloo terrane in northwestern North Carolina (Raymond et al., 1988; Raymond and Abbott, 1997; Raymond et al., 2003). The body is dominated by amphibole-chlorite schist, but contains a variety of rock types. Local zones contain porphyroclastic schist with relict orthopyroxene crystals and harzburgite microlithons in a chlorite matrix. A mappable body of magnetite-chlorite and magnetite-garnet-chlorite schist occurs at the southern end of the body; serpentinite veins and serpentinized zones are abundant in the north. Trace element chemistry in some rocks suggests altered gabbro protoliths (Vance and Raymond, 1994), whereas whole rock chemistries indicate that other rocks were ultramafic. No primary mineralogy or structures remain in the body.

Geologic mapping reveals that the Greer Hollow body forms a lens within pelitic schists and hornblende schists and gneisses and is tightly folded with them (Raymond et al., 2003). Contacts are sharp, but locally are marked by metasomatic reaction zones containing actinolite-talc assemblages. The surrounding region contains abundant metamafic rocks, typically with MORB chemistries (Misra and Conte, 1991) that are dominated by either plagioclase or hornblende. No ophiolite "stratigraphy" is recognizable.

It is difficult to match data from this body (Table 2) against the models in Table 1. The lack of dunite does not seem to fit any model, but the presence of harzburgite and absence of chromitite *seems* to best fit the harzburgite-type (HOT) ophiolites of Nicolas and Boudier (2003).

Hoots. The Hoots ultramafic body is a small, crudely lens-shaped, metadunite-rich mass of rock that crops out in triangular exposure near the base of the Gossan Lead (thrust) block of the Tugaloo terrane in northwestern North Carolina (Raymond et al., 2001). The body is surrounded by polymetamorphic, polydeformed amphibolite facies hornblende schist and gneiss derived from MORB protoliths (Misra and Conte, 1991). Contacts between the Hoots rocks and the surrounding mafic rocks are not exposed, but internal foliations in the metadunite are generally parallel to those in the surrounding gneisses and schists. Where the contacts of similar bodies in the Tugaloo terrane are exposed, they are tectonic and show blackwall metasomatic reaction zones against the country rocks (Raymond et al., 2003).

These characteristics (abundance of dunite, presence of harzburgite, lack of chromitite, Table 2) indicate the body is most similar to the harzburgite-type (HOT) model (Table 1) of Nicolas and Boudier (2003).

Day Book. The Day Book ultramafic body is enclosed in rocks of the Ashe Metamorphic Suite in the Spruce Pine (thrust) block of the Tugaloo terrane near Burnville, NC (Swanson, 1981, 2001; Raymond, 1995). The body is dominated by metadunite, with minor amounts of metaharzburgite and metachromitite. Locally hydrated lithologies include tremolite-chlorite-talc schist and serpentinite. Isolated pods and layers of metachromite represent primary igneous structures, but most of the rocks contain metamorphic fabrics. The ultramafic rocks are surrounded by amphibolite facies mafic and pelitic schists and gneisses. The contact between the ultramafic rocks and the enclosing schists and gneisses is sharp and characterized by a metasomatic "blackwall" of amphibole-talc-chlorite schist. A granitoid pegmatite of the Spruce Pine Plutonic Series intrudes the ultramafic rocks and the schists and gneisses (Swanson, 1981).

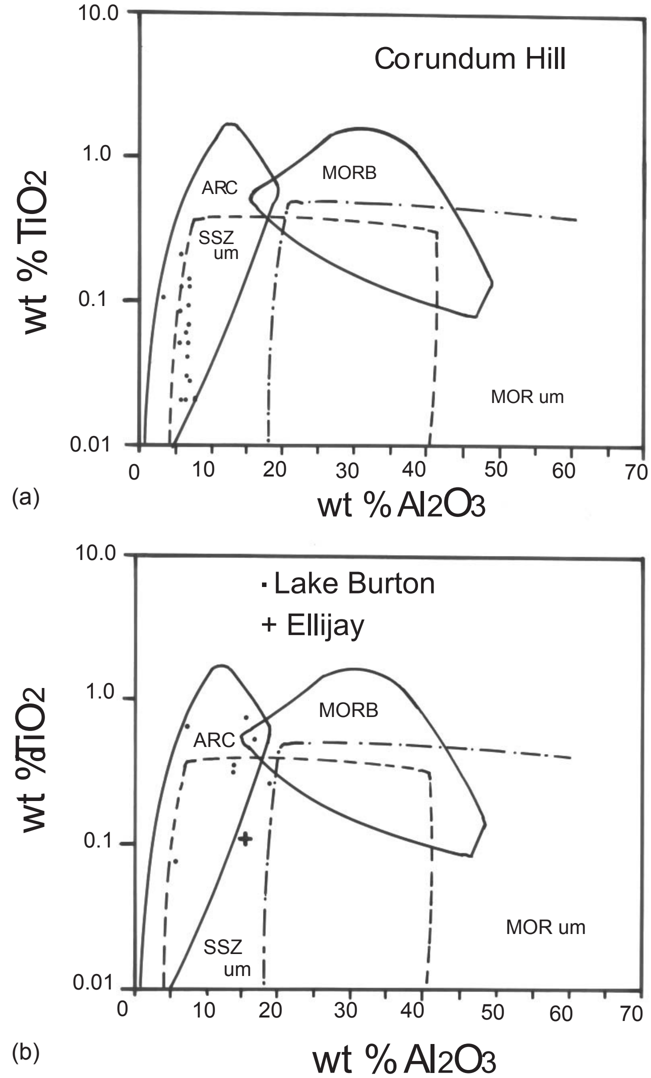


Figure 5. Chromite core compositions from metadunites in the Dahlonega gold belt plotted on the discriminant diagram of Kamenetsky et al. (2001) showing compositional fields for chromite from subduction-related (ARC) and ocean floor (MOR) volcanic rocks and mantle peridotites from oceanic (MOR) and suprasubduction zone (SSZ) settings. Chromites from Corundum Hill (a) have very uniform Al contents and plot along the margin of the SSZ field and in the center of the ARC field. The analyses from Lake Burton are all from a single sample and scatter within the ARC and SSZ fields (b). The single analysis from Ellijay is taken from Lipin (1984).

The ultramafic rocks are included as a pod-like mass in the enclosing schists and gneisses. Foliations within the schists and gneisses are generally conformable to the contacts with the ultramafic rocks. The polymetamorphic/polydeformational history of the schists, gneisses, and ultramafic rocks makes interpretation of any original stratigraphy difficult and none was recognized.

The abundance of dunite together with the presence of harzburgite and chromitite (Table 2) is most consistent with either the LHOT- type ophiolite of Nicolas and Boudier (2003), but a HOT classification is also possible.

Laurel Creek. Amphibolite, some containing garnet, encloses pods of ultramafic rock in the Laurel Creek complex in northeast Georgia. Host rocks for the Laurel Creek complex include the amphibolite facies schists and gneisses of the Tallulah Falls Formation (Hatcher et al., 1984). The ultramafic rocks are mostly hydrated lithologies including amphibole-chlorite-talc soapstone and serpentinite. Two bodies of metadunite (olivine with minor tremolite, magnesiocummingtonite/anthophyllite, chlorite talc, and corundum) represent less hydrated rocks, and these pods are surrounded by the soapstone. The Laurel Creek Mine was a major supplier of corundum in the United States in the 1890s. Asbestos was also produced from these rocks in the 1880s. Rare, small pods of coarse-grained chromitite occur in the metadunite.

The sinuous, elongate outcrop pattern of the amphibolite of the Laurel Creek complex is generally parallel to relict bedding in the enclosing Tallulah Falls metasedimentary rocks (Hatcher et al., 1984) suggesting an original sheet-like form of the complex. The sinuous outcrop pattern of both the amphibolite and metasediments reflects the polyphase deformation of these rocks and supports a predeformation emplacement of the Laurel Creek complex. A thin rind of soapstone separates ultramafic rocks from the enclosing garnet amphibolite (Vincent, et al., 1990) and reflects metasomatic reaction during regional metamorphism.

Compositions of the Laurel Creek amphibolites point to an oceanic origin for the basaltic protolith (Hatcher et al., 1984). Helms et al. (1987) reported the Laurel Creek amphibolites contained an assemblage of Al-rich minerals including gedrite, staurolite, and kyanite. Compositions of amphibolites from the adjacent Tallulah Falls Formation are different and suggest more of a continental affinity. An ophiolitic origin for the Laurel Creek Complex is proposed by Hatcher et al. (1984).

The abundance of dunite and the presence of chromitite (Table 2) is most similar to the LHOT model, but is also consistent with the harzburgite-type (HOT) model of Nicolas and Boudier (2003).

Amphibolite Compositions. In the central part of the Tugaloo terrane, amphibolite is found in association or in proximity to some of the ultramafic bodies, such as the Laurel Creek body, and as isolated layers or lenses enclosed in metasedimentary rocks. These layers probably represent flows of mafic lava and, based on limited data, have compositions consistent with continental basalts (Hatcher, 2002). The larger mafic/ultramafic complexes, such as the Laurel Creek complex, are the best candidates for ophiolites (Hatcher, 2002). Compositions of amphibolites near the Laurel Creek complex are variable with SiO_2 contents ranging from 47 to 60 percent, but most of the analyses fall in the rage of 47 to 52 percent SiO_2 (Hatcher et al., 1984). Petrotectonic discrimination diagrams suggest an oceanic or arc setting for the mafic magmatism (Hatcher et al., 1984). The Laurel Creek amphibolites plot as oceanic floor or island rocks on Figure 6a.

To the north, compositions of amphibolite in the Tugaloo Terrane, including the area around the Hoots and Greer Hollow ultramafic bodies, show that the rocks are metamorphosed tholeiitic basalts (Misra and Conte, 1991). The amphibolites range from 46 to 53 % SiO_2. Misra and Conte interpreted the amphibolite compositions in terms of N-type and T-type MORB basalts derived from a depleted source region and a third basalt type that was from a source region enriched in incompatible elements (perhaps due to interaction with a fluid phase). The authors concluded that the basalts formed in association with a spreading center, but could not distinguish between a oceanic ridge or back-arc spreading center for the tectonic setting (Misra and Conte, 1991). These rocks plot as oceanic floor or island rocks (Fig. 6b), in the same area as the Tugaloo amphibolites to the north (Fig. 6a).

Chromite Compositions. Chromite from metadunite in the Spruce Pine area of North Carolina preserves a complex pattern of recrystallization (Swanson and Raymond, in preparation). Individual spinel grains are commonly rimmed by chlorite, formed at the expense of the chromite. Zoning of individual grains is common and complex. Variation in Al from core to rim of chromite grains is related to the formation of the ubiquitous chlorite. As the size of the chlorite corona on the chromite increases, more Al is lost from the spinel. Eventually, a thick chlorite corona surrounds a homogeneous grain of low-Al chromite (Swanson and Raymond, 2004). Chromite grains with the highest core Al contents are the most likely to preserve igneous compositions (Swanson and Raymond, 2004) and these grains indicate a suprasubduction zone setting for the ultramafic rocks in this part of the Tugaloo terrane (as shown by the field on Fig. 7).

Chromite from Laurel Creek metadunite also has higher Al cores and common coronas of chlorite. Zoning patterns are as noted in the Spruce Pine chromites. Core compositions of the most Al-rich chromites from the Laurel Creek metadunites plot in the field defined by the Spruce Pine high-Al chromites (Fig. 7) and suggest a suprasubduction zone setting for formation of the Laurel Creek complex.

TERRANE HISTORY AND MAFIC/ULTRAMAFIC ROCKS

The distribution of ultramafic and associated mafic rocks varies among the terranes studied. Part of this variation is related to the outcrop area of the terranes, with larger terranes having more occurrences, but the density of occurrences does vary between terranes. The Cowrock and Cartoogechaye terranes (Fig. 1) are of similar size, but differ in the concentration of ultramafic/mafic rocks. The Cowrock terrane contains a few isolated pods and lenses of mafic/ultramafic rocks concentrated mainly along the Hayesville fault at the base of the terrane. The largest bodies of ultramafic rocks in the Blue Ridge, some associated with mafic rocks, occur in the Cartoogechaye terrane,

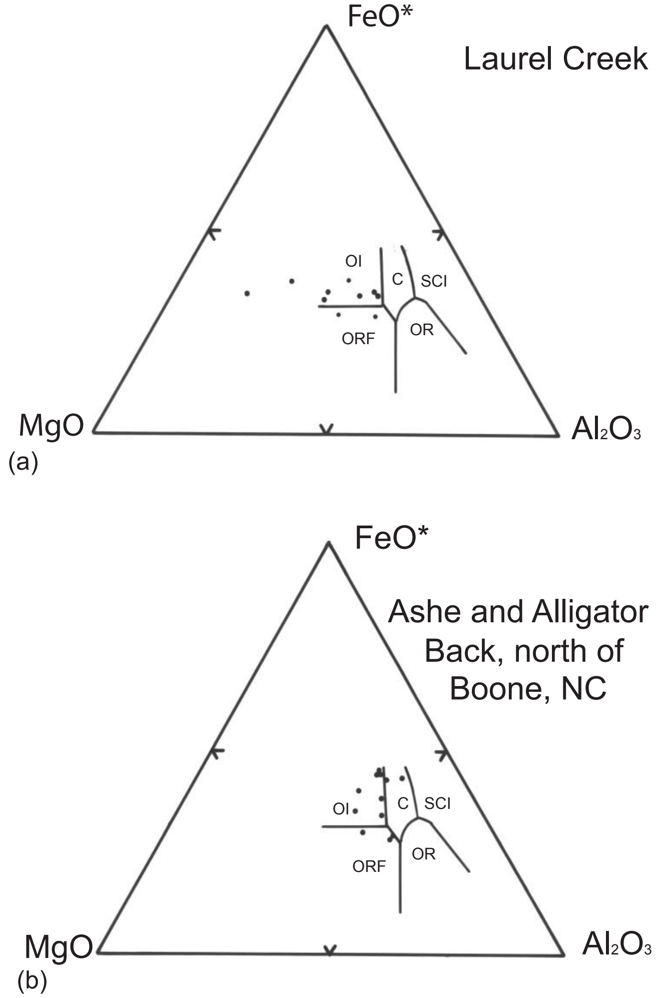


Figure 6. Amphibolite compositions from the Tugaloo terrane plotted on the discriminant diagram of Pearce et al. (1977) showing compositional fields for basaltic rocks from ocean floor and ridge ORF), oceanic islands (OI), continents (C) spreading center islands (SCI), and orogenic zones (OR). FeO* is total Fe expressed as FeO. Amphibolite data from Laurel Creek (a) are from Hatcher et al. (1984) and the amphibolite compositions from the Ashe and Alligator Back Formations north of Boone, NC are from Misra and Conde (1991).

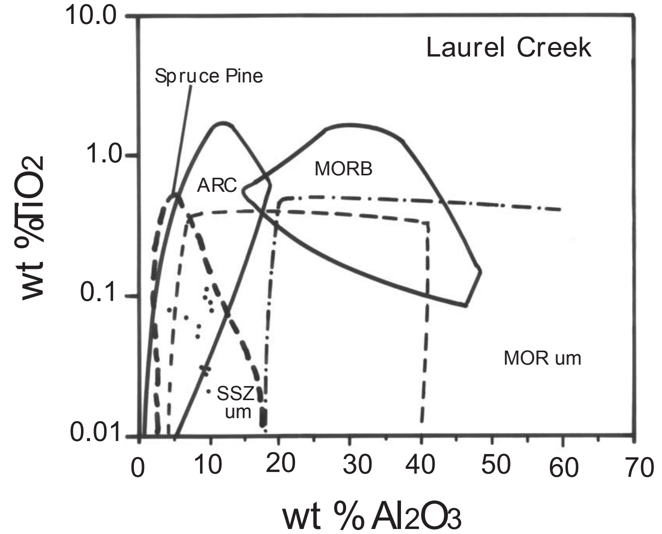


Figure 7. Chromite core compositions from metadunites in the Laurel Creek metadunite plotted on the discriminant diagram of Kamenetsky et al. (2001) showing compositional fields for chromite from subduction-related (ARC) and ocean floor (MOR) volcanic rocks and mantle peridotites from oceanic (MOR) and suprasubduction zone (SSZ) settings. The heavy dashed line encloses the field for the least recrystallized chromites from over 10 metadunite bodies in the area of Spruce Pine, NC (from Swanson and Raymond, in preparation). The data from Laurel Creek plot within the SSZ field and largely overlap the Spruce Pine field.

as do numerous smaller bodies. The Mars Hill terrane does not share the history of adjacent rocks in the eastern or western Blue Ridge; it is truly an exoctic terrane that only contains a few small bodies of mafic and ultramafic rock. Small bodies of ultramafic and mafic rocks are common in the Dahlonega gold belt. The Tugaloo terrane contains a large number of isolated bodies of mafic and ultramafic rocks.

Cowrock terrane. The Cowrock terrane formed outboard of the Laurentian continental margin (Hatcher et al., 2004), but contains Laurentian zircons Bream et al., 2004). Mafic and ultramafic rocks are associated with the accretion of the Cowrock terrane, but the accretion is not related to closing of any major (e.g., Iapetus) oceanic basin (Hatcher et al., 2004).

Small bodies of metadunite and a unique garnet clinopyroxenite are associated with the amphibolites and metagabbros of the Lake Chatuge complex (Hartley, 1973). The REE content of the garnet clinopyroxenite shows a depleted LREE pattern (Berger et al., 2001) consistent with the Nd isotope results reported by Shaw and Wasserburg (1984). The dominance of metagabbro in the Lake Chatuge complex along with the depleted character of the rocks suggests a HOT setting for the Lake Chatuge complex. The presence of olivine metaclinopyroxenite in the Dicks Creek suite suggests a relation to the rocks of the Lake Chatuge complex.

Mafic/ultramafic rocks are not abundant in the Cowrock terrane. The outboard setting of the terrane, cumulate textures and depleted chemistry suggest a MOR setting for the mafic/ultramafic rocks of the Cowrock terrane.

Cartoogechaye terrane. The Cartoogechaye terrane was also formed outboard of the Laurentian continental margin, but contains detrital zircons with Laurentian affinities (Hatcher et al., 2004, this guidebook). The accretion of the Cartoogechaye terrane to North America was accompanied by the obduction of ultramafic rocks and mafic rocks onto the continental margin (Hatcher et al., 2004). This suturing may be related to the closing of a forearc basin rather than the closing of Iapetus (Hatcher, 2001).

The largest mafic/ultramafic complexes in the Blue Ridge are in the Cartoogechaye terrane and Hatcher (2002) thought these large bodies are the best candidates for ophiolites. The Buck Creek complex contains metagabbro and a relatively thick section of amphibolite, all consistent with a HOT ophiolite. Amphibolite geochemistry and chromite compositions are also consistent with a MORB-related origin for these rocks.

The Webster-Addie complex is also large and has amphibolites in immediate proximity (Ryan et al., this volume), but ultramafic rocks are not surrounded by amphibolite and do not contain metagabbros as at Buck Creek. This, together with other properties, suggests the rocks are from a LHOT ophiolite.

The presence of some clinopyroxene-rich ultramafic rocks suggests a more enriched mantle source for these rocks. Such an enriched source is consistent with the proposed subduction-related source for the associated amphibolites (Peterson et al., 2004). A suprasubduction zone setting is proposed for the origin of the Webster-Addie complex based on amphibolite geochemistry (McElhaney and McSween, 1983; Quinn, 1991; Soraruf et al., 2002a, 2002b; Peterson et al., 2004) and chromite compositions.

The isolated ultramafic bodies at Balsam Gap and Dark Ridge are associated with the same subduction-related amphibolites as the Webster-Addie complex. The ultramafic bodies are consistent with a LHOT ophiolite.

Most of the mafic/ultramafic rocks of the Cartoogechaye terrane seem to have formed in association with a suprasubduction zone setting. The Buck Creek complex, however, appears to be more MOR-related.

Mars Hill terrane. Isotopic ages of rocks from the Mars Hill terrane indicate these rocks are older than adjacent rocks in both the eastern and western Blue Ridge (Ownby et al., 2004). Protoliths for the Mars Hill terrane include sedimentary and mafic igneous rocks, suggesting the terrane represents a continental fragment (Ownby et al., 2004). Docking of the Mars Hill terrane with the western Blue Ridge apparently was completed by the Neoproterozoic as mafic magmas of this age intruded both packages of rock (Ownsby et al., 2004). Thus, the Mars Hill terrane is part of the western Blue Ridge and not part of the central or eastern Blue Ridge. A similar history of Paleozoic metamorphism shows the amalgamation of the Mars Hill terrane with the western and eastern Blue Ridge was completed by the Paleozoic Era.

Small bodies of ultramafic rocks are a minor part of the Mars Hills terrane and none of the ultramafic rocks are well studied. The metadunite at Newfound Gap (Hirt et al., 1987) is the largest of these bodies in the Canton area of the Mars Hill terrane (Merschat and Wiener, 1988). Available data suggest the Newfound Gap body may be part of a LHOT ophiolite.

The enriched source region for the ultramafic rocks together with the subduction-related character of the amphibolites (Peterson et al., 2004) suggest a suprasubduction origin for the mafic/ultramafic rocks.

Dahlonega gold belt. The Dahlonega gold belt terrane lies between (and beneath) Laurentian-related Cowrock and Cartoogechaye terranes and the extensive Tugaloo terrane (Fig. 1). Rocks of the Dahlonega Gold belt are amphibolites, metagraywackes, and metapelites (German, 1985). Mafic rocks extend in a thin belt along the length of the Dahlonega Gold belt terrane. Hopson et al. (1989) noted that felsic gneisses and mafic rocks are more extensive in Georgia and thin to the northeast in North Carolina.

Small, isolated bodies of ultramafic rocks occur with the mafic rocks in the Dahlonega Gold belt. The Lake Burton mafic-ultramafic complex contains small pods and lenses of metadunite, metaclinopyroxenite, and serpentinite (Hopson, 1989), all surrounded by amphibolite. Further north, small bodies of metadunite in the Ellijay area and at Corundum Hill also have amphibolite in the immediate area. The favored setting for these rocks is that of a HOT-type ophiolite. Rocks of the Dahlonega gold belt probably formed in a suprasubduction zone environment associated with subduction-related arc magmatism.

Tugaloo terrane. Rocks of the Tugaloo terrane represent igneous rocks and sedimentary protoliths formed as, and deposited on, Iapetus oceanic crust (Hatcher, 2002). The most extensive unit in the Tugaloo terrane is the Ashe-Tallulah Falls Formation. Hatcher (2002) distinguished three parts to the Tallulah Falls Formation; a lower unit originally comprised of graywacke and mafic igneous rock, a middle pelitic unit, and an upper unit of graywacke. To the north, in the vicinity of Spruce Pine, North Carolina, the Ashe Metamorphic Suite consists of quartzo-feldspathic gneisses, pelitic schists, and amphibolites and shows the same stratigraphy as theTallulah Falls Formation to the south (Hatcher, personal communication, 2005). Associated with the more abundant rock types are smaller masses of ultramafic rock, metaconglomerate, metagabbro, marble, and massive sulfide deposits (Parker, 1952; Rankin et al., 1972). North of Boone, North Carolina, amphibolite is more abundant and, with quartzo-feldspathic gneiss, is a major lithology (Abbott and Raymond, 1984).

Ultramafic rocks in the Tugaloo terrane are not everywhere intimately associated with amphibolites. There are several kinds of amphibolites in the Tugaloo terrane: (1) isolated pods with and without ultramafic rocks that could represent dismembered ophiolites, (2) bodies of amphibolite and ultramafic rocks that are possibly dismembered ophiolites, and (3) amphibolites interlayered with the mica schists and quartzo-feldspathic gneisses. The ultramafic rocks are mostly metadunite with minor amounts of metaharzburgite. Small lenses of metapyroxenite occur in the metadunites (Raymond et al., 2001; Swanson, 2001). Some of the ultramafic rocks are intimately associated with amphibolites (Hoots, Laurel Creek), but most of the ultramafic rocks are surrounded by metasandstones and pelitic schists with some layers of amphibolite. The abundance of metadunite and metaharzburgite plus the MORB-like character of associated amphibolites suggest that the rocks represent a HOT-type ophiolite.

Raymond et al. (2003) considered the petrogenesis of Blue Ridge ultramafic rocks. Ultramafic rocks in the Tugaloo terrane do not commonly have associated gabbroic rocks and Raymond et al (2003) cited this as evidence for formation at a slow spreading center. Amphibolite compositions are consistent with an oceanic setting, either MOR or back-arc. Compositions of selected chromite grains suggest an origin for the ultramafic rocks in an arc of a suprasubduction zone setting (Raymond et al., 2003).

CONCLUSIONS

In spite of many years of study, the exact tectonic setting in which the ultramafic rocks were formed still eludes us. The variety of rock types and associations with mafic rocks coupled with the complex history of metamorphism and deformation, present a real challenge to unraveling the petrotectonic history of the rocks. The evolving terrane model of the Blue Ridge that is the subject of this field trip provides a new way to look for relations among the occurrences of mafic/ultramafic rocks.

Chromite is one of the few minerals in the Blue Ridge metaultramafic rocks that may retain some vestige of its igneous heritage. However, the chromite (and associated chromian magnetite and magnetite) was extensively recrystallized. Careful study of chromite zoning is required to determine if any potentially igneous signature remains in the rock. The spatial association of chlorite with chromite and the development of the "lattice texture" of Lipin (1984) both reflect recrystallization of chromite and such chromite grains do not preserve igneous compositions.

The available data on the mafic/ultramafic rocks in the various terranes are not uniform. Some terranes, such as the Tugaloo, are large and contain an abundance of mafic/ultramafic rocks, but even in this setting many of the ultramafic rocks are poorly characterized, especially in the southern part of the terrane. Other terranes, such as Cowrock, contain few mafic/ultramafic rocks and these are not well characterized. Indeed much of the petrologic work on these ultramafic bodies over the years has centered on the best-exposed outcrops or the bodies with the most diverse suite of rock types. The Buck Creek and Webster-Addie complexes in the Cartoogechaye terrane are well-characterized, and together have received more analytical attention than all complexes combined.

Based on this very uneven coverage and limited data base, some VERY TENTATIVE conclusions can be made. 1. The mafic/ultramafic rocks in the Cowrock terrane and the adjacent Buck Creek complex in the Cartoogechaye terranes have MORB-like affinities. 2. These bodies formed immediately outboard to the Laurentian continental margin. The remainder of the Cartoogechaye terrane (and adjacent Mars Hill terrane) appear to have formed in a suprasubduction zone setting. 3. The Tugaloo terrane has amphibolite compositions suggestive of a MOR setting, but the ultramafic rocks and chromite chemistry suggest a suprasubduction zone setting for the origin of these rocks. Together, these data suggest a back basin setting.

Clearly much more work needs to be done, but the terrane model provides a guide to future studies and shows us where our studies of mafic/ultramafic rocks should be focused. Studies are currently under way on the use of geochemistry and chromite chemistry to more fully define the setting of the mafic and ultramafic rocks, especially in the Cowrock, Dahlonega Gold belt, and Mars Hill terranes.

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Geology and petrogenesis of mafic and ultramafic rocks of the Willets-Addie area, central Blue Ridge, NC

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ABSTRACT

Rocks encountered in the Addie-Willets area of western North Carolina include migmatitic amphibolite gneisses, and highly deformed and variably altered dunites and pyroxenites of the Webster-Addie ultramafic complex. Detailed mapping results indicate that both mafic and ultramafic rocks occur as discrete cm- to km-scale bodies, locally in contact with each other, but more commonly enclosed in regionally extensive meta-sedimentary schists and gneisses. Evidence for high-temperature metamorphic conditions at moderate pressures (~ upper amphibolite facies) includes extensive migmatite development in the amphibolites and associated metasediments, extensive zones of talc+anthophyllite+vermiculite+chlorite±amphibole and/or antigorite alteration adjacent to trondjhemite dikes within the dunites, and pervasive ductile deformation.

