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GEOLOGIC HISTORY OF A PORTION OF THE EASTERN BLUE RIDGE, SOUTHERN APPALACHIANS: TRAY MOUNTAIN AND MACEDONIA 7 1/2 ' QUADRANGLES, GEORGIA

A Thesis

Presented for the

Master of Science

Degree

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Elizabeth A. McClellan

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ABSTRACT

The eastern Blue Ridge of northeastern Georgia and southwestern North Carolina is underlain by rocks of the Tallulah Falls Formation, Coweeta Group, and Otto Formation, separated from the Ocoee Supergroup in the western Blue Ridge by the Hayesville fault. The study area encompasses portions of the Tray Mountain and Macedonia 7 1/2 " quadrangles in northeast Georgia, and can be divided into three tectonic units. Tectonic unit 1 is the lowest, and includes the Otto Formation (mica schist and metasandstone) and intercalated amphibolite. It is bounded on the northwest by the Soque River fault. Tectonic unit 2, the next highest, comprises dominantly metasedimentary rocks of the Coweeta Group, underlain by possible Tallulah Falls Formation equivalents. The Coweeta Group can be divided into the Persimmon Creek Gneiss, Coleman River Formation, and Ridgepole Mountain Formation. Tectonic unit 3, the highest, includes an assemblage of garnetiferous mica schist, metasandstone, amphibolite, and hornblende gneiss, informally designated the Corn Ridge formation. This unit is separated from tectonic unit 2 by the Shope Fork fault.

All of the sequences above, with the exception of the Persimmon Creek orthogneiss, are dominated by psammitic and pelitic rocks that originated as thick clastic sequences of greywacke with interbedded shale. Textural evidence indicates that the rocks experienced two episodes of prograde mineral growth, the first episode reaching sillimanite or kyanite grade, and the second garnet grade. These events may be correlated with the Taconic and Acadian orogenies. Subsequent retrograde metamorphism affected only rocks of tectonic unit 3. The rocks are no younger than early Paleozoic, constrained by the age of Taconic metamorphism in the southern Appalachians (430-470 m.y.).

Multiple deformation resulted in six recognizable folding phases; earlier flowage folds resulted in transposition of layering into a regionally penetrative foliation, later overprinted by flexural flow and flexural slip buckle folds. In the study area, the Shope Fork fault juxtaposes rocks of different metamorphic grade and cuts steep, northwest-trending F_5 folds, indicating some post-metamorphic movement. The Soque River fault, recognized primarily by truncation of stratigraphy, has an uncertain movement history.

The Coweeta Group lies within the core of a regionally extensive syclinorium, in a structurally similar position to that of the Alligator Back Formation of North Carolina and Virginia and the Murphy belt rocks in North Carolina and Georgia. Based on lithologic comparisons, the Coweeta Group cannot be positively correlated with these units. If equivalent to one or both, the Coweeta Group must represent a considerably different facies. In addition, petrography and heavy mineral analysis suggest that rocks of this portion of the eastern Blue Ridge had a different source area from that of the Ocoee Supergroup immediately to the west, within the Great Smoky thrust sheet.

TABLE OF CONTENTS

SECTION		PAGE
I.	INTRODUCTION	1
	A. Study Area	3
	B. Objectives	4
	C. Methods	5
	D. Previous Investigations	6
II.	REGIONAL GEOLOGY	7
	A. Blue Ridge: Divisions	8
	B. Western Blue Ridge: Stratigraphy	10
	1. Basement Rocks	10
	2. Ocoee Supergroup	11
	3. Other Late Precambrian Stratified Rocks	12
	4. Chilhowee Group	12
	5. Murphy Belt	13
	6. Talladega Slate Belt	14
	B. Western Blue Ridge: Structure	14
	C. Eastern Blue Ridge: Stratigraphy	16
	1. Late Precambrian Stratified Rocks	
	2. Late Precambrian-Early Paleozoic (?) Stratified Rocks	18
	3. Other Eastern Blue Ridge Rocks, Georgia and North Carolina	21
	4. Basement Rocks, Mafic Rocks and Plutons	21
	D. Eastern Blue Ridge: Structure	23
	1. Folding	23
	2. Synclinoria	23
	3. Domes and Windows	24
	4. Faulting	25
III.	STRATIGRAPHY AND LITHOLOGIC UNITS	27
	A. Terminology	27
	B. Description of Lithologic Units	28
	1. Tectonic Unit 1	29
	2. Tectonic Unit 2	34
	3. Tectonic Unit 3	43
IV.	SEDIMENTARY PETROGRAPHY	46
	A. Sandstone Classification	46
	1. Methods	47
	2. Classification	48
	B. Discussion	55
	1. Biotite Content	63
	2. Absence of K-feldspar	64
	C. Heavy Mineral Analysis	66
	1. Methods	67

2. Detrital Heavy Minerals: Discussion 67
3. Comparison with other Blue Ridge Rocks 71

V. METAMORPHISM 74
A. Isograds 74
B. Metamorphic Assemblages and Reactions 76
1. Characteristic Assemblages and Textures 76
2. Reaction Relationships 84
3. Migmatization 86
F. Metamorphic Conditions 89
G. Retrograde Effects 91

VI. STRUCTURAL GEOLOGY 93
A. Mesoscopic Structures 93
1. Foliations 93
2. Mesoscopic Fold Styles 103
3. Fold Generations 110
4. Lineations 118
5. Dikes and Veins 118
6. Joints 121
B. Macroscopic Structures 121
1. Macroscopic Folds 121
2. Domain Analysis 124
3. Faulting 134
4. Cross Sections 137

VII. DISCUSSION AND INTERPRETATIONS 140
A. Significance and Interpretation of Rock Types 140
1. Metasedimentary Rocks 140
2. Mafic Rocks 145
3. Ultramafic Rocks 146
B. Significance and Interpretation of Metamorphism and Structure 146
1. Metamorphic Textures 146
2. Timing of Metamorphism 149
3. Comparison of Deformational Sequence with Previous Work in the Region 151
4. Timing of Faulting 153

VIII. CONCLUSIONS 155

LIST OF REFERENCES 157

APPENDICES 170

APPENDIX A 171

APPENDIX B 178

VITA 180

LIST OF TABLES

TABLE PAGE

1. Range (A) and average mineral composition (B) of metagreywackes from the Tallulah Falls Fm., Coleman River Fm., and Corn Ridge Fm., in volume percent..... 59

2. Range (A) and average mineral composition (B) of 88 samples of greywackes from the Harz Mountains, Germany..... 60

3. Range (A) and average mineral composition (B) of arkosic and feldspathic metasandstones from the Coleman River Formation, in volume percent..... 61

4. Mineral compositions of arkoses from the Auvergne in France..... 62

5. Heavy mineral concentrations of samples from the Coleman River and Tallulah Falls Formations..... 68

6. Comparison of primary heavy mineral suites in some rocks of the southern Appalachian Blue Ridge..... 72

7. Fold generations and associated foliations in the study area..... 114

8. Comparison of deformation sequence and timing (this study) with previous work in the region..... 152

A-1. Modal analyses of metasandstones..... 170

A-2. Quartz, feldspar, and mica contents of metasandstone samples..... 176

B-1. Mineral abbreviations used in text in discussion of metamorphic assemblages and reactions..... 178

LIST OF FIGURES

FIGURE		PAGE
1.	Index map covering portions of the Georgia, North Carolina, and South Carolina Blue Ridge and Inner Piedmont	2
2.	Crenulated muscovite-biotite schist of the Otto Formation	30
3.	Photomicrograph of crenulated actinolite schist	33
4.	Outcrop of metasandstone of the Coleman River Formation	36
5.	Photomicrograph of "biotite quartzite" of Tallulah Falls Formation	41
6.	Hand sample of garnet-chlorite schist of Corn Ridge Formation	44
7.	Classification of sandstones	49
8.	Distribution of sandstone types within the study area	56
9.	Stratigraphic relationships of rock sequences in tectonic unit 2	58
10.	Typical habits of zircons from metasandstone of the Coleman River and Tallulah Falls Formation (this study)	70
11.	Typical habits of zircons in sandstones (from Poldervaart, 1955)	70
12.	Distribution of aluminum silicates in the study area	75
13.	Photomicrograph of sub- to euhedral, inclusion-free garnets from the Otto Formation	78
14.	Radiating sillimanite fibers at quartz grain boundaries	79
15.	Photomicrograph of anhedral garnet with inclusions of biotite and quartz	82
16.	Photomicrograph of sillimanite gneiss from the Ridgepole Mountain Formation	83
17.	Hand sample of migmatitic gneiss from the Coleman River Formation	88
18.	Petrogenetic grid for the system Al_2SiO_5	90

19. Orientation and observed distribution of foliations in the study area..... 94

20. Biotite defining S_1 (pinstriping), developed as an axial planar foliation to F_1 fold..... 96

21. Typical metasandstone from the study area..... 97

22. Typical S_2 foliation..... 98

23. Development of axial planar schistosity S_3, within F_3 folds..... 100

24. S_3 foliation defined by biotite and fibrolitic sillimanite..... 101

25. Photomicrograph of pelitic rock from the Coleman River Formation... 102

26. Fold classification scheme based on shape and amplification..... 104

27. Dip isogon characteristics of main fold classes..... 104

28. Dip isogon analysis of type 1 fold, Tallulah Falls Formation..... 106

29. Dip isogon analysis of type 1 fold, Otto Formation..... 107

30. Dip isogon analysis of type 2 fold, Coleman River Formation..... 108

31. Dip isogon analysis of type 2 fold, Coleman River Formation, Tray Mountain..... 109

32. Flexural fold with fanning fractures..... 111

33. Ptygmatic folds in quartzofeldspathic veins..... 112

34. Dip isogon analysis of ptygmatic folds..... 113

35. Point diagram of (a.) poles to axial surfaces and (b.) axes of small kink folds within the study area..... 116

36. Type 2 interference pattern resulting from F_3 folds refolded by F_5..... 117

37. Point diagram of mineral lineations (dots), intersection lineations (squares), and sillimanite fibers (stars) from study area..... 119

38. Quartzofeldspathic dikes, earlier ones offset by later set..... 120

39. Point diagram of poles to joint surfaces from study area..... 122

40. Distribution of map-scale folds within the study area..... 123

41. Map of domain boundaries for domain analysis of study area..... 125

42. Domain I: Equal area plots of poles to axial surfaces and fold axes of mesoscopic folds..... 127

43. Domain II: Equal area plots of poles to axial surfaces and fold axes of mesoscopic folds..... 128

44. Domain III: Equal area plots of poles to axial surfaces and fold axes of mesoscopic folds..... 130

45. Domain IV: Equal area plots of poles to axial surfaces and fold axes of mesoscopic folds..... 131

46. Cylindrical best fit for foliations and mesoscopic fold axial surfaces from Domain IV..... 132

47. Domain V: Equal area plots of poles to axial surfaces and fold axes of mesoscopic folds..... 133

48. Results of Fry analysis of samples across the Soque River fault..... 138

LIST OF PLATES

PLATE

I. Regional Geologic Map of the Eastern and Western Blue Ridge, Southern Appalachians..... In Pocket

II. Geology of Portions of the Tray Mountain and Macedonia 7 1/2' Quadrangles, Georgia..... In Pocket

III. Form Surface Map for Dominant Foliation (S_2) in Study Area Including Orientation Data for S_2 in Individual Domains.... In Pocket

IV. Cross Sections Along Lines A-A', B-B', and C-C', on Plate II..... In Pocket

I. INTRODUCTION

The Appalachian Mountains of the eastern United States have been the laboratory and training ground for scores of geologists over the years, many of whom regard the chain as a classic orogen. A majority of early studies concentrated on the northern and central Appalachians, from Pennsylvania northward, although notable works by Safford (1856, 1869), Keith (1895, 1903, 1904, 1907) and others laid important groundwork in the geology of the southern Appalachians. In recent years, a number of studies have focused on both reconnaissance work and detailed mapping in the southern Appalachians. Many questions have been answered; however, just as many new questions have arisen as a result.

The southern Appalachians, for purposes of this thesis, are considered that portion from Virginia to Alabama. The study area for this thesis lies within the Blue Ridge province of the southern Appalachians, between the Valley and Ridge foreland fold and thrust belt and the Brevard fault zone (Fig. 1). The Blue Ridge, which contains unmetamorphosed to high-grade metamorphic rocks, is longitudinally divided into western and eastern segments. The boundary between these segments is a major fault system that marks an abrupt change in lithologies and tectonic environment. West of this boundary sequences of late Precambrian rifted margin clastic rocks or interlayered metavolcanic and clastic rocks that were deposited on 1.1 b.y. (Grenville) basement are overlain by Cambrian shelf clastic and carbonate deposits (Hatcher, 1978). These clastic and carbonate rocks can convincingly be linked to rocks of the foreland.

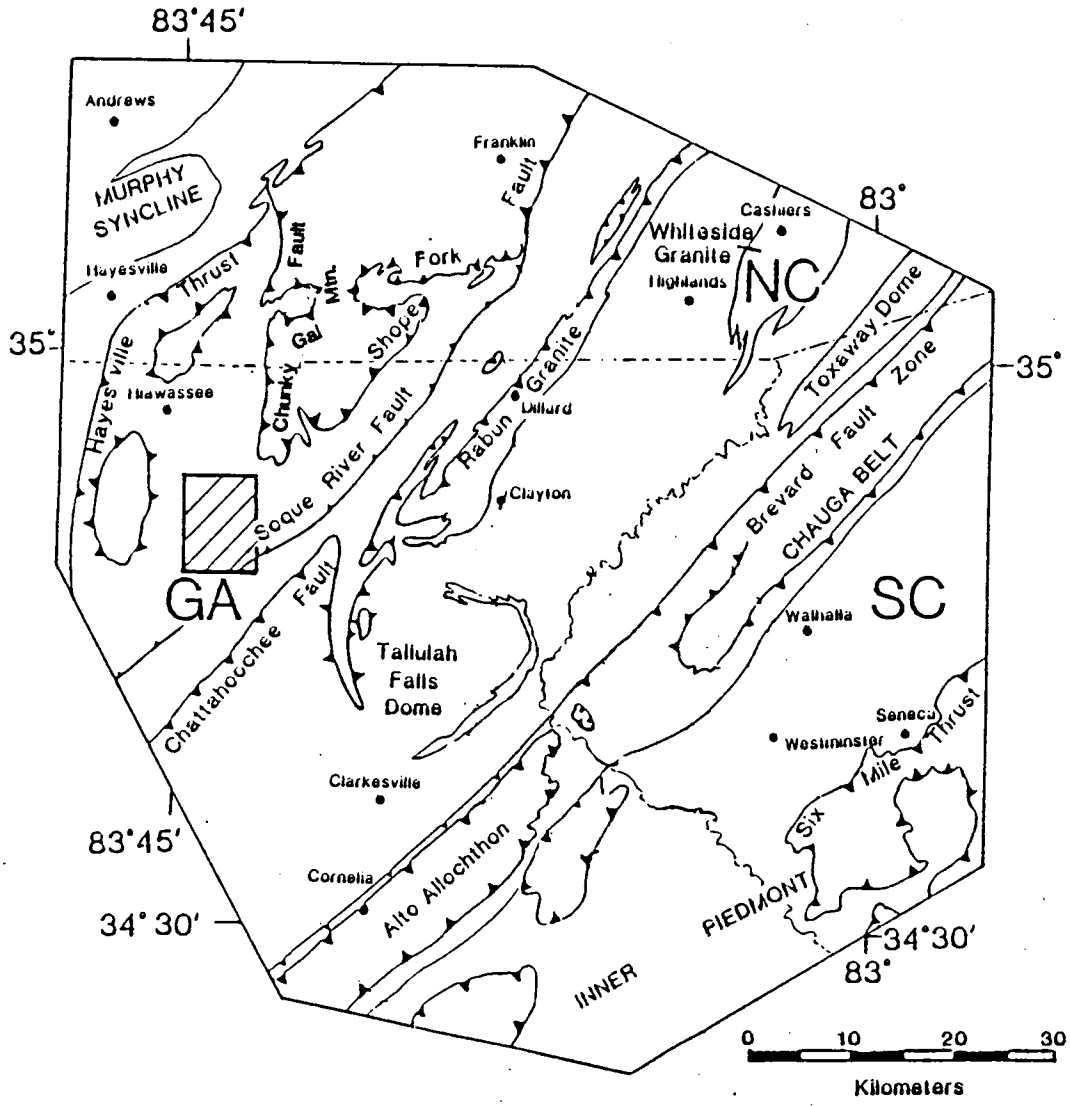


Figure 1: Index map covering portions of the Georgia, North Carolina, and South Carolina Blue Ridge and Inner Piedmont. Location of study area shown in block. (Modified from Hopson and Hatcher, 1987.)

East of the boundary, the rocks appear to be predominantly of deeper-water origin, and are commonly interlayered with mafic volcanic rocks. Numerous granitic plutons and ultramafic bodies are present. Stratigraphic relationships are highly obscured by metamorphism and deformation.

Despite the number of high-quality detailed studies that now exist, many more will be necessary before structures, local stratigraphic relationships and correlations both within the eastern Blue Ridge and across its boundaries can be described satisfactorily. This study strives to integrate detailed field mapping, petrography and structural analysis in order to investigate a small portion of that region, and to place that portion into a regional context.

Study Area

The study area encompasses portions of the Tray Mountain and Macedonia 7 1/2" quadrangles, located within the central part of the eastern Blue Ridge of northeastern Georgia (Fig. 1). The terrain is rugged and mountainous, with a maximum elevation of 4430 ft (1350 m) at the peak of Tray Mountain. A major part of the area lies within the Chattahoochee National Forest, where vegetation is abundant and roads are scarce. Narrow valleys that frame the area on the north, east, and west contain mostly farm and pasture land. A system of east-west striking ridges defines the Tennessee Valley Divide, which becomes coincident with the Eastern Continental Divide as the trend of the ridges bends northward. A portion of the Appalachian Trail follows this line of ridges. Streams are numerous and commonly have steep gradients. Dendritic or locally rectangular drainage patterns occur throughout the study area.

Lithologically, the area is dominated by metamorphosed clastic rocks. A sequence of muscovite-rich pelitic schist and metasandstone with interlayered amphibolite is overlain to the west by a thick sequence of garnetiferous metasandstone and biotite gneiss with minor biotite schist and calc-silicate layers or lenses. A quartz-feldspar-biotite-muscovite orthogneiss of variable thickness intruded the metasandstone. Exposures of sillimanite-bearing garnetiferous schist and gneiss with abundant calc-silicate layers cap ridges farther west, and this unit appears to overlie the metasandstone. Toward the north, chlorite-rich garnetiferous schists are intermixed with metasandstone, amphibolite and hornblende-bearing gneiss. Multiple phases of deformation have affected the area, and locally structural trends are quite diverse. Metamorphism ranges from kyanite to sillimanite grade, and retrograde effects are locally significant.

Objectives

Specific objectives of this investigation are to:

(1) Explore the possible sedimentary origins and environments represented by the metaclastic rocks;

(2) Discuss the metamorphic history deduced from mineral assemblages and textures observed in the rocks;

(3) Discuss the nature of the boundary between possibly lower grade (staurolite to kyanite ?) pelitic schist and sillimanite-grade metasandstone and schist to the northwest;

(4) Define the nature of the chlorite-rich rocks to the north, and explain relationships with a higher thrust sheet mapped to the north, bounded by the Chunky Gal and Shope Fork faults (Fig. 1);

(5) Delineate structural relationships on the meso- and macroscopic scales, and describe microstructures present in thin sections;

(6) Present possibilities for correlation of stratigraphy with that described in other portions of the Blue Ridge;

(7) Evaluate previous studies of the area, and attempt to integrate present data into the regional perspective.

Methods

Detailed field mapping at the scale of 1:24000 was begun in January, 1986, and continued through the spring. This work was completed during the spring of 1987. Approximately 1800 structural data stations were collected in the study area.

Seventy thin sections were made from samples of metaclastic, granitic and amphibolitic rocks collected in the field area. These were used for studies of mineral compositions, metamorphic assemblages, and both deformational and metamorphic textures. As the vast majority of rocks in the area are meta-psammites, point counts were made on 45 thin sections of this rock type to aid in determination of sedimentary protolith and definition of stratigraphic relationships. These were also compared with modal analyses of psammitic rocks from studies in adjacent areas. Heavy mineral separates of 5 samples were obtained for use in regional correlations.

Structural analysis was conducted both on the mesoscopic and macroscopic scales, through descriptions of structural elements, as well as through the use of equal area projections. As the orientation of penetrative mesoscopic structures is non-uniform

throughout the study area, the region was divided into more or less homogeneous domains which were analyzed separately.

Previous Investigations

Early studies that included rocks of the study area are a general description of the crystalline rocks of Georgia, by Crickmay (1952), and a discussion of the stratigraphy and structure of the Blue Ridge from Alabama to North Carolina by Hurst (1973). The general stratigraphy and structure of the "Tray Mountain Roadless Area" was described by Nelson (1982), who mapped most of the roads and trails in the Tray Mountain quadrangle. Nelson (1985) included these data in a more extensive study of the geology of the northwest part of the Greenville 2-degree quadrangle of Georgia. Other studies in adjacent and related areas include Hatcher (1971; 1979; 1987), Shellebarger (1980), Gillon (1982), Farrell (1986), Hopson (PhD dissertation, in progress), and Stieve (PhD dissertation, in progress).

II. REGIONAL GEOLOGY

The Appalachian orogen can be divided into three segments: northern, central, and southern, with each displaying unique structural characteristics. The southern portion extends from Virginia to central Alabama, where it is overlapped by Coastal Plain deposits (Plate I). Four major physiographic subdivisions have been recognized: the Cumberland Plateau, Valley and Ridge, Blue Ridge, and Inner Piedmont.

The foreland fold and thrust belt of the orogen is represented by the Cumberland Plateau and Valley and Ridge provinces. Here, platformal, miogeoclinal and foreland basin sediments were deposited along the cratonic margin throughout most of the Paleozoic. Collision of the North American continent (Laurasia) with Africa (Gondwanaland) occurred in the late Paleozoic (Hatcher, 1972), resulting in compression, folding and faulting. Thus, the foreland fold and thrust belt developed by thin-skinned deformation in the terminal stages of the Alleghanian orogeny (Rodgers, 1953). The pattern of sedimentation, however, reflects earlier tectonic events affecting the orogen (Hatcher, 1978). Rocks of the foreland are unmetamorphosed to very slightly metamorphosed on the eastern boundary of the Valley and Ridge province.

East of the foreland lies the Blue Ridge geologic province, part of a discontinuous chain of allochthonous uplifts (Wehr and Glover, 1985). The trace of axes of these anticlinoria forms the Blue-Green-Long axis of Rankin (1976), which extends from Newfoundland to Georgia. The northern portion of the Blue Ridge from Pennsylvania to central Virginia is an anticlinorium in which billion-year-old basement rocks are nonconformably overlain by late Precambrian to early Paleozoic metasedimentary and

metavolcanic rocks (Wehr and Glover, 1985). Southward, the Blue Ridge becomes more structurally complex, with extensive thrust faulting tending to obscure stratigraphic relationships. Metamorphism ranges from essentially none to upper amphibolite facies. As the Blue Ridge is the site of the study area, its major divisions will be discussed in greater detail in a following section.

The Brevard fault zone separates the Blue Ridge from the Piedmont, which consists of a number of subdivisions. Detailed description of the different belts will not be attempted here (see King, 1955 and Hatcher, 1972, for discussion of these subdivisions). Rock types present in the Piedmont are many and varied, and include metamorphosed sandstone, greywacke, carbonate, conglomerate, pelite, and volcanic rocks. Granitic to ultramafic plutons also occur here. Metamorphic grade ranges from sub-greenschist to amphibolite facies, and possibly three metamorphic episodes have been recognized (Hatcher and Odom, 1980; Wehr and Glover, 1985). The eastern half of the Piedmont, including the Charlotte belt and Carolina slate belt (Plate I) appears to be an exotic terrane (Williams and Hatcher, 1982) containing early Paleozoic European-African fauna (Secor and others, 1982). High-grade, migmatitic rocks of the Inner Piedmont resemble those of the Blue Ridge, although the metamorphic grade and multiple deformational events have hindered resolution of stratigraphic relationships.

Blue Ridge: Divisions

Recognition of major differences between the western and eastern portions of the Blue Ridge has led some geologists (Rankin, 1975; Hatcher, 1978) to postulate that a major fault system extends the entire length of the Blue Ridge (or possibly even the entire Appalachian chain). These differences include:

(1) In many areas, an abrupt change from rocks with shallow water (miogeosynclinal) or subaerial affinity to those which originated in deeper water (eugeosynclinal environment);

(2) Paucity of volcanic lithologies in the western Blue Ridge, except for the Mount Rogers and Catoctin Formations;

(3) Lack of Paleozoic granitic plutons in the western Blue Ridge;

(4) Relative abundance of basement exposures in the western Blue Ridge.

Hatcher (1978) suggested that a significant contact in the Blue Ridge of Georgia and North Carolina is a fault, and possibly part of such a major fault system. He named this the Hayesville fault, and suggested that movement on this fault began in the early to middle Paleozoic. Hatcher and others (1979) concluded that the Hayesville was largely pre-metamorphic.

The Hayesville may join other major faults to form the extensive fault system postulated above (Plate I). Hatcher (1978) proposed that the Fries fault, present northeastward from the Grandfather Mountain window, could be such an extension. The Allatoona fault near Cartersville, Georgia (Hurst, 1973) could be the southward continuation of this system (Hatcher, 1978), as it marks an abrupt change in lithology and tectonic environment in a similar fashion to the Hayesville (McConnell and Costello, 1980; McConnell and Abrams, 1986). Wehr and Glover (1985) suggested that a fault system exists from central Virginia southward, consisting of the Hayesville, Fries, and Rockfish Valley faults. More recently, Hatcher and others (1988) have suggested that the Gossan Lead overthrust, mapped by Stose and Stose (1957) may be the northeastward extension, rather than the Fries fault. The placement of the "Hayesville-

Fries" fault system in western North Carolina was disputed by Brewer (1986) and Brewer and Woodward (in press), based upon what they considered to be an erroneous correlation of amphibolite-bearing basement rocks with Late Precambrian (?) eugeosynclinal rocks of the eastern Blue Ridge.

These faults all differ somewhat in style and timing. The Fries fault is a major post-metamorphic thrust (Rankin, 1971) which cuts into basement, and appears to have overridden the Holston Mountain overthrust on the western edge of the Blue Ridge (Stose and Stose, 1957). The Rockfish Valley fault, on the other hand, developed contemporaneously with metamorphism and folding (Bartholomew, 1977) or possibly near the end or soon after a metamorphic episode (Gathright, and others, 1977). Movement along the Allatoona fault probably occurred soon after a peak of regional metamorphism (German, 1986). The Gossan Lead fault appears to be late Paleozoic, and is marked by shear zones and mylonitization (Stose and Stose, 1957). However, Rankin (1971) has concluded that this "fault" may actually be a nonconformity between basement rocks and late Precambrian metasedimentary rocks, on the basis of pebble and cobble conglomerates found with in the metasedimentary sequence near the contact. Undoubtedly, the exact nature and position of a major fault system in the Blue Ridge and beyond is still subject to debate.

Western Blue Ridge: Stratigraphy

Basement Rocks

Near its northern terminus in Maryland through North Carolina the Blue Ridge anticlinorium exposes older Precambrian granitic rocks in its core. Isolated outcrops are

also found farther to the southwest (Plate I). Various names have been assigned to these rocks; for example, the Cranberry Gneiss, and the Wilson Creek and Blowing Rock Gneisses exposed in the Grandfather Mountain window (Rankin, 1970). Quartz monzonite is dominant, although compositions range from diorite to granite. Texturally, the rocks are equigranular to porphyritic, and textural and compositional variety may occur in a single outcrop (Rankin, 1970). Collectively, Rankin included these basement rocks in the Elk Park Plutonic Group. They yield radiometric ages of 1050 to 1250 m.y. (Fullagar and others, 1978), which is about the same age as rocks of the Grenville province in Canada (Tilton and others, 1960).

Ocoee Supergroup

The Ocoee Supergroup (Plate I) is a thick sequence of predominantly clastic sedimentary rocks making up much of the western Blue Ridge from northeastern Tennessee to northwestern Georgia, and possibly into Alabama (Hadley, 1970; McConnell and Costello, 1982; Rast and Kohles, 1986). The rocks are late Precambrian (Hadley and Goldsmith, 1963), and unconformably overlie the basement (King, 1964). Metamorphism of the Ocoee ranges from sub-greenschist to upper amphibolite facies (Hadley, 1970).

The Ocoee has been divided into three broad lithologic units, the Snowbird Group, Great Smoky Group, and Walden Creek Group, with a combined thickness of at least 12,000 meters (Hadley, 1970). Many of the clastic sequences are monotonous, consisting of coarse-grained, poorly sorted rocks that are gradational both vertically and laterally (King, 1964). Minor limestone and dolostone are primarily confined to the Walden Creek Group. The Snowbird Group lies unconformably upon basement, and is

the oldest unit in the succession (King, 1964). The relative position of the Great Smoky and Walden Creek groups is unclear because of extensive faulting which has disrupted stratigraphic continuity. The contact between the two, however, may be conformable in the Ocoee River gorge (Costello and Hatcher, 1986). Detailed descriptions of the divisions within the Ocoee can be found in Hadley and Goldsmith (1963), and Hadley (1970).

Other Late Precambrian Stratified Rocks

The Ocoee basin terminates to the northeast. Late Precambrian units outcropping north of the Ocoee belt are characterized by interbedded clastic and volcanic rocks. These are the Grandfather Mountain Formation, found only within the Grandfather Mountain window of North Carolina, the Mount Rogers Formation of southwestern Virginia and northeastern Tennessee, the Swift Run Formation of Virginia, and the Catoctin Formation which occurs from central Virginia to southern Pennsylvania (Plate I). Volcanic rocks consist of a bimodal suite of basalts and rhyolites interpreted to be rift-related. These are dominant in the Catoctin and lower Mount Rogers formations, whereas clastic rocks are most abundant in the other formations. Common lithologies include feldspathic or lithic sandstone, conglomerate, siltstone, and argillite. Most of these units are interpreted to have been deposited in terrestrial environments (Wehr and Glover, 1985).

Chilhowee Group

The Chilhowee Group is preserved discontinuously along the entire northwest margin of the Blue Ridge, from Alabama to Pennsylvania (Plate I). In some areas, it

conformably (Whisonant, 1974) or disconformably (Wehr and Glover, 1985) overlies late Precambrian strata. Elsewhere the Chilhowee was nonconformably deposited upon Grenville basement rocks (Wehr and Glover, 1985). It has been divided into four formations in Alabama and Georgia (Mack, 1980), six in central-eastern and southern Tennessee (Keith, 1895; Rodgers, 1953), and three in the northern portions of the Blue Ridge (King and Ferguson, 1960). Despite variations in thickness and character of individual formations, an overall pattern is consistent along strike: basal conglomerate and feldspathic sandstone succeeded by a middle interval of siltstone and shale with interbedded sandstone, and an upper unit primarily consisting of quartz arenite. Early Cambrian fossils have been found in the upper and middle parts (Whisonant, 1974; Mack, 1980), although only trace fossils, particularly *Scolithus* burrows, have been recognized in lower formations. Therefore the Chilhowee is assigned a Cambrian or Cambrian (?) age (King, 1964; Neuman and Nelson, 1965; Mack, 1980).

Murphy Belt

Murphy belt rocks conformably overlie the Ocoee Supergroup in North Carolina and Georgia (Plate I). Metamorphosed equivalents of argillaceous siltstone, shale, and fine-grained sandstone comprise most of the belt. The most interesting and economically important unit is the Murphy Marble, which consists of calcite and dolomitic marble of variable purity (Hadley, 1970). Metamorphism reaches middle amphibolite facies in parts of the Murphy belt (McConnell and Costello, 1982). The basal unit of the Murphy belt may be conformable with the upper Great Smoky Group (Hadley, 1970). Correlations remain tentative, as various interpretations have equated parts of the Murphy sequence with the Chilhowee Group, Shady Dolomite, and rocks of the Talladega belt of Alabama and Georgia.

Talladega Slate Belt

The Talladega slate belt of eastern Alabama and western Georgia (Plate I) is a thick sequence of clastic, carbonate, and volcanic rocks that have undergone metamorphism to lower greenschist facies. Regional correlations of the Talladega belt are still uncertain, but Tull and Guthrie (1985) have considered that the history of the Talladega belt spans the interval from late Precambrian to Early or Middle Devonian, and various parts of the sequence may correlate with the Ocoee Supergroup, Chilhowee Group, Murphy belt rocks, and with Devonian rocks of the foreland. Rocks of the Talladega belt have been thrust over Lower Paleozoic rocks of the Valley and Ridge along the Cartersville-Talladega fault system in Georgia and Alabama (Bearce, 1978, 1982).

Western Blue Ridge: Structure

The transition from folded, thrusted and unmetamorphosed rocks of the Valley and Ridge to penetratively deformed and metamorphosed rocks to the east is exemplified by the western Blue Ridge of Tennessee, Georgia, and North Carolina. Metamorphism probably took place during the Taconic orogeny, while later deformation and over-thrusting toward the west occurred during the Alleghanian (Butler, 1973).

Two major thrust faults dominate the Ocoee Supergroup in the Great Smoky Mountains area (Plate I) (King, 1964): the Greenbrier fault which was emplaced prior to the peak of metamorphism and was folded and faulted after emplacement; and the Alleghanian Great Smoky and Miller Cove thrust faults that moved the entire Blue Ridge northwestward. Numerous minor faults of varying magnitude are also present (King,

1964; Neuman and Nelson, 1965). These include thrust faults, as well as late high-angle faults that parallel or cut across older structural features (Hadley and Goldsmith, 1963). Displacement on these high-angle faults has occurred by both dip-slip and strike-slip movements (King, 1964). The earliest folds recognized by Hadley and Goldsmith (1963) in this area are broad eastward-trending folds which predate movement on the Greenbrier fault. Later deformation produced northward-trending folds, as well as the high-angle faults discussed above.

The allochthonous nature of the Blue Ridge is demonstrated by the presence of the Grandfather Mountain window in northwestern North Carolina (Bryant and Reed, 1970), where basement and late Precambrian clastic and rift-related volcanic rocks of low metamorphic grade are exposed within the window, and also by windows within the Great Smoky thrust sheet that expose unmetamorphosed, fossiliferous Lower Paleozoic rocks within metamorphosed rocks of the Ocoee Supergroup (King, 1964; Neuman and Nelson, 1965).

Rankin and others (1973) described the structure of the western Blue Ridge in northwestern North Carolina as being dominated by large-scale folds (Stony Creek syncline, Blue Ridge anticlinorium) and faults (Holston-Iron Mountain fault, Stone Mountain fault system). Northward into Virginia, the Blue Ridge anticlinorium becomes an anticlinal fold within a single tectonic unit (Rankin and others, 1973). Granitic rocks are exposed in the core of the anticlinorium and are unconformably overlain by the Catoctin Formation and the Chilhowee Group, which can be traced around the hinge of the anticlinorium (Espenshade, 1970). At its northern end, the anticlinorium is asymmetric and overturned toward the northwest. Pervasive cleavage is present in all rocks, and gradually dies out westward.

Eastern Blue Ridge: Stratigraphy

In the unfossiliferous, often highly metamorphosed metasedimentary and metavolcanic rocks of the eastern Blue Ridge, age constraints are few and only broadly defined. Stratigraphic correlations are, for the most part, only tentative, but two major groups have been recognized (Wehr and Glover, 1985): (1) an extremely thick, unevenly distributed sequence of immature clastic and volcanic rocks, thought to be Late Proterozoic; and (2) an overlying sequence of predominantly clastic rocks which are probably Late Proterozoic to early Paleozoic. The older rocks may represent a rift environment, while the younger rocks may have been deposited in an environment of increasing stability and subsidence (Wehr and Glover, 1985).

Late Precambrian Stratified Rocks

The Tallulah Falls Formation, found in northeast Georgia into southwestern North Carolina and in northwestern South Carolina (Plate I), consists of metagreywacke and biotite gneiss interlayered with muscovite-biotite schist, aluminous schist, quartzite and amphibolite (Hatcher, 1971; 1973). It can be divided into four members (Hatcher, 1971):

(1) Greywacke-schist-amphibolite member, the basal unit;

(2) Garnet-aluminous schist member consisting of muscovite-garnet schist which commonly contains kyanite or sillimanite, interlayered with metagreywacke;

(3) Greywacke-schist member, metagreywacke with muscovite schist and

occasional amphibolite or quartzite. Graded beds are commonly recognizable;

(4) Quartzite-schist member (Tallulah Falls Quartzite), consisting of thick layers of quartzite alternating with muscovite schist. This member is in contact with basement rocks within the Tallulah Falls dome, a prominent structure in northeast Georgia (see following section on structure). A series of metaconglomerate beds are found near the basement contact.

The mineralogical composition of the metagreywackes is primarily quartz, plagioclase and biotite, with minor microcline and muscovite. Accessory minerals include epidote or clinozoisite, sphene, zircon, garnet and amphibole (Hatcher, 1973). In general, the quartzites have less total feldspar and are more quartz-rich. The accessory mineral suite is similar to the metagreywacke; traces of carbonate and tourmaline are also present. Metamorphic grade ranges from garnet or lower kyanite near the Brevard zone to sillimanite grade to the northwest, and the Tallulah Falls Formation becomes migmatitic at higher grades (Hatcher, 1973).

Probable correlatives of the Tallulah Falls Formation, based upon lithologic similarity and lateral continuity, are the Ashe Formation of northwestern North Carolina and southwestern Virginia, the Lynchburg Formation of northern and central Virginia, and the Sandy Springs Group, which is continuous with the Tallulah Falls Formation southward into Georgia.

Late Precambrian-Early Paleozoic (?) Stratified Rocks

Overlying the Tallulah Falls and Otto (see below) Formations in southern North Carolina and adjacent Georgia is the Coweeta Group, exposed in the trough of a synclinal structure known as the Coweeta syncline (Hatcher, 1979) (Plate I). The principal rock types are feldspar-quartz-biotite gneiss, metasandstone, aluminous schist and garnetiferous biotite schist, and quartzite. Amphibolite is present in minor amounts. Three subdivisions have been described by Hatcher (1979, 1987):

(1) Persimmon Creek Gneiss, the basal unit, is a coarse, even-grained feldspar-quartz-biotite (quartz diorite) orthogneiss, and is the most extensive unit in the Coweeta Group in North Carolina. It appears to have intruded the metasedimentary rocks, since xenoliths of metasandstone have been identified within it (Hopson, pers. comm., 1987). In some areas the gneiss is complexly interlayered with metasandstone and schist.

(2) Coleman River Formation overlies the Persimmon Creek Gneiss, and consists of feldspathic to quartzose metasandstone and quartz-feldspar gneiss interlayered with muscovite-biotite schist. Kyanite or staurolite occur in some schist layers. Calc-silicate lenses are common in some areas. A distinctive feature of metasandstones in the Coleman River Formation is "pinstriping", or thin laminae similar to that described for the Alligator Back Formation (see page 20).

(3) The Ridgepole Mountain Formation is a heterogeneous assortment of lithologies, including pelitic schist, coarse biotite-garnet schist, muscovite-chlorite quartzite, clean quartzite, and garnetiferous metasandstone. The highest unit in the sequence, it is preserved in

synclines throughout the region of outcrop. The contact with the Coleman River Formation appears to be gradational.

In its type area, the Coweeta Group is differentiated from the Tallulah Falls Formation on the basis of "cleaner" sediments and the paucity of amphibolite (Hatcher, 1979).

The Coweeta Group is the dominant unit within the area of this report. To the southwest, rocks equivalent to the Coleman River and Ridgepole Mountain Formations have been included in the Richard Russell Formation (Gillon, 1982; Nelson and Gillon, 1985). Nelson (1985) extended usage of this name to include rocks within the study area. The Richard Russell Formation consists of migmatitic biotite gneiss, metasandstone, aluminous biotite schist and biotite-muscovite schist, with minor calc-silicate granofels and amphibolite.

Within the trough of the Ararat synclinorium, the Alligator Back Formation overlies the Ashe Formation in apparent conformity (Rankin and others, 1973) (Plate I). The sequence consists of laminated metagreywacke and biotite gneiss with interlayered schist and amphibolite, and contains minor quartzite and marble. Abundant mafic and ultra-mafic rocks are present in portions of the Formation (Conley, 1987). Although the rocks were originally included within the Ashe, Rankin and his colleagues chose to distinguish them from the Ashe due to the following characteristics:

(1) More micaceous nature of the laminated metagreywacke;
(2) Presence of tourmaline as a common accessory mineral in the clastic rocks (uncommon in those of the Ashe);
(3) More calcic metagreywacke, with epidote as a common accessory mineral;

(4) Presence of thin quartzo-feldspathic laminae separated by micaceous parting ("pinstriping"). These laminae may represent original thin bedding and/or tectonic transposition of an earlier lamination. Pinstriping is rare in the Ashe Formation.

The Alligator Back Formation is probably Late Proterozoic or early Paleozoic because it appears to stratigraphically overlie the Ashe, although it must be noted that it is not in contact with any stratigraphically younger unit (Rankin and others, 1973). Rb-Sr radiometric ages obtained by Fullagar and Odom (1976) provided a minimum age of 350 to 371 m.y. for samples from the Alligator Back Formation.

Sequences of late Precambrian-Early Paleozoic (?) stratified rocks from Georgia to Virginia occupy similar stratigraphic positions along strike, although they are lithologically somewhat heterogeneous. The Evington Group can be traced into the Alligator Back Formation and is considered to be equivalent (Wehr and Glover, 1985). Hatcher (1979) recognized the similarity of the "pinstriped" texture in metasandstones of the Coleman River Formation (Coweeta Group) to that described by Rankin and others (1973) in the Alligator Back Formation, and later suggested that these units may be time equivalents (Hatcher, 1985). The Richard Russell Formation in Georgia is continuous with and equivalent to the Coweeta Group within the study area. Thick-bedded metasandstone and micaceous schist occurring southwest of Spruce Pine, North Carolina and mapped as Great Smoky Group (Hadley, 1970; Hadley and Nelson, 1971) may prove to be correlative with one or more of these "younger" units. This sequence is similar to the Coweeta Group in lithology and lack of abundant amphibolite.

Other Eastern Blue Ridge Rocks, Georgia and North Carolina

Directly underlying and possibly in fault contact with rocks of the Coweeta Group is the Otto Formation, a sequence of relatively quartz-rich two-mica feldspathic sandstones interlayered with aluminous schists that may contain kyanite or staurolite (Hatcher, 1987). These same lithologies extend as a narrow band to the southwest into Georgia, where they become complexly associated with amphibolite (Plate I). The Otto Formation can be traced into similar lithologies of the Helen Group, that consists of interlayered amphibolite and aluminous schist, micaceous metasandstone and feldspathic meta-siltstone, with minor biotite schist and pebbly quartz metasandstone. Age of the Helen Group may be late Proterozoic to early Paleozoic (Nelson and Gillon, 1985). Nelson and Gillon (1985) tentatively correlated a portion of the Helen Group with the New Georgia Group, a linear band of rocks mapped from near the Alabama-Georgia border northeast-ward in the Blue Ridge of Georgia. The New Georgia Group consists of a sequence of metavolcanic and metaplutonic rocks with interlayered metamorphosed volcaniclastic and sedimentary rocks, and amphibolite is ubiquitous.

Basement Rocks, Mafic Rocks, and Plutons

Two significant characteristics of the eastern Blue Ridge are; (1) the scarcity of unequivocal basement rock exposures; and (2) the presence of mafic and ultramafic rocks as well as Paleozoic (granitic) plutons, which are rare or absent in the western Blue Ridge (Hatcher, 1978).

Granitic gneisses thought to be basement orthogneisses occur in the core of the Toxaway dome of North and South Carolina (Hatcher, 1973, 1977, 1978; Fullagar and

others, 1978), and in the core and flanks of the Tallulah Falls dome in Georgia (Hatcher, 1973, 1978) (Plate I). Rb-Sr whole rock analysis produced a 1203±54 m.y. age for the Toxaway Gneiss (Fullagar, and others, 1979). Recent work by Stieve and others (1988) yielded Rb-Sr dates of 1179±15 and 1114±15 m.y. on two separate bodies of orthogneiss on the flanks of the Tallulah Falls dome.

Besides the abundant amphibolites which are interlayered with metasedimentary sequences in the eastern Blue Ridge, several discrete mafic/ultramafic bodies are also present. Examples are: the Lake Chatuge and Shooting Creek mafic/ultramafic complexes in northeastern Georgia and southwestern North Carolina, which frame the Brasstown Bald and Shooting Creek windows; the Laurel Creek complex in Georgia; and the Carroll Knob complex in North Carolina. These may all have been tectonically emplaced, and subsequently deformed and metamorphosed. Recent studies (Petty, 1982; Hatcher and others, 1984; McSween and Hatcher, 1985) point to a non-continental origin for the Laurel Creek and Carroll Knob complexes, and suggest that Laurel Creek may be an ophiolite.

