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Geologic Map of the Great Smoky Mountains National Park Region, Tennessee and North Carolina

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Introduction

The geology of the Great Smoky Mountain National Park (GSMNP) region of Tennessee and North Carolina was studied from 1993 to 2003 as part of a cooperative investigation with the National Park Service (NPS). This work has been compiled as a 1:100,000-scale map derived from mapping done at 1:24,000 and 1:62,500 scale. The geologic data are intended to support cooperative investigations with NPS, the development of a new soil map by the Natural Resources Conservation Service, and the All Taxa Biodiversity Inventory (http://www.discoverlifeinamerica.org/). At the request of NPS, we mapped areas previously not visited, revised the geology where stratigraphic and structural problems existed, and developed a map database for use in interdisciplinary research, land management, and interpretive programs for park visitors.

Available Products

- Map Plate: Geologic Map of the Great Smoky Mountains National Park Region, Tennessee and North Carolina [155-MB PDF]
- Explanation of Map Units [101-KB PDF]
- Microsoft Access database [26.5 MB]
- Report text [5 MB PDF]
- Report text, Section 508 compliant [5 MB PDF]
- Directory Listing of ArcInfo Shapefiles (map and park boundary, bedrock and surficial geology, faults, folds, structure, and mines and outcrops)
- Shapefiles Compressed [12.7 MB]

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GEOLOGIC MAP OF THE GREAT SMOKY MOUNTAINS
NATIONAL PARK REGION, TN/NC

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Introduction

The geology of the Great Smoky Mountain National Park (GSMNP) region of Tennessee and North Carolina was studied from 1993 to 2003 as part of a cooperative investigation of the U.S. Geological Survey (USGS) with the National Park Service (NPS). This work resulted in a 1:100,000-scale geologic map derived from mapping done at 1:24,000- and 1:62,500- scale. The geologic data are intended to support cooperative investigations with NPS, the development of a new soil map by the U.S. Natural Resource Conservation Service, and the All Taxa Biodiversity Inventory (http://www.discoverlifeinamerica.org/) (Southworth, 2001). At the request of NPS, we mapped previously unstudied areas, revised the geology where problems existed, and developed a map database for use in interdisciplinary research, land management, and interpretive programs for park visitors.

Previous Investigations

At the request of NPS, USGS geologists studied the rocks of the region from 1946 to 1954 and established the regional stratigraphic and structural framework (King and others, 1958) based on earlier work by Keith (1895, 1907). Detailed mapping at 1:24,000-scale was done by Hamilton (1961) (north central area), Hadley and Goldsmith (1963)(eastern area), King (1964)(central area), and Neuman and Nelson (1965)(western area). The previous studies were published as chapters of USGS Professional Paper 349, with geologic maps at scales of 1:24,000- and 1:62,500 (fig. 1), and they should be consulted for the extensive data on the petrology and petrography of the rocks. A general report and a geologic map compiled at 1:125,000-scale summarized the earlier work (King and others, 1968). Subsequent geologic maps of the region were published at 1:250,000-scale (Hadley and Nelson, 1971; Wiener and Mershat, 1992; Robinson and others, 1992). Some of the Mesoproterozoic rocks were studied by Cameron (1951) and Mershat and Wiener (1988), and Carter and Wiener (1999) studied similar rocks to the east of the park. The Neoproterozoic rocks were studied by Stose and Stose (1949), King

(1949), Rodgers (1953), King and others (1958), Espenshade (1963), and Lesure and others (1977). The surficial geology was studied locally by Hadley and Goldsmith (1963), King (1964), Neuman and Nelson (1965), Southworth (1995), Southworth and others, 1999), Schultz (1998; 1999), Schultz and others (2000), and summarized by Southworth and others (2003).

Studies of the surficial geology by Hamilton (1961), Hadley and Goldsmith (1963), King (1964), and Neuman and Nelson (1965), were incidental to the detailed investigations of the bedrock, although surficial geologists (James Gilluly, C.S. Denny, G.M. Richmond, J. T. Hack, and H.E. Malde) observed such deposits in the field and made insight to comments in their reports. These workers recognized that the three dominant types of surficial deposits in the region are residuum, colluvium, and alluvium, and that many units were deposited during climatic conditions very different than those of today. Although a previous small-scale map of the Park, at 1:125,000-scale (King and others, 1968), did not include surficial units, individual geologic maps of scales of 1:62,500 and 1:24,000 included these three general types of deposits. Rozanski (1943), King and Stupka (1950), King (1964), Clark (1968), Michalek (1968), Richter (1973), Reheis (1972), Torbett and Clark (1985), and Clark and Ciolkosz (1988) interpreted polygonal ground, block streams, fan deposits, and other features in the Great Smoky Mountains, as being periglacial in origin. Modern debris flows which occur in the highlands underlain by slate of the Anakeesta Formation were described by Bogucki (1970; 1976), Koch (1974), Moore (1986), and Clark (1987). Older fan deposits near Dellwood (Hadley and Goldsmith, 1963), were studied more recently by Mills (1982; 1986), and Kochel (1990). Mills and Allison (1994; 1995a and 1995b) mapped and studied similar young, intermediate, and old fan deposits in the Hazelwood, NC, area, southeast of the Park.

Methodology

Field Mapping

Bedrock and surficial geologic units were compiled on 1:24,000-scale topographic base maps having 40-ft contour intervals. Bedrock contacts from previous published maps were selectively revisited and revised where necessary. Geologic problems identified on the existing geologic maps were investigated and reinterpreted. Most new mapping was along selected field traverses along several thousands of miles of trails and roads that cross the area. Additional mapping was done along hundreds of miles off trail traverses. The discontinuous nature of bedrock units portrayed on the map is a function of the depositional interfingering of deposits, geologic structure, and poor along-strike exposure.

Surficial deposits were mapped along selected field traverses, by interpretation of aerial photographs, and by interpretation of landforms on topographic maps. Conceptual models generated from strategic field traverses were used to interpret landforms and recognize surficial units in lieu of field observations. Exposures of surficial deposits were provided by landslides and stream cut banks, as well as roadcuts, and sparse excavations. The surficial units were defined primarily by materials and by some reference to known or interpreted origin and age.

Digitization and base map

Bedrock units from the 1:24,000-scale maps were scanned and reduced to 1:100,000-scale. The units were drafted at 1:100,000-scale with ink on mylar, scanned, and edited. Surficial geologic map units were inked on mylar registered to latitude and longtitude coordinates of the 7.5-minute quadrangles, scanned, and edited. The 7.5-minute quadrangles were then mosaicked for presentation at 1:100,000-scale (Southworth and others, 2003).

The 1:100,000-scale base map is a mosaic generated by scanning the map separates of the 50-m contour interval map of most of the Knoxville, TN, and the northern part of the Fontana Dam, NC, 30- by 60- minute quadrangles. This base map was also overprinted on a 30-m digital elevation model (DEM) of the USGS National Elevation Dataset. Because the 30- by 60-minute, 50-m- contour maps were extrapolated from the 7.5-minute, 40-ft-contour maps, the polygons of 1:24,000-scale may not correspond directly with the generalized topographic contours on the 1:100,000-scale base map.

General Geologic Setting and Physiography

The study area is centered on the GSMNP in the western Blue Ridge province and includes a small part of the Tennessee Valley of the Valley and Ridge province (in the northwest), and the eastern Blue Ridge province (in the southeast) (fig. 2). The regional drainage is westward to the Tennessee River that drains into the Gulf of Mexico. The mountains rise more than 4600 ft (1500 m) above adjacent valley floors, and maximum relief of slopes, in places, reaches as much as 28 degrees. Slopes are mostly covered with soil and residuum 6 to 32 ft (2 to10 m) deep, and thick vegetation helps to stabilize the slopes.

Highlands Section of the Western Blue Ridge Province

The highlands section is underlain predominantly by Neoproterozoic metasedimentary rocks of the Snowbird and Great Smoky Group, but Mesoproterozoic basement rocks are exposed in the southeastern region in a series of tectonic windows and thrust slices (King and others, 1968). Altitudes range from about 1000 ft (305 m) along the Little Tennessee River, to about 6643 ft (2025 m) on Clingmans Dome, and steep topography is common. The highlands are bounded on the north by the Gatlinburg fault system. The southern boundary of the highlands is marked by the Hayesville fault and rocks of the eastern Blue Ridge on the southern slope of the Plott Balsams. The predominate structures in the western Blue Ridge highlands are the early Greenbriar fault and the Alum Cave synclinorium, that have been complicated by later folds and faults. The oldest structures are Mesoproterozoic foliation in the basement gneiss and Neoproterozoic synsedimentary folds in the rocks of the Ocoee Supergroup. The youngest structure is the Gatlinburg fault system that may have experienced post-Cretaceous motion (Naeser and others, 2004).

Foothills Section of the Western Blue Ridge Province

The foothills section is characterized by rolling hills having altitudes ranging from about 800 ft (244 m) along the Little Tennessee River to 3069 ft (~1000 m) on Chilhowee Mountain and 4077 ft (1243 m) on Cove Mountain. The bedrock is predominantly fine- to coarse-grained sedimentary rocks of the Neoproterozoic Walden Creek Group, fine-grained sedimentary rocks of the Neoproterozoic Snowbird Group, Lower Cambrian sandstone of the Chilhowee Group, Middle Ordovician Jonesboro Limestone, and metasandstone of the Great Smoky Group. In contrast to the higher-grade metamorphic rocks of the highlands, the rocks in the foothills section are either low- grade greenschist-facies or have not been metamorphosed. Coarse-grained and quartz-rich rocks form the high knobs, such as Webb Mountain, Shields Mountain, Green Mountain, and Chilhowee Mountain, whereas carbonate rocks and siltstones underlie the valleys and coves. The foothills are bounded on the north by the Great Smoky fault and they are bounded on the south by the Gatlinburg fault system.

The foothills can be subdivided into 3 distinct sections, from north to south; 1) the fault blocks of Chilhowee, English, and Green Mountains, 2) the northern foothills of rocks of the Walden Creek Group, and 3) the southern foothills of rocks of the Snowbird and Great Smoky Groups. The foothills are further divided east to west by the Pigeon Forge fault. The geology is different on either side of this fault.

Northern Foothills

From west to east, the northernmost foothills consist of Chilhowee Mountain, English Mountain, and Green Mountain, which are bounded by faults. Rocks of the Lower Cambrian Chilhowee Group underlie these mountains. The rocks that underlie Chilhowee Mountain and English Mountain are bounded by the Great Smoky fault on the north and lie as klippen above it. They have been overridden by rocks in the hanging-wall by the Miller Cove fault on the south. Green Mountain is in a tectonic window beneath a fault that may be either the Great Smoky fault or the Miller Cove fault.

Central Foothills

The central foothills consist of rocks of the Walden Creek Group that are bounded on the north by the Miller Cove fault and on the south by the Dunn Creek fault.

Southern Foothills

The southern foothills are underlain by rocks of the Snowbird and Great Smoky Groups. East of the Pigeon Forge fault, the northern boundary is the Dunn Creek fault.

The southern boundary is the Gatlinburg fault system. The rocks are exclusively of the Snowbird Group. West of the Pigeon Forge fault only a small area (south of Grindstone Ridge and near Line Spring) contains the Dunn Creek fault, as the later Gatlinburg fault system has overprinted the earlier fault. Here the rocks are predominately of the Great Smoky Group.

Tennessee Valley of the Valley and Ridge Province

The eastern part of the Tennessee Valley of the Valley and Ridge province is underlain by Cambrian, Ordovician, and Mississippian carbonate rocks that are interbedded with fine- and coarse-grained siliciclastic rocks, and a thin bed of Devonian and Mississippian shale. Carbonate rocks underlie the broad valleys and sandstone and shale underlie ridges, having altitudes ranging from about 900 ft (274 m) to 1400 ft (427 m) on Bays Mountain.

Rocks of the Eastern Blue Ridge

The Hayesville fault was proposed to be a major terrane boundary separating Laurentian basement and Neoproterozoic syn-rift deposits of the western Blue Ridge, from deep-water sedimentary rocks and mafic and ultramafic rocks of oceanic or mantle origin to the east (Hatcher and others, 1989). The Hayesville fault was interpreted to be a premetamorphic fault of the leading edge of an early Paleozoic accretionary wedge above an eastern-dipping subduction zone of the early phases of the Taconic orogeny (Hatcher, 1978). Recent studies of metamorphism and structure across the fault between Balsam Gap and Soco Gap along the Blue Ridge Parkway by Massey and Moecher (in press) suggests that in places, the contact is a lithologic boundary having no evidence of faulting.

Southeast of the Hayesville fault, Neoproterozoic biotite gneiss of the Ashe Metamorphic Suite (Zams) has been thrust onto Mesoproterozoic biotite gneiss (Ym) along the Holland Mountain fault that Carter and others (1999) interpreted to be a premetamorphic fault of Ordovician age. Mesoproterozoic biotite gneiss (Ym) is well-foliated, and is locally migmatitic, sillimanite-grade, biotite-quartz-plagioclase gneiss and biotite-hornblende gneiss. The leucosome-rich gneiss is interlayered and has gradational contacts with biotite-garnet gneiss, calc-silicate rock, marble, and amphibolite. This rock unit is correlated with biotite-hornblende gneiss of the Earlies Gap Gneiss of Mershat and Wiener (1988), to the northeast. The biotite-hornblende gneiss of the Earlies Gap Gneiss is here correlated with the 1117+-14 Ma biotite gneiss exposed in the Dellwood and Ela domes. The gneiss east of the Hayesville fault lacks coarse-grained muscovite seen in the rocks of the western Blue Ridge (Massey and Moecher, in press). Mesoproterozoic biotite gneiss contains mappable bodies of dunite and peridotite (Hadley and Nelson, 1971), probably remnants of ophiolites (oceanic crust) emplaced before metamorphism (Lipin, 1984). A circular body of metamorphosed dunite and peridotite (PzZud) within the biotite gneiss near Addie, NC, was mined for olivine and similar rocks in the area contain asbestos, chromite, corundum, nickel, platinum-group elements, soapstone-serpentinite, talc, and vermiculite (Robinson and others, 1992). Websterite, consisting of diopside and bronzite, is associated with the dunite body near Addie.

Biotite gneiss of the Ashe Metamorphic Suite (Zams) consists of muscovite-biotite gneiss, interlayered and gradational with metagraywacke, mica schist, amphibolite, and hornblende gneiss. The rocks are locally sulfidic and have been stained by a coating of secondary iron-oxide and iron-sulfate minerals. Like the rocks of the Ocoee Supergroup, these rocks are probably deep water sedimentary deposits, but they contain abundant lenses and bodies of amphibolite that formed by metamorphism of mafic volcanic rock. The metamorphosed rocks consist of quartz, plagioclase, biotite and muscovite. The folded rocks have been locally intruded by unfoliated dikes of vein quartz and trondhjemite.

Rocks of the Highlands of the Western Blue Ridge

Mesoproterozoic Basement Complex

Mesoproterozoic rocks in the study area occur in the southwestern-most part of the French Broad massif (Rankin and others, 1989). The Mesoproterozoic rocks are a polymetamorphic complex of paragneiss, migmatite, and orthogneiss that were locally mylonitized and partially melted during the Mesoproterozoic and Paleozoic. Orthogneiss grades into migmatitic gneiss (Cameron, 1951; Hadley and Goldsmith, 1963) that presumably was derived by partial melting and the local mobilization of the orthogneiss

(Hadley and Goldsmith, 1963; Kish and others, 1975). A suite of plutonic rocks, having the compositions of quartz monzonite, granodiorite, and granite, intrude the paragneiss and contain xenoliths of paragneiss.

Paleozoic deformation, uplift, and erosion have exposed the gneiss in 6 antiforms and along several late Paleozoic thrust faults. In the eastern part of the map area Mesoproterozoic gneiss crops out along the Cold Springs fault (to the north), and another belt of gneiss outcrops on Hurricane Mountain to the south (Hadley and Goldsmith, 1963). These structures merge to the northeast, and are herein called the Pigeon River belt. The Dellwood-Cherokee belt and the Straight Fork window (Hadley and Goldsmith, 1963) merge and herein are called the Qualla-Dellwood belt and the Cherokee-Raven Fork belt, respectively. The Qualla-Dellwood belt contains the antiforms in Maggie Valley and Dellwood, and the intervening synforms. The Ela and Bryson City domes are west of the Cherokee-Raven Fork belt. The Cove Mountain slice is a fault-bounded outlier of Mesoproterozoic rocks near Wear Cove, TN. Another small body of Mesoproterozoic rock along the Blue Ridge Parkway, northeast of Cherokee may also be a small dome (Hadley and Goldsmith, 1963).

The rocks have been subdivided into units based on petrology, foliation, and Super High Resolution Ion Microprobe (SHRIMP) U-Pb zircon geochronology (Southworth and Aleinikoff, in press). Paragneiss units include amphibolite (Ya), ultramafic rocks (Yu), hornblende-biotite gneiss (Yh), migmatitic biotite gneiss (Ym), and isolated bodies of calc-silicate granofels. Orthogneiss units include leucocratic metagranite, granitic gneiss (Yg), granodiorite (Ygd), biotite augen gneiss (Ybg), and porphyritic granitic gneiss (Ypg). Two structural/metamorphic units were recognized; 1) mylonitic gneiss derived from orthogneiss, and 2) migmatite derived from paragneiss. Ultramafic and mafic rocks, hornblende-biotite gneiss, biotite gneiss, Spring Creek Granitoid Gneiss, monzogranitic gneiss, biotite augen gneiss, and mylonitic gneiss characterize this unit within the park.

New work by Carrigan and others (2003) and Ownby and others (2004) on the crystallization and inheritance data of Mesoproterozoic rocks of the Blue Ridge in North Carolina and Tennessee was used to organize and correlate the rocks into several groups. The oldest (undated) rocks are the migmatitic gneiss group, called Group 1. Group 1 rocks contain the oldest dated meta-igneous rocks (Group 2) that have crystallization ages of 1194-1192 Ma. The Group 2 rocks were recycled during later magmatic events, as Group 3 rocks contain 1.190 and 1.180 Ma inherited zircons from Group 2 rocks. Granitic gneisses of Group 3 crystallized between 1178-1117 Ma. Orthogneisses of Group 4 crystallized between 1081-1029 Ma. A major deformation event occurred between the crystallization of Group 3 and Group 4 rocks, between 1117 and 1081 Ma.

Paragneiss

Paragneiss occurs mostly in the areas near Dellwood and the Ela dome, but some also crops out near Whittier (south end of Qualla-Dellwood belt) and at the southern end of the Bryson City dome. Paragneiss is characterized by mafic and felsic layers that may be migmatite segregations, and (or) primary compositional beds of the protolith (Hadley and Goldsmith, 1963; Kish and others, 1975). The protolith of the paragneiss has been interpreted as interstratified volcanic and sedimentary rocks (Hadley and Goldsmith, 1963). The dominant sedimentary protolith probably included argillaceous and quartzose graywacke. Rocks rich in hornblende were probably derived from iron-rich dolomitic rocks or andesitic to basaltic tuffs and calc-silicate rocks were probably derived from calcareous rocks (Hadley and Goldsmith, 1963). Amphibolite was probably derived from both intrusive rocks and mafic segregations in metamorphosed sedimentary or volcanic rocks.

The dominant paragneiss units are biotite gneiss (Yb) and hornblende-biotite gneiss (Yh). Biotite gneiss and hornblende-biotite gneiss, often migmatitic, occur together (east to west) in the Dellwood dome, the Ela dome, and in the southern part of the Bryson City dome. Amphibolite and ultramafic rocks are probably restites (oldest rock), the hornblende and biotite gneisses are migmatites of sedimentary protoliths, and the orthogneisses the product of melting (Cameron, 1951).

Ultramafic and Mafic Rocks

Mafic and ultramafic rocks occur in xenoliths within migmatites, gneiss, and granitoids. Metamorphosed ultramafic rocks (Yu) were found within biotite augen gneiss in the Pigeon River belt (Hadley and others, 1955), in the Qualla-Dellwood belt (Maggie

Valley and east of Soco Gap), near Smokemont in the Cherokee-Raven Fork belt, and in biotite gneiss and leucocratic metagranite of the Bryson City dome (Cameron, 1951; Hadley and Goldsmith, 1963). These rocks are shown on the map by “x’s”, labeled Yu. Hadley and Goldsmith (1963) described a body of ultramafic rock 1000 ft (305 m) across near the crest of Hurricane Mountain at the east border of the map area. At least four bodies of the ultramafic rocks are found immediately beneath the Greenbriar fault. These dark, rusty-weathered rocks are poorly exposed, so little is known about the contact relations. They are composed of chlorite, actinolite, tremolite, hornblende, biotite, garnet, quartz, and magnetite, and some iron-rich olivine. The high magnesium content and absence of feldspar suggests that they are altered peridotite (Hadley and Goldsmith, 1963). Two narrow belts of rock on the western margin of the Bryson City dome are massive to schistose, medium- to very coarse-grained metaperidotite composed of tremolite, anthophyllite, and chlorite that grades into biotite gneiss, hornblende schist, hornblende-biotite schist, and biotite schist (Cameron, 1951). A small body of massive, medium-grained, biotite-hornblende metagabbro in the Bryson City dome is comprised of plagioclase and clinopyroxene. This mottled rock is partly replaced by hornblende, biotite, and microcline, and may grade into biotite-hornblende gneiss and schist (Cameron, 1951). To the east, Mershat and Wiener (1988) and Carter and Wiener (1999) described mafic granulite (7 to 36 per cent hornblende, mostly retrograded to amphibolite and biotite schist), massive amphibolite, talc-bearing amphibolite, and talc bodies within paragneiss and orthogneiss.

Amphibolite

Amphibolite (Ya) occurs in dark pods and layers, inches (cm’s) - to 10 ft (several m)-thick, within the hornblende-biotite gneiss, biotite gneiss, and migmatite. Mesoscopic and map-scale bodies are within biotite gneiss and biotite augen gneiss. The largest body of amphibolite is near Dellwood and measures 900 ft (274 m) wide and 2000 ft (610 m) long. A slab-like body of amphibolite 5 ft (1.5 m) thick and 25 ft (7.5 m) long within biotite granite gneiss north of Big Cove was described by Hadley and Goldsmith (1963). The granoblastic, fine- to medium- grained amphibolite is composed of green to brownish-green hornblende and subordinate amounts of plagioclase, as well as biotite, quartz, and minor amounts of epidote, sphene, ilmenite, apatite, and almandine. A well-developed foliation is defined by streaks and lens of aligned hornblende, garnet, pyroxene, and biotite (retrograded from hornblende). Amphibolite in biotite gneiss at the south end of Ela dome consists of coarse-grained brownish-green hornblende (60 percent) and brown biotite (20 per cent). Clinopyroxene in the cores of hornblende grains suggests it may have originated as an intrusion of diorite or gabbro (Kish and others, 1975). Abundant amphibolite float occurs at the extreme north end of the Cherokee-Raven Fork belt.

Hornblende-biotite gneiss

Hornblende-biotite gneiss (Yh) includes dark hornblende-biotite gneiss, biotite-hornblende gneiss, quartz-bearing hornblende gneiss, layered amphibolite containing little or no biotite, and granoblastic calc-silicate granofels (Hadley and Goldsmith, 1963). The metamorphic layers are defined by thin selvages of subparallel olive-brown biotite and hornblende crystals that constitute as much as 20 to 50 percent of the rock, and leucosome rich in garnet. Thin layers and pods of amphibolite composed of plagioclase and hornblende (more than 50 percent) are within the hornblende-biotite gneiss. Hornblende-biotite gneiss is well exposed on Purchase Knob near Dellwood and on the east side of the Ela dome.

Migmatitic biotite gneiss

Migmatitic biotite gneiss (Ym) is light- to medium- gray, medium- to coarse- grained gneiss that contains biotite schist, muscovite-biotite gneiss and associated schist, and mica-feldspar quartz gneiss (Hadley and Goldsmith, 1963). Biotite gneiss is composed largely of quartz, plagioclase, and reddish-brown to olive-brown biotite (2 to 30 percent), with subordinate amounts of potassium feldspar and muscovite. The gneiss ranges from weakly- to well-foliated porphyroblastic gneiss, the foliation being defined by subparallel flakes and streaks of biotite, and lineated quartz and feldspar. Muscovite-biotite gneiss and associated schist occur as layers intercalated with quartzose gneiss within the biotite gneiss and seldom within the hornblende gneiss. The muscovite-biotite gneiss is light- to medium- gray, medium-grained, inequigranular and composed of quartz, calcic-oligoclase, and biotite, and less abundant muscovite, almandine, and kyanite. Mylonitic foliation is defined by muscovite and biotite concentrations around lenses of quartz and oligoclase. The muscovite-biotite gneiss and associated schist closely resemble the kyanite-bearing pelitic rocks of the Neoproterozoic Great Smoky Group. Mica-feldspar-quartz gneiss is abundantly interlayered with muscovite-biotite gneiss but is more quartzose and less micaceous. Biotite gneiss underlies much of Ela dome and Dellwood dome, and is found in the southern part of the Bryson City dome.

Migmatitic biotite gneiss is well exposed along the roadcuts at the south end of Ela dome. The migmatite was likely derived from biotite gneiss and biotite schist containing leucosomes of leucocratic metagranite. The leucocratic metagranite is a light gray to white, fine to medium grain, sugary textured, granitoid composed of microcline, microperthite, oligoclase, and quartz, and distinctive 0.5 inch (2 cm)-wide dark clots of biotite, muscovite and garnet. It ranges in composition from granite to granodiorite. The rock is massive with weak foliation defined by aligned clots of biotite. In the southern parts of the Bryson City and Ela domes, it constitutes the leucosomes in migmatitic biotite gneiss, and as was noted by Cameron (1951), the rocks are "mixed" together at all scales. Dikes of leucocratic granite, fine-grained granite, granite porphyry, and porphyritic quartz monzonite have intruded the leucocratic granite gneiss throughout the Bryson City dome. A sample from the western end of the Highway 74 cut at the south end of Ela dome yielded a SHRIMP U-Pb zircon age of 1194 Ma (Southworth and Aleinikoff, in press), therefore, this is the oldest dated rock in the western Blue Ridge. The migmatitic biotite gneiss is even older, however, as based on cross-cutting relations.

In the eastern part of the study area, younger plutonic rocks contain inclusions of migmatitic paragneiss (Hadley and Goldsmith, 1963). The composition of the granitic leucosomes is similar to that of the host rock, but varies with locality. This suggests that they were derived from a local source (anatexis) rather than from a common magma (Hadley and Goldsmith, 1963). In the Cherokee-Raven Fork belt and the Qualla-Dellwood belt west of Soco Gap, finer-grained biotite-bearing quartzofeldspathic gneiss and quartz-rich mylonitic gneiss are locally intercalated with (and locally contain inclusions of) metasandstone, quartz-feldspar-biotite-muscovite schist, amphibolite, biotite schist, and ultramafic rocks (Hadley and Goldsmith, 1963). Orthogneisses that have intruded the biotite gneiss have yielded a SHRIMP U-Pb zircon age of 1117 Ma (Southworth and Aleinikoff, in press).

Migmatitic biotite gneiss is correlated with rocks mapped to the east by Carter and Wiener (1999) as Earlies Gap Biotite Gneiss, an interlayered sequence of biotite granitic gneiss and biotite gneiss, locally interlayered with amphibolite and felsic gneiss. Quartz-feldspar leucocratic neosomes cross-cut the gneissic foliation and are sheared, folded, and mylonitized (Mershat and Wiener, 1988). Sandymush Felsic Gneiss is an interlayered sequence of granite gneiss, biotite granite gneiss, quartz diorite gneiss, and biotite gneiss, locally interlayered with amphibolite and sparse calc-silicate granofels (Mershat and Wiener, 1988).

Orthogneiss

Spring Creek Granitoid Gneiss

Spring Creek Granitoid Gneiss (Ysg) that has a protomylonitic to mylonitic foliation has been thrust onto mylonitic monzogranitic gneiss (Ymp) east of the Pigeon River (Carter and Wiener, 1999). It is a heterogenous meta-igneous unit dominated by biotite granite gneiss interlayered with biotite granodiorite gneiss, tonalitic gneiss, quartz monzodiorite gneiss, amphibolite, biotite gneiss, and biotite schist. A recent U-Pb SHRIMP analysis of zircon from a granitoid from this unit yielded an age of 1178 Ma (P. Berquist, Vanderbilt, 2004, oral comm.).

Granitic Gneiss

Granitic gneiss (Yg) is a well-foliated biotite granitic gneiss that occupies most of the Bryson City dome, where it is well exposed north and south of the Tuckaseegee River. Foliation is defined by planar blebs and streaks of biotite. Granitic gneiss is composed mostly of microcline, microperthite, plagioclase, quartz, biotite, and muscovite, that ranges in composition from granite to granodiorite. Locally, dikes of massive porphyritic quartz monzonite have intruded it (Cameron, 1951). The granitic gneiss yielded a SHRIMP U-Pb zircon age of 1163 Ma (Southworth and Aleinikoff, in press).

Similar rocks ranging from quartz diorite to granite in composition (Hadley and Goldsmith, 1963) crop out at the north end of the Qualla-Dellwood belt within the biotite gneiss and hornblende-biotite gneiss. It is medium- to coarse-grained, massive to foliated, with coarse augen of potassium feldspar, in a matrix of microcline, quartz, oligoclase, olive-brown biotite and muscovite. In some places, granitic gneiss appears to grade into and intertongue with biotite gneiss, but elsewhere it cuts sharply across it (Hadley and Goldsmith, 1963). Granitic gneiss from a new road cut in Cataloochee Estates, near the contact between augen gneiss granitoid gneiss as mapped by Hadley and Goldsmith (1963), yielded a SHRIMP U-Pb zircon age of 1168 Ma (Southworth and Aleinikoff, in press).

Monzogranite Gneiss

Monzogranite Gneiss (Ymp) is mylonitic and is characterized by pink and purple feldspar and grayish-blue quartz grains (Carter and Wiener, 1999). The protomylonitic to mylonitic phases of this unit were mapped from east of the Pigeon River westward into the Cataloochee area. A granitoid from this belt of rock yielded a preliminary U-Pb zircon age of 1148 Ma (P. Berquist, Vanderbilt, 2004, oral comm.).

Porphyritic Granite

Porphyritic granite (Ypg) is coarse-grained and has a rapikivi texture defined by 1 in (3-cm) long porphyroblasts of potassium feldspar. Sparse biotite (0 to 5 percent) and the coarse augen distinguish this unit from the biotite augen gneiss. Porphyritic granite intrudes biotite gneiss in the center of Ela dome. Porphyritic granite from the roadcut north of the bridge that crosses the Tuckaseegee River yielded a SHRIMP U-Pb zircon age of 1056 Ma (Southworth and Aleinikoff, in press).

