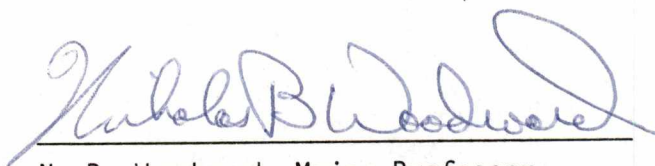


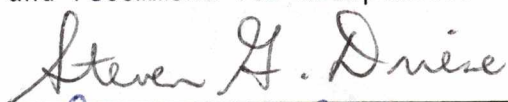
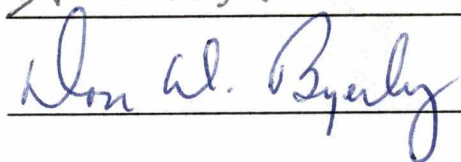
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STRUCTURAL ANALYSIS OF THE HUNTER VALLEY, WALLEN VALLEY,
AND WHITEOAK MOUNTAIN FAULT INTERSECTION
IN NORTHEASTERN TENNESSEE

A Thesis
Presented for the
Master of Science
Degree
The University of Tennessee, Knoxville

Jerry Wayne Beets

June 1985

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ABSTRACT

Detailed mapping at 1:12,000 scale was done in the area of the Hunter Valley, Wallen Valley, and Whiteoak Mountain thrust fault intersection in Anderson County, northeastern Tennessee. Strain analysis was done on hangingwall and footwall rocks using mesoscopic Rome Formation and Rockwood Formation folds, deformed Rockwood Formation brachiopods, a deformed Rome oolite, and balanced cross-sections.

The structural data indicate that (1) the Wallen Valley thrust is a footwall imbricate of the Hunter Valley thrust, (2) the Wallen Valley thrust transfers its displacement to the older Hunter Valley thrust and together they are renamed the Whiteoak Mountain thrust, (3) more intense macroscopic folding (shortening up to 29%) occurs in the hangingwall of the older Hunter Valley thrust due to folding by the younger Wallen Valley thrust below, (4) both the Hunter Valley and the Wallen Valley thrusts are folded by a footwall horse of Ordovician rocks along the Wallen Valley footwall flat, and (5) footwall deformation via trailing imbricates and mesoscopic folding is greater than hangingwall deformation via leading imbricates and mesoscopic folding.

The data from strain analysis indicate that (1) whole rock strains in the Rome Formation oolite (due to stratigraphic overburden) are greater than the strains which are recorded by the individual ooids, (2) strain markers (Rockwood Formation brachiopods) which are proximal to large displacement thrust faults show greater deformation

than strain markers (Rome oolite) which are proximal to small displacement thrust faults, and (3) the displacement and shortening on the Whiteoak Mountain thrust (9.87 km and 76%) is greater than the displacement and shortening on the Hunter Valley thrust (2.44 km and 53%) or the Wallen Valley thrust (8.5 km and 59%).

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CHAPTER I

INTRODUCTION

Reasons for the Study and Objectives

The aims of this study are to determine the geometry of the area of fault convergence, the timing for the emplacement of the different thrusts, the strains taken up by the hangingwall and footwall of each thrust sheet, and the amount of hangingwall and footwall duplication of stratigraphic units. Mapping at 1:12,000 scale is needed because previous mapping in the area was not sufficiently detailed and previous structural interpretations did not explain the geometry of the area of fault convergence. Two examples of curious map patterns are the outcrop widths of Rome in the hangingwall of the Hunter Valley thrust and Rockwood in the footwall of the Wallen Valley thrust. These outcrop widths are too wide given the published structural orientations and stratigraphic thicknesses of the units. One objective of this study is to look for evidence of Rome and Rockwood duplication via thrust faults. Although Rodgers (1953) speculated that major thrust duplication exists within the Rome of the Hunter Valley thrust sheet, previous mapping at 1:24,000 scale in the area found only minor duplication. Another unexplained map pattern is a slice of fault-bound Ordovician age rocks, that was mapped (Keith, 1896; Hardeman, 1966) northwest of the Whiteoak Mountain thrust within the Clinton quadrangle.

Until this study, the slice has never been related to the fault convergence northeast along strike.

The timing for thrust emplacement in the study area was first discussed by Harris and Milici (1977). They proposed a west-to-east sequence of younger thrusts overriding and burying older thrusts in order to explain the apparent truncation of the Wallen Valley thrust by the Hunter Valley and Whiteoak Mountain thrust faults. In contrast, Roeder et al. (1978) proposed an east-to-west sequence of younger thrust faults propagating from the footwall of older thrusts and folding these older thrusts above. Roeder et al. (1978) used a polyphase piggyback deformation theory to explain the thrust systems which Milici (1975) explained using the break-back theory. The testing of these two conflicting kinematic theories is another objective of this study. A final objective is to compare deformation intensities within the hangingwall and footwall rocks of the major thrust faults, using several strain analysis techniques.

Methods Used in the Study

The structure of the Valley and Ridge in Tennessee (Figure 1) is dominated by many sledrunner thrust faults, which bring the Cambrian Rome Formation to the surface in the hangingwall (Roeder, Gilbert, and Witherspoon, 1978). This thesis: (1) describes the microfabric, mesofabric, and macrofabric (thin section, outcrop, and map scale) deformation structures in the study area; (2) determines the timing of deformation; and determines the strain intensities and distributions. The study involved 1:12,000 scale mapping of the hangingwall and footwall

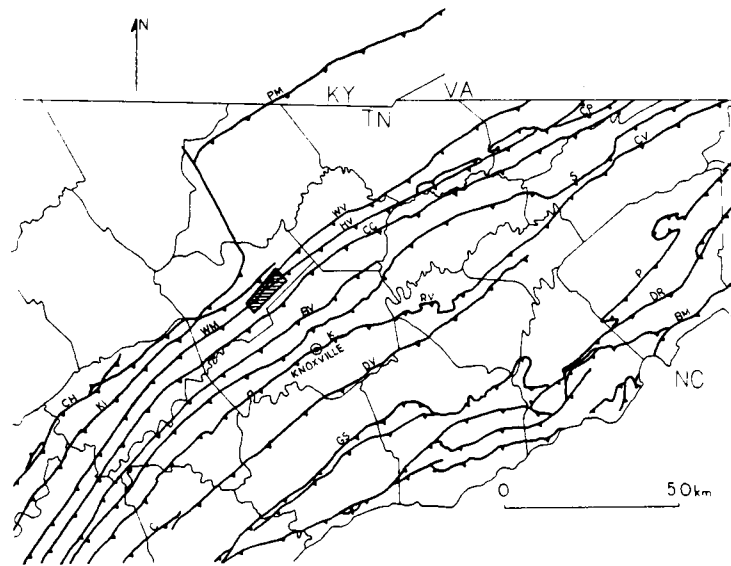


Figure 1. Geologic Map of East Tennessee and Location of the Study Area (after Rodgers, 1953; Roeder et al., 1978).

CH = Chattanooga thrust, KI = Kingston thrust, WM = Whiteoak Mountain thrust, PM = Pine Mountain thrust, WV = Wallen Valley thrust, HV = Hunter Valley thrust, BV = Beaver Valley thrust, CC = Copper Creek thrust, CP = Clinchport thrust, S = Saltville thrust, K = Knoxville thrust, RV = Rocky Valley thrust, DV = Dumplin Valley thrust, CV = Carter Valley thrust, GS = Great Smoky thrust, P = Pulaski thrust, DR = Dunham Ridge thrust, and BM = Buffalo Mountain thrust.

rocks of the Hunter Valley, Wallen Valley, and Whiteoak Mountain thrust faults in the area where these three faults merge in northeastern Tennessee. The specific aims of the study are to determine the geometry of the area of fault convergence, the timing of emplacement for the different thrusts, the strains taken up by the hangingwall and footwall rocks, and the amount of hangingwall and footwall duplication via thrust faulting (Plates 1-3). Mesoscopic folds and faults were studied in order to determine the deformation style and to derive information about the larger scale structures within each thrust sheet (Price, 1976). This mesofabric analysis involved detailed study of folds and faults in roadcut sections along I-75N and US-25W and thin section analysis of samples from these roadcuts. Deformed oolitic dolostones and fossiliferous limestones were subjects for strain analysis using the center-to-center, R_f/ϕ , and Wellman techniques (Fry, 1979; Dunnet, 1968; Wellman, 1962). Mesoscopic folds along roadcuts were also unfolded in order to determine bedlength shortening strains (Ramsay, 1969).

The results of this study are presented in two sections. Macroscopic, mesoscopic, and microscopic deformation structures and styles are discussed in section one. These observed structures are used to determine the geometry of the area of fault convergence, the timing for the emplacement of the different thrusts, and the amounts of footwall and hangingwall duplication. The second section deals with strain analysis of rocks in the study area. Different strain analysis techniques used are: (1) balanced cross sections; (2) Wellman's method for deformed fossils with original bilateral symmetry; (3) the Fry center

to center and Dunnet Rf/Ø methods for deformed originally spherical objects, and (4) the Ramsay method for unfolding mesoscopic folds.

Previous Work

Many previous workers have studied the geology and stratigraphy of the rocks in the Clinton, Powell, and Norris quadrangles. The earliest mapping was done at 1:125,000 scale by Arthur Keith (1896) for the U.S.G.S. Briceville folio. General reconnaissance mapping of the Powell and Norris quadrangles at 1:24,000 scale was done by Sniegocki (1949) and Lessig (1949). Their mapping very closely agreed with the geology of Keith (1896). Rodgers (1953) field checked the area and used soil maps to determine the geology compiled in his 1:125,000 scale Briceville Sheet. More detailed mapping of the Powell and Norris quadrangles at 1:24,000 scale was done by Murray (1960) and Johnson (1964). Their mapping closely agreed with the geology of Rodgers (1953). Swingle (1964) mapped the Clinton quadrangle at 1:24,000 scale for the U.S.G.S. and the Tennessee Division of Geology. All these studies attempted to define the geometry of the Hunter Valley, Wallen Valley, and Whiteoak Mountain faults in or near the area of convergence.

Problems with Previous Work and New Structural

Interpretations in This Thesis

The most recent mapping of the Powell, Norris, and Clinton quadrangles by Murray (1960), Johnson (1964), and Swingle (1964) presented problems in stratigraphy, stratigraphic thicknesses, and structural geometries. Murray (1960) mapped a series of slices

northwest of Eagle Bend which placed Ordovician Mascot and Silurian Rockwood on Cambrian Rome. This map pattern resulted in younger rocks being thrust over older rocks. The author believes that Murray mapped shales and siltstones of the lower Rockwood Formation as the lower Rome Formation. Mapping by the author indicates that the lower Rockwood is thrust over the upper Rockwood by a trailing imbricate of the Wallen Valley thrust (Plates 1-3). Murray also did not map the footwall horse of Mascot and Chickamauga on the southeast side of Lone Mountain near the Clinch River. He interpreted the dolomite and massive, brecciated, chert balls of the Mascot Dolomite and fossiliferous shaley limestones of the lower Chickamauga as lower Rome. A final problem with Murray's geology involves the great stratigraphic thicknesses which he used for the Rome, Sequatchie, and Rockwood Formations. Murray, after Rodgers (1953), mapped a high-angle reverse fault on the southeastern side of Lone Mountain which brought 137 m (450 feet) of Sequatchie to the surface in the hangingwall. This is 46 m more Sequatchie than the author measured along I-75N. The author has found Rockwood sandstones at the top of Lone Mountain and a thrust fault which places Sequatchie on Rockwood (Plate 1). Murray used improper thicknesses of Rockwood (198 m or 650 feet) and Rome (671 m or 2200 feet) in his structure sections, which curiously showed the Hunter Valley and Whiteoak Mountain thrusts cutting downsection forwards. The author explains the wide outcrop width of the Rockwood Formation by mapping several trailing imbricates of the Wallen Valley thrust which duplicate sections in the Lone Mountain and Pumpkin Hollow areas of Plate 1. These trailing imbricates allow the author

to maintain a Rockwood stratigraphic thickness of 152 m (500 feet). The author uses a Rome thickness of 457 m (1500 feet) in the hanging-wall of the Hunter Valley thrust. This thickness was measured along I-75N. Structural data indicate that the Hunter Valley thrust was folded by the younger Wallen Valley thrust below. A footwall horse of Mascot and Chickamauga below the Wallen Valley footwall flat folded the older thrusts above (Plates 1-3).

Johnson (1964) mapped the southern portion of the Norris Quadrangle. The four problems with Johnson's work are insufficient structural data, incorrect stratigraphy, incorrect stratigraphic thicknesses, and curious structural geometries. Johnson showed very few strikes and dips on his map for the Rockwood Formation and Conasauga Shale. His insufficient data allowed only inferred mapping of the trace of the Hunter Valley and Wallen Valley thrusts. The author has made a different interpretation in mapping these thrusts (Plate 1) and changed the position of contacts. Johnson (1964, p. 37) attempted to map the upper Silurian Sneedville limestone southwest of Norris Freeway (US 441). Although the fossiliferous Sneedville limestone and overlying Mississippian-Devonian Chattanooga Shale occurred together northeast of Norris Freeway, Johnson was unable to map the Chattanooga Shale southwest of Norris Freeway. He believed that the overlying Chattanooga Shale was completely faulted out by the Wallen Valley thrust as the fault cut downsection forwards. The author believes that the dolomitic limestones that Johnson mapped near Pumpkin Hollow are actually upper Rome limestones, which can be mapped for 5.5 km along strike in the hangingwall of the Wallen Valley thrust and

correlated to the hangingwall of the Hunter Valley thrust. These limestones are overlain and underlain by maroon Rome sandstones, siltstones, and shales and were used by the author to separate the Rome Formation into upper and lower portions. A final problem with Johnson's work was the incorrect stratigraphic thicknesses of Rockwood (457 m or 1500 feet) and Rome (762 m or 2500 feet) which he used in drawing his structure section. The author explains the wide outcrop width of the Rockwood Formation by mapping three trailing imbricates of the Wallen Valley thrust, which duplicate the Rockwood (Plates 1-3). Several unmapped folds in the hangingwall of the Hunter Valley thrust indicate that the Hunter Valley was folded by the younger Wallen Valley thrust below and explain the wide outcrop pattern of Rome given a stratigraphic thickness of 457 m (1500 feet).

Swingle (1964) mapped the Clinton quadrangle and produced a map that was basically correct. The two problems with Swingle's work are a curious structural geometry in his cross sections and his mapping of the trace of the Whiteoak Mountain fault. The author has adjusted the Whiteoak Mountain fault contact between the lower Rome and lower Chickamauga based on a persistent line of sinkholes and abundant fossiliferous Chickamauga limestone float. Mapping at 1:12,000 scale allowed the author to map several folds in the Rockwood Formation and a small displacement thrust in the upper Rome Formation. These structures are not shown on Swingle's 1:24,000 scale map. Swingle drew the Whiteoak Mountain thrust using a geometry which made the fault cut downsection forwards. The author makes a new interpretation by extending the footwall horse beneath the Whiteoak Mountain thrust and by having the younger

horse fold the Whiteoak Mountain thrust above. This geometry allows the author to maintain a maximum Rome thickness of 457 m (1500 feet).

Stratigraphy

Introduction

There are approximately 2700 m of Cambrian to Silurian age rocks in the study area. The different formations are shown in the generalized stratigraphic column in Figure 2. Thicknesses of the Rome, Sequatchie, and Rockwood Formations were measured along I-75N roadcuts. Only partial thicknesses of the Rome and Rockwood Formations were measured along US-25W. The thicknesses of all other formations are based on published data by Rodgers (1953), Swingle (1964), and Borowski (1982). Detailed lithologic descriptions and the depositional environments for the stratigraphic units are given below.

Rome Formation

The Cambrian Rome Formation has been called "a heterogeneous mixture of sandstone, siltstone, shale, limestone, and dolomite" (Rodgers, 1953, p. 43). It is divided into lower (€rl) and upper (€ru) portions by the author. The boundary between the two portions is drawn at the base of the lowest carbonate unit. Red, yellow, maroon, brown, gray, and olive-green shales make up most of the lower Rome Formation with the rest composed of thin bedded sandstones and siltstones. The sandstones and siltstones of the lower Rome are commonly laminated to thin bedded, fine grained, and hematitic. These sandstones and siltstones are maroon, brown, gray, and yellow in color. Bioturbation and

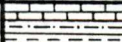
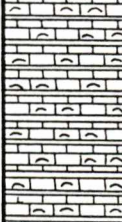
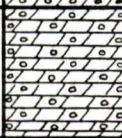
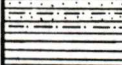
AGE	FORMATION	THICKNESS (meters)	COLUMNAR SECTION	DESCRIPTION	
SILURIAN	ROCKWOOD FORMATION	152		Sandstone, siltstone shale, and limestone	
ORDOVICIAN	SEQUATCHIE FORMATION	91		Siltstone, mudstone, and limestone	
	REEDSVILLE SHALE	61-76		Calcareous shale	
	CHICKAMAUGA GROUP	549		Limestone, shale, and oolitic chert	
	KNOX GROUP	MASCOT DOLOMITE	168		Cherty siliceous dolomite
		KINGSPORT FORMATION	107		Cherty siliceous dolomite
		LONGVIEW FORMATION	76		Very cherty dolomite
		CHEPULTEPEC DOLOMITE	213		Very cherty sili- ceous dolomite
CAMBRIAN	COPPER RIDGE DOLOMITE	305		Cherty siliceous dolomite	
	CONASAUGA GROUP	MAYNARDVILLE LIMESTONE	122		Shaley limestone
		CONASAUGA SHALE	366		Siltstone, shale, and limestone
	ROME FORMATION	UPPER ROME	274-335		Sandstone, siltstone shale, limestone, and dolomite
		LOWER ROME	122-183		Shale, siltstone, and sandstone

Figure 2. Generalized Stratigraphic Column for Rocks Exposed in the Study Area.

flaser bedding are commonly observed. The base of the lower Rome Formation is everywhere in fault contact with younger lithologies.

The dominant lithologies in the upper Rome are maroon sandstone and siltstone. Other lithologies present, in order from more to less common, are shale, limestone, and dolomite. Greenish-gray, yellowish-gray, yellow, pale brown, and olive-green sandstones are found interbedded with maroon sandstones in the upper Rome. These silica- and hematite-cemented sandstones are mostly fine-grained and quartzose. They may also be glauconitic and limonitic. The bedding thicknesses range from very fine laminations to medium beds. Ripple marks, small-scale unidirectional cross-beds, bidirectional cross-beds, flaser bedding, horizontal burrows, vertical burrows, mudcracks, and lenticular bedding are all common in the upper Rome sandstones. The upper Rome siltstones are yellow, gray, and greenish-gray. They are laminated to thin bedded and exhibit the same sedimentary structures as the sandstones. Shales in the upper Rome Formation are olive-green, pale brown, yellow, greenish-gray, and maroon in color. They are thin-bedded and commonly contain lenses of sandstone and siltstone. Upper Rome limestones are thin- to thick-bedded, light gray, and vuggy or honeycombed in appearance. The upper Rome dolostones are gray and olive-gray when fresh and weather to a grayish-yellow color. They are fine-grained and thin-bedded to laminated. A dolomitic oolite within the upper Rome is cross-bedded, hematitic, and glauconitic.

The Rome Formation has been interpreted as a tidal flat sequence, which becomes more clastic and coarser to the northwest (Samman, 1975). This pattern is indicative of deposition during a marine transgression

with a northwestern source. The Rome is dominated by sandstone and siltstone to the northwest and shale, limestone, and dolomite to the southeast. Sedimentary structures and trace fossils observable in outcrops within the study area were described by Samman (1975) and Spigai (1963). A detailed petrographic description of the Rome in Anderson County was given by Siribhakdi (1976). He estimated the depth of Rome burial by overlying stratigraphic units was greater than 6 km, due to the presence of sutured quartz grains.

Conasauga Group

The Conasauga Group is divided into lower clastic, (Ec) and upper carbonate (Emn) dominated portions within the study area. This division of the Conasauga Group into the lower Conasauga Shale (Ec) and upper Maynardville Limestone (Emn) is based on the division used by Rodgers (1953, pp. 47-49) and Swingle (1964). The Conasauga shale, from bottom to top, is made up of the Pumpkin Valley Shale, Rogersville Shale, and Nolichucky Shale. The boundary between the Pumpkin Valley Shale and the underlying upper Rome Formation is drawn at the top of the uppermost medium-bedded yellow-gray sandstone. Gray, olive green, yellow, and brown siltstones and shales are the dominant lithologies in all portions of the Conasauga Shale. These siltstones and shales are thin-bedded, fine- to medium-grained, quartzose, and limonitic. Scattered small lenses of gray limestones can be found in the Rogersville Shale and Nolichucky Shale. These limestones were interpreted as tongues of Rutledge Limestone and Maryville Limestone, which become thicker to the southeast (Rodgers, 1953).

Milici et al. (1973) interpreted the Conasauga Group as a deep-water lagoonal sequence to the west and a shallow-water carbonate shelf sequence to the east, which formed due to gradual basin subsidence. This environment produced the intertonguing of shales and limestones in the central phase, the shale-dominated northwestern phase, and the carbonate-dominated southeastern phase (Rodgers, 1953). Milici et al. (1973) interpreted these changing lithologies as indicative of a northwestern source.

Knox Group

The Knox Group in the study area is composed chiefly of thick-bedded siliceous dolomite, which produces abundant residual chert. Outcrops of the Knox Group are few, and therefore Rodgers (1953) and Swingle (1964) used residual chert and topographic expression to map the different formations. The Knox Group, from base to top, includes the Copper Ridge Dolomite (Ecr), Chepultepec Dolomite (Oc), Longview Dolomite (Olv), Kingsport Formation (Ok), and Mascot Dolomite (Oma). The thicknesses of these formations are given in Figure 2. Only the lower Knox Group, Copper Ridge Dolomite and the upper Knox Group, Mascot Dolomite crop out in the study area. Descriptions of the other formations of the Knox Group are based on those published by Rodgers (1953, pp. 55-61) and Swingle (1964).

Copper Ridge Dolomite. The Copper Ridge Dolomite is composed of medium- to thick-bedded, dark gray to brownish-gray, crystalline dolomite and light gray, finely- to medium-crystalline, siliceous dolomite. It is darker colored than the rest of the Knox Group due to a

high asphaltic content. Dark brown to black, dull luster, oolitic chert nodules are common in the lower and middle portions of the Copper Ridge Dolomite. Light brown, medium-grained, oolitic chert nodules are found in the upper portions of the Formation. The basal contact with the underlying Maynardville Limestone (6mn) of the upper Conasauga Group is drawn at the top of the lowest non-cherty, light gray, silty dolomite.

Chepultepec Dolomite. The Chepultepec Dolomite is a medium-bedded, light gray to light brown, finely- to medium-crystalline, siliceous dolomite. Dolomite-cemented, thin- to medium-bedded, white to yellow, medium grained sandstone layers are found in the lower third of the Chepultepec. Creamy white, porous, iron-stained, oolitic chert blocks are common in the residuum of the Formation. The basal contact with the underlying Copper Ridge Dolomite is drawn at the base of the lowest dolomite-cemented sandstone.

Longview Dolomite. The Longview Dolomite is composed of medium-bedded, light gray, finely- to medium-crystalline, siliceous dolomite and thin bedded, light gray to brown, fine grained limestone (which weathers to a bluish gray). These limestones are most commonly found in the upper portion of the Formation. Numerous massive blocks of light gray, fine-grained, non-porous, porcelaneous luster chert are characteristic of Longview residuum. The basal contact with the underlying Chepultepec Dolomite is commonly drawn based on the difference in chert float.

Kingsport Formation. The Kingsport Formation is predominantly thick-bedded, light gray, finely-crystalline, siliceous dolomite or less

commonly a medium-bedded, light gray, coarsely-crystalline, siliceous dolomite. Light gray to brown, fine-grained limestones occur in the basal fifty feet of the Formation. Small blocks of grayish-white, fine-grained, dull luster, oolitic chert and white, fine-grained, porous dull luster, iron-stained chert are found in the residuum. The Kingsport Formation has less residual chert than any other previously described formation in the Knox Group. The basal contact with the underlying Longview Dolomite is commonly drawn based on the difference in chert float.