Compositionally the amphibolites have predominantly andesitic protoliths with no clear petrogenetic relationships to the dunites. Ultramafic protoliths are geochemically and lithologically similar to smaller ultramafic bodies exposed in the Ellijay Creek and Balsam Gap areas, but distinct from ultramafic rocks found in "oceanic" mafic-ultramafic associations within the Buck Creek or Lake Chatuge complexes.

The random, "block-in-matrix" association of the units in the Addie-Willets area at a variety of scales and the arc-like character of the mafic rocks are similar to what is observed in subduction-related mélange deposits. Our results are consistent with convergent margin origins, and emplacement of these units into an accretionary sequence that was subsequently buried, heated, and strongly deformed.

INTRODUCTION

Isolated mafic and ultramafic are scattered throughout the central Blue Ridge, from Georgia to Virginia (Misra and Keller, 1978; Misra and McSween, 1984; Misra and Conte, 1991). While the petrogenesis of some of these rocks is well-documented (i.e., the Bakersville metagabbro, and mafic volcanic rocks of the Mt. Rogers and Grandfather Mountain formations: Goldberg et al, 1986; Rankin 1993), for most it is poorly understood. A primary scientific goal of the WCU-USF Blue Ridge REU Site research program has been to document the origins and tectonic significance of these units, through careful, detailed field, petrologic and geochemical characterization. Primary field targets of the 2001 Summer program were the mafic and ultramafic rock exposures in and around the hamlets of Addie and Willets, NC, along with the Balsam Gap ultramafic body (Figure 1). Twelve undergraduate researchers, with the guidance of five WCU and USF faculty, mapped and sampled four field areas, followed by targeted field geophysical studies of the ultramafic units, and geochemical characterization of all mafic and ultramafic rock types encountered (Soraruf et al.

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Ryan, J. G., Yurkovich, S., Peterson, V., Burr, J., and Kruse, S., 2005, Geology and petrogenesis of mafic and ultramafic rocks of the Willets-Addie area, central Blue Ridge, NC, *in* Hatcher, R. D., Jr., and Merschat, A. J., eds., Blue Ridge Geology Geotraverse East of the Great Smoky National Park, Western North Carolina: North Carolina Geological Survey, Carolina Geological Society Annual Field Trip Guidebook, p. 91-98.

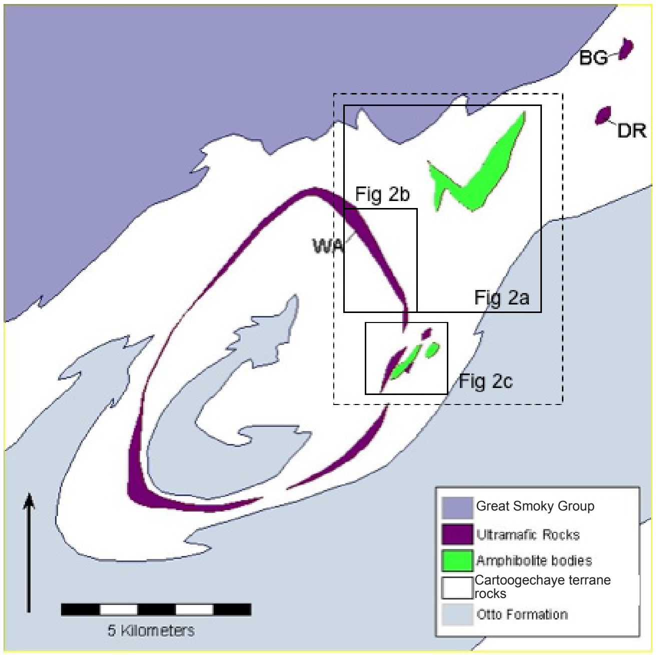


Figure 1. Schematic geologic map of the Webster-Addie Ultramafic body and associated mafic and ultramafic deposits. Addie-Willets field areas lie within the square. Modified from Brown et al 1995 (NC State Geological Map) and Quinn, 1991. BG: Balsam Gap dunite; DR: Dark Ridge dunite; WA: Webster-Addie ultramafic body.

2002a; Primm et al. 2002; Doughty et al. 2002). Post-summer studies of these sites included detailed analysis of pyroxenitic horizons in the dunites (McIlmoil et al. 2002) and rare-earth element characterization of key samples (Soraruf et al. 2002a)

BACKGROUND, FIELD AREA, AND PAST WORK

Earlier maps of the Addie/Willets area were produced as part of broader field campaigns focused on characterizing the Webster-Addie ultramafic deposits (Condie and Madison, 1969; Madison, 1968; Cronin, 1983; Quinn, 1991). The Webster-Addie ultramafic body, with its distinctive ring-shaped map pattern was mined extensively for olivine at several sites, most notably at Webster, Addie, and Chestnut Gap (Madison, 1968; Cronin, 1983; Quinn, 1991; Figure 1). Although highly recrystallized, metamorphosed dunite is the most common rock type in the unit, pyroxenites (orthopyroxenites, clinopyroxenites, and websterites) occur commonly throughout the body. Podiform pyroxenite bodies within the dunite vary in size from ~1-2 cm to several meters in length. Hydrothermal fluid exchanges, associated with crosscutting trondjhemitic dikes, have locally altered the fresh ultramafic rocks to zoned hydrous assemblages that include talc, anthophyllite, vermiculite, Mg-chlorite and locally tremolite or a green Mg-hornblende.

Amphibolite and ultramafic bodies are locally in contact, but are more typically enclosed within meta-sedimentary biotite and/or muscovite schists and gneisses. The amphibolites are also present as lenticular outcrop- to map-scale bodies. The most pervasive amphibolite exposures are in and around the Willets community along US Highway 23-76 (Figure 2a). These rocks and the associated metasediments show evidence for polyphase, high T° metamorphism and migmatite development – typically, the amphibolites consist of coarse-grained amphibole ± biotite± garnet melanosomes enclosed in coarse to spidery feldspathic leucosomes. Cross-cutting pegmatitic dikes within the migmatitic amphibolite may represent coalesced regions of the leucosome.

Past mapping results/inferences

Madison (1968) and Condie and Madison (1969) were the first to report on the structure and lithology of the Webster-Addie deposit in significant detail. Subsequent quadrangle-scale mapping studies by Cronin (1983) and Quinn (1991) refined the field geometry and outcrop pattern. Current structural interpretations suggest emplacement of the Webster-Addie dunite as a solid sheet, followed by deformation to produce a dome structure with a ring-shaped exposure pattern (Madison, 1968; Quinn, 1991). In detail the body is more extensively exposed in the northeast and southwest near Addie and Webster, respectively, and either very thin or dismembered along the southeastern and northwestern parts of the ring. Rocks exposed within the Webster-Addie "ring" are interpreted to be a different unit than those outside the ring and possibly correlative with rocks west of the Hayesville thrust (Cronin, 1983; Quinn, 1991). Previous geochemical data from the Webster-Addie complex, mostly trace element and isotopic studies of Webster pyroxenites, points to a non-depleted mantle source (i.e., εNd ~ 0; Shaw and Wasserburg, 1984). Amphibolites were reported in these studies (most recently by Quinn, 1991), and are included on maps where regionally significant, but have not been studied in detail.

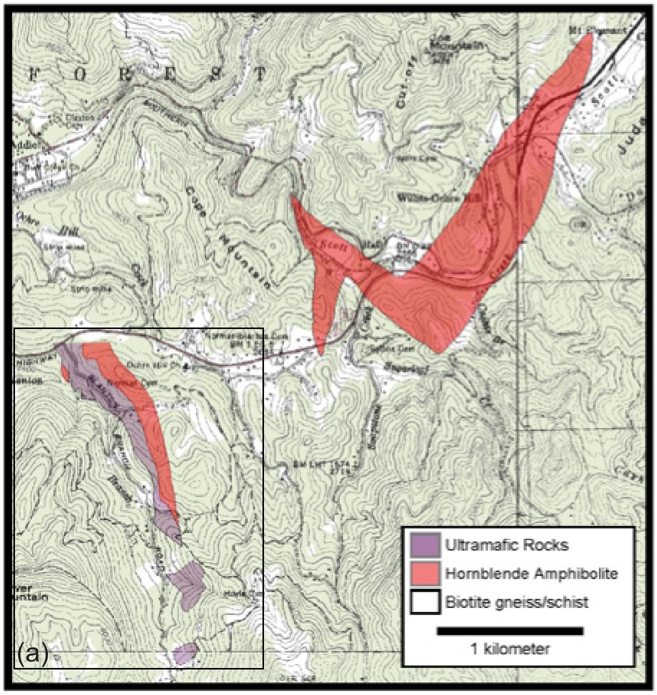
RESULTS AND DISCUSSION

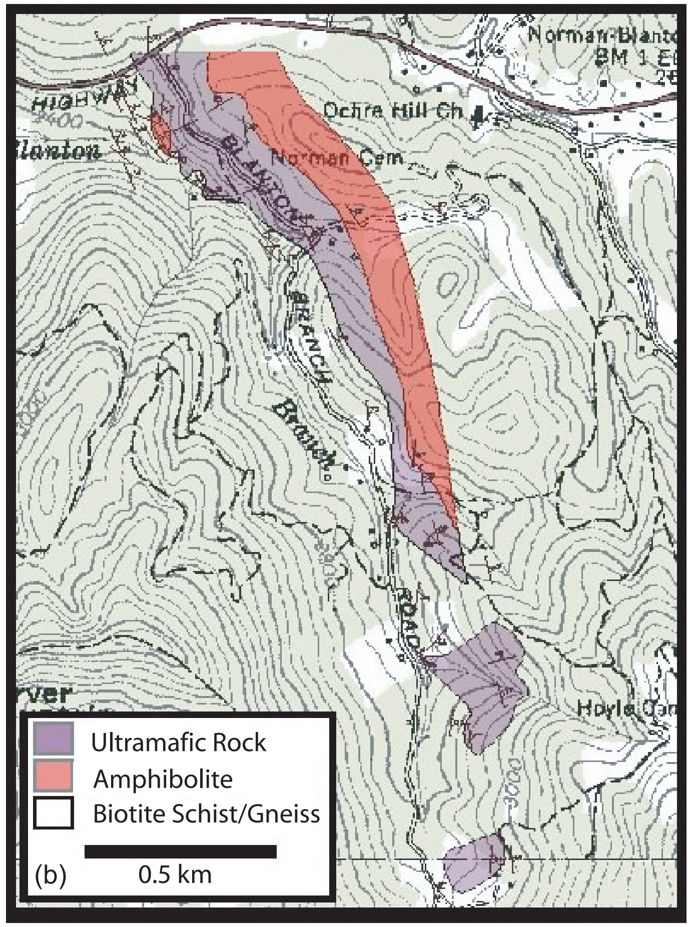
Detailed mapping, typically at 1:6000 scale and at outcrops scale in some quarries resulted in a slightly revised map pattern of the northeastern region of the Webster-Addie body and Willets area (Figure 2). Mapping was conducted by undergraduate participants in the 2001 Blue Ridge REU Site Summer program, closely mentored by the faculty authors of this contribution.

Field relations and lithologic descriptions

Ultramafic rocks. Our detailed mapping results point to an even more dismembered and lenticular character in the ultramafic rocks than previously recognized by Quinn (1991) and others in this area. REU field results show the fairly thick ultramafic body exposed in Addie pinching out southward with several small lenses defining the arc of the Webster-Addie ring to the south toward Chestnut Gap (Figure 2b). In the Chestnut Gap area, some small ultramafic lenses are offset from and oblique to the arcuate trend of the Webster-Addie ring (Figure 2c).

Dunite is the most common primary rock type encountered in the Addie and Chestnut Gap deposits. The fresh dunite (there is very little of this, as most has been at least partially altered to serpentine) has a classic recrystallized 'sugar' texture, with stringers of chromite (Figure 3). Pyroxenites occur as lenses in the dunite, ranging in size from a few cm to several meters across. Most of the smaller pyroxenite lenses are comprised of bronzitic enstatite, while the largest lenses contain significant diopsidic clinopyroxene, grading into websterite, which shows an almost gneissic banding of diopside and bronzite (Figure 4a). The pyroxenite lenses are sheathed in a tremolite+talc hydration assemblage, and pinch out into talc-rich stringers that define a pervasive foliation in the dunite that follows the regional patterns





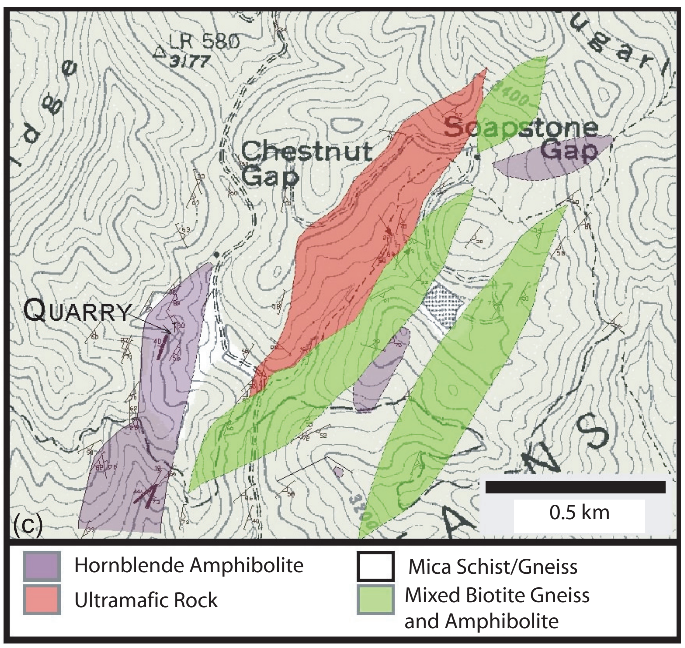


Figure 2. (a) Closeup of the northern part of the Addie-Willets field area. Ultramafic and amphibolite unit exposures based on detailed mapping results from the 2001 Blue Ridge REU Site research program. (b) Closeup of segment of the Webster-Addie ultramafic deposit in and around Blanton Branch Road. Note adjacent amphibolitic rocks, and the discontinuous nature of the ultramafic units. (c) Webster-Addie ultramafic and mafic exposures at Chestnut Gap.

(Figure 4b-c). Hydrothermal fluid interactions with the dunite produced serpentine-rich vein assemblages consisting largely of coarse antigorite. The most common metasomatic assemblage in the Addie dunites relates to the intrusion of trondjemitic dikes, and the dramatic chemical exchange reactions that occur in response to these intrusions. The contact assemblage is an

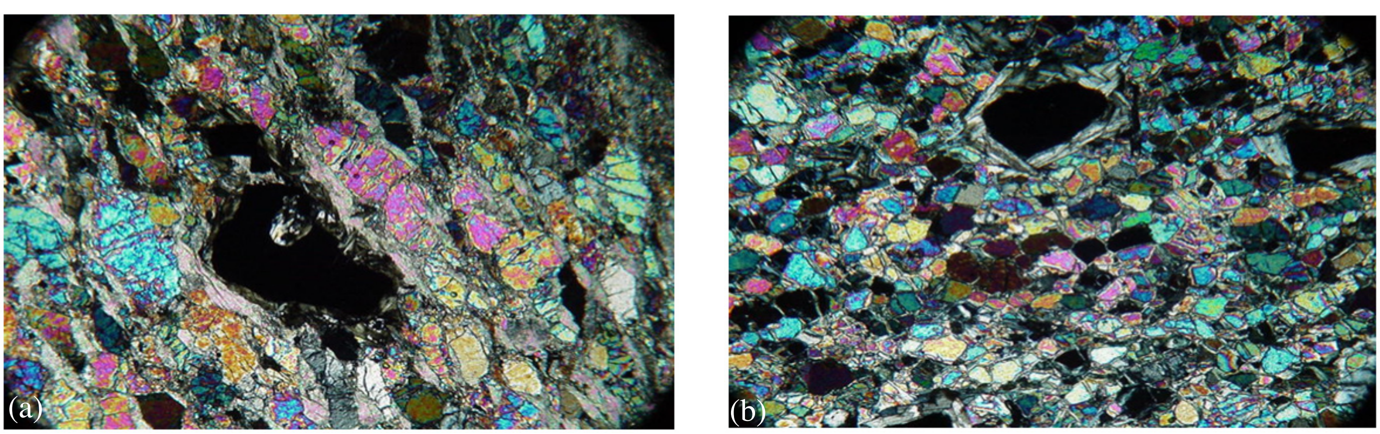


Figure 3. Thin sections of typical Webster-Addite dunite. (a) Sample WA01JH22B, crossed nicols Note elongated olivines that impart a subtle foliation. Field of view is 4.3 mm. (b) Sample WA01CB16B, crossed nicols. Fine grained polygonal texture common in Webster-Addie dunite. Chromite is surrounded by secondary kammererite, as is typical in these rocks. Olivine is partly serpentinized. Field of view is 4.3 mm.

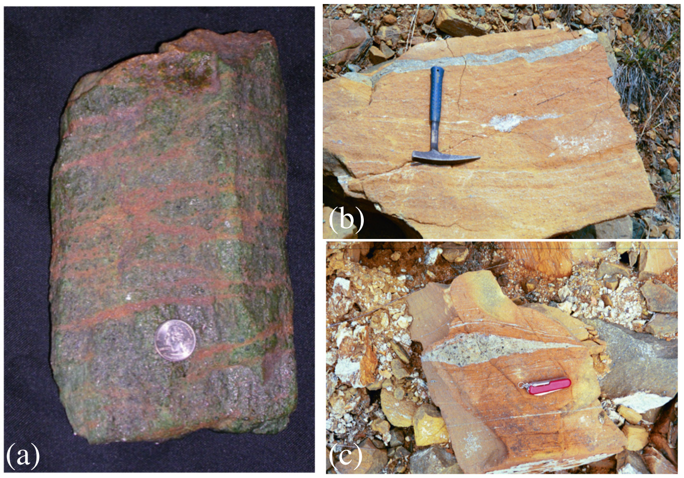


Figure 4. (a) Websterite with gneissic OPX and CPX banding, Blanton Branch Road. (b) Layers of pyroxenite in dunite, and talc-tremolite lens and laminations, from Chestnut Gap quarry (see Figure 2c). (c) An orthopyroxenite lens with tremolite-talc laminae, from Chestnut Gap quarry.

thophyllite+talc+chlorite +vermiculite, grading from chlorite-rich assemblages near the dunite, through anthophyllite+talc and into vermiculite-bearing assemblages nearest the intrusion. The intruding igneous rocks are typically completely reacted with the enclosing dunite, and consist of feldspar+hornblende assemblages. Exposed walls at the Chestnut Gap quarry preserve numerous dike reaction coronas, and large zones of dunite in this locality have been pervasively altered to a magnesian amphibole+chlorite granofelsic rock (Figure 5).

Amphibolites. An extensive body of amphibolite is exposed in and around the Willets community. Smaller map- to outcrop-scale lenses and pods of amphibolite are exposed adjacent to the Webster-Addie ultramafic rocks or enclosed within meta-sedimentary country rocks in the Addie-Chestnut Gap region. (Figure 2; Figure 6). The rocks consist of hornblende+feldspar±biotite ±quartz, locally with minor titanite (Figure 7). The rocks are variably migmatitic, with amphibole+biotite+titanite melanosomes and feldspar+quartz leucosomes. The best migmatite development is observed in a large roadcut along US 23-76 in Willets and in the surrounding area (Figure 6b; Figure 2). The amphibolites typically preserve the same regional folding and foliation patterns as the meta-sedimentary country rocks. Dikes and veins crosscut the amphibolites. Some ptygmatically folded dikes appear to be spatially associated with the migmatites and may represent anatectic melts.

Geochemistry and Chemical Petrology

Samples were analyzed for major element and lithophile trace element abundances by DC Plasma spectrometry at the

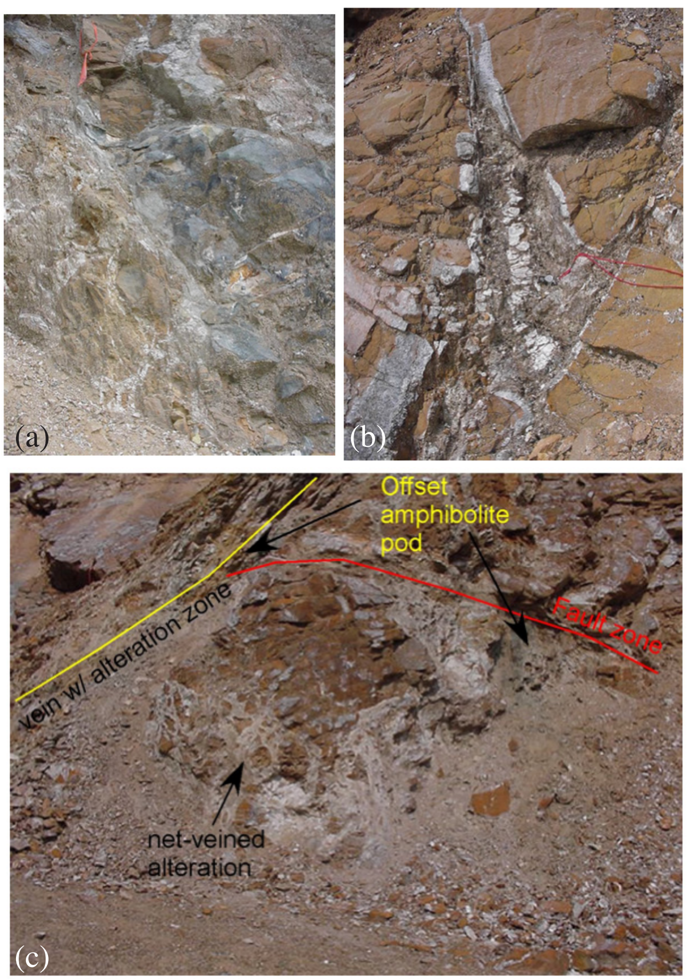


Figure 5. Ultramafic rock exposures at Chestnut Gap, typical of all Webster-Addie ultramafic outcrops in this field area. (a) Fine lattice-work anthophyllite-talc veins. (b) Thick anthophyllite-talc vein/dike, including abundant vermiculite, amphibole, and occasional feldspathic "trondjhemite" (c) Highly amphiolitized zones of ultramafic protolith in association with veins.

Figure 6. Characteristic exposures of amphibolites in the Addie-Willets area. (a) small amphibolite pod. (b) Migmatitic mafic amphibolite (grades from highly felsic to highly mafic).

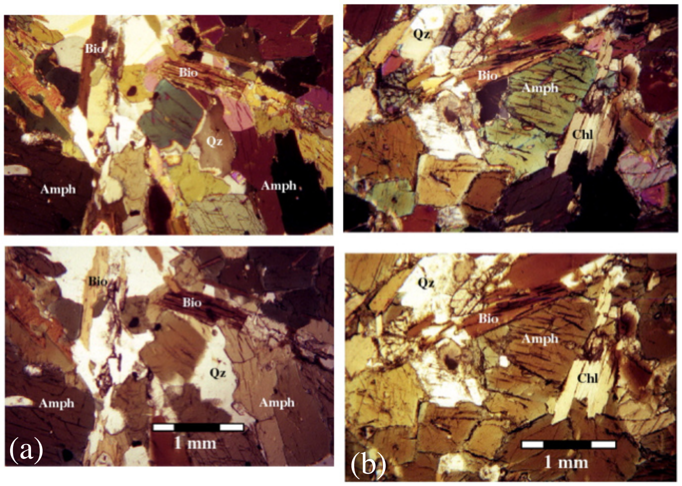


Figure 7. Typical thin sections of Willets/Addie amphibolites. (a) Sample WI01RF10B, crossed nicols (top) and PPL (bottom). Note the strong red/brown pleochroism of both the amphibole and biotite. (b) Sample WI01RF1D, crossed nicols (top) and PPL (bottom).

University of South Florida; see Tenthorey et al (1996) and Savov et al (2001) for details on the analytical protocols for mafic and ultramafic compositions. Rare earth element abundances were measured on a subset of amphibolites and pyroxenite samples by ICP-MS at Boston University. Table 1 includes typical compositions for dunites, pyroxenites and amphibolites from the Addie-Willets sites. Volatile-free geochemical data are used on all variation diagrams, to more reliably infer likely igneous protoliths.

Amphibolite samples are plotted on a K_2O vs. SiO_2 classification diagram (after Gill, 1981) in Figure 8, along with typical amphibolites from the Buck Creek and Carroll Knob mafic-ultramafic complexes in SW North Carolina. In general, Addie-Willets amphibolites are elevated in both silica and potassium, with compositions more consistent with basaltic andesites and andesites, or the intrusive equivalents thereof. By contrast, the amphibolites associated with the Buck Creek and Carroll Knob complexes are poorer in silica and much poorer in potassium, consistent with depleted basalts/gabbros from ocean floor settings. In Figure 9, an AFM diagram, the Addie-Willets samples plot along a distinctly calc-alkaline trajectory, in contrast to the definitive tholeiitic trajectory of the Buck Creek and Carroll Knob suites. While it may be possible to derive both calc-alkaline and tholeiitic melts from similar mantle sources (see Miller et al, 1992), dramatically different differentiation histories are required, pointing to different depths of melting and crystallization, and different mean volatile concentrations in the protolith melts for the Addie-Willets amphibolites relative to those of the other Blue Ridge suites. Figure 10 compares all the

Table 1. Typical bulk-rock compositions for amphibolites and ultramafic rocks in the Addie-Willets area, reported on a volatile-free basis.

Sample	Amphibolite WI01RF1E	Amphibolite WI01RF8	Dunite WA01CB016B	Clinopyroxenite WA01BA30A	Orthopyroxenite BG01DD013
SiO_2 (wt%)	60.9	52.6	42.7	52.3	64.2
TiO_2	1.11	1.17	0.01	0.02	0.01
Al_2O_3	15.7	16.1	2.31	0.89	0.54
Fe_2O_3	7.09	9.22	9.47	5.89	2.78
MgO	4.19	6.48	44.6	25.5	32.5
MnO	0.09	0.15	0.14	0.13	0.01
CaO	5.48	10.1	1.37	17.5	0.03
Na_2O	3.48	3.09	0.05	0.24	0.03
K_2O	1.95	0.83	0.02	0.03	0.02
Sr (ppm)	257	253	4	18	1
Ba	551	114	7	7	4
Ni	49	75	2940	1140	6270
Sc	14	29	11	25	4
Cr	110	228	2700	4000	1990
V	121	215	75	109	44
Zn	96	121	43	41	43
Cu	61	36	22	24	29
Totals	99.98	99.68	100.7	102.4	100.2
CIPW Norms					
Quartz	13.57	0.84	0.00	0.00	13.44
Anorthite	21.50	27.80	6.02	1.21	0.15
Albite	29.62	26.40	0.42	2.03	0.25
Orthoclase	11.58	4.96	0.12	0.18	0.12
Diopside	4.91	18.76	0.61	65.88	0.00
Hypersthene	14.65	16.31	12.66	6.76	85.42
Olivine	0.00	0.00	77.56	22.10	0.00
Chromite	0.03	0.04	0.57	0.85	0.43
Ilmenite	2.13	2.24	0.02	0.04	0.02
Corundum	0.00	0.00	0.00	0.00	0.41
Magnetite	2.07	2.70	2.75	1.68	0.81

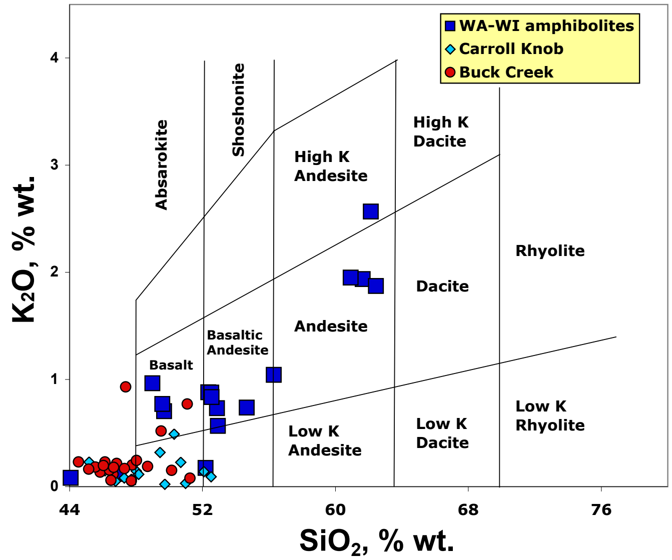


Figure 8. Plot of K_2O vs. SiO_2 for Addie-Willets amphibolites (large shaded squares). Buck Creek (small diamonds) and Carroll Knob (small circles) amphibolites included for comparison.

