OKLAHOMA GEOLOGICAL SURVEY

CARL C. BRANSON, *Director*

BULLETIN 101

STRUCTURE AND STRATIGRAPHY OF THE RICH MOUNTAIN AREA, OKLAHOMA AND ARKANSAS

by

DONALD R. SEELY

The University of Oklahoma
Norman
1963

CONTENTS

ILLUSTRATIONS

PLATES

FIGURES

Page

TABLE

STRUCTURE AND STRATIGRAPHY OF THE RICH MOUNTAIN AREA, OKLAHOMA AND ARKANSAS

Donald R. Seely

ABSTRACT

The Rich Mountain area comprises 200 square miles of rugged scenic country on either side of the Oklahoma-Arkansas border in the northern part of the Ouachita Mountain province. It lies between the Windingstair and Ti Valley faults and contains outcropping rocks that range in age from Mississippian to Early Pennsylvanian. Exposed strata are 13,500 feet thick. They include the Moyers and Chickasaw Creek Formations of the Stanley Group, the Jackfork Group and one of its formations (Game Refuge), the Johns Valley Shale, and the Atoka Formation. A tuff bed that occurs in the Chickasaw Creek Formation of the area is the youngest yet reported in the Ouachita Mountains.

Many of the exposed beds have current ripple marks and/or associated small-scale cross-bedding. Others are characterized by the presence of bottom casts. Lamination is common in sandstones of the Jackfork Group and the Johns Valley and Atoka Formations. Although graded bedding of some type is present locally, it is not megascopically obvious. Other features shown by various sandstones are ellipsoidal clay galls and their molds, load casts, contorted bedding (uneven surfaces due to squeezing while in a semiplastic state), trails of benthonic organisms, topsurface troughs and ridges, and molds of fragments of invertebrates.

Sandstones from the upper part of the Stanley, from the Jackfork, and from the lower part of the Atoka are classified petrographically as orthoquartzites. Excluding micas, the heavy-mineral suite is simple and consists of zircon, tourmaline, and garnet. Metamorphic-rock fragments and angular feldspar in small amounts are other significant constituents.

The source rocks were of sedimentary and low-rank metamorphic types. Volcanic debris was contributed locally to the basin.

Sandstones of the Jackfork-Atoka sequence have most of the characteristics of turbidites. Graded bedding is not a conspicuous feature, however, and many beds are relatively well sorted. Texture, lamination, good sorting, and absence of graded bedding indicate that many sandstones fit better into Lombard's description of laminites than into ten Haaf's description of turbidites, and that they probably were laid down by slow-moving bottom currents.

Paleocurrent studies and regional time correlations near the study area suggest a prevailingly westward or west-southwestward flow parallel to tectonic strike.

The Tenmile Creek Formation, Moyers Formation, and Jackfork Group sediments probably were deposited below wave base, and therefore not on a river flood plain, on top-set beds of a delta, on a tidal flat, or in shallow offshore water.

Major anticlinal and synclinal folds extend eastward through the area, dominating the gross structural pattern. They are displaced by eastward-trending faults, two of which, here named the Honess and Briery faults, were previously unmapped. The fault belt that lies north of the Windingstair fault in Oklahoma is thus extended into Arkansas, much farther eastward than was previously recognized.

The Honess, Briery, and Windingstair faults probably are thrust faults through most or all of the mapped area, judging from scanty field evidence and from the geometry of structures associated with the faults.

Dips of the major faults vary to an unknown extent along their traces. Near the west edge of the area the Briery fault has a south dip of about 55 degrees, and the Windingstair fault dips southward at an angle probably less than 40 degrees. Eastward the dips of both faults probably increase.

Thrust masses in the report area probably moved northward perpendicularly to the traces of major faults. Both right-lateral and left-lateral strike-slip components of movement occurred in the overriding masses.

On the basis of present knowledge it is believed that Ouachita structures originated with simple vertical movements of the basement or with vertical movements accompanying basement thrusting. The resulting gravitational gliding or spreading produced the major folds and thrust faults.

INTRODUCTION

LOCATION AND GENERAL DESCRIPTION OF STUDY AREA

The area of this study includes within its boundaries parts of Le Flore County, Oklahoma, and Polk and Scott Counties, Arkansas. The terrain is relatively rugged and possesses much scenic beauty. Mountains near Mena, Arkansas (pl. I), are some of the southernmost prominences north of the Gulf of Mexico. The top of Rich Mountain rises, in fact, more than 1,000 feet above the surrounding valleys, making it one of the higher mountains in Oklahoma. Because of this, the eastern summits of Rich Mountain, which may be reached via Skyline Drive (pl. I), provide excellent panoramas.

Another attraction on the crest of Rich Mountain is Wilhelmina State Park, a new and rapidly growing recreational area. The park surrounds partly rebuilt Wilhelmina Inn, an old hotel erected in the 1890's by the early owners of Kansas City Southern Railroad to stimulate business for the then newly laid line.

Geologically, the area is about midway in the east-west part of the Ouachita structural belt that is exposed from Little Rock, Arkansas, to Atoka, Oklahoma (fig. 2). Published geologic maps of the Ouachita Mountains show the area to be in a transition zone from tight folds and thrust faults in Oklahoma to open, unfaulted folds in Arkansas. However, results of the present research and of recent work by Reinemund and Danilchik (1957) in the Waldron quadrangle suggest that the transition is not nearly so abrupt as present maps imply. At least two major faults mapped as part of this investigation probably continue many miles eastward into Arkansas, and attitudes of Jackfork sandstone beds noted in reconnaissance to the east of the report area indicate steep dips which are not characteristic of open folds.

Windingstair fault marks the south boundary of the field-mapped area. Stanley beds to the south of the Windingstair fault are included on plate I only in order to allow the map to have geographic rather than geologic boundaries. To the north of the area is the Ti Valley fault. Mapped rocks, therefore, are part of a belt of rocks between the Windingstair and Ti Valley faults. This belt extends from near Atoka, Oklahoma, to the eastern edge of the area.

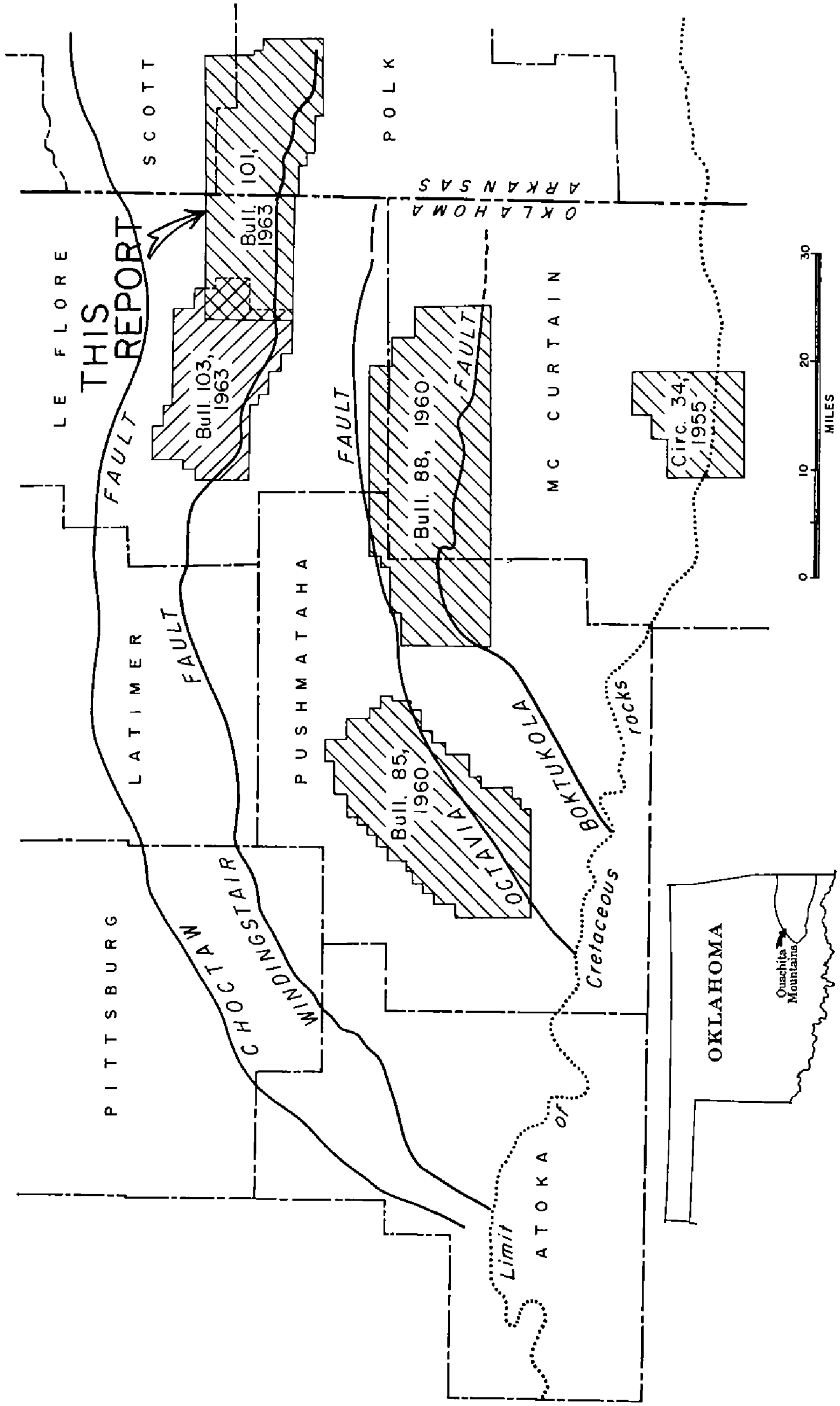


Figure 1. Index map showing area of this report and other related reports on the geology of the Ouachita Mountains.

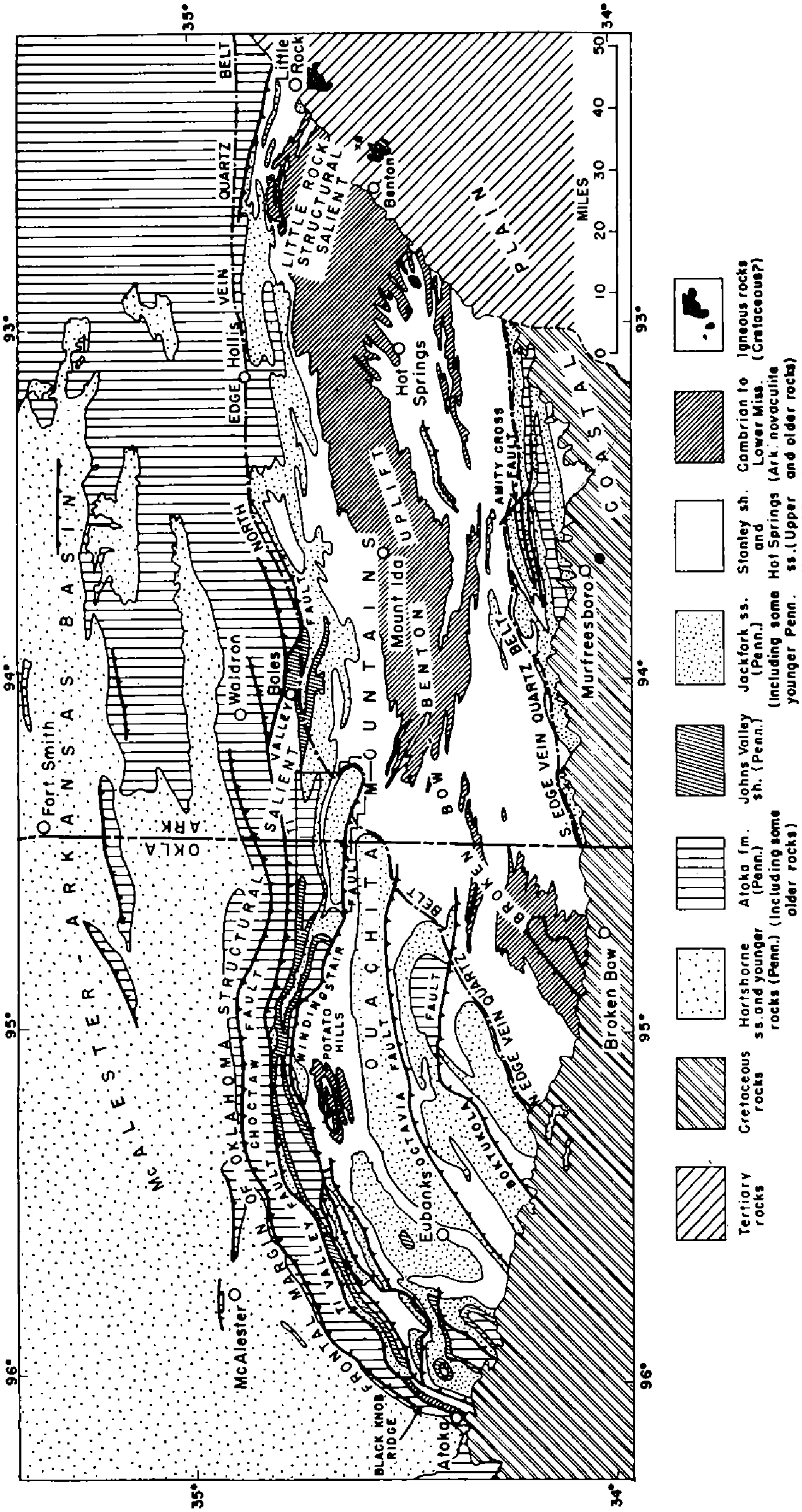


Figure 2. Geologic map of the Ouachita Mountains. (from Miser, 1959, fig. 3)

PAST AND PRESENT INVESTIGATIONS

The only published geologic maps which include the Oklahoma portion of the area are those of Honess' study of southern Le Flore and northwestern McCurtain Counties (Honess, 1924) and of Miser's interpretation of Ouachita geology shown on the 1954 edition of the *Geologic map of Oklahoma.* Honess' map has a small scale and includes too large an area for detailed mapping of Ouachita rocks, but it is the most accurate geologic map of part of the area heretofore published. Miser's interpretation was based upon the unpublished maps of J. A. Taff, who mapped large areas of the Ouachita Mountains near the turn of the century. Few of Taff's maps were published.

The *Geologic map of Arkansas,* published in 1929 by the Arkansas Geological Survey, does not indicate specific sources for the geology shown in the Arkansas portion of the study area. However, on it is stated that the geology of the Ouachita Mountains was obtained from maps, mostly unpublished, by L. S. Griswold, A. H. Purdue, H. D. Miser, and R. D. Mesler.

Reinemund and Danilchik (1957) included the northeast corner of the area in a small-scale, reconnaissance geologic map in their report. They also suggested the presence of a fault (herein named the Honess fault) in their small-scale, generalized structure map.

The geology of adjacent areas has not been published on detailed maps, but studies are being made to the north and west. P. H. Stark of the University of Wisconsin is investigating the microfauna of the Atoka Formation of Spring Mountain syncline; O. D. Hart, also a student at the University of Wisconsin, has mapped the geology directly to the west of the writer's mapping, and his work is being published by the Oklahoma Geological Survey (Bulletin 103).

Although the primary objective of the present research was to study the structural geology of the Rich Mountain area, stratigraphical problems were also investigated to identify mapping units and to learn something of their lithology and possible origin. Investigation of the many problems discussed in this report is not exhaustive and much research remains to be done.

ACKNOWLEDGMENTS

The writer is indebted to many people who have helped him complete this study. The late C. W. Tomlinson suggested the problem and W. D. Pitt, then assistant professor of geology, The University of Oklahoma, supervised the work. Ph. Kuenen, Rijks University, The Netherlands, supplied helpful literature and correspondence, as did Tjeerd van Andel of Scripps Institution of Oceanography. The Oklahoma Geological Survey, Carl C. Branson, director, provided financial assistance for field work. Charles J. Mankin, assistant professor of geology, The University of Oklahoma, visited the study area and provided stimulating discussions on various aspects of petrography and sedimentation. L. M. Cline, University of Wisconsin, and J. K. Arbenz, Shell Oil Co., Denver, Colorado, read the manuscript and gave helpful criticism.

STRATIGRAPHY

Stratigraphic Nomenclature

The rock-unit names in this report (fig. 3) are those used by Cline (1960). The units were distinguished on the basis of descriptions published by Harlton (1938), by Cline and Moretti (1956), by Cline and Shelburne (1959), and by Cline (1960). Although most of them were originally defined by Harlton (1938), definitions of several have been modified recently by Cline (appendix A).

The practice of subdividing the Stanley and Jackfork Groups into formations by using siliceous shale marker beds is considered unwise by T. A. Hendricks (personal communication). Hendricks believes that several formations are lithologically so similar that the units would be indistinguishable if the siliceous shales were not present. For this reason he suggests that the Stanley and Jackfork should be termed "formations" which contain siliceous shale "members."

Evidence to support Hendricks' viewpoint is the mapping of the upper part of the Wildhorse Mountain Formation, the Prairie Mountain Formation, and the Markham Mill Formation as a single undifferentiated unit by both Cline (1960) and Shelburne (1960). In the present report the Wildhorse Mountain, Prairie Mountain, Markham Mill, and Wesley Formations could not be differentiated owing to absence of the key beds. The approximate contact between the Wildhorse Mountain and the Prairie Mountain was mapped to show the structural geology more clearly, but this formation boundary is not based upon the siliceous shale present in the type localities.

Nomenclature is not the primary concern of this report. For clarity of usage and because of reliance on Cline's work for identification, the latter's definition of units is followed. The units and their present assigned ages are shown in figure 3. Ages are those assigned by Cline.

System	Series	Group	Formation	Member	Bed
PENNSYLVANIAN	MORROWAN - ATOKAN		Atoka		Fossiliferous sandstone and spicular siliceous shale near base
			Johns Valley		
MISSISSIPPIAN	MERAMECIAN - CHESTERIAN	Jackfork	Game Refuge		
			Wesley		
			Markham Mill	Siliceous Shale Member at base	
			Prairie Mountain	Siliceous Shale Member at base	
			Wildhorse Mountain	Prairie Hollow Shale Member below middle	
		Stanley	Chickasaw Creek		
			Moyers	Siliceous Shale Member at base	
			Tenmile Creek	Upper Member	Battiest Siliceous Shale Bed at base
				Lower Member	Tuskahoma Siliceous Shale Bed
					Lower Siliceous Shale Bed

Figure 3. Stratigraphic units and their ages.
(modified from Cline, 1960)

Thicknesses of units given herein are not rounded to significant figures because the accuracy of measurements cannot be determined. It is probable that the stated thicknesses are within ten percent of the true thicknesses.

Stanley Group

Major valleys of the Ouachitas have been eroded from shales which constitute the principal rock type of the Stanley Group. The writer has not mapped formations of the Stanley Group south of the Windingstair fault because to do so would involve detailed mapping of the Kiamichi Valley. More than 300 feet of undifferentiated Stanley south of the Windingstair fault is included in the descriptions of Ward Lake spillway and East Ward Lake measured sections (appendix H). These same sections include descriptions of the Chickasaw Creek Formation, which borders the Windingstair fault on the north at these localities.

Isolated exposures of the upper part of the Tenmile Creek, Moyers, and Chickasaw Creek Formations are present on the north slopes of Rich and Black Fork Mountains. Except for the lowermost beds of these units, they probably are of simple structure so that approximate positioning in the stratigraphic interval is possible. Because of the extensive cover of colluvium, however, units of the Stanley are poorly exposed and no attempt was made to map them.

MOYERS FORMATION

Rocks which are probably the basal siliceous shale of the Moyers Formation are present in a few outcrops at the foot of the north slope of Rich Mountain, where they appear on the dip slopes of a sandstone sequence which they overlie. Erosion of the overlying shale has exposed the siliceous shale. The unit consists of gray siliceous shale in beds a few inches and less thick. The beds make up a zone two to three feet thick, and weather into polygonal plates and blocks typical of siliceous shales. The shales overlying it are gray, gray green, and olive green in weathered exposures and compose a 30-foot section which is overlain by sandstone containing light-gray brittle shale and black shale.

The Moyers Formation on the north slope of Rich Mountain was determined to be about 1,000 feet thick. Sandstone beds observed in the upper part of the Moyers are lithologically similar to those of the lower part of the Jackfork.

CHICKASAW CREEK FORMATION

The siliceous shale of the Chickasaw Creek Formation is described both in the East Ward Lake and Ward Lake spillway measured sections. Its resistant beds are well exposed on the north slope of Black Fork Mountain (NE¼ SW¼ sec. 25, T. 1 N., R. 32 W.). Detailed descriptions of the Chickasaw Creek are included in appendix H and in the petrographic descriptions in this report. At the measured localities, the Chickasaw Creek is predominantly shale of varying shades of gray and of varying amounts of silica content. The highly siliceous zones are no more than a few feet thick and the large white specks characteristic of the Chickasaw Creek are not evident. Close inspection reveals specks to be present, but their diameters are generally less than 0.1 mm. Despite the small diameter of the specks, assignment of these beds to the Chickasaw Creek may be made upon the basis of topographic and stratigraphic position.

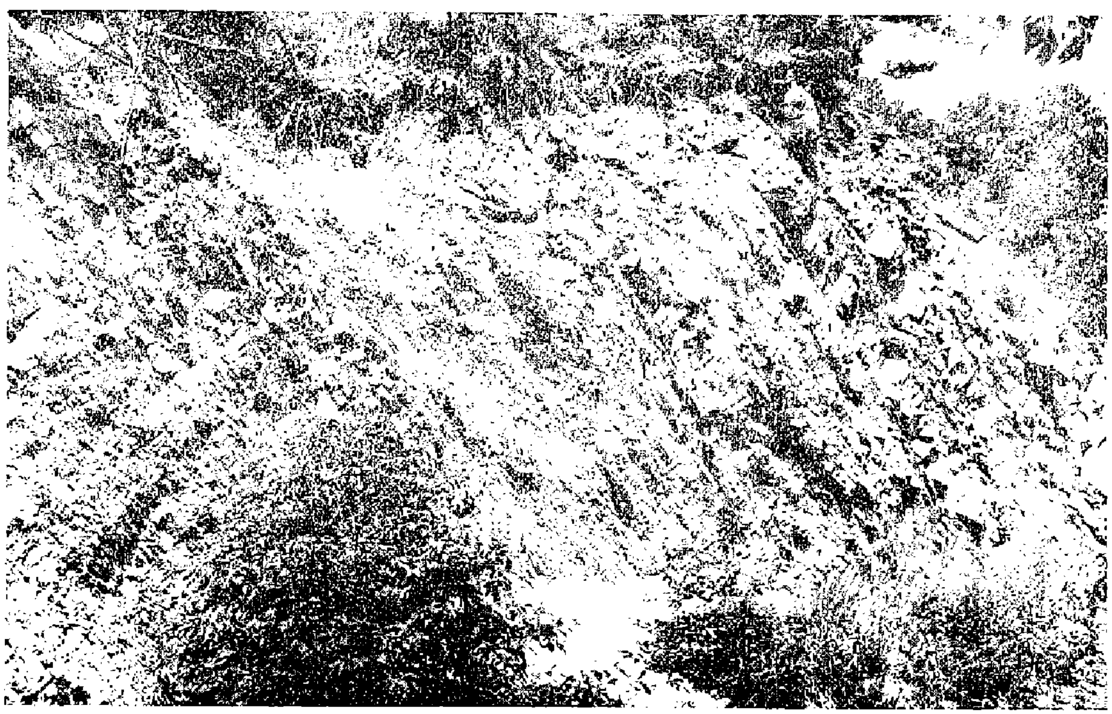


Figure 4. Ward Lake spillway exposure. Blocky weathering beds of a cherty zone near the top of the Chickasaw Creek are visible in left center. More readily eroded shales overlie the cherty zone and become interbedded upward with lowermost sandstones of the Wildhorse Mountain (of interval 5, Ward Lake spillway measured section).

Sandstone beds are a minor part of the Chickasaw Creek but increase in abundance toward the upper contact. Here they grade into the basal sandstone zone of the Jackfork (figs. 4, 5). At the East Ward Lake locality sandstone lenses up to six feet thick are present, but the normality of their stratigraphic position is made uncertain by the Windingstair fault.

An excellent exposure of beds of the upper part of the Moyers, Chickasaw Creek, and lower part of the Wildhorse Mountain Formations is in a stream valley on the north slope of Black Fork Mountain in NE¼ SW¼ sec. 25, T. 1 N., R. 32 W. Sandstones of the upper part of the Moyers appear to have a very fine-grained matrix in which larger quartz grains are embedded. White and black specks are visible and a small amount of mica is present. The sandstones are hard (for induration terminology see appendix H) and give rise to steep dip slopes extending several tens of feet downdip. A few sandstones are as much as 20 feet thick.

Above the upper sandstone zone of the Moyers Formation is a shale zone between 150 and 200 feet thick. The shales underlie

Figure 5. East Ward Lake measured section exposure. Diagonally across the picture from near the lower left corner to the transformer-bearing telephone pole in the upper right is a fault zone (probably that of the Windingstair fault). Directly to the right of this zone are white sandstones of the Stanley Group. To the left of the zone and underlying the topographic bench is the Chickasaw Creek Siliceous Shale. The Chickasaw Creek-Wildhorse Mountain gradational contact is exposed in the escarpment in the left background. The escarpment is capped by the lowermost resistant sandstone of the Wildhorse Mountain Formation.

a topographic bench and are poorly exposed. Two slightly elevated ridges are in the bench: one is due to an 8- to 10-foot siliceous shale zone near the base of the unit; the other is due to a 10-foot tuff bed about 100 feet above the siliceous shale. Bands and lenses of white specks are present in some of the 1- to 3-inch beds making up the siliceous shale zone. They are small, however, and generally are not visible on weathered surfaces. The beds are dark gray to black and most of them have a low percentage of silica, as is indicated by the absence of prominent conchoidal fracture, by the presence of earthy weathered surfaces, and by a low degree of hardness. The tuff bed is massive, medium dark gray, and hard in fresh samples. It contains many large light-colored grains, most of which have a talclike megascopic appearance but some of which are quartz grains. Dark grains that may be either siliceous shale fragments or carbonaceous matter also are present. Microscopic features are discussed in the microscopic petrography section of this report.

A few sandstones, each less than two feet thick, are in the thick shale sequence. These display good small-scale cross-bedding and groove casts in exposures near the top of the sequence. The sandstones are medium to dark gray and very fine grained and contain subellipsoidal clay galls. Carbonized plant fragments are concentrated near their upper surfaces. They weather to light gray or light greenish gray.

Overlying the shale sequence are two sandstone zones each approximately 50 feet thick, the upper zone being poorly exposed. Next above is an excellent exposure of 12 feet of highly siliceous shales and cherts, some of which possess the typical large white specks that characterize the Chickasaw Creek elsewhere. Individual cherty beds are less than eight inches thick, break into polygonal plates, have limonitic coatings on weathered surfaces and in fractures, and are interbedded with less siliceous subplaty shale. A few laminae of silt- or sand-size material are visible.

Directly overlying this upper siliceous shale is another sandstone zone. The sandstone beds are medium gray, very fine grained, and exceed two feet in thickness. In the stream bed up to and above this stratigraphic position are black asphaltic sandstone boulders and sandstone fragments containing a fauna of molds of invertebrates.

Exposures of the Chickasaw Creek at the extreme west end of Rich Mountain and on the eastern extremity of Winding Stair Mountain are of poor quality and difficult to find.

Thickness measurements by various workers at Chickasaw Creek localities in the Ouachitas are of questionable comparative significance because they are not based upon a consistent determination of top and bottom contacts. The thicknesses determined at the measured-section localities in the current study also are limited in value because of faulting. At the Black Fork Mountain exposure a question of where to place the top contact arises owing to the thick sandstone section below the upper cherty zone.

Jackfork Group

RICH MOUNTAIN MEASURED SECTION

Most of the following descriptions of stratigraphic units are based upon exposures of the Rich Mountain measured section. Data on location of these exposures and their description are in appendix H. The outcrops are not so good as those in the roadcuts south of Big Cedar on State Highway 103 that were used by Cline and Moretti (1956) in describing their measured section. Most of the outcrops are weathered, and there is much covered section. The effect of weathering on color is particularly difficult to assess. Examples of the difficulty are the white or nearly white sandstones on Rich Mountain. From surface exposures it is not possible to tell whether the color is primary or is due to leaching; however, O. B. Shelburne (personal communication) found that several Jackfork sandstones appear white in well samples.

Beds of the upper part of the Jackfork through the lower part of the Atoka are well exposed in the roadcuts on State Highway 103 where it traverses the western edge of the area.

WILDHORSE MOUNTAIN FORMATION

In the type localities this formation overlies the siliceous shale of the Chickasaw Creek and its top is marked by a siliceous shale (basal siliceous shale member of the Prairie Mountain Formation). Near the middle of the formation is a zone of variegated green and maroon shale (Prairie Hollow Shale Member) that has been used by Hendricks and others (1947), Shelburne (1960), and Cline (1960) as a key bed.

The siliceous shale which separates the Prairie Mountain Formation from the Wildhorse Mountain Formation in the type areas was not found by Cline in the Lynn Mountain syncline, nor by Shelburne in the Boktukola syncline (Cline, 1960, p. 53), nor by me in the report area. However, its apparent absence could be due to poor exposures. For mapping purposes the boundary between the Wildhorse Mountain and Prairie Mountain Formations was placed arbitrarily at the top of a sandstone zone in keeping with the practice of Cline and Moretti (1956, p. 13). The stratigraphic position of this zone is shown on plate II.

The Prairie Hollow Shale Member has not been positively identified in this study. If present in the stratigraphic position in which it has been found elsewhere in the western Ouachitas, it should crop out on the steep slopes of Rich and Black Fork Mountains. These slopes are covered with a flood of debris so that only the more resistant sandstones crop out. Under these circumstances it is to be expected that the Prairie Hollow would not be traceable in the Rich Mountain area. The only maroon shale exposed is in a small roadcut on the east side of Skyline Drive near NE cor. NW¼ sec. 6, T. 2 S., R. 30 W. This shale is only a few hundred feet above the base of the Wildhorse Mountain Formation.

The highest ridges of the Oklahoma Ouachitas are within the outcrop belt of the Wildhorse Mountain Formation because it has a higher percentage of sandstone and thicker and more indurated sandstone units than any other formation or zone of equivalent thickness in the Ouachitas. Cline (1960, p. 49) pointed out, however, that the Wildhorse Mountain Formation and the Jackfork Group contain sandstone and shale in nearly equal proportions. Because of the poor exposures on Rich Mountain, it is not possible to determine these proportions accurately there. The partial section of the Wildhorse Mountain described in the Rich Mountain measured section contains 511 feet of sandstone, 293 feet of sandstone and subordinate shale, and 230 feet of shale and subordinate sandstone. In addition, 2,344 feet of section is covered. Probably more than half of the covered sequence consists of weakly resistant shale. The total thickness of the Wildhorse Mountain is 3,378 feet in the Rich Mountain measured section. Not included in the section is an additional 270 feet of Wildhorse Mountain which would bring the complete thickness to about 3,600 feet.

Except for the erratics of the Johns Valley, the entire Stanley-Atoka sequence is composed alternately of sandstone and shale strata. The thickest sandstone beds are in the upper part of the

Moyers Formation, middle part of the Wildhorse Mountain Formation, the Game Refuge Formation, and the lower part of the Atoka Formation. Sandstone sequences unbroken by shale generally do not exceed 50 feet in thickness; the thickest bed in the Rich Mountain measured section is 31 feet (interval 21) and is near the crest of Rich Mountain and near the middle of the Wildhorse Mountain Formation. Sandstone zones, on the other hand, may be more than 100 feet thick; and several of these in the middle part of the Wildhorse Mountain have prominent topographic expression on the crests and south slopes of both Rich and Black Fork Mountains.

Upward from the base of the Wildhorse Mountain a change takes place from more poorly sorted sandstones containing varying percentages of clay and plant matter to better sorted, cleaner sandstones. The color also changes from dark shades of gray to white. Concentrations of carbonized plant matter are present, in the lower and upper parts of the formation in greater abundance than in the middle part.

Friable sandstones in the upper part of the Wildhorse Mountain Formation have a distinctive appearance where they crop out on the south slope of Rich Mountain (figs. 6, 7). Weathered surfaces

Figure 6. The dip slope of a friable sandstone of the Wildhorse Mountain Formation in interval 42, Rich Mountain measured section. Flaking due to weathering is present only at this stratigraphic level.

of these beds appear flaky and range from white to dark yellowish orange to reddish brown to grayish brown (National Research Council Rock Color Chart colors). This color gradation is from central parts of blocks outlined by joints and bedding planes to these bounding surfaces. The sandstones are particularly characteristic of the upper part of the Wildhorse Mountain Formation, although a few similar-appearing beds are present on the south slope of the ridge formed by the lower sandstone zone of the Atoka of the Rich Mountain measured section. It is possible, however, that their weathering features are a result of exposure on dip slopes rather than of distinctive primary lithology.

Most shales of the Wildhorse Mountain Formation are exposed as minor interbeds in outcrops of resistant sandstones. They range from olive gray to medium gray to grayish black. Most are fissile to flaky; a few are splintery. No thick shale section is free of thin beds or lenses of siltstone or fine sandstone. Many of these are deeply stained by iron oxide. Fracturing of the arenaceous layers is common, and iron oxide is concentrated along the fractures. The iron oxide is resistant to further weathering and stands out as a reticulate pattern of ridges where the nonimpregnated intervening material has been removed.

Figure 7. Wildhorse Mountain sandstones in the upper part of interval 42, Rich Mountain measured section. The bed shown in figure 6 is in left center. Southward dip is 56 degrees.

The occurrence of iron oxide is not dependent upon the presence of arenaceous layers. Most thick shale sections are thoroughly stained, probably as the result of oxidation of pyrite in the shale and its contained sandstones (see MRS 4-2 and MR 6-1, appendix B).

Trails left by benthonic organisms are conspicuous features on the tops of some sandstone beds of the Wildhorse Mountain Formation. Occurrences may be noted in the middle and upper parts of the unit (figs. 13, 14).

In the stream bed at and above the Chickasaw Creek locality on the north slope of Black Fork Mountain are sandstone boulders containing molds of invertebrates. These fossiliferous boulders indicate the presence of a fossiliferous sandstone in the lower Wildhorse Mountain beds exposed on the slopes above. Molds of invertebrates also occur near the base of the Wildhorse Mountain near Ward Lake. At about the same stratigraphic position is an abundance of *Calamites* stem fragments.

PRAIRIE MOUNTAIN, MARKHAM MILL, AND WESLEY FORMATIONS

In the type localities these units are identified by siliceous shales. The siliceous shales were not observed in the report area and the formations are undifferentiated.

This undifferentiated sequence, comprising the Prairie Mountain, Markham Mill, and Wesley Formations underlies a topographic low between the high ridge formed by sandstones in the middle part of the Wildhorse Mountain Formation and a lower ridge formed by Game Refuge sandstones. In outcrops of the Rich Mountain measured section it consists of 44 feet of sandstone, 113 feet of sandstone and subordinate shale, and 1,187 feet of shale and subordinate sandstone. A section of 413 feet is covered and the total thickness is 1,757 feet. The proportion of sandstone in the upper part of the sequence varies along strike. This variation is indicated locally by prominent sandstone ridges to the west of Skyline Drive.

It is evident that shale is the principal rock type of this sequence in the measured section. Except for sandstone near the base (pl. II), the sequence is composed entirely of shale containing minor sandstone interbeds. The basal sandstone could as well have been placed at the top of the Wildhorse Mountain Formation. If placed there, it would mark the boundary between a sandy sequence below and a shaly sequence above.

Sandstones in shale zones of the lower part of the Prairie Mountain-Markham Mill-Wesley sequence average one-half foot in thickness, are hard and planar to wavy bedded, with laminae generally not prominent, and have both bottom and ttop markings. The top markings consist of a central furrow with marginal, arcuate mounds, and small tubular grooves; the bottom markings are irregular and small and show no parallel alignment.

Higher in the sequence a few two-foot-thick sandstone beds are present in addition to the thinner beds. These thicker beds are topographically prominent. In several of them the top one-half inch has a high concentration of carbonized plant fragments. Carbonaceous material is also abundant in discoidal sandstone masses isolated in the shales (fig. 28).

Rare, but distinctive, features in the shale near the top of the sequence are ellipsoidal siliceous masses with a white weathered surface, and similarly shaped masses of limonite. Harlton (1938, p. 888) described "large rounded to subrounded chalcedonic masses" as "the most diagnostic feature of the Wesley siliceous shale." He

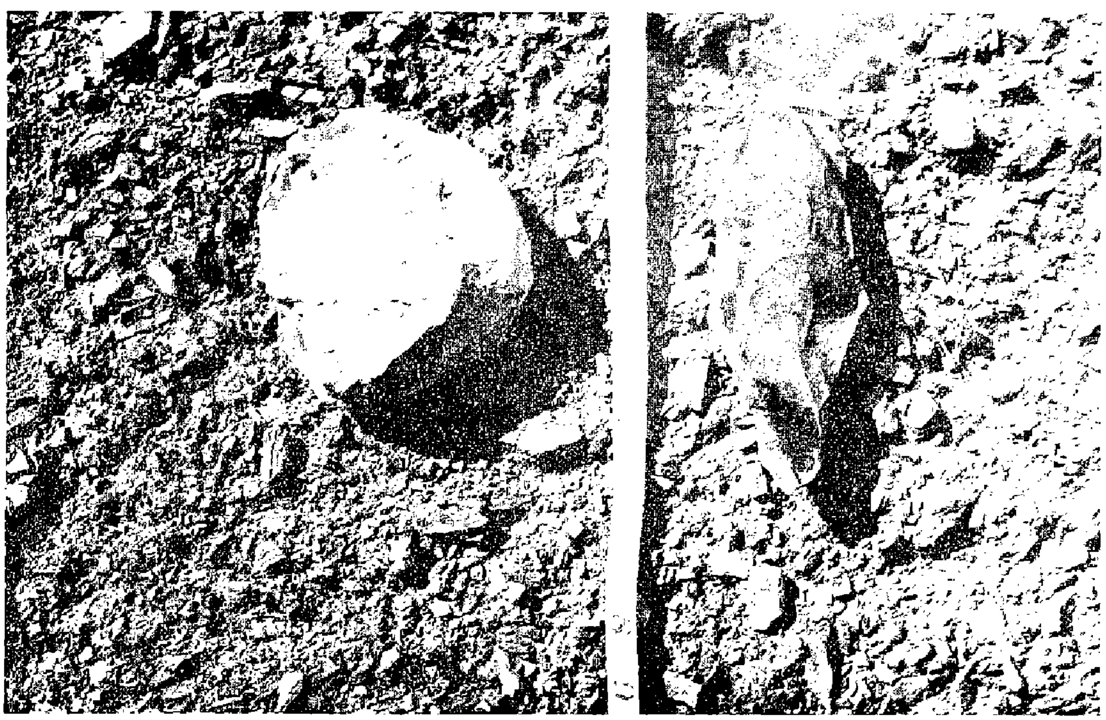


Figure 8. Black siliceous masses taken from the Prairie Mountain-Markham Mill-Wesley section. These are believed to be the "chalcedonic masses" which Harlton (1938, p. 888) described as the most diagnostic feature of the Wesley siliceous shale. The two pictured are the only such masses found on the outcrop.

stated that they "consist of dark-gray to black chalcedony with the color becoming darker toward the center." Because the siliceous masses were not found elsewhere in the stratigraphic section and because of their apparent similarity to Harlton's "chalcedonic masses," this feature was used to correlate the soft shale in which they occur with the siliceous shale of the Wesley Formation. The masses and the poor outcrop of siliceous Wesley Shale in the Rich Mountain measured section are shown in figure 8.

The lowermost siliceous mass is 175 feet below the base of the Game Refuge Sandstone, but the shales containing it continue lower in section. The location of the siliceous mass indicates that the Wesley is more than 175 feet thick. A few small exposures suggest that it is dominantly blocky and splintery dark-gray shale.

GAME REFUGE FORMATION

The Game Refuge Formation contains a higher proportion of resistant sandstone than do the rocks above and below. For this reason it forms a prominent ridge. In the Rich Mountain measured section its contact with the underlying Wesley Shale is gradational, for the transition from shale in the Wesley to sandstone in the Game Refuge spans an interval of 20 feet. Its upper contact is arbitrarily placed at the top of a thick sandstone zone. Between these two contacts are 172 feet of sandstone, 93 feet of sandstone and subordinate shale, and 247 feet of shale, for a total thickness of 512 feet.

Cline and Shelburne (1959, p. 192-193) stated that sandstones of the Game Refuge are characterized by molds of invertebrates; the plant, *Calamites;* ripple marks; small-scale cross-bedding; and, in some parts of the Ouachitas, by a thin siliceous shale. In the western Ouachitas it is 350 to 400 feet thick, but thickness decreases northward.

Beds assigned to the Game Refuge in the Rich Mountain measured section have small-scale cross-bedding and ripple marks, but, as discussed under the heading *Sedimentary Structures,* these are also found at other stratigraphic positions. Their greater abundance in the Game Refuge and overlying beds, however, is marked (interval 64, Rich Mountain measured section). The most impressive exposure of ripple marks in the area is in beds correlated with the Game Refuge at the foot of Fourche Mountain.

A zone of siliceous shale is present in the measured-section exposure. Its precise thickness could not be determined, but it is less than three feet thick. Blocks and plates which form the typical float of siliceous shales denote presence of the zone (fig. 9). The shale has laminae of various shades of gray and contains a microfauna including sponge spicules and possible radiolarians. In the few Game Refuge exposures near the western end of the Rich Mountain syncline, the siliceous shale was not observed.

In exposures to the east of Eagleton at the foot of Fourche Mountain a siliceous shale zone occurs in beds correlated with the Game Refuge. At some localities the siliceous shale reaches a thickness of six feet and consists of fissile 8- to 10-inch beds, the float of which is similar to that at the Rich Mountain occurrence.

Beds with invertebrate molds were not seen in place in the measured section, but rocks from such beds are present in the float on the south slope of the sandstone ridge. They also occur in beds of the Game Refuge exposed in NW¼ sec. 6, T. 2 N., R. 27 E. The sandstone containing them has granule-size quartz grains and abundant plant debris. In one exposure of the siliceous shale east of Eagleton (SW¼ sec. 10, T. 1 S., R. 31 W.) is a two-foot sandstone which has invertebrate molds defining planar laminae near its base.

Figure 9. Blocky float lying above a zone of siliceous shale in the Game Refuge Formation, interval 59, Rich Mountain measured section.

This lamination changes upward into cross-bedding at the top of the bed.

Cline (1960, p. 58) stated that flute casts and groove casts are few in the Game Refuge in the area in which he worked (south and west of the area of the current study). Beds included in the Game Refuge of this study commonly possess bottom casts. Randomly oriented (fig. 38) and/or parallel-oriented bottom casts are present at all exposures in the report area.

Casts of small animal tracks are at the base of some thin, cross-bedded and ripple-marked sandstones in the Rich Mountain measured section outcrop (fig. 37). Molds of clay galls pit the top of some of the thicker sandstones, and limonitic crusts cover their fracture surfaces and bedding planes.

Upper units of the Jackfork in the Spring Mountain syncline cannot be separated in the excellent roadcut exposures along State Highway 103 in sec. 26, T. 3 N., R. 25 E. Faulting is present but probably little duplication or elimination of section. The writer mapped about 2,000 feet of the upper part of the Jackfork as undifferentiated here. Beds directly below the Johns Valley Shale include two thick, apparently massive sandstones such as are present in the Game Refuge of other localities. These underlie cross-bedded and ripple-marked sandstones with flute and groove casts, which are interbedded with gray shale. The sandstone of the Jackfork grades into the shale of the overlying Johns Valley. Shale of the Wesley Formation is not readily apparent below the thick sandstones although siliceous shales are present as interbeds between thin sandstones.