Mascot Dolomite. The Mascot Dolomite is composed of thin to medium bedded, light gray, finely-crystalline, siliceous dolomite with the lower third made up of dolomite-cemented thin sandstone beds and a chert matrix sandstone base. Faulting in the study area has removed the lower, chert-matrix sandstone base, which was used by Rodgers (1953, p. 59) to divide the Mascot Dolomite from the underlying Kingsport Formation. The top of the Mascot Dolomite was mapped based on the presence of very distinctive chert balls, which are composed of brecciated and dolomite-recemented, dark brown, fine- to medium-grained porous, dull luster chert. These chert balls have diameters of as much as one meter and record erosion along the upper Knox Group unconformity. Borowski (1982) measured 63 m of relief on the upper Knox unconformity at the top of the Mascot Dolomite.

Knox Group depositional environment. Harris (1973) interpreted the Knox Group as a westward transgression of the Late Cambrian carbonate shelf sequence to the east, which earlier formed the shallow-water

carbonates of the Conasauga Group. The Knox Group is chiefly composed of dolomite with abundant chert on the northwest side of the Valley and Ridge and limestone with little chert on the southeast side of the Valley and Ridge (Rodgers, 1953, p. 61). Rodgers believed that the increasing chert content in the northwestern phase of the Knox Group was indicative of a northwestern silica source. According to Rodgers, the Pulaski fault marked the boundary between these two phases of the Knox Group. Harris (1973) believed that dolomite formed in the central and northwestern phases, due to the development of subtidal high-salinity environments which produced dolomitization via hypersaline magnesium-bearing solutions. The sandstone beds within the Knox Group represent minor regressions and transgressions in this dolomite-producing environment (Rodgers, 1953).

Chickamauga Group

Rodgers (1953) and Borowski (1982) discussed the stratigraphy of the Chickamauga Group in the study area. Rodgers (1953) and Swingle (1964) divided the Chickamauga Group into upper (Och₄) and lower (Olmc) portions based on two bentonite beds. The contact between the lower and upper portions of the Chickamauga Group was drawn 100 feet above the lowest bentonite bed. Rodgers (1953, pp. 88-95) described the lower portion of the Chickamauga (Olmc), which included both the lower and middle Chickamauga, as a light gray, to pinkish-colored, fine grained, locally cherty, silty, locally fossiliferous (bryozoans and brachiopods) limestone. The Olmc cherts were brown or black, porous, and oolitic or nodular. Dark gray, medium-grained, shaley, very

fossiliferous (brachiopods) limestones were described as characteristic of unit four (Och₄) of the upper Chickamauga Group.

Borowski (1982) did not agree with Rodgers and Swingle on the stratigraphy of the Chickamauga Group in the area between the Whiteoak Mountain/Wallen Valley thrusts and the Kingston thrust. Borowski measured two complete sections of the entire Chickamauga Group in the study area. He determined that these rocks were "dissimilar to the rocks located in the central Basin of Tennessee and south of the Whiteoak Mountain/Wallen Valley faults of the Tennessee Valley and Ridge" and "did not attempt to attribute the already established Central Basin and Valley and Ridge stratigraphic nomenclature to the middle Ordovician rocks near Clinton" (Borowski, 1982, p. 153). He recognized the Chickamauga Group as a stratigraphic unit between the Knox Group and the Reedsville Shale.

The author did not attempt to divide the Chickamauga Group into different formations during mapping. Stratigraphic contacts with the Mascot Dolomite below and the Reedsville Shale above allowed the author to determine that portion of the Chickamauga which was mapped. The author maps the lower Chickamauga as a light gray, fine-grained, and locally fossiliferous (brachiopods, bryozoans, echinoderms, and trilobites), shaley limestone. Small pieces of brown, iron-stained, fine- to medium-grained, porous, dull luster, oolitic chert and small pieces of fossiliferous limestone are commonly found in the residuum. The upper Chickamauga limestones are thin- to medium-bedded, dark gray to tan brown, very fine- to fine-grained, and fossiliferous. These rocks were examined petrographically by the author in the Carbonate Sedimentology

class at The University of Tennessee. The limestones are most commonly micritic mudstones and wackestones and contain abundant brachiopods, ostracods, trilobites, bryozoans, and echinoderms.

Borowski (1982) interpreted the Chickamauga Group as a shallow-water, tidal flat to carbonate shelf sequence that formed a southeastward-sloping carbonate ramp. Walker et al. (1983) described the entire Chickamauga Group in East Tennessee as shelf and shelf-edge carbonates (in the northwest), which grade into deep-water clastics and turbidites (to the southeast). These deeper water clastics and turbidites were deposited near their southeastern source in a foredeep basin, which formed during the Blountian Phase of the Taconic Orogeny (Rodgers, 1970). Carbonate production finally ended in the study area when a terrigenous influx from the northeast reached the shelf during a later phase of the Taconic Orogeny.

Reedsville Shale

The Reedsville Shale (Or) is composed of green and yellow silty shales and thin-bedded, light gray, locally fossiliferous (articulate brachiopods) limestones. Rodgers (1953, p. 94) correlated the Reedsville Shale with the upper portion of the Martinsburg Shale (Omb). Southeast of the Hunter Valley/Whiteoak Mountain thrusts the unit is dominantly shale and was mapped as the Martinsburg Shale (Omb). The Reedsville Shale had its source to the northeast from lands uplifted during a later stage of the Taconic Orogeny (Rodgers, 1970) and was deposited in a shallow subtidal environment (Thompson, 1984). Contacts between the upper Chickamauga Group and the overlying Reedsville

Shale are difficult to determine except along roadcuts. The boundary between the Chickamauga Group and Reedsville Shale is placed at the base of the lowest yellow or green silty shale.

Sequatchie Formation

The Sequatchie Formation, from bottom to top, is composed of thin- to medium-bedded maroon mudstones, maroon and yellow siltstones, and gray limestones. These upper limestones have a light gray color when fresh, and upon weathering turn maroon with green mottling. Characteristic features of the lower maroon mudstones are locally intense bioturbation (vertical and horizontal burrows) and minor ripple cross lamination. The maroon and yellow siltstones commonly have flaser bedding, unidirectional cross bedding, and bidirectional cross bedding. Locally fossiliferous zones occur in the gray limestones, which contain bryozoans, articulate brachiopods, trilobites, ostracods, and echinoderms. Thompson (1984) proposed a shallow subtidal, tidal flat depositional environment for the Sequatchie Formation in the study area. According to Rodgers (1953) and Hardemann (1966), the Sequatchie Formation is less calcareous southeast of the Hunter Valley/Whiteoak Mountain thrusts and was mapped as the Juniata Formation (Oj). The contact between the Sequatchie Formation and the underlying Reedsville Shale is drawn at the base of the lowest maroon-colored, silty mudstone.

Rockwood Formation

The Rockwood Formation, from bottom to top, is composed of light tan sandstones (up to 1.5 m thick) with interbedded thin-bedded, gray siltstones and gray-to-olive green shales; thin-bedded (less than .2 m

thick) ferruginous sandstones and siltstones; and gray-to-brown siltstones with interbedded olive or brown shales. The lowermost gray siltstones and shales locally have interbedded light-gray fossiliferous limestones, that contain bryozoa, articulate brachiopods, gastropods, and trilobites. These limestones lack the green mottling of the upper Sequatchie limestones and are much thinner bedded (.30 m thick). The interbedded gray silty shales contain abundant brachiopods, which allow correlation of this unit for 9.14 km along strike. Conodonts from these shaley limestones were examined by paleontologists at The University of Tennessee, Knoxville, and the presence of the species Neuroprionodus confirms that these limestones are Silurian in age. Small-scale bidirectional cross-bedding and larger-scale unidirectional cross-bedding are common in the thin-bedded siltstones and medium-bedded sandstones of the Rockwood Formation. Rodgers (1953, p. 99) correlated the clean, silica-cemented, basal Rockwood sandstones with the Clinch Sandstone (Sc), which thickens to the northeast. The contact between the Rockwood Formation and the underlying Sequatchie Formation is drawn at the base of the lowest gray siltstone--gray silty shale or the top of the uppermost maroon-weathering limestone which has green mottling. The upper Rockwood is in fault contact with the lower Rome, upper Rome, Mascot Dolomite, and lower Chickamauga Group.

Thompson (1984) interpreted the depositional environment of the Rockwood Formation, based on the trace fossils and sedimentary structures, as a siliciclastic tidal flat environment and an ancient sublittoral shelf environment. These environments were formed on the

eastern shoreline of a large epicontinental sea, which existed to the west during Silurian time.

Geographic Location and Topographic Expression of Stratigraphic Units and Thrust Faults

Ridge-Forming Stratigraphic Units

The four main ridges in the study area (Plate 1) are Pine Ridge, Bluebird Ridge, Lone Mountain/Lost Ridge, and the northwestern side of Chestnut Ridge. These ridges are underlain by sandstones and siltstones of the upper Rome Formation (Pine Ridge and Bluebird Ridge); sandstones, siltstones, and shales of the lower Rockwood Formation (Lone Mountain/Lost Ridge); siltstones and mudstones of the Sequatchie Formation (Lone Mountain/Lost Ridge); and the siliceous dolomite of the Copper Ridge Dolomite (Chestnut Ridge). The sides of the ridges are underlain by the silty shales of the Reedsville Shale (Lone Mountain/Lost Ridge), the siltstones and shales of the upper Rockwood (Lone Mountain/Lost Ridge), the cherty dolomite of the Mascot Dolomite (Lone Mountain/Lost Ridge), the siltstones and shales of the lower Conasauga Group (Pine Ridge and Bluebird Ridge), and the thin-bedded shales, siltstones, and sandstones of the lower Rome Formation (Pine Ridge and Bluebird Ridge).

Valley-Forming Stratigraphic Units

The five main valleys in the study area (Plate 1) are Wolf Valley, Brushy Valley, Blockhouse Valley, Pumpkin Hollow/Foust Hollow, and Big Valley. These valleys are underlain by silty shales of the

Conasauga Group (northwestern side of Wolf Valley and the northwestern side of Brushy Valley); the shaley limestones of the Maynardville Limestone (Wolf Valley) and Chickamauga Group (Big Valley and Blockhouse Valley); the shales of the lower Rome (Brushy Valley, Blockhouse Valley, and Pumpkin Hollow/Foust Hollow); and the shales and siltstones of the upper Rockwood (Pumpkin Hollow). Small hills within the valleys are formed by more silty shales (Conasauga), thin-bedded siltstones and shales (lower Rome and upper Rockwood), and more shaley limestones (Chickamauga and Maynardville).

Thrust Faults

The three main thrust faults in the study area are the Hunter Valley, Wallen Valley, and Whiteoak Mountain. Three trailing imbricates of the Wallen Valley thrust and five small-displacement thrusts in the hangingwalls of the Hunter Valley/Whiteoak Mountain and Wallen Valley thrusts are also mapped. The Hunter Valley thrust places lower Rome over upper Rome or lower Conasauga and parallels the southeastern side of Bluebird Ridge. Lower Rome, upper Rome, Mascot Dolomite, and lower Chickamauga are placed over Rockwood by the Wallen Valley thrust, which parallels the southeastern side of Lost Ridge/Lone Mountain (in the southwest) and the northwestern side of Bluebird Ridge (near Pumpkin Hollow and Foust Hollow Roads in the northeast). The Whiteoak Mountain thrust places lower Rome over lower Chickamauga and parallels Blockhouse Valley.

The three trailing imbricates of the Wallen Valley thrust place lower Rockwood over upper Rockwood and Sequatchie over Rockwood on the

southeastern side of Lone Mountain. Prominent hills on the southeastern side of Lone Mountain and southeast of Pumpkin Hollow/Foust Hollow Roads are underlain by the sandstones of the lower Rockwood. Maroon mudstones and siltstones of the lower and middle Sequatchie, which are thrust over the lower Rockwood, produce small hills on the southeastern side of Lone Mountain near the intersection of Brushy Valley and Pumpkin Hollow Roads.

Five small-displacement (30 m or less) thrusts are mapped in the upper and lower Rome Formations of Pine Ridge and Bluebird Ridge. These small displacement intraformational thrusts place older similar lithologies on younger similar lithologies and have little or no topographic expression. Three synthetic thrusts (southeast-dipping) are mapped on the northwest side of Pine Ridge (two along I-75N and one along US-25W). Two antithetic thrusts (northwest-dipping) are mapped on the northwestern side of Pine Ridge (along Dismal Valley Road) and the southeastern side of Bluebird Ridge (along I-75N).

CHAPTER II

STRUCTURAL GEOLOGY OF THE STUDY AREA

Introduction

This section deals with macroscopic, mesoscopic, and microscopic (map scale, outcrop scale, and thin-section scale) deformation structures, which were observed within hangingwall and footwall rocks of the Hunter Valley, Wallen Valley, and Whiteoak Mountain thrust faults. Keith (1896), who first mapped the area and drew the first structure sections, showed deformation caused by thrust faulting and minor folding. Rich (1934) and Rodgers (1953) proposed that the basal Rome shales serve as the master décollement for all Valley and Ridge thrust faults. The kinematic sequence of thrust faulting in the Valley and Ridge from east-to-west was proposed by Roeder et al. (1978). Roeder's students used the technique of meso-fabric analysis (Price, 1967) in determining the timing of different macroscopic-scale deformation episodes. Mesoscopic structural analysis was also used to determine the deformation mechanisms, which acted on the outcrop scale. Ramsay (1967) discussed the techniques of mesoscopic- and microscopic-scale strain analysis using objects with known original shapes or angular orientations. These strain values could be used to interpret and explain the larger-scale strain patterns derived from balanced cross-sections (Ramsay, 1969).

Regional Structural Setting

The Valley and Ridge of East Tennessee was formed by southeasterly dipping thrust faults, which brought Cambrian Rome or younger rocks to the surface within the hangingwall (Rodgers, 1953). Following a stairstep trajectory, these thrusts cut upsection across more competent stratigraphic units and parallel to bedding along weaker units (Dahlstrom, 1969, 1970). This stairstep trajectory of the thrust faults produced multiple upper-level décollements above the basal detachment (Roeder et al., 1978).

The major thrust faults in the study area are the Hunter Valley, Wallen Valley, and Whiteoak Mountain thrusts (Plate 1). The maximum displacements of the Hunter Valley (2.44 km), Wallen Valley (8.5 km), and Whiteoak Mountain (9.87 km) thrusts are determined by restoration of the Plate 2 deformed sections to the Plate 3 undeformed state. Displacement on the Hunter Valley thrust increases to the southwest as the displacement on the Wallen Valley thrust decreases. The Hunter Valley thrust is renamed the Whiteoak Mountain thrust, where the Wallen Valley thrust cuts upsection and rejoins the older Hunter Valley thrust. These major thrust faults have several mappable folds and faults associated with them (Plate 1). Commonly, the macroscopic folds have a concentric geometry in strong layers and a similar geometry in weak layers. They have maximum strike lengths of 11.5 km. Mappable minor thrusts have displacements of 30 m or less and maximum strike lengths of 3.5 km. These faults typically die out along strike into a fold

(Plate 1). Mesoscopic folds and faults parallel the orientations of the nearest macroscopic folds and faults.

Mapped lithologies in the study area range from Lower Cambrian to Silurian in age. Because the Valley and Ridge was deformed during the Late Permian (Rodgers, 1953) and given the cumulative thickness of the entire Paleozoic stratigraphic section (4.74 km) in the study area (Roeder et al., 1978), the maximum temperatures and pressures of overburden can be approximately calculated for the different units prior to thrust faulting. Assuming an average geothermal gradient of 25°C/km and a pressure gradient of 1 kb/3 km, the maximum temperatures and pressures of overburden range from 118.5°C and 1.6 kb at the base of the lower Rome to 51°C and .7 kb at the top of the Rockwood Formation. These lower temperatures and pressures explain the absence of cleavage within the rocks of the study area. Even a doubling of stratigraphic overburden via thrust-fault duplication places the rocks below the lowest temperature for cleavage development at approximately 250°C (Roeder et al., 1978).

Dominant Structures: Folds and Faults

Folds

Macroscopic, mesoscopic, and microscopic scale folds were produced by movement along large- and small-displacement contractional (thrust) faults. Macroscopic or map scale folds are shown in Plate 1, Plate 2, and Figure 3. Mesoscopic or outcrop-scale folds are shown in Figures 4-11, which are roadcut sections along I-75N and US-25W through the Rome (Gr), Conasauga (Gc), Sequatchie (Os), and Rockwood

Figure 3. Lithotectonic Map of the Study Area Showing the Location of Outcrops in Figures 4-11.

WM = Whiteoak Mountain thrust

HV = Hunter Valley thrust

WV = Wallen Valley thrust

Sr = Rockwood Formation

Os = Sequatchie Formation

Or = Reedsville Shale

Och = Chickamauga Group

Oma = Mascot Dolomite

Ecr = Copper Ridge Dolomite

Emn = Maynardville Limestone

Ec = Conasauga Group Shale

Gru = Upper Rome Formation

Erl = Lower Rome Formation

Outcrops 1-2 are along I-75N at Pine Ridge

Outcrop 3 is along Moore Gap Road at Pine Ridge

Outcrop 4 is along a Railroad Cut Through
Pine Ridge

Outcrop 5 is along Brooks Gap Road at Bluebird
Ridge

Outcrop 6 is along I-75N at Lone Mountain

Outcrop 7 is along US-25W at Lost Ridge

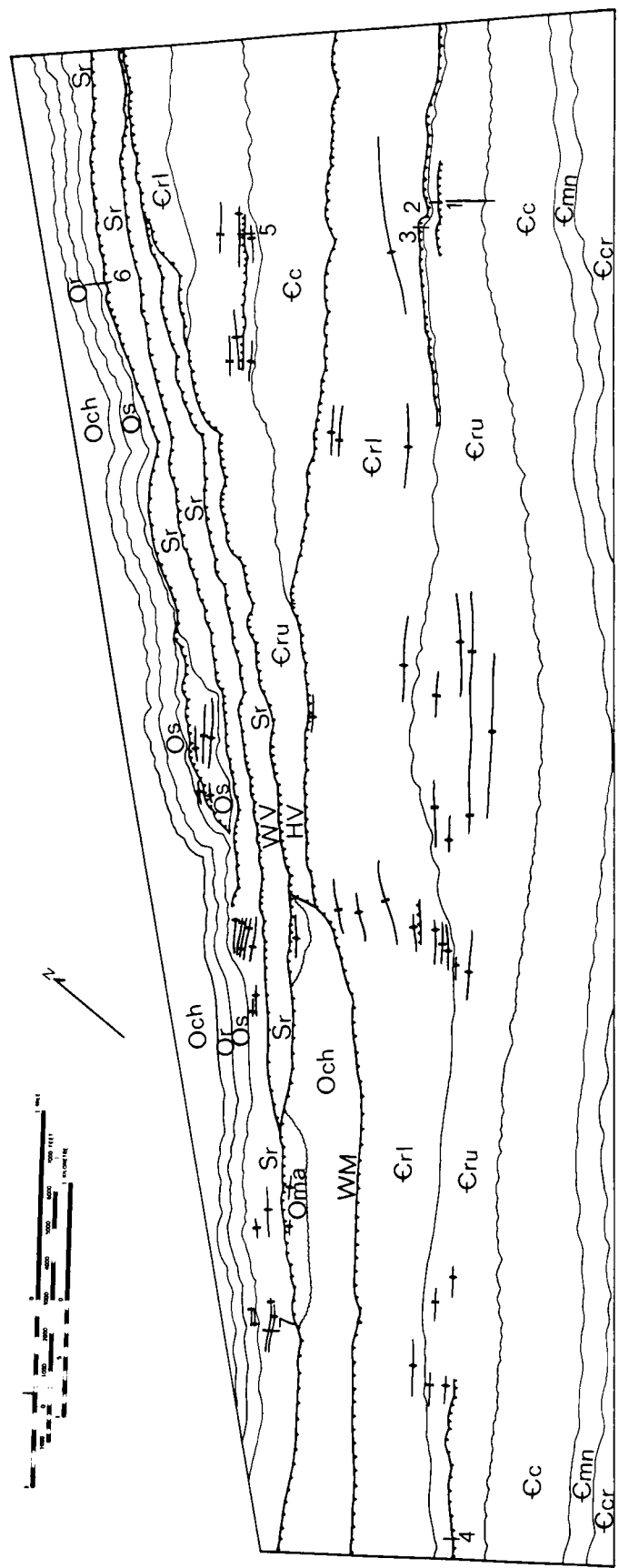


Figure 3

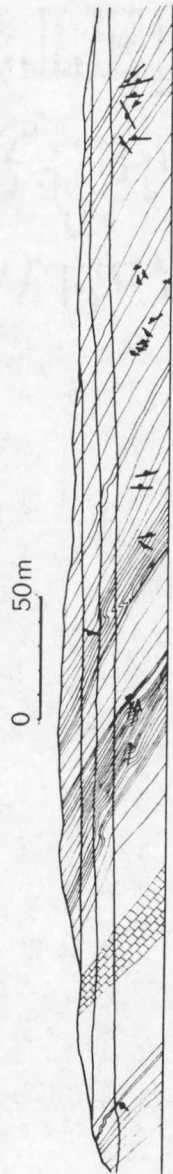


Figure 4. Outcrop Sketch of Upper Rome and Lower Conasauga Exposed Along I-75N at Pine Ridge.



Figure 5. Outcrop Sketch of Small-Displacement Thrusts in the Upper Rome Exposed Along I-75N at Pine Ridge.



Figure 6. Outcrop Sketch of Small-Displacement Thrust in the Lower Rome Exposed Along Moore Gap Road at Pine Ridge.

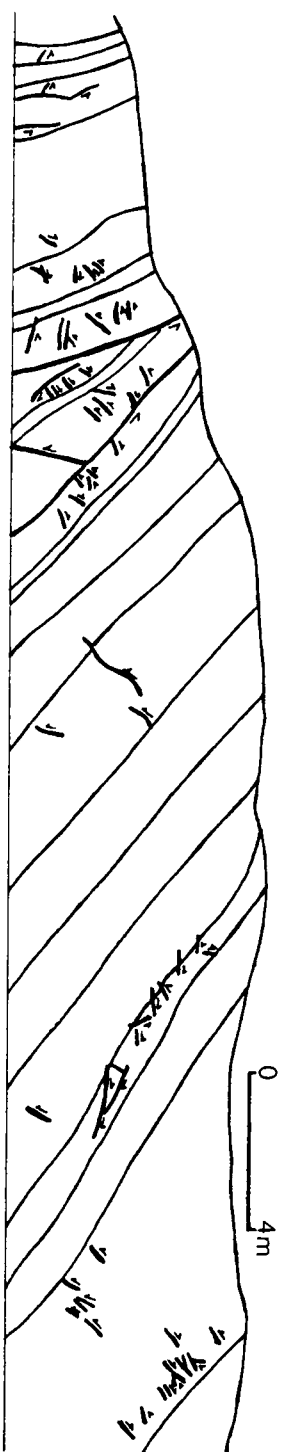


Figure 7. Outcrop Sketch of Small-Displacement Thrusts in the Upper Rome Exposed Along a Railroad Cut Through Pine Ridge.

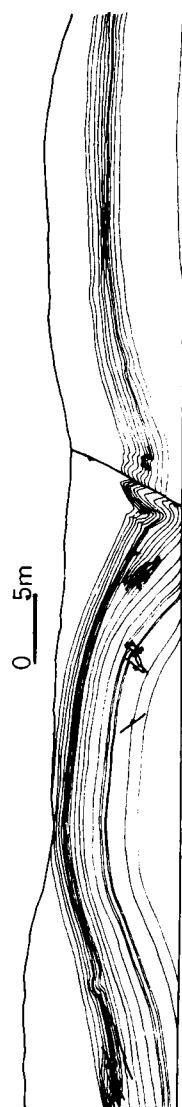


Figure 8. Outcrop Sketch of Macroscopic Folds and a Small-Displacement Antithetic Thrust in the Upper Rome Exposed Along Brooks Gap Road at Bluebird Ridge.

