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GEOLOGY OF THE WESTERN PART OF WINDING STAIR RANGE, LATIMER AND LE FLORE COUNTIES, OKLAHOMA

by

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Geology of the Western Part of Winding Stair Range, Latimer and Le Flore Counties, Oklahoma

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ABSTRACT

The area of this report is a part of the central Ouachita province between the Ti Valley and Windingstair faults in southern Latimer and southwestern Le Flore Counties, Oklahoma. Strata exposed are the Stanley and Jackfork Groups (Upper Mississippian), the Johns Valley Formation (Upper Mississippian and Lower Pennsylvanian), and the Atoka Formation (Pennsylvanian), all of which were deposited in the Ouachita geosyncline. Interbedded limestone lenses and exotic blocks with Wapanucka-like lithologies are present in the upper part of the Johns Valley Formation, proving that it is of Early Pennsylvanian age. During the Ouachita orogeny, which began after deposition of the Atoka Formation, these strata were thrust northward along the Ti Valley fault. They are now in contact with thinner and lithologically different shelf equivalents—the Caney Shale, Springer Formation, Wapanucka Limestone, Chickachoc Chert, and Atoka Formation.

A foreland-rim flexure was present at the northern margin of the Ouachita geosyncline during deposition of the Stanley-Jackfork-Johns Valley sequence. This structural feature probably served as a barrier of deposition that separated the Ouachita and Arbuckle facies. Interpretation of the existence and role of the foreland-rim flexure is based upon the thinning of pre-Atoka stratigraphic units over the frontal Ouachitas as well as upon the relationship of the Wapanucka Limestone and Chickachoc Chert to the limestone lenses and exotic blocks of Wapanucka-type rocks south of the Ti Valley fault.

During deposition of the Stanley-Jackfork-Johns Valley sequence, a hinge zone of tension developed along the southern flank of the foreland-rim flexure. Normal faults, probably submarine, subsequently developed along the hinge zone. Exposed in the southward-facing fault scarps were pre-Stanley strata of Arbuckle-like lithologies. Partially dissolved and pitted clasts of these strata accumulated at the bases of the fault scarps. Occasional penecontemporaneous submarine slumping occurred down the steep northern slope of the

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geosyncline. Slumping might have been triggered by earthquakes associated with normal faulting. By this mechanism, clasts of Arbuckle lithologies, ranging from granule size to boulder size, were emplaced in the Chickasaw Creek, Markham Mill, Wesley, and Johns Valley Formations. They were derived from local sources and are nonexotic. The association of exotics with black siliceous shales suggests that deposition of some of the black shales was also related to faulting along the hinge zone of tension.

Rolled sandstone masses of different origins are present within the Johns Valley and other formations. Some are segments of sandstone strata that were contorted during penecontemporaneous slumping. Others are interpreted as having been formed during the Ouachita orogeny, either along slippage planes that developed during folding or along minor bedding-plane faults.

During the Ouachita orogeny, strata were folded, the folds were overturned to the north, and reverse faults developed essentially parallel with the fold axes. The Ouachita province consists of a series of blocks bounded by major reverse faults and cut by minor reverse faults and cross faults. The Ti Valley fault, one of the major reverse faults, is directly related to the Stanley-Jackfork-Johns Valley hinge zone, the normal faults that developed along the hinge zone of tension, and the foreland-rim flexure that was present during deposition of pre-Early Pennsylvanian strata. Sources of the Johns Valley exotics are now beneath the Ti Valley fault.

INTRODUCTON

The area described in this report was selected for the following reasons. (1) It is an area of complex structure within which are several well-known localities, including Hairpin Curve, Coopers Cove, Bengal, and Compton Cut. (2) It is an area of transition between strata deposited in the Ouachita geosyncline during Mississippian and Early Pennsylvanian time and equivalent strata deposited on the adjacent northern shelf. (3) It is a key area for the study of the Johns Valley Formation because it contains excellent exposures of the controversial exotic-bearing horizons.

Special attention was given to the exotics in the Johns Valley Formation and their relationship to the enclosing shale in order to find some clue to their mode of emplacement. An areal geologic map and structure sections (pl. I) were constructed.

Location of Area

The area studied, located in the Winding Stair Range, Latimer and Le Flore Counties, Oklahoma, is in the central Ouachita structural province (fig. 1). The area mapped approximates 135 square miles, is bounded on the west by State Highway 2, on the south by State Highway 63, on the east by the St. Louis-San Francisco Railway, and on the north by the Ti Valley fault. The Potato Hills area adjoins on the south.

Previous Work

J. A. Taff of the U. S. Geological Survey, one of the first geologists to do detailed work in the Ouachita province in Oklahoma, named the Woodford Chert, Caney Shale, Stanley Formation, Jackfork Formation (1902), and the Atoka Formation (Taff and Adams, 1900). Unfortunately, most of Taff's work, including geologic maps of the McAlester, Tuskahoma, Winding Stair, Alikchi, and Antlers quadrangles, has not been published except as included in the Oklahoma geologic map of 1926 (Miser, 1926).

Additional contributions were made by Purdue (1909), Miser (1917, 1929, 1934), Honess (1923, 1924), Miser and Honess (1927), and Miser and Purdue (1929), to mention just a few. Harlton (1938), working in the Tuskahoma syncline, successfully subdivided the Stanley-Jackfork sequence on the basis of black siliceous shales.

The first detailed mapping in the frontal belt, that area between the Choctaw and Ti Valley faults, was done by Hendricks and others (1947). They observed that the frontal belt consists of several major fault blocks, each of which has a characteristic sequence of strata. The stratigraphic units and the geologic structure of the frontal belt and the central belt, which adjoins on the south, are considerably different. Hendricks and others identified most of Harlton's map units as far north as the Ti Valley fault but did not use his terminology. Field work for their report was done from 1936 to 1939.

After the field work by the U. S. Geological Survey in the frontal Ouachitas, little interest was shown in Ouachita geology for about ten years. In the summer of 1952, William D. Pitt, a graduate student at the University of Wisconsin, began remapping the core area under the direction of Dr. L. M. Cline. Results of this work were published by the Oklahoma Geological Survey in 1955. In 1953 Cline restudied Harlton's type sections of the Stanley and Jackfork Groups and succeeded in correlating the Stanley-Jackfork-Johns Valley-Atoka rocks of the central Ouachitas with those of the frontal Ouachitas. Cline (1956a, 1956b, 1956c) referred the thick Stanley-Jackfork stratigraphic succession to the Mississippian (it had previously been classed as Pennsylvanian) and showed that the lower part of the Johns Valley Formation has Caney (Mississippian) equivalents. Cline found two complete, unfaulted sections of the Jackfork Group in the Kiamichi Range, which he and Moretti described in detail in 1956. Reinemund and Danilchik (1957) mapped a portion of the frontal belt in the Waldron quadrangle, Arkansas. Shelburne (1960) published a geologic map of the Boktukola syncline, and Cline (1960) published a map of the western part of the Lynn Mountain syncline. Recent developments in the geology of the Ouachita province have been summarized in publications by Cline and Shelburne (1959), Cline (1960), and Shelburne (1960).

Selected areas in the central Ouachitas have been mapped by graduate students at The University of Oklahoma (Johnson, 1954,

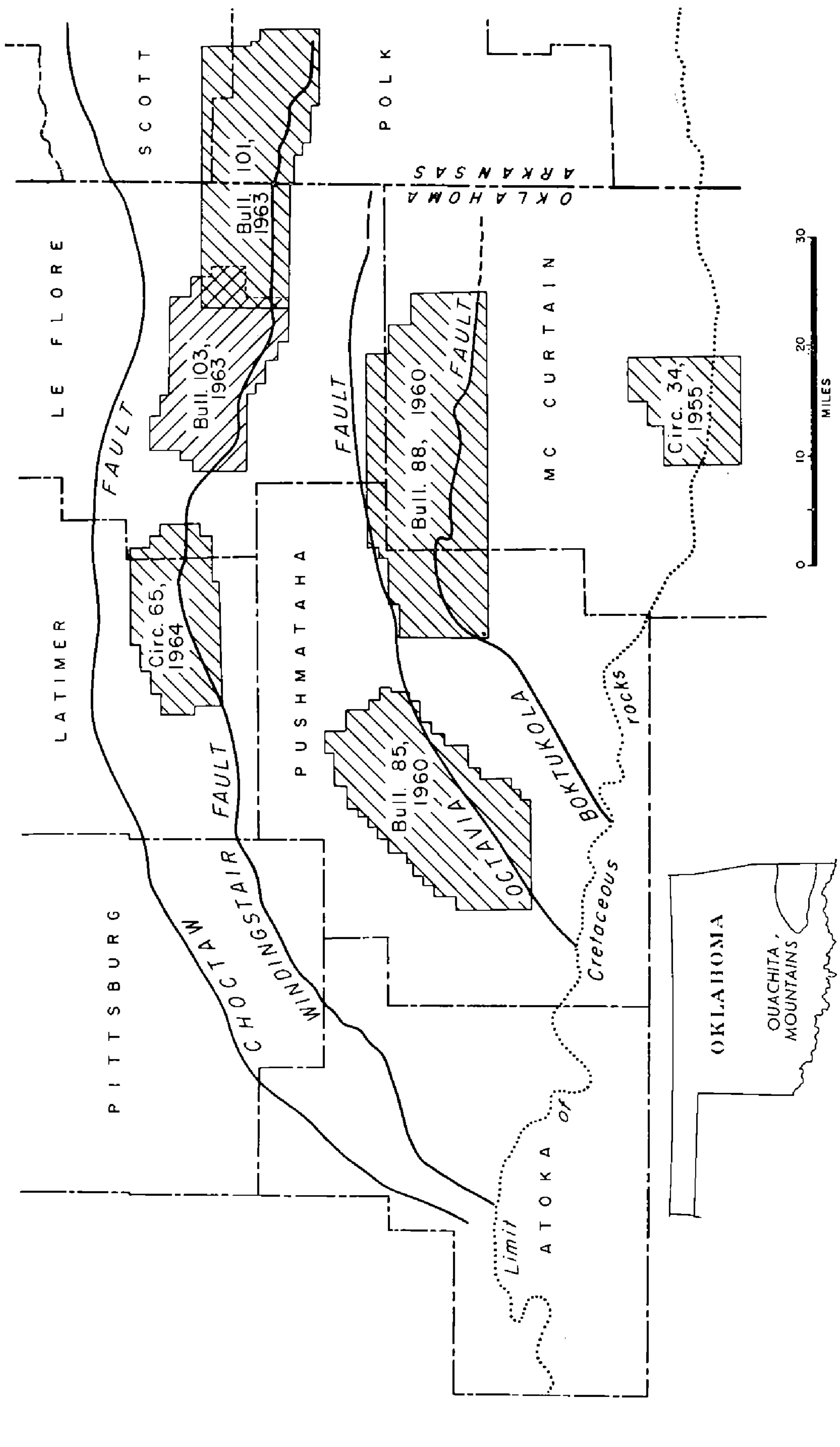

Figure 1. Index map of Ouachita province, Oklahoma, showing areas covered by this and other Oklahoma Geological Survey publications.

1959; McCullough, 1954; Willis, 1954; Miller, 1955; Roe, 1955; Seely, 1955, 1962) and at the University of Wisconsin (Laudon, 1959; Shelburne, 1959, 1960; Hart, 1962). Moretti (1958) made a petrographic study of Jackfork sandstones.

Neither the area of this investigation nor the areas adjoining on the north and east have previously been mapped in detail. However, Taff mapped the major structures in these areas. The area adjoining on the west was mapped in detail by Hendricks and others (1947). The area which adjoins on the south and includes the Potato Hills has also been mapped in detail (Miser, 1929; Miller, 1955; Roe, 1955; Seely, 1955). Several oil companies have done geological and geophysical mapping in the area of this report, as well as in other parts of the Ouachita province. However, little of this work has been published.

Field Methods

Field work for this study was done during the summer of 1960, from October 1961 to February 1962, and in April 1962. Mapping was done on aerial photographs with a scale of 1:20,000. A portion of the area of investigation is included on the topographic map of the Red Oak quadrangle (1:62,500), the scale of which precludes its use as a base map for this study.

Most of the geological information was obtained by traversing streams that flow at right angles to the regional strike. Slopes were avoided, except in critical areas, because of poor exposures, dense vegetation, and downslope movement of strata due to gravity. Geological stations were located stereoscopically. The abundant logging trails, most of which follow streams or ridges, were invaluable to this study. These trails, although almost impassable except by foot or on horseback, are visible on the aerial photographs and assist one in orienting himself.

Fault traces and stratigraphic contacts were interpolated between traverses by stereoscopic observation. The contact between the soft shales of the Johns Valley Formation and the resistant sandstones of the lower part of the Atoka Formation is quite distinct when viewed stereoscopically. Good exposures of the Johns Valley Formation are not common for two reasons. The Johns Valley is normally covered with colluvium derived from the Atoka Formation higher on the slope and is quite susceptible to downslope movement by gravity. The top of the Chickasaw Creek Formation, approximated by the highest position of the spotted siliceous shale on the slope, is easily identified on aerial photographs. For these reasons it was necessary to rely on photographic interpretations in certain localities. However, even the remote localities were visited.

The final map (pl. I) was made by reducing the aerial photographs and by enlarging the topographic map to a common scale, 2 inches per mile, with an opaque projector.

Acknowledgments

Dr. L. M. Cline of the University of Wisconsin suggested the problem and spent several days in the field during the course of the study. Dr. T. A. Hendricks and Boyd Haley of the U. S. Geological Survey spent two days in the field in April 1962. Dr. Hendricks gave helpful criticism while in the field and generously lent an unpublished report. Paul Condry of Bengal, Oklahoma, accompanied the writer in the field several days in the Bengal vicinity.

J. Gilmore Hill, D. Bradford Macurda, Harold Rollins, and Robert Scott, graduate students at the University of Wisconsin, gave instruction and advice on photographic processes. Thin sections were made by Don Fadness of the University of Wisconsin.

Financial support, which made this investigation possible, came from several sources. The Wisconsin Alumni Research Foundation provided a research assistantship from September 1959 to June 1961, inclusive. As a National Science Foundation Terminal Year Graduate Fellow during the academic year 1961-1962, the writer completed the study. The Oklahoma Geological Survey, directed by Dr. C. C. Branson, defrayed some field expenses and lent aerial photographs.

GEOGRAPHY

The area studied consists of a series of essentially parallel westward-trending ridges and valleys. The highest point in the area, 2,162 feet above sea level, on top of Buffalo Mountain, is approximately 1,400 feet above the valley of Buffalo Creek, which adjoins on the south. However, most of the area studied is north of the Windingstair fault and has an average relief of 500 to 600 feet. The Windingstair fault, which bounds Buffalo Mountain on the north, is a significant line with respect to topography, transportation, population, and areal geology.

The ridges, most of which have local names, are held up by resistant sandstone beds of the Jackfork Group or Atoka Formation that dip steeply to the south. Some of these sandstone ridges, or mountains as they are usually called, extend essentially unbroken for 10 miles or more.

The report area straddles the divide between the Arkansas River drainage basin on the north and the Red River drainage basin on the south. The rectangular drainage pattern illustrates the dominant influence of geologic structure in an area underlain by strata of unequal resistance to erosion.

The mean annual temperature in Talihina, at the southeastern corner of the area, is 63° F; the mean annual percipitation is 48 inches; the average frost-free period is 210 days. Plants thrive in this climate and form a dense cover throughout the entire area. The native pine forests were destroyed by lumbermen whose operations were based upon the philosophy: "cut out and get out." Scrub oaks and hickories have replaced the pines, which were not replanted.

The area of investigation is bordered by paved highways, most of which were constructed during the last 15 years. The broad, flat valley south of the Windingstair fault has a good system of improved roads. More than 90 percent of the population of the area is in this valley or in similar valleys.

The area north of the Windingstair fault is characterized by narrow valleys and steep, densely wooded slopes. Old logging trails have not been maintained and are virtually impassable, even in a jeep. The best methods of travel in this area are walking and horseback riding.

Bengal and Talihina are located along the St. Louis-San Francisco Railway, which runs between Ft. Smith, Arkansas, and Paris, Texas. This railroad was an extremely important outlet for logs and wood products when the logging industry was at its peak.

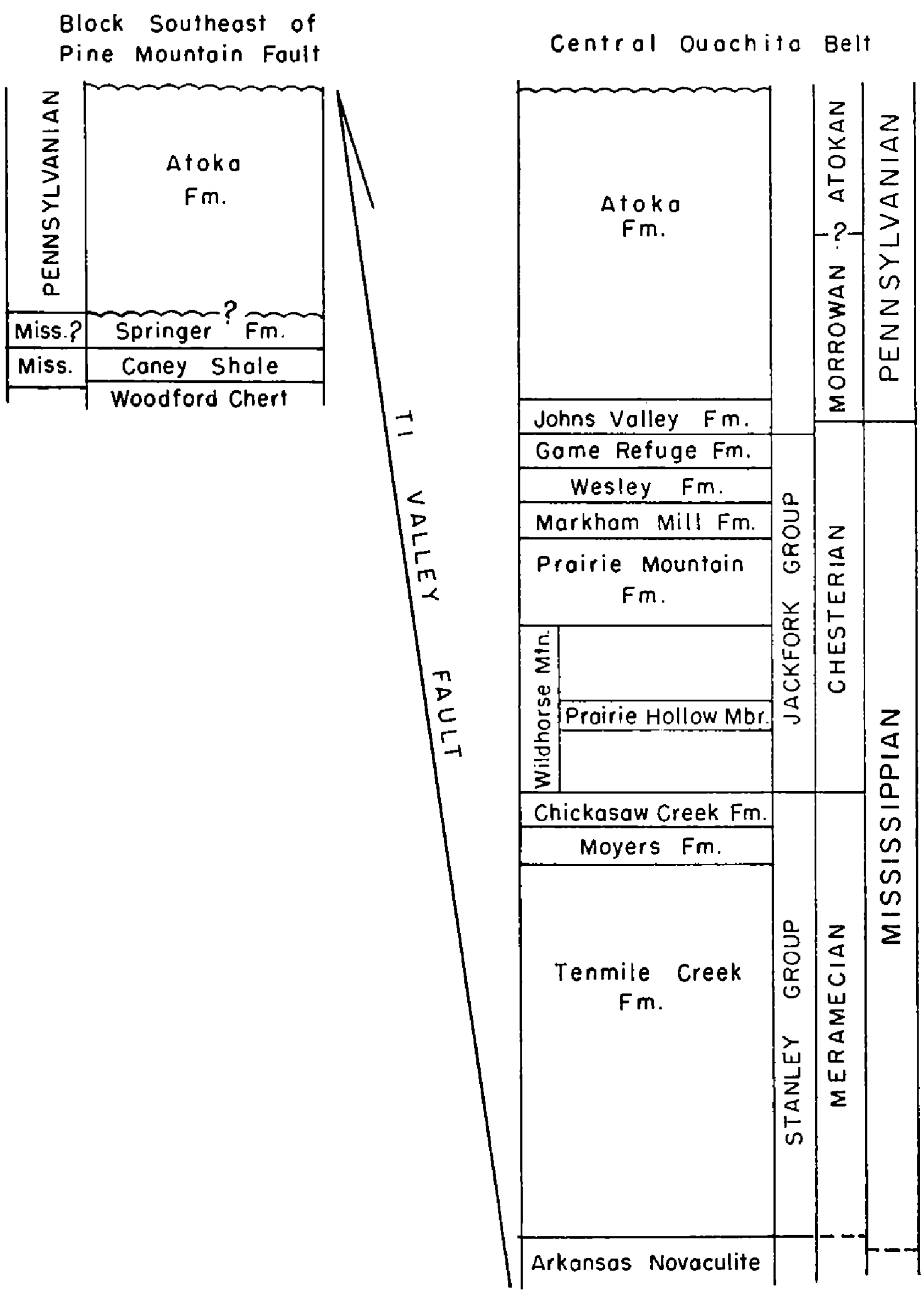

Figure 2. Generalized geologic columns for block southeast of Pine Mountain fault and for central Ouachita province.

STRATIGRAPHY

Stratigraphic Setting

OUACHITA STRUCTURAL PROVINCE

The Ouachita structural province can be traced northeastward from Chihuahua and Coahuila, Mexico, into the Marathon uplift, along the southern margin of the Llano uplift, into the Ouachita Mountains of Oklahoma and Arkansas, and under Cretaceous rocks of the Mississippi embayment (Flawn, 1959). The exact relationship of the Ouachita province to the Appalachian province and to the Laramide structures of northern Mexico is not known. Knowledge of the Ouachita province has been derived from observation of surface exposures in the Solitario uplift, the Marathon uplift, and the Ouachita Mountains of Oklahoma and Arkansas. Subsurface information is also available (Morgan, 1952).

The Ouachita province in southeastern Oklahoma, bounded on the north by the Choctaw fault, is overlapped on the south by rocks of Cretaceous age (fig. 1). That portion of the Ouachita province buried by Cretaceous rocks is known only from well samples, electric logs, and geophysical studies.

The Ouachita province in Oklahoma has been divided into the following units: (1) the frontal belt, (2) the central belt, and (3) the core area or the Broken Bow-Benton uplift (Hendricks, 1947). The frontal belt is between the Choctaw and Ti Valley faults. The central belt adjoins the frontal belt and extends to the core area, an anticlinorium in McCurtain County in which pre-Stanley strata are exposed. Hendricks (1947) subdivided the frontal belt into three major fault blocks: (1) the block southeast of the Choctaw fault, (2) the block southeast of the Katy Club fault, and (3) the block southeast of the Pine Mountain fault.

Most of the area of this investigation is in the central belt of the Ouachita province, as defined above. A small portion of the area is in the frontal belt, in the block southeast of the Pine Mountain fault.

TWO PALEOZOIC GEOSYNCLINES

In southeastern Oklahoma deposition occurred in geosynclinal proportions twice during the Paleozoic Era. The geosynclines not only received different types of sedimentary deposits but also had axes of deposition that intersected approximately at right angles. The thickest strata deposited in the Arbuckle geosyncline, predominantly carbonates, are of Late Cambrian and Ordovician ages. The strata deposited in the Ouachita geosyncline, predominantly clastics, are of Late Mississippian and Early Pennsylvanian ages.

PRE-STANLEY DEPOSITION

Pre-Stanley strata in the Arbuckle Mountains, Late Cambrian to Early Mississippian in age, are 6,800 to 11,500 feet thick, whereas their equivalents in the central Ouachita belt are 3,800 to 6,300 feet thick (Ham, 1959, p. 71). While the thick carbonate sequence was being deposited in the Arbuckle geosyncline during Late Cambrian and Ordovician times, a much thinner black shale and sandstone sequence was deposited in the Ouachita area.

Several times during the post-Ordovician pre-Late Mississippian interval, environmental conditions were similar in the Arbuckle and Ouachita areas (Hendricks and others, 1937, p. 18). The Viola, Sylvan, and Woodford Formations of the Arbuckle Mountains have the same ages and lithologic characteristics as their Ouachita equivalents, the Bigfork Chert, Polk Creek Shale, and Arkansas Novaculite, respectively (Ham, 1959, p. 83). At other times during this interval the rate of deposition and the type of sediment deposited in the two areas differed considerably.

Pre-Stanley strata in the Arbuckle Mountains are carbonates, cherts, and green shales. Equivalent strata in the central belt of the Ouachita province are black shales, cherts, and sandstones. Ham (1959) gave a more detailed discussion of pre-Stanley deposition in southeastern Oklahoma.

A significant observation made by Ham (1955, 1959) is that pre-Stanley rocks thin eastward from the Arbuckles and westward from central Arkansas as though they were merging toward a common shelf between them. This shelf is beneath the frontal belt of the Ouachita province. The significance of this observation is discussed on pages 75-76.

POST-WOODFORD DEPOSITION

After deposition of the Woodford Chert and the Arkansas Novaculite during Late Devonian and Early Mississippian time, the Ouachita geosyncline began to subside. Eventually it received at least 24,000 feet of clastic sediment. Deposited in the Ouachita geosyncline were the Stanley and Jackfork Groups (Middle and Upper Mississippian and Lower Pennsylvanian), and the Atoka Formation (Lower Pennsylvanian and lower Middle Pennsylvanian). Shelf equivalents of the geosynclinal units, 6,000 to 8,000 feet thick in the frontal belt, are somewhat thicker in the Arbuckle Mountains (fig. 2).