Granodiorite

Granodiorite (Ygd) is a weakly foliated to massive, medium- to coarse-grained, light- and dark-brown rock mottled by light green cm-wide clots of quartz, feldspar, and orthopyroxene. It occurs between the Thunderhead Sandstone (Zt) and Pigeon Siltstone (Zp) in a lens bounded by thrust faults on the west side of Cove Mountain just south of Wear Cove. The granulite-facies rock has been cut by brittle faults and fractures formed during the Paleozoic greenschist-facies metamorphism and deformation. The eastern part of the rock body contains hornblende-biotite gneiss (Yh). Granodiorite from a road cut east of King Hollow Branch yielded a SHRIMP U-Pb zircon age of 1040 Ma (Southworth and Aleinikoff, in press). This rock is lithologically correlated with the Max Patch Granite dated at 1050 Ma by Calvin Miller, Vanderbilt University (Carl Mershat, N.C. Geological Survey, 2004, oral comm.).

Biotite Augen Gneiss

Biotite augen gneiss (Ybg) is dark, mottled, coarse grained and consists of prominent white and pink microcline and oligoclase augen in a matrix of foliated quartz and plagioclase. Aggregates of biotite constitute as much as 4 to 24 percent of the rock. The augen are porphyroclastic rods and aggregates of feldspar and quartz several inches (3 cm) long. Its bulk composition is granite to quartz monzonite. This rock has been recognized from the Pigeon River west to the Cherokee-Raven Fork belt, and is also found in the Qualla-Dellwood belt. Road cuts on State Route 19 between Cherokee and Soco Gap have exposed biotite augen gneiss containing xenoliths of biotite gneiss and metagraywacke (Hadley and Goldsmith, 1963). Leucocratic granite and aplite have intruded the biotite augen gneiss, especially near the west margin of the Cherokee-Raven Fork belt along Oconaluftee River and near Cataloochee in the Pigeon River belt. Hadley and Goldsmith (1963) referred to these rocks as the “Ravensford body”. They are L-S tectonites having distinctive linear rods and knots of feldpar porphyroclasts that resulted from mylonitization, probably during Paleozoic deformation. Biotite augen gneiss on the north side of the bridge over Raven Fork at the head of Big Cove yielded a SHRIMP U-Pb zircon age of 1029 Ma (Southworth and Aleinikoff, in press), making it one of the youngest dated Mesoproterozoic rocks in the Blue Ridge.

Mylonitic Gneiss

Mylonitic gneiss (Ymg) is a strongly foliated and lineated biotite-rich quartzofeldspathic gneiss that ranges from 66 to 200 ft (20 to 60 m) thick along the contact of the rocks of the Great Smoky Group on the north and northwest margin of the Bryson City dome. Cameron (1951) recognized the sheared nature of the rocks and called them the “border gneiss”. Kish and others (1975) described the rock as a recrystallized protomylonite and suggested that the rocks may be related to a fault that has placed the younger cover rocks above the older basement rocks. The mylonitic gneiss has pink feldspar and blue quartz and resembles the mylonitic rock of the monzogranitic gneiss (Ymg) seen east of the Pigeon River.

Neoproterozoic Ocoee Supergroup

Introduction

Our current understanding of the Neoproterozoic stratigraphy of the region is based on King and others (1958), who subdivided these rocks into the Snowbird, Great Smoky, and Walden Creek Groups. They are restricted to the southern Appalachians from North Carolina south to Georgia, and are one of the largest Neoproterozoic rock assemblages in North America. Several "unclassified rock formations" (King and others, 1958), as well as rocks assigned to different formations by different authors working in different parts of the GSMNP region, were revised on our map.

Stratigraphic relations between the three Groups of rocks are not clear. In the east central part of the map area, Montes and Hatcher (1999) suggested that rocks of the Great Smoky Group may be conformable above rocks of the Snowbird Group. Northeast of the map area, Oriel (1950) interpreted the Pigeon Siltstone of the Snowbird Group to be overlain conformably by Wilhite Formation of the Walden Creek Group, with no intervening Great Smoky Group. To the southwest of the map area, rocks assigned to the Dean Formation of the Great Smoky Group grade upward into metasiltstone of the Wilhite Formation of the Walden Creek Group (Thigpen and Hatcher, 2004).

Snowbird Group

Introduction

The Snowbird Group consists of six formations. In ascending order, they are the Wading Branch Formation, Longarm Quartzite, Roaring Fork Sandstone, Pigeon Siltstone, Metcalf Phyllite, and Rich Butt Sandstone. The Wading Branch Formation (Zwb) and Longarm Quartzite (Zl) are exposed mostly in the southeastern part of the highlands where together they range from a veneer to about 2000 ft (610 m) thick (King and others, 1968). The Roaring Fork Sandstone (Zrf) and Pigeon Siltstone (Zp) are exposed mostly in the northeastern part of the highlands and southern foothills where both are as much as 17,000 ft (~5.2 km) thick (King and others, 1968).

These rocks are lithologically similar, intertongue and are current bedded. Rocks of the Wading Branch Formation grade up into the overlying Longarm Quartzite, which grades up into the Roaring Fork Sandstone, which in turn grades up into Pigeon Siltstone. Rich Butt Sandstone also grades upward into the Pigeon Siltstone. Locally, rocks of either the Wading Branch Formation, Longarm Quartzite, and Roaring Fork Sandstone unconformably overlie Mesoproterozoic basement gneisses. The rocks of the lower formations do not everywhere occur together. West of the Pigeon Forge fault, Metcalf Phyllite (Zm) is tectonized Pigeon Siltstone (King and others, 1968); it therefore, is part of the Snowbird Group.

Wading Branch Formation

The Wading Branch Formation (Zwb) consists of heterogeneous light- and dark-colored sedimentary rocks that rest unconformably on Mesoproterozoic gneiss. The type locality is Wading Branch Ridge northwest of Walters Lake (King and others, 1958). The basal rocks include phyllite, sandy slate, quartz-sericite-muscovite schist, and quartz-pebble conglomerate. The fine-grained rocks are either green because of chlorite, are green-speckled by small porphyroblasts of chloritoid, garnet, or magnetite, or are silver because of muscovite. The fine-grained rocks range from 10 to 1500 ft (3 to 458 m) thick, and grade upward into and become interbedded with coarse pebbly feldspathic metasandstone that contains lenses of quartz pebble conglomerate 1 to 3 ft (0.3 to 1 m) thick. The rocks are poorly sorted. Individual beds range from a few inches (4 cm) to several ft (~1 m) thick. Characteristic outcrops can be seen along the Pigeon River just west of Harmon Den.

Longarm Quartzite

Longarm Quartzite (Zl) is a sequence of predominantly light-colored, medium-grained, current-bedded quartzite and arkose that range from about 50 ft (15 m) to as much as 5000 ft (1525 m) thick. The type locality is along the Pigeon River at the north end of Longarm Mountain, where the resistant rocks underlie ridges. Longarm Quartzite is light-colored and has cross-beds that are pink, blue, and purple. The quartzite is interbedded and gradational into dark, fine-grained sandstone in the upper part. Where sandstone eventually predominates, the unit is called the Roaring Fork Sandstone. Sparse current ripples and slump folds suggest that sediment transport had been to the northwest. It weathers to a distinctive light-colored sandy soil. Characteristic outcrops can be seen along Ravenfork. The quartzite is a good dimension stone that was quarried near Ravens Ford and used in construction of the Oconaluftee Ranger Station, Headquarters, and other park structures.

Roaring Fork Sandstone

The Roaring Fork Sandstone (Zrf), named for the exposures along Roaring Fork southeast of Gatlinburg, is medium- to fine-grained, greenish-gray metasandstone interbedded with laminated metasiltstone and phyllitic to slaty metasiltstone. Prominent beds of fine-grained feldspathic metasandstone (Zrfs) in the Roaring Fork were mapped south of Gatlinburg (King, 1964). Fresh rocks are green because of chlorite, and weather to brown and yellow clayey soil. The lower part of the formation contains light-colored quartzite beds in transitional areas adjacent to the underlying Longarm Quartzite. It also intertongues laterally with the Longarm Quartzite. Thick-bedded feldspathic sandstone occurs in beds ranging from 1 to 80 ft (0.3 to 24 m) thick, although most are commonly 5 to 10 ft (1.5 to 3 m) thick. These beds contain sharply truncated cross beds and large scale slump folds (Hadley and Goldsmith, 1963). Lenticular laminated metasiltstones contain ripple marks and soft-sediment folds. The upper part of the Roaring Fork Sandstone contains muscovite-quartz phyllite, slate containing porphyroblasts of biotite, illmenite, and chloritoid, and abundant vein quartz. The formation grades upward into the Pigeon Siltstone. The Roaring Fork Sandstone ranges in thickness from several hundred feet (100 m) in the southeast to 7000 ft (2135 m) in the type area. Characteristic outcrops can also be seen along the Pigeon River south of Hartford, and southeast of Waterville. Accessible outcrops can be seen along the Roaring Fork Motor Trail and the Little Pigeon River on the road to Greenbriar Cove.

Pigeon Siltstone

The proportion of metasiltstone in the upper part of the Roaring Fork Sandstone increases upward to a point where metasiltstone predominates. These rocks are called the Pigeon Siltstone, named for the exposures along the Little Pigeon River. The Pigeon Siltstone (Zp) consists of very uniform, massive, greenish-gray metasiltstone characterized by dark- and light- colored laminae. It ranges in thickness from 10,000 to 15,000 ft (3050 to 4575 m). In the northern region, the rocks are green because of chlorite. To the south, they are brown because of biotite. The rocks weather to yellow and reddish-brown clayey soil as seen in the road cuts east of Gatlinburg. Everywhere the rocks have a pervasive cleavage, often at a high angle to bedding. The massive metasiltstone beds range in thickness from 1 to 30 ft (0.3 to 9 m). Current bedding and ripples, convolute bedding, load casts, slump folds and scour and fill structures are common. No beds are graded (Hamilton, 1961). The metasiltstone contains very fine-grained sand of quartz and feldspar that increase in abundance to the east-southeastward area of outcrop (Hamilton, 1961). Interbedded fine-grained feldspathic metasandstone beds (Zps), about 5 ft (1.5 m) thick east of Pigeon Forge, are similar to those in the underlying Roaring Fork Sandstone (King, 1964). The Pigeon Siltstone thins and coarsens eastward, and it intertongues with the Roaring Fork Sandstone. The upper-most rocks of the Pigeon Siltstone are overlain gradationally by the Rich Butt Sandstone at the east end of the park. Near the Dunn Creek fault, the metasiltstone contains abundant laminae of brown-stained iron-bearing ankerite and calcite. Near Bird Creek and the Dunn Creek fault it contains clasts of carbonate rocks intercalated with calcareous siltstone. These rocks are interpreted as slump blocks containing load structures. Characteristic outcrops of Pigeon Siltstone can be seen along the Little Pigeon River.

Metcalf Phyllite

The type locality of the Metcalf Phyllite (Zm) is Metcalf Bottoms on the Little River (King and others, 1958). King (1964) recognized that the Metcalf shared many lithologic characteristics with the Pigeon Siltstone. He suggested that the rocks were probably equivalent, although the Metcalf was well foliated, much contorted, and pervasively sheared over wide areas. Chemical analyses of the Metcalf Phyllite and Pigeon Siltstone are similar, therefore they were differentiated on the map primarily because the Metacalf was much more deformed (King, 1964).

In this study, rocks of the Pigeon Siltstone were traced west of the Pigeon Forge fault to Cove Mountain, where shearing along faults converted the finely-laminated dull green slaty-cleaved Pigeon metasiltstone into lustrous gray-green, sericite- and chlorite-rich phyllite containing several cleavages, that characterizes the Metcalf Phyllite. Along the Little River, well-developed shear-band cleavage (King, 1964) indicates top-to-the northwest motion. ^{39}Ar-^{40}Ar geochronology suggests that the white-mica grew in the shear-band cleavage about 350 Ma (M.J. Kunk, USGS, 2004, oral comm.). Early laminae (beds?) are perpendicular to northeast striking faults, but the penetrative cleavage and shear-band cleavage is parallel to faults. Where strain was less severe, laminated metasiltstone has been preserved, and it resembles Pigeon Siltstone. Characteristic outcrops can be seen along Little River from Metcalf Bottoms west to Cades Cove and along the road to Tremont and Tuckaleechee Cove. Several klippen of Metcalf Phyllite underlie hills in Tuckaleechee Cove where the Great Smoky fault has juxtaposed them onto the Ordovician rocks.

Rich Butt Sandstone

Coarse-grained rocks similar to rocks of both the Snowbird and Great Smoky Groups were not classified by King and others (1958). Here they are considered to belong to the Snowbird Group. The rocks on Rich Butt Mountain, a spur on the north side of Mount Cammerer, were termed Rich Butt Sandstone (Zr) by King and others (1958). The Rich Butt Sandstone intertongues laterally with, and grades up into, the Pigeon Siltstone. Rich Butt Sandstone consists of light-colored, medium- to fine-grained feldspathic metasandstone and conglomeratic arkose interbedded with thinly-laminated dark-colored metasiltstone. The metasandstone ranges from 1 to 20 ft (0.3 to 6 m) thick and is graded. It contains fewer current- bedding features than the underlying rocks. The metasandstone beds show graded bedding and locally contain fragments of the dark-colored metasiltstone. Local beds of ankeritic metasandstone and carbonaceous and sulfidic metasiltstone southeast of Big Creek resemble rocks of the Anakeesta Formation. Rocks on Webb Mountain and Big Ridge are also probably equivalent to the Rich Butt

Sandstone (Hamilton, 1961). At the east end of the park, the Rich Butt Sandstone is as much as 4000 ft (1220 m) thick.

In the southern foothills, there are two conspicuous subunits in the Rich Butt Sandstone (Hamilton, 1961). Thinly-bedded, fine-grained feldspathic metasandstones and laminated metasiltstones (Zrs), 600 to 1000 ft (183 to 305 m) thick characterize the lower part of the Rich Butt Sandstone. They can be seen along Warden Branch near Webb Mountain and Indian Camp Creek near Big Ridge. The upper unit is composed of thick, graded-beds of coarse-grained feldspathic metasandstone and dark-colored metasiltstone (Zr) and is about 3000 ft (915 m) thick. These rocks are best exposed along the west fork of Jones Branch at Webb Mountain and along Indian Camp Creek near Big Ridge. They contain detrital tourmaline similar to the rocks of the Great Smoky Group.

The composition, accessory minerals, and bed forms suggest that the Rich Butt Sandstone constitutes the vertical transition between the rocks of the Snowbird and Great Smoky Groups. Elsewhere, it has been mostly truncated by faults. Characteristic outcrops can be seen on the southeast side of Big Creek near Mount Sterling.

Great Smoky Group

Introduction

King and others (1958) defined the Great Smoky Group to consist of, in ascending order, the Elkmont Sandstone, Thunderhead Sandstone, and Anakeesta Formation, above the Greenbriar fault. The formational subdivisions were problematic and somewhat arbitrary because of the complex interbedding of coarse- and fine-grained rocks, their complex structural and metamorphic history, the poor exposure, and the difficult terrain. Rocks in North Carolina are structurally and stratigraphically above rocks in Tennessee. They have different names even though the rocks are lithologically somewhat similar. Throughout the Great Smoky Group, dominantly coarse-grained, clastic rocks are interbedded in the lower half with dark-colored, fine-grained rocks, and in the upper half with light- and dark-colored schist. The coarse-grained clastic rocks contain distinctive subspherical concretions.

The stratigraphic position of the Cades Sandstone has always been uncertain (King and others, 1958) as all its contacts are faults. Based on lithologic similarity, we assign it to the Great Smoky Group, and interpret it to be a facies of the Thunderhead Sandstone. The base of the Elkmont Sandstone is fault-bounded but the top is gradational into rocks of the overlying Thunderhead Sandstone (King, 1964). The transition can be either rapid as seen from Greenbriar Pinnacle west to Brushy Mountain (Hadley and Goldsmith, 1963), or gradual as in the western part of the study area (Neuman and Nelson, 1965). The upper part of the Thunderhead Sandstone grades into the overlying Anakeesta Formation (Hadley and Goldsmith, 1963). Dark-colored metasiltstone beds increase in abundance and become thicker toward the base of the Anakeesta, as seen on Mount Le Conte and near Mount Guyot. Rocks of the Thunderhead Sandstone grade into rocks of the Copperhill Formation where rocks of the Anakeesta Formation are absent. The Copperhill Formation contains graphitic and sulfidic slaty rocks in its lower part whereas the upper part contains muscovite schist. The base of the Wehutty Formation is sharp and is marked by graphitic and sulfidic rocks that overly metagraywacke of the Copperhill Formation. Rocks of the Grassy Branch Formation are conformable and gradational above rocks of the Wehutty Formation. Rocks of the Ammons Formation grade above rocks of the Grassy Branch Formation. The Ammons Formation grades up into rocks of both the Dean and Nantahala Formations (Mohr, 1973).

Elkmont Sandstone

The Elkmont Sandstone (Ze), named for exposures in the community of Elkmont, constitutes the base of the Great Smoky Group (King and others, 1958), in the sequence above the Greenbriar fault. Most of the Elkmont Sandstone is composed of feldspathic metasandstone interbedded with dark-colored metasiltstone (Ze). The metasandstone is typically fine-grained, thin-bedded, and is rusty where weathered. Some coarse-grained metasandstone and metaconglomerate (Zes) was mapped by King (1964) near Elkmont, and similar rocks occur throughout the formation. Elkmont Sandstone occurs from Greenbriar Pinnacle westward to Chilhowee Lake, and from Cove Mountain south to the top of the mountain south of Cades Cove. The Elkmont Sandstone holds up the steep mountain above Cades Cove although it is poorly exposed away from creeks, rivers, and roadcuts. It's range in thickness is from about 2000 ft (610 m) beneath Greenbriar Pinnacle, to about 5000 ft (1525 m) at Elkmont, and to about 9000 ft (2745 m) south of

Cades Cove. The base of the formation has been truncated by the Greenbriar and Gatlinburg faults southwest of Cades Cove. Southwest of Thunderhead Mountain to beyond Calderwood Lake, the massive, thick-bedded nature of the Thunderhead Sandstone diminishes, making the contact between the Elkmont and Thunderhead Sandstones approximate at best. Characteristic outcrops of Elkmont Sandstone are best seen along the Little River.

Thunderhead Sandstone

Named for exposures on Thunderhead Mountain (King and others, 1958), the Thunderhead Sandstone (Zt) consists of massive, thickly-bedded, fine-grained conglomerate and graded beds of coarse-grained metasandstone that are interbedded with dark-colored metasiltstone. The principle rock type is light- to medium- gray metasandstone composed of colorless quartz, blue quartz, and white potassium feldspar, commonly containing pebbles of leucogranite and clasts of dark-colored metasiltstone and slate. In many places the metasandstone contains subspherical concretions several inches (3 mm) in diameter. Where metamorphosed, these rocks were called pseudodiorite (Keith, 1913), because of the igneous appearing texture. Goldsmith (1959) described and classified these metamorphic rocks as granofels.

The Thunderhead Sandstone crops out better than other rocks of the Ocoee Supergroup and are best exposed on Mount Le Conte, as noted by King and others (1958). These thick, gently southward-dipping resistant rocks form prominent cliffs and ledges along the northern slope of the Great Smoky Mountains. Thunderhead Sandstone crops out from Mount Sterling west to Calderwood Lake, and from Cove Mountain south to Mount Le Conte. It is about 6000 ft (1830 m) thick on Mount Le Conte and thickens to about 8000 ft (2440 m) near the type locality.

Several distinct lithologic units were recognized within the Thunderhead Sandstone. Most of the formation consists of thick, graded-beds of coarse-grained feldspathic metasandstone and metaconglomerate interbedded with dark-colored, graphitic metasiltstone and slate (Zt). These rocks are best observed on the slopes of Sugarland Mountain and Mount Le Conte along and above Highway 411. Between Elkmont and Metcalf Bottom, the sandstone becomes thinner bedded and contains abundant metasiltstone interbeds which have been truncated by channel deposits of coarse-grained sandstone and conglomerate.

Thick beds of dark-colored graphitic metasiltstone and slate were mapped as dark-colored metasiltstone (Zts). These beds occur near Mount Guyot, Mount Le Conte, Sugarland Mountain, and Shuckstack. Dark-colored slate and laminated metasiltstone containing subordinate thin interbeds of metagraywacke, metasandstone, and metaconglomerate (Lesure and others, 1977) crop out in the extreme southwest part of the map area. These rocks were called the Boyd Gap Formation by Wiener and Mershat (1992), but here are assigned to slate of the Thunderhead Formation (Zts). .

Boulder conglomerate containing 3 ft (1 m) -diameter, rounded leucocratic granite clasts in a matrix of quartz and feldspar, and locally large boulders of dark-colored slate and dolomite, was mapped as a separate unit (Ztb). These distinctive boulder beds were found along Big Creek near Walnut Bottom and along the Appalachian Trail near Cosby knob, both in the east part of the Great Smoky Mountains National Park. The boulder bed in Big Creek is interbedded with dark-colored metasiltstone. Its near horizontal beds result in a wide outcrop area. The wide outcrop area of the boulder bed on the west side of Calderwood Lake mapped by Lesure and others (1977) may also result from low dips.

Cades Sandstone

The Cades Sandstone (Zc) (King and others, 1958) consists predominantly of medium-grained to conglomeratic metasandstone interbedded with dark-colored metasiltstone, as seen at Abrams Falls. Neuman and Nelson (1965) mapped sequences of dark-colored metasiltstone (Zcs), as much as 2000 ft (610 m) thick, in the extreme northwest limit of the outcrop belt above the Rabbit Creek fault. Similar rocks occur throughout the formation and may be as much as 500 ft (153 m) thick, but were not differentiated by Neuman and Nelson (1965). Distinctive beds of pebble conglomerate and boulder conglomerate (Zcb), as much as 50 ft (15 m) thick, occur locally. The conglomerate contains pebbles, cobbles, and boulders of quartzite, gneiss, and leucocratic granite as much as 20 in (10 cm) in diameter, that are crudely graded and oriented parallel to bedding. They are supported in a matrix of coarse sand and feldspar. The conglomerates are best seen in Panther Creek.

The medium- to coarse-grained metasandstone of the Cades Sandstone in part resembles the Elkmont Sandstone, but the Cades is generally thinner bedded, has a wider range of grain-sizes, and a finer-grained matrix (Neuman and Nelson, 1965). It contains more clasts of dark-colored slate than the Elkmont, but similar to the Elkmont Sandstone, it contains only sparse blue quartz. Along Little River, the Cades Sandstone contains less conglomeratic rock and more dark-colored, pyritic slate than the Thunderhead Sandstone. The Cades Sandstone contains fragments of granitic rock and mappable beds of dark-colored metasiltstone and boulder conglomerate, similar to the Thunderhead Sandstone. But the Cades Sandstone is finer-grained and thinner bedded than the Thunderhead Sandstone (King, 1964). In a klippe just south of Little River, belts of Cades Sandstone and Thunderhead Sandstone come together along an interpreted thrust fault (King, 1964). This klippe extends westward to the Cades Sandstone on Cades Cove Mountain. The large mass of Thunderhead Sandstone to the northeast has been detached along the Gatlinburg fault system from the main mass of rock that underlies Thunderhead Mountain. The Cades Sandstone is here interpreted to be a northwestern facies of the Thunderhead Sandstone. These rocks were thrust onto the Pigeon Siltstone along the leading edge of the Greenbriar fault. Reactivation of this fault resulted in formation of the Metcalf Phyllite. The Greenbriar fault has cut down-section in the hanging-wall to preserve the Elkmont Sandstone on the Snowbird Group as seen along the highlands. The Elkmont, Cades, and Thunderhead Sandstones could all be facies of clastic turbidites from different areas of the same sedimentary basin and, other than for some minor lithologic differences, could constitute the same unit.

Anakeesta Formation

The Anakeesta Formation (King and others, 1958) was named for dark-colored, fine-grained rocks that form craggy pinnacles that have steep slopes. It contains a greater variety of rocks than any other formation in the highlands. The rocks where fresh, are characteristically dark-colored because of graphite, but abundant contained sulfide minerals readily weather to stain exposures rusty-red. The dominant rock unit is dark-colored graphitic and sulfidic slate, metasiltstone, and phyllite, containing local thin beds of metasandstone and metagraywacke (Za). Depending on metamorphic grade and composition, the laminated metasiltstone and phyllite contain porphyroblasts of chloritoid, biotite, or garnet. The Anakeesta Formation contains mappable units of light, coarse-grained to conglomeratic, 1 to 10 ft- (0.3 to 3 m) thick, metagraywacke beds (Zag), interbedded with metasiltstone in sequences as much as 50 ft (15 m) thick. These metagraywacke beds are abundant in the middle part of the formation west of the Oconaluftee fault, and in smaller bodies in the upper part of the formation east of the Oconaluftee fault. These coarse-grained rocks can be seen near Morton Overlook north of Newfound Gap. Although resembling similar rocks in the underlying Thunderhead Sandstone and overlying Copperhill Formation, some of the coarse-grained beds are black similar to the enclosing metasiltstone. The dominantly fine-grained rocks are well exposed on The Chimneys, Alum Cave, and in road cuts leading to Newfound Gap, and along the Appalachian Trail east of Newfound Gap, and in places such as “the Jumpoff” and Charlies Bunyon.

The metasiltstone grades upward from the underlying metagraywackes and interbedded slate that characterize the upper part of the Thunderhead Sandstone. However, on the north side of Mount Le Conte, the base of the formation consists of a chloritoid slate unit (Zac) that is distinctively light- gray, fine-grained, and siliceous (quartz), containing dark-colored, thin, tablet-shaped porphyroblasts of chloritoid, small shiny plates of ilmenite, and minute sericite. The chloritoid porphyroblasts are randomly oriented and suggesting growth under static conditions. These rocks are well exposed near Myrtle Point and in areas east of the Oconaluftee fault.

Light-colored, fine-grained metasandstone, ankeritic metasandstone, chloritoid metasiltstone, and ankeritic sandy dolomite (Zas) constitute the base of the formation on the east side of Mount Le Conte near Laurel Top. In contrast to the metasandstone of the underlying Thunderhead Sandstone, these dominantly clastic rocks are finer-grained, and thinner bedded, and have a matrix containing abundant ankeritic dolomite. At the base of the formation just east of Thunderhead Mountain, ankeritic metasandstone is well-exposed in the bed of a creek. Elliptical voids, 0.2 to 0.4 in (5 to 10 mm) in the long dimension constitute as much as about 50 percent of the outcrop.

Thin bodies of dark fine-grained dolomite, sandy dolomite, and pisolitic dolomite (Zal) occur within the metasiltstone unit. They range from 1 to 3 ft (0.3 to 1 m) thick and are interbedded with either slate or metasandstone. These small bodies are shown on the map by an “x”. They can be seen along the north side of the Newfound Gap Road, in roadcuts downstream from the bridge over Walker Camp Prong (Hadley and Goldsmith, 1963), and in the cliff at Alum Cave. The dark-colored bodies of dolomite can be recognized readily by their dissolution texture. An unusual dolomite on Mount Le Conte, 8 to 12 in (4 to 6 cm) thick, contains pisolites of quartz and calcite 0.5 in (1-10 mm) in diameter (Hadley and Goldsmith, 1963). The pisolites resemble *Girvanella* algae, but the absence of any internal organic structures suggests instead that spheroidal inorganic precipitates were incorporated into the dominantly clastic material (Hadley and Goldsmith, 1963).

The rocks of the Anakeesta Formation were preserved in the Alum Cave syncline (Hadley and Goldsmith, 1963), but the internal and enclosing strata are different from fold limb to fold limb. The massive Thunderhead Sandstone on the northwest slope of Mount Le Conte is absent south of Newfound Gap. The basal units of the Anakeesta Formation are also absent. The stratigraphy and related structures suggest that the rocks are part of a complex intertonguing depositional prism of strata that was folded and faulted in a synclinorium. A mostly homoclinal, southeast-dipping sequence of rock crops out.

Copperhill Formation

The rocks of the Copperhill Formation (Zch) (Hurst, 1955), directly overlie the rocks of the Anakeesta Formation. Wiener and Mershat (1992) showed that the Copperhill Formation is stratigraphically higher than the Thunderhead Sandstone, with which they were previously correlated by King and others (1958) and Hadley and Nelson (1971). Locally, east of Thunderhead Mountain, King (1964) recognized rocks that were stratigraphically above the Anakeesta Formation that he called the “unnamed sandstone”.

Although most of the Copperhill closely resembles the turbidite deposits of the underlying Thunderhead Formation, they have a few distinct differences. The dominant rocks are massive –bedded and coarse-grained metagraywacke and metaconglomerate, which are interbedded with quartz-garnet-muscovite phyllites and schists, which are locally sulfidic. Lesser thin beds of metagraywacke also occur throughout the formation.

The Thunderhead Sandstone and Copperhill Formation are probably the same unit, distinguished only by the presence of intervening slaty rocks that intertongue laterally to the southwest.

Dark-colored slaty metasiltstone is interbedded with metagraywacke (Zchsl) west of Hazel Creek. These graphitic and sulfidic slaty rocks host the massive sulfide deposits of the Fontana copper mine. They are locally stained rusty-orange because of iron sulfides. The largest occurrence of this unit is within the Eagle Creek shear zone, where the rocks resemble those of the Anakeesta and Wehutty Formations.

Interbedded with metagraywacke east of Hazel Creek is light brown, quartz-muscovite schist and phyllite (Zchs). Locally the phyllite is graphitic and sulfidic. The schist contains porphyroblasts of garnet, staurolite, or kyanite, depending on the metamorphic grade. Garnetiferous metasiltstone west of Clingmans Dome is transitional between the dark-colored slaty rocks in the lower part of the formation and the schists in the upper part of the formation.

Wehutty Formation

Rocks of the Wehutty Formation (Zw) (Hernon, 1969) form the perimeter of the Murphy synclinorium, a large fold that plunges southwest from the area north of Bryson City. These rocks are predominantly muscovite-staurolite-kyanite schist interbedded with metagraywacke, metaconglomerate, and metasandstone. The rocks are dark-colored because they contain abundant graphite. They are rusty when weathered as can be seen in road cuts along Lakeshore Drive. These rocks are somewhat similar to those of the Anakeesta Formation, which led Hadley and Goldsmith (1963), King and others (1968), Hadley and Nelson (1971), and Mohr (1975), to assign them to that formation. Cross-sections (Hadley and Goldsmith, 1963; Hadley and Nelson, 1971), however, suggest that these rocks occupy a higher stratigraphic and structural position than that of the Anakeesta Formation at its type locality. Therefore, we assign these rocks to the Wehutty Formation as did Weiner and Mershat (1992) and Robinson and others (1992).