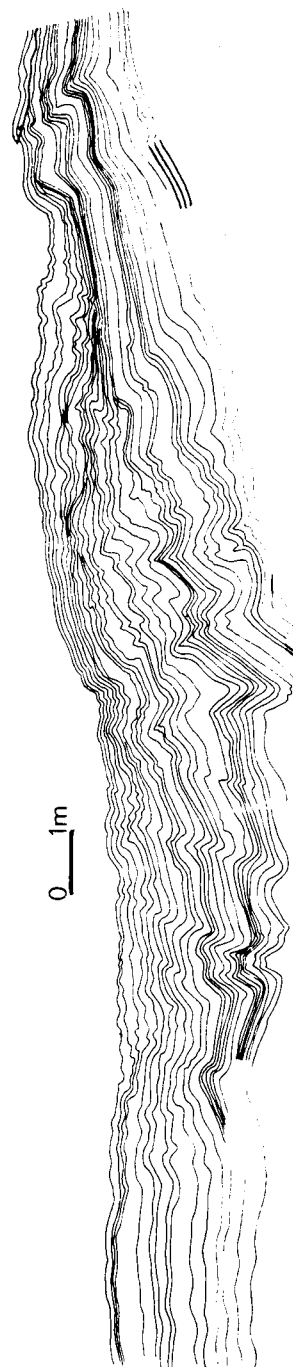


Figure 9. Detailed Outcrop Sketch of Minor Folds in the Macroscopic Anticline of Figure 8.

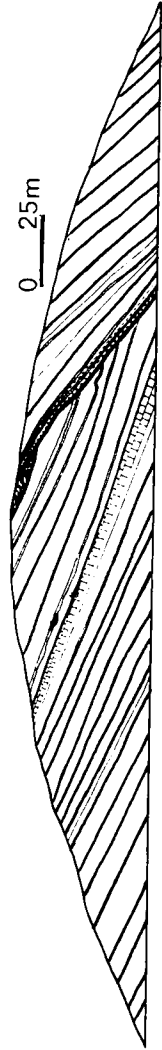


Figure 10. Outcrop Sketch of Trailing Imbricate of the Wallen Valley Thrust in the Lower Rockwood Exposed Along I-75N at Lone Mountain.

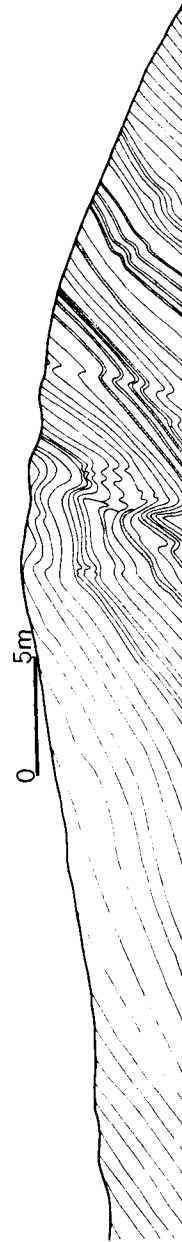


Figure 11. Outcrop Sketch of Chevron Folds in the Lower Rockwood Exposed Along US-25W at Lost Ridge.

(Sr) formations. Fold wavelengths were controlled by the thickness of competent layers, whereas amplitudes were controlled by strain (Donath and Parker, 1964; Currie et al., 1962). A maximum bedlength shortening value (Ramsay, 1969) of 21% was seen in the Rockwood footwall of the Whiteoak Mountain fault (Figure 11). In contrast, mesoscopic folds within the Rome hangingwall of the Wallen Valley thrust showed a maximum bedlength shortening of 14% (Figure 9). On a larger scale, the macroscopic folds in the Rome hangingwall of the Wallen Valley, Hunter Valley, and Whiteoak Mountain thrusts showed an increased amount of shortening along strike to the southwest. These maximum shortening values ranged from 4% in structure section A-A' to 29% in structure section D-D' (Plates 2 and 3). This increased shortening observed in both mesoscopic and macroscopic folds to the southwest represents the additive displacements of both the Hunter Valley and Wallen Valley faults to form the Whiteoak Mountain fault.

Three major types of folds (parallel, similar, and chevron) were observed in the study area. Chevron folds combined both inter-layered parallel and similar styles (Ramsay, 1974). Concentric parallel folding (Ramsay, 1967, p. 398) was seen in mesoscopic and macroscopic folds within the competent sandstone and siltstone beds of the Upper Rome (Figure 8), Rockwood (Figure 12), and Sequatchie (Plate 2) Formations. These folds had wavelengths of 2-670 m and amplitudes of .5-152 m. According to Ramsay's fold classification, parallel folds fell into class 1b or 1c and had a constant thickness normal to bedding (Ramsay, 1967, p. 365). Donath and Parker (1964) described parallel folds as flexural-slip folds, which were formed when mechanically



Figure 12. Parallel Fold in the Footwall of a Trailing Imbricate of the Wallen Valley Thrust Within the Lower Rockwood Exposed Along I-75N at Lone Mountain. Note hammer in the lower left corner for scale.

strong (competent) beds slipped along the interfaces with less competent beds above and below. These competent units commonly produced symmetric buckle folds (Figure 8) and boxfolds (Figures 6 and 13), which were two subclasses of parallel-style folds (Ramsay, 1967).

Similar folds were seen in mesoscopic and macroscopic folds within the shales of the lower Rome (Figures 6 and 9) and Rockwood (Figure 11) Formations. These folds had wavelengths of 3-366 m and amplitudes of 15-91 m. According to Ramsay's fold classification, similar folds fell into class 2 and had a constant thickness parallel to the axial plane (Ramsay, 1967, p. 365). Donath and Parker (1964) described similar folds as flexural-flow folds, which were formed when mechanically weak (incompetent) beds took up the shear strain internally. These incompetent units commonly produced asymmetric folds during flow into the voids formed by the folding of more viscous and competent layers (Ramsay, 1967, p. 416).

Chevron folds were seen in mesoscopic and macroscopic folds within the interbedded shales and sandstones or siltstones of the upper Rome (Figure 14), Conasauga Shale (Figure 15), and Rockwood (Figure 11) Formations. These folds had a small hinge surface relative to the limb surface, and the limb length was approximately ten times the thickness of the most competent layers involved in folding (Ramsay, 1974). Several bulbous hinge structures and limb thrusts were formed when the folding accommodated minor changes in the thickness of the more competent layers (Ramsay, 1974, p. 1745). Donath and Parker (1964) described the chevron folding process as layer-parallel compression of a more viscous (less ductile) layer in a less viscous

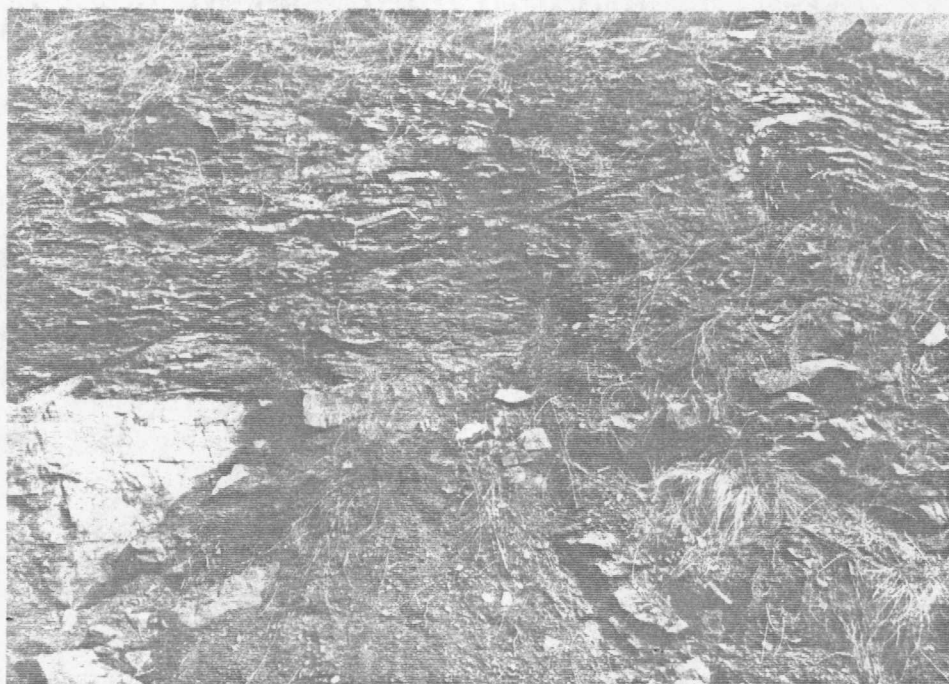


Figure 13. Boxfold in the Hangingwall of an Antithetic Thrust Within the Upper Rome Exposed Along Brooks Gap Road at Bluebird Ridge. Field of view is 10 m.

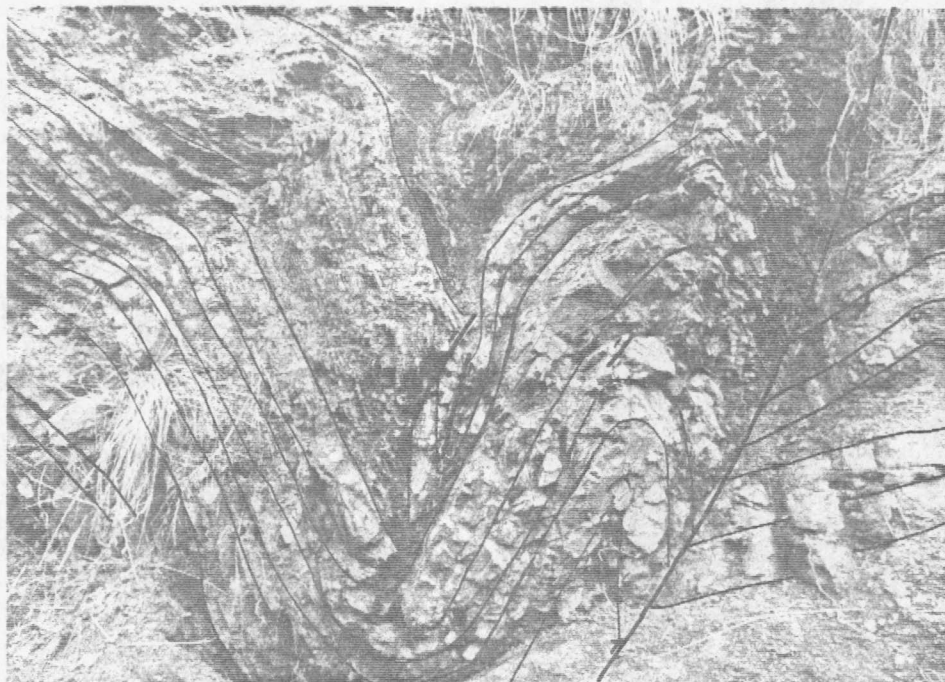


Figure 14. Chevron Fold in the Hangingwall of an Antithetic Thrust Within the Upper Rome Exposed Along Brooks Gap Road at Bluebird Ridge. Field of view is 6 m.

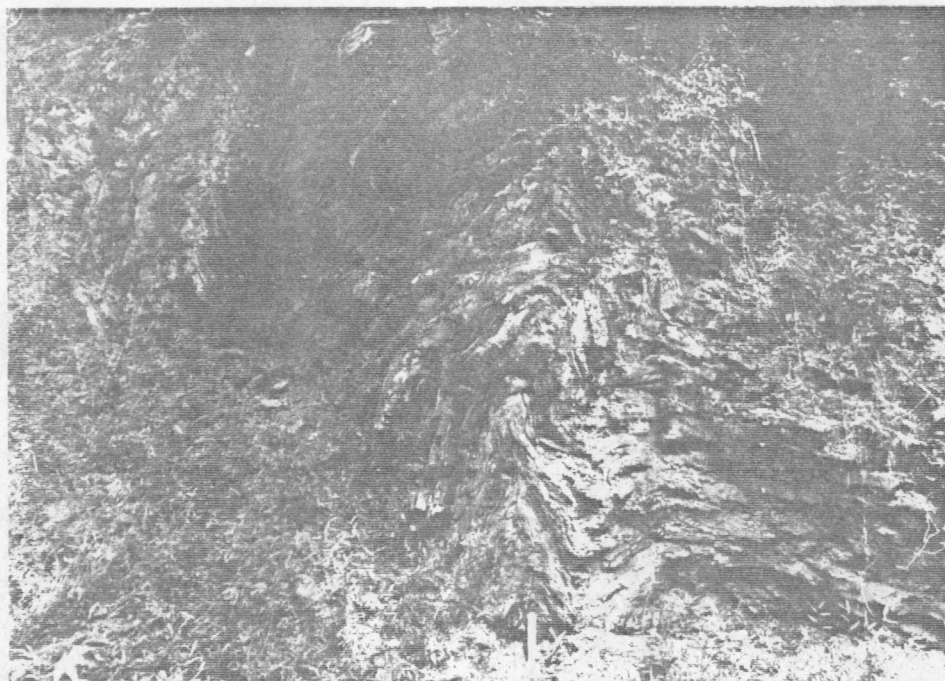
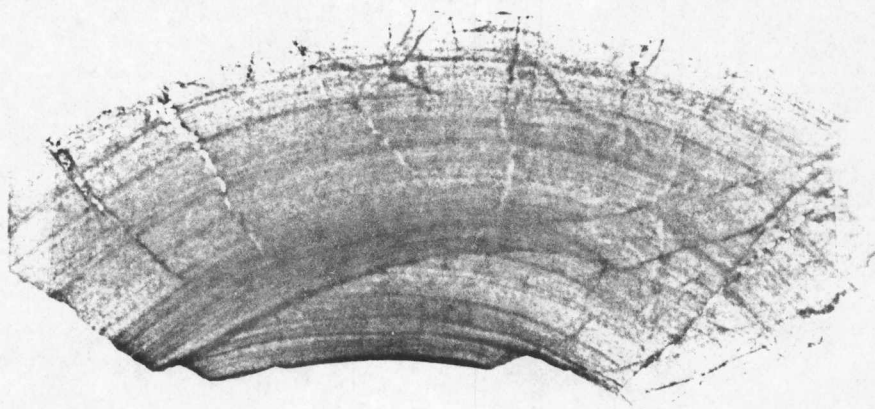


Figure 15. Chevron Folds in the Hangingwall of the Hunter Valley Thrust Within the Lower Conasauga Exposed Along Dismal Valley Road. Field of view is 3 m.

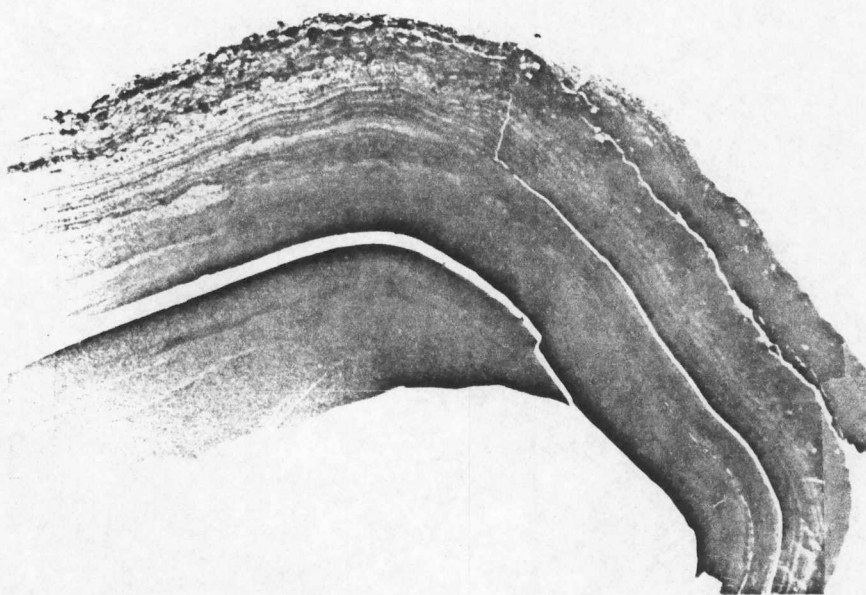
(more ductile) enveloping material. Similarly, Ramsay (1974) explained the formation of class 2 chevron-style folds by having class 1c competent layers surrounded by class 3 incompetent layers in a multilayer sequence.

Thin sections and polished slabs of microscopic and mesoscopic parallel folds (Figures 16, 17, and 18) showed extension in the outer arcs of folds (as evidenced by calcite-filled tension fractures or small displacement extension faults) and contraction in the inner arcs of folds (as evidenced by small-displacement contraction faults). These structures were formed by layer-parallel shortening of the more competent layers (Currie et al., 1962; and Ramsay, 1967, pp. 394-401). Commonly, greatest displacement was observed along southeasterly dipping microthrusts in the inner arcs of folds, which were formed by a dominantly northwesterly directed maximum compressive stress. The microthrusts commonly acted as small-scale bedding wedges (Cloos, 1964) and had less competent materials flowing into the voids created within fold cores (Figure 19). Similar- (Figure 19) and chevron- (Figure 20) style folds developed in less competent shales and shale-sandstone interbeds (Ramberg, 1964). Bulbous noses and second-order folds (Figure 19) were produced by flexural flow of material into the fold cores.

Mesoscopic and macroscopic fold axes trend northeast and southwest in the study area and are parallel to the strike of the nearest thrust faults. All folds were produced by movement along small- or large displacement thrust faults (Roeder et al., 1978; Dahlstrom, 1977, p. 424). The dominant orientations for macroscopic folds in the



16A



16B

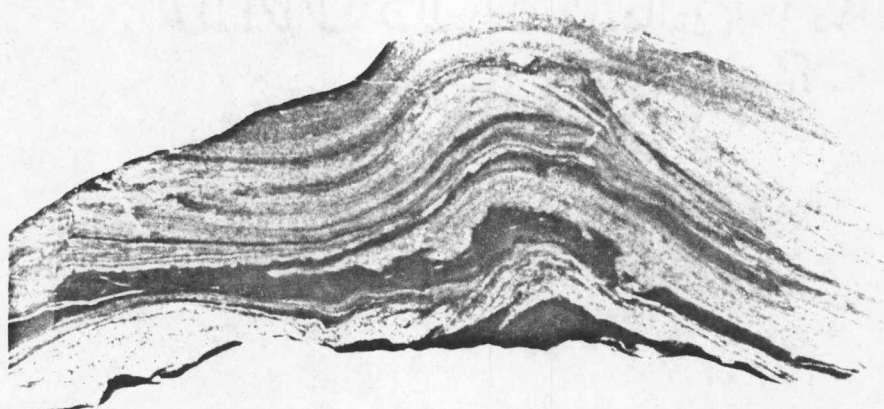
Figure 16 (A and B). Thin Sections of Parallel Folds Within the Lower Rockwood Exposed Along US-25W at Lost Ridge. Thin sections are 5 cm by 7.6 cm.



Figure 17. Thin Section of Parallel Fold Within the Upper Rome Exposed Along I-75N at Bluebird Ridge. Thin section is 5 cm by 7.6 cm.



Figure 18. Thin Section of Normal Fault Within the Upper Rome Exposed Along a Railroad Cut Through Pine Ridge. Thin section is 5 cm by 7.6 cm.



19A



19B

Figure 19 (A and B). Thin Sections of Bedding Wedges Within the Upper Rome Exposed Along I-75N at Bluebird Ridge. Thin sections are 5 cm by 7.6 cm.

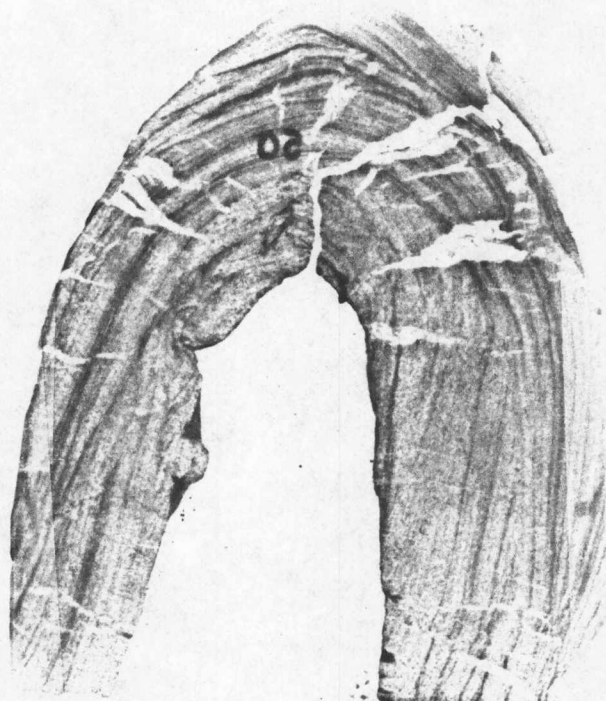


Figure 20. Thin Section of a Chevron Fold Within the Lower Rockwood Exposed Along US-25W at Lost Ridge. Thin section is 5 cm by 7.6 cm.

area are shown in Plate 1 and Figure 3 (page 27). The average mesoscopic fold axes orientations for each of the eight roadcut sections (Figures 4-11, pages 28-30) are shown in Figure 21. Typically, the fold axes plunge either northeast or southwest at low angles of 10-20°.

Faults

The largest displacement faults in the study area are contractional (thrust) faults. Extensional (normal) faults are common, but they have much less displacement (less than 2 m). The Hunter Valley (displacement 2.44 km in section A-A', Plates 2 and 3), Wallen Valley (displacement 8.5 km in section A-A', Plates 2 and 3), and Whiteoak Mountain (displacement 9.87 km in section D-D', Plates 2 and 3) thrusts are the three largest displacement faults in the study area. Structural and stratigraphic data indicate that the Wallen Valley fault folds the older Hunter Valley fault above (Plate 2). The basal decollement for all these thrust faults was the shaley Lower Rome Formation (Roeder et al., 1978). Other upper level décollements used by these faults are the base of the Conasauga Shale, the upper Mascot, the lower Chickamauga shaley limestone, the Reedsville shale, and the Rockwood shales (Plates 2 and 3). Trailing imbricates of the Wallen Valley thrust used the décollements in the Reedsville and Rockwood Formations to duplicate the thicknesses of these formations.

Detailed mapping of the rocks in the study area at 1:12,000 scale (Plate 1) shows that the Wallen Valley fault is a footwall imbricate of the Hunter Valley fault. Both structural and stratigraphic evidence support a Hunter Valley and then Wallen Valley sequence of

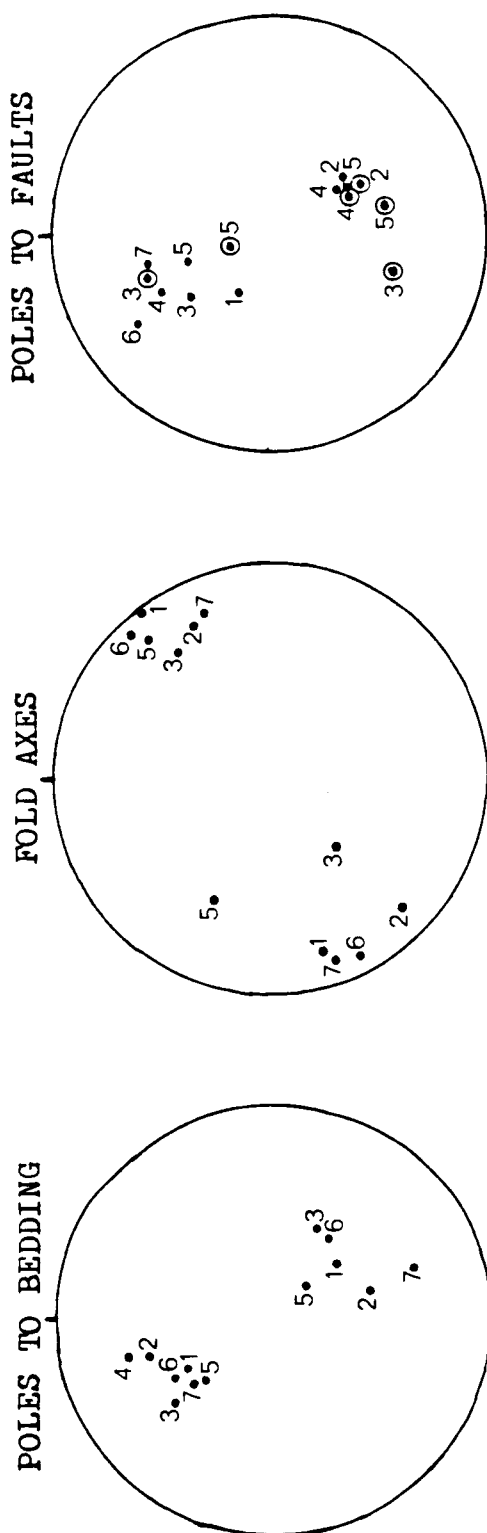


Figure 21. Stereonets (Schmidt nets) Showing the Dominant Orientations for Bedding, Fold Axes, and Faults Within Subareas of the Study Area. Circled poles to faults indicate extensional faults.

