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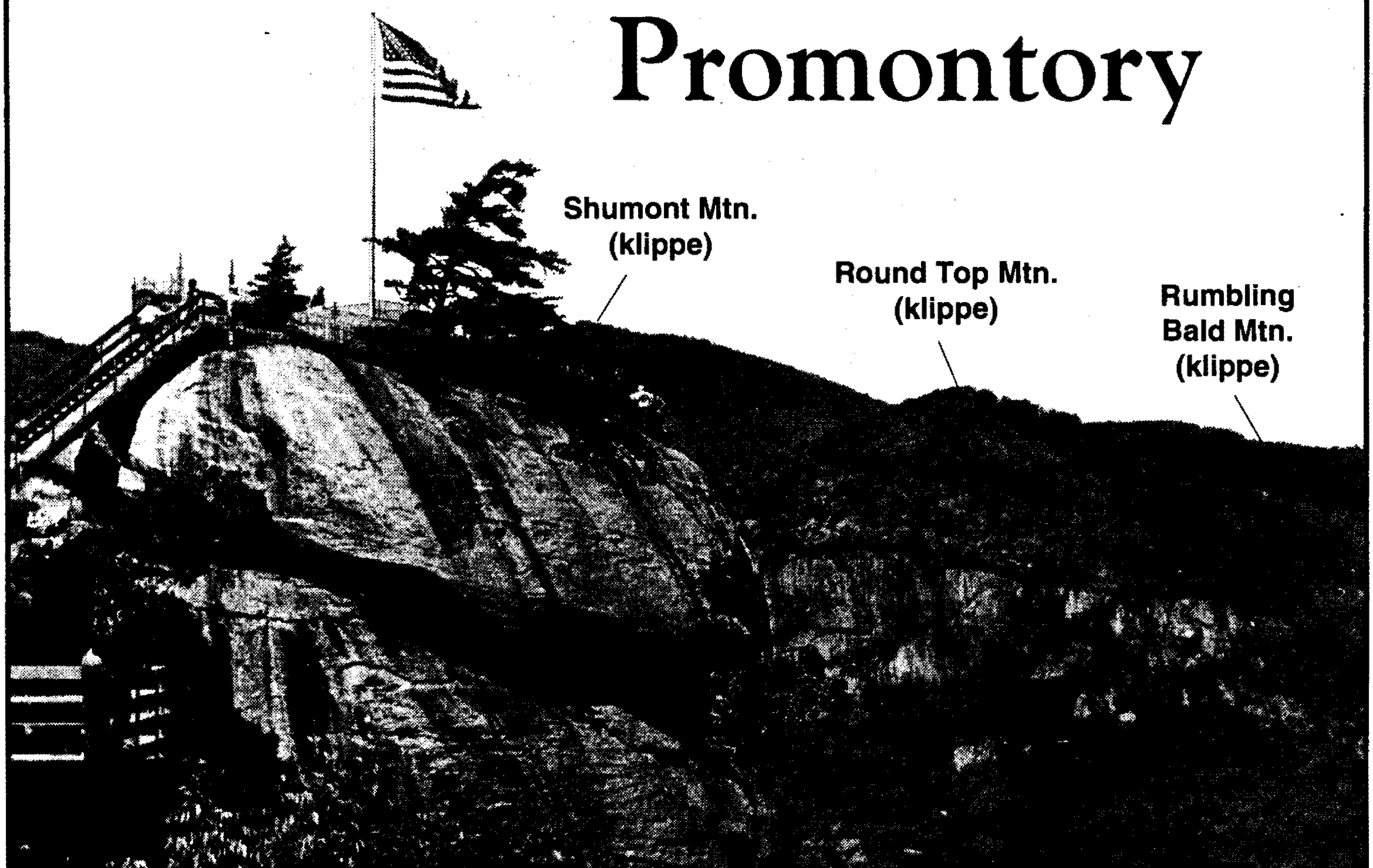
Studies of Inner Piedmont Geology with a focus on the Columbus Promontory



Carolina Geological Society Annual Field Trip

November 6-7, 1993

Guidebook edited by Robert D. Hatcher, Jr. & Timothy L. Davis
Field Trip Leaders: Timothy L. Davis & Gregory M. Yanagihara

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Cover Photograph:

View northeast from Chimney Rock Park. Chimney Rock, composed of 509 Ma Henderson augen gneiss is in the foreground, and the exfoliated cliffs across the valley are also Henderson Gneiss. The near-horizontal Sugarloaf Mountain thrust on Round Top Mountain that placed Poor Mountain Formation rocks over Henderson Gneiss is located in the trees about half way from the cliffs to the top of the mountain.

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PERSPECTIVE ON THE TECTONICS OF THE INNER PIEDMONT, SOUTHERN APPALACHIANS

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ABSTRACT

The purpose of the 1993 Carolina Geological Society field trip is to discuss structural and stratigraphic relationships in the western Inner Piedmont in the Columbus Promontory, in the Hendersonville-Tryon area, North Carolina, and to compare this with other parts of the Inner Piedmont and southern Appalachians. The stratigraphy making up most of the core and SE flank of the belt consists of an assemblage of metagraywacke, aluminous schist, and amphibolite, deposited on an unknown basement. This stratigraphic assemblage resembles the widespread lower unit in the eastern Blue Ridge that in places overlies Grenville basement, but mostly overlies an unknown—probably oceanic—basement. Overlying the lower assemblage on the NW and SE flanks of the Inner Piedmont is an upper sequence of metasiltstone/ metapelite, and amphibolite with some quartzite and marble (Poor Mountain and Chauga River Formations). The Henderson Gneiss, a 509 Ma granitoid, was probably tectonically emplaced onto the upper sequence in the western Piedmont near the metamorphic peak. These assemblages were intruded by numerous Paleozoic granitoids, along with some intermediate and mafic plutons. Metamorphism reached upper amphibolite grade conditions in the core and lower amphibolite grade conditions locally along the flanks.

The map-scale structure of the Inner Piedmont consists of a stack of westward-vergent fold-thrust nappes that are rooted along the SE flank. The western flank, called the Chauga belt, consists of an early syncline separating anticlinal structures in the eastern Blue Ridge and Inner Piedmont. The SE flank (part of the Kings Mountain belt) may also represent an early syncline truncated by the central Piedmont suture. The meso-scale structure of the Inner Piedmont is dominated by low dip of the dominant S_2 foliation that is folded into a regional antiform in the Carolinas and NE Georgia, but into a synform in Alabama and western Georgia. Extensive use of kinematic indicators in the Columbus Promontory has demonstrated that the dominant foliation throughout the western Inner Piedmont is a C-foliation that formed near the metamorphic peak. Kinematic indicators reveal W-directed transport dominated in the more internal parts of the Inner Piedmont that turns SW in the more western parts of the Inner Piedmont in both the Columbus Promontory and in NW South Carolina. The pervasive character of this C-surface suggests the entire western Inner Piedmont behaved as a crustal-scale shear zone that was buttressed against the present site of the Brevard fault zone and was decoupled from the eastern Blue Ridge. Based on the 438 Ma age of a deformed pluton in the Hendersonville area, the crustal-scale shear zone probably formed during the Acadian orogeny after the Taconian orogeny deformed the Blue Ridge and Inner Piedmont. This feature helps explain some of the previously unexplained differences between the Inner Piedmont and eastern Blue Ridge.

The value of suggestions by a number of geologists that the Inner Piedmont is either a suspect or an exotic terrane is diminished by evidence from several localities in the Carolinas and Georgia that the lower sequence in the Inner Piedmont is identical to that in the eastern Blue Ridge.

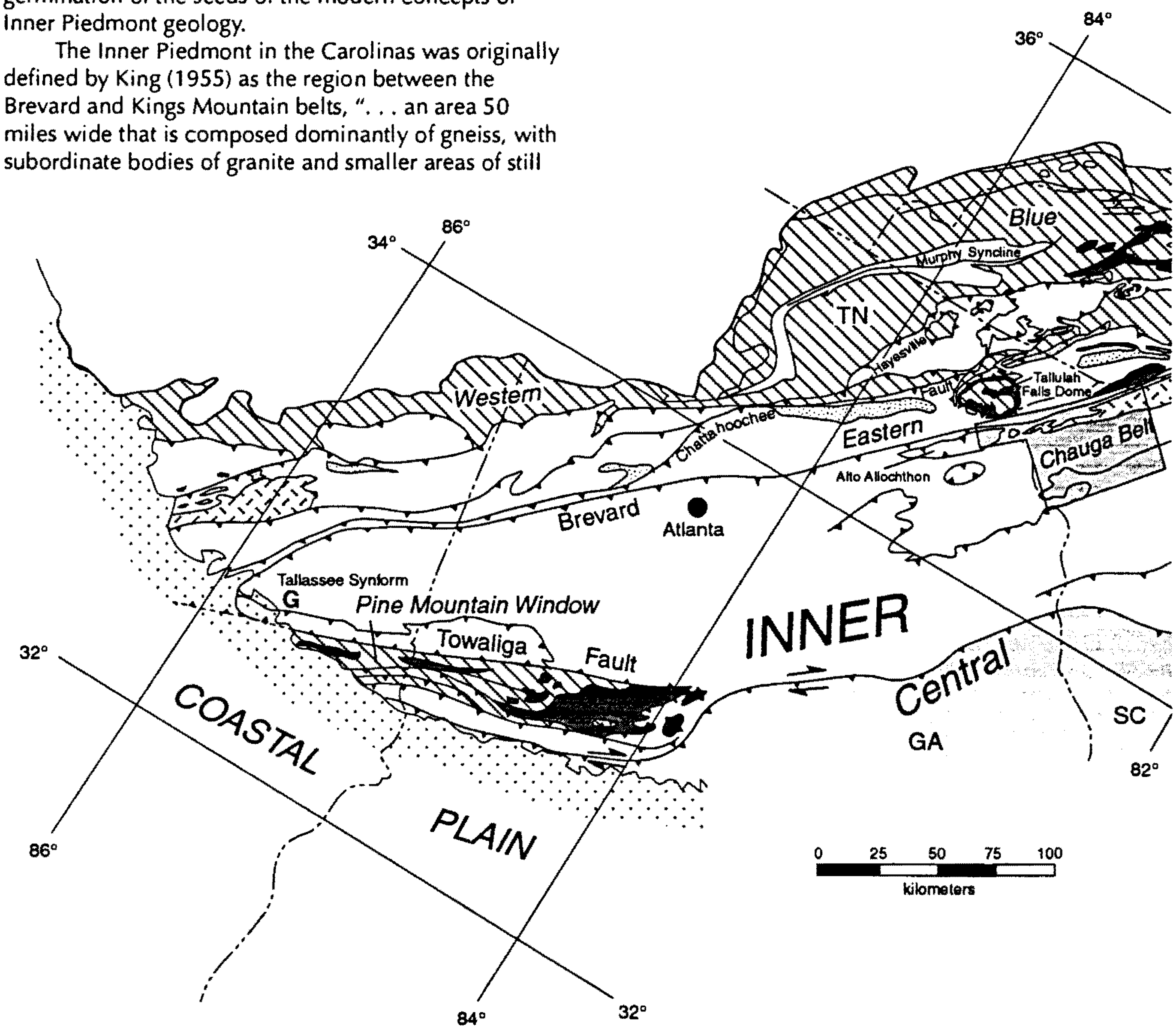
Post-Paleozoic history of the Inner Piedmont involved mid- to late Mesozoic extension that produced a brittle-ductile fault system of small displacement and intrusion of diabase dikes. This was followed by reversal in the Late Cretaceous of the extensional stress field to compression, uplift, and erosion throughout the Cenozoic.

*Managed by Martin Marietta Energy Systems, Inc. for the U.S. Department of Energy under contract DE–AC05–84OR21400.

INTRODUCTION

The Carolina Geological Society (CGS) Field Trip for 1993 will visit the western Inner Piedmont in the Columbus Promontory in southwestern North Carolina (Fig. 1). Previous and more recent CGS trips have focused on parts of both the Inner Piedmont and other tectonic units in Virginia and North Carolina (Price and others, 1980), North and South Carolina (Horton and others, 1981), and the southwestern end of the Sauratown Mountains window in North Carolina (Hatcher, 1988a). The last time similar rocks were visited by the CGS was in 1969 (Griffin, 1969; Hatcher, 1969), when we looked at the Inner Piedmont, Chauga belt, and Brevard fault zone in northwestern South Carolina, with the field trip headquartered in Clemson, South Carolina. The 1969 trip may have viewed the germination of the seeds of the modern concepts of Inner Piedmont geology.

The Inner Piedmont in the Carolinas was originally defined by King (1955) as the region between the Brevard and Kings Mountain belts, ". . . an area 50 miles wide that is composed dominantly of gneiss, with subordinate bodies of granite and smaller areas of still other rocks." Previously, Adams (1926, 1933) and Crickmay (1952, p. 6) called the same assemblage the Dadeville belt in Georgia and Alabama, and Crickmay indicated that the rocks of this belt are the same as those in the Tallulah belt (eastern Blue Ridge northwest of the Brevard fault zone). Bentley and Neathery (1970) subdivided the Inner Piedmont in Alabama and western Georgia into an upper series they called the Opelika complex. The Opelika complex consists of metagraywacke and pelitic schist thought to be separated unconformably (Stone Wall line) from the lower Dadeville complex of gneiss, schist, granitoids, mafic, and ultramafic rocks. Hatcher (1969, 1972) suggested the lower grade higher(?)-level rocks of the western Inner Piedmont in South Carolina should be separated into a different assemblage (metasiltstone, laminated amphibolite, quartzite, and marble—Poor Mountain and



Chauga River Formations—and the Henderson Gneiss) he suggested be called the Chauga belt, with possible equivalents in the Kings Mountain belt. Similar units have been recognized and traced throughout the western Inner Piedmont of Georgia and Alabama (Waresville formation, Jacksons Gap group, Ropes Creek amphibolite) by Bentley and Neathery (1970) and by Higgins and Atkins (1981) (parts of their Atlanta Group, a term subsequently abandoned by Higgins and others, 1988). Grimes and others (this guidebook) have demonstrated that the stratigraphic sequences in the Brevard fault zone and eastern Blue Ridge in Alabama wrap around the southwest end of the Inner Piedmont to connect with rocks in the Inner Piedmont northwest of the Pine Mountain window. The late (Alleghanian) brittle deformation in the Brevard fault zone, however, continues uninterrupted until it disappears beneath the Coastal Plain. Similar rocks also occur in the western Inner Piedmont of North Carolina (Goldsmith and others, 1988; Davis, 1993, this guidebook; Yanagihara, this guidebook).

Jonas (1932) and King (1955) both recognized the prevalence of gentle dip across much of the western Inner Piedmont, but neither put forth a plausible explanation for it. Although Clarke (1952) first suggested that the Brevard, Towaliga, and Goat Rock faults are all connected as a single large thrust nappe, an interpretation endorsed by Bentley and Neathery (1970), V. S. Griffin, Jr.'s incisive work should be considered the key to actual resolution of the internal structural style of the Inner Piedmont. Griffin (1967, 1969, 1974) suggested that this style consists of a stack of fold nappes analogous to those in the Pennine zone of the Alps. More recently, the polyphase nature of the ductile deformation has been better resolved by Bentley and Neathery (1970), Hatcher (1974, 1978a, 1989), Atkins

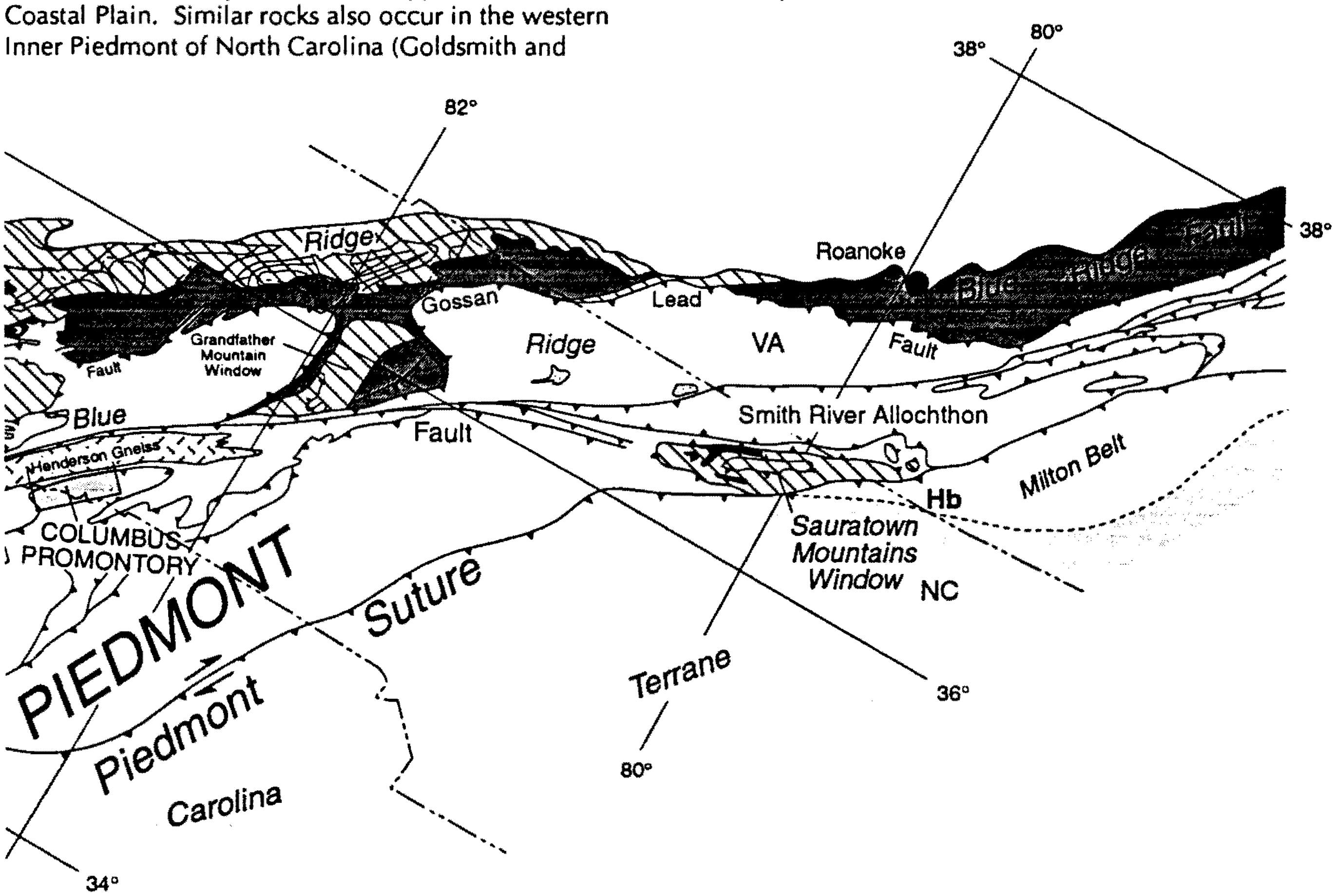


Figure 1. Index map of major tectonic units surrounding the Inner Piedmont, with the box in North Carolina indicating the Columbus Promontory, and that in South Carolina and northeastern Georgia the area of Figure 2. G – Location of contribution by Grimes and others (this guidebook). Hb – Location of the contribution by Hibbard (this guidebook). All other papers are in the Columbus Promontory.

and Higgins (1978, 1980), and McConnell and Abrams (1984), and a number of additional thrust sheets have been recognized (Nelson and others, 1987; Horton and McConnell, 1991). After more than two decades of scrutiny, however, the basic terrane-scale structural style suggested by Griffin remains unchanged in current reconstructions, and appears correct.

The purpose of the 1993 CGS field trip and guide is to present and discuss some of the geology worked out recently in the Columbus Promontory, point out some of the interesting unsolved problems that remain both in bedrock and surficial geology, and to present several additional contributions to the guidebook by others working elsewhere in the Inner Piedmont.

STRATIGRAPHY

The Inner Piedmont presents an initially level of complexity to resolution of the stratigraphy. This region is mostly metamorphosed to middle- and upper-amphibolite facies assemblages, is migmatitic, has been intruded by many deep- and shallow-level plutons, and has been multiply deformed. Despite what appear to be insurmountable difficulties facing any attempt to resolve the stratigraphy here, several geologists have recognized there is a stratigraphic order to this terrane. The western and eastern flanks are locally not as high grade (as low as garnet and staurolite grade) or migmatitic as the core, and continuous sequences of similar distinctive rock units have been recognized in the Inner Piedmont in widely separated parts of the terrane. Moreover, these more internal assemblages in the Inner Piedmont, because of the appearance of the same marker units and lithologies in almost the same position in the sequences, suggest they may be direct correlatives with the those immediately northwest of the Brevard fault zone from Alabama to Virginia (Hatcher, 1975, 1989). They (Fig. 2a) consist of various proportions of metagraywacke, aluminous schist, and amphibolite in different parts of the sequence that bear different names in different places: Lynchburg Formation in central Virginia; Tallulah Falls and Ashe Formation in southern Virginia, North and South Carolina, and northeastern Georgia; Sandy Springs Group in central and western Georgia; and Emuckfaw group in Alabama. Basement rocks are unknown in the Inner Piedmont.

Intact stratigraphic sequences identical to those in the eastern Blue Ridge have been recognized in the Inner Piedmont near the Sauratown Mountains (Heyn, 1988), in northwestern South Carolina (Hatcher, 1972, 1978a, 1989; Hatcher and Acker, 1984), in the Alto allochthon in northeastern Georgia and northwestern South Carolina (Hatcher, 1978b; Hopson and Hatcher, 1988), and in the Atlanta area (Higgins and Atkins, 1981; McConnell and Abrams, 1984). These contain recognizable and traceable marker units of aluminous schist and occasionally calcsilicate. Similar assemblages without the distinctive traceable stratigraphic markers have been recognized northwest of the Pine Mountain window in central Georgia (Hooper and Hatcher, 1989), in the Inner Piedmont in central South Carolina (Horkowitz, 1984; Willis, 1984), and in the Columbus Promontory (Davis, 1993, this guidebook; Yanagihara, this guidebook).

Overlying the graywacke-pelitic schist-amphibolite sequence in stratigraphic continuity in the western Inner Piedmont is another sequence consisting of a lower metasiltstone unit overlain by laminated to massive amphibolite and minor amounts of graphitic phyllite (Fig. 2a). The amphibolite is locally overlain by clean, to feldspathic, to calcareous quartzite, and impure to pure marble. All units grade vertically into each other and are extensively interlayered, so, while a map unit that is indicated as amphibolite is dominantly amphibolite, it commonly contains substantial amounts of metasiltstone, some quartzite, and locally some light-colored marble. This is the Poor Mountain sequence originally defined by Shufflebarger (1961), where he pointed out the similarities with the Evington Group in Virginia. The name Poor Mountain Formation, using the type area of Poor Mountain (now called Buzzard Roost Mountain on the Whetstone 7 1/2 minute quadrangle) near Walhalla, South Carolina, was suggested by Hatcher (1969) for this unit because of its mappable extent. The Poor Mountain Formation, as defined by Hatcher (1969), was recognized in the Hendersonville area by Lemmon (1973), and more recently has been traced throughout the Columbus Promontory (Davis, 1993, this guidebook; Yanagihara, this guidebook), and, because of similar lithologies mapped along strike in the Charlotte 1 x 2° sheet (Goldsmith and others, 1988), probably extends to the latitude of the Grandfather Mountain window. The Poor Mountain Formation has been traced as a continuous unit southwestward to the vicinity of Gainesville, Georgia (Hatcher, 1978b), and the amphibolite unit is chemically similar to the Ropes Creek amphibolite in Alabama and western Georgia (Davis, 1993).

Part of the Poor Mountain Formation, and Tallulah Falls Formation in the western Inner Piedmont of northwestern South Carolina has been termed the Walhalla metamorphic suite by Horton and McConnell (1991). I recommend that this term be immediately abandoned because it incompasses parts of several easily separated stratigraphic units by defining a petrologic unit based solely on metamorphic grade. Such units are very difficult to map over wide areas other than in reconnaissance.

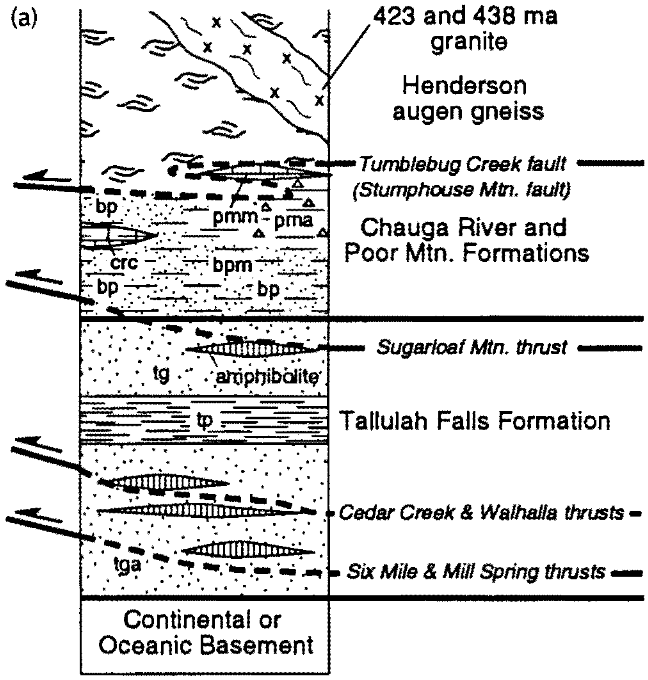
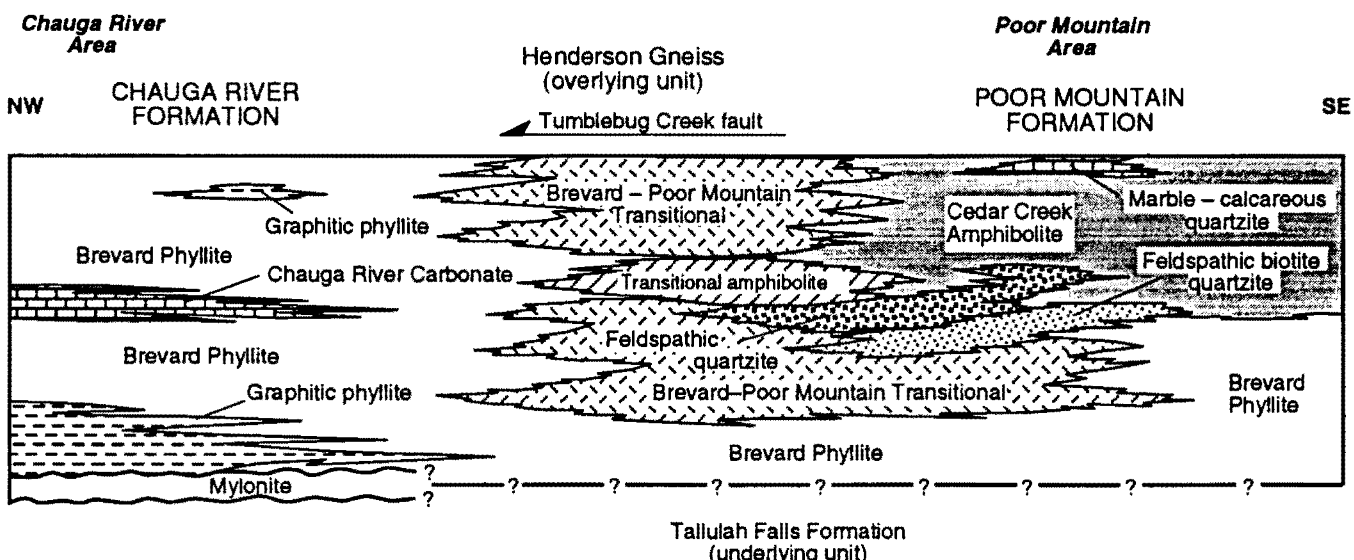


Figure 2. (a) Schematic diagram outlining the stratigraphy recognized in several localities in the Inner Piedmont using the terminology adopted (by Hatcher) for northwestern South Carolina. Tallulah Falls Formation terminology employed is that originally recognized in the eastern Blue Ridge. Tallulah Falls Formation units: tga–metagraywacke-schist-amphibolite member. tp–garnet-aluminous schist member. tg–metagraywacke-schist member. Chauga River and Poor Mountain Formations units: bp–Brevard phyllite. bpm– Brevard-Poor Mountain transitional member. crc–Chauga River carbonate member. pma–Poor Mountain Amphibolite. pmm–Poor Mountain Marble. (b) Stratigraphic relationships between the Chauga River and Poor Mountain Formations.

(b)



The sequence in the Brevard fault zone of graphitic phyllite, muscovite-chlorite phyllite, dark impure (quartz, muscovite) marble, and impure (feldspar, muscovite) quartzite, along with minor laminated amphibolite, has been correlated by Hatcher (1969) as facies equivalents of the Poor Mountain Formation (Fig. 2b). Hatcher (1969) named Chauga River Formation because of the excellent exposure of the unit within and southeast of the Brevard fault zone along the Chauga River and its tributaries in northwestern South Carolina. Equivalence of the Poor Mountain and Chauga River Formations is indicated by the dominance of the muscovite-chlorite metasiltstone in the lower parts of some sections of Poor Mountain Formation and the local occurrence of laminated amphibolite in the upper part of the Chauga River Formation. Recognition of other sites where these sequences may occur awaits additional detailed mapping (≥1:24,000 scale with all streams and ridges walked).

The Henderson Granite was named by Keith (1907) for exposures in Henderson County, North Carolina, and was subsequently renamed the Henderson Gneiss (Reed and Bryant, 1964). More recently, geologists have recognized that it is composed of several mappable subunits in the Chauga belt (for example, Cazeau, 1967; Hatcher, 1969; Roper and Dunn, 1973; Lemmon, 1973). The principal lithology throughout this area in North and South Carolina is coarse-grained microcline-plagioclase (oligoclase)-quartz-biotite-muscovite augen orthogneiss generally referred to as the Henderson augen gneiss. The augen are composed of microcline, some perthitic, and they are frequently surrounded by myrmekite. The Henderson Gneiss has a Rb-Sr whole-rock age of 509 Ma (Odom and Fullagar, 1973; recalculated by Sinha and others, 1989). It is located in thrust contact above the Chauga River and Poor Mountain Formations (Fig. 2).

Many large bodies of granitoid gneiss, including the Table Rock plutonic suite of Horton and McConnell (1991), have intruded Chauga belt and Inner Piedmont. These mostly consist of plagioclase-dominant, fine- to medium-grained equigranular foliated oligoclase-microcline-quartz-biotite granodiorite to leuco-quartz diorite bodies, but microcline-dominant megacrystic granitoids also occur here. These separated bodies of granitoid gneiss in the western Inner Piedmont in South Carolina and northeastern Georgia apparently are similar, because, on a single composite isochron plot, they yield a Rb-Sr age of about 423 Ma (Harper and Fullagar, 1981). Hatcher (1989) concluded they are probably also coeval and consanguineous with eastern Blue Ridge granitoids. Five regionally foliated granitoid plutons within the Inner Piedmont of Georgia and South Carolina and one undeformed pluton that straddles the Middleton-Lowndesville fault have preliminary $^{207}Pb/^{206}Pb$ zircon age dates. The late Paleozoic Cold Point pluton is traceable from the Inner Piedmont into the Carolina terrane and yielded has an age of 312 Ma, and the middle Paleozoic Inner Piedmont granitoids have ages ranging from 385 to 435 Ma (T. W. Stern, written communication, 1986, as cited in Nelson and others, 1987).

STRUCTURE

The meso-scale structure of the Inner Piedmont is dominated by low dip of the dominant foliation (S_2, or later), with locally steep dip, particularly along the eastern margin that Griffin (1974, 1978) called the SE flank. Even along the SE flank, however, shallow dip may dominate as the central Piedmont suture (Hatcher and Zietz, 1980) is folded exposing Inner Piedmont assemblages in windows southeast of the main boundary trace (Horkowitz, 1984; Willis, 1984). These rocks are foliated, indicating even most plutons were emplaced prior to or near the thermal peak when the rocks were plastic. Late granitoids, pegmatite dikes, and Mesozoic diabase dikes are either not foliated or only weakly so.

The dominant foliation in the Inner Piedmont in the northwestern South Carolina, southwestern North Carolina, and northeastern Georgia defines an antiformal structure (Griffin, 1971a, 1971b, 1974; Hatcher, 1972; Goldsmith, 1981; Hatcher and Hooper, 1992), but defines a synform that Bentley and Neathery (1970) called the Tallassee synform in western Georgia and Alabama (Crickmay, 1952; Clarke, 1952). The Inner Piedmont in part of North Carolina is a foliation synform that refolds earlier antiforms and sunforms related to nappe emplacement Goldsmith and others, 1988). The antiformal structure in the Carolinas is probably related to the primary kinematics of emplacement of the westward-directed fold-and-thrust nappes. The synformal character of the Inner Piedmont in Alabama and western Georgia is probably related to refolding by later (Alleghanian) deformation of the earlier fold-and-thrust nappes also present here. This later deformation initially involved strike-slip displacement along the Brevard, Ocmulgee (Hooper and Hatcher, 1989), Dean Creek (Hatcher and others, 1988), and other eastern Piedmont fault system faults, then was further modified by late Alleghanian brittle reactivation of the Brevard fault zone and emplacement of the Blue Ridge-Piedmont megathrust sheet. The structure of the southwestern end of the Inner Piedmont, originally suggested by Bentley and Neathery (1970, their Figs. 16 & 18) and earlier by Clarke (1952), is now confirmed to be a NE-plunging synform flanked by the sequence in the Brevard fault zone and eastern Blue Ridge that can be traced around the end. Additional evidence of separation of the major Inner Piedmont structure into early high-temperature ductile and later retrograde brittle phases has been reported from the southwestern exposed parts of the Inner Piedmont by Grimes and others (this guidebook).

The dominant S_2 near-thermal peak foliation present across much of the western Inner Piedmont has recently been demonstrated to be a penetrative C-foliation involving W-and SW-directed transport that transposed an earlier S-surface (Davis, 1993, this guidebook). The combination of a number of kinematic indicators (asymmetric porphyroclasts and boudins, S-C fabrics) and axes of sheath folds demonstrates the strong linear fabric in many rocks is a transport lineation, and was the key to this interpretation.

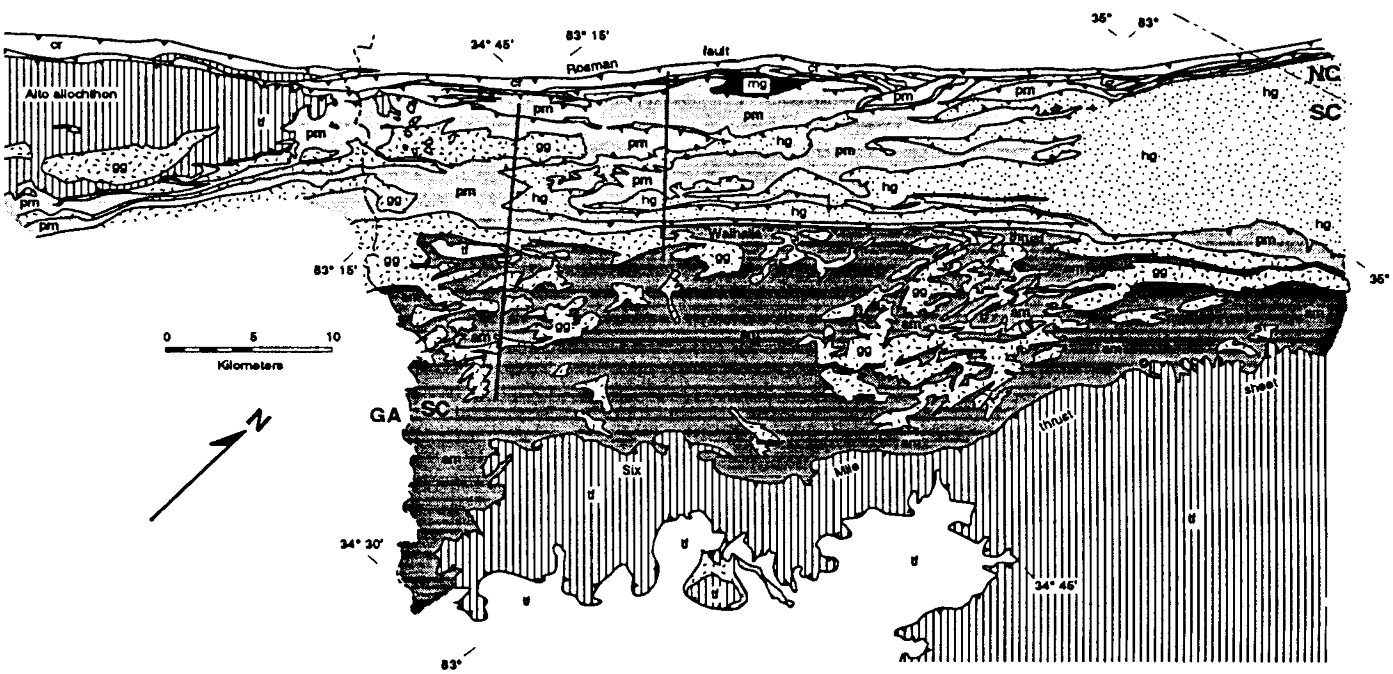
Several generations of folds are present in the western Inner Piedmont and Chauga belt. Early (plastic) folds are isoclinal to tight, with interlimb angles up to 70° (some sheath folds), and later (brittle) folds are more open, with interlimb angles that range from 60 to 120°. Fold interference patterns are recognizable on all scales involving interference of early folds with each other and with later folds (Hatcher, 1974a, 1978a, 1989). In addition, the recognition of sheath folds throughout the Inner Piedmont of the Carolinas helped pinpoint the early transport fabric and determine the orientation of slip lines (Yanagihara, this guidebook).

Late plutons intruding the Inner Piedmont are commonly less deformed and are Alleghanian. Examples include the Elberton Granite and other ~325 Ma plutons in the Atlanta area (McConnell and Abrams, 1984) and the 317 Ma Cold Point pluton (Horton and McConnell, 1991). The older plutons, like the Henderson Gneiss, the 423 Ma granitoid that intruded the Chauga belt in South Carolina, and the 438 Ma pluton that intruded the Henderson Gneiss in the Columbus Promontory area, are commonly polydeformed: superposed foliations provide evidence of at least two penetrative deformational events, and detailed map patterns indicate participation in multiple episodes of folding (for example, Fig. 3). The contacts of early plutons (mid-Paleozoic and older) are concordant with the enclosing rocks because of tectonic transposition, whereas those of the late (Alleghanian) plutons can usually be demonstrated to be more crosscutting.

Recognition of the overall mylonitic character of the progressively metamorphosed Henderson Gneiss led to the conclusion that the whole unit may have been emplaced along a premetamorphic thrust that accompanied the earliest deformation affecting the rocks of the Chauga belt, and was overprinted later by Alleghanian deformation and retrograde metamorphism. Stop 2 is an exposure of the fault contact between the Henderson Gneiss and Poor Mountain Formation.

The mylonitic overprint in the Henderson augen gneiss in the Brevard fault zone is significant. Both porphyroblasts and groundmass have been retrograded by Alleghanian deformation. Plagioclase porphyroblasts and muscovite formed at the expense of K-spar and perhaps earlier plagioclase. Chlorite locally formed at the expense of biotite, along with retrograde muscovite, define the mylonitic foliation in micaceous rocks. The resulting porphyroclastic blastomylonite has a distinctly different appearance from the original augen gneiss, which is an annealed mylonite. A gradual change from the prograde Henderson augen gneiss to the mylonitic variety may be observed in northwestern South Carolina and along U.S. 64 west of Rosman, North Carolina (see Horton and Butler, 1986). Mylonite derived from Henderson Gneiss near Rosman yielded Rb/Sr whole-rock ages of 355 Ma (Odom and Fullagar, 1973) and 273 Ma (Sinha and others, 1988), and a zircon Pb-Pb

Figure 3. Simplified geologic map of the Inner Piedmont, including the Chauga belt and Brevard fault zone, in northwestern South Carolina and part of northeastern Georgia. am–amphibolite, hornblende gneiss, and biotite gneiss. tf–Tallulah Falls Formation. gg–Paleozoic granitoids (cross-hatch pattern). hg–Henderson Gneiss (stippled pattern). pm–Poor Mountain Formation. cr–Chauga River Formation. mg– mylonite gneiss. Oc–Knox Group (Ordovician) horse in Brevard fault zone. Six Mile thrust sheet and Alto allochthon are vertical lined. Teethed lines are thrusts (teeth on upper plate). Modified from maps in Griffin (1974), Hopson and Hatcher (1988), and Hatcher (published and unpublished data). Section lines in Figure 4 are indicated by heavy lines.

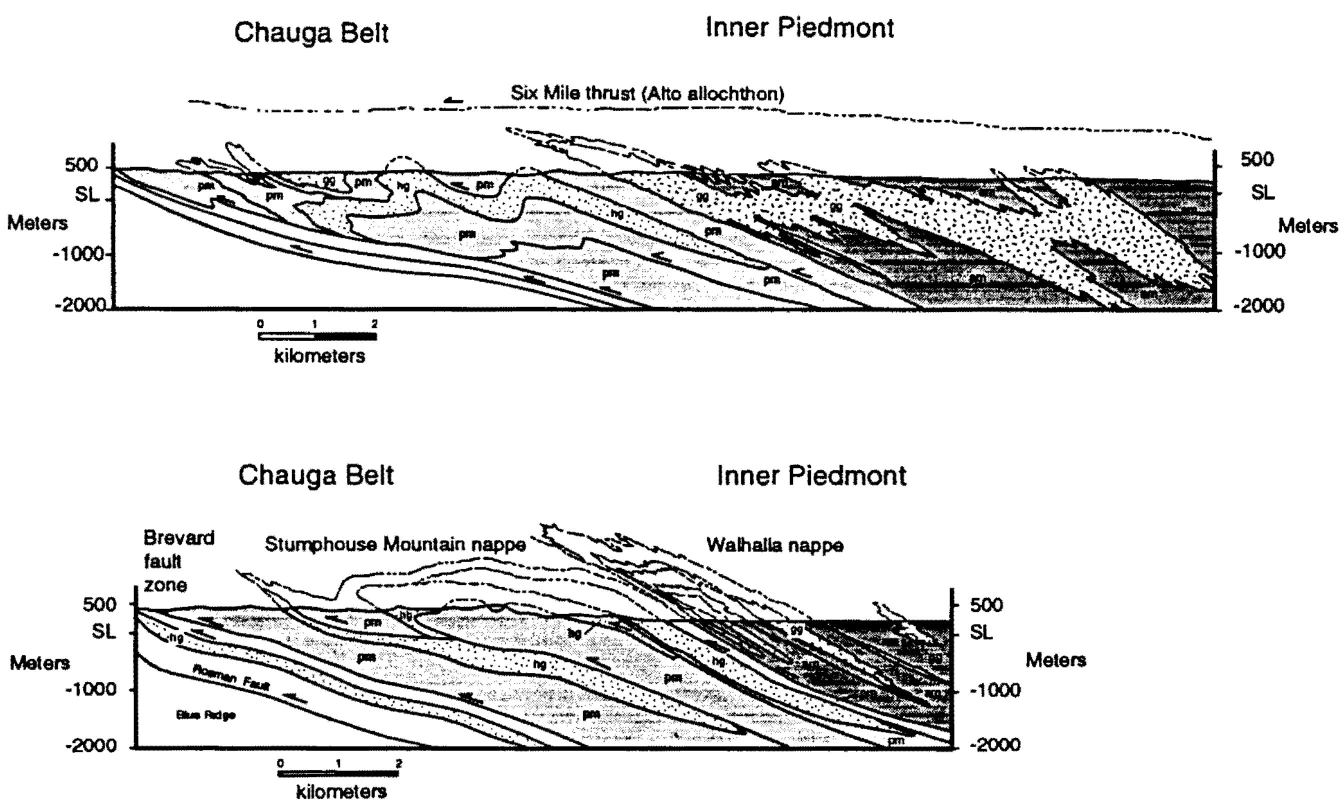


age of 480 Ma (Sinha and Glover, 1978). These ages probably indicate the Henderson Gneiss crystallized at 509 Ma, was metamorphosed at 480 (Taconic), and 355 (Acadian) Ma, and later retrograded during Alleghanian ductile deformation, although the 355 Ma event has been suggested by Sinha (pers. comm.) to be a mixing line. The appearance of late Paleozoic radiometric ages (see for example Goldberg and Fullagar, this guidebook) in the Inner Piedmont suggests either unroofing occurred at this time or a major thermal event affected the rocks during the Alleghanian.

Griffin (1967, 1969, 1974) first showed that large fold and thrust nappes exist in the Inner Piedmont, and suggested map-scale structure of the Inner Piedmont consists of a stack of westward-vergent fold-thrust nappes that are rooted along the SE flank. Griffin (1967, 1971a, 1974) recognized three major nappes in the Inner Piedmont of South Carolina and nearby Georgia: the Walhalla, Six Mile and Anderson nappes. The Walhalla nappe consists of the sheared-out limb of an antiform-synform pair, a fold nappe with the fault called a tectonic slide. Most of the thrust sheets in the Inner Piedmont and Chauga belt began as fold nappes, as Griffin suggested earlier, and comprise the Type F thrusts of Hatcher and Hooper (1992), using Griffin's Walhalla nappe as a type example. As the thrust formed along the common limb between the antiform and synform and transport progressed so that the two hinges became separated by a significant distance, some of the thermal energy employed to heat the rocks my have been transformed into mechanical energy, and the rock mass cooled imparting additional strength to the thrust sheet. This may have permitted thrust sheets like the sillimanite grade Six Mile thrust sheet to move several 10s of km over even cooler garnet-grade rocks of the Chauga belt and leave the dismembered Alto allochthon as the region was deeply eroded. The thrust sheets of the Columbus Promontory are also of this type. The western flank of the Inner Piedmont, called the Chauga belt (Hatcher, 1972), consists of an early syncline separating anticlinal structures in the eastern Blue Ridge and Inner Piedmont. The SE flank (part of the Kings Mountain belt) may also represent an early syncline truncated by the central Piedmont suture.

More recently, the extent of several of these structures has become better known and others, such as the Alto allochthon, have been recognized in both northeastern Georgia and northwestern South Carolina (Figs. 3 and 4; Hatcher, 1978b, 1989; Nelson and others, 1987; Hopson and Hatcher, 1988; Horton and

Figure 4. Cross sections through the Inner Piedmont in part of northwestern South Carolina. See Figure 3 for explanation of symbols and patterns, and locations of section lines.



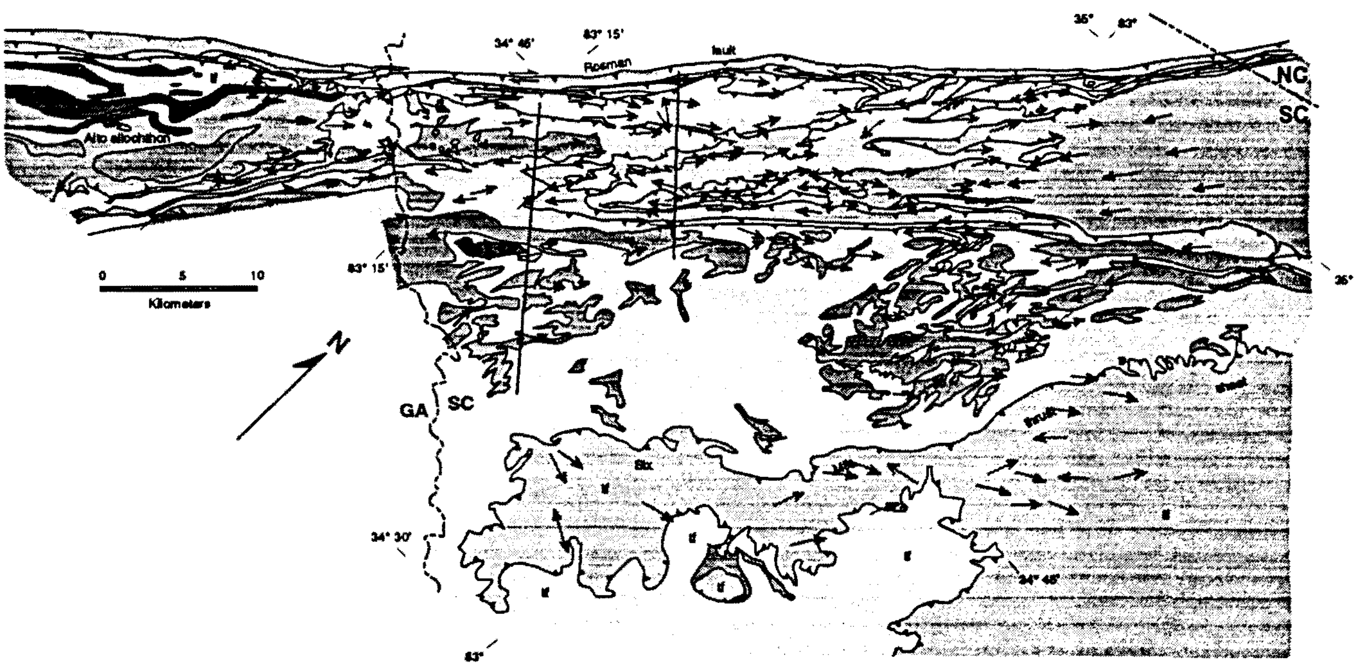
McConnell, 1991; Liu, 1991), a similar, but somewhat distinctive, stack in the Columbus Promontory (Davis, 1993, this guidebook; Davis and others, in review; Yanagihara, this guidebook), and the Smith River allochthon in North Carolina and Virginia (Conley and Henika, 1973).