Mohr (1973) mapped about 3937 ft (1200 m) of the following ascending sequence of rock in this unit: schist, metagraywacke, schist, metagraywacke, and schist. Dark-colored muscovite-biotite schist makes up about 90 percent of the section, while metagraywacke, tremolite schist, marble, quartz-chlorite schist and metasiltstone, and graphite-apatite schist constitute the remainder (Mohr, 1973). The metagraywacke is thinner bedded and darker than similar rocks in the underlying Copperhill Formation. Transitional between the schist and metasandstone facies are laminated schist and light-gray metagraywacke. These rocks are similar to the upper part of the Copperhill Formation, and they also contain abundant calc-silicate granofels. They are continuous over several mi (5 km) and are well exposed along Lake Shore Drive.

Grassy Branch Formation

The Grassy Branch Formation (Zgb) (Mohr, 1973) is metasandstone containing subordinate muscovite schist and metagraywacke that grades upward to dark-colored, porphyroblastic muscovite schist and metasandstone. The metagraywacke and laminated schist in the lower part of the unit are similar to rocks of the upper part of the Copperhill Formation and range in thickness from 330 to 984 ft (100 to 300 m). The schist contains porphyroblasts of garnet, biotite, and staurolite, commonly altered to sericite and chlorite (Mohr, 1973). These rocks are well exposed at the type locality on the north bank of Alarka Creek at its confluence with Grassy Branch.

Ammons Formation

The Ammons Formation (Zam) (Mohr, 1973) consists of metasandstone and muscovite schist that contains metasiltstone abundant in the lower part (Mohr, 1973). The upper part is dark-colored graphitic-sulfidic-mica schist and metasiltstone interbedded with metagraywacke, metasiltstone, muscovite schist, and local beds of metaquartzite and garnet-biotite porphyroblastic mica schist. The lower beds crop out in a cliff on the Little Tennessee River between Lemmons Branch and Battles Branch. The upper beds can be seen on Route 28 between the Nantahala River and the Swain-Graham County line. The uppermost beds crop out on Route 28 along Horse Branch, but were not differentiated on the map. The metasandstone is finer-grained than that metasandstone in the underlying formations of the Great Smoky Group and contains magnetite, pyrite, and epidote. These rocks have well-preserved soft sediment deformational features characteristic of turbidite deposits.

Dean Formation

Sericite schist, interbedded with metagraywacke and quartz-pebble conglomerate west of Roundtop, were assigned to the Dean Formation (Zd) of Hurst (1955), by Mohr (1975). Although the Dean and Nantahala/Tusquitee Formations are in fault contact in this area, they are in stratigraphic contact to the southwest (Mohr, 1975). East of the fault, rocks of the Dean Formation were not recognized between the Ammons Formation and the Nantahala and Tusquitee Formations. Therefore, the contact is probably unconformable. The fault that separates the Dean and Nantahala/Tusquitee Formations was interpreted as a normal fault by Mohr (1973), and a thrust fault by Weiner and Mershat (1992). The schist contains abundant biotite porphyroblasts and lesser porphyroblasts of staurolite and garnet. The coarse clastic beds are mostly in the middle and upper parts of the formation. The metaclastic rocks and mica schist are not intimately interbedded but constitute separate depositional units 33 to 330 ft (10 to 100 m) thick. The metagraywacke and conglomerate beds are similar to rocks of the underlying Copperhill and Wehutty Formations.

Rocks of the Murphy Belt

Tusquitee Quartzite and Nantahala Formation

The lowermost rocks of the Murphy belt are rocks of the Nantahala Formation, which are mostly black, graphitic metasiltstone and schist that are interbedded with white metaquartzite (Mohr, 1973). Tusquitee Quartzite is mostly white metaquartzite, interbedded with minor schist. Because metaquartzite and schist are interbedded throughout both formations, they are shown on the map as one unit (Znt), following Wiener and Mershat (1992). The Nantahala Formation (Hurst, 1955) is characterized by slaty rocks exposed in the Nantahala Gorge south of the map area. The Tusquitee Quartzite (Keith, 1907) was named for exposures on Tusquitee Mountain, about 6 mi (10 km) south of Andrews, NC. About 1000 ft (305 m) of these rocks crop out on Roundtop, west of Almond, NC, where they have been faulted onto rocks of the Dean Formation. The Tusquitee Quartzite and Nantahala Formation are gradational and conformable above rocks of the Ammons Formation (Mohr, 1973).

Igneous Rocks

Metadiabase, Metadiorite and Related Rocks

Metadiabase and metadiorite (PzZd) have intruded the metasedimentary rocks of the Snowbird and Great Smoky Groups in the highlands (Laney, 1907; Espenshade, 1963; Hadley and Goldsmith, 1963; King, 1964; Southworth, 1995). Twelve chemical analyses of these rocks indicate that they are subalkalic basalt or diorite (Southworth, 1995). They are metamorphosed and foliated, so they predate metamorphism (~440 to 415 Ma) and deformation. The dikes are medium- to coarse-grained, porphyritic to aphanitic, and contain large hornblende and plagioclase phenocrysts 1 in (2 cm) long. The northeast-trend of the dikes from near Fontana Dam to Clingmans Dome indicates that they have been transposed into the foliation within the Eagle Creek shear zone. There are multiple dikes, as can be seen on the north side of Clingmans Dome, and scattered throughout the highlands. The largest dike, about 350 ft (107 m) thick and 2 mi (3 km) long, parallels the schistosity in the metasedimentary rocks (Hadley and Goldsmith, 1963). These igneous rocks are seldom exposed. Their float consists of distinctive dark round cobbles that have pitted rinds. Cobbles and boulders rounded by streams have a diagnostic green variegated, megacrystic texture. Residuual soil and saprolite is conspicuously red and supports sparse vegetation.

Chloritic greenstone found in several places is probably altered metadiabase. Greenstone containing masses of epidosite (Southworth, 1995) can be seen adjacent to carbonate-chlorite schist and felsic rocks on the southeast bank of Ecoah Branch opposite the Fontana Copper mine. The fine-grained greenstone is composed principally of sausseritized plagioclase, quartz, actinolite, epidote, and chlorite. Locally the rock is vesicular containing quartz-filled amygdules. Epidosite masses of quartz and epidote nodules are as much as 1 ft (0.3 m) long. Float of greenstone occurs north of Mount Le Conte in the area underlain by Thunderhead Sandstone, and beneath the Greenbriar fault in three locations near Greenbriar Pinnacle (Hadley and Goldsmith, 1963). A greenstone dike in the Rich Butt Sandstone along Big Creek near Mount Sterling is 130 ft (40 m) thick and strikes east and dips 60 degrees south (Hadley and Goldsmith, 1963). The greenstone is probably metadiabase that was altered by metamorphic fluids in shear zones.

Carbonate-chlorite schist is found near the greenstone. Well-foliated carbonate-chlorite schist crops out in the Eagle Creek shear zone adjacent to thrust faults (Southworth, 1995), adjacent to metadiabase at Flint Gap and north of Clingmans Dome, and can be seen in float west of Fontana Copper mine and west of Soapstone Gap. The rock is chloritized and otherwise altered metadiorite (Espenshade, 1963; Hadley and Goldsmith, 1963). It is characteristically vuggy and contains rhombehedral cavities as much as 0.2 in (1 mm) in diameter resulting from the ankerite that can be in unweathered specimens. The rock is composed predominantly of chlorite and lesser quartz and talc, which produces a lustrous, greasy rock that can be scratched with a fingernail. "Soapstone Gap" on Pinnacle Ridge probably owes its name for this rock.

Light-gray, fine-grained, laminated, quartz-plagioclase rock that contains angular blocks and fragments as much as 1 ft (10 cm) across is near the greenstone along Ecoah Branch. The blocks and fragments are composed of medium- to coarse-grained porphyritic metavolcanic rock of intermediate composition that has equant feldspar phenocrysts as much as 0.2 in (1 mm) in diameter. Elsewhere angular blocks of the felsite occur as breccia in a fine-grained laminated quartz-plagioclase matrix. Fine-grained laminated tuff grades into crystal tuff containing abundant zoned plagioclase phenocrysts as much as 1 in (2 cm) across, lapilli tuff containing angular fragments of crystal tuff, and tuff breccia containing sub-rounded blocks of crystal tuff as much as 10 in (20 cm) long.

Higgins and others (1988) describe similar rocks that constitute the Ducktown assemblage within the Copperhill Formation in northern Georgia. Amphibolite dikes and sills, felsic metatuffs, coarse tuff breccias and coarse, poorly sorted volcanic-epiclastic conglomerates occur with the massive sulfide deposits at Ducktown, TN. Although volumetrically small, these igneous rocks are probably rift volcanic depsoits associated with the formation of the Great Smoky Group basin.

Pegmatite

Small pegmatites (Pzp) of two distinct ages are in the southern and southeastern parts of the map area. Foliated pegmatite, aplite, and pegmatitic granite resulting from plutonism and metamorphism during the Grenvillian orogeny are within the Mesoproterozoic gneisses. These foliated pegmatites contain pink potassium feldspar and green epidote. In contrast, younger tabular dikes and sills of white, massive, unfoliated pegmatite have intruded the Mesoproterozoic gneiss as well as rocks of the Neoproterozoic Great Smoky and Snowbird Groups. They are found only in rocks of staurolite to kyanite grade, especially east and southeast of Dellwood, and from Bryson City east to near Cherokee. In these areas, the pegmatite has intruded both the basement and cover rocks around the tectonic windows of the Greenbriar fault. It occurs in and along the margin of these windows, but is especially concentrated along the western margin of the Bryson City window. The pegmatite is homogeneous, zoned, and consists of white oligoclase, white perthite, quartz, muscovite, and at places biotite (Cameron, 1951); the muscovite books are as much as 2 to 4 in (4 to 8 cm) across. The pegmatites range in size from 1 in (2 cm) thick and 12 in (24 cm) long, to about 200 ft (61 m) thick and 500 ft (153 m) long. Most of the pegmatites are not foliated, and appear to have intruded the mylonitic rocks along the Greenbriar fault. However, some pegmatite must be pre- to syn-kinematic, because it is parallel to foliation and is folded. The northward elongation and steep westward dip (Cameron, 1951; Hadley and Goldsmith, 1963), suggests that the pegmatites were folded during the formation of the Bryson City dome and the Murphy synclinorium.

Trondhjemite

Dikes of trondhjemite occur in the Mesoproterozoic gneiss and Neoproterozoic rocks east and southeast of Dellwood, the Soco-Cherokee belt, and around Bryson City. These dikes range in width from 2 in (4 cm) to 10 ft (3 m), but most are 2-3 ft (1 m) wide. They are light-gray, have equigranular crystals of calcic oligoclase, quartz, biotite, and magnetite 0.25-0.5 in (1 cm) across. These rocks were too small to be shown on the map.

Vein Quartz

White vein quartz pods, dikes, lenses, and tabular bodies as much as 30 ft (9 m) thick, intruded all of the metamorphic rocks. Some of the largest concentrations are within the Eagle Creek shear zone (Southworth, 1995). There the vein quartz is isoclinally folded and boudinaged, whereas elsewhere in the region, vein quartz is massive and not foliated. In several places, brecciated vein quartz suggests local brittle faulting. Boulders of vein quartz as much as 10 ft (3 m) in diameter are common in creek beds in areas where fine-grained rocks of the Great Smoky Group and Metcalf Phyllite were sheared and fluids introduced during metamorphism. Bodies of vien quartz were too small to portray on the map.

Rocks of the Foothills of the Western Blue Ridge

Neoproterozoic Walden Creek Group

Introduction

The Walden Creek Group was named for rocks exposed along Walden Creek east of Pigeon Forge in the foothills of the western Blue Ridge (King and others, 1958). They subdivided this assemblage of sedimentary rocks, in ascending order, into the Licklog, Shields, Wilhite, and Sandsuck Formations. The stratigraphy seen east of Pigeon Forge (Hamilton (1961) were used to classify the rocks found west of Pigeon Forge, where the stratigraphy had been structurally disrupted. These rocks are mostly separated from the rocks of the Snowbird and Great Smoky Groups by faults. Northeast of the map area, Oriel (1950) and Ferguson and Jewell (1951), reported that rocks of the Walden Creek Group conformably overlie rocks of the Snowbird Group. Hamilton (1961) suggested that the Snowbird and Walden Creek Group rocks were transitional across the Dunn Creek fault.

The lower part of the Licklog Formation is a fault, so only a few hundred feet of it is present. It appears to intertongue with and be conformable with conglomerate of the Shields Formation (Hamilton, 1961). The Wilhite Formation is transitional above the coarse-grained rocks of the Shields Formation (King, 1964). The limestone and shale unit of the Wilhite Formation is concordantly overlain by siltstone and sandy limestone of the Sandsuck Formation. At the northeast end of Chilhowee Mountain, the upper sandstone and quartz-pebble conglomerate of the Sandsuck Formation appear to be channel deposits that were truncated and unconformably overlain by the Cochran Formation (King, 1964).

Rocks of the Walden Creek Group crop out along the Foothills Parkway between Wear Cove and Walland, near its west end near Chilhowee Lake, and in the western part of the National Park in the lower Abrams Creek drainage. This Group is perhaps the most diverse, heterogeneous, and controversial as to age in any rocks of the Ocoee Supergroup.

Licklog Formation

The Licklog Formation (Zll) (King and others (1958), consists of a poorly exposed sequence of siltstone and shale interbedded with some fine- to coarse-grained sandstone and pebble conglomerate a few feet thick. The rocks are best exposed along the Little Pigeon River near the Dunn Creek fault. Rocks of the Licklog Formation crop out near Shields Mountain east of Pigeon Forge and above the Great Smoky fault on the north side of Wear and Tuckaleechee Coves. North of these coves, the dominant siltstone unit contains beds of conglomeratic sandstone (Zllc).

Shields Formation

Named for rocks cropping out on Shields Mountain near the Little Pigeon River, the Shields Formation (King and others, 1958; Hamilton, 1961) consists of thick-bedded sandstone and conglomerate that are overlain by and intertongue with siltstone, shale, and slate. The rocks range from 2000 to 2500 ft (610-763 m) thick. The basal rocks of the formation are distinctive coarse conglomerate containing polymictic pebbles and cobbles that is interbedded with pebbly sandstone (Zsc). Most clasts are rounded milky quartz pebbles that average 1 in (2 cm) in diameter. In addition, there are also quartzite, granite, chert, siltstone, and limestone pebbles. Conglomeratic rocks previously assigned to the Wilhite Formation (Neuman and Nelson, 1965) and Thunderhead Sandstone (King, 1964), are here assigned to the Shields Formation. The conglomeratic rocks are best seen in road cuts along Shields Mountain, northwest of Kinzel Springs, and along the bluffs at Chilhowee dam. The coarse basal rocks grade up into laminated siltstone and shale (Zs), which constitute most of the formation. Locally, however, these rock types are interbedded. Coarse sandstone and siltstone beds (Zss) are transitional and intertongue with the conglomerate and silty rocks, especially east of Pigeon Forge. They are best seen along State Highway 321 at Kinzel Springs. Thin beds of argillaceous limestone within siltstone (Zsl) were mapped along the Dunn Creek fault near Bird Creek.

Wilhite Formation

The Wilhite Formation was named for rocks exposed along Wilhite Creek and Long Branch (King and others, 1958), in the eastern part of the northern foothills. In the type area, Hamilton (1961) mapped 2 members, the Dixon Mountain and Yellow Breeches. The Dixon Mountain Member consisted of about 1500 ft (458 m) of metasiltstone and metasandstone, which were overlain by about 2000 ft (610 m) of metasiltstone that grades up into limestone-conglomerate of the Yellow Breeches Member. West of the Pigeon Forge fault, King (1964) described a 1750 ft (534 m) thick lower unit of metasiltstone, that is interbedded with quartzite, metasandstone, and conglomerate, and a 1750 ft (534m) thick upper unit of shale and slate interbedded with limestone. Farther west, Neuman and Nelson (1965) mapped about 10,000 ft (3 km) of siltstone, interbedded with about 2500 ft (763 m) of quartz pebble conglomerate and quartzose sandstone.

Regionally, the Wilhite Formation was separated into 3 main units: 1) shale, siltstone, and slate (Zw), 2) clastic rocks interbedded with calcareous clastic rocks (Zwlc), and 3) limestone and shale (Zwl). Fine-grained metasiltstone (Zw) predominates. To the northeast, carbonate rocks are interbedded with the fine-grained rocks (Zwlc), and eventually carbonate rocks predominate (Zwl).

Most of the Wilhite Formation is composed of fine-grained shale, siltstone, and slate (Zw). In the eastern part of the map area, it is typically laminated metasiltstone containing thin interbeds of sandstone, limestone and dolomite. Typical outcrops can be seen along Dunn Creek. In the central area, the rocks are mainly laminated metasiltstone and slate containing interbeds of quartzite, sandstone, and conglomerate in the lower part, and shale and slate interbedded with limestone in the upper part. Typical outcrops can be seen along Cove Creek. In the western area, the dominant rocks are laminated metasiltstone, fine-grained sandstone, conglomerate, and carbonate rocks. The laminated metasiltstone has laminae of ankerite. Conglomerate and sandstones are interbedded with the finer-grained rocks. The conglomerates are graded and contain about 80 percent vein quartz clasts. They resemble rocks of the underlying Shields Formation. The largest clast observed was a sandstone boulder 5 ft (1.5 m) long and 3 ft (1 m) wide (Neuman and Nelson, 1965). Carbonate rocks range from layers a few inches to 150 ft (46 m) thick within both metasiltstone and conglomerate horizons, and include limestone breccia and fine-grained laminated limestone. Some limestone bodies are blocks and pebbles deposited in mud flows (Hanselman and others, 1974). Characteristic outcrops can be seen along Chilhowee Lake.

From near Crockettsville east to Sol Messer Mountain, clastic rocks are interbedded with carbonate rocks (Zwlc). Gray metasiltstone and sandstone contain interbeds of gray limestone, dolomite and ankeritic dolomite. About 1500 ft (460 m) of this sequence can be seen along Long Branch.

Limestone and shale (Zwl) were mapped near Pigeon Forge and Crockettsville. Thick gray limestone beds are commonly sandy and conglomeratic, and are interbedded with shale and siltstone. The limestone conglomerate consists of subangular chips and slabs of limestone, about 8 in (15 cm) long, as well as other types of rock fragments that are subparallel to bedding. Coarse conglomerate near Chavis Creek contains boulders of dolomitic limestone as much as 5 ft (1.5 m) long. Good exposures of the limestone and shale can be seen along Cosby Creek, in roadcuts near Jones Cove, and in old quarries near Pigeon Forge.

To the southwest of the map area, rocks of the Wilhite Formation grade up from the underlying Dean Formation that is considered to be part of the Great Smoky Group (Thigpen and Hatcher, 2004).

Sandsuck Formation

The Sandsuck Formation (Zss) (Keith, 1894) (King and others, 1958) was named for shale exposed near Sandsuck Branch near the northeastern end of Chilhowee Mountain. The rocks are best exposed on the northwest slope of Chilhowee Mountain near the heads of the North and South Forks of Ellijoy Creek, and south of English Mountain. Rocks of the Sandsuck Formation are lithologically similar to other rocks of the Walden Creek Group. They are predominantly gray, thin-bedded fissile siltstone interbedded with cross-bedded, fine- to coarse-grained feldspathic sandstone and conglomerate that are as much as 2500 ft (763 m) thick. Two units were mapped in the Sandsuck Formation; Greenish-gray siltstone, silty shale, sandstone, and conglomerate (Zss), and thick lenticular beds of coarse-grained sandstone and quartz-pebble conglomerate (Zssc). South of English Mountain, the coarse-grained sandstone and quartz conglomerate is in the middle part of the section (Hamilton, 1961), but it occurs at the top of the section on Chilhowee

Mountain (King, 1964). The quartz-pebble conglomerate beds of the Sandsuck are similar to quartz-conglomerate in the underlying Wilhite and Shields Formations. In contrast to rocks of the Wilhite Formation, the sandstone and conglomerate of the Sandsuck Formation contain fewer shale fragments and are generally thinner bedded. In addition, the siltstones lack the iron-rich layers and green color that are common to rocks of the Wilhite Formation.

Paleozoic Rocks

Paleozoic rocks crop out in the foothills of the western Blue Ridge province along the northern border above the Great Smoky fault, and within Calderwood, Cades, Tuckaleechee, and Wear Coves tectonic windows.

Lower Cambrian Chilhowee Group

The sequence of quartzite and interbedded siltstone and shale that are well exposed on Chilhowee Mountain, Green Mountain, and Stone Mountain, were called the Chilhowee Group. The Chilhowee Group extends over the entire length of the western Blue Ridge province, from northern Georgia into southern Pennsylvania. They were divided by Keith (1895) and King and others (1958) into (ascending order): the Cochran Formation, Nichols Shale, Nebo Quartzite, Murray Shale, Hesse Quartzite, and Helenmode Formation. The rocks on Chilhowee Mountain lie between the Great Smoky fault and Miller Cove fault. The rocks form a southeast-dipping homocline that has synclines formed in the footwall of the Miller Cove fault. The quartzite units hold up subparallel ridges and the intervening swales are floored by shale. The Hesse and Nebo Quartzites are especially prominent in cliffs along Chilhowee Mountain. Green Mountain is also underlain by a southeast-dipping sequence of rocks beneath the Dunn Creek fault. The rocks in the upper part of the Chilhowee Group (Ccu) (Nebo, Murray, Hesse, Helenmode Formations) were not differentiated.

The basal rocks of the Cochran Formation are conformably above and grade up from similar coarse-grained rocks of the Sandsuck Formation of the Walden Creek Group. The lower contact of the Nichols Shale with underlying quartzite of the Cochran Formation is sharp. Its upper contact with the overlying Nebo Quartzite is transitional from shale into quartzite. The contact between the Nebo Quartzite and the overlying Murray Shale appears to be sharp. The contact of the Murray Shale with the overlying Hesse Quartzite is abrupt but is locally transitional (Neuman and Nelson, 1965). In the upper Hesse Quartzite, quartzite interbedded with shale grades up into shale and siltstone interbedded with sandstone of the Helenmode Formation. At the top of the Chilhowee Group, the Helenmode grades up into dolomitic shale of the overlying Shady Dolomite.

Cochran Formation

The Cochran Formation (Cc), named for outcrops near Cochran Creek, consists of a basal gray conglomerate that is overlain by maroon pebbly arkose that is interbedded with maroon shale and siltstone. These rocks grade up into light-gray arkose, cross-bedded sandstone, and quartzite. Concretions of gray hematite, 1 to 4 in (2 cm) in diameter distinquish these quartzite beds from the beds in the overlying Nebo and Hesse Quartzites. The base of the formation was defined as the base of the thick conglomeratic feldspathic sandstone that overlies thin fissile siltstone of the Sandsuck Formation. The top of the formation is quartzite that is sharply overlain by Nichols Shale. The rocks of the Cochran Formation can be seen on the south side of Green Mountain, and extensively along the west slope of Chilhowee Mountain, and they also crop out on the east side of Chilhowee Mountain at the north end of Happy Valley.

Nichols Shale

The Nichols Shale (Cn), named for outcrops along Nichols Branch of Walden Creek, consists of greenish-gray, argillaceous to silty, well-laminated, fissile shale containing some layers of sandstone and quartzite. The sandstone is feldspathic and large flakes of detrital muscovite coat bedding surfaces. This unit is exposed at Chilagatee Gap and also in an old quarry north of Walland on the east side of Little River. Ribbon-like impressions in shale near Chilogatee Gap have been interpreted as trace fossil burrows (Neuman and Nelson, 1965).

Nebo Quartzite

The Nebo Quartzite (Cnb), named for rocks cropping out near Mount Nebo Springs, consists mostly of thin-bedded white quartzite. The basal rocks are quartz sandstone rich in chlorite. Clean quartzite beds, are mostly less than 1 ft (25 cm) thick and are cross-bedded. The quartzite ranges in thickness from 200 to 400 ft (61 to 122 m) over a distance of 15 mi (24 km). Closely-spaced, narrow, cylindrical tubes are perpendicular to bedding. These tubes terminate at bedding planes and are the trace fossil *Skolithos linearis,* which are intertidal to shallow subtidal burrows of worm-like organisms.

Murray Shale

The Murray Shale (Cm), named for Murray Branch of Walden Creek, is mostly greenish-gray argillaceous to silty shale. The shale is in sharp contact with the underlying Nebo Quartzite. The shale is interbedded with fine-grained feldspathic and glauconitic sandstone in the upper half of the formation. It ranges in thickness from 350 ft (107 m) at Murray Gap to 550 ft (168 m) at Chilogatee Gap, over a distance of about 5 mi (8 km).

Hesse Quartzite

The Hesse Quartzite (Ch), named for the Hesse Creek tributary of Little River, is mostly medium- to coarse-grained quartzite containing well-rounded grains. The quartzite beds are generally 2 to 4 ft (0.6 to 1.2 m) thick and are locally cross-bedded. The formation is about 500 to 600 ft (153 to 183 m) thick. *Skolithos linearis* tubes are abundant in some beds, but there are fewer and shorter than those seen in the Nebo Quartzite. Like the other quartzite units within the Chilhowee Group, the Hesse Formation forms dip slopes on Chilhowee Mountain along the Foothills Parkway.

Helenmode Formation

The Helenmode Member of the Erwin Formation, found about 40 mi (64 km) to the northeast (King and others, 1944), was renamed the Helenmode Formation (Chm) by King and others (1958). Rocks of the Helenmode Formation consist of silty shale and siltstone interbedded with thin quartzite at the base and coarse sandstone near the top.

Shady Dolomite

The Shady Dolomite (Cs) is a 1000 ft (305 m) thick sequence of thick-bedded dolomite that includes a few interbeds of dolomitic shale in the upper third of the formation (Neuman and Nelson, 1965). The dolomite can be either crystalline, thick bedded and massive, or thin-bedded and laminated. Where weathered, the residuum of Shady Dolomite is clay and contains massive boulders of jasperoid. The dolomite commonly contains irregular masses of fine-grained chalcedonic chert. The Shady Dolomite crops out north of Cosby where it underlies both broad valleys and ridges as much as 600 ft (183 m) high. Shady Dolomite also crops out on the southside of Chilhowee Mountain from its northeast end southwestward to Top of the World Estates. A small outlier of dolomite north of Walden Creek was interpreted to be Shady Dolomite exposed in a tectonic window (King, 1964). The shaly dolomite and dolomitic shale were mapped separately (Css) on either side of the Little River Gap on the south side of Chilhowee Mountain. Shady Dolomite is well exposed in roadcuts along State Highway 73 near the entrance to the Foothills Parkway near Miller Cove.

Rome Formation

The Rome Formation (Cr) consists of fissile, laminated, red and maroon shale, calcareous siltstone, and fine-grained sandstone (Neuman and Nelson, 1965). Although generally weathered, poorly exposed, and covered by surficial material, the red rocks are exposed in the cores of the synclines north and south of Miller Cove, and northeast of Cosby.

Paleozoic Rocks within the Tectonic Windows of the Foothills of the Western Blue Ridge

Six tectonic windows called coves were recognized in the region. From southwest to northeast, these are the Calderwood window, Cades Cove, Big Spring Cove, Whiteoak Sink, Tuckaleechee Cove, and Wear Cove. Limestone in the windows "is tectonically disordered, lacking distinctive fossils or lithologic features that would indicate their exact stratigraphic position" (King, 1964), so it has not been subdivided into mappable units (Neuman and Nelson, 1965), except for the Jonesboro Limestone (Oj) (Rodgers, 1953) and Blockhouse Shale (Obl). The underlying carbonate rocks dissolved by physiochemical processes to form closed basins surrounded by mountains. Whiteoak Sink is actually part of the Tuckaleechee window, but is separated from it by Scott Mountain. The western part of the cove at Calderwood was breached by the Little Tennessee River. Big Spring Cove was confirmed to be a tectonic window by drilling in 1951 (King, 1964) that penetrated limestone at a depth of 45 ft (15 m). The rocks in the Calderwood window, Cades Cove, and Big Spring Cove are all considered to be Jonesboro Limestone. The rocks in Tuckaleechee and Wear Coves are assigned to the Jonesboro Limestone, the Lenoir Limestone and the Blockhouse Shale. Tuckaleechee Cove contains several intraformational thrust sheets that locally duplicate strata, especially near the north and south margins of the window. In Wear Cove, Jonesboro Limestone has been thrust onto Blockhouse Shale along the margins of the window. Jonesboro Limestone crops out in the center of the cove in two anticlines.

The Jonesboro Limestone (Oj) is dominantly gray, fine-grained limestone in 0.5 (8 cm) to 3 ft (1 m) thick beds characterized by thin wavy clay and silt partings. Less abundant are massive limestone beds containing quartz sand grains, fossils, and dolomite containing cross hatch-patterns of calcite-filled joints. There are irregular masses of black and white nodular chert locally in the residuum. Disseminated chert stringers were seen in a few outcrops. Jonesboro Limestone is best observed in Tuckaleechee Cove, where about 2000 ft (610 m) of folded and faulted limestone crops out. The Jonesboro Limestone is overlain by the Middle Ordovician Lenoir Limestone (Safford and Killebrew, 1876) along an erosional unconformity that has as much as 140 ft (43 m) of relief (Bridge, 1955). The Lenoir Limestone (Ol) consists of gray, cobbly argillaceous limestone and limestone conglomerate. About 25 ft (8 m) of Lenoir Limestone in Tuckaleechee Cove was included with the Jonesboro. Locally the basal rocks of the Lenoir are limestone conglomerate composed of detritus eroded from older rocks. An 8-ft (2.5 m) thick bed of limestone conglomerate containing angular fragments of limestone 0.75 in (2 cm) long is exposed along the State Highway 73 at the south side of Tuckaleechee Cove.

The Blockhouse Shale (Obl) (Neuman, 1955) is dark-gray, finely-laminated, fissile calcareous shale. Beds of cobbly argillaceous limestone 3 to 5 ft (1 to 1.5 m) thick occur locally at the base. These cobble beds were named the Whitesburg Limestone Member by

Ulrich (1929). Local 10 ft (3 m) thick beds of calcareous sandstone also occur a little higher in the section. These were too thin to be shown separately on the map. The shale has been folded and contains abundant slickensided surfaces. The Whitesburg Limestone Member of the Blockhouse Shale disconformably overlies the Lenoir Limestone. The top of the Blockhouse Shale has been truncated by the Great Smoky fault, but in the Tennessee Valley it grades up into the Tellico Formation. Blockhouse Shale can be seen along the Rich Mountain Road at the crest of Rich Mountain, where it is exposed beneath the Great Smoky fault.