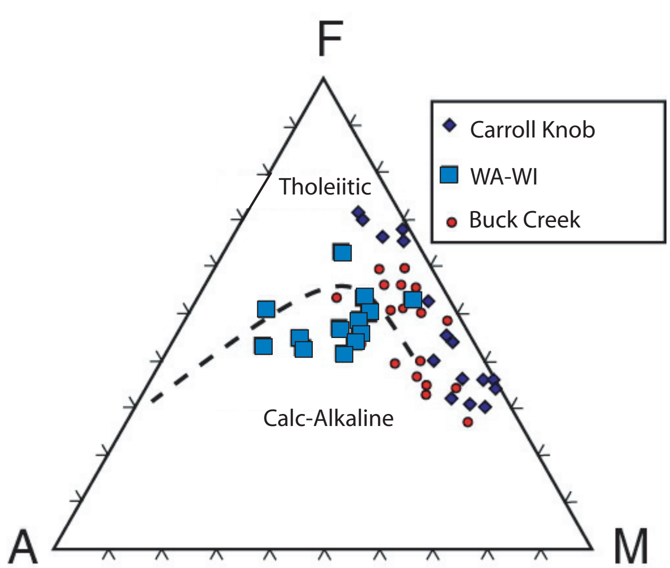


Figure 9. AFM diagram for Addie/Willets amphibolites, again with Buck Creek and Carroll Knob samples for comparison. Symbols as in Figure 8.

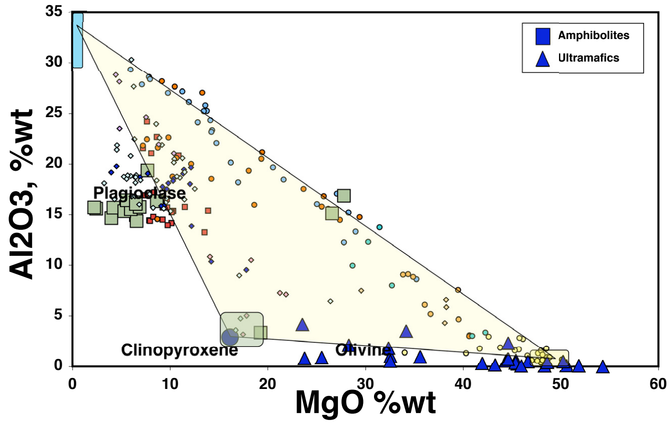


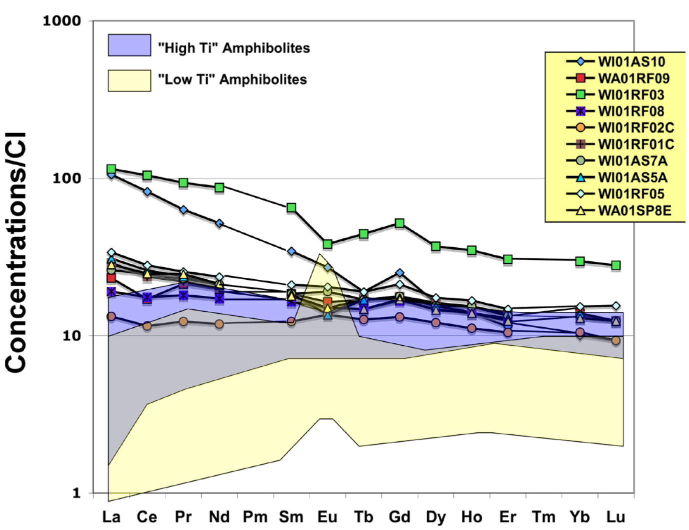
Figure 10. Plot of MgO vs. Al_2O_3 for the Webster-Addie ultramafic samples and associated amphibolites, with data from the Buck Creek and Carroll Knob complex for comparison. Symbols as in Figure 8, with the addition of the ultramafic data (large triangles)

mafic and ultramafic rocks of the Addie-Willets area to those of the Buck Creek and Carroll Knob complexes on a plot of MgO vs. Al_2O_3. On this diagram, the mineral endmembers of mafic cumulate assemblages (olivine, plagioclase and clinopyroxene) can be identified as the endpoints of a "cumulate triangle" of possible cumulate rock compositions. Most samples from the Addie-Willets area fall outside this "cumulate triangle": dunites and pyroxenites trend along the MgO axis at relatively low Al_2O_3 contents, while most of the amphibolites cluster at a relatively uniform Al_2O_3 content, consistent with lava compositions. Buck Creek and Carroll Knob complex rocks fall largely within the field for cumulate compositions, consistent with inferences by MacElhaney and McSween (1983), Hatcher et al (1984), Walter (1991), Tenthorey et al (1996), Berger et al (2001) and others that the rocks of these units represent mafic magma chamber sequences.

Figure 11 presents rare earth element (REE) data for the Addie-Willets amphibolites and some pyroxenites on Masuda-Coryell chondrite-normalized plots. All the amphibolites from this area show flat to modestly light rare-earth element enriched patterns at >10 x chondrites abundance levels. This contrasts markedly with the REE systematics of Buck Creek complex amphibolites (see Berger et al 2001), which show lower REE abundances overall, and distinctive light-REE depleted patterns consistent with ocean ridge volcanic protoliths (shaded fields, Figure 11). The overall abundances and patterns are similar to those observed in amphibolites further NE in the Blue Ridge by Misra and Conte (1991), though the low precision of these older data prevents a detailed comparison. Webster-Addie clinopyroxenite preserves a "V" shaped REE pattern at the ~1 x chondrite level. Orthopyroxenites from the Balsam Gap dunite to the NE, and Moores Knob dunite to the SW show similar REE patterns at slightly lower abundance levels, while a Buck Creek clinopyroxenite sample shows a MORB-like LREE depleted pattern at ~10 x chondrite levels, indicating a very different igneous origin, as a magmatic mafic cumulate. The similarity of the Webster-Addie, Balsam and Moore's Knob samples points to similar igneous origins, if not similar source regions, for the protoliths of these dunite bodies.

INTERPRETATION

Mineral chemistry data indicating olivine Mg numbers >90 (Madison 1968; Cronin 1983; Doughty et al. 2002), and our bulk chemical data for the ultramafic rocks in the Addie-Willets area point to residual mantle origins. By contrast, amphibolites exposed in this region preserve bulk chemical and trace element signatures consistent with igneous rocks of intermediate, calc-alkaline composition (i.e., andesites and diorites). The mafic rocks common in ophiolites are basaltic/gabbroic in character, and more evolved rocks are very rare or (more commonly) absent. Even in "arc ophiolites" such as Troodos or Thetford-Mines, the mafic rocks are commonly more silica-poor than is seen in the Addie-Willets region (Robinson et al 1983; Olive et al 1997). These geochemical differences argue against a direct petrogenetic connection between the adjacent mafic and ultramafic rock units: the andesitic protoliths of Willets amphibolites are not melts derived from the source that produced



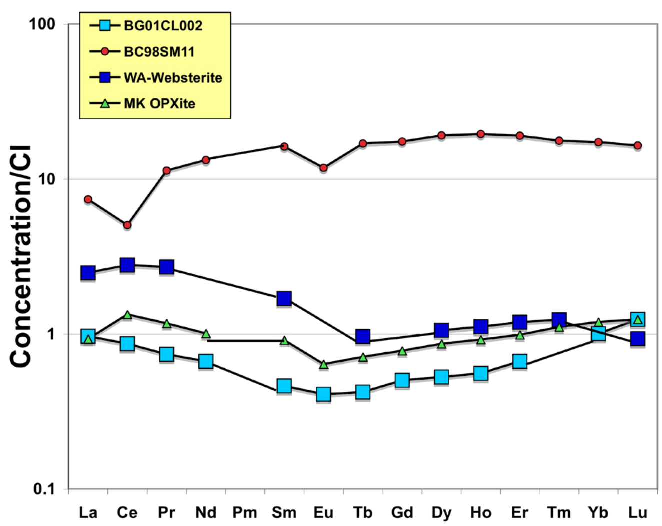


Figure 11. Masuda-Coryell chondrite normalized rare-earth element diagrams for Webster-Addie pyroxenites and Addie-Willets amphibolites. Samples are as identified on the diagrams. Fields for High Ti and Low Ti amphibolites are based on Buck Creek amphibolite data (Berger et al 2001). (a) Addie-Willets amphibolites compared to Buck Creek "high-Ti" amphibolites (which span a similar range to MORBs; Berger et al 2001), and "Low Ti" amphibolites. (b) REE patterns of pyroxenites from the Webster-Addie, Buck Creek, Balsam Gap, and Moores Knob ultramafic bodies. The Webster-Addie, Balsam, and Moores Knob samples have similar patterns and REE abundances.

the residual ultramafic rocks at Addie. The lack of regular contact relationships between mafic and ultramafic units in the field provides support to this contention: while amphibolites are in contact with ultramafic rocks at several locations in the area, these rocks are equally likely to occur as small pods within mica schists, or as larger bodies with no associated ultramafic rock. It is likely, however, that both units represent the outcomes of magmatism in a particular tectonic environment, i.e.: both may have been produced in a convergent plate boundary setting, as is typical of calc-alkaline igneous rocks today. Modern-day accretionary sequences, such as the Fransiscan Complex or the Catalina Schist may include well-preserved blocks of ultramafic rocks (dunites and pyroxenites) in close proximity with volcanic and sedimentary rocks, all reflecting different portions of a rubbleized upper plate. The rocks of the Addie-Willets area are part of what has been described as a regional olistostromal terrane, called the Cullowhee Terrane by Raymond et al (1989; more recently it has been described as part of the Cartoogachaye terrane; Hatcher and others 2004 and this volume; Swanson et al this volume). Amphibolite, ultramafic and metasedimentary blocks have been identified in the migmatitic, metasedimentary matrix in several described Cullowhee Terrane exposures lying to the south and east of Addie-Willets, along US 23 (Raymond et al 1989).

Our favored working hypothesis for the petrogenesis of the mafic and ultramafic rocks in the Addie-Willets area follow on the observations of Willard and Adams (1994) and Abbott and Greenwood (2001) to the northeast, and on the inferences of Raymond et al (1989): these rocks and their surrounding metasediments, like those of the Ashe Metamorphic Suite, represent the blocks and matrix of an ancient subduction-related accretionary sequence. The high P/T metamorphic signatures typically characteristic of such sequences have been pervasively overprinted at Addie and Willets by several of the high temperature collisional metamorphic events of the Blue Ridge, which provoked the development of migmatites and felsic dikes, domed and disaggregated the Webster-Addie ultramafic body, and produced the ductile deformation features observed pervasively in the Addie dunites and nearby schists and gneisses.

ACKNOWLEDGMENTS:

This work was supported by NSF REU Site awards EAR 9987985 (Peterson, Yurkovich and Burr) and EAR 9988077 (Ryan and Kruse). All of the Addie-Willets data presented herein was collected by the 2001 REU Site Program undergraduate participants. Buck Creek and Carroll Knob datasets were collected by undergraduate participants in the 1997, 1998, and 2000 Summer programs. JR thanks the Geochemistry Laboratory at the University of Maryland in College Park, where much of the first draft of this paper was written.

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Petrology of the Dark Ridge metaultramafic body, North Carolina

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ABSTRACT

The Dark Ridge metaultramafic body in Jackson County, North Carolina, is one of the larger metamorphosed ultramafic bodies in the central Blue Ridge Cartoogechaye terrane. The body is composed of metadunite, serpentinized in part, and metaharzburgite, much of which has been steatitized, with minor metachromitite. Textures and mineral compositions are primarily metamorphic. Hydration accompanied regional metamorphism of the body, resulting in the growth of hydrous minerals (talc, chlorite, calcic amphibole, serpentine) from the primary anhydrous assemblage of olivine + chromite ± orthopyroxene. The highest grade assemblage observed, olivine + chromite + chlorite + tremolitic hornblende ± orthopyroxene, corresponds to upper amphibolite/lower granulite facies metamorphism at 705 ± 35˚C and a pressure ~11 kb. The dominant metamorphic assemblage is olivine + chromite + chlorite + tremolite + talc and reflects retrograde lower amphibolite facies metamorphism. Talc is mainly derived from pseudomorphous replacement of orthopyroxene porphyroclasts. Serpentine-bearing assemblages formed later under greenschist facies conditions, as did magnesite-rich veins that penetrate the metaultramafic rocks. Metamorphic recrystallization of olivine and chromite affected their crystal habit and composition. Extensive polygonization of olivine occurred, producing mosaics of smaller (≤1 mm) grains with granoblastic texture. Growth of Mg-rich hydrous minerals caused olivines to become slightly more Fe-rich in hydrated rocks compared to relatively anhydrous metadunites. Primary Al-rich chromite reacted with olivine to form secondary Al-depleted chromites with poikiloblastic texture and a surrounding envelope of Cr-bearing chlorite.

INTRODUCTION

Small, metamorphosed ultramafic bodies occur along the entire length of the Appalachian Mountains (Larrabee, 1966). More than 200 are present in the southern Appalachian Blue Ridge (Raymond et al., 2003). These bodies have been of interest to geologists for years and help to give us a better understanding of the formation and metamorphic history of the southern Appalachians (Misra and Keller, 1978).

In many respects the origin of these metaultramafic bodies remains an enigma. This is partly due to their small size and often fragmented nature, but also is a consequence of their complicated metamorphic histories (Raymond et al., 2003; Swanson et al., 2005). The latter reflect Taconic, Acadian and Alleghanian metamorphism associated with the corresponding Paleozoic orogenies that affected the southern Appalachians. As many as five metamorphic recrystallization events can be recognized in individual ultramafic bodies (Raymond and Abbott, 1997).

Ultramafic rocks are highly susceptible to metamorphic recrystallization in the presence of water-rich fluids (Swanson, 2001). This can produce a variety of hydrous magnesian silicates; among them are talc, serpentine, chlorite, and anthophyllite and/or calcic amphibole. The occurrence of these minerals is sensitive to the physical conditions of metamorphism, especially temperature, and so provides an excellent guide for determining metamorphic grade (Trommsdorff and Evans, 1974; Evans, 1977). Thus, metaultramafic rocks can be useful for constraining the peak conditions attained during metamorphism. Products of retrograde metamorphic reactions are often well preserved in hydrated metaultramafic rocks, so they also potentially record assemblages developed during cooling (or reheating) post peak metamorphism.

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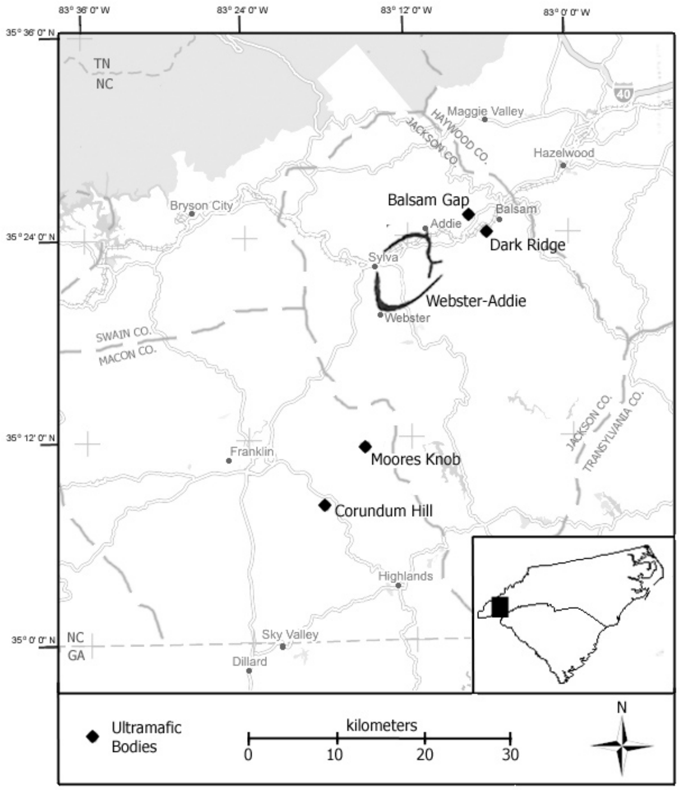

Figure 1. Map of a portion of western North Carolina giving location of Dark Ridge metaultramafic body in Jackson County. Other nearby ultramafic bodies are shown also, including the ring-like Webster-Addie complex.

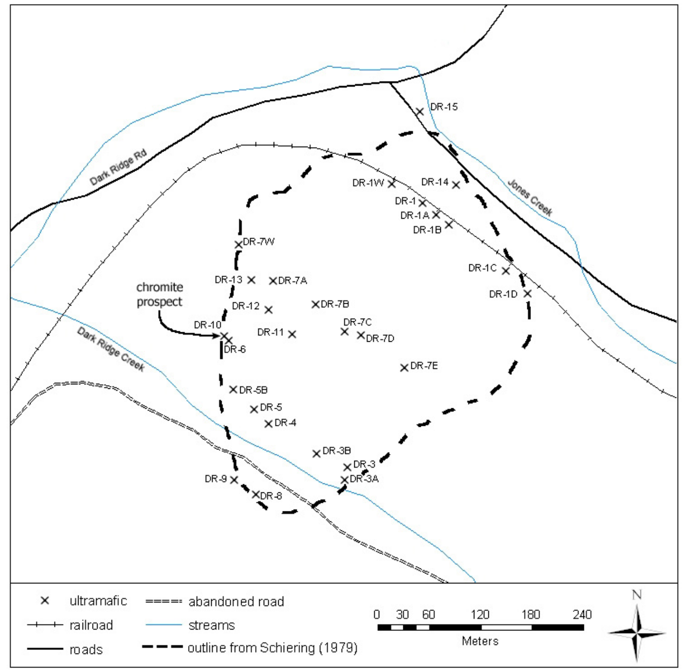

Figure 2. Map of Dark Ridge metaultramafic body (after Schiering, 1979) showing sampling locations for this study and location of the Dark Ridge chromite prospect.

In this paper we describe the metamorphic petrology of the Dark Ridge metaultramafic body. This body is located in Jackson County, North Carolina, about 6 km east of Addie and 2 km southwest of the community of Balsam (Fig. 1). It outcrops along a northwest-southeast trending ridge spur between Dark Ridge Creek to the south and Jones Creek on the north; tracks of the Southern Railway cut through the northern part of the body (Fig. 2). Dark Ridge is one of a dozen or so metaultramafic bodies contained within the Willits Tract (Furman, 1981), of which the largest body is the Webster-Addie ultramafic ring complex (Swanson et al., 2005). The tectonostratigraphic setting of the Dark Ridge metaultramafic body is within the Cartoogechaye Terrane, as recently defined by Hatcher (2001) and Hatcher et al. (2004). This terrane is bounded by the Hayesville fault to the west and the Soque River fault on the east (Settles, 2002).

PREVIOUS WORK

The earliest mention of the Dark Ridge metaultramafic body was as an unnamed peridotite body "cut by the railroad 1½ miles southwest of the (Balsam) station" (Pratt and Lewis, 1905, p. 99). They indicated that harzburgite is the chief constituent and also noted the occurrence of two chromite pits, on either side of Dark Ridge Creek, from which 15 tons of chromite ore had been removed. A more complete description, together with a map of the body, were given by Hunter (1941) in his report on olivine deposits in North Carolina and Georgia. He described the Dark Ridge deposit as "one of the largest and most outstanding olivine formations in the entire olivine belt" and estimated that it contained more than 16 million tons of "relatively unaltered granular olivine" (Hunter, 1941, p. 78). Hunter noted the presence of unusually coarse-grained olivine with crystals up to 6 inches in diameter. He also reported that the body contains seams of massive chromite and that saxonite – orthopyroxene-bearing peridotite – bordered the deposit's dunite core. The chromite deposits at Dark Ridge were described further in Hunter et al. (1942). Their map showed the location of the chromite prospect that occurs near the western border of the body, north of Dark Ridge Creek (see Fig. 2). The chromite is in a vein-like lens up to 2 ft thick and is thought to be related to a fault zone. Hunter et al. (1942) indicated that additional chromite veins were exposed in olivine ledges above the creek.

A petrofabric study of the Dark Ridge and nearby Balsam Gap metaultramafic bodies was undertaken by Astwood et al. (1972). These authors found that only a few olivine grains exhibit undulatory extinction and concluded that there is no preferred orientation shown by the olivine in either body. They interpreted the lack of preferred orientation to indicate crystallization or recrystallization in a low-strain environment and inferred initial emplacement as serpentine followed by thermal dehydration during regional metamorphism to form olivine. Dribus et al. (1976) studied the petrography of the Dark Ridge, Balsam Gap, and several other metaultramafic bodies in western North Carolina and observed that, although dominated by equant, polygonal grains of unstrained olivine, all of the bodies possessed deformational textures as indicated by the presence of strained olivine porphyroclasts. The porphyroclasts exhibit undulatory extinction and are often marked by strain bands. In contrast to Astwood et al. (1972), Dribus et al. (1976) inferred diapiric emplacement of upper mantle material that was recrystallized during or prior to ascent. According to these authors, hydrous alteration (producing serpentine, talc, chlorite, etc.) occurred during emplacement or in one or more post-emplacement events. Neither the Astwood et al. (1972) nor the Dribus et al. (1976) study mentions orthopyroxene-bearing

Figure 3. Outcropping of massive metaultramafics on slopes above Dark Ridge Creek.

Figure 4. Biotite gneiss exposure along Dark Ridge Road, west of Dark Ridge metaultramafic body.

rocks as being present at Dark Ridge; however, both indicate that talc is quite abundant (up to 19 % in modes reported in Dribus et al., 1976).

More recently a radiometric and a magnetic survey were conducted (Schiering, 1979; Schiering et al., 1982) to more accurately map the areal extent of the Dark Ridge metaultramafic body. Only a slight magnetic susceptibility contrast occurs between the metaultramafics and gneissic country rock; consequently, there is no magnetic anomaly associated with the Dark Ridge body. Gamma-ray counts were discovered to be significantly lower within the metaultramafic body, however, compared to adjacent country rocks, hence the radiometric survey was effective in delineating the body's outline. The results indicated that the body is crudely elliptical, measuring approximately 1000 ft by 1500 ft (Schiering, 1979).

FIELD SAMPLING AND ANALYTICAL METHODS

In this study samples were collected along several traverses across the Dark Ridge metaultramafic body (Fig. 2): (1) along the tracks of the Southern Railway in the northern part of the body; (2) along the ridge top near the middle of the body; and (3) south of the ridge, along a trail following the north side of Dark Ridge Creek. Good exposures of metaultramafic rock occur in several places on the north side of this trail and in the steep slopes above it (Fig. 3). In addition, samples were taken from outcrops below (south of) the ridge crest, from the aforementioned Dark Ridge chromite prospect (DR-10), and from both the northern end (DR-14) and southern end (DR-8 and DR-9, south of Dark Ridge Creek) of the Dark Ridge body (Fig. 2). Lastly, float samples were collected next to the railroad tracks (DR-2) and along the banks of Jones Creek near Dark Ridge Road (DR-15).

The outline of the Dark Ridge body shown in Fig. 2 was produced from the radiometric survey conducted by Schiering (1979). Generally, the contacts are poorly exposed. Thus, only one sample of country rock was collected, and that was of biotite gneiss outcropping in close proximity to DR-9, near the southern contact. Biotite gneiss seems to be the principal country rock as it is the dominant lithology exposed in roadcuts along Dark Ridge Road, about 300 meters to the west (Fig. 4). A large amphibolite layer occurs east of the Dark Ridge metaultramafic body, just west of the Soque River fault (Hatcher, personal communication).

Polished thin sections were prepared from the hand samples; these were examined with a Jenapol U polarizing microscope, using both transmitted and incident light. Modal abundances were obtained with a Swift Model F automatic point counter (average number of points counted per section was slightly less than 500). Mineral compositions were determined by analyzing with a JEOL JXA-8600 scanning electron microprobe at the University of Georgia. Analyses were made using a 15 nanoamp beam current at an accelerating voltage of 15 KV, and were corrected for differential matrix effects using the procedure of Bence and Albee (1968). Chromite analyses were recalculated on the basis of stoichiometry to determine Fe^{3+} and Fe^{2+} from total Fe.

Whole-rock geochemical analyses of selected samples were made at Acme Analytical Laboratories Ltd. in Vancouver, British Columbia. The analyses were obtained via inductively coupled plasma emission spectrometry (ICP-es) on 0.2 g samples following a $LiBO_2$ fusion and dilute nitric acid digestion. Loss on ignition (LOI) was determined by weight difference after ignition to 1000˚C.

METAULTRAMAFIC ASSEMBLAGES

Petrographic examination of the samples collected in this study reveals that all are clearly metamorphic rocks; that is, they have metamorphic textures and their mineral assemblages are those characteristic of metamorphosed ultramafic rocks (Evans, 1977). Therefore, metamorphic rock names are used in describing and classifying the Dark Ridge metaultramafic rocks (Table 1). Considerable variety is manifested, with at least a half dozen different rock types distinguishable based on (1) primary mineralogy and (2) relative degree of hydration.

Two populations of primary mineral assemblages are represented at Dark Ridge: (1) rocks in which only olivine and chromite were primary minerals; and (2) rocks containing the primary assemblage olivine + orthopyroxene + chromite (the so-called saxonites of Hunter, 1941). Where dominantly anhydrous, (1) is typically olivine-rich and thus classed as metadunite. Where chromite seams are encountered, as seen at the Dark

Table 1. Modal abundances in Dark Ridge metaultramafic rocks.

Sample	Olivine	Chromite	Opx	Calcic Amphibole	Talc	Chlorite	Serpentine	Carbonate
Metadunite								
DR-1C	97	1		Tr	Tr	1	<1	
DR-2	96	<1		1	<1	<1	2	
DR-3	97	1			<1	1	<1	
DR-4	97	<1			<1	<1	1	<1
Carbonated Metadunite								
DR-6	75	<1			4	2	1	16
Serpentinized Metadunite								
DR-7B	60	<1		3	3	2	32	
DR-7D	62	<1		4	Tr	2	31	1
DR-14	46	<1		Tr	9	1	43	<1
DR-15	65	<1		4	3	5	23	
Metaharzburgite								
DR-1	59	<1	7	15	6	5	7	
DR-1A	64	<1	10	11	<1	4	11	
DR-1W	86	<1	2	4	Tr	4	3	1
Steatitized Metaharzburgite								
DR-1B	77	<1		5	14	3	1	
DR-3A	75	<1	Tr	3	16	4	1	
DR-3B	80	<1	Tr	2	10	6	2	
DR-5	78	<1	Tr	9	7	4	2	
DR-5B	72	Tr	Tr	8	11	6	4	
DR-7W	60	<1	Tr	11	17	9	3	
DR-7A	60	<1	Tr	15	12	8	4	
DR-7C	84	<1		4	5	3	4	
DR-11	82	Tr	Tr	3	7	2	6	<1
DR-12	70	<1	Tr	5	16	5	4	
DR-13	71	<1		8	12	6	3	
Serpentinized Metaharzburgite								
DR-1D	63	<1	Tr	12	8	5	11	1
DR-7E	67	<1	Tr	3	8	2	20	<1
DR-8	43	2	Tr	6	18	5	24	2
Chlorite Schist								
DR-9*						99		

*Contains 1 % ilmenite

Ridge chromite prospect, (1) grades into olivine metachromitite. The latter rocks are very heterogeneous, consisting of alternating chromite-rich and olivine-rich layers (Fig. 5). Serpentine is the chief mineral introduced during hydration of metadunite at Dark Ridge, giving rise to serpentinized metadunite.

