GEOLOGIC INVESTIGATIONS IN THE BLUE RIDGE OF NORTHWESTERN NORTH CAROLINA

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Field Trip Guidebook 1983
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CAROLINA GEOLOGICAL SOCIETY

FIELD TRIP GUIDEBOOK

October 21 – 23, 1983

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Edited by

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Front cover: Southwest looking view, taken fro
Within the "collapse pit", of the main adit and
Stopes of the Cranberry mine

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DEFORMATIONAL HISTORY OF THE REGION BETWEEN THE GRANDFATHER MOUNTAIN AND MOUNTAIN CITY WINDOWS, NORTH CAROLINA AND TENNESSEE

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INTRODUCTION

The fieldwork on which this article is based was largely funded by the North Carolina Division of Land Resources between 1975 and 1978, during which time the Sherwood (Bartholomew and Wilson, in press), and Elk Park (Lewis and Bartholomew, in press) quadrangles were completed along with reconnaissance work in the Mountain City, Valle Cruise, Elk Mills and Boone quadrangles. Prior to completion of these quadrangles the area between the Mountain City and Grandfather Mountain windows (to the north and south, respectively) had received only cursory geological investigation since Keith (1903) completed the Cranberry Folio. Hamilton (in King and Gerguson, 1960) described and illustrated many of the crystalline rock types but, because the major emphasis of that work was on the Chilhowee Group and younger Paleozoic rocks and structures of Tennessee, only a very generalized map of these crystalline rocks was included. Likewise, Bryant and Reed (1970 a,b) dealt primarily with the rocks within the Grandfather Mountain window with only a perfunctory examination of those to the north of the window

These classic works were followed by work on the post-Grenville (~1000 my) cover rocks and related granitic intrusions (Rankin, 1970, 1975, 1976; Rankin and others, 1973), but the large-scale mapping (Rankin and others, 1972; Espenshade and others, 1975) which accompanied these articles still left a largely enigmatic region between the Grandfather Mountain and Mountain City windows. Jones (1976) mapped the Warrensville and West Jefferson quadrangles, just east of the area of present interest, and discussed both structure and petrology.

Recently Bartholomew and Lewis (in press) delineated the major Grenville-age plutonic suites and their country rocks, as well as the suite of later Precambrian peralkaline plutons, within the five crystalline massifs of the Blue Ridge geologic province of Maryland, Virginia and North Carolina. Based on their work, Bartholomew (1983b) constructed a palinspastic interpretation of Grenville-age metamorphic facies in the massifs prior to Paleozoic thrusting. Although Bartholomew and Lewis (in press) discuss both basement-cover rock relationships and the structural history in the region near the Grandfather Mountain window, their work is aimed at delineation of the Rockfish Valley-Fries-Fork Ridge-Linville Falls ductile deformation zone and the contrast of Grenville-age massifs and their cover rocks on opposing sides of this fault system.

Thus, the purpose of this article is to complement their regional study by discussing field relationships of the rocks and structural data in the area near the Grandfather Mountain window. Detailed structural studies have been done along other portions of the Rockfish Valley-Fries fault system in central (Bartholomew and others, 1981) and southwestern (Bartholomew and others, 1982) Virginia where detailed mapping also has been completed recently.

The locations of the four 7.5-minute quadrangles mapped in this study are shown on both a generalized tectonic map (Figure 1) and a generalized geologic map (Figure 2).

GRENVILLE OROGENESIS

According to Bartholomew and Lewis (in press) early (or pre-) Grenville-age layered gneisses form the country rocks of the Watauga, Elk park and Globe Massifs in north-

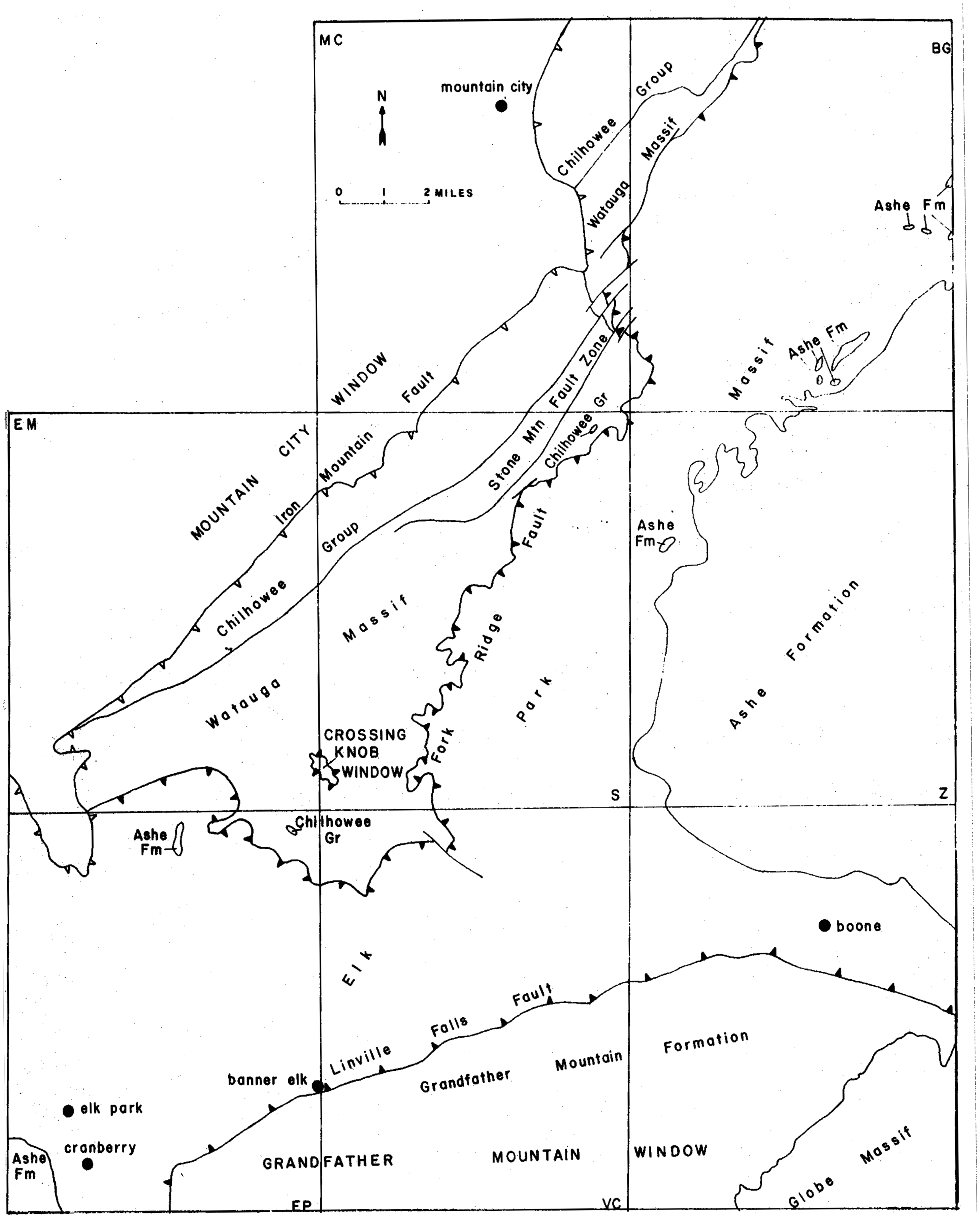


Figure 1: Generalized tectonic map showing quadrangle locations. Geology modified after Bartholomew (1983), Bartholomew and Gryta (1980), Bartholomew and Wilson (in press), Bryant and Reed (1970b), King and Ferguson (1960), Lewis and Bartholomew (in press); abbreviations for quadrangles: MC = Mountain City; BG = Baldwin Gap; EM = Elk Mills; S = Sherwood; Z = Zionville; EP = Elk Park; VC = Valle Crucis; B = Boone.

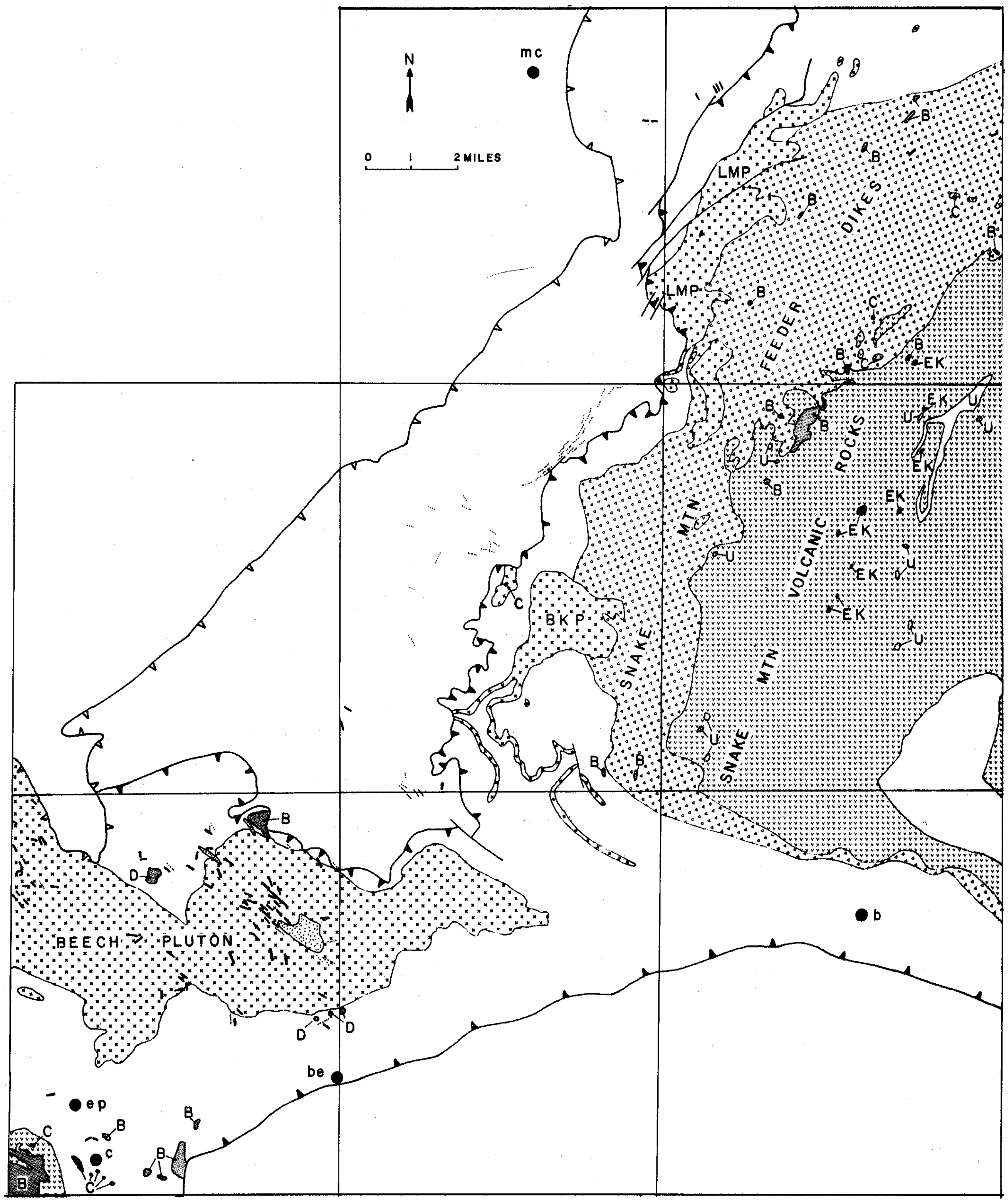


Figure 2: Generalized geologic map showing distribution of late Precambrian igneous rocks in and overlying the Elk Park Massif and intrusive dikes in the Watauga Massif; dotted lines and areas are felsic dikes related to Crossnore Suite and Mount Rogers volcanic rocks; heavy, short, dashed lines are intermediate to mafic dikes; B – Bakersville Gabbro; C – Cranberry magnetite ores; D – Dioritoid; EK – Elk Knob sulfide ores; SM – Snake Mountain volcanic rocks; u – ulframafic rocks; BKP – Buckeye Knob Pluton; LMP – Leander Mountain Pluton; b – Boone; be – Banner Elk; c – Cranberry; ep – Elk Park; mc – Mountain City; quadrangles outlined and major tectonic features are the same as in Figure 1.

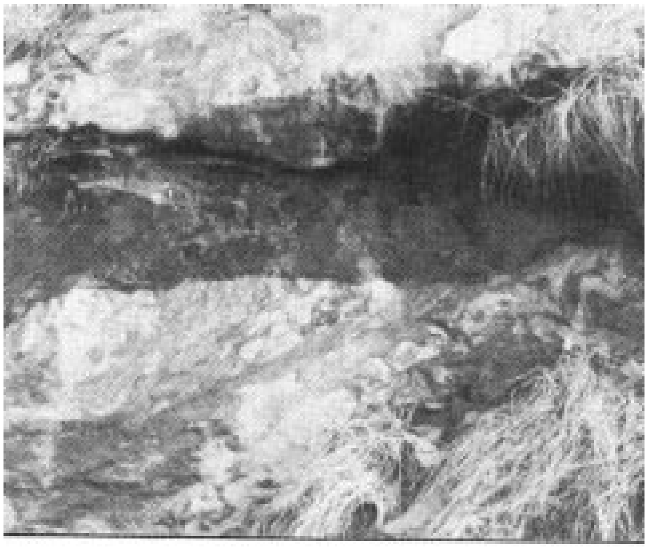


Figure 3: Exposure of Cranberry Gneiss along Roundabout Creek showing probable Grenville-age fold (lower right above 4-inch scale) refolded about younger Paleozoic(?) axial surfaces and classic Ramsey (1967) type-1 interference pattern (lower right-to left of scale). Both folds are cut by a late Precambrian (?) dike containing Paleozoic folds with axial planar foliation; outcrop is located on south side of North Carolina state road 1308 at Asheland in Ashe County within the Baldwin Gap quadrangle.

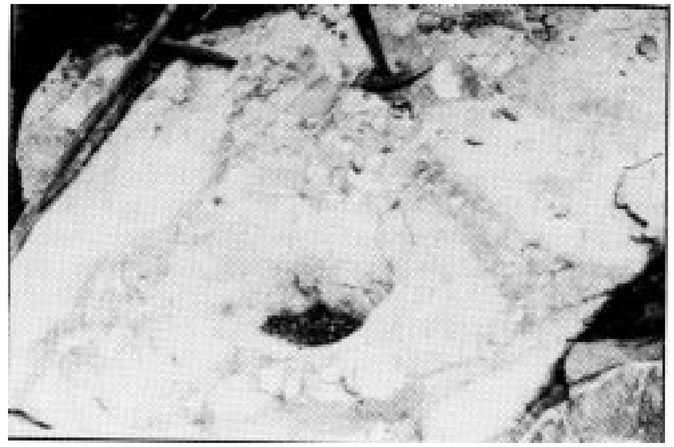


Figure 4: Exposure of basal boulder conglomerate in Chilhowee Group showing clasts derived from rocks of the Watauga Massif; outcrop is located in Creek bed in the northwestern corner of the Baldwin Gap quadrangle about 1.2 miles north of the Forge Creek School.

westernmost North Carolina (Figure 1). These layered gneisses, such as the Cranberry Gneiss (Keith, 1903), were intruded by suites of plutonic rocks and metamorphosed during the Grenville event placed by them at about 1000-1100 my (see also Fullagar and Bartholomew, this volume). Although the layered gneisses were undoubtedly folded (perhaps repeatedly) during the Grenville event, later Paleozoic deformation has rendered recognition of Grenville-age folds difficult. Examples of Grenville folds can be observed in outcrops where Paleozoic overprints are less intense and segregation layering, developed during Grenville metamorphism to upper amphibolite and lower granulite conditions, is well preserved (Figure 3).

LATE PRECAMBRIAN GEOLOGIC HISTORY

Following the Grenville event, extensive erosion of this crystalline terrane occurred during the late Precambrian. In the eastern part (Elk Park and Globe Massifs) of the reconstructed Grenville terrane (Bartholomew, 1983b), erosion was terminated by late Precambrian volcanism and sedimentation (about 750-700 my ago) whereas in the western part (Watauga Massif) erosion continued until deposition of Chilhowee Group sediments, about 650-600 my ago (Figure 4), in which rock clasts are found which are clearly recognizable as being derived from the Watauga Massif (Figure 4).

Because of the more extensive distribution of the Elk Park Massif and its cover rocks (Figure 2) within our study area a better sequential picture can be established for it from the Grenville event through late Precambrian time (Figure 5 and Table 1).

Crosscutting field relationships in the Elk Park quadrangle established that the older Grenville-age granitoid plutons (such as the laurel Creek Pluton, Figure 5) are cut by a series of younger granitoid plutons. Both these younger plutons and stocks of Bakersville Gabbro are truncated by the Beech Pluton of the Crossnore Suite (Figure 2) which contains rare xenoliths of older rocks (Fiugre 6). In the Elk Park, Zionville and Baldwin Gap quadrangles, dikes and stocks of Bakersville Gabbro (Figure 7) intrude the lower portion of the Snake Mountain volcanic pile that constitutes the basal portion of the Ashe Formation. In the Elk Park quadrangle mineralization associated with the Cranberry magnetite deposit also affected the basal portion of the volcanic pile (Figure 2). Similar mineralization is found near the unconformity in the Baldwin Gap quadrangle as well (Figure 2). Thus the Pb-Pb age of 700 my (Davis and others, 1962) for the Beech Pluton establishes minimum ages for Bakersville Gabbro and associated Cranberry mineralization as well as for the basal part of the Snake Mountain volcanic pile. In the Sherwood, Zionville, and Baldwin Gap quadrangles some of the Snake Mountain feeder dikes also intrude portions of plutons of the

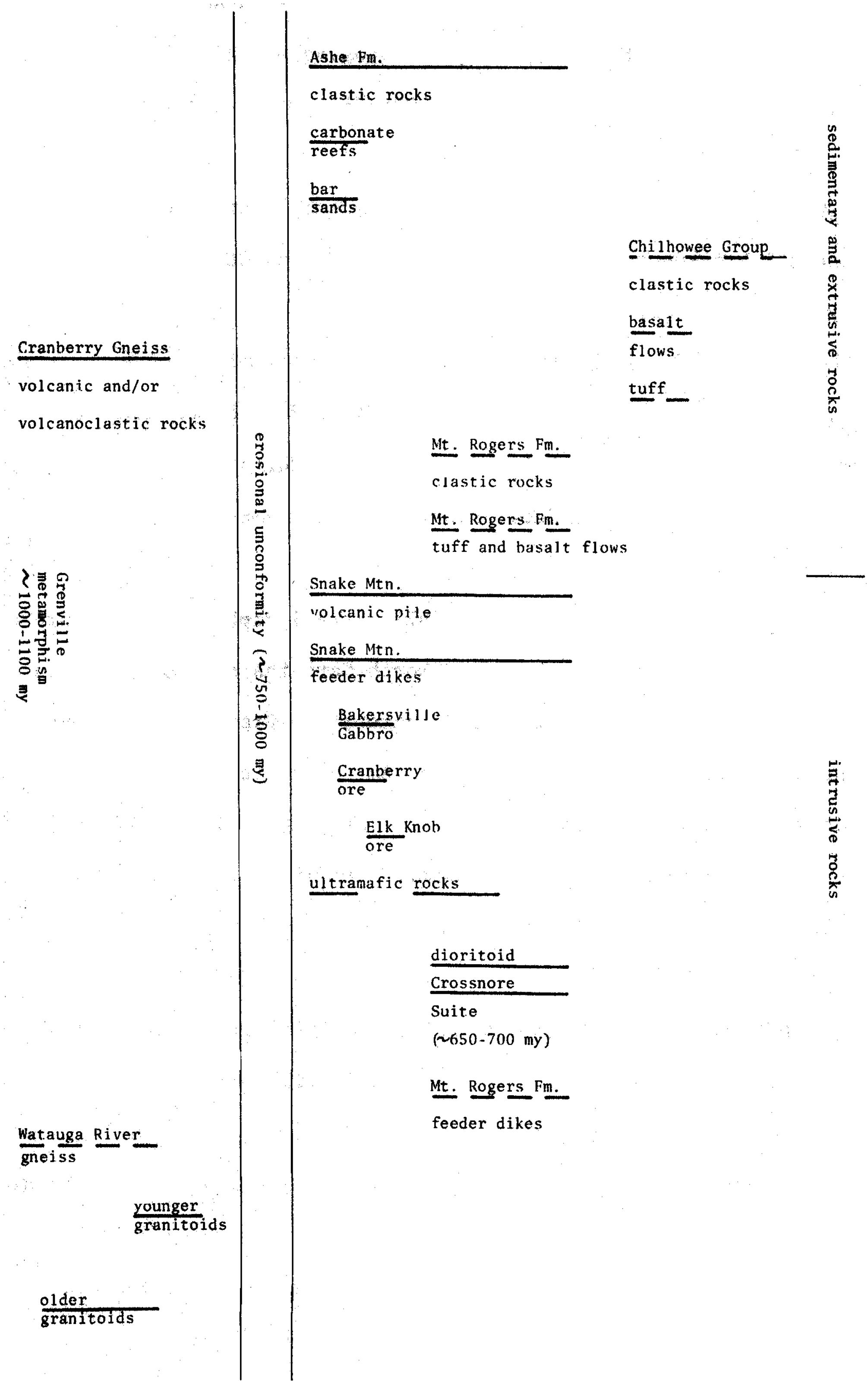


Table 1: Chart showing relative ages of rocks within and overlying the Elk Park and Watauga Massifs in northwestern North Carolina and adjacent states (Figures 2 and 5) solid line — rocks associated with the Elk Park Massif dashed line — rocks associatedwith the Watauga Massif.

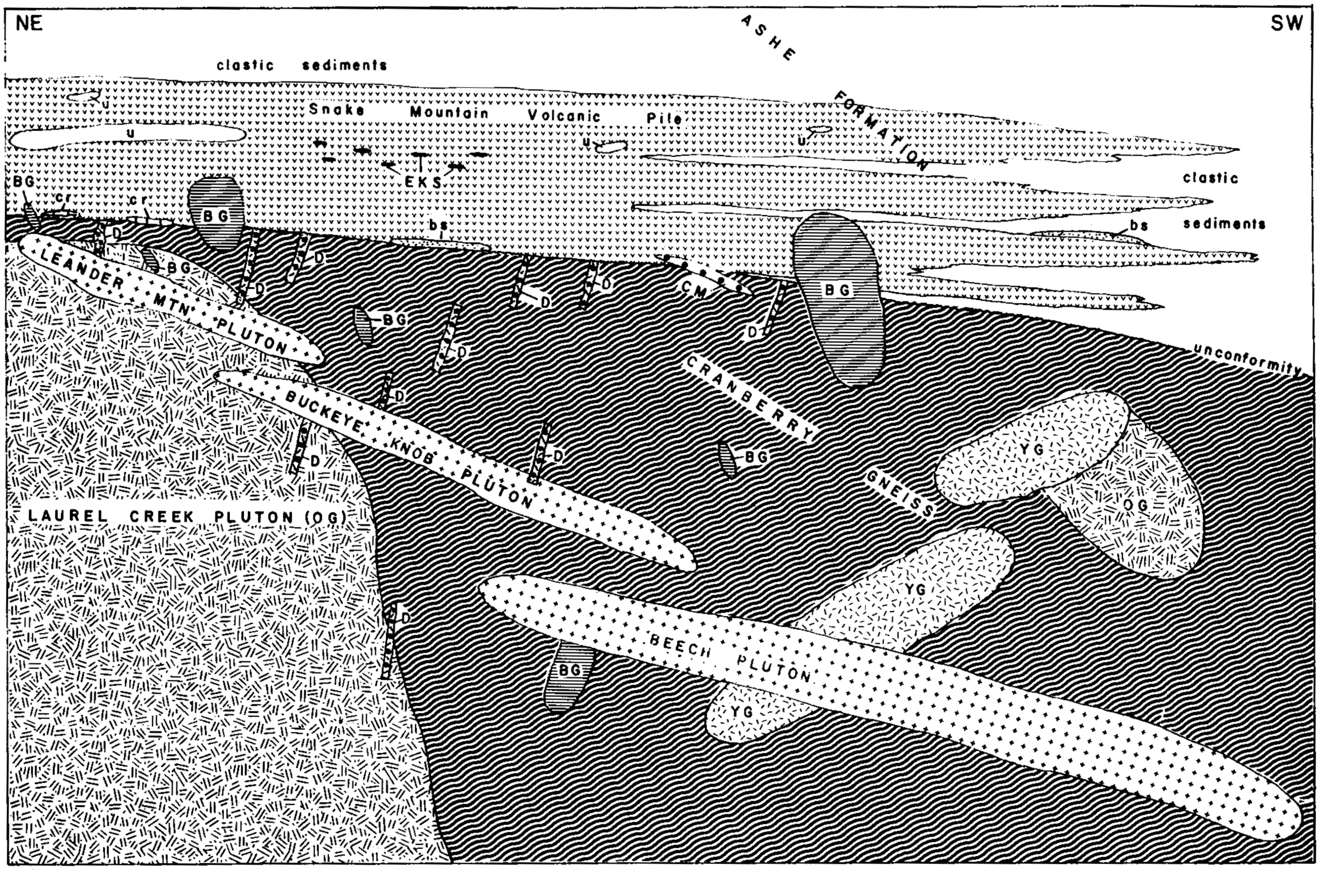


Figure 5: Diagrammatic sketch showing late Precambrian relationships among Grenville-age units of the Elk Park Massif and late Precambrian intrusive, volcanic and sedimentary rocks, BG – Bakersville Gabbro; bs – bar sands; cr – carbonate reefs; CM – Cranberry magnetite ore; D – Feeder dikes for Snake Mountain volcanic pile; EKS – Elk Knob sulfide ore; OG – older granitoid; u – ultramafic rocks; YG – younger granitoid.

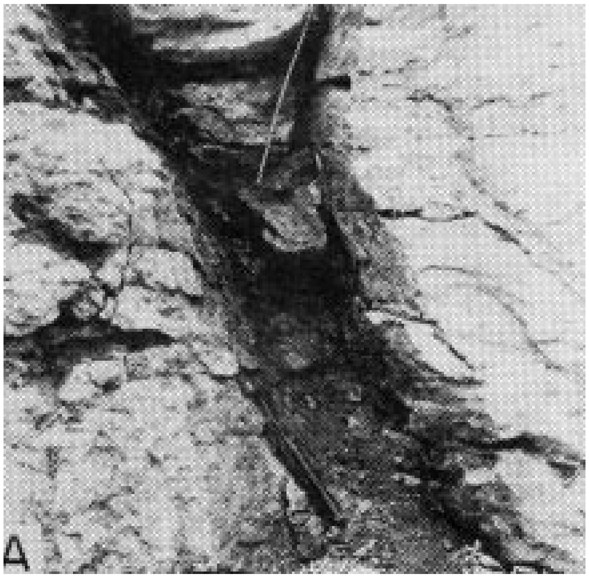
Crossnore Suite (Figure 2), thus indicating that Snake Mountain volcanism continued during and after emplacement of the Crossnore Suite.

Bartholomew and Lewis (in press) noted a close association of dioritoid rocks with border zone of several of the peralkaline granitic plutons of the Crossnore Suite. Among these dioritoid rocks are those that form the Cattron Pluton associated with the Striped Rock Pluton in nearby southwestern Virginia. We interpret the small bodies of dioritoid (Figure 2) which are mineralogically similar to Cattron diorite, to be similarly related to the Beech Pluton (Figure 2).

In the Zionville and Baldwin Gap quadrangles large, altered, ultramafic dikes and sills are common in both the upper portion and near the base of the Snake Mountain volcanic pile (Figures 2 and 5). Likewise, abundant copper mineralization at and near Elk Knob in the Zionville quadrangle apparently lies more or less at a stratigraphic level near the middle of the volcanic pile (Figures 2 and 5). Coarse-grained feeder dikes of the Snake Mountain volcanic field are found abundantly in the northern three quadrangles (Figure 2). These feeder dikes commonly contain disseminated sulfides. Within the Baldwin Gap quadrangle younger finer-grained dikes (mafic to intermediate composition) generally lacking sulfides, cut the older-feeder dikes (Figure 8). The finer-grained dikes are very abundant around, and intrusive into, the course-grained granitic rocks of the Beech) Pluton (Figure 2) suggesting an origin (like the dioritoid stocks and dikes) closely associated with the granites of the Crossnore Suite.

PALEOZOIC DEFORMATION

The principal period of Paleozoic metamorphism corresponds with development of the first generation of Paleozoic folds. These folds are developed in both the late Precambrian cover rocks structurally above the Fork Ridge-Linville Falls fault these earlier isoclinal folds (Figure 9A) have an associated axial planar foliation developed during the main Paleozoic metamorphism placed at about 450 my by Butler (1972,1973). Butler (1973) places emplacement of the Spruce Pine pegmatites at about 380 my during the latter stages of the main metamorphic event.



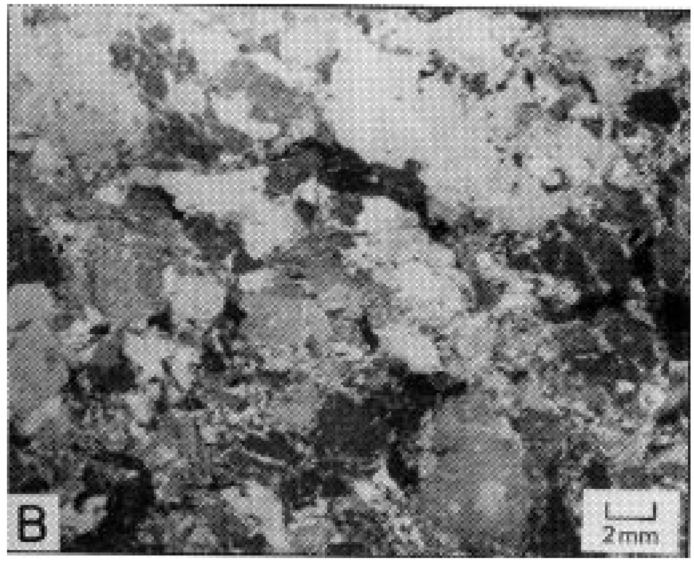


Figure 6: A. Rare xenolith of layered Cranberry Gneiss in coarser-grained granitic gneiss of the Beech Pluton. Outcrop is located along west side of a dirt road of Beech Mountain west of Buckeye Creek and about 1.5 miles northwest of the Pinnacles of the Beech in the Elk Park quadrangle.
B. Photomicrograph of typical granite of the Crossnore Suite. Sample collected from outcrop within small exposure of Beech granite at the head of Mill Creek in Carter County, Tennessee within the Elk Park quadrangle. Rock contains relict perthitic feldspar, plagioclase and quartz with epidote, chlorite, sericite and quartz present as part of the retrograde mineral assemblage.

Pegmatites of the Jefferson-Boone District (Lesure, 1968) are found in abundance in the Zionville and Baldwin Gap quadrangles and represent the extension of the Spruce Pine and Woodlawn pegmatite districts with the intervening portion of the belt above the Grandfather Mountain window having been removed by erosion. As also appears to be the case in the Spruce Pine and Woodlawn pegmatite districts (Lewis and Butler, in press) those of the Jefferson-Boone District are not found in the basement rocks and are generally found only in or near metasedimentary units in the cover rocks. This relationship suggests that the pegmatites resulted from partial melts of metasedimentary cover rocks during prograde metamorphism to upper amphibolite facies in this region. Their greater abundance and size in the Spruce Pine District probably reflects the greater abundance of metasedimentary rock there as compared with the Jefferson-Boone District where the cover rocks are largely of metavolcanic origin.

Structurally below the Fork Ridge-Linville Falls fault the main Paleozoic metamorphism only reached greenschist facies. Folds associated with this metamorphic event northwest of the Fork Ridge fault and within the Grandfather Mountain window are interpreted as approximately equivalent in age to the first generation of isoclinal folds (1, Figure 10) in the Ashe Formation above the fault.

As noted by Bartholomew and Lewis (in press) the relatively “dry” gneisses of the basement massifs show effects of the main Paleozoic metamorphic event to a considerably lesser degree than do the cover rocks. Evidence for this event in the basement gneisses typically consists of refolded segregation layering (Figure 3)in layered units like the Cranberry Gneiss and poorly developed foliation (typically biotite) in massive units like the Watauga River Gneiss and gneisses of the Crossmore Suite. Other massive Grenville-age granitic gneisses like the Laurel Creek Pluton commonly lack even a foliation, although xenoliths exhibit interference patterns indicative of polyphase folding (Figure 11).

The principal period of Paleozoic metamorphism was followed by thrusting along the Fries fault system (3, Figure 10). This thrusting, under ductile conditions, was associated with considerable retrograde metamorphism of basement rocks (Figure 12) as the Elk Park Massif was juxtaposed over both the Watauga Massif and rocks of the Grandfather Mountain window (Bartholomew and Lewis, in press). The prominent mylonitic foliation and layering which characterizes basement rocks in this region was developed during this event which is placed at about 350 my by Bartholomew and Lewis (in press) based on geochronological data of Dietrich and others (1969) and Odom and Fullagar (1973).

Tight to isoclinal folds, with poorly developed axial planar foliation, found in cover rocks of the Elk Park Massif (2, Figure 10) have been traced out and clearly show that earlier axial surfaces (1, Figure 10) have been refolded about these later folds in the Zionville quadrangle. Development of this second generation of folds (Figure 9c) is inferred to proceed thrusting along the Fries fault system. Butler (1973) rea-

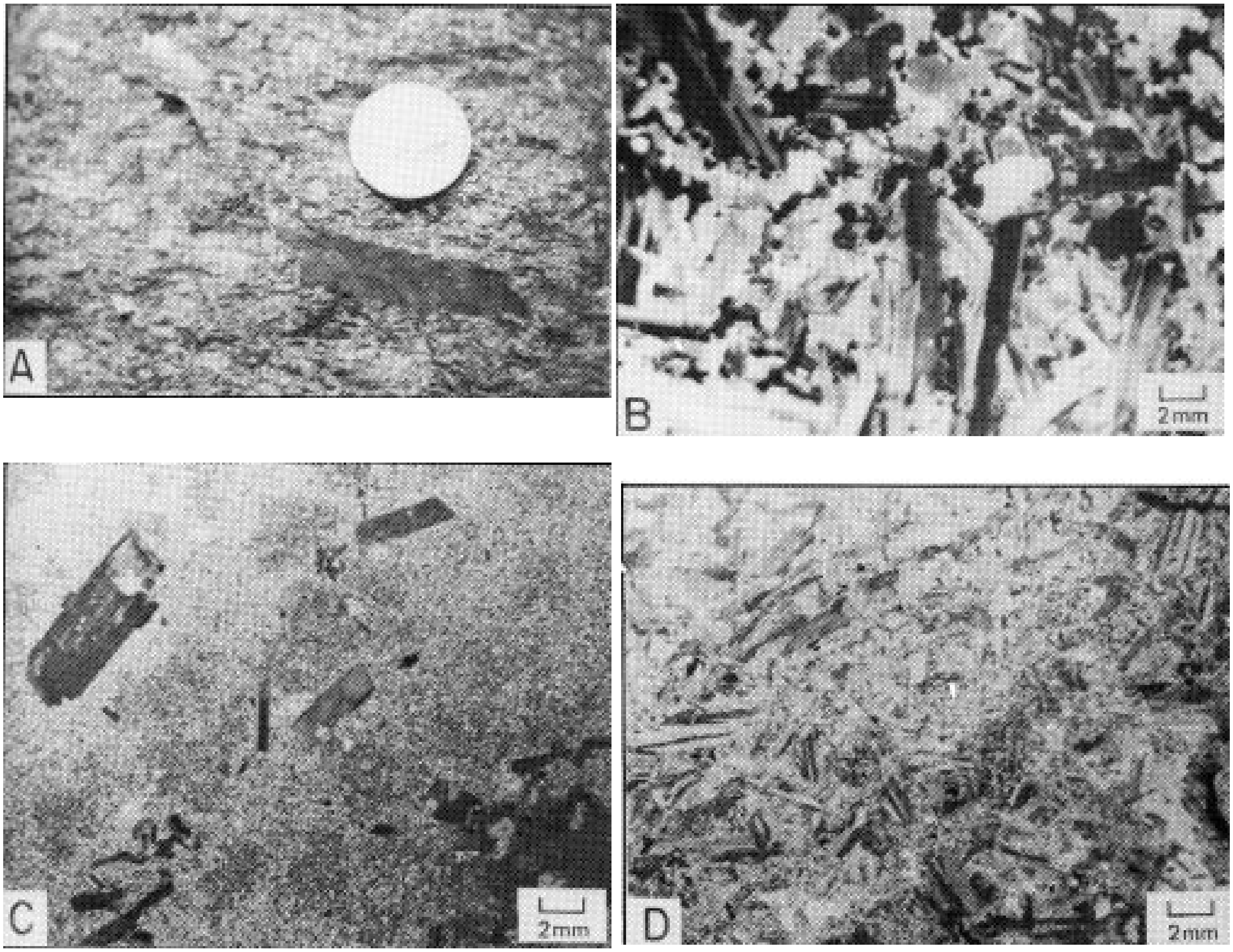


Figure 7: A. Close-up of foliated porphyritic Bakersville Babbro with mafic xenolith. Outcrop is located along north side North Carolina state road 1312 along 1.7 miles each of Whaley in Avery County within the Elk Park quadrangle.
B. Photomicrograph of unaltered, coarse-grained Bakersville Gabbro. Sample collected from dike along North Carolina state road 1311 about 0.8 miles due east of the junction of US 321 — US 421 in Watauga County within the Sherwood quadrangle. Rock contains plagioclase (An_{57}), olivine, hypersthene, biotite and magnetite.
C. Photomicrograph of fine-grained, porphyritic Bakersville metadiabase. Sample collected from outcrop at top of northern spur of Little Hump Mountain in Avery County within the Elk Park quadrangle. Rock consists of plagioclase laths and poikiloblastic garnet in a partially altered groundmass containing biotite, chlorite and hematite.
D. Photomicrograph of medium-grained Bakersville metagabbro. Sample collected from outcrop at the top of Hump Mountain in Avery County within the Elk Park quadrangle. Rock contains relict plagioclase laths and clinopyroxene within a partially altered groundmass containing biotite, quartz, hornblende, chlorite, hematite and apatite. Both C and D are from large stock of Bakersville Gabbro west of Cranberry.

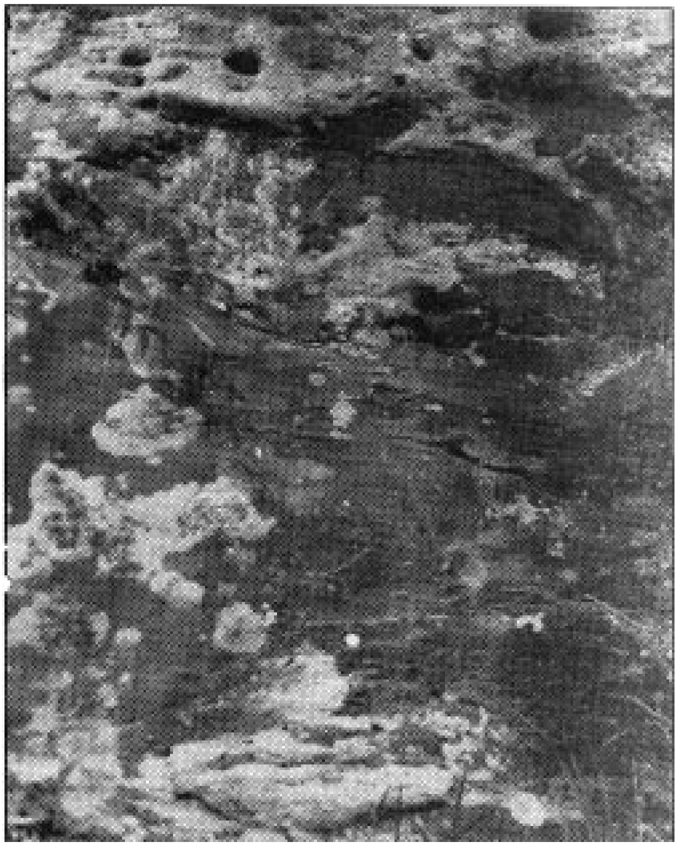


Figure 8: Exposure of Cranberry Gneiss invaded by coarse-grained mafic rock (feeder dikes of Snake Mountain volcanic pile) which is cut by younger, finer-grained dike. Outcrop is located on east side of North Carolina state road 1310 approximately 0.1 milessouth of the intersection with North Carolina state road 1315 along Laurel Creek in Ashe County within the Baldwin Gap quadrangle. Quarter and Knife mark the lower and upper contacts of the foliated fine-grained dike.

soned that these folds in the cover rocks are approximately the same age as Spruce Pine pegmatites (~380 my) emplaced under higher grade metamorphic conditions; thus, this second generation of folds would necessarily predate thrusting under lower grade metamorphic conditions. Small folds with axial planar mylonitic foliation did form in the recrystallized, mylonitic portions of the basement rocks during thrusting along the Fries fault zone.

The final Paleozoic deformational event in this region can be broken down somewhat into the sequential development of a series of structural features (4-8, Figure 10). However, all of these structures likely developed during the protracted Alleghanian event in late Paleozoic time when rocks of the Blue Ridge geologic-province were thrust over Paleozoic-age sedimentary rocks of the Valley and Ridge geologic province. Alleghanian deformation associated with thrusting along the Stone Mountain and then along the Blue Ridge fault systems typically progressed more toward the brittle realm (Figure 13) than is the earlier Paleozoic deformation which is totally within the ductile realm.

The sequence 4-8 portrayed on Figure 10 is developed from the following lines of reasoning:

Based on the map pattern in the Sherwood quadrangle, the doming (4, Figure 10) of the small Crossing Knob window preceded development of younger folds with axial planar crenulation cleavage (Figure 12). The dome probably developed as a fold associated with a structurally lower thrust beneath the window. Because it occupies a position structurally within the Stone Mountain thrust system (6, Figure 10) it is inferred to be slightly older than, or synchronous with, the thrusting.

Crenulation cleavage (Figure 9B) and associated folds (Figure 9A; 5, Figure 10) were preferentially developed in mylonitic gneiss and schist previously formed concomitantly with thrusting along the Fries fault system. Folds associated with the crenulation cleavage refolded earlier folds associated with development of the mylonitic gneiss and schist.

Thrusting along the Stone Mountain fault system (Figure 1; 6, Figure 10) is characterized by development of mylonite and ultramylonite fabrics (Higgins, 1971) in which mechanical processes clearly were dominant over recrystallization that characterized the earlier ductily formed mylonite

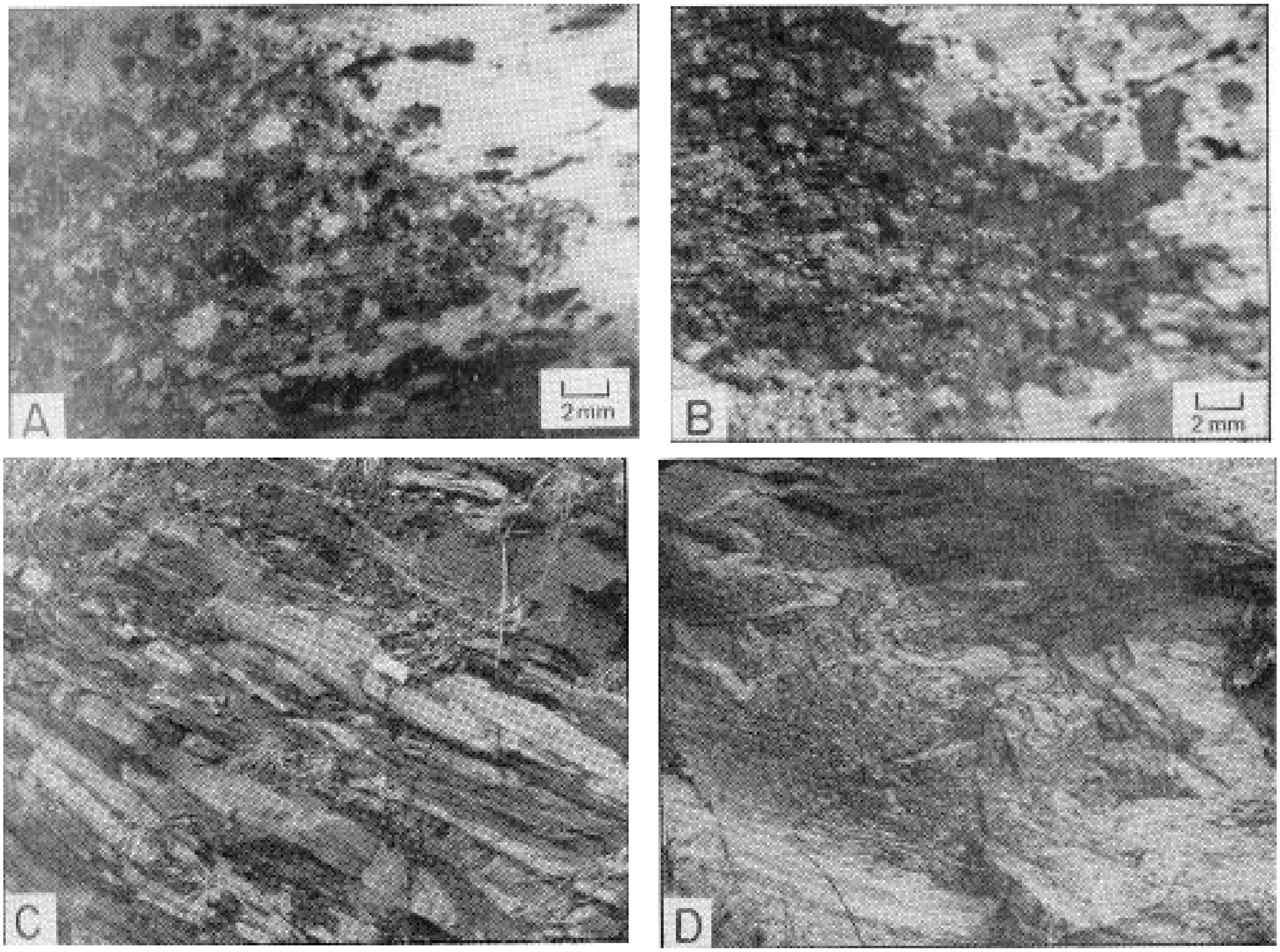


Figure 9: A. Photomicrograph of isoclinal fold with axial planar biotite foliation. Sample collected from outcrop of Ashe metasedimentary rock along north side of North Carolina state road 1306 about 0.6 miles south of Sugarloaf Mountain in Watauga County within the southern portion of the Zionville quadrangle. Rock contains feldspar, quartz, biotite, sphene and magnetite.

B. Photomicrograph of isoclinal fold with axial planar biotite foliation. Sample collected from outcrop on north side of North Carolina state road 1316 about 1.0 miles west of the junction with state road 1310 in Ashe County within the Baldwin Gap quadrangle. Rock contains a relict mineral assemblage of hornblende, biotite, perthitic feldspar, quartz and a retrograde metamorphic assemblage of biotite, clinozoisite, quartz, sphene and sericite.

C. Outcrop of interlayered siliceous gneiss and amphibolite along the east side of U.S. Highway 421 about 0.5 miles east of Zionville in Watauga County within the Zionville quadrangle. Earlier isoclinal folds (bottom-center) and most of the later tight of isoclinal folds appear nearly coaxial and thus produce Ramsey (1967) type-3 interference patterns as are also defined by the map pattern of 1 and 2 on Figure 10.

D. Outcrop of interlayered biotite gneiss and granitic gneiss of the Crossing Knob Gneiss (Bartholomew and Lewis, in press). Outcrop is on west side of North Carolina state road 1202 in Watauga County within the Sherwood quadrangle. A Ramsey (1967) type-3 interference pattern is produced by refolding of an older, steeply inclined, axial surface (center) about younger, nearly horizontal axial surfaces.

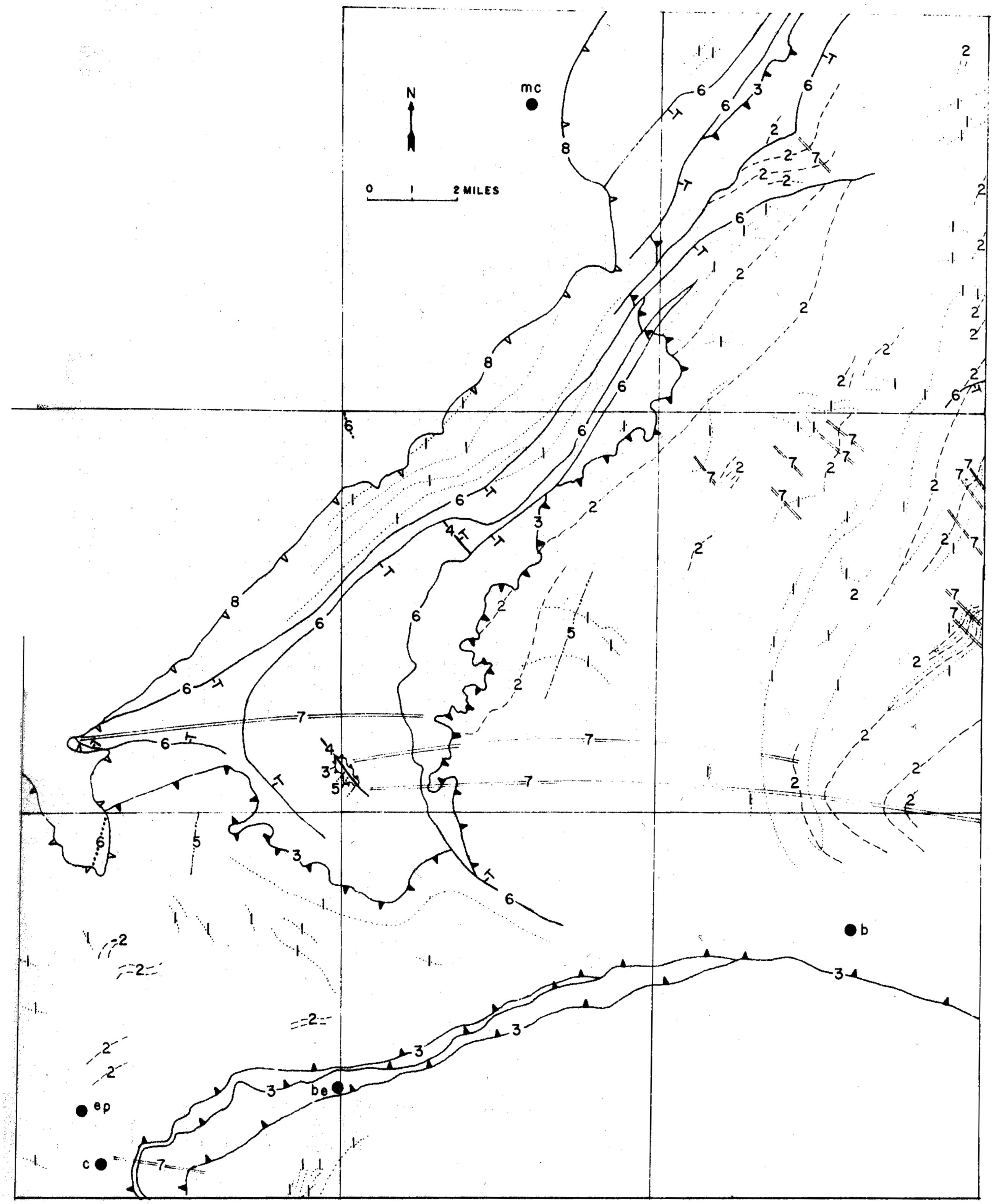


Figure 10: **Tectonic map showing sequential (1-oldest, 8-youngest) relationships among folds and thrusts in the area between the Grandfather Mountain and Mountain City windows; b – Boone, be – Banner Elk, c – Cranberry; ep – Elk Park; mc – Mountain City; quadrangles outlines and major tectonic features are the same as in Figure 1.**

Figure 11: Massive granitoid gneiss containing xenolith folded into a classic Ramsey (1967) type-2 interference pattern. Outcrop is along the northeast side of the North Carolina state road 1213 in Watauga County within the Sherwood quadrangle about 1.0 miles southeast of Georges Gap.