The proportion of sandstone in the upper part of the Jackfork sequence of the Spring Mountain syncline is relatively high and consequently the sequence forms a ridge south of the Johns Valley. The high proportion of sandstone is in marked contrast to the thick section of easily eroded shales in the upper part of the Jackfork of Rich and Black Fork Mountains. Because Black Fork Mountain lies along strike to the east-northeast of the State Highway 103 exposures, the difference in lithologic character must be due either to an abrupt east-west facies change or to faulting. As abrupt facies changes along strike are unknown elsewhere in the Ouachitas, faulting is the more probable cause. Possibly the Jackfork has thinned northward by convergence so that the 2,000 feet exposed on State

Highway 103 represents more than its upper portion; the 2,000 feet here could represent most of the group.

Post-Jackfork Rocks

Johns Valley Formation

The Johns Valley has been notable for its contained erratics and, more recently, for its record of the boundary between the Mississippian and Pennsylvanian Systems. Cline (1960, p. 60-85) gave an excellent discussion and general description of its occurrences in the western part of the Ouachita Mountains. A detailed investigation of the Johns Valley was not made during the course of this study.

The writer could not differentiate the Johns Valley Formation in exposures on Rich Mountain. In the Rich Mountain syncline, beds of the formation are mapped with those of the Atoka as a single unit and are discussed under the heading *Johns Valley-Atoka Undifferentiated.*

Outcrops of the Johns Valley are present in roadcuts of State Highway 103 and U. S. Highway 270 near the western border of the report area and on adjacent hillslopes. Part of a roadcut on U. S. Highway 270 near Stapp, Oklahoma, has been described by Harlton (1938, p. 897, 898) under the heading *Round Prairie Formation.* Beds exposed in this roadcut and the Stapp conglomerate facies (Harlton, 1938, p. 893-895) are here considered part of the Johns Valley Shale.

The roadcut exposure is in a structurally complex area so that the relative stratigraphic position in the Johns Valley of the conglomerate described by Harlton as part of the Round Prairie Formation was not determined. In order of increasing distance northward (which would be upward in section if the sequence is normal) from the conglomerate are gray shales containing (1) thin unfossiliferous sandstones, (2) chert and limestone erratics, and (3) sandstones with an abundant mold fauna. The fossiliferous sandstones are at the extreme north end of the cut. Limestone erratics reach a maximum dimension of about 15 feet. In a shale pit to the west along strike (in NE¼ sec. 12, T. 3 N., R. 25 E.) the erratics have been removed from shale matrix and their shapes can be observed. The largest limestone block thus uncovered is approximately 6 feet by 4 feet by 4 feet and is covered with a breccia crust of limestone cobbles and small boulders. Nearly as large is a 6-foot by 4-foot by 2-foot mass of conglomerate made up of limestone and chert boulders. Some of the erratic masses are angular, some are rounded. A correlation between size and roundness is not apparent.

To the south of the conglomerate in the roadcut exposure is a section of sandstone and interbedded shale which contains no erratics. Harlton (1938, p. 888) showed this sequence in fault contact with the Johns Valley of the north end of the roadcut. Although faulting is probable, the amount of fault movement is impossible to determined. Beds throughout the outcrop have similar attitudes.

The stratigraphic position of the sandstone-shale sequence to the south of the conglomerate is questionable. Sandstones in it bear little resemblance to those of the Game Refuge exposed elsewhere and appear more like some sandstones of the Atoka. Yet, assignment of them to the Atoka is less reasonable structurally. Perhaps they are part of a thickened section of Johns Valley but, with present knowledge, this cannot be proved. However, because placement in the Johns Valley leads to the simplest structural interpretation, the beds are mapped accordingly (pl. I).

The Johns Valley grades downward into Jackfork strata in roadcuts on State Highway 103 in sec. 26, T. 3 N., R. 25 E. It underlies Atoka beds below a fault contact. The faulting is probably local and is due, at least in part, to downslope movements so that the stratigraphic sequence is nearly normal. Two dismicrite ("birdseye limestone") boulders, the maximum dimensions of which are four to five feet, are present. The bedding of one is concordant with that of the enclosing shales, but bedding of the other is discordant. Smaller masses of chert occur in association with the limestone erratics near the top of the exposure. Molds of fragments of pelecypods, gastropods, bryozoans, brachiopods, and crinoid columnals occur in coarse sandstones overlying the Johns Valley.

The lower part of the Johns Valley is poorly exposed, being represented by a few outcrops of dark-gray to black shale containing a few thin white calcareous veinlets.

The approximate thickness of exposed beds of the Johns Valley at the State Highway 103 locality is 500 feet. The thickness of erratic-bearing beds on U. S. Highway 270 is also approximately 500 feet; however, inclusion of the lower sandstone-shale sequence increases the exposed thickness to about 900 feet. Because of faulting, the thickness of the formation cannot be determined at either locality.

STAPP CONGLOMERATE FACIES OF THE JOHNS VALLEY FORMATION

Harlton (1938, p. 889-895) referred to the Stapp Conglomerate Member of the Union Valley Formation. It is exposed in cuts of the Kansas City Southern Railroad (SW¼ sec. 7, T. 3 N., R. 26 E.) across the valley to the east of the Johns Valley exposures of U. S. Highway 270 (pl. I).

The conglomerate consists of rounded pebbles, cobbles, and boulders embedded in a sandstone matrix. The coarse fraction is composed of light-gray limestone and dark-gray to black chert. Sandstone fills the interstices and occurs in a few distinct lenses (fig. 10). The conglomerate appears to have a closed fabric because many of the larger fragments are in contact.

In most of the exposure, conglomerate is massive and the coarser material is randomly oriented (fig. 11). However, at one location stratification is indicated by sandstone lenses, and the long axes of associated coarse material are parallel to the bedding (fig. 10). The lenses are further stratified by laminae consisting of granules, pebbles,

Figure 10. The Stapp Conglomerate in the railroad cut in SW¼ sec. 7, T. 3 N., R. 26 E. The pick is on a sandstone lens (see text for description). Flute casts on the base of the Atoka Sandstone bed may be seen at the top of the picture.

and cobbles. At the top of the sandstone lenses some of these laminae continue laterally into the conglomerate.

At the top of the outcrop shown in figure 10 is a one-foot zone of gray shale which is overlain by a two-foot bed of sandstone. The sandstone has contorted bedding and, at its base, excellent flute casts. Above this bed are a siliceous shale and float of sandstone containing invertebrate molds.

Beds underlying the Stapp Conglomerate are buried beneath alluvium and the nearest exposure of the Jackfork Group is several miles away, so that it is not possible to determine its stratigraphic position by superposition. Harlton's (1938, p. 890) assignment of basal Morrowan age is the same as the age designation given by Cline to the upper part of the Johns Valley Shale (Cline, 1960, p. 85). The association of an overlying siliceous shale zone and beds containing molds of invertebrate shells is similar to the Johns Valley exposures at Hairpin Curve and other localities in the Ouachitas. Ridge-forming sandstone beds lie above the Stapp Conglomerate, as they do above the Johns Valley in the State Highway 103 exposure. For these reasons the Stapp Conglomerate is correlated with the upper part of the Johns Valley Shale as defined by Cline (1960, p. 60-85).

Figure 11. The Stapp Conglomerate at same locality as shown in figure 10. Compare the nonuniform orientation of long axes of cobbles with the parallelism of axes seen in figure 10.

JOHNS VALLEY-ATOKA UNDIFFERENTIATED

In the Rich Mountain syncline the Johns Valley Shale cannot be differentiated and it is mapped with the lower shales of the Atoka as a single unit (pl. I).

The problem of definition of Atokan beds is discussed in appendix A. Regardless of the solution to this problem, the Atoka Formation is a valid mapping unit in the western Ouachita Mountains. At many localities its base is marked by erratic-bearing shales of the Johns Valley in contact with lower sandstones of the Atoka. The basal sandstones commonly contain molds of invertebrate-shell fragments.

In the Rich Mountain measured section erratics of the Johns Valley were not identified, and the first thick sandstone section occurs more than 2,000 feet above the top of the Jackfork Group. For mapping purposes the sandstone is referred to as the basal sandy member and its lower contact is shown on the geologic map. (However, the zone is not separately named on the columnar section.) Below this contact are soft beds in which a valley has been carved (pl. I). This valley separates the two prominent ridges formed by the Game Refuge Formation and the sandstones of the lower part of the Atoka.

The 2,000 feet of the strata between the Game Refuge and the sandstones of the lower part of the Atoka consists of shale and a minor amount of thin sandstone and siltstone. Dark-gray to black, splintery to subplaty shale is exposed in outcrops about one mile to the east of Skyline Drive on the south slope of Middle Mountain. Interbedded with the shale are vertically spaced sandstone and siltstone beds, most of which are less than two inches thick. Poor road outcrops reveal medium- to dark- to olive-gray, splintery to fissile to massive shale that has siliceous zones near the middle and top of the sequence. Most of the sandstone beds in the shale are less than two feet thick, are laminated, and contain carbonized plant fragments. A local occurrence of fossil-mold-containing sandstone about 575 feet from the top of the sequence may be float.

Sandstones forming the lower Atoka ridge are light gray, laminated to massive, well sorted, fossiliferous, and occur as beds up to ten feet thick. Small-scale cross-bedding and ripple marks are common as are various types of bottom casts. Planar lamination is so well developed in a few beds that they present a sheaflike appearance upon weathering. Molds of clay galls occur on the tops of several beds. Above these sandstones near the top of Rich Mountain measured section are some olive-brown to olive-gray sandstones that are friable, micaceous, argillaceous, and massive, with a general appearance similar to sandstones of the Stanley Group.

Johns Valley-Atoka beds of the Rich Mountain measured section (including the basal sandstone member of the Atoka) consist of 343 feet of sandstone, 261 feet of sandstone and subordinate shale, 633 feet of shale, 868 feet of shale and subordinate sandstone, and 1,614 feet of covered section. The total thickness is 3,719 feet.

ATOKA FORMATION

The Atoka Formation is mapped in areas on and to the north of Spring Mountain and to the north of Black Fork and Fourche Mountains. In localities where the Johns Valley is present, the basal part of the Atoka contains a relatively high proportion of sandstone that is fossiliferous in most occurrences. These sandstones form a prominent ridge in the Spring Mountain syncline and are

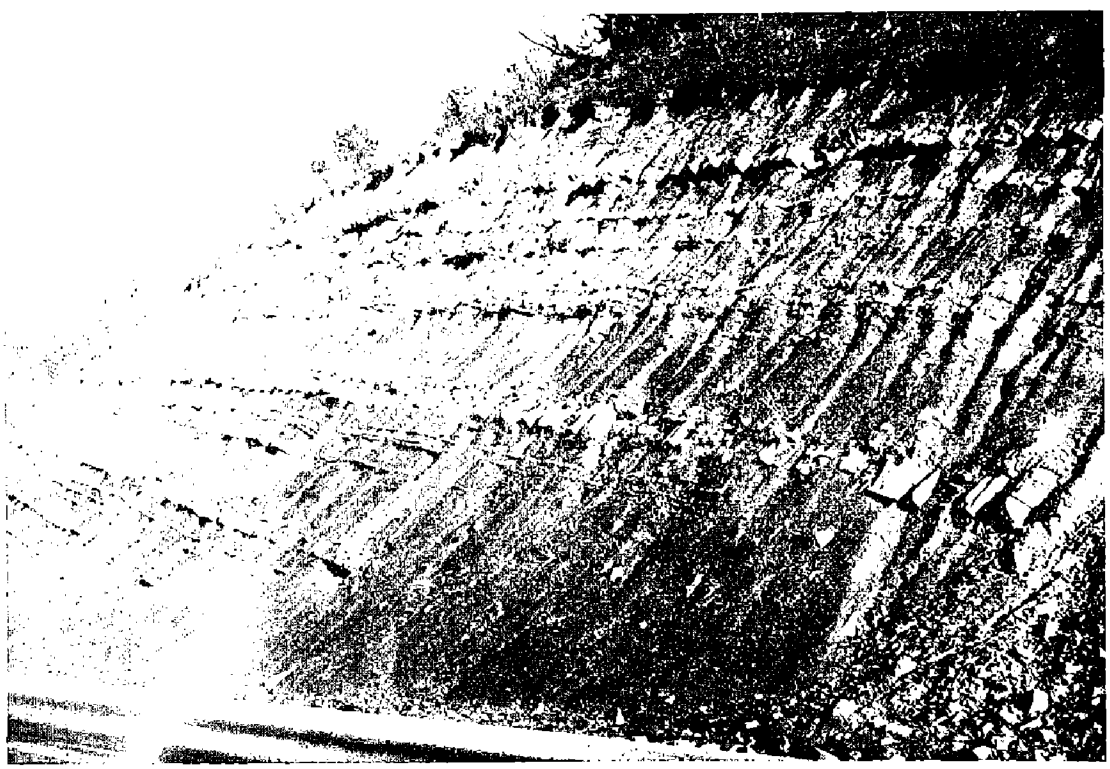


Figure 12. Shales and even-bedded sandstones of the Atoka Formation in south-central sec. 17, T. 3 N., R. 26 E. These beds are stratigraphically about 5,000 feet above the base of the formation.

present in a zone about 3,000 feet thick. A smaller ridge is present in the Stapp syncline.

Above the sandstone zone in the Spring Mountain syncline is a section composed predominantly of shale. This section is about as thick as the basal sandy zone and contains fossiliferous sandstone, such as the bed in hillside exposures directly to the west of State Highway 103 in NE¼ SE¼ sec. 17, T. 3 N., R. 26 E., and about 6,000 feet above the base of the Atoka.

Fossiliferous Atoka sandstones occur at several places to the north of Black Fork Mountain. Although their exact stratigraphic position cannot be determined, some probably are above the middle of the 18,500-foot section of the Atoka present in the Black Fork syncline (Reinemund and Danilchik, 1957). This upper Atoka placement seems particularly likely for beds cropping out in road-cuts near SE cor. NW¼ SW¼ sec. 21, T. 1 N., R. 31 W., and for those beds near the middle of the south edge of NW¼ SW¼ sec. 19, T. 1 N., R. 31 W. Localities of other fossiliferous sandstones at possibly the same stratigraphic level are near C sec. 20, T. 1 N., R. 32 W., and near C NE¼ NE¼ sec. 18, T. 3 N., R. 27 E. These beds, together with associated laminated, bottom-marked sandstones, make it possible to map the trace of the Briery fault north of Black Fork Mountain with fair precision (see section entitled *Briery Fault*).

COLLUVIUM

The steep slopes of Rich and Black Fork Mountains are covered with a mantle of weathered debris creeping downslope under the influence of gravity, especially on their north, or obsequent, slopes. The structure of the beds beneath these slopes is probably simple, and the bedrock can be determined beneath the mantle at most localities. This bedrock belongs to the upper part of the Tenmile Creek Formation or to the Moyers Formation.

The paucity of outcrops makes it impossible to map the basal siliceous shale of the Moyers and it is necessary to use aerial photographs and associated beds to map the approximate location of the siliceous shale of the Chickasaw Creek. Because of these factors, accurate presentation of field observations requires mapping of the creeping debris as colluvium. Downslope it is in gradational contact with alluvium, and upslope it is in gradational contact with bedrock exposures.

The colluvium consists of sandstone blocks in various stages of weathering, quartz sand, and clay. Downslope from resistant sandstone beds, talus has accumulated on topographic benches or in closed depressions and may spill over into stream valleys. These deposits greatly inhibit plant growth and thus are readily observable. Some local residents refer to them as rock rivers and this description seems appropriate, although geologists might be more inclined to refer to them as small rock glaciers. Their more common occurrence on south slopes appears due to the presence of a greater number of topographic traps which channel talus into concentrated streams. On north slopes the talus is spread out in sheetlike deposits.

ALLUVIUM

Two types of alluvial deposits were distinguished in the mapping: (1) terraces and (2) flood plains of present streams. These types intergrade with one another and with colluvium so that their margins locally are difficult to define.

These deposits are composed of fragments of the resistant beds exposed in the surrounding mountains. The fragments are from sand size up to boulder size and nearly all are composed of sandstone. Most fragments show some degree of rounding, but rounding varies with size, distance of transport and resistance to rounding of the constituent material. In older deposits on some of the higher terraces, the sandstone fragments have partially disintegrated to yield a sandy soil.

Terraces are most prominent near Mena, Arkansas. Mena itself is built upon several terrace levels. The maximum number of terraces in any one area is four. The highest terrace is about 200 feet above adjacent present-day streams. An attempt to correlate terraces from one area to another was not successful.

Most present-day streams are choked with debris and are not actively downcutting except in narrow mountain valleys and at knickpoints. Most good exposures of bedrock are found at the latter localities. In the broad valleys carved by the Kiamichi River and Mountain Fork of the Little River, streams have a characteristic braided pattern where they have established a wide flood plain. These streams and their ancestors have laid down an extensive cover over beds of the Stanley Group.

SEDIMENTARY STRUCTURES

RIPPLE MARKS AND CROSS-BEDDING

Impressed by the abundance of ripple marks and cross-bedding in the Stanley, Honess (1923, p. 197-198) noted that one of the outstanding facts of Stanley sedimentation is "the ripple-marked and cross-bedded structure of nearly all of the strata, sandstones and shales alike." Hendricks and others (1947) described the Atoka as being ripple marked but did not mention this characteristic in their descriptions of the Stanley and Jackfork; nor did Cline (1960, p. 58) mention ripple marks in his description of the Stanley, indicating only that they are not common in the Jackfork except in the Game Refuge. Cline (1960, p. 89) further stated, "Cross-bedding is rather uncommon in Stanley, Jackfork, and Atoka sandstones, with the exception of the Game Refuge Formation at the top of the Jackfork Group."

Honess (1923) apparently tried to make a distinction between ripple marks of current origin and undulatory surfaces of unknown origin. This is evident from a comparison of his section of a portion

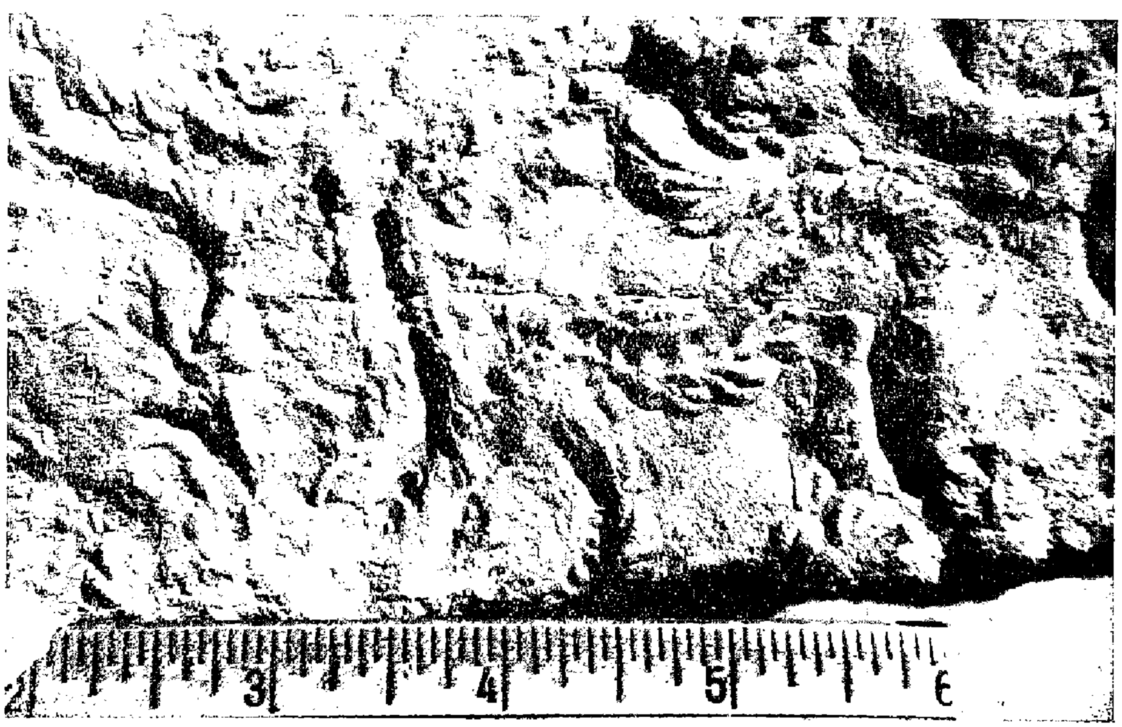


Figure 13. Trails of a bottom-dwelling organism preserved on the top of a 3-inch sandstone near the middle of the Wildhorse Mountain Formation. Sample taken from the railroad cut in SE¼ NE¼ sec. 8, T. 1 S., R. 31 W.

of the middle part of the Stanley (p. 152-163) and his section of the Stanley-Jackfork transition series (p. 165-173). In the present study, the term "ripple marked" is applied to any bed whose upper surface is uniformly undulatory in exposed segments. By this definition, ripple-marked surfaces are not necessarily of current origin. However, in many occurrences the ripple marks are associated with small-scale cross-bedding and were, therefore, formed by currents. Only where weathering has not etched the lamination pattern at the tops of beds is there a question as to the current origin of the ripple marks.

The upper surfaces of sandstone and siltstone beds of the Jackfork-Johns Valley-Atoka sequence in the report area possess a variety of configurations (pl. II). Because of the limited exposure of top surfaces, positive statements cannot be made about the comparative abundance of ripple marks at various horizons. Throughout the Ouachita Mountains the lower part of the Wildhorse Mountain Formation crops out on obsequent slopes above the Chickasaw Creek Formation. Top surfaces are exposed at few places on these slopes so that one would expect few ripple-mark observations even though the ripples might be present. On the other hand, dip slopes of sandstones at the top of the Wildhorse Mountain, in the Game Ref-

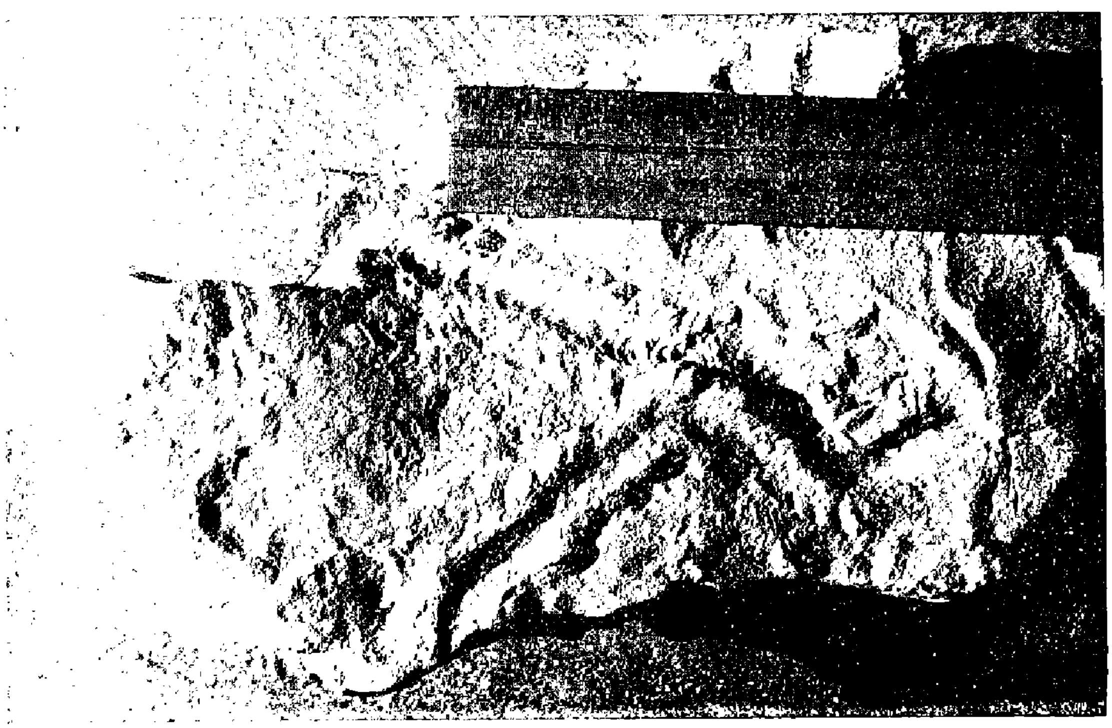


Figure 14. The upper surface of a $1\frac{1}{2}$-inch trail-bearing sandstone from interval 41, Rich Mountain measured section (near top of Wildhorse Mountain Formation).

uge, and in the lower part of the Atoka are exposed at many localities and ripple marks would be readily observable, if present. Incidence of bottom-surface exposures bears the opposite relationship to the topography and stratigraphic position so that observation of them would be less likely on dip slopes than on obsequent slopes.

More and better developed ripple marks were observed in the Game Refuge, Johns Valley, and Atoka Formations (pl. II) than lower in the stratigraphic section. Ripple marks and small-scale cross-bedding are also common near the Stanley-Jackfork contact. Cline and Moretti (1956, p. 18) described 300 feet of section at the base of the Jackfork as containing ripple-marked sandstones. They also mentioned small-scale cross-bedding of a few beds near the top of the Stanley (p. 10).

Ripple marks with linear subparallel crests and troughs are few. Prominent exceptions to this are found in well-exposed undulating surfaces of the Game Refuge in SW¼ NW¼ sec. 14, T. 1 S., R. 31 W., and in a limited outcrop of sandstones of the Wildhorse Mountain in SE¼ NE¼ sec. 8, T. 1 S., R. 31 W. (fig. 15). In both localities the ripples are asymmetrical, but the asymmetry is more pronounced in the exposure of Wildhorse Mountain beds. The crest-to-crest distance ranges from one to five inches, the shorter lengths

Figure 15. A small exposure of a ripple-marked bed of interval 45, Rich Mountain measured section (top Wildhorse Mountain Formation).

being in the Game Refuge exposure. At the Game Refuge outcrop are apparently larger ripples with crestal spacing of about three feet. These larger ripples have the same orientation as the smaller ripples, which are superimposed upon them. Low, elongate ridges about one-fourth inch high and two feet long lie in the troughs of some of the small ripples. Ripple-marked beds of the Game Refuge also possess small-scale cross-bedding, but it was not determined whether the ripple-marked bed of the Wildhorse Mountain is cross-bedded although other beds at the same exposure are cross-bedded. At both localities beds with flat upper surfaces are few.

A small exposure of ripple marks with linear, subparallel crests may be seen in figure 15. These have a crest-to-crest distance of two to three inches and have a stratigraphic position at the top of the Wildhorse Mountain Formation.

The more common ripples have short crest lengths or no well-defined linear crests. None of these ripples, however, has a geometric regularity allowing an interpretation of current direction. In several beds this absence of geometric regularity appears to be due to plastic flowage of their tops. Some of the thin cross-bedded sandstones have smooth low-relief mounds and depressions that appear similar to the periclinal undulations described by ten Haff

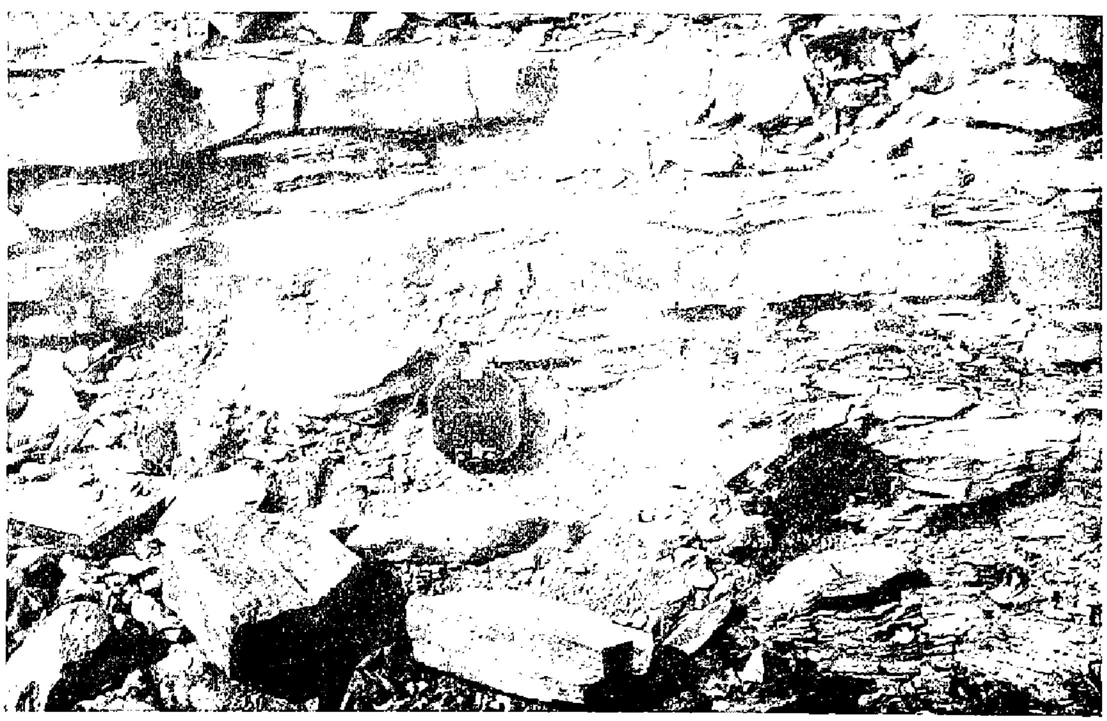


Figure 16. Thin, cross-bedded and ripple-marked sandstones possessing bottom casts (note upside-down blocks in the foreground) at base of interval 5, East Ward Lake measured section.

(1959a, p. 22), but are smaller and show well-developed cross-bedding.

The ripple profiles approximate smooth curves. No trochoidal profiles were observed. No ripple indices were determined because of the geometric irregularity of many upper surfaces.

Cross-bedded ripple-marked beds are quite distinctive and occur at many levels throughout the Stanley-Atoka sequence. Most are less than four inches thick and the topset-bottomset spacing measured perpendicularly to the bedding planes is less than 1.5 inches. The foresets are scoop shaped, resembling those that characterize festoon cross-beds as described by Pettijohn (1957, p. 169), except that these are smaller than are his examples. Many show the effect of post-depositional distortion that is apparently due to plastic flow. Such contorted beds do not preserve the shape of the original ripple-marked upper surfaces.

A cross-bedded zone no more than a few inches thick underlies the upper ripple-marked surfaces of many ½- to 3-foot sandstones. It differs from that found in thin beds only in its association with thicker beds.

Also occurring in a few of the thicker beds, particularly in the Atoka Formation, is a cross-bedded zone that is about midway between the top and the base of the bed. Bedding in this zone is contorted in the few observed occurrences.

The forementioned are the more common occurrences of cross-bedding observed by the writer. However, cross-bedding is not limited to these relationships. It may occur at any position in a bed and within beds of greatly differing thicknesses. Where cross-bedding has been definitely identified, it is small scale, as shown by a small topset-bottomset spacing.

WAVY AND PLANAR LAMINATION

Various sandstones of the Stanley-Atoka sequence show an upward change from planar to ripple-type cross lamination. Where this occurs, planar lamination (lamination in planes parallel to the bedding) may grade upward into wavy lamination which, in turn, grades into cross-bedding at the top of the bed. Normally, however, most of the bed is planar laminated, and an abrupt transition to a thin cross-bedded zone occurs at the top (fig. 17). This habit, common enough to be useful in top-bottom determinations, is modified in several beds by the presence of a massive zone at the base and, in a few beds, by planar lamination above the cross-bedded zone. Many beds do not have the upper part of the sequence and thus display only the massive lower zone with planar lamination above. Wavy lamination may occur at the top.

The sequence need not be present in its entirety for use in top-bottom determination. Fortunately for structural interpretation, sandstones of the Atoka Formation have more prominent lamination than have those stratigraphically lower. Some of the Atoka sandstones are so well laminated that their upper portions weather into sheaves of plates bounded by planar or subplanar parallel surfaces. Sandstones of similar habit are apparently absent, or are only sparingly present, lower in the section.

Laminae composition.—Laminae near the top of sandstone strata consist of mica, clay, clay galls, and widely differing amounts of plant fragments. The tops of many beds have countless impressions of plant fragments, most of which are flat. Some *Calamites* stems are preserved as molds or casts. In a few occurrences the carbonized plant fragments are compacted in thin seams that have vitreous luster and "coaly" appearance (fig. 28).

Figure 17. A 5-inch sandstone that is planar laminated except for a thin, cross-bedded zone at its top. Note the abrupt change from planar to cross lamination. This sandstone is typical of many in the Jackfork (sample taken from the lower beds of undifferentiated Jackfork exposed in roadcuts of State Highway 103).

The lower laminae of some beds contain molds of fragments of invertebrates, particularly crinoid columnals. If the beds are a foot or more thick, the molds normally are restricted to the lower portion of the bed. This habit applies also to many beds in which lamination is not apparent.

Diameters of the crinoid columnals are greater than the modal class of the bed in which they occur. Most fall in the sand- and granule-size ranges and a few fall in the lower pebble-size range (fig. 18). The smaller fragments occur in the finer grained sandstones; the larger fragments in the coarser grained sandstones.

Although molds of invertebrates are probably more common than previously was realized, they are absent from most beds. Lower laminae in many of these strata are indistinct, and, because the beds do not readily break along them, their composition is not so well known as that of the upper laminae. However, both lower and upper laminae appear to have similar compositions.

Lamination in the sandstones is also marked by a concentration

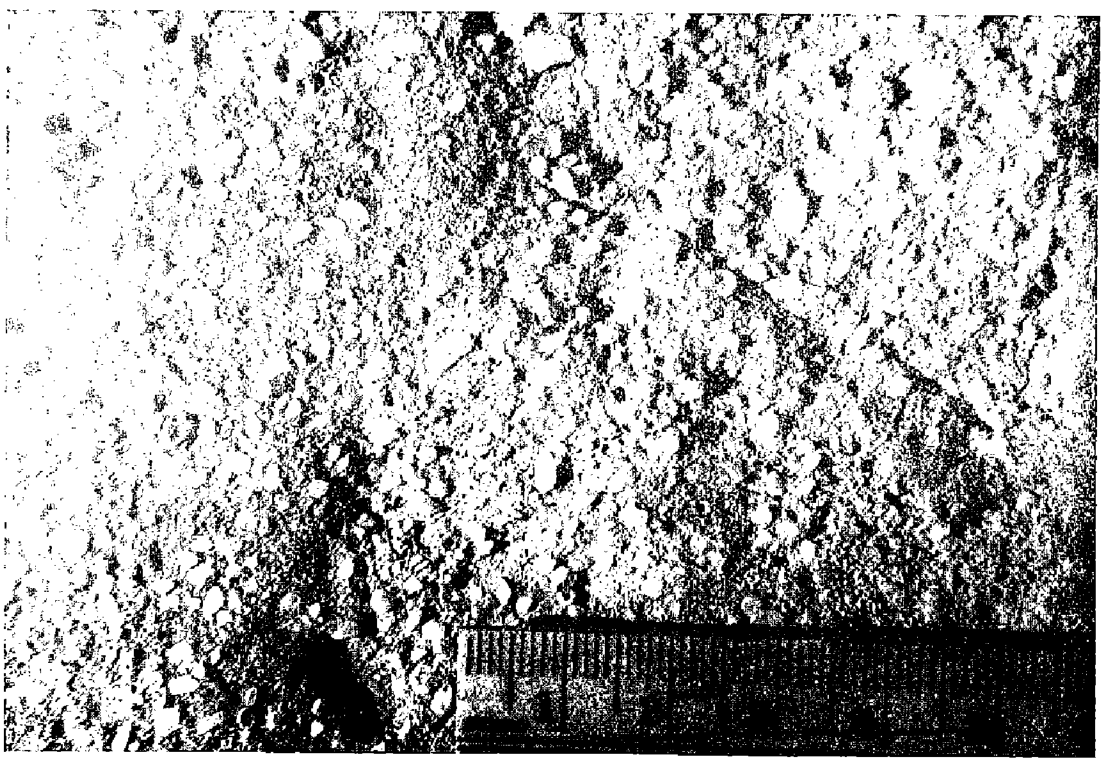


Figure 18. Granules and coarse sand composed of quartz and crinoid columnals etched from a matrix, the estimated median size of which lies in the medium sand-size range (a crinoid columnal lies 2 mm directly above the 1 cm mark on the ruler). Note the wide range of rounding. This sample was obtained from float of interval 5, East Ward Lake measured section, and, although conglomeratic throughout, it typifies the basal zone of fossiliferous sandstones of the Jackfork, Johns Valley, and Atoka. The surface viewed is probably the base of the bed.

of heavy minerals and by vertical variation of grain size. The relative abundance of these types of lamination is not known.

GRADED BEDDING

An upward decrease in grain size is not megascopically apparent in most of the sandstone beds of the Jackfork-Atoka sequence. How-

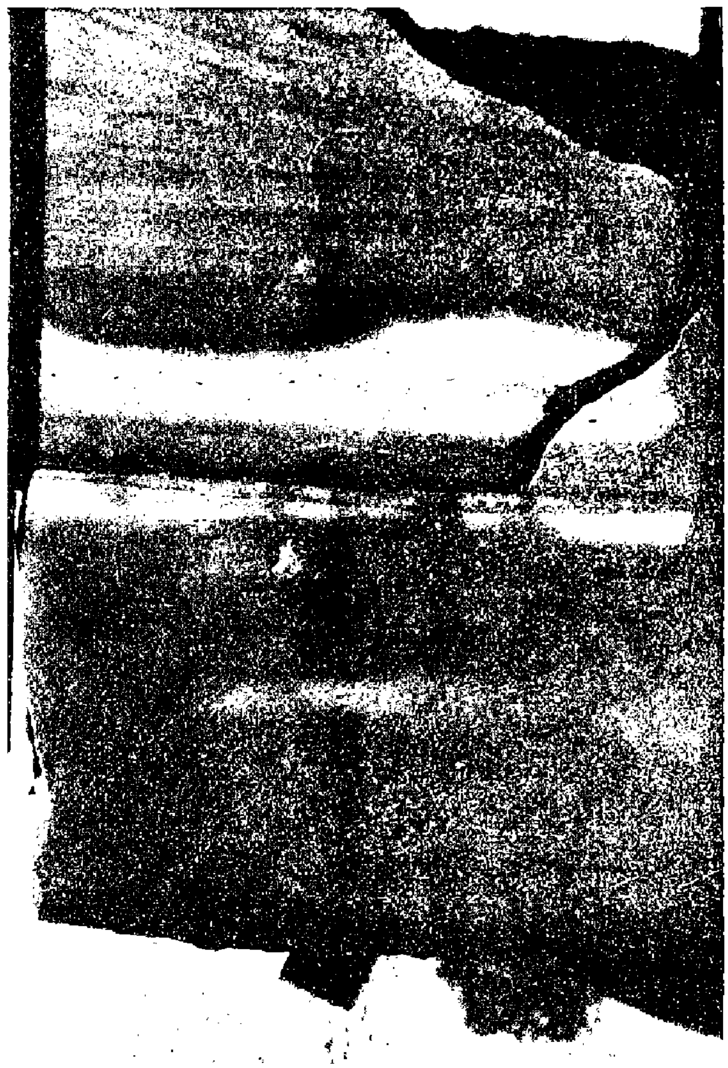


Figure 19. A "graded" bed composed of coarse, massive or planar-laminated sandstone below and fine, planar or cross-laminated sandstone above. The highest lamination containing soluble carbonate fossil fragments is in the upper white area (the rock has been deeply stained by iron oxide) and has been made visible by pitting. Pitting produced by solution identifies fossiliferous laminae on weathered surfaces (the pictured surface was saw-cut) of fossiliferous beds of the Jackfork, Johns Valley, and Atoka. Marks on the ruler to the left are 1/16-inch apart. The sample is of interval 77, Rich Mountain measured section (Johns Valley-Atoka undifferentiated).

ever, within the lower sandstone ridge in the Atoka of the Rich Mountain syncline are several beds in which the upper laminae have smaller grain sizes than have the lower laminae (fig. 19). The lower part of these sandstone beds generally contains abundant granule- or coarse-size grains of quartz or fossil fragments distributed in closely spaced planar laminae. These laminae may be poorly defined, and, if so, the rock appears massive.

An upward increase occurs in spacing of the coarse laminae. Changes in laminae composition to clay, mica, and, in some rocks, plant debris occur near the top of the bed. Upward these fine laminae become closely spaced. Small-scale cross-bedding may underlie the upper surface. Individual laminae and interlayers may be well sorted. The thickness of beds possessing this type of graded bedding ranges from 0.5 to 3 feet and is commonly between 1 and 2 feet.

Thin sections transverse to bedding taken near the top and the bottom of RM 6 and RM 8 show the same grain size. Thicknesses of beds from which these slides were made are 2 inches and 5 inches, respectively. An example of micrograding is shown in figure 45A.

CLAY GALLS

Molds of clay galls are common at various levels of the Stanley-Atoka sequence. Because exposure is necessary for observation and the softer clay is quickly removed from exposed sandstone surfaces, only molds remain as evidence of most of the clay galls. One exception was a partly filled cavity of a discoidal clay gall about one inch thick and three inches in diameter. This buried cavity was revealed when the bed containing it was broken. The cavity was partly filled with loose clay which X-ray analysis revealed to have an illite lattice.

A few freshly exposed sandstones have galls in which the clay is compacted but may be removed with a knife. Other cavities are lined with clay and are filled with sandstone. Exposures of filled cavities, however, are uncommon.

The cavities have many different shapes (figs. 20-22). Most of them have smooth, rounded, ellipsoidal outlines. Several have clamlike shapes. Their maximum dimension may exceed a foot, but in most instances is less than six inches. Clusters of them are more common at or near the tops of sandstone beds with long dimensions of the cavities parallel to the bedding planes. They also occur as sparsely distributed depressions on the tops of other beds. Where present within strata, the cavities are scattered and their long dimensions appear to be randomly oriented.

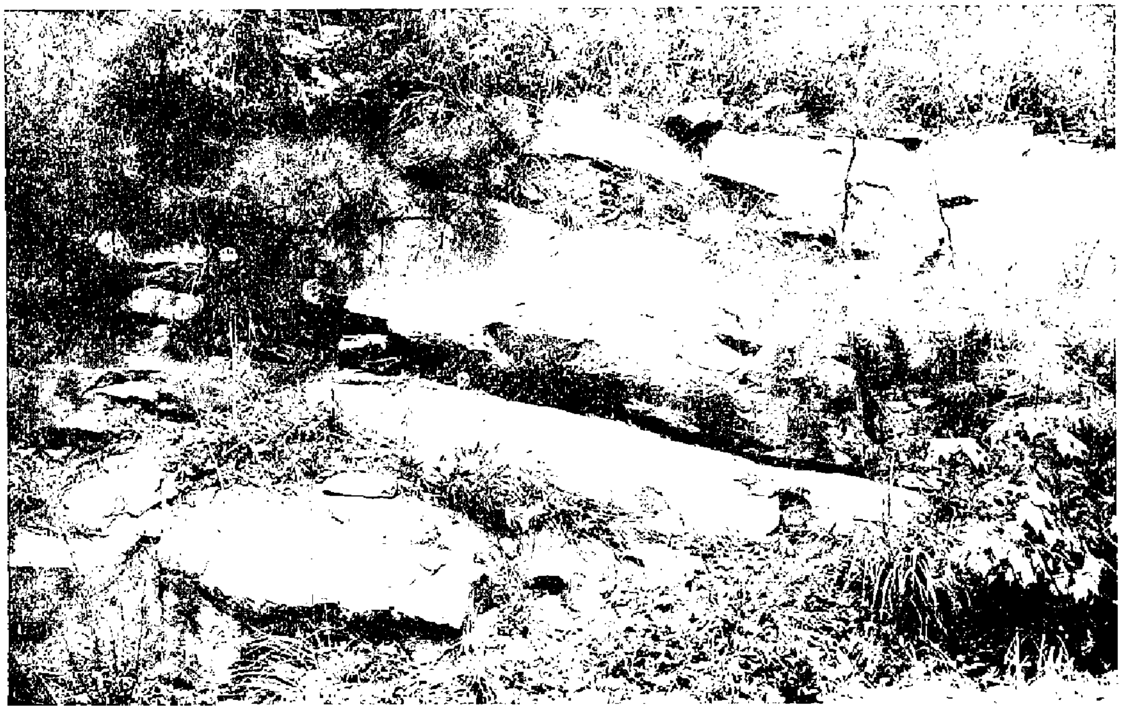


Figure 20. Large cavities caused by the removal of soft material (probably clay) during weathering. Looking northward at the dip slope of jointed sandstone in interval 45, Rich Mountain measured section (top Wildhorse Mountain).