The second group of rocks contains (or contained) primary orthopyroxene. Metaharzburgite is the name given to orthopyroxene-bearing rocks in which the amount of orthopyroxene is 10 percent or more. According to this definition, only one sample (DR-1A) is a metaharzburgite (Table 1). Orthopyroxene commonly is replaced by talc (discussed later), with the result that most of the talc now present in the Dark Ridge samples was derived from primary orthopyroxene. More than one third of the samples collected in this study contain $\geq$ 10 % talc (Table 1), so it is presumed that they were originally metaharzburgite that was subsequently hydrated (steatitized). In addition to steatitized metaharzburgite, several talc-bearing samples were moderately serpentinized.

More hydrated rocks locally occur around the periphery of the Dark Ridge metaultramafic body. Chlorite schist was found near the southern contact (sample DR-9, Fig. 1). In several

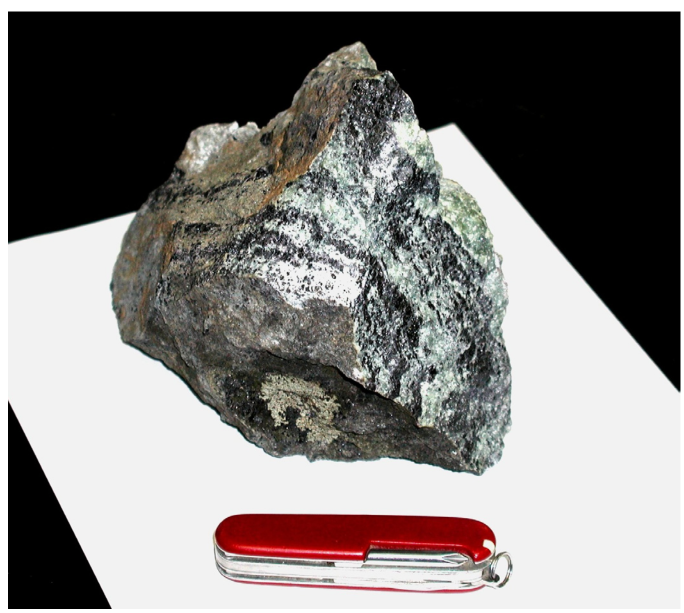

Figure 5. Sample of olivine metachromitite from chromite prospect north of Dark Ridge Creek (sample locality, DR-10).

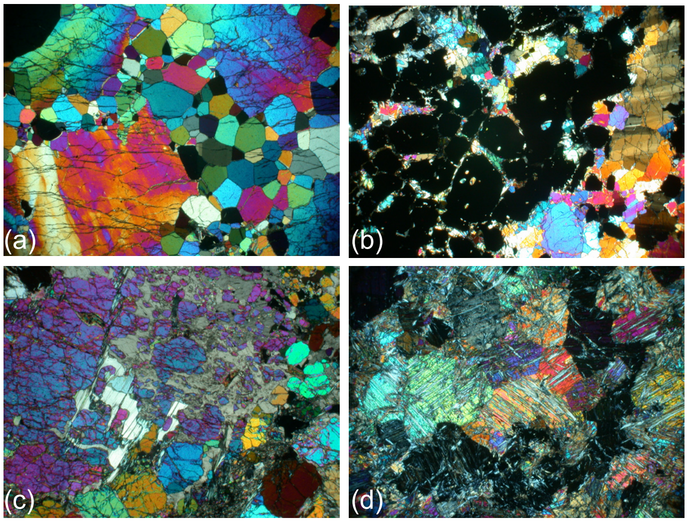

Figure 6. Photomicrographs (X nicols) illustrating metadunitic rock types. All are at the same scale, with long dimension of field of view ~7 mm. (a) Metadunite (DR-4) showing large strained olivine porphyroclasts (undulatory extinction) separated by finer-grained olivine with granoblastic polygonal texture. (b) Olivine metachromitite (DR-10) consisting of rounded chromite grains (dark under X nicols) and anhedral olivine (mostly strained). (c) Carbonated metadunite (DR-6) with olivine partially replaced by magnesite (M). Note antigorite (A) pseudomorph after chlorite (Chl). (d) Serpentinized metadunite (DR-7D) showing mosaic of polygonal olivine extensively replaced by antigorite along cleavage planes.

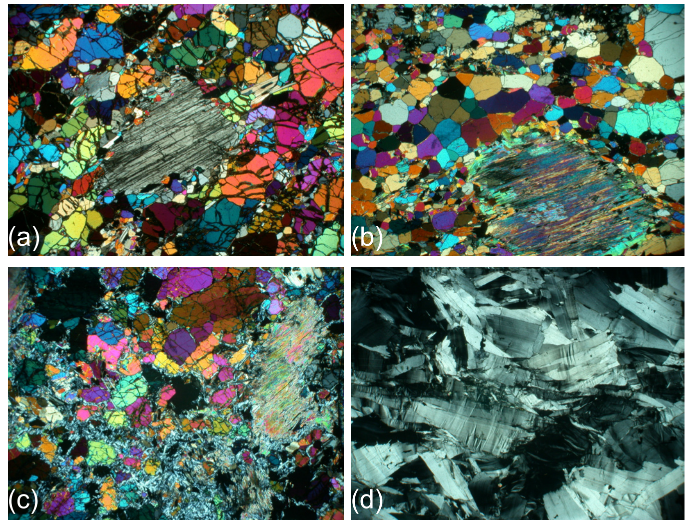

Figure 7. Photomicrographs (X nicols) illustrating metaharzburgitic rock types and chlorite schist. All are at same scale as in Figure 6. (a) Metaharzburgite (DR-1A) showing anhedral orthopyroxene porphyroclast (Opx) in a matrix of subequant olivine. (b) Steatitized metaharzburgite (DR-5B) containing large talc clot (T) surrounded by polygonal olivine. (c) Serpentinized metaharzburgite (DR-8) with olivine in a matrix of fine-grained antigorite; talc clot (at right) is partially replaced by antigorite. (d) Chlorite schist (DR-9) consisting almost entirely of intergrown chlorite laths, some showing kink banding.

places near the eastern contact serpentinite float is present, but no thin sections were made.

The various metaultramafic lithologies identified above are described in more detail below. Representative photomicrographs illustrating the mineralogy and textures encountered in the Dark Ridge rocks are shown in Figs. 6 and 7. Also, whole rock bulk analyses of selected samples are given in Table 2.

Metadunite. These rocks have the highest MgO and NiO and lowest Al_2O_3 (Table 2). They are essentially anhydrous, containing less than 4 percent hydrous minerals (Table 1). Olivine occurs both as recrystallized grains with granoblastic polygonal texture and as larger porphyroclasts (Fig. 6A). The porphyroclasts are strained and show undulatory extinction. Olivine crystals as large as 3.5 cm are present in sample DR-4, but even larger grains (up to ~15 cm) were reported by Hunter (1941). Unstrained polygonal olivines have an average size between 0.5 mm and 1 mm, but individual crystals may range down to 0.1-0.2 mm. They often are bounded by 120˚ triple junctions (Fig. 6A). Also present are olivine grains with curved boundaries and containing strain bands; these crystals are usually larger than the polygonal olivines and likely indicate incomplete recrystallization (Ragan, 1969; Dribus et al., 1976). Overall, olivine is quite fresh, with only minor alteration to mesh-textured serpentine (along fractures and grain boundaries). Metadunite samples contain one percent or less chromite, both as inclusions in olivine porphyroclasts and as interstitial crystals in polygonal olivines; chlorite laths are often associated with chromite crystals.

Olivine metachromitite. Chromite-rich samples from the Dark Ridge chromite prospect (DR-10) have a banded appearance (Fig. 5). Black-colored bands containing abundant chromite alternate with green olivine-rich bands. Thin sections were made of two olivine metachromitites, both too heterogeneous to point count. In one, chromite occurs as opaque clumps up to 6 mm long set in a confused mass of chlorite, serpentine, talc, and magnesite; part of the sample is olivine-rich with porphyroclasts up to 6 mm in diameter set in a matrix of 0.3-1 mm polygonal olivine. The second thin section consists of 1-4 mm anhedral crystals of deep reddish-brown chromite, variously zoned to nearly opaque rims, which are intergrown with similar-sized olivine (Fig. 6B). Many of the olivine crystals are strained (undulatory extinction or strain banding) and have irregular outlines; polygonal texture is mostly restricted to finer-grained areas with unstrained olivine. Chlorite laths are plentiful at chromite grain boundaries. Magnesite-serpentine veins cut both samples. A whole-rock analysis of olivine metachromitite shows that it is Al_2O_3-rich as well as Cr_2O_3-rich (Table 2), suggesting the chromite contains significant Al. Note also the higher carbon content, reflective of the magnesite present.

Carbonated metadunite. Sample DR-6, collected about 15 m southeast of the Dark Ridge chromite prospect (Fig. 2), is a metadunite containing abundant carbonate (Table 1). Magnesite is the principal carbonate mineral, with lesser dolomite. Olivine occurs as subequant porphyroclasts up to 1.5 cm long and as aggregates of smaller ($\leq$2 mm), equant crystals with poorly developed polygonal texture (Fig. 6C). Often the olivine crystals are rimmed by carbonate or, less commonly, talc. Widely scattered chromite crystals typically are partly or wholly enclosed in chlorite and/or magnesite. Some chlorite is pseudomorphously replaced by antigorite (Fig. 6C). Several magnesite-rich veins cut through the rock; these veins contain talc and antigorite as well.

Table 2. Whole-rock chemical analyses of Dark Ridge metaultramafic rocks.

	Metadunite		Olivine Metachromitite	Serpentinized Metadunite		Metaharzburgite	Steatitized Metaharzburgite			Serpentinized Metaharzburgite
	DR-1C	DR-4	DR-10*	DR-7D	DR-15	DR-1A	DR-5	DR-7A	DR-11	DR-8
SiO2 (wt%)	40.31	40.14	25.04	41.79	41.48	43.51	42.98	43.97	41.77	41.04
Al2O3	0.40	0.17	7.28	1.00	1.06	1.52	1.26	1.37	0.74	0.89
Cr2O3	0.37	0.51	15.37	0.41	0.38	0.40	0.39	0.37	0.35	0.34
FeO	8.49	7.15	11.71	9.06	8.27	8.00	7.94	7.71	7.85	7.95
MgO	48.45	49.46	35.96	40.47	43.46	40.81	42.99	43.57	45.13	43.82
MnO	0.11	0.09	0.14	0.11	0.12	0.11	0.11	0.11	0.11	0.11
NiO	0.36	0.38	0.30	0.29	0.31	0.33	0.30	0.30	0.31	0.30
CaO	0.29	0.11	0.05	0.20	0.56	1.63	1.28	1.26	0.28	0.64
Na2O	0.03	0.02	0.01	0.01	0.05	0.16	0.09	0.08	0.02	0.03
P2O5	0.02	0.02	0.03	0.02	0.01	0.03	0.03	0.02	<0.01	0.02
Total C	0.06	0.07	0.27	0.07	0.13	0.13	0.04	0.03	0.12	0.19
LOI	0.3	1.15	2.4	5.7	3.4	2.6	1.8	0.4	2.6	4.0
Sum	99.19	99.27	98.56	99.13	99.23	99.23	99.21	99.19	99.28	99.33
Mg#	91.0	92.5		88.8	90.4	90.1	90.6	91.0	91.1	90.8

*Includes 0.08 wt% TiO2 Mg# = 100 x Mg/(Mg+Fe2+)

Serpentinized metadunite. Hydrous minerals, dominantly serpentine, make up 35-53 % of these samples (Table 1). Talc, chlorite and calcic amphibole are present in various amounts and, combined, total 6-12 % of the modes. Serpentine (antigorite) most commonly replaces olivine along (010) cleavage planes; the resulting texture has the appearance of a lamellar intergrowth (Fig. 6D). Olivine grain size (average, 0.5-1 mm) is similar to that of polygonal olivine in anhydrous metadunite; although obscured by serpentinization, granoblastic polygonal textures with olivine grains meeting at 120˚ triple junctions are common. Olivine porphyroclasts were not observed in serpentinized metadunite samples. Chromite (<1 %) occurs as irregular crystals contained within chlorite lath aggregates; in some samples the enclosing chlorite is partially replaced by antigorite and/or talc. Serpentinized metadunite has lower MgO, and consequently a lower Mg-number, than anhydrous metadunite (Table 2).

Metaharzburgite. Metaharzburgite contains scattered orthopyroxene porphyroclasts (Fig. 7A) in a matrix of olivine, chromite, calcic amphibole, chlorite, and talc (Table 1). Sample DR-1W, despite containing only 2 percent orthopyroxene, is included in the metaharzburgite group because it is mineralogically (presence of primary orthopyroxene) and texturally more akin to this group than to metadunite. Rocks containing visible orthopyroxene porphyroclasts only outcrop along the railroad tracks in the northern part of the body (Fig. 2). Porphyroclasts range up to ~1 cm and stand out prominently in hand specimens because of their bronze-like luster. In thin section the orthopyroxenes are seen to contain tiny, oriented rod-like inclusions (schiller structure), which are responsible for the distinctive luster. Individual crystals show varying degrees of replacement by talc. Trace amounts of orthopyroxene also occur as small, rounded inclusions in olivine. Olivine crystals are mostly equant, and have a granoblastic polygonal texture, though not as well developed as in metadunite. They range from 0.2 mm to ~ 4 mm, with some larger grains exhibiting strain banding; olivine porphyroclasts are lacking, however. Serpentine alteration has taken place along olivine grain boundaries and along internal olivine fractures. Chromites are sparsely disseminated; most grains are enclosed by chlorite, but tiny, isolated inclusions in olivine plus minute lamellar inclusions in orthopyroxene also occur. Calcic amphibole constitutes up to 15 % of the mode (Table 1), resulting in higher bulk-rock CaO (Table 2). Calcic amphiboles are concentrated in areas with finer-grained olivines where they occur as bladed crystals ≤ 1 mm long (rarely up to 2 mm); basal sections are typically 0.25-0.5 mm across. Chlorite is present as widely scattered laths, usually between 0.5 mm and 1 mm long, and as aggregates of shorter laths that surround chromite. Talc is common in DR-1 where it mostly occurs around the margins of, and replacing, orthopyroxene crystals.

Steatitized metaharzburgite. In appearance these rocks are similar to metaharzburgite, but with the exception that orthopyroxene porphyroclasts are absent. Instead, large (3-8 mm) talc clots are peppered through the samples (Fig. 7B). The talc clots closely resemble, both in size and shape, orthopyroxene porphyroclasts in metaharzburgite and, indeed, were derived from them. Talc also occurs as isolated laths or clusters of laths scattered throughout the rocks: these likely formed by replacement of olivine or, in some cases, calcic amphibole. Olivine in steatitized metaharzburgite generally has a granoblastic polygonal texture; most crystals are unstrained, but a few have strain bands. The majority of crystals are between 0.3 and 1.0 mm, but range up to 2-4 mm. Porphyroclasts of olivine (maximum size ~6 mm) were observed only in DR-3A; they invariably exhibit undulatory extinction. Most samples of steatitized metaharzburgite contain traces of orthopyroxene (Table 1) as remnant inclusions in olivine. Crystal habits of chromite, chlorite and calcic amphibole are virtually the same as described above for metaharzburgite. Serpentine alteration has taken place to varying degrees, mostly along olivine grain boundaries or fractures in olivine grains (mesh texture), but antigorite replacement of olivine along (010) cleavage planes or of chlorite laths occurs as well.

Serpentinized metaharzburgite. Some metaharzburgite is characterized by more extensive serpentinization (Table 1). Serpentinized metaharzburgite contains comparable amounts of hydrous minerals (33-53 %) as serpentinized metadunite. However, the occurrence of serpentine differs: in serpentinized metadunite it primarily is present as antigorite laths oriented along (010) planes in olivine (Fig. 6D), but in serpentinized metaharzburgite antigorite tends to form a patchy fine-grained matrix that surrounds olivine (Fig. 7C). Disseminated magnetite is often associated with the latter serpentine. Talc

Table 3. Representative electron microprobe analyses of olivine in Dark Ridge metaultramafic rocks.

	Olivine Chromitite	Metadunite		Metaharzburgite	Steatitized Metaharzburgite		Serpentinized Metaharzburgite	Serpentinized Metadunite	
	DR-10	DR-4	DR-1C	DR-1A	DR-3A	DR-5	DR-8	DR-7B	DR-14
SiO2 (wt%)	41.06	41.01	40.73	40.84	40.93	40.43	40.51	39.99	39.51
MgO	53.17	51.75	51.08	49.63	49.61	49.00	48.53	47.84	46.33
FeO	4.98	6.86	7.9	8.98	9.47	9.72	10.21	11.83	13.74
MnO	0.11	0.12	0.11	0.18	0.16	0.17	0.11	0.23	0.21
NiO	0.60	0.43	NA	NA	0.39	0.40	0.47	0.38	0.32
Total	99.92	100.17	99.82	99.63	100.56	99.72	99.83	100.27	100.11
Si	0.991	0.994	0.991	1.000	0.999	0.997	1.000	0.991	0.989
Mg	1.913	1.869	1.853	1.812	1.805	1.801	1.786	1.766	1.729
Fe	0.101	0.139	0.161	0.184	0.193	0.201	0.211	0.245	0.288
Mn	0.002	0.003	0.002	0.004	0.003	0.004	0.002	0.005	0.005
Ni	0.012	0.008			0.008	0.008	0.009	0.008	0.006
Sum	3.019	3.013	3.007	3.000	3.008	3.011	3.008	3.015	3.017
Fo (%)	95.0	93.1	92.0	90.8	90.3	90.0	89.4	87.8	85.7

NA = not analyzed

clots are present in steatitized harzburgite, too, only here they are often partially replaced by antigorite. Antigorite also frequently replaces chlorite. Olivine habit and grain size variation are similar to those in metaharzburgite and steatitized metaharzburgite; so also are chromite, calcic amphibole, and chlorite. Traces of orthopyroxene are included in olivine. Carbonate (chiefly magnesite) is slightly more abundant in serpentinized metaharzburgite than in the other metaultramafic rock types (Table 1), and is usually associated with serpentine.

Chlorite schist. Sample DR-9 was collected south of Dark Ridge Creek near the southern contact of the metaultramafic body (Fig. 2). It is a coarse-grained, schistose rock composed almost wholly of chlorite (Table 1). The chlorite has undulatory extinction and frequently contains kink bands (Fig. 7D). Small anhedral ilmenite crystals are widely disseminated throughout the chlorite schist.

MINERAL COMPOSITIONS

Olivine. Olivine grains are unzoned and typically show a very narrow compositional range within individual thin sections, usually less than one % Fo, with porphyroclasts and polygonized crystals showing no discernible compositional difference. For the Dark Ridge metaultramafic body as a whole, there is about a 10 mol % variation in olivine composition (Table 3; Fig. 8). Within the specific rock types described in the preceding section, the variation is much less (2-3 % Fo). The most magnesian olivine occurs in olivine metachromitite and metadunite, while the most Mg-poor olivine is found in the serpentine-rich rocks. Metaharzburgite displays a slightly more restricted range in olivine composition relative to metadunite (Fig. 8).

Chromite. Chromite exhibits a much wider compositional variation than olivine or, indeed, any silicate with which it is associated. In large part this is related to its textural variety, which ranges from euhedral "clean" (Lipin, 1984) chromite with

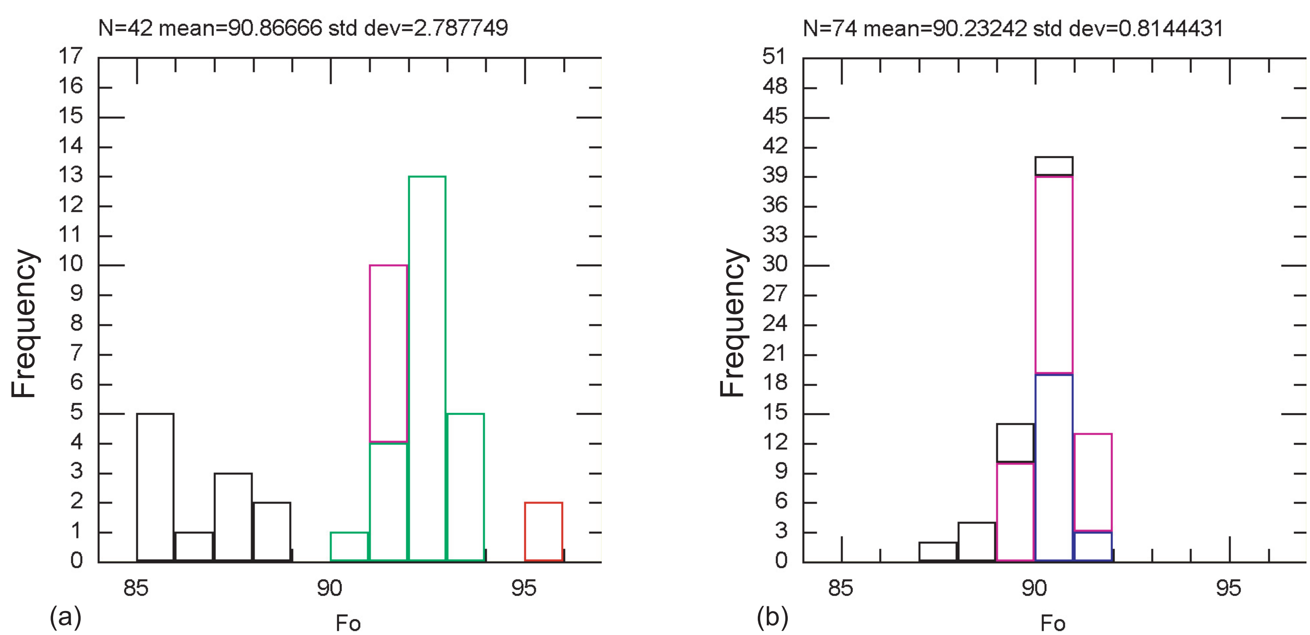

Figure 8. Histograms of Fo content of Dark Ridge olivine. Top, metadunitic samples: green, metadunite; red, olivine metachromitite; pink, carbonated metadunite; black, serpentinized metadunite. Bottom, metaharzburgitic samples: pale purple, metaharzburgite; bright pink, steatitized metaharzburgite; black, serpentinized metaharzburgite.

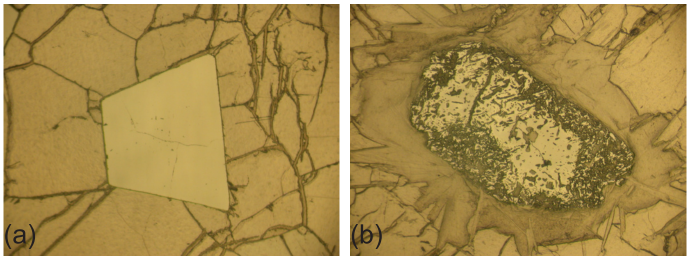

Figure 9. Photomicrographs (reflected light) illustrating typical chromite textures in Dark Ridge samples. Both are at the same scale (long dimension of field of view ~1 mm). (a) Euhedral, clean chromite associated with polygonal olivine, metadunite (DR-4). (b) Poikiloblastic chromite enclosed by chlorite laths, steatitized metaharzburgite (DR-1B).

little or no associated chlorite (Fig. 9A) to anhedral poikiloblastic and/or lattice chromite that is entirely surrounded by chlorite (Fig. 9B). The former is most common in anhydrous metadunite whereas the latter is prevalent in hydrated metaultramafic rocks. In addition, chromite crystals are often compositionally zoned. Compositional data for chromite (Table 4; Figs. 10-12) indicate the following: (1) the most Mg-rich chromite is from the chromite prospect (DR-10); (2) chromite in metaharzburgite and metadunite samples also tends to have higher Mg and is the most Al-rich; (3) chromite in hydrated (steatitized and/or serpentinized) samples has lower Mg and is characterized by high Cr-numbers (= 100 x Cr/(Cr+Al)) and also by higher Fe^{3+}; and (4) the trend in zoned crystals is for Cr-number to increase and Mg-number (=100 x Mg/(Mg+Fe^{2+})) to decrease from the core outward to the rim.

Table 4. Representative electron microprobe analyses of chromite in Dark Ridge metaultramafic rocks.

	Olivine Metachromitite	Metaharzburgite			Metadunite		Steatitized Metaharzburgite			Serpentinized Metadunite
	1	2	3	4	5	6	7	8	9	10
SiO2 (wt%)	0.03	BDL	BDL	BDL	BDL	0.07	0.03	0.05	0.03	0.03
TiO2	0.21	0.03	BDL	0.04	0.04	BDL	0.10	0.07	0.25	0.21
Al2O3	17.90	30.99	25.03	7.11	25.18	7.24	1.26	10.06	2.95	1.42
Cr2O3	50.66	35.32	42.07	58.83	40.62	61.21	59.23	51.39	54.06	55.52
Fe2O3	2.07	1.31	1.26	1.86	3.40	0.61	7.36	5.54	11.03	9.93
FeO	15.25	19.99	20.38	25.75	18.73	23.48	28.39	26.12	27.82	29.10
MnO	0.20	BDL	BDL	BDL	0.14	BDL	BDL	BDL	BDL	BDL
MgO	12.72	11.08	9.96	4.72	10.60	6.17	2.57	4.48	2.97	1.71
Total	99.04	98.72	98.70	98.31	98.71	98.78	98.94	97.71	99.11	97.92
Si	0.001					0.002	0.001	0.001	0.001	0.001
Ti	0.005	0.001		0.001	0.001		0.003	0.002	0.007	0.006
Al	0.668	1.111	0.926	0.296	0.928	0.297	0.055	0.417	0.127	0.063
Cr	1.268	0.849	1.044	1.644	1.004	1.683	1.723	1.430	1.556	1.644
Fe3+	0.053	0.039	0.030	0.058	0.079	0.016	0.215	0.147	0.302	0.280
Fe2+	0.401	0.499	0.534	0.753	0.491	0.683	0.863	0.769	0.847	0.912
Mn	0.005				0.004					
Mg	0.600	0.502	0.466	0.249	0.494	0.320	0.141	0.235	0.161	0.095
Sum	3.001	3.001	3.000	3.001	3.001	3.001	3.001	3.001	3.001	3.001
Cr#	65.5	43.3	53.0	84.7	52.0	85.0	96.9	77.4	92.5	96.3
Mg#	59.9	50.1	46.6	24.9	50.2	31.9	14.0	23.4	16.0	9.4

Cr# = 100 x Cr/(Cr+Al) Mg# = 100 x Mg/(Mg+Fe2+)

Columns: 1 - chromite from chromite-rich layer, DR-10; 2 - small, euhedral chromite inclusion in olivine, DR-1W; 3,4 - core and rim of zoned poikiloblastic chromite, DR-1W; 5 - core of zoned chromite, DR-1C; 6 - euhedral chromite surrounded by olivine, DR-4; 7 - small, irregular chromite with associated pentlandite, DR-7C; 8,9 - core and rim of zoned poikiloblastic chromite, DR-11; 10 - poikiloblastic chromite, DR-14

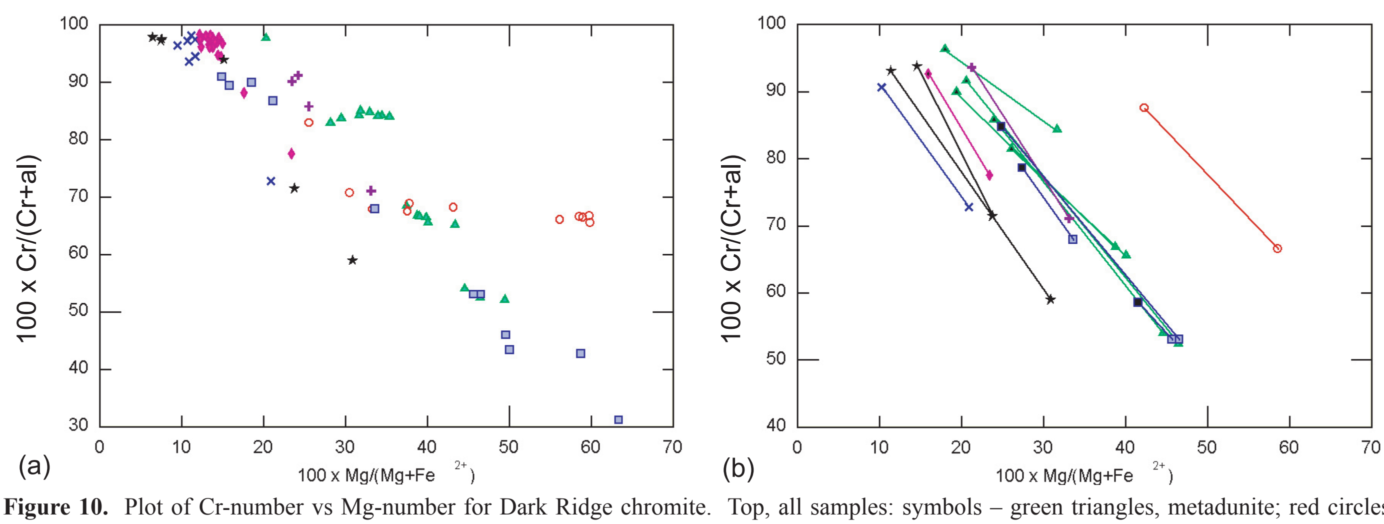

Figure 10. Plot of Cr-number vs Mg-number for Dark Ridge chromite. Top, all samples: symbols – green triangles, metadunite; red circles, olivine metachromitite; purple crosses, carbonated metadunite; blue 45˚ crosses, serpentinized metadunite; pale purple squares, metaharzburgite; pink diamonds, steatitized metaharzburgite; black stars, serpentinized metaharzburgite. Bottom, zoned crystals: filled symbols, cores; open symbols, rims; rock designations same as above.