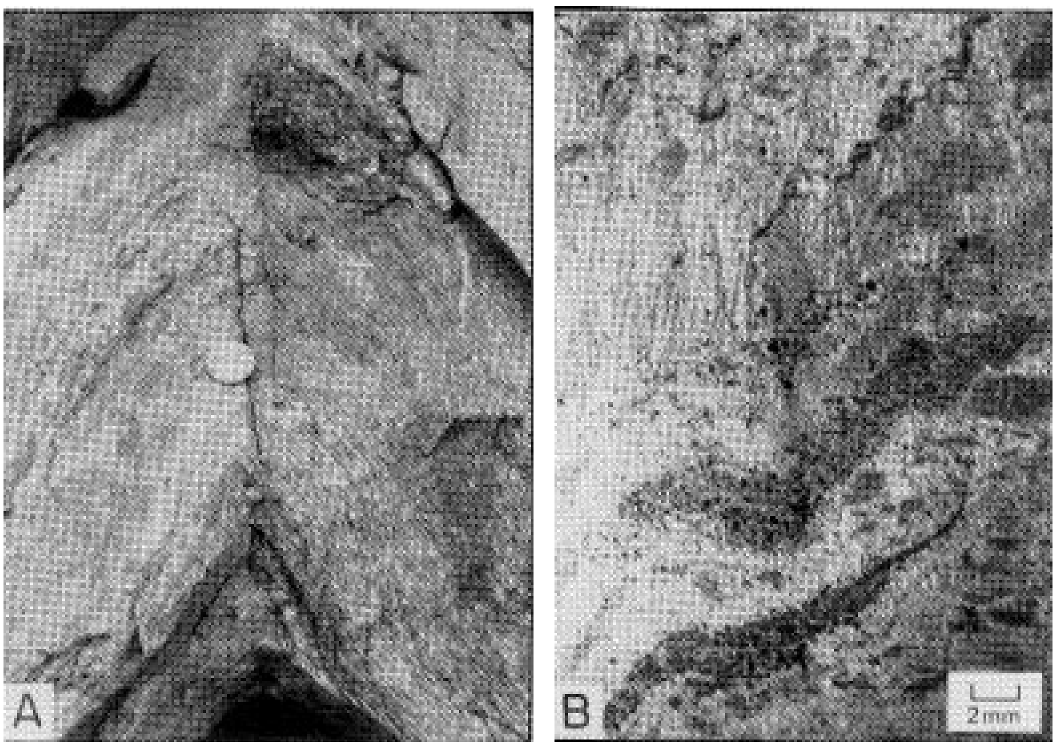


Figure 12: A. Fold with axial planar crenulation cleavage in mylonitic schist derived by ductile deformation of the Watauga River Gneiss (Bartholomew and Lewis, in press). Outcrop is along northside of North Carolina state road 1201 about 1.3 miles north ofthe bridge over the Watauga River.

B. Photomicrograph of mylonitic Watauga River Gneiss showing younger crenulation cleavage (vertical) cutting older mylonitic foliation (subhorizontal) metamorphic mineral assemblage consists of feldspar, microcrystalline quartz, sericite, chlorite, epidote, magnetite and sphene. Sample is from outcrop along northeastern flank of Big Ridge in Avery County within the northeastern corner of the Elk Park quadrangle.

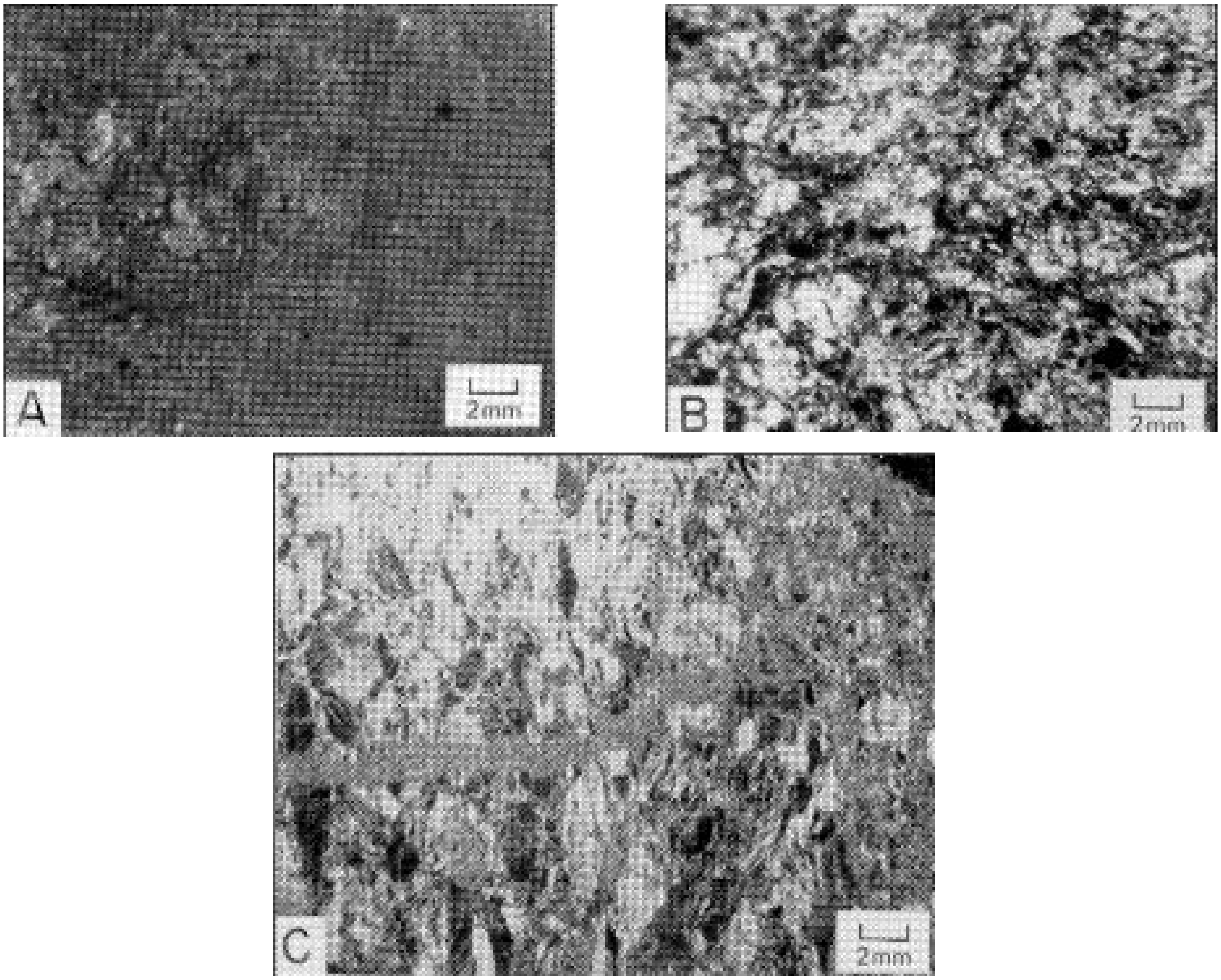


Figure 13: A. Photomicrograph of microbreccia from small microbreccia dike along north side of North Carolina state road 1306 at Silverstone in Watauga County within the Sherwood quadrangle.
B. Photomicrograph of microbreccia from large microbreccia dike about 0.3 miles northeast of the junction of North Carolina state roads 1305 and 1233 in Watauga County within the eastern portion of the Sherwood quadrangle.
C. Photomicrograph of microbreccia dikelet cutting older mylonitic foliation. Sample from outcrop of mylonitic gneiss (derived fom the Beech granite) along the northeastern flank of Dark Ridge about 0.3 miles east of the Tennessee-North Carolina border in Avery County within the Elk Park quadrangle. Sample is located about 1 mile east of the trace of a segment of the Iron Mountain fault system.

gneiss and schist (Higgins, 1971). Crenulation cleavage locally cuts this mylonite fabric (Figure 12); but folds associated with the crenulation cleavage appear to change trends across the Stone Mountain faults, suggesting, perhaps, synchronous or repeated development of thrusts, folds and cleavage.

Folds (6, Figure 10) in the Mountain City window preceded emplacement of the Blue Ridge thrust sheet along the Iron Mountain thrust (8, Figure 10).

Development of broad open folds (7, Figure 10) are associated with doming of the Grandfather Mountain window. These structures appear to be the youngest folds in this region and are inferred to have developed as folds associated with the structurally lower Iron Mountain thrust (8, Figure 10) or other deeper thrusts presumably related to Alleghanian deformation in the Valley and Ridge geologic province.

Microbreccia (Higgins, 1970) zones (Figure 13A, B) cut across the mylonitic foliation in the Sherwood quadrangle and are found associated with the Iron Mountain thrust in the Elk Park quadrangle (Figure 13C) suggesting that, like the Blue Ridge thrust further to the northeast (Bartholomew and others, 1982, Their Figure 43), the Iron Mountain Fault is characterized by brittle, not ductile, deformation.

DISCUSSION

The sequence of Paleozoic structural events (Table 2) developed for this region suggests the following larger-scale development of the Appalachians. First, regional Paleozoic metamorphism and deformation was followed (and perhaps terminated) by ductile deformation under greenschist facies metamorphic conditions associated with thrusting along the Fries-Linville Falls fault system. Second, younger thrusting along the Stone Mountain fault system which originated from a structurally lower position, is characterized by deformation more intermediate between the ductile and brittle realms. Third, primarily brittle deformation characterizes thrusting along the Blue Ridge fault system (Iron Mountain fault) which originated from a still lower structural position. Fourth, doming of the Mountain City window in the north suggests the presence of still younger thrusts at structurally lower levels.

From this summary of the Paleozoic deformational sequence the conclusion seems warranted that thrusting in the area near the Grandfather Mountain window generally progressed from structurally higher to structurally lower levels from mid- to late Paleozoic time. This conclusion agrees with conclusions concerning the same area reached by Harris and others (1981) and Boyer and Elliot (1982) using different lines of reasoning.

A similar evolution, from structurally higher to lower thrust sheets, was documented recently by Bartholomew and others (1982) for the Blue Ridge-Valley and Ridge transition near Roanoke, Virginia. This structural evolutionary model (east to west) is in marked contrast to the west to east model (from structurally lower to higher thrusts) advocated by Harris and Milici (1977) for the Valley and Ridge province. Bartholomew and others (1982) show that, overall, the main thrusts developed from east to west (Fries, Blue Ridge, Pulaski, St. Clair). Their data suggests, though, that once a major sheet is emplaced it may "break back" if the toe locks, thus producing a west to east internal sequence (St. Clair, Narrows, Saltville).

Furthermore, a progressive change occurs in the rheorological conditions of deformation from mid- to late Paleozoic time. Deformation progresses from ductile conditions at high-grade to low-grade metamorphic conditions, then continues as intermediate ductile-brittle deformation under nonmetamorphic conditions, and finally is followed by brittle deformation. These progressive changes suggest deformation of a gradually cooling crustal block as, and after, it was emplaced over a considerably cooler lower block and probably accompanied by considerable uplift and erosion. Observations by Harris and others (1981) and Boyer and Elliot (1982) are also consistent with this interpretation.

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Age	deformational event (Fig. 10)	Elk Park Massif	Grandfather Mtn. window	Crossing Knob window	Watauga Massif	Mountain City window
~ 300 my	9- NE-SW folds					doming of window by lower thrust NE-SW folds
	8-Iron Mtn. thrust	microbreccia				
	7- E-W folds (NW-SE bends)	7- E-W folds NW-SE folds	7-doming of window by lower thrusts E-W folds		E-W folds	
	6-Stone Mtn. faults	mylonites			mylonite cut by crenulation cleavage	NW-SE folds
	5- open folds	upright folds - crenulation cleavage		folded window (NE-SW folds)	upright folds- crenulation cleavage	
	4- NW-SE folds			doming of window by lower thrust	(older thrust; mylonite cut by crenulation cleavage)	
~ 350 my	3- ductile deformation; greenschist metamorphism	Fork Ridge-Linville Falls thrust	Linville Falls thrust	Crossing Knob thrust	Fork Ridge thrust	
~ 380 my	2-tight to isoclinal folds (amphibolite metamorphism)	tight to isoclinal folds; pegmatites				
~ 450 my	1- isoclinal greenschist to amphibolite metamorphism	isoclinal folds; amphibolite metamorphism	tight to isoclinal folds greenschist metamorphism		open to tight folds; greenschist metamorphism	
~ 750 1000 my	unconformity					
~ 1000 my	Grenville metamorphism	amphibolite metamorphism	amphibolite metamorphism	(amphibolite metamorphism)	(granulite metamorphism)	

Table 2: Chart showing relative ages of deformational events (Figure 1) in the area between the Grandfather Mountain City windows. () indicates a higher degree of uncertainty.

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RUBIDIUM-STRONTIUM AGES OF THE WATAUGA RIVER, CRANBERRY, AND CROSSING KNOB GNEISSES, NORTHWESTERN NORTH CAROLINA

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ABSTRACT

The Watauga River, Cranberry, and Crossing Knob Gneisses have Rb-Sr whole-rock ages of 1177 ± 29 m.y., 1018 ± 19 m.y., and 947 ± 57 m.y., respectively. Field relationships plus Sr isotopic compositions suggest that the Watauga River Gneiss was part of a plutonic suite intruded into older gneisses; the Cranberry and Crossing Knob Gneisses formed during the Grenville Orogeny by remobilization of older basement rocks.

INTRODUCTION

Relatively little information exists regarding the ages of the “basement” rocks in western North Carolina and adjacent states. Because of the general lack of detailed geologic mapping, most geochronological studies in this region have been reconnaissance in nature. One of us (M.J.B.) has mapped portions of Ashe, Avery, and Watauga counties in the Blue Ridge of North Carolina (e.g., Bartholomew, 1982; Bartholomew and Gryta, 1980; Bartholomew and Wilson, in press; Lewis and Bartholomew, in press). Based on these field investigations, we collected samples from three major lithologic units: Watauga River Gneiss, Cranberry Gneiss, and Crossing Knob Gneiss. Thirty-five samples from these units were selected for Rb-Sr whole-rock age determinations; these results are presented and discussed in this paper.

GEOLOGIC SETTING

In the northwestern portion of North Carolina, “basement” rocks beneath the late Precambrian unconformity have recently been subdivided into three groups by Bartholomew and Lewis (in press). According to them layered gneisses form the pre- (or early) Grenville-age country rocks into which Grenville-age plutonic suites were emplaced more or less concomitant with Grenville metamorphism (Figure 1). Following extensive erosion of this high-grade metamorphic terrane during the late Precambrian, the region was covered beneath a volcanic sequence and the Grenville basement was cut by numerous mafic dikes and plutons of peralkaline granite.

Following the principal Paleozoic metamorphic event, placed by Butler (1983) at about 450 m.y. ago, extensive thrusting during middle to late Paleozoic time produced the structurally complex terrane that is presently exposed. According to Bartholomew and Lewis (in press) the Fries-Fork Ridge-Linville Falls fault displaced the Elk Park Massif and its cover rocks (Ashe Formation) over rocks within the Grandfather Mountain Window as well as over rocks northwest of the present-day trace of the Fries-Fork Ridge fault zone (Figure 1). These latter rocks constitute the Watauga Massif and its cover rocks (Chilhowee Group and Mount Rogers Formation). The displacement on the Fries fault system in this region is estimated at about 50 km (Bartholomew, 1983) based on a palinspastic reconstruction of Paleozoic-age metamorphic zones. Thus, the Grenville-age Watauga (footwall) and Elk Park (hanging wall) Massifs were originally separated by about 50km with rocks of the Grandfather Mountain Window and the tiny Crossing Know Window (Figure 1) originally occurring in between.

This paper provides geochronological data on three units, each from a different Massif: 1) the Watauga River Gneiss (Bartholomew and Lewis, in press) of the Watauga Massif; 2) the Cranberry Gneiss (Keith, 1903) of the Elk Park Massif; and 3) the Crossing Knob Gneiss (Bartholomew and Lewis, in press) of the Crossing Knob Window

The Watauga River Gneiss, along with the associated Comers and Grayson Gneisses, forms part of the Forge Creek Suite (Bartholomew and Lewis, in press). This plutonic suite intruded older layered gneisses which are only poorly exposed in the Watauga Massif (Figure 1). The Watauga River Gneiss is typically a massive, coarse-grained, medium-gray to greenish-gray or pinkish-green granodioritic gneiss. It consists of 50 to 75 percent feldspar [microcline, plagioclase (An_{35}), ± perthitic feldspar], 10 percent large biotite flakes, and quartz. Xenoliths, small pegmatite and aplite dikes, and quartz veins are common. Within the Watauga River Gneiss are protomylonite zones and zones of

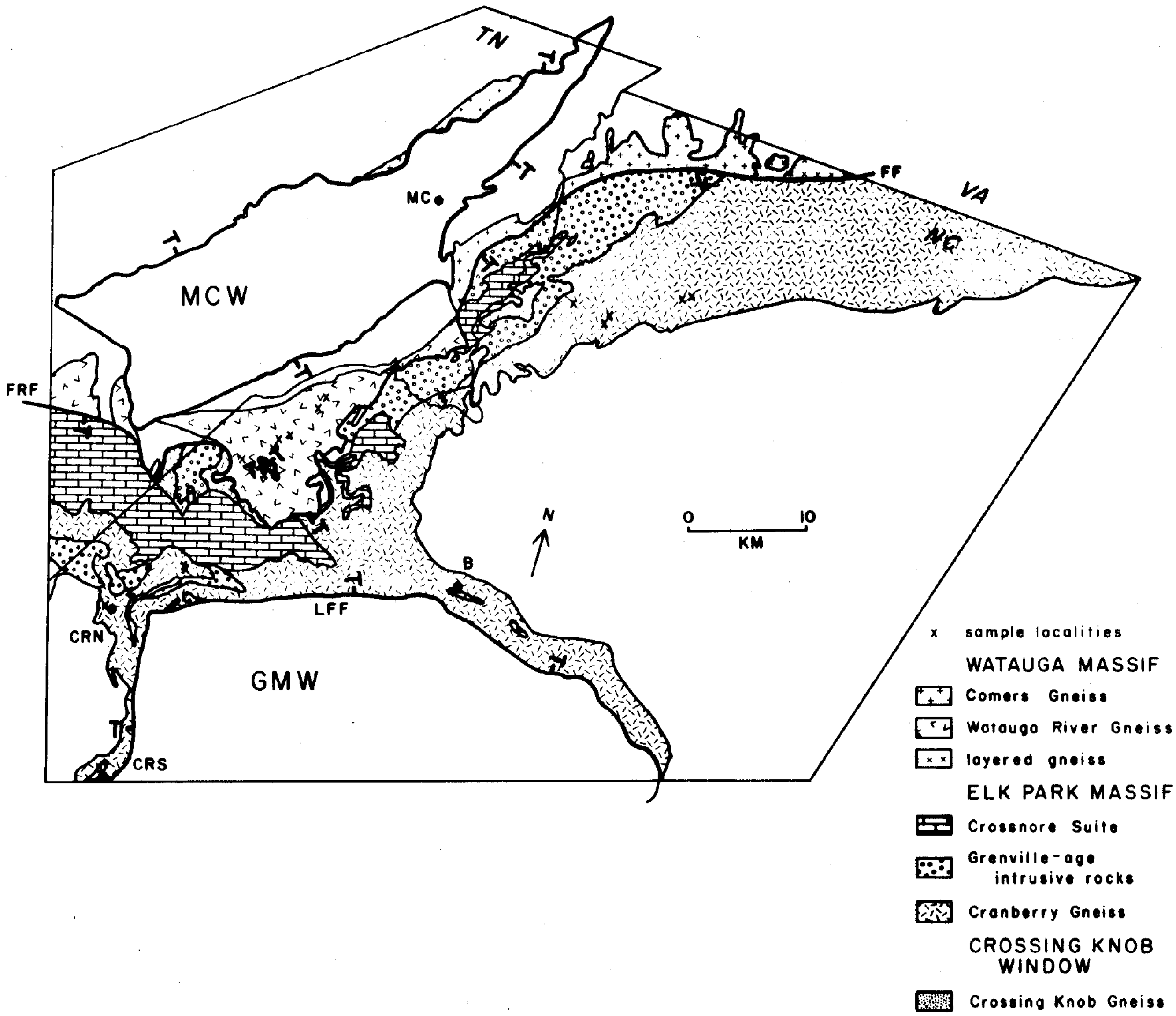

Figure 1: Generalized geologic map of basement rocks between Mountain City (MCW) and Grandfather (GMW) Windows (modified after Bartholomew and Lewis, in press). Sample locations (indicated by "X") also are shown on Table 3 (page 9). Town abbreviations: MC, Mountain City, TN; B, Boone, NC; CRS, Crossnore, NC; CRN, Cranberry, NC. Faults: FF, Fried fault; FRF, Fork Ridge Fault; LFF, Linville Falls fault.

mylonitic gneiss and schist.

The Cranberry Gneiss is a rather heterogeneous sequence of layered quartzofeldspathic and biotite-rich gneisses which forms the country rock of the Elk Park Massif. This rock unit is a well-layered, medium to dark gray, quartzofledspathic and biotite gneiss with alternating segregation bands ranging from a few cm to several tens of cm in thickness. Cranberry Gneiss is commonly coarse-grained and consists of feldspar (perthitic and potassic), quartz, biotite, and sphene. Layers of medium-grained, dark gray biotite schist are common. The Cranberry Gneiss is usually highly deformed and, near fault zones, mylonitic. Hornblende-rich phases are locally present, particularly near Cranberry. Plutons of granitic rock intruded the Cranberry Gneiss during Grenville orogenesis. Subsequently, numerous, large peralkaline plutons of the Crossnore Suite intruded the Elk Park Massif (Figure 1).

The Crossing Knob Gneiss is found only in an allochthonous tectonic slice exposed in a window beneath the Watauga Massif. Lithologically this gneiss is somewhat similar to the Cranberry Gneiss of the Elk Park Massif and the Wilson Creek Gneiss found within the Grandfather Mountain Window. The Crossing Knob Gneiss is a layered, medium-gray, coarse-grained biotite gneiss with scattered

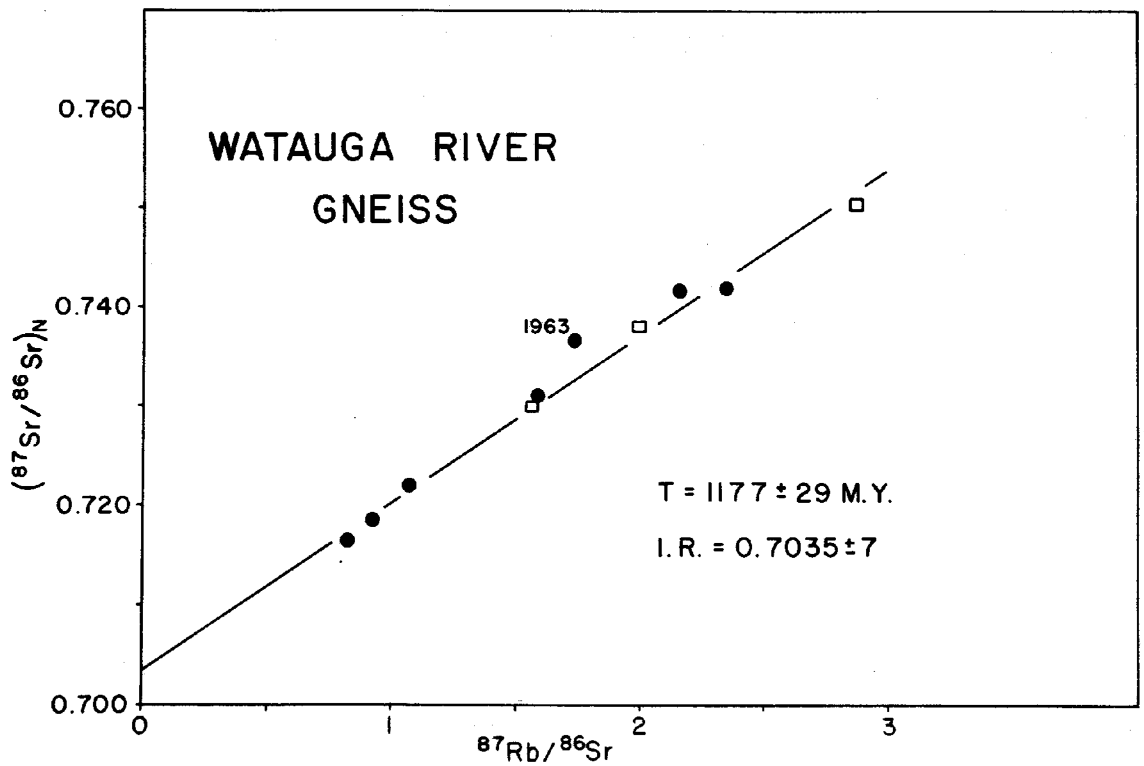

Figure 2: Plot of $^{87}Rb/^{86}Sr$ *vs* $(^{87}Sr/^{86}/Sr)_N$ for samples of Watauga River Gneiss. Solid dots represent analyses by Odom (1971); these were reported in Fullagar and Odom (1973). See text for discussion.

lenses (xenoliths?) of medium-grained, more mafic biotite gneiss. The layering is due to development of coarse-grained quartzofeldspathic segregations. The Crossing Knob Gneiss is commonly highly deformed and is mylonitic near faults.

PREVIOUS GEOCHRONOLOGICAL STUDIES

The oldest age reported for rocks of the North Carolina Blue Ridge is approximately 1300 m.y. for paragneiss near Deyton Bend, Yancey County (Davis and others, 1962). However, since this U-Pb age was obtained on detritial zircons, rocks 1300 m.y.-old need not be present in North Carolina.

Rb-Sr ages of approximately 1200 m.y. were reported for Cranberry Gneiss from central Ashe County (Odom, 1971; Fullagar and Odom, 1973), and for migmatite from the Mars Hill Quadrangle, Madison and Buncombe Counties (Fullagar and others, 1979). As is discussed in the next section of this paper, we now question the existence of Cranberry Gneiss with a Rb-Sr age as old as 1200 m.y. Granitic gneisses of this age are present in adjacent portions of South Carolina and Virginia (Fullagar and Odom, 1973; Fullagar and others, 1979).

Several studies indicate the wide-spread occurrence in the North Carolina Blue Ridge of gneisses with U-Pb zircon and Rb-Sr whole-rock ages of 1050 to 1000 m.y. (Davis and others, 1962; Fullagar and Odom, 1973). Rock units reported to have ages in this range include the Wilson Creek, Blowing Rock, and Cranberry Gneisses.

Peralklaine granites of the Crossnore Suite have U-Pb zircon ages of approximately 800 m.y. (Rankin and others, 1969) and Rb-Sr whole-rock ages of approximately 700 m.y. (Odom and Fullagar, in press). Based on analyses of several zircon fractions from each of the Crossnore, Beech, Brown Mountain, and Striped Rock (Virginia) plutons. Odom and Fullagar concluded that these plutons contain older xenocrystic zircons. The zircon ages reported by Rankin and others (1969) reflect the presence of these older zircons; the Rb-Sr ages give the times of crystallization of the plutons.

Other geochronological studies on samples from the North Carolina Blue Ridge are mainly analyses of minerals such as biotite and hornblende (see for example, Odom and Fullagar, 1973; Van Camp, 1982; Van Camp and Fullagar 1982).

NEW GEOCHRONOLOGICAL DATA

Thirty-five whole-rock samples were analyzed for Sr isotopic composition, and Rb and Sr concentration (Table 1). Analytical procedures are described in Harper and Fullagar (1981). Ages and initial Sr isotopic compositions $[(^{87}Sr/^{86}Sr)_o)]$ were calculated using the regression method of York

Table 1. Rb-Sr Data

Unit and Sample	$(^{87}Sr/^{86}Sr)_N$	Rb ppm	Sr ppm	$^{87}Rb/^{86}Sr$
Watauga River Gneiss				
1855	0.71836	138.2	432.6	0.926
1956	0.71641	133.1	467.1	0.826
1960	0.72200	154.0	415.8	1.073
1961	0.73083	125.3	239.2	1.519
1962	0.73425	288.8	139.9	5.987
1963	0.73667	198.9	331.9	1.739
1964	0.74159	150.7	202.4	2.162
1965	0.74197	172.7	213.7	2.346
Cranberry Gneiss				
1858	0.71332	36.26	302.8	0.347
1859	0.71581	64.16	315.9	0.588
1861	0.71396	39.55	300.9	0.381
1931	0.72892	159.0	285.0	1.618
1932	0.72748	221.1	472.0	1.364
1933	0.72189	154.8	397.8	1.128
1934	0.73108	122.8	226.7	1.571
1935	0.72944	156.1	291.1	1.555
1943	0.73162	160.0	266.0	1.744
1944	0.73017	159.5	264.7	1.747
1945A	0.73676	124.6	180.1	2.008
1945B	0.71802	108.6	470.0	0.669
1946	0.73640	114.6	173.0	1.922
1952B	0.72726	120.2	254.1	1.371
1953	0.72037	141.8	486.8	0.844
1954	0.71760	105.9	466.5	0.657
1955	0.74222	168.3	208.1	2.347
Cranberry Mine (Gneiss)				
1967	0.73082	91.89	137.2	1.942
1968	0.70929	20.64	160.6	0.372
1970	0.70876	4.02	303.5	0.038
1974	0.71815	13.39	57.29	0.677
1975	0.70944	10.12	95.83	0.305
1977	0.72978	21.37	19.14	3.238
79U-22*	0.70478	7.2	545.2	0.038
Crossing Knob Gneiss				
1848	0.74165	137.7	189.5	2.110
1850	0.74007	164.9	235.7	2.032
1851	0.73526	164.3	287.0	1.662
1852	0.74623	174.1	204.2	2.478
*Ussler (1980)				

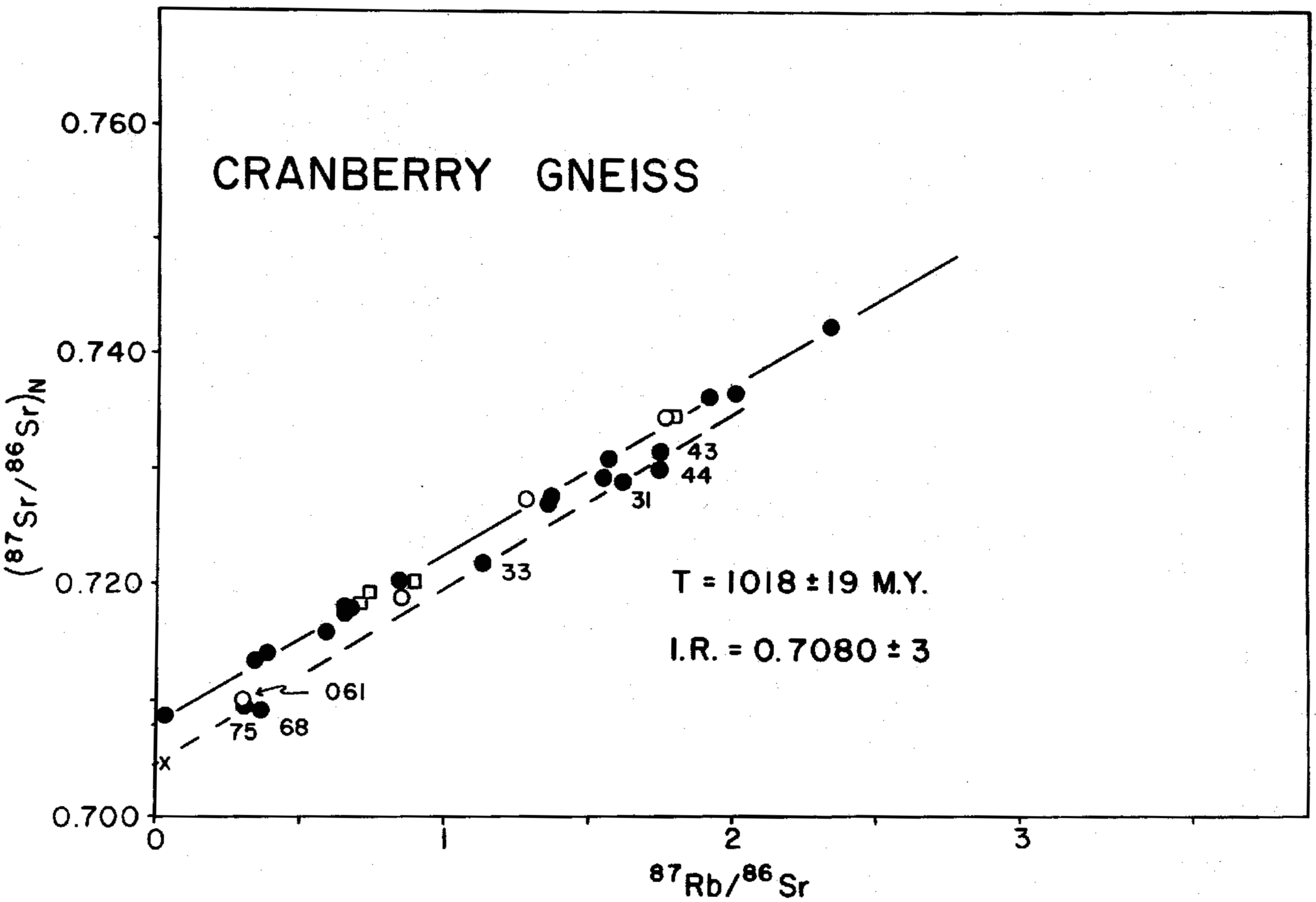

Figure 3: Plot of $^{87}Rb/^{86}Sr$ *vs* $(^{87}Sr/^{86}Sr)_N$ for samples of Cranberry Gneiss. Solid dots represent new analyses; data in Table 1. All sample members except 061 are prefixed by 19 (i.e., 75 is 1975). Open symbols represent analyses by Odom (1973). Symbol x represents a sample analyzed by Ussler (1980); data for this sample are in Table 1. The age and initial Sr ratio (I.R.) refer to theupper line on the diagram. See text for discussion.

(1969), and a 87Rb decay constant of 1.42×10^{-11} yr.$^{-1}$.

Eight samples of Watauga River Gneiss were analyzed and these results (except for 1962) are plotted in Figure 2. In addition, results for three samples analyzed by Odom (197a) and reported in Fullagar and Odom (1973) are included in Figure 2; these samples are Odom's 037, 038, and 039. These samples originally were considered to be Cranberry Gneiss, but now are considered to be Watauga River Gneiss.

Nine Watauga River Gneiss samples define an isochron (Figure 2) corresponding to an age of 1177 ± 29 m.y. (1 σ), and an $(^{87}Sr/^{86}Sr)0$ of 0.7035 ± 0.0007 (1σ). Most is not all of the scatter of the data points can be attributed to analytical uncertainty as the regression treatment yields an MSWD (mean square of weighted deviates) value of 2.6. If the value is about 2.5 or less, the scatter of data points is due to analytical uncertainties; if the MSWD value is significantly greater than 2.5 the data points scatter due to geological problems (rocks chemically altered, different lithologic units, etc.).

Plot of $^{87}Rb/^{86}Sr$ vs. $(^{87}Sr/^{86}Sr)$ N for samples of Watauga River Gneiss. Solid dots represent new analyses; data in Table 1. Open squares represent analyses by Odom (1971); these were reported in Fullagar and Odom (1973). See text for discussion.

Two samples of Watauga River Gneiss were excluded from the regression analysis: 1962 and 1963. These samples are from the same outcrop and both are very mylonitic. Presumably the chemistry of these samples was significantly altered during mylonitization. Sample 1962 contains much more Rb and less Sr than the other samples and thus it cannot be shown on Figure 2. If we assume that the age of sample 1962 was completely resent during mylonization we can calculate when this event occurred—provided we also assume a correct value for the initial $^{87}Sr/^{86}Sr$ of the mylonitic rock. If the ratio were 0.715, the age would be about 230 m.y.; if 0.710, an age of 280 m.y. is obtained. Unfortunately, the assumptions may not be valid, and this late Paleozoic model age of 280 or 230 m.y. may have no significance.

Cranberry Gneiss data are shown in Figure 3. Included

Table 2. Summary of Ages for North Carolina and Adjacent Areas.

Unit	Age, m.y./Method	$(^{87}Sr/^{86}Sr)_0$	Reference
Crossing Knob Gneiss, NC	947±57/Rb-Sr	0.7128±0.0017	This paper
Blowing Rock Gneiss, NC	1006±35/Rb-Sr	0.7077±0.0007	Fullagar and Odom, 1973
Blowing Rock Gneiss, NC	1050/U-Pb	-	Davis and others, 1962
Cranberry Gneiss, TN*	1050/U-Pb	-	Davis and others, 1962
Wilson Creek Gneiss, NC	1050/U-Pb	-	Davis and others, 1962
Cranberry Gneiss, NC	1018±19/Rb-Sr	0.7080±0.0003	This paper
Grayson Gneiss, V	1150±14/Rb-Sr	0.7044±0.0004	Fullagar and others, 1973
Watauga River Gneiss, NC	1177±29/Rb-Sr	0.7035±0.0007	This paper
Migmatite, Mars Hill Quad, NC	1183±65/Rb-Sr	0.7069±0.0009	Fullagar and others, 1979
Toxaway Gneiss, NC-SC	1203±54/Rb-Sr	0.7016±0.0021	Fullagar and others, 1979

*Correlation with type Cranberry Gneiss is uncertain.

are results for 23 new samples (Table 1), 8 samples analyzed by Odom (1971), and 1 sample analyzed by Ussler (1980) which is listed in Table 1. The samples analyzed by Odom (1971) and reported in Fullagar and Odom (1973) are shown as open squares and circles. The squares are Odom's samples 046, 047, 048, and 067. These Cranberry samples originally were included with 3 others now considered Watauga River Gneiss to have an age of approximately 1250 m.y.

Most of the data points in Figure 3, a total of 22, yield an isochron (upper line on figure) with these values: 1018 ± 19 m.y. (1σ); $(^{87}Sr/^{86}Sr)_0$ = 0.7080 ± 0.0003 (1σ); MSWD = 0.8. Several of the 8 data points that plot below this line represent some of he samples from the Cranberry Mine: amphibole gneiss 79U-22 (Ussler, 1980), 1968, and 1975. The formation of the ore deposit could have affected the Rb-Sr system of these samples. There is no obvious explanation as to why the other 5 samples (1931, 1933, 1943, 1944, and 061) plot significantly below the 1018 m.y. isochron. Except for 061, the other samples were collected within 7 km of the Cranberry Mine; perhaps they too were affected when magnetite deposits formed in the area. Or, proximity to mylonitic zones could have resulted in chemical alteration of these samples of Cranberry Gneiss.

The eight data points that plot below the 1018 ± 19 m.y. isochron do plot on or close to a line. This line is shown as a light-weight, dashed line in Figure 3. However, for this line to have significance, all of the 8 samples must have had the same initial $^{87}Sr/^{86}Sr$ ratio (about 0.7045) approximately 1018 m.y. ago. As noted above, we see no obvious reason for this to have been the case. Thus, we consider this distribution of these data points to be fortuitous.

Plot of $^{87}Rb/^{86}Sr$ vs. $(^{87}Sr/^{86}Sr)$ N for samples of Cranberry Gneiss. Solid dots represent new analyses; data in Table 1. All sample numbers except 061 are prefixed by 19 (i.e., 75 is 1975). Open symbols represent analyses by Odom (1971); these analyses were reported in Fullagar and Odom (1973). Symbol x represents a sample analyzed by Ussler (1980); data for this sample are in Table 1. The age and initial Sr ratio (I.R.) refer to the upper line on the diagram. See text for discussion.

Sample 061, from Ashe County (Odom, 1971) clearly is anomalous. Except for samples from the Cranberry Mine, it plots by itself (Figure 3). The other 3 samples from Ashe Country reported by Fullagar and Odom (1973) as having an age of about 1250 m.y., actually plot on or reasonably close to the 1018 m.y. isochron. Three new samples, 1858, 1959, and 1861 (analyses in Table 1), were collected about 4 k west of Odom's samples 062 and 064. The new samples also plot on the 1018 m.y. isochron. This means that in the rocks we sampled we have found no confirmation of an approximately 1250 m.y. age for typical samples of Cranberry Gneiss. More analyses do need to be done from the specific locations sampled by Odom.

Two additional samples from the Cranberry Mine were analyzed: 1967 and 1977. These samples are from pegmatites which cut the Cranberry Gneiss. Assuming an initial Sr ratio of 0.715, these samples have modal ages of 570 m.y. and 320 m.y., respectively. Ussler (1980) obtained similar results for additional pegmatite samples. The pegmatites probably are Paleozoic in age, but the uncertainty of the initial Sr ratio makes it impossible to be more specific.

Four samples were analyzed of the Crossing Knob Gneiss. These resulted in an isochron (Figure 4) which represents an age of 947 ± 57 m.y. (1σ), an initial Sr ratio of 0.7128 ± 00017 (1σ) and an MSWD value of 0.3.

Plot of $^{87}Rb/^{86}Sr$ vs. $(^{87}Sr/^{86}Sr)$N for samples of Crossing Knob Gneiss.

DISCUSSION

Table 2 summarizes our new ages and initial $^{87}Sr/^{86}Sr$ ratio plus previously published results. These results indicate two groups of ages: approximately 1200 to 1150 m.y., and

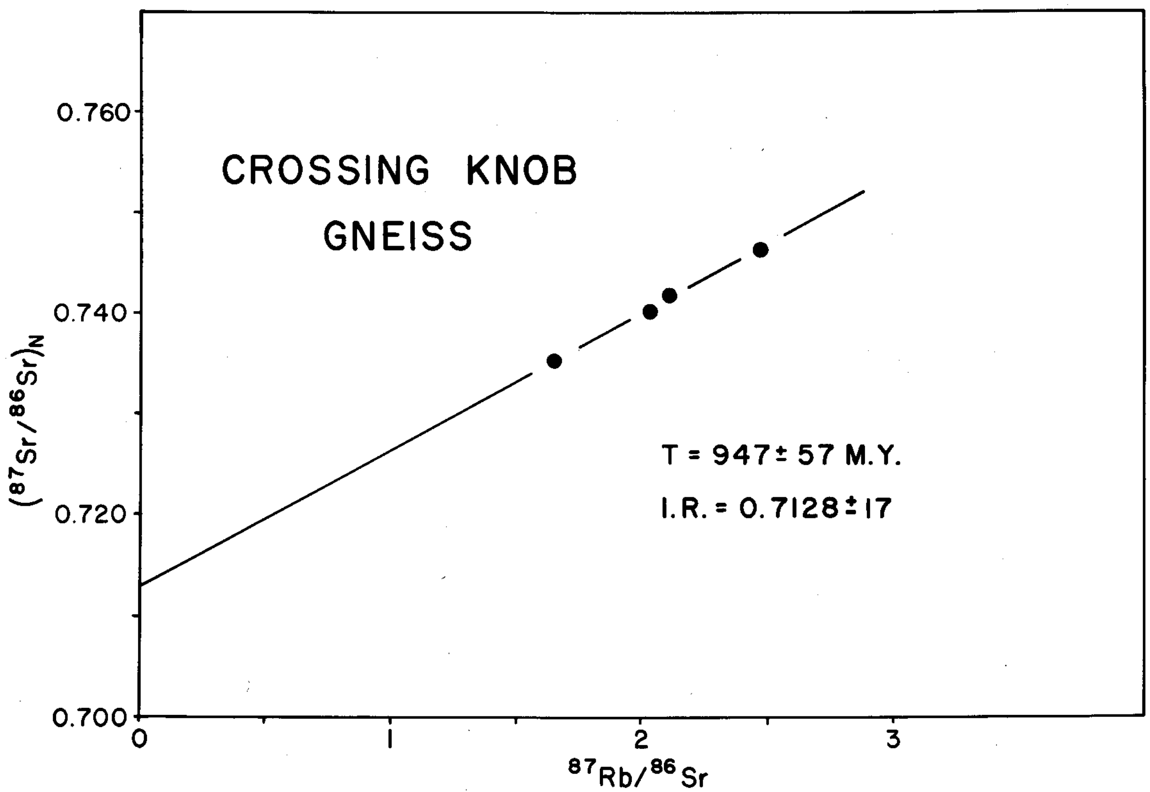

Figure 4. Plot of $^{87}Rb/^{86}Sr$ *vs* $(^{87}Sr/_{86}Sr)_N$ for samples of Crossing Knob Gneiss.

1050 to 950 m.y. Considering the uncertainty of the age for the Crossing Knob Gneiss, it probably is appropriate to state that the younger group has ages of about 1050 to 1000 m.y.

Most of the rocks belonging to the older group of gneisses have relatively low initial $^{87}Sr/^{86}Sr$ ratios. The initial ratio of 0.7035 for the Watauga River Gneiss is typical of values expected for magmas derived from the lower crust or upper mantle approximately 1200 m.y. old. Thus, the Sr isotopic data are consistent with the field observations that this rock unit is part of a plutonic suite that intruded older, poorly exposed layered gneisses. The Grayson Gneiss which is also part of this suite has a similar age of initial ratio (Table 2).

The Cranberry Gneiss and the Crossing Knob Gneiss are distinctly younger than the Watauga River Gneiss. The younger gneisses probably formed in response to Grenville amphibolite facies metamorphism approximately 1000 m.y. ago. Odom (1971) and Fullagar and Odom (1973) suggested that older basement rocks were reworked or remobilized during the Grenville orogeny; this would result in the younger reworked rocks having higher initial $^{87}Sr/^{86}Sr$ ratios than the original basement rocks. The results tabulated in Table 2 show that the younger rocks do have higher initial $^{87}Sr/^{86}Sr$ ratios than the older rocks.

ACKNOWLEDGMENTS

We thank Sharon E. Lewis for assistance with geologic mapping of the area and for helpful discussions regarding interpretation of field and age data. Financial support for field work and laboratory investigations was provided in part by the Geological Survey Section of the North Carolina, Department of Natural Resources and Community Development.

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Table 3. Location of samples used for Rb-Sr analyses.

Unit and sample	7.5” quadrangle	location	NC Coordinates	
Watauga River Gneiss				
1855	Sherwood	N.C. Road 1222	948,700N	1,161,200E
1856	“	“	“	
1960	“	N.C. Road 1201	932,700N	1,154,600E
1961	“	“	“	
1962	“	“	933,600N	1,155,300E
1963	“	“	“	
1964	“	“	934,700N	1,157,100E
1965	“	“	935,200N	1.157.300E
Cranberry Gneiss				
1858	Baldwin Gap	N.C. Road 1308	990,000N	1,217,900E
1859	“	“	“	
1861	“	“	“	
1931	Elk Park	N.C. Highway 194	895,800N	1,138,000E
1932	“	“	“	
1933	“	“	“	
1934	“	“	896,400N	1,137,400E
1935	“	“	“	
1943	“	N.C. Highway 194	886,500N	1,126,300E
1944	“	“	“	
1945A	“	“	885,800N	1,127,500E
1945B	“	“	“	
1946	“	“	“	
1952B	“	N.C. road 1312	918,700N	1,142,700E
1953	“	“	‘	
1954	“	“	“	
1955	“	“	“	
1967	“	Cranberry mine	882,900N	1,122,300E
1968	“	“	“	
1970	“	“	“	
1974	“	“	“	
1975	“	“	“	
1977	“	“	“	
Crossing Knob Gneiss				
1848	Sherwood	N.C. Road 1202	927,800N	1,152,900E
1850	“	“	“	
1851	“	“	926,100N	1,153,700E
1852	“	“	“	

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THE CRANBERRY MAGNETITE DEPOSIT, AVERY COUNTY, NORTH CAROLINA

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INTRODUCTION

The Cranberry Magnetite deposit of Avery County, North Carolina, is the largest of a regionally extensive set of discrete, lens-shaped magnetite concentrations that extend in a broad arc from southwest of Roan Mountain, Tennessee north and east through the town of Cranberry, North Carolina (see Figure 1). The entire district may, in fact, be more than 40 km in linear extent.

The Cranberry deposit is the only deposit to have been exploited to any significant degree, with perhaps as much as 1.5 million tons of magnetite ore extracted between 1882 and 1930 (Kline and Ballard, 1948). The deposit consists of a large number of discrete, independent ore masses ranging in thickness from a few centimeters to 60 m and as extensive as 275 m along strike. They lie concordantly in the plane of a strong cataclastic foliation which dips 30-40° southwest. The ore lenses are elongate in a roughly N80°E direction, roughly parallel to the axial trace of the Spruce Pine-Snake Mountain Synform. The deposit was classed by Bayley (1921) as nontitaniferous, in contrast to the titaniferous magnetite bodies associated with fine-grained mafic rocks in Avery and Mitchell Counties, North Carolina.