Thrust transport apparently was initially west-directed, but, either by buttressing or otherwise, was redirected toward the southwest, producing a dextral sense of transport against the Brevard fault zone (Davis, 1993, this guidebook; Davis and others, in review). Interestingly, although the evidence from the linear fabric data for dextral transport is not as clearcut in South Carolina, the detailed map pattern in the Chauga belt and westernmost part of the higher grade Inner Piedmont (Fig. 5) is best explained as a series of SW-directed sheath folds. Most of the linear fabrics here trend SW. This episode of structural development may have occurred during the Acadian orogeny (Davis and others, in review). These structures later may have functioned as a weak zone to localize the early Alleghanian dextral strike-slip deformation along the Brevard fault zone, thus accounting for the origin and maintenance of a listric strike-slip fault surface. Bobyarchick (1984) was the first to firmly document dextral Alleghanian motion on the Brevard fault, although Reed and Bryant (1964) suggested that it had 135 miles of dextral displacement. They (Reed, Bryant, and Myers, 1970) later reinterpreted the motion sense, however, suggesting that it was sinistral, with a thrust component. A number of summaries of Brevard fault zone history and the multitude of interpretations have been written (see Bobyarchick and others, 1988; Hatcher, 1989; and Hatcher and others, 1989, and references cited in each), but most of the alternative interpretations address only the Alleghanian deformation.

INNER PIEDMONT AS A SUSPECT OR AN EXOTIC TERRANE

Partly because of the presumed fundamental nature of the Brevard fault zone (e.g., Burchfiel and Livingston, 1967; Butler, 1971), its lengthy extent across the region, linear character, pervasive mylonitization extending several km to either side of the main contact, and differences in rocks between the Blue Ridge and Piedmont, the Inner Piedmont has appeared in numerous compilations during the last decade and a half as a terrane or composite terrane separated from the Blue Ridge (Williams, 1978; Rankin, 1975; Rankin and others, 1989; Horton and others, 1989, 1991), with the Brevard fault zone treated as a suture. The Inner Piedmont thus becomes at least a suspect terrane, and has been considered exotic by some geologists. In

Figure 5. High-temperature (pre- or syn-Alleghanian) mineral lineation patterns plotted on the simplified geologic map in Figure 3. Arrows indicate orientation of mineral lineation and direction of plunge. Sources of information are the same as those for Figure 3 (and sources cited therein), with the addition of Roper and Dunn (1970).



contrast, stratigraphic data cited above from the eastern Blue Ridge and western Piedmont require that the Brevard fault not be a suture (see Hatcher and others, 1990).

The interpretation proposed by Davis (this guidebook) accounts for all of the differences and similarities between the eastern Blue Ridge and Inner Piedmont rocks. The differences in migmatitic character (actually relatively small), structural style, and timing of deformation and metamorphism (if any) in both domains can all be accounted for by this new model (see Davis, this guidebook). The crustal shear zone model requires early (but probably mid-Paleozoic—Acadian) dextral decoupling of the Inner Piedmont from the Blue Ridge along the primordial Brevard fault zone after Taconian folding, thrusting, and suturing of the distal North American assemblages described here, along with some rifted Laurentian basement, to the rifted margin of Laurentia along the Allatoona-Hayesville-Gossan Lead fault, which, together with the eastern Blue Ridge, formed the Piedmont terrane of Williams and Hatcher (1983). The central Piedmont suture bounds the Inner Piedmont on the southeast, joining the Carolina terrane to Laurentia. This probably occurred during the early to middle Paleozoic, during either the Taconian or Acadian events, or possibly during the early phases of the Alleghanian, as suggested by the new Goldberg and Fullagar age dates published herein. Suitably oriented segments of this boundary have been reactivated by dextral motion as part of the Alleghanian orogeny.

Most of the Inner Piedmont and eastern Blue Ridge structures presently exposed, except for the Alleghanian elements in the Brevard fault zone and along the central Piedmont suture, were beheaded during late Alleghanian collision of Laurentia with Africa, producing long-distance westward transport of the sheared-off segments. The Blue Ridge-Piedmont megathrust sheet was shoved as a cool slab onto the North American margin, pushing the Valley and Ridge thrusts in front of it. Maximum transport was on the order of several hundred km. The Inner Piedmont has thus functioned initially as part of the eastern Blue Ridge-Inner Piedmont block (Piedmont terrane) during the Taconian orogeny, then was partially decoupled and mobilized, producing transport on the order of a few tens of km on thrusts and a crustal-scale dextral shear zone (Acadian? or early Alleghanian?), and finally functioned as part of the great slab of middle Paleozoic and older basement that was transported during the Permian (Alleghanian) onto the North American platform.

POST-PALEOZOIC HISTORY OF THE INNER PIEDMONT

The Inner Piedmont became an element of the new Appalachian crust on the Laurentian side of the supercontinent Pangaea at the end of the Paleozoic. This crust was overthickened because of the great collision that had occurred during the Permian and immediately began isostatic readjustment toward normal thickness. The plate regime that had brought the continents together by Late Triassic time had reorganized into an extensional regime that initially rifted Pangaea and later (Middle to Late Jurassic) opened the Atlantic Ocean. While the opening process continued to widen the Atlantic, the stress regime in eastern North America again became compressional by Late Cretaceous time producing many small thrust faults and folds in Upper Cretaceous and Tertiary Coastal Plain sediments derived from continued erosional unroofing of the Appalachians (Prowell, 1988). A number of small thrust, strike-slip, and normal faults have been observed along the wave-scoured shores of reservoirs and in other exposures in the Inner Piedmont (e.g., Hatcher and Acker, 1984).

Two important phenomena appeared in the Inner Piedmont during the extensional phase of Mesozoic tectonism: formation of an extensive system of multiply reactivated quartz-filled fractures forming siliceous cataclasite and mylonite involving a regional fault system of small displacements, and a system of olivine-normative diabase dike intrusions. Diabases were originally thought to be intruded at two different times: first in a NW-trending set about 190 Ma and a more N-S, but radially oriented, set thought to be intruded about 170 Ma (Ragland and others, 1983), but Sutter (1988) dated both sets using the $^{40}Ar/^{39}Ar$ technique and both appear to have been intruded ~200Ma. All of the dikes in the southern Appalachians are olivine normative (Ragland, 1991). Both the diabase dikes and quartz-filled fractures commonly have near vertical dip. The trends of the siliceous cataclasite bodies range from near EW and NS to NE (Conley and Drummond, 1965; Garihan and others, 1988, this guidebook; Hooper and Hatcher, 1989). Displacement sense ranges from normal to dextral to sinistral (Garihan and others, 1988, this guidebook; Hooper and Hatcher, 1989). Several prominent lineaments of the Columbus Promontory (Pax Mountain and related faults of Garihan and others, this guidebook) and the region farther west in Georgia (Warwoman and Lake Rabun lineaments of Hatcher, 1974b) formed by weathering and erosion of siliceous cataclasite bodies. In Blue Ridge topography siliceous cataclasites mostly form valleys, whereas in the Piedmont they hold up low ridges.

Hooper (1989) described a set of 030- and 070-trending siliceous cataclasite and quartz mylonite zones in central Georgia that he traced for more than 50 km in a single zone. They are interrelated and yield a dominant sinistral displacement sense. Minor dextral segments are also present, but have a different orientation (~300?). Interestingly, much of the siliceous cataclasite is concentrated in rhomb-shaped pull-apart segments connected by linear fault zones composed dominantly of quartz mylonite, but with minor amounts of siliceous cataclasite.

The segment of the Blue Ridge physiographic front in the Carolinas is the location of the Blue Ridge fault of White (1950). Although White based much of his conclusions on orientations of manganese-coated slickensides in saprolite, he may have inadvertently incorporated some of the Mesozoic fault system into his data and analysis, despite statements to the contrary in his paper. He concluded that the Blue Ridge escarpment is a fault scarp that involved down-to-the- southeast displacement. Detailed geologic mapping along the escarpment in both Carolinas has failed to confirm the existence of such a fault, as has landform analysis which suggests a long complex history of drainage development in the southern Appalachian Blue Ridge and Piedmont (Hack, 1982; Clark, this guidebook). We thus may be seeing only a temporary location for an escarpment that is retreating westward at a fairly rapid rate as Piedmont streams with steeper headwaters gradients behead Blue Ridge streams with gentler gradients (Acker and Hatcher, 1970). Streams are clearly fracture controlled, and, although Paleozoic structure is not expressed in the topography as it is in the Valley and Ridge, once the bedrock geology is mapped in detail, many subtle controls of topography become evident (e.g., Hatcher, 1988b).

Subtleties in control of the present-day topography by fractures and rock units will be evident in the field trip stop at Chimney Rock Park (Stop 6). Here the valley trends nearly EW and is probably related to a localized zone of Mesozoic fractures in the Henderson Gneiss. Increased fracture density also localized streams that drain into the valley. The massive and largely unfractured character of the Henderson Gneiss produced the scenic exfoliation surfaces that rim the Chimney Rock valley where sheeting (release) joints produce large blocks that spall from the existing surfaces, many of which produce talus caves (Bat Cave is also an example of such a feature) (also see Clark, This guidebook for additional discussion). From 5 to 15 m into the forest above the exfoliation surfaces at Chimney Rock valley is the thrust contact of the Henderson Gneiss (below) with the Poor Mountain Formation schist and amphibolite (above). This change in rock type and tectonic unit is also quite evident in a subtle change in slope that occurs along the ridge crest at the thrust contact on the north side of the valley where the slope steepens in the Poor Mountain rocks then flattens into the Henderson Gneiss. Boulder talus and colluvium are quite well exposed on the upper slopes below Chimney Rock. Some of the colluvium here contains deeply weathered clasts of Henderson Gneiss and Poor Mountain Formation lithologies, attesting to the lengthy history of development of the exfoliation surfaces and slope deposits here.

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GEOLOGY OF THE COLUMBUS PROMONTORY, WESTERN PIEDMONT, NORTH CAROLINA, SOUTHERN APPALACHIANS

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ABSTRACT

The geology of the Columbus Promontory is dominated by a crystalline thrust stack composed of three thrust sheets, which in ascending order includes the: 1) Tumblebug Creek thrust sheet, 2) Sugarloaf Mountain thrust sheet, and 3) Mill Spring thrust sheet. Each contains a distinct lithostratigraphy that defines its outcrop pattern. The Tumblebug Creek thrust sheet contains the 509 Ma Henderson Gneiss and a 438 Ma granitoid intruded into the Henderson Gneiss; the Sugarloaf Mountain thrust sheet contains the Sugarloaf gneiss, Poor Mountain Formation, and rocks of the upper Mill Spring complex; the Mill Spring thrust sheet contains the mafic-rich lower Mill Spring complex. Rocks in the Columbus Promontory have a polyphase deformation history (D_1, D_2, D_3). D_2 was the most pervasive and directly related to thrust sheet emplacement, associated internal deformation, and was synchronous with upper amphibolite facies first–sillimanite zone metamorphism (M_1). The kinematics of thrust emplacement was complex and involved contemporaneous W–directed thrusting and SW–directed stretching. Thrust stacking occurred in a foreland-to-hinterland (west to east) progression beginning with the structurally lowest Tumblebug Creek thrust sheet followed by successive emplacement of the structurally higher Sugarloaf Mountain and Mill Spring thrust sheets. A model for ductile deformation in the Columbus Promontory is presented that suggests that D_2 transpression was caused by interaction of W–NW directed thrust sheets butted against the Brevard fault zone. This resulted in partitioned SW- and W- directed flow regime and a "break-back" thrusting sequence. Total displacement in this transpressional regime was accommodated by thrusting and penetrative intracrystalline processes suggesting that Columbus Promontory can be considered a shear zone with the primordial Brevard fault zone as its western boundary.

INTRODUCTION

The focus of the 1993 Carolina Geological Society field trip is a crystalline thrust stack exposed in the Inner Piedmont in western North Carolina in the Columbus Promontory (Figs.1 and 2; Plate I). This crystalline thrust complex comprises three thrust sheets that in ascending order include the Tumblebug Creek thrust sheet, the Sugarloaf Mountain thrust sheet, and the Mill Spring thrust sheet. The purposes of this paper are to present 1) an overview of the lithostratigraphic, structural, metamorphic characteristics of thrust sheets in the Columbus Promontory, and 2) a model for crystalline thrusting applicable to the Columbus Promontory and elsewhere. Hopefully, this paper will provide the necessary background for this year's Carolina Geological Society field trip and for critical evaluation of the interpretations presented in the field.

The Columbus Promontory is located in a high relief (700–1000 m) area of Blue Ridge topography within the Inner Piedmont that produces continuity of outcrop atypical of this terrane, and thus is an outstanding area for study of the complex geologic history of crystalline thrust sheet development. The Columbus Promontory, as defined by Davis (1993), extends from the Brevard fault zone approximately 40 km into the Inner Piedmont belt and sits astride the boundary between the Chauga and Inner Piedmont belts. This area encompasses seven 7 F(1,2) minute quadrangles including the Bat Cave, Cliffield Mountain, Fruitland, Hendersonville, Lake Lure, Mill Spring, and Saluda quadrangles (Fig. 3).

Previous detailed investigations of the geology of the Columbus Promontory include those of Lemmon (1973), Lemmon and Dunn (1973a, 1973b), and the unpublished work of Conley and Drummond (1975). Lemmon (1973) and Lemmon and Dunn (1973a) made the first major contributions to resolution of the internal lithostratigraphy of the Columbus Promontory as part of their study of the Henderson Gneiss and adjacent Inner Piedmont rocks in the Bat Cave quadrangle Fig. 2; Plate

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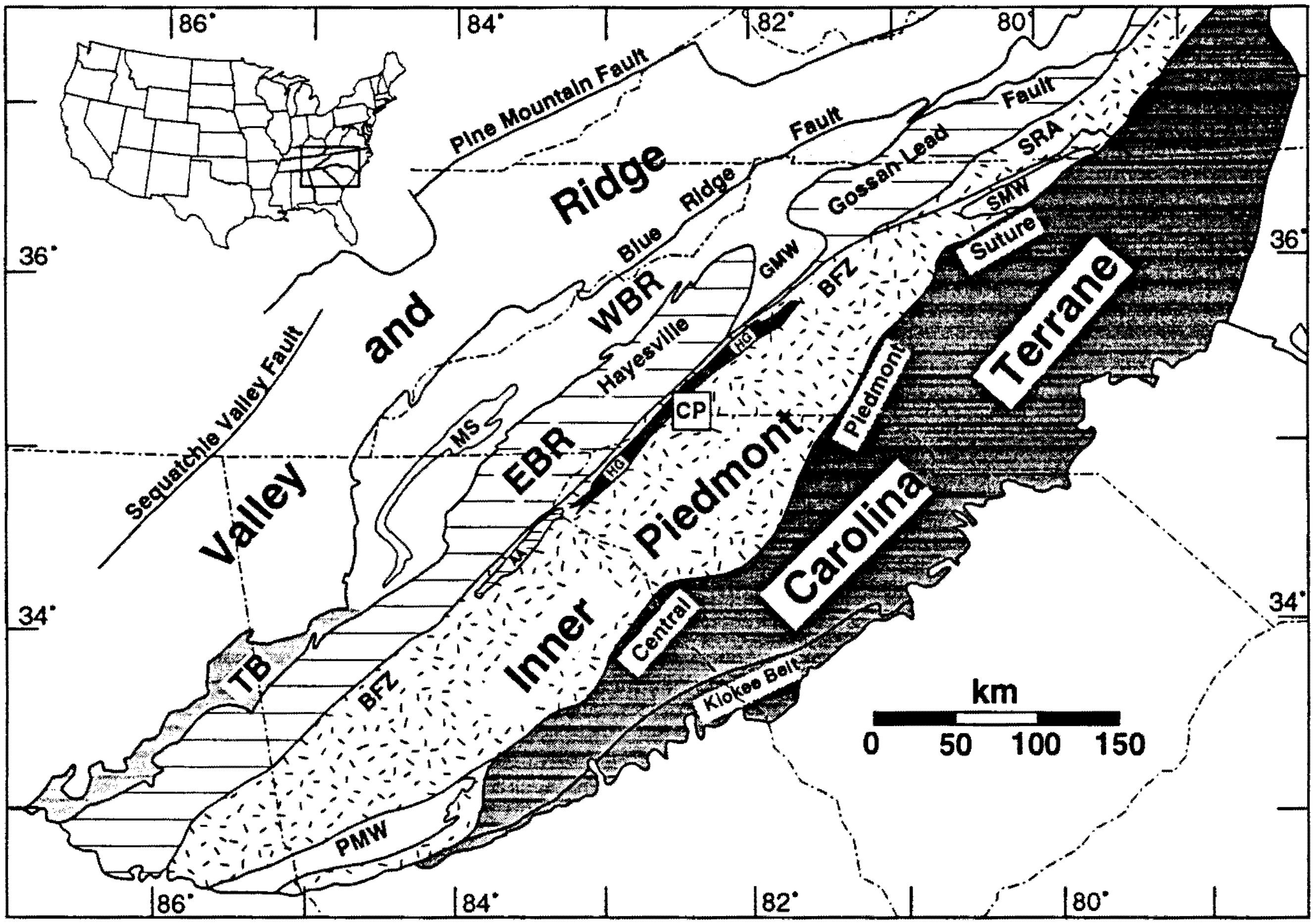


Figure 1. Tectonic subdivisions of the southern Appalachians showing the locations of the Columbus Promontory (CP).

l). Lemmon (1973) and Lemmon and Dunn (1973a) defined two rock units within the Henderson Gneiss outcrop belt and subdivided the overlying Inner Piedmont rocks into three mappable units as part of a sequence termed the Sugarloaf Mountain group contained in an unnamed thrust sheet. Lemmon (1973) and Lemmon and Dunn (1973a) also recognized a thick biotite gneiss unit beneath the eastern extent of the Sugarloaf Mountain group, but were unable to definitively determine the contact relationships between these two units.

Conley and Drummond (unpublished, 1975) working south of Lemmon's area in Polk County, North Carolina, described a lithostratigraphic unit informally called the Tryon formation for the town of Tryon, North Carolina. The Tryon formation included rock units similar to those in Lemmon's Sugarloaf Mountain group as well as the underlying biotite gneiss. Conley and Drummond also defined a sequence of migmatitic granitic gneiss, amphibolite, and amphibole gneiss they informally termed the Mill Spring group for the town of Mill Spring, North Carolina.

The first major contribution towards the understanding of the structural and metamorphic history of the Columbus Promontory was also by Lemmon (1973) and subsequent work of Lemmon and Dunn (1973a, 1973b). Lemmon (1973) and Lemmon and Dunn (1973a) suggested that the Sugarloaf Mountain group underwent isoclinal to recumbent folding (F_1) and almandine amphibolite facies metamorphism (M_1) during the Taconian orogeny. This event was followed by intrusion of coarse-grained Ordovician (438 Ma, Odom and Russell, 1975) granitoid gneiss into the Henderson Gneiss (535 Ma, Odom and Fullagar, 1973). Subsequent regional metamorphism (M_2) and isoclinal folding (F_2, nearly coaxial and coplanar to F_1) affected these rocks during the late Taconian or Acadian(?) orogeny. Lemmon also recognized emplacement of the Sugarloaf Mountain group over the Henderson Gneiss occurred during M_2.

These earlier studies provided the critical framework for the recent work in the Columbus Promontory by Davis (1993), from which much of this paper is extracted. Although many of the original interpretations

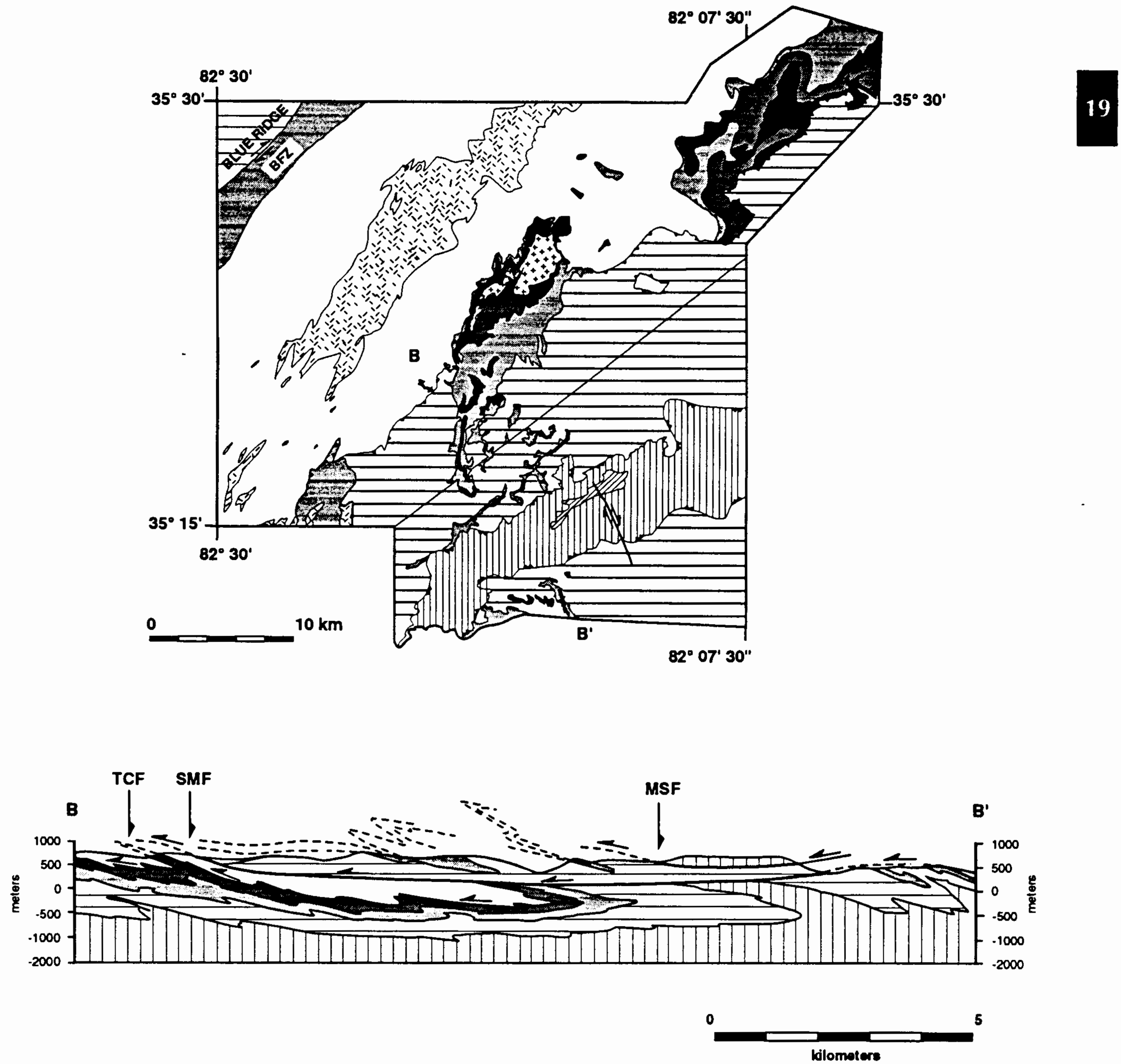


Figure 2. Geologic map (a) and cross section (b) of the Columbus Promontory. Unpatterned–Henderson Gneiss. cross hatch–438 Ma granitic gneiss. crosses– Sugarloaf gneiss. black–Poor Mountain Formation quartzite. dark gray–Poor Mountain amphibolite. light gray–Poor Mountain pelitic schist. horizontal lines–Mill Spring complex migmatitic biotite gneiss-metagraywacke. vertical lines–Mill Spring complex migmatitic metagraywacke-amphibole-amphibole gneiss. BFZ–Brevard fault zone; TCT–Tumblebug Creek thrust; SMT–Sugarloaf Mountain thrust; MST–Mill Spring thrust. Teethed lines are thrust faults (teeth on hanging wall). Map compiled from Lemmon (1973), Lemmon and Dunn (1973a, 1973b), unpublished data of Lemmon (1975), unpublished data of Davis (1987-1990), unpublished data of Tabor (1988-1989), and unpublished data of Yanagihara (1990-1992).

of the previous studies are essentially unchanged by this new work, several new observations have been made that not only refine the geologic understanding of this area, but also resulted in new ideas about crystalline

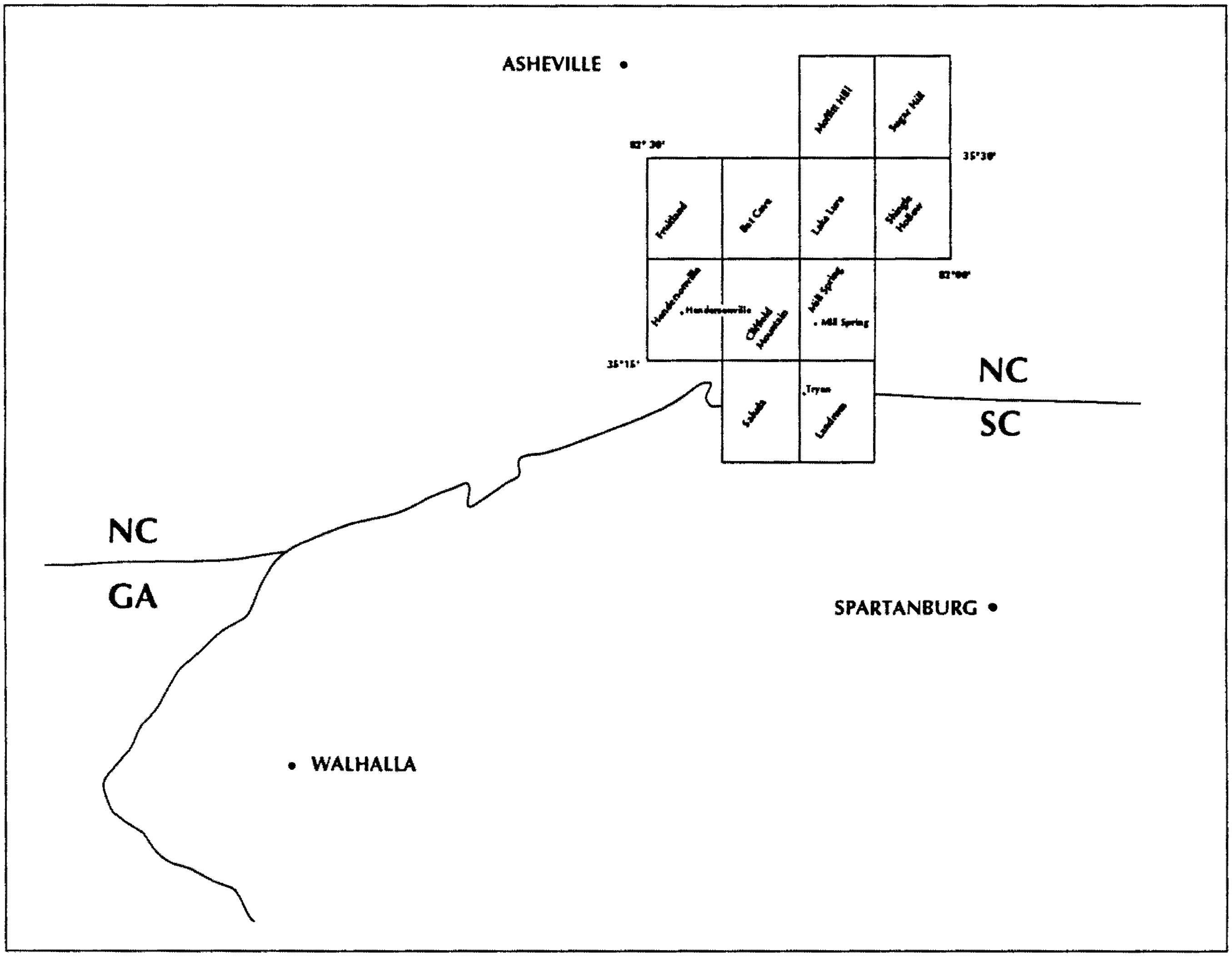


Figure 3. Index to topographic quadrangles included as part of the Columbus Promontory.

thrusting in the Inner Piedmont.

REGIONAL TECTONIC SETTING

The Inner Piedmont represents the part of the highly deformed and metamorphosed core of the southern Appalachian orogen (Fig.1). This extensive terrane includes the Inner Piedmont belt of King (1955) and Chauga belt of Hatcher (1972), and extends from the Sauratown Mountains and Smith River allochthon in North Carolina and Virginia to the Coastal Plain onlap in Alabama. The Inner Piedmont is bounded on the NW by the Brevard fault zone and on the SE by faults and shear zones that define the central Piedmont suture of Hatcher and Zeitz (1980).

The Inner Piedmont has long been considered one of the most enigmatic geologic features in the southern Appalachians. Over the course of the last twenty years, however, numerous geologic investigations have established that the Inner Piedmont is a composite stack of NW–directed crystalline thrust sheets. Significant studies supporting this interpretation include Griffin (1971, 1974), Hatcher (1972, 1978a, 1987), Lemmon (1973), Conley and Henika (1973), Hatcher and Odom (1980), Heyn (1984), Nelson and others (1987), Hopson and Hatcher (1988), Goldsmith and others (1988), McConnell (1988), and Steltenpohl and others (1990).

The lithostratigraphy of the Inner Piedmont is one of the most poorly understood aspects of the geology of the crystalline southern Appalachians. In general, Inner Piedmont lithostratigraphy consists of an assemblage of medium- to high-grade ortho– and paragneisses intruded by pre–, syn–, and postkinematic plutons. The protoliths of the Inner Piedmont rocks consisted predominantly of immature quartzofeldspathic and pelitic sediments, and mafic lavas. The internal stratigraphy of the Inner Piedmont has been resolved in detail in only a few areas including the Smith River allochthon (Conley and Henika, 1973), the Alto allochthon (Hatcher, 1978b; Hopson and Hatcher, 1988), west of the Sauratown Mountains (Heyn, 1984, 1988; Hatcher, 1988; McConnell, 1988), the Chauga belt of South Carolina (Hatcher, 1969, 1970; Hatcher and Acker, 1984), and adjacent to the Brevard fault zone in Alabama (Bentley and Neathery, 1970; Steltenpohl and others, 1990; Neilson and others, 1990a, 1990b).

The Inner Piedmont represents part of the crystalline metamorphic core of the southern Appalachians (Fig.4). In the Carolinas, the early to middle Paleozoic metamorphic pattern of the western Inner Piedmont, as recorded by pelitic units, is characterized by an extensive high-grade (sillimanite–muscovite) core, bounded on the NW by lower-grade (garnet to kyanite) rocks of the Chauga belt (Fig. 4). This pattern of decreasing metamorphic grade, from the deeper Inner Piedmont westward towards the Chauga belt, is considered to be the result of retrograde emplacement of higher-grade thrust sheets (e.g., Six Mile thrust sheet and Alto allochthons) and folds nappes (e.g. Walhalla and Anderson nappes) of the Inner Piedmont belt, over the lower-grade rocks of the Chauga belt (Griffin, 1969, 1971a, 1971b, 1974a, 1974b; Hatcher, 1972, 1978a, 1987, 1989; Lemmon, 1973, 1982; Lemmon and Dunn, 1975; Roper and Dunn, 1973; Nelson and others, 1987; Hopson and Hatcher, 1988; Htm and McConnell, 1990). Towards the central part of the Inner Piedmont pelitic rocks consistently record first-sillimanite zone conditions (Fig. 4). Butler (1990) noted that isogradic surfaces in the central Inner Piedmont of the Carolinas must be nearly horizontal, because over an area of more than 60 km the metamorphic grade is constantly first-sillimanite zone (sillimanite–muscovite) and does not increase to second sillimanite zone (sillimanite + K–feldspar) or drop to kyanite grade.

Inner Piedmont thrust sheets are bounded to the northwest by the Brevard fault zone, traceable 600 km from Alabama to Virginia (Fig. 1). The Brevard fault zone separates the eastern Blue Ridge from the Inner Piedmont (Reed and Bryant, 1964), and has long been considered as a major crustal break (Jonas, 1932; King 1955; Reed and Bryant, 1964; Butler, 1971; Bryant and Reed; 1970; Bird and Dewey, 1970; Odom and Fullagar, 1973; Horton 1974, 1982; Horton and Butler, 1986; Hatcher, 1971b, 1972, 1978b, 1987, 1989). At least three major deformational phases have been recognized in the Brevard fault zone: early thrusting (Jonas, 1932; Bentley and Neathery, 1970; Hatcher 1971b, 1972; Stirewalt and Dunn, 1973; Roper and Justus, 1973); dextral strike slip (Reed and Bryant, 1964; Bobyarchik, 1984; Edelman and others, 1987; Evans and Mosher, 1986; Bobyarchik and others, 1988); and brittle thrusting during the Rosman phase (Horton, 1974,1982; Horton and Butler, 1986; Edelman and others, 1987; Hatcher and others, 1989). A listric thrust geometry has been corroborated by seismic reflection profiles (Clark and others, 1978; Cook and others, 1979; Harris and Bayer, 1979; Harris and others, 1981; Çoruh and others, 1987; Costain and others, 1989).

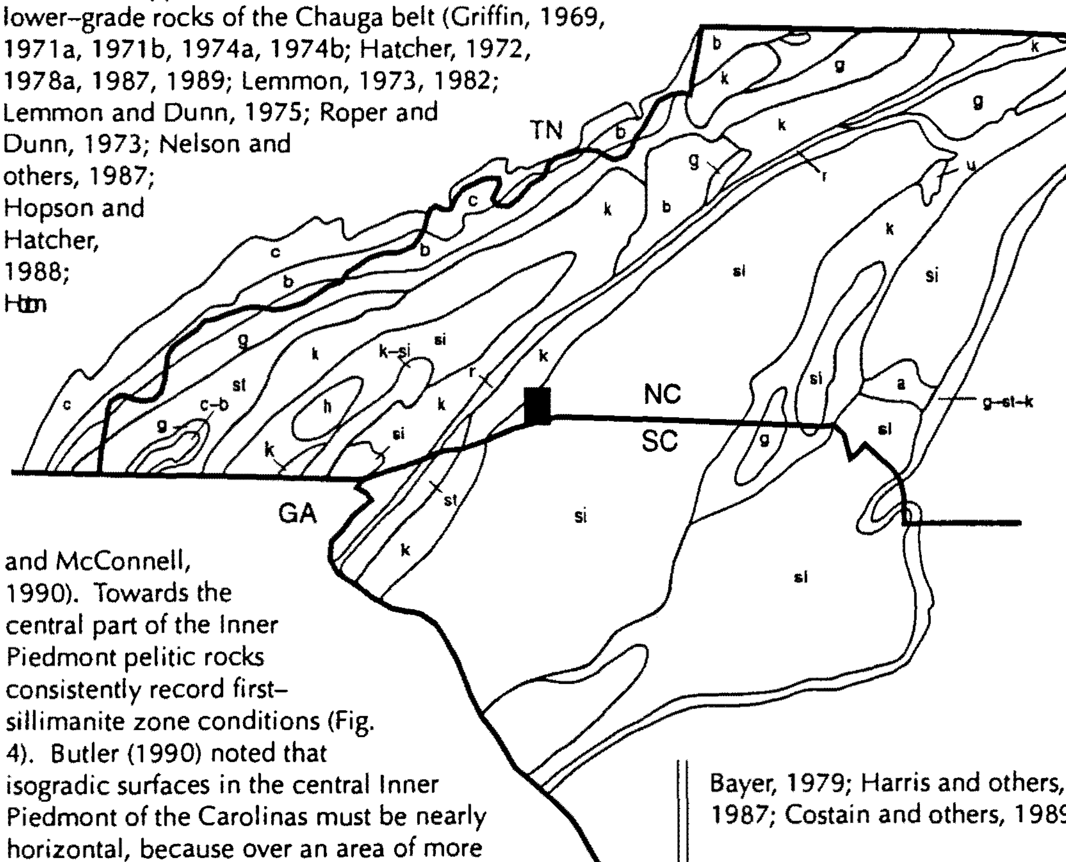


Figure 4. Distribution of Paleozoic metamorphic isograds in the southern Appalachian orogen. Study area is shown by black box. k–kyanite; si–sillimanite; a–andalusite; st–staurolite; g–garnet; b–biotite; c–chlorite; u–unmetamorphosed; h–hypersthene; r–retrograde. Modified after Butler (1990).

LITHOSTRATIGRAPHIC FRAMEWORK

Davis (1993) redefined the lithostratigraphy of the Columbus Promontory to include four major lithostratigraphic units including the Henderson

Gneiss, Sugarloaf gneiss, Poor Mountain Formation, and the Mill Spring complex. The rationale for this redefinition was based on new field, petrographic, and geochemical data presented in Davis (1993). In part, this new framework is consistent with many original ideas postulated by Lemmon (1973).

Lemmon (1973) and Davis (1993) recognized that rocks of the Sugarloaf Mountain group are nearly identical to the Poor Mountain Formation, defined to the southwest in South Carolina by Sloan (1907), Shufflebarger (1961), and Hatcher (1969 and 1970). Davis (1993) suggested assignment of these rocks to the Poor Mountain Formation. Lemmon (1973) also included a granitoid gneiss at the top of Sugarloaf Mountain in the Sugarloaf Mountain group, but also postulated that it could be igneous origin. Based on additional petrographic work, Davis (1993) designated this granitoid gneiss as a distinctive unit within the lithostratigraphy of the Columbus Promontory called the Sugarloaf gneiss.

Davis (1993) also included the thick biotite gneiss recognized by Lemmon (1973) and Conley and Drummond beneath the Sugarloaf Mountain group, and the migmatitic gneiss and interlayered amphibolite, and amphibole gneiss of Conley and Drummond into a lithostratigraphic assemblage termed the Mill Spring complex. The Mill Spring complex represents the areally most extensive rock unit in the Columbus Promontory and can be divided into upper and lower subunits based on the relative abundance of amphibolite. The upper Mill Spring complex is relatively amphibolite–poor whereas the lower Mill Spring complex contains abundant amphibolite.

These lithostratigraphic units, defined by Davis (1993) and discussed here, comprise the three thrust sheets recognized in the Columbus Promontory (Fig. 2; Plate I) and their description below is organized according to the thrust sheets (structurally lowest to highest) in which they occur. The Tumblebug Creek thrust sheet contains the 509 Ma Henderson Gneiss and a 438 Ma granitoid intruded into the Henderson Gneiss; the Sugarloaf Mountain thrust sheet contains the Sugarloaf gneiss, Poor Mountain Formation, and upper Mill Spring complex; the Mill Spring thrust sheet contains the mafic–rich lower Mill Spring complex. The lithostratigraphic and structural relationships between these units are shown in Figure 5 and Plate I.

Tumblebug Creek Thrust Sheet

Henderson Gneiss. Keith (1905, 1907) defined and delineated the Henderson Granite, with the type section located in Henderson County, North Carolina. Reed and Bryant (1964) redefined the Henderson Gneiss and restricted the outcrop unit to southeast of the Brevard fault zone. As defined by Reed and Bryant (1964), the Henderson Gneiss extends in the Piedmont from the South Carolina-Georgia border northeastward to the southeastern flank of the Grandfather Mountain window (Fig.1).

The most extensive exposures of the Henderson Gneiss within the Columbus Promontory occur in the Fruitland, Bat Cave, Hendersonville, Lake Lure and the NW corner of the Cliffield Mountain quadrangles (Plate I). The Henderson Gneiss was also mapped in the Mill Spring quadrangle through either a window or reentrant in the Sugarloaf Mountain thrust sheet. The composition of the Henderson Gneiss varies from that of granite to quartz monzonite and is composed of microcline, oligoclase, quartz, and biotite with accessory muscovite, garnet, allanite, zircon, sphene, and opaque minerals (Lemmon 1973). One of the most distinctive characteristics of the Henderson Gneiss is the presence of distinct K–feldspar augen up to 3 cm long. The augen are white microcline, ovoid to asymmetric in cross section and elongate within the foliation plane. The augen are commonly rimmed by quartz, plagioclase, and embayments of myrmekite. The Henderson Gneiss, in the Columbus Promontory and throughout much of the outcrop length, contains a pronounced NE–SW–trending mineral lineation defined by quartz ribbons, elongate K-feldspar porphyroclasts, flakes of biotite, and occasional muscovite, and in places is an L–tectonite.

The Henderson Gneiss has been the focus of

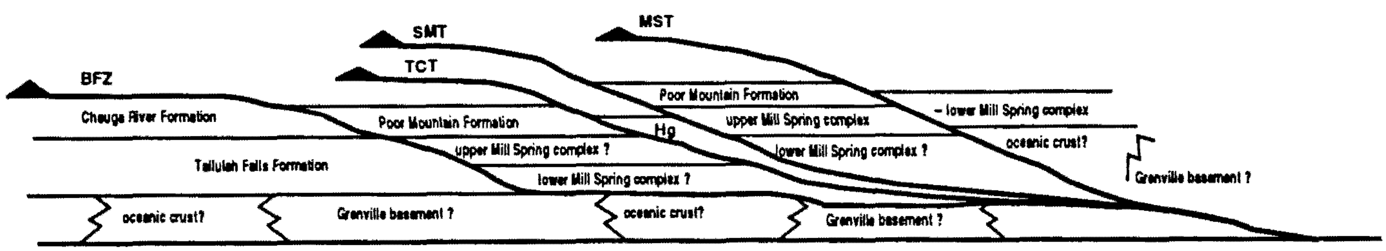


Figure 5. Conceptual stratigraphic and structural cross section for rock units in the Columbus Promontory (Inner Piedmont) and the adjacent eastern Blue Ridge. MST–Mill Spring thrust; SMT–Sugarloaf Mountain thrust; TCT–Tumblebug Creek thrust; BFZ–Brevard fault zone; Hg–Henderson Gneiss.

many geochronological studies, but, unfortunately, many discrepancies exist in the results. Odom and Fullagar (1973) reported a Rb–Sr whole–rock age of 535 Ma for the Henderson Gneiss with an initial $^{87}Sr / ^{86}Sr$ ratio of 0.70309 and a U–Pb age of 538 Ma. They interpreted these to represent crystallization ages and the low initial $^{87}Sr / ^{86}Sr$ ratio indicative of an igneous lower crustal origin. Odom and Fullagar (1973) also reported an age of 356 Ma for mylonitic Henderson Gneiss in the Brevard fault zone. Sinha and Glover (1978) reported an U–Pb age of about 600 Ma based on a different common lead constant. Sinha and others (1989) reported a Rb–Sr age of 509 Ma (recalculated from Odom and Fullagar, 1973), and Sinha and others (1988) obtained a Rb–Sr age of 273 Ma from Henderson Gneiss ultramylonite in the Brevard fault zone. They attributed this young age to reopening of the system during extensive fluid flow and mylonitization related to movement on the Brevard fault zone during the late Paleozoic Alleghanian orogeny.

438 Ma granitoid. Detailed mapping by Lemmon (1973) and Lemmon and Dunn (1973a and 1973b) in the Bat Cave and Fruitland quadrangles indicate that the Henderson Gneiss was intruded by a younger granite prior to the development of the dominant foliation (Fig. 2; Plate I). The contact between the Henderson Gneiss and the granitic gneiss is now concordant and parallel to the regional southeast-dipping foliation. This unit is distinguished from the Henderson Gneiss by the lack of K–feldspar augen, less biotite, and generally lighter color. The composition of this granitoid gneiss varies from granite to granodiorite. Odom and Russell (1975) reported a Rb–Sr whole–rock age of 438 Ma, with an initial $^{87}Sr / ^{86}Sr$ of 0.7045 for the granitoid gneiss.

Sugarloaf Mountain Thrust Sheet

Sugarloaf Gneiss. The Sugarloaf gneiss is confined to the top of Sugarloaf Mountain in the Bat Cave quadrangle (Fig 2; Plate I). It represents the structurally highest unit in the Columbus Promontory contained in the Sugarloaf Mountain thrust sheet. This unit was originally included with the Sugarloaf Mountain group by Lemmon (1973). Based on zircon morphology and scatter of Rb–Sr isotopic data, Lemmon suggested that the Sugarloaf gneiss could be a recycled sedimentary rock, but did not rule out the possibility of an igneous protolith for this unit. New isotopic work in progress (Goldberg and Fullagar, this guidebook) should help elucidate the protolith history and age (metamorphic of crystallization?) of this unit. As noted previously, because of the possibility of an igneous origin for this unit, Davis (1993) redefined it as a separate lithologic unit called the Sugarloaf gneiss. A nearly identical group of unnamed gneiss bodies present in the Chauga belt of South Carolina have a Rb–Sr age of 423 Ma (Harper and Fullagar, 1981). Both rock units maintain a similar structural (stratigraphic?) position above the Poor Mountain Formation.

The modal composition of the Sugarloaf gneiss is granitic to granodioritic in composition and contains predominantly microcline, plagioclase, biotite and muscovite (Lemmon, 1973). Accessory minerals include zircon, sphene, apatite, allanite, chlorite, epidote, and garnet. The rock is light gray to white, massive to well foliated, and contains a pronounced NE-SW mineral lineation defined by oriented micas, and quartz ribbons and elongate feldspar. The contact relationship with the underlying amphibolite is concordant and parallel to the southeast–dipping foliation. Locally, the contact is folded.

Poor Mountain Formation. The Poor Mountain Formation in the Columbus Promontory consists of three mappable units, equivalent to the Poor Mountain amphibolite, Brevard–Poor Mountain transitional member, and the quartzite member recognized in South Carolina by Shufflebarger (1961) and Hatcher (1969, 1970). In descending order, these include: (1) laminated amphibolite and hornblende gneiss; (2) garnet–mica schist and quartzite; and (3) interlayered amphibolite and quartzite. Lemmon (1973) also noted discontinuous lenses of marble interlayered within the mica schist in the Bat Cave quadrangle.

The Poor Mountain Formation crops out structurally below and above the Henderson Gneiss in the study area (Fig. 2; Plate I). Structurally below the Henderson Gneiss, only Poor Mountain amphibolite has been observed in a window through the Henderson Gneiss along the Tumblebug Creek thrust (Fig. 2; Plate I). Structurally above the Henderson Gneiss, all units of the Poor Mountain Formation have been observed in the Sugarloaf Mountain thrust sheet. Stratigraphic and petrologic characteristics of the Poor Mountain Formation of the Columbus Promontory contained within the Sugarloaf Mountain thrust sheet are described below.

Amphibolite–hornblende gneiss. The amphibolite unit crops out beneath the Sugarloaf Gneiss and is present in the Cliffield Mountain, Saluda, Mill

Spring, and Lake Lure quadrangles (Fig. 2; Plate I). The map unit is fine to medium grained, dark gray to black, and is commonly laminated with well-defined quartzofeldspathic layers. Where folded, this laminated amphibolite produces some of the most spectacular mesoscopic folds in the Columbus Promontory. Mineralogically, the unit contains 22–70 percent dark-green pleochroic hornblende, 7–61 percent plagioclase (An $_{25-37}$), 0–22 percent quartz, occasional diopside, and small flakes of pleochroic biotite (Davis, 1993). Other accessory minerals include garnet, sphene, tremolite, zircon, apatite, and opaque minerals. In handspecimen, the amphiboles have a readily discernible nematoblastic shape and define a weak to strong linear fabric in the rock.

Garnet-mica schist and quartzite. The garnet–mica schist and quartzite unit generally crops out below the amphibolite–hornblende gneiss unit, although in some areas these units are interlayered by folding. In the northwestern part of the Columbus Promontory, in the Bat Cave, Cliffield Mountain, and Lake Lure quadrangles, it rests directly on the Henderson Gneiss along the Sugarloaf Mountain thrust, forming one of the sharpest fault contacts in the entire study area (Fig. 2; Plate I). In many cases, this unit occurs on the topographically highest areas capping the ridge tops. This rock is purplish–red, to brown, to light gray, with the color related to the amount of sillimanite, biotite, or muscovite in the rock and the degree of weathering. The schist consists of folia of strongly aligned grains of biotite (21–40 percent), muscovite (2–38 percent), and fibrolitic sillimanite (0–13 percent) alternating with ribbons or layers of recrystallized quartz (22–50 percent), and minor amounts of K-feldspar (Davis, 1993). Accessory minerals include zircon, apatite, magnetite, ilmenite, and graphite. Garnets have several morphologies: some are elongated parallel to the dominant foliation; others (up to 5mm) are anhedral to subhedral with inclusion–rich cores and clear rims, while others (1-2mm) have sub- to euhedral outlines and lack inclusions. Commonly, a second foliation can be observed that is defined by asymmetric mica grains and sillimanite bundles, and asymmetric quartz–feldspar pods.

Other minor components of this map unit include quartzite and marble. Mappable quartzite layers also occur folded in with the garnet–mica schist. The stratigraphic position of the quartzite within varies. The quartzite varies from light yellow to brown. Mineralogically it contains predominantly quartz and accessory amount of muscovite, and garnet. Lemmon (1973) also reported pods and discontinuous lenses of marble within the garnet-mica schist unit in the Columbus Promontory. A single chemical analysis of this marble (Lemmon, 1973) reveals the rock is a high–calcium marble.