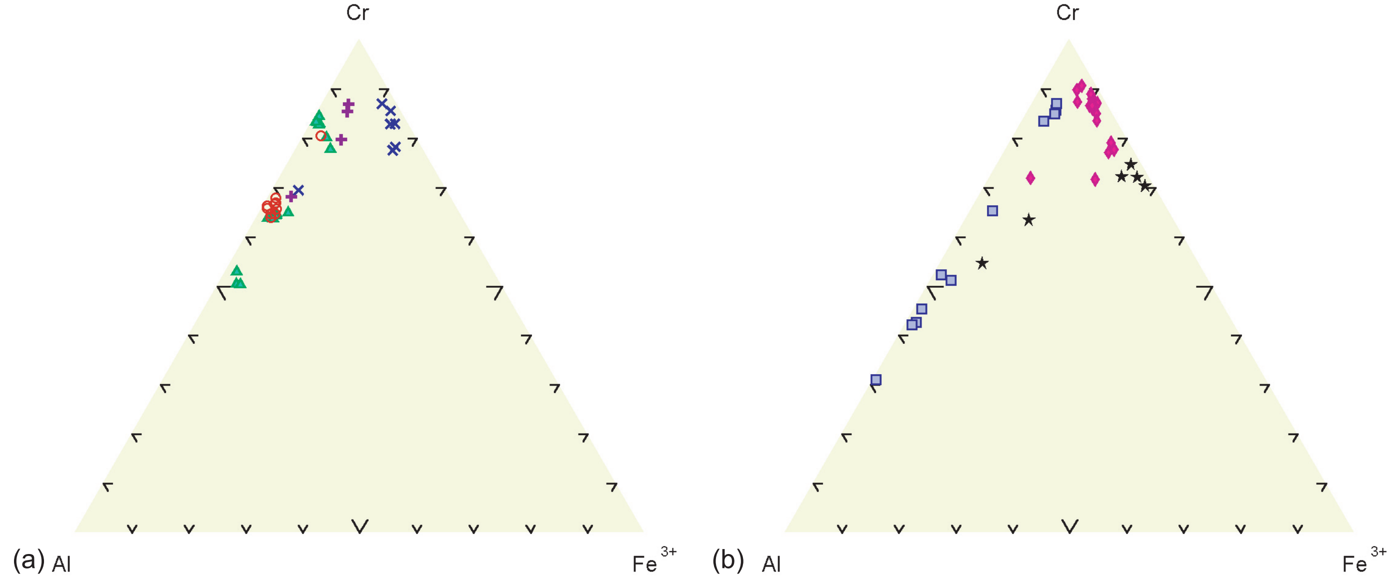

Figure 11. Compositions of Dark Ridge chromite plotted on ternary Al-Cr-Fe^{3+} diagram. Top, metadunitic samples; bottom, metaharzburgitic samples. Symbol designations are same as in Figure 10.

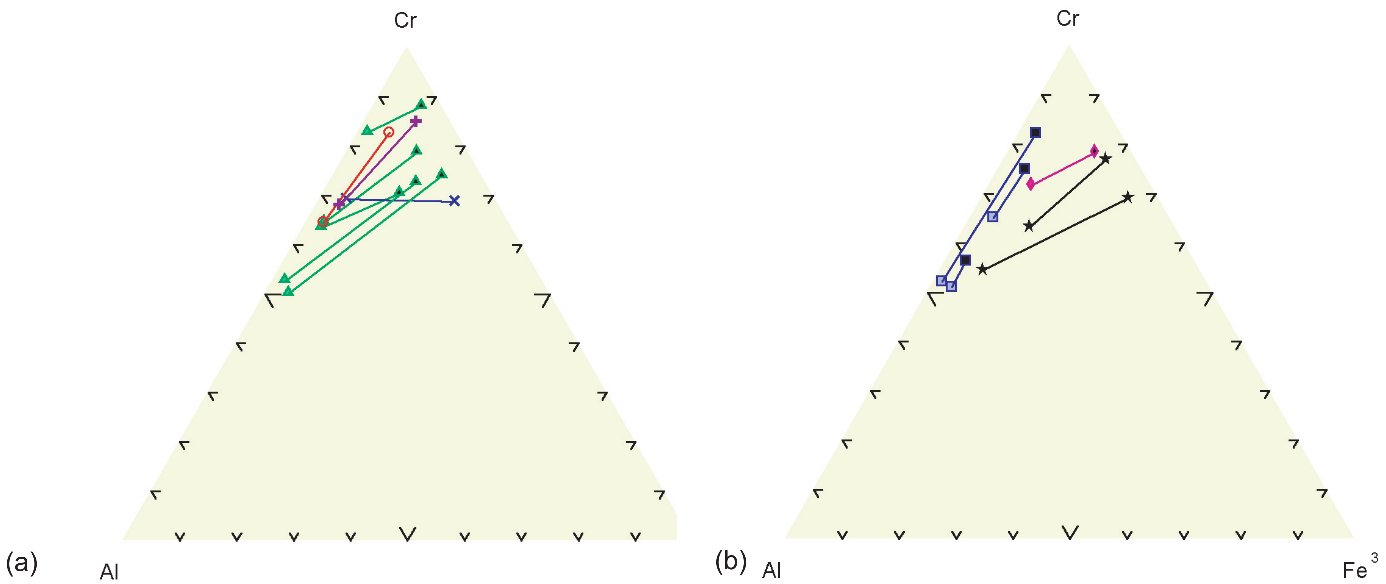

Figure 12. Core (filled symbols) – rim (open symbols) compositions of zoned chromite plotted on ternary Al-Cr-Fe^{3+} diagram. Symbol designations same as in Figure 10.

Orthopyroxene. Orthopyroxene primarily occurs as porphyroclasts in metaharzburgite (Fig. 7A). The porphyroclasts contain oriented rod-like inclusions of Al-rich chromite, giving rise to schiller structure. Extremely fine lamellae of calcic pyroxene (identified from EDS spectra) occur in orthopyroxene porphyroclasts in sample DR-1W. Serpentine alteration oriented parallel to the schiller inclusions is common. Orthopyroxene is preserved in steatitized metaharzburgite (e.g., DR-11) only where it occurs as inclusions (small) in olivine. Orthopyroxene exhibits a very narrow range in composition with no apparent difference between porphyroclasts and small inclusions in olivine (Table 5); Mg-numbers are virtually identical to that of coexisting olivine.

Calcic amphibole. Calcic amphibole is a common secondary mineral found in the metaultramafic rocks at Dark Ridge; it is particularly abundant in the various types of metaharzburgite (Table 1). Two populations of amphibole compositions occur (Table 5): tremolitic hornblende with 4-7 wt% Al_2O_3, and nearly pure tremolite with higher Mg-numbers. Frequently, crystals are inhomogeneous with tremolitic hornblende cores and tremolite rims. Anthophyllite is conspicuously absent; only one crystal was found in the 30 thin sections examined.

Chlorite. Chlorite is the variety chlinochlore and mainly occurs as aggregates of small lath-shaped crystals that surround chromite (Fig. 9B). Such chlorite contains measurable Cr, usually in excess of 1 wt% Cr_2O_3 (Table 6). More rarely, larger platy crystals are present, often enclosing irregular chromite grains. Chlorite in samples from the Dark Ridge chromite prospect contains the highest amount of Cr (nearly 3 wt%, comparable to the Cr_2O_3 concentration reported for chromian

Table 5. Representative electron microprobe analyses of pyroxene and amphibole in Dark Ridge metaultramafic rocks.

	Orthopyroxene			Calcic Amphibole				
	DR-1A	DR-1W	DR-11	DR-1A	DR-1W	DR-12*		DR-14
SiO2 (wt%)	56.11	56.29	56.65	53.42	53.85	51.04	58.83	58.27
Al2O3	1.81	1.61	1.03	6.45	5.05	7.16	0.08	0.15
Cr2O3	0.24	0.23	0.25	NA	NA	NA	NA	NA
MgO	34.76	35.30	34.88	21.36	22.26	20.01	22.69	23.82
FeO	6.37	6.14	5.94	2.44	1.98	2.47	1.37	1.91
MnO	0.21	0.14	0.11	BDL	BDL	BDL	0.08	0.05
CaO	0.18	0.18	0.17	12.68	13.29	12.92	13.67	13.46
Na2O	BDL	BDL	BDL	0.86	0.63	1.87	0.08	0.16
K2O	BDL	BDL	BDL	0.16	0.08	0.18	BDL	BDL
Total	99.68	99.89	99.03	97.37	97.14	95.65	96.80	97.82
Oxygens	6	6	6	23	23	23	23	23
Si	1.943	1.943	1.969	7.371	7.443	7.225	8.074	7.949
Al	0.074	0.066	0.042	1.049	0.823	1.195	0.013	0.024
Cr	0.007	0.006	0.007					
Mg	1.795	1.816	1.807	4.393	4.586	4.223	4.642	4.843
Fe	0.185	0.177	0.173	0.282	0.229	0.293	0.158	0.218
Mn	0.006	0.004	0.003				0.009	0.006
Ca	0.007	0.007	0.006	1.874	1.969	1.959	2.010	1.967
Na				0.229	0.168	0.514	0.020	0.043
K				0.028	0.014	0.032		
Sum	4.017	4.019	4.007	15.226	15.232	15.441	14.926	15.050
Ca (%)	0.3	0.3	0.3	28.6	28.0	30.3	29.5	28.0
Mg	90.4	90.8	91.0	67.1	68.7	65.2	68.2	68.9
Fe	9.3	8.9	8.7	4.3	3.3	4.5	2.3	3.1
Mg#	90.7	91.1	91.3	94.0	95.2	93.5	96.7	95.7

*Analyses are from core and rim of same crystal BDL = below detection limit NA = not analyzed

DR-1A, DR-1W = metaharzburgite; DR-11, DR-12 = steatitized metaharzburgite; DR-14 = serpentinized metadunite

Table 6. Representative electron microprobe analyses of sheet silicates in Dark Ridge metaultramafic rocks.

	Chlorite					Talc		Serpentine	
	DR-10	DR-1A	DR-7C	DR-8	DR-9	DR-3A	DR-1D	DR-14	DR-1D
SiO2 (wt%)	33.04	30.51	29.61	31.70	28.27	65.23	64.59	43.03	45.13
Al2O3	12.67	19.91	18.74	16.06	21.93	0.09	0.14	1.58	0.06
Cr2O3	2.74	1.07	1.12	1.23	BDL	BDL	0.05	0.95	0.10
FeO	1.05	2.83	3.56	3.35	7.65	0.82	1.37	4.52	5.83
MgO	34.79	32.45	32.24	33.40	28.95	30.82	30.87	36.10	37.33
NiO	0.21	0.28	0.21	0.24	0.05	NA	NA	0.10	NA
Total	84.50	87.05	85.48	85.98	86.85	96.96	97.02	86.28	88.45
Oxygens	14	14	14	14	14	11	11	7	7
Si	3.275	2.906	2.891	3.074	2.731	4.047	4.019	2.041	2.090
Al	1.480	2.235	2.157	1.835	2.497	0.007	0.010	0.089	0.003
Cr	0.314	0.118	0.127	0.138			0.003	0.036	0.004
Fe	0.087	0.225	0.291	0.272	0.618	0.043	0.071	0.179	0.226
Mg	5.140	4.608	4.691	4.828	4.169	2.849	2.864	2.552	2.577
Ni	0.017	0.021	0.017	0.019	0.004			0.004	
Sum	10.313	10.113	10.174	10.166	10.019	6.946	6.967	4.901	4.900
Mg#	98.3	95.3	94.2	94.7	87.1	98.5	97.6	93.4	91.9

BDL = below detection limit NA = not analyzed

DR-1A = metaharzburgite; DR-3A, DR-7C = steatitized metaharzburgite; DR-1D, DR-8 = serpentinized metaharzburgite; DR-9 = chlorite schist; DR-10 = olivine metachromitite; DR-14 = serpentinized metadunite chlorite in the Day Book metaultramafic body by Phillips et al., 1980). In contrast, chlorite in the chlorite schist (DR-9) does not contain detectable Cr; chlorite from this sample is also more Fe-rich.

Talc. One of the prominent petrographic features of the Dark Ridge metaultramafic rocks is the abundance of talc (up to 18 %, Table 1). Although isolated crystals of talc are not uncommon, the most distinctive occurrence is as large talc clots (Fig. 7B). Often these clots contain a schiller structure resembling that characteristic of orthopyroxene porphyroclasts. Thin section examination, confirmed by EDS spectra obtained with the electron microprobe, reveals that the schiller structure, like that in orthopyroxene, is due to minute chromite inclusions; lamellae of calcic amphibole, chlorite, and serpentine are oriented parallel to the schiller inclusions. Microprobe analyses of talc indicate that it is nearly pure hydrous Mg-silicate with higher Mg-numbers than any other metaultramafic mineral (Table 6).

Serpentine. Serpentine exhibits a variety of crystal habits. In serpentinized metadunite antigorite occurs as lamellar intergrowths oriented along (010) cleavage planes in olivine (Fig. 6D). In contrast, antigorite in serpentinized metaharzburgite forms a patchy, fine-grained matrix (Fig. 7C). The two are compositionally distinct (Table 6): antigorite in the first association typically contains about 1.5 wt% Al_2O_3 and up to 1 wt% Cr_2O_3, whereas in serpentinized metaharzburgite the concentrations of these oxides in antigorite are negligible. Talc clots in the latter rocks are partially replaced by antigorite and, in nearly all samples of metaultramafic rock, antigorite occasionally replaces chlorite pseudomorphously. Serpentine also occurs as either chrysotile and/or lizardite, formed by late-stage replacement of olivine along grain boundaries and internal fractures (mesh texture).

DISCUSSION

Contrasting conclusions have been reached about the distribution of hydrated rocks in the Dark ridge metaultramafic body. The map published by Hunter (1941) showed serpentinized dunite forming a peripheral border around a core of relatively unaltered olivine. Schiering (1979) found just the opposite – that hydrated rocks are more prevalent near the center of the body (i.e., along the top of the ridge separating Dark Ridge and Jones Creeks). Our sampling indicates that hydrated rocks occur throughout the entire body, interspersed with relatively anhydrous metadunite. Thus, serpentinized metadunite was found both near the northern contact (DR-14) and at two localities (DR-7B; DR-7D) towards the center of the Dark Ridge body (Fig. 2). The largest exposures of coarse, anhydrous metadunite occur in the vicinity of DR-4, along the trail north of Dark Ridge Creek, but are flanked by steatitized metaharzburgite (DR-3B and DR-5) containing 20-22 percent hydrous minerals. We suspect that the distribution of hydrated rocks is controlled by the presence of joints and small faults, which provided pathways for the access of hydrating fluids.

One of the major findings of this study is that hydrated metaharzburgite is fairly common in the Dark Ridge metaultramafic body: it was found along all three traverses plus south of the ridge (DR-11 ® DR-13; Fig. 2). Although Hunter

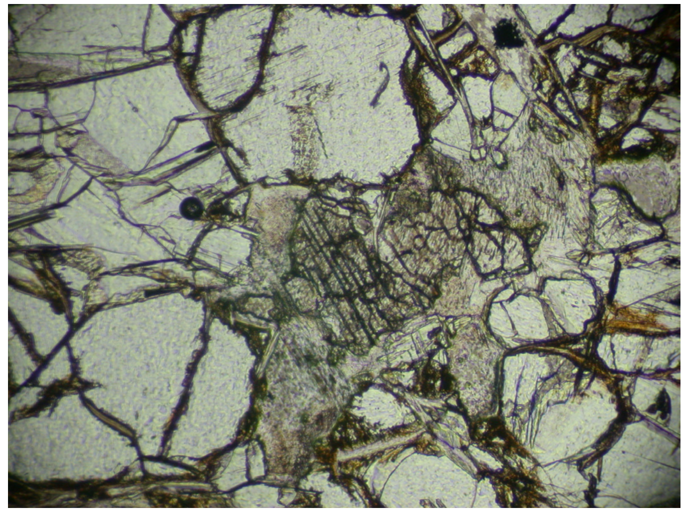

Figure 13. Photomicrograph (plane polarized light) of metaharzburgite (DR-1) showing talc (T) replacing orthopyroxene (Opx). Long dimension of field of view is ~1 mm.

(1941) reported that saxonite – rocks containing about 10 percent bronzite (orthopyroxene) crystals – form a significant portion of the Dark Ridge olivine deposit, subsequent investigators (Astwood et al., 1972; Dribus et al., 1976; Schiering, 1979; Schiering et al., 1982) failed to find primary orthopyroxene and, hence, referred to the body exclusively as the Dark Ridge dunite. These authors did note, however, that the body had been extensively steatitized, producing abundant talc. The talc was ascribed by Schiering et al. (1982) to have originated through alteration of olivine. Furman (1981), however, proposed that alteration of primary bronzite was the source of the abundant talc at Dark Ridge as well as in several smaller metaultramafic bodies less than 2 km distant.

Petrographic evidence indicates that much of the talc in the Dark Ridge metaultramafic body had an origin similar to that proposed by Furman (1981). In one sample (DR-1) talc clearly replaces orthopyroxene (Fig. 13). Also, talc clots in thin section appear to pseudomorph orthopyroxene porphyroclasts (Fig. 7A,B); they are similar in size and shape and possess a schiller structure that closely resembles the schiller structure in orthopyroxene. Oriented inclusions of chromite occur in both talc and orthopyroxene and are, in part, responsible for the schiller effect seen in the two minerals. The similarity in appearance of talc and orthopyroxene crystals extends to field outcrops: Pratt and Lewis (1905) stated that it was very difficult to distinguish between them in hand specimen. We conclude, therefore, that orthopyroxene was originally present in the now steatitized metaharzburgite, but was subsequently converted to talc.

Another characteristic feature of the Dark Ridge metaultramafic rocks is the presence of secondary calcic amphibole (both tremolite and tremolitic hornblende), which is most abundant in metaharzburgite assemblages (Table 1). This is reflected by higher CaO in metaharzburgite whole-rock analyses (Table 2). Unless calcium was added during hydration and metamorphism of the rocks (Scotford and Williams, 1983), a primary Ca-bearing mineral must have been present. Fine-scale exsolution lamellae (too small for quantitative analysis) of calcic clinopyroxene occur in orthopyroxene porphyroclasts, and this could account for some of the Ca inferred to have been in the protolith. Nevertheless, it appears likely that a few percent (probably ≤5) of diopside must originally have been present in the harzburgitic parent rocks.

Olivine and chromite are the two primary minerals that have been retained in the metamorphosed ultramafic rocks. Both minerals recrystallized during metamorphism, and this has affected their crystal habit and composition. Much of the olivine has recrystallized to a mosaic of strain-free crystals exhibiting granoblastic polygonal texture and having grain sizes generally ≤1 mm. Relict olivine grains (porphyroclasts) are larger (some exceed 1 cm) and invariably strained, with undulatory extinction or strain banding. Incompletely recrystallized olivine contains strain-bands and often has curved grain boundaries. These textures are all typical of recrystallized dunites (Ragan, 1969). So, too, is the narrow range in olivine composition (usually <1 % Fo) in individual samples. However, the overall range in olivine composition in the Dark Ridge body (~10 % Fo) found in this study is much larger than reported by previous investigators. Nearly anhydrous metadunite has an average olivine composition of $Fo_{92.2}$, virtually identical to the average composition ($Fo_{92.4}$) for southern Appalachian ultramafics reported by Carpenter and Phyfer (1975) and typical of Alpine-type ultramafics worldwide (Trommsdorff and Evans, 1974; Evans, 1977). As Fig. 8 and Table 3 indicate, olivine from chromite-rich rocks is more magnesian (~Fo_{95}), presumably due to partial equilibration with chromite wherein Fe was preferentially incorporated into the latter mineral, while olivine in hydrated rocks is more Fe-rich. In fact, a very good correlation exists between the degree of hydration, as measured by the percentage of hydrous minerals in the mode (Table 1), and the % Fo in olivine (Fig. 14): the more hydrated the metaultramafic rock, the lower the Fo content of the olivine. This simply reflects the higher preference for Mg of coexisting metamorphic minerals such as talc, chlorite, serpentine, and calcic amphibole (all have higher Mg-numbers than coexisting olivine). Hence, olivine lost magnesium as these Mg-rich hydrous minerals formed.

Chromite was strongly affected by the hydration and recrystallization of the Dark Ridge metaultramafic rocks. In most samples chromite has reacted with adjacent olivine producing an envelope of chlorite (Fig. 9B). In only a few essentially anhydrous metadunite samples (e.g., DR-2 and DR-4) has this not occurred to any great extent. The composition of chromite crystals varies widely (Table 4 and Figs. 10-12). Those from the chromite prospect are the most magnesian. This is consistent with the findings of Carpenter and Fletcher (1979) and Lipin (1984), both of whom found that massive (segregated) chromite from pods and lenses is richer in Mg than disseminated chromite. Raymond et al. (2003) noted that Cr-numbers and Mg-numbers range from 40 to 100 and <10 to 65, respectively, in chromite from Blue Ridge metaultramafic rocks. Fig. 10 shows that this entire range is displayed in the Dark Ridge metaultramafic body. As noted for olivine, chromite composition correlates with hydration. Steatitized or serpentinized assemblages typically contain chromite having Cr-numbers above 90 and Mg-numbers ≤15; this chromite invariably has poikiloblastic texture and is enclosed by chlorite (Fig. 9B). Chromite in more anhydrous rocks displays much greater compositional variation, with much of the range encompassed by individual zoned crystals (Figs. 10 and 12). This wide range suggests the possibility that a vestige of primary igneous compositions may have been preserved (i.e., in the cores of some zoned crystals and in chromite inclusions in olivine and orthopyroxene porphyroclasts).

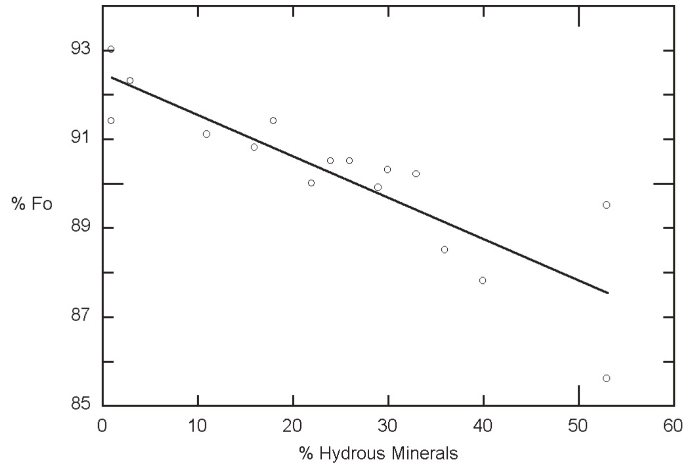

Figure 14. Plot of average Fo content in sample versus percentage of hydrous minerals.

The primary assemblage of olivine+chromite±orthopyroxene has undergone at least three episodes of metamorphism. The highest grade assemblage that is found in the Dark Ridge metaultramafic rocks is olivine + chromite + chlorite + tremolitic hornblende ± orthopyroxene. This assemblage is preserved in metaharzburgite containing orthopyroxene porphyroclasts and in steatitized metaharzburgite with orthopyroxene remnants in olivine, and by cores of tremolitic hornblende in calcic amphibole crystals. Chlorite in this assemblage is derived from reactions between primary Al-chromite and adjacent olivine; the resulting metamorphic chromite is richer in Cr and Fe^{3+} and depleted in Al and Mg relative to primary chromite. Presence of orthopyroxene indicates a minimum temperature for this assemblage of about 670˚C (Bucher and Frey, 1994; Spear, 1993). Chlorite plus calcic amphibole stably coexist in metaultramafic rocks to at least 740˚C, or possibly somewhat higher (Bucher and Frey, 1994). Thus, a peak metamorphic temperature of 705 ± 35˚C is inferred. The virtually complete absence of anthophyllite in the Dark Ridge metaultramafic rocks indicates that pressures must have exceeded 6 kb, as that is the upper pressure stability limit for anthophyllite in the system MgO-SiO_2–H_2O (Bucher and Frey, 1994). The presence of Fe in amounts typical for ultramafic rock compositions extends the anthophyllite stability field to about 11 kb (Evans and Guggenheim, 1988), so it is probable that pressures were at or slightly above 11 kb.

The most common metamorphic assemblage at Dark Ridge, characteristic of steatitized metaharzburgite, is olivine + chromite + chlorite + tremolite + talc. Talc is mainly derived from breakdown of orthopyroxene (Fig. 13). Nearly pure tremolite replaces Al-bearing tremolitic hornblende that was stable at higher temperature, and is often found rimming the latter (Table 5). This assemblage is stable from 570-670˚C at conditions along the kyanite geotherm (Bucher and Frey, 1994), but this 100˚C range narrows considerably at higher pressure (to <50˚C at ~11 kb (Evans and Guggenheim, 1988; Winter, 2001)).

Retrograde serpentine-bearing assemblages formed at lower (<570˚C) temperatures. Antigorite does not appear to be in equilibrium with other silicate minerals (Swanson, 2001) as it replaces olivine (Fig. 6D), chlorite (Fig. 6C), and talc (Fig. 7C).

Magnesite + antigorite ± talc (ophimagnesite assemblage) veins are present in several samples. The coexistence of antigorite and magnesite in these veins indicates that X_{CO2} of the fluid phase responsible for their formation was very low (<0.03), as more CO_2–rich fluids stabilize the assemblage talc + magnesite (Trommsdorff and Evans, 1977; Berman, 1991; Bucher and Frey, 1994). Mesh-textured serpentine (chrysotile ± lizardite) formed at still lower temperatures by alteration of olivine along cracks and grain boundaries.

In summary, the Dark Ridge metaultramafic rocks record a history of upper amphibolite facies metamorphism (Taconic event?) followed by lower amphibolite facies metamorphism (Neoacadian event, or cooling from peak Taconic metamorphism?) and still later retrograde metamorphism under greenschist facies conditions (Alleghanian?), all accompanied by increasing hydration. The peak metamorphic conditions estimated in this study are in pretty good accord with those (725˚C and 9 kb) determined for rocks of the Tallulah Falls Formation in the vicinity of Sylva, North Carolina by Quinn (1991). Comparable P-T conditions have been reported for the nearby Cullowhee gneiss (Davidson, 1995).

