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

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STRUCTURE AND STRATIGRAPHY OF THE RICH MOUNTAIN AREA,

OKLAHOMA AND ARKANSAS

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[illegible] Springer

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

INTRODUCTION

Location and General Description of Study Area

Location of the area that was mapped as a part of this study is shown on the Index Map (fig. 1). Included within its bounds are parts of Le Flore County, Oklahoma; Polk and Scott Counties, Arkansas. The terrain here is relatively rugged and possesses much scenic beauty (pl. III). Mountains near Mena, Arkansas, (pl. I) form some of the southernmost prominences present to the 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 highest mountains in Oklahoma. Because of this, the eastern summits of Rich Mountain, which may be reached via Skyline Drive (pls. I and XXVI), 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 partially rebuilt Wilhelmina Inn, an old hotel erected by the early owners of Kansas City Southern Railroad to stimulate business for the then newly laid line. The hotel was constructed in the 1890's and named for Queen Wilhelmina of the Netherlands. Subsequently the hotel was abandoned and, until the recent renovation, was in a state of decay.

*Reinemund, J. H. and Danilchik, Walter, 1957, Preliminary geologic map of the Waldrön quadrangle and adjacent areas, Scott County, Arkansas: U.S. Geol. Survey, Oil and Gas Investigations, Map OM 192.

**Shelburne, O. B., Jr., 1960, Geology of the Boktukola Syncline, Southeastern Oklahoma: Oklahoma Geol. Survey, Bull. 88, 84 p.

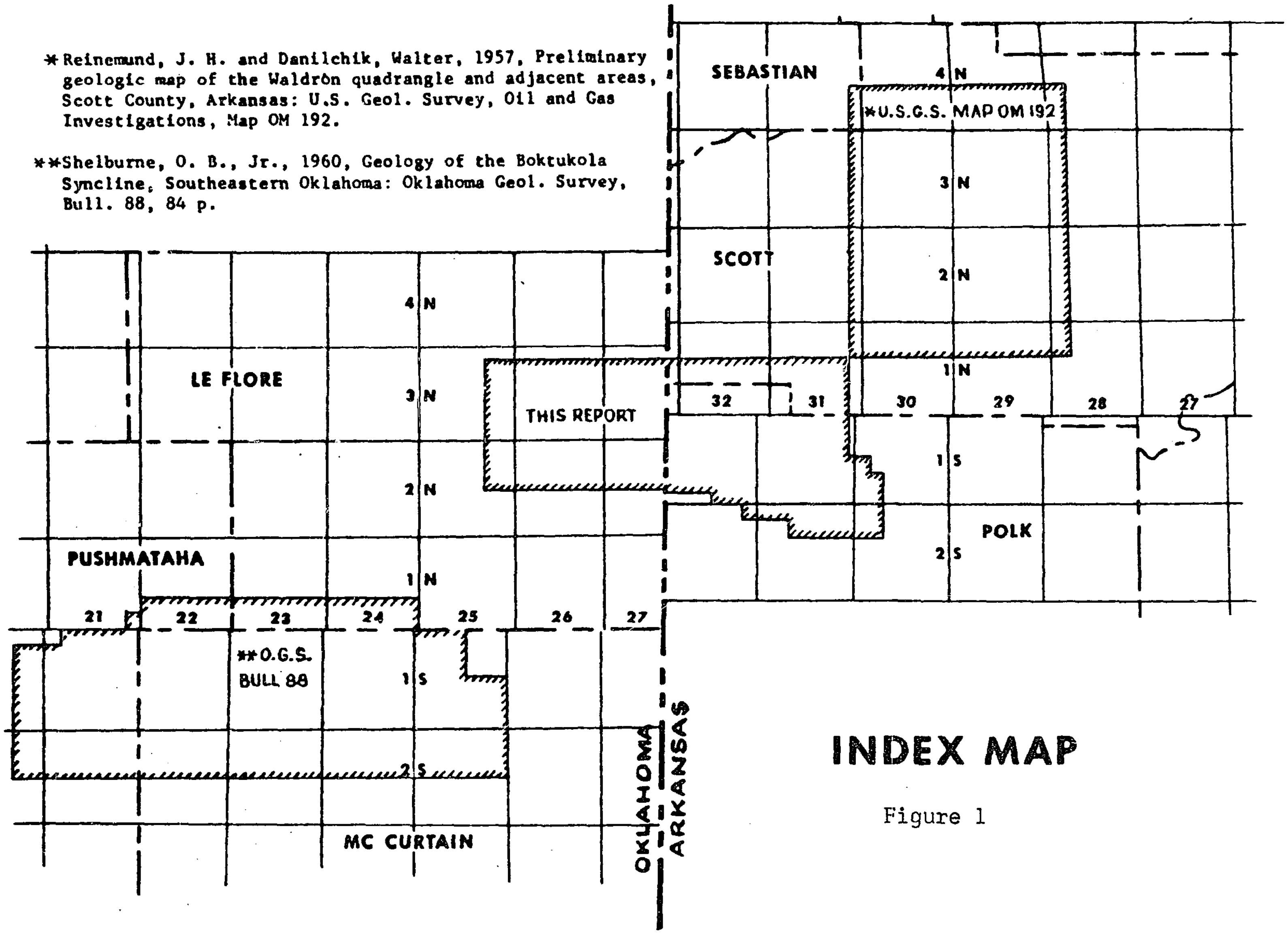
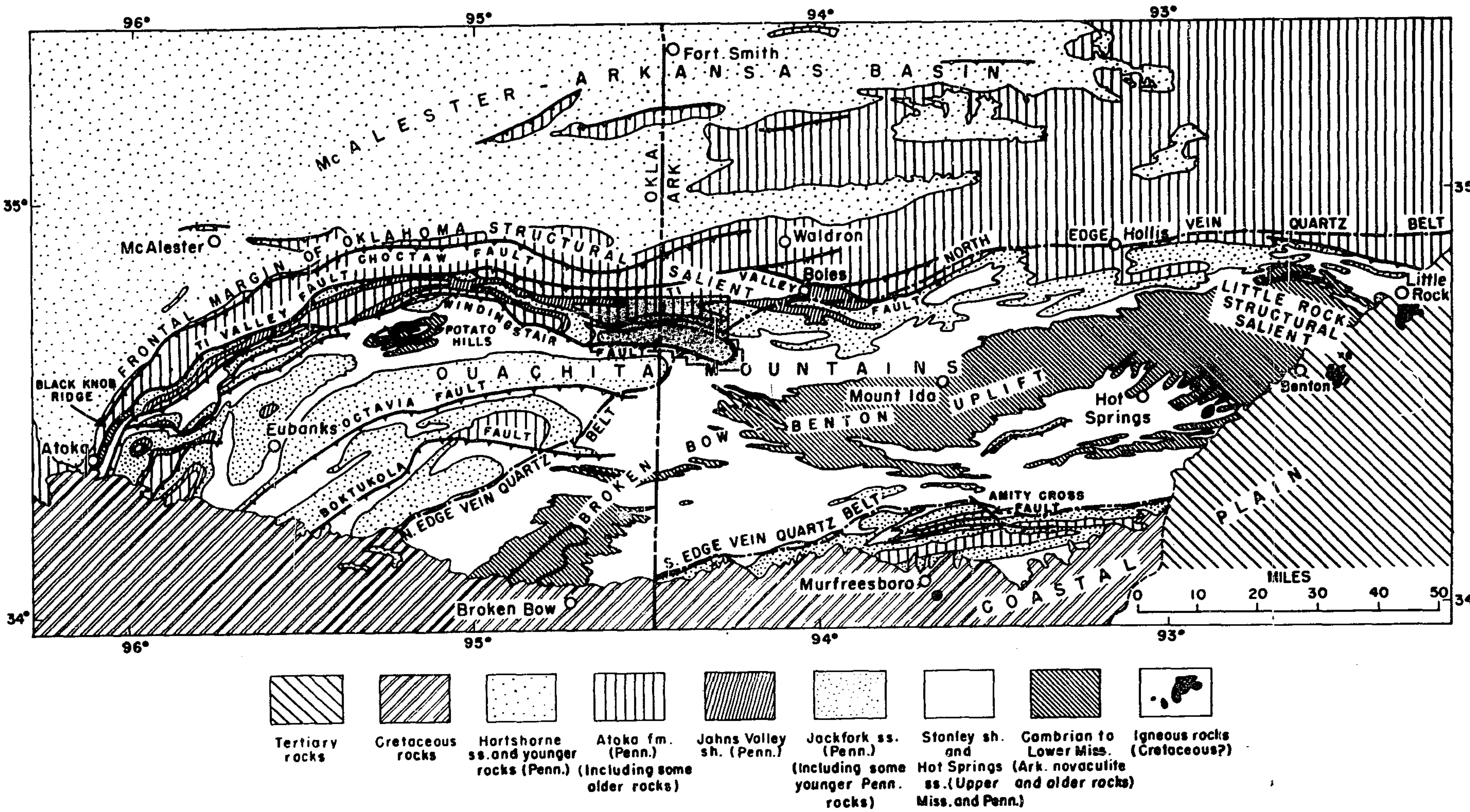


Figure 1



GEOLOGIC MAP OF THE OUACHITA MOUNTAINS

This map is as shown by Miser (1959, Figure 3) except that location of the study area (colored) has been added. Interpretations of the geology in Oklahoma by various workers in the 1950's and 1960's are not incorporated in the map.

Figure 2

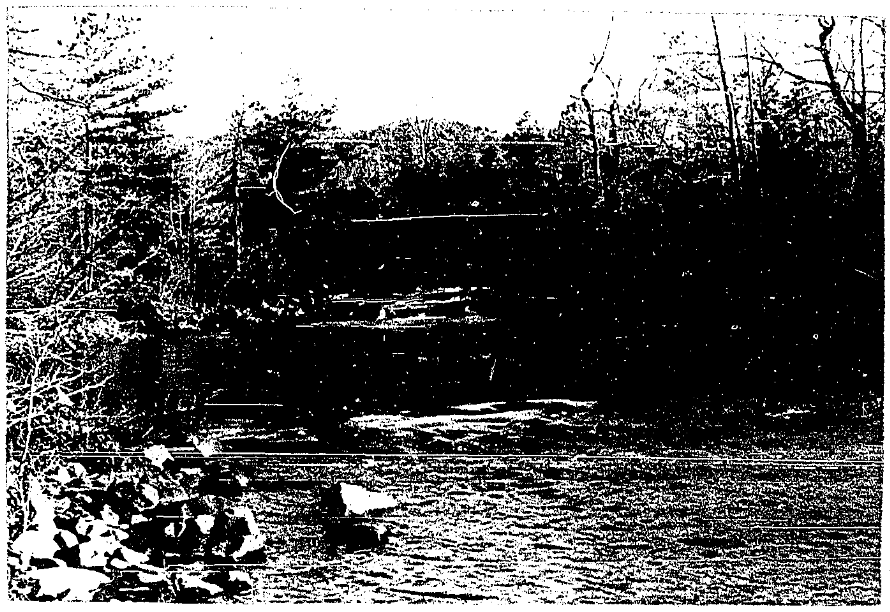
PLATE III

Scenic Locations in Rich Mountain Area

1. Looking southeastward across the headwaters of the Kiamichi River at the lumber road crossing, SW$\frac{1}{4}$ sec. 1, T. 2 N., R. 26 E. The falls are caused by resistant sandstones of the Johns Valley - Atoka, dipping southward (away form the observer).

2. Looking northeastward up Pashubble Creek at beds near the axis of Rich Mountain syncline at the NW cor. SW$\frac{1}{4}$ sec. 4, T. 2 N., R. 26 E.

PLATE III



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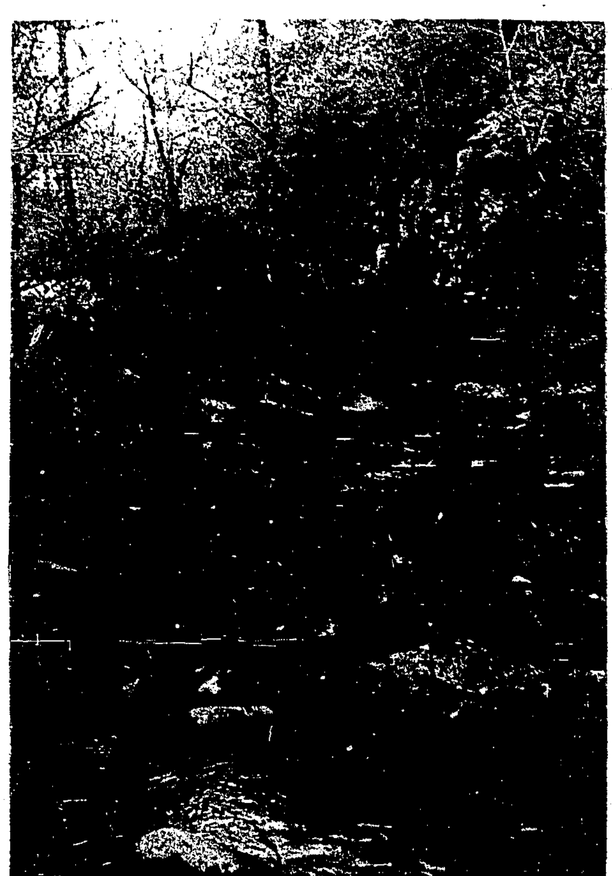


2

The area is covered with a dense growth of many types of plants. Reputedly, Rich Mountain has an unusual variety of vegetation. According to a pamphlet written by G. C. Konkler and revised in 1957 by A. W. Dodson (distributed by the merchants of Mena, Arkansas), a naturalist determined that a square mile of land on Rich Mountain has more kinds of wild fruit, timber, flowers, medicinal plants, ferns, weeds, and other small plants than can be found on any similar tract in a natural, wild state anywhere in the world. He was awarded first prize in a contest sponsored by Ripley's "Believe It or Not" for his itemized list of plant varieties.

Geologically, the area is positioned about midway in the east-west part of the Ouachita Structural Belt that is exposed from Little Rock, Arkansas, to Atoka, Oklahoma (fig. 2). According to the impression given by prior geologic maps of the Ouachita Mountains, the area is also located in a transition zone from tight folds and thrust faults in Oklahoma to open, unfaulted folds of Arkansas. However, results of the writer's research and of recent work by Reinemund and Danilchik (1956) 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 study area indicate steep dips which are not characteristic of open folds (pl. I).

Windingstair fault marks the south boundary of the field-mapped area. Stanley beds to the south of 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 study area.

Investigations: Past and Present

The only published geologic maps which include the Oklahoma portion of the study 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. However, it is the most accurate portrayal of the geology in part of the study area published prior to the present work. Miser did not map the area in person, but based his interpretation upon the unpublished results of the mapping of J. A. Taff. Taff mapped large areas of the Ouachita Mountains near the turn of the century, but only a few of his 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, it does state 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 northeastern corner of the present study area in a small-scale, reconnaissance geologic map that is a part of their report. They also appeared to suggest the presence of a fault (herein named the Honess fault) in their small-scale, generalized structure map.

In the winter of 1959, the writer was searching for a suitable subject for a Ph. D. dissertation. Dr. W. D. Pitt of the University of Oklahoma advised writing to Dr. C. W. Tomlinson, now deceased, for suggestions of good geologic problems. Dr. Tomlinson responded with a letter detailing the discrepancies he had noted in the vicinity of Blackfork Mountain (see plate I for location). He felt that these discrepancies needed to be checked and that study of the area might make a worthwhile undertaking for a Ph. D. dissertation.

Dr. Tomlinson's intimations led the writer to a preliminary survey of the few aerial photographs then in the files of the Oklahoma Geological Survey, which were taken in the vicinity of Blackfork Mountain. The presence of major faults both to the north and to the south of Blackfork Mountain was clear from the photo-reconnaissance. Delineation of these faults and determination of their nature seemed a worthwhile project, and investigation was initiated.

The field work was done during the summers of 1959 and 1960, and on week-ends or vacations when the writer could take time from his teaching duties. The best mapping times were in the fall and spring when vegetative cover was sparse and animal pests dormant. Field transportation was first a motorcycle, which proved inadequate; and, in the final few weeks, a jeep provided by the Oklahoma Geological Survey. The jeep was invaluable; it should have been used throughout the investigation.

During the early stages of research, it became evident that both Rich Mountain and Blackfork Mountain are composed of strata of the same stratigraphic interval. Because exposures on Blackfork Mountain are few and inaccessible, it seemed advisable to identify the beds on Rich Mountain and then work northward. This led to the description of Rich Mountain measured section (Appendix H) and the eventual mapping of Rich Mountain syncline.

Although the subject area is presently surrounded by lands whose geology has not been published on detailed maps, there are current studies being carried on to the north and west. Philip H. Stark of the University of Wisconsin is investigating the microfauna of the Atoka formation of Spring Mountain syncline as a part of his Ph. D. dissertation. Dorwin Hart, also a student at the University of Wisconsin, is mapping the geology directly to the west of the writer's mapping. J. R. McGinley, Jr. of Tulsa University is understood to be mapping beds of the Atoka formation to the north of the subject area in Oklahoma.

It seems unfortunate that mapping of the Atoka formation of the Frontal Belt has proceeded from the west eastward. Ideally the best approach would seem to have been to work first the stratigraphy of the Atoka in the structurally simple areas of thick sections, such as Blackfork syncline (shown in Reinemund and Danilchick's map of the Waldron quadrangle), and then to work westward into the areas of more complex structure. The complex structure is extremely difficult to map accurately with our present incomplete knowledge of Atoka stratigraphy.

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

Order of Presentation

This report is divided into two main parts. The first part is a presentation of observational data gathered during the course of the investigation. The second section discusses hypotheses based upon these data and upon data accumulated by other writers. The division is an attempt to distinguish clearly between the objective and subjective content of this dissertation. However, it is realized that the distinction between the two is relative and that there are degrees of objectivity and subjectivity.

The twofold organization may be somewhat artificial in some respects, but it is hoped that its advantages will outweigh its shortcomings. When hypotheses and observational data are intermingled, it is often difficult to differentiate fact and fancy. It is the writer's desire that this paper represent an effective separation of the two.

PART I

OBSERVATIONAL DATA

CHAPTER I

MEGASCOPIC STUDY

Stratigraphic Units

Definition

The rock unit names in this study (fig. 3) are those employed by Cline (1960). They 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 marker beds composed of siliceous shale is contested by T. A. Hendricks (personal communication). Hendricks feels that the lithology of several formations is so similar that the units would be indistinguishable if the siliceous shales were not present. For this reason he believes that the Stanley and Jackfork would better be termed, "formations", which contain siliceous shale "members". Evidence to support this 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 study, the Wildhorse Mountain, Prairie Mountain, Markham

System	Series	Group	Formation	Member	Bed
Pennsylvanian	Morrowan-Atokan		Atoka		Fossiliferous sandstone and spicular siliceous shale near base
			Johns Valley		
Mississippian	Meramecian - Chesterian	Jackfork Group	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 Group	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

STRATIGRAPHIC UNITS AND THEIR AGES
(After Cline, 1960)

Figure 3

Mill, and Wesley formations could not be differentiated due to absence of the key beds. The approximate contact between the Wildhorse Mountain and the Prairie Mountain was mapped in order 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 the writer. 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 presently assigned ages are shown in figure 3. Ages are those assigned by Cline, as no reason to change them was discovered.

Thicknesses of the following units are not rounded to significant figures because there is no way to determine the accuracy of measurements. It seems probable that the stated thicknesses are within five percent of the true thicknesses, but there is no objective way to verify this estimate.

Stanley Group

Major valleys of the Ouachitas have been eroded from soft shales which comprise the principal rock type of this group. Traces of major faults are found on the north side of many Stanley shale strike valleys. The valley normally lies on the upthrown block. The writer has not mapped formations of the Stanley group south of the Windingstair fault because detailed mapping of the Kiamichi Valley would be necessary just to determine if it would be possible to delineate them. Slightly more than 300 feet of undifferentiated Stanley south of the Windingstair fault are 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 Blackfork Mountains. Except for lowermost beds of these units, they probably are of simple structure so that approximate positioning in the stratigraphic interval is possible. Because of the flood of colluvium, however, units of the Stanley are poorly exposed and the writer made no attempt to map them.

Moyers formation. What is probably the basal Moyers siliceous shale is present in a few outcrops at the foot of the north slope of Rich Mountain where it appears on the dip slopes of a sandstone sequence which it overlies. Erosion of the overlying shale has caused the siliceous shale to be exposed. It consists of gray, siliceous shale in beds a few inches and less in thickness. 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-blue-gray, brittle shale and black shale.

Systematic investigation of the Moyers formation was not made. However, its thickness on the north slope of Rich Mountain was determined to be about 1,000 feet. Sandstone beds observed in the upper part of the Moyers are lithologically similar to those of the lower Jackfork.

Chickasaw Creek formation. The Chickasaw Creek siliceous shale is described in both the East Ward Lake and Ward Lake Spillway measured sections. Its resistant beds also are well exposed on the north slope of

Plate IV

Outcrops of Chickasaw Creek Siliceous Shale

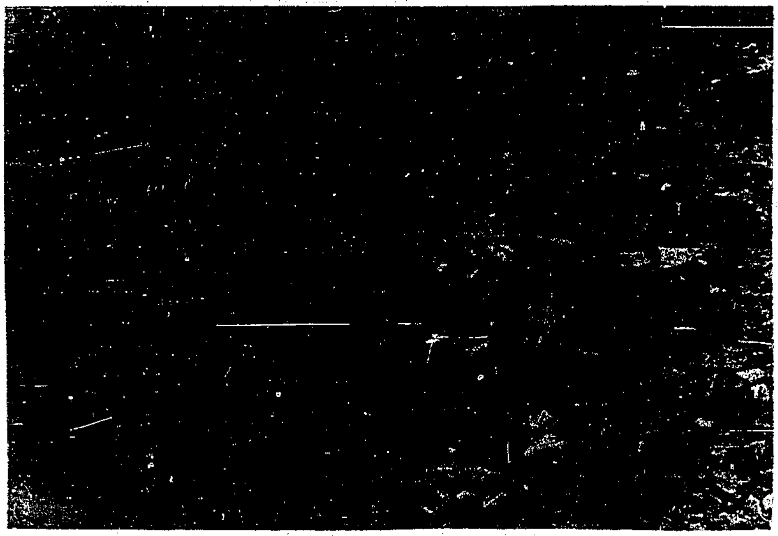
1. Ward Lake spillway exposure. The base of the Chickasaw Creek (top Interval 3, Ward Lake spillway measured section) is at the top of the resistant sandstone exposed in the lower left corner. The pick lies on the blocky-weathering beds of a typical cherty zone.

2. Ward Lake spillway exposure. Blocky-weathering beds of a cherty zone near the top of the Chickasaw Creek are visible in left center. Softer, 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).

3. 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 lowest resistant sandstone of the Wildhorse Mountain formation.

PLATE IV

1



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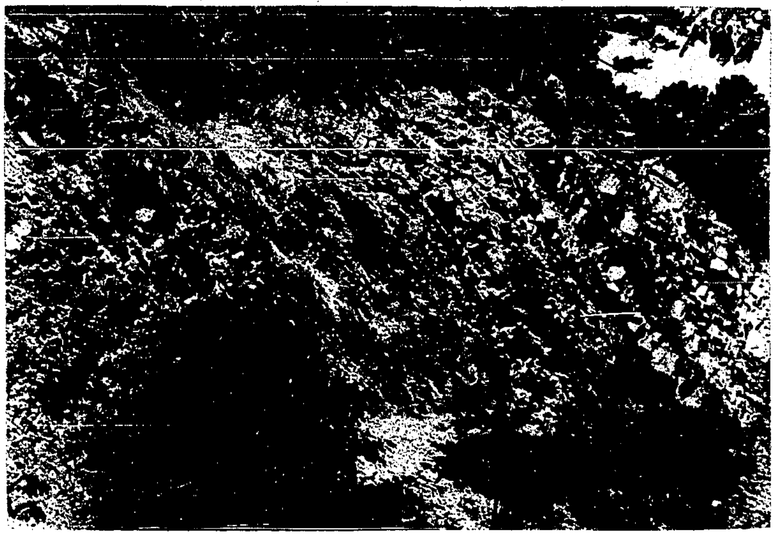

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Blackfork Mountain (NE$\frac{1}{4}$ SW$\frac{1}{4}$ sec. 25, T. 1 N., R. 32 W.). For detailed descriptions of the Chickasaw Creek the reader is referred to the measured sections that are included in Appendix H and to the microscopic petrography section of this paper. At the measured localities, the Chickasaw Creek is composed predominantly of shales 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 which one would expect to find in them, are not evident. Close inspection reveals specks to be present, but their diameters are generally less than 0.1 mm. The smallness of the specks causes assignment of these beds to the Chickasaw Creek to be based upon topographic position and sequence correlation.

Sandstone beds make up a minor part of the Chickasaw Creek but increase in abundance as the upper contact is approached. Here they grade into the basal sandstone zone of the Jackfork (2 and 3, pl. IV). 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.

There is an excellent exposure of beds of the upper part of the Moyers, Chickasaw Creek, and lower part of the Wildhorse Mountain formations in a stream valley on the north slope of Blackfork Mountain in NE$\frac{1}{4}$ SW$\frac{1}{4}$ 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 and give rise to steep dip slopes extending several tens of feet downdip. A few sandstones reach a thickness of 20 feet.

Above the upper Moyers sandstone zone is a shale zone whose thickness is estimated to be between 150 and 200 feet. The shales underlie a topographic bench and are poorly exposed. There are two slightly elevated ridges in the bench: one is due to an 8- to 10-foot siliceous shale zone near the base of the interval; the other is due to a 10-foot tuff bed that is about 100 feet above the siliceous shale. Bands and lenses of white specks are present in some of the one- to three-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 do not contain a high percentage of silica as 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 and the ridge it forms is about 10 feet thick. Extent of tuffaceous beds above and below the ridge is not evident. The tuff is medium-dark-gray and hard (for induration terminology see Appendix H) in fresh samples. It contains many large light-colored grains, most of which have a talc-like megascopic appearance, but some 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 paper.

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

Overlying the shale sequence is a sandstone zone with an estimated thickness of 50 feet. This sandstone zone in turn is overlain by a poorly exposed sequence of about the same thickness also dominantly composed of sandstone. 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 sub-platy shale. A few laminae of silt- or sand-sized material are visible.

Directly overlying this upper siliceous shale is another sandstone zone. The sandstone beds are medium-gray, very-fine-grained, very hard, 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.

Other exposures of the Chickasaw Creek were observed at the extreme west end of Rich Mountain and on the eastern extremity of Windingstair Mountain. These are of poor quality, however, and require careful search for their location.

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 the presence of faulting. At the Blackfork Mountain exposure a question of where to place the top contact arises due 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. For location of these exposures and their description the reader is referred to Appendix H. Outcrop quality does not compare with the excellent roadcuts south of Big Cedar on 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 due to leaching; however, O. B. Shelburne (personal communication) found that several Jackfork sandstones appear white in well samples.

Highway 103 has good roadcut exposures where it traverses the western edge of the study area. Beds of the upper part of the Jackfork through the lower Atoka are well exposed in them. Philip H. Stark is currently measuring and describing them as a part of his Ph. D. dissertation at the University of Wisconsin.

Wildhorse Mountain formation. In the type localities this formation overlies the Chickasaw Creek siliceous shale and its top is marked by a siliceous shale (basal Prairie Mountain siliceous shale member). Near the middle of the formation is a zone of variegated green and maroon shale (Prairie Hollow maroon shale member) that has been used by Hendricks

PLATE V

Outcrops of the Wildhorse Mountain Formation

1. Quartz vein directly above hammer head between sandstones of Interval 5, East Ward Lake measured section.

2. The dip slope of a friable sandstone of Interval 36, Rich Mountain measured section. The sandstone is distinctive because of the smooth mounds on its weathering surface and because of yellow, yellow-brown, and red-brown staining.

3. A northward view of sandstones of Interval 38, Rich Mountain measured section. The sandstones are dipping 60 degrees southward (toward the observer), except where they have slumped as in the upper left part of the picture.

PLATE V

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PLATE VI

Outcrops of the Wildhorse Mountain Formation

1. The dip slope of a friable sandstone of Interval 42, Rich Mountain measured section. Flaking due to weathering is present only at this stratigraphic level.

2. Looking westward at sandstones in the upper part of Interval 42, Rich Mountain measured section. The bed shown in 1, plate VI, is in left center. Southward dip is 56 degrees.

3. Looking westward at a thick zone of massive sandstones of Interval 44, Rich Mountain measured section. The beds are dipping 60 degrees southward.

PLATE VI

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et al. (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 located by Cline in the Lynn Mountain syncline or by Shelburne in the Boktukola syncline (Cline, 1960, p. 53) and was not found in the present study 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 may be viewed on plate II.

The Prairie Hollow maroon 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 Blackfork 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 the NE cor. NW$\frac{1}{4}$ sec. 6, T. 2 S., R. 30 W. This shale is only a few hundred feet above the base of the Wildhorse Mountain formation.

Within the outcrop belt of the Wildhorse Mountain formation are the highest ridges of the Ouachita Mountains in Oklahoma. These highest ridges are the result of the large percentage of sandstone in the Wildhorse Mountain, of the thickness of sandstone units, and of their induration. All of these factors are high when compared to those of any other formation or zone of equivalent thickness exposed in the Ouachitas.

Cline (1960, p. 49) has 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 Wildhorse Mountain contains 511 feet of sandstone, 293 feet of sandstone and subordinate shale, and 230 feet of shale and subordinate sandstone. In addition, there is 2,344 feet of covered interval in Rich Mountain measured section. Probably more than half of the covered interval consists of shale that has been covered due to its susceptibility to erosion. The total thickness of the Wildhorse Mountain is 3,378 feet.

Except for the erratics of the Johns Valley, the entire Stanley through Atoka sequence is composed of alternating sandstone and shale strata. The thickest sandstone beds are present in the upper Moyers formation, middle Wildhorse Mountain formation, the Game Refuge formation, and the lower Atoka formation. Individual beds rarely exceed 50 feet in thickness; the thickest bed in Rich Mountain measured section is 31 feet (Interval 21) and is located 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 Wildhorse Mountain have prominent topographic expression on the crest and south slopes of both Rich and Blackfork Mountains (2, pl. XXVII).

Progressing upward from the base of the Wildhorse Mountain one notices a change from more poorly sorted sandstones containing varying percentages of clay and plant matter to better sorted, cleaner sandstones. There is also a color change from dark shades of gray to white. Concentrations of carbonized plant matter were noted especially in the lower and upper portions of the formation. The middle part of the Wildhorse Mountain also possesses carbonaceous material, but not in the abundance observed elsewhere.

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 (pls. V and VI). Weathered surfaces of these beds appear "flaky" and have a color 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 in color from olive gray to medium gray to grayish black. Most are fissile to flaky; a few are splintery. No thick shale sections are 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 when the non-impregnated intervening material has been removed.

The occurrence of iron oxide is not dependent upon the presence of arenaceous layers. Most thick shale sections are thoroughly stained. This may be 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 were noted in the middle and upper parts of the unit (1 and 2, pl. X).

In the stream bed at and above the Chickasaw Creek locality on the north slope of Blackfork Mountain are sandstone boulders containing a fauna consisting of molds of invertebrates. These fossiliferous boulders indicate the presence of a fossiliferous sandstone in the lower Wildhorse Mountain beds exposed on the slopes above. Another occurrence of molds of invertebrates in the lower part of the Wildhorse Mountain is near its base. It is noted in Interval 5 of the East Ward Lake measured section (Appendix H). 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 study area and the formations are undifferentiated.

This undifferentiated sequence representing the Prairie Mountain, Markham Mill, and Wesley formations underlies a topographic low between the high ridge formed by the middle Wildhorse Mountain sandstones and a lower ridge formed by Game Refuge sandstones (1, pl. XXVII). 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

PLATE VII

Rocks of the Prairie Mountain - Markham Mill - Wesley Unifffferentiated Interval

1. Sandstone bed in shale section of Interval 57, Rich Mountain measured section. Looking eastward in the direction of Skyline Drive which is about 100 yards away.

2. Looking westward at the poor exposure of shale underlying the Game Refuge formation in Interval 58, Rich Mountain measured section. The shale probably is that of the Wesley formation, but no basal contact is evident.

3. Black siliceous masses taken from the outcrop pictured in 2, plate VII. These masses are believed to be the "chalcedonic masses" which Harlton (1938, p. 888) describes as the most diagnostic feature of the Wesley siliceous shale. The two pictured are the only such masses found on the outcrop.

PLATE VII

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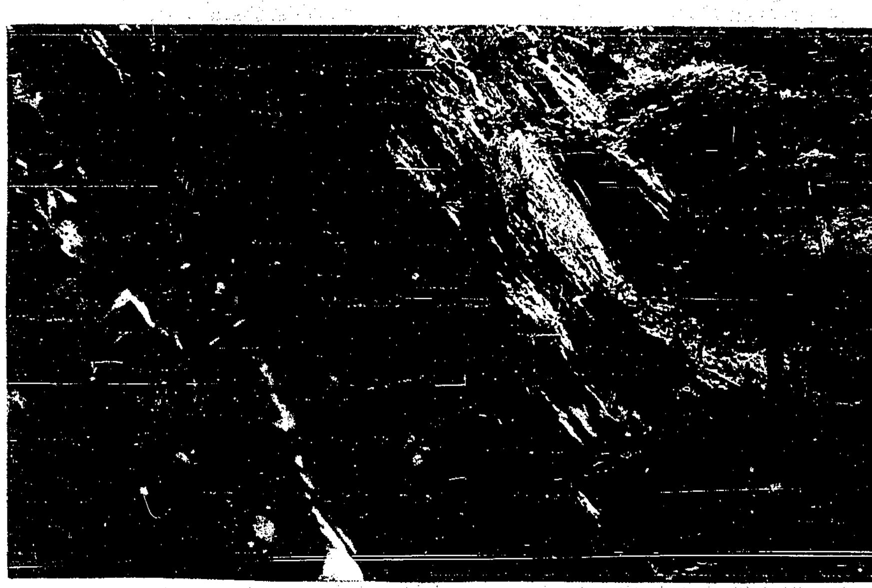

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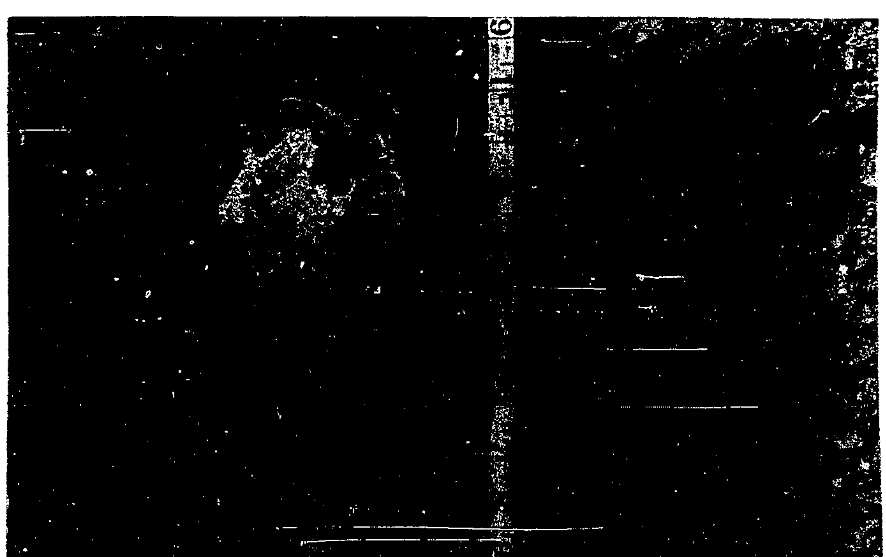

subordinate sandstone. Four hundred thirteen feet of the section are covered. Total thickness is 1,757 feet. The proportion of sandstone in the upper part of the sequence varies along strike. This variation is evidenced locally by prominent sandstone ridges to the west of Skyline Drive.