Slices of Jonesboro Limestone along the Great Smoky Fault

Slices of Jonesboro Limestone occur between the Blockhouse Shale and older rocks of the Ocoee Supergroup or Chilhowee Group along the Great Smoky fault. These slivers of rock range from a few feet (0.5 m) thick and a hundred feet (30 m) long to several thousand feet (1 km) thick and several miles (4 km) long. A well-exposed slice 5 to 25 ft (1.5 to 7.6 m) thick occurs at the southern boundary of Tuckaleechee Cove along State Highway 73. Small slices of limestone and dolomite occur along the leading edge of the Great Smoky fault in the Tennessee Valley. The largest slice of limestone is just south of Cedar Bluff.

Paleozoic Rocks of the Tennessee Valley

Introduction

The youngest Paleozoic rocks are Carboniferous clastic rocks beneath the Great Smoky fault in the easternmost Tennessee Valley. Most of the rocks in the Tennessee Valley are Cambrian and Ordovician in age. These are carbonate rocks interbedded with calcareous sandstone and shale. There are no Silurian age rocks. About 10 ft (3 m) of black shale at the base of the Mississippian strata is considered to be Devonian in age. They however, could be Mississippian in age (Neuman and Nelson, 1965). The lack of Silurian rocks and possibly Devonian rocks is difficult to explain. Either they were never deposited or they were eroded prior to Devonian and (or) Mississippian sedimentation. The Paleozoic rocks were defined and described by Rodgers (1953), Neuman (1955;

1960), Cattermole (1955; 1962), and Neuman and Wilson (1960), and are only discussed briefly herein.

Conasauga Group

The Conasauga Group consists of the Rome Formation, Pumpkin Valley Shale, Rutledge Limestone, Rogersville Shale, Maryville Limestone, Nolichucky Shale, and Maynardville Limestone. These rocks crop out near Maryville in a northeast-trending anticlinorium along a thrust fault north of Alcoa that has placed them on rocks of the Knox Group. The clastic rocks of the Rome Formation and Pumpkin Valley Shale were combined on this map into one unit (Cpr). The Rome Formation consists of maroon sandstone, siltstone, and shale, and lesser quartzite and limestone. These rocks grade up into greenish-gray and red siltstone, silty shale, and thin layers of limestone of the Pumpkin Valley Shale. The Rutledge Limestone (Crt) is limestone with argillaceous partings at the base and fine-grained dolomite at the top. The Rogersville Shale (Crg) is variegated purple and green fissile shale containing a 20 ft (6 m) thick bed of limestone in its middle part. The Maryville Limestone (Cml) has greenish-gray shale near both its base and the top. The Nolichucky Shale (Cno) is greenish-gray calcareous shale interbedded with oolitic limestone.

Knox Group

In this area the Knox Group (OCk) consists of the Copper Ridge Dolomite, the Chepultepec Dolomite, the Longview Dolomite, and the Newala Limestone (Neuman and Nelson, 1965). The Copper Ridge Dolomite is Late Cambrian in age whereas the other units are Early Ordovician in age. The Copper Ridge Dolomite is cherty. Locally, Upper Cambrian Maynardville Limestone was included at the base of the Copper Ridge Dolomite on the map because of their small size. The Chepultepec Dolomite is dolomite and limestone with sandstone beds in its upper part. The Longview Dolomite is cherty. The Newala Limestone consists of limestone interbedded with dolomite and sandstone. Rocks of the Knox Group crop out north of Pigeon Forge in the core of several anticlines, as well as in two belts northwest and southeast of Maryville.

Middle Ordovician Rocks

Middle Ordovician rocks include the Lenoir Limestone (Olh), the Blockhouse Shale (Obl), the Chapman Ridge Sandstone (Ocr) and marble (Ocm), the Tellico Formation (Ot), the Ottosee Shale (Oo), the Chota Formation (Oc), the Sevier Formation (Os), and the Bays Formation (Ob). The stratigraphy differs across the Dumplin Valley thrust fault. To the northwest the Blockhouse Shale is overlain by Chapman Ridge Sandstone and marble and Ottosee Shale. Southeast of the thrust fault, the Blockhouse Shale is overlain by Tellico Formation, Chota Formation, Sevier Formation, and Bays Formation. The Lenoir Limestone (Ol) is a cobbly and argillaceous and locally contains the fine-grained Mosheim Limestone Member and basal clastic rocks. Small areas of thick-bedded marble of the overlying Holston Formation were included with Lenoir Limestone on the map. These rocks crop out in the extreme northwest part of the map area north of Rockford. Similarly, locally small slivers of Lenoir Limestone (Olh) were included at the top of the Copper Ridge Dolomite. The Blockhouse Shale (Obl) is dark-colored, fissile, and calcareous containing local beds of sandstone in its lower part as well as thin limestone beds locally at its base. Blockhouse Shale occurs in several anticlines between Sevierville and Pigeon Forge, and east of Maryville. The Chapman Ridge Sandstone (Ocr) is gray calcareous sandstone and argillaceous shale that underlies High Top and Red Mountain, northeast of Rockford in the northwestern part of map area. The Chapman Ridge marble (Ocm) is thickly bedded to massive, gray to red, coarsely crystalline marble in lenses at the top of the sandstone. The marble crops out along the lower slopes of ridges supported by the underlying sandstone. The Ottosee Shale (Oo) has calcarenite and sandy limestone in its lower part and grades up into calcareous shale. Ottosee Shale underlies most of the lowland around Fort Loudon Lake in the extreme northwest part of the map area.

Gray, sandy and silty calcareous shale contains beds of calcareous sandstone, fine-grained sandstone, and impure limestone of the Tellico Formation (Ot) (Keith, 1895; Neuman, 1955). The light-gray shale is interbedded with the darker-colored Blockhouse Shale at its base. It transitionally overlies the Blockhouse Shale at the town of Blockhouse and southeast of the Dumplin Valley thrust fault. These rocks underlie a broad area from Sevierville southwest to beyond Blockhouse, and east of Pigeon Forge, north of the Great Smoky fault. Along the Great Smoky fault west of Pigeon Forge, the rocks assigned to the Tellico Formation may include rocks of the Blockhouse Shale, tectonically mixed along the fault (Neuman and Nelson, 1965). These clastic rocks underlie a series of low ridges (Black Sulphur Knobs and Chestnut Ridge) near Blockhouse. Transitionally above the rocks of the Tellico Formation are blue-gray calcareous sandstone and sandy limestone of the Chota Formation (Oc) (Neuman, 1955). The Chota Formation contains fossil debris, especially abundant brachiopods. These rocks extend from the west side of the map area near Sixmile, northeast to just beyond the Blount/Sevier County line, where they are truncated by the Guess Creek fault. Rocks of the Sevier Formation (Os) (Neuman, 1955), are gray, calcareous, silty and sandy shale and locally, sandstone. The basal shale conformably overlies rocks of the Chota Formation on a sharp contact. The top of the Sevier Formation is gradational between the Sevier and the overlying Bays Formation (Ob). The Sevier Formation also extends from west side of the map near Sixmile, northeast to beyond the Blount/Sevier County line. Overlying the Sevier Formation, the Bays Formation (Ob) (Keith, 1895; Rodgers, 1953; Neuman, 1955) consists of red calcareous mudrock and siltstone, locally containing coarse-grained feldspathic and light-gray quartzitic sandstone interbedded with red fine-grained sandstone. The uppermost quartzite of the Bays Formation is discontinuous. These characteristic red rocks extend from the west side of the map near Sixmile Creek, northeast to near Rocky Branch.

Chattanooga Shale

Keith (1895) called the thin, dark carbonaceous shale that overlies the rocks of the Bays Formation the Chattanooga Shale (Dc). The shale contains sulfide minerals and rusty concretions and is not calcareous. Locally at the base, the shale is sandy, contains sandstone beds a few inches thick, and is an unconformity. However, the shale has been mostly deformed into pods about 10 to 25 ft (3 to 7.6 m) thick with slickenlines. The Chattanooga Shale extends from the west side of the map area near the base of Chilhowee Mountain, where it underlies a swale marked by small borrow pits and is often covered by colluvium. These black rocks were locally exploited as coal, which they are not. Conodonts, linguloid brachiopods, and marine megafossils collected from the black shale in the region suggest that some of these rocks may be Early Mississippian and are not restricted to the Upper Devonian (Neuman and Nelson, 1965).

Grainger and Greasy Cove Formations

The Lower Mississippian Grainger Formation (Mg) (Keith, 1895) consists of a monotonous sequence of siltstone and fine-grained sandstone that grade up into coarser-grained feldspathic sandstone and pebble conglomerate interbedded with silty shale. Light-gray siltstone at the base of the Grainger sharply overlies the dark-colored shale of the underlying Chattanooga Shale. The coarse-grained rocks in the upper part of the Grainger are similar to and grade into rocks of the overlying Greasy Cove Formation.

The lower Upper Mississippian Greasy Cove Formation (Mgc) (Neuman and Wilson, 1960) consists of interbedded gray, calcareous shale, argillaceous limestone, fine-grained sandstone, and red shale. The shale and limestone readily weather to a greasy, waxy clay residuum. Only the lower part of the formation is exposed along creeks because of the abundant colluvium along the base of Chilhowee Mountain. Its upper part has been truncated by the Great Smoky fault. The belt of Mississippian rocks along the Guess Creek fault have been so structurally deformed that they cannot be diffentiated. Therefore, they are shown as Greasy Cove and Grainger Formations (Mgg).

Interpretation of Sedimentary Environments

Ocoee Supergroup

Presented here is the interpretation of the origin of the rocks of the Ocoee Supergroup as suggested by King and others (1958; 1968), De Windt (1975), and Rast and Kohles (1986). Sedimentologic data suggests that the rocks of the Snowbird Group were deposited in a basin that was both horizontally and vertically asymmetric. These rocks consist of alluvial (Wading Branch Formation) and fluvio-deltaic-offshore (Longarm Quartzite, Roaring Fork Sandstone, Pigeon Siltstone, and Rich Butt Sandstone) deposits that had a source to the east-southeast. Deposition of these sediments waned and pinched-out eastward, and sediments of the Great Smoky Group were deposited above them.

Rocks of the overlying Great Smoky Group were deposited in an asymmetric basin on the east side of the Snowbird basin. Great Smoky Group rocks are turbidites that were deposited into a deep elongated basin with a source to the northeast. The basal Elkmont Sandstone is transitional above the prodeltaic sediments of the Rich Butt Sandstone of the underlying Snowbird basin. The graphitic, sulfidic, slaty rocks throughout the Great Smoky Group were deposited in deep euxinic subbasins within a larger basin, which was continuously covered with turbidites.

The strata of the Walden Creek Group were deposited above the rocks in the Snowbird basin, and were derived from a source to the west. Synchronously, rocks of the Murphy belt were being deposited in the Great Smoky basin. The discontinuous, lithologically heterogeneous rocks of the Walden Creek Group were deposited on a tectonically unstable shallow marine shelf. The lenticular deposits of polymictic conglomerate and carbonate breccias indicate reworked shallow water deposits during either tectonic events or during storms. A passive margin then developed and blanket sands and shale sequences of the Chilhowee Group were deposited onto the Walden Creek rocks.

Snowbird Group

The rocks of the Snowbird Group lie unconformably on Mesoproterozoic gneisses and are considered the oldest rocks of the Ocoee Supergroup. In the southeastern part of the region, the Snowbird Group is thinnest, locally absent, and is overlain by rocks of the Great Smoky Group. Rocks of the Snowbird Group are fine-grained and are thickest to the northwest of the outcrop area. Here they were possibly overlain by rocks of the Walden Creek Group (King and others, 1958). The rocks of the Snowbird Group thicken to the northwest, a result from the deepening basin related to the formation of normal growth faults (Montes and Hatcher, 1999). Mineralogical and chemical analyses suggest that these rocks are the reworked residuum of granitic rocks (Hadley and Goldsmith, 1963). The lack of coarse debris and the dominance of reworked sediment suggest that the relief of the source area was relatively low. Basal rocks of the Wading Branch Formation were local channel fill deposits as well as residual soil that were deposited on an irregular surface of granitic rocks (Hadley and Goldsmith, 1963). The current-bedded Longarm Quartzite indicates river deposits of quartz sand, which are thickest where it was deposited in valleys. Slump-folds in the Roaring Fork Sandstone suggest that the land subsided to form a deep water basin. Further deepening of the water is indicated by the fine-grained laminated rocks of the Pigeon Siltstone. The metasiltstone contains syn-depositional normal faults and gravity slumps that suggest rapid deposition of reworked sediments. The overlying rocks of the Rich Butt Sandstone mark a dramatic change in sedimentation as turbidity currents were introduced. The high feldspar content of the sandstone reflects the rapid erosion of granitic rocks in the source region and rapid deposition within the basin.

Great Smoky Group

Rocks of the Great Smoky Group are flysh that was deposited by turbidity currents. The change from the massive, thick-bedded metagraywacke of the Thunderhead Sandstone, to the medium-bedded metagraywacke of the Copperhill Formation, to the thin-bedded metagraywacke in the Dean Formation, records the decrease in energy as a function of time. Episodes of faulting and earthquakes helped to generate the turbidites. Between the deposition of these large submarine turbidites, quiet water and oxygen-poor conditions resulted in the precipitation of carbonaceous and sulfidic sediments. This reducing environment allowed the minor influx of carbonate debris and fine- to coarse-grained, more oxidized material (Mohr, 1973). The best examples of the non-turbidite deposits are the Anakeesta Formation, Wehutty Formation, and the Ammons Formation. Incorporation of black slaty rocks occurred at all scales, within outcrops, within formations, and within the group. Regionally, the relative amount of argillaceous sediments in the Great Smoky Group increases to the south (Fairly, 1965). The formations of the Great Smoky Group then, are in essence local deposits in a larger basin.

Rip-up clasts of black slaty material containing soft-sediment deformation features, as well as lithified rocks, occur in the coarse-grained turbidites of the Elkmont Sandstone, Thunderhead Sandstone, and Cades Sandstone. This suggests that euxinic conditions existed within the basin prior to the deposition of the Anakeesta Formation. The lack of graphite and iron sulfide minerals in the schist of the upper part of the Copperhill Formation suggests a temporary change from anoxic to aerobic conditions during sedimentation.

Walden Creek Group

Laminated metasiltstone is the dominant rock of the Walden Creek Group, followed by fine-grained sandstone, quartzite, conglomerate, and carbonate rocks. These rocks of contrasting composition, lenticular nature, and texture suggest an extreme alternation of depositional conditions, source material, and relief. Cross-beds, asymmetric loadcasts, and pebble imbrication suggest transport from the north (Neuman and Nelson, 1965). Fine sediment deposited under calm conditions was followed by heterogeneous gravel deposits (conglomerates) of both foreign and local source. Detrital grains of feldspar, biotite, and muscovite, and clasts of well-rounded vein quartz and leucogranite suggest derivation from igneous and metamorphic rocks and reworked sediment. Erosion and transport were rapid and deposition under widespread heterogeneous conditions. Rounded vein quartz pebbles suggest transport of reworked materials. Locally derived fragments, however, are subangular, suggesting that they were not transported far. Conglomeratic rocks in the lower (Shields), middle (Wilhite), and upper (Sandsuck) parts of the Group suggest debris flows that had been deposited near a shallow water shelf. Slumps, debris flows, and submarine deposits (King, 1964) of clastic carbonate, blocks of carbonate rocks, limy sands, intraformational breccias, and storm deposits, suggests a local source near sea level (Hanselman and others, 1974). The influx of fine to coarse clastic debris was episodic. Local channel and basinal deposits of quartz-pebble conglomerate were probably partly eroded prior to the deposition of the overlying Cochran Formation.

Chilhowee Group

Rocks of the Chilhowee Group are terrigeneous clastic deposits that record the stabilization of the continental margin after sedimentation in the Ocoee basin. Feldspathic sandstone and conglomerate of the lower part of the Chilhowee Group are transitional above the coarse, poorly sorted, arkosic rocks of the Walden Creek Group. Traditionally, they have been interpreted as fluvial deposits (Walker and Driese, 1991). The top of the Cochran Formation has the first occurrence of quartz arenite. The overlying Nichols Shale and the lower half of the Nebo Quartzite consist of feldspathic sandstone interbedded with siltstone and quartz arenite. The first trace fossils were found in these rocks. The overlying Murray Formation is siltstone, which in turn is overlain by quartz arenite of the Hesse Formation. Arkosic sands decrease upward into the middle part of the section, where quartz sand is interbedded with silty deposits, and is finally succeeded by mature weathered deposits of pure, well sorted quartz sands. These rocks reflect early transgressive shallow-marine deposits on the newly formed foreshore, shoreface, and shelf of the Iapetus shore of North America (Simpson and Eriksson, 1989). These rocks are succeeded by the carbonate bank deposits of the Shady Dolomite. Walker and Driese (1991) used trace fossils and body fossils to suggest that the Neoproterozoic-Early Cambrian boundary probably occurs within the lower to middle part of the Nichols Shale on Chilhowee Mountain.

Cambrian, Ordovician, Devonian and Mississippian Rocks

Cambrian and Lower Ordovician rocks, from the Shady Dolomite to the top of the Knox Group, constitute a thick sequence of limestone and dolomite interbedded with calcareous shale of the Conasauga Group. The Rome Formation, between the Shady Dolomite and the carbonate rocks of the Knox Group, resulted from the influx of terrigenous sediment onto the carbonate bank. A subsiding Appalachian basin created an intrashelf basin for deposition of rocks of the Conasauga Group. These shallow-water marine carbonate shelf units were interbedded with fine-grained clastic rocks, followed by deposition of carbonate rocks of the Knox Group. The top of the Knox Group was truncated by a regional unconformity resulting from the Taconic orogeny. The carbonate shelf was unstable. Existing strata were eroded and flysh derived from a highland was deposited.

These marine rocks are overlain by red calcareous mudrock and sandstone of the Bays Formation that was subaerial in part (Neuman, 1955). This erosional debris marks a major change in the environment of deposition in the southern Appalachian basin. Uplift and erosion prior to Late Devonian (?) and early Late Mississippian time removed all rocks above the Middle Ordovician Bays Formation producing a major unconformity. The unconformity represents about 100 My of time. There are no Silurian rocks and only 10 ft (3 m) of the Devonian Chattanooga Shale were preserved.

Minerals

Sulfide Minerals

Metamorphosed stratabound massive sulfide deposits of probable Besshi-type (Fox, 1984) were mined for copper and zinc, which contain trace amounts of lead, silver, and gold, between 1926 and 1944. The ore was taken from the Fontana copper mine, north of the confluence of Ecoah Branch and Eagle Creek (Espenshade, 1943 and 1963; Robinson and others, 1992), on the southeast nose of Pinnacle Ridge. The host rock of the Fontana mine is the graphitic and sulfidic metasiltstone of the Copperhill Formation (Wiener and Merschat, 1992). Besshi-type massive sulfide deposits formed in extensional marine basins that were filled by clastic sediments containing variable amounts of tholeiitic basalt (Fox, 1984). Like the meta-igneous rocks near similar sulfide deposits (Fox, 1984; Misra and Lawson, 1988), metadiabase and metadiorite dikes and sills are common at the Fontana copper mine area, and at mines and prospects along strike to the northeast. However, metadiabase dikes and greenstone were not reported at these mines (Espenshade, 1963). The association of igneous dikes, carbonate-chlorite schists, and massive sulfide deposits is well documented (Laney, 1907; Espenshade, 1963), but their association is still unresolved. They occur within the Eagle Creek shear zone that has transposed the bedding or layering of all units into the foliation. Specimens of folded sulfide deposits show that the mineralization preceded deformation. The Redmond mine exploited a small lead-zinc deposit near Sheldon Laurel at the east margin of the map (Espenshade, 1943; Espenshade and others, 1947; Hadley and Goldsmith, 1963). The ore deposit was explored and mined between 1905 and 1943. It consists of a series of sulfide-bearing quartz lenses within sheared and faulted basement gneiss and garnet-actinolite-chlorite schist of the Wading Branch Formation and Longarm Quartzite.

Pegmatite

Paleozoic pegmatites, which were intruded into the basement gneiss and Copperhill Formation near Bryson City were mined for clay and feldspar during the 1930's and 1940's (Cameron, 1951).

Secondary Minerals

Fine-grained soluble, secondary, salt minerals currently precipitate from the evaporation of acidic metal-bearing groundwater seeps. The minerals copiapite, alunogen, gypsum, halotrichite, pickeringite, melanterite, rozenite, slavikite, epsomite, and starkeyite, have been identified at the bluffs of Alum Cave (Flohr and others, 1995; Hammarstrom and others, 2003). The occurrence of alum, a potassium aluminum sulfate mineral, at Alum Cave is questionable, although shoveled earth and water was mixed in hoppers at Alum Cave during the Civil War to extract “alam” to tan leather, salt peter for gun powder, copper for dye, and salts for medicine. The Epsom Salts Manufactoring Company exploited magnesium sulphate for use as a cure-all. Neither alum nor salt peter were reported to be present at Alum Cave in 1837.

Structure

Introduction

The rocks of the GSMNP region were structurally deformed, tectonically assembled, and metamorphosed, during several episodes over a period of about 900 My, beginning in the Mesoproterozoic (about 1194 Ma) and ending in the late Paleozoic era (about 285 Ma). The deformation and metamorphism of the rocks occurred tens of kilometers deep within the crust and at least 60 mi (100 km) to the southeast of the park (Hatcher, 1978). They were transported westward, on the Great Smoky fault subsequent to the early Mississippian, and later were folded and faulted. Several phases of faulting, folding, and cleavage development, occurred throughout the Paleozoic. The predominant foliations shown on the map are bedding in the sedimentary rocks, schistosity and cleavage (slaty, crenulation, and shear band varieties) in the metamorphic rocks, and gneissic banding in the Mesoproterozoic rocks. Other structural elements in the rocks are lineations, folds, and faults. The bedding, foliation, cleavage, and fold data shown on the map were compiled from published maps (see index) and include some new data. These data represent only a small part of the existing structural data. Outcrop-scale folds are shown on the map, but regional-scale folds were shown separately on the tectonic figure on the plate. The style and time of deformation was different in each region, from the highlands and foothills of the western Blue Ridge, to the Tennessee Valley of the Valley and Ridge province.

Four major structural systems constitute the geologic framework of the GSMNP region (fig. 2)(from southeast to northwest): 1) the Greenbriar fault, Dunn Creek fault, and related faults and folds in the Blue Ridge highlands, 2) the Great Smoky fault, Miller Cove fault, and related faults and folds in the Blue Ridge foothills, 3) the late Gatlinburg fault system and Pigeon Forge fault also in the foothills, and 4) the thrust sheets of the Tennessee Valley bounded by the Pine Mountain thrust fault (floor) and the Great Smoky fault (roof). The Greenbriar, Dunn Creek, and related faults were early and pre-dated metamorphism. The rocks above these faults have east-trending folds. The Greenbriar fault was folded about northeast-striking axes. The Dunn Creek fault truncated northeast-trending folds and cleavage in rocks of the Walden Creek Group. The Miller Cove fault has juxtaposed deformed rocks of the Walden Creek Group above unmetamorphosed rocks of the Chilhowee Group. The Pigeon Forge fault may be a tear fault above a northeast-dipping (?) lateral ramp formed during transport of the Great Smoky fault. The Gatlinburg fault system and the related Oconoluftee fault, have truncated the Pigeon Forge fault and have deformed rocks of the highlands and foothills by a series of steep northeast-trending right-lateral strike slip faults.

The chlorite-biotite zone marker is along the Gatlinburg fault system. The easterly-trending staurolite and kyanite zone markers at the east end of the park were deflected along northeast-trending folds parallel to late shear zones associated with the folds. These north-northeast-trending folds and faults have overprinted the early east-trending folds in the eastern half of the highlands. The north-northeast-trending cleavage has transected more easterly structures throughout the park and may have formed during this late folding event. Similarly north-northeast-trending folds in the Tennessee Valley in the northwest part of the map area have been truncated by thrust faults having a more easterly trend. Early tectonic transport was to the north in the hinterland (Blue Ridge). Later transport, probably in two different phases, was to the north-northwest and later to the northwest.

Gneissic Foliation

The oldest planar structures are gneissic foliation in the Mesoproterozoic rocks. From west to east, these rocks crop out in the Bryson City dome, Ela dome, Ravenfork anticlinorium, Cataloochee anticlinorium, the Soco Gap antiform, and the Maggie Valley-Dellwood domes. Paleozoic deformation and metamorphism at staurolite-kyanite grade has largely overprinted Mesoproterozoic foliations and folds. The foliation and folds can be recognized locally in outcrop but could not be used to define map-scale folds. Foliation formed during the Grenville orogeny was locally preserved in the interior of the Paleozoic domes where it is defined by aligned biotite, hornblende, orthopyroxene, large garnet, granoblastic quartz, and feldspar. The mineral assemblage hornblende-orthopyroxene-microcline in textural equilibrium indicates granulite- facies metamorphism. Migmatitic biotite gneiss and hornblende-biotite gneiss contain isoclinal folds of granitic leucosome and melanosome of biotite-hornblende schist and amphibolite.

U-Pb geochronology (Southworth and Aleinikoff, in press) suggests that the foliation formed during several episodes of Grenvillian deformation. Migmatitic gneiss, 1194+-7 Ma, suggests partial melting during a deformation event prior to the intrusion of the granitoids of Group 2 between 1178-1117 Ma. Rocks of Group 1 and Group 2 were both deformed between 1117 and 1056 Ma (Southworth and Aleinikoff, in press).

Bedding

Bedding in the sedimentary rocks was recognized by a grading, fine laminations marked by graded microscopic grains, color differences related to primary minerals and cement, cross-beds, ripple marks, and accumulations of rock fragments. Concretions are a false indicator of bedding as they are locally aligned parallel to cleavage. The younging direction of beds is difficult to determine in many of the fine-grained and conglomeratic rocks.

Schistosity

A pervasive schistosity is the dominant foliation in the garnet- to kyanite- grade schists of the highlands as well as the metadiabase dikes. Bedding in schist was locally recognized by color changes related to mineral differences.

Cleavage

Cleavage is pervasive in rocks of the highlands and much of the foothills; it is best developed in fine-grained rocks. Cleavage formed during several different deformation events making the definition of specific cleavages to separate events problematic. Most of the cleavage was related to folding but some was related to faulting. Most cleavage is axial planar to folds, but other cleavage transects fold axes. The earliest fold phase has no recognized axial planar cleavage. Bedding could locally have been transposed into it. The early folds however, are transected by later cleavage. This later cleavage cannot be attributed to any fold event. Later folds have axial planar cleavage that conforms to the fold limbs. This cleavage is marked by chlorite and sericite crystals that cut and replace porphyroblasts of garnet, muscovite, and biotite. Iron-oxide in the cleavage suggests that it may have formed by pressure solution. Slaty cleavage related to later folds and faults transects the schistosity. Early high-grade minerals are retrograde to chlorite and sericite of greenschist-facies conditions.

The relations of bedding and cleavage were used to determine the position of fold geometry, but locally however, folds and cleavage have been arched into broad, open folds. This is the case on the margins of the tectonic windows through the Great Smoky fault (Neuman and Nelson, 1965).

Cleavage associated with shearing and faulting was recognized in the Eagle Creek shear zone, the Cold Springs fault, the Metcalf Phyllite in the Little River shear zone, and the Ordovician Tellico Shale between the Great Smoky and Guess Creek faults in the Tennessee Valley.

Phyllite, slate, and schist locally have a spaced, crenulation cleavage that has crinkled the earlier cleavage and schistosity. It has various orientations but is mostly vertical. Crenulation cleavage in Metcalf Phyllite around Cades Cove was related to the late Paleozoic doming of the Great Smoky fault.

Sheared fine-grained rocks such as the Metcalf Phyllite have a phyllonitic foliation and shear band cleavage that formed as the result of northwest-directed thrust faulting. The asymmetric top-to-the-northwest shear band cleavage overprints the bedding that is perpendicular to the cleavage. Thin layers of vein quartz have been transposed into the foliation. The Wading Branch Formation near Harmon Den also contains phyllonitic foliation and shear band cleavage near the Cold Spring fault.

Mylonitic Foliation

Mylonitic foliation contains recrystallized and anhealed minerals that were reduced to grain-size by shearing. Mylonitic foliation is seen locally in Mesoproterozoic rocks, Longarm Quartzite, and metasandstone of the Copperhill Formation. Some mylonitic foliation in the Mesoproterozoic gneiss in the region probably formed during the Mesoproterozoic Grenville orogeny (Mershat and Wiener, 1988). Similar mylonitic rocks in the Cherokee-Ravenfork belt, the Qualla-Dellwood belt, and in the Bryson City dome, strike into the contact between the Mesoproterozoic and Neoproterozoic rocks, which was not cut. They too may have formed during the Grenvillian deformation. Other mylonitic rocks, such as the monzogranitic gneiss and border gneiss along the west margin of the Bryson City dome (Cameron, 1951), result from Paleozoic deformation. Mylonitic foliation in biotite augen gneiss, Longarm Quartzite, and Copperhill Formation near the Ravenfork anticlinorium suggests that these occurrences are related to formation of the antiform.

Lineations

Lineations observed are both intersection and stretching types. Fine-grained metasedimentary rocks commonly contain a conspicuous lineation of bedding-cleavage intersection, which parallel the plunge of folds. The intersection of cleavage and (or) schistosity with crenulation cleavage marks the attitude of crinkles in the fine-grained rocks. Mineral stretching lineations are common in highly deformed rocks such as the Eagle Creek shear zone and Ravenfork antiformal window. They parallel the fold axes. Lineations and mylonitic foliation in the granitic gneiss of the Ravenfork antiform define an L-S tectonite that probably formed during the Paleozoic folding event.

Folds

Introduction

All of the rocks are folded, but their age and style varies across the region (fig. 3). The oldest folds formed during the high temperature ductile deformation ~1000 Ma Grenvillian orogeny. The youngest folds are broad arches that post-date the 280 Ma emplacement of the Great Smoky fault during the late Paleozoic Alleghanian orogeny.

Additional folds formed by soft sediment deformation during deposition during the Neoproterozoic ~830 to 545 Ma (Aleinikoff and others, 2004), and several other phases formed during Paleozoic deformation.