- 1 = Conasauga along Dismal Valley Road
- 2 = Rome Formation along Dismal Valley Road through Pine Ridge
- 3 = Rome Formation along I-75N through Pine Ridge
- 4 = Rome Formation along Railroad Cut through Pine Ridge
- 5 = Rome Formation along Brooks Gap Road through Pine Ridge
- 6 = Rockwood Formation along I-75N through Lone Mountain
- 7 = Rockwood Formation along US-25W through Lost Ridge

faulting (section A-A', Plates 2 and 3). Stratigraphic evidence is based on the correlation of two carbonate units (16.8 m and 11.4 m thick) in the Rome hangingwall of both the Hunter Valley and Wallen Valley thrust sheets. This correlation indicates that a nearly complete section of Rome (457 m) is brought to the surface by the Hunter Valley thrust. In contrast, the Rome thickness brought to the surface by the Wallen Valley thrust is less (396 m) and decreases to the southwest as the Wallen Valley thrust cuts upsection in its hangingwall (section B-B', Plates 2 and 3). When the Wallen Valley thrust completely loses its hangingwall Rome by cutting upsection (section C-C', Plates 2 and 3), the Wallen Valley encounters and rejoins the older Hunter Valley thrust. The Hunter Valley thrust now has the combined displacement of both itself and the Wallen Valley thrust. This new fault has a Cambrian Rome hangingwall and an Ordovician Chickamauga horse (fault-bound slice of rock) footwall, and the fault is renamed the Whiteoak Mountain thrust (sections C-C' and D-D', Plates 2 and 3). The structural evidence for a Hunter Valley and then Wallen Valley sequence of thrusting is based on the tightness (short wavelength and high amplitude) of folds, which appear to the southwest in the Rome hangingwall of the Hunter Valley thrust. These higher strain (tighter) folds are produced as the Hunter Valley thrust takes up the displacement of both itself and the Wallen Valley thrust. Maximum bed-length shortening values range from 4% in section A-A' to 29% in section D-D' (Plate 2).

The intersection of the Hunter Valley, Wallen Valley, and Whiteoak Mountain thrusts produces a transference of displacement from one fault to another in an area which Dahlstrom (1970) called a transfer

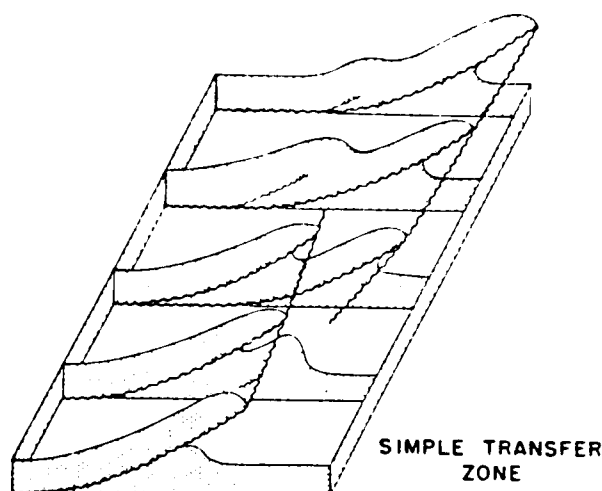


Figure 22. Simple Transfer Zone.
(from Dahlstrom, 1970)

zone (Figure 22). In this transfer zone, the last remnant of the Wallen Valley hangingwall appears as an Ordovician-age horse northwest of the Whiteoak Mountain thrust. The transference of displacement from one fault to another along strike and the formation of horses during footwall weakening was discussed by Dahlstrom (1970, p. 751) and Boyer and Elliot (1982, p. 1198). Footwall weakening and failure of the upper Mascot and lower Chickamauga, within the footwall horse beneath the Wallen Valley and Whiteoak Mountain thrusts, represents the formation of two upper-level décollements along pre-existing zones of incompetent or mechanically weak rocks (Plates 2 and 3). The upper Mascot was probably weakened by dissolution channels, which formed during subaerial exposure and erosion associated with the upper Knox unconformity (Rodgers, 1953). Abundant shaley limestones within the lower Chickamauga (Borowski, 1982) explain the pre-existing zones of weakness within this lithology.

Faults are important in producing mesoscopic and microscopic deformation as well as macroscopic deformation. Mesoscopic contraction faults in the Rome Formation dip both southeast and northwest (Figures 5, 6, 7, and 8, pages 28, 29) and can often be mapped (Plate 1). These mesoscopic thrust faults have maximum displacements of 30 m and when symmetrically paired accommodated the room problem in the cores of macroscopic parallel folds (Gwinn, 1964; Serra, 1977; Morse, 1977). Other smaller-displacement contractional (thrust) and extensional (normal) faults are found in the hangingwall and footwall folds, which are formed by larger-displacement thrust faults (Figures 6 and 8, pages 28, 29). Commonly, paired symmetric contractional faults are found in the inner arcs of mesoscopic folds and extensional faults are found on the outer arcs (Figure 6, page 28).

Thin sections and polished slabs made from the footwall and hangingwall rocks of mesoscopic contractional faults show that mesoscopic and microscopic deformation mechanisms are the same. There are contractional micro-thrusts and extensional micro-faults, which are seen in the inner and outer arcs of small-scale folds (Figures 16, 17, 18, 19, and 20, pages 38-41). These extensional fractures are commonly filled with calcite and may have a sigmoidal shape, due to flexural slip between layers during fracture-filling (Figures 17 and 20, pages 39 and 41). Calcite also re-cements fault gouge along small-displacement (30 m) thrust faults into a coherent material (Figure 23).

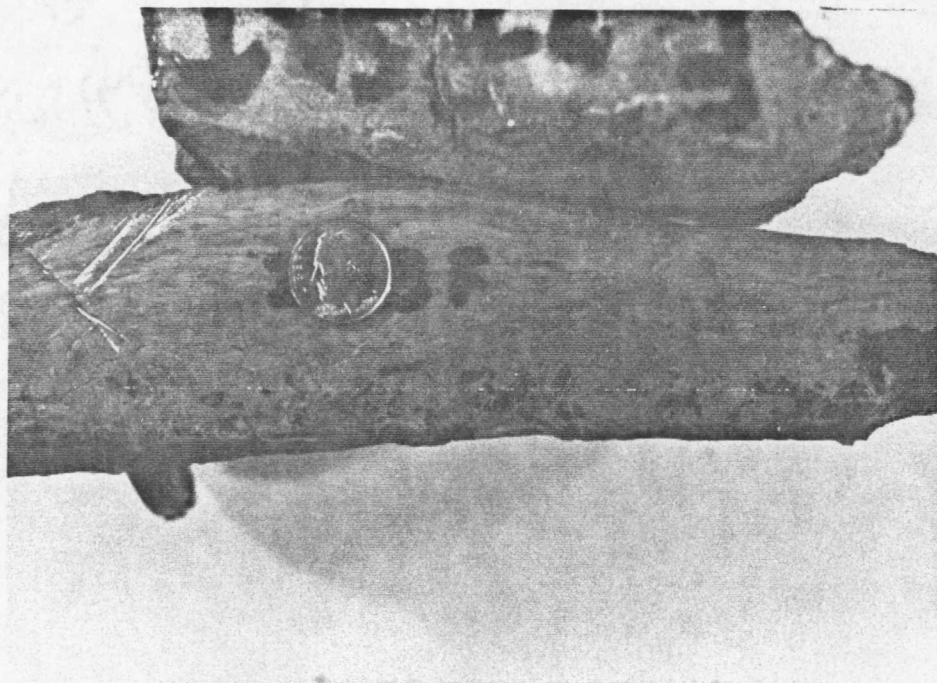


Figure 23. Calcite Re-cemented Fault Gouge from the Surface of the Figure 6 Thrust Fault.

Summary and Conclusions

The most important deformation features in the study area are folds and faults. Three major types of folds (parallel, similar, and chevron) are observed at all scales of deformation. These folds produce shortening by flexural slip between layers and flexural flow within layers. Mechanically strong layers tend to deform via flexural slip into parallel folds, and mechanically weak layers tend to deform via flexural flow into similar folds. Multilayers of weak and strong interbeds deform via both flexural flow and flexural slip. These multilayered sequences form chevron-style folds. Folding on mesoscopic and macroscopic scales can produce maximum bedlength shortening

values of 21% and 29%. All folds are produced by movement along major or minor displacement contractional (thrust) faults or extensional (normal) faults.

Contractional and extensional faults are observed in all scales of deformation. The three largest displacement thrust faults in the study area (the Hunter Valley, Wallen Valley, and Whiteoak Mountain) were mapped at 1:12,000 scale. Stratigraphic and structural evidence both support a Hunter Valley and then Wallen Valley kinematic (timing) sequence of thrust faulting. A transfer zone is produced where the Wallen Valley fault merges with the older Hunter Valley thrust. The new fault, which has the displacement of both the Hunter Valley and Wallen Valley thrusts, is named the Whiteoak Mountain thrust. Mesoscopic and microscopic scale faults are important in outcrop and thin-section deformation. Typically, extensional (normal) faults and contractional (thrust) faults are produced in the outer arcs and inner cores of mesoscopic or microscopic folds.

CHAPTER III

STRAIN VARIATIONS WITHIN ROCKS IN THE STUDY AREA

Introduction

The techniques of strain analysis used in this study involved determining the changes in known shapes and known angles during faulting and folding of rocks. Strain markers, which had a known length, shape, or angular relationship between parts (Ramsay, 1967), were used to determine the strains imposed upon rocks in different parts of the study area. The observed strain orientations were compared to macroscopic and mesoscopic fold and fault orientations in order to infer cause-and-effect relationships. Similarly, the intensity and distribution of strains were compared to the displacements on the major thrust faults in the area. According to Price (1967), the greater displacement thrust faults should show greater strains within hangingwall and footwall rocks. The comparison of strains over a wider area was also used to characterize the dominant deformation mechanisms. Four strain analysis techniques were used on rocks in the study area. These techniques were: (1) balancing cross sections (Dahlstrom, 1969 and 1970); (2) determining the strain ratio (R_f) for deformed fossils with original bilateral symmetry (Wellman, 1962; Ragan, 1973); (3) determining the strain ratio (R_s) and original ellipticity (R_i) of deformed originally spherical ooids (Dunnet, 1968;

Fry, 1979; Ramsay and Huber, 1983); and (4) the unfolding of mesoscopic folds (Ramsay, 1969). Each of these methods and strain analysis results are discussed below.

Balanced Cross-Sections

The geologic structure sections in Plate 2 were restored by taking the hangingwalls of all thrust sheets and pulling them back to their pre-thrusting positions. This is the technique that was called "balancing" by Dahlstrom (1969, 1970). Balancing allows the calculation of both displacement along a fault surface and shortening within a sheet, due to folding and faulting. The method also assures that the structure sections are geometrically correct, although other interpretations may also be geometrically correct.

All geologic structure sections (Plate 2) were drawn and restored (Plate 3) using seven basic assumptions concerning thrust geometries and the proper methods for balancing cross-sections. The first assumption was that bedding thickness and lithology control the ways that rocks respond to stress in thrust-faulted terrains (Dahlstrom, 1969). Competent (mechanically strong), thick-bedded rocks tend to form ramps during thrust faulting and concentric (parallel) folds during folding. Incompetent (mechanically weak), thin-bedded rocks tend to form flats (décollements) during thrust faulting and similar folds during folding. Ramps and flats (décollements) are formed when thrust faults cut up-section across bedding and parallel to bedding. The second assumption was that anticlines and synclines on the surface in thrust areas are indicative of ramps and flats at

depth (Boyer and Elliot, 1982). A third assumption was that horses (fault-bound slices of rock) may be broken off the footwall of thrust faults and brought to the surface by the motion of the hangingwall (Boyer and Elliot, 1982). The fourth assumption was that younger thrusts fold older inactive thrusts above by propagating from underneath and in front of the older thrusts (Boyer and Elliot, 1982). A fifth assumption was that deformed geologic structure sections must be restored in the direction opposite to the direction of tectonic transport or thrust displacement (Dahlstrom, 1969). The sixth assumption was that plane strain existed during thrust-fault deformation. Plane strain was necessary for two-dimensional area balancing of deformed sections (Rich, 1934). Finally, the seventh assumption was that no volume loss occurred during deformation in the study area. Low temperature deformation here resulted in the absence of cleavage and negligible volume loss.

The results of restoring the Plate 2 deformed sections are shown in Plate 3. These restored sections show maximum displacements of 2.44 km, 8.5 km, and 9.87 km for the Hunter Valley (section A-A'), Wallen Valley (section A-A'), and Whiteoak Mountain (section D-D') thrusts. Maximum shortening values for the Hunter Valley, Wallen Valley, and Whiteoak Mountain faults are 53%, 59%, and 76% (sections A-A' and D-D'). The large amount of displacement and shortening on the Whiteoak Mountain fault is caused by the transference of the Wallen Valley displacement to the older Hunter Valley fault (Plate 1 and Plate 2, section C-C'). Displacement and shortening on the Wallen Valley fault decrease to the southwest as the Wallen Valley cuts upsection

in its Rome hangingwall and finally rejoins the older Hunter Valley thrusts (Plates 2 and 3, sections A-A', B-B', and C-C'). To the northeast, the Wallen Valley thrust has trailing-edge imbricates, which repeat the thicknesses of the Sequatchie and Rockwood Formations (Plate 1). The Hunter Valley thrust cuts upsection in its footwall to the northeast as evidenced by the appearance of Conasauga Shale (Plate 1). Therefore, both the Hunter Valley and Wallen Valley faults have greater displacement and deformation to the northeast.

Deformed Rockwood Brachiopods

Deformed fossils with original bilateral symmetry were used to determine the strain ratio ($R_s = 1 + e_1 / 1 + e_2$) within the bedding plane (Wellman, 1962; Ramsay, 1967, pp. 239-242; Ragan, 1973). Fifty brachiopods (Figure 24), within the folded and faulted Rockwood hangingwall of a trailing Wallen Valley imbricate, were used to determine the strain ratio. The bearing of the brachiopod hinge line, the bearing of the symmetry line, and the small angle between the hinge and symmetry line were measured for each specimen. These orientations were plotted on a map view which is parallel to bedding (Figure 25). If the brachiopods are undeformed, then a rectangle or square is produced by the symmetry and hinge lines and the mirror image of the symmetry and hinge lines (Figure 26). Deformed brachiopods will produce a rhomboid (unequal side lengths) or rhombus (equal side lengths), because the small angle subtended by the hinge and symmetry lines will be less than 90°. The plots of the hinge and symmetry line

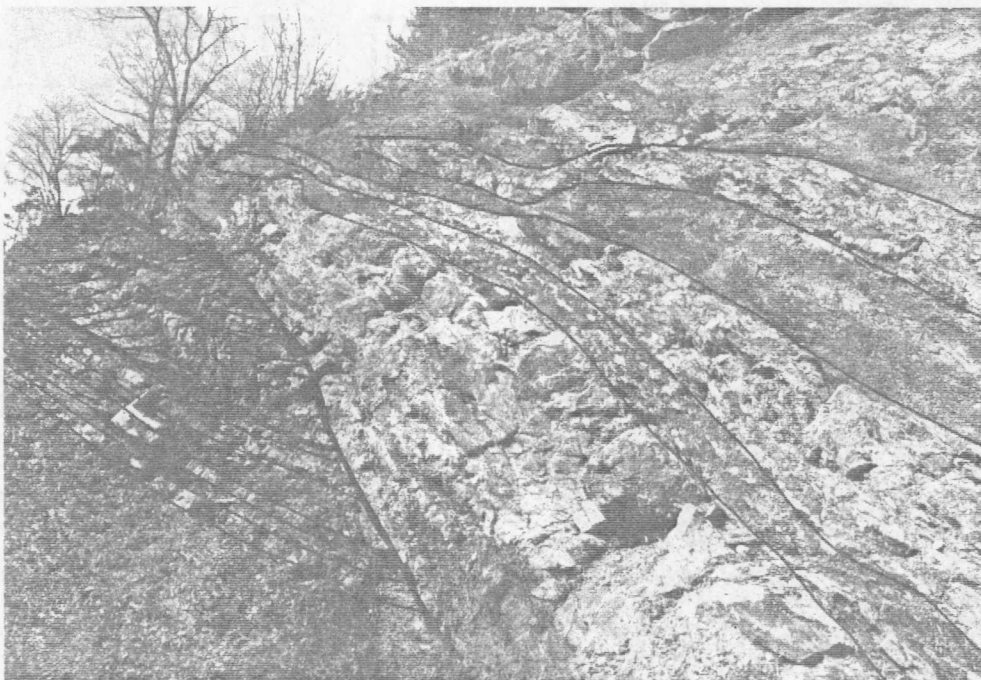


Figure 24. Parallel Folds in the Hangingwall of a Trailing Imbricate of the Wallen Valley Thrust Within the Lower Rockwood Exposed Along I-75N at Lone Mountain. Note hammer in the lower left-hand corner for scale.

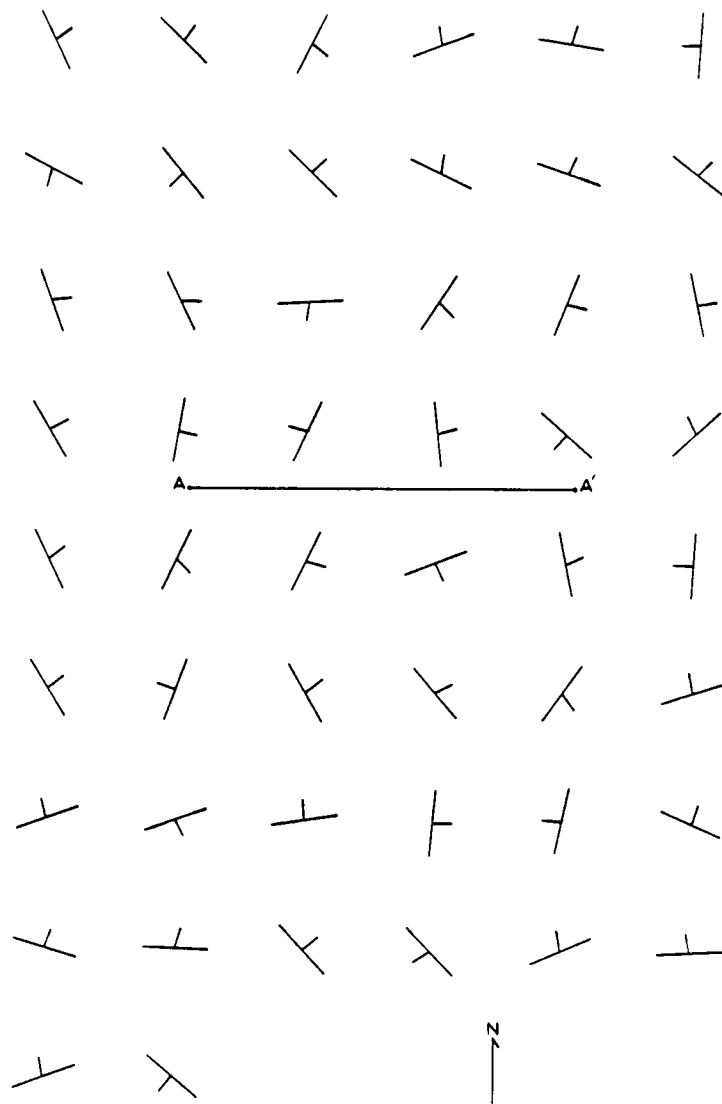


Figure 25. Map View of Rockwood Brachiopod Hinges and Symmetry Lines.

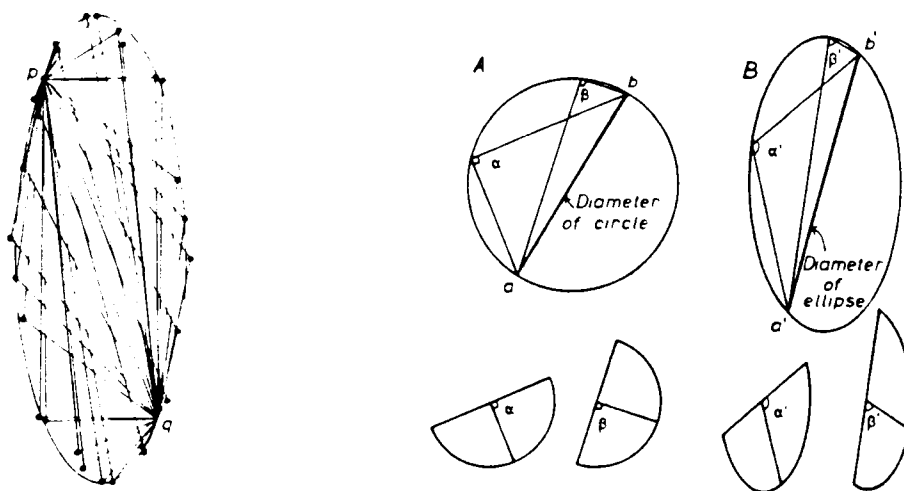


Figure 26. Theoretical Basis for Wellman's Method (from Ramsay, 1967, p. 242).

intersections and their mirror images define the orientation and shape of the best-fit strain ellipses (Figures 27 and 28).

The results of strain analysis on the folded and faulted Rockwood brachiopods indicate a minimum extension of $R_s = 1.15$ subparallel to strike and a maximum extension of $R_s = 1.43$ subnormal to the strike of bedding (Figure 28). These strains are attributed to extension in the outer arcs and contraction in the cores of concentric minor folds within the 3.5 m thick zone of interbedded limestone and shale, which overlies the thrust fault (Figure 24).

Deformed Rome Oolite

The strain ratios (R_s) and maximum original ellipticities ($R_{i_{max}}$) of deformed but originally spherical ooids were determined from six samples of an Upper Rome (Gru) oolite, within the hangingwall of the

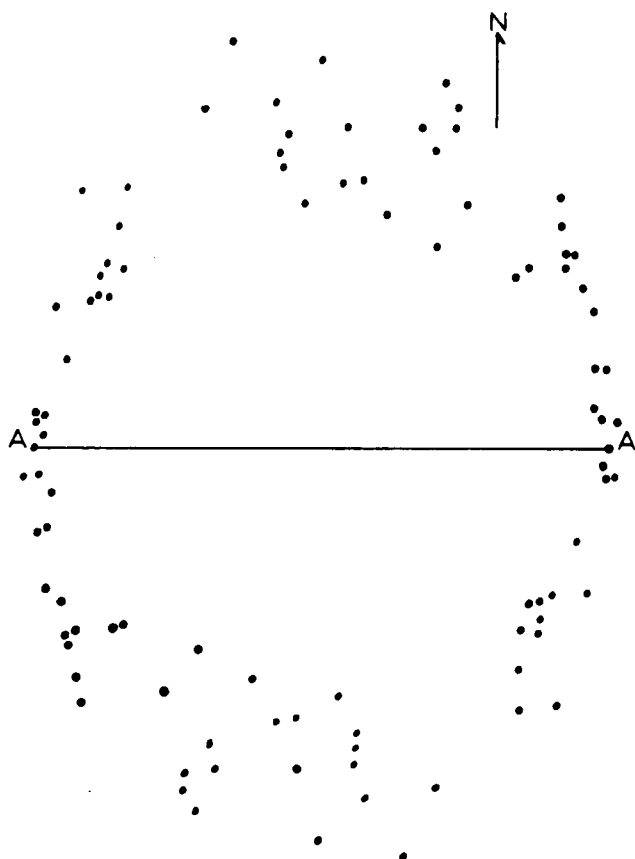


Figure 27. Plot of Points Defined by the Intersection of the Symmetry and Hinge Lines and Their Mirror Images.

