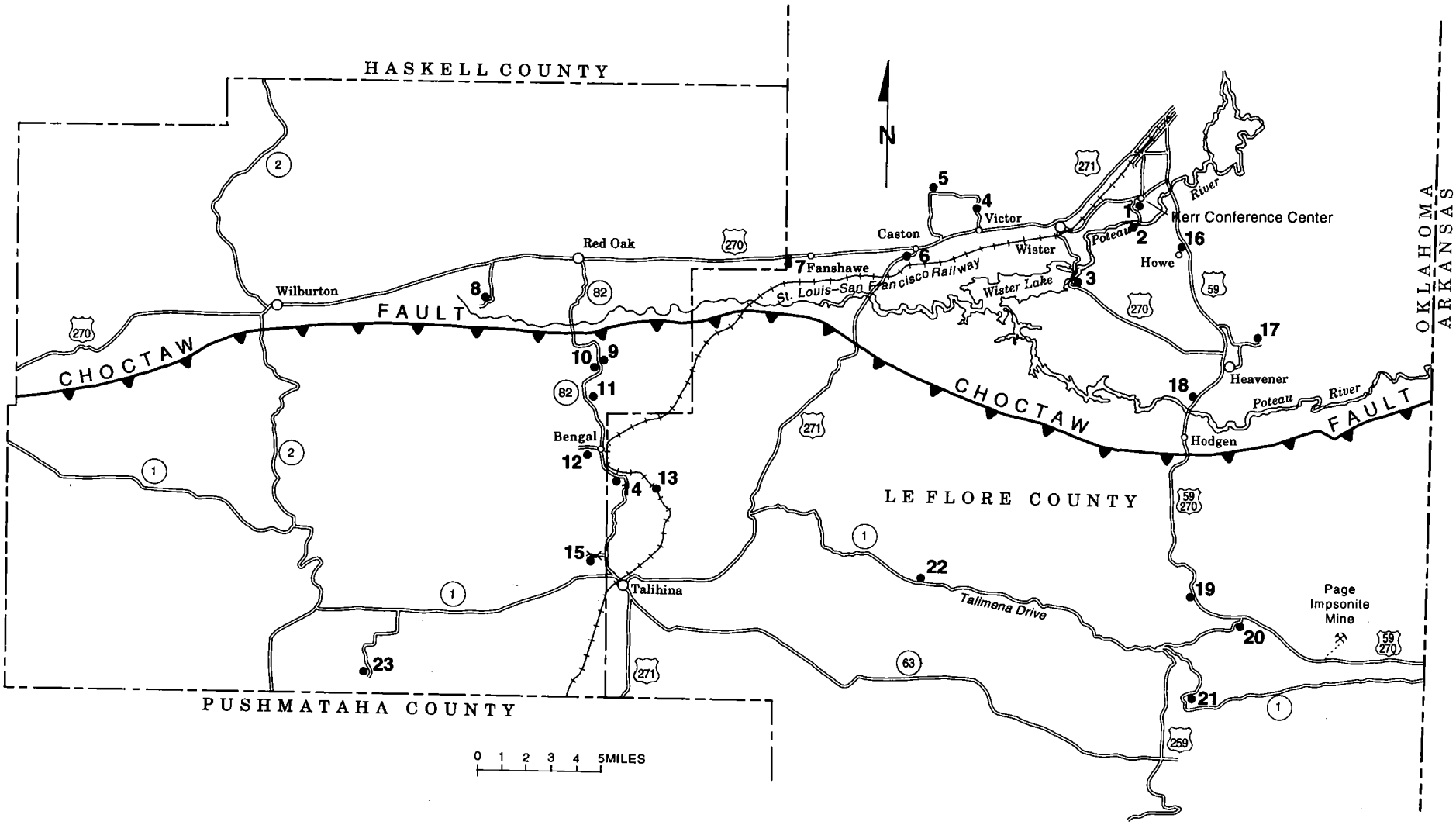
OKLAHOMA GEOLOGICAL SURVEY GUIDEBOOK 29



Geology and Resources of the Eastern Ouachita Mountains Frontal Belt and Southeastern Arkoma Basin, Oklahoma

MAP OF FIELD - TRIP STOPS



Oklahoma Geological Survey
Charles J. Mankin, *Director*

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Guidebook 29

Geology and Resources of the Eastern Ouachita Mountains Frontal Belt and Southeastern Arkoma Basin, Oklahoma

Neil H. Suneson and LeRoy A. Hemish
Editors

Proceedings of a one-day workshop (November 15, 1994) held in Poteau, Oklahoma, and a two-day field trip (November 16–17, 1994) to the Ouachita Mountains and southeastern Arkoma Basin, Oklahoma.

The University of Oklahoma
Norman

1994

DEDICATION

We dedicate this guidebook to our families, who graciously tolerated our extended absences from home to produce the maps on which much of this book is based.

Peter Suneson
Ryan Suneson
Judy Reynolds

Jesse Hemish
Janice Hemish

Thanks.

Noel

Lee Roy

Front Cover

Side-looking radar (SLAR) image of the Ouachita Mountains and southern Arkoma Basin of Oklahoma.

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PREFACE

This guidebook provides an introduction to new research on the geology and mineral resources of the southern part of the Arkoma Basin and the northern part of the Ouachita Mountains in southeastern Oklahoma and adjacent Arkansas. The guidebook is divided into two parts: a road log for a two-day field trip that begins and ends at the Robert S. Kerr Conference Center in Poteau, Oklahoma, and contributed research papers. The focus of the field trip is the area within the perimeter marked by Poteau, Red Oak, Talihina, and the Talimena Drive in Oklahoma (see map, inside front cover). The structure and stratigraphy of Carboniferous strata within the area, as well as energy resources in the region, are emphasized. Contributed papers by industry, academic, and government geologists represent the variety of important new research that is being done in the Arkoma Basin and Ouachita Mountains. Some of the ideas and interpretations presented in these papers differ from ours. We hope that these differences will lead to discussions and to a better understanding of the geology of the field-trip area.

This guidebook includes several of the region's historical and archaeological aspects. They are included because we believe that most geologists are curious about the landscape and its history, whether that history goes back 100 years or 100 million years.

Most of the field-trip area has been mapped recently at a scale of 1:24,000 by the Oklahoma Geological Survey (OGS). The mapping was funded by the OGS and the U.S. Geological Survey (USGS) as part of the USGS COGEOMAP (Cooperative Geologic Mapping) program that involved the OGS, USGS, and the Arkansas Geological Commission. More recently, the mapping was completed as part of the USGS STATEMAP program.

We thank all the local landowners who have given us permission to examine the rocks on their property. Special thanks go to Bob Cooper of the Farrell-Cooper Mining Company for allowing us to visit the Wister Coal Mine. Several individuals helped us with our Oklahoma history: Edward Connie Shoemaker of the Oklahoma Historical Society (History of the Old Military Road); Don L. Hofsommer, Dept. of History, St. Cloud University (The St. Louis–San Francisco Railway); and Kenneth L. Smith, Fayetteville, Arkansas (History of Logging in the Oklahoma Ouachitas). We are especially grateful to Christie Cooper, OGS editor, and her staff (particularly Frances Young), and to Charlotte Lloyd, Jim Anderson, and T. Wayne Furr (OGS Cartographic Section) for their efforts in helping us to produce a well-designed and well-illustrated guidebook.

NEIL H. SUNESON AND LEROY A. HEMISH
Editors and Field-Trip Leaders

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Part I



Road Logs and Stop Descriptions

Geology and Resources of the Eastern Ouachita Mountains Frontal Belt and Southeastern Arkoma Basin, Oklahoma

INTRODUCTION

The Arkoma Basin and the Ouachita Mountains in Oklahoma are part of a long, mostly buried, foreland basin/fold-and-thrust belt pair that forms the southern margin of the North American craton. Prior to the Carboniferous, the Arkoma Basin area was part of an epicontinental shelf that received a relatively thin sequence of shallow-water carbonates and clastics. Similar strata also underlie most, if not all, of the exposed Ouachita Mountains in Oklahoma. Coeval deep-water facies strata, consisting mostly of black shales and cherts, accumulated in the Ouachita trough to the south (see stratigraphic column, inside back cover).

Similar patterns of deposition persisted in the Arkoma Basin through the Morrowan. In the middle Atokan, the Arkoma Basin developed into a major foreland basin. This transition is marked by the abrupt change from deposition of relatively thin, shelfal, Morrowan and early Atokan sandstones and carbonates to deposition of thick, middle Atokan, deep-water clastic sediments. To the south in the Ouachita Mountains, Upper Mississippian strata mark the onset of rapid subsidence of the Ouachita trough and voluminous clastic sedimentation. The Late Mississippian–Early Pennsylvanian depocenter of the Ouachita Mountains migrated northward into the Arkoma Basin during the early Atokan through Desmoinesian.

The beginning of Ouachita deformation is difficult to establish; it is possible that the earliest folding and thrust faulting was synchronous with the Late Mississippian subsidence of the Ouachita trough, but located south of it. Cretaceous strata unconformably overlying Ouachita strata are undeformed. At the end of the early Desmoinesian, Ouachita-related folding of Arkoma Basin strata ceased and uplift of parts of the Ouachita Mountains occurred.

In Oklahoma, the Choctaw fault (map, inside front cover) separates the frontal belt of the Ouachita Mountains (which consists of steeply tilted, imbricately thrust-faulted, and complexly folded strata) from the Arkoma Basin (a region characterized by broad, open folds). In general, the topography of the Arkoma Basin is an inverse reflection of the underlying strata: the mountainous areas typically mark the surface traces of synclines and the valleys are underlain by anticlines. Cavanal Mountain, visible just to the north of the Robert S. Kerr Conference Center, is an example of a synclinal mountain. In contrast, the topography of the Ouachita Mountains is controlled by rock type. The mostly low, gently rolling northern part of the frontal belt is underlain by the Atoka Formation and Johns Valley Shale, which are dominantly shale. The imposing ridgeline of Winding Stair Mountain, the dominant topographic feature in this part of the frontal belt, is held up by the thick sandstones of the Jackfork Group.

The principal mineral resources of the region are coal and natural gas. Building stone and shale are relatively minor resources. The coal beds occur only in the Arkoma Basin. From oldest to youngest, they are: Lower Hartshorne, Upper Hartshorne, McAlester, Upper McAlester, Cavanal, Secor, and Secor Rider. The coals are predominantly low- and medium-volatile bituminous and they have medium to high sulfur contents. Methane gas is being produced from the Hartshorne coal. Seven coal beds have commercial value; currently, five are being mined in the field-trip area. The Lower Hartshorne coal is being mined in the Pine Mountain Mine, near Heavener; the McAlester and Upper McAlester coals are being mined in the Red Oak Mine, near Red Oak; and the Secor and Secor Rider coals are being mined in the Wister Mine, near Wister. In Oklahoma, all production from these coal beds is from surface operations. However, the Lower Hartshorne coal is being mined in an underground operation near the town of Hartford, Arkansas, about 15 miles southeast of Poteau.

Natural gas in the Arkoma Basin and Ouachita Mountains has been exploited since the early 1900s. Peaks in drilling activity occurred in the 1960s and from the late 1980s to the early 1990s. The major gas reservoirs in the field-trip area range in age from Ordovician to Pennsylvanian and include (from oldest to youngest): Arbuckle Group carbonates, Spiro sandstone, Red Oak sandstone, and Hartshorne Sandstone. There are many minor reservoirs that range in age from Devonian to Pennsylvanian. At present, the Oklahoma Nomenclature Committee of the Mid-Continent Oil and Gas Association has recognized seven gas fields in the field-trip area; they range in size from the one-well, 640-acre Talihina NW field to the giant Red Oak–Norris field.

As in the early part of this century, the Arkoma Basin and Ouachita Mountains continue to be fertile ground for research designed to further our understanding of the tectonics of the mountain belt and the sedimentology of the associated foreland basins. In addition, our increasing understanding of the geology will aid the ongoing exploration for and development of energy resources in this part of Oklahoma.

Most locations in this guidebook use standard section-township-range abbreviations. In addition, the letter L refers to line; for example, C WL is "center of the west line" and FEL is "from the east line." Rock-color terms are those shown on the rock-color chart (Rock-Color Chart Committee, 1991).

PRELIMINARY DAY

ROAD LOG

Cumulative mileage	*Interval*	
0.0	0.0	Start trip from the Robert S. Kerr Conference Center, located in the NE¼NE¼ NE¼NE¼ sec. 21, and the NW¼NW¼ NW¼NW¼ sec. 22, T. 6 N., R. 25 E., Le Flore County, Oklahoma (Wister 7.5′ quadrangle).

Robert S. Kerr Conference Center

The Kerr Conference Center in Poteau was built originally as a home for the late Robert S. Kerr and his family. It was designed by Jean Bolls and constructed by Gatle Frank Construction Company. Work started in 1957 and was completed in 1960.

Robert S. Kerr unquestionably was one of Oklahoma's most prominent and well-known citizens. He was born in 1896 in a log cabin near Ada, Oklahoma, and went on to serve as governor of the State. He was elected to the U.S. Senate in 1948 and was serving his third term when he died on January 1, 1963, at the age of 66. At the time, he was considered by some to be the leading possibility for the U.S. presidency.

The Kerr home was presented as a gift from the Kerr family to the Oklahoma State Board of Regents for Higher Education in 1978. It was remodeled for its current use in 1983 at a cost of $500,000. Since 1968, the Kerr Mansion also has housed the Kerr Museum, which now is located in the former carport, remodeled for its present use. The museum is operated by the Eastern Oklahoma Historical Society.

Visitors to the Kerr Mansion have included such famous people as Carl Albert, George Bush, Gerald Ford, Lyndon B. Johnson, John F. Kennedy, and Norman Schwartzkopf.

The valley to the south and east of the Kerr Conference Center was carved by the Poteau River, a stream that enters the Arkansas River at Fort Smith, Arkansas. The area drained by the Poteau River receives extremely heavy rainfall (~50 in. annually), and, historically, the stream has been noted for flooding. Since construction of Wister Dam in the 1940s, the lower reaches of the Poteau have been protected from those floods that originate in the Ouachita Mountains and upstream to the west along the Fourche Maline.

Walk ~150 yd southwest from Conference Center and descend bluff.

STOP 1A

Spaniard(?) Limestone—Contact between McAlester and Savanna Formations

Location: SE¼NE¼NE¼NE¼ sec. 21, T. 6 N., R. 25 E., Le Flore County

A section of the lower part of the Savanna Formation and the upper part of the underlying McAlester Formation is exposed in the bluff overlooking the Poteau River Valley just southwest from the Kerr Conference Center (map, inside front cover). The base of the Savanna Formation is picked here at the base of a thin, fossiliferous marine limestone interpreted to be the Spaniard(?) Limestone (Hemish, 1993b, p. 84). Occurrences of marine limestones are uncommon in the area of the field trip. Recognition of the Spaniard(?) Limestone in this part of the Arkoma Basin is critical because it provides a definitive marker for mapping the McAlester/Savanna contact. The convention of mapping the McAlester/Savanna contact at the base of the oldest sandstone in the Savanna Formation was established by Taff (1899), and in the absence of the Spaniard(?) Limestone, that convention was followed in recent Arkoma Basin mapping by Hemish (1991,1992), Hemish and Mazengarb (1992), Hemish and Suneson (1993,1994), and Hemish and others (1990a,b,c). Oakes and Knechtel (1948, p. 51, fig. 7) showed that the base of the Spaniard Limestone in Muskogee County occupies approximately the same stratigraphic position as does the base of the lowest sandstone of the Savanna Formation in the Arkoma Basin. Throughout the shelf area of northeastern Oklahoma, the base of the Spaniard Limestone Member marks the contact between the McAlester Formation and the Savanna Formation (Branson and others, 1965, p. 26; Oakes and Koontz, 1967, p. 15).

The outcrop belt of the Spaniard(?) Limestone in the field-trip area extends westward from Stop 1A for ~9 mi to just west of Caston Creek (Hemish, 1993b, fig. 1). Extent of the limestone to the east from Stop 1A is unknown.

Measured Section, Stop 1A gives a description of the strata exposed in the bluff; Figure 1 shows a graphic columnar section. Unit 1 is typical of shales in the McAlester Formation, and Unit 8 is representative of many of the sandstones in the Savanna Formation. (The Savanna Formation is discussed in detail at Stop 17A.)

Measured Section, Stop 1A

Spaniard(?) Limestone — Contact between McAlester and Savanna Formations

SE¼NE¼NE¼NE¼ sec. 21, T. 6 N., R. 25 E., Le Flore County (Wister 7.5′ quadrangle). Measured from top of bluff down steep slope ~150 yards southwest from Kerr Conference Center (from Hemish, 1993b, p. 110–111).

	Thickness (feet)
KREBS GROUP:	
SAVANNA FORMATION:	
8. Sandstone, light olive gray (5Y5/2) to moderate brown (5YR4/4) and dark reddish brown (10R 3/4), very fine grained, noncalcareous, medium- to thick-bedded; irregular, nonparallel- to nodular-bedded; indistinctly cross laminated; blocky; micaceous; contains small fragments of black (N1) carbonized plant material; trace fossils on sole; top eroded; base sharp	3.0
7. Shale, light olive gray (5Y5/2), weathers grayish orange (10YR7/4) and light brown (5YR5/6); includes some shaly siltstone stringers; moderately bioturbated; weathers to small, blocky flakes on outcrop; base gradational (poorly exposed)	17.4
6. Sandstone, light olive gray (5Y5/2), weathers grayish orange (10YR7/4) and moderate brown (5YR4/4), very fine grained, noncalcareous, very thin bedded, shaly, weathers to flakes on the outcrop, moderately to strongly bioturbated; base gradational	2.3
5. Shale, olive gray (5Y4/1) to light olive gray (5Y6/1), stained light brown (5YR5/6) in part, silty, fissile, flaky, moderately bioturbated; includes rare dark reddish brown (10R3/4) and moderate reddish brown ironstone concretions as well as rare, 4-in.-thick, dusky brown (5YR2/2), ferruginous concretions containing poorly preserved fossil shells; base sharp	18.4
4. Sandstone, grayish brown (5YR3/2) to moderate reddish brown (10R4/6), very fine grained, noncalcareous, medium-bedded, strongly bioturbated; contains rare, poorly preserved fossil marine shells; thickness variable laterally; base sharp	1.1
3. Shale, medium dark gray (N4) to olive gray (5Y4/1), fissile, flaky, noncalcareous; contains abundant, dark yellowish orange (10YR6/6) and moderate brown (5YR4/4) clay-ironstone concretions occurring in 0.5-in.-thick, discontinuous layers; weakly bioturbated; base sharp	12.9
Spaniard(?) Limestone Member:	
2. Limestone (skeletal lime mudstone), medium gray (N5), weathers moderate yellowish brown (10YR5/4), impure, shaly, contains abundant broken and well-preserved marine fossils—predominantly brachiopods; thins laterally; base gradational	0.6
MCALESTER FORMATION:	
1. Shale, medium dark gray (N4), weathers moderate reddish brown (10R4/6), brittle, flaky, noncalcareous; contains scattered moderate brown (5YR3/4; 5YR4/4) ironstone concretions of varying sizes—generally 1–2 in. in diameter; includes concretionary shale structures; base covered	27.5
Total thickness of section	83.2

STOP 1A

Spaniard(?) Limestone—Contact between McAlester and Savanna Formations

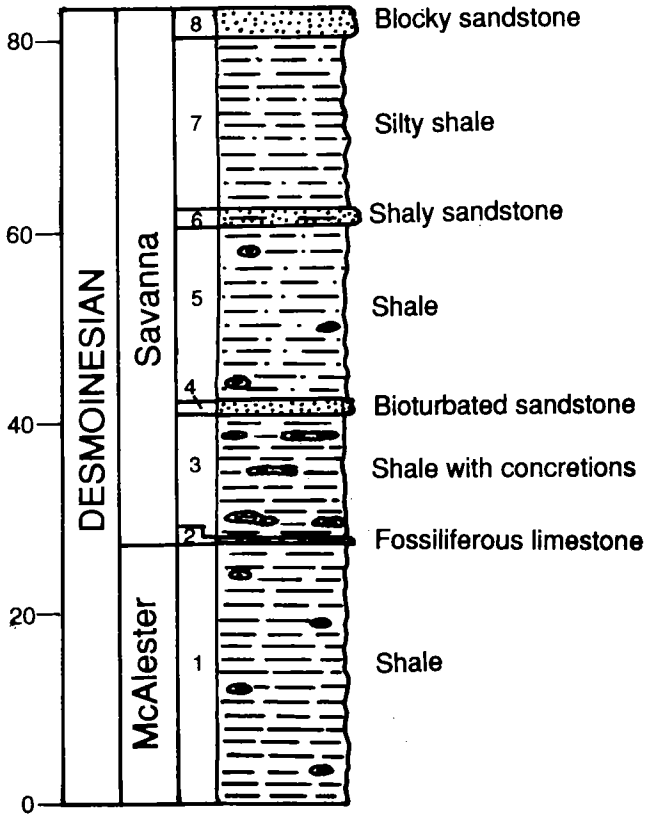


Figure 1. Graphic columnar section of upper part of McAlester Formation and lower part of Savanna Formation at Stop 1A.

		Return to parking lot. Enter vehicle; turn left on blacktop road and travel 0.5 mi west.
0.5	0.5	Intersection with gravel road. Turn left (south) and drive ~0.4 mi to shale pit.
0.9	0.4	STOP 1B. Shale pit.

STOP 1B

Spaniard(?) Limestone—Contact between McAlester and Savanna Formations

Location: SW¼NE¼SW¼NE¼ sec. 21, T. 6 N., R. 25 E., Le Flore County

Measured Section, Stop 1B gives a description of the strata from the curve in the road at top of bluff to covered area at bottom of shale pit; Figure 2 shows a graphic columnar section.

Figure 3 is a photograph of the Spaniard(?) Limestone taken in a small gully near the base of the shale pit. Where observed along its outcrop belt, the unit is similar lithologically and does not exceed more than a few inches in thickness.

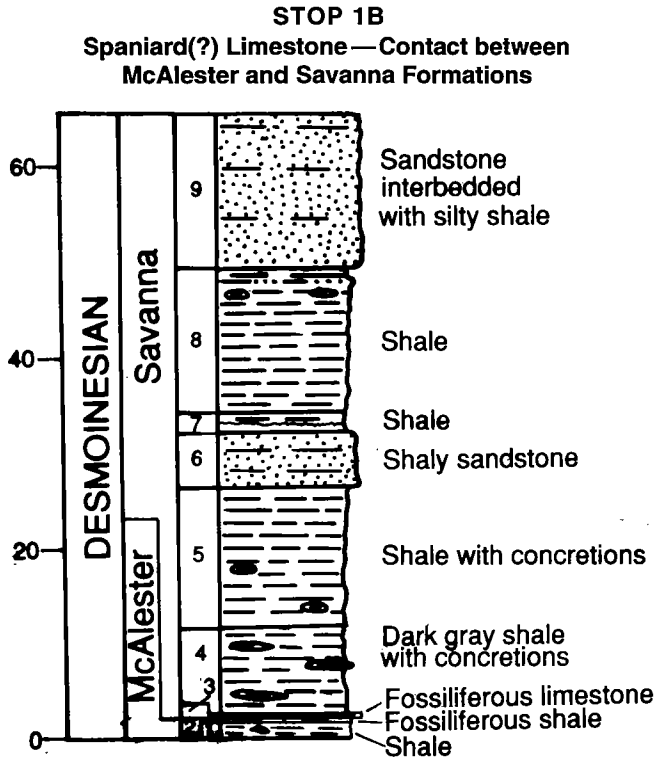


Figure 2. Graphic columnar section of upper part of McAlester Formation and lower part of Savanna Formation at Stop 1B.

Measured Section, Stop 1B

Spaniard(?) Limestone — Contact between McAlester and Savanna Formations

SW¼NE¼SW¼NE¼ sec. 21, T. 6 N., R. 25 E., Le Flore County (Wister 7.5′ quadrangle). Measured in shale pit and in road ditch (from Hemish, 1993b, p. 109–110).

Thickness (feet)

KREBS GROUP:

SAVANNA FORMATION:

9. Sandstone, light olive gray (5Y5/2) to moderate yellowish brown (10YR5/4) to light brown (5YR 5/6), with some moderate reddish brown (10R 4/6) staining, very fine grained, micaceous, noncalcareous, thin- to medium-bedded; bedding irregular to parallel and nodular in places; blocky fracture; bioturbation features abundant; interbedded with light olive gray (5Y5/2), silty, flaky shale; base sharp, with trace fossils on sole 16.0
8. Shale, grayish orange (10YR7/4) to moderate yellowish brown (10YR5/4), to light olive gray (5Y 5/2), weathers to small flakes on outcrop; includes a 2-in.-thick layer of thin-bedded, bioturbated, light olive gray (5Y5/2), sandy siltstone 2 ft below top of unit; also contains dark reddish brown (10R3/4), sandy, ferruginous concretions 1–8 in. in diameter in upper 4 ft of unit; weakly to moderately bioturbated; base gradational.... 15.2

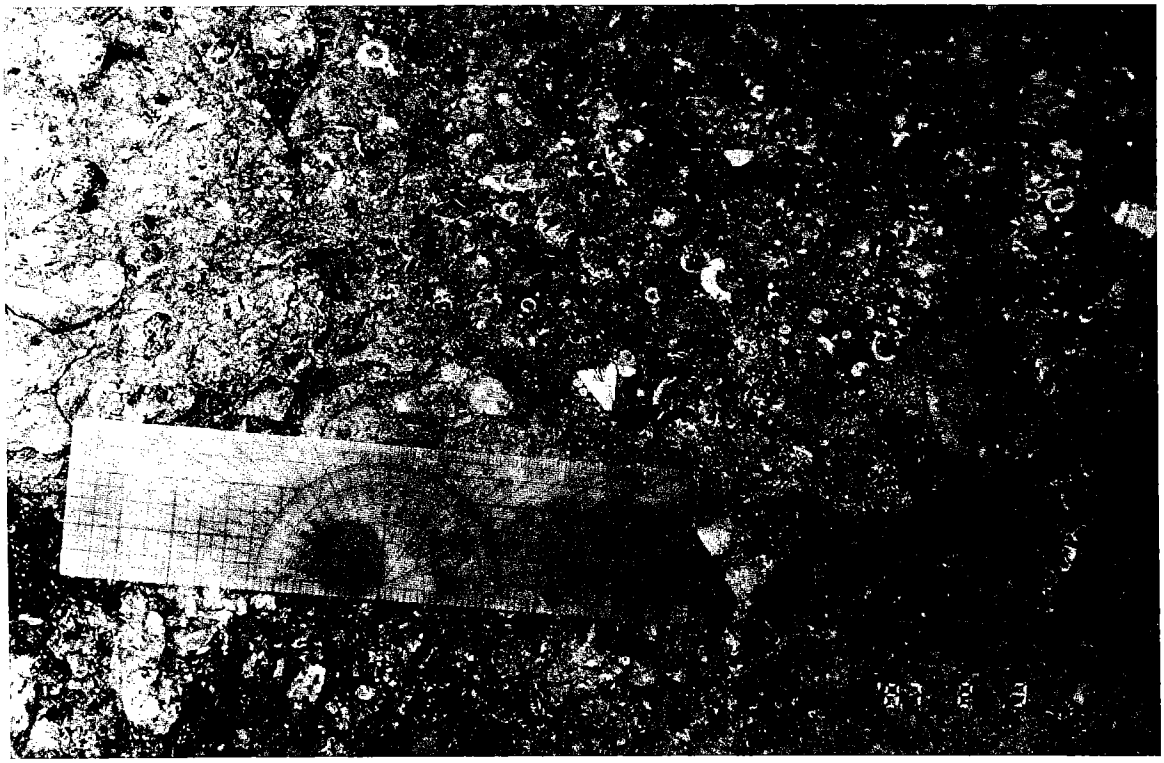


Figure 3. Outcrop photo of the Spaniard(?) Limestone at the location of Measured Section, Stop 1B (from Hemish, 1993b, fig. 6). Note the shelly nature of the rock at this site. Scale is 6 in. long.

7. Shale, olive gray (5Y4/1) to medium dark gray (N4), flaky, moderately bioturbated, carbonaceous in some thin layers; base sharp 2.5
6. Sandstone, light olive gray (5Y5/2) to moderate yellowish brown (10YR5/4), stained light brown (5YR5/6) in part, very fine grained, noncalcareous, shaly, indistinctly parallel-wavy-bedded, thin-bedded, moderately bioturbated, base gradational 5.2
5. Shale, olive gray (5Y4/1) with light olive gray (5Y6/1) tint in part; stained grayish brown (5YR 3/2) in places, silty, brittle, fissile, contains rare dark yellowish orange (10YR6/6) and grayish red purple (5RP4/2) clay-ironstone concretions in lower part, includes concretionary shale structures several feet in diameter; base gradational 14.8
4. Shale, dark gray (N3), weathers light brown (5YR 5/6) along joints, silty, brittle, fissile, contains concretionary shale structures continuous with overlying unit; also contains abundant dark yellowish orange (10YR6/6) and blackish red (5R2/2) clay-ironstone concretions occurring in irregular, rounded discoidal shapes that form ledges 0.5–1.0 ft apart; moderately bioturbated; base sharp 8.6

Spaniard(?) Limestone Member:

3. Limestone (skeletal lime mudstone), brownish gray (5YR4/1) to grayish black (N2), weathers moderate brown (5YR4/4) to light brown (5YR 5/6), impure, shaly, abundantly fossiliferous, contains numerous well-preserved brachiopod valves and crinoid ossicles; base sharp 0.1

McAlester Formation:

2. Shale, grayish black (N2), very calcareous, crumbly, contains abundant well-preserved brachiopod fossils, carbonaceous; contact between fossiliferous shale and underlying nonfossiliferous shale sharp 0.2
1. Shale, medium dark gray (N4), weathers moderate reddish brown (10R4/6), noncalcareous, soft, clayey; contains rare, dark gray (N3), irregular-shaped, 1-in.-diameter nodules; base covered 2.4

Total thickness of section 65.0

		Proceed south on winding gravel road across flood plain of Poteau River and cross river on concrete bridge.
1.6	0.7	STOP 2. Stop at south side of river. Low-water crossing of Poteau River.

STOP 2

Keota Sandstone Member of McAlester Formation

Location: SW¼SW¼SE¼SE¼ sec. 21, T. 6 N., R. 25 E., Le Flore County

The Keota Sandstone is well exposed at the south end of the bridge over the Poteau River (Fig. 4). The Keota



Figure 4. Keota Sandstone exposed at Poteau River crossing, SW¼SW¼SE¼SE¼ sec. 21, T. 6 N., R. 25 E., Le Flore County (from Hemish, 1993b, fig. 14). Note the flaggy and shaly nature of the member.

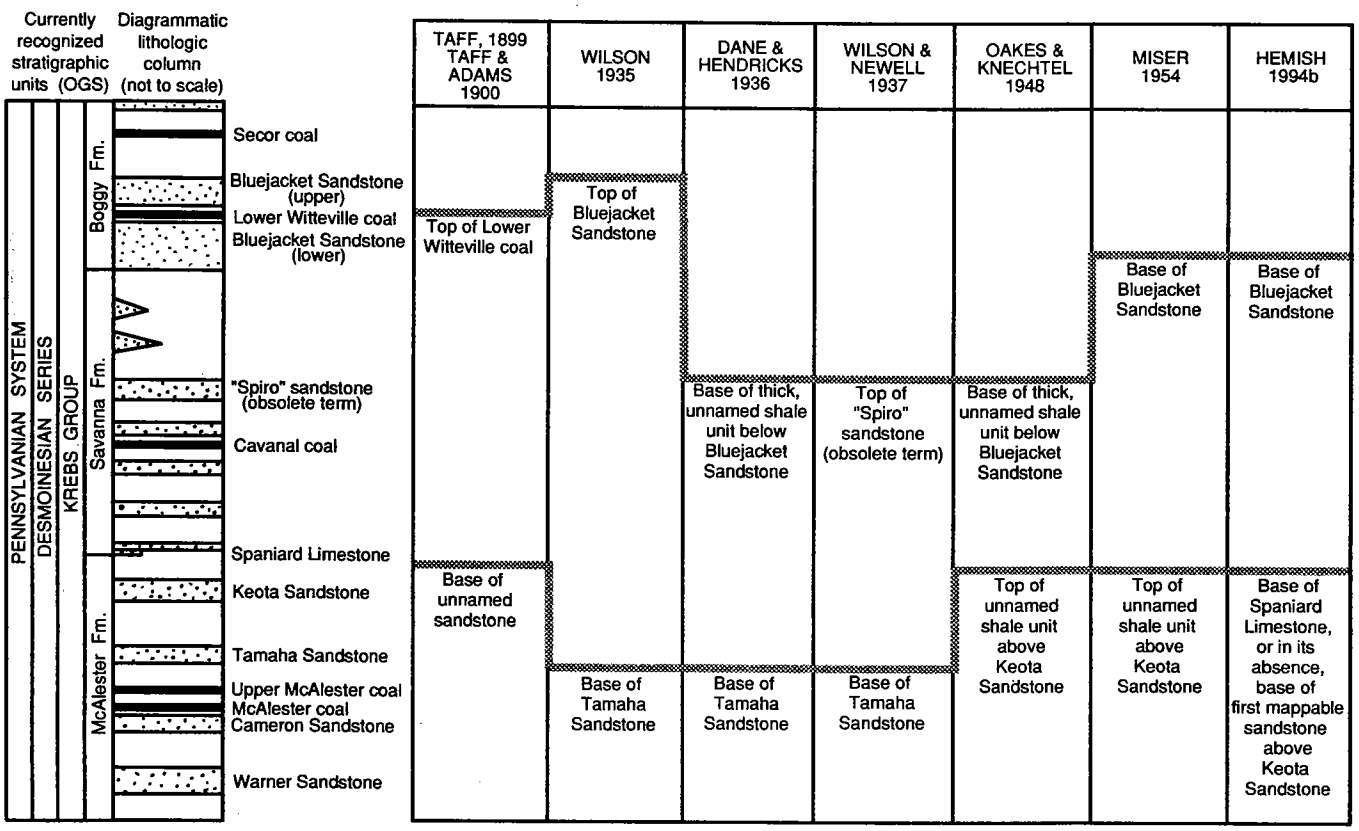


Figure 5. Concepts of the McAlester/Savanna and Savanna/Boggy boundary positions in Oklahoma (from Hemish, 1994b, fig. 7).

Sandstone is the youngest named member in the McAlester Formation.

Taff (1899) first used the name McAlester for outcrops of the formation in the vicinity of McAlester, Pittsburg County, Oklahoma. He did not designate a formal type locality. Common usage of the term McAlester Formation has been applied to strata between the top of the Hartshorne Formation, below, and the base of the Savanna Formation, above. The McAlester Formation in the field-trip area consists predominantly of dark-colored shales with abundant sideritic concretions, several prominent ridge-forming sandstones, and two commercially important named coal beds, as well as other thin, discontinuous, unnamed coals. The upper and lower contacts of the McAlester Formation have been redefined at various times. The boundary positions of the contact with the underlying Hartshorne Formation are discussed at Stop 6. Figure 5 shows changing concepts of the McAlester/Savanna contact through time.

One shale unit and five sandstone units have been named in the McAlester Formation. They are, the McCurtain Shale Member at the base, the Warner Sandstone Member, the Lequire Sandstone Member, the Cameron Sandstone Member, the Tamaha Sandstone Member, and the Keota Sandstone Member.

Thom (1935) named the Keota Sandstone from outcrops in the vicinity of the town of Keota, Haskell County, Oklahoma. Wilson and Newell (1937, p. 43) classified the Keota as a member of the Savanna Formation (Fig. 5), whereas it was shown by Oakes and Knechtel (1948, p. 42) to be in the McAlester Formation. The Keota Sandstone Member consists predominantly of sandstone and shale beds, but locally may contain thin limestones or coals. The number and thickness of beds in the Keota varies; thus, measured sections may range from 20 to 200 ft or more (Oakes, 1977, p. 20). It is persistent and mappable over a large area of northeastern Oklahoma.

The outcrop of Keota Sandstone exposed at Stop 2 is typical. It is commonly rippled, platy, and shaly. Of particular interest at this exposure are the dewatering structures on some beds, which might be mistaken for large-scale interference ripples.

Owing to discontinuous occurrences and difficult access to outcrops, stops are not scheduled for viewing either the Tamaha Sandstone Member or the Lequire Sandstone Member of the McAlester Formation. In general, both units are composed predominantly of olive gray, very fine grained, thin-bedded, shaly sandstone, rarely exceeding 4–8 ft thick. The Tamaha is characterized by an abundance of well-preserved trace fossils at most outcrops in the field-trip area. The Lequire is absent over most of the area but, in places, is a low ridge former. Although the Tamaha does form ridges in some places, it is usually too thin and shaly to be a ridge former.

3.2 1.6 Return to Kerr Conference Center.

DAY ONE

ROAD LOG

Cumulative mileage	*Interval*	
0.0	0.0	Start trip from Kerr Conference Center parking lot. Turn left (west) onto blacktop road. Descend to valley on dip slope of Savanna sandstone, then drive on shales of the Savanna Formation. (The Savanna Formation is discussed in detail at Stop 17A.)
1.6	1.6	Turn left on old U.S. Highway 271 and continue due southwest on Savanna shale and Quaternary alluvium (in low-lying areas along streams) toward the town of Wister.
2.9	1.3	Cross Rock Creek and ascend onto Quaternary terrace deposits associated with Mountain Creek and Caston Creek. The main part of Wister is built on terrace deposits and alluvium.
4.4	1.5	Turn left (south) in downtown Wister onto U.S. Highway 270 and proceed through town, crossing railroad tracks and following the bends in the road.
4.9	0.5	Cross Caston Creek and drive southeast on its flood plain. The ridges to the right are formed by the Cameron and Warner Sandstone Members of the McAlester Formation.
5.9	1.0	Dip-slope outcrop of the upper unit of the Warner Sandstone on side of road. Continue around broad curve in road, enter Wister State Park and ascend to crest of Wister Dam. McCurtain Shale Member of McAlester Formation crops out in ditch on northwest side of highway.
6.7	0.8	North end of Wister Dam. Turn left (south) and drive on dam.

WISTER, OKLAHOMA

The town of Wister, Oklahoma, was founded (more correctly, the post office was established) on June 30, 1890, as a railroad junction. Between 1905 and 1920, the town thrived as a lumber, mining, and stock center. In addition, a cotton gin and the railroad were important sources of income for local residents.

Perhaps the most memorable events in the town's history relate to disasters. In the spring of 1909, a fire started in a room of the Brown Boarding House between the bank and the Frisco railroad. At the time, all the buildings in town were made of wood, and there was no water to fight fires. The flames spread quickly, and the entire northern part of Wister was destroyed. Later that same day, a clerk at the Shipley Drugstore, located on the south side of the Rock Island railroad tracks, accidently ignited a can of wood alcohol. There was an explosion and eight nearby buildings were destroyed within a matter of hours.

On April 18, 1927, at about 12:20 a.m., heavy rains caused Mountain Creek to overflow its banks and flood Wister. The Reverend Buck Thomas and his wife were trapped in their home; when rescued, they were clinging to their mattress as it floated to the ceiling of their bedroom. The flood was blamed on new highway construction that turned the water from Mountain Creek into the center of town. The State settled $80,000 in claims for damages. Another flood hit Wister in 1961. On May 18, at about 3:00 a.m., nearly 6 inches of rain fell in the Mountain Creek and Caston Creek drainages. Rapidly, the part of Wister north of the Rock Island railroad was inundated by about 8 feet of water. This storm occurred less than two weeks after a tornado had devastated the town of Howe (see Day Two, p. 90).

WISTER LAKE AND DAM

Wister Lake was authorized by the Flood Control Act of 1938 for the purposes of flood control and conservation. The project was designed and built by the U.S. Army Corps of Engineers, Tulsa District. Construction of the dam began in April 1946, and the dam was placed in full flood-control operation in December 1949. The dam impounds waters of Fourche Maline Creek and the Poteau River.

The total project includes: a 5,700-ft-long, 99-ft-high embankment (Wister Dam) across the Poteau River Valley; an earth dike 2,400 ft in length; and a 600-ft-long, uncontrolled, concrete, chute-type spillway located between the dike and the embankment. Six vertical lift gates regulate flood-control releases through the outlet works.

The 4,000-acre lake stores 27,100 acre-ft of water and has a shoreline of 115 mi. The conservation-pool elevation is 472 ft above sea level (1 Dec. to 1 June). The flood-pool elevation is 509.5 ft above sea level. There are 400,800 acre-ft of flood-water storage with a surface area of 23,070 acres (U.S. Army Corps of Engineers, undated brochure).



Figure 6. Thick, monotonous sequence of steeply dipping, olive black to dark gray silty shales of the Atoka Formation exposed at edge of discharge channel below Wister Dam (from Hemish, 1993a, fig. 18).

7.0	0.3	Quarry Isle and Wister State Park Headquarters on the right (west). Quarry Isle is formed by a hogback of moderately dipping sandstone in the Hartshorne Formation. Dips here are ~25°, increasing to the south toward the axis of the Heavener anticline. Continue south across dam.
7.5	0.5	Control tower and lift gates on the right (west); discharge chute on the left (east) with outcrops of steeply dipping, dark gray shales of the upper part of the Atoka Formation visible in channel walls downstream (Fig. 6).
7.6	0.1	Entrance to rest stop and scenic overlook at south end of embankment. To the west is a view of Wister Lake and Wister Valley. Continue on U.S. 270 and drive due east. The dam's 600-ft-wide spillway is to the left (north) and ahead to the right (south) is the 2,400-ft-long, 40-ft-high dike.
8.1	0.5	Turn sharply left (west) at U.S. Army Corps of Engineers "Lake Wister" sign and drive to parking area at east side of spillway.
8.2	0.1	Stop 3.

Prehistory of the Wister Valley

Larry Neal
Oklahoma Archaeological Survey
Norman, Oklahoma

The combined drainages of the Poteau River and its major tributary, Fourche Maline Creek, comprise a distinct archaeological region called the Wister Valley. Much of the upper part of the basin in Oklahoma is filled by Wister Lake, and a good part of its drainage is from the north face of the Ouachita Mountains. The terraces and low ridges of the region contain evidence for some 10,000 years of human use. Good sites of these early occupants are rare due to the drastic effects of Holocene climatic changes. These changes affected not only the people of the time but also the preservation of traces of their way of life. Much of the Fourche Valley was eroded and flushed out sometime after 3000 B.C., eliminating early intact deposits. Only at such sites as Mackey (in process of analysis) and Scott (Bell, 1953; Galm and Flynn, 1978), are there traces of these early people in the Wister Valley; even then the traces are not in clear stratigraphic separation. Deposits from later occupations of the valley are more intact and contribute to an understanding of the transition made by a prehistoric people over a span of some 2,000 years—from hunter-gatherers to pottery-using horticulturists to full-fledged farmers.

The archaeological sites containing this evidence line the banks of the two major streams of the Wister Valley. These sites are concentrated at or near the confluence of a tributary and a major stream or overlooking an old meander. They are easily recognized by very dark, gray-brown dirt accumulated in horizontal sheet "middens" (dirt and trash left by humans) or small (20-m diameter) midden mounds. These dark brown middens contrast clearly with the normal reddish yellow soils. In contrast to the normally acid local soils, these midden accumulations are quite basic and contain perishable artifacts not preserved in the normal soils. Ash, charred nut hulls, charcoal, mussel shells, animal bone, and other organic trash have combined with colluvial silt to produce these unusual sites. This fertile soil also permits the growth of different vegetation; in particular, trees such as black locust and Bois d'Arc help mark these sites. Early investigators interpreted these intensive accumulations of trash as evidence that the people who caused them lived on and around the middens and occupied the same sites over a period of years. Unfortunately, the early workers were unable to identify stratigraphy clearly, to separate cultural components, or to recognize seasonal use.

The initial archaeological work under the auspices of the Works Projects Administration (WPA) ended with the beginning of WWII, and, except for a few preliminary descriptions, the recovered tools and debris from the extensive excavations remained unanalyzed for some time. Nevertheless, a construct (model) known as the Fourche Maline Focus was described in the latter part of the 1940s; it encompassed the transition from a hunting-gathering lifeway to a pottery-making, horticultural way of life (Krieger, 1946). Sites of this construct were recognized by the presence of the "black midden mounds" and were believed to be spread over a large area of southeast Oklahoma and northeast Texas. A few short surveys and several moderate excavations in the Wister Valley in the late 1940s revealed a wealth of sites and materials, but did little to explicate the transitional cultures (Watson, 1947; Bell and Baerreis, 1951; Bell, 1953).