Figure 21. Clay gall molds in the top of a sandstone in interval 80, Rich Mountain measured section. (Johns Valley-Atoka undifferentiated).

Figure 22. Large clay gall molds on the top of an Atoka sandstone exposed in roadcuts just south of the crest of Blue Mountain (not in study area). Imprints of *Calamites* stem segments are also present.

PLASTIC-FLOW STRUCTURES

Plastic-flow structures are ubiquitous at many levels of the Stanley-Atoka sequence. Perhaps the most impressive example is illustrated in figure 23. This sandstone dike is six to eight inches thick in the lower part of the exposure and tapers upward. It cuts shales and thin sandstones of the Stanley south of the Windingstair fault. Drag of the intruded beds adjacent to the dike is not consistent along its length. Some beds bend upward, others bend downward. The tortuous configuration of the dike is marked, and its contact with intruded beds is well defined. Megascopically the sandstone of the dike appears similar to that of the thin beds it intrudes except that it is massive. Cone-in-cone structures are present in calcareous masses isolated in the shale of the same exposure. Discussion of a possible origin of the dike is presented under the heading *The Hubbert-Rubey Hypothesis and Sandstone Dikes.*

Contorted bedding and ripple distortion have been discussed under the heading *Ripple Marks and Cross-bedding.* In the few places where noted, the small internal contortion folds possess axial surfaces with opposing dips. Overturning in both directions was observable (fig. 24).

Another type of plastic deformation is illustrated in figure 25. The surface of this rock shows the effect of apparent small-scale faulting. Several generations of apparent faulting may be discerned. One might justifiably criticize the use of the term "faulting" here because it normally implies nonplastic deformation. However, nonplastic and plastic deformation can be gradational, as they are in this example. The "pseudofaults" do not extend through the rock, so that the rock does not consist of a series of offset blocks of equal thickness as would be the case in true faults. Fault planes are not visible, although they should be if these were true faults.

Figure 23. Sandstone dike in Stanley strata a short distance south of Windingstair fault in east-center sec. 4, T. 2 S., R. 31 W., beside Rock Creek. Note that the lowest sandstone bed cut by the dike maintains its thickness up to the dike walls, but the second sandstone from the bottom pinches out to the right of the dike. Northward view; beds dipping away from observer.

Because of the plastic deformation, the "downfaulted" portions of the block are actually thin segments of the bed. The upper right-hand margin of the block in figure 25 is its thinnest part. Irregular sandstone beds and isolated boulder-size masses in thick shale sections also have surfaces with small escarpments. The boulders, however, are somewhat rounded and the escarpments may be folded.

Also in thick shale sections are discoidal or roughly ellipsoidal isolated sandstone masses in which internal laminae are contorted to various degrees. Some have laminae that are planar and continuous to the margin of the mass; for example, the laminae composed of carbonized plant matter seen in figure 26. In others the laminae are warped or folded, as may be seen in the lower boulder in figure 27. Plant imprints on the outer surface of this boulder are similar to those found on many others. Few thick shale sections are free of isolated sandstone masses or discontinuous sandstone beds.

At the top of interval 5 of the East Ward Lake measured section is a sandstone bed ranging in thickness from 2 to 3 inches.

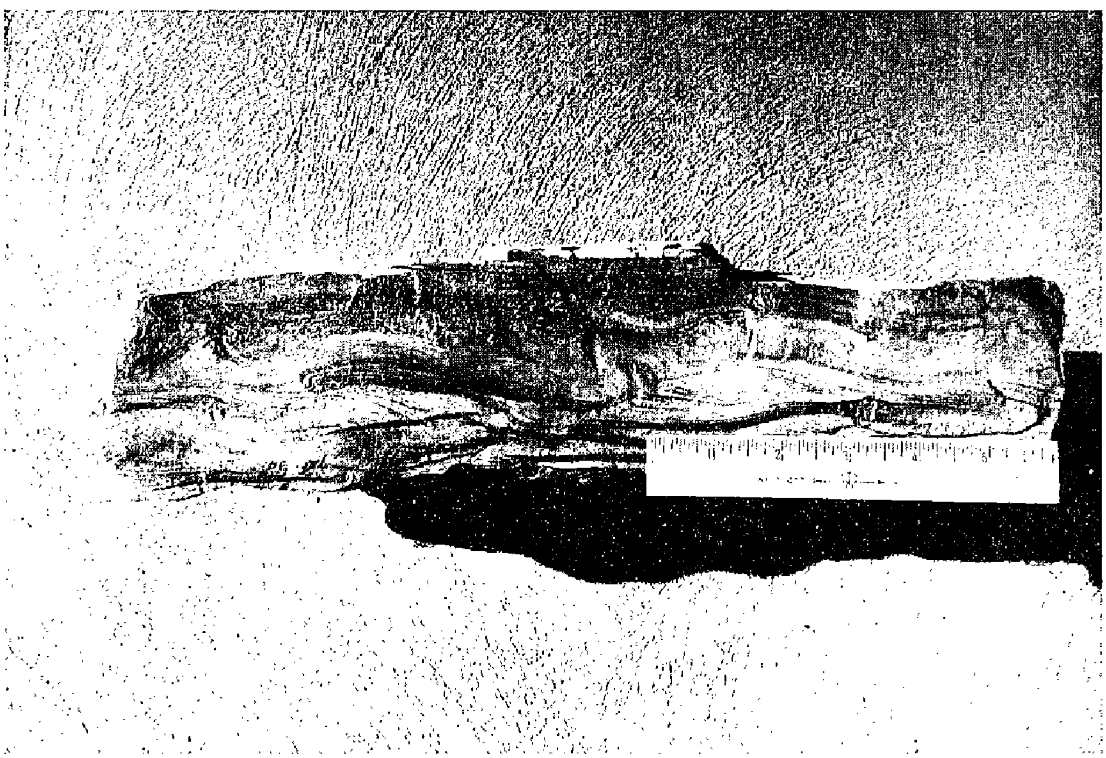


Figure 24. Example of contorted bedding in a cross-bedded thin sandstone of the Stanley Shale in SW¼ SW¼ sec. 7, T. 2 N., R. 22 E. Note that contortion folds at the left end of the block are overturned in the opposite direction to that in upper right center.

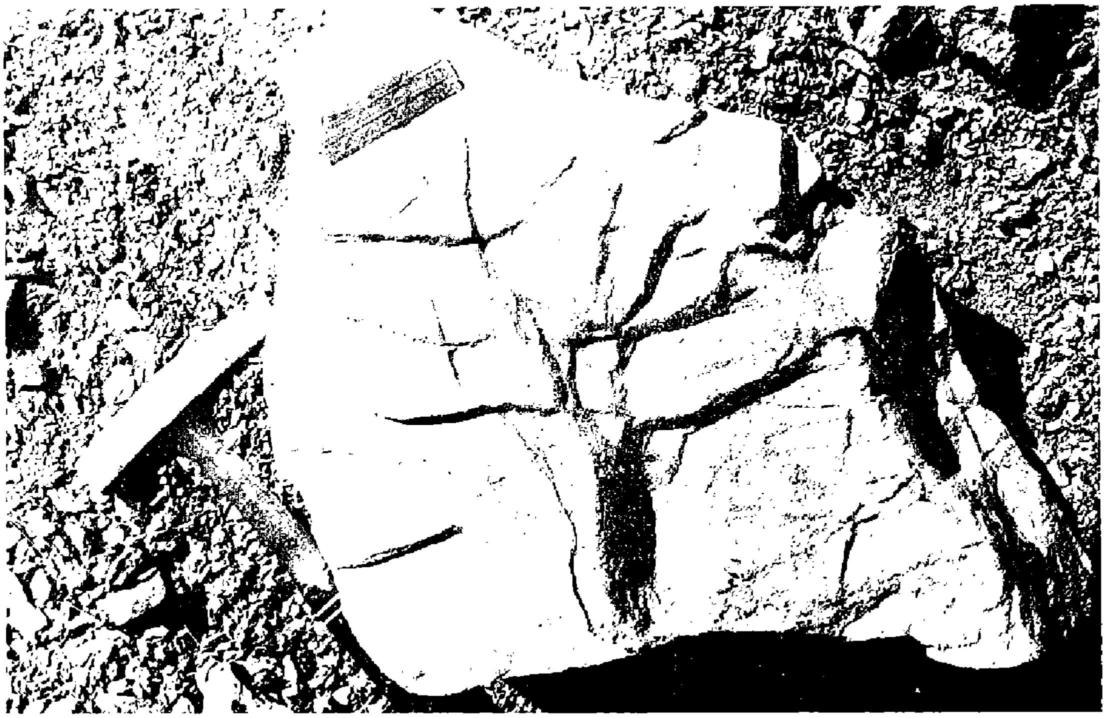


Figure 25. Escarpments caused by "faulting" of the bottom of a sandstone bed while the sandstone was in a semiplastic state. Note that some escarpments offset others, thereby indicating that they must have formed later. The thickest (6 inches) part of the block is left of the center of the photograph; the thinnest (1½ inches) part is near the right-hand edge. The sandstone is well sorted and has planar lamination near its base and cross-bedding at its top. The top surface of the sandstone has a few low escarpments which could not be correlated with bottom-surface escarpments. Taken from interval 8, Ward Lake Spillway measured section (Wildhorse Mountain Formation).

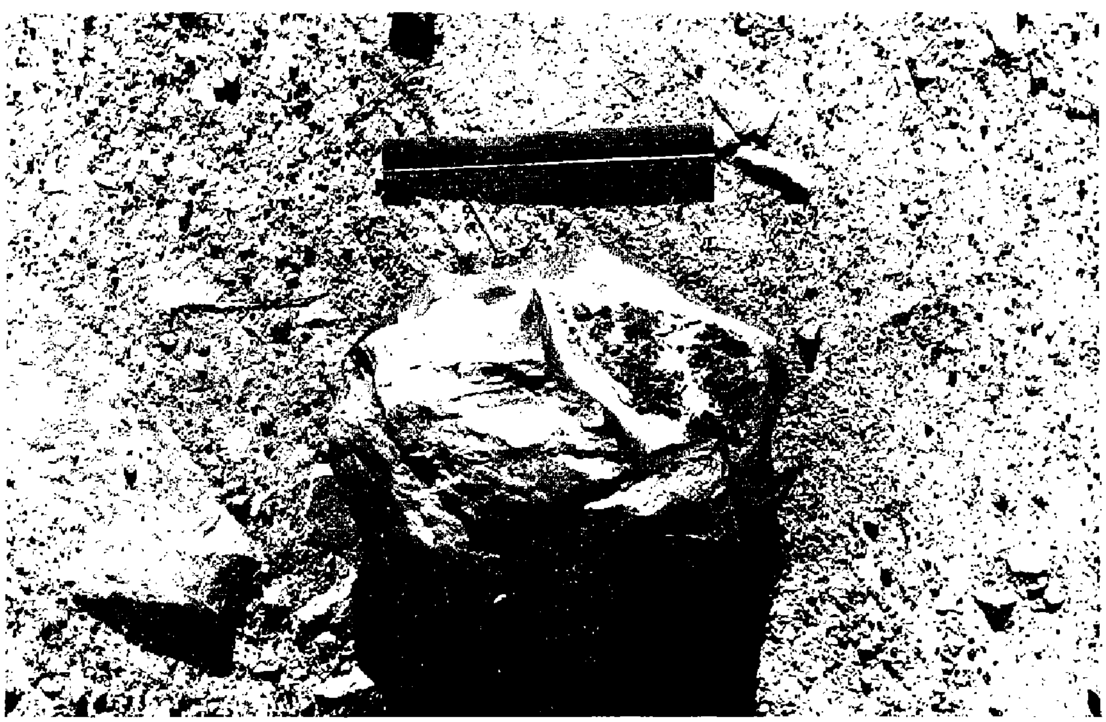


Figure 26. Discoidal sandstone mass containing planar laminae of carbonized plant fragments. Taken from float of interval 69 (Johns Valley-Atoka undifferentiated).

Figure 27. The lower rock has contorted laminae. Plant imprints are on its upper surface. The upper rock is composed of limonitic shells surrounding a small sandy core and is 7 inches in diameter and 3½ inches thick. Both rocks are from beds of interval 58, Rich Mountain measured section (upper Prairie Mountain-Markham Mill-Wesley; probably Wesley).

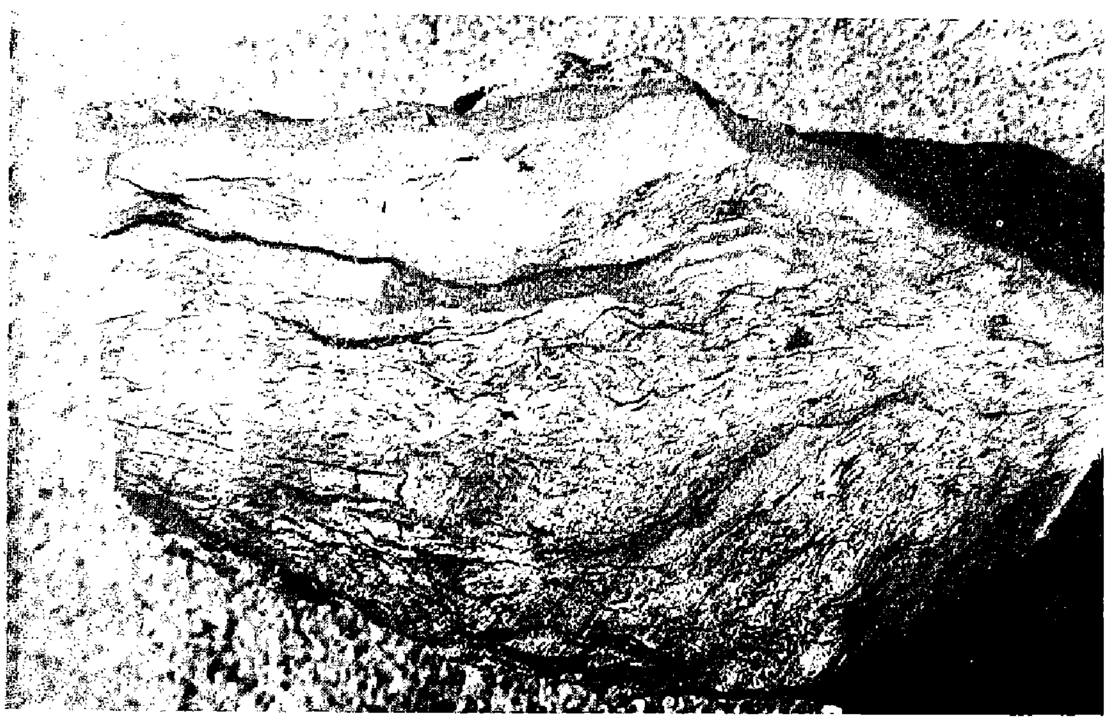


Figure 28. Plastic deformation of this sample is shown by abundant carbonized plant fragments. A vitreous luster imparts a "coaly" appearance to seams at the base of the rock. The sample is from interval 57, Rich Mountain measured section (Prairie Mountain-Markham Mill-Wesley undifferentiated) and is pictured at true scale.

Rising from this bed and protruding several feet upward into the overlying shale are large contorted sandstone masses, some of which exceed seven feet in their long dimension (fig. 29). These sandstone masses resemble the isolated boulders found elsewhere in shale intervals except that their attachments to a parent bed is here preserved. They have abundant plant fragments, and one has molds of crinoid columnals.

Cline (1960, p. 74, 76, 77, 82) referred to "rolled sandstones," which he considered to be characteristic of the Johns Valley Shale. Shelburne (1960, p. 36, 38, 55, 57) referred to the same feature as "balled sandstones." Apparently these are what the writer has described above as "isolated sandstone masses," common to many thick shale sections in the Stanley-Atoka sequence.

The bottom surfaces of several sandstones show bulbous or sinuous downward protuberances into the underlying shale. Shale is pinched into infolds of these protuberances and, locally, between

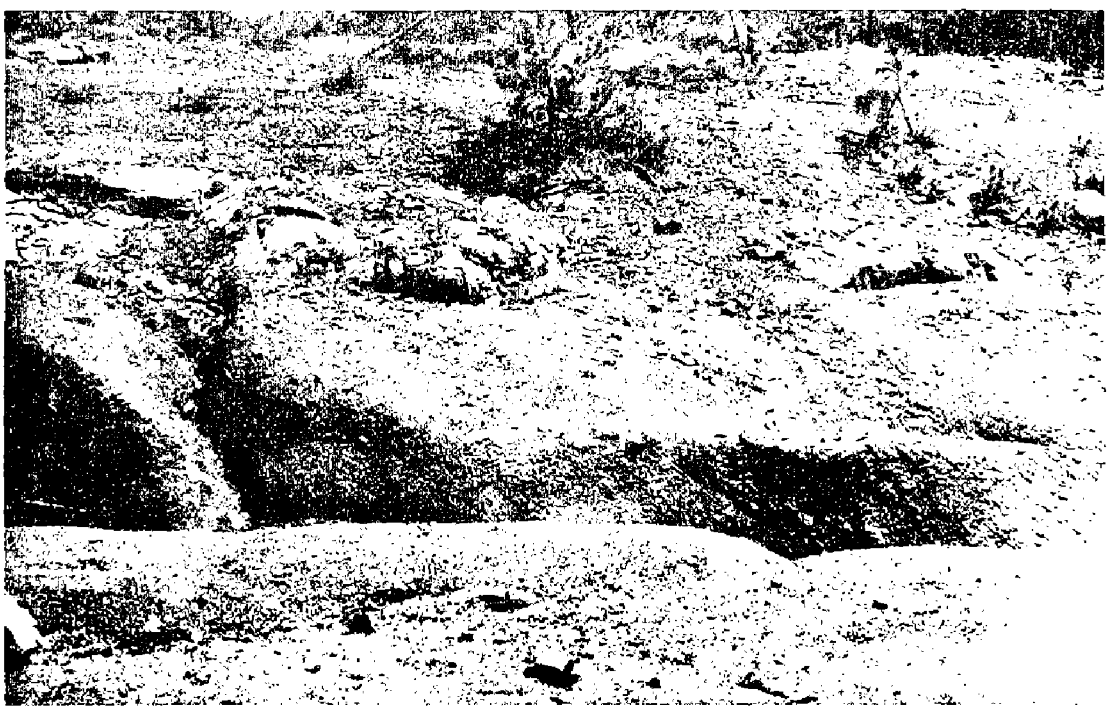


Figure 29. In the foreground is shale of interval 6, East Ward Lake measured section. The base of interval 6 is marked by the top of the sandstone in the upper middle of the picture, and the dip slope of a sandstone within interval 5 may be seen at the top of the picture. The middle sandstone is 2 to 3 inches thick except where large bulbous masses project into the overlying shale of interval 6. These masses are more than 7 feet long and are approximately 3 feet thick. They make up all but the left end of the middle sandstone of the picture. Near the left end, connection of the masses with the thin parent bed (whose dip slope is present at the left margin) is visible. A mass just to the right of the field of view contains molds of crinoid columnals.

the protuberance and the base of the bed. The protuberances show varying relief and appear to grade into flute casts, groove casts, or other bottom irregularities. Similar features have been described as load casts by Kuenen (1953, p. 1058) and are best developed on sandstone beds that are more than a foot thick.

BOTTOM-SURFACE MARKINGS

Perhaps because of their possible significance for interpretation of depositional environments, markings on the bottom surfaces of sandstones of the Stanley-Atoka sequence have been the most studied sedimentary structures of Ouachita investigators. Markings due to current action may be divided into two main groups: flute casts and groove casts. Flute casts are parallel-oriented, elongate, downward protuberances at the base of a sandstone bed, which have a blunt or rounded bulbous terminus opposite a gradually tapering and flaring end that imperceptibly merges with the lower surface of the bed. The blunt end has been interpreted as pointing upcurrent. Groove casts are parallel-oriented, elongate, downward protuber-

Figure 30. Large flute casts indicating a down-dip current direction. Faulting has offset sole markings in the foreground. Note small casts on the largest flute cast near the center of the picture. A sandstone of the Johns Valley Shale (?) exposed in the U. S. Highway 270 roadcut near Stapp, Oklahoma.

ances that do not possess a bulbous terminus. Examples of flute and groove casts are shown in figures 30-34.

Sole markings of random orientation are not the product of current action but probably represent the work of bottom organisms or depressions formed by bottom debris removed prior to deposition (fig. 38). However, normally these casts cannot be attributed definitely to organic activity. Exceptions to this generalization are well-preserved tracks and burrows (figs. 36, 37). Most nonoriented sole markings are smaller than flute or groove casts, protruding only

Figure 31. Northward view of large flute and groove casts on the base of a Jackfork sandstone exposed in the east roadcut of State Highway 103 just north of Briery fault. Note that small casts are present on the large casts as in figure 30. Current flowed from right to left.

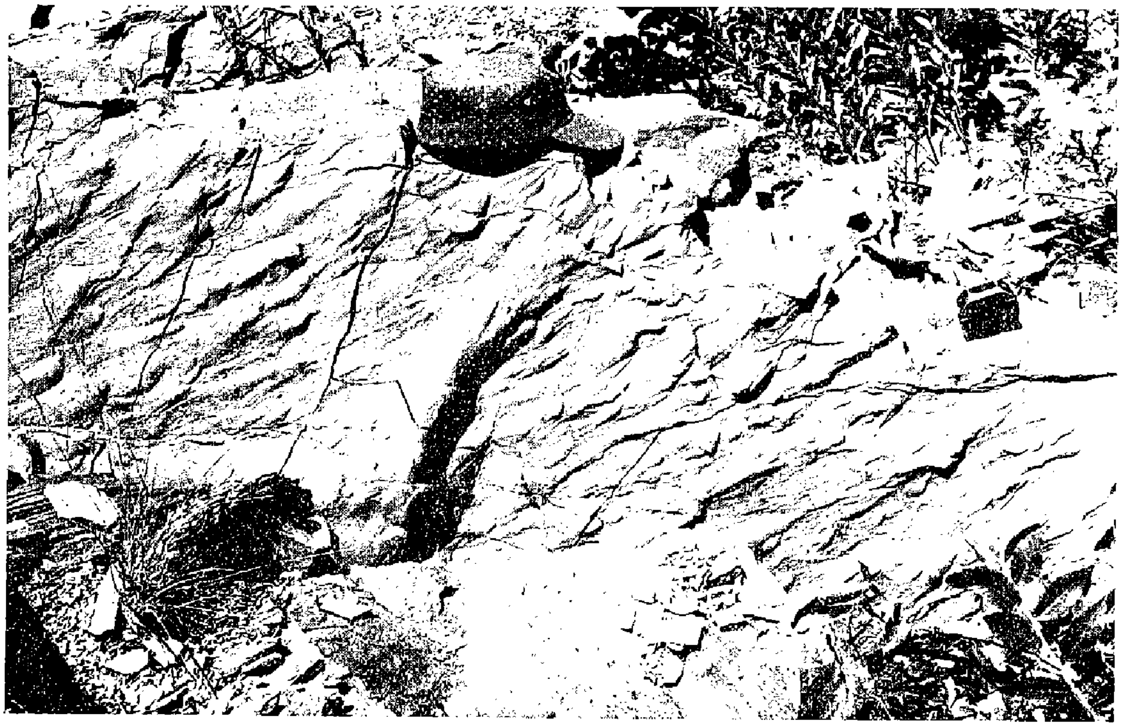


Figure 32. Flute casts on a sandstone in the Johns Valley Shale (?) exposed in the U. S. Highway 270 roadcut near Stapp, Oklahoma. View is toward the northwest; current flow was from upper right to lower left.

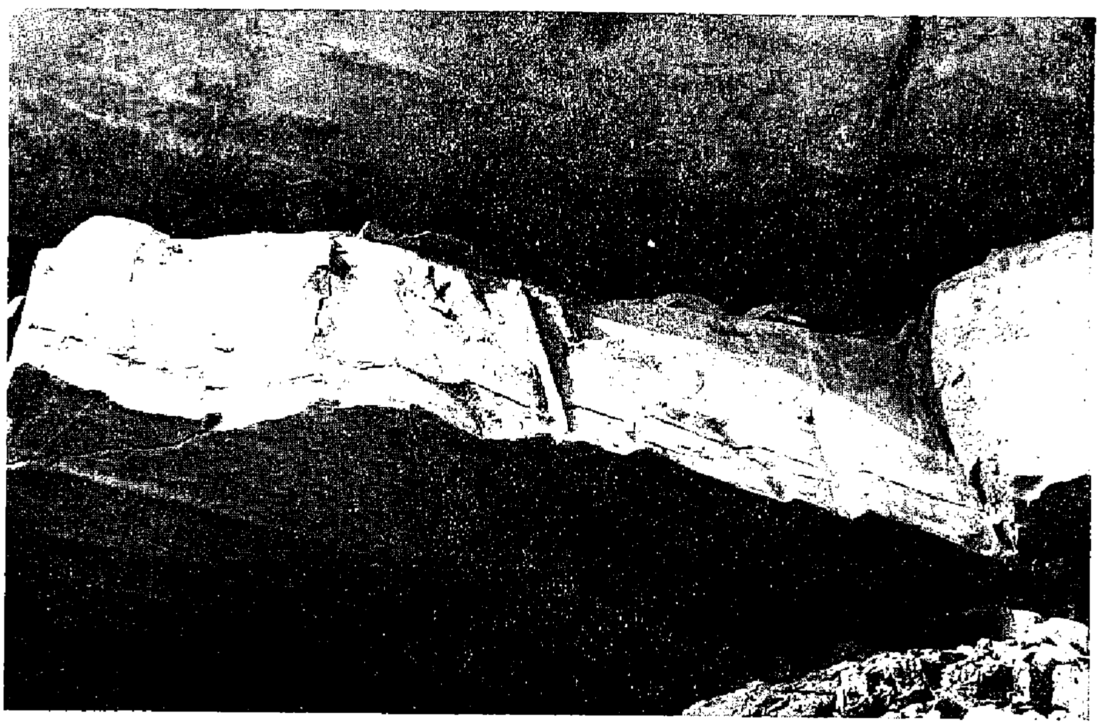


Figure 33. Groove casts on the bottom of sandstones at the base of interval 5, East Ward Lake measured section (base Wildhorse Mountain). View is toward the northwest; cross-bedding in the sandstones indicates that current flow was from right to left.

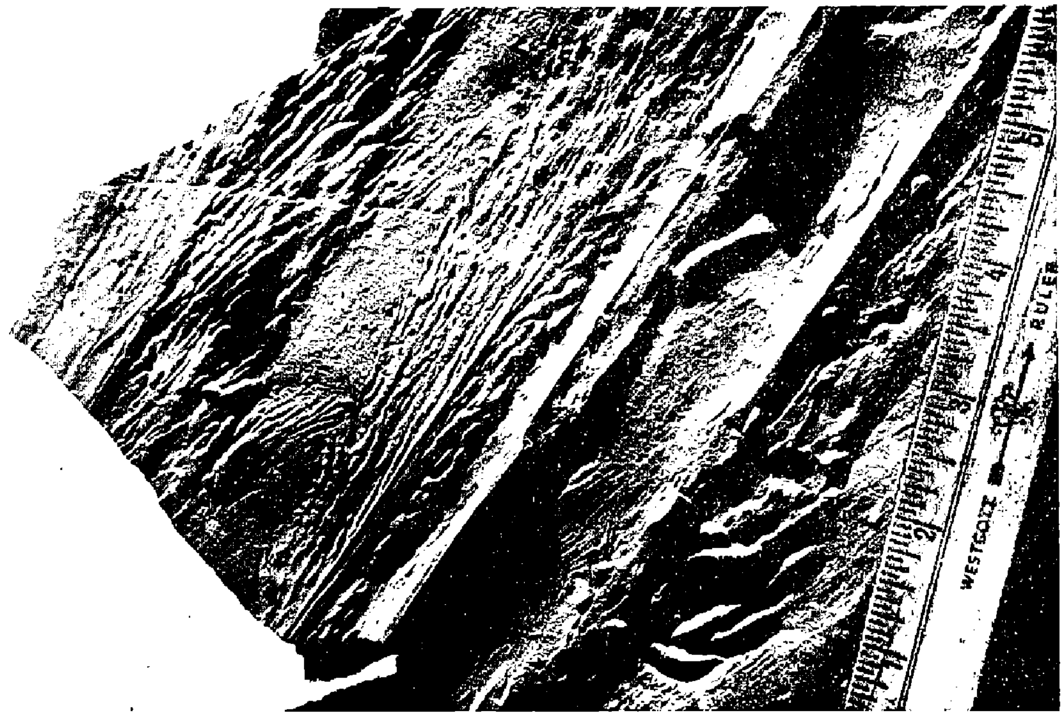


Figure 34. Unusual groove casts on the base of a thin (1½-inch) Atoka sandstone from SE¼ SE¼ sec. 18, T. 1 N., R. 32 W. The large groove cast running diagonally across the middle of the slab was broken in transportation of the sample.

Figure 35. Sinuous, tubular bottom casts on the base of a sandstone in the Johns Valley Shale (?) in the U. S. Highway 270 roadcut near Stapp, Oklahoma. Numerous faults cut the bed.

a small fraction of an inch below the bottom of sandstone beds. A few, however, are quite large (fig. 35).

A rather elaborate classification of bottom-surface markings is possible, and proposed terminologies are making their way into the literature. The broad classification used above has exceptions but serves the ends of the present report adequately.

Not enough observations were made to determine the relationship between the thickness of a bed and the size of its flute and/or groove casts. Generally the larger casts are found at the base of beds more than a foot thick. A prominent exception to this, however, is illustrated in figure 39A-C. This cast protrudes 1¼ inches below

Figure 36. Burrow casts on the bottom of a 1-inch sandstone bed. Sample from interval 41, Rich Mountain measured section (upper Wildhorse Mountain Formation).

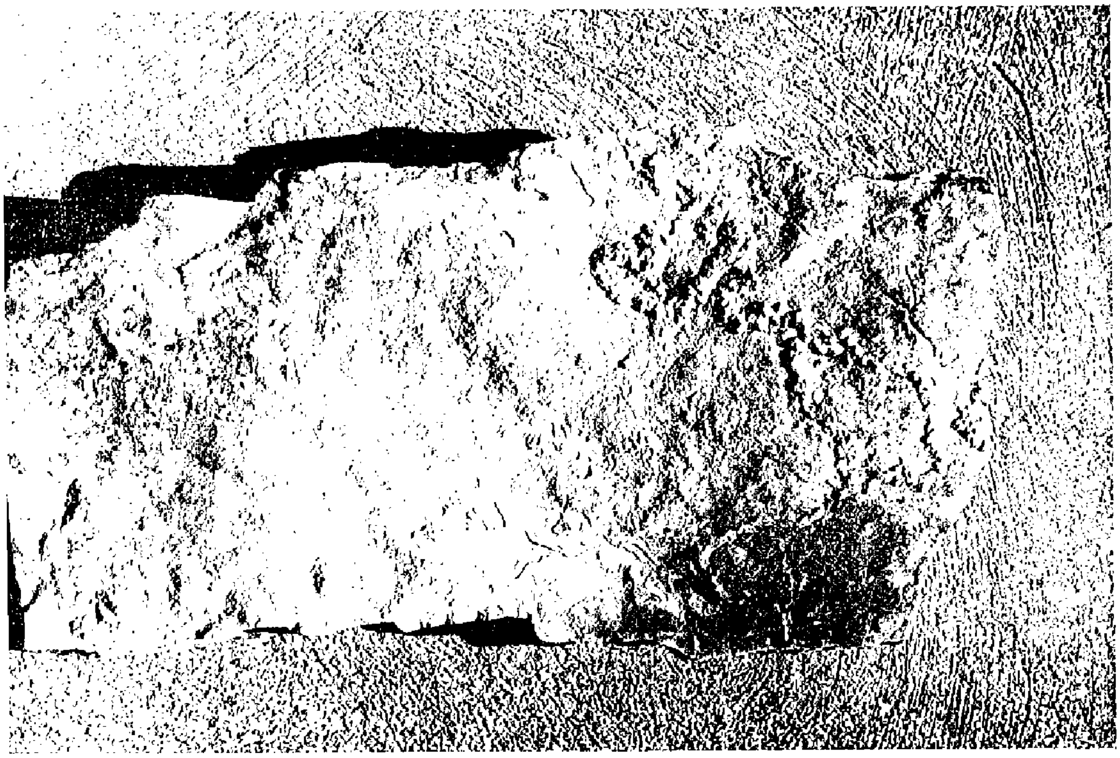
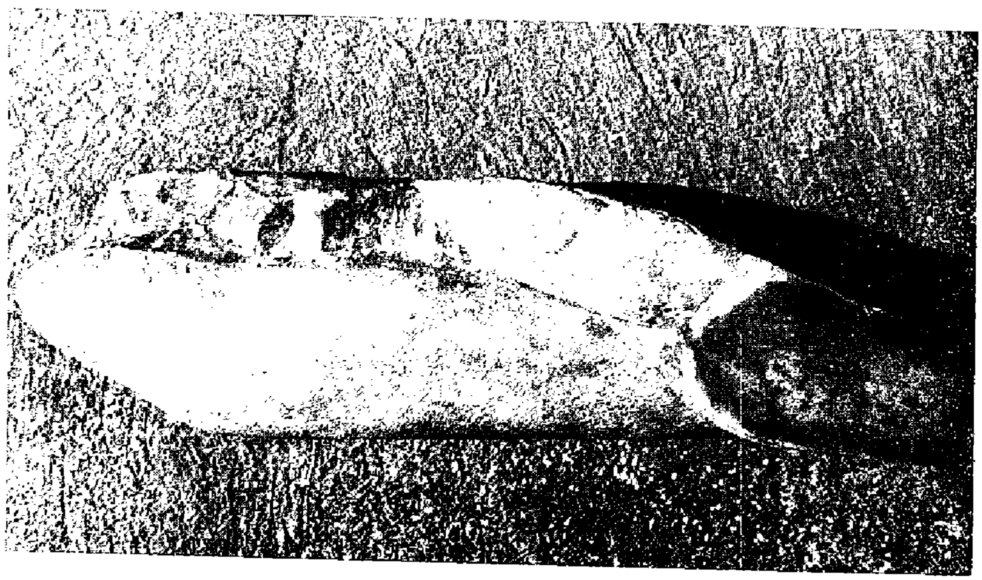


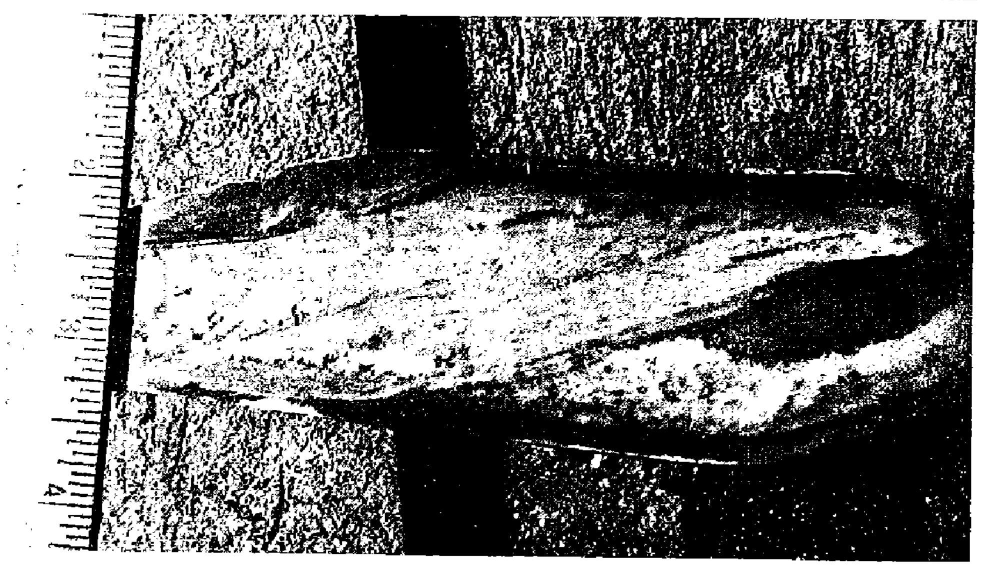
Figure 37. Casts of regularly offset depressions found on the base of a 1½-inch, cross-bedded and ripple-marked sandstone from interval 61, Rich Mountain measured section (Game Refuge Formation). The regularity of offset is best seen in the trail nearest the right edge of the slab.

Figure 38. Nonoriented bottom casts possibly caused by activities of benthonic organisms or by bottom debris removed prior to deposition. Casts are on the bottom of a 1½-inch, cross-bedded and ripple-marked sandstone bed of interval 59, Rich Mountain measured section (Game Refuge Formation).

A



B



C

the base of a ½-inch bed and is especially interesting because of the cross-bedding within it. Topset beds are clearly inclined away from the bulbous end of the cast. A section at right angles to the length of the cast shows the laminae to be concave upward with the outermost laminae parallel to the bottom periphery of the cast. Cross-bedding in the thin bed above the cast is not concordant with that within the cast and indicates a change in current direction during its filling. The depression represented by the cast was filled from the deepest portion (represented by the bulbous end) to the shallowest portion (represented by the flaring end). The bulbous end points upcurrent. The more common association of big casts with beds more than one foot thick is shown in figures 30, 31. Thick beds, however, may have only small flute and/or groove casts.

Thin beds showing abundant current-ripple cross-bedding generally have small bottom-surface markings. These beds are described in the section on ripple marks and cross-bedding, and an example of one showing current casts is given in figure 16. Casts of a thicker bed at the same locality are shown in figure 33.

TOP-SURFACE MARKINGS

The ridges shown on the top surface of a thick sandstone bed in figure 40 were not observed on other strata of the Stanley-Atoka sequence. A few taper uniformly from a large end. The largest of this type is crossed by the tape at its lower end and has an arm-like extension near its bulbous end. Other ridges are subparallel and taper towards both ends.

Channels with pitted bottoms enlarge downdip, reaching widths of more than a foot and depths of two inches at the lower edge of the exposed surface. The surface of the thick sandstone bed on which they occur has profuse plant impressions. Prominent asym-

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Figure 39. A cross-bedded flute cast from the Atoka Formation, north-center sec. 7, T. 2 N., R. 27 E.

A Lower surface of cast. Note presence of smaller casts.

B Upper half of cast shown in A, showing saw-cut surface of the small cross-bedded flute cast appearing at the top of A. The cast makes up the bottom 1 inch of the bed and the remaining part (at the top) is ½ inch thick.

C Saw-cut surface of piece shown in B. The specimen has been stained by weathering, but cross-bedding is clearly visible. The cross-bedding indicates a current flow from right to left, which is the flow direction indicated by the shape of the cast (ends of cast reversed from their position in A and B).

metric ripple marks are present in the same exposure. Assuming that the current direction did not change from the time the ripple marks were formed to the time the channels were eroded, the axes of the channels parallel the current direction and the channels widen and deepen downcurrent.

PALEOCURRENT INDICATORS

Flute and groove casts are considered by many writers as reliable indicators of current direction during the deposition of the sandstone bed of which they are a part. The sectioned flute cast discussed under the heading *Bottom-Surface Markings* has cross-bedding, indicating that the bulbous end pointed upcurrent for the time during which the mold was filled. However, the cross-bedding in the thin bed above the cast suggests a slightly different current direction from that in the cast, a component of flow still being in the direction indicated by the cast. The writer considers the cross-bedding in the cast as corroborating the evidence presented by other writers that the bulbous ends of flute casts point upcurrent.

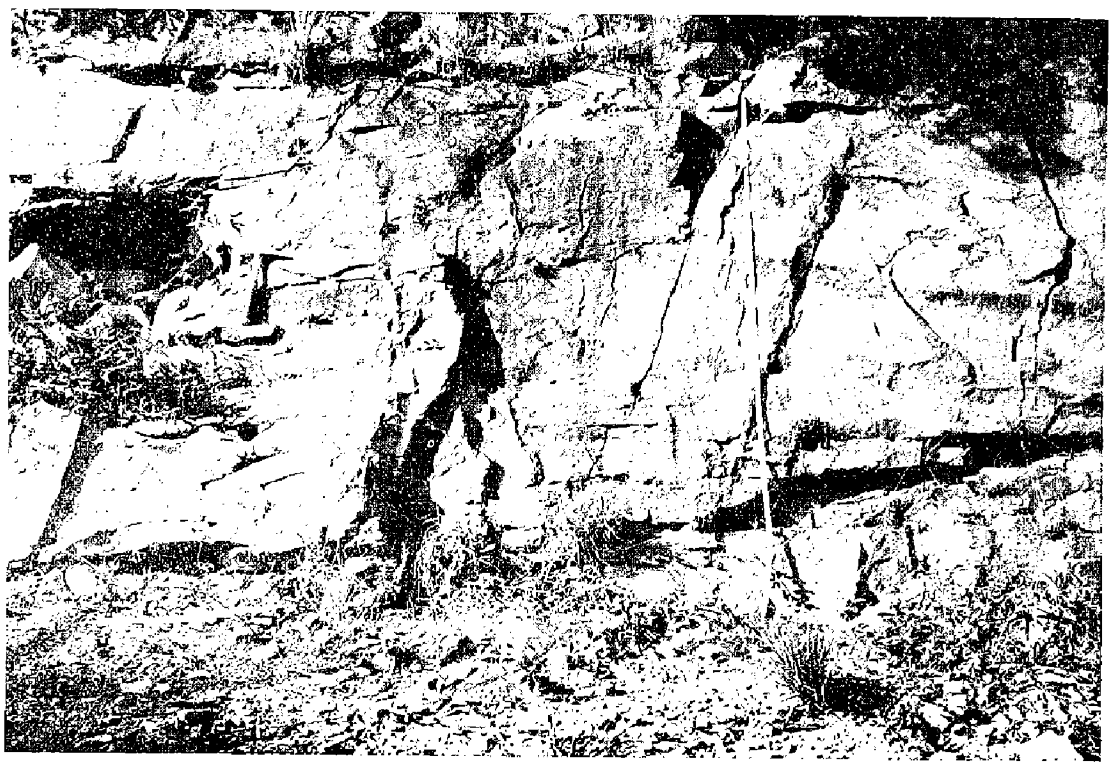

Figure 40. The dip slope of a sandstone in interval 62, Rich Mountain measured section (Game Refuge Formation). The shape of some ridges suggests that they may be casts of animals or plants, but, if so, the matter composing them has not been fossilized.

Groove casts may or may not occur with flute casts. Where the casts occur together, the long axes of groove casts approximately parallel those of the flutes. This fact and additional evidence cited by others suggest that their lengths parallel the current direction. Where groove casts occur in the absence of flute casts and cross-bedding, the current direction cannot be determined.

In the present study, orientations of flute and groove casts were recorded at several localities without attempting to compute a statistical average of the several casts in a single exposure. Several writers have pointed out the presence of variations in cast orientation at a single outcrop, but these variations are small in comparison to regional flow patterns. In the report area, orientations in the outcrops observed did not vary by more than a few degrees and the average was estimated. The data are recorded along with those of other current-direction indicators in table I and plotted on figures 41 and 42.

Ripple marks with associated cross-bedding are also useful indicators of ancient currents. However, exposures are too few for reliable determination of current direction. Figures 41 and 42 show that most current directions determined from ripple marks differ from current directions determined by use of sole markings.

Where cross-bedding occurs in the absence of ripple marks with subparallel linear crests, current-direction determination is dependent upon good exposure and lack of marked contortion. The scoop shape of the forseset beds requires that several readings be taken at a single exposure. Because of these limitations, cross-bedding was not used as a current-direction indicator in the absence of ripple marks or sole markings.

The current directions listed on table I are corrected for rotation of bedding planes about strike. If strike changed during deformation, this correction is indequate.