In addition to Keith's early studies in the region (Keith, 1903), Bayley (1921, 1922, 1923), Ross (1935), and Bryant and Reed (1970) briefly describe the Cranberry deposits. Kline and Ballard (1948) report on the U.S.B.M. directed wartime drilling program and magnetic surveys. Pack (1976), in an unpublished M.S. thesis at the University of Tennessee-Knoxville, presents petrographic descriptions and microprobe analyses of Cranberry minerals and the country rocks.

Various theories for the origin of the deposits have been presented over the years. Among the suggested mechanisms are 1)hydrothermal, post-magmatic fluids derived from the Bakersville gabbro (Keith, 1903), 2) a late, pegmatitic differentiate of Precambrian granites (Bayley, 1923), 3)replacement veins on faults (Ross, 1935), and 4) metasedimentary iron formation which was subjected to post-depositional partial melting (Pack, 1976). In this study, we propose that the Cranberry magnetite ores are products of an iron-rich immiscible melt from a Bakersville-type magma which experienced extreme plagioclase and possibly, pyroxene fractionation under extremely low fO_2 and fH_2O conditions. We will refer to this process as "late-stage magmatic differentiation," though we wish only to imply by the use of the term "late-stage" that the cranberry magnetites were derived following and interdeterminant amount of fractional crystallization. The term is *not* meant to imply a second-boiling or post-magmatic process.

GEOLOGICAL SETTING

The Cranberry-type deposits lie wholly within the cranberry gneiss of the Elk Park Plutonic Group, near the Grenvillian basement-Late Precambrian contact. 1.25 k southwest of the town of Cranberry is the largest body of the Bakersville Gabbro outside of the type-locality at Bakersville, North Carolina. An extensive body of the Beach Granite outcrops 5.0 km to the north.

Other papers in this volume will provide the details of the geologic and structural setting of the Cranberry deposits.

ORE GEOLOGY AND PETROLOGY

The ore is composed of either equigranular to foliated magnetite plus hedenbergite or pure magnetite veins as much as 30 cm in thickness. The equigranular to foliated magnetite-cliopyroxene rock is locally called "lean ore". The pure magnetite veins cross-cut pods of lean ore at high angles and appear to be more abundant near quartz –plagioclase-microcline pegmatites which also cross-cut the ore. Other silicate phases in the lean ore include plagioclase, amphiboles (commonly as an alteration product of clinopyroxene), epidote (commonly as an alteration product of the plagioclase), biotite, garnet, and quartz. Calcite is present, mainly as late cross-cutting veinlets.

The Cranberry ores were popular exploitable deposits in the eighteenth century because they were low in P_2O_5 (<0.04%), but petrographic studies show that apatite is present in amounts of 1-2% in a third of the samples studied. We can only surmise that the low P_2O_5 was a function of mine control or hand-cobbing of the ore. Rare ilmenite (as exsolutoin lamellae in magnetite), pyrite, pyrrhotite, sphalerite, and chalcopyrite are present. Molybdenum was reported from the Wilder Mine (Bayley, 1923). Hematite overgrowths on magnetite suggest the occurrence of a late-stage oxidation event. A probable paragentic sequence is shown in Figure 2. The "thermal event" indicated in Figure 2 was responsible for only small-scale mineralogical changes, and associated pure magnetite veins. The regional metamorphism succeeded by a fluid-rich epidote-amphibolite facies event.

In general, the ore bodies have a lens-shaped configuration, as shown in Figure 3. Table 1 lists the mineral composition of each of the units shown in Figure 3. In effect, the ore bodies appear to be large mega-boudins enclosed in a thin, mylonitic zone of the Cranberry Gneiss.

Efforts by one of us (W.U.III) to radiometrically deter-

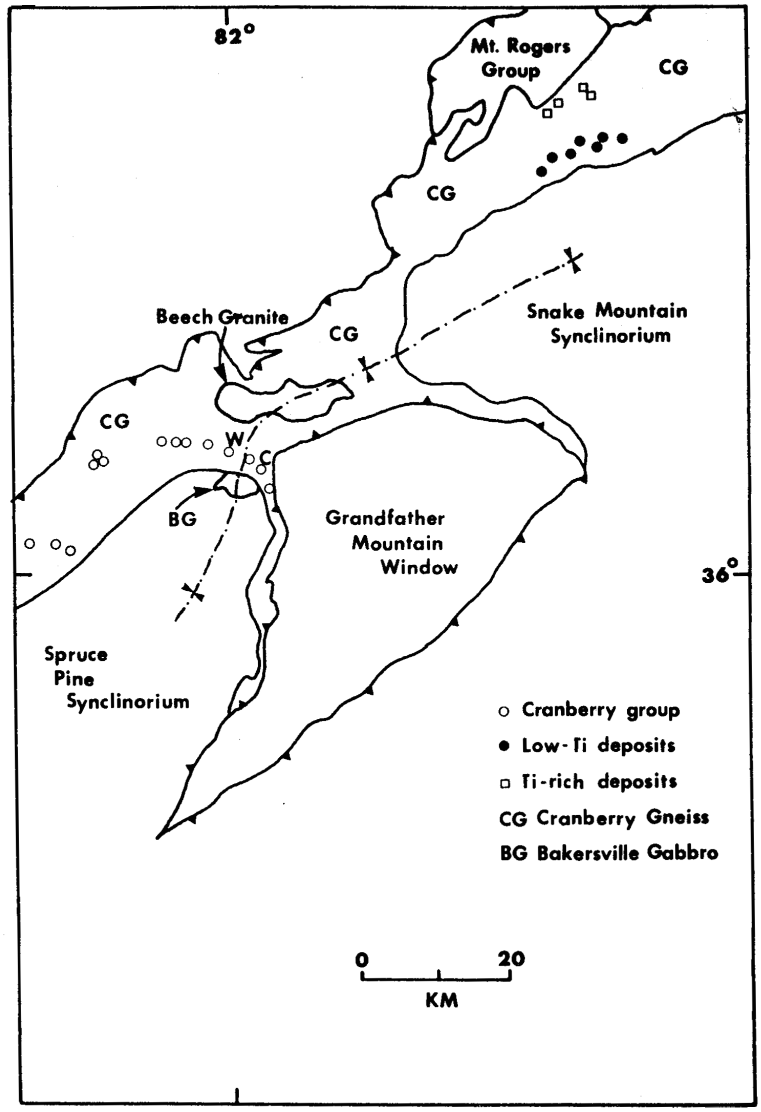

Figure 1: Generalized geological map of the northwestern Blue Ridge Province of North Carolina showing the location of the Cranberry magnetite and other magnetite deposits (from Ussler, 1980). "C" is the Cranberry Magnetite Deposit, "W" is the Wilder Mine.

mine the age of the cross-cutting pegmatites were a qualified failure. Two not-very-impressive Rb-Sr isochrons were obtained, based on the analysis of eight whole rock samples. These isochrons (T1 = 579 ± 51 m.y. and T2 = 524 ± 145 m.y.) strongly suggest that extensive mixing of pegmatite and wall-rock (Cranberry Gneiss), as well as post-crystallization alteration of the pegmatites, has taken place. An estimated $^{87}Sr/^{86}Sr$ initial ration of 0.730, and a Sr content of 104 ppm have been calculated for the uncontaminated pegmatite melt phase using a two-component mixing model (see Figure 4). The high initial ration suggests that the pegmatite was derived by anatexis of a metamorphosed or old (highly evolved) source. The Cranberry Gneiss is the likely source for the pegmatites.

A pre-deformation origin for the Cranberry ore is inescapable, if only because the foliation within the ore is parallel to the foliation in the country rock. However, such a pre-metamorphic origin leaves a number of plausible explanations for the origin of the Cranberry magnetite deposits.

These include:

1. Contact metasomatic (skarn)
2. Sedimentary (possibly volcano-sedimentary)
3. Post-magmatic hydrothermal
4. Magmatic cumulate (Skaergaard-type)
5. Late stage magmatic differentiate (Kiruna-type)

The Cranberry Gneiss contains no recognized calc-silicate units which would provide a favorable host for a skarn-type deposit. In addition, the absence of a nearby intrusive to act as a source of mineralizing fluids and the lack of any local calc-silicate units or remnants of such units argues

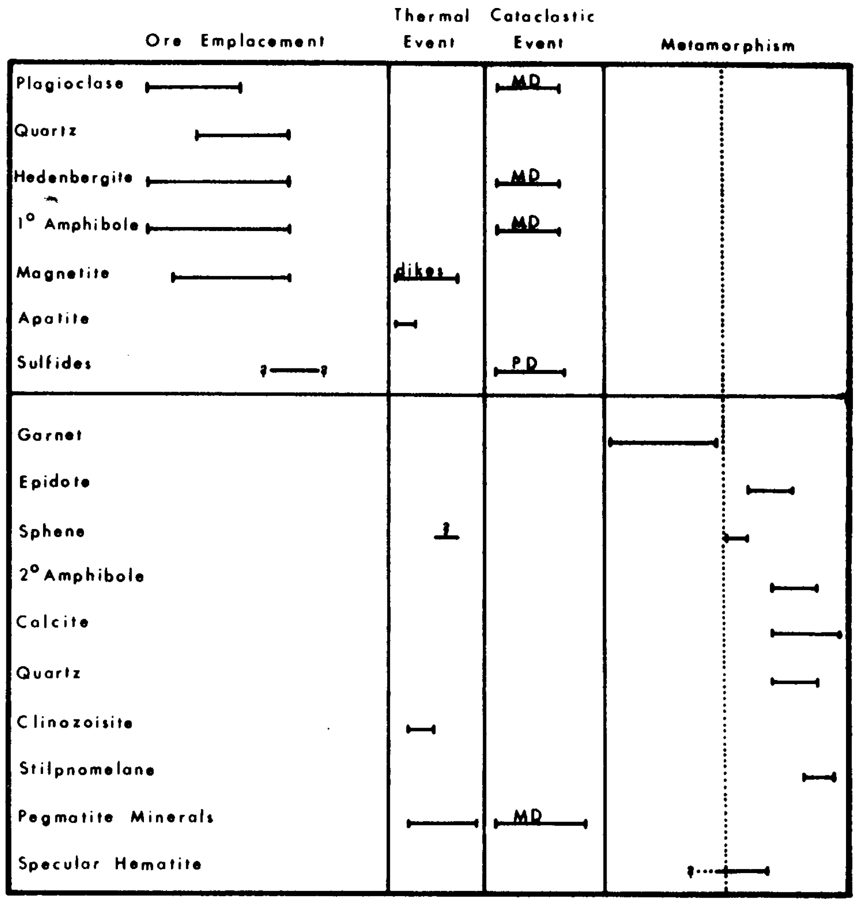

Figure 2: Paragenetic sequence of minerals in the Cranberry Mine. MD = mylonitic deformation, PD = plastic deformation. The dual metamorphic paragenesis distinguishes between an early, anhydrous garnet grad event and a latter fluid-rich epidote amphibolite facies metamorphism.

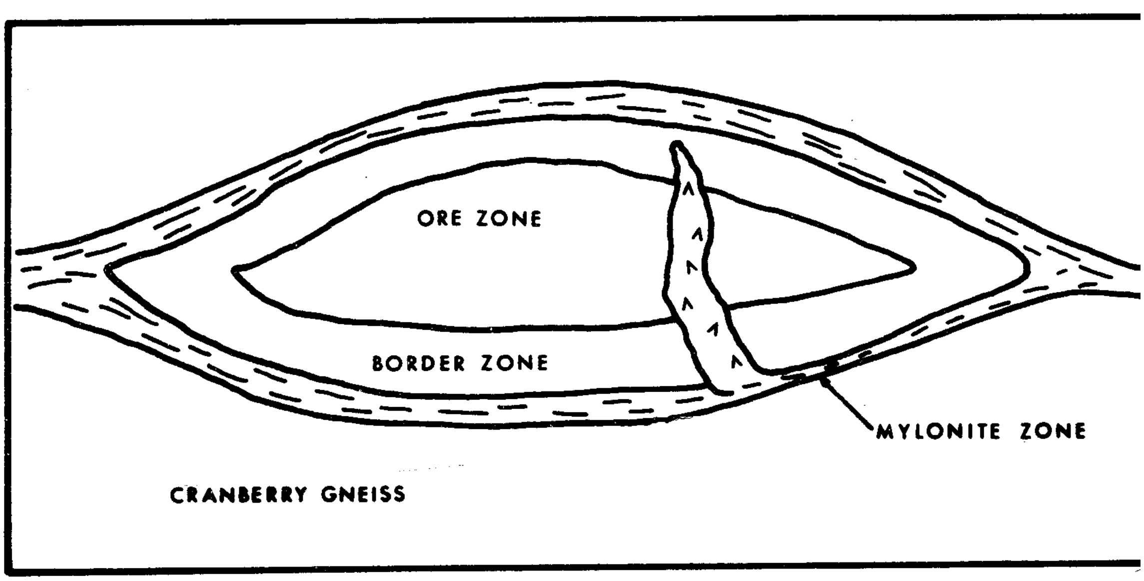

Figure 3: Idealized representation of a typical ore lens at the Cranberry Mine. The horizontal scale can vary from a few metersto >275 m and the vertical from centimeters to 60 m.

strongly against a contact metasomatic origin. The absence of any evidence for a pervasive wallrock alteration or for a nearby plutonic source and the generally sulfide-poor nature of the ore assemblage make s a post-magmatic hydrothermal origin seem unlikely

Table 1. Mineral assemblages of rock types shown in Figure 3(P= principal mineral phase) (M= minor mineral phase)

ORE ZONE:
P: Clinopyroxene, magnetite, quartz, apatite, sulfides
M: Amphibole, garnet, calcite, hematite
BORDER ZONE:
P: Clinopyroxene, hornblende, Ca-plagioclase, magnetite, sulfides
M: Epidote, calcite, quartz, sphalerite, stilpnomelane
MYLONITIC BORDER ZONE:
P: Plagioclase, microcline, biotite, quartz
M: Muscovite, epidote
PEGMATITES:
P: Microcline, Na-plagioclase, quartz
CRANBERRY GNEISS:
P: Na-plagioclase, microcline, quartz, biotite, muscovite

ORE GEOCHEMISTRY

In order to distinguish between several possible genetic models for origin of Cranberry magnetite ore (magmatic cumulate, late-stage magmatic differentiate and sedimentary), two geochemical techniques were utilized. They are a whole rock rare earth element (REE) analysis, and a trace element study of the ore magnetites.

REE Characteristics of Cranberry Magnetite Ore

Because it is likely that the ore originated by a magmatic or volcano-sedimentary process, REE date should exhibit characteristics of one of these processes. Given the extensive literature on REE partitioning between various mineral phases under magmatic conditions and the somewhat less extensive, but no less significant analyses that have been carried out on sediments, one can infer that magnetite ores of such varied origins should have a characteristic REE signature. Specifically, and probably somewhat simplistically, any magnetite-rich rock formed in equilibrium with sea water which has not previously been affected by the removal of authigenic or biogenic phases, should have

A REE pattern resembling that of shale (piper, 1974) (see Figure 5). In contrast, magnetite-rich rocks which equilibrated with a magma should differ in their trace element signature, owing to the differences created by high temperature silicate liquid partitioning. This may be especially true if plagioclase was a liquidus phase, as the ore petrology suggests.

REE analyses were carried out by isotope dilution techniques in the Isotope Laboratory at the University of North Carolina at Chapel Hill. Techniques, rock and mineral analyses, and discussion of accuracy and precision are available in

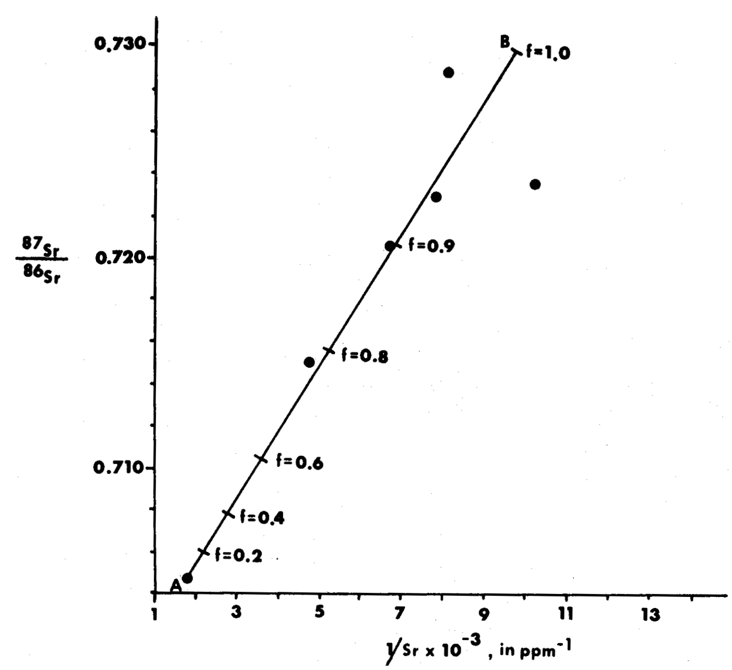

Figure 4: Two-component mixing model for the cross-cutting pegmatite veins in the Cranberry Mine, based on the model of Faure (1977). "f" is the degree of mixing between a hypothetical Cranberry Gneiss (end member B) and assimilated wall rocks (end member A). Filled circles are actual analyses of pegmatite samples from the Cranberry Deposit.

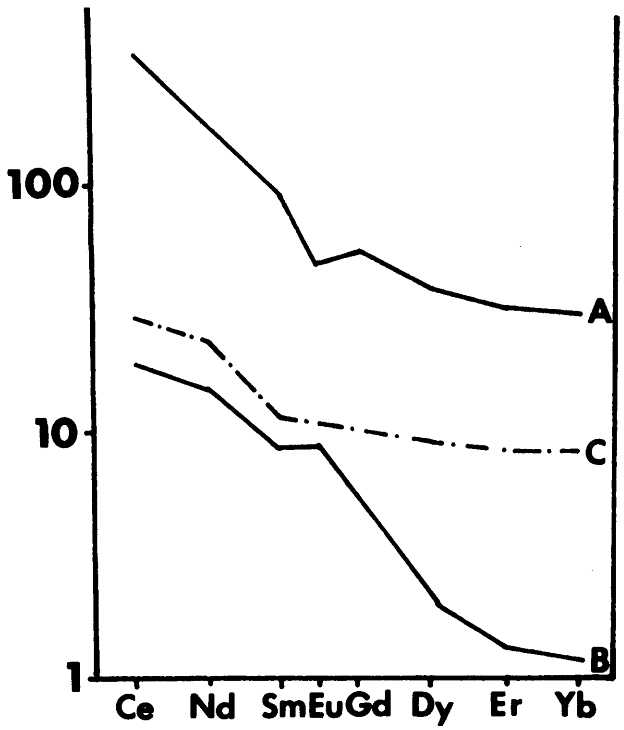

Figure 5: Examples of chondrite-normalized rare earth element patterns for a trachyte (A), a quartz diorite (B), and a Precambrian shale (C). Igneous analyses from Hansen (1978) and shale analysis fro Wildeman and Condie (1973).

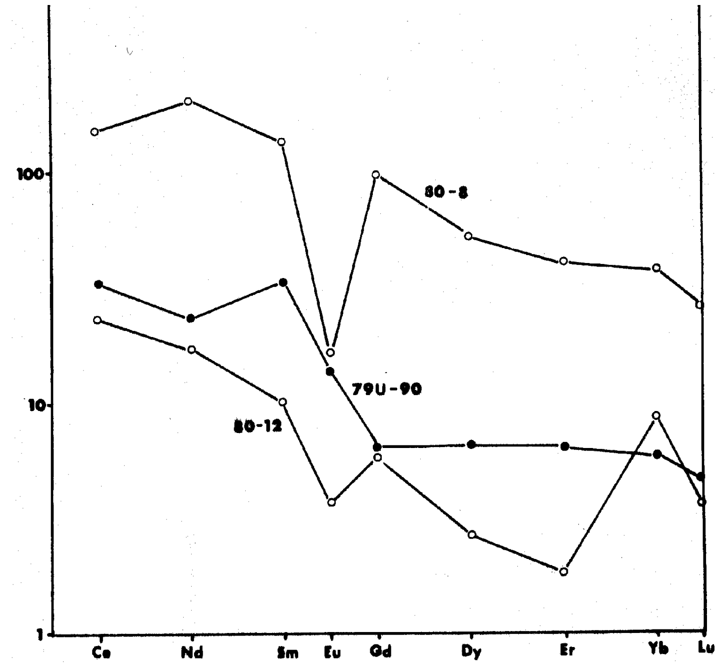

Figure 6: Chondrite-normalized rare earth element patterns for three whole rock samples from the Cranberry Mine. Open circles are CM80-8 and CM80-12, oxide-rich samples, and filled circles are for 79U-90, a garnet-amphibole rock presumed to be from the immediate hanging or footwall of the magnetite lenses.

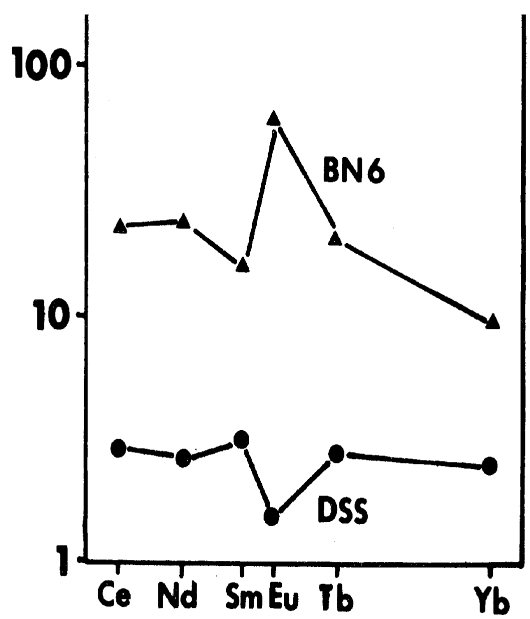

Figure 7: Chondrite-normalized rare earth element patterns for iron formations. DSS (circles) is from the Dellwood Seamount (Piper et al., 1975) and BN6 (triangles) is an oxide-rich iron formation from the Brunswick Number Six mine (Graf, 1977).

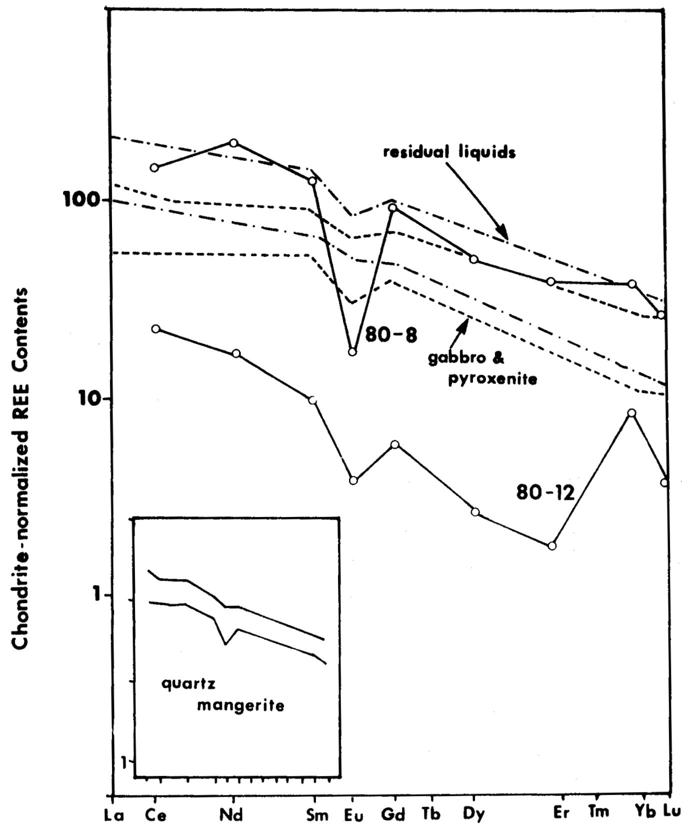

Figure 8: Chondrite-normalized rare earth element patterns for iron-rich residual liquids from the Bjernkear-Soguedal massif, Norway (dot-dash line from Roelandts and Duchesne, 1979) and oxide-rich gabbro and pyroxenite from the Adirondacks (dashed line from Ashwal and Seifert, 1980). Also shown are two of the Cranberry magnetite samples shown in Figure 6. The inset shows the relative rare earth element content of quartz mangerites from Duchesne and Demaiffe (1978).

Ussler (1980) and can be obtained from the authors on request. In general, errors were less than 4% RSD, except below the 1 ppm range, where errors may be as high as 7% for the more difficult to resolve elements, such as Lu. Chondrite normalized data for Cranberry ore (CM80-8) are shown in Figure 6. The sample is a slightly foliated magnetite-clinopyroxene rock ("lean ore"), and was taken from the northeast wall of the 3200 level, about 0.1 km east of the Fortner Tunnel.

Figure 7 shows the REE patterns for an iron-rich sediment from the Dellwood seamount (Piper at al., 1975) and for two iron formations, one chlorite-rich and one oxide-rich, from the Brunswick Number Six (Graf, 1977). These samples are depleted in total REE contents, and are generally light REE depleted relative to Cranberry magnetite ore (Figure 6). The negative Eu anomaly of the Dellwood Seamount pattern is characteristic of some submarine altered volcanic rocks in that plagioclase, which preferentially partitions Eu relative to other common rock-forming minerals in volcanics, is the most readily altered phase. Plagioclase-destructive alteration, subsequently, releases Eu to the fluid phase, resulting in a negative Eu anomaly in the residual phases. The positive Eu anomaly observed by Graf (1977) may indicate that these iron formations are the product of transport by a fluid phase that had equilibrated with submarine volcanics and preferentially extracted Eu during plagioclase-destructive alteration. In any case, the Cranberry REE pattern is distinctly different when compared to those of volcano-sedimentary iron deposits. The Cranberry ore is LREE-enriched with a large negative Eu anomaly and, in general, is enriched 100 times chondrites.

REE data for Cranberry magnetite ore are similar to those observed in some igneous rocks (Hanson, 1978; Nagasawa, 1970), and mamatic iron ore deposits (Roelandts and Duchesne, 1976). Fractional crystallization of a continental tholeiitic magma under conditions of low oxygen fugacity can result in production of iron-rich residual liquids (Osborne, 1962). Iron-rich residual liquids have been inferred to produce ferrogabbros and magnetite ore associated, for example, with massif-type anorthosite (Roelandts and Duchesne, 1979; Ashwal and Seifert, 1980). Other models envision fractional crystallization of plagioclase to produce anorthosite, felsic rocks such as quartz mangerite, and an iron-rich residual liquid (Duchesne and Demaiffe, 1978). Figure 8 shows REE patterns for ferrogabbro, estimated liquids which produced iron-rich ore, oxide-rich gabbros, and quartz mangerites. It is noteworthy that the REE patterns from these rocks are strikingly similar to the REE pattern of Cranberry ore. This analogy is not necessarily meant to imply a genetic relationship between Cranberry ore and anorthosite, but rather to show possible examples where fractional crystallization may have generated iron-rich residual liquids.

REE data for Cranberry ore differ from those of a volcano-sedimentary environment, implying that this environment was not a source for the ore. It would appear more likely that an Fe- and LREE-enriched residual liquid, derived during fractional crystallization of plagioclase and clinopyroxene under low fo2 conditions, was the parent for the clinopyroxene-magnetite rock at Cranberry. REE data for residual liquids (Figure 8) exhibit small La/Sm ratios, and large total REE abundances. Such feature are consistent with the presence of an initially LREE-enriched parental magma which experienced extensive plagioclase fractionation. Cranberry ore and residual liquid data are similar to REE data from apatite mineral separates. Apatite exhibits very large REE distribution coefficient values, and it will therefore be a sink for the prevailing REE composition of the coexisting Ferich residual liquid. Variability in the REE content of the two oxide-rich Cranberry ore samples correlates well with the modal apatite of these samples.

Trace Element Characteristics of Cranberry Magnetites:

Trace element behavior in natural systems may be described by the partitioning between the mineral(s) under consideration and the remainder of the system with which the mineral is in equilibrium. In this study, magnetite separates were analyzed and, thus, the partitioning can be described by a bulk distribution coefficient K_D:

$$KD = \frac{C_{(X)mag}}{C_{(X)sys}}$$

Where $C_{(X)mag}$ = the concentration of component X in magnetite, and

$C_{(X)sys}$ = the concentration of component X in the system

The term "system" refers to the remainder of the material with which the crystallizing magnetite is in thermodynamic equilibrium. Typically, for a liquidus phase, this would be the silicate melt and co-crystallizing minerals, but it could also be sea water and any authigenic minerals for a magnetite crystallizing in a sedimentary environment. When K_D values for a given element are greater than 1, magnetite is a net concentrator of that element with respect to the remainder of the system with which it is in equilibrium.

In general, magnetite K_D values are >>1 for most transition metals. Specifically, Ti^{+4}, V^{+3}, Cr^{+3}, and Sn^{+3} substitute readily for the ferric iron in the magnetite structure. Mg^{+2}, Mn^{+2}, and Ni^{+2} substitute for Fe^{+2}. First order transition metals, such as Cu^{+2} and Zn^{+2}, enter the magnetite structure in much more limited amounts. It is not the goal of this study to evaluate the K_D's in any systematic sense, but rather to observe the trace element content of the magnetites in hopes that, by analogy with other deposits, some restrictions can be placed on the environment of deposition. The working hypothesis is that any magnetite which crystallized in thermodynamic equilibrium with a low temperature-low pressure aqueous brine (a volcano-sedimentary environment), a high temperature mafic to ultramafic magma (Skaergaard analogy), or a lower temperature, more evolved mafic differentiate (Kiruna) would show different trace element contents due to the highly varied K_D's of those systems as well as the different trace element contents due to the highly varied K_D's of those systems as well as the different bulk metal contents. We wish to emphasize that this is a comparative study and that any conclusions which we draw are only as good as the data base of the analog systems.

Magnetite was separated with a Frantz magnetic separator. The magnetic fraction was hand-picked to insure than silicate fragments were not included. All samples were analyzed for $A1_2O_3$ and SiO_2 as a check against potential silicate contamination of the magnetite separate. All chemical analyses were carried out in the geochemistry laboratory at the University of North Carolina. SiO_2 analyses were determined by colorimetric methods. $A1_2O_3$, TiO_2, MgO, MnO, V, Cr, and Zn analyses were determined by flame atomic absorption spectrophotometry on a Perkin-Elmer Model 603. Cu and Ni were analyzed on the same instrument with a HGA-2200 graphite furnace. All analytical curves were established with rock standards. Analytical data, standard curves, and details of the precision and accuracy of analyses are available upon request from the authors.

Whole rock analyses were not attempted in this study, for several reasons. The degree of alteration of the silicate phases, uralitization and epidotization in particular, made fresh samples difficult to obtain on a sufficiently large scale to insure minimal sampling error. Furthermore, the most useful data to be obtained from such as whole rock study in conjunction with the magnetite analyses would be bulk K_D's between magnetite and the system in which it was in equilibrium. There is no assurance in our minds that the magnetite is still in physical contact with unaltered portions of the system from which it crystallized. This separation could have been the result of either tectonic remobilization, or mobility during genesis. We view the relatively unaltered, crystalline magnetite as a reasonable refractory phase capable of preserving original trace element compositions.

Table 2 gives the results of this analytical work. For each element, the number of samples above detection limits, the mean, the range and the detection limit is given. Table 3 gives a two variable correlation matrix for those elements for which five or more analyses are available. The reason for this limitation of N > 5 is the statistical uncertainty for such small sample populations (at N = 5, a correlation of 0.86 is necessary at the 95% confidence level). As it is, with a maximum N = 11, a correlation coefficient of >0.760 is necessary at the 99% confidence level.

Table 2. Trace Element Analyses of Cranberry Magnetites

Element	# of anal.	Mean	Range	Detection limit
SiO_2	11	0.91%	0.265–1.47%	
$A1_2O_3$	0	n.a.	n.a.	
TiO_2	8	0.40%	BDL-1.18%	
MnO	11	0.08%	0.060-0.146%	
V	1	150 ppm	BDL-150 ppm	50 ppm
Mg	5	82 ppm	BDL-488 ppm	
Cr	0	n.a.	n.a.	10 ppm
Cu	9	40 ppm	BDL-179 ppm	
Zn	11	50 ppm	32–105 ppm	
Ni	5	4 ppm	BDL–9 ppm	1 ppm

The absence of any detectable $A1_2O_3$ in the magnetite separates strongly suggests that whatever silicate contamination is present is not due to plagioclase inclusions in the magnetite. The negative correlation between SiO_2 and MgO also suggests that clinopyroxene inclusions are not present in significant amounts. The most significant correlations are, predictably, among the transition elements.

In general, the Cranberry magnetites appear to be characterized by relatively low concentrations of trace elements. Interestingly, magnetites separated from the Bakersville Gabbro are more enriched in trace elements relative to Cranberry samples. Table 4 shows these tow analyses of magnetite from the Bakersville Gabbro.

The data can be used to suggest that the Cranberry magnetites cannot be modeled as being a cumulate from the Bakersville Gabbro or a Bakersville-type magma. Parental liquids which crystallized Cranberry-type magnetite must contain relatively smaller transition element abundances. This conclusion is consistent with the results of magnetite analyses from other magmatic segregations, such as those from the Skaergaard intrusion. Figure 9 shows a comparison of the Mg, Ti, V, Cr, and Mn content of magnetites fro Cranberry and from cumulates elsewhere in the world. Because a mafic to ultramafic magma of this type would be relatively enriched in sideophile elements and because the magnetite/magma bulk K_D's will be greater than one for these elements, both the theoretical and empirical evidence for the trace element distributions imply that the Cranberry magnetites are not a magmatic cumulate phase.

A literature search for analysed magnetites from ore deposits in other parts of the world has yielded a fairly limited set of data for comparison with the Cranberry suite. Table 5 lists the types of deposits, the regions, the number of analyses in the literature and the source. In general, a statistically significant suite of analyses is available only for Ti, V, Cr, Cu, Mg, and Mn.

In addition to these published analyses, samples of magnetite from a magnetite granite in the midwestern U.S., fro a magnetite cumulate in diabase from Arizona, and fro Sanford Lake, New York were analysed for comparative purposes.

Table 3. Correlation Matrix for Cranberry Magnetites [N ≥ 5]

	Fe_3O_4	TiO_2	MnO	SiO_2	MgO	Cu	Zn	Ni
Fe_3O_4	1.000	-0.393	0.410	-0.823	0.583	0.110	-0.041	0.182
TiO_2		1.000	-0.259	0.516	-0.395	-0.170	-0.122	-0.555
MnO			1.000	-0.547	0.878	0.745	0.809	-0.760
SiO_2				1.000	-0.831	-0.517	-0.117	-0.383
MgO					1.000	0.970	0.829	0.000
Cu						1.000	0.782	0.084
Zn							1.000	-0.540
Ni								1.000

Table 4. Magnetite Analyses — Bakersville Gabbro [TiO2 and MnO in per cent, all others in ppm]

	TiO_2	MnO	Mg	Cu	Zn	V	Cr	Ni
T-NC-7	BDL	0.042	996	>2000	497	<50	14	>3000
LCBG	1.833	1.021	>1000	34	924	~2500	377	-

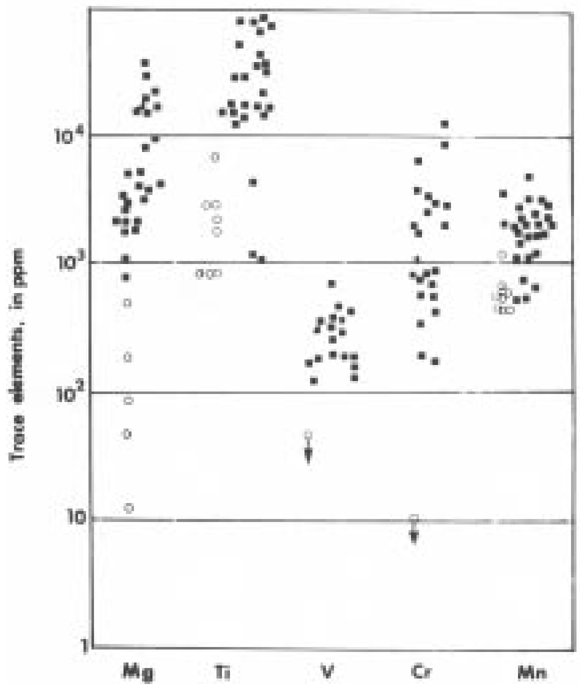

Figure 9: Selected trace elements of Cranberry magnetites (open circles) and typical magnetites of cumulate origin (filled squares) elsewhere in the world. Source of the data given in Table 5.

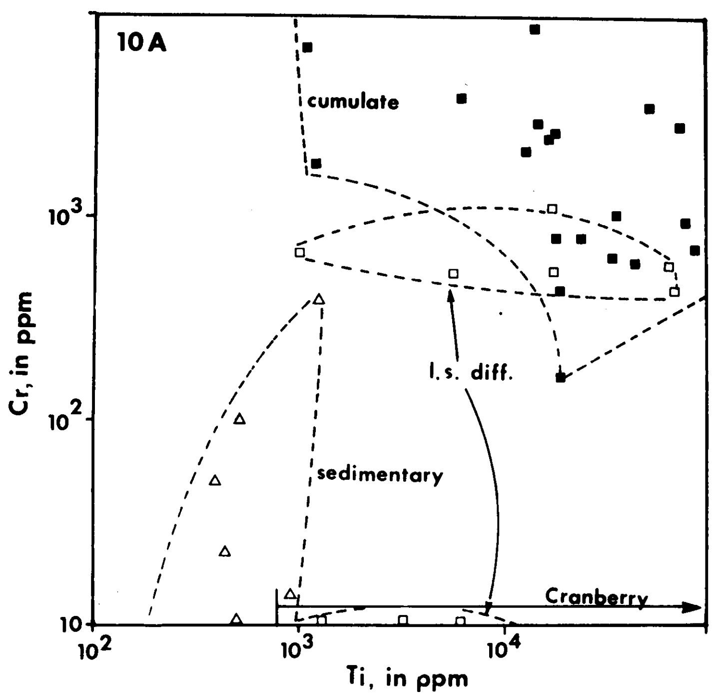

Figure 10A: Cr-Ti plot of Cranberry magnetites (solid arrow) compared to sedimentary (open triangles), cumulate (filled squares), and late stage differentiate (open squares) magnetites. Fields are hand-drawn on analyses and do *not* represent a statistical discrimination of the classes of deposits. Sources of data given in Table 5.

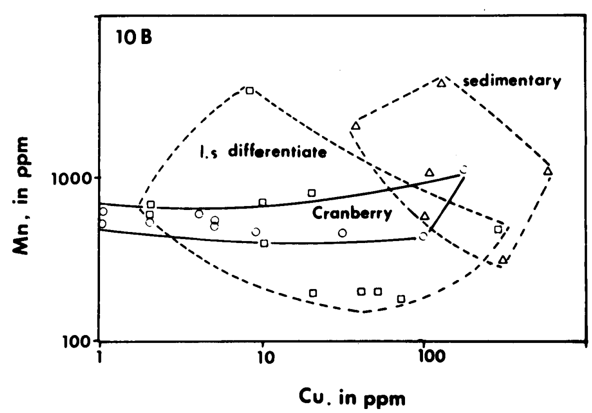

Figure 10B: Mn-Cu plot of Cranberry magnetites compared to late stage differentiates and sedimentary magnetites. Symbols and sources of data the same as Figure 10A.

Applications of a discriminant analysis to data from the suites in Table 5, as well as data from skarn magnetites, yields two discriminant functions which each contain >96% of the variance in the groups. The first function is heavily weighted towardV, Ti, Cr, and Mn and the second toward Cu and Mn. Results of the discriminant analysis indicate that 80% of all cases are correctly classified. The average Cranberry magnetite (see Table 2) is classified as a late-stage magmatic differentiate.

Table 5. Sources of Magnetite Analyses for Comparison with Cranberry [If N = A, only an average is reported]

	# of analyses	Literature source
Sedimentary Magnetite		
Worldwide	A	
Sierra Leone	A	Frietsch, 1970
Australia	A	Frietsch, 1970
Sweden	A	Frietsch, 1970
Korea	6	So, 1970
Magmatic Segregation		
Norway	10	Thy, 1982
Eastern Canada	5	Liset, 1966
Korea	1	So, 1970
Russia	9	Borisenko, et al., 1968
Skaergaard, Greenland	5	Wager and Mitchell, 1951
Late Differentiate		
St. François Mtns., MO	3	Kisvarsanyi and Proctor 1967
Sweden	14	Frietsch, 1970
Finland	4	Frietsch, 1970
Eastern Canada	5	Lister, 1966

Only one Cranberry sample contained V in excess of the I ppm detection limit, further distinguishing them from sedimentary and magmatic segregation-type magnetites. Data from these suites are presented in the form of covariation diagrams to demonstrate the distinction between three types of magnetite deposits, considered as possible analogs of the Cranberry deposits. Figure 10 shows Ti vs. Cr and Cu vs. Mn. The fields on this figure were derived from the published analyses referred to in Table 5. It is noted that in some cases the points plotted were averages of many analyses, while in other cases the raw data was used, producing a greater spread in the relative fields.

Figure 10a shows that the Cranberry magnetites are significantly Cr-depleted relative to magmatic cumulates, and Ti-enriched relative to sedimentary magnetites. Cranberry magnetites are similar to low Cr, late-stage differentiate magnetites such as those from Pea Ridge, Iron Mountain, and Bourbon, Missouri.

Figure 10b shows the co-variation of Cu and Mn for Cranberry magnetites, and also for data of sedimentary and the late-stage magmatic differentiate magnetites of our reference suite. The magmatic cumulate group is not shown in order to simplify the diagram because we feel that the Cr content on Cranberry magnetites is sufficient to distinguish between various genetic types. It is noted that the field of magmatic segregation magnetites would plot on the high Mn and Cu side of the Cranberry analyses. In general, sedimentary magnetites are also more Cu- and Mn-rich than those derived by a late-stage magmatic differentiation process. The Cranberry magnetites in terms of Cu and Mn content.

DISCUSSION

The rare earth and trace element characteristics of the Cranberry magnetite deposit would seem to suggest a late-stage magmatic differentiation model for the origin of the magnetite concentrations. The regionally extensive, linear band of magnetite bodies would then be explained by either the tectonic modification of an essentially continuous dike of magnetite-rich rocks, or a series of discrete dike-like intrusives along a major regional linear trend. The boudin-like form of many of the deposits as well as the mylonitic haloes around many of the bodies suggests that tectonic processes may have modified the original form of the deposits. The Pea Ridge and Iron Mountain magnetite/hematite deposits in the Precambrian of the St. François Mountains in eastern Missouri have been modelled as iron-rich dikes and have similar forms and mineralogy (Murphy and Ohle, 1968; Emery, 1968).

It remains to determine whether a petrologic model can be developed for the generation of an iron-rich magma in an environment compatible with the regional setting of the Cranberry deposits. In our discussion of the REE data, a somewhat cautious analogy with anorthosite genesis was made. The classic association of magnetite-ilmenite concentrations with anorthosite massifs makes such an analogy all the more attractive. Philpotts (1981) summarizes a relevant model for the origin of anorthosites and the cogenitors. Figure 11 from Philpotts (1979) shows that a mafic magma undergoing plagioclase and clinopyroxene fractionation *may* move into an immiscibility field, with liquid-liquid tie-lines running approximately parallel to the SiO_2 – (FeO, TiO_2, CaO, P_2O_5) join. Thus, the two liquids generated will be an

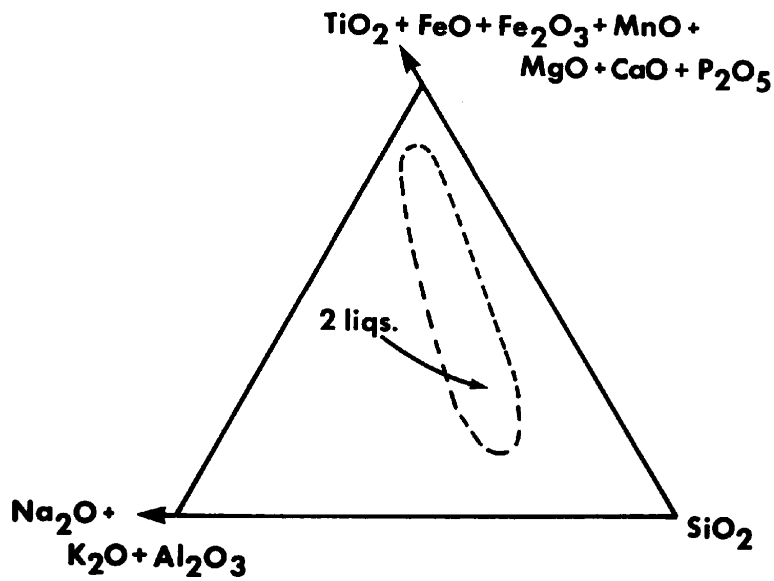

Figure 11: Plot f the liquid miscibility gap in the system SiO_2-Na_2O-K_2O-Al_2O_3-TiO_2-Fe_2O_3-FeO-MnO-MgO-CaO-P_2O_5 from Philpotts (1977).

iron-rich ferrodiorite and a silica-rich quartz mangerite. Table 6 lists analyses of Philpott's proposed parental magma and two experimentally derived immiscible liquids as well as two whole rock analyses of a jotunite (ferrodiorite) and a quartz mangerite. Noteworthy is the fact that the quartz mangerite phase differs from typical calc-alkaline or mildly alkaline granites only in its K_2O content. Such a variation might be easily obscured during later metamorphism.

Table 6. Composition of Immiscible Liquids in an Andesite System [all analyses from Philpotts, 1981]

		103°			
	Starting Material	**Fe-rich**	**Si-rich**	**Quartz mangerite**	**Ferrodio rite**
SiO_2	61.31	41.50	72.24	68.6	48.0
TiO_2	0.86	2.61	0.62	1.0	3.3
$A1_2O_3$	13.35	7.74	11.76	13.1	10.9
FeO(tot)	10.31	22.1	6.20	4.6	16.9
MnO	0.11	0.20	0.00	0.1	0.1
MgO	0.81	4.02	0.50	5.5	0.7
CaO	6.72	13.37	3.41	2.8	9.1
Na_2O	1.90	0.60	1.60	3.1	2.2
K_2O	3.54	0.90	2.82	5.4	2.6
P_2O_5	0.94	6.53	0.77	0.3	2.5

REE data for the Cranberry ore suggest that the Cranberry "magma" equilibrated with a plagioclase fractionating system, as the model would propose. This inference is drawn from the presence of a negative Eu anomaly and a La/Sm ratio of approximately 1 in Cranberry magnetite ore. Continuing fractional crystallization of the ferrodiorite will precipitate clinopyroxene with resulting iron-enrichment toward the Fe, Ti, Ca, P oxide apex of Philpott's phase diagram. The presence of significant amounts of TiO2 and P2O5 enhances this iron enrichment by fluxing the ferrodioritic magma to temperature well below 1000oC. The trace element evidence for the magnetites supports a model in which magnetite was in equilibrium with a magma relatively depleted in Cr, V, Mn, Cu, and Ni. Typical gabbros contain 5-6% Fe-Ti oxides. If only 2% Fe-Ti oxides crystallized during the fractional crystallization of the Cranberry parental magma (a necessary condition for the liquid path to intersect the two-liquid field), there would still be a significant depletion of the remaining magma in Cr, V, Mn, Cu, and Ni. Using a Rayleigh fractionation model with a magnetite/melt $K_D = 50$, a reasonable value for transition metals in magnetites and ilmenites, the fractionated magma would become depleted in these trace elements by approximately 49-63%, depending on whether one assumes Rayleigh-or Nernst-type equilibration of oxides with the melt. With a $K_D = 100$, the proportion of trace elements left in the melt drops to 16-36%. Thus, when the melt unmixes on intersecting the miscibility gap, there is already a very significant depletion in Cr, V, Mn, Cu, and Ni. The Cranberry parental liquid is, then a trace element depleted system, more analagous in its transition element abundances to a granitic melt than a mafic-ultramafic one.

One would like to test this transition metal depletion model by utilizing magnetite/melt K_D's obtained from similar natural systems. However, distribution coefficient data are not plentiful for the elements and phases analysed in this study. Alternatively, by ratioing incompatible to compatible elements which were affected by the crystallization process, it is possible to determine if the trace element content of the Cranberry magnetites corresponds to that expected in mafic or felsic systems. A review of the data presented by Haggerty (1976) and Rollinson (1980) suggests that the Mg/Mn ratio of magnetites precipitated from basaltic to andesitic liquids is in the range of 1-15, while those from granitic liquids are <<1. The Cranberry magnetites show Mg/Mn ratios of 0.2 to <.02. Furthermore, the two Bakersville magnetite samples are ≥ 3.0. This suggests that the Cranberry magnetites were in equilibrium with a more silica-rich liquid than the Bakersville gabbro.

The Bakersville, then, appears not to be compositionally similar to the magma which crystallized to form Cranberry magnetite ore, but *is* a possible co-genitor of the Cranberry parental magma. The Bakersville gabbro is a clinopyroxene-plagioclase-magnetite rock with significant apatite (Gulley, 1983). Philpotts (1981) describe the typical plagioclase of anorthosite suites as An_{60-30} and the clinopyroxene as $Ca_{47}Mg_{43}Fe_{10}$. Relict igneous plagioclase in the Bakersville is reported as An_{43-53} (Gulley, 1983) and the clinopyroxene is variously $Ca_{40-48}Mg_{33-51}Fe_{9-23}$ (Wilcox and Poldevaart, 1958).

Thus, we would suggest that Bakersville-type magmas intruded the Cranberry gneiss and differentiated, in rare situations, into an iron-rich and SiO_2-rich two-liquid system by plagioclase and clinopyroxene fractionation. To approach the two-liquid miscibility field it is necessary that negligible amounts of magnetite be extracted during this fractionation process. Herz (1982) has suggested a liquid immiscibility process as the most likely mechanism for generating the ilmenite-apatite (nelsonite) rocks associated with anorthosites and charnokites in the granulite terrains in the Blue Ridge and Piedmont of the Roseland district, Nelson and Amherst Counties, Virginia. If magnetite was a volumetrically significant liquids phase at Cranberry, then the remaining liquid composition will miss the two-liquid field in all but the most P_2O_5-TiO_2 rich systems. Thus, the typical plagioclase-clinopyroxene-magnetite-apatite-bearing Bakersville gabbro or diabase does not appear to represent an iron-rich liquid capable of producing a Cranberry-type accumulation of magnetite.