A decision by the U.S. Army Corps of Engineers to increase the permanent pool of Wister Lake created the opportunity to use more recently developed techniques of excavation and analysis. A phased program of site discovery, initial testing and evaluation, and data recovery was carried out between 1974 and 1978 (Neal and Mayo, 1974; Mayo, 1975; Galm, 1978a,b; Galm and Flynn, 1978). This work confirmed that the sites, especially the midden mounds, were occupied serially, probably seasonally, and many had occupations spanning the period from the Late Archaic into the early pottery-using Woodland cultures. Further, it was recognized that pottery making was a dimension added to a way of life already developed by the Late Archaic Period, and that these cultures represented a long-term use of the area between 1500 B.C. and A.D. 700–800 by the same people, ancestors and descendants. This chronology was confirmed through some 74 radiocarbon dates from stratified contexts at eight sites. Bell (1980) proposed that this transitional period be divided between the prepottery people of the Late Archaic Period and the early pottery makers of the Woodland Stage.

The people of the Late Archaic Period are identified as the Wister Phase in the accompanying table. (The term "Phase" is used to identify a group of people using a distinct, shared set of cultural materials and way of life during the same time). These were the non-pottery-using people that occupied the valley at least from 1500 B.C., and developed the basic pattern for exploitation of plants, animals, and minerals that came to mark the cultures of the region. Non-ceramic components are often the lower portions of the midden-mound accumulations. The numerous sites indicate a sizable population that frequently reused the same locales on a seasonal basis, probably in the fall (Galm, 1981, 1984). Paleoclimatic data indicate that the climate of the Late Archaic period was wetter and slightly warmer than at present. Such conditions facilitated the recovery of forest vegetation after the 3,000-year interval of aridity attributed to the Altithermal climatic event of the mid-Holocene (7,000 to 4,000 years ago) (Albert and Wyckoff, 1984). The more amenable climate provided an improved food/water resource base, which supported a larger population.

Wister Phase people made compound short spears (darts) that were used also as knives; scrapers; ground-stone atlatl (spear/dart thrower) weights; bone-atlatl spurs; pear-shaped grubbing hoes attached to short wooden handles; occasional ground-stone

grooved axes; and grinding bowls and mullers to process seeds. Other items made of bone included antler and bone knapping tools, bone "hair pins" or bodkins, mat- and basketry-manufacturing tools, and shell and bone ornaments. Deer, fish, mussels, terrapin, and other small game were gathered and eaten. Many plants, particularly hickory, were exploited, and the debris from seasonal processing of plants is believed to be the major contributor of the organics that formed the middens. The primary season of use, which can be identified by the remains of the deer and plants, was late summer and early fall.

Sometime between A.D. 1 and A.D. 300, the Fourche Maline Phase of the Arkansas River Period replaced the Wister Phase. The Fourche Maline Phase was marked by the introduction of pottery making. Possibly, limited horticulture of plants like *Chenopodium* and *Amaranthus* became part of this culture at about the same time, but slight difference in seed sizes is the sole evidence. The people of the Fourche Maline Phase continued to use the earlier Wister Phase sites. Thicker, more organically enriched midden soils may indicate that they lived at the sites longer, or returned to them more often. This, in turn, may reflect another increase in population. It is apparent, however, that the Fourche Maline Phase people continued the way of life of the Wister Phase people, with a few added features.

There are some stylistic changes in the material culture that attest to the passage of time and to the cardinal rule of cultures: they always change. For these early pottery makers, not only the pottery style changed; there also were changes (in frequency and even presence or absence) in other tools and ornaments in their culture. These changes occurred gradually, and it is by studying the changes that archaeologists identify distinct cultures. The Wister Phase people used many expanding stemmed-corner-notched dart-points and fewer contracting-stemmed dart-point/knives. Over time, the relative frequency of the forms reversed until the contracting-stemmed form became commonplace, and the corner-notched points, rare. Bone tools and ornaments, some more elaborate than before, continued to be manufactured. The pear-shaped hoe became double-bitted. The ground-stone tools included pitted stones that were anvils, and they were often made on mullers. Their use extended back into the Wister Phase, and it has been suggested that they were used, at least in part, as anvils to crack hickory nuts for processing. With only a few additions to the cultural inventory (better and slightly more elaborate pottery and the bow and arrow), this cultural pattern remained essentially the same until the introduction of the Mississippian cultural pattern (Galm, 1984).

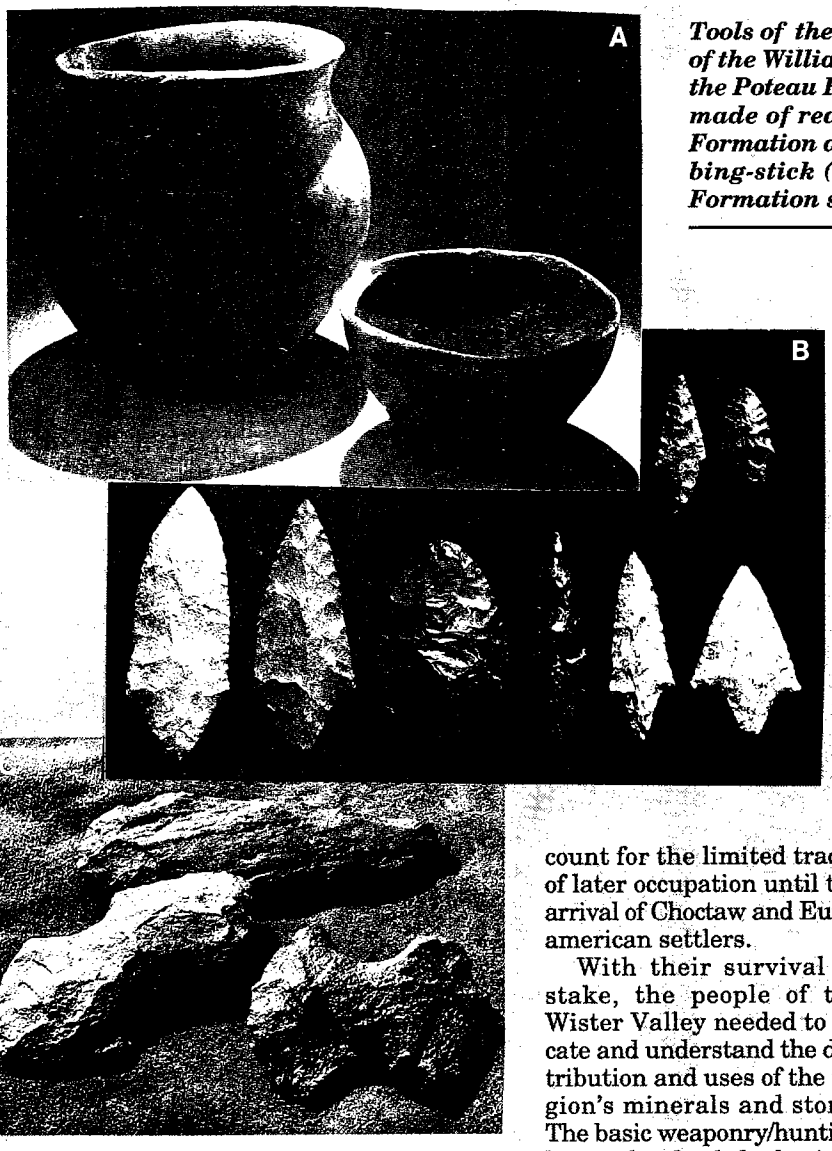
Archaeological Chronology in the Wister Valley
(after Galm, 1984)

Time	Phase	Period	Stage
1830	Historic	Choctaw	Indian Territory/ Removal Euroamerican
1700	Caddo/Wichita	Exploration	
	Fort Coffee	Coalescent	Ethnographic
1400			
	Spiro	Habi'ukut	Mississippian
1200	Harlan		
1000	Evans		
700			
300 A.D.	Fourche Maline	Arkansas River	Southern Woodland
B.C.	Wister	Late Archaic	Archaic
1,500	Tom's Brook	Middle Archaic	
6,000	Dalton	Early Archaic	
8,000			
12,000 +	?	Paleoindian	Early Foraging

This Mississippian Stage in the Arkansas Basin is referred to as the Habi'ukut Period. This period lasted for about 600 years (A.D. 800–1400), and included three recognizable cultures or phases: the Evans, Harlan, and Spiro Phases. The Mississippian pattern featured agriculture and a ranked society with cultural elites (Bell, 1984; Brown, 1984). These elites led the society through prestige engendered by supposed power over nature (they knew the proper ceremonies to grow corn) and prowess in warfare (Brown, 1984). They lived in and around ceremonial/civic centers, such as Spiro Mounds, and participated in a rather widespread ceremonial/exchange network known as the Southern Cult, which evidently also enhanced their power and prestige. The elites' right to command the labor to construct earthen burial and temple mounds, as well as differences in burial elaboration and in nutrition (reflected in the skeletons) between the elites living at the ceremonial centers and the outlying populations, all support the existence of a social gap. There is considerable discussion about what these people of the large Arkansas Basin were contributing to the Southern Cult, (bison, warfare alliances?), but the role of the people of the Wister Valley in this period was primarily economic support for the elite. To Spiro went the fruits of the hunt and the gathered resources, while the common folk ate a lot of corn and little meat.

There are a few clearly Mississippian sites in the Wister Valley, such as Victor Landing and Holson Creek, where status burials and artifacts link the valley to the ranked society at the Spiro Mounds. However, many of the sites of the Mississippian Stage located in the Wister Valley consist of thin veneers on the older Fourche Maline middens, good evidence for very transitory use in ways different from the Fourche Maline Phase. The dominant site types of the Fourche Maline are the midden mounds with frequent and long-term seasonal reoccupation, whereas the Mississippian occupants' settlements are concentrated at major stream junctures and higher terraces overlooking fertile flood plains suitable for low-technology agriculture (Galm, 1981). After about A.D. 1350, the mound makers' control in the Arkansas Basin seems to have collapsed. This loss of control possibly was the result of a drier period that expanded the grasslands (and bison) and made agriculture more unpredictable.

The Wister Valley seems to have been used only in a very transitory fashion after the end of the Spiro Phase, during a span described as the Coalescent Period, which lasted to



Tools of the Fourche Maline Phase: (A) early pottery of the Williams Plain type, made of clays mined from the Poteau River terraces; (B) gary points and knives made of redeposited cherts from the Johns Valley Formation and Atoka Formation siltstones; (C) grubbing-stick (short-handled hoe) tips made of Atoka Formation siltstone.

Historic contact. During this time, there are only a few traces of use that can be separated from the earlier Spiro Phase sites. If the people of the area became bison hunters primarily, as seems likely based on work along the Arkansas River, then it makes sense that the wooded terraces of Fourche Maline Creek and Poteau River would be used in a limited fashion. What this fashion was is not clear and is a matter for further research. We do know that at the time of LaHarpe's 1719 journey through the region to meet with the Wichita near Tulsa, the area was largely abandoned (Wedel, 1971,1978). The primary uses were for hunting and lithic-resource extraction by Caddos, Osage, and other people. Extended use of the Ouachita Mountains in this manner could account for the limited traces of later occupation until the arrival of Choctaw and Euroamerican settlers.

With their survival at stake, the people of the Wister Valley needed to locate and understand the distribution and uses of the region's minerals and stone. The basic weaponry/hunting kit, and indeed the basis of most of the tools used in the manufacture of the non-stone pieces, were derived from formations containing cryptocrystalline materials. The gravels of the streams leading into Fourche Maline Creek contain chert cobbles and pebbles suitable for making stone tools, but these are small in size and often fractured. The formations that were most important and that were exploited directly were: the Johns Valley Formation, with its residual cobbles of chert and silicified sandstone from many sources; a blue-gray silicified siltstone from the Atoka Formation; and a banded silicified sandstone from Winding Stair Mountain.

Projectile points, knives, drills, scrapers, woodworking tools, and hafted grubbing-stick/hoe blades were made by exploiting the brittleness and conchoidal fracture characteristic of these cherty materials. Many of the stone tools were used to create spear shafts, bows, arrow shafts, handles, fishhooks, canoes, and a variety of tools and ornaments of bone and wood. Some of the bone and wood tools were then used to make items such as nets and mats and baskets, from even more perishable materials, such as vegetal fibers. Conveniently shaped creek cobbles, hand sized and larger, composed of tough silicified sandstones or quartzite, were modified minimally and made into hafted axes. Stream-rolled pebbles were notched roughly to manufacture weights for nets to catch fish and small game. Sandstones, particularly coarser forms, were used to manufacture grinding bowls for processing plant materials and marrow-bearing animal bones, as well as to the manufacture ground-stone tools.

Other minerals were important to the people of the Wister Valley. The Stanley Shale provided soft to hard hematite and limonite concretions for pigment (and ground-stone tools if it was durable enough), and the Johns Valley provided phosphate nodules for drilling into beads and other small ornaments. Galena may have been obtained from sources in the Ozarks. Asphalt, to be used as an adhesive, was collected from natural exposures, especially in the Kiamichi Valley south of the Wister Valley.

We don't know to what degree the indigenous people of the Wister Valley understood the geological concepts used by modern geologists. Their different culture and different relationship to their environment may have led them to different beliefs about why things were where they were. There is little doubt, however, that if you could have shown a piece of culturally important mineral or tool stone to one of these prehistoric people, he could have told you where to find more. From this standpoint, could we consider them among Oklahoma's early practical geologists?

STOP 3

Sedimentology of the Upper Part of the Atoka Formation (Pennsylvanian), Wister Lake Spillway, Le Flore County, Oklahoma

James R. Chaplin
Oklahoma Geological Survey

Location: SW¼NW¼NE¼SW¼ sec. 6, T. 5 N., R. 25 E., Le Flore County

Permission to enter the spillway must be obtained at the U.S. Army Corps of Engineers' office located in Wister State Park.

SIGNIFICANCE

- spectacular exposures of the Atoka Formation (upper part), a stratigraphic interval rarely exposed
- superb sedimentary structures, particularly ripple bed forms that represent a range of flow velocities and directions
- overall thick, fining-upward transgressive shale sequence containing thin, coarsening-upward regressive sandstone sequences
- well-exposed tear fault on northern flank of the Heavener anticline that can be examined directly
- storm-influenced, mud-dominated shelf depositional setting
- first published petrographic or lithofacies study, of either a local or regional nature, on the upper part of the Atoka Formation in Oklahoma.

GEOLOGIC SETTING

Wister State Park is located in the southern part of the Arkoma Basin. The most prominent structural feature within the park is the Heavener anticline, which trends westerly across the area and extends well beyond the park boundaries. Wister Dam is located on the northern flank of the Heavener anticline where beds have a northward dip averaging ~26° and an average strike of N. 68° W. The Heavener anticline is broken on both flanks by numerous tear faults. One of these faults can be observed at this stop along the west side of the spillway where surface rocks (Unit 13) are offset ~10 ft (Fig. 7).

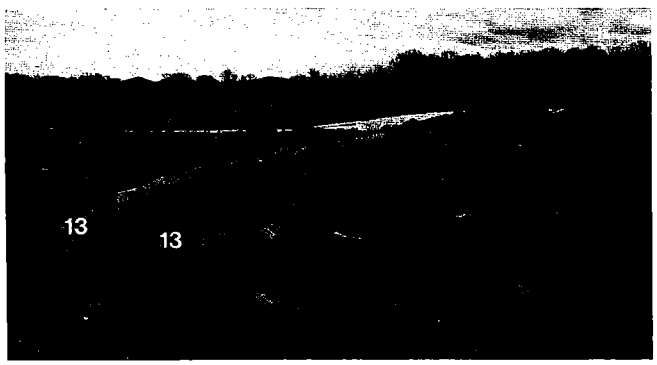


Figure 7. West side of Wister Dam spillway where a 3.3-ft.-thick sandstone bed (Unit 13, 199–202.3 ft interval) is offset ~10 ft by a fault. The Heavener anticline is broken on both flanks by numerous faults, but these faults rarely have surface expressions.

GENERAL STRATIGRAPHY

The rocks that crop out in Wister State Park are assigned to, in ascending order, the Atoka Formation (Atokan Series?) (~11,000 ft exposed in the vicinity of the park) and the Hartshorne (250–300 ft) and McAlester (900–950 ft) Formations (Desmoinesian Stage). Lithologically, the succession is composed dominantly of dark gray (N3) to medium dark gray (N4), noncalcareous, silty shale and subordinate amounts of ripple-marked, locally cross-bedded, very fine grained sandstones, some very discontinuous (Fig. 8). A detailed lithologic description of the upper part of the Atoka Formation is given in Measured Section, Stop 3. Figure 8 shows unit thicknesses, thin-section and shale-sampling intervals, lithologic succession, depositional textures, level of bioturbation, sedimentary features, and thin-section photomicrographs from certain intervals.

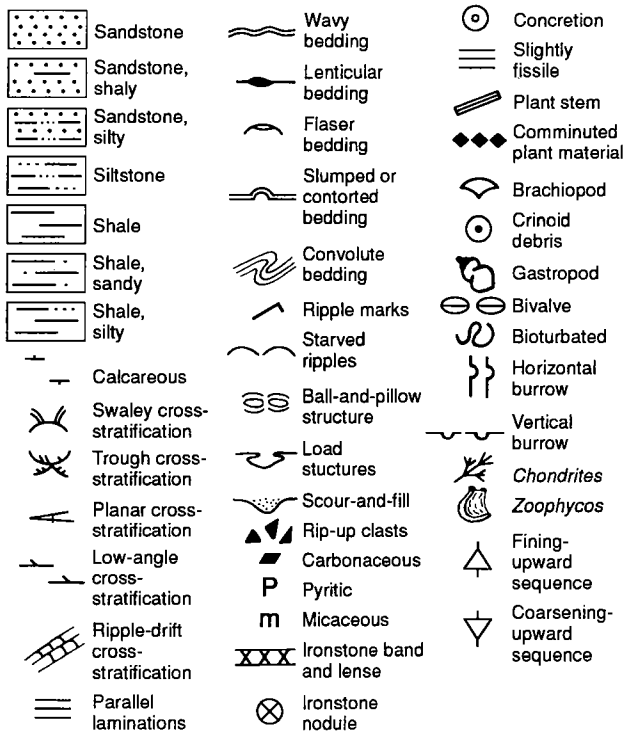


Figure 8 *(p. 15–22)*. Diagram of measured stratigraphic section of the Atoka Formation (upper part) at Wister Dam spillway. Unit number and thickness, lithology, depositional texture, level of bioturbation, and sedimentary features are shown. Explanation of symbols is shown above.

The approximate stratigraphic positions from which shale samples (*) and thin sections (e.g., 26-1) were taken are marked. Thin sections have two numbers: the first number identifies the unit from which the sample was collected, and the second number designates the stratigraphic position of the sample within that particular unit (e.g., of the three thin sections collected from Unit 26, 26-1 was collected from the lowest stratigraphic position within that unit). Not all thin sections are represented by photomicrographs.

The level of bioturbation may vary slightly throughout any unit; therefore, the assigned ichnofabric index only represents the general overall intensity of bioturbation throughout that particular unit.

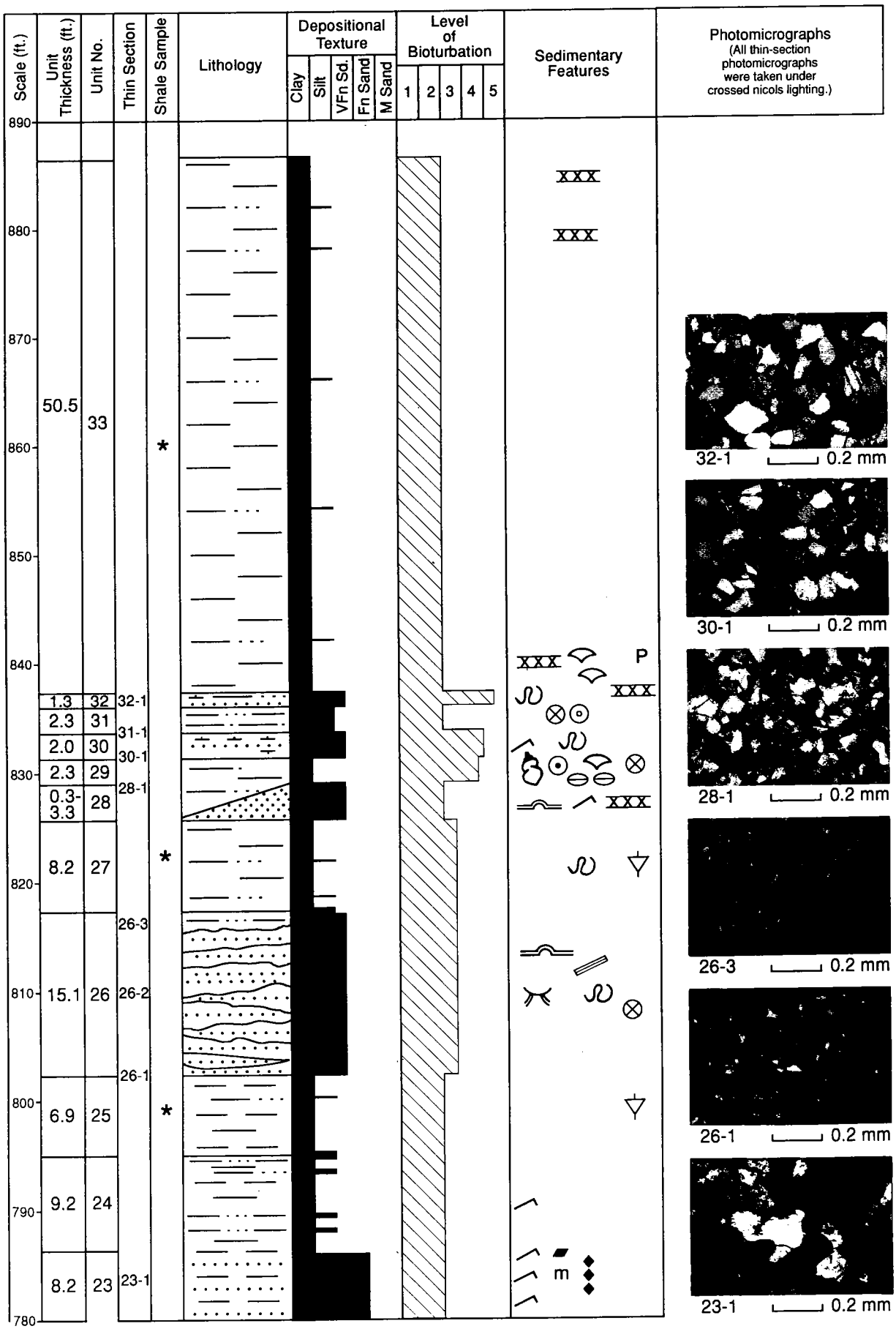


Figure 8 *(continued)*.

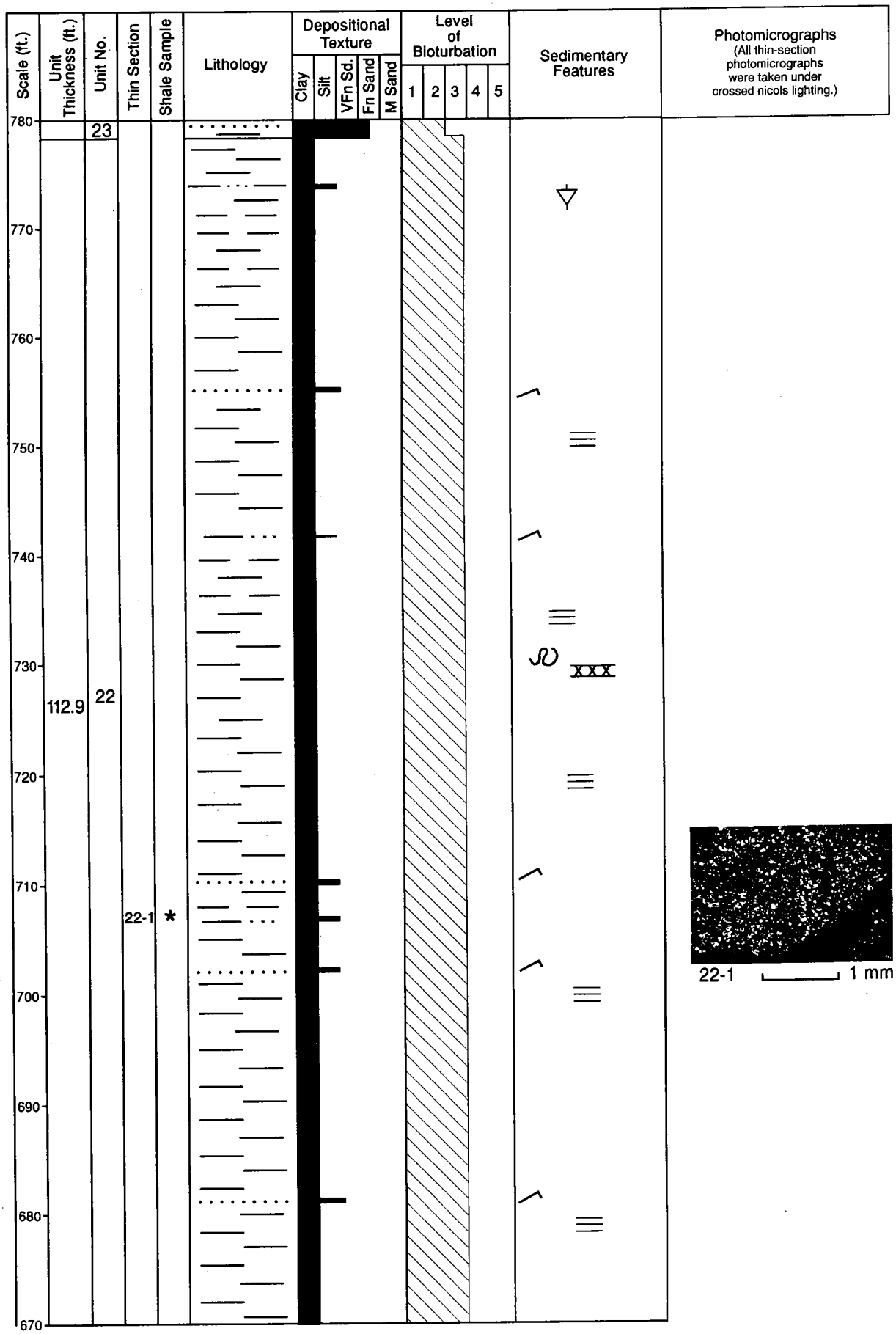


Figure 8 *(continued)*.

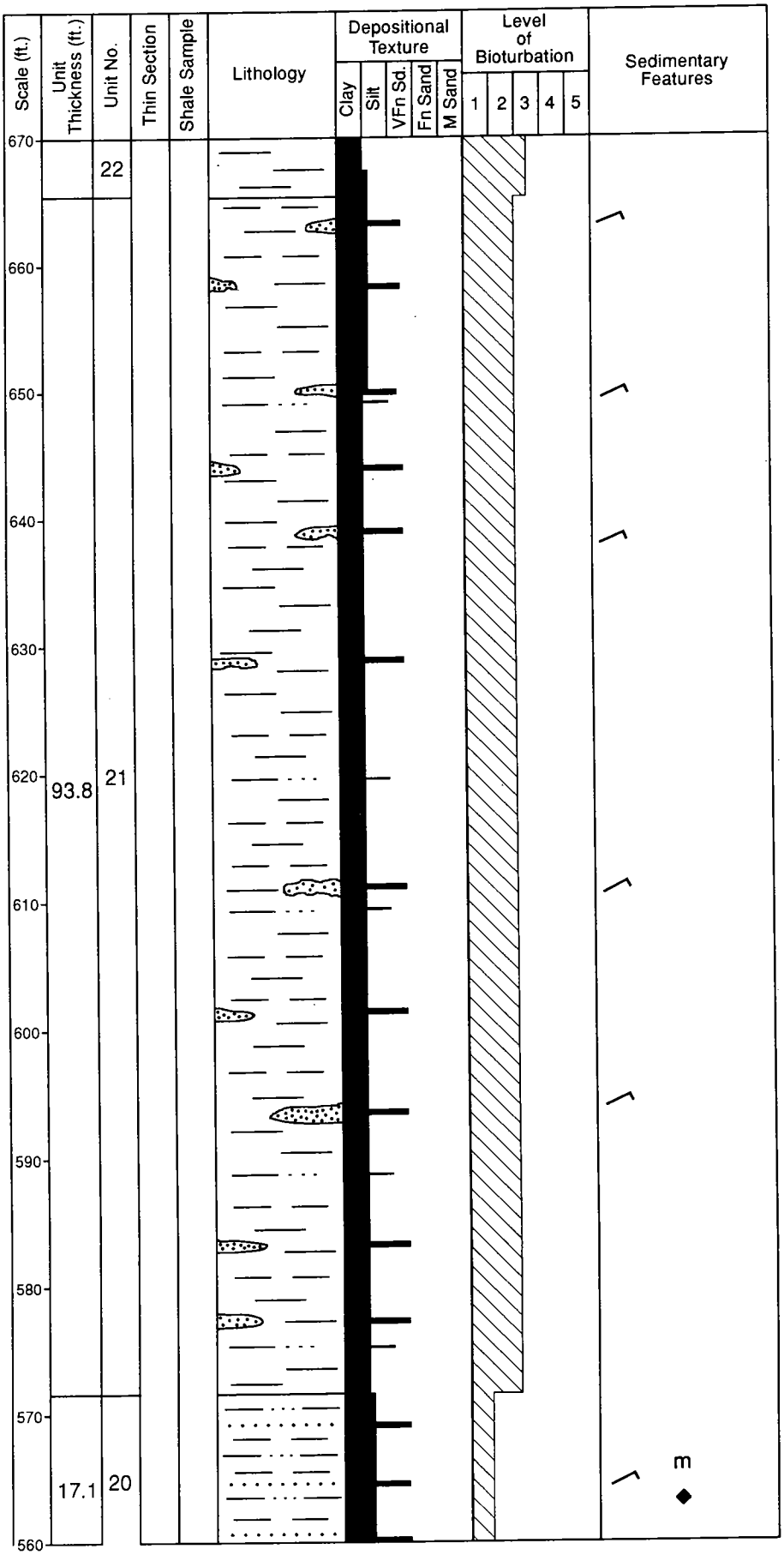


Figure 8 *(continued)*.

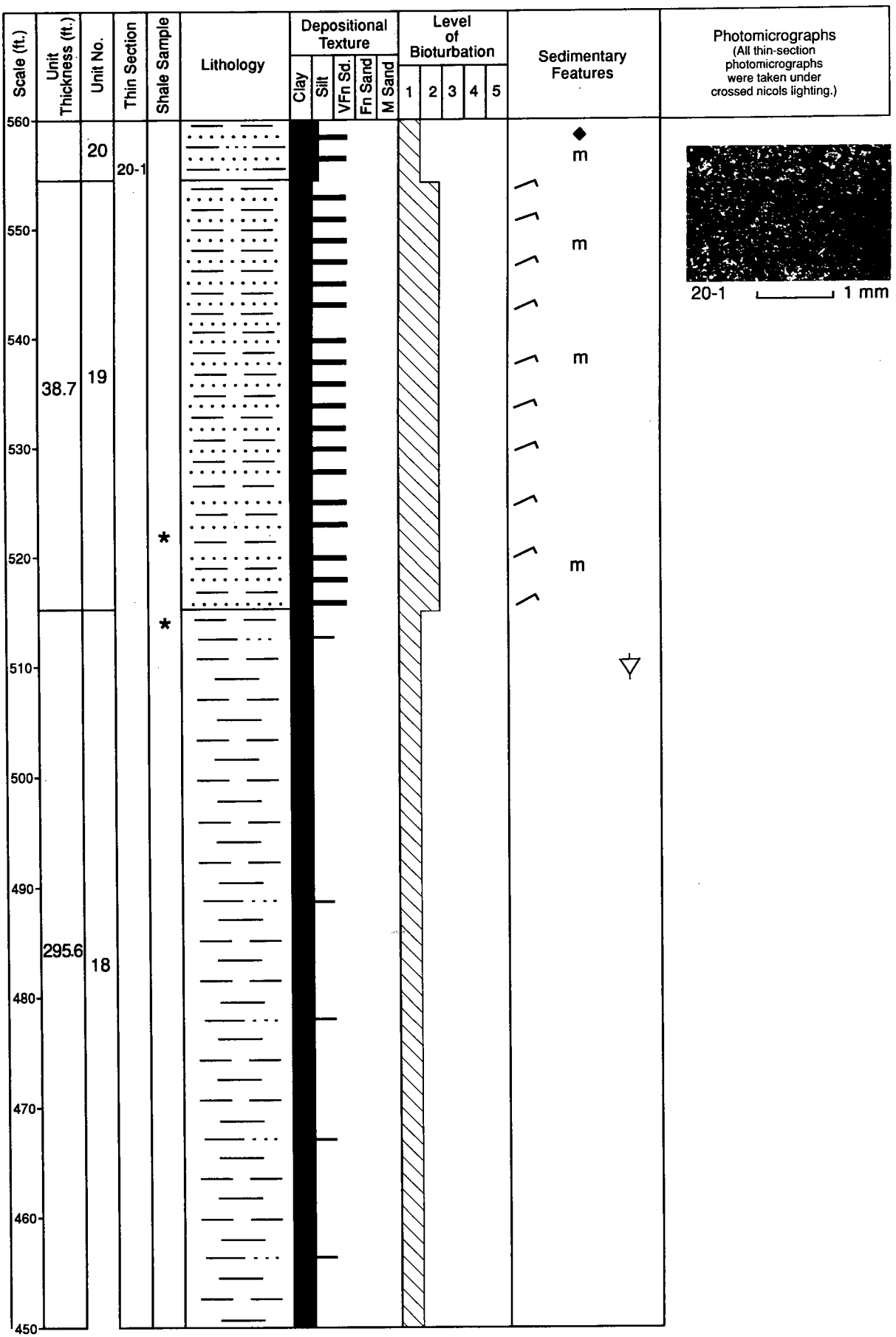


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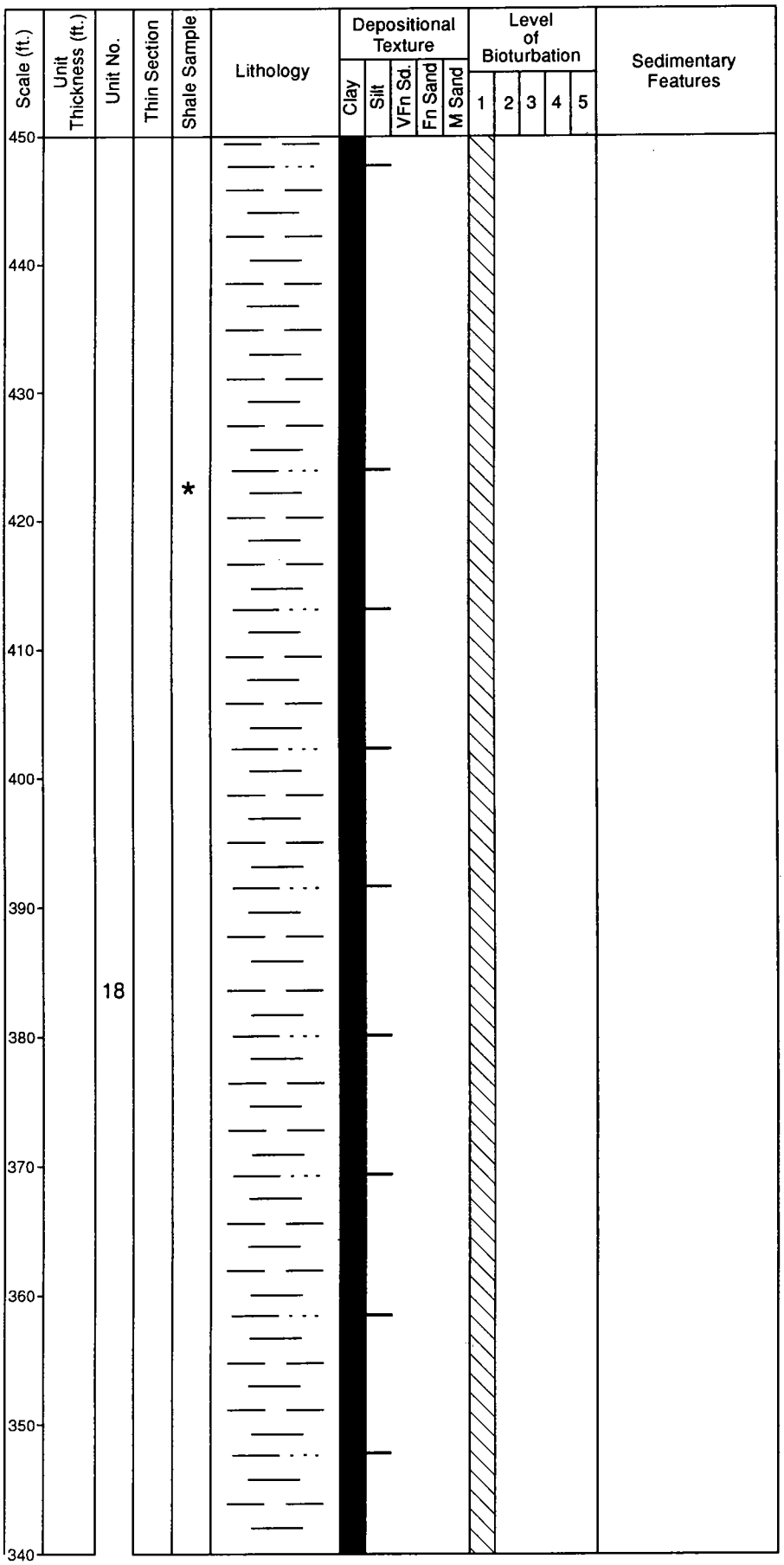


Figure 8 *(continued).*

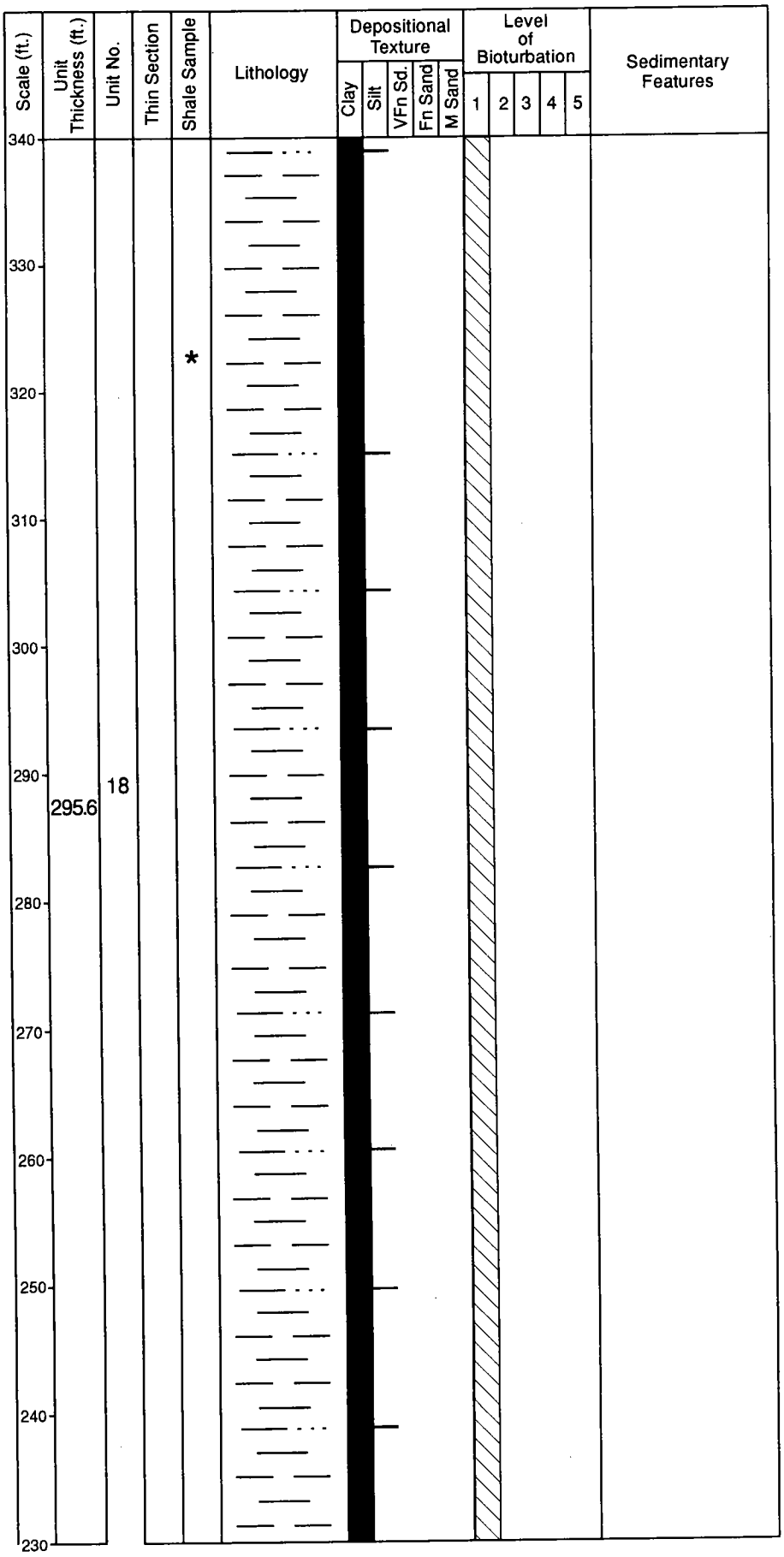


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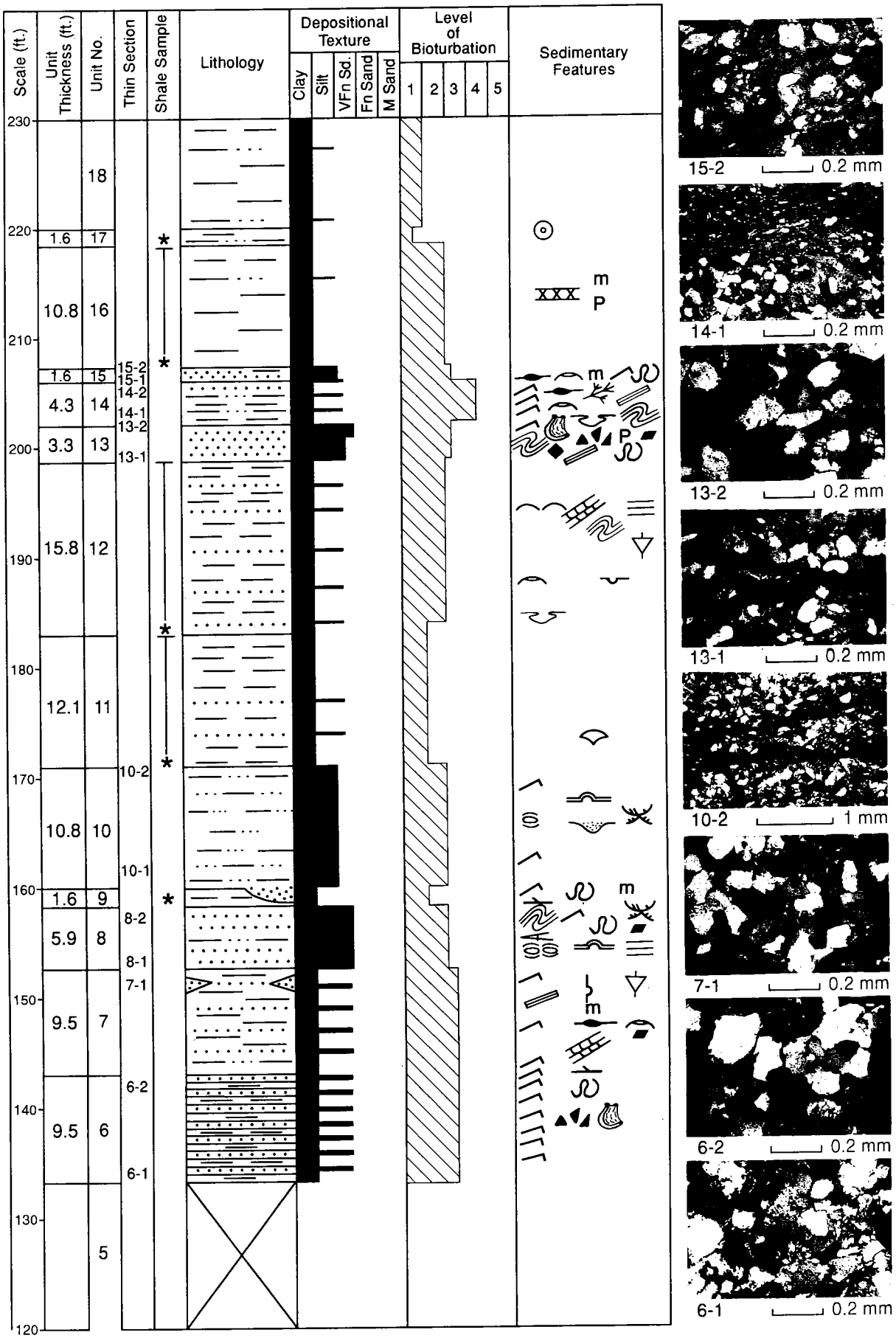


Figure 8 *(continued)*.