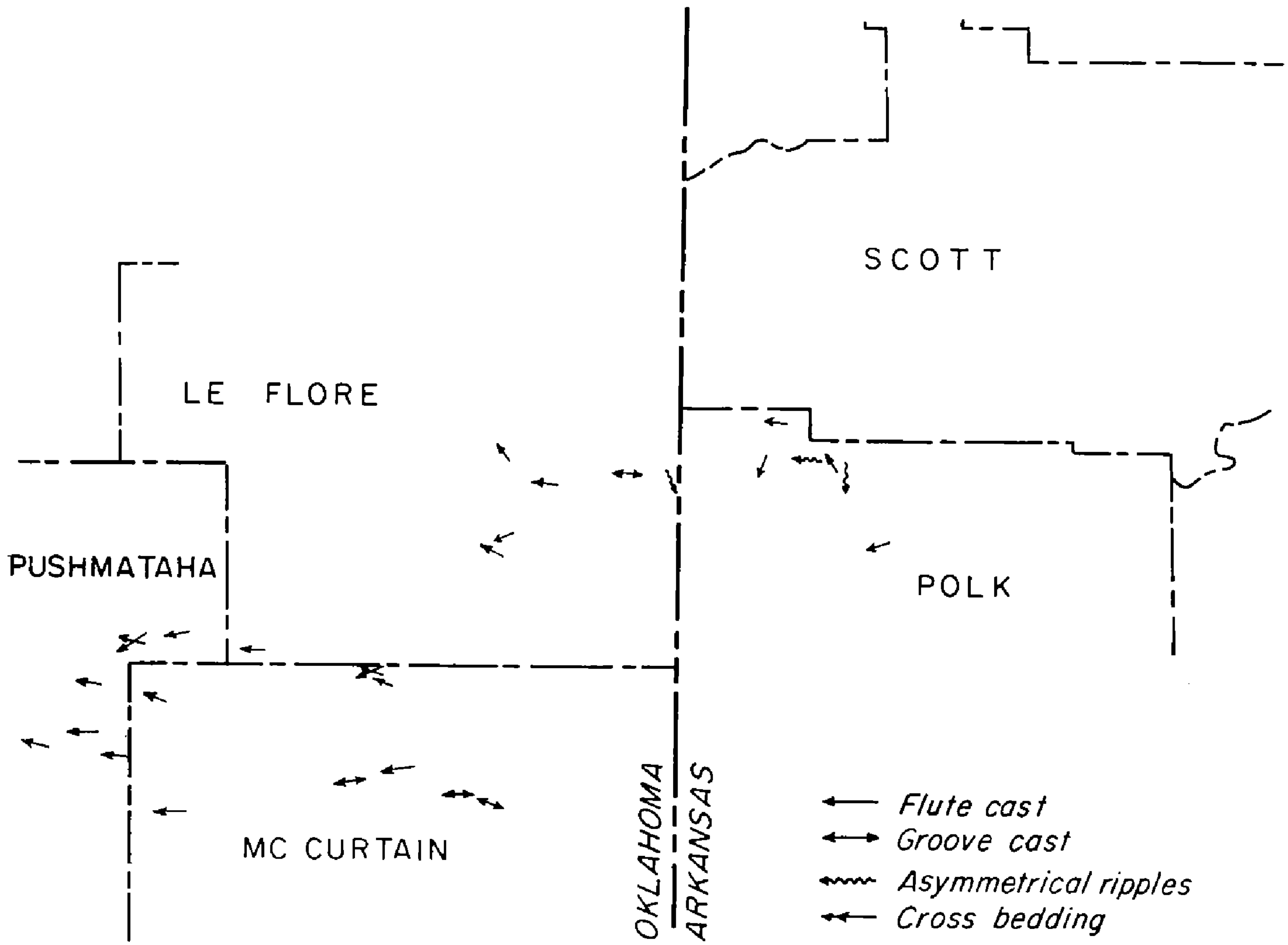

Figure 41. Map of Chickasaw Creek-Jackfork current indicators. Each symbol represents a single observation. Observations are those of the author and of Shelburne (1960).

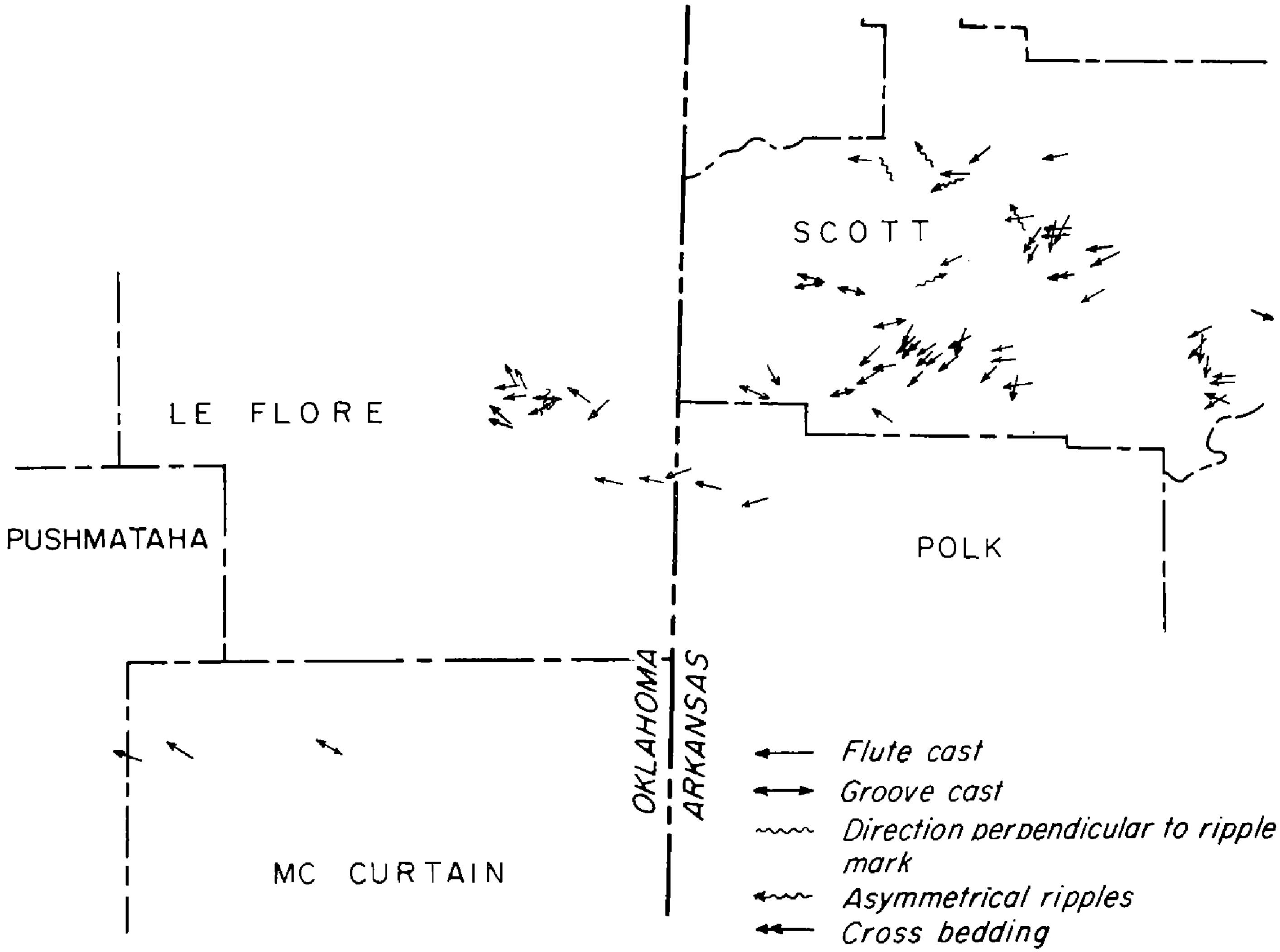

Figure 42. Map of Atoka current indicators. Each symbol represents a single observation. Observations are those of the author, of Shelburne (1960), and of Reinemund and Danilchik (1957).

TABLE I.—PALEOCURRENT DIRECTIONS INFERRED FROM SEDIMENTARY STRUCTURES

Stratigraphic Position	Location	Type of[1] Marking	Bedding Attitude	Corrected[2] Current Direction	Structural[3] Complexity	Reliability[4]
Atoka?	NE NE 12, T 3 N, R 25 E	Flute and groove casts	N57E, 66S (overturned)	S87W	High	Low
Atoka	SE NE 1, T 3 N, R 25 E	Flute casts	N73W, 85S	N30W	Low	Low
Atoka	NW SW 11, T 3 N, R 25 E	Flute casts	N90W, 72N	N60W	High	Low
Atoka	NW 13, T 3 N, R 26 E	Flute casts	N54W, 75S (overturned)	S40W	High	Low
Atoka	SW 6, T 3 N, R 26 E	Flute casts	N70W, 47S	N25W	High	Low
Atoka	SE cor. 13, T 1 N, R 32 W	Flute casts	N71E, 83S (overturned)	S30E	High	Low
Atoka	NW NE 23, T 1 N, R 32 W	Groove casts	N60W, 48S	N75W or S75E	High	Low
Atoka	SW NW 21, T 1 S, R 32 W	Flute casts	N82W, 90N	N82W	Low	High
Atoka	SE NW 18, T 1 S, R 32 W	Flute casts	N87W, 54S	S73W	Low	Low
Atoka	NE cor. 22, T 1 N, R 31 W	Groove casts	N57E, 75N	N87E or S87W	High	Low
Lower Atoka	SE 23, T 3 N, R 25 E	Flute casts	N40W, 90N (approx.)	N50W	High	Low
Lower Atoka	NE SW 24, T 3 N, R 25 E	Groove casts and cross-bedding	N77W, 66S (overturned)	N60W	Low	Moderate
Lower Atoka	NW SE 17, T 3 N, R 26 E	Flute casts		S65W	Low	High
		Ripple marks with no pronounced asymmetry		N10E or S10W	Low	High
Lower Atoka	NE NE 17, T 3 N, R 26 E	Groove casts	N88W, 26N	N88W or S88E	Low	High
Lower Atoka	NW 7, T 2 N, R 27 E	Flute and groove casts	N48E, 20S	N80W	Low	High
Lower Atoka	NE 9, T 2 N, R 27 E	Groove casts	N88E, 41S	N80W	High	Moderate

Lower Atoka	NE 9, T 2 N, R 27 E	Cross-bedding	N78W, 82N	W	High	Moderate
Lower Atoka	NW 25, T 1 S, R 32 W	Flute casts	N76W, 47N	S74W	Moderate	Low
Overlying Stapp Conglomerate	SW SW 7, T 3 N, R 26 E	Flute casts	N40W, 58N	S85W	Low	High
Game Refuge	NW 6, T 2 N, R 27 E	Flute casts	N73E, 42S	E or W	Low	High
Game Refuge	SW 3, T 2 N, R 27 E	Cross-bedding and ripple marks	N86W, 56S	S13E	Low	High
Game Refuge	SW 10, T 1 S, R 31 W	Flute and groove casts	N40W, 50N	N27W	High	Low
Game Refuge	SE NW 14, T 1 S, R 31 W	Cross-bedding and ripple marks with no pronounced asymmetry	N60W, 24S	S	Low	High
Upper Jackfork	SE 26, T 3 N, R 25 E	Flute and groove casts	N50W, 80S	N35W	High	Low
Wildhorse Mountain	SW 5, T 2 N, R 26 E	Flute and groove casts	N85W, 60N	N75W	Moderate	Moderate
Wildhorse Mountain	SE NW 12, T 1 S, R 32 W	Flute casts	N70W, 52S	S30W	Low	High
Wildhorse Mountain	SE NE 8, T 1 S, R 31 W	Asymmetric ripple marks	N20W, 39W	W	Low	High
Base Wildhorse Mountain	SW SW 6, T 2 S, R 30 W	Groove casts and cross-bedding	N88E, 34N	S75W	Low	High
Chickasaw Creek	NE SW 25, T 1 N, R 32 W	Ripple mark, cross-bedding, flute and groove casts	N76W, 48S	W	Low	High
Upper Stanley	SE 3, T 1 S, R 31 W	Ripple mark and cross-bedding	N80E, 28S	S50W	High	Low

[1] Flute casts as here applied are any sole markings of common orientation with a bulbous terminus.

[2] Corrected only for rotation of the bedding plane about the strike as an axis. The angle between the bedding strike and current direction measured in the plane of the bedding may be determined by subtracting bedding strike from corrected current direction.

[3] High structural complexity indicates that local folding or faulting was observed or inferred to be present.

[4] A rating of the corrected current direction as an accurate indicator of what would have been obtained had the beds not been deformed.

PETROGRAPHY

Data obtained from the thin-section study are tabulated in appendices B, C, and D. Supplemental remarks, a cross-reference index, and descriptions of the rocks from which the thin sections were obtained are included in appendices E, F, and G. Photomicrographs of most of the study slides may be seen in figures 43 through 49.

The 23 thin sections described in the tables were selected from a stratigraphic interval of some 9,500 feet, of which perhaps 5,000 feet of rocks is composed of sandstone. The samples were chosen on the basis of freshness of hand specimen, stratigraphic unit represented, or some particularly interesting feature illustrated.

Composition*

QUARTZ

Based upon Folk's empirical classification of quartz types (Folk, 1959, p. 72), the dominant quartz type of all the sandstones sampled would be classified 2d. This class includes single quartz grains which require a stage rotation of one to five degrees for the deepest part of the extinction shadow to sweep from one side of the grain to the other and which contain few vacuoles and no microlites. The second most abundant quartz type falls in class 3d and differs from 2d in having more strongly undulose extinction. Of significance, but quantitatively minor, are grains falling into classes 2a and 6d. The first of these is similar to 2d except that it has abundant vacuoles. Class 6d is composed of composite grains with strongly undulose extinction. Most individuals of the composite grains share crenulate boundaries in the thin sections studied (fig. 47c, H). Of rare occurrence are single grains with rutile needles or microlites which would fall into Folk's classes 2b and 2c, and composite grains with noncrenulate internal boundaries falling into class 5d (fig. 46E.

When lighted by reflected light, quartz grains having many vacuoles are milky and nonvacuolized grains are colorless. Some quartz grains have an illusory weathered appearance that causes

* Compositions of the thin sections are summarized in appendix B.

them superficially to resemble feldspars. One such grain may be seen in figure 46C. The altered portions are linear seams of sericite or a clay mineral of similar appearance.

FELDSPAR

Only in the tuff (thin section 7-20-5F, appendix G) is there adequate plagioclase for an accurate determination of composition; it is andesine (An_{45}). Plagioclase is only sparsely distributed in thin sections of the sandstones present in the measured sections and its composition generally cannot be determined. Thin section 21A has two grains showing combined Carlsbad-albite twinning, each indicating an andesine composition.

The presence of orthoclase is assumed in many of the slides, but no certain identification was made. Most of the possible orthoclase is weathered so that no satisfactory interference figure was obtained from it. Carlsbad twinning was observed on a few grains thought to be orthoclase. Only a few microcline grains were observed in the entire set of thin sections, and these were not identified with certainty. One of these is shown in figure 46A.

CHERT

Chert occurs in several varieties, some of which are gradational and some of which appear distinct. The gradation is evident primarily in the coarseness of texture and may be observed in a single chert grain in some thin sections (fig. 46H). A distinct variety of chert occurs in elongate masses that contain what is apparently carbonaceous material and that have a relatively coarse uniform texture (fig. 46F).

In several slides it could not be determined definitely whether the chert was of terrigenous, allochemical, or orthochemical derivation (Folk, 1959, p. 1). However, most chert is probably orthochemical or allochemical in origin and contains varying amounts of clay. Much of the chert shown in figure 45 appears to be filling pore spaces and is probably an orthochemical constituent. The chert grain shown in figure 46H is of uncertain derivation. The pore-filling appearance of some of the chert may be largely due to en-

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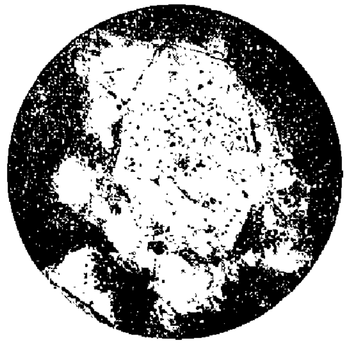

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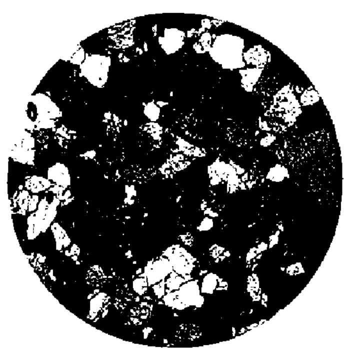

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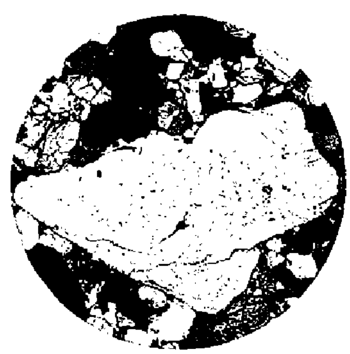

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croachment of quartz into the chert grains (note the relative sharpness of the chert-quartz contact at different points around the periphery of the chert grain of fig. 46H).

As the grain size of the chert becomes smaller, the chert masses become more nearly isotropic. Most, if not all, of the apparently isotropic silica in the thin sections is inferred to be cryptocrystalline.

Finely crystalline chert could not be reliably separated from clay in some thin sections. For this reason chert and clay are tabulated together.

METAMORPHIC ROCK FRAGMENTS

The metamorphic rock fragments seen in thin-section photomicrographs in figures 43 and 44 have a mircocrystalline texture revealed by mica flakes. The fine texture and absence of porphyroblasts suggest that these are slate or phyllite. They represent

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Figure 43. Photomicrographs of rocks in the Rich Mountain measured section.*

A Thin section RM 40, crossed nicols, field diameter is 0.416 mm. Crushed metamorphic rock fragments and crystallized quartz grains, some of which share mutual straight boundaries. The black areas are holes in the section. Atoka Formation.

B Thin section RM 40, crossed nicols, field diameter is 0.208 mm. Secondarily enlarged quartz grain showing the relict outline of the original angular grain and the overgrowth, which has grown in optical continuity with that grain. The grain is the large area in the central part of the photograph. Atoka Formation.

C Thin section RM 40, crossed nicols, field diameter is 0.416 mm. Portion of a metamorphic rock fragment engulfed by secondary quartz. The engulfed portion may be seen in the lower central part of the photomicrograph. Atoka Formation.

D Thin section RM 40, crossed nicols, field diameter is 1.04 mm. A detrital siliceous shale fragment or fine-textured metamorphic rock particle surrounded by recrystallized quartz. The black areas are holes in the section. Atoka Formation.

E Thin section RM 40, crossed nicols, field diameter is 1.04 mm. Glauconite and metamorphic rock fragments in recrystallized quartz. The glauconite may be located by comparison with F. Atoka Formation.

F Thin section RM 40, crossed nicols, field diameter is 0.208 mm. Finely crystalline glauconite forms the central portion of this photomicrograph. Atoka Formation.

G Thin section RM 40, crossed nicols, field diameter is 0.416 mm. A mosaic fabric which has resulted from extensive recrystallization. Atoka Formation.

H Thin section RM 29A, crossed nicols, field diameter is 4.16 mm. Porous sandstone which contains a mold fauna (not visible). Atoka Formation.

I Thin section RM 29A, crossed nicols, field diameter is 4.16 mm. Large quartz grain which is a probable approximate hydraulic equivalent to the invertebrate fragments that have been dissolved from other portions of the rock. H and I have been greatly fractured during preparation of the thin section. Atoka Formation.

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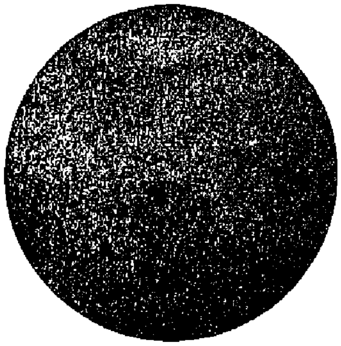
* Intraformational stratigraphic positions are given in appendix E.



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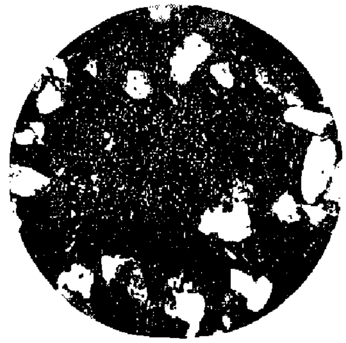

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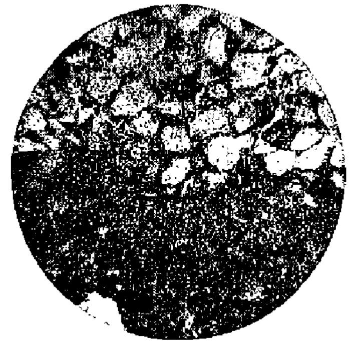

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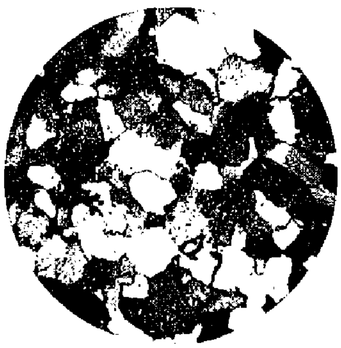

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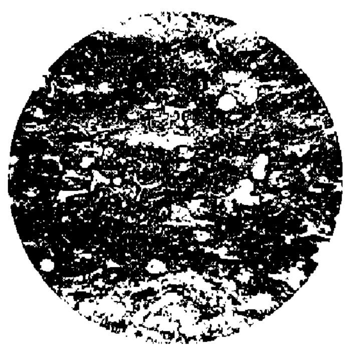

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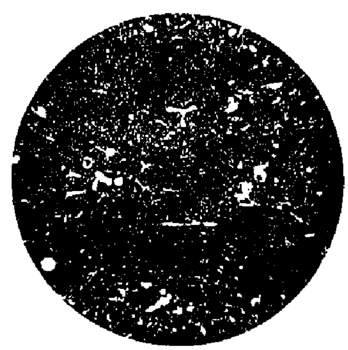

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the most common type of metamorphic rock observed in the thin sections.

According to Folk (1959, p. 70), stretched metamorphic, composite grains of quartz, such as are shown in figure 47C, H, have been derived from low-rank metamorphic rocks in which there has been an absence of recrystallization. These grains are abundant in a few strata but are absent elsewhere in the geologic section. Possibly of similar origin are grains such as those shown in figure 46I and figure 47G. These, however, are more finely microcrystalline than the grains described above.

CHALCEDONY

Chalcedony is not common but was found in thin sections RM 24A, RM 19, RM 20, MR 6-1, and MR 4-2. It occurs as a replacement or in-filling of microfossil molds in these slides.

Figure 44. Photomicrographs of rocks in Rich Mountain measured section.*

A Thin section RM 29, ordinary light, field diameter is 10.4 mm. Cross-bedding and an unconformity on a small scale. The unconformity is at the base of the photomicrograph (see C below) where the sandstone lies upon nearly isotropic siliceous shale (or shaly chert, depending upon the dominant constituent). The dark laminations marking the cross-bedding are made up of fragments of the same siliceous shale. Atoka Formation.

B Thin section RM 29, crossed nicols, field diameter is 1.04 mm. Close-up of a siliceous shale particle in one of the cross laminations. Atoka Formation.

C Thin section RM 29, ordinary light, field diameter is 1.04 mm. A close-up of the unconformity shown in A. Zircon grains are concentrated along the unconformity. The black area at the far right of the unconformity is a black opaque heavy mineral. Atoka Formation.

D Thin section RM 24A, crossed nicols, field diameter is 1.04 mm. Mosaic fabric due to the intergrowth of quartz crystals during recrystallization. Game Refuge Sandstone.

E Thin section RM 24A, crossed nicols, field diameter is 0.416 mm. Mosaic fabric. Game Refuge Sandstone.

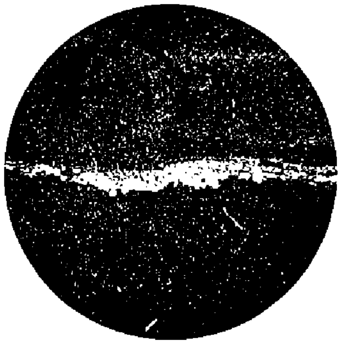
F Thin section RM 24A, ordinary light, field diameter is 0.416 mm. Rounded and euhedral zircon grains that are part of a thin heavy mineral stratum. Game Refuge Sandstone.

G Thin section RM 23A, ordinary light, field diameter is 10.4 mm. Stratification due to varying concentrations of carbonized plant matter in siliceous shale. Game Refuge Sandstone.

H Thin section RM 23A, ordinary light, field diameter is 0.416 mm. Cross-sections of sponge spicules (circular), organic debris, and possible radiolarians (3-rayed) in a stratum containing a relatively high amount of carbonized plant matter. Game Refuge Sandstone.

I Thin section RM 23A. Same as H, crossed nicols.

* Intraformational stratigraphic positions are given in appendix E.



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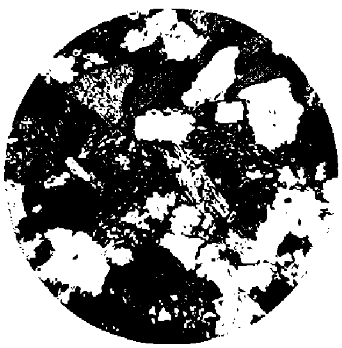

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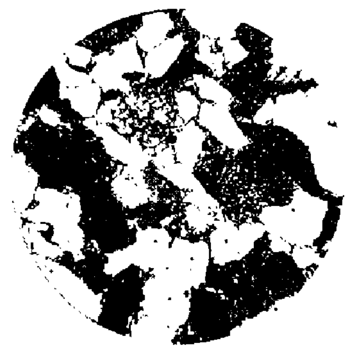

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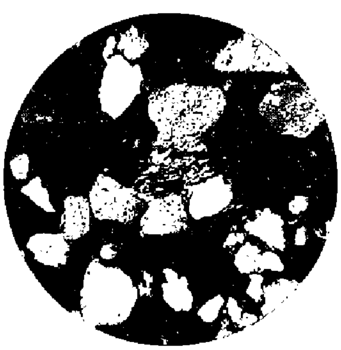

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HEAVY MINERALS

In order of decreasing abundance, the heavy minerals observed are zircon, tourmaline, and garnet. Zircon is far more abundant than are the other two combined, but, even so, no more than a trace of zircon was found in any thin section. The heavy-mineral grains normally are well rounded (fig. 44C, F), but may be angular. Several grains show lesser degrees of rounding and a few are euhedral (fig. 44F). Because of the small number of grains, no attempt to differentiate the varieties of each heavy mineral was made.

Biotite, muscovite, and penninite are the only micas that could be identified confidently. Other micas and clay minerals are grouped with muscovite under the general heading *Colorless Mica* in appendix B. Actually, some of the flakes so grouped have a pale color, such as green or purple. Vermiculite is probably one of the minerals in this group.

Figure 45. Photomicrographs of rocks in Rich Mountain measured section.

A Thin section RM 22A-1, crossed nicols, field diameter is 4.16 mm. A thin, graded, silt stratum in siliceous shale (or shaly chert, depending upon the dominant constituent). Wesley Shale.

B Thin section RM 21A, ordinary light, field diameter is 4.16 mm. Uneven seams of carbonized plant material in one of the discoidal masses in interval 57 of Rich Mountain measured section. Markham Mill-Prairie Mountain undifferentiated.

C Thin section RM 21A, crossed nicols, field diameter is 0.208 mm. Plagioclase, muscovite, and chert surrounded by quartz grains. Markham Mill-Prairie Mountain undifferentiated.

D Thin section RM 21A, crossed nicols, field diameter is 1.04 mm. Cluster of metamorphic rock fragments, plagioclase, and chert in quartz grains. Markham Mill-Prairie Mountain undifferentiated.

E Thin section RM 21A, crossed nicols, field diameter is 0.416 mm. A perthitic feldspar grain may be seen in the center of the field. Markham Mill-Prairie Mountain undifferentiated.

F Thin section RM 20, crossed nicols, field diameter is 1.04 mm. Metamorphic rock fragment, coarsely crystalline chert, and finely crystalline chert separating quartz grains. The black areas are holes in the slide. Markham Mill-Prairie Mountain undifferentiated.

G Thin section RM 20, crossed nicols, field diameter is 1.04 mm. A metamorphic rock fragment may be seen near the center of the field. Markham Mill-Prairie Mountain undifferentiated.

H Thin section RM 20, crossed nicols, field diameter is 0.416 mm. A close-up of the metamorphic rock fragment seen in G. Note that it has been squeezed into inter-grain spaces. Markham Mill-Prairie Mountain undifferentiated.

I Thin section RM 20, crossed nicols, field diameter 0.416 mm. A close-up of a portion of the field shown in F. Markham Mill-Prairie Mountain undifferentiated.

* Intraformational stratigraphic positions are given in appendix E.

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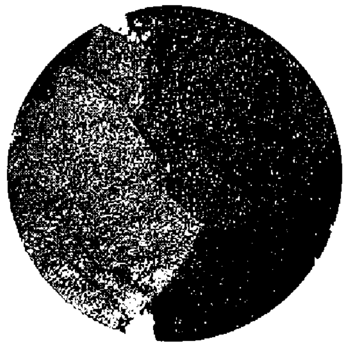

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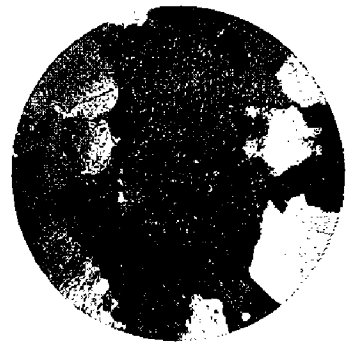

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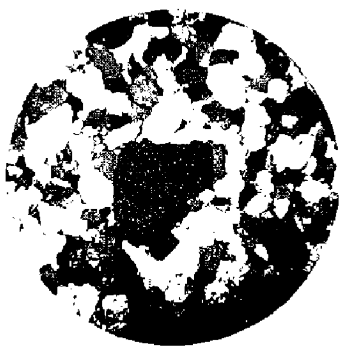

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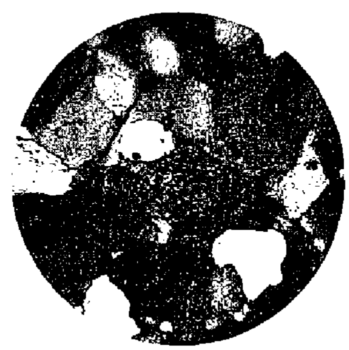

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In order of decreasing abundance, the large mica flakes are muscovite, biotite, and penninite. A muscovite flake is visible in the right center of figure 45c. Lepidomelane, a dark-gray to black variety of biotite, is present in thin sections RM 8-2 and RM 21A.

CALCITE

Calcite is practically absent from the geologic section. Traces are present in slides MR 6-1 and MRS 4-2, but it is a major constituent only in 7-20-5F. In all of these instances it is replacing feldspar and/or quartz, and in 7-20-5F it is also replacing volcanic rock fragments. Calcite occurs in isolated patches in MRS 4-2 and 7-20-5F; however, more even distribution is present in MR 6-1. The association of calcite with pyrite in MR 6-1 and MRS 4-2 should be noted; both minerals are probably secondary in origin.

Figure 46. Photomicrographs of rocks in Rich Mountain measured section.

A Thin section RM 19, crossed nicols, field diameter is 1.04 mm. A tentatively identified microcline grain tightly hemmed in by recrystallized quartz. Markham Mill-Prairie Mountain undifferentiated.

B Thin section RM 19, crossed nicols, field diameter is 1.04 mm. Chert surrounded by quartz grains of a wide size range. Markham Mill-Prairie Mountain undifferentiated.

C Thin section RM 19, crossed nicols, field diameter is 1.04 mm. In the center of the field is a quartz grain which is weathering along linear seams giving a "pseudo feldspar" appearance. Markham Mill-Prairie Mountain undifferentiated.

D Thin section RM 19, crossed nicols, field diameter is 10.4 mm. A large grain of strained quartz with irregular extinction is shown in the center of the field. Markham Mill-Prairie Mountain undifferentiated.

E Thin section RM 14, crossed nicols, field diameter is 1.04 mm. A large angular composite quartz grain surrounded by metamorphic rock fragments and smaller single-crystal quartz grains. Wildhorse Mountain Formation.

F Thin section RM 11A, crossed nicols, field diameter is 0.416 mm. A coarse-textured particle of chert. Wildhorse Mountain Formation.

G Thin section RM 11A, crossed nicols, field diameter is 1.04 mm. Two particles of chert, one finely crystalline and the other coarsely crystalline, may be seen for comparison. Wildhorse Mountain Formation.

H Thin section RM 11A, crossed nicols, field diameter is 0.208 mm. The finely crystalline chert particle shown in G. Note the variation of crystal size within the grain. Wildhorse Mountain Formation.

I Thin section RM 11A, crossed nicols, field diameter is 1.04 mm. A metamorphic rock fragment may be seen in the center of the field. Wildhorse Mountain Formation.

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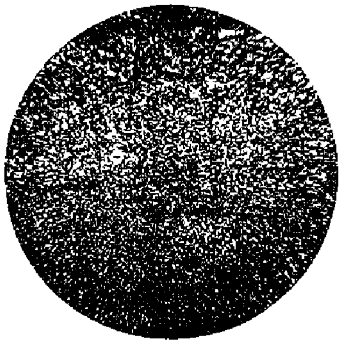
* Intraformational stratigraphic positions are given in appendix E.



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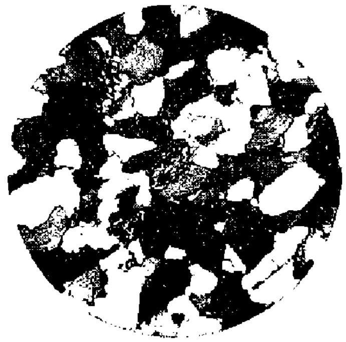

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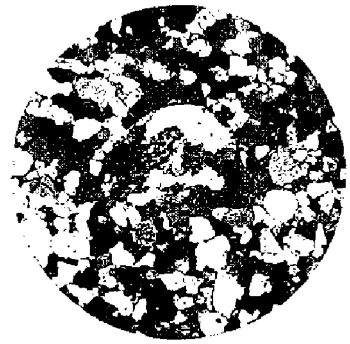

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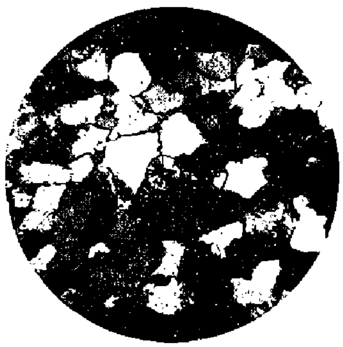

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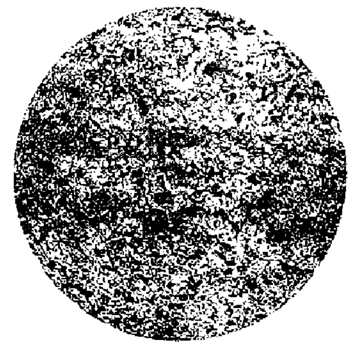

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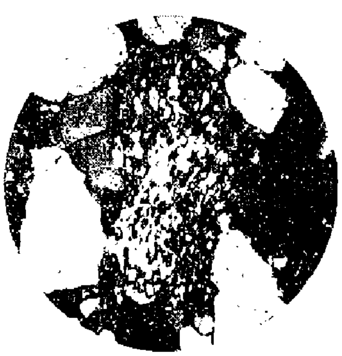

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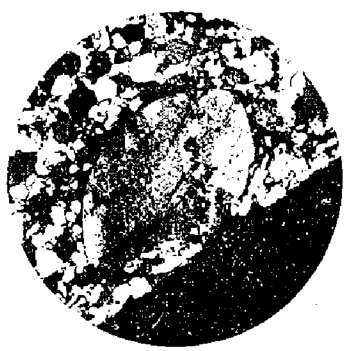

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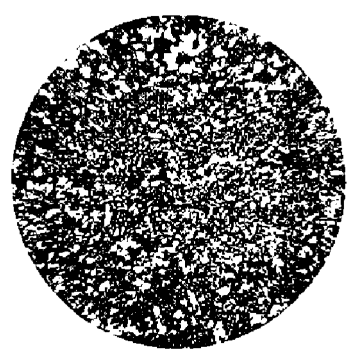

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MICROFOSSILS

Two distinct forms of microfossils are recognizable in RM 23A (fig. 44H, I). One form, probably sponge spicules, is also abundantly present in MR 4-2. Transverse cross sections of them are circular and are 15 to 35 microns in diameter. Longitudinal cross sections reveal the central canal, which appears as a circular darkened area in most transverse cross sections. Longitudinal fragments up to 250 microns long are present, but these are only segments of the entire spicule, which must be longer yet. They possess a maximum of three rays observable in thin section.

The other distinctive microfossil form which is present in RM 23A has three rays in an apparently planar arrangement. The largest fossil has a ray length of 30 microns and a ray diameter that is 10 microns at the bulbous end of the ray but 5 microns elsewhere. This form cannot be reliably identified. Morphologically it resem-

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Figure 47. Photomicrographs of rocks in Rich Mountain and Ward Lake Spillway measured sections.*

A Thin section RM 11, crossed nicols, field diameter is 10.4 mm. Very well-sorted very fine sand overlying very well-sorted coarse silt. Wildhorse Mountain Formation.

B Thin section RM 8-2, crossed nicols, field diameter is 1.04 mm. A metamorphic rock fragment and a chert particle may be seen near the center of the field. Wildhorse Mountain Formation.

C Thin section RM 7, crossed nicols, field diameter is 4.16 mm. A large stretched metamorphic composite quartz grain (see Folk, 1959, p. 69) may be seen near the center of the field. Dark areas within the boundaries of the grain are extinguished portions. Wildhorse Mountain Formation.

D Thin section RM 6-1, crossed nicols, field diameter is 1.04 mm. A replaced crinoid columnal is at the center of the field. Wildhorse Mountain Formation.

E Thin section RM 6-1, crossed nicols, field diameter is 0.208 mm. A close-up of the replaced columnal seen in D. Wildhorse Mountain Formation.

F Thin section RM 2-1, ordinary light, field diameter is 4.16 mm. Stylolites outlined by carbonized plant matter. Wildhorse Mountain Formation.

G Thin section MRS 6-1, crossed nicols, field diameter is 1.04. A fine-textured, stretched metamorphic quartz grain is in the center of the field. Wildhorse Mountain Formation.

H Thin section MRS 6-1, crossed nicols, field diameter is 4.16 mm. A large, stretched metamorphic, composite quartz grain (see Folk, 1959, p. 69) that is well rounded may be seen in the center of the field. The straight border is the edge of the slide. Wildhorse Mountain Formation.

I Thin section MRS 4-2, crossed nicols, field diameter is 10.4 mm. Size stratification parallel to the bottom of the page may be seen in this photomicrograph. Chickasaw Creek Formation.

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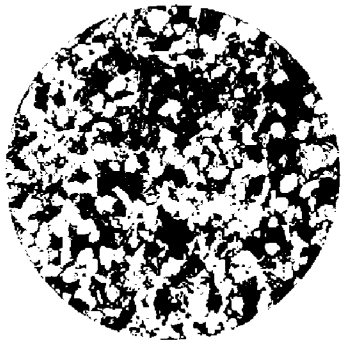
* Intraformational stratigraphic positions are given in appendix E.



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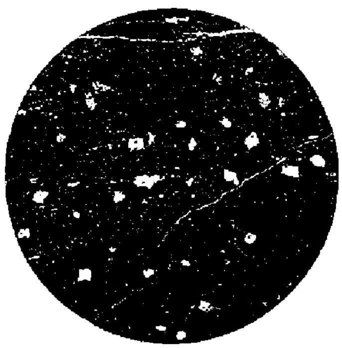

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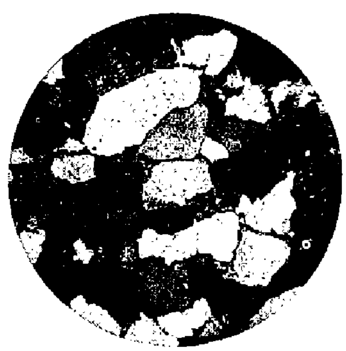


E

bles a radiolarian but it may be a planar three-rayed megasclere from the sponge species that yielded the spicules described above or from another sponge.

In RM 23A are fragments of several other microfossils in addition to the two forms described above. Some of these resemble parts of bryozoans; others are similar in size to the three-rayed form, but they have up to four rays.

Fragmentary siliceous organic debris that is too small for identification and fusiform silica deposits, such as that pictured in figure 48D, are listed as unidentifiable microfossils.

OPAQUE COMPONENTS

Carbonized plant material, the prevalent opaque component observed in the study slides, ranges widely in concentration. Clean sandstones have practically no carbonaceous material; sandstones with a siliceous shale or chert matrix may have abundant large fragments and/or macerated matter.

Authigenic pyrite is present in MR 6-1 and MRS 4-2 (figs. 47, 48). Both slides were obtained from sandstone in a sequence dominantly composed of shale and contain a relatively high proportion of matrix. As pointed out earlier, these are the only slides containing calcite other than the tuff (7-20-5F).

No attempt was made to distinguish magnetite from ilmenite.

Figure 48. Photomicrographs of rocks in East Ward Lake measured section.*

A Thin section MR 6-1, crossed nicols, field diameter is 4.16 mm. Square-edged pyrite silhouettes are present in the central portion of the photomicrograph. Siliceous shale and chert make up an unusually high proportion of this rock. Wildhorse Mountain Formation.

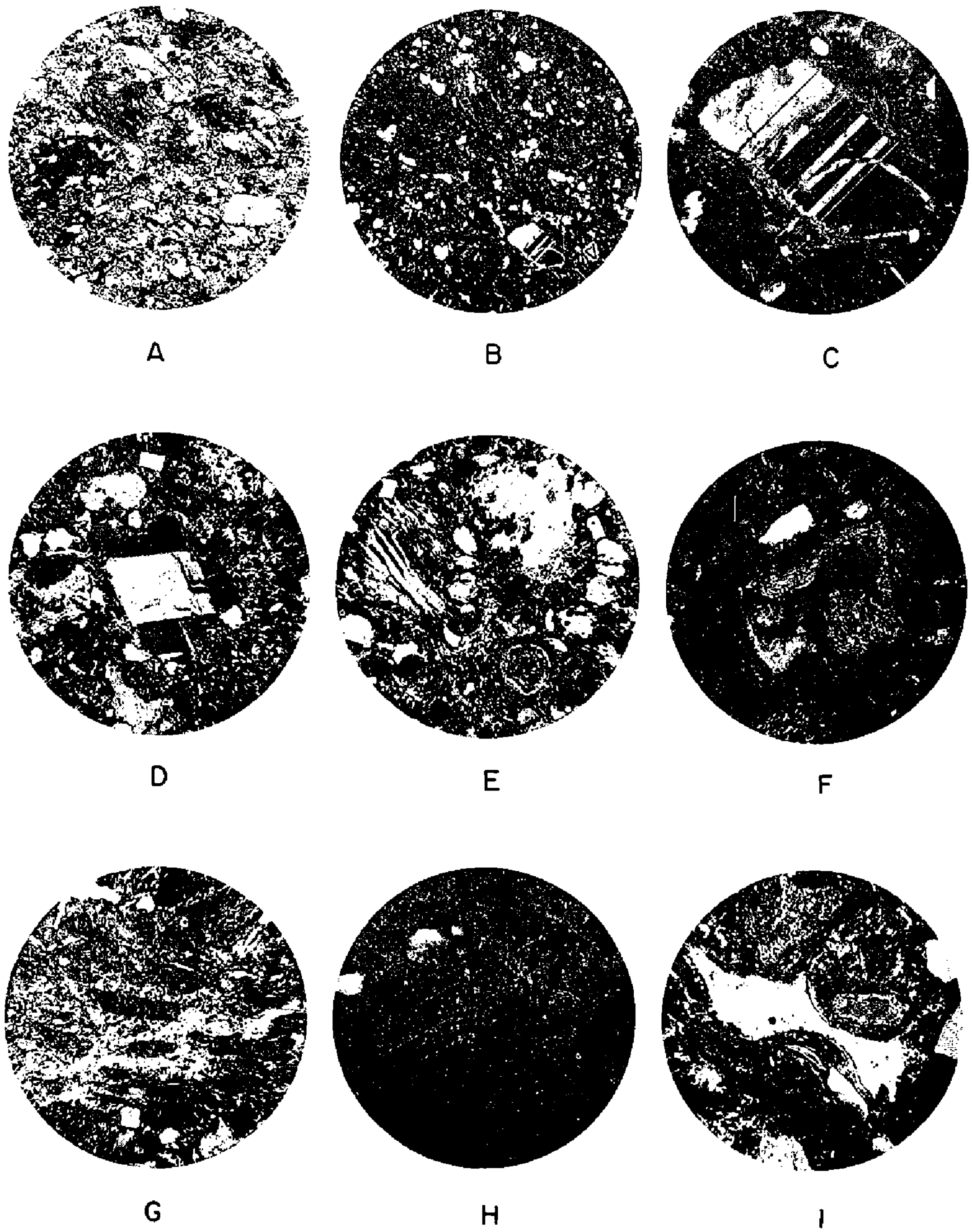
B Thin section MR 4-2, ordinary light, field diameter is 10.4 mm. Siliceous shale faintly laminated by varying proportions of carbonized plant matter. The white spots are holes in the section. Chickasaw Creek Formation.