Quartzite–amphibolite. The stratigraphically (?) lowest lithology of the Poor Mountain Formation in Sugarloaf Mountain thrust sheet is a discontinuous sequence of interlayered impure quartzite and amphibolite Fig. 2; Plate I). The quartzite varies from light yellow or white to dark brown or black. Mineralogically it contains mostly quartz, although in some cases it does contain muscovite, sillimanite, amphibole, garnet, and other accessory minerals.

The amphibolite unit is mineralogically and texturally identical to the main body of Poor Mountain amphibolite described above. This quartzite–amphibolite unit is discontinuous and is commonly found at the contact between the Poor Mountain Formation and the underlying rocks of the upper Mill Spring complex. At some localities both rock types are present, while at others only one of the rock units is visible. The contact between the Poor Mountain Formation and the underlying Mill Spring complex is interpreted to be primarily stratigraphic, although this is not entirely clear. The best exposures of this contact occurs on Long Ridge in the Cliffield Mountain quadrangle and at Melrose Mountain in the Saluda quadrangle (Plate I). At these localities, however, it is difficult to determine if it is a stratigraphic or fault contact. On Long Ridge the contact is sharp and the lowermost quartzite–amphibolite unit of the Poor Mountain Formation has mylonitic characteristics and the amphibolite is occasionally intensely folded. On Melrose Mountain there is a stratigraphic interleaving of the biotite gneiss of the upper Mill Spring complex with the overlying Poor Formation, although the contact could represent either a transposed stratigraphic or early(?) premetamorphic fault contact.

Upper Mill Spring Complex. The upper Mill Spring complex (Davis, 1993) is the stratigraphically lowest unit the Sugarloaf Mountain thrust sheet (Fig 2; Plate I). The upper Mill Spring complex is dominantly a thick sequence of migmatitic biotite gneiss and metagraywacke, and is distinguished from the lower Mill Spring complex by the lower relative abundance of mafic rocks. It commonly produces massive exposures and forms the cliffs and balds throughout the study area. The mineralogy of the biotite gneiss–

metagraywacke is quite variable, but on average contains 12-60 percent plagioclase (An $_{20-35}$), 12–56 percent quartz, 0-25 percent muscovite, 2-19 percent biotite, and minor amounts of sillimanite, garnet, and sphene (Davis, 1993). Accessory amounts of epidote, zircon, and opaque minerals are also present. The biotite gneiss–metagraywacke is light to medium gray, equigranular, fine- to medium–grained, and massive to slightly banded. Locally the grain size of the mica is quite large and the unit resembles mica schist. Foliation is produced by oriented micas and elongated quartz–feldspar layers. In many areas, this unit appears very migmatitic with marked segregations of the felsic, more micaceous, and mafic–rich layers.

The biotite gneiss–metagraywacke contains pods and lenses of amphibolite parallel to the regional foliation. These lenses commonly have a sill-like geometry and in most cases are parallel to the regional S_2 foliation. This unit also contains pods and lenses of pegmatite parallel to the dominant foliation. In the western part of the study area, along the Green River, the biotite gneiss of the upper Mill Spring Formation is a porphyroclastic biotite gneiss. This rocks generally has a matrix identical to the biotite gneiss, but contains gray to pink, carlsbad–twinned microcline porphyroblasts.

Mill Spring Thrust Sheet

The Mill Spring thrust sheet (Fig. 2; Plate I) contains the mafic–rich rocks of the lower Mill Spring complex of Davis (1993). The lower Mill Spring complex consists of a migmatitic sequence of biotite–granitic gneiss–metagraywacke, coarse amphibolite gneiss, and fine- to medium–grained amphibolite. The complex interlayering of these rocks types makes it very to difficult to subdivide the individual units.

Like the biotite gneiss–metagraywacke of the upper Mill Spring complex, the mineral composition of the biotite gneiss–metagraywacke in the lower Mill Spring complex is also quite variable. On average it consists of 10–45 percent plagioclase, 28–35 percent quartz, 0-29 percent K–feldspar, 11–27 percent biotite, and 6–26 percent muscovite. Zircon, apatite, sphene and opaque minerals are generally present in accessory amounts (Davis, 1993). Epidote occurs in veins and as fillings in late brittle fractures. This unit is generally a mesocratic, light–gray, segregation banded, inequigranular, biotite gneiss–metagraywacke. It is permeated by migmatite and concordant pegmatite layers.

Amphibolite and amphibole gneiss in the lower Mill Spring complex occur as both large pods and tabular or sill–like stringers. The large pods are commonly permeated with leucogranite or pegmatite layers, as in the biotite gneiss–metagraywacke units. The foliation in the amphibolite and amphibole gneiss is defined by alternating mafic and felsic layers. The mineralogic makeup of the amphibolite in the lower Mill Spring complex is also variable, but on average includes 35–60 percent dark green pleochroic hornblende, 25–50 percent plagioclase (An $_{20-30}$), 2–20 percent quartz, 0–30 percent biotite, and minor amounts of epidote, garnet, and opaque minerals (Davis, 1993). Zircon, sphene, and chlorite occur as accessory minerals. Amphibolite of the lower Mill Spring complex is generally more massive and coarser grained than the amphibolite of the Poor Mountain Formation, although this is not always the case. Individual amphibole minerals also have a nematoblastic shape and define a weak to strong linear fabric.

On the top of White Oak Mountain, the migmatitic granitic gneiss and amphibolite of the lower Mill Spring complex grades into a porphyroclastic biotite gneiss similar to that in the upper Mill Spring complex with porphyroclasts composed of white microcline. This unit commonly contains abundant mica and in many cases is very schistose. Thin amphibolite stringers are rare, but do occur within this lithology. Towards the western boundary of the lower Mill Spring Formation in the Saluda and Cliffield Mountain quadrangles, migmatitic amphibolite and granitic gneiss grade into a more amphibolite-poor biotite-gneiss similar to that in the upper Mill Spring complex. Here the lower Mill Spring complex can be seen overlying the garnet-mica schist of the Poor Mountain Formation and the biotite gneiss—metagraywacke of the upper Mill Spring complex along the Mill Spring thrust.

METAMORPHISM

Lemmon (1973) postulated on the possibility of two prograde metamorphic events (M_1 and M_2) in the Columbus Promontory area based on the presence of two episodes of garnet growth in the Sugarloaf Mountain group rocks. Lemmon recognized M_1 only in the Sugarloaf Mountain group rocks (Poor Mountain Formation). M_2 affected all rocks in the area. Lemmon also recognized that the M_2 event produced the pervasive S_2 foliation that occurs throughout the Columbus Promontory. Based on garnet composition and zoning patterns, reaction relationships, and microtextures, Davis (1993) suggested that the two stage garnet growth was more easily explained by a single (M_1) upper amphibolite facies sillimanite-grade metamorphism. M_1 affected all rock units in the area and accompanied development of the S_2 foliation and is equivalent to M_2 of Lemmon (1973). The discussion here relates to the M_1 event of Davis (1993).

Mineral Assemblage and Reaction Relationships

Metamorphism accompanying crystalline thrusting in the Columbus Promontory was determined from pelitic rocks in the Sugarloaf Mountain thrust sheet. As previously described, pelitic units represent a significant part of the stratigraphy of the Sugarloaf Mountain thrust sheet and occur within the Poor Mountain Formation and the upper part of the Mill Spring complex. Pelitic schist of the Poor Mountain Formation defines a distinct mappable stratigraphic unit, whereas pelitic rocks of the upper Mill Spring complex occur throughout the thick sequence of biotite gneiss–metagraywacke that crops out over a significant part of the study area.

Pelitic rocks in both the Poor Mountain Formation and the upper Mill Spring complex, invariably contain a metamorphic assemblage consisting of the mineral phases

$$\text{gar} + \text{bio} + \text{mus} + \text{si} + \text{plag}\ (An_{13-38}) + \text{qtz} \pm \text{il} \qquad (A_1)$$

This assemblage is characteristic of the first–sillimanite zone of regional metamorphism (Winkler, 1978; Yardley, 1989). This mineral assemblage defines the dominant planar (S_2) and linear (L_2) elements in these pelites.

The metamorphic reaction history recorded in these pelites involved a continuous reaction between phases in A_1. Evidence for discontinuous reactions was not observed in any of the pelite samples examined. Textural evidence indicates that sillimanite growth in A_1 occurred as a result of continuous reactions that involved consumption of garnet and muscovite. Pelitic schists containing garnets embayed by intergrown fibrolitic sillimanite, biotite and quartz, and replacement of muscovite by fibrolitic sillimanite, biotite, and quartz provide strong textural evidence that the general garnet consuming reaction

$$\text{gar} + \text{mus} = 2\ \text{si} + \text{bio} + \text{qtz} \qquad (1)$$

was operative during the metamorphic development of these pelites (Fig. 6a).

These textural features (e.g., garnet embayed by sillimanite) described above are consistent with garnet having reacted with the decreasing temperature from the peak metamorphic conditions. The metamorphic significance of reaction 1 was recently discussed by Mohan and others (1989) and Spear and others (1990). In these studies, the garnet–consuming and sillimanite–producing reaction (reaction 1) was interpreted to be a continuous reaction during retrograde metamorphic conditions. Tracy (1982) discussed a similar garnet–consuming reaction from pelitic rocks of New England, but with K–feldspar and H_2O in place of muscovite. Similarly, Tracy (1982) indicated that this reaction also occurs during the initial stages of cooling following the metamorphic peak. These relationships suggest that assemblage A_1 in pelites of the Sugarloaf Mountain thrust sheet also represents a retrograde metamorphic mineral assemblage developed during initial cooling from the metamorphic peak.

Metamorphic Conditions

The metamorphic conditions attained by pelites of the Sugarloaf Mountain thrust sheet are constrained by several field and petrographic observations. In all hand samples and thin sections examined, sillimanite is the only aluminum silicate polymorph observed in the Poor Mountain Formation and upper Mill Spring complex pelitic rocks. Kyanite, recognized in many thrust sheets elsewhere in the western Piedmont, was not observed in any of the thin sections examined or during field mapping, thus qualitatively restricting the P–T conditions to occur below the kyanite–sillimanite boundary. Although a few samples examined contained minor amounts of K–feldspar, reaction textures indicative of second–sillimanite zone metamorphic conditions, generally characterized by the reaction

$$\text{mus} + \text{qtz} = \text{Kfs} + \text{si} + H_2O \qquad (2)$$

of Chatterjee and Johannes (1974) were not observed. This suggests that metamorphic conditions occurred on the lower P–T side of this reaction. In many areas within the Sugarloaf Mountain thrust sheet, rocks of the upper Mill Spring complex are migmatitic suggesting metamorphic conditions were, at least locally, higher than the minimum granite melting curve. Migmatitic textures are much more commonly observed in the interlayered biotite gneiss–metagraywacke and schist of the upper Mill Spring complex than in the Poor Mountain Formation. This suggests that, at least locally, the minimum melting reaction for pelitic rocks (Thompson and Algor, 1977)

$$\text{mus} + \text{Kfs} + \text{ab} + \text{qtz} + H_2O = \text{liquid} \qquad (3)$$

may have been operative. These observations qualitatively suggest metamorphic conditions for these pelites are within the sillimanite stability field at P–T

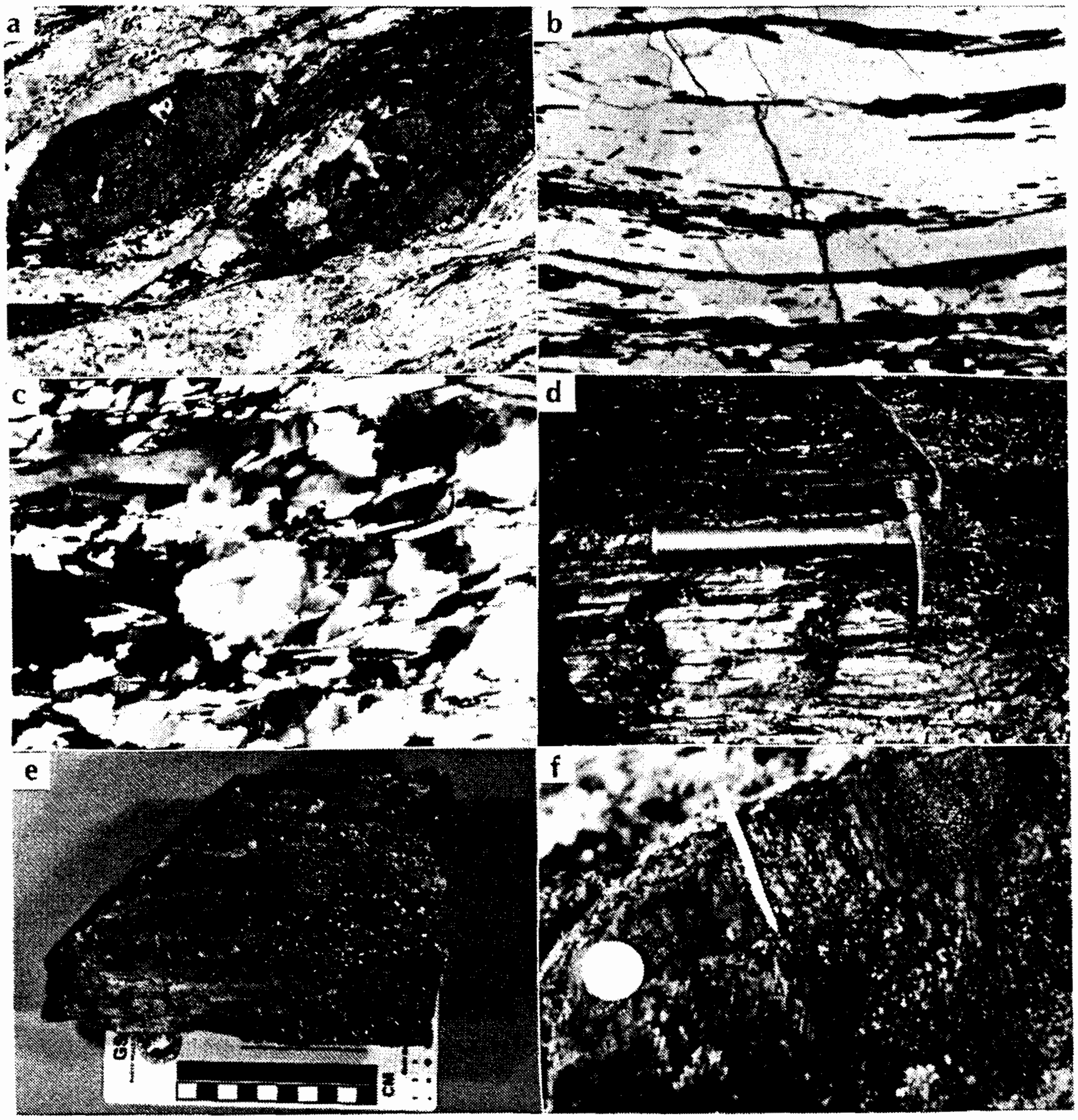


Figure 6. a) Sheared garnets within S_2, showing top-to-SW (right) shear sense. Note sillimanite growth paralleling garnet boundaries, suggesting synkinematic sillimanite growth during shearing. Field of view is 4 mm. Plane light. b) Quartz ribbons defining S_2. Also note sillimanite parallel to quartz ribbon. Field of view is 4 mm. Plane light. c) Type II S–C mylonite (Lister and Snoke, 1984) defined by highly strained quartz microfabric and asymmetric white-mica fish. Shear sense is top-to-W (right). Field of view if 4 mm. Plane light. d) NE–SW oriented mineral lineation in interleaved Poor Mountain amphibolite and Henderson Gneiss. Hammer is 40 cm. e) E–W oriented lineation from metagraywacke of Mill Spring complex in the MST. f) Sample containing two mineral stretching lineations. Toothpick oriented 18°/243°, the other orientation is 22°/272°.

conditions between the that of reactions 2 and 3 (Fig. 7).

Geothermobarometry

Additional constraints on the metamorphic conditions are provided by the results of geothermobarometric estimations. Temperature-pressure estimates were determined from seven pelitic samples from across the Sugarloaf Mountain thrust sheet. The detailed electron microprobe analysis are

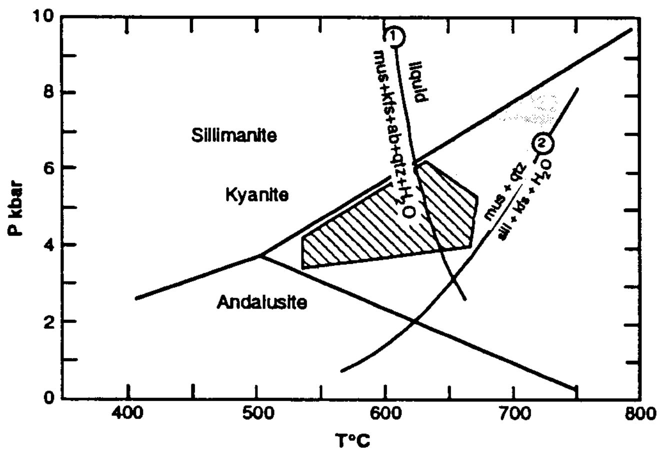


Figure 7. Metamorphic conditions of the pelites in the Sugarloaf Mountain sheet, Columbus Promontory. Shaded area is P–T range estimated from field and petrographic relationships. Striped area P–T range estimated from geothermobarometry. Reaction (1) after Thompson and Algor (1978); Reaction (2) after Chatterjee and Johannsen (1974). Al_2SiO_5 triple point after Holdaway (1971).

listed in Davis (1993). Temperature–pressure values were estimated using rim compositions of garnet and the compositions of matrix biotite, plagioclase, and muscovite (Table 1). Based on the retrograde reaction textures discussed above, P–T estimates are related to post metamorphic peak conditions.

Table 1. Mineral compositions and estimated temperature and pressures for pelitic schist of the Sugarloaf Mountain thrust sheet.

	cm 175	cm 501	cm 126	cm 677	cm 427	12–2–4
Garnet						
X_{Alm}	0.771	0.778	0.673	0.853	0.721	0.642
X_{Pyr}	0.105	0.104	0.119	0.100	0.084	0.110
X_{Sps}	0.094	0.080	0.150	0.026	0.142	0.179
X_{Gro}	0.030	0.039	0.060	0.020	0.052	0.069
Biotite						
X_{Ann}	0.567	0.563	0.601	0.576	0.595	0.514
XPhl	0.433	0.437	0.399	0.424	0.404	0.486
Plagioclase						
X_{An}	0.185	0.178	0.331	0.117	0.303	0.320
Muscovite						
X_K	0.798	0.850	0.838	0.759	0.854	0.902
X_{Na}	0.109	0.136	0.069	0.117	0.068	0.080
$X_{Al^{IV}}$	0.938	0.930	0.917	0.948	0.927	0.923
Temperature (°C)						
F & S (1978)	565	555	640	535	545	580
G & S (1984)	580	560	670	545	590	610
H & S (1982)	590	570	665	545	565	610
Pressure (kbs)						
N & H (1981)	4.1	4.9	5.5	4.0	4.0	5.2
H & S (1982)	3.8	4.5	5.2	3.5	4.0	5.1
G & S (1981)	3.5	4.1	4.1	3.6	4.1	4.0
H & C (1985)	4.5	5.0	7.2	4.2	4.2	5.2

Representative mineral compositions; X_{Alm} = (Fe/Fe+Mg+Mn+Ca), X_{Pyr} = (Mg/Fe+Mg+Mn+Ca), X_{Sps} = (Mn/Fe+Mg+Mn+Ca), X_{Gros} = (Ca/Fe+Mg+Mn+Ca); X_{Ann} = (Fe/Fe+Mg) X_{Phl} = (Mg/Fe+Mg); X_{An} =Ca; X_K= K, X_{Na}=Na, $X_{Al^{IV}}$= Al ^{IV}; all Fe is assumed Fe^{2+};

Temperatures were estimated from the Fe–Mg exchange reaction between garnet and biotite based on the relationship of Ferry and Spear (1978)

$$Mg_3Al_2Si_3O_{12} + KFe_3AlSi_3O_{10} = Fe_3Al_2Si_3O_{12} + KMg_3AlSi_3O_{10}(OH)_2 \quad (4)$$

The experimental calibration of Ferry and Spear (1978) and the empirical calibrations of Hodges and Spear (1982) and Ganguly and Saxena (1984) were used to estimate temperatures. The refinement of Hodges and Spear (1982), based on the experimental data of Ferry and Spear (1978), corrects for non–ideal mixing of Ca in garnets. The Ganguly and Saxena (1984) model corrects for the effects of both Mn and Ca on Fe–Mg mixing in garnet. Both of these calibrations generally result in higher estimated temperatures than the Ferry and Spear calibration (Spear and Peacock, 1989).

Pressures were estimated using equilibrium relationships that involve net exchange reactions between the grossular component in garnet and the anorthite component in plagioclase. The techniques used include the garnet–plagioclase–sillimanite–quartz (GASP) barometer of Ghent and Stout (1981) which is based on the reaction

$$3CaAl_2Si_2O_8 = 2Al_2SiO_5 + Ca_3Al_2Si_3O_{12} + SiO_2 \quad (5)$$

For this barometer, the calibrations of Newton and Haselton (1981) and Hodges and Spear (1982) were used. For comparative purposes, pressures were also estimated using the garnet–plagioclase–muscovite–biotite (GPMB) barometer which is based on the reaction

$$Fe_3Al_2Si_3O_{12} + Ca_3Al_2Si_3O_{12} + KAl_3Si_3O_{10}(OH)_2 = 3CaAl_2Si_2O_8 + KFe_3AlSi_3O_{10}(OH)_2 \quad (6)$$

For this relationship, the calibrations of Ghent and Stout (1981) and Hodges and Crowley (1985) were used.

The results of geothermobarometric estimates suggest post peak temperatures in the range of 535° to 690°C and pressures in the range of 3 to 6 kb (Table 1). These P–T estimates fall within the sillimanite stability field, consistent with the observation of sillimanite being the only Al_2SiO_5 polymorph present in any of the samples analyzed (Fig. 7). For individual samples, temperatures estimated by the different calibrations of the Fe–Mg exchange display differences of less than 35°C. The Ferry and Spear (1978) calibration consistently yielded the lowest temperatures. Temperatures estimated using the Hodges and Spear (1982) and Ganguly and Saxena (1984) calibrations yielded temperatures generally between 20° to 30° C higher than the Ferry and Spear (1978) calibration. The highest temperatures are recorded by sample cml26a, consistent with other petrologic and compositional evidence indicating that this sample may have attained the highest temperature of the sample suite; sample cml26a contains the best–developed prismatic sillimanite. In a review of the Al_2SiO_5 polymorphs, Kerrick (1986) suggested that the presence of fibrolitic versus prismatic sillimanite in pelitic rocks may be a function of P–T conditions: prismatic sillimanite being favored at higher temperatures and pressures. Furthermore, biotite in cml26 contained the highest Ti content of all samples analyzed (Davis, 1993). Studies by Guidotti (1984) and Spear and others (1990) have suggested that Ti content in minerals such as biotite and amphibole increases with increasing metamorphic grade.

Pressures calculated using the GASP barometer indicate values for the entire sample suite ranging from 3 to 5 kb. Pressures estimated using the calibration of Newton and Haselton (1981) are generally 0.5 to 1 kb higher than those calculated using the calibrations of Hodges and Spear (1982). Pressures estimated by the empirically derived GPMB barometer using the calibrations of Ghent and Stout (1981) and Hodges and Crowley (1985), yielded values generally consistent with pressures determined by GASP. Values determined by the Ghent and Stout (1981) calibration range from 3 to 4 kb, those using the Hodges and Crowley (1985) are about 1 kb higher and range from 4 to 5 kb.

STRUCTURAL GEOLOGY

Tocks contained within the crystalline thrust sheets of the Columbus Promontory record a polyphase deformation history phase. The focus of this section is on the three ductile deformation phases (D_1, D_2, D_3) important to the understanding of crystalline thrusting in this part of the Inner Piedmont. These phases are separated based on their timing relative to M_1 and to the formation of the pervasive S_2 that occurs throughout the western Inner Piedmont. The relative chronology of deformation phases is as follows: D_1 was pre–M_1 and pre–S_2; D_2 was pre– to syn–M_1 and resulted in S_2; D_3 was late–M_1 and post–S_2.

D_1 Deformation. D_1 structures are rarely observed in the study area. The earliest structure observed in the Columbus Promontory are F_1 folds that fold compositional layering (S_0?) preserved in the Poor Mountain Amphibolite beneath the Henderson Gneiss in the Tumblebug Creek thrust sheet (Fig. 8a). The kinematics of D_1 deformation is difficult to determine because of the lack of exposure (or recognition?) of these structures in the Columbus Promontory and overprinting of penetrative D_2 structures discussed below.

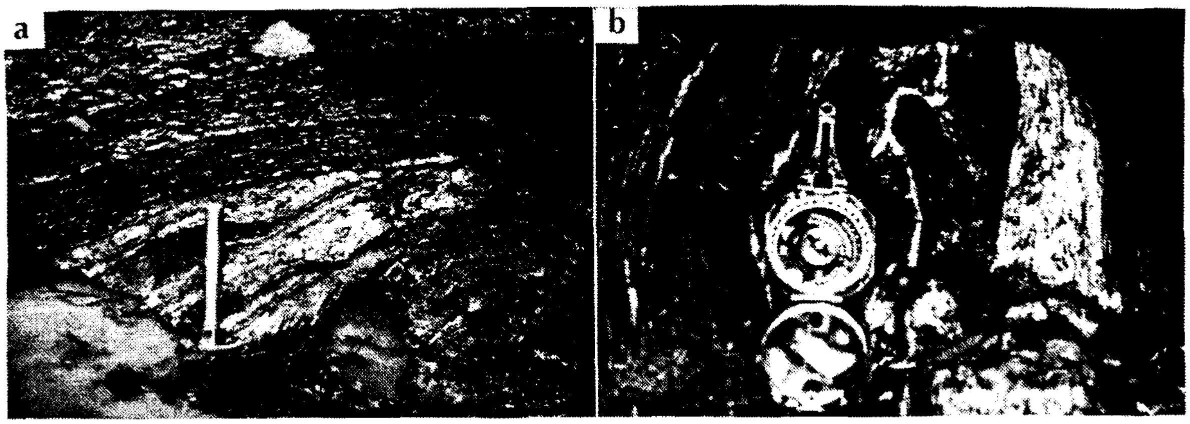


Figure 8. a) Exposure of Tumblebug Creek thrust (TCT) showing truncation of F_1 folds in Poor Mountain amphibolite (PMA) by Henderson Gneiss (HG). View looking NW> Hammer is 40 cm. b) Folded TCT with thrust contact transposed into S_2 foliation. Compass is 20 cm.

D_2 Deformation. D_2 represents the most pervasive deformation phase in the study area and is divided into two subphases designated D_{2a} and D_{2b}. D_{2a} deformation includes pre– to syn-M_1 emplacement of the Tumblebug Creek thrust sheet, development of F_{2a} isoclinal, recumbent folds in the Henderson Gneiss, and initial development of the S_2 foliation across the area. D_{2b} was a penetrative deformation characterized by extensive tight to isoclinal, recumbent to reclined F_{2b} folding , formation of the S_2 mylonitic foliation, and a mineral stretching lineation (L_2) within the plane of S_2. D_{2b} deformation also includes emplacement of the Sugarloaf Mountain thrust sheet and Mill Spring thrust sheet. Metamorphic textures indicate that D_{2b} was coeval with upper amphibolite facies metamorphism in the western Piedmont. Because of the importance of D_2 structures to the structural development of this area they are discussed in greater detail below.

D_3 Deformation. D_3 represents syn- to post-M_1 deformation and is characterized by folds and faults. F_3 folds that deform S_2 occur in the Columbus Promontory, but are not penetrative. F_3 folds are distinguished from F_2 folds by their distinctive geometry. F_3 synform–antiform pairs in many cases have a common limb that is sheared or faulted. This faulted limb generally parallels the S_2 mylonitic foliation and records NW–W displacement. D_3 faulting in the Columbus Promontory includes a near vertical, northwest–trending, ductile strike–slip fault that cuts the Mill Spring thrust sheet (Fig. 9b). A NW–SE horizontal mineral stretching lineation and deflection of wall–rock units into the fault indicates a sinistral displacement sense. The areal extent of this structure has not been determined, despite detailed mapping. Metamorphic textures and crosscutting relationships indicate this feature also developed during M_1, but after emplacement of the Mill Spring thrust sheet and development of S_2 and F_2.

D_2 Structures

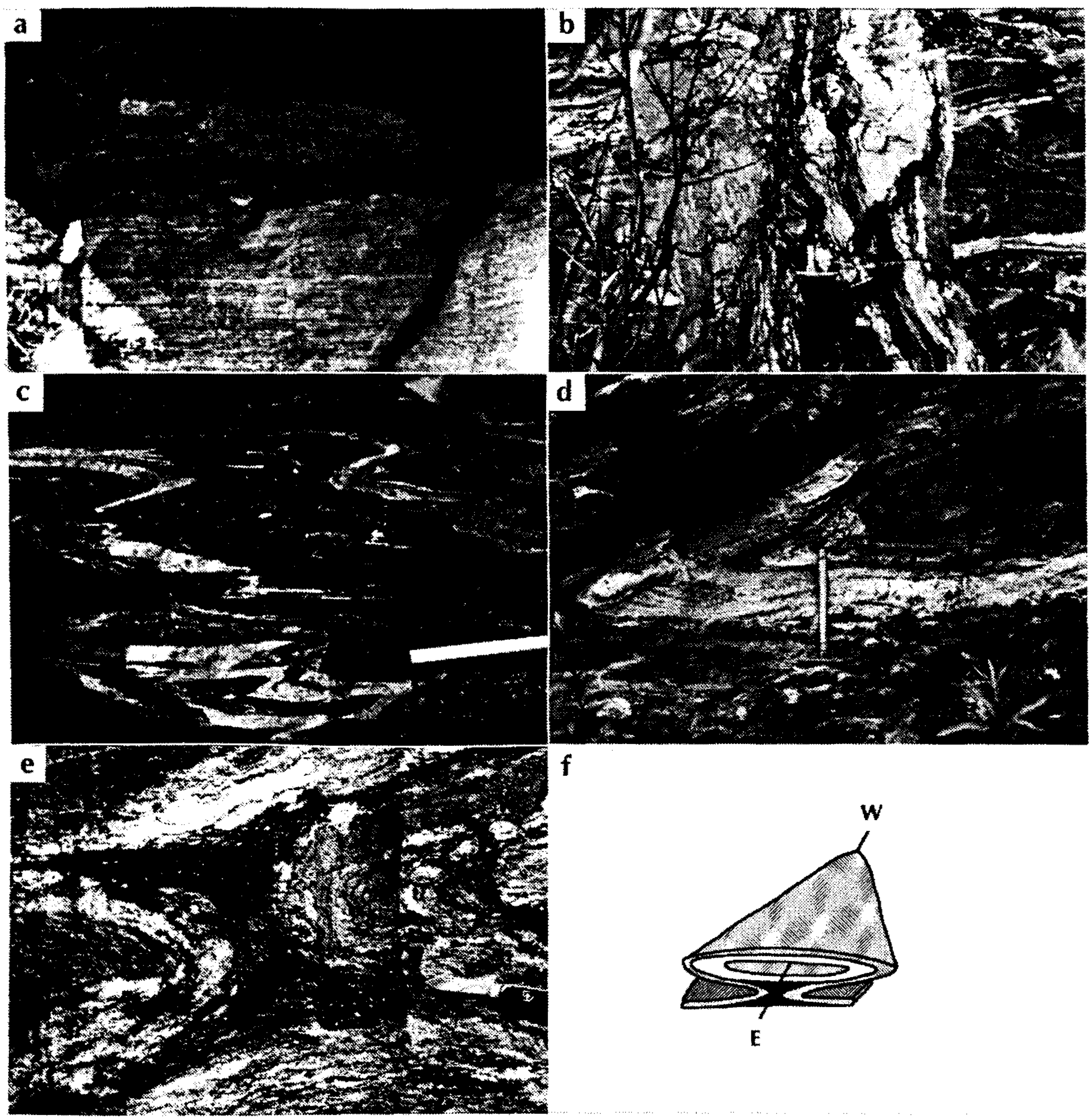
Faults. The map–scale structure of the Columbus Promontory is dominated by D_2 crystalline thrust sheets including the Tumblebug Creek thrust sheet, Sugarloaf Mountain thrust sheet, and Mill Spring thrust sheet (Fig. 2; Plate I). D_2 thrust faults dip gently southeast parallel to S_2 and strike northeast roughly parallel to the orogen, although some local discordance is present. D_2 thrusting is separated into two phases designated D_{2a} and D_{2b}.

D_{2a} faulting involved emplacement of the Tumblebug Creek thrust sheet in the Columbus Promontory (Figs.2, 8a, and 8b; Plate I). The Tumblebug Creek thrust sheet is correlative with the Stumphouse Mountain thrust sheet of Liu (1990) and similarly contains only the Henderson Gneiss and later intrusives. Several fundamental structural relationships have been observed along the Stumphouse Mountain thrust and the Tumblebug Creek thrust: (1) footwall rocks and earlier F_1 isoclinal folds in the Poor Mountain Formation are sharply truncated by these faults; (2) emplacement of these faults occurred before the development of pervasive S_2; and (3) the fault contact has been transposed into S_2 (Figs. 8a and 8b).

D_{2b} thrusting involved emplacement of the Sugarloaf Mountain thrust sheet and Mill Spring thrust sheet. The Sugarloaf Mountain thrust sheet, originally identified by Lemmon (1973) and Lemmon and Dunn (1973a), emplaced sillimanite–bearing pelitic schist, quartzite, and amphibolite of the Poor Mountain Formation and biotite gneiss of the upper Mill Spring complex over the Henderson Gneiss. The contact between the Henderson Gneiss and the Sugarloaf Mountain thrust sheet and represents one of the most abrupt structural contacts in the Columbus Promontory (Fig. 9a). Everywhere this fault has been observed, the contact is knife sharp, subhorizontal, and does not appear to have been appreciably folded after emplacement. Extreme grain–size reduction of the augen in the Henderson Gneiss occurs along this contact. The Mill Spring thrust sheet is the structurally highest thrust sheet in the Columbus Promontory and placed migmatitic biotite–granitic, amphibole gneiss, and amphibolite of the lower Mill Spring complex over rocks of the upper Mill Spring complex and Poor Mountain Formation.

Folds. Recumbent F_2 folds (Figs. 9c and 9d) are the dominant folds preserved throughout the Columbus Promontory and are subdivided into two coaxial phases designated F_{2a} and F_{2b}. Both phases are typically isoclinal, reclined to recumbent with thickened hinges and attenuated limbs. F_{2a} folds have been recognized

Figure 9. a) Contact of Sugarloaf Mountain thrust (SMT) showing Poor Mountain schist structurally overlying the Henderson Gneiss. Knife if 10 cm. b) NW–trending strike–slip in the Mill Spring thrust sheet. View is toward the SE. Hammer is 40 cm. c) F_{2b} folds in Poor Mountain amphibolite. Pen is 15 cm. d) F_2 folds in Mill Spring complex. Hammer is 60 cm. e) Part of W–vergent sheath fold in the Sugarloaf Mountain thrust sheet. View is facing E parallel to E–W trending L_2 mineral stretching lineation. Knife is 10 cm. f) Interpreted of geometry of sheath fold in (e).



primarily in the Henderson Gneiss within the Tumblebug Creek thrust sheet. F_{2a} folds are truncated by the Sugarloaf Mountain thrust. F_{2b} folds are the dominant folds recognized in the Columbus Promontory and are defined by S_2 and generally do not contain an axial planar foliation. F_{2b} folds exhibit a wide variation in orientation of hinges (Fig. 10) and sense of vergence. In some cases, F_{2b} folds have long straight hinge lines parallel to L_2, whereas in other cases, F_{2b} hinges are at a high angle to L_2. F_{2b} also includes sheath folds (Figs. 9e and 9f) observed in the Sugarloaf Mountain thrust sheet and the Mill Spring thrust sheet in areas dominated by E–W oriented L_2. These are recognized in outcrop by the characteristic eye, anvil, and arrowhead shapes, and small-scale folds with highly curvilinear hinges; the long axis of the sheath folds parallel L_2. A detailed discussion of the development and kinematic significance of F_2 and F_3 folds in the Columbus Promontory is presented by Yanagihara (this guidebook).

Foliation. S_2 is the most pervasive and penetrative mesoscopic structural feature observed in the Columbus Promontory and throughout much of the Inner Piedmont (Fig 10; Plate I). S_2 generally strikes slightly oblique (N–NE) to the orogen and has a shallow (< 30°) southeast dip, although variations exist. Towards the Brevard fault zone, the strike of foliation parallels the fault zone and dips gently to moderately southeast. Within the Brevard fault zone S_2 foliation dips moderately (~40°). The shallow dip of foliation in the Inner Piedmont with steepening along the Brevard fault zone has also been clearly imaged in seismic reflection profiles (Cook and others, 1979; Costain and others, 1989).

Mesoscopic–components of the S_2 foliation across the Columbus Promontory, including compositional layering, schistosity, and gneissic banding, as well as intrafolial folds and boudinage, indicate that S_2 is composite, in the usage of Tobisch and Paterson (1988), and represents an end–product stage of a strongly transposed, evolved fabric.

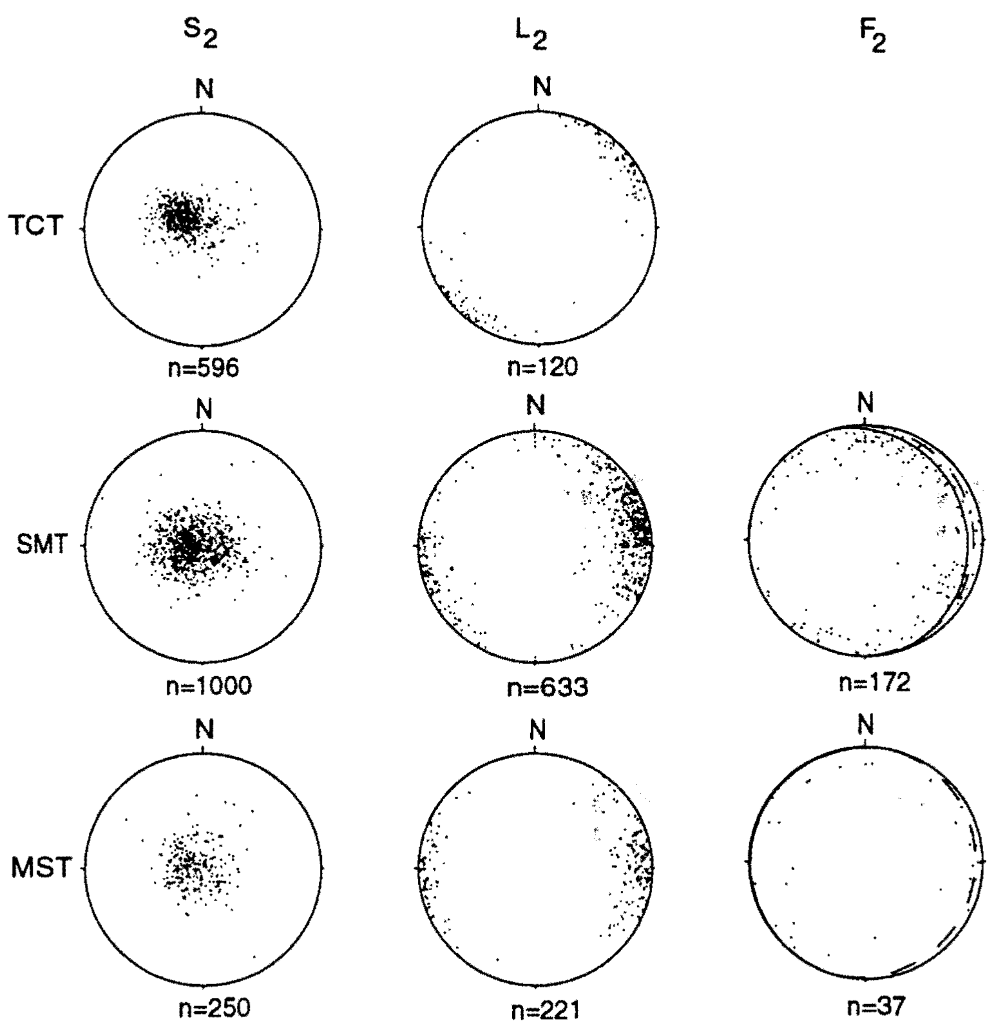


Figure 10. Lower–hemisphere equal–area projections of structural data of D_2 mesoscopic structures Columbus Promontory; TCT–Tumblebug Creek thrust sheet; SMT–Sugarloaf Mountain thrust sheet; MST–Mill Spring thrust sheet. Dashed great circles on equal–area projection of F_2 folds represent plane normal to mean vector of poles to S_2 mylonitic foliation; solid great circles are the best fit plane to F_2 hinges (fold–hinge girdle). Columbus Promontory data compiled from Lemmon (1973) and unpublished data of Davis (1987-1990) and Tabor (1988-1989).

Mesoscopic and microscopic attributes of the S_2 fabric also indicates that this foliation is mylonitic (commonly annealed). This fact is manifested by the strong planarity of the foliation, the presence of mineral stretching lineations on nearly all foliation surfaces, and the widespread occurrence of structures indicative of noncoaxial deformation, such as composite–planar fabrics, asymmetric intrafolial folds, σ and δ type porphyroclasts and porphyroblasts (Simpson, 1986; Simpson and Schmid, 1983), and asymmetric garnets (Figs. 6a, 6b, and 6c).

The mylonitic characteristics of S_2 are best observed in quartz-rich pelitic units. In these rocks, S_2 comprises a well–developed micro– and mesoscopic composite–planar fabric. Components of this fabric are designated S_{2c}, S_{2s}, and S_{2e}, based on their geometric and kinematic significance (Fig. 11). These S_2 foliation components are defined by parallel orientation of the minerals of assemblage A_1 and impart an anastomosing

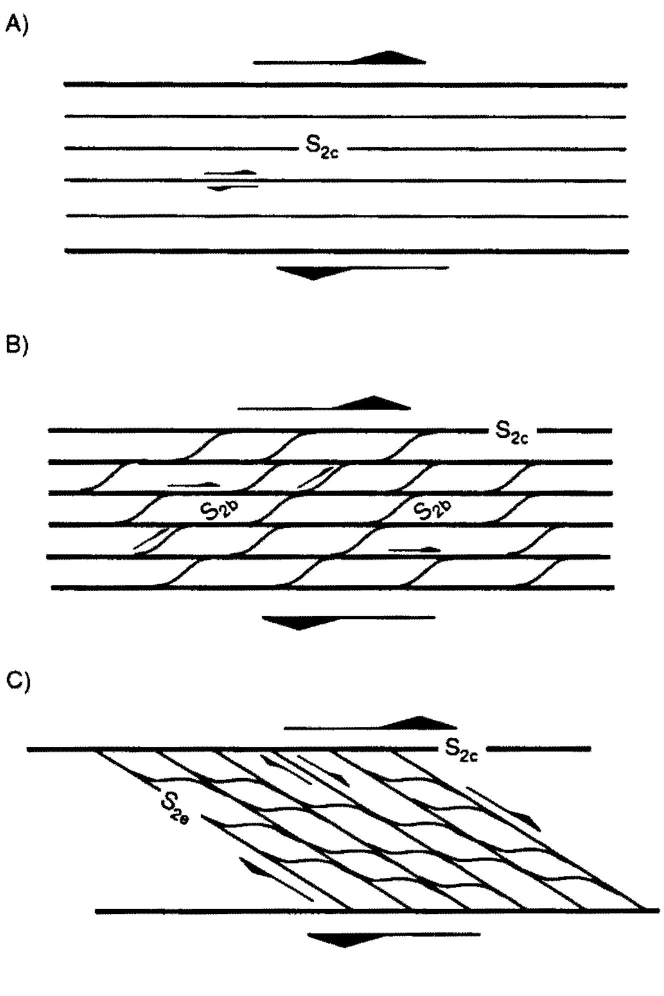


Figure 11. Schematic representation of the geometric and kinematic relationships between S_2 foliation components a) S_{2c}, S_{2s}, and S_{2e}.

microscopic texture to the pelites. These components essentially outline a series of microscopic–scale shear zones within the pelites.

S_{2c} is the best developed foliation component in the pelites and defines the shear zone boundaries. S_{2c} parallels the direction of shear and is analogous to C–surfaces of Berthé and others (1979). S_{2c} is defined by quartz ribbons, muscovite, biotite, fibrolitic sillimanite, and elongate garnets (Fig. 6b).

S_{2s} dips at angles between approximately 10° and 45° from S_{2c} in a direction opposite to the overall direction of shear (Figs. 6a, 6b, and 11). S_{2s} is most commonly defined by sigmoidal–shaped mica (buttons) or mica fish of Lister and Snoke (1984), fibrolitic sillimanite, asymmetric garnets, and quartz pods. S_{2s} is interpreted to represent a component of pure shear or flattening across the microscopic scale shear zones formed similar to that defined by Lister and Snoke (1984). The combination of S_{2c} and S_{2s} define an S–C fabric (Berthé and others, 1979; Simpson and Schmid, 1983; Lister and Snoke, 1984) in the pelites of the Sugarloaf Mountain thrust sheet.

S_{2e} also dips at angles between approximately 10° and 45° from S_{2c}, but in the direction of overall shear (Fig. 11). S_{2e} disrupts S_{2c} and defines a set of extensional shear surfaces in the pelites. S_{2e} is analogous to C' surfaces of Berthé and others (1979), extensional crenulation cleavage (Platt and Vissers, 1980), shear bands (White and others, 1980; Dell'Angelo and Tullis;1989) and normal–slip crenulations (Dennis and Secor, 1988). S_{2e} is defined primarily by parallel alignment of biotite and muscovite as well as by fibrolitic sillimanite and quartz ribbons.

Lineation. Penetrative subhorizontal mineral lineations (L_2) lying in the plane of the S_2 are also a widespread and penetrative mesoscopic structural feature in the study area (Figs. 6d, 6e, 6f, 10, and 12). Associated kinematic indicators including ribboned quartz, stretched and boudinage K–feldspar, and composite–planar fabric suggest the L_2 is a stretching lineation. Across the Columbus Promontory from northwest to southeast and structurally lowest to highest there is a distinct change in the trend of L_2 mineral lineations from NE–SW to E–W (Figs. 10 and 12).

In the Tumblebug Creek thrust sheet, L_2 has a ubiquitous NE–SW orientation (Figs. 10 and 12). L_2 is defined by elongate K–feldspar porphyroclasts, commonly with recrystallized tails, quartz ribbons, and preferentially oriented biotite and muscovite. This orientation is persistent into the Brevard fault zone (Fig. 12). It is important to emphasize that the linear fabric discussed here formed under distinctly different P–T conditions from the coaxial linear fabric related to late Paleozoic retrograde greenschist facies mylonitization in

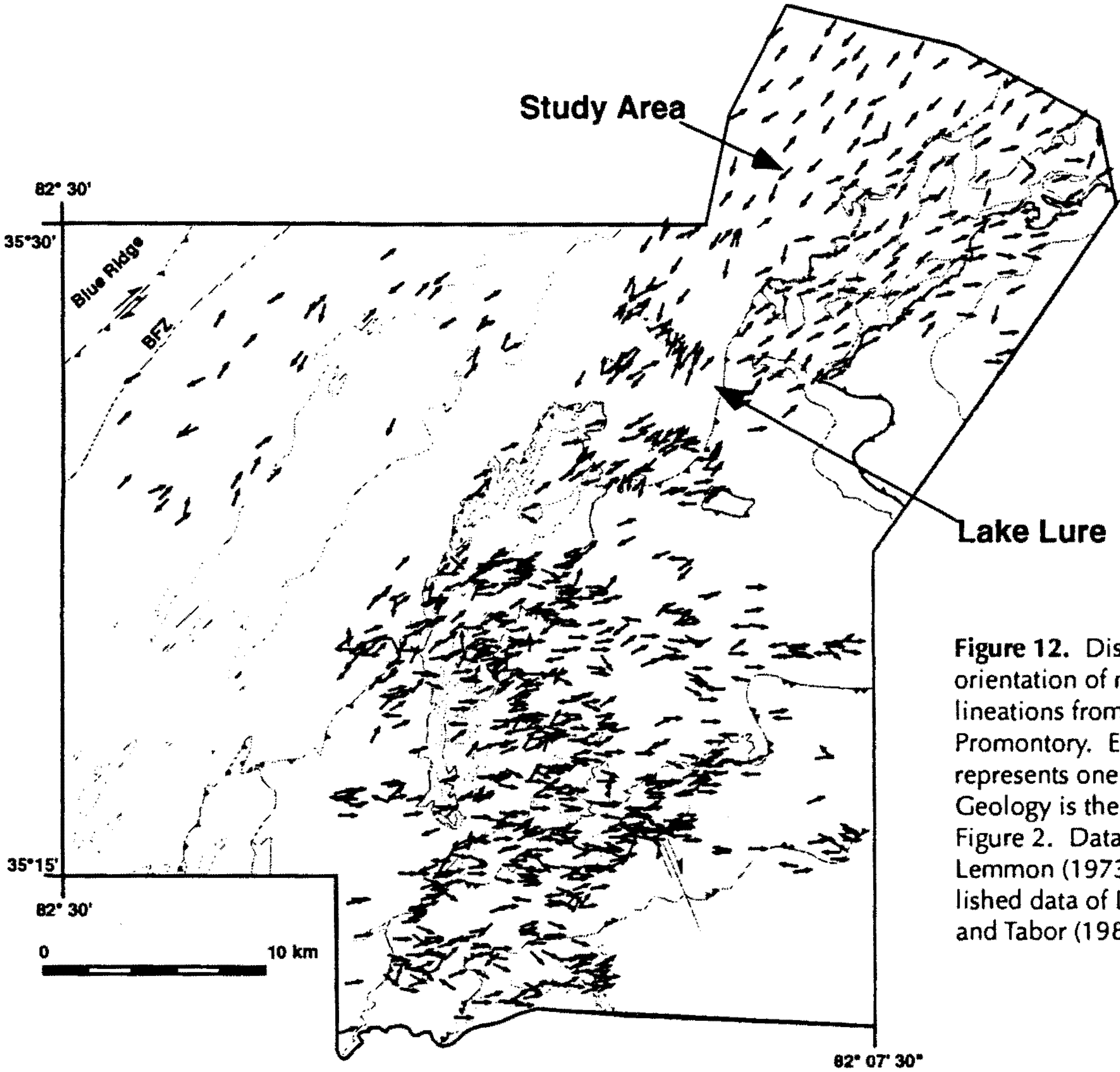


Figure 12. Distribution and orientation of mineral stretching lineations from the Columbus Promontory. Each arrow represents one measurement. Geology is the same as shown in Figure 2. Data compiled from Lemmon (1973) and unpublished data of Davis (1987-1990) and Tabor (1988-1990).

the Brevard fault zone discussed by Reed and Bryant (1964), Bryant and Reed (1970), Hatcher (1969), Griffin (1974), Bobyarchik and others (1988), and Vauchez and Brunel (1988).