It is evident that shale is the principal lithology of this sequence in the measured section. Except for sandstone near the base (pl. II) it is entirely composed of shale containing minor sandstone interbeds. The basal sandstone could just 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, planar to wavy-bedded with laminae generally not prominent, and have both bottom and top markings. The top markings consist of a central furrow with marginal, arcuate-shaped mounds, and small tubular grooves. The bottom markings are irregular and small. They show no parallel alignment.

Higher in the sequence a few vertically-spaced, two-foot sandstone beds are present in addition to the thinner beds. These thicker beds have prominent topographic expression (1, pl. VII). The top one-half inch of several of them has a high concentration of carbonized plant fragments. Carbonaceous material is also abundant in sub-discoidal sandstone masses isolated in the shales (1, pl. XIX).

Rare, but distinctive, features in the shale near the top of the sequence are sub-ellipsoidal siliceous masses with a white weathered surface, and similar-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." In addition he states 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", the writer used them to correlate the soft shale in which they occur with the Wesley siliceous shale. The masses and the poor outcrop of Wesley siliceous shale on Rich Mountain measured section are shown on 3, plate VII. According to the definition of chalcedony as fibrous microcrystalline silica, the siliceous masses of the study area fail to follow Harlton's "chalcedonic" classification; rather, they appear to consist of cryptocrystalline silica (1, pl. XXX and the discussion of siliceous shale fabrics in Chapter II).

The lowest siliceous mass was found 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 greater than 175 feet thick. A few small exposures (2, pl. VII) suggest that its dominant lithology is blocky and splintery, dark-gray shale.

Game Refuge formation. The Game Refuge formation contains a relatively high proportion of resistant sandstone in comparison to the rocks above and below. For this reason it forms a prominent ridge. In Rich Mountain measured section its contact with the underlying Wesley shale is gradational, for the transition from the dominantly shale lithology of the Wesley to the sandstone lithology of the Game Refuge takes place within 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) indicated that sandstones of the Game Refuge are characterized by molds of invertebrates; the plant, Calamites; ripple marks; cross-bedding; and, in some parts of the Ouachitas, by a thin siliceous shale. In the western Ouachitas its thickness is 350 to 400 feet, but this thickness decreases northward.

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

A zone of siliceous shale is present in the measured section exposure. Its precise thickness could not be determined, but is less than three feet. Blocks and plates which form the typical float of siliceous shales denote presence of the zone (1, pl. VIII). The shale shows laminae of varying shades of gray and contains a microfauna including sponge spicules and possible radiolarians (see Chapter II). In the few Game Refuge exposures near the western end of 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 Rafuge.

At some localities the siliceous shale reaches a thickness of six feet and consists of fissile 8- to 10-inch beds whose float is similar to that at the Rich Mountain occurrence.

Rocks with invertebrate molds were not seen in place in the measured section, but some are present as 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-sized quartz grains and an abundance of 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. 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 rare in the Game Refuge in the area in which he worked (south and west of the current study). Beds included in the Game Refuge of this study commonly possess bottom casts. Randomly oriented (2, pl. XXIII) and/or parallel-oriented bottom casts were noted at all exposures in the study 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 (1, pl. XXIII). 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 Spring Mountain syncline cannot be separated in the excellent roadcut exposures along Highway 103 in sec. 26, T. 3 N., R. 25 E. Faulting is present but there is probably little

PLATE VIII

Outcrops of the Game Refuge Formation and Stapp Conglomerate

1. Blocky float lying above a zone of siliceous shale in the Game Refuge formation, Interval 59, Rich Mountain measured section.

2. The Stapp conglomerate in the railroad cut in SW¼ sec. 7, T. 3 N., R. 26 E. The pick is located on a sandstone lens (see text for description). Note the excellent flute casts on the base of the Atoka sandstone bed seen at the top of the picture.

3. The Stapp conglomerate at same locality as 2, plate VIII. Compare the non-uniform orientation of long axes of cobbles with the parallelism of axes seen in 2, plate VIII.

PLATE VIII

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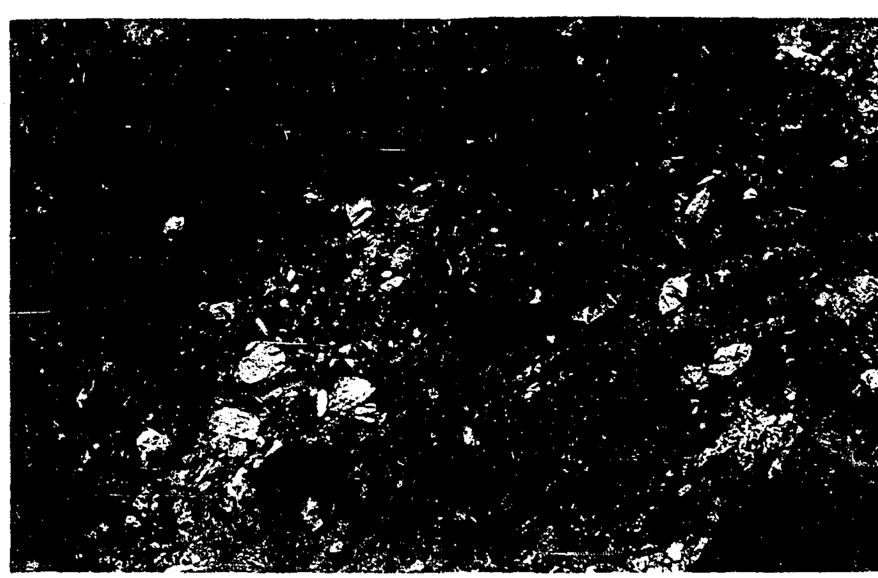


duplication or elimination of section. The writer has mapped about 2,000 feet of upper Jackfork as undifferentiated here. Beds directly below the Johns Valley shale include two thick, apparently massive sandstones such as present in the Game Refuge of other localities. These underlie cross-bedded and ripple-marked sandstones with flute and groove casts, that are interbedded with gray shale. By gradation, shale becomes the dominant lithology, and the Jackfork grades into the overlying Johns Valley. Wesley shale lithology is not readily apparent below the thick sandstones although siliceous shales are present as interbeds between thin sandstones.

There is a relatively high proportion of sandstone in the upper Jackfork sequence of Spring Mountain syncline, and consequently it 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 Jackfork of Rich and Blackfork Mountains. Because Blackfork Mountain lies along strike to the east-northeast of the Highway 103 exposures, the different lithologies require either an abrupt east-west facies change or modification of geographic relationships by tectonism. The absence of known abrupt facies changes along strike elsewhere in the Ouachitas and the presence of evidences for tectonism support the latter hypothesis (see "The Cause of Strike-slip Faulting"). It is possible that the Jackfork has thinned northward by convergence so that the 2,000 feet exposed on Highway 103 represents more than its upper portion: the 2,000 feet here could represent most of the group.

Johns Valley Formation

This unit has been famous historically for its contained erratics and, more recently, for its record of the boundary between the Mississippian and

Pennsylvanian systems. Cline (1960, p. 60-85) gives 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 in exposures on Rich Mountain. Here 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 Highways 103 and 270 near the western border of the study area and on adjacent hillslopes. Part of a Highway 270 roadcut 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 considered by the writer as 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; (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 15 feet. In a shale pit to the west along strike (in the NE¼ sec. 12, T. 3 N., R. 25 E.) the erratics have been removed from shale matrix and their shape may be observed. The largest limestone block thus uncovered has dimensions of 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 rounding is not apparent.

To the south of the conglomerate in the roadcut exposure is sandstone and interbedded shale which contains no erratics. Harlton (1938, p. 888) shows 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 difficult to determine. Beds throughout the outcrop have similar attitudes.

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. They 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 proven. However, because placement in the Johns Valley leads to the simplest structural interpretation, the writer has chosen this designation and the beds are mapped accordingly (pl. I).

The Johns Valley grades downward into Jackfork strata in roadcuts on 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 due, at least in part, to downslope movements so that the stratigraphic sequence is nearly normal. Two dismicrite ("birdseye limestone") boulders whose maximum dimensions 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 portion of the Johns Valley is poorly exposed. There are a few outcrops of dark gray to black shale containing rare thin white calcareous veinlets.

The approximate thickness of exposed beds of the Johns Valley at the Highway 103 locality is 500 feet. Thickness of erratic-bearing beds on Highway 270 is also approximately 500 feet; however, inclusion of the lower sandstone-shale sequence increases the exposed thickness to about 900 feet. Faulting prohibits a reliable determination of formation thickness at both places.

Stapp Conglomerate Facies of the Johns Valley Formation

This facies is described by Harlton (1938, p. 889-895), who referred to it as the Stapp conglomerate member of the Union Valley formation. It is exposed in cuts of the Kansas City Southern Railway (SW$\frac{1}{4}$ sec. 7, T. 3 N., R. 26 E.) across the valley to the east of the Johns Valley roadcuts of 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 (2, pl. VIII). The conglomerate appears to have a closed fabric due to contact between many of the larger fragments.

In most of the exposure, conglomerate is massive due to the absence of recognizable bedding and to random orientation of the coarse material (3, pl. VIII). However, at one location stratification is indicated by sandstone lenses, and associated coarse material shows a parallelism between its long dimension and bedding (2, pl. VIII). The lenses are further stratified by laminae consisting of granules, pebbles, and cobbles. At the top of the sandstone lenses some of these laminae continue laterally into conglomerates, thus indicating a depositional rather than an erosional contact.

At the top of the outcrop shown in 2, plate VIII 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 is 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. Faunal studies were not a part of the current study, but Harlton's (1938, p. 890) assignment of basal Morrowan age is the same as the age designation given by Cline to the upper Johns Valley shale (Cline, 1960, p. 85). The association of an overlying siliceous shale zone and beds containing invertebrate molds is similar to the Hairpin Curve Johns Valley exposure and other Johns Valley outcrops in the Ouachitas. Ridge-forming sandstone beds lie above the Stapp conglomerate, as they do above the Johns Valley in the Highway 103 exposure. For these reasons the writer correlates the Stapp conglomerate with the upper Johns Valley shale as defined by Cline (1960, p. 60-85).

Johns Valley - Atoka Undifferentiated

In Rich Mountain syncline the Johns Valley shale could not be differentiated. Its beds are mapped with lower shales of the Atoka as a single unit.

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

In 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 lower Atoka sandstone 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 Atoka.

The 2,000 feet of strata between the Game Refuge and lower Atoka sandstones consist of shale and a minor amount of thin sandstone and siltstone. Dark-gray to black, splintery to sub-platy 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 interval. Most of the sandstone beds in the shale are less than two feet thick, laminated, and contain carbonized plant fragments. A local occurrence of fossil-mold-containing sandstone about 575 feet from the top of the interval may be float.

Sandstones forming the lower Atoka ridge are light gray, laminated to massive, well sorted, fossiliferous, and in beds up to ten feet thick. 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 sheaf-like appearance on weathering. Molds of clay galls occur on the top of several beds. Above these sandstones near the top of Rich Mountain measured section are several sandstones that have an olive-brown to olive-gray color. They are friable, micaceous, argillaceous, and massive with a general appearance similar to sandstones of the Stanley group.

Johns Valley - Atoka beds of Rich Mountain measured section (including the lower Atoka sandstone member) consist of 343 feet sandstone, 261 feet sandstone and subordinate shale, 633 feet shale, 868 feet 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 Blackfork and Fourche Mountains. In locali-

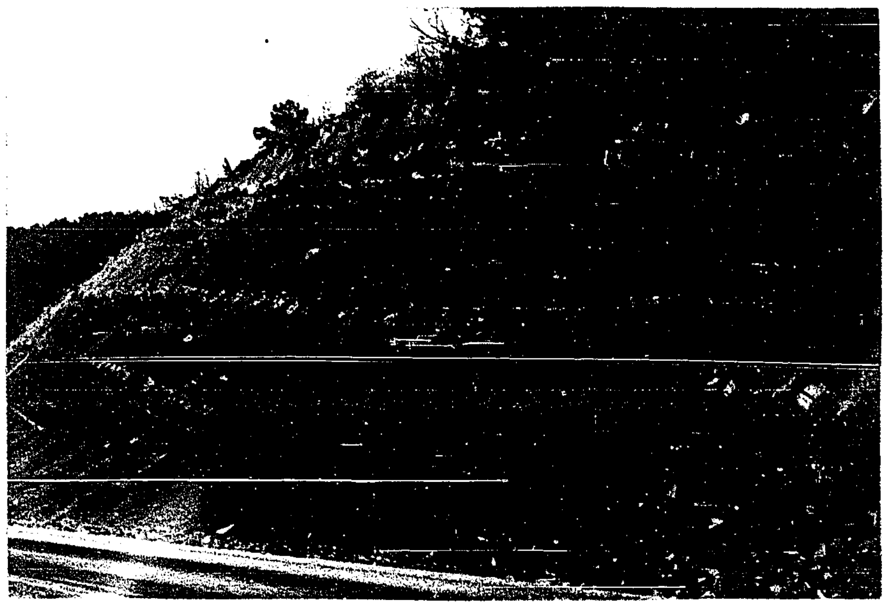
PLATE IX

Outcrops of the Atoka Formation
and Terrace Deposits

1. Looking eastward at shales and even-bedded sandstones of the Atoka formation in south center sec. 17, T. 3 N., R. 26 E. These beds are stratigraphically about 5,000 feet above the base of the formation.

2. Terrace deposits in the north roadcut of Highway 8 at Rocky, Arkansas.

PLATE IX



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ties where 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 Spring Mountain syncline and are present in a zone about 3,000 feet thick. A smaller ridge is present in Stapp syncline.

Above the sandstone zone in Spring Mountain syncline is a section composed predominantly of shale. This section is about as thick as the basal sandy zone. It also contains fossiliferous sandstone, such as the bed in hillside exposures directly to the west of Highway 103 in NE¼ SE¼ sec. 17, T. 3 N., R. 26 E. and about 6,000 feet above the base of the Atoka. Philip H. Stark of the University of Wisconsin is currently describing as part of his Ph. D. dissertation Atoka stratigraphy exposed in the excellent roadcuts of Highway 103.

Fossiliferous Atoka sandstones were noted at several places to the north of Blackfork Mountain. Although their exact stratigraphic position cannot be determined, some probably are above the middle of the 18,500-foot section of Atoka present in Blackfork syncline. This upper Atoka placement seems particularly likely for beds cropping-out in roadcuts near the 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 the center sec. 20, T. 1 N., R. 32 W. and near the center of 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 Briery fault north of Blackfork Mountain with fair precision (see section entitled "Briery Fault").

Colluvium

The steep slopes of Rich and Blackfork Mountains are covered with a mantle of weathered debris creeping downslope under the influence of gravity. This is particularly true of their north, or obsequent, slopes. The structure of the beds beneath these slopes is probably simple, and one can determine the bedrock 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 Moyers siliceous shale and it is necessary to use aerial photographs and associated beds to map the approximate location of the Chickasaw Creek siliceous shale. 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 sheet-like deposits.

Alluvium

Three types of alluvial deposits were distinguished in the mapping: (1) fans, (2) terraces, and (3) floodplains 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 found on some of the higher terraces, the sandstone fragments have partially disintegrated to yield a sandy soil (2, pl. IX).

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 terraces are about 200 feet above adjacent present-day streams. An attempt to correlate terraces from one area to another was not successful.

Where present or former streams have built up deposits having the surface configuration characteristic of alluvial fans, these are so mapped. The fans differ from terraces only in their present-day morphology. Some of the fans are elevated above valley floors whereas others are being formed by existing streams.

Most present-day streams are choked with debris and are not actively down-cutting except in narrow mountain valleys and at knickpoints. Only at the latter localities may one expect to find consistently good exposures of bedrock. In the broad valleys carved by the Kiamichi River and Moun-

PLATE X

Fossils

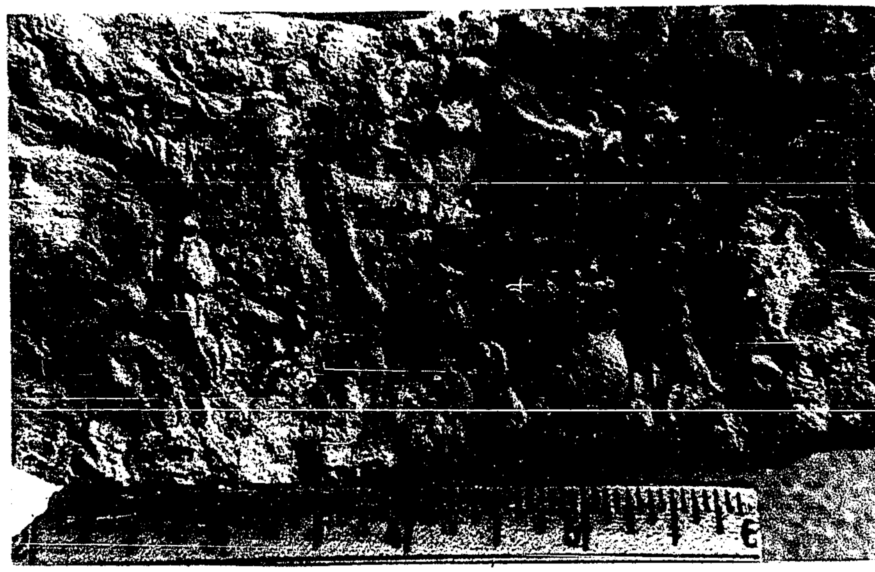
1. Trails of a bottom-dwelling organism preserved on the top of a 3-inch sandstone near the middle of the Wildhorse Mountain formation. Sample taken from the railroad cut in SE¼ NE¼ sec. 8, T. 1 S., R. 31 W.

2. The upper surface of a 1½-inch sandstone from Interval 41, Rich Mountain measured section (near top of Wildhorse Mountain formation) showing trails.

3. A Calamites fragment preserved "in the round." Pictured where it was uncovered in Interval 5, Ward Lake Spillway measured section (base Wildhorse Mountain formation).

PLATE X

1



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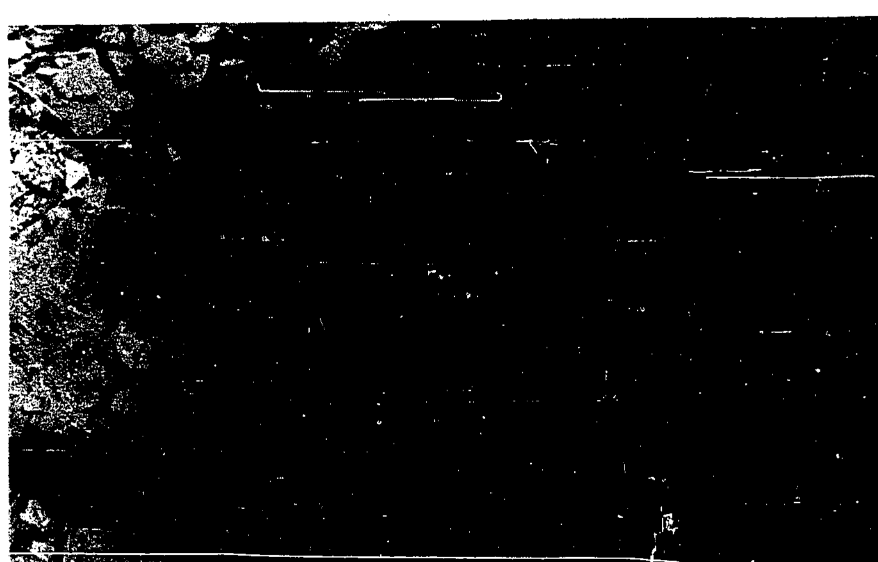


tain Fork of the Little River, streams have a characteristic braided pattern where they have established a wide floodplain. These streams and their ancestors have laid down an extensive cover over beds of the Stanley group.

Sedimentary Structures

Ripple Marks and Cross-Bedding

Honess (1923, p. 197-198) was impressed with the abundance of ripple marks and cross-bedding in the Stanley: "The outstanding facts with regard to the sedimentation of the Stanley are: (1).......(5) the ripple-marked and cross-bedded structure of nearly all of the strata, sandstones and shales alike." Hendricks, Gardner, Knechtel, and Averitt (1947) describe the Atoka as being ripple-marked but do not mention this in their descriptions of the Stanley and Jackfork; nor does Cline mention ripple marks in his description of the Stanley, only indicating that they are not common in the Jackfork except in the Game Refuge (Cline, 1960, p. 58). Cline (1960, p. 98) 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 of the middle Stanley (p. 152-163) and his section of the Stanley-Jackfork transition series (p. 165-173). In the present study, the expression, "ripple-marked", is applied to any bed whose upper surface is uniformly undulatory in exposed segments. Under this definition, ripple-marked surfaces

PLATE XI

Ripple Marks

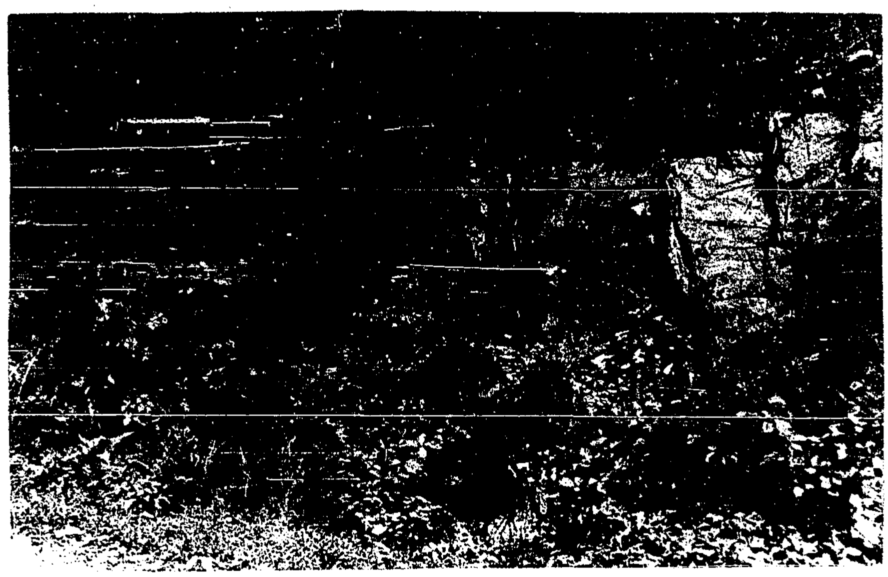
1. Dip slopes of thin-bedded, ripple-marked, cross-bedded sandstones possessing bottom casts. These sandstones in Interval 38, Rich Mountain measured section (upper Wildhorse Mountain formation) are typical of many thin sandstones of the Jackfork group.

2. Asymmetric ripples exposed in the railroad cut in SE$\frac{1}{4}$ NE$\frac{1}{4}$ sec. 8, T. 1 S., R. 31 W. (2, pl. XXV). Current flow was from left to right. Bed is in the Wildhorse Mountain formation.

3. A small exposure of a ripple-marked bed of Interval 45, Rich Mountain measured section (top Wildhorse Mountain formation).

PLATE XI

1



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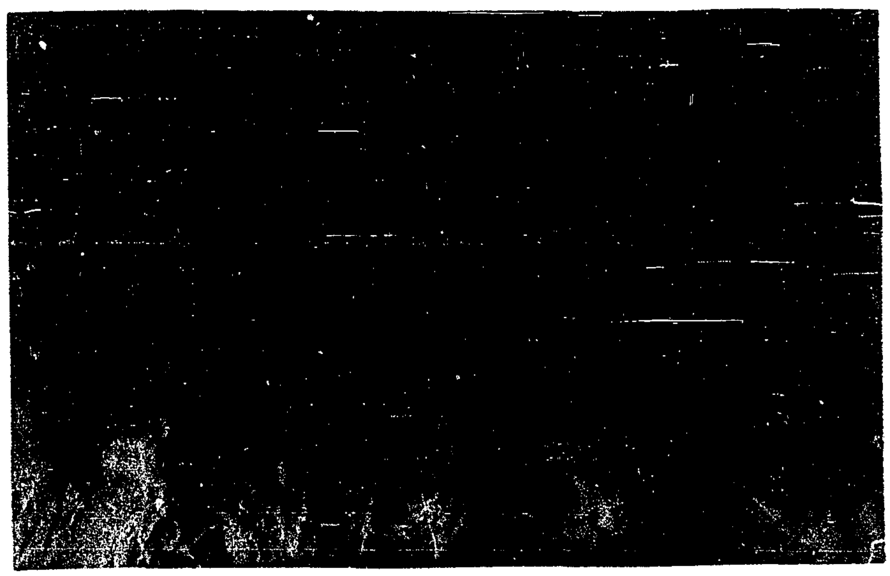

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are not necessarily of current origin. However, in many occurrences the ripple-marks occur in association with cross-bedding and were, therefore, formed by currents. Only where weathering has not etched the lamination pattern at the tops of beds is there a question as to the current origin of the ripple marks.

The upper surface of sandstone and siltstone beds of the Jackfork - Johns Valley - Atoka sequence in the study area possesses a variety of configurations (pl. II). Because of the limited exposure of top surfaces, it is not possible to make positive statements about the comparative abundance of ripple marks on 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 Refuge, 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.

The writer estimates that he observed more and better developed ripple marks in the Game Refuge, Johns Valley, and Atoka formations (pl. II) than lower in the stratigraphic section. Ripple marks and cross-bedding are also common near the Stanley-Jackfork contact (1, pl. XII). These were noted by Cline and Moretti (1956, p. 18) who described 300 feet of section at the base of the Jackfork as containing ripple-marked sandstones.

PLATE XII

Ripple Marks and Cross-bedding

1. Irregular upper surface of a 3-foot sandstone in Interval 7, Ward Lake Spillway measured section. The irregularity of the surface may be partially or entirely due to plastic flow of the sandstone. Maximum relief of the depressions is $1\frac{1}{4}$ inches.

2. Short-crested, current-generated ripple marks above cross-bedded upper parts of sandstones in Interval 64. These beds are arbitrarily assigned a stratigraphic position 60 feet above the Game Refuge formation, but the abundant ripple marks could be evidence that the beds should be included in the Game Refuge.

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

PLATE XII

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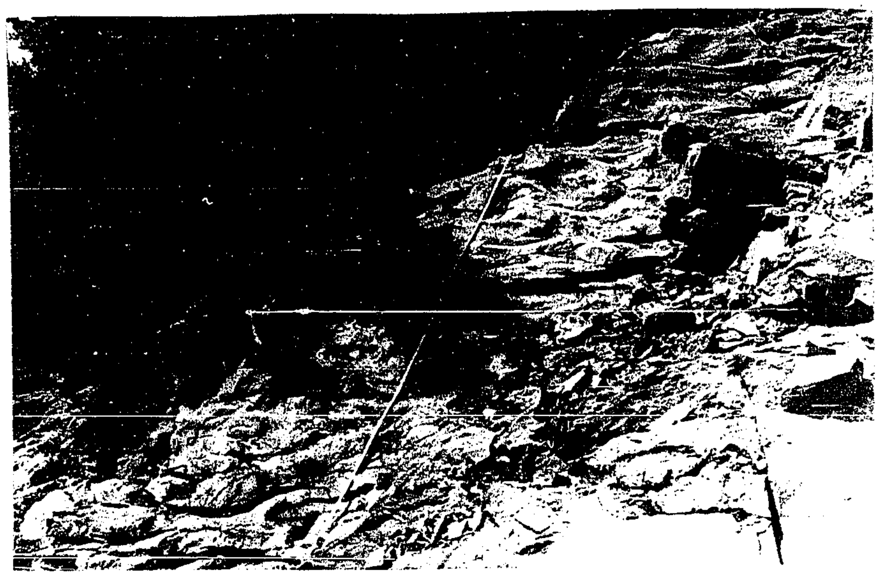

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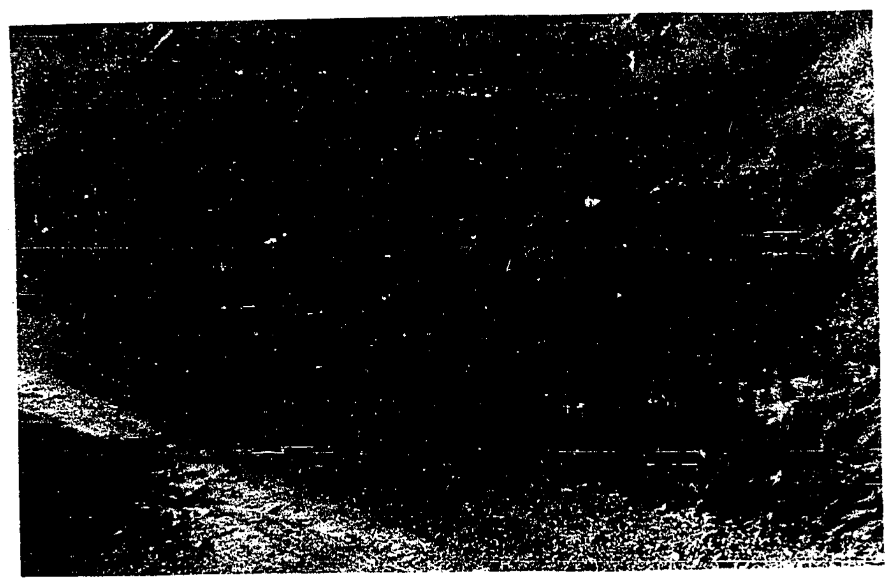

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They also mentioned small-scale cross-bedding of a few beds near the top of the Stanley (p. 10).

Ripple marks with linear sub-parallel crests and troughs are few. Prominent exceptions to this are found in exceptionally well-exposed undulating surfaces of the Game Refuge in SW$\frac{1}{4}$ NW$\frac{1}{4}$ sec. 14, T. 1 S., R. 31 W. and in a limited outcrop of sandstones of the Wildhorse Mountain in SE$\frac{1}{4}$ NE$\frac{1}{4}$ sec. 8, T. 1 S., R. 31 W. (2 and 3, pl. XI). In both localities the ripples are asymmetrical, but the asymmetry is most pronounced in the exposure of Wildhorse Mountain beds (2, pl. XI). The crest-to-crest distance is one to five inches; the shorter lengths being found in the Game Refuge exposure. At the Game Refuge outcrop are apparent 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 rare.

A small exposure of ripple marks with linear, sub-parallel crests may be seen in 3, plate XI. 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 most common ripples have a short crest length or no well-defined linear crests (1 and 2, pl. XII). None of these ripples, however, has a geometric regularity allowing an interpretation of current direction. In

PLATE XIII

Ripple Marks, Cross-bedding,
and Planar-Lamination

1. A cross-bedded and ripple-marked sandstone from Interval 32, Rich Mountain measured section (upper Wildhorse Mountain formation). Note the symmetry of this current-formed ripple.

2. A 5-inch sandstone that is planar-laminated except for a thin, cross-bedded zone at its top. Note the abrupt transition 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 Highway 103).

3. A sandstone that has lamination similar to 2, plate XIII, except that cross-bedding also occurs within the bed (about 1½ inches from the top) as well as at the top and there are groove casts on its base. The sample came from Interval 8, Rich Mountain measured section (lower Wildhorse Mountain formation).

PLATE XIII

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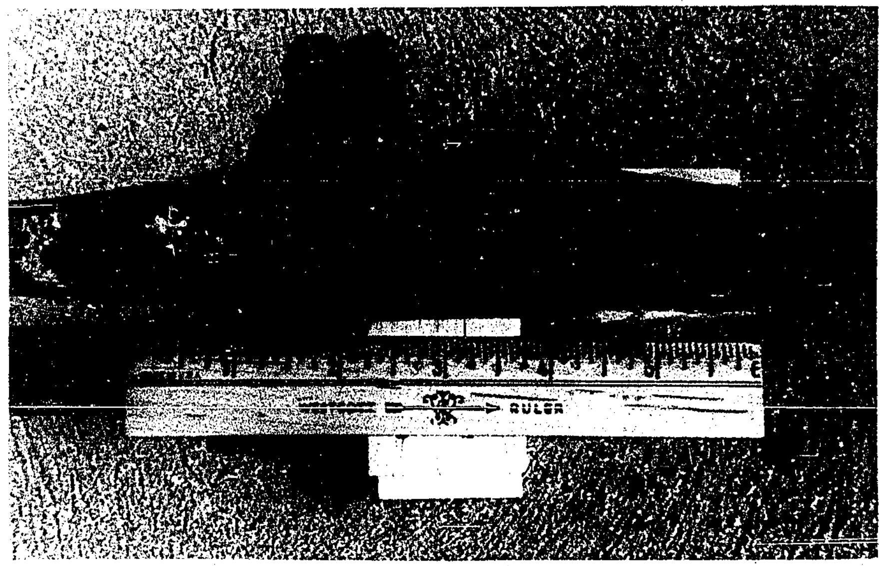

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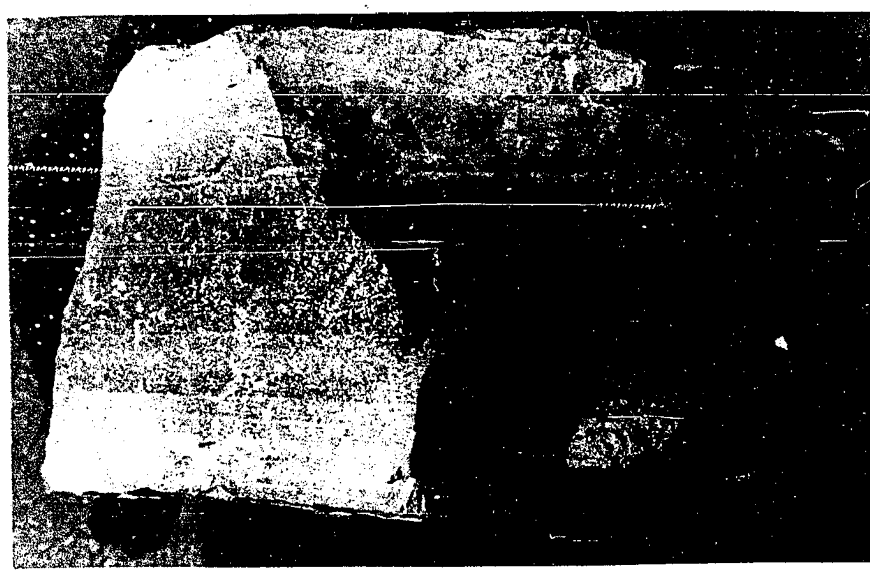

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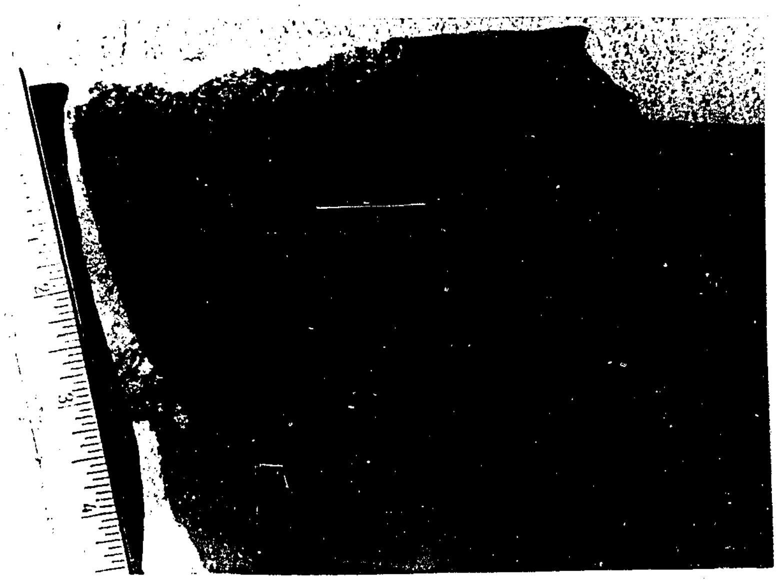


several beds this lack of geometric regularity appears to be due to plastic flowage of their tops. Some of the thin cross-bedded sandstones have smooth low-relief mounds and depressions that appear similar to the periclinal undulations described by ten Haaf (1959, p. 22), but are smaller and show well-developed 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.