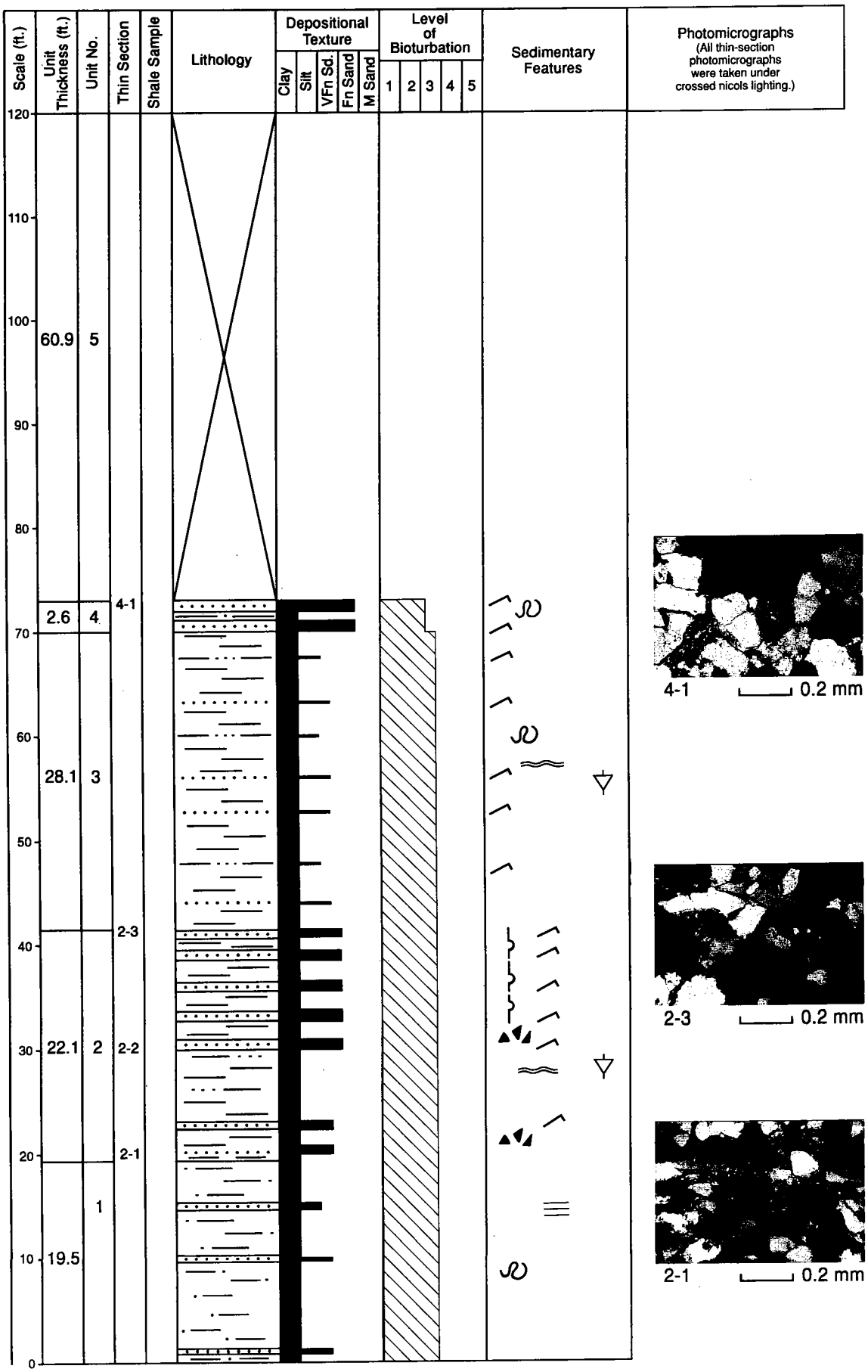


Figure 8 *(continued).*

Measured Section, Stop 3

Atoka Formation (upper part)

SW¼NW¼NE¼SW¼ sec. 6, T. 5 N., R. 25 E., Wister quadrangle, Le Flore County. Measured from base of exposure just south of spillway overlook northward below spillway to gate at lower end of cut. Section measured perpendicular to east-west strike. Average strike N. 68° W., average dip 26° north-northeast. (Measured by James Chaplin, LeRoy Hemish, and Mark Keys.)

Thickness (feet)

33. Shale, dark gray (N3) to very dusky red purple (5RP 2/2), silty, noncalcareous; contains some dark reddish brown (10R3/4) ironstone layers at top of unit; includes a 2–4 in. thick ironstone layer ~1 ft above the base containing abundant chonetid- and spiriferid-type brachiopods; some brachiopods are pyritized, others are calcareous; weakly (2) bioturbated; gradational base; nonresistant unit; top not exposed (shale sample A 33-1, middle of unit) 50.5
32. Interstratified medium gray (N5), very fine grained sandstone and siltstone, very shaly; calcareous in certain intervals; contains very abundant trace fossils; strongly (4) bioturbated to churned (5), top weathers to shaly siltstone with abundant brachiopods; gradational base; (thin section sample A 32-1) 1.3
31. Shale, dark gray (N3), sandy to silty, noncalcareous; includes some 1–2-in.-thick siltstone/sandstone layers and dark reddish brown (10R3/4) and dark yellowish orange (10YR6/6) ironstone lenses/nodules/concretions; weakly (2) bioturbated; gradational base (thin section sample A 31-1, very fine grained sandstone/siltstone stringer ~1 ft above base) 2.3
30. Interstratified medium light gray (N6) and medium dark gray (N4) siltstone/sandstone and grayish purple (5P4/2), silty shale, calcareous; highly ripple bedded; sandstone, very fine grained, noncalcareous; moderately (3) to strongly (4) bioturbated; gradational base; resistant unit; (thin section sample A 30-1, at base of unit) 2.0
29. Shale, medium dark gray (N4), sandy to silty, soft, noncalcareous; very fossiliferous, particularly at top with brachiopods, crinoidal debris, bivalves (some encrusting), and gastropods; includes some dark reddish brown (10R3/4), ironstone nodules; moderately bioturbated (3); base sharp; nonresistant unit; (shale sample A 29-1, continuous) 2.3
28. Sandstone, light bluish gray (5B7/1), medium light gray (N6), noncalcareous, very fine grained, thin- to medium-bedded, rippled; includes rare soft-sediment deformation feature (contorted bedding); contains some dark reddish brown (10R3/4) ironstone bands/lenses; weakly (2) bioturbated; irregular base; lenticular sandstone body ~26.3 ft wide; resistant unit; (thin section sample A 28-1, rippled sandstone lens near top) 0.3–3.3
27. Shale, medium bluish gray (5B5/1) and medium dark gray (N4), silty, noncalcareous; includes minor siltstone stringers; weakly (2) to moderately (3) bioturbated; weathers to flakes/chips; nonresistant unit; sharp base; (shale sample A 27-1, middle of unit) 8.2
26. Sandstone/siltstone, medium light gray (N6), light bluish gray (5B7/1), noncalcareous, very fine grained; grades into siltstone at top; irregular bedded; some small-scale soft-sediment deformation features; shaly near top; rare swaley cross-stratification; contains trace fossils, plant fossils, and ironstone nodules; fractured; dark yellowish orange (10YR6/6) staining on outcrop surface; weakly (2) to moderately (3) bioturbated in certain intervals; sharp base; very resistant unit; (thin section sample A 26-1, base of unit; A 26-2, middle of unit; A 26-3, just below top of unit) 15.1
25. Shale, medium bluish gray (5B5/1) and medium dark gray (N4), noncalcareous, silty, soft, weathers smoothly; includes some siltstone stringers/lenses; weakly (2) bioturbated; gradational base; (shale sample A 25-1, middle of unit, continuous) 6.9
24. Interstratified dark gray (N3) to grayish purple (5P 4/2), noncalcareous, micaceous, silty shale (60%) and shaly siltstone/sandstone (40%); 1–2-in.-thick, rippled, shaly siltstone/very fine grained sandstone beds throughout; weakly (2) bioturbated; gradational base 9.2
23. Interstratified silty shale (50%) and shaly siltstone/sandstone (50%); shale, dark gray (N3), silty, noncalcareous; sandstone, medium gray (N5) and grayish red purple (5RP4/2), noncalcareous, fine- to very fine grained, dominantly fine grained, rippled, highly micaceous and rippled unit; contains very abundant comminuted plant material; top of unit contains wavy, discontinuous, dark gray (N3) carbonaceous laminations; weakly (2) bioturbated; gradational base; (thin section sample A 23-1, rippled sandstone 4.9 ft above base) ... 8.2
22. Shale (80%), dark gray (N3), noncalcareous, silty, slightly fissile; includes rippled shaly siltstone (20%) laminations/stringers; contains more silty shale than Unit 21; includes rare ironstone stringers; weakly (2) to moderately (3) bioturbated; gradational base; base of unit approximately at end of concrete spillway apron; (shale sample A 22-1, 41.6 ft above base; thin section sample A 22-1, siltstone stringer 41.6 ft above base) 112.9
21. Shale (70%), dark gray (N3) and brownish gray (5YR 4/1) and shaly siltstone/sandstone (30%) stringers; shale, noncalcareous, micaceous, sandy; sandstone, noncalcareous, very fine grained, rippled, occurs mainly as lenses; white (N9) oxidized staining on outcrop surface; weakly (2) bioturbated; gradational base 93.8
20. Siltstone/silty shale (60%) and very fine grained sandstone (40%); shale, medium dark gray (N4) and grayish red (10R4/2), micaceous, noncalcareous except in certain intervals; includes some rippled, light gray (N7) sandstone streaks; contains some comminuted plant material; weakly (2) bioturbated to nonbioturbated; gradational base; (thin section sample A 20-1, 1.4 ft above base) 17.1
19. Alternating beds of grayish black (N2), micaceous, noncalcareous, silty shale (50%) and grayish red purple (5RP4/2) and brownish gray (5YR4/1), very fine grained sandstone/siltstone (50%); rippled sandstone/siltstone beds occur in 1–2-in.-thick layers/lenses; weakly (2) bioturbated; noncalcareous except where micro-size calcite crystals coat shale surfaces; gradational base; (shale sample A 19-1, 6.2 ft above base) 38.7
18. Shale, dark gray (N3), and brownish gray (5YR4/1), silty, noncalcareous, monotonous shale sequence; poor to moderate exposure due to slump blocks;

weakly (2) bioturbated to nonbioturbated; nonresistant unit; gradational base; offset to the east of spillway floor to continue measurements; (shale sample A 18-1, just below contact with Unit 17; A18-2, ~102.7 ft above base; A18-3, ~202.4 ft above base; A 18-4 top of unit). 295.6

17. Shale, brownish gray (5YR4/1), silty, noncalcareous; deeply weathered limonitic concretionary horizon; contains abundant dark yellowish orange (10YR6/6) and dark reddish brown (10R3/4) ironstone concretions; spheroidal-like weathering; nonbioturbated; nonresistant; gradational base .. 1.6
16. Shale, grayish black (N3), brownish black (5YR2/1), and grayish red purple (5RP4/2), silty, micaceous, noncalcareous; includes some grayish red (10R 4/2) ironstone stringers/bands/ lenses, some with pyrite; weakly (2) bioturbated; nonresistant unit; sharp base; (shale sample A 16-1, continuous) 10.8
15. Sandstone/siltstone, medium light gray (N6), noncalcareous, micaceous, very fine grained, slabby, rippled, parallel-bedded; weathers dark yellowish orange (10YR6/6); interbedded with pale red (10R 6/2) rippled sandy shale/shaly sandstone showing small-scale ripple, lenticular and flaser bedding; ripple types include a mixture of current and interference ripples but current ripples are more common; unit becomes more shaly at the top; weakly bioturbated (2) to moderately (3) bioturbated in certain intervals; gradational to sharp base; (thin section A 15-1, just above base of unit; A 15-2, top of unit) 1.6
14. Interstratified shaly sandstone/sandy shale grading into very fine grained sandstone at top; small-scale heterolithic bedding; similar in lithology to Unit 12; shaly sandstone/sandy shale alternating with siltstone in 1–4-in.-thick beds; siltstone, light gray (N7), noncalcareous, occurs as streaks and rippled lenses; sandy shale, medium light gray (N6) and medium gray (N5); wavy and lenticular bedded; includes dark reddish brown (10R3/4) ironstone beds/lenses; highly rippled unit; certain intervals contain small-scale flaser/lenticular bedding and soft-sediment deformation features (convolute bedding, load casts); contains well-preserved, slightly coalified *Calamites* stems, some 30 in. in length; moderately (3) to strongly (4) bioturbated, contain some *Chondrites*; gradational to erosional base; (thin section sample A 14-1, ~1 ft above base of unit; A 14-2, top of unit) 4.3
13. Sandstone, yellowish gray (5Y7/2) weathers dark yellowish orange (10YR6/6), noncalcareous, micaceous, very fine grained; highly rippled unit, interference ripples dominant; convolute bedding at base; includes dark reddish brown (10R3/4) shale rip-up clasts, particularly near the top; some pyrite nodules with white (N9) halos; contains dark gray (N3) comminuted plant material and rare scoured *Calamites*; top highly rippled, contains indistinct scoured *Zoophycos*; weakly (2) to moderately (3) bioturbated, contains horizontal grazing/feeding trails; single very resistant bed; unit offset ~10 ft by faulting along west side of spillway; erosional base; (thin section sample A 13-1, base of unit; A 13-2, top of unit). 3.3
12. Shale, medium gray (N5), brownish gray (5YR4/1) and medium bluish gray (5B5/1), noncalcareous, very sandy, slightly fissile; small-scale heterolithic bedding; includes some 1–4-in.-thick, dark reddish brown (10R3/4), very fine grained, noncalcareous sandstone/siltstone stringers/lenses; starved ripples, flaser bedding, and truncated sets of small-scale ripple-drift laminations in certain intervals; some pin-stripe millimeter-thick laminations; certain intervals of very fine grained sandstone show small-scale soft-sediment deformation features including flame structure, convolute bedding and load casts; outcrop surface has white (N9) oxidized blotches; nonresistant unit; weakly (2) bioturbated; gradational base; (shale sample A 12-1, continuous) 15.8
11. Shale, dark gray (N3), brownish black (5YR2/1) and dark reddish brown (10R3/4), noncalcareous; includes light gray (N7) and grayish red purple (5RP 4/2) millimeter-thick very fine grained sandstone/siltstone laminations/streaks; vertical plug; sandstone includes delicate, finely ribbed, thin-shelled brachiopods; weakly (2) bioturbated to nonbioturbated; nonresistant unit; gradational base; (shale sample, A 11-1, continuous) 12.1
10. Interstratified siltstone/sandstone and sandy shale; heterolithic bedded unit; highly contorted unit; siltstone, light gray (N7) and pale red (10R6/2) weathering dark yellowish orange (10YR6/6); flat bedded, thin-bedded; dominantly siltstone unit contains some rippled sandy shale/shaly sandstone; small- and large-scale soft-sediment deformation features include slumped sandstone bodies, contorted bedding, and ball-and-pillow structures; some scour channels filled with high-angle trough cross-stratified sandstone; rapid lateral changes in facies/thickness; weakly (2) bioturbated; gradational to erosional base; (thin section sample A 10-1, ~1.4 ft above base; A 10-2, top of unit) 10.8
9. Shale, medium dark gray (N4), brownish gray (5YR 4/1), and medium bluish gray (5B5/1), sandy, noncalcareous; includes light gray (N7), very fine grained, rippled sandstone lenses/streaks; slightly micaceous; shale unit pinches out laterally due to scouring by Unit 10; weakly (2) bioturbated to nonbioturbated; nonresistant unit; gradational base; (shale sample, A 9-1, continuous) 1.6
8. Sandstone, light gray (N7), very fine grained, micaceous, noncalcareous; weathers dark yellowish orange (10YR6/6); some siltstone in certain intervals; highly deformed unit; soft-sediment deformation features include pinch-and-swell bedding, ball-and-pillow structures, and convolute bedding; highly rippled unit, current and interference ripples most common; small-scale truncated sets of low-angle trough cross-stratification, particularly at the top; interlaminated, rippled sandy shale in certain intervals; rare low-angle planar cross-stratification; faint horizontal, even parallel, laminations in certain intervals where flat-bedded; some carbonaceous streaks/laminae; weakly (2) bioturbated; unit highly variable in facies/thickness laterally; base sharp, planar to highly contorted; (thin section sample A 8-1, base of unit; A 8-2, top of unit) 5.9
7. Interstratified sandstone/siltstone and sandy shale, grayish red (10R4/2) and medium light gray (N6), noncalcareous; highly rippled unit; includes 1–2-in.-thick, very fine grained sandstone beds interlaminated with sandy shale; lenticular sandstone beds more common in upper 2.0 ft; micaceous, very carbonaceous, contain some comminuted plant material (*Calamites*); certain intervals con-

tain flaser/lenticular bedding and small-scale ripple-drift cross-stratification; weakly (2) to moderately (3) bioturbated, contains dominantly horizontal grazing trails; base gradational; (thin section sample A 7-1, ~1.4 ft below top of unit) 9.5

6. Sandstone, medium light gray (N6), fine- to very fine grained, noncalcareous; weathers to dark yellowish orange (10YR6/6); contains sandy shale partings/streaks; dominantly a highly rippled unit in which stacked bedding plane surfaces show a diversity of ripple types; includes certain horizons of shale rip-up clasts and low-angle cross-stratification; individual sandstone beds 2–6-in.-thick; certain intervals contain very abundant, scoured *Zoophycos* on top bedding-plane surfaces; irregular- to ripple-bedded; rippled surfaces stained dark reddish brown (10R3/4); weakly (2) bioturbated, certain intervals moderately (3) bioturbated; very resistant unit, exposed at edge of concrete apron; base not exposed; (thin section sample A 6-1, basal sandstone bed exposed; A 6-2, top of wavy-bedded sandstone) 9.5
5. Covered interval (concrete apron of spillway) 60.9
4. Sandstone, light brown (5YR5/6) to moderate brown (5YR4/4), dominantly fine grained but contains certain intervals of very fine grained sandstone, noncalcareous, porous; interval consists of two ~1-ft-thick sandstone beds separated by 0.8-ft-thick light brown (5YR5/6), noncalcareous, sandy shale; impure, contains abundant clay-alteration products; thin-bedded; low-angle cross-stratification; surfaces ripple-marked; weakly (2) bioturbated; (thin section sample A 4-1, uppermost rippled sandstone bed) 2.6
3. Shale, light brown (5YR5/6) interbedded with light gray (N7), grayish orange (10YR7/4), and light brown (5YR5/6), very fine grained sandstone and siltstone; noncalcareous; thin- to very thin bedded; wavy, parallel-bedded; surfaces ripple-marked; weathers to small, broken fragments on the outcrop; weakly (2) to moderately (3) bioturbated; gradational base 28.1
2. Sandstone, light brown (5YR5/6), moderate brown (5YR4/4), and very pale orange (10YR8/2), rhythmically alternating with dusky yellowish brown (10YR2/2) sandy shale that weathers light brown (5YR6/4, 5YR5/6); sandstone, fine- to very fine grained, noncalcareous, thin- to medium-bedded; wavy, parallel-bedded; sandstone beds are generally about 8–15 in. thick, particularly in upper part; sandstone beds are more common and thicker in upper part; certain intervals contain shale rip-up clasts; surfaces marked with interference ripples; top bedding plane surfaces covered with trace fossils (horizontal locomotion/feeding trails); weakly (2) to moderately (3) bioturbated; gradational base; (thin section sample A 2-1, base; A 2-2, top of 5.2-ft-thick rippled sandstone bed; A 2-3 top of unit) 22.1
1. Shale, very pale orange (10YR8/2) to light brown (5YR5/6) to brownish gray (5YR4/1), slightly fissile, sandy, noncalcareous; includes some moderate brown (5YR4/4), very fine grained, bioturbated sandstone beds; base covered 19.5

Total thickness, including covered interval ~886.0

STRATIGRAPHIC POSITION/THICKNESS

The thickness of the upper part of the Atoka Formation exposed at this stop is ~886 ft. The Hartshorne/Atoka contact is exposed ~1.3 mi northwest of this stop on the east point of Wister Ridge, just east of the picnic area at Area 2 campground (Fig. 9). On Wister Ridge, the uppermost erosion-resistant sandstone in the Atoka Formation is ~475 ft below the Atoka/Hartshorne contact. The same resistant sandstone (Unit 26, Fig. 8) is exposed in the lower part of the spillway section below ~69 ft of silty, sandy shale. Therefore, ~406 ft of the uppermost part of the Atoka Formation is not exposed in the spillway section. However, Hemish (1993a, p. 19) described 113.5 ft of this 406-ft interval on Wister Ridge. Lithologically, the interval is composed predominantly of olive gray (5Y4/1) to medium gray (N5) silty shales that are interstratified with light olive gray (5Y5/2) siltstones in the upper 25 ft.

GENERAL DESCRIPTION OF SUCCESSION

The succession can be divided informally into the following general lithologic subdivisions (Fig. 8).

Lower 207 ft (Units 1–15)—Composed of four coarsening-upward lithofacies. The lowermost (finest grained) part of each coarsening-upward sequence begins with an approximately 10–20-ft-thick shale or (less frequently) siltstone/sandstone interval. The uppermost (coarsest grained) part of each lithofacies usually is dominated by 5–20-ft-thick, very fine grained sandstone/siltstone beds or, less commonly, by fine-grained sandstone (Fig. 10). This part of the overall succession has the greatest contrast in lithologies and contains the most abundant and diverse sedimentary features, including ripple marks, scour-and-fill marks, low-angle cross-stratification, trough cross-beds, planar cross-beds, parallel laminations, lenticular and flaser bedding, contorted bedding, convoluted bedding, load structures, ball-and pillow structures, rip-up clasts, comminuted plant material, plant stems, carbonaceous matter, pyrite, mica, trace fossils, and brachiopods (Fig. 8).

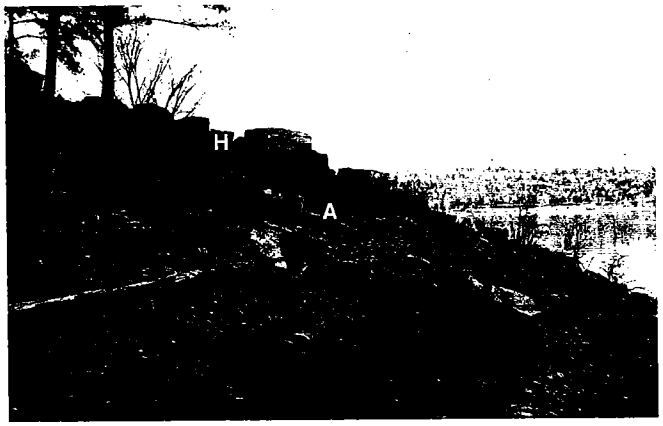


Figure 9. Contact of Atoka Formation (A) and Hartshorne Sandstone (H). Located at Area 2 campground in Wister State Park ~1.3 mi from spillway section (Fig. 8).

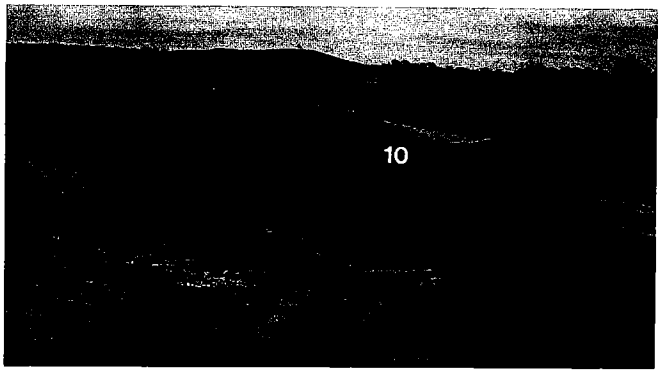


Figure 10. Northward dipping (26°) beds of the upper part of the Atoka Formation (133- to ~886-ft interval) exposed in Wister Dam spillway. Note the overall mud-dominated succession. Two small-scale coarsening-upward sequences can be seen, one in the immediate foreground and the other at about the middle of the photograph (Unit 10).

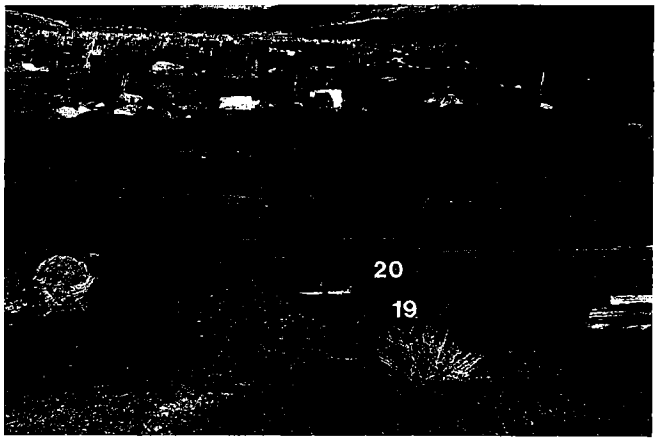


Figure 11. Lower part of mud-dominated part of succession. Contact of shale Unit 19 with shale Unit 20 (516–572-ft interval). Hoe pick head marks contact. Unit 20 is composed of medium dark gray (N4) and grayish red (10R4/2) siltstone and silty shale (60%) and 1–2-in.-thick, rippled, very fine grained sandstone streaks/lenses (40%). Hoe pick is 2.2 ft long.

Middle 207–516 ft (Units 16–18)—Consists mostly of a very thick, dark gray (N3), silty, noncalcareous, monotonous shale succession with streaks of siltstone and ironstone stringers locally. Sedimentary features are rare to absent in this interval. Ironstone bands, concretions, mica, and pyrite occur in the lower 10 ft.

Upper 516–886 ft (Units 19–33)—Composed of four coarsening-upward lithofacies.

516–572 ft (Units 19 and 20)—Represents the upper (coarsest grained) part of the lowest coarsening-upward lithofacies in this interval. Interval is composed of alternating beds of medium dark gray (N4) silty, micaceous, noncalcareous (except in a few intervals) shale and 1–2-in.-thick laminations/streaks/lenses of rippled, noncalcareous, very fine grained, light gray (N7) sandstone and siltstone (Fig. 11). Sedimentary features include ripple marks, mica, and comminuted plant material (Fig. 8). The lowermost part of this coarsening-upward succession is represented by Unit 18, a monotonous, 295.6-ft-thick shale sequence (Fig. 8).

572–787 ft (Units 21–23)—The lowermost (finest grained) part of this second coarsening-upward sequence is composed of 206.7 ft of dark gray (N3), silty, micaceous, slightly fissile, noncalcareous shale (65%) containing 1–2-in.-thick, light gray (N7), noncalcareous, rippled, very fine grained sandstone (35%) lenses and streaks. Sedimentary features include isolated ripple marks, bioturbation, and ironstone bands (Fig. 8).

The uppermost (coarsest grained) part of this coarsening-upward sequence is composed of 8.2 ft of interstratified dark gray (N3), noncalcareous, silty shale (50%) and medium gray (N5) and grayish red purple (5RP4/2), noncalcareous, micaceous, fine- to very fine grained (mostly fine grained) sandstone (50%) (Fig. 12). Sedimentary features include very abundant comminuted plant material, ripple marks, and wavy, discontinuous, carbonaceous laminations.

787–818 ft (Units 24–26)—The lowermost (finest grained) part of this third coarsening-upward lithofacies consists of 16.1 ft of interstratified dark gray (N3), noncalcareous, micaceous, silty shale (dominant) and 1–2-in.-thick beds of shaly siltstone and very fine grained sandstone. Sedimentary features consist mostly of ripple marks in the lower 9.2 ft.

The uppermost (coarsest grained) part of the lithofacies is composed of 15.1 ft of medium light gray (N6), noncalcareous, very fine grained sandstone. Sedimentary features, confined to the more sandy part of the interval, include small-scale, soft-sediment deformation features, rare swaley cross-stratification, ironstone nodules, trace fossils, and plant fossils.



Figure 12. Upper part of mud-dominated part of succession. Contact of shale Unit 22 with shale Unit 23 (770–787-ft interval). Note the increase in rippled shaly siltstone streaks/lenses in Unit 23. Unit 23 contains abundant comminuted plant material. Hoe pick is 2.2 ft long.

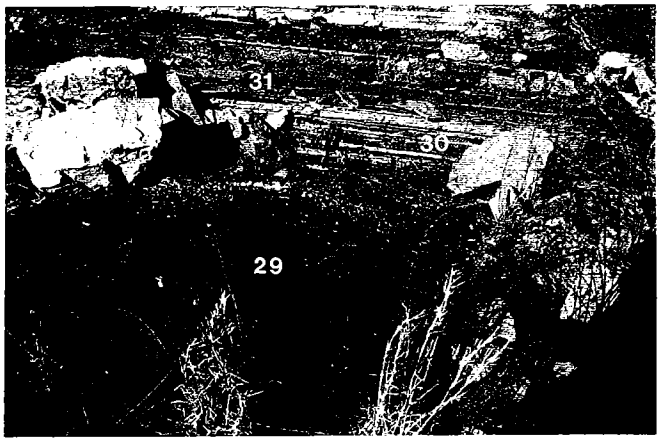


Figure 13. Contact of marine shale Unit 29 with interstratified siltstone/shale Unit 30 (830–834-ft interval). Unit 29 contains a good marine fauna of brachiopods, crinoid ossicles/stems, bivalves, and gastropods. Geologic pick is 1.1 ft long.

818–836 ft (Units 27–32)—The lowermost part of the fourth coarsening-upward sequence is composed of 12.4 ft of medium dark gray (N4), silty, noncalcareous shale (Fig. 8). A lenticular (0.3–3.3-ft-thick), very fine grained sandstone bed with contorted bedding, ironstone bands, and ripple marks occurs in this interval.

The uppermost (coarsest) part of this lithofacies is composed of 5.6 ft of mostly interstratified medium gray (N5), calcareous to noncalcareous, very fine grained sandstone and dark gray (N3), silty, noncalcareous, shale (Fig. 13). Sedimentary features include ripple marks, ironstone lenses/nodules/concretions, pyrite, contorted bedding, strong bioturbation, and marine fossils (e.g., brachiopods, crinoid ossicles, gastropods, and bivalves) (Fig. 8).

836–886 ft (Unit 33)—This 50-ft-thick interval represents only the lowermost (finest grained) portion of another, overlying coarsening-upward sequence. The interval is composed of dark gray (N3), silty, noncalcareous shale. Sedimentary features consist mostly of ironstone layers. One ironstone layer, near the base, contains an abundant brachiopod fauna (Fig. 8).

SEDIMENTOLOGY

Petrography

Twenty-seven thin sections from outcrop samples were prepared and studied to determine textural and mineralogical characteristics (Fig. 8). Sandstones are represented by 19 of the 27 thin sections; siltstones, by seven; shales by one. The coarsest sediment fraction observed in the thin sections was fine sand (0.12–0.16 mm).

Primary sedimentary structures are well developed in outcrop, but some also are discernible at thin-section scale. Some thin sections show stratification or grain fabric (Fig. 8, thin sections 14-1 and 20-1). Laminations, wispy clay layers, and bioturbation are visible in a few (Fig. 8, thin section 10-2). In one shale sample, centimeter-scale convolute bedding was noted (Fig. 8, thin section 22-1). Heterogeneities in grain size and color are present on a centimeter scale. One sample shows graded bedding on a millimeter scale (Fig. 8, thin section 14-1).

Sandstones

Sandstones are represented by 19 of the 27 thin sections. Most of the sandstones are classified as quartz arenites. Sandstones are mostly very fine grained (0.12–0.16 mm); some contain as much as 20% silt-size grains, whereas others may contain as much as 10% fine-sand-size grains. Some of the fine-grained sandstones may contain as much as 25% very fine sand-size grains. Grains are poorly to moderately well sorted in most samples.

Monocrystalline quartz is the dominant framework grain in all sandstone samples. A few of the monocrystalline quartz grains are brown and slightly turbid. Some grains appear welded together by quartz overgrowths (Fig. 8, thin section 2-3). Apparent suturing of grains is the result of quartz overgrowth cementation (Fig. 8, thin section 2-3). Iron oxides form pin-point rims around many of the quartz grains. Feldspar is present in small amounts (<8%) in most samples. Both K-feldspar (orthoclase, microcline) and plagioclase are present; plagioclase is more common. Some feldspar is twinned. Rock fragments, predominantly metamorphic(?), are scarce or absent in the samples.

Accessory minerals are present in trace amounts (<1%) in most samples. Mica, especially muscovite, is present in most samples and is a significant component, as much as 2–3%, in a few samples. Mica content increases with decreasing grain size. Organic matter and pyrite are the most abundant opaque grains.

Quartz is the most common and significant cement in many samples. Quartz cement occurs as well-developed crystal overgrowths on detrital quartz grains.

Clay generally is not abundant but is visible in most thin sections as delicate, brownish grain-coatings and pore-linings. Clay also is visible as a pseudomatrix, produced by compactional deformation of clay- or mica-rich rock fragments between less-ductile grains. In the very fine grained sandstones and in siltstones, detrital clay is present as continuous and discontinuous laminae. Opaque clay-sized particles, including organic matter, could not always be distinguished.

Siltstones

Siltstones are represented by seven of the 27 thin sections. The silt grains range in size from 0.02 to 0.06 mm. Grains are poorly to moderately well sorted in most samples. Grain shapes are generally subangular to angular (Fig. 8, thin section 10-2). The siltstone samples contain 5–30% very fine sand-size quartz grains. Detrital silt-size quartz grains and feldspar are mixed with larger, platy grains of muscovite and organic matter. Organic matter and pyrite constitute opaque grains. Laminations, consisting of millimeter-thick, closely packed silt and very fine sand grains of quartz, occur in some samples. Detrital clay is visible as diffuse brown patches and dark laminae.

Shale

Shale is represented in only one of the 27 thin sections. (Fig. 8, thin section 22-1). Principal constituents of the shale are terrigenous silt and clay. The terrigenous silt fraction is composed of subangular to angular grains of monocrystalline quartz and lesser amounts of mica, feldspar, and accessory minerals similar to those in the sandstones. Convolute bedding on a centimeter scale was noted in the sample. Part of that micro-scale soft-sediment deformation feature is shown in Figure 8, thin section 22-1. Carbonaceous organic matter is present, and pyrite occurs in trace amounts.

Sedimentary Features

Ripple Marks

The Atoka succession contains a variety of current and wave ripples representing a range of flow velocities and directions (Fig. 14). Some ripple patterns (interference) suggest the combined influence of waves and currents. In some cases, thin mud drapes occur on ripple foresets, or ripples have a more symmetrical form, which suggests wave and, possibly, tidal influence. Paleocurrents measured from current ripples on bedding-plane surfaces and from a limited number of cross-laminations tend to be highly variable and are of limited value in interpreting prevailing current directions. Superimposed ripple-bed forms show rapid lateral changes in size and orientation over short distances (Fig. 15). Ripple-bed forms typically are <2 in. high. In this study, small ripples are not regarded as distinct from the larger megaripples, as I believe that the two classes show considerable overlap. However, the bed-form scale of some of the ripple trains in the upper part of the Atoka Formation may well warrant their classification as megaripples. This possibility will be considered in ongoing studies.

Both symmetrical and asymmetrical wave ripples are present. Symmetrical wave ripples are essentially straight-crested and show, in part, a tuning-fork-like bifurcation. This type of ripple typically is produced by slow-moving waves on the margins of a water body in water a few inches deep and does not need a fetch (Reineck and Singh, 1980) (Fig. 16A,B).

Asymmetrical wave ripples are similar to straight-crested current ripples in possessing a steep lee side and a gentle stoss side (Reineck and Singh, 1980). These wave ripples may develop as climbing ripples or ripple laminae in-phase if sufficient sediment is available in suspension. Crests of both asymmetrical and symmetrical wave ripples show repeated bifurcation in the shape of a tuning fork, and the profile of crests is rather regular (Fig. 16C).

Climbing ripples and ripple-drift cross-laminations occur in the coarsening-upward sequences, particularly at, or near, the tops of sandstone beds. They represent current deposition in an environment where intermittent interaction with, or reworking by, waves produced the ripple laminae (Fig. 16D–F). Evidence for wave-influenced ripple laminations in the succession includes opposite ripple-foreset directions within single ripple horizons and laminations out-of-phase with ripple forms. Climbing-wave ripple laminations are highlighted by the presence of quartz-rich laminae distributed among clay-rich laminae. The climbing- ripple laminae are produced by low flow-velocity currents saturated with sediments (probable flood event). The high sedimentation rate from suspension produces local hydrodynamic conditions similar to those of waning turbidity currents. Some unusually thick ripple-drift cross-laminations observed suggest low-flow regime, but excessive bed-load sediment supply.



Figure 14. View (taken looking southward) of the lower part of the section, Units 6–8 (133–158-ft interval). Note the diversity and abundance of ripple-bed forms on the 26° northward-dipping bedding-plane surfaces.

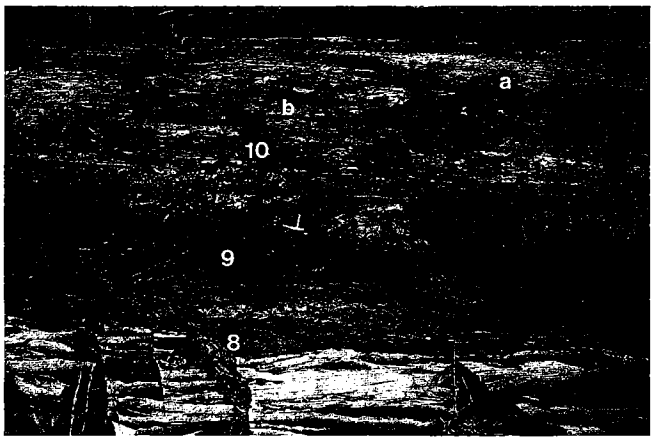


Figure 15. Top of highly deformed Unit 8, shale Unit 9, and lower part of heterolithic-bedded Unit 10 (156–165-ft interval). Note the stacked ripple-bed forms in Unit 8 and the cross-sectional view of planar cross-beds produced by migration of straight-crested current ripples into relatively deeper water during storm-events. In Unit 10, note the concave-upward erosional scour (a) and contorted bedding (b). Geologic pick is 1.1 ft long.

Current ripples are most common and are represented by both unidirectional and multidirectional bed forms. *Straight-crested current ripples,* with almost straight and parallel crests, are the most common form of current ripple (Fig. 17A). These ripples may show small, tongue-like projections in the down-current direction. Straight-crested current ripples form at relatively low velocities and, upon migration, may produce small-scale units of planar cross-bedding (Reineck and Singh,

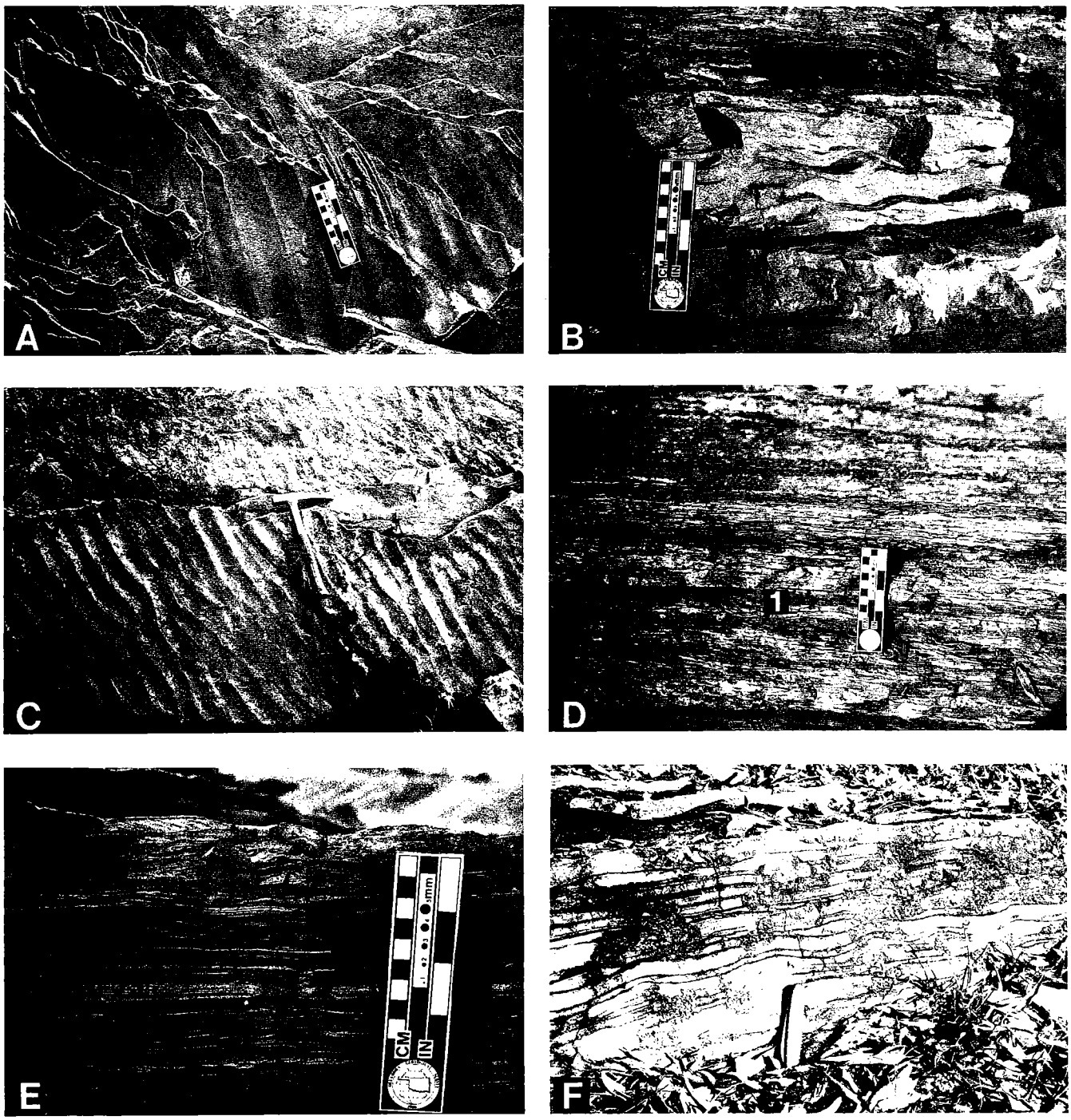


Figure 16

A. Unit 8, top. Straight-crested symmetrical wave ripples showing rounded troughs and parallel crests. Note bioturbation in troughs. Scale is 5 in. long.

B. Unit 13, upper part. Flaser bedding in wave-rippled (small-scale, symmetrical) sandstone suggests alternating deposition of sand (in turbulent water) and mud (in slack water). Note compressed (elongated), shale rip-up clasts at top. Scale is 5 in. long.

C. Unit 8, top. Asymmetrical wave ripples showing bifurcation of crests in the shape of a tuning fork. Note the rather regular profile of the crests. Geologic pick is 1.1 ft long.

D. Unit 14, lower part. Small-scale heterolithic bedding (i.e., centimeter-thick alternating couplets of siltstone and shale). Note starved ripples (flaser bedding) at top of scale and isolated, truncated sets of ripple-drift cross-laminations (1). Scale is 5 in. long.

E. Unit 7. Vertical sequence of small-scale sedimentary features: starved ripples (lenticular bedding) at the base; laminated small-scale planar bedding in the middle; and small-scale trough cross-beds at the top produced by small-scale climbing ripples. Scale is 5 in. long.

F. Unit 2. Climbing-ripple lamination with laminae in-phase. Ripples are asymmetrical and were produced under current action. Rippled-bed forms are preserved as wavy bedding. Knife is 3 in. long.