C Thin section MR 4-2, ordinary light, field diameter is 4.16 mm. A close-up of the white spots shown in B. The sharp-cornered rhombic shape of several of the holes suggests that some of them were originally filled by authigenic mineral grains, by pyroclastics, or by angular detrital debris. Chickasaw Creek Formation.

D Thin section MR 4-2, ordinary light, field diameter is 1.04 mm. The particle in the center of the field is probably of organic origin and vestiges of the original structure may be seen. Chickasaw Creek Formation.

E Thin section MR 3-1, crossed nicols, field diameter is 1.04 mm. Quartz grain in center is being replaced by a clay mineral along its borders. An elongate metamorphic rock fragment may be seen to the upper left. Stanley shale.

* Intraformational stratigraphic positions are given in appendix E.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One (or both) of these is present in MRS 4-2, MRS 6-1, and RM 21A. In describing the thin sections, any white opaque particle was labelled leucoxene. Its association with other heavy minerals in several slides makes this assignment seem reasonable; however, in most slides there is no such association, and the classification of all white opaque masses as leucoxene seems less justifiable.

Not included in appendix B is limonite. This substance is a minor constituent of many thin sections. Limonite in RM 29 appears to have been present when the secondary silicification of the rock took place. Where it occurs in the other thin sections, it is a product of weathering.

Figure 49. Photomicrographs of tuff.*

A Thin section 7-20-5F, ordinary light, field diameter is 4.16 mm. Subangular andesine, volcanic rock fragments (the one in left center is being replaced by calcite), and quartz (there is embayed quartz adjacent to the large andesine grain in the lower right) embedded in a finely microcrystalline and cryptocrystalline matrix of quartz and clay mineral(s).

B Thin section 7-20-5F. Same view as A, crossed nicols.

C Thin section 7-20-5F, crossed nicols, field diameter is 1.04 mm. A close-up of the large andesine grain seen in the lower right of A and B. Replacement by calcite may be seen around the margins of the grain. The ribbonlike feature curving across the grain is Canada balsam.

D Thin section 7-20-5F, crossed nicols, field diameter is 1.04 mm. A very angular particle of plagioclase showing Carlsbad twinning. It, also, is being replaced by calcite. Calcite patches may be seen in the left and lower center.

E Thin section 7-20-5F, crossed nicols, field diameter is 1.04 mm. The mold of what was probably a crinoid columnal may be seen in the lower right center. The grain with a frayed appearance in the left center is probably a volcanic rock fragment that has been altered to clay. At its lower end is a small teardrop-shaped quartz grain that may have originally been a glass shard. In the upper right is a calcite mass that may be replacing a volcanic rock fragment, the dark spots in the mass representing vesicles.

F Thin section 7-20-5F, crossed nicols, field diameter is 0.416 mm. Calcite replacement of this feldspar (?) grain has progressed further than it has in other grains.

G Thin section 7-20-5F, crossed nicols, field diameter is 1.04 mm. Cellular structure of what was probably a pumice fragment. The fragment has been altered to clay.

H Thin section 7-20-5F, crossed nicols, field diameter is 0.416 mm. Calcite replacing andesine.

I Thin section 7-20-5F, crossed nicols, field diameter is 0.416 mm. A devitrified shard of volcanic glass.

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* This thin section is from the Chickasaw Creek Formation or from its approximate position, depending upon how the Chickasaw Creek is defined. The rock from which the thin section was made was collected in NE¼ SW¼ sec. 25, T. 1 N., R. 32 W.

Texture

SANDSTONES

Grain-size analysis.—The method used to obtain the statistical parameters of grain size is that described by Folk (1959, p. 44). The graphic mean was determined by the method used by Inman (cited by Folk, 1959, p. 44). This method has the weakness of assuming a symmetrical size distribution. Sorting is represented by the graphic standard deviation as defined by Folk.

Two factors of importance affect the validity of the mean and the standard deviation as thus determined: (1) both parameters are dependent upon accurate estimation of the sizes of the 16th and 84th area percentiles so that any error in this estimation will make the parameters incorrect, and (2) the effect of thin sectioning on size parameters must be estimated.

The analysis was made by viewing slides through a petrographic microscope, the parameters thus obtained being corrected to their sieve equivalents by the method developed by Friedman (1958, p. 408, fig. 4). Both the original thin-section data and the sieving equivalents are given in appendix C. As may be noted in the appendix, the effect of the corrections is to lower slightly the mean grain size and to increase the sorting by lowering the standard deviation. Except for the coarser grain sizes, the correction is quite small and may be ignored in most instances.

Recrystallization of many of the quartz grains has proceeded to the point that original grain outlines are not visible. The sizes of the 16th and 84th percentiles were determined by using the present enlarged grain boundaries. However, in areas of the thin sections containing no chert cement or clay matrix the quartz grains were probably tightly packed originally, and overgrowth has been relatively small as compared with original long diameters. Where chert cement or clay matrix separates grains, little modification of grain shape has occurred.

The mean size in most thin sections falls within the range of fine sand. Second in frequency of occurrence are thin sections with means in the very fine-sand range. The few thin sections, or layers within thin sections, that have means falling outside of these two size ranges are no coarser than coarse sand and no finer than fine silt. The average of the thin-section means uncorrected to sieve equivalents is 2.67 phi, a value which falls within the lower fine-sand-size range.

The standard deviation does not exceed 0.5 phi in 16 of the 20 slides listed in appendix C. Five-tenths phi is the boundary between moderately sorted and well-sorted sediments, according to Folk (1959, p. 103). It also divides submature and immature sediments from mature and supermature sediments. Based upon this terminology, the 16 slides fall into the better sorted and more mature categories. The remaining four thin sections have a standard deviation less than 1.0 phi, which classes them all as moderately sorted.

Grain roundness.—Extensive recrystallization in the sandstones has effectively masked the original roundness of most quartz grains. Relatively few grains show relict outlines. On the other hand, where a chert cement or clay matrix is present between grains, original outlines appear to be only slightly modified.

The difficulty in estimating roundness prohibited the distinction between mature and supermature (appendix D) for most of the orthoquartzites. A roundness of 3.0 (Folk, 1959, p. 101) separates these two classes of sediment. Generally where it was possible to estimate the original roundness of quartz grains, the roundness was close to this borderline value. Feldspar grains have a size near the thin-section mean and roundness that differs little from that of the quartz.

Composite, stretched metamorphic quartz granules are well rounded (fig. 47C,H). These are the only well-rounded, granule-size particles visible in the thin sections. Most other unusually large grains are composed of quartz and have a roundness similar to the sediment average (figs. 43I, 46D,E). Fine-textured, stretched metamorphic quartz grains are few. The grain shown in figure 47G is rounded.

Most metamorphic rock fragments are predominantly composed of mica and consequently are soft. Their morphology has been so affected by diagenetic processes that their roundness in most thin sections is of little significance. However, the few less-deformed grains show practically no modification by abrasion.

The roundness of tourmaline and zircon has an extreme range, from well rounded to euhedral (fig. 44F).

Fabric.—Stratification is revealed in three ways in the sandstone thin sections: layered variation in quartz grain size, laminae of heavy minerals, and laminae of carbonaceous matter, mica, and clay. The effect of quartz grain-size variation in thin sections MRS 4-2, MRS 6-1, and RM 11 may be seen in the several mean and sorting values shown in appendix C. Examples of this as it exists in RM 11 and MRS 4-2 are observable in figure 47A, I . The other thin sections show little or no size lamination.

Stratification by heavy minerals occurs in RM 8-2 and RM 20. In RM 20 one stratum is slightly inclined and marks small-scale cross-bedding. Perhaps related in significance is the concentration of zircon grains above the unconformity at the base of RM 29 (fig. 44c.)

Recrystallization has produced an intricate intergrowth of quartz grains where no chert cement or clay matrix is present. This mosaic fabric characterizes most sandstone thin sections. The mutual boundaries between grains are straight locally, but most of their peripheries are uneven.

In portions of several slides a chert-cement or clay matrix is present and may compeltely surround grains so that they appear to float in it. As the chert-clay matrix decreases in abundance, it surrounds progressively smaller portions of grain peripheries. In several slides it occurs only in isolated patches. Where only the patches are present in a thin section, it is difficult to differentiate them from allochemical or terrigenous constituents. Most of the patches appear to be intergrain fillings having no shape of their own and, for this reason, are considered to be of orthochemical or allochemical origin.

Quartz grains, having a prominent long dimension, and muscovite flakes show alignment with bedding. Although not quantitatively determined, the degree of alignment appears to be higher in the better sorted rocks which have little or no matrix.

Most holes present in the thin sections are probably the result of weathering or of breakage during preparation of the slide. The smallest holes in the thin sections are apparently the result of incomplete filling of pores by quartz overgrowths (fig. 43G). Unfilled pores appear to make up a relatively small proportion of the rock so that porosity is low.

SILICEOUS SHALES

Fabric.—Thin sections MR 4-2, RM 22A-1, and RM 23A fall in the general category of siliceous shale. MR 4-2 and RM 23A are from well-bedded sequences of similar lithology, whereas RM 22A-1 is from a siliceous nodule or concretion within a soft shale section. The modes of occurrence cause MR 4-2 and RM 23A to have similar textures which differ markedly from RM 22A-1. The first two are well laminated with carbonaceous material, whereas lamination in the third is practically absent and bedding is made evident only by a thin silt stratum and slight variations in the concentrations of organic matter. Most laminations in MR 4-2 are between 0.05 mm and 1 mm thick. Laminations in RM 23A are about 0.1 mm thick.

ENVIRONMENT OF DEPOSITION

The environment of deposition of the Upper Paleozoic rocks of the Ouachita Mountains has been a controversial subject for several years. Although many viewpoints have been expressed, most may be roughly classified into two main categories: (1) below-wave-base deposition by turbidity currents and (2) shallow-water deposition.

Honess (1923, p. 196-202) analyzed the depositional environment of the Stanley Shale and considered the Stanley and Jackfork as having been derived from the same source area (p. 196). He was one of the early advocates of a shallow-water depositional environment for the Stanley, believing that "the ripple marks, rill marks, and cross-bedding throughout the succession" indicate the existence of mud flats at the mouth of a large river where the Stanley was deposited.

Bokman (1953, p. 161) considered that the bedding-plane features observable on many Jackfork sandstones were formed in the littoral zone as rill or drag markings. For the bedding-plane features of other beds, he stated, "Local and ephemeral scouring action of turbidity currents may have been partly responsible . . . . the Stanley-Jackfork sequence represents an over-all trend from deep- to shallow-water deposition."

More recently Scull, Glover, and Planalp (1959, p. 168) postulated that the Atoka Formation of the frontal Ouachitas and Arkoma basin is the deposit of a westward-building delta. Among evidences cited for this theory are the eastward increase in the sand-shale ratio and the local presence of coal in Arkansas.

Cline (1960, p. 87-100) compared the Stanley-Atoka sequence with the black-shale flysch facies. He concluded (p. 100) ". . . the environment was that of a deep trough in which dark muds were the prevailing or most characteristic type of sediment and into which sands were introduced from time to time."

Stanley-Atoka Sediments and Turbidites

During the past 20 years Philip H. Kuenen of the University of Gröningen has been a leading proponent of the turbidity-current concept. He and his students have described sedimentary sequences at widely separated localities and identified in them strata deposited by turbidity currents. One of his students, E. ten Haaf, summarized what he considered to be diagnostic characteristics of turbidites (ten Haaf, 1959b). The properties he enumerated follow:

(1) Graded bedding, combined with poor sorting
(2) Continuous, parallel stratification
(3) Interstratification with nonturbidites
(4) Oriented sole markings mainly restricted to soles of medium-grain-size beds
(5) Aligned inclusions and, in coarse beds, aligned grains
(6) Unidirectional cross and ripple lamination
(7) Burrows and reworked fossils only; fossils generally scarce
(8) Hydroplastic load casting and convolution
(9) Constant current trend
(10) Slump structures
(11) Absence of shallow-water phenomena

For purposes of comparison, properties of beds in the report area are discussed in following paragraphs.

(1) Graded bedding, combined with poor sorting.—Graded bedding is not a conspicuous megascopic feature of Jackfork, Johns Valley, or lower Atoka strata. However, the presence of a coarse, commonly fossiliferous zone at the base of several sandstones, particularly in the lower part of the Atoka, may be thought of as a type of grading. A fossil breccia at the base of turbidite beds is referred to by ten Haaf (1959b, p. 219) and the implication is that these beds are graded.

The tops of many strata have closely spaced laminae of clay, mica, and plant material that are sparse or absent at lower levels of the same bed. This finer grained sediment at the tops of beds also may be thought of as representing a type of grading. However, if the layers between laminae are of uniform grain size throughout the bed (this was not obvious on megascopic inspection and needs to be checked microscopically), the beds may be described as bimodal—the modes indicating the contrast in grain size between the fine laminae and the interlayers.

The sorting of twenty sandstone thin sections (appendix C) at different levels of the Jackfork-lower Atoka sequence in the report area is good; most sandstones are well or very well sorted and a paste matrix is either absent or sparse. However, this sample may be biased because most thin sections were obtained from thin beds (appendix F). More poorly sorted sandstones occur in the Jackfork elsewhere in the Ouachitas (Shelburne, personal communication; Goldstein, 1959, p. 105). Moretti studied 140 thin sections of Jackfork sandstones south and west of the report area and found that most are fine grained. Sorting in them ranges from 1.28 to 1.72, with sorting values being less than 1.5 (Moretti, 1958, p. 26). These values are uncorrected for thin-section bias and were determined by using Trask's definition: sorting equals the square root of the quotient derived by division of the third quartile by the first quartile. If one assumes a normal size distribution, these values may be converted to standard-deviation values for the same curve so that a comparison with the writer's results may be obtained. The converted standard-deviation values are 0.52, 1.15, and 0.86 phi, respectively. Based upon Folk's classification (1959, p. 103), nearly all of Moretti's thin sections are moderately sorted.

Bokman (1953, p. 163) concluded that the average arithmetic mean of Stanley graywackes is 3.6 phi (very fine sand), whereas that of the Jackfork is 2.8 phi (fine sand). The average standard deviation (sorting) is 0.85 phi for both the Stanley and the Jackfork. The above figures do not include the paste matrix. If the matrix is included, the mean is about 4.8 phi and the standard deviation is 1.7 phi for both groups.

Early in his investigations, Kuenen (1953, p. 8) indicated that the sorting values of typical turbidites are greater than 1.5, using Trask's coefficient of sorting; this value corresponds to a standard deviation greater than 0.86 phi. Pettijohn (1957, p. 171) listed the standard deviations for different levels of a turbidite bed of the Ventura River, and these exceed 0.87 phi. Ericson and others (1961,

p. 216, 217) listed the grain-size values of 84 graded beds, most of the medians of which fall in the fine- and very fine-sand-size ranges. Eleven beds have a standard deviation equal to or less than 0.50 phi (computed from quartile deviation and assuming normal distribution) and are well sorted. Fifty-six of the remaining beds are moderately sorted and 17 are poorly sorted (sorting terminology from Folk, 1959, p. 103).

Compared with these values, the sorting of Jackfork sandstones is better than that of the typical turbidite of Kuenen and Pettijohn but is similar to that of several graded beds described by Ericson and others. Several sandstone beds of the lower part of the Atoka are also well sorted, but some are argillaceous and probably more poorly sorted.

(2) *Continuous, parallel stratification.*—Sandstone zones of the Jackfork are continuous for many miles along strike, but it is difficult to determine the extent of individual beds. The continuity of sandstones is clearer in an east-west (or southwest-northeast in the westernmost Ouachitas) direction simply because this is the dominant strike. Continuity in a north-south direction is unknown. A typical exposure of Jackfork strata shows parallel stratification of all units. This is also true of beds in the lower part of the Atoka in the report area. Except where there has been plastic flow or local deformation, even the thinner sandstones appear continuous. Sandstones of the Moyers Formation have a continuity similar to that of sandstones of the Jackfork. Units of the Tenmile Creek Formation are of unknown continuity because of the widespread structural complexity of this formation.

On aerial photographs the sandstone zones of the Wildhorse Mountain Formation are traceable for several miles and a distinctive shale zone is in the lower part of the Wildhorse Mountain (see intervals 8 through 11 of Rich Mountain measured section), which may be traced the full length of both Rich and Black Fork Mountains. Some of the individual sandstone beds do lose their identity, however, within the lengths of the two mountains.

Although the sandstone zones of Rich Mountain appear relatively continuous along strike, a zone-to-zone correlation across the strike with beds of Black Fork Mountain could not be accomplished for most zones. However, the similar topographic characteristics of the two mountains suggest little difference in their over-all stratigraphy.

In the upper part of the undifferentiated Prairie Mountain-Markham Mill-Wesley sequence of the Rich Mountain syncline, sandstone ridges are locally present along strike. The limited extent of these ridges suggests a lateral discontinuity of units at that stratigraphic level, a level predominantly composed of shale in the exposures of the Rich Mountain measured section. The Jackfork strata in State Highway 103 exposures in the northwestern part of the area have a high proportion of sandstone making up the section below the Johns Valley Shale and, in this respect, differ markedly from the Jackfork of Rich or Black Fork Mountains. The beds along State Highway 103 are here considered to have been laid down relatively farther north at a different depositional site than were the Rich and Black Fork Mountain beds, having acquired their present geographic position through deformation.

As stated by ten Haaf, regular stratification is a distinctive feature if it is found in coarsely clastic beds, which commonly tend to have discordant and discontinuous stratification. The quartz- and fossil-granule conglomerates of the upper part of the Jackfork and lower part of the Atoka are either massive or planar laminated parallel to bedding planes. Beds containing conglomerates maintain a uniform thickness within the limits of a single exposure, but individual beds probably do not extend over large areas. Zones, however, are widespread.

The topographic position and expression of the Chickasaw Creek and associated beds are similar from the western margin of the Ouachitas eastward to, at least, the eastern edge of the area, a distance of about 80 miles. The Chickasaw Creek is above fluted slopes of the Moyers Formation which join with the Stanley-floored valleys. This consistent topographic position suggests a continuity of sandstone zones above and below the Chickasaw Creek throughout this distance.

(3) *Interstratification with nonturbidites.*—Sandstone beds of the Stanley-lower Atoka sequence are interbedded exclusively with shales. The only exceptions to this known to the writer are rare thin limestone beds exposed in the Finley syncline near the western end of the Ouachitas. Interstratification with shale is typical of turbidite sequences as well as of other sequences.

(4) *Oriented sole markings mainly restricted to soles of medium-grain-size beds.*—Oriented sole markings are common at several horizons within the Jackfork, the Johns Valley, and the lower part

of the Atoka. Although only a limited number of observations have been recorded, they suggest a uniformity of orientation over a wide area. In general, but not everywhere, the sole markings are more prominent on thick beds. The relationship of sole markings to grain size, if any, was not determined.

(5) *Aligned inclusions and, in coarse beds, aligned grains.*—Swarms of shale inclusions (referred to here as clay galls) are found at many levels in the Jackfork-lower Atoka sequence. These are more common on upper surfaces of relatively thick sandstone beds where they are parallel to bedding planes. They are present in lesser concentration and with less obvious orientation within several beds. Their maximum dimension is more than a foot, although most are less than six inches in diameter.

(6) *Unidirectional cross and ripple lamination.*—The orientation of cross-bedding is less uniform over a wide area than is the orientation of sole markings, judged by the measurements thus far made in the area. In a single exposure, however, all cross-bedding has approximately the same orientation; there appear to be no criss-cross arrangements vertically spaced. Such uniform orientation is characteristic of turbidites. Also characteristic of turbidites is the restriction of ripple structures to silty or fine sandy levels.

(7) *Burrows and reworked fossils only; fossils generally scarce.*—Other than tracks and burrows, megafossils of organisms buried in situ are not known from the Stanley-lower Atoka sequence. Invertebrate fossils are restricted to the lower coarser parts of sandstone beds, where they occur as detrital fragments of about the same size as terrigenous mineral grains in the same beds. Also fragmented is plant debris, which is especially noticeable near the tops of many sandstone beds.

(8) *Hydroplastic load casting and convolution.*—Hydroplastic features of several types are common in the Stanley-lower Atoka sequence. As noted by ten Haaf, convolute lamination, continuous and dying out towards the flat bedding planes, is presumably restricted to turbidites (1959b, p. 219). Such lamination is found at many levels of the Ouachita sequence.

(9) *Constant current trend.*—The readings of sole-marking orientations thus far recorded indicate a relatively constant current trend over a stratigraphic interval exceeding 10,000 feet. This trend is parallel to the apparent axis of the depositional trough and to the

present tectonic strike. These are common features of turbidite sequences (Kuenen, 1957).

(10) *Slump structures.*—No attempt was made to determine the directional nature of hydroplastic features and their relation to current directions. Therefore slump structures as such were not identified.

(11) *Absence of shallow-water phenomena.*—According to ten Haaf (1959b, p. 219), none of the turbidite sequences observed has such phenomena as wave ripples, tidal scour and fill, fossils of neritic animals or plants buried in place, or is in association with such rocks as reefs or clean, well-sorted sandstones. Beds of the Stanley-lower Atoka sequence are like turbidites in that they do not have wave ripples, tidal scour marks, or neritic fossils; but they are unlike turbidites in that they contain clean, well-sorted sandstones.

The foregoing paragraphs briefly compare features of the Jackfork-lower Atoka sequence with those thought characteristic of turbidites. Many, if not all, Jackfork sandstones differ from the typical turbidite in possessing better sorting and being relatively clean. Grading of some type is present in some beds but is not obvious in most. Many sandstone beds of the lower part of the Atoka are similar to those of the Jackfork, but also included are argillaceous sandstones that are probably more poorly sorted.

Sandstone beds of the Stanley have been described by workers such as Cline (1960) and Shelburne (1960). Considered as a group, the sandstones are thought to be more poorly sorted than those of the Jackfork and to include graywacke. Sorting improves upward in the Stanley, and Cline (1960, p. 37) classified many beds of the Moyers Formation as subgraywacke. (Cline used the definitions of graywacke and subgraywacke given by Pettijohn, 1957.)

Problem of Sorting Variation in a Turbidite Sequence

Present knowledge suggests that sandstone sorting generally improves upward from the Stanley into the Jackfork; although Bokman's results (1953, p. 163) do not support this generalization. Lower strata of the Atoka in the report area have a mixture of well-sorted and more poorly sorted sandstone layers. What could cause variation of sorting in a turbidite sequence?

Moretti (1958, p. 70) concluded that variation in sorting of beds in the Stanley-Atoka sequence is a direct reflection of differences in sorting in the source sediments of the shallow shelf. However, C. J. Mankin (personal communication) thinks that even if only well-sorted sand were available at the source, there should be turbulent mixing with the bottom material, and thus turbidites generally should not be well-sorted and clean. A question suggested by Moretti's hypothesis is why are the thicker beds of the Jackfork apparently more poorly sorted than the thinner beds (Cline, 1960, p. 46). If the degree of sorting is inherited from the source, why should thick beds accumulate when the shelf sediments are poorly sorted and thin beds accumulate only when they are well sorted? The thin beds are finer grained and at places display excellent planar, cross, or contorted lamination. Some of this lamination consists of alternating well-sorted layers (some of which are only a few millimeters thick) that have a varying mean grain size. Why are the mold faunas associated with granule- or pebble-size quartz grains? Are these associations due to characteristics of the transporting current, or are they acquired from shelf sediment?

It seems more reasonable to assume that texture, lamination, and good sorting are primarily due to the dynamics of transportation and deposition rather than to the characteristics of the source. But, if this be true, we are faced with two questions: (1) how do these dynamics fit into the over-all picture of a turbidite environment and (2) did a turbidite environment exist at all?

Pertinent to this problem is the recent work of A. E. Lombard of the University of Geneva. Lombard (personal communication) distinguished beds he referred to as "laminites" from turbidites and, though they occur in the same sequence of strata, he considered turbidites and laminites to be the products of two extremely different processes acting in the same or similar environments. Laminites are laid down by slowly moving currents that closely follow the bottom. Turbidites, on the other hand, are the result of submarine mass movement down a gentle slope at a high velocity.

Laminites, according to Lombard, are characterized by: (1) clay laminae that are planar, wavy, or inclined; (2) a thickness less than three feet, and, in some instances, of only an inch; (3) a general absence of grading; (4) a common group of heavy minerals throughout the sequence; (5) the possible presence of sole markings. Sorting in the laminites typically is better than that of the turbidites, and laminites are finer grained. Laminites may occur in shale sections between turbidite beds (fig. 50). Lombard considered an idealized sequence as having a turbidite bed at its base, and this as being overlain by interbedded laminites and shales with the sandstone-shale ratio decreasing upward. At the top of the sequence is a relatively thick shale section which is topped by the basal turbidite of the next series. Lombard distinguished several imperfectly represented idealized sequences in the basal Jackfork exposed on the north slope of Kiamichi Mountain south of Big Cedar.

Lombard defined laminites on a purely descriptive basis. In particular, he did not associate them with a particular depth of water. Some writers may consider them the distal deposits of turbidity currents, but it is this writer's understanding that turbidity currents moving slowly enough to deposit the laminites would quickly dissipate. Ericson and others (1961, p. 217) tacitly recognized laminites as distinct from turbidites. They referred to the laminites as "sand and silt layers due to winnowing by deep-current scour." The layers accumulated "under the influence of a continuous or nearly continuous current which has prevented accumulation of particles below a certain grade size, [and] accumulation under such conditions should be extremely slow."

Of interest is Lombard's comment that the isolated sandstone masses in shale sections may be former channel deposits that were linear rather than tabular. These channel deposits were subsequently

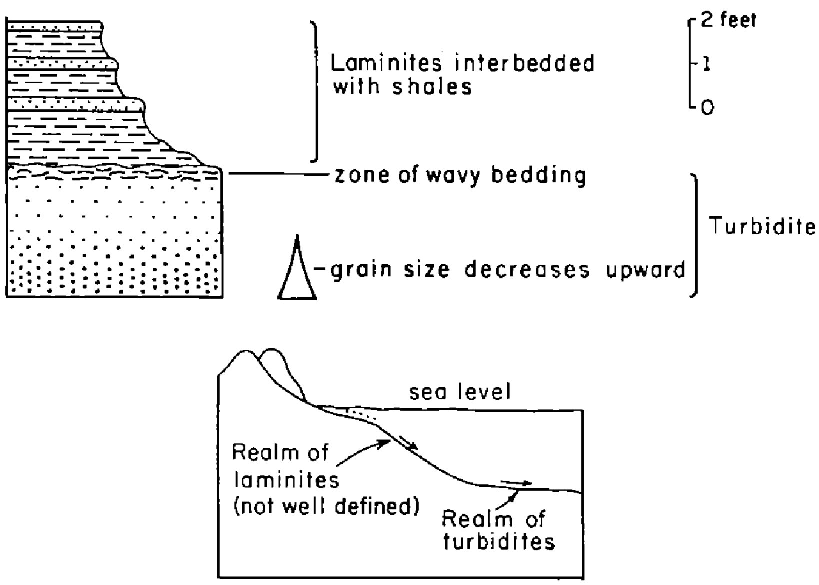


Figure 50. Turbidite-laminite relationships according to A. E. Lombard.

squeezed, causing their present discontinuous distribution. They could have formed on a tidal mud flat. However, they could also have been derived from tabular, rather than linear, beds (see *Plastic-Flow Structures*).

The descriptions and concepts set forth by Lombard have a direct application to the apparent difference between sandstones of the Stanley and the Jackfork. Those of the Stanley are dominantly turbidites; those of the Jackfork, laminites. What environmental changes are necessary to produce this difference, however, is not clear if, indeed, these changes took place at all. Does the apparent interbedding of laminites and turbidites indicate that the origin of both is related to a submarine slope? If so, why the variation in relative abundance of each to the other at different stratigraphic levels? To explain this nonuniformity of relative abundance, Lombard hypothesized that turbidites are the result of active subsidence and tilting, whereas laminites accumulate during intervening intervals of tectonic stability. Under this hypothesis the Jackfork would have been laid down during a relatively quiet interlude between a time of active subsidence during deposition of the Stanley and overlying Johns Valley-Atoka beds. The volcanism of Stanley time might be considered as corroborating evidence for this hypothesis. However, the work of Ericson and others (1961) on Atlantic deep-water sediments suggests that laminites are deposits of deep-sea currents and that these currents may have no relation to tectonic activity in the depositional basin.

Longitudinal Current Direction

Regardless of whether one accepts or rejects the turbidity-current hypothesis, it is evident that Kuenen has demonstrated the genetic relationship of rock sequences in different parts of the world. If similar causes give rise to similar effects, then many of the causal factors operating during the deposition of these sequences must have been similar. One of the more interesting effects is the dominance of a longitudinal current direction parallel to the axis of the depositional trough throughout deposition of most turbidite sequences (Kuenen, 1957). Transverse current directions have been detected along the margins of some troughs, but longitudinal directions dominate. Although further measurements are needed, a longitudinal current direction also appears to have been present throughout the depositional interval of the Jackfork and Atoka sediments of the western Ouachitas.

What can produce such a current trend through a long interval of time? According to the turbidity-current hypothesis, it is produced by the same bottom slope that caused the continued flow of the current. It was suggested by ten Haaf (1959b, p. 219) that the average slope is about 1 percent. This assumption requires that the water depth at one end of the trough is deeper than at the other and gives a means of approximating the minimum water depth at the deeper end. Consequently if the Jackfork beds near Mena, Arkansas, were deposited at sea level, the water depth near Antlers, Oklahoma, would have been about 4,000 feet. (This is based upon the assumption that the same turbidity currents deposited the Jackfork beds at both localities.) Because the Jackfork near Mena is probably a submarine deposit, the western depth should have been even greater. If Jackfork strata farther east in Arkansas also show predominantly westward current flow, the western depths would be further increased.

Other currents which may maintain a constant trend for relatively long intervals of time are longshore currents and currents of the open ocean. If we seek an answer to the characteristics of turbidites in these mechanisms, we are faced with the problem of how to "turn them on and off" or shift their course rather abruptly and repeatedly for the alternating deposition of sandstone and shale.

Can a major river maintain essentially unidirectional currents during the deposition of sediment sequences several thousand feet thick? Today's distributaries which deposit the coarser sediments of delta topsets fan out in divergent paths. Extrapolated into the past, their composite pattern should present an intricate maze. If the submarine slope upon which the river's sediments were deposited was maintained, unidirectional currents might be expected if gravity is the dominant force, but this reasoning returns us to the realm of turbidity currents.

Perhaps undue emphasis is being placed upon this seemingly dominant current direction and that further investigations will prove that the direction has been more variable. However, Kuenen (1957, p. 190-191) cited at least nine different sequences giving some evidence of prevailingly longitudinal currents; and preliminary data show that the same may be true for the Stanley-Atoka sequence of the western Ouachitas.

Tectonics, Turbidites, and the Stanley-Atoka Sequence

Directly involved in the problem of the origin of the Stanley-Atoka sequence of the Ouachita Mountains is the concept of a geosyncline. Does marginal uplift result from rock moving from below a subsiding trough? Are the flanking borderlands the major source of materials deposited in the trough? Can a major river system "create" a geosyncline by dumping its materials and causing the crust to subside? Are deep-sea trenches associated with island arcs of the present oceans typical or atypical geosynclines? Does the locus of deposition migrate toward the craton as the geosyncline evolves? These and other controversial questions arise from a study of the Ouachita geosyncline.

What is the common cause of all the effects which characterize turbidite sequences? A general explanation has been sought in the rapid synorogenic but preparoxysmal deposition of coarse and fine detritus in a thick marine series thought to characterize the flysch facies (ten Haaf, 1959b, p. 220). Cline (1960, p. 87) concluded that the Stanley-Atoka sequence is "comparable to the typical black-shale flysch facies of the Cretaceous and Eocene of the Alps and Carpathians of Europe." Bokman (1953, p. 169) determined by petrographic studies that the Stanley and Jackfork were laid down in a northward-migrating trough with an orogenic belt on its southern margin. He concluded that the older sediments of the Stanley were eroded and redeposited in the Jackfork. This origin is in harmony with the above-described concept of the flysch facies but requires extensive exposures of the Stanley in nearby borderlands. These exposures would have to yield huge quantities if the sandstone-poor Stanley were to be an adequate source for the sandstone-rich Jackfork. Cline considered that the Stanley is considerably less than 25 percent sandstone (Cline, 1960, p. 27) and the Jackfork is about 60 percent sandstone (Cline, 1960, p. 46). If one uses the convenient figures of 18 percent sandstone for the Stanley (10,000 feet thick) and 60 percent sandstone for the Jackfork (6,000 feet thick), then about 12,000 cubic miles of Stanley strata would have been required to supply all the sand for Jackfork sandstones in Oklahoma alone.

In Oklahoma, southward thickening of the Stanley and Jackfork has been demonstrated in the narrow frontal belt of the Ouachitas but does not necessarily continue into the central and southern parts of the mountains. In nearly all areas the structural complexity of the Stanley defies reliable measurement. There is no perceptible southward thickening of the Jackfork from the Rich Mountain syncline to the Boktukola syncline, a north-south distance of about fifteen miles which does not take into account shortening caused by thrusting. The presence of a southern borderland should, therefore, no longer be assumed because of thickening in that direction. In fact, there is little direct evidence of any kind for a southern source for nonvolcanics of the Stanley-Atoka sequence.

Bokman's hypothetical northward-migrating depositional belt is supported by the movement of the hingeline from the frontal Ouachitas in Stanley-Jackfork time northward into the Arkoma basin in Atoka time. It is possible, however, that the Stanley and Jackfork share a common hingeline in Oklahoma so that a similar migration did not occur during Stanley-Jackfork time.

Several researchers have stated that subsidence was rapid during deposition of the Stanley-Atoka sequence. What is the evidence for this hypothesis? True, the sequence is much thicker than its correlatives to the north and west. But is this, per se, evidence of subsidence? Are correlations across the strike well enough established to indicate thinning by convergence rather than by onlap? An alternate hypothesis would be that the site of the Ouachitas was a deep trench during deposition of the Arkansas Novaculite and this trench was filled with little further subsidence by clastics of the Stanley-Atoka sequence.

Stanley-Atoka Sediments and Shallow-Water Deposits

Could the Stanley-Atoka sequence have been laid down in predominantly shallow water? If one answers this question in the affirmative, problems arise just as they do with the turbidite-laminite interpretation. Of first importance is the absence of autochthonous megafossils. Abundant fossils are present in correlative beds exposed to the north in the Ozark uplift and even in the frontal Ouachitas. Detrital fragments of these or similar animals are preserved as molds in sandstones at widely spaced intervals of the sequence. The presence of these molds suggests that an autochthonous fauna should have been preserved if it existed. However, only tracks and burrows are found.

Perhaps the absence of a benthonic fauna is due to toxic bottom conditions in a shallow-water body. The black shales, carbonized plant matter, and localized occurrence of pyrite indicate a non-oxidizing environment such as would limit bottom life; but can a shallow restricted basin be maintained throughout the deposition of such a thick sequence?

The absence of a bottom fauna may be due to conditions similar to present-day tidal flats. However, although sparse, a fauna does exist on many, if not all, present-day flats (McKee, 1957, p. 1739). These flats also have abundant ripple marks, drag marks, and animal trails. Coquina, white sand, gray sand with black organic matter, and black muds are the dominant rock types. The rate of sediment accumulation is rapid according to McKee (1957, p. 1739) and reaches five to ten feet per year on several flats.

Except for coquina, the above features are also common to beds of the upper part of the Stanley, the Jackfork, and the lower part of the Atoka. The absence of an autochthonous fauna might be explained by the fact that forms capable of living on these flats had not yet evolved. Although ripple marks are present, they are not so abundant as they are on present-day tidal flats. However, McKee (1957, p. 1742) could find no evidence of ripple lamination in the deposits of the flat at Cholla Bay, Sonora, Mexico. Ripple marks on the surface of the tidal flat at Sonora are dominantly parallel crested. Short-crested ripples are limited to channels where tidal movement is concentrated. This relationship is in contrast to that observed in the Jackfork, where short-crested ripples are dominant and long parallel-crested ripples are relatively few.

The preservation of laminae in most sandstones of the Jackfork suggests that burrowing organisms were not active or that the laminae were buried to a depth greater than the organisms could penetrate. Burrowing organisms, however, are common inhabitants of today's tidal flats. If conditions appropriate to the survival of bottom organisms existed during several intervals of deposition, why did they not flourish throughout the entire depositional interval? Is it because there were two different types of deposition which gave rise to laminites and turbidites? Periods between turbid flows may have allowed burial of the sandstones before migrating animals could return to populate a sandy bottom. The laminites with their well-preserved laminae presumably were deposited by slower currents than were the turbidites. Slow-moving currents probably would not exterminate a bottom fauna or bury sand beds to an impenetrable depth, and therefore trails are present at several different levels of some laminite beds.

The presence of molds of fragments of invertebrates at the bases of several sandstones, particularly in the upper part of the Jackfork, in the Johns Valley, and in the lower part of the Atoka might be explained by extreme tides which brought in coarse offshore debris and laid it down as a basal deposit. This is the explanation given for them by Honess (1923, p. 198), who considered the Stanley to be the deposit of a large delta.

The reddish-brown color of the maroon shale member of the Wildhorse Mountain Formation and other less prominent shale beds present in the upper parts of the Stanley and the Jackfork indicates the possibility of a change to oxidizing conditions at intervals during deposition. The color change is apparently a primary feature, as indicated by the continuity of such zones as the maroon shale. Cline and Moretti (1956, p. 14) expressed this opinion. Oxidizing conditions might have been brought about by a decrease in the rate of supply of organic material while the oxygen supply remained constant or by a constant supply rate and variation in oxygen supply.

Serious problems arise with consideration of the tidal-flat interpretation. Foremost among these is the presence of lenticular bodies in today's tidal flats and their lack of lamination (McKee, 1957, p. 1742, 1746). In order to fit the Jackfork into a tidal-flat environment, it would require a flat of much larger areal extent than any known today, and the flat must persist with neither marine nor continental encroachments for a long time. The flat would have to be inundated long enough for sorting and lamination to be accomplished by currents, but not long enough for a fauna to develop or for symmetrical wave ripples to form. If long-shore currents were not present, there would have to be an eastward-sloping sea floor so that the incoming tides, which cause the strongest currents, would advance westward and continue to do so during the deposition of at least 6,000 feet of sediment.

The problem of a paucity of fossils might be explained by assuming that they existed but that the environment was not favorable for their preservation. In so doing we erect another problem: why was the detrital fauna preserved? We might postulate a series of smaller migrating tidal flats as a substitute for a single larger one. But then we must explain the apparent continuity of sandstone zones for many miles parallel to possible time-rock units if siliceous shales such as the Chickasaw Creek have time-parallel significance. A large tidal flat might have been more likely if a greater portion of the continental margin were flooded in Mississippian time than is flooded now. Why then were continuous sandstones produced rather than channel sandstones such as those of the Permian? Sandstone beds of shallow-water origin may be continuous for many miles, but, to the writer's knowledge, they have not previously been known to make up an unfossiliferous section as thick as the Stanley-Atoka sequence.

The accumulation of the Stanley-Atoka sequence in shallow water or on mud flats within the continental margin might be considered more likely than in deep water because modern depositional rates may be high in these localities. However, much remains to be learned about deep-water depositional rates. Although beds of the Stanley-Atoka sequence are commonly thought to be the result of rapid deposition, actually there is little evidence for the determination of absolute depositional rates.

Turbidites or Shallow-Water Sediments—A Summary of the Problem

The turbidity-current hypothesis for the depositional environment of Jackfork beds exposed in the report area is weakened by the absence of a proved mechanism for deposition of laminated, well-sorted, slightly graded or nongraded sandstone beds which make up a significant, but unknown, proportion of the total sandstone complement. Shallow-water hypotheses are weakened by the absence in the Jackfork of a record of localized energy concentrations found in shallow-water environments, by the absence of autochthonous fossils, and by the difficulty of maintaining uniform conditions throughout the deposition of thousands of feet of strata.

INTERPRETATION OF TEXTURE AND MINERALOGY

Literature on the interpretation of thin-section data gathered from sedimentary rocks has increased recently. Some interpretative criteria applied to thin sections are more reliable than others. One criterion, the significance of which is in dispute, is the presence of strained quartz. In the past, strained quartz has been thought to be primarily metamorphic in origin, but some present researchers find it more characteristic of igneous rocks. Whatever the outcome of this dispute, other criteria which are less controversial can be used for interpretation of source and depositional environment.

INTERPRETATION OF SORTING AND MEAN GRAIN SIZE

Bokman (1953, p. 163) stated that the average sandstone mean of the Jackfork is 2.80 phi and the average standard deviation, 0.85 phi. Moretti (1958, p. 25, 26) determined that the medians for Jackfork sandstones fall between 2 and 3 phi and that most standard-deviation values lie between 0.52 and 0.86 phi (converted from his Trask sorting coefficient values of 1.28 and 1.50). Both Bokman and Moretti based their calculations upon uncorrected raw thin-section data obtained by grain counts.

In the current study the standard deviation and mean were determined by estimation. The average of the thin-section means is 2.67 phi and the average standard deviation is 0.43 phi. These calculations are based upon 20 thin sections consisting of 1 from the top of the Stanley, 16 from the Jackfork, and 3 from the Atoka (appendix C).

The above figures indicate that the average Jackfork sandstone is fine grained and moderately to well sorted (sorting terminology from Folk, 1959, p. 103). The significance of the lower average standard-deviation values obtained by the writer is probably due to sampling differences, but it could be due to an actual difference in sorting or to estimation errors although an attempt to reduce the latter was made by use of a comparison chart and use of estimates made by several observers. Laminations due to variations in mean size occur in several of the thin sections. If these had not been treated separately, poorer sorting values would have been obtained.

Moretti (1958, p. 26) concluded that sandstone beds of the Jackfork show no correlation between median grain size and sorting. Results of the current study are not adequate to confirm or refute this. Observations of thin sections suggest that the finer grain sizes tend to be better sorted and that the coarser grain sizes are more poorly sorted. However, the few coarse-grained slides observed appear to be bimodal, and their sorting values have a different significance from those of unimodal slides.

Folk (1959, p. 5) stated that probably in every environment sorting is strongly dependent upon grain size. In general, the better sorted sediments have a mean size between 2 and 3 phi, and these have been derived from the stable residual products liberated from weathering of granular rocks like granite, schist, phyllite, metaquartzite, or older sandstones in which the grains were derived ultimately from one of these sources.

This concept is based upon the hypothesis that certain grain-size populations are much more abundant in nature than are others. A degree of sorting is therefore accomplished by magmatic, metamorphic, and weathering processes. This sorting establishes size limits of source materials and imposes limitations upon the sorting effected by sedimentary processes.

The Stanley-Atoka sequence is composed of alternating rock types of two of the major populations: the sand-coarse silt population and the clay population. The pebble population is unrepresented except for local occurrences of granules and small pebbles.

One might expect the grains of the granule-containing beds of the sequence to be more poorly sorted than those of the finer beds because the coarser beds represent an intermixing of two populations. Generally this appears to be true in the small sample observed by the writer.

After studying seven beds from the Stanley and the Jackfork, Bokman (1953, p. 164) concluded that bed thickness and mean grain size are not systematically related and that the small amount of grading present is of questionable significance.