Examples of ripple-mark cross-bedding may be seen in plate XIII. Such 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 one and one-half 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 (1, pl. XII).

A cross-bedded zone no more than a few inches thick underlies the upper ripple-marked surfaces of many $\frac{1}{2}$- to 3-foot sandstones (2 and 3, pl. XIII). 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 mid-way between the top and the base of the bed. Bedding in this zone is contorted in the few observed occurrences.

The foregoing occurrences of cross-bedding are the more common observed by the writer. However, cross-bedding is not limited to these relationships. It may occur at any position in a bed and in 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 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 there is an abrupt transition to a thin cross-bedded zone at the top (2 and 3, pl. XIII). This habit is common enough to be useful in top-bottom determinations. It 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 those stratigraphically lower. Some of the Atoka sandstones are so well laminated that their upper portions weather into sheafs of plates bounded by planar or sub-planar 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 varying amounts of plant fragments. The top of many beds possesses countless impressions of plant fragments. Most of the impressions are of flat or flattened fragments. Some *Calamites* stems are preserved "in the round" (3, pl. X) as molds or casts. In a few occurrences the carbonized plant fragments are compacted in thin seams that have a vitreous luster and "coaly" appearance (1, pl. XIX).

The lower laminae of some beds are composed of fragments of invertebrates, particularly crinoid columnals. Due to the solution of these fragments, only their molds may be observed now. 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 (1, pl. XIV). The smaller fragments occur in the finer-grained sandstones; the larger fragments in the coarser-grained sandstones.

Although invertebrate molds are probably more common than previously 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 as well known as that of the upper laminae. However, both lower and upper laminae appear to have similar compositions.

As noted in Chapter II, lamination in the sandstones is also marked by a concentration of heavy minerals and by vertical variation of grain size.

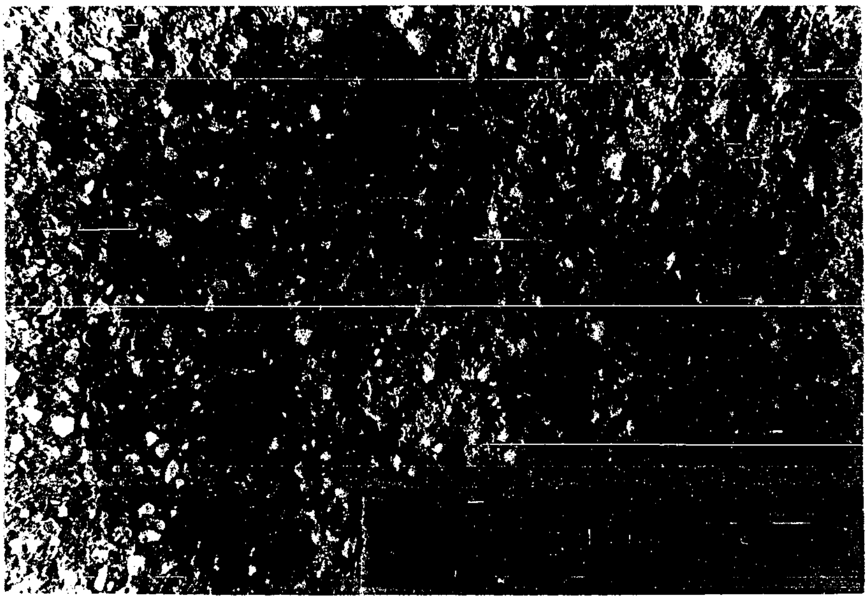
PLATE XIV

Graded Bedding

1. Granules and coarse sand composed of quartz and crinoid columnals etched from a matrix whose estimated median 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.

2. A "graded" bed composed of coarse, massive or planar-laminated sandstone below and fine, planar or cross-laminated sandstone above. The highest lamination which contained 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 1/16th inch apart. The sample is of Interval 77, Rich Mountain measured section (Johns Valley - Atoka undifferentiated).

PLATE XIV



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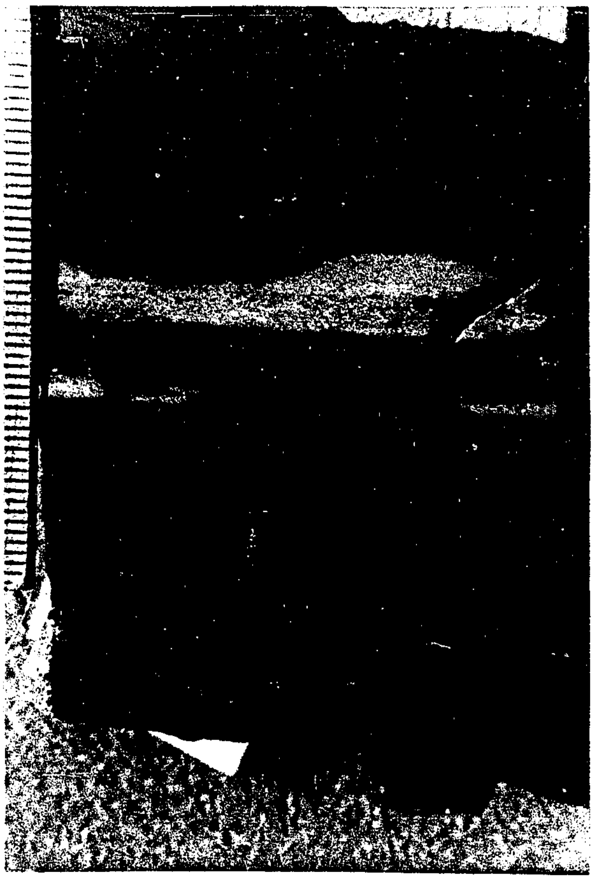

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The relative abundance of these types of lamination is not known.

Graded Bedding

An upward decrease in grain size is not megascopically apparent in most of the sandstone beds of the Jackfork-Atoka sequence. There are, however, several beds within the lower Atoka sandstone ridge of Rich Mountain syncline whose upper laminae have smaller grain sizes than their lower laminae (2, pl. XIV). The lower part of these sandstone beds generally contains abundant granule- or coarse-sand-sized 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.

There is an upward increase in spacing of the coarse laminae. A change in laminae composition to clay, mica, and, in some rocks, plant debris occurs near the top of the bed. Upwards these fine laminae become closely spaced. There may be cross-bedding underlying the upper surface. Individual laminae and interlayers may be well-sorted. The thickness of beds possessing this type of graded bedding lies between one-half foot and three feet, and is commonly between one and two feet.

Thin sections transverse to bedding taken near the top and the bottom of RM 6 and RM 8 (see Chapter II) show the same grain size. Thicknesses of beds from which these slides were made are two inches and five inches, respectively. An example of micro-grading is shown on 1, plate XXX.

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

PLATE XV

Clay Gall Molds

1. 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).

2. Looking down into the cavity seen in the lower left center of 1, plate XV. The cavity occupies the right half of the picture and, in lower right center, are structures suggestive of dentition. Early in his work, the writer considered it possible that some of the cavities were molds of pelecypods. However, failure to find molds preserving shell detail and lack of known large Carboniferous pelecypods caused the writer to consider all cavities as clay gall molds. The breadth of the horseshoe-shaped steel rod is about 5 inches (measuring tape is wired to it).

3. This cavity in the top of a 4-foot sandstone exposed in the railroad cut in SE¼ NE¼ sec. 8, T. 1 S., R. 31 W. (middle part of Wildhorse Mountain formation) also has shape suggestive of a large pelecypod. It was entirely chipped out by the writer, and it consisted of a 1/8th-inch thick layer of dark-gray shale and a sand-sandstone filling. Prior to chipping, the cavity was filled to a level flush with the surface of the sandstone bed.

PLATE XV

1

2

3

PLATE XVI

Clay Gall Molds

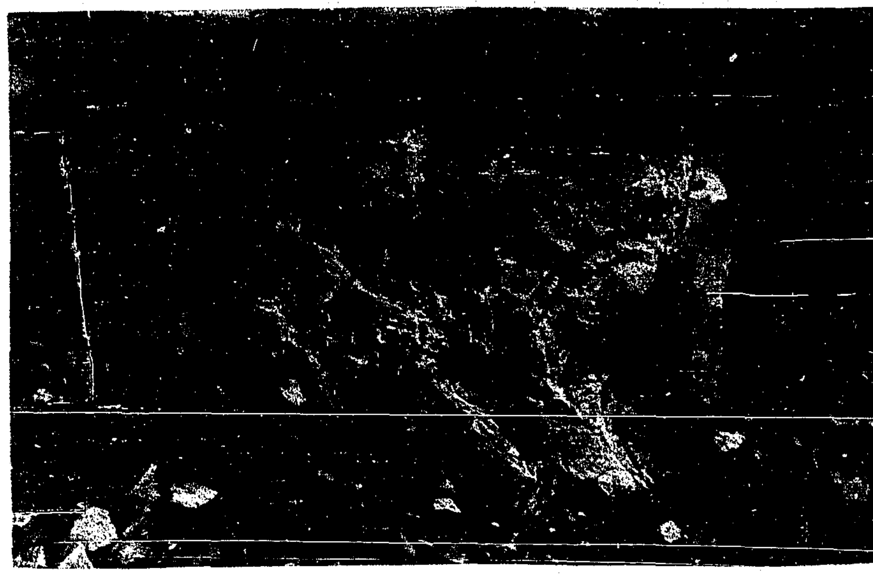
1. Molds of clay galls in the top of a sandstone in Interval 53, Rich Mountain measured section (lower Prairie Mountain - Markham Mill - Wesley undifferentiated). The largest molds are about 2 inches in maximum dimension.

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

3. Large clay gall molds on the top of an Atoka sandstone exposed in roadcuts of Highway 2 just south of the crest of Blue Mountain (not in study area). Imprints of Calamites stem segments are also present.

PLATE XVI

1



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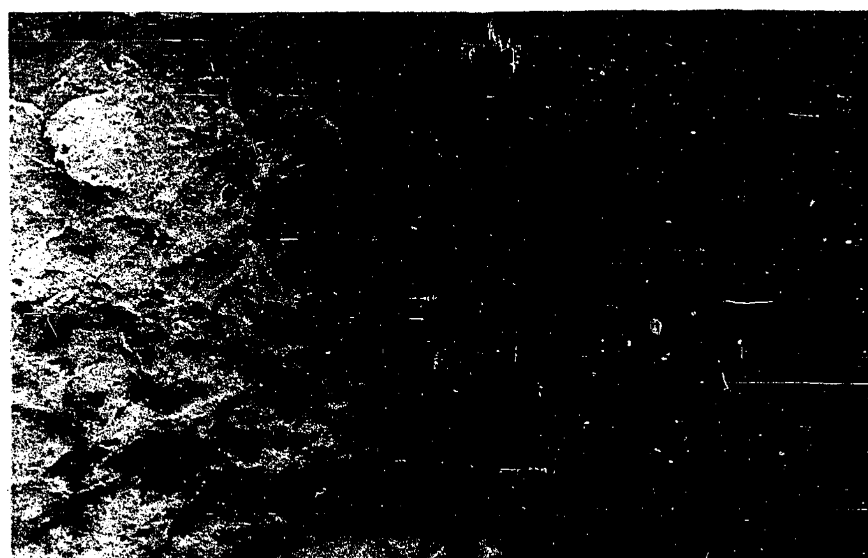

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molds remain as evidence of most of the clay galls. One exception was a partially filled cavity of a discoidal clay gall about one inch thick and three inches in diameter. This buried cavity was revealed when the bed containing it was broken. The cavity was partially 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 occur in many different shapes (pls. XV and XVI). Most of them have smooth, rounded, ellipsoidal outlines. Several have clam-like shapes. Their maximum dimension may exceed a foot, but in most instances is less than six inches.

Clusters of them are most likely found 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 found within strata, the cavities are scattered and their long dimensions show no consistent relationship with bedding planes. However, a statistical study might disprove this apparent random orientation.

Plastic Flow Structures

These structures are ubiquitous at many levels of the Stanley-Atoka sequence. Perhaps the most impressive example may be seen in 1, plate XVII. 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

PLATE XVII

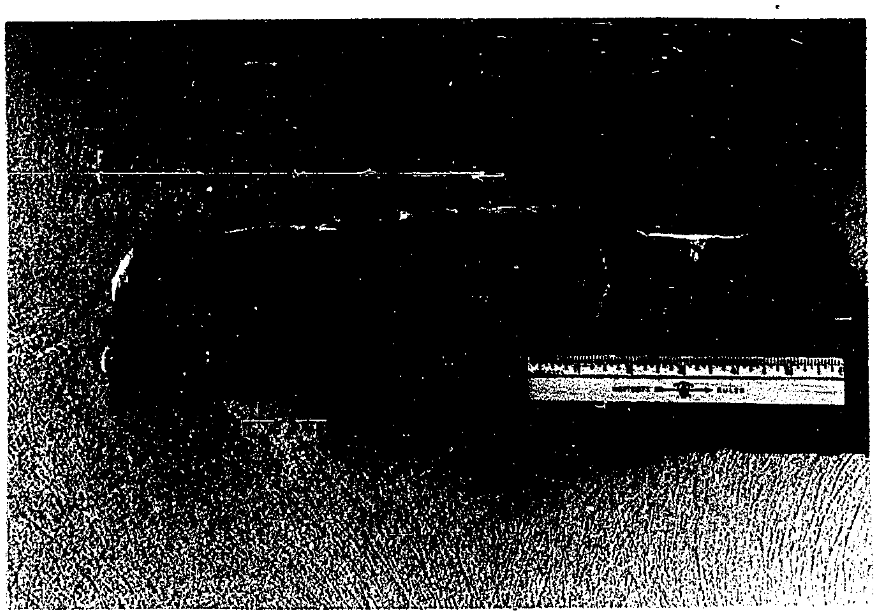
Plastic Flow Structures

1. 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.

2. Example of contorted bedding in a cross-bedded thin sandstone of the Stanley shale found in SW$\frac{1}{4}$ SW$\frac{1}{4}$ sec. 7, T. 2 N., R. 22 E. and collected for the writer's Master's thesis (Seely, 1955, p. 23). Note that contortion folds at the left end of the block are overturned in the opposite direction to that in upper right center.

PLATE XVII

1



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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 and abrupt. 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; Sandstone Dikes; Cementation" in Chapter III.

Contorted bedding and ripple distortion have been discussed under the heading, "Ripple Marks and Cross-bedding." In the few instances where notes were taken, the small internal contortion folds possess axial surfaces with opposing dips. Overturning in both directions was observable (1, pl. XVII).

Another type of plastic deformation is illustrated in 1, plate XVIII. 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 as faults normally imply non-plastic deformation. However, non-plastic and plastic deformation can be gradational, as they are in this example. The pseudo-"faults" 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 if these were faults in the ordinary sense of the term. "Fault planes" are not visible, although they should be if these were true faults. 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

PLATE XVIII

Plastic Flow Structures

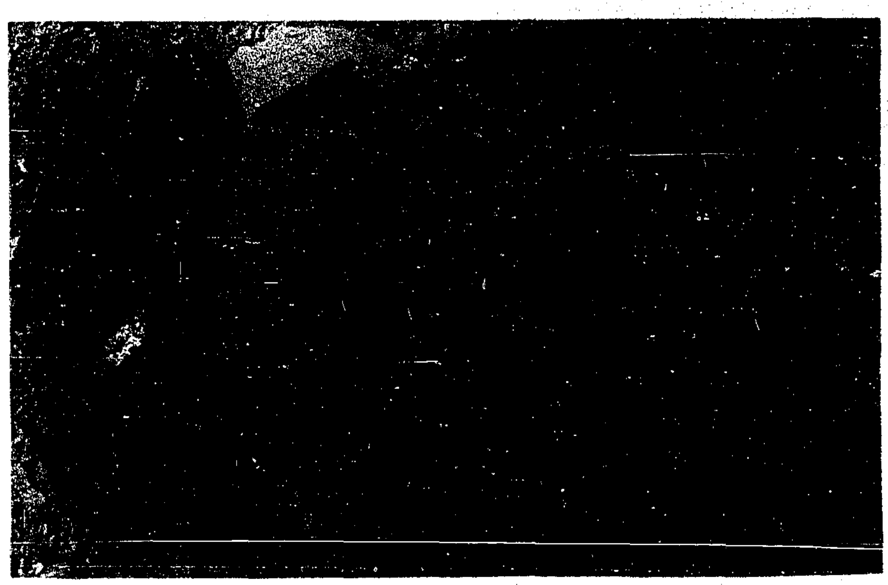
1. Escarpments caused by "faulting" of the bottom of a sandstone bed while the sandstone was in a semi-plastic state. Note that some escarpments offset others, thereby indicating that they must have formed later. The thickest (6 inches) part of the block is left of the photograph's center; the thinnest (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).

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

3. 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).

PLATE XVIII

1



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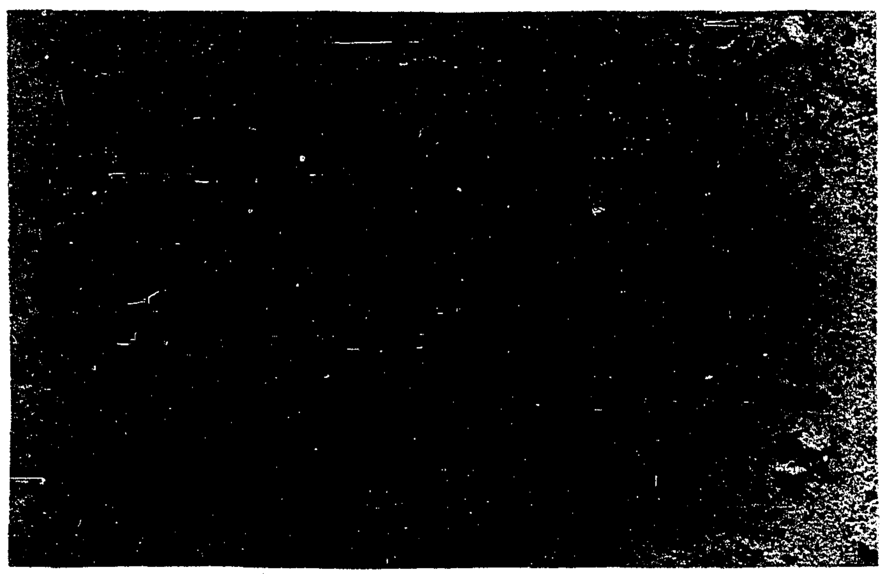

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block in 1, plate XVIII is its thinnest part. Irregular sandstone beds and isolated boulder-sized masses in thick shale sections also have surfaces with small escarpments. The boulders, however, are somewhat rounded and the escarpments may be folded.

Also in thick shale sections are discoidal or roughly ellipsoidal isolated sandstone masses whose internal laminae show widely varying contortion. 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 2, plate XVIII. In others the laminae are warped or folded as may be seen in the lower boulder of 3, plate XVIII. 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 East Ward Lake measured section is a sandstone bed whose thickness ranges from two to three inches. 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 (2, pl. XIX). These sandstone masses resemble the isolated boulders found elsewhere in shale intervals except that their attachment to a parent bed is here preserved. Plant fragments are abundant in them 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 intervals in the Stanley-

PLATE XIX

Plastic Flow Structures

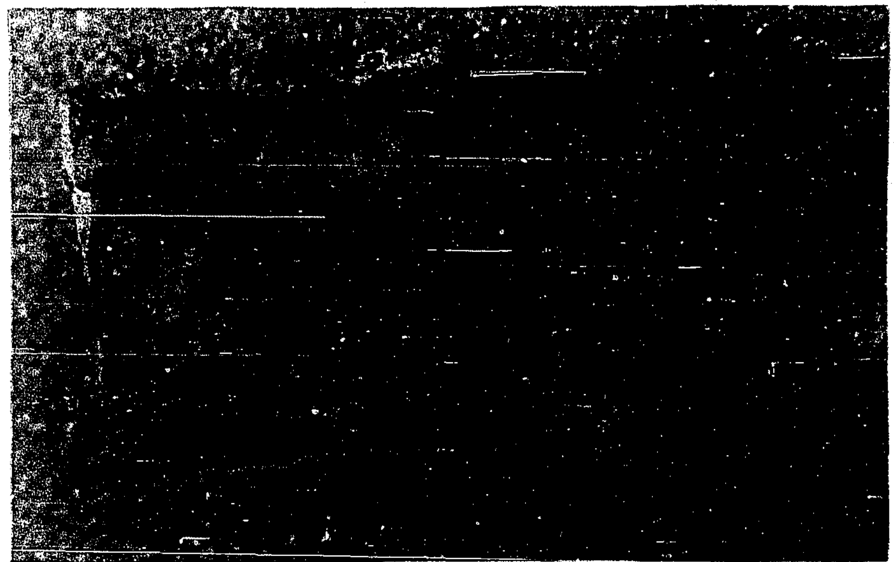
1. Plastic deformation of this sample is shown by abundant carbonized plant fragments. A vitreous luster imparts a "coaly" appearance to seams at the base of the rock. The sample is from Interval 57, Rich Mountain measured section (Prairie Mountain - Markham Mill - Wesley undifferentiated) and is pictured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

2. 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 over 7 feet in long dimension 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.

3. Load casts at the base of a sandstone bed of the Jackfork group exposed just north of Briery fault in the east roadcut of Highway 103.

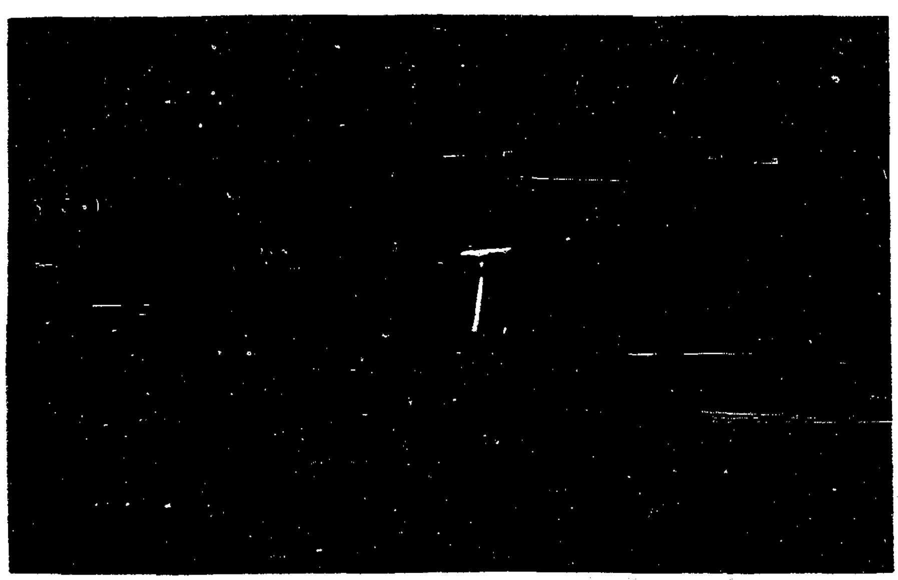
PLATE XIX

1



2

3



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 the protuberance and the base of the bed. An example of the sinuous type of downward protuberance is shown on 3, plate XIX. The protuberances show varying relief and appear to grade into flute casts, groove casts, or other bottom irregularities. They have been described as load casts by Kuenen (1953, p. 1058) and are best developed on sandstone beds that are greater 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 publicized sedimentary structure 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 which imperceptibly merges with the lower surface of the bed. The blunt end has been interpreted as pointing up-current. Groove casts are parallel-oriented, elongate, downward protuberances that do not possess a bulbous terminus. Examples of flute and groove casts are shown on plates XX and XXI.

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 (2, pl. XXIII). How-

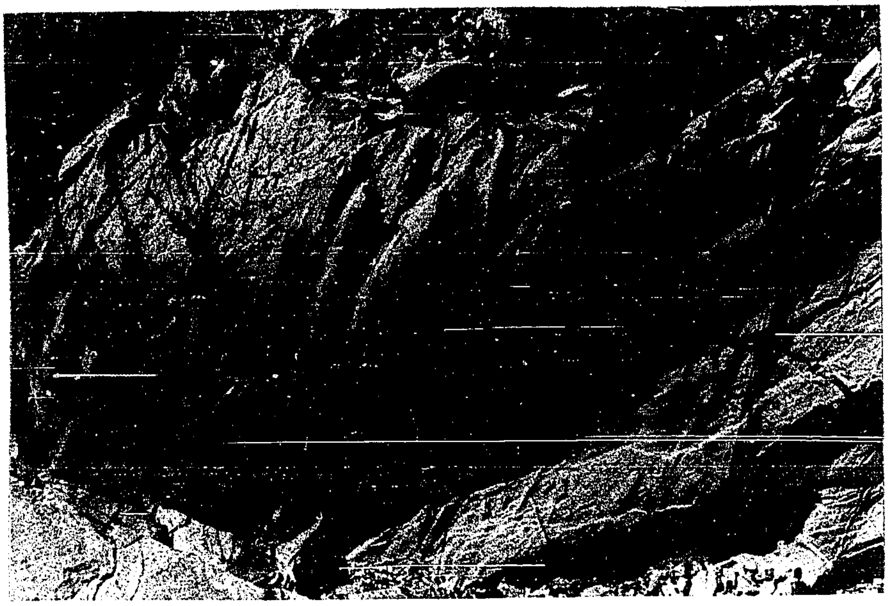
PLATE XX

Large Flute Casts

1. 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. If these small casts were formed by organisms, they may be evidence for a time interval between the eroding and filling of the bottom depression represented by the large cast. Looking northwestward at a sandstone of the Johns Valley shale (?) exposed in the Highway 270 roadcut near Stapp, Oklahoma.

2. Northward view of large flute and groove casts on the base of a Jackfork sandstone exposed in the east roadcut of Highway 103 just north of Briery fault. Note that small casts are present on the large casts as in 1, plate XX. Current flowed from right to left.

PLATE XX



1

2

PLATE XXI

Flute and Groove Casts

1. Looking northwestward at flute casts on a sandstone in the Johns Valley shale (?) exposed in the Highway 270 roadcut near Stapp, Oklahoma. Current flow was from upper right to lower left.

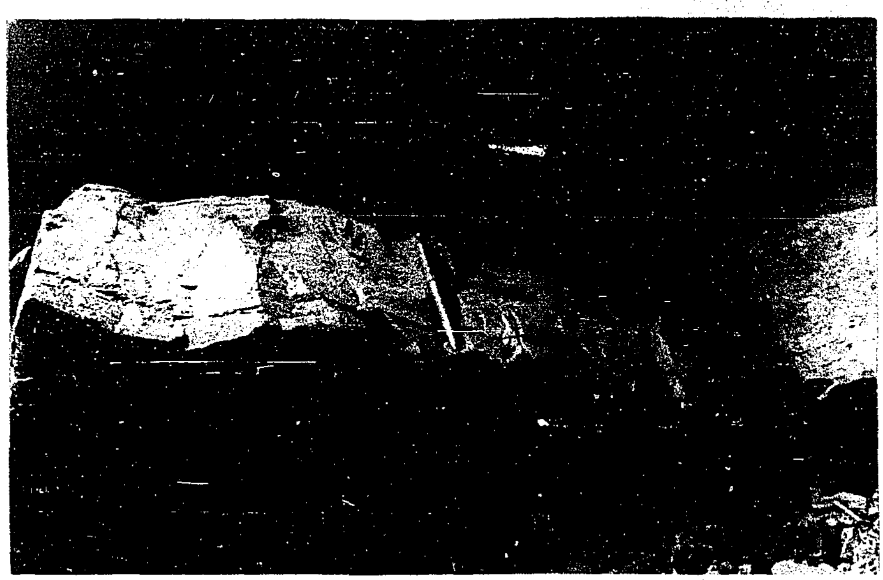
2. Looking northwestward at groove casts on the bottom of sandstones at the base of Interval 5, East Ward Lake measured section (base Wildhorse Mountain). The sandstones are visible in 3, plate IV. Cross-bedding in the sandstones indicates that current flow was from right to left.

3. Unusual groove casts on the base of a thin ($1\frac{1}{2}$-inch) Atoka sandstone from SE$\frac{1}{4}$ SE$\frac{1}{4}$ 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.

PLATE XXI

1

2



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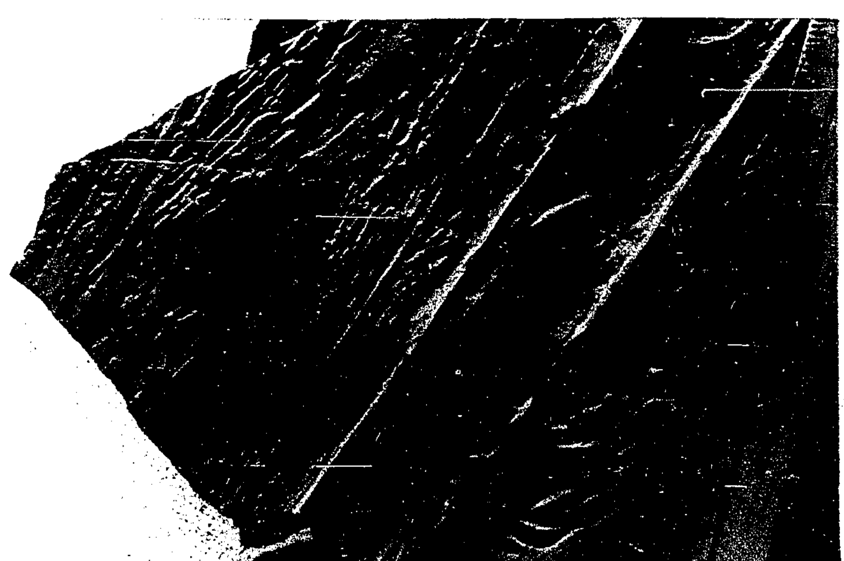


PLATE XXII

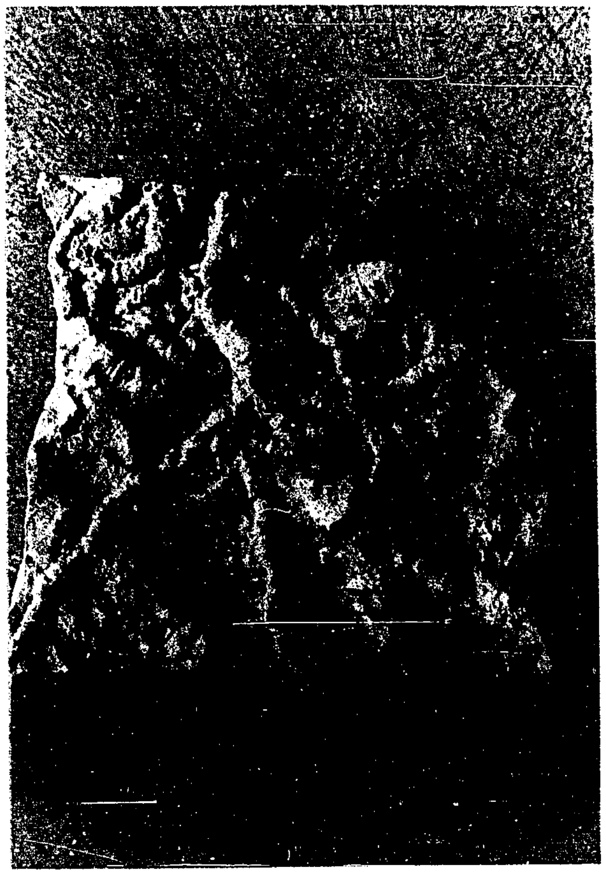
Burrow (?) Casts and Other Sinuous Bottom Casts

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

2. Burrow (?) casts on the bottom of a 1-inch sandstone bed. Sample from Interval 41, Rich Mountain measured section (upper Wildhorse Mountain formation).

PLATE XXII

1



2

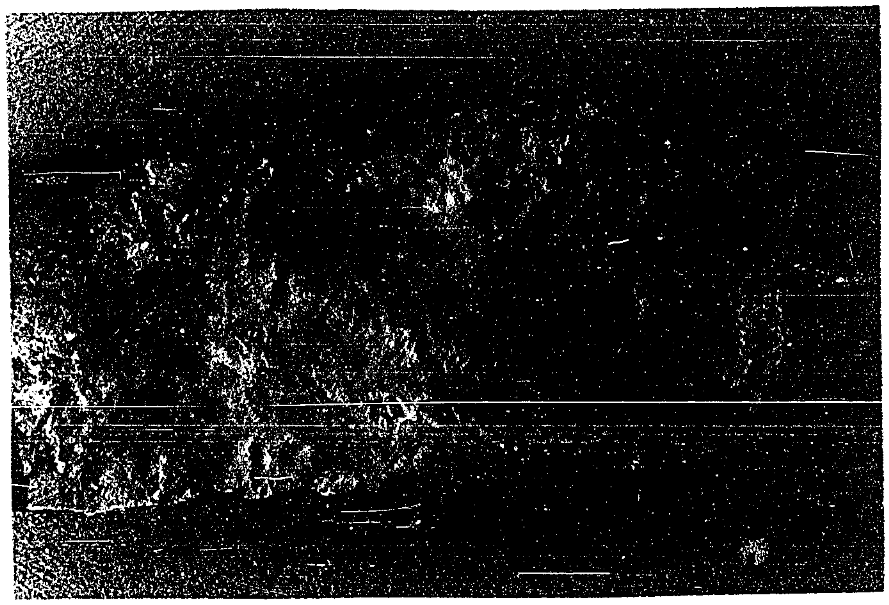
PLATE XXIII

Bottom Casts probably made by Organisms

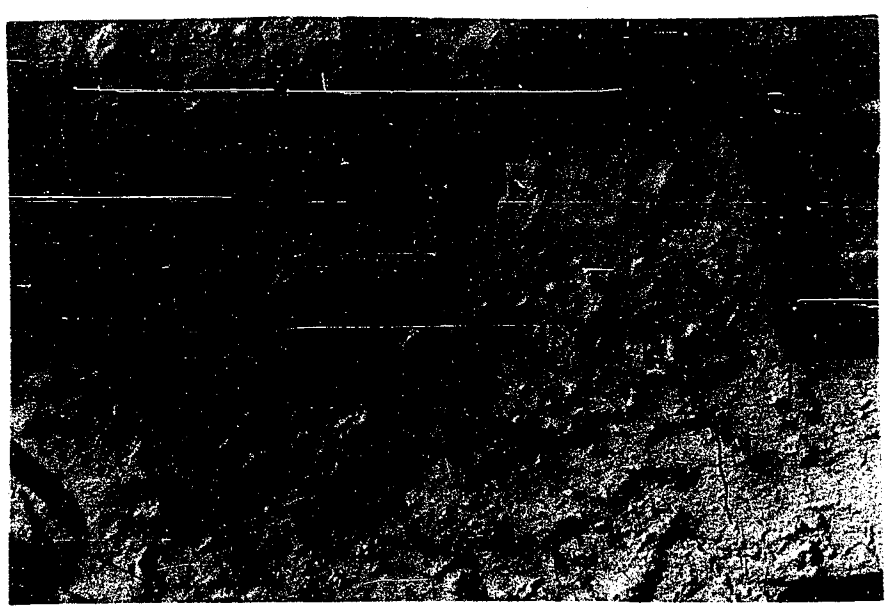
1. Casts of regularly offset depressions found on the base of a $1\frac{1}{2}$-inch, cross-bedded and ripple-marked sandstone from Interval 61, Rich Mountain measured section (Game Refuge formation). The regularity of offset is best seen in the trail nearest the right edge of the slab.

2. Non-oriented bottom casts probably caused by activities of benthonic organisms or by bottom debris removed prior to deposition. Casts are on the bottom of a $1\frac{1}{2}$-inch, cross-bedded and ripple-marked sandstone bed of Interval 59, Rich Mountain measured section (Game Refuge formation).