To prevent magnetite precipitation, it is necessary that f_{O2} and f_{H2O} be maintained at low values. Gulley (1983) has described granulite facies rocks in the Elk Park Plutonic Group. This granulite facies event predates a Bakersville-type metadiabase which outcrops near Roan Mountain, Tennessee. This area is within the Blue Ridge thrust sheet about 20 km west of Cranberry. Presumably, other intrusives which were co-genetic with the Bakersville gabbro (s.s.) may have encountered areas of granulite facies rocks and thus may have lost significant H2O to the surrounding country rocks. A decrease in the fH2O during the fractional crystallization of plagioclase and, possibly, clinopyroxene, lowers fO2 and permits the magma to enter the two-liquid field and to generate ferrodioritic magmas. These ferrodiorities, possibly because of their significantly lower viscosity with respect to the co-genetic quartz mangeritic liquid (estimated by Philpotts, 1981, as differing by a factor of 25,000), intruded as dike-like bodies which, in turn, continued to fractionate plagioclase and clinopyroxene to produce the magnetite-apatite-clinopyroxene bodies. The plagioclase-clinopyroxene haloes which surround the magnetite-rich rocks may represent, then, a complementary liquid phase that was in equilibrium with magnetite. It is possible that this final crystallization episode was enhanced by an increase in f_{O2} of f_{H2O} after separation from the felsic liquid. The question of the whereabouts of the corresponding felsic fraction is moot, as it is not clear that such a rock would be easily distinguished from known granites in the Crossnore Plutonic Group.

CONCLUSIONS

We make the following conclusions on the origin of the Cranberry Magnetite deposits.

1. The geological evidence argues strongly against a skarn of post-magmatic, hydrothermal origin for the magnetites.
2. The REE patterns in the magnetite rocks are similar to the patterns of many intrusives of mafic affinities, and similar to iron-rich rocks considered to be derived fro residual liquids associated with massif anorthosite bodies. The REE data suggest that the magnetites at some stage equilibrated with a plagioclase-rich restite phase. The REE patterns are distinctly different from typical sedimentary rocks or sedimentary iron formations.
3. The trace elements in magnetite separates from the Cranberry ore deposit are most similar to those fro late-stage differentiates, such as Kiruna or Pea Ridge/ Iron Mountain. The Cranberry magnetites contain, as a group, significantly less MgO, MnO,V, Cr, and Ni than cumulate magnetites from stratiform intrusives or than several magnetites from the Bakersville gabbro. The trace element ratios are similar to those from magnetites that formed on the liquids of mafic melts. These ratios are also distinctly different from those of magnetites in the Bakersville gabbro.
4. The origin of the Cranberry magnetite deposit is best explained by the low f_{O2}-low f_{H2O} emplacement of a mafic magma of Bakersville-type. During fractional crystallization of plagioclase and clinopyroxene without significant Fe-Ti oxide crystallization, the fractionating magma is driven into a field of liquid immiscibility, forming a felsic and an iron-rich magma. The less viscous iron-rich magma, a ferrodiorite intrudes as dike-like masses which then continue to fractionate in place to produce the magnetite-rich ore surrounded by a pla-

gioclase-clinopyroxene restite. The low f_{H2O} and f_{O2} essential to this model may be a result of intrusion of some Bakersville magma bodies into "dry" granulite terrains of the Blue Ridge thrust sheet.

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AGE AND P-T CONDITIONS DURING METAMORPHISM OF GRANULITE-FACIES GNEISSES, ROAN MOUNTAIN, NC-TN

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ABSTRACT

Polymetamorphic basement gneisses on Roan Mountain, in the western Blue Ridge province of North Carolina and Tennessee, may be divided into two metamorphic suites: 1) felsic to mafic, massive to layered, orthopyroxene-bearing rocks ("Carvers Gap gneiss") with occasional quartzo feldspathic segregations ("granitoid pods"); and 2) massive to layered garnet-kyanite gneiss ("Cloudland gneiss"). Both gneiss suites were metamorphosed tot he granulite facies in the Precambrian under estimated P-T conditions of 680-760 C and 10-12 kbar. The Carvers Gap gneiss records an Rb-Sr whole-rock age of 1815 m.y. (I.R.= 0.7058), interpreted to represent either 1) the time of metamophism of continental crust, or 2) the time of formation of crustally-derived felsic igneous material. The Cloudland gneiss and granitoid pods yield an age of 807 m.y. (I.R.=0.7149) possibly corresponding to granulite-facies metamorphism of the gneiss during the waning stages of the Grenville orogeny. The granitoid pods formed at this time by limited partial melting, under dry conditions, of the enclosing Carvers Gap gneiss.

INTRODUCTION

The existence of hypersthene-bearing gneisses in the western Blue Ridge province of North Carolina has been known for over two decades (Bryant, 1962; Bryant and Reed, 1970a; Merschat, 1977; Kuchenbuch, 1979). Recent field work and petrologic studies by Gulley (1981,1982)have confirmed the existence of a Proterozoic granulite-facies terrane on Roan Mountain, along the North Carolina-Tennessee state line (Fig. 1).

Two distinct granulite-facies lithologies occur on Roan Mountain: 1) felsic to mafic, massive to layered, meta-igneous (?) gneisses characterized by orthopyroxene ± clinopyroxene ± hornblende; and 2) quartzofeldspathic to pelitic gneisses characterized by garnet ± kyanite (paramorphous after sillimanite). The pyroxene-bearing gneisses are informally designated Carvers Gap gneiss; the garnet/kyanite-bearing gneisses are termed Cloudland gneiss.

Despite later intrusion by diabase dikes associated with the Bakersville Gabbro and overprinting by Taconic garnet-grade metamorphism, the Carvers Gap and Cloudland gneisses display only minor retrogression and deformation. These lithologies therefore provide a unique opportunity to study the physical and chemical evolution of the southern Blue Ridge belt. In this paper we present preliminary RB-Sr whole-rock age data and discuss the metamorphic history of the granulite-facies gneisses on Roan Mountain.

GEOLOGIC SETTING

Roan Mountain lies within the Hayesville thrust sheet (Hatcher, 1978) 15 km west of the Grandfather Mountain window and 10 km north of the Spruce Pine synclinoriu (Fig. 1). The granulitic gneisses on Roan Mountain, consisting of the Carvers Gap and Cloudland gneisses (Gulley, 1982), have been mapped as part of the Elk Park Plutonic Group (Tankin and others, 1973; Rankin, 1975), although geologic relationships between the Roan Mountain granulites and Elk Park plutonic rocks are by no means well-established. The Elk Park plutonic suite was apparently stabilized as continental crust by at least 1200-1300 million years ago (m.y.) (Fullagar and Odom, 1973). Igneous rocks of the Elk Park group were metamorphosed during the Grenville orogeny approximately 1000m.y. (Rodgers, 1970; Fullagar and Odom, 1973; Dallmeyer, 1975a; Sutter and Dallmeyer, 1982).

The contact between the Elk Park Plutonic Group and the overlying metasedimentary and metavolcanic rocks of the Ashe Formation may be either a nonconformity (Rankin, 1970) or a fault (Bryant and Reed, 1970a, p. 36). South of Roan Mountain near Bakersville NC (fig. 1), the Elk Park Plutonic group appears to be nonconformably overlain by the Ashe Formation (J.R. Butler, personal communication). The Ashe Formation is apparently absent in the vicinity of

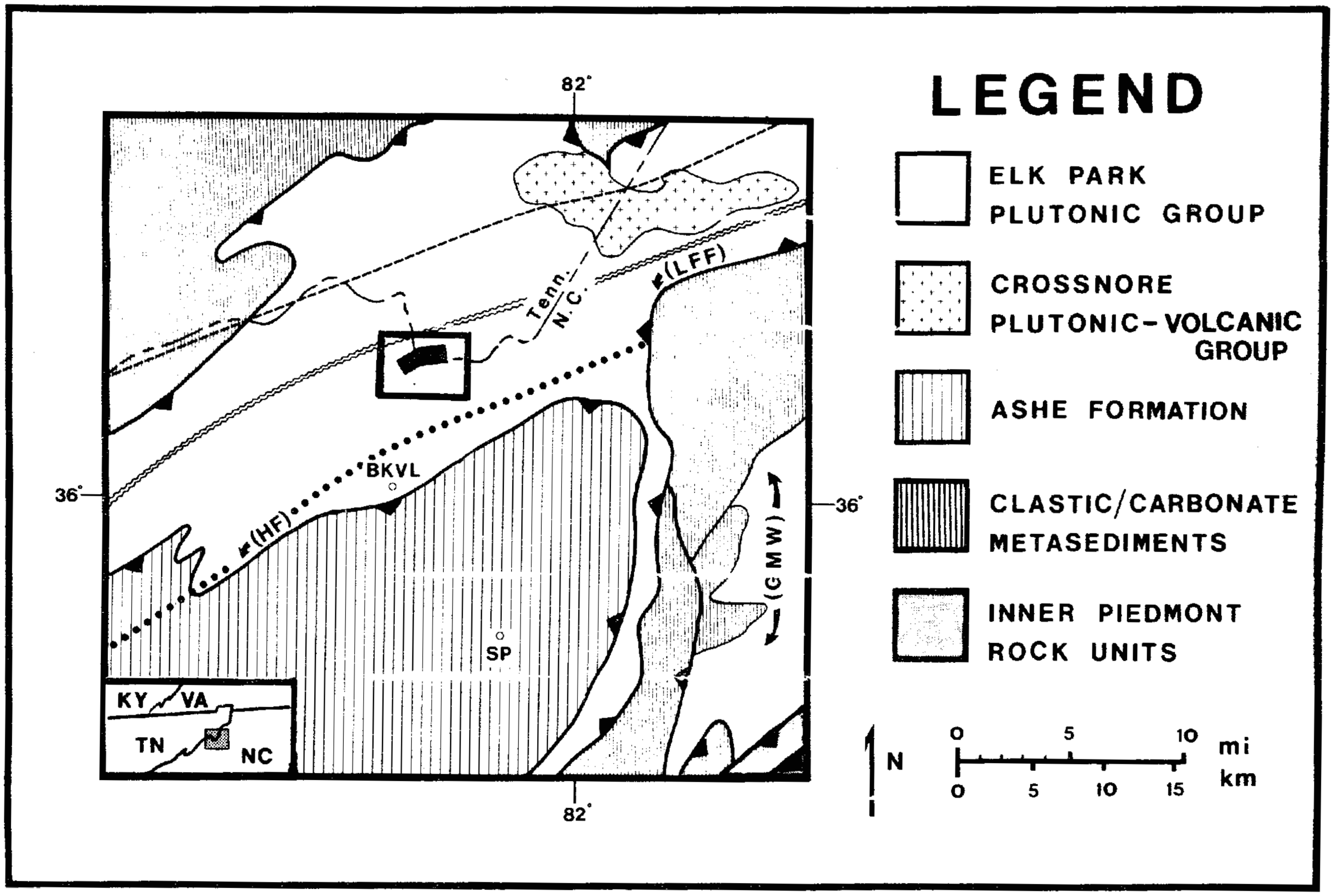

Fig. 1 Generalized geology of the Roan Mountain, NC.-TN., area (after Williams, 1978). Roan Mountain is located in the enclosed area in the central portion of the figure. (HF), Hayesville thrust fault; (GMW), Grandfather Mountain Window; (LFF). Linville Falls fault; (BKVL) and (SP), Bakersville and Spruce Pine, NC., respectively. Dashed line = first appearance of Paleozoic garnet; doubled sawtoothed line = first appearance of Paleozoic staurolite – kyanite in pelitic lithologies (Dallmeyer, 1975a).

Roan Mountain. However, Carvers Gap and Clouldland gneisses are both intruded by diabase dikes of the Bakersville Gabbro (Wilcox and Poldervaart, 1958), which are perhaps intrusive equivalents of basal Ashe volcanics. These dikes are considered to be mafic members of the Crossnore Plutonic-Volcanic Group (Rankin and others, 1973; Rankin, 1975), interpreted to have been emplaced during the late Precambrian opening of the Iapetus Ocean (Odon and Fullagar, 1973, 1982).

The major Paleozoic metamorphism in the North Carolina Blue Ridge occurred approximately 470-480 m.y. during the Taconic orogeny (Dallmeyer, 1975b; Sutter, 1982; Sutter and Dallmeyer, 1982). Taconic metamorphic conditions ranged from greenschist facies in the northwestern Blue Ridge belt to granulite facies in the southeastern Blue Ridge belt (Carpenter, 1970; Bryant and Reed, 1970b; Hatcher and others, 1979; McElhaney and McSween, 1982, 1983; Absher and McSween, 1983).

OUTCROP LOCATIONS

Carvers Gap gneiss is named after the topographic feature at the top of Roan Mountain, found on the U.S.G.S. 7 ¾-minute Carvers Gap topographic map. The best exposures of Carvers Gap gneiss occur along 1) a paved road from Carvers Gap to the top of Roan Mountain; 2) an unpaved access road ("Balsam Road"); and 3) the Appalachian Trail (Fig. 2). Carvers Gap gneiss is also exposed at an overlook on Roan High Bluff, which may be reached by a partially-paved footpath that leads to the overlook.

Cloudland gneiss is named for Cloudland resort (Keith, 1907), a late 19th-century hotel once situated on Roan Mountain. The most accessible exposures of Cloudland gneiss occur next to a parking lot between Roan High Knob and Roan High Bluff, at the end of the paved road extending from Carvers Gap.

Both the Carvers Gap and Cloudland gneisses were sampled for Rb-Sr isotopic analysis. Sample locations in the gneisses are shown on Figure 2.

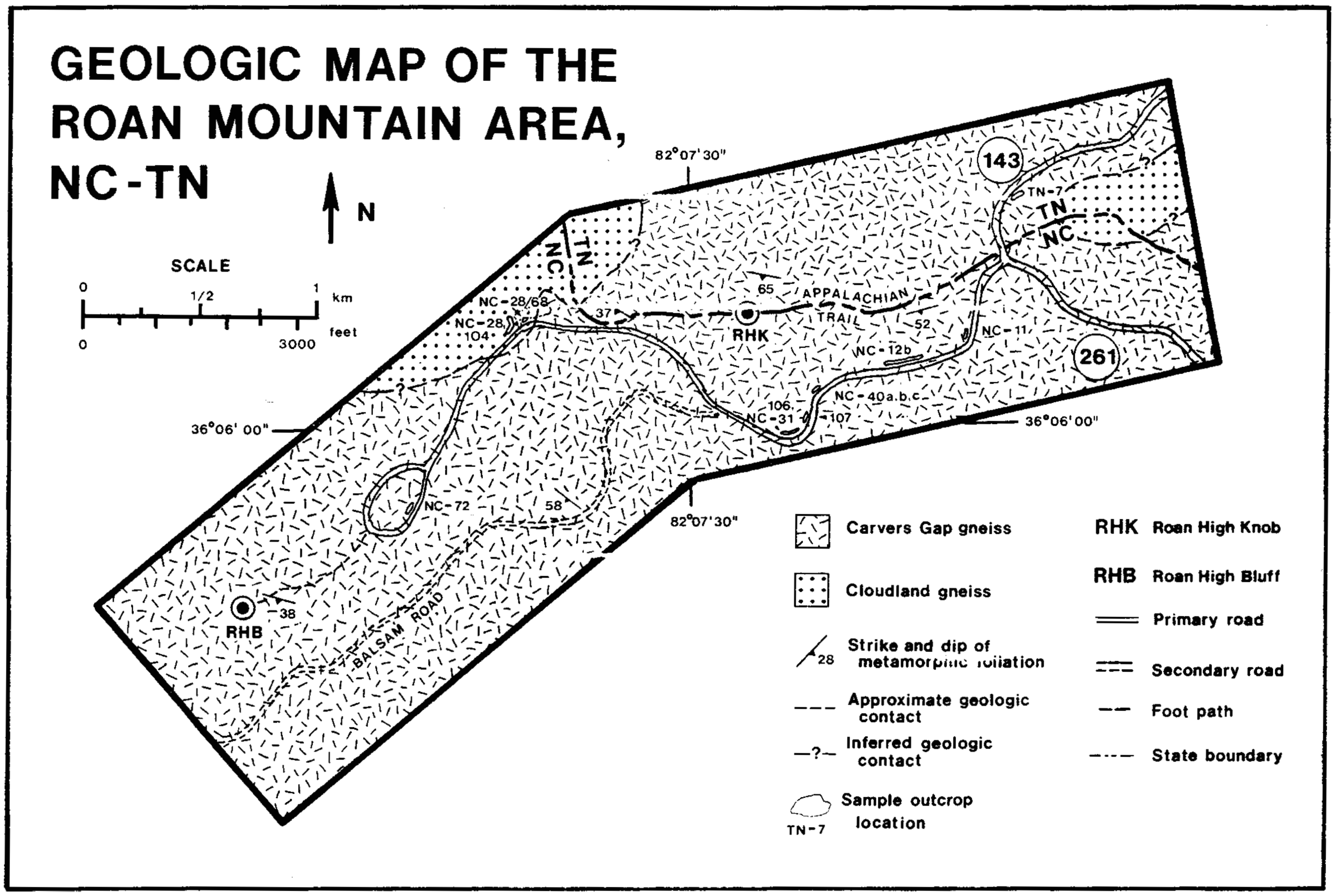

Fig. 2 Generalized geologic map of Roan Mountain. Carvers Gap is located along the NC.-TN. State line at the intersection of state highways 261 and 143.

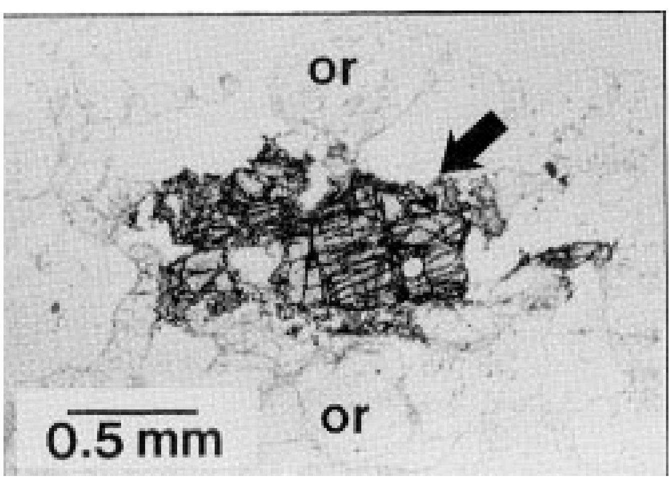

Fig. 3 Photomicrograph of orthopyroxene (see arrow) in a massive Carver's Gap gneiss (plane-polarized light). Grains of perthitic orthoclase (or) are also present.

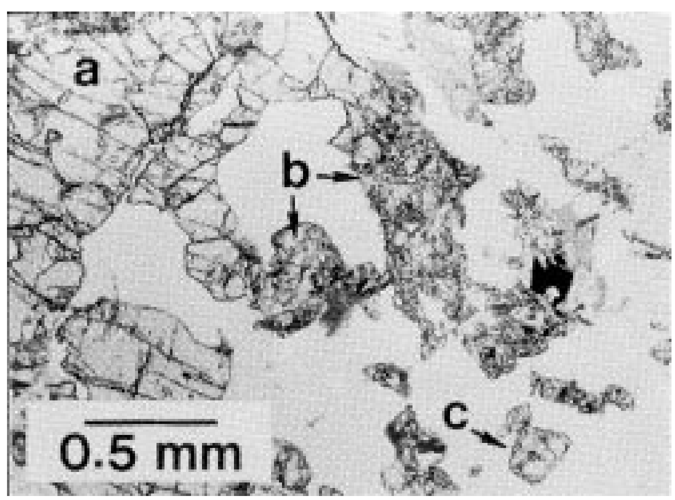

Fig. 4 Photomicrograph of a high-alumina domain in Cloudland gneiss (plane-polarized light). "a" is unaltered granulite-facies garnet; "b" refers to pseudomorphous replacement of the garnet by aggregates of quartz-aluminum silicate-opaque mineral; "c" is a former sillimanite grain paramorphopusly replaced by grains of kyanite.

PETROGRAPHY

Most Carvers Gap gneiss has a fine- to medium-grained, massive to faintly layered, felsic lithology (Appendix 1). The gneiss is often intergradational between the massive and layered varieties perhaps as result of metamorphic differentiation. Amphibolitic, retrogressed equivalents of these gneisses comprise a small portion of the total gneisses observed. In general, Carvers Gap gneisses have a granoblastic texture. Locally, nematoblastic (in layered pyroxene-bearing gneiss) or isoblastic (in layered amphibolitic gneiss) textures are present. Massive mafic gneiss, nodules of hornblende-bearing pyroxenite (up to 25 cm in diameter), and "pods" or segregations of coarse-grained granitoid rock (Appendix 1) constitute a minor volume of Carvers Gap gneiss.

Orthopyroxene is the characteristic granulite-facies mineral in massive Carvers Gap gneiss; clinopyroxene and hornblende additionally occur in layered felsic and massive mafic gneisses and hornblende pyroxenites. Biotite or orthopyroxene is the characteristic mafic mineral in the granitoid segregations. Quartz, potassium feldspar, and plagioclase feldspar are the other major minerals present in massive and layered gneisses and granitoid segregations (Appendix 2). Potassium feldspars are perthitic orthoclases; some plagioclases have antiperthitic structure. Garnet and amphibole form retrograde alteration coronas around mafic granulite-facies minerals in the gneisses (Fig. 3).

Cloudland gneiss consists of lithologically segregated high-alumina and normal-alumina domains (Appendix 1). The former are medium-grained and foliated, and contain garnet and kyanite; the latter are coarse-grained, granoblastic, and have garnet but no (or only trace amounts of) kyanite (Appendix 2). Both domains of Cloudland gneiss have perthitic orthoclase and abundant, rounded zircons. Retrogression of garnet produced biotite ± plagioclase. All kyanite is paramorphous after sillimanite (Fig. 4).

METAMORPHIC PRESSURE-TEMPERATURE CONDITIONS

Two-pyroxene geothermometry (Wood and Banno, 1973; Wells, 1977) combined with mineral stability curves provide temperature and pressure estimates of granulite-facies metamorphism (Fig. 5). Coexisting orthopyroxenes and clinopyroxenes in massive mafic and layered granulitic gneisses have been used in the geothermometry calculations (Gulley, 1982). Mineral reactions which limit pressures and temperatures include 1) garnet stability in Cloudland gneiss and mafic Carvers Gap gneiss; and 2) the stability of hypersolvus feldspars. Estimated pressure-temperature conditions for the granulite-facies metamorphism are 750-847 C and 6.5-8.0 kbar.

Temperature and pressure estimates of amphibolite-facies metamorphism (Fig. 6) are based on four criteria: 1)the presence of garnet and plagioclase in mafic Carvers Gap gneiss; 2) the breakdown of orthopyroxene ± quartz to biotite + potassium feldspar; 3) the inversion of sillimanite to kyanite; and 4) the presence of kyanite + garnet vs. staurolite + quartz. These mineral reactions provide approximations of Taconic metamorphic conditions at 710-765 C and 9.8-11.6 kbar. The pressure limits of 9.8-11.6 kbar old for anyhydrous conditions (Green and Ringwood, 1967), and would be slightly lower for hydrous Taconic metamorphism.

GEOCHRONOLOGY

Samples from three rock units were selected for Rb-Sr isotopic analyses: Carvers Gap massive and layered gneiss, granitoid segregations (or "pods") contained within the Carvers Gap gneiss, and Cloudland gneiss. Sample selection within each suite was based on whole-rock chemical and petrographic similarities. These criteria have provided geologically reasonable ages for identifiable litholo0gic unites in Archean greenschist/amphibolite facies layered gneisses in India (Monrad, 1983) and Proterozoic granulite-facies gneisses in Australia (Gray and Compston, 1982). Each sample for isotopic analysis weighed at least 5 kg in order to minimize the effects of possible isotopic mobility or inhomogeneity on age determinations (Fullagar and Odom, 1973). Despite this precaution, a few samples from each suite contained portions of different lithologic units upon thin section examination. These petrographically mixed samples were excluded from the isochon age determinations. Details of the isotopic analyses and Rb-Sr data are listed in Appendix 3. Analytical errors are within the area of the circles representing each sample on the isochrons.

Five samples of the Carvers Gap gneiss yield a whole-rock Rb-Sr isochron age of 1815 ± 31 m.y. and an initial ^{87}Sr r/^{86}Sr ratio of 07058 ± 0.0003 (1? errors; Fig. 7). This age is about 500 m.y. older than that recorded for plutonic rocks considered part of the Elk Park Group within the Grandfather mountain window (Fullagar and Odom, 1973), but approaches a U-Pb zircon age of about 1870 m.y. obtained by Sinha and Bartholomew (1982) for Blue Ridge basement paragneiss in central Virginia. The sample with the highest ^{87}Rb/^{86}Sr ratio (TN-7; Fig. 7) was obtained about 1/2-k east of the general area of the other four samples (Fig. 2). If this sample is excluded from the isochron age determination, the age and initial ^{87}Sr/^{86}Sr ratio is consistent with the protolith or precursor to the Carvers Gap gneiss having a significant crustal residence time prior to about 1800 m.y.

An isochron constructed by using samples from both the Cloudland gneiss and granitoid segregations (pods) from the Carvers Gap gneiss yields an age of 807 ± 26 m.y. and an initial ^{87}Sr/^{86}Sr ratio of 0.7140 ± 0.0004 (Fig. 8). The M.S.W.D. for this isochron (3.3) suggests a relatively crude fit of the data to the indicated isochron; however, the residuals determined from the York (1969) regression indicate a

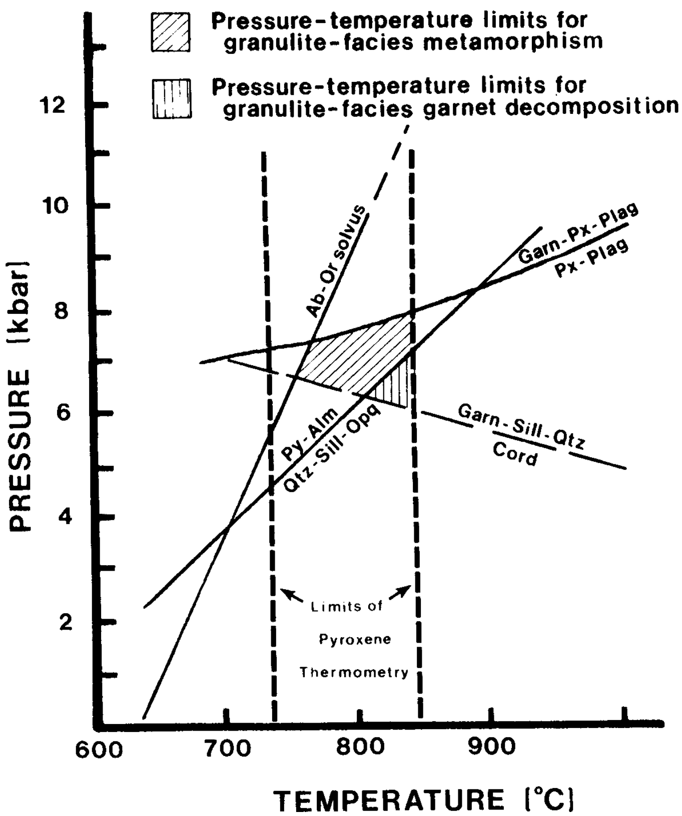

Fig. 5 Limiting pressure-temperature curves for granulite-facies metamorphism. The given reactions are (1) ab-or solvus (Morse,1970); (2) garnet + pyroxene + plagioclase ⇔ pyroxene + plagioclase (Green and Ringwood, 1967); (3) garnet + sillimanite + quartz ⇔ cordierite (Thompson, 1976); and (4) pyrope - almandine garnet ⇔ quartz + sillimanite + opaque mineral (Yoder and Chinner, 1960).

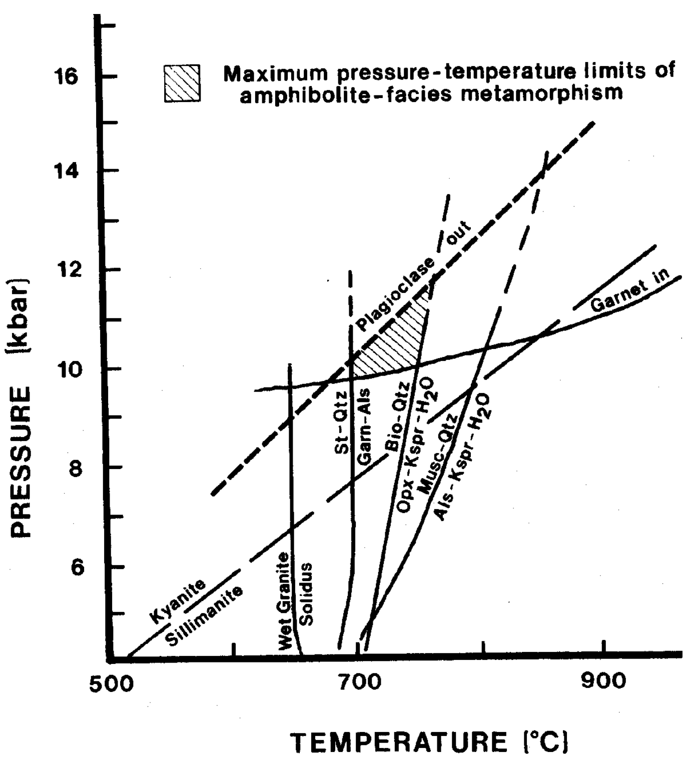

Fig. 6 Limiting pressure-temperature curves for amphiboli5e-facies metamorphism. The given reactions are (1) plagioclase out, garnet in (Green and Ringwood, 1967); (2) kyanite ⇔ sillimanite (Holdway, 1971) (3) wet granite solidus (Luth and others, 1964); (4) staurolite + quartz ⇔ garnet + aluminum silicate (Hoschek, 1967); and (5) biotite + quartz ⇔ orthopyroxene + K-feldspar + H_2O (Hess, 1969). Also shown for reference is (6) muscovite + quartz ⇔ aluminum silicate + K-feldspar + H_2O (Day, 1973).

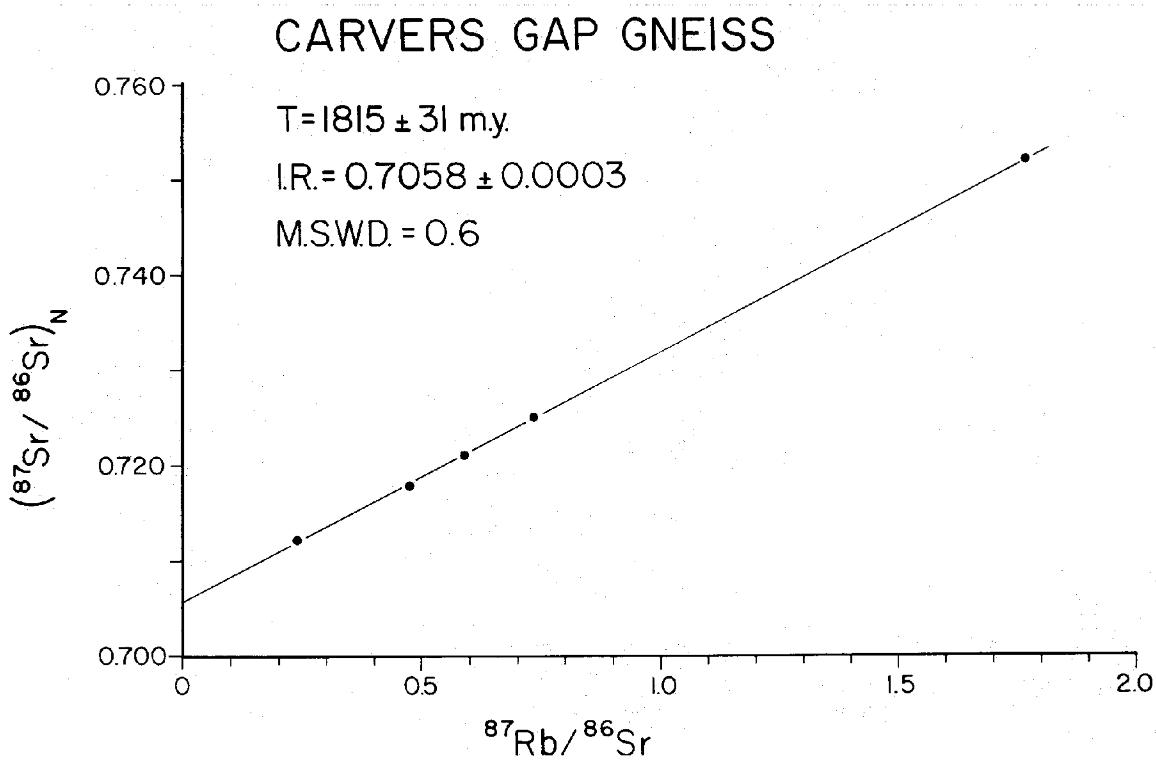

Fig. 7 Isochron for Carvers Gap gneiss. Errors are 1 sigma; M.S.W.D.=mean square of weighted deviates.

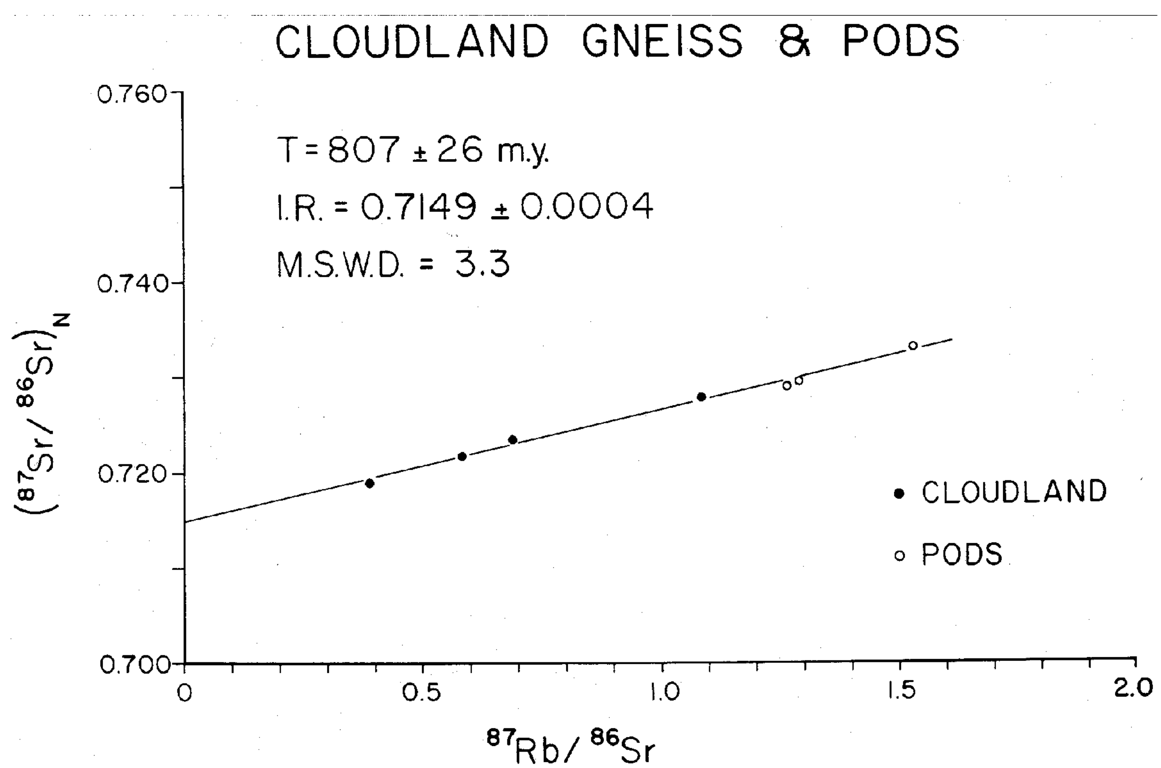

Fig. 8 Isochron for Cloudland gneiss and granitoid pods.

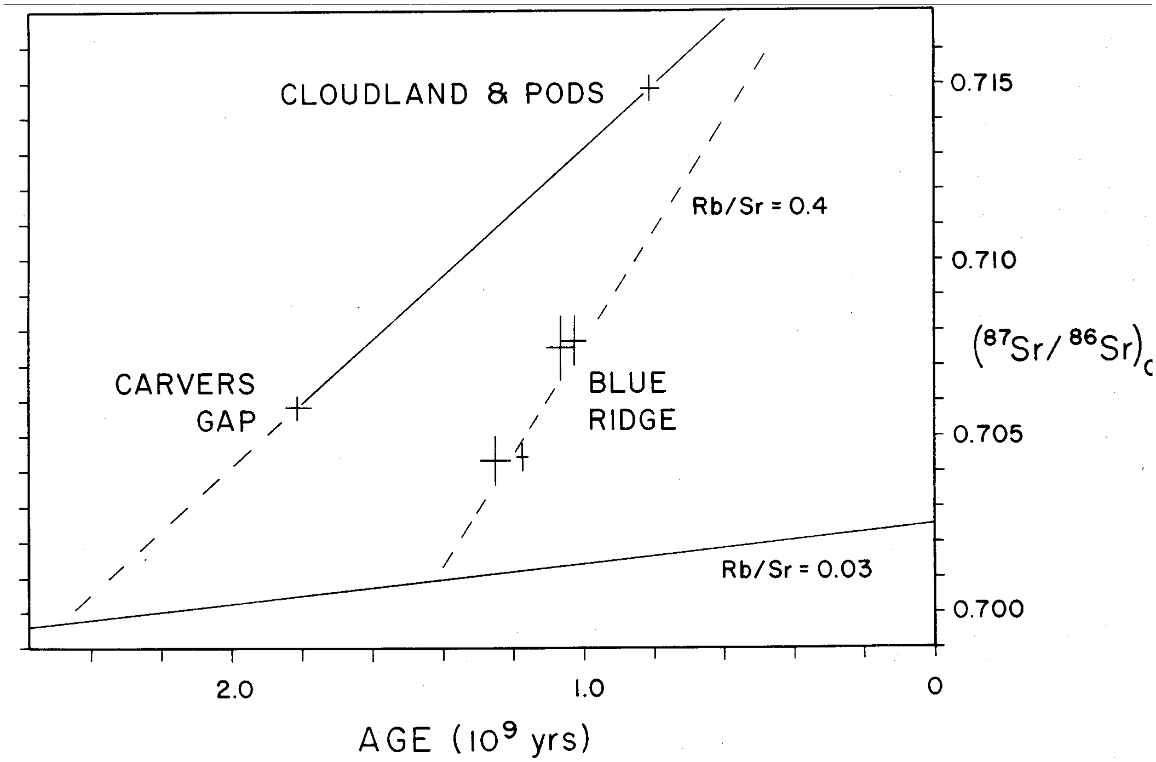

Fig. 9 Sr-evolution diagram for Roan Mountain gneisses ("Carvers Gap" and "Cloudland and Pods") and members of the Elk Park Plutonic Group ("Blue Ridge"; Fullagar and Odom, 1973). Average Rb/Sr of Carvers Gap gneiss=0.21. Mantle growth curve shown for reference (Rb-Sr = 0.03).

random, rather than systematic, scatter of data points. If samples of the granitoid pods are excluded, the isotopic age of the Cloudland gneiss increases to about 825 m.y. with an initial $^{87}Sr/^{86}Sr$ ratio of about 0.714. The high initial $^{87}Sr/^{86}Sr$ ratio indicates a relatively long crustal residence time for the protoliths or sources of the Cloudland gneiss and granitoid pods.

DISCUSSION AND CONCLUSIONS

The petrographic and isotopic data suggest the following metamorphic history for the Roan Mountain granulites. The 1800 m.y. age associated with the Carvers Gap gneiss reflects Sr isotopic homogenization during a mid-Proterozoic orogenic event, due to either 1) metamorphism of continental crust, or 2) formation of crustal-derived, dominantly felsic igneous material. Both interpretations require the existence of continental crust in the present southern Appalachians prior to 1800 Ma. We interpret the 800-m.y. age to represent limited isotopic equilibration during granulite-facies metamorphism of the Carvers Gap and Cloudland gneisses. Although this age is somewhat younger than that generally ascribed to the Grenville orogeny, it may reflect isotopic homogenization during the waning stages of Grenville metamorphism.

The presence of granulite assemblages in the Cloudland gneiss requires its emplacement prior to 800 Ma, the time of the granulite event. Abundant aluminum silicates in the Cloudland gneiss suggest a sedimentary protolith, the sedimentary material perhaps derived form the adjacent Carver Gap gneiss. By contrast, the lack of a high grade mineral suite in the granitoid pods indicate that they were formed at about 800 Ma, most likely by partial melting during the granulite metamorphism. The occurrence of the pods within the Carvers Gap gneiss and their granitic composition are consistent with derivation of the pads by relatively dry partial melting of the Carvers Gap gneiss. Such dry anatexis would yield a highly potassic, viscous liquid that would remain essentially in place. In this regard, the Cloudland gneiss and granitoid pods fall directly on the Sr growth curve (Fig. 9) from the Carvers Gap gneiss (average Rb/Sr of 0.21), which is required for their derivation from the Carvers Gap gneiss. Metamorphism under relatively dry conditions also reduces the likelihood that Sr isotopic mobility would occur within the Carvers Gap gneiss, thus allowing retention of its 1900 m.y. age.

The amphibolite-faces event most likely occurred during the Taconic orogeny (Gulley, 1982). We are currently performing analyses on mineral separates to determine the actual age of this event. Nevertheless, metamorphism followed emplacement of the Bakersville Gabbro dikes, which generally display amphibolite facies assemblages. The relatively high pressures obtained (Fig. 6) may reflect metamorphism during subduction.

If the interpretation outline about is correct, correlations between the Roan Mountain gneisses and members of the Elk Park Plutonic Group are suspect. It is likely that the Carvers Gap gneiss, and perhaps other granulites exposed further west in Tennessee or contained within other Blue Ridge thrust sheets, represent lithologies emplaced in to continental crust at least 500 m.y. before intrusion of the Elk Park suite. We therefore suggest that remnants of continental crust in the Southern Appalachians have a minimum age of about 1800 m.y.

AKNOWLEDGEMENTS

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APPENDIX 1. WHOLE-ROCK GEOCHEMISTRY AND NORMATIVE MINERALOGY.

Whole-rock chemical analyses were performed at the Department of Geology, University of North Carolina at Chapel Hill. SiO_2, TiO_2, Al_2O_3, Fe_2O_3(T), CaO, MnO, K_2O, Rb, Sr, Y, and Zr were measured by X-ray fluorescence; P_2O_5 and duplicate SiO2 were determined colorimetrically; and MgO, Na_2O, and Al_2O_3 were measured by atomic absorption spectrophotometry (Drez, 1977). FeO and Fe_2O_3 were determined by titration (Shapiro, 1975). Normative mineralogies were calculated after Morse (1976). All major oxides are recorded in weight percent; trace elements are reported as parts per million.

Representative Whole Rock Analyses

	Carvers Gap Gneiss			(pod)	Cloudland Gneiss	
	NC-12b	NC-31	TN-7	NC-40c	NC-28c	NC-67b
SiO_2	74.40	73.03	73.21	71.27	66.36	58.78
TiO_2	0.29	0.34	0.37	0.61	0.87	0.89
Al_2O_3	13.09	13.39	13.13	14.50	15.50	24.91
Fe_2O_3	0.59	0.87	0.93	0.20	1.10	1.67
FeO	1.54	1.86	2.58	1.67	2.98	6.14
MgO	0.75	0.96	0.39	0.86	1.54	2.21
CaO	2.95	3.84	1.18	1.39	3.15	0.29
MnO	0.05	0.07	0.08	0.02	0.05	0.08
Na2O	3.13	3.20	2.13	2.37	3.26	0.38
K_2O	3.23	1.95	4.39	6.27	2.94	2.46
P_2O_5	0.09	0.13	0.09	0.06	0.08	0.09
TOTAL	100.11	99.64	98.48	99.22	97.84	97.90
Rb	72	15	115	199	59	67
Sr	276	346	183	421	352	21
Y	15	21	92	15	48	59
Zr	103	161	497	13	325	348
Normative Mineralogy						
Q	36.03	36.83	39.32	28.20	25.64	39.34
Or	19.06	11.51	25.91	37.01	17.35	14.52
Ab	26.46	27.05	18.01	20.04	27.56	3.21
An	12.12	16.40	5.85	6.89	15.62	1.42
Di	2.07	2.17	--	--	--	--
Hy	3.13	3.45	4.33	4.19	6.94	11.87
Mt	0.86	1.26	1.35	--	1.61	2.43
Il	0.55	0.65	0.70	1.16	1.66	1.69
C	--	--	2.73	1.28	1.23	21.10

APPENDIX 2. MODAL MINERALOGY

Representative samples of the Carvers Gap and Cloudland gneisses were point-counted using a mechanical stage and thin sections stained for potassium feldspar. Average major mineral analytical error was ± 2.00%, with 1,000-1,500 points counted on each sample (Chayes, 1956).

	Carvers Gap gneiss		(pod)	Cloudland gneiss	
	NC-12b	TN-7	NC-40a	NC-28	NC-676
Orthopyroxene	0.70	--	--	--	--
Clinopyroxene	Tr	--	--	--	--
Hornblende	Tr	1.39	--	--	--
Garnet	0.99	4.66	--	15.92	15.40
Biotite	0.56	0.20	5.58	4.15	7.14
Quartz	49.86	36.07	25.86	23.64	45.61
K-feldspa	21.37	36.87	61.41	33.04	10.14
Plagioclase	25.39	20.32	6.86	21.86	0.19
Kyanite	--	--	--	1.19	19.69
Apatite	Tr	--	--	--	0.10
Zircon	Tr	Tr	--	0.20	0.19
Rutile	--	Tr	--	--	0.10
Perovskite	--	--	--	Tr	0.29
Opaques	0.85	0.49	0.29	Tr	0.88

APPENDIX 3. ISOTOPIC ANALYSES

Isotopic analyses for Rb and Sr were performed at Chapel Hill using a 30-cm radius, solid source, triple filament mass spectrometer. A Nuclide CA/CS-111 automation and data reduction computer system was used for data collection and analysis. The samples were prepared by standard chemical and isotope dilution techniques (Fullagar an d Odom, 1973). Blank values proved insignificant. During the period of analysis, the Eimer and Amend $SrCO_3$ standard yielded an average value of $(^{87}Sr/^{86}Sr)_N$ = 0.70763 ± 0.00010 (1σ) when normalized to $^{86}Sr/^{86}Sr$ = 0.1194. Therefore, 0.00037 has been added to each measure $(^{87}Sr/^{86}Sr)_N$ ratio in order to standardize them relative to a value of 0.70800 for the Eimer and Amend $SrCO_3$. Calculation of the isochron age and initial $^{87}Sr/^{86}Sr$ ratio utilized the York (1969) regression treatment and $\lambda_{87Rb} = 1.42 \times 10^{-11} yr^{-1}$. Rb-Sr data are listed in the table below.

Sample	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$
Carvers gneiss		
NC-11	0.71776	0.475
NC-12b	0.72123	0.595
NC-31	0.71233	0.241
NC-107	0.72504	0.734
TN-7	0.75203	1.769
Cloudland gneiss		
NC-28b	0.72782	1.085
NC-28c	0.72170	0.582
NC-28/68	0.72365	0.691
NC-104	0.71886	0.390
Grantoid pod		
NC-40a	0.73310	1.530
NC-40b	0.72902	1.290
NC-40c	0.72884	1.268

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DEBRIS-AVALANCHE TYPES FEATURES IN WATAUGA COUNTY, NORTH CAROLINA

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INTRODUCTION

During August of 1969, Hurricane Camille stalled over Nelson County, Virginia and literally dumped 27 to 28 inches of rain within 8 hours. As a result, and without warning, at least several hundred debris avalanches were initiated, mobilizing significant volumes of rock, soil and vegetation downslope and consequently causing extensive damage (Williams and Buy, 1971). These avalanches likely contributed to the loss of lives in the region. The fact that most mountainous areas in the Appalachians of eastern United States are both susceptible to debris-avalanche activity induced by high rainfall and have a moderate to high susceptibility for landslide development due to soil and rock characteristics (Radbuch-Hall and others, 1976), creates a potential threat to human life and property which necessitates the need for geologists and urban planners to recognize regions of past debris-avalanche occurrences as well as areas potentially susceptible to future avalanching.

Our main purpose in this study is to describe and compare several generations of ancient debris-avalanche occurrences in Watauga County, northwestern North Carolina, where future occurrences are likely. In addition, by comparing these older avalanches to younger, well-documented debris avalanches formed in central Virginia in 1969, where the mechanics of debris-avalanche development have been related to geology and topography, we can better document the potential for future occurrences.

DEBRIS AVALANCHE SETTING

Geologic conditions favorable for potential debris-avalanche activation are long, steep slopes thinly mantled by unconsolidated, permeable soils and colluvium, and underlain by relatively less permeable rock. Debris-avalanche movement could then be triggered during short term (usually summer) heavy rainfall which recharge the permeable soils on slopes with water. If water cannot escape rapidly enough, pore pressure builds up immediately below the surface and will overburden the slope. Eventually, slope shear strength will drop due to a reduction of intergranular friction, triggering a nearly instantaneous release of material down-slope. A debris-avalanche, then, is a rapid downslope flowage of shallow surficial soils, vegetation (when present), saprolite and/or rock mixed with major amounts of water introduced into the system from heavy storm rainfall (modified from Sharpe, 1938, p. 61).

Well documented debris avalanches have occurred approximately 200 miles northeast of Watauga County in Nelson County, Virginia, where long, steep slopes are composed of permeable, shallow, surficial soils and underlain by less permeable granitoid and mylonitic gneisses. Williams and Guy (1971) measured 12 linear debris-avalanche scars scoured during 1969. Scar dimensions range from 200 to 800 feet from top to bottom and are about 27 to 75 feet wide, while scour depths range from 1 to 3 feet and are typically limited to the depth to bedrock. Although scour depths are minor, Williams and Guy estimated that 3 to 4 thousand tons of debris were displaced in one average debris avalanche in Nelson County, thereby making the debris avalanche an important geomorphic process in slope erosion within this area.

Williams and Guy provide several references to locations of other historical debris avalanches that have occurred along the mountainous reaches of eastern United states. Figure 1 shows debris-avalanche locations nearer to Watauga County. During the summer of 1976, we examined about 100 debris avalanche scars and the related deposits which formed in the northern portion of the devastated area caused by Hurricane Camille (Figure 2a). By comparison, in Watauga County, debris-avalanche features (Figure 2b) develop in topographic basins smaller in size and with less relief than those basins in central Virginia. Also in Watauga County, debris-avalanche deposits commonly coalesce along bottoms of the valleys in which they develop, whereas in central Virginia debris-avalanche material, in many cases, bypassed local valley bottoms and pass into larger streams flowing along valley bottoms whose sides were no directly scoured by debris avalanche movement. The very large volume of water generated by Hurricane Camille likely was responsible for long debris-avalanche tracks which bypassed the first and second order valleys. Thus, by inference, the Watauga County debris avalanche probably developed during storms of lesser magnitude than Camille.