Folk (1959, p. 4) suggested that mean size is a function of (1) the size range of available materials and (2) the amount of energy imparted to the sediment. Sorting depends upon (1) size range of material supplied, (2) type of deposition, (3) current characteristics, and (4) time.

The over-all limited range of grain size in the sandstones of the Stanley, Jackfork, and lower part of the Atoka might be explained by assuming only a limited size range of available materials or by assuming a wide size range of available materials and a long enough distance of transport for effective sorting. The size of the erratics of the Johns Valley and their restriction to a narrow belt along the frontal Ouachitas suggest that their source was nearby and different from that of the source of associated sandstones and shales.

The presence of fresh feldspar and various other rock fragments suggests that part of a single source area, or one of several source areas, was eroded rapidly. The streams of the source area probably supplied a broad size range of material; however, the larger sizes did not reach the Stanley-lower Atoka depositional basin. Apparently the small mean grain size of the sandstones in the basin is due to sorting over an adequate distance of transport.

According to Folk, sorting is essentially a function of the mechanical energy exerted upon a sediment after it reaches the site of final deposition. Currents of a constant velocity that are neither very rapid nor very slow will produce well-sorted sediments if the rate of supply is not too high and the grain-size range of supplied materials is of a suitable magnitude.

If thicker beds of the Jackfork are more poorly sorted and contain a higher percentage of matrix (Cline, 1960, p. 46) but have a mean grain size that is similar to thin beds, their texture could be a result of depositional rate. The higher rates would typify thick beds and the lower rates, thin beds. Under this hypothesis the deposition of each bed would require approximately the same amount of time and mechanical energy. The sorting differences would be due to the changes in mechanical energy per unit volume of sediment deposited. This hypothesis requires that the size range of material supplied, type of deposition, and current characteristics would remain constant while depositional rate would vary.

Under another hypothesis the size range of material supplied would remain constant while the type of deposition, current characteristics, and depositional rate would vary. This would result from having two different types of deposition, that of turbidity currents and that of normal bottom currents. The thick, more poorly sorted beds would be deposited by turbidity currents, whereas the thin, well-sorted beds (laminites) would be laid down by normal bottom currents.

INTERPRETATION OF GRAIN ROUNDNESS AND MINERAL SUITE

Grain roundness is particularly informative in the reconstruction of source areas. Although recrystallization has modified original grain outlines in many sandstone beds, less altered beds give a general picture of roundness.

Bokman (1953, p. 163) classified 45 percent of the Stanley grains as angular, 35 percent as worn, and 20 percent as rounded. Moretti

(1958, p. 79) estimated that perhaps one-half the grains of Jackfork sandstones can be considered to be angular or only slightly rounded. In the current study, it was noted that most of the plagioclase grains and many of the quartz grains show a high degree of angularity. This, alone, indicates that little rounding took place in the depositional basin. Confirming this belief are the size and shape of the micaceous metamorphic rock fragments in the sandstones. These are as large or larger than the mean of the quartz grains and are too soft to withstand extensive abrasion. Their shape commonly has been distorted by diagenetic processes, but less deformed grains show practically no modification by abrasion.

Assuming that little rounding occured during Stanley-Atoka deposition, rounded grains in the same sediment must have been abraded during a previous cycle of deposition. Large composite stretched metamorphic quartz grains and some of the zircon and tourmaline are well rounded. If the rounding occurred in the last sedimentation cycle, the incorporation of products of two different energy levels into the same sediment would be indicated. Beach dune-sand grains blown into the adjoining lagoonal environment illustrate this type of phenomenon. However, well-rounded quartz, zircon, and tourmaline in the sand-size range are not likely to result from a single sedimentation cycle and therefore probably have been derived from a sedimentary rock source.

Euhedral zircon occurs with rounded zircon in some laminae. This extreme difference in roundness indicates that the zircon has been derived from at least two different sources. Folk (1959, p. 94) stated that euhedral zircon is indicative of volcanism. If this be true, the volcanoes which caused the tuffs of Stanley time may have been the source of the zircon.

Volcanoes may have been the source of the small quantity of angular plagioclase grains in sandstones of both the Stanley and Jackfork. According to Honess, plagioclase of the lower Stanley tuff has an oligoclase-andesine composition (1923, p. 180), and the tuff at the top of the Stanley discovered in the present study has an andesine composition. Unfortunately the compositions of the plagioclases of the Stanley and Jackfork sandstones are not available for comparison with that of the tuffs. Some writers have stated that acidic or sodic plagioclase occurs in the sandstones, but no further differentiation has been made. A few grains in thin sections of the present study permitted tentative identification as andesine. Of interest is Folk's conclusion (1959, p. 80) that more plagioclase than potassium feldspar in a formation is suggestive of a volcanic source. This predominance appears to be present in the Jackfork, but further evidence is needed before the plagioclase can be assigned definitely to a volcanic source.

Miogeosynclinal Graywacke

In a discussion of miogeosynclinal graywacke, Krynine (Folk, 1959, p. 118) described rocks with many features similar to those of the Stanley-lower Atoka sequence. Conditions and features are idealized, but they provide an interesting basis of comparison.

The miogeosyncline (or exogeosyncline) is between the craton and a welt of metasediments offshore. Seaward from the metasediments is a eugeosyncline and then a belt of metamorphics, granitized rocks, and some volcanoes. In the miogeosyncline, sediments being deposited become increasingly more mature toward the craton. Several other sedimentary changes occur in a direction from the axis of the miogeosyncline toward the craton: (1) the percentage of soft metamorphic rock fragments declines relative to the percentage of hard fragments, (2) the amount of clay matrix declines, and (3) the percentage of quartz or carbonate cement increases. Lithologic character changes from that of a graywacke to that of a subgraywacke. Clean, white, well-sorted, mature, subgraywacke beds are laid down either far from the metasedimentary welt or near it along local temporary strand lines. During brief periods of tectonic quiescence these sands may be spread across the entire basin as blankets of supermature subgraywacke or orthoquartzite.

The metasedimentary welt is composed of low-rank metamorphics such as metaquartzite, slate, phyllite, and schist, and its formation is due to the same forces which form the elongate eugeosynclines and miogeosynclines. The metamorphics provide a source for a flood of clay and mica which produces a thick shale section. Because metamorphism is low rank, the heavy-mineral suite is simple. Heavy minerals may include zircon and tourmaline which were present in the sediments prior to metamorphism. Potassium feldspar is absent or is present in minor quantities except where the rocks were eroded enough to expose high-rank metamorphics and the granitized core. Volcanism may be represented by the presence of angular grains of plagioclase, as well as by other kinds of pyroclastics.

From the above brief description it is apparent that the Stanley-lower Atoka sequence could have been laid down in such a miogeosyncline. Stanley sediments would have resulted from relatively rapid burial by the flood of material adjacent to the metasedimentary welt or, more probably, from relatively rapid subsidence within the geosyncline farther from the welt. Jackfork beds were deposited in the geosyncline during relatively slow subsidence when several blanket sands were spread over the floor of the geosyncline. Some of these sands contain granules, small pebbles, and fossil fragments derived from environments nearer shore. The Moyers Formation and the lower part of the Wildhorse Mountain Formation represent a transition from conditions of relatively rapid subsidence to slow subsidence. The Atoka Formation was the final unit laid down in the geosyncline and is transitional (marine to nonmarine).

The major source area of the Stanley-Atoka sequence would lie to the south and today is revealed in the belt of slate, phyllite, and metaquartzite in subsurface beds of Texas (Flawn, 1959, p. 22). (However, there is little direct evidence of a southern source.) The dominant currents distributing sediments within the geosyncline would be from east to west in eastern Oklahoma. The volume eroded was enormous so that the source area probably consisted of either a single large uplift or several smaller uplifts. It was probably subjected to repeated uplift and compression. Volcanoes existed within the metasedimentary welt or outside of it and contributed pyroclastics which decrease in volume northward and, in Oklahoma, westward. Pyroclastics also decrease in volume upward from the base of the Stanley.

Tectonic activity ultimately responsible for compression causing the Ouachita thrust belt began prior to depositon of the Stanley shale, probably in the Early Paleozoic. By Stanley time the locus of deposition had migrated northward to a location including the presently exposed Ouachitas, and the earlier Paleozoic sediments had been subjected to repeated northward-migrating deformations and had become metamorphosed. They provided one source of Stanley, Jackfork, and, perhaps, Atoka sediments, but other sources were probably of equal or greater importance. What is commonly referred to as the Ouachita orogeny is but the final paroxysm of an orogenic belt with a long history of deformation. Mesozoic and Cenozoic sedimentation has taken place seaward of this belt.

The general picture given in the preceding paragraph is similar to that described by King (1959, p. 66) for the Appalachian Mountains. It is highly speculative but has many appealing features. Bordering the United States along the east and south during Paleozoic time was a belt having many tectonic and sedimentational features in common throughout its length.

King (1959, p. 66) suggested that it would be better to refer to the series of orogenies that deformed the sediments in the Appalachians at various times in the Paleozoic as the Appalachian Revolution and use the term "Allegheny Orogeny" for the final deformation which occurred near the end of the Paleozoic. A similar procedure might be advisable for the Ouachita Mountains; the "Ouachita Revolution" could refer to Paleozoic orogenies within the geosyncline and the "Choctaw Orogeny" could refer to the final deformation which resulted in the structures of the Ouachita Mountains.

The absence of sedimentary structures indicative of oscillating currents produced by wave action suggests that the Stanley-lower Atoka sequence of the Ouachita trough was laid down below wave base. If this be true, the statement that the better sorted, cleaner sediments of the Jackfork indicate shallower water deposition than that of the Stanley seems questionable. Where wave action is the dominant source of mechanical energy during deposition, it would be expected that, in a general way, the sediment would become more mature as the water becomes shallower. However, because the mechanical energy does not appear to be of wave origin, other indices of depth must be sought. In present-day seas mechanical energy is influenced by such factors as configuration of the ocean floor, salinity and temperature gradients, and prevailing wind directions. Mechanical energy may have no regular variation with depth. Mechanical energy supplied by turbidity currents has been evident in water as shallow as the lakes of Switzerland and as deep as the open ocean, but its broad lateral extent seems to occur only in deeper water bodies. Turbidites, therefore, are not indicators of depth.

It is possible that the westward-flowing currents of Stanley-lower Atoka time were due to a dominant circulation pattern such as that in an east-west trough having a restricted inlet and a restricted outlet; or the westward-flowing currents could be related in some yet unknown way to turbidity currents. Present research on environments below wave base should eventually allow more definite conclusions. At the present time, it appears safe to state that the

Jackfork Group, Moyers Formation, and, probably, the Tenmile Creek Formaton were not deposited on river flood plains, on the topset beds of a delta, on a tidal flat, or in shallow offshore water above wave base. They were probably deposited below wave base.

STRUCTURE

Structural Features

The base map upon which the geology is plotted was obtained from the U. S. Geological Survey. It consists of parts of three 15-minute quadrangle maps: Mena, Arkansas; Potter, Arkansas-Oklahoma; and Page, Oklahoma.

Information added to the base map was derived from two primary sources: field observation and aerial photographs. Initial interpretations were based upon examination of stereoscopic pairs of aerial photographs. Then areas where field relationships are not clear were selected for field checking. Most of these areas are along the southern margin of the Rich Mountain syncline and to the north and west of the Eagle Gap syncline. Elsewhere the structure is relatively simple and, where possible, attitudes were determined directly from the topographic map by transferring information from the aerial photographs. Several of the attitudes determined in this way may be considered more accurate than those obtained in the field because this method minimizes the effect of local slump. In most instances elevation control for these determinations was improved by the use of 7.5-minute quadrangle sheets with a 1:24,000 scale and a contour interval of 20 feet. Topographic expression of resistant sandstones near the middle of the Wildhorse Mountain Formation may be seen as southward-pointing V's where these beds are incised by southward-flowing streams.

Aerial photographs in stereoscopic pairs were useful both in the office and in the field. Subtle differences in the vegetation and topographic patterns, which are not evident in the field, may be clearly seen on the photographs. The dense overgrowth and actively creeping regolith effectively mask many such features on the ground. The photographs also show local structure in relation to the over-all structural pattern.

Although data acquired in the field and from aerial photographs during the course of this study are adequate to distinguish major structures of the area, they do not show the small features. Many small features are mantled by colluvium or alluvium, but others may be mapped successfully. Future, more detailed investigations should yield more useful information in structurally complex areas.

Early in the investigation it was noted that Middle Mountain* and Self Mountain* in the south-central part of Rich Mountain syncline, have nearly identical topographic forms and vegetation patterns. Their similar appearance is obvious in views from the crest of Rich Mountain and upon aerial photographs. This similarity suggested the possibility that a strike thrust fault, upthrown on the south, follows the valley separating the two ridges. This fault would join the Windingstair fault in sec. 12, T. 2 N., R. 26 E. and sec. 25, T. 1 S., R. 32 W. There is adequate complexity and cover to map the fault trace through these two sections.

If such a fault exists, Self Mountain is supported by the same upper Jackfork beds as is Middle Mountain and little or no Atoka is present in the Rich Mountain syncline. The Johns Valley-Atoka sequence described in the text on plate II of this bulletin would then be upper Jackfork in most part, although some beds younger than Jackfork may be present. Such an interpretation of the structure and stratigraphy of Rich Mountain syncline was not used because (1) Middle Mountain contains a clearly defined siliceous shale zone whereas Self Mountain contains only poorly defined, thinner, siliceous zones and these occur in a different stratigraphic position and (2) the sandstones exposed at the crest of Self Mountain are much more fossiliferous than the sandstones exposed in Middle Mountain. According to this interpretation the siliceous shale on Middle Mountain is typical of the Game Refuge at other exposures in Rich Mountain area and at other localities in the Ouachitas (Cline, 1960, p. 58). The siliceous zones and fossiliferous sandstones of Self Mountain are then correlatable with rocks of the lowermost Atoka such as those exposed on Spring Mountain, near the town of Stapp, and at the Hairpin Curve on State Highway 2 in sec. 3, T. 3 N., R. 19 E.

The stratigraphic evidence against a strike fault between Middle

* Middle and Self Mountains are not labeled on plate I. The crest of Middle Mountain coincides with the outcrop of the Game Refuge Formation between Mill Creek and Bufram Creek. The crest of Self Mountain is at the contact between the Atoka-Johns Valley Formations (undifferentiated) and the basal sandy member of the Atoka Formation south of Kiamichi River and Bufram Creek.

Mountain and Self Mountain is not conclusive, however, especially considering our inability to correlate many lithologic zones across the strike of Ouachita structure. The possibility that such a fault is present should, therefore, not be ruled out.

WINDINGSTAIR FAULT

The approximate trace of the Windingstair fault may be determined from its topographic expression. The truncation of internal structures of the Rich Mountain syncline along its southern border is readily seen at its eastern and western ends. Just to the west of the report area, Simmons Mountain is also sharply truncated by the Windingstair fault.

The fault trace, however, cannot be precisely located at any particular point. A fault zone is exposed in the excellent outcrops at the east end of Ward Lake dam, sec. 6, T. 2 S., R. 30 W., and is described in the East Ward Lake measured section (interval 3). The zone probably marks the trace of the Windingstair fault. Faulting on an indeterminate scale is evident in beds adjacent to the zone on the south. The dip of the Windingstair fault is not clear here, but the best defined of the faults to the south dips northward at 30 to 45 degrees. Here, and west along the foot of Long Mountain, beds on each side of the fault have concordant attitudes. Positioning of the fault at the base of Long Mountain is based upon the presence of a zone of small folds and faults.

Farther to the west, location of the fault trace is based primarily upon its topographic expression. Colluvium and alluvium at the foot of the south slope of the mountains cover the trace throughout most of its extent.

HONESS FAULT

The trace of the Honess fault is followed by Big Creek down the valley between Rich and Black Fork Mountains. Honess discovered the fault and showed it in one of his cross sections (1924, p. 21). For some reason the fault trace was not shown on the geologic map of the same publication. Perhaps because of this, it was not included in the 1954 edition of the *Geologic map of Oklahoma* by H. D. Miser.

Honess was a pioneer geologist who mapped large areas of the

Ouachitas in southeastern Oklahoma between 1916 and 1923. It seems appropriate to give honor to this man. Accordingly, on August 10, 1961, the Domestic Names Commitee of the U. S. Board of Geographic Names, at the writer's request, gave the name Honess Mountain to a previously unnamed mountain, the summit of which lies in sec. 29, T. 3 N., R. 26 E. This mountain lies approximately at the west end of the fault herein named Honess fault.

The recognition of the Honess fault solves a problem posed by the late C. W. Tomlinson. He noted in the valley between Rich and Black Fork Mountains that beds making up Black Fork Mountain appear to be dipping south, whereas the latest editions of geologic maps of both Oklahoma and Arkansas indicate that north dip is to be expected.

Along most of the fault trace in the report area, beds of the upper part of the Tenmile Creek Formation are in contact with Jackfork strata younger than the Wildhorse Mountain Formation; consequently, the stratigraphic throw is 5,000 to 7,000 feet. Isolated fault exposures occur in the stream beds of Big Creek and the Ouachita River, but it could not be determined whether they are of Honess fault or of associated faults. The trace of Honess fault rather closely follows the courses of these two streams as indicated by the exposures of the Stanley to the south of them and the Jackfork to the north. No control is available for placement of the trace of the fault near its juncture with the Briery fault south and west of Page, Oklahoma.

East of the area, topography clearly indicates a fault at the base of the south slope of Irons Fork Mountain. This fault may be a continuation of the Honess fault and, if so, it probably extends many more miles eastward.

BRIERY FAULT

The Briery fault is continuous across the area and beyond it both to the east and west. It is named for Briery Creek, which follows its trace for a short distance along the north base of Black Fork Mountain. As with other major faults, much of the trace of the Briery fault is covered with alluvium and colluvium, but its effect on topography is readily observable. A small exposure of the fault trace is seen where it crosses State Highway 103.

Fossiliferous Atoka sandstones form low ridges north of Black

Fork Mountain. These contrast sharply with Stanley shale strata which underlie the lower north slopes of the mountain and reveal the approximate location of the fault trace. Maximum stratigraphic throw of the Briery fault occurs here and could be as great as 25,000 feet because beds of the Tenmile Creek Formation are in contact with beds in the core of the Black Fork syncline that are probably near the top of the Atoka Formation. On the south edge of the Spring Mountain syncline, the stratigraphic throw reaches a minimum of no more than a few thousand feet.

RICH MOUNTAIN SYNCLINE

The Rich Mountain syncline is a large structure between the Windingstair fault on the south and the Honess and Briery faults on the north. Its southern flank has been modified by several folds and faults. In an east-west direction, it extends from Mena, Arkansas, to Big Cedar, Oklahoma. The name is taken from Rich Mountain, which forms a prominent ridge along the north flank of the syncline.

The syncline is asymmetrical; portions of its south flank are nearly vertical to overturned; dips of beds in its north limb do not exceed 60 degrees. The steepest dips are midway along the length of the structure near the Arkansas-Oklahoma state line. At the western end of the syncline, the dip of the axial surface is about 40 degrees south. Eastward this dip appears to increase to about 50 degrees south. These determinations are subject to error because they are based upon constructions of planar segments to represent an axial surface that may be slightly curved (see *Dip of the Windingstair Fault,* p. 117).

The axial trace of the Rich Mountain syncline is not continuous along the length of the structure. The folded south flank of the syncline has been thrust northward across the trace in sec. 7, T. 2 N., R. 27 E., and sec. 25, T. 1 S., R. 32 W. The major syncline to the west of sec. 7, T. 21 N., R. 27 E., is actually a separate fold from that to the east. The axial traces of both folds are referred to as representing the Rich Mountain syncline because they form the dominant synclinal structures along its length and have a common north flank.

The south flank of the Rich Mountain syncline has many smaller structures, only a few of which appear important enough to require naming at the present time. Horsepen anticline is named for Horsepen Creek, which flows through sec. 9, T. 2 N., R. 27 E. In this area the anticline separates a syncline which underlies Pine Mountain from the axial fold of the Rich Mountain syncline. At its eastern end, where it is cut by the Windingstair fault, the anticline is overturned. To the west, the tight fold passes into a thrust fault which will be referred to as the Horsepen fault, which probably dips between 15 and 40 degrees to the south.

Many small-scale structures are in the general area of sec. 34, T. 1 S., R. 31 W., and are exposed on Round Mountain, Middle Mountain, and Long Mountain. Only the larger features are mapped, however. One of these is an east-west fault, the trace of which coincides with the north base of Round Mountain and continues westward to Pilot Mountain where it joins the Windingstair fault. To the east it is continuous with the axial trace of a syncline, the axial surface of which dips southward. It is here named the Round Mountain fault, and the anticline bordering it on the south is the Round Mountain anticline. The anticline is asymmetrical and has a southward-dipping axial surface. Much of its north limb is nearly vertical. Surface control for the fold is best in secs. 34, 35, T. 1 S., R. 31 W., and is poor at the east and west ends of the structure.

To the south of the Round Mountain anticline is a syncline, the south flank of which is exposed on Long Mountain and is here named the Long Mountain syncline. It is an open fold that is relatively symmetrical. Several unmapped faults and small folds are present in the syncline. Such features are particularly noticeable on the crest of Long Mountain and near the axial trace of the fold in the valley below. Although possibly possessing steeper dips, attitudes of beds to the south of the mapped trace of the Windingstair fault are concordant with those of beds in the Long Mountain syncline.

EAGLE GAP SYNCLINE

Black Fork Mountain is the next ridge to the north of Rich Mountain, and rocks composing it make up the north flank of a syncline, the south flank of which is almost entirely missing in surface exposures. The syncline is herein named Eagle Gap syncline for the pass between Black Fork and Fourche Mountains. (The more appropriate name Black Fork Mountain is rejected because of the possibility of confusion with Black Fork syncline named by Reinemund and Danilchik (1957) in the Waldron quadrangle a few miles to the north.)

Evidence that the structure of Black Fork Mountain is synclinal is the abbreviated south flank at the western end of the mountain near Page and the change in strike to the west of Eagle Gap. Isolated exposures along the axial trace at the western end of the structure allow tracing of some beds of one flank across the axis and up the other flank.

To the east of Black Fork Mountain, Fourche Mountain is considered a continuation of the Eagle Gap syncline. No suggestion of a south limb was observed on Fourche Mountain.

SPRING MOUNTAIN SYNCLINE

The Spring Mountain syncline is the third largest syncline that lies almost entirely within the report area. It is crossed by State Highway 103 near the western edge of the area. Spring Mountain, from which the syncline derives its name, is the highest ridge in the syncline, and is made up of resistant sandstones at the base of the Atoka Formation that are part of the south flank of the syncline.

State Highway 103 follows the axial trace of the structure in secs. 17, 18, 19, T. 3 N., R. 26 E. The steepest dips of the north limb are about 50 degrees, whereas the south limb is vertical or slightly overturned. The attitudes along the south limb have been markedly affected by downslope movements and by small faults. However, it is clear that the syncline is asymmetrical and probably has a southward-dipping axial surface.

The eastern end of the syncline is cut by several faults and the axis there has a westward plunge. To the west the structure narrows in breadth as folding appears to become tighter, and some of the section may be eliminated by faulting.

STAPP SYNCLINE AND STAPP FAULT

The Stapp Conglomerate was named by Harlton (1938, p. 893) for beds exposed in the Kansas City Southern Railroad cut in SW¼ SW¼ sec. 7, T. 3 N., R. 26 E. These beds form part of the south flank of a north-northwest-trending syncline herein named the Stapp syncline. At the west end the syncline plunges eastward, but in the east the south limb is truncated by faulting, and beds of its north limb are folded to form an eastward-plunging anticline. The anticline is probably asymmetrical with a southward-dipping axial surface. The Stapp syncline appears symmetrical, but the inconsistency of dips along its south flank suggests that they may have been reduced by downslope movement, and, therefore, it may be asymmetrical.

The Stapp syncline is in fault contact with all surrounding structures except the small anticline described above. The most prominent fault borders the syncline on the north and continues westward along Shawnee Creek and the north base of Tram Ridge. This fault is here named Stapp fault.

BLACK FORK SYNCLINE

The Black Fork syncline was mapped and named by Reinemund and Danilchik (1957) in their study of the Waldron quadrangle. It is a large structure that includes about 18,500 feet of the Atoka Formation in surface exposures.

A short segment of the south flank is present in the report area. It is represented by the southwestern tip of Horseshoe Mountain in sec. 13, T. 1 N., R. 32 W. The north flank borders the area to the north of the Potter 15-minute quadrangle.

SHUT-IN ANTICLINE

Shut-in Mountain is a high ridge near the southern edge of T. 1 N., R. 31 W. On a small-scale extended-coverage map included in their report, Reinemund and Danilchik (1957) showed its underlying structure to consist of a doubly plunging anticline. The name Shut-in anticline is derived from the mountain on which it is exposed.

The present interpretation of the geology of the area differs from that of Reinemund and Danilchik. The Shut-in anticline is bordered on the north by a syncline that they did not map. Outcrops of Johns Valley Shale mapped by Reinemund and Danilchik were not found to the north. The chert that is present is believed to be that of the Wesley Shale rather than that of the Johns Valley Formation.

Interpretation

NORTHWARD ELIMINATION OF SOUTH FLANKS OF SYNCLINES

A striking feature of Ouachita structure in Oklahoma is the change in pattern of folding and faulting from the central parts of the mountains northward toward the Choctaw fault (fig. 2). Central folds have more gently dipping limbs than have those to the north and west, and they have a greater areal extent. In the frontal belt of the mountains, no large synclines are present and imbricate structure prevails. At no other place in the Ouachitas is the transition northward from open folds between widely spaced thrusts to imbricate structure better illustrated than near the Arkansas-Oklahoma state line. In order from south to north, the Lynn Mountain, Rich Mountain, and Eagle Gap synclines illustrate progressive elimination of south flanks.

The Lynn Mountain syncline lies to the south of the Rich Mountain syncline. Both synclines are composed of the same stratigraphic interval. The writer did not map the Lynn Mountain syncline, but the topography indicates that most of its southern flank is present as far west as sec. 19, T. 1 N., R. 26 E. (pl. I). To the west of this point, some of the surface section is eliminated by the Octavia fault. Even so, the Lynn Mountain syncline is twice as broad as the Rich Mountain syncline in north-south dimension, and dips within the Lynn Mountain syncline are more gentle.

In the Lynn Mountain syncline upper beds of the Jackfork Group are probably present at least as far west as sec. 17, T. 1 N., R. 25 E., on the south limb of the structure. Jackfork strata are not represented at the surface on the south limb of the Rich Mountain syncline, however, except at its eastern and western ends. Beds of the Atoka Formation make up the middle part of the limb.

North of the Rich Mountain syncline the Eagle Gap syncline has no south limb except at its extreme western end where a few beds of the Wildhorse Mountain Formation swing around to form an abbreviated south flank. Uppermost Jackfork, Johns Valley, and Atoka strata are not represented in the structure.

Based upon the hypothesis that the dominant deforming-force components were horizontal, the pile-up of synclines to the south of the Briery fault could be explained by assuming the existence of a barrier to the north and a gradual southward decrease in horizontal stresses south of the Briery fault. The barrier could have been the thick section of the Atoka Formation present in the Black Fork syncline and along the southern margin of the Arkoma basin, which may have been at the foot of a northward-sloping land surface at the time of thrusting. Another barrier within the Ouachita frontal belt may have been formed by basement relief associated with the Stanley-Jackfork hingeline. This relief was probably accentuated by up-to-the-north faulting during Johns Valley time.

DIP OF THE WINDINGSTAIR FAULT

The angles at which Ouachita faults are inclined is a controversial detail. Some geologists insist that all of the major faults must have a high angle of dip, whereas others feel that at least some of the faults dip at low angles. Generally the trace of the Windingstair fault is of no use for determination of fault dip. The sinuosity of the trace is the result of the intersection of a curving fault surface with the ground surface. This is indicated by the irregularly curving nature of the trace where it traverses low-relief terrain. However, the Windingstair fault borders the Rich Mountain syncline on the south, and structures within the syncline should reveal something of the nature of the fault in the report area.

The axial trace of the Rich Mountain syncline traverses terrain with 500 to 800 feet of relief for a distance of more than six miles at its west end. The attitude of the axial surface may thus be approximated by using the three-point method where segments of the surface are probably planar. If the surface segments are actually concave northward, the calculated dip will be greater than the true dip; if, instead, they are convex northward, the calculated dip will be less. The first case is probably illustrated by the extreme west end of Rich Mountain; the second, possibly by Wilton Mountain*. Quentin Peak* lies between these two and the axial surface appears to be least curved there. The axial surface dips southward at an angle of 40 degrees as determined by computations based upon the axial trace on Quentin Peak. To the east in Wilton Mountain it appears to increase to about 50 degrees.

If the major Ouachita faults were formed by the rupture of growing asymmetrical or overturned folds, it should be possible to predict something of the nature of these faults by studying the syn-

* Wilton Mountain and Quentin Peak are not labled on plate I; they are in sections 6 and 34, respectively, of T. 2 N., R. 26 E.

clinal fold segments that remain (fig. 51). The 40-degree dip of the axial surface of the west end of the Rich Mountain syncline suggests that the Windingstair fault dips 40 degrees or less southward in this area.

Farther eastward in the Rich Mountain syncline, the Horsepen fault dips south at an angle of between 15 and 40 degrees. Eastward the fault is replaced by the axial surface of the Horsepen anticline, which has an approximate southward dip of 50 degrees. The Horsepen fault and anticline are smaller structures than the west end of the Rich Mountain syncline, but their presence suggests that the Windingstair fault dips southward at an angle of 50 degrees or less in their vicinity.

In the east end of the Rich Mountain syncline, the Round Mountain fault has an unknown angle of dip, but its direction of dip must be southward to account for the steep dip of the north flank of the Round Mountain anticline. The Round Mountain fault has a decreasing stratigraphic throw eastward and ultimately dies out in a syncline. It represents thrusting of the folded south flank of the Rich Mountain syncline over the north flank.

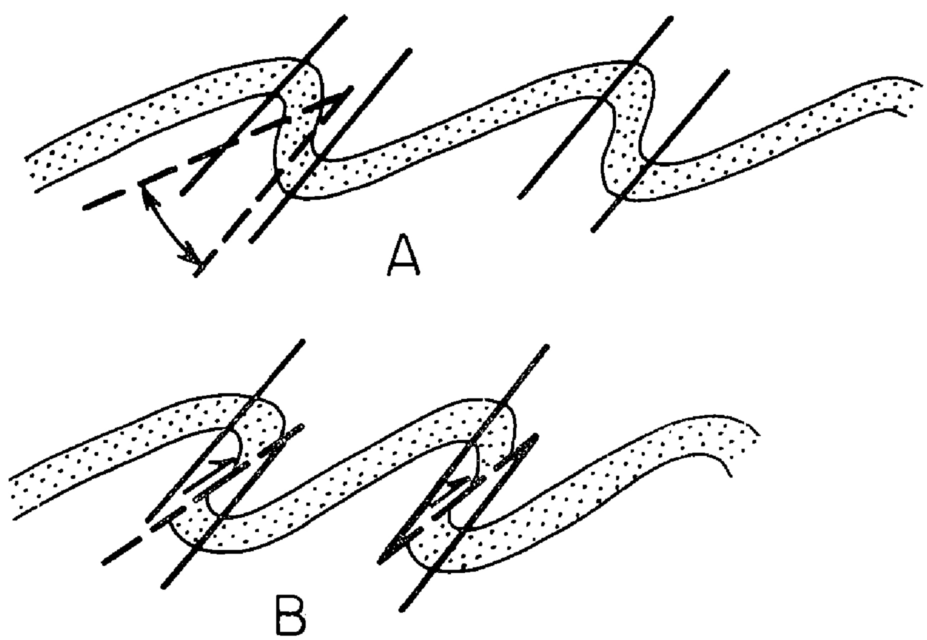


Figure 51. Relationship of fault surfaces to axial planes of folds.
A Overturned folds before faulting. Faulting will probably occur with a dip angle within the limits of the arc indicated by the dashed lines, which represent traces of incipient faults. The angle of dip of a fault will be equal to or less than that of the axial surfaces (solid lines) of the folds.
B Overturned folds after faulting. Dashed lines are traces of faults. Solid lines are traces of the axial surfaces of the folds.

Farther to the east the structure of Long Mountain syncline appears continuous across the mapped trace of Windingstair fault. The continuity of this structure, in fact, suggested the possibility that the fault cuts across the crest of Long Mountain where the structure is complex (not shown on pl. I). Faults near the west end of Long Mountain, however, made it seem more reasonable to show the Windingstair fault as mapped.

The continuity of Long Mountain syncline across Windingstair fault indicates that the fault has a low net slip at this point. The open, nearly symmetrical nature of the syncline and the high angle of intersection between its trace and that of Windingstair fault suggest that the fault dips steeply at this point (cross section E-E′, pl. I).

Far to the west of Rich Mountain in Rs. 14, 15 E., the dip of the Windingstair fault was inferred by Hendricks (1959, p. 45) to be approximately 20 degrees. Previous work on beds adjoining the Windingstair fault north and east of Talihina, Oklahoma (Seely, 1955), indicated that these beds dip southward at angles between 30 and 50 degrees. If the Windingstair is a bedding-plane fault in this locality, then its dips would also lie within the same range.

The above data give scattered but scant information about dips of the Windingstair fault. These range from 20 to 50 degrees. It may be that this range eventually will be found to apply to the fault throughout most of its exposed extent in the Ouachita Mountains.

DIP OF THE BRIERY FAULT

The trace of the Briery fault is exposed in a roadcut of State Highway 103 on the south border of the Spring Mountain syncline. Only a small part of the fault surface is visible, and this has a dip of approximately 55 degrees south. Inclination of the fault plane is apparently greater north of Black Fork Mountain. Strata on both sides of the fault there dip 60 to 70 degrees southward, suggesting that the fault surface dips similarly.

FAULTS WITH STRIKE-SLIP COMPONENTS

Faults in the report area, interpreted to have a strike-slip component of their net slip, show no prevalent direction of movement.

Both right-lateral and left-lateral strike-slip components are present. This is illustrated by the recess in the trace of the Briery fault south of the Spring Mountain syncline.

The north-south breadth of the Spring Mountain syncline is 2.5 miles at its maximum, but westward it narrows to about 1.5 miles. This narrowing is due to the change in strike of the south limb of the syncline, as the north limb shows no change in strike westward. The westward decrease in breadth implies that the south limb has been pushed northward against the north limb of the fold. Further evidence of this northward movement of the south limb lies in the continued parallelism of current indicators and strike as the strike changes. If one assumes that current direction is constant and that direction changes of current indicators are due to tectonism, then a persistent parallelism between strike and indicator can only be accomplished by rotation of strike about a vertical axis. This rotation was probably produced by the western end of the south limb moving farther northward than parts of the limb farther east.

Strike-slip movement accompanying this crushing of the west end of Spring Mountain syncline is suggested by the large drag fold produced in rocks of Winding Stair Mountain south of the Briery fault. This fold is west of the study area in secs. 21, 22, 27, 28, T. 3 N., R. 25 E. It consists of a syncline-anticline couplet in the northern limb of the major syncline directly west of the Rich Mountain syncline. This couplet indicates right-lateral movement of the Briery fault on the southwestern limb of the Spring Mountain syncline which probably resulted from northward thrusting.

The Eagle Gap syncline lies directly east of the Spring Mountain syncline. At the west end of Black Fork Mountain there is a slight change in strike of the ridge-forming beds. However, the dips of these beds appear to remain as steep as they are farther east. The uniformity in degree of dip despite strike changes suggests that the strike change is caused by drag against the Briery fault rather than by the nearness of these beds to the eastward-plunging axis of the Eagle Gap syncline. If the strike change was caused by drag, the Briery fault probably had a left-lateral component of movement to the west of Black Fork Mountain.

As does the west end, the east end of the Spring Mountain syncline appears to be crushed. One evidence of this is the east-west fault, the trace of which is near the south edge of sec. 15, T. 3 N., R. 26 E. Its upthrown south block appears to have offset the warped axial trace of the Spring Mountain syncline to the east and, in so doing, has formed an S-fold in beds of the north block. This fault joins the Briery fault and probably reflects the same left-lateral movement.

The fact that the Honess and Windingstair fault traces are not parallel to the Briery fault trace south of the Spring Mountain syncline also suggests that dip slip alone cannot explain the recess in the trace of the Briery fault. Honess and Windingstair faults are probably dip-slip faults above which the upper plates have moved northward. If the upper plate of Briery fault has also moved northward, then any segment of it not striking parallel to the Honess and Windingstair faults will have strike-slip components of movement associated with it. The reasonable conclusion is that the net slip of the Briery fault has a right-lateral component on the western side of the recess and a left-lateral component on the eastern side. The two components may be of about the same magnitude, but this cannot be proved.

Another occurrence of strike-slip faulting is apparently due to the crushing of the south flank of the Black Fork syncline. Results of this crushing are visible north and west of Shut-in Mountain in T. 1 N., R. 31 W. Price Creek marks the trace of a left-lateral strike-slip fault which separates the relatively undisturbed ridge formed by the youngest Atoka sediments in the Black Fork syncline from older folded Atoka sediments that have been moved relatively in an east-northeasterly direction. The undisturbed inner ridge of the syncline is known as Horseshoe Mountain (sec. 13, T. 1 N., R. 32 W.), and the folded sediments are present east of Saddle Gap. The folds consist of an anticline and a syncline that are overturned. Their axial surfaces probably dip southwestward at angles not exceeding 50 to 55 degrees. The southwest limb of the syncline shows an abrupt strike change adjacent to the fault trace, a change resulting from drag due to left-lateral movement along the fault (see NE cor. sec. 24, T. 1 N., R. 32 W.). Part of a small drag fold in Horseshoe Mountain may be seen at the edge of the map. This fold also indicates left-lateral strike-slip movement of the fault. The non-parallelism of structures on either side of Price Creek indicates that movement along the fault was essentially strike slip in nature; dip-slip movement was minor.

The fault trace along which Price Creek is developed (Dry Creek to the north in Bates quadrangle, Arkansas-Oklahoma, is also along this trace) probably joins with another along which Clear Fork flows. Clear Fork is east of Price Creek. Each fault appears to have resulted from strike-slip movement. The strike-slip faults are evidence of northeastward and eastward movement of beds making up the south limb of the Black Fork syncline. This movement may have been caused by wedge action of the overthrust sheet of the Briery fault. The salient of the Briery fault to the west of Shut-in Mountain may have acted as a plow in turning blocks of rock aside from its leading edge.

JOHNSON CREEK FAULT

Reinemund and Danilchik (1957) mapped a fault called the Johnson Creek fault. It was considered to approximate a bedding-plane fault that follows the Johns Valley-Atoka contact. Realizing the complexity of the western part of the Black Fork syncline, they did not consider a reliable interpretation of the fault possible until that area was mapped. Using evidence then available to them, they tentatively concluded that the Atoka rocks of the Black Fork syncline had been thrust eastward and northward with respect to adjacent rocks.

As mentioned above, evidence for eastward and northeastward thrusting of Atoka beds in the southwestern exposed part of the Black Fork syncline is abundant. Although this evidence does not confirm Reinemund and Danilchik's original interpretation, it conforms with the direction of thrusting they postulated.

The fault trending east-west through the middle of sec. 27, T. 1 N., R. 31 W., and through the lower middle of sec. 26, T. 1 N., R. 31 W., is on strike with the Johnson Creek fault. Here, however, it is not interpreted as being the Johnson Creek (bedding plane) fault (cross section E-E′, pl. I). It seems more likely that this fault continues eastward between Fourche and Mill Creek Mountains and does not follow the Atoka-Johns Valley contact. Perhaps it is the westward continuation of a fault mapped by Reinemund and Danilchik as branching off the Johnson Creek fault in secs. 26, 27, 28, T. 1 N., R. 30 W.

POSSIBLE ORIGIN OF OUACHITA STRUCTURE

Although this investigation is primarily concerned with the structure of the Rich Mountain area, some aspects of the investigation can be related to the regional Ouachita structure. This discussion is not intended to be a comprehensive treatment of the posssible origins of Ouachita structure, as it is limited to only a few aspects.

Two features which bear upon Ouachita tectonics are found in the Rich Mountain area. These are (1) the strike-slip components of movement shown by some faults and (2) the progressive northward change in the characteristics of the Lynn Mountain, Rich Mountain, and Eagle Gap synclines. The strike-slip components indicate that, in this area, the overthrust masses were moved northward perpendicular to axial traces of the major synclines. The change from the broad, open folding of the eastern end of the Lynn Mountain syncline to incipient imbrication of the Eagle Gap syncline suggests a northward increase in north-south stress components. This northward change is common to the Oklahoma part of the Ouachita salient. Changes in the stratigraphy from the interior to the peripheral parts of the salient also may have resulted in the change of structural style. However, in the report area it is assumed that the stratigraphy remains relatively unchanged in a north-south direction; thus a northward increase in north-south stress components may be postulated.

Implicit in this postulate is the assumption that the forces responsible for the creation of the Ouachita structure acted horizontally and were not due to deformation of the basement rocks which now underlie the structure. Evidence of the horizontal component of the deforming forces is abundant. Beginning with the Boktukola syncline south of the Lynn Mountain syncline and moving northeastward therefrom to include the Lynn Mountain, Rich Mountain, and Eagle Gap synclines, two facts stand out: each is asymmetrical with a southward-dipping axial surface and each has its south flank cut by a major fault with upthrown side to the south. With the exception of the Eagle Gap syncline, these synclines also have similar stratigraphic sections.

Whereas fold asymmetry can be caused by forces acting in a vertical direction, overturned beds and isoclinal folds require force components acting in a horizontal plane. Such beds and folds are present on the south flank of the Rich Mountain syncline and, probably, on the south flanks of the other synclines to the south as well.

If vertical forces, presumably originating in the basement, were directly involved in formation of the Boktukola, Lynn Mountain, Rich Mountain, and Eagle Gap synclines, why is it that the stratigraphic throw of all major faults bordering them on the south is nearly the same? Assuming that these forces have produced vertical movement, is it just a coincidence that all these apparent fault blocks have had the same stratigraphic displacement? Why has the southern block been upthrown in all cases, and why do axial surfaces of the synclines dip southward? These questions are difficult to answer if one assumes that the principal deforming-force components were more nearly vertical than horizontal, but are readily explained if the opposite assumption is applied. It seems reasonable to conclude, therefore, that structures of the Rich Mountain area and its immediate surroundings were caused by horizontal forces acting in a north-south direction and that the magnitudes of these forces increased northward.

The Oklahoma part of the Ouachita salient occurs where the east-west structural trends of Arkansas turn to the north-south trends of northeast Texas. If movement within the salient was everywhere in a direction perpendicular to the trends of major faults and folds, it must have been northwestward near the town of Atoka, Oklahoma, westward in northeastern Texas, and northward in the report area.

Assuming this to be true, it is interesting to speculate on the origin of Ouachita structure. If the deformed rocks of the Ouachitas have been propelled into place by rigid crustal slices overriding one another, the direction of propulsion has been radial. The source of these slices can hardly have been any farther to the southwest than the southwestern extension of the Broken Bow-Benton uplift in northeastern Texas; otherwise they would have arisen from a point source (the point where lines drawn perpendicularly to Ouachita structural trends would intersect).

Hendricks (1959, p. 50) estimated a minimum cumulative displacement of 53 miles along faults north of the Octavia fault in the westernmost Ouachita Mountains. Additional displacement occurs between the Octavia fault and the Broken Bow-Benton uplift. If this total displacement also applies to the Ouachitas in the subsurface of northeast Texas as well as to the exposed salient in Oklahoma a crustal-shortening hypothesis would require radial movement in the crust of this magnitude (crustal shortening is here defined as the distance between two crustal points adjacent to the deformed belt and across its strike before deformation minus their distance after deformation).

It is difficult to visualize a crustal displacement of such magnitude taking place within the radius of the structural arc. One would expect large radial grabens and strike-slip zones caused by the crustal slices moving apart and being independent of one another. And if such radial movement has taken place, what mechanisms of crustal shortening could have caused it?