Mesoproterozoic Folds

Isoclinal and transposed folds recognized in outcrops in migmatitic gneiss at the south end of the Ela dome formed during the Grenville orogeny. These folds were refolded by upright structures during Paleozoic doming. Map relations and outcrop data near Dellwood (Hadley and Goldsmith, 1963) suggests that there are many Grenville folds in that region, but they have been obscured by Paleozoic deformation.

Neoproterozoic Soft Sediment Folds

Rocks of the Ocoee Supergroup contain folds that formed by gravitational slumping during deposition and prior to lithification from about ~830 to 545 Ma (Aleinikoff and others, 2004). The slump folds are in rocks of the Longarm Quartzite, Roaring Fork Sandstone, Thunderhead Sandstone, Cades Sandstone, Anakeesta Formation, and Copperhill Formation. They range from 2 in (4cm)-scale to about 6 ft (2 m) wavelength as seen in Cades Sandstone at Abrams Falls. Isoclinal folds of metasandstone and dolomite with black slate in the bluffs of Alum Cave are interpreted to be soft sediment folds because the cleavage transects them and they have no evident axial plane cleavage.

Paleozoic Folds in the Highlands

Alum Cave Synclinorium

Hadley and Goldsmith (1963) described the east-west trending Alum Cave syncline as a complex syncline east of the Oconaluftee fault, centered on Mount Kephart. It is named herein the Alum Cave synclinorium, because it contains many faults and subordinate folds. Motion along thrust faults, such as the Mingus fault, folded rocks in their footwalls during displacement. The truncation of beds within the Anakeesta Formation on the northern upright limb of the synclinorium suggests an unconformity or a fault. The rocks on the upright limb do not occur on the overturned south limb. Massive, thick-bedded metasandstone of the Thunderhead on the north limb grades into thinner-bedded metasandstone that is interbedded with schist on the south limb. The beds on the south limb dip steeply and are locally overturned in the footwall of a fault that has juxtaposed rocks of the Copperhill Formation.

West of the Oconoluftee fault, the rocks of the Anakeesta Formation form a homoclinal sequence, and are interbedded with metasandstone, and overlain by metasandstone and slate. Within the main syncline east of the Oconoluftee fault, cleavage strikes N40-70E (Hadley and Goldsmith, 1963), and includes both an axial planar cleavage and a later transecting cleavage. Folds plunge 5-35 degrees east-northeast, but many folds plunge east-southeast and west-southwest (Hadley and Goldsmith, 1963).

Murphy Synclinorium

The Murphy synclinorium plunges southwest from the area near Deep Creek, north of Bryson City, NC. In the map area it contains rocks of the uppermost Great Smoky Group (Wehutty Formation) and the lower part of the Murphy belt (the Tusquitee Quartzite and Nantahala Formation). The synclinorium extends southwest almost 124 mi (200 km) into Georgia (Higgins and others, 1988), where it contains rocks as young as Ordovician age, presumably correlative to those in the Valley and Ridge (Tull and others, 1998). It is a composite of early folds that have been overprinted by later north-northeast trending folds (Mohr, 1973). Map patterns and foliation attitudes of the Copperhill Formation suggest that the synclinorium may extend northward to the Oconaluftee fault. Secondary, parasitic folds in rocks of the Wehutty Formation in the synclinorium are abundant along Lakeshore Drive, northwest of Bryson City.

Highly folded rocks of the Copperhill Formation along Fontana Lake contain an axial planar cleavage related to a single fold phase. At least three orders of upright to inclined folds that plunge gently to the northeast and southwest are exposed here. Large, second-order folds verge westward to the syncline at Hazel Creek. West of Hazel Creek, rocks in the Eagle Creek shear zone contain ptygmatic, isoclinal, intrafolial, and upright folds (Southworth, 1995).

Basement Domes and Anticlinoria

Domes that expose basement gneiss are associated with young north-northeast-trending folds. The contacts between the mylonitic gneiss and cover rocks of the domes are interpreted as faults (Hadley and Nelson, 1971) forming tectonic windows similar to Cades Cove. They are probably early faults related to the Greenbriar fault, but they could be younger Paleozoic faults. The early faults are folded with the dome, and locally they are cut by later faults, particularly around the Ravenfork window. These later faults were either synchronous with folding or post-date it. An example is the Oconaluftee fault which cuts the axial surface of the anticlinoria. Older east-northeast-trending folds are overprinted by younger, north-northeast-trending folds; the fold interference produced warped attitudes of foliations.

Paleozoic Folds in the Southern Foothills

Cartertown, Copeland Creek and Chestnut Mountain Anticlines

The Cartertown, Copeland Creek, and Chestnut Mountain anticlines east of Gatlinburg have folded the Roaring Fork Sandstone and the overlying Pigeon Siltstone. The folded rocks strike generally N80E, and have been transected by cleavage that strikes between N45-60E. Hadley and Goldsmith (1963) and Hamilton (1961) interpreted these relations to indicate that early east-west- trending folds were transected by northeast-trending cleavage formed during later shearing. Woodward and Connolly (1992) interpreted the folds to be pre-metamorphic structures that were transected by cleavage formed during Ordovician deformation. These folds have been locally cut by later northeast-striking thrust faults, including those of the Gatlinburg fault system.

The Chestnut Mountain anticline at the northeast end of the park between Snag Mountain and Midway may be part of the Copeland Creek anticline. The Chestnut Mountain anticline has also been truncated by faults of the Gatlinburg system as well as by the Snag Mountain fault. The Snag Mountain fault may be part of the late Gatlinburg system or somehow associated with the early Greenbriar fault.

Bedding form lines suggest that the Copeland Creek anticline, the Greenbriar fault, and the overlying rocks of the Great Smoky Group, have been folded into an anticline that plunges east. These relations were interpreted to result from the Greenbriar fault truncation of early folds by the Greenbriar fault (Woodward and Connolly, 1992).

Alternatively, the relations can be taken to suggest that the Greenbriar fault has cut upsection in both plates, and was later folded.

Paleozoic Folds in the Northern Foothills

A broad northeast-plunging synclinorium of rocks of the Walden Creek Group is east of the Pigeon Forge fault. Two large synclines on the southeast limb of Chilhowee Mountain, west of the Pigeon Forge fault, probably formed as footwall synclines during motion along the Miller Cove fault and Bogle Spring fault. Two broad synclines containing rocks of the Wilhite Formation, northeast of Miller Cove, may also be related to footwall deformation along thrust faults within the Walden Creek Group. Along the west margin of the study area along Chilhowee Lake are recumbent folds of rocks of the Wilhite Formation. These folds have axial planar cleavage and have been refolded into broad domes and swales. The folded rocks and associated cleavage in this region were broadly arched above the tectonic windows (Neuman and Nelson, 1965).

Paleozoic Folds in the Tennessee Valley of the Valley and Ridge

The Guess Creek thrust sheet southeast of Sevierville contains the Fair Garden anticline and Rock Quarry Dome. The Fair Garden anticline can be aligned with the broad arches centered on the tectonic windows through the Great Smoky fault in the foothills if the displacement along the left lateral Pigeon Forge fault is restored. This suggests blind thrust faults and duplex beneath the Guess Creek and Great Smoky thrust sheets.

Rocks of the Knox Group have been folded into anticlines above thrust faults from Blockhouse northwest to Maryville. Near Alcoa, a structural culmination containing rocks as old as the Early Cambrian Rome Formation lies above the Dumplin Valley thrust near Bays Mountain. Folds in the Alcoa region trend north-northeast and have been truncated by later northeast-striking faults.

Faults

Introduction

Faults were identified by the omission of strata, mylonitic foliation, gouge and (or) breccia. Premetamorphic faults are characterized by the omission of strata and discordant beds; they have been folded and cut by later faults and cleavage, making their recognition difficult. Post-metamorphic and post-early deformation faults are associated with discontinuities in folds, truncated strata and cleavage, and by retrograde minerals such as chlorite and sericite. Some late faults formed cleavage that is co-planar to the fault. Some of the latest faults, like the Oconaluftee and Gatlinburg faults, have a pronounced topographic expression and may be Mesozoic or younger in age (Rast and Kohles, 1986; Naeser and others, 2004).

Regional faults have divided the region into distinct geologic terranes (fig. 4). The Greenbriar fault in the highlands is an early, premetamorphic low-angle fault that has been folded. Related faults are the Rabbit Creek and Dunn Creek faults in the foothills and the basement-cover faults bounding the windows in the highlands. The Great Smoky and related low-angle thrust faults have transported all of the rocks westward and mark the northwestern boundary between the western Blue Ridge and the Tennessee Valley. The Pigeon Forge fault is a transverse fault that had left lateral and down-to-the southwest displacement. It divides the foothills into western and eastern blocks, and it may have been synchronous with or post-dated the Great Smoky fault. The Gatlinburg fault system consists of a series of steep thrust faults that had a component of right-lateral displacement. Faults of this later system cut all the other faults.

Neoproterozoic Faults in the Highlands

The Neoproterozoic extensional deformation that formed the basins containing rocks of the Ocoee Supergroup likely resulted in faults that were active during sedimentation as evidenced by soft-sediment folds and microfaults. Small outcrop-scale offsets in graded-beds within the Thunderhead Sandstone suggest such normal faults. The truncation of stratigraphic units along contacts, such as the basal units of the Anakeesta Formation on Mt. Le Conte, could mark an unconformity or be the result of syndepositional faulting. Recognition of such faults is obviously difficult given the subsequent complex history of the rocks, but some evidence suggests that they exist. The Caldwell Fork fault and related faults near Cataloochee are later steep thrust faults that are marked by retrograde fabrics

(Montes and Hatcher, 1999). They interpreted the Caldwell Fork fault to be a late thrust fault having minor displacement formed by reactivation of a Neoproterozoic fault that had been active during sedimentation of the Longarm Quartzite. Stratigraphic facies changes and changes in thickness across the fault suggest that the down-to-the-northwest normal growth fault was reactivated by top-to-the-northwest motion during later contractional deformation. The Greenbriar fault could have had a similar origin. Rather than being an early Paleozoic contractional fault (Kin and others, 1968; Woodward and Connolly, 1992), it could have been a Neoproterozoic detachment fault (submarine slide?) active during Ocoee basin sedimentation.

Early Paleozoic Faults in the Highlands

Greenbriar Fault

Perhaps the most enigmatic fault in the entire Blue Ridge province is named after exposures on Greenbriar Pinnacle. The Greenbriar fault (Hadley and Goldsmith, 1963) has placed rocks of the Great Smoky Group onto older rocks of the Snowbird Group. From the Sugarland Valley east, Elkmont Sandstone and Thunderhead Sandstone are structurally above rocks of the Roaring Fork Sandstone and Pigeon Siltstone. Within a horizontal distance of 300 ft (100 m), beds both above and below the fault can rotate as much as 90 degrees. The fault can be traced eastward around Mount Cammerer and Mount Sterling and eventually beyond Cataloochee. At the east end of the Park near Big Creek, the fault was later folded into a series of synclines and anticlines. Metamorphic zone markers cross the fault without deflection or truncation, therefore, the Greenbriar fault was early and predated metamorphism and folding. West of Gatlinburg, the Greenbriar fault was dismembered by later faults and folds, so there are few exposures of it. Thunderhead Sandstone above Pigeon Siltstone on the north end of Cove Mountain, and Elkmont Sandstone above Metcalf Phyllite south of Cades Cove, probably define the location of the original Greenbriar fault. The elongate body of metasandstone along Little River to White Oak Sink is a klippe of the hangingwall rocks above the Greenbriar fault that were folded and later faulted (King and others, 1965), into a duplex within the belt of Metcalf Phyllite, herein called the Little River shear zone. Early foliation strikes into the faults and later shear band cleavage is parallel to the faults. This suggests that the early faults truncated strata and pervasive cleavage, and that shear band cleavage formed during later faulting. Several later faults, probably members of the Gatlinburg fault system reactivated segments of the Greenbriar fault in this area. The contact between Thunderhead Sandstone and Metcalf Phyllite, from the Big Spring Cove area to the north side of Cove Mountain, was called the Greenbriar fault by King (1964). The contact between the Thunderhead Sandstone and the Pigeon Siltstone at the northeast end of Cove Mountain was folded and cut by faults of the late Gatlinburg fault system. Field relations, however, suggest that parts of the Greenbriar fault were here reactivated. Along Little River, both hanging-wall and footwall rocks have strong cleavage that is parallel to the fault, suggesting that faulting and cleavage development was synchronous. Along the fault to the northeast, strong cleavage along the fault suggests later reactivation. Slices of Cades Sandstone and Metcalf Phyllite around Cades Cove also are early thrust sheets that were cut by later faults.

The restoration of the right-lateral strike slip faults of the Gatlinburg fault system allows the reconstruction of a large Greenbriar thrust sheet. The fault cuts downsection to the north in the upper plate, and cuts upsection northward in the lower plate. The rocks of the two plates were interpreted to have originated in separate depositional basins (King and others, 1968; Rast and Kohles, 1986) that were juxtaposed along the fault, metamorphosed, folded, and later faulted.

Faults Slices beneath the Greenbriar Fault

Some faults predate and (or) were synchronous with the Greenbriar fault. North of Greenbriar Pinnacle and Maddron Bald, fault-bounded rocks of the Rich Butt Sandstone are beneath the Greenbriar fault (Hadley and Goldsmith, 1963). They interpreted these to be fault slivers derived from the Roaring Fork Sandstone that crops out near Mount Cammerer, during the northwestward transport of the Greenbriar thrust sheet. The relation between the Snag Mountain fault and the Greenbriar fault warrants further study. Hadley and Nelson (1971) suggested that the Snag Mountain fault cuts the Chestnut Mountain anticline, the continuation of the Copeland Creek anticline. This interpretation suggested that the anticlines and the Snag Mountain fault predate, or were synchronous with the Greenbriar fault, whereas the map relations suggest that the Greenbriar fault cuts the Snag Mountain fault. Alternatively, the Snag Mountain fault may be part of the late Gatlinburg fault system that overprints the Greenbriar fault at Greenbriar Pinnacle.

The northern part of the Ravensford anticlinorium has a lower thrust sheet of gneiss that encloses a thin slice of Longarm Quartzite in a window through the bounding (Greenbriar?) fault. Hadley and Goldsmith (1963) assumed the gneiss was thrust westward over rocks of the Snowbird Group, which were in turn overridden by rocks of the Greenbriar thrust sheet, and were subsequently folded and faulted. Along the Greenbriar fault, on the north side of Cove Mountain, Mesoproterozoic rocks between rocks of the Thunderhead Sandstone, Pigeon Siltstone, and Metcalf Phyllite, may be a horse similar to the structures in Ravensford.

Windows through the Greenbriar and Related Faults

The contact between the rocks of the Great Smoky Group and the Mesoproterozoic gneiss and rocks of the Snowbird Group in the southeastern part of the highlands has long been interpreted as the Greenbriar fault (Hadley and Goldsmith, 1963; Hadley and Nelson, 1971). The contacts are interpreted in this report as faults related to the Greenbriar fault. These contacts may also have been unconformities that were faulted and folded.

Late Paleozoic Faults and Shear Zones in the Highlands

Introduction

A series of thrust faults post-dated the peak metamorphism in the Paleozoic and are associated with later folding. The Eagle Creek shear zone (Southworth, 1995) and Mingus fault (King, 1964), and several unnamed faults, are characterized by truncated units and discordant folds within slate of the Anakeesta and Copperhill Formations. Garnet and biotite were retrogressively metamorphosed during movement of these faults. Faults between the Thunderhead Sandstone, Metcalf Phyllite, and Cades Sandstone, define the broad Little River shear zone. Thrust faults in the eastern part of the study area, like the Cold Springs fault, Caldwell Fork fault, and the unnamed thrust fault near Cherokee, have involved Mesoproterozoic rocks and have cut the Cataloochee anticlinorium and the Ravenfork anticlinorium. The results of contractional deformation along these shear zones helps to better define the late structural front first portrayed by Hadley and Goldsmith (1963) in the eastern half of the study area. This area of late north-northeast–trending folds and cleavage are parallel to and bounded by the late shear zones.

Eagle Creek Shear Zone

Late northwest-directed thrust faults in the Eagle Creek shear zone are within the Copperhill Formation. Folds, thrust faults, cleavage, and mylonitic, phyllonitic, and transposition foliations, are probably a continuum of deformation that post-dated peak metamorphism. Early foliation in the garnet- and biotite-grade rocks has been crenulated. Garnet and biotite have been retrogressed to sericite and chlorite. Mappable bodies of vein quartz and retrogressively metamorphised carbonate-chlorite schist along faults (Southworth, 1995) suggests a substantial migration of fluids during deformation.

Little River Shear Zone

West of the Pigeon Forge fault, the Pigeon Siltstone has been placed on rocks of the Walden Creek Group. West of Starkeytown, the Pigeon Siltstone was sheared to produce the Metacalf Phyllite along a late fault zone that King (1964) called “the Raven Den slices”. The Raven Den slices consist of conglomeratic rocks of the Shields Formation, not Thunderhead Sandstone as depicted by King (1964). He noted the complex structure here and that “the intensity of deformation was remarkable and must be largely cataclastic and dynamic”. The Line Spring fault of King (1964) placed Metcalf Phyllite on rocks of the Walden Creek Group, and is probably a remnant of the Dunn Creek fault, which was truncated along strike by the Great Smoky fault. The Metcalf Phyllite defines the broad Little River shear zone, within which the original Greenbriar fault was folded and faulted by the later Gatlinburg fault system.

Cold Spring and Caldwell Fork Faults

A series of late thrust faults are along the east side of the map area from approximately Harmon Den south to near Shelton Laurel, NC. A pair of these late faults forms a duplex of basement gneiss and rocks of the Snowbird Group, which was called the Cold Spring fault at Harmon Den (Hadley and Goldsmith, 1963). Schist of the

Wading Branch Formation is shear banded phyllonite that superficially resembles rocks of the Metcalf Phyllite. Mylonitic Mesoproterozoic rocks superficially resemble the Longarm Quartzite. The mylonite contains abundant porphyroclasts of pink potassium feldspar and chlorite/epidote, indicating metasomatic mineralization and hydrothermal alteration. From west of the Pigeon River to Cataloochee Creek, mylonitic quartzofeldspathic gneiss has been transected by pervasive cleavage that is parallel to the fault. These faults were not traced to the southwest, but they may continue to the Ravenfork anticlinorium near Cherokee. The Caldwell Fork and related faults near Cataloochee are steep later thrust faults that contain retrograde fabrics (Montes and Hatcher, 1999).

Gatlinburg Fault System

The Gatlinburg fault system is the latest to cut rocks of the highlands. It consists of numerous fault strands that define the northwestern boundary of the highlands. The rocks of the fault zone are brittle, highly fractured iron oxide stained cataclasites and breccias. The Oconaluftee fault has splayed from the system and transects the central part of the park at Newfound Gap. It has experienced right-lateral and down-to-the-southwest motion and can be traced east to the northwest limb of the Ravenfork anticlinorium where it is a northwest-dipping back thrust. The fault has profoundly affected the geometry of the rocks and structures of the Great Smoky Group. Fission track data suggests that strands of the Gatlinburg fault system were active at around 285 Ma, and experienced subsequent post-Cretaceous motion (Naeser and others, 2004). Many of these late faults have cut the rocks of the Pigeon Siltstone, but mapping them is difficult due to poor exposure and the monotonous nature of the rocks.

Faults of the Foothills

Much of the northern foothills are underlain by rocks of the Walden Creek Group that are bounded on the north by the Miller Cove fault and associated splays. The foothills are different west and east of the Pigeon Forge fault. The southern boundary of the foothills is the Dunn Creek fault, east of the Pigeon Forge fault. The Dunn Creek, Line Springs, Rabbit Creek, and Great Smoky faults mark the southern boundary west of the Pigeon

Forge fault. The region west of the Pigeon Forge fault is more complex than the area east of the fault in that it contains many early folds, cleavage, and faults that have been cut by later folds and faults. A broad northeast-plunging synclinoria containing rocks of the Walden Creek Group is east of the Pigeon Forge fault. These rocks have been cut by east-trending faults that have been in turn cut by northeast-trending faults. The folds and cleavage in these rocks were folded and faulted during the emplacement of the Great Smoky thrust sheet.

West of the Pigeon Forge fault, the rocks of the Walden Creek Group are in a series of thrust sheets in which the Shields Formation has been placed on the Wilhite Formation. From northeast to southwest, these are floored by the Walden Creek fault, the Happy Hollow fault, and the Capshaw Branch fault. The Capshaw Branch sheet is a klippe (Neuman and Nelson, 1965) that has been cut by the Carr Creek fault. The Carr Creek fault has reactivated the western part of the Rabbit Creek fault. The broad Bates Mountain syncline and the Little Rocky syncline containing rocks of the Wilhite Formation are north of the Capshaw Branch fault and Happy Hollow fault (King, 1964). The Walden Creek fault has truncated existing faults in the thrust sheet. In summary, early faults have been cut by later faults, which in turn were cut by the Great Smoky fault (King, 1964). Even later, the folds, cleavage, and faults, were arched over the tectonic windows.

Dunn Creek Fault

East of the Pigeon Forge fault, Hamilton (1961) described the contact of Pigeon Siltstone above rocks of the Walden Creek Group to be the premetamorphic Dunn Creek fault. Pigeon Siltstone is in the hanging-wall of the fault, and the fault truncated rocks of numerous formations as well as folds and faults in the Walden Creek Group. Bedding and cleavage in the footwall are mostly parallel to the fault. The Dunn Creek fault was cut at an oblique angle by later thrust faults, and was folded at its eastern termination near Green Mountain. Like the Greenbriar fault, the Dunn Creek fault is a complex reactivated fault. The Dunn Creek fault cuts earlier folds and faults, and was folded and cut by later faults and transected by cleavage. West of the Pigeon Forge fault, the Pigeon Siltstone was thrust onto rocks of the Walden Creek Group as late as 280 Ma (Naeser and others, 2004), so it is not the Dunn Creek fault.

Rabbit Creek Fault

The Rabbit Creek fault (Neuman and Nelson, 1965) placed the Cades Sandstone on rocks of the Walden Creek Group northwest of Cades Cove. Because the Cades Sandstone is here considered to be a facies of the Thunderhead and Elkmont Sandstones, the Rabbit Creek fault is probably part of the early Greenbriar fault system (King and others, 1968). The loop at the northern part of the Rabbit Creek fault suggests that it is a low angle thrust fault. The southeastern part of this fault was cut by the Great Smoky fault at Tuckaleechee Cove. The northwest part of the Rabbit Creek fault was cut by the younger Carr Creek fault (Neuman and Nelson, 1965).

Capshaw Branch Fault

The Capshaw Branch fault bounds a klippe of rocks of the Shields Formation resting on rocks of the Wilhite Formation. The southeast side of this fault is truncated by the Rabbit Creek fault, and the younger Carr Creek fault juxtaposes part of the klippe. The structural discordance along the boundary of the klippe can be seen in outcrops along Route 73 at Kinzel Springs.

Webb Mountain-Big Ridge Faults

The rocks that underlie Webb Mountain and Big Ridge have always been considered to be “unclassified” as they are unique in the southern foothills. Hamilton (1961) discussed evidence for both stratigraphic continuity and structural discordance, and he placed the rocks in the hanging walls of early thrust faults. Hadley and Nelson (1971) adopted the interpreted structure and considered the rocks to be undifferentiated Cades Sandstone and Rich Butt Sandstone, that stratigraphically overlie the Pigeon Siltstone. Connolly and Woodward (1992) interpreted the rocks as a klippe of Great Smoky Group rocks, and R.D. Hatcher, Jr. (Univ. of TN, 2003, oral comm.) suggested that the rocks belong to the Great Smoky Group that are exposed in tectonic windows through the Dunn Creek thrust sheet.

These rocks are here classified as Rich Butt Sandstone that is stratigraphically and structurally above the Pigeon Siltstone. The contact with the Pigeon Siltstone is herein interpreted as a fault. The faults within the surrounding Pigeon Siltstone that mark thrust sheets are not shown on the map because of a lack of agreement of previous workers and non-reproduceability (Hamilton, 1961; King and others, 1968; Hadley and Nelson, 1971; Mark Carter and R.D. Hatcher, Jr., 2002, written comm..).

Great Smoky and Related Faults

The Great Smoky fault (Keith, 1895, 1927; Gordon, 1920; Wilson, 1935) is a low angle thrust fault that carried rocks of the Ocoee Supergroup westward over Ordovician carbonate rocks. It is usually about horizontal, but locally is inclined, and has been broadly folded, and cut by later thrusts. These carbonate rocks are exposed in the tectonic windows of the foothills, such as Cades Cove. The faults bounding the coves and the fault on the northwest slope of Chilhowee Mountain have always been called the Great Smoky fault. The Miller Cove fault on the east side of Chilhowee Mountain has placed rocks of the Walden Creek Group on rocks of the Chilhowee Group; Neuman and Nelson (1965) and King (1964) suggested that the Great Smoky and Miller Cove faults are related. In that scenario, the rocks of the Chilhowee Group are a horse between the Great Smoky and the Miller Cove faults. East of the transverse Pigeon Forge fault, the Great Smoky and Miller Cove faults change. Hamilton (1961) interpreted the fault on the south side of English Mountain and Green Mountain as the Great Smoky fault (Walden Creek Group rocks thrust onto Chilhowee Group rocks). He called the fault on the north side of the mountains the English Mountain fault (Chilhowee Group rocks thrust onto Ordovician rocks of the Tennessee Valley). Hamilton’s (1961) rationale for this interpretation was that the hanging-wall rocks in the tectonic windows west of the Pigeon Forge fault are Walden Creek Group, as in his area east of the Pigeon Forge fault. The rocks of the Chilhowee Group, east of the Pigeon Forge fault, were either 1) truncated by the overriding fault having rocks of the Walden Creek Group in its hanging wall or, 2) the Great Smoky fault climbed northeastward along a lateral ramp. In this scenario, the Great Smoky fault at the northwest base of Chilhowee Mountain is a splay from the Miller Cove fault. The rocks of the Walden Creek Group are intensely folded, and have a pervasive cleavage that is folded at many places. Stratigraphic displacement, however, was not great (Sandsuck Formation on Chilhowee Group). The stratigraphic displacement was greatest on the Great Smoky fault along the northwest base of Chilhowee Mountain (Sandsuck Formation on Tellico Shale). The rocks in the hanging-wall were only broadly folded into a southeast-dipping homocline of right-side-up strata with synclines at Top of the World Estates, Miller Cove, and Dupont Springs. The footwall synclines and related anticlines resulted from motion on the Miller Cove fault on the south slope of the mountain. There is little spaced cleavage and the rocks were not metamorphosed and were never at depths to recrystallize U/Pb ratios in detrital zircons (Naeser and others, 2004). Where exposed along Highway 321, the Great Smoky fault dips about 45 degrees to the southeast, and it cuts upsection both to the north and south into the Helenmode Shale near the middle of Chilhowee Mountain. Folds were truncated by transverse faults that had thrust, normal and strike slip displacement. These faults were later cut by the Miller Cove, Bogle Springs, and Great Smoky faults. Therefore, folding of the rocks was probably synchronous with transport and was followed by minor faulting.

Tectonic Windows of the Great Smoky Thrust Sheet

The tectonic windows in the Great Smoky fault are, from southwest to northeast, the Calderwood, Cades Cove, Big Spring Cove, Tuckaleechee Cove, and Wear Cove windows. They are west of the Pigeon Forge fault in the southern foothills, and probably formed on a structural culmination within a duplex in the foothills (Hatcher and others, 1989). The exposed Ordovician Jonesboro Limestone is in a duplex, therefore, it cannot be correlated from window to window. For example, the Jonesboro Limestone exposed in the Cades Cove window is an upper thrust sheet. The Jonesboro Limestone in the Tuckaleechee Cove and Wear Cove windows are part of the lower thrust sheet. The Jonesboro Limestone forms a regional structural dome having its culmination in Tuckaleechee Cove. In contrast, the Great Smoky fault is in a structural depression in Tuckaleechee Cove. Erosion by the Little River has lowered the floor of Tuckaleechee Cove more than 700 ft (200 m), so that the Jonesboro Limestone exposed in the floor of Cades Cove, is high on the margins of Tuckaleechee cove. The broad northeast trending antiform is reflected in cleavage that has been folded into an arch in the surrounding rocks of the Walden Creek Group (King, 1964; Neuman and Nelson, 1965). Hatcher and others (1989) suggested that this antiform extends eastward into the Fair Garden anticline in the Tennessee Valley. These arches, however, are en echelon when the late Pigeon Forge fault displacements are restored. The Great Smoky fault is well exposed at several places in Calderwood, Tuckaleechee Cove, White Oak Sink, along Highway 74 between Townsend and Tremont, and along the road to from Townsend to Wear Cove.

English Mountain, Green Mountain and the English Mountain Fault

English Mountain is underlain by rocks of the Chilhowee Group on the south-dipping limb of an anticline. The fault on the north slope of English Mountain juxtaposed rocks of the Chilhowee Group onto Ordovician rocks of the Tennessee Valley. This structure was called the English Mountain fault by Hamilton (1961) and King and others (1968), but has since been recognized as the Great Smoky fault by Hadley and Nelson (1971). The fault on the south side of English Mountain that has placed rocks of the Walden Creek Group on rocks of the Chilhowee Group was called the Great Smoky fault (Hamilton, 1961; King and others, 1968). Cross sections by Hadley and Nelson (1971) suggested that the Great Smoky fault and Miller Cove fault are the same. Northwest–dipping strata of the Chilhowee Group and Rome Formation on Green Mountain are exposed in a tectonic window.

Pigeon Forge Fault

The Pigeon Forge fault (King, 1964) creates a profound structural break across the central part of the foothills. Erosion along the Pigeon River has removed diagnostic exposures, but offset map patterns provide clues to its interpretation. The fault has left-lateral strike slip motion that had some down to the southwest displacement. The Great Smoky fault, Miller Cove fault, Dunn Creek fault, and other faults cannot be connected across it. Therefore, the Pigeon Forge fault may be a lateral ramp along which the Great Smoky fault climbed up section during transport.

Other Faults

Highly fractured metasandstone containing deformed vein quartz is abundant near the Gatlinburg fault system and are probably late, brittle, intraformational thrust faults that had insignificant offset. These late faults are in the Roaring Fork Sandstone along the Little Pigeon River near Greenbriar. Many outcrops of Roaring Fork Sandstone and Elkmont Sandstone along the Gatlinburg fault system contain similar brittle structures. Lustrous phyllonite at places is brecciated. Micro-brecciated vein quartz near the Greenbriar fault in topographic swales resulted from late fault movements. East striking normal faults that experienced down-to-the-south motion have cut beds and cleavage along Fontana Lake (Southworth, 1995). These small faults are subparallel to the large east trending topographic lineaments in the region (such as Fontana Lake and Yellow Creek to the south), that are interpreted as Mesozoic faults (R. D. Hatcher, Jr., Univ. of TN, 2004, oral comm.) and as post-Mesozoic faults (K. Stewart, Univ. of NC, 2004, oral comm.) faults.