The NE–SW oriented L_2 is also present in the western part of the Sugarloaf Mountain thrust sheet (Figs. 10 and 12). Here L_2 , depending on rock type, is defined by recrystallized quartz ribbons, elongate amphibole and micas, and preferentially oriented sillimanite. Farther east in the Sugarloaf Mountain thrust sheet, L_2 changes to a more E–W orientation. Both orientations have been observed at a few localities (Fig 6f); here the curvilinear geometry of the lineations indicate a progressive change in orientation.

In the structurally highest Mill Spring thrust sheet, L_2 is oriented east–west (Figs 10 and 12). Depending on rock type, L_2 is defined by quartz rods, elongate amphibole and micas, and occasional preferentially oriented sillimanite. Abundant kinematic indicators including micro– and mesoscale shear bands, S–C fabrics, sheath folds, and amphibolite layers extended to form boudinage indicate intense east–west stretching and slip.

Thrusting Sequence and Flow Kinematics

Crosscutting relationships of rocks units and mesoscopic structural features, and the progressive changes in orientation of L_2 and associated kinematic indicators, suggests that thrust sheets in the Columbus Promontory were stacked in a foreland– to hinterland progression (west to east) with the structurally lowest Tumblebug Creek thrust sheet being emplaced first, followed by successive emplacement of the Sugarloaf Mountain thrust sheet and Mill Spring thrust sheet. The Tumblebug Creek thrust sheet truncates F_1 folds in the structurally underlying Poor Mountain Amphibolite (Fig. 8a). The Tumblebug Creek thrust sheet has been folded and transposed into the S_2 suggesting emplacement prior to development of the S_2. The Sugarloaf Mountain thrust sheet truncates the outcrop pattern of the Henderson Gneiss and earlier F_{2a} folds within the Henderson Gneiss and parallels S_2. S_2 and other kinematic indicators (primarily L_2 and associated kinematic indicators) within pelites of the Sugarloaf Mountain thrust sheet are defined by mineral phases in assemblage A_1 indicating formation synchronous with D_2 and M_1. The structurally highest Mill Spring thrust sheet truncates outcrop patterns and S_2 within the Sugarloaf Mountain thrust sheet. S_2 and L_2 in the Mill Spring thrust sheet are also defined by M_1 mineral phases. This suggests emplacement of the Mill Spring thrust sheet during D_2 and M_1, but following development of S_2 in the Mill Spring thrust sheet.

The kinematics of flow within Columbus Promontory thrust sheets is recorded by the orientation of L_2 and abundant shear sense indicators including σ and δ type porphyroclasts, composite–planar fabric, and asymmetric and sheath folds. These structures indicate a change in orientation of flow in the PC from west to east and from structurally lowest to highest levels. In the CT, L_2 is oriented NE–SW and shear sense indicators consistently record a top–to–SW shear sense. In the western part of the Mill Spring thrust sheet, L_2 is oriented NE–SW and kinematics indicators record a top–to–SW shear sense similar to those in the structurally underlying Tumblebug Creek thrust sheet. Towards the eastern boundary of the Mill Spring thrust sheet and structurally higher, however, L_2 undergoes a progressive change in orientation to E–W and kinematics indicators record top–to–W shear sense. In the structurally highest Mill Spring thrust sheet, L_2 is E–W and shear sense is almost invariably top–to–W.

MODEL FOR CRYSTALLINE THRUSTING IN THE COLUMBUS PROMONTORY–INNER PIEDMONT

An important result of this investigation is the recognition of a change in the kinematics of flow across the Columbus Promontory. The fact that L_2 and associated shear-sense indicators for both flow directions (W and SW) are defined by the M_1 mineral assemblage indicate that these structures are part of the same kinematic event. These observations suggest the contemporaneous operation of W–directed thrusting and SW–directed stretching in the Columbus Promontory. I believe this kinematic association reflects a D_2 transpressional flow regime caused by the interaction of W–NW directed Columbus Promontory thrust sheets and the buttress effect of a primordial (pre–greenschist facies) Brevard fault zone (Fig. 13).

An important component of the model presented here is that the primordial Brevard fault zone had a ramp or listric geometry during D_2. A listric geometry for the Brevard fault zone has been corroborated by seismic reflection profiles (Clark and others, 1978; Cook and others, 1979; Harris and Bayer, 1979; Harris and others, 1981; Çoruh and others, 1987, Costain and others, 1989). In these profiles, it can be seen that the relatively steep (~40°) dip of the Brevard fault zone flattens to the east where it parallels S_2. The evolution of this listric geometry is not fully understood. It can be argued that this listric geometry is primarily a function of movement of the Brevard fault zone (e.g., Rosman phase of Horton and Butler, 1986) later than D_2. For this model, a listric geometry for the Brevard fault zone during D_2 is assumed.

During D_2 overall flow was W–directed and as

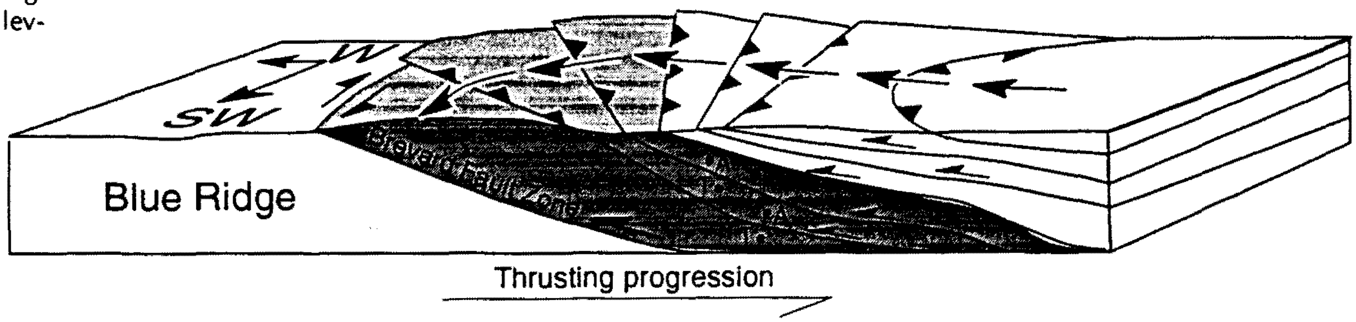


Figure 13. Schematic configuration of the thrust stacking and deformational sequence envisioned for Columbus Promontory and elsewhere in the western Inner Piedmont. Configuration shows dominance of west–directed displacement in the deeper Inner Piedmont and dominance of SW–directed displacement and stretching near the Brevard fault zone. Configuration also shows linkage of the Brevard fault zone and Inner Piedmont thrust sheets

the Columbus Promontory thrusts sheet encountered the NE–striking and relatively steeply SE–dipping Brevard fault zone, displacement of, and flow within the approaching thrust sheets rotated towards the SW (Fig 13). This change in orientation was caused by development of a large component of W–NW shortening across the listric or ramp geometry of the Brevard fault zone that was accommodated by penetrative SW– stretching in the western part of the Columbus Promontory. SW–stretching was most intense in the westernmost thrust sheets, in this case, the Tumblebug Creek thrust sheet and the western part of the Sugarloaf Mountain thrust sheet. To the east at structurally higher levels, strain accommodation was not by SW–stretching, the SW–flow regime essentially was locked, but by break-back thrusting and related W–directed penetrative deformation (eastern Sugarloaf Mountain thrust sheet and Mill Spring thrust sheet) that reflected the larger–scale W–directed flow regime of the Inner Piedmont terrane (Fig. 13). W–directed deformation was most intense in the structurally highest and easternmost Mill Spring thrust sheet.

The most intense part of this transpressional flow occurred during D_2, but it also included D_3 structures in the later stages of this deformation which primarily record W–NW displacement. Total displacement in the Columbus Promontory during this deformation was accommodated not only by crystalline thrusting, but also by intracrystalline processes exemplified by the penetrative nature of the S_2 mylonitic foliation and L_2 across the Columbus Promontory. The geometric characteristics of S_2 foliation, including the extreme planarity and abundance of evidence indicative of noncoaxial deformation, suggest that during D_2, S2 was active as a C–surface (Berthé and others, 1979) and accommodated a considerable amount of the inhomogeneous strain. These characteristics, commonly observed in shear zones (e.g., Ramsay and Graham, 1980; Lister and Snoke, 1984), suggest that the Columbus Promontory can essentially be considered a shear zone with the primordial Brevard fault zone as the western boundary.

The model presented here can be extended to include the complex of thrust sheets in the Inner Piedmont of South Carolina and NE Georgia including the Stumphouse Mountain thrust sheet (Liu, 1990), the Alto allochthon (Hopson and Hatcher, 1988), and the Walhalla nappe and Six Mile thrust sheet of (Griffin, 1971 and 1974). This crystalline thrust stack records an nearly identical kinematic pattern as the Columbus Promontory. Expansion of the shear zone concept to include this area and perhaps much of the Inner Piedmont suggests that this terrane represents a shear zone of orogenic or crustal dimensions. Similar to thrust sheets in the Grenville orogen described by Nadeau and Hanmer (1992), ductile thrust sheets in the Inner Piedmont appear to record changes in rheological behavior during their emplacement history related to the M_1 metamorphic event that affected the western Inner Piedmont. The Inner Piedmont thrust sheets had to have cooled enough to have sufficient strength to remain coherent during transport, but were sufficiently ductile to become penetratively deformed. The details of this more regional model are discussed by Davis and others (in review).

The model presented here is similar to that proposed by Nadeau and Hanmer (1992) to explain the penetrative nature of deformation and complex patterns of flow during amphibolite granulite to upper amphibolite facies ductile thrusting in the Central Gneiss belt of the Grenville orogen. Nadeau and Hanmer (1992) recognized break back-thrusting as an important mechanism in the tectonothermal development of the Grenville orogen and described a transpressional flow regime in the Huntsville thrust zone that similar to that recognized in the Columbus Promontory. Nadeau and Hanmer (1992) also discussed the penetrative nature of deformation in the Grenville orogen and suggested that the scale of this penetrative deformation is comparable with the largest examples of high–grade thrust belts recognized elsewhere. Similarities in structural style and deformation processes (e.g., ductile thrust and penetrative deformation) between the Inner Piedmont and the Grenville orogen suggests that the Grenville orogen may be a good analogue by which dutile deformation can be framed.

Structural patterns like those in the Columbus Promontory have also been recognized in other orogenic belts. Studies of finite strain in ductile thrust zones in the Scandinavian and British Caledonides (Lisle, 1984; Coward and Potts, 1983), and in the Variscan belt (Brun and Burg, 1982), reveal large areas with longitudinal mineral stretching lineations subparallel to thrust fronts and normal to the regional thrust transport directions. Coward and Potts (1983), using the Moine thrust as an example, proposed that longitudinal strain may be explained by differential movement and are related to complex strain patterns developed at the frontal tips of thrust zones. LaGarde and Michard (1986) interpreted longitudinal strain, along the frontal tip of the Central Meseta thrust in the Rehamna massif in the Moroccan Meseta to be the product of thrust–wrench shearing, combining ductile thrusting and wrenching during progressive synmetamorphic shortening. Oldow (1990) has shown displacement compatibility for coeval transcurrent and contractional faults requires that these faults share a common décollement system. Such a linked décollement system can produce a complex array of transport directions within a single orogen that include dip–slip and strike–slip motion.

Transpressional models to explain the Alleghanian (retrograde) transcurrent displacement along the Brevard fault zone, the Brookneal shear zone, and faults of the eastern Piedmont fault system (Hatcher and others, 1977) have been proposed by LeFort (1984), Gates and others (1986, 1988), and Vauchez and others (1987). These models generally discuss only the retrograde movement on the Brevard fault zone, and thus do not discuss the possible synkinematic relationship between the Brevard fault zone and Inner Piedmont thrust sheets. The transpressional model presented in this paper is significantly different in that it suggests a kinematic connection between early Brevard fault zone and Inner Piedmont thrust sheets during the latest (?) upper amphibolite facies tectonothermal event that affected these rocks.

The kinematic linkage between the Brevard fault and Inner Piedmont thrust sheets was, in some respects, implied by the earlier studies of Clarke (1952), Bentley and Neathery (1970), Griffin (1971b, 1974a), Hatcher (1969, 1972, and 1978a), Stirewalt and Dunn (1973), and Edelman and others (1987), however, these studies lacked supporting kinematic data. I suggest that the new data presented in this paper provides strong kinematic evidence for such a flow history in the Inner Piedmont.

CONCLUSIONS

The Columbus Promontory represents a unique area along the Blue Ridge physiographic front in North Carolina where geologic features characteristic of the southern Appalachian Inner Piedmont are well exposed. This area provides an outstanding opportunity to examine the crystalline thrusting process the Inner Piedmont terrane. This study adds new data to the limited, but continually growing, data base for the Inner Piedmont terrane as well as some new thoughts on the process of crystalline thrusting in this complex terrane.

Three crystalline thrust sheets are present in the Columbus Promontory that in ascending order include the Tumblebug Creek, Sugarloaf Mountain, and Mill Spring thrust sheets. These thrust sheets contain a distinct lithostratigraphy that helps define their extent. The Tumblebug Creek thrust sheet contains the 509 Ma Henderson Gneiss and 438 Ma granitoid intruded into the Gneiss; the Sugarloaf Mountain thrust sheet contains rocks of the Poor Mountain Formation, the Sugarloaf gneiss, and upper Mill Spring complex; the Mill Spring thrust sheet contains rocks of the mafic–rich lower Mill Spring complex.

Three phases of ductile deformation (D_1, D_2, D_3) are present in the Columbus Promontory which place constraints on the history of emplacement and internal deformation of this crystalline thrust stack. Detailed geologic mapping and mesoscopic suggests that the most significant amount of thrust displacement and internal deformation occurred during D_2 which was synchronous with upper amphibolite sillimanite–grade metamorphism. Detailed mapping and mesoscopic analysis also indicates that the kinematics of flow during D_2 was more complex than previously recognized in the Inner Piedmont. This involved a partitioned flow regime involving contemporaneous W-NW directed thrusting and SW-directed stretching caused by the interaction of Inner Piedmont thrust sheets and the listric geometry of the primordial Brevard fault zone; the flow regime in the Columbus Promontory was transpressional. An important consequence of the model presented in this paper is the kinematic linkage between the Brevard fault zone and Inner Piedmont thrust sheets during upper amphibolite facies conditions. The penetrative nature of this D_2 deformation in the Columbus Promontory and throughout the Inner Piedmont terrane suggests that the Inner Piedmont may be considered an orogenic or crustal shear zone with the Brevard fault zone as the western boundary.

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EVOLUTION OF FOLDS ASSOCIATED WITH D_2 AND D_3 DEFORMATION AND THEIR RELATIONSHIP WITH SHEARING IN A PART OF THE COLUMBUS PROMONTORY, NORTH CAROLINA

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ABSTRACT

Five major deformation events are recognized in the Inner Piedmont and, of these, D_2 and D_3 are the most significant. Of the two events, D_2 accounts for virtually all the shear strain observed in the Inner Piedmont and involved penetrative deformation during peak-M_2 metamorphism. D_3 involved considerably less shear strain and probably occurred following peak-M_2 metamorphism but prior to complete cooling of the rocks. Most F_2 folds are probably the result of perturbations within the shear plane that were passively amplified to produce sheath folds. The high shear strains involved with D_2 deformation allowed the limbs of the sheath folds to rotate into parallelism with the transport direction and the L_2 mineral lineation. This interpretation is supported by the geometry of the majority of F_2 folds, which are isoclinal recumbent and are oriented parallel to L_2, and probably represents the limbs of sheath folds whose axial culminations remain unexposed. D_3 folds occur in a variety of geometries ranging from open upright to nearly isoclinal recumbent. This variety reflects decreasing strain intensity and gradually decreasing temperatures that increased the competency of the rocks.

Hansen analysis indicates the direction of shearing, which varied from W to SW, did not change significantly between D_2 and D_3 deformation. The analysis and textural relationships suggest that D_2 and D_3 represent a single, continuous episode of deformation during which the ductility of the rocks decreased. This change in rheology led to the distinctive geometries associated with the two deformation events.

INTRODUCTION

The western Inner Piedmont was affected by multiple episodes of deformation and at least five have been identified in the Columbus Promontory (see Fig. 1 in Hatcher, this guidebook). The first episode, D_1, involved isoclinal recumbent folding and is synchronous with an episode of early, M_1, metamorphism (Lemmon, 1973; Hopson and Hatcher, 1988; Davis, 1993). Subsequent metamorphism and deformation have transposed and reoriented D_1 structures to make it the most poorly understood event in the western Inner Piedmont. The two latest events, D_4 and D_5, involved gentle, upright folding that interfered to produce a dome-and-basin pattern (Davis, 1993).

The two most important deformation events, D_2 and D_3, are interpreted to represent a single continuous event (Hopson and Hatcher, 1988; Liu, 1991; Davis, 1993). They have been subdivided on the basis of the distinctive geometries of their associated structures: D_2 structures are penetrative whereas D_3 structures are nonpenetrative at all scales. Both events are coeval with M_1 metamorphism of Davis (this guidebook) which reached the sillimanite-muscovite zone in pelitic schist and middle- to upper-amphibolite facies in amphibolite. The most pervasive structural element in the Columbus Promontory is the S_2 foliation. It is defined by the parallel orientation of M_2 mineral assemblages indicating that D_2 deformation is contemporary with peak M_2 metamorphism. D_3 structures deform the S_2 foliation and are associated with a later, weakly-developed, nonpenetrative foliation (S_3) defined by chlorite and sericite (Hopson and Hatcher, 1988). Such deformation is interpreted to have occurred during post-peak M_2 metamorphism but prior to complete cooling of the rocks (Griffin, 1969, 1971, 1974; Hatcher, 1969; Hopson and Hatcher, 1988; Liu, 1991; Davis, 1993). Of the two events, D_2 accounts for virtually all the shear strain observed in the Inner Piedmont whereas D_3 involved considerably less shear strain. A more detailed treatment of the associated structures for both deformation events and their relationship with M_2 metamorphism in this area is presented by Davis (this guidebook).

Transport directions during D_2 deformation, determined using winged porphyroclasts, mineral lineations, S-C mylonites, and snowball garnets, are W-directed in areas to the southeast changing gradually to SW-directed in areas to the NW (Fig. 1, Yanagihara and Davis, 1992; Davis, 1993). Because deformation during D_3 was nonpenetrative, shear-sense indicators associated with this episode are rarer, but the orientation of boudinage neck lines and strike-slip faults suggest that transport directions remained W to SW directed. Associated with both episodes are folds that have a variety of geometries and orientations. The origin of these folds is interpreted to be shear-related (Davis, 1993), but their use in determining shear-sense has largely been ignored. This paper addresses this aspect of deformation in a part of the Columbus Promontory by first documenting the evolution and geometries of folds associated with D_2 and D_3, and then uses fold vergence to determine the shear sense during both folding events.

More than 100 folds were measured during field mapping in the Lake Lure, Shingle Hollow, Sugar Hill, and Moffitt Hill quadrangles. The study area, covering approximately 180 km^2 (70 mi^2), is located in northern Rutherford and southern McDowell counties, northwest of Lake Lure in western North Carolina (Fig. 1). The majority of these folds were measured in the Sugarloaf Mountain thrust sheet (Lemmon, 1973; Davis, 1993). The Sugarloaf Mountain thrust sheet contains rocks of the Poor Mountain Formation and upper Mill Spring complex (see Davis, this guidebook). The data set also includes unpublished data by Christopher Jayne measured immediately south of the study area. In the entire data set, the vergence of 51 folds could be determined.

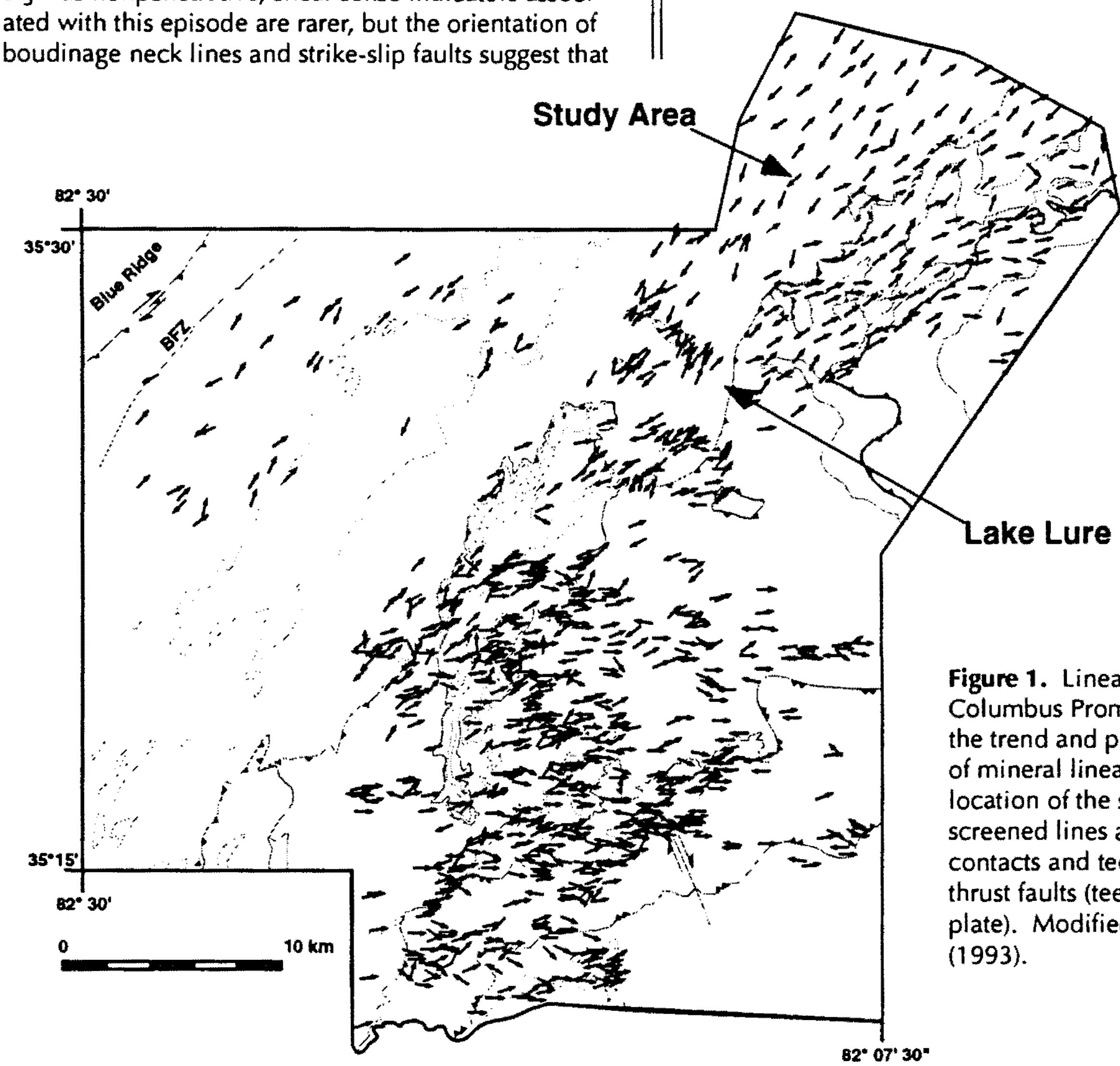

Figure 1. Lineation map of the Columbus Promontory showing the trend and plunge direction of mineral lineations and the location of the study area. The screened lines are geologic contacts and teethed lines are thrust faults (teeth on upper plate). Modified from Davis (1993).

FOLDS ASSOCIATED WITH D_2 AND D_3

Folds associated with D_2 and D_3 deformation exhibit a variety of geometries and orientations (Fig. 2), and differ in several aspects. F_2 folds are typically penetrative at all scales, are isoclinal recumbent, exhibit thickened hinges and attenuated

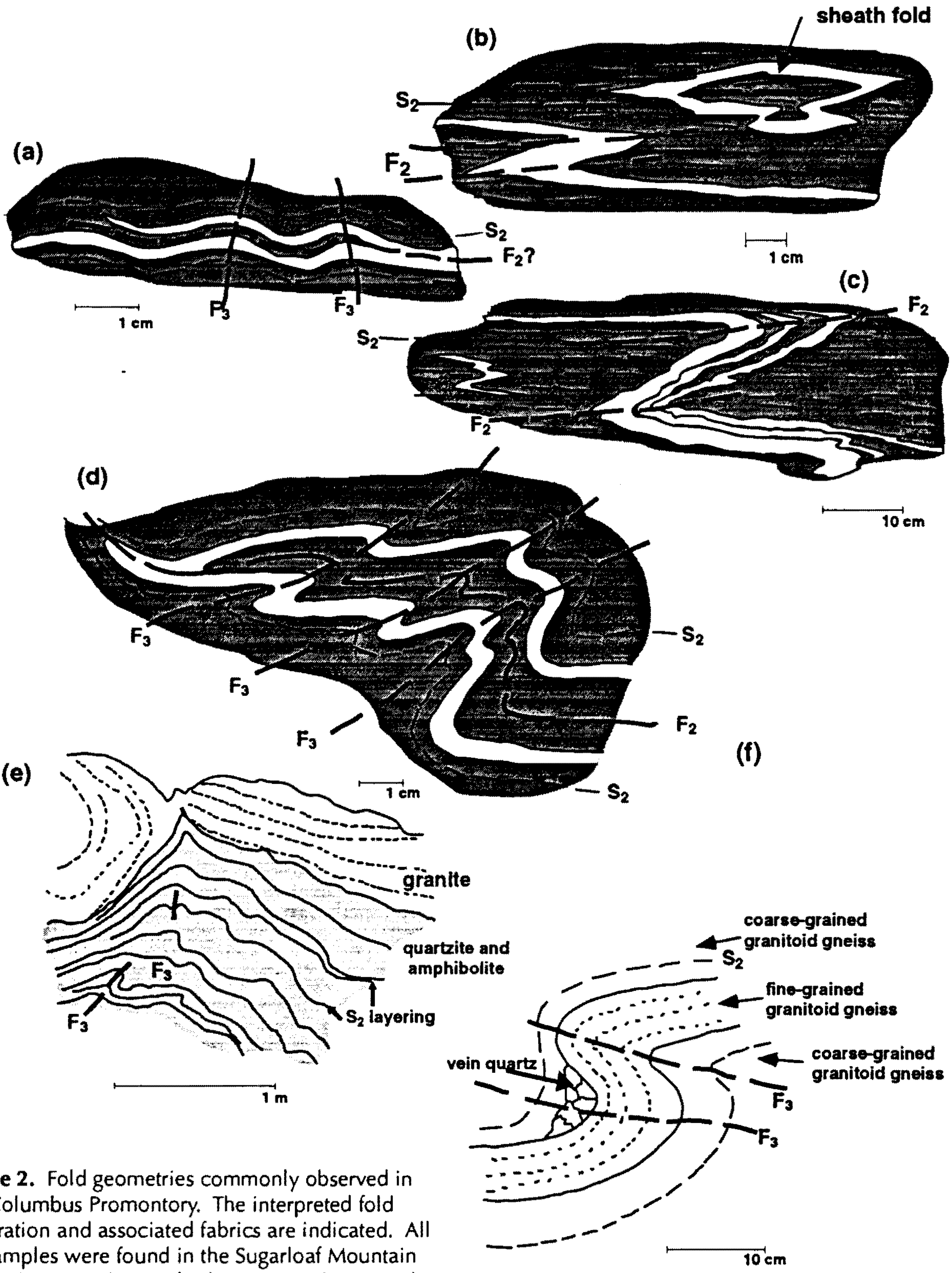

Figure 2. Fold geometries commonly observed in the Columbus Promontory. The interpreted fold generation and associated fabrics are indicated. All six samples were found in the Sugarloaf Mountain thrust sheet, northeast of Lake Lure. a), b), c), and d) occur in the Poor Mountain Amphibolite, e) occurs in Poor Mountain Quartzite with minor amphibolite, and f) occurs in granitic gneiss of the upper Mill Spring formation.

nism. F_3 folds are typically nonpenetrative, variably inclined, and tight to open. Early F_3 folds tend to have thickened noses and attenuated limbs, like F_2 folds, but are classified as F_3 because they are nonpenetrative.

Table 1. Comparison of F_2 and F_3 folds on the Columbus Promontory.

	F_2	F_3
Interlimb angle	Isoclinal	Open to tight
Axial planes	Parallel axial planar with foliation	Variable: upright to gently inclined
Hinge lines	Variable: mostly NE to ESE (parallel to mineral lineation)	Variable: mostly NW to ESE
Morphology	Thick hinges, attenuated limbs	Variable: Older folds resemble F_2 folds, deformed layers in younger folds maintain constant thickness
Associated fabrics	Penetrative foliation (S_2) and lineation (L_2)	Weakly developed non-penetrative foliation (S_3) and boudinage

The layers in later folds tend to maintain their original thickness around the nose and limbs. Their geometry and close association with boudinage suggest that adjacent rock layers were beginning to acquire a competency contrast during D_3 deformation (Table 1).

Stereonet analysis of the folds indicates the hinge lines of both generations plot along a great circle that nearly approximates the orientation of the S_2 foliation (Figs. 3 and 4). Likewise, the great circle defined by the mean vector for the axial plane poles also nearly parallels the foliation. These relationships suggest a contemporary origin for the folds and the S_2 foliation and will be discussed later in this paper.

F_2 Folds. The majority of F_2 folds are intrafolial indicating they underwent high shear strains. The parallel axial planar orientation of the folds and the foliation, and the parallelism of the fold axes and D_2 mineral lineations (L_2) suggests a contemporary origin for the folds, foliations, lineations, and other D_2 structures (Fig. 4). These relationships are common throughout the Inner Piedmont in northeast Georgia and northwest South Carolina (Griffin, 1967, 1969, 1971, 1974; Hatcher, 1969, 1970; Hopson and Hatcher, 1988; Liu, 1991) and in the eastern Blue Ridge (Hatcher and Butler, 1979; Quinn, 1991). In addition, this evidence, as well as the presence of sheath folds, suggests that rocks in the Inner Piedmont underwent a high degree of shear strain during D_2.

Sheath folds are common in the Columbus Promontory, probably even penetrative, but the three-

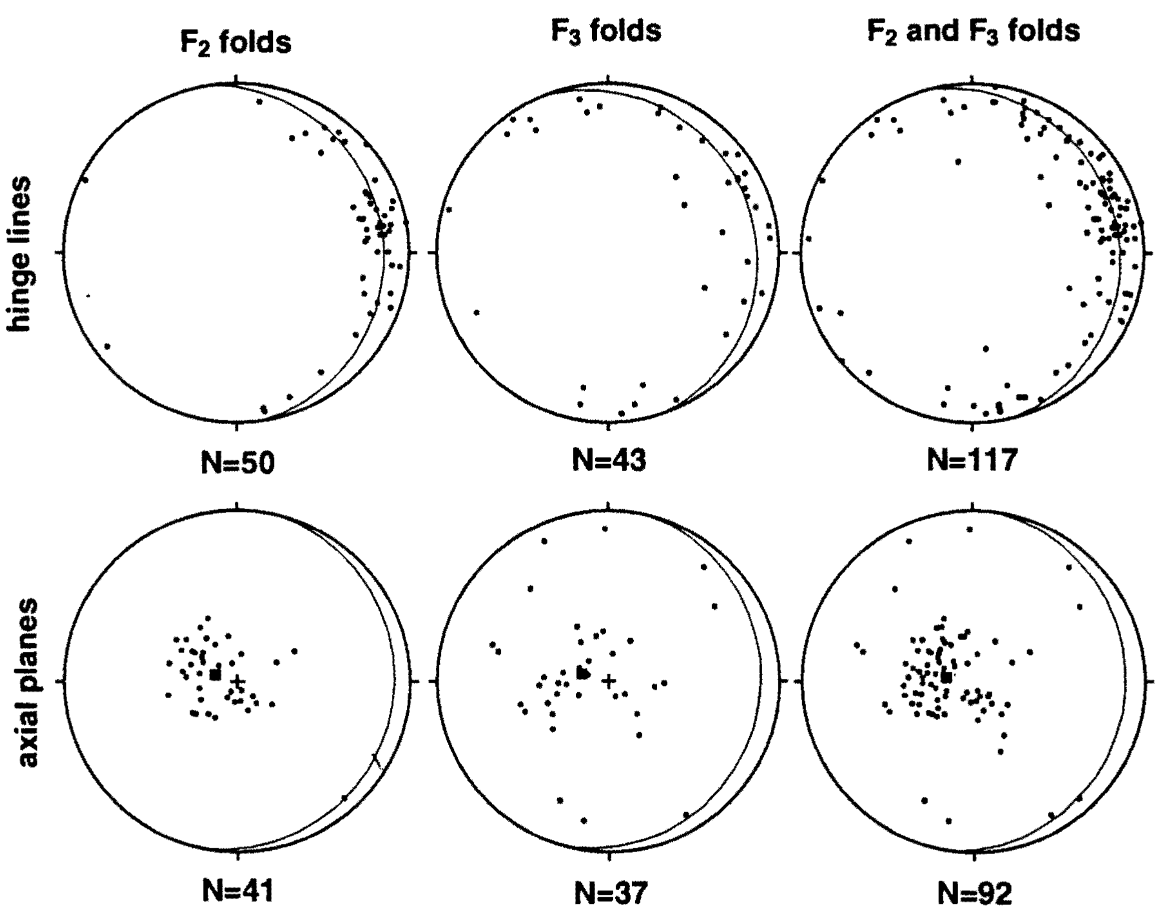

Figure 3. Equal-area plots of folds measured in the Columbus Promontory. The plots include the best-fit great circle for the hinge lines and the mean plane of poles to axial planes and its pole (indicated by the square). Stereonet analysis of the foliations indicates that the effect of F_4 and F_5 folding on the orientation of earlier folds is insignificant. Additional unpublished data is provided by T. L. Davis and Christopher C. Jayne.

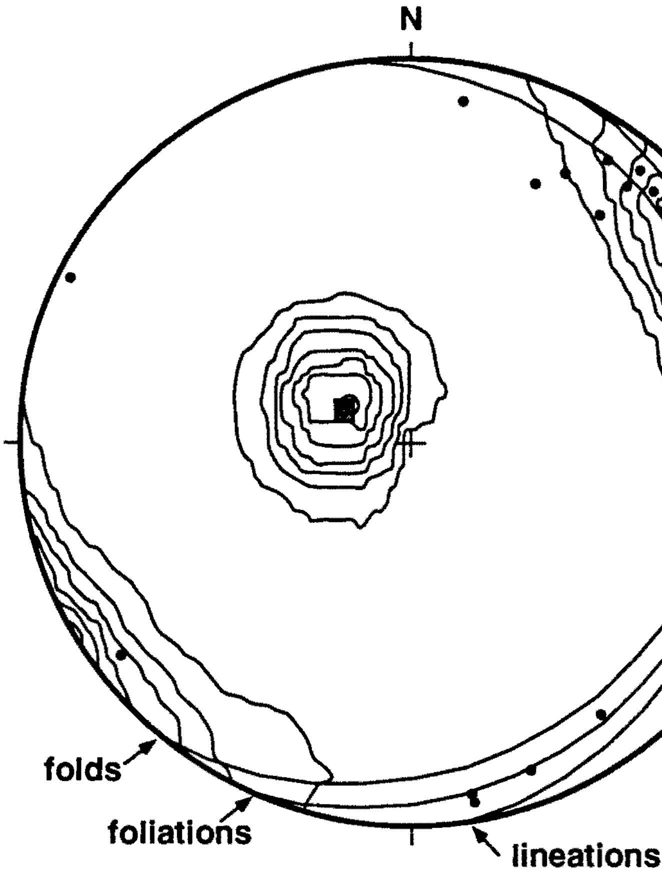

Figure 4. Equal-area projection of D_2 foliations(n=919), lineations (n=333), and fold hinges (n=50). Contours for the foliations are in the center of the plot, while contours for the lineations are located along the NE and SW rims. Contouring was performed using the Kamb method. Contour intervals are 3σ for foliations and 5σ for lineations. Folds are not contoured. Best-fit great circles for the three fabrics are shown and labeled. The great circle for foliations is based on the best-fit center of the contoured poles, whereas those for lineations and fold hinges are based on the best-fit of their points. Note the nearly parallel orientation of the three great circles, suggesting a contemporary origin for the three structures.

dimensional complexity of these folds makes assessment of their penetrative nature difficult. Evidence for their existence includes bull's-eye, anvil, and mushroom patterns and elongated folds with very small interlimb angles (Fig. 5). Bull's-eye patterns can be easily confused with type 1 fold interference patterns or dome and basin structures in upright folds created by inhomogeneities during a single folding event. Type 1 interference patterns are produced where one set of folds is superposed at right angles by another set of folds (Ramsay, 1962, 1967). These patterns are present in the Inner Piedmont but they involve very broad, upright F_4 and F_5 folds and are therefore, not likely to have produced the bull's eye patterns. Bull's eye patterns in the Columbus Promontory typically are intrafolial and elongated so that their long axes lie in the plane of the S_2 foliation (Fig. 2b). These relationships suggest a contemporary origin for the bull's-eye patterns and the foliation. Mushroom and anvil patterns can be confused with type 2 interference patterns but such patterns are very rare in the Columbus Promontory, especially among F_2 folds. Furthermore, several folds that have sections parallel and perpendicular to the mineral lineations are exposed. In sections perpendicular to the lineation, a bull's-eye pattern can be observed whereas in sections parallel to the lineation, an elongate fold with a narrow interlimb angle appears (Fig. 5).

Complex outcrop patterns in the Chauga belt suggest that sheath folds may also exist at map scale (Davis, 1993). Similar relationships were observed by Carreras and others (1977) in the Cap de Creus mylonites, Evans and White (1984) in the Moine Nappe, Meneilly and Storey (1986) in the South Orkney Islands, and Mies (1991) along the basement-Ashe boundary in the eastern Blue Ridge.

Davis (1993, this guidebook) argues that the S_2 foliation represents a mylonitic C-fabric because of its strong planarity, the presence of mineral lineations on virtually all foliation surfaces, and its association with structures indicative of noncoaxial deformation such as S-C fabrics, σ-type tails on porphyroclasts (Simpson, 1986), and intrafolial folds. Furthermore, Davis (1993) argues that most F_2 folds represent an artifact of high shear strains within S_2 and are probably the result of perturbations within the shear plane. Such folds probably formed by a process similar to that described by Cobbold and Quinquis (1980) where the perturbations in the shear plane are passively amplified to produce sheath folds. Under continued shearing, the limbs of the sheath folds will approach parallelism with the shear direction. Many F_2 folds, which are typically highly cylindrical at the outcrop scale and parallel the mineral lineation, may represent the limbs of sheath folds whose axial culminations remain unexposed. This suggests that rocks in the Columbus Promontory probably underwent very high shear strains (γ>>10) in order for the limbs of the sheath folds to approach parallelism with the mineral lineations (Cobbold and Quinquis, 1980; Skjernaa, 1989; Mies, 1993).

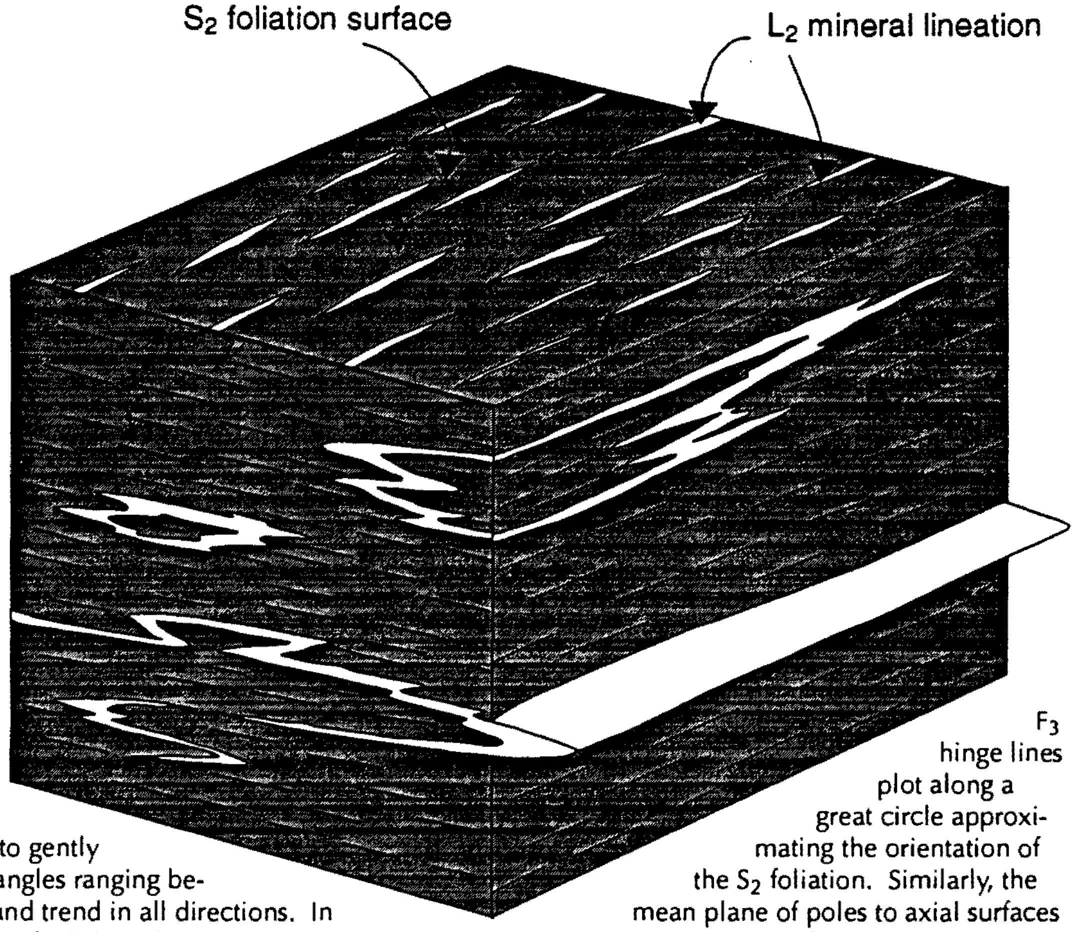

Figure 5. Block diagram showing typical F_2 morphologies. In sections cut perpendicular to the L_2 mineral lineation, bull's-eye and anvil patterns are common. In sections cut parallel to the lineation, folds with very small interlimb angles or highly cylindrical folds are typically observed.

F_3 Folds. While F_2 folds generally exhibit a narrow range of geometries and orientations, F_3 folds occur in a wide variety of geometries and orientations (Table 1). They range from upright to gently inclined, with interlimb angles ranging between 10° to near 120° and trend in all directions. In general, steeply dipping F_3 folds have large interlimb angles whereas gently dipping folds have small interlimb angles. Fold interference patterns between the tight, gently-inclined folds and the open upright folds indicate that the tight, gently inclined folds are older. Furthermore, relationships between F_3 folds and S_2 layering suggest that tight, gently inclined folds formed under highly ductile conditions whereas the open upright folds formed in relatively competent rock. These observations suggest a relationship between the relative age of the F_3 fold and its geometry (orientation of axial surface and interlimb angle)— the more gently inclined and tighter it is, the older it is. This relationship, however, does not apply to F_2 folds since virtually all of these folds have identical geometries.

The wide range of geometry and orientation displayed by F_3 folds reflects the relatively low shear strains sustained during D_3 deformation, and gradually decreasing temperatures increasing the competency of the rock layers. Folds were probably nucleating continuously throughout D_3. Under continued shearing, the geometries of these folds were modified toward smaller interlimb angles and more gently-dipping axial planes. The low shear strains involved with D_3 deformation, however, did not allow F_3 folds to take on similar forms and orientations as F_2 folds.

F_3 hinge lines plot along a great circle approximating the orientation of the S_2 foliation. Similarly, the mean plane of poles to axial surfaces also parallels the S_2 foliation (Fig. 3). It was suggested earlier that these relationships indicate a contemporary origin for F_3 folds and the S_2 foliation. Textural relationships, however, indicate that F_3 folding postdates development of S_2. Evidence includes F_3 folds deforming S_2 along with development of a new retrograde fabric (S_3) in F_3 hinge zones that overprint S_2. The parallel orientation of the fold girdle and the mean plane of poles to axial surfaces to S_2 does indicate, however, that the orientation of the strain ellipsoid did not change significantly between D_2 and D_3 deformation.

USE OF HANSEN'S METHOD TO DETERMINE SHEAR SENSE

The vergence of asymmetric folds is a common criteria used to define the shear direction, but the wide variety of orientations of both F_2 and F_3 folds in the Inner Piedmont, ranging over 360° (Griffin, 1967, 1969, 1974; Hatcher, 1969; Lemmon, 1973; Hopson and Hatcher, 1988; Liu, 1991; Davis, 1993), makes the use of fold vergence, on a fold-by-fold basis, ambiguous. In the Columbus Promontory, folds with similar orientation can verge in opposite directions, even in the same outcrop. While many of the opposite vergences are created by the fold-nappe style geometry of the F_2 folds, many other opposite verging folds have an anvil-like form. These forms, as discussed earlier, are probably indicative of sheath folding (Ramsay and Huber, 1987; Skjernaa, 1989).

Hansen (1971) demonstrated that groups of asymmetric folds formed during a single deformational event can be used to estimate the regional slip-line orientation. For example, imagine a thrust sheet dipping to the east where asymmetric folds develop within the sheet as a result of west-directed shear. Folds that develop first will trend N-S and plunge in both directions. With continued shearing, north-plunging folds will rotate counterclockwise whereas south-plunging folds will rotate clockwise (note that vergence is always described in a down-plunge view). Plotted on an equal-area net, the fold hinges plot along a great circle that parallels the orientation of the thrust fault. This great circle can be separated into two domains of clockwise- and counterclockwise-verging folds separated by a narrow gap. The gap between these two fields is called the separation arc and the slip-line direction lies within this arc (Hansen, 1971; Marshak and Mitra, 1988).

Figure 6 shows an equal area plot of 51 F_2 and F_3 folds measured in the study area. The plot shows considerable overlap of the clockwise and counterclockwise fields. The zone of overlap occurs over an arc of about 40-45° for both fold generations, defined in this study as the overlap arc. This arc is interpreted to represent the range of orientations of the separation arc within the study area and indicates that the slip-line direction varied from SW to WNW.

The overlap arc is probably related to the fact that the analysis was conducted over a large area rather than at a single locality. It is also possible some of this scatter is attributable to F_4 and F_5 folding but stereonet analysis of F_2 axial planes and S_2 foliations (Fig. 4), which show little scatter, suggests that such deformation had little effect on the orientation of F_2 and F_3 folds. Shear-sense indicators and mineral transport lineations associated with D_2 deformation, in general, indicate that transport directions ranged from SW to WNW within the study area (Fig. 1). These observations are consistent with the orientation of the overlap arc for F_2 folds. While similar D_3 shear-sense indicators are absent, the nearly identical orientation of the overlap arcs for both F_2 and F_3 folds strongly suggest that the same shearing relationships that existed during D_2 deformation also existed during D_3.