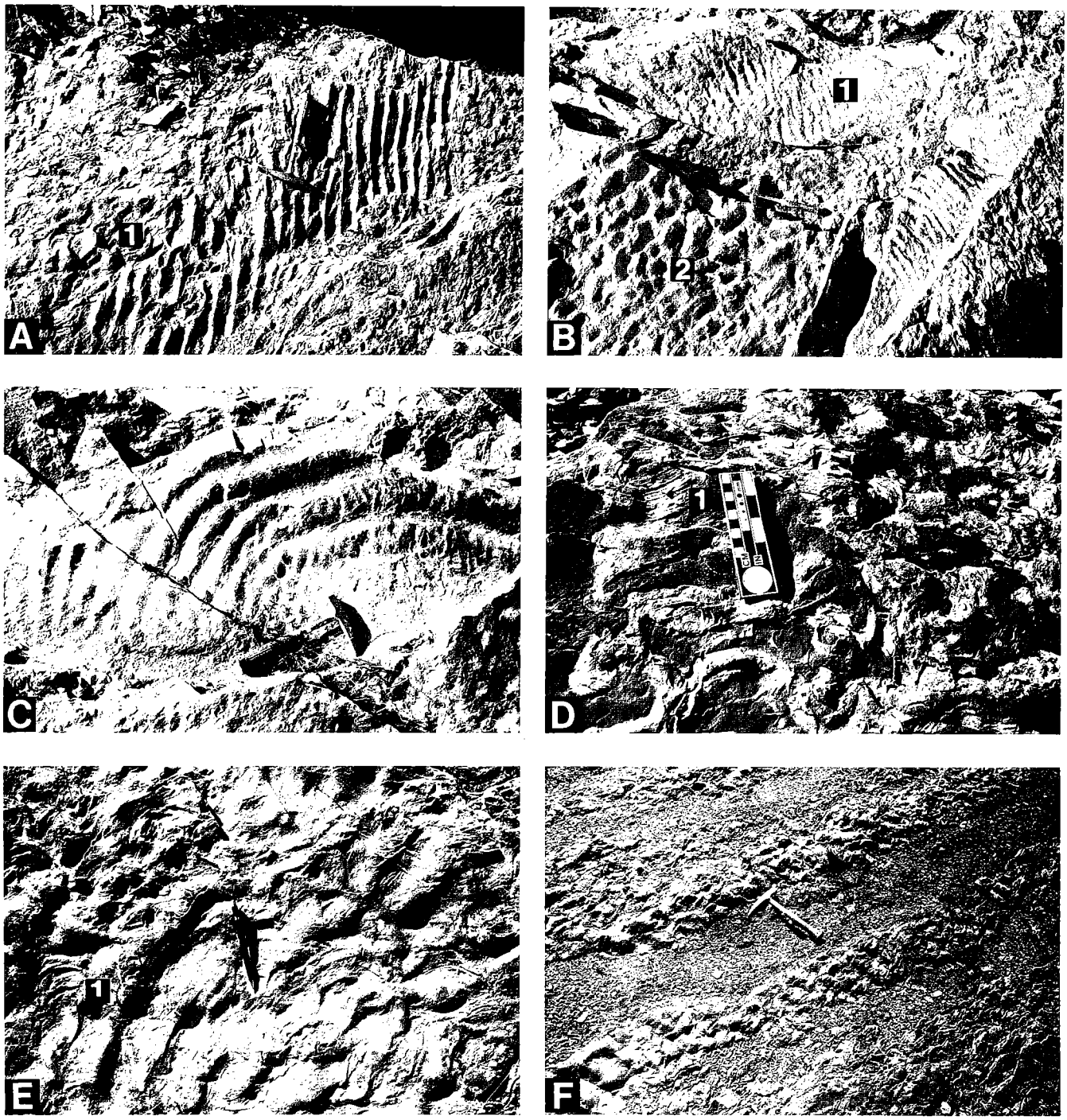


Figure 17

A. Unit 7, near top. Straight-crested current ripples with almost straight and parallel crests. Note superimposed ripple-bed forms and change in ripple pattern to strongly undulatory, almost linguoid-like, small-current ripples on left side of photo (1). Geologic pick is 1.1 ft long.

B. Unit 6. Bedding-plane surfaces showing weakly undulatory small-current ripples (1) and strongly undulatory, almost linguoid-like, small-current ripples (2). Geologic pick is 1.1 ft long.

C. Unit 6. Undulatory current-ripple train. Ripples have long, continuous, wavy or undulatory crests that project forward into tongue-like extensions. Geologic pick is 1.1 ft long.

D. Unit 13. Top bedding plane surface showing strongly undulatory, almost linguoid-like, current ripples with out-of-phase crests. Note the horizontal grazing (feeding) pattern of *Zoophycos* (1). Scale is 5 in. long.

E. Unit 6. Grazing (feeding) pattern of *Zoophycos* (1) on a strongly undulatory current-rippled bedding-plane surface. Probable trace maker of the spreiten was a surface deposit-feeding annelid. Pencil is 6 in. long.

F. Unit 23. Interstratified silty shale (50%) and shaly siltstone (50%). Siltstone beds show strongly undulatory, almost linguoid-like, current ripples. Geologic pick is 1.1 ft long.

1980) (Figs. 15,16E). *Undulatory small-current ripples* possess long, continuous, wavy or undulatory crests. The undulations may be arranged in-phase or out-of-phase, but usually they are out-of-phase. The crests are continuous but project forward into tongue-like extensions. These common ripple-bed forms represent a transition form between low-energy straight-crested, small-current ripples and higher-energy, linguoid, small-current ripples (Fig. 17B–F) (Reineck and Singh, 1980). The migration of undulatory current ripples may produce weakly festoon-shaped cross-bedding in side view and strongly trough-shaped cross-bedding in front view. *Linguoid current ripples* consist of small curved crests that extend forward in a tongue-like or lobe-like forms (Fig. 17C). The tongues are out-of-phase. The crests of the ripples are discontinuous and broken and have forward closures; thus, they cannot be traced over long distances. Migrating linguoid current ripples may produce strongly festoon-shaped cross-bedded units. In the coarsening-upward sequences within the upper part of the Atoka Formation, there is a gradual transition from straight-crested to undulatory, and undulatory to linguoid-like ripples, and all transition forms are found (Fig. 17A,B, D,F).

Some ripple patterns suggest the combined influence of waves, currents, and, possibly, tides. These longitudinal ripples (interference) are relatively straight-crested, and usually show no bifurcation of crests (Figs. 18,19A). The ripple crests ran parallel to the current and wave propagation was at right angles to the current. These ripple-bed forms may have originated in cohesive, muddy sediments with very low sand content. The sharp interbedding of wave-rippled siltstone/sandstone and shale suggests that deposition took place in conditions alternating between wave-influenced and quiet water (Fig. 16B). Discontinuous siltstone/sandstone lenses reflect the development of small-scale ripple bed forms (starved ripples) produced by stronger flows associated with storm waves (Fig. 16D,E). The starvation of the ripples suggests that sediment supply was limited, at least temporarily, and probably was episodic.

Stratification

Planar cross-bedding is not common. In some places, scour channels are filled with high-angle planar cross-stratified sandstone (Fig. 19B). The presence of planar cross-beds in solitary sets suggests deposition from straight-crested current ripples that migrated into deeper water during storm events (Figs. 15;16D,E).

Trough cross-stratification, present in some intervals, is usually small-scale and commonly truncated (Figs. 16E,19C). Trough cross-beds record the migration of sinuous-crested ripples from relatively low-flow velocity currents.

Thinner couplets of sandstone/siltstone and shale, without cross-laminations, suggest deposition below storm-wave base. These couplets are particularly characteristic of the heterolithic-bedded facies (i.e., Units 7, 10, 12, and 14) (Figs. 15,16D,19D).

Flaser/lenticular bedding is most common in the heterolithic-bedded units (Figs. 16D,E;19E). All variants of flaser bedding are present: simple flaser-bedding, bifurcated flaser-bedding, wavy flaser-bedding, and bifurcated, wavy flaser-bedding. Flaser bedding is associated with straight-crested current ripples, small current ripples with curved crests, and wave ripples. Lenticular bedding is associated both with sand lenses of current origin and with sand lenses produced by both symmetrical and asymmetrical wave ripples (Figs. 16E,19E). It results when incomplete ripples form on a muddy substrate that is covered later by another mud layer and then sand in the form of ripples. Thin sandstone beds contain wavy, discontinuous, clayey laminae, which developed as clay drapes that filled ripple troughs and outlined ripple lenses (Figs. 16B,D,E;19E). Flaser/lenticular bedding provides evidence for: (1) availability of both sand and mud, (2) alternating turbulent-water and slack-water intervals for deposition of sand and mud, respectively, and (3) high loads of suspended clay/silt/sand. The occurrence of streaks and lenticular siltstone/sandstone laminations in a shale-dominated interval, observed in certain intervals as load-casted ripples, suggests a depositional environment where the hydraulic regime is fluctuating constantly between periods of increased and decreased fluid motion and/or where sediment supply is episodic (Figs. 11,12).

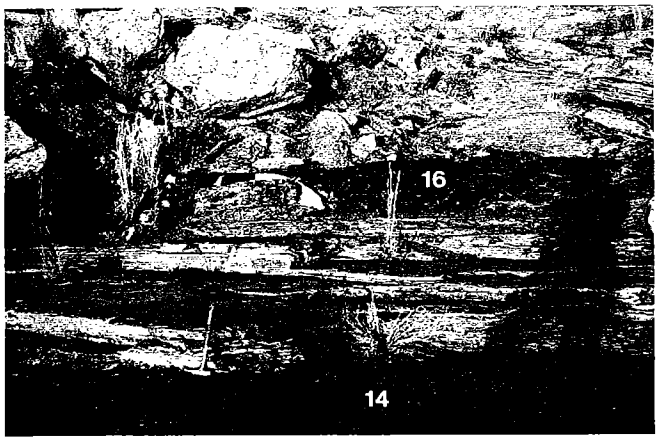


Figure 18. Tops of highly rippled (interference ripples) Unit 14, Unit 15, and lower part of grayish black (N3) shale Unit 16 (207–210-ft interval). Note sharp contacts. Geologic pick is 1.1 ft long.

Horizontal, parallel, continuous laminations are rare, but when present they are preserved as small-scale planar-bedding (Figs. 15,16E). The dominantly horizontal laminations are formed by grain-size and sorting variations, silt or mud laminations, or carbonaceous material. Such laminations are upper flow-regime sedimentary features probably formed by deposition of sediment from suspension in response to wave action, which suggests aggradation of the facies into a more energetic zone.

Rare intervals of intersecting, low-angle to slightly undulatory, subparallel laminations with a slight upward convexity suggest swaley cross-stratification as described by Leckie and Walker (1982) (Fig. 19F). The difference between swaley and hummocky cross-stratification, in my opinion, has not been resolved fully. Swaley cross-stratification is produced by large, storm-generated waves that scour and then deposit very fine to fine sand below the fair-weather breaker zone in broad, shal-

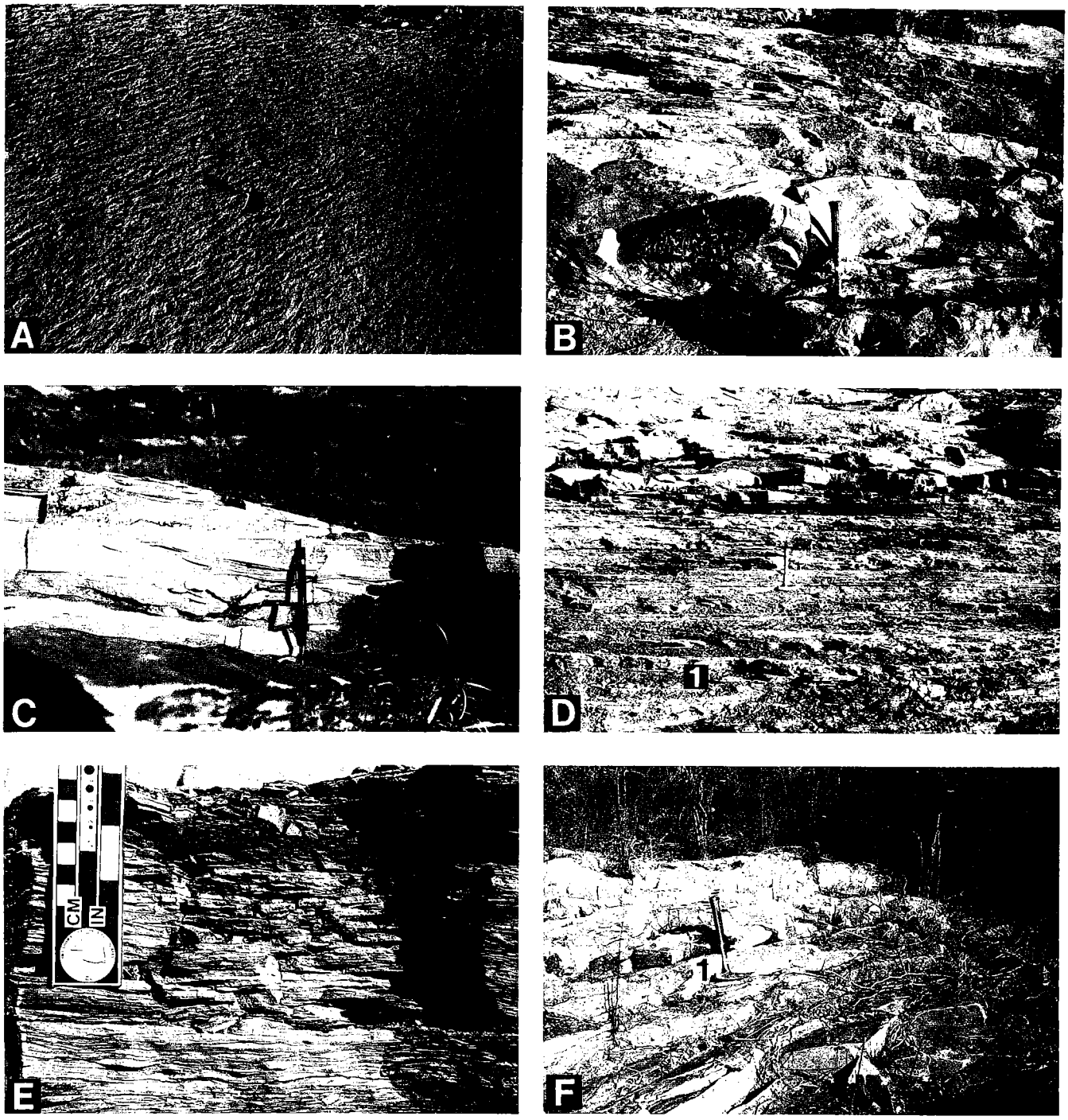


Figure 19

A. Unit 12, near top. Interference ripple pattern suggestive of both current and wave influence. Geologic pick is 1.1 ft long.

B. Unit 10. High-angle planar cross-stratified sandstone body filling in scour channel. Cross-bedding probably was produced by the migration of straight-crested current ripples into deeper water during storm-events. Geologic pick is 1.1 ft long.

C. Unit 8, top. Truncated set of low-angle, trough cross-stratified, very fine grained sandstone produced by the migration of sinuous-crested ripples from low-flow velocity currents. Pencil is 6 in. long.

D. Unit 7. Small-scale, coarsening-upward cycle with heterolithic bedding (alternating couplets of sandstone and shale) in lower part. Note concave-upward erosional scour (1). Geologic pick is 1.1 ft long.

E. Unit 12, upper 1 ft. Flaser/lenticular bedding in starved-ripple marked siltstone/sandstone. Clayey laminae developed as clay drapes that filled ripple troughs and outlined ripple lenses. Scale is 5 in. long.

F. Unit 26. Rare swaley cross-stratification (1) in very fine grained sandstone succession. Cross-stratification is produced by deposition in shallow swales related to increased wave energy. Geologic pick is 1.1 ft long.

low swales. This sedimentary structure is preserved best between fair-weather base and storm-wave base where low-energy conditions dominate (e.g., outer shelf, lower delta front).

Soft-Sediment Deformation Features

Penecontemporaneous, soft-sediment deformation structures include contorted/slumped beds, convolute bedding, ball-and-pillow structures, load-casted ripples, pinch-and-swell bedding, and small-scale recumbent folds. The abundance of the features suggests rapid sedimentation (such as that caused by storm waves or by gravity-induced slumping or sliding on oversteepened depositional slopes), which produced unstable substrates. Soft-sediment deformation also may occur where there is abundant disseminated organic material: trapped methane formed during organic decay may trigger sediment failure by helping to raise the pore-fluid pressure.

Contorted bedding (noted in Units 8, 10, 12, and 14 [Fig. 8]) is a product of syndepositional slumping of sandstone bodies due to loading and oversteepening of the depositional surface during periods of peak water discharge (Fig. 15). Contorted bedding indicates sufficient sand to induce load deformation and sufficient mud to initiate sediment failure. Contorted bedding results in considerable disruption of the original bedding (Fig. 20A,B).

Load casts and ball-and pillow structures (noted in Units 8, 10, 14, 26, and 28 [Fig. 8]) indicate rapid deposition of sand onto a water-rich "soupy" substrate. Discontinuous ball-and-pillow structures (pinch-and-swell bedding) are produced when a sand layer is broken up into several isolated ellipsoidal (pillows) masses (Fig. 20C,D). Sandstones soled by ball-and-pillow structures and/or other soft-sediment deformation features indicate in-place sediment failure after deposition (Fig. 20E). These soft-sediment features are evidence for intermittent storm cycles or shoaling events. In addition, these features are indicative of slope instability due to: (1) angle of slope, (2) sediment/water pore-pressure differences, and (3) sudden sediment influx.

Some beds pass laterally into a slumped horizon (contorted massive sandstone) that has been fluidized (Fig. 20D). The lateral variability in bed thickness produces a pinch-and-swell type bedding (ball-and-pillow structures) (Fig. 20D) similar to that in Units 8 and 10 (Fig. 8). Others beds have been deformed syndepositionally, which suggests that episodic depositional events (e.g., storms) caused rapid influx of sediments and subsequent dewatering and/or slumping of water-saturated sands/silts.

Another type of small-scale slump structure, a recumbent fold, was formed in a very fine grained sandstone in Unit 26 (Fig. 20F). The recumbent fold formed in place from original bedding failure. The single occurrence in Unit 11 of a sandstone plug suggests sand mobilization following burial.

Scours

Locally, isolated broad erosional scours, some as much as 8 ft wide and 2 ft deep, contain either low-angle trough cross-beds or high-angle planar cross-beds of very fine grained sandstone (scour-and-fill) (Figs. 19B, 21A). The scours and cross-bedded sandstones usually are encased within planar-bedded siltstone/sandstone. The erosional scours are concave-upward and sandstone beds commonly pinch out laterally above the erosional surfaces (Figs. 15;20A,C). Because of erosional scours and soft-sediment deformation, sandstone/siltstone beds generally are discontinuous laterally at outcrop scale. The sharp basal contacts of several sandstone bodies suggest that some erosion of the underlying strata may have occurred prior to deposition of the basal sand.

Shale/Mud Clasts

Shale/mud rip-up clasts are not common and are present only in Units 2, 6, and 13. Most of the clasts are deformed or "squeezed" and are oblong in shape, which suggests that the clasts were semi-consolidated during deposition and suffered plastic flow after burial (Fig. 16B). Some rounded clay/shale clasts occur at the tops of rippled sandstone beds and indicate reworking and redeposition of partially consolidated mud during a period of relatively higher flow-velocity. The occurrence of mud/shale clasts suggests that the sediment was broken up periodically by wave/current action in water probably shallower than the depth of the effective wave base.

Pyrite/Siderite

Pyrite occurs in both disseminated and nodular form and also as a replacement mineral for brachiopod shells. The presence of pyrite indicates deposition under reducing conditions, probably produced by organic decay.

Ironstone, commonly in the form of siderite, occurs as nodules, lenses, and bands. Its presence suggests anoxic reducing conditions due to organic decay. The presence of siderite also suggests that pore fluids lacked sulphate which, if present, would have produced pyrite.

Plant Material

Carbonaceous material is common at several stratigraphic levels and in several different forms. Comminuted plant material is common throughout the succession. Laminations highlighted by comminuted plant material are common. Partially coalified *Calamites* stems are present in Units 7, 13, and 14 (Fig. 21B). Some of the *Calamites* stems show evidence of scouring. The presence of comminuted plant material coincides with an increase in micaceous-rich sandstones/siltstones/shales.

Coalified plant stems, common comminuted plant material, and the dark gray (N3) color of the shales suggest a heavily vegetated and poorly drained, low-energy, marginal-marine source area as well as transportation by storm-influenced currents/waves over a mud-dominated shelf into a relatively low-energy, dysaerobic to anaerobic, relatively deeper-water setting.

Body Fossils

Body fossils are extremely rare in the succession. A dark gray (N3) noncalcareous shale (Unit 11) contains delicate, finely ribbed, thin-shelled, pyritized chonetid-type brachiopod shells. Brachiopods, crinoidal hash, bivalves, and gastropods were recovered from a medium

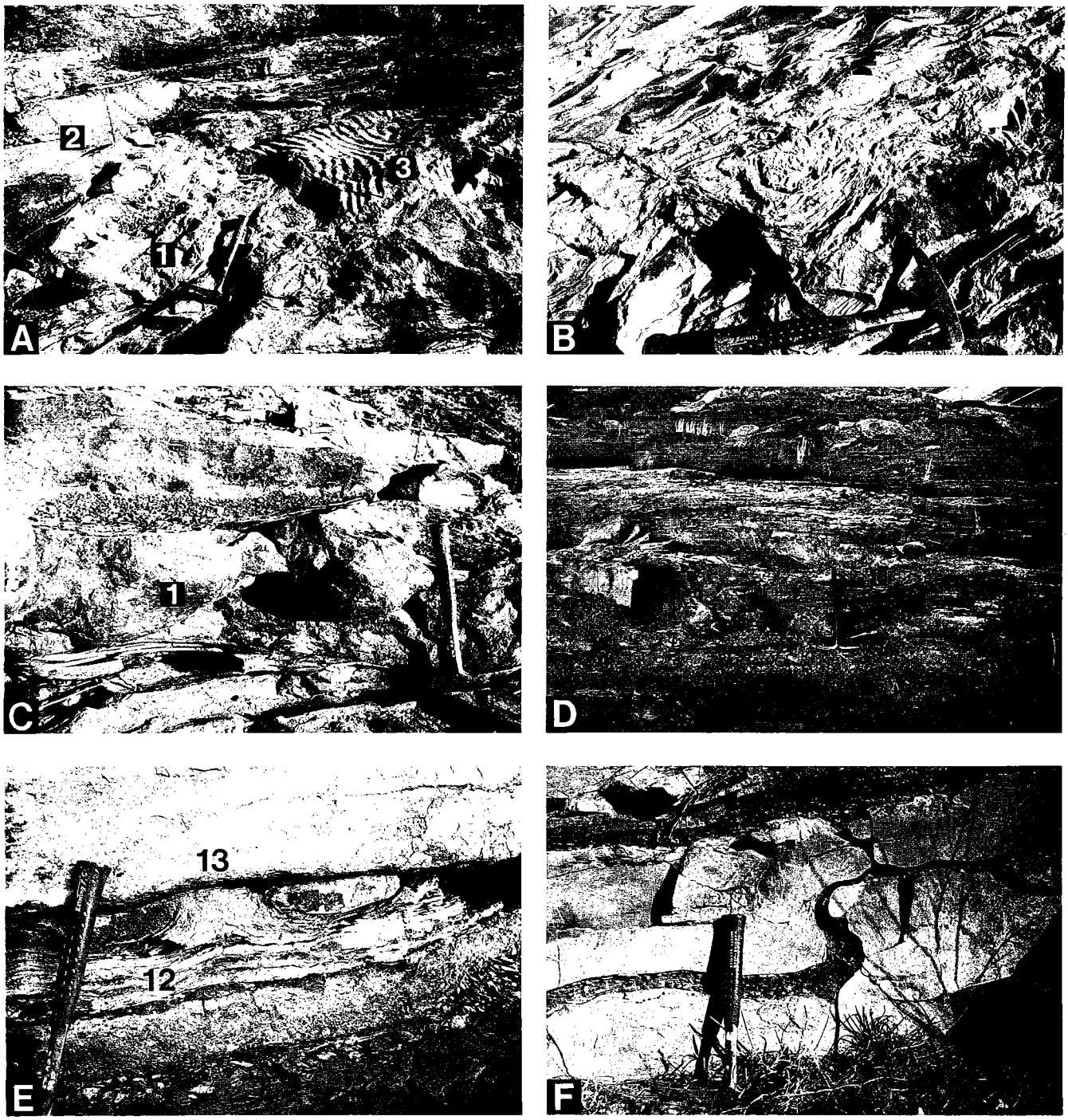


Figure 20

A. Unit 10. Highly contorted unit. Note contorted bedding (1), concave-upward erosional cut-and-fill scour (2), and weakly undulatory small-current ripple trains (3). Geologic pick is 1.1 ft long.

B. Unit 8. Penecontemporaneous soft-sediment deformation structure showing slumped or contorted bedding. Note strongly rippled bedding-plane surface at the top. Geologic pick is 1.1 ft long.

C. Unit 10. Ball-and-pillow structure (1), a product of soft-sediment deformation that indicates rapid deposition of sand onto an unstable water-rich "soupy" substrate. Note concave-upward erosional scour at the top of the photo. Geologic pick is 1.1 ft long.

D. Unit 10. Foundered sandstone body resulting from in-place sediment failure after deposition on a water-rich "soupy" substrate. Note the lateral variability in bed thickness. Hoe pick is 2.2 ft long.

E. Contact between Unit 12 and Unit 13. Note soft-sediment deformation feature (truncated flame structure) at contact. In-place sediment failure after deposition is suggested by soft-sediment deformation features on the sole of Unit 13. Geologic pick is 1.1 ft long.

F. Unit 26. Recumbent fold, a type of slump structure, in a very fine grained sandstone, formed in place from original bedding failure. Geologic pick is 1.1 ft long.

dark gray (N4) sandy to silty, noncalcareous shale in the upper part of the mud-dominated succession (Unit 29, Fig. 13). In addition, a 2–4-in.-thick ironstone layer in shale Unit 33 contains chonetid- and spiriferid-type brachiopods in which some of the shells are pyritized. The representative fossils collected are from organisms adapted, for the most part, to soft substrate and relatively turbid water.

The overall paucity of body fossils in the upper part of the Atoka Formation suggests that sedimentation rates were very high and/or that the environment was hostile (e.g., oxygen-deficient) to benthic organisms. Thirteen shale samples are currently being processed for microfossil data (Fig. 8).

Biogenic Structures

The overall intensity or level of bioturbation is assigned an ichnofabric index of 1 to 5. These indices indicate the degree of bioturbation according to the visual scale of Droser and Bottjer (1986,1989): 1 = nonbioturbated (1–10%); 2 = weakly bioturbated (10–30%); 3 = moderately bioturbated (30–60%); 4 = strongly bioturbated (60–95%); 5 = churned (95–100%). The percentages arbitrarily assigned to each ichnofabric index above are the

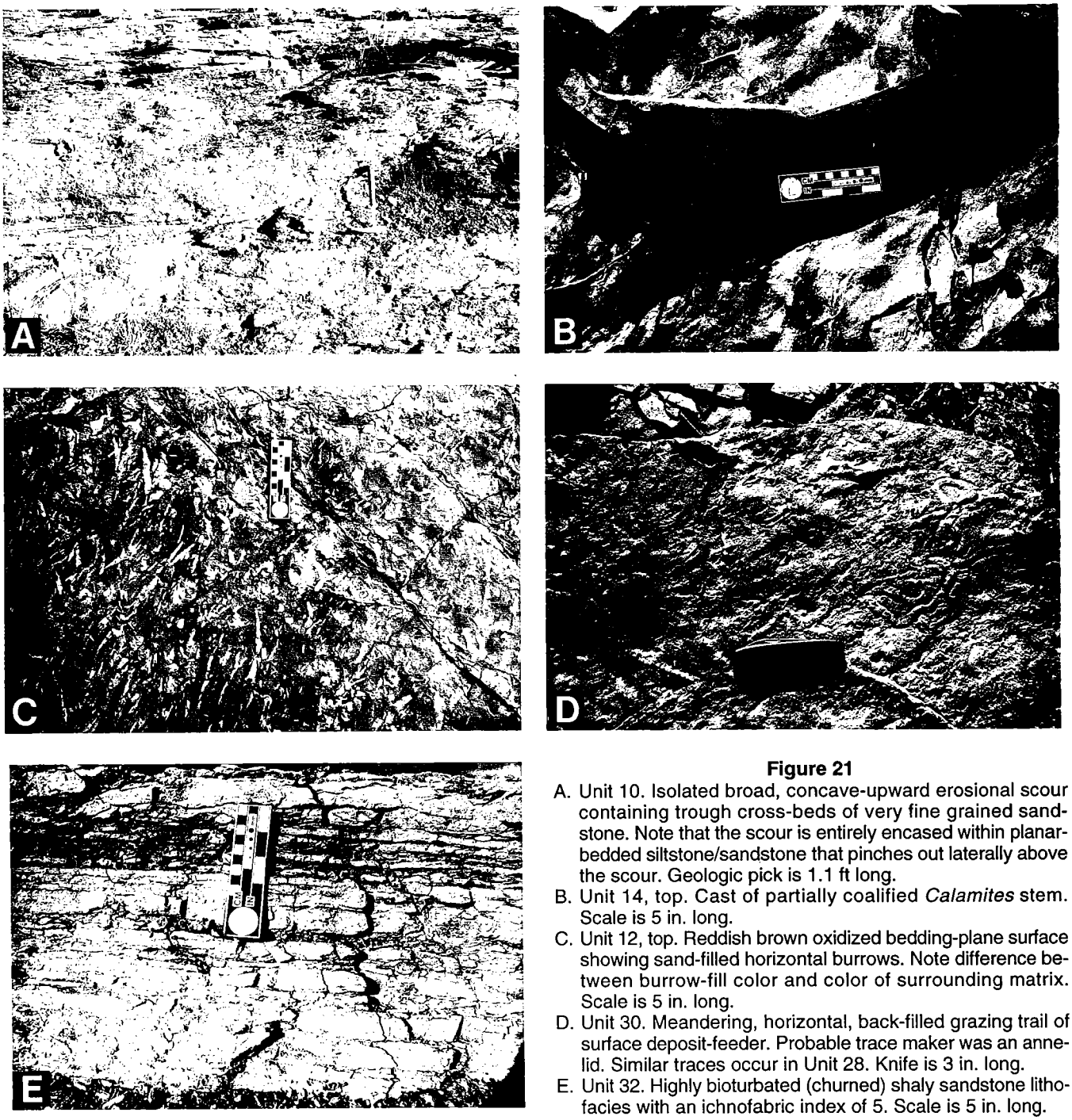


Figure 21

A. Unit 10. Isolated broad, concave-upward erosional scour containing trough cross-beds of very fine grained sandstone. Note that the scour is entirely encased within planar-bedded siltstone/sandstone that pinches out laterally above the scour. Geologic pick is 1.1 ft long.
B. Unit 14, top. Cast of partially coalified *Calamites* stem. Scale is 5 in. long.
C. Unit 12, top. Reddish brown oxidized bedding-plane surface showing sand-filled horizontal burrows. Note difference between burrow-fill color and color of surrounding matrix. Scale is 5 in. long.
D. Unit 30. Meandering, horizontal, back-filled grazing trail of surface deposit-feeder. Probable trace maker was an annelid. Similar traces occur in Unit 28. Knife is 3 in. long.
E. Unit 32. Highly bioturbated (churned) shaly sandstone lithofacies with an ichnofabric index of 5. Scale is 5 in. long.

author's (JRC). Some caution should be used when noting the indices assigned in Figure 8. The intensity of bioturbation may vary slightly throughout any unit; therefore, the assigned index only represents the general overall intensity of bioturbation throughout that particular unit.

Bioturbation structures are preserved best in the heterolithic-bedded units, that is, units with contrasting rock types (e.g., interstratified sandstone and shale). The most common bioturbation pattern is the alternation of weakly bioturbated (sparse, restricted fauna) intervals with nonbioturbated intervals (Fig. 8). This pattern suggests that, in general, deposition probably took place near the dysaerobic/anaerobic boundary and that there was periodic alternation between dysaerobic (low oxygen: dissolved oxygen between 1.0 and 0.1 mL/L) and anaerobic (dissolved oxygen <0.1 mL/L) conditions. Overall, biogenic structures are rare, and identifiable ichnogenera are rare to absent. This overall paucity of biogenic sedimentary structures probably indicates that there was an unusually rapid rate of fair-weather suspension sedimentation and/or that the environment was hostile (e.g., oxygen-deficient) to benthic organisms. The relatively high abundance of relatively few ichnogenera (e.g., *Zoophycos* and *Chondrites*) in certain intervals also suggests stressful environmental conditions (Fig. 8). These conditions may have been due to variable rates of sedimentation, fluctuating salinity, unstable substrate, or other associated environmental parameters.

The relative abundance of horizontal feeding-burrows of deposit feeders (e.g., *Zoophycos*) in some intervals of the succession indicates an intermittent low wave-energy environment and a relatively slow rate of deposition that permitted benthic organisms to colonize the substrate (Fig. 21C–E).

The occurrence of burrows throughout the coarsening-upward sandstone sequences suggests a marine setting that must have been relatively shallow in order to have had steady-to-episodic strong currents. The *Zoophycos* feeding traces, most common in sandstone beds at the top of the coarsening-upward sequences, are related to deposit-feeding organisms that lived in the overlying mud and burrowed down into the buried sand (Fig. 17D,E). Increased biogenic activity from the base to the top of the sandstone sequence is in response to the aggradation of the seafloor (i.e., moderately slow sedimentation), the development of more oxygenated marine conditions, and the relative absence of direct fluvial influence or marine reworking. The tracemaker of *Zoophycos* was most likely a deposit-feeding vermiform organism (e.g., annelid, sipunculid) that developed a complex feeding strategy. Some workers suggest that the occurrence of *Zoophycos* indicates middle- to outer-shelf conditions (Bottjer and others, 1984). More recently, other workers have shown that *Zoophycos* is a facies-crossing ichnogenus that may be found in many different depositional settings at all water depths, although it is associated generally with relatively quiet water conditions (Marintsch and Finks, 1978).

Chondrites was identified in Unit 14 (Fig. 8). *Chondrites* consists of dense, branching feeding probes probably of a worm-like animal. *Chondrites* may be a response to local anoxic conditions caused by the decay of storm-buried organic material (Vossler and Pemberton, 1988). The presence of pyrite, siderite, and the trace fossil *Chondrites* suggests that low oxygen levels existed locally in the interstitial waters within the sediment (Bromley and Ekdale, 1984).

DISCUSSION

This exposure of the upper part of the Atoka Formation contains a diverse suite of sedimentary features associated with current and/or wave processes. These include (1) small-scale cross-laminations, (2) current ripples, (3) wave ripples, (4) scour-and-fill features, (5) erosional truncations, (6) climbing wave-ripple laminations, and (7) swaley bedding. Lulls or pauses in current activity preceding deposition of rippled sandstone are shown by shaly partings lining basal scour surfaces. Periods of lower flow-velocities and/or shallowing of water, probably associated with fair-weather conditions, are indicated by the presence of ripple-bed forms on top of trough cross-bedded sets.

Evidence of deposition by infrequent (episodic) high-energy events of short duration (storms) characterizes the coarsening-upward sandstone lithofacies found throughout the overall mud-dominated lithofacies (Fig. 22). Some evidence for storm-influenced, episodic sedimentation includes: (1) laminated siltstone/sandstone lenses in silty shale intervals; (2) alternating flow-velocity characteristics of vertical-bedding sequences (e.g., basal rippled sandstone units [low-flow regime] overlain by sandstone cross-beds [higher flow regime] overlain by rippled sandstone units at the top [low-flow regime]); (3) asymmetric ripple laminations with flaser and thin, wavy shale beds; (4) symmetrical and climbing ripples; (5) rip-up clasts in cross-bedded sandstone facies; (6) sharp-based sandstone beds separated by shale interbeds; (7) rapid lateral variation in stratification; (8) horizontal trace fossils common, but restricted to rippled surfaces in upper part of sandstone facies; and (9) alternating variations in grain size vertically. Some units have been syndepositionally deformed, which suggests that epi-



Figure 22. Basal part of section, south of spillway overlook, showing Units 1, 2, and 3 (0–70-ft interval). Note the coarsening-upward, small-scale depositional packages within Unit 2. Note the increase in frequency and thickness of sandstone beds upward in Unit 2. Hoe pick is 2.2 ft long.

sodic depositional events (e.g., storms) caused rapid sediment influx and subsequent dewatering or slumping of water-saturated sands and silts.

Periods of moderate to intense sediment reworking are recorded by (1) abundance of storm deposits, (2) swaley cross-stratification, and (3) contorted/slumped beds.

High rates of sedimentation are suggested by (1) a truncated flame structure, (2) high-angle ripple-drift cross-laminations, (3) abundant comminuted plant material, (4) abundant soft-sediment deformation features, (5) relatively poor sorting of sandstones, and (6) paucity of body fossils and biogenic structures.

Dysaerobic to anaerobic environments during shale deposition are indicated by (1) weak bioturbation, (2) paucity of body fossils, (3) organic-rich shales, and (4) abundant pyrite and ironstone (siderite).

INTERPRETATION

Environmental interpretations regarding the probable depositional setting of this exposure of the upper part of the Atoka Formation are constrained by a limited data base. However, the overall succession of sediments are related genetically and are part of a single depositional system.

The textures, vertical succession of facies, and sedimentary features suggest conditions ranging from suspension fallout and, possibly, weak density flows below storm wave base (e.g., slightly fissile shale, graded sandstone/shale couplets, etc.) through the progressively higher energy and greater sediment supply of storm-wave influence (e.g., wave-rippled siltstone, swaley cross-stratification, etc.). All of the vertical stratigraphic trends of the succession, considered together, suggest deposition on a storm-influenced, mud-dominated shelf.

The lower part (Units 1–15) of the succession (less distal shelf) is mud-dominated, but it is more sand-rich and has higher sand/shale ratios than the upper part (Units 16–33) (Figs. 10,11,14,22). The lower part also shows more frequent indications of oscillatory flow and of sedimentary features recording deposition by less distal processes.

The upper part (Units 16–33) of the succession (more distal shelf) is mud-dominated and has lower sand/shale ratios (Figs. 12,13,23). In addition, the upper part shows less frequent evidence of oscillatory flow and records deposition by more distal processes as evidenced by the dominance of finer grained sediments and paucity of shallow-water features. The upper part contains rare isolated sand lithofacies, which suggests only localized transport of sand (Fig. 23).

The gradational facies transition from lower, relatively shallow-water, more sand-rich, storm-dominated shelf deposits to an upper, relatively deeper-water, below-wave-base, mud-rich succession, is supported by: (1) the up-section increase in shale; (2) a progression up section to less frequent, thinner sandstone beds; and (3) the rare occurrence of swaley cross-stratification in the upper part of the section.

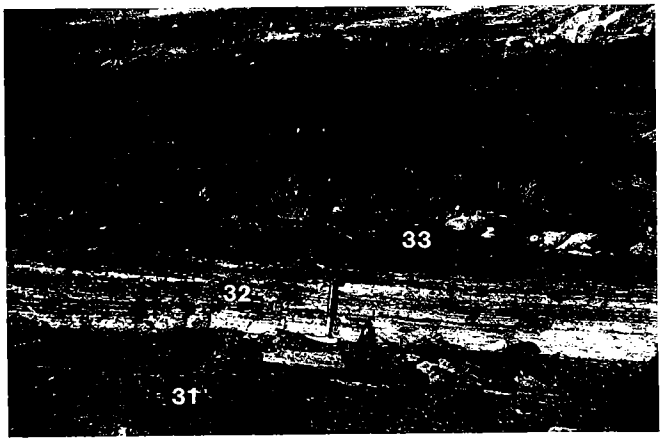


Figure 23. Upper part of Unit 31, Unit 32, and lower part of Unit 33 (836–860-ft interval). A 1.3-ft-thick sandstone lithofacies (Unit 32) is encased by two shale lithofacies, Units 31 and 33. Unit 32 contains abundant trace fossils and brachiopods. Geologic pick is 1.1 ft long.

Transgressive Facies

On the whole, it is thought that the succession was deposited as part of a transgressive system. The vertical facies succession implies upwards deepening. The thick, mud-dominated (i.e., dark-gray, noncalcareous, slightly fissile shales) part of the succession records deposition from suspension in a relatively low-energy, moderate- to deep-marine setting. A subsiding basin during this relative sea-level rise provided accommodation space for accumulation and preservation of thick shale lithofacies within the upper part of the Atoka Formation. Continued subsidence and/or transgression resulted in a repetition of the infilling cycle. Storm-activity periodically interrupted the background shale sedimentation in the basin. The transgressive event was probably relatively slow overall as indicated by the preservation of sandstone beds encased in marine-shale lithofacies (Figs. 23,24).

As transgression continued, thick, dark gray, organic-rich, essentially nonbioturbated shales were deposited. The overall increase in organic-carbon content and the absence of burrowing organisms suggest that the bottom of the water column may have been oxygen-deficient. However, these conditions may have been present only at the sediment-water interface. Sparse horizontal burrowing within thin siltstone/sandstone interbeds is interpreted to represent brief oxygenated events resulting from storm-influenced currents. The oxygenated waters allowed a short-lived benthic community to colonize the substrate.

Regressive Facies

During a relative sea-level fall, base level is lowered and shelf sequences may record a shallowing-upward (coarsening-upward) sequence. At maximum sea-level fall, storms may effectively transport sediment far offshore. Sandstone lithofacies within the upper part of the Atoka Formation, at least at this stratigraphic level and geographic setting, are interpreted to record regressive intervals deposited during sea-level fall within an overall deepening-upward transgressive (flooding) event. Sandstones were deposited during sea-level lowstands when sand prograded across a mud-dominated shelf.



Figure 24. Unit 28. Lenticular, very fine grained sandstone body encased by bounding shale packages (i.e., Units 27 and 29). Hoe pick is 2.2 ft long.

Storm activity influenced transportation and deposition of sediment on the shelf within the basin. One possible mechanism of sand transportation, storm-induced geostrophic flows, would be consistent with the evidence for oscillatory currents. These currents were probably too weak to transport much sand except when enhanced by storm currents. Periodic flooding (storm events) during which sediments were transported seaward out of brackish, marginal-marine bays and marshes is indicated by: (1) the presence of crinoid stems and disarticulated brachiopod and bivalve shells, (2) wood fragments, and (3) organic fossil hash in sandstones.

The increasing frequency and upward thickening of sandstone beds, the diversity of ripple-bed forms, and the presence of small-scale cross-stratification suggest deposition in a relatively shallow marine-shelf setting in a zone seaward of fair-weather wave base, but within storm-wave base. Relatively deeper water, more distal shelf shales abruptly overlie each of these shallow-water, more proximal shelf sandstone lithofacies.

An inferred chronological development of the succession can be proposed and might include: (1) an initial transgressive event with a relative sea-level rise and concomitant, moderate reworking of sand in a shallow-marine setting; (2) a major transgression with subsequent flooding of the mud-dominated shelf and deposition of thick, open-marine shales; and (3) periodic interruption of background shale deposition by regressive events with relative sea-level falls and deposition in certain intervals of coarsening-upward sandstone lithofacies by storm-influenced transportation of sand across a mud-dominated shelf.

EXPLORATION SIGNIFICANCE

The key to effective exploration and to production and ultimate recovery of hydrocarbons from these sandstone reservoirs is the recognition of depositional controls and facies. Deep-sea fan deposits, deltaic sequences, and shelf deposits result from very different processes; they also have vastly different paleogeographic significance. But, because the three share a wide variety of common sedimentary features, distinguishing among these depositional controls and facies may be difficult. Therefore, the ability to identify the differences among turbidite-sandstone facies, deltaic-sandstone facies, and storm-influenced shelf sandstones will be a significant aid in successfully predicting the occurrence and trend in the Arkoma Basin/Ouachita Mountains frontal belt of sandstones that have hydrocarbon-production potential.

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		Return to vehicles and retrace route back through Wister on U.S. 270 to the intersection with U.S. 271.
12.3	4.1	Turn left (west) and proceed along U.S. 270/271.
12.9	0.6	Cross Mountain Creek and continue due west on U.S. 270/271. Low ridges on both sides of the road are formed by unnamed sandstone units in the Savanna Formation. (The Savanna Sandstone is described at Stop 17A.)
13.7	0.8	Leave Wister 7.5′ quadrangle, enter Summerfield 7.5′ quadrangle.
16.4	2.7	Turn right (north) at community of Victor.
16.5	0.1	Spoils from abandoned strip pits in the Cavanal coal are visible to the right (east) of the vehicle. Coal was mined from these pits at some unknown time prior to 1931. Hendricks (1939) did the field work for his publication on the geology and fuel resources of this area during the summer of 1931. His map shows the mine as abandoned at that time. However, he did measure sections of the Cavanal coal in mines that were active in 1931 (Hendricks, 1939, pl. 35). The coal varies in thickness from 1 ft 8 in. to 3 ft 2 in. in the area. Named mines in the vicinity mapped (and 1931 status noted) by Hendricks (1939, pl. 35) include: the Lewis Slope (SW¼ sec. 28, T. 6 N., R. 23 E., abandoned); Wister Coal

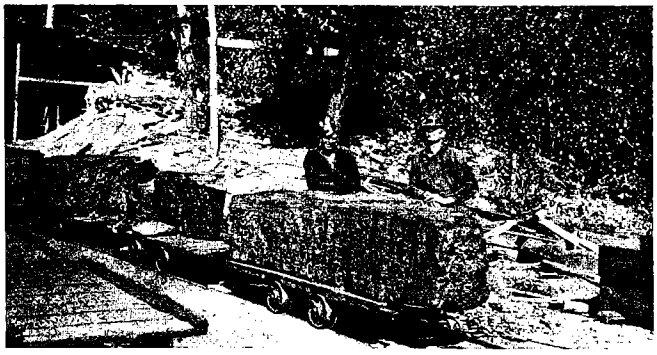


Figure 25. Blocks of Cavanal coal on mine flat cars (circa 1931). Oakland Coal Co.'s mine, sec. 4, T. 6 N., R. 25 E. (from Hendricks, 1939, pl. 32B).

Co. (NE¼ sec. 26, T. 6 N., R. 24 E., abandoned); Walls Slope (SE¼ sec. 8, T. 6 N., R. 25 E., active); Oakland Coal Co. (SE¼ sec. 4, T. 6 N., R. 25 E., active); and Poteau Coal Corp. (NW¼ sec. 35, T. 7 N., R. 25 E., active). Figure 25 is a photograph showing blocks of Cavanal coal being hauled from the Oakland Coal Co.'s mine. Analytical data from Friedman (1974, p. 112) show that the Cavanal coal in Le Flore County is medium-volatile bituminous and has an average sulfur content of 3.3%.