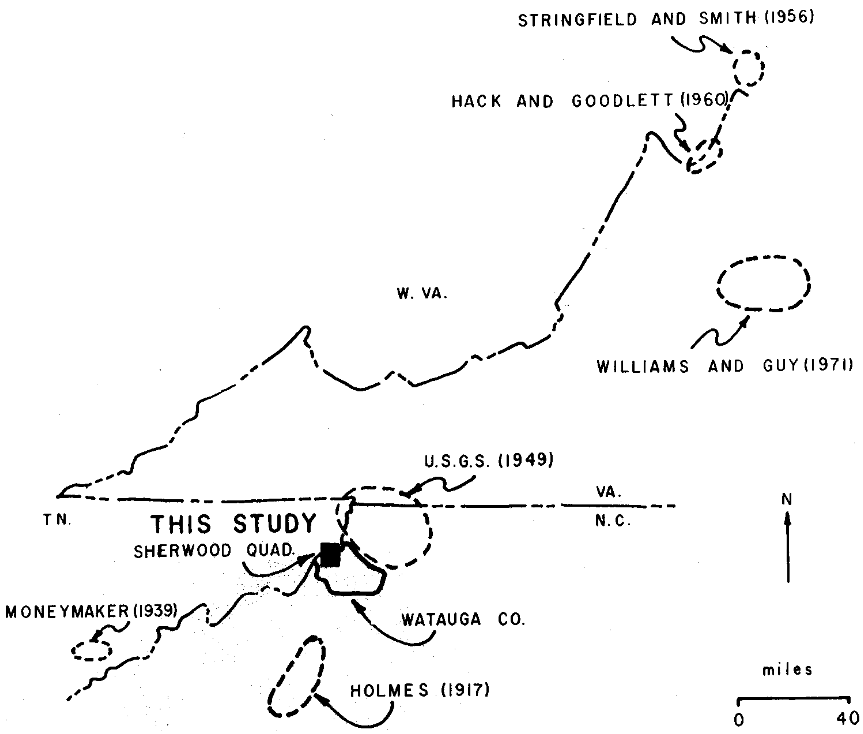
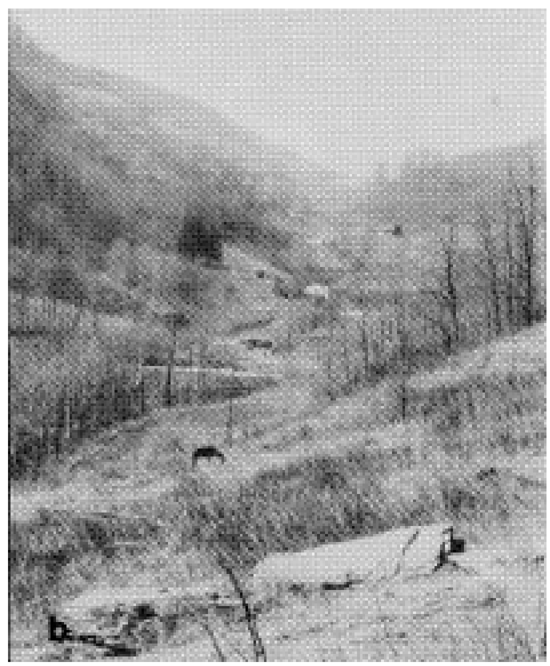


Figure 1: Location map showing Watauga County, North Carolina, in relation to documented occurrences of debris avalanches.

Figure 2 a. Eastward-viewing photograph of the Virginia Blue Ridge Mountains from Blackrock Mountain at the Wintergreen recreation area. Arrows point to debris avalanche scars scoured along naturally steep slopes during Hurricane Camille in 1969.



b. Northwestward viewing photograph of George Gap, located in the Sherwood Quadrangle, North Carolina (see Figure 3 for location), where some of the Watauga County debris avalanche features are found. In the foreground is the bouldery strewn surface of an old debris avalanche.

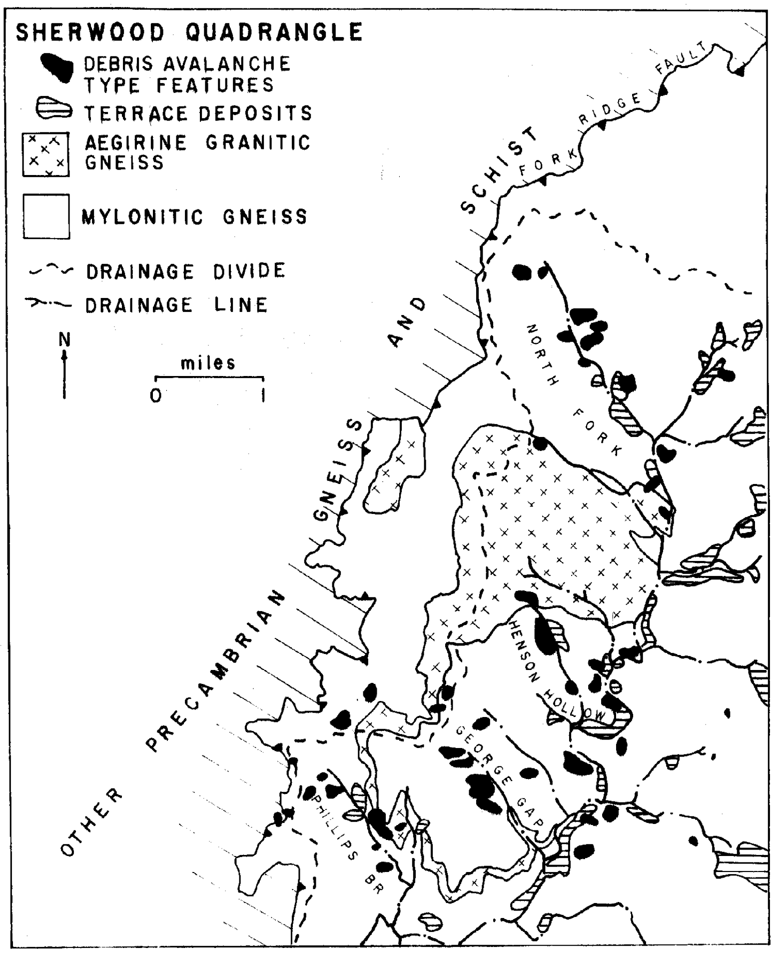


Figure 3. Bedrock geology, terrace deposits, and areas where debris-avalanche features occur in a portion of the Sherwood Quadrangle, generalized after Bartholomew and Gryta, 1980.

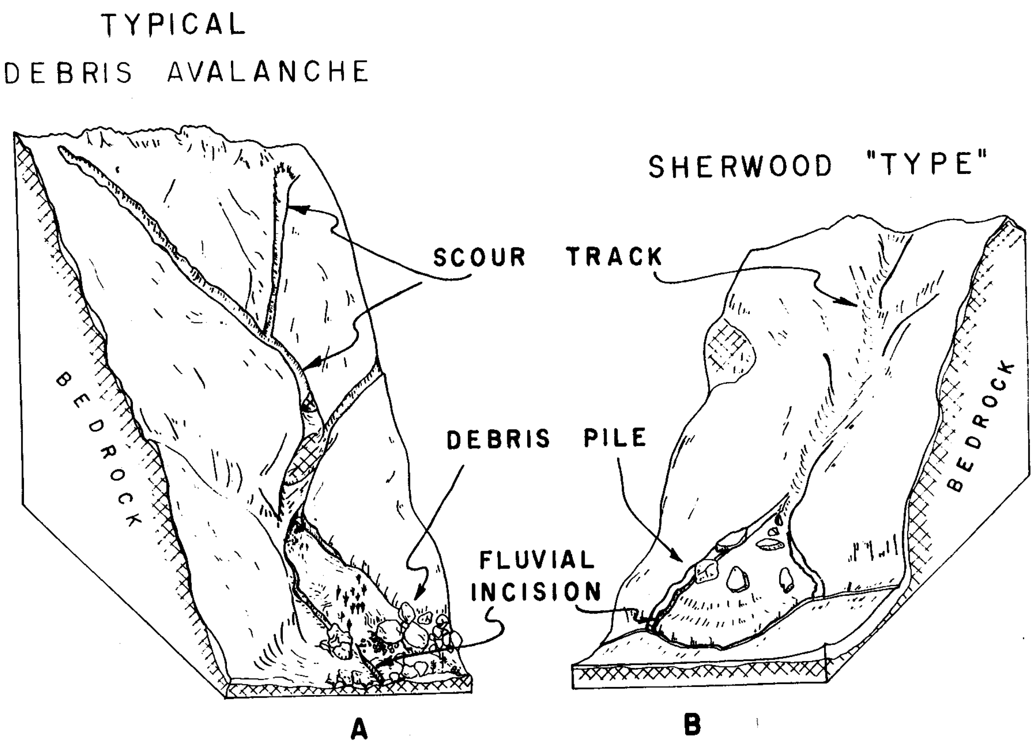


Figure 4. Schematics comparing morphology of a typical debris avalanche (A) found in central Virginia associated with Hurricane Camille to that of the Sherwood type (B) found in Watauga County, northwestern North Carolina.

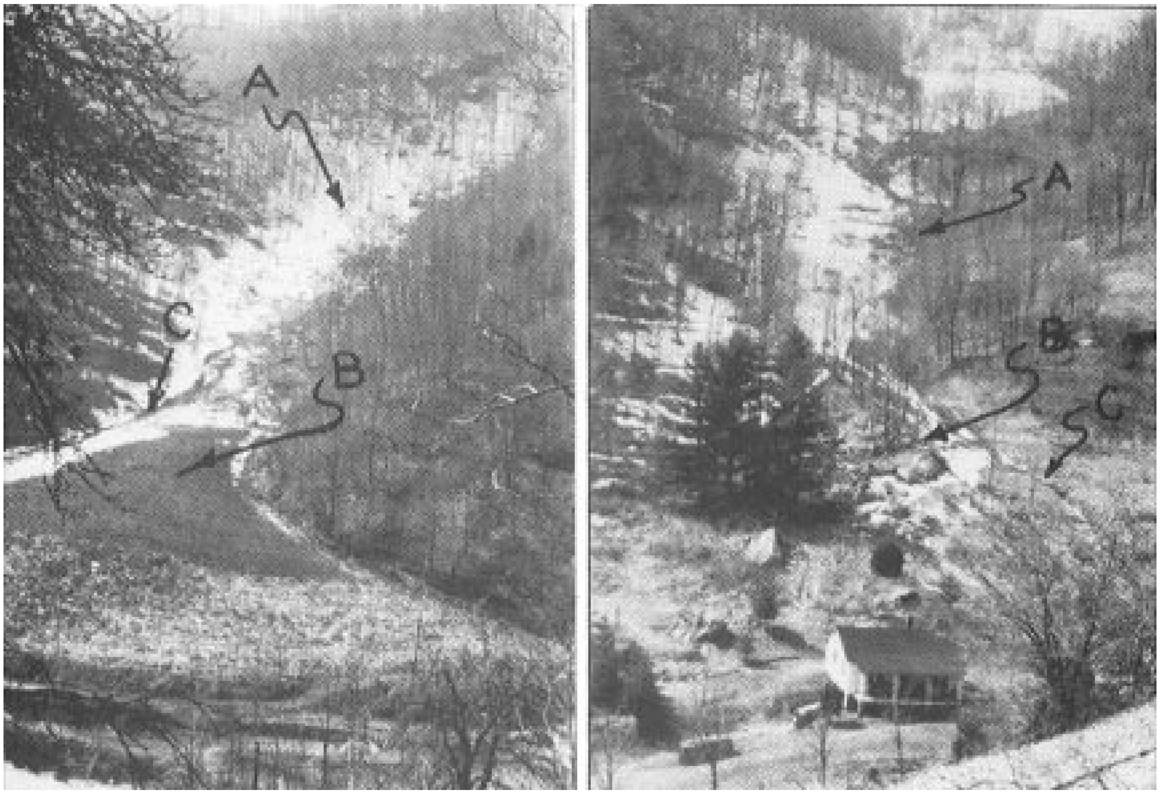


Figure 5. Two younger Sherwood debris avalanches found in George Gap showing a u-shaped track (A), a conical shaped toe debris pile (B) and post-movement fluvial incision down along the debris (C). Building construction in the Sherwood Quadrangle commonly occurs on top of or against toe debris deposits.

WATAUGA COUNTY DEBRIS-AVALANCHE FEATURES

Watauga county is located in the Blue Ridge Mountains of northwestern North Carolina; valley sides are steep and underlain by Precambrian schists and gneisses with Precambrian and lower Cambrian metavolcanic and metasedimentary rocks forming many of the higher peaks. Watauga county debris avalanches described here largely occur in a portion of the Sherwood 7.5-minute quadrangle (also shown in Figure 1). In this quadrangle, more than 70 mappable debris avalanches occur along valley slopes and are mostly concentrated in a region underlain chiefly by mylonitic Cranberry Gneiss with a small amount of aegirine granitic gneiss (Figure 3). These landslide features mainly occur within several narrow, northwest-trending valleys drained by first and second order streams. The northwest-trending valleys are perpendicular to the regional gneissic foliation. Many of these debris avalanches follow first order stream courses and do not frequent dip slopes, a fact that indicates lack of bedrock structural control in landslide development. However, their high concentration in the Cranberry Gneiss suggests a higher susceptibility for this lithology with its extensive saprolite overburden. Typically, sporadic first order stream development along these slopes punctuated by debris avalanches suggests that fluvial processed are subordinate to landslide processes. Fluvial processes, however, dominate in lower elevations where larger. Lower-gradient streams with floodplains occur.

Morphology of Sherwood debris avalanches differ fro the more typical type found in central Virginia (Figure 4). Typical Sherwood debris-avalanche features include a singular, narrow, cross-sectionally u-shaped track or chute and a conical shaped toe-debris-pile incised by post movement fluvial erosion down along the toe (Figure 5). Although the latter two features were not discussed by Williams and guy, we examined similar features in central Virginia where incision of toe-debris occurs quickly after formation of the toe-debris pile. The morphological differences which distinguish Watauga County types from those of central Virginia are that Watauga County debris avalanches each have only one main, short track down first order valleys and thus have coalesced and bypassed first and second order valleys. Consequently, the Watauga debris avalanches appear merely to be compressed versions of the typical debris avalanche of central Virginia. However, fluvial incision of the debris pile is common to both younger debris avalanches of Watauga County and the 1969 central Virginia types caused by Hurricane Camille. This incision reveals the typical non-stratified, non-

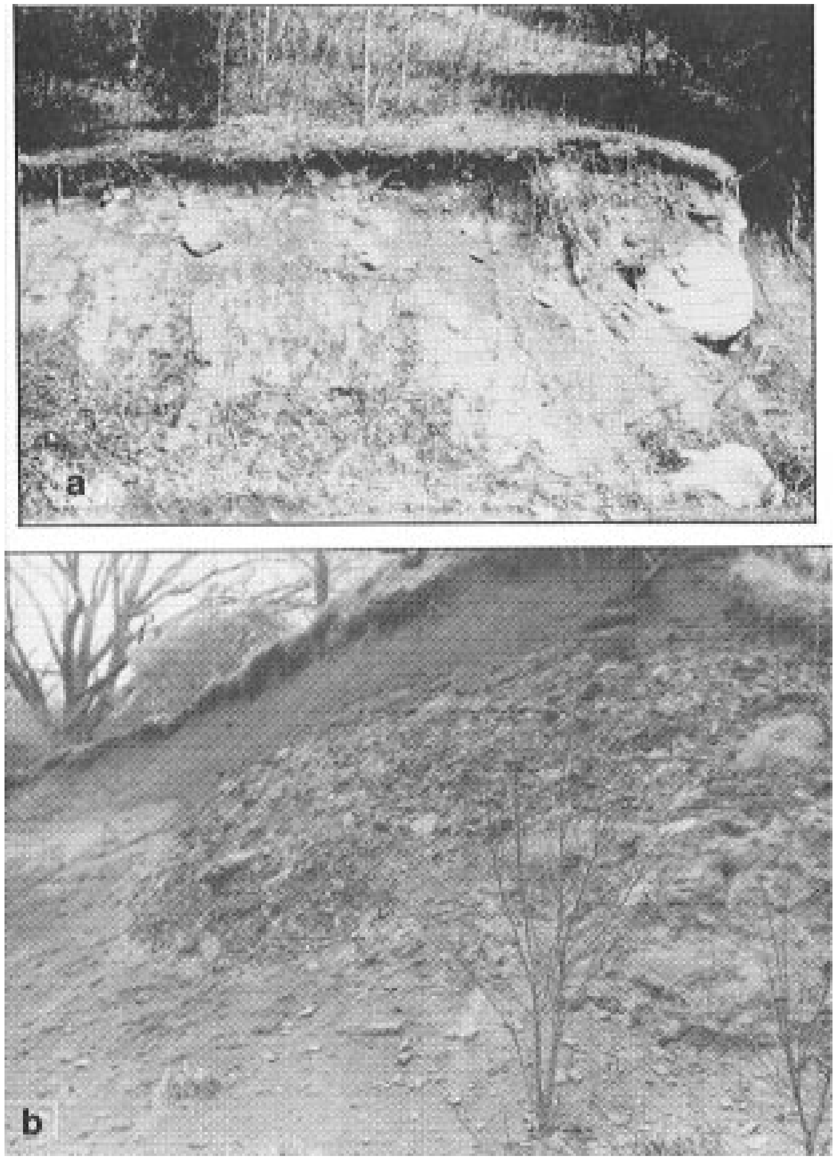


Figure 6. Internal nature of toe debris found among debris avalanches in Watauga County. 6a is along the east side of the Georges Gap road northwest of the Gap in the Sherwood quadrangle and 6b is along the north side of U.S. Highway 421/321 a few miles east of the Sherwood quadrangle. 6a is representative of the more youthful debris avalanche deposits found in the Sherwood quadrangle whereas 6b is typical of older deposits which commonly are highly modified by fluvial processes and soil development.

sorted, bouldery-clay internal nature of the toe debris (Figure 6). Toe incision probably occurs soon after and possible during the waning stages of the debris-avalanche episode while the material is loosely packed and before it is forested. Seven years later in 1976, we noticed approximately 6 feet of incision among toe deposits emplaced by Hurricane Camille debris avalanches. In this same region, we observed older toe deposits which have abandoned channels cut to similar depths. This observation suggests that once channel incision is established, soon after avalanching, it will increase little until the next storm.

Precise activation dates for debris avalanching within Watauga County have not been established. All the well defined debris avalanches in this area are probably Holocene in age. Two relative generations of movement occurred in Watauga County based upon (a) local historical and rainfall records, (b) expression of morphological features, (c) spatial relationships of toe deposits to stream deposits, and (d) degree of post-movement fluvial dissection and reworking of toe deposits. These two generations are: (1) a younger avalanche generation probably activated sporadically from several hundred to several thousand years ago; and (2) an older avalanche generation probably in excess of several thousand years in age. Preserved morphological features are more

Figure 7. Toe deposits of younger debris avalanches in Watauga County. Each show bouldery strewn surface and conical shape. A, c and d are deposits fund in Phillips Branch (see Figure 3 for location), while b occurs in the adjacent Zionville quadrangle. Figure d is another view of the toe deposits shown in c and displays post-movement fluvial incision down along the deposit. This toe also deflects the stream flowing beside the road in the foreground and rests against an older deposit found on the opposite side of the road.

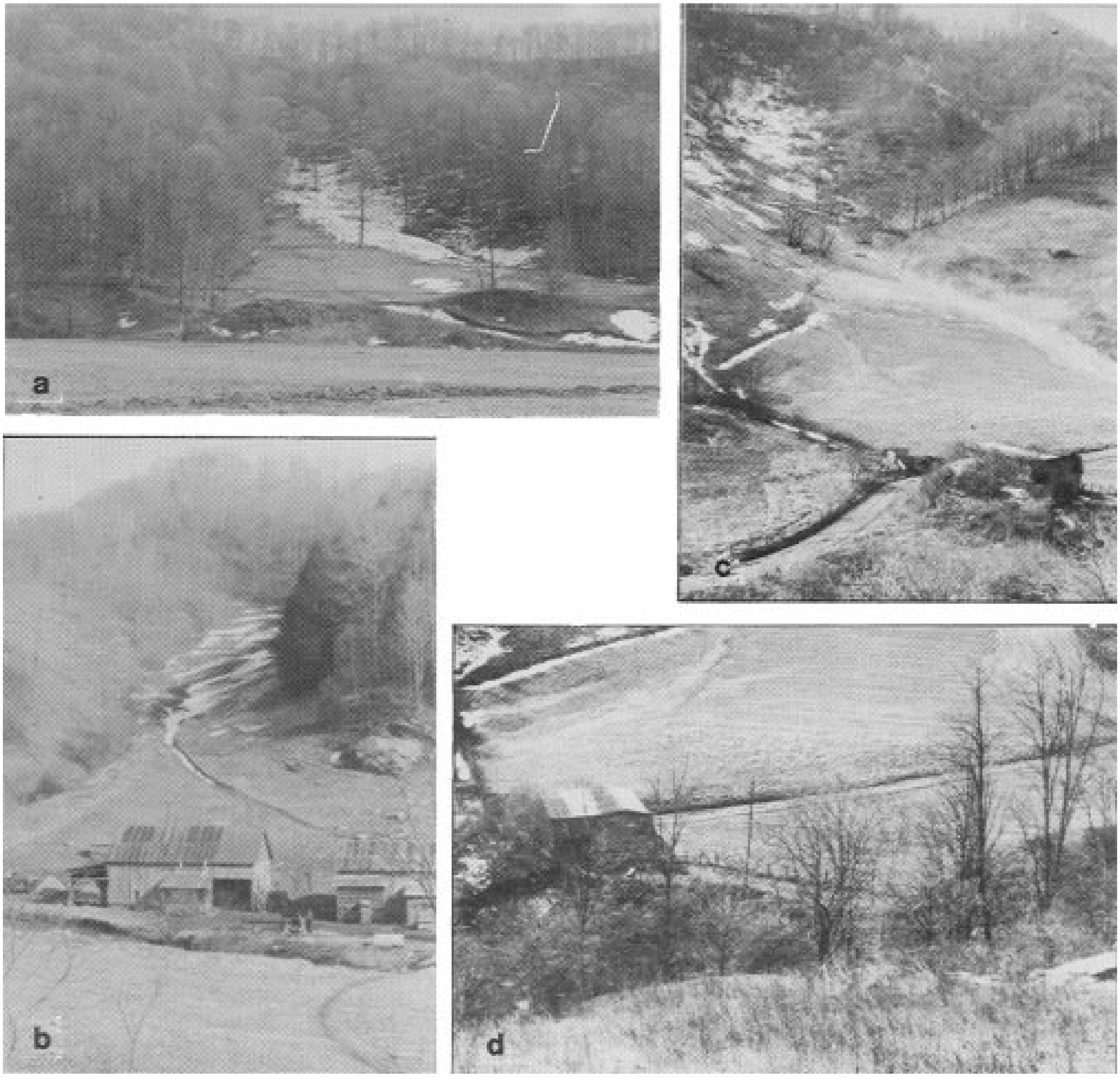


Figure 8. Older debris-avalanche features of Watauga County. A and b occur in the northeastern portion of debris avalanche areain the Sherwood quadrangle; c and d occur in George Gap; e occurs near the town of Bethel, located over 4 miles northwest of George Gap; and f occurs along the east side of Route 421 in the southeast corner of the Sherwood quadrangle. A, b, c, e show fluvial truncation by a valley bottom stream. Photograph d is another view of c and in the foreground shows another toe deposit (boulder by shed) of an older debris avalanche which came down from the opposite side of the valley and has been dissected by the main stream. NOTE: PHOTOGRAPHS E ANF F ARE NOT IN THE ORIGINAL GUIDEBOOK. DH 2-2000.

definitive among the younger deposits and become progressively more subdued with older deposits because of post movement colluvial and fluvial modification. Also, toe deposits of younger debris avalanches, showing post movement fluvial incision along their debris piles (Figure 7), can deflect smaller valley-bottom streams (Figure 7d) and rest either on top of or against toes of older debris avalanches. Toes of older deposits, on the other hand, generally do not show fluvial incision down along their debris piles do show extreme fluvial dissection (Figure 8a, d) and truncation and reworking laterally along their basses by larger valley streams (Figure 8). Local Watauga records indicate that toe

Table 1. Area precipitation events associated with debris avalanche activity. Dashes (---) indicate a lack of data. Not reported here are other intense storms for which no debris avalanches were reported.

Date	Precipitation	Location	Reference
Winter 1901	---	Western N.C.	Holmes, 1917
Summer 1916	10-23 in./15 hrs.	Western N.C.	Holmes, 1917
Summer 1938	12 in./4 hrs.	Tennessee	Moneymaker, 1939
Summer 1940	8-13 in./20-23 hrs.	Western N.C.	U.S.G.S.,1949
Winter 1942	15 in./24 hrs.	Watauga Co., N.C.	---
Summer 1979	---	Va./W. Va.	Hack and Goodlett, 1960
Summer 1969	27-28 in./8 hrs.	Central Va.	Williams and Guy, 1971

incision among these younger avalanches has maintained its present form for more than 150 years. Since this incision probably occurs soon after emplacement of toe debris, and because we did not note any older channels in the central Virginia deposits, these younger Watauga deposits likely resulted from debris-avalanche activation more than 150 year ago. Older debris avalanche features commonly are deeply dissected by first order streams (Figure 8a) and truncated (Figure 8a, b) or incised (Figure 8c, d) by larger streams transverse to debris-avalanche tracks. Modification to crudely sloped terraces (Figure 8a) is less common. Among these older deposits the external form typically is subdued, but the associate wide track heading the conical shaped deposits suggest a landslide origin.

DISCUSSION

Generally, those factors, which rapidly increase external slope shear stress and/or decrease internal slope shear strength, can cause debris-avalanche activity in the area. Debris avalanches typically develop during summer months characterized by periods of intense rainfall. During such precipitation events, slopes lose their shear strength due to a reduction of intergranular friction. Undermining slopes in this area (either naturally by lateral stream incision or artificially by construction) and/or overburdening slope tops (either by construction or increasing soil moisture content by contour farming) could promote premature debris-avalanche activity in the region. The timing and frequency of individual storm events also affect future debris-avalanche activity. Debris avalanches are less likely to occur during winter when rainfall is less and slope materials either more permeable or frozen. Frequent, closely spaced spring or summer storms can gradually increase surficial pore water pressure (especially during the spring thaw), increasing the likelihood of future debris-avalanche activity even if precipitation values are less than amounts received earlier in the season along the same slope or are less that local, recorded rainfall amounts which have been directly responsible for past debris-avalanche activity.

Landslide-types in Watauga County are not restricted to debris avalanches inasmuch as smaller debris slides also occur. Local residents testified that numerous smaller landslides occurred during the winter of 1942 when the area received 15 inches of precipitation in 24 hours. Periods of lesser rainfall in the region have not produced abundant landslides, thus suggesting that at least 15 inches of precipitation occurring over 2 hours are needed to promote minor sliding in the Watauga County are. Considering the above discussion and the precipitation values known responsible for debris avalanches in the region (Table 1), rainfall intensities greater than 20 inches occurring over a 24-hour period could trigger future debris avalanches in Watauga County.

Debris avalanche activity has played an important role in both valley side scour and valley bottom fill in the first and second order drainage basins of Watauga County, North Carolina. Debris avalanches repeatedly have scoured valley sides and subsequently filled valley bottoms with thousands of tons of debris during intense summer storms. Although inter-storm fluvial processes dissected and reworked this valley bottom debris, much of this fill still remains within valley bottoms draining these small topographic basins. However, future valley fill is likely to be less than earlier volumes, since younger debris avalanches appear smaller in magnitude than the older ones. As a result, fluvial processes will become progressively more important because they will scour greater amounts of valley fill from the valley bottoms of the smaller basins into adjacent larger basins.

Future debris-avalanche activity is likely to occur in Watauga County because local geology, topography and climatology favor their development. It would be difficult to predict when and exactly where the next generation of debris avalanches will occur in this region. Certainly if catastrophic rainfalls like Hurricane Camille should strike Watauga County, avalanching of slope debris is likely to occur in, but not necessarily limited to, the same valleys of the Sherwood Quadrangle where they have occurred in the past that is: valleys cutting through the Cranberry Gneiss upon which thick saprolite has developed. Unfortunately, although past debris avalanches occurred in relatively unpopulated areas, today more houses are being built directly on top of or against past debris avalanche deposits as well as on slopes where the saprolite is easily excavated. Therefore, future debris avalanched in this region will probably cause considerably more property damage and loss of life.

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PRELIMINARY RESULTS FROM DETAILED GEOLOGIC MAPPING STUDIES IN THE WESTERN SAURATOWN MOUNTAINS ANTICLINORIUM, NORTH CAROLINA

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ABSTRACT

Geologic mapping in the southwestern portion of the Sauratown Mountains anticlinorium basement massif has yielded some interesting preliminary results. A sequence of Grenville basement (?) orthogneiss is overlain by high grade (amphibolite facies) mylonitic cover metasedimentary rocks suggesting that the cover may be thrust over the basement here. These rocks either grade stratigraphically upward, or are separated by a higher premetamorphic thrust or an unconformity from a kyanite or lower grade assemblage of metasandstone, quartzite, pelitic schist (graphitic to the northwest), and amphibolite (presently called Ashe Formation) toward the northwest and southeast. Cover sequences are compostionally somewhat different on the NW and SE flanks of the anticlinorium suggesting that if they are the same, considerable shortening has occurred. However, these sequences may be connected around the SW end of the anticlinorium. The Inner Piedmont boundary in this area, like many others, appears to be a metamorphic gradient, not a tectonic contact. The same rocks of the Ashe (Tallulah Falls) Formation are present here, only at higher grade. All the rocks and contacts in this area are polydeformed in the usual sequence of two or more sets of isoclinal folds overprinted by crenulation cleavage, then more brittle structures. The entire sequence is cut by brittle Mesozoic faults (Sony Ridge fault zone with small displacement) and diabase dikes.

INTRODUCTION

The purpose of this study is to present some preliminary results of a field investigation in progress in the western portion of the Sauratown Mountains anticlinorium in North Carolina (Fig. 1). The area being studied is a roughly 15' area northwest of Winston Salem, North Carolina, which has not been studied in detail before. The area has been the subject of reconnaissance work in the past (Dunn and Weigan, 1969; Espenshade and others, 1975). Lewis (1980) made a more detailed study of the Elkin North and Copeland quadrangles to the west of the present study area.

Fullagar and Butler (1980) summarized the radiometric age data for the Sauratown Mountains anticlinorium. Our preliminary studies indicate that a great deal more systematic age dating must be carried out before any firm conclusions may be drawn about which rocks are older basement and which are younger cover rocks.

ROCK UNITS

The rock units present in this area include orthogneisses, which are in part basement, but are also in part metamorphosed Paleozoic granitic plutons. He basement rocks are overlain by a cover sequence of metasedimentary rocks which are interpreted by other workers as basement metasedimentary rocks, along with a sequence belonging to the Ashe (Tallulah Falls) Formation of presumed late Precambrian age.

The basement rocks consist of a group of orthogneisses composed of feldspar, quartz, and varying amounts of biotite which were polydeformed and contain anatectic masses of quartzofeldspathic material. Basement orthogneisses appear to have been derived from metamorphism and deformation of a series of plutonic bodies. Their assignment to the basement is based upon the complexity of deformation to which they appear to have been subjected and the age dates fro the Pilot Mountain Quarry (Fullagar and Butler, 1980). These at present include rocks called Crossnore by Espenshade and others (1975) and Lewis (1980).

Metasedimentary sequences which appear to lie above the orhogneiss basement are separated based upon their lesser degree of deformation and their differences in composition. The metasedimentary rock units are obviously more uniformly layered and contain higher quartz and mica contents with a correspondingly lower percentage of feldspar. Amphibolites and some ultramafic rocks occur within the metasedimentary sequence. Ultramafic rocks are commonly associated with the amphibolites.

It is possible to subdivide the metasedimentary unit into an upper amphibolite, ultramafic rocks, aluminous schist-bearing sequence and a monotonous lower sequence of metagraywacke interlayered with muscovite-biotite schist. The former is approximately the Ashe Formation of Espenshade and others (1975) and the latter corresponds to their basement metasedimentary rock sequence. The lower sequence also contains some marble. One possible explanation of this relationship is that the lower sequence is attached to the orthogneiss basement and has a continental source

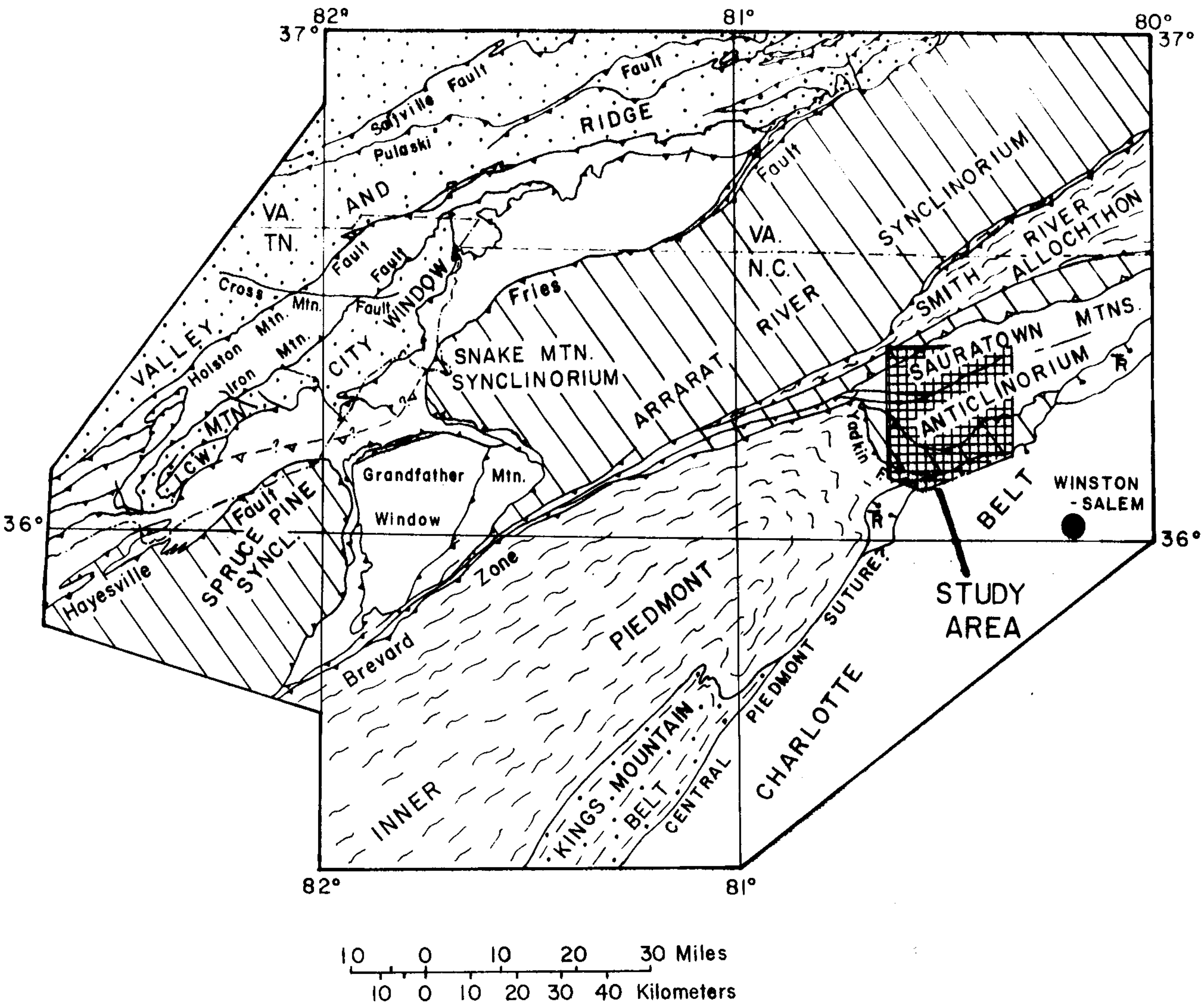

Figure 1. Map showing the major tectonic features of northwestern North Carolina, northeasternmost Tennessee and part of Virginia, as well as the study area. LCW - Limestone Cove window. TR - Triassic-Jurassic sedimentary rocks.

whereas the upper sequence was deposited on oceanic crust and now is separated from the lower sequence by a premetamorphic thrust (Hayesville?) which juxtaposed the two units subsequent to their deposition.

The upper sequence (Tallulah Falls/Ashe) is present in the Inner Piedmont as a higher grade (upper kyanite) assemblage of metagraywacke, amphibolite and aluminous schist. The metagraywackes are more feldpathic here but aluminous schists which are traceable through part of the Inner Piedmont in this area become more interlayered and indistinct as separate units along strike. They are still part of the same original assemblage. Rocks originally mapped here as Henderson Gneiss by Espenshade and others (1975) are metagraywacke layers containing coarse feldspar porphyroblasts (or porphyroclasts). WE therefore find no evidence for a "Yadkin fault."

A feldspar-quartz-biotite augen orthogneiss occurs concordantly within the sedimentary sequence. Additional foliated granitic bodies composed of more even grained feldspar-quartz-(mica) gneiss appear to be material which was intruded and then deformed and metamorphosed, possibly late in the deformational sequence, since these rocks do not exhibit the multiple fabrics observed in the adjacent metasedimentary rocks.

These rocks were intruded by Mesozoic diabase dikes which range up to 7 m. thick. Dike orientations are NS, EW and NE.

STRUCTURE

The rocks of this area were subjected to several deformational events beginning before Paleozoic regional meta-

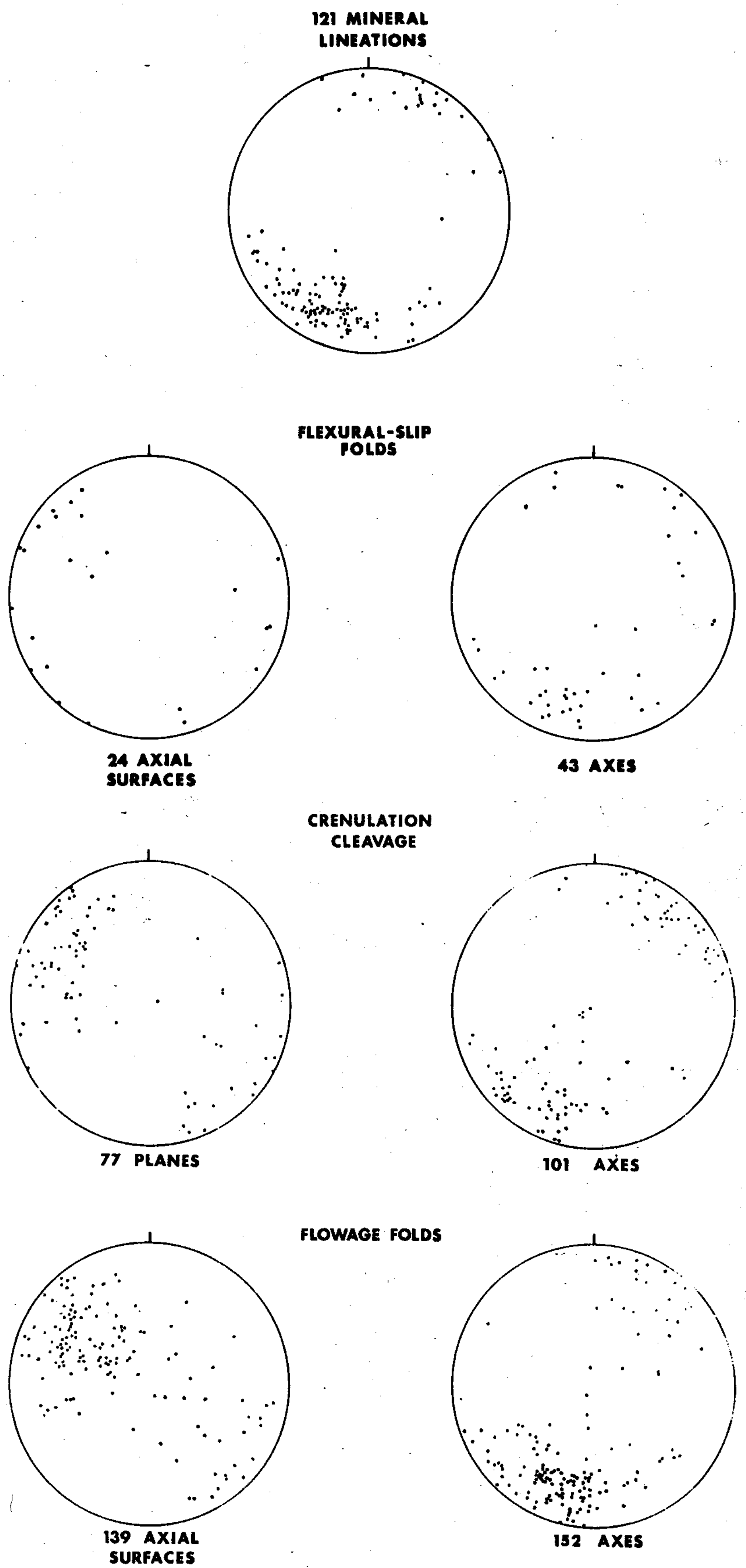

Figure 2. Equal area plots of linear fabric elements from the western Sauratown Mountains area.

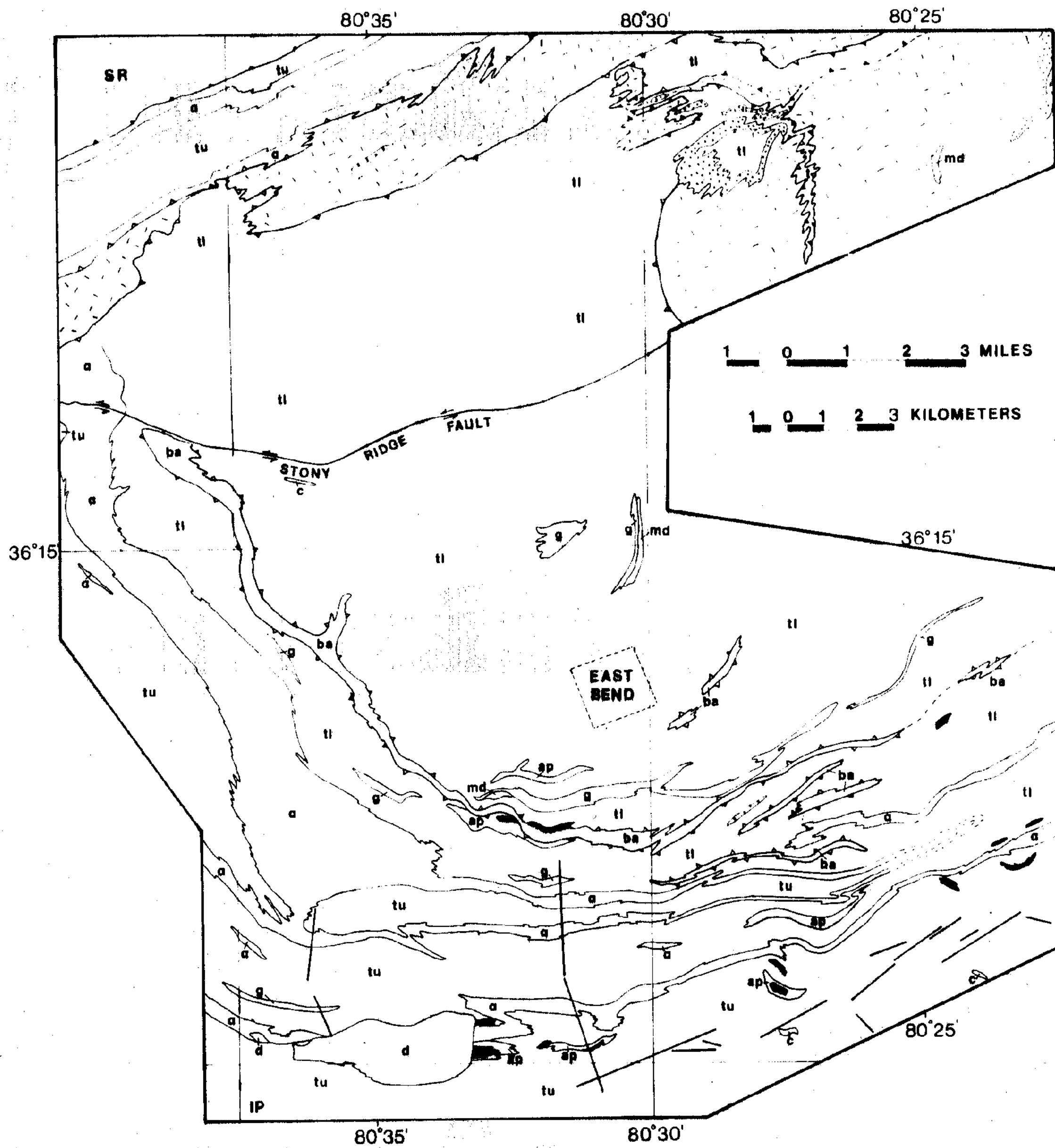

Figure 3. Preliminary geologic map of the western Sauratown Mountains anticlinorium. The area west of 80°37'30" is modified slightly from Lewis (1980). Basement orthogneiss is cross-hatched. Ultramafic rocks are black. Quartzite is indicated by a stipple pattern. tl – metagraywacke, schist, minor amphibolite, lower sequence. tu – metagraywacke, schist and amphibolite, upper sequence. ap – amphibolite. a – aluminous (kyanite and/or staurolite-bering) schist. c – carbonate rocks. g – granitic gneiss. md – metadiorite. d – diorite. ba – biotite augen gneiss (basement). Heavy straight lines are Mesozoic diabase dikes. SR – Smith River allochthon. Solid teeth on contacts indicate known fault. Open teeth on contacts indicate speculated fault. IP – Rocks of the conventionally known Inner Piedmont (actually part of the upper sequence).

morphism and continuing into at least the Mesozoic. A set of fabric diagrams showing the relationships among several different fabric elements which were measured thus far is shown in Figure 2.

The rocks possess a dominant foliation on which most of the foliation measurements were made. Additionally, older foliations were observed and measured in a few places but for the most part these were transposed by the younger, dominant foliation. The latter is overprinted by at least one crenulation cleavage which locally began to transpose some of the earlier foliation into a new orientation. The crenulation cleavage is associated with open to tight folds whose axial surfaces dip moderately to the southeast and occasionally to the northwest.

The contacts between the metasedimentarty sequences and the orthogneiss units have a strong mylonitic character where they have not been obliterated by intrusives. One possible explanation then is that the lower metasedimentary sequence is overthrust onto rocks of the craton. Also, the cover sequence appears to have not reached the uppermost grades of the amphibolite facies or the granulite facies during metamorphism.

INTERPRETATION

The pattern of distribution of rock units on the map (Fig. 3) may be interpreted to indicate rocks were subjected to at least one event of isoclinal folding during the Paleozoic, then refolded several times by more open folds. Truncation of units in the metasedimentary cover sequence by the augen gneiss unit indicates either the augen gneiss was intruded into the sequence as sills, then multiply deformed, or, the cover sequence was thrust over the augen gneiss and the linear outcrop pattern of the augen gneiss is a series of elongate windows. If the latter is the correct interpretation, the augen gneiss may be a basement unit. However, this can be determined only by additional radiometric age dating studies.

The possibility of two metasedimentary sequences juxtaposed by a premetamorphic thrust was discussed above. The argument that the lower metasedimentary unit is part of the basement is favored by relationships in the Pilot Mountain Quarry where a metasedimentary unit has been intruded by granitic rocks that probably yielded pre-Paleozoic ages. However, the complex relationships and alteration of rocks dated as basement in the quarry raise questions about the validity of any of the age dates and whether specific lithologies can on radiometric bases be called basement. It is also likely that zircons from this locality are inherited.

If could also be argued that the lower sequence is a basement unit because it appears to be more deformed than the sequence above (Lewis, 1980). This is a correct approach because the basement should contain a greater number of deformations than the cover. However, it is very difficult to prove, at least here, because the character of the sedimentary rocks in the upper sequence is different from that in the lower. The metasandstone and schists of the lower sequence contain better strain markers than the rocks of the upper sequence. Consequently, this criterion for recognition of basement may be difficult to use successfully here. We regard the question unresolved at present.

Faulting occurred after all the folding and other deformational events forming the Stony Ridge fault zone (Butler and Dunn, 1968). This fault is primarily a fissure having little displacement. Its multiple history of reactivation is recorded in the siliceous cataclasite found along its outcrop extent. However, Lewis (1980) indicates the fault could have at least 2km of apparent left lateral displacement (Fig. 3).

Our interpretation of this area involves initial deposition of the metasedimentary cover sequence partly on oceanic crust (represented only as fragmented pieces of ultramafic and mafic rocks) and partly on continental crust. A large premetamorphic thrust formed which emplaced the sequence which was deposited on oceanic crust onto the exposed basement and autochthonous cover. Then the whole sequence was isoclinally folded. Later folds were formed which served to modify the outcrop patterns and create some of the interference patterns which exist on the map.

Later fracture and fissure systems, such as the Stony Ridge fault zone, developed during the Mesozoic but did not serve to offset the existing contacts to any great degree. The motion on this fault was primarily extensile so that, through a series of stages, opening produced some minor offset of contacts. However, it primarily served as a conduit syste into which silica-bearing fluids were emplaced, solidified, then were broken again as the fault was reactivated. Toward the southeast and northeast, Triassic/Jurassic normal faulting produced small basins which received terrigenous sediments and serve as indicators of later tectonic activity, again though of an exsional or possibly strike-slip character.