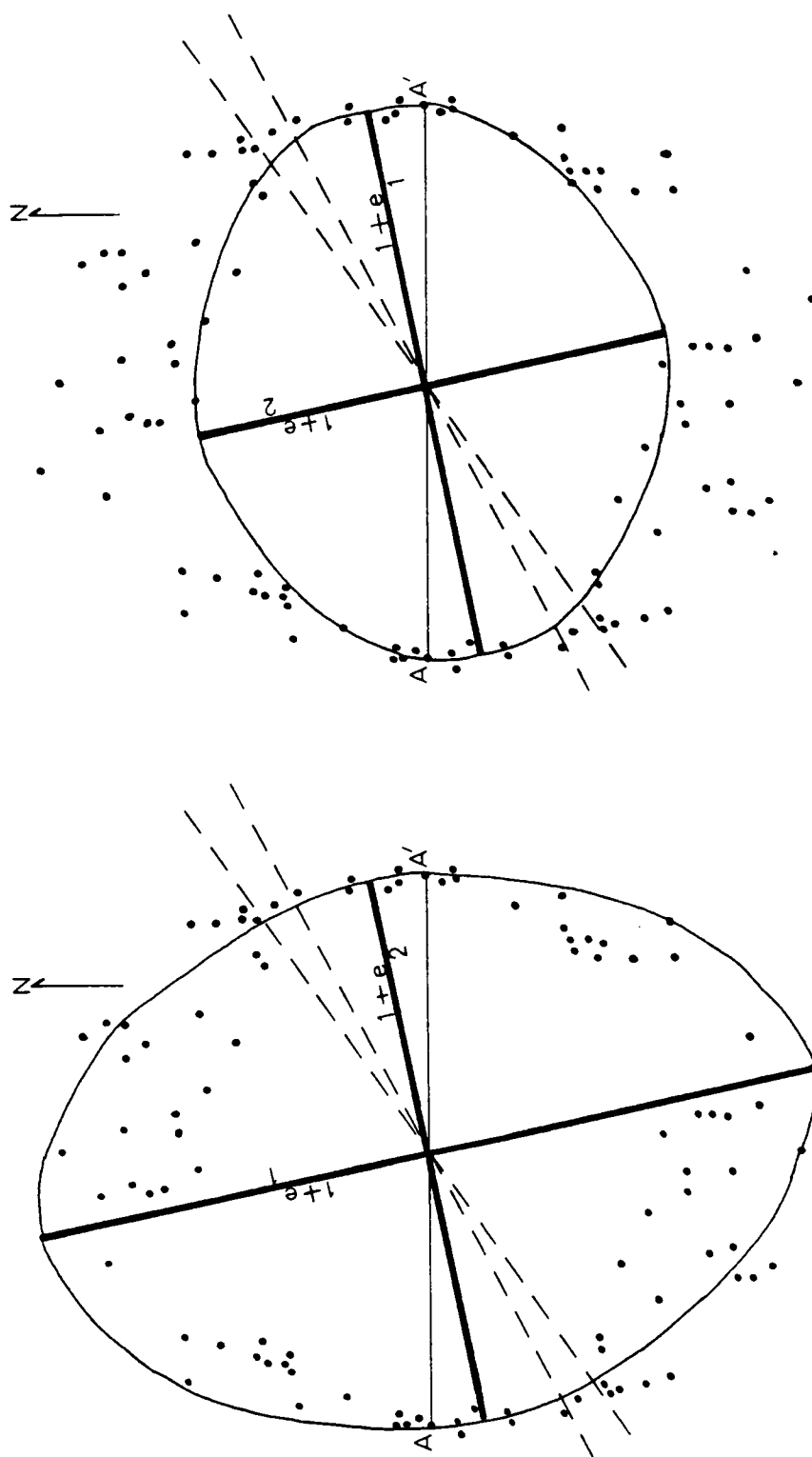


Figure 28. Plots of the Two Best-Fit Strain Ellipses for the Figure 27 Points.

Hunter Valley thrust. Each sample was analyzed using the Dunnet (1968) R_f/ϕ and the Fry (1979) center-to-center techniques. The Fry method gives values of whole rock or bulk rock strain (R_f), which includes both matrix and ooid strain. In contrast, the R_f/ϕ method gives values of ooid strain (R_s) and maximum original ooid ellipticities ($R_{i_{max}}$). The Fry (1979) and Dunnet (1968) methods and results of strain analysis are discussed below. Note that the three axes of the strain ellipsoid were not determined, because the computer program was unavailable.

Fry Method

The Fry (1979) center-to-center method is a rapid and accurate graphical solution of bulk rock (whole rock) strain (R_f) that derives the strain ellipse in the plane of a thin section (Ramsay and Huber, 1983, pp. 111-113). Three mutually perpendicular thin sections (R = parallel to strike, F = normal to strike, and T = parallel to bedding) were made from each of six oriented oolite samples. Assuming an undeformed regular hexagonal packing of the nearly circular ooids (circular in the plane of the thin section), one can calculate the strain ellipse by changes in the length of lines which connect neighboring ooid centers (Figure 29). Lines connecting neighboring centers which are parallel to the strain ellipse axis of maximum extension ($1 + e_1$) will be lengthened during oolite deformation. In contrast, lines connecting neighboring centers which are parallel to the strain ellipse axis of minimum extension ($1 + e_2$) will be shortened during oolite deformation. Therefore, the centers of ooids which lie upon lines parallel to the $1 + e_1$ axis will be moved farther apart whereas the centers of ooids

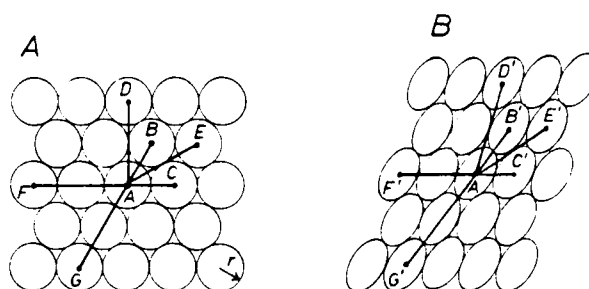


Figure 29. Theoretical Basis for the Fry Center-to-Center Technique (from Ramsay and Huber, 1983).

which lie upon lines parallel to the $1 + e_2$ axis will be moved closer together. This is the theoretical basis for the Fry center-to-center method of strain analysis.

The first step of the Fry (1979) method involved measuring the shortest and longest axes of fifty ooids from photomicrographs of each thin section in order to determine the location of the fifty ooid centers. These fifty ooid centers and axes were then recorded on a piece of tracing paper with an x-y coordinate grid. A central reference ooid was picked for each thin section and another x-y coordinate grid was set up on a transparent overlay (Figure 30). The central ooid was made the center of the transparent overlay, and the x-y coordinate axes were made parallel on both pieces of paper. All other ooid center positions were recorded on the transparent overlay. This process was continued until all fifty ooid centers had served as the center of the transparent overlay and all other ooid centers were recorded on the transparent overlay. A central vacancy with an elliptical shape was produced around the center of the transparent

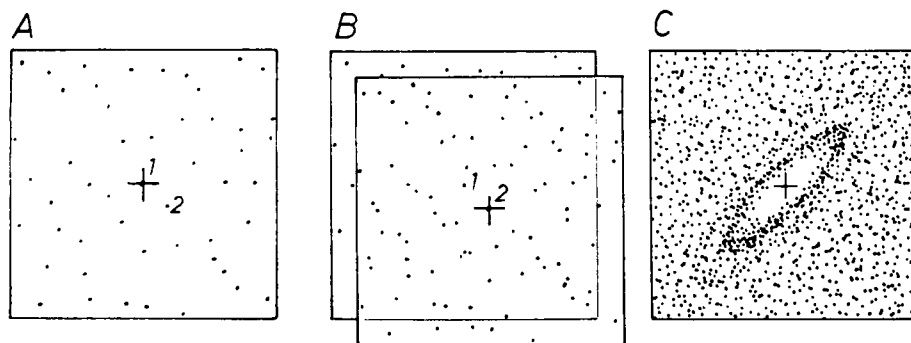


Figure 30. Steps in the Fry Center-to-Center Technique (from Ramsay and Huber, 1983, p. 113).

overlay, when the oolite was deformed. If the oolite was undeformed, the central vacancy produced around the center of the transparent overlay was circular. The axes of maximum extension ($1 + e_1$) and minimum extension ($1 + e_2$) were directly measured from the best fit ellipse, which could be drawn in the elliptical central vacancy. Using these axes lengths, the ratio of bulk rock strain ($Rf = 1 + e_1 / 1 + e_2$) was calculated.

The Fry method strains, which were determined for each of the three thin sections of each of the six samples, are presented in graphical and numerical forms (Figure 31 and Table 1). All of the strain ellipses were transferred to unfolded cubes with front, right, and top views which correspond to planes that are normal to strike, parallel to strike, and parallel to bedding. The ellipses and strain axes were rotated until the trace of bedding in the front and right views (as defined by layering of ooids and glauconite) was parallel to the base of the front and right sides of the unfolded cube (Figure 31). This

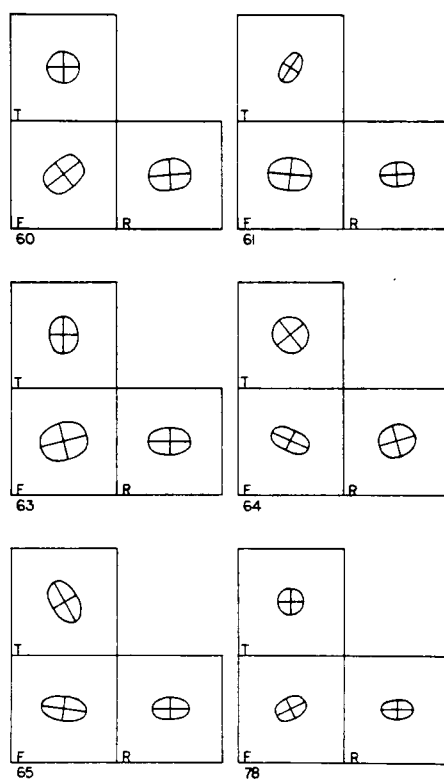


Figure 31. Unfolded Cubes Showing the Strain Ellipses Determined Using the Fry Center-to-Center Technique for Six Samples of the Rome Oolite.

Table 1. Table of Fry Method vs. Rf/Ø Method Results for Six Samples of the Rome Oolite.

Sample	Bulk Rock Strain (Rf) $Rf = 1 + e_1/1 + e_2$	Ooid Strain (Rs) and Maximum Original Ellipticity (Ri_{max})	Angle Between Bedding and the $1 + e_1$ Direction	
			Center-to-Center	Rf/Ø
60F	1.60	1.50	38°	9°
60R	1.36	1.20	5°	3°
60T	1.0	1.14	/	/
61F	1.40	1.41	4°	6°
61R	1.53	1.32	4°	17°
61T	1.71	1.16	/	/
63F	1.29	1.40	15°	6°
63R	1.55	1.29	0°	3°
63T	1.29	1.10	/	/
64F	1.87	1.30	24°	4°
64R	1.17	1.28	16°	13°
64T	1.08	1.46	/	/
65F	1.84	1.32	8½°	4°
65R	1.68	1.42	0°	0°
65T	1.70	1.19	/	/
78F	1.41	1.45	25°	7°
78R	1.66	1.38	0°	6°
78T	1.0	1.08	/	/

orientation produced three mutually perpendicular planar views of the samples, which were oriented relative to bedding. The calculated bulk rock strains (R_f) for the six samples and the angle measured between bedding and the direction of maximum extension ($1 + e_1$) are presented in Table 1.

The results of the Fry (1979) method are that bulk rock strains in the Rome oolite show maximum extension in planes parallel and perpendicular to strike and minimum extension in planes parallel to bedding (Figure 31 and Table 1). Average strain ratios for the planes normal to strike (F), parallel to strike (R), and parallel to bedding (T) are 1.57, 1.5, and 1.3. The average angular deviation of the trace of bedding from the long axis of the strain ellipse ($1 + e_1$) in planes parallel and perpendicular to strike is 11.6 degrees (Table 1). There are also numerous calcite-filled extension fractures within the oolite which are oriented sub-normal to the $1 + e_1$ axes (Figure 32). Based on these data, the oolite deformation in the planes parallel and perpendicular to the strike of bedding is attributed to stratigraphic overburden. The direction of maximum overburden stress is oriented sub-perpendicular to bedding. A small contractional bedding wedge which duplicates the oolite in outcrop appears to have no effect on bulk rock strain (Figure 33). Samples from above and below the bedding wedge consistently show that the directions of maximum extension are within bedding.



Figure 32. Calcite Filled Bedding Normal Extension Fractures in the Rome Oolite. Field of view is 3.2 mm by 2.15 mm.



Figure 33. Bedding Wedge in the Rome Oolite and Location of the Six Samples. Field of view is 30 m.

Rf/ $\emptyset$ Method

The Dunnet (1968) Rf/ $\emptyset$ method gives values of ooid strain independent of the surrounding matrix (R_s) and the maximum original ellipticity of the ooids prior to deformation ($R_{i_{max}}$). Three mutually perpendicular thin sections (parallel to strike, normal to strike, and parallel to bedding) were used for strain analysis of each of the six oriented Rome oolite samples. The first step of the method involved measuring the shortest and longest axes of fifty ooids from photomicrographs of each thin section. These measurements of the long ($1 + e_1$) and short ($1 + e_2$) axes were used to calculate the finite strain ratio ($R_f = 1 + e_1 / 1 + e_2$) for each ooid. Next, the angular orientation ($\emptyset$) of the long axis of each ooid was measured relative to some fixed axis for each thin section. This step was made possible by first tracing the ooids from the photomicrographs and onto a transparent overlay with an x-y grid. These values of Rf and $\emptyset$ for each of the fifty ooids per thin section were plotted on the Rf/ $\emptyset$ graphs (Figure 34) for six samples.

The next step of the Dunnet (1968) method involved finding the average orientation of the long axes of all fifty ooids in a thin section. This average long-axis orientation was found by constructing frequency histograms of $\emptyset$ at 5 degree intervals plotted against the frequency of ooid orientation (Figure 35). If the $\emptyset$ vs orientation frequency histograms showed a symmetric distribution, then the initial orientation of the ooid long axes was random prior to deformation (Ramsay and Huber, 1983, p. 82). A non-random initial orientation of

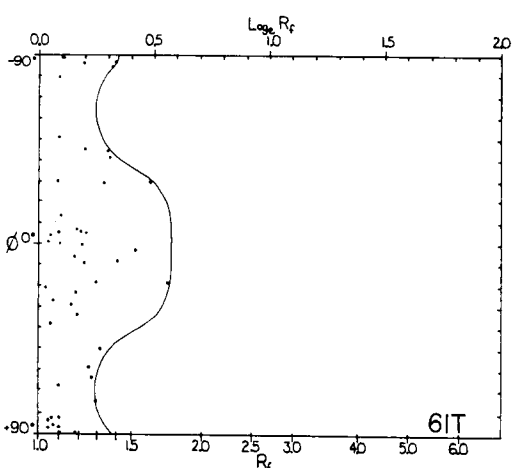
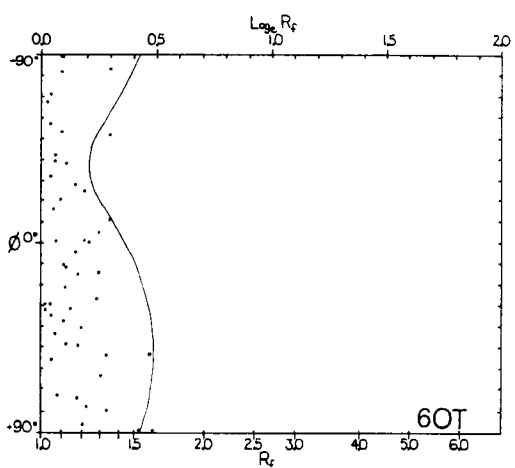
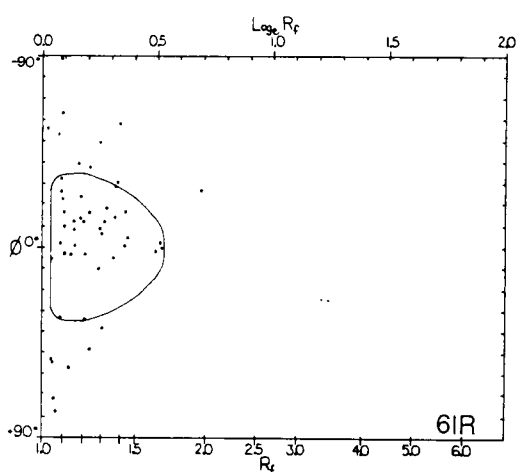
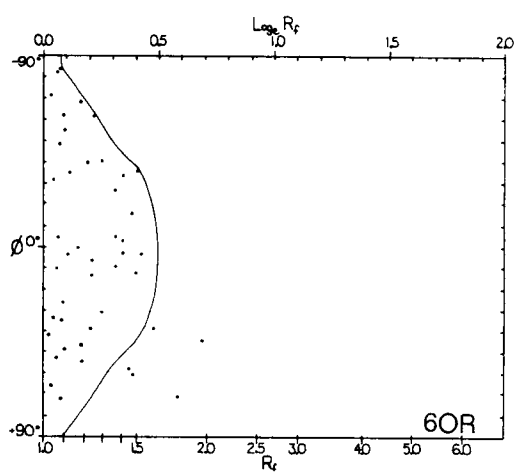
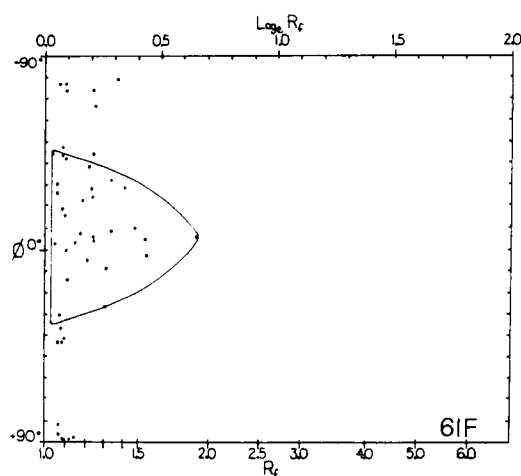
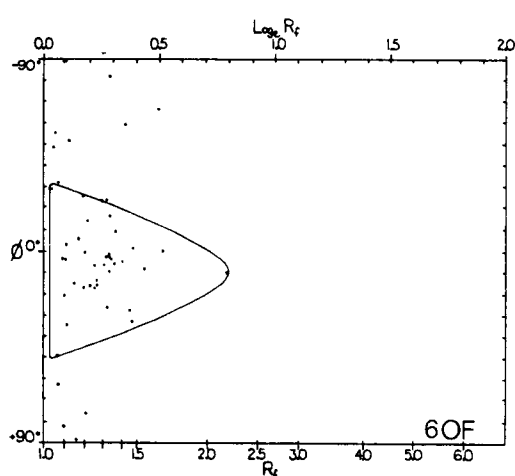


Figure 34. R_f/ϕ Plots for Oolite Samples 60, 61, 63, 64, 65, and 78.

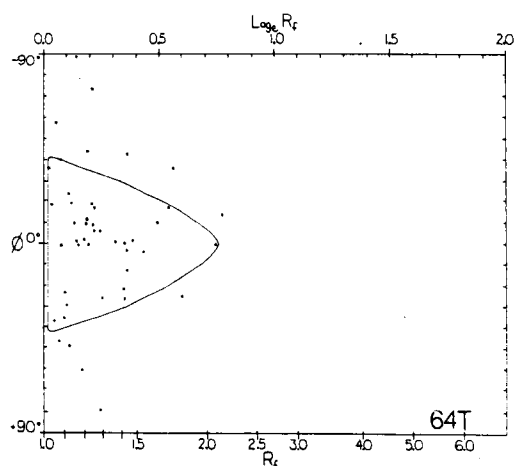
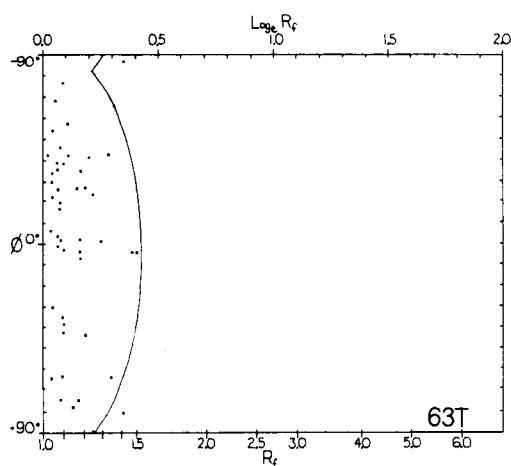
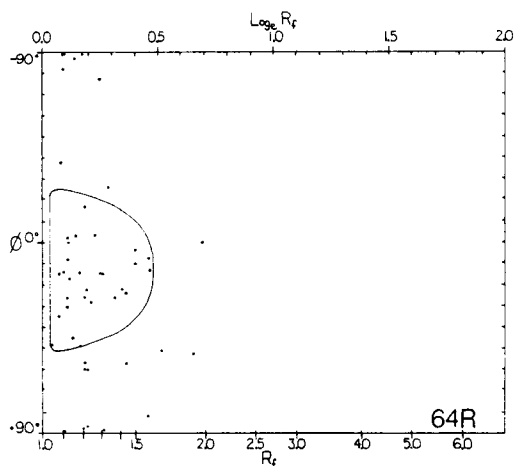
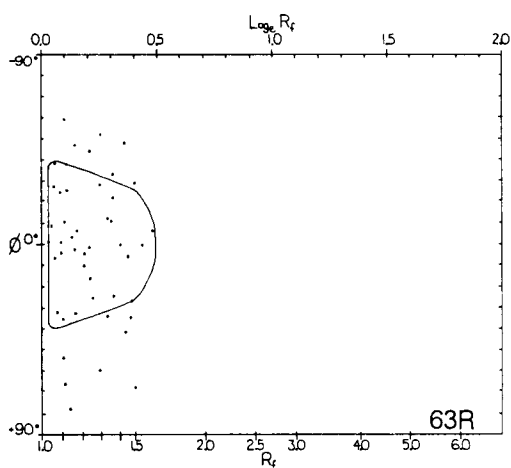
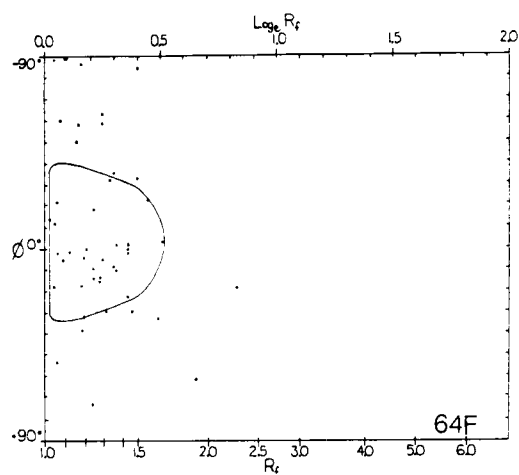
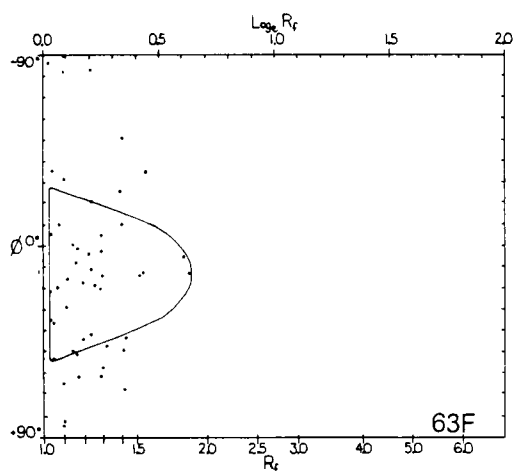


Figure 34 (continued)