During the Late Mississippian-Early Pennsylvanian interval, as during pre-Late Mississippian time, greater deposition occurred both in the Arbuckle area and the Ouachita area than in the belt between. Pre-Late Mississippian deposition was greater in the Arbuckle area than in the Ouachita area. Late Mississippian-Early Pennsylvanian deposition was greater in the Ouachita area than in the Arbuckle area.

DEFINITION OF OUACHITA FACIES AND ARBUCKLE FACIES

In this report Ouachita facies refers to all Paleozoic strata in the central Ouachita province, south of the Ti Valley fault. Arbuckle facies refers to the Paleozoic strata west and northwest of the Ouachita province and in the Arbuckle Mountains. Strata in the frontal Ouachita belt, the belt bounded by the Choctaw and Ti Valley faults, are transitional between the Arbuckle facies and the Ouachita facies. These strata were referred to the Arbuckle facies by Ulrich (1927), Powers (1928), and Miser (1929). The relationship of strata in the frontal belt to the Arbuckle and Ouachita facies was described by Goldstein and Hendricks (1962, p. 387):

> The frontal zone lies between the Choctaw and Ti Valley faults and consists of at least three major fault blocks, each with distinctive stratigraphic features, that separate rocks typical of the central part of the Ouachita Mountains from their equivalents in the McAlester basin to the north and the Arbuckle Mountains to the west. The rocks in these fault

blocks have distinctive stratigraphic features, but all are lithologically intermediate between rocks of typical Ouachita facies and typical Arbuckle facies. Furthermore, these differences are progressive, as the fault block nearest the McAlester basin most closely resembles rocks of the Arbuckle facies, whereas those in the fault block adjacent to the central part of the Ouachitas most closely resemble typical sedimentary rocks of the Ouachita facies.

RELATIONSHIP OF POST-WOODFORD OUACHITA AND ARBUCKLE FACIES

The Caney Shale is interpreted as the direct lateral equivalent of the Stanley and Jackfork Groups and the lower part of the Johns Valley Formation south of the Ti Valley fault (Cline and Shelburne, 1959). The Springer Formation is considered to be the direct lateral equivalent of the upper part of the Johns Valley Formation south of the Ti Valley fault. The Wapanucka Limestone grades southward into the Chickachoc Chert. However, no Wapanucka or Chickachoc equivalents are recognized in the block southeast of the Pine Mountain fault (Hendricks, 1947). Wapanucka equivalents have been described from the Johns Valley Formation between the Ti Valley and Windingstair faults (Cline and Shelburne, 1959) and in the Boktukola syncline (Shelburne, 1960). Interbedded lenses of Wapanucka-like limestone and Wapanucka exotics in the upper part of the Johns Valley Formation occur in the area of this investigation. These limestone lenses are described on pages 52-53. The Atoka Formation, the youngest unit in the Ouachita province, was deposited throughout the entire province.

AREAL DISTRIBUTION OF STRATA

The Ti Valley fault is an important stratigraphic and structural boundary (Cline, 1960, p. 16). Strata deposited in the Ouachita geosyncline (Mississippian and Early Pennsylvanian) were thrust northward along the Ti Valley fault and are in contact with thinner, lithologically different shelf equivalents.

Pre-Stanley strata are exposed only in the following isolated areas: (1) core area, or Broken Bow-Benton uplift, (2) Potato Hills, and (3) inliers north of the Ti Valley fault.

Stratigraphic Problems

POOR EXPOSURES

Best exposures in the area of investigation are along streams or in road cuts. Other exposures are poor because of colluvial cover and dense vegetation. The Johns Valley and Atoka Formations, exposed at the surface in most of the area studied, are particularly susceptible to downslope movements such as rock creep. The few good exposures of the Johns Valley Formation are badly slumped or contorted. Because of this situation, attitudes were not measured on slopes, except in critical areas.

FACIES CHANGES

Different post-Woodford facies are separated by the Ti Valley fault. Strata deposited in the Ouachita geosyncline are absent north of the Ti Valley fault. Shale and carbonates were deposited on the northern shelf, whereas the geosyncline received much thicker equivalents of shale and sandstone.

Much of the northward change in rock type and thickness occurs between the Ti Valley and Windingstair faults. Tomlinson (1959, p. 1) described the sharp convergence in post-Woodford pre-Wapanucka strata toward Black Knob Ridge in Atoka County. Cline and Shelburne (1959, p. 207) stated that Stanley and Jackfork Groups thicken from nearly zero near the Ti Valley fault to 16,200 feet in the Tuskahoma syncline, approximately 13.5 miles to the southeast.

The relationship of northward thrusting to northward thinning is a subject of controversy. Cline and Shelburne (1959, p. 207) believed that, although horizontal movement along thrust faults might be responsible for some northwestward thinning, the rates of thinning of individual units are not significantly interrupted by thrusting. Goldstein and Hendricks (1962, p. 389) believed that rocks of Ouachita facies were thrust over rocks of Arbuckle facies for considerable distances.

INCOMPLETE SECTIONS NORTH OF WINDINGSTAIR FAULT

No complete stratigraphic sections were found in the portion of the area north of the Windingstair fault because of complex folding and faulting and of erosion of the top of the Atoka Formation, the youngest unit. The only complete sections in the area studied are those of the Chickasaw Creek Formation exposed on Buffalo Mountain. Because of the lack of complete sections, the rate of northward thinning toward the adjoining shelf could not be determined.

LACK OF STRATIGRAPHIC MARKERS

As previously stated, the Johns Valley and Atoka Formations are exposed at the surface in most of the area. Both the Game Refuge Formation and the Prairie Hollow Member of the Wildhorse Mountain Formation were recognized at only one locality. The Moyers, Chickasaw Creek, and Wildhorse Mountain Formations are exposed on the south side of Buffalo Mountain.

The Johns Valley Formation can be identified by the clasts of Arbuckle-like lithologies, commonly gray Ordovician limestones. Several exotic-bearing horizons, none of which is unique, are present. Probably these horizons are discontinuous. Lenses of Wapanucka Limestone in the upper part of the Johns Valley Formation are thin and discontinuous. A tongue of black shale of Caney lithology is present near the base of the formation.

Approximately 100 to 200 feet above the base of the Atoka Formation is a black siliceous shale unit. Although persistent, this 3-foot unit is at most places difficult to find. Other thin siliceous shale units are also present in the lower part of the Atoka Formation. No better stratigraphic markers in the Atoka Formation, which ranges in thickness from 1,000 to 7,000 feet, occur in the area.

DEVONIAN-MISSISSIPPIAN TRANSITIONAL ROCKS

WOODFORD CHERT

Nomenclature, classification, and correlation.—The Woodford Chert was named by Taff (1902, p. 4) for exposures in the northwestern part of the Atoka quadrangle near the town of Woodford in Carter County, Oklahoma. It is Late Devonian and Early Mississippian (Kinderhookian) in age (Hass, 1956). In the area of this investigation the Woodford Chert, which has not been subdivided, is overlain by the Caney Shale (Mississippian). The base of the Woodford is not exposed.

The Woodford Chert is the partial equivalent of the Arkansas Novaculite. Hass (1956), on the basis of a conodont study, stated that it is equivalent to the middle division of the Arkansas Novaculite at Caddo Gap, Arkansas, which is Late Devonian and Early Mississippian (Kinderhookian) in age. Ham (1959, p. 75) stated that both the lower and middle divisions of the Arkansas Novaculite are represented at Black Knob Ridge. The middle division is the equivalent of the Woodford Chert. The relationship of the lower division and its equivalents, if any, is not known (Ham, 1959, p. 75). Ham considered the Woodford Chert and the Arkansas Novaculite to have been deposited continuously between and within the Arbuckle and Ouachita Mountains.

Distribution, thickness, and boundaries.—The Woodford Chert, present throughout the Arbuckle Mountains, is exposed in the frontal belt of the Ouachita province in a series of fault slices, or inliers. In the area of this investigation it is exposed at only two places (NE¼ NE¼ SW¼ sec. 10, T. 4 N., R. 21 E.; NE¼ NW¼ SW¼ sec. 11, T. 4 N., R. 21 E.), both of which are in fault slices mentioned in the preceding paragraph. The latter exposure, the better of the two, is referred to as the Bengal locality. At this locality are exposed the upper 36 feet of the Woodford Chert, the Caney-Woodford contact, and the lower 45 feet of the Caney Shale.

As does the entire pre-Stanley sequence in southeastern Oklahoma, the Woodford Chert and its equivalents thin eastward from the Arbuckle Mountains and westward from central Arkansas toward the frontal belt of the Ouachita province (Ham, 1959). In most of the Arbuckle Mountains the Woodford is 300 to 400 feet thick. It attains a known maximum thickness of 560 feet in the Wapanucka syncline on the southeastern edge of the Arbuckle Mountains (Ham, 1955, p. 33).

In the frontal belt of the Ouachita province the Woodford ranges in thickness from 67 to 360 feet. At Black Knob Ridge, approximately 20 miles east of the west flank of the Wapanucka syncline, the Woodford is 234 to 360 feet thick (Hendricks and others, 1937). The Woodford is only 67 feet thick at the Pinetop School locality (Tulsa Geol. Soc., 1947, p. 23).

Southeastward the lower and middle divisions of the Arkansas

Novaculite thicken, reaching a maximum thickness at Caddo Gap, Arkansas. Harlton (1953, p. 791) stated that the upper division of the Arkansas Novaculite is present in the Potato Hills, but he did not give its thickness. Miller (1955, p. 19) stated that the Arkansas Novaculite is approximately 595 feet thick in the western Potato Hills. At Caddo Gap the lower and middle divisions are 466 feet and 347 feet thick, respectively, and the upper division, not recognized in the frontal belt, is 127 feet thick. The total thickness of the Arkansas Novaculite at Caddo Gap is 940 feet (Miser and Purdue, 1929).

If Ham's interpretation (1959, p. 75) is correct, the lower division of the Arkansas Novaculite thickens from 130 feet at Black Knob Ridge to 466 feet at Caddo Gap and the middle division thickens from 230 feet to 347 feet. Definite thinning of the Arkansas Novaculite occurs over the frontal belt of the Ouachita province. The relationship of the post-Hunton pre-Woodford un-

Figure 3. Contact of the Woodford Chert (right) and the Caney Shale at the Bengal locality (NW¼ SW¼ sec. 11, T. 4 N., R. 21 E.). Point of the geology pick is on the contact.

conformity (Maxwell, 1959) to the thickness of the Woodford Chert is not known.

The Woodford Chert, unconformable on older units (Maxwell, 1959), overlies the Missouri Mountain Shale (Silurian-Devonian) at Black Knob Ridge and the Pinetop Chert (Middle Devonian) at the Pinetop School locality. The Caney Shale (Mississippian) is unconformable upon the Woodford at both Black Knob Ridge and the Pinetop School locality (Hendricks, 1947). Ulrich (1927, p. 27) concluded that the Woodford-Caney contact is unconformable at Brushy Creek (W½ sec. 5, T. 2 N., R. 15 E.), largely because phosphatic nodules are present in the basal unit of the Caney. At the Bengal locality the Caney appears to be conformable upon the Woodford (figs. 3, 4).

Lithologic characteristics.—At the Bengal locality (NE¼ NW¼ SW¼ sec. 11, T. 4 N., R. 21 E.) the Woodford Chert consists of alternating laminated black shale and black chert in beds 1 to 5 inches thick (fig. 4). Phosphatic nodules are abundant both in the

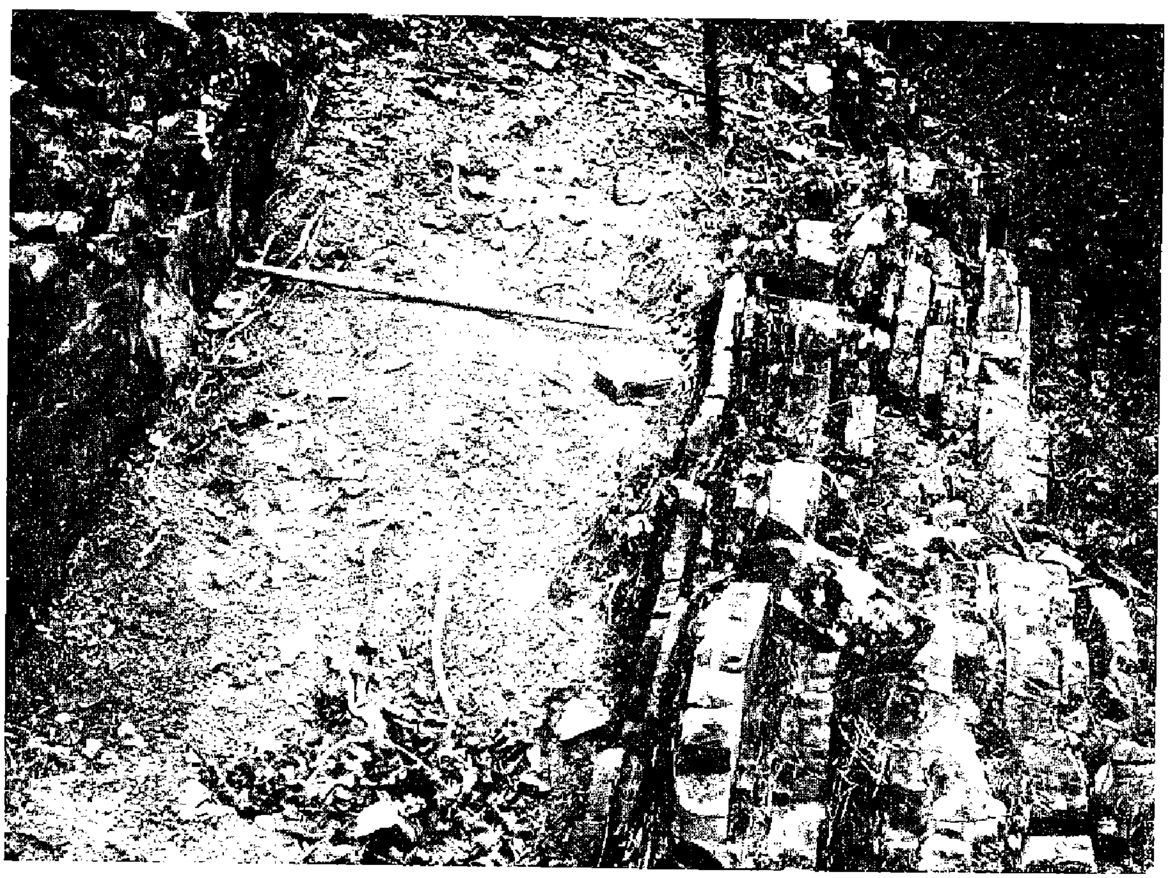

Figure 4. Alternating beds of black chert and shale of the Woodford Chert at the Bengal locality. The stick, three and one-half feet long, is lying on green silty shale (unit 3 of measured section, appendix A) that is similar to some of the shales of the Stanley Group. Woodford-Caney contact is to the right.

shale and in the chert. Rocks exposed at the Bengal locality are described in detail in appendix A.

Lithologically the Woodford Chert at the Bengal locality and that at the Pinetop School locality are almost identical. The guidebook for Tulsa Geological Society field conference in western part of the Ouachita Mountains in Oklahoma (1947, p. 23) gives a detailed description of the Woodford Chert at the Pinetop School locality.

Mississippian System

CANEY SHALE

Nomenclature, classification, and correlation.—The Caney Shale was defined by Taff (1902, p. 4). Much confusion resulted from subsequent attempts to subdivide the Caney and to correlate it in the Ouachita province. Elias and Branson (1959) discussed the history of nomenclature of the Caney Shale, designated a type section in an area of good exposures, subdivided it into three members (Ahloso, Delaware Creek, and Sand Branch, in ascending order), and interpreted it as being Meramecian and Chesterian. In the southern Arbuckles, Elias (1956) correlated the Goddard Shale with the Sand Branch Member. He later (1957, p. 425) assigned the lower part of the Goddard to the latest Chesterian (Clore-Kinkaid) interval. Therefore, the upper part of the Goddard might also be post-Kinkaid Mississippian. In the area of this investigation the Caney Shale has not been subdivided. The classification used in this report is the same as that of Elias and Branson (1959, p. 22).

Miser and Honess (1927) presented two possible interpretations of the relationship of the Caney Shale to the Stanley and Jackfork Groups: (1) the Caney overlaps the Stanley and Jackfork Groups and (2) the Caney is a northwesterly development of the same age as the Stanley and Jackfork Groups. Cline and Shelburne (1959) favored the latter interpretation but also stated that a tongue of Caney shale is in the lower part of the Johns Valley Formation. This was pointed out earlier by Cline (1956b, p. 104). The present interpretation of Cline (1960, p. 18) is that the Caney Shale north of the Ti Valley fault is the direct lateral equivalent of the Stanley and Jackfork Groups and the lower part of the Johns Valley Formation south of the Ti Valley fault.

Cline (1960, p. 10) diagrammatically illustrated the possibility that the siliceous shales of the Stanley and Jackfork Groups and the lower part of the Johns Valley Formation are actually tongues of Caney shale. He stated (p. 45) that "the type section of the Wesley is almost certainly a tongue of the Caney." The writer shares the interpretation that the siliceous shales in the Stanley and Jackfork Groups and the lower part of the Johns Valley Formation are tongues of the Caney that extend basinward.

Distribution, thickness, and boundaries.—The Caney Shale is present in the Arbuckle Mountains and in the frontal belt of the Ouachita province, where it is exposed in a series of fault slices. In the report area the Caney is exposed at only two localities (NE¼ NE¼ SW¼ sec. 10, T. 4 N., R. 21 E.; NE¼ NW¼ SW¼ sec. 11, T. 4 N., R. 21 E.), both of which are in one of the fault slices mentioned above. At the latter exposure, the Bengal locality, the Woodford-Caney contact and the lowermost 45 feet of the Caney are exposed.

The only complete section of the Caney Shale that has been described is at the Pinetop School locality, where the Caney is 524 feet thick (Tulsa Geol. Soc., 1947, p. 20). At the type locality to the west it is 386.5 feet thick (Elias and Branson, 1959). In the Arbuckles, along the Arbuckle anticline, the Caney Shale is approximately 1,200 feet thick (Ham, 1955, p. 31). Ham stated that this westward thickening is similar to the thickening of the pre-Stanley sequence. The Caney Shale and its equivalents are thicker in the Arbuckle Mountains and in the central Ouachita province than within the frontal belt.

The Caney Shale overlies the Woodford Chert (Upper Devonian) and is overlain by the Springer Formation (uppermost Mississippian). According to Hendricks and others (1937), Hendricks (1947), and Goldstein and Hendricks (1962), the Caney is unconformable on the Woodford at Black Knob Ridge and at the Pinetop School locality, even though they appear to be gradational. The Caney is conformable on the Woodford at the Bengal locality (figs. 3, 4). In the frontal belt the Caney-Springer contact is conformable (Hendricks, 1947; Goldstein and Hendricks, 1962).

Lithologic characteristics.—The Caney Shale consists predominantly of laminated black shale, some of which is siliceous. Dense black limestone concretions, sideritic concretions, and marblelike phosphatic nodules are common. The basal unit of the Caney is green, silty, and has phosphatic nodules at both the Pinetop School and the Bengal localities. This unit resembles the Stanley Shale. At the Bengal locality a 3-foot bed of green silty shale is interbedded with black shale and chert of Woodford lithology (fig. 4). Appendix A gives a detailed description of the rocks exposed at the Bengal locality.

The Caney Shale at the Bengal locality is almost identical with that at the Pinetop School locality. A detailed description of the Caney Shale at the Pinetop School locality was published by the Tulsa Geological Society (1947, p. 24).

SPRINGER FORMATION

Nomenclature, classification, and correlation.—The Springer Formation was first defined by Goldston (1922, p. 7) as a member of the Glenn Formation (Pennsylvanian) in Carter County, Oklahoma. Tomlinson (1928, p. 245) subdivided it. In this report the Springer is assigned to the Upper Mississippian, as was done by Branson (1959, p. 114). The Springer Formation in the Arbuckle Mountains has been subdivided into several members, none of which has been identified in the frontal belt of the Ouachita province. The equivalent of the Springer south of the Ti Valley fault is the upper part of the Johns Valley Formation.

Distribution, thickness, and boundaries.—The Springer Formation is exposed throughout the Arbuckle Mountains and in the frontal belt of the Ouachita province. In the report area exposures at two localities were mapped as Springer (SE¼ sec. 2, T. 4 N., R. 21 E.; SE¼ NW¼ sec. 2, T. 4 N., R. 21 E.).

According to Hendricks (1947), the Springer Formation is approximately 2,500 feet thick in the block southeast of the Choctaw fault, but only 100 to 500 feet thick in the block southeast of the Pine Mountain fault. Tomlinson (1929, p. 16) stated:

> Sediments between the Sycamore and Wapanucka limestones are reported to have a maximum thickness of only 1,600 feet north and east of the Arbuckle Mountains, where they are all assigned to the Caney shale; as compared with fully 5,000 feet in the Ardmore basin, where the corresponding strata include the Caney and Springer formations. This great increase in thickness calls to mind the

> even greater increase in thickness of supposed early Carboniferous formations eastward from the Atoka-Coalgate area, to the Stanley-Jackfork section in the Ouachita Mountains.

These observations indicate that the Caney-Springer sequence thins eastward from the Arbuckle Mountains, and its equivalents thin westward from the central Ouachita province. Ham (1959) described a similar thinning of pre-Stanley strata.

Southeast of the Choctaw fault the Springer Formation overlies the Caney with no evidence of an angular unconformity, whereas in the block southeast of the Pine Mountain fault it is unconformable upon the Caney Shale (Hendricks, 1947).

Goldstein and Hendricks (1962, p. 410) diagrammatically showed the Atoka Formation to be unconformable upon the Caney Shale in the block southeast of the Pine Mountain fault. They interpreted the Wapanucka and Springer as having been removed by erosion.

Lithologic characteristics.—The Springer Formation, which consists predominantly of gray shale and minor thin beds of sandstone and siltstone, is lithologically similar to the upper part of the Johns Valley Formation, but does not contain exotics (Cline and Shelburne, 1959, p. 204).

STANLEY GROUP

Nomenclature, classification, and correlation.—The Stanley Shale was named by Taff (1902, p. 4) for exposures near the village of Stanley in Pushmataha County, Oklahoma. Harlton (1938, p. 856) elevated the Stanley to group rank and subdivided it. Cline (1960, p. 23) gave a more detailed discussion of the history of nomenclature.

In the classification used in this report, essentially the same as that used by Cline and Shelburne (1959, p. 179), the Stanley Group is regarded as Meramecian(?). The classification differs from that of Harlton (1938) only in age assignments. Because the precise ages of the Stanley and Jackfork Groups are controversial, the Stanley is arbitrarily classified as Meramecian(?) and the Jackfork as Chesterian(?).

The sparse fauna of the Stanley and Jackfork Groups is nondiagnostic. A few conodonts collected from the lower part of the Stanley were assigned a Meramecian age by Hass (1950). Cline

(1956a) illustrated that the lower part of the Johns Valley Formation has a tongue of the Caney Shale. Upon this basis the Stanley and Jackfork Groups, stratigraphically below the Johns Valley Formation, cannot be younger than Late Mississippian. The entire Stanley Group is considered to be Meramecian on the basis of the conodonts collected from its lower part. The Jackfork Group has been assigned to the Chesterian.

Elias (1959, p. 158) concluded that there is strong evidence that the conodonts in the lower part of the Stanley Group are older than Meramecian and that there must be a re-evaluation of conodont classification. Elias and Branson (1959, p. 22) interpreted the lower part of the Johns Valley as probably containing an Ahloso equivalent (Meramecian).

In this report the Stanley Group is considered to be the direct lateral equivalent of the lower part of the Caney Shale north of the Ti Valley fault.

Distribution, thickness, and boundaries.—The Stanley Group is absent north of the Ti Valley fault. In the area studied it was not identified in the block between the Ti Valley and Windingstair faults. Although exposures of the Stanley Group are widespread in the central Ouachita belt, all of the recognized subdivisions of the Stanley Group could not be identified in the area studied. However, Stanley rock types could easily be distinguished from other units. Hendricks (1947), Johnson (1954), and Roe (1955), who had similar problems in their areas of study, classified the Stanley and Jackfork as formations and their constituent units as members.

The Stanley Group, 10,000 to 11,000 feet thick in the central belt of the Ouachita province, appears to thin rapidly northward. No complete unfaulted section of the Stanley has been measured.