Pegmatite bodies yielding Paleozoic ages have been found in the eastern Blue Ridge near Franklin, North Carolina (Butler, 1972) and have intruded rocks of the Ashe Formation in the Spruce Pine area of western North Carolina (Butler, 1972, 1973). Butler (1973) suggested that the Spruce Pine pegmatites were emplaced during an episode of regional metamorphism which occurred around 380 Ma.

Eastern Blue Ridge: Structure

Folding

Most workers in the eastern Blue Ridge have recognized at least three phases of folding (Rankin and others, 1973; Butler, 1973; Hatcher and Butler, 1979; Hatcher, 1987); in some areas as many as six phases can be deduced (Hatcher, 1976). A typical pattern, which may be representative throughout the eastern Blue Ridge was recognized by Hatcher (1977) in and around the Toxaway dome of South and North Carolina. Early folds are pre- to syn-metamorphic flowage folds, the first set isoclinal and recumbent, the second set more upright, isoclinal to open. Later folds are represented by crenulations, and at least two sets of upright flexural folds.

The dominant foliation in some areas has been produced by the D_2 episode (Hatcher, 1977; Nelson 1985), and this may be the typical case. In the southern Blue Ridge, major folds predominantly trend northeast and verge northwest. Superposed folding probably produced the outcrop patterns of many large structural features (Hatcher and Butler, 1979; Nelson, 1985).

Synclinoria

An interesting feature of the eastern Blue Ridge is a linear pattern along much of its extent of synformal structures, most of which expose younger rocks. From north to south, these are: the James River synclinorium, which contains the Evington Group; the Ararat River synclinorium, where the Alligator Back Formation overlies the Ashe

Formation; the Spruce Pine synclinorium; and the Coweeta syncline, which preserves rocks of the Coweeta Group (Plate I). No unequivocally younger rocks have been separated in the Spruce Pine synclinorium (Butler, 1973), although Hadley and Nelson (1971) recognized two general metasedimentary units, the upper of which they correlated with the Great Smoky Group of the Ocoee Supergroup.

An intriguing question is whether these structures are temporally and/or structurally related to one another. Hatcher (1979) described the Coweeta syncline as being a polyphase structure, overturned toward the east, which was formed by interference of F_1 and F_2 folds. Later flexural folding had little effect on the structure. On the other hand, the synclinal nature of the Spruce Pine area is considered a result of late-stage deformation (Rankin and others, 1973; Butler, 1973). The Ararat River synclinorium is inferred from the distribution of stratigraphic units, and appears to be overturned toward the northwest (Rankin and others, 1973). It contains major thrust faults internally, and is dextrally offset from the James River synclinorium (Rankin, 1975). Therefore, it must be older than the latest stage of deformation.

Domes and Windows

A major concept regarding the structural history of the Blue Ridge is that the entire belt is allochthonous, and has been thrust over sedimentary rocks of the platform along a master décollement (Bryant and Reed, 1970; Hatcher, 1972). Recent seismic reflection data have aided in resolution of much of the deeper structure of the southern Appalachians, and have supported this conclusion. It is interesting to note here that two prominent structures in the eastern Blue Ridge, the Shooting Creek window in North Carolina and Georgia and the Tallulah Falls dome in Georgia (Plate I), have been

suggested to be the result of large antiformal stack duplexes developed within the platform rocks (Hatcher and others, 1987). It appears that the basal crystalline thrust sheet was arched above the stacks to produce these late stage structures (Hatcher, 1987). The Brasstown Bald window in Georgia may have resulted from the same process (Hatcher and others, 1986).

Faulting

Faulting is a significant structural element in the eastern Blue Ridge. Movement histories of the major faults are varied, and in some cases complex. Evidence for post-metamorphic faulting may include offset of stratigraphic markers, offset of metamorphic isograds, and mylonitization or cataclasis along the fault zone. Pre-metamorphic faults are much more difficult to recognize, as subsequent metamorphism and deformation may obscure physical evidence of faulting, and isograds are not truncated. As more stratigraphy is broken out in the eastern Blue Ridge, more pre-metamorphic faults are deduced from truncations of these stratigraphic sequences.

Besides the Hayesville fault, discussed on page 9, three other faults that have been mapped in or around the study area should be mentioned. The Shope Fork fault is an extensively folded pre- to syn-metamorphic fault which cuts the west limb of the Coweeta syncline (Hatcher, 1979) (Plate I). Nelson (1982, 1985) and Gillon (1982) extended this fault into Georgia as the contact between their Helen Group and Richard Russell Formation. This may be an erroneous correlation, although the contact may be a separate fault. The Chunky Gal Mountain fault in North Carolina (Plate I) separates amphibolites from a metasedimentary unit tentatively correlated with the Coweeta Group (Hatcher, 1985). It is interpreted as a synmetamorphic thrust. The Chattahoochee fault

in Georgia (Plate I) separates migmatites of the Sandy Springs Group (and equivalent Tallulah Falls Formation) from non-migmatitic rocks of the New Georgia Group (and equivalent Otto Formation). It has been mapped by German (1986) from Habersham County, Georgia southwestward, extending to Douglas County where it is overridden by the Blairs Bridge fault (McConnell and Abrams, 1984). The Chattahoochee fault may be traced northeastward to near the Georgia-North Carolina border.

III. STRATIGRAPHY AND LITHOLOGIC UNITS

Terminology

Metasedimentary rocks dominate all major rock units of the study area, and an attempt has been made to identify protoliths wherever possible. Therefore, the following descriptions must necessarily incorporate both metamorphic and sedimentary terminology. Below is a list of terms commonly used in the study of sandstones, and which generally follow the classifications of Pettijohn (1944; 1949), or terminology in Pettijohn and others (1972). Sedimentary rock names will frequently be prefixed by "meta-" to indicate the metamorphic equivalent.

Arkose: a sandstone containing 25% or more feldspar;

Feldspathic Sandstone: sandstone with 10 to 25 percent feldspar but with appreciable matrix;

Greywacke: a type of sandstone marked by detrital quartz and feldspar in a prominent clay matrix. Originally a field term to denote sandstones with a grey or dirty appearance due to presence of abundant phyllosilicate minerals (Blatt and others, 1973);

Sandstone: any detrital rock composed predominantly of quartz grains 1/16 to 2 mm. In this study, will be used as a general term;

Subgraywacke: a sandstone with over 20% matrix and less than 10% feldspar;

Quartzose Sandstone: a sandstone of 90% or more detrital quartz.

In addition, the following terms will be used as adjectives to denote broad compositional classes of rocks (Mason, 1978): *pelitic*, for rocks having the chemical

composition of shales or slates, and *psammitic,* for rocks having the composition of sandstone. The following metamorphic terms will also be employed:

Schist: any metamorphic pelitic rock with aligned platy minerals;

Biotite Gneiss: a rock in which the minerals are segregated into distinct bands of quartz and feldspar and bands of mica (predominantly biotite);

Calc-silicate Quartzite: a fine-grained metamorphic rock containing conspicuous calcium or magnesium silicate minerals.

Plagioclase compositions were determined from thin sections through the use of the Michel-Levy method, in which the maximum extinction angles of albite twins were determined in sections normal to (010).

Description of Lithologic Units

The study area may be divided into three major lithotectonic units, separated by fault boundaries, each bearing a distinctive assemblage of rocks (Plate II):

(1) Tectonic unit 1 (southeast of the Soque River fault): Otto Formation and associated amphibolites; with minor ultramafic rocks;

(2) Tectonic unit 2 (northwest of the Soque River fault): rocks of the Coweeta Group and a metasandstone-schist unit of uncertain affinity, possibly correlative with the Tallulah Falls Formation;

(3) Tectonic unit 3 (north of the Shope Fork fault): garnet-chlorite schist, amphibolite and metasandstone.

Tectonic Unit 1:

Otto Formation: Bodies of silvery, pelitic muscovite-biotite schist and quartz-rich metasandstone with intercalated amphibolite occur in the southeastern corner of the study area, and are correlated with the Otto Formation of Hatcher (1987). The unit is continuous with that mapped as Otto Formation to the southwest by Farrell (1986) and to the northeast by Hopson (in progress).

Rocks of the Otto Formation generally occupy valleys and low, gentle slopes, and many exposures consist of saprolite. The best exposures in this area are found along streams near the base of Chimney Mountain, where medium-layered sandstone alternates with thick to thinly-layered schist. Sandstone layers commonly contain thin micaceous laminae consisting of biotite and/or muscovite. Where fresh, the sandstone is light to medium grey, compact and fine-grained, with a silvery sheen on foliation planes due to the presence of muscovite. Quartz veins up to 0.5 cm wide commonly parallel the compositional layering of sandstone, schist and micaceous laminae, and are affected by mesoscopic folding which also folds the layering. Schist layers may be dominated by either muscovite or biotite, although muscovite is especially prevalent in thicker layers, and forms the distinctive crenulated silvery schist characteristic of the Otto Formation (Fig. 2). Garnets up to 0.5 cm in diameter are found in some horizons. In recent road cuts, the schist commonly weathers to a deep red and purplish-black streaked saprolite.

In thin section, the dominant minerals in sandstone layers are, in order of abundance, quartz, plagioclase (An_{25-30}), green-brown biotite, muscovite and microcline. A small amount of garnet is present in two sections, and one sample

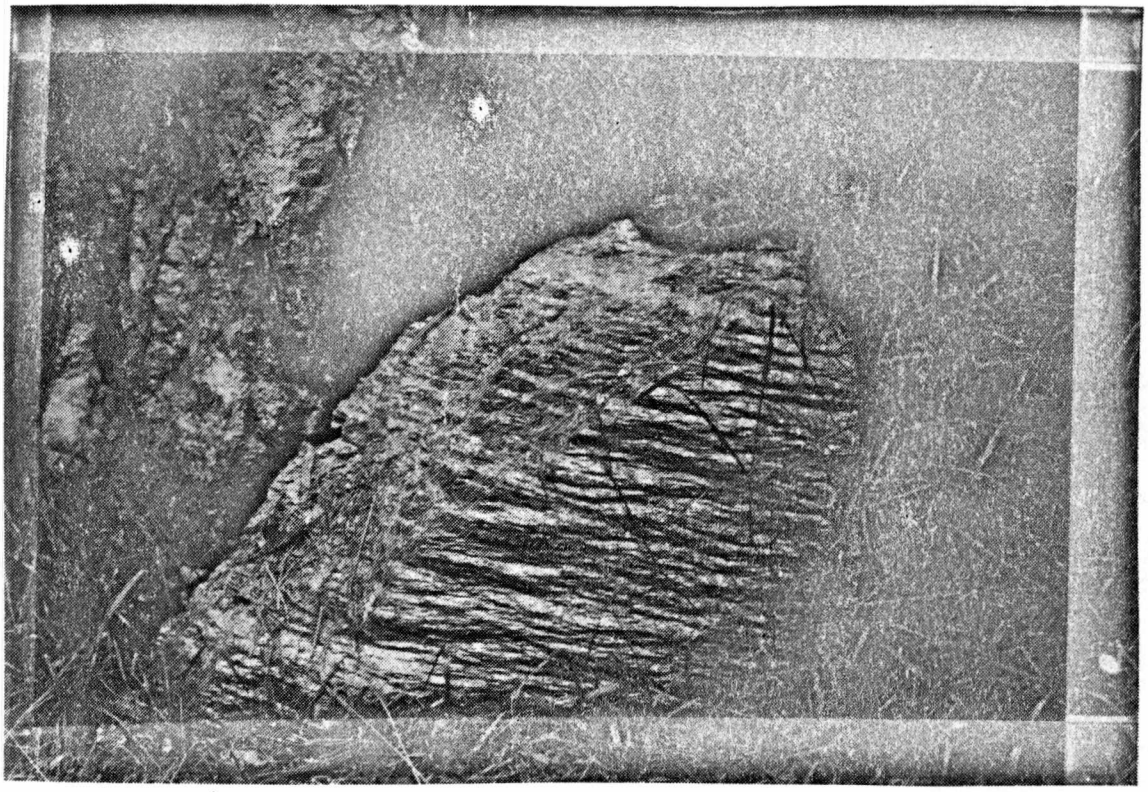


Figure 2: Crenulated muscovite-biotite schist of the Otto Formation. Outcrop located on Wildcat Creek Road, approximately 1 1/2 mi east of study area.

contains a few tiny fibers of sillimanite. Accessory minerals include apatite, zircon, edipote and calcite. Allanite was identified in one thin section.

No thin sections are available of the schist. The dominant minerals are muscovite, biotite, and quartz, sometimes with abundant garnet present. Where the Otto Formation was originally described, in the vicinity of Otto, North Carolina (Hatcher, 1980), kyanite and staurolite are abundant in the schistose rocks (Hatcher, 1987). Farrell (1986) also recognized these minerals in schists along strike directly to the southwest of the study area.

Although the Otto Formation as a whole does not display a "migmatitic" character, in some exposures within the study area thin to thick (up to 25 cm) continuous layers of medium to coarse-grained muscovite-quartz-microcline granitoid parallel the compositional layering. These layers can be massive, or a weak foliation may be developed, especially in the finer-grained examples. Coarse muscovite clasts are common.

Amphibolite: Intercalated with the schist and metasandstone of the Otto Formation are discrete horizons of laminated to massive amphibolite. The massive rock is fine-grained, dark greyish-green to greenish-black, and dense in appearance. Laminated amphibolite is by far the most prevalent, and consists of fine laminae (approximately 0.1 cm) of alternating green-black amphibole and light grey to white feldspar. In some cases, these laminae are crenulated.

The amphibolite is recognized as discrete bodies in outcrop, as yellow-orange residue in saprolite, and in abundant rectangular float blocks. Also present, but found

only as rounded, lumpy float blocks is a coarse-grained black and white amphibole-plagioclase rock (metadiorite?). Although distinctive, this rock is relatively uncommon, and due to lack of exposure, its relationship to other rocks of the Otto Formation is unknown.

In the type area of the Otto Formation, amphibolite is not a prominent rock type of the unit. It is widespread, however, to the northeast and southwest of the study area, intermixed with schist and metasandstone (Hopson, in progress; Farrell, 1986), and is also ubiquitous in the equivalent Helen and New Georgia Groups.

Ultramafic Rocks: Only one possible occurrence of ultramafic rocks was encountered during the course of this study. A discrete layer of crenulated grey-green actinolite schist, approximately 2 m thick, is sandwiched between layers of muscovite schist of the Otto Formation near the eastern boundary of the study area. A smaller exposure of the same rock is found approximately along strike in near proximity to the first. This exposure appears to be associated with massive, non-foliated greenish-black amphibolite, although outcrop is discontinous and the contact is not exposed. The actinolitic schist is interpreted to be the product of alteration of an ultramafic body (see Section VII).

A thin section reveals that the rock is composed almost entirely of fine-grained actinolite, with pale green to colorless pleochroism (Fig. 3). A strongly crenulated foliation is defined by the typically elongate grains. Larger grains of variable orientation also occur, and are parallel or oblique to the foliation. Diamond-shaped cross sectionsare common. Opaque minerals and rare grains of plagioclase are also present.

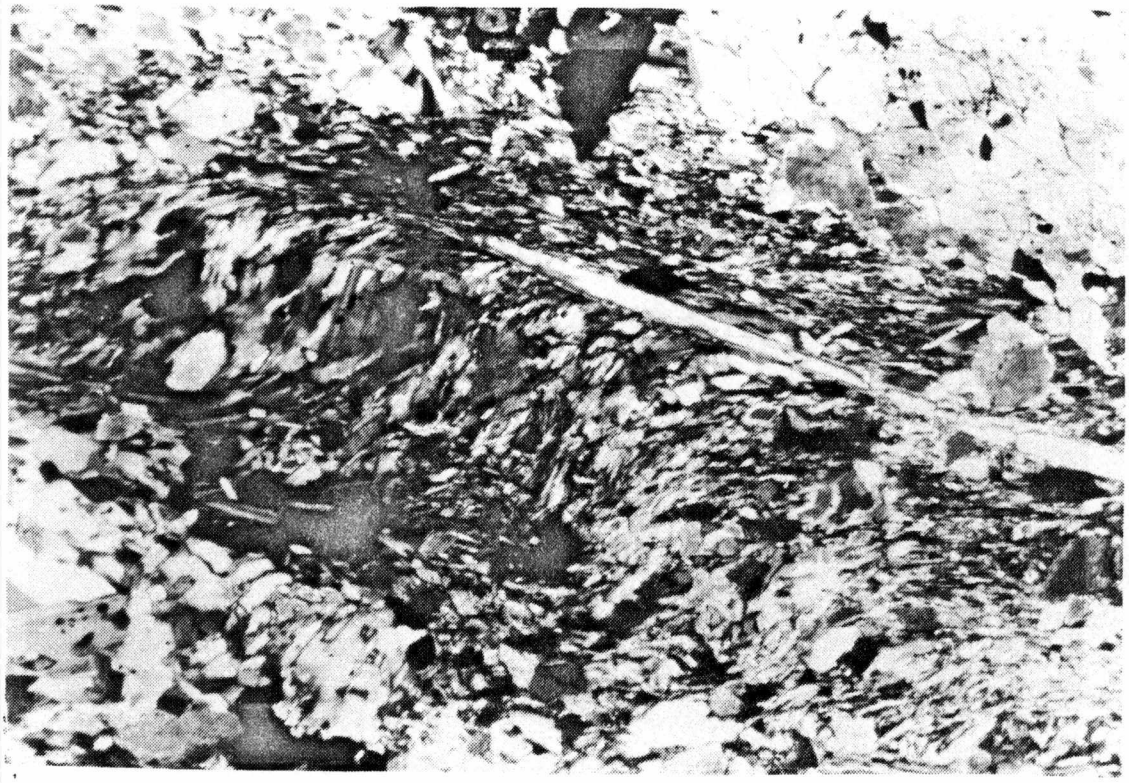


Figure 3: Photomicrograph of crenulated actinolite schist. Interlayered with Otto Formation metasandstone and schist; sample from SE portion of study area, near Wolfpen Branch.

Tectonic Unit 2:

Coweeta Group: The Coweeta Group, described in an earlier section, was subdivided into the Persimmon Creek Gneiss, Coleman River Formation, and Ridgepole Mountain Formation (Hatcher, 1979). A comparable sequence of rocks is recognized northwest of the Soque River fault in the area of this report, and is correlated with the Coweeta Group based on the similar lithologic sequence, as well as on analogous structural and stratigraphic positions.

Persimmon Creek Gneiss: Immediately northwest of the Soque River fault are exposures of a massive to foliated orthogneiss, correlated with the Persimmon Creek Gneiss of Hatcher (1979, 1987). In the study area, the gneiss is fine to medium-grained, and appears to consist of feldspar (primarily plagioclase) and quartz with varying amounts of muscovite and biotite, although biotite is generally dominant. Where developed, the foliation is defined by thin micaceous laminae. No thin section of this rock is available.

In places the gneiss appears alone, but elsewhere it is complexly interlayered with muscovite-biotite schist or feldspathic metasandstone that commonly contains a pinstriped texture. The percentage of metasandstone increases toward the contact with the Coleman River Formation, resulting in a "transition zone" of mixed Persimmon Creek Gneiss and Coleman River metasandstone.

The Persimmon Creek Gneiss is an orthogneiss, derived from an intrusive igneous body, on the basis of the high feldspar content and massive, igneous appearance of the rock, and on reports of xenoliths of metasandstone ocurring in the same body to the

northeast of the study area (Hopson, in progress). The thickness of the gneiss in the study area is estimated to be 500-700 m, less than its thickness in the type area (Hatcher, 1979). This difference may be due partially to the internal deformation affecting the rocks, but also to the irregular thickness and lateral thinning of the intrusive body, which pinches out completely immediately to the southwest (Farrell, 1986). To the northwest within the study area, the Coleman River Formation rests directly upon metasandstones and schists of the Tallulah Falls Formation, and the Persimmon Creek Gneiss is entirely absent.

Coleman River Formation: By far the most widely exposed unit in the study area is a thick sequence of metasandstone and biotite gneiss which appears correlative with the Coleman River Formation (Hatcher, 1979). Fine to medium-grained, light to medium grey metasandstone of variable composition is the most abundant rock type (Fig. 4). Layers may be thick (up to 1m), but characteristically contain thin (less than 0.5 cm) biotite-rich laminae. Quartzofeldspathic, coarse-grained to conglomeratic (?) horizons occur sporadically, but are uncommon. Small lilac garnets (<0.25 cm) are locally abundant in the metasandstones, but are absent over much of its outcrop. Biotite gneiss, also locally garnetiferous, is interlayered and gradational with the metasandstone. It consists of light-colored layers of quartz and feldspar alternating with layers of dark grey, extremely biotite-rich metasandstone or black micaceous schist. The thickness of these layers is variable.

Discrete layers of micaceous schist (generally <20 cm thick) are common, but subordinate to metasandstone and biotite gneiss. Calc-silicate quartzite is present as minor pods, lenses, or thin layers in outcrop. Discontinuous pegmatite veins and pods are ubiquitous, although much more extensive toward the southwest, in the vicinity of

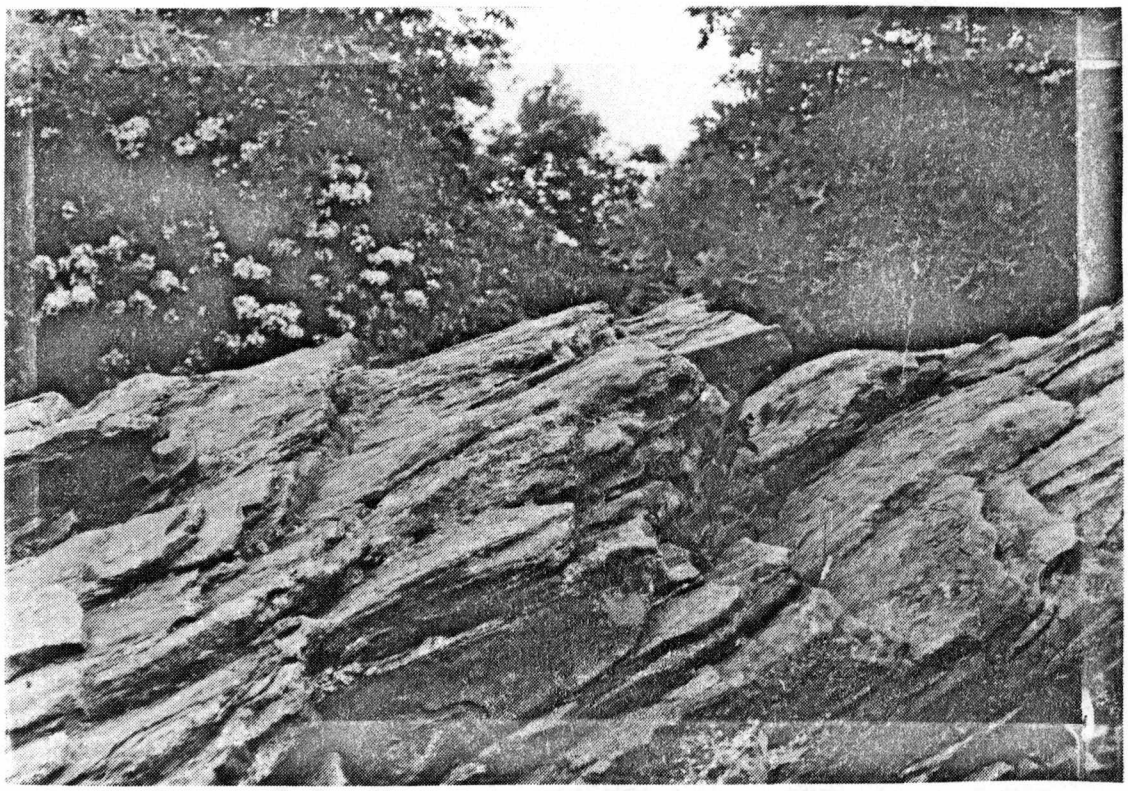


Figure 4: Outcrop of metasandstone of the Coleman River Formation. Located at the peak of Tray Mountain.

Andrew's Creek. Amphibolite is rare in the Coleman River Formation, and is usually found only as rectangular float blocks. Only one possible continuous horizon of laminated amphibolite was encountered, associated with the pegmatitic granites near Andrew's Creek. The nature of this horizon could not be confirmed, however, due to discontinuity of outcrop. In one exposure near the Coleman River-Persimmon Creek Gneiss contact, a lens, approximately 0.5 m thick, of laminated, garnetiferous amphibolite is interlayered and folded with metasandstone.

Thin sections of Coleman River metasandstones reveal that the dominant minerals are quartz, plagioclase (An_{25-32}), and brown biotite. Muscovite is present in most sections, but typically makes up less than 10% of the total mica. A small amount of garnet is present in most sections. Microcline was identified in only two specimens. The accessory mineral suite consists of zircon, apatite, allanite, and epidote or clinozoisite, with a trace of carbonate observed in two sections. Opaque minerals (magnetite- ilmenite?) are ubiquitous. A small amount of sillimanite is present.

Two thin sections were prepared of schistose rocks from the Coleman River Formation. In both samples muscovite forms fine-grained "sericitic" clots between biotite laminae, possibly resulting from later replacement or weathering. One section contains abundant tiny euhedral garnets. Garnet is also present in the other sample, but the grains are much larger and contain abundant inclusions.

Hatcher (1979) described a characteristic texture of the Coleman River Formation as "pinstriping", formed by thin laminations that may be parallel or oblique to the dominant foliation, similar to that described in the Alligator Back Formation of northwestern North Carolina by Rankin and others (1973). Pinstriping is present, although not ubiquitous,

in the Coleman River metasandstones of the study area, formed by thin biotite-rich laminae. Where present, it is commonly oblique to the dominant foliation. Possible origins of this texture will be discussed in a later section. The Coleman River Formation is estimated to be approximately 1830 meters thick in the study area, but thickness estimates are likely to be imprecise because of the intense penetrative deformation which has affected the rocks.

Ridgepole Mountain Formation: Overlying the Coleman River Formation in gentle synforms is a body of interlayered metasandstone and garnet-sillimanite schist (± kyanite), sillimanite gneiss, and muscovite-biotite schist. The metasandstone is light grey, and may be massive or thinly interlayered with micaceous schist. Muscovite is more common than in the Coleman River Formation and many of the sandstones, especially those associated with micaceous schist, contain abundant garnets. The sillimanite schist occurs in layers up to 13 m thick. Reddish garnets up to 0.5 cm are abundant, and sillimanite occurs as white patches of fibrolite on foliation surfaces. In some outcrops it appears that the sillimanite and micas are parallel to axial planes of tight to isoclinal folds.

The sillimanite gneiss consists of layers of quartz-rich metasandstone alternating with thick laminae of sillimanite schist. Kyanite was tentatively identified in some hand samples which exhibit a bluish coloration on foliation surfaces. The presence of kyanite was confirmed in at least one thin section of this rock type.

Calc-silicate quartzite is abundant in outcrop and as float. A rare but distinctive rock type, found only as float, is an "aluminous quartzite", consisting almost entirely of quartz, muscovite, and radiating laths of sillimanite and kyanite.

This sequence is correlative with the Ridgepole Mountain Formation (Hatcher, 1979), on the basis of (a) its stratigraphic position in relation to the Coleman River Formation, and (b) the presence of a variety of lithologies, including pelitic schist, garnet-rich schist, metasandstone, and muscovite quartzite. Sillimanite was not reported in the Ridgepole Mountain Formation by Hatcher (1979). However, this may be due to metamorphic grade rather than compositional differences.

Thin sections of the Ridgepole Mountain lithologies in this area show that the metasandstones are, as a whole, more micaceous than those of the Coleman River Formation, and generally contain slightly less quartz. Plagioclase (An_{28-32}) is abundant, but no potassium feldspar was identified. Small amounts of garnet are present in all but one section. The sillimanite gneiss is similar, but contains abundant sillimanite and a greater amount of garnet.

Garnet-sillimanite-mica schists are quartz-poor, and the foliation is defined by elongate grains of reddish-brown to tan biotite. Muscovite is abundant in all sections, occasionally as well-formed grains but more commonly as fine-grained "sericitic" masses. Some parallels the foliation defined by biotite, but other muscovite (especially well-formed grains) cross-cuts this foliation at an oblique to perpendicular angle. In one section, muscovite may have formed pseudomorphs after sillimanite. Kyanite was confirmed in some sections, and sillimanite occurs in all as fibrolitic mases which are subparallel to the foliation.

Zircon is ubiquitous as an accessory mineral in rocks of the Ridgepole Mountain Formation, generally as tiny rounded grains enclosed in biotite. Epidote is locally

abundant, and opaque minerals (magnetite/ilmenite?) occur in all sections. Other accessory minerals are scarce; apatite and carbonate were identified in only one sample.

A thin-section of calc-silicate quartzite from the Ridgepole Mountain Formation shows that it consists of quartz, plagioclase and garnet, with abundant intergrowths of epidote/clinozoisite. Grains of carbonate and sphene are also present.

Tallulah Falls Formation: To the north in the study area, the Persimmon Creek Gneiss is absent and metasandstones of the Coleman River Formation are underlain by a sequence of similar but distinct biotite gneiss, metasandstone, and biotite schist. The contact appears conformable but fairly sharp, marked by an increase in biotite in psammitic rocks and as discrete layers of black biotite schist. Garnetiferous biotite gneiss is also more abundant in this unit, as is muscovite-biotite schist. A distinctive rock type that occurs sporadically but is unique to this sequence, is an odd "biotite quartzite" consisting almost entirely of quartz and biotite (Fig. 5). It occurs interlayered with other lithologies, particularly with biotite gneiss. Amphibolite and metagabbro are present but rare.

Compared to the Coleman River Formation, metasandstones of the Tallulah Falls Formation have a darker, "dirtier" appearance due to the presence of a greater amount of biotite (see below), and may be referred to more accurately as metagreywackes. Muscovite appears primarily as large clasts on foliation surfaces. Thick biotite schist layers (up to 0.5 m) commonly appear to consist of almost nothing but biotite.

Hatcher (1987) noted that the Persimmon Creek Gneiss may pinch out west of the Coweeta syncline, and there the Coleman River Formation is underlain by a lower

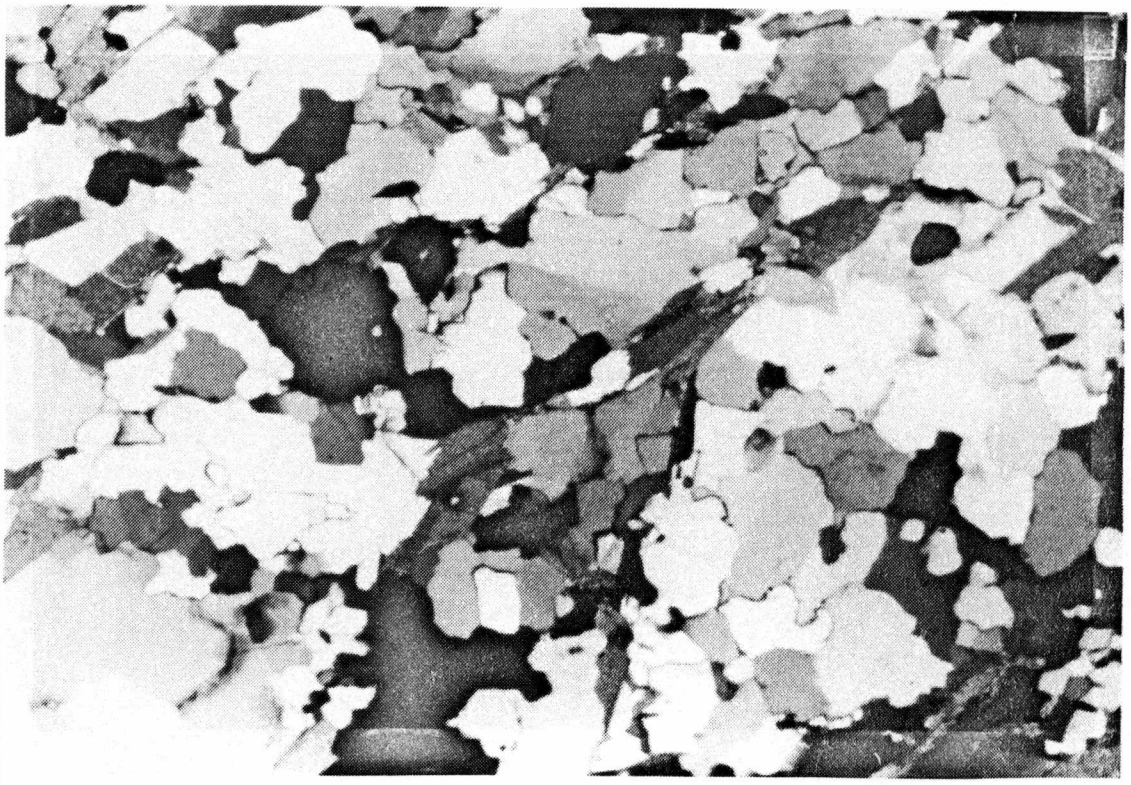


Figure 5: Photomicrograph of "biotite quartzite" of Tallulah Falls Formation. Sample from Appalachian Trail near Addis Gap (Tray Mountain quadrangle). Scale bar is 1 mm.

metasandstone unit tentatively correlated with the Tallulah Falls Formation. Amphibolite is much less abundant in this region than in the Tallulah Falls Formation to the east (Hatcher, unpublished data). This same relationship is present in the study area, and the biotite-rich metasandstone-schist unit described above may correlate with the upper greywacke-schist member of Tallulah Falls Formation (Hatcher, 1971). A thickness estimate for this unit is impossible, because the base is cut out by faulting.

Thin sections of the metagreywacke/biotite gneiss lithologies of this unit indicate that they consist primarily of quartz, plagioclase (An_{24-32}) and biotite, with subordinate muscovite. Potassium feldspar is absent. The quartz/plagioclase ratio is variable, but the biotite content is consistently higher by 10-12% than in the Coleman River metasandstones. Muscovite is present in all samples, but makes up <10% of the total mica.

Small amounts of garnet occur in every sample. Zircon and apatite are ubiquitous in the heavy mineral suite, and opaque minerals (magnetite/ilmenite) are present in most secitons. Allanite was tentatively identified in some, as well as epidote/clinozoisite. Radiating sillimanite fibers are present at plagioclase grain boundaries in two samples.

Muscovite-biotite schists examined in thin section are quartz-poor, with moderate amounts of plagioclase and a muscovite-biotite ratio of approximately 1 to 3. Muscovite does exceed biotite in one section. Garnet is abundant, except in one section in which traces of sillimanite were identified. Zircon and apatite are ubiquitous, as in the metagreywackes. A thin section of the crenulated biotite schist confirms that the dominant mineral is biotite, with minor amounts of quartz, zircon, apatite, and opaque minerals.

Tectonic Unit 3:

A complex assemblage of rocks occurs in the Shope Fork thrust sheet which overrides the Tallulah Falls Formation. This includes garnet-chlorite schist, muscovite-biotite metagreywacke, amphibolite, metadiorite (?), and hornblende gneiss. The garnet-chlorite schist is a distinctive grey-green color, and contains abundant red garnets up to 0.75 cm in diameter (Fig. 6). The schist is interlayered with metagreywacke and minor muscovite-biotite schist. Laminated amphibolite occurs as discrete horizons as well as interlayered with the lithologies above. Hornblende gneiss is light grey, and spotted with grains of greenish-black amphibole throughout. It is also interlayered with the rocks above, although not as abundant as amphibolite. Weathered, coarse-grained metadiorite was found in only one location. The rocks have been metamorphosed to kyanite grade (see below) but show the effects of retrograde metamorphism as well. The extent of outcrop of this sequence in the study area is limited to the area between Berrong Mountain (locally known as Corn Ridge) and Highway 76. The same lithologies are present to the northeast in the Hightower Bald quadrangle (Hopson, in progress) and are here included within the Shope Fork thrust sheet as a continuation of mapping in that area. The abrupt appearance of abundant amphibolite and other mafic/intermediate lithologies, and as equally sudden appearance of the effects of extensive retrograde metamorphism suggest that a fault contact may be feasible. At this time it is not known if this sequence correlates with any previously described unit in the eastern Blue Ridge. Therefore for purposes of this report, the sequence will be informally referred to as the Corn Ridge formation.

In thin section, schists from this unit are composed mostly of muscovite, biotite, and garnet, with minor amounts of quartz and plagioclase (An_{24-30}). The dominant

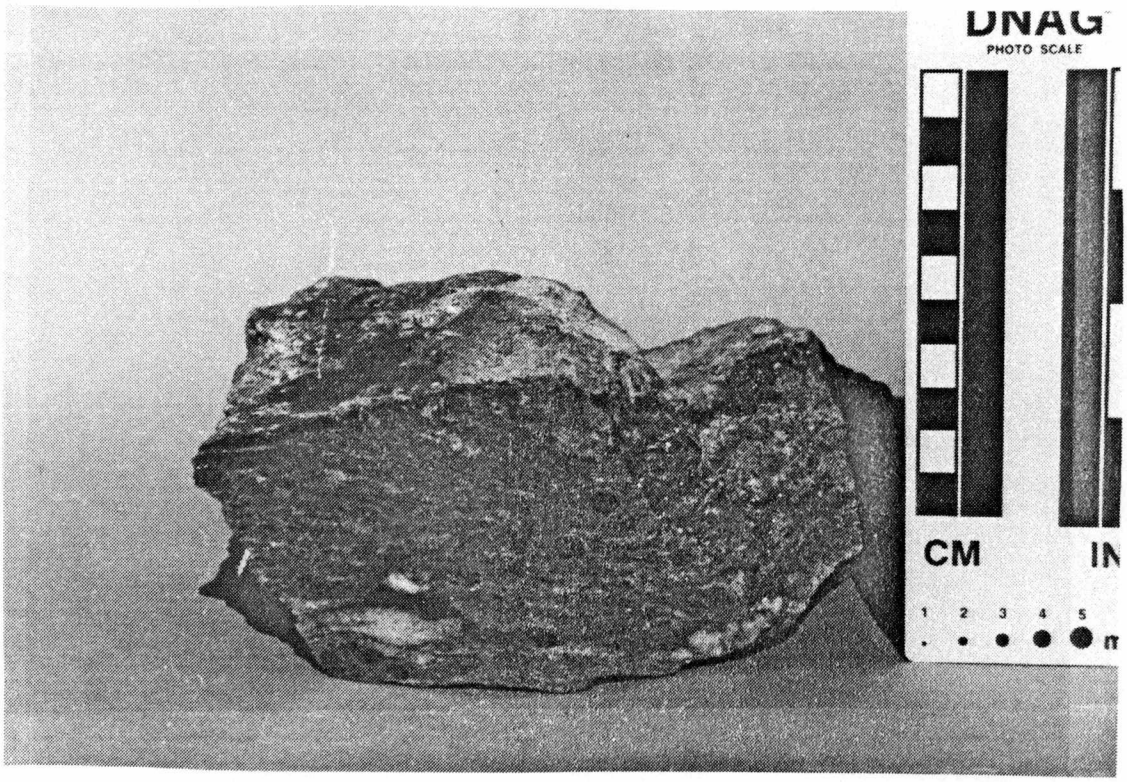


Figure 6: Hand sample of garnet-chlorite schist of Corn Ridge formation. Sample from Corn Ridge (Berrong Mountain), Macedonia quadrangle.

foliation is defined by biotite, and muscovite may parallel this foliation, although many grains are at a high angle to it. Kyanite was identifed in all thin sections, and generally parallels the dominant foliation. Columnar to anhedral grains of epidote are common. In most samples, large radiating masses of chlorite replace biotite and are present within fractures in garnet, indicating that it grew as a retrograde mineral. Its orientation is either random, or highly oblique to the dominant foliation (similar to the muscovite discussed above).

Metagreywackes contain quartz and plagioclase in varying ratios, and biotite is abundant. The muscovite content is generally higher than in the Coleman River or Tallulah Falls psammitic rocks. Garnet is abundant in some samples, although not present in others. Only a minor amount of chlorite was identified in the metagreywackes. Zircon and apatite characterize the heavy mineral suite in both schists and metagrey-wackes. Epidote/clinozoisite and allanite are common, and opaque minerals are present in all samples.

One thin section of the fine-grained hornblende gneiss reveals that it is composed of quartz and plagioclase in subequal amounts, with grains of a blue-green amphibole scattered throughout. Epidote/clinozoisite is common, and often appears intergrown with the amphibole. Opaque minerals and zircon comprise the heavy mineral suite.

IV. SEDIMENTARY PETROGRAPHY

Metamorphosed psammitic rocks are by far the most abundant in the study area, and make up a substantial portion of every unit described in the previous section, except the Persimmon Creek Gneiss. The mineralogy is consistent from unit to unit; essential minerals in all the metasandstones are quartz, plagioclase, biotite, and (generally) muscovite. Other metamorphic minerals (garnet, kyanite, sillimanite) and accessory minerals such as zircon, epidote, or opaque minerals constitute a small proportion of the total rock in almost every case.

Differences are subtle. In the field, only slight diferences in color and texture distinguish metasandstones of the various units from one another, and it is often necessary to use the relationships with other lithologies as a basis for identifying the rock unit. Laboratory study of thin sections and determination of modes, however, suggests that these subtle differences in texture and coloration often relate to the relative proportions of the predominant minerals, particularly in the amount of mica and the muscovite/biotite ratio. Dissimilarities are also seen in the quartz/feldspar ratio, and in the accessory mineral suite.

Sandstone Classification

Numerous sandstone classification schemes have been devised, but almost all are based on either the quartz-feldspar-clay ratio or the quartz-feldspar-lithic fragment ratio. Lithic fragments are generally impossible to recognize in highly recrystallized rocks, but

classification of metasandstones in the study area using the quartz-feldspar-clay ratio may be justified. The following assumptions are made:

(1) Metamorphic mica represents the clay fraction of the original sediments (Hadley and Goldsmith, 1963).
(2) Metamorphic reactions involving quartz and feldspar have been minimal, and the present abundance of these minerals reflects the original detrital composition.
(3) The effects of mechanical or chemical segregation of material, possible mechanisms for formation of gneissic banding (discussed in Section V), are minor in the samples used for this study.
(4) Effects of volume change are assumed to be neglibible.

Methods

A total of 45 thin sections were chosen for this study. The samples are predominantly metasandstone, with minor schistose metasandstone, biotite gneiss, and sillimanite gneiss. For each rock, an attempt was made to pick the most homogeneous and representative area from which to make a thin section.

Point counting was performed on all samples in order to estimate the volume of each constituent in the total volume of the rock, as demonstrated by Chayes (1956). Grain size of all samples is relatively uniform, and a point interval of 0.3 mm was chosen, which generally slightly exceeds the largest grain diameter (Van der Plas and Tobi, 1965). In almost every case, with a few noted exceptions, the quartz (Q) + feldspar (F) + mica (M = biotite and muscovite) constitutes greater than 95% of the total composition of

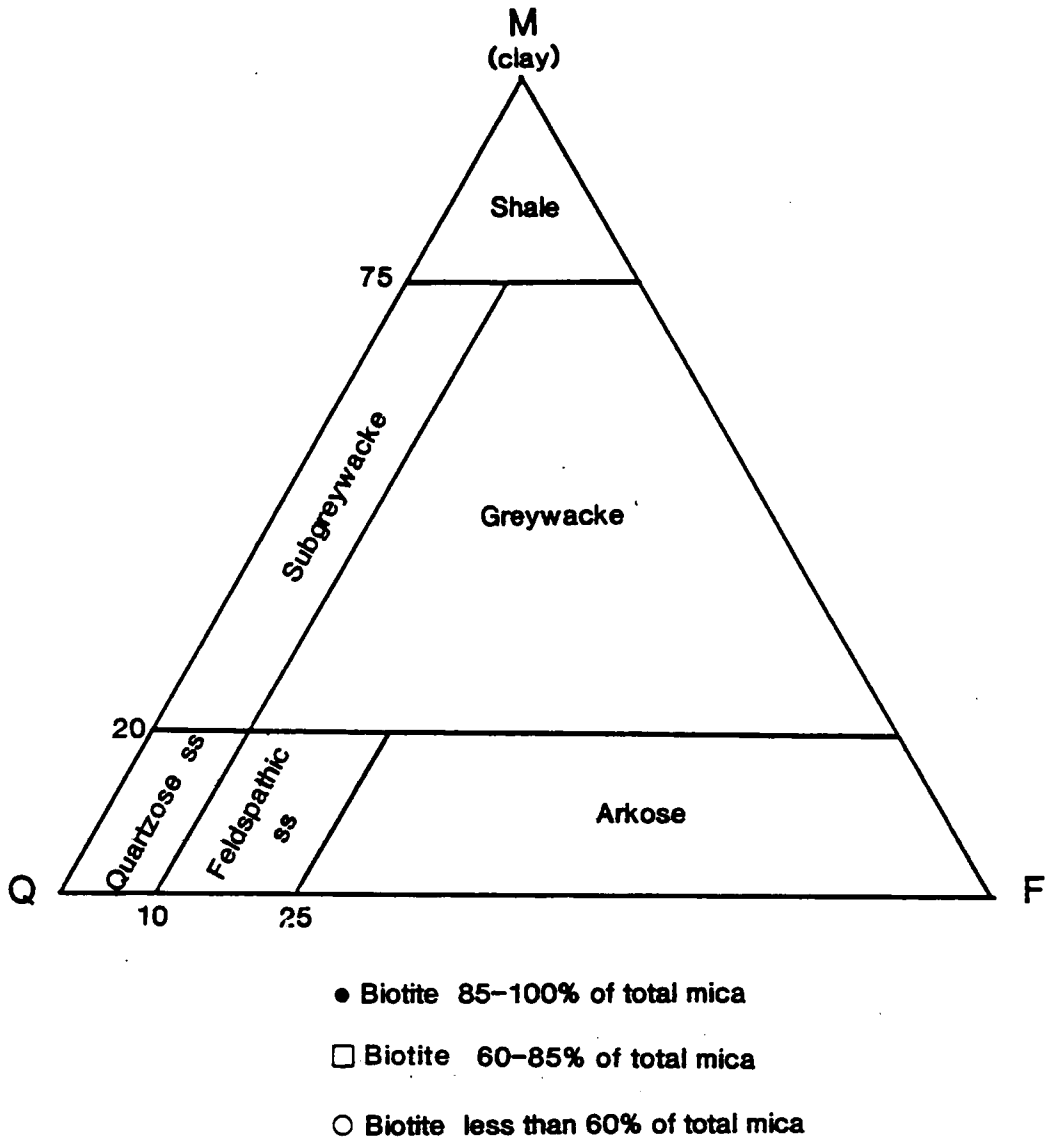
the sample. Results of the point counts were recalculated such that (Q + F + M) equals 100% (See Appendix A). Percentages of Q, F, and M were then plotted on ternary diagrams using a classification scheme after Pettijohn (1949) (Figs. 7a - 7f). The muscovite/biotite ratio is also shown graphically on the diagrams as biotite/biotite + muscovite.