PLATE XXIII



1



2

PLATE XXIV

A Cross-bedded Flute Cast

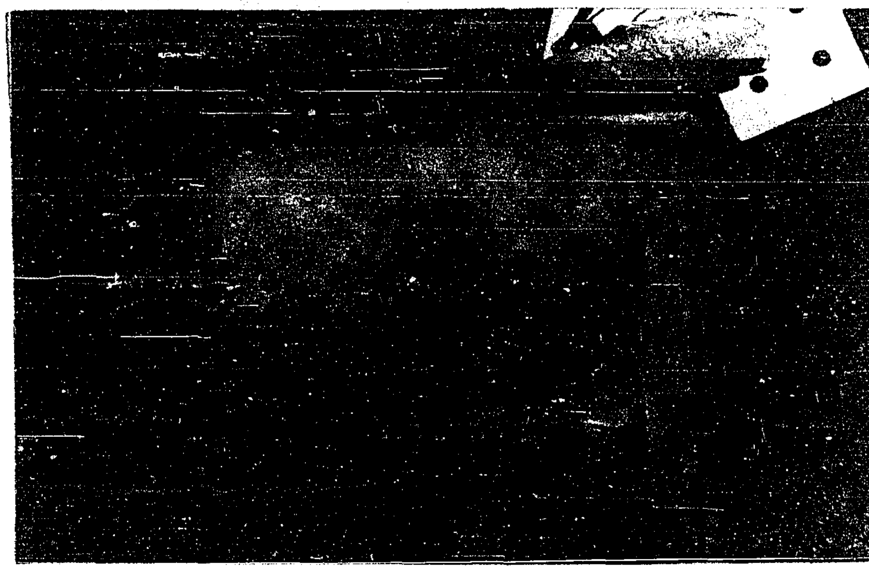
1. The cast's lower surface. Note the presence of smaller casts. This cast was taken from the Atoka formation, north-center sec. 7, T. 2 N., R. 27 E.

2. The upper half of the cast shown in 1, plate XXIV, showing the saw-cut surface of the small flute cast appearing at the top of 1, plate XXIV. The cast makes up the bottom 1 inch of the bed and the remaining part of the bed (at the top) is one-half inch thick. Cross-bedding in the small flute cast is apparent on the saw-cut surface.

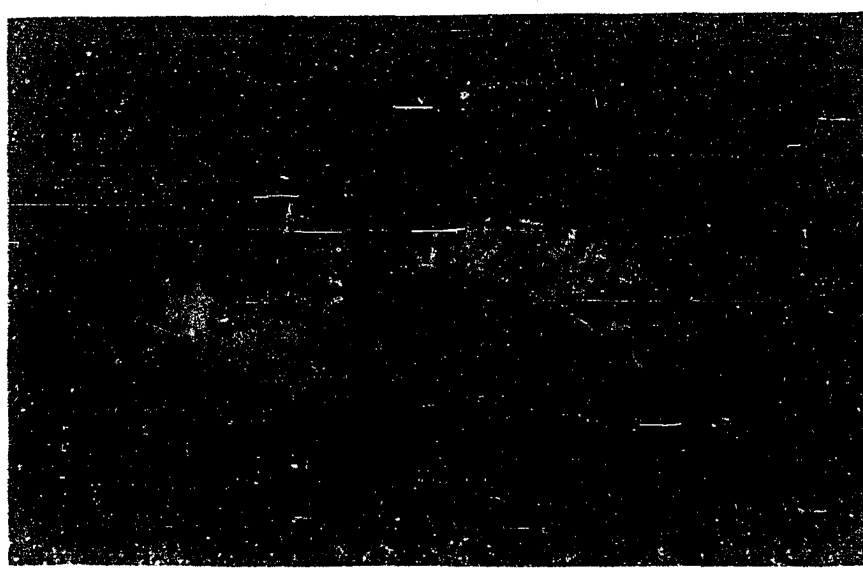
3. Saw-cut surface of 2, plate XXIV. The specimen has been stained during weathering, but cross-bedding is clearly visible. The cross-bedding indicates a current flow from right to left, which is in agreement with the flow direction shown by the flute cast shape.

PLATE XXIV

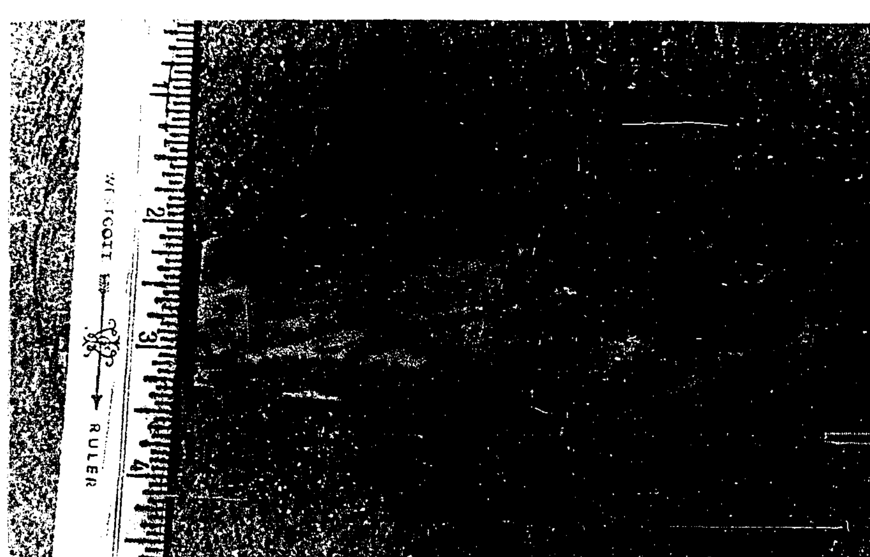
1



2



3



ever, normally these casts cannot be traced definitely to organic activity. Exceptions to this generalization are well-preserved tracks and burrows (2, pl. XXII and 1, pl. XXIII). Most non-oriented sole markings are smaller than flute or groove casts, protruding only a small fraction of an inch below the bottom of sandstone beds. A few, however, are quite large (pl. XXII).

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 study adequately.

Not enough observations were made to make a statistical statement about 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 exceeding a foot in thickness. A prominent exception to this, however, may be seen in plate XXIV. This cast protrudes 1¼ inches below the base of a ½-inch bed. It is especially interesting because of the cross-bedding which is apparent 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 greater than one foot thick is shown on plate

XX. Thick beds, however, may have only small flute and/or groove casts.

Thin beds showing abundant ripple-mark 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 3, plate XII. Casts of a thicker bed at the same locality are pictured on 2, plate XXI.

Unusual Top Surface Markings

The ridges shown on the top surface of a thick sandstone bed in 1, plate XXV 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 sub-parallel and taper towards both ends.

Channels with pitted bottoms are pictured in 2, plate XXV. These channels enlarge downdip, reaching a width of more than a foot and a depth 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 asymmetric ripple marks are present in the same exposure and may be seen in the background of 2, plate XXV. 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 down-current.

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

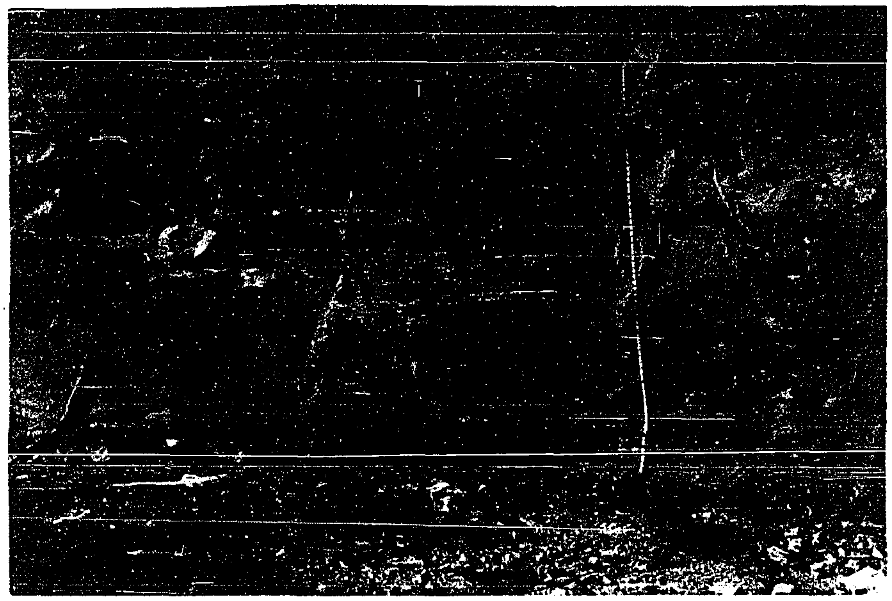
PLATE XXV

Unusual Top Surface Markings

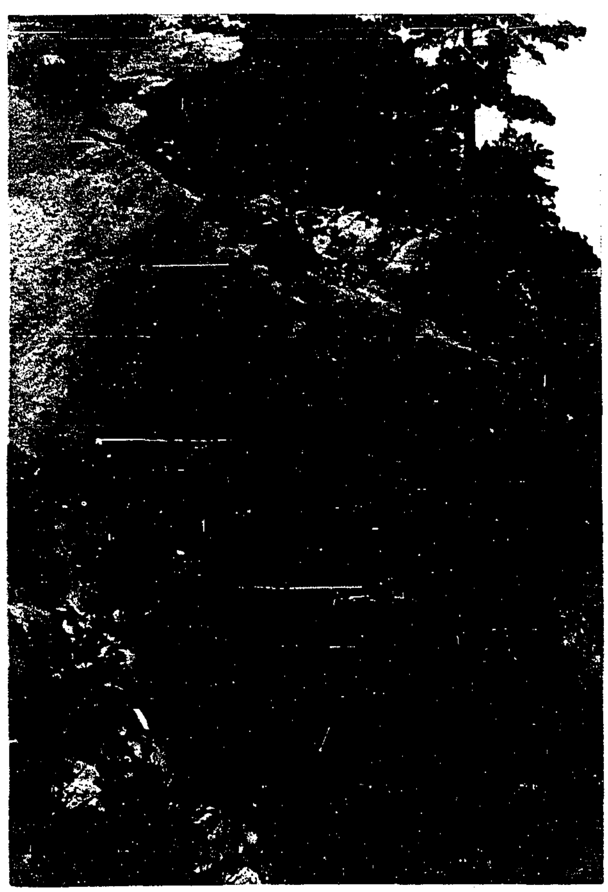
1. The dip slope of a sandstone in Interval 62, Rich Mountain measured section (Game Refuge formation). The shape of some ridges suggests that they may be casts of animals or plants, but, if so, the matter composing them has not been fossilized.

2. Two channels are present on top of the thick sandstone of the Wildhorse Mountain formation in the foreground (the pick lies in one channel). In right background, the ripple-marked upper surface pictured in 2, plate XI, is visible. Current direction indicated by the ripples is down-dip (toward the observer). Orientation of the channels parallel to dip suggests that they are also of current origin and they parallel current direction. The exposure is in the railroad cut in SE¼ NE¼ sec. 8, T. 1 S., R. 31 W.

PLATE XXV



1



2

the heading "Bottom Surface Markings" has cross-bedding indicating that the bulbous end points 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 than 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 end of flute casts points upcurrent.

Groove casts may or may not occur with flute casts. Where occurring with flute casts, the long axes of groove casts approximately parallel those of the flutes. This 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 direction which the current travelled along their length cannot be determined.

In the present study, spacial orientation of flute and groove casts was recorded at several localities. No attempt was made to compute a statistical average of the several casts in a single exposure. Several writers have pointed out the existence of variations in cast orientation at a single outcrop. In general, these variations have been small in comparison to regional flow patterns. In the area here under study, orientations in the outcrops observed did not vary by more than a few degrees and the average was estimated. They are recorded along with other current direction indicators in table I and plotted on figures 4 and 5.

Ripple marks with associated cross-bedding are also useful indicators of ancient currents. However, as pointed out in the discussion of ripple marks, exposures of them from which current direction may be deter-

TABLE I

INFERRED CURRENT DIRECTIONS FROM RIPPLE-MARKS, CROSS-BEDDING, AND SOLE MARKING ORIENTATIONS

Stratigraphic Position	Location	Type of[1] Marking	Bedding Attitude	Corrected[2] Current Direction	Structural[3] Complexity at Locality	Reliability[4]
Atoka (?)	NE NE 12, T 3 N, R 25 E	Flute and groove casts	N 57 E, 66 S (overturned)	S 87 W	High	Low
Atoka	SE NE 1, T 3 N, R 25 E	Flute casts	N 73 W, 85 S	N 30 W	Low	Low
Atoka	NW SW 11, T 3 N, R 25 E	Flute casts	N 90 W, 72 N	N 60 W	High	Low
Atoka	NW 13, T 3 N, R 26 E	Flute casts	N 54 W, 75 S (overturned)	S 40 W	High	Low
Atoka	SW 6, T 3 N, R 26 E	Flute casts	N 70 W, 47 S	N 25 W	High	Low
Atoka	SE cor. 13, T 1 N, R 32 W	Flute casts	N 71 E, 83 S (overturned)	S 30 E	High	Low
Atoka	NW NE 23, T 1 N, R 32 W	Groove casts	N 60 W, 48 S	N 75 W or S 75 E	High	Low
Atoka	SW NW 21, T 1 S, R 32 W	Flute casts	N 82 W, 90 N	N 82 W	Low	High
Atoka	SE NW 18, T 1 S, R 32 W	Flute casts	N 87 W, 54 S	S 73 W	Low	Low
Atoka	NE cor. 22, T 1 N, R 31 W	Groove casts	N 57 E, 75 N	N 87 E or S 87 W	High	Low
Lower Atoka	SE 23, T 3 N, R 25 E	Flute casts	N 40 W, 90 N (Approx.)	N 50 W	High	Low
Lower Atoka	NE SW 24, T 3 N, R 25 E	Groove casts and cross-bedding	N 77 W, 66 S (overturned)	N 60 W	Low	Moderate
Lower Atoka	NW SE 17, T 3 N, R 26 E	Flute casts	. . . .	S 65 W	Low	High
		Ripple marks with no pronounced asymmetry	. . . .	N 10 E or S 10 W	Low	High
Lower Atoka	NE NE 17, T 3 N, R 26 E	Groove casts	N 88 W, 26 N	N 88 W or S 88 E	Low	High
Lower Atoka	NW 7, T 2 N, R 27 E	Flute and groove casts	N 48 E, 20 S	N 80 W	Low	High
Lower Atoka	NE 9, T 2 N, R 27 E	Groove casts	N 88 E, 41 S	N 80 W	High	Moderate

TABLE I—Continued

Stratigraphic Position	Location	Type of[1] Marking	Bedding Attitude	Corrected[2] Current Direction	Structural[3] Complexity at Locality	Reliability[4]
Lower Atoka	NE 9, T 2 N, R 27 E	Cross-bedding	N 78 W, 82 N	E to W	High	Moderate
Lower Atoka	NW 25, T 1 S, R 32 W	Flute casts	N 76 W, 47 N	S 74 W	Moderate	Low
Overlying Stapp Conglomerate	SW SW 7, T 3 N, R 26 E	Flute casts	N 40 W, 58 N	S 85 W	Low	High
Game Refuge	NW 6, T 2 N, R 27 E	Flute casts	N 73 E, 42 S	E or W	Low	High
Game Refuge	SW 3, T 2 N, R 27 E	Cross-bedding and ripple marks	N 86 W, 56 S	S 13 E	Low	High
Game Refuge	SW 10, T 1 S, R 31 W	Flute and groove casts	N 40 W, 50 N	N 27 W	High	Low
Game Refuge	SE NW 14, T 1 S, R 31 W	Cross-bedding and ripple marks with no pronounced asymmetry	N 60 W, 24 S	South	Low	High
Upper Jackfork	SE 26, T 3 N, R 25 E	Flute and groove casts	N 50 W, 80 S	N 35 W	High	Low
Wildhorse Mountain	SW 5, T 2 N, R 26 E	Flute and groove casts	N 85 W, 60 N	N 75 W	Moderate	Moderate
Wildhorse Mountain	SE NW 12, T 1 S, R 32 W	Flute casts	N 70 W, 52 S	S 30 W	Low	High
Wildhorse Mountain	SE NE 8, T 1 S, R 31 W	Asymmetric ripple marks	N 20 W, 39 W	West	Low	High
Base Wildhorse Mountain	SW SW 6, T 2 S, R 30 W	Groove casts and cross-bedding	N 88 E, 34 N	S 75 W	Low	High
Chickasaw Creek	NE SW 25, T 1 N, R 32 W	Ripple mark, cross-bedding, flute and groove casts	N 76 W, 48 S	West	Low	High
Upper Stanley	SE 3 T 1 S, R 31 W	Ripple mark and cross-bedding	N 80 E, 28 S	S 50 W	High	Low

TABLE I—Continued

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

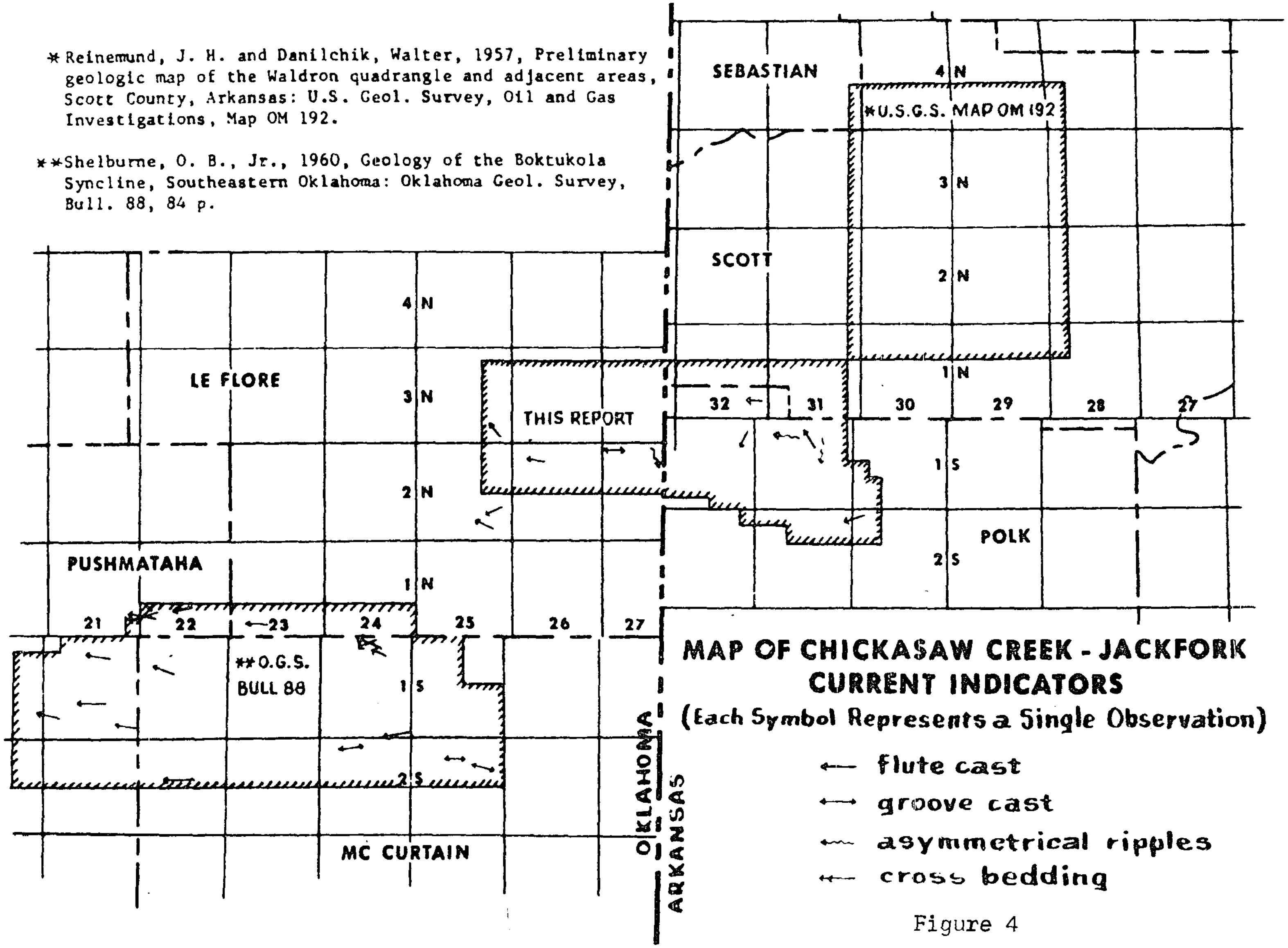
2 Corrected only for rotation of the bedding plane about the strike as an axis. The angle between the bedding strike and current direction measured in the plane of the bedding may be determined by subtracting bedding strike from corrected current direction.

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.

*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 Investigations, Map OM 192.

**Shelburne, O. B., Jr., 1960, Geology of the Boktukola Syncline, Southeastern Oklahoma: Oklahoma Geol. Survey, Bull. 88, 84 p.

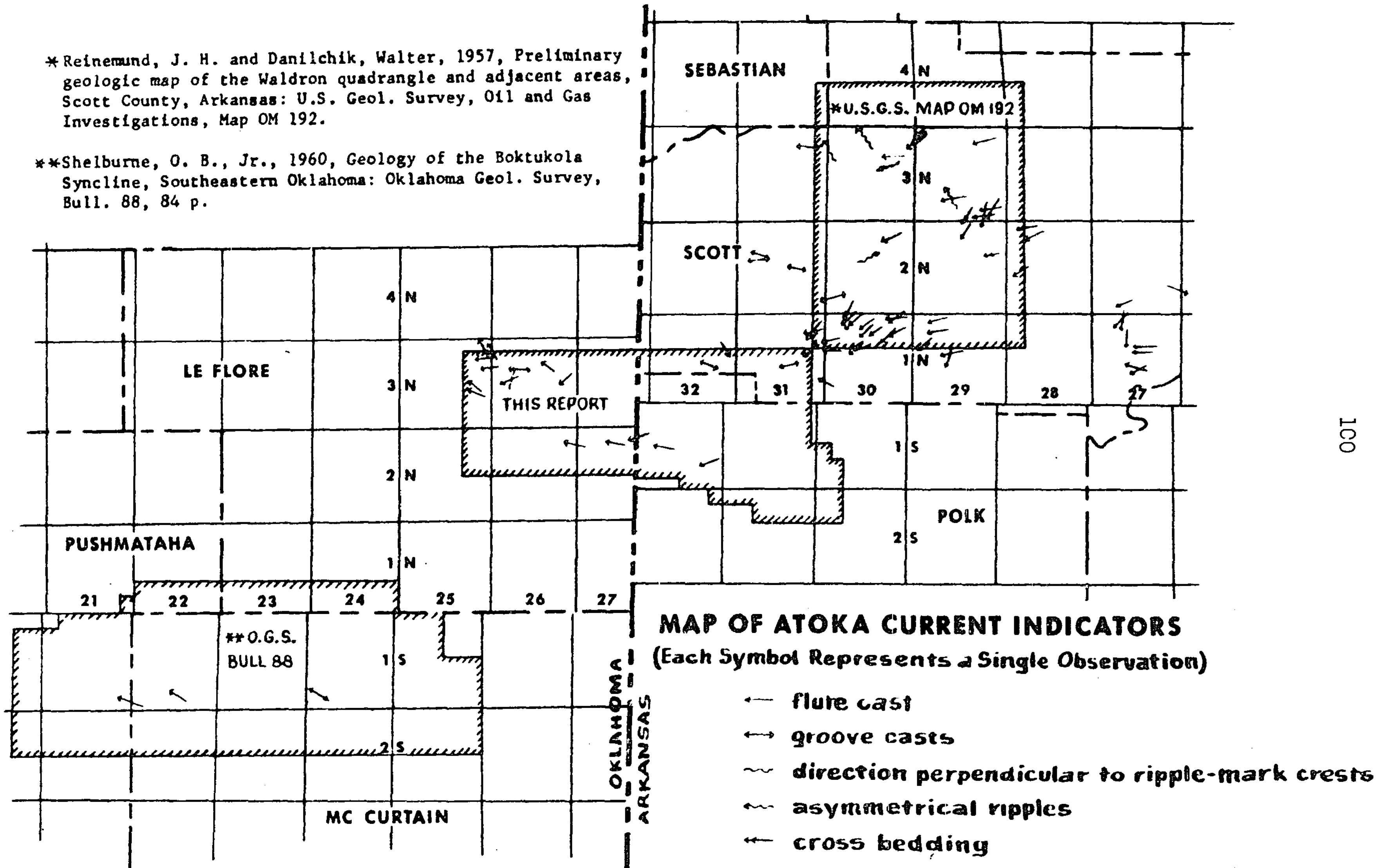


MAP OF CHICKASAW CREEK - JACKFORK CURRENT INDICATORS

Figure 4

*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 Investigations, Map OM 192.

**Shelburne, O. B., Jr., 1960, Geology of the Boktukola Syncline, Southeastern Oklahoma: Oklahoma Geol. Survey, Bull. 88, 84 p.



MAP OF ATOKA CURRENT INDICATORS
(Each Symbol Represents a Single Observation)

Figure 5

mined are uncommon. It should be noted on figures 4 and 5 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 foreset beds requires that several readings be taken at a single exposure. Because of these limitations, the writer did not use cross-bedding 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 inadequate.

Geologic Map

Base Map

The base upon which the geology is plotted was obtained from the United States Geological Survey. It consists of three 15-minute quadrangle maps that have been joined together. From east to west these quadrangles are Mena, Arkansas; Potter, Arkansas-Oklahoma; and Page, Oklahoma. The Page quadrangle is an advance print and is, therefore, subject to correction. However, the other two quadrangles are in final form.

The quadrangles were supplied in sets of two sheets each in the form of autopositive prints. In this way the topography was separated from other information shown on a topographic map. This was done in order to print the topography as a light background on the finished map.

Sources of Plotted Information

Information added to the base map was derived from two primary sources: (1) field observation; (2) aerial photographs. Reconnaissance was first carried out with the use of aerial photograph stereographic pairs. Then areas where field relationships are not clear were picked out for field checking. Most of these areas are along the southern margin of Rich Mountain syncline and to the north and west of Blackfork Mountain 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, in the sense that they are based upon a much greater portion of the bedding surface and, thus, are not affected by local slump. In most instances elevation control for these determinations was improved by the use of 7½-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. A good example may be seen in the south half of sec. 12, T. 1 S., R. 32 W.

Aerial photographs in stereographic pairs were found to be an extremely useful tool both in the office and in the field. Subtle 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 to the ground observer. The photographs also allow one to see local structure in relation to the overall structural pattern—an immense aid in the formulation of hypotheses in the field for immediate checking.

Whereas 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.

Windingstair Fault

The approximate trace of the Windingstair fault may be clearly determined from its topographic expression, as may be seen by a quick glance at the geologic map, plate I. The truncation of internal structures of Rich Mountain syncline along its southern border is readily seen at its eastern and western ends. Just to the west of the study area, Simmons Mountain is also sharply truncated by the Windingstair fault.

Precise location of the fault trace, however, is a much more difficult problem. At no spot can the writer point with certainty at the trace of the fault. 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 Windingstair fault. Faulting on an indeterminate scale is evident in beds adjacent to the zone on the south. Dip of Windingstair fault is not clear here, but the most well 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 both sides 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 this fault follows Big Creek down the valley between Rich and Blackfork Mountains. Honess discovered the fault and shows 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 (Honess, 1924). 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 during the "teens" and early "twenties." It seems appropriate to the writer to honor this man by giving his name to a major Ouachita fault originally discovered by him.

At its August 10, 1961, meeting the Domestic Names Committee of the United States Board of Geographic Names, Department of the Interior, approved the writer's request to give the name of Honess Mountain to a previously unnamed mountain whose summit 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 discovery of Honess fault solves a problem posed by the late C. W. Tomlinson. On trips through the valley between Rich and Blackfork Mountains, he noted that beds making up Blackfork Mountain appear to be dipping south rather than north. The latest editions of geologic maps of both Oklahoma and Arkansas indicate that north dip is to be expected.

Tomlinson suggested that this and other problems of the area should be investigated when the writer wrote him in search of a possible dissertation topic.

Along most of the fault trace in the study area, beds of the upper part of the Tenmile Creek formation are brought into contact with beds of the upper Jackfork younger than the Wildhorse Mountain formation; consequently, the stratigraphic throw is 5,000 to 7,000 feet. Isolated fault exposures occur in the streambeds 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 Stanley to the south of them and Jackfork to the north. There is no control for placement of the trace of the fault near its juncture with Briery fault south and west of Page, Oklahoma.

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

Briery Fault

This fault is continuous across the study area and extends out of the area on both its east and west sides. It is named for Briery Creek, which follows its trace for a short distance along the north base of Blackfork Mountain. As with other major faults, much of the trace of 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 Highway 103.

PLATE XXVI

Panoramas seen from Rich Mountain Fire Tower

1. Looking southeastward across rocks of the north flank of Rich Mountain syncline, Round Mountain anticline, and Long Mountain syncline (in order of increasing distance). Tree-covered cuestas formed by southward-dipping beds of the north flank of Rich Mountain syncline are present in the middle distance. These cuestas appear to abut the ridge of Round Mountain (which runs parallel to the top of the picture in the background). The trace of Round Mountain fault lies at the north (side nearest observer) base of Round Mountain. Ridges formed by resistant older Paleozoic beds of the Benton Uplift (Core Uplift of Arkansas) underlie the horizon.

2. Looking eastward across the settlement of Eagle Gap, Arkansas. Fourche Mountain is in the right middle distance and Shut-in Mountain is in the left middle distance. The eastern end of Blackfork Mountain projects from the middle left margin. Ridges of resistant beds in Blackfork syncline are dimly visible beyond Shut-in Mountain.

PLATE XXVI

1

2

Fossiliferous Atoka sandstones form low ridges north of Blackfork 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 Briery fault occurs here and could be as great as 25,000 feet as beds of the Tenmile Creek formation are in contact with beds that are probably near the top of the Atoka formation. On the south edge of Spring Mountain syncline, the stratigraphic throw reaches a minimum of no more than a few thousand feet.

Rich Mountain Syncline

This name is applied to the large synclinal 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 mid-way along the length of the structure near the Arkansas-Oklahoma state line. At the western end of the syncline, dip of the axial surface is 42 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", Chapter III).

The axial trace of 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. 27 E., R. 2 N. and sec. 25, T. 1 S., R. 32 W. The major syncline to the west of sec. 7, T. 27 E., R. 2 N. is actually a separate fold from that to the east. The axial traces of both folds are referred to as representing Rich Mountain syncline because they form the dominant synclinal structures along its length and have a north flank common to both of them.

The south flank of 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 a creek bearing that name that 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 Rich Mountain syncline. At its eastward termination by the Windingstair fault, the axial surface of Horsepen anticline dips about 50 degrees southward and the anticline is overturned. To the west, the tight fold passes into a thrust fault which will be referred to as "Horsepen fault". Its dip is difficult to determine accurately, but probably is between 15 and 40 degrees.

There are many small-scale structures in the general area of sec. 34, T. 1 S., R. 31 W. and exposed on Round Mountain, Middle Mountain, and Long Mountain. Only the larger features are mapped, however. One of these is an east-west fault whose trace occurs at the north base of Round Mountain and continues westward to Pilot Mountain where it joins Windingstair fault. To the east it is continuous with the axial trace of a syncline whose axial surface dips southward. It is here named the "Round Mountain fault," and the anticline bordering it on the south is "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

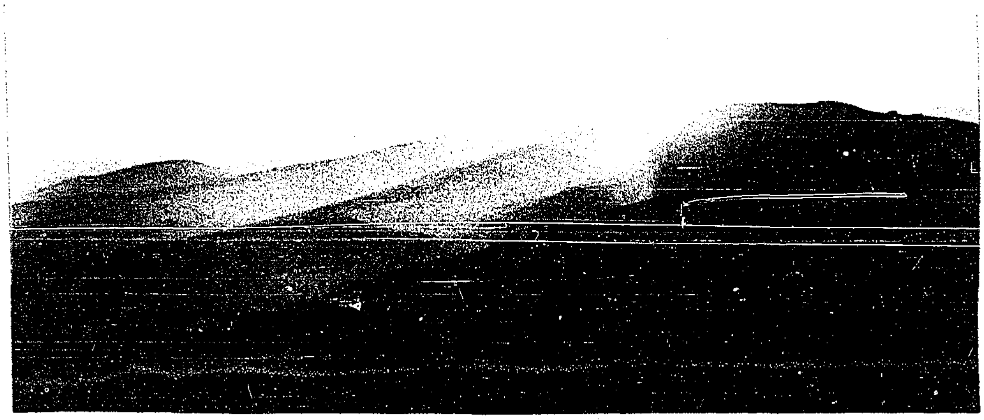
PLATE XXVII

Scenes from the Crest of Rich Mountain

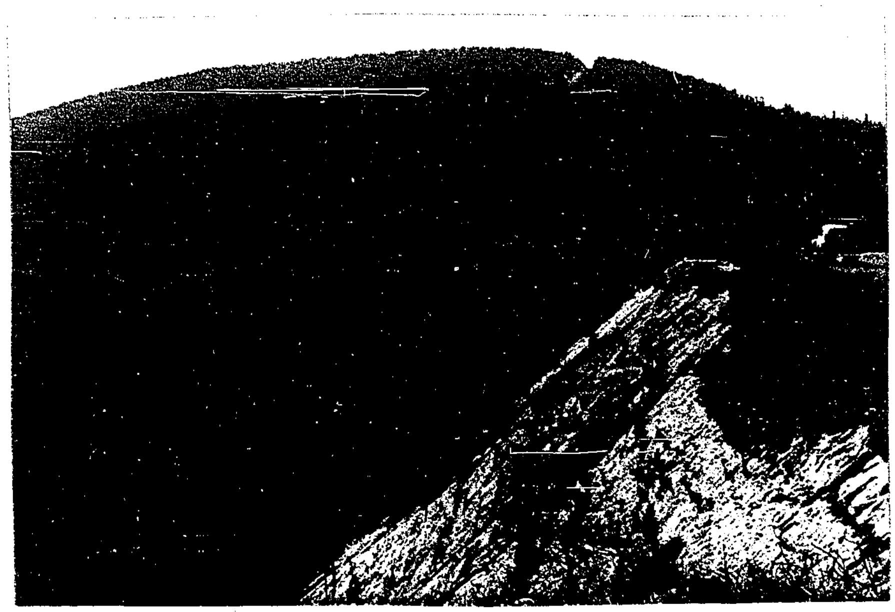
1. Looking westward from Rich Mountain fire tower. Topographic expression of units described in Rich Mountain measured section is well shown in this picture. The high ridge on the right is underlain by beds of Wildhorse Mountain formation. The valley to the south of it has been carved in the soft beds of the Prairie Mountain - Markham Mill - Wesley undifferentiated sequence. The contact of this sequence with the Game Refuge formation appears as a vegetation change (deciduous to evergreen) near top of the valley's south slope. The Game Refuge caps this slope forming the first low ridge south of the high, Wildhorse Mountain ridge. Easily eroded beds of the Johns Valley - Atoka sequence underlie the valley separating the Game Refuge ridge from a similar ridge to the south that is underlain by fossiliferous sandstones of the Atoka formation. To the south (left) of the Atoka ridge are ridges formed by rocks of the truncated south flank of Rich Mountain syncline. Bordering these ridges on the south is Windingstair fault and the Kiamichi Valley.

2. Looking westward at dip slopes of southward-dipping sandstones underlying the crest of Rich Mountain. Picture taken on Skyline Drive just west of Rich Mountain fire tower.

PLATE XXVII



1



2

best in secs. 34 and 35, T. 1 S., R. 31 W. and is poor at the east and west ends of the structure.

To the south of Round Mountain anticline is a syncline whose south flank is exposed on Long Mountain. The name "Long Mountain syncline" seems appropriate for this structure. 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 Long Mountain syncline.