Continue north on Savanna Formation.

17.2	0.7	Foot of Goat Ridge, a 150–190-ft escarpment formed by the lower unit of the Bluejacket Sandstone. Ascend escarpment. Park at edge of road at top of ridge.
17.8	0.6	Stop 4.

STOP 4

Bluejacket Sandstone Member of Boggy Formation

Location: top of Goat Ridge, C WL NW¼NW¼ sec. 21, T. 6 N., R. 24 E., Le Flore County

The Bluejacket Sandstone (Bartlesville sand of the subsurface) in the field-trip area generally consists of two major sandstone units separated by a shaly interval of variable thickness. The sandstone units contain numerous interbedded sandy shale and siltstone layers. Measured Section, Stop 4, and Figure 26 give a description of the lower unit of the Bluejacket Sandstone in the Goat Ridge escarpment.

The Boggy Formation was named by Taff (1899), probably from extensive outcrops in the valley of Clear Boggy Creek in Pontotoc, Coal, and Pittsburg Counties, Oklahoma. He set the upper limit at the base of the Thurman Sandstone. The upper contact of the Boggy Formation is of no consideration in the field-trip area because the top of the Boggy is everywhere eroded. The thickest section preserved nearby is on Cavanal Mountain, where the Boggy is ~3,000 ft thick (Webb, 1960, p. 25). The Boggy Formation consists of alternating units of very fine to fine-grained, medium- to thick-bedded, moderate brown to moderate yellowish brown sandstones, brownish gray siltstones, dark gray shales, and three named coal beds—the Lower Witteville, the Secor, and the Secor Rider.

The Bluejacket Sandstone is the basal member of the Boggy Formation. It was named for the town of Bluejacket, Craig County, Oklahoma, in an unpublished manuscript by Ohern (1914, p. 28). He said that its typical development is found in the hills west of the town, where its total thickness is 50–60 ft. There, it may occur as a solid mass of sandstone, but usually it is broken up into several beds by intervening shales.

Measured Section, Stop 4

Lower unit of the Bluejacket Sandstone Member of Boggy Formation

SE¼SE¼ sec. 20, T. 6 N., R. 24 E., and E½NW¼ sec. 21, T. 6 N., R. 24 E., Le Flore County. Measured from crest of ridge in ditches along winding road in south-facing escarpment of Goat Ridge to a point ~120 ft north from driveway west of road at foot of slope.

	Thickness (feet)
BOGGY FORMATION:	
Bluejacket Sandstone Member:	
13. Sandstone, pale yellowish brown (10YR6/2), very fine grained, very thin bedded, micaceous and flat-bedded at base, grading upward to fine-grained, grayish orange (10YR7/4) to very pale orange (10YR 8/2), medium- to thick-bedded, irregular- to wavy-bedded sandstone (80%) interbedded with shaly sandstone; base gradational; top eroded	9.7
12. Shale (60%) interbedded with very fine grained, thin-bedded, wavy-bedded sandstone and siltstone; moderate reddish orange (10R6/6) to moderate reddish brown (10R4/6); poorly exposed	6.5
11. Sandstone, grayish orange (10YR7/4), stained dusky yellowish brown (10YR2/2) and light brown (5YR 5/6), very fine grained, medium- to thin-bedded; irregular- to wavy-bedded in part; trace fossils on some soles; includes rare plant impressions; base gradational	5.4
10. Shale (80%), very pale orange (10YR8/2) to grayish orange (10YR7/4) with light brown (5YR5/6) staining, noncalcareous, very silty, interstratified with siltstone, coarsens upward	10.5
9. Covered interval	6.5
8. Sandstone (75%), moderate yellowish brown (10YR 5/4) to pale yellowish brown (10YR6/2), very fine grained, micaceous; contains black (N1) carbonized, comminuted plant material; thin- to medium-bedded; wavy-, parallel-bedded; shaly; interference ripple marked; base sharp	4.6
7. Siltstone to very silty shale, light gray (N7) to grayish orange (10YR7/4), noncalcareous, micaceous, contains small ovate concretionary structures, base sharp	20.9
6. Shale (60%), dark yellowish orange (10YR6/6) to moderate reddish orange (10R4/6), silty, noncalcareous; interbedded with thin-, wavy-, lenticular-bedded, micaceous, very fine grained sandstone and siltstone; base gradational	37.2
5. Siltstone to very fine grained sandstone (75%), light gray (N7) to grayish orange (10YR7/4) with light brown (5YR5/6) staining; interbedded with yellowish gray (5Y7/2) silty shale, stained light brown (5YR5/6); very thin to thin-bedded; wavy-, parallel-bedded; interference ripple marked; micaceous; coarsens upward; some trace fossils on soles and asymmetric ripple marks on sandstone beds in upper part of unit; base gradational	5.2
4. Shale (90%), medium dark gray (N4) to yellowish gray (5Y7/2) with light brown (5YR5/6) staining, noncalcareous; interlaminated with lenticular-bedded siltstone and very fine grained, micaceous sandstone; weathers to small flakes on the outcrop; basal contact sharp	3.1
3. Sandstone (90%), pale yellowish brown (10YR6/2) to	

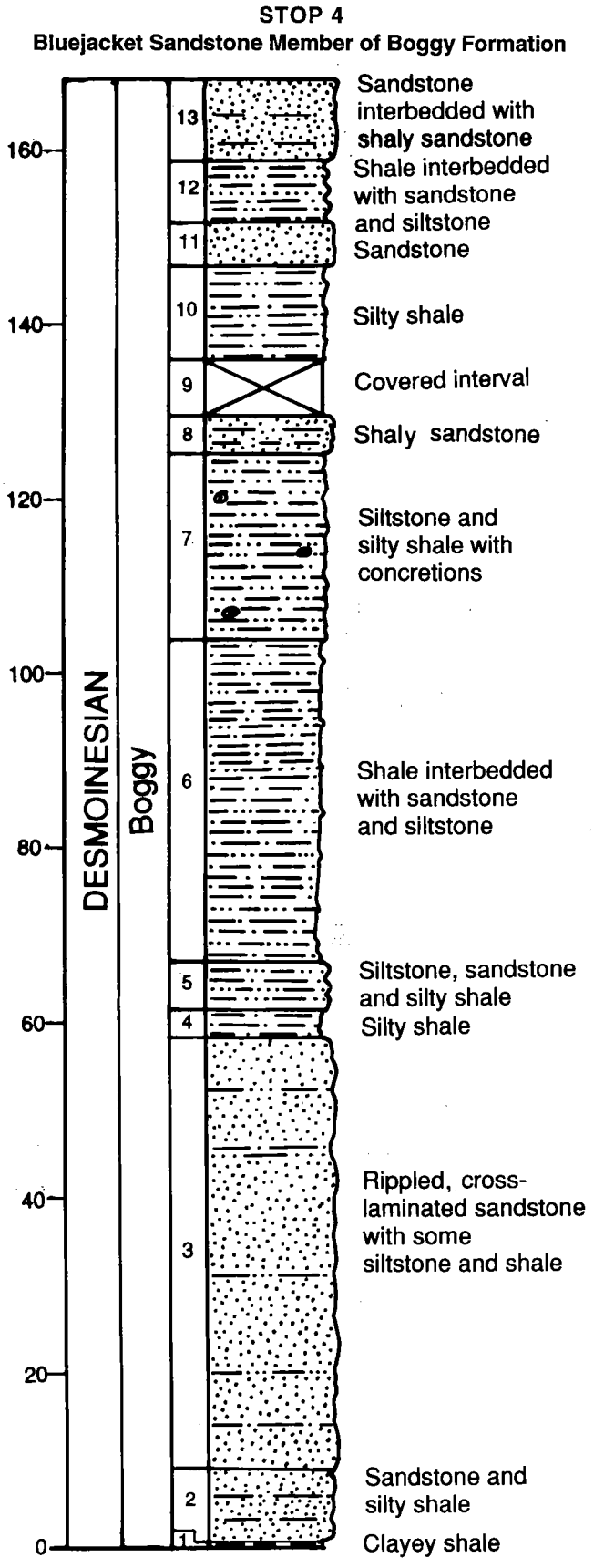


Figure 26. Graphic columnar section of the lower part of the Bluejacket Sandstone Member of the Boggy Formation at Stop 4.

light brown (5YR5/6), very fine grained, noncalcareous, very micaceous, carbonaceous flecks concentrated in some ripple troughs; interference ripple marked; contains low-angle, small-scale, cross-laminations; includes rare trace fossils; interbedded with shale and siltstone in some intervals; mostly thin-, parallel-bedded; base gradational .. 49.4

2. Sandstone (75%), grayish orange (10YR7/4) to very pale orange (10YR8/2) with moderate reddish orange (10R6/6) mottling; very fine grained; micaceous; noncalcareous; contains small-scale cross-laminations; ripple-marked; interference ripples common; interbedded with yellowish gray (5Y8/1) silty shale with moderate reddish orange (10R6/6) mottling; base gradational 8.4

1. Shale, clayey, yellowish gray (5Y8/1), noncalcareous; interlaminated with moderate reddish brown (10R4/6) ferruginous siltstone. (Unit exposed by excavating alluvium in road ditch below Unit 2; underlain by a poorly exposed sandstone ledge that appears to be in place, but slumped. May be base of Bluejacket Sandstone) 0.6

Total thickness of section 168.0

In the field-trip area, the lower unit of the Bluejacket Sandstone is ~168 ft thick (Fig. 26). The middle unit, exposed to the northwest at Mile 19.3, is ~22 ft thick and consists mostly of shale (Fig. 27). It includes the Lower Witteville coal, which is not of minable thickness in this area. However, the same coal bed is 4 ft 8 in. thick on the east side of Cavanal Mountain (Taff and Adams, 1900, p. 294). The upper unit of the Bluejacket Sandstone is ~39 ft thick and consists of two resistant sandstones separated by ~20 ft of nonresistant shale (Fig. 27).

Measured Section, Mile 19.3, Day One

Part of Boggy Formation

E½SE¼ sec. 18, T. 6 N., R. 24 E., and W½NW¼ sec. 17, T. 6 N., R. 24 E., Le Flore County. Measured along Little Caston Creek from low waterfall to just south of bridge on county road.

Thickness (feet)

KREBS GROUP:

BOGGY FORMATION:

25. Sandstone, moderate brown (5YR3/4) to moderate yellowish brown (10YR5/4), very fine grained, medium- to thick-bedded; contains dish-and-pillar structures; cross-bedded, locally fills 6-ft-wide channels incised ~10 in. into underlying strata; thickness variable laterally; lower contact sharp 1.5
24. Siltstone, light brownish gray (5YR6/1), with light brown (5YR5/6) staining, micaceous; contains black comminuted plant material; interlaminated with very fine grained sandstone 2.7
23. Shale, dark gray (N3), brittle, flaky; contains ironstone concretions; poorly exposed 17.0
22. Covered interval 21.0
21. Coal, black (N1), finely cleated, moderately friable; exposed in stream bed; thickness difficult to measure accurately (Secor coal) 1.6
20. Underclay, medium gray (N5) with dark yellowish

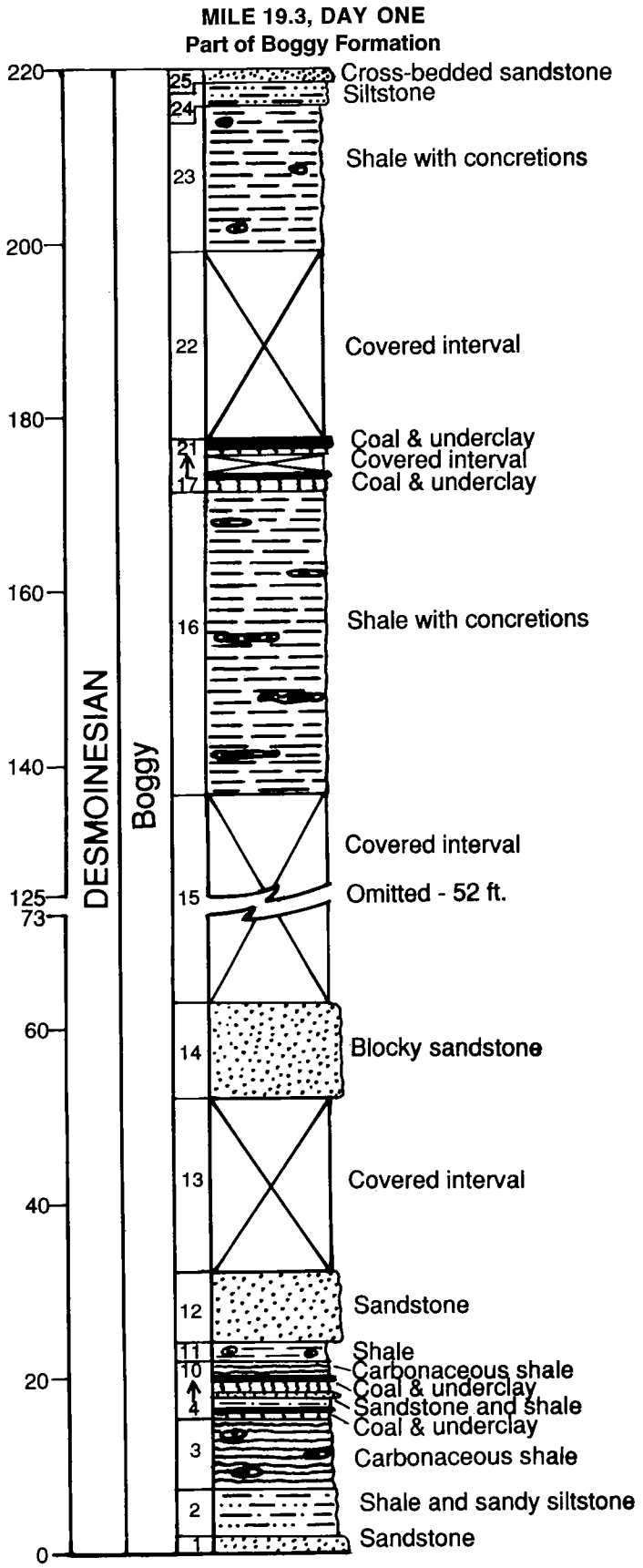


Figure 27. Graphic columnar section of lower part of the Boggy Formation including Lower Witteville and Secor coals at Mile 19.3.

orange (10YR6/6) streaks; contains carbonized plant remains and coaly material 1.0

19. Covered interval 1.8
18. Coal, black (N1), finely cleated; contains dense, pyritic coal masses in upper part that are as much as 9 × 7 in. and contain well-preserved plant fossils (Secor coal) 0.8
17. Underclay, olive black (5Y2/1) with light brown (5YR 5/6) staining, compact, contains abundant carbonized plant material 1.2
16. Shale, olive gray (5Y4/1) to medium dark gray (N4), brittle, flaky, weakly bioturbated; contains grayish red (10R4/2) ironstone concretions in discontinuous 1-in.-thick layers 35.0
15. Covered interval 74.0
14. Sandstone, grayish orange (10YR7/4) to grayish red (10R4/2), very fine grained, medium- to thick-bedded, irregular-bedded, blocky (upper part of Bluejacket Sandstone Member) 10.5
13. Covered interval 20.0
12. Sandstone, grayish orange (10YR7/4) with light brown (5YR5/6) and moderate reddish orange (10R6/6) staining, very fine grained, thin-bedded, cross-laminated in part; includes fossil-plant casts and ironstone concretions; breaks into blocky, irregular masses in upper part 8.2
11. Shale, medium gray (N5), soft, flaky; contains abundant dark yellowish (10YR6/6) orange clay-ironstone nodules 2.3
10. Shale, brownish black (5YR2/1) with pale brown (5YR 5/2) and dark yellowish orange (10YR6/6) bands, very carbonaceous, soft, flaky; includes well-preserved *Lepidodendron* compressions as well as other plant fossils 2.0
9. Coal, grayish black (N2) with light brown (5YR5/6) bands, impure, shaly (Lower Witteville coal) 0.5
8. Underclay, light brownish gray (5YR6/1) with moderate reddish orange (10R6/6) streaks; slickensided; contains carbonized plant fragments 1.1
7. Sandstone, moderate reddish brown (10R4/6) to dark yellowish orange (10YR6/6), very fine grained, no visible stratification; occurs in pod-like layers varying from 2 to 6 in. thick; strongly bioturbated; root casts abundant 0.4
6. Shale, medium dark gray (N4) with abundant dark reddish brown (10R3/4) and dark yellowish orange (10YR6/6) ferruginous siltstone layers, weathers to brittle flakes on the outcrop 1.4
5. Coal, black (N1), impure, very shaly, soft 0.3
4. Underclay, light gray (N7) with dark yellowish orange (10YR6/6) bands 0.8
3. Shale, olive black (5Y2/1), weathers medium light gray (N6), flaky, carbonaceous; contains well-preserved fossil plant compressions such as fern leaves and *Annularia*; includes abundant dark reddish brown (10R3/4) ironstone concretions. 8.0
2. Shale, dark yellowish brown (10YR4/2), noncalcareous, hard, brittle, interbedded with sandy siltstone layers and lenses 1–2 in. thick 5.5
1. Sandstone, light olive gray (5Y6/1), weathers grayish orange (10YR7/4) and grayish red (10R4/2), very fine to fine-grained, thin- to medium-bedded; contains black carbonized plant fragments; mostly irregular-bedded, but includes some ripple-marked layers; contact with overlying unit sharp; base covered (upper part of lower unit of Bluejacket Sandstone Member) 1.5

Total thickness of section 220.1

A Brief History of Coal Mining in Oklahoma

Nuttall (1821, p. 146–177) recorded the presence of coal in what is now Oklahoma as early as 1821, but mining on a commercial scale did not begin until railroads were built in 1872. McAlester became the hub city for the Missouri–Kansas–Texas Railroad. Branch lines were built to haul coal from nearby mines to the main line at McAlester; similarly, commercial-scale mining in other parts of the coal field was made possible by the arrival of railroad lines (Trumbull, 1957).

In 1872, the Choctaw, Oklahoma, and Gulf Railroad was built eastward from McAlester through Hartshorne, Wilburton, Howe, and other points. Later, extension of the line eastward to Memphis, Tennessee, and westward across Oklahoma, widely increased the market for coal. The St. Louis–San Francisco Railway was built across the east side of Indian Territory (Oklahoma) about 1885. Building of the Kansas City Southern Railway followed, and the coal field was linked to the Gulf of Mexico at Port Arthur, Texas (Trumbull, 1957).

Statistics on the coal trade in Oklahoma (Indian Territory) in two of the early years, 1887–88, were published by Ashburner (1890, p. 124). He listed coal production as 685,911 short tons for 1887 and 761,986 short tons for 1888. In 1888, 1,700 men were employed by the industry in the Indian Territory, according to Ashburner (1890, p. 137), and the value at the mine for "soft coal" was $1.95 per ton.

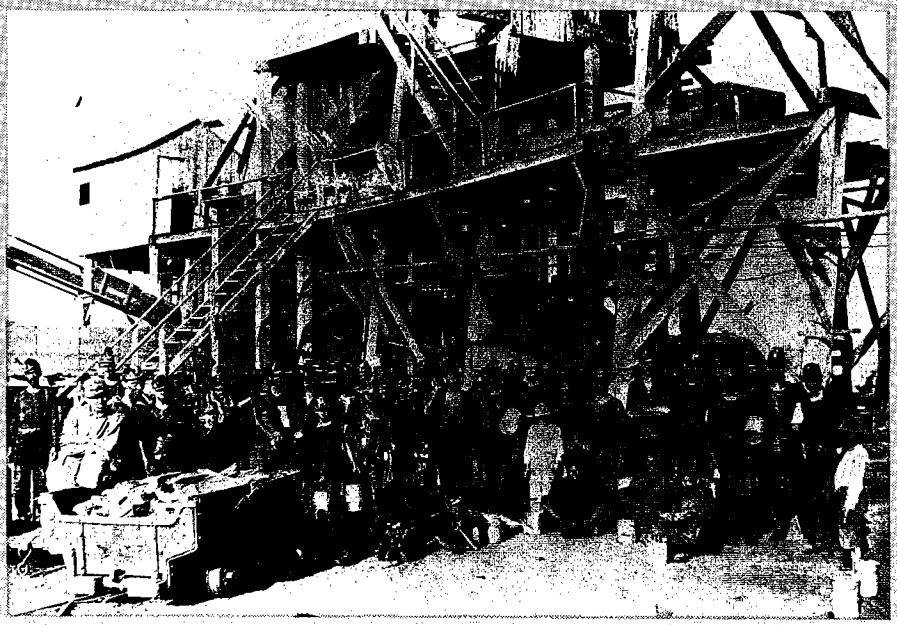
Even at that time, wages paid for mining coal were "the subject of an almost constant dispute between the employer and the employee" (Ashburner, 1890, p. 135). Wages were based primarily on the market price received for the product of the mine. If the employer did not meet wage demands—and mines could not be kept in active operation at a direct loss to the operator—miners would strike. Ashburner (1890, p. 135) eloquently expressed the difficulties of the times:

> The inevitable result of such a strike is that while the operator loses much money during these periods, in case there is a legitimate demand for his coal, yet the actual personal suffering is at all times infinitely greater to the miner and his family, whose distress for the want of proper food and clothing, and even at times shelter from the weather, is frequently heartrending.

Average wages in 1888 were 80¢ per net ton per miner. The coal miners worked an average of 300 days per year. Their work consisted of removing the coal from the bed and placing it on railroad cars and wagons at the mouth of the mine for shipment to market (Ashburner, 1890, p. 136–137).

As a result of the completion of the four new railroad lines in the late 1880s, and of the increased market for coal, coal production in Oklahoma had reached a million tons per year by 1891. By 1903, production from 117 mines throughout the coal field exceeded 3.5 million tons per year (Trumbull, 1957, p. 362).

Between 1873 and 1883, there were about 1,000 coal miners in the Indian Territory. Ten years later, there were twice as many, and, by 1904, the number of coal miners stood at more than 8,000. In the early history of coal mining in the Indian Territory, most of the miners were immigrants from the British Isles. However, around the turn of the century, more and more immigrants from southern and eastern Europe arrived to work in the coal mines and find a home (Hightower, 1985).



Deep-shaft coal mining in the Indian Territory lured three generations of immigrants to a life of coal dust, back-breaking labor, and constant danger. (Photograph courtesy of the Oklahoma Historical Society.)



In the early days of coal mining, miners used canaries to detect deadly gases. A dead bird meant extreme danger! (Photograph courtesy of the Oklahoma Historical Society.)

Mining towns in the Indian Territory were built to imitate those in Pennsylvania. Miners were paid in scrip that was used as legal tender and backed by the mining companies, most of which were owned by the railroads. The economic system has been described as semi-feudal. Communities were totally dependent on the production of coal. It was not until the 1920s that company stores began to decline in importance and a measure of free enterprise was introduced (Hightower, 1985).

From 1900 through WWI, Oklahoma was an important coal producer. Coal was a major fuel in Oklahoma and a major ingredient in steel production in adjacent states (Friedman, 1974, p. 44). During the past 120 years, coal production in Oklahoma has gone through a series of cycles, generally controlled by demands of steel manufacturing for fuel and coke.

Prior to 1920, production of coal by strip mining was insignificant. In 1920, 95% of all coal mined was produced by underground methods. Since then, the trend toward surface mining has increased steadily; at present, only one underground mine is operating.

Oklahoma's underground mines—where the men were surrounded by crude machinery and worked by maneuvering in near darkness—were among the most dangerous in the country. The bituminous coal produced a great deal of fine dust that was inhaled by the miners. Explosions were common. Gas seeps in mine shafts could be deadly, so miners used birds as an early-warning system.

The immigrant miners were thought of as no better than pit mules. They went into the coal mines at daybreak and came out after dark. There were no eight-hour days. It was not unusual for the miners to walk 2 or 3 miles to the mines, put in a day's work, then—sweaty and grimy—walk home. Beginning wages were $1.15 per day. Experienced miners were paid somewhat better (Hightower, 1985).

The coal mining industry in Oklahoma has had a history of disasters, in which a number of men have lost their lives. Much of the blame for coal mining accidents rests with mine operators whose lack of safety precautions contributed to hundreds of injuries and deaths in the early days of mining. Gas and dust explosions in underground mines were the primary cause of the disasters. The earliest records of deaths go back to 1885 when 13 men lost their lives in a mine at Krebs. Other disasters followed; a few of the major ones are listed here: in 1892 another underground mine explosion at Krebs killed 96 men; in 1912, 73 men were killed in a mine at McCurtain; and, in 1926, 91 men were killed in a mine at Wilburton (Oklahoma Department of Mines, 1988).

Earliest published mapping and discussion of coals in the field-trip area were by Chance (1890, p. 653–661). He wrote a description of the Choctaw coal field, which, in general, included the area from just west of McAlester to the Arkansas state line. It was bounded on the south by the Choctaw fault and, on the north, by the San Bois Mountains and Cavanal Mountain. As would be expected, subsequent, more detailed work revealed many errors in his mapping. Names he had given coal beds were changed or abolished as geologic work progressed. For example, the name "Grady" coal was dropped in favor of "Hartshorne" coal, and the "Mayberry" coal was shown to be equal to the "Secor" coal. The "Norman" coal could not be identified, and the name was never used again (Oklahoma Geological Survey, 1954).

In 1899, 1901, and 1903, the U.S. Geological Survey mapped the geology of parts of the Choctaw coal field. In 1900, the report, "Geology of the Eastern Choctaw Coal Field, Indian Territory," was published (Taff and Adams, 1900). A second report on the geology of coal in Indian Territory, Arkansas, and Texas, entitled "The Southwestern Coal Field," was published in Part III of the Twenty-Second Annual Report of the U.S. Geological Survey (Taff, 1902b).

Many familiar stratigraphic names, such as "Atoka Formation," "Hartshorne Sandstone," "McAlester Shale," "Savanna Formation," and "Boggy Shale," had their origins in these reports (Taff, 1902b; Taff and Adams, 1900). Considering the limited resources available at the time to aid field workers in their research, as well as the adverse working conditions in the region, the quality of this early work is truly remarkable.

18.2	0.4	Continue north on dip slope of lower part of Bluejacket Sandstone to four-way intersection and turn left (west) at corner.
19.3	1.1	Cross Little Caston Creek. Measured section, Mile 19.3, Day One (see p. 40) was measured along the creek on both sides of the road. The Lower Witteville coal, which is only 0.5 ft thick here, crops out along the road just west of the bridge. Rocks dip north-northwest at 5–7°, toward the axis of the Cavanal syncline in this area.
20.2	0.9	Four-way intersection. Turn right (north) and drive to mine office of active strip mine operated by Farrell-Cooper Mining Co.
20.4	0.2	Stop 5.

STOP 5

Wister Mine — Secor and Secor Rider Coals

Location: N½ sec. 18, T. 6 N., R. 24 E., Le Flore County

For permission to visit the mine contact Bob Cooper, Farrell-Cooper Mining Co., (501) 646-4366.

Coal is being produced from three beds at the Wister Mine, operated by Farrell-Cooper Mining Co., Fort Smith, Arkansas. Composite Measured Section, Stop 5 (from Hemish, 1994a, p. 46) (Fig. 28) was taken from the highwalls of the mine in three different cuts on October 21, 1993.

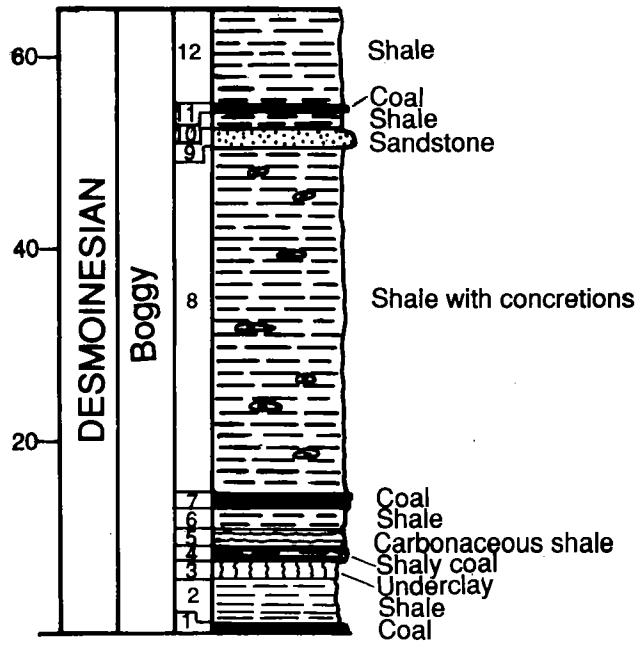


Figure 28. Graphic columnar section of the part of the Boggy Formation that contains the Secor and Secor Rider coals at the Wister Mine (Stop 5).

Figure 29 shows the mining operation as it appeared in October 1993. Topsoil is removed and stockpiled prior to the start of the stripping operation. The overburden is "shot" prior to removal to break the consolidated rock and to facilitate handling. In places, the overburden can be removed without shooting, which reduces mining costs. A dragline is used to remove the thickest overburden, which consists predominantly of shale.

In October 1993, production from the mine was averaging ~30,000 tons/month. The coal is hauled by truck directly from the mine to the AES Shady Point co-generation plant where it is used to generate electricity and CO_2. Each round trip requires ~2 hours (Hemish, 1994a).

In the Farrell-Cooper operation, reclamation of mined-out areas keeps pace with the mining. Overburden material from each new cut is used to fill the preceding cut after the coal has been removed. After the mined-out area has been restored as nearly as possible to its original contour, the topsoil is redistributed, fertilizer is spread, and the area is revegetated. Grasses commonly used for reclamation include fescue and Bermuda.

Composite Measured Section, Stop 5

Part of Boggy Formation that contains Secor and Secor Rider Coals

SE¼NW¼NE¼ sec. 13, T. 6 N., R. 23 E., and SW¼SW¼NW¼ sec. 18, T. 6 N., R. 24 E., Le Flore County. Measured in highwalls of active strip mine operated by Farrell-Cooper Mining Co.

	Thickness (feet)
KREBS GROUP:	
BOGGY FORMATION:	
12. Shale, medium dark gray (N4), upper 3 ft weathered to dark yellowish orange (10YR6/6)	10.0
11. Coal, black (N1), finely cleated, iron oxide staining on cleats (Secor Rider coal)	0.7
10. Shale, dark gray (N3), carbonized plant fragments in upper part	1.8
9. Sandstone, medium light gray (N6), weathers moderate yellowish brown (10YR5/4)	2.0
8. Shale, dark gray (N3) with medium gray (N5) siderite nodules and stringers	36.0
7. Coal, black (N1), finely cleated, moderately friable; includes a 4-in.-thick, dark gray (N3), carbonaceous shale parting in middle part of unit (upper split, Secor coal)	1.6
6. Shale, dark gray (N3)	2.1
5. Shale, grayish black (N2), very carbonaceous	1.8
4. Coal, black (N1), impure, very shaly, finely cleated (middle split, Secor coal)	1.7
3. Underclay, medium gray (N5), slickensided	2.0
2. Shale, dark gray (N3)	4.4
1. Coal, black (N1), finely cleated, moderately friable (lower split, Secor coal)	0.9
Total thickness of section	65.0

The name Secor was first used by Chance (1890, p. 658–660, pl. 1), but the origin of the name is not known. Possibly, the coal bed was named for a "Secor family that lived in the area where the coal was first mined in Pittsburg County, Oklahoma" (Hemish, 1987, p. 105). Taff and Adams (1900, p. 294) described two beds of coal mined at the east end of Cavanal Mountain near the town of Witteville, which they called the "lower Witteville" coal and the "upper Witteville" coal. Subsequently, it was shown by C. C. Branson (Oklahoma Geological Survey, 1954, p. 131) that their "upper Witteville" was stratigraphically equivalent to the Secor, and the name "upper Witteville" was dropped. Although only one of the two original Witteville names survived, the term "lower" was never dropped, so the name "Lower Witteville" persists in published literature (Hemish, 1994b, p. 7).

On the east end of Cavanal Mountain, ~250 ft of shale and sandstone separate the Lower Witteville coal from the Secor coal (Taff and Adams, 1900, p. 294). The Secor coal is 3–4 ft thick there and is separated into two nearly equal benches by a thin parting of shale. Splits develop in the Secor coal to the west along the south flank of Cavanal Mountain. In the Wister Mine, the lower split is 0.9 ft thick and is mined. It is 6.4 ft below the middle split. The middle split is 1.7 ft thick and is too shaly and impure to mine. The upper split is separated from the middle split by ~4 ft of dark gray shale. The upper split is 1.6 ft thick, includes a 4-in.-thick carbonaceous shale parting in the middle of the unit, and is mined.

The third bed that is mined is ~40 ft above the Secor. It is called the Secor Rider and generally varies from 0.5 to 0.8 ft thick. In spite of its thinness, the Secor Rider is valuable because of its superior quality. Its low ash content, low sulfur content, and high heat content make it useful for blending with the lower-quality Secor coal (Table 1).

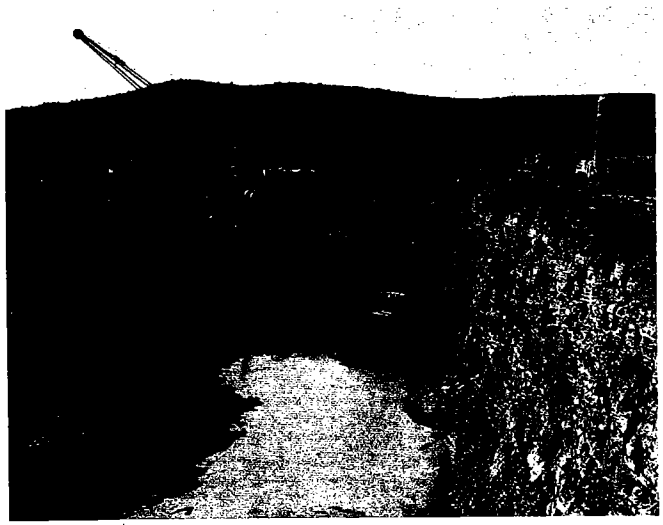


Figure 29. View of the Wister Mine, showing part of the strip-mine operation as it appeared in October 1993. At right is a shothole rig drilling blast holes in consolidated overburden; at base of highwall is a front-end loader stockpiling coal; and at center is a dragline removing overburden.

TABLE 1. — WISTER MINE COAL ANALYSES[a,b]

	Secor			Secor Rider
	Lower Split	Upper Split		
		Lower Bench	Upper Bench	
Moisture	1.73	2.17	2.69	3.95
Ash	13.90	15.30	19.60	3.91
Sulfur	2.40	5.37	5.20	0.92
Btu/lb	12,956	12,892	12,017	14,213

[a]Analytical data, from grab samples, supplied by Farrell-Cooper Mining Co.
[b]Values other than calorific value in wt%.

		From the mine office drive south, ascend the dip slope of the north-dipping Bluejacket Sandstone to crest of Goat Ridge; continue down the curving road to base of the escarpment (Boggy/Savanna contact), cross Savanna Formation strata onto alluvium of the flood plain of Caston Creek to right-angle bend in road.
22.4	2.0	Turn left (east), drive on alluvium and Savanna Formation.
22.9	0.5	Turn right (south) and drive on Savanna Formation to stop sign.
23.0	0.1	Intersection with U.S. 270/271. Turn right (west) onto highway.
23.4	0.4	Cross Caston Creek. Keota Sandstone, Tamaha Sandstone, McAlester coal, and Cameron Sandstone are exposed in creek bed south of bridge. Beds dip north at 22–28° in this vicinity.
24.3	0.9	Intersection of U.S. 270 and U.S. 271. Community of Caston.
24.5	0.2	Turn left (south) on U.S. 271. Follow curving highway to the southwest. Poorly developed Cameron Sandstone exposed in ditch below farm buildings on low ridge east of road. Ascend ridge on dip slope of Warner Sandstone.
25.2	0.7	Stop 6. Park at turnout.

STOP 6

Warner Sandstone Member of the McAlester Formation

Location: crest of ridge, C NW¼SE¼ sec. 35, T. 6 N., R. 23 E., Le Flore County

The Warner Sandstone Member of the McAlester Formation consists of three units over most of the field-trip area. (East and southeast of Heavener, only one unit is mappable). The basal unit is a resistant, 21.5-ft-thick sandstone. The 223-ft-thick middle unit is predomi-

nantly shale but includes minor siltstones and sandstones. The upper unit is a 35-ft-thick, resistant sandstone. Except for a 70-ft-thick covered interval (assumed to be shale) in the middle unit, the entire member is exposed at this stop. Measured Section, Stop 6 gives a detailed description of the rock units shown in Figure 30. The strata dip moderately north (35°) in this vicinity, so the observer must keep in mind that the rocks lowest stratigraphically are those highest topographically. The lower Warner Sandstone is at the crest of the ridge adjacent to the turnout.

The Warner Sandstone Member of the McAlester Formation was described first by Wilson and Newell (1937, p. 37–38). The type locality is ~0.5 mi east of the northwest corner of sec. 21, T. 12 N., R. 19 E., 1 mi north of Warner, Muskogee County, Oklahoma. Wilson and Newell (1937, p. 38) recognized that the Warner "division" consisted of two sandstones and an intervening shale. However, they believed the upper sandstone to be relatively local and nonresistant and that more than one depositional event was involved, so they restricted the term Warner to only the scarp-forming, massive, lower sandstone. Oakes and Knechtel (1948, p. 33) disagreed with this view and stated that "a member is fundamentally a mappable unit, born of expediency, convenience, or necessity," and on that basis they included both sandstones and the intervening shale in the Warner Member.

Mapping by Hemish and others (1990a,b,c) demonstrated that Oakes and Knechtel showed sound judgment in their interpretation. In the Wilburton area, the Warner Member can be mapped only as a single, ridge-forming sandstone. However, to the east in the Panola 7.5′ quadrangle, the member splits into upper and lower units with an intervening shale. Both sandstone units form prominent ridges eastward to the Arkansas state line (and beyond, toward Hartford). East and southeast of Heavener, the two Warner sandstones appear to merge. The Warner ridge on which the Heavener water tower sits (see Mile 12.3, Day Two) is supported by a single sandstone unit. This single unit crops out all along the southern flank of Poteau Mountain and is about 25–30 ft thick (Hemish and Suneson, 1994).

Busch and Link (1985, p. 171–178) showed that the Warner Sandstone was the lower member of a 2,000-mi² delta complex in east-central Oklahoma, which they referred to as the "Booch delta." The Booch sands of the McAlester Formation were interpreted as a series of delta-distributary channels with associated crevasse splays. They mapped a deep channel trending southeast from T. 15 N., R. 11 E., to T. 11 N., R. 14 E. with a channel width of 3–5 mi and a maximum depth of ~240 ft (Busch and Link, 1985, fig. 8-35). Direction of transport was from the north. Their study area was T. 13–15 N., R. 14–16 E., considerably northwest of the present field-trip area. In their opinion, the channel extended farther southeast, into the Arkoma Basin, but they lacked the data for mapping it.

Hemish (1993a, p. 16) examined the Warner Sandstone in Wister State Park, within the field-trip area. He stated that, because of its bedding characteristics, he believed the sandstone was deposited in a high-energy, deltaic-marine, inner-fringe environment. The Warner is typically very fine grained, thin- to medium-bedded, and

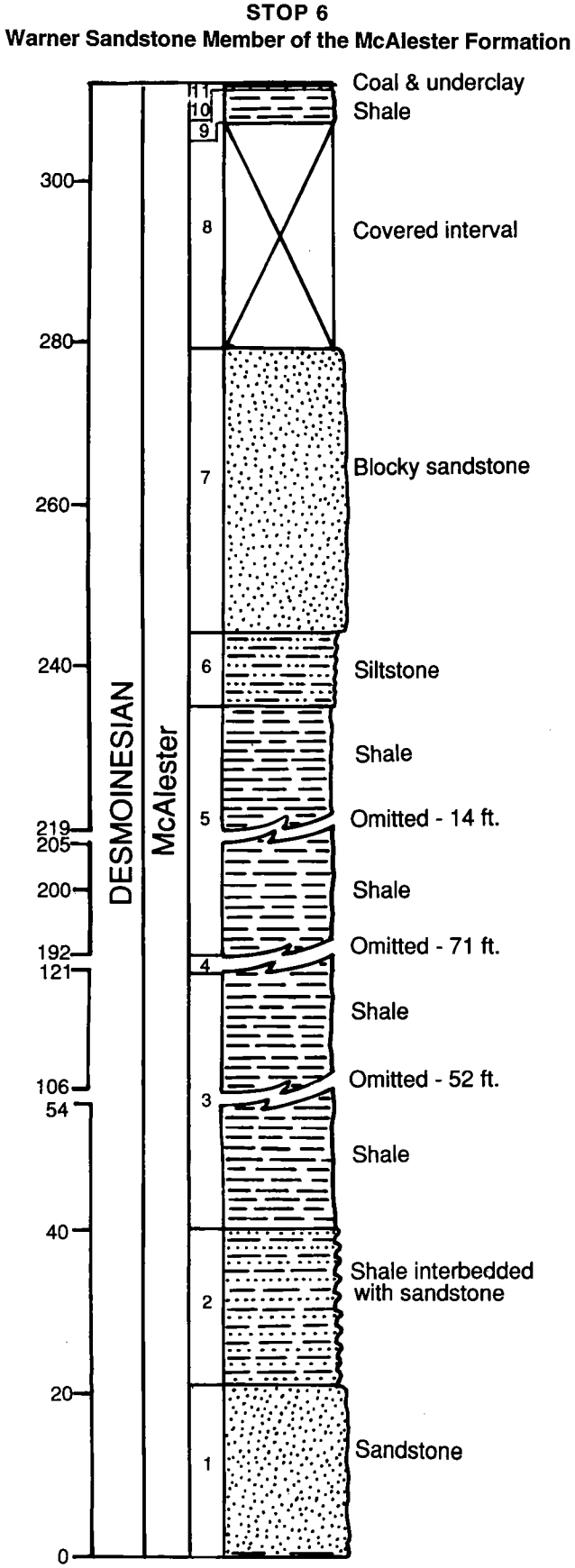


Figure 30. Graphic columnar section of the Warner Sandstone Member of the McAlester Formation at Stop 6.

Measured Section, Stop 6

Warner Sandstone Member of the McAlester Formation

N½N½SE¼ and SE¼SE¼NE¼ sec. 35, and SW¼SW¼NW¼ sec. 36, T. 6 N., R. 23 E., Le Flore County. Measured from driveway northwest side of U.S. Highway 271, across road to outcrop of upper Warner Sandstone, then due southwest obliquely up slope in exposures southeast of road to crest of ridge at observation turnout.

	Thickness (feet)
McAlester Formation:	
Unnamed Shale Member:	
11. Coal, black (N1), very soft, smutty, weathered, thickness uncertain	0.1
10. Underclay, yellowish gray (5Y8/1), poorly exposed, thickness uncertain	0.1
9. Shale, pale yellowish brown (10YR6/2) with grayish orange (10YR7/4) mottling, very weathered; poorly exposed	5.0
8. Covered interval (to top of sandstone exposed in slope southeast across highway)	27.0
Warner Sandstone Member:	
7. Sandstone, moderate yellowish brown (10YR5/4), to moderate reddish brown (10R4/6), very fine grained, noncalcareous, medium-bedded; obscurely wavy bedded, with most surfaces indistinctly ripple marked; blocky; fossil-plant impressions common; becomes thin-, parallel-bedded with rectangular jointing in lower 4 ft; base moderately sharp (upper Warner Sandstone)	35.3
6. Siltstone, grading downward to silty shale, moderate yellowish brown (10YR5/4) to moderate brown (5YR4/4; 5YR3/4), sandy, ripple-marked; wavy-, lenticular-bedded (exposed in south edge of low ridge adjacent to ravine southeast of road cut)	9.0
5. Shale, light gray (N7) to olive gray (5Y4/1), weathers light brown (5YR5/6), flaky, noncalcareous; well-exposed in road cut southeast of highway	42.8
4. Covered interval	70.6
3. Shale (same description as Unit 5)	81.3
2. Shale (80%), interbedded with sandstone. Shale is grayish orange (10YR7/4) to dark yellowish orange (10YR6/6), clayey, very weathered. Sandstone beds are light olive gray (5Y5/2), weathered to moderate reddish orange (10R6/6) and moderate reddish brown (10R4/6), very fine grained, noncalcareous, ferruginous; beds are 2–4 in. thick; contacts with shale are irregular; contains low-angle cross-stratification; poorly exposed	19.3
1. Sandstone, moderate reddish brown (10R4/6), grayish red (10R4/2), moderate yellowish brown (10YR 5/4), and moderate brown (5YR4/4), very fine grained, thick- to medium-bedded, mostly massive with irregular parting surfaces; contains obscure soft-sediment deformation features such as flow rolls and large-scale contorted beds; includes rare, obscure, low-angle cross-bedding; at base of unit, ~1 ft of shaly, lenticular-bedded sandstone is overlain by ~2 ft of thin-, parallel-, wavy-bedded, ripple-marked sandstone (Lower Warner Sandstone)	21.5
Total thickness of section	312.0



Figure 31. Lower unit of Warner Sandstone exposed in the NW¼NW¼NE¼SW¼ sec. 31, T. 6 N., R. 25 E., Le Flore County, in Wister State Park. Note parallel bedding, characteristic of deltaic-marine, inner-fringe sandstones (from Hemish, 1993a, fig. 21).

wavy-, parallel-bedded; it contains low-angle cross-stratification and includes rare plant impressions, as well as trace fossils, on the soles of some beds (Fig. 31).