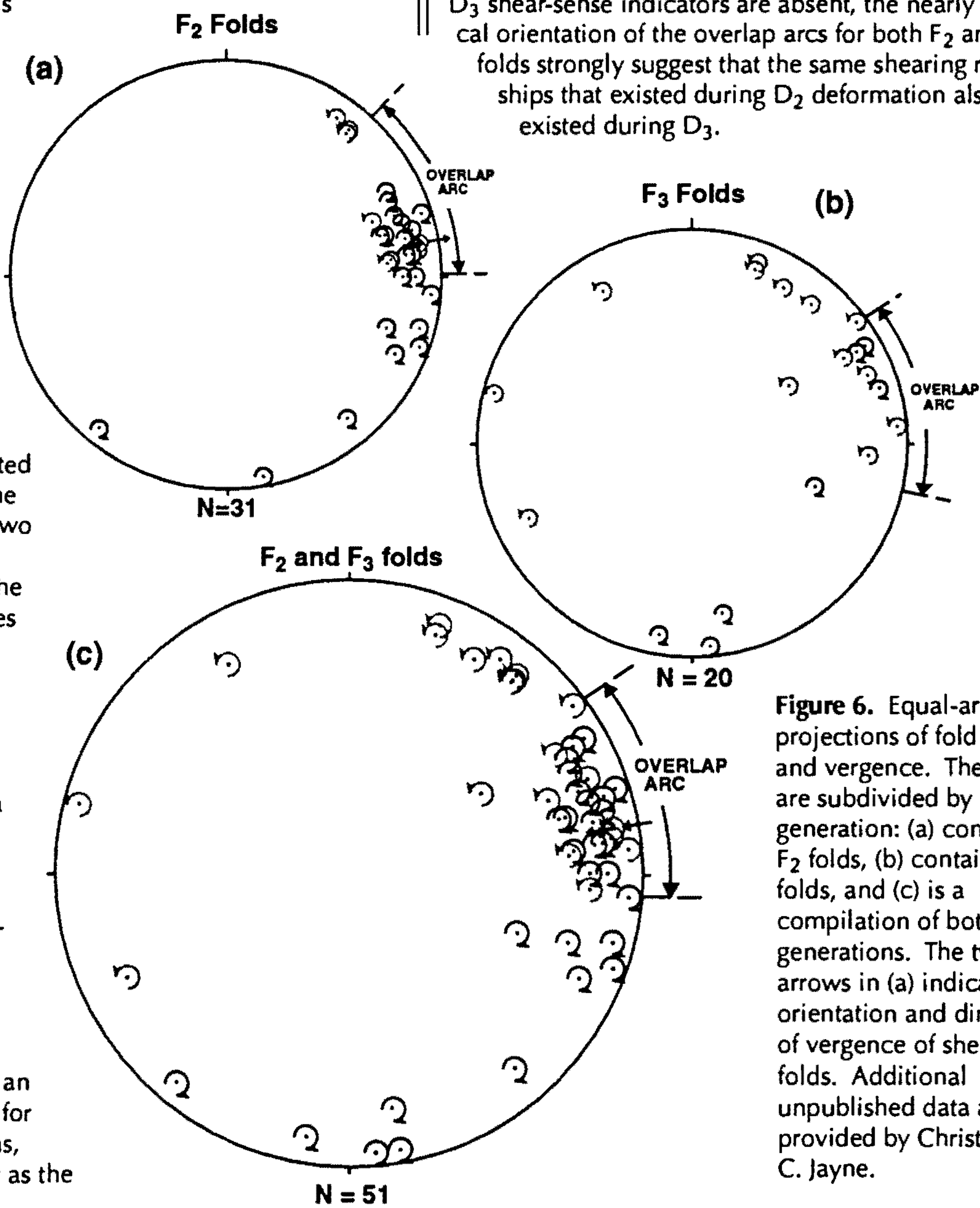

Figure 6. Equal-area projections of fold axes and vergence. The data are subdivided by fold generation: (a) contains F_2 folds, (b) contains F_3 folds, and (c) is a compilation of both generations. The two arrows in (a) indicate the orientation and direction of vergence of sheath folds. Additional unpublished data are provided by Christopher C. Jayne.

DISCUSSION

Hansen analysis of F_3 folds yields a nearly identically oriented overlap arc as F_2 folds thereby indicating W to SW-directed shearing also existed during D_3 deformation. Unlike, D_2 deformation, however, there are no other independent structures associated with D_3 that yield the sense of shear, and the results of the Hansen analysis do not yield information on the spatial distribution of shear direction— that is, the analysis cannot confirm if shearing was W-directed in areas to the SE changing gradually to SW-directed in areas to the NW as it was during D_2. This can be resolved by applying the Hansen method at individual outcrops in selected areas throughout the study area, but the widely scattered distribution of F_3 folds makes such an analysis impossible. Based on stereonet analysis of hinge lines and axial planes of F_2 and F_3 folds, which both plot along a plane approximating S_2 and results of the Hansen analysis that yield nearly identical overlap arcs, I believe that the shear direction and the spatial distribution of shearing did not change significantly between D_2 and D_3.

While others have distinguished D_2 and D_3 as separate events, they have long recognized a textural link between them (Hatcher, 1969, 1970, 1971; Griffin, 1969; 1974; Lemmon, 1973; Hopson and Hatcher, 1988; Liu, 1991; Davis, 1993). In northwestern South Carolina and northeastern Georgia, crosscutting relationships between D_3 faults and D_2 folds, as well as the relationships between D_3 structures and the S_2 foliation, strongly suggest that temperatures remained elevated subsequent to formation of D_2 structures and during D_3 deformation (Hopson and Hatcher, 1988). Furthermore, Liu (1991) concluded that some F_3 folds formed by continued progression of F_2 folding without significant time break. In the Columbus Promontory, the wide variety of F_3 fold geometries reflects gradually decreasing shear strains and decreasing temperatures in the rocks. Older F_3 folds have a more isoclinal recumbent geometry and tend to be axial planar with the S_2 foliation. Many of these folds are difficult to distinguish from F_2 folds suggesting that F_2 and F_3 represent a continuous series of folds.

Based on structural analysis and textural relationships of the folds, I conclude that D_2 and D_3 represent a single continuous episode of deformation, synchronous with peak to post-peak M_2 metamorphism, and during which the shear sense did not change. During later stages of this single episode, rocks began to cool thereby changing the rheological properties of the rocks. This change in rheology created the distinct geometric and textural styles associated with the two events.

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SUMMARY OF MID-MESOZOIC BRITTLE FAULTING IN THE INNER PIEDMONT AND NEARBY CHARLOTTE BELT OF THE CAROLINAS

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ABSTRACT

Brittle faults and cataclastic zones (trending N50°-70°E, N15°-20°E, N35°-55°W, and E-W) developed across the Inner Piedmont and Charlotte belt of the Carolinas during mid-Mesozoic continental rifting. Cataclasites and quartz microbreccias display a spectrum of textures indicating multiple episodes of brecciation, silicification, and dilational vein and extension fracture development. Left normal-oblique movements dominated the N50°-70°E fault system, producing the Marietta-Tryon graben (725 km^2) in upstate South Carolina; younger (?), right normal-oblique movements affected ENE-trending faults. Brittle faults developed by shearing along a widespread, steep, systematic joint set at N60°E, accompanied by oblique (~S40°E) crustal extension. Regionally, the emplacement of NW-trending diabase dikes overlapped displacements along the NE-trending cataclastic zones. Paleozoic ductile faults, such as the Lowndesville and Kings Mountain shear zones, experienced Mesozoic reactivation locally.

Orientations of syntaxial veins and extension fractures indicate two stages of regional mid-Mesozoic crustal dilation: initially SE directed, but later oriented NE-SW and post-dating a regional N-trending diabase dike set. Regional σ_3 stress trajectories for the two stages display consistently oriented or broadly arcuate map patterns for the minimum principal stress. The shift in extension direction (from S40°E to S30°W), involving uplift, arching, brittle fault reactivation (right normal-oblique motions), and northeastward tilting across the Carolinas, developed subsequent to the regional NW-SE crustal rifting which obliquely opened the Marietta-Tryon graben.

REGIONAL SETTING

Brittle deformation at upper crustal levels accompanied mid-Mesozoic rifting of the southern Appalachian continental margin 40 to 50 million years (Olsen and others, 1989) after Alleghanian collisional orogenesis. Involved were multiple episodes of silicification, left- and right-oblique slip along normal faults, and late kinematic dilational vein filling. Steep (>75° dip), brittle Mesozoic faults lined with quartz microbreccia and cataclasite are widespread in Georgia and the Carolinas from the Blue Ridge (e.g., the Warwoman lineament of Hatcher, 1974) across the Inner Piedmont to the Charlotte belt (e.g., the Cold Spring, Watts, and Enochville cataclastic zones of, respectively, Griffin, 1979; Snipes and others, 1983; and Preddy, 1991). Cataclastic zones, from <1 m up to several hundred m wide (the New Liberty Church roadcut through the Pax Mountain fault zone), trend N50°-70°E, less typically N15°-20°E, N35°-55°W, and E-W.

Brittle faults and associated cataclastic rocks transect moderate to high grade metamorphic units in the Carolina counties represented by Figure 1 (the stippled area of the index map). Our study area lies between Westminster, SC and the Sauratown Mountains area, NC, a strip 350 km long and 20-100 km wide.

Along the southeastern margin of the Inner Piedmont, some regional Mesozoic brittle faults are spatially related to earlier ductile faults. Locally, cataclasis affecting Paleozoic mylonitic rocks, for example along the Lowndesville shear zone (LSZ) (Griffin, 1979; Nelson 1981), may have developed as early as the waning thermal stages of Alleghanian ductile deformation. Orientations of slickenlines and normals to extension veins and fractures suggest mid-Mesozoic shearing and dilation also affected portions of the LSZ (Garihan and others, 1990) and the Kings Mountain shear zone (KMSZ) (see below). Numerous cataclastic zones of the Sispring fault system (Horkowitz, 1984; Niewendorp, 1992) follow the LSZ mylonite and phyllonite northeastward along the Inner Piedmont-Charlotte belt terrane boundary to the vicinity of Clinton, SC, where the Cross Anchor fault-central Piedmont suture (Dennis, 1988, 1991) swings north to join or offset (Maybin, oral commun., as an Alleghanian thrust?) the KMSZ. At that point the Sispring brittle faults continue 30 km northeast into the Charlotte belt. NW-trending Jurassic(?) diabase dikes cut the Sispring cataclastic zones, and are therefore younger; in northern

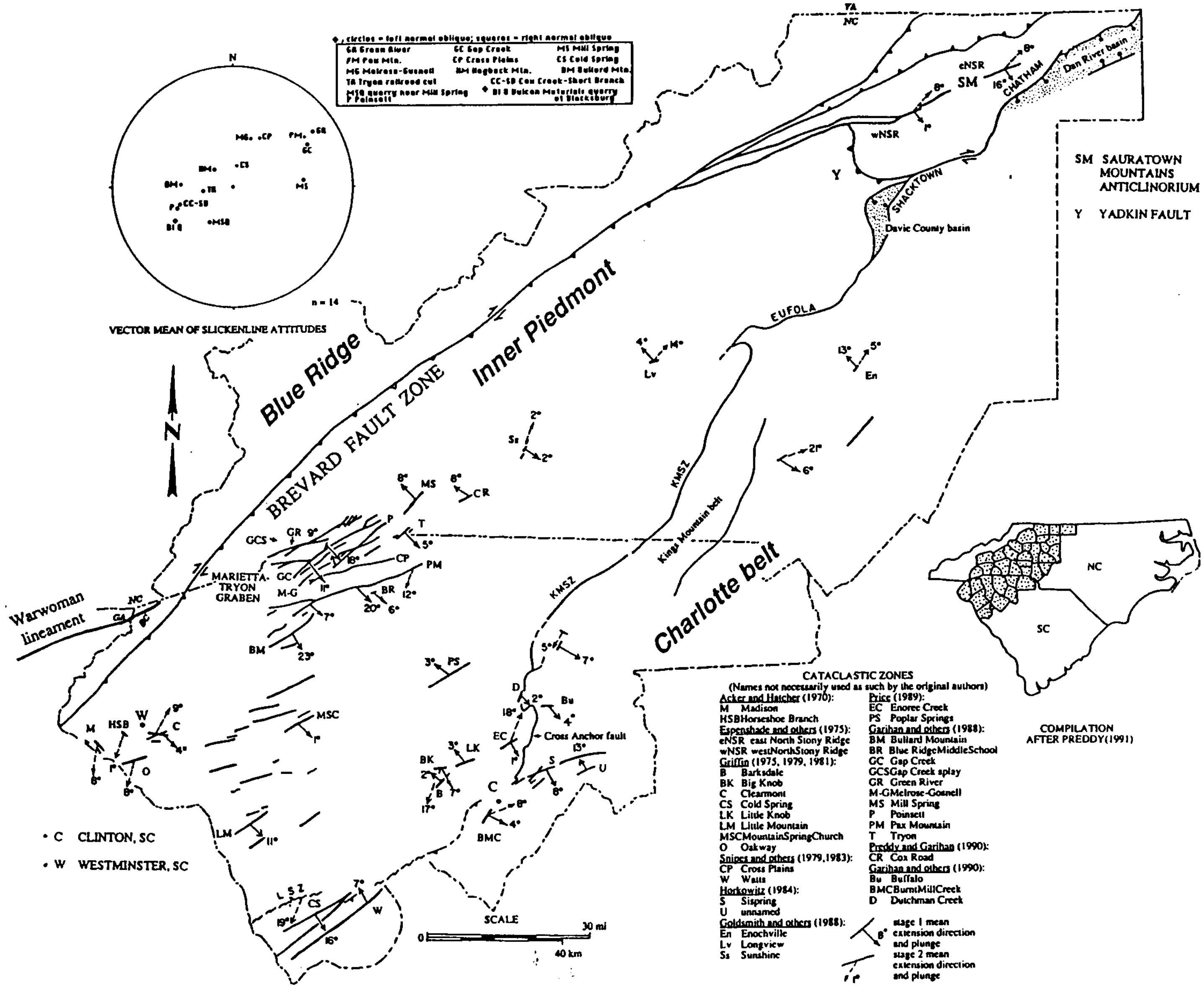


Figure 1. Compilation of cataclastic zones in the Inner Piedmont and Charlotte belt of the Carolinas. For clarity, the Forbush and Danbury faults in the Sauratown Mountains anticlinorium area are not shown.

Greenville County 75 km away, a dike displays offset with left lateral separation across two cataclastic zones. By implication, the NW-trending diabases and the NE-trending cataclastic zones regionally overlap in age.

The Eufola fault (Fig. 1) (Milton, 1981; Goldsmith and others, 1988), truncating the northern segment of the KMSZ, has a cataclastic character. Northeastward, siliceous microbreccia of the Shacktown fault (Heyn, 1988) overprints mylonitic fabric. Near the NC-VA state line, movements along the Chatham fault on the northwest margin of the Dan River basin involve rejuvenation of a late Paleozoic mylonite zone (Horton and McConnell, 1991). Mesoscopic brittle deformation zones (0.5-1 km wide, cut by diabase) were caused by Mesozoic reactivation of the Hylas mylonized thrust in eastern Virginia (Venkatakrishnan and Watkins, 1992). Characteristics of brittle over ductile deformation and retrograde metamorphism occur southwest of the junction of the Yadkin fault and the North Stony Ridge fault, where superimposed cataclastic effects are developed along the fault contact (Lewis, 1980). The structural heredity of Mesozoic fault positioning, if not strictly including attitude, from Paleozoic thrusts is probably widespread in the Virginia Piedmont (Glover and others, 1980) and the central Virginia Blue Ridge, where thrust ramps were properly oriented in the Mesozoic stress field to accommodate extension ("backslip" of Bartholomew and others, 1991). Elsewhere, development of Mesozoic faults probably involved utilization of pre-existing regional joint sets (Garihan and Ranson, 1992).

Rb-Sr isotopic analysis of multiply brecciated and silicified feldspathic rocks along the North Stony Ridge fault provides data for an isochron yielding a Middle Jurassic age of 180 +/- 3 Ma (Fullagar and Butler, 1980; Fullagar, 1992). Based on this single age, it has been tacitly assumed by all workers that widespread occurrences of similar cataclastic rocks are also Mesozoic in age.

The significance of "ultramylonite" zones in North Carolina postulated to be Triassic in age was noted initially by Conley and Drummond (1965). In the Knoxville 1° x 2° quadrangle southeast of the Brevard zone, Hadley and Nelson (1971) mapped zones of silicified breccia in the Piedmont (20-60 km long and 4-8 km apart), corresponding to parts of the Pax Mountain, Gap Creek, Poinsett, and Cross Plains faults (Garihan and others, 1988) (Fig. 2). Birkhead (1973) catalogued "flinty crush rock" exposures across the SC Piedmont, providing useful locality information.

MARIETTA-TRYON FAULT SYSTEM

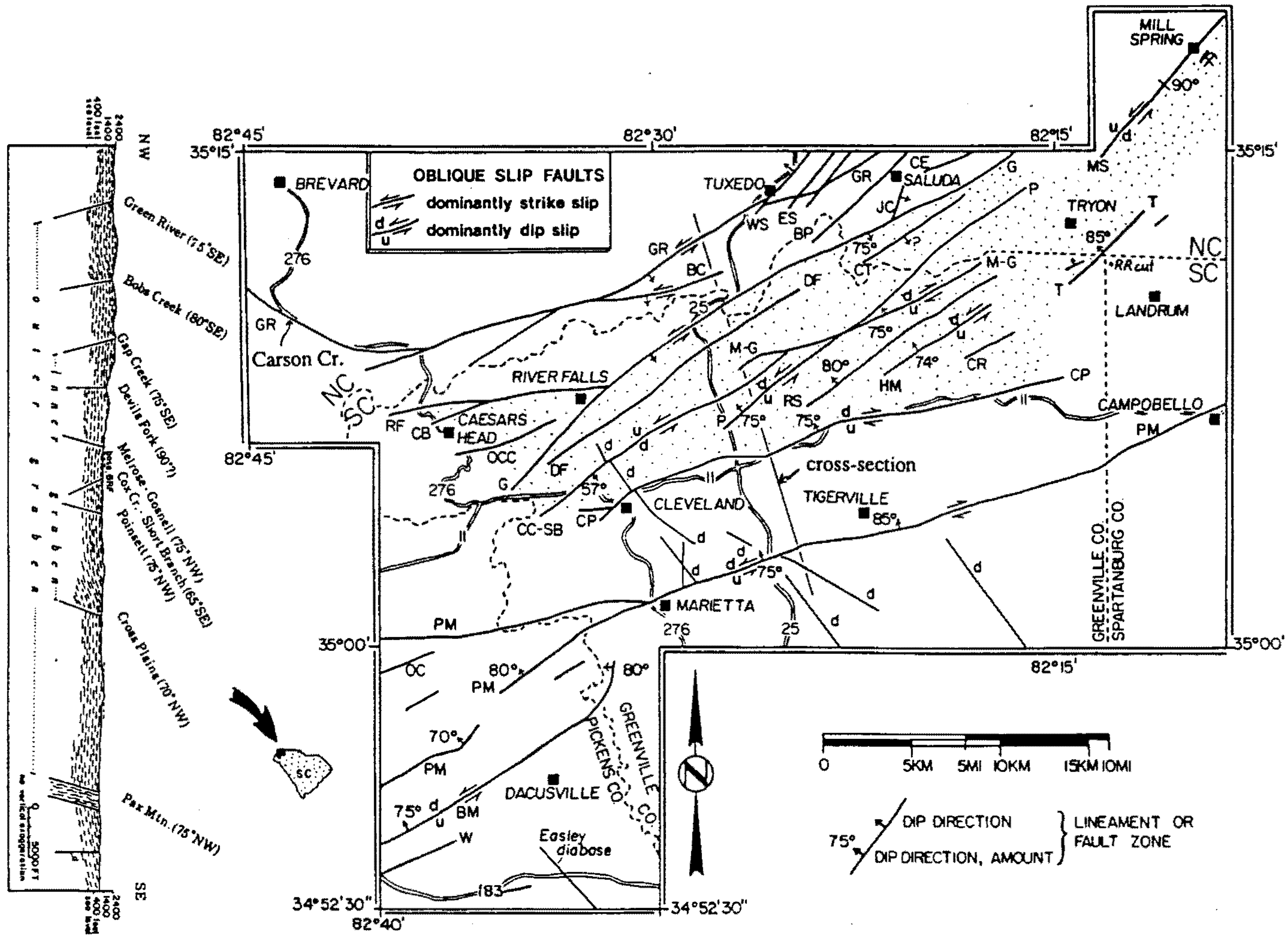
Fourteen major faults and at least seven associated splays (trending N50°-70°E) plus north-northeast cross faults comprise the Marietta-Tryon fault system in northwest South Carolina, a relatively narrow graben of at least 725 km^2 bound by the Green River and Pax Mountain faults. A westward-tapering, highly faulted inner graben of about 235 km^2 lies between the Gap Creek and Cross Plains faults (Fig. 2, cross section). Beneath a reservoir, the NE-trending Cox Creek-Short Branch fault appears to terminate against the ENE-trending, younger(?) Melrose-Gosnell fault. Approximately 30-100 m and 125 m of left strike separation of a single diabase dike occur across the Cox Creek-Short Branch and the Cross Plains (Snipes and others, 1979) faults, respectively. In the immediate vicinity of the Cross Plains fault, hand samples of the displaced diabase display fractures due to intense shearing (D. Snipes, oral commun.).

Topographically the brittle faults are expressed as: 1) prominent linear or arcuate valleys (up to 500 m relief, viz. across the Gap Creek fault), with the faults coincident with portions of major drainages; 2) aligned linear segments of lower-order stream tributaries; 3) very prominent to very subtle NE-trending ridges; 4) aligned headwater tributaries, saddles, and ridges; 5) prominent linear, NE-trending slope breaks; 6) linear shorelines (Lake Summit, North Saluda reservoir); and 7) aligned springs.

Our regional structural studies (Garihan and others, 1988, 1990) indicate left normal-oblique motions dominated the major NE-trending faults of the Marietta-Tryon system. In contrast, east-northeast faults, such as the Melrose-Gosnell and Cross Plains faults, and part of the Pax Mountain fault, display a distribution of slickenlines on fracture surfaces as a result of predominantly right normal-oblique movements. The vector means for slickenline attitudes from various fault, railroad cut, and quarry localities is shown in Figure 1, with different symbols indicating primarily left- or right-oblique slip. Dominant strike-slip or dip-slip components of Mesozoic oblique slip for the faults are differentiated in Figure 2.

Episodes of recurrent motion along brittle faults are indicated by multiple shearing, silicification, and extension veining-fracturing textures of many cataclastic rocks. Hallman and others (1987) have identified and described the spectrum of brittle textures associated with increasing degrees of cataclasis along many faults of the Marietta-Tryon graben. Photomicrographs of these textures are found in Garihan and Ranson (1992).

A kinematic scheme similar to the one proposed here is found north of the study area. In the Waynesboro-Charlottesville, Virginia area southward to



the Danville basin, patterns of *en echelon* diabase dikes in a NNW-trending swarm have been used to infer sinistral shear parallel to NE-trending brittle faults (like the Chatham Hill-North Stony Ridge fault zones) and dextral shear parallel to NW-trending faults (Bartholomew and others, 1991). Butler and Dunn (1968) postulated 370 m of right strike separation along the E-trending North Stony Ridge fault. Lewis (1980, Figure 3) has suggested some 11-12 km of left strike separation of the Crossnore gneiss across the North Stony Ridge fault at its western end. More recent detailed mapping, however, indicates minimal to zero displacement across the Stony Ridge brittle fault zone (Hatcher, 1988).

Hooper (1989) has recognized in the central Georgia Piedmont an extensive NNE-trending sinistral shear system, called the Piedmont brittle fault system. The faults are mostly strike slip, but locally oblique slip and normal slip. Mesozoic (?) sinistral offsets in Grenville basement stratigraphy and mylonites of the Towaliga fault zone are mapped in the vicinity of the Pine Mountain window. The Georgia brittle faults, which developed after diabase dike intrusion and which post-date Paleozoic ductile faults, display prominent trends of N30°E (sinistral), N60°W (sinistral), and N70°E (antithetic in character to the regional kinematic scheme). Rhombic-shaped pods of siliceous cataclasite are formed locally as left-steping tensional bridges along the NNE-trending Barnes Mountain fault system of the Pine Mountain window (Hooper and Hatcher, 1989). Similarly, a minor N15°E-trending cataclastic zone (the right normal oblique Joels Creek fault) occurs in a transtensional zone of overlap between two regional left normal oblique brittle faults of the Marietta-Tryon graben south of Saluda, NC (Fig. 2) (Garihan and Ranson, 1992).

REGIONAL SYSTEMATIC JOINTS

In northwesternmost South Carolina, Acker and Hatcher (1970) recognized rectangular stream drainage patterns produced by strong N40°-

Figure 2. Mesozoic brittle faults and lineaments of the Marietta-Tryon fault system, with cross-section. Stippled region is the inner graben. Abbreviations for the faults and lineaments are:

BP	Blakes Peak lineament
BC	Bobs Creek fault
BM	Bullard Mountain
CR	Chestnut Ridge lineament
CB	Coldspring Branch fault
CT	Colt Creek fault
CE	Cove Creek fault
CC-SB	Cox Creek-Short Branch
CP	Cross Plains fault
d	diabase
DF	Devils Fork fault
ES	East Lake Summit fault
G	Gap Creek fault
GR	Green River fault
HM	Hogback Mountain fault
JC	Joels Creek fault
M-G	Melrose-Gosnell fault
MS	Mill Spring fault
OCC	Oil Camp Creek lineament
OC	Oolenoy Church fault
PM	Pax Mountain fault
P	Poinsett fault
RF	River Falls fault
RS	Rocky Spur lineament
T	Tryon fault
WS	West Lake Summit fault
W	Wolf Creek fault

50°W and N40°-50°E joint control on headward erosion and stream piracy. Reactivation of that joint trend is suggested by several cataclastic zones of that orientation in the Tugaloo Lake and Holly Springs quadrangles. In the Holly Springs quadrangle, a set of approximately N-S cataclastic zones is also recognized (R. Hatcher, personal commun.), although this trend is not common elsewhere in our study area. N30°-80°W, N30°-70°E, and E-W-trending joints are common in the Piedmont. The extreme western segment of the Green River fault along Carson Creek south of Brevard, NC is probably controlled by the same NW-trending joint set (Figure 2). In the immediate vicinity of the Marietta-Tryon graben, excluding sheeting joints, N50°-70°E, N80°E-N80°W, and N50°-70°W oblique joint sets are present (n=1683 joints) (Preddy and others, 1987). A N15°E joint trend near Saluda, NC controlled the development of the N15°E Joels Creek fault in the zone of left overlap and transtension between the Gap Creek and Cove Creek faults. We interpret the consistent regional N50°-70°E trend of brittle faults to have developed by left-oblique shearing along the most widespread, steep joint set at N60°E, accompanying oblique (~S40°E) mid-Mesozoic crustal extension.

BRITTLE FAULT FEATURES IN THE TRYON - MILL SPRING AREA, NC

Unusually good bedrock exposures displaying slickenlines related to Mesozoic brittle faulting are found in a railroad cut south of Tryon, NC and a stone quarry near Mill Spring, Polk County, NC (locations, Figs. 2 and 3). The trends of faults in these exposures mimic the trends of regional joints. At the railroad cut, 14 minor faults generally strike northeasterly or easterly, dipping steeply northward (65°-80°); slickenlines on these surfaces plunge moderately(24°-71°). In general, the northeast set of faults in the railroad cut displays dip slip to left-oblique slip.

A limited number of slickenlines from NE-trending fracture surfaces developed in and adjacent to the Mill Spring cataclastic zone (MS in Figure 2) indicates left normal-oblique motions, ranging from nearly dip slip to essentially strike slip. The Green River, Bobs Creek, and Gap Creek faults of the Marietta-Tryon fault system are other faults that display principally strike-slip movement. The Mill Spring fault-cataclastic zone is linear and SE-dipping to vertical.

At the Mill Spring quarry, in hanging-wall rocks 1 km southeast of the Mill Spring fault, a strongly developed, E-trending fault set (n=7) with steep northerly dips (75°-90°) displays moderately plunging slickenlines consistent with left-oblique slip. This easterly set of quarry normal faults is antithetic to faults of the main Mill Spring zone. A NE-striking set of minor faults in the quarry is more poorly developed and scattered in orientation than the easterly set. A summary of vector means of slickenline attitudes for the Mill Spring fault, quarry, and Tryon railroad localities is shown in the inset of Figure 1.

BRITTLE FAULT DATA FROM VULCAN MATERIALS QUARRY, BLACKSBURG, SC

The Gaffney Marble of the Blacksburg Formation (Late Proterozoic?) crops out along a 25 km length of the Kings Mountain belt within a kilometer or so of the N65°E-trending KMSZ (Horton, 1981). Our study of 26 NE-trending faults of probable Mesozoic age in the Gaffney Marble exposed in the Blacksburg quarry (operated by Vulcan Materials Company) demonstrates they are unquestionably normal and dominated by left-oblique slip. Two populations are present: 1) a set (n=6) averaging N41°E, 83°SE; vector mean for slickenlines = 62°/N54°E; and 2) a set

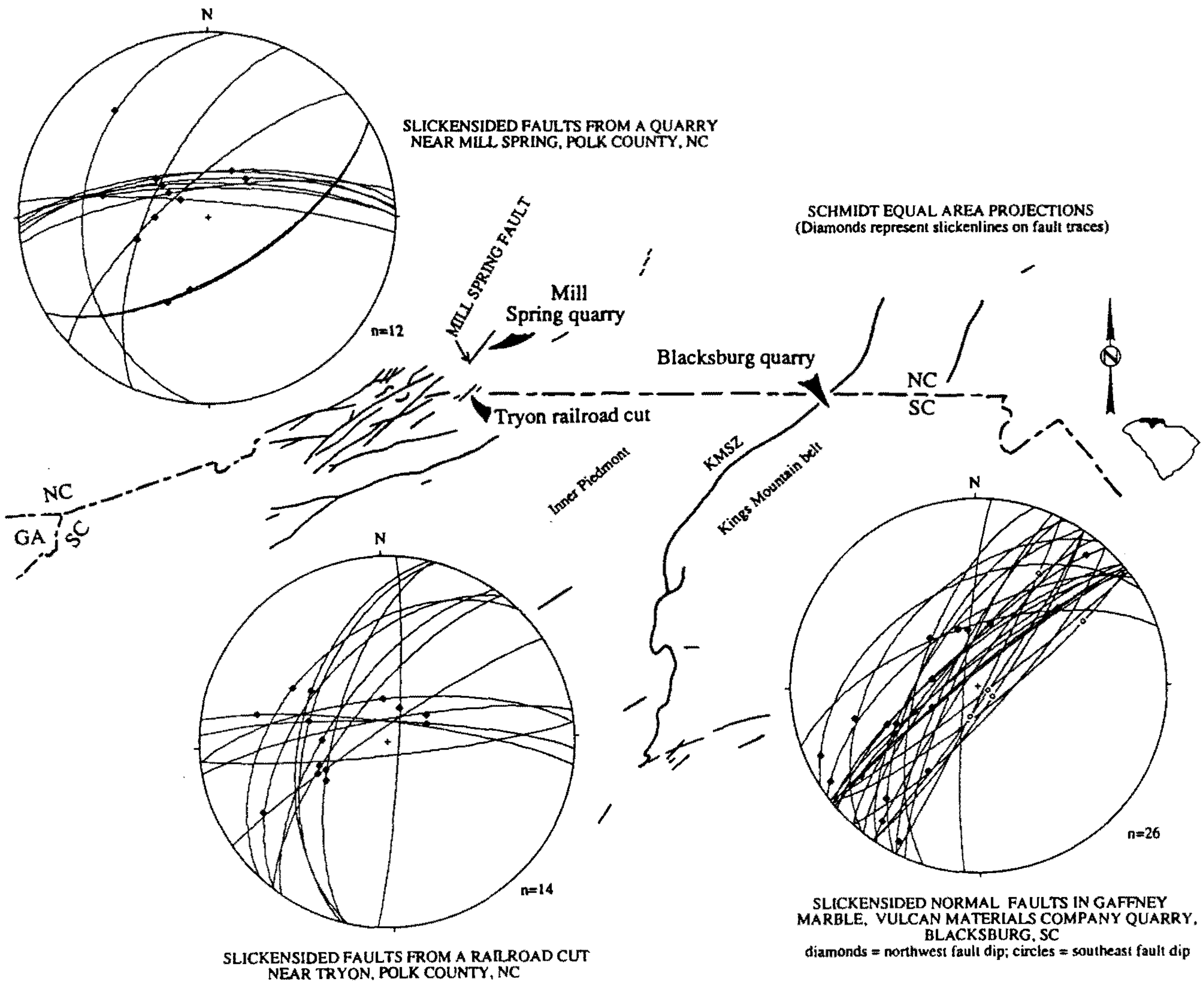


Figure 3. Brittle fault features in the Mill Spring, NC - Blacksburg, SC area.

(n=20) averaging N42°E, 73°NW; vector mean for slickenlines = 43°/N60°W (Fig. 3). Just three normal faults are dominantly right-normal oblique. We interpret all minor faults to be part of a population of regional Mesozoic brittle faults, based chiefly on their similarity in mean slip direction to vector means of thirteen regional Mesozoic faults from across the entire length of the Inner Piedmont of the Carolinas (Figs. 1 and 3). Moreover, we infer the cause to be Mesozoic reactivation along the KMSZ. Heck (1989) suggested the KMSZ was at least the northwestern limit of Mesozoic crustal extensional reactivation affecting Paleozoic thrusts of the type visible on COCORP seismic reflection profiles.

As an aside, we offer the suggestion that, in the Blacksburg quarry, the mineral stretching lineation on foliation surfaces (= relict sedimentary bedding) oriented 25°-30°/N50°-55°E (Horton, 1981) and normals to carbonate extension veins we measured (n= 6) oriented 2°/N47°E may be the result of Alleghanian dextral strike-slip motion along the adjacent KMSZ.

LATE KINEMATIC EXTENSIONAL FEATURES

Planar to lenticular to anastomosing, open to healed fractures (generally <1 cm wide) lined with drusy "comb" quartz crystals and syntaxial extension veins of similar size filled with anhedral quartz are present in most good exposures of quartz microbreccia and cataclasite. Typically these widespread features crosscut most shear fabrics in the cataclastic rock along the brittle faults. Hence their timing is interpreted as late kinematic. Displaced features of the metamorphic fabric of the country rock, as well as offset breccia-clast margins and local quartz veins in the cataclastic rock, indicate dilation has occurred normal to vein walls, with virtually no shear component. We interpret the normals to these features to be the extension direction parallel to the minimum principal stress that produced them.

Locally a complex dilational fracture-vein chronology exists. Nonetheless, in the majority of exposures where both are present, we recognize an earlier set of dilational features of northeast trend, indicating mid-Mesozoic southeast extension (stage 1), and a later crosscutting NW-trending set, consistent with NE-SW extension (stage 2) (Garihan and others, 1993). We emphasize that we are not referring to the trends of the cataclastic zones, but rather to orientations of sets of specific features within the cataclastic zones. Indeed, any NW or NE cataclastic zone could display either or both NW- and NE-oriented dilational features.

In Figure 1, the mean extension direction for all field data from each of 37 brittle fault zones is plotted at a mid-fault position. Solid NW and dashed NE arrows, respectively older (stage 1) and younger (stage 2), represent the azimuth and amount of plunge of the mean normal of dilational features for each fault. At the stage in our investigation where we had studied 24 faults in an eight-county area of upstate South Carolina (Garihan and others, 1990, Figure 7), the mean extension orientation of 24 mean extension orientations (i.e., 24 faults, each *fault* weighted equally) was 5°/S38°E; the mean extension orientation of normals to 383 dilational features (each *feature* weighted equally) was 10/S37°E. Interestingly, roughly doubling the data (see below) has modified the results only a few degrees. Taken alone, extension fracture and vein data from five Charlotte belt faults yielded a substantially similar result: a 2°/S34°E minimum principal stress direction (σ_3).

The shallow plunges of mean extension directions allow the construction of minimum principal stress trajectories across the study area which are not greatly distorted when projected onto map view (Fig. 4). The minimum stress trajectories spatially are drawn parallel to the mean normal orientations, and the spacings are arbitrary. Stage 1 southeast extension (solid lines in the diagram) is exceedingly consistent across the 350 km length of the Carolinas, which we consider remarkable for such apparently minor features. There is little deflection of the pattern of trajectories across the Inner Piedmont and the LSZ, KMSZ, and Eufola fault at its southeast margin. The vector mean of 696 stage 1 extensional features from 31 cataclastic zones (each *fault* weighted equally) is 4°/S40°E. A relatively homogeneous regional stress field is indicated for stage 1.

The vector mean of stage 2 extensional features from 15 cataclastic zones between Westminster, SC and the Sauratown Mountains area, NC is 1°/N30°E. Data for stage 2 extension come from NW-, NE-, and ENE-trending cataclastic zones. In Figure 4, σ_3 trajectories in map pattern (dashed lines) are broadly curved across the region, plunging gently S20°-30°W near the SC-Georgia state line to plunging gently N40°-45°E near the NC-Virginia state line. The σ_3 stress trajectories therefore arc in map view (about 25° change in strike) and broadly arch in cross section (about 16° by reversing the plunge direction).

DISCUSSION

The S40°E regional pull-apart direction for the Mesozoic southern Appalachian crust compares favorably with a S40°-50°E opening direction for the Newark basin (Ratcliffe and others, 1986), a S50°E direction for the Hartford basin (Olsen and others, 1989), and the S55°E orientation of the Blake Spur and Carolina fracture zones in the Atlantic. The acute angle (~80°) between the N60°E cataclastic zones inherited from steep regional systematic joints and the S40°E extension direction may be responsible for recurrent left normal-oblique movements on many regional brittle faults, including those of the Marietta-Tryon graben. Regional geologic relationships suggest the intrusion of NW-trending Jurassic (?) diabases regionally overlapped with slip along the NE-trending cataclastic zones.

Diagrams constructed to show the regional minimum principal stress patterns for two stages of mid-Mesozoic rifting indicate a significant 70° clockwise shift of the regional stress field, from S40°E to S30°W. One result may have been recurrent right normal-oblique motions on favorably oriented ENE-striking brittle faults.

A compilation of Mesozoic diabases by Ragland and others (1983) and Ragland (1991) shows two swarms in the Carolinas: NW- and N-trending dikes. In the central Piedmont of South Carolina the stage 2 extension directions lie some 15° to the normals to NW-trending diabase dikes in map view, and contemporaneity seems unlikely. We suggest the stage 2 extensional features in fact post-date NW dike emplacement. The orientations of stage 2 extensional fractures and veins we have described appear to bear a more systematic spatial relationship to the entire group of diabases. Stage 2 arcing trajectories tend to cross dikes everywhere at approximately 65°. One possible explanation is that the later dilational features (stage 2) developed in a stress regime influenced by preexisting dikes and divergent or "radial" crustal weakness directions.

We infer that the shift in extension direction between the two dilational stages, from S40°E to S30°W, involving uplift, arching, brittle fault reactivation (right normal-oblique motions), and northeastward tilting across the Carolinas, developed subsequent to the regional NW-SE crustal rifting which obliquely opened the Marietta-Tryon graben.

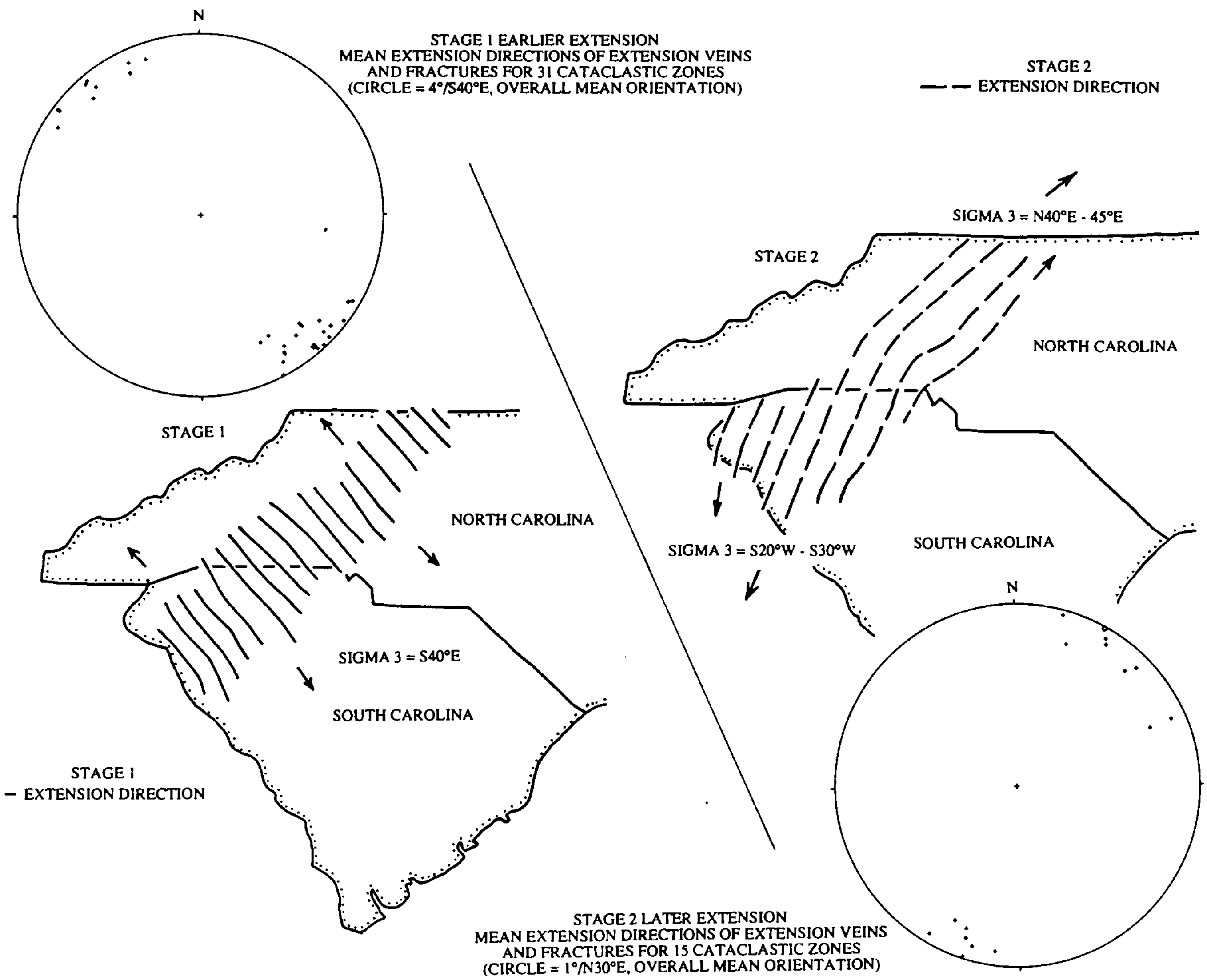


Figure 4. Minimum principal stress trajectories for two stages of Mesozoic extension in the Carolinas

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QUATERNARY GEOLOGY AND GEOMORPHOLOGY OF PART OF THE INNER PIEDMONT OF THE SOUTHERN APPALACHIANS IN THE COLUMBUS PROMONTORY UPLAND AREA, SOUTHWESTERN NORTH CAROLINA AND NORTHWESTERN SOUTH CAROLINA

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ABSTRACT

The Columbus Promontory Upland, a newly defined subsection of the Inner Piedmont block geomorphic section, is a relatively high-elevation, high-relief, terrain underlain by rocks of parts of the Chauga belt and Inner Piedmont terranes in North and South Carolina. During the Cenozoic Era, this area is interpreted to have experienced major changes in tectonism and climate that have profoundly affected the development of landscapes and geomorphic materials. Cenozoic tectonic activity appears to have been mainly epeirogenic; the present-day stress field has persisted from Late Cretaceous to the present, based upon both the sedimentary record in the Atlantic Coastal Plain and in offshore areas, and geomorphic topographic and drainage criteria that suggest subregional upwarping and/or upfaulting. Many of the landform, landscape, and drainage elements and their boundaries are known to be spatially related to characteristics and variations in the underlying bedrock lithologies and structures, thus demonstrating probable cause-and-effect structural geomorphic relationships. The effects of climatic change are most apparent for events that probably occurred during the last 2.4 Ma. These changes produced major episodes of weathering and erosion of rock and regolith, transportation of these earth materials, deposition of colluvial and alluvial sediments, and development of weathering and soil properties seen today.

INTRODUCTION

The Inner Piedmont block geomorphic section (Figures 1, 2; Tables 1, 2) occupies a critically important position in the Appalachian Highlands with respect to a number of major geomorphic landform and drainage features, materials, and process-oriented problems. Among the landscape features are the Blue Ridge Escarpment, the mountain masses, spurs, and isolated mountains (monadnocks or inselbergs) outboard from the Escarpment, and relict upland levels. Drainage developments include effects of bedrock structure and fracture on stream patterns, history of stream piracy, origins and ages of older and younger alluvium, and the evolution of the Atlantic Ocean-Gulf of Mexico drainage divide. Geomorphic materials include saprolite, high-level gravels, and residual, alluvial and colluvial soil parent materials, and the soils developed on them. Process-oriented problems include the nature and timing of epeirogenic movements (for example, uplift), rates of saprolite production and removal, and the effects of Quaternary climatic change on geomorphic materials and landforms. There are many localities in and near the Columbus Promontory Upland where field relationships bearing on these topics can be seen and discussed, and where research on these problems could be initiated.

REGIONAL GEOMORPHOLOGY

Early overall treatises on, and maps of, the regional geomorphology of the southern Appalachians include works by Hayes and Campbell (1894), Fenneman (1928, 1938), and Fenneman and Johnson (1946). Subsequent works that include the field trip region include the land-surface-form map by Hammond (1963) and the report by Redington (1978). But, other than to identify, describe, analyze, and name several new sectional subdivisions, primarily in the Appalachian Plateaus province, little basic research or synthesis has been done with regional geomorphology in the Appalachians since Thornbury (1965).