The Stanley Group, the first stratigraphic unit to be deposited in the Ouachita geosyncline, is conformable upon the Arkansas Novaculite (Upper Devonian-Osagean) and is conformably overlain by the Jackfork Group (Chesterian?). At many places in the Potato Hills the Arkansas Novaculite-Stanley transition occurs over a 100-foot zone, making location of a contact quite difficult. The Stanley-Jackfork contact, also gradational, is arbitrarily placed at the top of the Chickasaw Creek Formation by some (Harlton, 1938; Johnson, 1954; Cline, 1956a) and at the base of the Chickasaw Creek Formation by others (Hendricks, 1947). In this report the top of the Chickasaw Creek is considered to be the top of the Stanley Group.

Lithologic characteristics.—The Stanley Group consists predominantly of olive-weathering clay shales that are dark gray in fresh exposures. Green, micaceous, argillaceous, quartzose sandstones, of minor importance in the lower four-fifths of the group, are abundant in the upper 1,500 feet (Cline, 1960, p. 22).

Many of the sandstones have a pronounced bimodal grain-size distribution. This observation is based on thin-section analysis of sandstones collected in the report area as well as examination of published photomicrographs. The largest grains, predominantly quartz of medium-sand size, are commonly subrounded or rounded. The matrix consists predominantly of angular to subrounded quartz grains of fine-sand or coarse-silt size. Most of the sandstones are poorly sorted but are not graded. The coarse grains are randomly distributed in a matrix of finer grains (fig. 5).

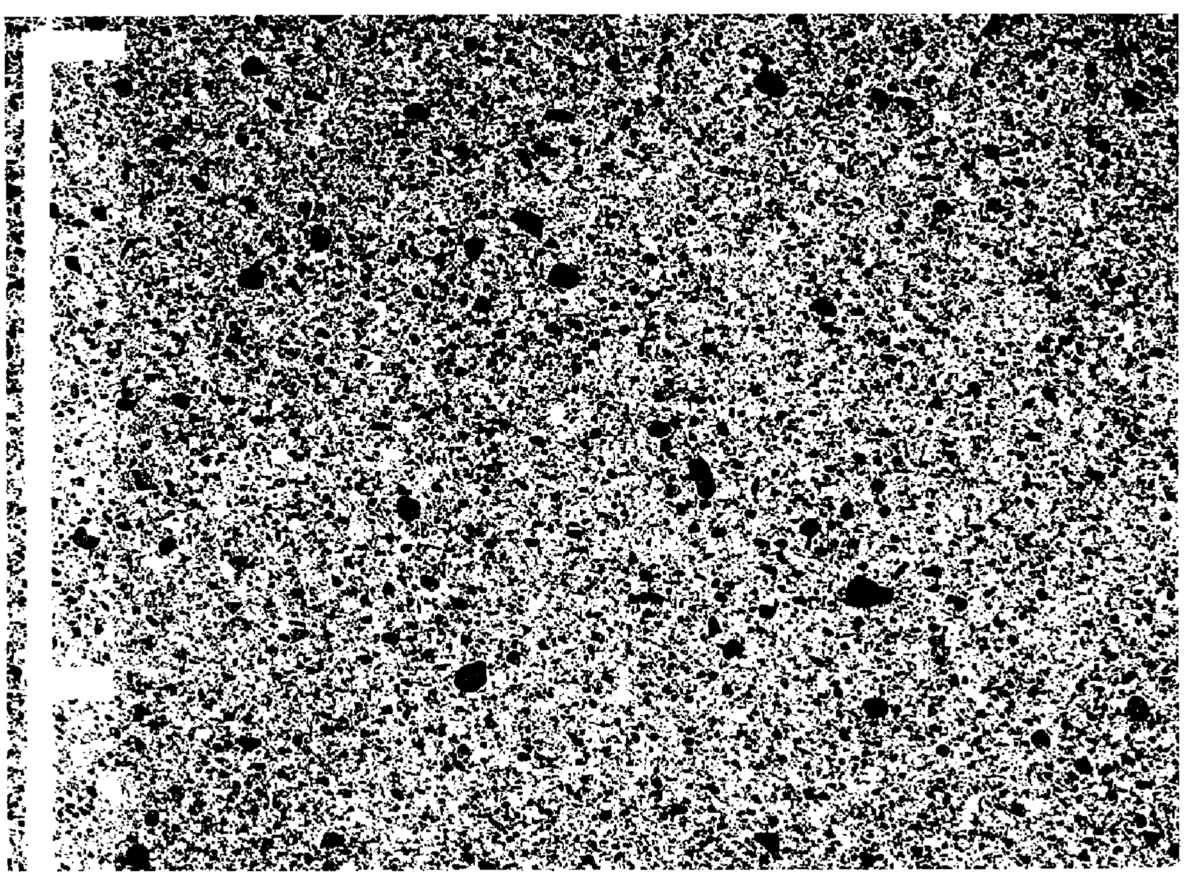

Figure 5. Negative print showing grain-size distribution of a Stanley sandstone. Large quartz grains (black), many of which are well rounded, are in a matrix of smaller grains. Scale is in ½-inch units. Specimen 25, ordinary light.

Chert, rock fragments, feldspar, and heavy minerals are present in the sandstones in minor amounts. Clay-size particles are common in most of the sandstones. Some sandstones have an abundance of macerated plant fragments. Several laterally persistent black siliceous shale and chert units are present. Honess (1921), Bokman (1953), Cline (1960), and Goldstein and Hendricks (1962) gave more detailed descriptions of lithologic characteristics.

That many of the sandstones in the Stanley Group have sole marks was noted by Bokman (1953) and Johnson (1954). Cline (1960) described them in more detail.

Structurally the Stanley Group is incompetent. It is easily eroded and generally forms floors of broad, flat valleys, such as the Kiamichi Valley or Buffalo Valley. The more resistant Jackfork Group and pre-Stanley units rise above the Stanley valleys.

Stratigraphic units.—Upon the basis of black siliceous shales, Harlton (1938) subdivided the Stanley Group into the Tenmile Creek, Moyers, and Chickasaw Creek Formations.

TENMILE CREEK FORMATION.—The Tenmile Creek Formation, the oldest subdivision of the Stanley Group, lies conformably upon the Arkansas Novaculite (Devonian-Osagean) and is conformably overlain by the Moyers Formation. The Tenmile Creek Formation, approximately 9,000 feet thick in Jumbo Valley (Laudon, 1959) makes up more than four-fifths of the Stanley Group (Cline, 1960). It was from the base of this unit that Hass (1950) described Meramecian conodonts. The Tenmile Creek consists predominantly of olive-weathering shale, minor interbedded argillaceous sandstone, and a few thin black siliceous shales. Many of the sandstones have sole marks.

In the report area only the uppermost part of the Tenmile Creek Formation is exposed.

MOYERS FORMATION.—The Moyers Formation, 1,100 feet thick in the type area, is conformable upon the Tenmile Creek Formation and below the Chickasaw Creek Formation. A fairly persistent black siliceous shale is present at the base of the Moyers Formation.

The Moyers Formation consists predominantly of olive-weathering shale and minor argillaceous sandstones, some of which are massive. Sole marks are common on the bases of many of the sandstone beds. Small-scale cross-lamination is well developed locally.

In the area studied the basal siliceous shale member of the

Moyers is well exposed near the tuberculosis sanatorium (NW¼ sec. 2, T. 3 N., R. 21 E.). These exposures were described and photographed by Seely (1955, p. 33).

The Moyers Formation, which has a greater proportion of sandstone than the underlying Tenmile Creek Formation, is more resistant to erosion. Best exposures of the Moyers in the area studied are at the base of Buffalo Mountain.

Chickasaw Creek Formation.—The Chickasaw Creek Formation, approximately 350 to 400 feet thick, according to estimates made from aerial photographs, is conformable upon the Moyers Formation and below the Wildhorse Mountain Formation of the Jackfork Group. The Stanley-Jackfork contact is arbitrarily placed at the top of the Chickasaw Creek Formation.

The Chickasaw Creek consists of alternating olive-weathering shale, argillaceous sandstone, and black siliceous shale and chert. White-spotted black chert is characteristic of the formation. Massive sandstone beds as much as five feet thick are common.

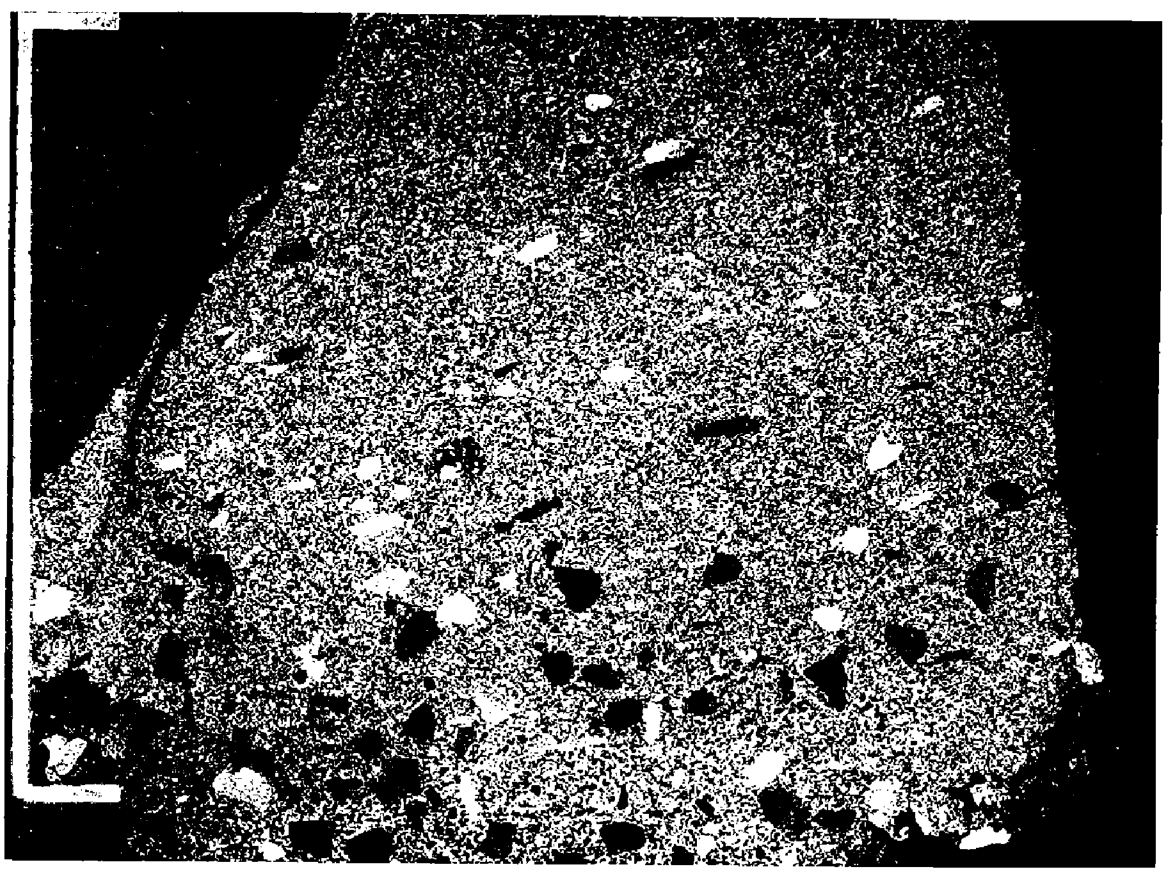

Figure 6. Negative print showing progressive vertical decrease in size of quartz pebbles and rock fragments in a sandstone of the Chickasaw Creek Formation. Scale is in ½-inch units. Specimen 114, ordinary light.

A conglomerate sandstone containing pebbles of Arbuckle-like rocks is present within the Chickasaw Creek (Hendricks, 1947). Locally the pebbles in this conglomerate are graded (fig. 6).

Along the base of Buffalo Mountain, the base of the Chickasaw Creek Formation is marked by two 10-foot units of black siliceous shale and chert, separated by approximately 35 feet of olive-weathering clay shale. The best exposures of the Chickasaw Creek in the area are at the northeastern end of Buffalo Mountain, less than 100 yards above the west end of the dam on Talihina Lake (NW¼ sec. 35, T. 4 N., R. 21 E.).

JACKFORK GROUP

Nomenclature, classification, and correlation.—The Jackfork Formation was named by Taff (1902, p. 4) for exposures on Jackfork Mountain in Atoka County, Oklahoma. Harlton (1938, p. 856) elevated the Jackfork to group rank and subdivided it. Cline (1960, p. 42) gave a more detailed discussion of the history of nomenclature.

Although both the Stanley and Jackfork Groups are considered to be Mississippian, their precise ages are controversial. In this report the Stanley is classified as Meramecian(?) and the Jackfork as Chesterian(?), essentially the same classification used by Cline and Shelburne (1959, p. 190).

In this report the Jackfork Group is considered to be the direct lateral equivalent of the upper part of the Caney Shale north of the Ti Valley fault (Cline and Shelburne, 1959, p. 190). This interpretation was also presented by Miser and Honess (1927, p. 12).

Distribution, thickness, and boundaries.—The Jackfork Group, absent north of the Ti Valley fault, is widely distributed in the central Ouachita belt. In the area studied, the Jackfork is not fully represented. It appears to be thinner between the Windingstair and the Ti Valley faults than in the remainder of the central Ouachita belt. Only one exposure of the Prairie Hollow Member of the Wildhorse Mountain Formation was observed (sec. 10, T. 3 N., R. 19 E.). Only one exposure of the Game Refuge Formation was identified (secs. 1, 2, T. 3 N., R. 19 E.). A portion of the lower part of the Wildhorse Mountain Formation is exposed on Buffalo Mountain.

Exotic-bearing horizons in the Johns Valley Formation exposed in road cuts along State Highway 2 (SW¼ sec. 2, NE¼ sec. 10, and NW¼ sec. 11, T. 3 N., R. 19 E.) are here mapped as Johns Valley Formation. These were mapped as Wesley Formation by Hendricks and others (1947) before the highway was widened and new road cuts were constructed. This locality is also described as Wesley at mile 49.35 in a recent guidebook (Cline and others, 1959, guidebook, p. 42).

The Jackfork Group, 5,000 to 6,000 feet thick in the central belt of the Ouachita province, apparently thickens toward west-central Arkansas. Miser and Purdue (1929) found the Jackfork to be 5,000 to 6,600 feet thick in the De Queen and Caddo Gap quadrangles, Arkansas. Cline and Moretti (1956) measured two sections of Jackfork in the Kiamichi Range and found them to be 5,600 and 5,800 feet thick. They believed that this is a good average thickness for the Jackfork in the central Ouachitas.

The Caney Shale, the shelf equivalent of the Jackfork and Stanley Groups, is 524 feet thick at the Pinetop School locality. Therefore the shelf equivalent of the Jackfork Group cannot be more than a few hundred feet thick.

The Jackfork Group is conformable upon the Stanley Group (Meramecian?) and below the Johns Valley Formation (Chesterian and Lower Pennsylvanian).

Lithologic characteristics.—The Jackfork Group consists of sandstone and subequal amounts of gray shale. Several black siliceous shale and chert units, similar in lithology to those in the Stanley Group, are also present. One persistent unit, the Prairie Hollow Member of the Wildhorse Mountain Formation, consists of maroon and green shale and friable gray sandstone.

Bimodal grain-size distribution of sand grains, similar to that observed in Stanley sandstones, was also observed in many, but not all, of the Jackfork sandstone units.

The sandstones of the Jackfork Group are much more resistant to erosion than are the shales of the underlying Stanley Group. Structurally, the Jackfork Group is a competent unit.

Bokman (1953), Cline and Moretti (1956), Moretti (1958), Cline and Shelburne (1959), and Goldstein and Hendricks (1962) gave more detailed descriptions of the lithologic characteristics of the Jackfork Group.

Stratigraphic units.—The Jackfork Group, subdivided largely on the basis of siliceous shales, consists of the following formations, in ascending order: Wildhorse Mountain, Prairie Mountain, Markham Mill, Wesley, and Game Refuge. Only the units present in the area of investigation—the Wildhorse Mountain Formation, the Prairie Hollow Member of the Wildhorse Mountain Formation, and the Game Refuge Formation—are discussed below.

Wildhorse Mountain Formation.—The Wildhorse Mountain Formation, the oldest unit of the Jackfork Group, is conformable upon the Chickasaw Creek Formation (Meramecian?) and below the Prairie Mountain Formation (Chesterian?). It is approximately 3,550 feet thick at the type locality on Wildhorse Mountain in Pushmataha County, Oklahoma (Harlton, 1938).

The Wildhorse Mountain Formation consists of alternating sandstone and black shale. In some zones sandstone predominates, whereas in others black shale is predominant. The sandstone beds, which range in thickness from one inch to several feet, consist of poorly sorted quartz grains of medium to fine size.

The Wildhorse Mountain Formation, exposed at the top of Buffalo Mountain, has been eroded and the Prairie Hollow Member is absent.

Prairie Hollow Shale Member of the Wildhorse Mountain Formation.—In the middle of the Wildhorse Mountain Formation is a unit, approximately 300 feet thick, consisting of alternating maroon and green shale and friable gray sandstone.* This unit was recognized by Hendricks (1947) between the Ti Valley and Windingstair faults. Hendricks and others (1947) mapped it as the maroon shale member of the Jackfork Formation. Cline (1960, p. 49) gave a more detailed description of the lithology of the Prairie Hollow Shale Member.

The Prairie Hollow Shale Member, easily eroded, forms a narrow strike valley that can be traced on aerial photographs.

Game Refuge Formation.—The Game Refuge Formation, named by Harlton (1959, p. 135), is 350 to 400 feet thick in the western Ouachitas (Cline and Shelburne, 1959, p. 193). It is conformable upon the Wesley Formation (Chesterian?) and below the

* On plate I, the geologic map of the area, the Prairie Hollow Shale Member is erroneously shown to be at the top of the Wildhorse Mountain Formation. The only outcrop of the formation in which the member was mapped covers less than 0.25 square mile.

Johns Valley Formation (Chesterian-Lower Pennsylvanian). As presently defined, the Game Refuge Formation is the youngest subdivision of the Jackfork Group.

The Game Refuge consists of gray quartzose sandstone and gray shale. Sandstones containing a mold fauna or abundant plant fragments are common. Many sandstones have interference ripple marks (fig. 7). Cline stated that the Game Refuge Sandstone is lithologically similar to other sandstone units in the Jackfork Group (Cline and others, 1959, guidebook, p. 42). Cline and Shelburne (1959, p. 192) gave a more detailed description of the lithologic characteristics of the Game Refuge Formation.

Figure 7. Ripple marks on a sandstone bed of the Game Refuge Formation exposed along State Highway 2 (SE¼ NW¼ sec. 2, T. 3 N., R. 19 E.) southeast of the Hairpin Curve locality.

MISSISSIPPIAN-PENNSYLVANIAN TRANSITIONAL ROCKS

JOHNS VALLEY FORMATION

More has been written about the Johns Valley Formation than about any other stratigraphic unit in the Ouachita province. The origin of the exotic clasts of Arbuckle-type rocks in place at several horizons in the Johns Valley is one of the more challenging problems encountered in the Ouachita province. The different theories explaining the emplacement of the exotics in the Johns Valley Formation were discussed by Miser (1934, p. 933), Rea (1947, p. 48), Tomlinson (1959, p. 10), Cline and Shelburne (1959, p. 203), Cline (1960, p. 82), and Goldstein and Hendricks (1962, p. 400). The exotic-bearing facies of the Johns Valley Formation crops out within the area studied. Therefore, a study of this area is critical to an interpretation of the emplacement of the exotic clasts. The evidence obtained from this field study has given insight into the mode of emplacement of the exotics. However, most of the conclusions reached, although derived from this field study, have previously been presented.

Nomenclature, classification, and correlation.—The Johns Valley Formation was named by Ulrich (1927, p. 21) for exposures in the center of the Tuskahoma syncline. However, as pointed out by Miser and Honess (1927, p. 22-23), this was also Taff's type area for the Caney Shale. Cline and Shelburne (1959, p. 193) gave a more detailed discussion of the history of nomenclature.

In this report the Johns Valley Formation is considered to be of Late Mississippian and Early Pennsylvanian age, as it was by Miser and Honess (1927, p. 12). Cline (1956b) and Cline and Shelburne (1959) also used this classification. Harlton (1933) described a Pennsylvanian fauna from the Johns Valley. Moore (1934, p. 441) observed Morrowan fossils in the Johns Valley at the Compton Cut locality. Cline (1956b, p. 104) concluded that the Pennsylvanian fauna described from the Johns Valley by Harlton (1933) was from the upper part of the formation and that a tongue of Caney Shale (Mississippian) is in place in the lower part of the formation. Elias and Branson (1959, p. 22) stated that the Johns Valley contains in its lower part "probably an Ahloso equivalent (Meramecian), definitely a Delaware Creek equivalent (late Meramecian and early Chesterian), and possibly a Sand Branch equivalent (late Chesterian)." Cline and Shelburne (1959, p. 204) concluded that the upper part of the Johns Valley is equivalent to some part of the Wapanucka Formation. Thin, discontinuous lenses of Wapanucka-like limestone, interbedded in the upper part of the Johns Valley Formation, and exotics of Wapanucka rock types occur at several localities in the area studied. The limestone lenses, interpreted here as Wapanucka equivalents, are described in more detail in the section on the Wapanucka Limestone.

In reference to the age and correlation of the Johns Valley Formation, Tomlinson (1959, p. 14) stated:

> This problem, therefore, now appears to be near the same solution which was stated by Girty (1927)[1] thirty years ago;—that the "Caney" shale of the Ouachita region (later called Johns Valley shale—Johns Valley shale *and* Round Prairie formation of Harlton, 1959) is Mississippian in its lower part and Pennsylvanian in its upper part; and that the Stanley and Jackfork are both of Mississippian age. If Cline is right, Taff and Girty were also right (1902, 1909) in treating the Caney shale of the Arbuckle region as one and the same formation as the Caney shale of the Ouachitas.

The Johns Valley Formation, as presently defined, has equivalents of the Caney Shale (Mississippian), Springer Formation (uppermost Mississippian), and the Wapanucka Limestone (Lower Pennsylvanian). In much of the central belt of the Ouachita province the Jackfork Group, Johns Valley Formation, and the Atoka Formation are conformable and have similar lithologies (Cline, 1956b, p. 105).

Distribution, thickness, and boundaries.—The Johns Valley Formation, absent north of the Ti Valley fault, is widely distributed in the central belt of the Ouachita province; however exotic-bearing horizons are restricted to a narrow belt bounded on the north by the Ti Valley fault. The exotic-bearing belt, nowhere more than 25 miles wide, extends from Atoka, Oklahoma, to near Boles, Arkansas, a distance of approximately 130 miles. Cline and Shelburne (1959, p. 195) observed that, although no exotics have been found in the Kiamichi Range east of State Highway 2, a shale unit in the stratigraphic position of the Johns Valley can be traced eastward on aerial photographs. Shelburne (1960) mapped the Johns Valley Formation in the Boktukola syncline but found no exotics.

Hendricks (1947) stated that the Johns Valley Formation is

[1] Miser and Honess, 1927, p. 23, 24.

200 to 1,000 feet thick in the western Ouachita province. According to Cline and Shelburne (1959, p. 194), it ranges from 425 to 900 feet in thickness. Shelburne (1960, p. 36) reported that it is 600 to 750 feet thick in the Boktukola syncline. Goldstein and Hendricks (1962, p. 398) concluded that the Johns Valley thickens from 200 feet in the extreme northwestern part of the central Ouachitas to as much as 1,000 feet farther east.

No unfaulted section of the Johns Valley Formation is present in the report area. At the Hairpin Curve locality, the best exposure observed by the writer, it is at least 425 feet thick. The Hairpin Curve section was measured by Hendricks and Averitt in 1939 (Tulsa Geol. Soc., 1947, p. 32) and by Cline and Laudon in 1958 (Cline and others, 1959, guidebook, p. 35) after State Highway 2 had been widened and new road cuts had been made. Judged from the width of outcrop bands and the steepness of dip, the maximum thickness of the Johns Valley in the report area is estimated to be from 500 to 700 feet.

The Johns Valley Formation is conformable upon the Jackfork Group (Chesterian?) and is overlain by the Atoka Formation (Pennsylvanian). The Johns Valley-Atoka contact is conformable in the central belt and in most of the frontal belt. However, Goldstein and Hendricks (1962, p. 410) diagrammatically showed the Atoka to be unconformable upon the Caney in the block southeast of the Pine Mountain fault.