CONCLUSIONS

1) The Dark Ridge metaultramafic body consists of both metadunite, in part serpentinized, and metaharzburgite, much of which has been steatitized. Chromitite is a minor constituent and is mostly confined to a small area near the western edge of the body, north of Dark Ridge Creek.

2) Metamorphism and accompanying hydration of the primary assemblage olivine + chromite ± orthopyroxene has led to the formation of metamorphic assemblages containing talc, chlorite, calcic amphibole, and serpentine. Upper amphibolite/lower granulite facies metamorphism (T = 705 ± 35˚C) resulted in a peak metamorphic assemblage of olivine + chromite + chlorite + tremolitic hornblende ± orthopyroxene. Retrograde lower amphibolite facies metamorphism produced the dominant assemblage at Dark Ridge, olivine + chromite + chlorite + tremolite + talc. Disequilibrium assemblages containing serpentine (antigorite) formed still later under greenschist facies conditions and were accompanied by the intrusion of veins containing magnesite + antigorite ± talc.

3) Olivine and chromite recrystallized during metamorphism, causing changes in both texture and mineral composition. Much of the olivine has been polygonized to smaller (≤1 mm) granoblastic-textured grains; olivine lost magnesium as Mg-rich hydrous minerals formed. Within individual samples olivine exhibits a very narrow range in composition, usually < 1% Fo. Primary Al-rich chromite reacted with olivine to form Al-depleted chromite with poikiloblastic texture and a surrounding envelope of Cr-bearing chlorite. The composition of chromite varies widely; it is possible that the cores of zoned crystals and chromite inclusions in olivine porphyroclasts may reflect primary igneous compositions.

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Field Guide

Field Trip Stop Descriptions for Carolina Geological Society Field Trip November 5–6, 2005

INTRODUCTION

This field guide contains 12 stops in the North Carolina Blue Ridge (Fig. 1). On day 1 we will examine representative exposures of Tugaloo terrane, Dahlonega gold belt, Cartoogechaye terrane, and western Blue Ridge rocks along the Blue Ridge Parkway (see Hatcher et al., this guidebook, their Fig. 2). Day 2 will concentrate on rocks in the Dahlonega gold belt and the Cartoogechaye terrane in the area around Sylva. We have not constructed a detailed roadlog for the trip, but have provided instructions and distances for geting from one stop to another, along with GPS (Global Positioning Satellite) locations at each stop to aid those wanting to use the guidebook on future field trips. Most of the stops are on public roads, but a few are on private property. Please honor property owners' requests when entering privately owned localities in the future, and ask permission where needed so that others can also enjoy these exposures in the future. **PLEASE NOTE THAT NO ROCK HAMMERS ARE ALLOWED ON ANY OF THE STOPS ON THE BLUE RIDGE PARKWAY.**

SAFETY

Please be very cautious at each field trip stop. Board and leave the buses carefully, look very carefully crossing highways, and exercise caution walking along highways. Do not go close to overhanging ledges, particularly where there is loose rock, and do not attempt to climb roadcuts or other rock faces at any of the stops. Each stop presents different kinds of hazards, so please look around at all of the stops before looking at geology.

We wish everyone a safe, educational, and enjoyable 2005 Carolina Geological Society field trip.

Figure 1. Location of 2005 Carolina Geological Society field trip stops and headquarters at Maggie Valley, North Carolina.

__DAY 1.__ Depart Quality Inn, Maggie Valley at 8 am, turn left onto U.S 276 and drive 3.8 mi E to turn onto U.S. 23–74S. Drive S to Blue Ridge Parkway, get onto Parkway and drive E to Caney Fork Overlook. Optional stop at rest area on U.S. 23–74 before getting onto Parkway.

STOP 1–1: Plane and wedge rock failure modes in the Ashe–Tallulah Falls Formation at Caney Fork Overlook.

Leaders: Rebecca Latham and Rick Wooten. **20 minutes.**

GPS location: 35° 19' 38.4" N, 82° 57' 55.4" W.

PURPOSE: To observe brittle and ductile structures and compositional layering related to rock slope stability in metasedimentary rocks of the Ashe-Tallulah Falls Formation. Analysis of rock slope stability at this location is part of ongoing geologic and geohazards studies by the North Carolina Geological Survey (NCGS) along the North Carolina segment of the BRP, and in landslide hazard mapping in western North Carolina.

WARNING: Be extremely careful around the steep rock cut. Rockfalls may occur suddenly without warning.

Description. Two types of rock slide failure modes are represented at this location, a plane failure and a wedge failure (Fig. 1–1–1). In both cases the failed material is gone, exposing the planes of weakness in the rock slope formed by brittle (joints) and ductile (foliation) structures, and compositional layering. Migmatitic metagraywacke is the dominant lithology here, with subordinate amounts of biotite-muscovite schist. Both lithologies are cut by more than one generation of pegmatite dike and cross cutting relationships are well exposed. The dominant foliation subparallels the migmatitic and compositional layering of the rock, and generally dips steeply northwest (out of the cut slope). This foliation parallels the limbs of tight to isoclinal, upright folds, and envelops pods and boudins of calc-silicate, pegmatite, and migmatite. Migmatitic layering is folded indicating that these are at least second generation (F_2) folds. Kinematic (stereonet) and factor of safety stability analyses were performed for the plane and wedge failures using Rockpack III software (Watts and others, 2003). Rock weathering descriptors are from Williamson (1984). Note: The convention for left and right in describing slope failure features is looking from the top of the failure toward the

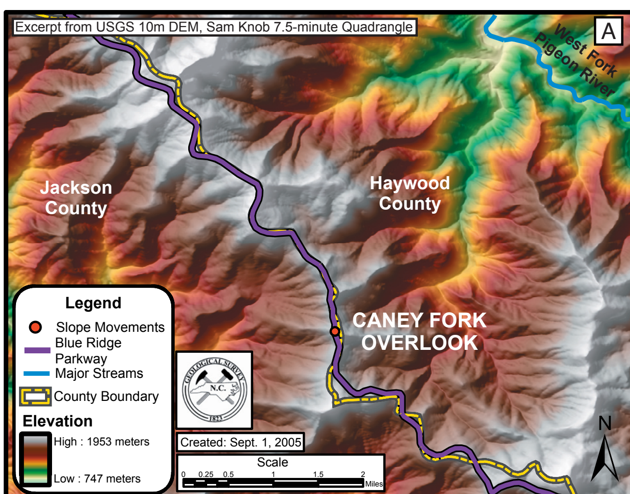
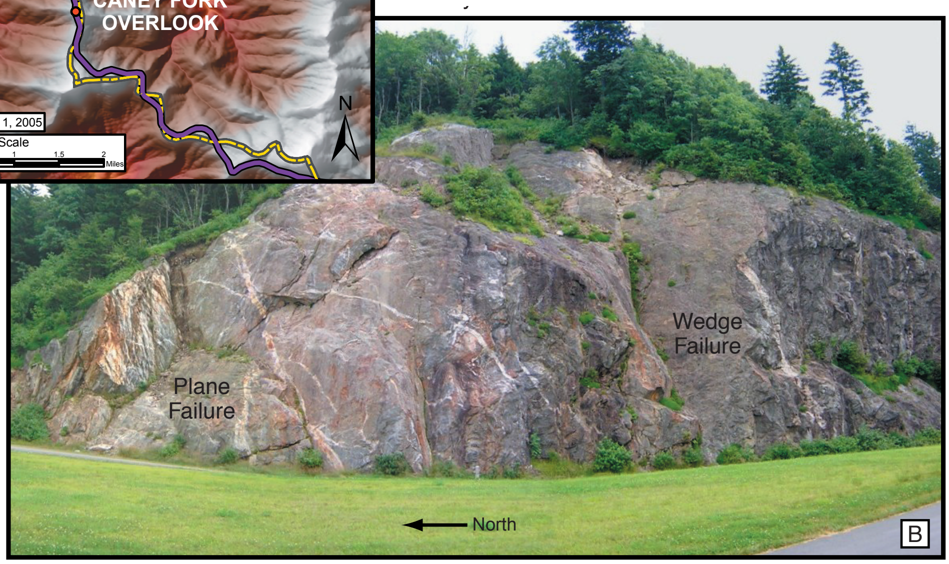


Figure 1–1–1. Location of Caney Fork Overlook. A. Shaded relief map showing the location of the Caney Fork Overlook. B. Panoramic view of the outcrop at the Caney Fork Overlook. Dominant lithology is migmatitic metagraywacke with subordinate biotite, muscovite schist and cross cutting pegmatites. Stability analyses were performed at this location on a plane failure and a wedge failure. Such rock slope failure modes are common along the Blue Ridge Parkway.



direction of movement.

Plane Failure. This block of material (~1,500 m^3, 55,000 ft^3) was removed about seven years ago over concerns about a potential damaging failure, possibly because the block had shown signs of movement. Our kinematic (stereonet) and plane failure factor of safety analyses confirm that the block was unstable with a 0.87 factor of safety (Fig. 1–1–2). Intersecting exfoliation joints and dominant foliation (S_2) planes in stained state migmatitic metagraywacke are the critical failure surfaces and form the undulating basal sliding surface. The average dip (weighted by dip segment length) of the sliding surface is about 44°. A preferentially weathered, biotite, muscovite schist layer parallel to the S_2 foliation and migmatitic layering forms the back release surface of the block. A subordinate S_3 (?) foliation oblique to the S_2 foliation is also present in the schist. The partly to completely decomposed schist layer, now removed from the back scarp, is visible where the right flank release surface abuts the back scarp (Fig. 1–1–2F). In the factor of safety analysis we modeled the schist layer as a tension crack with pore water pressure, because of the preferential weathering and evidence of seepage. Intersecting S_2 foliation planes and high angle tectonic joint surfaces (100°/70°–122°/62°) form the stepped, right flank release surface. Also exposed along the right flank release surface are numerous, tectonic slip surfaces (72°/65° to 78°/70°) with subhorizontal slickenlines indicating right-lateral, strike-slip movement (Figs. 1–1–2C and 1–1–2D). The plane failure analysis does not account for any frictional resistance afforded by the stepped right

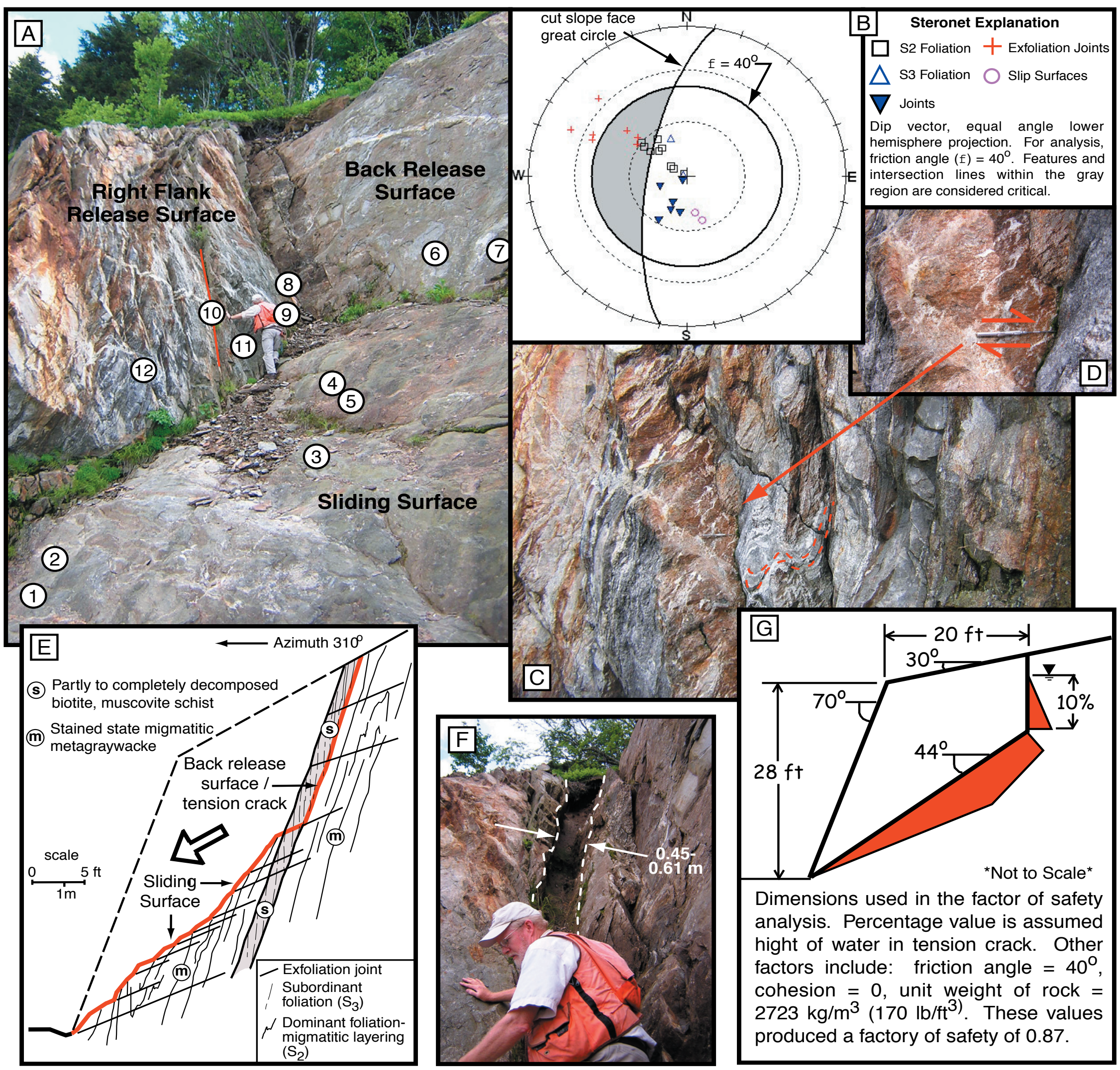


Figure 1–1–2. Details for plane failure. A. Structural details for top left photograph. 1. S2 foliation -- 216/67. 2. Exfoliation joint -- 202/36. 3. Exfoliation joint -- 205/35. 4. Exfoliation joint -- 214/59. 5. S2 foliation -- 216/63. 6. S2 foliation in migmatitic metagraywacke -- 230/70. S3 foliation in schistose layer -- 205/88. 7. Just off the photograph, exfoliation joint -- 220/57. 8. Weathered biotite muscovite schist layer (See F). 9. Joint -- 122/62. 10. Fault subparallel to foliation -- 235/87. Offset is approximately 15-23 cm (6-9 in). 11. Joint -- 100/70. 12. Slip surface with subhorizontal slickenlines on surface -- 72/65. Movement is right lateral (See C and D). B. Dip vector stereonet plot of structural features. Projection is equal angle, lower hemisphere. C. View of brittle and ductile features on right flank release surface (Datapoint 12 in A). Folds are outlined in red. D. Subhorizontal slickenlines shown in C. Arrows indicate right lateral movement. Scrive is 15 cm (~6 in) long. E. Schematic geologic cross section through the plane failure. F. View of the partly to completely decomposed biotite, muscovite schist exposed along the right flank release surface. G. Plane failure factor of safety analysis of slope failure. Calculated factor of safety = 0.87.

flank release block; therefore the 0.87 factor of safety may be somewhat conservative. The analysis does indicate the potential for future movement of the block remaining along the right flank.

Wedge Failure. The curvilinear intersection of a steeply dipping joint (158°/75°) and an S_2 foliation plane (230°/70°) is exposed at the location of a past rock wedge failure in migmatitic metagraywacke (Fig. 1–1–3). As expected, the stereonet and factor of safety analyses indicate the missing rock wedge was unstable. In the stereonet analysis (Fig. 3B) the great circles of the joint and S_2 foliation planes intersect in the gray area determined by the assumed friction angle (μ = 40°), and the cut face orientation. Intersections and poles to planes that plot within this gray area are considered critical.

Acknowledgements. Funding for the geologic inventory of the North Carolina segment of the Blue Ridge Parkway is provided in part by the National Park Service. Anne Witt, NCGS, prepared Figure 1–1–1A.

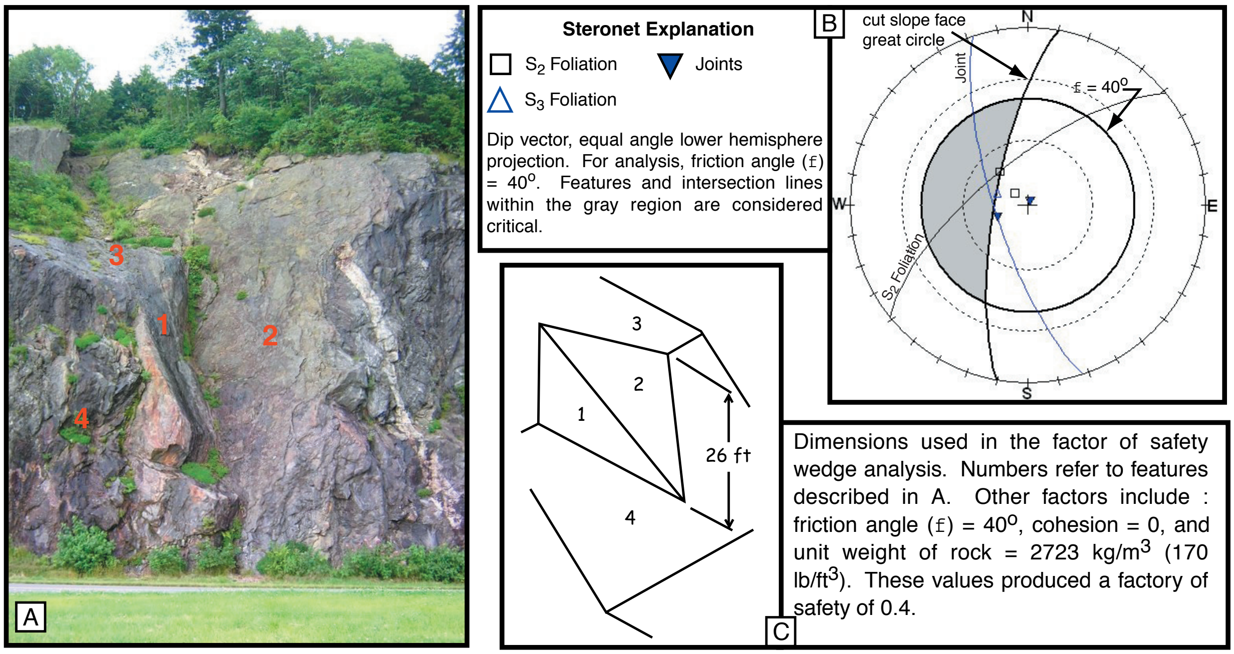


Figure 1–1–3. Details for wedge failure. A. View from the overlook of past wedge failure in migmatitic metagraywacke. 1. Joint (158/75) forms the right side of the wedge. 2. S_2 foliation (230/70) forms the left side of the wedge. 3. Upper slope—190/25. 4. Cut slope face —190/75. B. Dip vector stereonet plot of the data. Intersection of the S_2 foliation and joint great circles indicates the critical orientation of the line of intersection of the two planes along which sliding occurred. Plot is an equal angle lower hemisphere projection. C. Wedge failure factor of safety analysis. Calculated factory of safety = 0.4.

Continue W on the Blue Ridge Parkway past Beartrap Gap Overlook (0.6 mi, elev. 5,545 ft) and Reinhart Gap (0.4 mi, elev. 5,455 ft). Some 0.4 mi W of Reinhart Gap is an exposure of migmatitic biotite gneiss (Ashe-Tallulah Falls Formation) folded with Otto metasandstone (Fig. 2). This exposure thus probably contains the Chattahoochee-Holland Mountain fault. Continue W past Beautiful Ridge Overlook (1 mi, elev. 5872 ft.) into the Great Balsam Mountains window; Cowee Mountain overlook (0.4 mi, elev. 5,950 ft); Haywood–Jackson Overlook (0.3 mi, elev. 6020 ft) and an exposure of Otto Formation metasandstone and schist; Richland Balsam, highest point on the Blue Ridge Parkway (0.4 mi, elev. 6,410 ft), and exposures of Otto Formation garnet-muscovite schist; Lone Bald Overlook (1.4 mi, elev. 5,635 ft); Roy Taylor Forest Overlook (0.5 mi, elev. 5,580 ft); Doubletop Mountain Overlook (2.0 mi, elev. 5,365 ft); Licklog Gap Overlook (0.6 mi, elev. 5,135 ft), with Grassy Ridge Mine Overlook (0.9 mi, elev. 5,250 ft); Steestachee Bald Overlook (2.2 mi, elev. 4,780 ft), with exposures of Otto Formation metasandstone and schist across from it; a total of 9.4 mi to Cove Field Ridge Overlook (0.5 mi, elev. 4,620 ft). Pull into parking area.

STOP 1–2. Otto Formation Metasandstone at Cove Field Ridge Overlook. Leaders: Arthur Merschat and Bob Hatcher Elevation 4,620 ft. **25 minutes.**

GPS Location: 35° 25.85’ N; 83° 02.14’ W.

Purpose: To examine a large exposure of Otto Formation metasandstone in the Great Balsam Mountains window (Dahlonega gold belt).

Description. This exposure was visited previously on the 1975 CGS field trip (Kish et al., 1975) and revisited here to discuss

Figure 2. Complex folds along the Blue Ridge Parkway near the Chattahoochee-Holland Mouuntain fault ~0.4 mi W of Reinhart Gap—the boundary between the Tugaloo terrane and Dahlonega gold belt. Ashe-Tallulah Falls Formation rocks of the Tugaloo terrane are more migmatitic than the rocks of the Dahlonega gold belt, providing evidence that this fault is post-metamorphic, but still being emplaced at sillimanite grade conditions.

the multiple interpretations of the statrigraphic position of the rocks, recent detailed geologic mapping, and geochronology. Hadley and Nelson (1971) originally correlated these rocks with the Dahlonega gold belt, but others have correlated these rocks with Ashe/Tallulah Falls Formation (e.g. Rankin, 1975; Brown, 1985). Recent detailed geologic mapping suggest that these are Otto Formation rocks exposed in the Great Balsam Mountains window, which extends from south of Sylva to Cruso, NC (Quinn, 1991; Steven H. Edelman unpublished, A. J. Merschat unpublished data). Otto Formation is dominated by migmatitic metasandstone with minor interlayers of schist, metaconglomerate, and calc-silicate. Otto metasandstone is fine- to medium grained sandstone dominated by quartz, with minor biotite, muscovite, plagioclase, K-spar, and minor accessories (Fig. 1–2–1). Layers generally range from 10 cm to >1m. Percentages of K-spar and muscovite are comparable in modal analyses of Quinn (1991) from the Sylva area. Metaconglomerate layers can be recognized by the presence of 0.5 – 1 cm long quartz and feldspar porphyroclasts and grade into metasandstone layers, likely representing original bedding. Schist layers vary from a <1cm to >1m and are composed of sillimanite, garnet, biotite, muscovite, and quartz. Minor calc-silicate occurs as lenses and layers usually closely associated with metasandstone. Foliation dips moderately to steeply north. Large probably 3rd-generation folds are present farther west toward the tunnel in migmatitic metasandstone. Trondhjemite dikes (age estimated ~400 Ma by Kish, 1975, using Rb-Sr, and determined by Mapes, 2002, 402 Ma using the SHRIMP) crosscut the foliation in the enclosing rocks. Older pegmatite veins both parallel and cross the foliation. A sample of Otto metasandstone from this locality yielded detrital zircons with SHRIMP ages of 1.1, 1.2, 1.3, and 1.4 Ga, and metamorphic rims of ~455 Ma (A. J. Merschat, unpublished data; see Hatcher et al., this guidebook Tables 3 and 5).

Continue driving W on the BR Parkway. We cross the W-dipping Soque River fault into the Cartoogechaye terrane ~1000 ft E of the Waynesville Overlook, ~ 0.75 mi E of Stop 1–3 (from Steven H. Edelman unpublished geologic map of Hazelwood quad), then

Figure 1–2–1(a). East-vergent folds in Otto Formation metasandstone at Cove Field Ridge Overlook cut by a small trondhjemite body and later quartz-feldspar veins that both crosscut and parallel the dominant foliation.

Figure 1–2–1(b). Close view of the east end of the cut at Cove Field Ridge Overlook showing the west dip of foliation and quartz-feldspar veins the parallel the dominant foliation.

pull into Standing Rock Overlook some 1.75 mi W of Stop 1–2.

STOP 1–3. Standing Rock Overlook. Fan deposit landforms and Cartoogechaye Terrane Migmatite.
Elevation 3,915 ft. **Leaders:** Rick Wooten and Rebecca Latham **20 minutes.**

GPS location: 35°26'18.86" N, 83°03'16.76" W

Purpose: To observe landforms characteristic of colluvial and alluvial fan deposits along mountain foot slopes. The North Carolina Geological Survey (NCGS) is mapping and compiling the locations of these fan deposits as part of geologic and geohazards studies along the North Carolina segment of the BRP, and landslide hazard mapping in western North Carolina

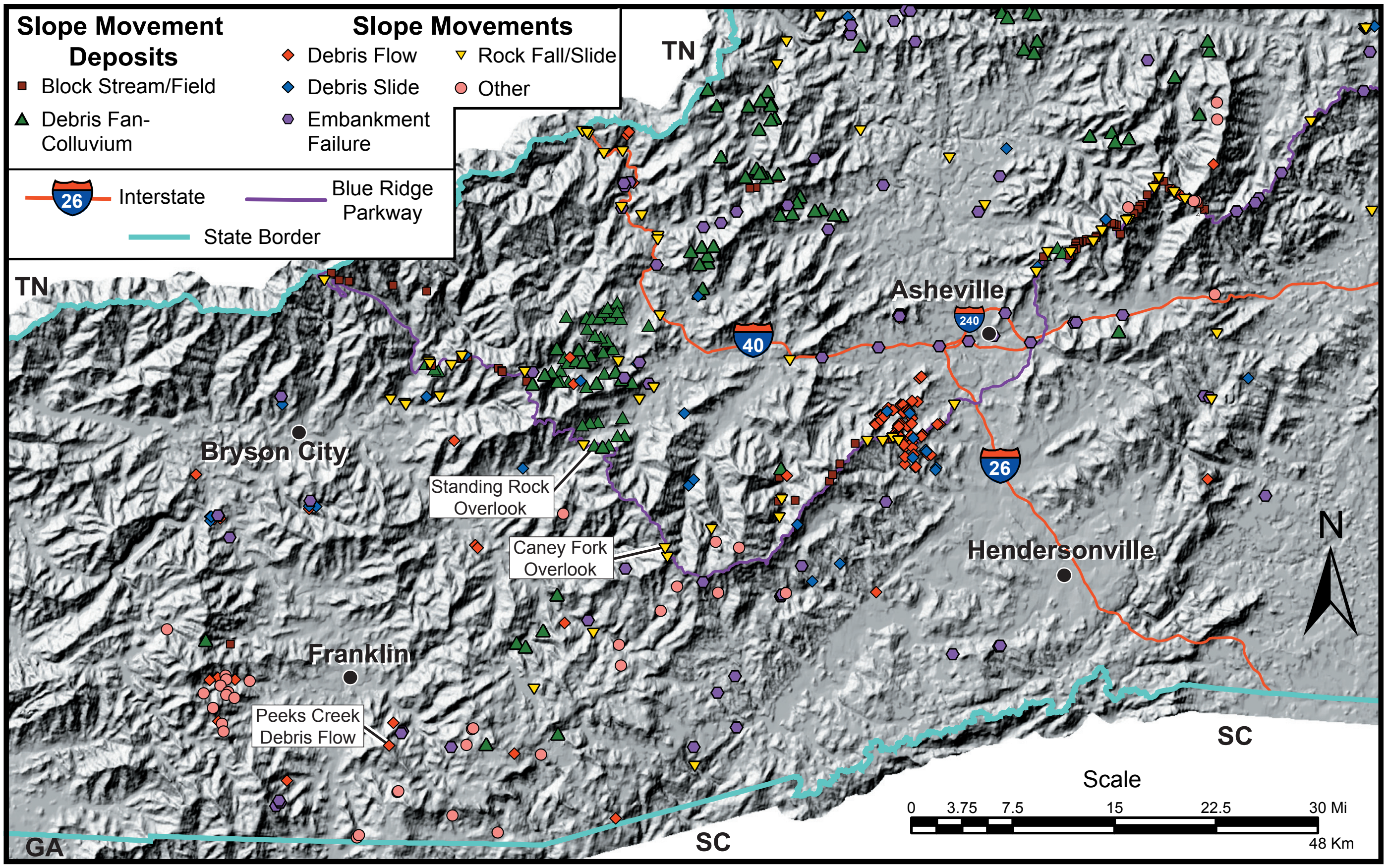


Figure 1–3–1. Shaded relief map showing a portion of western North Carolina and locations of slope movements and slope movement deposits currently in the North Carolina Geological Survey database. These data were compiled from numerous literature sources and mapped by NCGS staff. Field trip stop locations at the Caney Fork and Standing Rock overlooks are shown along with the location of the Peeks Creek debris flow. The remnants of Hurricane Ivan triggered the Peeks Creek debris flow on September 16, 2004 resulting in five fatalities and over 20 severely damaged or destroyed homes.