Blackfork Mountain Syncline

Blackfork Mountain is the next ridge to the north of Rich Mountain, and rocks composing it make up the north flank of a syncline whose south flank is almost entirely missing in surface exposures. The syncline is named for Blackfork Mountain. It is a different feature from the Blackfork syncline, which was named by Reinemund and Danilchik (1957) in the nearby Waldron quadrangle.

Evidence that the structure of Blackfork Mountain is synclinal in nature 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 Blackfork Mountain, Fourche Mountain is considered a continuation of Blackfork Mountain syncline. No suggestion of a south limb was observed on Fourche Mountain.

Spring Mountain Syncline

This is the third largest syncline that lies almost entirely within the study area. It is crossed by Highway 103 near the western edge of the area. There is a good northward outlook from Spring Mountain summit on the highway because Spring Mountain, from which the syncline derives its name, is the highest ridge in the syncline. This ridge is made up of resistant sandstones at the base of the Atoka formation and is part of the south flank of the syncline.

Highway 103 follows the axial trace of the structure in secs. 17, 18, and 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 it 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 Railway cut in SW¼ SW¼ sec. 7, T. 3 N., R. 26 E. These beds form part of the south flank of a syncline trending WNW-ESE. Because of this relationship between the conglomerate and the structure of which it is a part, the structure will be referred to as the Stapp syncline.

To the west the syncline plunges eastward, but to 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 asymmetrical and probably has a southward-dipping axial surface. Stapp syncline appears symmetrical, but the inconsistency of dips along its south flank suggests that they may have been reduced by downslope movement. If this is true it may be asymmetrical also.

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.

Blackfork Syncline

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

A short segment of the south flank is believed to be present in the study area. It is represented by the southwest tip of Horseshoe Mountain in sec. 13, T. 1 N., R. 32 W. The north flank borders the study 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) show 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 writer's interpretation of the geology of the area differs from that of Reinemund and Danilchik. Shut-in anticline is bordered on the north by a syncline which they did not map. Outcrops of Johns Valley shale mapped by Reinemund and Danilchik were not found to the north. Chert present is believed to be that of the Wesley shale rather than that of the Johns Valley formation.

CHAPTER II

MICROSCOPIC 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 also included in Appendices E, F, and G. Photomicrographs of most of the study slides may be seen on plates XXVIII through XXXIV.

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 is composed of sandstone. They were chosen on the basis of freshness of hand specimen, stratigraphic unit represented or some particularly interesting feature illustrated. If the thin sections were laid out in contiguous sequence, they would represent a total stratigraphic interval of slightly less than two feet. Such a sample is a small sample in comparison to a population about 2,500 times its size from which it was obtained.

The standard deviations of many of the sample attributes in this study are small. A uniformly small standard deviation may be due either to sampling accident or to inherent small standard deviations of these attributes in the population sampled. Studies elsewhere of the same portion of the geologic section have indicated that standard deviations of compositional and some textural properties are relatively small in many respects;

however, the definition of "small" depends upon the significance level one is seeking.

Composition

Compositions of the thin sections are summarized in Appendix B.

Quartz

Using 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 possession of 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 (above) except that there are abundant vacuoles. Class 6d is comprised of composite grains with strongly undulose extinction. Most individuals of the composite grains share crenulate boundaries in the thin sections studied (3 and 8, pl. XXXII). 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 non-crenulated internal boundaries falling into class 5d (5, pl. XXXI).

When seen by reflected light, quartz grains having many vacuoles are milky and non-vacuolized grains, colorless. Some quartz grains have an illusory weathered appearance that causes them superficially to resemble feldspars. One such grain may be seen in 3, plate XXXI. 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 normally could not 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 1, plate XXXI.

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 (8, pl. XXXI). A distinct variety of chert occurs in elongate masses that contain what is apparently carbonaceous material and have a relatively coarse uniform texture (6, pl. XXXI).

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 on plate XXX appears to be filling pore spaces and is probably an orthochemical constituent. The chert grain in 8, plate XXXI, is of uncertain derivation. The pore-filling appearance of some of the chert may be largely due to encroachment 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 8, plate XXXI).

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 on plates XXVIII and XXIX have a microcrystalline texture revealed by mica flakes. The fine texture and absence of porphyroblasts suggest that these are slate or phyllite. They represent 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 3 and 8, plate XXXII, 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 of photomicrographs 9, plate XXXI, and 7, plate XXXII. 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.

Heavy Minerals

In order of decreasing abundance, the heavy minerals observed are zircon, tourmaline and garnet. Zircon is far more abundant than 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 (3 and 6, plate XXIX) but may be angular. Several grains show lower degrees of rounding and a few are euhedral (6, pl. XXIX). Because of the small number of grains, no attempt to differentiate the varieties of each heavy mineral was made.

The Micas

Biotite, muscovite and penninite are the only micas of sufficient grain size and diagnostic optical properties 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.

In order of decreasing abundance the large mica flakes should be listed muscovite, biotite, and penninite. A muscovite flake is visible in the right center of 3, plate XXX. 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.

Spicules

Two distinct forms of microfossils are recognizable in RM 23A. One of these may also be observed in lesser abundance in MR 4-2. This form that is present in both slides probably consists of sponge spicules. Transverse cross-sections of them are circular in outline and 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 in length 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. Reliable identification of this form could not be made. Its external shape appears similar to that of the Tertiary radiolarian, <u>Rhopalodictyum</u> (<u>Rhopalodictyum</u>) <u>malaganese</u> Campbell and Clark (Moore, Lalicker, and Fischer, 1952, p. 74). On the other hand, it may be a planar three-rayed megasclere from a species other than that whose spicules were described first above or a microsclere from that same sponge. The two microfossil forms may be seen on photomicrographs 8 and 9, plate XXIX.

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.

Unidentifiable Microfossils

Fragmentary siliceous organic debris that is too small for identification is placed in this category. Also included here are fusiform silica deposits such as that pictured in 4, plate XXXIII.

Opaque Components

Carbonized plant material is the prevalent opaque component found in the study slides. There is a great variation in its 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 (pls. XXXII and XXXIII). Both of the 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. 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.

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.

There are at least two factors of importance that affect the validity of the mean and 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; (2) the affect of thin-sectioning on size parameters must be estimated.

All analysis was done by viewing slides through a petrographic microscope. There have been various publications concerning the reliability of size analysis accomplished in this way. In an effort to determine the possible error involved in the present investigation, thin section parameters were corrected to their sieve equivalents by the method developed by Friedman (1958; see Figure 4, p. 408). Both the original thin-section data and the sieving equivalents are shown in Appendix C. As may be noted from 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 coarsest 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 where there is no chert cement or clay matrix the quartz grains were probably tightly packed originally and overgrowth has been relatively small when compared to original long diameters. Where chert cement or clay matrix separates grains there has been little modification of grain shape.

The mean size in most thin sections falls in the range of fine sand. Second in frequency of occurrence are thin sections whose means fall 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 in 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. Using 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 causes them all to be classified as moderately sorted.

Grain roundness. Extensive recrystallization in the sandstones has effectively masked the original roundness of most quartz grains. Only a relatively few grains show relict outlines of their border prior to overgrowth. On the other hand, where a chert cement or clay matrix is present between grains, original outlines appear to be only slightly modified. Unfortunately slight modification is all that is required in order that roundness be greatly affected.

The difficulty in estimating roundness prohibited the distinction between mature and supermature (Appendix D) for most of the orthoquartzites. A roundness of three (Folk, 1959, p. 101) separates these two classes of sediment. Normally 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 (3 and 8, plate XXXII). These are the only well rounded, granule-sized particles visible in the thin sections. Most other unusually large grains are composed of quartz and have a roundness similar to the sediment average (9, pl. XXVIII; 4 and 5, pl. XXXI). Fine-textured, stretched metamorphic quartz grains are rare. The grain shown in photomicrograph 7, plate XXXII, 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. This may be observed for zircon in photomicrograph 6, plate XXIX.

Fabric. Stratification is revealed in three ways in the sandstone thin sections: layered variation in quartz grain size; laminae of heavy minerals; 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 1 and 9, plate XXXII. 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 (3, pl. XXIX).

Recrystallization has produced an intricate intergrowth of quartz grains where there is no chert cement or clay matrix. This mosaic fabric characterizes most sandstone thin sections. The mutual boundaries between grains are straight locally, but most of their peripheries are uneven and probably reflect their original silhouettes with varying degrees of faithfulness.

In portions of several slides a chert cement or clay matrix is present. The matrix may completely 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 highest in the better sorted slides which have little or no matrix.

Holes present in the thin sections are probably of several origins. Some represent loosely held grains that have been plucked out during making of the thin section. Others probably represent softer grains that disintegrated when the thin section was made. This latter feature is evident where patches of argillaceous chert were only partially removed. Because most of the slides were obtained from rocks that show some degree of weathering, there is a question as to how much of the more easily weathered materials has been removed in this manner. The smallest holes in the thin sections are apparently the result of incomplete filling of pores by quartz overgrowths (7, pl. XXVIII). 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 while 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 in thickness.

PLATE XXVIII*

Photomicrographs of Rocks in Rich Mountain Measured Section

1. Thin section RM 40, crossed nicols, field diameter is 0.416 mm (X125). Crushed metamorphic rock fragments and recrystallized quartz grains, some of which share mutual straight boundaries. The black areas are holes in the slide. Atoka formation.

2. Thin section RM 40, crossed nicols, field diameter is 0.208 mm (X250). 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.

3. Thin section RM 40, crossed nicols, field diameter is 0.416 mm (X125). 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.

4. Thin section RM 40, crossed nicols, field diameter is 1.04 mm (X50). A detrital siliceous shale fragment or fine-textured metamorphic rock particle surrounded by recrystallized quartz. The black areas are holes in the slide. Atoka formation.

5. Thin section RM 40, crossed nicols, field diameter is 1.04 mm (X50). Glauconite and metamorphic rock fragments in recrystallized quartz. The glauconite may be located by comparison with 6, plate XXVIII. Atoka formation.

6. Thin section RM 40, crossed nicols, field diameter is 0.208 mm (X250). Finely crystalline glauconite forms the central portion of this photomicrograph. Atoka formation.

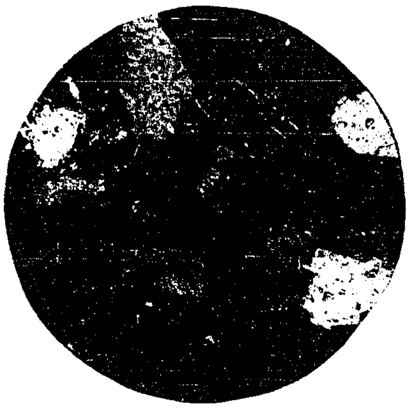
7. Thin section RM 40, crossed nicols, field diameter is 0.416 mm (X125). A mosaic fabric which has resulted from extensive recrystallization. Atoka formation.

8. Thin section RM 29A, crossed nicols, field diameter is 4.16 mm (X12.5). Porous sandstone which contains a mold fauna (not visible). Atoka formation.

9. Thin section RM 29A, crossed nicols, field diameter is 4.16 mm (X12.5). Large quartz grain which is a probable approximate hydraulic equivalent to the invertebrate fragments that have been dissolved from other portions of the rock. Both 8 and 9, plate XXVIII, have been greatly fractured during manufacture of the thin section. Atoka formation.

* For a more exact determination of stratigraphic position see the Cross Reference Index for Thin Sections and Measured Sections, Appendix E.

PLATE XXVIII



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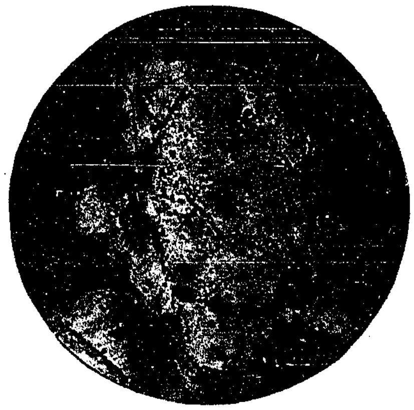

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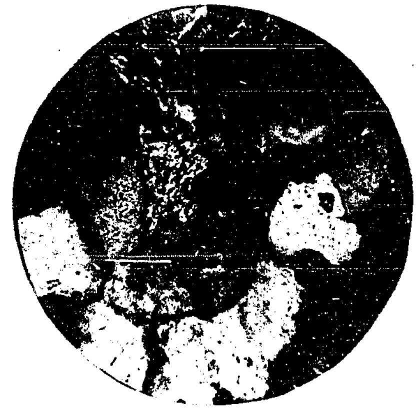

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PLATE XXIX*

Photomicrographs of Rocks in Rich Mountain Measured Section

1. Thin section RM 29, ordinary light, field diameter is 10.4 mm (X5). Cross-bedding and an unconformity on a small scale. The unconformity is at the base of the photomicrograph 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.

2. Thin section RM 29, crossed nicols, field diameter is 1.04 mm (X50). Close-up of a siliceous shale particle in one of the cross-laminations. Atoka formation.

3. Thin section RM 29, ordinary light, field diameter is 1.04 mm (X50). A close-up of the unconformity showing the concentration of zircon grains along it. The black area at the unconformity at the far right is a black opaque heavy mineral. Atoka formation.

4. Thin section RM 24A, crossed nicols, field diameter is 1.04 mm (X50). Mosaic fabric due to the intergrowth of quartz crystals during recrystallization. Game Refuge sandstone.

5. Thin section RM 24A, crossed nicols, field diameter is 0.416 mm (X125). Mosaic fabric. Game Refuge sandstone.

6. Thin section RM 24A, ordinary light, field diameter is 0.416 mm (X125). Rounded and euhedral zircon grains that are part of a thin heavy mineral stratum. Game Refuge sandstone.

7. Thin section RM 23A, ordinary light, field diameter is 10.4 mm (X5). Stratification due to varying concentrations of carbonized plant matter in siliceous shale. Game Refuge sandstone.

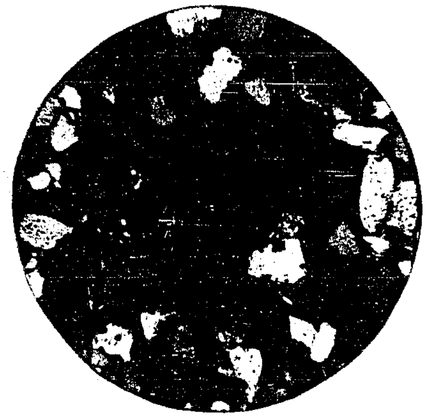
8. Thin section RM 23A, ordinary light, field diameter is 0.416 mm (X125). Cross-sections of sponge spicules (circular), organic debris and possible radiolarians (3-rayed) in a stratum containing a relatively high amount of carbonized plant matter. Game Refuge sandstone.

9. Thin section RM 23A, crossed nicols, field diameter is 0.416 mm (X125). The same view as 8, plate XXIX, except that nicols are crossed. Game Refuge sandstone.

* For a more exact determination of stratigraphic position see the Cross Reference Index for Thin Sections and Measured Sections, Appendix E.

PLATE XXIX

1



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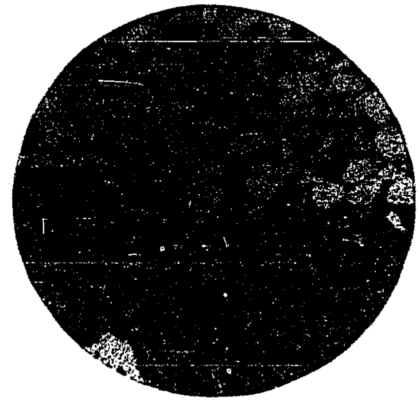

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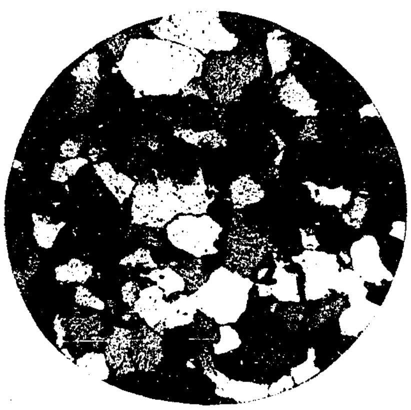

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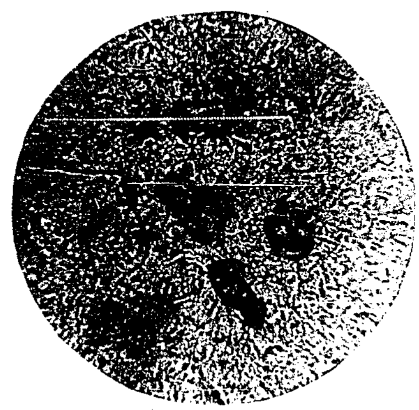

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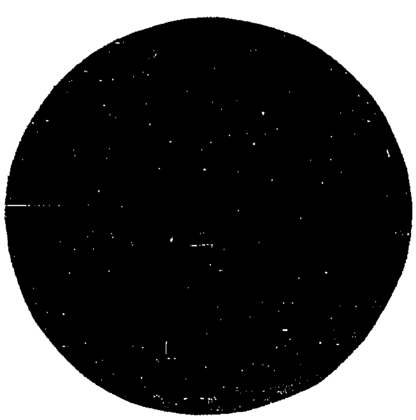

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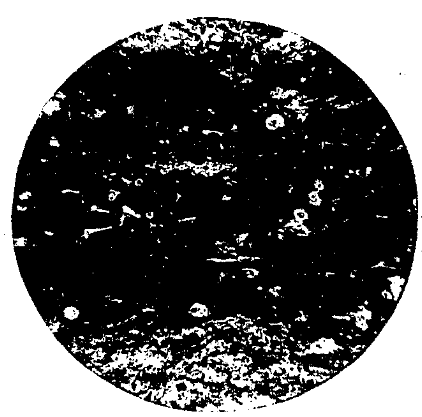

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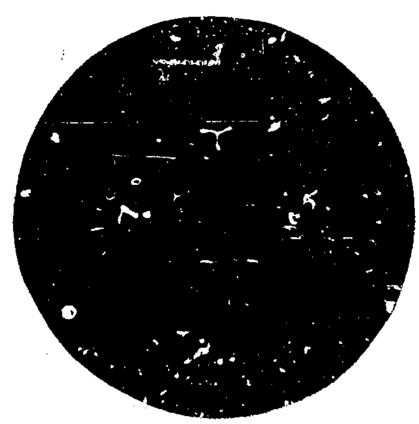

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PLATE XXX*

Photomicrographs of Rocks in Rich Mountain Measured Section

1. Thin section RM 22A-1, crossed nicols, field diameter is 4.16 mm (X12.5). A thin, graded, silt stratum in siliceous shale (or shaly chert, depending upon the dominant constituent). Wesley shale.

2. Thin section RM 21A, ordinary light, field diameter is 4.16 mm (X12.5). 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.

3. Thin section RM 21A, crossed nicols, field diameter is 0.208 mm (X250). Plagioclase, muscovite, and chert surrounded by quartz grains. Markham Mill - Prairie Mountain undifferentiated.

4. Thin section RM 21A, crossed nicols, field diameter is 1.04 mm (X50). Cluster of metamorphic rock fragments, plagioclase and chert in quartz grains. Markham Mill - Prairie Mountain undifferentiated.

5. Thin section RM 21A, crossed nicols, field diameter is 0.416 mm (X125). A perthitic feldspar grain may be seen in the center of the field. Markham Mill - Prairie Mountain undifferentiated.

6. Thin section RM 20, crossed nicols, field diameter is 1.04 mm (X50). 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.

7. Thin section RM 20, crossed nicols, field diameter is 1.04 mm (X50). A metamorphic rock fragment may be seen near the center of the field. Markham Mill - Prairie Mountain undifferentiated.

8. Thin section RM 20, crossed nicols, field diameter is 0.416 mm (X125). A close-up of the metamorphic rock fragment seen in 7, plate XXX. Notice that it has been squeezed into inter-grain spaces. Markham Mill - Prairie Mountain undifferentiated.

9. Thin section RM 20, crossed nicols, field diameter is 0.416 mm (X125). A close-up of a portion of the field shown in 6, plate XXX. Markham Mill - Prairie Mountain undifferentiated.

* For a more exact determination of stratigraphic position see the Cross Reference Index for Thin Sections and Measured Sections, Appendix E.

PLATE XXX

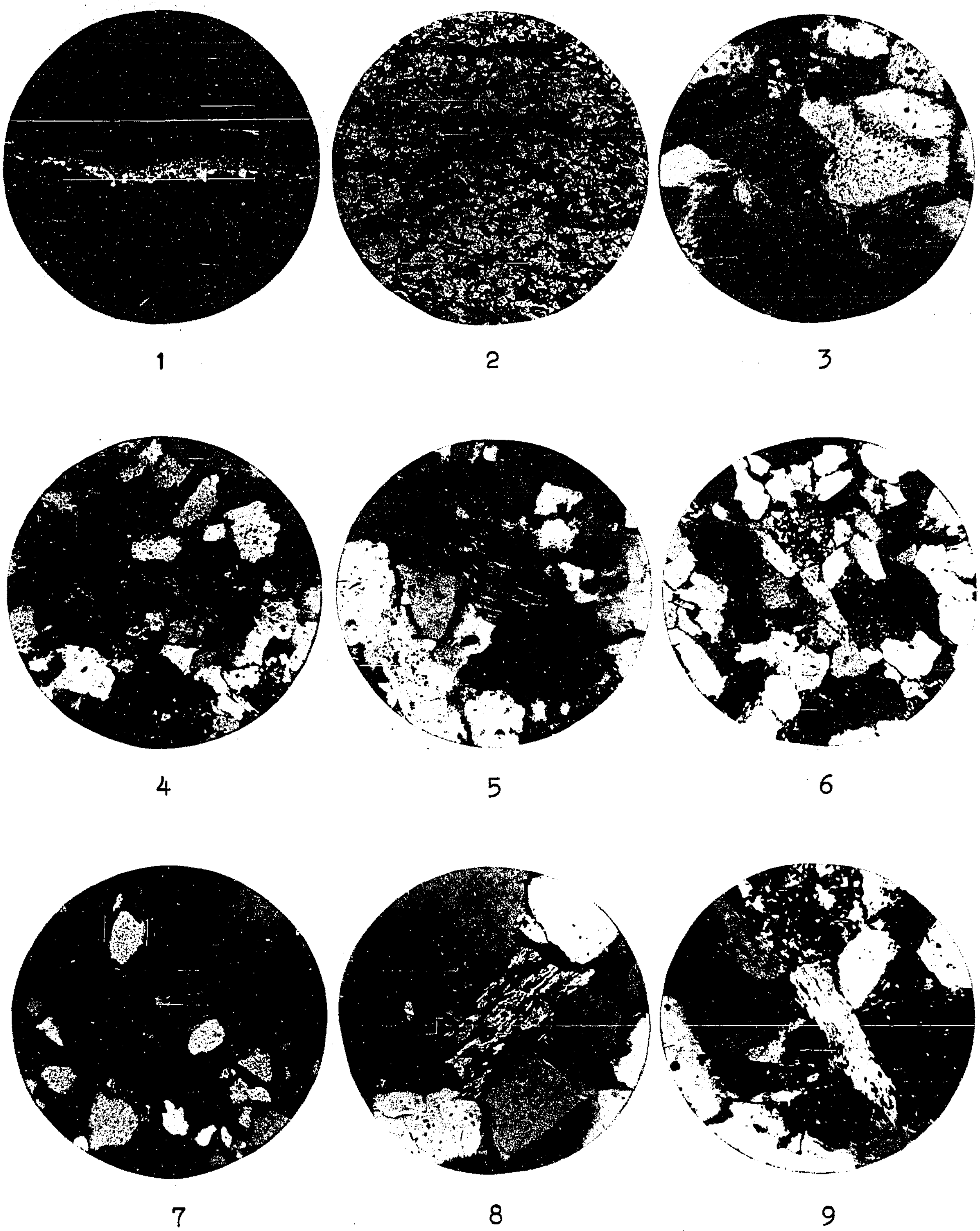


PLATE XXXI*

Photomicrographs of Rocks in Rich Mountain Measured Section

1. Thin section RM 19, crossed nicols, field diameter is 1.04 mm (X50). A tentatively identified microcline grain tightly hemmed-in by recrystallized quartz. Markham Mill - Prairie Mountain undifferentiated.

2. Thin section RM 19, crossed nicols, field diameter is 1.04 mm (X50). Chert surrounded by quartz grains of a wide size range. Markham Mill - Prairie Mountain undifferentiated.

3. Thin section RM 19, crossed nicols, field diameter is 1.04 mm (X50). 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.

4. Thin section RM 19, crossed nicols, field diameter is 10.4 mm (X5). A large grain of strained quartz with irregular extinction is shown in the center of the field. Markham Mill - Prairie Mountain undifferentiated.

5. Thin section RM 14, crossed nicols, field diameter is 1.04 mm (X10). A large angular composite quartz grain surrounded by metamorphic rock fragments and smaller single-crystal quartz grains. Wildhorse Mountain formation.

6. Thin section RM 11A, crossed nicols, field diameter is 0.416 mm (X125). A coarse-textured particle of chert. Wildhorse Mountain formation.

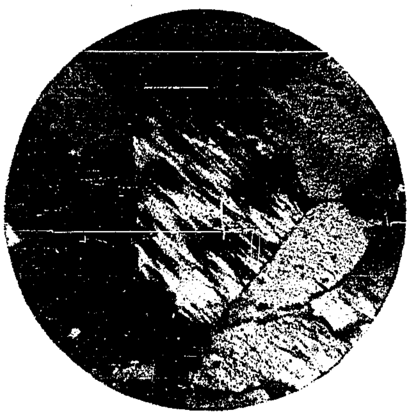
7. Thin section RM 11A, crossed nicols, field diameter is 1.04 mm (X50). Two particles of chert, one finely crystalline and the other coarsely crystalline, may be seen for comparison. Wildhorse Mountain formation.

8. Thin section RM 11A, crossed nicols, field diameter is 0.208 mm (X250). The finely crystalline chert particle of 7, plate XXXI. Note the variation of crystal size within the grain. Wildhorse Mountain formation.

9. Thin section RM 11A, crossed nicols, field diameter is 1.04 mm (X50). A metamorphic rock fragment may be seen in the center of the field. Wildhorse Mountain formation.

* For a more exact determination of stratigraphic position see the Cross Reference Index for Thin Sections and Measured Sections, Appendix E.

PLATE XXXI



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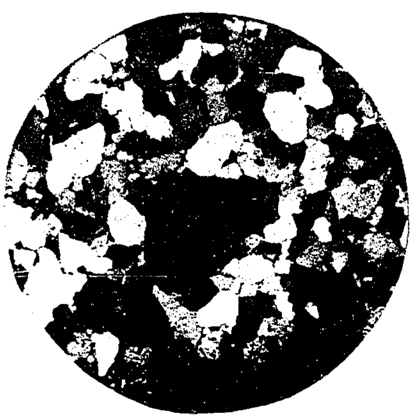

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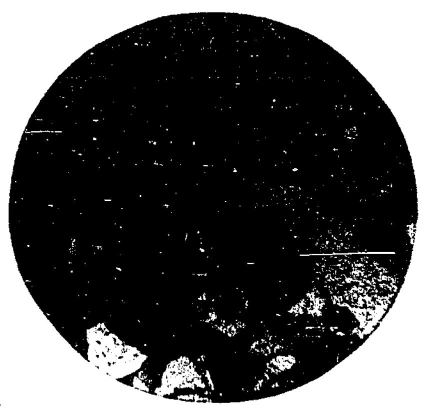

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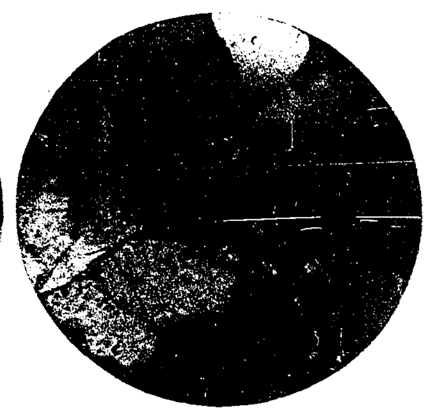

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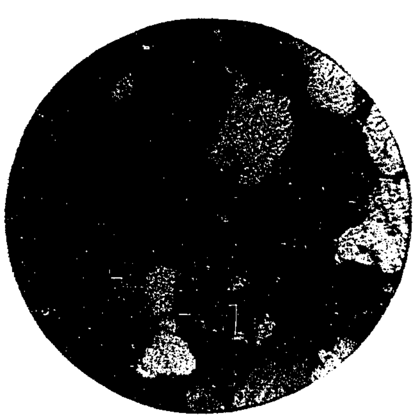

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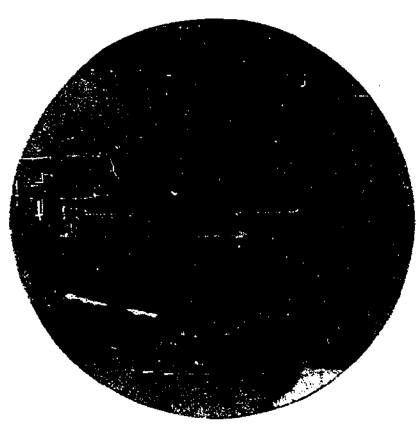

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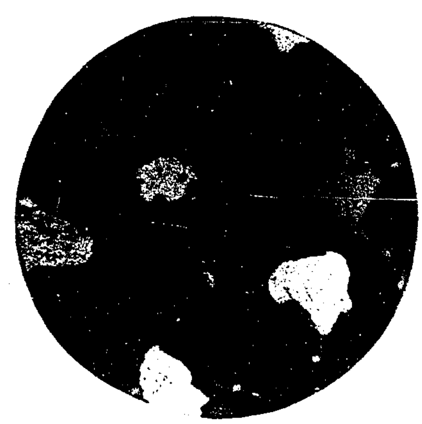

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PLATE XXXII*

Photomicrographs of Rocks in Rich Mountain and Ward Lake Spillway Measured Sections

1. Thin section RM 11, crossed nicols, field diameter is 10.4 mm (X5). Very well sorted very fine sand overlying very well sorted coarse silt. Wildhorse Mountain formation.

2. Thin section RM 8-2, crossed nicols, field diameter is 1.04 mm (X50). A metamorphic rock fragment and a chert particle may be seen near the center of the field. Wildhorse Mountain formation.

3. Thin section RM 7, crossed nicols, field diameter is 4.16 mm (X12.5). 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.

4. Thin section RM 6-1, crossed nicols, field diameter is 1.04 mm (X50). A replaced crinoid columnal is present at the center of the field. Wildhorse Mountain formation.

5. Thin section RM 6-1, crossed nicols, field diameter is 0.208 mm (X250). A close-up of the replaced columnal seen in 4, plate XXXII. Wildhorse Mountain formation.

6. Thin section RM 2-1, ordinary light, field diameter is 4.16 mm (X12.5). Stylolites outlined by carbonized plant matter. Wildhorse Mountain formation.

7. Thin section MRS 6-1, crossed nicols, field diameter is 1.04 (X50). A fine textured, stretched metamorphic quartz grain is present in the center of the field. Wildhorse Mountain formation.

8. Thin section MRS 6-1, crossed nicols, field diameter is 4.16 mm (X12.5). 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.

9. Thin section MRS 4-2, crossed nicols, field diameter is 10.4 mm (X5). Size stratification parallel to the bottom of the page may be seen in this photomicrograph. Chickasaw Creek siliceous shale.

* For a more exact determination of stratigraphic position see the Cross Reference Index for Thin Sections and Measured Sections, Appendix E.

PLATE XXXII

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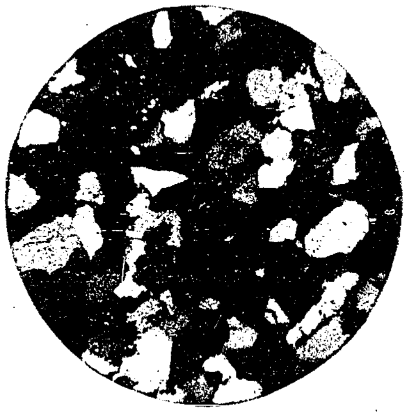

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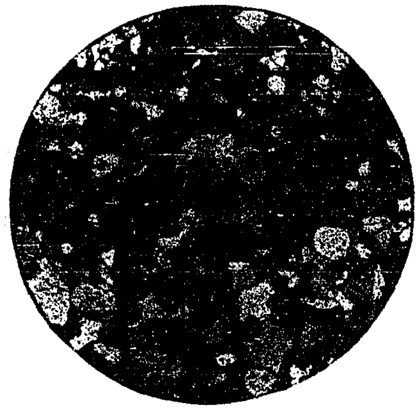

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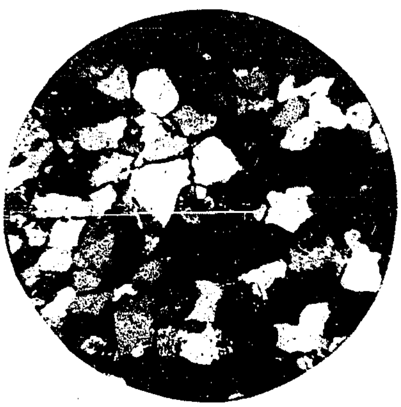

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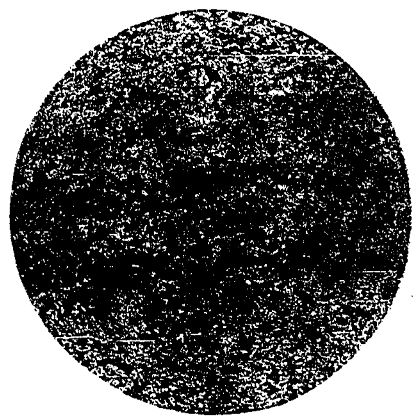

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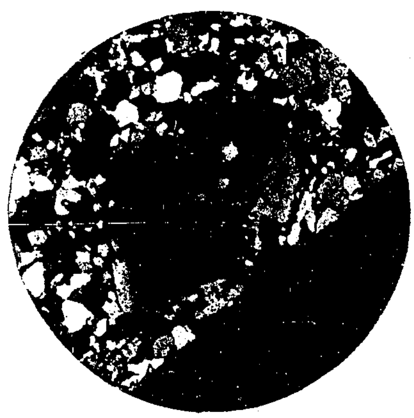

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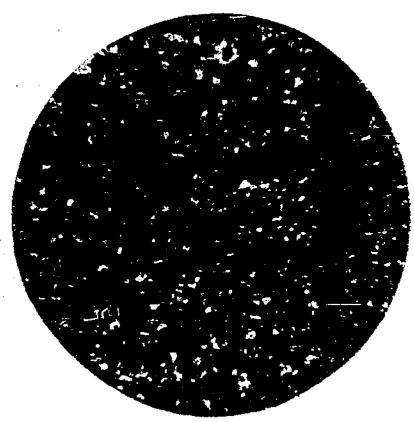

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PLATE XXXIII*

Photomicrographs of Rocks in East Ward Lake Measured Section

1. Thin section MR 6-1, crossed nicols, field diameter is 4.16 mm (X12.5). 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.

2. Thin section MR 4-2, ordinary light, field diameter is 10.4 mm (X5). Siliceous shale that is faintly laminated by varying proportions of carbonized plant matter. The white spots are holes in the slide. Chickasaw Creek siliceous shale.