Lithologic characteristics.—The Johns Valley Formation consists predominantly of gray-green clay shale and interbedded sandstone. Near its base in the area studied is a black shale member of typical Caney lithology. Localities at which Caney Shale is exposed in the lower part of the Johns Valley Formation are listed in appendix B. The better exposures of Caney lithologies in the report area are in Coopers Cove (NE¼ NE¼ sec. 32, T. 4 N., R. 20 E.) and at the Hairpin Curve locality. Exotic-bearing horizons are present throughout the formation. Rolled sandstone masses are present within the exotic-bearing horizons. Thin beds of conglomeratic sandstone, although not common, are in the upper part of the Johns Valley Formation. Rusty weathering sideritic concretions occur throughout the formation. Thin, discontinuous lenses of limestone, lithologically similar to those of the Wapanucka and interpreted as Wapanucka equivalents, are interbedded with the shale and sandstone of the upper unit. Wapanucka exotics were also observed.

Because of the predominance of clay shale, the Johns Valley is susceptible to slumping. Structurally, it is an incompetent unit. The sandstone of the overlying Atoka Formation, more resistant to erosion, forms ridges, whereas the Johns Valley is a valley former.

Exotic-bearing horizons, clasts of Arbuckle-like rocks, conglomeratic sandstone, interbedded sandstone, and rolled sandstone masses, each of which gives some clue to the origin of and mode of emplacement of the Johns Valley exotics, are described in more detail below.

Exotic-bearing horizons.—Although the Johns Valley Formation is widely distributed in the central belt of the Ouachita province, exotic-bearing horizons within it are restricted to a narrow belt 10 to 25 miles wide. This belt, bounded on the north by the Ti Valley fault, extends from Atoka, Oklahoma, to Boles, Arkansas, a distance of approximately 130 miles. The exotic-bearing belt is approximately 25 miles wide in the western Ouachitas, whereas in

Figure 8. Clasts of Arbuckle-like rock in a channel, or pocket, in the Johns Valley Formation at the Hairpin Curve locality (NW¼ SE¼ sec. 3, T. 3 N., R. 19 E.).

west-central Arkansas, as mapped by Reinemund and Danilchik (1957) and Morris (1962), it is less than 10 miles wide.

Several exotic-bearing horizons, none of which is unique, are present. Probably none of these horizons is laterally persistent. An exotic-bearing horizon is present below the tongue of the Caney Shale at Hairpin Curve. In Johns Valley, exotics are present locally in the tongue of the Caney Shale (Cline, 1956b). In the area studied exotics are abundant in the Caney Shale at Coopers Cove (NE¼ sec. 32, T. 4 N., R. 20 E.). Several exotic-bearing horizons are present in the upper member of the Johns Valley Formation at the Hairpin Curve locality. In 1939 six exotic-bearing horizons at the Hairpin Curve locality were described by Hendricks and Averitt (Tulsa Geol. Soc., 1947, p. 32). However, after new road cuts were made at the same locality, only three exotic-bearing horizons were noted (Cline and others, 1959, p. 35).

The exotics are embedded in gray clay shale from which Harlton (1933) described a marine fauna. This shale, gray in fresh exposures, weathers olive gray. At some localities the Johns Valley

Figure 9. Pocket of exotics in the Johns Valley Formation along U. S. Highway 59 south of Stapp, Oklahoma. The mass of exotics is completely surrounded by shale.

Shale resembles the Stanley Shale in color. The shale is soft, plastic, easily eroded, and highly susceptible to downslope creep.

The exotic-bearing horizons are not ordinary basal conglomerates. Exotics are normally randomly distributed in the shale matrix and have no preferred orientation or size distribution. Pockets of exotics are present at the Hairpin Curve locality and near Stapp. Cline and Laudon (Cline and others, 1959, p. 39) described the pocket of boulders at Hairpin Curve (fig. 8) as a channel. They stated that the channel, 11.5 feet deep, is filled with boulders that are coarsely graded. At Stapp the pocket of exotics is completely surrounded by shale (fig. 9).

Rolled sandstone masses are also present in the exotic-bearing horizons.

Clasts of Arbuckle-like rocks.—Clasts of Arbuckle-like rocks, in place at several horizons in the Johns Valley Formation, have been referred to as Johns Valley boulders, erratics, and exotics. In this report these clasts have been consistently referred to as exotics

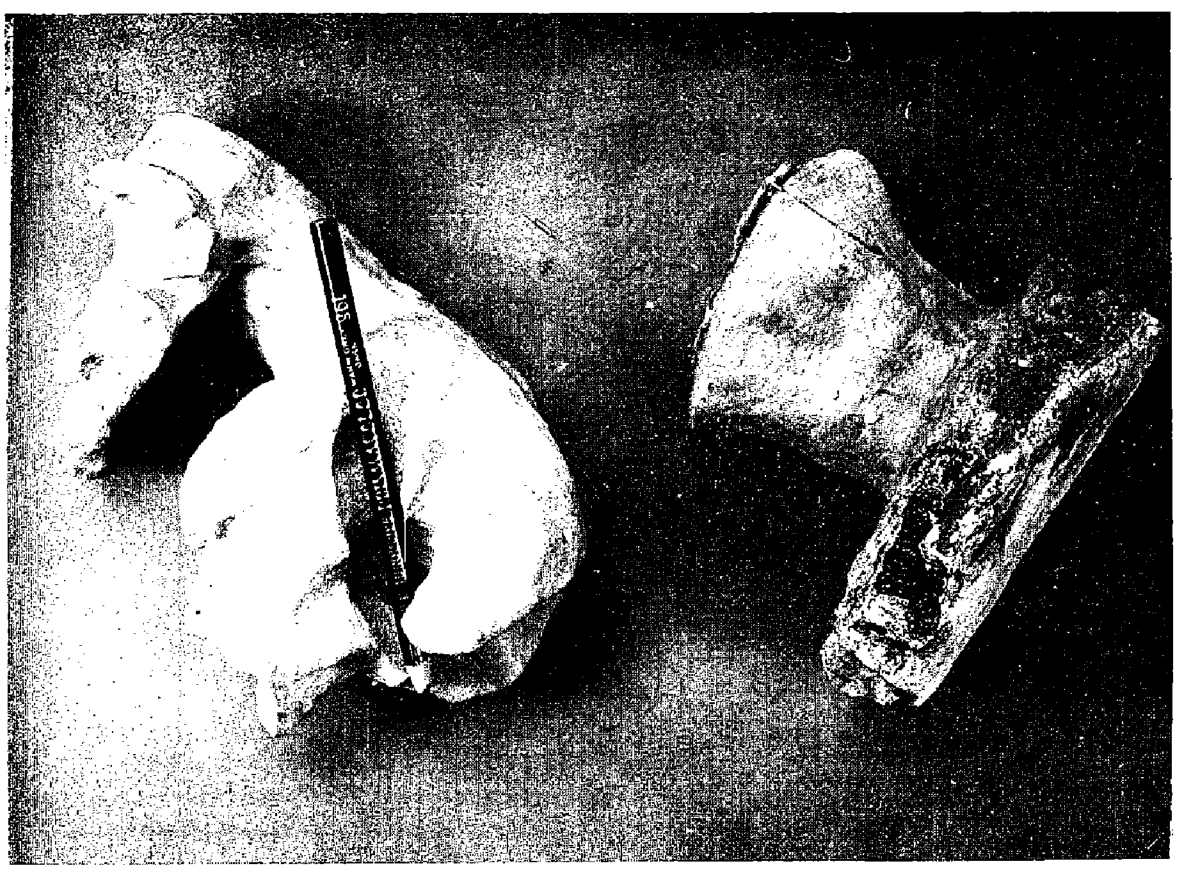

Figure 10. Irregularly pitted limestone exotics of Arbuckle-type rocks in the Johns Valley Formation.

or clasts of Arbuckle-like rocks. Reasons for the use of this terminology are given on page 85.

The Johns Valley exotics are not of tectonic origin, but are primary depositional features (Cline and Shelburne, 1959, p. 203). Exotics range in size from granules to boulders. The largest boulder observed in the report area, more than 100 feet long, is in Coopers Cove (center sec. 32, T. 4 N., R. 20 E.). Most of the exotics in the area range from 1 inch to 1 foot in diameter, although boulders 3 to 5 feet in diameter are common.

Ulrich (1927, p. 7) studied the exotics and concluded that there are "boulders of at least a dozen readily distinguishable kinds of limestone, many of silicified oölite, four or more kinds of sandstone, and others of arenaceous or calcareous shale." Some of the limestone exotics have been partially or completely silicified. Exotics ranging in age from Ordovician to Early Pennsylvanian, inclusive, are present in the Johns Valley Formation. Similar exotics have been described from the Chickasaw Creek (Harlton, 1938; Hendricks,

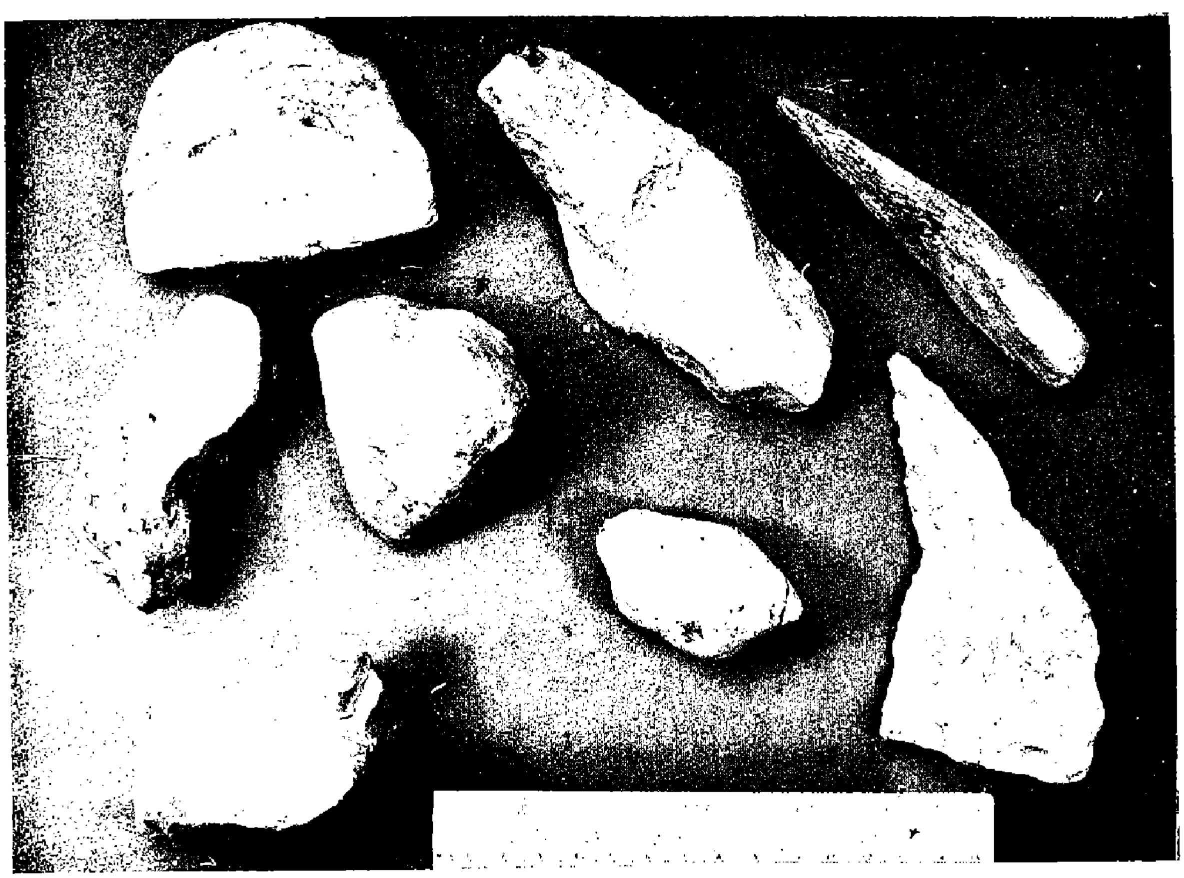

Figure 11. Pitted exotics of Ordovician limestone in the Johns Valley Formation.

1947), Markham Mill (Harlton, 1938), Wesley (Hendricks, 1947), Wapanucka (Wallis, 1915, p. 72), and the lower part of the Atoka (Hendricks and others, 1937, p. 22). Powers (1928, p. 1043) reported finding exotics stratigraphically above the Wapanucka Limestone at Lamberson Spur (NE¼ sec. 29, T. 4 N., R. 22 E.).

The best exposure of Wapanucka Limestone exotics observed by the writer is in a road cut along U. S. Highway 271, six miles northeast of Talihina, Oklahoma (NW¼ sec. 25, T. 4 N., R. 22 E.). Rounded exotics up to 1 foot in diameter were collected from this locality. At first glance, one block, 6 feet long, gives the appearance of being a bed in place in the enclosing Johns Valley Shale. A block of Wapanucka Limestone, 10 feet in diameter, is exposed 2 miles southeast of Bengal (NE¼ sec. 24, T. 4 N., R. 21 E.). Less than half a mile east is a bedlike block of Wapanucka Limestone approximately 7 feet long. All of the Wapanucka blocks described above are south of the Ti Valley fault. In general, the Wapanucka

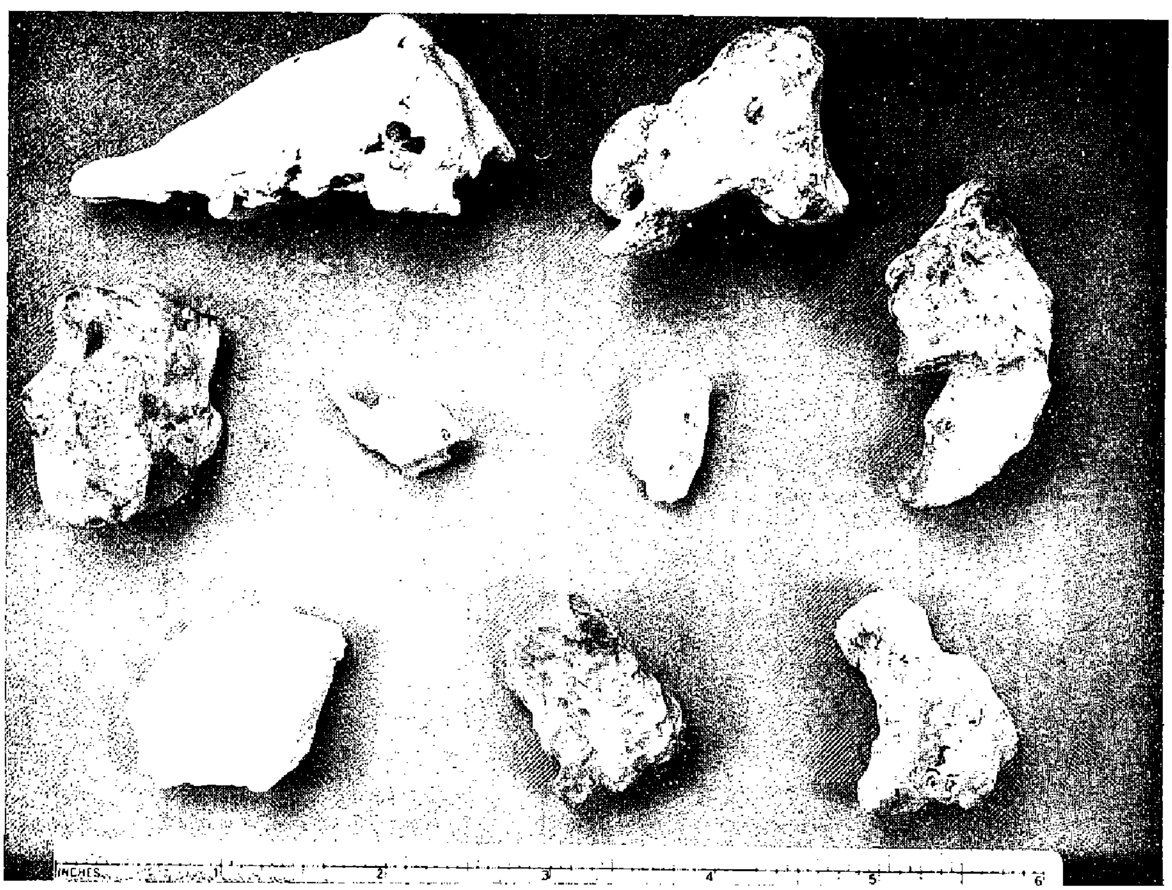

Figure 12. Pitted exotics of Ordovician limestone in the Johns Valley Formation.

exotics observed by the writer are more rounded and less pitted than are the exotics of other lithologic types.

Ulrich (1927) and Miser (1934) concluded that the black shale of Caney lithology in the lower part of the Johns Valley Formation is exotic. However, Cline (1956a) showed that this shale is in place. The same interpretation was given by Miser and Honess (1927, p. 12), although Miser later changed his opinion (1934, p. 984).

In the report area the more common exotics are gray sublithographic limestone of Bromide lithology. Ulrich (1927, p. 7) observed, however, that the predominant lithologic type varies from locality to locality.

Almost all of the exotics appear to be fresh and unweathered, as though they were washed before being deposited. Many have smooth, unstriated surfaces, whereas others are irregularly pitted (figs. 10, 11, 12). Sharp edges and points, as well as delicately

Figure 13. Negative print showing well-rounded pebbles of limestone in a matrix of quartz grains (black). The limestone pebbles are lithologically similar to the exotics in place in Johns Valley shale. Carbonate grains of sand size are also present in the matrix. Thin section was made from a 1½-inch bed in the Johns Valley Formation at the Hairpin Curve locality (NW¼ SE¼ sec. 3, T. 3 N., R. 19 E.). Quartz grains are black; limestones are gray or white. Scale is in 1-inch units. Specimen 144, ordinary light.

etched fossils, are present. The exotics are not associated with stream deposits but are embedded in marine shale. They have not been rounded or abraded during transportation. Delicate shapes of many exotics, presence of etched-out fossils, and absence of evidence of abrasion preclude long-distance transportation.

Striated exotics, described from several localities, were observed by the writer at the Compton Cut (NW¼ sec. 20, T. 4 N., R. 22 E.). These striations could easily have been formed during folding and faulting of the strata in which the exotics are embedded. A few striated rocks in areas of complex structure are not necessarily indicative of transportation by drifting ice.

That pitting occurred prior to deposition is clearly indicated. Depressions in the exotics are filled with the same shale that surrounds them. Had pitting taken place during exposure to the atmosphere after the Ouachita orogeny, one would expect to find a weathered crust around the exotics. This would also be expected

Figure 14. Exotic clasts of Arbuckle-like rocks in a sandstone matrix at Compton Cut (NW¼ NW¼ sec. 20, T. 4 N., R. 22 E.).

of clasts that had undergone subaerial weathering before deposition. Almost all of the exotics have fresh, unweathered surfaces.

Pitting appears to be the result of solution, which might have occurred along a beach or under water. However, if the exotics were shaped by exposure to the atmosphere, the weathering products were removed by wave or current action.

Pebbles and boulders transported in present-day streams are subjected to intense chemical weathering. Many of them have been transformed into spongy, whitish, clayey masses, some of which have less weathered limestone cores.

CONGLOMERATIC SANDSTONE.—Exotics are normally restricted to clay shales; however some thin beds of conglomeratic sandstone or sandy conglomerate are interbedded in the upper part of the Johns Valley. The clasts in these units are lithologically similar to those embedded in clay shale, but are smaller and well rounded

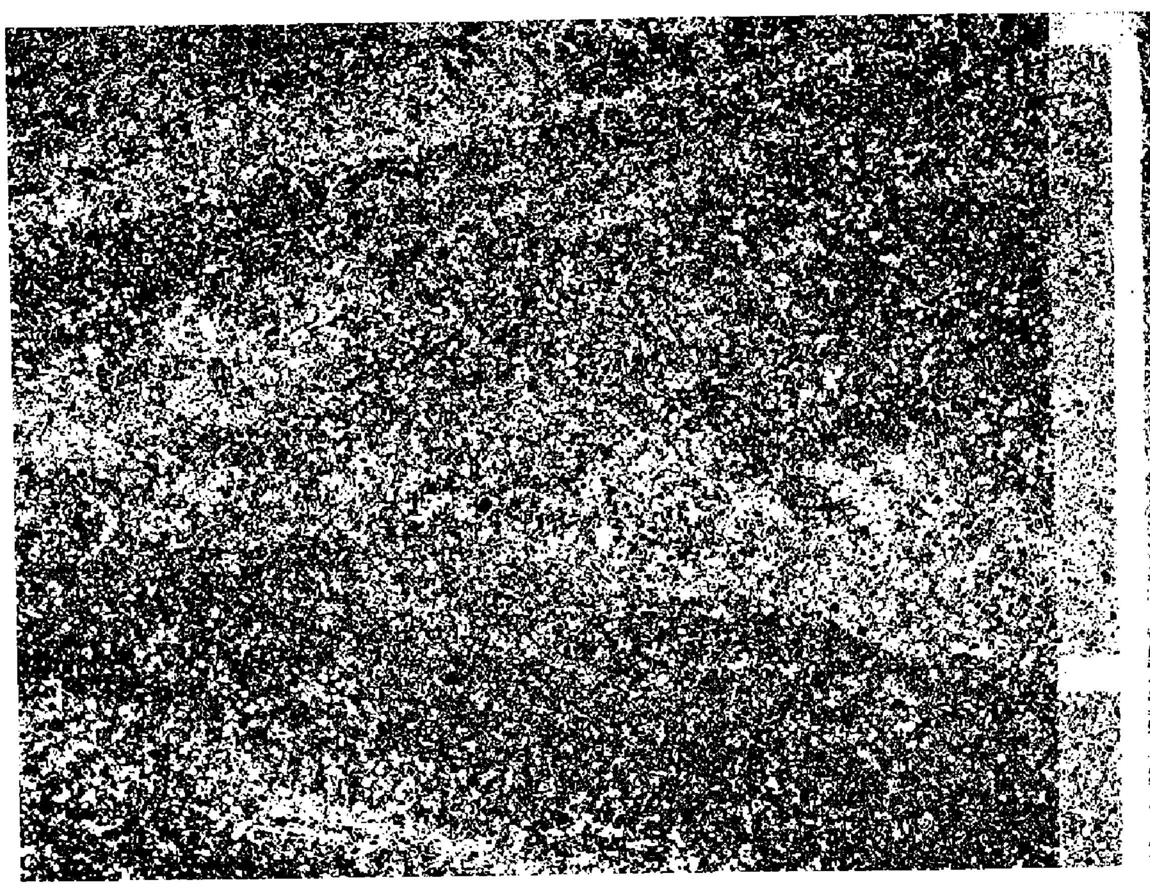

Figure 15. Negative print of a cross-laminated sandstone interbedded in the Johns Valley Formation. Quartz grains are black. Scale is in ½-inch units. Specimen 64, ordinary light.

(fig. 13). Some of these conglomeratic units, only 1 or 2 inches thick, have a red, sandy clay matrix.

Several large boulders of conglomeratic sandstone are present at Compton Cut (fig. 14). The component clasts are of Arbuckle-like rock types, but are embedded in sandstone. Boulders of this size and lithology were not observed elsewhere in the area.

INTERBEDDED SANDSTONE.—Green sandstone beds ranging in thickness from 1 to 6 inches are interbedded in the upper part of the Johns Valley Formation. Some of these are also present within one of the exotic-bearing horizons at Hairpin Curve. The sandstone, slightly argillaceous, consists predominantly of fine, angular quartz grains (fig. 15). These sandstone beds commonly have sole marks, small-scale cross-lamination, and marks made by organisms on tops of beds. Trails similar to those attributed to the gastropod *Paleobulla* are abundant (Cline, 1960, p. 92).