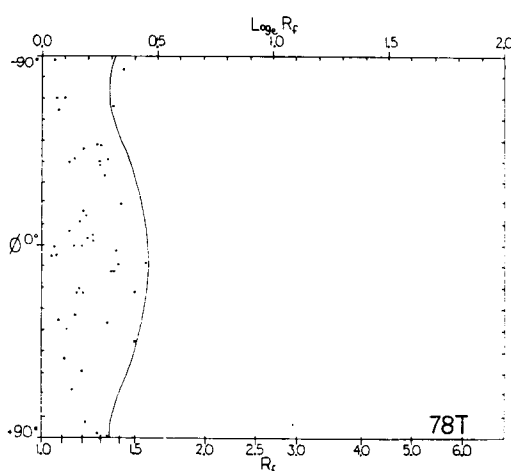
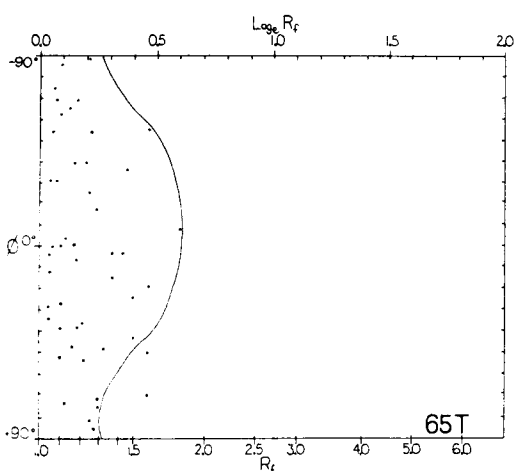
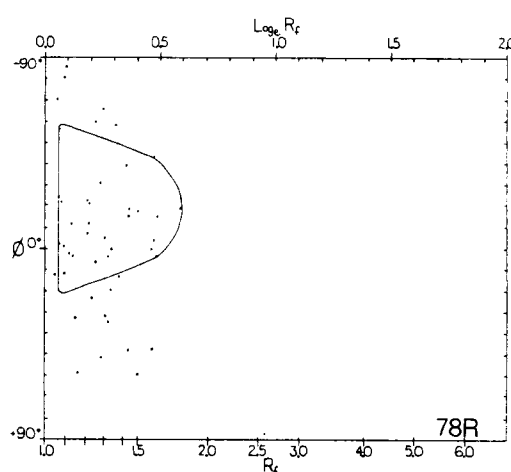
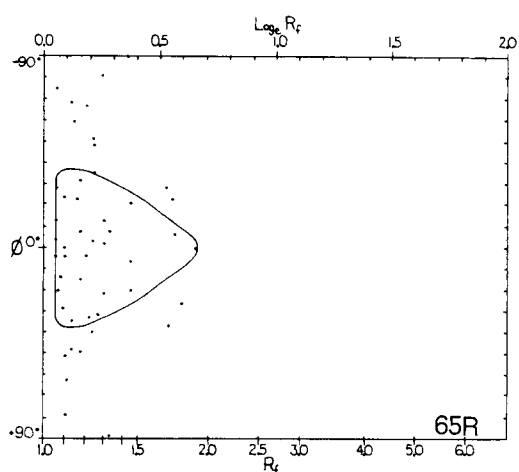
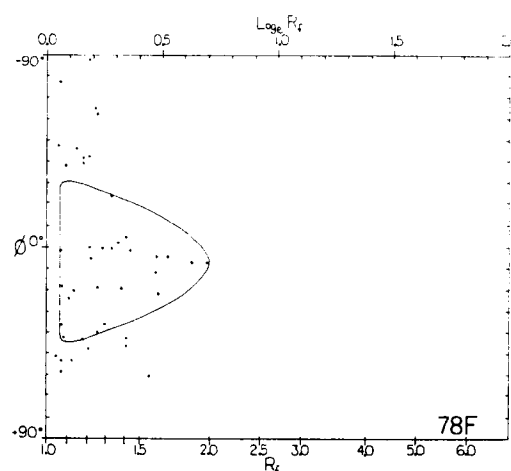
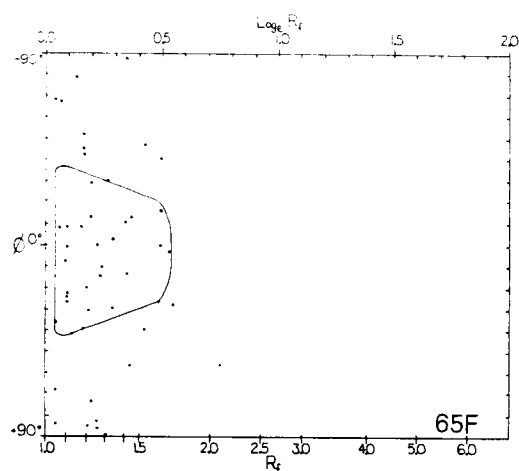


Figure 34 (continued)

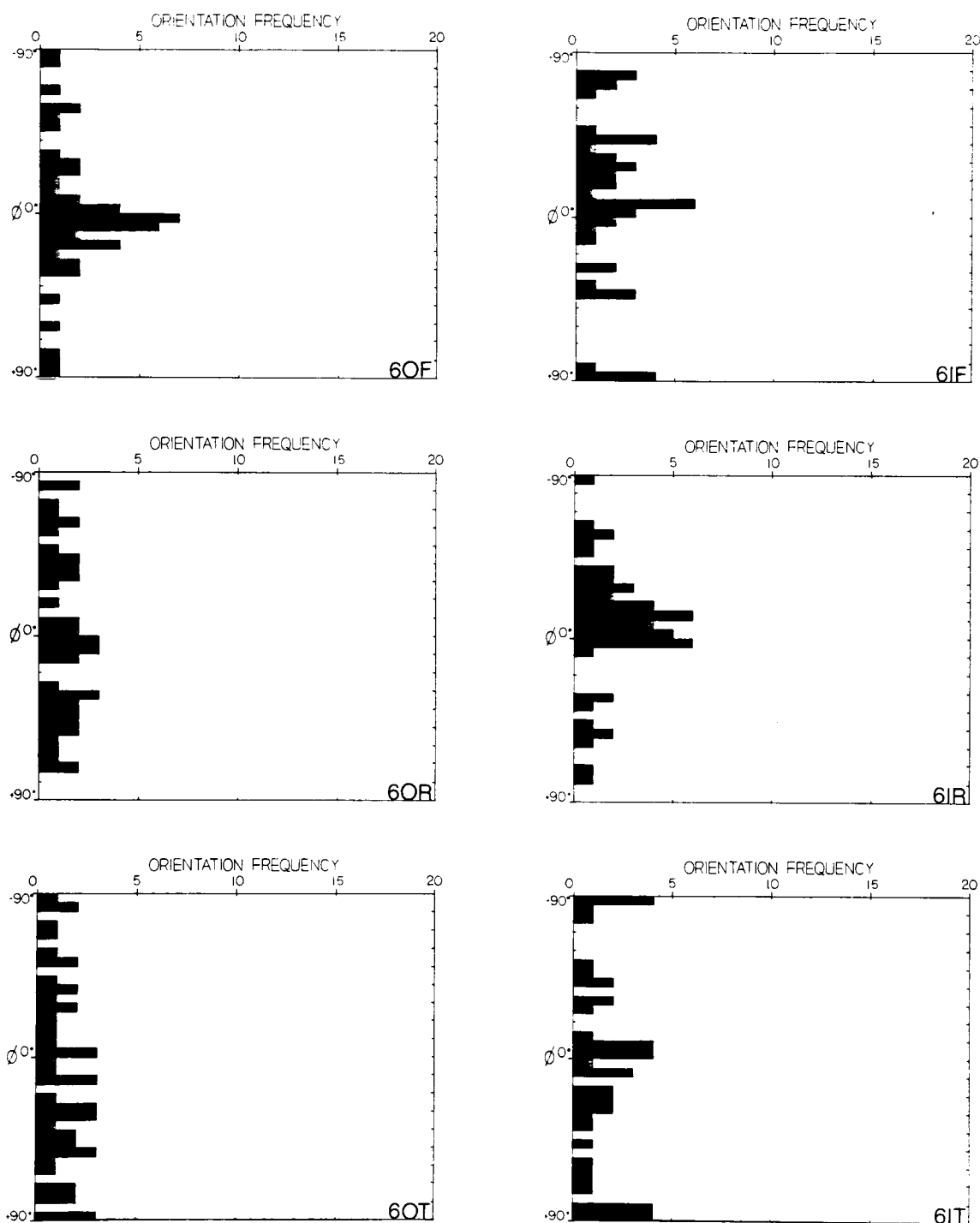


Figure 35. Orientation Frequency vs $\emptyset$ Plates for Oolite Samples 60, 61, 63, 64, 65, and 78.

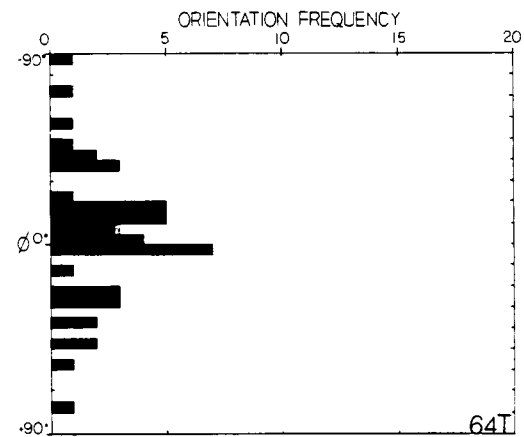
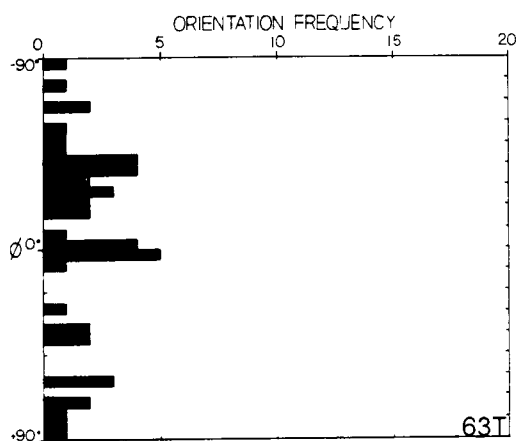
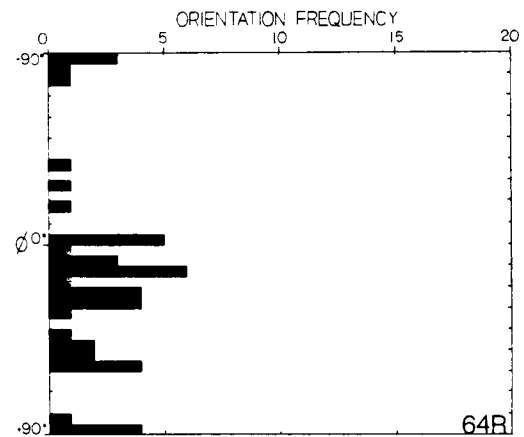
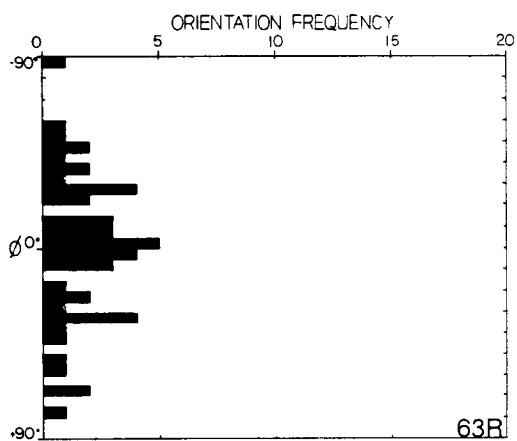
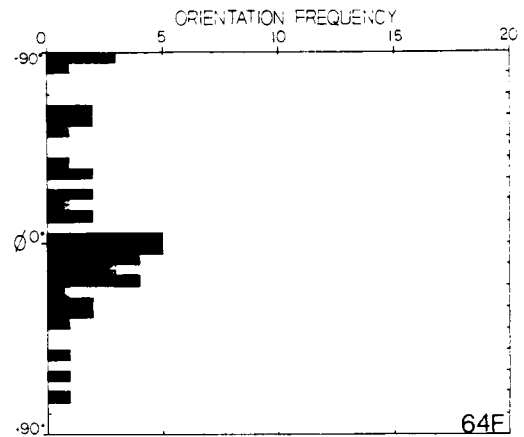
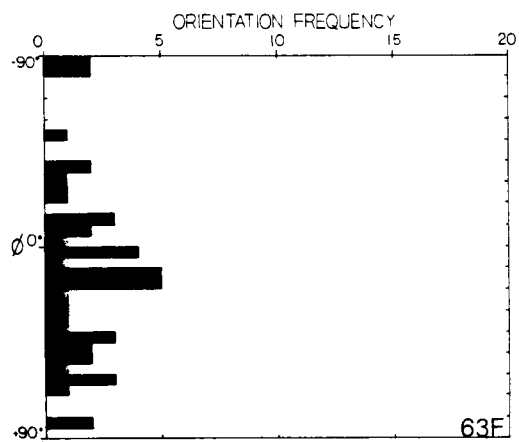


Figure 35 (continued)

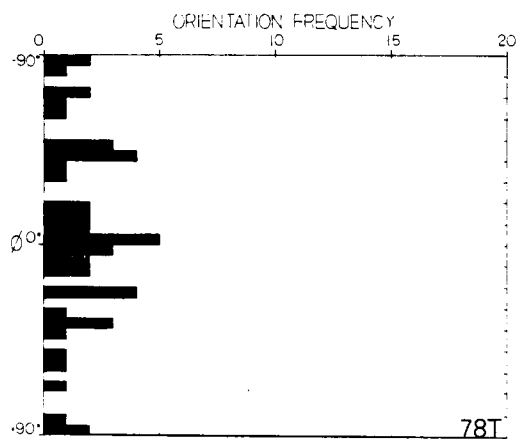
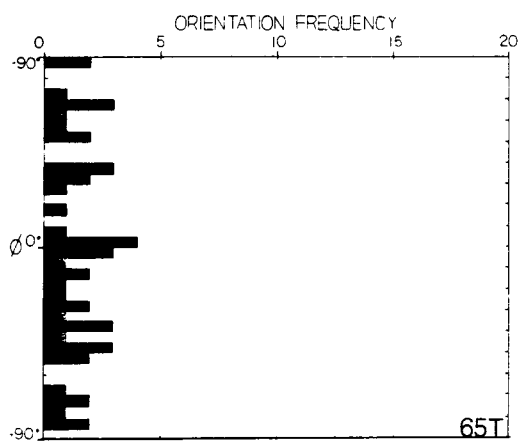
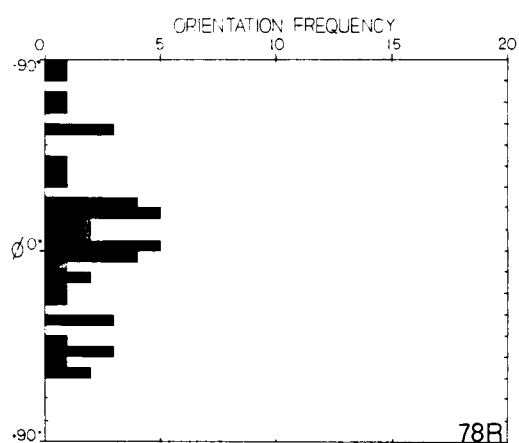
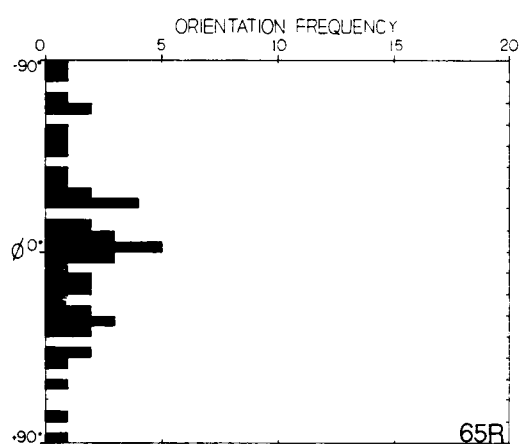
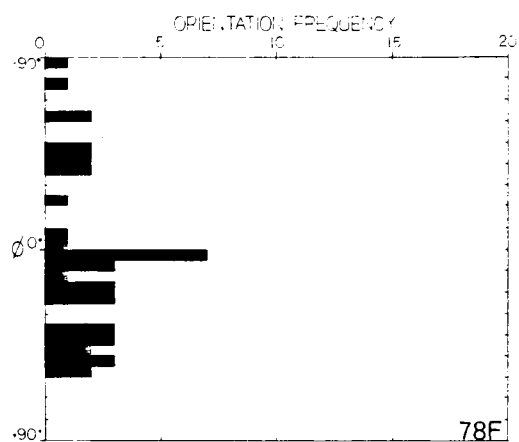
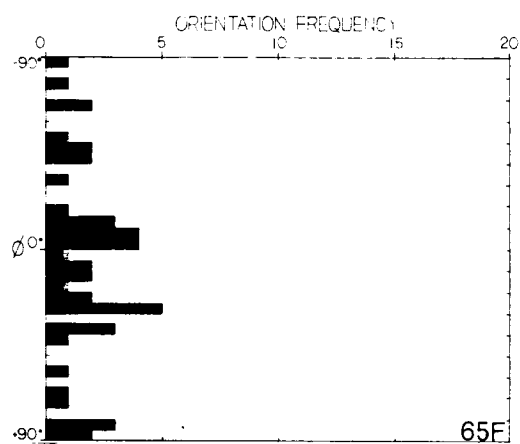


Figure 35 (continued)

oid long axes, either due to sedimentation or diagenesis, was indicated by an asymmetric $\emptyset$ vs orientation frequency plot. The strain analysis of ooids with preferred-orientation long axes is a complicated process, which was discussed by Ramsay (1967, pp. 218-221). Fortunately, all samples of the Rome oolite showed a random initial orientation of ooid long axes. The average long axis orientation for each thin section was plotted on its $Rf/\emptyset$ graph (Figure 34). These $Rf/\emptyset$ graphs were then compared with mathematically derived $Rf/\emptyset$ graphs given by Ramsay and Huber (1983, p. 76). The mathematically derived $Rf/\emptyset$ graph whose $Ri_{\max}$ curves best fit the Rome oolite $Rf/\emptyset$ data points was determined, and the maximum original ooid ellipticity ($Ri_{\max}$) and ooid strain (Rs) were read from the Ramsay and Huber (1983) graph.

An alternative method for directly calculating the values of $Ri_{\max}$ and Rs was given by Ramsay and Huber (1983, p. 77). This method involved the construction of $Rf/\emptyset$ curves, which enclosed the Rome oolite $Rf/\emptyset$ data points. The advantage of this method was that one did not have to interpolate $Ri_{\max}$ and Rs values from mathematically derived $Rf/\emptyset$ graphs. This technique was more useful when very accurate values of Rs and $Ri_{\max}$ were desired and was used in this study to check the values of Rs and $Ri_{\max}$, which were determined using mathematically derived graphs. In this method, the values of $Rf_{\max}$ and $Rf_{\min}$ as defined by the Rome oolite $Rf/\emptyset$ graphs were substituted into one set of two sets of equations (Figure 36). The range of orientation of long axes (F = fluctuation) determined which set of equations was used. If the fluctuation was less than 90 degrees, then $Rs > Ri_{\max}$ and the

proper equations to use were $Rs = \sqrt{(Rf_{\max})(Rf_{\min})}$ and $Ri_{\max} = \sqrt{\frac{Rf_{\max}}{Rf_{\min}}}$.

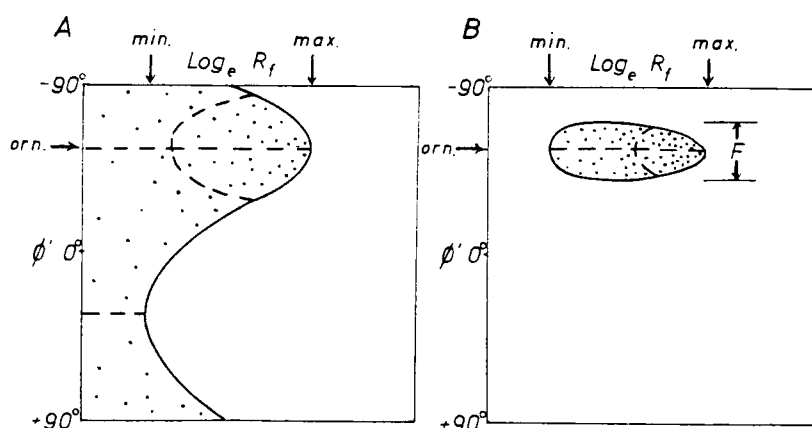


Figure 36. Theoretical Basis for the R_f/ϕ Technique (from Ramsay and Huber, 1983, p. 77).

When the fluctuation was 90 degrees, then $R_s = R_{i_{\max}}$ and the oolite deformation could not be calculated. If the fluctuation was greater than 90 degrees, then $R_{i_{\max}} > R_s$ and the proper equations to use were

$$R_s = \sqrt{\frac{R_{f_{\max}}}{R_{f_{\min}}}} \quad \text{and} \quad R_{i_{\max}} = \sqrt{(R_{f_{\max}})(R_{f_{\min}})}. \quad \text{These equations were}$$

based on the fact that individual strains on a body which act in the same direction are multiplied and strains which act in opposite directions are divided. Therefore, the maximum finite strain ratio ($R_{f_{\max}}$) observed in an ooid was the product of the maximum original ellipticity ($R_{i_{\max}}$) and the strain imposed in the same direction upon the ooid during deformation (R_s). Similarly, the minimum finite strain ratio ($R_{f_{\min}}$) observed in an ooid was the quotient of the maximum original ellipticity ($R_{i_{\max}}$) divided by the strain imposed in the opposite direction during deformation (R_s). These relationships were stated

mathematically (Ramsay and Huber, 1983, p. 77) using the equations

$$Rf_{\max} = (Rs)(Ri_{\max}) \text{ and } Rf_{\min} = (Ri_{\max}/Rs).$$

The results of the $Rf/\emptyset$ method of ooid strain analysis (Table 1) show that the ooid strains (Rs) are smaller than the bulk rock strains (Rf), which are determined using the Fry (1979) method. However, both the Fry (1979) and Dunnet (1968) methods have maximum extension in the planes that are parallel and perpendicular to strike. The average $Rf/\emptyset$ ooid strains (Rs) for the planes normal to strike (F), parallel to strike (R), and parallel to bedding are 1.4, 1.3, and 1.0. Angular deviation of the long axis of the strain ellipse from the trace of bedding, in the planes parallel and perpendicular to strike, averages 6.5 degrees. Based on these data, the oolite deformation in the planes parallel and perpendicular to the strike of bedding is attributed to stratigraphic overburden. This interpretation is supported by the orientation of the long axes of most ooids sub-normal to numerous bedding-perpendicular calcite-filled extension fractures (Figure 32, page 62). A small contractional bedding wedge which duplicates the oolite in outcrop appears to have no effect on the ooid strains (Figure 33, page 62). The apparent deformation in the plane parallel to bedding which was determined using the Fry center-to-center method is attributed to original ellipticity (Table 1, page 60).