Paleozoic Faults in the Tennessee Valley

The rocks in the Tennessee Valley are mostly folded and faulted carbonates interbedded with fine-grained clastics. The Pine Mountain thrust fault marks the leading edge of a series of imbricate thrust sheets that have transported all of the rocks in this region several hundred mi (300 km) westward (Harris and Milici, 1977). At least 9 major thrust sheets are present between the Great Smoky fault on the east and the Pine Mountain on the Cumberland Plateau on the west.

The map area can be subdivided into several belts and thrust sheets which are the Guess Creek, Dumplin Valley, and Knox thrust sheets. The Guess Creek thrust sheet is bounded by the Guess Creek fault (floor) and the Great Smoky fault (roof). The Guess Creek fault has cut upsection into the hanging-wall rocks from Ordovician Tellico Shale into Mississippian Grainger Formation (to the southwest). The Guess Creek thrust sheet has overridden rocks of the Dumplin Valley thrust sheet. The Dumplin Valley thrust sheet has overridden rocks of the Knoxville thrust sheet. The Knoxville thrust fault is not exposed in the map area. Faults and associated folds in the Knoxville thrust sheet near Alcoa trend north-northeast and have been truncated by later northeast-trending faults.

Metamorphism

Mesoproterozoic

The Mesoproterozoic rocks contain the mineral assemblage hornblende-orthopyroxene-microcline in a textural equilibrium that indicates granulite-facies metamorphism. Migmatitic biotite gneiss and hornblende-biotite gneiss contain isoclinal folded granitic leucosome and melanosome of biotite-hornblende gneiss and schist, and amphibolite. U-Pb geochronology suggests that the migmatitic gneiss (1194+-7 Ma) formed during a partial melting event that preceeded the intrusion of granitoids between 1178-1117 Ma (Southworth and Aleinikoff, in press). A thermal event at 1044 Ma coincided with the crystallization of granodiorite exposed on Cove Mountain. The biotite augen gneiss (1029+-6 Ma) crystallized during the peak Grenvillian metamorphism (Ottawan phase), at about 1028+-9 Ma (Ownby and others, 2004). A younger thermal event is indicated by unfoliated leucocratic granitic dikes that have intruded the 1028 Ma granitoid.

Paleozoic

The Neoproterozoic rocks in the western Blue Ridge record a Barrovian-type metamorphism that increased to the southeast, from sub-chlorite grade near Gatlinburg, to kyanite-grade near the southern end of the Blue Ridge Parkway (Hadley and Goldsmith, 1963; Connolly and Dallmeyer, 1993). The location of the zone markers was generalized based on exposure, bulk rock composition, samples collected, and thin sections studied (Hamilton, 1961; Hadley and Goldsmith, 1963; King, 1964; Neuman and Nelson, 1965; Lesure and others, 1977; Mohr and Newton, 1983). The eastward deviation of the northeast-trending zone markers near the north end of the Murphy synclinorium eastward, suggests post-metamorphic folding (Hadley and Goldsmith, 1963), but faults have not offset the zone markers. Metamorphic P-T conditions were calculated at ~580 C and 6.6 kbar (Mohr and Newton, 1983) for one kyanite-grade rock in the Noland Creek 7.5-minute quadrangle. Kohn and Malloy (2004) analyzed monazite in order to calculate the staurolite-in reaction at ~600 C and 7 kbar occurring roughly at ~400 Ma. Using a modern geothermal gradient of 25 C/km, these rocks were buried at depths of 23 to 24 km at that time.

Metamorphic Grades and Minerals

Some of the rocks in the sub-chlorite zone have slaty cleavage but there is no evidence of new mineral growth. The rocks in the chlorite zone are green because of chlorite porphyroblasts, and some rocks have a lustrous sheen because of the presence of minute crystals of chlorite and sericite. The rocks in the biotite zone seldom contain visible porphyroblasts of biotite. Rocks in the garnet zone contain visible porphyroblasts of garnet, biotite, muscovite, and chloritoid. Most of the garnets are 1 mm to 2 cm diameter. Small dark green laths and discs of chloritoid are readily seen in the light gray quartzose slate of the Anakeesta Formation. Rocks in the staurolite and kyanite zones contain visible albite, garnet, bioitite, muscovite, staurolite, and kyanite. Kyanite crystals as long as 5 cm occur locally. Porphyroblasts are confined to the bedding. Peak thermal metamorphism produced randomly oriented porphyroblasts that grew statically. The metasedimentary rocks within the kyanite zone were locally partially melted during peak metamorphic conditions. Along the southeast margin of the Ela dome beneath the spillway dam on the Tuckasegee River, the metagraywacke and schist of the Copperhill Formation is migmatitic containing lenticular and podiform granitic leucosomes.

In rocks higher in grade than those in the garnet zone, metasandstone and metagraywacke typically have concretions called calc-silicate granofels by Goldsmith (1959). These so-called "pseudo-diorite" of Keith (1913), have preserved clusters of visible garnet, hornblende, and biotite crystals. These crystals are commonly randomly oriented, but locally have been aligned in syn- to post-metamorphic shear zones.

Abundant evidence of retrograde metamorphism is in the southern and southeastern part of the highlands (Mohr, 1973; Southworth, 1995; Montes and Hatcher, 1999). Pressure solution cleavage which is axial planar to folds has cut across and the garnet and biotite porphyroblasts have been retrograded to chlorite and sericite. Carbonate-chlorite schist and greenstone resulted from retrograde metamorphism.

Isotopic Data

K-Ar and $^{40}Ar/^{39}Ar$ radiometric dating of biotite, hornblende, muscovite, and whole rock specimens suggested a complex metamorphic history (Kish, 1991). The polymetamorphic history consisted of a 460 to 440 Ma (Middle to Late Ordovician) event (Taconian orogeny) and a 380 to 360 Ma (Middle to Late Devonian) event (Acadian orogeny)(Connolly and Dallmeyer, 1993). Garnet porphyroblasts overgrew an earlier foliation, and one or more retrograde events was evident (Mohr, 1973). Kohn and Malloy (2004) used U-Th-Pb electron microprobe analyses of monazite from three samples from the east limb of the Murphy synclinorium near Bryson City to determine an age that ranged from 437 Ma to 357 Ma. They concluded that metamorphism occurred at ~400 Ma. Moecher and others (2004) determined a 480 Ma age for zircon from schist of the Copperhill Formation (exposed along the Blue Ridge Parkway) using the thermal ionization microprobe spectrometer technique. Therefore, the results of Ordovician metamorphism (Taconian orogeny) were overprinted by those formed by Devonian metamorphism (Acadian orogeny). These were further overprinted by retrograde metamorphism and deformation during the Mississippian to Permian (Alleghanian orogeny). Good field relations locally support these three events; high-grade ductile deformation and partial melting in the southeast, later non-coaxial folding and faulting under retrograde conditions in the east-central region; and discrete brittle faults formed during the late transport of the metamorphic rocks. Similar complex Paleozoic histories are being unraveled in the Piedmont of northern Virginia and Maryland (Kunk and others, 2004), for example.

^{39}Argon/^{40}argon hornblende cooling ages of 430 Ma (M.J. Kunk, U.S. Geological Survey, 2004, oral comm.) and 425- 415 Ma (Dallmeyer, 1975), U-Pb age of sphene from basement gneiss dated at 440 Ma (J. A. Aleinikoff, U.S. Geological Survey, 2004, oral comm.), and U-Pb age of monazite from Neoproterozoic pelitic rocks of the Great Smoky Group dated at between ~480 Ma (Moecher and others, 2004) and ~400 Ma (Kohn and Malloy, 2004), supports the concept that the rocks above the garnet zone were deformed and metamorphosed over a period of about 80 My from the Ordovician to the Devonian. Between the Gatlinburg fault system (biotite zone) and the Hayesville fault, the mean zircon fission track age of 365 Ma (Naeser and others, 2004), and argon-argon white mica ages of 377 to 354 Ma (Connolly and Dallmeyer, 1993) and 350 Ma (M.J. Kunk, U.S. Geological Survey, 2004, oral comm.), suggests a deformation and metamorphic event occurred between 377 and 350 Ma. Most of the rocks in the foothills were not metamorphosed, and contain detrital zircons that were not reset. In this, they are similar to rocks of the Chilhowee Group that contain the oldest detrital zircons (900 Ma) (Naeser and others, 2004).

Tectonic Summary

Faults, foliations, fission track analysis of zircon, and U-Pb and $^{39}Ar/^{40}Ar$ geochronology were used to refine our understanding of the tectonic history. Overall, these data suggest that the highlands and foothills are distinct fault blocks that had different metamorphic histories. They were tectonically transported throughout the Paleozoic, and finally assembled by the Great Smoky fault subsequent to the Mississippian. The similar cooling history of the rocks of the foothills and highlands from at least the Late Triassic or Early Jurassic (~200 Ma) to the present, supports the concept of a very slow cooling rate (~17.7 m/My)(Naeser and others, 2004). Between the Gatlinburg fault system and the Cold Springs fault, the mean fission track zircon age of ~280 Ma suggests that the highlands constitute a horst block formed during the latest Alleghanian or younger time. The present-day altitude difference between the highlands and foothills coincides with the apatite fission track ages. This suggests that the post Mesozoic-Cenozoic uplift has continued, with possibly Late Cretaceous or younger uplift along the Gatlinburg fault system and the Cold Spring fault.

Surficial deposits and Landforms

The surficial deposits and landforms in the GSMNP result from three main agents and processes: running water (alluvium and terrace deposits), chemical and physical weathering (sinkholes and residuum), and gravity movement on slopes (colluvium, debris flows, and prehistoric debris fans) (Hadley and Goldsmith, 1963; King, 1964; Neuman and Nelson, 1965). The map portrays only the major surficial deposits of alluvium, terrace deposits, sinkholes, residuum, colluvium, debris flows, and debris fans. Most surficial deposits in the map area are saprolite and residual soils derived locally from the weathering of the underlying bedrock. Exceptions are floodplain and terrace deposits along the rivers, and debris fans. This transported material is different from the underlying bedrock units.

Colluvium, debris fans, coarse alluvium, and terrace deposits, constitute a continuum and are differentiated from one another by material, slope, and distance of transport by gravity and water. Abundant examples of such deposits can be seen in the valleys of (from east to west), Cosby Creek, Rocky Grove, Little Pigeon River at Greenbriar Cove, Le Conte Creek, and the Little River near Elkmont. Along the valley of Le Conte Creek from Mount Le Conte north to Gatlinburg, colluvium (Qc) occurs high on the slopes associated with modern talus and older talus, boulder fields, and boulder streams derived from weathering of the Thunderhead Sandstone. On lower slopes the material was transported by floods, creep, and solifluction, into debris fans (Qd), such as seen at Cherokee Orchard. The Cherokee Orchard debris fan has been incised by Twin Creeks. The fine material has been eroded leaving only coarse alluvium that has been terraced. These fluvial terraces have been incised by modern streams that flow through coarse alluvium and, locally, bedrock.

Fluvial Deposits and Landforms

Fluvial deposits and landforms include alluvium (Qa) that underlies valleys and modern flood plains, bedrock terraces, strath terraces, and upper level strath terraces with or without associated alluvial deposits (QTt). Alluvium (Qa) consists of unconsolidated deposits of stratified silt, sand, gravel, and cobbles as much as 20 ft (6 m) thick that was transported and deposited by running water. Alluvium is typically exposed along the banks of creeks and rivers and in road cuts. The large valleys in areas like Cades Cove, Tuckaleechee Cove, and Wear Cove, are broad alluvial plains that have not been inundated by modern floods. Larger drainages such as the Little River, West Prong of the Little Pigeon River, Oconaluftee River, Tuckasegee River, and Jonathan Creek have narrow flood plains that are subject to high-water floods and the deposition of alluvium. Most rivers at lower altitudes occupy broad valleys. Extensive alluvium was deposited here in the past, but modern streams have incised the alluvium and exposed bedrock. The particle size of alluvium is a function of the parent bedrock and transport distance. Erosion-resistant, silica-rich bedrock produces abundant boulders and cobbles, whereas finer-grained bedrock produces mostly silt and clay. Although alluvium (Qa) is ubiquitous on the map, its volume is not great.

A characteristic feature of the GSMNP region are creeks, streams, and rivers that contain very coarse alluvium (King, 1964; Southworth and others, 2003). Coarse alluvium occurs along drainages within the Blue Ridge highlands and consists of boulders of metasandstone and metaconglomerate as much as 33 ft (10 m) across. Boulders as much as 5 ft (16 m) across are actively transported during modern storms (Moneymaker, 1939; King, 1964). The size and abundance of boulders are a function of the local bedrock source. The predominant boulders were derived from the massive, thick bedded, coarse metasandstone of the Thunderhead Sandstone. Well-foliated biotite gneiss constitutes rectangular slabs in coarse alluvium in "The Gorge" of Raven Fork, upstream of Big Cove, NC. Much of the coarse alluvium is a lag deposit of pre-existing material in a debris fan from which the fines were removed as modern drainages incised into it. Thus, they are the product of the modern erosion of a relict deposit. Coarse alluvium occurs in the upper reaches of creeks and rivers such as Hazel Creek, Little River, Little Pigeon River, Big Creek, Oconaluftee River, Cataloochee Creek, and Straight Fork. Some areas of coarse alluvium have broad flood plains containing fine alluvium that was preserved on top, as is the case at Rough Fork of Cataloochee Creek and at the Pioneer Village at the Oconaluftee Visitor Center.

Terraces

Terraces (QTt) are benches that have been cut by rivers into bedrock that contain remnant deposits of unconsolidated sand, gravel, and cobbles. Terraces, called straths, range from 10 to 120 ft (3 to 36 m) above modern water levels. The majority of clasts consist of resistant quartz-rich rocks that are in a fine-grained matrix of silt and clay. Terrace deposits are well developed along the Tuckasegee River in North Carolina, and in Tennessee along the Little River, West Prong Little Pigeon River, Little Pigeon River, Dunn Creek, and Cosby Creek.

Terraces mark abandoned fluvial meanders at "The Sinks" along Little River (between Metacalf Bottoms and the Townsend "Y") and along Camp Prong, about 2.5 mi (4 km) north-northwest of Clingmans Dome. The meander at "The Sinks" is about 80 ft

(24 m) above present water level whereas the Camp Prong meander is more than 200 ft (61 m) above present water level. Airport Road in Gatlinburg is on a terrace developed on the lower part of a debris fan about 80 ft (24 m) above modern drainage.

Terraces near Glade and Emerts Cove near Pittman Center are related to hardness of bedrock and subsequent stream capture. Terraces at the lower reach of Greenbrier Cove are well developed on Pigeon Siltstone but not on the Roaring Fork Sandstone. To the west, remnants of terrace deposits have been preserved in “The Glades”, where there is no modern drainage. A north-flowing stream was captured by Dunn Creek, which flows west to Gatlinburg, leaving “the Glades” as an abandoned “dry” valley. On terraces near Pigeon Forge, cobbles and pebbles of coarse metasandstone derived from the highlands were transported a distance of more than 7 miles (12 km).

Upper level terraces have been cut into bedrock as much as 200 ft (61 m) above present stream levels (Southworth and others, 2003), they, however, were not differentiated on this map. Much of the unconsolidated material once lying on the bedrock has been completely eroded, leaving a few remnant boulders and cobbles. Upper level terraces occur along the Tuckasegee River between Ela and Bryson City, along Little River in Tuckaleechee Cove, and along West Prong of the Little Pigeon River and Little Pigeon River, between Pigeon Forge and Sevierville.

Landforms and Deposits formed by Physiochemical Weathering

The physical and chemical weathering of rocks and minerals under different climatic regimes contributed greatly to the landscape evolution of the GSMNP region over several tens of millions of years. All rocks of the region have been chemically weathered and saprolite and (or) residuum that have been highly oxidized (red) and leached (yellow) is common. With the exception of isolated outcrops of bedrock, the majority of the landscape is underlain by saprolite, residuum, and (or) surficial deposits that commonly contain transported residuum. Residuum mantles much of the area, but only a few deposits are shown on the map because they are difficult to define and portray at map scale. Hamilton (1961) and King (1964) noted extensive residuum and saprolite as much as 100 ft (30 m) thick in the foothills, and Hadley and Goldsmith (1963) described residuum as much as 100 ft (30 m) thick on ridges transected by the Blue Ridge Parkway.

Where contacts are exposed, residuum is unconformably overlain by colluvium, debris, and alluvium. This residuum may form in situ beneath the overlying deposit or it may have developed during a climate different from today. Residuum on upper slopes is unconformably overlain by debris, whereas debris is unconformably above fresh bedrock in the valleys (Hadley and Goldsmith, 1963). They suggested that the matrix of the debris was recycled residuum. This suggests that the earlier landscape of the region was dominated by the production of residuum and saprolite during a warm and humid climate. The saprolite and residuum were eroded and recycled into a variety of slope deposits perhaps during a colder and wetter climate. The deposits are currently being eroded to form the present landscape.

Karst

Carbonate rocks dissolve to form caverns and sinkholes. Insoluble minerals in the limestone and dolomite, such as quartz and chert, remain as lag deposits in the clay-rich residual matrix. The carbonate rocks and associated residuum are restricted to the tectonic windows (from west to east in the west central part of the map area) at Calderwood, Cades Cove, Big Spring Cove, Tuckaleechee Cove, and Wear Cove, as well as near Cosby, Walland, and northwest of Chilhowee Mountain in the Valley and Ridge province.

Tuckaleechee Caverns, south of Townsend, TN, is the only commercial cave in the study area. Within the park, but closed to the public, are Bull Cave and unnamed caves in White Oak Sink (both in the southern part of Tuckaleechee Cove), and Gregory Cave on the north side of Cades Cove. Bull Cave is near the crest of Rich Mountain at an altitude of about 1900 ft (579 m). The depression that led to the opening is about 140 ft (43 m) deep and the cave shaft is more than 500 ft (152 m) deep. The stream that flows into the cave at White Oak Sink at an altitude of about 1700 ft (518m), emerges about 3 mi (5 km) to the northwest at Dunn Spring, at an altitude about 640 ft (195 m) lower. Numerous smaller caverns and cavities in limestone demonstrate that the process is still active at all scales.

In the western Blue Ridge province, sinkholes (QTs) are locally common within parts of Cades Cove, Tuckaleechee Cove and White Oak Sink, and Big Spring Cove. They range in diameter from about 6 ft (several meters) to as much as 1000 ft (305 m), south of Townsend, TN, near Red Bank. Active sinkholes can be seen in the northern part of the White Oak Sink near the cliffs of limestone. In Big Spring Cove, mounds of boulders and intervening pits and depressions in a debris fan of metasandstone often contain water that are interpreted to be collapse sinkholes. In Cades Cove, “Lake in the Woods” and two small ponds northeast of Carter Shields Place are also considered to be water-filled sinkholes. Sinkholes are very abundant in Tuckaleechee Cove but few were recognized in Wear Cove. Sinkholes are also found within areas of the Cambrian Shady Dolomite in Miller Cove near Walland, and in the valley between Cosby and Wilton Springs. Numerous valleys in the western foothills are underlain by limestone of the Walden Creek Group and contain sinkholes; examples are (west to east) Happy Valley, Walden Creek valley north of “Sinkhole Mountain”, and near Jones Cove,

Sinkholes are most abundant in the Tennessee Valley. Although sinkholes are presently subsiding, material deposited within them suggests that they have formed over an extended period of time. About 75 mi (40 km) northeast of the study area near Gray, TN, Miocene vertebrate fossils that could be 5 to 23 My were deposited in a probable ancient sink (Kohl, 2000; Wallace and others, 2002).

Residuum

Accumulations of subrounded to angular cobbles and boulders of quartz and chert that originated in veins or beds in limestone or dolomite were mapped as residuum (QTr). These silica minerals are very resistant to erosion and remain as lag deposits in clay-rich terra rosa formed by the chemical disintegration of the carbonate rock. Small deposits containing small clasts of quartz and chert were locally preserved in road cuts in Tuckaleechee Cove, and on the tops of small hills in fields near Cosby. In Tuckaleechee Cove north of Townsend at an altitude of 1200 ft (366 m), residuum contains limonite concretions. Similar residuum containing concretions of iron and manganese oxides were prospected in Cades Cove (Southworth and others, 1999), Tuckaleechee and Wear Coves

(King, 1964), and in the foothills underlain by limestone of the Walden Creek Group (Hamilton, 1961).

Accumulations of cobbles and boulders of jasper were mapped locally as residuum in the Cambrian Shady Dolomite in the valleys near Walland. Neuman and Nelson (1965) suggested that the jasper was a residual precipitate from the percolation of water derived from the overlying silica-rich slope material. If this interpretation is correct, the jasper residuum is a very old deposit formed during a climate that favored the development of laterites. Similar laterite deposits of bauxite, kaolinite, and lignite in the Appalachians may be Late Cretaceous to Eocene in age (Overstreet, E. F., 1964; Pierce, 1965; Tschudy, 1965; Hack, 1979; McLaughlin and Darrell, 1972; McLaughlin and McBroom, 1980; Wallace and others, 2002; and Bearce and Carroll, 2003).

Areally extensive gravel deposits as much as 40 ft (12 m) thick at an altitude of 1420 ft (433 m) near Sixmile in the Tennessee Valley were mapped as residuum. Neuman and Wilson (1960) suggested that this gravel was Tertiary alluvium that was deposited as a sheet, that was later warped, modified by dissolution of the underlying carbonate bedrock, and locally mixed with residual gravel.

Residuum of rocks of the Great Smoky Group must have been extensive, because the abundant debris deposits have a clay-rich matrix that may have been recycled and derived from these rocks. Residuum of Pigeon Siltstone can be seen along Rt. 73, and broad areas of residuum of biotite granite gneiss are found south of Cherokee along Rt. 441.

Slope Deposits

Three general types of slope deposits (listed in increasing abundance) are 1) historical and recent debris flows in the upper highlands resulting from intense rain storms, 2) colluvium in hollows on slopes below bedrock escarpments, on side slopes above valley bottoms, and depressions on the upper slopes of the highlands, and 3) coarse boulder debris that forms broad, gently-sloping fans on the lower slopes that formed in the past in a different climate.

Debris Flows

Debris flows (Qdf) are landslides that are common in the highlands north of Newfound Gap, along the Boulevard Trail, and along the Appalachian Trail from Charlies Bunion to Laurel Top. They consist of masses of bedrock, soil, and vegetation that catastrophically moved rapidly down slope, usually incorporating more material as they moved. The debris flows occur on steep mountain slopes (35 to 44 degrees; Bogucki, 1970; 1972) underlain by soil and regolith that generally is less than a meter-thick and developed on slate of the Anakeesta Formation. The head of some debris flows have very narrow "V"-shaped ravines that were called wedge-failures (Moore, 1986), that formed at the intersection of cleavage and joints or cleavage and bedding. Elsewhere, the flows originated on very steep slopes as small slips on the cleavage or bedding. Debris flows in the study area resulted from prolonged rainfall events in 1938 (Moneymaker, 1939; Koch, 1974), 1940 (USGS, 1949), and 1942, 1943, 1951, 1956, 1967, 1971, 1975, 1984, and 1993 (Bogucki, 1970, 1972, and 1976; Clark, 1987; Schultz and others, 2000). Water-logged soil and regolith usually fail when large trees topple and begin to move down very steep slopes. These slope failures are major geomorphic agents that have formed the steep craggy summits in the central part of the park. The map mostly shows scars of historical debris flows that have been slow to revegetate because of the lack of soil.

Local concentrations of debris flows formed during isolated cloudbursts. Charlies Bunion, for example, located west of Newfound Gap along the Appalachian Trail, is the site of an isolated debris flow event. A forest fire burned the vegetation cover and later, a rainstorm resulted in debris flows that stripped the burned trees to expose bedrock. Debris flow deposits of trees and rock can be seen north of Arch Rock where the Alum Cave Trail crosses Styx Branch, and along the tributaries along the upper part of Newfound Gap Road in Tennessee.

Small rock slides, slumps, and landslides are common along road embankments that have been cut into inclined bedrock and(or) unconsolidated soil and debris. They are very common along the Foothills Parkway where sandstone and quartzite beds dip into the road, and along steep roadcuts around Gatlinburg, but they were not depicted on the map.

Colluvium

Colluvium (Qc) includes talus, boulder streams, and boulder fields of cobbles and boulders derived from the weathering of coarse metasandstone on the upper slopes and hollows in the highlands. Boulders are typically 3 to 10 ft (1 to 3 m) long, but can be as much as 50 ft (15 m) long (King, 1964). Talus is forming today below steep bedrock escarpments. Recent talus merges down slope into older deposits. Large boulders abutt one another on the surface and have no preferred orientation. The deposits contain little or no soil; thus, they support little to no vegetation. Subsurface drainage and intermittent streams have modified the boulder deposits. Individual boulders may be covered with moss and (or) lichen, suggesting little transport in the last several decades. Bent and leaning trees suggest minor recent movement of the deposits. Areas of bedrock cliffs near waterfalls, like Rainbow Falls, Grottoe Falls, and Buckeye Cove, have blocks that were detached from the cliff along bedding planes and joints. Fresh piles of jumbled blocks are below and downslope of the escarpments. Colluvium occurs on all hard rock units including the clastic rocks of the foothills, especially on Chilhowee Mountain and Green Mountain, as well as on quartzose gneiss in the eastern highlands. Colluvium is best developed on the rocks of the Great Smoky Group in the highlands. Excavations of some colluvium suggest that it is unrelated to modern topography. It fills swales on ridges (King, 1964) and noses of slopes, and roadcuts reveal steep angular unconformities that are perpendicular to modern slopes.

Although colluvium and talus is being produced today, the majority of the mapped deposits are relicts of an earlier (Pleistocene) cold, periglacial climate. Many boulder streams and boulder fields probably formed in periglacial environments during the Pleistocene (Delcourt and Delcourt, 1985). During cold phases of the Pleistocene, ridge crests in the GSMNP may have been above the forest limit in an active periglacial frost-climate environment (King and Stupka, 1950). King (1964) described 10 to 15 ft (3 to 5 m) of "mantle" on bedrock exposed during construction on the Clingmans Dome Road. Rozanski (1943), Clark (1968), Michalek (1968), Reheis (1972), Richter (1973), and Torbett and Clark (1985), Clark and Ciolkosz (1988), and Clark and others (1989) have interpreted polygonal ground, block streams, block fields, fan deposits, and other features in the park to have had a periglacial origin. Block fields and block streams are colluvial deposits that were long considered periglacial deposits. Block fields are large sheet-like accumulations of blocks that commonly mantle upland surfaces. Block streams typically extend farther downslope than along a contour, which suggests transport by gravity, solifluction, freeze-thaw, ice-wedging, and ice rafting (Clark and Ciolkosz, 1988). Differences between valley forms developed on north- and south-facing slopes in the Great Smoky Mountains, probably had a periglacial origin as well (Richter, 1973). For example, coarse boulder deposits in the north-facing drainage basins, such as Le Conte Creek, probably formed in Pleistocene periglacial environments. Erosion over-steepened the bedrock highwalls of a south-dipping homocline of thick, massive Thunderhead Sandstone and solifluction moved the rock debris down the valley.

Debris Fans

Fan-shaped and irregular sheet-like accumulations of debris constitute the dominant and most prominent Cenozoic deposit on the middle and lower altitudes in the unglaciated highlands of the Appalachians (Mills, 2000a; 2000b). These deposits characterize a significant part of the landscape of the Great Smoky Mountains. The debris fans (Qd) are very poorly sorted deposits consisting mostly of matrix-supported diamicton, containing boulders and cobbles in a fine-grained matrix of sand, silt, and clay. These debris fan deposits were classified according to the dominant rock type, size of the clasts, and the matrix material (Southworth and others, 2003), but here were combined into one unit. The size, shape, and topographic setting of the debris fans are distinctive. For example, debris fans of gneiss near Dellwood and Hazelwood (Mills and Allison, 1994), differ from fans of metasandstone boulder debris near Gatlinburg (Schultz, 1998), fans of fine-grained metasandstone above carbonate rocks (Southworth and others, 1999), and fans of cobbly sandstone debris on Chilhowee Mountain. The abundance of large and durable boulders contributed to the size of the fans by armoring their surface from weathering and erosion. Many of the fans occur in coves and hollows away from erosive rivers and are well preserved. The debris fans are complex assemblages reflecting a long history of deposition and modification. They are relict deposits that are now undergoing chemical weathering and stream incision. The oldest deposits are residuum containing few clasts in a clay rich matrix. Fan material was reworked into coarse alluvium by the removal of fine-grained matrix leaving a bouldery surface as a lag concentrate. Early settlers modified these deposits as land was cleared for pasture, as evidenced by piles, terraces, and fence lines of boulders. Lateral migration of streams and stream capture, were the dominant secondary processes that contributed to the present patchwork assemblage of deposits on any given fan.

Debris fans are thought to have formed by several different processes. Earlier workers called the deposits colluvium (Hamilton, 1961; Hadley and Goldsmith, 1963; Neuman and Nelson, 1965), bouldery alluvial deposits (Hadley and Goldsmith, 1963), and coarse bouldery alluvium in Piedmont coves (King, 1964). Deposits formed dominately by debris-flow processes are now called debris fans, and those deposits formed dominately by fluvial processes are called alluvial fans (Mills, 2000b).

A combination of processes probably created the fans in the GSMNP region. Debris flows were common during Pleistocene glacial/interglacial transitions as warm and cold cycles fluctuated and storms were common. A build-up of debris and colluvium on the upper slopes during cold periods was transported down slope as debris flows during warm periods of increased precipitation (Mills, 2000b). Many fans are in hollows and valleys, suggesting a pre-existing depression. Fluvial erosion must have formed a basin, cove, hollow, or valley by incision, which filled with debris, and were subsequently modified by running water.

Boulder debris fans are the dominant slope deposit in the Blue Ridge Highlands. Thunderhead Sandstone was the dominant source, but these fans also contain metasandstone of the Elkmont Sandstone, Anakeesta Formation, and Copperhill Formation as well as locally blocks of Longarm Quartzite and rocks of the Snowbird Group. These deposits make extensive, broad, convex-upward fans and aerially extensive sheets that have been modified by erosion. Boulder debris fans are especially abundant where the Thunderhead Sandstone is massive, thick bedded, and coarse-grained, and forms large cliffs that face north-northwest. Individual cliffs are as much as 250 ft (76 m) high, and range over 4000 ft (1220 m) in total relief. These massive escarpments extend approximately from Blanket Mountain near Elkmont, TN, eastward to Cosby, TN. The north-facing cliffs, the abundant source material, and the orographic setting for storms has provided a favorable setting for the formation of these deposits. The amount and size of boulder debris deposits was directly related to the eastward thickening wedge of Thunderhead Sandstone (King and others, 1958).