26.2	~1.0	Return to U.S. 270 and turn left (west) at Caston. Continue west on McAlester Formation shales. Ridges to the north are Savanna Formation sandstones. About 2 mi west of the Caston corner, the Cameron Sandstone forms the first prominent ridge south of the highway.
28.6	2.4	Leave Summerfield 7.5′ quadrangle, enter Leflore 7.5′ quadrangle.
30.5	1.9	Town of Fanshawe, Oklahoma.

FANSHAWE, OKLAHOMA

The post office at Fanshawe was established on March 13, 1891, but it was not until 1902 that the town was surveyed and staked. In the early 1900s, Fanshawe was an active lumbering, coal-mining, sawmilling, farming, and stock-raising center, as well as being a stop on the railroad. In 1919, a fire destroyed all but one store and the town never recovered.

31.4	0.9	Le Flore/Latimer county line—turn left (south) on county line road (blacktop) and drive to first ridge crest.
31.7	0.3	Stop 7.

STOP 7

Cameron Sandstone Member and Associated Coal Beds of the McAlester Formation

Location: C WL SW¼ sec. 31, T. 6 N., R. 23 E., on county line road, Le Flore County

The Cameron Sandstone is well exposed in the road cut to the east of the road, where it is mostly flat bedded, slabby and flaggy, and nearly 40 ft thick. Measured Section, Stop 7 gives a description of the member shown in Figure 32, and Figure 33 is a view of the sandstone at the contact with the underlying shale. Beds dip northwest at 26° at this location. The couplet of subdued ridges ~0.5 mi to the southeast is formed by the upper and lower Warner Sandstones. The next ridge, ~0.6 mi to the south, is formed by the Hartshorne Formation.

The Cameron Sandstone was named for the town of Cameron, northern Le Flore County (Wilson, 1935, p. 507). Wilson (1935, p. 507) described the Cameron in its type locality as a "relatively uniform buff, unfossiliferous sandstone ranging ordinarily from 3 to 5 ft, relatively platy and calcareous and sufficiently resistant to make a distinct low escarpment." The McAlester coal lies a few feet above the Cameron Sandstone wherever it is present. The McAlester (Stigler) coal was named by Chance (1890, p. 656, 658) from the city of McAlester, Pittsburg County, Oklahoma, where the coal was extensively mined in the late 1800s.

In the western part of the field-trip area, the McAlester coal is overlain by a shale section about 25–40 ft thick, at the top of which is another coal bed called the "Upper McAlester." These two beds have been mined on a very small scale for local use in the past, as evidenced by small abandoned strip pits near Fanshawe (Hemish, 1991). The McAlester and Upper McAlester coals have been surface-mined on a much grander scale by the Farrell-Cooper Coal Co. north and east of the town of Red Oak during the 1970s, 1980s, and 1990s. The McAlester coal bed is generally ~2 ft thick in that area and the Upper McAlester, ~1.5 ft thick. The McAlester coal crops out just west of Caston Creek near the town of Wister (Hemish and Suneson, 1993) and probably subcrops to the southeast under the thick alluvium associated with Caston Creek and the Poteau River. It surfaces again near the city of Poteau, in sec. 28, T. 1 N., R. 26 E. (Knechtel, 1949, pl. 1). There is no evidence of occurrences of the McAlester coal (or the underlying Cameron Sandstone) around the base of Poteau Mountain, east of Heavener (Hemish and Suneson, 1994). Elsewhere in the field-trip area, the Cameron generally forms a prominent, but discontinuous, ridge and is ~10–40 ft thick.

Turn around and return to U.S. 270.

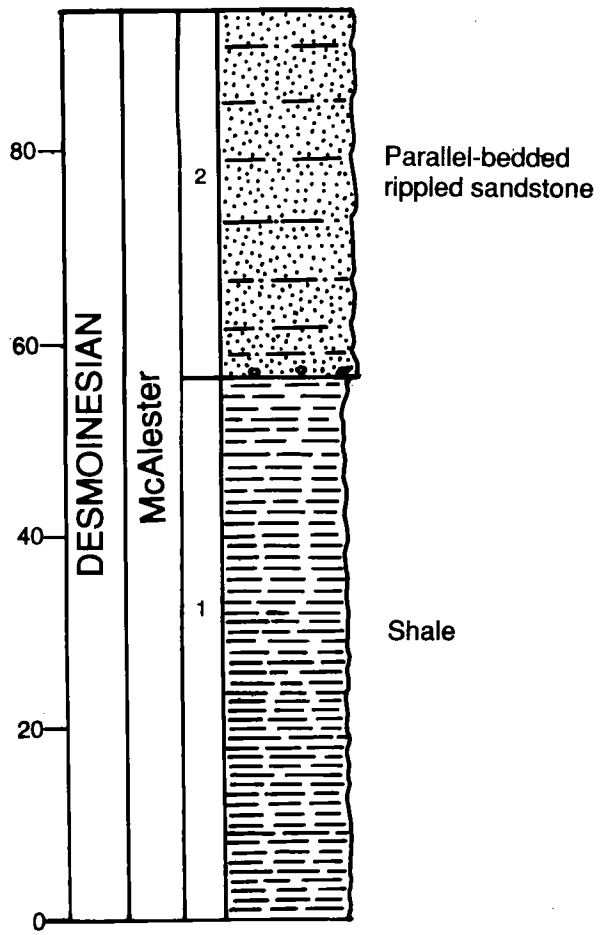


Figure 32. Graphic columnar section of the Cameron Sandstone Member of the McAlester Formation at Stop 7.

Measured Section, Stop 7

Cameron Sandstone Member of the McAlester Formation

SW¼SW¼NW¼SW¼ and NW¼NW¼SW¼SW¼ sec. 31, T. 6 N., R. 23 E., Le Flore County. Measured at crest of ridge and down south slope, east side of blacktop road on county line.

	Thickness (feet)
KREBS GROUP:	
McALESTER FORMATION:	
2. Sandstone, grayish orange (10YR7/4) to light brown (5YR6/4) to very pale orange (10YR8/2), very fine grained, micaceous, includes black macerated plant fragments on some stratification planes, thin- to medium-bedded, generally flat-parallel bedded; ripple marks on some bedding planes; contains low-angle cross-bedding in part; load casts on some soles, shale partings common; slump structures rare; contains some fine-grained conglomerate near base; basal contact sharp (Cameron Sandstone Member)	38.4
1. Shale, moderate yellowish brown (10YR5/4) to pale brown (5YR5/2) in upper part, noncalcareous, clayey, weathers to soft flakes; includes fossil-plant compressions in upper 1 ft, base covered	56.5
Total thickness of section	94.9



Figure 33. Basal part of the Cameron Sandstone Member of the McAlester Formation exposed at Stop 7. Note the flat, parallel-bedded nature of the sandstone and the sharp basal contact with the underlying shale.

32.0	0.3	Turn left (west) and drive westward on McAlester shale.
36.1	4.1	Leave Leflore 7.5′ quadrangle, enter Red Oak 7.5′ quadrangle.
38.0	1.9	Red Oak Peak, capped by Savanna sandstone #6 (see Stop 17A) (Hemish and others, 1990b) on the right (north).
39.5	1.5	On the left (south) is the loading dock for McAlester coal mined by Farrell-Cooper Coal Co. northeast of Red Oak. Isolated peak (Red Oak Mountain), capped by Savanna sandstone #3 (Hemish and others 1990b), lies to the north of the town of Red Oak ~0.5 mi ahead. The axis of the Cavanal syncline trends east-west through the center of the mountain range. The gap between Red Oak Peak and Red Oak Mountain is an area of historical significance. Butterfield overland stage coaches traveled through the pass in the mid-19th century.
40.5	1.0	Intersection with State Highway 82 within the town of Red Oak. McAlester coal was strip mined on a small scale at some unknown time in the past at the southeast edge of town. Continue west through town on U.S. 270. Low ridge just south of the highway is formed by the Tamaha Sandstone. The mountain that abruptly rises more than 1,000 ft just north of the highway is Second Mountain. It is capped by Savanna sandstone #5 (Hemish and others, 1990a, b). The Red Oak–Norris gas field also lies north of the highway (Fig. 34).

RED OAK, OKLAHOMA

A post office was established here on March 11, 1868. The town took its name from a well-known red oak tree that for many years stood in the center of town.

§

RED OAK–NORRIS GAS FIELD

The Red Oak–Norris gas field is the largest gas field in the Arkoma Basin. It is located in northeastern Latimer County, southeastern Haskell County, and west-central Le Flore County, Oklahoma (Table 2). At the present time, the field covers 105.5 mi^2. Production is from several Middle Pennsylvanian sandstones (Fig. 35). Table 3 shows the cumulative gas production from single-completion wells in the Red Oak–Norris field through October 1993. The history of the field is typical of many other Arkoma Basin gas fields: initial discovery on surface structure; later exploratory drilling on the structure based on nearby discoveries of deeper reservoirs; still later drilling off-structure based on improved interpretation of seismic data, sedimentology, and diagenesis.

The discovery well for the field was drilled by the Gladys Belle Oil Co. in 1912 near the crest of the surface expression of the east-west elongated Brazil anticline in sec. 10, T. 6 N., R. 21 E., ~4 mi north of the town of Red Oak. The well followed earlier Arkoma Basin discoveries: the Mansfield gas field in 1901, the Poteau-Gilmore gas field in 1910, and the Cameron gas field in 1911, all of which were drilled on surface anticlines. The Gladys Belle well discovered gas in the Hartshorne Formation at a depth of ~1,500 ft. That well and a few development

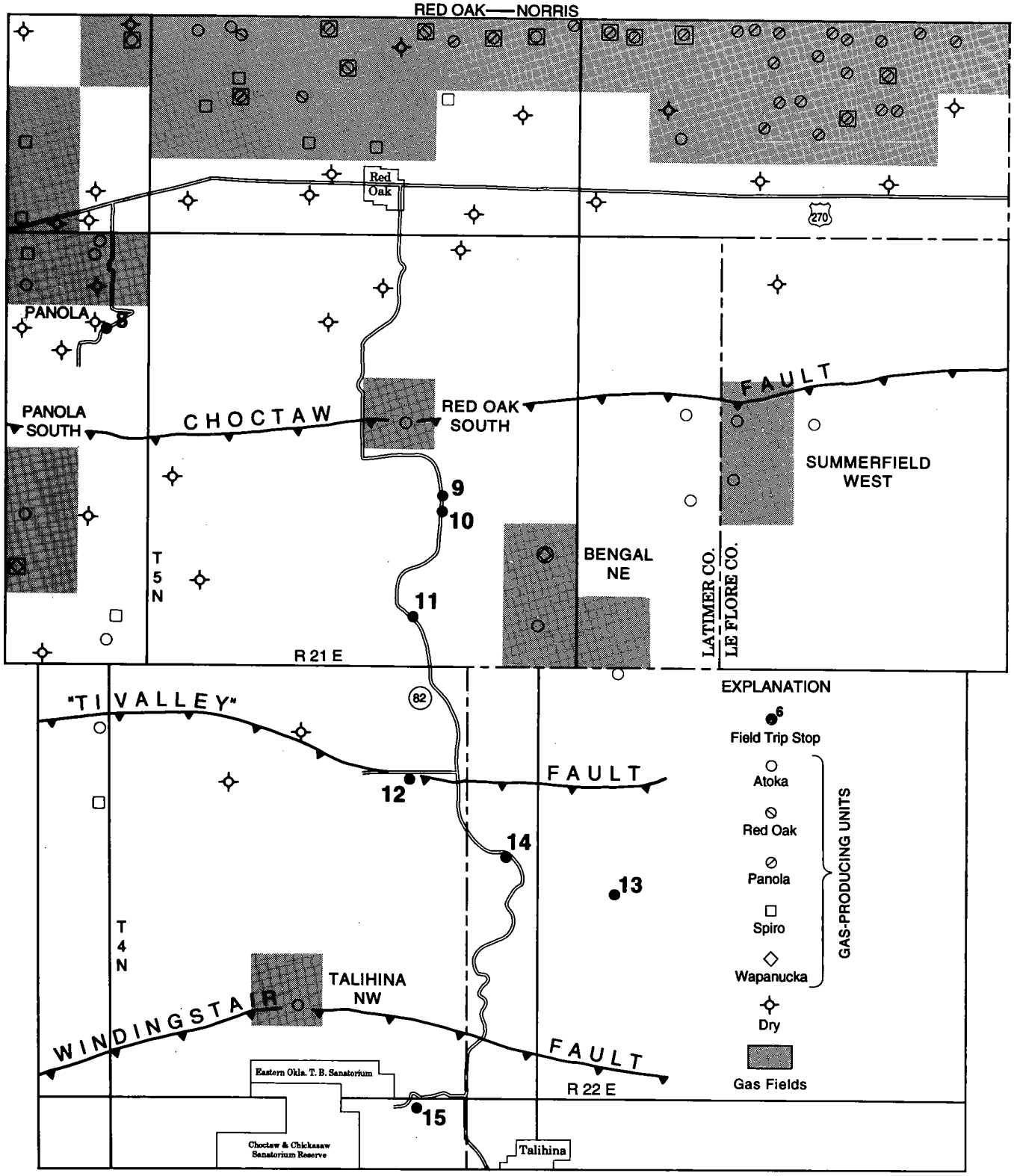


Figure 34. Map showing locations of wells, gas fields, and field-trip stops near Red Oak.

TABLE 2. — LOCATION OF RED OAK–NORRIS GAS FIELD

T. 7 N., R. 21 E.	sec. 36
T. 7 N., R. 22 E.	secs. 25, 26, 31–36
T. 7 N., R. 23 E.	secs. 8, 9, SW¼ 10, 13–24, 26–36
T. 7 N., R. 24 E.	secs. W½ and SE¼ 18, 29, 31
T. 6 N., R. 20 E.	secs. 12–14, 24
T. 6 N., R. 21 E.	secs. 1–3, 7–24, 27–30
T. 6 N., R. 22 E.	secs. 1–24, 26–29
T. 6 N., R. 23 E.	secs. 2–7, NW¼ 8, 9, NW¼ 10, 11, 13, 14, 18

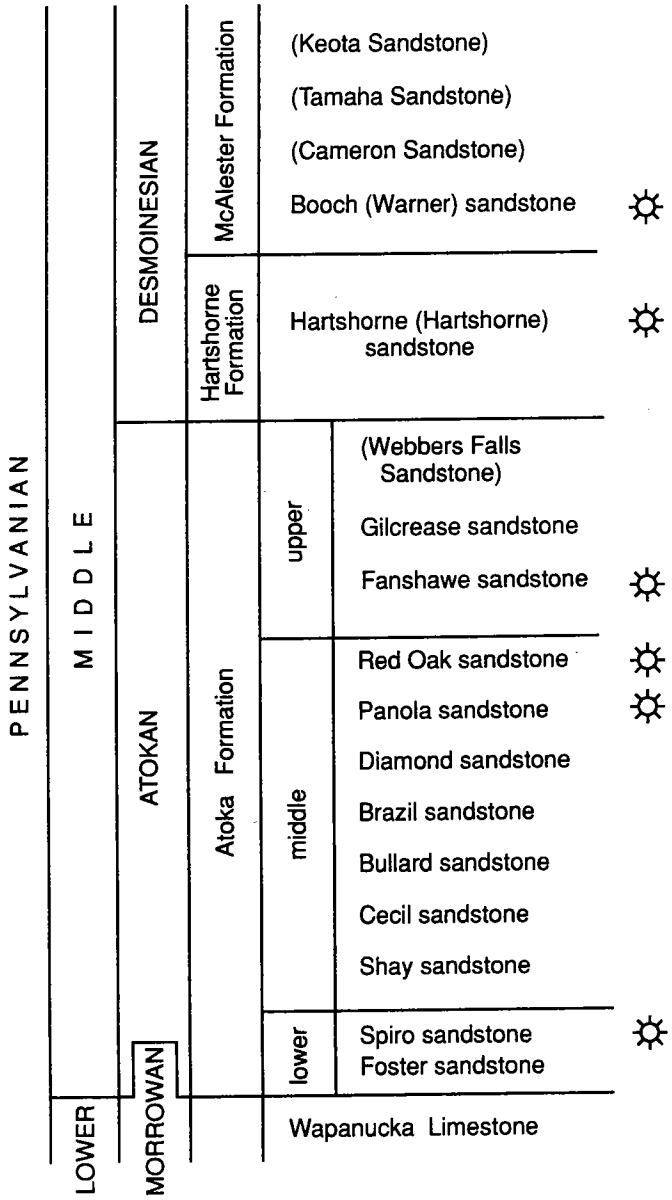


Figure 35. Composite lower Middle Pennsylvanian stratigraphy and principal gas-producing sandstones (denoted by ☼) in the Red Oak–Norris gas field, Arkoma Basin, Oklahoma. Names in parentheses denote surface sandstones.

TABLE 3. — CUMULATIVE GAS PRODUCTION, SINGLE-ZONE COMPLETIONS, RED OAK–NORRIS GAS FIELD

Unit	No. of single completion wells[a]	Cumulative production through October 1993 (MMcf)[b]
Hartshorne	24 (1)	4,551
Fanshawe	21 (3)	60,360
Red Oak	143 (11)	658,552
Panola	4 (1)	2,268
Spiro	33 (6)	116,320

[a]Number in parentheses is number of dual-completion wells that include unit shown.

[b]Production figures from Natural Resources Information System (NRIS), Geological Information Systems, University of Oklahoma.

wells supplied gas to the town of Red Oak for several years, but no other development immediately took place (Houseknecht and McGilvery, 1990).

The Le Flore County Gas and Electric Co. acquired the leases in 1928 and began to develop the Hartshorne along the eastern part of the Brazil anticline. Fifty-five wells were drilled: 43 were completed as gas wells at 1,500–2,100-ft depths. Locally, production was established from sandstones in the McAlester Formation (Booch sandstones) above the Hartshorne (Fig. 35). The Hartshorne still produces gas in the Red Oak–Norris field.

Between 1928 and 1950, several deep tests (maximum 6,510 ft deep) were drilled on the Brazil anticline into Atoka sandstones below the Hartshorne Formation. Gas shows were reported, but all were completed as dry holes (McClain and Planalp, 1961). Between 1924 and 1958, several gas fields north of the Red Oak–Norris field produced gas from the Spiro sandstone at the base of the Atoka Formation along the Milton anticline and subsurface Kinta fault. In 1955, production from middle Atoka sandstones was established at the Gragg field in west-central Arkansas, nearly 50 mi east of the Red Oak–Norris field. This production demonstrated that reservoir-quality sandstones existed in the lower and middle Atoka Formation in Arkoma Basin anticlines (Houseknecht and McGilvery, 1990).

The Red Oak–Norris deep-field discovery was made by the Midwest Oil No. 1 Orr in sec. 8, T. 6 N., R. 22 E. The surface location of the well was on the crest of the Brazil anticline, and drilling began in May 1959. At 7,190 ft (drilled depth), the top of the Red Oak sandstone was penetrated and gas was discovered; the sandstone in the well was 145 ft thick and flowed 11.7 MMcf gas per day on a 0.5-in. choke with 1850 psi tubing pressure on a subsequent production test (McClain and Planalp, 1961). The well continued to drill and the top of the basal-Atoka Spiro sandstone was penetrated at 11,510 ft (drilled depth). The well drilled 73 ft of gross sandstone and a later production test gauged 5.5 MMcf gas per day on a 0.25-in. choke with 3009 psi flowing tubing pressure (McClain and Planalp, 1961).

The Midwest No. 1 Orr was completed officially in April 1960. Within two years, 25 development wells had been completed and ultimate gas recovery was estimated at >1.1 Tcf: the Red Oak sandstone contributed 84% and the Spiro sandstone 16% (Houseknecht and McGilvery, 1990). By the end of 1965, the main production limits of the field had been established. Of the 80 wells drilled, 71 were completed as gas producers: the Red Oak sandstone was productive in 56 wells; the Spiro in 19; the Fanshawe (a local upper-Atoka sandstone) in 11; the Panola (a local middle-Atoka sandstone underlying the Red Oak) in two; and unnamed Atoka sandstones in three (Six, 1968) (Fig. 35).

In the mid-1960s, Amoco acquired most of the Red Oak–Norris field production. In the mid-1980s, Amoco received state approval to increase drilling density from one to two wells per 640 acres. Between 1985 and 1988, 42 infill wells were completed and the gas deliverability was expected to increase from 70 MMcf gas per day to 160 MMcf.

The geology of the Red Oak–Norris field has been described by Houseknecht and McGilvery (1990) and the following discussion is based on that report. The Spiro sandstone represents the final phase of stable-shelf sedimentation that began in the Cambrian. It is a blanket sandstone that is most productive in northwest-trending, fluvially and tidally dominated channels. Interchannel deposits were deposited in shallow, subtidal to tidal-flat environments and are thinner, finer grained, and less porous than the channel sandstones.

Following Spiro deposition, the Arkoma Basin subsided along a series of south-side-down normal (growth) faults. The major fault of this type in the Red Oak–Norris field is the San Bois fault, which has ~7,000 ft of offset at the Spiro level. The Red Oak sandstone accumulated on the footwall of the San Bois fault, adjacent to the fault. The sandstone forms a multistoried, slope-channel deposit trending approximately east-west to northeast-southwest, parallel to the San Bois fault, but perpendicular to the overall slope of the Arkoma Basin.

Movement on the Arkoma Basin growth faults stopped in the late Atokan, but downwarping at a much slower rate continued. Deltaic deposits (Hartshorne Formation) advanced from east to west across the shallow basin. Most Hartshorne production is from distributary-channel sandstones that cap prodelta and delta-front shales and are laterally equivalent to interdistributary-bay deposits, which include coal beds.

Following deposition of the early Desmoinesian Krebs Group, the Red Oak area was mildly compressed. North-directed, bedding-plane thrusts associated with Ouachita tectonism ramped up-section near the San Bois fault but did not offset the Hartshorne. Thrusting caused local repetition of strata in the Atoka and folding at the Hartshorne level (Brazil anticline).

In summary, Hartshorne, Red Oak, and Spiro production at Red Oak–Norris is controlled by structural features superposed on channel location and geometry. In the case of the Red Oak sandstone, channel location was controlled by syndepositional structures.

43.3	2.8	Leave Red Oak 7.5′ quadrangle, enter Panola 7.5′ quadrangle. U.S. 270 bends slightly to the southwest, following a valley underlain by shale in the McAlester Formation. Prominent ridge south of the highway is formed by the Tamaha Sandstone. Museum-quality sole markings are abundant along the Tamaha outcrop belt for ~3 mi from here to where the ridge is truncated by Horse Creek to the west.
44.5	1.2	Craven Corner—turn left (south) onto Craven Road and cross a series of closely spaced ridges. The first ridge south of Craven Corner is formed by the Tamaha Sandstone: dip is 14° northwest here. The next prominent ridge is formed by the Cameron Sandstone (see Stop 7): dip is 16° northwest. The low ridge just south of the Cameron escarpment, <0.5 mi west of the road, is formed by the Lequire Sandstone.
45.3	0.8	Road curves southwest at the base of the dip slope of the upper unit of the Warner Sandstone (see Stop 6). Follow road to crest of ridge and cross shale unit of the Warner Sandstone Member of the McAlester Formation; then cross the lower unit of the Warner Sandstone. Descend into valley formed in the McCurtain Shale Member of the McAlester Formation; then cross low ridge formed by unnamed sandstone in the McCurtain Shale. Dip is 20° northwest here.
46.0	0.7	Road bends sharply left (east) at base of dip slope of Hartshorne Formation. A core hole was drilled by the OGS in the pasture just north of the road in 1984. The lithologic column for the hole is shown in Figure 36, and detailed description is given in Core-hole Log, Mile 46.0, Day One. The Lower Hartshorne coal was penetrated at a depth of ~165 ft. Its thickness at this location is 1.8 ft.

Core-Hole Log, Mile 46.0, Day One

Lower Hartshorne Coal

NE¼SW¼NW¼NW¼NE¼ sec. 34, T. 5 N., R. 20 E., Latimer County. Drilled 50 ft north from edge of gravel road. Surface elevation, estimated from topographic map, 590 ft (from Hemish, 1988, p. 16–17).

	Depth to unit top (feet)	*Thickness of unit (feet)*
Sand, dark yellowish brown, very fine grained; contains organic material	0.0	1.0
Sand, dark yellowish orange, very fine grained, clayey; contains gravel-sized, dusky brown concretion fragments and subangular clasts of moderate reddish brown sandstone	1.0	5.7
Sand, as above, but becoming dark yellowish brown	6.7	3.8

KREBS GROUP:		
McALESTER FORMATION:		
Shale, light olive gray, weathered	10.5	1.5
Shale, olive gray to dark gray with moderate brown staining along fracture surfaces	12.0	9.0
Shale, dark gray, noncalcareous; includes dark reddish brown siderite concretions	21.0	5.5
Shale, dark gray, pyritic, moderately hard; contains abundant sideritic concretions ~2 in. thick, some of which are cut by calcite-filled veinlets; also includes beds of soft, grayish black carbonaceous shale 1–2 in. thick from 26.5 to 28.5 ft	26.5	17.2
HARTSHORNE FORMATION:		
Shale, grayish black, carbonaceous, soft, interbedded with harder dark gray shale	43.7	1.5
Sandstone, medium gray to light olive gray, very fine grained, silty, shaly in part; contains well-preserved, carbonized and pyritic fossil-plant remains, including abundant seed-fern compressions	45.2	5.8
Shale, medium dark gray, silty, interbedded with very fine grained sandstone and siltstone; contains black carbonized plant fossils	51.0	5.2
Shale, medium dark gray; contains black carbonized plant fragments	56.2	4.6
Shale, medium dark gray with light gray bands, highly silty; contains abundant layers of very fine grained sandstone 1/64 to 4 in. thick; includes black carbonized plant fragments, pyrite, and minor streaks of coal	60.8	16.2
Sandstone, medium gray with light gray bands, very fine grained, silty, shaly, wavy-bedded and cross-bedded in part; contains plant fossils, streaks and flecks of black carbonized plant material, as well as minor coal inclusions; includes some pyrite, rare 0.5-in.-thick layers of intraformational conglomerate, burrows filled with medium-grained sandstone and sandy concretionary lenses up to 4 in. thick	77.0	24.4
Sandstone, very light gray to white with medium dark gray bands 1/64 to 4.5 in. thick, very fine to fine-grained, micaceous, cross-bedded; abundant sedimentary structures, such as scour-and-fill, plume, soft-sediment deformation, and cross-lamination; black macerated plant materials common on stratification surfaces; contains discoidal sideritic concretions about 2–3 in. in diameter	101.4	27.8
Sandstone, very light gray, fine-grained, massive, noncalcareous	129.2	1.3
Sandstone, very light gray, with dark gray laminae spaced ~1 in. apart, fine-grained, micaceous, noncalcareous; contains black macerated plant material; fractured; includes some 1- to 3-in.-thick, dark gray, shaly sandstone layers below 152 ft; abundant sedimentary structures, such as scour-and-fill, plume, soft-sediment deformation, and cross-lamination; includes some sandy, sideritic concretions averaging ~1 in. thick; contact with underlying unit sharp	130.5	34.8
Coal, black, bright, banded, moderately friable; contains veinlets and small nodules of pyrite (Lower Hartshorne coal)	165.3	1.8
Underclay, dark gray to grayish black, shaly, highly carbonaceous; plant compressions abundant	167.1	1.2
Total Depth		168.3

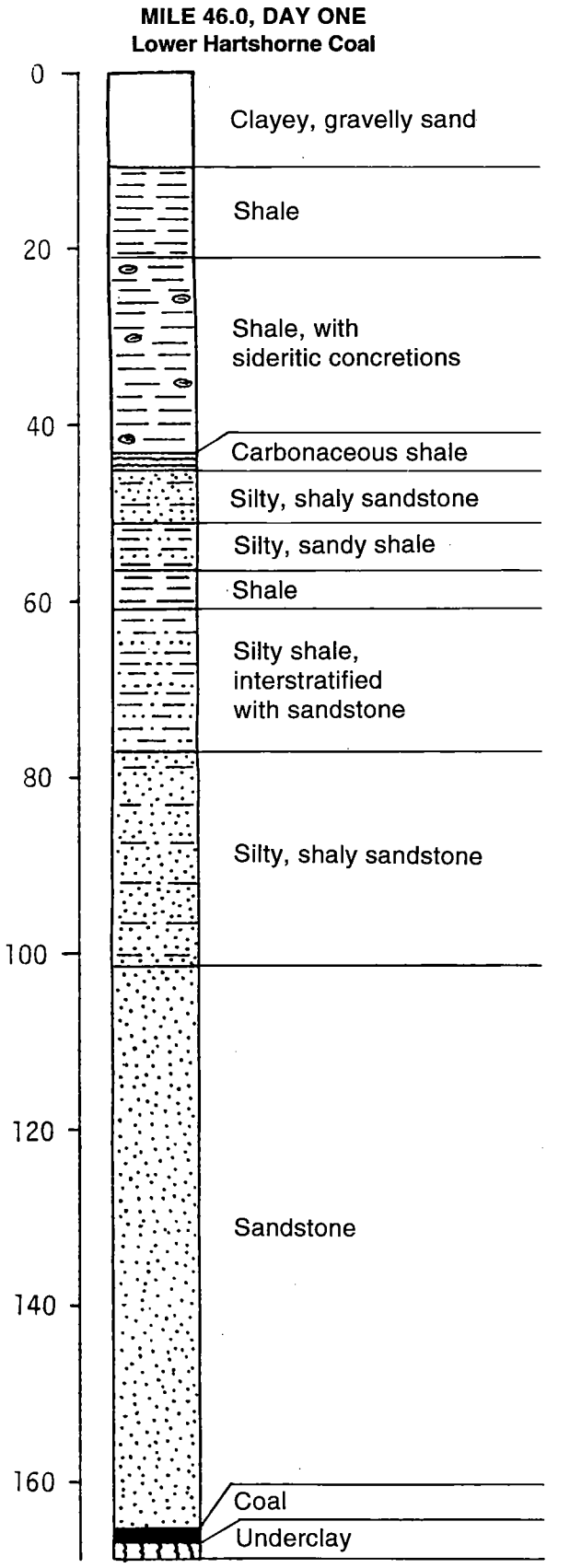


Figure 36. Lithologic column for core hole drilled near Mile 46.0, Day One, in Latimer County by the OGS in 1984 (from Hemish, 1988, p. 16.)

THE BUTTERFIELD OVERLAND MAIL

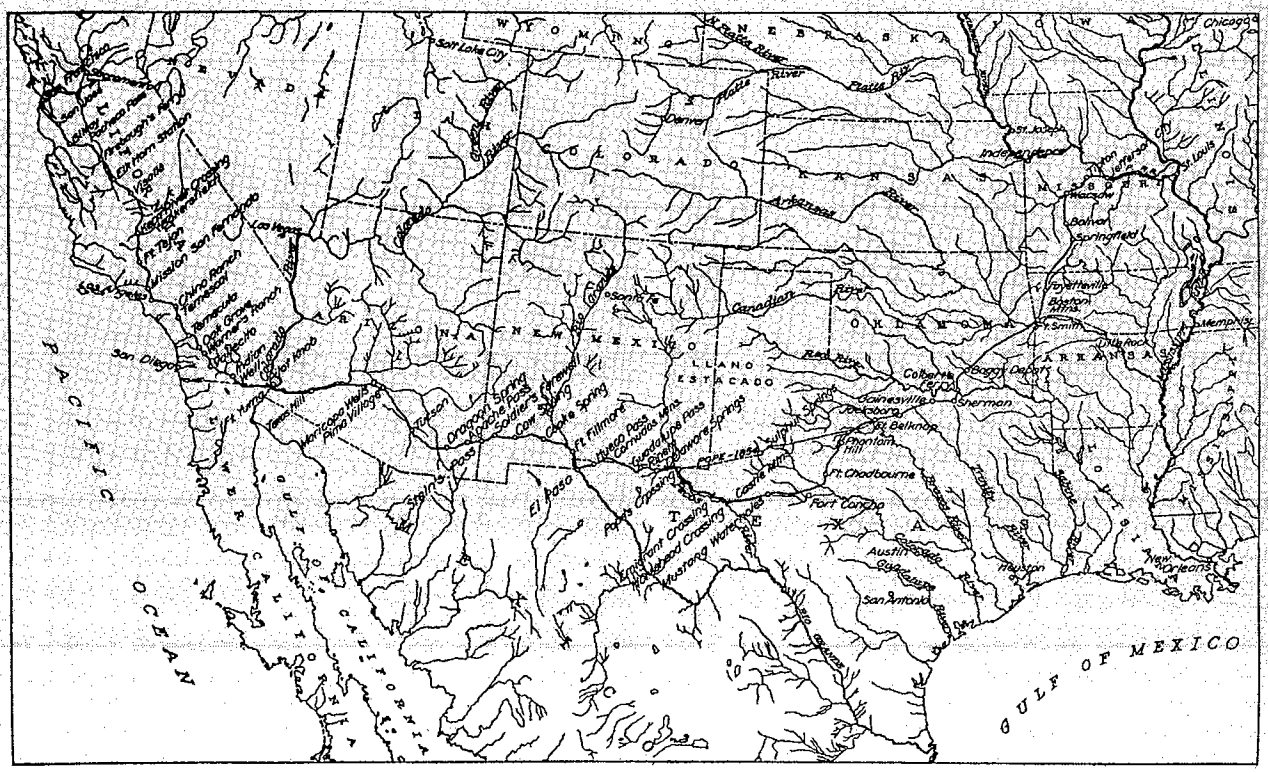
As the population of the western United States increased during the 1840s and 1850s, particularly after the discovery of gold in California, the need for better east-west communications grew. In April 1857, the postmaster-general advertised for bids to operate an overland stage line between the Mississippi River and San Francisco. In September, John C. Butterfield (of Utica, New York) and associates were awarded the contract. The route they chose to follow started in Tipton, Missouri (the railroad terminus at that time), and extended west to Fort Smith, Arkansas; Preston, Texas; El Paso, Texas; Fort Yuma, California; and on to San Francisco—a trip of about 2,800 miles. The terms of the contract required that mail and passengers be carried safely to and from San Francisco twice a week and that the total time for the one-way trip was not to exceed 25 days. For this, Butterfield and his associates would be paid $600,000 per year. An additional stipulation of the contract was that stations were to be built along the route; they would serve as "restaurants" for the passengers on the stageline, and also as stables, blacksmith shops, and local post offices. In Indian Territory (now Oklahoma), these stations were run by citizens of the Choctaw and Chickasaw Nations.

The stage ran 24 hours a day so passengers had to sleep on the move, but "the seat backs would let down for beds at night" (Wooldridge, 1976, p. 9). The one-way fare was $200, not including meals, and passengers were allowed 40 pounds of luggage at no extra cost. A letter from Missouri to California cost 10¢.

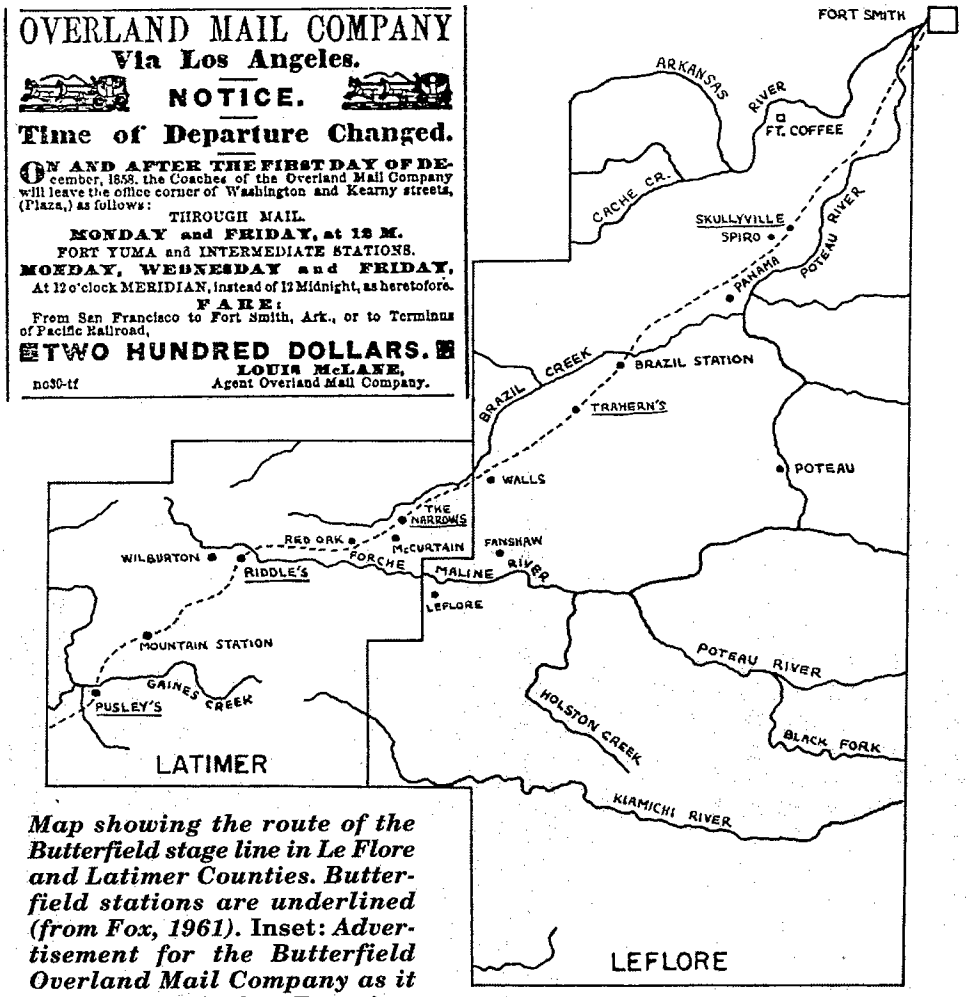
The Butterfield Mail operated in southeastern Oklahoma between 1858 and 1861. The 192-mile trip through Indian Territory started in Fort Smith, Arkansas, and ended at Colbert's Ferry (near present-day Colbert) on the Red River. The stage passed through 12 stations in Oklahoma: Walker's (Skullyville) and Trahern's in Le Flore County; Holloway's (the Narrows), Riddle's, and Pusley's in Latimer County; Blackburn's in Pittsburg County; Waddell's, Geary's, and Boggy Depot in Atoka County; and Nail's Crossing, Fisher's (Carriage Point), and Colbert's Ferry in Bryant County. The Choctaw Nation recognized the importance of permanent highways through their lands, and the Choctaw General Council passed laws requiring all free men (excluding school teachers, farmers, students, and doctors) to work six days per year on public roads. However, the size of the territory (18,220 square miles) and low population (24,000 persons) precluded effective highway maintenance (Wright, 1933, p. 805).

The operation of the Butterfield stages through Indian Territory was of such importance that the Choctaw and Chickasaw Nations each granted eight toll-gate privileges along the road from Fort Smith to the Red River. The tolls (as granted by the Choctaw Council) were as follows:

> For each four wheeled wagon, or other vehicle, drawn by four or more horses, mules, or oxen with driver, the sum of *Fifty cents*; For each four wheeled wagon, or other vehicle, drawn by one or two horses, mules or oxen, the sum of *Twenty-five cents*; For each man and horse, the sum of *Ten cents*; and for each animal in every drove of cattle, horses, mules, hogs, or sheep, *One cent.* (Wright, 1933, p. 806)



Map showing the route of the Butterfield stage line from St. Louis to San Francisco, September 1858 to March 1861 (from Lang, 1945).



Map showing the route of the Butterfield stage line in Le Flore and Latimer Counties. Butterfield stations are underlined (from Fox, 1961). **Inset:** ***Advertisement for the Butterfield Overland Mail Company as it appeared in the San Francisco Bulletin on January 11, 1859.***

The tolls were collected by citizens of the Choctaw and Chickasaw Nations who guaranteed to build bridges over the larger streams and to maintain roads over the more difficult parts of the countryside.

Because the Butterfield Mail route traverses a part of our field-trip area, a description of its course is given here:

> About one mile east of present-day Red Oak, highway 270 and the tracks of the Rock Island Railroad cross the abandoned mail road where it curved west after traversing the old McCurtain place. From this point the road continued on a course almost due west, south of the present main lines of traffic, to [Capt. John] Riddle's keeping along the edge of the southern foothills, that border the narrow valley now traversed by the railroad and highway. Both Little and Big Fourche Maline creeks were crossed....An iron bridge now spans the stream [Big Fourche Maline] near the old crossing which may have been about seventy-five feet north of the bridge.
>
> The west course of the Overland Mail, south of Red Oak, was no accident. It was determined by the geology of the country. From a point three miles south of Red Oak to a point south of Wilburton, there is a line of east-west hills, rising to a height of over eight-hundred and fifty feet, that would have been real "horse-killers" to cross. The Overland Mail in choosing this route took advantage of the fine engineering work of nature and crossed these hills at a point where Fourche Maline Creek had laboriously cut through the hills. (Wooldridge, 1976, p. 7)
>
> [Riddle, whose mother was Choctaw,]...was a prominent leader...in the Indian Territory.... [Riddle Station] was...a little over two miles east of Wilburton...near the old [Lutie] cemetery just to the south of the present highway, on the side of the hill a few hundred yards west of the Fourche Maline. (Wright, 1933, p. 807–808)
>
> The station was a two room log house with a breezeway and stone chimneys....Some of the land was bottom, some prairie, and some hillside....Wood, water, and shade were abundant here in the station area and seams of good coal outcropped in the hillside to the south of the station. This coal was mined and used for fuel and forge. There was considerable blacksmithing to do in the Butterfield business as well as for the Choctaw farmers and ranchers, and the coal was a boon to the blacksmith forges. It is known that much coal was mined and shipped by wagon to other blacksmith shops east and west of the Riddle station along the Butterfield Road. The company itself operated blacksmith shops at intervals along the route.
>
> The stable at the Riddle Station was south of the house on a grassy plat and Riddle's two sons took care of the horses, as well as harnessing them and bringing them out to hitch to the stage as the coach approached the station.... The Riddle station was completely demolished at the death of George Riddle [John Riddle's son] by people digging for gold, because of a story that he had buried gold on his premises. (Wooldridge, 1976, p. 7–8)

Conkling and Conkling (1947, p. 253) noted that, "There was an intermediate station between the Narrows and Riddle's on later established mail routes between the years 1864–1878. This was known as Austin's, four miles east of Riddle's, and it is believed to have been located on the mail road near the crossing on Little Fourche Maline creek, or about a mile east of Panola."

After the Civil War, the Butterfield Overland Mail changed its name to Wells-Fargo Express Company and new routes were established. The company became known as the American Railway Express Company when railroads replaced stagecoaches. Later, it was split into the American Express Company (credit cards, traveler's checks) and the Railway Express Company (Fox, 1961, p. 41).

Analytical data from the OGS chemistry laboratory show that on an as-received basis the proximal values are: moisture content, 4.6%; ash, 8.8%; volatile matter, 34.5%; fixed carbon, 52.1%. Sulfur content is 2.0%, and the heat value is 12,862 Btu/lb.

46.2	0.2	Hairpin turn in road—ascend dip slope of Hartshorne Formation.
46.5	0.3	Stop 8.

STOP 8

Contact between Atoka Formation and Hartshorne Formation

Location: NE¼NW¼SE¼NW¼ sec. 12, T. 5 N., R. 20 E., crest of Red Oak Ridge along Craven Road, Latimer County

The contact between the Atoka Formation (below) and the Hartshorne Formation (above) is well exposed at this stop. The contact appears to be gradational (Fig. 37), and its placement probably will elicit some discussion.

The criteria for placing the contact at the base of Unit 4 (Measured Section, Stop 8; Fig. 38) are: (1) the change from a heterolithic unit of siltstone, sandstone, and shale to a comparatively thick package of shaly sandstone; and (2) the presence of local, small channels cutting into the heterolithic unit at the base of the sandstone unit. In some places elsewhere in the field-trip area, the channeling is much more pronounced and a pebble conglomerate is present at the base of the Hartshorne Formation (Hemish, 1993a, stop 6).

A short hike through the woods to the east along Red Oak Ridge reveals a pronounced change in character of the basal part of the Hartshorne Formation. Here the sandstone is ~50 ft thick, coarser grained, massive, blocky, and contains soft-sediment deformation features as well as cross-bedding in places (Fig. 39). At the base of the cliff, which is inaccessible from above, several sedimentological features suggest a (distributary) channel environment: (1) thick, irregularly bedded sandstones (upper sequence) fill a channel eroded to a depth of 2 ft into a lower sequence of parallel-stratified sandstones (Fig. 40); (2) the lower sequence shows large-scale (50-ft-



Figure 37. Outcrop of Atoka and Hartshorne Formations in abandoned shale pit showing interbedded siltstone, sandstone, and shale overlain by parallel-bedded shaly sandstone.