ACKNOWLEDGMENTS

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DISTRIBUTION AND RELATIONSHIPS OF LATE PRECAMBRIAN AND UNDERLYING GRENVILLE(?)-AGE ROCKS, SAURATOWN MOUNTAINS AREA, NORTH CAROLINA

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INTRODUCTION

Data and ideas presented in this paper derive heavily from the mapping of two 7.5 minute quadrangles, the Copeland and Elkin North (Appendix I – plates 1 and 2, respectively) in Surry, Wilkes, and Yadkin Counties in the northwestern North Carolina Piedmont. An additional 24 km^2 in the southeastern quarter of the Dobson quadrangle (figure 1) also was mapped in detail; and this geologic mapping is incorporated in small-scale regional maps used in this paper. The recently completed study of Grenville basement rocks of the southern Appalachian Orogen by Bartholomew and Lewis (in press) and the study of basement-cover rock relationships near Boone (Bartholomew and others, this volume) also contributed to ideas expressed in this paper.

Detailed mapping originally was begun by the writer (Lewis, 1980a) in the Copeland quadrangle in order to investigate the nature of the relationship between the Smith River allochthon of Conley and Henika (1973) and the Brevard Zone (Lewis, 1980b). Reconnaissance mapping in surrounding areas led to additional detailed mapping to the west in the Elkin North quadrangle. This paper is intended to complement previous work by presenting some regional stratigraphic and structural relationships in the southwestern Sauratown Mountains anticlinorium. Comparisons of the Ashe Formation in this area with the Ashe Formation in its type area in Ashe County is considered in light of the emphasis of the field trip. The discussion of basement-cover rock relationships also is included because of the significantly different interpretations being advanced by Hatcher and others (this volume) based on their very recent, preliminary work in the region to the southeast of the Copeland quadrangle.

Previous regional studies and reconnaissance geologic mapping includes work by Butler and Dunn (1968) in the Pilot Mountain area of the Sauratown Mountains anticlinorium, a study along the Brevard zone by Justus (1971), and

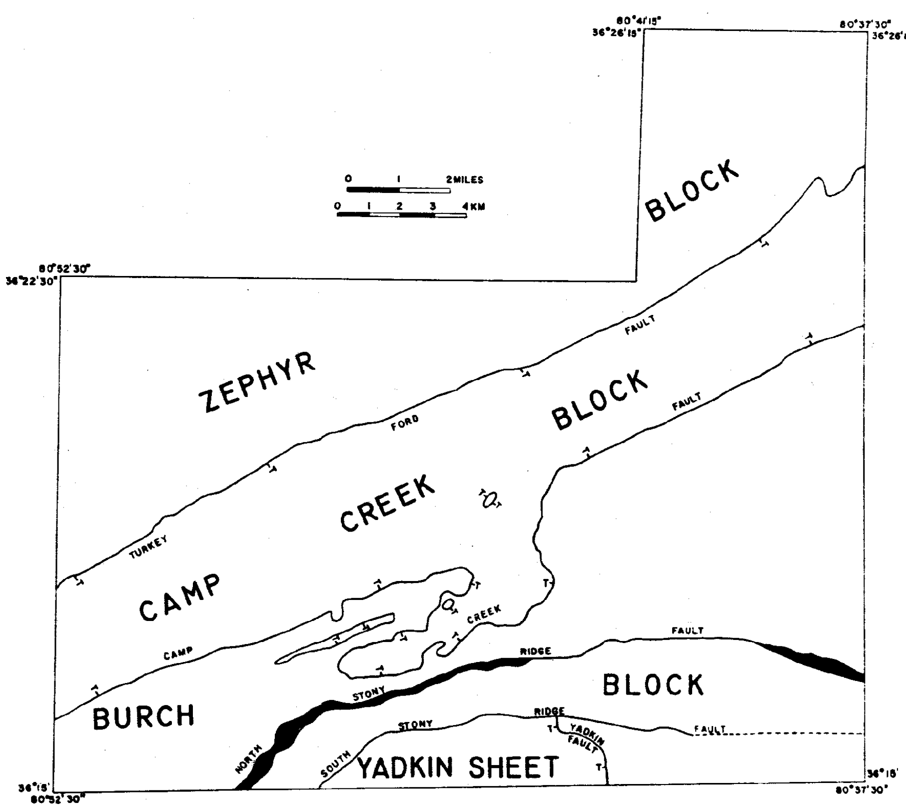

Figure 1Generalized tectonic map of study area showing major faults and structural blocks. T – on upper thrust plate.

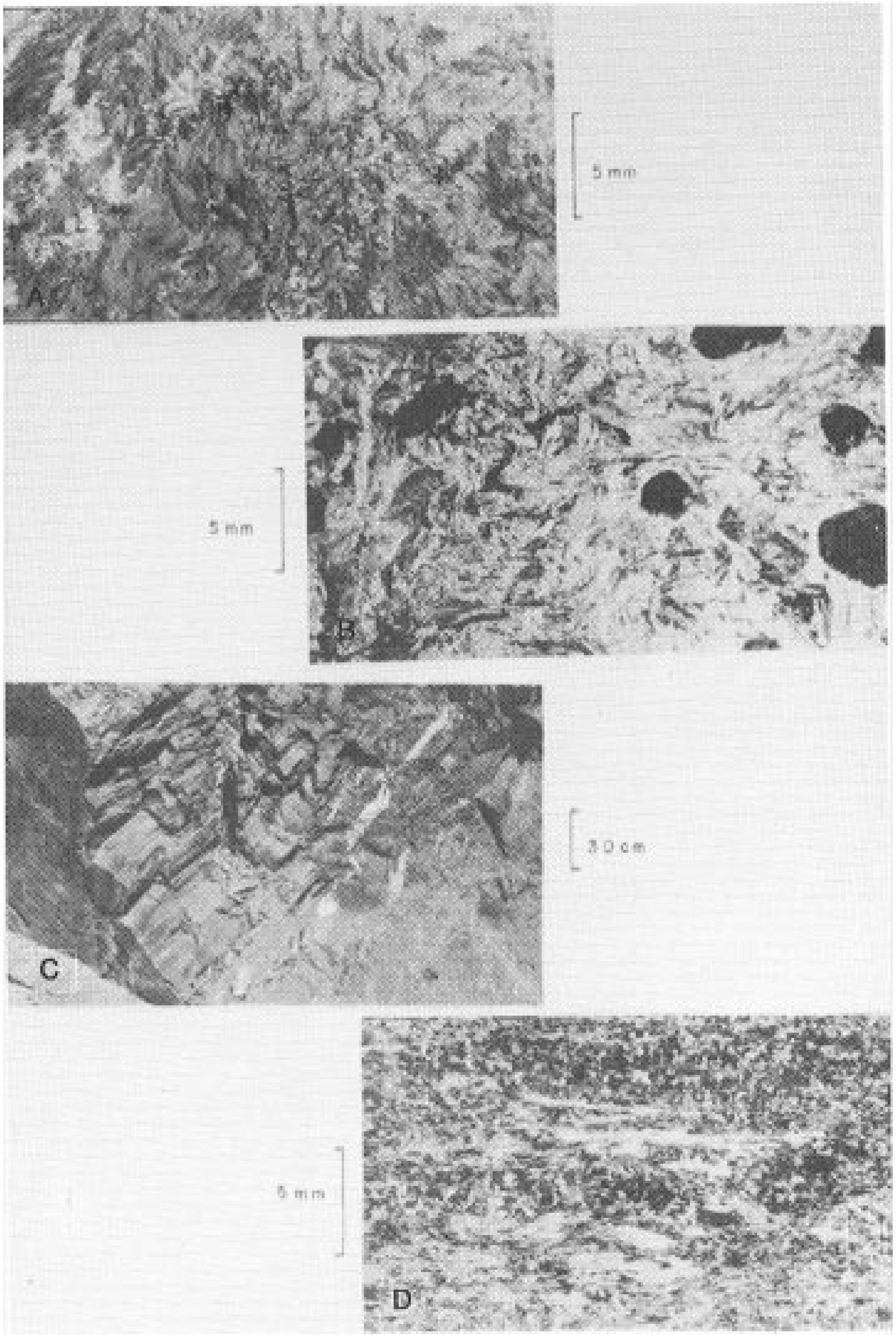

Figure 2. A - Photomicrograph of coarse muscovite and quartz schist associated with ggms-1 unit. As is typical of this litholog , throughout its sporadic occurrences, mineralogy consists essentially entirely of muscovite, quartz and lesser amounts of opaques.
B - Photomicrograph of garnet-muscovite schist of ggms-1 unit. Muscovite foliation is highly contorted. Garnets exhibit varying degrees of post M_1 distortion.
C - Biotite gneiss (bg-2). Outcrop at Fisher River bridge at intersection of SR 2221 and SR 2227. Pencil at center of photograph is approximately parallel to F_1 fold axis.
D - Photomicrograph of biotite gneiss (bg-2).As exhibited by this photograph, coarse muscovite commonly parallels the biotite foliation. Quartz and palgioclase (An_{24}) commonly form distinct layers.

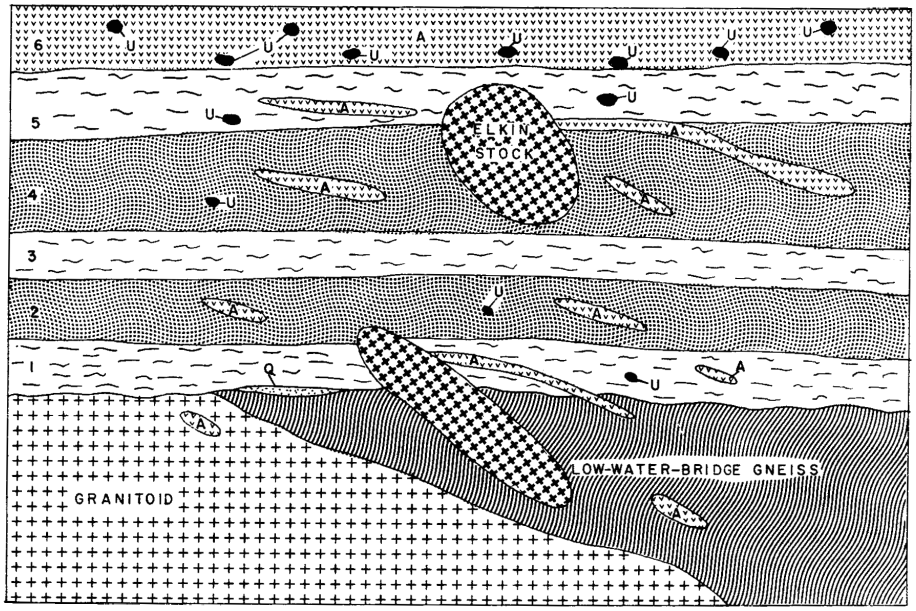

Figure 3. Diagrammatic sketch showing relationships among units of the basal Ashe Formation, mafic intrusions, granites of the Crossnore Suite and Grenville-age rocks in the Sauratown Mountain area. Units 1, 3 and 5 are pelitic (garnet mica) schists, units 2 and 4 are biotite gneisses; unit 6 is amphibolite; A-amphibolite; q-quartzite; u-ultramafic.

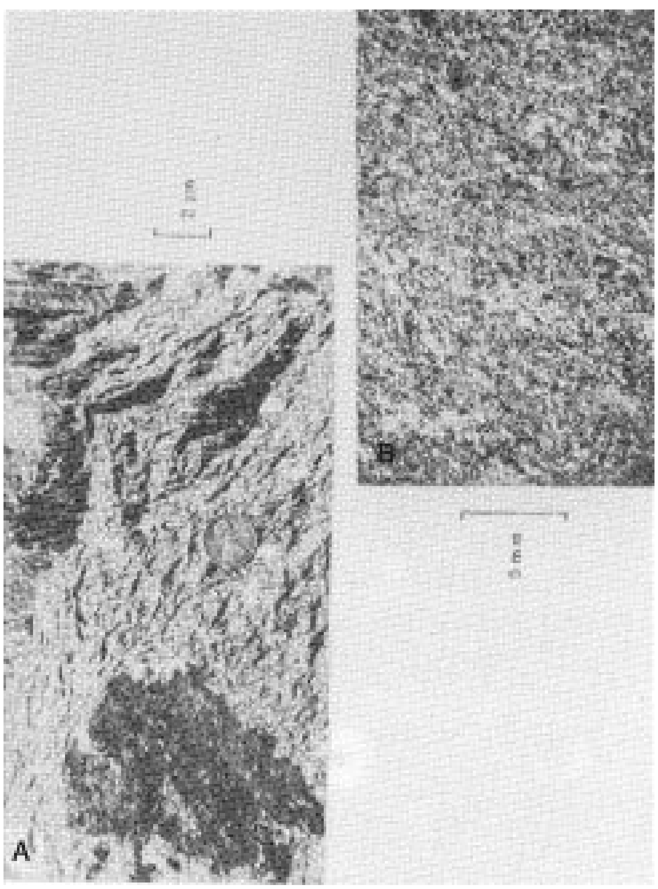

Figure 4. A - Outcrop at Elkin Quarry of "Crossnore" lithology. Note mafic xenolith at left of penny and "rodding" effect formed by intersection of layering and later foliation.
B - Photomicrograph of quartz-monzonitic "Crossnore" lithology. Mineralogy is quartz, K-feldspar, alkali amphibole, plagioclase with minor fluorite as an accessory

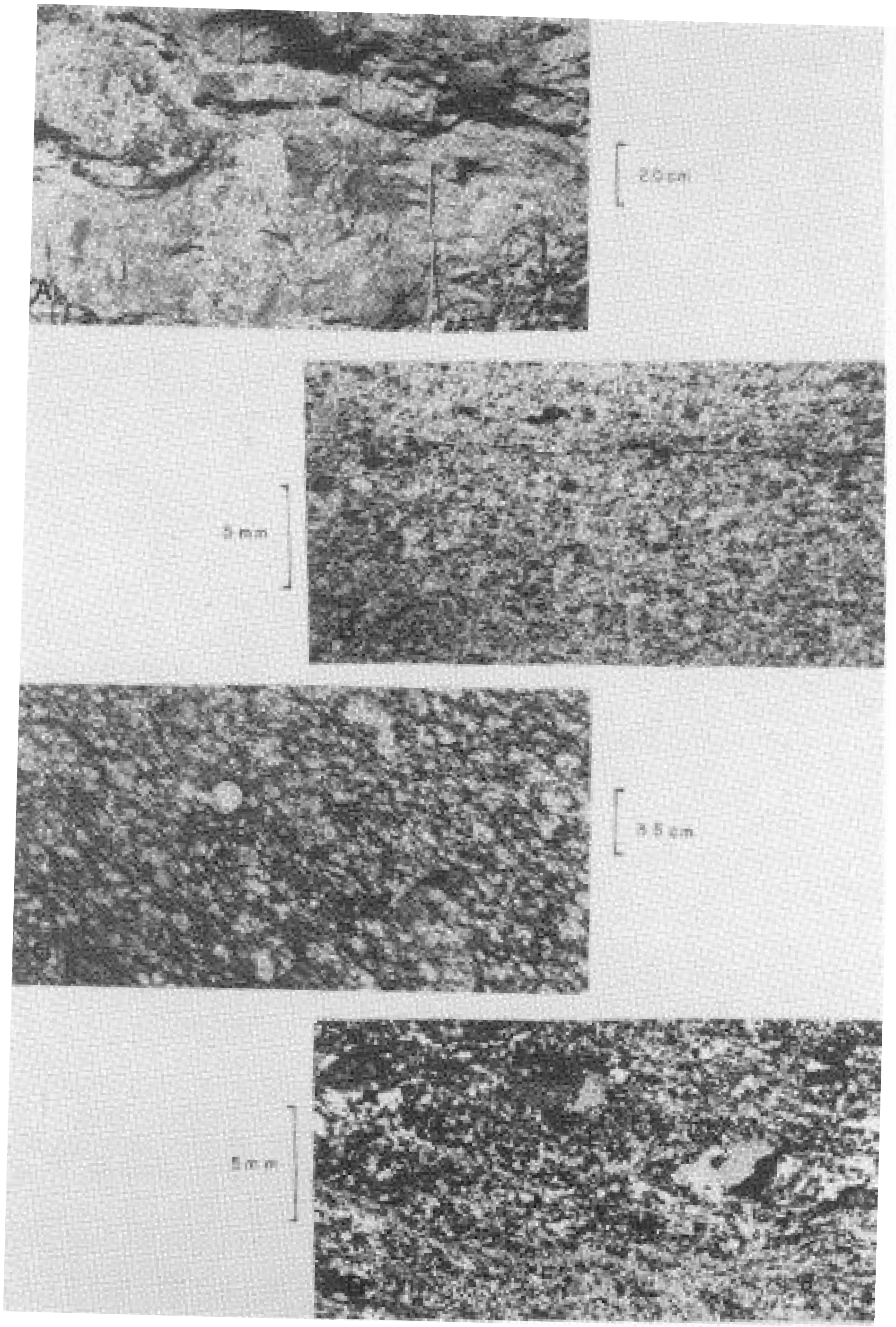

Figure 5. A - Layered biotite gneiss. Outcrop on southeast side of Yadkin River,one-half kilometer northeast of the Rockford bridge along dirt road. Dime in center of picture rests on folded quartzofeldspathic layers. Close-up of folds is shown in Figure 7A.
B - Photomicrograph of mylonitic layered biotite gneiss. Dominant mineralogy is biotite, muscovite, quartz, K-feldspar, plagioclase feldspar and epidote.
C - Augen gneiss. Outcrop of augen gneiss on railroad track two kilometers northeast of Rockford. One to four cm K-feldspar augen occur with 5 mm to 1 cm blue quartz augen.
D - Photomicrograph of augen gneiss.Large augen of perthitic K-feldspar dominate in a matrix of biotite, smaller quartz augen, muscovite and minor chlorite.

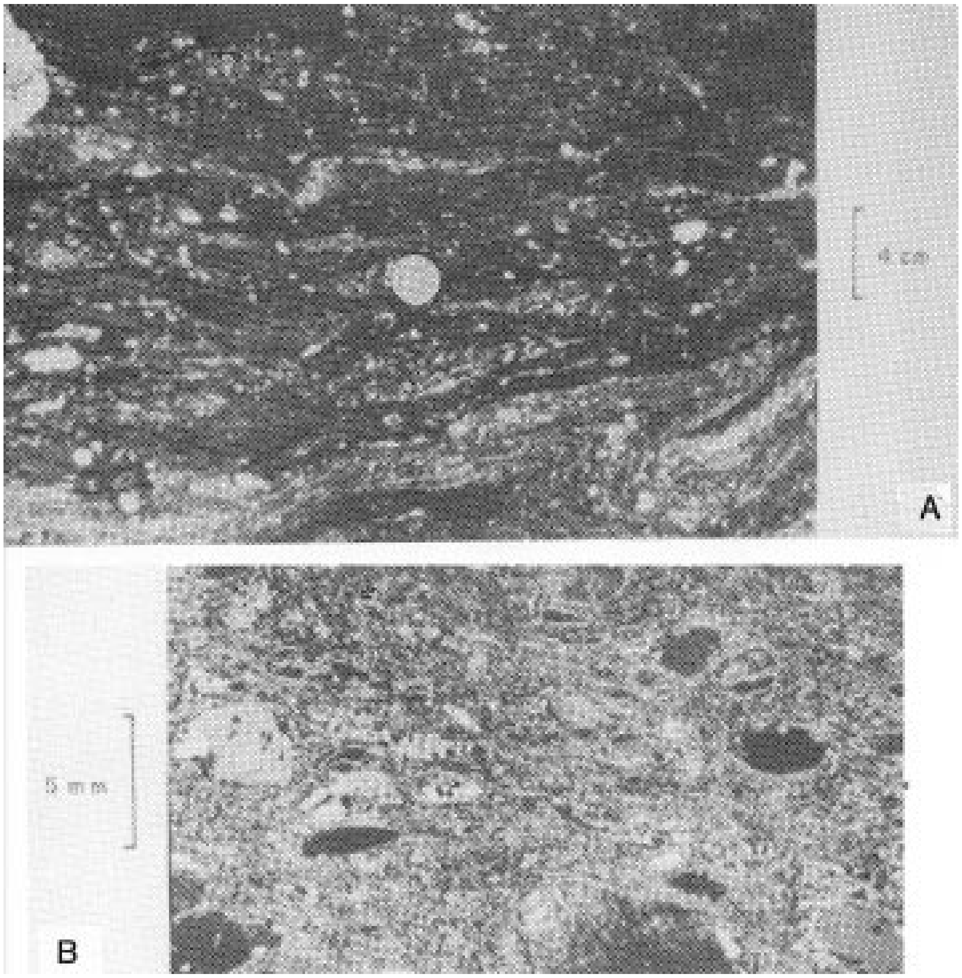

Figure 6. A - "Henderson" gneiss. Outcrop of biotite-rich "Henderson" gneiss with "floating" K-feldspar "augen." Note F2 fold in lower right corner. Outcrop approximately 1 km south of Crutchfield on U.S. 601.
B - Photomicrograph of mylonitic "Henderson" gneiss. Small, rectangular late biotite, typical of this lithology, is visible superimposed on some feldspar grains.

mapping of the eastern half of the Winston-Salem 2^o sheet by Espenshade and others (1975). The Smith River allochthon was identified and named by Conley and Henika (1973) and was extended into North Carolina by Espenshade and others (1975). Dunn and Weigand (1964) mapped the Pinnacle and Pilot Mountain quadrangles and J.R. Butler has done detailed mapping (unpublished) in the central Sauratown Mountains anticlinorium and one of his students (Centini, 1968) did some detailed mapping in the Hanging Rock area. Regional structural analyses which concentrated on the northeastern end of the Sauratown Mountains anticlinorium were done by Stirewalt and Dunn (1973). Roper and Justus (1973) collaborated on a structural study along the northeastern portion of Brevard zone. A ^{207}Pb – ^{206}Pb zircon age of 1192 m.y. (Rankin and others, 1973) on a granitoid body near the town of Pilot Mountain, North Carolina is the oldest age obtained for any of the basement rocks of the Sauratown Mountains anticlinorium. Fullagar and Butler (1980) did Rb-Sr model age determinations for three Crossnore-type granitoid plutons (572 m.y., 627 m.y. and 674 m.y.), some schists (354 m.y.) and metaquartzites (281 m.y.) as well as an isochron age for microbreccias along the Stony Ridge fault zone (180 m.y).

Although in the Piedmont physiographic province, the rocks of the southwestern Sauratown Mountains anticlinorium have lithologic affinities to the late Precambrian metasedimentary/metavolcanic rocks, as well as to both older layered gneisses and younger intrusive Grenville-age basement rocks, of the Blue Ridge geologic province. Granitoid orthogneisses included (Espenshade and others, 1975) in the Crossnore Suite (Bartholomew and Lewis, in press; Rankin and others, 1973) intrude the lowermost metasedimentary rocks of the Ashe Formation as well as the crystalline basement rocks. Bartholomew and others (this volume)

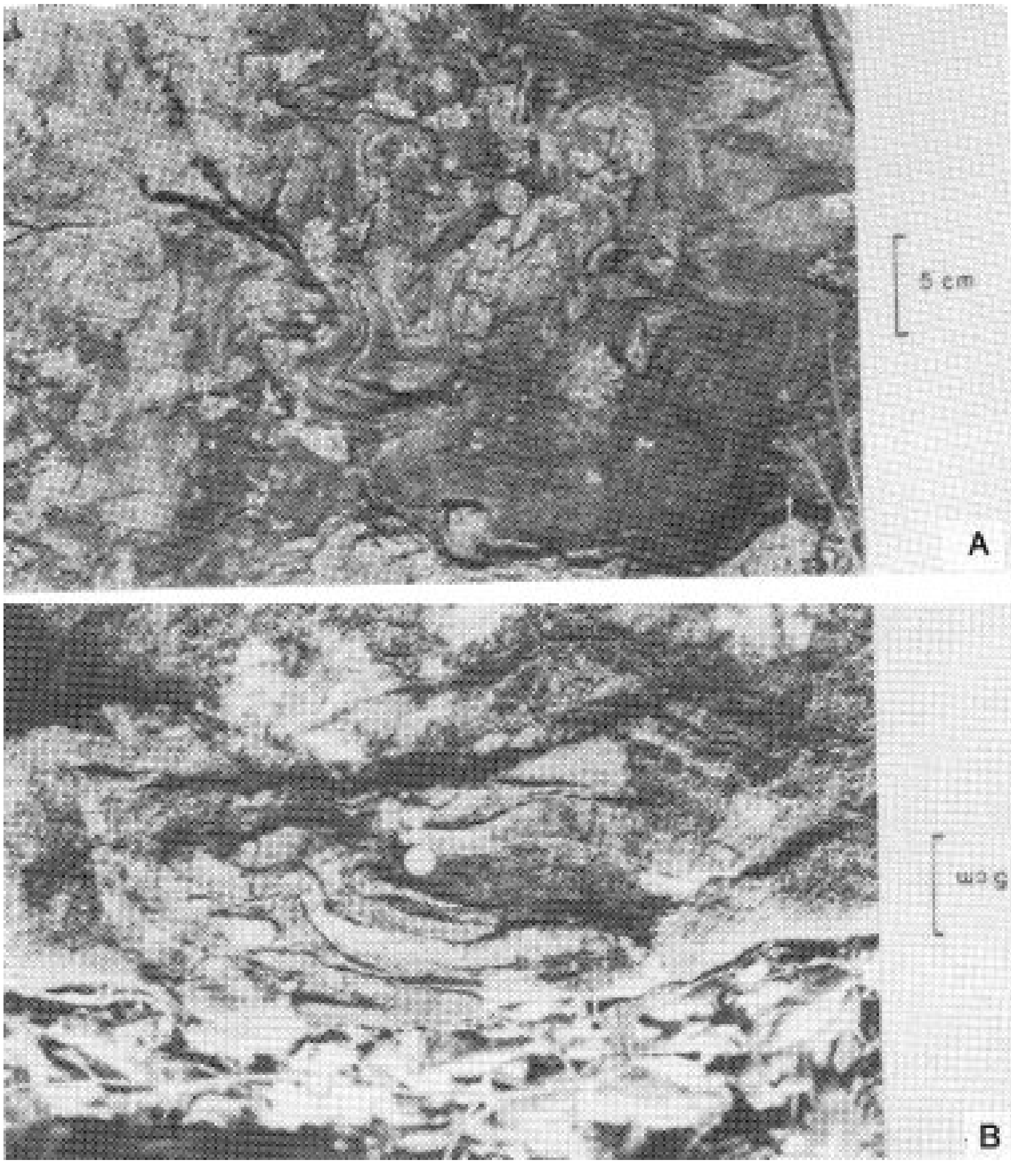

Figure 7. A - Close-up of outcrop shown in Figure 5A showing interference pattern of folds in gneiss near Rockford bridge. Ramsay Type 3 interference pattern, produced by less than 30^0 angular difference between respective axial surfaces of two different folding events. Approximately 3 cm diagonally below the quarter is the nose of a possible pre-F_1 (Grenville?) fold. The lithology is the layered biotite gneiss (lbg).
B - Isoclinal folds in biotite gneiss. Small-scale nappe-like F_1 isoclinal folds outlined by quartzose layers within a biotite gneiss. The outcrop is located in a stream valley in the southwestern Copeland quadrangle approximately 2 km east of SR 2230 and 0.5 km north of the railroad.

discuss in more detail the stratigraphy of the Ashe Formation in the northwestern North Carolina Blue Ridge and Bartholomew and Lewis (in press) have outlined both a large scale regional interpretation and nomenclature for Grenville crystalline rocks in northwestern North Carolina, Virginia, and Maryland.

ASHE FORMATION

The most complete mappable stratigraphic sequence of the Ashe Formation in the area mapped by Lewis (1980a) occurs southeast of the Camp Creek fault (Figure 1; Appendix – Plates 1 and 2) and northeast of the layered biotite gneisses and granitoid gneisses of the basement core of the Sauratown Mountains anticlinorium. A metaquartzite similar to the one which caps the Pilot Mountain topographic feature rests nonconformably on the basement core and is overlain by various interlayered garnet-bearing pelitic schists (Figure 2)and a few calc-silicates, biotite gneiss/metagraywackes

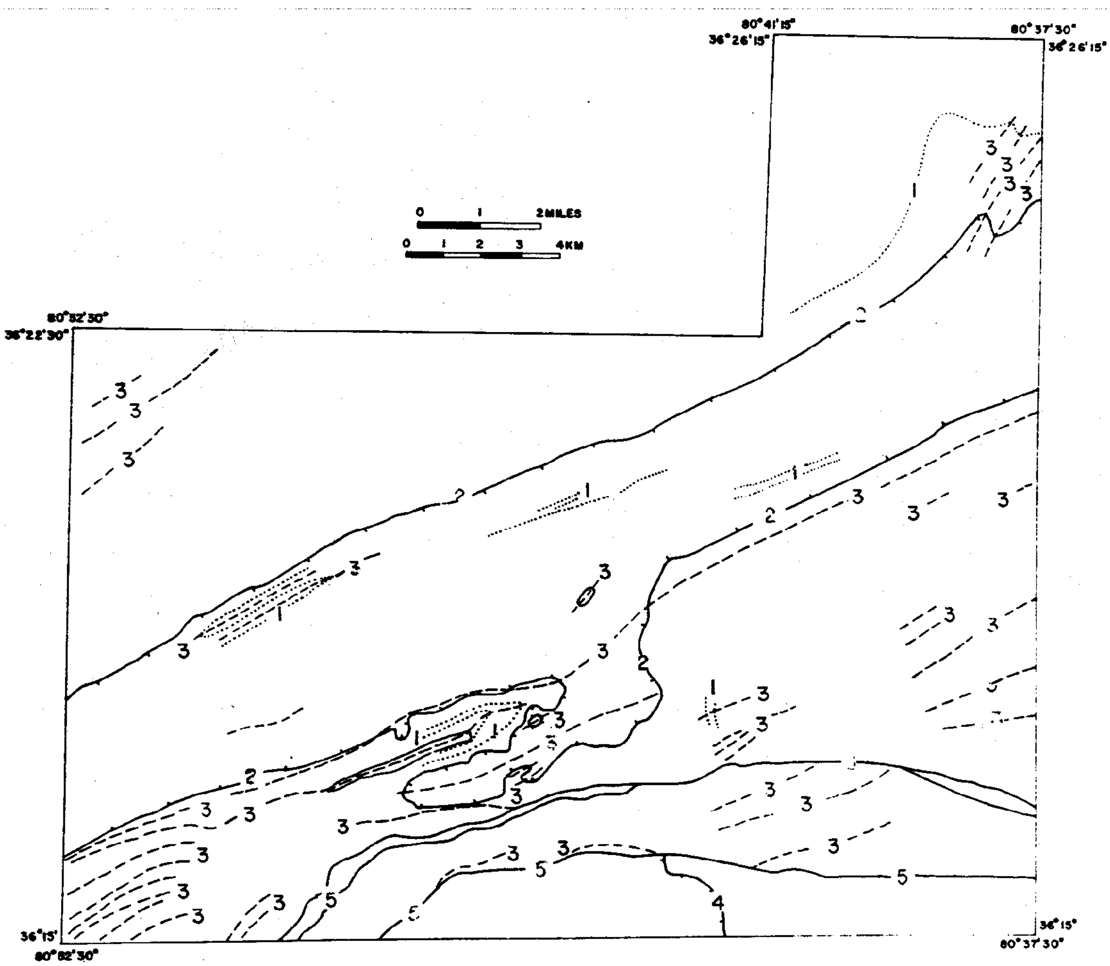

Figure 8. Diagrammatical representation of F_1 and F_2 fold axial traces within the study area.
….. F_1 fold axis ----- F_2 axis

(Figure 2). Metavolcanic rocks (amphibolites) are uncommon within this sequence until the uppermost unit (a-6, Plates 1 and 2) which is truncated by the Camp Creek fault (Figure 1; Plate 2). This unit (a-6) contains predominantly amphibolite with some interlayered mica schist and numerous pods of serpentinized to unaltered dunite. This mafic flow (or sill) probably also marks the time of intrusion of the numerous mafic dikes which are found in both basement and the cover rocks below the mafic flow (or sill).

A number of similarities exist between rocks of the Ashe Formation in the southwestern Sauratown Mountains area (Figure 3) and the Ashe Formation in its type area (Bartholomew and others this volume). Both begin at the uncomformity with rocks of probable shallow water origin which include metaquartzite (possibly bar sands), calc-silicate biotite gneisses and lensoidal marbles (possibly *Archeocyathid*-type reefs). The metaquartzites appear to be laterally more extensive in the Sauratown Mountains anticlinorium (see, for instance, Dunn and Weigand, 1974; Centini, 1968) where they crop out mostly as erosional remnants above, and surrounded by, crystalline rocks of the basement. In the Copeland quadrangle (Plate 1) small metaquartzite outcrops occur both as an isolated erosional remnant and as a small metaquartzite at the base of the continuos metasedimentary sequence described above.

Whereas following deposition of the metaquartzite in the type area of the Ashe Formation, the metasedimentary rocks were engulfed by the thick mafic volcanic (amphibolite) sequence of the Snake Mountain volcanic pile (Bartholomew and others, this volume), in the southwestern Sauratown Mountains area, sedimentation continued with alternating pelitic and metagraywackes (?) until emplacement of the amphibolite (a-6) which is the uppermost unit in the continuos sequence. Possibly this area was farther east of the rifting center (but both were on continental crust) which was sited in Ashe County; hence, volcanism in the Sauratown Mountains area begins at a higher relative stratigraphic level. IN the Elikin area, the lowermost Ashe Formation, as well as the basement core of the Sauratown Mountains anticlinorium and, indeed the unconformity itself, are intruded

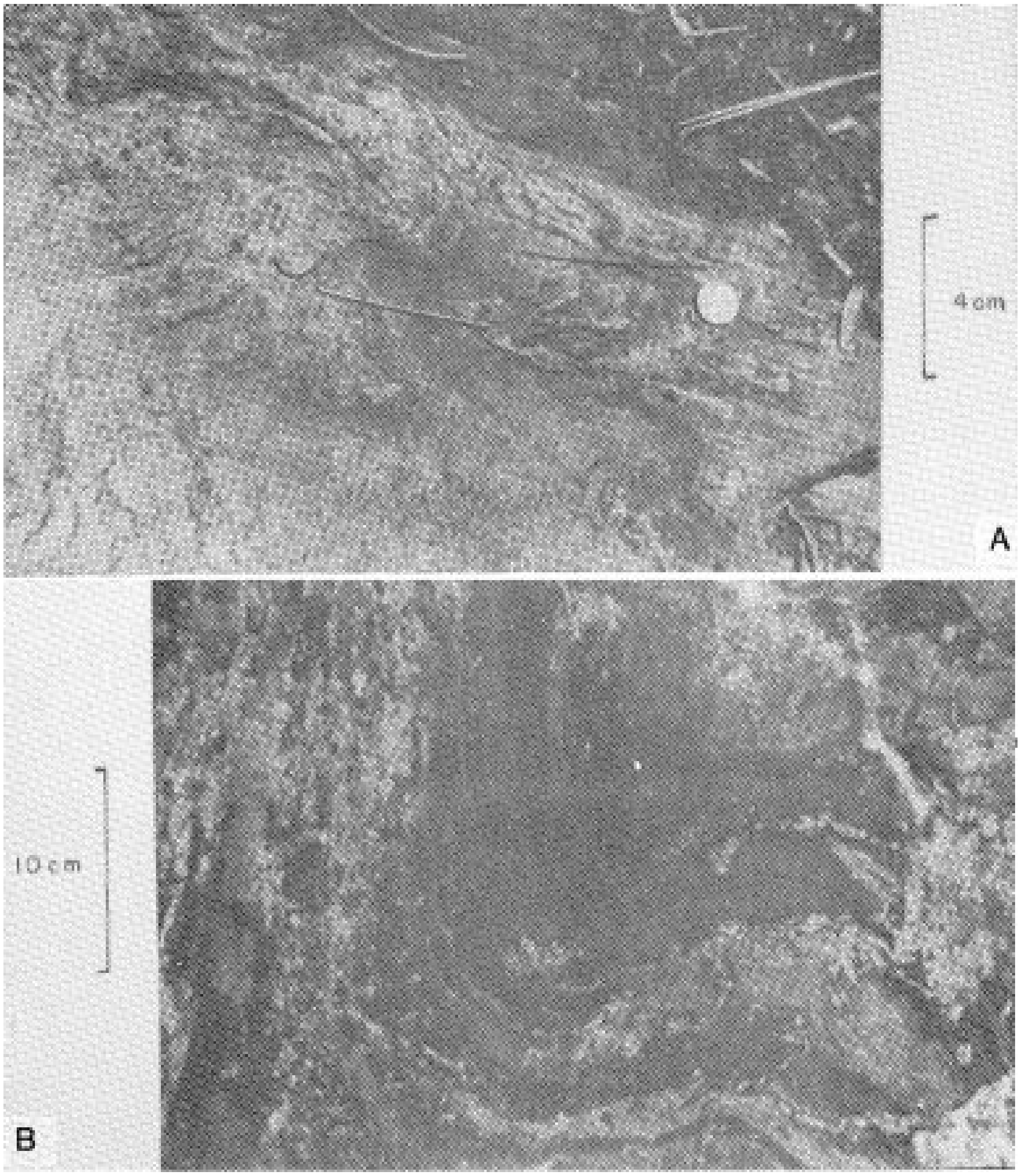

Figure 9. A - F_1 refolded about F_2 axial surface (msg). Outcrop of garnet-mica gneiss (msg) unit, Camp Creek block, containing F_1 fold being refolded about F_2 axial surface. Penny rests on F_2 nose with stick in foreground approximately parallel to the F_1 axis on that limb of the F_2 fold.
B - Asymmetric F_2 fold. Outcrop of asymmetric F_2 folds with southeast limb vertical and northwest limb horizontal. Note small rootless isocline at right center of photograph. The outcrop is located on Grassy Creek approximately 100 m north of SR 1307.

by peralkaline granitic gneisses (Figure 4) of the Crossnore Suite which, in this region, is known to range from 572 m.y. to 674 m.y. (Fullagar and Butler, 1980).

BASEMENT GNEISSES

Basement lithologies of the southwestern Sauratown Mountains anticlinorium include a layered, variably biotite-rich gneiss (Figure 5): the Low-Water Bridge Gneiss (Bartholomew and Lewis, in press). It is lithologically similar to the Cranberry Gneiss of the Elk Park massif (Bartholomew and Lewis, in press and the Stage Road Gneiss (Sinha and Bartholomew, in press) of the Lovingston massif. Those writers interpret these layered gneisses as being of probably volcanic/volcanoclastic origin. The results presented by Fullagar and Bartholomew (this volume) and Monrad and Gulley (this volume are consistent with this interpretation. Biotite augen gneisses (Figure 5), similar to the intrusive Little River and Blowing Rock gneisses of the Lovingston and

Globe massifs, respectively (Bartholomew and Lewis, in press) are probably intrusive into the Low-Water-Bridge Gneiss. The Low-Water-Bridge Gneiss also is intruded by generally leucocratic granitoid gneisses which are very similar to the Laurel Creek Pluton that intrudes the Cranberry Gneiss in Ashe County. Bartholomew and Lewis (in press) chose not to show the layered biotite gneiss in the Sauratown Mountains as Cranberry (as did Espenshade and others, 1975) only because it is not in physical continuity with the Cranberry in its type area; however, the two gneisses are lithologically very similar as noted Espenshade and others (1975).

DEFORMATIONAL HISTORY

Although most certainly a Grenville tectonic history exists in the basement rocks of the Sauratown Mountains area, generally poor exposure and intense Paleozoic metamorphic and deformational overprint obscure Precambrian structures (Figure 7a) in the crystalline rocks. The earliest widespread Paleozoic structural event in the Elkin area resulted in isoclinal folding visible from microscopic to megascopic scale ()Figure 7B). This deformation culminated with emplacement of large-scale nappe-like structures with gently southeast-dipping axial surfaces exemplified by the allochthon (camp Creek Block, Figure 1) structurally above the Camp Creek fault (2, Figure 8).

During this early period of folding (1, Figure 8), compositional layering in non-massive rocks was folded about these originally nearly horizontal, low-angle axial surfaces (Figure 7B), and in many instances, transposition of the compositional layering parallel to, or nearly parallel to, the axial surfaces of these folds occurred.

Metamorphism reached upper amphibolite grade and peaked during this deformational event. Emplacement of the Camp Creek nappe under these conditions of relatively high temperature and ductile conditions resulted in widespread shearing and disruption of fabric elements. Previously-formed isoclinal folds were sheared, and may, for instance, appear as isolated intrafolial folds from the thin section to the outcrop scale (Lewis, 1980a). Inasmuch as most of the southwestern Sauratown Mountains anticlinorium area was the upper footwall block of this nappe, the associated deformational effects occur over a wide area.

Subsequent to emplacement of the Camp Creek Block, rocks of the southwestern Sauratown Mountains anticlinorium were subjected to another folding event (3, Figure 8), producing tight to isoclinal folds (Figures 6A and 9B) which refolded the previous Camp Creek generation (Figures 7B and 9A) and the Camp Creek fault (Figure 8). This set of folds is strongly asymmetric to, and commonly overturned to, the northwest. Ductile conditions continued to prevail during this even t and into the movement of the Yadkin fault (4. Figure 8), which juxtaposed the Inner Piedmont with the Sauratown Mountains anticlinorium. Ductile deformation concomitant with the Yadkin fault is more localized in intensity near the trace of the Yadkin fault than the widespread deformation of the Camp Creek nappe.

Axial surfaces of the first and second folding events discussed above generally strike nearly parallel with one another and are nearly coaxial (Figure 9A). Most of the axes plunge gently southwest. Figure 7B is a mesoscopic view of what probably is a similar relationship between the first generation nappes and the succeeding fold event.

Within this portion of the Sauratown Mountains anticlinorium, at least, the two above-described fold types are the most prominent folds in the region. Outcrops which show both the relationships of the aforementioned folds plus the later folds (not shown on Figure 8) are uncommon. The youngest folds in the region tend to be broad open structures with upright, or very nearly so, axial surfaces.

The last major tectonic event in this region is movement on the Stony Ridge fault 95, Figure 8) which is characterized by brittle, rather tan ductile, deformation. Hatcher and others (this volume) suggest that no movement occurred on this 100 km-long fault zone. If this were indeed the case this zone would undoubtedly be an extremely long fault zone with no movement. The extensive zones of cataclasites, the abrupt termination of lithologic units and the re-equilibration of the Rb-Sr system to yield a 180 m.y. isochron all argue for significant movement on this fault zone.

DISCUSSION

Although more geochronology is sorely needed to resolve problems of relative ages, particularly on suspect Grenville-age rocks (Low-Water-Bridge Gneiss), in the Sauratown Mountains anticlinorium, the similarity of the older lithologic package to those of Grenville basement terranes nearby in the southern Appalachians is striking. The association of scattered augen gneisses and granitoids both intruded into layered biotite gneiss and metamorphosed to upper amphibolite grade approximately 1000 m.y. is typical of the entire eastern part of the Blue Ridge Province (Bartholomew and Lewis, in press). The Sauratown basement rocks appear to fall into a similar tectonic framework. Also, mapping by Lewis (1980) argues strongly for a nonconformable relationship (established before intrusion of the Crossnore Suite) between the Ashe metasedimentary sequence as mapped by Lewis (1980) and the felsic and layered biotite gneisses below the lowest metasedimentary unit of the Ashe Formation. The crosscutting Crossnore-type granitoid would indeed preclude a pre-metamorphic thrust as proposed by Hatcher and others (this volume) inasmuch as the thrust would also have to be pre-Crossnore (572-674). Plate 1 shows truncation of the contact, between the layered biotite gneiss and the granitoid, against the lowermost schist unit, in the northcentral portion of the Copeland quadrangle, which would be typical at an unconformity but requires multiple faults under Hatcher and others' interpretation. Moreover the fact that metaquartzite occurs both within the central basement area structurally above the gneisses (as also on Pilot Mountain), and as a basal unit along the garnet mica schist-basement core contact argues against a fault relationship between the basement core area and the overlying cover rocks. Likewise to suggest (Hatcher and others, this volume) that the "Henderson Gneiss" (Figure 6) found southeast of the Yadkin thrust, is lithologically distinguishable neither from the Low-Water-Bridge Gneiss (Figures 5 and 7) nor the biotite gneisses (Figure 2C) of the Ashe is to ignore obvious textural and mineralogical differences. Hatcher and others (this volume) suggest that the augen gneisses and granitoid intrusives (more or less acknowledged as basement) are coincidentally everywhere in fault contact (their Figure 3) with the surrounding layered biotite gneiss (Low-Water-Bridge Gneiss). This results in a number of odd-shaped "windows" to account for what are herein considered normal Grenville-age intrusive contacts. The conclusions of their "in progress" study seem to follow the train of Hatcher's previous interpretation (1978) that the Ashe lithologies must be underlain almost everywhere by oceanic (not continental) crust. This does not appear to be the case in Ashe and Avery counties (Bartholomew and others, this volume; Bartholomew and Lewis, in press) where he previously had projected the Hayesville fault along the Ashe-basement unconformity. Moreover it also does not appear to be the case in the Sauratown Mountains anticlinorium. Finally, what these two contrasting interpretations hinge on is the Low-Water-Bridge Gneiss: if it is the typical, Grenville-age (or older) amphibolite-facies, layered gneiss then it is basement and no fault is necessary to separate the Ashe Formation from the basement core. If it is a late Precambrian (post-Grenville) unit more structurally complex relationships, like those proposed by Hatcher and others (this volume), would be required. There does appear to be an additional (earlier) folding event in the Low-Water-Bridge Gneiss (Figure 7A), however, more work needs to be done on this unit.

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APPENDIX I

Key to Map Symbols

Lithologic Symbols

Burch Block

Cg – Crossnore Gneis
p – pegmatite
lg – leucocratic gneiss

a – amphibolite
u;ut;ud – ultramafic rocks; talc schists; dunite
a-6 – amphibolite and associated ultramafic rocks
gms-5 – heterogeneous garnet mica schist
bg-4 – biotite gneiss and associated ultramafics
gms-3 – garnet muscovite schist
bg-2 – biotite gneiss and associated ultramafic rocks
ggms-1 – graphitic garnet mica schist
q – quartzite
hg – hornblende gneiss
lgg – laminated granitic gneis
ag – augen gneiss
lbg – layered biotite gneiss

Zephyr Block

a – amphibolite
lg – leucocratic gneis
c – metaconglomerate
msp – mica schist and phyllite
fgms – finer-grained mica schist

Camp Creek Block

lg – leucocratic gneis
msg – micaschist and gneiss
bfg – biotite feldspathic gneis
fbg – feldspathic biotite gneis

Yadkin Thrust Sheet

Hg – Henderson Gneiss

Structural Symbols

Contacts

______ - exposed
---------- - approximate
………. - alluvium or terrace deposit

Faults

high-angle fault, sense of movement unknown
thrust fault, teeth on upper plate

Folds

direction and angle of plunge of early (F_1) fold showing strike and dip of axial surface
direction and angle of plunge of later (F_2) fold showing strike and dip of axial surface.

Foliation

strike and dip of compositional layering
strike and dip of early foliation (S_1)
strike of vertical early foliation
strike and dip of early foliation parallel to compositional layering
strike of vertical early foliation parallel to compositional layering
strike and dip of later foliation ($_2$)
strike of vertical S_2 foliation
strike and dip of later foliation, relative age uncertain
strike and dip of layering in amphibolite

Mines, Quarries and Prospects

active quarry
inactive or abandoned mine or quarry
inactive or abandoned quarry
prospect

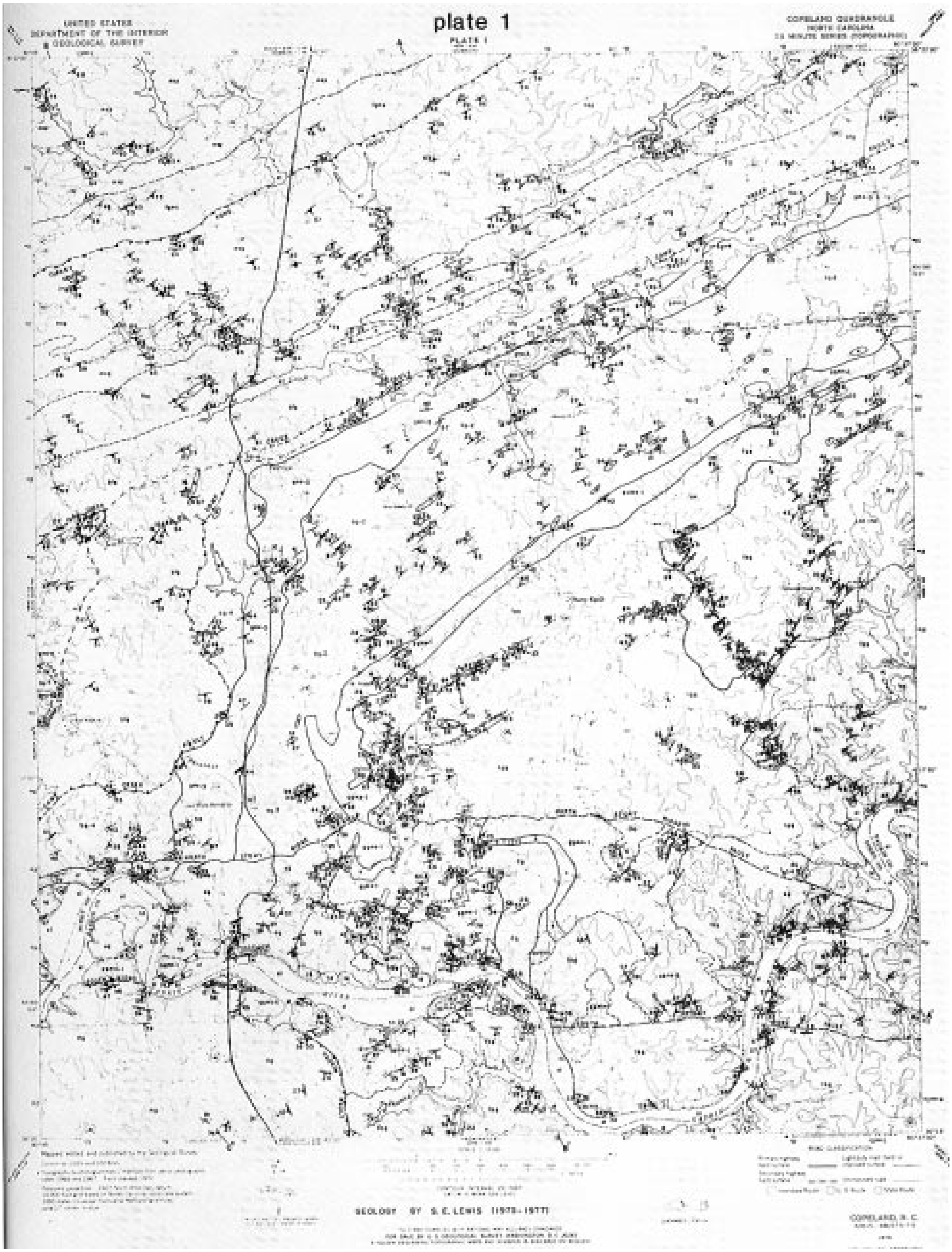


plate 1
UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY
COPELAND QUADRANGLE
NORTH CAROLINA
GEOLOGY BY S. E. LEWIS
COPELAND, N.C.


plate 2
UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY
ELKIN NORTH QUADRANGLE
NORTH CAROLINA
7.5 MINUTE SERIES (TOPOGRAPHIC)
GEOLOGY BY S.E. LEWIS (1977-1978)
ELKIN NORTH, N.C.