Boulder debris deposits consists mostly of matrix-supported diamicton containing locally stratified silt and clay that supporting sub-rounded boulders of metasandstone. Blocks of metasandstone and metaconglomerate 40 ft (12 m) long in Cherokee Orchard, and blocks 20 ft (6 m) high, 25 ft (7.6 m) wide, and 45 ft (14 m) long are 1.3 mi (2 km) from the source of Greenbriar Pinnacle (Hadley and Goldsmith, 1963). Some have been incised more than 70 ft (21 m) with no bedrock exposure. Thickness of material is highly variable, as some fans appear to have filled concave-up depressions having 20 ft (6 m) of debris at the head and toe and more than 70 ft (21 m) in between.

Noteworthy areas of boulder debris fans are in Tennessee (from west to east), at the campground in Cades Cove, Big Spring Cove, Sugarland Valley, Cherokee Orchard, Roaring Fork Valley, Greenbriar Cove, Albright Cove, and the campground in the park at Cosby. Good vertical exposures of the diamicton can be seen in road cuts along the trail south of Elkmont along Little River, in small landslides on Albright Grove and the head of Cosby Creek, along Route 19 north of Soco Gap at the head of Maggie Valley, and along Big Creek. Deposits of boulder debris were also found outside of the region of massive Thunderhead Sandstone, such as near the Oconoluftee Valley in North Carolina, and along the tributaries that drain Fontana Lake.

Some of the largest ancient debris fans in the Appalachian highlands occur near Cosby, TN. Small hills underlain by bedrock are surrounded by boulder debris are as much as 200 ft (61 m) above the valley, suggesting a long and complex history of erosion and deposition. Exposures of the lower parts of these fans are deeply weathered and have oxidized matrixes. Near Cosby, boulders more than 5 ft (1.5 m) across have been transported more than 6 mi (9.7 km) from their bedrock source (Hamilton, 1961). The fan to the west at Albright Grove supports an old growth forest more than 500 years old. This fan forms the drainage divide between two tributaries over a distance of 2 mi (3.2 km), that diverge at the base of the fan and enter the French Broad River 25 mi (40 km) apart. This boulder debris was deposited in a valley that today was topographically inverted, by gully gravure, to form a convex upland (Mills, 1981. There are abundant smaller examples of this type of topographic inversion in other debris deposits throughout the region, that have as much as 60 ft (18 m) of relief. Debris from Greenbriar Pinnacle was transported northward to the valley of Webb Creek, where creeks modified the fans to form terraces.

Boulder debris fans above carbonate rock occur at the southeast end of Cades Cove and to the east at Big Spring Cove. These types of deposits may include residuum of carbonate rock in the matrix at depth, as the landforms have been modified by sinkholes. Debris in Big Spring Cove is 45 ft (14 m) thick (King, 1964).

Fans of metasandstone debris are above carbonate rock in Cades Cove, Tuckaleechee Cove, and Wear Cove. Fine-grained, thin-bedded metasandstone of the Cades Sandstone was the source of material in the diamicton. Excavated pits in Cades Cove expose stratified to non-stratified, rounded to subrounded fine-grained metasandstone in a fine-grained matrix of silt, suggesting significant alluvial transport. Sinkholes were developed on the fans. Bedrock outcrops along the margins of the fans and sinkholes suggest that the deposits are locally as little as 3 ft (1 m) thick (Southworth and others, 1999). Fans of cobbles of sandstone and quartzite derived from the Chilhowee Group on Green Mountain were deposited on carbonate rocks northeast of Cosby. The cobbles of sandstone here were mixed with a residuum of Shady Dolomite. In these settings, the underlying karst in the center of the deposits may have allowed a significant amount of material to accumulate, as in similar settings in central Virginia near Elkton (King, 1950; Hack, 1979; Whittecar and Duffy, 2000).

In Tuckaleechee Cove, metasandstone boulders as much as 8 ft (2.4 m) long are concentrated in incised drainages more than a mi (1.1 km) from their source on Rich Mountain. Exposures in Wear Cove have several feet of mature soil overlying as much as 10 ft (3 m) of metasandstone debris on limestone residuum (Neuman and Nelson, 1965). The lower altitude distal parts of the debris fans in Wear Cove and Cades Cove were terraced and were modified by water. The same process (debris fans modified by alluvial processes) probably affected the fans along the Little River in Tuckaleechee Cove.

Sandstone debris of the Lower Cambrian Chilhowee Group constitutes an extensive series of fans on the northwest slope of Chilhowee Mountain. The angular to sub-rounded cobble-size clasts of friable fine-grained, sugary-textured sandstone of the Cochran

Sandstone, Nebo Sandstone, and some Hesse Quartzite that litter the surface and have weathering rinds. Excavations reveal a clay-rich, ruby-red matrix that has been highly oxidized. These fans were incised by modern streams to form terraces. The angular cobbles of sandstone grade down-slope into sub-angular to well-rounded cobbles that average about 6 in (10 cm) across, the largest boulder is about 3 ft (1 m) in diameter. These cobbles and boulders armour the underlying shale. The gentle slope and the rounding of the cobbles suggests a large alluvial component of either primary deposition or secondary modification. These landforms and deposits are best seen in road cuts between Camp Montvale and the Foothills Parkway, west of Look Rock. As noted by Mills and Whisner (2000), these old weathered deposits are unique to the Blue Ridge province.

Debris fans of gneiss form extensive deposits in North Carolina (from west to east), at the north end of the Ela dome, at the head of Big Cove, and in Maggie Valley, Dellwood, and Saunook. Elongated slabs of gneiss have broken along the foliation. Feldspar has decomposed to clay, so the clasts are friable, have thick weathering rinds, and the surrounding red, clay-rich matrix has been oxidized. Mills and Allison (1995a) determined the relative age by the amount of clay, their Munsell red hue, and percent of weathered clasts in the Dellwood and Saunook area south of Hazelwood, to differentiate old and intermediate fans amongst mostly young fans. Impressive fans of gneiss debris can be seen at the head of Big Cove where the incision of Raven Fork has formed "The Gorge".

In the foothills section of the western Blue Ridge, small debris fans of metasandstone were mostly derived from rocks of the Cades Sandstone, Walden Creek Group, and Chilhowee Group. They are in isolated hollows where there was a source of coarse-grained, quartz-rich bedrock.

Some of the oldest landforms and deposits in the region are the upper level debris fans near Cosby, Townsend, and along Chilhowee Mountain. At Cosby, the deposit occupies a transitional area between debris fans and alluvial terraces, and they are about 160 ft (49 m) above the modern drainage of Cosby Creek. Upper level fans of metasandstone debris above carbonate rock in the Dry Valley and the White Oak Sink part of Tuckaleechee Cove are the remnants of the oldest preserved fill in the coves. The upper level fan in White Oak Sink is about 100 ft (30 m) above the present bottom of the sink. Red, residual soil underlies the upper level fan, and gray, cobbly soil constitutes the lower, younger fan deposit (Neuman and Nelson, 1965). The upper level fans are elongate because they were incised by streams to depths as much as 50 ft (15 m), and expose bedrock along their margins. The majority of fans of sandstone debris on the northwest slope of Chilhowee Mountain are upper level types that are as much as 80 ft (24 m) above the lower fans. They are about 100 ft (30 m) above the modern drainages along their incised margins.

Isotopic Ages of Debris Fans

A few samples of organic material have been derived from debris deposits and terraces in this area and "Lake of the Woods", a water-filled depression on a terraced debris fan in Cades Cove (Southworth and others, 1999). Davidson (1983) got a ^{14}C age of 6,600 years before present (ybp) at "Lake of the Woods". Near Dellwood, NC, Kochel (1990) obtained ten samples from 5 fans that had ^{14}C ages ranging from 1,000 to 25,000 ybp, with 16,000 to 18,000 ybp being the most consistent age. He proposed that the summer polar front retracted several thousand years earlier in North Carolina than New England, so that post-glacial debris flows and fan deposition began around 16,000 years ago. Around Shenandoah National Park, VA, 39 ^{14}C samples of material from fans, slope, and fluvial deposits had a range of age from more than 51,000 to 2000 ybp (Eaton and others, 2003a). Although some material may have been recycled, the age range suggests debris flows formed over at least 25,000 years, recurring, on the average, at least every 2500 years since the onset of the Wisconsinan glacial maximum. Whittecar and Duffy (2000) and Eaton and others (2003b) contend that debris fans around Shenandoah National Park, VA, formed during the late Pleistocene and have since been eroded by Holocene incision.

Some fans in the western Blue Ridge highlands may be hundreds of thousands of years old, and paleomagnetic reversal of iron oxides suggests that a minimum age of 1,000,000 years is likely (Mills and Allison, 1995a, 1995b). Mills and Granger (2002) also analyzed cosmogenic ^{10}Be and ^{26}Al in quartz, to derive a 1.45 +-0.17 Ma age for a debris fan on Rich Mountain, in Watauga County, NC, approximately 78 mi (125 km) northeast of this area. This cosmogenic age corresponds to a 1.5 +-0.3 Ma cosmogenic age obtained from the southern advance of the ice sheet to the Ohio River (Mills and Granger, 2002). Therefore, debris fans are composite landforms that have formed throughout the Pleistocene.

Isotopic Ages of Terraces

The study of the Little Tennessee River by Delcourt (1980) identified nine discontinuous terraces as much as 100 feet (30 m) above river level. Consistent ^{14}C ages from organic remains from the lowest terrace were 15,000 to 7,000 ybp and an average of about 31,000 ybp was obtained from the next highest terrace. He suggested that these terraces have recorded deposition of material destabilized from upland surfaces during the transition from cold-phase maxima to interglacial or interstadial conditions. Frost-bounded debris provided large sediment loads, the remnants of which were preserved in the terraces.

Relative Ages of Debris Fans and Terraces

Landforms incised by modern streams and rivers can be used to calculate a minimum relative age of such features by using modern erosion rates of 28m/My, that was determined from cosmogenic exposure ages (Matmon and others, 2003a and b). Possibly the oldest features are the abandoned meander at Camp Prong, a few upper level terraces, and the bedrock knoll within the boulder debris fan near Cosby campground. These elevated landforms suggest incision over a period of about 2.18 Ma, perhaps since the late Pliocene. Boulder debris fan deposits were incised over a period of 750,000 years. The debris fans on Chilhowee Mountain were incised about 214,000 years ago, whereas the upper level fans were incised over a period of about 857,000 years. Terraces along the major rivers have been incised over a period of about 1.29 Ma, whereas the terraces along modern floodplains were incised over a period of about 429,000 years.

References Cited

Aleinikoff, J.N., Southworth, C.S., Yacob, E.Y., and Fanning, C. M., 2004, Isotopic constraints on the time of deposition of the Thunderhead Sandstone, Ocoee Supergroup: U-Pb geochronology of monazite and xenotime overgrowths. [abst]: Geological Society of America Abstracts with Programs, v. 37 no. 6, p. A312.

Bearce, D. N. and Carroll, R. E., 2003, Geology of the Heeble Mine, St. Clair County, Alabama, in Holcomb, F. W. and others, Geology and operations at National Cement Company's Ragland Quarry: Alabama Geological Society Workshop and Field Trip, Feb. 22, 2003, 15 p.

Bogucki, D.J., 1970, Debris slides and related flood damage associated with the September 1, 1951, cloudburst in the Mt. Le Conte-Sugarland Mountain area, Great Smoky Mountains National Park: Knoxville, University of Tennessee, Unpublished Ph.D dissertation.

Bogucki, D.J., 1972, Intense rainfall in the Great Smoky Mountains National Park: Journal of Tennessee Academy of Science, v. 47, p. 93-97.

Bogucki, D.J., 1976, Debris slides in the Mt. Le Conte area, Great Smoky Mountains National Park, U.S.A.; Geografiska Annaler, v. 58A, p. 179-191.

Bridge, Josiah, 1955, Disconformity between the Lower and Middle Ordovician series at Douglas Lake, Tennessee. Geological Society of America Bulletin, v. 66, p. 725-730.

Cameron, E.U., 1951, Feldspar deposits of the Bryson City District, North Carolina. Division of Mineral Resources, Bulletin no. 62, 100 p., and 1:24,000-scale map.

Carrigan, C.W., Miller, C.F., Fullagar, P.D., Bream, B.R., Hatcher, R.D., Jr., and Coath, C.D., 2003, Ion microprobe age and geochemistry of southern Appalachian basement, with implications for Proterozoic and Paleozoic reconstructions. Precambrian Research, v. 120, p.1-36.

Carter, M.W., and Wiener, L.S., 1999, Bedrock geologic map of the Fines Creek 7.5-minute quadrangle, North Carolina. North Carolina Geological Survey, Asheville, scale 1:24,000.

Carter, M.W., Mershat, C.E., and Wilson, W.F., 1999, A Geologic Adventure along the Blue Ridge Parkway in North Carolina. North Carolina Geological Survey Bulletin 98, 60p.

Cattermole, J. M., 1955, Geology of the Shooks gap quadrangle, tenneessee: U.S. Geological Survey Geologic Quadrangle Map GQ-76, scale 1:24,000.

_________, 1962, Geology of the Maryville quadrangle, Tennessee: _GQ-163, scale 1:24,000.

Clark, G.M., 1968, Sorted patterned ground: New Appalachian localities south of the glacial border: Science, v. 161, p. 355-356.

Clark, G.M., 1987, Debris slide and debris flow historical events in the Appalachians south of the glacial border in: Geological Society of America Reviews in Engineering Geology, v.7, p. 125-138.

Clark, G.M., and Ciolkosz, E.J., 1988, Periglacial geomorphology of the Appalachian Highlands and Interior Highlands south of the glacial border—a review: Geomorphology, v.1, p.191-220.

Clark, G.M., Ciolkosz, E.J., Kite, J.S., and Lietzke, D.A., 1989, Central and southern Appalachian geomorphology-Tennessee, Virginia, and West Virginia: 28th International Geological Congress Field trip Guidebook T150. American Geophysical Union, Washington, D.C., 105p.

Connelly, J.B., and Dallmeyer, R.D., 1993, Polymetamorphic evolution of the western Blue Ridge: Evidence from $^{40}Ar/^{39}Ar$ whole-rock slate/phyllite and muscovite ages: American Journal of Science, v. 293, p. 323-359.

Connelly, J.B., and Woodward, N.B., 1992, Taconian foreland-style thrust system in the Great Smoky Mountains, Tennessee: Geology, v. 20, no. 2, p. 177-180.

Dallmeyer, R.D., 1975, Incremental Ar/Ar ages of biotite and honblende from retrograded basement gneisses of the southern Blue Ridge: Their bearing on the age of Paleozoic metamorphism: American Journal of Science, v. 275, p. 444-460.

Davidson, J.L., 1983, Paleoecological analysis of Holocene vegetation, Lake in the Woods, Cades Cove, Great Smoky Mountains National Park [M.S. thesis]: University of Tennessee, Knoxville, 100p.

Delcourt, P.A., 1980, Quaternary alluvial terraces of Little Tennessee River valley, east Tennessee: University of Tennessee Department of Anthropology Report of Investigations, v.29, p.110-121 and 175-212.

Delcourt, P.A., and Delcourt, H.R., 1985, Dynamic Quaternary landscapes of East Tennessee: an integration of paleoecology, geomorphology, and archaeology, Field trip 7, in Field trips of the Southern Appalachians, SEGSA 1985: University of Tennessee, Department of Geologic Sciences, Studies in Geology 9, p. 191-220.

DeWindt, J.T., 1975, Geology of the Great Smoky Mountians, Tennessee and North Carolina, with road log for field excursion, Knoxville-Clingmans Dome-Maryville: Compass, v. 52, p. 73-129.

Eaton, L.S., Morgan, B.A., Kochel, R.C., and Howard, A.D., 2003a, Role of debris flows in long-term landscape denudation in the central Appalachians of Virginia: Geology, v. 31, no. 4, p. 339-342.

Eaton, L.S, Morgan, B.A., Kochel, R.C. and Howard, A.D., 2003b, Quaternary deposits and landscape evolution of the central Blue Ridge of Virginia: Geomorphology, v. 1351, p. 1-16.

Espenshade, G.H., 1943, Geology of the Hazel Creek copper mine area, Swain County, North Carolina: U.S. Geological Survey Open File Report.

Espenshade, G. H., Staatz, M.H., and Brown, E.A., 1947, Preliminary report, Redmond lead-zinc mine, Haywood County, North Carolina: U.S. Geological Open File Report.

Espenshade, G.H., 1963, Geology of some copper deposits in North Carolina, Virginia, and Alabama: U.S. Geological Survey Bulletin 1142-i, p. i1-i50.

Fairly, W. M., 1965, The Murphy syncline in the Tate quadrangle, Georgia: Georgia Geological Survey Bullettin 75, 71p.

Ferguson, H.W., and Jewell, W.B., 1951, Geology and barite deposits of the Del Rio district, Cocke County, Tennessee: Tenn. Dept. Conserv., Div. Geol. Bull. 57, 235p.

Flohr, M.J.K., Dillenburg, R.G., and Plumlee, G.S., 1995, Characterization of secondary minerals formed as the result of weathering of the Anakeesta Formation, Alum Cave, Great Smoky Mountains National Park, Tennessee: U.S. Geological Survey Open-File Report 95-477, 22 p.

Fox, J.S., 1984, Besshi-type volcanogenic sulfide deposits--a review: Canadian Institute of Mining and Metallurgy Bulletin, v. 77, p. 57-68.

Goldsmith, Richard, 1959, Granofels: a new metamorphic rock name: Journal of Geology, v. 67, p. 109-110.

Gordon, C.H., 1920, Notes on the geology of the cove areas of east Tennessee [abst.]: Science, v.51, p.492-493.

Hack, J.T., 1979, Rock control and tectonism—Their importance in shaping the Appalachian highlands: U.S. Geological Survey Professional Paper 1126-B, 17p.

Hadley, J.B., 1970, The Ocoee Series and its possible correlatives, in Fisher, G.W., Pettijohn, F.J., Reed, J.C., Jr., and Weaver, K. N., Studies in Appalacjian Geology: Central and southern: New York, Intersci. Publ., p. 247-259.

Hadley, J.B., and Goldsmith, Richard, 1963, Geology of the eastern Great Smoky

Mountains, North Carolina and Tennessee: U.S. Geological Survey Professional Paper 349-B, 118 p.

Hadley, J.B., and Nelson, A.E., 1971, Geologic map of the Knoxville quadrangle, North Carolina, Tennessee, and South Carolina: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-654, scale 1:250,000.

Hamilton, Warren, 1961, Geology of the Richardson Cove and Jones Cove quadrangles, Tennesseee: U.S. Geological Survey Professional paper 349-A, p. A1-A55.

Hammarstrom, J.M., Seal, R.R., Meier, A.L., and Jackson, J.C., 2003, Weathering of sulfidic shale and copper mine waste: secondary minerals and metal cycling in Great Smoky Mountains National Park, Tennessee, and North Carolina, USA. Environmental Geology, v. 45, p. 35-57.

Hanselman, D.H., Conolly, J.R. , and Horne, J.C., 1974, Carbonate environments in the Wilhite Formation of Central Eastern Tennessee. Geological Society of America Bulletin vol. 85, p. 45-50.

Harris, L.D., and Milici, R.C., 1977, Characteristics of thin-skinned style of deformation in the southern Appalachians, and potential hydrocarbon traps: U.S. Geological Survey Professional Paper 1018, 40p.

Hatcher, R.D., Jr., 1978, Tectonics of the western Piedmont and Blue Ridge, southern Appalachians: Review and Speculation. American Journal of Science, Vol. 278, p. 276-304.

Hatcher, R.D., Jr., Thomas, W.A., Geiser, P.A., Snoke, A.W., Mosher, Sharon, and Wiltshko, D.V., 1989, Alleghanian orogen in The Geology of North America, v. F-2, The Appalachian-Ouachita Orogen in the United States: The Geological Society of America, Boulder, CO, p. 233-318.

Hatcher, R.D., Jr., Larson, K.W., and Neuman, R.B., 1989, Day 6, Western Great Smoky Mountians Windows: The Foothills Duplex *in* T167, p. 49-60.

Hatcher, R.D., Jr., and Goldberg, S.A., 1991, The Blue Ridge Province, in Horton, J.W., Jr., and Zullo, V.A., eds., The Geology of the Carolinas--Carolina Geological Society 50th Anniversary Volume: Knoxville, University of Tennessee Press, p. 11-35.

Hernon, R.M., 1969, The Wehutty Formation of North Carolina in Changes in Stratigraphic Nomenclature by the U.S. Geological Survey 1967: U.S. Geological Survey Bulletin 1274-A, p. A49-A52.

Higgins, M.W., Atkins, R.L., Crawford, T.J., Crawford, R.F., III, Brooks, R., and Cook, R.B., 1988, The Structure, Stratigraphy, Tectonostratigraphy, and Evolution of the

Southernmost Part of the Appalachian Orogen. U.S. Geological Survey Porfessional Paper 1475, 173 p.

Hurst, V.J., 1955, Stratigraphy, structure, and mineral resources of the Mineral Bluff quadrangle, Georgia: Georgia Geological Survey Bulletin 63, 137p.

Keith, Arthur, 1895, Description of the Knoxville quadrangle (Tennessee-North Carolina): U.S. Geological Survey Geologic Atlas Folio 16, 6 p.

_____1907, Description of the Nantahala quadrangle (North Carolina-Tennessee): U.S. Geological Survey Geologic Atlas Folio 143, 11 p.

_____1913, Production of apparent diorite by metamorphism: Geological Society of America Bulletin, v. 24, p. 684-685.

_____1927, Great Smoky Overthrust [abst.]: Geological Society of America Bulletin v. 38, p.154.

King, P.B., 1949, The base of the Cambrian in the southern Appalachians: American Journal of Science, v. 247, p. 513-530, 622-645.

King, P.B., 1950, Geology of the Elkton area, Virginia: U.S. Geological Survey Professional Paper 239, 82 p.

King, P.B., 1964, Geology of the central Great Smoky Mountains, Tennessee. U.S. Geological Survey Professional Paper 349-C, 148 p.

King, P.B., Hadley, J.B., Neuman, R.B., and Hamilton, W.B., 1958, Stratigraphy of the Ocoee Series, Great Smoky Mountains, Tennessee and North Carolina: Geological Society of America Bulletin, v. 69, no. 8, p. 947-966.

King, P.B., Neuman, R.B., and Hadley, J.B., 1968, Geology of the Great Smoky Mountains National Park, Tennessee and North Carolina: U.S. Geological Survey Professional Paper 587, 23 p., and 1:125,000-scale map.

King, P.B., and Stupka, A., 1950, The Great Smoky Mountains—their geology and natural history. Science Monogram, v. 71, p. 31-43.

Kish, S.A., 1991, Potassium-argon dating in the western Blue Ridge of North Carolina and Tennessee, in Kish, S.A., ed., Studies of Precambrian and Paleozoic Stratigraphy in the western Blue Ridge: Carolina Geological Society Field Trip Guidebook, p.69-77.

Kish, S.A., Merschat, C.E., Mohr, D.W., and Wiener, L.S., 1975, Guide to the Geology of the Blue Ridge South of the Great Smoky Mountains, North Carolina. Carolina Geological Society Field Trip Guidebook, Raleigh, 49 p.

Koch, C.A., 1974, Debris slides and related flood effects in the 4-5 August 1938 Webb Mountain cloudburst; Some past and present environmental geomorphic implications [M.S. thesis]: Knoxville, University of Tennessee, 112p.

Kochel, R.C., 1990, Humid fans of the Appalachian Mountains, in Rachocki, A.H., and Church, M., eds., Alluvial fans: a field approach: New York, Wiley, p. 109-129.

Kohl, M., 2000, Welcome to the Gray Site: Tennessee Geological Survey, http://www.state.tn.us/environment/tdg/gray/

Kohn, M.J. and Malloy, M.A., 2004, Formation of monazite via prograde metamorphic reactions among common silicates: Implications for age determinations. Geochimica et Cosmochimica Acta, Vol. 68, No. 1, p. 101-113.

Kunk, M.J., Wintsch, R.P., Southworth, S., Mulvey, B., Naeser, C., and Naeser, N., 2004, Multiple Paleozoic metamorphic histories, fabrics, and faulting in the Westminster and Potomac composite terranes, central Appalachian Piedmont, northern Virginia and southern Maryland *in* : Geology of the National Capital Region—Field trip Guidebook. Southworth, S., and Burton, W., eds., USGS Circular 1264, p. 163-188.

Laney, F.B., 1907, Copper deposits of Swain County in Pratt, J.H., The mining industry in North Carolina during 1906: North Carolina Geological and Economical Survey, Economic Paper 14, p. 72-79.

Lesure, F.G., Force, E.R., Windolph, J.F., and Hill, J.J., 1977, Mineral resources of the Joyce Kilmer-Slickrock Wilderness, North Carolina-Tennessee: U.S. Geological Survey Bulletin 1416, 89 p.

Lipin, B.R., 1984, Chromite from the Blue Ridge province of North Carolina: American Journal of Science, v. 284, n.4-5, p. 507-529.

Massey, M.A., and Moecher, D.P., in press, Metamorphic, structural, and fabric evolution of a segment of the eastern Blue Ridge-western Blue Ridge boundary in central western North Carolina. Tectonics.

Matmon, A., Bierman, P.R., Larsen, J., Southworth, S., Pavich, M., and Caffee, M., 2003a, Temporally and Spatially Uniform Rates of Erosion in the Southern Appalachian Great Smoky Mountains: Geology, v. 31, no. 2, p.155-158.

Matmon, A., Bierman, P.R., Larsen, J., Southworth, S., Pavich, M., Finkel, R., Caffee, M., 2003b, Erosion of an ancient Mountain Range, The Great Smoky Mountains, North Carolina and Tennessee: American Journal of Science, v. 303, p. 817-855.

McLaughlin, R. E., and Darrell, J. H., II, 1972, Age determination of the bauxite associated geomorphic surface on the Paleozoic old land in Northwest Georgia [abs.]; Geological Society of America, Abstracts with Programs, v. 4, no. 2, p. 90.

McLaughlin, R. E., and McBroom, M. N., 1980, Geographic distribution of early Cenozoic palynomorph assemblages in West Tennessee [abs.]: Geological Society of America, Abstracts with Programs, v.12, no. 4, p. 201.

Mershat, C.E., and Wiener, L.S., 1988, Geology of the Sandymush and Canton quadrangles, North Carolina. North Carolina Geological Survey Bulletin 90, 66 p., and scale 1:24,000.

Michalek, D.D., 1968, Fanlike features and related periglacial phenomena of the southern Blue Ridge [PhD. Thesis]: Chapel Hill, Univeristy of North Carolina, 198 p.

Mills, H.H., 1981, Boulder deposits and the retreat of mountain slopes, or, "Gully Gravure" revisited: Journal of Geology, v.89, p. 649-660.

Mills, H.H., 1982, Piedmont-cove deposits of the Dellwood Quadrangle, Great Smoky Mountians, U.S.A.: Morphometry: Zeitschrift fur Geomorphologie, v.26, p. 163-178.

Mills, H.H., 1986, Piedmont-cove deposits of the Dellwood quadrangle, Great Smoky Mountains, North Carolina, U.S.A.: Some aspects of sedimentology and weathering: Biuletyn Peryglacjalny, v.30, p. 91-109.

Mills, H.H., 2000a, The relationship of slope angle to regolith clast size: A study based on surficial mapping in the southern Blue Ridge province, western North Carolina: Southeastern Geology, v. 39, nos. 3 and 4, p. 243-258.

Mills, H.H., 2000b, Controls on the form, process, and sedimentology of alluvial fans in the central and southern Appalachians, Southeastern U.S.A. Southeastern Geology, v. 39, nos. 3 and 4, p. 281-313.

Mills, H.H., and Allison, J.B., 1994, Controls on the variation of fan-surface age in the Blue Ridge Mountains of Haywood County, North Carolina: Physical Geography, v.15, no. 5, p. 465- 480.

Mills, H.H., and Allison, J.B., 1995a, Weathering and soil development on fan surfaces as a function of height above modern drainage ways, Roan Mountain, North Carolina: Geomorphology, v. 14, no.1, p. 1-17.

Mills, H.H., and Allison, J.B., 1995b, Weathering rinds and the evolution of Piedmont slopes in the southern Blue Ridge Mountains: The Journal of Geology, v. 103, p. 379-394.

Mills, H.H., and Granger, D.E., 2002, Cosmogenic isotope burial dating reveals 1.5 million-year-old fan deposit in Blue Ridge Mountains of North Carolina. Geological Society of America Abstracts with Programs, v.34, no.2, p. A-32.

Mills, H.H., and Whisner, S.C., 2000, Old fan deposits on the northwestern flanks of Chilhowee Mountain, southeastern Tennessee: Indicators of an inactive landscape?: American Quaternary Association program and Abstracts of the 16th Biennial Meeting, 2000, Fayetteville, Arkansas, p. 84.

Misra, K.C., and Lawson, J.S., 1988, Paleotectonic setting of massive sulfide deposits of the Ducktown mining district, Tennessee, in Zachrisson, Ebbe, ed., Proceedings of the Seventh Quadrennial IAGOD Symposium: Stuttgart, E., Schweizerbait'sche Verlagsbuchhandlung, p. 685-692.

Moecher, D., Tracy, R. J., and Miller, B.V., 2004, Taconian metamorphism alive and well in the southern Blue Ridge. Geological Society of America Abstracts with Program, vol. 36, no. 1, p. 61.

Mohr, D.W., 1973, Stratigraphy and structure of part of the Great Smoky and Murphy belt Groups, western North Carolina: American Journal of Science, v. 273-A, p. 41-71.

_____1975, Geologic map of the Noland Creek quadrangle, North Carolina: North Carolina Geological Survey Geologic Map GM 158-NE, scale 1:24,000.

Mohr, D.W., and Newton, R.C., 1983, Kyanite-staurolite metamorphism in sulfidic schists of the Anakeesta Formation, Great Smoky Mountains, North Carolina: American Journal of Science, v. 283, p. 97-134.

Moneymaker, B.C., 1939, Erosional effects of the Webb Mountain (Tennessee) cloudburst of August 5, 1938: Journal of Tennessee Academy of Science, v.14, p.190-196.