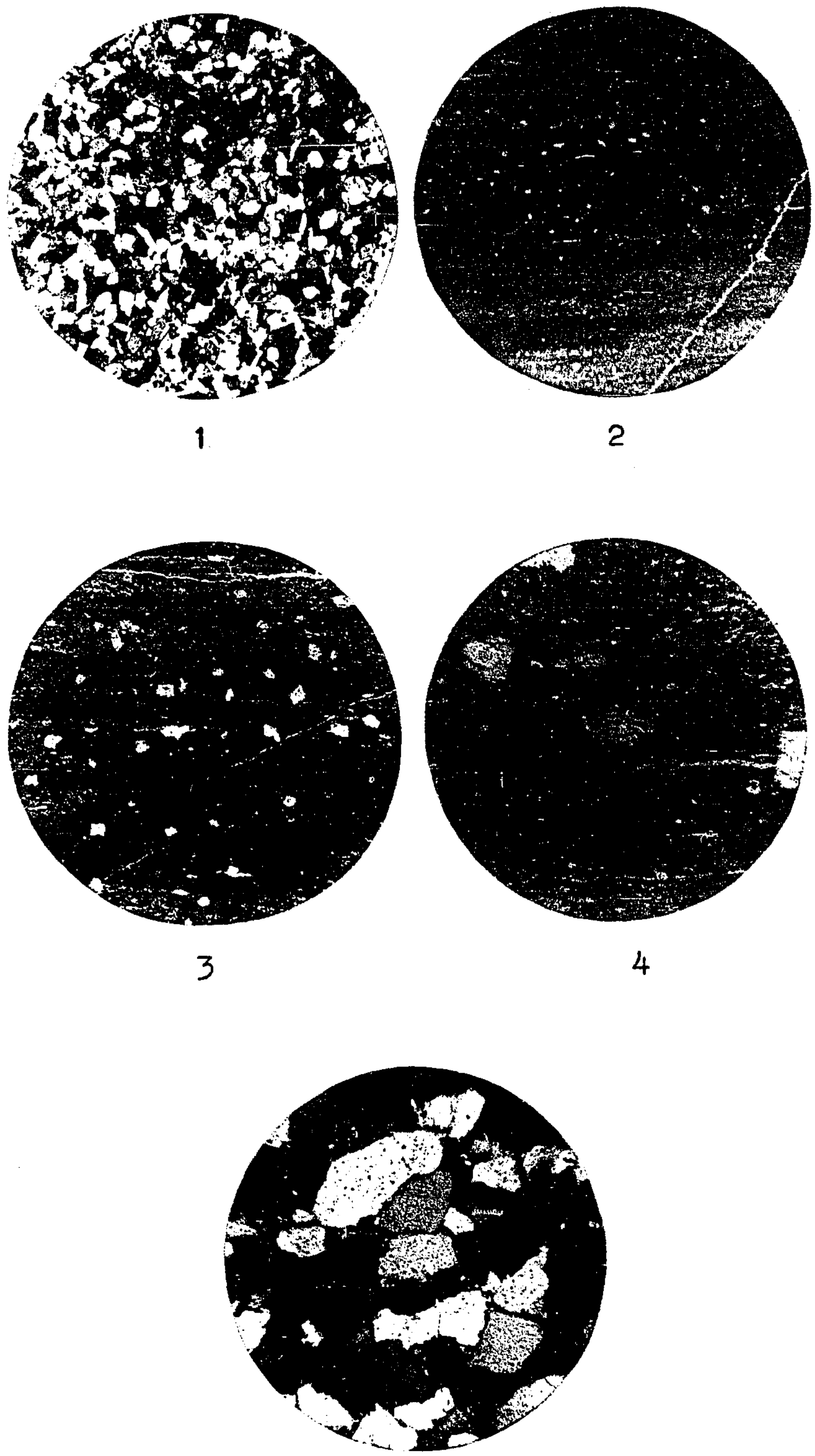
3. Thin section MR 4-2, ordinary light, field diameter is 4.16 mm (X12.5). A close-up of the white spots of 2, plate XXXIII. The sharp-cornered rhomb shape of several of the holes suggests that some of them were originally filled by authigenic mineral grains, by pyroclastics, or by angular detrital debris. Chickasaw Creek siliceous shale.

4. Thin section MR 4-2, ordinary light, field diameter is 1.04 mm (X50). The particle in the center of the field is probably of organic origin and vestiges of the original structure may be seen. Chickasaw Creek siliceous shale.

5. Thin section MR 3-1, crossed nicols, field diameter is 1.04 mm (X50). 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.

* For a more exact determination of stratigraphic position see the Cross Reference Index for Thin Sections and Measured Sections, Appendix E.

PLATE XXXIII



1 2 3 4 5

PLATE XXXIV*

Photomicrographs of Tuff

1. Thin section 7-20-5F, ordinary light, field diameter is 4.16 mm (X12.5). 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).

2. Thin section 7-20-5F, crossed nicols, field diameter is 4.16 mm (X12.5). The same view as 1, plate XXXIV, except that nicols have been crossed.

3. Thin section 7-20-5F, crossed nicols, field diameter is 1.04 mm (X50). A close-up of the large andesine grain seen in the lower right of 1 and 2, plate XXXIV. Replacement by calcite may be seen around the margins of the grain. The ribbon-like feature curving across the grain is Canada balsam.

4. Thin section 7-20-5F, crossed nicols, field diameter is 1.04 mm (X50). 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.

5. Thin section 7-20-5F, crossed nicols, field diameter is 1.04 mm (X50). The mold of 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.

6. Thin section 7-20-5F, crossed nicols, field diameter is 0.416 mm (X125). Calcite replacement of this feldspar (?) grain has progressed farther than it has in other grains.

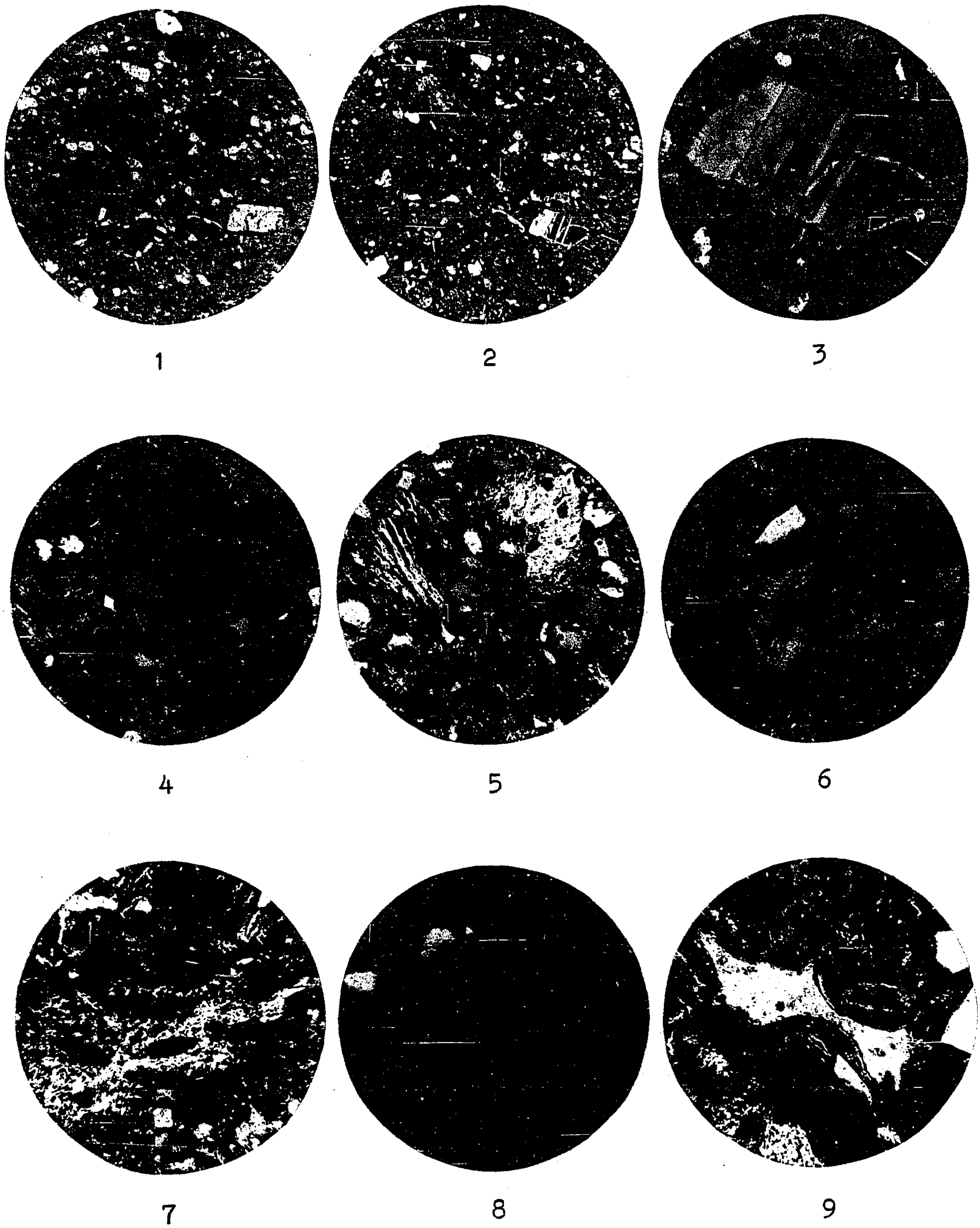
7. Thin section 7-20-5F, crossed nicols, field diameter is 1.04 mm (X50). Cellular structure of what was probably a pumice fragment. The fragment has been altered to clay.

8. Thin section 7-20-5F, crossed nicols, field diameter is 0.416 mm (X125). Calcite replacing andesine.

9. Thin section 7-20-5F, crossed nicols, field diameter is 0.416 mm (X125). A devitrified shard of volcanic glass.

* This thin section is from the Chickasaw Creek siliceous shale or from that approximate position, depending upon how the Chickasaw Creek is defined. The rock from which the thin section was made was collected in NE SW 25, T. 1 N., R. 32 W.

PLATE XXXIV



PART II

HYPOTHESES

CHAPTER III

STRUCTURAL GEOLOGY

Northward Elimination of South Flanks of Synclines

A striking feature of Ouachita structure in Oklahoma is the change in style of folding and faulting from the central parts of the mountains northward towards the Choctaw fault (fig. 2). Central folds have more gently-dipping limbs than 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. To the writer's knowledge, 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, Lynn Mountain, Rich Mountain, and Blackfork Mountain synclines illustrate progressive elimination of south flanks.

Lynn Mountain syncline lies to the south of Rich Mountain syncline. Both synclines are composed of the same stratigraphic interval. Although the writer did not map Lynn Mountain syncline, the topography indicates that most of its southern flank is present as far west as section 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, Lynn Mountain syncline is twice as broad as Rich Mountain syncline in north-south dimension, and dips within Lynn Mountain syncline are more gentle.

In Lynn Mountain syncline upper beds of the Jackfork 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 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 Rich Mountain syncline Blackfork Mountain 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 southern flank. Uppermost Jackfork, Johns Valley, and Atoka strata are not represented in the structure.

Evidences of the horizontal component of the deforming force are abundant. Beginning with the Boktukola syncline south of Lynn Mountain syncline and moving northeastward therefrom to include the Lynn Mountain, Rich Mountain, and Blackfork Mountain synclines, two facts stand out: all are asymmetrical with southward-dipping axial surfaces, and all have their south flank cut by a major fault whose upthrown side is to the south. With the exception of Blackfork Mountain syncline, all of 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 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 Blackfork Mountain 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. For a discussion of these questions, the reader is referred to "The Origin of Northward Thrusting."

Under the hypothesis that the dominant deforming force components were horizontal, the pile-up of synclines to the south of 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 Briery fault. The barrier could have been the thick section of Atoka present in Blackfork syncline and along the southern margin of the Arkoma basin. If one assumes that the Atoka thins south of Blackfork syncline, then this would cause a north-south physical change in rock masses and provide a possible barrier to northward moving thrust sheets. Presence of this barrier to the north of Briery fault could account for the dying-out of the Choctaw fault in the Waldron quadrangle and a southward shift in the belt of major thrusts in Arkansas. Such a hypothesis would involve westward as well as southward thinning of the Atoka in the Oklahoma portion of the Ouachitas, with the westward and southward thinning suggested by the northward swing of the Windingstair fault from Big Cedar to Talihina.

Thickness variations of the Atoka formation in the Ouachita Mountains cannot be conclusively demonstrated because it is the youngest unit present and because its stratigraphy is not yet adequately known. Hendricks (1947) states that the formation attains an exposed thickness of 9,000 feet in the westernmost Ouachitas. To the writer's knowledge, this is the greatest published thickness of Atoka beds in the Oklahoma portion of the Mountains. It occurs in the outermost belt of thrusts. An isopach map of Atoka strata in the Arkoma Basin and in the Ouachita Mountains would show the greatest thicknesses in a belt approximately following the Choctaw fault. In the large synclines of the central Ouachitas, exposed thicknesses of the Atoka formation apparently do not exceed 7,000 feet. Compared with the 19,000-foot thickness of Atoka measured in the Waldron quadrangle by Reinemund and Danilchik (1957), thicknesses in Oklahoma represent a marked westward and southward thinning. This thinning could be depositional or erosional in origin; however, according to the above hypothesis it is dominantly depositional. The deepest part of the Atoka trough lies in the frontal belt of the Ouachita Mountains in Arkansas. Thinning takes place to the west, south, and north of this location.

Implicit in the assumption of a barrier due to the thick section of Atoka strata is a steepening of the northward slope of the basement to the south of the Blackfork syncline and an absence of a northward slope of the basement to the north of Blackfork syncline. Under these conditions the thickened Atoka should have acted as a barrier rather than as a mass gliding northward under the influence of gravity (see section of this paper entitled "The Origin of Northward Thrusting"). Maximum horizontal stresses due to gravitational gliding would have occurred at the south margin of the barrier.

Dip of the Windingstair Fault

The angles at which Ouachita faults are inclined is a controversial subject. 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. The Windingstair fault borders Rich Mountain syncline on the south, and structures within the syncline should reveal something of the fault's nature in the study area.

The axial trace of Rich Mountain syncline traverses terrain with 500 to 800 feet of relief for a distance of over six miles at the west end of the syncline. 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 segments are assumed to be planar, but are actually concave northward, the dip of the axial surface will be overstated. If, instead, they are convex northward, the dip of the axial surface will be understated. 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 42 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 same force field caused the major synclines of the Ouachitas and the faults bordering them, it should be possible to predict something of the nature of the bordering faults by studying the synclines. The 42-degree dip of the axial surface in western Rich Mountain syncline suggests that the Windingstair fault dips 42 degrees or less southward in this area. If the fault had a higher dip, it would be difficult to account for the low dip of the axial surface.

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

In eastern Rich Mountain syncline, 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 Round Mountain anticline. 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 Rich Mountain syncline over the north flank.

Faults with sinuous traces may have a low dip angle. A sinuous trace results if the gently-dipping fault surface is planar and it intersects an uneven ground surface. Under these conditions the trace should move updip where traversing hills and downdip where crossing valleys.
If the Windingstair fault dips gently southward in the study area, and if it has a planar surface, then its trace should be likewise affected by hills and valleys. However, as mapped by the writer, the trace, although sinuous, does not show this behavior; in fact, the fault would have to

PLATE XXXV

Faults near Ward Lake

1. Looking eastward at Stanley beds of Interval 2, East Ward Lake measured section. Terrace gravels cap the outcrop. In the right hand part of the picture is the trace of a northward-dipping fault. The trace of a fault believed to be the Windingstair is a few tens of feet out of the picture to the left.

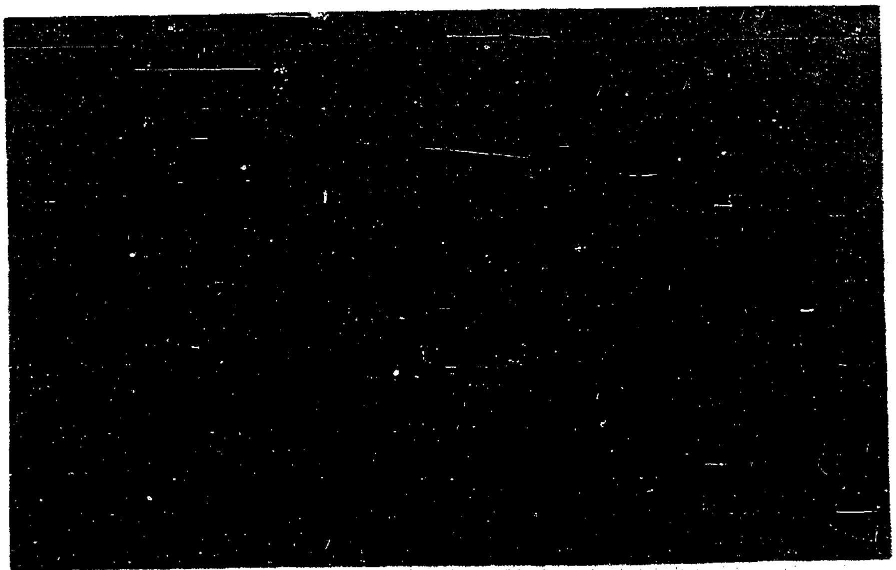
2. Looking westward at a small reverse fault cutting beds of Interval 3, Ward Lake Spillway measured section. A synclinal axis is out of the picture to the north (right) and the displacement by the fault corresponds to that predicted by theory.

3. Looking northward at 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. The writer interprets these faults as 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 of 2, plate XXXV, 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.

PLATE XXXV

1

2



3

be horizontal to explain the behavior of the trace over most of its length. Because the trace does not follow topographic contours into the internal parts of the syncline, the fault cannot be horizontal. Thus, the sinuous trace of Windingstair fault cannot be attributed to intersection of a horizontal or gently-dipping planar fault surface with the ground surface.

If the fault surface dips gently or steeply but is curved rather than planar, a sinuous trace would result. But, then, under these conditions a sinuous trace is no indication of dip angle. If cross-faults do not play a part in the apparent sinuosity of the Windingstair fault, then the fault must have a curved surface and its trace is of no use in determining dip angle. It is probable that a cross-fault cuts the Windingstair fault near Big Cedar, but the presence of cross faults causing all bends of the Windingstair fault is improbable; therefore, it must have a curved surface.

Far to the west of Rich Mountain in Ranges 14 and 15 East, the dip of Windingstair fault is inferred to be approximately 20 degrees by Hendricks (1959, p. 45). The writer's previous work on beds adjoining the Windingstair fault north and east of Talihina, Oklahoma (Seely, 1955), indicates 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 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.

PLATE XXXVI

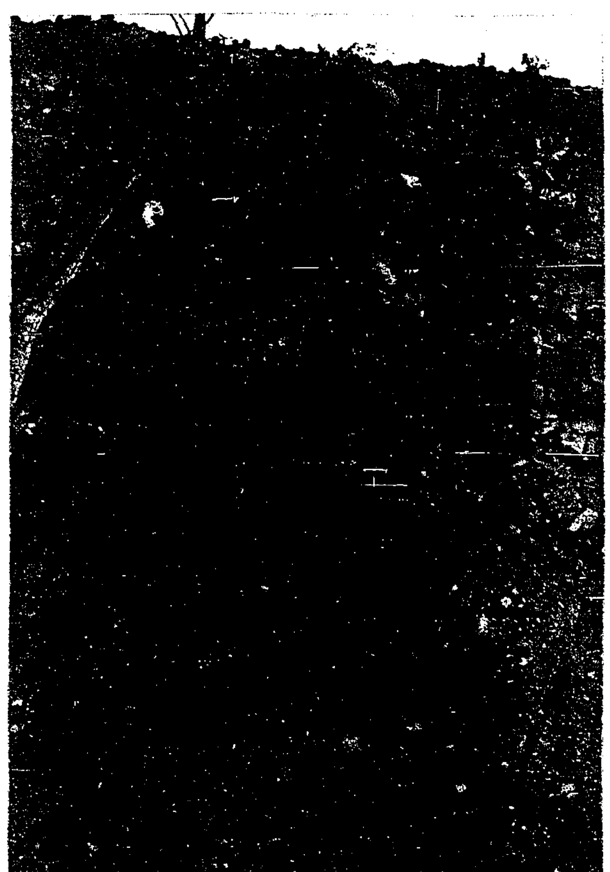
Faulted beds of Spring Mountain Syncline

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

2. Small reverse faults cutting lower Atoka sandstones exposed in the east roadcut of 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 plate XXXVIII.

PLATE XXXVI

1



2

Dip of the Briery Fault

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

The Cause of Strike-slip Faulting

Faults in the study area, interpreted to have a strike-slip component of their net slip, show no prevalent direction of movement. The writer attributes their origin to tears produced when northward moving thrust sheets encountered resistant blocks of the foreland. This is best illustrated by the recess produced by Spring Mountain syncline in Briery fault.

The north-south breadth of Spring Mountain syncline is two and one-half miles at its maximum, but westward it narrows to about one and one-half 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 thrusting 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 thrusting is suggested by the large drag fold produced in rocks of Windingstair Mountain south of Briery fault. This fold is west of the study area in sections 21, 22, 27, and 28, T. 3 N., R. 25 E. and is shown by aerial photographs and the topography. It consists of a syncline-anticline couplet in the northern flank of the major syncline directly west of Rich Mountain syncline. This couplet indicates right lateral movement of Briery fault on the southwestern flank of Spring Mountain syncline in addition to northward thrusting.

Blackfork Mountain syncline lies directly east of Spring Mountain syncline. At the western end of Blackfork Mountain there is a 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 while strike changes suggests that the strike change is caused by drag against Briery fault rather than by the nearness of these beds to the eastward plunging axis of Blackfork Mountain syncline. If the strike change was caused by drag, it indicates that Briery fault probably had a left lateral component of movement to the west of Blackfork Mountain.

Like the western end, the eastern end of Spring Mountain syncline also appears to be crushed. The east-west fault whose trace is near the south edge of section 15, T. 3 N., R. 26 E. shows the effects of shearing. Its upthrown south block appears to have offset the warped axial trace of Spring Mountain syncline to the east and, in so doing, it has formed an S-fold in beds of the north block. This fault joins Briery fault and probably reflects the same left lateral movement.

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

Under this hypothesis, Spring Mountain and Stapp synclines form a resistant element in the foreland against which beds above the Briery fault were thrust. This buttress is also proposed as the cause for the recess in this fault. Directly south, the western end of Rich Mountain syncline has been crushed against it, but just to the west, Simmons Mountain shows much less crushing of the syncline of which its rocks are a part. North of the resistant element the Stapp fault forms a small salient, which also attests to the relatively high compressive strength of the block.

Upper Jackfork beds of Spring Mountain syncline have a much higher proportion of sand than correlative beds of Blackfork Mountain syncline. The sand may represent a facies nearer the northern shore of the Jackfork sea. This relationship is favored by the tectonic interpretation given above.

A condition which appears somewhat similar to that just described, but that is probably significantly different, is evidenced by Shut-in anticline and by rocks near the town of Eagle Gap. A recess in Briery fault occurs south of Shut-in Mountain, the causes of which appear different from those causing the much larger recess to the west.

Eagle Gap is located on an axial culmination of Blackfork Mountain syncline between Blackfork and Fourche Mountains. Strike swings of the major ridges in this area, therefore, are not the result of drag. The curvature of Shut-in anticline may be a reflection of the weakness created in the overriding sheet by the axial culmination. This weakness caused the recess in Briery fault, and its trace parallels the strike of beds on either side. Although the resistance to northward movement presented by Shut-in anticline is not obviously reflected by Briery fault, it is possibly shown by the strike-slip fault at the western end of Fourche Mountain. Beds to the east of this fault have moved northward relative to those to the west to form a left-lateral strike-slip fault.

The curvature of Shut-in Mountain and its relation to Blackfork and Fourche Mountains may be similar to the curvature of Rich Mountain and its relation to Kiamichi Mountain. Eastward the trend of Rich Mountain syncline changes from east-west to more nearly northwest-southeast. This change in trend causes the syncline to appear draped around the eastern end of Lynn Mountain syncline. The draping could be caused by the absence of competent Jackfork beds in the upper plate of the Windingstair fault south of the eastern half of Rich Mountain syncline on a level parallel with the present topographic surface. The shear strength of shales, exposed east of Jackfork outcrops at the east end of Lynn Mountain syncline, would be less than where sandstones of the Jackfork are present within the syncline. The shales east of the Jackfork outcrop would, therefore, allow the eastern end of Rich Mountain syncline to lag behind the western end during northward movement of the thrust sheet.

Another occurrence of strike-slip faulting is apparently due to the crushing of the south flank of Blackfork 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 Blackfork syncline from older folded Atoka sediments that have been moved relatively in an east-northeast 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 syncline that are overturned. Their axial surfaces probably dip southwestward at angles not exceeding 50 to 55 degrees. The southwestern 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. 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 which follows Price Creek (the trace also follows Dry Creek to the north in Bates quadrangle, Arkansas-Oklahoma) probably joins with another which follows Clear Fork. Clear Fork is to the east of Price Creek. Both faults appear to have resulted from strike-slip movement. Still farther to the east is Heath Creek. A fault trace is also present in the valley formed by Heath Creek. The trace is apparently the result of a thrust fault whose west block has moved relatively upward. Together with the strike-slip faults, the thrust fault is evidence of northeastward and eastward thrusting of beds making up the south limb of Blackfork syncline. This thrusting may have been caused by wedge action of the overthrust sheet of Briery fault. The salient of Briery fault to the west of Shut-in Mountain may have acted as a plow in turning blocks of earth aside from its leading edge.

Johnson Creek Fault

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

As previously mentioned, there is abundant evidence for eastward and northeastward thrusting of Atoka beds in the southwestern exposed part of the Blackfork syncline in the writer's study area. While this evidence does not confirm Reinemund and Danilchik's original interpretation, it fits in nicely with the direction of thrusting they postulated.

The fault trending east-west through the middle of section 27, T. 1 N., R. 31 W. and through the lower middle of section 26, T. 1 N., R. 31 W. is on-strike with the Johnson Creek fault. The writer, however, does not interpret it as being the Johnson Creek (bedding plane) fault in his area of investigation (see 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 sections 26, 27, and 28, T. 1 N., R. 30 W.

The Origin of Northward Thrusting

In previous paragraphs the writer has interpreted structure of the study area to be the result of northward moving thrust sheets. A hypothesis for the origin of these sheets may be constructed using the work of Hubbert and Rubey (1959) as a basis. This work gives a possible explanation for the origin of overthrusts by invoking the presence of interstitial fluid pressure that approaches overburden pressure at some level in a rock sequence. This condition is most likely to occur in thick sections of shale buried within a geosyncline (Hubbert and Rubey, 1959, p. 201). Many major thrusts of the United States are associated with thick geosynclinal shale sections.

A detailed application of the hypothesis to the Ouachita Mountains is beyond the scope of this paper, but a few points will be discussed. The three thickest shale sections in the western Ouachita Mountains are represented by the Mazarn-Womble shales, the Stanley shale, the Atoka formation, and the Caney-Springer sequence. Thinner shale sequences are the Collier shale, the Missouri Mountain - Polk Creek shales, and the Johns Valley shale. Windingstair, Jackfork Mountain, Octavia, and Boktukola faults are bordered on the south throughout most of their length by the Stanley shale. The overthrust plates of the Windingstair and Jackfork Mountain include beds older than the Stanley shale. If the Core Area south of the Boktukola fault is omitted, the oldest known beds in the thrust plates north of the Core belong to the Mazarn-Womble shale section. This is also true of the Ti Valley fault, which lies to the north of the Windingstair fault. It thus appears that several of the thrust surfaces were formed, at least locally, in the Mazarn-Womble shale section (or possibly, stratigraphically lower).

According to the Hubbert-Rubey hypothesis, permeability must be reduced in beds above and below the stratum ultimately responsible for the development of a thrust plane in order to prevent dissipation of pressure built up in the entrapped water. This reduction in permeability is accomplished in shale by compaction. Because at any given depth, pressure due to overburden is greater than pressure due to a liquid column equal in height to the thickness of the overburden, liquids tend to migrate upwards as the sediment containing them is buried and compacted. If burial is slow enough, liquid migration may keep pace with compaction so that abnormal pressures cannot build up. The limestones and cherts of the Bigfork chert and Arkansas novaculite may have inhibited vertical migration of liquids during burial of the Mazarn-Womble shale section. Even if they did not, thin sandstones in the shale section provided avenues for local migration from the highly compactable shales into the less compactable sandstones. Shales adjacent to the sandstones would have become impermeable with compaction preventing migration of water from more distant shale. Abnormal interstitial pressure would then have built up in this "sealed-off" distant shale as burial progressed. In addition to the Mazarn-Womble sequence, the Stanley shale, Atoka formation, and Caney-Springer sequence also have sandstones in thick shale sections and, consequently, the development of thrust planes within them would not be surprising.

It is apparent from the hypothetical structure cross-sections published by Hendricks (1947) that he considered the Mazarn-Womble shales as the interval in which several of the major thrusts originated. Parts of the upper plates of the Choctaw, Ti Valley, and Windingstair faults are shown with this interval at their base. Near the northwestern end of the Choctaw sole fault, Hendricks shows many of the imbricate thrusts which join the sole faults, as originating in the Caney-Springer shale section. Other shale sections he shows lying above thrusts are those of the Stanley shale and Johns Valley shale.

The same general method of analysis of the thrusting in Idaho and Wyoming as used by Hubbert and Rubey may be applied to the Ouachitas. In so doing an interesting point to investigate is the applicability to the Ouachitas of their conclusion that the thrusts may be explained as the result of gravitational gliding if gaps at the rearward edge of the sheets can be found (Hubbert and Rubey, 1959, p. 196-197). Circumstances favorable for gravitational gliding can also be proposed for the Ouachita Mountains; however, more investigation is needed to check the hypothesis. The slope necessary for gliding may have been established (1) by northward migration of the axis of maximum deposition to the frontal belt by Atoka time and/or (2) by tilting of the basement associated with uplift in the Ouachita core areas.

The axes of maximum deposition were south of the frontal belt in Stanley-Jackfork time, but presently it is not known how far south. Beds of the Morrowan and Atokan epochs have not been separated satisfactorily within the Ouachitas; it is not possible, therefore, to locate the axis of maximum deposition during each of those epochs. Further complicating the problem of locating axes of maximum deposition is the absence of beds younger than Atokan. However, if one assumes that the present eroded thicknesses of the Atoka formation reflect the original thickness distribution, then it is evident that the axis of maximum deposition lay in or near the frontal Ouachitas during Atoka time, which probably includes part of the Morrowan and Atokan epochs. Thus, there may have been northward migration of depositional troughs after Stanley-Jackfork time.

Hubbert and Rubey (1959, p. 196) appeal to a similar migration of the axes of maximum deposition with time to cause a movement toward the foreland of deep, abnormally high, fluid pressure zones. New thrusts evolved with the migrating troughs, and thus the youngest thrusts were associated with the youngest trough. Uplifts formed by the older faults were a source area for sediments being deposited in the younger troughs.

Recently, Pitt (1955) restudied the Ouachita Core Area in McCurtain County, Oklahoma (western end of Broken Bow - Benton uplift, fig. 2) and found that its structure is primarily that of a large anticline or anticlinorium. He did not find the window hypothesized by Miser (1929, p. 22) on the basis of Honess' work. The parallelism of fold axial traces on the northeastern, northern, and northwestern margins to the periphery of the uplift is striking. It is noteworthy that the traces of faults and fold axes of the entire Oklahoma salient of the Ouachita mountains parallel the trends shown by folds and faults of the older Paleozoic rocks surrounding the core. Such parallelism would be expected if gravitational gliding from the Core Area of southeastern Oklahoma was the primary cause of all Ouachita folding and thrusting in Oklahoma. Because fold axial traces and fault traces do not parallel the southeast margin of the central uplift it is possible (1) that the uplift swings eastward under the Cretaceous beds to the south, or (2) that another uplift farther south has imposed a general northward slope on the basement. Information plotted on Flawn's map (1959) suggests that the second explanation should be chosen and that this distant southern uplift may be associated with the belt of low grade metamorphic rocks whose presence to the south may be inferred from Flawn's map.

Two structural features of the northern and northwestern border of the Core Area have puzzled geologists for a long time. These are northward and northwestward-dipping normal faults and similarly inclined axial surfaces of overturned folds. Miser (1929, p. 22) explained the faults as due to motion of the overriding plate of the Boktukola thrust and used them as one of the evidences for existence of the plate. To the writer's knowledge, an explanation for the southward overturning of the folds is not in print.

As pointed out by Miser (1929, p. 22) normal faults are uncommon in the Ouachita Mountains so that their presence about the Core Area is of particular note. The writer interprets the faults as a possible result of gravitational gliding off of the Core Area uplift. The southward-overturned folds with associated small faults whose upthrown side is to the north might have been caused by movement of deep rock masses northward relative to overlying beds. This relative movement could have occurred during formation of the essentially synclinal structure between the Core Area and the trace of Boktukola fault. Drag is often associated with folds, and in synclines, younger beds move away from the fold axis over underlying strata during folding because of forces required by the geometry of the fold.

On the other hand, it could be argued that the size and number of Core Area drag folds are too great to be explained by simple movement of younger beds over older in such a synclinal structure. Dr. Jean Goguel has suggested (personal communication) that the relatively complex structure of older Paleozoic beds in the Core Areas of the Ouachita Mountains in both Oklahoma and Arkansas may be a result of an early deformation and may exist where the older beds are deeply buried as well as in the Core Areas. Larger features, such as the Boktukola and Lynn Mountain synclines, would have resulted from a later deformation. This hypothesis is handicapped by the lack of a known widespread angular unconformity in Ouachita rocks which could delineate the two, separately deformed, sequences.

Other evidence that gravitational gliding was an important mechanism in deformation of Ouachita rocks is discussed under the heading, "Northward Elimination of South Flanks of Synclines." The writer pointed out the change from open folds of the central Ouachitas to tight folds and imbricate faults along their outer margin. This change implies that deforming stresses increased from the Core Area outward. Increase in this direction would be predicted by the gravity-gliding hypothesis whereas a northward push on the rearward end of thrust sheets would cause stress increase in the opposite direction.

It should be noted that the recent mapping of Johnson (1954) and Shelburne (1960) has shown that the trace of Boktukola fault differs from that shown on the Geologic Map of Oklahoma (Miser, 1954). The fault trace is only 30 miles long and maximum stratigraphic displacement of the fault is about 14,000 feet (Shelburne, 1960). The fault can hardly underlie a thrust sheet that extends southward an unknown distance beneath Cretaceous strata. This view is also expressed by Shelburne (1960, p. 49, 50) who stated that the Boktukola fault "is not a reasonable vehicle for the low-angle overthrusting proposed by some workers."

Stratigraphic displacement is greatest mid-way along the trace of the fault opposite a deep axial depression in the Boktukola syncline. It is probable that net slip is also greater here and that it decreases eastward and westward from this midpoint. Such behavior would be predicted by the hypothesis of gravitational gliding off the Core Area uplift. Twenty-eight thousand feet of strata are present in the Boktukola syncline that are absent from the crest of the Core Area uplift. The distance from the uplift to the syncline is about 20 miles. If the slope of the basement is assumed to be uniform over this distance, the present slope angle would be 16 degrees. According to the Hubbert-Rubey hypothesis (1959, p. 197), a fluid pressure-overburden ratio of about 0.50 is all that would be required for gravitational gliding to occur on such a steep slope. Indeed, it is difficult to envision circumstances which could prevent gliding of Ouachita rocks on such a slope. The slope of the basement from the Core Area northward to the eastern terminus of the Boktukola fault is probably close to the average slope from the Core Area to Rich Mountain. This slope is approximately six degrees if the thickness of older Paleozoics lying upon the basement is constant. Stresses built up because of the slope were apparently not great enough to cause rupture to extend the plane of Boktukola fault farther eastward.

Hubbert and Rubey (1959, p. 196) consider a slope of two and one-half to three degrees as an adequate cause for gravity gliding and formation of the Wyoming overthrusts. If the writer's estimate of a six-degree average slope from the Core Area northward to Rich Mountain syncline is correct and if this slope existed at the time Ouachita thrusts were forming, gliding could have occurred in the Ouachitas with a lower fluid pressure-overburden ratio than possible in Wyoming. The northward slope should continue north of Rich Mountain syncline as northward thickening of the Atoka section probably more than makes up for thinning of the Stanley-Jackfork sequence. A basement low might then be located beneath the maximum thickness of Atoka beds north and east of Blackfork Mountain.