ROAD LOG – FIRSTDAY

Field trip log starts the first day at the intersection of US 321/US 421 bypass (N.C. road 1107) and N.C. Highway 105. NOTE: **Buses will be boarded at the CGS headquarters motel in Boon and not at the starting point of the log.**

incremental mileage	total miles	
0.0	0.0	proceed southwestward on N.C. Highway 105 to town of Foscoe.
5.6	5.6	Town of Foscoe, continue southwestward on N.C. Highway 105 south.
3.5	9.1	Cross Watauga County/Avery County line, continue southwestward on N.C. Highway 105 south.
1.6	10.7	Junction of N.C. Highways 105 and 184; turn northwestward (right) onto 184 north and proceed toward Banner Elk.
4.3	15.0	Town of Banner Elk; junction of N.C. Highways 184 north and 194; turn westward (left) on 194/184 toward Beech Mountain.
0.3	15.3	Junction of N.C. Highways 184 north and 194 south; turn northward (right) onto 184 north toward Beech Mountain.
2.5	17.8	Stop at dirt road on west side (left) of 184 part way up Beech Mountain; **STOP** 1 is a 0.5mile steep hike to the west up the dirt road. (NOTE: Buses will drive up to the top of Beech Mountain on the paved road and return to this point before passengers get off.)
2.5	20.3	Turn around and proceed southward on 184 south toward Banner Elk to junction of N.C. Highways 184 and 194; turn west (right) on 194 south.
4.4	24.7	Junction of N.C. Highway 194 and state road 1308 to Whaley; continue westward on 194 toward Elk Park.
2.1	26.8	Junction of N.C. Highway 194 and U.S. Highway 19E; turn east (left) on 194/19E south.
0.4	27.2	Junction U.S. Highway 19E and N.C. Highway 194; turn southwest (right) on 19E south toward-Cranberry.
0.7	27.9	Turn west (right) onto dirt road entrance to Cranberry mine at town of Cranberry.
0.2	28.1	Park at business office and proceed on foot approximately 0.1 miles to the west to **STOP 2** in the Cranberry mine. (NOTE: Permission in writing should be obtained from the Cranberry Magnetite Corporation before attempting to visit the Cranberry mine.)
0.2	28.3	Turn around and proceed eastward on dirt road back to the intersection with U.S. Highway 19E at town of Cranberry. Turn north (left) onto 19E north.
0.7	29.0	Junction of U.S. Highway 19E and N.C. Highway 194; turn northwest (left) on 19E/194 north.
0.4	29.4	Junction of U.S. Highway 19E and N.C. Highway194; continue northwestward on 19E north toward Elk Park.
0.7	30.1	Community of Elk Park; continue northwestward on 19E north toward Roan Mountain, Tennessee.
1.4	31.5	Cross North Carolina (Avery County)/Tennessee (Carter County) state line.
4.6	36.1	Junction of U.S. Highway 19E and Tennessee Highway 143 near town of Roan Mountain; turn South onto 143 toward Bakersville, N.C. (NOTE: Tennessee Highway 143 becomes North Carolina Highway 261 at the Tennessee (Carter County)/North Carolina (Mitchell County) state line at the crest of Roan Mountain.)
12.8	48.9	Junction of TN 143 (N.C. 261) with access road to Roan High Bluff at Carvers Gap on the crest of Roan Mountain at the Tennessee/North Carolina state line; turn west (right) onto Access road up Roan Mountain.
1.7	50.6	Turn into first park lot on the right; after getting off buses proceed to the southwestern portion of lot. **STOP 3A**. (NOTE: If the weather is nice, field trip participants will walk back down the access road and reboard the buses at the junction of the access road and TN 143 (N.C. 261) at Carvers Gap; for bad weather the buses will also stop at STOP 3B at the sharp bend along the access road (0.8 miles from A) and at STOP 3C (between 1.0 and 1.3 miles from A).
1.7	52.3	Pickup point for STOP 3 at Carvers Gap; turn south (left) onto TN 143 (N.C. 261) toward town of Roan Mountain.
12.8	65.1	Junction of Tennessee Highway 143 and U.S. Highway 19E; turn southeast (right) onto 19E South.
4.6	69.7	Cross Tennessee (Carter County)/North Carolina (Avery Count) state line.
1.4	71.1	Community of Elk Park; continue southwestward on 19E south.
0.7	71.8	Junction of U.S. Highway 19E and N.C. Highway 194; turn north toward Banner Elk on 194.

2.1 73.9 Junction of N.C. Highway 194 and state road 1308; turn north (left) on 1308 (Beech Mountain road) toward Whaley.

3.0 76.9 Junction of NC state roads 1308 and 1310; continue north on main paved road (now 1310).

1.9 78.8 Junction of NC state roads 1310, 1311, and 1316 at Dark Ridge; continue north on main paved road (now 1316).

3.0 81.8 Junction of NC state roads 1316 and 1312 at Whaley; bear to northeast (right) onto 1312 (main paved road).

3.6 85.4 Junction NC state roads 1312 and 1314; continue northward on main paved road (now 1314).

0.6 86.0 Junction of NC state road 1314 with U.S. Highway 321 at the Avery County/Watauga County line; turn east (right) onto U.S. 321 and cross county line into Watauga County.

2.1 88.1 Junction of U.S. Highway 321 and N.C. state road 1202 (Bethel Church road) turn north (left) onto 1202.

0.5 88.6 **STOP 4 –** limited parking available on western shoulder of road. (NOTE: If weather is nice field trip participants will leave buses at this point (STOP 4A) and walk along the road for the next 1.0 mile and reboard buses at the bridge over Beaverdam Creek; for bad weather the buses will also stop at STOP 4B at the pullout just north of the bridge over the Watauga River (0.4 miles from A) and at STOP 4C just south of the bridge over Beaverdam Creek (0.9 miles from A).

1.0 89.6 Pickup point for STOP 4C just north of bridge over Beaverdam Creek at junction of N.C. state roads 1201 and 1202; proceed northward on main paved road (now 1201).

0.7 90.3 **STOP 5 –** passengers will please leave buses and allow buses to clear this stretch of narrow road before viewing exposures at curve; passengers will then walk to the buses which will be parked along more open visible stretches of the road; continuing northeastward on 1201.

1.6 91.9 Junction of N.C. state roads 1201 and 1222 at Bethel; proceed eastward on 1201.

1.6 93.5 Junction of N.C. state roads 1201 and 1221; continue eastward on 1201.

0.2 93.7 Junction of N.C. state roads 1201 and 1213; continue southeast on main paved road (now 1213) toward Georges Gap.

2.0 95.7 Crest of Georges Gap; continue southeastward on 1213 toward Sugar Grove.

0.7 96.4 **STOP 6 –** limited parking on west (right) shoulder, please allow buses to drop off passengers and move to more visible parking areas before viewing outcrop at STOP 6; if weather and time permit we may walk 0.5 miles on down the road to view some of the debris-avalanche deposits in this valley; reload buses and continue southeastward on 1213 toward Sugar Grove.

2.0 98.4 Junction N.C. state road 1213 with old U.S. Highway 421; turn southwestward (right) onto Old U.S. 421.

0.4 98.8 Junction of old U.S. Highway 421 and U.S. Highway 321 at Sugar Grove; turn east (left) Onto U.S. 321.

1.0 99.8 Junction of U.S. Highways 321 and 421; turn southeast (right) onto U.S. 321/421.

3.6 103.4 Junction of U.S. Highways 321/421 and Bypass (321/421) around Boone; turn southeast onto 421 Bypass.

1.8 105.2 Junction of U.S. 321/421 Bypass and N.C. State Highway 105.

END ROAD LOG – FIRST DAY.

Buses will turn east (left) onto N.C. 105/U.S. 321/421 Bypass and return to CGS Headquarters motel in Boone.

ROAD LOG – SECOND DAY

Field trip log starts the second day at the intersection of U.S. 321/U.S. 421 Bypass (N.C. road 1107) and N.C. Highway 105. NOTE: Buses again will be boarded at the CGS headquarters motel in Boone and not at the starting point of the log.

incremental total
mileage miles

0.0 0.0 Proceed northwestward on U.S. 321/421 Bypass.

1.8 1.8 Junction of US 321/421 Bypass and U.S. 321/421; turn west (left) onto U.S. 321/421.

3.6 5.4 Junction of U.S. 321 and U.S. 421; proceed northward (bear right) onto U.S. 421 north.

6.8 12.2 Cross North Carolina (Watauga County)/Tennessee (Johnson County) state line; continue northward on U.S. 421.

0.8 13.0 Junction of U.S. Highway 421 and Tennessee state road 67 (N.C. 88) at Trade, Tennessee turn northeast (right) on TN 67.

1.6 14.6 Cross Tennessee (Johnson County)/North Carolina (Watauga County) state line. (NOTE: Road changes from TN 67 to N.C. Highway 88);continue northeastward on N.C. 88.

1.9 16.5 Cross Watauga County/Ashe County boundary.

2.1 18.6 Junction of N.C. Highway 88 and N.C. state road 1118; continue northeastward (main road) on N.C. 88.

5.2 23.8 Junction of N.C. Highway 88 and N.C. state road 1308; turn northwest onto 1308 toward Asheland.

0.6 24.4 **STOP 7** – Pull off into limited parking area along north (right) side of road; exposures across the road are STOP 7 at Asheland; after viewing STOP 7, turn around and proceed southeastward on N.C. 1308.

0.6 25.0 Junction of N.C. Highway 88 and N.C. state road 1308; turn northeast onto N.C. 88.

2.3 27.3 Junction of N.C. Highway 88 and N.C. state road 1100 at Creston; continue northeastward (main road) on N.C. 88.

1.2 28.5 Junction of N.C. Highway 88 and N.C. state road 1315 (Big Laurel road); turn northwest (left) onto 1315 toward Mountain City.

1.6 30.1 Junction of N.C. state roads 1315 and 1310; continue northeastward on 1315 toward Mountain City.

6.1 36.2 Cross North Carolina (Ashe County)/Tennessee (Johnson County) state line.

3.1 39.3 **STOP 8** – Parking space is limited along shoulder of road; exposures to be viewed from the outcrop along the north side of the road; after STOP 8 reboard buses and continue northwestward on the Forge Creek road toward Mountain City.

2.7 42.0 Junction of Forge Creek road with U.S. Highway 421; turn south (left) at stoplight just south of Mountain City, Tennessee on U.S. 421 south toward Boone, N.C.

7.9 49.9 STOP 9 – Park in parking are on west (right) side of highway in front of the firecracker shop; outcrop to be examined is directly across highway; after STOP 9 reboard buses and continue southward on U.S. 421 south, toward Boone, N.C.

0.4 50.3 Junction U.S. Highway 421 and Tennessee state road 67 (N.C. 88); continue southward on U.S. 421 south toward Boone, N.C.

0.8 51.1 Cross Tennessee (Johnson County)/North Carolina (Watauga County) state line.

0.4 51.5 **STOP 10** – Park on shoulder on west (right) side of highway; principal outcrop to be viewed is on east (left) side of highway; after STOP 10 reboard buses and continue southward on U.S. 421 south toward Boone. Junction U.S. Highways 421 and 321; continue southeastward on U.S. 421/321 toward Boone.

3.6 55.1 Junction of U.S. Highways 321/421 and 321/421 Bypass; turn southeast (right) onto 321/421 Bypass around Boone.

1.8 56.9 Junction of U.S. 321/421 Bypass and N.C. Highway 105.

END OF ROAD LOG – SECOND DAY.
Buses will turn east (left) onto N.C. 105/U.S. 321/421 Bypass and return to CGS Headquarters Motel in Boone.

STOP 1 – ROCKS OF THE ELK PARK MASSIF AT BEECH MOUNTAIN (FIGURE 1)

Leaders: Sharon E. Lewis, Mervin J. Bartholomew
7.5-minute quadrangle: Elk Park (Lewis and Bartholomew, in press)
References: Bartholomew and others (this volume)
Bartholomew and Lewis (in press)

Walk up the dirt road, to the west of the parking area, to the sharp curve (about 0.5 miles) at the top of the hill. For the next 0.5-0.75 miles a variety of rocks on the Elk Park Massif (Bartholomew and Lewis, in press) are exposed in the roadcuts. To the west at the bottom of the incline both the medium-grained and coarse-grained granite of the Beech Pluton are well exposed (A, Figure 1; Figure 2). The coarse-grained granite forms the prominent bluffs at the top of the Beech Mountain along the road. Returning eastward, the coarse-grained granite overlies the medium-grained granite in which biotite is considerably more abundant. These two rock types comprise the bulk of the Beech Pluton in this quadrangle. The Beech Pluton overlies the adjacent country rock along a gently inclined contact as first noticed by Keith (1903). By contrast, the contact along the northern side of the Beech Pluton is steeply dipping. Thus we have inferred that the Beech Pluton is actually more of a tabular-shaped body which forms the lower limb of a large recumbant fold similar to the folded Buckeye Knob Pluton in the adjacent Sherwood quadrangle (Bartholomew and Gryta, 1980). The upper limb of the gently-eastward-plunging Beech Mountain nappe largely has been removed by erosion in this area.

As can be seen from the map (B, Figure 1) of the Elk Park quadrangle, Beech Pluton cuts older granitoid plutons. These older, smaller, plutons resemble the Crossnore Suite (Bartholomew and Lewis, in press) of late Precambrian peralkaline granites defined by Rankin and others (1973) as well as rocks of the older (Grenville-age) Forge Creek Suite (Bartholomew and Lewis, in press) found in a structurally

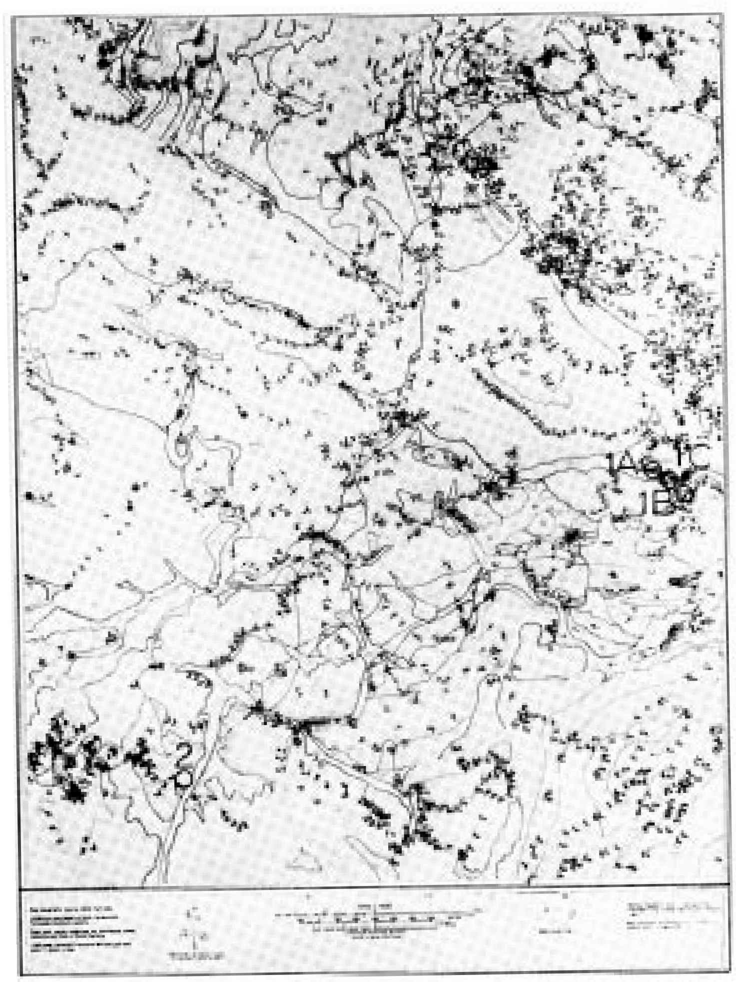


Figure 1: Preliminary geologic map of the Elk Park 7.5-minute quadrangle (Lewis and Bartholomew, in press) showing location of STOPS 1 and 2. Map is reproduced here with permission of the N.C. Division of Land Resources.

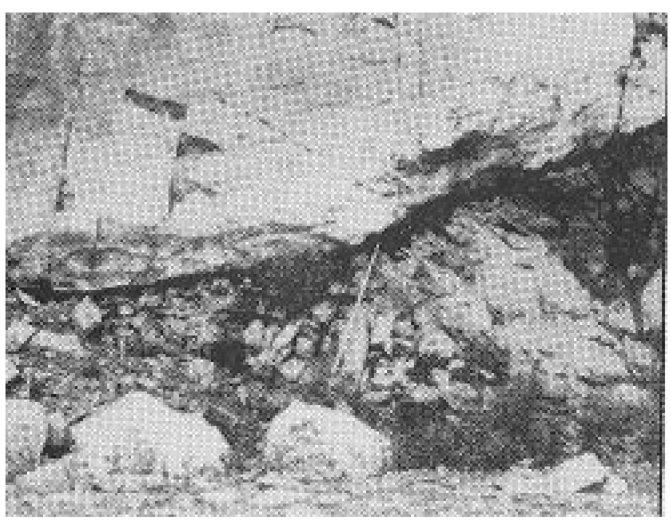


Figure 2: Gently dipping contact between the coarse- and medium-grained granites of the Beech Pluton at STOP 1A. Walking staff is about 5 feet.

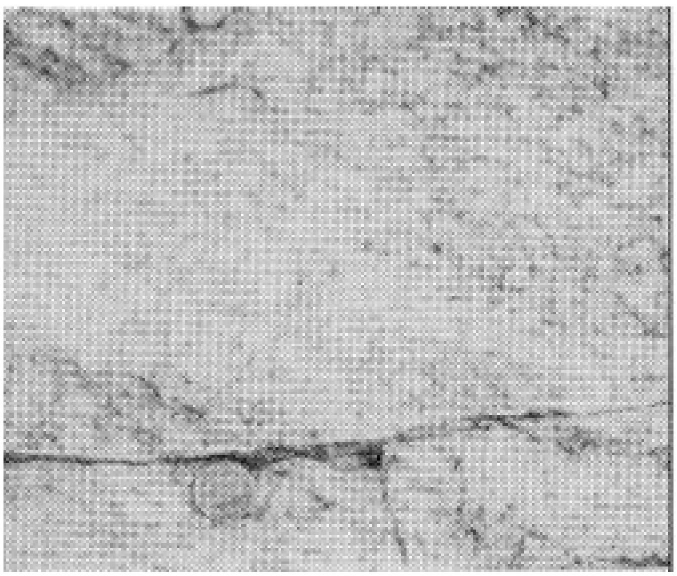


Figure 3: Typical younger granitoid (pre-Beech Pluton) exposed at STOP 1B.

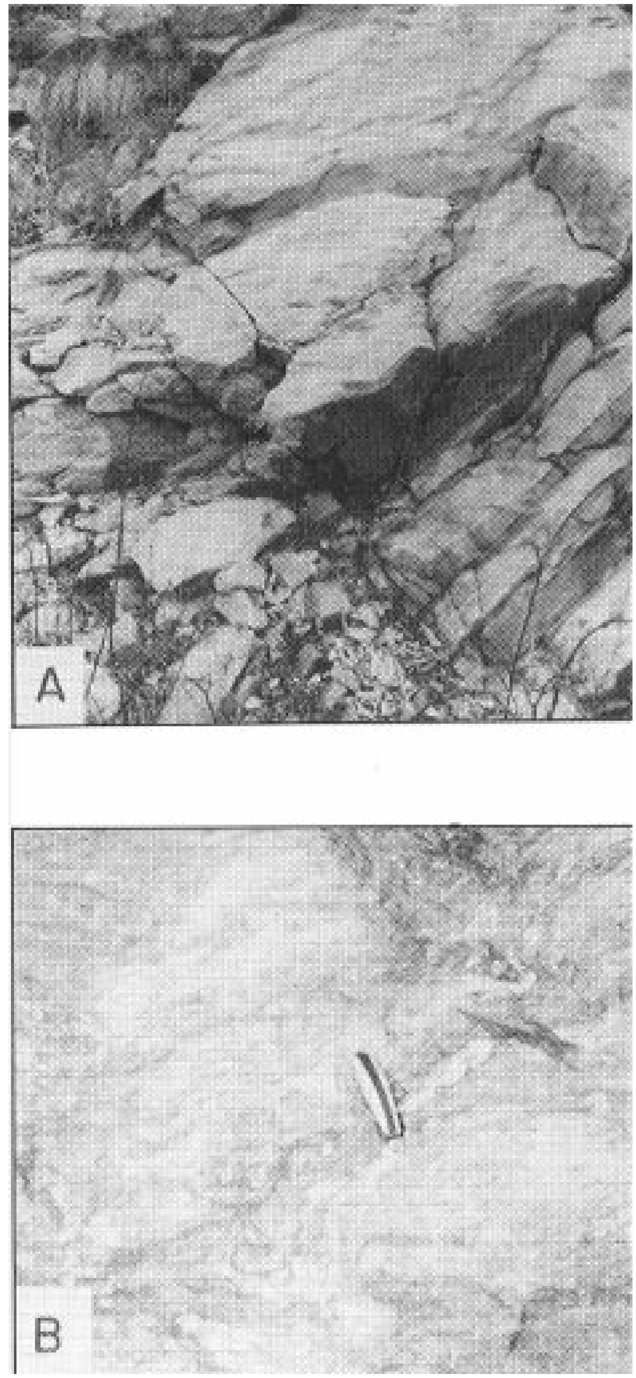
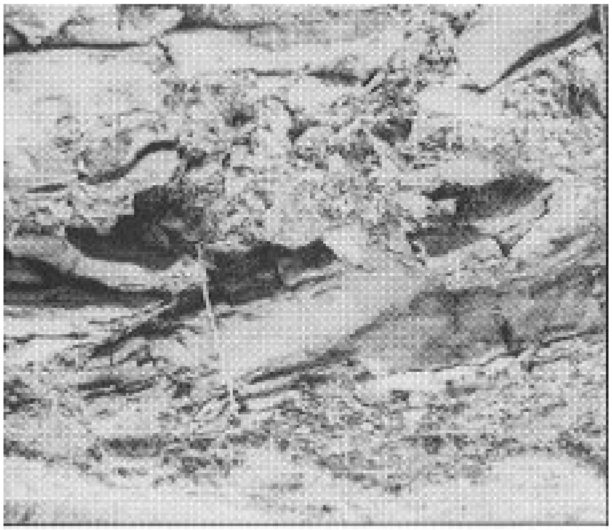


Figure 4: Cranberry Gneiss exposed at STOP 1; A - shows the thick lithologic layering that characterizes the Cranberry; folds are visible in upper portion of outcrop; biotite gneiss layers at lower right are about 0.5 to 1 foot thick.
B – Closeup of highly deformed layer within Cranberry Gneiss.



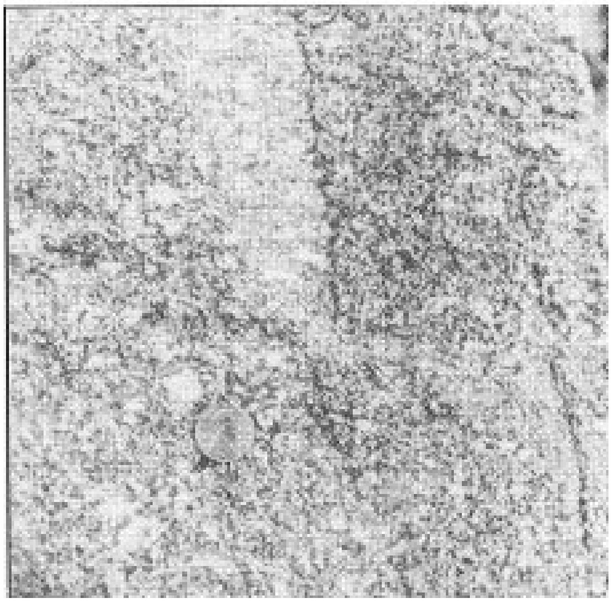


Figure 5: (Top) Fine-grained felsic dike exposed at STOP 1B. These dikes represent the last intrusive rocks of the Beech Pluton. Staff is about 5 feet.

Figure 6: (Bottom) Typical coarse-grained dioritoid exposed at STOP 1B; such dioritoids are commonly associated with the peralkaline granitic plutons of the Crossnore Suite.

lower thrust sheet to the northwest. The age of these small, pre-Beech granitoid plutons has yet to be determined. The prominent exfoliated knob along the road to the east is part of one of these pre-Beech granitoid plutons. The rock (Figure 3) typically contains both green and pink feldspar and biotite forms a pronounced foliation. This granitoid unit is well exposed near the base of the exfoliated knob.

The Cranberry Gneiss (Keith, 1903) forms the country rock into which the various plutons were intruded. The Cranberry in this area (C, Figure 1) is largely a coarsely layered biotite gneiss (Figure 4A) which is highly deformed (Figure 4B).

Other rock types which also occur along this dirt road include the late-stage, fine-grained felsic dikes (Figure 5) and associated darker more mafic dikes; these are both considered to be related to the Beech Pluton. Also float of coarse-grained dioritoid (Figure 6) is found near the border of the Beech Pluton in this area as well.

STOP 2 – CRANBERRY MAGNETITE DEPOSIT

Leaders: William Ussler,P. Jeffries, Geoffrey Feiss, and Stephen Goldberg

Introduction:

The Cranberry Magnetite Mine is located on U.S. Highway 19E, less than a mile south of the intersection of U.S. 19E and N.C. 194. The mine is located just west of the Cranberry Post Office. Permission to enter the mine must be obtained from the mill office on the premises (office phone number 919-733-5748).

For details on the geology of the Cranberry Deposit, the reader is referred to the paper by Feiss, et al. in this volume. Figure 7 shows the location of the mine. A similar deposit, the Wilder Mine, is also shown on Figure 7. The Wilder is located one mile south of U.S. 19E on a dirt road which turns off 3.4 miles west of the intersection of U.S. 19E and N.C. 194. Should you plan to visit the Wilder on your own, permission to cross the property (always granted) should be asked at the house southwest of the intersection of the dirt road with U.S. 19E. The lithologies and lithologic relations are easily observed at the Wilder Mine and are similar to those at the Cranberry Mine.The mill is currently in operation processing magnetite ores from upstate New York. The resulting magnetite concentrate is sold to coal companies and refractory producers for heavy metal flotation. The mine was surveyed in detail in 1979 and, in the future, may become an alternate source of magnetite for the mill.

Mine Geology:

Magnetite ore was mined both from surface workings and from an adit on the 3200’ level. At present, the portal of this adit is caved and entry is made through the “collapse pit”, see Figure 8. The 3200 level was driven into the hanging wall of the deposit and from the adit a series of open stopes were driven. Large pillars of ore remain. Declines below the 3200 level were run down

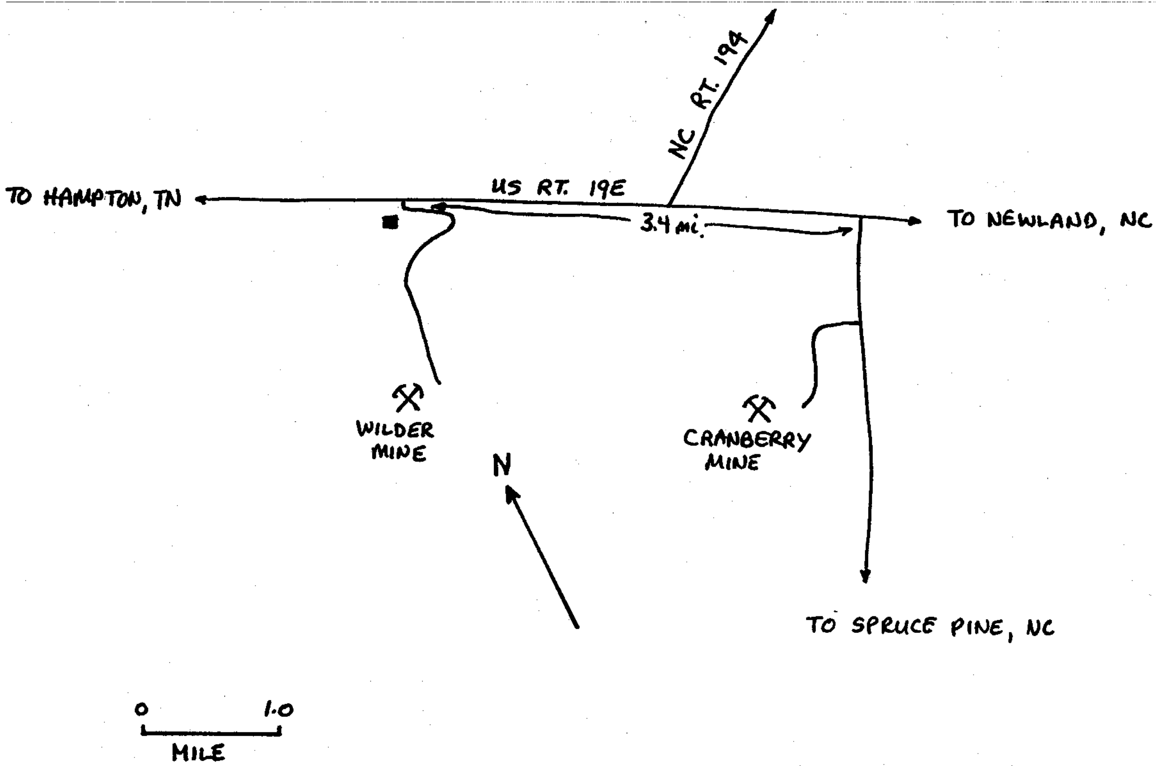


Figure 7: General location of Cranberry Magnetite deposit.

dip, the local name for these declines is "slopes". Several drifts appear to have been driven along the base of the slopes on what would be approximately the 3100 level. All workings below the 3200 level are presently flooded.

The 3200 adit was driven 1106 m into the hillside parallel to the strike to the ore zone (N45^{o}W). The ore zone consists of individual, discontinuous lenses of magnetite-rich rock within the host Cranberry Gneiss. They vary greatly in size and lie in the plane of the regional cataclastic foliation which dips 30 to 40^{o} to the southwest. The ore lenses are elongate in a roughly N80^{o}E direction, approximately parallel to the axial trace of the Spruce Pine-Snake Mountain Synform.

Entering the adit is extremely dangerous. An extensive roof fall occurred near the portal in 1976 or 1977.

Ore Petrology:

In the collapse pit (see Figure 8), one can see coarse-grained pegmatites which cross-cut the cataclastic foliation. The pegmatites consist of microcline, Na-plagioclase, and quartz. Within the pegmatites are numerous inclusions of the host rock, especially hedenbergite and Ca-plagioclase rocks which show strong epidote alteration.

The best area for collecting samples is the south mine dumps. Six major lithologies can be collected. These include:

Ore Zone Materials

1. Ore – principally composed of clinopyroxene and magnetite, sometimes pure magnetite. Note the variation in texture from equigranular to foliated.
2. Garnet + calcite ± clinopyroxene – associated with ore in an unclear manner; this assemblage has not been observed in place underground.

Border Zone Materials

3. Foliated clinopyroxene + Ca-plagioclase – most of this assemblage is altered to epidote.

Host Rock

4. Cataclastic Cranberry Gneiss – Na-plagioclase + microcline + biotite + quartz.
5. Coarse-grained pegmatites – microcline + Na-plagioclase + quartz with inclusions of clinopyroxene.
6. Calcite + quartz veins and veinlets – associated with a post-deformational retrograde metamorphism.

ACKNOWLEDGEMENTS:

The authors which to thank the Cranberry Magnetite Corporation for providing access to the Cranberry Mine during the study and for their willingness to permit us to visit the Mine today. A number of individuals have provided interest and support for this study. Noteworthy are Paul Fullagar, Jim Craig, and especially Jerry Bartholomew. Ussler's

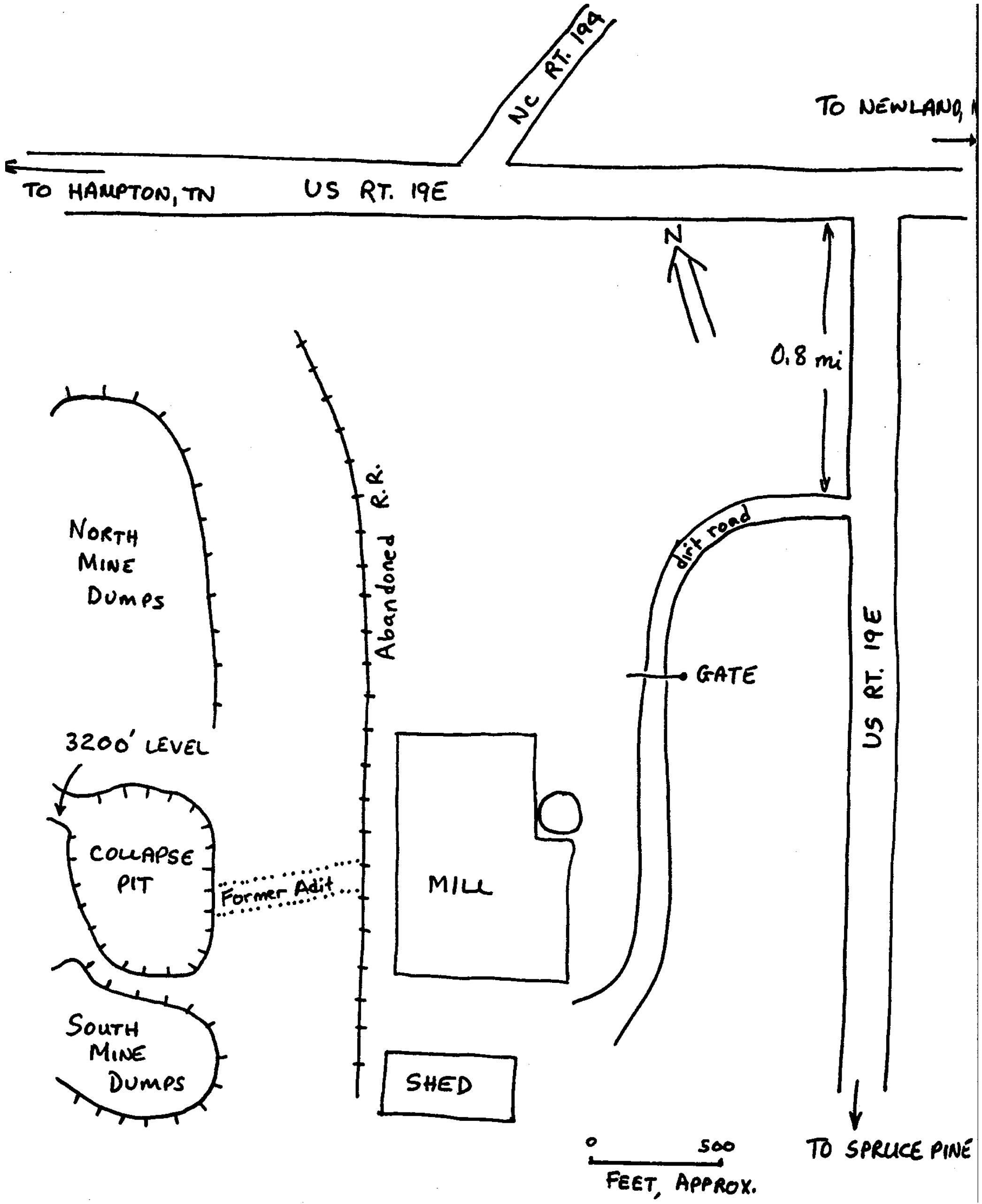


Figure 8: Sketch map of Cranberry Magnetite deposit.

work was partially supported by a grant from the Division of Land Resources.

STOP 3 – GRANULITE-FACIES GNEISSES ON ROAN MOUNTAIN

Leaders: Gerald L. Gulley, Jr., John R. Monrad
7.5-minute quadrangle: Carvers Gap (Gulley, 1982)
References: Monrad and Gulley (this volume)
Gulley (1982)

Cloud Land Gneiss is exposed on the southwestern and southern margins of the parking lot and along the trail beginning at its western end. The gneiss is generally coarse-grained and granoblastic, but locally contains discontinuous, subhorizontal, reddish-brown lenses of medium-grained mafic minerals. Felsic minerals are pink, perthitic orthoclase, quartz and white plagioclase. Mafic minerals are primarily medium-grained biotite and garnet in mafic layers, and isolated garnet crystals up to 2 cm diameter. Rounded zircons are common, while kyanite is rare or absent.

Between the parking lot and access road are boulders of Cloud Land Gneiss in colluvium. The gneiss at this outcrop contains bluish-gray, locally foliated and folded, kyanite-rich domains up to 15 cm wide. In addition to kyanite (20%), prominent minerals in the domains are quartz (45%), garnet (15%), perthitic orthoclase (10%) and biotite (7%). Rounded zircons are also present. These "high-Al" (>20% Al_2O_3) domains lie within pink to white granoblastic regions that contain garnet as the sole mafic phase.

The Al_2O_3-rich domains are approximately pelitic in composition, the high- Al_2O_3 domains in the Cloud Land Gneiss are locally intergradational over a scale of several centimeters. One possible explanation for this feature is that the pelitic domains arose from incorporation of "sedimentary xenoliths" into felsic igneous material, and were subsequently flattened due to deformation. Another possibility is that the high- Al_2O_3 and normal-Al_2O_3 regions in the Cloud Land Gneiss reflect original alternation of pelitic and quartzofeldspathic layers of a sedimentary sequence. Because of the apparent intergradational nature of these regions in this lithology, and because there is no evidence of any reaction between pelitic layers and any likely igneous material, the latter possibility of the two is more attractive; i.e., the protolith of the Cloud Land Gneiss was a stratified sedimentary sequence. This may explain the chemical variation within the homogeneous-looking quartzofeldspathic domains, as well as the local occurrence of irregular biotite-garnet layering in outcrop. Furthermore, the presence of numerous rounded zircons may represent a detrital mineral fraction within the "metasedimentary" sequence.

At STOP 3B and down the road to the northwest are representations of three lithologic units: Carvers Gap granulitic gneiss; granitoid segregations or "pods" and metadiabase dikes of the Bakersville Gabbro. Two types of Carvers Gap granulitic gneiss may be distinguished in outcrop: (1) massive graunulitic gneiss, in which foliation is faint to absent; and (2) layered granulitic gneiss which exhibits prominent gneissic layering. The massive and layered varieties are commonly intergradational.

Massive granulitic gneiss is fine- to medium-grained, granoblastic, and predominantly felsic in composition. Gneissosity, when visible, is defined by very thin (1-2 mm), discontinuous mafic layers spaced approximately 1 cm apart. The dominant felsic mineral is quartz, with lesser amounts of plagioclase and perthitic orthoclase. Mafic minerals (garnet, orthopyroxene and biotite) generally comprise less than 5% of felsic, massive granulites. The layered granulitic gneiss is characterized by alternating mafic and felsic layers ranging in thickness of 0.5-10 cm. The layers are discontinuous and commonly display an undulous, subhorizontal attitude. Felsic layers consist of plagioclase and quartz with potassium feldspar rare or absent. Common mafic minerals are orthopyroxene, garnet, clinopyroxene, biotite and two generations (brown and blue-green) of hornblende. The layered granulites are more common between STOP 3B and 3C.

Both varieties of Carvers Gap Gneiss contain granitoid segregations or pods. The segregations consist of abundant K-feldspar (up to 3 by 10 cm), quartz, plagioclase and biotite. In the massive granulites, the segregations commonly occur as pale-red, medium-grained, feldspar-rich dikes and sills that are 1-2 cm thick, and as blotchy, ill-defined regions rich in medium-grained quartz and feldspar. Granitoid segregations in the layered granulites are medium- to coarse-grained and generally concordant, although they locally crosscut or appear to deform the gneissic layering.

Between STOPS 3B and 3C are numerous dikes of the Bakersville Gabbro, which may be confused with rare mafic varieties of the Carver Gap Gneiss. The metadiabase dikes range from fine-grained and phaneritic with visible tiny laths of plagioclase, to aphanitic and flinty looking. Poikiloblastic garnet up to 3 cm diameter give the appearance of pink "snowflakes" in the dikes. These "snowflake" garnets, along with the presence of acicular plagioclase laths in less-recrystallized samples, may be used as field criteria to distinguish the dikes from mafic granulites. The dikes display sharp contacts with enclosing gneisses. Locally, evidence for multiple intrusions is exhibited.

STOP 4 – CROSSING KNOBAND WATAUGA GNEISSES ALONG THE BETHEL CHURCH ROAD (FIGURE 9

Leader: Mervin J. Bartholomew
7.5-minute quadrangle: Sherwood (Bartholomew and Gryta, 1980)
References: Bartholomew and others (this volume)
Fullagar and Bartholomew (this volume)
Bartholomew and Lewis (in press)

Bartholomew and Lewis (in press) designated these exposures (4A, Figure 9) as the type locality for the Crossing Knob Gneiss (Figure 10) and it is from outcrops along this road that samples were collected for radiometric age determinations (Fullagar and Bartholomew, this volume). The Crossing Knob Gneiss lies structurally beneath mylonitic rocks derived from the Watauga River Gneiss. Because of both its structural position and its distinct lithologic contrast with rocks of the Watauga Massif (Bartholomew and Lewis, in press), the Crossing Knob Gneiss was interpreted (Bartholomew and Gryta, 1980) to lie within a small window. The fault surrounding the window is probably of similar age as the Fork Ridge fault and is folded (post-ductile deformation) by folds with axial planar crenulation cleavage.

Walking northward on the Bethel Church road we cross out of the Crossing Knob window into the structurally overlying Watauga River Gneiss (Bartholomew and Lewis, in press). The mylonitic rocks derived from the massive grano-

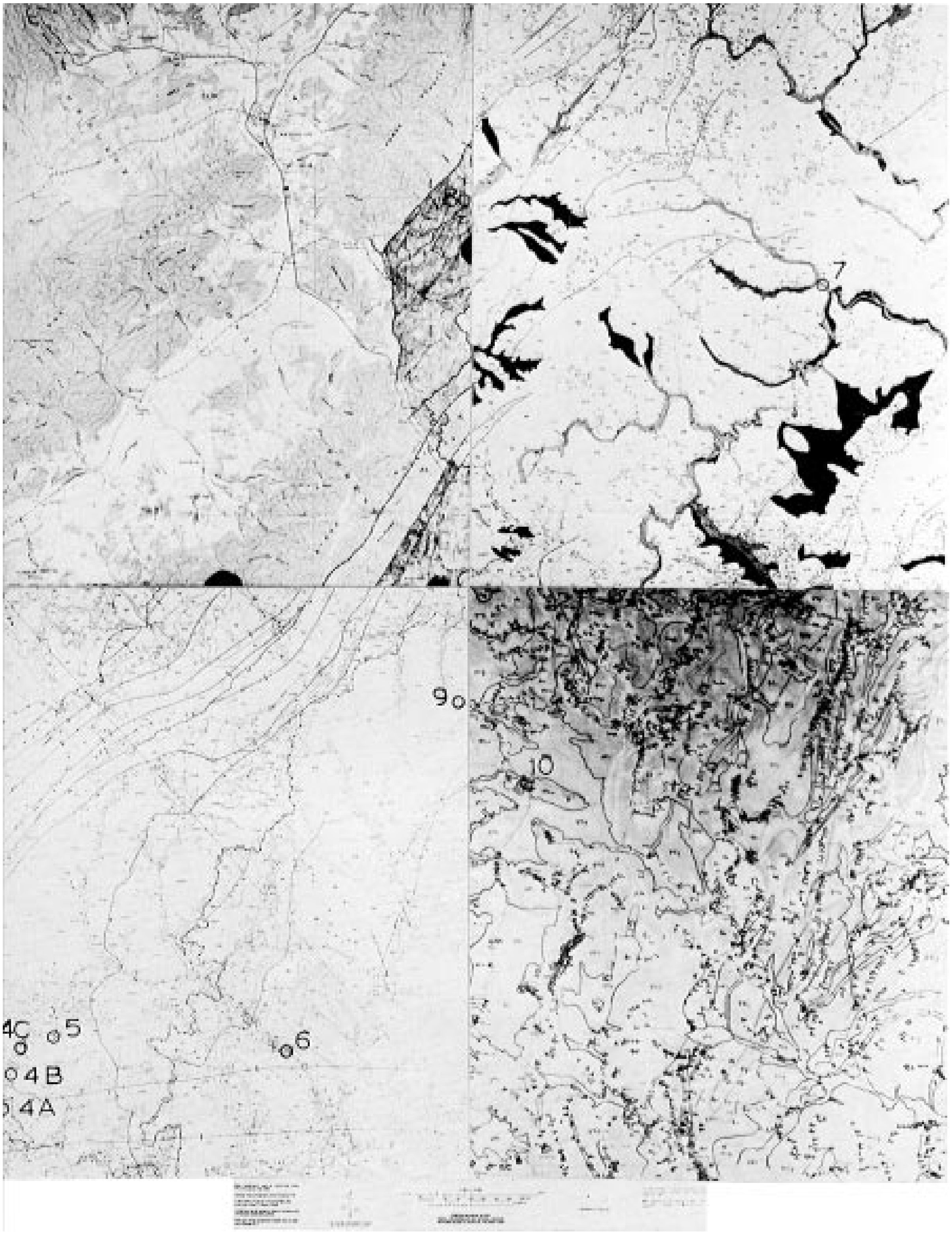


Figure 9: Geologic maps of the area showing locations of field trip STEOP 4,5,6,7,8,9, and 10; SW-published map of the Sherwood 7.5-minute quadrangle (Bartholomew and Gryta, 1980); SE-preliminary map of the Zionville quadrangle (Bartholomew and Wilson, in press). NW-reconnaissance map of basement rocks in the Mountain City quadrangle; NE-published map of the Baldwin Gap quadrangle (Bartholomew, 1983).

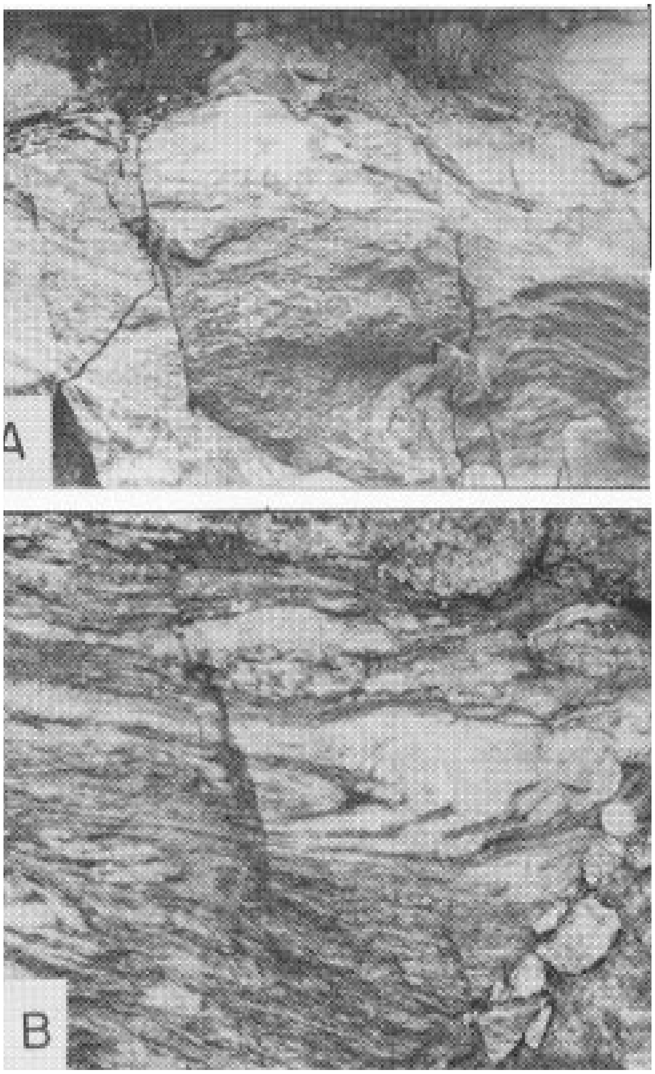


Figure 10: Crossing Knob Gneiss exposed at STOP 4A: A shows typical thick lithologic layering;
B – shows closeup of folds.

diorite are well exposed near the bridge over Watauga River (4B, Figure 11A). Here the distinctive layering (Figure 11B) is well developed with zones of quartzofeldspathic mylonite gneiss (Higgins, 1971) alternating with zones of mylonite schist (Higgins, 1971) much richer in micaceous minerals. A mafic dike has been transposed during the ductile deformation so that it is now sub parallel to the mylonitic foliation. Younger folds with axial-planar crenulation cleavage are also visible in these exposures.

Continuing northward on the Bethel Church road we cross out of the more mylonitic rocks into protomylonite (Higgins, 1971) and massive gneiss (Figure 11C,D) of the Watauga River Gneiss. A late-Precambrian dike (Figure 11C) cuts these massive rocks near the bridge over Beaverdam Creek (4C, Figure 9). Samples for the radiometric age determination of the Watauga River Gneiss (Fullagar and Bartholomew, this volume) were collected both from these outcrops near this bridge and from others along the Bethel Church road for several miles northeastward from this bridge. Typical relict igneous textures, xenoliths (Figure 11D), and numerous small apalitic and pegmaitic dikes in the massive portions of the Watauga River Gneiss all suggest an igneous origin for this Grenville-age unit.

STOP 5 – MYLONITIC ROCKS OF THE WATAUGA RIVER GNEISS ALONG THE BETHEL CHURCH ROAD (FIGURE 9)

Leaders: Mervin J. Bartholomew, Sharon E. Lewis
7.5-minute quadrangle: Sherwood (Bartholomew and Gryta, 1980)
References: Bartholomew and others (this volume)
Bartholomew and Lewis (in press)

The rocks here at these exposures (5, Figure 9) illustrate the changes which occurred in the Watauga River Gneiss (Figure 12A) as it was subjected to ductile deformation under greenschist-facies metamorphic conditions during movement along the Fork Ridge thrust. Clearly identifiable Watauga River Gneiss forms massive cores within the larger massive gneiss and protomylonite boudin, as well as most of the smaller boudin. All exhibit some degree of recrystallization and the surrounding mylonite schist (Higgins, 1971) is largely recrystallized to a greenschist-facies metamorphic assemblage (Figure 12B) with scattered, relic grains of feldspar.