Classification

Otto Formation: Only four samples from the Otto Formation are available, and these prove to have highly variable Q, F, and M ratios (Fig. 7b). Two plot as greywackes, one as feldspathic sandstone and one as quartzose sandstone. Biotite ranges from 57 to 93% of the total mica. The feldspar content is dominantly plagioclase but contains small amounts of microcline. Variability in these samples appears to be due more to mica vs. quartz content; in every sample the feldspar content is equal to or less than 15%.

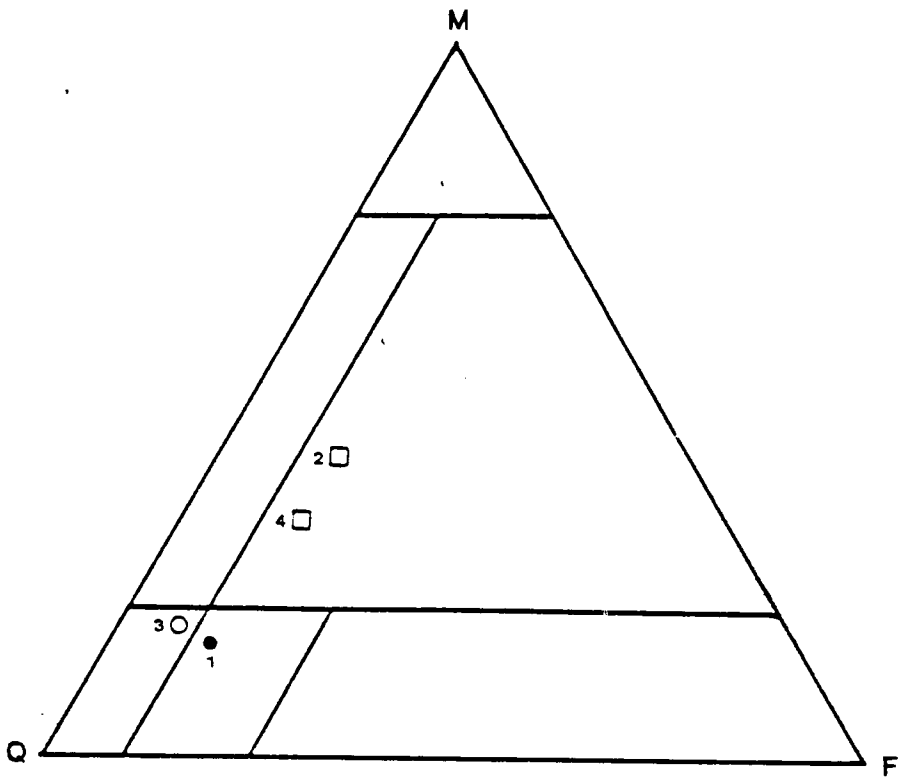
Coleman River Formation: Sixteen samples from the Coleman River Formation plot as an assemblage of greywacke, arkose, feldspathic sandstone, and shale (Fig. 7c). In only three samples does the biotite content constitute less than 89% of the total mica. One of these (#2763) plots as a shale, and contains abundant muscovite. Plagioclase makes up nearly 100% of the feldspar content in all samples.

Ridgepole Mountain Formation: Samples from this unit exhibit a wide variation, but all plot within the greywacke field (Fig. 7d). Sample #955, a schistose metasandstone, plots very near the shale field. Some samples contain abundant muscovite, up to 55% of the total mica. No potassium feldspar was identified in the feldspar fraction.



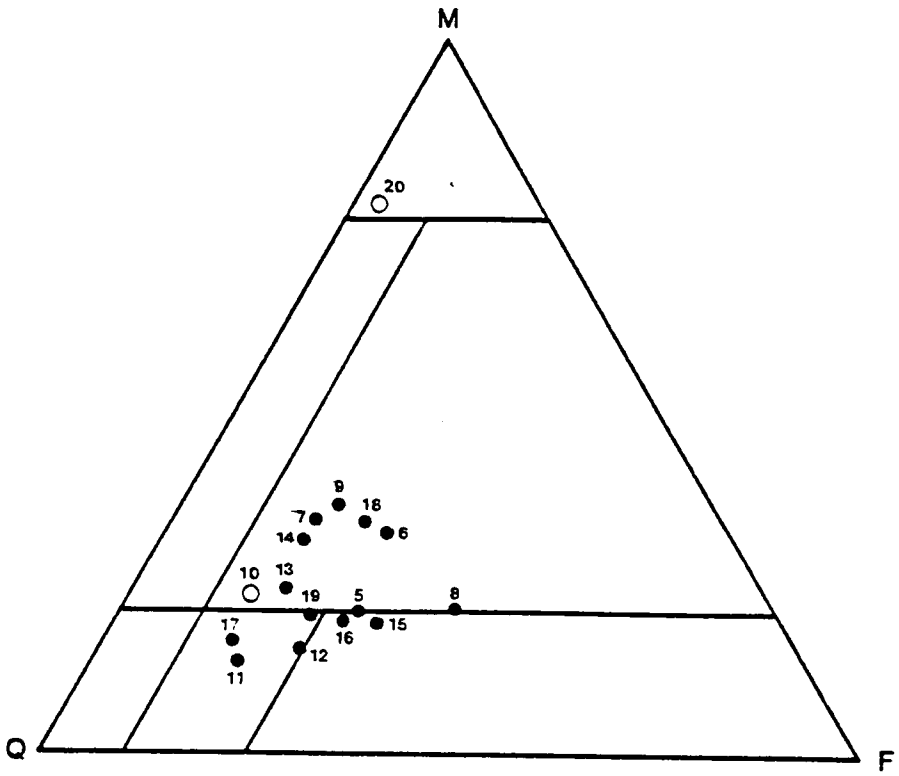
(a) Key for sandstone classifications shown in Figures 7b-7f. (Modified from Pettijohn, 1949.)

Figure 7: Classification of sandstones.



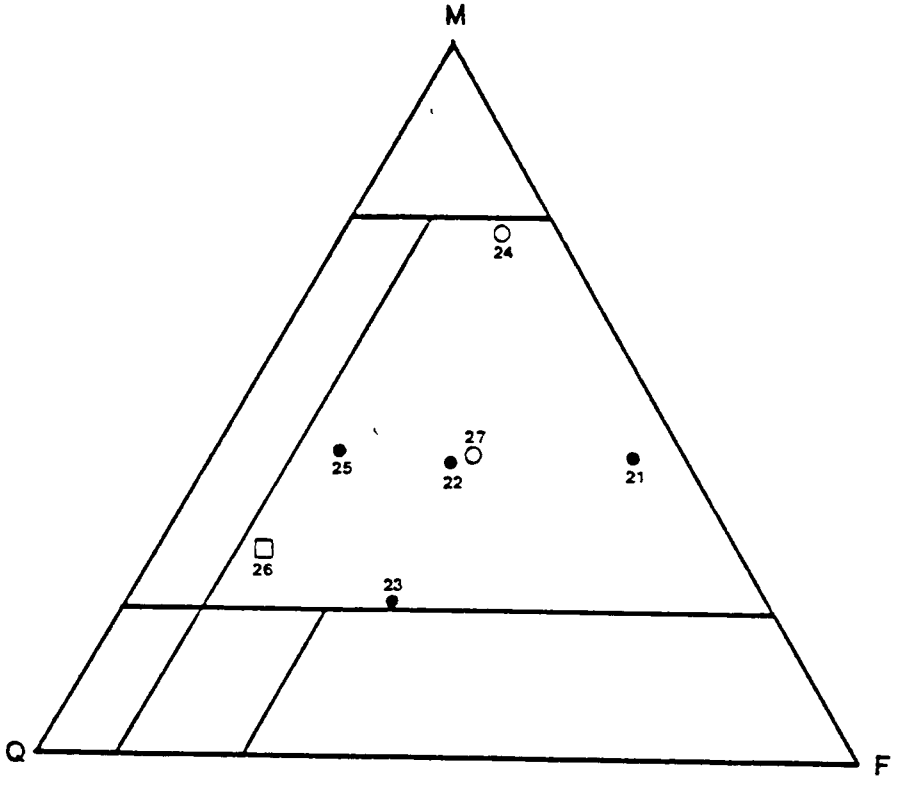
(b) Classification of 4 sandstones from the Otto Formation.

Figure 7 (continued)



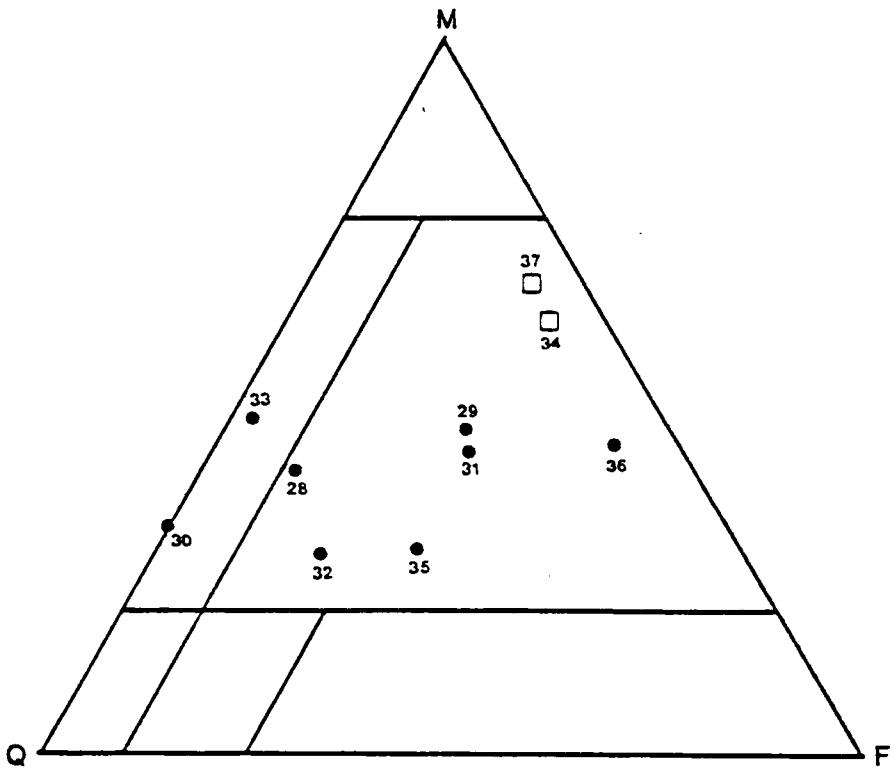
(c) Classification of 16 sandstones from the Coleman River Formation.

Figure 7 (continued)



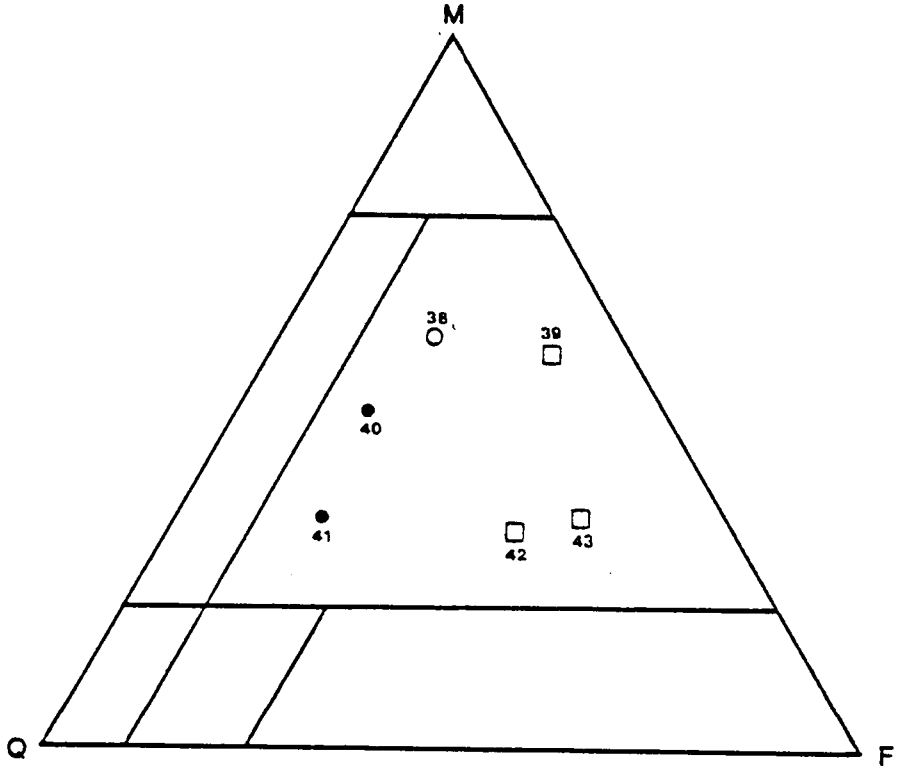
(d) Classification of 7 sandstones from the Ridgepole Mountain Formation.

Figure 7 (continued)



(e) Classification of 10 sandstones from the Tallulah Falls Formation.

Figure 7 (continued)



(f) Classification of 6 sandstones from the Corn Ridge formation.

Figure 7 (continued)

It must be noted that two samples (#882 and #2740) each contain abundant sillimanite and garnet, and Q + F + M for both rocks is only approximately 75% of the total. Therefore, these samples may not fairly represent the original sedimentary composition when plotted on a diagram of this type.

Tallulah Falls Formation: Results of analysis of ten thin sections from the Tallulah Falls Formation are shown in Fig. 8e. All plot within the greywacke or subgreywacke fields. The subgreywackes are examples of the "biotite quartzite" discussed earlier. Two samples of schistose metasandstone plot near the shale boundary, and both contain abundant muscovite, but in all other sections the total mica content consists of at least 85% biotite. Feldspar content is 100% plagioclase, and is consistently high in all rocks other than the subgreywackes.

Corn Ridge Formation: All samples of metasandstone from the Corn Ridge Formation plot as greywackes (Fig. 8f). Total mica content is difficult to estimate accurately, however, because of the abundant retrograde chlorite in some samples. Even so, the muscovite/biotite ratio appears to be variable. No potassium feldspar was identified in these samples.

Discussion

Major trends identified in this study are illustrated in Figure 8. The following relationships were recognized:

(1) If the interpretation of an upright sequence for tectonic unit 2 is correct, a transition occurs from greywacke and subgreywacke with a high

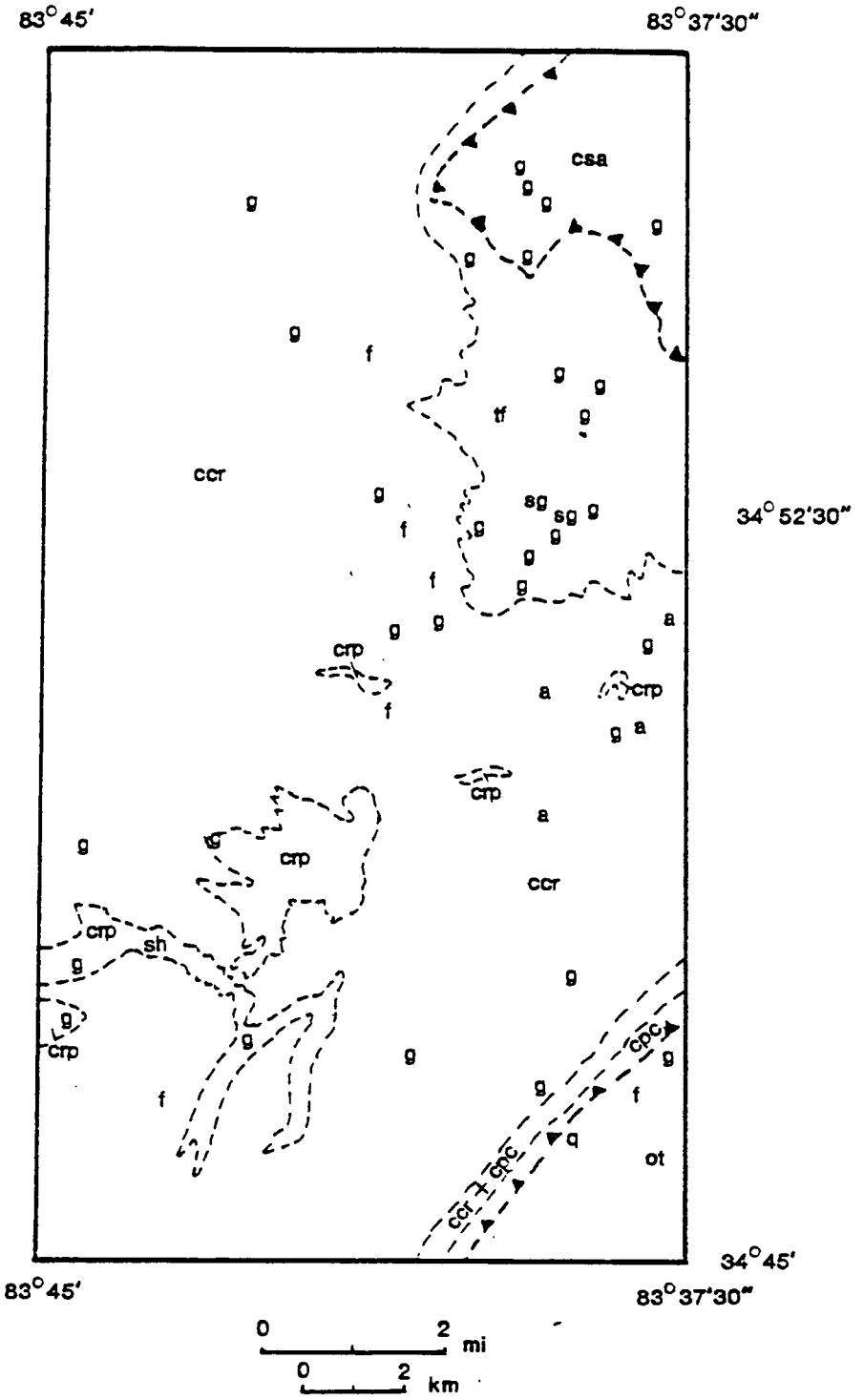


Figure 8: Distribution of sandstone types within the study area. g = greywacke; sg = subgreywacke; a = arkose; f = feldspathic sandstone; q = quartzose sandstone; sh = shale. Rocks units as in Plate II.

matrix content (Tallulah Falls Formation), to greywacke intermixed with "cleaner" arkosic and feldspathic sandstone (Coleman River Formation), to greywacke intermixed with alumina-rich sediments (Ridgepole Mountain Formation) (Fig. 9).

(2) Metagreywacke is associated with amphibolite (basaltic lava flows?) in the Shope Fork thrust sheet.

(3) The Otto Formation consists of an association of quartz-rich sandstone and highly aluminous shale with intercalated mafic rocks, possibly representing basaltic lava flows.

How well does the present composition of the rocks support this attempt at a sedimentary rock classification? Tables 1 and 2 compare average greywacke compositions and ranges from the Tallulah Falls, Coleman River, and Corn Ridge formations with those for "classical" greywackes from the Harz Mountains of Germany (data compiled by Huckenholz, 1963). Rocks from this study have a higher mica (clay) content; otherwise the analyses are in good agreement. The present samples also compare well with the data of Pettijohn and others (1972), although their modal analyses include a "rock fragment" category, which is impossible to determine in these highly metamorphosed rocks.

Comparison of the average composition of arkose and feldspathic sandstone from the Coleman River Formation with three samples from the "classic" region of Auvergne in France (Huckenholz, 1963) is shown in Tables 3 and 4. One striking difference stands out. In only one of the Auvergne samples is there any appreciable plagioclase; K-feldspar is dominant in the others. This appears to hold true for analyses of arkoses

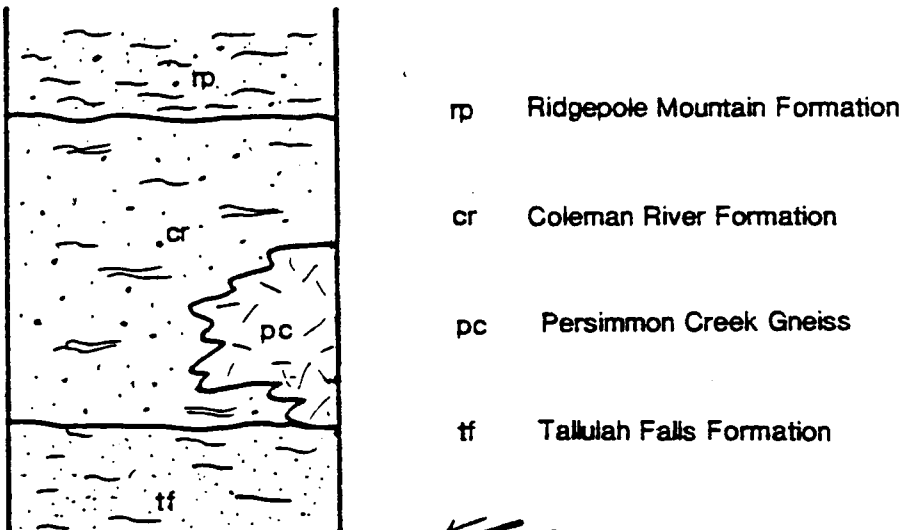


Figure 9: Stratigraphic relationships of rock sequences in tectonic unit 2.

TABLE 1: **Range (A) and average mineral composition (B) of metagreywackes from the Tallulah Falls Fm., Coleman River Fm., and Corn Ridge Fm., in volume percent.**

	Tallulah Falls Formation (5 samples)		Coleman River Formation (8 samples)		Corn Ridge Formation (4 samples)	
	A	B	A	B	A	B
Quartz	23-51	38	41-63	51	21-49	33
Plagioclase	11-32	25	13-26	19	16-42	23
K-Feldspar	-----	---	0-1.1	0.2	-----	---
Mica	26-42	35	19-33	28	31-55	41
Accessories	O.6-9	3	0.4-2.6	1.7	0.4-6.6	2.1

TABLE 2: **Range (A) and average mineral composition (B) of 88 samples of greywackes from the Harz Mountains, Germany (from Huckenholz, 1963).**

	A	B
Quartz	28-53	40
Plagioclase	20-42	31
K-Feldspar	2 - 6	3
Biotite	1 -10	5
Muscovite	1 - 7	3
Chlorite	4 -25	14
Carbonate minerals	0 - 6	2
Accessories	1 - 3	2

TABLE 3: **Range (A) and average mineral composition (B) of arkosic and feldspathic metasandstones from the Coleman River Formation, in volume percent.**

	Arkose (4 samples)		Feldspathic Sandstone (3 samples)	
	A	B	A	B
Quartz	38-51	47	56-67	61
Plagioclase	27-40	32	17-25	22
K-Feldspar	-----	---	-----	---
Mica	18-20	19	13-18	15
Accessories	1.8-3.7	2.6	1.7-2.8	2.4

TABLE 4: Mineral compositions of arkoses from the Auvergne in France (from Huckenholz, 1963).

	Sample 1	Sample 2	Sample 3
Quartz	53.1	51.0	37.7
Plagioclase	0.4	tr.	45.4
K-Feldspar	18.5	18.9	0.7
Kaolinite	17.0	19.3	-----
Montmorillonite	-----	-----	12.0
Mica	6.9	5.1	4.2
Barite	0.5	3.8	-----
Accessories	3.6	1.9	-----

from various regions, cited by Pettijohn, Potter, and Siever (1972), as well. The modes compiled by Pettijohn and others also show a notable sparsity of clay and mica.

In terms of the sedimentary aspect of the rocks, two questions stand out:

(1) What is the significance of the generally high biotite/muscovite ratio, in terms of the original clay content?

(2) Does the presence of plagioclase as the dominant feldspar provide any information about the original rock?

Biotite Content

Reactions which lead to the formation of biotite in metamorphic rocks are not well-understood (De Segonzac, 1970; Winkler, 1979), but two common reactions that may account for the bulk of biotite in pelitic rocks are given by Winkler (1979):

phengite + chlorite = biotite + Al-richer chlorite + quartz (1)

and

phengite + chlorite + microcline = biotite + phengite with less Fe, Mg (2).

Reaction (2) may also be significant in metagreywackes (Mather, 1970; Winkler, 1979).

According to Harker (1950), biotite is formed largely at the expense of muscovite and chlorite, and it is always chlorite which provides the starting point. Phengite is a mica which is very similar to muscovite but contains less Al, is richer in SiO_2, and

contains (Mg, Fe). It replaces illite at very low grades of metamorphism, and is the common white mica of the epizone (De Segonzac, 1970).

Therefore, the presence of abundant biotite in the metasedimentary rocks appears to indicate that the protolith contained abundant chlorite and illite-phengite as matrix material. The presence of a fine-grained matrix consisting of chlorite, white mica, and silt-size quartz and feldspar is characteristic of greywackes (Pettijohn and others, 1972), although the origin of this matrix is still subject to debate. The paucity of matrix material in Tertiary to Recent sediments accumulating in the same environments interpreted for ancient greywackes has been noted by several authors (Kuenen, 1966; Whetten and Hawkins, 1970). Cummings (1962) suggested that the bulk of the matrix in greywackes is diagenetic in origin, and this view has been supported by the work of Whetten and Hawkins (1970) and Brenchley (1969). It was suggested by both of these studies that the matrix formed by alteration of unstable clasts, particularly volcanic grains, during diagenesis.

Absence of K-feldspar

Potassium feldspar is rare to absent in all samples of metasedimentary rocks from the study area. Tartan-twinned microcline appears as a minor constituent in only a few samples. Plagioclase occurs as both twinned and untwinned grains. Five representative thin sections were stained by etching with hydrofluoric acid and immersing in Na cobaltinitride solution. None of the untwinned grains proved to be potassium feldspar. Three possible explanations for this absence are:

(1) Potassium feldspar was rare or absent as a detrital mineral in the original sediments, as a result of composition of the source rocks or weathering in the source area;

(2) K-feldspar was altered in composition during diagenesis; or

(3) K-feldspar was lost to metamorphic reactions such as (2) on page 63.

Plagioclase, commonly albite, is the dominant feldspar in greywackes, and K-feldspar may be present in small amounts or entirely absent (Pettijohn and others, 1972). This may be due to the albitization of K-feldspar during diagenesis by the action of Na-rich interstitial waters (Middeton, 1972). Modal analyses of unconsolidated "greywacke-type" sediments (Whetten and Hawkins, 1970), however, have a high plagioclase/K-feldspar ratio, contributed to a largely volcanic source terrain. Oligoclase is the dominant feldspar in metasandstones of this study. If albite was originally present, this is consistent with numerous observations of an increase in anorthite content in plagioclase with increasing metamorphism, typically from An_{6-8} to An_{28-32} (Goldsmith, 1982).

If reaction (2) on page 63 is correctly interpreted to have formed the bulk of the biotite present in the rocks, then microcline most certainly had to have been present in the original rock (Mather, 1970). This loss to metamorphism leads to the possibility that feldspar content was underestimated (and clay content overestimated) during point counting, and that the results do not exactly represent the original sedimentary rocks. These effects were probably similar for metasandstones throughout the study area, however, so that the overall trends would remain accurate.

Heavy Mineral Analysis

Studies of detrital heavy minerals in clastic rocks have been used in the past to aid regional and/or local correlations, and in interpretation of provenance or paleogeography of sediments. For metamorphic rocks, the heavy mineral suite is more complex because of the presence of heavy minerals of metamorphic origin. Some "ultrastable" minerals (for example, zircon, and tourmaline) are considered to have a detrital origin, however. Both experimental and empirical studies cited by Poldervaart (1955) show the following accessory minerals to be the most resistant to both chemical and mechanical weathering: zircon, tourmaline, rutile, and monazite. Saxena (1966) postulated that zircon could form completely by authigenesis, or be greatly altered in form by metamorphism, but several researchers (Kalsbeek, 1966; Marshall, 1966) refuted his conclusions, presenting arguments supporting the long-held confidence in the stability of zircons during sedimentary and metamorphic processes.

The heavy mineral composition of detrital sediments may be altered by the effects of weathering and abrasion (van Andel, 1959) as well as by post-depositional solution (van Andel, 1959; Blatt and Sutherland, 1969), but van Andel (1959) suggested that the effects of all the factors listed above are minimal for basins where the sediment supply is the result of rapid erosion and deposition (i.e., tectonically active areas such as that postulated for rocks of this study).

Preliminary analysis of heavy minerals from the Coleman River and Tallulah Falls Formations is included in this report. The use of this study for purposes of local correlation is limited, since only two (probably stratigraphically continuous) units are represented. Results of this analysis, however, may be used for:

(1) Confirmation of a sedimentary, as opposed to igneous, origin for rocks of this study.

(2) Comparison with other rocks in the eastern and western Blue Ridge (i.e., regional correlation).

Methods

Five samples of stream bed sediment were used for this study. Each sample was taken from near the head of a stream, on a ridge underlain by only one rock unit. The Tallulah Falls Formation is represented by three samples, the Coleman River Formation by two. Heavy liquid separation incorporated tetrabromoethane (S.G. = 2.95) and methods follow those described by Milner (1952), and used by Petree (1971). The magnetic separation procedure follows Friedman and Johnson (1982). Table 5 summarizes the occurence and relative abundance of heavy minerals in the five samples.

Detrital Heavy Minerals: Discussion

Results of this analysis show that the following heavy minerals of probably detrital origin are present:

Coleman River:

Zircon + Apatite + Sphene + Rutile + Opaques

Tallulah Falls:

Zircon + Apatite + Allanite + Sphene + Rutile + Opaques

TABLE 5: **Heavy mineral concentrations of samples from the Coleman River and Tallulah Falls Formations. R = rare, C = common, A = abundant.**

	Coleman River Formation			Tallulah Falls Formation	
	# 2706	# 2693	#2671-C	# 2771	# 2772
Zircon	A	A	A	A	A
Apatite	--	R	--	C	C
Tourmaline	--	R	--	--	R
Sphene	R	C	C	--	C
Rutile	R	R	R	R	--
Hornblende	R	--	--	--	R
Epidote	--	R	--	R	--
Clinozoisite	C	C	A	A	A
Zoisite	R	--	--	--	--
Allanite	--	--	--	R	--
Magnetite	C	C	C	C	A
Opaques (unidentified)	--	R	--	C	A
Sillimanite	R	C	C	C	C
Kyanite	R	R	--	--	--
Garnet	A	A	A	A	A
Biotite	A	A	A	A	A
Muscovite	A	A	A	--	C

Rutile and sphene, which are uncommon to rare, may be secondary. Tourmaline also rarely occurs; one grain was identified in each of two samples. Opaque minerals include magnetite and an unidentified variety, possibly ilmenite. Allanite was identified in thin sections from the Coleman River Formation, although not in the heavy mineral separates.

The heavy mineral suites of the two formations are similar, with the main difference being only the proportions of certain species. Particularly, anhedral rounded grains of apatite are much more common in the Tallulah Falls samples, and these samples also contain more opaque minerals.

Zircons are abundant in all samples, and most occur as colorless, somewhat cloudy grains in a variety of shapes (Fig. 10). In order of decreasing abundance, the shapes are: rounded oval to elliptical grains, anhedral "compound" grains with outgrowths, and elongate grains with pyramidal terminations. The least common type is clear, transparent oval to drum-shaped grains, some of which appear to be zoned.

The shape of zircons in metamorphic rocks is regarded as an indication of the original rock type. Paragneisses generally contain abundant well-rounded zircons with average elongation ratios (length divided by breadth) under or near 2; orthogneisses contain more euhedral grains with elongation ratios well over 2 (Kalsbeek, 1966). Others (Marshall, 1966; and other studies cited therein) have also considered well-rounded zircons diagnostic of metasedimentary rocks.

The anhedral, irregular or compound grains present in the samples may represent authigenic outgrowths. Saxena (1966) considered the clear, transparent grains to be

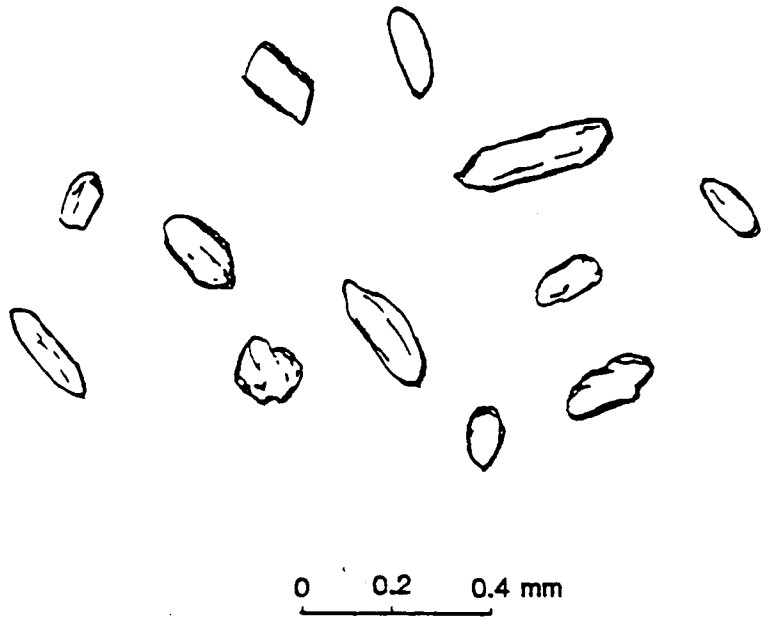


Figure 10: Typical habits of zircons from metasandstone of the Coleman River and Tallulah Falls Formations (this study).

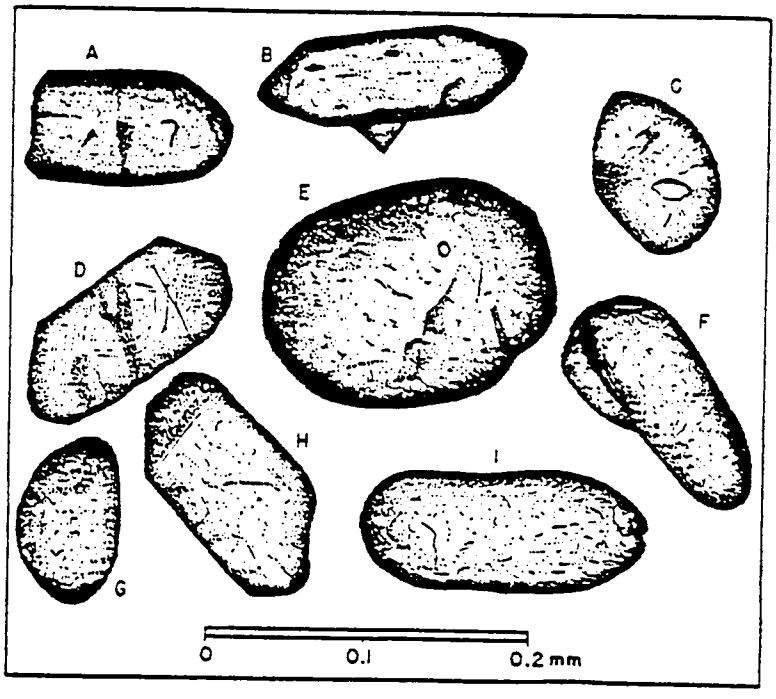


Figure 11: Typical habits of zircons in sandstones (from Poldervaart, 1955).

authigenic in origin; however, as discussed earlier, others have questioned the occurrence of completely authigenic zircons.

Figure 11 shows some typical habits of zircons in sandstones (Poldervaart, 1955). These compare well with the most abundant shapes present in the samples from this study (Fig. 10). This, along with the rounded nature of apatite grains, provides a good indication of the sedimentary origin of the rocks.

The association of heavy minerals in clastic sediments is often used as an indication of provenance. The association of zircon + apatite + rutile + opaques most closely typifies a source comprised of acid igneous rocks, or reworked sediments (Pettijohn and others, 1972). It must be kept in mind, however, that heavy minerals typical of other sources (for example, mafic or earlier metamorphic sources), are much less stable under conditions of chemical or physical weathering, as well as under metamorphic conditions, and may well have not survived.

Comparison with Other Blue Ridge Rocks

Table 6 compares the detrital heavy mineral suites from the Coleman River and Tallulah Falls formations of this study to those found in the Tallulah Falls to the east of the study area (Hatcher, 1973), to two units within the Ocoee Supergroup (Hadley and Goldsmith, 1963), and to rocks of the Alligator Back Formation of North Carolina and Virginia (Rankin and others, 1963; Conley, 1987). Some biotite and epidote were recognized as detrital minerals in rocks of the Snowbird Group which are of a relatively lower metamorphic grade (Hadley and Goldsmith, 1963). Otherwise the suites are quite similar in most respects, but a major difference in the suites appears to be the amount of

TABLE 6: Comparison of primary heavy mineral suites in some rocks of the southern Appalachian Blue Ridge.

Coleman River Formation: (this study)	Zircon + Magnetite + Opaques (unidentified) + Apatite + Sphene + Rutile (tr) + Tourmaline (tr)
Tallulah Falls Formation: (this study)	Zircon + Magnetite + Opaques (unidentified) + Apatite + Allanite + Sphene + Rutile (tr) + Tourmaline (tr)
Tallulah Falls Formation: (Hatcher, 1973)	Zircon + Opaques + Sphene + Allanite + Apatite (tr) + Tourmaline (tr)
Snowbird Group (Ocoee): (Hadley and Goldsmith, 1963)	Zircon + Apatite + Magnetite + Sphene + Epidote + Biotite + Tourmaline (tr)
Great Smoky Group (Ocoee): (Hadley and Goldsmith, 1963)	Zircon + Apatite + Magnetite + Tourmaline (common)
Alligator Back Formation, NC and VA: (Rankin and others, 1963; Conley, 1987)	Zircon + Magnetite + Ilmenite + Graphite + Tourmaline (common) + Sphene

tourmaline. It occurs in only trace amounts in all units except the Great Smoky Group and the Alligator Back Formation, in which tourmaline is common.

It is tempting to interpret the metasedimentary rocks of the eastern Blue Ridge as high-grade (deeper-water?) equvalents of those in the western Blue Ridge, because of their proximity and to their similarity as thick clastic sequences (particularly the Great Smoky Group). This interpretation has been suggested by several workers (e.g., Hatcher, 1973). The limited scope of this study does not invalidate this interpretation by any means. It does suggest, however, that the source area for Great Smoky Group sediments was somewhat different than that of the Coweeta Group and Tallulah Falls Formation, based upon the abundance of tourmaline. Supporting evidence is found in the abundant potassium feldspar of Great Smoky rocks (Hadley and Goldsmith, 1963), which have been metamorphosed to as high as kyanite grade, and also in petrographic descriptions of the Great Smoky Group vs. the "Tray Mountain Gneiss" (Coleman River Formation) by Shellebarger (1980).

V. METAMORPHISM

Metamorphism in rocks of the study area is classic Barrovian, as first described by Barrow (1893) in the Dalradian series of Scotland, and reached the first sillimanite zone of the amphibolite facies (Eskola, 1939; Turner, 1968). In the study of metasedimentary rocks, index minerals and mineral assemblages used to interpret metamorphic conditions are most commonly found in pelitic rocks, and may be rare or absent in psammitic or calcareous rocks due to the paucity of unstable constituents (i.e. clays) in the original rock. Pelitic rocks are in the minority throughout the study area, but enough samples exist to provide a number of clues about the metamorphic history. In the following sections, characteristic mineral assemblages and textural relationships will be described and used to interpret minimum and maximum temperatures and pressures that may have occurred.

Isograds

Sporadic distribution of sillimanite throughout the study area indicates that most of it is within the lower sillimanite zone (Fig. 12). Samples from the Corn Ridge formation within the Shope Fork thrust sheet, contain kyanite, which appears abruptly as the trace of the fault is crossed. Because of the lack of laterally continuous mapping in this area, it is not clear whether the Shope Fork fault has placed kyanite grade rocks onto sillimanite grade, or whether the kyanite-sillimanite isograd obliquely crosses the fault. Recent mapping in the adjacent Hightower Bald quadrangle (Hopson, pers. comm.) suggests that the fault may not truncate the isograd, as it appears to do in this area.

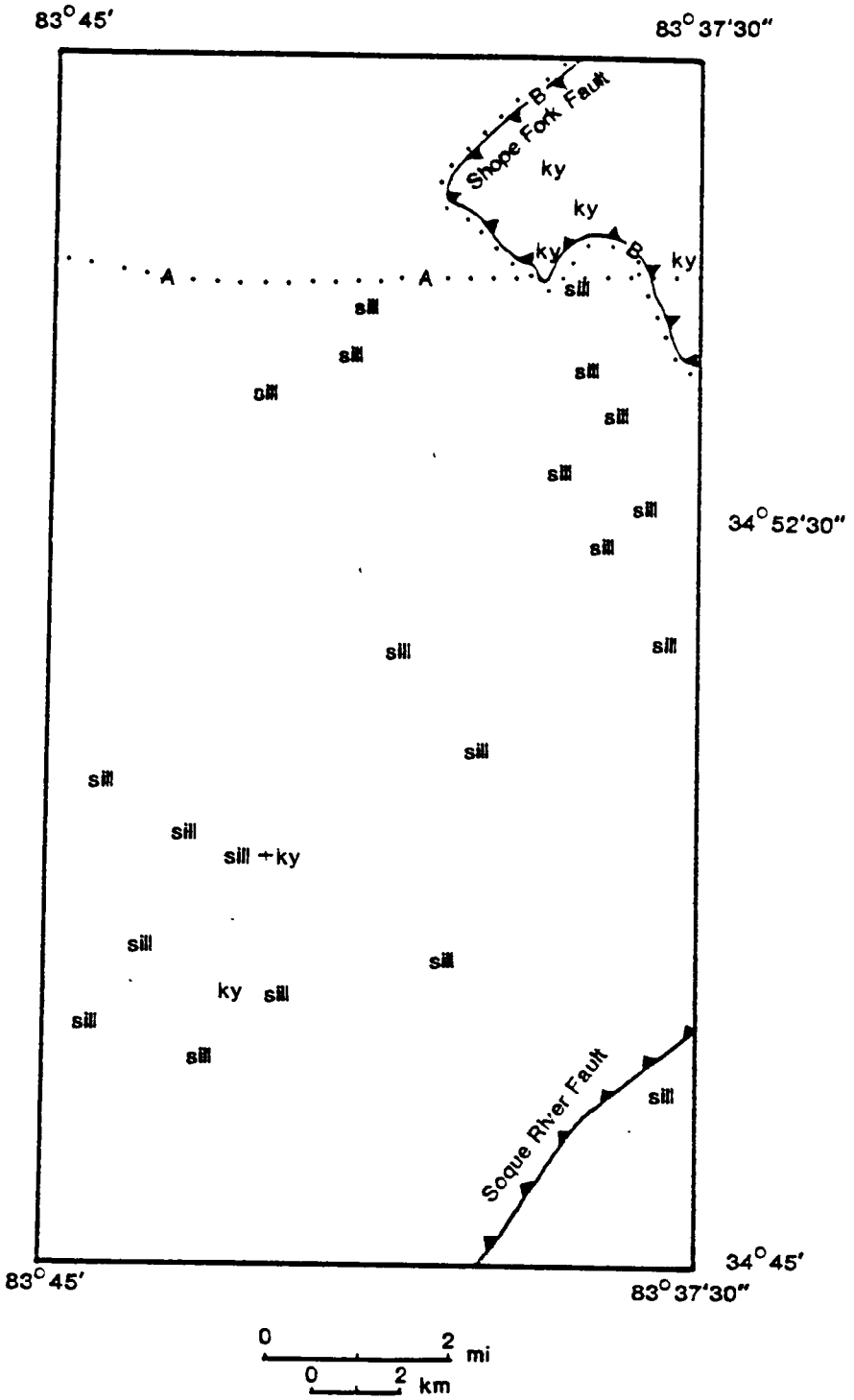


Figure 12: Distribution of aluminum silicates in the study area. (Based upon thin sections or hand samples.) Included are alternative interpretations of the location of the kyanite-sillimanite isograd. Interpretation A: isograd crosses Shope Fork fault. Interpretation B: isograd parallels Shope Fork fault. sill = sillimanite; ky = kyanite.