Unfolding Mesoscopic Folds

A large outcrop of mesoscopic folds can be used as a small-scale, detailed structure section when the outcrop is viewed normal to the fold axes (Ramsay, 1969). Like all structure sections, the mesoscopic

folds were pinned on one side and unfolded in order to determine the bedlength-shortening during flexural-slip folding (Ramsay, 1967, p. 387; and Dahlstrom, 1969, p. 746). Ramsay (1969, p. 62) gave the conditions under which the directly measured length of a folded bed in cross-section records the original length of the bed. These conditions were: (1) folding via flexural slip between layers; (2) simple shear between bedding planes; (3) no finite strain within the plane parallel to the folded axes; (4) and no volume loss during plane strain. Flexural slip folding via simple shear between beds was assumed because the mesoscopic folds had parallel or chevron styles (Figure 37). Those beds

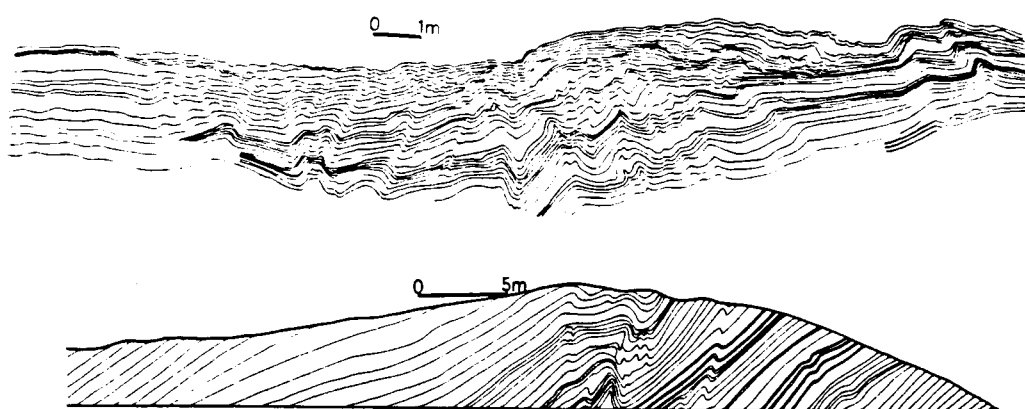


Figure 37. Mesoscopic Folds Used in Bedlength-Shortening Calculations.

which exhibited similar folding via flexural flow were not used in bedlength-shortening calculations. Plane strain was assumed for each outcrop, and therefore the finite strain within the plane parallel to

fold axes was zero. The lack of cleavage within the folded sandstones, siltstones, and shales made negligible the problem of volume loss.

The strain analysis technique involved: (1) measuring all fold axes in the outcrop; (2) determining the average fold axis orientation on a Schmidt net; (3) photographing the outcrop on a line normal to the average fold axis orientation; (4) transferring the folded layers from a photo-mosaic onto a piece of tracing paper; (5) and restoring the beds to their original position using a planimeter. Bedlength-shortening strain within the plane normal to the average fold axis was calculated as a percentage using the equation percentage shortening = $\frac{(L_i - L)}{L_i} \times 100$, where L_i and L were the undeformed and deformed lengths occupied by the layers.

The method above was used to unfold mesoscopic folds along roadcuts of similar lithologies in the hangingwall of the Wallen Valley thrust and the footwall of the Whiteoak Mountain thrust. Unfolding of the mesoscopic folds in the Rome hangingwall of the Wallen Valley fault (maximum displacement of 8.5 km in section A-A', Plates 2 and 3) shows maximum and minimum bedlength-shortening values of 14 and 3.8 percent (Figure 38). In contrast, unfolding the mesoscopic folds in the Rockwood footwall of the Whiteoak Mountain fault (maximum displacement of 9.87 km in section D-D', Plates 2 and 3) shows maximum and minimum bedlength shortening values of 21 and 1.4 percent (Figure 38). These data indicate that the mesoscopic folds, which were caused by thrust sheet movement in the study area, have greatest shortening associated with the greatest displacement faults. Therefore, the greater

displacement of the Whiteoak Mountain thrust, which was calculated using restored sections, is reflected in footwall deformation intensity.

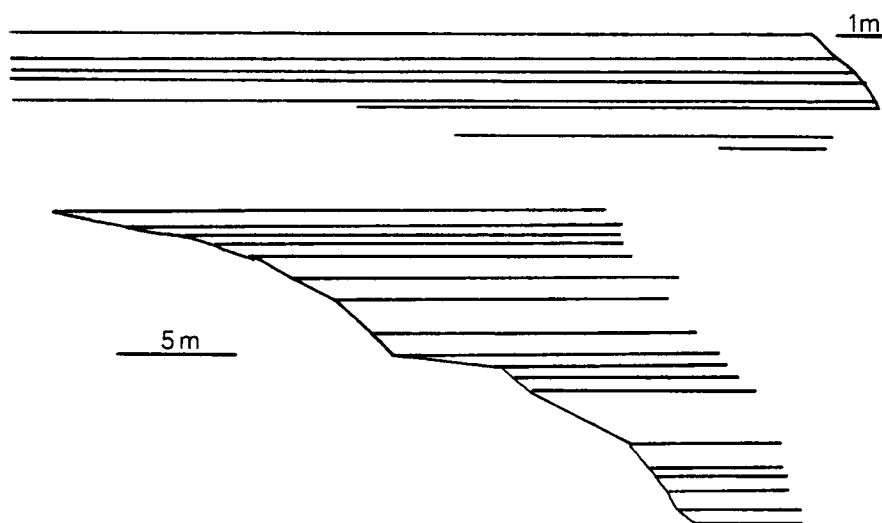


Figure 38. Folded Beds Restored to Original Horizontality.

Summary and Conclusions

The conclusions which can be drawn from strain analyses on rocks in the study area are (1) the maximum displacements on the Hunter Valley, Wallen Valley, and Whiteoak Mountain thrusts are 2.44 km, 8.5 km, and 9.87 km; (2) strain markers (Rockwood brachiopods) which are proximal to large displacement thrust faults show greater deformation than strain markers (Rome oolite) which are proximal to small displacement thrust faults; and (3) deformation via macroscopic and mesoscopic folding is greatest in the hangingwall and footwall rocks of the

greatest displacement thrust faults. In summary, deformation in the study area is controlled by the number of thrust faults and the displacement on these thrusts. The maximum strains determined from unfolding macroscopic folds, unfolding mesoscopic folds, strain analysis on ooids, and strain analysis on brachiopods are 29%, 21%, 40%, and 43%.

CHAPTER IV

SUMMARY AND CONCLUSIONS

Structural Interpretation and Results of Strain Analysis

Structural Interpretation

Detailed mapping of the rocks in the area of the Hunter Valley, Wallen Valley, and Whiteoak Mountain fault intersection indicates that the Wallen Valley thrust is a footwall imbricate of the Hunter Valley thrust (Plates 1-3). These two faults merge to the southwest and form the Whiteoak Mountain thrust. A footwall horse of Ordovician age is brought to the surface where the Wallen Valley thrust joins the Hunter Valley thrust. Many trailing and leading imbricates occur within the footwall and hangingwall of the major thrusts. The trailing-edge imbricates (Dahlstrom, 1970; Boyer and Elliot, 1982) produced duplication in the mapped thicknesses of the Sequatchie and Rockwood formations (Plates 1 and 2).

Stratigraphic and structural evidence support a Hunter Valley and then Wallen Valley sequence of thrust faulting. Correlation of two carbonate units (16.8 m and 11.4 m thick) of the Cambrian Rome Formation in the hangingwall of both thrust sheets indicates that a nearly complete section of Rome (457 m) is brought to the surface by the Hunter Valley thrust. In contrast, the thickness of Rome brought to the surface by the Wallen Valley thrust is less (396 m) and

decreases to the southwest as the Wallen Valley cuts upsection in its hangingwall (Plate 2 sections A-A' and B-B'). Folding of the Hunter Valley thrust by the younger Wallen Valley thrust below is indicated by large wavelength (336 m) and amplitude (61 m) folds in the hangingwall of the Hunter Valley thrust (Plate 2, section B-B'). This folding of the Rome in the hangingwall of the Hunter Valley thrust produces a very wide outcrop width, without major thrust duplication.

The Wallen Valley fault cuts upsection in its hangingwall from northeast to southwest (Hardeman, 1966). Upon cutting upsection to the pre-existing Hunter Valley décollement, the Wallen Valley thrust rejoins the Hunter Valley thrust, and together they are renamed the Whiteoak Mountain thrust (Plate 2, section C-C'). The last remnant of the Wallen Valley hangingwall is a footwall horse of Ordovician rocks northwest of the Whiteoak Mountain fault (Plates 1 and 2, sections C-C' and D-D'). This horse (fault bound slice) dies out southwest of Oak Ridge, Tennessee (Hardeman, 1966). All along strike, the horse folds the footwall flat of the Whiteoak Mountain thrust above into tight (short wavelength and high amplitude) folds. In this geometry, the displacements of the Hunter Valley and Wallen Valley thrusts are transferred to the Whiteoak Mountain thrust (Plate 3).

Strain Analysis Results

Strain analysis was done on hangingwall and footwall rocks in the study area using mesoscopic Rome and Rockwood folds, deformed Rockwood brachiopods, a deformed Rome oolite, and balanced cross sections. These techniques are discussed fully in the section entitled Strain

Variations Within Rocks in the Study Area. The strain values derived using these four methods are given below.

Mesoscopic folds along roadcuts of similar lithologies in the Cambrian Rome and Silurian Rockwood Formations were unfolded normal to the fold axes in order to determine bedlength-shortening values (Ramsay, 1967, p. 387). Unfolding of mesoscopic folds in the upper Rome (Eru) hangingwall of the Wallen Valley thrust derives maximum and minimum bed shortening values of 14% and 3.8%. Similar lithology mesoscopic folds in the Silurian Rockwood (Sr) footwall of the Whiteoak Mountain thrust have maximum and minimum bedlength-shortening values of 21% and 14%. These strains suggest that mesoscopic folds in similar lithologies have greater bedlength-shortening values associated with greater displacement faults.

The deformed Rockwood brachiopods in the folded and faulted hangingwall of a small imbricate which duplicates the Rockwood section along I-75N were analyzed for strain using the Wellman (1962) method. These brachiopods show a maximum extension of $R_f = 1.43$ sub-normal to the strike of bedding and a minimum extension of $R_f = 1.15$ sub-parallel to the strike of bedding. The observed strains are attributed to extension in the outer arcs and contraction in the cores of concentric minor folds. These parallel-style folds (Ramsay, 1967, p. 365) deform the 3.5 m thick zone of interbedded limestone and shale that overlies the fault surface.

The deformed upper Rome (Eru) oolite in the hangingwall of the Hunter Valley thrust along I-75N was analyzed for bulk rock strain (R_f), ooid strain (R_s), and maximum original ooid ellipticity ($R_{i_{max}}$).

These values were determined using the Fry (1979) center-to-center method and the Dunnet (1968) $R_f/\emptyset$ method. Strain analysis indicates that bulk rock strains (R_f) are consistently greater than the ooid strains (R_s). Bulk rock strains show maximum extension in planes parallel (R) and perpendicular (F) to the strike of bedding and minimum extension in the plane parallel to bedding (T). These strains average 1.57, 1.5, and 1.31 within planes F, R, and T. The ooid strains are smaller, but they show the same maximum extension in planes F and R. These strains average 1.40, 1.3, and 1.0 within planes F, R, and T. Both bulk rock and ooid strains show maximum extension in directions sub-parallel to bedding and sub-normal to numerous calcite-filled extension fractures within the oolite. These data suggest that deformation was caused by the weight of stratigraphic overburden. A small bedding-wedge fault (Cloos, 1964), which duplicates the oolite in outcrop, appears to have no effect on either bulk rock or ooid strains.

Balanced cross-sections (Plate 3) show maximum displacements of 2.44 km, 8.5 km, and 9.87 km for the Hunter Valley, Wallen Valley, and Whiteoak Mountain thrusts. Maximum shortening values for the Hunter Valley, Wallen Valley, and Whiteoak Mountain faults are 53%, 59%, and 76%. The large amount of displacement and shortening on the Whiteoak Mountain thrust results from the merging of the Hunter Valley and Wallen Valley thrusts (Plates 2 and 3, section C-C'). Displacement and shortening values for the Hunter Valley and Wallen Valley thrusts increase northeast of the intersection zone (Roeder et al., 1978). Similarly, the displacement and shortening of the Wallen Valley thrust

decreases southwesterly toward the fault intersection zone, where the Wallen Valley transfers its displacement to the older Hunter Valley thrust.

Conclusions

The conclusions which can be drawn from the synthesis of both strain and structural data are (1) footwall deformation via trailing imbricates and mesoscopic folding is greater than hangingwall deformation via leading imbricates and mesoscopic folding, (2) more intense macroscopic folding occurs in the older Hunter Valley thrust hangingwall due to folding by the younger Wallen Valley thrust below, (3) the Wallen Valley thrust transfers its displacement to the Hunter Valley thrust and forms the Whiteoak Mountain thrust, (4) whole rock strains are greater than the strains which are recorded by the individual ooids in the Rome Formation oolite, and (5) strain markers (Rockwood Formation brachiopods) which are proximal to large-displacement thrust faults show greater deformation than strain markers (Rome oolite) which are proximal to small-displacement thrust faults.

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LIST OF REFERENCES

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APPENDIXES

APPENDIX A

GUIDE TO SELECTED OUTCROPS IN THE STUDY AREA

Stop 1: Contact Between Conasauga Group and Upper Rome Formation at Pine Ridge

There are 290 m of upper Rome Formation exposed at Stops 1-2 in the hangingwall of the Hunter Valley thrust. The contact between the Conasauga and upper Rome is drawn at the top of the uppermost medium-bedded yellow-gray Rome sandstone. The most common to least common lithologies in the upper Rome are sandstone, siltstone, shale, limestone and dolomite. Samman (1975) determined the sandstone and siltstone to shale ratio (3:1) and the clastic to carbonate ratio (8:1) for this outcrop. The upper Rome at this stop is on the southeastern limb of an anticline in the hangingwall of the Hunter Valley thrust. Deformation in the outcrop is restricted to minor extensional (normal) faults and contractional (bedding wedge) faults. Minor folding occurs above the bedding wedges. The dolomitic limestones (16.8 m thick) which occur 143 m below the Rome-Conasauga contact were mapped for 8 km along strike and were correlated across strike to the hangingwall of the Wallen Valley fault (a distance of 4.3 km based on restored sections). These limestones in the hangingwall of the Wallen Valley fault were mapped along strike for 5.5 km. A dolomitic oolite which occurs 60 m below the Rome-Conasauga contact is duplicated by a

small bedding-wedge fault. Strain analysis on the ooids shows that deformation is within the planes normal to bedding. The average strains in planes parallel to strike (R), normal to strike (F), and parallel to bedding (T) are 1.40, 1.30, and 1.0. Average angular deviation of the $1+e_1$ direction from the trace of bedding is 6.5 degrees. These data suggest that deformation was caused by the weight of stratigraphic overburden.

Stop 2: Small Displacement Thrusts and Folds in the
Upper Rome Formation at Pine Ridge

There are two small displacement contractional (thrust) faults exposed in this outcrop, which is on the southeastern limb of an anticline in the hangingwall of the Hunter Valley thrust. The uppermost fault within the sandstones (strike = N47E, dip = 65° SE) can be mapped for .9 km along strike. Note the fold within the maroon shale footwall of the lowermost fault. Several calcite-filled extension fractures occur on the outer arc of this fold. The sandstone and siltstone to shale ratio for this outcrop is 1:1.

Stop 3: Exposure of a Trailing Imbricate of the
Wallen Valley Thrust Which Places Rockwood
on Rockwood at Lone Mountain

This trailing imbricate is the most northwestern of three imbricates which duplicate the Rockwood Formation (true stratigraphic thickness of 152 m or 500 ft) to a maximum structural thickness of 457 m or 1500 ft. The décollement surface is the interbedded dark gray shales

and limestones at the base of the Rockwood Formation. Note the small footwall horse in the upper portion of the outcrop below the main thrust fault and the minor folds in the hangingwall and footwall. The main thrust strikes N55E and dips 60SE and was mapped for 6.7 km along strike. Fold axes in the hangingwall and footwall trend N30-40E and plunge 7-19° SW and NE. Strain analysis was done on brachiopods within the shaley limestones above the thrust fault. The brachiopods showed a minimum extension of 1.15 subparallel to strike and a maximum extension of 1.43 subnormal to strike. These strains are attributed to extension in the outer arcs and contraction in the cores of concentric minor folds.

Stop 4: Antithetic Thrust Fault and Folds in
the Upper Rome Formation Within the
Hangingwall of the Wallen Valley
Thrust

The small-displacement antithetic (northwest-dipping) thrust fault, which places upper Rome on upper Rome, strikes N70E and dips 50° NW. This thrust can be mapped for 1.4 km along strike. Other deformation features in this exposure are extensional (normal) faults and contractional (bedding wedge) faults. The shales, siltstones, and sandstones of the upper Rome form chevron, parallel, and similar style folds. Note the bulbous hinge structures and limb thrusts in the chevron folds above the antithetic thrust. These structures are formed when folding accommodates minor changes in the thickness of the more competent layers. Parallel folds in the upper portion of

the outcrop have maximum bedlength-shortening values of 14 percent and are produced by the numerous bedding wedges below.

Stop 5: Small-Displacement Thrust in the Lower

Rome Formation at Pine Ridge

The small-displacement thrust (30 m) which duplicates a 10 m interval of lower Rome Formation shales, siltstones, and sandstones strikes N41E and dips 71° SE. This thrust can be mapped for 3.4 km along strike. Several extensional (normal) faults, contractional (thrust) faults, and minor folds occur in the hangingwall and footwall. There are 82 m of lower Rome Formation exposed in this outcrop. Note that the three outcrops at Pine Ridge (Stops 1, 2, and 5) expose a total of 417 meters of Rome Formation. A sandstone and siltstone to shale ratio of 2:1 was determined for this outcrop. The boundary between the upper Rome and lower Rome Formation is drawn at the base of the thick-bedded limestone in the uppermost portion of the outcrop. This dolomitic limestone (11.4 m thick), which occurs 324 m below the Rome-Conasauga contact, was mapped for 5 km along strike and was correlated across strike to the hangingwall of the Wallen Valley thrust (a distance of 4.3 km based on restored sections). The limestone was mapped for 5.5 km along strike in the hangingwall of the Wallen Valley thrust.

Stop 6: Chevron Folds in the Silurian Rockwood
Formation at Lost Ridge

Chevron folds in the Rockwood Formation were mapped for a distance of 3.7 km along strike to the northeast. These folds were produced by trailing imbricates of the Wallen Valley fault which cut to the surface to the northeast and cause structural duplication of the Sequatchie and Rockwood Formations. Note the numerous calcite-filled extension fractures on the outer arcs of the folded sandstones and siltstones. Strain analysis on these folds indicates a maximum bedlength shortening value of 21 percent. The contact between the Rockwood and the underlying Ordovician Sequatchie Formation is drawn at the top of the uppermost bed of maroon Sequatchie limestone. This contact is visible along the railroad outcrop which parallels Clinton Highway.

Stop 7: Small-Displacement Thrusts in the Upper
Rome Formation at Pine Ridge

The two small-displacement thrusts, which twice duplicate a 3 m interval of upper Rome Formation sandstones and shales, strike N60E and dip 86° and 58° SE. These two thrusts merge in the top of the outcrop and produce a small horse. The thrust zone made by these two thrusts was mapped for 1.5 km along strike to the northeast and then died out into an anticline. Extensional (normal) faults occur in both the hangingwalls and footwalls of these two thrusts. These normal faults have displacements of a few centimeters and in thin

section are filled with calcite. The entire exposure is 46 m thick. Samman (1975) determined a 1:1 sandstone and siltstone to shale ratio for this outcrop. Based on field observations, the shale content of the Rome Formation increases along strike from 2-3:1 in the northeast to 1:1 in the southwest.

APPENDIX B

MILEAGE LOG FOR APPENDIX A

<u>Interval</u>	<u>Cumulative (miles)</u>	
0.0	0.0	University of Tennessee Geology and Geography Building
0.1	0.1	Turn left on Cumberland Avenue (US 11-70).
0.3	0.4	Turn right onto 17th Street.
0.5	0.9	Turn right onto Dale Avenue. Take I-275N to I-75N.
3.0	3.9	Exposure of the Saltville thrust on the left at Sharps Gap. Cambrian Rome Formation is thrust over Ordovician Knox Group.
3.9	7.8	Middle Ordovician Moccasin Formation exposed on the right.
3.5	11.3	Upper Cambrian Maynardville Formation exposed on the right.
1.0	12.3	Cambrian Conasauga Group exposed on the right.
0.4	12.7	Contact between Conasauga Group and the underlying Rome Formation on the right.
0.1	12.8	Exposure of the Copper Creek thrust on the right at Bull Run Ridge. Cambrian Rome Formation is thrust over Middle Ordovician Moccasin Formation.
1.1	13.8	Mascot Dolomite of the upper Knox Group exposed on the right at Flint Ridge.
.35	14.15	Copper Ridge Dolomite of the Knox Group exposed on the left at Chestnut Ridge.

<u>Interval</u>	<u>Cumulative (miles)</u>	
1.35	15.5	STOP 1. Contact between the Conasauga Group and the underlying upper Rome Formation at Pine Ridge.
0.2	15.7	STOP 2. Small displacement thrusts and folds in the upper Rome Formation at Pine Ridge.
0.1	15.8	Small displacement thrust in the lower Rome Formation on the left.
0.35	16.16	Cross unexposed Hunter Valley thrust which places Cambrian lower Rome Formation over Cambrian lower Conasauga Group.
0.8	16.6	Upper Rome Formation with small folds and antithetic thrust on the right.
0.40	17.0	Cross unexposed Wallen Valley thrust which places Cambrian Rome Formation over Silurian Rockwood Formation.
0.30	17.30	STOP 3. Exposure of a trailing imbricate of the Wallen Valley thrust which places Rockwood on Rockwood at Lone Mountain.
0.20	17.50	Contact between the lower Silurian Rockwood Formation and the upper Ordovician Sequatchie Formation.
0.20	17.70	Upper Ordovician Reedsville Shale on the left.
0.9	18.6	Take Clinton/Norris exit 22 to Highway 61. Turn left onto westbound 61.
0.6	19.2	Turn left at Git-n-Go Market onto Hillvale Road.
0.5	19.7	Middle Ordovician Chickamauga Group limestones in old quarry.
0.5	20.2	Cross Mountain Road.
0.2	20.4	Reedsville Shale on the right.
0.1	20.5	Sequatchie Formation on the right.

<u>Interval</u>	<u>Cumulative (miles)</u>	
0.2	20.7	Rockwood Formation on the right. The thrust which duplicates the Rockwood Formation is exposed on I-75N to the left.
0.2	20.9	Junction with Pumpkin Hollow and Foust Hollow Roads. Turn left onto Foust Hollow Road.
0.1	21.0	Turn right onto Brooks Gap Road.
0.15	21.15	Cross the unexposed Wallen Valley thrust which places Cambrian Rome on Silurian Rockwood.
0.45	21.60	STOP 4. Antithetic thrust fault and folds in the upper Rome Formation within the hangingwall of the Wallen Valley thrust at Bluebird Ridge.
0.05	21.65	Continue southbound on Brooks Gap Road and cross I-75.
0.35	22.0	Cross the unexposed Hunter Valley thrust which places Cambrian Rome on Cambrian Conasauga.
0.55	22.2	Junction of Brooks Gap Road and Brushy Valley Road. Turn left.
0.10	22.3	Turn to the right onto Moore Gap Road.
0.30	22.6	STOP 5. Small displacement thrust in the lower Rome Formation at Pine Ridge.
0.20	22.8	Upper Rome Formation and Conasauga Group contact.
0.60	23.4	Junction of Moore Gap Road and Wolf Valley Road. Turn right.
0.8	24.2	Exposure of upper Cambrian Maynardville Formation and the overlying Copper Ridge Dolomite (lower Knox Group) on the left.
1.5	25.7	Conasauga Shale on the right.

<u>Interval</u>	<u>Cumulative (miles)</u>	
3.0	28.7	Turn to the right onto Yarnell Road.
0.4	29.1	Conasauga Group and upper Rome Formation contact on left.
0.30	29.4	Cross the unexposed axial plane of a macroscopic anticline in the upper Rome Formation.
0.60	30.0	Continue right on Yarnell Road.
0.30	30.3	Cross the unexposed Whiteoak Mountain fault which places Cambrian lower Rome Formation over Ordovician lower Chickamauga Group.
0.35	30.65	Junction with Clinton Highway (US-25W). Turn right onto northbound 25W.
0.15	30.8	Cross unexposed fault which places Ordovician lower Chickamauga over Silurian Rockwood Formation.
0.20	31.0	STOP 6. Chevron folds in the Rockwood formation at Lost Ridge.
0.0	31.0	Turn around and travel southbound on Clinton Highway.
0.85	31.85	Cross the unexposed Whiteoak Mountain fault.
0.75	32.6	STOP 7. Small displacement thrusts in the upper Rome Formation at Pine Ridge. Park near the orange 25W Ready Concrete Company sign. Cross Clinton Highway and climb the ridge to the outcrop along the railroad tracks.
0.0	32.6	Continue southbound on Clinton Highway.
2.8	35.4	Rome Formation in the hangingwall of the unexposed Copper Creek fault on the left.
9.5	44.9	Junction of Clinton Highway and I-75. Take I-275S straight ahead.