Although our structural analysis of this region is not yet completed one of the more obvious relationships is the (past-ductile-deformation) development of upright folds (Figure 13) with an associated axial planar crenulation cleavage (Figures 12B,C,D). Mapping in this region confirms that development of these folds and associated cleavage is dependent on lithology. Here at STOP 5 the boudin of massive gneiss seem to strongly influence both location and orientation of larger folds. If this influence exists on the map scale, as well, it may account for variations in cleavage and fold orientations without necessitating a subsequent folding event.

Typically small folds developed during the ductile deformation are observable in both thin sections and hand specimens; however, rarely are they observable on a larger scale. Here at STOP 5 a quartz vein (Figure 13) appears to be folded about an axial surface coincident with the mylonitic foliation and refolded by one of the younger folds. We believe that the lack of numerous folds associated with the mylonitic foliation is attributable to a lack of any planar fabric in the Grenville-age plutons prior to the ductile deformation associated with thrusting along the Fork Ridge fault.

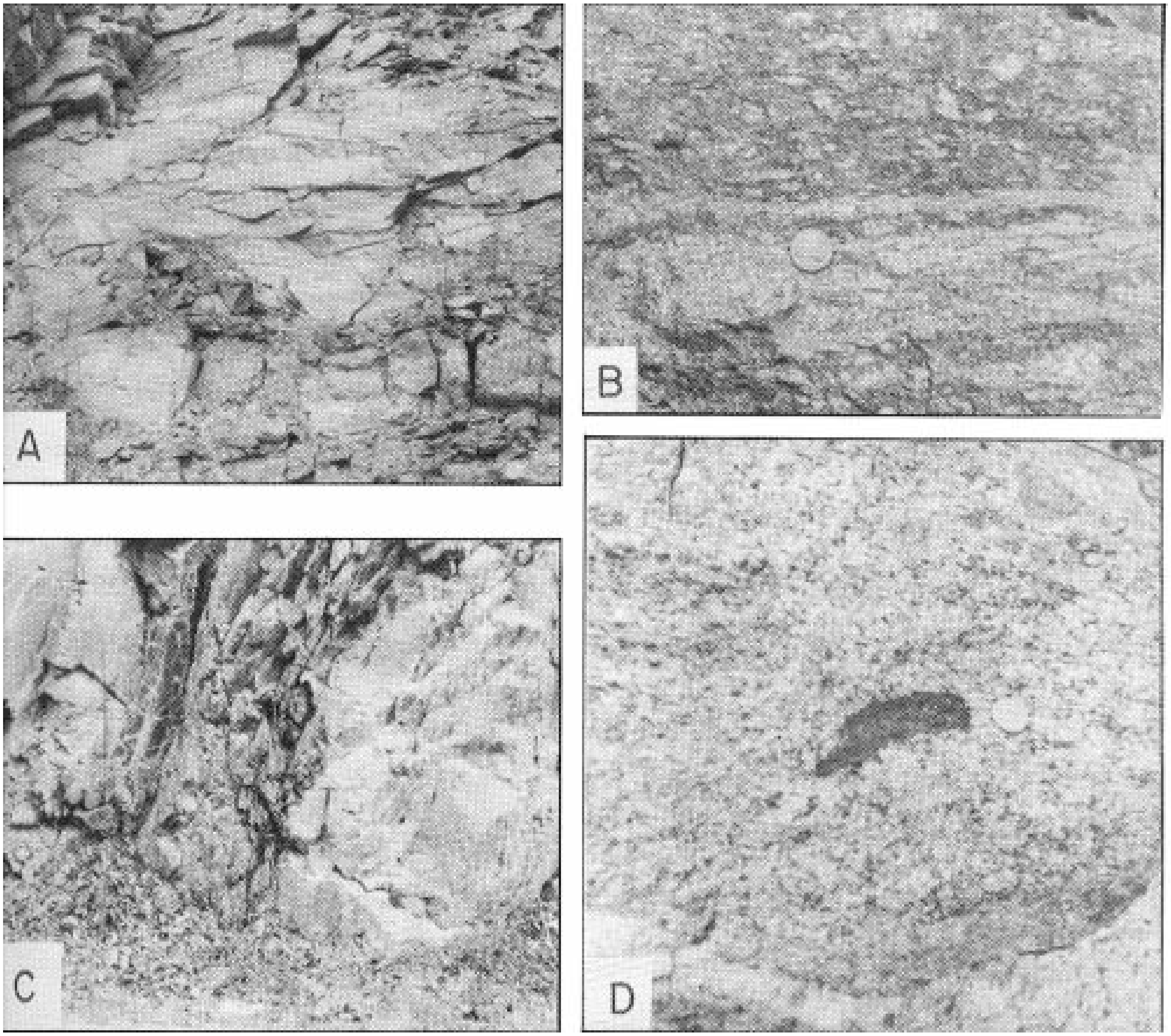


Figure 11: Mylonitic rocks of the Watauga River Gneiss exposed at STOP 4B;
A – shows typical thick (0.5-2 feet) layering of mylonite gneiss and schist.
B – shows closeup of contact between quartzofeldspathic mylonite gneiss (bottom) and mylonite schist (top) with small feldspar augen.
C – Massive granodiorite gneiss of the Watauga River Gneiss with a small xenolith at STOP 4C.
D – Late Precambrian dike (to left of hammer) cutting massive gneiss of Watauga River Gneiss at STOP4C.

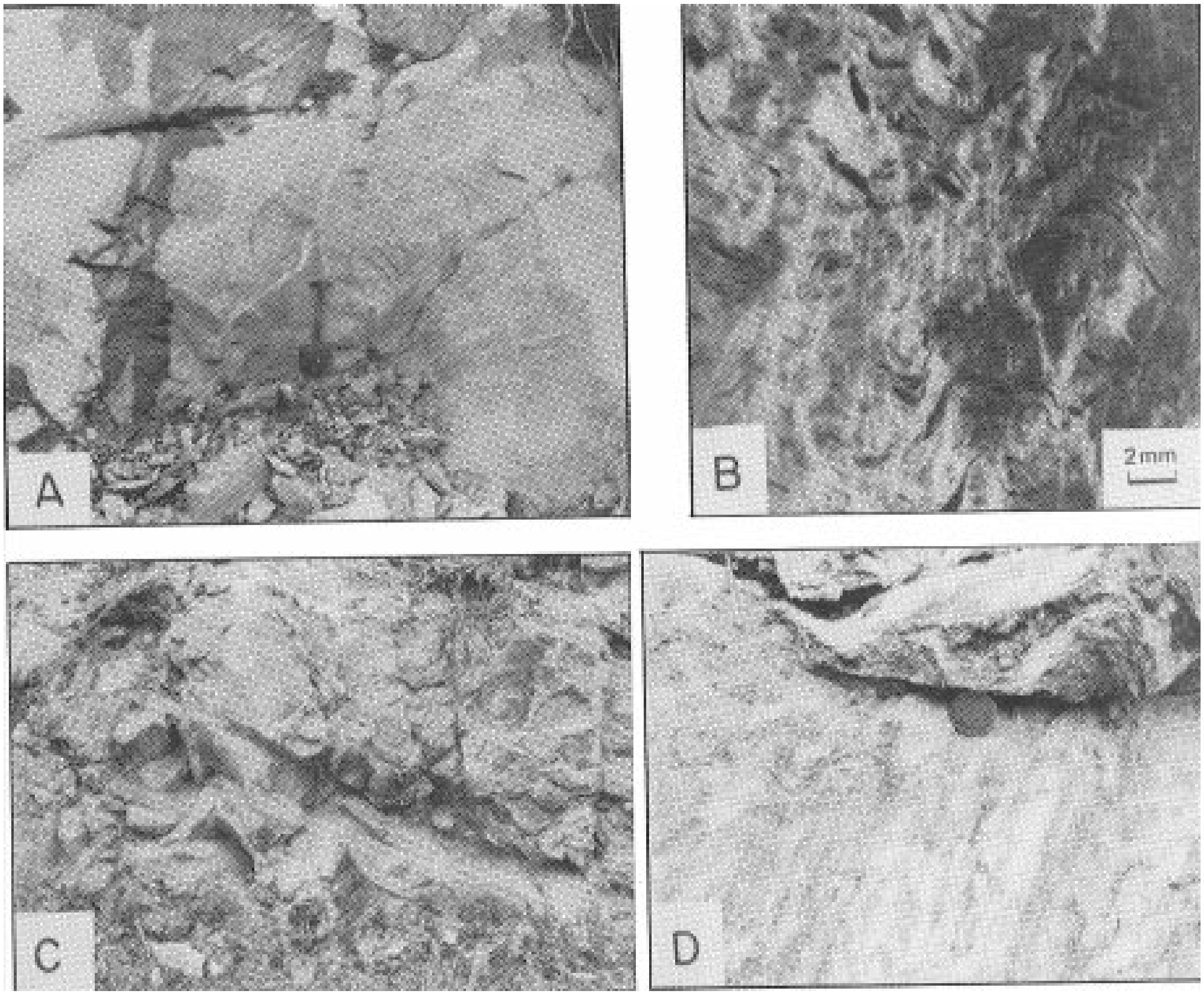


Figure 12: Rocks of the Watauga River Gneiss: A – show the massive granodiorite cut by pegmatitic and apalitic dikelets at the t ype locality at the junction of N.C. state roads 1207 and 1222, 1.2 miles north of Bethel;
B – is photomicrograph of sample collected near STOP 5 showing deformation of mylonitic fabric by crenulation cleavage;
C and D – show large scale folds with axial planar cleavage at STOP 5.

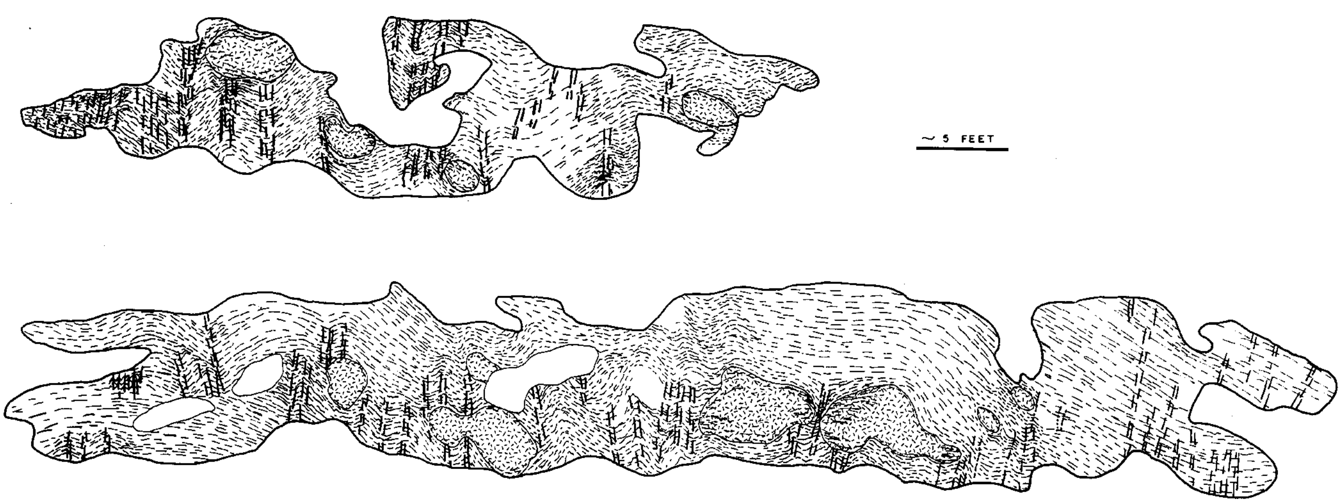


Figure 13: Sketch of outcrops at STOP 5 showing relationships among mylonitic fabric, crenulation cleavage, folds and massive-gneiss boudins.

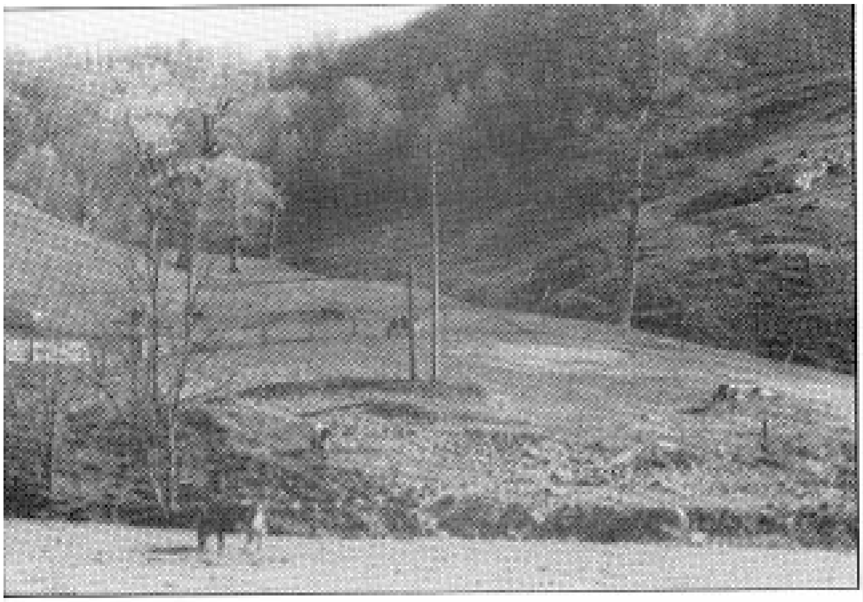


Figure 14: Debris-avalanche deposit on the south side of state road 1201 east of Bethel along the field trip route at about mile 93. Although not plentiful in areas underlain by Watauga River Gneiss, this toe deposit occurs just north of the Georges Gap area. Shows typical conical shape; stream deflection by toe; and longitudinal incision by stream.

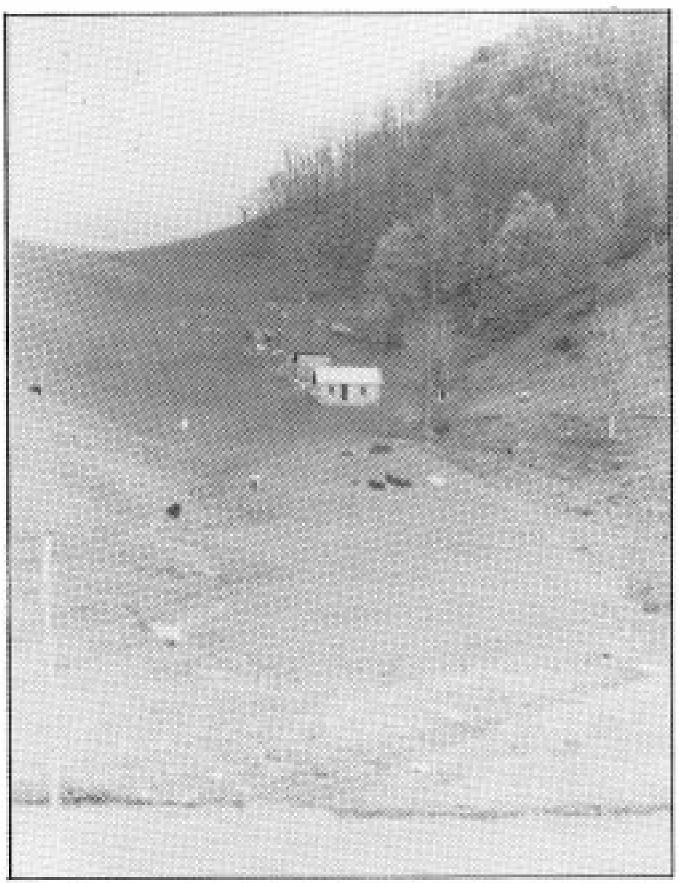


Figure 15: Debris avalanche on the south side of N.C. Highway 88 at about mile 27 along the field trip route (second day); shows: typical shape; longitudinal incision by streams; and exhumed boulder train in upper portion of toe debris.

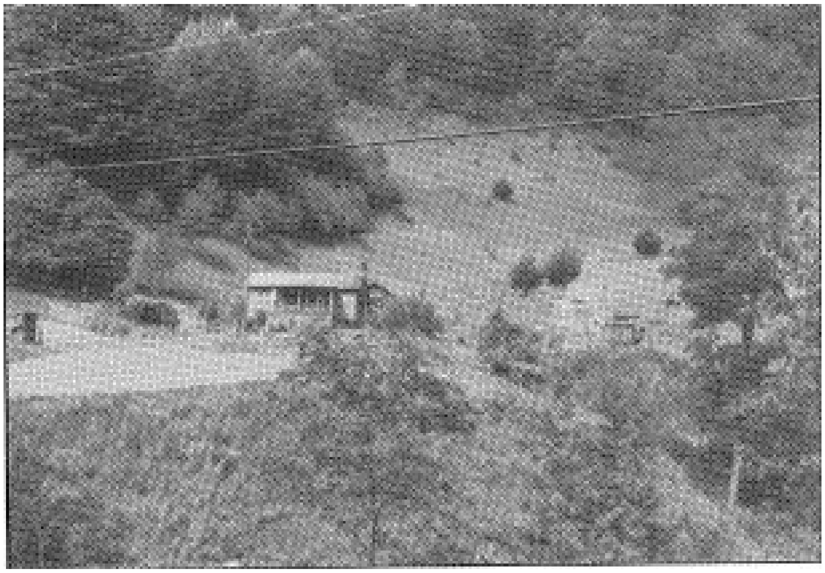


Figure 16: Multiple debris-avalanche track on the east side of U.S. Highway 421 at about mile 45.7 along the field trip route; shows: older, isolated toe-deposit (house and garden); younger debris deposit (in valley) with exhumed boulder train; and deep incision of younger deposit by stream.

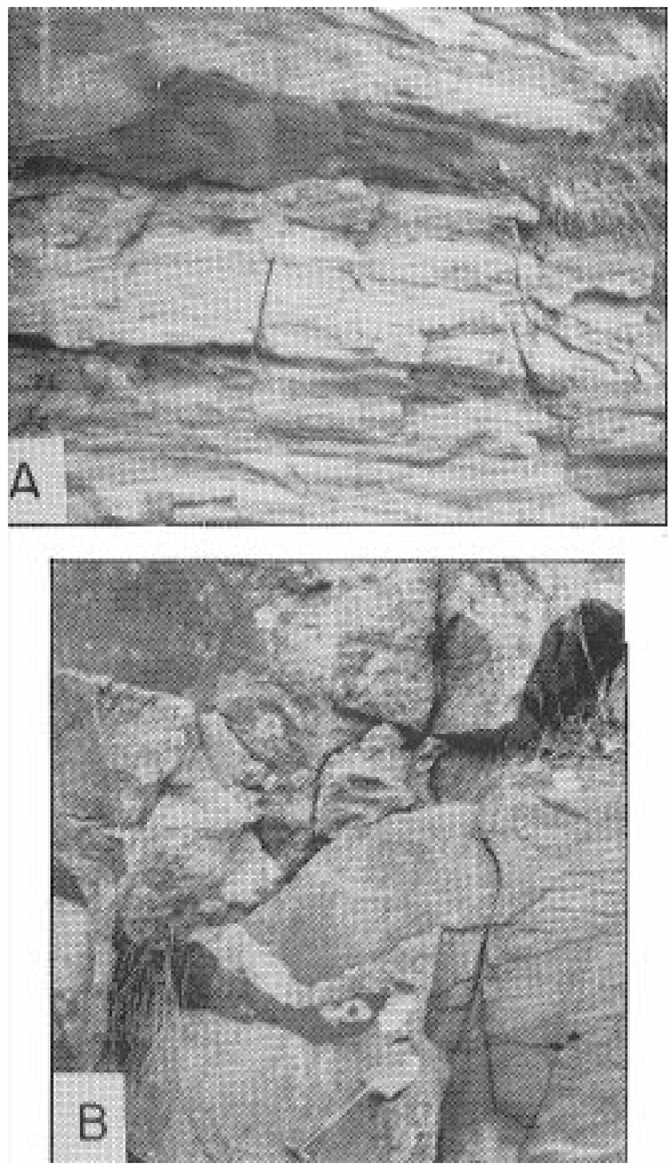


Figure 17: Rock types associated with debris avalanches:
A – Mylonitic gneiss at STOP 6 showing layers of biotite gneiss and quartzofeldspathic gneiss, derived from the Cranberry Gneiss, and dark green, altered mafic dike; fold in extreme upper right;
B – Slightly foliated, coarse-grained, porphyritic granitic gneiss (of the Buckeye Knob Pluton) exposed at mile 95.9.

STOP 6 – DEVELOPMENT OF DEBRIS-AVALANCHE FEATURES IN THE GEORGES GAP AREA (FIGURE 9)

Leaders: Jeffrey J. Gryta, Mervin J. Bartholomew
7.5-minute quadrangle: Sherwood (Bartholomew and Gryta, 1980)
References: Gryta and Bartholomew (this volume)

Although many debris-avalanche features are easily recognizable throughout the mountainous Blue Ridge, few have received any intensive study because of the traditional low population density of the mountain. However, the recreational industry, over recent years, has prompted a steadily increasing number of people to build fairly large, expensive homes in the remote mountains of western North Carolina; and with this sustained development, the geologic community should gradually shoulder more of the responsibility for locating building sites which are relatively "safe" from natural geologic hazards such as debris avalanches and landslides.

Within the region covered on this field trip numerous debris-avalanche deposits of various ages can be observed (Figures 14,15,16); however, within the Georges Gap area a higher than normal concentration of debris-avalanche features exists. The bedrock in this, and adjacent, valleys consists largely of mylonite gneiss and schist derived from the Cranberry Gneiss (Figure 17A) interlayered with altered mafic dikes (Figure 17A) and massive granitic gneiss of the Buckeye Knob Pluton (Figure 17B). These subhorizontal, layered, mylonitic rocks seem to be more susceptible than other basement rocks to development of a thick cover of saprolite. The thick saprolite, in turn, is quite susceptible to development of debris avalanches during periods of intense rainfall.

Typical features which are easily recognizable on these debris-avalanche features are: the wedge or conical-shaped toe deposit; the U-shaped track in the upper part of the valley; boulder-trains partially exhumed during subsequent erosion; and incision or truncation of toe-deposits by streams flowing transverse to the main transport direction of the debris avalanche.

STOP 7 – DEFORMATION OF THE CRAN BERRY GNEISS ALONG ROUNDABOUT CREEK (FIGURE 9)

Leaders: Mervin J. Bartholomew, Sharon E. Lewis
7.5-minute quadrangle: Baldwin Gap (Bartholomew, 1983)
References: Bartholomew and others (this volume)
Fullagar and Bartholomew (this volume)
Bartholomew and Lewis (in press)

The outcrops (Figure 18) here at Stop 7 are one of those often-hoped-for-but-rare exposures where crosscutting relationships and multiple folding events are reasonably distinguishable. The dike is lithologically similar to those associated with, and cutting, the Beech granite. Thus it is inferred to be late Precambrian in age. The dike contains folded quartz veins and a mica foliation which is axial planar to these folds. This generation of folds and foliation is inferred to be Paleozoic in age based on the inferred age of the dike. Likewise, folds and foliation of similar orientation in the adjacent country rocks are clearly associated with the retrograde greenschist metamorphism which accompanied the ductile deformation during movement on the Fries fault system. The dike truncates older folds whose axial surfaces are parallel to the segregation layering in the Cranberry Gneiss. Based on the interpretation of the radiometric age determination (Fullagar and Bartholomew, this volume) for the Cranberry Gneiss we believe that this early generation of folds and segregation layering were developed during Grenville metamorphism about 1000 my ago. This locality was among those sampled for the geochronology work.

STOP 8 – DEVELOPMENT OF LIT-PAP-LIT FEATURES IN A LATE-PRECAMBRIAN DIKE (FIGURE 9)

Leader: Sharon E. Lewis
7.5-minute quadrangle: Mountain City (reconnaissance)
References: Bartholomew and others (this volume)
Bartholomew and Lewis (in press)
Fullagar and Odom (1973)

This unusual dike (Figure 19) was noted in 1977 during reconnaissance mapping in the Mountain City quadrangle. In addition to the lit-par-lit injections this dike also differs both texturally and mineralogically from the more typical, mafic to intermediate, fine-grained dikes believed to be associated with the Mount Rogers volcanic rocks that overlie the Watauga Massif. Its coarser grain size and the presence of large relict(?) amphibole suggest that it was perhaps relatively "dry" and hence less susceptible to Paleozoic retrograde metamorphism than the other dikes which were extensively retrograded to greenschist-facies mineral-assemblages. Several features observed here suggest a probable origin for the lit-par-lit features. First, the fine-grained granitoid material forming the lit-par-lit features (Figure 20A) is mineralogically similar to the enclosing coarse-grained granitoid gneiss of the Forge Creek Suite (Bartholomew and Lewis, in press). Second, the fine-grained granitoid material is confined either to the dike or to the border area of the dike (Figures 19,20). Third, no fine-grained granitoid rocks were found within this portion of the Forge Creek Suite.

An explanation which is consistent with these observations is that the dike was intruded at a relatively high temperature and/or high pH_2O. Thus the enclosing granitoid rock

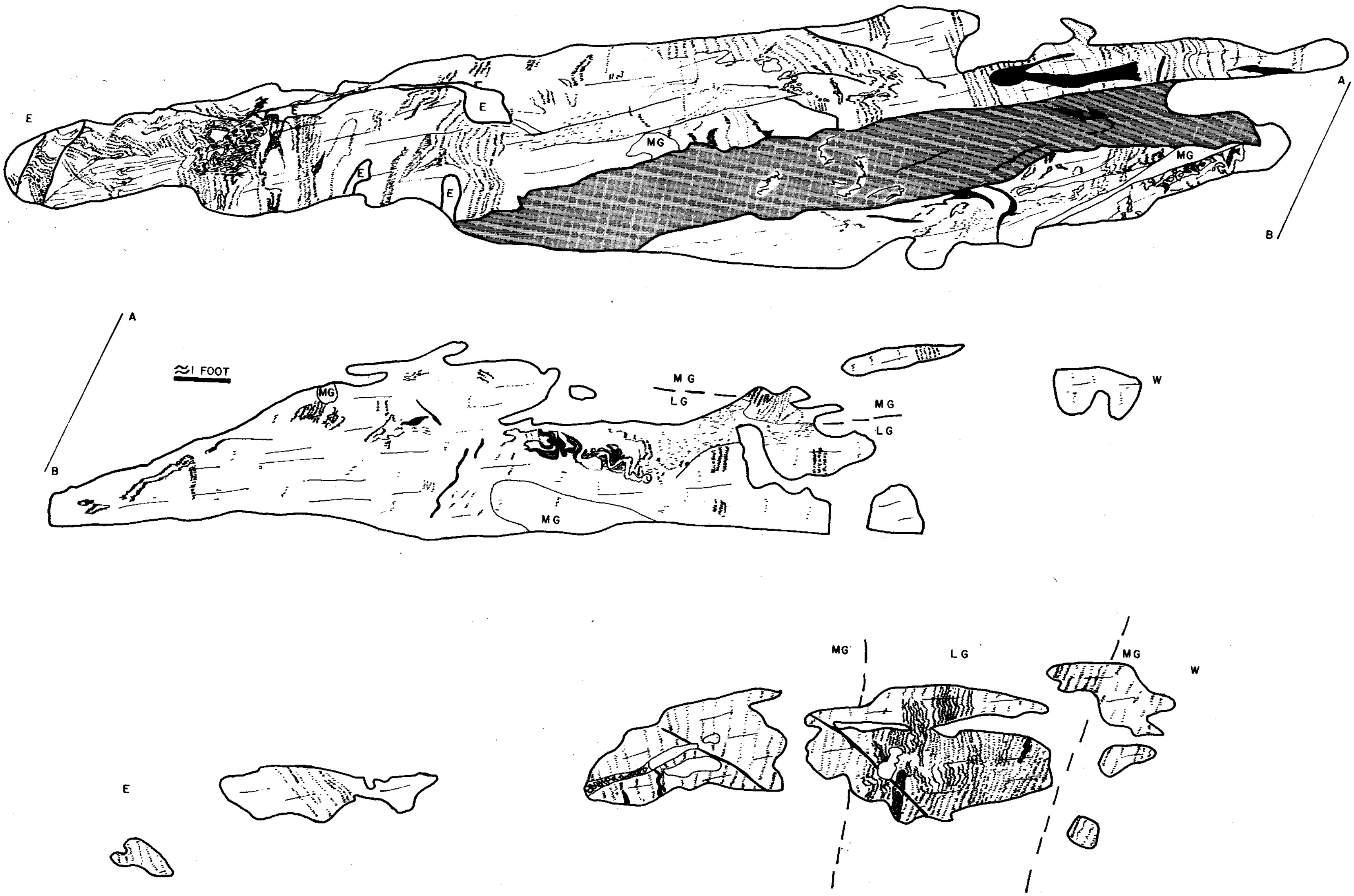


Figure 18: Sketches of outcrops of complexly folded Cranberry Gneiss at STOP 7 along Roundabout Creek. Shaded area is late Precambrian dike; lower portion of diagram is to the west of the outcrop with the dike. MG - massive to slightly foliated gneiss; LG – layered gneiss; e – epidotized regions.

was partially mobilized and then was injected into the dike and chemically reacted with the dike to produce the present rocks. These rocks were folded (Figure 20) and altered subsequently during Paleozoic metamorphism. If this type of lit-par-lit development was extensive (but largely unrecognized) in this region it also would provide a possible explanation for the anomalous 871 my Rb-Sr age reported by Fullagar and Odom (1973) for nearby rocks from the Forge Creek Suite. This age is interpreted as anomalous due to partial re-equilibration of the Rb-Sr system during late-Precambrian igneous activity associated with Mount Rogers volcanic rocks.

STOP 9 – MYLONITIC ROCKS OF THE ELK PARK MASSIF ALONG U.S. 421 (FIGURE 9)

Leaders: Mervin J. Bartholomew, Sharon E. Lewis
7.5-minute quadrangle: Sherwood (Bartholomew and Gryta, 1980)
References: Bartholomew and others (this volume)
Bartholomew and Lewis (in press)

Ductile deformation of basement rocks of the Elk Park Massif produced coarsely layered mylonitic gneiss in which a variety of rock types, that originally had crosscutting relationships, were transposed to such a degree that virtually all contacts are now subparallel (Figures 21,22A). At this STOP, three distinctive rock types are recognizable discrete layers. The layered biotite gneiss is easily recognizable as derived from one of the lithologies of the Cranberry Gneiss. The lighter-colored granitoid lenses (Figure 22A) are part of the Laurel Creek Pluton (Bartholomew and Lewis, in press). This Grenville-age pluton intruded the Cranberry Gneiss and today forms a large portion of the Elk Park Massif northeast of STOP 9. The dark green rocks (Figure 22A,B) are some of the coarse-grained feeder dikes for the Snake Mountain volcanic pile.

STOP 10 – THE ASHE-BASEMENT CONTACT ALONG U.S. 421 (FIGURE 9

Leaders: Mervin J. Bartholomew, Sharon E. Lewis
7.5-minute quadrangle: Zionville (Bartholomew and Wilson, in press)
References: Bartholomew and others (this volume)
Bartholomew and Lewis (in press)

Here at STOP 10 is one of the few places where the actual contact between the Ashe Formation and the underlying basement rocks of the Elk Park Massif is visible (Figure 23A). A short walk along both the sketched portion of the outcrop (Figure 24) and the unsketched portion of the exposures (to the south) should be sufficient to demonstrate that the basement rocks of the Elk Park Massif are intimately

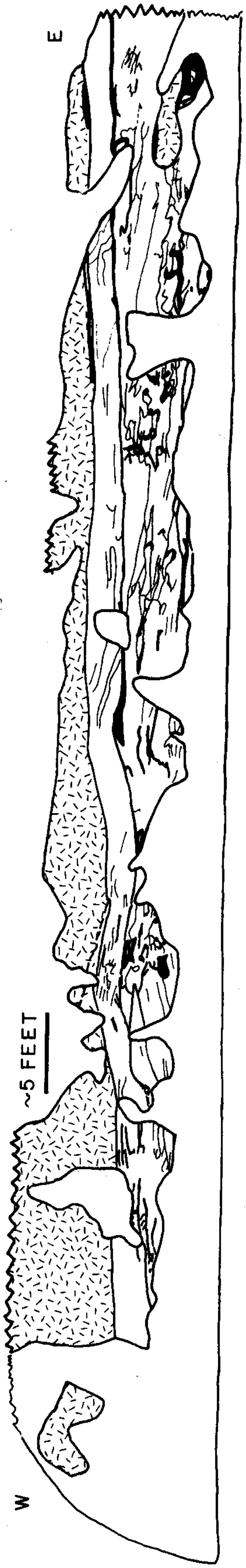


Figure 19: Sketch of a portion of the outcrop at STOP 8 showing mafic dike, with lit-par-lit features, cutting granitoid of the Forge Creek Suite.

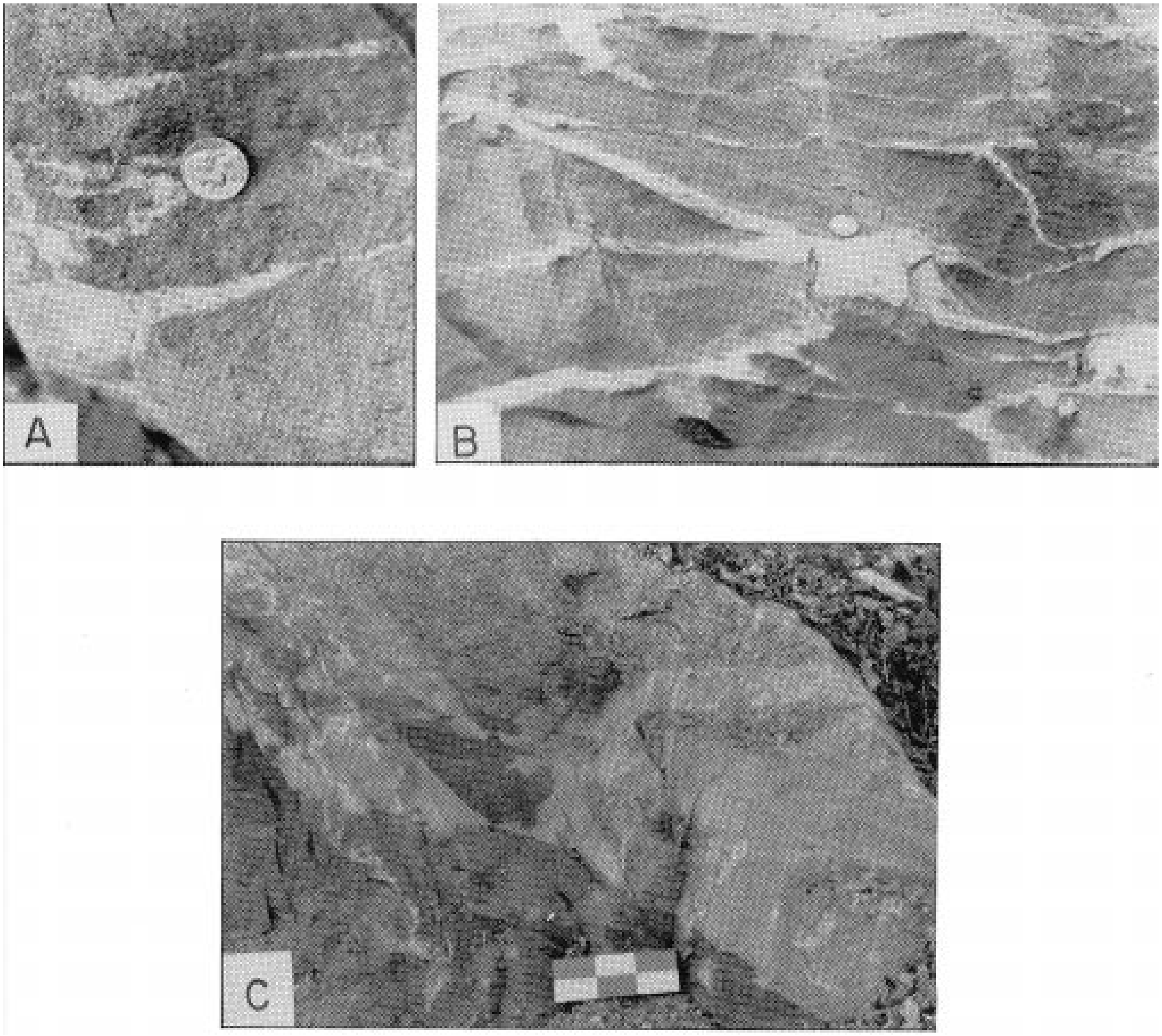


**Figure 20: Mafic dike with lit-par-lit features at STOP 8: A – shows folded granitoid lense with axial planar-foliation;
B – shows folds (upper left, lower right) and lit-par-lit features within dike;
C – shows fine-grained granitoid material cutting coarse-grained granitoid gneiss at dike margin.**

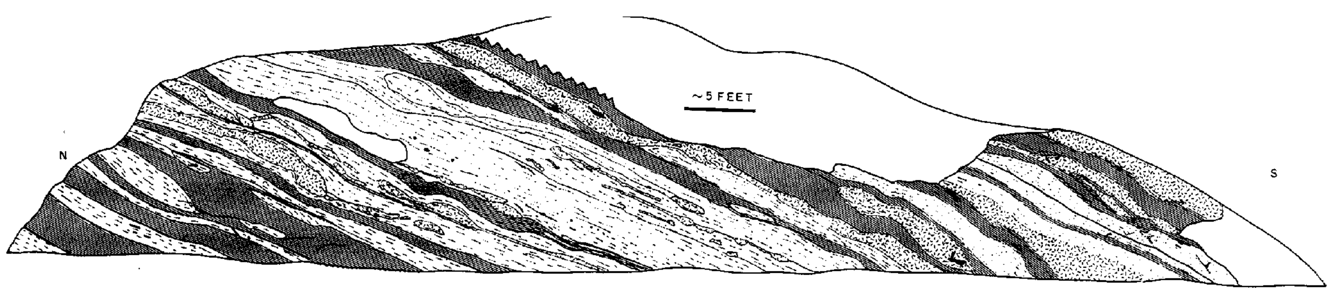


Figure 21: Sketch of mylonitic rocks at STOP 9 showing altered amphibolite (shaded), biotite-gneiss (dashes) and granitoid lenses in subparallel layers.

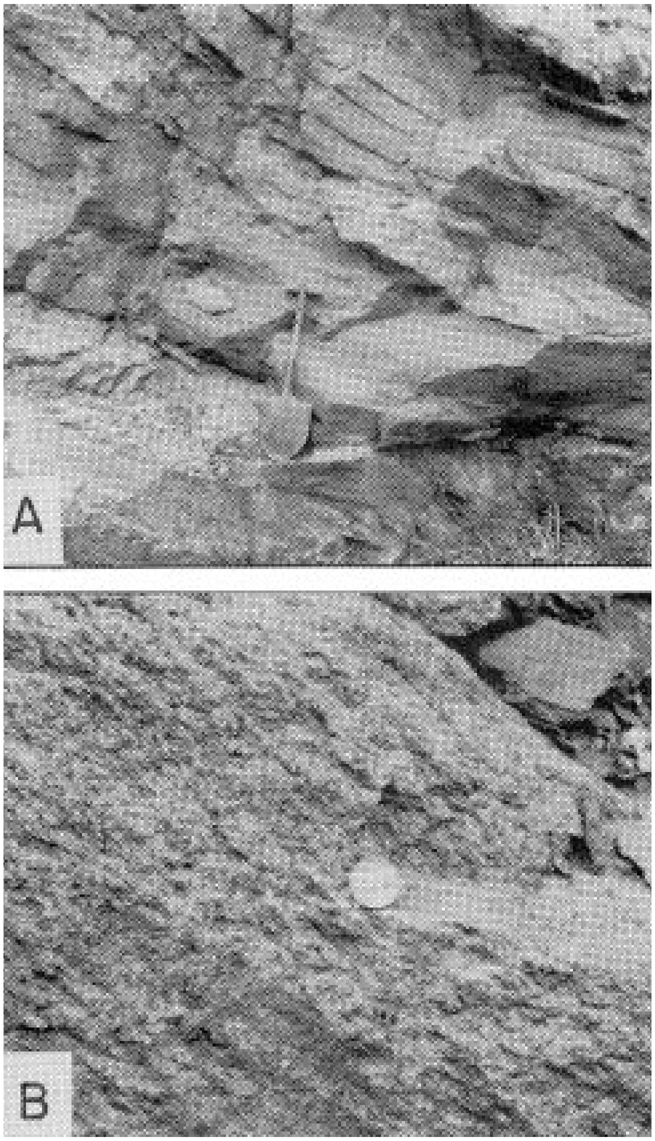


Figure 22: Rocks exposed at STOP 9: A – shows dark green altered amphibolite and amphibole gneiss, granitoid lenses and layered Cranberry Gneiss;
B – shows altered mafic dike with relict coarse-grained amphibole.

folded along with the overlying Ashe Formation. Thus both the inferred trace of the Fries fault by Rankin and others (1972) and the inference by Hatcher (1978) that the Fries thrust sheet lacks Grenville-age basement are both in serious error! Our mapping in these quadrangles indicates that the Ashe Formation rests nonconformably on a significant portion of Grenville-age crystalline rocks (Elk Park Massif). It is these Grenville-age rocks, and not the Ashe Formation, which are in thrust contact along the Fries fault with the Watauga Massif and its cover rock.

The dominant folds visible in these outcrops (Figures 24, 23B,C) were mostly formed during the second phase of Paleozoic folding (Fp_2). These folds are easily recognized in the siliceous layers where an older mica foliation can be traced around individual fold hinges. Although the dominant, visible folds may be Fp_2 we believe that the innumerable repetitions of the siliceous layer are actually due to earlier Paleozoic folding (Fp_1). This earlier phase of isoclinal folding produced greatly attenuated limbs, thus many sharp contacts between siliceous layers (some are marked by thin layers of amphibolite) are inferred to actually be Fp_1 axial surfaces. A few Fp_1 hinges also can be seen in this outcrop.

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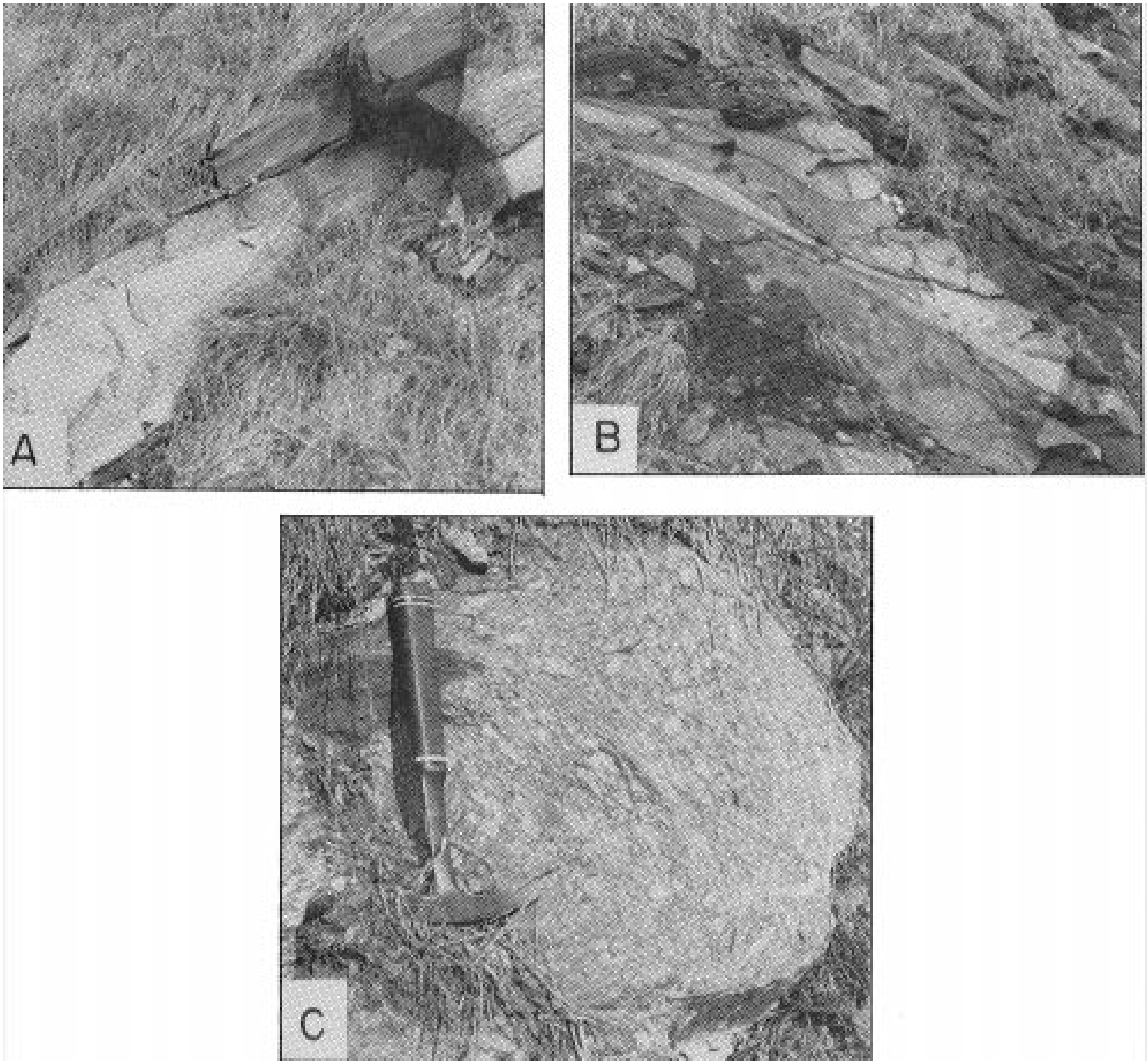

Figure 23: Rocks exposed at STOP 10: A – shows outcrop on west (right) side of U.S. 421 with hammer on contact between amphibolite (top) and layered augen gneiss (Cranberry Gneiss) with lenses of granitoid (Laurel Creek Pluton); Paleozoic-age folds about which earlier foliation is folded; C – fold in augen gneiss of Cranberry Gneiss in outcrop to south of sketch (Figure 24).

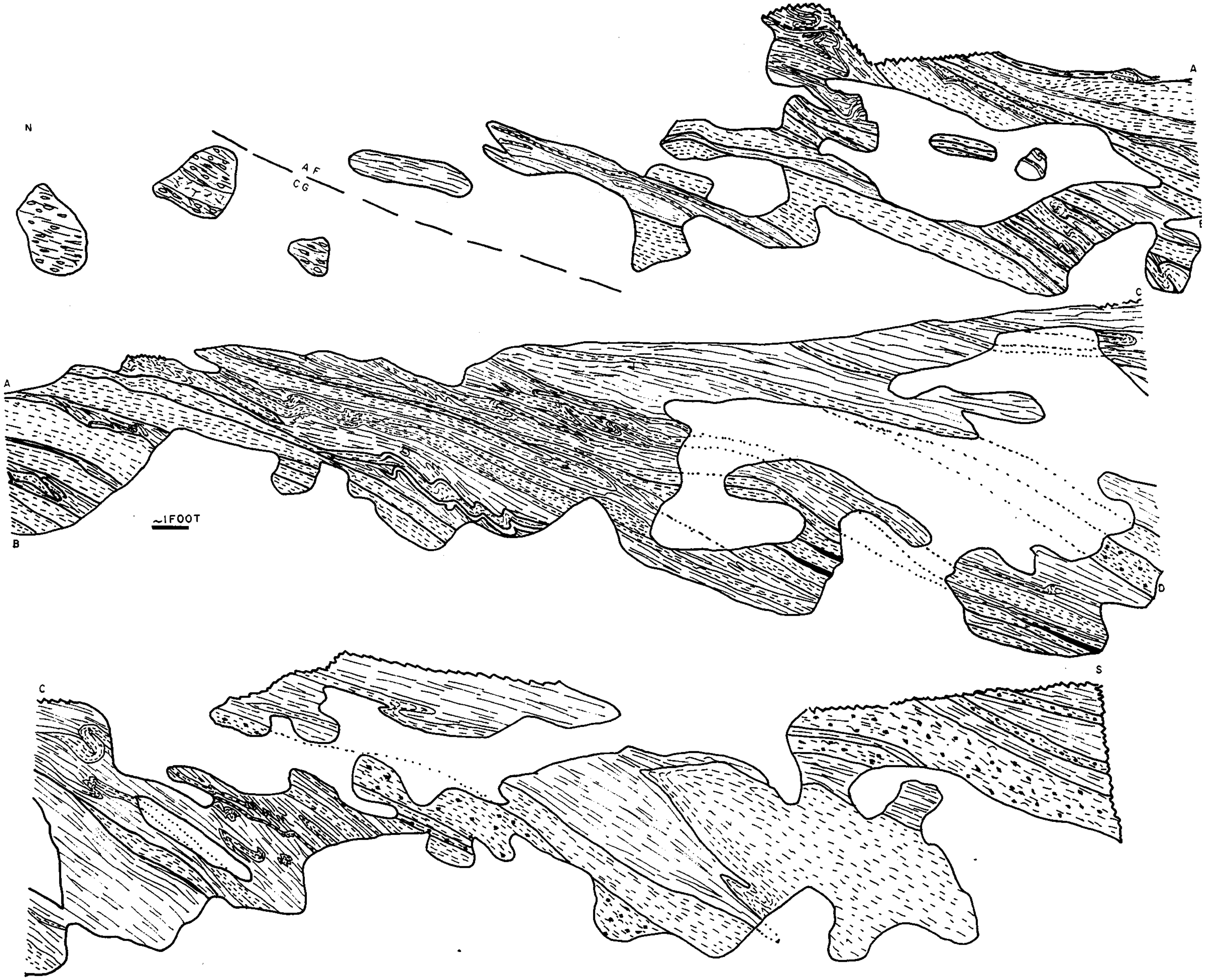

Figure 24: Sketch of outcrop on east (left) side of U.S. Highway 421 showing Cranberry augen gneiss (CG), with granitoid lense, below Ashe Formation (AF); siliceous gneiss (short dashes); biotite gneiss with small feldspar augen (dashes with small dots); amphibolite and amphibole gneiss (long dashes); dashes outline foliation.

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