Metamorphic grade in the Otto Formation, southeast of the Soque River fault, reached sillimanite grade. Staurolite and kyanite were reported from Otto schists immediately to the southwest (Farrell, 1986), and by Hatcher (1980) to the northeast. Garnet and biotite are the only index minerals observed in the majority of hand samples and thin sections from the study area, but tiny radiating fibers of sillimanite were identified in one thin section from a sample correlated with the Otto Formation in this study. Albino and Gillon (1987) noted the presence of sillimanite, although rarely preserved, in correlative rocks of the Dahlonega belt along strike southwest of the study area. They suggest that the kyanite and/or staurolite assemblages result from retrograde effects.

Metamorphic Assemblages and Reactions

Characteristic Assemblages and Textures

Characteristic metamorphic assemblages and textures identified from thin section analyses are described below. All the assemblages are assumed to be essentially products of prograde metamorphism, although some samples exhibit textural or mineralogical evidence for disequilibrium. Appendix B contains a list of mineral abbreviations used in the following sections.

Otto Formation: Four samples of psammitic rocks from the Otto Formation contain the following assemblages:

(1) qtz+plag+bio+musc

(2) qtz+plag+bio+musc+gar

(3) qtz+plag+bio+musc+sill

The dominant foliation in the samples is defined by subparallel biotite and muscovite. However, some muscovite grains cross this foliation in every sample, and one section (#2734) contains cross-cutting biotite as well. Garnet occurs as both anhedral grains with inclusions of biotite and quartz, and also as tiny euhedral grains (Fig. 13). Sillimanite, which is present in only one sample (#49), occurs as radial fibers in contact with biotite and quartz grains (Fig. 14).

Tallulah Falls Formation: The following metamorphic assemblages were identified in pelitic and psammitic rocks from the Tallulah Falls Formation:

(1) qtz+plag+bio+musc

(2) qtz+plag+bio+musc+gar

(3) qtz+plag+bio+musc+gar+sill

(4) qtz+bio+musc+sill.

The dominant foliation in all sections is defined by aligned biotite grains. Muscovite occurs both parallel to the biotite and as cross-cutting, typically porphyroblastic grains. Garnet is subhedral to euhedral, and generally appears to be homogeneous except for rare inclusions of biotite and quartz. Sillimanite occurs in two forms, as needles enclosed within porphyroblastic muscovite, and as radial fibers concentrated at boundaries of plagioclase grains. The latter type is occasionally in contact with biotite.

Coleman River Formation: Metamorphic assemblages present in samples of metasedimentary rocks of the Coleman River Formation are:

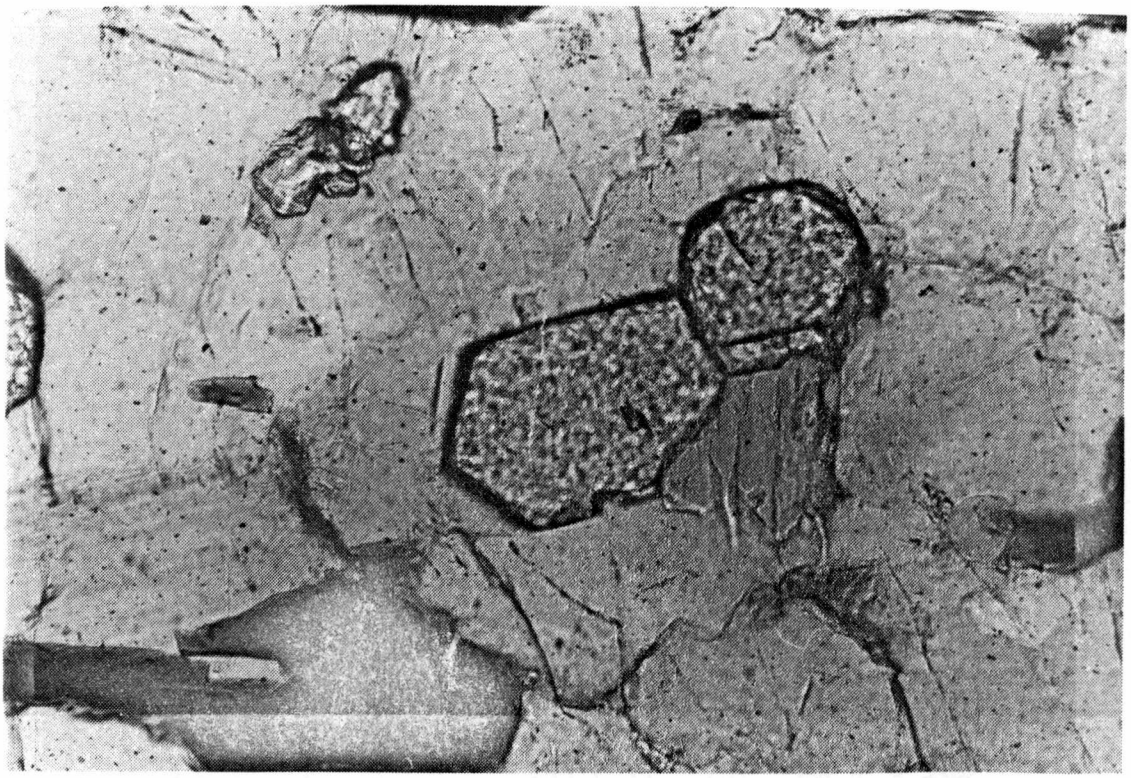


Figure 13: Photomicrograph of sub- to euhedral, inclusion-free garnets from the Otto Formation. Scale bar is 1 mm.

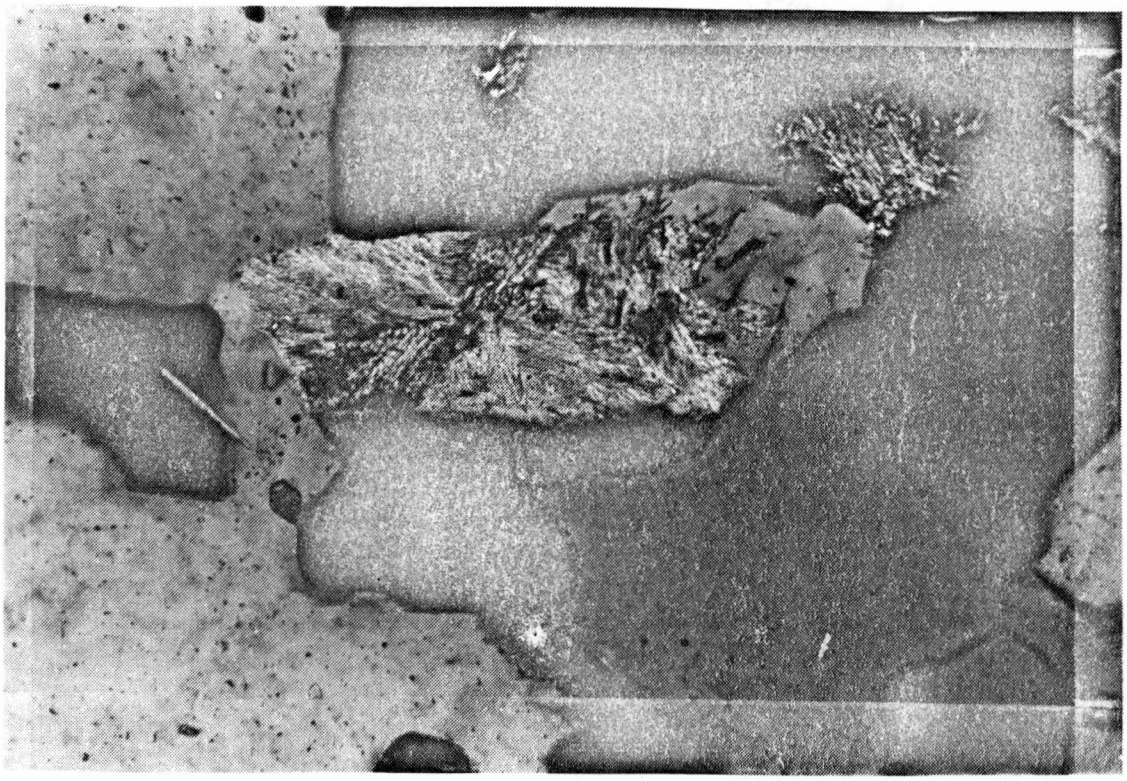


Figure 14: Radiating sillimanite fibers at quartz grain boundaries. Sample from metasandstone of the Otto Formation. Scale bar is 1 mm.

(1) qtz+plag+bio+musc

(2) qtz+plag+bio+musc+gar

(3) qtz+plag+bio+musc+gar+sill

(4) qtz+plag+bio+gar

Biotite exhibits two preferred orientation fabrics in many samples, and muscovite is parallel to either or both, although it seldom displays the porphyroblastic habit described in the previous section. Garnet is commonly subhedral to anhedral, and contains inclusions of biotite, muscovite, and quartz. Two samples (#645 and #2296) contain tiny euhedral garnets that have overgrown biotite and muscovite. Sillimanite was identified in two samples. In #294 it occurs as a fibrous mass in contact with biotite, plagioclase, and quartz. Sample #2241 contains anastomosing bands of fibrous sillimanite surrounded by muscovite.

Ridgepole Mountain Formation: The following assemblages are present in pelitic and psammitic rocks of the Ridgepole Mountain Formation:

(1) qtz+plag+bio+musc

(2) qtz+plag+bio+musc+gar

(3) qtz+plag+bio+musc+gar+sill

(4) qtz+plag+bio+musc+gar+ky

(5) qtz+plag+bio+musc+gar+ky+sill

(6) qtz+plag+musc+ky+sill

The dominant foliation is defined by biotite, but muscovite exhibits variable orientations. These include muscovite parallel to the dominant foliation, cross-cutting

the foliation at a high angle, and random, with no preferred orientation. Garnets are typically large (5 mm), sub- to anhedral, with inclusions of biotite, quartz, or opaque minerals (Fig.15). Fractured garnets are common. Sillimanite occurs as radial fibers at plagioclase or quartz grain boundaries, as fibrous aggregates interleaved with biotite, and as fibers, needles, and prismatic grains surrounded by muscovite. Kyanite occurs in one sample (#778A) as subhedral, porphyroblastic grains which contact muscovite, biotite, and garnet. Kyanite and sillimanite coexist in two sections (#882 and #884). In sample #884, kyanite is typically porphyroblastic, and is surrounded by muscovite and sillimanite. In sample #882 kyanite clearly cross-cuts the sillimanite fibers (Fig.16).

Corn Ridge Formation: Pelitic and psammitic rocks from the Corn Ridge formation yield the following metamorphic assemblages:

(1) qtz+plag+bio+musc

(2) qtz+plag+bio+musc+gar

(3) qtz+plag+bio+musc+gar+ky

(4) qtz+bio+musc+gar.

In most samples the dominant foliation is defined by biotite, with parallel or subparallel muscovite. Muscovite also cross-cuts the foliation at a high angle. Garnet is generally subhedral, contains inclusions of biotite, and is highly fractured, but one sample contains relatively small euhedral garnets. Kyanite occurs most commonly as small grains surrounded by muscovite, or entirely enclosed in porphyroblastic muscovite grains.

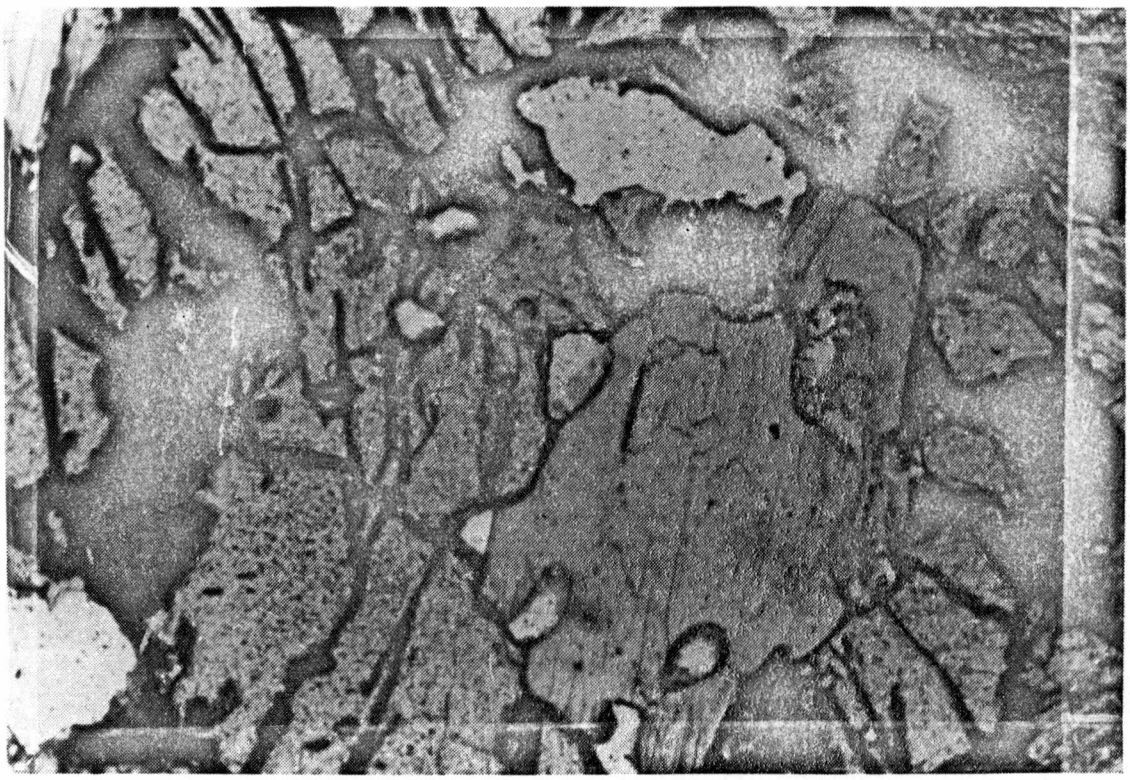


Figure 15: Photomicrograph of anhedral garnet with inclusions of biotite and quartz. Sample from the Ridgepole Mountain Formation. Scale bar is 1 mm.

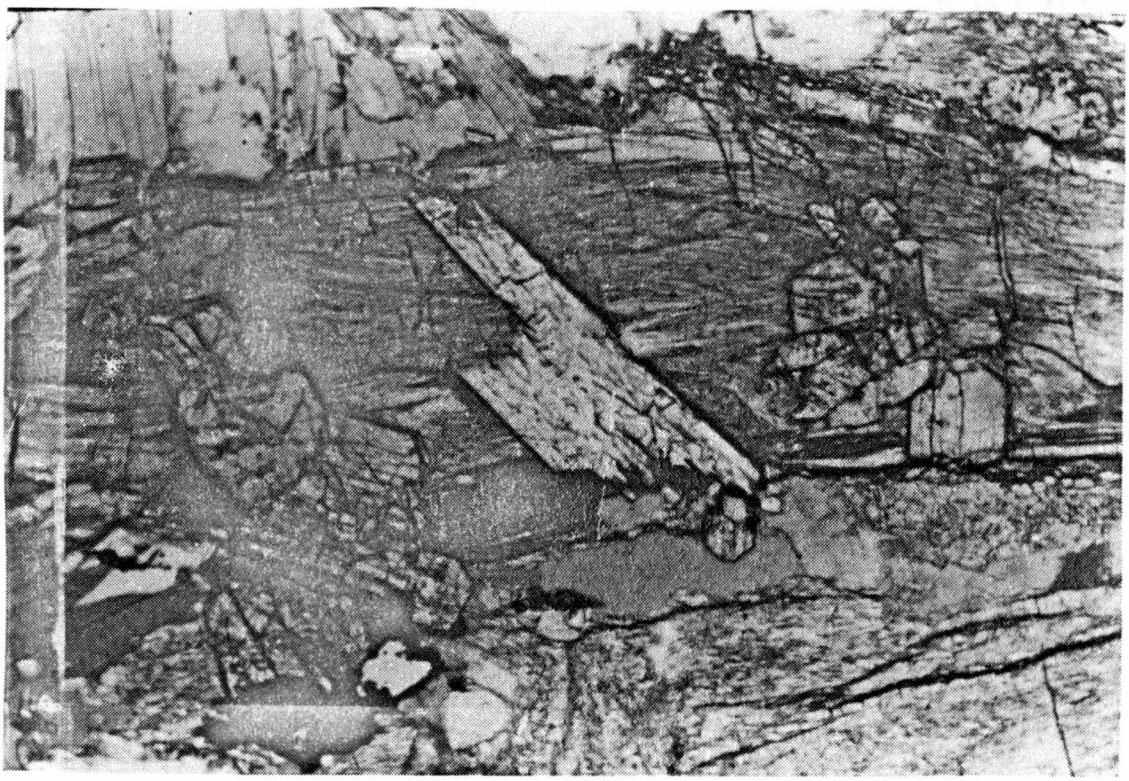


Figure 16: Photomicrograph of sillimanite gneiss from the Ridgepole Mountain Formation. Kyanite grains cross-cut the foliation containing sillimanite. Scale bar is 1 mm.

Retrograde chlorite (pennine), abundant in pelitic rocks and occasionally observed in rocks of psammitic composition, is not included with the prograde assemblages above. The chlorite displays distinct retrograde textures, occurring within fractures in garnet, or as replacement of biotite.

Reaction Relationships

Possible reactions for formation of biotite were discussed in Chapter IV. Garnet is abundant in both psammitic and pelitic rocks of all units. This garnet is assumed to be almandine, although no chemical analyses were made to confirm this assumption. Almandine is the common garnet found in metasedimentary schists and gneisses affected by regional metamorphism (Harker, 1932; Phillips and Griffin, 1981), and garnets of the present study normally range from pale pink to red, which is characteristic of almandine. In many cases, garnet is the only metamorphic index mineral (other than micas) present in the rocks, although the sporadic occurrence of aluminum silicate minerals assures us that the rocks are well above garnet grade. In the study samples, garnet is commonly in contact with quartz, plagioclase, or muscovite, but is pervasively associated with biotite, with no reaction textures. This suggests that biotite may have formed simultaneously. A reaction that satisfies the observed textural relationships is the following:

chl+musc+qtz = alm+bio+H_2O (Thompson and Norton, 1968; Hollister, 1969).

In the study samples which contain kyanite as the only aluminum silicate, kyanite is always in contact with muscovite; muscovite commonly completely encloses it. Other minerals observed in contact with kyanite are biotite, garnet, and epidote. Coexisting

kyanite and biotite exhibit no reaction textures at grain contacts, and appear to be in equilibrium. In one sample (#2400), some kyanite occurs as an aggregate of small euhedral grains at the edge of a corroded garnet, suggesting a reaction involving both minerals.

The relationship between kyanite and muscovite is difficult to discern, especially in samples from the Corn Ridge formation. These commonly show kyanite completely enclosed by porphyroblastic muscovite. This texture is ambiguous, and could represent either older kyanite engulfed by younger muscovite, or kyanite as a younger phase growing in the muscovite (Spry, 1969). In other samples where kyanite is not completely enclosed in muscovite, textures more clearly suggest that kyanite formed from a reaction involving muscovite. The possibility exists for two phases of muscovite growth, resulting in muscovite that is both older and younger than kyanite.

In this study, sillimanite occurs as 3 distinct types:

(a) Aggregates of fibrolite interleaved with biotite and/or muscovite;
(b) Acicular grains completely enclosed in muscovite;
(c) Tiny radial fibers concentrated at boundaries of plagioclase, or less commonly, quartz grains.

Types (a) and (b) may be coincident, and very possibly share a common origin. Type (c), however, is isolated from the other types, and is generally of sporadic occurrence in rocks which otherwise contain no aluminum silicates. Types (a) + (b) and type (c) may represent two generations of sillimanite growth, possibly involving different reactions, as suggested by Hopson (1984). In one sample (#2740), there is good

evidence for garnet breaking down to produce sillimanite + biotite. The presence of coexisting kyanite and sillimanite, described on page 81, appears to be indicative of disequilbrium rather than inversion of kyanite to sillimanite.

Numerous reactions have been cited as significant in producing aluminum silicate minerals, but it is difficult to speculate upon which ones may have occurred in rocks of this study because of ambiguous textural relationships. The following reaction offers a possible explanation for the concentration of sillimanite at plagioclase grain boundaries, as in type (c), above:

Na-musc+gar = bio+sill+ab+H_2O (Guidotti, 1970).

In samples where garnet appears to be a reactant, such as in sample #2740 described above, a viable reaction may be:

gar+musc = AS+bio+qtz (Yardley, 1977; McLellan, 1985).

Migmatization

Localized migmatitic layers interspersed with metasandstone are sporadically distributed within rocks of tectonic unit 2 (Coleman River, Ridgepole Mountain, and Tallulah Falls Formations). These are composed of light-colored layers (leucosomes) of quartz and plagioclase (no K-feldspar has been identified) separated from the surrounding rock by narrow dark selvedges consisting primarily of biotite with or

without garnet (Fig. 17). These are of interest as they may give some indication of the maximum conditions reached during metamorphism.

The most commonly proposed mechanisms for the existence of granitic to granodioritic layers within pre-existing metamorphic rocks are metasomatism, or the transformation of rocks in the solid state by addition of elements from an external source; injection of granitic material by intrusions, often parallel to the foliation; and partial melting, or anatexis, giving rise to a melt of granitic composition and a crystalline residue composed mainly of mafic minerals, such as biotite. Precise determination of the origin of the migmatitic layers in rocks of this study was not possible. The selvedges could have formed during incipient partial melting (Vernon, 1976), and the commonly contorted structures observed within migmatite layers may be evidence for a semi-molten state of the material (Mason, 1978; Winkler, 1979), also suggesting that anatexis could have occurred.

Granitic horizons within rocks of the Otto Formation, described on page 31, are thick, continuous non-contorted layers comprised of quartz + plagioclase + microcline + biotite + muscovite. These do not have the same character as the migmatites described above. Although partial melting cannot be ruled out, I suggest that these layers resulted from injections of granitic material parallel to the foliation. Some granitic layers exhibit a weak foliation, so the material must have been emplaced, at the latest, prior to the close of the latest metamorphic episode.

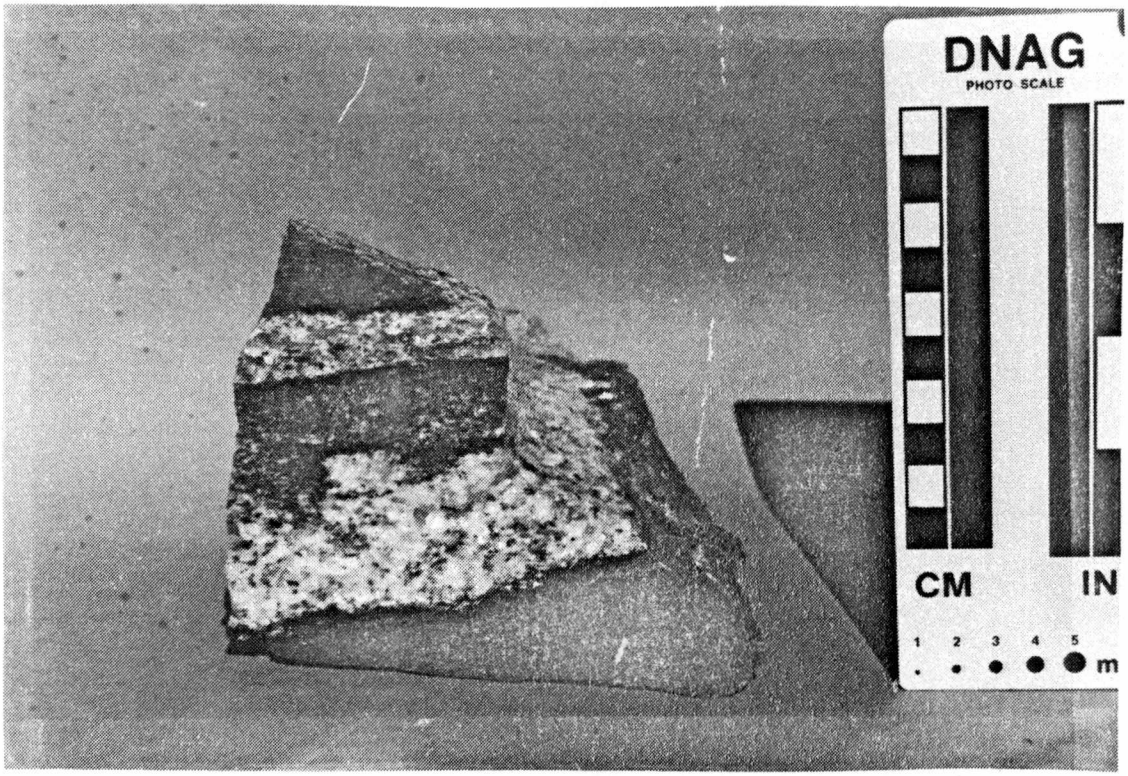


Figure 17: Hand sample of migmatitic gneiss from the Coleman River Formation.

Metamorphic Conditions

Because of the apparently complex metamorphic history of this area and the limited extent of data, only unconstrained inferences may be made on the maximum temperatures and pressures attained during metamorphism. Comparison of the data with standard PT diagrams, however, should yield useful information, even if somewhat broad.

Figure 18 is a simple petrogenetic grid showing relevant reactions with respect to experimental PT conditions. The problems of variable results in the study of the Al_2SiO_5 system are well illustrated by Vernon (1976), and discussed by numerous others (i.e., Turner, 1981). The limits, shown in Figure 18, represent a combination of two well-documented studies. The band designated "anatexis" (Curve 1) represents the beginning of anatexis in gneiss (Winkler, 1979). Curve 2 represents the second sillimanite, or "muscovite-out", zone, signifying the formation of K-feldspar and sillimanite at the expense of muscovite.

Comparison of the data with Figure 18 provides upper and lower limits of temperature and pressure for most of the region under investigation. Most of this area is within the first sillimanite zone, as indicated by the presence of sillimanite + muscovite, with no evidence for K-feldspar formation. Therefore, the lower limit is the kyanite-sillimanite boundary, at temperatures of 500-600° C and pressures of 4-6 kbar (or depths of 10-20 km; see Turner, 1981). Minimum conditions in areas of localized migmatization, if it resulted from anatexis, probably reached 625-650° C, within the same pressure range as above. The sporadic development of migmatitic layers could

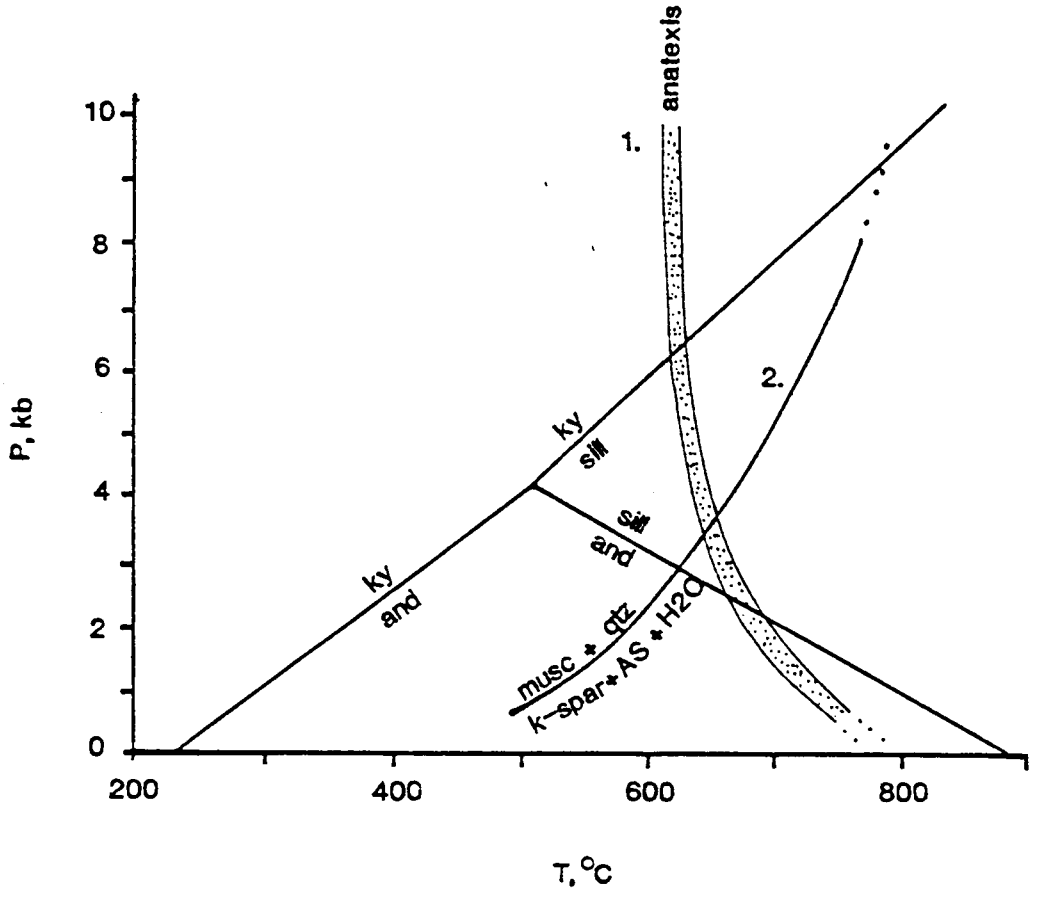


Figure 18: Petrogenetic grid for the system Al_2SiO_5. Modified from Vernon (1976), after Richardson, Gilbert, and Bell (1969) and Holdaway (1971). Curve 1 (anatexis) from Winkler (1979). Muscovite dehydration curve (2) from Turner (1981).

have resulted from variations of P_{H20} that may be common in areas of extensive deformation in response to stress and chemical potential gradients (Eckert, 1984). The maximum conditions are defined by Curve 3, representing the reaction: musc + qtz = K-spar + sill + H_2O, and corresponding to temperatures of approximately 650-725° C, and pressures of at least 4 kbar. The crystallization of kyanite subsequent to sillimanite in rocks that contain both Al_2SiO_5 polymorphs could indicate a rise in pressure. Alternatively, kyanite grew as temperatures waned, with the metastable persistence of sillimanite.

Limits on conditions for rocks of the Shope Fork thrust sheet, that are of kyanite grade in the study area, are more difficult to narrow down. These rocks exhibit no evidence of migmatization; therefore the maximum possible temperature is approximately 625° C, as inferred from Figure 26. Pressures greater than 3.5 kbar are indicated by the presence of kyanite rather than andalusite as the Al_2SiO_5 polymorph (Turner, 1981). In a study of pelitic rocks from the Grampians of Scotland, McLellan (1985) calculated temperatures of 540-625° C and pressures of 4.5-6.5 kbar for kyanite-bearing rocks with no sillimanite overprint.

Retrograde Effects

Clear evidence for retrograde metamorphism is observed to any extent only in rocks of the Shope Fork thrust sheet (Corn Ridge formation). The most distinct feature is the presence of abundant chlorite (pennine) replacing garnet and biotite. Chlorite occurs as radial clusters, randomly oriented grains, and occasionally appears to parallel the foliation due to pseudomorphous replacement of biotite. Garnets tend to be highly

fractured, and chlorite commonly surrounds garnet and fills these fractures. Tiny grains of muscovite occur within fractures in one sample.

Evidence for retrogression in units other than the Corn Ridge formation is present but rare; only a sporadic grain of chlorite, or possible replacement of an aluminum silicate mineral by muscovite in one sample. Localization of retrograde effects is common (Vernon, 1976). In some cases this may be related to deformation, such as folding during metamorphism, resulting in a local concentration of water in fold hinges and local development of hydrous assemblages (Carpenter, 1968; *in* Vernon, 1976). Abundant folds do occur throughout the area, and this cause cannot be ruled out. An alternative suggestion is that the retrogression is related to emplacement of the Shope Fork fault, because extensive effects occur only in the Shope Fork thrust sheet.

VI. STRUCTURAL GEOLOGY

The study area is characterized by a complex structural pattern resulting from polyphase, superposed folding and faulting. During the course of field study, the orientations of approximately 1900 foliations and 400 folds were measured (Plate II). Through observation of the relationship of foliations to folds, and of overprinting relationships between folds, as many as six generations of folding may be resolved. In the following section, evidence for this polyphase history is examined, along with evidence for large-scale faulting.

Mesoscopic Structures

Foliations

Four separate foliations were observed in rocks of the study area, and are described below. S_2, the dominant surface, is well-preserved and observable in almost every outcrop, whereas the other surfaces occur sporadically. Seldom were more than two foliations observed in any one outcrop. Therefore, correlations were made primarily on the mineralogy associated with the surfaces, and on cross-cutting relationships where they could be deduced. Figure 19 is a generalized form surface map of the orientation and distribution of the foliations. Plate III is a more detailed form surface map illustrating the orientation of the dominant foliation, S_2.

S_1: The earliest foliation is preserved as a "pinstriping" texture, most commonly observed in metasandstones of the Coleman River Formation, and is defined by thin,

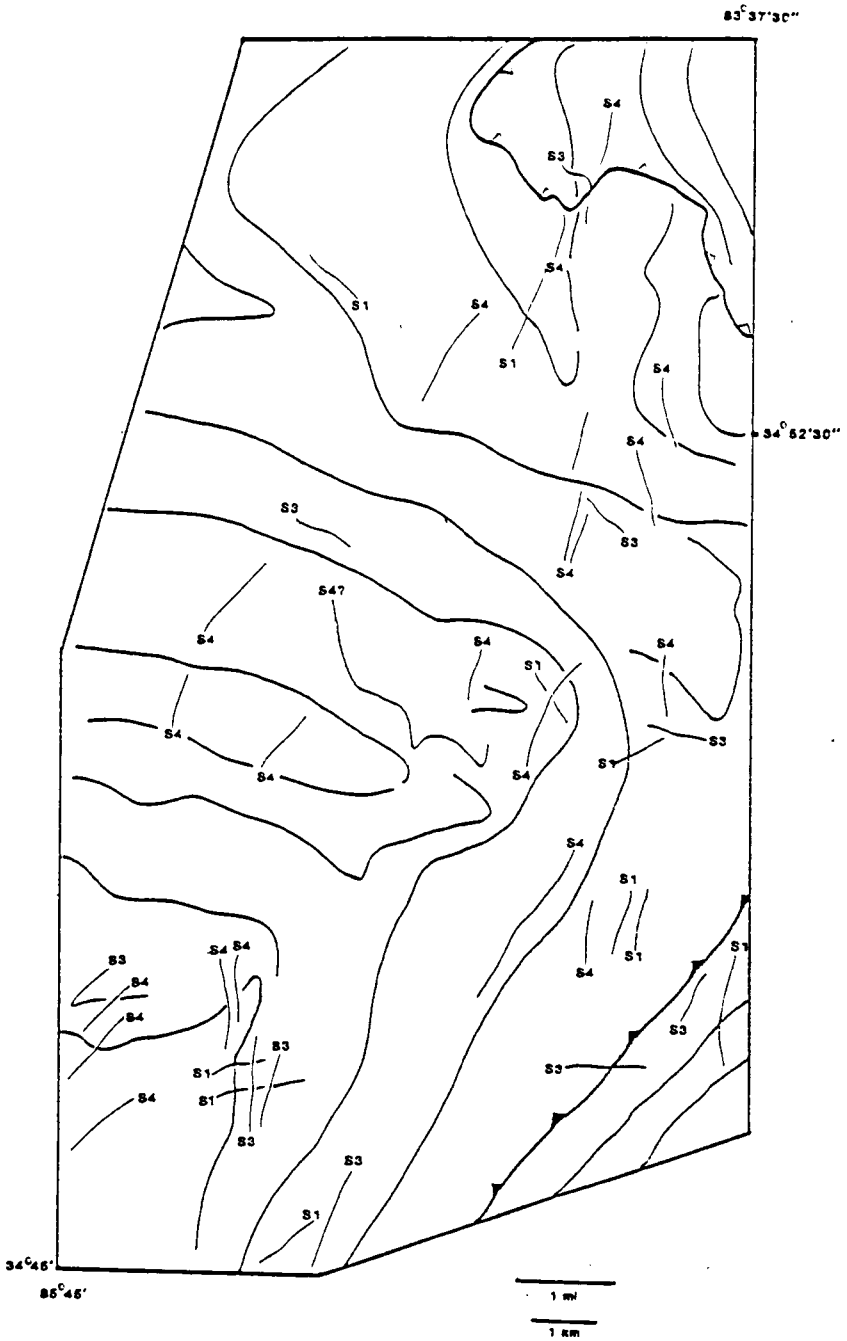


Figure 19: Orientation and observed distribution of foliations in the study area. Heavy, unlabeled lines represent the general form lines of the dominant foliation, S_2. Lighter, labeled lines represent subordinate foliations. Heavy line with solid teeth = Soque River fault. Heavy line with open teeth = Shope Fork fault.

parallel biotite laminae in the otherwise homogeneous metasandstone. Although often parallel to the dominant foliation, the laminations are also commonly oblique to, and overprinted by, S_2. Alternative interpretations can be proposed for their origin; the laminations may represent relict cross-bedding or other sedimentary structures, or they may be the product of transposition of an earlier surface or bedding, as suggested by Hatcher (1979). In one thin section from an adjacent area (Hopson, in progress) the biotite that defines the pinstriping appears to have developed parallel to the axial surfaces of nearly-microscopic folds (Fig. 20), indicating a tectonic origin. Evidence exists in nearby areas within the eastern Blue Ridge for development of an early foliation prior to formation of the dominant surface (Hatcher, 1977). Using these lines of evidence, the pinstriping is designated S_1.

S_2: The dominant foliation throughout the study area, designated S_2, is compositional layering consisting of alternating quartzofeldspathic and micaceous layers. The thickness and degree of differentiation of these layers is variable. In metasandstones, the quartzofeldspathic layers are typically peppered with biotite, producing monotonous medium to dark grey layers of variable thickness, and segregated by thin biotite-rich laminae (Fig. 21). Differentiation is more pronounced in some rocks, resulting in well-defined gneissic banding. The dominant foliation resulted from transposition of pre-existing layering, based upon the presence of tight, rootless intrafolial folds preserved within the dominant foliation and of structures resembling crossbedding in some layers, that may actually be pseudo-crossbedding resulting from discontinuities developed parallel to layering which eventually cut the fold closures during transposition (Hobbs and others, 1976). Microscopically, S_2 is defined by layers of biotite with a preferred orientation, separating layers of primarily quartz and feldspar (Fig. 22). In

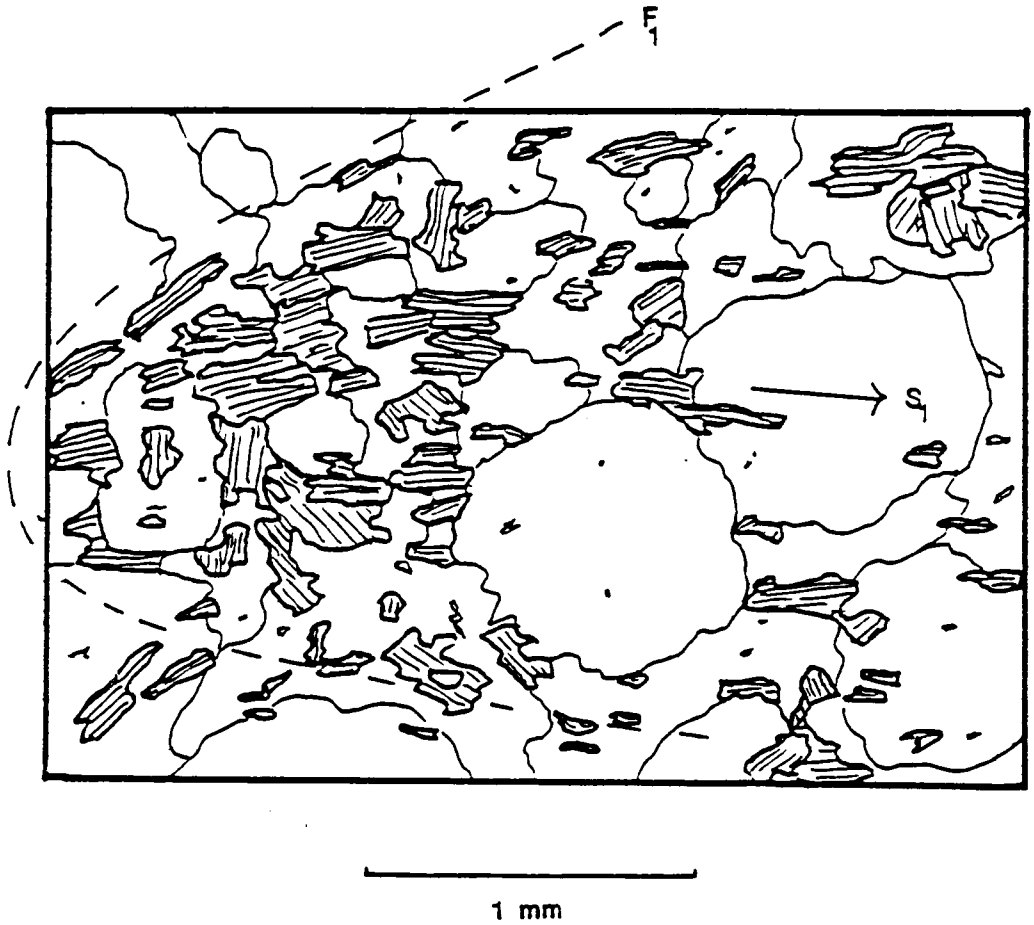


Figure 20: Biotite defining S_1 (pinstriping), developed as an axial planar foliation to F_1 fold. Sketch from photomicrograph. Sample from Coleman River Formation, northeast of study area.

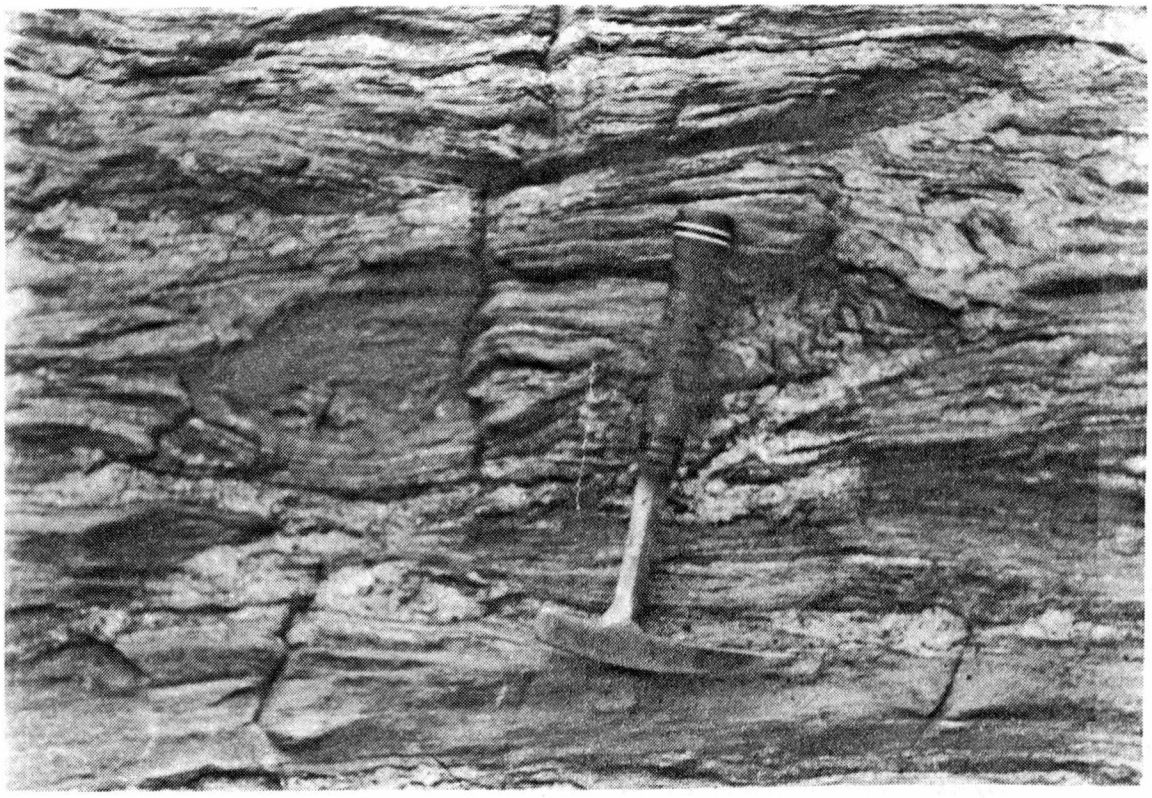


Figure 21: Typical metasandstone from the study area. Dominant foliation formed by quartzofeldspathic layers segregated by biotite laminae. (Coleman River Formation, Tray Mountain.)