(Fig. 1–3–1). Also is a view of the high mountains to the NW, and opportunity to examine blocks of Cartoogechaye terrane metasandstone migmatite that make up Standing Rock.

Looking north from the Standing Rock overlook are good views of two distinct landforms resulting from colluvial and alluvial deposition along mountain foot slopes. Figure 1–3–2, modified from Mills and Allison (1995), shows the generalized outlines of the Winchester Creek (WCF) and Barber Orchard (BOF) fans, and other nearby fan deposits. The WCF is a mountain cove deposit visible across the valley (Fig. 1–3–3A), whereas the BOF, immediately below the overlook, displays a more classic fan morphology (Fig. 1–3–3B). These composite fans record past sediment transport in the form of flood, debris flow, and colluvial deposits, and as such can be indicators of potential future slope movements originating from their source areas upslope. Deep, unconsolidated fan deposits can also be unstable in high, steep excavations. As a side note, the Environmental Protection Agency recently cleaned up soil contamination on parts of the BOF from pesticide and herbicide use during earlier fruit tree growing operations.

Mills and Allison (1995) demonstrated the composite nature of the deposits by delineating three relative fan-surface age groups (young, intermediate and old) within the individual fans based on topographic position and weathering characteristics. Figure 1–3–4 shows examples of deposits generally characteristic of the three fan-surface age groups. Young fan deposits (Holocene to Late Pleistocene?) typically have hard rock clasts and brown matrix (Munsell hue 7.5-10 YR) (Fig. 1–3–4A). Intermediate age fan-surface deposits (Late to Middle Pleistocene?) tend to contain more weathered rock clasts in a yellow brown matrix (Munsell hue 5-7.5 YR) (Fig. 1–3–4B). The old deposits (Middle to Early Pleistocene?) generally contain partly, to completely decomposed rock fragments in a characteristic red, to red-brown matrix (Munsell hue 2.5-5YR) (Fig. 1–3–4C). Mills (2005) provided an overview of index characteristics used for relative-age dating of fan deposits.

The Balsam Gap II landslide (Fig. 1–3–2), located on the north side of U.S. 23-74 immediately west of the Blue Ridge Parkway access road, is developed in saprolite overlain by colluvium. The southwest lobe of the slide failed in 1969, and the larger northeast lobe failed during the 1980's (Lambe and Riad, 1990).

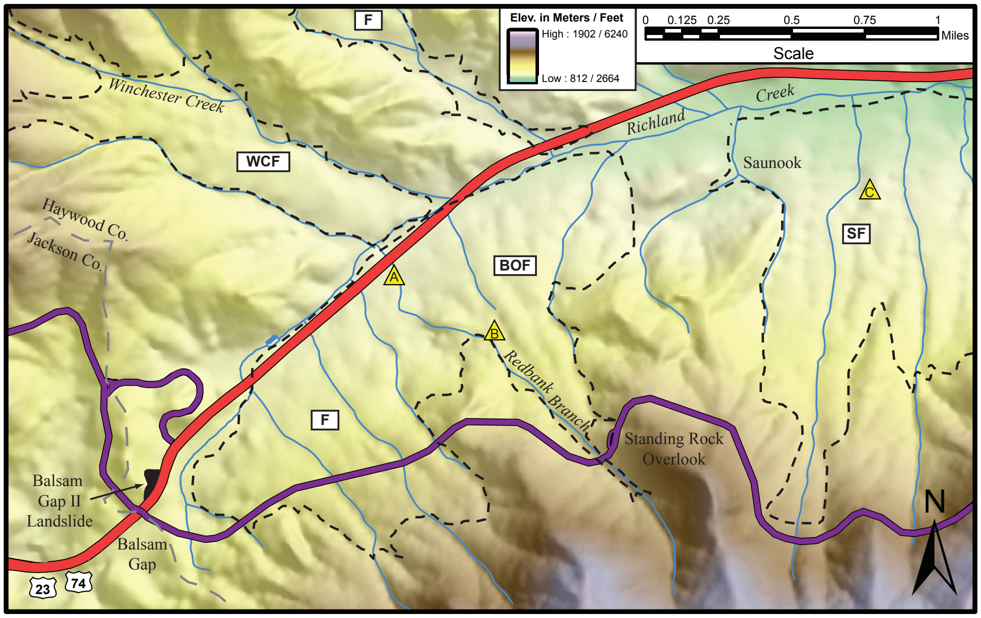


Figure 1–3–2. Shaded relief map showing generalized outlines of fan deposits (dashed black line) in the vicinity of Standing Rock Overlook (modified from Mills and Allison, 1995). WCF - Winchester Creek fan, BOF - Barber Orchard fan, SF –Saunook fan, F–undifferentiated fan deposits. Shaded relief map created from U.S.G.S. 10-m digital elevation model (DEM).

Figure 1–3–3. Views of fan deposits (dashed white outlines). A. View across the valley from Standing Rock overlook toward the Winchester Creek fan (WCF). Toe of the Barber Orchard fan (BOF) is visible at photo bottom. B. View of the Barber Orchard fan (BOF) looking toward Standing Rock overlook.

Figure 1–3–4. Fan deposits, letters correspond to lettered locations in Figure 1–3–2. **A.** Young fan deposits exposed at the toe of the BOF. Downstream is to right as indicated by imbricated clasts near hammer (handle is 28 cm (11 in) long). **B.** Intermediate age fan deposits with yellowish brown matrix exposed on an interfluve midslope on the BOF. **C.** Old fan deposits with red matrix exposed in the lower slope of the Saunook fan (SF). Scribe is 15 cm (6 in) long.

Acknowledgements. Funding for the geologic inventory of the North Carolina segment of the Blue Ridge Parkway is provided in part by the National Park Service. Anne Witt, NCGS, provided geographic information system (GIS) support and prepared Figure 1–3–1.

Blocks of banded, migmatitic metasandstone of the central Blue Ridge (Cartoogechaye terrane) contains tight folds, probably 3rd generation, on the parking lot side of the block, but also a number of earlier refolds and possible sheath folds are exposed on the northwest side of the exposure (downhill side away from the BRP). RDH.

Continue 2.1 mi W on the Blue Ridge Parkway to Balsam Gap (35° 26.31' N; 83° 04. 54' W, elev. 3,395 ft), exit the Parkway, and drive north 0.5 mi to the rest area on right. ***Morning Break.***

HISTORICAL SIGNIIFICANCE OF BALSAM GAP (35° 26.31' N; 83° 04. 54' W, elev. 3,395 ft)

In 1776, American General Griffith Rutherford brought 2,400 troops (probably irregulars) through Balsam Gap from the south and destroyed some 36 Cherokee Indian villages in the valleys just north of here forcing the Cherokees to move into the Great Smoky Mountains. The Cherokees had previously been aiding the British by providing resistance in the upper Piedmont and mountain region. Rutherford's action put an end to Cherokee activities in the Revolutionary War, except for minor skirmishes, and permitted the colonists to focus on defeating the British along the Eastern Seaboard. (From historical marker at Rabb Knob Overlook.)

Return to the Blue Ridge Parkway and drive 2.6 mi W to Woodfin Valley Overlook. Pull into parking area.

STOP 1–4. Structural and Metamorphic evolution of the Western–Central Blue Ridge boundary Exposed at Woodfin Valley Overlook. **Leaders:** Matthew A. Massey and David P. Moecher. Elevation 4,120 ft. **20 Minutes**

GPS Location: 35° 27.15' N; 83° 05.58' W.

Purpose: To examine Hayesville fault and terrane boundary contact between Cartoogechaye terrane and Great Smoky Group rocks of the Laurentian rifted margin sequence.

Description: Central Blue Ridge rocks here consist of dark, biotite-dominant metasandstone and amphibolite, making up those in the E part of the exposure (and exposures up the hill), whereas muscovite-bearing metasandstone and schist make up the W part of the cut (poorly exposed during the summer). The photo in Fig. 1–4–1 was taken during the winter when the leaves were down and other vegetation was subdued (Massey and Moecher, 2002). This boundary is recognizable from Alabama to Newfoundland in the west-central Appalachians. The fault that forms the boundary transported ophiolites onto the North American margin, preserved as remnants today in the Bay of Islands, Newfoundland, and at Thedford Mines, Québec. It is possible that the rocks at Stops 2–3 and 2–4 are also remnants of an ophiolite, but they are unrecognizable as such today because they enjoyed plastic deformation and high-grade metamorphism after emplacement.

The boundary between the western Blue Ridge (WBR) and central Blue Ridge (CBR, Cartoogechaye terrane) was defined as the Hayesville fault (Hatcher et al., 1979), and proposed to represent a pre-metamorphic suture established during the early phases of Taconian orogenesis, possibly representing the up-dip leading edge of an early Paleozoic accretionary wedge above an east-dipping subduction zone (Hatcher, 1987; Goldberg et al., 1989). Although mapping and lithologic descriptions have permitted precise location of the Hayesville fault in North Carolina and northern Georgia (Hatcher et al., 1979; Eckert et al., 1989; Brumback, 1990; Quinn, 1991; Adams et al., 1995; Davidson, 1995; Montes, 1997), details regarding the nature of the boundary at a variety of scales remain to be elucidated and disseminated. The road cut at Woodfin Valley Overlook on the Blue Ridge Parkway is a fortuitous exposure of the Hayesville fault. This easily accessible locality affords an excellent opportunity to examine, describe, and discuss these details. Additional details are provided in Massey and Moecher (2005, in press).

Lithologies. Exposures at Woodfin Valley Overlook are comprized of WBR metapelitic and metapsammitic paragneisses and migmatites, and CBR migmatitic mafic orthogneisses and amphibolite. WBR lithologies are mapped as "Great Smoky Group, undivided" by both Hadley and Nelson (1971) and the North Carolina Geological Survey (1985). The CBR is mapped as "Carolina Gneiss" (biotite and hornblende gneiss, amphibolite) by Hadley and Goldsmith (1963), and layered gneiss and migmatite by both Hadley and Nelson (1971) and the North Carolina Geological Survey (1985). Biotite-hornblende paragneisses to the north, south, and east of Woodfin Valley Overlook have been correlated with the Ashe–Tallulah Falls Formation (Quinn, 1991; Montes, 1997), but today are recognized as being part of the separate Cartoogechaye terrane.

Structures. The dominant structure at Woodfin Valley Overlook is the NW-SE striking gneissic banding, which correlates with the regional S_2 foliation (Fig. 1c). Here, S_2 is defined by parallelism of 001 planes of fine- to coarse-grained muscovite (WBR/GSG only) and biotite, leucocratic layers of quartz and feldspar, long axes of elliptically shaped garnet, hornblende (CBR orthogneisses only), titanite, and locally the 010 planes of kyanite blades and c-axes of sillimanite (WBR GSG only; see Figs. 3c, d, and e). Intrafolial, isoclinal folds (F_2) coplanar to S_2 can be found on both sides of the boundary.

Locally, CBR amphibolite boudins containing isoclinal folds with axial planes oblique to the surrounding S_2 foliation provide evidence for pre-S_2 deformation (D_1; Fig. 1f). Evidence for the D_1 event is also preserved locally as the S_1 fabric in

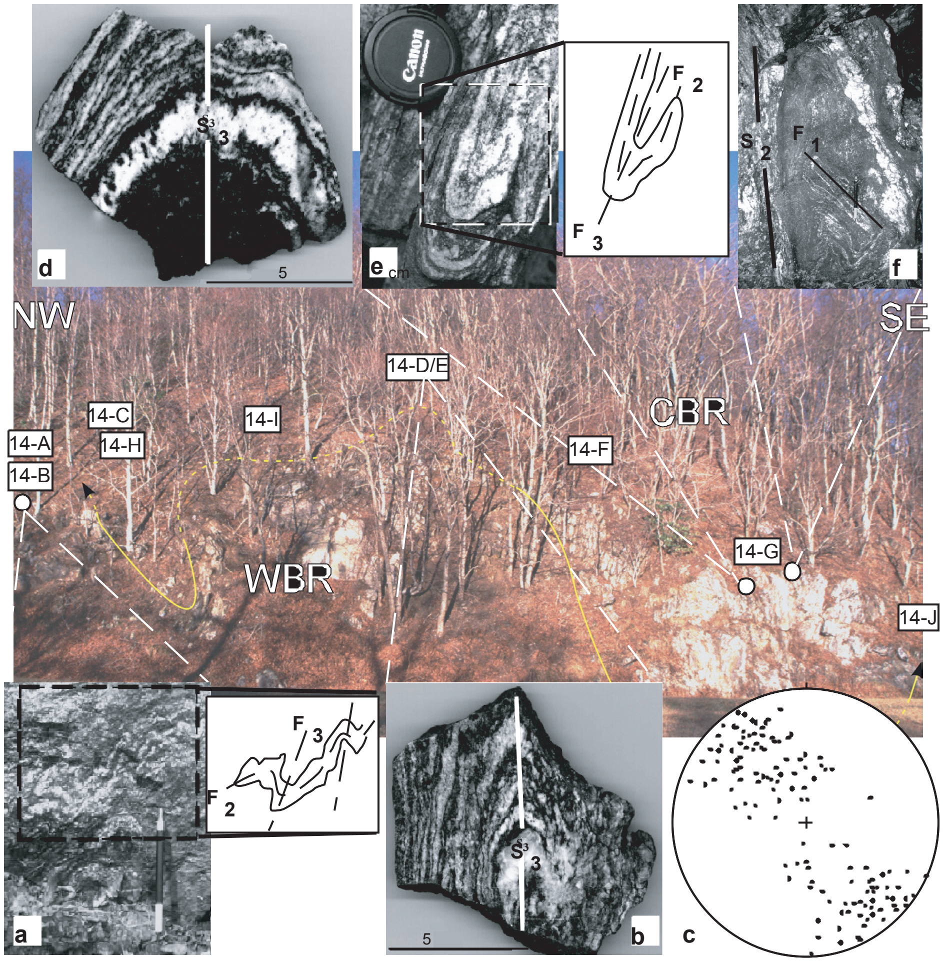


Figure 1–4–1. Road cut at Woodfin Valley Overlook locality (center) with folded WBR-CBR contact delineated by solid-dashed line; sample localities marked for reference. Figures a through f show details exposed in outcrop. (a) F_2 isoclinal fold refolded by inclined F_3 open folds in WBR; pencil for scale. (b) F_3 fold in WBR garnet paragenesis showing development of S_3 parallel to F_3 axial plane. (c) Equal area net shows poles to foliation along Blue Ridge Parkway northwest and southeast of Woodfin Valley Overlook. (d) F_3 fold in EBR migmatitc biotite orthogneiss showing development of S_3 parallel to F_3 axial plane. (e) F_2 isoclinal fold refolded by isoclinal F_3 fold in EBR; lens cap for scale. (f) F_1 isoclinal folds defined by leucocratic layers preserved in EBR amphibolite boudin, surrounded by sub- vertical S_2 foliation in EBR orthogneiss.

garnet-bearing WBR metapsammites, and as inclusion trails in WBR garnets to the northeast. There is no direct correlation of S_1 fabrics preserved in WBR garnet and F_1 folds of the CBR. Consequently, these folds could represent an earlier phase of deformation before emplacement of the Hayesville fault.

Open to isoclinal F_3 folds with NNE-trending fold axes largely control the structure of the area. F_3 folds commonly refold the S_2 foliation and F_2 folds (Figs. 1a and e), and are associated with a disjunctive, anastomosing S_3 axial planar cleavage (Figs. 1–4–1b and 1–4–1d) characterized by dynamic recrystallization of quartz grains and quartz ribbons, folding and rotation of isolated muscovite flakes, and neocrystallized aggregates of fine-grained biotite and muscovite in thin folia. Syn-kinematic recrystallization of feldspars is variable, but where present produces core-mantle structures (Fig. 1–4–2a). Rims of feldspar porphyroclasts contain bulging microstructures and few subgrains equivalent in size (~10 µm) to recrystallized margins, indicating a predominance of recrystallization by grain boundary migration. The prevalence of feldspar recrystallization in quartzofeldspathic layers, and lack of recrystallization in more micaceous layers suggests strain partitioning (Fig. 1–4–2b); strain was preferentially accommodated by micas, probably by basal slip. Crystal-plastic deformation of feldspar indicates temperatures attending D_3 deformation of at least 500˚-550˚C (amphibolite facies; Tullis, 1983; Voll, 1976). Kinked kyanite, muscovite, and biotite, fractured garnet, and quartz ribbons and undulose extinction are also prototypical of D_3 microstructures (see Figs. 1–4–2b and 1–4–3e).

Localized micro-shear zones and pressure solution characterize a distinct low-grade deformation (D_4). Dynamic recrystallization of quartz and neocrystallization of muscovite and biotite are typical in micro-shear zones, although D_4 microstructures are distinguished from D_3 by fracturing and cataclasis of feldspars (Fig. 1–4–2c). Fracturing, cataclasis, and retrogression of garnet porphyroclasts to biotite + chlorite (Fig. 1–4–2d) are also products of D_4. Pressure solution of biotite, muscovite, and feldspar is often localized along S_2/S_3 folia near D_4 micro-shear zones (see Fig. 1–4–3f). Microstructures and mineral assemblages are consistent with a distinct lower greenschist facies overprint.

Metamorphism. Lithologies at Woodfin Valley Overlook contain mineral assemblages that we interpret to have developed during four metamorphic episodes. Two distinct episodes of prograde metamorphism, relative to S_2, define the peak regional event, designated M_2; M_{2A} is pre- to syn-kinematic reaching kyanite-grade, and M_{2B} was a static, syn- to post-kinematic, sillimanite-grade event. M_2 corresponds to the regional Barrovian progression mapped by Hadley and Goldsmith (1963) throughout the Great Smoky Mountains. M_3 and M_4 metamorphism distinctly post-date M_2; M_3 is the result of D_3 deformation and M_4 is a static retrograde event associated with D_4.

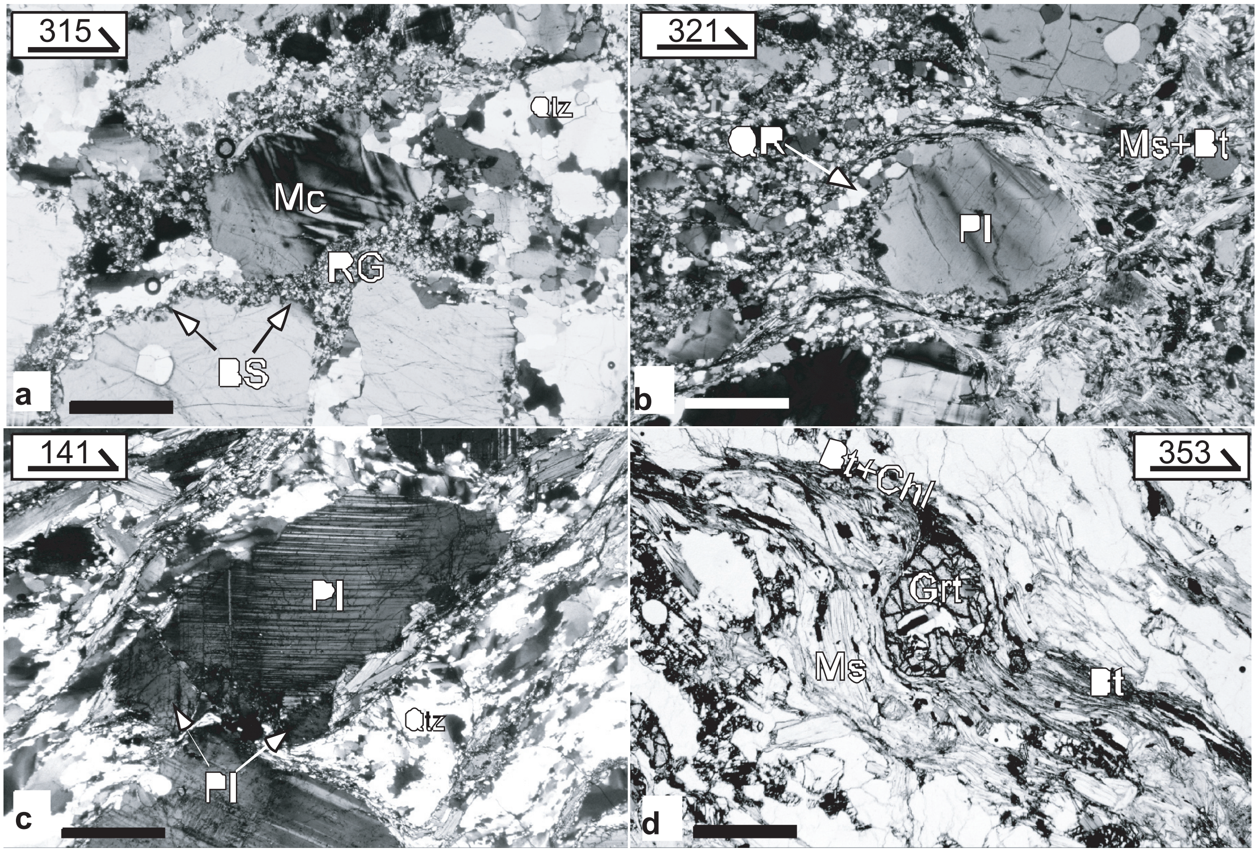


Figure 1–4–2. D_3 and D_4 microstructures at Woodfin Valley Overlook; scale bars are 1mm, directions of dip are indicated. See Figure 1–4–1 for sample locations. (a) D_3 deformation of microcline (Mc) porphyroclast. Bulging structures (BS) of equivalent size as recrystallized grains (RG) in rim indicate deformation by grain boundary migration. MM14-G, cross polars. (b) Variable D_3 deformation in WBR paragneiss. Plagioclase porphyroclast exhibits localized recrystallization by grain boundary migration along with quartz ribbons (QR) on left side of photomicrograph. Lack of plagioclase recrystallization on right side indicates strain accommodated by muscovite-biotite matrix (Ms+Bt) probably by basal slip. MM14-D, cross polars. (c) Plagioclase porphyroclast in D_4 micro-shear zone. Notice smaller, angular fragments extending from tail and fractures within plagioclase porphyroclast. MM14-C, cross polars. (d) Garnet (Grt) sigma clast in D_4 micro-shear zone. Notice retrogression of garnet to biotite + chlorite (Bt+Chl) in tails. MM14-A, plane light.

In addition to M_2 index minerals (garnet + kyanite ± staurolite ± sillimanite), WBR parageneses contain biotite + muscovite + plagioclase + quartz + rutile + ilmenite + apatite + monazite. Garnet porphyroblasts are poikiloblastic (Figs. 1–4–3a and 1–4–3b), but fine-grained, inclusion-free garnet, and inclusion-free rims on poikiloblastic garnet suggest a second phase of garnet growth associated with M_{2B}; additionally, garnet compositional zoning patterns exhibit Ca-enriched rims surrounding unzoned garnet cores (Fig. 3a). Although M_{2A} kyanite defines the S_2 foliation (see Fig. 1–4–3e), M_{2B} sillimanite is commonly randomly oriented and included in post-kinematic, coarse-grained muscovite (M_{2B} or M_3?).

CBR mafic orthogneisses most commonly display the M_{2A} assemblage plagioclase + biotite + epidote + clinozoisite + rutile + titanite + ilmenite + apatite + hornblende ± garnet. Sporadic inclusion-free garnet is wrapped by the S_2 foliation defined by layers of biotite, epidote, apatite, titanite, and hornblende (Figs. 1–4–3c and 1–4–3d). M_{2B} is recognized by epidote, clinozoisite, and hornblende porphyroblasts that cut S_2; hornblende and clinozoisite locally contain S_2 inclusion trails (Fig. 1–4–3d). Amphibolite boudins contain weak to no internal fabrics, but coarse-grained garnet includes hornblende, epidote, and biotite, assumed to correlate with M_{2B}.

A large proportion of both WBR and CBR lithologies at Woodfin Valley Overlook are migmatites. Mesoscopically, migmatites appear as stromatites (Ashworth and McLellan, 1985) deformed during D_2. F_2 folding (locally F_3 refolding; Figs. 1a and e), boudinage and flattening of leucosomes parallel to S_2 attest to D_2 deformation implying anatectic conditions during pre- to syn-D_2 conditions. Microstructurally, leucosomes vary in extent of deformation due to D_3 deformation. Least deformed leucosomes show no sign of crystal-plastic deformation in feldspars, quartz is characterized by semi-polygonal to lobate grain boundaries with little or no undulose extinction and textures are generally equigranular and medium- to coarse-grained. Most deformed leucosomes display characteristic D_3 microstructures. Melanosome selvages are predominantly biotite and minor muscovite parallel to surrounding S_2 fabrics, further suggesting pre-/syn-S_2 migmatization. Locally, leucosomes cutting S_2 are observed, indicating some post-kinematic migmatization occurred. Observations indicate timing of migmatization is in accord with both M_{2A} and M_{2B} metamorphism as discussed above.