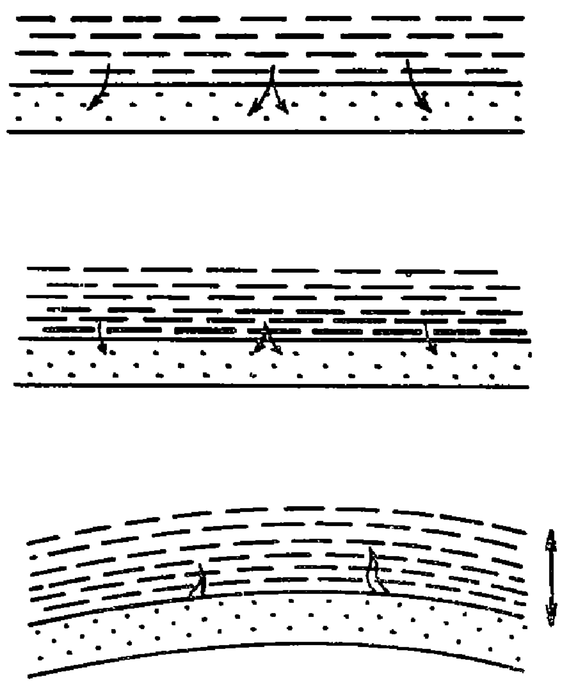
An alternate explanation for the origin of Ouachita structure is that it has resulted from gravitational spreading and/or sliding from the belt of low-grade metamorphics exposed at the surface in the Broken Bow-Benton uplift (and, perhaps, from other metamorphic trends yet to be discovered in the subsurface). The gravitational origin of Ouachita structure has already been suggested by Jacobs and others (1959, p. 300, 315-320) and by Lyons (1961, p. 97). In addition, Ouachita structure has been likened to the Valley and Ridge Province of the Appalachians, and Carey (1962, p. 139-143) has proposed a gravitational origin for that province.

An increase in horizontal stresses toward the outer margin of the Ouachitas is compatible with a gravitational origin of Ouachita structure. Under such an hypothesis this margin is near the foot of a former surface sloping outward from above the rising metamorphic welt. Horizontal stresses would have ranged from a minimum on top of the welt to a maximum at the foot of the slope. Therefore "deformational intensity" should increase toward the margin of the salient. The opposite relationship would be expected if Ouachita rocks had been propelled into place by a "push from the rear."

Under the hypothesis of a gravitational origin for Ouachita structure, the radial nature of movement in Oklahoma and northeast Texas would have been caused by outward slopes conforming to the curvature of the metamorphic welt. The absence of a wide, tectonically denuded strip along the top of the welt could be explained by assuming a dominance of plastic spreading over rigid gliding. This dominance might also explain the lack of radial grabens and strike-slip zones. Fault dips would decrease with increasing depth and displacement would die out downdip.

THE HUBBERT-RUBEY HYPOTHESIS AND SANDSTONE DIKES

The Hubbert-Rubey hypothesis (1959) has many interesting sidelights. Relatively dry shale sections adjacent to sandstone beds should have a greater strength than either the "wet" shales farther away (which have a high interstitial fluid pressure) or the sandstone beds themselves (which have provided avenues for escape of water expelled from adjacent shales). Given sufficient strength, these dry shales would yield to deformation by rupturing. The rupture would be most likely to occur where the shale has the greatest rigidity, which would be where it could most readily get rid of its contained water. This would be in the immediate vicinity of the sandstone bed. As distance from the sandstone aquifer is increased, the rigidity of the shale should decrease and the rupture would close (fig. 52). Thus there would be a gradual increase of rigidity as the sandstone aquifer is approached and then an abrupt decrease across the shale-sandstone interface. For this reason it is not shale that fills the rupture, but sandstone. At an unknown (and probably variable) distance from the aquifers the shale is more mobile than the sandstone because of its higher content of interstitial fluid, but this shale does not fill the rupture in the dry shale because fissures close before reaching the highly mobile shale zone (fig. 52).



Early stage of compaction. Flow rate of water from shale into sand is at its maximum.

Late stage of compaction. Flow of water from shale into sand has nearly stopped.

"Wet" shale yields to deformation by plastic flow.
"Dry" shale yields to deformation by rupture.
Fissures formed during deformation are filled by sand from fluid-filled aquifer.

Figure 52. Hypothetical origin of sandstone dikes.

The sandstone dike (fig. 23) described under the title *Plastic Flow Structures* was probably injected upward into a small fissure. Some sandstone beds transected by the dike maintain a uniform thickness up to the dike margins, whereas others appear to taper near the dike as though they had contributed some of their substance to it. The beds of uniform thickness must have been indurated at the time of intrusion. This induration and the rigidity of the shale suggest that intrusion of the dike took place under considerable overburden.

STRUCTURE DUE TO DOWNSLOPE MOVEMENTS

Sandstone beds in shale sections that are exposed on steep slopes may show marked deformation due to downslope movement. This is most obvious in Atoka strata of the study area, but large sandstone blocks several hundred feet long in the Wildhorse Mountain Formation near the crest and on the south slope of Black Fork Mountain show extensive slumping.

State Highway 103 roadcuts in Atoka beds of the Spring Mountain syncline clearly reveal the effects of downslope movement. Most beds in the south flank of the Spring Mountain syncline are probably nearly vertical at shallow depths, but their surface dips depend mostly upon the steepness and direction of the slope upon which they are exposed, and their position upon that slope. Generally, strata exposed on northward-facing slopes dip southward, whereas beds exposed on southward-facing slopes dip northward; and the steeper the slope, the more gentle the dip.

In roadcuts the effect or downslope movement on attitudes is clearly seen. However, off the road where Atoka exposures are commonly few and far between, care must be taken to separate attitudes due to downslope movement from those representative of true structural attitude. The attitudes least affected by downslope movement are more likely to be found in stream valleys and on hill crests.

A rather unusual possible effect of downslope movement is the pair of low-angle thrust faults visible in the west roadcut of State Highway 103 at the crest of the hill in SE¼ NE¼ sec. 24, T. 3 N., R. 25 E. (fig. 53). These thrusts have an apparent dip of nine degrees south (which probably also approximates the true dip) in the lower portions of the roadcut, but the dip decreases upward to the

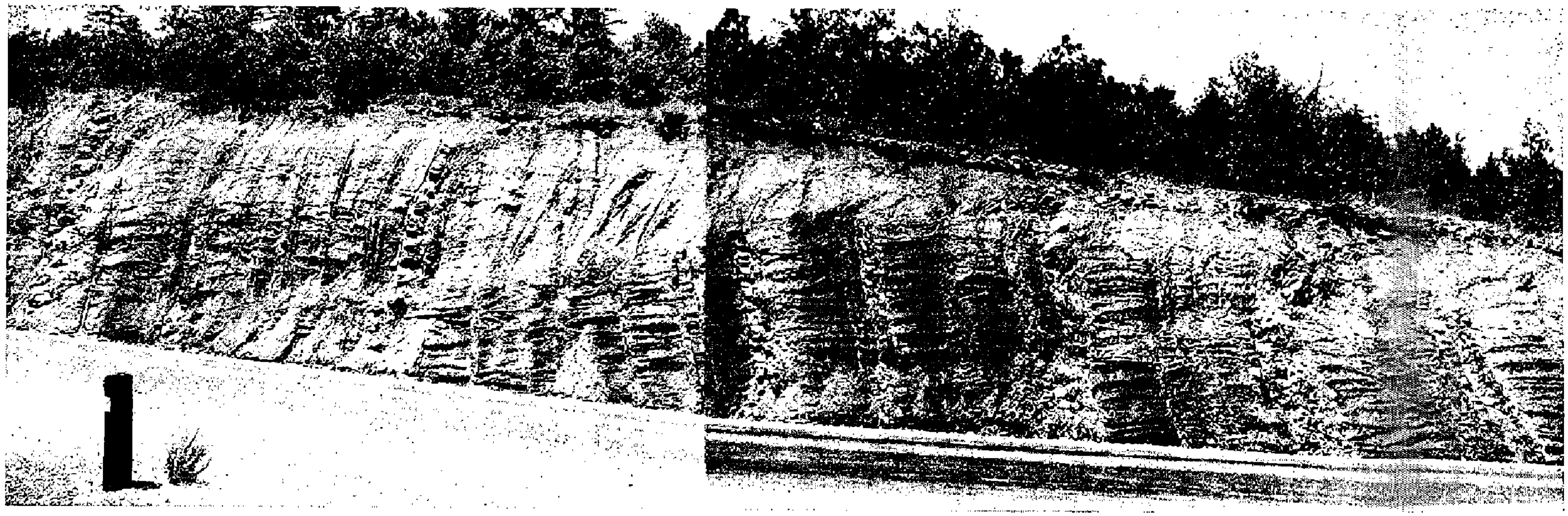


Figure 53. Gravity (?) induced thrust faults. Westward view of Highway State 103 roadcut in east-center sec. 24, T. 3 N., R. 25 E. The trace of one fault begins at the foot of the roadcut above the post in left foreground and displacement of beds cut by it increases up the trace of the fault to the right. Another fault trace begins at the base of the roadcut near the splice of the two photographs and displacement along it also increases up the trace. These faults are discussed in the text under *Structure due to Downslope Movements*.

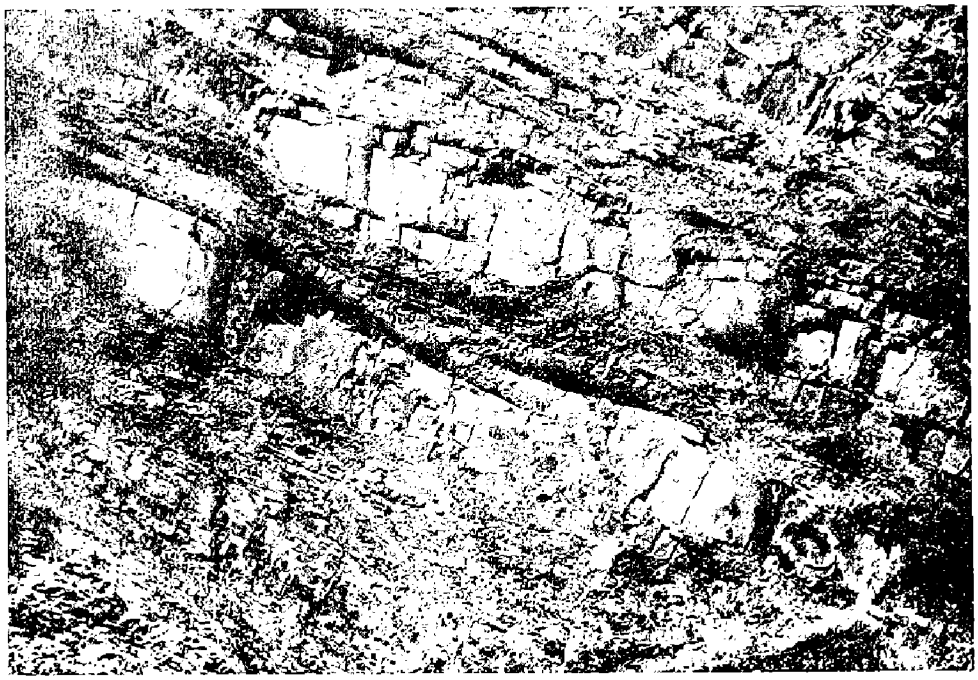


Figure 54. A small reverse fault cutting beds of interval 3, Ward Lake Spillway measured section. A synclinal axis is out of the picture to the north (right) and the displacement by the fault corresponds to that predicted by theory.

Figure 55. Lower Wildhorse Mountain beds of interval 5, East Ward Lake measured section. A normal fault is visible in lower left foreground and similar normal faults of the same bed are visible near the middle right margin. These faults are interpreted as having been caused by movement of beds above the pick to the right relative to beds below. They are "drag faults" in that their origin was probably similar to the origin of drag folds. As in the case of the fault shown in figure 54, the movement may have been caused by synclinal folding in which younger beds towards the center of curvature moved away from the fold axis over older beds.

zone of most active creep. The apparent displacement of beds increases up the dip of the fault plane from a small displacement at the foot of the embankment. Toward the surface, beds of the upper plates become increasingly overturned and bent in the direction of the ground slope. It seems probable that the movement has been caused by the downslope creep of surface beds under the influence of gravity, but it is possible that the apparent updip increase in displacement is due to tectonism.

The two thrusts have at least two possible origins. One is that the faults formed during folding of the Spring Mountain syncline. Faults formed during the folding might be expected to have a southerly dip as have the major faults of the area. The dip, however, is lower than that thought to characterize most Ouachita faults. It contrasts with the steeper dipping fault planes caused by small faults of probable tectonic origin visible in State Highway 103 roadcuts near the crest of Spring Mountain (figs. 56, 57).

Figure 56. Trace of Briery fault in the east roadcut of State Highway 103. The trace slants diagonally across the middle of picture separating strata of the Stanley (right) from strata of the Jackfork (left). Dip of the fault plane is about 55 degrees southward. Jackfork beds of the over-ridden block are nearly vertical and Stanley beds of the over-riding block nearly parallel the fault plane.

A second possible origin of the faults involves the force of gravity. If one extrapolated the rate at which displacement along the faults decreases, he would infer that the faults die out at a shallow depth beneath the base of the roadcut. The first unfaulted bed might have a bend at a position corresponding to the point where the fault surface, if extended, would intersect it. The bend divides beds above, which have moved under the influence of gravity, from beds below, which have been little affected. The attitudes of beds above and below this bend would correspond to those visible above and below the upper thrust trace shown in figure 53.

Figure 57 Small reserve faults cutting lower Atoka sandstones exposed in the east roadcut of State Highway 103 near Spring Mountain summit. The south side (to the right) is upthrown. Compare the dip of these faults with that of the faults shown in figure 53.

SUMMARY AND CONCLUSIONS

This study was undertaken because of discrepancies between existing geologic maps and observed reconnaissance data. Stratigraphical investigations were made to accompany the structural studies.

Stratigraphic units identified in Rich Mountain area are Moyers siliceous shale and Moyers Formation, Chickasaw Creek Formation; Jackfork Group and one of its formations, the Game Refuge Formation; and Johns Valley Shale, which may be locally distinguished from the Atoka Formation. For mapping purposes arbitrary contacts were chosen between (1) the Wildhorse Mountain Formation and the overlying Prairie Mountain-Markham Mill-Wesley sequence, and (2) the basal sandy zone of the Atoka and overlying shaly zone. A tuff bed was found in the Chickasaw Creek Formation—stratigraphically the highest tuff yet reported in the Ouachita Mountains.

Many beds throughout the stratigraphic section have current ripple marks and/or associated small-scale cross-bedding. These features are particularly common in thin beds and near the tops of thick beds. Their occurrence and that of bottom casts do not appear mutually exclusive because both features may be found in the same bed.

Lamination is common in sandstones of the Jackfork Group and of the Johns Valley and Atoka Formations. In many beds, however, it is not obvious except on surfaces etched by weathering or in thin section. Although graded bedding of some type is present in some beds, it is not megascopically obvious in most. Other features represented by various sandstones are ellipsoidal clay galls and their molds, load casts, contorted bedding (uneven surfaces due to squeezing while in a semiplastic state), trails of benthonic organisms, top-surface troughs and ridges, and molds of fragments of invertebrates.

The mineralogy of a small sample of sandstones from the upper part of the Stanley, Jackfork, and the lower part of the Atoka was studied with a polarizing microscope. All sandstones are classified as orthoquartzites, according to Folk's classification. The heavy-mineral suite is simple and consists of zircon, tourmaline, and garnet in order of decreasing abundance. Metamorphic rock fragments and angular feldspar are other significant constituents, also present in small amounts.

The source area (or areas) contributing material to the Jackfork-Atoka sedimentary basin was probably composed of sedimentary and low-rank metamorphic rocks. Its location is not yet known. Little rounding took place in the basin. Volcanoes were also present.

Sandstones of the Jackfork-Atoka sequence have most of the characteristics of turbidites. Graded bedding is not a conspicuous feature, however, and many beds are relatively well sorted. The degree of sorting is probably due to the dynamics of transportation and deposition rather than to sorting of the source materials. The texture, lamination, good sorting, and absence of graded bedding indicate that many sandstones fit better into Lombard's description of laminites than into ten Haaf's description of turbidites and probably were laid down by slow-moving bottom currents.

Paleocurrent studies and regional time correlations are not advanced enough for reliable conclusions about current direction during deposition of beds of the Jackfork-Atoka sequence. Preliminary results in and near the study area suggest a prevailingly westward or west-southwestward flow parallel to tectonic strike.

The Tenmile Creek Formation, Moyers Formation, and Jackfork Group probably were not deposited on a river flood plain, on top-set beds of a delta, on a tidal flat, or in shallow offshore water. They were probably deposited below wave base.

Results of the study indicate that existing geologic maps are incorrect. Two major, previously unmapped, faults cross the area. These faults trend east-west and are herein named Honess and Briery Creek faults. Therefore the fault belt north of the Windingstair fault in Oklahoma extends farther eastward than has heretofore been realized.

Although direct field evidence of the type of fault movement is scanty, this evidence and the geometry of structures associated with the faults suggest that the Honess, Briery Creek, and Windingstair are all thrust faults over most or all of their field-mapped extents.

Dips of the major faults vary to an unknown extent along their traces. Near the west edge of the area the Briery Creek fault has a southerly dip of about 55 degrees, and the Windingstair fault dips southward at an angle probably less than 40 degrees. Eastward the dips of both faults probably increase. The southward dip of the

Horsepen fault is steepest at its outcrop and may decrease to 15 degrees downdip.

It appears that, in the area, thrust masses moved northward perpendicularly to the trends of major faults. Both right- and left-lateral strike-slip components of movement occurred in the overriding masses.

On the basis of present knowledge it is postulated that Ouachita structures originated with simple vertical movements of the basement or with vertical movements accompanying basement thrusting. The resulting gravitatonal gliding or spreading produced the major folds and thrust faults.

REFERENCES

BILLINGS, M. P., 1954, Structural geology: Englewood Cliffs, New Jersey, Prentice-Hall, Inc., 514 p.

BOKMAN, JOHN, 1953, Lithology and petrology of the Stanley and Jackfork Formations: Jour. Geology, vol. 61, p. 152-170.

BRANSON, C. C., 1959, Regional relationships of Ouachita Mississippian and Pennsylvanian rocks, *in* The geology of the Ouachita Mountains, a symposium: Dallas Geol. Soc. and Ardmore Geol. Soc., p. 118-121.

CAREY, S. W., 1962, Folding: Alberta Soc. Petroleum Geologists, Jour., vol. 10, no. 3, p. 95-144.

CLINE, L. M., 1960, Stratigraphy of the Late Paleozoic rocks of the Ouachita Mountains, Oklahoma: Okla. Geol. Survey, Bull. 85, 113 p.

CLINE, L. M., AND MORETTI, F. J., 1956, Two measured sections of Jackfork Group in southeastern Oklahoma: Okla. Geol. Survey, Circ. 41, 20 p.

CLINE, L. M., AND SHELBURNE, O. B., 1959, Late Mississippian-Early Pennsylvanian stratigraphy of the Ouachita Mountains, Oklahoma, *in* The geology of the Ouachita Mountains, a symposium: Dallas Geol. Soc. and Ardmore Geol. Soc., p. 175-207.

ELIAS, M. K., 1956, Upper Mississippian and Lower Pennsylvanian formations of south-central Oklahoma, *in* Petroleum geology of southern Oklahoma, vol. 1: Tulsa, Amer. Assoc. Petroleum Geologists, p. 56-134.

ELIAS, M. K., AND BRANSON, C. C., 1959, Type section of the Caney Shale: Okla. Geol. Survey, Circ. 52, 24 p.

ERICSON, D. B., EWING, MAURICE, WOLLIN, GOESTA, AND HEEZEN, B. C., 1961, Atlantic deep-sea sediment cores: Geol. Soc. America, Bull., vol. 72, p. 193-286.

FLAWN, P. T., 1959, the Ouachita structural belt, *in* The geology of the Ouachita Mountains, a symposium: Dallas Geol. Soc. and Ardmore Geol. Soc., p. 20-29.

FOLK, R. L., 1959, Petrology of sedimentary rocks: Austin, Texas, Hemphill, 154 p.

FRIEDMAN, G. M., 1958, Determination of sieve-size distribution from thin-section data for sedimentary petrological studies: Jour. Geology, vol. 66, p. 394-416.

GODDARD, E. N., chm., and others, 1948, Rock-color chart: Washington, D. C., Natl. Research Council.

GOLDSTEIN, AUGUST, JR., 1959, Petrography of Paleozoic sandstones from the Ouachita Mountains of Oklahoma and Arkansas, *in* The geology of the Ouachita Mountains, a symposium: Dallas Geol. Soc. and Ardmore Geol. Soc., p. 97-109.

HAAF ERNST TEN, 1959a, Graded beds of the northern Apennines; sedimentary structures and directions of supply, diss., Gröningen, Rijksuniv., 102 p.

__________1959b, Properties and occurrence of turbidites, *in* Sedimentology of recent and old sediment, a symposium: Geologie en Mijnbouw, new ser., vol. 21, p. 217-222.

HAM, W. E., 1959, Correlation of pre-Stanley strata in the Arbuckle-Ouachita Mountain region, *in* The geology of the Ouachita Mountains, a symposium: Dallas Geol. Soc. and Ardmore Geol. Soc., p. 71-85.

HARLTON, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains: Amer. Assoc. Petroleum Geologists, Bull., vol. 22, 852-914.

HENDRICKS, T. A., 1959, Structure of the frontal belt of the Ouachita Mountains, *in* The geology of the Ouachita Mountains, a symposium: Dallas Geol. Soc. and Ardmore Geol. Soc., p. 44-56.

HENDRICKS, T. A., GARDNER, L. S., KNECHTEL, M. M., AND AVERITT, PAUL, 1947, Geology of the western part of the Ouachita Mountains of Oklahoma: U. S. Geol. Survey, Oil and Gas Inv., Prelim. Map OM 66.

HONESS, C. W., 1923, Part I, Stratigraphy, structure, and physiographic history, *in* Geology of the southern Ouachita Mountains of Oklahoma: Okla. Geol. Survey, Bull. 32, 278 p.

__________1924, Geology of southern Le Flore and northwestern McCurtain Counties, Oklahoma: Bur. Geology [Okla.], Circ. 3, 22 p.

HUBBERT, M. K., AND RUBEY, W. W., 1959, Role of fluid pressure in mechanics of overthrust faulting: Geol. Soc. America, Bull., vol. 70, p. 115-206.

JACOBS, J. A., RUSSELL, R. D., AND WILSON, J. T., 1959, Physics and geology: New York, McGraw-Hill Book Co., Inc., 424 p.

JOHNSON, R. H., 1954, Geology of the Medicine Springs area, Pushmataha County, Oklahoma: Okla., Univ., unpublished Master of Science thesis.

KING, P. B., 1959, The evolution of North America: Princeton, New Jersey, Princeton University Press, 189 p.

KUENEN, PH. H., 1953, Graded bedding with observations on the lower Paleozoic rocks of Britain: K. Nederlandsche Akademie van Wetenschappen, Verh., Afd. Natuurkunde, vol. 20, no. 3, 47 p.

__________1957, Longitudinal filling of oblong sedimentary basins: K. Nederlandsche Geologisch Mijnbouwkundig Genootschap, Verh., Geologische serie, vol. 18, p. 189-195.

LYONS, P. L., 1961 [1962], Geophysical background of Arkoma basin tectonics, *in* Symposium on the Arkoma basin: Tulsa Geol. Soc., Digest, vol. 29, p. 94-104.

McKEE, E. D., 1957, Primary structures in some recent sediments:

Amer. Assoc. Petroleum Geologists, Bull., vol. 41, p. 1704-1747.

MISER, H. D., 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas: Okla. Geol. Survey, Bull. 50, 30 p.

________1934, Carboniferous rocks of Ouachita Mountains: Amer. Assoc. Petroleum Geologists, Bull., vol. 18, p. 971-1009.

________1954, Geologic map of Oklahoma: Okla. Geol. Survey and U. S. Geol. Survey.

________1959, Structure and vein quartz of the Ouachita Mountains of Oklahoma and Arkansas, *in* The geology of the Ouachita Mountains, a symposium: Dallas Geol. Soc. and Ardmore Geol. Soc., p. 30-39.

MISER, H. D., AND HONESS, C. W., 1927, Age relations of the Carboniferous rocks of the Ouachita Mountains of Oklahoma and Arkansas: Okla. Geol. Survey, Bull. 44, 28 p.

MORETTI, F. J., 1958, Petrographic study of the sandstones of the Jackfork Group, Ouachita Mountains, southeastern Oklahoma: Wis., Univ., unpublished doctoral dissertation.

PETTIJOHN, F. J., 1957, Sedimentary rocks: New York, Harper and Brothers, 718 p.

PITT, W. D., 1955, Geology of the core of the Ouachita Mountains of Oklahoma: Okla. Geol. Survey, Circ. 34, 34 p.

REINEMUND, J. H., AND DANILCHIK, WALTER, 1957, Preliminary geologic map of the Waldron quadrangle and adjacent areas, Scott County, Arkansas: U. S. Geol. Survey, Oil and Gas Inv., Map OM 192.

SCULL, B. J., GLOVER, G. D., AND PLANALP, R. N., 1959, The Atoka of the McAlester Basin-Arkansas Valley region, *in* The geology of the Ouachita Mountains, a symposium: Dallas Geol. Soc. and Ardmore Geol. Soc., p. 166-172.

SEELY, D. R., 1955, Geology of the Talihina area, Latimer, Le Flore, and Pushmataha Counties, Oklahoma: Okla., Univ., unpublished Master of Science thesis.

SHELBURNE, O. B., JR., 1960, Geology of the Boktukola syncline, southeastern Oklahoma: Okla. Geol. Survey, Bull. 88, 84 p.

SITTER, L. U. DE, 1956, Structural geology: New York, McGraw-Hill Book Co., Inc., 552 p.

TAFF, J. A., 1901, Description of the Coalgate quadrangle [Indian Terr.]: U. S. Geol. Survey, Geol. Atlas, Folio 74.

________1902, Description of the Atoka quadrangle [Indian Terr]: U. S. Geol. Survey, Geol. Atlas, Folio 79.

TAFF, J. A., AND ADAMS, G. I., 1900 Geology of the eastern Choctaw coal field, Indian Territory: U. S. Geol. Survey, 21st Ann. Rept., pt. 2, p. 263-311.

ULRICH, E. O., 1927, Fossiliferous boulders in the Ouachita "Caney" shale and the age of the shale containing them: Okla. Geol. Survey, Bull. 45, 48 p.

APPENDICES

APPENDIX A

History of Nomenclature of Stratigraphic Units

Stanley Group

The Stanley Group was first defined by Taff (1902, p. 4): "The formation takes its name from the village of Standley, in the Kiamitia Valley, where it is extensively exposed." In the intervening years, the definition has been modified, the name of the village has been changed to Stanley, and the valley has come to be known as the Kiamichi Valley.

Harlton published a paper (Harlton, 1938) upon which subsequent work involving beds of Stanley or younger age has been based. Harlton raised the Stanley to group status and subdivided it into three formations, using siliceous shale beds as the stratigraphic boundaries of these formations. In ascending order these formations are the Tenmile Creek, Moyers, and Chickasaw Creek. The type localities for these units are along the southwestern edge of the Ouachita Mountains in Oklahoma, specifically the flanks of the Tuskahoma and Round Prairie synclines.

Taff originally placed the Stanley-Jackfork contact several hundred feet below the top of the Chickasaw Creek Siliceous Shale (Harlton, 1938). Honess (1923, p. 173-174) tried to place the contact in a position in agreement with Taff but had difficulty in finding it. According to Harlton, however, the Stanley-Jackfork contact is at the top of the Chickasaw Creek. Hendricks and others (1947), like Taff and Honess, placed the Chickasaw Creek in the Jackfork Group. Students working under the direction of J. K. Arbenz at The University of Oklahoma in 1954 and 1955 followed Harlton in placing the contact at the top of the Chickasaw Creek. L. M. Cline, who has made an extensive study of the stratigraphy of the western Ouachitas, also followed Harlton (Cline, 1960, p. 24).

Jackfork Group

The type locality of the Jackfork Sandstone, Jackfork Mountain, about 15 miles northwest of Clayton, Oklahoma, was described by Taff (1902, p. 4). Location of exposures on Jackfork Mountain is responsible for its name.

It has been only in recent years, however, that the Jackfork has become a consistent, relatively well-defined stratigraphic unit. In terms of present nomenclature, Honess (1924, p. 10-12) included the uppermost Stanley, Jackfork, Johns Valley, and lower Atoka in a unit he called the Jackfork Sandstone. Harlton (1938, p. 878-889) redefined the Jackfork but was forced to use two type sections which he miscorrelated (Cline, 1960, p. 42). Cline restudied Harlton's type sections, modifying Harlton's description and definition of the unit.

As originally proposed by Harlton and modified by Cline, the Jackfork is considered to be a group comprising the Wildhorse Mountain, Prairie Mountain, Markham Mill, Wesley, and Game Refuge Formations in ascending order. Like the Stanley, the Jackfork is subdivided principally by means of siliceous shales. The beds are useful as markers in the type areas about the Tuskahoma and Round Prairie synclines and in nearby regions. However, workers in other areas of the Ouachitas have had difficulty in finding some of them (Cline, 1960, p. 53).

Johns Valley Shale

The name Johns Valley Shale was first applied by Ulrich (1927, p. 21, 22) to beds cropping out in the center of the Tuskahoma syncline near the southwestern margin of the Ouachita Mountains in Oklahoma. In this location is a bowl-shaped depression called Johns Valley, so named for a stream draining it, now known as Johns Creek. At the time that Taff named and defined many of the stratigraphic units, this stream was known as Cane Creek, and Taff gave the name Caney Shale (Taff, 1901) to the same strata later renamed Johns Valley Shale by Ulrich.

Subsequent to Taff's original definition the name Caney Shale was widely applied to rocks of both the Arbuckle and Ouachita facies as Taff had apparently intended it to be (Cline, 1960, p. 60). However, when Ulrich renamed beds of Taff's Caney shale cropping out in the Ouachita Mountains, use of the name Caney became restricted to beds of the Arbuckle facies. Thus arose the rather confusing situation in which the type locality of a prominent unit present in many of Oklahoma's oil-producing provinces and associated with the Arbuckle facies, was found in a supposedly contrasting suite of rocks called the Ouachita facies. To compound the confusion, Ulrich maintained that the strata exposed at the type locality are not correlative to beds of the Arbuckle facies named "Caney" by most geologists. When Ulrich renamed the formation Johns Valley Shale, the peculiar situation arose in which two stratigraphic units were defined by the same type locality.

In recent years the work of Cline has done much to eliminate this confusion. The name Johns Valley Shale is retained as defined by Ulrich. Arbuckle facies correlative to the Johns Valley Shale are the Caney Shale at its base, the Goddard Formation in its middle and upper portions, and at least the lower part of Wapanucka Formation at its top (Cline 1960, p. 85). Thus defined, the Caney Shale is not a unit of the Ouachita facies, but is restricted to strata equivalent to lower beds of the original shale named by Taff.

The problem of defining the Caney Shale has been attacked again by Elias (1956) and Elias and Branson (1959). A type section has been chosen and is augmented by three additional localities. According to these workers, the Johns Valley Shale, in its lower portion, contains a probable Ahloso equivalent, definite Delaware Creek equivalent, and a possible Sand Branch equivalent (Elias and Branson, 1959, p. 22). These are the three members of the Caney Shale, in ascending order, as they are exposed in the northern Arbuckle Mountains. Thus, these geologists appear to agree with Cline in considering the lower part of the Johns Valley as an equivalent of the Caney of their definition in the Arbuckle facies.

Atoka Formation

The Atoka Formation was named for exposures near the town of Atoka, Oklahoma, by Taff and Adams (1900, p. 273). Honess (1924) mapped beds he considered to be a shoreward phase of the Atoka as Jackfork. However, in 1927 Miser and Honess (p. 21) noted that these strata are Atokan in age and should therefore be called Atoka. They were so mapped on the 1926 edition of the geologic map of Oklahoma edited by Miser.

Branson (1959, p. 121) questioned the presence of Atokan beds in the Ouachita province. Of first order of importance in determining whether this is true is definition of the Atoka itself. Taff's original description of the Atoka is vague and exposures of the unit near the town are inadequate. The Oklahoma Geological Survey is working on a new, carefully described type area.

Further complicating matters in the Ouachita Mountains is the absence of a Johns Valley-Atoka contact at the type locality of the Johns Valley Shale. Some of the localities described under *Noteworthy Johns Valley Outcrops* by Cline (1960, p. 66-82) would be more suitable type sections because there the Johns Valley Shale, including the upper and lower contacts, is well exposed. If the contact is drawn at the top of a shale section, then the overlying sandstones may or may not be Morrowan in age. With present knowledge, it seems wise to place the sandstones in the Atoka Formation as they have been in the past, but, in so doing, one should realize that they may be of Morrowan age.

APPENDIX B

Thin-Section Compositions*

(Estimated percentages of each component)

Thin Section	Quartz	Feldspar	Chert and/or Clay	Metamorphic Rock Fragments	Chalcedony	Zircon	Tourmaline	Garnet	Biotite	Chlorite	Colorless Mica	Glauconite	Calcite	Carbonized Plant Matter	Ilmenite, Magnetite, or Leucoxene	Pyrite	Spicules	Unidentifiable Microfossils
RM 40	96	..	..	3	..	Tr	..	..	..	Tr	Tr	0.5	..	Tr	Tr	..	..	..
RM 29A	95	..	3	2	..	..	..	..	..	..	..	..	..	..	..	..	..	..
RM 29	95	Tr	4	Tr	..	Tr	Tr	..	Tr	..	Tr	Tr	..	Tr	Tr	..	..	..
RM 24A	99	..	Tr	Tr	Tr	Tr	..	..	Tr	..	Tr	..	..	Tr	Tr	..	..	..
RM 23A	..	..	x	..	..	..	..	..	..	..	..	..	..	x	..	..	x	x
RM 22A-1	1	..	99	..	..	..	..	..	..	..	..	..	..	Tr	..	..	..	..
RM 21A	93	0.5	2	2	..	Tr	Tr	Tr	Tr	..	0.5	..	..	2	Tr	..	..	..
RM 20	90	Tr	9	Tr	Tr	Tr	Tr	..	Tr	..	Tr	..	..	Tr	Tr	..	..	..
RM 19	98	Tr	1	Tr	Tr	Tr	Tr	..	Tr	..	Tr	..	..	Tr	..	..	..	..
RM 14	98	..	1	0.5	..	Tr	..	..	..	..	Tr	..	..	Tr	Tr	..	..	..
RM 11A	98	Tr	1	Tr	Tr	Tr	..	..	Tr	..	Tr	..	..	..	..	..	..	..
RM 11	98	..	1	..	..	Tr	..	..	..	..	Tr	..	..	Tr	Tr	..	..	..
RM 8-2	97	..	1	1	..	Tr	..	..	Tr	..	..	..	..	..	..	..	..	..
RM 7	99	..	Tr	..	..	Tr	..	..	Tr	..	..	..	..	Tr	Tr	..	..	..
RM 6-1	99	Tr	Tr	0.5	..	Tr	Tr	..	Tr	..	Tr	..	..	Tr	Tr	..	..	..
RM 2-1	95	4	Tr	..	..	Tr	..	..	Tr	..	Tr	..	..	..	Tr	..	..	..
MRS 7-3	98	..	1	Tr	..	Tr	..	..	..	..	Tr	..	..	..	Tr	..	..	..
MRS 6-1	97	Tr	2	Tr	..	..	..	..	..	..	..	..	..	Tr	Tr	..	..	..
MRS 4-2	93	0.5	5	..	..	Tr	Tr	Tr	Tr	Tr	Tr	..	Tr	Tr	Tr	Tr	..	..
MR 6-1	84	0.5	10	1	Tr	Tr	..	..	Tr	..	1	..	Tr	Tr	..	3	..	..
MR 5-2	97	1	0.5	1	..	Tr	..	..	..	..	Tr	..	..	..	Tr	..	..	..
MR 4-2	1	..	96†	..	Tr	..	..	..	..	..	..	..	..	2	..	..	1†	x
MR 3-1	98	Tr	Tr	1	..	Tr	..	..	..	..	Tr	..	..	Tr	..	..	..	..

* For composition of thin section 7-20-5F see appendix G.
x Indicates component is present in indeterminate amounts.
† Figure based upon recognizable fraction. A high proportion of the slide may consist of unrecognizable organic debris.

APPENDIX C

THIN-SECTION TEXTURES

Thin Section	Sorting* ϕ Thin Section	Sorting* ϕ Sieving Equivalent	Mean* ϕ Thin Section	Mean* ϕ Sieving Equivalent	Textural Classification† Based on Uncorrected Thin Section Data	Textural Classification† Corrected to Sieving Equivalent
RM 2-1	0.35	0.30	3.30	3.35	Well- to very well-sorted very fine sandstone	Very well-sorted very fine sandstone
MRS 7-3	0.43	0.40	1.80	2.00	Well-sorted medium sandstone	Well-sorted medium to fine sandstone
MRS 6-1	0.45	0.40	3.25	3.30	Well-sorted very fine sandstone	Well-sorted very fine sandstone
	0.45	0.40	2.45	2.60	Well-sorted fine sandstone	Well-sorted fine sandstone
	0.50	0.45	1.80	2.05	Slightly granular, moderately to well-sorted medium sandstone	Slightly granular, well-sorted fine sandstone
MRS 4-2	0.45	0.40	2.25	2.40	Well-sorted fine sandstone	Well-sorted fine sandstone
	0.35	0.35	3.65	3.65	Well- to very well-sorted very fine sandstone	Well- to very well-sorted very fine sandstone
	0.45	0.40	2.55	2.70	Well-sorted fine sandstone	Well-sorted fine sandstone
MR 6-1	0.40	0.35	2.70	2.85	Well-sorted fine sandstone	Well- to very well-sorted fine sandstone
MR 5-2	0.65	0.60	2.65	2.80	Moderately sorted fine sandstone	Moderately sorted fine sandstone
MR 3-1	0.40	0.35	2.40	2.55	Well-sorted fine sandstone	Well- to very well-sorted fine sandstone
RM 40	0.40	0.35	2.70	2.85	Well-sorted fine sandstone	Well- to very well-sorted fine sandstone
RM 29A	0.70	0.60	0.70	1.00	Slightly granular, moderately sorted coarse sandstone	Slightly granular, moderately sorted medium to coarse sandstone
RM 29	0.30	0.30	3.30	3.40	Very well-sorted very fine sandstone	Very well-sorted very fine sandstone
RM 24A	0.30	0.30	3.00	3.10	Very well-sorted fine to very fine sandstone	Very well-sorted very fine sandstone
RM 21A	0.30	0.30	3.00	3.10	Very well-sorted fine to very fine sandstone	Very well-sorted very fine sandstone
RM 20	0.50	0.45	2.80	2.95	Moderately to well-sorted fine sandstone	Well-sorted fine sandstone
RM 19	0.50	0.45	1.50	1.75	Moderately to well-sorted medium sandstone	Well-sorted medium sandstone
RM 14	0.60	0.50	2.10	2.30	Moderately sorted fine sandstone	Moderately to well-sorted fine sandstone
RM 11A	0.45	0.40	2.55	2.70	Well-sorted fine sandstone	Well-sorted fine sandstone
RM 11	0.20	0.18	3.80	3.80	Very well-sorted very fine sandstone	Very well-sorted very fine sandstone
	0.25	N. A.	4.85	N. A.	Very well-sorted coarse siltstone	Very well-sorted coarse siltstone
RM 8-2	0.20	0.18	2.80	2.90	Very well-sorted fine sandstone	Very well-sorted fine sandstone
RM 7	0.90	0.85	2.10	2.25	Moderately sorted fine sandstone	Moderately sorted fine sandstone
RM 6-1	0.30	0.30	2.80	2.90	Very well-sorted fine sandstone	Very well-sorted fine sandstone

* Sorting is the graphic standard deviation of Folk (1959, p. 44); the mean is as computed by Inman (Folk, 1959, p. 44).
† Classification based upon Folk (1959, p. 103).
N.A. Conversion method not applicable.

APPENDIX D

Sandstone Classification

(According to Folk, 1959, p. 136)

Thin Section	Sandstone Classification
RM 40	Fine sandstone: highly siliceous, mature or supermature, glauconitic orthoquartzite
RM 29A	Slightly granular coarse sandstone: siliceous, submature, fossiliferous orthoquartzite
RM 29	Very fine sandstone: chert-cemented, mature, glauconitic orthoquartzite
RM 24A	Very fine sandstone: highly siliceous, mature or supermature orthoquartzite
RM 21A	Very fine sandstone: siliceous, chert-cemented*, mature or supermature orthoquartzite
RM 20	Fine sandstone: chert-cemented, supermature orthoquartzite
RM 19	Medium sandstone: highly siliceous, supermature orthoquartzite
RM 14	Fine sandstone: siliceous, chert-cemented*, submature orthoquartzite
RM 11A	Fine sandstone: highly siliceous, mature or supermature orthoquartzite
RM 11	Very fine sandstone: siliceous, chert-cemented*, mature or supermature orthoquartzite
	Coarse siltstone: siliceous, chert-cemented*, mature or supermature orthoquartzite
RM 8-2	Fine sandstone: highly siliceous, mature or supermature orthoquartzite
RM 7	Fine sandstone: siliceous, submature orthoquartzite
RM 6-1	Fine sandstone: highly siliceous, mature or supermature orthoquartzite
RM 2-1	Very fine sandstone: siliceous, chert-cemented*, mature or supermature orthoquartzite
MRS 7-3	Medium sandstone: siliceous, supermature (chert-bearing) orthoquartzite
MRS 6-1	Slightly granular medium sandstone: siliceous, chert-cemented*, mature or supermature orthoquartzite
	Fine sandstone: siliceous, chert-cemented*, mature or supermature orthoquartzite
	Very fine sandstone: siliceous, chert-cemented*, mature or supermature orthoquartzite
MRS 4-2	Fine sandstone: siliceous, supermature orthoquartzite
	Very fine sandstone: chert-cemented, mature or supermature orthoquartzite
	Fine sandstone: chert-cemented, mature or supermature orthoquartzite
MR 6-1	Fine sandstone: chert-cemented, mature or supermature, pyritic orthoquartzite
MR 5-2	Fine sandstone: chert-cemented, submature orthoquartzite
MR 3-1	Fine sandstone: highly siliceous, supermature orthoquartzite

* "Siliceous, chert-cemented" is the description used when a quartz overgrowth mosaic and a shaly chert matrix are both present in the same thin section.

APPENDIX E

Cross-Refference Index for Thin Sections and Measured Sections

Thin Section	*Stratigraphic Position*	
	Measured Section	Interval
RM 40	Rich Mountain	87
RM 29A	Rich Mountain	72
RM 29	Rich Mountain	72
RM 24A	Rich Mountain	61
RM 23A	Rich Mountain	59
RM 22A-1	Rich Mountain	58
RM 21A	Rich Mountain	57
RM 20	Rich Mountain	56
RM 19	Rich Mountain	55
RM 14	Rich Mountain	44
RM 11A	Rich Mountain	38
RM 11	Rich Mountain	38
RM 8-2	Rich Mountain	25
RM 7	Rich Mountain	21
RM 6-1	Rich Mountain	20
RM 2-1	Rich Mountain	12
MRS 7-3	Ward Lake Spillway	7
MRS 6-1	Ward Lake Spillway	6
MRS 4-2	Ward Lake Spillway	4
MR 6-1	East Ward Lake	6
MR 5-2	East Ward Lake	5
MR 4-2	East Ward Lake	4
MR 3-1	East Ward Lake	3
7-20-5F	Collection locality: NE SW 25, T. 1 N., R. 32 W.	

APPENDIX F

Descriptions of Sandstones from which Thin Sections Were Obtained

RM 40: From a planar-laminated, fine sandstone that breaks along laminae leaving flat surfaces with many muscovite flakes.

RM 29A: From the top of a 14-inch bed of apparently massive, hard, fossiliferous, granule conglomerate that is deeply iron stained along fracture surfaces and in cavities from which fossil fragments have been removed by solution.

RM 29: From the base of a two-inch, ripple-marked and cross-bedded, very hard sandstone, and including the upper part of a half-inch black chert layer. The sandstone and chert are tightly welded to one another and thus appear as a single physical unit.