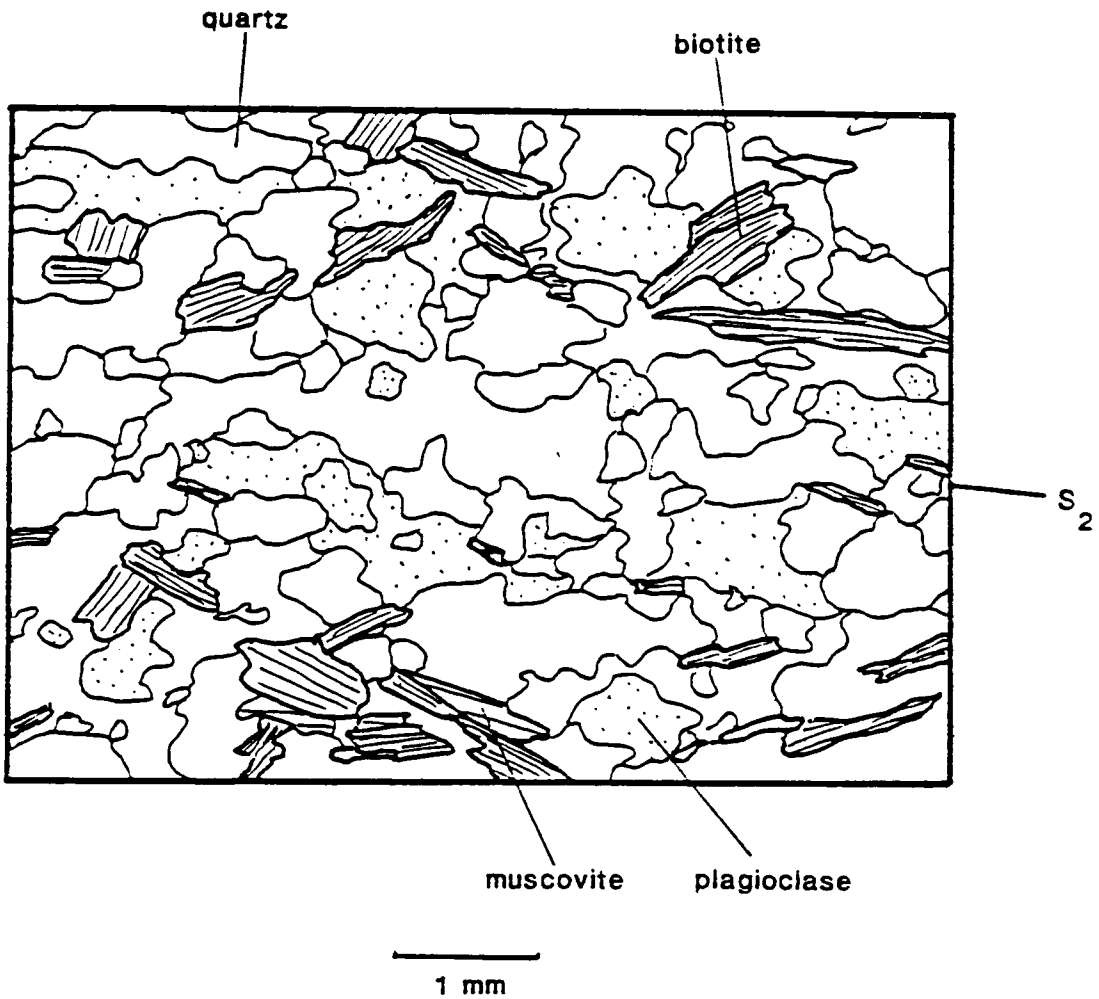


Figure 22: Typical S_2 foliation. Sketch from photomicrograph. Sample from Coleman River Formation, Chimney Mountain.

some samples, grains of muscovite and/or sillimanite exhibit the same orientation as biotite.

S_3: A locally penetrative foliation, designated S_3, is best developed in the more pelitic rocks of the area. It is a schistosity defined by micas and sometimes fibrolitic sillimanite, aligned parallel to axial planes of tight F_3 folds that fold the dominant layering. The schistosity cross-cuts S_2 within the noses of these folds, but becomes parallel or subparallel to S_2 within the limbs (Fig. 23). A photomicrograph from the same location as Figure 28 shows parallel biotite and fibrolitic sillimanite cross-cutting biotite that defines S_2 (Fig. 24).

Although sillimanite occurs parallel to both the dominant foliation and S_3, the possibility exists that sillimanite growth took place only during formation of S_3. If so, S_2 has been transposed into parallelism with this surface in most cases, or sillimanite mimetically replaced other minerals (muscovite?) within S_2. Alternatively, both surfaces were formed at sillimanite-grade conditions, with S_3 developed only locally.

S_4: Grains of muscovite cross-cutting the dominant foliation at a high angle were observed in several thin sections (Fig. 25). This fabric could not be positively correlated with S_3, which more commonly contains biotite; therefore, it was tentatively designated S_4. On the outcrop scale, a weak schistosity or a poorly developed spaced cleavage was occasionally observed, and a number of pegmatite bodies in the area contain a weak foliation oblique to that in the surrounding metasedimentary rocks. The orientation of

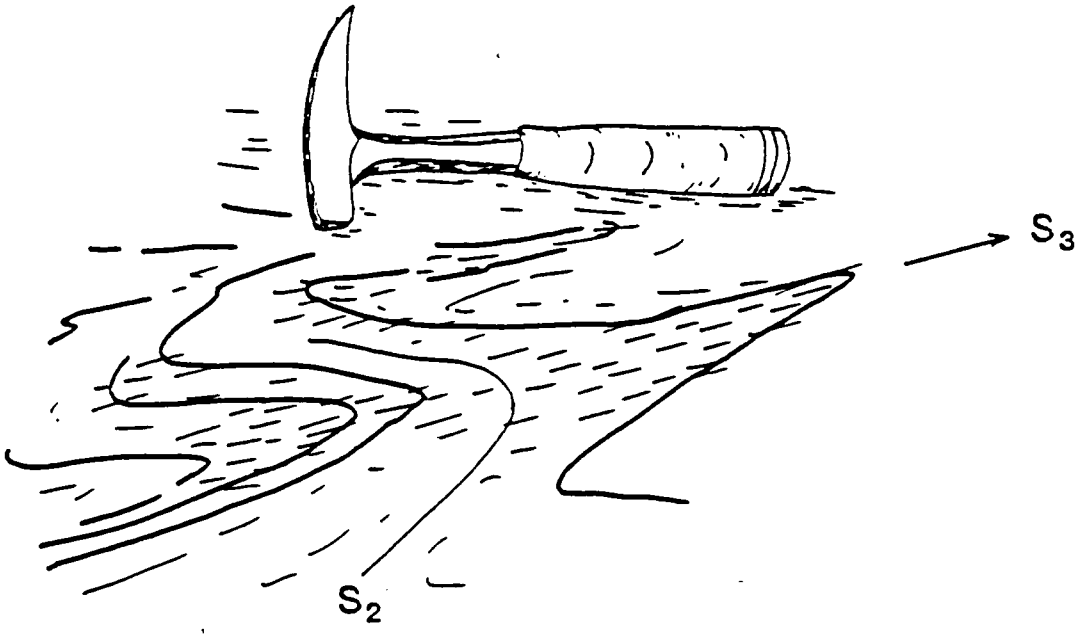


Figure 23: Development of axial planar schistosity S_3, within F_3 folds. Outcrop located near Indian Grave Gap, Tray Mountain quadrangle. View approximately due west. (Sketch from photo.)

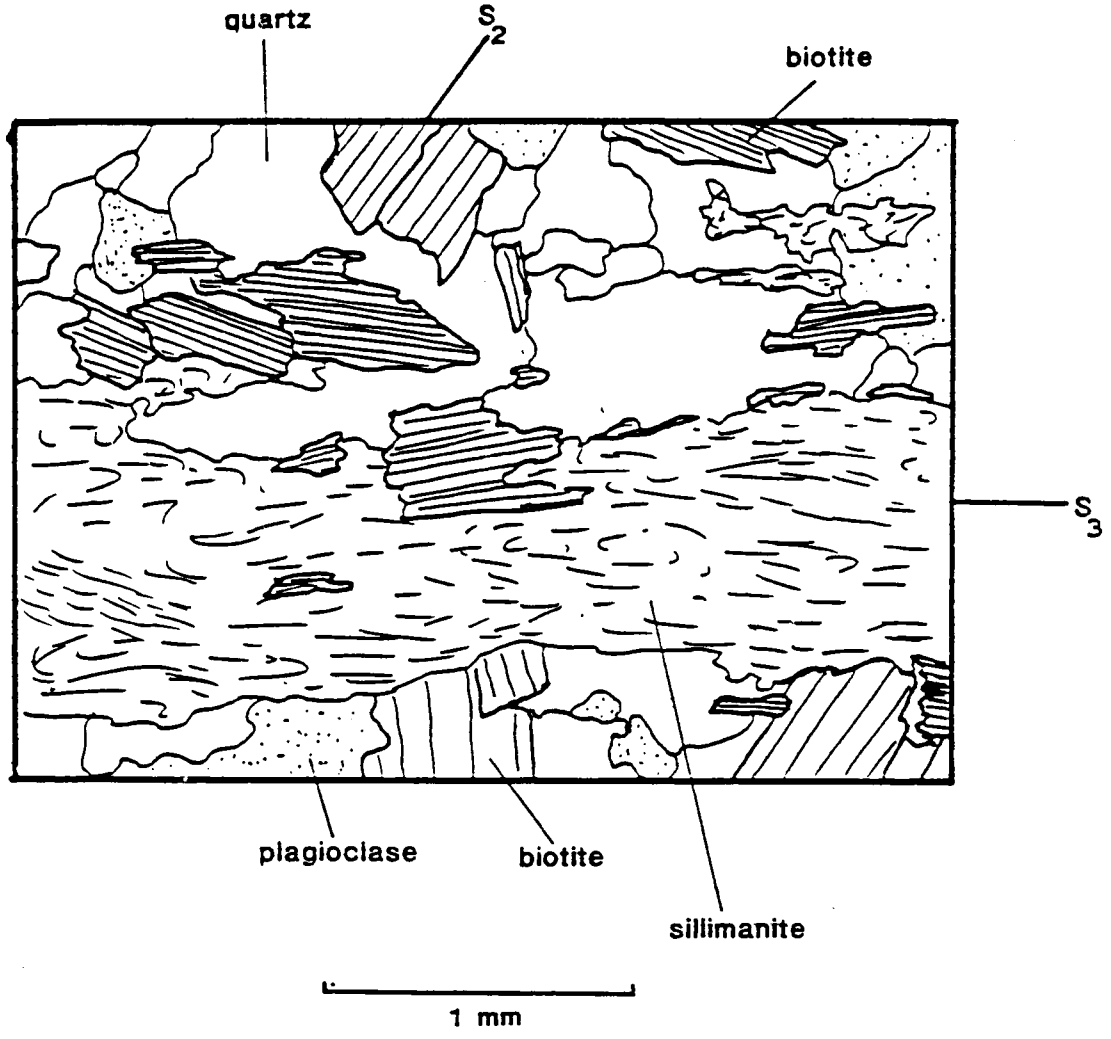


Figure 24: S_3 foliation defined by biotite and fibrolitic sillimanite. Sketch from photomicrograph; sample from the Ridgepole Mountain Formation. Location same as Figure 23.

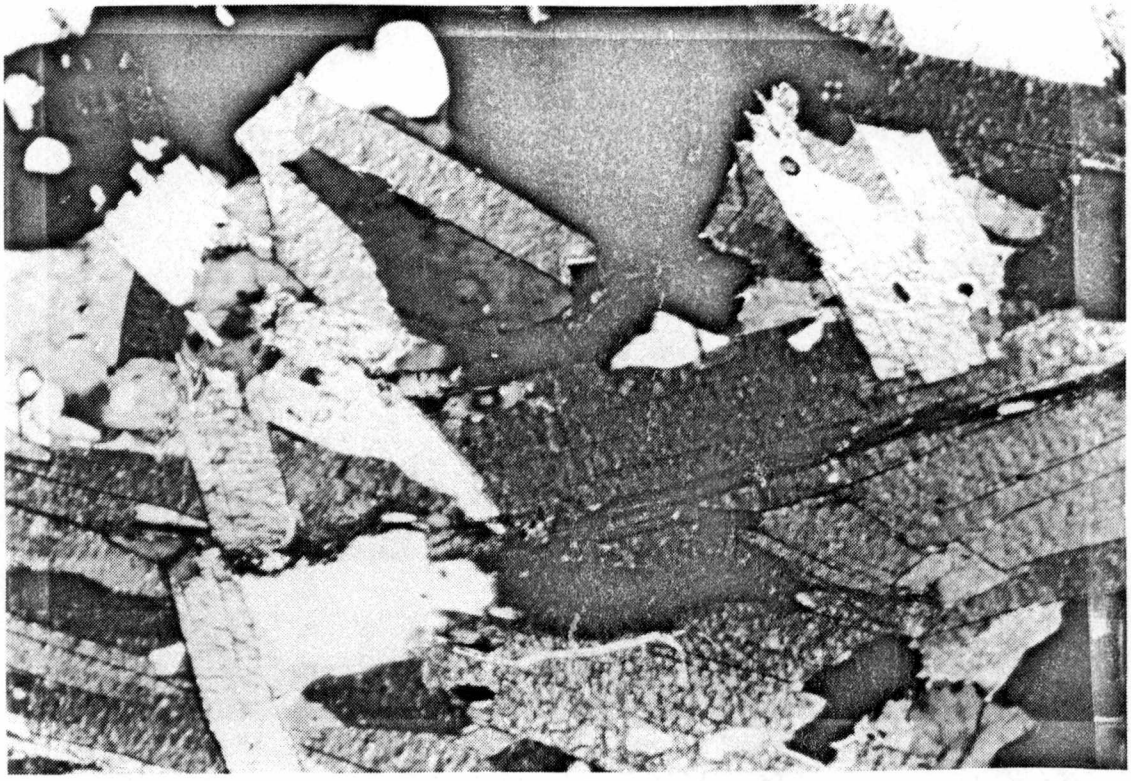


Figure 25: Photomicrograph of pelitic rock from the Coleman River Formation. Biotite defines dominant foliation. Muscovite cross-cuts this at a high angle. Scale bar is 1 mm.

these structures commonly parallels that of macroscopic F_4 folds (page 115). The mineralogical relationship shown in Figure 25 is tentatively interpreted as the microscopic manifestation of these structures.

Mesoscopic Fold Styles

A combination of fold classifications will be used to describe mesoscopic folds in the study area. By using fold shape vs. fold amplitude, Hudleston (1973) devised the shape classification scheme illustrated in Fig. 26. A fold may be designated by a letter and a number corresponding to its morphology. Dip isogons (Elliott, 1965) are lines that connect points of equal slope on adjacent surfaces of a folded layer. In the classification of Ramsay (1967), folds may be classified according to the pattern of convergence or divergence of dip isogons, which compares the variations in slope of adjacent surfaces in a fold. This classification is shown in Figure 27. Dip isogons were constructed for selected folds from the study area.

The geometry of a fold may be a reflection of the mechanism(s) that produced it (Donath and Parker, 1964). Mechanisms that have been suggested to be important in fold formation include passive flow or slip and flexural flow or slip (Donath and Parker, 1964), buckling, and bending. Ideal passive flow folds should exhibit a similar geometry and correspond to Ramsay's Class 2 folds (Fig. 27), while folds formed by both buckling and flattening generally fall within Class 1C, or a combination of classes (Ramsay, 1967). Complex and non-ideal geometries may be due to irregular thickness of a folded layer or to flattening oblique to the axial surface trace (Hudleston, 1973). Donath and Parker (1964) recognized that, in real rocks, all gradations exist between ideal examples of passive to flexural folds. Combined fold mechanisms may operate

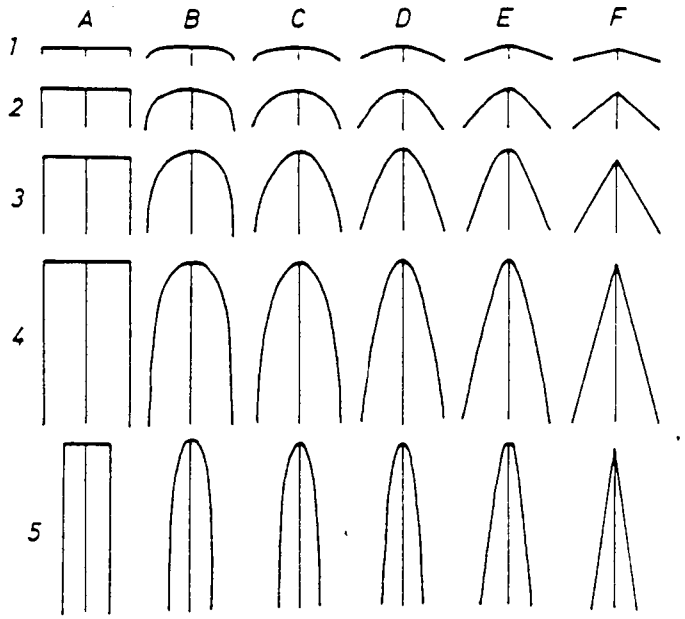


Figure 26: Fold classification scheme based on shape (A-F) and amplification (1-5) (from Ramsay and Huber, 1987; after Hudleston, 1973.)

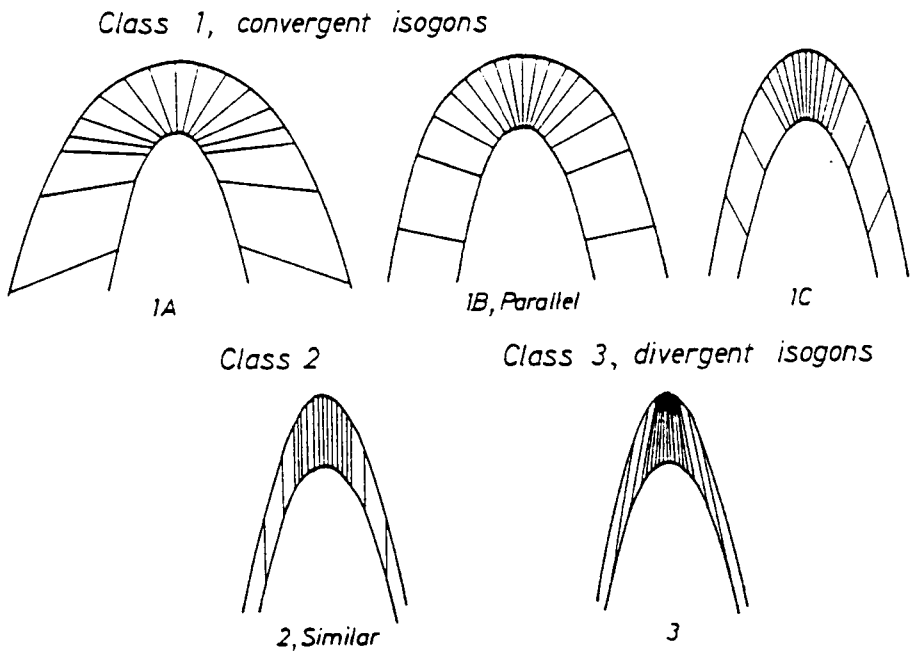


Figure 27: Dip isogon characteristics of main fold classes (after Ramsay, 1967).

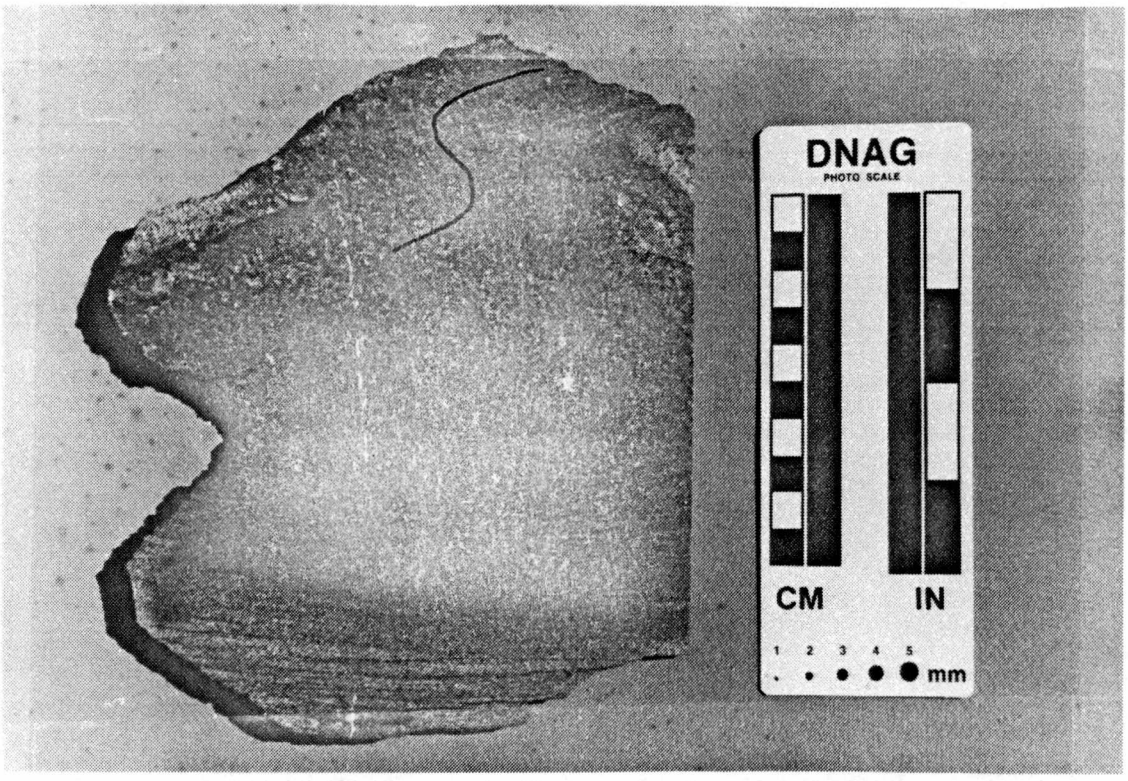
during the folding of a layered sequence to produce non-ideal types (Hatcher, in press), but one mechanism will likely control fold geometry (Donath and Parker, 1964).

Folds in the study area can be divided into three basic styles:

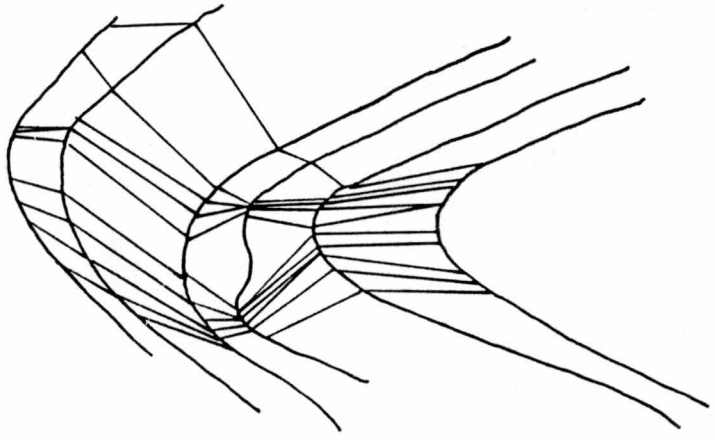
Type 1: Tight to isoclinal (D4, E4 or E5 of Fig. 26) folds; recumbent or reclined in orientation, with horizontal to inclined axial surfaces. Isogon analysis of two folds (Figs. 28 and 29) indicates a combination of Classes 1C, 2, and 3. Passive flow was probably the dominant mechanism, although flexural flow and buckling are also evident. This may be due to the presence of high ductility contrast between layers, in which case passive flow may not be possible even in a state of high mean ductility (Donath and Parker, 1964).

Type 2: Upright to overturned (rarely recumbent) folds; shapes vary from open and parabolic to chevron or cuspate (D2, D3, F3 or F4 of Fig. 26), and are typically asymmetric. Isogon analysis of two examples (Figs. 30 and 31) again indicates a combination of classes ranging from 1A to 1C to 3; note, however, that the synform in Figure 31 approaches an ideal Class 2 fold. Flexural flow with a buckling component was dominant, but passive flow was operative in some instances.

Type 3: Upright to inclined, rarely recumbent flexural folds (C2, C3, D2 of Fig. 26) with horizontal to gently plunging axes. These folds are

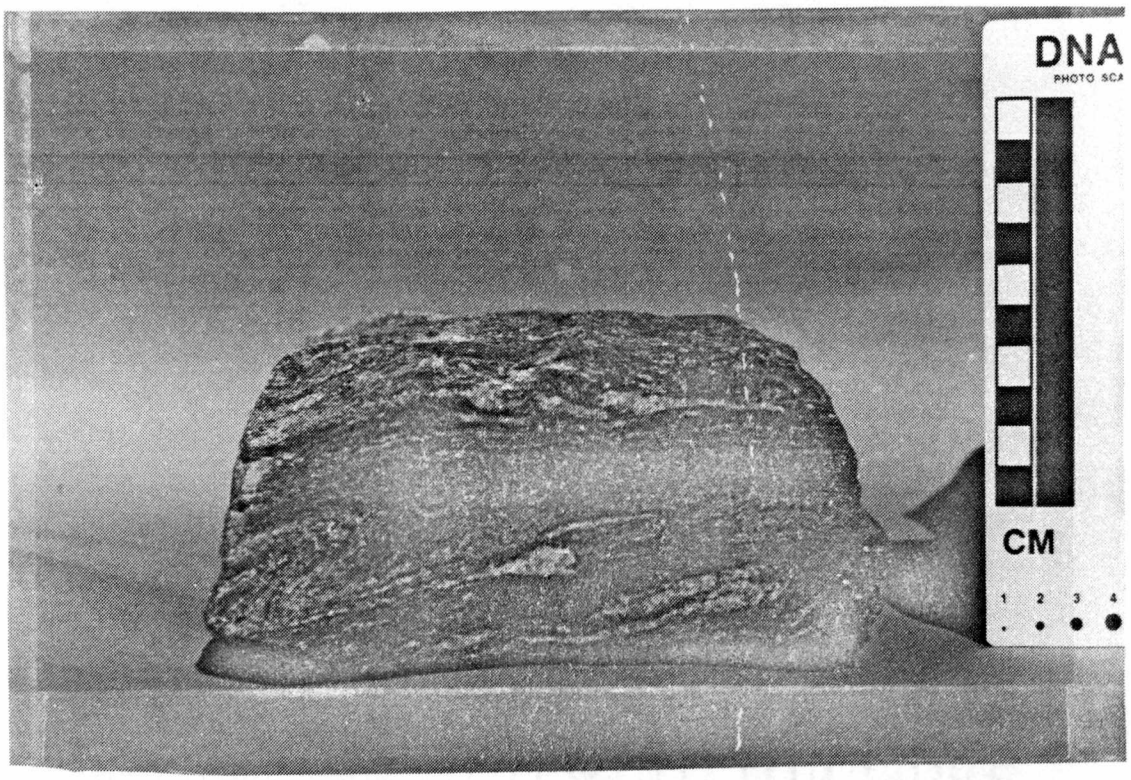


(a) Type 1 fold from Tallulah Falls Formation, Dismal Gap (Macedonia quadrangle).

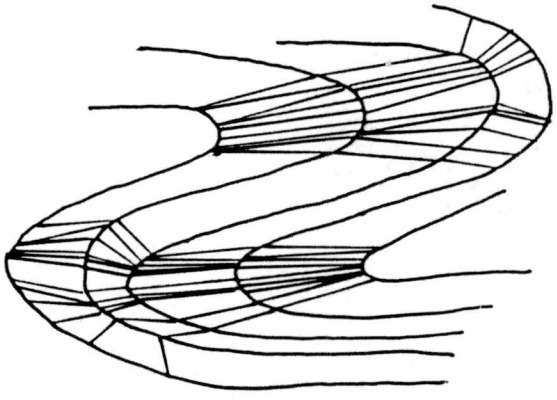


(b) Dip isogon analysis of same fold.

Figure 28: Dip isogon analysis of type 1 fold, Tallulah Falls Formation.

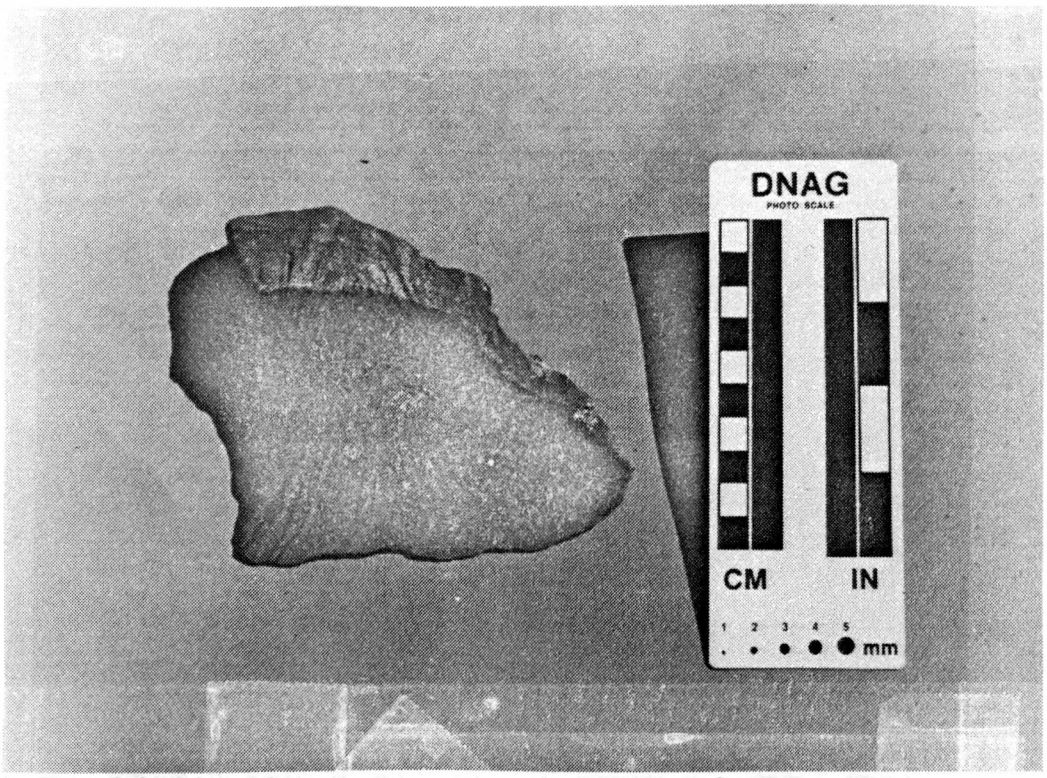


(a) Type 1 fold from Otto Formation, Wildcat Creek Road (just east of study area).

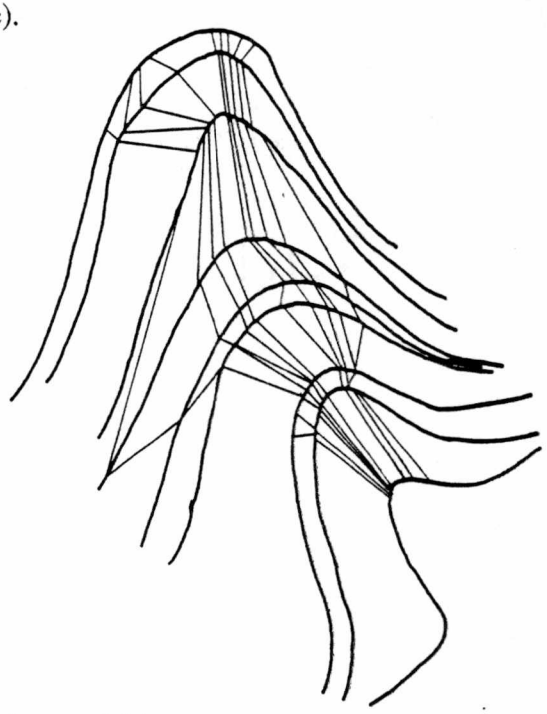


(b) Dip isogon analysis of same fold.

Figure 29: Dip isogon analysis of type a fold, Otto Formation.

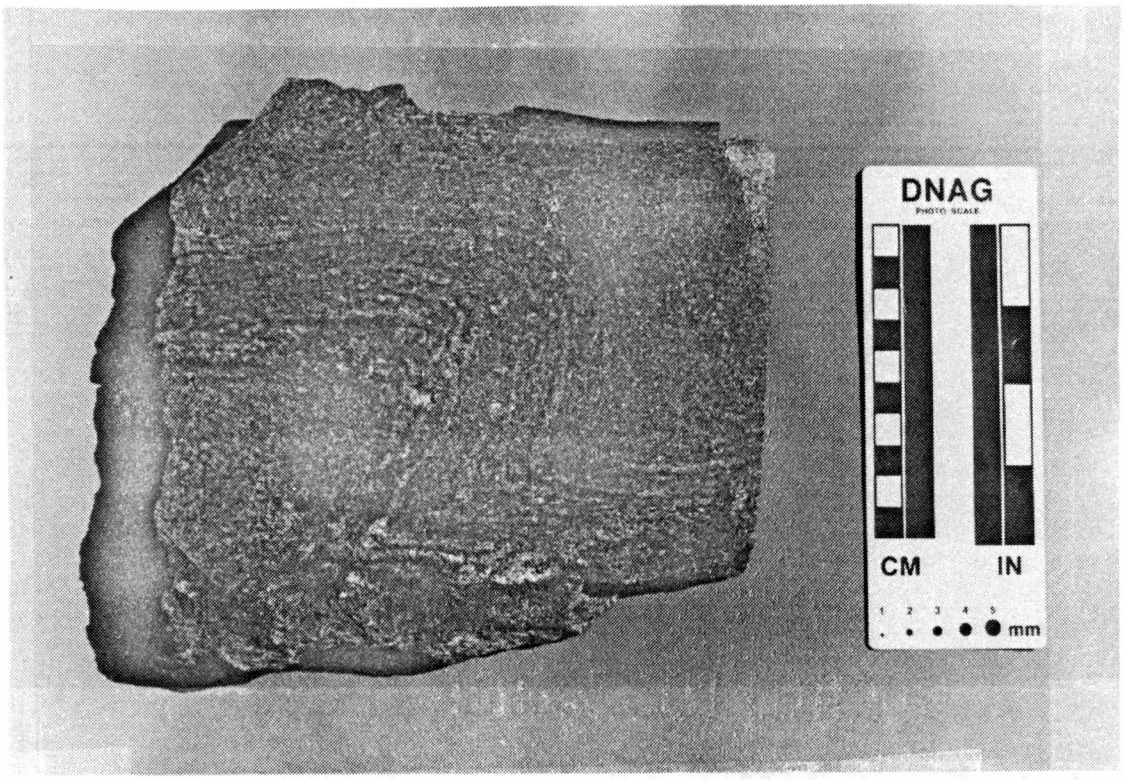


(a) Type 2 fold from Coleman River Formation, Wildcat Creek Road (Tray Mountain quadrangle).

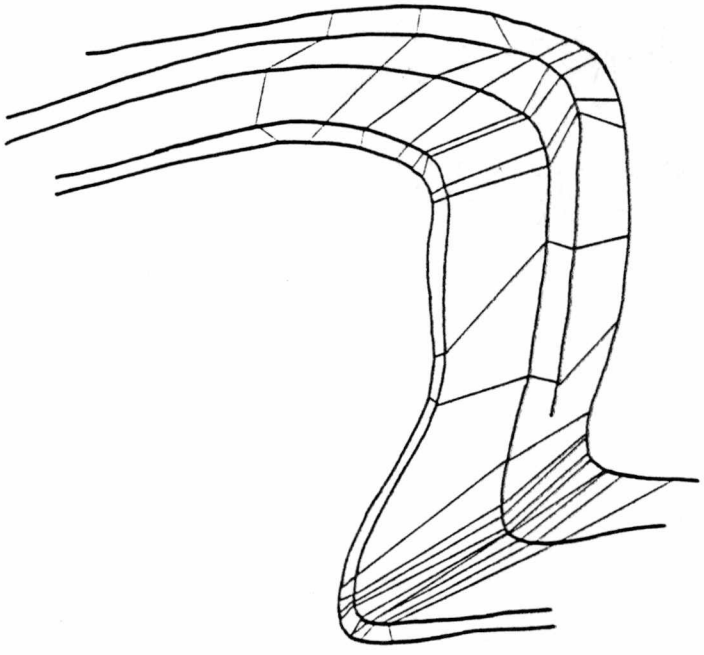


(b) Dip isogon analysis of same fold.

Figure 30: Dip isogon analysis of type 2 fold, Coleman River Formation.



(a) Type 2 fold from Coleman River Formation, Tray Mountain.



(b) Dip isogon analysis of same fold.

Figure 31: Dip isogon analysis of type 2 fold, Coleman River Formation, Tray Mountain.

most easily recognized in thickly-layered metasandstone with relatively low mean ductility. Widely spaced, fanning fractures are commonly developed in psammitic layers (Fig. 32), and the thickness of these layers remains relatively constant, with a minor amount of thickening in hinge zones. However, where metasandstone occurs with interlayered schist, the schistose layers behave more ductilely, indicating that flexural flow may be the overall mechanism for formation of these folds.

In addition, two less common fold types occur. These are small-scale to microscopic kink folds, observed only locally in schistose layers; and ptygmatic folds, often polyclinal, developed in isolated quartzofeldspathic veins (Fig. 33). Ptygmatic folds are generally Class 1A to 1B (Fig. 34), and formed by buckling of isolated competent layers within a matrix of much lower competence (Ramsay and Huber, 1987).

Fold Generations

Each of the three basic styles described above may form during more than one fold generation. Fold style and orientation should be used with caution in deducing fold generations (Williams, 1970), but style remains the best single characteristic for correlation of structures (Turner and Weiss, 1963; Williams, 1985). The combination of style and orientation provide usesful criteria for correlation in the absence of clear overprinting relationships on the outcrop scale.

As many as six generations of folding have been recognized, the characteristics of which are summarized in Table 7. F_1 folds are inferred from the presence of the

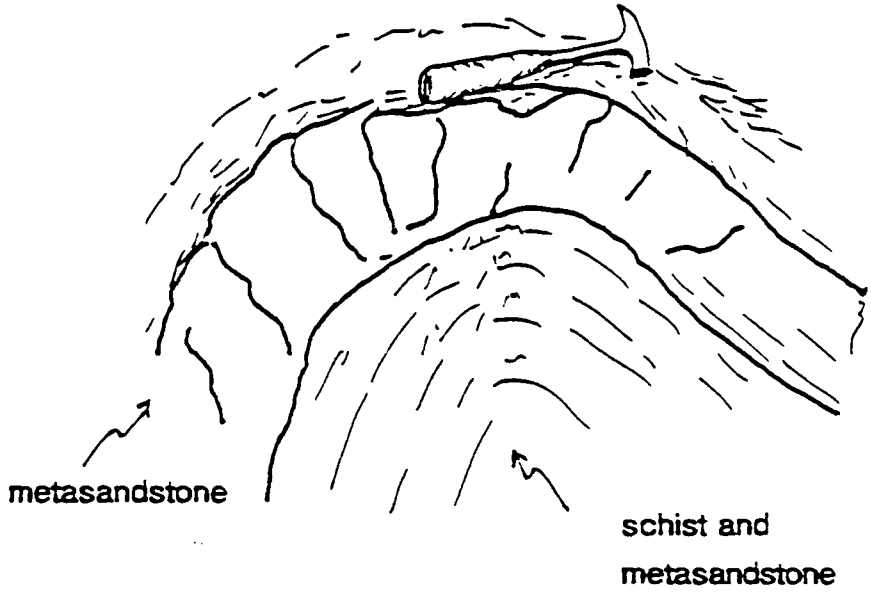


Figure 32: Flexural fold with fanning fractures. In metasandstone of Tallulah Falls Formation, York Ridge (N Tray Mountain quadrangle). Sketch from photo.

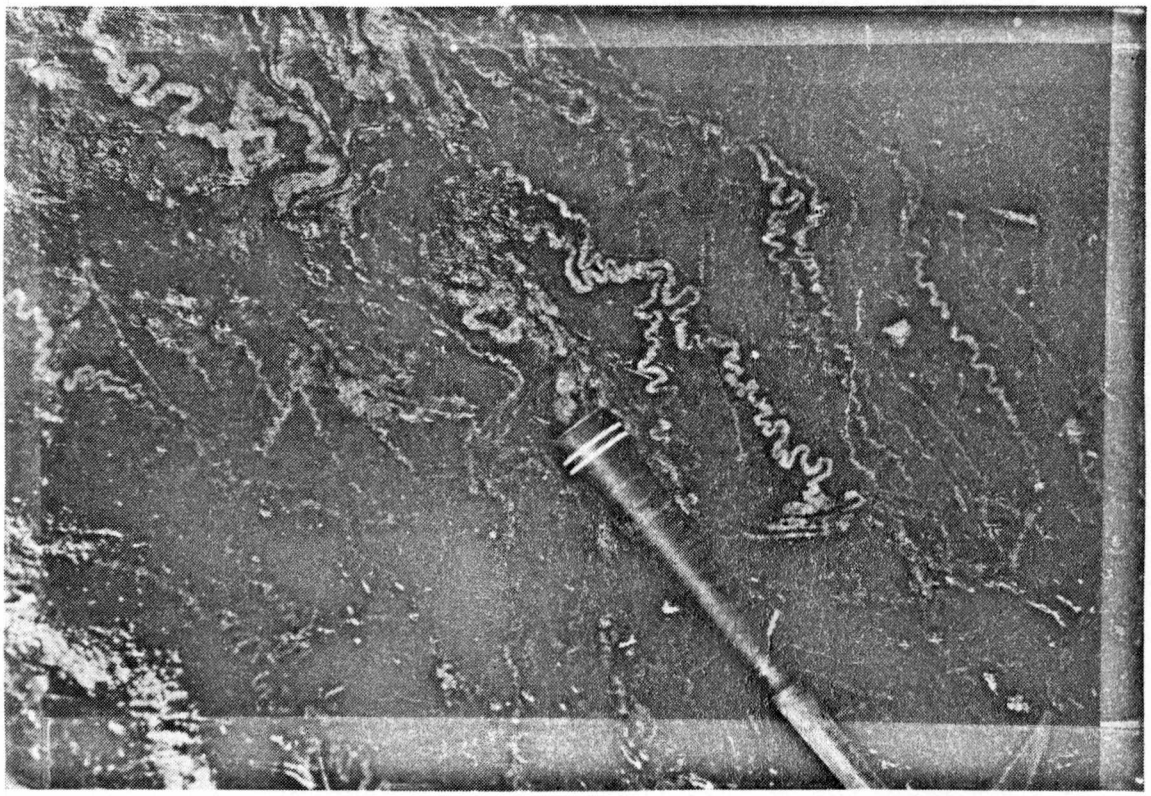
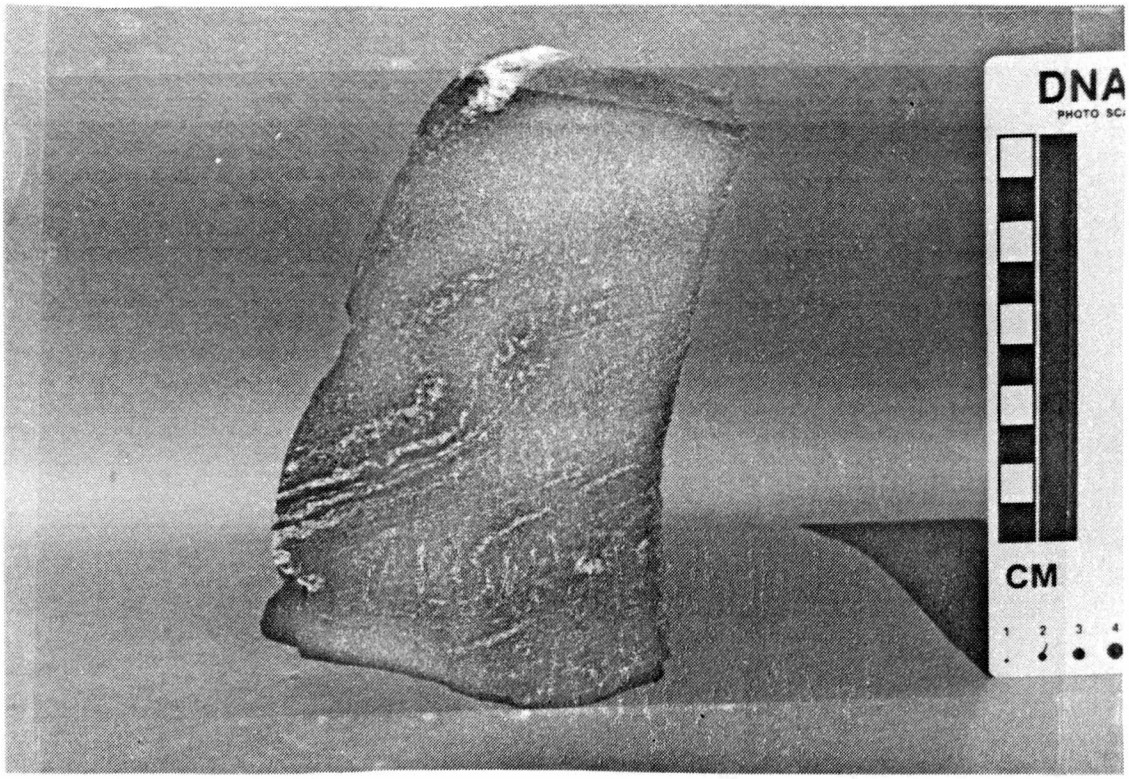
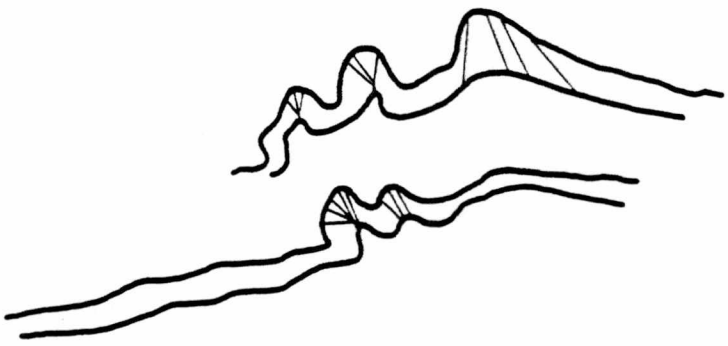


Figure 33: Ptygmatic folds in quartzofeldspathic veins (Tallulah Falls Formation, Tray Mountain quadrangle).