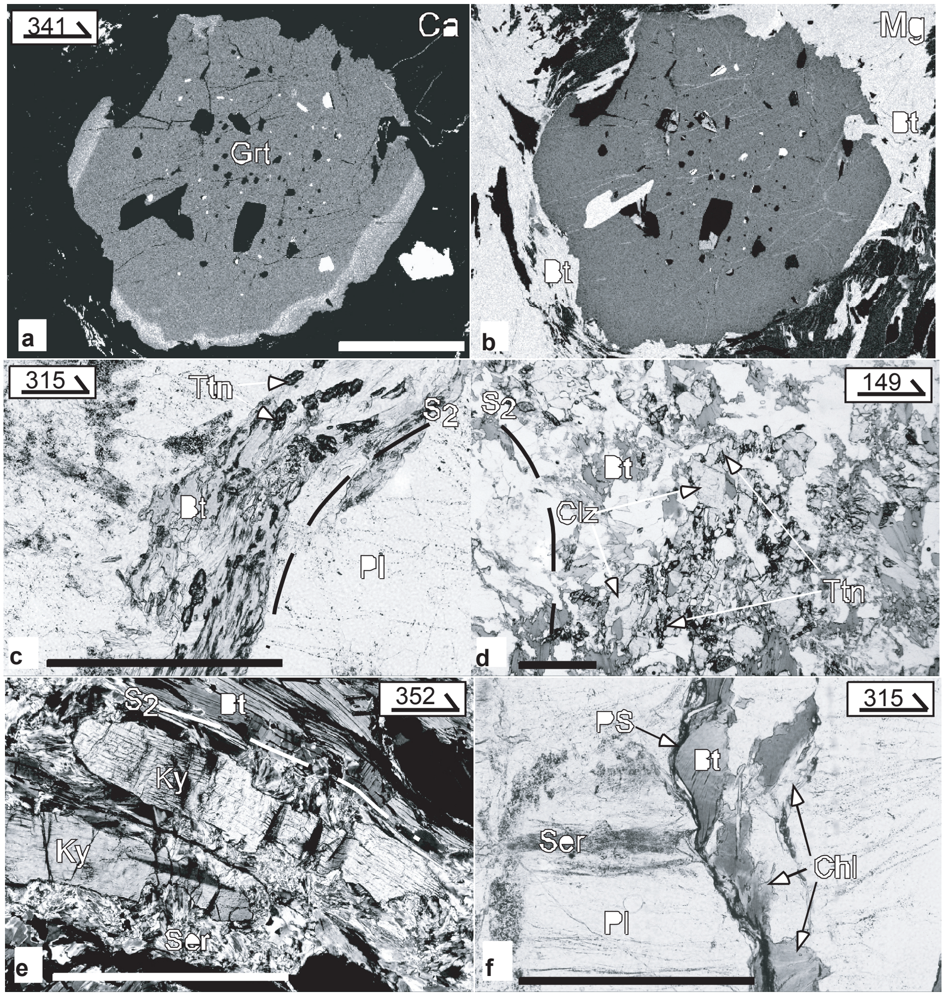


Figure 1–4–3. Mineral parageneses, and porphyroblast-matrix relationships for M_2, M_3, and M_4 metamorphism; scale bars are 1mm, directions of dip are indicated. See Figure 1–4–1 for sample locations. (a-b) Ca and Mg X-ray compositional maps illustrating zoning pattern in poikiloblastic garnet (Grt) from sample MM24-A (~100 m west of Woodfin Valley Overlook). Notice second generation garnet growth (M2B) marked by inclusion-free, Ca-enriched rim, late resorption of garnet rims, and truncation by Mg-rich biotite (Bt). (c) S_2 foliation defined by M2A biotite and elongate titanite (Ttn), alternating with layers of plagioclase (Pl). MM14-G, plane light. (d) M2B clinozoisite (Clz) and biotite porphyroblasts overgrowing and including M2A titanite and quartz defining S2 foliation. MM14-H, plane light. (e) S2 foliation defined by M2A biotite and kinked kyanite (Ky). M4 static retrogression of kyanite to sericite (Ser). Notice random orientation of sericite. MM14-B, cross polars. (f) Pressure solution (PS) of biotite and plagioclase representative of D4 deformation. Also notice M4 retrogression of plagioclase to sericite, and biotite to chlorite (Chl) associated with pressure solution. MM14-G, plane light.

M_3 was primarily a fold and foliation-forming event (F_3/S_3) that resulted in new mica growth, recrystallization of quartz and feldspar, and deformation or slight retrograde reaction of M_2 metamorphic index minerals. X-ray compositional maps show that the M_3 event resulted in variable resorption of garnet margins (see Fig. 1–4–3a). The Ca-rich band characteristic of garnet interpreted to have formed during peak M_2 P-T conditions, is truncated by biotite or muscovite that grew during M_3 and define S_3 (see Fig. 1–4–3b). A final generation of Al_2SiO_5 (recognized within ~100 m of Woodfin Valley Overlook) distinct from the acicular fibrolite and granular sillimanite of M_{2B}, occurs as bundles and clusters of 10 µm long, randomly oriented, stubby needles mainly concentrated along plagioclase grain boundaries (see Moecher and Massey, this volume). Although the paragenesis of this sillimanite is not clear at present, its presence corroborates the amphibolite facies grade for D_3/M_3.

Retrograde M_4 metamorphism is associated with D_4 micro-shear zones and pressure solution, however, fabric development is localized to shear zones. Besides phase alteration associated with D_4 deformation (discussed above), M_4 is characterized by sericitization of kyanite and feldspars, and replacement of biotite and garnet by chlorite (Figs. 1–4–2d, 1–4–3e and 1–4–2f). Static retrograde reactions are consistent with D_4 deformation mechanisms at lower greenschist facies.

Tectonic Interpretation. Exposures at Woodfin Valley Overlook allow comparison of structures on each side of the WBR-CBR boundary. At least two fold periods (F_2 and F_3), three periods of foliation development (S_1, S_2, and S_3), two periods of upper amphibolite facies metamorphism and associated partial melting (M_{2A} and M_{2B}), and lower greenschist facies retrograde metamorphism (M_4) can be identified and correlated across the boundary in both the WBR and CBR. While regional terrane relationships imply that this boundary developed as a suture (Eckert and Hatcher., 2003), we do not see the meso- or micro-scale evidence (e.g., relic annealed mylonites) that support the inference of faulting, shearing, or emplacement of a thrust sheet along the contact prior to regional Taconian M_2 metamorphism. If the WBR-CBR contact was originally a fault, subsequent deformation and metamorphism has completely eliminated fault fabrics where exposed on the Blue Ridge Parkway.

Amphibolite facies D_3 deformation clearly post-dates M_2 metamorphism; however, there is no relation to an Acadian orogen-scale, dextral fault zone as documented to the northeast along this same boundary (Adams et al., 1995; Stewart and Miller, 2001; Troupe et al., 2003). Although the possibility does remain that the regional D_3 event is a late manifestation of Taconian orogenesis, detailed geochronology of post-peak Taconian deformation and metamorphism to the northeast suggests an Acadian age (see Moecher and Massey, this volume). Static, retrograde lower greenschist facies M_4 metamorphism and localized D_4 deformation were presumably Alleghanian effects.

Continue W along the BR Parkway 5.6 mi to Waterrock Knob Overlook and visitor center. ***Lunch Stop.***

STOP 1–5. Great Smoky Group Rocks at Waterrock Knob Overlook. Elevation 5,820 ft. **Leaders:** Scott Southworth, Bob Hatcher, and those who distribute the lunches. **40 minutes.**

GPS Location: 35° 27.60' N; 83° 08.48' W.

Purpose: To view Great Smoky Mountains to W, and numerous 6,000-ft mountaintops nearby; fresh Great Smoky Group rocks.

Description: A number of 6,000+-ft. mountaintops are visible from here, both close by and in all directions. The overlook is just below the crest of Waterrock Knob (6,292 ft). Before you start up the trail to the top, realize that you will have a climb of 472 ft to get there. Folded (tight to sheath folds) Great Smoky Group metasandstone and schist, cut by small pegmatite dikes, are exposed near the National Park Service Visitors Center.

Return to the Parkway and continue W 1.5 mi to Woolyback Overlook.

STOP 1–6. Folds at Woolyback Overlook. Elevation 5,425 ft. **Leader:** Bob Hatcher. **15 minutes.**

GPS Location: 35° 28.07' N; 83° 08.51' W.

Purpose: To examine noncylindrical folds in Great Smoky Group rocks.

Description: Rusty-weathering Great Smoky Group metasandstone and schist are folded into tight, noncylindrical folds (Fig. 1–6–1). Alternating thin and medium beds of metasandstone with schist doubtlessly influenced the deformation style here. These folds are not unique in this area; others are present at Stop 1–5, so the deformation plan for this region includes heterogeneous strain. The rusty appearance of the rocks is related to weathering of pyrite, pyrrhotite, and possibly other sulfide minerals.

Continue 3.8 mi W to Soco Gap exit and turn onto U.S. 19S (toward Cherokee). Drive 0.6 mi and pull off to right along U.S. 19S.

Figure 1–6–1. Noncylindrical folds in Great Smoky Group metasandstone and schist at Woolyback Overlook.

***Stop 1–7.* Longarm Quartzite (Snowbird Group) at New Roadcut on U.S. 19S.** Elevation ~3,900 ft. **Leader:** Scott Southworth. **15 minutes.**

GPS Location: 35°29.422'N; 83º 09.835' W

Purpose: To examine some of the Longarm Quartzite that makes up the thin section of Snowbird Group rocks above basement and below Greenbrier fault.

Description: Light-colored, thin to medium-bedded sandstone sets the Longarm Quartzite apart from the darker Great Smoky Group rocks to the E. The premetamorphic Greenbrier fault that thrust Great Smoky Group rocks over Snowbird Group and basement rocks crosses the Parkway ~0.5 mi E of Soco Gap, and the contact with the 1.1 Ga Grenville basement is ~0.25 mi W.

Return to Blue Ridge Parkway and drive 2.3 mi. W past Jonathan Creek Overlook (0.7 mi, elev, 4,460 ft) and an exposure on left of basement orthogneiss (0.9 mi), and pull into Plott Balsam Overlook.

***Stop 1–8.* Granitoid gneiss at Plott Balsam Overlook.** Elevation 5,020 ft. **Leader:** Scott Southworth **15 minutes.**

GPS location: 35° 30' 30" N; 83° 10' 24" W

Purpose: To examine an exposre of basement orthogneiss.

Description: Fine- to coarse-grained granite, granitic orthogneiss, and sheared granite that forms a major part of the basement in this area (Fig. 1–8–1). This fine- to medium-grained granite is part of a pluton belonging to a large group of 1.15 Ga plutons that were intruded during the middle stages of the Grenville orogeny (Carrigan et al., 2003) (Fig. 1–8–1). Metamorphism and deformation took place ~1.0 Ga.

Turn around and return to Soco Gap and turn N. Drive through Maggie Valley to the Quality Inn. End of Day 1.

Figure 1–8–1. Strongly foliated, medium- to coarse-grained felsic orthogneiss basement that is representative of the basement rocks exposed in the eastern Great Smoky Mountains. The other principal lithology that occurs in the basement in thiis area is biotite augen gneiss.

Day 2 Field Guide

Depart Quality Inn, Maggie Valley at 8 am, turn left onto U.S 276 and drive 3.8 mi E to get onto U.S. 23–74S. Drive S to Sylva, bypass the town to Dillsboro, and continue S on U.S. 441-23 5.3 mi to Savannah Church exposure. Optional stop at rest area on U.S. 23–74 before getting onto Parkway.

STOP 2–1. Block-in-Matrix Structure at Savannah Church. Elevation ~2,400 ft. **Leaders:** Shawna Cyphers, Steve Yurkovich. **25 minutes.**

GPS location: 35°17.569' N; 83°16.515' W

Purpose: To examine a fresh exposure of block-in-matrix structure in the Dahlonega gold belt.

Description: The rocks at this locality were described by Raymond et al. (1989). They consist of a folded series of amphibolite, calc-silicate, and biotite gneiss blocks in a matrix of finer–grained feldspar-quartz-biotite partial melt (Fig. 2–1–1). They were cut by several prominent late pegmatite dikes that appear largely undeformed.

Return to Dillsboro and turn right at traffic light into the main part of town. Continue through town and turn onto NC 107 toward Cullowhee and pull into Arbys parking lot.

STOP 2–2. Cartoogechaye Terrane Rocks behind Arbys Restaurant in Sylva. Elevation: 2,120 ft. **Leaders:** Bob Hatcher, Steve Yurkovich, and Arthur Merschat. **25 minutes.**

GPS location: 35°22.287'N; 83°12.310' W.

Purpose: To examine some complexly deformed migmatitic biotite gneiss and amphibolite in hanging wall of Soque River fault.

Description: The rocks at this stop are high grade (Ordovician upper amphibolite to granulite(?) facies) Cartogechaye terrane biotite gneiss, calc-silicate layers, garnet amphibolite, and amphibolite cut by early to late quartz-feldspar veins. The dominant foliation is probably S_2, but could be S_3. Both isoclinal and later folds are present, along with boudins of amphibolite (Fig. 2–2–

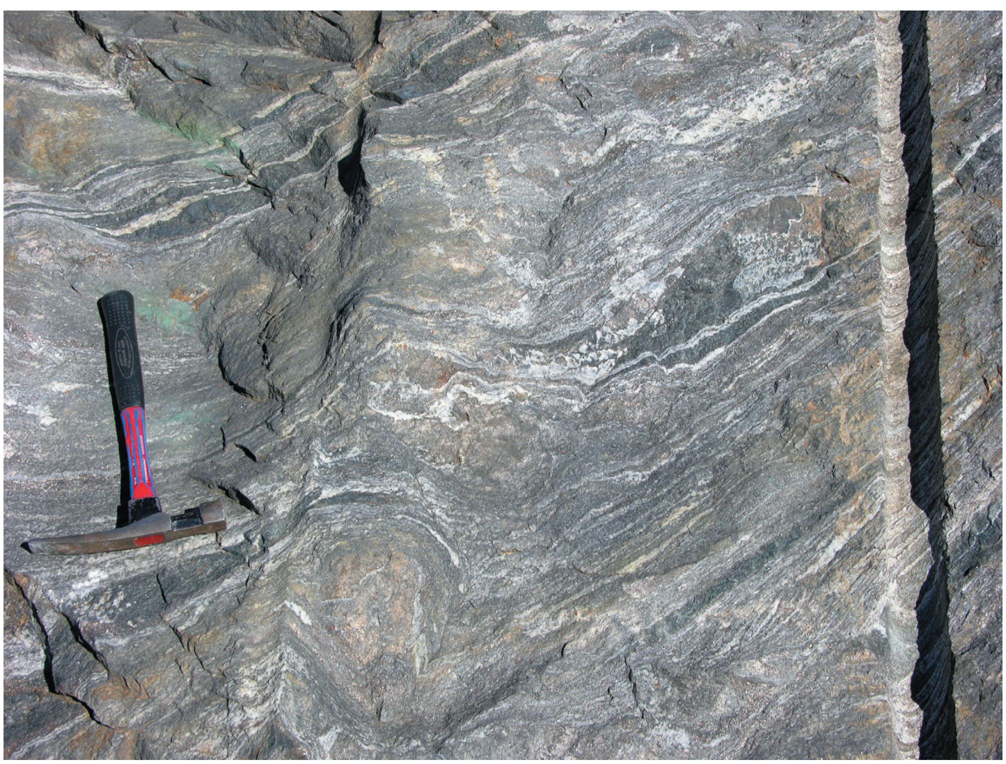


Figure 2–1–1. Complexly deformed, partially melted amphibolite "blocks" at Savannah Church on U.S. 23–441 south of Dillsboro. The matrix is Otto Formation(?) biotite gneiss.

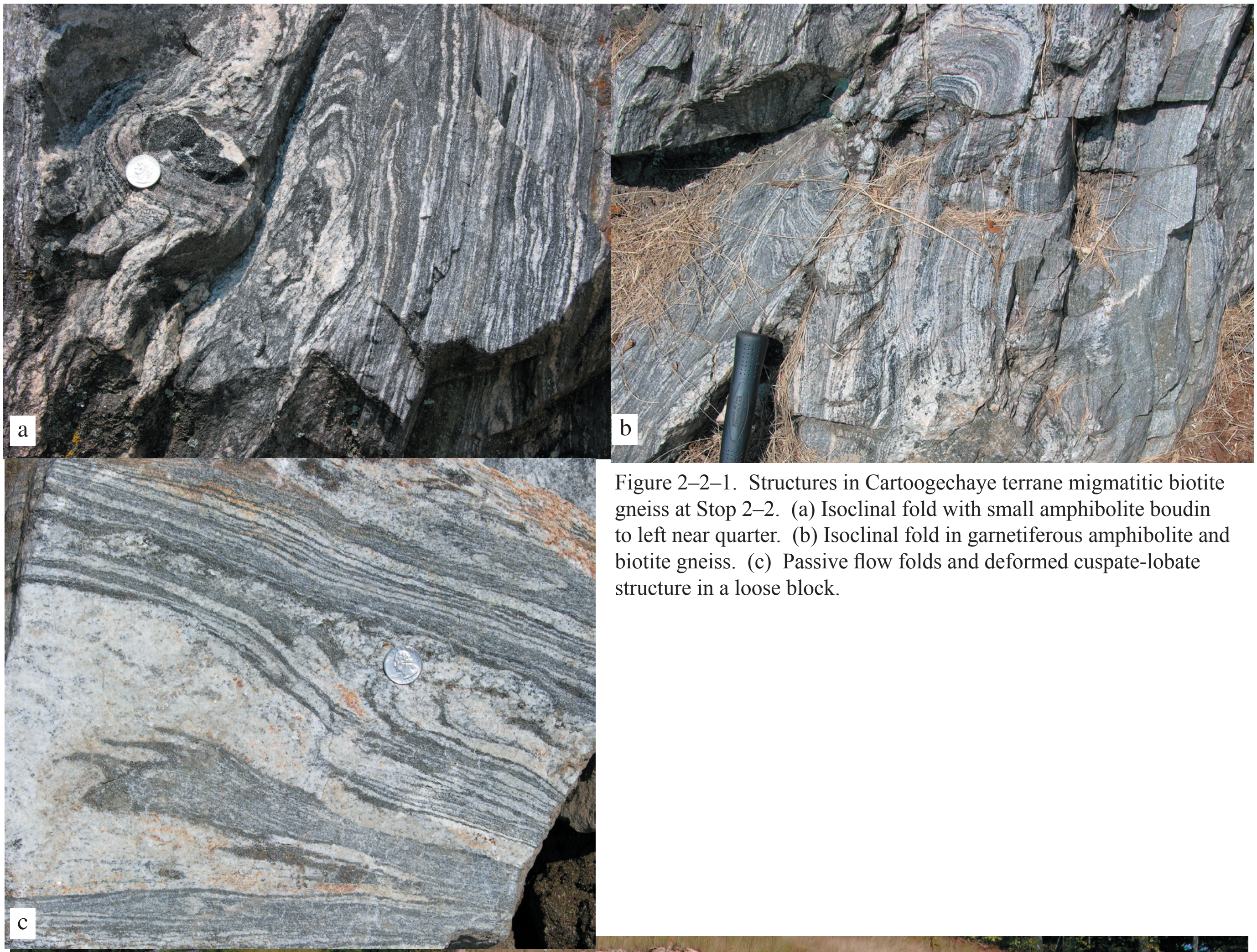


Figure 2–2–1. Structures in Cartoogechaye terrane migmatitic biotite gneiss at Stop 2–2. (a) Isoclinal fold with small amphibolite boudin to left near quarter. (b) Isoclinal fold in garnetiferous amphibolite and biotite gneiss. (c) Passive flow folds and deformed cuspate-lobate structure in a loose block.

Figure 2–2–2. Profile from fresh to partially weathered to completely decomposed Cartoogechaye terrane migmatitic biotite gneiss at Stop 2–2.

1). An excellent weathering profile is also exposed here (Fig. 2–2–2).

Turn around and return to downtown Sylva; turn north onto U.S. 23 toward Waynesville and return to the bypass. Drive ~4.5 mi N on US 23-74 and turn left on Mineral Springs Rd SR1456. Drive ~0.9 mi and turn right into the quarry st the bottom of a steeep hill.

STOP 2–3. Addie Quarry: Webster-Addie Mafic-Ultramafic Complex Dunite off Mineral Springs Road. Elevation: 2,360 ft
Leaders: Richard Warner and Steve Yurkovich. 30 minutes

GPS Location: 35º 23.987' N; 83º 09.592' W.

Purpose: To examine lower cumulate part of upper mantle component of an ophiolite, or possibly the bottom of a differentiated pluton.

Description: The Webster-Addie ultramafic complex is a discontinuous series of metaultramafic rocks folded into a dome shape and conformably enclosed by gneissic metagraywacke and amphibolite of the Cartoogechaye terrane (Quinn, 1991). The outcrop pattern is ring-shaped with a diameter of about 10 km. The Webster-Addie complex has been described in the older literature as an example of a "ring" dike," but the body was reinterpreted in the 1960s as a concordant tabular pluton arched into a dome (Madison, 1968), because the contacts parallel the foliation in the enclosing rocks. Metadunite is the principal rock type, but metaharzburgite and pyroxene-rich rocks (metawebsterite, metaorthopyroxenite) are locally abundant. The Addie Quarry is located on the northern tip of the complex. The metadunite was mined for olivine at this stop and also at Chestnut Gap, approximately 4 km SSE of this stop.

Rocks exposed within the quarry are primarily metadunite and its alteration products (Fig. 2–3–1). The original dunite mineralogy is represented by olivine, chromite, and, locally, orthopyroxene; however, these minerals have all been recrystallized during the several episodes of Paleozoic metamorphism that affected the southern Appalachians. The addition of H_2O-rich, CO_2-bearing fluids during metamorphism produced secondary minerals such as anthophyllite, serpentine, chlorite, talc, tremolite, and magnesite. The northeast face of the quarry illustrates the compositional layering common in this body. Serpentine occurs as an alteration product of the olivine, as veins that surround and/or cross cut olivine, and as concentrated masses along joint or foliation surfaces (Cronin, 1983). Talc is found along the contact of the body with the country rock and adjacent to intrusions of more felsic rocks (pegmatite here).

In thin section the metadunite is has a granoblastic polygonal texture with olivine grains often meeting in 120˚ triple junctions (Fig. 2–3–2). The average olivine grain size is about 0.5mm; the largest olivine grains are ~2mm. In some samples there is a poor banding marked by layers of very fine-grained (average ~0.2mm) olivine alternating with coarser olivine layers. The olivine is for the most part fresh with some incipient alteration to serpentine along internal cracks and grain boundaries; in some samples antigorite replacement has been concentrated along olivine cleavage planes. Modes of thin sections from the Addie Quarry and nearby outcrops at the intersection of Blanton Branch Road and US 19/23 indicate that the metadunite contains 65-93% olivine. Olivine compositions, determined from electron microprobe analyses, range between Fo_{89} and Fo_{91}, typical of "Alpine-type" ultramafic bodies. Chromite constitutes 1-2% of the metadunite. Secondary minerals present include serpentine (5-18%), talc

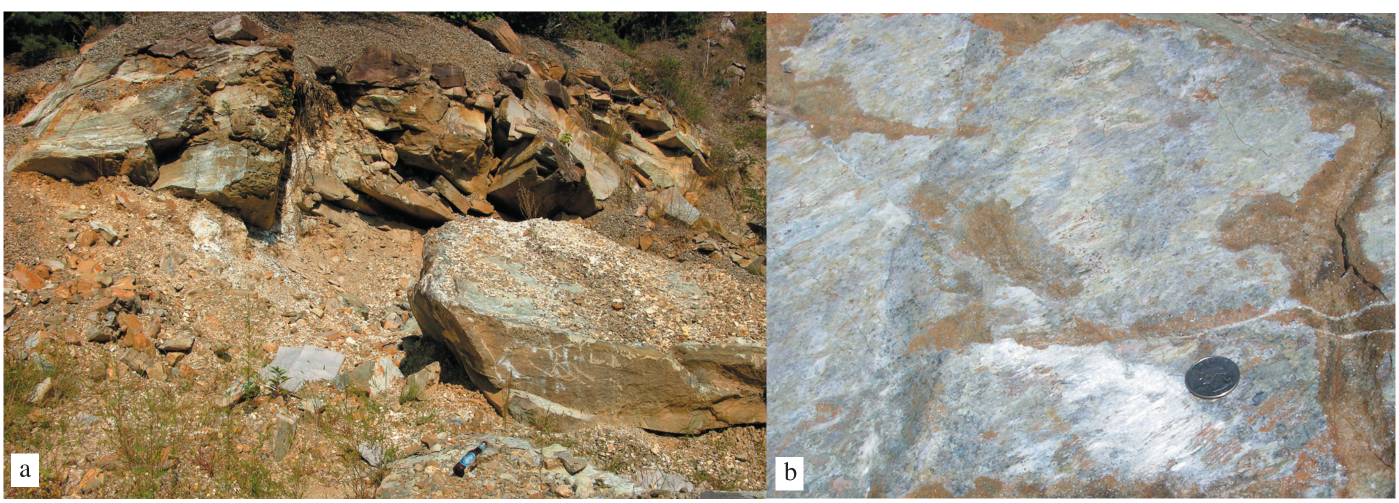


Figure 2–3–1. Dunite and alteration products at Addie Quarry. (a) Dunite with fractures coated with talc, anthophyllite, and minor antigorite. The common accessory mineral in fresh dunite is chromite. (b) Anthophyllite-talc-coated surface on a dunite block.

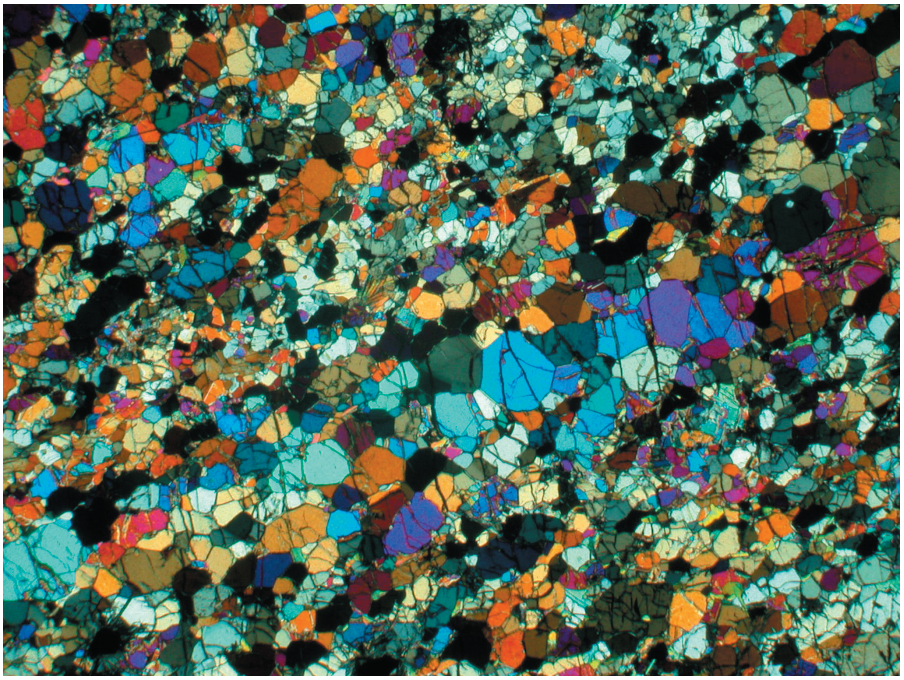


Figure 2–3–2. Photomicrograph of dunite from Blanton Branch Road intersection with US 23-74, ~2 km east Stop 2–3. Polygonal olivine dominates the fabric of this sample, with minor amounts of chromite, and alteration products. Serpentine alteration can be seen along cracks in the maroon-colored grain near the right boundary of the photo.

(0-16%), chlorite (0-3%) and calcic amphibole (0-3%). Chlorite forms lath-like aggregates surrounding chromite crystals and separating the chromite from contact with olivine. Chromite in this association typically has lattice or poikiloblastic texture (Lipin, 1984); where chlorite is absent, "clean" chromite crystals occur.

A whole-rock chemical analysis of fresh metadunite from the Blanton Branch Road outcrop is as follows (in wt%): SiO_2, 41.98; Al_2O_3, 0.46; Cr_2O_3, 0.35; FeO, 7.84; MgO, 44.56; MnO, 0.11; NiO, 0.31; CaO, 0.50; Na_2O, 0.04; LOI (loss on ignition), 3.0.

Return to US 23–74 and turn left (north), drive 1.6 mi and turn left onto SR1432 Skyline Drive, then turn right in ~0.1 mi onto Willets Rd SR1537. Drive ~0,6 mi to the intersection with Dark Ridge Rd. Park and walk up the hill on an abandoned segment of old U.S.23–74.

STOP 2–4. Cartoogechaye Terrane Amphibolite on Old U.S. 23–74 at Willits–Ochre Hill. Elevation: 2,520 ft **Leader:** Steve Yurkovich. 25 minutes

GPS Location: 35º 23.987' N; 83º 09.592' W.

Purpose: To examine some high-grade amphibolite in the Cartoogechaye terrane.

Description: Cartoogechaye terrane amphibolite exposed here consists predominantly of coarse-grained amphbolite, with some interlayered garnet-bearing biotite gneiss (Fig. 2–4–1). These rocks are very migmatitic, and although this is one of the block-in-matrix localities described by Raymond et al., (1989), they emphasized standard migmatite texturees in their deacription, identifying blocky migmatite (agmatite and rafts or schollen), and less common layered migmatite (stromatic migmatite).

End of field trip. Return to the Quality Inn in Maggie Valley.

Figure 2–4–1. Blocky migmatite (agmatite) at Stop 2–4, with blocks of coarse amphibolite in a more felsic biotite, garnet matrix. Much of the amphibolite at this locality is massive, but the large strain in these rocks is evident wherever there is enough layering to delineate the strain.

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