Montes, Camillo, and Hatcher, R.D., Jr., 1999, Documenting Late Proterozoic Rifting in the Ocoee Basin, Western Blue Ridge, North Carolina. Southeastern Geology, v. 39, no. 1, p. 37-50.

Moore, H. L., 1986, Wedge failures along Tennessee highways in the Appalachian region: their occurrence and correction: Bulletin Association of Engineering Geologists, v. 23, no. 4, p. 441-460.

Naeser, C.W., Naeser, N.D., Kunk, M.J., Morgan, B.A., III, Schultz, A.P., Southworth, C.S., and Weems, R.E., 2001, Paleozoic through Cenozoic uplift, erosion, stream capture, and deposition history in the Valley and Ridge, Blue Ridge, Piedmont, and Coastal Plain provinces of Tennessee, North Carolina, Virginia, Maryland, and

District of Columbia [abs.]: Geological Society of America Abstracts with Programs, v. 33, no. 6, p. A312.

Naeser, N., Naeser, C., Southworth, S., Morgan, B., and Schultz, A., 2004, Paleozoic to recent tectonic and denudation history of rocks in the Blue Ridge province, central and southern Appalachians— evidence from fission-track thermochronology. [(abs.): Geological Society of America Abstracts with Programs, v.36, no.2, p.114.]

Neuman, R. B., 1955, Midddle Ordovician rocks of the Tellico-Sevier belt, eastern Tennessee: U.S. Geological Survey Professional paper 274-F, p. F141-178.

Neuman, R. B., 1960, Geology of the Wildwood quadrangle, Tennessee: U.S. Geological Suvey Geologic Quadrangle GQ-130, scale 1:24,000.

Neuman, R.B., and Nelson, W.H., 1965, Geology of the western Great Smoky Mountains, Tennessee. U.S. Geological Survey Professional Paper 349-D, 81 p., map scales 1:24,000 and 1:62,500.

Neuman, R.B., and Wilson, R.L., 1960, Geology of the Blockhouse Quadrangle, Tennessee: U.S. Geological Survey Geologic Quadrangle Map GQ-131, scale 1:24,000.

Oriel, S. S., 1950, Geology and mineral resources of the Hot Springs window, Madison County, North Carolina: N.C. Dept. Conserv. Cevel., Div. Min. Res., Bull. 60, 70p.

Overstreet, E. F., 1964, Geology of the southeastern bauxite deposits: U. S. Geological Survey Bulletin 1199-A, 19 p

Ownby, S.E., Miller, C.F., Berquist, P.J., Carrigan, C.W., Wooden, J.L., and Fullagar, P.D., 2004, U-Pb geochronology and geochemistry of a portion of the Mars Hill terrane, North Carolina-Tennessee: Constraints on origin, history, and tectonic assembly, in, Tollo, R.P., Corriveau, L., McLelland, J., and Bartholomew, M.J., eds., Proterozoic tectonic evolution of the Grenville orogen in North America: Boulder, Colorado, Geological Society of America Memoir 197, p. 609-632.

Pierce, K.L., 1965, Geomorphic significance of a Cretaceous deposit in the Great valley of southern Pennsylvania, *in,* Geological Survey Research 1965: U.S. Geological Survey Professional Paper 525-C, p. C152-C156.

Rankin, D.W., Drake, A.A., Jr., and Ratcliffe, N.M., 1989, Geologic map of the U.S. Appalachians showing the Laurentian margin and the Taconic orogen, plate 2, in Hatcher, R.D., Jr., Thomas, W.A., and Viele, G.W., eds., The Appalachian-Ouachita orogen in the United States: Geological Society of America, The Geology of North America, Boulder, v. F-2, scale 1:2,000,000.

Rast, N., and Kohles, K.M., 1986, The origin of the Ocoee Supergroup. American Journal of Science, v. 286, p. 593-616.
Reheis, M.J., 1972, Periglacial features in the Southern Appalachian Mountains. [B.S. Thesis]: University of Georgia, Athens, Ga., 26p.

Richter, D.M., 1973, Periglacial features in the Central Great Smoky Mountains. [Ph.D. thesis]: University of Georgia, Athens, Ga., 156 p.

Robinson, G.R., Jr., Lesure, F.G., Marlow, J.I., II, Foley, N.K., and Clark, S.H., 1992, Bedrock geology and mineral resources of the Knoxville $1^{o}x2^{o}$ quadrangle, Tennessee, North Carolina, and South Carolina: U.S. Geological Survey Bulletin 1979, 73 p. with 1:250,000-scale map.

Rodgers, John, 1953, Geologic map of east Tennessee, with explanatory text: Tenn. Dept. of Conserv. Bull. 58, pt. 2, 168 p.

Rozanski, George, 1943, Stone-centered polygons: Journal of Geology, v.51, p. 330- 341.

Safford, J. M., and Killebrew, J.B., 1876, The elementary geology of Tennessee: Nasheville, Tenn., 255p.

Schultz, Art, 1998, Geologic map of the Mount LeConte 7.5 min. quadrangle, Great Smoky Mountains National Park, Tennessee and North Carolina: U.S. Geological Survey Open-File Report 98-32, 10p., map scale 1:24,000.

Schultz, Art, 1999, Geologic maps of the Mount Guyot, Luftee knob, and Cove Creek Gap 7.5-minute quadrangles, Great Smoky Mountains National Park, Tennessee and North Carolina: U.S. Geological Survey Open-File Report 99- 536, 8p., and 4 1:24,000-scale maps.

Schultz, A.P., and Southworth, C.S., 2000, Geologic history and geologic map of Great Smoky Mountains National Park: Special Publication of the Great Smoky Mountains Natural History Association, map scale 1:200,000 http://www.smokiesstore.org/browse.cfm/4,58.htm

Schultz, Art, Southworth, Scott, Fingeret, Carrie, and Weik, Tom, 2000, Digital Geologic Map and Web site of the Mount LeConte 7.5-minute quadrangle, Great Smoky Mountains National Park, Tennessee and North Carolina, U.S.Geological Survey Open File Report 00-261 http://geology.er.usgs.gov/eespteam/Mtleconte/

Simpson, E.L., and Eriksson, K.A., 1989, Early Cambrian progradational and transgressive sedimentation patterns in Virginia: An example of the early history of a passive margin. Journal of Sedimentary Petrology. Vol. 60, No. 1, p. 84-100.

Southworth, Scott, and Aleinikoff, J., N., in press, Mesoproterozoic rocks of the Great Smoky Mountains National Park Region, Tennessee and North Carolina. Precambrian Research.

Southworth, Scott, 1995, Preliminary geologic map of the Great Smoky Mountains National Park within the Fontana Dam and Tuskeegee quadrangles, Swain County, North Carolina: U.S. Geological Survey Open-File Report 95-264, 33 p. and map scale 1:24,000. http://geology.er.usgs.gov/eespteam/smoky/

Southworth, Scott, 2001, The link between Geology, GIS, and ATBI plots, ATBI Quarterly Spring Newsletter, Discover Life In America, Gatlinburg, TN, p. 6-7.

Southworth, Scott, Chirico, Peter, and Putbrese, Trevor, 1999, Geologic story and map of parts of the Cades Cove and Calderwood quadrangles, Tennessee and North Carolina, Great Smoky Mountains National Park: U.S. Geological Survey Open- File Report 99175, map scale 1:24,000.

Southworth, S., Schultz, A., Denenny, D., and Triplett, J., 2003, Surficial geologic map of the Great Smoky Mountains National Park Region, Tennessee and North Carolina. U.S. Geological Survey Open File Report 03-381, scale 1:100,000.

Stose, G.W., and Stose, A.J., 1949, Ocoee Series of the southern Appalachians: Geological Society of America Bulletin, v. 60, no. 1, p. 267-320.

Thigpen, J. R., and Hatcher, R.D., Jr., 2004, New digital geologic map of the western Blue Ridge, SE Tennessee, SW North Carolina, northern Georgia: A tool for regional analysis of the Ocoee basin [(abs.): Geological Society of America Abstracts with Programs, v.36, no.5, p.503.]

Torbett, C.A., III, and Clark, G.M., 1985, Morphology, pedology, and origin of selected sorted patterned ground, Great Smoky Mountains National Park, Tennessee/North Carolina (abs.): Geological Society of America Abstracts with Programs, v. 17, no. 2, p. 139.

Tschudy, R. H., 1965, An Upper Cretaceous deposit in the Appalachian Mountains: U. S. Geological Survey Professional Paper 525-B, p. B64 - B68.

Tull, J.F., Kish, S.A., Groszos, M.S., and Campbell, S.K., 1998, Laurentian magmatism of the southern Appalachian Blue Ridge: Post-Iapetan rifting. Geological Society of America Bulletin v. 110, no. 10, p. 1281-1303.

Ulrich, E.O., 1929, Ordovician troilobites of the Family Telephidae and concerned stratigraphic correlations: U.S. National Museum Preoced., v.76, art. 21, 101p.

U.S. Geological Survey, 1949, Floods of August 1940 in the southeastern states: U.S. Geological Survey Water Supply Paper 1066, 554p.

Walker, D., and Driese, S.G., 1991, Constraints on the position of the Precambrian Cambrian boundary in the southern Appalachians. American Journal of Science, Vol. 291, p. 258-283.

Wallace, S.C., Nave, J.W., Burdick, K.M., 2002, Preliminary report on the recently discovered Gray Fossil Site (Miocene), Washington Co., Tennessee: With comments on observed paleopathologies - The advantages of a large sample: Journal of Vertebrate Paleontology, v. 22, supplement to no. 3, p. 117.

Whittecar, G.R., and Duffy, D.B., 2000, Geomorphology and stratigraphy of late Cenozoic alluvial fans, Augusta County, Virginia, U.S.A. Southeastern Geology, v. 39, no.2 and 3, p. 259-279.

Wiener, L.S., and Merschat, C.E., 1992, Geologic map of southwestern North Carolina including adjoining southeastern Tennessee and northern Georgia: North Carolina Geological Survey, scale 1:250,000.

Wilson, C.W., Jr., 1935, The Great Smoky thrust fault in the vicinity of Tuckaleechee, Wear, and Cades Cove, Blount and Sevier Counties, Tennessee: Journal of the Tennessee Academy of Science, v.10, no.1, p. 57-63.

INDEX TO GEOLOGIC MAPPING

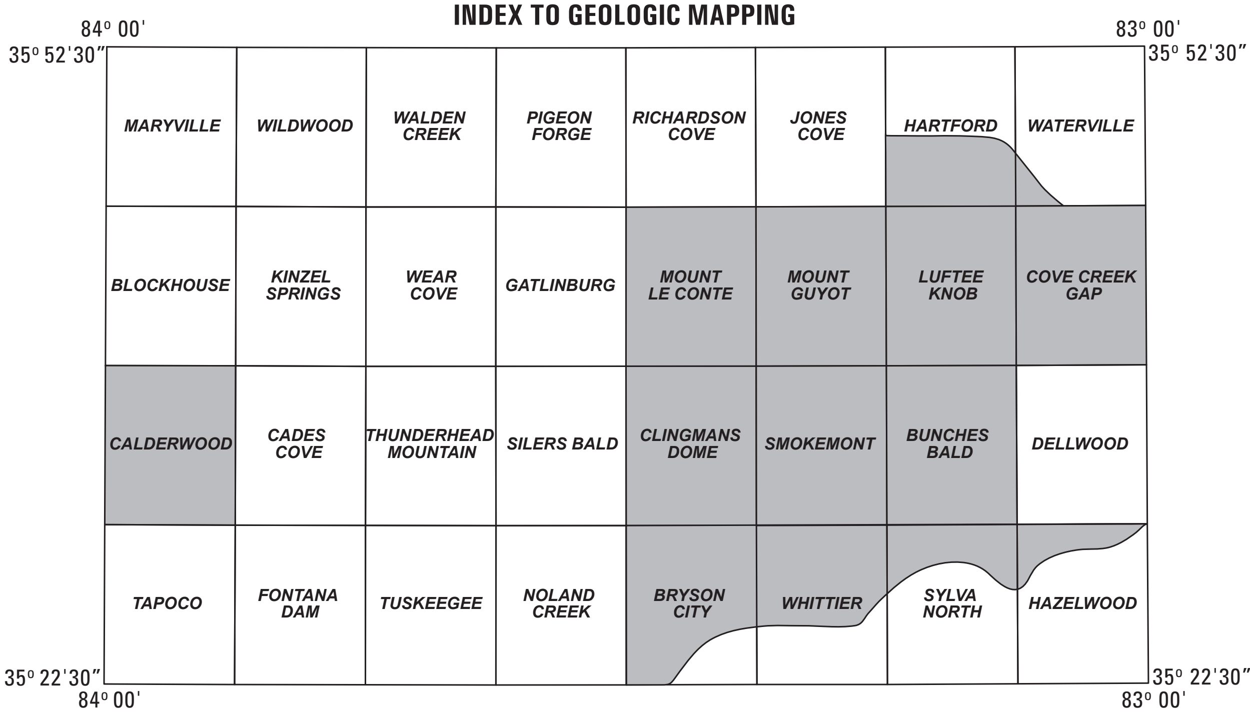

Fig. 1 SOURCES OF DATA FOR 7.5-MINUTE QUADRANGLES

Geologic map data are published at 1:24,000-scale, unless noted. Light shaded area is data published at 1:62,500-scale (Hadley and Goldsmith, 1963). Dark shaded area is reconnaissance data at 1:24,000-scale (Southworth, unpub. data). Data revised by Southworth (unpub. data).

Blockhouse:	Neuman and Wilson (1960)
Bryson City:	Cameron (1951); Hadley and Goldsmith (1963)(1:62,500)
Bunches Bald:	Hadley and Goldsmith (1963)(1:62,500)
Cades Cove:	Southworth, Chirico, and Putbrese (1999)
Calderwood:	Neuman and Nelson (1963)(1:62,500)
Clingmans Dome:	Hadley and Goldsmith (1963)(1:62,500)
Cove Creek Gap:	Hadley and Goldsmith (1963)(1:62,500); Keller (1981)
Dellwood:	Hadley and Goldsmith (1963)
Fontana Dam:	Southworth (1995) (north half); P.P. Fox (unpub. data)(southeast quadrant); Wiener and Mershat (1992)(1:250,000)(southwest quadrant)
Gatlinburg:	King (1964)
Hartford:	Hadley and Goldsmith (1963)(1:62,500)
Hazelwood:	Hadley and Goldsmith (1963)(1:62,500); Hadley and Nelson (1971)(1:250,000)
Jones Cove:	Hamilton (1961)
Kinzel Springs:	Neuman and Nelson (1965)
Luftee Knob:	Hadley and Goldsmith (1963)(1:62,500)
Maryville:	Cattermole (1962)
Mount Guyot:	Hadley and Goldsmith (1963)(1:62,500)
Mount Le Conte:	Hadley and Goldsmith (1963)(1:62,500); Schultz and others (2000)
Noland Creek:	Mohr (1975); Wiener and Mershat (1992)(1:250,000)
Pigeon Forge:	King (1964)
Richardson Cove:	Hamilton (1961)
Silers Bald:	King (1964) (north half)
Smokemont:	Hadley and Goldsmith (1963)(1:62,500)
Sylva North:	Hadley and Goldsmith (1963)(1:62,500)
Tapoco:	Lesure and others (1977)(western half); Wiener and Mershat (1992)(1:250,000)(eastern half)
Thunderhead Mountain:	King (1964)(north half)
Tuskeegee:	Southworth (1995)(north half); P.P. Fox (unpub. data)(south half)
Walden Creek:	King (1964)
Waterville:	Hadley and Goldsmith (1963)(1:62,500); Keller (1981)
Wear Cove:	King (1964)
Whittier:	Hadley and Goldsmith (1963)(1:62,500)
Wildwood:	Neuman (1960)

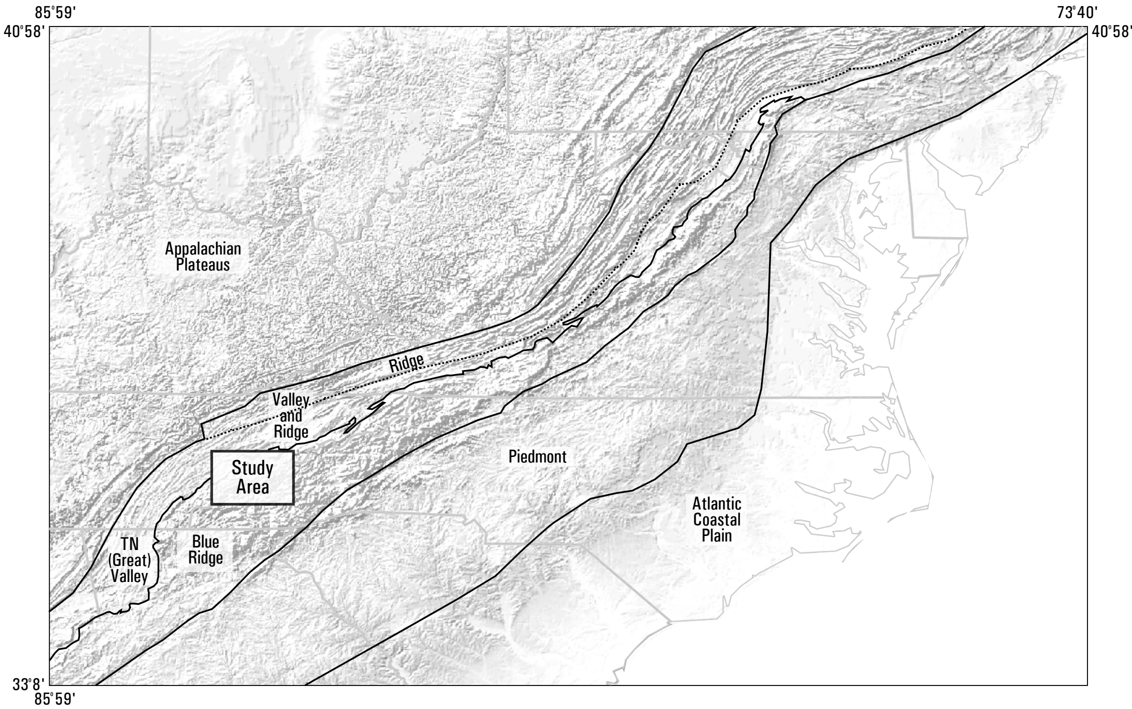

Fig. 2*A* Physiographic provinces of the Appalachian region. Study area is indicated.

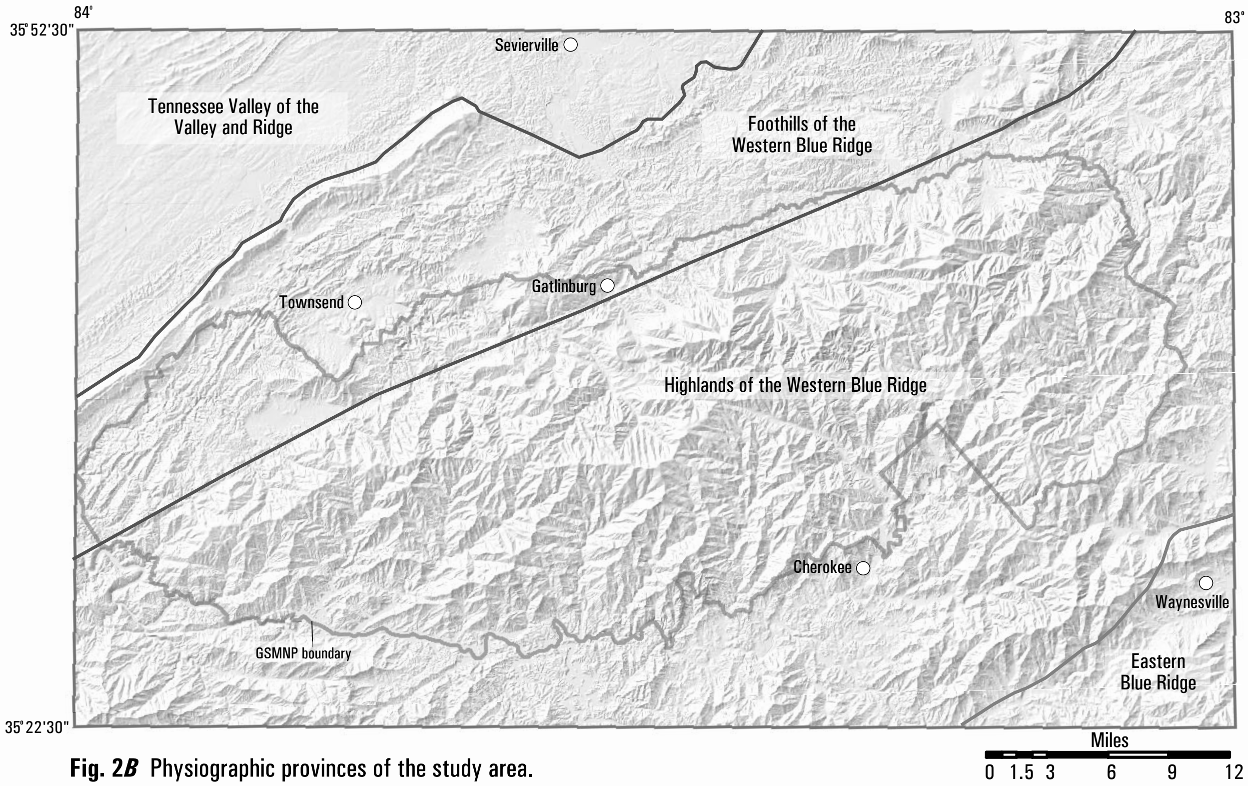

Fig. 2*B* Physiographic provinces of the study area.

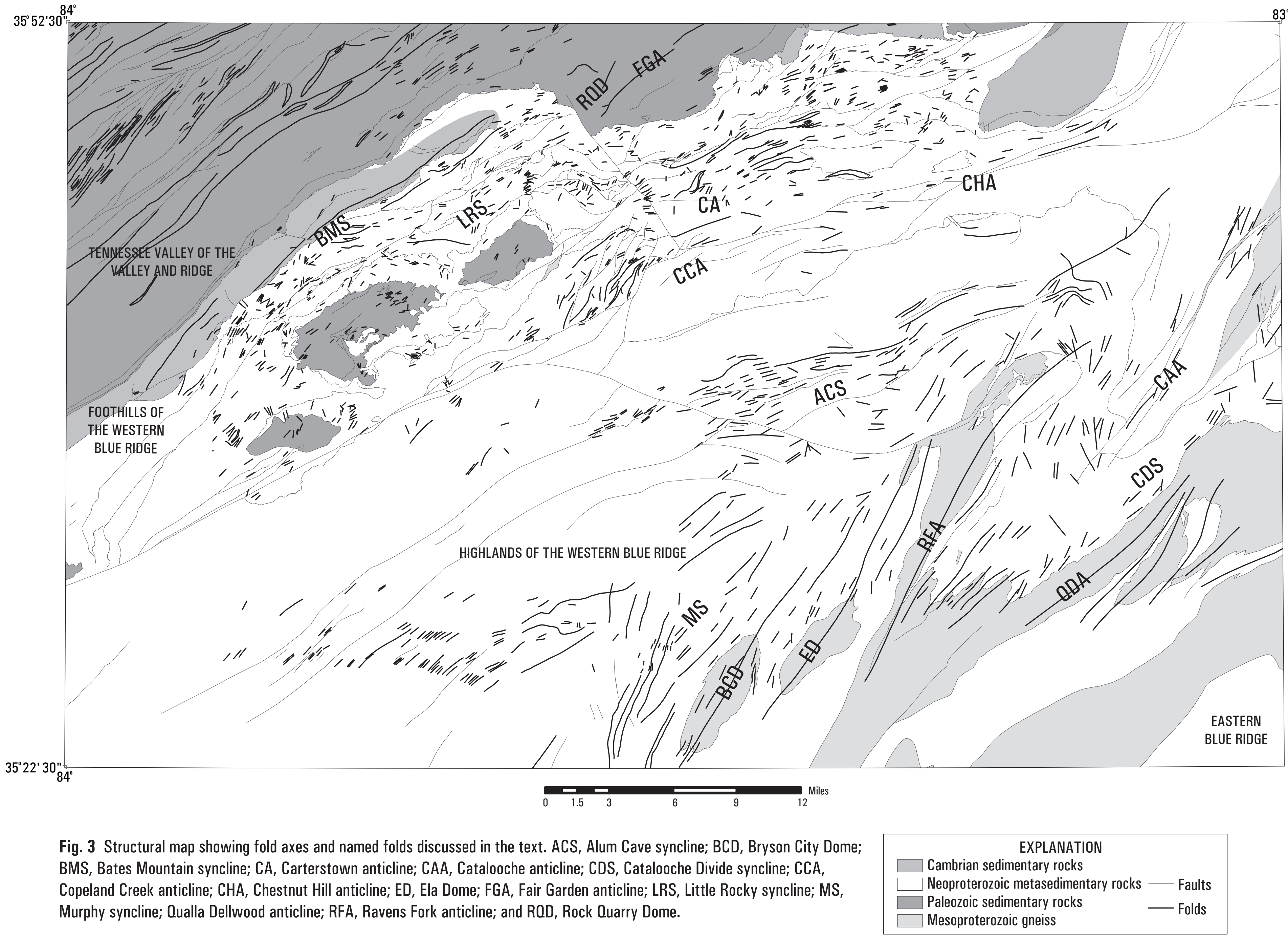

Fig. 3 Structural map showing fold axes and named folds discussed in the text. ACS, Alum Cave syncline; BCD, Bryson City Dome; BMS, Bates Mountain syncline; CA, Carterstown anticline; CAA, Catalooche anticline; CDS, Catalooche Divide syncline; CCA, Copeland Creek anticline; CHA, Chestnut Hill anticline; ED, Ela Dome; FGA, Fair Garden anticline; LRS, Little Rocky syncline; MS, Murphy syncline; Qualla Dellwood anticline; RFA, Ravens Fork anticline; and RQD, Rock Quarry Dome.

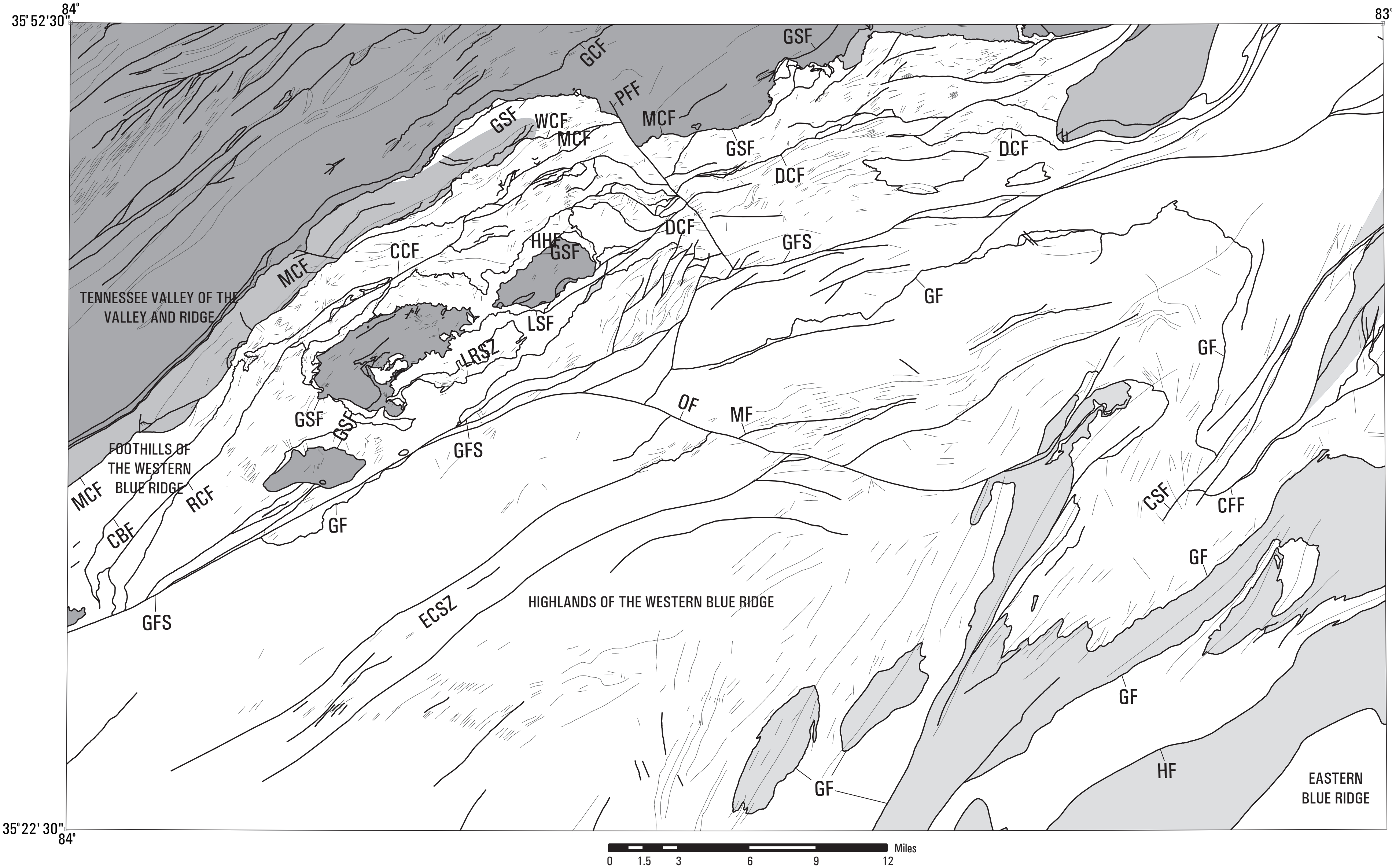

Fig. 4 Structural map showing faults and named faults discussed in the text. BSF, Bogle Spring fault; CBF, Capshaw Branch fault; CCF, Carr Creek fault; CFF, Caldwell Fork fault; CSF, Cold Spring fault; DCF, Dunn Creek fault; DVF, Dumplin Valley fault; ECSZ, Eagle Creek shear zone; GCF, Guess Creek fault; GF, Greenbriar fault; GFS, Gatlinburg fault system; GSF, Great Smoky fault; HF, Haysville fault; HHF, Happy Hollow fault; LRSZ, Little River shear zone; LSF, Line Spring fault; MF, Mingus fault; OF, Oconaluftee fault; PFF, Pigeon Forge fault; RCF, Rabbit Creek fault; WCF, Walden Creek fault; MCF, Miller Cove fault.

EXPLANATION

- Cambrian sedimentary rocks
- Neoproterozoic metasedimentary rocks
- Paleozoic sedimentary rocks
- Mesoproterozoic gneiss
- Faults
- Folds