(a) Ptygmatic folds in hand sample of Coleman River metasandstone.



(b) Dip isogon analysis of same folds.

Figure 34: Dip isogon analysis of ptygmatic folds.

TABLE 7: Fold generations and associated foliations in the study area.

Mesoscopic Folds		Foliations		
Designation	Style	Orientation	Designation	Description
F_6	Open, upright to recumbent (C2, C3, D2).	NW	None	
F_5	Open to chevron (D2, D3, F3, F4); upright to overturned.	NW to WNW	None	
F_4	Tight to open, upright (D3, D4); also small-scale kink folds (F2).	NNE	S_4	Localized schistosity, weak foliation in pegmatite bodies.
F_3	Tight to isoclinal, recumbent (D5, E4, E5).	NE	S_3	Localized axial planar schistosity.
F_2	Isoclinal, recumbent (D5 of Fig. 29)	E-W?	S_2	Alternating quartzo-feldspathic and micaceous layers; gneissic banding.
F_1	(Not observed)	?	S_1	"Pinstriped" laminae

pinstriped laminations, and can be observed on the microscopic scale in at least one sample (Fig. 20, pg. 96). F_2 folds are isoclinal folds which resulted in the transposition of S_1 and formation of S_2. These folds are rarely preserved, but are occasionally observed as rootless isoclines within the dominant foliation (S_2). F_3 folds are commonly observed as tight to isoclinal folds that fold S_2. Locally, an axial planar foliation (S_3) is developed (see pg. 99), but many folds occur where no new foliation is evident. The average orientation of F_3 folds appears to be northeast, although overall this is quite variable, likely due to later overprinting.

The effects of F_4 folding are best observed in the southwest portion of the field area. Mesoscopic folds are difficult to recognize. A weak foliation or cleavage is associated with this event (page 99). The major trend of F_4 is north to north-northeast. Localized kink folds in schistose layers may belong to this fold generation, because their axial surfaces generally display this orientation (Fig. 35), and in some cases they were deformed later.

F_5 folds produce a strong mesoscopic and macroscopic overprint over a substantial portion of the field area. These upright, open asymmetric folds or chevron folds commonly refold F_3 isoclines into Type 2 interference patterns (Ramsay, 1967) (Fig. 36). No axial planar foliation was identified to be associated with these folds. The dominant orientation of F_5 is northwest to west-northwest, and in some areas this trend appears to have completely reoriented many earlier structures (see page 129).

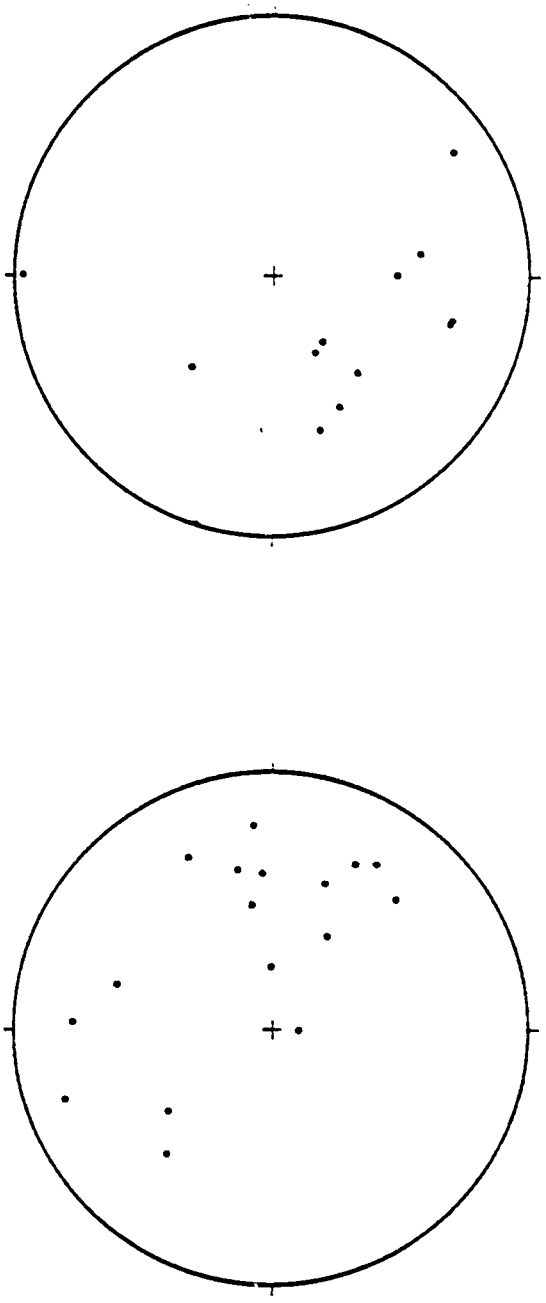


Figure 35: Point diagram of (a.) poles to axial surfaces and (b.) axes of small kink folds within the study area.

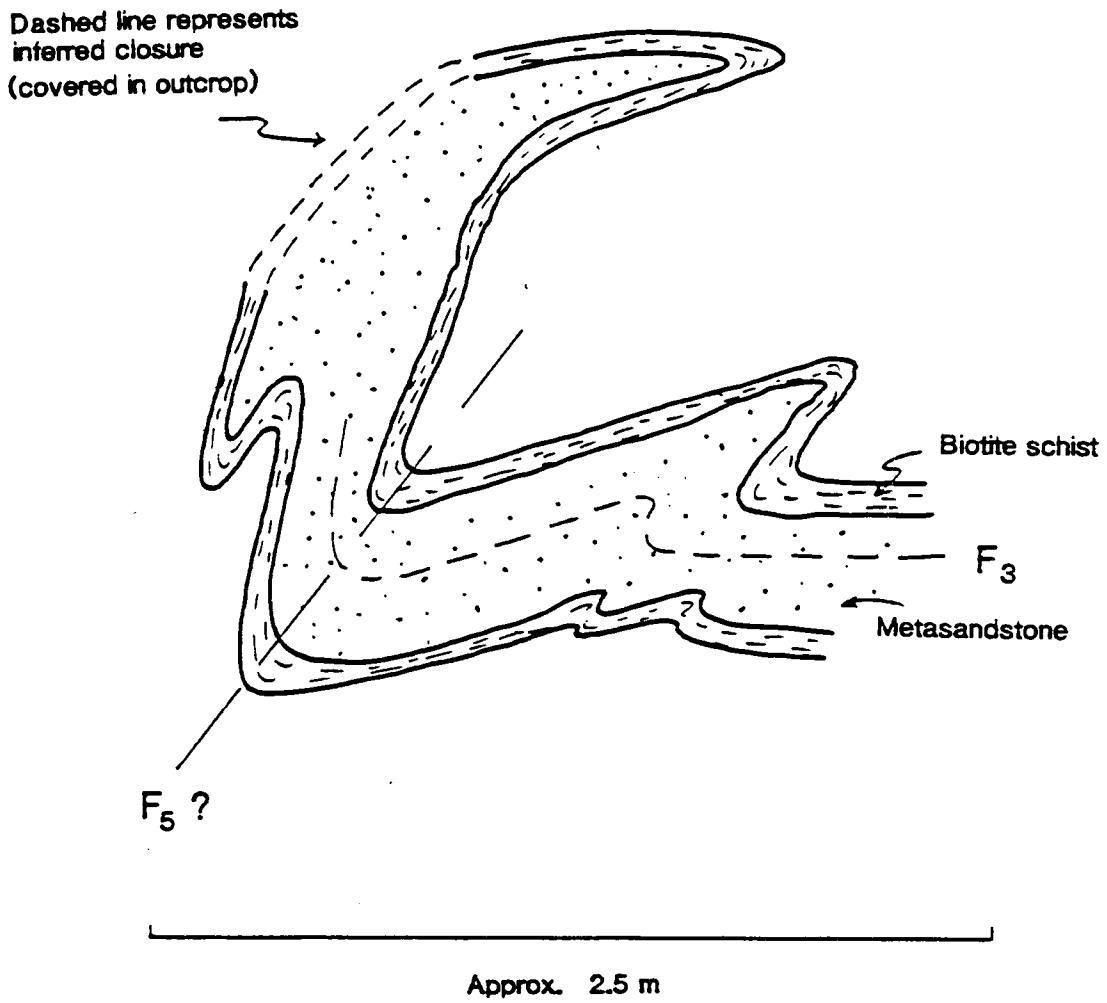


Figure 36: Type 2 interference pattern resulting from F_3 folds refolded by F_5. Outcrop located near Alice Branch, Tray Mountain quadrangle. View looking N70W. (Field sketch.)

The youngest folding episode (F_6) is represented by open, upright to recumbent flexural flow to flexural slip folds with no axial planar foliation. Mesoscopic F_6 folds are sporadically distributed over the field area and are concentrated in particular domains (page 124), although this event resulted in broad northwest-oriented warping throughout much of the area.

Lineations

Significant lineations other than fold axes are not a prominent feature in the study area. Mineral lineations defined by aggregates of mica $\pm$ sillimanite are the most common, and typically parallel fold axes. Intersection lineations of layering with axial surfaces of kink folds (F_4 ?) were observed, but only rarely. A fabric diagram of primarily mineral lineations (Fig. 37) shows a wide scatter over the northwest and southwest quadrants.

Dikes and Veins

Thin quartzofeldspathic dikes and quartz veins are numerous, and occur in a vast array of orientations. One set of dikes, however, appears to dominate and cross-cut earlier sets, commonly producing small-scale (up to approximately 15 cm) offsets (Fig. 38). The geometry is similar to that for brittle-ductile shears, as described by Ramsay and Huber (1987). The orientation of these structures is consistently north-northeast.

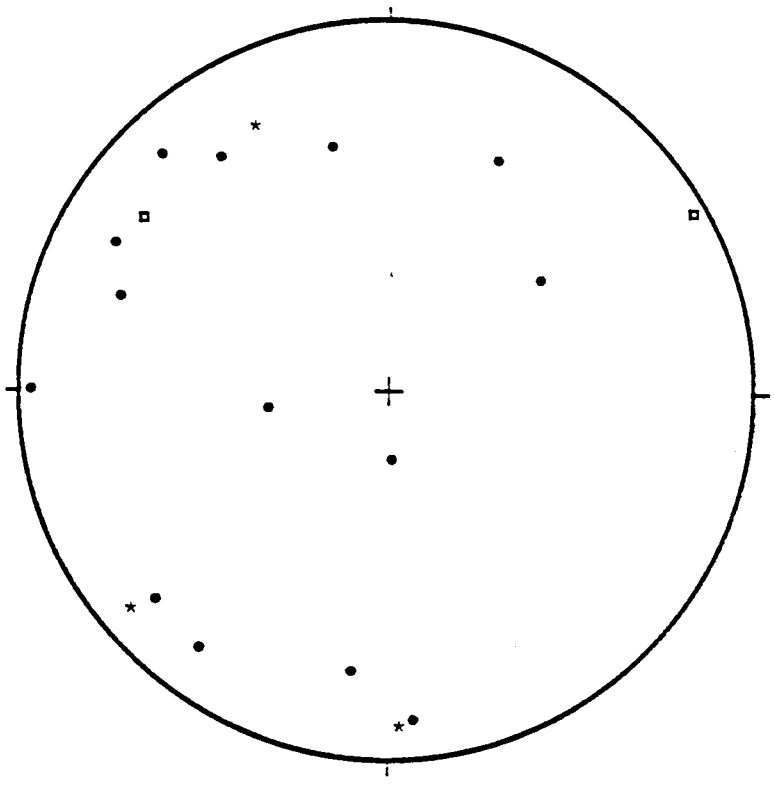


Figure 37: Point diagram of mineral lineations (dots), intersection lineations (squares), and sillimanite fibers (stars) from study area.

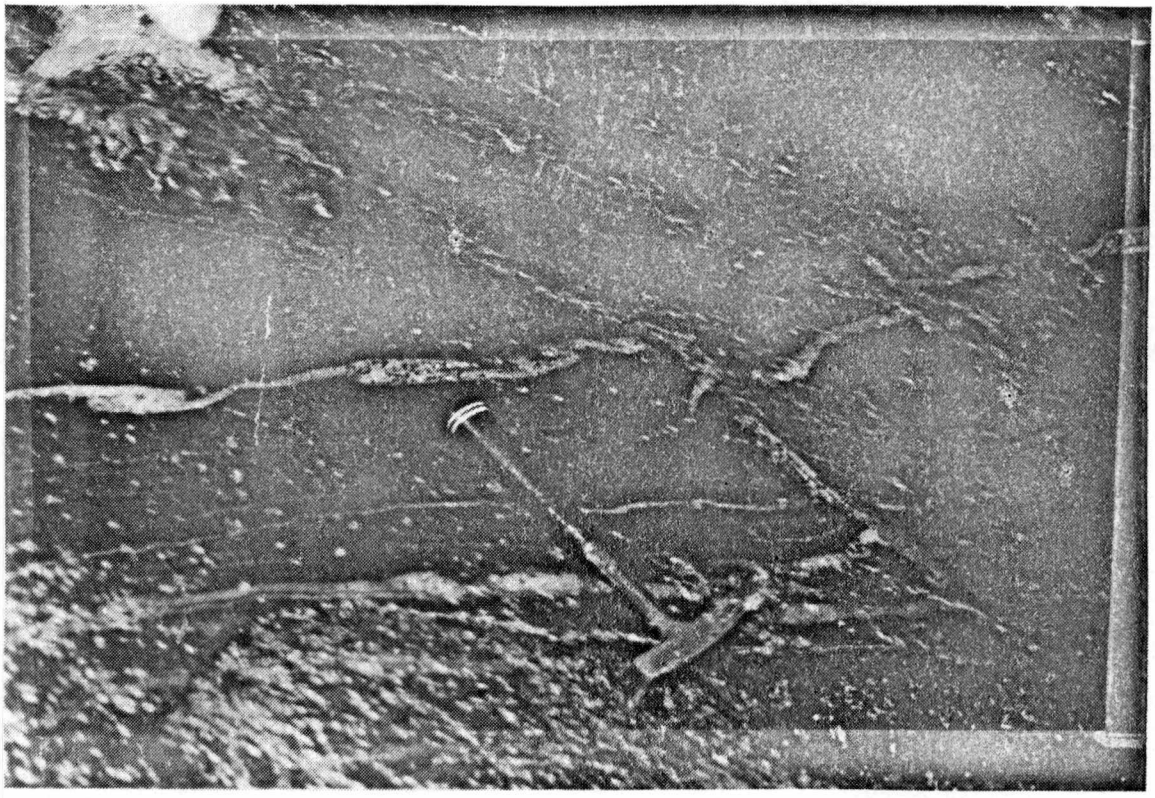


Figure 38: Quartzofeldspathic dikes, earlier ones offset by later set. Outcrop in Sassafra Branch, Tray Mountain quadrangle. Hammer handle points due north.

Joints

Joints are fairly common in metasandstones of the field area. These are open fractures, generally spaced at 20 to 60 cm intervals. Two joint sets are indicated by Figure 39; a prominant set striking approximately N45E, and a less conspicuous set striking N35-40W. Dip is generally 90^0 or nearly so, regardless of the dip of foliation in surrounding rocks. Joints are the youngest structural feature recognized in the study area, although it could not be determined whether one set is younger than the other because the two sets were not observed together in the same outcrop. The structures are probably analogous to vertical unloading joints (Engelder, 1985) that form due to thermal-elastic contraction during erosion and uplift.

Macroscopic Structures

Macroscopic Folds

A number of map-scale folds occur within the study area (Fig. 40). The Ridgepole Mountain Formation serves as a marker unit from which early structures (F_3?) can be deduced. The orientation of these folds is variable because of later overprinting, although the mean orientation appears to be northeast to east-west. Macroscopic F_4 folds are observed in the southwestern and central portions of the area, where early structures are refolded around north-northeast to due north-trending axes. The "boomerang-shaped" outcrop pattern of the Ridgepole Mountain Formation in the southwest (Plate II) likely resulted from refolding of inclined to recumbent F_3 folds by more upright F_4 folds.

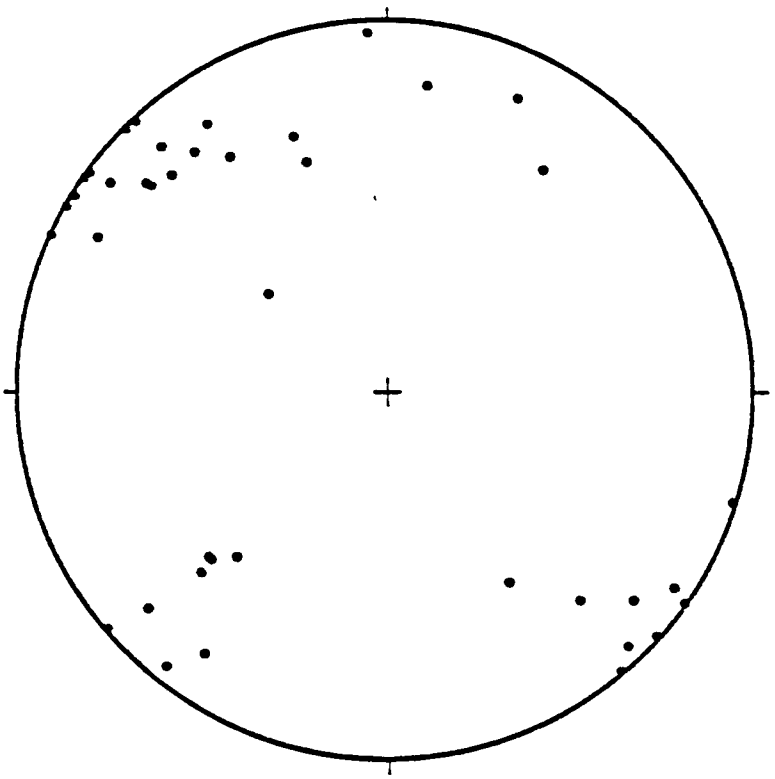


Figure 39: Point diagram of poles to joint surfaces from study area.

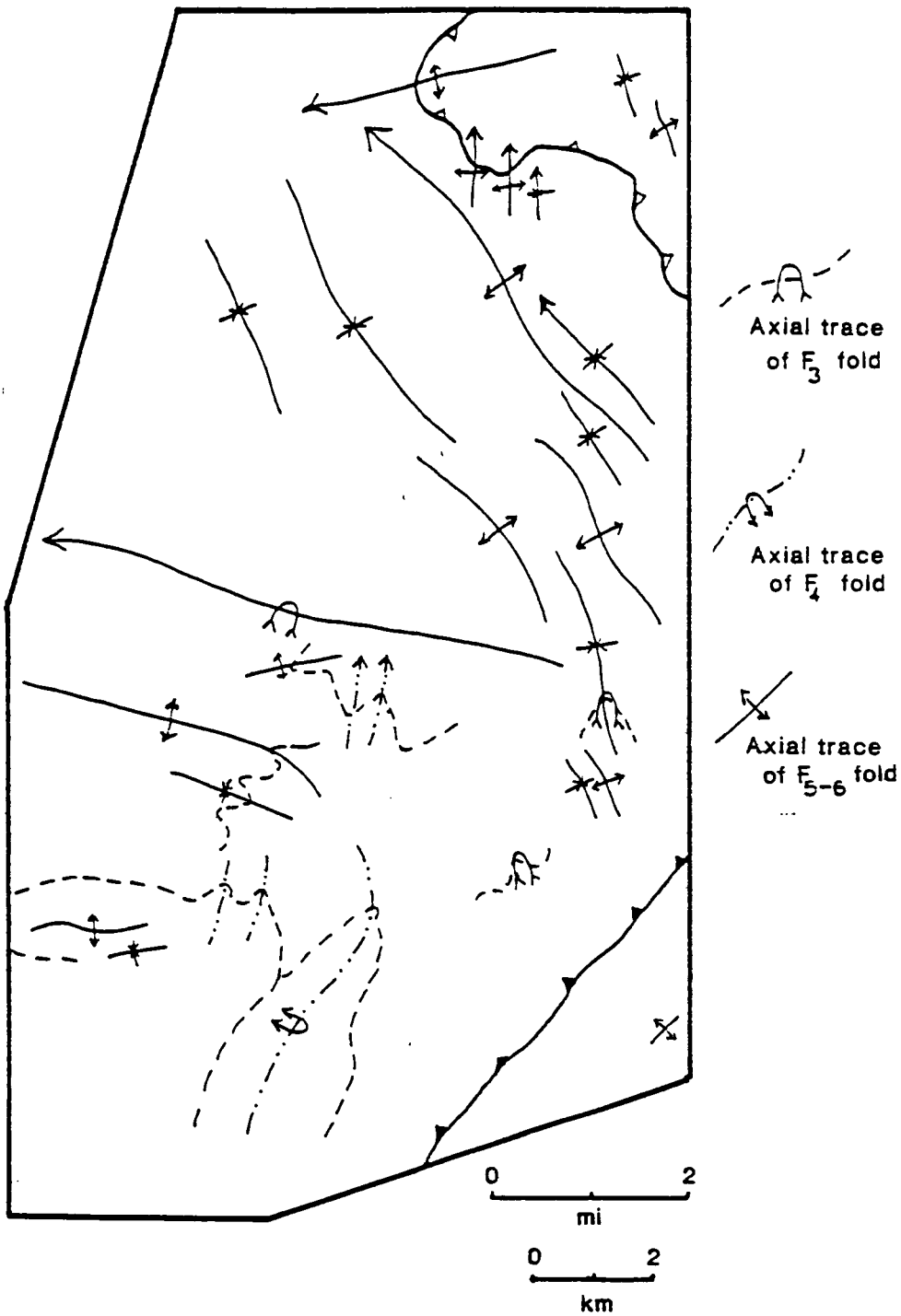


Figure 40: Distribution of map-scale folds within the study area. Heavy line with solid teeth = Soque River fault. Heavy line with open teeth = Shope Fork fault.

Late folds, tentatively correlated with the F_5 event, are steep, upright to slightly overturned structures that trend northwest to east-west. An extensive, slightly overturned synform, plunging to the west (Fig.40), appears to be the eastern hinge of the Chattahoochee basin, a large structural basin mapped in this and adjacent areas by Nelson (1985). The dominant foliation, S_2, wraps around the hinge of this basin (Fig. 19, pg. 94; Plate III), and upright, northwest-trending antiforms and synforms are abundant on the northern limb. The relationship of upright, north-trending folds that cross the Shope Fork fault (Fig. 40) to the other late folds is unclear.

Domain Analysis

The study area is structurally heterogeneous, and has been divided into five more or less homogeneous domains (Turner and Weiss, 1963) (Fig. 41) in order to more coherently describe the variety of fabrics present in the area. The domains are defined by segments of broadly uniform strike of the dominant foliation, and by orientation of predominant mesoscopic and macroscopic folds. Specific fold generations were not used to classify domains because in many cases clear cross-cutting relationships were absent, and many folds do not have an observable axial planar foliation; therefore, the generation of all folds could not accurately be deduced. For each domain, fabric diagrams were constructed for the dominant foliation and for axial surfaces and axes of the three types of mesoscopic folds described on page 105, briefly defined below:

Type 1: Tight to isoclinal, passive to flexural flow folds;

Type 2: Upright to overturned, dominantly flexural flow folds;

Type 3: Upright to inclined, flexural slip or flexural flow folds.

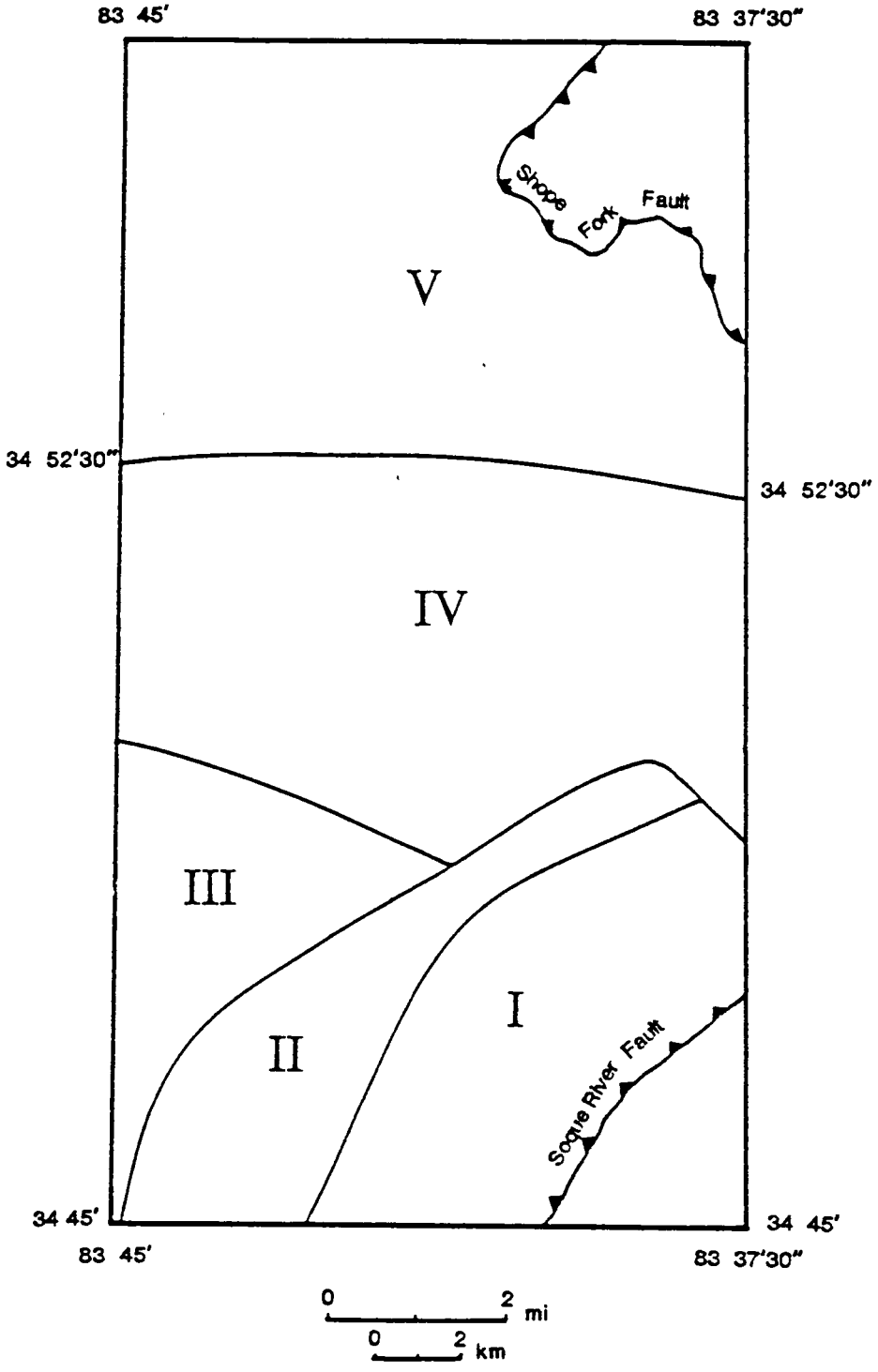


Figure 41: Map of domain boundaries for domain analysis of study area.

Domain I: Domain I occupies the southeast portion of the study area. The dominant foliation consistently strikes NE and dips moderately to steeply NW (Plates II and III). Poles to axial surfaces of Type 1 folds, plotted on an equal area net, are shown in Figure 42. Axial surfaces of these folds strike NE and dip shallowly to steeply NW. Axes (Fig. 42) generally plunge SW at a shallow angle, although a few which plunge northwest indicate the reclined orientation observed in some folds. Axial surfaces to Type 2 folds (Fig. 42) show considerable scatter. Type 3 folds are rare in Domain I. Only two of these were measured, both with steep axial surfaces (Fig. 42).

Domain II: A narrow strip designated Domain II represents a transition between a primarily NE strike of the dominant foliation in Domain I to a NW strike in other domains (Plate III). S_2 shows a maximum with a NNE strike, dipping west at a shallow to moderate angle. Axial surfaces of Type 1 folds strike northeast (Fig. 43), but a number are reclined with axes plunging northwest. Trends of Type 2 folds are variable from N to NW (Fig. 43). Again, few Type 3 folds were observed in this domain; those which were recognized have variable trends, and are inclined with shallowly-dipping axial surfaces (Fig. 43).

Domain III: Domain III lies in the southwestern portion of the study area. Strikes of the dominant foliation, S_2, are scattered (Plate II, Plate III), but dips are consistently shallow, apparently due to the location within the hinge of a macroscopic overturned synform, inferred to be an F_3 structure (Plate IV). The scattered strikes probably result from the tendency of variation to be produced by small pertubations in very shallowly dipping rocks.

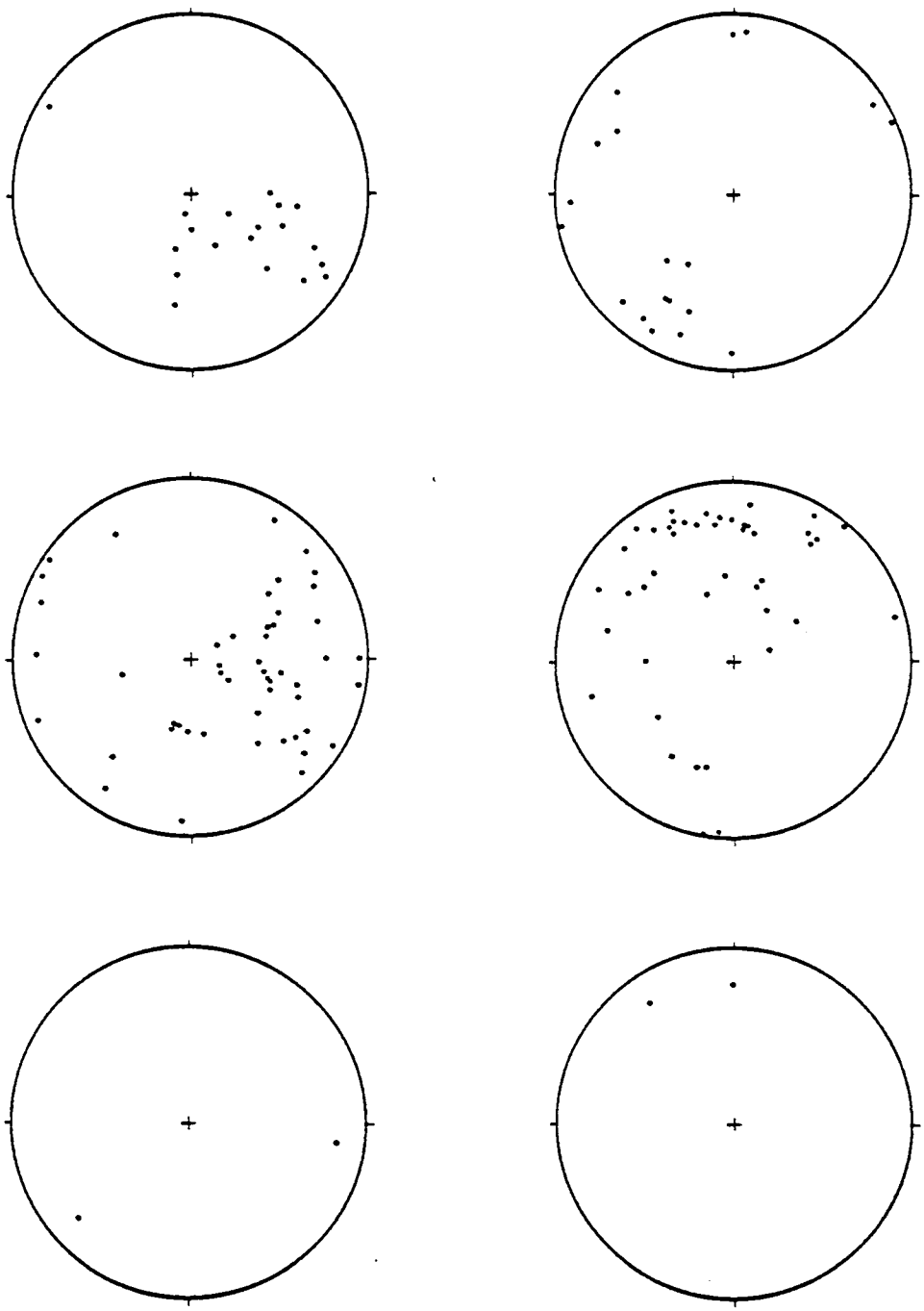


Figure 42: Domain I: Equal area plots of poles to axial surfaces (left column) and fold axes (right column) of mesoscopic folds. a. = Type 1 folds. b = Type 2 folds. c = Type 3 folds.

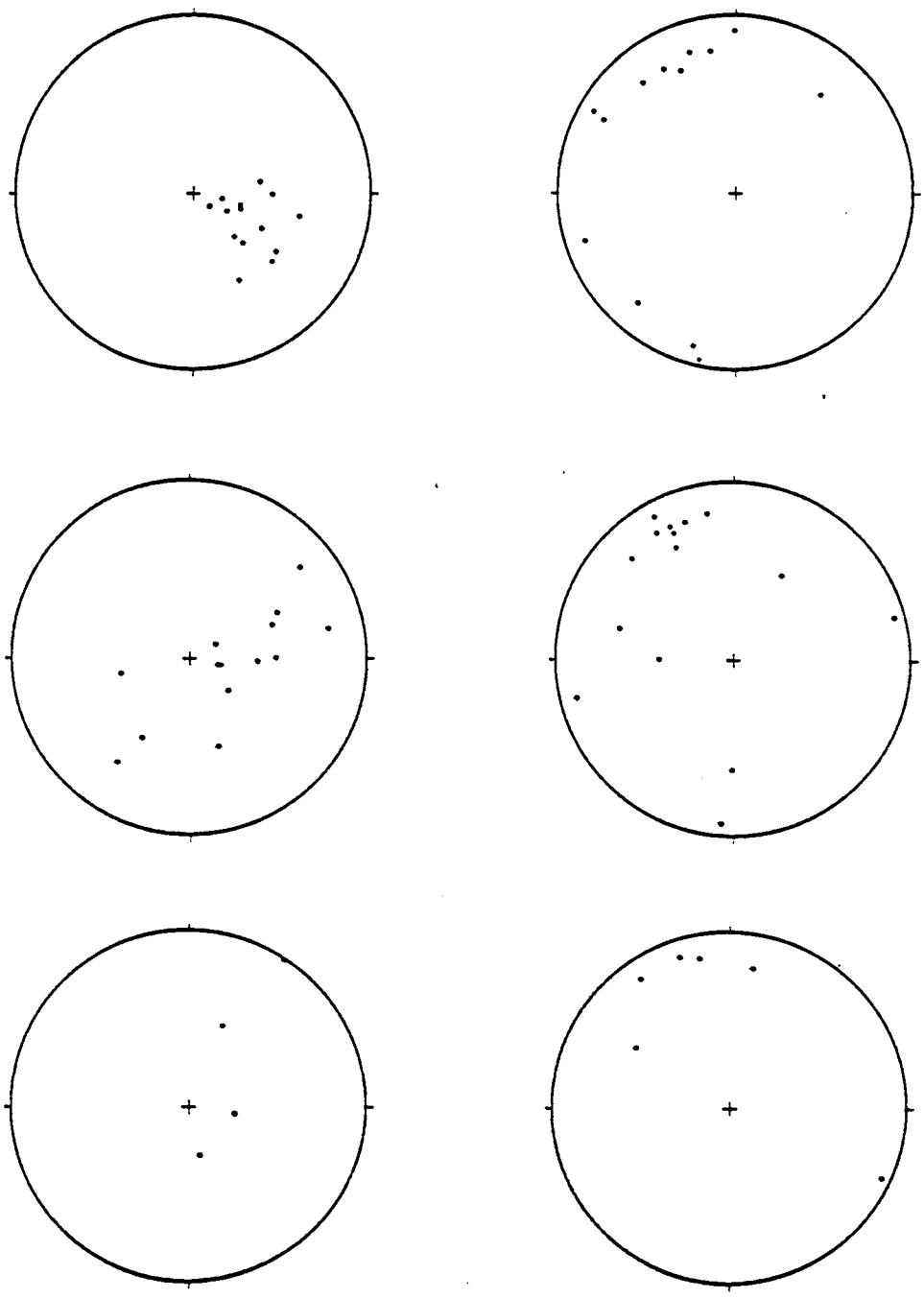


Figure 43: Domain II: Equal area plots of poles to axial surfaces (left column) and fold axes (right column) of mesoscopic folds. a. = Type 1 folds. b = Type 2 folds. c = Type 3 folds.

Most Type 1 axial surfaces strike NE to ENE (Fig. 44), with shallow S to SE dips. Southeastward-dipping surfaces are rarely observed in other domains. Axial surfaces of Type 2 folds lie on a great circle whose ß axis plunges 29° S87W. Type 3 folds are not abundant in Domain III. Where observed, axial surfaces are typically steep, but strikes are quite scattered.

Domain IV: Domain IV contains the axis and part of the northern limb of the Chattahoochee basin, where late, northwest-trending folds are abundant. This is reflected in the dominant foliation (Plate II, Plate III) as well as in axial surface orientations of all folds (Fig. 45), and all earlier structures have sustained reorientation during formation of the basin. Steep upright folds expose the Tallulah Falls Formation, underlying the Coweeta Group (Plate IV). Cylindrical best fit for foliations and all fold axial surfaces yield a similar ß axis, plunging approximately N48W at 20° (Fig. 46).

Domain V: The northern portion of the study area was designated Domain V. Foliations are oriented NW, but the maximum occurs NNW rather than WNW as in Domain IV. Fold axial surfaces are quite variable; Type 1 folds (Fig. 47) generally parallel the dominant foliation, whereas Type 2 folds display a more E-W trend, although with considerable scatter (Fig. 47). Axial surfaces of Type 3 folds strike NNE, and primarily dip NW.

Northwest-trending macroscopic folds dominate most of Domain V (Fig. 40, page 123), although some due north trends are evident near the trace of the Shope Fork fault. The fault does not truncate the dominant foliation, but the tight NW folding of Domain IV and a portion of Domian V is absent in the thrust sheet. This area may define a separate domain, but is included in Domain V because of a lack of abundant data.

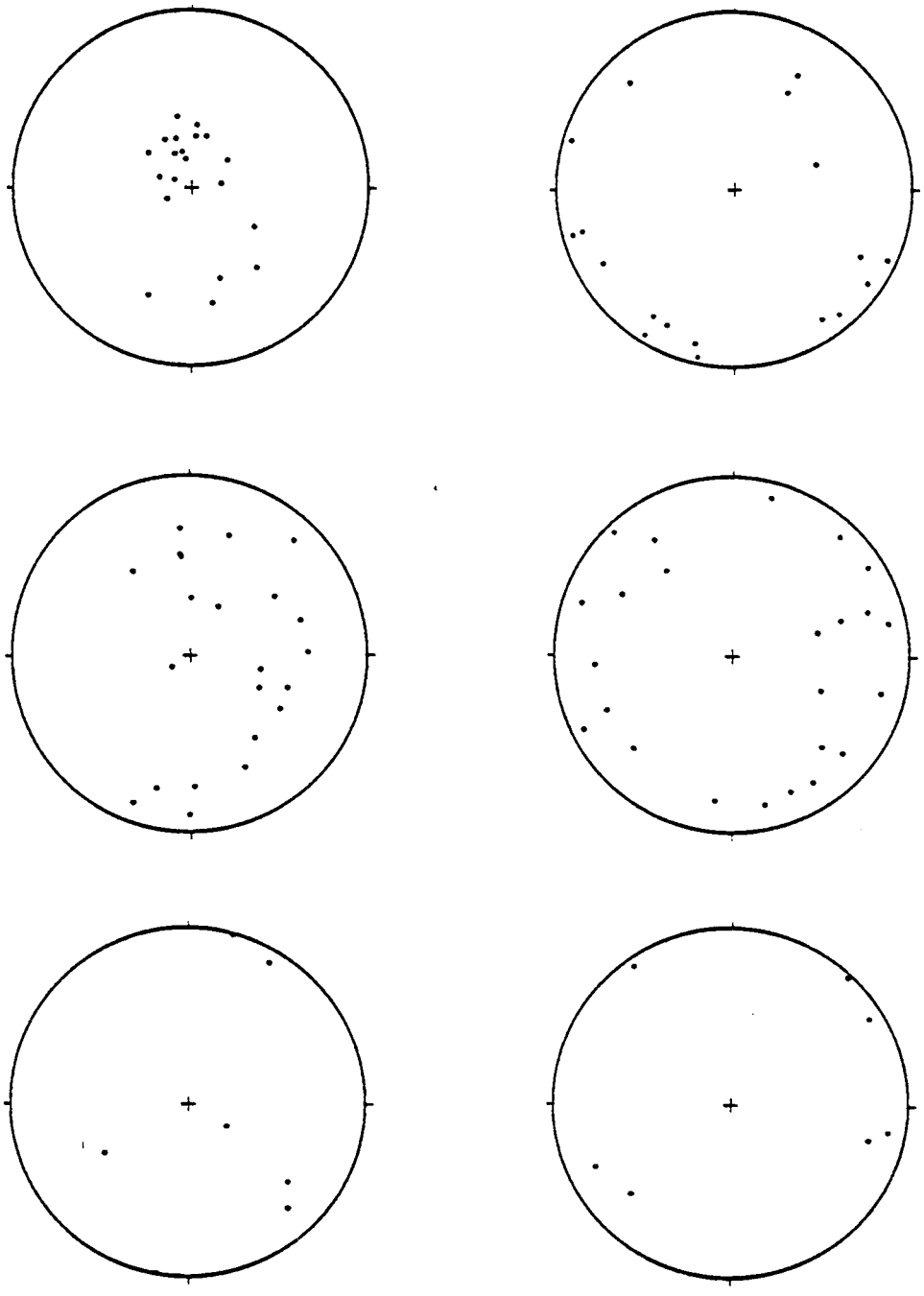


Figure 44: Domain III: Equal area plots of poles to axial surfaces (left column) and fold axes (right column) of mesoscopic folds. a. = Type 1 folds. b = Type 2 folds. c = Type 3 folds.