<u>Interval</u>	<u>Cumulative (miles)</u>	
4.1	49.0	Take the 17th Street exit (387). Turn right from Ailor Avenue onto 17th Street.
1.4	50.4	Turn left from 17th Street onto Cumberland Avenue.
0.30	50.7	Turn right to Circle Drive and The University of Tennessee Geology and Geography Building.

APPENDIX C

MINOR STRUCTURES AT APPENDIX A STOPS

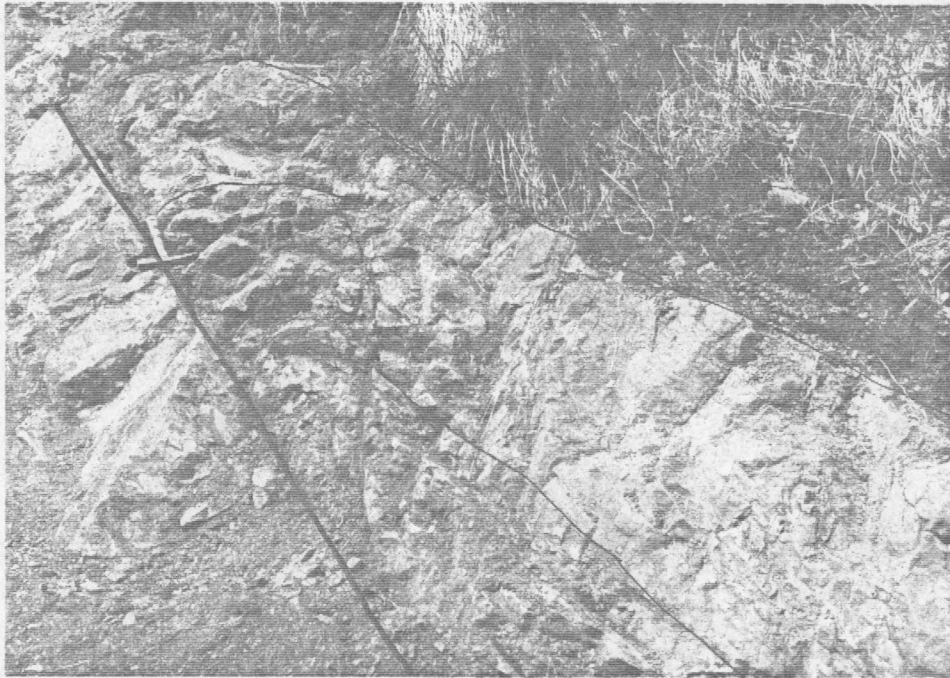


Figure C-1. Bedding wedge in the upper Rome dolomitic oolite within the hangingwall of the Hunter Valley thrust at Stop 1. Note hammer in the upper left-hand corner for scale.



Figure C-2. Small fold in the Rockwood sandstones and shales within the footwall of a trailing imbricate of the Wallen Valley thrust at Stop 3. Note hammer in the lower center for scale.



Figure C-3. Small footwall horse in the Rockwood siltstones and shales within the footwall of a trailing imbricate of the Wallen Valley thrust at Stop 3. Note hammer in the lower center for scale.



Figure C-4. Bedding wedges and normal faults in the upper Rome sandstones within the hangingwall of the antithetic (northwest dipping) thrust at Stop 4. Note 15 cm pen in the upper center for scale.

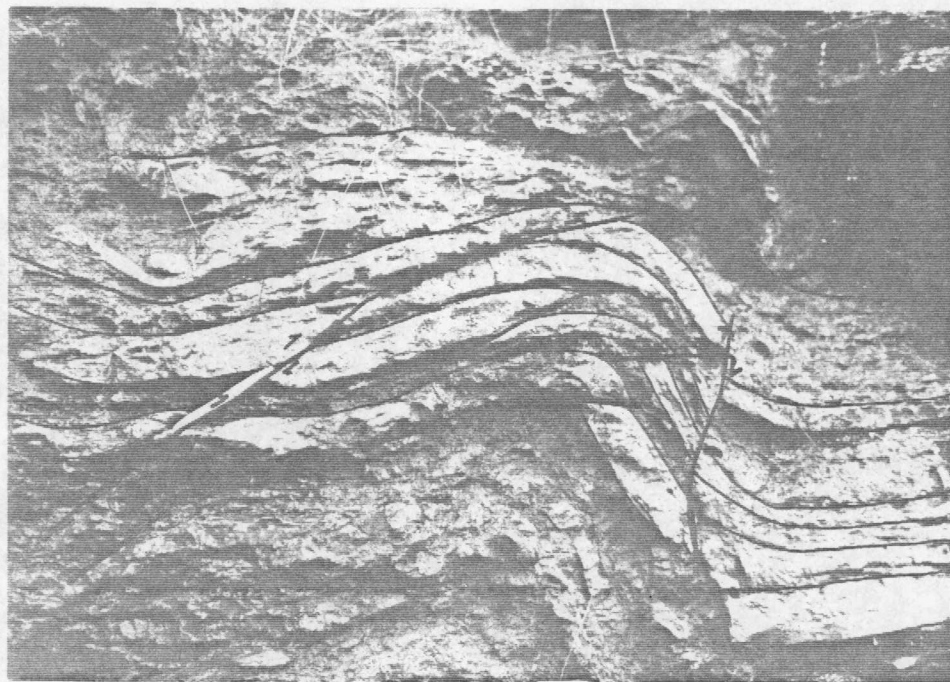


Figure C-5. Small scale limb thrusts in the upper Rome sandstones and shales within the footwall of the antithetic thrust at Stop 4. Note 15 cm pen along the left limb thrust for scale.

APPENDIX D

STRUCTURAL ORIENTATION DATA FOR OUTCROPS IN THE STUDY AREA

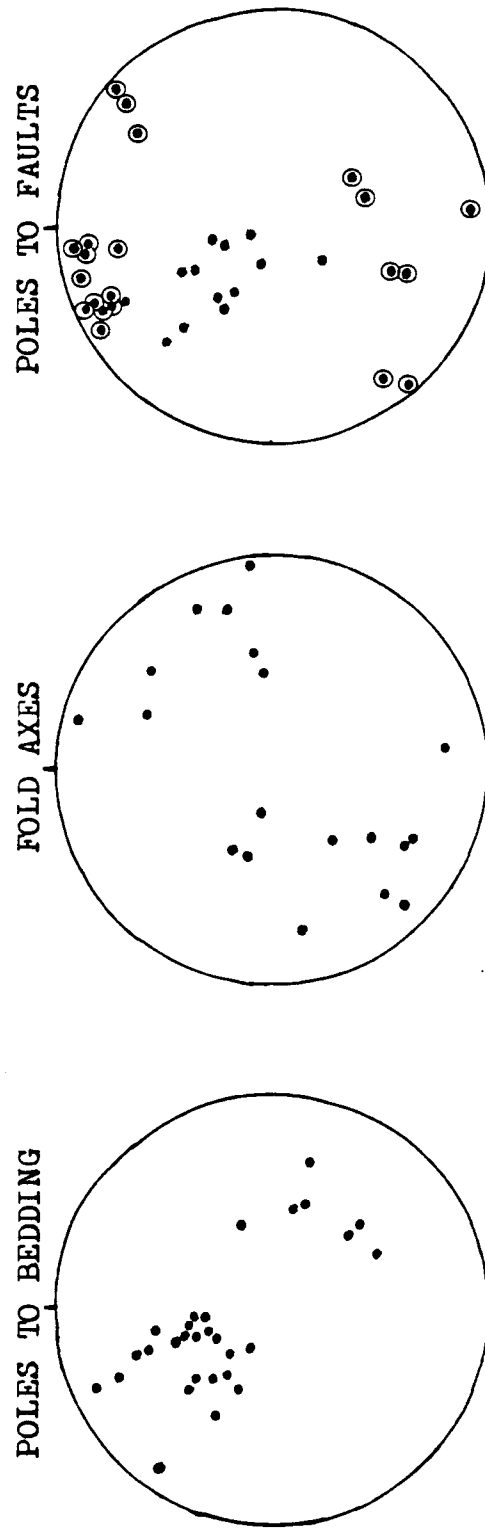


Figure D-1. Lower Conasauga Group and Rome Formation along I-75N at Pine Ridge (Stops 1 and 2 of Appendix A).

All data are plotted on Schmidt nets.
Circled poles to faults are extensional faults.

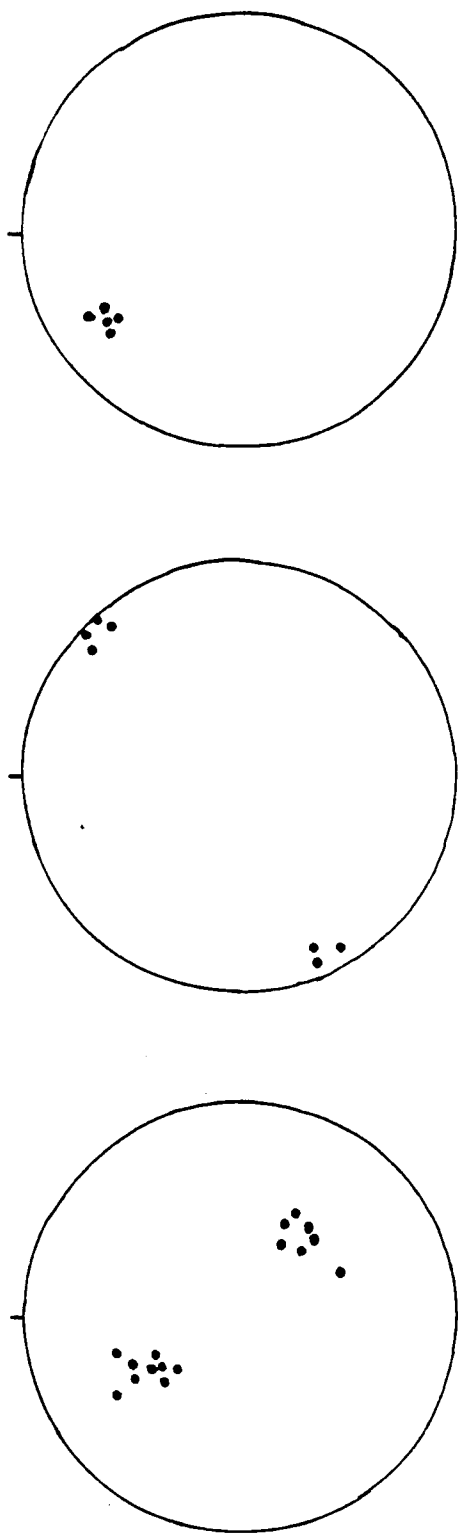


Figure D-2. Rockwood and Sequatchie Formations along I-75N at Lone Mountain (Stop 3 of Appendix A).

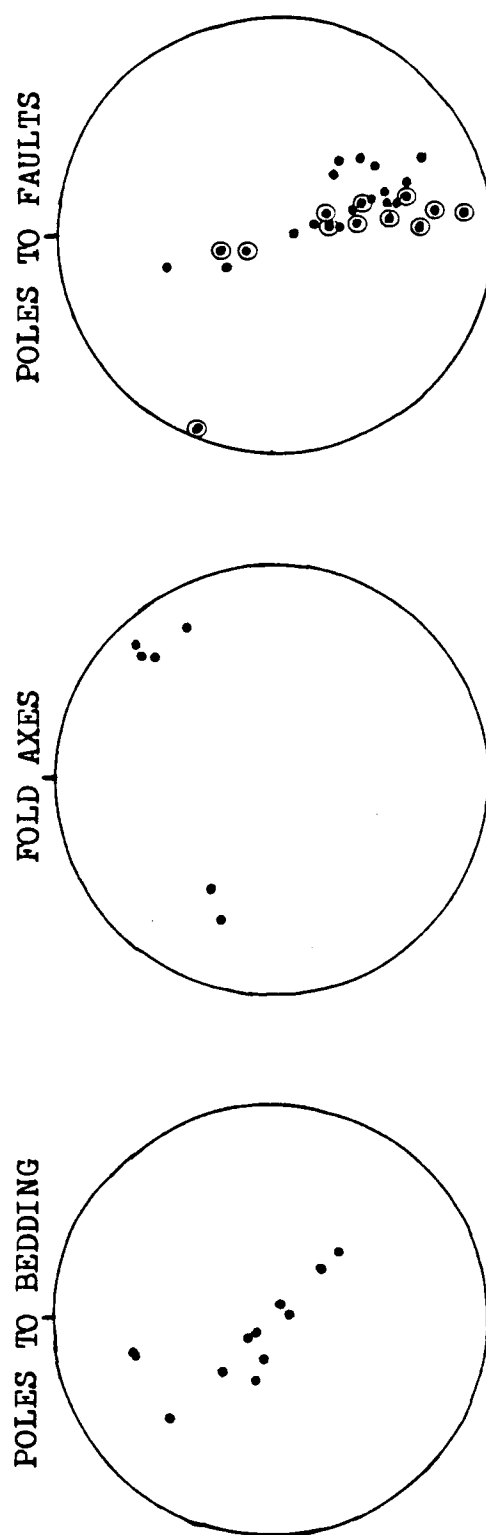


Figure D-3. Upper Rome Formation along Brooks Gap Road at Bluebird Ridge (Stop 4 of Appendix A).

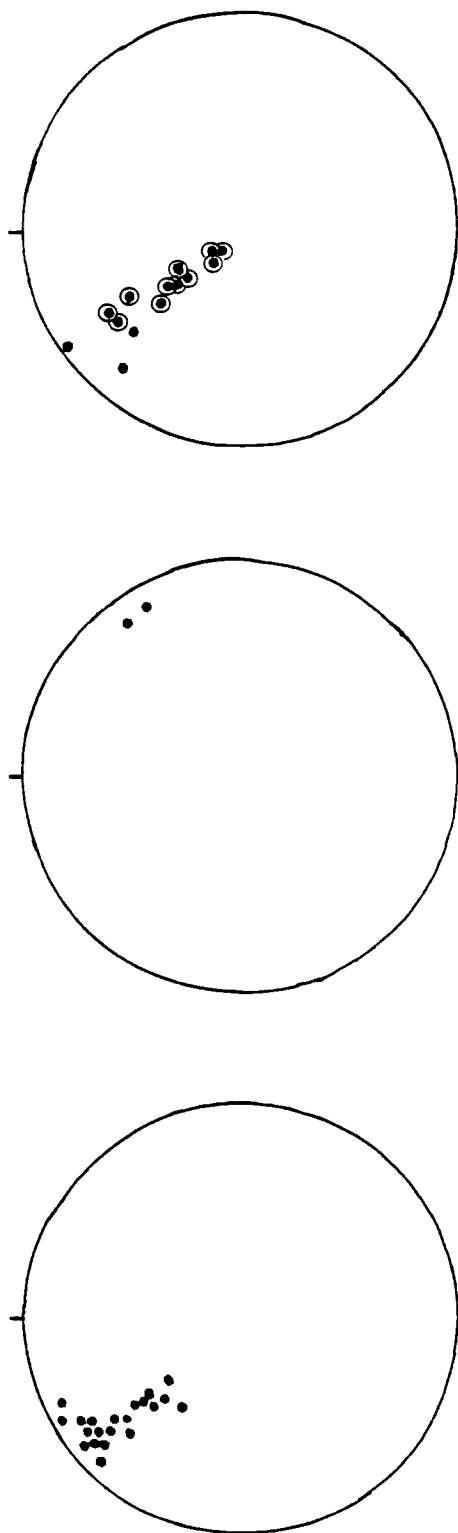


Figure D-4. Lower Rome Formation along Moore Gap Road at Pine Ridge (Stop 5 of Appendix A).

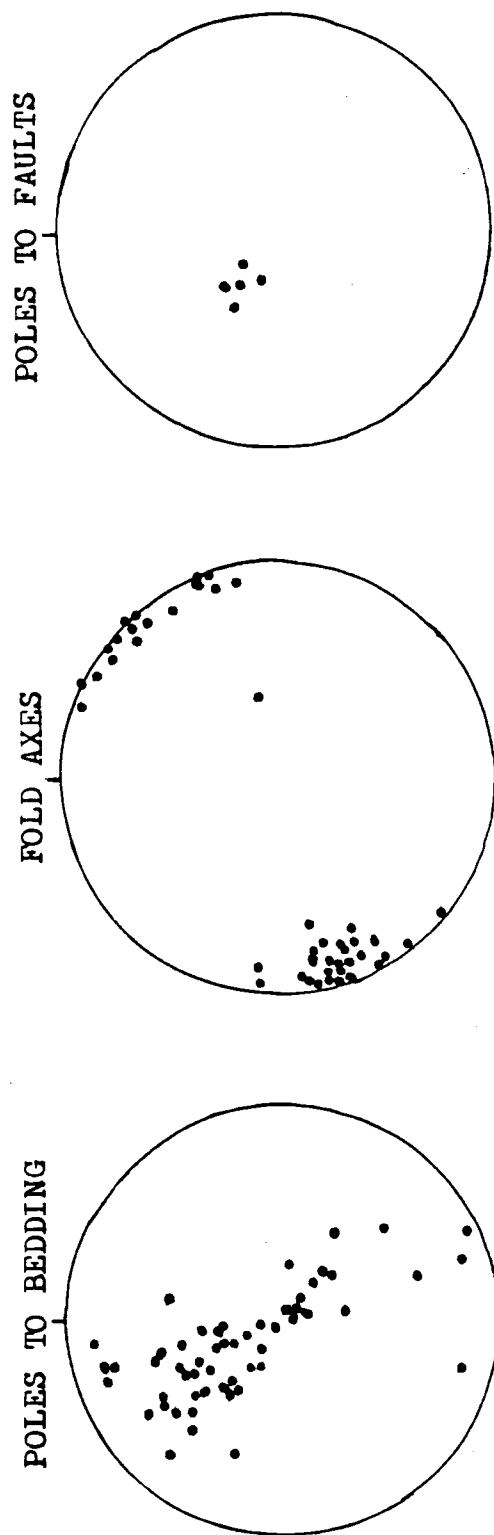


Figure D-5. Conasauga Group along Dismal Valley Road at Pine Ridge.

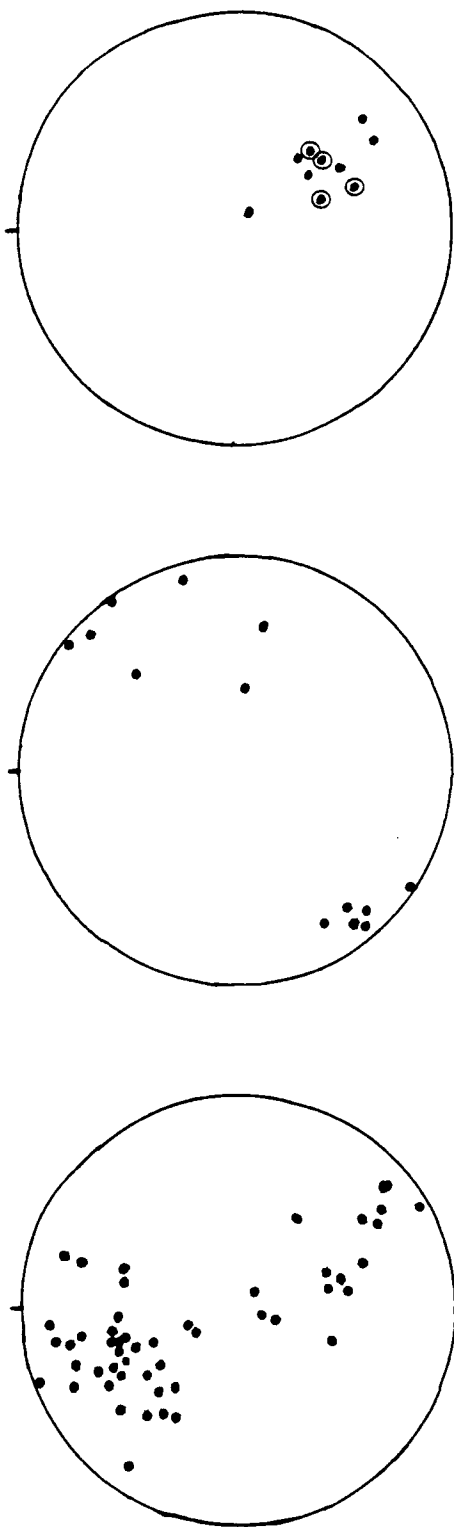


Figure D-6. Rome Formation along Dismal Valley Road at Pine Ridge.

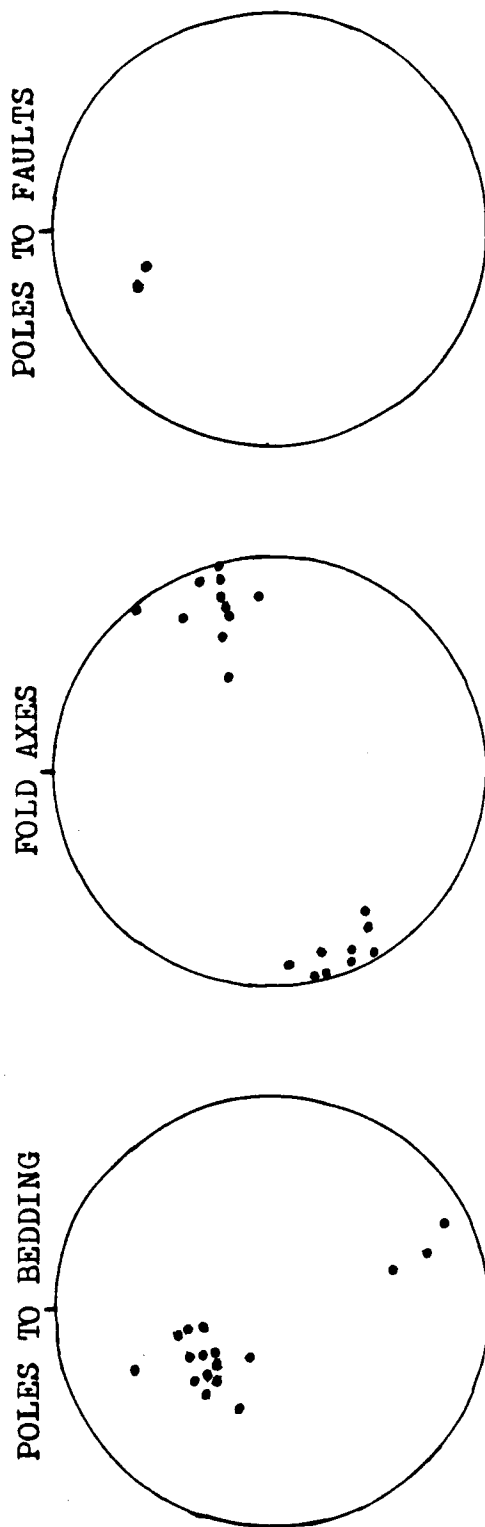


Figure D-7. Rockwood Formation along Clinton Highway US-25W at Lost Ridge (Stop 6 of Appendix A).



Figure D-8. Upper Rome Formation along a railroad cut through Pine Ridge (Stop 7 of Appendix A).

VITA

Jerry W. Beets was born in Knoxville, Tennessee, on November 29, 1959. He attended public schools in Knoxville and was graduated from Bearden High School in June 1978. The following September he entered The University of Tennessee, Knoxville under a four year scholarship. He was graduated with honors June 1982 with a Bachelor of Arts degree in Geology. In the fall of 1982 he entered graduate school at The University of Tennessee, Knoxville under a two year teaching assistantship. He was graduated with honors June 1985 with the degree Master of Science in Geology.

The author is a member of the Geological Society of America and the American Association of Petroleum Geologists. He coauthored and presented a paper at the Southeastern Section Meeting of the Geological Society of America in Knoxville on March 22, 1985 (Geological Society of America Abstracts with Programs, v. 17, no. 2, January 1985). Mr. Beets resides in Houston, Texas, and he is employed as an Exploration Geologist with the Mid-Continent Division of Shell Western Exploration and Production Inc.