Measured Section, Stop 8

Upper Part of Atoka Formation and Lower Part of Hartshorne Formation

NE¼NW¼SE¼NW¼ sec. 12, T. 5 N., R. 20 E., Latimer County. Measured in shale pit west side of road on south slope of Red Oak Ridge.

	Thickness (feet)
KREBS GROUP:	
HARTSHORNE FORMATION:	
4. Sandstone, moderate brown (5YR4/4) to light brown (5YR5/6) to grayish orange (10YR7/4), very fine grained, noncalcareous, thin- to medium-bedded, mostly parallel bedded, blocky; contains abundant small-scale, low-angle cross-bedding; surface ripple-marked; in places contains sparse ironstone pebbles; lower bed thins laterally, in places fills small channels cut into underlying unit; shale partings common, with sandstone beds generally ~1 ft thick; basal contact conformable, but locally disconformable	18.0
ATOKA FORMATION:	
3. Siltstone, sandstone, and shale, interstratified, moderate yellowish brown (10YR5/4) to grayish orange (10YR7/4), micaceous, noncalcareous; contains abundant black, comminuted plant material on stratification surfaces; mostly lenticular bedded; very thin bedded to laminated, parallel-bedded; sandstone units cross-bedded and ripple-marked; sandstone layers more abundant in upper half of unit; entire unit is heterolithic; base gradational	15.5
2. Siltstone, brownish gray (5YR4/1) to grayish red (10R4/2), interlaminated with light gray (N7), very fine grained sandstone and dark gray (N3) shale; very thin bedded, wavy-, parallel-bedded; micaceous; black comminuted plant material common; horizontal trace fossils abundant on bedding planes; low-angle cross-laminations common; includes scattered grayish orange (10YR7/4), very fine grained sandstone lenses about 6–12 in. thick and 6–10 ft wide that fill channels cut into underlying heterolithic units; basal contact gradational	17.6
1. Shale, olive gray (5Y4/1) to dark gray (N3) with moderate brown (5YR3/4) to dark yellowish orange (10YR6/6) iron-stained layers; includes minor siltstone and sandstone laminae as well as clay-ironstone stringers; base covered	25.0
Total thickness of section	76.1

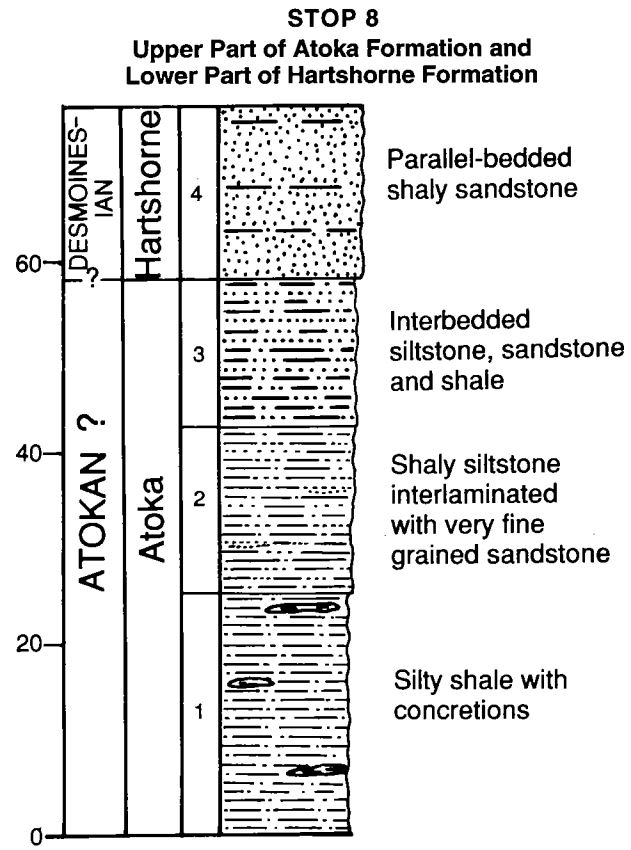


Figure 38. Graphic columnar section of upper part of Atoka Formation and lower part of Hartshorne Formation at Stop 8.

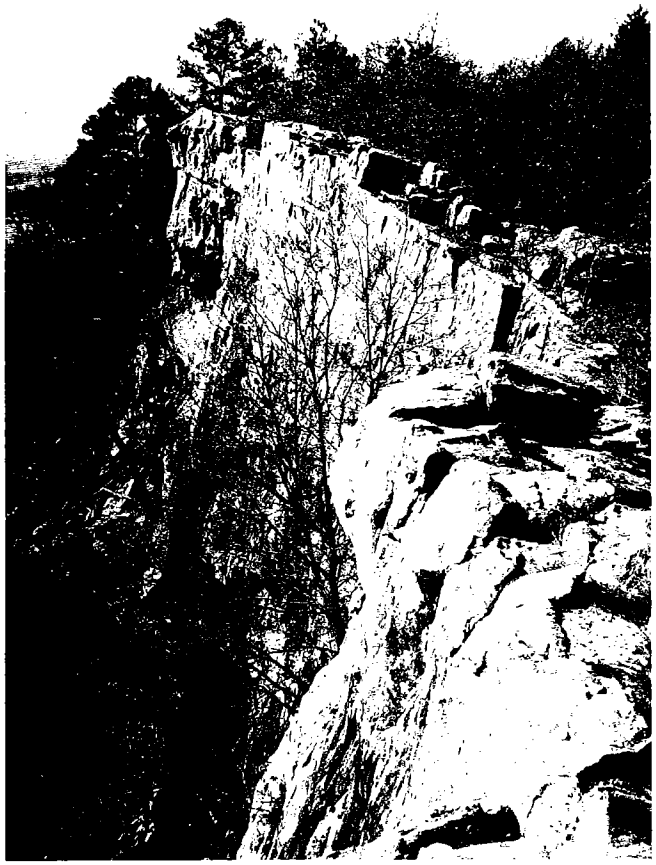


Figure 39. Massive, blocky, 50-ft-thick section of the lower Hartshorne Sandstone exposed on Red Oak Ridge ~0.25 mi east of shale pit.

long and 5-ft-thick) lenticular bedding; (3) shale (Atoka Formation) below the lower sequence is eroded locally to 2 ft deep and shale rip-up clasts are abundant locally near the base of the overlying sandstone (Fig. 41); (4) sandstone beds near the base of the lower sequence pinch and swell and locally show large-scale cross-stratification.

Our interpretation of this outcrop is that it is a large distributary channel in a deltaic environment and that the shale-pit outcrop is marginal to the channel. Stop 8 typifies exposures of the Hartshorne Formation laterally along its outcrop belt in the field-trip area.

The Hartshorne Sandstone (Formation) was renamed by Taff (1899) for exposures near the town of Hartshorne, Oklahoma. Chance (1890) had earlier described and mapped the same unit as the "Tobucksy" sandstone in the Heavener, Oklahoma, area. Taff and Adams (1900) subsequently included in the formation the Lower Hartshorne coal and some overlying shale and sandstone. Oakes and Knechtel (1948) redefined the Hartshorne Formation and placed the upper boundary at the top of the Upper Hartshorne coal. This removed the Upper Hartshorne coal from the overlying McAlester Formation.

Branson (1956) subsequently suggested that the term "Tobucksy" be resurrected for the lowest sandstone member, while supporting the definition of Oakes and Knechtel (1948) for the boundaries of the Hartshorne Formation. McDaniel (1961) agreed with the defined boundaries and proposed that the formation be divided into an upper and lower member. The lower member extends from the base of the lowest sandstone



Figure 40. Thick-bedded sandstone filling channel eroded into underlying parallel-bedded sandstone in exposure of the lower Hartshorne Sandstone, same location as Figure 39.



Figure 41. Sharp, erosional contact between lower Hartshorne Sandstone and underlying Atoka Formation, about the same location as Figure 39.

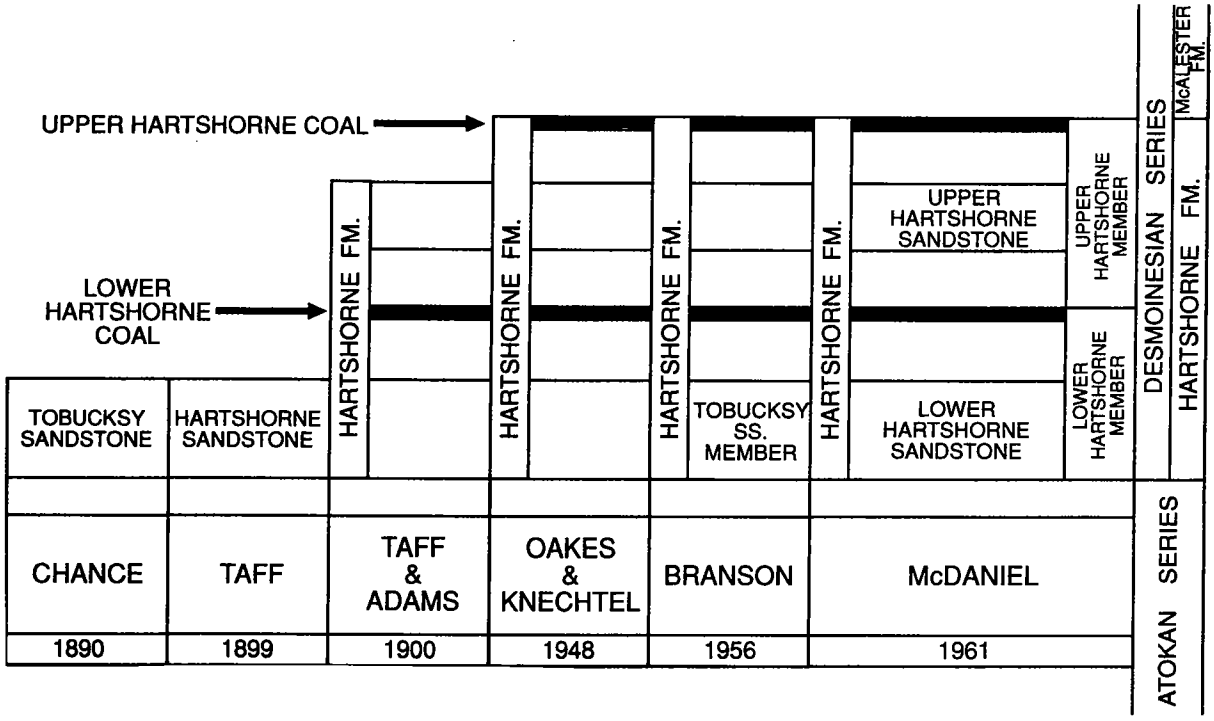


Figure 42. Definitions of the Hartshorne Formation, 1890–1961 (modified from McDaniel, 1961).

unit to the top of the Lower Hartshorne coal, and the upper member includes the strata from the top of the Lower Hartshorne coal to the top of the Upper Hartshorne coal. In this scheme the sandstone in the lower member is named the Lower Hartshorne sandstone, and the sandstone in the upper member is named the Upper Hartshorne sandstone. Figure 42 shows concepts of the Hartshorne Formation through time.

In the field-trip area, the Hartshorne Formation is generally about 250–300 ft thick. Stop 18 will provide further opportunity for observing the contact between the Atoka Formation and the Hartshorne Formation. The contact is well exposed at that stop in the road cut on U.S. Highway 59, ~3 mi south of Heavener.

52.5 6.0 Turn around and retrace route (reentering Red Oak 7.5′ quadrangle) to the intersection of U.S. 270 and S.H. 82 on east side of the town of Red Oak. Turn right (south) on S.H. 82.

52.7 0.2 Cross poorly exposed Tamaha Sandstone Member of the McAlester Formation. Here, the Tamaha dips ~10° north and is ~1,800 ft above the base of the McAlester Formation. Southward, toward the Choctaw fault, the northward dip of the McAlester and underlying Hartshorne and Atoka Formations gradually, but steadily, increases. Rocks directly south of the Choctaw fault dip to the south.

52.9 0.2 Cross poorly exposed Upper McAlester coal beds. The lower bed was surface mined ~0.4 mi to the east at some unknown time in the past, but after 1931 when Hendricks (1939) mapped the area.

53.7 0.8 Cross ridge of Warner Sandstone Member of McAlester Formation. In this area, the Warner Sandstone dips 20–25° north and consists of an upper bed ~60 ft thick and a lower bed ~150 ft thick, separated by ~150 ft of shale.

53.9 0.2 The flat valley on both sides of the road trends east-west and is underlain by the McCurtain Shale Member of the McAlester Formation. Just west of the road, Mitchell Energy spudded the No. 1 Russel Albin (K.B. 626 ft) in the C SW¼ sec. 3, T. 5 N., R. 21 E., on March 10, 1984. Log tops (drilled depths) reported to the Oklahoma Corporation Commission are: Atoka upper—2,810 ft; Panola—7,333 ft; Diamond—7,988 ft; Brazil—8,612 ft; Four Fingers—8,916 ft; Brazil second—9,085 ft; Four Fingers second—9,167 ft; Cecil—10,912 ft; Cecil second—11,421 ft; Spiro upper—12,940 ft; Spiro channel—13,022 ft; Wapanucka—13,069 ft; Cromwell—13,708 ft. Two blind thrusts cause repetition of the strata and a lower bedding-plane thrust may be present in the shale ~600 ft above the Spiro (Hemish and others, 1990b).

54.1 0.2 Road turns right (southwest) and begins to climb ridge of Hartshorne Formation. Covered Upper Hartshorne coal at top of Hartshorne Formation at bend in road.

54.5 0.4 Top of Red Oak Ridge. The Lower Hartshorne coal is present, but covered, immediately north of the top of the ridge. In this area, the Hartshorne Formation is ~300 ft thick and dips 30° north. The base of the Hartshorne Formation is between the bend in the road at the top of the ridge and the very steep hairpin curve just south of the ridge crest. Compare Measured Section, Mile 54.5, Day One (measured here from the summit of Red Oak Ridge southward) and its graphic columnar section (Fig. 43) with Measured Section, Stop 8, and Figure 38.

Measured Section, Mile 54.5, Day One

Upper Part of Atoka Formation and Lower Part of Hartshorne Formation

SE¼NW¼NW¼ sec. 10, T. 5 N., R. 21 E., Latimer County, Oklahoma. Measured along winding road (S.H. 82) from summit of Red Oak Ridge southward. Estimated elevation at top of ridge 780 ft.

	Thickness (feet)
KREBS GROUP:	
HARTSHORNE FORMATION:	
9. Sandstone, moderate reddish orange (10R6/6) to grayish orange (10YR7/4), very fine grained, noncalcareous, well-indurated, burrowed, thin- to medium-bedded; shaly, with low-angle cross-lamination in lower 8 in.	4.0
8. Shale, medium light gray (N6) with moderate red (5R4/6) mottling, silty, noncalcareous, weathers light brown	4.2
7. Sandstone, light gray (N7), weathers moderate reddish orange (10R6/6) to dark reddish brown (10R 3/4), very fine grained, noncalcareous; irregularly thin to medium bedded, cross-bedded and rippled in part; includes some thin shaly layers; trace fossils abundant on soles of some beds; contains scattered pebbles of clay-ironstone and clay galls in places; fossil-plant material common; contact with underlying unit sharp	11.5
6. Shale and sandstone, interbedded, very pale orange (10YR8/2) to grayish orange (10YR7/4), wavy-bedded and thin-bedded, noncalcareous, micaceous; contains abundant trace fossils; locally includes lenses of very fine grained, cross-laminated sandstone up to 10 in. thick	13.0
5. Sandstone (same description as Unit 7)	9.3
4. Shale and sandstone, interbedded, poorly exposed (same description as Unit 6)	30.0
3. Sandstone, light olive gray (5Y5/2), very fine grained, noncalcareous, very hard, contains low-angle, small-scale cross-stratification; contact with underlying unit sharp	0.7
ATOKA FORMATION:	
2. Shale, medium gray (N5) to pale yellowish brown (10YR6/2), silty and sandy, noncalcareous, bioturbated, includes abundant dusky brown (5YR2/2) staining on stratification surfaces; rippled in places; finer grained downward; grades into underlying unit	8.0
1. Shale, medium dark gray (N4) with dark reddish brown (10R3/4) staining, silty, micaceous in places, noncalcareous; includes some sandstone-filled burrows and black (N1) plant fragments ..	19.3
Total thickness of section	100.0

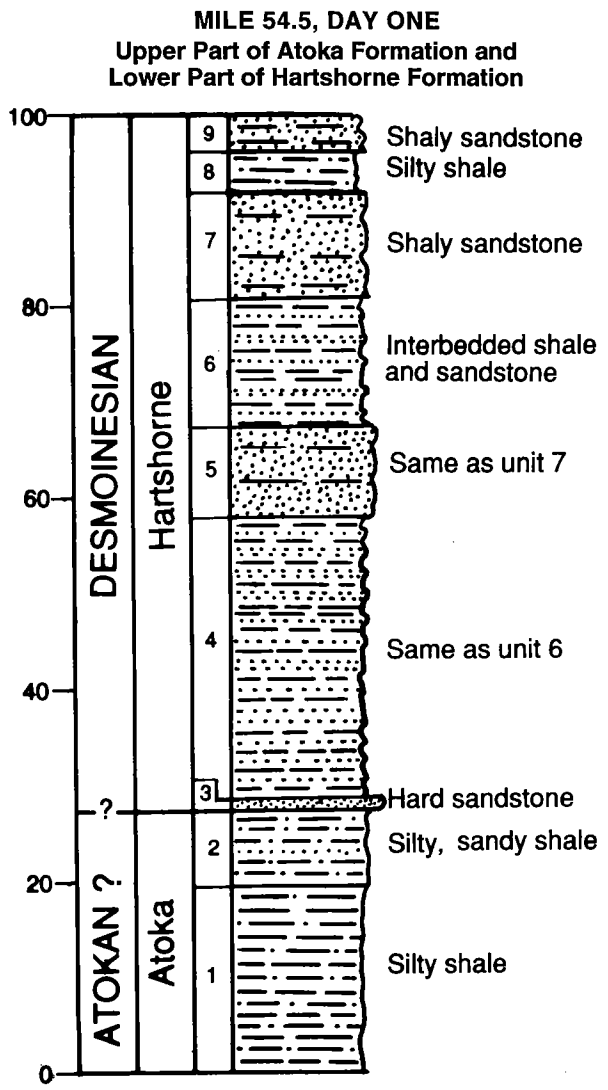


Figure 43. Graphic columnar section of the upper part of the Atoka Formation and lower part of the Hartshorne Formation at Mile 54.5, Day One.

Continue south on S.H. 82.

54.7 0.2 Excellent outcrop of uppermost Atoka Formation on right (north) side of road. Bedding characteristics (Measured Section, Mile 54.5, Day One) suggest that this part of the Atoka Formation was deposited in relatively shallow water.

55.0 0.3 Road to left (east). Continue straight (south). This flat area, known as Long Prairie, is underlain by shale in the upper part of the Atoka Formation.

55.5 0.5 Low ridge (Little Ridge on Red Oak 7.5′ quadrangle) is underlain by sandstone that is 3,200 ft below the top of the Atoka Formation, about the same interval occupied by the Fanshawe sandstone (Fig. 35) in wells to the north. Here, the Fanshawe(?) is ~370 ft thick and dips 45–50° north. The sandstone is yellowish brown, silty and shaly, thin to medium bedded, and extensively ripple marked. To the east and west, the sandstone on the ridge is offset by several northwest-striking strike-slip faults; most show left slip, but some show the opposite sense of movement.

56.0 0.5 Cross Fourche Maline.

56.1 0.1 Cross buried (by alluvium) trace of Choctaw fault, the leading edge of the Ouachita fold and thrust belt at the present level of exposure. Here, the fault juxtaposes shale in the upper part of the Atoka Formation (in the footwall) against shale in a lower part of the Atoka Formation (in the hanging wall).

About 0.6 mi to the east, in C NW¼SE¼ sec. 15, T. 5 N., R. 21 E., Anadarko Petroleum drilled the No. 1-15 Alford "A" (D.B. 568 ft). The well was designated a 17,000-ft Arbuckle test. It spudded on August 7, 1989, and reached 17,470 ft total depth on February 20, 1990. Log tops (drilled depths) reported to the Oklahoma Corporation Commission are: Panola—4,240 ft; Brazil—5,915 ft; Cecil—8,862 ft; Shay—10,600 ft; Wapanucka—16,439 ft; Cromwell—17,022 ft. The well was completed for 335 Mcf gas per day, after hydraulic fracturing, from the Shay (Fig. 35) at 10,601–10,636 ft.

56.7 0.6 Turn left (east) on S.H. 82 (paved road). Ridge immediately to south is Limestone Ridge, which is capped by the Spiro sandstone, considered by some to be the basal sandstone of the Atoka Formation (Groves and Grayson, 1984; Sutherland and Manger, 1984).

57.2 0.5 Powerlines cross road. Atoka Formation mudstones and siltstones are exposed under the powerlines immediately south of the road. The beds dip south, and poorly developed cross-laminations suggest that the strata are overturned. A major(?) north-directed thrust fault about half-way up the north side of the ridge juxtaposes the Atoka Formation (exposed here in the footwall) against Spiro-Wapanucka (in the hanging wall) (Fig. 44). The Spiro-Wapanucka on Limestone Ridge forms the north limb of a north-vergent overturned syncline. (The axis and part of the south limb are described at Stop 10.) About 0.5 mi to the east, the thrust fault ramps up-section, cutting off the syncline until, in the southeast corner of the Red Oak 7.5′ quadrangle, the Atoka Formation is present in the footwall and the hanging wall.

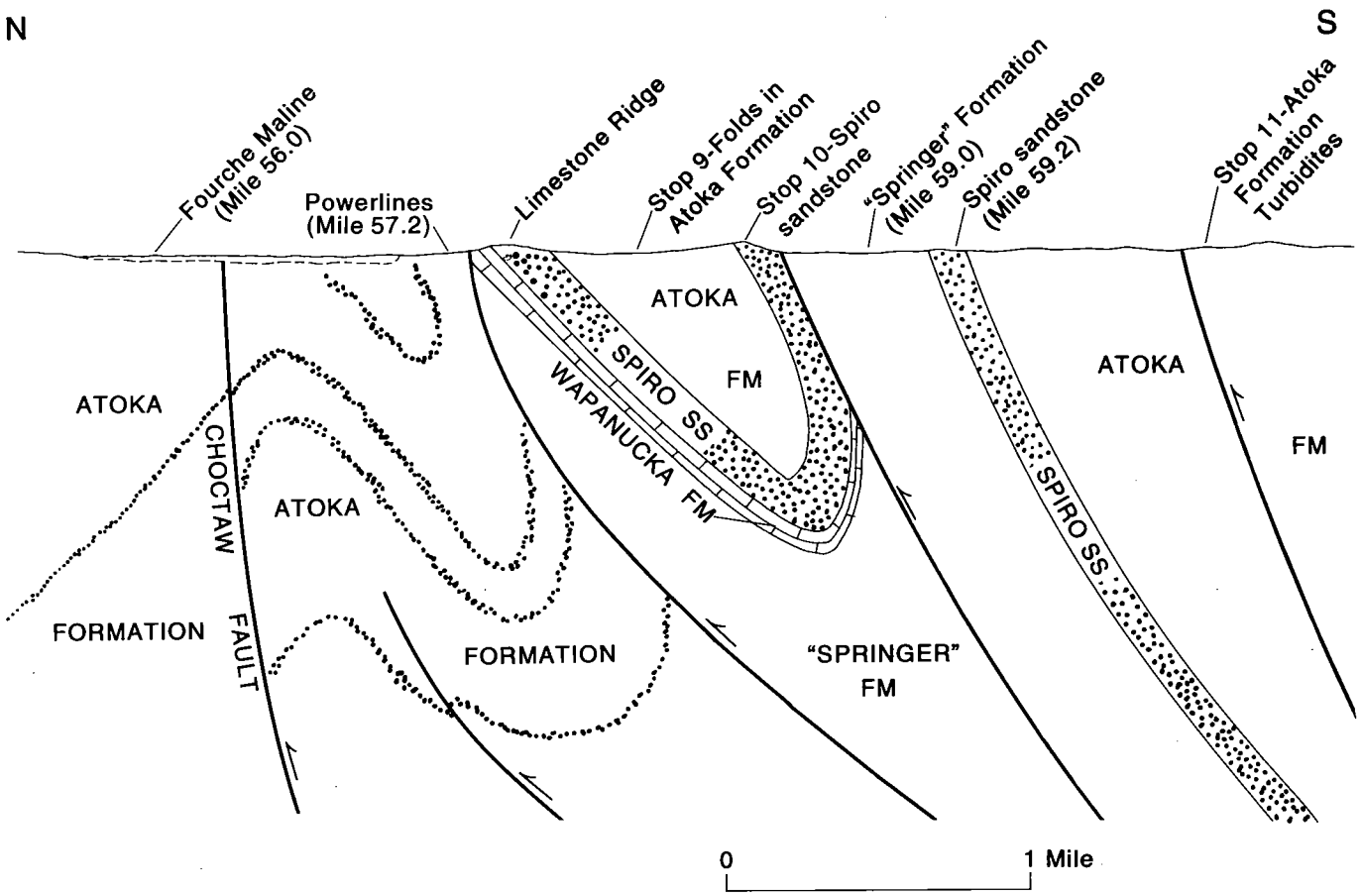


Figure 44. North-south cross section across Limestone Ridge showing structural position of the Atoka Formation at Stop 9, the Spiro sandstone at Stop 10, and outcrops of the "Springer" Formation at mile 59.0 (modified from Hemish and others, 1990b).

The Wapanucka Formation, as exposed high on the slope, consists of hard gray limestone, soft reddish and gray fossiliferous sandstone, and shale (mostly covered). The Spiro sandstone at the top of the ridge is a medium-grained, stratified, locally vuggy sandstone that consists of 5–20-ft-thick amalgamated sequences.

Continue east on S.H. 82.

57.7	0.5	County road to Salonia and Leflore continues straight ahead (east). S.H. 82 to Talihina bends right (south). Follow S.H. 82.
58.0	0.3	Cross thrust fault described at mile 57.2. Evidence for the fault is that the Spiro sandstone, which caps Limestone Ridge and extends as an almost continuous outcrop for >16 mi to the west, does not cross the road here.
58.5	0.5	Stop 9.

STOP 9

Folds in Atoka Formation

Location: NW¼SW¼ sec. 23, T. 5 N., R. 21 E., Latimer County, large road cut on east side of S.H. 82

This is the first well-exposed outcrop of turbidite sandstones and shales in the Atoka Formation south of the Choctaw fault in this area (Fig. 45). These strata are mapped as lower Atoka shale by Hemish and others (1990b) and are preserved in the core of a north-vergent overturned syncline (Fig. 44). These strata are ~550 ft above the top of the Spiro sandstone (see Stop 10), which is exposed in a road cut ~0.1 mi south of this stop.

This outcrop has been described in several publications and undoubtedly has been visited by many geologists and geology students on field trips. Chamberlain (1978, p. 45) described the trace-fossil assemblage at this exposure and identified the following: *Asterichnus, Chondrites, Paleodictyon,* large *Paleodictyon* (cf. *Megagrapton*), *Scalarituba, Paleodictyon* (= ? *Squamodictyon*), and *Sustergichnus.* He interpreted these as belonging to the *Chondrites* subfacies and suggested that the sediments were deposited in an outer-shelf to slope environment. Sutherland (1979, p. 63–64) noted that the Atoka sandstones at this exposure represent mostly incomplete



Figure 45. Folded turbidite sandstones and shales in the lower part of the Atoka Formation at Stop 9. The anticline-syncline pair just to the left of the center of the photograph is fold V/VI/VII on Figure 47.

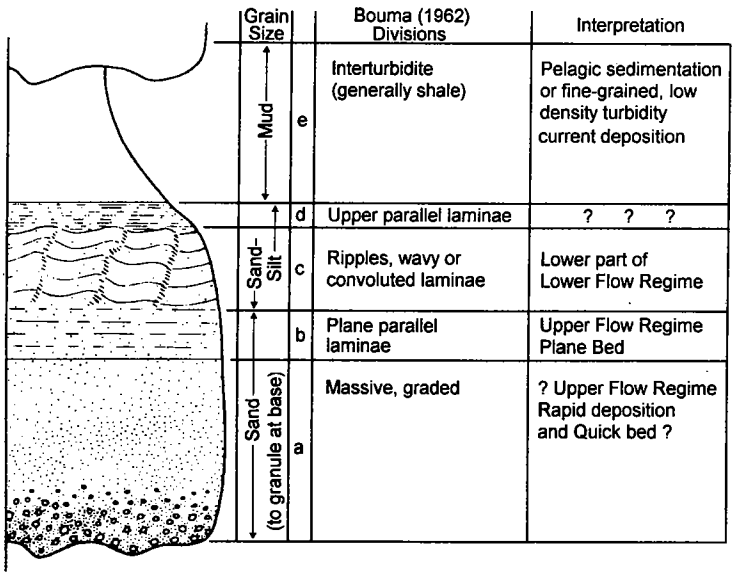


Figure 46. Idealized Bouma sequence Ta–e showing grain size, sedimentary structures, and flow regime of deposit from a turbidity flow (modified from Middleton and Hampton, 1976).

Bouma sequences (Fig. 46) lacking transport features on the base and he interpreted the sequence as distal turbidites. More recently, Poole (1985, p. 103) interpreted this exposure in terms of melange formation and recognized contrasting styles of deformation (ductile vs. brittle).

Figure 47 is a detailed sketch (looking east) of the strata exposed in the road cut on the east side of the highway. The dominant structural feature consists of south-dipping, north-facing strata on the south limb of a north-vergent, overturned syncline. The strike of the overturned strata varies from N. 55° E. to N. 85° E., and the dip varies from 68° SE. to 84° SE.; the average attitude of the overturned strata is N. 68° E. 77° SE. The axis of the syncline is near the north end of the exposure and strata immediately north of the axis strike N. 45° E., dip 26° SE., and face southeast.

Parts or all of 10 out-of-the-syncline parasitic folds and fold pairs, labeled with Roman numerals on Figure 47, deform the strata on the overturned south limb. Most of these small folds plunge gently (5–15°) to the east-northeast or west-southwest, approximately parallel to the synclinal axis. The parasitic fold nearest the core of the syncline (labeled I,II) differs from the others—it plunges moderately to the east. The average axial-plane attitude of these small folds is about N. 60° E. 40° SE. These minor folds probably are flexure-slip folds formed at the same time as the major fold.

Many of the parasitic folds and several of the overturned beds on the south limb of the syncline are offset by 27 mostly steeply north-dipping, high-angle reverse faults (Fig. 47, capital letters). Most of these faults strike about N. 75° E. and dip about 65° NW.; however, some faults strike about N. 85° W. Where observed, most slickensides show that the latest movement was dip slip; some, however, show oblique-slip movement on the fault. The offset of the fold labeled IV by one of these reverse faults (Fig. 47) suggests that the faults formed after the folds. The faults probably represent out-of-the-syncline "accommodation" faults formed in response to the volume problem in the core of the syncline. Their origin, therefore, is the same as that of the small folds, but the fact that most of them juxtapose younger against older strata suggests that this fold limb was steeply dipping or overturned when the faults formed.

Other features that are related less clearly to the major folding and overturning include: (1) south-side-up thrust faults; (2) south-side-up high-angle faults; (3) south-side-down high-angle "normal" faults; and (4) subhorizontal top-to-south faults. Some of (1) and (2) are small, "accommodation" faults associated with the small folds. The large type (1) faults (e.g., northern "a" on Fig. 47) offset the small folds but themselves are offset by the north-dipping reverse faults (e.g., "C" on Fig. 47). The south-side-up thrust faults originally may have been north-directed high-angle reverse faults formed in the early stages of folding (but after formation of the small folds) that subsequently were rotated to low angles as folding progressed. Similarly, the type (3) "normal" faults may be early-formed, out-of-the-syncline "accommodation" reverse faults that were rotated slightly to their present position. The origin of the type (2) faults is less clear—they may be very late-stage north-directed reverse faults.

Data Points–
Out-of-the-Syncline Reverse (Accommodation) Faults

A. N. 75° E. 75° N. Slickensides rake 69° W.
B. N. 70° E., dips steeply south. 2 ft separation.
C. North-directed thrust offset by reverse fault.
D. N. 75° E. 52° N.
E. N. 75° E. 75° N.
F. No match across large reverse fault.
G. N. 85° W. 62° N. Slickensides indicate dip slip.
H. N. 50° E. 51° NW. Small horse block. Slickensides indicate dip slip.
I. N. 75° W. 52° N. Slickensides rake 60° W. and indicate oblique slip.
J. N. 85° W. 79° N.
K. N. 75° E. 62° N. Slickensides indicate dip slip.
L. N. 65° E. 59° NW. Slickensides indicate oblique slip.
M. N. 80° E. 61° N. ~4 ft separation.
N. Shattered horse block.
O. Fault dips north, 7 in. separation.
P. N. 80° E. 55° N., 3 in. separation.
Q. 1 in. separation.
R. N. 90° E. 70° N., ~7 in. separation.

Data Points–
Out-of-the Syncline Parasitic Folds

I. Axial surface N. 55° E. 31° SE.
II. Axis plunges 47° to E.
III. Axis plunges 10° to S. 65° W.
IV. Axis plunges 10° to S. 70° W. Hinge thickened.
V. Axial surface N. 65° E. 46° SE.
VI. Axis plunges 03° to N. 70° E.
VII. Axis plunges 13° to N. 85° E.
VIII. Axis plunges 10° to S. 65° W.
IX. Axis plunges 14° to S. 65° W.
X. Axis plunges 09° to S. 65° W.; axial surface dips ~60° S.
XI. Axis plunges 06° to S. 70° W.; axial surface dips steeply south.
XII. Axis plunges 08° to S. 55° W.; axial surface dips steeply south.
XIII. Axis plunges 16° to N. 60° E.
XIV. Axis plunges 10° to S. 55° W.; axial surface dips ~45° S.
XV. Axis plunges 13° to N. 70° E.; axial surface dips steeply south; south limb thinned.

Other Faults (See Text)

a. North-directed low-angle thrusts (2)
b. North-directed bedding-plane thrust.
c. Subhorizontal south-directed thrust.

Soft-Sediment Deformation and Extensional (?) Features

1. Bed thinned.
2. "Tail."
3. Sandstone phacoid.
4. Sandstone phacoid.
5. Phacoid trail.

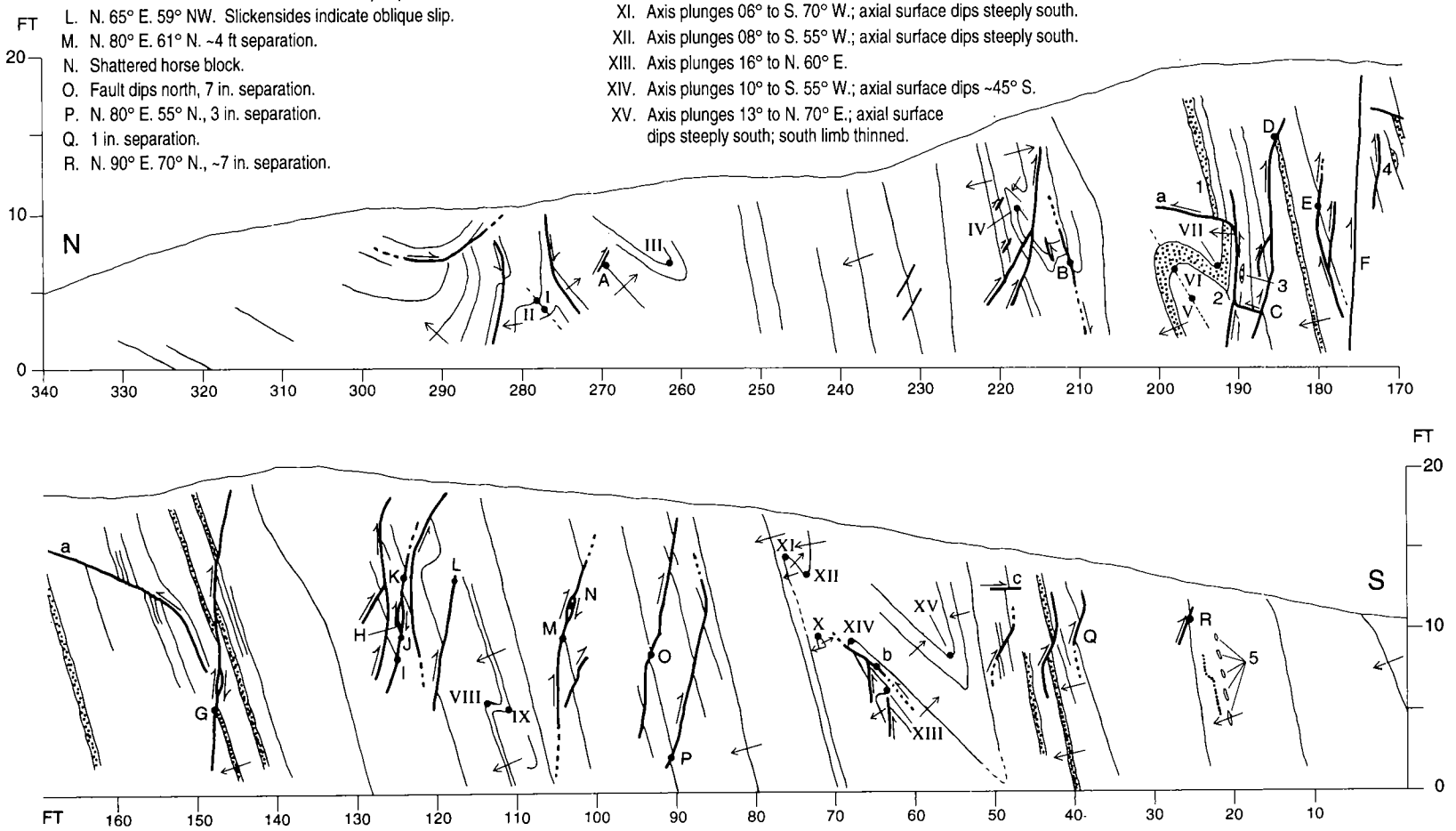


Figure 47. Detailed sketch of Atoka Formation at Stop 9. Attitudes of major structural features are labeled: reverse faults (capital letters), parasitic folds (Roman numerals). Minor structural features are labeled: other faults (lower-case letters), soft-sediment deformation and extensional features (Arabic numerals). Arrows show facing direction.

The single type (4) south-directed thrust fault (labeled "c" on Fig. 47) may be a back-thrust.

There is a high degree of structural complexity in this exposure; nevertheless, most of the strata were lithified during folding. Soft-sediment-deformation features related to dewatering of sand beds are common, but they formed prior to folding. There is evidence (labeled 1, 2, 3... on Fig. 47) that some sandstone beds behaved less coherently than others during folding; this evidence includes phacoid trails (labeled 5), isolated phacoids (3, 4), thinned beds (1), and the conspicuous "tail" (2) on small fold V/VI/VII. Except for the "tail," these features are extensional in origin and probably formed relatively late in the folding.

In summary, evidence for the following sequence of events can be seen at this exposure:

1) Deposition of turbidite sands and mud, dewatering of some sands;

2) Burial and partial to complete lithification;

3) Broad folding about east-northeast axis and development of parasitic small folds;

4) Continued folding and development of south-dipping, high-angle reverse faults;

5) Continued folding; rotation of reverse faults; development of early out-of-the-syncline, north-dipping reverse faults;

6) Continued folding; rotation of all previously formed features; and development of late out-of-the-syncline, north-dipping reverse faults and extensional features (phacoids, thinned beds).

Continue south on S.H. 82.

58.6 0.1 Stop 10.

STOP 10

Spiro Sandstone

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MASERA Corp.
Tulsa, Oklahoma

Location: C SW¼SW¼ sec. 23, T. 5 N., R. 21 E., Latimer County

INTRODUCTION

The Spiro–Wapanucka interval is a mixed clastic-carbonate sequence that is complicated by the presence of several unconformities. The Wapanucka Formation is upper Morrowan in age and usually is separated from the overlying Spiro by the so-called "sub-Spiro shale" that appears to be uppermost Morrowan in age (Fig. 48). The Spiro is lowermost Atokan and rests unconformably on the "sub-Spiro shale" or, where the latter is absent, on the Wapanucka Formation.

The Spiro can be subdivided into upper and lower units. The lower unit is a valley-fill sandstone that typically is found north of the Ouachita Mountains. The upper unit is a complex of cyclic shallow-marine sandstones and carbonates that are found in both the Arkoma Basin and the Ouachita Mountains in Oklahoma.

OUTCROP DESCRIPTION

This outcrop is located ~6 mi south of Red Oak on the west side of S.H. 82 (Fig. 49A). Strata strike N. 80° E., dip 75° S., and are overturned. Although the Spiro in this outcrop is composed predominantly of sandstone, it does contain a limestone bed near the top. The Spiro has a gradational basal contact with the sub-Spiro shale. The Wapanucka does not occur at this outcrop.

The Spiro sandstone is 228 ft thick at this outcrop and can be divided into six intervals.

94–228 ft

106–228 ft—The upper part of the uppermost interval consists of five, thick, covered intervals separated by clean, tan to red, fine- to medium-grained sandstone.

94–106 ft—The basal part of the uppermost interval is composed of red, medium-grained, friable sandstone that contains abundant fossil molds and cross-bedding.

79–94 ft—The second interval from the top consists of gray, massive, micritic limestone.

57–79 ft—The third interval down from the top is composed of fine- to medium-grained, poorly to well-sorted sandstone. Fossil molds are common from 57 to 66 ft (Fig. 49B). Cross-bedding occurs near the base. Burrows occur from 66 to 79 ft as do echinoderm and other fossil fragments.

35.5–57 ft—The fourth interval from the top contains red to tan, fine-grained, well-sorted sandstone. It is thin to medium bedded with shale laminae, shale wisps, and current bedding. Ripples and burrows are common sedimentary structures. Glauconite occurs in trace amounts throughout.

26–35.5 ft—The second interval from the base is composed of tan to gray, fine- to medium-grained, poorly to moderately well sorted sandstone. Clay and shale intraclasts are abundant in the section from 26 to 32 ft (Fig. 49C). Fossils are also common in this interval, as is glauconite. The dominant sedimentary structures are cross-bedding in beds between 26 and 32 ft (Fig. 49D) and ripples and burrows from 32 to 35.5 ft. This interval has a sharp basal contact.

7–26 ft—The lowermost interval consists of tan to gray, fine-grained, well-sorted sandstone. Traces of glauconite are present throughout. Burrows and ripples are common sedimentary features. The basal part of the sandstone at 10 ft is intensely bioturbated and burrowed (Fig. 49E). It has a gradational contact (7–10 ft) with the sub-Spiro shale (0–7 ft).

PETROGRAPHIC ANALYSIS

Quartz is the primary detrital constituent in the sandstones. Grains, which are mainly monocrystalline, have straight extinction. Quartz grains with undulatory extinction and polycrystalline grains also occur. Feldspar is present as a trace in one sample. Rock fragments include

chert, shale, and siltstone. Muscovite, zircon, and tourmaline are present as accessories.

Fossil fragments are common in some intervals. Many grains are partially dissolved. Echinoderm fragments are the most common type.

Glauconite and chamosite occur in small amounts. Most glauconite grains are deformed and show dissolution. Some grains are altered to pseudomatrix or brown glauconite. A few chamosite peloids are recognizable in some samples. Oolitic chamosite coatings on quartz grains also occur, but either they are partially dissolved or they are replaced almost entirely by calcite.