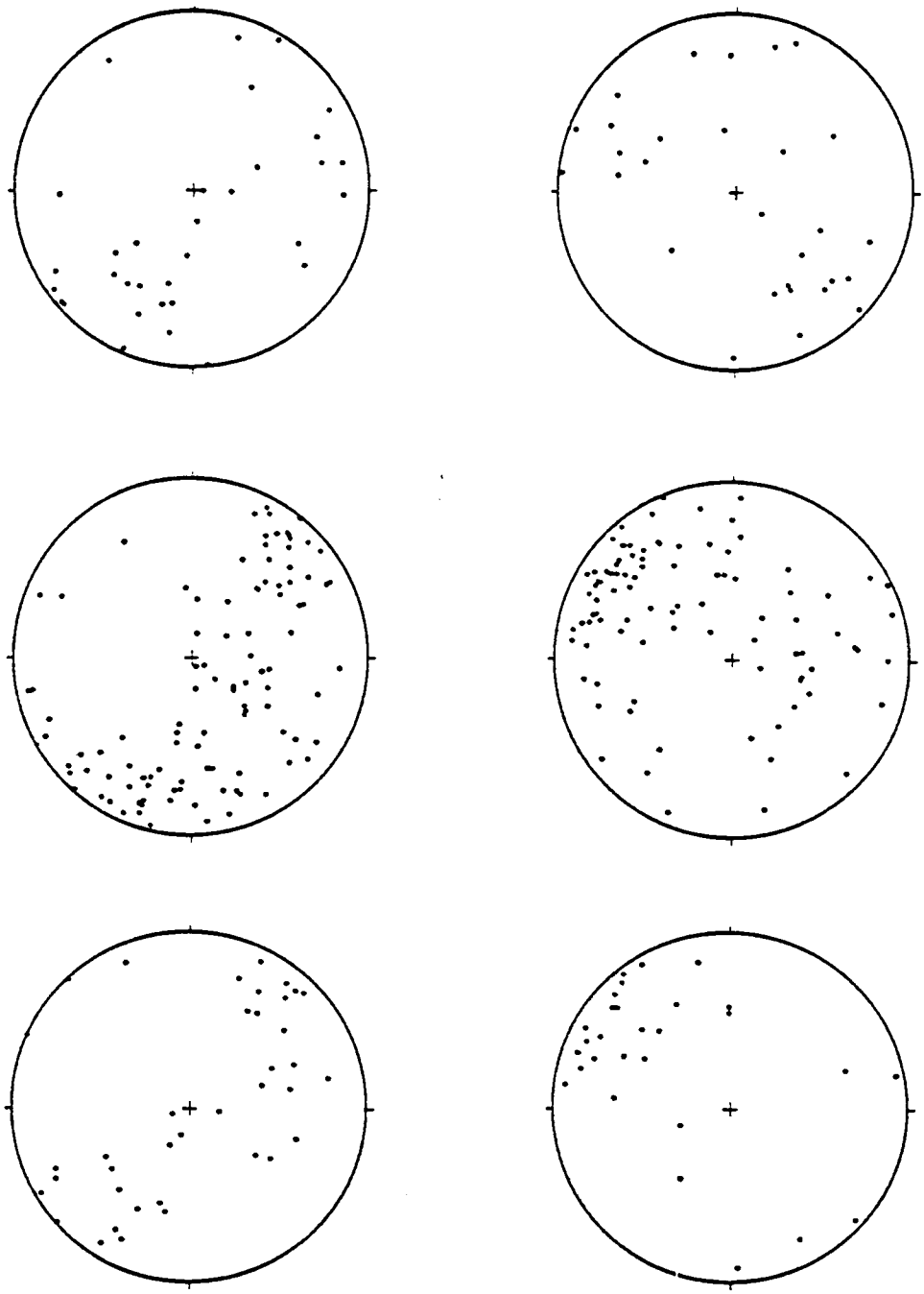


Figure 45: Domain IV: Equal area plots of poles to axial surfaces (left column) and fold axes (right column) of mesoscopic folds. a. = Type 1 folds. b = Type 2 folds. c = Type 3 folds.

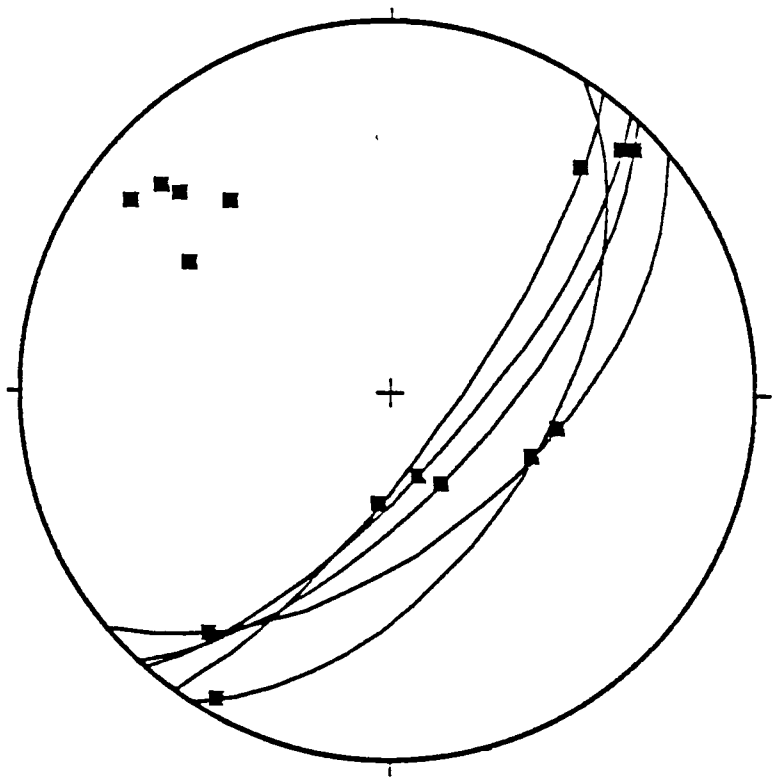


Figure 46: Cylindrical best fit for foliations and mesoscopic fold axial surfaces from Domain IV.

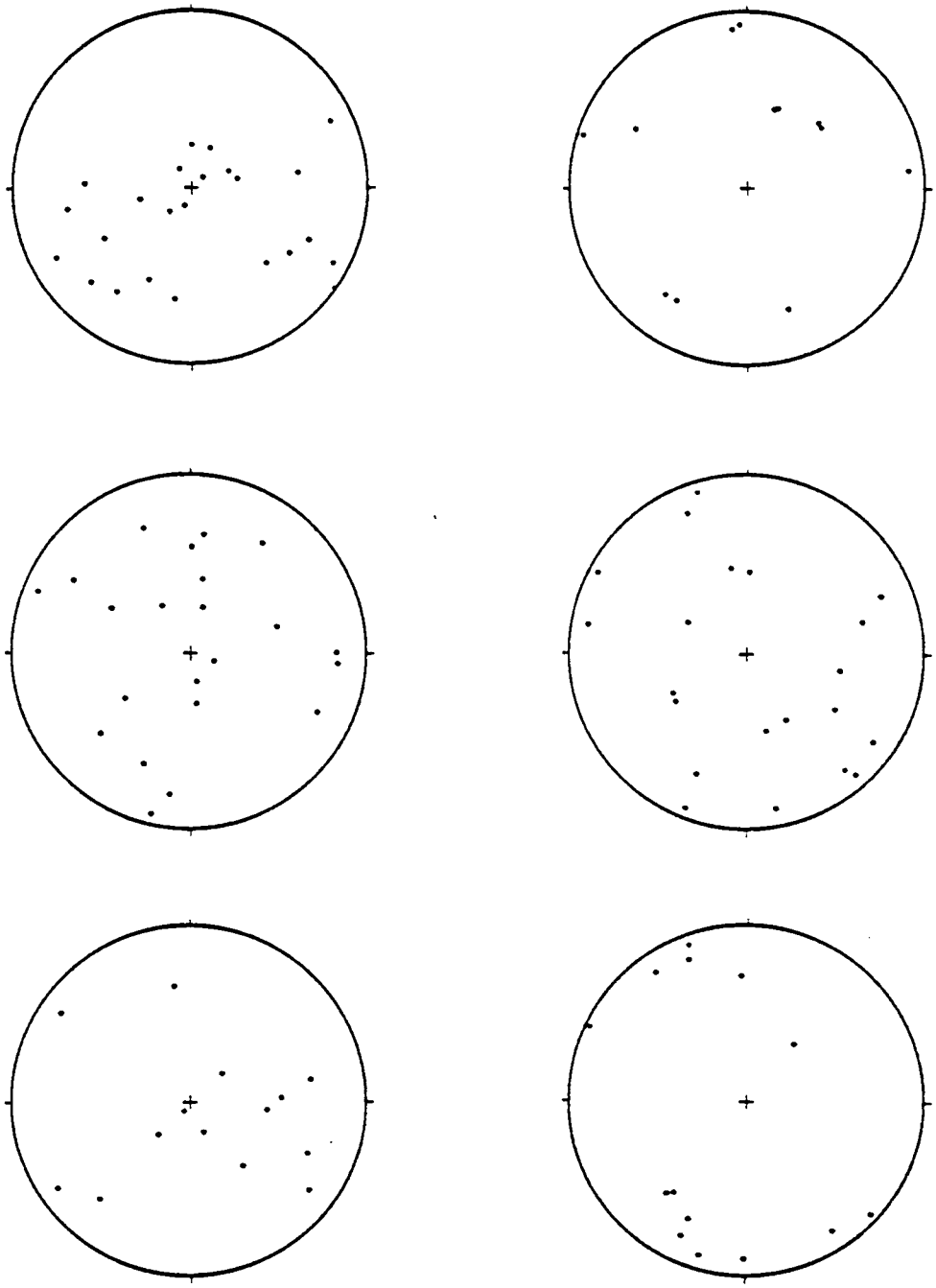


Figure 47: Domain V: Equal area plots of poles to axial surfaces (left column) and fold axes (right column) of mesoscopic folds. a. = Type 1 folds. b = Type 2 folds. c = Type 3 folds.

Faulting

Two thrust faults mapped in adjacent areas by previous workers have been extended into the study area (Plate II). These are the Shope Fork fault (Hatcher, 1976, 1979), and the Soque River fault, which was previously mapped as a continuation of the Shope Fork fault (Gillon, 1982; German, 1985; Nelson, 1985). Subsequent work has shown that these two are not the same contact (Hatcher and others, 1987). Recognition of the faults has been based primarily on stratigraphic criteria, and no indication of brittle movement was observed in association with either during this study. Evidence for both faults will be considered below.

Shope Fork Fault: The Shope Fork fault cuts the northwest limb of the Coweeta syncline in its type area (Hatcher, 1979), placing rocks of the Tallulah Falls Formation (biotite gneiss and schist with amphibolite pods) upon rocks of the Coweeta Group. The fault was mapped primarily on truncation of stratigraphy, and does not offset isograds. It is extensively folded, and may have formed as a tectonic slide during early folding (Hatcher, 1979). The Shope Fork fault apparently joins the Chunky Gal Mountain fault, the trace of which is exposed to the west (Fig. 1, page 2) (Hatcher and others, 1986). This fault juxtaposes similar sequences; however, there is evidence for later, post-metamorphic peak movement along the Chunky Gal Mountain portion of the fault system, including annealed mylonites and abrupt change in orientation of dominant foliations (Hatcher, 1978).

In the study area, the contact mapped as the Shope Fork fault is somewhat different in character; the footwall is occupied by mafic-poor rocks of the upper Tallulah Falls Formation (underlying the Coweeta Group); the hanging wall contains mafic-rich rocks

of uncertain affinity (Corn Ridge formation). There is tentative evidence for movement subsequent to peak metamorphism in the possible truncation of isograds (page 74), and the fault surface appears to cut steep, northwest-trending folds in the Tallulah Falls Formation (Plate IV). Also, extensive retrogressive effects are localized in this portion of the thrust sheet.

Soque River Fault: The Soque River fault (Plate II) was originally mapped as the Shope Fork fault by Nelson (1985) in the Tray Mountain quadrangle, and by Gillon (1982) in the adjacent Helen quadrangle to the southwest. This was considered to be a continuation of the Shope Fork fault of Hatcher (1976, 1979); the Shope Fork fault, however, juxtaposes different lithologies, and is apparently limited to an area to the north and west. Therefore, the former structure was renamed the Soque River fault by Hatcher and others (1987).

The Soque River fault separates the Otto Formation, with interlayered amphibolites and isolated ultramafic pods, from the essentially mafic and ultramafic-free Coweeta Group. In adjacent areas, it appears to place rocks of higher metamorphic grade (Coweeta Group) upon lower grade rocks (Otto Formation equivalent) (Nelson, 1985), although this criterion is not supported by data from the present investigation. Foliations are not truncated by the fault, and it appears to be isoclinally folded toward the northeast. The fault surface now dips northwest as a result of folding subsequent to emplacement.

Two possibilities exist for the configuration of the Soque River fault: the hanging wall is occupied by the Coweeta Group and underlying Tallulah Falls Formation; or the Otto Formation occupies the hanging wall, in which case the fault surface must have been tightly folded and overturned, producing the northwest dip. Implicit in the latter

interpretation is large-scale overturning of the Coweeta Group as well. Although it is certainly possible to produce such a structure in a ductilely-deformed terrain, no evidence has been recognized to suggest that the Coweeta Group does not represent an essentially upright (although extensively folded) sequence. Therefore, the former interpretation is favored. Since neither the Otto Formation or the Coweeta Group has been recognized east of the present outcrop belt of the Otto Formation, it is difficult to pinpoint a source in which to root the fault.

Evidence for a fault contact between the two units is not overwhelming in the study area, although regional relationships support the existence of such a structure. The Fry technique (Fry, 1979) of strain analysis was performed on three metasandstone samples, in order to test the possibility of higher strains near the fault surface. This technique is a graphical approximation of the center-to-center method (Ramsay, 1967), in which the bulk rock strain is calculated by utilizing the redistribution of points in the deformed rock. Fry diagrams were constructed by the method described by Ramsay and Huber (1983), for the following samples:

#605: Otto Formation (footwall), approximately 150 meters from the trace of the fault;

#999: Coleman River Formation (hangingwall), approximately 600 meters from the trace of the fault;

#1028: Coleman River Formation (hangingwall), approximately 3300 meters from the fault trace.

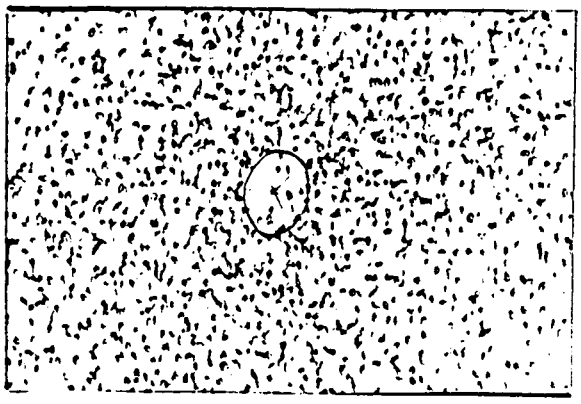
For each sample, only one thin section was available, cut across the foliation (preserving the XZ plane).

Results are illustrated in Figure 48. The ellipticity of the strain ellipse (R) was calculated for each sample. This value shows a small variation between samples, but provides little information about strains near the fault; if anything, it increases with greater distance from the fault. A major problem with this particular analysis stems from the amount of recrystallization apparent in the samples, which surely has modified the original grains. From this short exercise, I conclude that higher strains may have existed near the fault but, if so, have been partially annealed through recrystallization. Simpson (1985) concluded that recovery and recrystallization processes dominate in quartz, feldspar, and micas at epidote-amphibolite facies and above. This suggests that the Soque River fault was emplaced prior to or during these conditions.

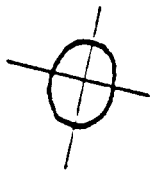
Cross Sections

Three cross sections are presented in Plate IV, corresponding to section lines on Plate II. Cross sections constructed for this study are designed to illustrate the dominant structural form in areas that are relatively homogeneous. Intensely deformed bodies of metamorphic rocks often are not good subjects for depiction in cross section (Turner and Weiss, 1963). Different generations of folds with divergent axes, and inconsistent and unpredictable thickness changes related to penetrative deformation may preclude one from making accurate predictions of the structure in the subsurface. Despite these pitfalls, cross sections are still the best available means to portray the general concept of structural relationships.

Two assumptions were made in order to construct the cross sections. First, the stratigraphic sequence pictured in Fig. 9 (page 58) is thought to be upright (Hatcher, 1985; see also page 135). Second, a thickness estimate of approximately 1830 meters

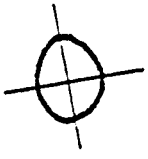


a. Sample # 605

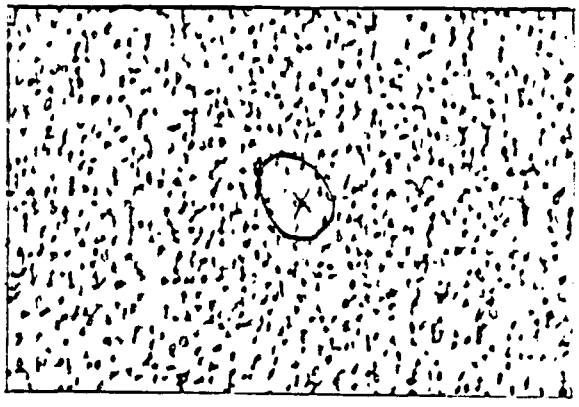


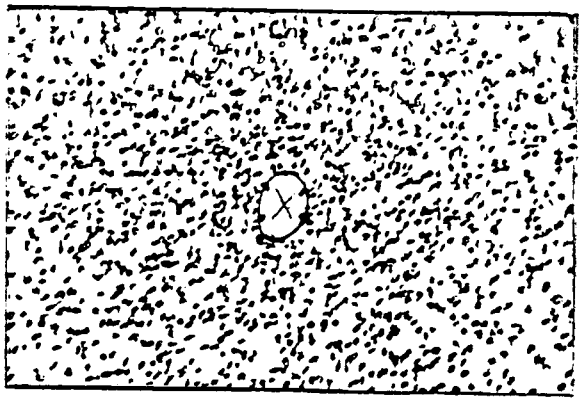
R 1.36

b. Sample # 999



R 1.40





c. Sample #1028



R 1.48

Figure 48: Results of Fry analysis of samples across the Soque River fault.

for the Coleman River Formation, calculated from section B-B', was utilized in all three sections, with the assumption that it does not change drastically over the area. Thicknesses of other units cannot be determined because of faulting or erosion.

Section A-A' is drawn through the southeastern portion of the study area, where the dominant strike of the foliation is northeast. The major structure is an overturned syncline cored by the Ridgepole Mountain Formation, at Middle Ridge. Another syncline at Taylor Ridge is inferred because of an outcrop of Ridgepole Mountain Formation southwest of the section line. The Persimmon Creek Gneiss is thought to pinch out so that the Tallulah Falls Formation underlies the Coleman River Formation to the west.

Section B-B' crosses a major portion of the study area, where the dominant strike is northwest. The syncline depicted in Section A-A' is folded by upright northwest-trending folds, exposing the Tallulah Falls Formation. In Section C-C', these folds are cut by the gently warped Shope Fork fault.

VII. DISCUSSION AND INTERPRETATIONS

Significance and Interpretation of Rock Types

Metasedimentary Rocks

Sedimentation and Tectonic Setting: On the basis of petrographic data, rocks of the study area are interpreted to be primarily turbidites of greywacke composition. Greywackes are characteristic of geosynclinal tracts of orogenic belts (Pettijohn and others, 1972), and typically occur in thick, rhythmically interbedded sand-shale sequences (Blatt and others, 1973). They are generally considered texturally immature sands, although one study (Brenchley, 1969) suggested that at least some greywackes are derived from diagenesis of mineralogically unstable volcanic arenites. Most geologists agree, however, that the term "greywacke" does connote a specific tectonic environment in which erosion, transportation, and deposition are rapid enough to prevent complete chemical weathering of detrital meterials. According to Pettijohn and others (1972) the majority of greywackes in the geologic record appear have a marine origin, and were deposited as turbidites. Mineralogical composition generally indicates a varied provenance.

Unfortunately, sedimentary structures in rocks of this study have been virtually obliterated during complex deformation and high-grade metamorphism. Overall, the units do have the appearance of thick clastic sequences of interbedded sand and shale layers, and some layers appear to exhibit graded bedding. The rocks may represent incomplete Bouma sequences that are common in modern-day turbidites (Walker, 1978), in which thick-bedded or high-density turbidites are characterized by a dominance of

massive and parallel laminated sands (A and B divisions of the sequence) (Leeder, 1982).

Metasandstones of the Tallulah Falls Formation may have been matrix-rich greywackes, with both abundant plagioclase and chlorite possibly derived from a volcaniclastic source. Volcanic material could have come from waning volcanic activity which is reflected in more easterly belts of the Tallulah Falls Formation and equivalents or from rapid erosion and redeposition of material from a slightly earlier volcanic episode. A partially continental source is postulated as well, for the quartz content and (possibly) microcline which has been lost to metamorphism.

Greywackes of the Coleman River Formation are similar to those of the Tallulah Falls Formation but contain less matrix. The presence of feldspathic sandstone and arkose may reflect a lessening of volcanic input and a larger continental influence. Those rocks designated "arkose" may not be true arkoses in the strict sense, indicating a rock which is specifically derived from a granitic source (Pettijohn and others, 1972). Rather, they are "cleaner" feldspathic greywackes which show influence of, but not total control by, continent-derived sedimentation.

Psammitic rocks of the Ridgepole Mountain Formation are difficult to interpret because of the presence of abundant garnet and, in some cases, sillimanite, but the unit is characterized by a relatively higher aluminum content and a more pelitic nature than the previous units. The higher alumina rocks may reflect proximity to a granitic source, as suggested by Hadley and Goldsmith (1963).

Greywackes of the Corn Ridge formation are interlayered with mafic rocks suggestive of basaltic volcanic activity. Mica content is high, and may indicate an original matrix abundance, possibly derived from alteration of material from a proximal volcanic source. Associated lithologies of the Otto Formation are similar to those of the Corn Ridge formation, but sandstone composition is highly variable, and more data would be necessary for a viable interpretation.

Thick sequences of metagreywacke and associated lithologies are found throughout the Blue Ridge. These are commonly interpreted to have been deposited in basins formed during continental rifting (see Rankin, 1975; Hatcher, 1978; Wehr and Glover, 1985). For some units, such as Ashe Formation (Conte, 1986) and the Alligator Back Formation (Conley, 1987), this interpretation is supported by geochemistry of intercalated amphibolites, if it can be assumed that the amphibolites represent lavas extruded during deposition of the associated sediments. Few, if any, data of this kind exist for rocks of the study area, with the exception of the Otto Formation (see page 145).

An alternative possibility for accumulation of these thick turbiditic sediments is an outer-arc trough formed near a volcanic arc in a subduction related setting, somewhat analogous to the depositional environment of the Great Valley sequence of northern California (Dickinson, 1971; Dickinson and Seely, 1972; Sucheki, 1985). Outer-arc trough successions can be recognized by great thickness, predominance of clastic facies, and elongate basin shape (Mitchell and Reading, 1978). They typically have three sediment sources: the outer arc (often dominantly turbidites), the volcanic arc, and a longitudinal source from the adjacent continent. The inferred composition of metagrey-wacke protoliths in this study may include both continental and volcaniclastic material

(see page 141), which would accumulate in such an environment. Although basin characteristics are difficult to reconstruct because of intense deformation, the original sediments were thick, predominantly clastic, and appear to have been laterally continuous.

In this interpretation, sediments of the Otto Formation, or the Corn Ridge formation, may have accumulated in close proximity to the volcanic arc, and consist of interlayered basaltic lava flows and terrigenous clastic sediments. The Tallulah Falls-Coleman River-Ridgepole Mountain Formation sequence was deposited either at a greater distance from the arc or during waning volcanic activity. The trend toward "cleaner" sediments upward in the section may represent closing of the basin, and possibly also a lower relief of the source area, as suggested by Hatcher (1979). The abundant pelitic sediments of the Ridgepole Mountain Formation may be explained by the slowing of sedimentation as the source area was lowered, and possibly the advent of low-density turbidity currents which can deposit appreciable mud (Rupke, 1978). If this interpretation is correct, the entire volcanic arc-outer arc trough sequence has been telescoped by subsequent faulting.

Comparison of the Coweeta Group with the Richard Russell Formation: In previous work, rocks of the Coweeta Group in the study area have been included in the Richard Russell Formation (Gillon, 1982; Nelson, 1985; Nelson and Gillon, 1985). Gillon (1982) suggested that the Richard Russell Formation might be correlative with the Coweeta Group, although the two units were separated by his placement of the Shope Fork fault. Subsequent compilation has shown this boundary as a separate fault (Hatcher and others, 1987; McClellan and Hatcher, 1987), (discussed on page 135). In this

interpretation, rocks of the study area are continuous with the Coweeta Group in its type locality.

The Richard Russell Formation as described by Gillon (1982) and Nelson (1985) consists of an essentially undivided assemblage of primarily biotite gneiss, with mica schist, metasandstone, quartzite, hornblende gneiss, amphibolite, and calc-silicate layers. Gillon defined highly schistose horizons as the Rice Cabin Member. In this study, I was able to divide the rocks into three mappable units, correlative with the Persimmon Creek Gneiss, Coleman River Formation, and the Ridgepole Mountain Formation of the Coweeta Group. In addition, a subtly unique underlying unit, tentatively correlated with the upper Tallulah Falls Formation, was recognized. It is important to note that mafic rocks are described as "abundant" in the Rice Cabin Member of the Richard Russell Formation to the southwest of the study area (Gillon, 1982), although they are rare in the Coweeta Group of this study.

Regional Correlations of the Coweeta Group: The succession of "dirty", matrix-rich greywackes to cleaner sandstones is characteristic of much of the eastern Blue Ridge belt. In particular, it has been suggested that the Tallulah Falls-Coweeta Group sequence may be equivalent to the Ashe-Alligator Back and/or the Ocoee-Murphy belt sequences (Hatcher, 1979). Lithologic comparisons of the Coweeta Group with the Alligator Back Formation and the Murphy belt rocks show a number of dissimilarities to either unit. The Alligator Back Formation contains abundant microcline-bearing metagreywacke and mica schist with accessory tourmaline; these lithologies are interlayered with graphite schist, calcareous quartzite, and impure marble. More importantly, mafic and ultramafic rocks make up as much as 40% of the Formation (Conley, 1987). Murphy belt rocks consist

of laterally continuous units of graphitic argillite, schist, quartzite, marble, and calc-silicate granofels. Therefore, if the Coweeta Group is equivalent to either the Alligator Back Formation or the Murphy sequence, it must represent a quite different facies.

Based on modal and heavy mineral analyses, I suggest that the Great Smoky Group of the Ocoee Supergroup in the western Blue Ridge was derived from a somewhat different source area than rocks of this study. The Great Smoky Group is commonly thought of as having been deposited in basins formed upon intial rifting of the continent (see Rankin, 1975; Hatcher, 1978). Rocks of this study may represent coeval sediments deposited in separate basins with different source areas, or a younger sequence deposited during formation and subsequent destruction of an island arc.

Mafic Rocks

Discrimination of the tectonic setting of amphibolites associated with the Otto Formation and the possibly correlative New Georgia Group has been attempted in several studies, although results are, as yet, somewhat ambiguous. Geochemical data for New Georgia Group amphibolites (McConnell and Abrams, 1984) suggest an extensional environment for formation, interpreted as a back-arc or marginal basin (McConnell and Abrams, 1986). In contrast, geochemical analyses of amphibolites associated with the Otto Formation just northeast of the study area by Hopson (1988) indicate a predominant island arc affinity. Analyses suggesting both ocean floor and island arc types in the Helen terrane (Otto equivalent) were reported by Nelson (1988). It is feasible that more than one body of amphibolite is represented by these analyses, although Nelson did not address this possibility. Results of prelimary analyses of amphibolites associated with the Corn Ridge formation are inconclusive at this time (Hopson, in progress).

Ultramafic Rocks

The actinolite schist found within the Otto Formation is interpreted as an altered ultramafic body. Actinolite is a common product of low-grade regional metamorphism (greenschist facies) of ultramafic or carbonate rocks (Phillips and Griffin, 1981), or of mafic rocks (Winkler, 1979), but as metamorphic grade increases to amphibolite facies, actinolite changes to hornblende. According to Winkler (1979), this transformation occurs at about the same pressure and temperature conditions as for the formation of almandine garnet in metapelitic rocks. Because of the presence of garnet in some of the surrounding pelitic schists, and the mutual occurrence of the actinolite schist and massive (hornblende-rich) amphibolite within a few meters, the actinolite does not appear to be a product of the metamorphic processes described above. In addition, metamorphic minerals such as talc, epidote, carbonates or chlorite, commonly associated with actinolite formed during low-grade metamorphism (Phillips and Griffin, 1981), are absent in the actinolite schist. An alternative explanation for the presence of actinolite is as the product of hydrothermal alteration of mafic rocks containing pyroxene. The resulting pale green amphibole, considered to be actinolitic in composition (Deer and others, 1963), is given the name "uralite".

Significance and Interpretation of Metamorphism and Structure

Metamorphic Textures

A number of interesting textures were described from the study samples, and are reiterated below:

(1) Two preferred orientations of micaceous minerals are present in many samples:

   (a) The dominant foliation is nearly always defined by biotite; parallel to subparallel muscovite may or may not be present.

   (b) A subordinate foliation results from muscovite which cross-cuts the dominant one, often at a high angle. Porphyroblastic muscovite is frequently observed. Rarely, biotite is parallel to this subordinate foliation as well.

(2) Muscovite sometimes occurs also as fine-grained aggregates with random orientation in addition to the oriented grains described above.

(3) Garnet occurs in two forms:

   (a) Sub- to anhedral grains, typically large, and often containing inclusions of biotite, quartz, and muscovite. In many cases these are highly fractured.

   (b) Tiny, euhedral grains, in some cases growing over porphyroblastic muscovite. These typically contain no inclusions.

(4) Textures exhibited by the Al_2SiO_5 polymorphs are quite enigmatic. For example:

   (a) Kyanite ranges from small, subhedral to porphyroblastic euhedral grains. More often than not it possesses no distinct preferred orientation. Where it coexists with sillimanite, the relationship is ambiguous; in one sample, porphyroblasts of kyanite coexist with sillimanite with no evidence for reaction, whereas in another, idiomorphic kyanite clearly cross-cuts the foliation containing sillimanite.

(b) Sillimanite occurs in two forms: needles, prismatic grains, and masses of fibrolite interleaved with biotite and/or muscovite, commonly within a foliation which wraps around large garnet; and radial fibrolitic masses within plagioclase and quartz grains.

What interpretations can be made from these textures and relationships? Some lines of evidence strongly suggest two phases of mineral growth. The two sets of preferred orientation fabrics in the micaceous minerals (especially muscovite) may indicate syntectonic growth or recrystallization in response to different stress regimes. The difference in preferred orientations (Spry, 1969), combined with two size fractions of muscovite, porphyroblastic vs. finer-grained (Guidotti, 1970), suggest two generations of the mineral.

Large, subhedral, fractured garnets may be pretectonic (Spry, 1969). Inclusions of biotite, muscovite, and quartz indicate that garnet postdated at least some biotite and muscovite formation. Tiny, euhedral garnets overprint porphyroblastic muscovite in at least one sample, and these may represent a second generation of garnet growth.

As discussed earlier, the kyanite-sillimanite relationship is ambiguous. At any rate, the coexistence of two Al_2SiO_5 polymorphs generally indicates disequilibrium, with one mineral existing metastably (Spry, 1969; Mason, 1978). Kyanite rarely, if ever, crystallizes syntectonically (Spry, 1969), and appears to be randomly oriented in the study samples, but sillimanite which is intergrown with biotite often forms a foliation. Whether or not the latter two minerals formed synchronously (under the influence of a stress field) or sillimanite is replacing biotite is unclear. The previous discussion of

possible reactions suggests synchronous growth. Kyanite clearly crystallized later than sillimanite in sample #882.

Timing of Metamorphism

Three distinct Paleozoic orogenic episodes have been recognized in the southern Appalachians. Traditionally, these have been correlated with the Taconic, Acadian, and Alleghanian episodes of the central and northern Appalachians. In recent years, workers have utilized isotopic ages of regionally metamorphosed rocks and their structural position in relation to plutons of known age to bracket timing of these orogenies.

The Taconic orogeny produced regional metamorphism and ductile deformation throughout much of the Piedmont and Blue Ridge. On the basis of K-Ar dates on metamorphic minerals, Butler (1972, 1973) concluded that the peak of regional metamorphism was reached in the Blue Ridge about 430-470 Ma, as part of this episode. Glover and others (1983) suggested that Taconic metamorphism was a protracted event, beginning as early as 200 million years before the thermal peak. A number of studies cited by Glover and others (1983) provide mineral cooling ages which broadly concur with Butler's conclusions.

Evidence for Acadian metamorphism in the Blue Ridge is less well-recognized. It appears to be the only metamorphic event affecting rocks of the Talladega Belt (Tull, 1982), but elsewhere it is generally limited to a lower grade overprint of the Taconic event. Butler (1973) suggested the occurrence of a post-Taconic metamorphic event based upon radiometric dates of 390-320 m.y. from pegmatites of the Spruce Pine district, North Carolina. On the assumption that pegmatite formation requires T, P

conditions similar to those expected to produce medium to high-grade metamorphism, he surmised that these bodies were emplaced during a regional metamorphic episode. Butler's proposal was supported by Rb-Sr whole rock ages of 402-350 m.y. from samples of the Ashe and Alligator Back Formations in North Carolina and southern Virginia (Fullagar and Dietrich, 1976), although the authors cautioned that Rb-Sr ages have been interpreted by some to represent time of deposition or dewatering of sediments rather than time of metamorphism. Glover and others (1983) cited studies of undeformed to mildly deformed intrusive rocks in the eastern Piedmont, primarily granites, which yield "Acadian" ages, but indicated that the Blue Ridge and western Piedmont offered little evidence for Acadian effects. However, Acadian metamorphism was a major event in the Inner Piedmont of Georgia, and in the Talladega Belt and Ashland-Wedowee belt of eastern Alabama. Also, some mylonites from the Brevard fault zone yielded Acadian ages (studies in Glover and others, 1983). Effects of an Acadian event, however sporadic, clearly exist, and it is probably reasonable to assume that these effects extend into the area of this study.

The Alleghanian deformation was accompanied by metamorphism in the eastern Piedmont, but is characterized by brittle deformation in the eastern Blue Ridge. An age range of 330-230 m.y. is given for this orogenic episode by Glover and others (1983).

In light of the above discussion, it seems likely that kyanite to sillimanite-grade metamorphism occurred during the Taconic episode, some 430 to 470 million years ago. The second phase of lower-grade mineral growth conforms to the character of the Acadian event as recognized in much of the Blue Ridge. Whether the "Taconic" and "Acadian" can actually be separated into two episodes, or are both phases of one protracted event is still subject to speculation.

Comparison of Deformational Sequence with Previous Work in the Region

Hatcher and Butler (1979) recognized six phases of folding in rocks of the eastern Blue Ridge. Nelson (1985) observed at least four phases in the Greenville 2 degree quadrangle, which includes the study area. Table 8 compares this study with the previous work above, and summarizes the interpreted deformational sequence and timing of events from this study.

Nelson (1985) considered F_3 folds to belong to a conjugate pair, with one set oriented NNE and the other WNW. In this study, F_4 folds, oriented NNE, and F_5 folds, oriented WNW, may correspond to Nelson's F_3 folds. No evidence was observed here to substantiate the interpretation of a conjugate set, however. F_4 folds locally form an axial planar foliation, observed as a schistosity or weak banding in pegmatite bodies, while F_5 folds apparently have no associated fabric.

Although Nelson (1985) designated the regional foliation as S_1 (Table 8), he did suggest the possibility of a pre-S_1 deformation, based on the presence of foliated clasts in metasedimentary gneiss. These clasts may have been derived from basement rocks, or they could represent vestiges of folding that produced the pinstriped lamina in rocks of the Coleman River Formation.

TABLE 8: Comparison of deformation sequence and timing (this study), with previous work in the region

Hatcher and Butler, 1979 Blue Ridge and W. Piedmont	Nelson, 1985 NW part of Greenville 2^0 Quadrangle	This Study Tray Mountain and Macedonia	Interpreted Deformational Event (this study)	Interpreted Timing (this study)
F_6--Upright, open. Orientation--NW.	Post-F_3--Gentle, open folds, around an E-W or NNE axis.	F_6--Open, upright to recumbent. NW.	D_4	Alleghanian (?)
F_5--Upright, open. Orientation--NE.	F_3--Tight to open, upright to overturned flexural folds or crenulations. Conjugate set, WNW and NNE.	F_5--Open to chevron, upright to overturned. NW to WNW.		
F_4--Crenulation cleavage. Orientation--NE.		F_4--Tight to open, upright folds or kink folds. Local S_4 schistosity. NNE.	D_3	Acadian
F_3--Upright, isoclinal to open (ductile). Orientation--NE.	F_2--Steeply inclined to recumbent. Major fold set. NE to SW. Localized S_2.	F_3--Tight to isoclinal, recumbent. Major fold set. NE. Local S_3 (schistosity).	D_2	Taconic
F_2--Isoclinal, recumbent. Formed dominant s-surface. Orientation--EW to NE.	F_1--Recumbemt isoclines, rootless, intrafolial. Formed dominant foliation (S_1).	F_2--Recumbent isoclines, rootless, intrafolial. Formed dominant foliation (S_2). E-W?	D_1	Taconic
F_1--Isoclinal, recumbent. Formed S_1. Rarely observed. Orientation E-W to NE.	Pre-S_1 deformation (?). Presence of foliated clasts in metasedimentary rocks. (Foliation oblique to S_1.)	F_1--folds not observed. Formed S_1 (pinstriping).		

Timing of Faulting

Shope Fork Fault: In the study area, the Shope Fork fault does not truncate the dominant foliation, but does appear to juxtapose rocks of different metamorphic grade. Tight northwest-trending folds (F_5) do not extend across the fault, and effects of retrograde metamorphism are concentrated in the Shope Fork thrust sheet. Thus, the bulk of the data indicate post-metamorphic movement along the fault, occurring either syn- or post-D_4.

The Shope Fork fault is interpreted to be a pre- to syn-metamorphic structure in its type area (Hatcher, 1979). The Chunky Gal Mountain fault, which either joins or truncates the Shope Fork fault (Fig. 1, page 2), has a long and complex movement history, and contains retrograde mylonites (Lacazette, 1986). Therefore, I suggest that the Shope Fork fault as mapped in the study area may actually be more analogous to the Chunky Gal Mountain portion of the fault system. Movement of the thrust sheet may have begun prior to D_4, although later than peak metamorphism, and subsequently experienced reactivation and retrograde metamorphism during D_4.

Soque River Fault: Timing of movement on the Soque River fault (which has been erroneously referred to as the Shope Fork fault by earlier workers) is ambiguous. According to German (1985), the fault represents the earliest recognizable deformational event in the Dahlonega gold belt, and any associated fabric has been destroyed by later overprinting. Farrell (1986) did not consider the boundary between the Otto Formation and the Coweeta Group to be a fault at all, but rather a continuous stratigraphic contact. The fault has been reported by others, however, to truncate foliations (Nelson, 1985),

or to juxtapose rocks of differing metamorphic grade (Gillon, 1982). Neither of these latter criteria hold true in the study area. In addition, the brief exercise in strain analysis described in Section VII indicates that if a strain gradient were present across the fault surface at one time, the higher strains must have subsequently been annealed. Evidence from this study is in agreement with either emplacement prior to amphibolite facies metamorphism, or with an unfaulted contact between the two units. The trace of the Soque River fault occupies a very limited portion of the investigation site; additional work is clearly necessary to resolve the uncertainties associated with this contact.

VIII. CONCLUSIONS

1. The study area can be divided into three tectonic units. Tectonic unit 1 includes the Otto Formation (mica schist, metasandstone, and abundant amphibolite), and is bounded on the northwest by the Soque River fault. Tectonic unit 2 comprises dominantly metasedimentary rocks of the Coweeta Group and possible Tallulah Falls Formation equivalents. The Coweeta Group can be divided into the Persimmon Creek Gneiss, Coleman River Formation, and Ridgepole Mountain Formation. Tectonic unit 3 includes an assemblage of garnetiferous mica schist, metasandstone, amphibolite, and hornblende gneiss, informally designated the Corn Ridge formation. This unit is separated from tectonic unit 2 by the Shope Fork fault.

2. Metasedimentary rocks of the study area were deposited as thick clastic sequences of greywacke with interbedded shale, probably as marine turbidites. Possible tectonic environments of deposition include basins formed during continental rifting, or an outer-arc basin adjacent to an eastward-dipping subduction zone.

3. Metamorphism reached lower sillimanite grade of the amphibolite facies throughout much of the study area. Rocks of the Shope Fork thrust sheet (tectonic unit 3) reached only kyanite grade, however, and were also affected by retrograde metamorphism.

4. Two episodes of prograde mineral growth are suggested from textural evidence. The first episode reached sillimanite or kyanite grade, while the second resulted in secondary growth of muscovite and garnet. These episodes may correspond to Taconic and Acadian metamorphism.

5. The Coweeta Group, the most extensive lithologic unit in the study area, cannot be positively correlated with other structurally similar units within the Blue Ridge. If equivalent to units such as the Murphy Group or the Alligator Back Formation, it must represent a considerably different facies. In addition, petrography and heavy mineral analysis suggest that rocks of this portion of the eastern Blue Ridge had a different source area from that of the Ocoee Supergroup to the west.

6. The area is structurally complex. Six folding phases were recognized; earlier flowage folds resulted in transposition of layering into a regionally penetrative foliation, while later flexural flow to flexural slip buckle folds have little or no associated fabric.

7. The Shope Fork fault in the study area is part of the Shope Fork-Chunky Gal Mountain fault system. Possible early, although post-metamorphic peak, movement may have occurred, although later reactivation and retrograde metamorphism occurred along this portion of the system. Timing of movement along the Soque River fault is uncertain, although evidence from this study suggests that it is a pre-metamorphic structure.

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APPENDICES

APPENDIX A

TABLE A-1: Model analyses of metasandstones

MINERAL	*SAMPLE # 1-TM49	2-TM605	3-TM2734	4-TM2764	5-TM1388	6-TM162	7-TM210	8-TM237
Quartz	70.10	42.44	72.63	50.99	50.85	40.82	48.23	38.01
Plagioclase	11.10	15.21	8.04	14.42	28.47	25.79	17.72	40.32
Biotite	14.50	30.23	10.43	24.69	18.18	29.31	29.82	19.86
Muscovite	1.10	11.11	7.75	7.54	0.70	1.62	3.54	0.00
Chlorite	–	–	–	–	–	–	–	–
Microcline	1.30	0.20	–	0.66	–	–	–	–
Garnet	0.60	–	0.29	–	0.10	1.17	0.10	1.50
Sillimanite	tr	–	–	–	–	–	–	–
Kyanite	–	–	–	–	–	–	–	–
Opaque	–	–	0.48	–	–	0.45	–	–
Zircon	–	–	tr	0.75	–	–	–	–
Apatite	–	–	0.29	0.19	–	0.09	–	–
Allanite	–	–	0.09	0.09	–	0.09	–	–
Epidote/ Clinozoisite	–	–	–	0.66	–	–	–	–
Zoisite	–	–	–	–	–	–	–	–
Myrmekite	–	–	–	–	–	0.63	–	–
Sphene	–	–	–	–	–	–	–	–
Carbonate	0.50	–	–	–	–	–	–	–
Tourmaline	–	–	–	–	–	–	–	–
**Accessory	0.80	0.80	--	--	1.70	–	0.59	0.30
Total	100.00	99.99	100.00	99.99	100.00	99.97	100.00	99.99

* First digit(s) refers to sample number used in sandstone classification, Section IV. "TM" and last digits are actual field sample number.
** May include zircon, apatite, epidote, opaque.

TABLE A-1 (continued)

***SAMPLE #**

MINERAL	9-TM239	10-TM294	11-TM330	12-TM510	13-TM645	14-TM999	15-TM1028	16-TM2101
Quartz	45.41	62.54	67.00	60.11	57.21	51.35	47.14	51.09
Plagioclase	19.45	14.25	17.42	24.95	22.74	16.55	31.38	26.84
Biotite	31.98	11.80	12.90	12.95	18.44	26.52	18.31	17.99
Muscovite	0.79	10.23	0.00	0.28	0.38	3.29	0.00	0.40
Chlorite	–	–	–	–	–	–	–	–
Microcline	–	–	–	–	–	1.10	–	–
Garnet	2.07	0.39	1.57	0.95	tr	0.19	2.09	2.88
Sillimanite	tr	–	–	–	–	–	–	–
Kyanite	–	–	–	–	–	–	–	–
Opaque	–	–	–	0.47	–	tr	0.54	–
Zircon	–	–	–	0.19	–	–	0.18	–
Apatite	–	–	–	–	–	–	0.18	–
Allanite	–	–	–	–	–	–	–	–
Epidote/ Clinozoisite	–	–	–	–	–	–	0.27	–
Zoisite	–	–	–	–	–	–	–	–
Myrmekite	–	–	–	–	–	–	–	–
Sphene	–	–	–	–	–	–	–	–
Carbonate	–	–	–	0.09	–	–	–	–
Tourmaline	–	–	–	–	–	–	–	–
**Accessory	0.30	0.80	–	–	1.20	1.00	–	0.80
Total	100.00	100.01	98.89	99.99	99.97	100.00	100.09	100.00

* First digit(s) refers to sample number used in sandstone classification, Section IV. "TM" and last digits are actual field sample number.
** May include zircon, apatite, epidote, opaque.