An accurate determination of average basement slope is not possible because of the unknown thickness variations of pre-Stanley strata and because the position of the Stanley-Jackfork pinch-out cannot be determined. These unknowns are probably less of a problem in the determination of slope from the Core Area northward than from the Core Area westward and northwestward to the Ti Valley fault. Some pre-Stanley units are known to thin to the west and northwest (Ham, 1959), but behavior of the oldest formations is not known. If the pre-Stanley interval is assumed to maintain a constant thickness, and if the thinning of Stanley-Jackfork beds in the outermost belt is ignored, an estimated average west and northwest slope of three and one-half degrees seems reasonable. However, the angle of slope is probably greatest near the Core Area.

A cross-section from the Boktukola syncline to the Potato Hills would probably show a southward slope of the basement. Yet no major faults with the north block upthrown are present between these two features: all major faults have their upthrown side to the south. If the assumed southward basement slope is correct and gravity is the basic force causing faulting, then the northward thrusting requires special explanation. Two hypotheses which might explain it are: (1) the Potato Hills uplift post-dates the period of active thrusting and is, thus, a younger feature than the Core Area uplift; (2) both uplifts occurred at the same time, but the surface elevation above the Core Area uplift was always greater than that above the Potato Hills.

The first hypothesis seems more likely due to the absence of structures about the Potato Hills which show evidence of having influenced gliding and due to the possibility that the present basement elevations are the same beneath both uplifts. If the Potato Hills anticlinorium post-dates active thrusting, the tighter folds of older Paleozoic beds it exposes probably existed prior to its uplift. This would suggest that the folds must continue beneath the younger Paleozoics of the Ouachitas.

One of the greatest problems arising in the gravitational gliding hypothesis as applied above is location of a vacant space left at the rearward edge of the gliding sheets. Hendricks (1959) has estimated a minimum displacement of 53 miles along faults north of the Octavia fault in the westernmost Ouachita Mountains. If it is assumed that this displacement is primarily horizontal and it has been accomplished by rocks moving <u>en masse</u> off the Core Area uplift dome, then one's first impression is that the vacant space at the top of the dome should be at least 53 miles measured at right angles to fold axial traces and fault traces. However, there are several factors whose importance cannot be evaluated at the present time. These are: (1) Hendricks' figure does not separate vertical from horizontal displacement; (2) as Hendricks stated many assumptions enter in, in determining displacement—combined displacement of the Pine Mountain and Ti Valley faults makes up 30 miles of the 53-mile total; an error in this figure could greatly affect the total; (3) slip planes probably developed in the Stanley as well as in the Mazarn-Womble and the vacancy of the former unit is considerably broader at the top of the Core Area uplift than that of the latter; (4) thrust planes may die out and not connect to form a few sole thrusts. If the thrust planes die out downdip there may be a stretching parallel to bedding of strata above the thrust plane with respect to those below as is suggested, for instance, by King's cross-section of the Appalachian Valley and Ridge Province (King, 1959, p. 47).

A hypothesis which has the attractive capability of explaining both northward thrusting where it is difficult to explain by gravity and of solving the problem of finding a rearward vacant space, combines basement thrusting with gravity gliding. Goguel (personal communication) feels that basement thrusting is a necessary accompaniment to deep convection currents. Such thrusting will also result in basement uplift which may cause gravitational gliding.

It is probable that the Benton (Core) uplift of Arkansas (eastern part of Broken Bow - Benton uplift, fig. 2) is similar in nature to the Core Area uplift of Oklahoma. If structures produced by gravity gliding from the Benton uplift are different from those caused by gliding from the Oklahoma Core uplift, their difference may be due to location of Benton uplift at a different position in the Ouachita geosyncline. Western folds of the Benton uplift are on structural strike with the Potato

Hills of Oklahoma. This suggests that both features may be due to uplifts along the same basement trend.

From the foregoing discussion, it may be seen that the writer considers Ouachita structures to be superficial in nature: Their primary origin was simple vertical movements of the basement or vertical movements accompanying basement thrusting; their secondary origin was gravitational gliding. Under this hypothesis, Ouachita rocks have been thrust over strata of the Arbuckle facies only along the frontal belt in Oklahoma. Elsewhere Precambrian basement directly underlies Ouachita rocks.

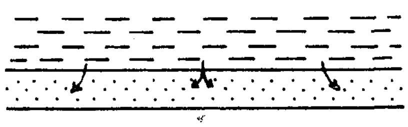
Although Goguel explains vertical movements of the basement by convection currents, Jacobs, Russell, and Wilson (1959, p. 290-361) explained these movements in another way. The Ouachita Mountains, along with other salients of the Appalachian system, are classified by these co-authors as the folded part of a secondary arc (p. 317). Secondary arcs composed of ridges and troughs occur on the continental side of primary island arcs, and tectonism within the arcs is not deep-seated. Ridge-uplifts or welts occur within the secondary arcs and may expose basement rocks. These uplifts are convex toward the continent and "it is evident in some cases that the uplift has caused much of the folding in the adjacent basin by slumping under the influence of gravity" (p. 300). The ridges are best formed between the junction of two primary island arcs and the continent. Sedimentary basins lying between the ridges and the continent are cut off from the primary arcs and are characterized by the miogeosynclinal facies with nonvolcanic sediments.

Ridge-uplifts of secondary arcs are a theoretical necessary accompaniment to the deep fracture system occurring off the edges of the continents when this system produces two intersecting primary arcs (Jacobs, et al., p. 358). According to theory, they should be convex toward the continent as is the Ouachita salient. The distance between them and the primary arcs is a function of the angle at which the primary arcs intersect.

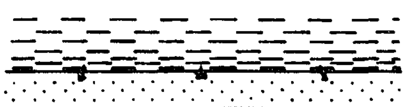
The Blue Ridge is considered to be a ridge-uplift of the Appalachian Mountains which marks the outer edge of the pre-Paleozoic continent (p. 318). East of the Blue Ridge are much younger igneous rocks and metasediments of the Paleozoic geosyncline. West of the Blue Ridge is the Valley and Ridge Province whose structure the writer pictures as sharing some features in common with the Ouachitas. The Core Areas of the Ouachita Mountains are also ridge-uplifts, but are located at various positions within the secondary arc. Metasediments that were originally laid down in the geosyncline adjacent to the primary arcs are separated from the exposed Ouachita Core Areas. These metasediments, presumably metamorphosed and uplifted prior to Stanley-Jackfork time, are farther south beneath Mesozoic and Cenozoic sediments of the Gulf Coast. They were the source of metamorphic detritus found in the Stanley-Jackfork sandstones.

The Hubbert-Rubey Hypothesis; Sandstone Dikes; Cementation

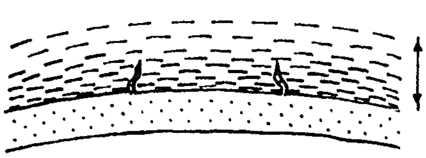
The Hubbert-Rubey hypothesis has many interesting sidelights. Relatively dry shale sections adjacent to sandstone beds should have a greater strength than either the shales farther away (which have a high interstitial fluid pressure) or the sandstone beds, themselves (whose fluid pressure has been increased by the receipt of water from the now dry 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 read-



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.

HYPOTHETICAL ORIGIN OF SANDSTONE DIKES

Figure 6

ily 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. 6). 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 interstitial fluid pressure, but this shale does not fill the rupture in the dry shale because fissures close before reaching the highly mobile shale zone (fig. 6).

The sandstone dike (1, pl. XVII) described under the title, "Plastic Flow Structures," shows evidence of having been injected upward into a small fissure which it forcefully enlarged by pressure on the fissure walls. Shale beds abutting the dike are bent, but not in a uniform direction. Their warping is probably not the result of drag; instead it was caused by pressure directed outward from the 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, the rigidity of the shale, and the evidence of pressure suggests that intrusion of the dike took place under considerable overburden.

Another effect which the hypothesis points up, but which was recognized previously, is the introduction of foreign materials into porous strata in shale. As suggested by C. J. Mankin (personal communication), perhaps most of the silica causing secondary enlargement of quartz grains within clean sandstones of the Jackfork group was brought in by water injected from adjoining shale. It is a significant fact that thin sandstone beds in sections dominantly composed of shale commonly show a high degree of induration as would be expected if they had been the recipients of large volumes of silica-bearing water from neighboring shale. The known siliceous shale zones suggest the possibility that silica was also present, but in more dilute concentration, in interstitial water within the great thicknesses of non-indurated shale. This silica-bearing water could conceivably carry into some porous sandstones small quantities of clay. If its derivation is not recognized, this clay might cause a misinterpretation of sedimentary environments.

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 Blackfork Mountain show extensive slumping.

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 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 on that slope. Generally, strata exposed on northern slopes dip southward while beds exposed on southern slopes dip northward and the steeper the slope, the more gentle the dip.

In roadcuts the effect of 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 initial structural position. The attitudes least affected by downslope movement are most 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 Highway 103 at the crest of the hill in SE$\frac{1}{4}$ NE$\frac{1}{4}$ sec. 24, T. 3 N., R. 25 E. (pl. XXXVIII). 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 upwards to the zone of most active creep. The apparent displacement of beds increases up the dip of the fault plane from a very small displacement at the foot of the embankment. Towards 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 up-dip increase in displacement is due to tectonism. Dr. Jean Goguel viewed the faults briefly during a visit to the Ouachita Mountains in April, 1961, and considered the latter explanation more likely (personal communication). An appeal to differing horizontal movement (that is, rotation) up the dip of the fault plane was more reasonable to him.

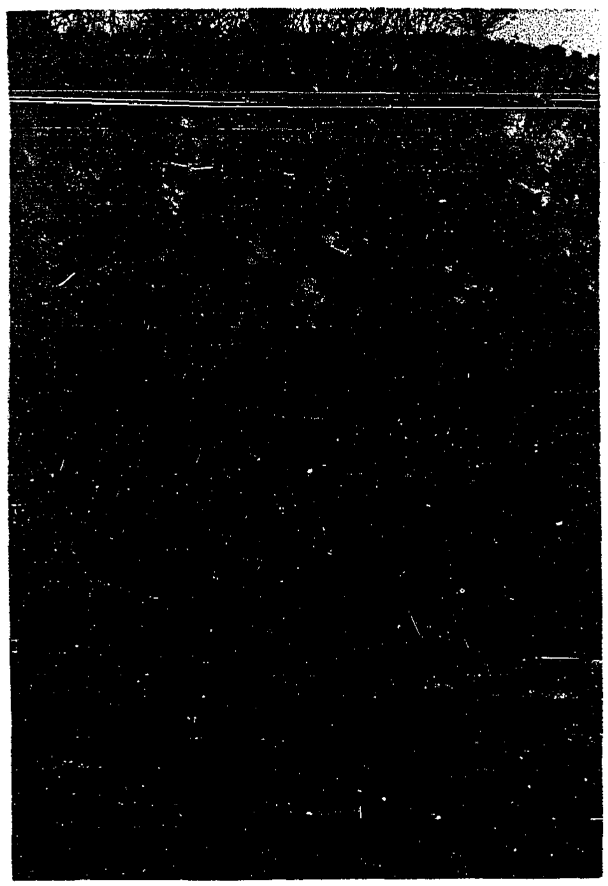
There are at least these two possible origins for the two thrusts. One of them is that the faults formed during folding of the Spring Moun-

PLATE XXXVII

Faulted beds in Spring Mountain Syncline

This fault is exposed near the faults shown in 2, plate XXXVI and cuts lower Atoka sandstones exposed in the east roadcut of Highway 103 at the summit of Spring Mountain. The southeastward view shows the change in dip of the fault plane with depth. Near the top of the roadcut the plane must have a southward dip component, but at the base of the cut a northward component is present.

PLATE XXXVII



tain syncline. Faults formed during the folding might be expected to have a southward dip as do 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 Highway 103 roadcuts near the crest of Spring Mountain (pls. XXXVI and XXXVII).

A second possible origin of the faults involves the force of gravity (fig. 7). If one extrapolates 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 may have a bend at a position corresponding to where the fault surface, if extended, would intersect it. The bend divides beds above that 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 in plate XXXVIII.

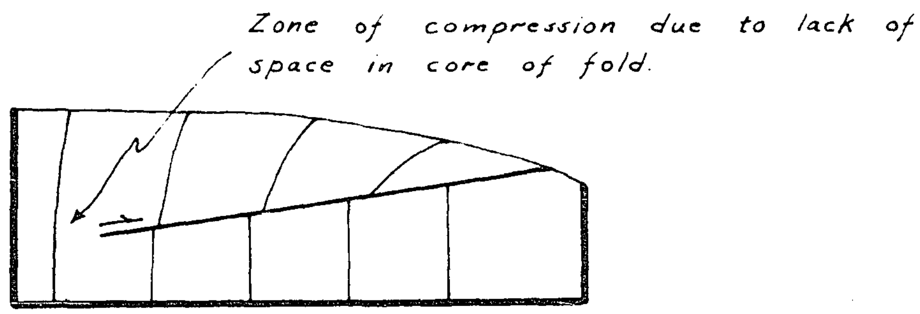
The bend marks the trough of a recumbent synclinal fold that opens to the right in plate XXXVIII and in the upper drawing of figure 7. During formation of the fold there would have been compression of beds near its center of curvature (upper drawing of fig. 7) if folding was of the concentric type (de Sitter, 1956, p. 198). If the fold had been buried at the time of formation, this compression probably would have resulted in shale flowage and similar folding instead of concentric folding. On the other hand, if the fold were formed near its present near-surface position and the axial surface were more nearly vertical rather than approximately horizontal, the competent sandstones would have moved parallel to bedding planes relative to the shales as in flexure folding (Billings,

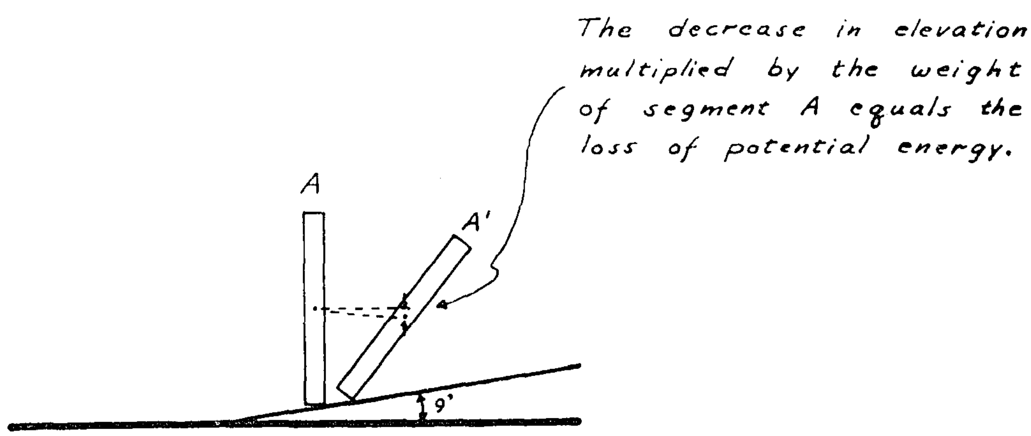
PLATE XXXVIII

Gravity (?) Induced Thrust Faults

Westward view of Highway 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 mutual boundary 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."

PLATE XXXVIII





FACTORS TO CONSIDER IN THE GRAVITY-THRUSTING HYPOTHESIS

Figure 7

1954, p. 89). However, the probable near-surface conditions under which the faults were formed suggest that the thrust surfaces developed in response to compression near the center of a fold that was caused by downslope movement. The principle is indicated by the sketches shown in figure 7.

Perhaps the most difficult idea to grasp in such an explanation is how gravity can cause beds to move upward and thus gain potential energy. However, this can be explained by assuming that only some mass gains potential energy by upward movement, and that this is overcompensated for by the greater loss of potential energy by other mass. Thus, in figure 7, bed segment A has lost potential energy when its position has been changed from A to A', even though its base has gained potential energy by upward movement. Some of the lost potential energy may have been used to produce thrusting. The net result suggests that thrusting may occur even where there is the potential energy loss of the system required when motion is due to the force of gravity.

Under a genetic classification, these faults might be referred to as gravity faults, but under a descriptive classification they should be referred to as thrust faults. The same terminology could, perhaps, be applied to the major faults of the Ouachita Mountains. According to definition, a gravity fault is one in which the hanging wall has moved down relative to the footwall (Billings, 1954, p. 143). Thus, we have a genetic term defined by a descriptive elucidation. This is an unfortunate kind of classification and illustrates the hazards in prematurely basing descriptive terminology upon the presumed origins of features.

CHAPTER IV

ENVIRONMENT OF DEPOSITION

In studying the stratigraphy of Rich Mountain Area the writer accumulated data bearing upon depositional environments. Because of the interest in this subject by students of the Ouachitas these data are discussed below in relation to various environmental concepts.

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) deep water deposition by turbidity currents; (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." He concluded (p. 161): "the Stanley-Jackfork sequence represents an over-all trend from deep- to shallow-water deposition."

More recently Scull, Glover, and Planalp (1959, p. 168) have 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 tenet is 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 blackshale 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."

What factors distinguish sediments laid down in a neritic environment from those laid down in a bathyal environment? How do both of these differ from deltaic sediments and other kinds? A detailed investigation of Quaternary sediments deposited in these environments is just beginning. This study should contribute greatly to the ability of geologists to reach more reliable conclusions about ancient depositional environments. Any present-day hypotheses regarding depositional environments will face severe testing as more observational data are accumulated. Ripple marks, small-scale cross-bedding, and rill marks, suggested a mud flat environment to Honess (1923, p. 198). Yet, we now know that ripple marks and small-scale cross-bedding occur on the floors of water bodies thousands of feet deep as well as in shallow water.

Whereas some conditions of neritic and bathyal environments are similar on all coasts, others clearly are not; in fact, sediments laid down in one environment may share more features in common with sediments laid down in another environment than with other sediments laid down in the same environment. This line of reasoning carried to its extreme leads one to wonder whether we should abandon our present environmental classification schemes because no two environments are identical in every way. Perhaps a better approach to the study of environments is to classify them by the physical, chemical, and biological conditions that prevailed during deposition of the sediment, and then to compare these inferred conditions with the known conditions of present areas receiving sediments.

As an example, cross-bedding of a given bottom sediment is caused by currents moving at a given velocity; this velocity is common in several depositional environments. Similarly, clay deposition may occur at any depth and in nearly any environment.

Stanley-Atoka Sediments and Turbidites

During the past decade Philip H. Kuenen of the University of Groningen has been the world's leader in development 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 enumerates follow:

(1) Graded bedding, combined with poor sorting

(2) Continuous, parallel stratification

(3) Interstratification with non-turbidites

(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 found in the study 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 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 top 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 top 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 consisting of 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 study area is good; most thin sections are well or very well sorted and a paste matrix is either absent or a minor constituent. 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 study area and found that most are fine-grained. Sorting in them varies from 1.28 to 1.72 with most sorting values below 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 third quartile divided 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 (Chapter II). The converted standard deviation values are 0.52, 1.15, and 0.86 phi respectively. Using 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) while 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 method of sorting computation, which would correspond to a standard deviation greater than 0.86 phi. Pettijohn (1957, p. 171) lists the standard deviations for different levels of a turbidite bed of the Ventura River and these exceed 0.87 phi. Ericson et al. (1961, p. 216, 217) lists the sorting of 84 graded beds most of whose medians fall in the fine- and very-fine-sand size ranges. Eleven beds have a standard deviation equal to or less than 0.50 phi (converted from quartile deviation and assuming normality) and are well sorted. Fifty-six of the remaining beds are moderately sorted and 17 are poorly sorted (sorting terminology after Folk, 1959, p. 103).

When 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 et al. Several sandstone beds of the lower 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. The 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 study area. Except where there has been plastic flowage or local diastrophic activity, even the thinnest sandstones appear continuous. Sandstones of the Moyers

formation have a continuity similar to those of the Jackfork. Units of the Tenmile Creek 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 there is a distinctive shale zone in the lower Wildhorse Mountain (see Intervals 8 through 11 of Rich Mountain measured section) which may be traced the full length of both Rich and Blackfork Mountains. Some of the individual sandstone beds do lose their identity, however, within the lengths of the two mountains.

While the sandstone zones of Rich Mountain appear relatively continuous along strike, a zone-to-zone correlation across the strike with beds of Blackfork Mountain could not be accomplished for most zones. However, the similar topographies of both mountains suggest that there is little difference in their overall stratigraphy.

In the upper part of the undifferentiated Prairie Mountain - Markham Mill - Wesley sequence of Rich Mountain syncline, sandstone ridges are locally present along strike. The local presence of these ridges suggests a lateral discontinuity of units at that stratigraphic level, a level predominantly composed of shale in the exposures of Rich Mountain measured section. The Jackfork strata in Highway 103 exposures in the northwest part of the study 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 Blackfork Mountains. The writer considers the beds along Highway 103 to have been laid down relatively farther north on a different depositional site than the Rich and Blackfork Mountain beds. The present geographic relationships are a result of diastrophism.

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 Jackfork and lower 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 is similar from the western margin of the Ouachitas eastward to, at least, the eastern edge of the study area, a distance of about 80 miles. It occupies a position above fluted slopes on the Moyers formation which join with the Stanley-floored valleys. This consistent topographic position of the Chickasaw Creek suggests a continuity of sandstone zones above and below it throughout this distance.

(3) Interstratification with non-turbidites. 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 (and other sequences as well).

(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 Atoka. Although only a limited number of observations have been recorded, these suggest a uniformity of orientation over a wide area. In general, but not every-

where, the sole markings are most 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 as clay galls by the writer) are found at many levels in the Jackfork - lower Atoka sequence. These are most 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 over 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 the orientation of sole markings, judged by the measurements thus far made by the writer. 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, autochthonous megafossils 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 top 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 in thickness. 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. It is stated by ten Haaf (1959b, p. 219) that none of the turbidite sequences observed has been found to display such phenomena as wave ripples, tidal scour and fill, autochthonous neritic fossils—fauna or flora—, or to occur 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. It may be seen that 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 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) classifies many beds of the Moyers formation as subgraywacke. (The writer understands that Cline uses 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 (Bokman, 1953, p. 163) do not agree with this generalization. Lower strata of the Atoka in the study area have a mixture of both 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 sorting changes in the source sediments of the shallow shelf. However, C. J. Mankin of the University of Oklahoma Geology School (personal communication) feels that even if only well sorted sand were available at the source, there should be turbulent mixing with the bottom material and, thus, there should generally not be well sorted, clean turbidites. 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 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 also finer grained and may display excellent planar-, cross-, or contorted-lamination. Some of this lamination consists of alternating well sorted layers (which may be only a few millimeters thick) that have a varying mean grain size. Why are the mold faunas associated with granule- or pebble-sized quartz grains? Are these associations due to characteristics of the transporting current, or are they acquired from shelf sediment?

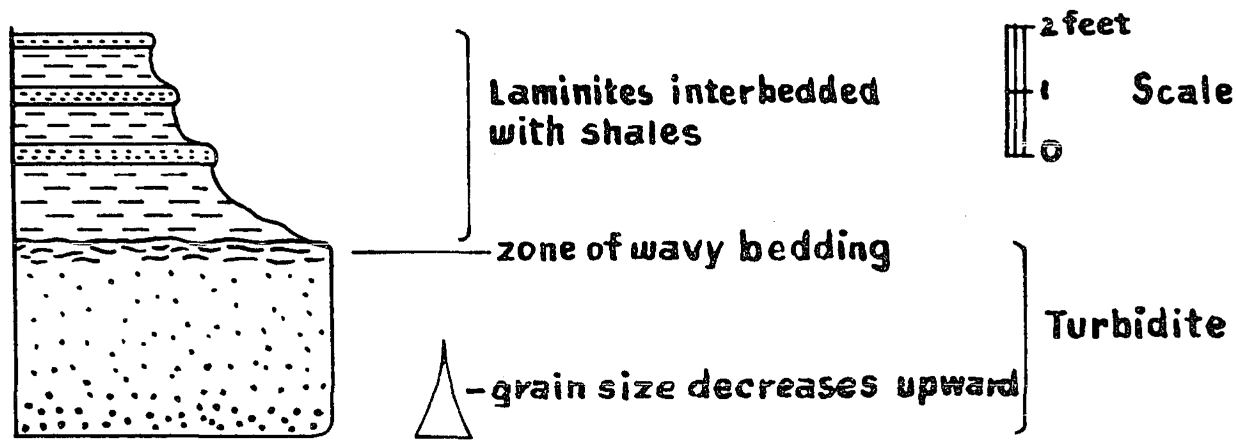
To the writer it seems more reasonable to assume that texture, lamination, and good sorting are primarily due to the dynamics of transportation and deposition rather than to source character. But, if this be true, we are faced with two questions: (1) how do these dynamics fit into the overall picture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

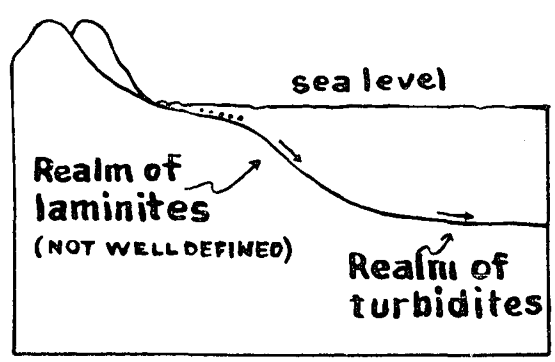
Pertinent to this problem is the recent work of A. E. Lombard of the University of Geneva. Lombard (personal communication) distinguished beds he referred to as "laminites" from turbidites and, though they occur in the same sequence of strata, he considered turbidites and laminites to be the products 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.

In the present study trails of bottom organisms were found only on or within thin (less than six inches thick), laminated, very-fine-grained sandstones. Absence of trails from thick beds may indicate that these beds were deposited rapidly, and suspended material covered their tops before crawling organisms repopulated the sea floor.

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, it may be 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. 8). Lombard considered an idealized sequence to have a turbidite bed at its base, and this is 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 felt he could distinguish 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 wished not to 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 transporting sand-sized sediment require relatively high velocities in order to exist. Ericson et al. (1961, p. 217) tacitly recognize laminites as distinct from turbidites. The laminites are what they refer to as "sand and silt layers due to winnowing by deep current scour." They accumulated "under the influence of a continuous or nearly continuous current which has prevented accumulation of particles below a certain grade size," and





The above two figures are based upon notes taken during A. E. Lombard's visit to the University of Oklahoma in April, 1960.

TURBIDITES AND LAMINITES

Figure 8

"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 squeezed causing their present discontinuous distribution. They could have formed on a tidal mud flat. However, the writer believes that 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 Jackfork. Those of the Stanley are dominantly turbidites; and those of the Jackfork, laminites. The environmental changes necessary to produce this difference, however, are not clear if, indeed, they exist 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 non-uniformity of relative abundance, Lombard hypothesized that turbidites are the result of active subsidence and tilting, while 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 et al. (1961) on Atlantic deep water sediments suggests that laminites are deposits of deep-sea currents and 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 most 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. Current markings thus far found indicate a prevailing direction of transport within the trough, but tell little of the source area position with respect to the geosyncline. The latter might be determined by a regional study of the ratio of metamorphic rock fragments to quartz.

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 cuased the continued flow of the current. It is suggested by ten Haaf (1959b, p. 219) that the average slope is about one percent. This requires that the water depth at one end of the trough be deeper than at the other and gives a means of approximating the minimum water depth at the deepest 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. When 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 returns us to the realm of turbidity currents.

It may be that 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) cites at least nine different sequences giving some evidence of prevailingly longitudinal currents; and preliminary data show 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. Is marginal uplift necessary to accomodate rock "flowing" from beneath the subsiding trough? Is the major source of materials deposited in the trough, the flanking borderlands? 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 when one studies 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. The older sediments of the Stanley were therefore 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 the sand for exposed Jackfork sandstones in Oklahoma alone.

It should be pointed out that, in Oklahoma, southward thickening of the Stanley and Jackfork has been demonstrated in the narrow frontal belt of the Ouachitas but that this 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 Rich Mountain syncline to Boktukola syncline, a north-south distance of about fifteen miles which does not take into account shortening caused by thrusting. A southern borderland should, therefore, no longer be hypothesized because of thickening in that direction. In fact, there is little direct evidence of any kind for a southern source for non-volcanics 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. The apparent common hingeline for both the Stanley and the Jackfork, however, could be considered as an argument against a similar migration during Stanley-Jackfork time.

Several researchers have stated that there was rapid subsidence 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 was filled with little further subsidence by clastics of the Stanley-Atoka sequence. Evidences for this interpretation might be: (1) the bedded cherts of the Arkansas novaculite interpreted in the traditional sense as of deep water origin; (2) the nearly complete absence of a benthonic fauna; (3) interpretation of the Atoka as a deltaic deposit antecedent to the coal swamps preserved in the overlying beds of the Arkoma basin; (4) the upward change from argillaceous, more poorly sorted sands of the Stanley to cleaner better sorted sands of the Jackfork; (5) correlation of the fossiliferous lower Atoka sandstone ridge of Rich Mountain syncline with that of Spring Mountain and Stapp synclines, thus causing the Atoka to be an onlapping unit northward.

Stanley-Atoka Sediments and Shallow Water Deposits

Could the Stanley-Atoka sequence have been laid down in predominantly shallow water? If one answers this in the affirmative, problems arise just as they do with the turbidite-laminite hypothesis. Of first importance is the absence of autochthonous megafossils. An abundant fauna is 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 lack 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 an abundance of ripple marks, drag marks, and animal trails. Coquina, white sand, gray sand with black organic matter, and black muds form 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 Stanley, Jackfork, and lower 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. Whereas ripple marks are present, they are not in the abundance that might be expected from their ubiquity on present tidal flats. However, McKee (1957, p. 1742) could not find evidence of ripple lamination in pits dug into 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 is the opposite relationship from that observed in the Jackfork: it possesses mostly short-crested ripples and only minor long parallel-crested ripples.

Tracks and trails found on mud flats are of particular interest. Similar features are preserved on the upper surface of several thin sandstones (laminites) of the Jackfork and also on their bottom surface as casts of depressions in the underlying shale. The sharpness with which these tracks are set apart from the surface of the sandstone and the preservation of detail is marked in some occurrences. How can one account for such preservation? There must have been some cohesion between the sand grains or clay flakes in order for the trail to have retained its identity for even a short time after it was formed. This cohesion is favored by alternate wetting and drying of the bottom. Trails formed in fine-textured sediments when they are wet may be hardened with the drying of the sediment and survive subsequent wettings prior to burial. Preservation is favored under these circumstances. Similar submarine sediments never exposed to drying may also gain some cohesive strength if deposition is not continuous, but the degree of their cohesion and consequent degree of detail preservation is not clear.

Lamination in the top of the bed directly underlying the trails is not disturbed except where more trails are present at lower levels. The preservation of laminae in most sandstones of the Jackfork indicates that burrowing organisms were not active. 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? Long periods between turbid flows may have allowed burial of the sandstones before migrating fauna 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, and for this reason, trails are present at several different levels of some laminite beds.

The presence of molds of invertebrate fragments at the base of several sandstones, particularly in the upper Jackfork, Johns Valley, and lower Atoka might be explained by extreme tides which brought in coarse off-shore 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 Stanley and 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) express this opinion. The oxidizing conditions may result from a decrease in the supply rate of organic material while oxygen supply remains constant or a steady supply rate and variation in oxygen brought in contact with the sediments.

Red clays are known to be accumulating over wide areas of the present deep sea floor where oxygen (and carbon dioxide) supply by cold deep currents is high in relation to the amount of organic matter settling to the sea floor. The depositional rate of red clays is thought to be quite low over most of the sea floor where they are present. Red shales are found associated with coarser clastics in the Alps, Barbados, and Timor but their origin in these localities is controversial (Gilluly, Waters, and Woodford, 1959, p. 346).

Another environment in which red beds accumulate is one in which alternate wetting and drying of the sediments occurs, such as on alluvial plains and mud flats. Plains and flats of this type are thought to have bordered the Permian seas of west Texas and New Mexico, but the red beds deposited upon them contain channel sandstones with large-scale cross-bedding, remains of amphibians, and associated evaporites.

Serious problems arise with consideration of the tidal flat hypothesis. 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, we must have a flat probably much larger than any known today, and it must persist without marine or continental ingressions for a long interval of 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 ripples to form due to wave action. 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.

We might try to evade the problem of a paucity of fossils by assuming that they existed, but the environment was not favorable for their preservation. In so doing we erect another problem: why was the detrital fauna preserved? Likewise we can try to 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 may have been more likely if a greater portion of the continental margin was flooded in Mississippian time than is flooded now, but why 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. There is, however, still much to be learned about deep water depositional rates. Although beds of the Stanley-Atoka sequence are commonly thought to be the results of rapid deposition, actually there is little evidence for determining depositional rate in an absolute rather than relative sense.

<u>Turbidites or Shallow Water Sediments: A Summary of the Problem</u>

The turbidity current hypothesis for the depositional environment of Jackfork beds exposed in the study area is handicapped by the lack of a proven mechanism for deposition of laminated, well-sorted, slightly graded or non-graded sandstone beds which make up a significant, but unknown, proportion of the total sandstone complement. Shallow water hypotheses are handicapped 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

In recent years there has been increasing literature on the interpretation of thin section data gathered from sedimentary rocks. Some interpretative criteria applied to thin sections are more reliable than others. At present, one criterion whose significance is in dispute is the meaning of 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, there are other criteria which are less controversial and can be used for interpretation of source and depositional environment.

Interpretation of Sorting and Mean Grain Size

Bokman found in his study of the Stanley and Jackfork (1953, p. 163) 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 (Chapter II). The average of the thin section means is 2.67 phi and the average standard deviation is 0.43 phi. This is based upon

20 thin sections consisting of one from the top of the Stanley, 16 from the Jackfork, and three 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 not clear. It 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 writer's 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. Of the thin sections studied, there is a suggestion that the finer grain sizes tend to be better sorted; and the coarser grain sizes, more poorly sorted. However, the few coarse-grained slides observed appear to be bi-modal, and their sorting values have a different significance than those of uni-modal slides.

Folk (1959, p. 5) stated that it is probable that in every environment sorting is strongly dependent upon grain size. In general, the best 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 whose 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 others. A degree of sorting is therefore accomplished by magmatic, metamorphic, and weathering processes. This establishes size limits of source materials and imposes limitations upon the sorting effected by sedimentation 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 the granules and small pebbles found locally.

One might theorize that grains of the granule-containing beds of the sequence should be more poorly sorted than those of the finer beds because the coarser beds should represent an intermixing of two populations just as do argillaceous sandstones. In a general way this appears to be true in the small sample observed by the writer.

After studying seven beds from the Stanley and Jackfork, Bokman (1953, p. 164) concluded that bed thickness and mean grain size are not systematically related. In addition he concluded that although there appears to be slight grading of the beds, it is of such small magnitude as to be 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.

One may explain the overall limited range of grain size in the sandstones of the Stanley, Jackfork and lower Atoka 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 selective sorting to have been effective. 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 other than the source of associated sandstones and shales.

The presence of fresh feldspar and various rock fragments suggests that part of a single source area, or one of several source areas, was undergoing rapid erosion. These rapidly eroded areas are thought to have down-cutting streams and possible topographic relief. The streams probably supplied a broad size range of material, the larger sizes of which, however, did not reach the Stanley-lower Atoka depositional basin. Apparently the small mean grain size of the sandstones in the basin is due to a distance of transport adequate to produce selective sorting.

The factors that determine the degree of sorting according to Folk are those operating at the final site of deposition. Sorting is essentially a function of the mechanical energy exerted upon a sediment after it reaches this location. 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 the correct 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 general length of depositional time and the same total expenditure of mechanical energy. The sorting differences would be due to the changes in mechanical energy per unit volume of sediment deposited. This hypothesis would require that the size range of material supplied, type of deposition, and current characteristics be constant while depositional rate varied.