ROLLED SANDSTONE MASSES.—Rolled sandstone masses, spher-

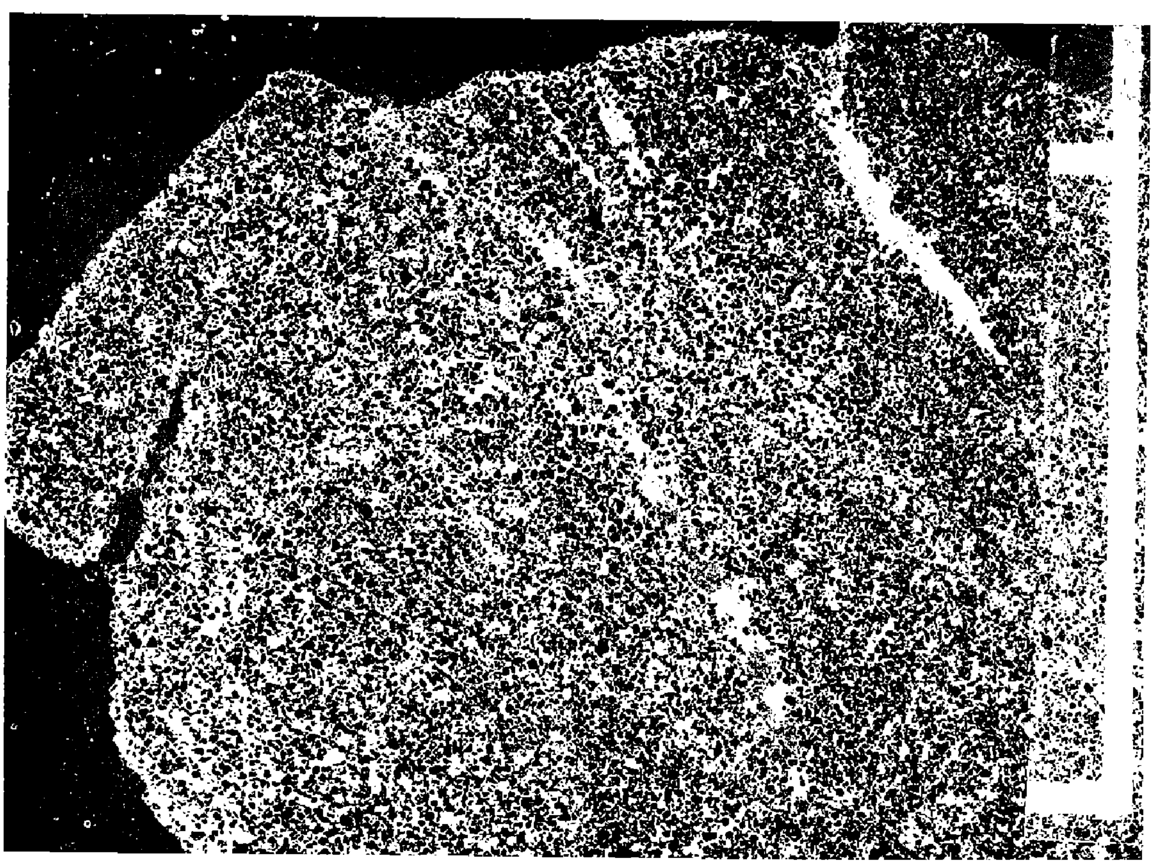

Figure 16. Negative print of a section from a rolled sandstone mass in the Johns Valley Formation. Quartz grains are black. The sandstone has been fractured. Scale is in ½-inch units. Specimen 8, ordinary light.

odial, oblate spheroidal, or irregular, are present within the exotic-bearing horizons of the Johns Valley Formation. Some of these masses, which range in diameter from less than 1 inch to more than 4 feet, have been intricately fractured and step faulted (figs. 16-20). Many of the faults do not extend through the masses but die out within. Some of the fractures are filled with the same material that composes the masses themselves (fig. 18), whereas other rolled sandstone masses have smooth surfaces and are unfaulted (Cline, 1960, fig. 34). Some of the masses are striated and appear to have been pinched off (fig. 20).

Many of the rolled sandstone masses have distorted sole marks on their outer surfaces. Some have exotic pebbles embedded at the surface. Similar pebbles were described by Crowell (1957, p. 999, 1002) as being contorted sandstone masses in the Pico Formation (Pliocene) and in the Chico Formation (Upper Cretaceous) in California. Harlton (1934, p. 1042) noted that exotic boulders at Compton Cut

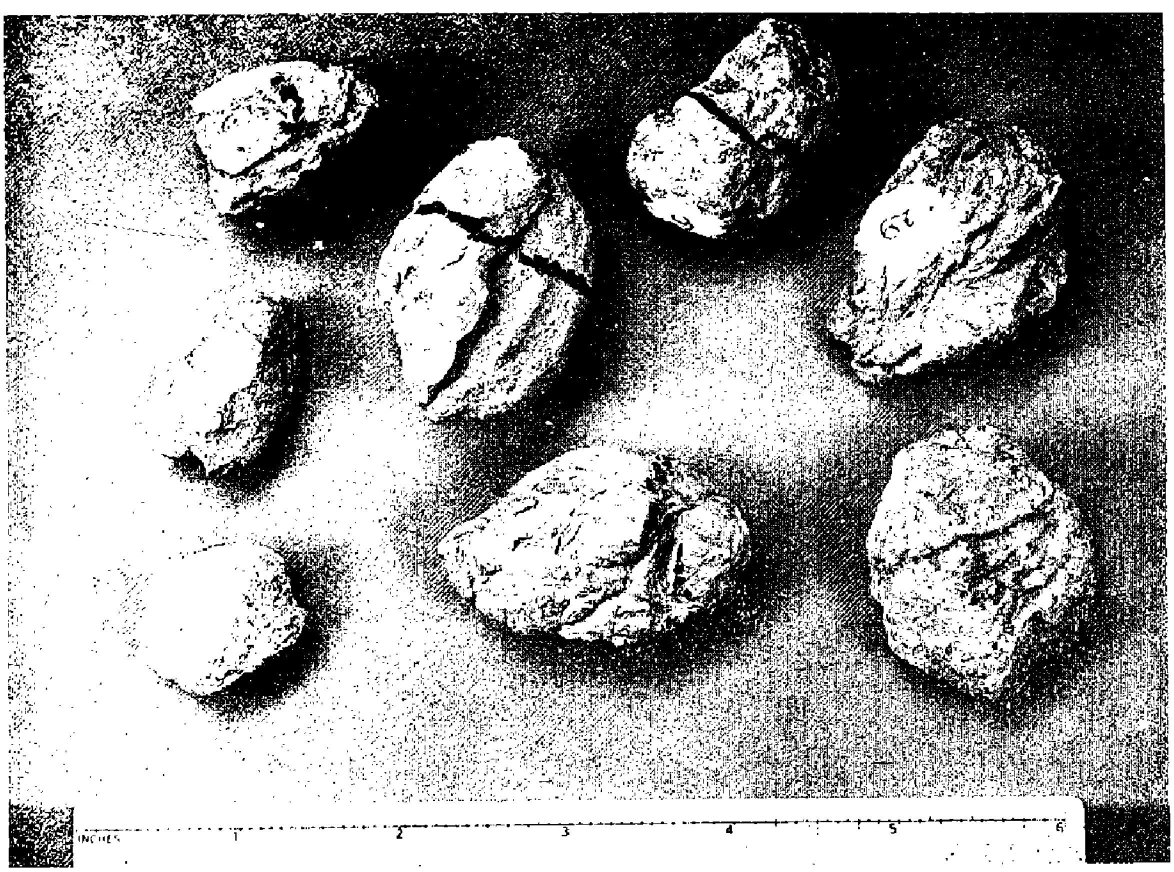

Figure 17. Small rolled sandstone masses of the Johns Valley Formation. Sections were cut along the black lines. Notice the intricate fracture pattern.

are normally surrounded by a fringe of pebbles, but none of these was observed by the writer.

Lithologically, the rolled sandstone masses are identical with the interbedded green sandstones described above. Compare the thin section of an interbedded sandstone (fig. 15) with the section of a rolled sandstone mass (fig. 16), both of which are from the Hairpin Curve locality.

In the report area, rolled sandstone masses were observed only in the exotic-bearing horizons. However, Shelburne (1960, p. 36) described them in the Johns Valley Formation in the Boktukola syncline, where no exotics are present. Goldstein and Hendricks (1962, p. 399) stated that rolled sandstone masses are also present in the Jackfork and the Atoka. Hill (1962, p. 37) stated that they are common in the Stanley Group.

Figure 18. Rolled sandstone mass of the Johns Valley Formation that has fractures filled with material of the same composition as the mass itself. Sole marks are faintly visible on the surface.

Figure 19. Cross-laminated sandstone bed in the Johns Valley Formation that has been bent into a hollow ball.

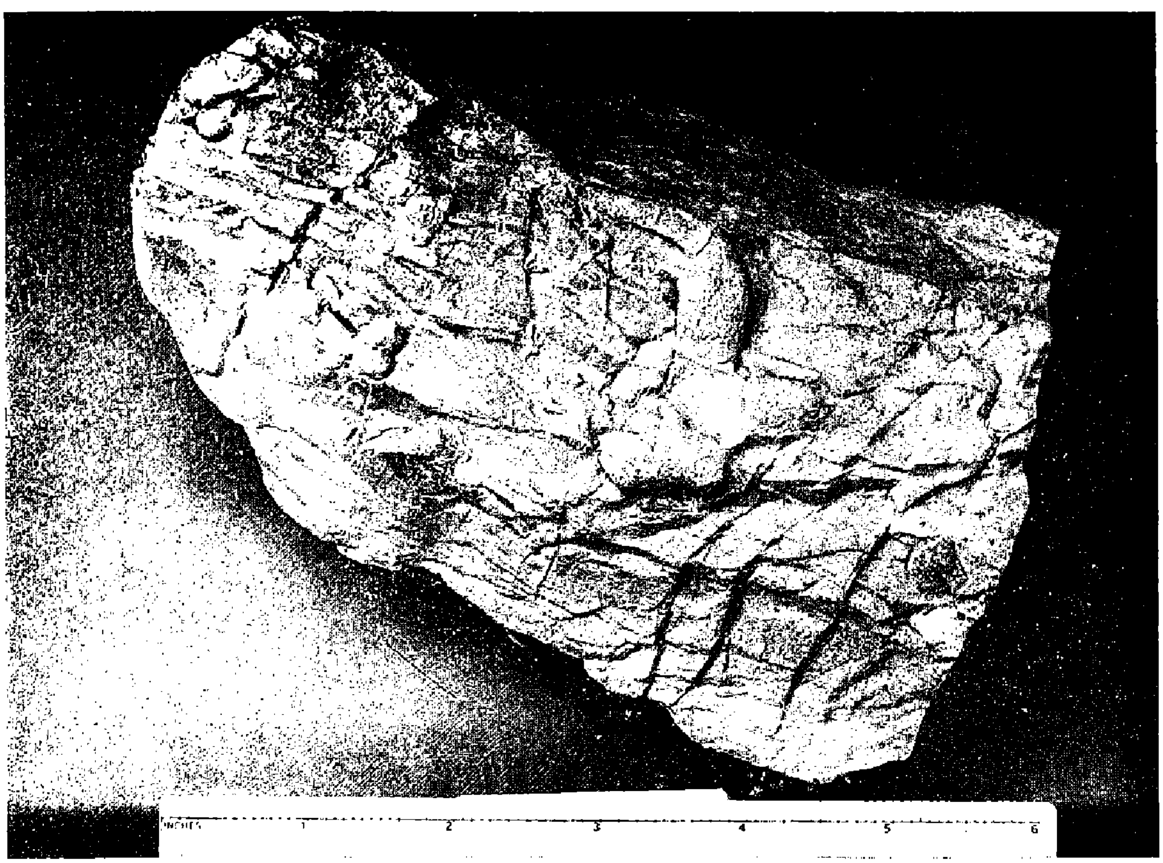

Figure 20. Fractured sandstone mass from the Johns Valley Formation that appears to have been pinched off. Sole marks are present on the surface.

One of the first to mention the presence of rolled sandstone masses in the Ouachita province was Bokman (1953, p. 159), who described them as "balled up zones in which bedding is locally wrapped around a core." Several specimens plainly show this relationship (fig. 19). Rolled sandstone masses were described in more detail by Cline (1956c, p. 58; 1960, p. 74), Cline and Shelburne (1959, p. 200), Shelburne (1960, p. 36), Seely (1962, p. 71), and Goldstein and Hendricks (1962, p. 399).

Stratigraphic units.—Although the Johns Valley Formation has not been formally subdivided, the following distinct stratigraphic units are present in the report area: (1) a lower exotic-bearing member, possibly transitional with the underlying Game Refuge Formation, (2) a black shale member considered to be a tongue of the Caney Shale, and (3) an upper member consisting of gray-green shale, green sandstone, and exotic-bearing horizons.

Pennsylvanian System

WAPANUCKA LIMESTONE

Nomenclature, classification, and correlation.—The Wapanucka Limestone, named by Taff (1901, p. 3) for exposures in Johnston County, Oklahoma, was subdivided by Harlton (1938, p. 902). It is Morrowan in age. The Chickachoc Chert is the lateral equivalent of the Wapanucka Limestone. Wapanucka equivalents are also present in the upper part of the Johns Valley Formation south of the Ti Valley fault.

Distribution, thickness, and boundaries.—The Wapanucka ranges from 270 to 720 feet in thickness in the block southeast of the Choctaw fault (Hendricks, 1947). In the next block south, the block southeast of the Katy Club fault, it is represented by the Chickachoc Chert, the thickness of which is not known. No Chickachoc or Wapanucka equivalent is present in the block southeast of the Pine Mountain fault. However, south of the Ti Valley fault thin, discontinuous lenses of Wapanucka-like limestone, considered by the writer to be Wapanucka equivalents, are interbedded with the shale and sandstone of the upper part of the Johns Valley Formation. Boundaries of the Wapanucka equivalents could not be determined. Localities at which interbedded limestone lenses are present are listed in appendix C.

Honess (1923, p. 17) correlated sandstones containing a mold fauna in northern McCurtain County with the Wapanucka Limestone of the Arbuckle Mountains. Cline and Shelburne (1959, p. 204), who noted that sandstone containing a Morrowan mold fauna is present at the Hairpin Curve locality and in the Boktukola syncline, concluded that the upper part of the Johns Valley is equivalent to some part of the Wapanucka Formation. Cline (oral communication, March 1959) reported that a thin bed of limestone, which he believed to be a tongue of the Wapanucka, is present in the upper part of the Johns Valley near the center of the west line of the SE¼ sec. 1, T. 1 S., R. 12 E. This locality, 3¾ miles northeast of Stringtown, is about 1 mile southeast of the Ti Valley fault. Goldstein and Hendricks (1962, p. 399) described lenses of impure

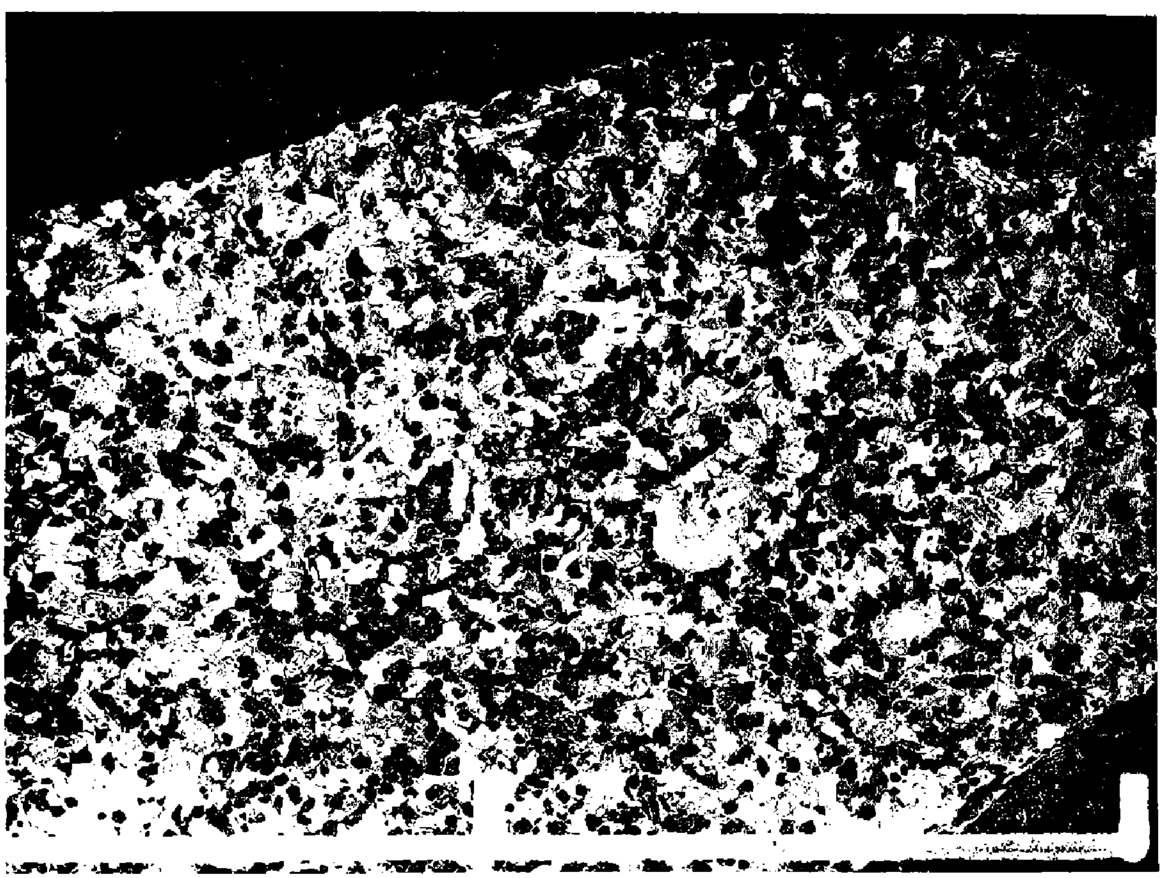

Figure 21. Negative print of a thin section from the Wapanucka Limestone along State Highway 2 south of Wilburton. Quartz (black) and fossil fragments (gray) are predominant. Many of the quartz grains are well rounded. Scale is in ½-inch units. Specimen 19, ordinary light.

limestone in the upper part of the Johns Valley. Hendricks (oral communication, April 1962) reported seeing limestone lenses interbedded in the Johns Valley Formation along Boggs Creek (SE¼ sec. 11, T. 3 N., R. 19 E.). Although limestone blocks are still present at this locality, recent slumping has obscured the exact relationship of the limestone to the enclosing shale. Morris (1962, p. 28), who described similar beds of limestone in the Johns Valley in the Y City and Acorn quadrangles, Arkansas, interpreted these beds as indigenous in the Johns Valley. Interbedded limestone lenses up to 4 inches thick in the Johns Valley Formation occur at several localities between the Ti Valley and Windingstair faults.

Lithologic characteristics.—The Wapanucka-type limestone lenses and exotics in the area studied consist almost entirely of rounded fossil fragments, many of which are encrusted with carbon-

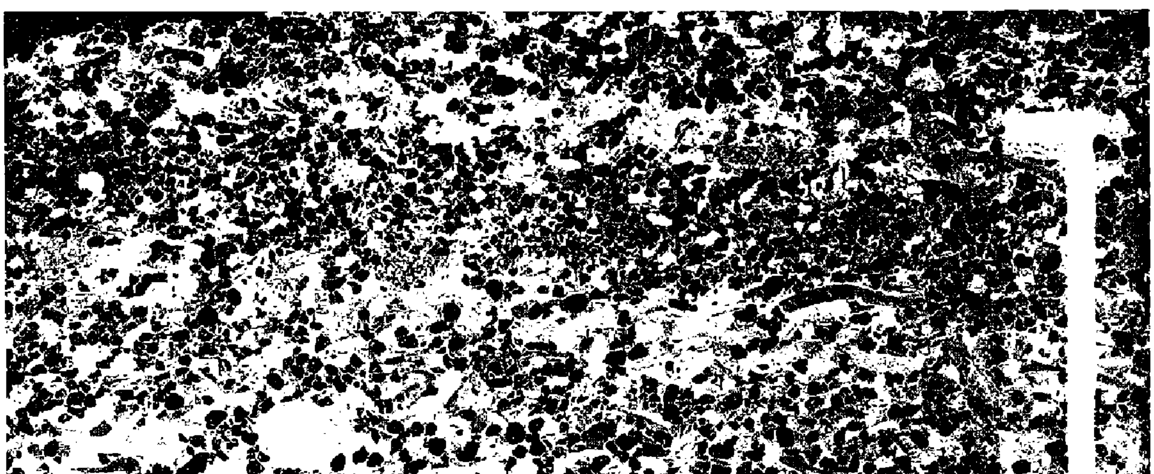

Figure 22. Negative print of a thin section of a limestone interbedded in the Johns Valley Formation southeast of Bengal (NE¼ sec. 24, T. 4 N., R. 21 E.). Quartz (black) and fossil fragments (gray) are predominant. Most of the quartz grains are well rounded. Rounded pebbles of limestone are also present. Scale is in ½-inch units. Specimen 32, ordinary light.

ate. Fragments of crinoid columnals, bryozoans, and fusulinids are common. These limestones, apparently shallow-water deposits, are calcarenites. Because most of the lenses and exotics have been partially oxidized, weathered surfaces are various shades of brown; however, unweathered surfaces are dark gray or black. Other limestone exotics in the Johns Valley Formation are not so deeply weathered.

The Wapanucka Limestone at the Limestone Gap locality near Chockie and at the Bandy Creek locality along State Highway 2 south of Wilburton was sampled and thin-sectioned. Lithologically and petrographically these limestones are almost identical with the lenses and exotics in the mapped area, south of the Ti Valley fault (figs. 21-24). Southeast of Bengal the limestone is quite sandy and at one locality is cross-laminated. A similar eastward lithologic change in the Wapanucka Limestone occurs north of the Ti Valley fault (Harlton, 1938).

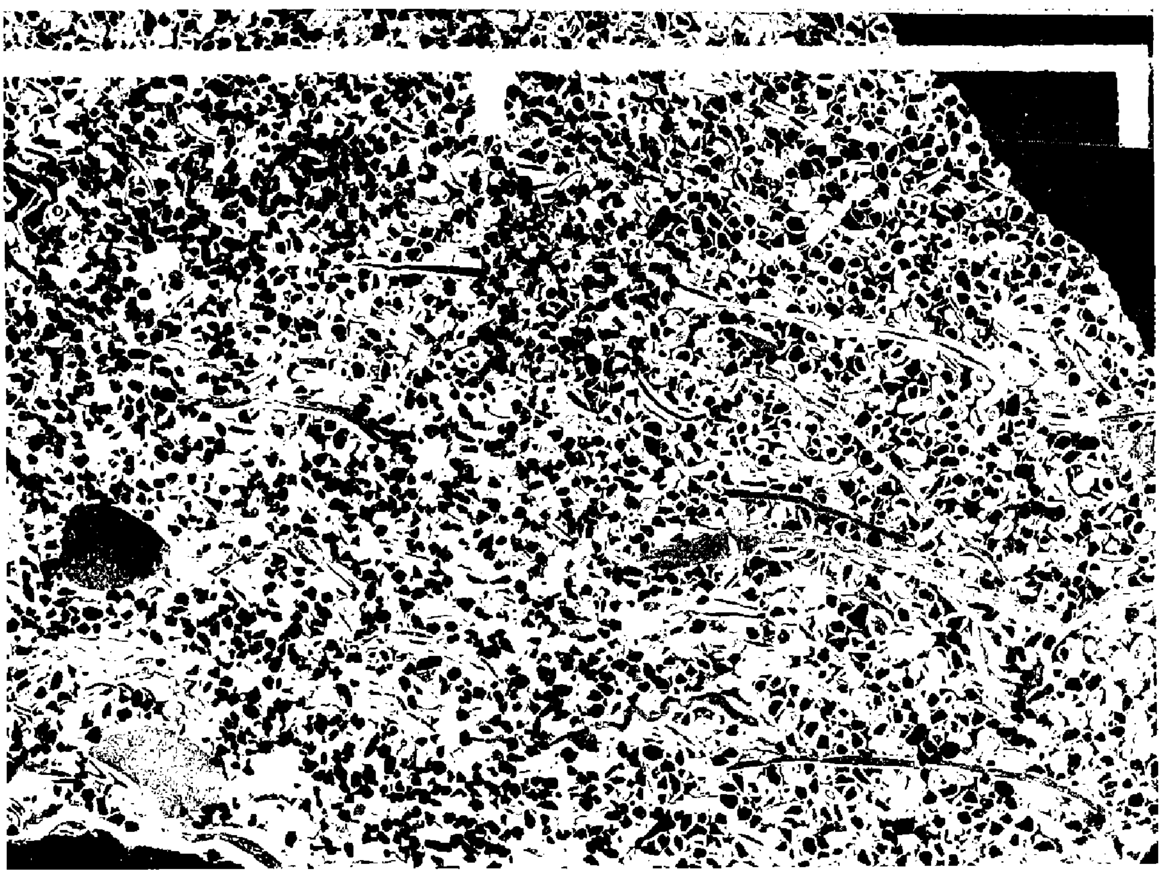

Figure 23. Negative print of a thin section of Wapanucka Limestone exposed along State Highway 2 south of Wilburton. Many of the quartz grains (black) are encrusted with carbonate. Scale is in ½-inch units. Specimen 20, ordinary light.