TABLE A-1 (continued)

***SAMPLE #**

MINERAL	17-TM2451	18-TM2737C	19-TM2744	20-TM2763	21-TM496	22-TM882	23-TM950A	24-TM955
Quartz	67.79	42.95	55.82	2.46	6.44	22.65	46.56	6.59
Plagioclase	15.56	23.38	23.49	18.69	38.57	22.46	31.85	19.78
Biotite	6.44	31.02	17.89	31.20	31.16	29.08	19.76	31.97
Muscovite	9.51	tr	–	43.38	17.12	2.24	1.36	38.26
Chlorite	–	–	–	–	–	–	–	–
Microcline	0.30	–	–	–	–	–	–	–
Garnet	–	1.17	1.40	3.20	2.29	6.34	0.17	1.20
Sillimanite	–	–	–	–	2.82	14.44	–	0.60
Kyanite	–	–	–	–	–	1.02	–	–
Opaque	0.30	0.78	0.09	0.53	0.09	1.77	–	0.60
Zircon	0.10	0.20	0.28	0.32	0.09	–	0.26	0.10
Apatite	–	0.30	0.28	0.21	–	–	–	–
Allanite	–	0.20	0.19	–	–	–	–	–
Epidote/ Clinozoisite	–	–	0.19	–	–	–	–	–
Zoisite	–	–	0.37	–	–	–	–	–
Myrmekite	–	–	–	–	1.41	–	–	0.90
Sphene	–	–	–	–	–	–	–	–
Carbonate	–	–	–	–	–	–	–	–
Tourmaline	–	–	–	–	–	–	–	–
**Accessory	–	–	–	–	–	–	–	–
Total	100.00	100.00	100.00	99.99	99.99	100.00	99.99	100.00

* First digit(s) refers to sample number used in sandstone classification, Section IV. "TM" and last digits are actual field sample number.
** May include zircon, apatite, epidote, opaque.

TABLE A-1 (continued)

***SAMPLE #**

MINERAL	25-TM2740	26-TM2250	27-TM2750A	28-TM370B	29-TM558	30-TM709	31-TM836	32-TM1099
Quartz	32.19	59.10	25.99	49.02	22.71	67.24	25.63	51.53
Plagioclase	9.76	12.43	29.95	11.44	26.80	–	31.84	21.22
Biotite	32.88	20.45	23.28	37.18	40.35	26.51	35.33	22.70
Muscovite	–	8.02	16.91	1.76	0.78	3.07	6.31	3.80
Chlorite	–	–	tr	–	–	–	–	–
Microcline	–	–	–	–	–	–	–	–
Garnet	10.63	–	0.97	0.20	7.70	–	0.19	0.18
Sillimanite	14.15	–	0.10	–	0.10	–	–	–
Kyanite	–	–	–	–	–	–	–	–
Opaque	0.39	–	0.10	–	1.26	2.59	0.19	0.18
Zircon	tr	tr	tr	0.10	0.10	0.48	0.19	–
Apatite	–	–	1.64	0.29	–	0.10	0.10	0.28
Allanite	–	–	–	–	–	–	–	0.09
Epidote/ Clinozoisite	–	–	–	–	–	–	0.10	–
Zoisite	–	–	–	–	–	–	–	–
Myrmekite	–	–	–	–	0.19	–	0.10	–
Sphene	–	–	–	–	–	–	–	–
Carbonate	–	–	1.06	–	–	–	–	–
Tourmaline	–	–	–	–	–	–	–	–
**Accessory	–	–	–	–	–	–	–	–
Total	100.00	100.00	100.00	99.99	99.99	99.99	99.98	99.98

* First digit(s) refers to sample number used in sandstone classification, Section IV. "TM" and last digits are actual field sample number.
** May include zircon, apatite, epidote, opaque.

TABLE A-1 (continued)

*SAMPLE #

MINERAL	33-TM2005	34-TM2120	35-TM2166	36-TM2762	37-TM2770	38-TM2178A	39-TM2180	40-TM2405
Quartz	49.25	6.51	38.88	7.58	6.93	21.19	9.73	35.79
Plagioclase	3.19	30.28	31.93	48.57	28.01	17.44	34.03	16.37
Biotite	42.40	41.04	16.74	40.69	39.27	27.85	35.06	42.41
Muscovite	1.50	14.24	10.70	2.26	24.83	26.85	19.85	4.87
Chlorite	–	0.47	–	–	–	0.55	–	–
Microcline	–	–	–	–	–	–	–	–
Garnet	0.19	5.19	0.36	0.19	0.19	4.29	0.38	–
Sillimanite	–	–	–	0.10	0.09	–	–	–
Kyanite	–	–	–	–	–	–	–	–
Opaque	2.81	1.79	0.27	0.10	–	0.46	–	0.37
Zircon	0.37	0.28	0.09	0.30	0.38	0.18	0.09	–
Apatite	0.28	0.09	0.27	0.19	0.29	0.55	0.28	–
Allanite			0.09			0.36		
Epidote/ Clinozoisite	–	–	–	–	–	–	–	0.09
Zoisite	–	–	–	–	–	–	–	0.09
Myrmekite	–	–	–	–	–	0.27	0.57	–
Sphene	–	–	0.64	–	–	–	–	–
Carbonate	–	–	–	–	–	–	–	–
Tourmaline	–	0.09	–	–	–	–	–	–
**Accessory	–	–	–	–	–	–	–	–
Total	99.99	99.98	99.97	99.98	99.99	99.99	99.99	99.99

* First digit(s) refers to sample number used in sandstone classification, Section IV. "TM" and last digits are actual field sample number.
** May include zircon, apatite, epidote, opaque.

TABLE A-1 (continued)

***SAMPLE #**

MINERAL	41-TM2724A	42-TM2728A	43-TM2769
Quartz	48.60	26.36	16.45
Plagioclase	17.78	41.93	43.96
Biotite	28.02	26.07	23.79
Muscovite	4.73	5.25	6.04
Chlorite	–	–	–
Microcline	–	–	–
Garnet	–	–	6.69
Sillimanite	–	–	1.30
Kyanite	–	–	–
Opaque	0.58	0.19	1.76
Zircon	0.10	tr	–
Apatite	0.19	–	–
Allanite	–	0.19	–
Epidote/ Clinozoisite	–	–	–
Zoisite	–	–	–
Myrmekite	–	–	–
Sphene	–	–	–
Carbonate	–	–	–
Tourmaline	–	–	–
**Accessory	–	–	–
Total	100.00	99.99	99.99

* First digit(s) refers to sample number used in sandstone classification, Section IV. "TM" and last digits are actual field sample number.
** May include zircon, apatite, epidote, opaque.

TABLE A-2: Quartz, feldspar, and mica contents of metasandstone samples

Sample #	Location	Q+F+M (Total)	Q (Recalculated to 100%)	F	M	Biotite as % of Total M
49	Quarry-Abercrombie Property	98.10%	71	13	16	93
605	F.S.Rd 166 (to Wolfpen)	99.19%	43	15	42	73
2734	Creek Below Chimney Mt.	98.85%	74	8	18	57
2764	Beginning of Wildcat Cr. Rd.	98.30%	52	15	33	77
645	North Prong-Soque	98.77%	58	23	19	98
999	Top of Chimney Mtn.	98.81%	52	18	30	89
2737C	Tray Mtn Rd.	97.35%	44	24	32	100
882	Hanson Mtn.	76.43%	30	29	41	93
950A	Hwy 17 S. of Unicoi Gap	99.56%	47	32	21	93
955	App Trail, From Hwy 17	97.50%	7	21	72	45
2250	F.S. Rd Near Lost House Rd.	100.00%	59	12	29	72
2740	Near Indian Grave Gap	74.83%	43	13	42	100
2750A	Hwy 17 S. of Unicoi Gap	96.23%	27	31	42	58
162	Waterfall-Mill Cr. Rd.	98.20%	42	27	31	95
294	Sassafra Br.	98.82%	63	15	22	54
330	Sassafra Br.	97.32%	69	18	13	100
496	Nell Knob	94.70%	7	42	51	65
510	Corbin Cr. Rd.	98.29%	61	25	14	98
1028	Below Young Lick	96.73%	49	32	19	100
2744	Pheasant Branch	97.20%	57	24	19	100
134A	Quarry-Wildcat Cr. Rd.	97.80%	62	14	24	79
138B	Fuller Gap	98.20%	52	29	19	96
210	Wildcat Cr. Rd. Area	99.31%	49	18	33	89
237	Alice Br.	98.19%	39	41	20	100
2101	Mary Anne Br.	96.32%	53	28	19	98
370B	App. Trail to Roundtop	99.40%	49	12	39	95
558	Trail Near York Ridge N.	90.83%	25	30	45	98
709	App. Trail to Kelly Knob	96.82%	69	—	31	90
836	Below Rattlesnake Knob	99.21%	26	32	42	85
1099	Stream Near Addis Gap	99.25%	52	21	27	86
2005	Stream Near Addis Gap	96.34%	51	3	46	96
2770	Below Wolfstake Knob	99.04%	7	28	65	61
2120	Dismal Creek	92.54%	7	33	60	74
2166	Rd. By Lovell Br.	98.25%	40	32	28	61
2451	Ridge by Rocky Br. (Kelly R.)	99.60%	68	16	16	40
2762	Rice Bridge Rd.	99.10%	8	49	43	95
2763	Hwy 76, Bearmeat Gap	95.73%	3	19	78	42
2178A	To Corn Ridge	94.15%	23	19	58	51
2179	To Corn Ridge	96.52%	59	—	41	9
2180	To Corn Ridge	99.24%	10	35	55	64
2405	Corn Ridge	99.44%	36	16	48	90
2769	North of Deep Gap	90.24%	18	49	33	80
2728A	Joe Cove	99.61%	27	42	31	83
2724A	Joe Cove	99.13%	49	18	33	86
239	Alice Br.	97.63%	46	20	34	97

APPENDIX B

TABLE B-1: Mineral abbreviations used in text in discussion of metamorphic assemblages and reactions.

Abbreviation	Mineral Name
ab	albite
alm	almandine garnet
an	anorthite
AS	aluminum silicate
bio	biotite
cal	calcite
chl	chlorite
clino	clinozoisite
epi	epidote
gar	garnet
hbl	hornblende
ky	kyanite
mt	magnetite
micro	microcline
musc	muscovite
oligo	oligoclase
plag	plagioclase
qtz	quartz
sill	sillimanite
st	staurolite

VITA

Elizabeth A. McClellan was born in Donaldsonville, Georgia on April 16, 1953. At the age of seven, she moved with her family to Mobile, Alabama where she attended elementary and high schools. She was graduated from Theodore High School in June 1971, and received a Bachelor of Science degree in Biology, with a minor in Geology, from the University of South Alabama in 1980.

In 1982 she entered the University of Alabama at Birmingham to pursue a degree in Geology, and also worked part-time at Vulcan Materials Company. While at UAB, she was awarded the George F. Brockman Scholarship, and was voted "Outstanding Student of the Year" in the Geology department. She served as a teaching assistant for several terms. The Bachelor of Science degree in Geology was awarded in 1985.

During the summer of 1985, she attended The University of Tennessee's Geology Field Camp in Dayton, Tennessee. In the fall of that year she accepted a teaching assistantship at the University of South Carolina, Columbia and began study toward a Master's degree. When her advisor accepted a job at The University of Tennessee, Knoxville in 1986, she transferred to that school in order to continue work on her chosen project, serving as a teaching assistant in that department and for the Field Camp in 1986. After receiving the Master's of Science degree in Geology in August, 1988, Ms. McClellan plans to continue work toward a doctorate degree at The University of Tennessee.



Western Blue Ridge

- Pz — Paleozoic rocks
- thg — Hillabee Greenstone
- ttg — Talladega Gp.
- tsm — Sylacauga Marble Gp.
- tkm — Kahatchee Mountain Gp.

  (thg, ttg, tsm, tkm: Talladega Slate Belt (late PC–to Dev.))
- Ꞓc — Chilhowee Gp. (locally includes Shady or Rome Fms.)
- pCoc — Ocoee Supergroup, undivided
- pCw — Walden Creek Gp.
- pCgs — Great Smoky Gp.
- pCs — Snowbird Gp.

  (pCw, pCgs, pCs: Ocoee Supergroup)
- pCm — Murphy Gp.
- pCgm — Grandfather Mountain Fm.
- pCcg — Catoctin Greenstone
- pCmr — Mount Rogers Fm.
- pCb — Basement rocks

Eastern Blue Ridge

- Mz — Mesozoic rocks
- PzpCcg — Coweeta Gp.
- PzpCab — Alligator Back Fm.
- PzpCev — Evington Gp.
- pCot — Otto Fm.
- pCng — New Georgia Gp.
- pCtf — Tallulah Falls Fm.
- pCss — Sandy Springs Gp.
- pCa — Ashe Fm.
- pCly — Lynchburg Gp.
- pCb — Basement rocks
- Black = Mafic or ultramafic rocks
- Unlabeled areas = Paleozoic granites

PLATE I

REGIONAL GEOLOGIC M
WESTERN AND EASTERN E
SOUTHERN APPALACH

Modified from Hatcher and oth
Geology compiled by E.A. McCle

Hadley, 1970
Hadley and Nelson, 19
Rankin and others, 197
Tull and others, 1985
Wehr and Glover, 198
McConnel and others,
Hatcher and others, 1

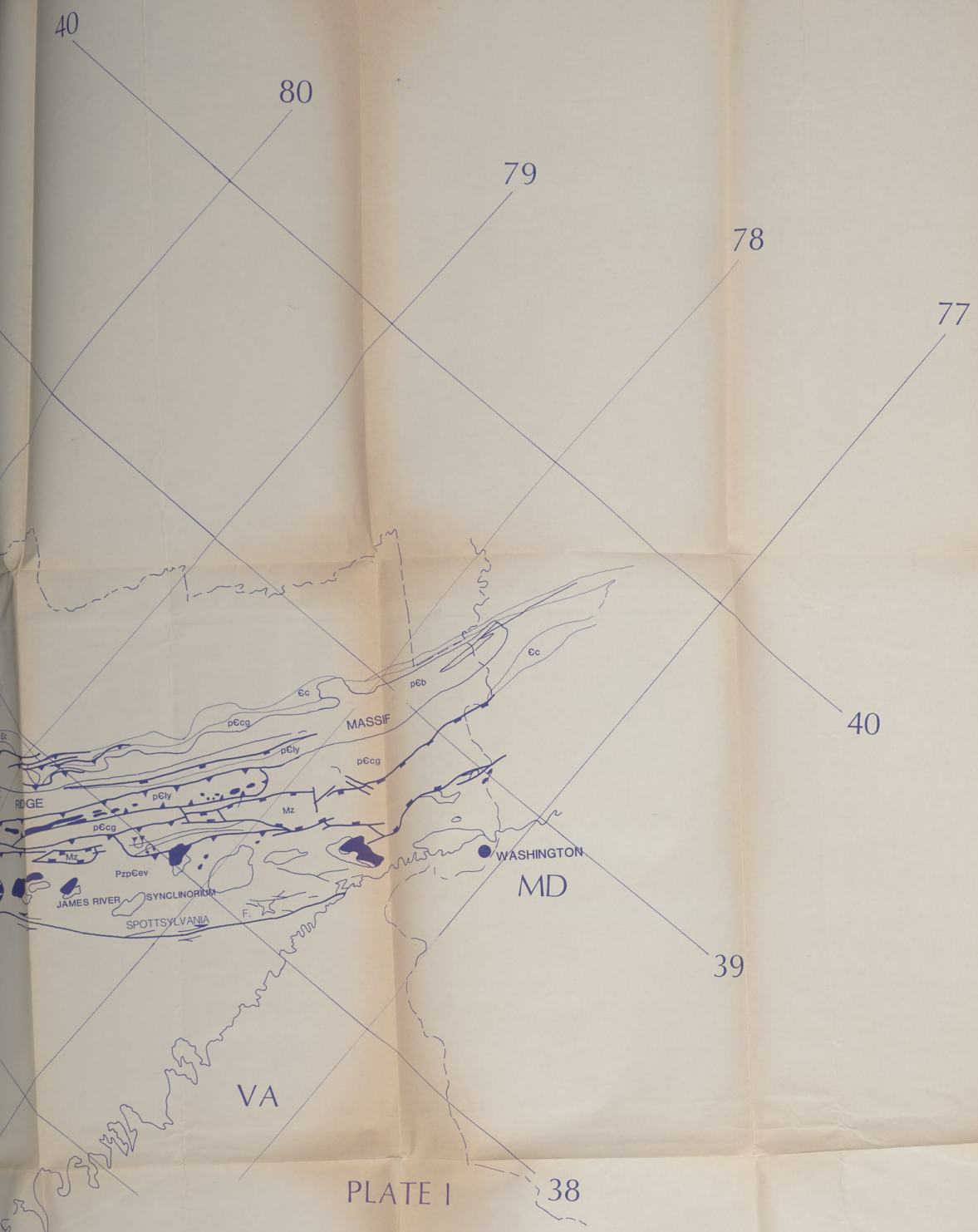


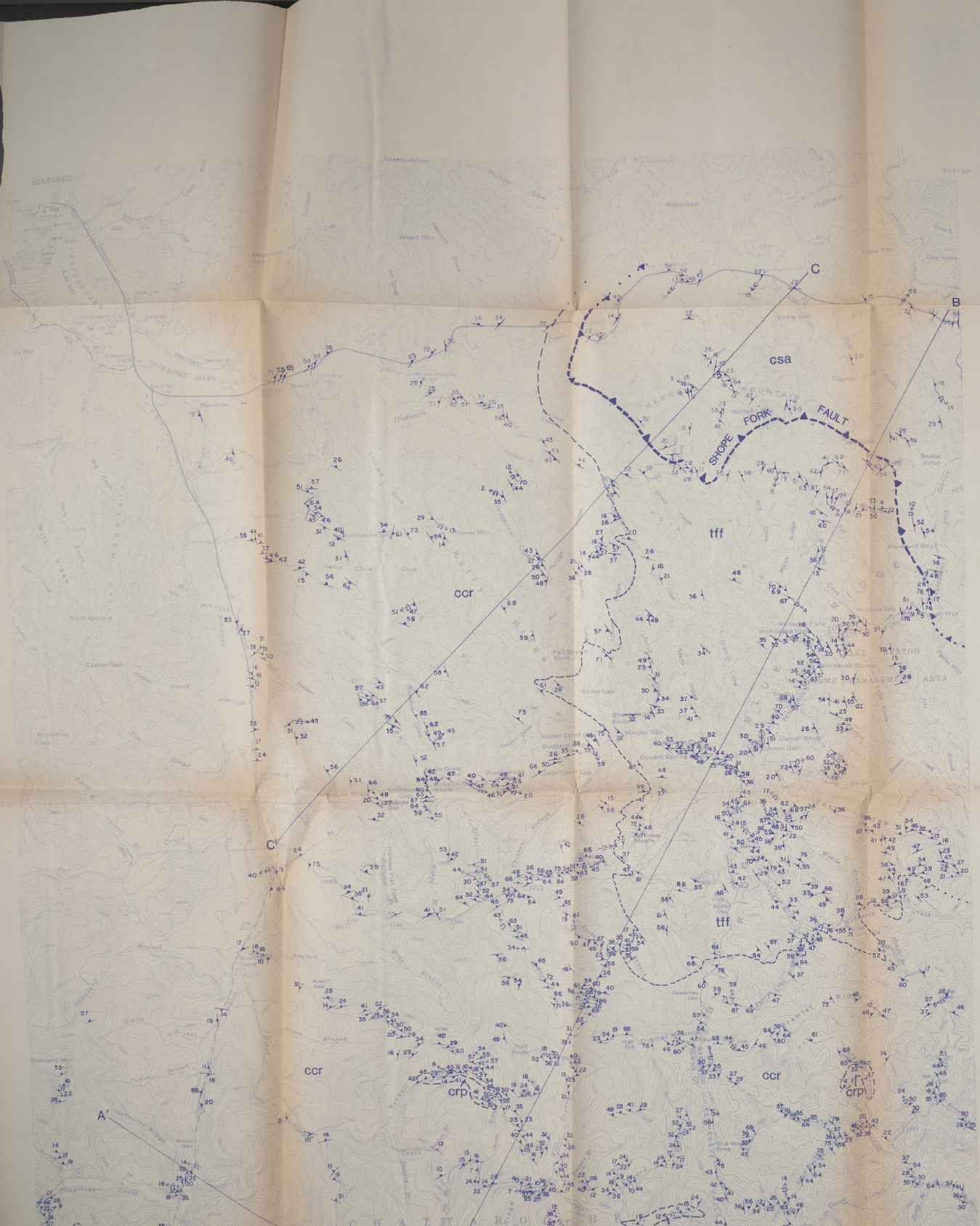
PLATE 1

REGIONAL GEOLOGIC MAP OF THE WESTERN AND EASTERN BLUE RIDGE, SOUTHERN APPALACHIANS

Modified from Hatcher and others, 1988.

Geology compiled by E.A. McClellan, from:

Hadley, 1970
Hadley and Nelson, 1971
Rankin and others, 1973
Tull and others, 1985
Wehr and Glover, 1985
McConnel and others, 1986
Hatcher and others, 1987



EXPLANATION

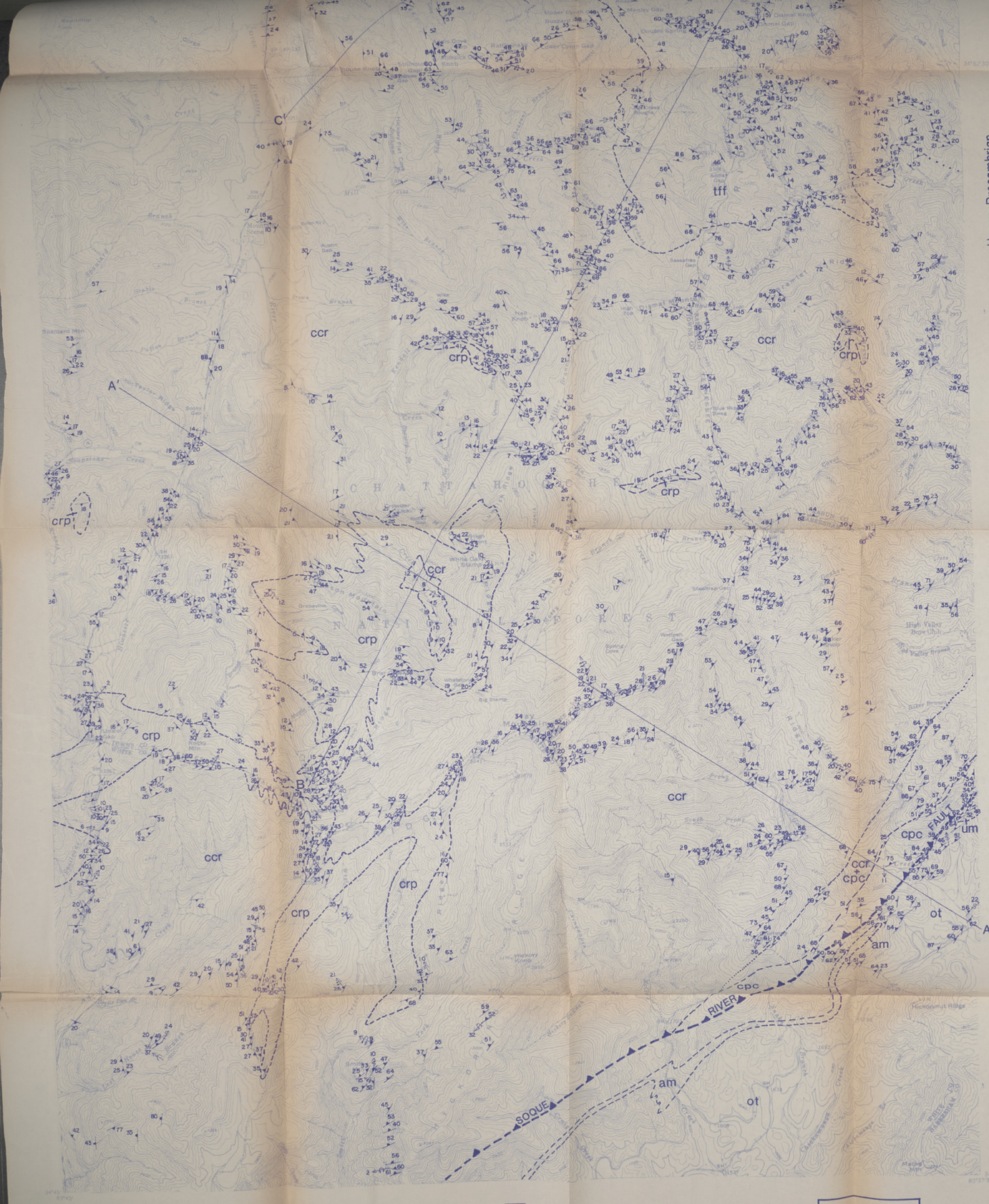
Upper Precambrian or Lower Paleozoic

Coweeta Group

- crp — Ridgepole Mountain Formation
- ccr — Coleman River Formation
- cpc — Persimmon Creek Gneiss
- tff — Tallulah Falls Formation(?)

Upper Precambrian(?)

- um — Ultramafic rocks
- ot — Otto Formation
- am — Amphibolite
- csa — chlorite garnet schist and amphibolite



EXPLANATION

Upper Precambrian or Lower Paleozoic

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Contacts:

Lithologic — exact — approximate - - - inferred . . . .

Thrust Faults — exact — approximate — inferred

teeth on hanging wall

Structural Symbols:

- 30 — Strike and dip of foliation
- 10 — Trend and plunge of lineation
- 15 — Trend and plunge of mesoscopic flexural antiform
- 15 — Trend and plunge of mesoscopic flexural synform
- 45 — Trend and plunge of mesoscopic flowage antiform
- 45 — Trend and plunge of mesoscopic flowage synform
- 20 — Trend and plunge of crenulation axis

Scale 1:24,000

0 1 mile

0 1000 0

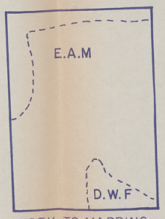


Plate II

GEOLOGY OF PORTIONS OF THE TRAY MOUNTAIN AND MACEDONIA 7½' QUADRANGLES, GEORGIA

by

Elizabeth A. McClellan



INDEX TO MAPPING
E.A.M Elizabeth A. McClellan
D.W.F David W. Farrell

Contacts:

Lithologic — exact — approximate — inferred

Thrust Faults — exact — approximate — inferred
teeth on hanging wall

Structural Symbols:

30 Strike and dip of foliation

10 Trend and plunge of lineation

15 Trend and plunge of mesoscopic flexural antiform

15 Trend and plunge of mesoscopic flexural synform

45 Trend and plunge of mesoscopic flowage antiform

45 Trend and plunge of mesoscopic flowage synform

20 Trend and plunge of crenulation axis

Scale 1:24,000

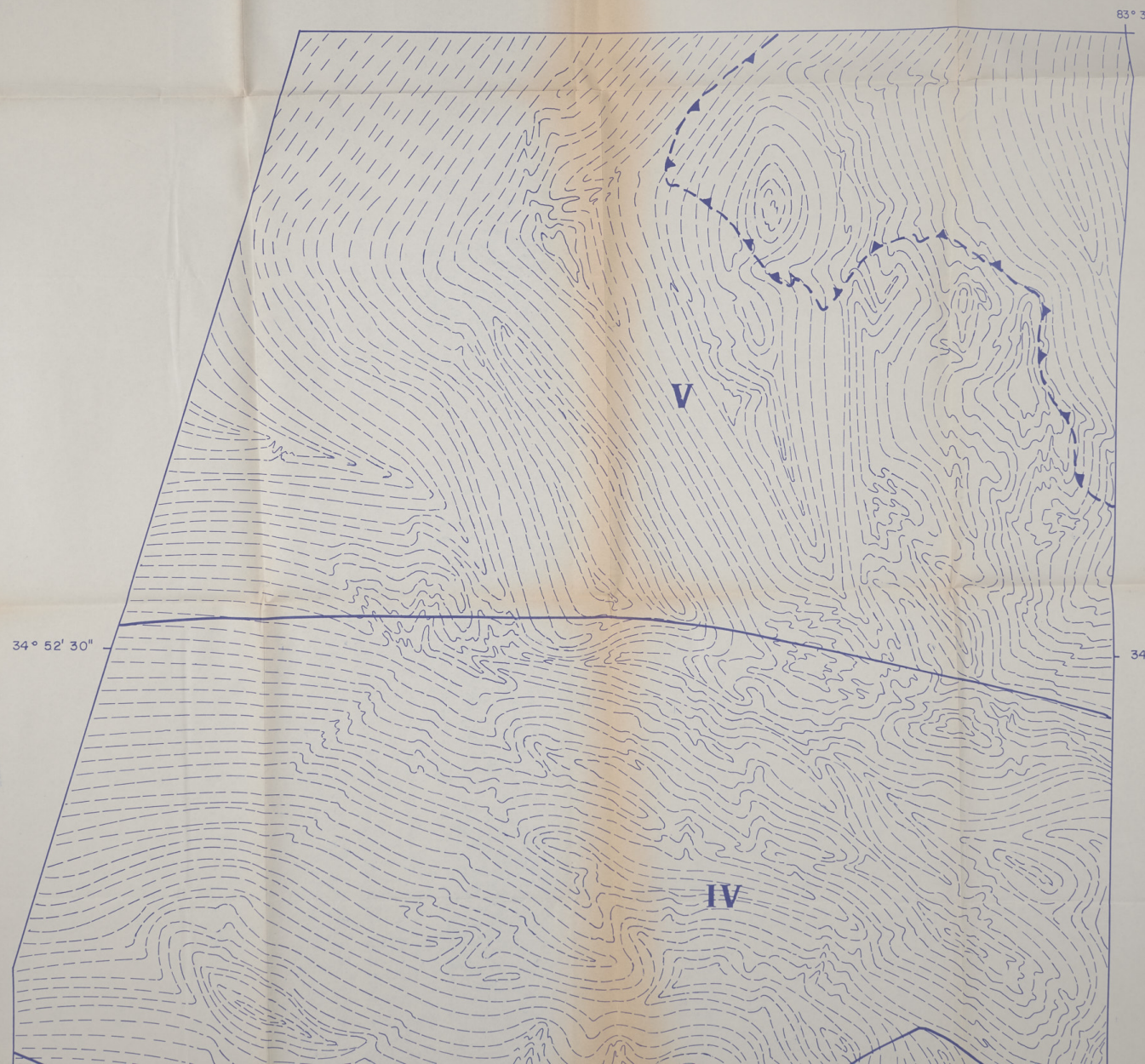
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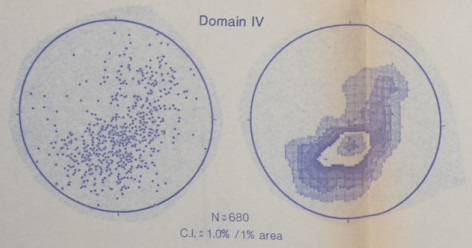
1000 0 7000 feet

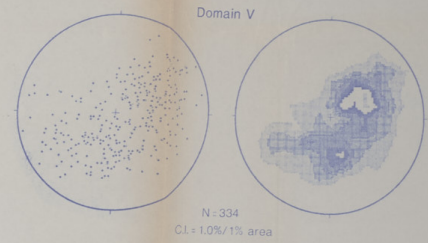
1 5 0 1 kilometer

PLATE III

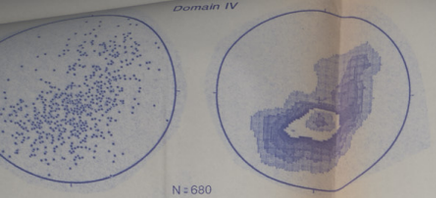
FORM SURFACE MAP FOR DOMINANT FOLIATION (S_2) IN STUDY AREA INCLUDING ORIENTATION DATA FOR S_2 IN INDIVIDUAL DOMAINS





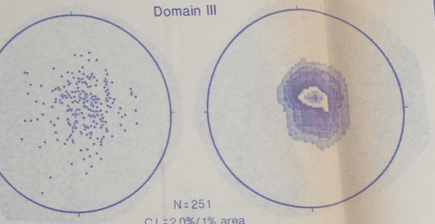


Domain IV



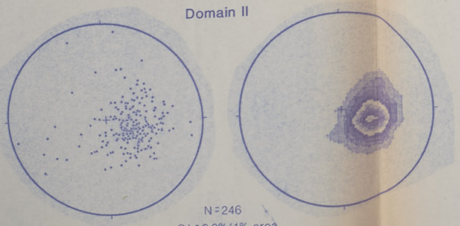
N = 680
C.I. = 1.0% /1% area

Domain III

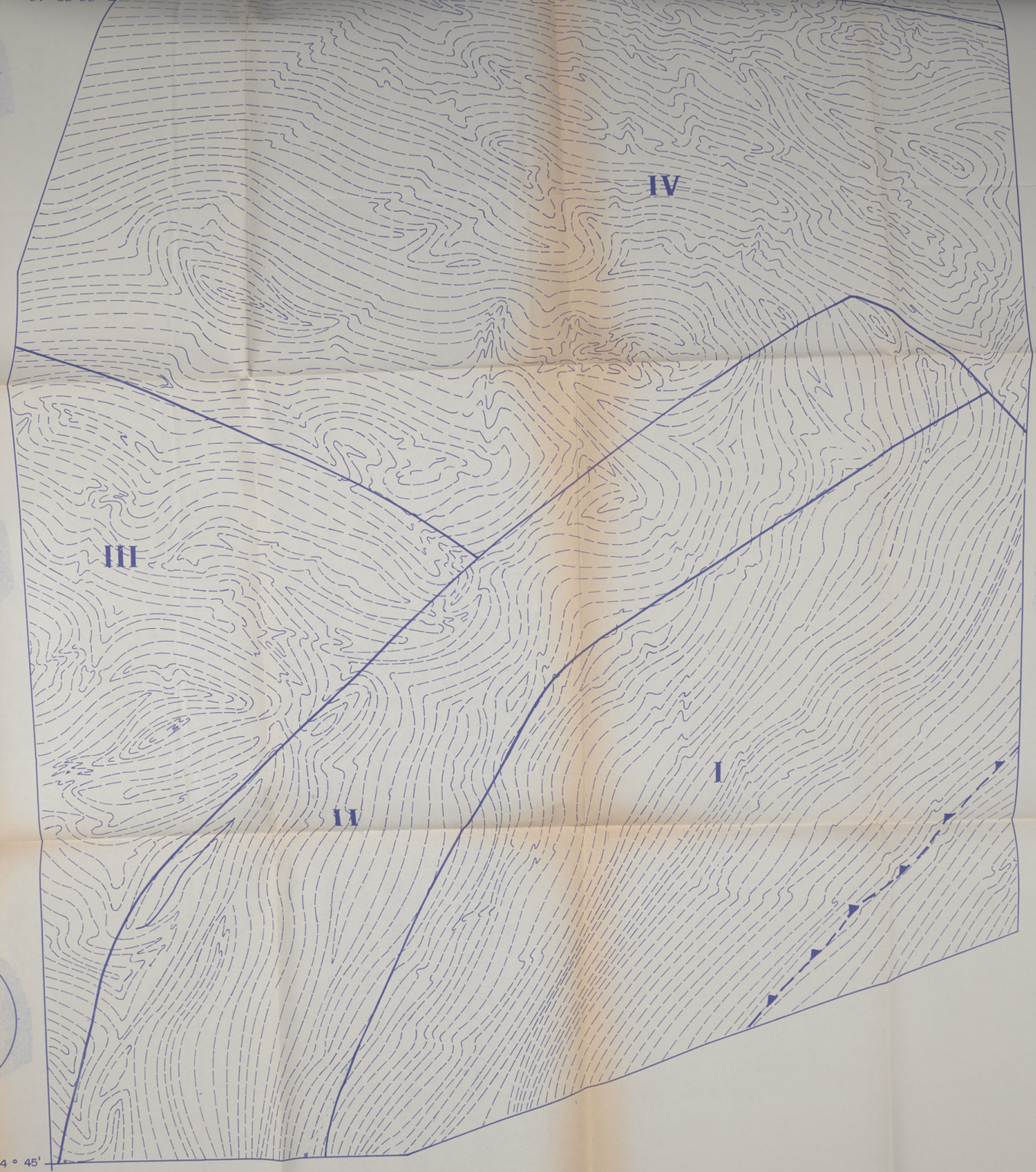


N = 251
C.I. = 2.0%/1% area

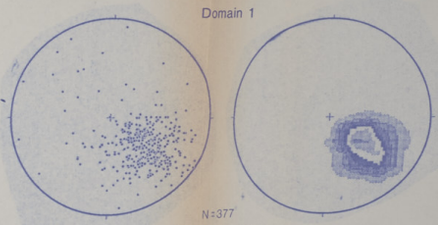
Domain II



N = 246
C.I. = 2.0%/1% area



Domain I



N = 377
C.I. = 2.0%/1% area

Fabric diagrams plotted and contoured using Stereonet, v. 2.8,
a plotting program for Macintosh computer (Allmendinger, 1987).

SCALE 1:24,000

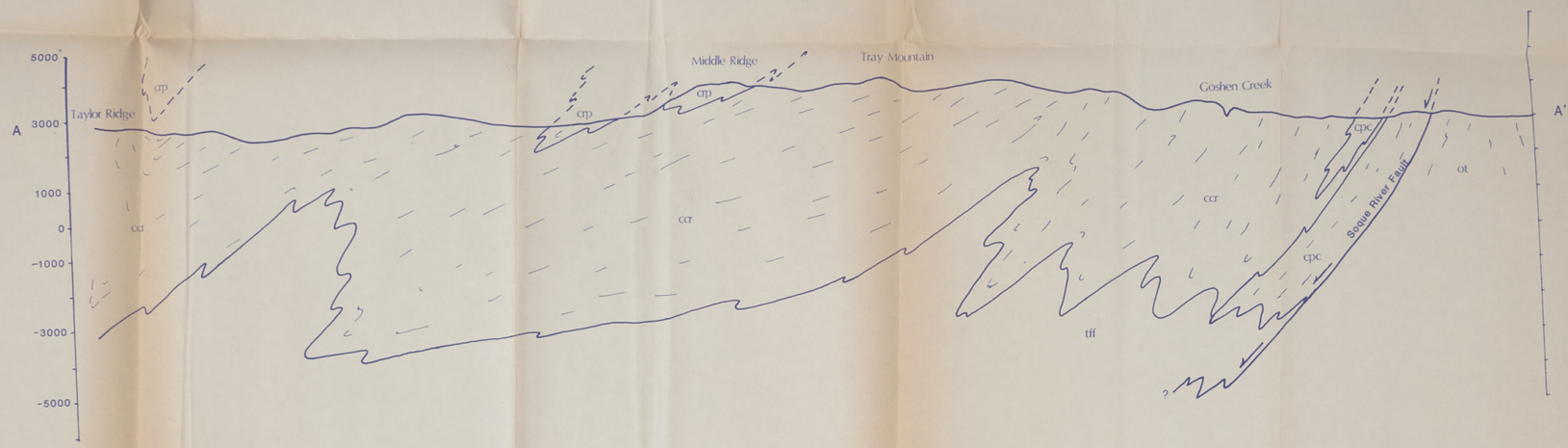
0 2
miles

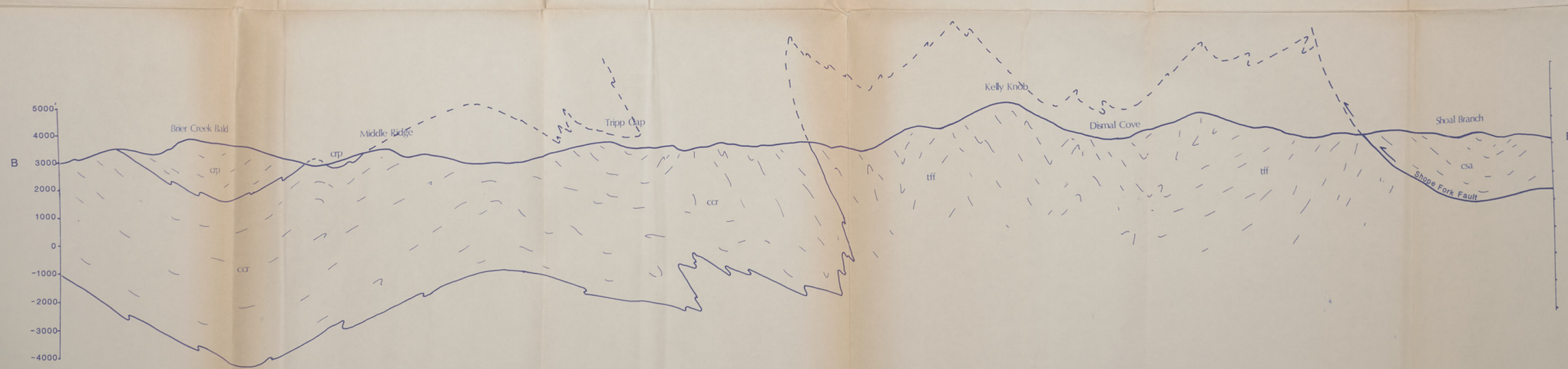
0 2
kilometers

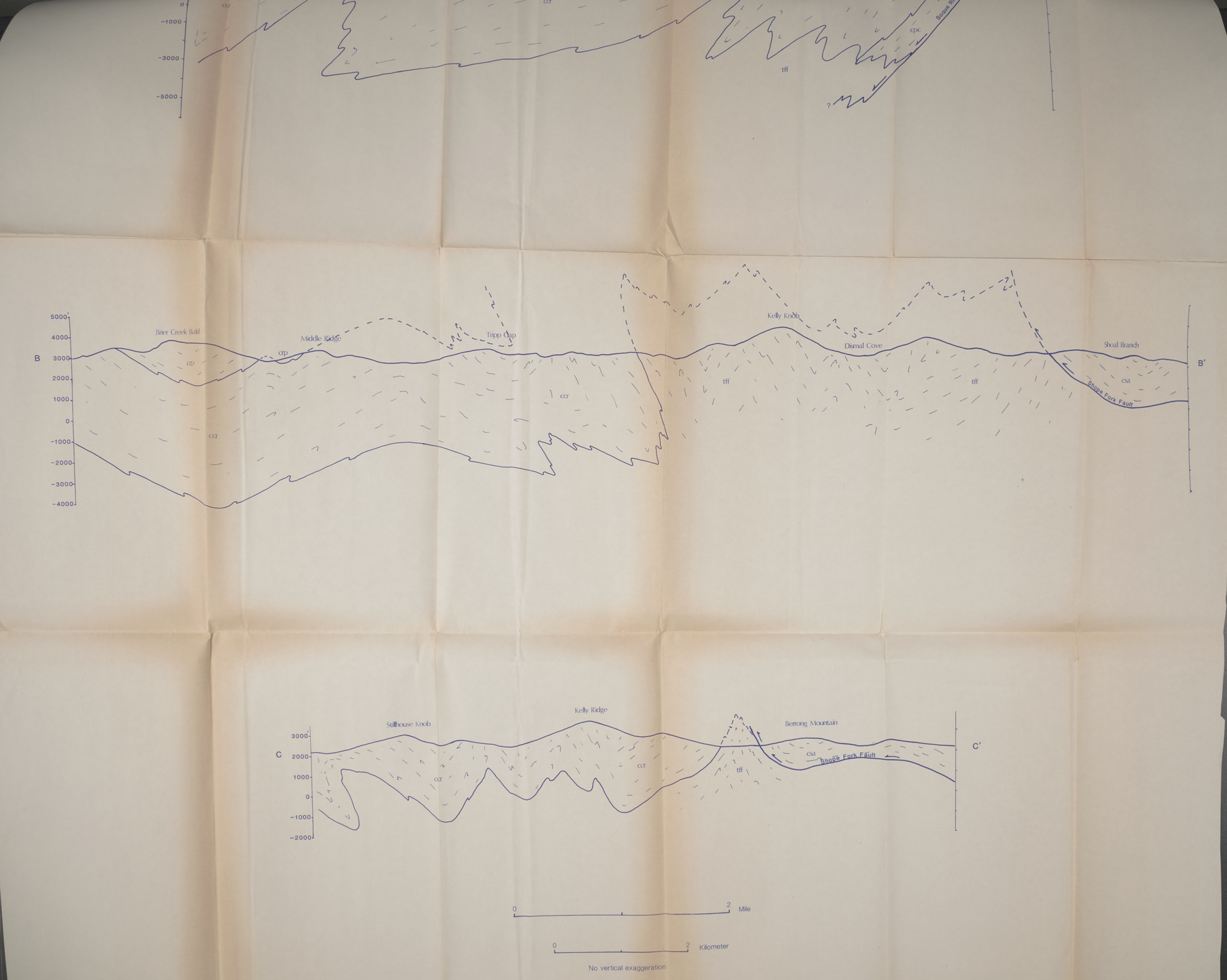
PLATE IV

CROSS SECTIONS ALONG LINES A-A', B-B', AND C-C', ON PLATE II.

(Rock units same as Plate II.)






0
-1000
-3000
-5000
ccr
ccr
tff
cpc
Sopue R
5000'
4000
3000
2000
1000
0
-1000
-2000
-3000
-4000
B
B'
Brier Creek Bald
Middle Ridge
Tripp Gap
Kelly Knob
Dismal Cove
Shoal Branch
cp
crp
ccr
tff
tff
csa
Shope Fork Fault
C
C'
Stillhouse Knob
Kelly Ridge
Berrong Mountain
3000
2000
1000
0
-1000
-2000
ccr
ccr
tff
csa
Shope Fork Fault
0
2
Mile
0
2
Kilometer
No vertical exaggeration