RM 24A: From a few inches above the base of a light-gray to white, very fine-grained, very hard, faintly laminated, 12-inch sandstone bed.

RM 21A: From a one-inch, highly carbonaceous, plastically deformed sandstone float fragment. The carbonaceous material appears vitreous in some of the thicker (approximately 1 mm) seams. Similar thicker sandstone fragments are nearby.

RM 20: From the basal portion of a hard, light-gray, two-inch sandstone that is planar laminated with thin layers of shale in the lower half and cross-bedded in the upper half. Vertical distance from top set to bottom set is three-fourths of an inch. The only laminae well enough developed to cause the rock to split along their surface are near the base of the bed. The bed occurs with others of similar thickness and character in a shale section.

RM 19: From the upper part of a hard block of sandstone float that is 14 inches thick. The upper surface of the boulder is pocketed with an intricate maze of depressions having a nonuniform shape. These depressions appear to be caused by the weathering out of formerly clay-filled cavities. Embedded, smaller clay blebs aligned with the bedding are present in the upper one inch of the block. Below this one-inch zone is another more profuse concentration of clay pebbles and cobbles. These clay pebbles cause the upper one-inch zone to break off the block as a slab. The thin section comes from just below this second (lower) zone.

RM 14: From hard, light-gray sandstone. The rock is apparently massive except for faint, dark-gray, curving laminations that are concave upwards. The bed from which it was obtained is probably more than one foot thick as are most of the sandstone beds of the interval, but this information was not recorded.

RM 11A: From a massive, (except for faint laminae as in RM 14) light- gray sandstone bed that is about 12 inches thick. The thin section was obtained about three inches from the top of the bed. The upper half-inch of the bed is poorly laminated by abundant plant fragments up to 1 cm in length concentrated along curving inclined surfaces. Smaller fragments are present in zones of lower concentration throughout the rock.

RM 11: From the base of a 2½-inch sandstone bed that is planar laminated below and becomes cross-bedded upwards. The cross-bedding locally is contorted. The upper surface has tracks of invertebrates and the lower surface has small-scale, nonoriented bottom casts.

RM 8-2: From the base of a five-inch, hard, gray sandstone. Planar laminae that are spaced one to two mm apart are present in the lower four inches of the bed. The upper one inch has cross-laminae with a one-fourth-inch top-set to bottom-set distance. (An undescribed thin section, RM 8-1, was constructed from this.) Groove casts protrude one-eighth inch below the base of the bed and the surface displays low-relief ripple marks.

RM 7: From a massive, light-gray, hard, 30-foot sandstone bed containing scattered white specks. Position in bed not recorded.

RM 6-1: From the base of a two-inch, hard, olive-gray sandstone that possesses faint planar lamination in the upper one inch and has groove casts protruding one-sixteenth inch below the base. (An undescribed thin section, RM 6-2, was cut from the top of this bed.)

RM 2-1: From a 1½-inch, hard, faintly cross-bedded sandstone. Where the rock has broken along the undulating laminae, there is a rough, pitted surface reflecting stylolitic development.

MRS 7-3: From a medium-gray, hard, massive sandstone whose fractures contain drusy quartz and halloysite. The thickness of the bed was not recorded, but most beds of the interval are one-half to one foot thick in zones three to ten feet thick.

MRS 6-1: From a hard, massive, discordant mass embedded in a friable, two-inch, cross-bedded siltstone bed that is underlain and overlain by shale. The mass measured two inches by five inches in the plane of the outcrop and appeared to taper into the hillside. Cross-bedding of the siltstone is abruptly cut off by the mass and is not apparent within it. The lower half of the mass contains abundant quartz granules floating in a fine-sand matrix.

MRS 4-2: From a very-hard two-inch sandstone that is massive except for a color change from dark gray to light gray in the middle of the bed. White specks are profuse.

MR 6-1: From a two- by six-inch, very-hard, medium-light-gray sandstone disc embedded in a thick, gray shale section.

Plant imprints are present on the surfaces of other similar discs. Some show small-scale cross-bedding. All appear similar to semicontinuous thin sandstone beds also present.

MR 5-2: From a medium-hard, massive sandstone of unrecorded thickness.

MR 4-2: From a fractured, three-inch, siliceous shale bed with irregular light- and dark-gray laminae. White specks in the bed average approximately 0.1 mm in diameter.

MR 3-1: From a hard, light-gray sandstone that is faintly laminated by closely spaced (1- to 2-mm spacing) planar concentrations of dark material.

APPENDIX G

Supplemental Remarks about Thin Sections

RM 40: The occurrence of glauconite in this slide is noteworthy.

RM 29A: This thin section was obtained from a porous rock containing molds of fragments of invertebrate fossils. It contains quartz grains, some of which are granule in size. Most grains are subangular to rounded and do not show well-developed rounding; several show straining under crossed nicols; a few are rounded, composite, stretched metamorphic grains (see Folk, 1959, p. 69).

RM 29: Small-scale cross-bedding is clearly shown by this slide. The cross laminations are defined by fragments of siliceous shale or shaly chert that is of the same composition as the siliceous layer at the base of the slide. The layer and fragments contain circular patches of quartz or chalcedony of possible organic origin. There is a concentration of zircon grains and a black opaque heavy mineral along the contact between the cross-bedded sandstone and the underlying siliceous layer. Glauconite is present.

RM 24A: A mosaic texture due to extensive recrystallization is apparent in this thin section. This is associated with good sorting and stratification that is only faintly defined by linearly scattered leucoxene grains.

RM 23A: The relative amount of carbonized plant matter in this slide is difficult to estimate because the fragments are finely divided and dense. It is probable that the carbonized material makes up a minor proportion of the rock even though it may largely cause its dark color. There are many spicules and other fragments of organic origin, including an abundant form that may be a radiolarian. It is possible that most of the rock consists of similar organic debris that is very finely crystalline. The thin section is laminated by the carbonized plant matter; many of the laminations are spaced 0.1 mm apart.

RM 22A-1: Although this slide is also dominantly composed of chert, it is distinctly different from 23A in several ways: lamination is practically absent; spicular or other organic material is not recognizable; a thin silt layer is present. The few scattered cavities have been filled with finely crystalline silica. The crystals increase in size towards the center of one of the larger cavities.

RM 21A: The plagioclase in this slide probably is andesine. Orthoclase is also present. The biotite present is pleochroic in shades of gray and is probably the variety, lepidomelane.

RM 20: Cross-bedding in this slide is defined by the heavy minerals, zircon, leucoxene, garnet, and tourmaline. A few quartz grains have a recrystallized outline that is euhedral in part.

RM 8-1 and RM 8-2: These two slides were cut from the top and bottom of the same bed (see Appendix F). Except for a change from planar lamination at the base to cross-bedding at the top, no difference in textural or compositional properties was noted.

RM 6-1 and RM 6-2: Silica has replaced the carbonate of a crinoid columnal and shreds of plant matter have been compressed into the boundary areas between grains in RM 6-1. RM 6-2 was cut from the top of the same bed (see Appendix F). Except for the stylolitic development of RM 6-2, the thin sections appear identical.

MRS 6-1: The texture of the rock represented by this thin section has a "dumped" character where large, rounded granules of stretched metamorphic quartz grains (see Folk, 1959, p. 69) are present.

MRS 4-2: This slide shows distinct size stratification.

MR 4-2: Mostly cryptocrystalline and finely microcrystalline silica with embedded clay flakes compose this slide. Varying concentrations of carbonized, macerated plant matter cause laminations that are generally greater than 0.05 mm and less than 1 mm thick. Holes in the slide are probably due to the weathering of large grains, some of which were apparently angular. These grains may have been the white specks that are megascopically observable.

MR 3-1: Lamination is shown on this slide by concentrations of alternating grain sizes.

7-20-5F: Details of this thin section are not listed in the other appendices.

Composition:

- 13% Calcite
- 10% Quartz
- 75% Matrix of microcrystalline quartz and clay minerals

1% Volcanic rock fragments
1% Andesine (An_{45})
Tr Chlorite
Tr Biotite
Tr Muscovite
Tr Microcline
Tr Zircon
Tr Garnet
Tr Leucoxene
Tr Metamorphic rock fragments
Tr Carbonized plant fragments

Texture:

Microcrystalline matrix with embedded pyroclasts and normal sedimentary debris. Perlitic and bogen structures are present.

Calcite is replacing andesine and the volcanic rock fragments. All andesine grains are angular and fresh. Quartz fragments are angular to well rounded and some are embayed. Their estimated average diameter is 0.3 mm. The clay mineral flakes in the matrix are too small for positive identification. They are pale green and may be a species of chlorite.

APPENDIX H

Notes on Measured Sections

General

Rock types in each interval are listed in order of relative thickness; for example, ''Sandstone, light-gray and gray shale'' indicates that sandstone is dominant. The word minor added to the description indicates that the rock type comprises less than 10 percent of the interval.

The colors used are those of the *Rock-Color Chart* (Goddard and others, 1948) except that a generalized term is substituted if the precise color could not be determined: for example, ''gray shale'' means that the shale could be any or several of the gray values of the chart. Used in this sense the generalized term designates the hue. Many more distinct colors are present in the intervals than are listed in the chart. The colors stated are intended to describe unweathered rock surfaces; however, it was difficult to determine megascopically the relative freshness of the surfaces.

An arbitrary hardness scale is used: friable, firm, medium hard, hard, and very hard, in order of increasing hardness. Friable rocks may be crumbled in the hand while very hard rocks have a subconchoidal fracture. The intermediate grades of hardness fall between these two extremes. The Wentworth size classification is used. Sorting terminology is that of Folk (1959).

Interval numbers are given for cross reference to the columnar section (pl. II). Where desired, cumulative thicknesses may be determined from the columnar section. At several points on the columnar section and inset map, control locations such as U. S. Geol. Survey unchecked elevation (U. E.) stations are noted. These stations are identified by white paint on road-side rocks.

Rich Mountain Measured Section

This and the following sections were described during the summer, 1959. Road improvements associated with the development of Wilhelmina State Park have subsequently provided more and better exposures of the lower 26 intervals of the Rich Mountain measured section.

Thicknesses were computed from measurements taken by brunton compass and steel tape. Strike is nearly east-west throughout the entire section. From a glance at the inset map on plate II, it should be apparent that many measurements were made nearly parallel to strike and therefore that the possibility of error in thickness determinations was increased. In order to minimize the cumulative error, corrections have been made to fit thicknesses indicated by cross sections.

The base of Interval 1 is 650 feet inside the Wilhelmina State Park eastern boundary sign on the entrance road from the town of Rich Mountain. All descriptions are based upon rocks exposed along this road and Skyline Drive. Nearly all exposures are poor, the poorest being indicated on the columnar section. The shales are exposed only where they are interbedded with resistant sandstones. Here they comprise a minor part of the section, thus giving an erroneous impression as to their relative abundance in the total stratigraphic interval.

Ward Lake Spillway Measured Section

Strata of this section are well exposed in the spillway of Ward Lake. The lake provides water for Mena, Arkansas, and is in NW¼ SW¼ sec. 6, T. 2 S., R. 30 W.

The base of the measured interval is below the lowest beds of the continuously exposed sequence in the spillway walls. This is at the lower end of the spillway. The easily identified siliceous shales of the Chickasaw Creek Formation crop out near the base of the section and also may be used to relate the descriptions to field exposures.

East Ward Lake Measured Section

Outcrops of beds described in this section are south of the east end of the Ward Lake dam beside the short road between Skyline Drive and the purification plant. Siliceous shale zones of the Chickasaw Creek Formation provide a convenient reference location point. They and the other beds described occur in excellent exposures.

Interval	*Thickness in feet*	
Top of Rich Mountain measured section in the Atoka Formation		
88	128	Sandstone, light-gray and gray shale; firm to friable, massive to faintly planar-laminated sandstones between 1 and 3 feet thick at the base and top of the interval; hard, cross-bedded and ripple-marked, 3- to 6-inch sandstones with casts on bottom surface, interbedded with fissile shale and thin siltstones averaging less than one-half inch thick in the middle of the interval. The top of the measured section is drawn at the top of this interval because of the absence of exposures above and the beginning of structural complexity. Poorly exposed shale and friable sandstone directly overlie this interval.
87	123	Sandstone, light-gray and gray shale; hard, fine-grained relatively well-sorted and clean, massive to faintly planar-laminated, 2- to 3-foot sandstones mostly at the top of the interval; hard, fine- to very fine-grained, moderately sorted, slightly micaceous, ripple-marked and cross- to contorted-bedded or planar-laminated and cleavable into plates, ½- to 6-inch sandstones interbedded with fissile shale and thin siltstones in the lower portion of the interval. Also in the lower portion are a few friable to firm, micaceous, massive sandstones that weather to olive-brown or olive-gray shades.
86	126	Covered. The road and road-cuts are sandy near the base of the interval and shaly near the top. The sandy basal half may be due to weathering of sandstone float, in which case the entire interval would consist of shale.
85	97	Shale, poorly exposed.
84	89	Sandstone, very light-gray and minor medium-gray and very light-gray shale; firm to medium-hard, moderately sorted and somewhat argillaceous, micaceous, carbonaceous (with rare coal in lenses up to ½ inch thick), wavy and cross-bedded sandstone in beds that average 6 to 12 inches in thickness, although a few are up to 30 inches thick; very light-gray shale that is plastic when wet; medium-dark-gray silty shale and thin siltstone with much iron oxide impregnation. A few of the firm sandstone beds display only olive-brown and olive-gray colors that are assumed to be the result of weathering. These appear more poorly sorted, argillaceous and micaceous than the lighter colored beds.

Interval	Thickness in feet	
83	219	Covered except for a 20-foot section exposed near the middle of the interval. This midportion consists of very light-gray, friable to firm, moderately sorted argillaceous massive sandstone that is deeply iron stained and has clay galls; very light-gray, hard to very hard, cross-bedded and ripple-marked thin (2- to 4-inch) sandstone with prominent nonoriented bottom casts; very light-gray, plastic shale; silty shale and interbedded thin siltstone. The shale has abundant disseminated carbonized plant fragments.
82	21	Slumped section. No beds in place. Float consists of thin (1- to 4-inch), hard, cross-bedded and ripple-marked sandstone with one 1-foot bed. The thicker beds have linear, current-formed bottom casts modified by nonoriented casts, but the thinner beds have only the nonoriented type. The regolith consists of sandy clay.
81	108	Covered.
80	69	Sandstone, very light-gray to white and light-gray shale; sandstone is similar to that of the lower portion of Interval 79, and possesses ripple marks with discontinuous curved crests in no discernable pattern, clay gall molds on top surfaces, and raised limonite crusts; shale is very poorly exposed. Sandstones directly to the south of the road have scattered molds of invertebrate fragments including crinoids and bryozoans.
79	81	Sandstone, very light-gray to white and minor gray shale; hard to very hard, cross-bedded and ripple-marked, 2- to 12-inch sandstones with nonoriented bottom casts and prevalent in the basal portion of the interval; hard to very hard, apparently fairly well sorted, fine-grained, massive, thick (beds up to 10 feet thick) sandstone in the upper part of the interval; light-gray poorly exposed shale interbedded with the thin basal sandstones. Slumpage has overturned the upper portion of this interval.
78	68	Covered. Topographic expression suggests this interval is underlain by relatively soft beds.
77	45	Shale, light-gray and very light-gray to white sandstone; shale, interbedded thin siltstones and a few thin, ripple-marked and cross-bedded sandstones with profuse nonoriented bottom casts, all in the upper portion of the interval; very hard to medium-hard, very fine- to medium-grained, fairly clean

Interval	*Thickness in feet*	
		to argillaceous sandstones that are 1 foot and less in thickness and prevalent in the lower portion of the interval. The laminae of some of the lower sandstones are so well formed that they display a sheaflike appearance upon weathering, the sheaves consisting of thin sandtone plates. Where not as well developed the laminae may cause the sandstone to be cleavable. These laminae consist of clay, mica, carbonaceous material and, in a few beds, molds of crinoid columnals. They are planar near the base and wavy- to cross-bedded near the top of several beds.
76	46	Sandstone, very light-gray and shale; sandstone similar to that at the base of Interval 77 and in which small (less than 2 mm) clay galls and a suggestion of graded bedding were observed; shale presence inferred. The lower half of this interval is very poorly exposed.
75	18	Sandstone, very light-gray to white and minor dark-gray to grayish-black shale; hard, very fine- to fine-grained, fairly well sorted, laminated sandstone with current-formed bottom casts; fissile to splintery shale with interbedded thin siltstone. Spacing of the laminae in the sandstone is roughly proportional to the thickness of the bed. The thinner beds may have laminae spaced less than 1 mm apart whereas the thicker beds may have a spacing of 8 inches. Wavy- and cross-bedding are more apparent in beds less than 6 inches thick. Planar lamination causes fissility in the thicker beds.
74	63	Shale, dark-gray to grayish-black and minor light-gray sandstone; fissile to splintery shale and silty shale with thin siltstone interbeds; rare grayish-black, cherty shale that weathers very light-gray; hard to very hard, very fine-grained, planar-, cross-, and contorted-bedded sandstones averaging a foot and less in thickness. In the upper middle portion of the interval is a 2½-foot sandstone bed that is underlain by a thin (6-inch), very fine-grained, planar-laminated bed which, in turn, is underlain by a wavy-bedded zone about 3 inches thick. Below this is a 1-inch, medium-grained sandstone succeeded by a 1½-inch, coarse-grained sandstone, both containing abundant invertebrate molds, especially molds of crinoid columnals. Continuing downward, shale becomes the dominant constituent and the occurrence of invertebrate molds terminates a short distance below the sand-shale contact.

Interval	*Thickness in feet*	
73	30	Sandstone, very light-gray to white; very hard, very fine-grained, planar- to wavy-laminated sandstone in the lower part of the unit; friable to hard, fine- to coarse-grained, massive to cross-bedded sandstone in the upper part. In the upper middle portion of the interval is a 1½-foot bed with a crumbly, limonitic, quartz-granule conglomerate zone in the lower 2 inches that contains an invertebrate mold fauna. Grain size decreases upward so that the uppermost layers are fine grained and also cross-bedded.
72	43	Shale grayish-black and minor gray sandstone; splintery shale interbedded with 3- to 6-inch, medium-hard to very hard, very fine-grained, laminated sandstones,several of which have bottom casts. At the base of the interval is a 14-inch, massive bed of quartz-pebble conglomerate that contains invertebrate molds and is deeply iron stained. A thin, grayish-black, siliceous siltstone is present near the lower fossiliferous sandstone. Associated with this is a 3-inch bed consisting of cross-bedded, very fine-grained sandstone in the upper 2½ inches and black chert in the lower ½ inch. Near the top of the interval is a 2-inch, quartz-granule conglomerate also containing invertebrate molds.
71	575	Shale, dark-gray to grayish-black and light-gray sandstone; shale and thin siltstone interbedded with 1- to 3-inch, cross-bedded sandstones. Most of the interval is either covered or poorly exposed. The base is marked by a fine- to coarse-grained, fossiliferous sandstone that may not be in place. A few discoid-shaped, contorted, squeezed, carbonaceous sandstone masses are present in the road bank. In float at the top of the interval is a thin, grayish-black, siliceous siltstone that may have come from the overlying interval.
70	248	Shale and thin sandstone are suggested by the regolith and topographic expression of this interval. Most of the interval is covered, but at the base is a small exposure of a 1½-foot, massive sandstone bed that is overlain by gray shale.
69	99	Shale, gray and minor thin siltstone and sandstone; olive-gray to medium-gray, soft, splintery to poorly fissile to massive shale dominates the interval; dark-gray to olive-gray, fissile, slightly siliceous shale is present in a 2-foot zone near the base of the interval; light-gray, thin (less than 6 inches thick),

Interval	*Thickness in feet*	
		hard to very hard, cross-bedded and ripple-marked sandstones with current-generated bottom casts are also present in the lower portion of the interval. The sandstones are deeply iron stained. A few discoid-shaped, highly carbonaceous sandstone masses with contorted laminae and irregular surfaces due to plastic squeezing were noted in the float near the top of the interval.
68	377	Mostly covered. One hundred fifty feet above the base is a poor exposure of a 12-foot section consisting of 1- to 3-inch, firm to hard, very fine-grained, cross-bedded sandstone; thin siltstone; an unknown amount of shale. Near the top of the interval is a 10-foot zone consisting of light-gray, hard, fine-grained, 3- to 12-inch sandstones. The thicker beds are planar laminated at the base and wavy laminated at the top, but the thinner beds tend to be wavy or cross laminated. Carbonized plant fragments are the major constituent of the laminae and are especially abundant near the tops of beds where the laminae are most closely spaced.
67	331	Shale, light-grayish-white, light-brownish-gray, light-yellowish-brown, and light-whitish-gray siltstone; massive shale that weathers into layers averaging $\frac{1}{8}$- to $\frac{1}{4}$-inch in thickness; thin and lenticular, micaceous, brittle siltstones, some of which display wavy- or cross-bedding. These form a float of thin plates and rodlike blocks. The shale and siltstone is thoroughly iron-oxide impregnated thus suggesting that the colors listed above may be largely due to weathering. Carbonized plant fragments are present in the shale. The foregoing description applies to well-exposed beds in the upper half of the interval and is inferred to apply to the very poorly exposed lower half. One mile to the east of the road along strike are good exposures of dark-gray to black, splintery to sub-platy to massive shale and spaced, thin (mostly less than 2-inch) siltstones.
66	239	Mostly covered. About 50 feet from the top of the interval and appearing as float elsewhere are very light-gray, hard, very fine-grained, 1- to 3-inch, cross-bedded and ripple-marked sandstones with nonoriented bottom casts and grooves possibly left by invertebrates on the top surface. Some fragments have small well-defined escarpments which offset bottom casts, but do not terminate them.

Interval	*Thickness in feet*	
		In float near the base of the interval are lens-shaped, 6-inch, iron-oxide-encrusted discs and similarly shaped sandstones with moderately contorted surfaces.
65	396	Covered. An estimated 25 to 30 feet of erosion-resistant sandstone occurs at the top of the interval and is poorly exposed on adjacent hill slopes. This consists of light-gray to white, very hard, very fine-grained, well-sorted, clean, massive, ripple-marked, ½ to 1½-foot sandstones and light-gray to white, medium-hard to hard, micaceous, cross-bedded sandstones with carbonaceous laminae and containing clay galls.
64	20	Sandstone, light-gray, very fine-grained, clean, apparently well-sorted, very hard to hard; in beds greater than ½ foot thick that are planar laminated at the base and cross-bedded at the top. Carbonized plant fragments and mica are most abundant along the closely spaced cross-laminae at the tops of the beds. Ripple marks are prominent on upper surfaces.
63	60	Covered. Topographic expression suggests that this interval is underlain by easily eroded beds. Float of sandstone containing mold fauna probably from Interval 61 or 62 found in this interval.
Top of Game Refuge Sandstone and Jackfork Group		
62	6.5	Sandstone, light-gray; similar to that of Interval 61. This interval is distinctive because of the ridges on the upper surface of a 2-foot, massive bed present at the top of the interval. Their irregular arrangement and shape are suggestive of organic origin.
61	86	Sandstone, light-gray to white and gray shale; very hard, very fine-grained, well-sorted, clean sandstones that are generally cross-bedded and ripple marked where beds are less than ½ foot thick and planar laminated at the base to cross-bedded at the top where beds are greater than ½ foot thick; poorly exposed gray shale that is concentrated in the lower midportion of the interval. On the under side of some of the thin, cross-bedded strata are casts of trails made by some organism, possibly worms. These trails wander randomly across the lower bedding surface, are ¼ to ½ inch wide, and the casts consist of low mounds that are transverse-

Interval	*Thickness in feet*	
		ly offset. The wider trails have a slightly depressed axial channel.
60	247	Shale, gray and minor very light-gray sandstone. The interval is poorly exposed.
59	172	Sandstone, very light-gray and minor gray shale, hard to very hard, very fine-grained, planar-, wavy-, and cross-bedded sandstone weathering to shades of grayish-orange to dark-yellowish-orange to brownish-black, fissile, splintery or siliceous shale that is iron-oxide impregnated. The contact with underlying Wesley shales is gradational and is arbitrarily drawn at the base of the lowest sandstone. Most of the sandstones are less than 1½ feet thick and possess laminae formed by carbonized plant matter. In the upper midportion of the interval is a medium-dark-gray to dark-gray, fissile, siliceous shale that is less than 3 feet thick, and it contains white material as laminae and specks. Its float consists of straight-sided, polygonal slabs that are a fraction of an inch to 2 inches thick and up to 8 inches on a side.
Top of undifferentiated Wesley, Markham Mill, and Prairie Mountain Formations		
58	547	Shale, dark-gray to grayish-black and minor light-gray to white sandstone; fissile, splintery, blocky or siliceous shale and silty shale; hard, very fine-grained sandstone containing clay galls and carbonized plant fragments. Most of this interval is poorly exposed. Two subdiscoidal black cherty masses were noted, one about 175 feet below the top of the interval and the other near the top. These are 6 inches in maximum diameter and have a thin white surface coating. The few sandstones in the interval are a foot or less thick and occur as beds or discoidal masses. The discoidal masses have laminae of carbonaceous material that may be either planar or highly contorted, the geometry of the laminae seemingly having no relationship to the presence or absence of a contorted outer surface. Limonite occurs in masses that are also discoidal, are less than a foot in maximum diameter, and that possess an internal structure of concentric shells.
57	448	Shale and minor thin sandstone; similar to Interval 58, but without the black, cherty shale or limonite masses; poorly exposed. Sandstones at the top of the interval are observable in an isolated exposure to the west of the road. These are light gray and

Interval	Thickness in feet	
		fine grained with laminae consisting of clay flakes and abundant carbonaceous material. Imprints of plant fragments are present on the tops of some beds. Discoidal sandstone masses also exposed to the west of the road are highly carbonaceous and have coaly seams. Sandstone blocks with sinuous, tubular grooves on their upper surfaces are in float near the interval base.
56	192	Shale, dark-gray and minor light-gray sandstone; fissile shale and hard to very hard, fine- to very fine-grained, well-sorted, clean, faintly planar- to cross-bedded sandstones that are less than 10 inches thick. Some beds have nonoriented bottom casts and/or tubular grooves 1/16 to 1/8 inch in diameter and an inch or less in length on upper bedding surfaces. Also on the upper surfaces are trails consisting of a central furrow and marginal arcuate mounds. Most sandstone is in the upper third of the interval and the remainder of the interval is poorly exposed.
55	236	Covered; topographic low: sandy regolith.
54	44	Sandstone, very light-gray, friable to hard, fine-grained, moderately sorted, carbonaceous; containing shale chips. Beds are 6 feet and less thick, and a few have subellipsoidal depressions in their upper surfaces.
53	113	Sandstone, medium light-gray to white and dark-gray shale; friable to very hard, medium-grained, moderately sorted, clean, carbonaceous sandstone in beds less than 6 feet thick; shale in a 6-inch zone just above the poorly exposed basal quarter of the interval. Several of the sandstone beds that are less than ½ foot thick are ripple marked and cross-bedded, as are the upper few inches of some thicker beds. Subellipsoidal depressions are present on the upper surfaces of several beds and a few similarly shaped cavities are present within the beds. These are up to a few inches in maximum diameter and appear to be the molds of large clay galls.
52	177	Covered.
51 to 46		Northward reentrant in Rich Mountain; description of beds exposed included in other intervals.
Approximate top of Wildhorse Mountain Formation		
45	67	Sandstone, light-gray to white and gray shale; friable to hard, fine- to medium-grained, carbonaceous, massive to cross-bedded sandstones, some

Interval	*Thickness in feet*	
		of which have bottom marks and/or a concentration of carbonized plant fragments and shale chips in a thin laminated zone underlying their upper surface; poorly exposed shale making up less than 1/3 of the interval.
44	47	Sandstone, light-gray to white, medium-hard to hard, fine- to medium-grained, massive; in beds 1 to 4 feet thick.
43	29	Covered.
42	20	Sandstone, white, friable to firm, medium- to fine-grained, clean, massive.
41	134	Covered; abundant as float are 1- to 6-inch sandstone plates and blocks with prominent nonoriented bottom marks. These consist of casts of invertebrate tracks and sinuous cylindrical ridges about $\frac{1}{8}$ inch in diameter, parts of which have transverse segmentations spaced at 1/16-inch intervals.
40	37	Sandstone, white; interbedded with soft covered strata that make up an estimated 25 percent of the interval; hard to very hard, fine-grained sandstone in beds up to 10 feet thick near the base and less than a foot near the top. Upper surface of some beds exhibits imprints of plant fragments, up to 8 inches in length, and subellipsoidal depressions that are probably the molds of clay galls. Carbonized plant material is concentrated in the uppermost inch or so of several beds. Current-ripple cross-bedding is present on several of these beds and is particularly noticeable on those that are relatively thin. Uniformly distributed, small-scale pitting apparently due to stylolitic development is observable on bedding surfaces.
39	25	Covered.
38	14.5	Sandstone, very light-gray and gray shale; hard to very hard, fine-grained sandstone whose thicker beds are generally massive and whose thinner beds are wavy-, cross-, and contorted-bedded; poorly exposed shale that weathers light gray and composes less than 50 percent of the interval. Nonoriented bottom casts are present.
37	95	Covered.
36	155	Sandstone, very light-gray; interbedded with soft covered strata that make up an estimated 50 percent of the interval; friable to hard, fine- to medium-grained sandstone in beds from $\frac{1}{2}$ foot to greater than 3 feet thick. There is an apparent

Interval	*Thickness in feet*	
		correlation between grain size and hardness, the finer-grained beds being harder.
35	86	Covered.
34	40	Sandstone, white, friable to hard, fine- to medium-grained; fairly well sorted with the exception of a moderately sorted bed; massive and containing a small amount of disseminated carbonaceous material. Fractures in the friable beds are strikingly delineated in the outcrop by concentrations of yellow and reddish-brown oxidation coloration which stand out from the white nonfractured interareas. Subellipsoidal molds were noted.
33	86	Covered.
32	5	Sandstone, white, friable to firm, fine- to very fine-grained; with limonitic staining.
31	8.5	Covered.
30	55	Sandstone, white to medium-gray, friable to hard, very fine-grained; clean or containing abundant disseminated black (carbonaceous?) material, exhibiting cross-bedding in some of the thinner beds; poorly exposed.
29	41	Covered.
28	84	Sandstone, medium dark-gray to white, fine grained; hard and friable zones alternating; well sorted with sorting especially noticeable in the lighter colored beds, containing disseminated carbonaceous material especially in the poorly cemented beds.
27	560	Covered. This is the estimated thickness of covered beds underlying vegetative cover along Skyline Drive on the ridge west of Wilhelmina Inn.
26	261	Covered. The top of this interval is in the middle of the Wilhelmina Inn turn-off from Skyline Drive on the crest of the ridge.
25	110	Shale, grayish-black and olive-gray sandstone; fissile shale that is poorly exposed and that, together with other erosion susceptible beds in an unknown proportion, makes up an estimated 70 percent of the interval; hard, very fine-grained, cross-bedded, ripple-marked, sole-marked sandstone.
24	23.5	Covered.
23	5	Sandstone, olive-gray, very hard, very fine-grained, massive; a 3-foot bed overlain by a 2-foot bed.
22	6	Covered.
21	31	Sandstone, light-gray, very hard, very fine to fine-grained, moderately sorted; a single massive bed.

Interval	*Thickness in feet*	
20	36	Shale, medium dark-gray and olive-gray sandstone; fissile, silty shale with intercalated medium dark-gray, thin siltstone and white to light-gray, very fine-grained sandstone in beds less than 2 inches thick and containing abundant wavy laminae of black carbonaceous matter; very hard, fine-grained, sole-marked sandstone containing planar laminae of carbonaceous material and in beds 2 to 12 inches thick. Estimated total percentages of the lithologies are: shale and siltstone, 60 percent; sandstone less than 2 inches thick, 10 percent; sandstone from 2 to 12 inches thick, 30 percent.
19	42	Sandstone, yellowish-gray, medium-hard, medium-to coarse-grained, moderately sorted; containing a zone of quartz granule conglomerate about 7 feet above the base.
18	221	Covered. There are a few limited exposures of sandstone similar to that of Interval 17.
17	27	Shale, medium-gray and medium dark-gray sandstone; fissile shale with intercalated siltstone and very fine-grained sandstone in beds about ¼ inch thick; very hard, carbonaceous sandstone in beds averaging 6 inches in thickness and possessing nonoriented bottom casts.
16	36	Sandstone, light-brownish-gray and minor medium-gray siltstone; hard to very hard, very fine- to fine-grained, moderately sorted sandstone containing disseminated carbonaceous specks and rounded quartz grains up to 1 mm diameter; fissile, sandy siltstone comprising an estimated 10 percent of the interval.
15	198	Covered.
14	96	Sandstone, very light-gray to white and medium-gray shale; hard to very hard, very fine-grained, well-sorted sandstone with basal surfaces displaying excellent oriented and nonoriented bottom casts; fissile shale comprising less than 5 percent of the interval. Seams, ½ inch and less in thickness, of carbonized plant fragments are present in the sandstone.
13	88	Mostly covered. This interval underlies a topographic bench and, therefore, probably contains a high proportion of shale. A small exposure of medium dark-gray, splintery and flaky shale was noted. Float of medium-gray, very fine-grained, wavy- and cross-laminated, thin sandstone is present.

Interval	Thickness in feet	
12	19.5	Sandstone, medium light-gray and medium dark-gray shale; medium hard to very hard, fine-grained sandstone with stylolitic parting surfaces and in beds less than 4 feet thick; fissile shale. Concentrations of clay, mica, and carbonaceous material are present as laminae in the thinner sandstone beds and as thin zones at the tops of these beds.
11	44	Covered.
10	16.5	Shale, gray; containing discoidal masses of siltstone and very fine sandstone occurring at distinct horizons.
9	334	Covered.
8	13	Shale, gray and light-gray sandstone; poorly exposed shale and very hard, fine-grained, well-sorted, clean sandstone.
7	13	Sandstone, light-gray, medium-hard to hard, medium-grained, moderately sorted; in beds 1½ to 4 feet thick.
6	84	Covered.
5	9	Shale, gray and light-gray sandstone; splintery shale making up about 85 percent of the interval and very hard, fine-grained, well-sorted, thin (1 inch to 6 inches thick) sandstone with distinct wavy laminae of carbonaceous material. A 6-inch bed exhibits flute casts.
4	8.5	Sandstone, light-gray, medium-grained, moderately sorted, micaceous; containing abundant carbonized plant fragments.
3	20	Covered.
2	18	Shale, gray and gray sandstone; splintery shale with interbedded, fine-grained, thin sandstone making up 70 percent of the interval; medium-grained, moderately sorted, hard sandstone in beds ½ to 1½ feet thick making up the remainder. Carbonized plant fragments are abundant in both the sandstone and the shale.
1	29	Sandstone, light-gray, medium-grained, moderately sorted with relatively large, rounded quartz grains evident, massive. White specks are prominent.

Base of Rich Mountain measured section in the lower Wildhorse Mountain Formation.

Top of Ward Lake Spillway measured section in the lower Wildhorse Mountain Formation of the Jackfork Group

Interval	Thickness in feet	
9	12	Sandstone, medium dark-gray to dark-gray and minor medium-gray shale; very hard, very fine- to

Interval	Thickness in feet	
		medium-grained, moderately sorted, massive sandstone cut by quartz-lined joints and in beds 1 to 3 feet thick; shale and very light-gray, friable, medium-grained, micaceous, carbonaceous, white-speckled sandstone make up the remaining 10 percent of the interval. Also present is a small amount of light-gray-weathering, crumbly clay shale that is plastic when wet. Stylolitic seams are present between several of the hard sandstones.
8	28	Sandstone, medium- to dark-gray or pinkish-gray and grayish-orange siltstone; medium-hard, very fine-grained, fairly well-sorted, cross-bedded, ripple-marked, 2- to 10-inch sandstones alternating with pinkish-gray, firm, medium-grained, moderately sorted (containing many frosted, rounded, coarse, sand-size quartz grains), massive sandstones containing an abundance of disseminated, carbonized plant fragments. The grain size of each of the latter beds appears to decrease upward, grading ultimately into thin intervals of gray shale. Most of these shales are directly overlain by sandstones. The foregoing lithologies compose the lower half of this interval. The upper half is made up mostly of grayish-orange to olive-gray, friable, shaly, carbonaceous, micaceous siltstone also containing frosted, rounded, quartz grains. Included in the siltstone are very hard sandstones similar to the medium-hard sandstones described above; however, these sandstones occur as isolated masses, as well as in continuous beds, and display intersecting escarpments on their surfaces.
7	44	Sandstone, white to very light-gray, friable to firm, micaceous, massive, or medium-gray, hard, white-speckled, even-bedded to ripple-marked; containing stylolitic seams. All beds are moderately sorted: most are medium grained although some are fine grained and several are a coarse-sand conglomerate. The strata are ½ foot to 4 feet thick and are cut by many drusy-quartz-lined joints. Enveloping the quartz grains in several of the joints is halloysite, first observed by Dr. C. J. Mankin on a visit to the area.
6	44	Shale, medium-gray to dark-gray and minor very light-gray to white siltstone; well-bedded shale that has float consisting of flakes about 1/16 inch thick; firm to medium-hard, thin, wavy bedded siltstones intercalated in the shale. A 2-inch siltstone bed near the top of the interval contains a 2-

Interval	*Thickness in feet*	
		inch by 5-inch by ? mass of hard sandstone that is similar in general appearance to the sandstones of Interval 7, but separated from them by a thin shale section. The lower portion of this mass has a "dumped" texture with well-rounded quartz granules, clay galls, and "coaly" grains embedded in a fine-grained matrix. The coarse fraction terminates rather abruptly against a sloping contact and the fine-grained matrix is alone in the upper portion. Interval 6 is separated from Interval 5 by a fault contact.
5	48	Sandstone, medium-gray to medium dark-gray, hard, fine-grained, moderately sorted; containing many coarse-sand-sized quartz grains; a few beds are ripple marked. Carbonized plant matter is profuse throughout the interval and a portion of a *Calamites* stem about 4 inches long and averaging ¾ inch in diameter was found in place (pl. X). Also abundant in the sandstones are subellipsoidal masses of clay or carbonaceous siltstone which are aligned both parallel to and across bedding surfaces. These are 2 inches to 8 inches in maximum dimension. Less than 5 percent of the interval is made up of dark-gray shale separating the sandstone beds. Quartz-filled joints are numerous.
Top of Chickasaw Creek Shale and Stanley Group		
4	34	Shale, light-gray to grayish-black, light-gray siltstone and sandstone; moderately siliceous to highly siliceous shale and silty shale containing abundant white specks, most of which appear to be less than 0.1 mm in diameter but a few with diameters up to about 0.5 mm. The two most siliceous zones are 1 foot and 2 feet thick respectively, and in these zones beds are 1 inch to 6 inches thick, either dark gray with light-gray streaking or light gray with dark-gray streaking, dissected by limonite-stained joints which make it difficult to break the shale in order to obtain a fresh surface, and have planar upper and lower surfaces that cause float to be polygonal plates and blocks bounded by former joint surfaces. The less siliceous shales of the interval are light gray to grayish black to olive gray, platy and thinner bedded. These have intercalations of hard to very hard, micaceous, carbonaceous, shaly, cross-bedded siltstones and very fine-grained sandstones. Some of the sandstones change upward from dark gray and hard to light gray and friable. If the faults present in the interval are ignored,

Interval	Thickness in feet	
		the transition from the Chickasaw Creek to overlying beds of the Jackfork appears gradational and takes place by an increase in the sandstone-shale ratio.
Top of Moyers Formation		
3	18	Sandstone, medium light-gray, and medium light-gray to dark-gray shale; firm to medium-hard, very fine-grained sandstone containing white and black specks and in beds averaging a foot and less in thickness; platy, brittle shales and gray siltstones breaking into plates less than ¼ inch thick and an inch in maximum dimension. Some of the thicker sandstone beds grade upward into medium-gray to light-gray, friable, carbonaceous, micaceous, shaly siltstone which, in turn, grades into silty shale and then into shale. The base of the overlying sandstone lies clearly defined upon the shale. Small up-to-the-north reverse faults cut beds of this interval.
Base of Ward Lake Spillway measured section in the Moyers Formation of the Stanley Group		
Top of East Ward Lake measured section in the lower Wildhorse Mountain Formation		
6	47	Shale, medium-gray to olive-gray with minor medium light-gray siltstone and sandstone; flaky shale cut by numerous joints and small faults whose surfaces are limonite-stained; thin siltstone beds less than 1 inch thick; very hard, massive to cross-bedded, pyritic sandstone in discontinuous beds and discs up to 3 inches thick. The uppermost ½ inch of the cross-bedded sandstones is medium hard, micaceous, dirty, and very fine grained. Carbonized plant fragments are present in several of the finer grained sandstones and siltstones.
5	88	Sandstone, medium-gray and medium-gray shale; friable to very hard, thick (up to 3 feet) and massive to faintly laminated or thin (a few inches) and cross-bedded, very fine- to medium-grained sandstone with prominent flute and groove casts especially on the lower beds; poorly exposed, platy shale and silty shale. Quartz veins and drusy-quartz-lined joints are numerous. At the top of the interval directly underlying the shales of Interval 6 is a 3-inch medium-hard sandstone bed with contorted upward protrusions into the shale. These have dimensions of a few feet perpendicular to the bedding and up to 7 feet parallel to the bedding. They contain abundant plant fragments and at

Interval	*Thickness in feet*	
		least one contains the molds of many crinoid columnals.
Top of Chickasaw Crek Shale and Stanley Group		
4	70	Shale, dark-gray and minor light- to medium-gray sandstone and siltstone; slightly to moderately siliceous shale with very light-gray streaking and white specks averaging a small fraction of a millimeter in diameter; firm to hard, very fine- to medium-grained, massive to cross-bedded, carbonaceous sandstone in beds 1 inch to 2 feet thick; firm to medium-hard, even- to cross-bedded siltstone in beds ½ inch to 2 inches thick. Quartz veinlets and drusy-quartz-lined joints are common. At the top of the interval thin siltsone and very fine-grained, cross-bedded sandstones make up a gradually greater part of the stratigraphic section and grade into the basal Jackfork, where they dominate. This interval is on the margin of a fault zone (see Interval 3) so that many small faults, joints, and folds are present. Limonite has stained the joint surfaces and filled other joints. Locally where joints have been filled the shale has been eroded leaving a raised meshwork of limonite ridges outlining polygonal cavities.
Top of undifferentiated Stanley Shale		
3	12	Fault zone. The thickness of the interval ranges from about 2 feet to 15 feet in this exposure. The zone is characterized by an absence of uniform bedding planes in the shales, slickensides, lenses, and abruptly terminated thin beds of sandstone, a few quartz veins, and general discontinuity of all units. Lithologic sequence varies along the strike. Shales are crumbly, contorted, and dissected by myriad joints which are limonite stained or filled. Thin, cross-bedded sandstones in the shales have limonite concentrations and quartz veinlets along enlarged former laminae, and contain associated solution vugs.
2	158	Sandstone, pinkish-gray to very light-gray and very light-gray, moderate olive-brown, dark-gray, grayish-pink, grayish-orange-pink, or pale-red shale; faulted, friable, massive, very fine- to medium-grained, moderately sorted sandstone that contains clay galls, particularly above thin shale interbeds; shale that ranges from "papery" to massive beds greater than ¼ inch thick. The sandstone comprises an estimated 80 percent of the interval and is in beds up to 20 feet thick that are transect-

Interval	Thickness in feet	
		ed by joints, faults, and quartz veins. Float of black discs with cone-in-cone structure was noted in this interval with other debris washed down from slope-capping terrace deposits.
1	181	Shale and sandstone as in Interval 2; however, shale makes up approximately 70 percent of Interval 1 and the sandstone is in beds less than 30 inches thick where it is present near the base of the interval. Carbonized plant fragments are abundant in and around some of the shale lenses in the sandstones and in some of the sandstones themselves. One firm, very fine-grained, 3-inch sandstone near the middle of the interval is clearly cross-bedded with a topset-bottomset spacing of about 1½ inches. Siltstone occurs as thin interbeds in the shale.

INDEX

(Boldface indicates main reference)