Horton and Zullo (1991) discussed the history of geologic belt terminology used in the southern Appalachians. Within the region that includes the 1993 Carolina Geological Society field trip area, King (1955, p. 337-338) subdivided rocks seen during a geologic transect across Tennessee, North Carolina, and South Carolina into persistent, narrow, long linear "geologic belts" (terranes) on the basis of similarities in bedrock lithology and structure, and compared them with the physiographic provinces (terrains). He noted both a general spatial correspondence of the major tectonic units with the classical physiographic provincial boundaries and that there are deviations in detail. Such

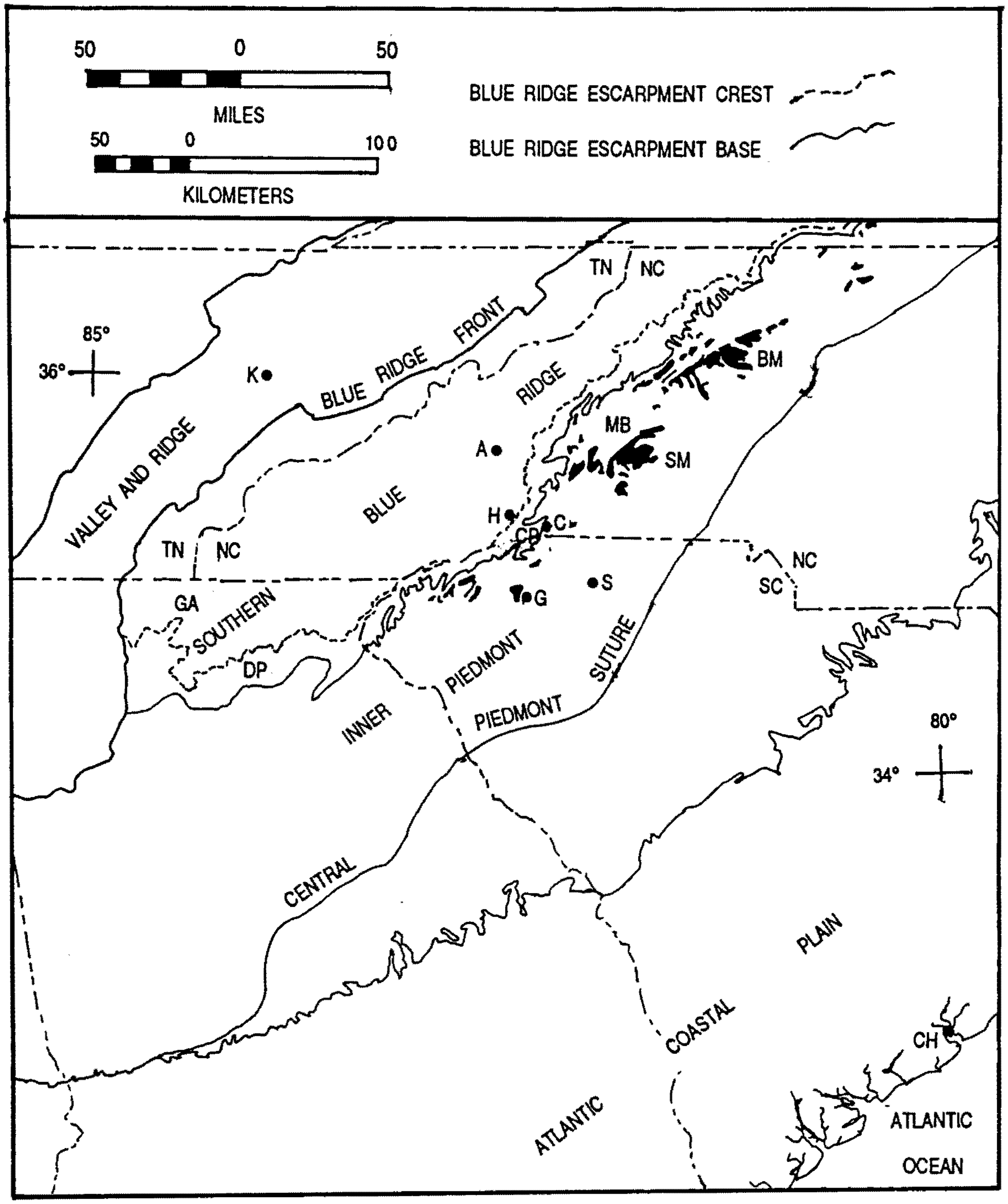

Figure 1. Index map for parts of the southern Appalachians and Atlantic Coastal Plain, showing locations of major features noted in text: A = Asheville, C = Columbus, CH = Charleston, G = Greenville, H = Hendersonville, K = Knoxville, S = Spartanburg; CP = Columbus Promontory, MB = Morganton Basin, DP = Dahlonega Plateau. Black areas denote isolated mountain masses near the Blue Ridge Escarpment in North and South Carolina: BM = Brushy Mountains, SM = South Mountains (*cf.* Figure 2). Data from Hatcher and others (1990), Kesel (1974), and LaForge and others (1925).

relationships are scale dependent (Stein and Linse, 1993). For example, on small-scale maps at the major divisional level, coincidence of landform subdivisions with tectonic units would seem great; at progressively finer subdivisions more and more differences would emerge. Godfrey and Cleaves (1991) focused on the scale aspects of landscape subdivisions, and proposed a hierarchical subdivision of landscapes according to

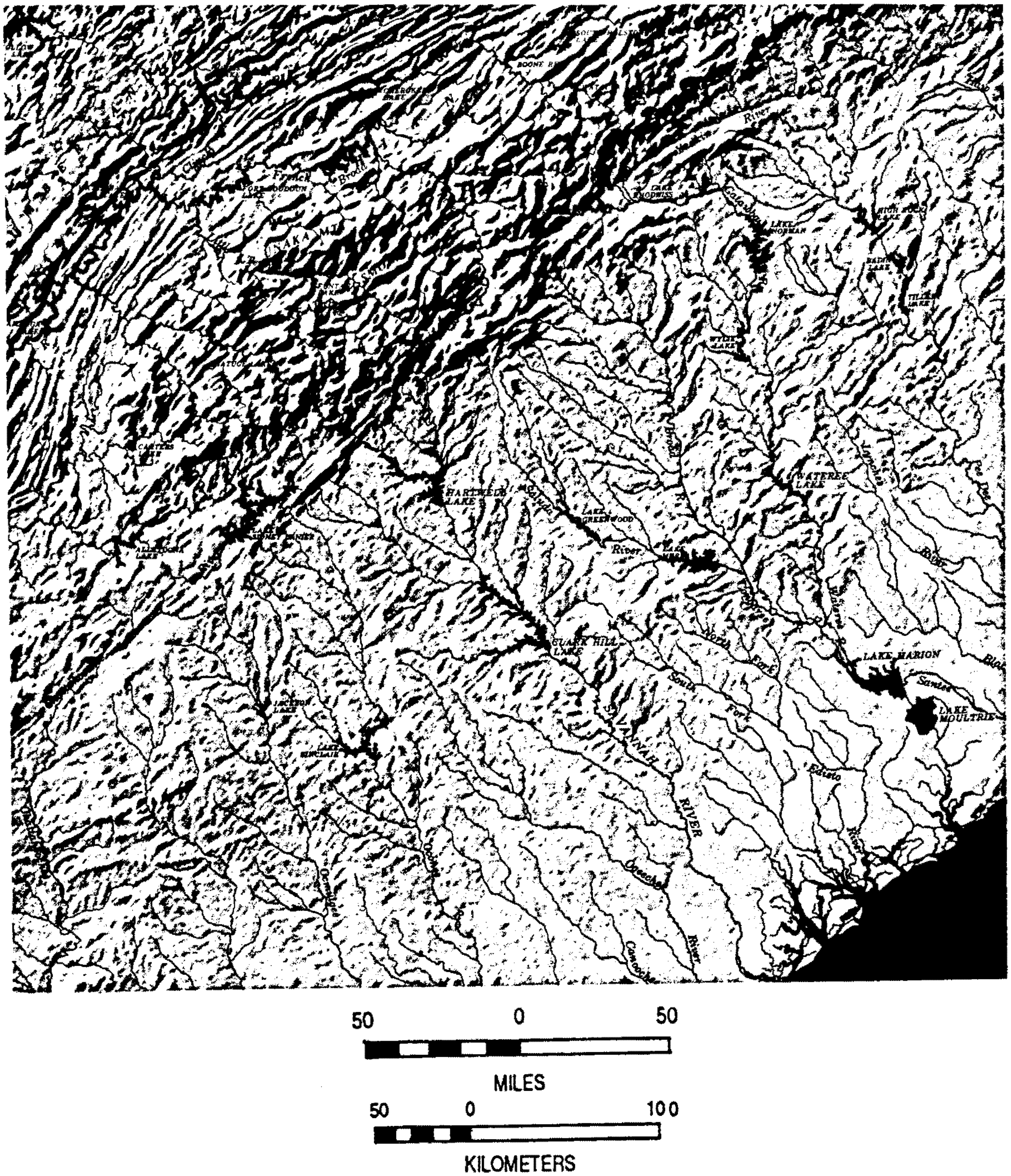

Figure 2. Portion of digital shaded-relief map of the conterminous U.S. showing map expression of the Blue Ridge Escarpment, isolated mountain masses, the Columbus Promontory Upland subsection and many other features. Rivers that drain the Columbus Promontory and the isolated mountain masses in the Carolinas (*cf.* Figure 1) are tributary to major Atlantic-Slope rivers that display downriver convergence to the Atlantic Coast in the Charleston, SC, area. From "Landforms and drainage of the conterminous United States." Reproduced with permission of Raven Maps and Images, 34 North Central Avenue, Medford, OR, 97501. For techniques used in production of such maps see Thelin and Pike (1991).

decreasing orders of magnitude of map area (Table 1).

The traditional method of identifying and mapping natural geomorphic subdivisions in the U.S. is based on similarities or differences in: geologic structure, lithology, topography, and geologic history (Thornbury, 1965). Also, in terrains where landmasses can be

Table 1. Ranking of landscape units by decreasing map area occupied, with examples appropriate for the Columbus Promontory Upland. Modified from Godfrey and Cleaves (1991).

Rank	Area (km^2)	Basis (Dominant Entity)	Example(s)
Realm	10^7	Largest Plate-Tectonic Units	North American Plate
Major Division	10^6	Sub-Continental Entities	Appalachian Highlands
Province	10^5	Regional Similarity	Piedmont province
Section	10^4	One Tectonic-Landscape Style	Inner Piedmont block
Subsection	10^3	Structure-Landform Similarity	Columbus Promontory Upland
District	10^2	Form-Material Relationships	Hendersonville Basin
Subdistrict	10^1	Direct Material-Form Linkage	Small Mountain Range
Zone	10^0	Few Form-Relief Parameters	Mountain Summit; Large Fan
Locale	10^{-1}	Individual Landforms	Cliff or Gully
Compartment	10^{-2}	Single Form-Relief Units	Typical Outcrop
Feature	10^{-3}	Specific Microform	Differential Weathering Features

identified as parts or collages of microplates, tectonic attributes of terranes can often be used effectively as regional geomorphic criteria. Landscapes also bear some stamps of the various formative climatic environments under which they have evolved, although until recently the climatic factor has largely been disregarded in American geomorphology except by workers familiar with such work in Europe and elsewhere. Finally, given enough geomorphic time, the operation of similar geomorphic process groups working on similar earth materials should logically be expected to result in similar erosional and depositional landforms, so that terrain genesis would also seem to be a highly desirable criterion. Many process geomorphologists, however, would no doubt argue that to implement such a scheme successfully would require data and genetic understanding far in excess of those available at present.

Davis (1993) defined the Columbus Promontory as an irregularly-shaped, high-relief area (700-1000 m) of Blue Ridge-type topography that extends from the Brevard fault zone on the northwest approximately 40 km to the southeast (Figures 1, 2) and that includes bedrock geology of parts of both the Chauga belt and the Inner Piedmont (see also Hatcher, 1993). From a regional geomorphic standpoint, the internal consistency of the Columbus Promontory Upland can be characterized and delimited by statistics of elevation and topographic slope (Pike and Thelin, 1989), and by relief ratio, and other, measurements. The Columbus Promontory Upland also displays well-demarcated boundaries; the Brevard fault zone on the northwest and other sharp surrounding landscape changes. Many of these breaks are hillslope discontinuities that are often surmounted by peaks such as Wildcat Spur, Chimney Rock, Tryon Peak, and Glassy, Hogback, Big and Little Warrior, and White Oak Mountains. As a regional geomorphic unit, the Columbus Promontory Upland stands out on digital shaded relief maps (Thelin and Pike, 1991), and remote sensing imagery such as SKYLAB photography (Griffin, 1974) and various examples of mosaicked radar imagery. The geomorphic subdivision terminology for this part of the southern Appalachians that recognizes the Columbus Promontory Upland as a geomorphic subsection is shown in Table 2.

STRUCTURAL GEOMORPHOLOGY

Introduction

Structural geomorphology is the branch of geomorphology that deals with the effects of bedrock structure and lithologic variation on the development of overlying landforms and landscapes. Important overall quests of structural geomorphic research are: to determine the exact relations between bedrock and geomorphology, to ascertain the chronology of topographic and drainage development, and to construct detailed geomorphic histories.

Out of necessity, the early geomorphic research in the southern Appalachians consisted largely of physiographic description and interpretation of the fluvial and erosional surface history from topographic maps and field reconnaissance. Many of the concepts and much of the terminology were imported—without modification—from classical study areas in the Appalachians of the Middle Atlantic and New England States. Early workers in the southern Appalachians assumed that erosional surfaces formed in a discrete cyclical

Table 2. Provisional geomorphic subdivisions used for the Piedmont province of the Appalachian Highlands major geomorphic division in North and South Carolina.

HIERARCHY (modified from Thornbury, 1965)	NOTATIONS
Appalachian Highlands major division	Thornbury (1965)
Piedmont province	Thornbury (1965)
Piedmont Lowlands subprovince	Modified from Hack (1982)
Danville Basin	Hack (1982)
Durham–Sanford Basin	Hack (1982)
Wadesboro Basin	Hack (1982)
Outer Piedmont	Hack (1982)
Carolina Slate belt	Hack (1982)
Uwharrie Mountains	Hack (1982)
Kings Mountain belt	Hack (1982)
Foothill zone subprovince	Modified from Hack (1982)
Smith River Allochthon section	
Sauratown Mountains Window section	
Inner Piedmont block section	Horton and McConnell (1991)
Chauga belt subsection	Hatcher (1972)
Inner Piedmont belt subsection	King (1955)
Columbus Promontory Upland subsection	Davis (1993)

mode of uplift-standstill-erosion, so that the highest surfaces were always the oldest and the lowest surfaces were always the youngest. Inter-regional correlation of these inferred erosion surfaces was therefore accomplished mainly on the basis of elevation. With respect to the origin and evolution of the master drainage lines, complicated schemes were invoked in order to fit the developmental paths of southern Appalachian rivers into the youth-maturity-old age paridigm of Davis (1889), a technique that had been employed extensively in the central Appalachians.

Today, results of modern bedrock-geologic research, and the availability of large-scale maps and high-resolution remote-sensing imagery, permit geomorphic studies that link geologic structures with landforms. The relatively high topographic relief in the Columbus Promontory Upland is associated with good bedrock exposures that permit much better spatial association of rocks and their structures with the overlying topography than is possible in many other areas of the Piedmont province. For examples, Hatcher (1993) and Garihan and others (1988) refer to a number of areas in the Inner Piedmont where structural control by fractures and rock units has been demonstrated.

Structure and Topography

A number of researchers have noted the commonality of upland levels and summit elevations in both the Blue Ridge and Piedmont provinces. Whether observed from the land surface at an overlook, or on topographic maps, these collages of upland surface elements seem to visually dominate the topography. Several elevation ranges (Table 3) occur and have been recognized by modern workers, and in earlier reports not referenced herein. For example, there is much current interest in the origin and age of the characteristic broad, flat, and concordant interflueves that are dissected by present drainage lines in the classical "Piedmont Upland" section (Thornbury, 1965), which encompasses most of the Piedmont province in the Fenneman classification (Fenneman, 1928; Fenneman, 1938; Fenneman and Johnson, 1946). As a specific example, LaForge and others (1925) described the Dahlonega Plateau in Georgia as such an area. There seem few detailed measurements of hillslope properties, although the works of Eckoff (1960) and Kesel (1974) are exceptions. Earlier workers described such uplands as peneplain remnants; Pavich (1985, 1989) however, considered that the present topography has evolved in an equilibrium fashion, with the rate of saprolite production roughly equal to the rate of removal. Soller and Mills (1991)

Table 3. Selected structural-topographic-drainage relationships in the Inner Piedmont block section of southwestern North Carolina and northwestern South Carolina from selected literature sources.

CATEGORY	LITERATURE SOURCE		
AUTHORS	Acker and Hatcher (1970)	Garihan and others (1993)	Haselton (1974)
JOINTS	N40-50°W N30-50°E	N50-70°W N50-70°E N80°E-N80°W N15°E	N50°W N50°E E-W N27°E
BRITTLE FAULTS	- -	N41-70°E	- -
FOLIATION AND/OR STRUCTURAL GRAIN	N35-70°E MAXIMUM: N50-60°E	- -	NE-SW
STREAM TRENDS	N50-70°W N30-40°W N-N10°W N10-N70°E, W/ STRONG N20-30°E	- -	NE-SW E-W
TOPOGRAPHIC GRAIN	- -	- -	N50-60°E
UPLAND LEVEL ELEVATIONS (m)	701 457-518 290	- -	609 "Asheville"

provided a well-balanced discussion of this controversy.

Many linear topographic trends are either parallel to, or bear some other consistent relationship with, trends of foliation, joints (and/or their intersections), fracture traces, lineaments, and faults (Table 3). The lowland areas that overlie much of the trace of the Brevard fault zone are examples of features that together produce a lineament zone of regional extent. Features of lesser length, but that are visible throughout the Columbus Promontory area include parallel-trending ridges and valleys, elongate dip slopes (Acker and Hatcher, 1970; Haselton, 1974), and single lineaments and fracture traces.

Location and shape of landscapes on the subdistrict and zonal levels (Table 1) may be strongly influenced by structural elements. For example, Kesel (1974) studied the kinds and characteristics of isolated mountains on the piedmont of Virginia, North Carolina, and South Carolina. He found that joint direction, joint intensity, and thickness and attitude of lithologic units were important factors in controlling size and orientation of inselbergs he studied.

Structure and Drainage

Trends of foliation, joints, fracture traces, lineaments, and faults are related to control of stream patterns (Table 3), although the exploiting mechanisms (*cf.* enhanced hydraulic plucking, physical abrasion, and chemical and biochemical corrosion) have been little studied in nature. Acker and Hatcher (1970) measured 2,078 stream channel directions and compared them with 1,169 joint measurements in northwest South Carolina. They found that drainage control by joints and joint intersections (and by foliation) is responsible for angulate and rectangular stream patterns. Several river areas (e.g. the Green River, and Hickory Creek-Broad River Gorge) have long, straight valleys interspersed with short reaches having angular bends. In a nearby area in northwestern South Carolina and adjacent North Carolina, Haselton (1974) found similar relationships among brecciated zones, jointing, foliation, and drainage patterns of both major and minor streams.

NEOTECTONICS AND MORPHOTECTONICS

Introduction

The term neotectonics can refer either to tectonic activity during different geologic-time slices of the Cenozoic or to activity during the length of time on a particular tectonic plate since the development of the present-day stress regime (see Gardner, 1989, for a discussion of neotectonics in the Appalachians). Morphotectonics (Morisawa and Hack, 1985) is the branch of geomorphology that investigates the relationships between tectonics and surface features. These fields have received increasing attention in geomorphic research for several reasons. There is increasing recognition of bedrock terranes on microplate-scale levels, coupled with growing usage of a wide variety of remote sensing imagery, that permit the identification and mapping of more rigorously defined areas with coherent bedrock geology, form, and relief. Practical and environmental considerations such as land use planning, waste disposal siting and mitigation, and earthquake hazards assessment have also added impetus to morphotectonic and neotectonic research.

A central question that is difficult to address in the Inner Piedmont is the relationship between neotectonic epeirogenic deformation and landscape development. The Atlantic Coastal Plain province, however, contains a wealth of stratigraphic information about Cenozoic events in those parts of the Appalachian Highlands that provided influxes of clastic sediments (*cf.* Nystrom, 1986). Offshore, the depositional record in the Atlantic Continental Margin contains a valuable and often much less discontinuous long-term record of former conditions in the field trip region perhaps especially in the Charleston Embayment-Southeast Georgia Embayment areas on the Carolina Platform and Hatteras Basin-Carolina Trough regions (Poag and Valentine, 1988; Ward and others, 1991, p. 274-275).

Deformation

Hack (1982) identified a number of subprovinces in the Piedmont using the criteria of bedrock geology, topography, and stream development (Table 2). He also delineated a line (the Tallapoosa-Rappahannock line), which separates the Piedmont Lowlands from terrains to the northwest. This major dividing line is expressed in the regional topography by a slight steepening of the regional topographic slope, which is accompanied by a faint lineation in the topography (Hack, 1982, p. 24). Hack suggested that, although both tectonic forces and rock resistance to weathering and erosion have had effects on producing the conspicuously high areas northwest of the Tallapoosa-Rappahannock line, tectonic processes have played the greater role. He reached this conclusion because many of the resistant rock units also occur in areas of low relief in the Piedmont Lowlands.

Prowell (1986) summarized the nature of faulting and related tectonism during the Cretaceous and Cenozoic. He concluded that the results of research on reverse faults demonstrate that regional compression in the crust has existed from at least the Early Cretaceous through the Pleistocene. Faults in Holocene sediments have not been recognized, however, perhaps due to the low slip rates calculated. Prowell calculated an average fault motion for the last 110 Ma that is close to 0.5 *m/* Ma. The tectonic effects of compression on the Columbus Promontory might have produced some deformation, but recognition of such late-stage deformation may prove difficult, and no geomorphic effects are known.

In order to constrain the geometry, magnitude, and timing of epeirogenic pulses, numerical age data on rock burial depth and cooling history are needed. Fullagar and Kish (1981) used both published and new mineral age data from samples along three traverses across the Piedmont of North and South Carolina to study rock burial and cooling history. In conjunction with estimates of the temperature and pressure conditions from metamorphic mineral assemblages, they determined the average geothermal gradient and estimated the depths where different minerals of determined radiometric ages reached blocking or closure temperatures. Their data show a general pattern of the youngest mineral ages (240-260 Ma) for the Inner Piedmont as contrasted with older ages (*ca.* 300 Ma) for the Piedmont to the east and southeast, suggesting that the Inner Piedmont was uplifted, eroded, and cooled later than the eastern and southeastern belts. Because of an unfortunate lack of age data for different minerals from individual rock samples, Fullagar and Kish did not attempt to calculate average uplift rates. Durrant (1978) used K/Ar thermal chronology to arrive at times of burial depth for an area on the Piedmont near Richmond, Virginia. Data from fission track analysis (Zimmermann, 1979) suggest that isostatic uplift in response to erosion is post-Triassic. More mineral age data would permit tighter constraints on the nature and timing of uplift, allowing rigorous evaluation of cyclic (*cf.* Davis, 1889) *versus* steady state, or equilibrium (Hack, 1960) models of denudation.

CLIMATOLOGY AND VEGETATION

Paleoclimatology and Vegetational History

Introduction. Because most estimates of rates of uplift, stream incision, and denudation (*cf.* Ahnert, 1970; Hack, 1979) suggest that modern landscapes produced in conjunction with tectonic events probably are no older than Tertiary, it is probably productive to concentrate on Cenozoic climatic history. Even within this time slice, there was, until recently, not much evidence that could be used for guidance. Such sources of information include paleoclimatic inferences from sediments derived from the Appalachians (*cf.* Poag and Sevon, 1989) and computer modeling (*cf.* Barron, 1989). Frakes and others (1992) discussed the complexity of Cenozoic climate changes and described major unresolved research problems. Cenozoic climates in southeastern U.S. apparently varied considerably during the first half of the Era, but followed a major trend of increasing warmth and rainfall accompanied by a lack of pronounced seasonality. Frakes and others (1992) indicated that a major change in climate started in the Middle Miocene with trends of cooling and precipitation change that culminated in the Pleistocene.

Tertiary Period. Tiffney (1985) discussed the vegetational changes that occurred in eastern North America during the Cenozoic. He noted that the warm-temperate to subtropical vegetation, which gradually developed over much of North America during the Eocene, was gradually replaced as the climate began world-wide cooling and increased seasonality. Barron (1989) indicated that the Appalachians would have been an area of focused precipitation throughout the Cenozoic, but with gradually decreasing rates.

A speculative scenario that can be created from the above is that, during the first part of the Cenozoic (to the Middle Miocene), the climate was sufficiently wet and hot to support a cover of abundant vegetation. Such an overall environment would have enhanced deep *in situ* chemical and biochemical weathering of rock, but would have inhibited physical erosion. As both climate and associated vegetation changed, a critical geomorphic threshold triggered by uplift was reached in the Middle Miocene. Large amounts of clastic sediment were then eroded from the Piedmont and transported to the Coastal Plain and even to offshore basins. Erosion slowed during the Pliocene, but was renewed in the Pleistocene, especially during cold intervals, which fostered vigorous glacial and periglacial erosion (Braun, 1989).

Quaternary Period. A wealth of information about climatic and vegetational history is contained in Quaternary terrestrial sediments in the Appalachians south of the glacial borders. These data indicate that episodes of cold climates were interspersed with interglacial occurrences of paleoclimates that were warmer than now. This land record, however, is often weathered, fragmented, and lacking in continuous sequences containing easily datable materials. The marine record must still be referred to for a long-term and nearly-continuous record of events on land. Using the marine record from several offshore Atlantic areas as a guide, the last major glaciation and related cold climates of pre-Wisconsinan age (> 130 Ka) is inferred by many workers to have been correlative with the Late-Illinoian glaciation of midwestern U.S. In like manner, the succeeding warm interval (130-75 Ka) is inferred to have been correlative with the midwestern Sangamon interglacial. The last major Wisconsinan refrigeration in the Appalachians is divided into a long, but not severely cold Early- and Middle-Wisconsinan interval (75-25 Ka; *cf.* Eyles and Westgate, 1987) and a shorter (25-12.5 Ka) Late-Wisconsinan that had extremely cold conditions in its earlier phases. The later phases were less cold, for example, the Late-Wisconsinan, Late-Glacial interval (*ca.* 16.5-12.5 Ka). Pollen records suggest that many upland areas in the American Southeast—especially in the Southern Blue Ridge geomorphic section—were above the forest limit during Late Wisconsinan time (Delcourt and Delcourt, 1985).

At the Pleistocene-Holocene boundary in the American Southeast (~12.5 Ka), dramatic changes in climate, vegetation, and geomorphic process-response mechanisms occurred. Environments, processes, and their effects rapidly approached essentially modern aspects in Early Holocene time. The Middle Holocene time interval—termed the Altithermal or the Hypsithermal—had elevated temperatures and decreased effectiveness of precipitation, as compared to the present. The Late Holocene time interval began with climatic conditions similar to those of today. It was followed by the Neoglacial geologic-climatic time unit (Porter and Denton, 1967), an episode of minor climatic deterioration that terminated with the end of "The Little Ice Age" (Grove, 1991). Subsequent climates in the excursion area have approximated those shown in historical records.

During Neoglacial time, environments at the higher elevations in the Appalachians were severe enough to produce minor remobilization of regolith that had apparently been stabilized throughout earlier Holocene time. On floodplains, accumulation of sediments suggest that one or more episodes of intensified aggradation may have occurred in the Neoglacial. Some of these events may predate the effects of European settlement. It is thus necessary for researchers to apply rigorous field and laboratory criteria to the study of forms and materials that could have developed under marginal periglacial conditions such as those that occurred during Neoglacial time.

Paleoclimatic Summary. Research to date therefore defines a number of "intermediate-member and end-member extremes" of Cenozoic paleoclimates for the southern Appalachian Highlands. Paleogene climates were hot and humid, and fostered the development of deep regoliths and extensive vegetational assemblages over the Appalachian Highlands. Marked climatic deterioration began in the Neogene and culminated with the rapidly alternating Pleistocene paleoenvironments, which fluctuated between cold-and-dry and warm-and-humid in the periglacial and interglacial phases, respectively. Suggested effects of climatic change on landscapes are given in the final section of this text.

Present-Day Climate, Vegetation, and Land Use

Climate and Vegetation. The Columbus Promontory Upland subsection is within the belt of humid subtropical climate. This climate occurs at the southern edge of the middle-latitude zone of conflict between polar and tropical air masses where tropical air masses dominate. In this zone, masses of maritime tropical air are infrequently interrupted by occurrences of cold winter weather air. Because of the influence of large warm water bodies, rainfall is plentiful except during droughts and is distributed fairly uniformly throughout the year. During the summer, heating of stationary to slow moving air masses brings high temperatures. Precipitation and temperature data for a station that has relatively long-term records in the Inner Piedmont are given in Table 4. The overall climate therefore has a moderate annual range of temperature, high humidity, and a long duration of frost-free days. The average time period between the last spring frost and the first fall frost, or "growing season" can range from less than 160 days in mountainous areas at the higher elevations to more than 220 days at low elevation stations with favorable exposures and air drainage. Mean annual air temperatures range from about 58 to about 61°F (14-16°C). For example, the E-W trend of the S side of the Columbus Promontory created a "thermal belt" (*cf.* Peattie, 1928-1929, p. 110-111)—Blue Ridge topography, but with warm winters and more moderate summers. Mean annual precipitation varies widely. Windward high elevation sites may receive 75" or more (1905 mm) per year, and valley areas may be rain-shadow areas with much less than about 45" (1143 mm) per year. For example, there are xeric areas in western North Carolina such as leeward slopes on the Blue Ridge Escarpment that are rain shadows, and that supported forest stands composed largely of *Castanea* (chestnut, a drought-tolerant genus) before decimation by combined effects of logging and chestnut blight (Braun, 1950, p. 220). Another example is that laurel

Table 4. Precipitation and temperature data from Tryon, Polk County, North Carolina (35° 12'; 82° 14'; 328 m). Years of record as of 1980 were 56 for temperature and 62 for precipitation. Compiled from NOAA Climatological Data, National Climatic Center, Asheville, NC.

MONTH	AVERAGE TEMPERATURES (°F/°C)	TOTAL PRECIPITATION (inches/mm)
JANUARY	43.2/06.2	4.93/125.2
FEBRUARY	45.6/07.6	5.64/143.3
MARCH	51.5/10.8	6.66/169.2
APRIL	60.8/16.0	5.09/129.3
MAY	68.3/20.2	4.51/114.6
JUNE	74.4/23.6	5.00/127.0
JULY	76.9/24.9	5.94/150.9
AUGUST	76.1/24.5	5.81/147.6
SEPTEMBER	70.5/21.4	4.94/125.5
OCTOBER	61.3/16.3	4.40/111.8
NOVEMBER	51.5/10.8	4.12/104.7
DECEMBER	43.4/06.3	5.24/133.1
ANNUAL	60.3/15.7	62.28/1582.2

and rhododendron thickets are concentrated on N- and NE-facing slopes, and deciduous genera such as poplar are more common on west- and south-facing slopes. Such variety in detail is a result of complex interactions among: differences in elevation, exposure, latitude, distance from water bodies, and other factors.

The factors listed above produce considerable local variations in climate (*i.e.*, microclimates), especially in the more mountainous parts of the subsection. When dealing with potential microclimate differences, the following factors should be borne in mind. The ridge crests, especially on their windward edges, will receive more and higher velocity winds than the valleys. Second-order, air-density- and topographic-driven winds may be channeled downslope, particularly in hollows and ravines. Especially during the earlier parts of the day, the valleys may be cold sinks, with the warmest temperatures occurring on the shoulders of the slope. The slope orientation (aspect) is very important on clear sunny days, when the difference of light between north- and south-facing slopes can amount to 46 units or more (say, in g cal/cm^2 hr^{-1}). Of course, for weathering studies, it would be desirable to know the total amounts on all slopes of interest. In diffuse light, all slopes should receive the same amount.

Unfortunately, there are insufficient meteorological stations in the mountainous areas of this part of the Appalachians to provide data on microclimates, average summit temperatures, or to calculate probable altitudes of treelines (Leffler, 1981a, b). Because of the shortage of mountain weather data and the high interest in climatic conditions on Appalachian summits (Leffler, 1981a), methods of approximation have been used. Leffler (1981b) analyzed temperature records from eight summit-level stations (topographic crests at least 300 m above surrounding land) from New Hampshire to South Carolina at elevations from 524 to 2022 m. He calculated lapse rates for summits in New England and developed interregional linear equations for computing 30-year average monthly and average daily maximum and minimum summit temperatures as functions of elevation and latitude. Leffler computed an annual lapse rate of 5.8°C/km that he believed to be representative of the Appalachian Mountain areas he studied and that includes the Columbus Promontory Upland area (Leffler, 1981b, Figure 1, p. 637). Schmidlin (1982) evaluated Leffler's equations in an area where they had not previously been tested and found that the estimated average monthly temperatures were within 0.6°C of the averaged long-term weather records.

The above summary indicates that the Columbus Promontory Upland subsection experiences a climate with elevated temperatures and abundant precipitation sufficient to foster rapid chemical and biochemical weathering through almost the entire year. Except at the higher elevations, frost action in soils and exposed rocks during the winters should be limited to short-term (mostly diurnal) effects except for rare invasions of extremely cold polar air masses.

Vegetation and Land Use. As an area with Blue Ridge topography but Inner Piedmont soil parent materials, the Columbus Promontory might be expected to have hosted forests of diverse compositions. Peattie (1928-1929) reported that few of the early botanists visited this part of the southern Appalachians. He reviewed the early records known to him and concluded that the pre-settlement vegetation in the Tryon region (a distinctive forest unit) consisted of a highly-endemic oak-pine forest with some 70 deciduous tree species widely diffused and intermixed in the association. Braun (1950, Chapter 7) included the Tryon region in the Oak-Chestnut Forest region. In some areas, chestnut was the dominant tree, often occurring in nearly pure stands.

In areas E, SE, and S of the Columbus Promontory, Braun (1950, Chapter 8) described an Atlantic Slope section of the Oak-Pine Forest region, and indicated that the Piedmont of North Carolina is most representative of this forest section. She noted that—even at the time of her observations—only small isolated stands of relatively little disturbed old age forests (200-300 years) existed. Except for isolated or protected areas, generations of land use practices resulted in a vegetational mosaic of cleared land, old-field successions with pine canopies, second-growth forests, and culled hardwood stands.

Stephenson and others (1993) provided an updated overview of the Appalachian Oak Forest (Oak-Chestnut Forest of Braun in the southern Appalachians) and Oak-Hickory-Pine Forest (Oak-Pine Forest region of Braun) of the southeastern U.S. The Appalachian Oak Forest contained a number of species of oak along with chestnut as the most common trees. Chestnut was a very important tree in mountainous areas of western North Carolina. These trees grew with hickory, locust, and pine on the more xeric sites, and commonly with poplar, beech, hemlock, basswood, birch, ash, and maple in more mesic sites such as the valleys and coves. On the Piedmont, Stephenson and others (1993) indicated that forested parts of the region are in third growth since European settlement, and that pines probably had a lesser, long-term role in overall forest composition except in disturbed or poor-quality sites. Thus, both the Appalachian Oak and Oak-Hickory-Pine Forest descriptions may reflect former disturbance histories and land use practices. Of course chestnut is gone from its former important role in the Appalachian Oak forest of the mountains, as well as in the Piedmont, and the steady-state typal forest of the southern Piedmont may actually be an oak-hickory-yellow poplar association.

A number of authors described quantitative effects of land misuse following European settlement of the Piedmont. Both early (*cf.* Ireland and others, 1939; Happ and others, 1940) and modern (*cf.* Trimble, 1975) workers have documented extensive soil erosion and gullying of agricultural land, and accelerated downvalley stream and floodplain sedimentation. Although the effects of land-use on erosion have become less severe since the 1930s, the effects of sediment reentrainment and deposition in fluvial systems continue (Trimble, 1975).

The most striking change in modern land use in the Columbus Promontory area has been the abandonment of land once in small farms and its reversion to forest (Stephenson and others, 1993). Other changes include increases in land used for retirement and "second home" development, recreation, and for commercial fruit orchards. The last usage is favored by the long growing season and excellent exposures and air drainage of the "thermal belt."

WEATHERING, REGOLITH, AND SOIL DEVELOPMENT

Exfoliation surfaces are important features in this geomorphic section, and are related to an interaction between the bedrock nature in thick-layered to massive rocks (*e.g.*, Henderson Gneiss) and the release of lithostatic confining pressure during evolution of the current relatively high topographic relief. These conditions lead to the development of structures such as sheeting (release) joints, and features such as talus caves, talus slopes, and sliderock (or screes). Excellent exposures in Chimney Rock Park, the bordering gorge of Hickory Creek-Broad River, and at Bat Caves illustrate the above features.

Effects of physical, biochemical, and chemical weathering on the alteration of fresh bedrock to regolith in the Inner Piedmont are profound. The term regolith is used here in its original sense to refer to the total range of unconsolidated materials underlying the land surface (*cf.* Gale, 1992). Regoliths of various types (as opposed to fresh bedrock) constitute the overwhelming category of parent materials that have been used for development of the modern soils in the Inner Piedmont.

By volume, the major kind of regolith in this section is saprolite. Becker (1895) proposed the term "saprolite" as a general term for thoroughly decomposed untransported bedrock where texture and structure are preserved *in situ*. Many publications contain information about saprolite characteristics in local study areas, but there seems to be no modern regional treatment that covers the Foothill zone subprovince. Overstreet and others (1968), however, provided excellent descriptions of saprolite including a major area studied for placer monazite in the Carolinas. They estimated that 95 percent of the land surface area of the Inner Piedmont is underlain by saprolite and that thicknesses are highly variable (from <1 to >50 m). The basal contact zone of saprolite with fresh bedrock is very irregular, usually thin to sharp, and does not conform to the overlying land surface. The upper contact of saprolite with colluvium and alluvium is generally sharp and does not conform to present-day topography (Overstreet and others, 1968).

Little published information seems available on transported regolith in the Inner Piedmont section. On a broader scale, Soller and Mills (1991) provided a summary of work on colluvial and alluvial sediments in the Piedmont and Blue Ridge provinces. In his last report, Eargle (1977) indicated that the parent materials of Piedmont soils in South Carolina were saprolite of crystalline rock, colluvium, and alluvium. He also noted that the percentage of colluvial parent materials had been underestimated by earlier workers, and was perceptive in this respect.

Soils in the Columbus Promontory area are predominantly Entisols, Inceptisols, and Ultisols (Soil Orders). Floodplain soils are mapped as Entisols, indicating that their parent materials have only been in place for pedogenically short periods of time. Soils on moderately to steeply sloping topographic noses, sideslopes, and hollows have one or more quickly forming horizons and are mapped as Inceptisols. Most of the hillslope soils are probably developed in colluvial parent materials. Many soils on gentle upland surfaces are Ultisols indicating sufficient landform stability for long-term soil development. A few soils in the area have been mapped as Alfisols. Modern pedogenic studies have concentrated on the development of soil properties that shed light on soil morphogenesis. For example, Rebertus and Buol (1985a) studied the distribution of Fe in a chronosequence of soils developed from mica gneiss and schist parent materials, and Rebertus and Buol (1985b) have determined that illuviation has been intermittent in Dystrochrepts and Hapludults in the Piedmont and Blue Ridge provinces of North Carolina.

SPECULATION ON LONG-TERM GEOMORPHIC EVOLUTION

Evolution of Topography

Several lithologies of Tertiary age on the Upper Coastal Plain reflect time intervals of high relief in Appalachian Highland areas to the northwest. An area of the Upper Coastal Plain in central South

Carolina contains the type area of the Sawdust Landing Formation, which rests unconformably on Upper Cretaceous sediments (Colquhoun and Muthig, 1991). Nystrom and others (1991) described the Sawdust Landing Formation as consisting primarily of clay but containing quartzose sand with common quartz granules and quartz pebbles, some rose quartz, and abundant angular feldspar grains up to 2.4 cm in length. Nystrom and others (1991) assigned the Sawdust Landing Formation an early Paleocene age.

By probable Middle Miocene time, the preserved Coastal Plain sedimentary record of clastic sediments indicative of source-area uplift had shifted southward. A major Piedmont-Blue Ridge uplift episode is recorded in an assemblage of sediments termed the "upland unit" by Nystrom and others (1991). This sedimentary package contains a clayey grit with the grit composed of cross-bedded, poorly sorted, poorly packed, angular to subangular, very coarse grained and granular quartz. Quartz-pebble stringers and cross-bedded quartz cobble deposits are also found in the upland unit.

The above findings suggest that at least some of the uplift in this part of the Appalachians—and hence the gravitational energy to power landscape development—may have been episodic and varied in nature from place to place. Hack (1982) marshalled evidence to indicate that topographic development over the crystalline rock terranes in this region required differential, and late geologic, uplift of both the Piedmont and Blue Ridge. Modeling studies of uplift can provide useful insight into possible uplift scenarios. Ahnert (1970) modeled several types of uplift, including cyclical episodic motion, for river basins comparable in size to those in and around the field trip area. He concluded that his results can be used to constrain hypotheses, but that irregularities in uplift rates and regional spatial variations in uplift nature probably complicate actual reconstructions of past uplift events.

Although origin of the Blue Ridge Escarpment is difficult to address at present, several authors have discussed its evolution. I suggest here that the Escarpment may not be of one geomorphic age, but, at least in North Carolina, developed progressively from northeast to southwest. The lines of evidence for this progression are only suggestive at present, and include both direct inferences from topographic and drainage features and indirect lines of reasoning about escarpment evolution. Landscape features that suggest sequential development include the progressive spatial separation of both isolated mountains and mountain masses to the northeast in South and North Carolina (Kesel, 1974, Figure 2) and the higher-then-lower, more rugged, and then increasingly imperfect escarpment development to the southwest in North Carolina. Indirect lines of reasoning include an extension of the consequences of White's (1953) idea of progressive dismemberment through time of an ancestral, northeast-flowing Inner Piedmont drainage. In this scenario, Atlantic Slope rivers progressively captured the old longitudinal drainage from northeast to southwest, successively providing new higher-gradient, Atlantic-Slope drainage to erode headward.

In the above context, use of the ergodic hypothesis (substitution of space for time) would allow us to use the Columbus Promontory as a field laboratory to study the dynamic processes by which the escarpment evolved. Caution, however, will be needed because only present-day processes can be studied directly, and the nature and effects of past environments (*e.g.*, periglacial, tectonic) will have to be interpreted from the study of the geomorphic record. Almost everything remains to be learned about the geomorphic history of the Escarpment, including the timing, rates, and mechanisms of scarp production, and the climatic and tectonic environments under which it has developed.

Drainage Development

Origin and evolution of regional drainage patterns on the Inner Piedmont have been complicated. Both the effects of structural control and the results of river piracy on drainage development have probably been underestimated except by a few workers. For examples, White (1953) hypothesized regional dismemberment of ancient northeast-flowing drainage along the Inner Piedmont by headward-working Atlantic Slope rivers to produce the present-day drainage systems, Acker and Hatcher (1970) emphasized the importance of stream capture in drainage development in northwest South Carolina. More recently, Whittecar (1984, 1988) documented the presence of high volumes of debris in terraces that still partially clog second- and third-order stream valleys in the Inner Piedmont of Western North Carolina (Figure 3), further documenting the earlier findings of Eargle (1940, 1977) and Overstreet and others (1968) that the effects of Quaternary morphogenetic systems on mountain fluvial systems were pervasive.

Understanding of long-term drainage evolution will be necessary in order to comprehend the origin and evolution of landscape features such as the Blue Ridge Escarpment and mountainous areas on the Inner Piedmont. The Columbus Promontory is a area where headwater streams are actively dismembering a landscape, and can serve as an outdoor laboratory for the study of processes of topographic development that are occurring along such features as the Blue Ridge Escarpment. Long-continued dissection should result in the production of a mountain-mass core that will become isolated from the Escarpment. In this interpretation, the

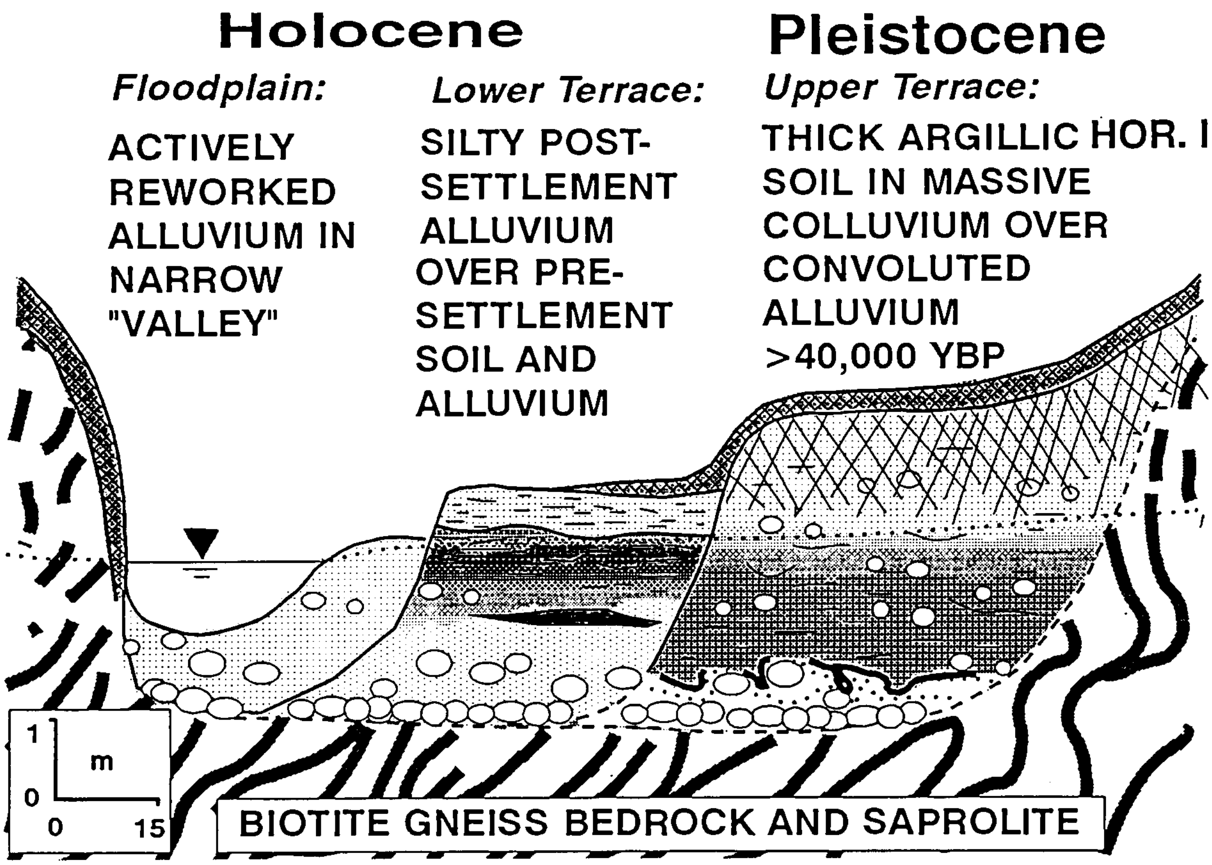

Figure 3. Schematic cross section of stratigraphy beneath geomorphic surfaces in valley of "Lutz Creek" (local informal name for a tributary to Henry River in Catawba County, NC). Floodplain alluvium (channel gravel and point-bar sand) is actively reworked within the narrow entrenched valley of the modern stream. The lower terrace contains a fining-upward sequence of alluvium capped by a variety of fluventic and histic soil profiles; over the pre-settlement soil are up to 1 m of silty post-settlement overbank deposits and sheetwash fans from sideslopes. The upper terrace contains basal sandy alluvium intermixed with contorted (frost-heaved?) beds of silty and organic-rich sediment (>40,000 Ka); a thick and massive silty-stony sand diamict dominates the section, grading upward from organic-rich gray colors to whitish at the water table. A well-drained, truncated argillic Bt horizon (>1 m thick) is present in soils at the top of many terraces; some terraces contain multiple truncated argillic horizons formed in stacked diamicts. Diagram and description prepared by G. R. Whittecar, Department of Geological Sciences, Old Dominion University, Norfolk, VA, 23529-0496. (see Whittecar, 1984, 1988).

South Mountains constitute a remnant of a once-more-continuous range, which has been separated from the Brevard fault zone and the Blue Ridge through valley development by headwaters of the Broad and Catawba Rivers. In like manner, the Brushy Mountains have become isolated by valley development of the Catawba and Yadkin Rivers. Such may become the fate of the Columbus Promontory as headward extension of the Broad, Saluda, and Savannah Rivers continues at the expense of Gulf of Mexico drainage.

Quaternary Landscape Evolution

By 2.4 Ma (Braun, 1989), the combined effects of global climatic deterioration and other factors produced the first great Quaternary ice sheets in the Northern Hemisphere, and the earliest associated geomorphically-effective cold climates south of them. As documented by Braun (1989), the Appalachian Highlands south of the glacial borders experienced episodes of intense periglacial weathering, mass wasting, and fluvial activity that resulted in widespread production of new landforms and geomorphic materials.

Literature on the geomorphic history of the Inner Piedmont in the Carolinas during Quaternary cold phases is surprisingly scarce, however. Clark and Ciolkosz (1988) summarized the periglacial geomorphology of the Appalachians south of the glacial border and noted several reports of features interpreted to have had probable cold-climatic origins. In particular, Eargle (1940, 1977) described thick colluvial and alluvial deposits that overlie discontinuous organic-rich sediments rich in pollen and macrofossils that indicate much cooler climatic conditions. Eargle reconstructed some of the paleotopography in the Spartanburg area, South Carolina, and concluded that the relief and sharpness of Late Pleistocene drainage divides and valleys were greater. The modern landscapes have topography with much more rounded and subdued valleys, sideslopes, and divides. Eargle inferred that the processes of topographic modification included extensive colluviation and gully gravure (in this case, side-to-side migration of third-order streams on hillslopes), producing lateral divide migration on hillslopes. Many of the high-level gravel deposits noted in the region (cf. White, 1953; Overstreet and others, 1968; Haselton, 1974) may have been deposited by streams heavily loaded with debris during Quaternary cold phases (Clark and Ciolkosz, 1988; Whittecar, 1984, 1988).

Environments similar to the present-day became established during Holocene time; the major effects on landscapes are interpreted to have been those associated with cataclysmic hillslope and fluvial events, establishment of Holocene vegetational assemblages, and regolith weathering and development of the modern soils. One geomorphically important effect of the Early and Late Holocene climates during spring, summer, and fall seasons was the increased availability of moisture-laden air masses from the Atlantic Ocean and the Gulf of Mexico (Delcourt and Delcourt, 1981). Such moisture supplies, coupled with the increased moisture capacity of warmer air columns, provided conditions permitting increased likelihood of catastrophic precipitation events that could modify landscapes rapidly. Newson (1980) distinguished two types of floods, based upon their different kinds of geomorphic effectiveness: "slope floods" that produce severe hillslope and toeslope modification, and "channel floods" that mainly impact floodplain areas. Both types of storms can be observed in the same geographical area during a short span of time under the same overall climate. Historic and stratigraphic evidence indicates that both "slope flood" and "channel flood" events occur and are common in southern sections of the Blue Ridge and Piedmont provinces, although the visible effects of slope floods that result in debris slide and debris flow activity are most evident in the Blue Ridge (Clark, 1987; Mills and others, 1987).

The hillslope and channel flood effects discussed above need to be borne in mind when discussing the survivability of supposed relict geomorphic deposits and landforms, especially those as old as Pleistocene. For example, there may be few to no relict features on certain steep hillslope areas and on affected floodplains.

CONCLUSIONS

Many of the hindrances to advancement of geomorphic knowledge in the Inner Piedmont (and bordering Blue Ridge) are associated with the lack of knowledge about: details of the bedrock geology, geochemical and geophysical constraints on uplift and covermass unroofing, the stratigraphy and sedimentation of regolith, and the inability to date geomorphic materials and landforms. There is now sufficient detailed bedrock geologic mapping in several areas so that surficial geologic and geomorphologic research have solid bases to proceed upon. New methods for dating geomorphic surfaces (Phillips and Dorn, 1991) provide techniques that will allow numerical age dating of exposed surfaces, and the calculation of erosion rates from these exposures. A bright future for geomorphic research on both old and new problems is thus possible.

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THE MILTON BELT - CAROLINA SLATE BELT BOUNDARY: THE NORTHERN EXTENSION OF THE CENTRAL PIEDMONT SUTURE ?

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ABSTRACT

The central Piedmont suture is a fundamental crustal boundary in the southern Appalachians, separating probable continental-based crust of the Inner Piedmont from volcanic arc rocks of the Carolina terrane from Georgia to north central North Carolina. The northern continuation of the suture is problematical and depends upon the affinity of metamorphic tectonites of the Milton belt, situated northeast of the Inner Piedmont and northwest of the Carolina slate belt, the heart of the Carolina terrane. Compilation of available regional data as well as new field work on the Milton belt-Carolina slate belt boundary indicates that the this boundary is a viable, if not favored, extension of the central Piedmont suture.