Under another hypothesis the size range of material supplied would be kept constant while the type of deposition, current characteristics, and depositional rate varied. 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 are deposited by turbidity currents while the thin, well-sorted beds (laminites) are 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 reconstituted 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 there was little rounding taking place in the depositional basin. Confirming this belief is the size and shape of the micaceous metamorphic rock fragments found 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 generally has been distorted by diagenetic processes, but less deformed grains show practically no modification by abrasion.

Under the hypothesis that little rounding occurred in the Stanley-Atoka basin, rounded grains in the same sediment must have had a previous abrasion history. Large composite stretched metamorphic quartz grains, and some of the zircon and tourmaline show excellent rounding. The previous abrasion history may have been proximal, but was probably distant in time. If the rounding occurred in the last sedimentation cycle, it would indicate the incorporation of products of two different energy levels into the same sediment. Beach dune sand grains blown into the adjoining lagoonal environment illustrate this type of phenomenon. However, it is doubtful that well rounded quartz, zircon, or tourmaline in the sand size range can result from a single sedimentation cycle; therefore, it is probable that it has been derived from a sedimentary rock source.

Euhedral zircon occurs with rounded zircon in some laminae. This extreme difference in abrasion history 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. Further evidence for a volcanic source of the zircon is the absence of potassic or sodic feldspar and biotite which should be present if the euhedral zircon came from a granitic source.

The 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 plagioclase composition of Stanley and Jackfork sandstones does not appear in the literature so that a comparison with that of the tuffs is not possible. Some writers state that acid of sodic plagioclase occurs in the sandstones, but no further differentiation is 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 definitely assigned to a volcanic source.

Miogeosynclinal Graywacke

Under this heading, 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 located between the craton and a welt of metasediments off shore. 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 percent of soft metamorphic rock fragments declines relative to the percent of hard fragments; (2) the amount of clay matrix declines; and (3) the percent of quartz or carbonate cement increases. Lithology 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 metasediment 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 metasediment 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 eu- 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 lacking 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 (the interested reader should refer to Folk for further details), 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 metasediment 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 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 non-marine).

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 metasediment welt or outside 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 deposition 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 the dominant source of Stanley, Jackfork, and, perhaps, Atoka sediments. What is commonly referred to as the "Ouachita Orogeny" is but the final paroxysm of a dying tectonic belt. Mesozoic and Cenozoic sedimentation has taken place seaward of this belt.

The general picture given in the last 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 give the title of Allegheny Orogeny to 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, sediment maturity would increase as 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 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, except the very widespread ones (whose existence is yet to be proved), are therefore 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. Now, it appears safe to state that the Jackfork group, Moyers formation and, probably the Tenmile Creek formation were not deposited on a river floodplain, on the topset beds of a delta, on a tidal flat, or in shallow off-shore water above wave base. They were probably deposited below wave base.

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 siliceous shale; Jackfork group and one of its formations, the Game Refuge formation; Johns Valley shale—which may be locally separated 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; (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 ripple marks and/or small-scale cross-bedding. These features are particularly common in thin beds and near the top 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, Johns Valley and Atoka formations. In many beds, however, it is not obvious except on surfaces etched by weathering or when seen in thin section. Whereas graded bedding of some type is present in some beds it is not megascopically obvious in most. Other features represented by various sandstones are sub-ellipsoidal clay galls and their molds; load casts; contorted bedding, uneven surfaces due to squeezing while in a semi-plastic state; trails of benthonic organisms; top-surface channels and ridges; molds of invertebrate fragments.

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. They extend the fault belt to the north of Windingstair fault in Oklahoma farther eastward than had heretofore been realized.

Although direct field evidence of the type of fault movement is scant, 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 extent; however, Windingstair fault may not be a thrust fault at the extreme southeastern edge of Rich Mountain syncline.

Dips of the major faults vary to an unknown extent along their traces. Near the west edge of the area Briery Creek fault has a south dip of about 55 degrees, and Windingstair fault dips southward at an angle probably less than 42 degrees. The low dip angle of Windingstair fault may be due to apparent crushing of the west end of Rich Mountain syncline against Spring Mountain syncline. Eastward the dip of both faults probably increases. The southward dip of Horsepen fault is steepest at its outcrop and may decrease to 15 degrees downdip.

The thick section of Atoka strata present to the north of Blackfork

Mountain may have presented a physical barrier to northward movement of thrust masses and could be responsible: (1) for the southward swing of the Windingstair fault from Talihina, Oklahoma, to Big Cedar, Oklahoma; (2) for the dying out of the Choctaw fault in the Waldron quadrangle.

There is no evidence in the study area that faults probably having a strike-slip component to their movement are predominantly right-lateral. Yet they should be right-lateral if the part of the Ouachitas in Oklahoma was formed by northwestward-moving thrust sheets. Instead, it appears that, in the study area, thrust masses moved northward prependicularly to the traces of major faults. Both right- and left-lateral strike-slip movement occurred along tears in the overriding masses.

On the basis of present knowledge it seems probable that Ouachita structures originated with simple vertical movements of the basement or with vertical movements accompanying basement thrusting. This resulted in gravitational gliding and it was this process which produced the major folds and thrust faults.

A possible explanation for the origin of sandstone dikes involves: (1) movement of water from shale into adjoining sand during compaction with resultant increase in shale rigidity; (2) tectonism causing rupture of "dry" (rigid) shale beside sand; (3) injection of fluid-bearing sand into the rupture. Sand is injected into the rupture rather than shale because decreasing shear strength prohibits the rupture from forming in localities where the shale could flow if a fissure were present.

The optical mineralogy of a small sample of sandstones from the upper Stanley, Jackfork and lower Atoka was studied with a polarizing microscope. All sandstones are classified as orthoquartzites under

Folk's classification scheme. 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. We do not yet know its location. Volcanoes were also present. Little rounding took place in 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. 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 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 writer's study area suggest a prevailingly westward flow parallel to tectonic strike.

The Tenmile Creek formation, Moyers formation, and Jackfork group probably were not deposited on a river floodplain, on top-set beds of a delta, on a tidal flat or in shallow off-shore water. They were probably deposited below wave base.

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APPENDIX A

HISTORY OF NOMENCLATURE OF STRATIGRAPHIC UNITS

As explained at the beginning of Chapter I, the writer subdivided the stratigraphic column in the same manner as L. M. Cline (1960). The reader is referred to the publication by Cline for a more thorough review of stratigraphic nomenclature.

Stanley Group

This unit 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, p. 371). Honess (1923, p. 173-174) tried to place the contact in a position in agreement with Taff but had difficulty in locating it. According to Harlton, however, the Stanley-Jackfork contact is at the top of the Chickasaw Creek. Hendricks, Gardner, Knechtel, and Averitt (1947), like Taff and Honess, placed the Chickasaw Creek in the Jackfork group. Students working under the direction of Kaspar 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 comprised of 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

This name 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 located 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 L. M. Cline has done much to eliminate this confusion. The name Johns Valley shale is retained as defined by Ulrich. Arbuckle facies correlatives to the Johns Valley shale are the Caney shale at its base, the Springer formation in its middle and upper portions, and at least the lower 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 new 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, a 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 workers appear to agree with Cline in considering the lower 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 thus be called Atoka. They were so mapped on the 1926 edition of the geologic map of Oklahoma edited by Miser.

Recently Branson (1959, p. 121) has questioned the presence of any 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 of Atoka are inadequate. The Oklahoma Geological Survey is presently 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 the present state of 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	..	..

APPENDIX B—Continued

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
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	..	..	..	..

x Indicates component is present in indeterminate amounts.

* For composition of thin section 7-20-5F (a tuff) see Supplemental Remarks about Thin Sections, Appendix G.

** These figures based upon recognizable fractions. A high proportion of the slide may consist of unrecognizable organic debris.

APPENDIX C

THIN SECTION TEXTURES

Thin Section	Sorting* (ø) Thin Section	Sorting* (ø) Sieving Equivalent	Mean* (ø) Thin Section	Mean* (ø) Sieving Equivalent	Textural Classification** Based on Uncorrected Thin Section Data	Textural Classification** Corrected to Sieving Equivalent
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 sand-stone	Very well sorted very fine sand-stone
RM 24A . .	0.30	0.30	3.00	3.10	Very well sorted fine to very fine sandstone	Very well sorted very fine sand-stone
RM 21A . .	0.30	0.30	3.00	3.10	Very well sorted fine to very fine sandstone	Very well sorted very fine sand-stone
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 sand-stone	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 sand-stone
	0.25	N.A.	4.85	N.A.	Very well sorted coarse silt-stone	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

APPENDIX C—Continued

Thin Section	Sorting* (ϕ)		Mean* (ϕ)		Textural Classification**	
	Thin Section	Sieving Equivalent	Thin Section	Sieving Equivalent	Based on Uncorrected Thin Section Data	Corrected to Sieving Equivalent
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
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

* The sorting is σ_g of Folk (1959, p. 45) and is the phi standard deviation determined by estimating the sizes of the 16th and 84th percentiles based on grain area. The difference between these in phi units is divided by 2 to determine σ_g. The mean is the midpoint between the 16th and 84th percentiles. This method for determining the mean was used by Inman and is cited by Folk (1959, p. 44).

** The sorting classification is 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 REFERENCE 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 21A: 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 non-uniform 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 greater 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 two and one-half-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, non-oriented bottom casts.

RM 8-2: From the base of a five-inch, hard, gray sandstone. Planar laminae that are spaced one to two millimeters 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 upper 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 one and one-half-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 vermiculite. 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 semi-continuous 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 larger 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 have a pale green color and may be a species of chlorite.

APPENDIX H

NOTES ON MEASURED SECTIONS

General

The order of presentation of lithologies for each interval is in the order of relative thickness of each lithology; for example, "Sandstone, light-gray and gray shale" indicates that sandstone is the dominant lithology. If, in the example above, the word minor is added so that it would read "minor gray shale", then this lithology comprises less than 10 percent of the interval.

The colors used are those of the Rock Color Chart published by the National Research Council in 1948 with the exception 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. There are many more distinct colors represented in the intervals than are listed in their description due to the impracticality of listing each one. 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 sub-conchoidal 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 (see Plate 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. G. S. 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 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. Due to this the possibility of error in thickness determinations is 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 Queen 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 the city of Mena, Arkansas, and is located in NW¼ SW¼ sec. 6, T. 2 S., R. 30 W.

The base of the measured interval is located below the lowest beds of the continuously exposed sequence seen 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 located south of the east end of 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.

Interval	Thickness in feet	
		The top of the measured section is drawn at the top of this interval because of the lack 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, $\frac{1}{2}$- 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; very 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 $\frac{1}{2}$ 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.
83	219	Covered except for a 20-foot section exposed near the middle of the interval. This mid-portion 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 non-oriented bottom casts; very-light-gray, plastic shale; silty shale and interbedded thin siltstone. The shale has abundant disseminated carbonized plant fragments.

Interval	Thickness in feet	
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 non-oriented casts, but the thinner beds have only the non-oriented 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 non-oriented 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 non-oriented bottom casts, all in the upper portion of the interval; very hard to medium hard, very-fine- to medium-grained, fairly clean 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 sheaf-like appearance on weathering, the sheaves consisting of thin sandstone 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.

Interval	Thickness in feet	
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 thinnest beds may have laminae spaced less than 1 mm apart while the thickest beds may have a spacing of 8 inches. Wavy- and cross-bedding are most 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.
73	30	Sandstone, very-light-gray to white; very hard, very-fine-grained, planar- to wavy-laminated sandstone in the lower part of the interval; 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

Interval	Thickness in feet	
		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 $2\frac{1}{2}$ inches and black chert in the lower $\frac{1}{2}$ 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 is 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 $1\frac{1}{2}$-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), 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

Interval	Thickness in feet	
		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 1/8- to 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 rod-like 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 non-oriented 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. 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, $\frac{1}{2}$- to $1\frac{1}{2}$-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 $\frac{1}{2}$ foot thick that are planar-laminated at the base and cross-bedded at the top. Carbonized plant frag-

Interval	Thickness in feet	
		ments 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 is 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 mid-portion 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 transversely 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 mid-portion 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-shaped

Interval	Thickness in feet	
		slabs that are a fraction of an inch to 2 inches in thickness 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 sub-discoidal 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 in thickness 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 in shape, 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 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 non-oriented 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-shaped mounds. Most sandstone is in the upper third of the interval and the remainder of the interval is poorly exposed.

Interval	Thickness in feet	
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 possess sub-ellipsoidal 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. Sub-ellipsoidal 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 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 non-oriented bottom marks. These consist of casts of invertebrate tracks

Interval	Thickness in feet	
		and sinuous cylindrical ridges about 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, sub-ellipsoidal 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. Non-oriented 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 ½ foot to greater than 3 feet thick. There is an apparent 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 non-fractured interareas. Sub-ellipsoidal molds were noted.
33	86	Covered.

Interval	Thickness in feet	
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, hard and friable zones alternating, fine-grained, well sorted with sorting especially noticeable in the lighter colored beds, containing disseminated carbonaceous material especially in the more 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.
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

Interval	Thickness in feet	
		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 and 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 non-oriented 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 non-oriented 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.
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

Interval	Thickness in feet	
		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 and 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 and 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

Interval	Thickness in feet	
Top of Ward Lake Spillway measured section in the lower Wildhorse Mountain formation of the Jackfork group		
9	12	Sandstone, medium-dark-gray to dark-gray and minor medium-gray shale; very hard, very-fine- to medium-grained, moderately sorted, massive sandstone cut by quartz-lined joints and in beds 1 to 3 feet thick; shale and very-light-gray, firable, 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-sized quartz grains), massive sandstones containing an abundance of disseminated, carbonized plant fragments. The grain size of each of the latter beds appears to decrease upwards, grading ultimately into thin intervals of gray shale. Most of these shales are directly overlain by sandstones. The foregoing lithologies comprise 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 and 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 megascopically identified vermiculite first observed by Dr. C. J. Mankin on a visit to the writer's area.

Interval	Thickness in feet	
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-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 ripple-marked. Carbonized plant matter is profuse throughout the interval and a portion of a <u>Calamites</u> stem about 4 inches long and averaging 3/4 inch in diameter was found in place (see plate X). Also abundant in the sandstones are sub-ellipsoidal 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 inter-

Interval	Thickness in feet	
		calations of hard to very hard, micaceous, carbonaceous, shaly, cross-bedded siltstones and very-fine-grained sandstones. Some of the sandstones change upwards from dark-gray and hard to light-gray and friable. If the faults present in the interval are ignored, 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 $\frac{1}{4}$ 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 $\frac{1}{2}$ 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.

Interval	Thickness in feet	
		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 least one contains the molds of many crinoid columnals.

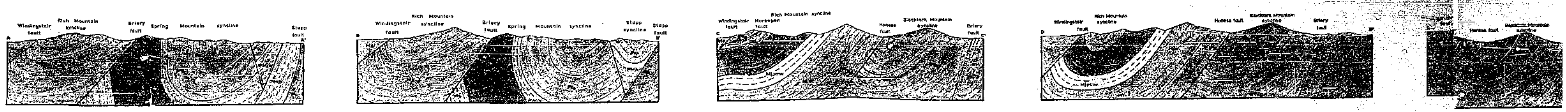
Top of Chickasaw Creek shale and Stanley group

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 $\frac{1}{2}$ inch to 2 inches thick. Quartz veinlets and drusy-quartz-lined joints are common. At the top of the interval thin siltstone 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 varies from about 2 feet to 15 feet in this exposure. The zone is characterized by a lack 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, mod-

Interval	Thickness in feet	
		erately sorted sandstone that contains clay galls, particularly above thin shale interbeds; shale that varies from "papery" to massive beds greater than $\frac{1}{4}$ inch thick. The sandstone comprises an estimated 80 percent of the interval and is in beds up to 20 feet thick that are transected 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 $1\frac{1}{2}$ inches. Siltstone occurs as thin interbeds in the shale.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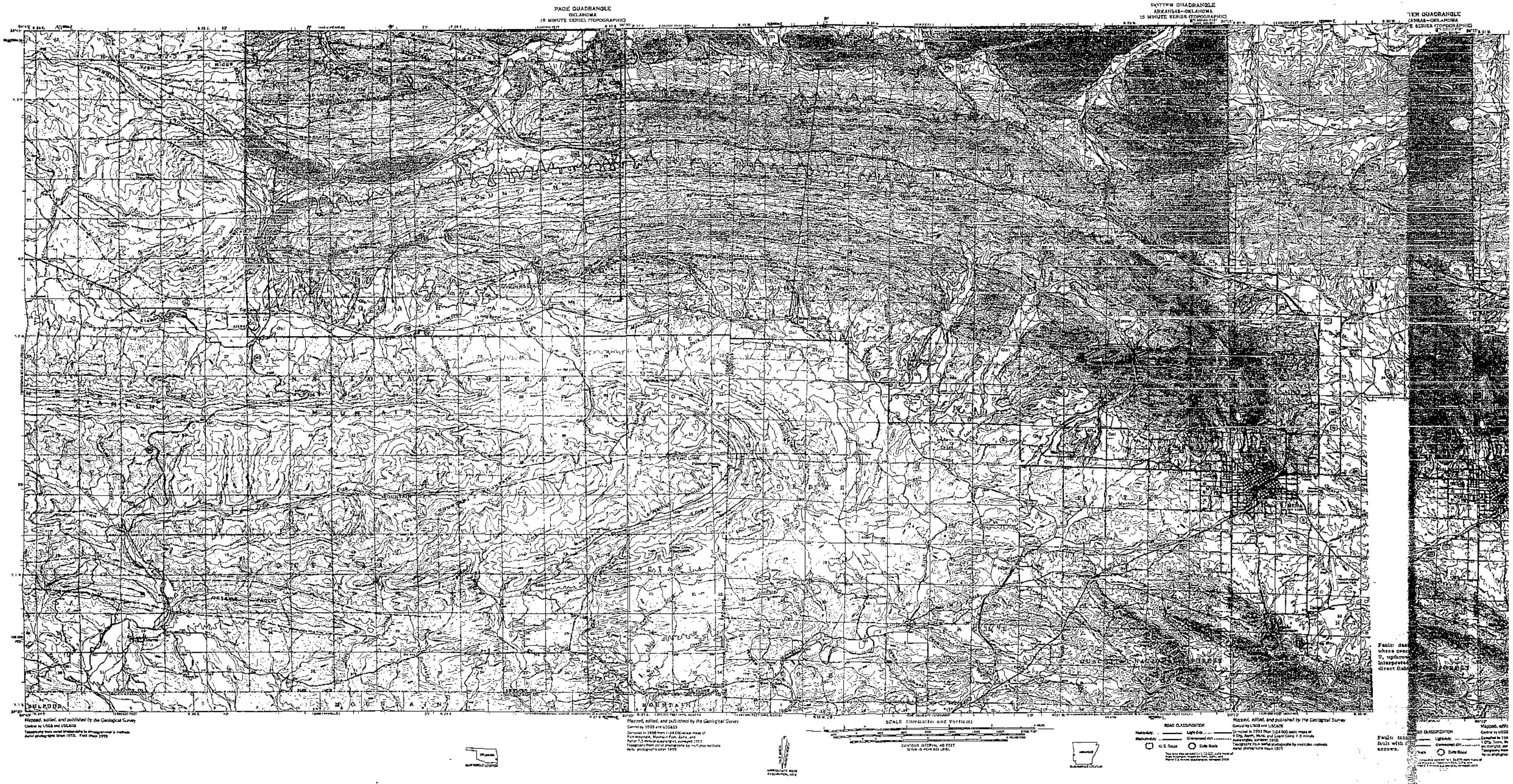
Briery fault
Spring Mountain syncline
Stapp fault
Stapp syncline
Horsepen fault
Honess fault
Blackfork Mountain syncline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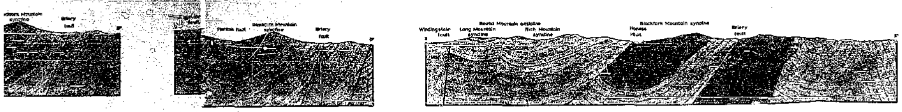
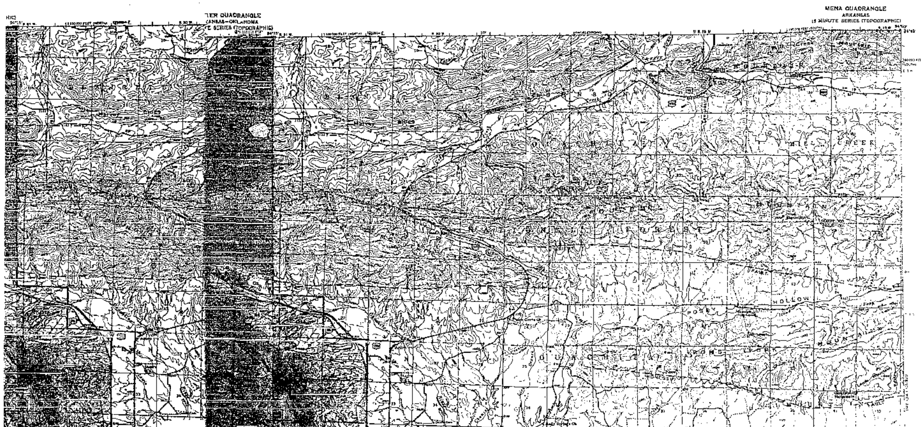
OKLAHOMA
15 MINUTE SERIES (TOPOGRAPHIC)
POTTER QUADRANGLE
ARKANSAS-OKLAHOMA
15 MINUTE SERIES (TOPOGRAPHIC)
Mapped, edited, and published by the Geological Survey
SCALE (Horizontal and Vertical)
CONTOUR INTERVAL 40 FEET
ROAD CLASSIFICATION





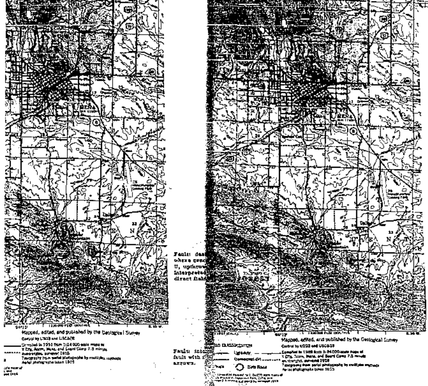


PLATE I

GEOLOGIC MAP AND CROSS-SECTIONS OF THE RICH MOUNTAIN AREA

LE FLORE COUNTY, OKLAHOMA

SCOTT AND POLK COUNTIES, ARKANSAS

by D. R. SEELY, 1962

Fault: dashed where inferred; dotted where concealed. U, upthrown side D, downthrown side Interpreted to be a dip-slip fault, but direct field evidence generally absent.

Fault: interpreted to be a strike-slip fault with relative movement shown by arrows.

Syncline: showing trace of axial surface; plunge direction of axis; dip of axial surface.

Anticline

Overturned syncline

Overturned anticline

Strike and dip of beds from aerial photograph and topographic map information.

Strike of vertic[illegible]ds. 90 placed on same side of symbol [illegible] top.

Stratigraphic contact: dashed where inferred; dotted where concealed.

Plunge angle and direction of small anticline.

Strike and dip of beds from field observation.

Dip varies within this approximate range.

Strike varies within this approximate range.

Strike and dip of overturned beds.

EXPLANATION

QUATERNARY

Qal

ALLUVIUM

Gravel, sand and clay of present stream floodplains.

Qc

COLLUVIUM

Sandstone blocks, quartz sand and clay in downslope transport.

Qt_1 Qt_2 Qt_3 Qt_4

TERRACE DEPOSITS

Alluvium of former streams that has an elevation higher than present river floodplains. Numbered according to local relative elevation.

Qt

TERRACE DEPOSITS

Where only one terrace level is locally present.

PENNSYLVANIAN

Morrowan Series

Pa Pa_{sh} Pa_s

ATOKA FORMATION

Light-gray, laminated to massive, sandstones and fossiliferous sandstones interbedded with gray shale. Locally divided into a basal sandy member (Pa_s) and an overlying shaly member (Pa_{sh}). In good exposures of Spring Mountain syncline the estimated thicknesses are: Pa 6,000 feet (including Pa_s 3,000 feet; Pa_{sh} 3,000 feet).

PMjv

JOHNS VALLEY FORMATION

Gray shale containing cobbles and boulders of limestone, chert, and limestone-chert conglomerate; and shale interbedded with locally fossiliferous sandstones.
Thickness: At least 500 feet

ATOKA FORMATION
and
JOHNS VALLEY FORMATION
(Undifferentiated)

Gray shale interbedded with thin sandstones and siltstones. Thickness: 2300 feet.

MISSISSIPPIAN

Meramecian and Chesterian Series

Mjg

GAME REFUGE FORMATION

Gray shale and a siliceous shale zone separating thick sandstone zones containing locally fossiliferous beds. Thickness: 510 feet.

Mjpmw

WESLEY, MARKHAM MILL
and
PRAIRIE MOUNTAIN FORMATIONS
(Undifferentiated)

Gray shale and minor, thin, light-gray carbonaceous sandstones. Thickness: 1,760 feet.

Mjw

WILDHORSE MOUNTAIN FORMATION

Gray shales alternating with gray, carbonaceous sandstones at the base; white, well-sorted, massive sandstones near the middle; and light-gray, friable, carbonaceous sandstones near the top.
Thickness: 3,380 feet.

Mjwp

PRAIRIE MOUNTAIN FORMATION
and
WILDHORSE MOUNTAIN FORMATION
(Undifferentiated)

Mj

JACKFORK GROUP
(Undifferentiated)

CHICKASAW CREEK FORMATION
and
MOYERS FORMATION
(Undifferentiated)

Gray and olive-gray sandstones and shale overlying a 3-foot zone of gray siliceous shale, and including sparsely speckled gray siliceous shale and local tuff at the top. Thickness: About 1100 feet.

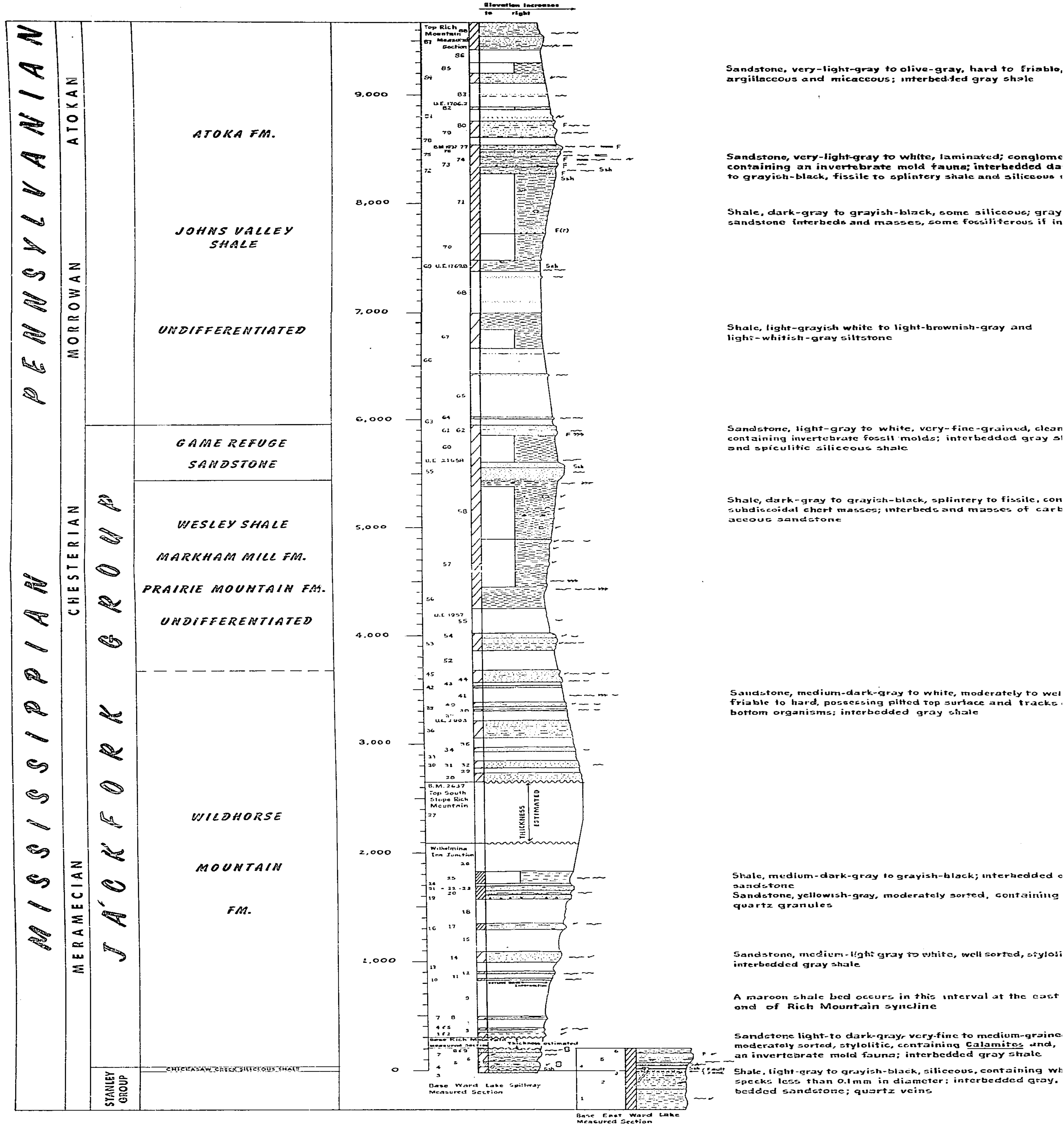
Ms

STANLEY GROUP
(Undifferentiated)

Mst

TENMILE CREEK FORMATION

Gray shale and subordinate sandstone. Only topmost few hundred feet identified in map area.



PENNSYLVANIAN
MISSISSIPPIAN
ATOKAN
MORROWAN
CHESTERIAN
MERAMECIAN
JACKFORK GROUP
STANLEY GROUP
ATOKA FM.
JOHNS VALLEY SHALE
UNDIFFERENTIATED
GAME REFUGE SANDSTONE
WESLEY SHALE
MARKHAM MILL FM.
PRAIRIE MOUNTAIN FM.
UNDIFFERENTIATED
WILDHORSE MOUNTAIN FM.
CHICKASAW CREEK SILICEOUS SHALE
9,000
8,000
7,000
6,000
5,000
4,000
3,000
2,000
1,000
0
Elevation increases to right
Top Rich Mountain Measured Section
B.M. 2437 Top South Slope Rich Mountain
Wilhelmina Inn Junction
Base Rich Mountain measured section
Thickness estimated
THICKNESS ESTIMATED
Base Ward Lake Spillway Measured Section
Base East Ward Lake Measured Section
Sandstone, very-light-gray to olive-gray, hard to friable, argillaceous and micaceous; interbedded gray shale
Sandstone, very-light-gray to white, laminated; conglome containing an invertebrate mold fauna; interbedded da to grayish-black, fissile to splintery shale and siliceous
Shale, dark-gray to grayish-black, some siliceous; gray sandstone interbeds and masses, some fossiliferous if in
Shale, light-grayish white to light-brownish-gray and light-whitish-gray siltstone
Sandstone, light-gray to white, very-fine-grained, clean containing invertebrate fossil molds; interbedded gray s and spiculitic siliceous shale
Shale, dark-gray to grayish-black, splintery to fissile, con subdiscoidal chert masses; interbeds and masses of carb aceous sandstone
Sandstone, medium-dark-gray to white, moderately to wel friable to hard, possessing pitted top surface and tracks bottom organisms; interbedded gray shale
Shale, medium-dark-gray to grayish-black; interbedded sandstone
Sandstone, yellowish-gray, moderately sorted, containing quartz granules
Sandstone, medium-light gray to white, well sorted, styloli interbedded gray shale
A maroon shale bed occurs in this interval at the east end of Rich Mountain syncline
Sandstone light-to dark-gray, very-fine to medium-graine moderately sorted, stylolitic, containing Calamites and, an invertebrate mold fauna; interbedded gray shale
Shale, light-gray to grayish-black, siliceous, containing wh specks less than 0.1mm in diameter; interbedded gray, bedded sandstone; quartz veins

Sandstone, very-light-gray to olive-gray, hard to friable, clean to argillaceous and micaceous; interbedded gray shale

Sandstone, very-light-gray to white, laminated; conglomerate, containing an invertebrate mold fauna; interbedded dark-gray to grayish-black, fissile to splintery shale and siliceous shale

Shale, dark-gray to grayish-black, some siliceous; gray sandstone interbeds and masses, some fossiliferous if in place

Shale, light-grayish white to light-brownish-gray and light-whitish-gray siltstone

Sandstone, light-gray to white, very-fine-grained, clean, containing invertebrate fossil molds; interbedded gray shale and spiculitic siliceous shale

Shale, dark-gray to grayish-black, splintery to fissile, containing subdiscoidal chert masses; interbeds and masses of carbonaceous sandstone

Sandstone, medium-dark-gray to white, moderately to well sorted, friable to hard, possessing pitted top surface and tracks of bottom organisms; interbedded gray shale

Shale, medium-dark-gray to grayish-black; interbedded olive-gray sandstone

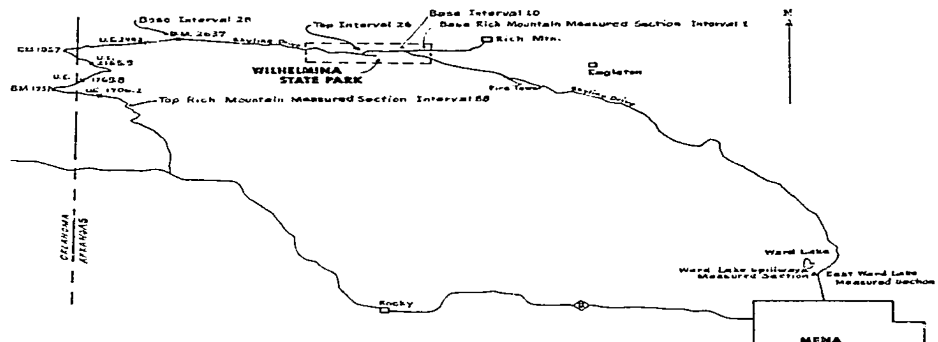
Sandstone, yellowish-gray, moderately sorted, containing quartz granules

Sandstone, medium-light gray to white, well sorted, stylolitic, interbedded gray shale

A maroon shale bed occurs in this interval at the east end of Rich Mountain syncline

Sandstone light-to dark-gray, very-fine to medium-grained, moderately sorted, stylolitic, containing Calamites and, locally, an invertebrate mold fauna; interbedded gray shale

Shale, light-gray to grayish-black, siliceous, containing white specks less than 0.1mm in diameter; interbedded gray, cross-bedded sandstone; quartz veins



LOCATION MAP FOR MEASURED SECTIONS

0 1 mile
Scale

EXPLANATION

✓	Layers of carbonized plant fragments prominent in sandstone
F	Molds of invertebrate fragments, especially crinoid or blastoid columnals, present in sandstone
⊕	Calamites stems in sandstone
	Trails of benthic organisms on top or bottom surface of sandstone
~~	Undulating upper surface of sandstones. These overlie a zone of wavy-or cross-lamination that is no more than a few inches thick
	Contorted bedding in sandstone
=	Even-bedded sandstone breaking into plates along shaly, carbonaceous or fossiliferous laminae
	Sandstone has oriented and/or non-oriented bottom surface markings
	Sub-ellipsoidal depressions on top surface of sandstone and/or sub-ellipsoidal cavities within beds
	Quartz veins
Ssh	Siliceous shale
	Most distant spacing indicates white sandstone. Color is increasingly darker gray with increasingly close spacing of hachures. Closest spacing indicates dark gray
34	Measured section interval number. See Appendix for detailed interval description
	The most poorly exposed intervals are indented
U.E. 2155.9	Unchecked elevation points which are marked with white paint on roadside rocks and whose positions are plotted on the Location Map are present at the stratigraphic levels shown

COLUMNAR SECTION OF THE UPPER STANLEY - LOWER ATOKA INTERVAL

PLATE II

D. R. SEELY, 1962