ATOKA FORMATION

Nomenclature, classification, and correlation.—The Atoka Formation was named by Taff and Adams (1900, p. 273) for exposures near Atoka in Atoka County, Oklahoma. No type section was designated. In this report the Atoka Formation is considered to range in age from Morrowan to Atokan (Cline, 1960, p. 85). The formation has not been subdivided. The only persistent unit recognized in the area studied was a spicular black siliceous shale 100 to 200 feet above the base of the formation (Hendricks, 1947). The Atoka Formation, the youngest stratigraphic unit in the Ouachita province, is present throughout the frontal and central belts of the Ouachita province and in the eastern part of the Arbuckle Mountains. In the western Arbuckles it has been removed by erosion.

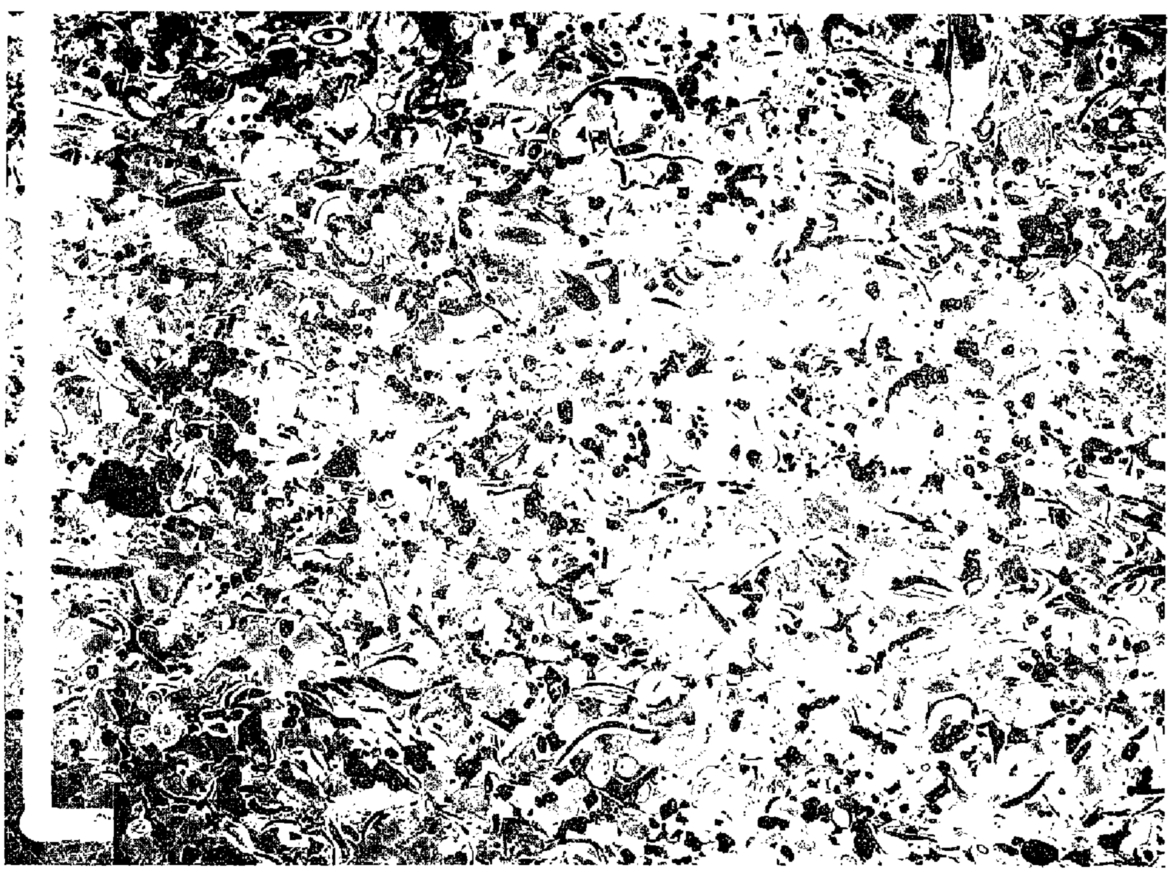

Figure 24. Negative print of a thin section of limestone in the upper part of the Johns Valley Formation southeast of Bengal (SW¼ sec. 18, T. 4 N., R. 22 E.). Fossil fragments (gray) and quartz grains (black) are abundant. Some of the quartz grains have carbonate "crusts." Scale is in ½-inch units. Specimen 38, ordinary light.

Distribution, thickness, and boundaries.—The complete thickness of the Atoka Formation in the Ouachita province is not known. The maximum reported thickness of the Atoka is in Arkansas, where it is 19,000 feet thick (Reinemund and Danilchik, 1957). Shelburne (1960, p. 41) stated that the maximum thickness of the Atoka in the Boktukola syncline is 6,800 feet. In the report area the maximum thickness is approximately 6,000 to 7,000 feet along Winding Stair Mountain northeast of Talihina Lake. In most of the area it is 1,000 to 3,000 feet thick.

The Atoka Formation is conformable upon the Wapanucka Limestone and the Chickachoc Chert north of the Pine Mountain fault. It is conformable upon the Johns Valley in the central belt. Goldstein and Hendricks (1962, p. 410) showed it to be unconformable upon the Caney Shale in the block southeast of the Pine Mountain fault. Kramer (1933, p. 593) stated that the Atoka overlies

Figure 25. Current ripples on the top of an Atoka sandstone bed (NE¼ SW¼ sec. 27, T. 4 N., R. 21 E.). Currents that produced these ripples flowed from left to right (west to east). Most of the current features observed indicated that current flow was from east to west.

the Woodford at two localities in the Pine Mountain-Ti Valley block. At the Bengal locality the Woodford is interpreted as in fault contact with the Atoka.

The top of the Atoka Formation is eroded in the Ouachita province.

Lithologic characteristics.—The Atoka Formation consists of alternating gray shale and sandstone. At certain horizons, such as near the base of the formation in the area studied, sandstone predominates. Many of the sandstone units are massive and are as much as 10 or more feet thick. A large proportion of the formation consists of alternating 1- to 3- inch beds of sandstone and somewhat thicker beds of black or gray shale.

The sandstone, made up almost entirely of quartz grains, is micaceous. Plant fragments are abundant near the top of the sandstone units at many places. Thin black siliceous shales or siltstones are common (Shelburne, 1960, p. 41). Also common are zones containing a mold fauna.

Many of the sandstone units have well-preserved convolute lamination, normally present in the top half of the bed. A crude type of graded bedding is common. Clay galls are common at both the tops and bottoms of beds. Well-preserved current ripples are found in places (fig. 25).

In the report area the basal Atoka sandstone, much more resistant to erosion than the underlying Johns Valley Shale, forms a series of essentially parallel ridges.

SEDIMENTARY STRUCTURES

Several types of sedimentary structures are present in sandstone units in the report area. Specimens were collected, thin-sectioned, and examined microscopically in order to determine how the sedimentary structures were formed. All of the sedimentary structures discussed are present in strata of the Johns Valley or Atoka Formations. Sedimentary structures from the Stanley and Jackfork Groups were not described. The sedimentary structures discussed are sole marks, cross-lamination, convolute lamination, graded bedding, and marks on bed tops.

Sole Marks

Features on the bottoms of sandstone beds, also known as sole marks, are common in the Johns Valley and Atoka Formations. They were also observed at many horizons in the Stanley Group. Rich (1950, p. 724) noted that they are abundant in the Jackfork of the Ouachita region. Bokman (1953, p. 158) described these features, noting that they are on the bottoms of beds, but attached no other significance to them. Johnson (1954) made use of them in determining whether beds are overturned.

Sole marks are also valuable as paleocurrent-direction indicators. Flow markings were described by Rich (1950, p. 725) as having been cut by currents that flowed along the bottom. Rich noted that at Aberystwyth, Wales, the "flow markings" are aligned in the same direction even though beds may lie 100 or more feet apart stratigraphically.

Paleocurrent-direction studies were made in Oklahoma by Cline and Shelburne (1959), Cline (1960), Shelburne (1960), and Seely (1962). Similar studies were made in Arkansas by Reinemund and Danilchik (1957). As stated by Cline and Shelburne (p. 206) and substantiated by subsequent studies, currents flowed approximately W 12° S down the axis of the Ouachita geosyncline during deposition of the Stanley-Atoka sequence. No systematic or detailed paleocurrent study was made by the writer, but most of the sole marks examined are considered to have been formed by southwestward-flowing currents, as was formerly observed by Cline and Shelburne. However, at one locality near C sec. 27, T. 4 N., R. 21 E., ripple marks on the top of a sandstone bed in the Atoka Formation indicate that currents flowed from west to east (fig. 25).

No attempt has been made in this report to classify the many types of sole marks. Those on the bottoms of sandstone beds are casts made by infilling of previously formed depressions in the underlying mud (Kuenen, 1953). Some of the depressions, or molds, were presumably made by currents. Others are due to the activity of burrowing organisms, to dragging or sliding of objects across the mud bottom, or to loading phenomena. In this study the primary concern is with depressions that were scoured by currents and filled with sand. Both scouring and filling have been attributed to turbidity currents (Kuenen, 1953). No agreement has been reached as to whether the depressions in the underlying mud were scoured and filled by the same turbidity current (Kuenen, 1957)

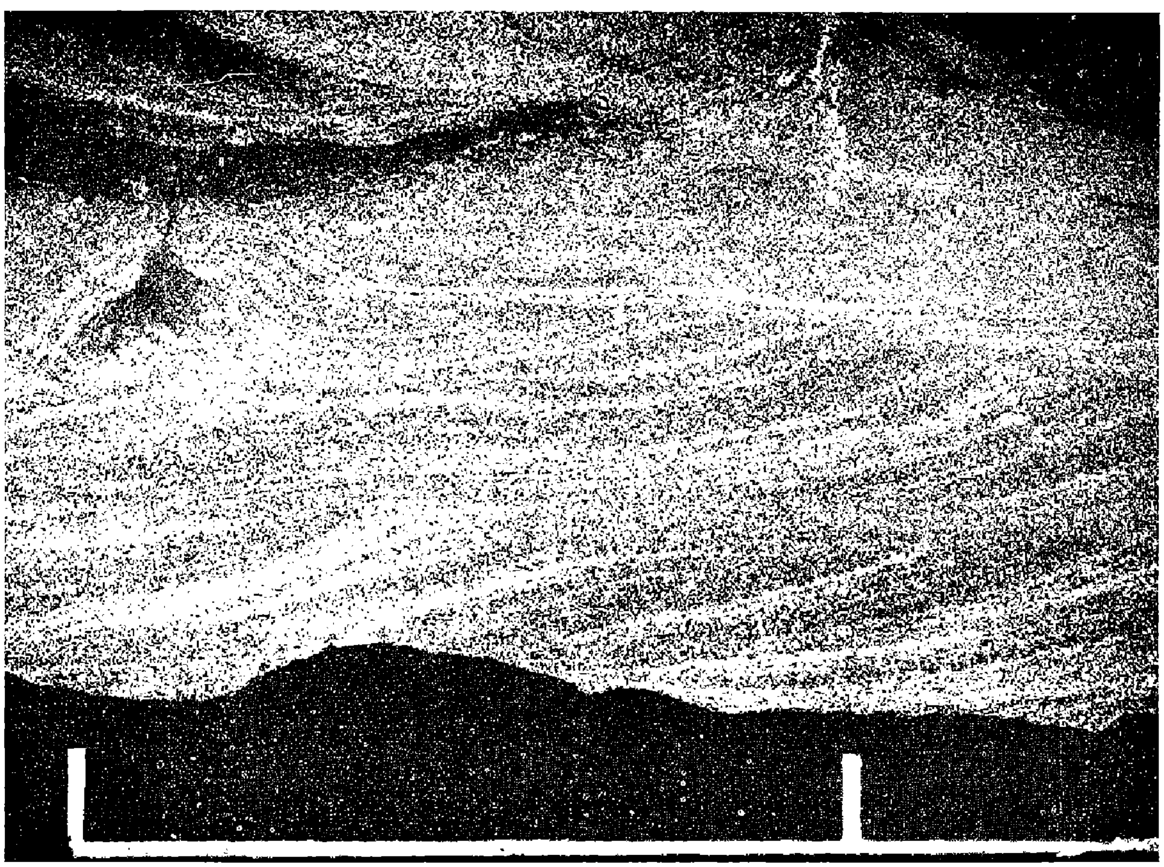

Figure 26. Negative print showing small-scale cross-lamination in a sandstone bed of the Atoka Formation. Quartz grains are black. Scale is in 1-inch units. Specimen 75, ordinary light.

or were scoured by one turbidity current, left open, and filled by a later current (Crowell, 1955, 1958).

Sole marks do not necessarily represent deep-water deposition (Glaessner, 1958) or deposition in a geosyncline. They have been reported in rocks of the Eastern Interior basin, in the Devonian of New York, in the German Muschelkalk, and in other shallow-water deposits.

Cross-Lamination

Although large-scale cross-lamination was not observed in the area of investigation, small-scale cross-lamination is common in sandstones of the Johns Valley and Atoka Formations. It is the rule rather than the exception in beds that have sole marks. Cross-lamination is not obvious megascopically, but shows up clearly in thin section (figs. 15, 26). At some horizons in the Moyers Forma-

Figure 27. Convolute lamination in the upper portion of a sandstone bed of the Atoka Formation.

tion cross-lamination is so well developed that the lines of intersection of individual laminae with the upper plane of stratification can be used to identify bed tops. Current sense can be measured by reading the bearing of a line at right angles to the surface trace of the laminae.

Comparison of the bearing of a line at right angles to the surface trace of the laminae with the bearing of oriented sole marks showed that, in many cases, the direction of flow of the currents that formed the initial depressions and of the currents that filled them were not parallel. In some instances these directions are almost at right angles. Kuenen (1957, p. 242) stated that "the current lamination filling many flute casts and other bottom cavities is commonly oriented more or less obliquely to the trend of the markings. This may be due to vortices or to a slight change in current direction."

Glaessner (1958, p. 6) stated that "flute casts showing cross-lamination in section were formed by gradual infilling of precut

Figure 28. "Ouachita graded bedding" in a sandstone unit of the Atoka Formation. Pencil is at the top of the unit.

scours. The turbulent conditions in which the hollows were formed were followed by nonturbulent conditions in which laminated sediments were deposited."

Convolute Lamination

Convolute lamination (fig. 27), well-preserved in many sandstone beds of the Atoka Formation, occurs in the top half of the stratification unit. The writer has reached no conclusions about the origin and significance of convolute lamination. Apparently it can be formed in several ways. Ten Haaf (1956) concluded that convolute lamination is formed hydraulically during deposition by turbidity currents. Williams (1960) attributed it to spontaneous liquefaction and lateral flow as a result of loading. The examples observed by the writer do not appear to have been formed by current drag as described by ten Haaf. Some sandstone beds with sole marks and undistorted cross-lamination also have convolute cross-lamination, indicating that convolute lamination did not form by plastic flow unless only the top part of the bed flowed. It is possible that the convolute lamination observed was formed by spontaneous liquefaction and lateral laminar viscous flow as grains attain a more stable packing arrangement.

Graded Bedding

Graded bedding of one type is well developed throughout the Stanley-Atoka sequence. Most of the sandstone beds have sharp contacts with the underlying shale, are massive at the base, become progressively more shaly and more thinly bedded upward, and are transitional with the overlying shale (fig. 28). This type of grading, which can be used to indicate bed top, was described by Cline (1960, p. 88). It is often referred to as "Ouachita graded bedding" or simply "Ouachita grading." The significance of this type of grading is not known.

Another type of graded bedding is also present. Commonly coarser sand grains were the first to be deposited in the depressions in the underlying shale. However, the coarser grains did not fill the depressions and there is no progressive upward decrease in grain size.

"True" graded bedding was observed only in the pebble conglomerates in the Chickasaw Creek Formation (fig. 6).

Turbidity-current deposits characteristically have graded bedding (Kuenen and Carozzi, 1953), although graded bedding can be formed in other ways (Dott and Howard, 1962, p. 120). As Dott and Howard concluded (p. 114), unless repeated graded bedding occurs, one cannot be sure that he is dealing with turbidity-current deposits.

Whether the Stanley-Atoka sequence consists of turbidity-current deposits depends upon the mechanism by which "Ouachita graded bedding" was formed. If it is related to "true" graded bedding, the Stanley-Atoka sequence might have been deposited by turbidity currents. Most of the sandstones are poorly sorted, but the grains are distributed randomly throughout the bed.

Marks on Tops of Sandstone Beds

Organisms have made burrows or trails on many of the sandstone beds of the Johns Valley Formation. These features were described and photographed by Cline (1960, p. 92). One type of trail, attributed to the gastropod *Paleobulla,* is common in the Johns Valley sandstones.

Many sandstone beds with top marks also have bottom marks and cross-lamination. However, convolute lamination was not observed.

Rolled Sandstone Masses

Although reported to be present in other formations as well, rolled sandstone masses have been found only in the exotic-bearing horizons of the Johns Valley Formation in the report area. These masses are described in detail in the section on lithologic characteristics of the Johns Valley Formation. Their origin is discussed later in this report in the section on clues to emplacement of exotics.

STRUCTURE

Structural Setting

Structures in the Ouachita province were formed during the Ouachita orogeny, which began after deposition of the Atoka Formation. Several major reverse faults, essentially parallel to the structural strike, were formed. The Ti Valley fault, one of the more important reverse faults, separates the frontal belt and the central belt of the Ouachita structural province (fig. 1). Strata deposited in the Ouachita geosyncline were moved northward along the Ti Valley fault and are in contact with their shelf equivalents. The area described in this report is bounded on the north by the Ti Valley fault.

Post-Woodford strata in the Ouachita province consist of alternating competent and incompetent units. The Arkansas Novaculite, Jackfork Group, and the lower part of the Atoka Formation are competent, whereas the Stanley Group, the Johns Valley, Caney, and Springer Formations, and the upper part of the Atoka Formation are largely incompetent. Faulting has been concentrated in the incompetent units, several of which have served as glide planes (Hendricks, 1959, p. 53). The structure of the frontal belt of the Ouachita province is similar to the structure of the Valley and Ridge province in the southern Appalachians.

Age of Ouachita Orogeny

Regional uplift of the Ouachita province began in late Atokan time (Hendricks, 1937; Hendricks and Parks, 1937, 1950) and continued until Late Pennsylvanian or Early Permian time. Clastics eroded from this uplift were deposited in a Desmoinesian basin north of the Ouachita province. The Boggy Shale (Clawson, 1928) and possibly beds as young as Thurman (Hendricks, 1937) were folded during the deformation, indicating that structural movement in the Ouachita province continued until at least as late as middle Desmoinesian time (Hendricks, 1959, p. 54). Cheney (1929), Melton (1930), and van der Gracht (1931) concluded that pulses of the Ouachita orogeny occurred in Middle and Late Pennsylvanian time and as late as Early Permian time. Hendricks (1959, p. 54) gave a more detailed discussion of the age of the Ouachita orogeny.

Major Structures

Five major structures were mapped in the report area: Buffalo Mountain syncline, Windingstair fault, Peachland anticline, Ti Valley fault, and Bengal inlier. With the exception of the Buffalo Mountain syncline and the Bengal inlier, portions of these structures were also mapped by Hendricks and others (1947) in the area adjoining on the west (fig. 29). The Jackfork Creek fault, also a major structure, is present between the Buffalo Mountain syncline and the Potato Hills. The Coopers Cove and Hairpin Curve localities, along the axis of the Peachland anticline, are discussed because they are well known, somewhat controversial, and quite interesting. Cross faults, although not considered to be major structures, are also discussed.

As is indicated on the geologic map (pl. I), the area has been complexly folded and faulted. Most of the faults are high-angle reverse faults that formed as a result of compression and rupture of major anticlines (pl. I). Although numerous, these faults are considered to be minor structures along which predominant vertical movement has occurred.

The following structures are discussed, in order of location, from south to north: (1) Jackfork Creek fault, (2) Buffalo Mountain syncline, (3) Windingstair fault, (4) Peachland anticline, (5) Ti Valley fault, (6) Bengal inlier, and (7) cross faults.

Jackfork Creek Fault

Between the southern edge of the Potato Hills and the Kiamichi Range, the Stanley Group is approximately 10,000 feet thick (Laudon, 1959). However, along the north side of the Potato Hills, outcrops of Arkansas Novaculite are only half a mile south of those of the basal Chickasaw Creek Formation. Post-Arkansas Novaculite pre-Chickasaw Creek strata, which dip gently northward, are much less than half a mile thick along the north edge of the Potato Hills, whereas they are more than 9,500 feet thick between the Potato Hills and the Kiamichi Range. The Stanley Group has either thinned

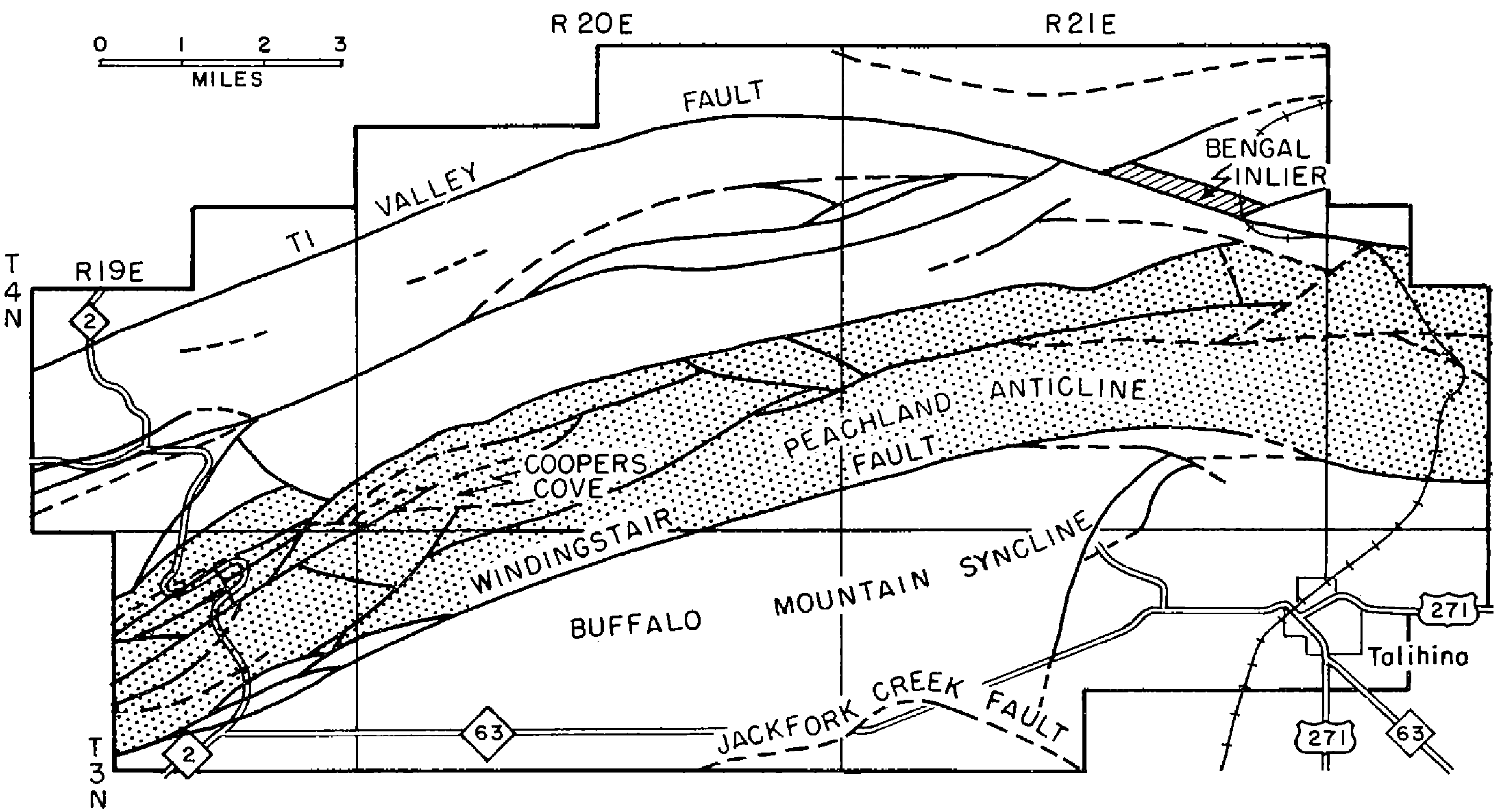

Figure 29. Index structure map.

rapidly to the north or is in fault contact with the pre-Stanley strata of the Potato Hills.