INTRODUCTION

The central Piedmont suture (CPS) is a first-order crustal boundary in the southern Appalachians; it was defined as separating volcanic arc rocks of the Carolina terrane (Ct) to the east from continental-based crust of the Inner Piedmont (IP) to the west (Hatcher and Zietz, 1980) (Fig. 1). The suture extends for the length of the IP, from Georgia to north central North Carolina; however, north of Winston-Salem, NC, the extent and location of this fundamental structure is uncertain and controversial. In this area, some workers have assigned metamorphic tectonites of the Milton belt that lie immediately east of the Mesozoic Dan River basin (Fig. 1) to the Ct and imply that the basin overlies the northward extension of the CPS (Glover and others, 1983; Hatcher and others, 1990; Baird, 1991). In contrast, other workers consider the Milton belt as either an extension of the IP (Kish, 1983; Sinha and others, 1989; Hibbard, 1991) or a distinct crustal block (Horton, and others, 1989, 1991), thus placing the northern extension of the central Piedmont suture at the eastern margin of the belt, truncating the Charlotte belt and abutting the heart of the Ct, the Carolina slate belt.

The purpose of this brief interim report is to present some preliminary results of a new ongoing field study of the Milton belt - Carolina slate belt boundary in the context of the regional geologic framework of the IP and Ct. On the basis of regional compilation and the new work, the Milton belt-Carolina slate belt boundary is viewed as a viable northern extension of the CPS.

THE CENTRAL PIEDMONT SUTURE

The CPS was defined on the basis of potential field data (Hatcher and Zietz, 1980) and is clearly expressed in the surface geology (e.g. Hatcher, 1989). In particular, the suture was defined on the basis of the steep Appalachian gravity gradient, displaying increasing values eastward from the IP into the Ct and an abrupt change in magnetic patterns from uniform broad wavelength patterns in the IP to choppy, high frequency patterns in the Ct (Hatcher & Zietz, 1980).

These geophysical changes correspond to a continuous zone of ductile-brittle shear that separates distinct lithologic packages in the surface geology (e.g. Griffin, 1978; Horton & Butler, 1981). The IP is characterized by biotite gneiss, granitoid gneiss, mica schist, amphibolite, quartzite and marble (e.g. Griffin, 1974, 1978; Davis and others, 1990); granitic gneisses from the IP have yielded radiometric ages ranging from 420-480 Ma interpreted as ages of intrusion (Odom and Russell, 1975; Harper, 1977; Fullagar, 1971). The Carolina terrane comprises Late Proterozoic to Cambrian metaigneous and metasedimentary, volcanic arc rocks (e.g. Butler and Secor, 1991).

Perhaps the most striking contrast across the CPS is the abrupt change in structural style from the Ct to the IP. The Ct is generally characterized by upright folds and a steep foliation whereas IP units are disposed in regional scale recumbent nappes with a prominent gently dipping foliation (Griffin, 1974; 1978). The CPS lies within a broader structural zone that has been interpreted to be the root zone for the IP nappes (Griffin, 1974; 1978).

In addition, isotopic and petrochemical contrasts are documented across the CPS (LeHuray, 1986; Misra and others, 1990). Pb- and Sr- isotope data indicate that IP rocks are probably continentally-derived, whereas the Ct are mantle-derived, with little continental contamination (LeHuray, 1986). Amphibolites from the IP and Ct have been shown to be texturally and petrochemically distinct (Misra and others, 1990).

THE MILTON BELT - CAROLINA SLATE BELT BOUNDARY

Previous Work

A compilation of previous work indicates that the contact between the Milton belt and the Carolina slate belt is marked by contrasts in lithic character, potential field expression, and structural style. These contrasts bear strong resemblance to those encountered across the CPS further south. To the west, the Milton belt constitutes a distinct lithologic package of mainly sillimanite grade biotite gneiss, granitoid gneiss, mica schist, amphibolite, quartzite, and marble (Butler and Secor, 1991); granitoid gneiss within the belt has yielded radiometric ages of ~ 425-465 Ma (Henika, 1980; Hund, 1987, reported in Sinha and others, 1989) similar to those of the IP, yet distinct from any suite of Ct granitoids. To the east, the Carolina slate belt is composed of dominantly volcanic arc-related, greenschist grade Late Proterozoic to Cambrian igneous and sedimentary rocks.

In addition to this lithologic-metamorphic change, the Milton belt-Ct boundary coincides with the northeastern extension of the abrupt, first-order geophysical gradients that define the CPS. In the vicinity of Winston Salem, NC, where the IP appears to terminate (Figure 1), the sharp potential field gradients swerve eastward to faithfully outline the Milton belt-Ct boundary (Hatcher and Zietz, 1980).

The Milton belt-Ct boundary is also marked by a major structural contrast that is highly reminiscent of relationships along the CPS. The Milton belt is characterized by regional scale recumbent nappes with a gently dipping foliation (Tobisch & Glover, 1971; Henika, 1977; Kreisa, 1980; Baird, 1991). This style contrasts sharply with the upright folds with subhorizontal axes accompanied by a steep foliation typical of the Carolina slate belt.

In spite of these contrasts in lithology, metamorphism, potential field data, and structural style, the Milton belt-Carolina slate belt boundary has been considered to be a conformable and gradational infrastructure-suprastructure boundary within rocks now designated as Ct (Tobisch & Glover, 1971; Baird, 1991), although some workers have locally identified the contact as a structural zone of uncertain significance (Kreisa, 1980; Baird, 1991). Ongoing new work along the boundary in the area of the North Carolina-Virginia border undertaken during the present study also indicates that the boundary is tectonic; recent isotopic work confirms that the boundary is likely a significant tectonic break.

Preliminary Results

In the area of the North Carolina-Virginia border, new mapping and structural analysis has led to the preliminary recognition of a complex heterogeneous, multiphase structural zone along the Milton belt-Carolina slate belt boundary (Hibbard, 1991, unpublished data; Shell and Hibbard, 1993) (Fig. 2). This zone is at least 7 km wide, involves three major deformational events superimposed onto mainly mafic and granitoid protoliths, and appears to truncate structures in the Carolina slate belt. The zone was a locus for early granitoid pegmatites that in most places have been transposed into a gneissic banding. During generation of the gneissic banding, most of the pegmatites have been grain-size reduced, leaving only relict, coarse porphyroclasts of feldspar up to 25 mm across in a fine-grained foliated felsic matrix. Sense-of-shear indicators are lacking in this apparently mylonitic rock, probably due to the intense, multi-phase deformational overprint involving two phases of tight to isoclinal folding and foliation formation. The later of these two events appears to be related to dextral oblique normal shear that heterogeneously spans the entire zone. Structures change within this zone from dominantly flat-lying in the Milton belt to upright in the Carolina slate belt (Fig. 2).

In summary, the Milton belt-Carolina slate belt boundary in the area of the North Carolina-Virginia border is a wide composite tectonic zone, involving an early mylonitic gneiss-forming event that has been overprinted by intense, multi-phase folding and dextral oblique normal shear. The timing of these events is largely unconstrained, although they appear to post-date a ~ 575 Ma granitoid (Glover and Sinha, 1973) at the Carolina slate belt edge of the zone.

The significance of this tectonic zone at the Milton belt-Carolina slate belt boundary is not clear in the local area, but on the basis of the regional compilation outlined above, it appears that the zone is a profound tectonic break that has all of the 'earmarks' of the CPS. In order to further test this hypothesis, we have initiated a Nd isotopic study of the Milton belt and Carolina slate belt. Initial results on 23 samples appear to indicate that the Milton belt has a distinct continental isotopic influence that is lacking in the more juvenile Carolina slate belt (Samson and others, 1992a, b). This isotopic pattern appears to mimic that for lithologic packages on either side of the CPS (LeHuray, 1986) and thus, further supports a correlation of the Milton belt-Carolina slate

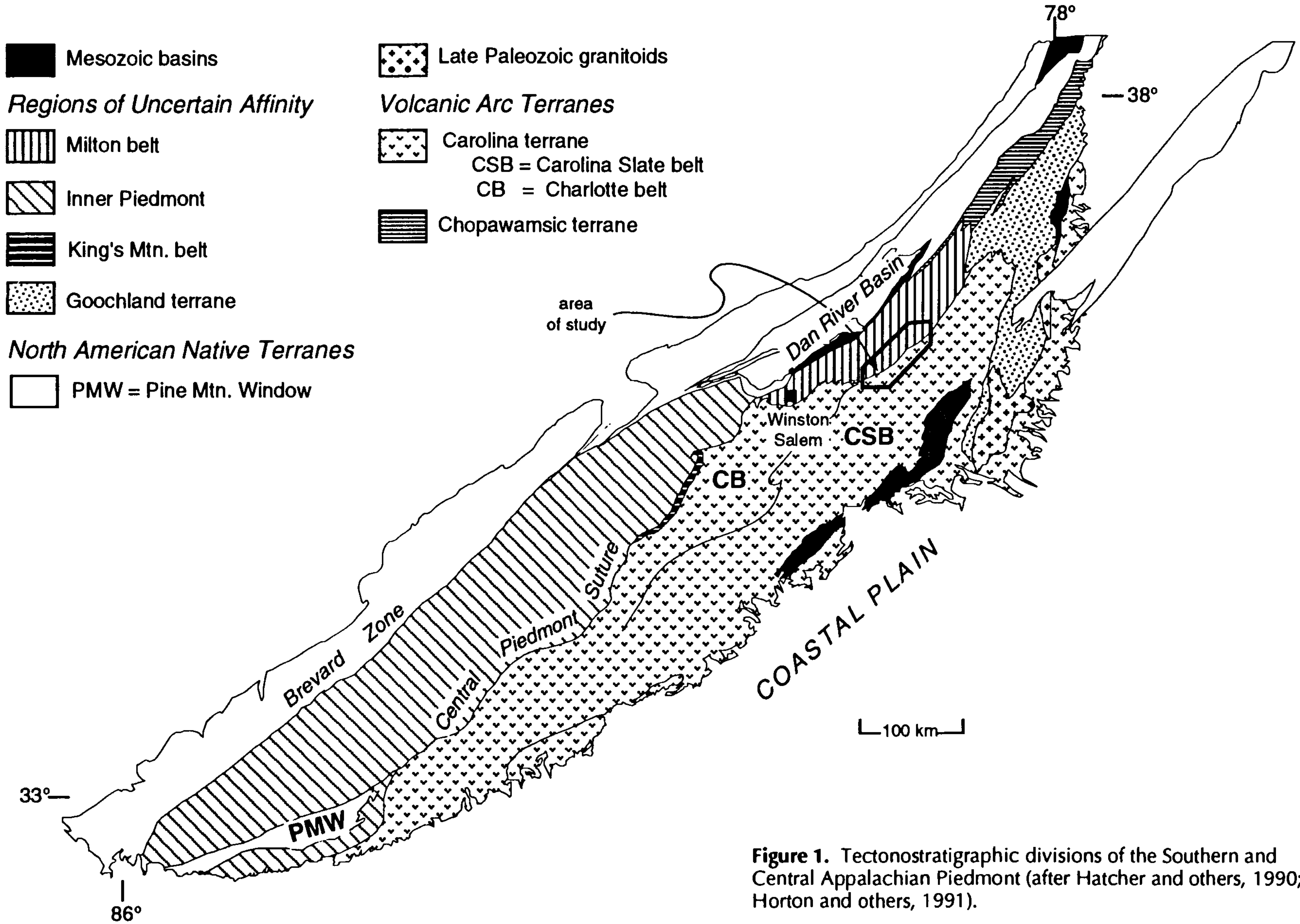

Figure 1. Tectonostratigraphic divisions of the Southern and Central Appalachian Piedmont (after Hatcher and others, 1990; Horton and others, 1991).

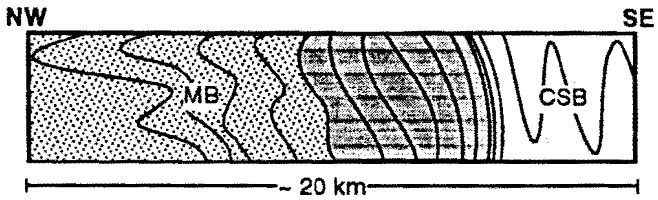

Figure 2. Simplified schematic cross section across the Milton belt (MB)-Carolina slate belt (CSB) boundary; based on field data collected during the present study. The dense stipple pattern denotes the structural zone separating two belts; linework indicates attitude of compositional layering, and depicts the youngest fold events in respective belts.

belt boundary with the CPS.

CONCLUSIONS

The compilation of existing data concerning the Milton belt-Carolina slate belt boundary strongly indicates that the boundary bears striking resemblance to the CPS. Both structures form the eastern boundary of similar lithologic packages (IP and Milton belt) and these packages are identical in regional structural style (recumbent nappes) and potential field expression; moreover, they contrast sharply with Ct rocks and structures to the east. Preliminary results of this study indicate that the Milton belt-Carolina slate belt contact is not simply a metamorphic gradient, but a multiphase tectonic zone involving early mylonitization overprinted by at least two intense phases of deformation, the later of which involved heterogeneous dextral oblique normal faulting. In addition, preliminary Nd isotopic data indicate that the Milton belt and Carolina slate belt have experienced different paths of crustal evolution. On the basis of these preliminary observations, it is proposed that the Milton belt-Carolina slate belt boundary is a viable, if not favored, candidate for the northward extension of the CPS.

ACKNOWLEDGMENTS

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GEOLOGY OF THE SOUTHERNMOST BREVARD FAULT ZONE, ALABAMA, AND ITS IMPLICATIONS FOR SOUTHERN APPALACHIAN TECTONOSTRATIGRAPHY

Jonathan E. Grimes, Mark G. Steltenpohl, Robert B. Cook, and William D. Keefer

ABSTRACT

Mapping in east-central Alabama documents lithologies of the Brevard fault zone continue eastward around the hinge of the Tallassee synform into the Opelika Complex of the Inner Piedmont. This dictates a reevaluation of the structural configuration of the southernmost Brevard fault zone. Quartzites of the Opelika Complex on the east limb of the Tallassee synform have been mapped westward into quartzites of the Jacksons Gap Group along the west limb. These quartzites occur with structurally interleaved garnet-kyanite schist, graphite schist, and felsic gneiss. This lithologic package is overlain by the Ropes Creek Amphibolite of the Dadeville Complex but is underlain by the eastern Blue Ridge. Retrogressive, greenschist-facies, right-slip mylonite zones of the Brevard fault zone do not accompany the Jacksons Gap Group - Opelika Complex through the Tallassee synform hinge zone. Rather, higher-temperature, amphibolite-facies metamorphic and annealed mylonitic fabrics characterize units in the hinge zone and along the east limb. The retrogressive Brevard fault zone, does not bend south near Jacksons Gap, Alabama, but diverges from Jacksons Gap Group to continue southwestward. The retrogressive Brevard fault zone may merge with the Alexander City fault, through right-slip splays. These findings require a reevaluation of tectonostratigraphy and possible correlation of the eastern Blue Ridge and Inner Piedmont.

INTRODUCTION

The Brevard fault zone is an extensive fault zone that extends from Mt. Airy, North Carolina, to Jacksons Gap, Alabama (Fig. 1). Brevard fault zone lithologies make a change in attitude from a northeast strike and moderate southeast dip to a northerly strike and a shallow eastward dip near Jacksons Gap (Fig. 2). Farther south, workers (Stose, 1926; Neathery, 1968) suggested that the strike changes from northwest, to north, and finally to east-west, with a shallowing of the dip, before becoming covered by the Gulf Coastal Plain. These relations led to the idea that the southern Appalachian Piedmont is allochthonous along a west-directed thrust comprising the Brevard fault zone and faults framing the Pine Mountain basement window (Clarke, 1952; Fig. 1). COCORP developed a similar interpretation based on seismic-reflection profiling (Cook and others, 1979). The early history of the Brevard fault zone remains the subject of debate (Bobyarchick and others, 1988).

Herein we summarize our geologic and structural analysis of rocks south of Jacksons Gap, Alabama, which has focused on the mapping of stream drainages that penetrate beneath the Coastal Plain (Fig. 3). Significant findings are summarized as follows: (1) Lithologic units constituting the Brevard fault zone in the study area, called the Jacksons Gap Group, do not disappear beneath the Coastal Plain as previously described (Bentley and Neathery, 1970). Rather, these units can be mapped eastward around the hinge zone of the Tallassee synform where along the eastern limb they correspond to the Opelika Complex. (2) The tectonostratigraphy, from top to bottom, the Dadeville Complex - Jacksons Gap Group - eastern Blue Ridge, wraps around the hinge of the synform as well. This results in a panel of eastern Blue Ridge rocks in contact with rocks of the Pine Mountain basement massif along the eastern limb of the synform, a structural position equivalent to the Hollins Line fault in other parts of the orogen, which requires a reevaluation of the tectonostratigraphic framework of the southernmost part of the orogen. (3) A distinction must be made between lithologies of the Brevard fault zone (i.e., Jacksons Gap Group) and the structural features that characterize what we consider to be the Brevard fault zone. This is critical because the late retrogressive structures and fabrics that characterize the Brevard fault zone northeast of our study area do not accompany the Jacksons Gap Group eastward around the Tallassee synform hinge zone. Rather, higher temperature annealed mylonitic/metamorphic fabrics wrap around the hinge, implying that the late structures continue southwestward along the structural grain of the Brevard fault zone, whereas the

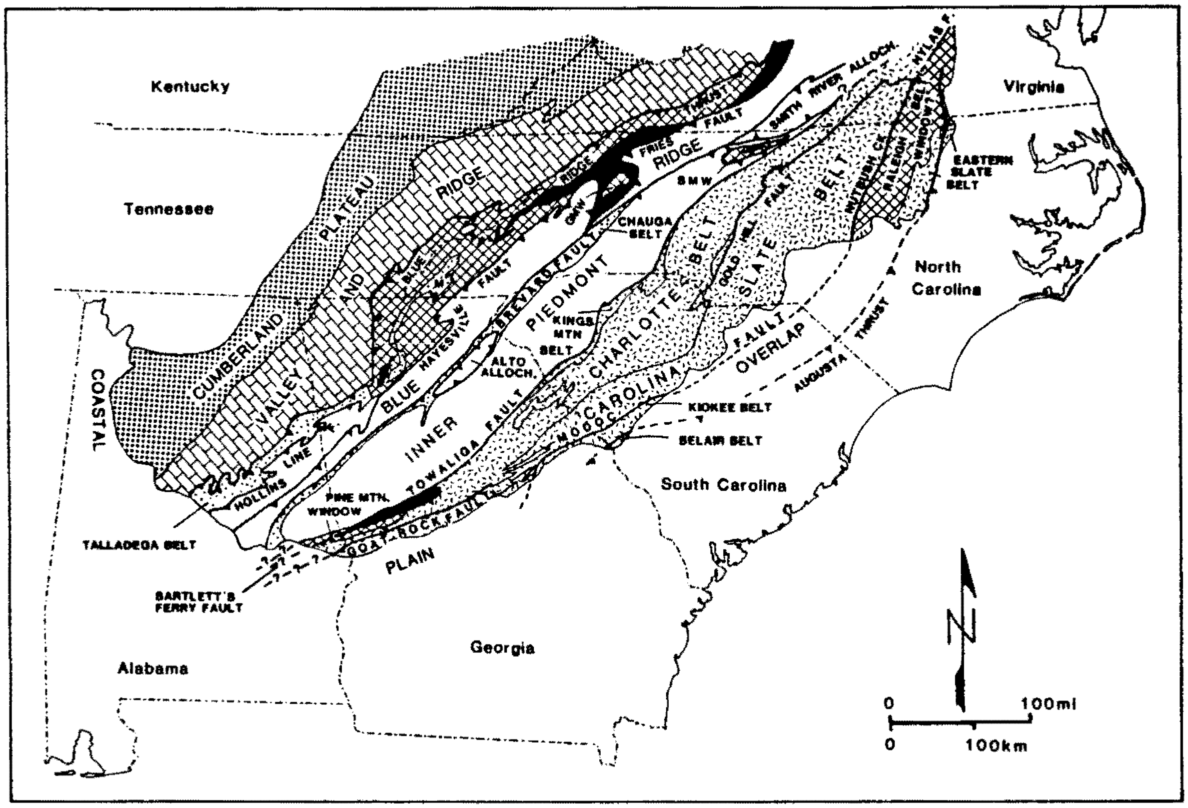


Figure 1. Generalized tectonic map of the southern Appalachians illustrating the location of the study area, major fault zones, and tectonostratigraphic subdivisions (taken from Hopson and Hatcher, 1988).

early pre-retrogressive mylonitic rocks do not. This area therefore offers the unparalleled opportunity to investigate the early history of the Brevard fault zone.

RECENT FINDINGS

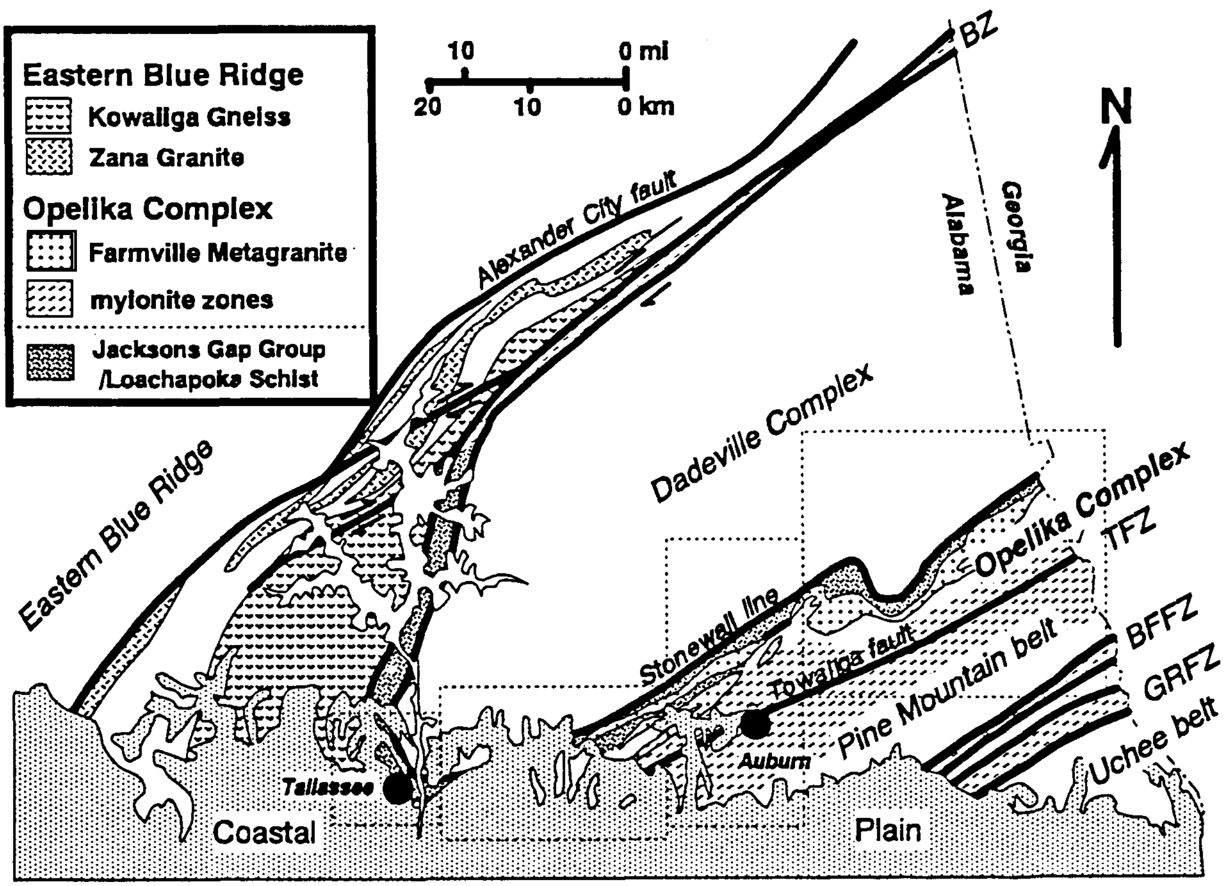
Contributions to understanding the geology along the Fall Line were made by Keefer (1992). His mapping along the west limb of the Tallassee synform, focused on the units of the Jacksons Gap Group within the Brevard fault zone. In mapping these units eastward, he recognized the misplacement of the Tallassee Metaquartzite (Raymond and others, 1988) on earlier maps. He also described the Stone Creek imbricate zone, a complex tectonic interleaving of quartzite and orthogneiss (Keefer, 1992). Contrary to earlier workers' maps that indicate the disappearance of the Jacksons Gap Group beneath the Coastal Plain cover directly southeast of Tallassee (e.g., Bentley and Neathery, 1970; Osborne and others, 1988), Keefer demonstrated that exposures of the Tallassee Metaquartzite and the Stone Creek imbricate zone continued eastward around the Tallassee synform hinge (Fig. 3). He suggested a correlation of the Jacksons Gap Group to the Opelika Complex. Keefer also noted that retrogressive structures of the Brevard fault zone elsewhere do not occur in the Jacksons Gap Group of the hinge zone of the synform. Rather, the deformational fabrics of rocks in the hinge zone formed under higher grade deformational conditions than did the fabrics of the Brevard fault zone (i.e., lower to middle amphibolite facies, kyanite-staurolite zone, and greenschist facies; Fig. 4). Keefer also recognized an amphibolite-facies ductile deformation zone along Stonewall line (i.e., the base of the Dadeville Complex) which he interpreted as a fault.

Recently, Grimes and Steltenpohl (1993) mapped the area between those mapped by Keefer (1992) and Steltenpohl (1990). Mapping within stream drainages (Fig. 3), Grimes and Steltenpohl (1993) demonstrated a connection of Keefer's Stone Creek imbricate zone (i.e., the Jacksons Gap Group) with the kyanite-staurolite schist, graphitic schist, Saugahatchee quartzite, and sheared metagranite of Sears and others' (1981) Loachapoka formation (i.e., Loachapoka Schist of Bentley and Neathery, 1970; Fig. 5). The same authors recognized Sears and others' (1981) Auburn formation which comprises the Auburn schist (metapelite) and Auburn gneiss (metagraywacke), both of which were previously considered Auburn Gneiss by Bentley and Neathery (1970)(Fig. 5).

Grimes and Steltenpohl (1993) also reported felsic intrusives in the Loachapoka Schist and Auburn Gneiss are more abundant and texturally variable than was previously thought and may not be restricted to the Devonian Farmville Metagranite. They mapped a body of eastern Blue Ridge augen gneiss, which Keefer (1992) correlated with the Zana Granite (Osborne and others, 1988), eastward around the hinge of the Tallassee synform into augen gneiss of the Opelika Complex previously mapped as Farmville Metagranite. Grimes and Steltenpohl (1993) reported that felsic intrusives of the Opelika Complex can be subdivided into augen gneiss, lineated orthogneiss, and metagranite. This relationship is similar to the problem of distinguishing the Kowaliga Gneiss from Zana Granite in the eastern Blue Ridge (Bieler and Deininger, 1987). The same authors also reported a number of thin (<10 m thick) amphibolites within the Auburn Gneiss. These observations combined with the metapelite/metagraywacke character of the Auburn Gneiss led Grimes and Steltenpohl (1993) to suggest correlation with the Emuckfaw Group (Bentley and Neathery, 1970) of the eastern Blue Ridge.

Figures 2 and 3 illustrate our understanding of the geology along the Fall Line in this area. The Jacksons Gap Group can be mapped eastward out of the retrogressive Brevard fault zone around the hinge of the Tallassee synform and into the Loachapoka Schist (Fig. 3). Structural form-lines (Fig. 6) for the Loachapoka Schist are simpler and provide good structural control. These form-lines document weak, open, upright, late-phase folds modifying S_1. Form-lines away from the Loachapoka Schist are complex, interpreted as competency contrasts between the rigid granitic rocks and weaker phyllosilicate rocks. Additionally, post-M_1 ductile deformation zones, such as those described by Sears and others (1981) and Steltenpohl and others (1990), further disrupt the form-lines within Loachapoka

Figure 2. Generalized geologic map of the crystalline rocks along the Alabama fall line. Dashed blocks illustrate areas studied, from west to east, by Keefer (1992), Grimes and Steltenpohl (1993), Steltenpohl (1988), and Sears and others (1981). Abbreviations are BFZ = Brevard fault zone, TFZ = Towaliga fault zone, BFFZ = Bartletts Ferry fault zone, and GRFZ = Goat Rock fault zone.



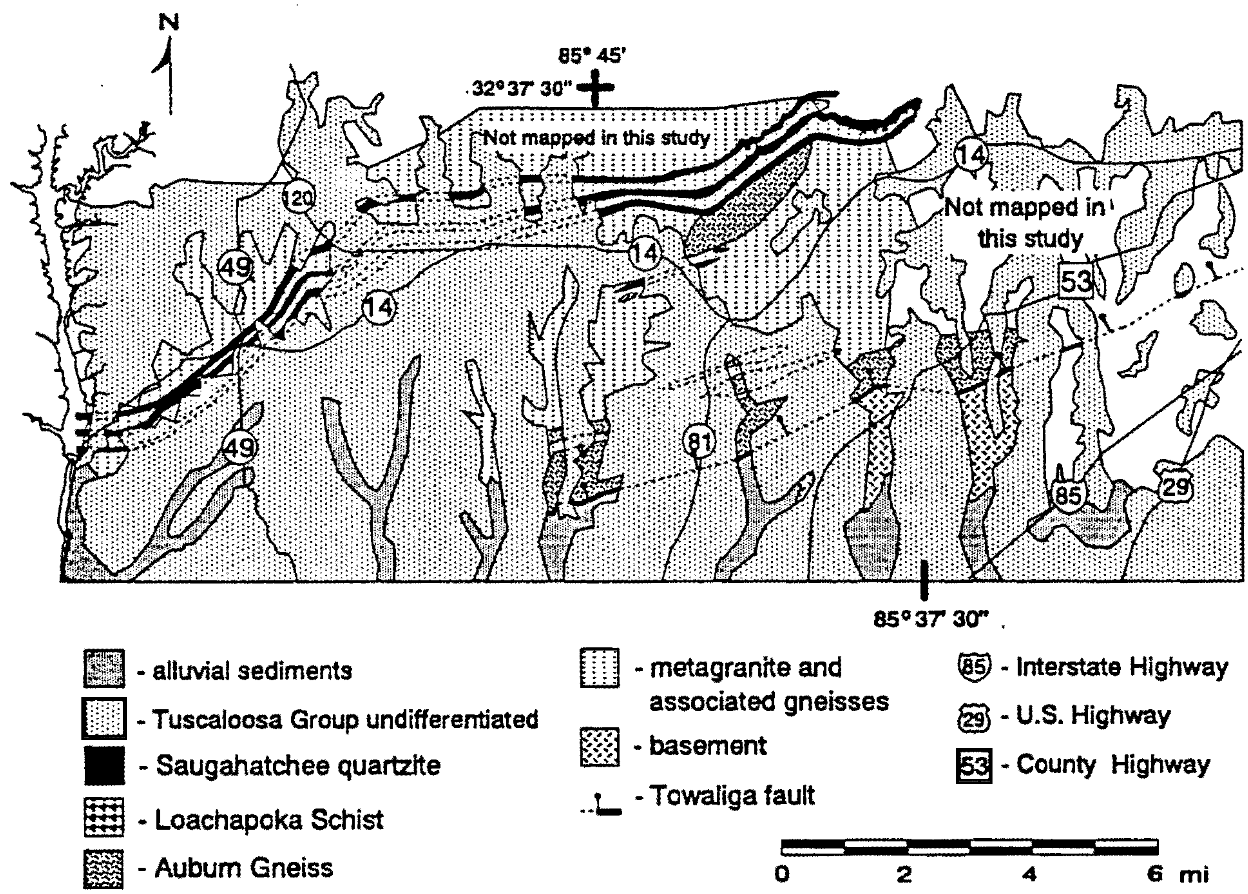


Figure 3. Geologic map of the areas mapped by Keefer (1992) and Grimes and Steltenpohl (1993).

Schist. Shear-sense indicators from these deformation zones document oblique, right-slip, top-down-to-the-northwest normal displacements (Steltenpohl, 1988; Keefer, 1992; Grimes and Steltenpohl, 1993). As described by Steltenpohl and others (1990), these shear zones root into large-scale, 6-7 km amplitude, late-phase folds that resulted from this oblique, right-slip motion. Orientations, kinematics, deformational conditions based on microstructures, timing, and geographic position of these shear zones document their formation coeval with the Towaliga fault zone (Steltenpohl, 1988; Steltenpohl and Kunk, 1993).

DISCUSSION: TECTONOSTRATIGRAPHIC CORRELATIONS

An important finding of our study is that the entire tectonostratigraphic package of the Dadeville Complex, Jacksons Gap Group, Loachapoka Schist, and eastern Blue Ridge wrap around the hinge of the Tallassee synform. This requires a re-evaluation of the tectonostratigraphy in this part of the orogen. Stratigraphic correlations within the southern Appalachian Piedmont are difficult due to metamorphism and deformation coupled with sparse absolute-age data on metasedimentary protoliths. Therefore, the following correlations are based on similarities in lithology, ages (where available), and tectonostratigraphic positions.

Units structurally above the Jacksons Gap Group and Loachapoka Schist correspond to the Dadeville Complex (Fig. 7). The Dadeville Complex contains metaplutonic and metavolcanic rocks intruded by meta-igneous rocks (Steltenpohl and others, 1990), some which are reported to be Ordovician in age (Seal and Kish, 1990). The Dadeville Complex is a true igneous complex and is not unlike the Inner Piedmont in other parts of the orogen (Steltenpohl and others, 1990). The Inner Piedmont is a stack of thrust sheets containing schist, gneiss, amphibolite, and rare ultramafic bodies (Higgins and others, 1988). It is interpreted as either a microcontinent formed during pre-Appalachian rifting of the Laurentian margin, the reworked and thrust emplaced remnants of this margin (Hatcher, 1978), or as several amalgamated disrupted terranes (Horton and others, 1989).

(a)

(b)

Figure 4. (a) Greenschist-facies microstructures in granitic mylonites characteristic of the retrogressive Brevard structural zone along the west limb of the Tallassee synform. Note crystal-plastic quartz (ribbons and LPO), mica halo, and delta-type feldspar porphyroclast. (b) Amphibolite-facies mylonitic/ metamorphic microstructures of granitic Jacksons Gap Group zone lithology from the hinge zone of the Tallassee synform. Note well-equilibrated, triple-point grain boundaries in quartz, and late coarse-grained biotite, muscovite, and feldspar indicating a lack of unrecovered strain in this rock. Field of view is 3.00 mm for both photomicrographs.

In contrast to the Dadeville Complex, the underlying Loachapoka Schist and Auburn Gneiss are continentally derived metasedimentary rocks that have been intruded by Devonian felsic plutons. The quartzites, metapelites, and schists of the Jacksons Gap Group and Loachapoka Schist are distinct. Quartzites like the Devils Backbone/ Tallassee and Saugahatchee quartzite are rare in the Piedmont; the only other quartzite in this region is the Hollis Quartzite of the Pine Mountain Group which overlies the Grenville basement massif.

The Jacksons Gap Group has been mapped northeastward from Jacksons Gap, Alabama, into Georgia where Medlin and Crawford (1973) called lithologically similar units associated with sparse marbles 'Brevard fault zone lithologies'. Northeastward along strike Higgins and McConnell (1978) and McConnell and Abrams (1984) called these units the Sandy Springs Group (Fig. 7). On many maps, Hatcher and coworkers included these units in the Chauga belt (Hatcher, 1978; Hopson and Hatcher, 1988). Hatcher (1970) reported that the base of the Chauga belt is the western margin of the Brevard fault zone and provides stratigraphic control of the fault, which is similar to what we have observed along the base of the Jacksons Gap Group northeast of Jacksons Gap, Alabama. Workers agree that rocks corresponding to Chauga belt -Brevard lithologies - Jacksons Gap Group contain lower metamorphic grade mineral assemblages than the adjacent units (e.g., greenschist facies vs middle amphibolite facies, kyanite and/or sillimanite grade); whether or not these assemblages reflect prograde or retrograde metamorphism is debated (e.g., Higgins and others, 1988). Higgins and others (1988) implied these units to reflect a rift-basin fill assemblage mixed with volcanic material.

Along the east limb of the Tallassee synform, the Loachapoka Schist projects northeastward into rocks that Higgins and others (1988) assigned to the Sandy Springs thrust sheet. They considered the southern Appalachians to comprise thin, west-directed, thrust sheets rather than broader lithotectonic 'belts.' The Sandy Springs Group of Higgins and McConnell (1978) was interpreted by Higgins and others (1988) to be confined to the Sandy Springs thrust sheet. The Chattahoochee Palisades Quartzite is a quartzite within this group and is associated with schists of the Factory Shoals Formation (Higgins and others, 1988). Our reconnaissance mapping along the Georgia - Alabama border, allows connection of the Saugahatchee quartzite with the Chattahoochee Palisades Quartz-

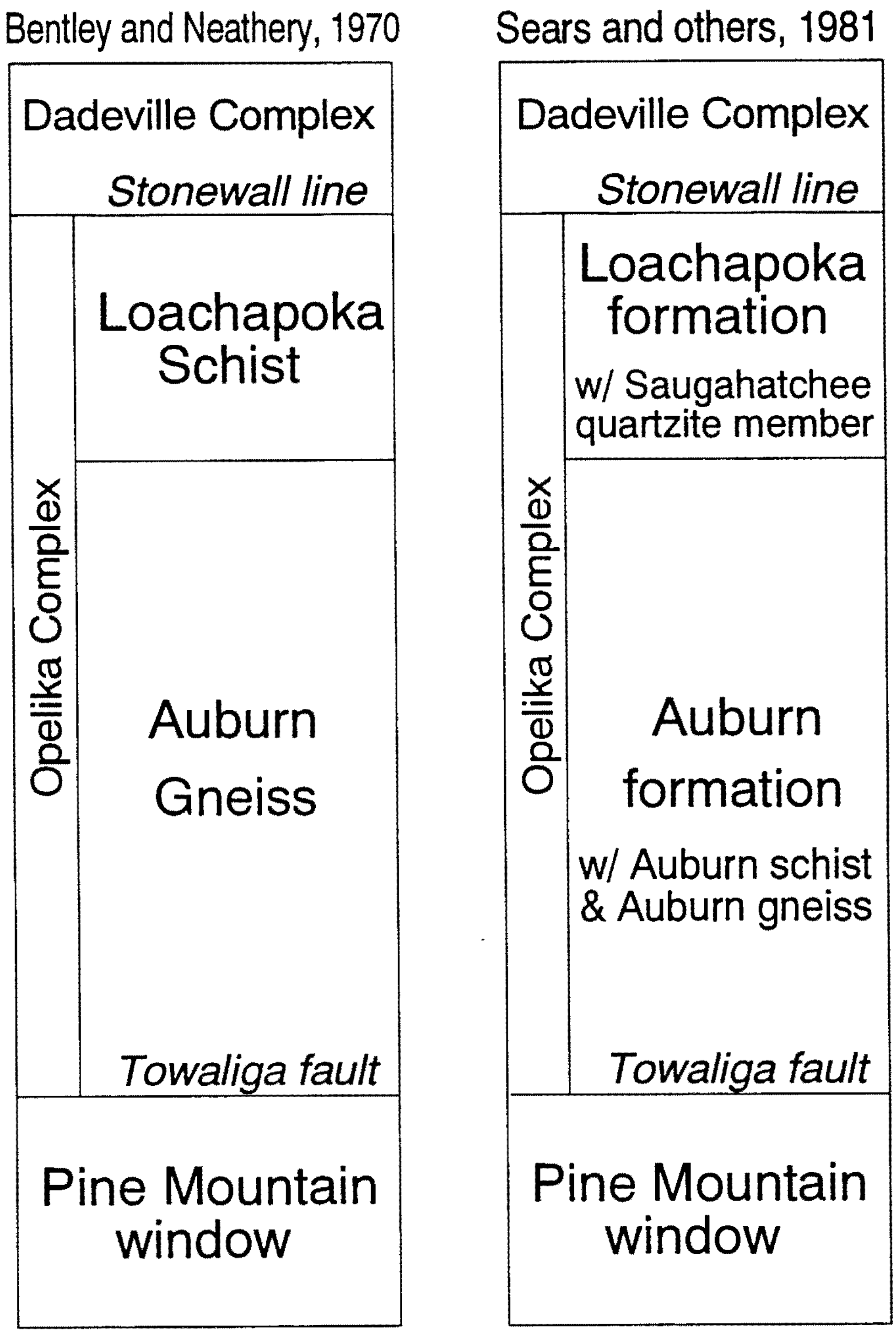


Figure 5. Schematic tectonostratigraphic columns comparing terminology used for rocks in the study area.

ite and associated schists. On their tectonostratigraphic map, Higgins and others (1988) indicated that klippen of Sandy Springs Group rocks occur farther northward in Georgia where these rocks approach the Brevard fault zone north of Atlanta (Fig. 8). Kline (1981) reported that these units are lithologically similar to those occurring within the Brevard fault zone near Atlanta.

Higgins and others (1988) reported that north of Atlanta, the Sandy Springs Group can be mapped across the Brevard fault zone. The same is implied by our lithologic correlations, as can be seen by the conver-

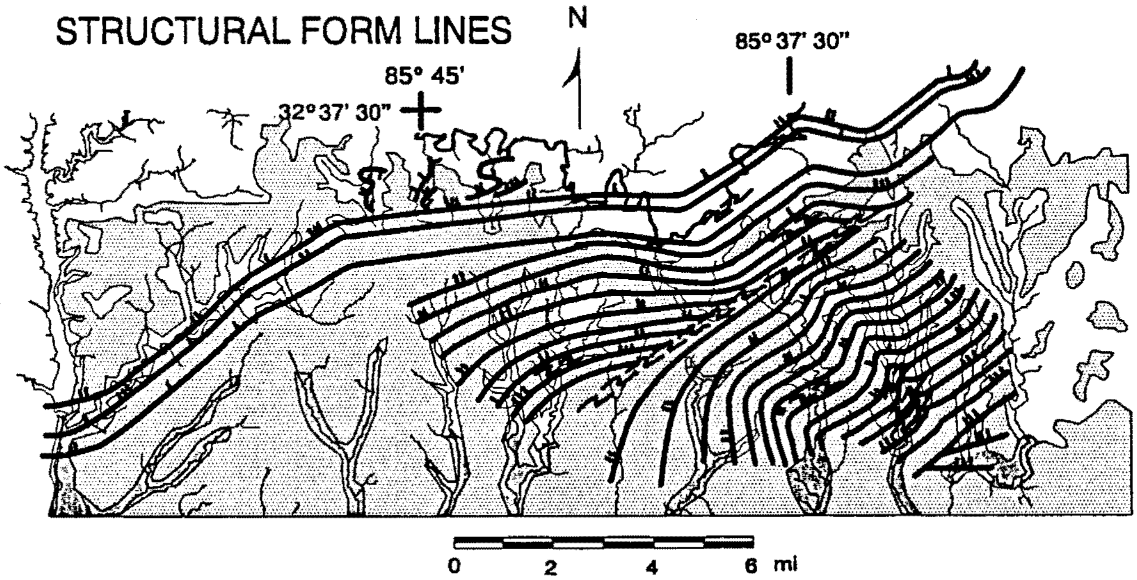


Figure 6. Form-line map of the dominant S_0/S_1 foliation for the area depicted in Figure 3. Tick marks indicate magnitude of dip (one tick mark < 30°, two tick marks 30° <60°, and three tick marks > 60°).

gence of the Jacksons Gap Group -Loachapoka Schist - Sandy Springs Groups in the Tallassee area (Fig. 7). The implication is that lithologies associated with the Jacksons Gap Group can be mapped southwestward from near Atlanta to Tallassee, eastward around the hinge of the Tallassee synform, and northeastward back to the Atlanta area. The Dadeville Complex therefore appears to be a large klippe in the core of the Tallassee synform. This structural setting is not unlike that of the Alto allochthon (Hopson and Hatcher, 1988) which occurs northward along strike in Georgia. The lithologies of the Alto allochthon, metagraywacke, schist, biotite gneiss, and minor amphibolite (Hopson and Hatcher, 1988), contrasts, however, with the of amphibolite and ultramafic rock of the Dadeville Complex.

The Stonewall line may be a major fault zone separating the Jacksons Gap Group - Loachapoka Schist sequence from the Dadeville Complex. The position of the Loachapoka Schist is variable due to structural interleaving and the present authors' observations indicate that the actual fault boundary lies along the structural top of the Loachapoka Schist. Keefer (1992) and Grimes and Steltenpohl (1993) reported high-temperature, syn-M_1, annealed mylonitic/metamorphic rocks from along the Stonewall line. These fabrics locally are overprinted by retrograde D_2 mylonitic shear zones indicating reactivation of the Stonewall line.

The location of the trace of the Stonewall line

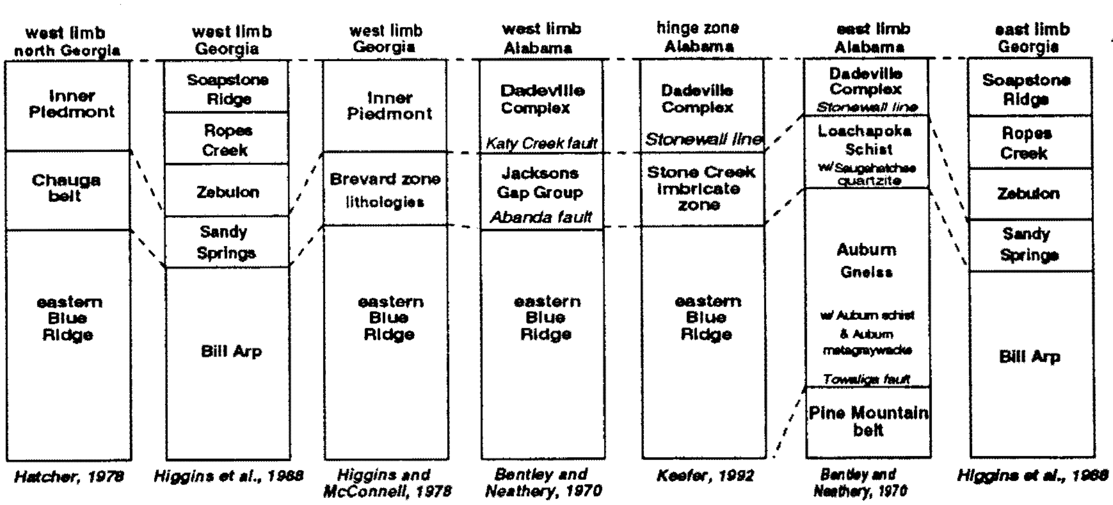


Figure 7. Tectonostratigraphic correlations suggested for crystalline rocks in the hinge of the Tallassee synform along the Alabama fall line with other parts of the southern Appalachians (not to scale).

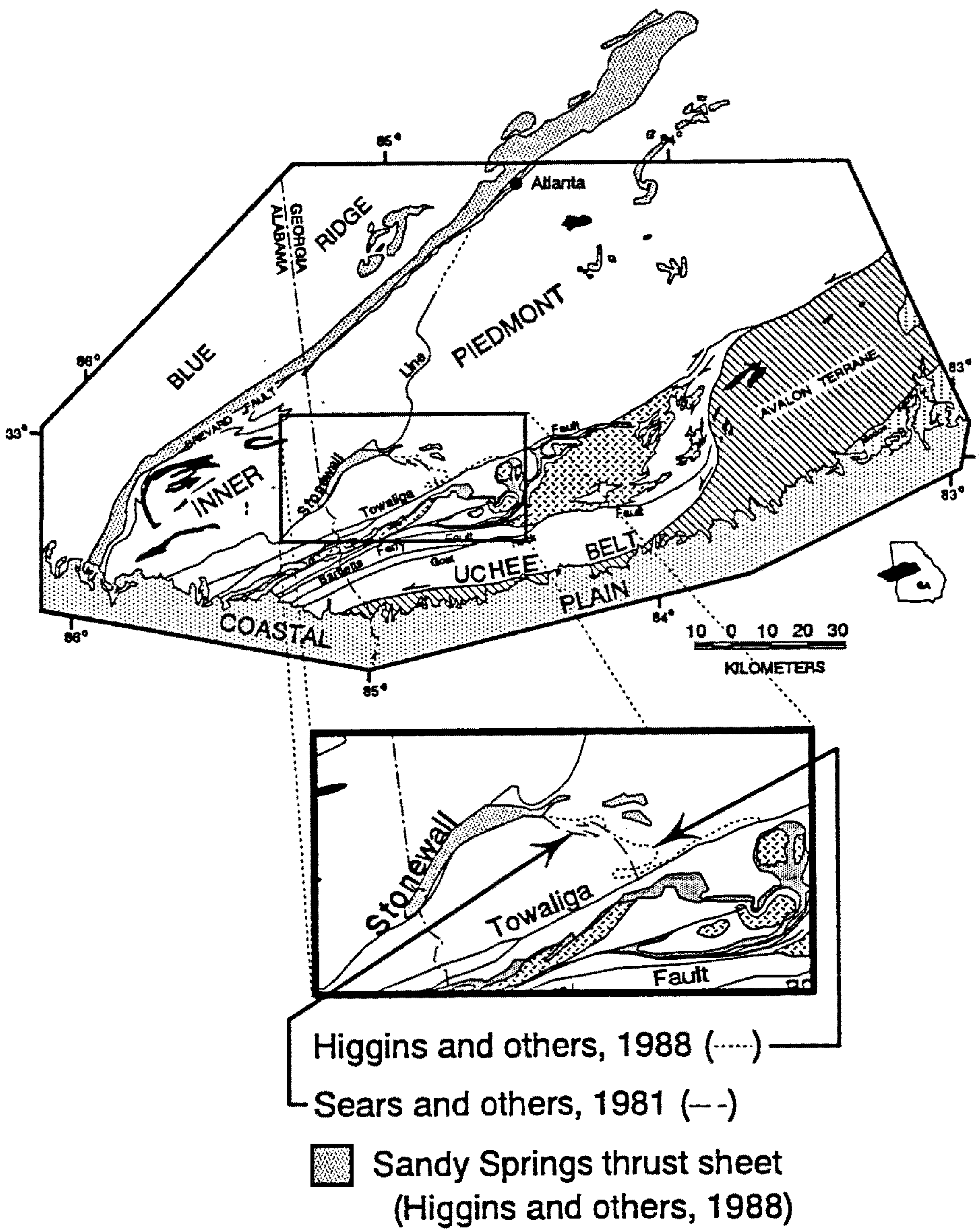


Figure 8. Comparison of various workers' placement ofthe Stonewall line and the Sandy Springs thrust sheet (see text). Base map is modified from Hooper and Hatcher (1988).

northeast of the study area is problematic. Bentley and Neathery (1970) suggested that the Stonewall line separates meta-igneous rocks of the Dadeville Complex from metasedimentary rocks of their "Randolph" lithologies. The same authors drew the Stonewall line northeast to merge with the Brevard fault zone, southwest of Atlanta (Fig. 8). In contrast, Sears and others (1981) reported that northeast of the Alabama - Georgia State line rocks associated with the Stonewall line become assimilated by granitic plutons and are no longer recognizable. Nonetheless, Sears and coworkers indicated that the Stonewall line strikes eastward and then southeast where it is truncated by the Towaliga fault (Fig. 8). In Georgia, existing maps (Georgia Geological Survey, 1976; Higgins and others, 1988) indicate that units corresponding to the Sandy Springs lithologies are discontinuously preserved northeast of the State line as klippen (Fig. 8). Higgins and others (1988) depicted the klippen to bend toward the southeast, and to the north where they merge with the Brevard fault zone northeast of Atlanta, farther east than indicated by other workers.

In Alabama, units lying to the northwest, structurally beneath the Jacksons Gap Group have been called the Ashland/Wedowee belt (Adams, 1926; Neathery, 1975), the northern Piedmont (Bentley and Neathery, 1970), or the eastern Blue Ridge (Tull, 1978; Steltenpohl and Moore, 1988). Osborne and others (1988) reported that the Emuckfaw Group (Bentley and Neathery, 1970), the part of the eastern Blue Ridge underlying the Jacksons Gap Group west of the Tallapoosa River in Alabama, contains metapelite, metagraywacke, and minor amphibolite that are intruded by the Ordovician (Russell, 1978) Kowaliga Gneiss and the Zana Granite. Hatcher (1978) first suggested the subdivision of the Blue Ridge in western North Carolina and northern Georgia into western, with rare meta-igneous rocks, and eastern halves, with abundant metavolcanic and metaplutonic rocks. He called the contact between them the Hayesville-Fries fault system. The eastern Blue Ridge is defined to lie above and east of the western Blue Ridge and below and west of the Brevard fault zone. Hatcher (1978) considered the eastern Blue Ridge to reflect a late Precambrian (?) rift-related basin. Horton and others (1989) used the term Jefferson terrane for the same group of rocks and interpreted it to be a disrupted terrane. Hopson and Hatcher (1988) pointed out similarities between rocks of the Inner Piedmont and the eastern Blue Ridge and interpreted them to represent a marginal basin developed upon attenuated North American crust.

Several lines of evidence indicate to us that units of the Auburn Gneiss are part of the eastern Blue Ridge rather than the Inner Piedmont. Grimes and Steltenpohl (1993) report that the Auburn Gneiss contains a metagraywacke and metapelite sequence with thin (<10 m thick) amphibolite. These units are similar to those of the Emuckfaw Group (Bentley and Neathery, 1970), which underlies the Jacksons Gap Group along the west limb of the Tallassee synform. This lithologic package also is similar to parts of the Tallulah Falls Formation of Hatcher (1971) and the Sandy Springs Group (Higgins and McConnell, 1978) which compose part of the eastern Blue Ridge to the north.

The metasedimentary rocks in the Auburn Gneiss become less abundant southwestward from Auburn, Alabama, giving way to various types of metagranite and granitic gneisses. On earlier maps (e.g., Osborne and others, 1988), these rocks were undifferentiated and shown as Bottle Granite (revised by Steltenpohl and others, 1990, to the Farmville Metagranite). Grimes and Steltenpohl (1993), however, reported that these felsic units are characterized by a complex suite of metagranite, orthogneiss, and augen gneiss that may or may not be textural variations of the Farmville Metagranite. Although the existing data do not require that these intrusives be unrelated to the Farmville Metagranites, their gradational nature and difficulty in separating them on a geologic map are similar to the problem of distinguishing the Kowaliga Gneiss from the Zana Granite of the eastern Blue Ridge (Bieler and Deininger, 1987). The felsic intrusives of the Emuckfaw Group are similar to the Farmville Metagranite in their petrography, whole-rock geochemistry, outcrop and hand specimen appearance, textures, and their occurrence as sill-like bodies. A problem with this correlation lies with the determined isotopic ages. Felsic intrusives in the Emuckfaw Group yield Ordovician U-Pb zircon dates (Russell and others, 1987), but a Silurian Rb-Sr whole-rock age for the Kowaliga Gneiss and a Devonian Rb-Sr whole-rock age for the Zana Granite were also suggested by the work of Russell (1978). Felsic intrusives in the Auburn Gneiss (i.e., Farmville Metagranite) have yielded a Devonian Rb-Sr whole-rock age (Goldberg and Burnell, 1987). Augen gneiss of the eastern Blue Ridge, which Keefer (1992) correlated with the Zana Granite, forms the southernmost exposure along the Tallapoosa River and projects into augen gneisses of the Opelika Complex, previously mapped as Bottle Granite (Osborne and others, 1988; i.e., Farmville Metagranite).

The nature of the contact between the Loachapoka Schist and Auburn Gneiss is not understood because it is obscured by pervasive injection of granitic material and ductile shearing. Higgins and others (1988) considered a thrust to separate the Sandy Springs thrust sheet from the underlying Bill Arp thrust sheet. Sears and others (1981) thought that the granites intruded as sills along intraformational contacts within the quartzites. Higgins and others (1988) interpreted the quartzites within the Sandy Springs thrust sheet to have been thrusted upon granites in the Georgia Inner Piedmont and that the granites themselves were emplaced along thrusts. Keefer (1992) suggested a similar structural style in rocks he observed in the Stone Creek imbricate zone. Given the sheared nature of the granite margins throughout the Loachapoka Schist and Auburn Gneiss, this replication is tectonic. Steltenpohl and others

(1990) suggested that these magmas were syntectonically injected within a ductile deformation zone. The same authors suggested that melt-enhanced shearing (Hollister and Crawford, 1986) may account for features associated with the Farmville intrusions. This deformation zone would have formed during the Devonian on the basis of the 369 Ma crystallization date for the Farmville Metagranite. Given the uncertain origin of the boundary between the Loachapoka Schist and Auburn Gneiss in Alabama, Grimes and Steltenpohl (1993) suggested caution be used when considering the Loachapoka Schist and Auburn Gneiss as part of a single Opelika Complex.

The nature of the contact at the base of the Auburn Gneiss is potentially significant. As noted by Grimes and Steltenpohl (1993), we also believe that the Auburn Gneiss units are part of the eastern Blue Ridge and that the Pine Mountain basement gneiss and feldspathic schist east of the Towaliga fault are Laurentian basement-cover units. Eastern Blue Ridge rocks therefore appear to continue eastward around the Tallassee synform and are juxtaposed with Laurentian units along the east limb, a relationship that is tectonostratigraphically equivalent to that of the Hayesville-Fries fault in other parts of the orogen. This contact has been excised in our study area by the Towaliga normal fault (Schamel and others, 1980).

CONCLUSIONS

The Jacksons Gap Group diverges from the northeast trend of the Brevard fault zone south of Jacksons Gap, Alabama, and strikes southward and then eastward around the Tallassee synform. The retrogressive mylonites of the Brevard fault zone record right-slip motion (Bobyarchick and others, 1988, and references therein) and we suggest that late folds in the Jacksons Gap Group units where it diverges from the Brevard fault zone reflect drag as a result of this dextral shear. Some right-slip displacement may have been transferred to the Alexander City fault, which along its length the sense of movement has not been determined but is locally reported to be right-slip (Guthrie and Dean, 1989). Splays connecting the two structures in the Alexander City/Jacksons Gap area are shown as right-slip faults on some maps (e.g., see Osborne and others, 1988) and may have accommodated dextral motion between the two fault zones (Fig. 2). If in the future right-slip motion becomes documented along the Alexander City fault, this would be the only major dextral shear zone in the Appalachians northwest of the Brevard fault zone.

There are discrepancies in the location of the Stonewall line northeast into Georgia. Our findings show that the Stonewall line is an important fault that separates the Dadeville Complex from the Jacksons Gap Group - Loachapoka Schist and marks the top of the structurally interleaved Loachapoka Schist. Its placement northeast into Georgia has similar litho- and tectonostratigraphic implications as suggested in this study.

Our findings indicate that the Jacksons Gap Group units lying in the hinge of the Tallassee synform provide the unique opportunity to investigate the early, pre-right-slip, pre-retrograde history of this zone. Studies show (Reed and Bryant, 1964; Wielchowsky, 1983; Bobyarchick, 1983; Vauchez, 1987) that retrogressive, right-slip movements have obscured this earlier history in other parts of the orogen. The amphibolite-facies assemblage, schistosity/gneissosity, in the eastern Blue Ridge and Dadeville Complex formed during D_1, and is cut by the D_2 retrogressive mylonites. We cannot yet demonstrate whether S_1 in these lithotectonic elements is crosscut by the higher temperature mylonites in the Jacksons Gap Group in the hinge of the synform (Keefer and others, 1993; Grimes and Steltenpohl, 1993). Although the absolute timing of D_1 is not known (Steltenpohl and Kunk, 1993), M_1 must postdate emplacement of the Ordovician granitoids, which contain S_1, in both the eastern Blue Ridge and Dadeville Complex. The retrogressive D_2 structures and fabrics are attributed to Alleghanian deformation (Steltenpohl and Kunk, 1993) along the length of the Brevard fault zone (Bobyarchick and others, 1988). Workers argue the timing of the early movements resulted from either the Taconian or Acadian tectonometamorphic events (see Bobyarchick and others, 1988, for a summary). Future isotopic studies of the high temperature mylonitic fabric in the hinge zone of the Tallassee synform may provide data on the timing of earlier Brevard fault zone movements.

Our conclusions can be summarized as follows:

1) The Jacksons Gap Group diverges south out of the Brevard fault zone south of Jacksons Gap, Alabama, wraps around the hinge of the Tallassee synform to become the Loachapoka Schist. The Loachapoka Schist along the east limb of the synform continues northeastward into Georgia where it corresponds to the Sandy Springs thrust sheet of Higgins and others (1988). North of Jacksons Gap, Alabama, the Jacksons Gap Group has a northeast trend and lies within the Brevard fault zone where, near Atlanta, it is correlative with Sandy Springs Group exposed in the Sandy Springs thrust sheet of Higgins and others (1988). Consequently, the Saugahatchee quartzite of the Loachapoka Schist, the Tallassee and Devils Backbone quartzites of the Jacksons Gap Group, and the Chattahoochee Palisades Quartzite are correlative. We suggest that the name Chattahoochee Palisades Quartzite of Higgins and

others (1988) be applied to these quartzites. Similarly, we recommend that the schists of the Loachapoka Schist and Jacksons Gap Group be referred to as the Factory Shoals Formation of the Sandy Springs Group (Higgins and others, 1988).

2) Felsic gneisses, including the Kowaliga Gneiss, Zana Granite, and Farmville Metagranite, underlie or associate with the Sandy Springs Group both northwest and southeast of the hinge of the Tallassee synform. Our work suggests that the Kowaliga Gneiss and Zana Granite of the eastern Blue Ridge wrap around the hinge of the synform. These units are involved with the Stone Creek imbricate zone and continue to the northeast as lithologic units of the Auburn Gneiss, previously called the Farmville Metagranite.

3) Rocks of the Opelika Complex that are assigned to the Inner Piedmont belong to the eastern Blue Ridge. The problem of distinguishing between the Opelika Complex and eastern Blue Ridge units in north Georgia and the Carolinas has been the topic of several papers (e.g., Higgins and others, 1988; Hopson and Hatcher, 1988) and this study as well. These new tectonostratigraphic findings have significance for southern Appalachian paleogeography, particularly in their relation to the Laurentian units within the Pine Mountain window. The Towaliga fault is crucial because it marks the boundary between the eastern Blue Ridge and the Pine Mountain window in this region. Steltenpohl and Kunk (1993) have suggested that the Towaliga fault/fault zone marks the boundary between the amphibolite-facies Alleghanian tectonothermal zone to the east and an earlier deformed and metamorphosed zone to the west. The proximal occurrence of this zone to the Alleghanian suture with Gondwanan crust in the Suwannee terrane is crucial to establishing a lithospheric cross section that relates faults exposed at the earth's surface to the suture.

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ALLEGHANIAN RB-SR MINERAL AGES FROM THE INNER PIEDMONT OF SOUTHWESTERN NORTH CAROLINA

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Rb-Sr isotopic data are presented for three muscovite-bearing amphibolite facies rocks from the Inner Piedmont of southwestern North Carolina (Table 1). Rocks include garnet-mica gneiss of the lower Mill Spring complex, Sugarloaf granitic gneiss, and mylonitic Henderson Gneiss. Geological relations in the Columbus Promontory region of the Inner Piedmont are based on the field and petrologic study of Davis (1993).

Sugarloaf gneiss and the Mill Spring complex are components of the Sugarloaf Mountain thrust sheet, structurally overlying the Tumblebug Creek thrust sheet containing Henderson Gneiss (Fig. 1). A single penetrative fabric with similar orientation is recognized in both thrust sheets. The single foliation in pelitic rocks contains the assemblage muscovite + biotite + sillimanite ± garnet. Based on thermobarometry, Davis (1993) estimated temperatures of 550 to 650 °C and pressures of approximately 4 kb for the above assemblage. Davis (1993) also noted homogeneous compositional profiles in garnet, consistent with a single high-grade tectonothermal event.

Ages are obtained by regression of data for whole-rock + muscovite + feldspar ± garnet, but are essentially defined by the whole-rock - muscovite pair (Fig. 2). Calculated ages are 293, 300, and 308 Ma. Feldspars plot near whole-rock data points and appear to be in isotopic equilibrium with both whole-rock and white mica based on linear isochron relationships. This assumption may not be correct because of the similarity of the whole-rock and feldspar Rb-Sr isotopic data and the relatively low closure temperature of 300 to 400 °C for Sr in feldspar compared to muscovite (500 °C) or garnet (700 °C). The large differences in measured $^{87}Rb/^{86}Sr$ ratios between muscovite and whole-rock are reflected in the small errors associated with calculated regression ages. Muscovite Rb-Sr mineral ages have been used successfully to interpret events in other orogens, in particular the Alpine orogen. We interpret the mineral assemblage as having formed at or above

Table 1. Rb-Sr isotopic and concentration data for Inner Piedmont gneisses §.

Sample	87Rb/86Sr	87Sr/86Sr	ppm Rb	ppm Sr
CP5(3) WR	1.4113	0.724386	113.24	228.04
CP5(3) FSP	0.0975	0.718608	4.10	119.06
CP5(3) MUSC	25.576	0.830254	263.09	29.54
CP9(1) WR	1.6941	0.719364	139.69	234.23
CP9(1) FSP	0.2772	0.712424	21.97	225.03
CP9(1) MUSC	25.895	0.819568	235.51	26.09
CPSMG1 WR	4.8849	0.737672	118.24	68.88
CPSMG1 FSP	3.8491	0.734276	80.45	59.46
CPSMG1 MUSC	166.40	1.42824	427.87	7.81
CPSMG1 GT	2.3698	0.727105	3.00	3.60

§ Rb and Sr were analyzed on a VG Sector 54 8-collector mass spectrometer. Sr isotopic ratios were exponentially corrected for mass fractionation using $^{86}Sr/^{88}Sr = 0.1194$. Within-run errors (± 2σ SE) are less than 0.000025 for whole-rock and feldspars, and 0.000075 for muscovite. Repeated measurement of SRM 987 during the course of analyses yielded a $^{87}Sr/^{86}Sr$ ratio of 0.710241 ± 9 (2σ). Concentrations of Rb and Sr were determined by isotope dilution using well calibrated ^{87}Rb and ^{84}Sr spikes. Elemental concentrations are known to within 0.3 percent.

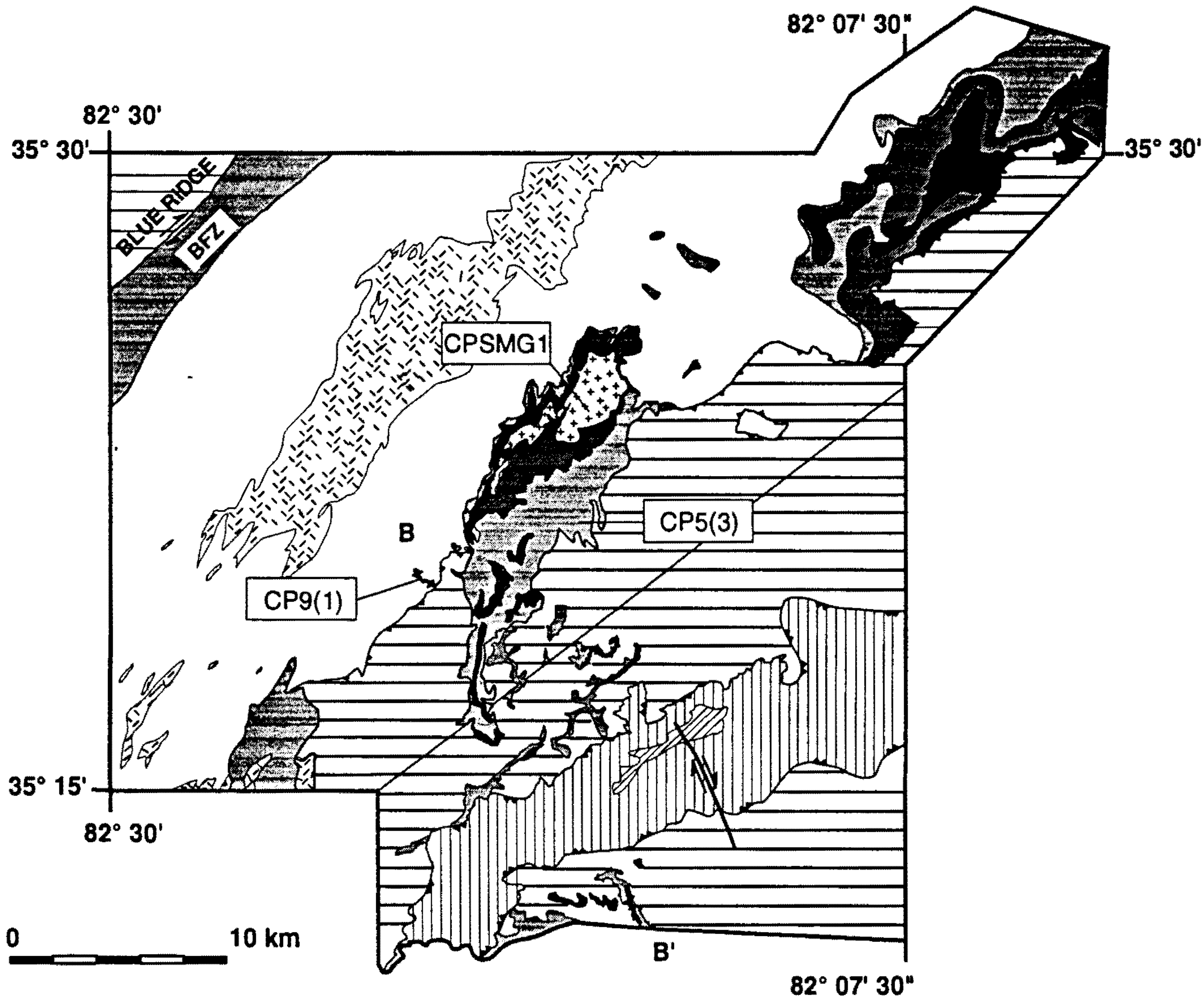


Figure 1. Simplified geologic map of the Columbus Promontory area showing the locations of the samples and analyses reported herein. Explanation of details of the rock units may be found in Davis (this guidebook).

550 °C, based on estimates from reaction curves and thermobarometry. Ages calculated from garnet -whole-rock and muscovite–whole-rock pairs in sample CPSMG1 are similar, suggesting that Sr closure at 300 Ma occurred at relatively high temperatures. In conjunction with petrographic observations and thermobarometric data, these radiometric ages indicate that moderate–to high–grade metamorphic conditions affected these thrust sheets during the Late Pennsylvanian. Further radiometric data are needed, and a U-Pb study is in progress, to confirm these preliminary Alleghanian ages.

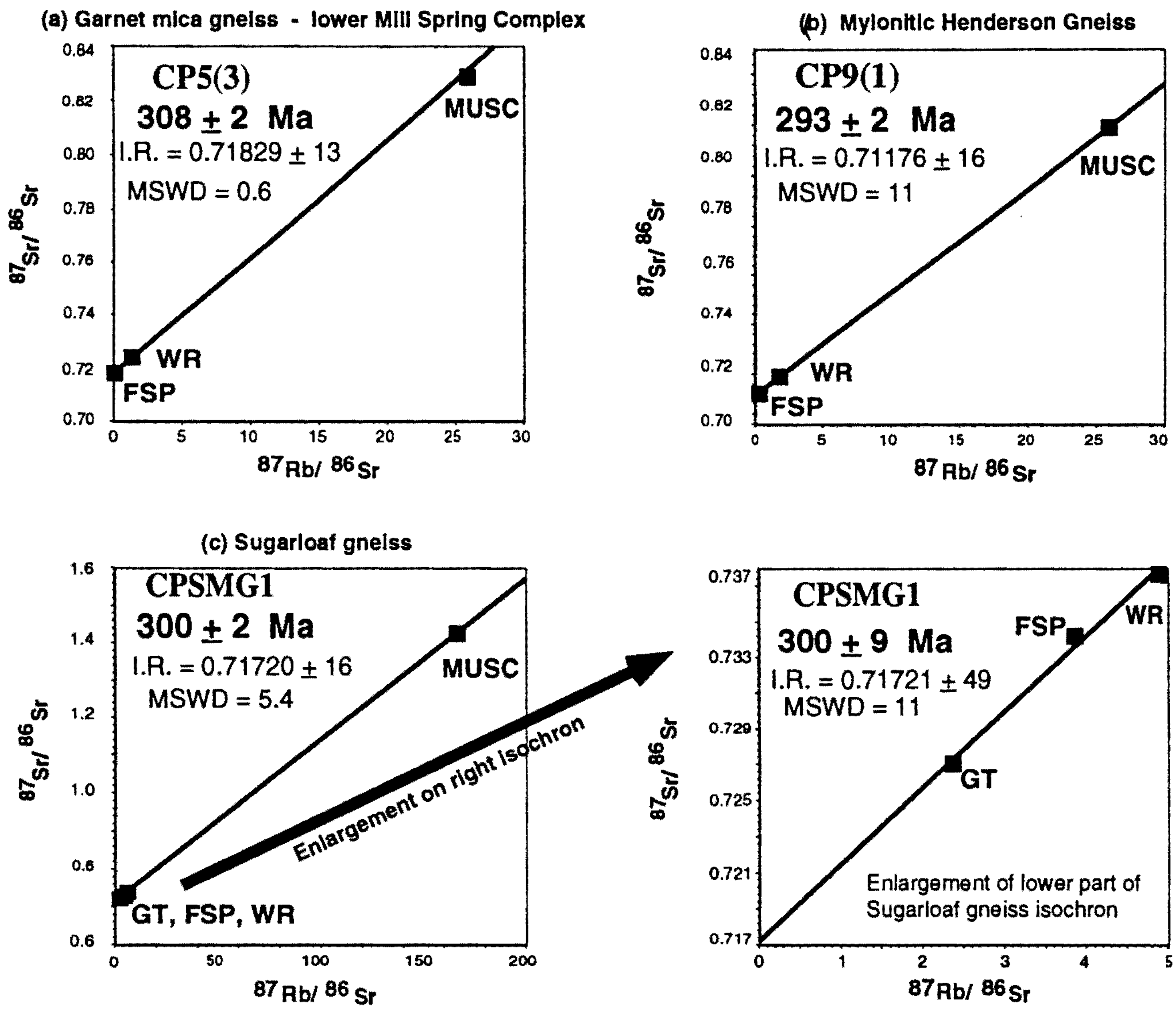


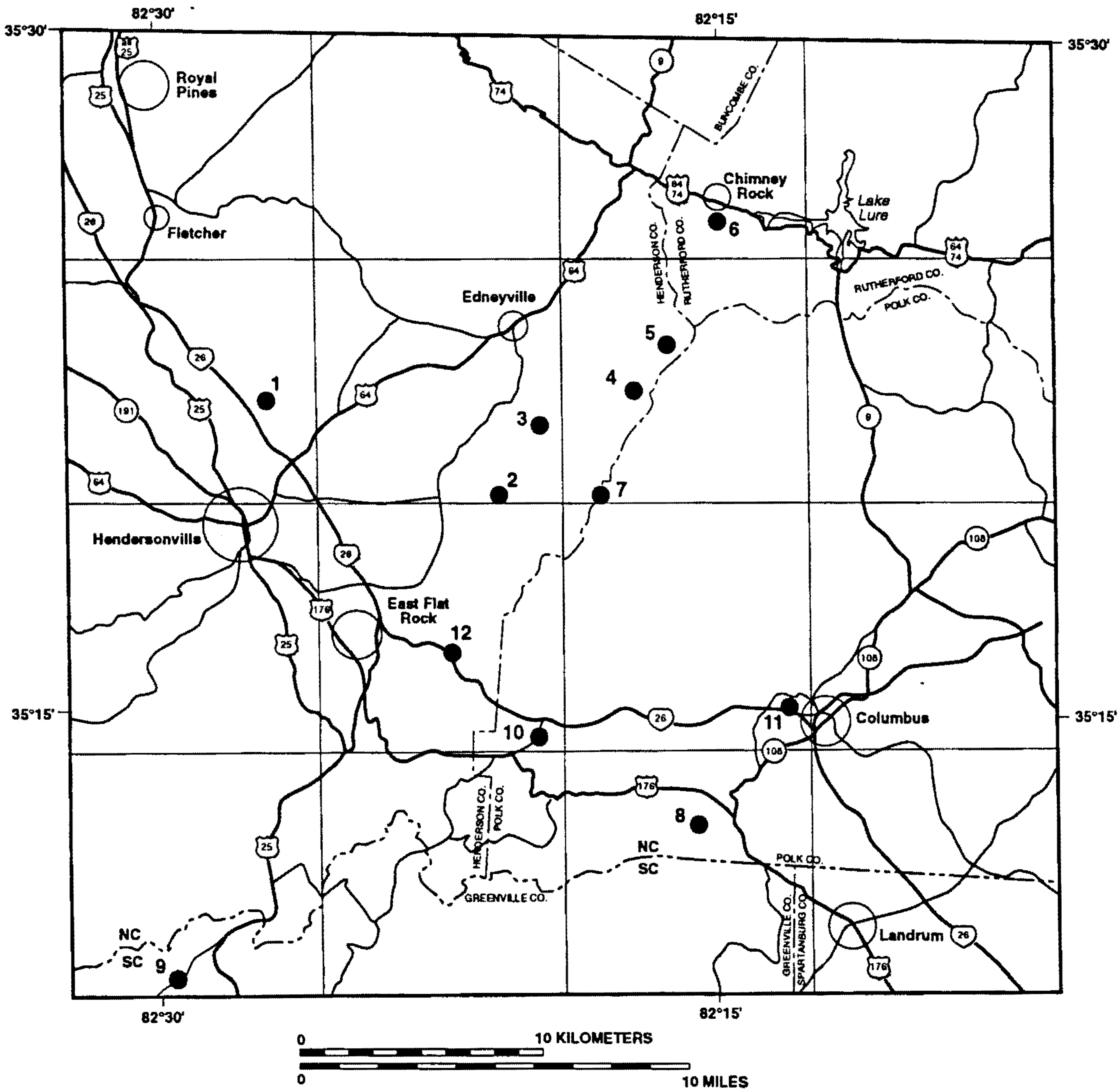
Figure 2. Rb-Sr isochrons for (a) Mill Spring Complex, (b) Henderson gneiss, and (c) Sugarloaf gneiss.

REFERENCE CITED

Davis, T. L., 1993, Geology of the Columbus Promontory, North Carolina *in* Hatcher, R. D., Jr., and Davis, T. L., editors, Studies of Inner Piedmont geology with a focus on the Columbus Promontory: Carolina Geological Society Guidebook, p. 17-44.

FIELD TRIP STOP DESCRIPTIONS

The purpose of the 1993 CGS field trip and guide is to present and discuss some of the geology worked out recently in the Columbus Promontory, and point out some of the interesting unsolved problems that remain both in bedrock and surficial geology. We have planned a series of stops to illustrate the different lithologies, tectonic styles, and rock units (Fig. S1), as well as elements of the early Mesozoic structure and late Tertiary to Holocene surficial geology.



DAY 1

STOP 1: Henderson Gneiss (Please bring hard hats)
Location: Vulcan Materials Quarry, County Road 1503 Hendersonville, NC (HQ)

The purpose of this stop is to examine the lithologies and the mesoscopic structural features within the Henderson Gneiss. The Henderson Gneiss is a critical marker unit for interpreting the sequence of development of crystalline thrusts and related deformation in the Columbus Promontory. The quarry contains numerous structural features characteristic of the Henderson Gneiss including: 1) the penetrative NE–SW striking and SE–dipping S_2 mylonitic foliation, 2) a pervasive NE–SW mineral lineation, 3) ovoid–to asymmetric and tailed feldspar augen, and 4) mesoscopic SW–vergent folds. On NW wall of the quarry is a large (~ 20 m) SW–vergent isoclinally folded quartzofeldspathic layer within the Henderson Gneiss. Also present here are more fine–grained lithologic units within the Henderson Gneiss including quartzofeldspathic veins, and dikes that have been transposed into the S_2 mylonitic foliation. **For obvious safety reasons you must stay approximately 75 feet away from the quarry walls. It will be possible to closely examine and sample the Henderson Gneiss along the road leading to the quarry floor.**

STOP 2: Tumblebug Creek Thrust Fault
Location: Tumblebug Creek , County Road 1525 near Dana, NC (CMQ)

This stop examines the thrust contact between the Henderson Gneiss and Poor Mountain Amphibolite. The Tumblebug Creek thrust emplaced Henderson Gneiss over rocks of the Poor Mountain Formation. This exposure occurs along Tumblebug Creek in part of the type area for the fault (Davis, 1993). A strongly developed NE–SW trending linear fabric is present here. Kinematic indicators suggest top–to–SW sense of shear associated with this penetrative linear fabric. Approximately 1 km to the W– NW of this exposure along Tumblebug Creek are additional excellent exposures of this contact that reveal the presence of 1) intensely mylonitized Henderson Gneiss; 2) truncation of folds within Poor Mountain Formation by the Tumblebug Creek thrust fault; and 3) folding of this contact and transposition of it into the penetrative S_2 mylonitic foliation, suggesting emplacement prior to development S_2.

STOP 3: Sugarloaf Mountain Thrust Fault
Location: Little Hungry River, County Road 1719 near Union Hill, NC (CMQ)

The thrust contact between the Henderson Gneiss and rocks of the structurally overlying Sugarloaf Mountain thrust sheet is exposed at this stop. The Sugarloaf Mountain thrust sheet was originally defined and described by Lemmon (1973) north of this locality in the Bat Cave quadrangle. The name Sugarloaf Mountain thrust sheet was suggested by Davis (1993). Lemmon originally described the sequence within the thrust sheet to include in ascending order, garnet–mica schist, amphibolite–hornblende gneiss, and lineated granitic gneiss. Lemmon also described local occurrences of marble interleaved with the garnet–mica schist at the base of World's Edge in the Bat Cave quadrangle. Recent detailed geologic mapping by Davis (1993), unpublished work of J. R. Tabor, and the work of G. Yanagihara (this guidebook) indicates that the Sugarloaf Mountain thrust sheet extends into Cliffield Mountain, Saluda, Mill Spring, Lake Lure, and Sugar Hill quadrangles. Based on this additional geologic mapping Davis (1993), redefined the Sugarloaf Mountain thrust sheet to include rocks of the Poor Mountain Formation, the Sugarloaf gneiss, and rocks of the upper Mill Spring complex. At this locality, a traverse from N–NW to S–SW along county road 1719, reveals the sequence that, in descending order, includes Poor Mountain amphibolite, Poor Mountain schist (Brevard–Poor Mountain transitional member), and Henderson Gneiss. The Sugarloaf Mountain thrust sheet is parallels to the S_2 mylonitic foliation. Along the thrust contact are asymmetric quartzofeldspathic (pegmatite) pods. Both the Henderson Gneiss and the rocks of the structurally overlying Sugarloaf Mountain thrust sheet contain a penetrative NE–SW oriented mineral stretching lineation, and related shear–sense indicators that indicate a top–to–SW sense of shear during emplacement of the Sugarloaf Mountain thrust sheet.. Also present along the contact here and elsewhere in the Columbus Promontory , the Henderson Gneiss exhibits evidence of extreme grain–size reduction.

STOP 4 (ALTERNATE): Poor Mountain Formation schist (Brevard–Poor Mountain Transitional Member), Sugarloaf Mountain Thrust Sheet
Location: Hungry River on County Road 1713 near Horace, NC (CMQ)

Stops 4 and 5 provide an opportunity to examine Poor Mountain Formation rocks within the Sugarloaf Mountain thrust sheet. Stop 4 is an exposure of the sillimanite–bearing, garnet–mica schist unit within the Sugarloaf Mountain thrust sheet. The lithology at this exposure is a quartz and muscovite–rich schist with a foliation dominated by the S_2 mylonitic foliation. Garnet and ilmenite are visible in many hand samples.

In thin section the S_2 mylonitic foliation comprises a well–developed composite–planar fabric defined by muscovite, biotite, fibrolitic sillimanite, and asymmetric garnets. Sillimanite is most commonly observed in thin section, although at several localities where this unit is exposed, sillimanite needles are readily visible in hand samples. Textural evidence indicates that sillimanite growth was synkinematic with formation of S_2. In this area the mineral stretching lineation, defined by rodded quartz, mica, and sillimanite is oriented dominantly NE–SW. This mineral lineation and composite planar fabric indicate a top–to–SW shear sense.

STOP 5 (ALTERNATE): Poor Mountain Amphibolite, Sugarloaf Mountain Thrust Sheet
Location: Hungry River on County Road 1715 near Ottanola, NC (BCQ)

This stop examines Poor Mountain Amphibolite within the Sugarloaf Mountain thrust sheet. Here the Poor Mountain Amphibolite contains the characteristic fine laminations, parallel to the S_2 mylonitic foliation, defined by alternating dark amphibole–rich and light quartzofeldspathic layers. These layers parallel the S_2 mylonitic foliation. Poor Mountain Amphibolite in the Columbus Promontory is almost identical to that observed at the type section of this unit at Poor Mountain, South Carolina described by Sloan (1907), Shufflebarger (1961) and Hatcher (1969, 1970). Lemmon (1973) also described small pods of marble within the rocks observed here and at the preceding stop that have petrographic and chemical similarity to Poor Mountain Marble described by Hatcher (1973) in South Carolina. The fine–laminated characteristic of the Poor Mountain Amphibolite produces some of the most spectacular mesoscopic–scale folds in the southern Appalachians. The Poor Mountain Amphibolite also contains a well–developed NE–SW mineral stretching lineation defined by elongated amphiboles and rodded quartz. The mineral lineation and the asymmetry of folds record top–to–SW shearing.

The high ridge to the north of this stop is Sugarloaf Mountain, which is capped by a penetratively lineated (NE–SW) granitic gneiss. This unit is the structurally highest within the Sugarloaf Mountain thrust sheet– originally described by Lemmon (1973) as the upper unit in his Sugarloaf Mountain group. Lemmon (1973) speculated on whether this unit has a sedimentary or igneous origin. Davis (1993) interpreted this unit as igneous body either intruded into or thrust (?) over Poor Mountain Formation rocks, and suggested it be mapped separately as the Sugarloaf gneiss.

STOP 6: Henderson Gneiss and Sugarloaf Mountain Thrust Fault
Location: Chimney Rock Park , US 64 , Chimney Rock, NC (BCQ)

Chimney Park provides an opportunity to examine an impressive array of geologic, geomorphic, and botanical features along with a panoramic view of the western Piedmont. The geology of Chimney Rock Park was mapped and described by Lemmon (1973). It represents an important locality where structural features within the Henderson Gneiss and the contact with the overlying Sugarloaf Mountain thrust sheet can be observed. The traverse will begin by boarding the elevator from the upper parking lot of the park to ascend 26 stories to begin the hike on the Skyline–Cliff Loop Trail. As you wait for the elevator, note the magnificent folds (SW–vergent), rotated porphyroblasts, deformed quartz–feldspar dikes, and S–C fabric in the Henderson Gneiss. We have allotted approximately 2. 5 hours to complete the loop to provide ample opportunity to examine and appreciate the many outstanding features present. Geologic descriptions of several stops included in the guide to the Skyline–Cliff Loop Trail were written by R.E. Lemmon and we recommend that field trip participants use this guide (provided in your registration packet) while walking the trail.

Several (many more than described here) geologic features are particularly interesting on the trail are important to unraveling the geologic history of this area. Near the top of the Skyline–Cliff Loop Trail (Stop 11–Exclamation Point) you will be walking on the thrust contact (Sugarloaf Mountain thrust) between the Henderson Gneiss, and Poor Mountain Schist and Amphibolite of the overlying Sugarloaf Mountain thrust sheet. Spectacular cliffs and balds produced by exfoliation of Henderson Gneiss are visible to the north and east from the trail at Exclamation Point. The Sugarloaf Mountain thrust occurs near the base of the tree line directly above the exfoliation surfaces on these ridges. The Henderson Gneiss is multiply deformed here and throughout the Columbus Promontory, and the dominant structures exposed are the pervasive D_2–D_3 structures related to thrusting in the western Inner Piedmont. Included are the pervasive S_2 mylonitic foliation, the NE–SW trending mineral lineation, mesoscopic shear zones, boudinage, and SW–vergent folds. A significant number of folds within the Henderson Gneiss are bounded by shear zones. As with other exposure of Henderson Gneiss, the majority of kinematic indicators here suggest top–to–SW shear sense. A possibly later (?) set of W to NW–vergent tight to open folds and crenulations is present here along with numerous pre–, syn–, and post– S_2 mylonitic foliation quartzo–feldspar intrusions (e.g., Grotto stop). Several locations contain all generations of structures and provide an excellent opportunity to work out crosscutting relationships.

Chimney Rock Park also contains numerous interesting geomorphic features characteristic of this part of the Inner Piedmont. Along the road to the parking lots are extensive colluvial deposits filling stream valleys and large colluvial benches on many hillsides. Some of the colluvium contains decomposed clasts attesting to the antiquity of many of the tongues of colluvium that occur downslope from the exfoliation surfaces in the western Piedmont. The upper parking lot near the tunnel entrance is situated on a former landslide composed of exfoliated Henderson Gneiss. A recent landslide scar is visible above the parking lot. Another common feature observed is talus caves by opening of sheeting and other fractures (e.g., Moonshiners Cave) produced by exfoliation of the Henderson Gneiss. Throughout the western Inner Piedmont stream valleys are controlled by jointing and fracturing. An excellent example of joint control on drainage development can be observed at Peregrine's Rest on the Skyline–Cliff Trail Loop, to the north across the Broad River where closely spaced joints provide a ready conduit for water and thus control the drainage. Similarly, several drainages are fracture controlled.

Please strictly obey the rules of Chimney Rock Park. Do not carry your rock hammers or leave the trail.

STOP 7 (ALTERNATE): Poor Mountain Formation, upper Mill Spring complex, Sugarloaf Mountain Thrust Sheet
Location: Deep Gap on Cliffield Mountain near County Road 1799 (CMQ)

The purpose of this stop is to examine internal deformation of Poor Mountain Formation and Mill Spring complex rocks within the Sugarloaf Mountain thrust sheet. The traverse begins along the Hungry River at Deep Gap and follows an unimproved road (N–NE) to a Public Service North Carolina (PSNC) gas line that crosses the top of Cliffield Mountain. Geologically, the traverse begins in biotite–granitic gneiss of the upper Mill Spring complex succeeded by the ascending sequence of Poor Mountain Formation schist, amphibolite, and quartzite. At the top of Cliffield Mountain, biotite–granitic gneiss of the upper Mill Spring complex is again encountered. This repetition in the stratigraphic succession and mesoscopic folds suggest a traverse across a macroscopic, synformal, isoclinal fold cored by rocks of the Poor Mountain Formation with upper Mill Spring complex rocks on the limbs. To the east along the gas line on Cliffield Mountain are two excellent exposures of upper Mill Spring complex rocks containing an abundance of mesoscopic D_2–D_3 structures including W and SW–vergent folds, shear zones, and W–SW oriented mineral lineations. The first exposure to be encountered occurs in an excavation along the gas line beneath the peak of Cliffield Mountain. Here is a series of D_2–D_3 folds and faults that deform quartz–rich biotite gneiss and an interlayered mafic dike of the upper Mill Spring complex. In blocks bounded by the D_2–D_3 3 structures are earlier (?) folds with various orientations.

The second exposure occurs topographically above (north and up the slope ~200m) the first exposure at the base of the crest of Cliffield Mountain. This is a 30m high exfoliation surface that contains numerous mesoscopic folds, deformed quartz–feldspar dikes, and shear zones within quartz–rich biotite gneiss of the upper Mill Spring complex. The most spectacular is a large (~ 10 m) refolded(?) quartzofeldspathic layer. At these two exposures, and in the surrounding area, mineral stretching lineations are dominantly oriented NE–SW and E–W and this area are interpreted to occur in the transition between SW–directed and W–directed shearing during emplacement of thrust sheets in the western Inner Piedmont.

STOP 8 (ALTERNATE): Poor Mountain Formation and Mill Spring complex, window in Sugarloaf Mountain thrust sheet
Location: Southern Railway tracks south of Tryon CC, Tryon, NC (SQ)

This stop is in a relatively small (~ 30m) railroad cut containing Poor Mountain Formation schist and amphibolite beneath biotite–gneiss of the lower Mill Spring complex. The structure here is interpreted as a small window that exposes Poor Mountain Formation beneath the Sugarloaf Mountain thrust sheet(See plate 1). This further indicates duplication of Poor Mountain Formation and Mill Spring complex rocks. Detailed mapping in this area reveals truncation Poor Mountain Formation rocks at the contact. This contact between the biotite gneiss of the upper Mill Spring complex and the Poor Mountain Formation rocks occurs on the east end of the railroad cut. Here is more steeply dipping S_2 mylonitic foliation, a well–developed composite planar fabric, and several large (1–2m) asymmetric quartzofeldspathic pods. These kinematic indicators and an E–W trending mineral lineation suggest a top–to–W shear sense.

END OF DAY 1 – RETURN TO HOLIDAY INN, HENDERSONVILLE, NC

DAY 2

STOP 9: Quartz Microbreccia, Cataclasite, Extension Fractures along Gap Creek Splay
Location: US 25 south at Gap Creek Road (ZQ)

Quartz microbreccia and cataclasite are well exposed along the crest of a NE–trending ridge adjacent to the Gap Creek Road (0.7 mi southwest of the intersection with Route 25) near the NC–SC state line. The resistant outcrops mark the position of the N35° E–trending Gap Creek splay, a minor fault approximately 1 mi in length that splits off the main Gap Creek fault. The latter is oriented N 55° E. In a nearby, steep roadcut at the northeast terminus of the brittle splay, biotite gneiss and augen gneiss are sheared and occur as visible clasts in microbreccia. Country rock gneiss locally display drusy quartz–coated extension fractures.

The dominant features of the quartz microbreccia and cataclasite outcrops are unusually well–developed, regularly spaced, drusy quartz–lined, open extension fractures. In the immediate area, the mean extension direction represented by normals to the dilational fractures (n=18) is 18°/S42°E, consistent with the regionally determined direction for mid–Mesozoic crustal fracturing.

STOP 10: Mill Spring thrust fault
Location: Intersection of County Roads 1140 and 1103, Saluda, NC (SQ)

The Mill Spring thrust separating mafic–rich rocks of the lower Mill Spring complex in the hanging wall from Poor Mountain Formation and upper Mill Spring rocks of the Sugarloaf Mountain thrust sheet in the footwall is exposed at stop 10. The Mill Spring thrust sheet is the structurally highest thrust sheet identified thus far in mapping of the Columbus Promontory. Our interpretation of the Mill Spring thrust sheet is based on several lines of evidence: 1) truncation of Poor Mountain Formation rocks at this exposure and in the southeast corner of the Mill Spring quadrangle; 2) rocks that define the Mill Spring thrust sheet outline a distinct outcrop pattern; 3) intense mylonitic fabrics near the base of the map unit; and 4) a pervasive suite of thrust and related structures within the lower Mill Spring complex rocks (Stop 11). Here, facing north, rocks of the Mill Spring thrust sheet are to the right (NE) and Poor Mountain Formation schist are to the left (SW). Present here is a series of imbricate thrust faults and imbricated mafic dikes also interpreted as D_2–D_3 features. Truncation of migmatitic layering and fault related folding both occur along these imbricate features. Mineral stretching lineations are oriented nearly E–W and associated kinematic indicators suggest top–to W shear sense.

STOP 11: Lower Mill Spring Complex
Location: Roadcut on westbound I26 at the exit to Columbus, NC (MSQ)

Exposed here are migmatitic mafic–rich lower Mill Spring complex rocks that constitute the Mill Spring thrust sheet. This roadcut contains a spectacular suite of mesoscopic structures characteristic of the internal deformation within the Mill Spring thrust sheet. Particularly evident at this exposure is the penetrative deformation that characterizes the emplacement related structures of the Mill Spring thrust sheet. Structures present include the penetrative S_2 mylonitic foliation, numerous shear zones, W–vergent isoclinal folds, refolded folds, and sheath folds. This exposure is completely within the domain of W–directed shearing. Mineral stretching lineations are oriented E–W and associated kinematic indicators suggest a top–to W shear sense. Within the lower Mill Spring rocks are pelitic lenses that contain sillimanite. Textural evidence indicates that sillimanite growth was synkinematic with penetrative deformation present in this exposure. Near the middle of the roadcut is a well– exposed vertical, ductile strike slip fault. Kinematic indicators suggest a sinistral sense of movement. The extent of this fault has not been determined despite detailed mapping in the area. Immediately south across I26 at the east end of the roadcut (eastbound side) are several large (~10m) asymmetric bodies interpreted as either shear zone– bounded horse block or boudins. At the E–SE end of this exposure are epidote– coated and slickensided brittle faults and fractures.

WE ASK THAT YOU NOT CROSS THE INTERSTATE AT STOPS 11 AND 12 TO EXAMINE THE EXPOSURE ON THE OTHER SIDE. ALL PERTINENT STRUCTURES CAN BE OBSERVED ON THE WESTBOUND SIDE!

STOP 12 : Upper Mill Spring Complex
Location: Roadcut on westbound I26 east of Green River bridge (CMQ)

This large roadcut contains quartz–rich, biotite gneiss, and interlayered mafic (dikes–sills) rocks of the upper Mill complex within the Sugarloaf Mountain thrust sheet. This exposure provides another excellent opportunity to examine internal deformation within the rocks of the upper Mill Spring complex. This exposure also occurs in the domain dominated by west–directed structural features. Present in this exposure is the S_2 mylonitic foliation, mesoscopic west–vergent recumbent isoclinal folds, nappe–style structures with parasitic S, M, and Z folds, and shear zones. On the NW end of the exposure are two opposing fold hinges interpreted as a cross section through a mesoscopic sheath fold; the axis of this fold parallels the E–W mineral stretching lineation that occurs in this exposure. This exposure also contains numerous mafic layers nearly oriented subparallel to the S_2 mylonitic foliation. In several places are reclined to isoclinal folds that have limbs truncated by or sheared out along the contacts of these mafic layers. Also present is evidence for later (?) open folding of the penetrative E–W mineral stretching lineation.

END OF FIELD TRIP — RETURN TO HOLIDAY INN, HENDERSONVILLE, NC

BCQ – Bat Cave quadrangle; CMQ – Cliffield Mountain quadrangle;
SQ – Saluda quadrangle; MSQ–Mill Spring quadrangle;
ZQ– Zirconia quadrangle