The Jackfork Creek fault, between the Buffalo Mountain syncline and the Potato Hills, was mapped by Miller (1955), Roe (1955), and Seely (1955). Its trace is poorly exposed 1 to 2 miles south of the southern boundary of the area mapped by the writer. Within the mapped area, direct evidence of faulting was found at two localities, both of which are in road cuts along State Highway 63 (NW¼ NW¼ sec. 17, T. 3 N., R. 21 E.; SW¼ NW¼ sec. 18, T. 3 N., R. 21 E.). At these localities the soft shale and thin sandstone beds of the Stanley Group are contorted and almost vertical. Less than 1/4 mile to the north undeformed strata of the Stanley Group dip approximately 25°N.

The eastern end of the Buffalo Mountain syncline is truncated by a major fault. The relationship of the Jackfork Creek fault to this fault is not known.

BUFFALO MOUNTAIN SYNCLINE

The Buffalo Mountain syncline, approximately 10 miles long and 3 miles wide, is capped by resistant sandstone of the lower part of the Jackfork Group. The top of Buffalo Mountain is approximately 1,400 feet above the floor of the valley below. This syncline, the axis of which plunges gently eastward, is bounded on the north by the Windingstair fault. A major fault, the Jackfork Creek fault, is between the Buffalo Mountain syncline and the Potato Hills, which adjoin on the south. At the east end of Buffalo Mountain is a major fault, along which the upper part of the Tenmile Creek and the lower part of the Moyers Formations have moved upward and northward over the Moyers, Chickasaw Creek, and Wildhorse Mountain Formations (pl. I). A minor fault, which Seely (1955) named the Sanatorium fault, is also present in this vicinity. The major fault, which was not mapped by Seely and has not been named in this report, is probably a tear fault connected with the Jackfork Creek fault on the southwest and with the Windingstair fault on the northeast.

Exposed on the flanks of Buffalo Mountain syncline are Moyers, Chickasaw Creek, and lower Wildhorse Mountain Formations, which dip 25° to 35° toward the synclinal axis.

WINDINGSTAIR FAULT

The Windingstair fault, one of the major faults of the central belt of the Ouachita province, can be traced from near Atoka, Oklahoma, eastward across the area studied, and on into Arkansas. In the area of this investigation, strata of the Stanley Group have been moved northward over the Atoka Formation along the Windingstair fault. Thin slices, or slivers, of younger strata, present in the fault zone, were dragged upward along the reverse fault from the underlying block. Fault slices are exposed at two localities (secs. 11, 12, 14, T. 3 N., R. 19 E.; NW¼ NW¼ sec. 35, T. 4 N., R. 21 E.). At the latter locality, a cut at the west end of the dam at the south end of Talihina Lake, a sliver of the Johns Valley is between strata of the Chickasaw Creek Formation and undifferentiated Stanley Group. Wallis (1915, p. 39), who studied the Wapanucka Limestone in the frontal belt of the Ouachita province in Oklahoma, stated that "the structure consists of a series of narrow, closed folds which have been overturned toward the northwest and broken by strike thrust faults. . . . . .The faulting probably occurred in the axis of the anticlines."

The trace of the Windingstair fault is straight, indicating that the fault dips steeply. At no place was the fault surface actually observed.

PEACHLAND ANTICLINE

Peachland anticline is the name proposed here for the structure that adjoins the Windingstair fault on the north. The area of this complexly faulted anticline is indicated by shading on the index structure map (fig. 29). South Peachland Creek, after which the Peachland anticline was named, flows along the axis of the anticline for approximately 4 miles (secs. 25, 26, T. 4 N., R. 20 E., and secs. 17, 19, 20, T. 4 N., R. 21 E.). The Peachland anticline can be traced southwestward into the area mapped by Hendricks and Averitt (Hendricks and others, 1947).

Because of continued compression, the Peachland anticline was overturned to the north and ultimately ruptured. A series of high-angle reverse faults, essentially parallel to the anticlinal axis, were formed.

Soft shales of the Johns Valley Formation are exposed along the axis of the anticline, whereas resistant sandstones of the lower part of the Atoka Formation form the limbs. Strata in the north limb dip steeply to the north, are vertical, or are overturned to the north. Strata in the south dip 50° to 60° S. Several fault slices containing rocks of the Jackfork Group are present along the axial portion of the anticline. The Hairpin Curve and Coopers Cove localities are along the axial portion of the Peachland anticline.

Between the Ti Valley fault and the Peachland anticline is the south limb of another anticline. The Johns Valley Formation served as the gliding plane along which the south limb was thrust northward, obscuring the north limb in all but one locality (secs. 13, 14, 15, 16, T. 4 N., R. 20 E.).

Hairpin Curve locality.—State Highway 2 makes a hairpin turn approximately 13 miles south of Wilburton (SE¼ sec. 3, T. 3 N., R. 19 E.). Portions of the Game Refuge and Johns Valley Formation and the lower part of the Atoka Formation are exposed in road cuts at this locality, referred to as the Hairpin Curve locality. Structurally, these strata are part of the north limb of the Peachland anticline described above (pl. I).

Traveling southwestward, toward the hairpin turn, one goes down in the stratigraphic section from lower Atoka sandstone into the upper part of the Johns Valley Formation. Strata of the lower part of the Atoka are nearly vertical. Strata of the upper part of the Johns Valley exposed in road cuts dip 25° to 40°N. and are not overturned. Traveling northeastward, away from the hairpin turn, one goes down in the stratigraphic section from lower Johns Valley shales with interbedded black shale of typical Caney lithology into ripple-marked Game Refuge sandstone beds (fig. 7). Strata of part of the Johns Valley exposed on the south side of the hairpin turn dip 25° to 40°S. and are overturned. The Game Refuge sandstone is nearly vertical and has northward-facing bed tops (ripple-marked).

Exposures at the Hairpin Curve locality were first mapped by Hendricks and others (1947) and described by Hendricks, Fitts, Russom, and Jones (Tulsa Geol. Soc., 1947, p. 28-38). The map by Hendricks and others (1947) shows the Johns Valley to be in normal stratigraphic contact with the lower part of the Atoka Formation north of the hairpin turn. This sequence is shown on the map as being vertical or overturned toward the south. Bed tops face north.

The writer made the same observations and essentially the same structural interpretation.

Both the lower part of the Johns Valley south of the hairpin turn and the upper part of the Johns Valley north of the turn dip gently into the slopes. The Game Refuge Sandstone, in contact with the lower part of the Johns Valley Formation on the south side of the hairpin, and the lower part of the Atoka Formation, in contact with the upper part of the Johns Valley Formation on the north side of the hairpin, are almost vertical. Tops of both Game Refuge and Atoka strata face north. Anomalous dips of the Johns Valley strata have been produced by rock creep due to gravity. This is revealed by a close examination of strata of the lower part of the Atoka north of the hairpin turn. The 3-foot bed of black siliceous shale 100 to 200 feet above the base of the Atoka Formation clearly shows the effects of rock creep. In the road ditch this unit is overturned and dips 84°S. In the bank it dips 43°N. and is not overturned. In less than 20 feet the siliceous shale unit has undergone a 53° change in dip because of rock creep. Similar creep has occurred in the shales in the lower part of the Johns Valley south of the hairpin turn.

The Johns Valley Formation and the lower part of the Atoka Formation are particularly susceptible to rock creep, which makes more difficult structural interpretations of an area that has rather complex structure to begin with. The hairpin turn locality, at first glance, appears to be an unbroken anticline with Johns Valley strata exposed along its axis. However, a careful examination of sole marks, anomalous dips, and the stratigraphic succession proves that it is not.

Coopers Cove locality.—Coopers Cove is a topographic basin formed by erosion of the Johns Valley Formation and the overlying Atoka Formation along the axis of the Peachland anticline.

Evidence of high-angle reverse faulting along the anticlinal axis is present at the Slate Dump locality (NE¼ NE¼ sec. 32, T. 4 N., R. 20 E.). Younger Johns Valley strata have been moved northward over the Caney Shale in the lower part of the Johns Valley of the north limb of the Peachland anticline. Most of the movement along the high-angle reverse faults has been vertical (pl. I, section B-B′).

The relationship of the Caney Shale in the lower part of the

Johns Valley to the enclosing shale has been controversial. Ulrich (1927) and Miser (1934, p. 1000) believed that the masses of Caney Shale are actually exotics. Ulrich postulated that they were emplaced by drifting ice, whereas Miser attributed their emplacement to submarine slumping. Howell (1947), who noted a similarity between the structure of the Scottish Highlands and the belt between the Choctaw and Windingstair faults, concluded that slices of lubricant members and pre-Stanley rocks are present in the fault zone of sole thrusts. According to Hendricks (1958, p. 2761), Howell considered the Johns Valley Shale to be a stratigraphic unit that contains erratic boulders that are part of the depositional unit. However, Howell interpreted the large exotic limestone blocks and Caney Shale masses in the lower part of the Johns Valley at a number of localities to be part of a "friction carpet." Cline (1956b) pointed out that the Caney-type shale occupies a constant stratigraphic position in the lower part of the Johns Valley Formation.

Shale of Caney lithology was observed at three localities in Coopers Cove. No evidence that this shale is in a fault slice or that it was emplaced by slumping during deposition was found. The shale appears to occupy a constant stratigraphic position in the lower part of the Johns Valley Formation.

Several 100-foot exotic boulders of Arbuckle-like lithology, all located along the axis of the Peachland anticline, are present in the Johns Valley Formation in Coopers Cove.

TI VALLEY FAULT

The Ti Valley fault, along which rocks of Arbuckle facies and Ouachita facies are in contact, forms the northern boundary of the area of investigation. The trace of this fault has several large reentrants, indicating that it does not dip so steeply as do the other faults in the area.

The area between the Ti Valley fault and the Peachland anticline has been isoclinally folded and faulted. Isoclinal folds are well exposed along State Highway 2 in NE¼ sec. 28, T. 4 N., R. 19 E. (pl. I). No attempt was made to map the minor folds in this belt.

South of Bengal the Ti Valley fault truncates several high-angle reverse faults associated with anticlines.

BENGAL INLIER

The Bengal inlier, one of the pre-Stanley inliers north of the Ti Valley fault, is similar to those at Wesley, Ti, and Brushy Creek. Strata of Arbuckle-like facies are exposed in these inliers. Lithologically these strata are transitional between typical Ouachita and Arbuckle facies, but more closely resemble the latter.

The Bengal inlier is bounded on all sides by faults. On the north, Woodford Chert is in fault contact with the Atoka Formation. On the south, the Johns Valley Formation has been thrust northward over the Caney Shale along the Ti Valley fault.

CROSS FAULTS

Although not considered to be major faults, cross faults are common. In most of these northwestward-trending faults the east side has moved upward and northward. Hendricks (1959, p. 46) gave a more detailed description of cross faults.

Rock Creep

Interpretation of the complex structure of the report area was complicated by rock creep, the downslope bending of strata due to gravity. Rock creep produces anomalous structures, visible both in the field and on aerial photographs. Strata in the belt between the Ti Valley and Windingstair faults, as well as in other parts of the Ouachita province, have two characteristics which make them extremely susceptible to rock creep. First, alternating shale and sandstone units are present. Secondly, many of these sequences of alternating rock types are steeply dipping. Rock creep in areas of less steeply dipping strata is not so pronounced.

Erosion of steeply dipping stratigraphic sequences of alternating layers of unequal resistance to erosion results in a series of essentially parallel ridges and valleys. Soft shale that forms the slopes and valleys is most susceptible to creep. However, in many places even the resistant sandstone exposed near the ridge crests will creep downslope. The intensity of rock creep progressively increases toward the base of the slope. As a result, strata exposed on slopes dip into the slopes, at least near the surface. Strata exposed along the ridge crests may be unaffected by creep and, therefore, reliable attitudes can be obtained only along ridge crests and in valleys. In the area studied, the Johns Valley shales and portions of the Atoka Formation were commonly affected by rock creep.

Excellent examples of rock creep can be observed at the following localities: (1) Hairpin Curve locality (SE¼ sec. 3, T. 3 N., R. 19 E.), (2) along both the north and south slopes of Eight Mile Mountain, (3) in road cuts along State Highway 103 where it crosses Spring Mountain north of Big Cedar, and (4) in road cuts along U. S. Highway 71, approximately four miles southeast of Waldron, Arkansas (NE¼ sec. 10, T. 2 N., R. 29 W.).

The amount of rock creep varies along strike. As a result, false structures are produced. Clearly shown on aerial photographs, these "structures" resemble small, doubly plunging anticlines or drag folds, but field checking reveals no such structure.

INTERPRETATIONS

Tectonic Framework and Source of Sediment

Ham (1959, p. 72) interpreted the pre-Stanley strata of the Ouachita province as having been deposited in a shallow offshore basin, or shelf, that was steadily being depressed. According to Cline and Shelburne (1959, p. 205), these strata represent "a period of very slow sedimentation in essentially a starved trough."

The Stanley-Jackfork-Johns Valley sequence was deposited in the Ouachita geosyncline, a rapidly subsiding trough at the southern margin of the craton. The source of the sediment making up these units is not known.

Potter and others (1958, p. 1041) concluded that "a southwesterly regional slope existed in the early Mississippian, prevailed throughout Chester sedimentation in the Illinois basin, persisted throughout the long erosional interval of Mississippian-Pennsylvanian unconformity of the Illinois basin, and extended into the Pennsylvanian." Paleocurrent-direction studies by Reinemund and Danilchik (1957), Cline and Shelburne (1959), Cline (1960), Shelburne (1960), and Seely (1962) indicate that currents flowed toward the southwest in western Arkansas and southeastern Oklahoma during Mississippian-Early Pennsylvanian time. On the basis of these studies, it is conceivable that the Appalachian and Ouachita troughs might have been connected during this interval.

If the Appalachian and Ouachita troughs were connected during the Mississippian-Early Pennsylvanian interval and if currents flowed southwestward from the Appalachian region into the Ouachita region, long-distance transportation of clastics might have occurred. Positive areas in the southern Appalachian area might have served as sediment sources. Siever (1951, p. 575) suggested that the sediment eroded from the Illinois basin and surrounding areas such as the Ozark dome during the post-Kinkaid pre-Pennsylvanian interval might have been deposited in a remnant Mississippian sea on the southwest. If a southwesterly regional slope occurred in the Illinois basin during much of the Early Mississippian-Pennsylvanian interval (Potter and others, 1958), sediment might have been swept out of the Illinois basin by currents that flowed toward the southwest. Scull (1961), who concluded that much of the Atoka Formation was derived from the north and northeast, inferred the presence of a large delta in north-central Arkansas.

Clastics might have been derived from weathering of positive areas that adjoined the Ouachita trough on the south and southeast, in southern Arkansas, northern Louisiana, and northeastern Texas. These clastics might have merely been reworked and distributed by currents that flowed down the axis of the trough. Honess (1921, p. 78), who studied an area in the central Ouachitas in McCurtain County, Oklahoma, concluded that the Stanley Group represents the large delta of a stream that flowed north from a southern landmass, Llanoria. Morgan (1952) concluded that the southern sediment source consisted of a series of welts.

Doubt has recently been cast on the existence of a southern sediment source (Branan, 1961, p. 7; Seely, 1962, p. 202). However, it is difficult to ignore the conclusions of Honess, the increase in the sand-shale ratio toward the south (Miser, 1921, p. 70), the presence of thick beds of tuff in the lower part of the Stanley Group near the core area (Miser, 1921, p. 71), and the thickening of the Stanley-Jackfork-Johns Valley-Atoka sequence toward the south and southeast.

Therefore, a major unsolved problem of Ouachita geology is whether sediment was transported great distances from northeastern sources by currents that flowed toward the southwest parallel to the axis of maximum deposition or whether the sediment was derived from closer southern and southeastern sources and merely reworked and distributed by currents. The solution of this problem must involve petrologic studies. One must be cautious in working only with paleocurrent features such as sole marks. According to present interpretation, currents that flowed across mud bottoms produced scour features, which were later filled up with sand and preserved. The resulting protuberances on the bottoms of sandstone beds are known as sole marks. It is not known whether sand was deposited by the same current that scoured the mud bottom. To determine the mode of transportation of the sand that filled the scours, one must examine features within the sandstone units—grain orientation or cross-lamination. Most of the sandstone units are cross-laminated, although the laminae are inclined at a small angle. In many cases the current direction indicated by cross-lamination differs from that indicated by sole marks. This difference could be due either to derivation of sediment from marginal sources and reworking by currents or to transportation by anastomosing currents that flowed down the axis of the trough.

Environment of Deposition

None of the sedimentary structures observed gave positive indication of the depth of water in which they were formed. However, if the sandstone units had not been deposited below wave base, sedimentary structures would not have been preserved.

Apparently this was an environment in which slow deposition of clay-sized particles was frequently interrupted by a rapid influx of sand, possibly in turbidity currents. Intermittent periods of widespread quiescence occurred during which very little clastic material was deposited in the trough. Refer to the discussion of the deposition of siliceous shale and chert.

This environment was unfavorable for organisms. Shales in the Stanley-Jackfork-Johns Valley-Atoka sequence contain a microfauna consisting largely of Foraminifera. Brachiopods and crinoid columnals are present in some of the sandstone units but are rare. Plant fragments are abundant locally. Fossils in the sandstone units may have been transported to their deposition sites from some other environment. Some organisms left burrows and trails on the upper surfaces of the Johns Valley sandstone units.

Depositional Barrier Along Foreland Margin of Ouachita Geosyncline

The presence of a depositional barrier at the northern margin of the Ouachita geosyncline was suggested by Ulrich (1927, p. 26), Kramer (1933, p. 612), Rea (1947, p. 49), Flawn (1959, p. 26), and Ham (1959, p. 83). The barrier postulated by Kramer consisted of a chain of islands, which he called Bengalia. Kramer (p. 619) concluded that the inliers of pre-Stanley strata north of the Ti Valley fault are approximately on the axis of Bengalia. Rea theorized that "the presence of Arbuckle type rocks so far from their present source can be explained by a barrier formed of these type rocks that existed during Johns Valley time and separated the Arbuckle and Ouachita basins of deposition. This barrier was probably located along what is now the zone between the Ti Valley and Choctaw faults." Flawn suggested the possibility that the Ouachita geosyncline may have developed as two or more troughs separated by essentially continuous or en echelon submarine welts. He implied that the submarine welt might have served as a depositional barrier and theorized that ". . . the trough nearest the active source area receives the great load of rapidly deposited clastic sediments while the frontal trough receives only fine clastics and possibly siliceous sediments." Ham suggested that a depositional barrier may have separated the Arbuckle and Ouachita provinces during deposition of pre-Stanley strata.

EVIDENCE FOR DEPOSITIONAL BARRIER

Thinning of stratigraphic units along northern margin of Ouachita geosyncline.—Pre-Wapanucka stratigraphic units are much thicker in the central Ouachita province and the Arbuckle Mountains than in the area between, the frontal belt of the Ouachita province. This thinning, in addition to areal distribution of facies patterns, suggests that during pre-Middle Pennsylvanian deposition a flexure was present along the northern margin of the Ouachita geosyncline.

Tomlinson (1929, p. 16) described thickening of strata between the Sycamore (pre-Woodford) and Wapanucka Limestones toward the Ardmore basin. Hendricks (1947) stated that the Springer is 2,500 feet thick in the block southeast of the Choctaw fault but only 100 to 500 feet thick in the block southeast of the Pine Mountain fault. According to Ham (1955, p. 31),

> The maximum thickness of Caney-Sycamore rocks is nearly 1,200 feet in the Arbuckle anticline, compared with a normal thickness of about 350 feet in the Hunton anticline and Wapanucka syncline. The regional distinction between basin and shelf established in Cambrian and Ordovician time thus reappears in Mississippian time, although during the Silurian-Devonian interval this distinction is not marked.

Ham (1959, p. 84) concluded that "the pre-Stanley rocks thin markedly eastward in the Arbuckles and markedly westward in the Ouachitas, as though they were merging toward a common shelf between them."

Whether the Atoka Formation is thinner in the frontal belt cannot be determined because of complex structure and erosion of the top of the formation.

Branan (1961, p. 13) interpreted a pre-Woodford positive feature beneath Black Knob Ridge, but did not show that the overlying Woodford Chert also thins in this area.

Relationship of Wapanucka Limestone, Chickachoc Chert, and Wapanucka equivalents south of Ti Valley fault.—Goldstein and Hendricks (1962, p. 404, 406) concluded that the Wapanucka Limestone was deposited above wave base in a clear, open, well-aerated sea. The Chickachoc Chert, its equivalent in the block southeast of the Katy Club fault, was probably deposited in clearer and deeper water, possibly on a part of the ancient continental shelf that was deeper than normal wave base. Hendricks (1947) stated that no Wapanucka or Chickachoc equivalent is in the block between the Pine Mountain and Ti Valley faults.

Honess (1924, p. 20) correlated the beds containing Morrow faunas in northern McCurtain County, Oklahoma, with the Wapanucka Limestone. Cline and Shelburne (1959) described similar lenses of mold-fauna sandstone in the Johns Valley Formation between the Ti Valley and Windingstair faults and in the Boktukola syncline. They interpreted these lenses as equivalents of Wapanucka Limestone washed in from shallower areas. The writer found thin, discontinuous lenses of limestone of shallow-water origin interbedded in the upper part of the Johns Valley Formation between the Ti Valley and Windingstair faults and interpreted them as Wapanucka equivalents. Apparently the deeper water environment of the Chickachoc Chert was bounded on the north and on the south by shallower water environments.

The presence of limestone lenses of shallow-water origin south of the Ti Valley fault is significant. If Goldstein and Hendricks' interpretation of the environment of deposition of the Wapanucka Limestone and Chickachoc Chert is correct, and if the present interpretation that the limestone lenses are in place in the upper part of the Johns Valley Formation south of the Ti Valley fault is correct, then an area of shallow-water deposition was south of the deeper water environment in which the Chickachoc Chert was deposited. Some sort of depositional barrier must have been present along the northern margin of the Ouachita geosyncline at this time.

Today the crest of this barrier is apparently represented by the area between the Pine Mountain and Ti Valley faults.

Whether Wapanucka equivalents thinned over this feature and were later removed by pre-Atoka (formation) erosion is not known. Goldstein and Hendricks (1962, p. 410) diagrammatically showed a positive feature beneath the block southeast of the Pine Mountain fault. Interpretation of this feature is apparently based upon the absence of Wapanucka equivalents in this block. Their diagram indicates that the Springer and Wapanucka Formations were removed by pre-Atoka (formation) erosion and that the Atoka Formation is unconformable upon the Caney in the block southeast of the Pine Mountain fault.

Some structural feature served as a barrier that separated Wapanucka and Chickachoc deposition on the north from upper Johns Valley deposition on the south. Eastward near Le Flore, Oklahoma, the Wapanucka becomes sandy. Limestone is not present between Le Flore and the Arkansas state line (Wallis, 1915, p. 42). Morris (1962, p. 28) concluded that thin limestone beds are indigenous in the Johns Valley Formation in the Y City and Acorn quadrangles, Arkansas. Perhaps the depositional barrier extended this far eastward.

DESCRIPTION OF DEPOSITIONAL BARRIER

The present interpretation is that a submarine welt or flexure was located along the northern margin of the Ouachita geosyncline during deposition of the Stanley-Jackfork-Johns Valley sequence and possibly at times during the deposition of pre-Stanley strata. The flexure may have extended from the western Ouachitas into western Arkansas. It was not a strongly positive feature. The writer visualizes it merely as having been slightly less negative than the foreland trough that adjoined on the north and the Ouachita geosyncline that adjoined on the south. No evidence is available to indicate that the flexure, or welt, beneath the area now known as the frontal belt, was ever mountainous or highly folded, as was suggested by Powers (1928, p. 1045), Cheney (1929, p. 570), Kramer (1933, p. 611), Moore (1934, p. 447), and Miser (1934, p. 1003).

The present interpretation is that this submarine flexure served as a barrier between deposition of the Arbuckle and the Ouachita facies. A transitional sequence of strata, more closely resembling the Arbuckle facies, was deposited over the flexure. The two facies were not deposited in widely separated areas and have not been brought together by long-distance thrusting. Strata originally deposited south of the flexure have been moved northward along the Ti Valley fault, the location of which is considered to be directly related to the flexure.

Strata of Ordovician-Early Pennsylvanian age thin over the flexure, indicating that it was present during much of this depositional interval. It might have been accentuated in post-Woodford pre-Atoka (formation) time, particularly after deposition of the Johns Valley and Springer Formations.

The foreland rim flexure postulated by the writer is essentially the same structural feature described by Kramer (1933, p. 611) and Flawn (1959, p. 26). This feature was also shown diagrammatically by Goldstein and Hendricks (1962, p. 410), and by Hendricks (unpublished report, 1962).

Exotics of Arbuckle-like Rocks in Ouachita Province

Present knowledge of the exotic-bearing horizons in the Johns Valley Formation was summarized by Tomlinson (1959, p. 12):

> At present, it appears that both the location of the source area or areas from which the boulders came, and the manner of their transportation, are still in doubt. It seems likely, however, that a closer approach to solution of both questions may be achieved by more thorough analysis of the character of the boulders and their relation to the enclosing shale.

The present field study was begun with these words in mind. The interpretations below are based largely on observations made in the report area.

CLUES TO EMPLACEMENT OF EXOTICS

Discussed below are features or relationships that give some clue to the mode of emplacement of exotic clasts of Arbuckle-like rock types present within the Johns Valley Formation in the report area.

Areal distribution of exotic-bearing horizons.—The exotic-bearing portion of the Johns Valley Formation has the configuration of a narrow band bounded on the north by the Ti Valley fault. This band extends from near Atoka, Oklahoma, to east of Boles, Arkansas, a distance of more than 130 miles. Such a distribution pattern is suggestive of nearshore deposits. Because no pre-Woodford strata of Arbuckle-like rock types crop out in the Ouachita province, it has been theorized that exotics were derived from a distant source, the Arbuckle Mountains, by ice rafting. However, no exotics were rafted into the intermediate area. Because of this fact and because of the reasons listed by Tomlinson (1929, p. 26) and Bailey and others (1928, p. 606) ice rafting is not considered as a possible mechanism for the transportation of exotics.

The unusual distribution of exotics in the Johns Valley and other formations can be explained by deposition along a fault scarp or a series of scarps.

Distribution of exotics within stratigraphic units.—The distribution of exotics within stratigraphic units at the Hairpin Curve and Stapp localities is extremely significant. At the Hairpin Curve exotics are present in a "channel" that has cut into 11.5 feet of underlying strata (Cline, 1960, p. 72). The shale in which the channel is located also contains exotics. At the Stapp locality exotics are in pockets in clay shale, as well as in the clay shale (fig. 9). Harlton (1938, p. 893) described this locality.

The exotic-bearing clay shale that contains the channels or pockets of exotics is marine. No stream deposits are associated. The distribution of exotics at these localities precludes normal deposition by currents or streams.

Size of exotics.—Exotics in the Johns Valley Formation range from a fraction of an inch to several tens of feet in diameter. The large size of many of the exotics indicates that a special method of transportation must have been involved. Blocks of such sizes could not have been transported far from their source.

Shape and surface texture of exotics.—The exotics, pitted before deposition, were not abraded and show no evidence of having been transported by streams or ocean currents. The irregular pitting on many of the exotics (fig. 10) precludes long-distance transportation. Had the exotics been subjected to wave action on a beach, they would have been rounded and abraded. If they were formed by subaerial weathering processes, they should have partially weathered surfaces. However, they all appear to have been washed clean before deposition and to have no such weathered outer layers.

The exotics are interpreted as having been formed by submarine weathering processes and transported only short distances from their source.

Rolled sandstone masses within exotic-bearing horizons.—Rolled sandstone masses present within the exotic-bearing horizons of the Johns Valley Formation provide a significant clue to the emplacement of these units. Shape, fracture pattern, and filling of fractures in the sandstone masses indicate that they were formed while the sand was in a plastic state. Presence of sole marks, small-scale cross-lamination, and top marks indicates that the rolled sandstone masses were derived from sandstone beds that had been previously deposited by currents. Thin-section analysis showed that the green sandstone interbedded in the upper part of the Johns Valley Formation and the rolled sandstone masses are lithologically alike.

Cline (1956c, p. 58), who concluded that the rolled sandstone masses were deformed and rounded prior to and perhaps during fracturing, also suggested that they might be tectonic. Cline and Shelburne (1959, p. 74) stated that the sandstone masses represent lenses of sand torn from the bottom by a submarine slide, or flow, and rolled along the flow. Shelburne (1960, p. 55) concluded, "Balled sandstone masses in the Johns Valley Formation are structures formed by penecontemporaneous slumping of the beds of sandstone within a thick shale sequence." He noted (p. 36) that the size of the "rounded sandstone boulders" is related to the thickness of the parent bed, and he even identified a parent bed at one locality.

Rolled sandstone masses of different origins are present within the Johns Valley and other formations. Some are tectonic, having been formed either along slippage planes during folding or along minor bedding-plane faults. The writer has observed localities where the rolled sandstone masses resemble boudinage structure. Other rolled sandstones are undoubtedly primary depositional features formed by contortion of sand lenses or strata during penecontemporaneous slumping. The rolled sandstone masses in the Johns Valley Formation at the Hairpin Curve locality are primary depositional features. No faulting is present in the sequence which contains these sandstone masses.

SOURCE OF EXOTICS

In place at several horizons within the Ouachita province are exotic clasts of Arbuckle-like rocks. However, no rocks lithologically similar to those of the exotics crop out within the Ouachita province. Therefore, some geologists have postulated that the exotics were transported to their depositon sites from distant sources. The exotics are here interpreted as having been derived from local sources along the northern margin of the Ouachita geosyncline and transported only short distances to their deposition sites. If this interpretation is correct, Arbuckle-like rocks must have been present beneath the northern margin of the geosyncline and exposed at or near the surface during deposition of the Johns Valley and other exotic-bearing units. This idea was previously presented by Tomlinson (1929, p. 26), Cheney (1929, p. 570), Dixon (1931, p. 343), and Hendricks and others (1937, p. 23). Referring to the differences between strata of the Arbuckle and Ouachita areas, Ham (1959, p. 72) concluded that "although certain facies differences are indeed real, they are not yet sufficiently well established in space to compel acceptance of the concept of large-scale thrusting."

SLUMPING FROM FAULT SCARPS

The exotics in the Johns Valley Formation are interpreted as having been emplaced by submarine slumping from the base of normal fault scarps that developed along a hinge-line zone of tension at the northern margin of the Ouachita geosyncline during deposition of the Stanley-Jackfork-Johns Valley sequence. Strata of pre-Stanley Arbuckle-like rocks exposed in the scarps were weathered by submarine, or possibly subaerial, processes. The unconsolidated strata that accumulated at the base of the scarps periodically slumped basinward along the steep northern margin of the geosyncline. Exotics were transported within the slumping masses. Earthquakes associated with the faulting might have triggered the slumps or slides.

The faults were formed along the southern margin of the submarine flexure during deposition of the Stanley-Jackfork-Johns Valley sequence. These faults, therefore, are related to the development of the Ouachita geosyncline. They did not form as a result of compressive forces associated with the Ouachita orogeny.

No single, high fault scarp extended along the entire margin of the geosyncline. Instead, a series of normal faults probably developed along the hinge-line zone of tension. Harlton (1938) proposed the concept of subsidence along a hinge zone, which developed into a zone of tension and normal faulting.

Hendricks (1947), on the other hand, visualized an Early Pennsylvanian fault overthrust from the north, which brought to the surface rocks of the Arbuckle province on the north side of the geosyncline in which the Stanley, Jackfork, and Johns Valley rocks were deposited. Hendricks believed that the fault scarp supplied the exotics now found in the Jackfork and Johns Valley but made no mention of how they were transported into their enclosing strata. He stated that the fault was later concealed by overthrusts from the south, and he named the fault after Sidney Powers to whom he attributed the origin of the idea.

Cline (1960, p. 22) stated that deposition of the Stanley-Jackfork sequence occurred along a hinge line but did not postulate faulting along the hinge line. In a diagram of Goldstein and Hendricks (1962, p. 410), a hinge zone and a flexure adjoining on the north are shown. They stated (p. 400) that presumably the exotic boulders were transported into place from southward-facing fault scarps by submarine landslides and sediment flows triggered by earthquakes associated with faulting along the northern margin of the geosyncline.

A description of submarine slumping from fault scarps along a geosynclinal hinge zone was given by Kuendig (1959, p. 464-465) in his classification of post-Paleozoic geosynclines:

> The steep shelf edge marks the landward edge of the geosynclinal hinge zone, the zone of strong differential vertical movements. We assume that this zone, as far as the outer crust is concerned, is considerably affected by step faulting with periods of tension, rather than by downfolding.
>
> The hinge or slope zone is morphologically a zone of scarp forming, and its high mobility is accompanied by frequent shallow earthquakes. The angle of slope may vary from 90° to 2°. Sediments deposited on such a slope, as well as those laid upon the bordering shelf edge, will periodically glide and slump into the nearby abyss, and turbidity currents will be a frequent phenomenon. Not only the unconsolidated sediments, but sometimes also the hard bedrock

> of the scarps will glide into the eugeosyncline and accumulate near the foot of the slope, particularly when they are strongly brecciated by fault movements. . .
>
> Such block accumulations, often of great extent and including giant boulders of even more than a cubic kilometer, represent a submarine eugeosynclinal screen. They have been described in many places. . . and were generally designated as exotic blocks. The term ''olistostrome'' has been introduced for these accumulations by Flores; the individual blocks are then ''olistoliths.''
>
> The most notable effect of the geosynclinal slope. . .has hitherto scarcely been recognized. The slope is a facies divide of the first order and separates, by submarine erosion as a full or partial gap, the shallow shelf sediments from the isochronous deep eugeosynclinal sediments which are deposited at a topographically much lower level.

The postulated faults have not been observed in the field and might never be observed. The Atoka Formation, as much as 5,000 feet thick, was deposited over the scarps. During the Ouachita orogeny the strata deposited south of the fault scarps were detached from their source, or sources, and were thrust northward along the Ti Valley fault. These opinions were expressed by Powers (1928), Harlton (1934, p. 1029), and Hendricks (1947).

The interpretation of Cline and Shelburne (1959, p. 198) that the "channel" at the Hairpin Curve locality is the product of a single submarine slide is probably correct. Rolled sandstone masses formed within the slumping mass. Some lenses of sand were undoubtedly torn from the bottom, rolled along in front of the slumps, and incorporated into the slumping masses. The present interpretation is that the exotics moved within slumps and did not slide across the surface of the mud bottom. The soft underlying strata were distorted, grooved, or channeled by the sliding or slumping masses. Crowell (1957) described similar slumping at four localities in California and at two localities in the Alps.

Exotics were emplaced at several horizons in the Stanley-Jackfork-Johns Valley sequence. They were described from the Chickasaw Creek Formation (Harlton, 1938; Hendricks, 1947), siliceous shale of the Markham Mill Formation (Harlton, 1938), and the Wesley Formation (Hendricks, 1947). Cline (1960, p. 80) described them from the Caney Shale in the lower part of the Johns Valley Formation. The presence of exotics at several horizons in the Stanley-Jackfork-Johns Valley sequence indicates periodic faulting and submarine slumping during this interval. However, emplacement of the exotics was most pronounced during deposition of the Johns Valley Formation.

That exotics were derived from sources along the northern margin of the Ouachita geosyncline has been postulated by Powers (1928, p. 1045), Cheney (1929, p. 569), Dixon (1931, p. 344), Kramer (1933, p. 613), Miser (1934, p. 1003), Moore (1934, p. 447), Harlton (1938, p. 861), Hendricks (1947), Goldstein (Flawn and others 1961, p. 38), and Goldstein and Hendricks (1962, p. 400).

TERMINOLOGY APPLIED TO CLASTS OF ARBUCKLE-LIKE ROCK

Clasts of Arbuckle-like rock in place in the Johns Valley Formation are considered to be nonexotic. They are interpreted as having been derived from local sources at the northern margin of the Ouachita geosyncline. The clasts, which range from sand size to boulder size, should not be described as boulders. Neither should they be called erratics, because of the implication of glacial origin. In the past these clasts have been described as exotics because they were believed to have been derived from some other geologic province. In this report the clasts are called exotics only because their composition is considerably different from the matrix which surrounds them. The term olistolith has also been used to describe similar clasts.

Nonexotic clasts, described elsewhere in this report as rolled sandstone masses, are also present in the exotic-bearing horizons of the Johns Valley Formation. These clasts are interpreted as segments of sand units that were deformed by submarine slumping before they were lithified.

Deposition of Siliceous Shales and Cherts

Miser and Honess (1927, p. 12, fig. 2B), Cline (1960, p. 10), and Goldstein and Hendricks (1962, p. 410) interpreted the black siliceous shales and cherts of the Stanley-Jackfork-Johns Valley sequence as tongues of the Caney Shale. However, there is little agreement about the source of the silica and the depth of water in which the black siliceous shales and cherts were deposited. The widespread deposition of these thin units is difficult to explain, as was stated by Laudon (1959, p. 89):

> The origin of the silica is still a major problem. The thinness of the siliceous shales relative to their large areal extent suggests that their deposition was controlled by some event which took place almost instantaneously over the entire basin. . . The tectonic mechanism whereby the source area for a rapidly filling geosyncline can be made periodically quiescent for a period of time remains entirely beyond the comprehension of the present writer. . .Yet it seems that such an event occurred and essentially no clastics were deposited over the entire basin during the time of deposition of each of the siliceous shales of the Stanley.

Shelburne (1960, p. 57) stated:

> The wide distribution of these cherts indicates that uniform conditions prevailed over large areas in the geosyncline. . . The distribution of siliceous shales suggests that the formation of these deposits is affected by proximity to the slope or marginal shelf environment. The number of siliceous shales and thickness of individual siliceous shales decreases from the frontal Ouachitas to the central Ouachitas.

Harlton (1938, p. 861) described exotics in the siliceous shale of the Markham Mill. Hendricks (1947) noted that boulders up to seven feet in diameter are present in two of the siliceous shale beds of the Jackfork Group. One of these beds is at the base of the Jackfork (the Chickasaw Creek) and the other is in the upper part (Wesley). Cline (1956b) described the presence of exotics in the black siliceous shale of Caney lithology in the lower part of the Johns Valley Formation in Johns Valley.

Evidence that the exotics were derived from nearby fault scarps, coupled with the common association of exotics with black siliceous shales, suggests that the origin of the black siliceous shales in the Stanley and Jackfork Groups might be directly related to faulting along the margin of the geosyncline. Perhaps this faulting is the tectonic mechanism whereby the steadily subsiding Ouachita geosyncline was periodically and almost instantaneously made quiescent.

Deformation During Ouachita Orogeny

MECHANICS OF DEFORMATION

Compression of the Ouachita geosyncline after deposition of the Atoka Formation resulted in elevation of the thick Stanley-Jackfork-Johns Valley sequence. The uplift was accompanied by folding and reverse faulting along the northern margin of the geosyncline, where deformation was most severe.

In the area studied, two anticlines were formed, compressed, overturned, and finally faulted. Wallis (1915, p. 39) described the same type of deformation in the frontal belt of the Ouachita province. The faults related to the anticlines are high-angle reverse faults with relatively small horizontal components of movement. The south limbs of these southwestward-striking anticlines moved upward and northward along the reverse faults, many of which are confined to the incompetent Johns Valley Formation. Finally, several major reverse faults developed, dividing the province into a series of fault-bounded blocks, each major block, in turn, being cut by reverse faults related to anticlines and by cross faults. Two of these major faults, the Ti Valley and Windingstair faults, were mapped in the report area. Although early movement might have occurred along these faults, they were the ones along which latest movement occurred (Hendricks, 1959, p. 51). The structure of the area studied is almost a replica of that of the Valley and Ridge province of the southern Appalachians.

Intensity of deformation increases northward from the Windingstair fault. A transition occurs from the imbricate blocks adjacent to the Ti Valley fault, to tightly folded and faulted anticlines between the Ti Valley and Windingstair faults, to the broad, open synclines south of the Windingstair fault.

DISPLACEMENT ALONG MAJOR REVERSE FAULTS

Maximum northward movement along major reverse faults in the Ouachita province occurred at the western end of the province (Wallis, 1915, p. 40; Miser, 1934) and diminished eastward. Reinemund and Danilchik (1957) mapped the Choctaw fault as dying out and other faults as having only minor displacements in the Waldron quadrangle, Arkansas (Tomlinson, 1959, p. 8). Apparently movement was clockwise around a vertical axis in west-central Arkansas.

The amount of northward movement along reverse faults in the area studied must be less than that in the western Ouachitas and more than that in west-central Arkansas.

No fault surfaces were observed in the area. However, the straight surface traces of the Ti Valley and Windingstair faults indicate that they dip steeply. The large reentrants in the trace of the Ti Valley fault indicate that it dips less steeply than do the Windingstair fault and the reverse faults related to anticlines. The present interpretation is that movement along the Ti Valley and Windingstair faults was predominantly vertical in the area studied.

RELATIONSHIP OF TI VALLEY FAULT TO DEPOSITIONAL BARRIER AND SOURCE OF EXOTICS

During the Ouachita orogeny the Ti Valley fault developed along what was the hinge-line zone of tension and normal faulting during deposition of the Stanley-Jackfork-Johns Valley sequence. This was also a zone of weakness during portions of the pre-Stanley depositional interval. Strata deposited at the base of the normal faults were moved upward and northward along the Ti Valley fault. Cline (1960, p. 22) described the possible relationship of the Ti Valley fault to the hinge line. Inliers of pre-Stanley strata north of the Ti Valley fault represent slices that have been moved upward along the Ti Valley fault from the underlying block or foot wall. The contact of the Ouachita and Arbuckle facies along the Ti Valley fault can be explained by rapid northward convergence of post-Arkansas Novaculite units and a small amount of northward displacement along the fault.

Several features indicate that strata of the Johns Valley Formation bounded on the north by the Ti Valley fault were deposited relatively short distances from their sources. These features are exotics, exotics in a sandstone matrix, thin-bedded conglomeratic sandstones, and beds and blocks of Wapanucka Limestone in the upper part of the Johns Valley Formation.

Exotics, interpreted as having been emplaced by submarine slumping from fault scarps, were not moved far from their sources.

Exotics, normally restricted to clay shale, are not commonly embedded in a sandy matrix. Clay, much more susceptible to slumping than sand, can slide or flow farther. However, one would expect to find sand along the base of fault scarps if sandstone were exposed in the scarps. Therefore, one would expect to find exotics in a sandstone matrix adjacent to the source or sources. Such exotic-bearing sandstone and boulder conglomerate is present at the Compton Cut locality, less than one mile south of the Ti Valley fault (fig. 14).

The thin-bedded conglomeratic sandstones previously described were deposited by currents and not by submarine slumping. However, the pebbles were probably derived from the same source or sources as the exotics in the clay shale. Pebbles the size of those in the conglomeratic sandstone were probably not transported far from their source.

The presence of Wapanucka Limestone of shallow-water origin in the upper part of the Johns Valley Formation not only indicates that Johns Valley deposition continued into the Pennsylvanian period, but also gives a clue to the environment of deposition. Both interbedded limestone lenses and bedlike exotic blocks were observed. Occasionally thin beds of limestone were deposited in the upper part of the Johns Valley, whereas at other times previously deposited limestone beds slumped or glided basinward, perhaps in the same manner as described by Kuendig (1959, p. 464-465).

SUMMARY

A submarine flexure, or welt, along the northern margin of the Ouachita geosyncline served as a depositional barrier that separated the Ouachita and Arbuckle facies during the Mississippian-Early Pennsylvanian interval. Deposition in the Ouachita geosyncline was accompanied by subsidence along a hinge zone located along the southern margin of the depositional barrier. During deposition of the Stanley-Jackfork-Johns Valley sequence, occasional normal faulting, probably submarine, occurred along the marginal hinge zone, a zone of tension and weakness. Strata with Arbuckle-like lithologic composition, which ranged in age from Ordovician to Early Pennsylvanian, were exposed in the south-facing normal-fault scarps that developed in the hinge zone of tension.

Pitted and partially dissolved clasts of Arbuckle-like rocks, ranging from sand size to boulder size, accumulated at the bases of the fault scarps. Clasts of Arbuckle-like rocks that accumulated at the base of the fault scarps moved basinward in submarine slumps, probably triggered by earthquakes associated with the normal faulting, and were deposited with strata of the Ouachita facies. Unconsolidated sand layers, interbedded with the clasts that accumulated at the base of the fault scarps, were twisted and deformed into rolled sandstone masses during submarine slumping. Clasts of Arbuckle-like rocks, in place in the Johns Valley Formation, were derived from local fault scarps and are nonexotic.

The presence of tongues of Caney Shale in the lower part of the Johns Valley and interbedded lenses and exotics with Wapanucka rock types in the upper part of the Johns Valley Formation proves that deposition of the Johns Valley Formation began sometime during the Mississippian and continued into the Early Pennsylvanian.

During the Ouachita orogeny the Ti Valley fault developed at what had been the hinge zone during deposition of the Stanley-Jackfork-Johns Valley sequence. Inliers of pre-Atoka Arbuckle-like strata along the Ti Valley fault are fault slices that were brought up along the Ti Valley fault from the block below. Fault slices similar in origin to those mentioned above, but composed of strata of the Ouachita facies, are exposed along the Windingstair fault.

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APPENDICES

APPENDIX A

Measured Stratigraphic Section

Section measured along tributary of Long Creek west of Bengal, Oklahoma, in NE¼ NW¼ SW¼ sec. 11, T. 4 N., R. 21 E.

	Feet
Caney Shale:	
9. Shale, black, fissile; with dense, black petroliferous, limestone concretions	30
8. Siltstone, gray, thin-bedded; with plant-fragment impressions	12
7. Shale, green, silty	2
6. Shale, green; with phosphatic nodules	0.5
Total Caney exposed	44.5
Woodford Chert:	
5. Shale, black; with phosphatic nodules	0.5
4. Chert, black; with some white bands; gray on weathered surface; in beds one-half inch to five inches thick; and shale, black, fissile, with common, but not abundant, phosphatic nodules	11
3. Shale, olive-green, silty, easily eroded; resembles some shales of the Stanley Group	3.3
2. Chert, black; in beds one-half inch to six inches thick; and shale, black, fissile; in beds less than two inches thick. Phosphatic nodules are present in both chert and shale	14
1. Shale, black, fissile	7
Total Woodford exposed	35.8

APPENDIX B

Observed Exposures of Caney Shale in the Lower Part of the Johns Valley Formation

1. NW¼ SE¼ sec. 3, T. 3 N., R. 19 E.
2. NW¼ NE¼ sec. 32, T. 4 N., R. 20 E.
3. NE¼ NE¼ sec. 32, T. 4 N., R. 20 E.
4. SE¼ NW¼ sec. 33, T. 4 N., R. 20 E.
5. SW¼ NE¼ sec. 23, T. 4 N., R. 21 E.
6. SE¼ NE¼ sec. 13, T. 4 N., R. 21 E.

APPENDIX C

Observed Exposures of Wapanucka Limestone Lenses or "Exotic" Blocks in the Upper Part of the Johns Valley Formation

1. NE¼ NE¼ sec. 14, T. 3 N., R. 19 E.
2. NW¼ NE¼ sec. 14, T. 3 N., R. 19 E.
3. NE¼ NE¼ sec. 24, T. 4 N., R. 21 E.
4. NW¼ NE¼ sec. 24, T. 4 N., R. 21 E.
5. S½ SW¼ sec. 18, T. 4 N., R. 22 E.
6. NE¼ NW¼ sec. 14, T. 4 N., R. 21 E.
7. SW¼ SE¼ sec. 10, T. 4 N., R. 21 E.
8. SW¼ SW¼ sec. 11, T. 4 N., R. 21 E.

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