Clays include illite and chlorite. Illite is predominantly recrystallized matrix and pseudomatrix. Authigenic chlo-

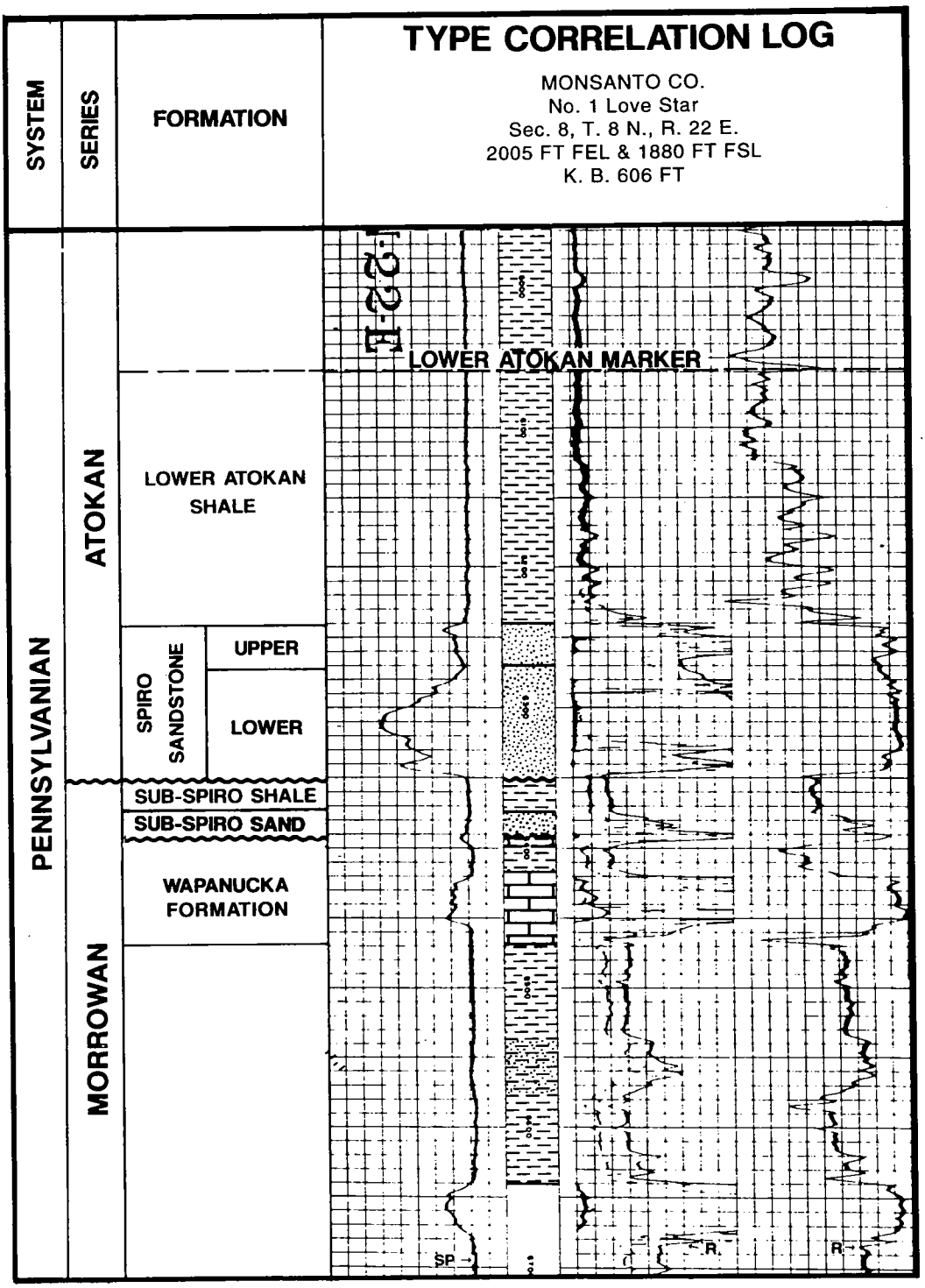


Figure 48. Type correlation log for Spiro sandstone showing age relations, nomenclature, lithologic character, and log character of upper Morrowan and lower Atokan units in Arkoma Basin and northern Ouachita Mountains.

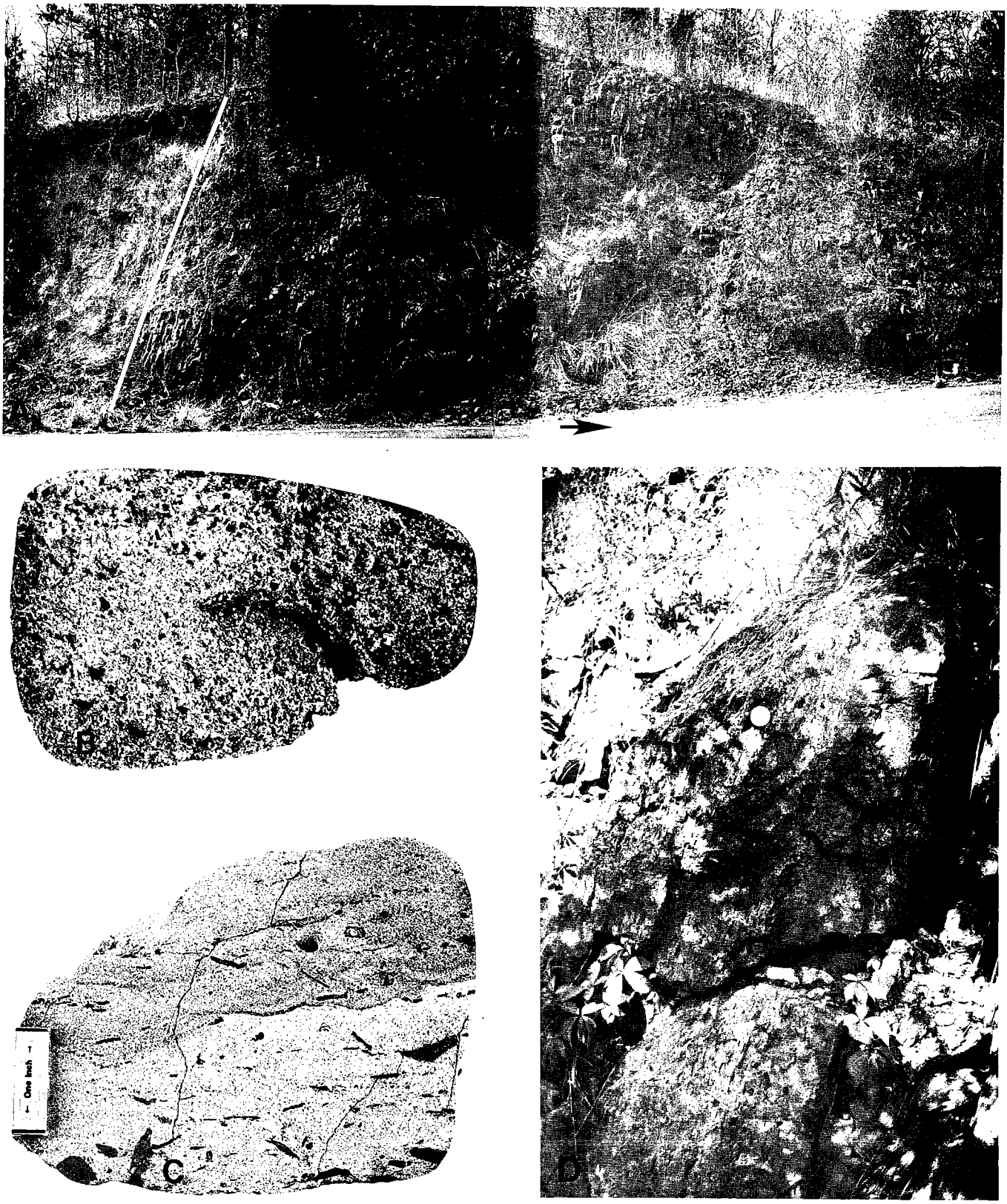


Figure 49 *(above and opposite page)*. *A*—Road cut of the Spiro sandstone at Stop 10. White line marks the transition from the sub-Spiro shale to the Spiro sandstone (arrow shows facing direction). *B*—Medium- to coarse-grained, friable sandstone with abundant fossil molds, 64 ft above the sub-Spiro shale. *C*—Medium-grained sandstone containing abundant shale and clay intraclasts and scattered fossils, 28 ft above the sub-Spiro shale. *D*—Outcrop of cross-bedded sandstone containing abundant shale and clay intraclasts, 28 ft above the sub-Spiro shale. Coin (quarter) in upper-center for scale. *E*—Intensely bioturbated sandstone near the base of the Spiro sandstone, 10 ft above the sub-Spiro shale.

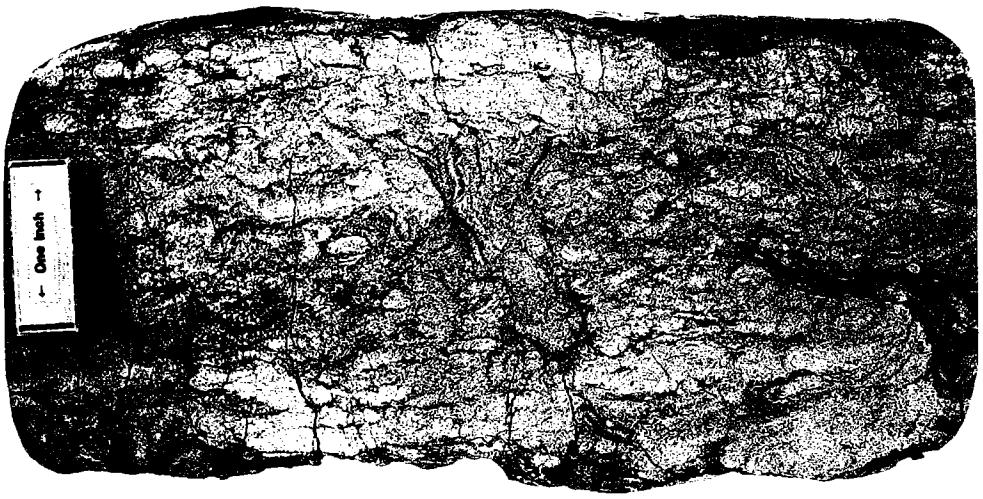


Figure 49 *(continued)*.

rite, probably iron-rich, coats quartz grains in a few samples. In the more porous intervals, limonite and hematite are common as grain coatings.

Silica-quartz overgrowth, the most common cement type, is developed most extensively where there was little clay to inhibit its growth. Calcite cement, which is less common, occurs mainly in rocks that contain some fossils.

Porosity is characteristic of the Spiro at this locality. Some intergranular, primary porosity occurs at grain contacts and in zones where it was preserved by early-chlorite grain coatings. The dissolution of glauconite, matrix, chamosite, fossils, and quartz contributes to secondary porosity.

INTERPRETATION

The occurrence of burrows, bioturbation, fossils, and glauconite suggests that the Spiro sandstone at this outcrop probably was deposited in a shallow-marine environment. The cross-bedding and coarser grain size at the base of the uppermost interval, in the third unit from the top, and in the second interval from the base suggest possible channeling. The limestone in the second interval from the top apparently represents a period when the sand source was cut off.

Continue south on S.H. 82.

58.7	0.1	Cross trace of north-directed thrust fault juxtaposing "Springer" Formation (equals Morrow shale in petroleum-industry parlance) in the hanging wall against Spiro sandstone in the footwall.
59.0	0.3	Good exposure of "Springer" Formation in bar ditch on east side of road (Fig. 44). This structure is called the Pigeon Creek anticline on the geologic map of the McAlester 2° sheet published by the OGS (Marcher and Bergman, 1983). Detailed mapping in this area by Hemish and others (1990b) shows that this structure does not exist.
59.2	0.2	Fair exposure of Spiro sandstone on west side of road. This is the structurally highest and, presumably, the most-basinward outcrop of Spiro sandstone in this part of the Ouachita frontal belt. The Spiro pinches out ~2 mi east of here and turbidites of the Atoka Formation are deposited on the "Springer" Formation.
59.6	0.4	Leave Red Oak 7.5′ quadrangle, enter Talihina 7.5′ quadrangle.
60.1	0.5	Most of the Atoka Formation (best exposed on east side of highway) faces south and dips south, but small areas with north-facing and overturned strata indicate that the Atoka is complexly folded.
60.6	0.5	Stop 11. Excellent road cut on northeast side of highway.

STOP 11

Atoka Formation Turbidites

Location: C NE¼ sec. 34, T. 5 N., R. 21 E., Latimer County

A well-exposed, continuous, and almost unfaulted section of the Atoka Formation is located on the east side of S.H. 82, ~2.5 mi north of Bengal (Fig. 50). The strata exposed in this outcrop are typical of the Atoka Formation in this part of the frontal belt. Fine-grained sandstone comprises ~17% of the section (Fig. 51A,B), and shale and minor siltstone the remainder. Most of the sandstone beds have sedimentary features typical of sediments deposited by turbidity currents, such as partial (no Ta was observed) Bouma sequences (Fig. 46),



Figure 50. Turbidite sandstone and shales (mostly grass-covered) of the Atoka Formation at Stop 11. Facing direction is to the right. Thick sandstone bed on left side of photo is Unit 24, Measured Section, Stop 11.

sole marks (trace fossils, flute and groove casts), dewatering (dish-and-pillar) structures, and soft-sediment slump structures (convolute stratification). Some sandstone beds (e.g., Units 20 and 34, Measured Section, Stop 11) that do not have identifiable Bouma sequences may have been deposited by mass-flow processes (e.g., debris flows) and may not be "true" turbidites. Paleocurrent azimuths based on groove casts range from N. 53° W. to S. 51° W. Paleocurrent bearings based on flute casts range from N. 77° W. to S. 68° W. Fining-upward or -downward megasequences are not present (Fig. 51B).

The purpose of this stop is to examine features typical of turbidites in the Atoka Formation. In addition, some discussion may focus on distinguishing between turbidite and debris-flow sandstones. Without looking at the descriptions in the Measured Section, Stop 11, can you identify the north-facing sandstone beds? (Hint: they're near the south end of the outcrop.)

Continue south on S.H. 82.

61.2 0.6 Road to left (east) to Cities Service No. 1 Smith-English A (~1.5 mi).

The Cities Service No. 1 Smith-English A is located near the C S½N½ sec. 36, T. 5 N., R. 21 E. It spudded in Atoka Formation on October 20, 1983, and reached a total depth of 14,902 ft on August 6, 1984. Interpreted log tops (K.B. 811 ft) are: Spiro—14,333 ft; Wapanucka—14,649 ft; "Springer"—14,759 ft. Based on neutron-density/density-porosity crossover, mudlog shows, and sidewall-core analyses, Cities Service perforated and fractured several Atoka sandstones at ~8,000 ft, ~8,700 ft, and ~12,000 ft. Initial-test data released to the Oklahoma Corporation Commission indicated a combined flow-rate of 190 Mcf gas per day. The well has been abandoned, but gas can still be heard bubbling inside the surface casing during the rainy season when the water table is high.

Two recently drilled wells are also located just east of here. The Arkla No. 1 Smith-English (SW¼SW¼NE¼ sec. 25, T. 5 N., R. 21 E.) spudded on August 17, 1990, and reached a total depth of 10,702 ft on September 27, 1990. Reported log tops are: Hartshorne [*sic*]—1,660 ft; Atoka M—3,770 ft; Wapanucka—4,320 ft; Shay—9,940 ft. Test data indicated some production from the Wapanucka and 300 Mcf gas per day from the middle Atoka, although mapping by Hemish and others (1990b) suggests lower rather than middle Atoka. The Amoco No. 1-31 Talihina (NE¼NW¼NE¼ sec. 5, T. 4 N., R. 22 E.) spudded on March 25, 1991, and reached a total depth of

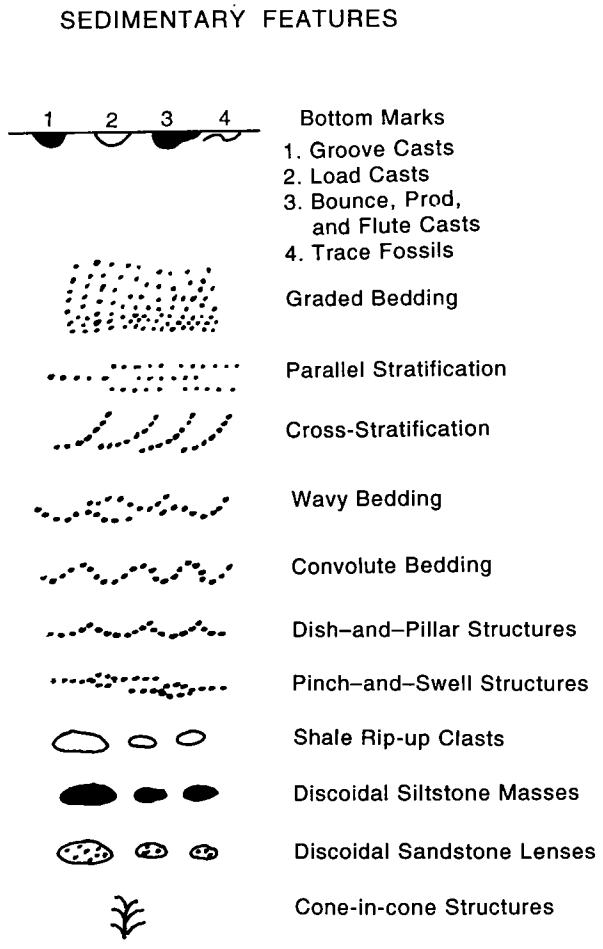


Figure 51. *A (above)*—Explanation of symbols for sedimentary features shown in Figures 51B, 56, 73, and 75. *B (see p. 69–70)*.

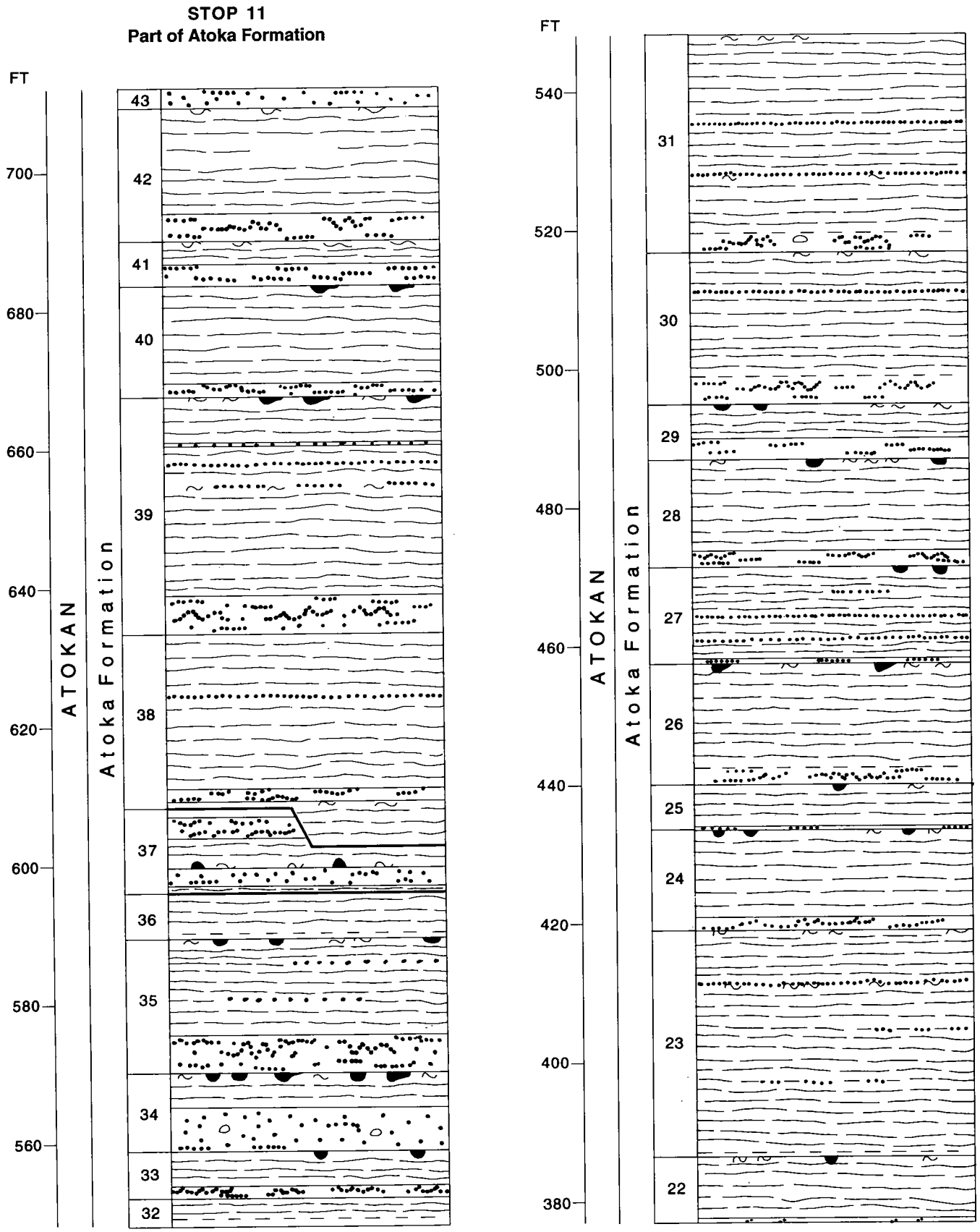


Figure 51. *B*—Graphic columnar section of part of Atoka Formation at Stop 11.

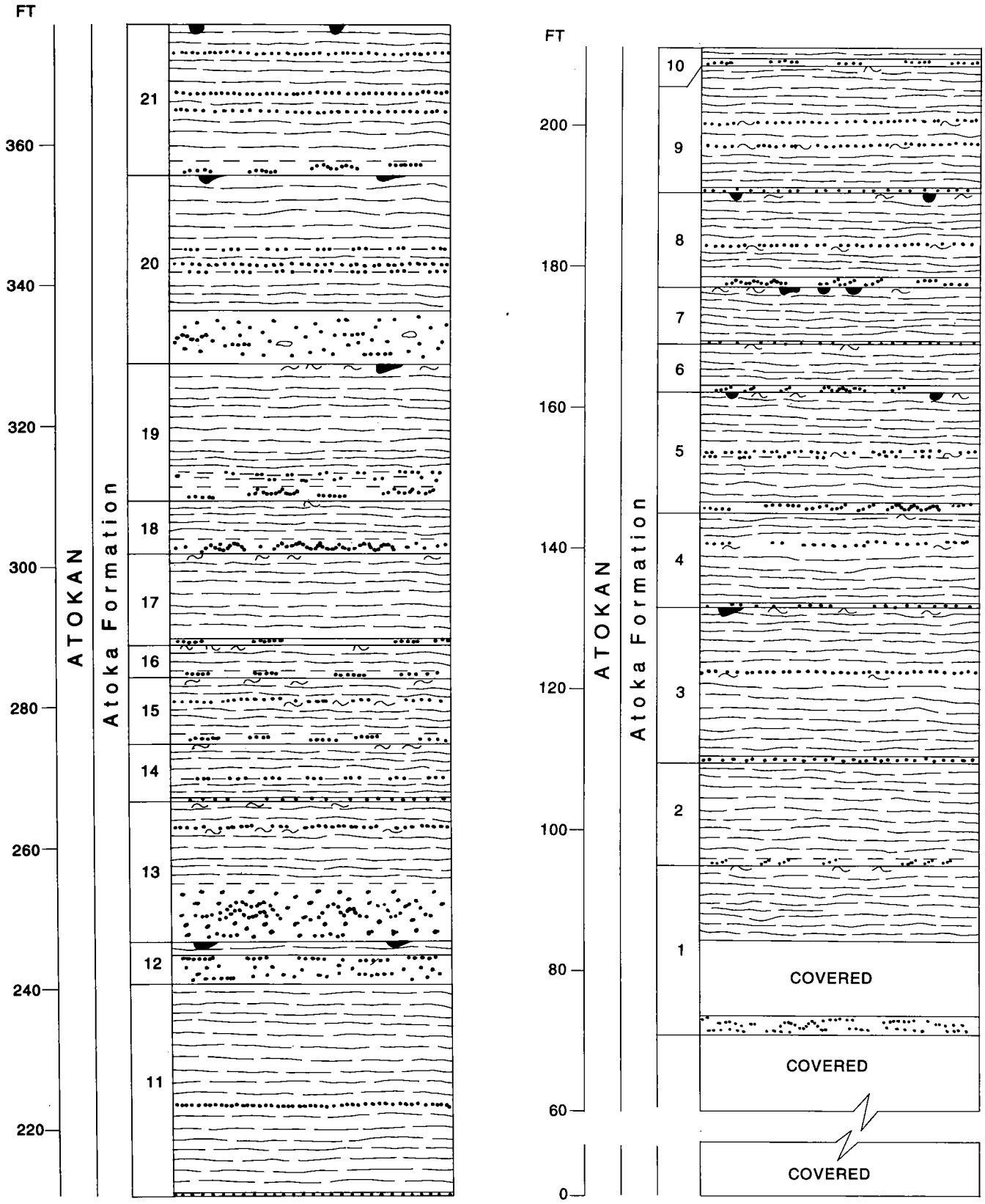


Figure 51. *B (continued).*

Measured Section, Stop 11

Part of Atoka Formation

C NE¼ sec. 34, T. 5 N., R. 21 E., Latimer County, Oklahoma. Base of measured section begins at intersection of S.H. 82 and driveway to small church on east side of highway.

Thickness (feet)

43. Sandstone. 2.3 ft thick, olive gray (5Y4/1) to light olive gray (5Y5/2). Lower 2 ft massive, blocky, weathered. Upper 0.5 ft platy. Base with trace fossils. Sandstone fine-grained, quartzose, abundant limonitized plant debris at top. Top covered (top of measured section). 2.3
42. Sandstone. 4.5 ft thick, light olive gray (5Y5/2) to yellowish gray (5Y7/2). Basal 0.5 ft Tb, remainder Tb or convolute with locally overturned folds. Soft-sediment deformation features result of dewatering and shearing. Base with trace fossils. Top 0.5 ft weathered, soft, probably grades into overlying shale. Sandstone fine-grained, silty, quartzose, abundant mica, plant compressions parallel to laminations. Overlain by shale, largely covered 18.6
41. Sandstone. 3.3 ft thick, olive gray (5Y3/2). Parallel-stratified throughout, possibly Tbd. Lower and middle parts blocky, upper part flaggy. Base with flute casts (paleocurrent N. 80° E. to S. 80° W.). Sandstone fine-grained, quartzose, micaceous, local concentrations of carbonized plant debris parallel to stratification in middle and towards top of bed. Overlain by shale 6.7
40. Sandstone. 2.3 ft thick, light olive gray (5Y5/2). Mostly Tb, top 0.5 ft slightly wavy bedded (Tc). Base with trace fossils, flute casts (paleocurrent N. 68° E. to S. 68° W.). Sandstone fine-grained, micaceous, locally with plant compressions parallel to stratification. Overlain by shale 16.0
39. Sandstone. 5.5 ft thick, light olive gray (5Y5/2). Basal 1.5 ft faintly parallel stratified, overlain by 1.5 ft-thick convolute layer with abundant dish-and-pillar structures. Layers pinch against pillars but are not truncated. Convolute layer overlain by 1-ft-thick platy, parallel-stratified layer, the basal part of which is draped into underlying convolutions. Upper 1 ft is poorly exposed. Sandstone fine-grained. Overlain by shale interbedded with three 0.3–0.5-ft-thick sandstone beds, one of which has trace fossils on its base.. 33.8
38. Sandstone. 2 ft thick, consists of 1-ft-thick Tb overlain by 1-ft-thick Tc. Tb locally wavy bedded, grades up to Tc. Base with trace fossils. Overlain by slightly silty, olive gray (5Y3/2) shale interbedded with a 0.4-ft-thick sandstone 24.4

FAULT

37. Sandstone. 3 ft thick, consists of 2-ft-thick Tb overlain by 1-ft-thick Tc. Faces north. Does not extend to road; may be truncated by fault. Overlain by shale. Shale overlain by sandstone, 2.5 ft thick. Lower part faintly parallel stratified. Base with trace fossils, randomly oriented groove casts. Top with abundant limonitized plant stems, debris. Faces north. Overlain by shale 12.4

FAULT

36. Sandstone. 0.7 ft thick, olive gray (5Y3/2). Thin basal Tb overlain by locally convolute Tc overlain by thin, draped Td. Base with trace fossils, groove casts (paleocurrent N. 51° E.-S. 51° W.). Sandstone very fine grained, abundant mica parallel to laminations. Top covered. Overlain by olive gray (5Y3/2) silty shale that is fissile where weathered 7.2
35. Sandstone. 5.5 ft thick, light olive gray (5Y5/2). Consists of basal 2-ft-thick blocky, faintly parallel stratified Tb overlain by 0.5–1-ft-thick Tc. Tc is overlain by an ~2-ft-thick convolute layer that is overlain by wavy-bedded Tc. Base with trace fossils, prod and bounce marks, rare flute casts (paleocurrent N. 75° E. to S. 75° W.). Sandstone fine-grained, hard. Overlain by shale with two 2-in.-thick sandstone beds 19.5
34. Sandstone. 6 ft thick. Basal 0.7 ft faintly parallel stratified. Remainder unstratified to faintly parallel stratified, locally with shale rip-up clasts as large as 2 in., weathered. Base with groove casts (paleocurrent S. 77° E.-N. 77° W.). Overlain by shale 11.3
33. Sandstone. 1.8 ft thick. Basal 0.3 ft Tb, gradually develops convolutions so that upper part of bed is convolute to very convolute and has overturned to locally detached folds. Possible dish-and-pillar structures. Appears to have sharp top. Overlain by shale 6.5
32. Sandstone. 0.7 ft thick, olive gray (5Y4/1). Consists of 0.2 ft-thick Tc overlain by Td. Lower part flaggy, upper part platy. Base with trace fossils, top grades upwards to siltstone and shale. Very fine grained. Overlain by shale 4.0
31. Sandstone. 3 ft thick, light olive gray (5Y5/2). Consists of 0.8 ft-thick Tb overlain by unstratified to locally convolute Tb or Tc with shale rip-up clasts. Base with trace fossils, top poorly exposed. Upper part with locally abundant limonitized plant debris as large as 1 in. Sandstone fine-grained, quartzose, hard lower part, soft upper part. Overlain by olive gray (5Y3/2) shale with 2 thin T(c)d sandstone beds, one with trace fossils on base, that grade up to shale 31.6
30. Sandstone. 4 ft thick, olive gray (5Y3/2). Lower part parallel-stratified (Tb), upper part wavy-bedded to convolute, locally with overturned folds. Base with trace fossils, groove casts (paleocurrent N. 70° E.-S. 70° W.). Character of base changes abruptly from planar with small groove casts to irregular with large trace fossils. Sandstone hard, quartzose. Overlain by shale with one 0.4 ft-thick Tcd sandstone bed 21.8
29. Sandstone. 3 ft thick, light olive gray (5Y5/2). Slabby Tb in lower part, platy Tb in upper part. About 1.7 ft above base is a 0.1 ft-thick, continuous convolute layer. Base with trace fossils and groove casts (paleocurrent N. 75° E.-S. 75° W.). Sandstone fine-grained, quartzose, micaceous. Overlain by shale 8.1
28. Sandstone. 2.2 ft thick. Basal 1.2 ft Tb, remainder wavy-bedded Tc with very gently inclined cross-stratification. Small dish-and-pillar structures locally. Base with small groove casts (paleocurrent N. 68° E.-S. 68° W.). Carbonized plant stems as long as 5 in. at top of bed. Overlain by shale 15.4
27. Sandstone. 0.6 ft thick. Tcd with trace fossils and flute casts (paleocurrent N. 78° E. to S. 78° W.) on base. Abundant, thoroughly iron-oxide-stained plant debris concentrated parallel to laminations near top. Overlain by shale with two, thin Tcd sandstone beds that grade upward to siltstone and shale. Also a 0.5-in.-thick sandstone lens 13.7
26. Sandstone. 2.3 ft thick, light olive gray (5Y5/2). Consists of 0.5 ft-thick Tb overlain by Tc with large-scale wavy-, locally convolute stratification and with dish-and-pillar structures. Tc overlain by about 0.3 ft-thick Td. Base generally planar and featureless, although with faint, randomly oriented groove casts. Top covered. Overlain by silty, olive gray (5Y3/2) shale 17.8
25. Sandstone. 0.5 ft thick, olive gray (5Y4/1). Consists of 0.1 ft-thick Tc overlain by 0.4 ft-thick Td. Base with trace fossils and groove casts (paleocurrent S. 63° E.-N. 53° W.). Sandstone fine-grained with disseminated carbonized organic matter throughout. Overlain by shale .. 6.5

24. Sandstone. 2 ft thick, olive gray (5Y4/1). Consists of 1.5-ft-thick Tb overlain by 0.5-ft-thick Tc. Upper part of Tb through Tc slighty convolute to wavy-bedded. Base with trace fossils. Upper surface of Tc forms interference ripple marks, thin Td draped over ripples. Sandstone hard, fine-grained, quartzose, 0.1-in. mica. Overlain by olive gray (5Y3/2) shale 14.6
23. Sandstone. 0.9 ft thick, light olive gray (5Y5/2). 0.1-ft-thick Tc overlain by platy Td. Base with trace fossils, faint groove casts that vary in rake from 5° E. to 49° E. (paleocurrent N. 75° E.-S. 75° W. to S. 61° E.-N. 61° W.). Top grades into overlying shale. Quartzose, mica parallel to laminations. Overlain by flaky-weathering, olive gray (5Y3/2) shale. Shale interbedded with rare, thin Td siltstones and one 0.3-ft-thick Td sandstone with trace fossils on base 32.3
22. Sandstone. 0.5 ft thick. Consists of 0.3-ft-thick Tc with low-inclination cross-stratification overlain by 0.2-ft-thick Td. Base with groove casts (paleocurrent N. 76° E.-S. 76° W.) Overlain by shale 9.7
21. Sandstone. 1.8 ft thick, light olive gray (5Y5/2). Basal 0.5 ft Tb, grades up to faintly cross stratified and slightly convolute Tc. Tc appears wavy bedded and grades up to mottled Td. Td draped over wavy-topped Tc. Base with rare tool marks and unusual 0.5–2-in.-deep "dimples." Sandstone fine-grained, mica conspicuous. Lower part hard, upper part softer (weathered). Grades up to siltstone and shale. Overlain by slightly silty, olive gray (5Y3/2) shale. Shale interbedded with two thin siltstone and one thin, platy, light olive gray (5Y5/2) Tcd sandstone beds 21.1
20. Sandstone. 7.5 ft thick, light olive gray (5Y5/2). Lower 2 ft faintly parallel stratified, grades upwards to mostly unstratified to faintly convolute or parallel-stratified sandstone. Base with trace fossils and flute casts (paleocurrent S. 78° E. to N. 78° W.). Sandstone fine-grained, quartzose, locally with shale rip-up clasts. Middle and top parts weathered. Overlain by flaky-weathering, olive gray (5Y3/2) shale that shows slight pencil structure. Interbedded in shale is one 0.3-ft-thick, flaggy-weathering Td sandstone that grades up to siltstone and shale. Two thin siltstone beds also interbedded in shale 26.8
19. Sandstone. 2 ft thick, lower part with faint parallel stratification (Tb), becomes less faint and more convolute upwards. Top part with large dish-and-pillar structures. Base with rare and small trace fossils. Top poorly exposed, but appears to grade upward to ~2 ft of siltstone that grades upward to shale. Overlain by very silty, olive gray (5Y3/2) shale 19.4
18. Sandstone. Approximately 2 ft thick, top covered. Light olive gray (5Y5/2). Basal 1 ft massive with faint convolute (locally overturned) stratification, top mottled, weathers to slabs, but also shows convolute stratification and local dish-and-pillar structures. Base with trace fossils. Fine-grained, micaceous, locally abundant carbonized plant debris as long as 0.5 in. Overlain by shale 7.6
17. Sandstone. 1 ft thick. Basal 0.2 ft Tc, overlain by 0.8 ft of flaggy to platy Tb or Td. Base with trace fossils. Overlain by olive gray (5Y3/2) shale that weathers to flaky appearance on surface 12.6
16. Sandstone. 1.3 ft thick, light olive gray (5Y5/2). Consists of 0.8-ft-thick Tb overlain by 0.3-ft-thick Tc overlain by 0.2-ft-thick Td. Base with trace fossils, top grades into overlying shale. Mica conspicuous. Overlain by shale 5.1
15. Sandstone. 1.4 ft thick, light olive gray (5Y5/2). Top 0.3 ft platy Td, underlain by 0.2 ft Tc, remainder Tb. Base with trace fossils. Top not exposed, may grade up into shale. Sandstone fine-grained with abundant mica and small pieces of carbonized plant debris. Overlain by shale with a 0.4-ft-thick, olive gray (5Y4/1), Tbcd sandstone bed. 0.1-ft-thick Td appears draped between wavy Tc. Trace fossils on base, top grades into overlying shale. Sandstone very fine grained, quartzose, with abundant mica that is conspicuous on laminations. Weathers to 1–2-in.-thick "flagstones".. 9.2
14. Sandstone. 0.6 ft thick. Basal 0.1 ft Tb; remainder convolute with locally overturned and truncated syndepositional folds. Base with trace fossils. Overlain by shale interbedded with one 0.1-ft-thick Td siltstone. 8.5
13. Sandstone. 8 ft thick, medium light gray (N6). Basal 1 ft faintly parallel stratified, middle part slightly to very convoluted, top 1 ft mottled, stratification poorly preserved. Carbonized plant debris common in upper half of sandstone. Base with flute casts (paleocurrent N. 79° E. to S. 79° W.). Fine-grained, quartzose. Overlain by slightly silty, olive gray (5Y3/2) shale. Shale interbedded with one 0.4-ft-thick, olive gray (5Y4/1) sandstone. Sandstone consists of 0.1-ft-thick Tc overlain by 0.3-ft-thick Tb overlain by 0.1-ft-thick Tc. Cross-laminations locally show slight overturning due to shear while sediment was soft. Base with trace fossils, top wavy. Sandstone very fine grained 19.4
12. Sandstone. 4 ft thick, dusky yellow (5Y6/4), mostly weathered. Lower part massive, faintly parallel stratified; top part platy, parallel-stratified. Top and base not exposed. Locally contains carbonized plant debris. Overlain by shale 5.9
11. Sandstone. 0.5 ft thick, Tcd. Poorly exposed. Overlain by slightly silty, well-stratified, olive gray (5Y3/2) shale. Weathers to flaky appearance. Shale interbedded with one 0.3-ft-thick Td sandstone that is overlain by siltstone that grades up to shale. Sandstone very fine grained with abundant mica. Top of interval poorly exposed 30.4
10. Sandstone. 1.3 ft thick, light olive gray (5Y5/2) where fresh, grayish brown (5YR3/2) where weathered. Locally "bleached" to a very light color. Faintly parallel stratified (Tbd?), locally shows slight soft-sediment deformation. Base planar with rare trace fossils. Weathered. Overlain by shale 2.5
9. Sandstone. 1 ft thick, olive gray (5Y4/1). Consists of 0.8-ft-thick Tc overlain by 0.2-ft-thick platy-weathering Td. Tc locally shows soft-sediment deformation. Base with large and small trace fossils, faint groove casts (paleocurrent S. 83° E.-N. 83° W.). Abundant mica and carbonized plant debris parallel to laminations. Overlain by shale interbedded with two thin, very fine grained, silty, Td sandstone beds with trace fossils on base 17.7
8. Sandstone. 1.5 ft thick, olive gray (5Y4/1) where fresh, pale brown (5YR5/2) where weathered. 1-ft-thick Tb (locally showing soft-sediment deformation) underlying 0.3-ft-thick Tc with rare dish-and-pillar structures. Top is platy Td. Dish-and-pillar structures <2 in. high and resemble wavy beds; pillars locally poorly developed. Base with large and small trace fossils, flute and groove casts (paleocurrent S. 80° E. to N. 80° W.). Local accumulations of carbonized plant debris on laminations. Overlain by fissile, olive gray (5Y4/1) shale. Surface of shale outcrop appears contorted, but is evenly bedded beneath. Thin Tc sandstone showing slight pinch-and-swell and with trace fossils on base interbedded in shale 13.5
7. Sandstone. 0.5 ft thick, olive gray (5Y4/1). 0.4-ft-thick Tc overlain by platy Td. Tc shows some soft-sediment deformation and dish-and-pillar structures that are locally truncated by parallel laminations of Td. Base with some trace fossils. Overlain by shale 8.4
6. Sandstone. 1.2 ft thick, light olive gray (5Y5/2). 0.8-ft-

thick low-inclination Tc overlain by platy Td. Tc locally shows soft-sediment deformation. Base with abundant trace fossils and groove casts (paleocurrent S. 71° E.-N. 71° W.). Top of Tc slightly undulatory. Overlain by shale 6.8

5. Sandstone. 1.4 ft thick, light olive gray (5Y5/2). Basal 0.5-ft-thick Tb overlain by 0.3-ft-thick flaggy- to platy-weathering Tc; both locally show soft-sediment deformation. Top of sandstone platy, poorly exposed, weathered; probably Td. Base planar with small trace fossils. Overlain by olive gray (5Y3/2), silty, fissile shale with a 0.3-ft-thick, fine-grained, olive gray (5Y 4/1), Tcd sandstone showing low-inclination cross-stratification, trace fossils on base, and mica and abundant carbonized plant debris parallel to laminations. A 0.5-in.-thick Td(?) siltstone bed is present below the thin sandstone 16.8
4. Sandstone. 0.8 ft thick, light olive gray (5Y5/2) to olive gray (5Y4/1). Basal 0.1 ft Tb, overlain by 0.5 ft Tc, grades into overlying 0.2 ft Td siltstone. Base with abundant trace fossils, flute casts (paleocurrent S. 77° E. to N. 77° W.). Abundant mica, moderate carbonized plant debris parallel to stratification. Overlain by shale interbedded with a thin Tcd sandstone-siltstone bed with abundant trace fossils on its base 13.5
3. Sandstone. 1 ft thick, very poorly exposed. Overlain by shale and 0.2-ft-thick sandstone with trace fossils on base. Shale olive black (5Y2/1), silty, poorly exposed. 22.0
2. Sandstone. 0.9 ft thick, light olive gray (5Y5/2). Basal 0.8 ft is cross-stratified (low inclination) (Tc), overlain by parallel-stratified Td. Base with small and large trace fossils. White mica parallel to stratification. Grades up to siltstone. Overlain by shale 14.3
1. Sandstone. Poorly exposed, 3.1 ft thick. Parallel-stratified (Tb) basal 0.5 ft, overlain by 0.5-ft-thick convolute layer (soft-sediment deformation). Top part Tb. Weathers flaggy to platy. Overlain by covered interval (probably shale) and shale. Shale olive black (5Y2/1) to olive gray (5Y3/2), silty, fissile 24.0

Base of section (covered) 70.8

Total thickness of section 710.5

15,112 ft on June 4, 1991. The bottomhole location is ~3,000 ft north of the surface location. Log tops reported to the Oklahoma Corporation Commission are: Atoka M—4,559 ft; Atoka L—11,168 ft; Spiro—14,656 ft; Wapanucka—14,822 ft. The well was completed in the lower Atoka (12,214–12,240 ft) for an initial flow rate of 2.6 MMcf gas per day.

61.7	0.5	Cross Long Creek.
62.2	0.5	Steeply south-dipping, south-facing strata of Atoka Formation on west side of road.
63.0	0.8	Enter Bengal, Oklahoma. Bengal was established as a post office on January 24, 1890. Its name comes from a state in eastern India and was used to name a supposed ancestral Ouachita mountain range (Bengalia). Turn right (west) on paved county road.

BENGALIA

Powers (1928, p. 1044) was the first geologist to suggest that a high landmass ("pre-Ouachita uplift") existed in the northern part of the Ouachita Mountains during Stanley–Jackfork sedimentation. Kramer (1933, p. 612) named this landmass "Bengalia" and identified it as a long-lived "land area along the north side of the Oklahoma Ouachitas." Bengalia was named after the town of Bengal where Kramer (1933, p. 597) identified outcrops of Arkansas Novaculite which represented (to him) "inliers" (remnants?) of Bengalia. Bengalia supposedly was located between the present traces of the Choctaw and Ti Valley faults in the central and eastern frontal belt of the Ouachita Mountains and just east of the trace of the Windingstair fault in the western Ouachitas. Kramer (1933) believed that the Bengalia landmass was necessary to explain (1) the juxtaposition of foreland- and Ouachita-facies strata, (2) the restriction of the thick Stanley and Jackfork Groups to south and east of the Ti Valley fault, and (3) the presence of foreland-facies boulders in the Johns Valley Shale. Moore (1934) agreed with the need for a highland source terrane for the Johns Valley; he called this terrane the "Ancestral Ouachita Mountains" and suggested it was a relatively short-lived, Early Pennsylvanian feature. Fellows (1964, p. 75–79) reviewed the evidence for Bengalia and added more evidence supporting the existence of Bengalia: (4) Wapanucka Formation–Chickachoc Chert–Johns Valley Shale facies relations. More recently, Gatewood and Fay (1991) attempted to identify Bengalia based on gravity and magnetic data. They suggest that it is a buried horst block 3 mi wide and 102 mi long that was high from the Ordovician to Middle Pennsylvanian.

All the stratigraphic arguments for the existence of the Bengalia landmass assume no large-scale thrusting. Ham's (1959, p. 83) observations regarding the lower and middle Paleozoic strata in the Arbuckle and Ouachita Mountains are valid also for the Carboniferous:

> Stratigraphic correlations and regional interpretations of pre-Stanley rocks of the Arbuckle–Ouachita provinces have been [*sic*] advanced remarkably in the half-century during which these regions have been investigated. In perspective view, it appears reasonable that there are no major differences in age or in lithology during Viola–Bigfork, Sylvan–Polk Creek, and Woodford–Arkansas novaculite times; and that within these time periods, entirely similar sediments were deposited in open seaways that connected the Arbuckle and Ouachita provinces. Moreover, it is equally reasonable that considerably different sediments were being deposited simultaneously in the two provinces, but at different rates, first during Lower and early Middle Ordovician time and later during Silurian and Early Devonian time. At such times the respective provinces were (a) either widely separated, the change in facies between them taking place gradually, in which case large-scale thrusting is called for, or (b) the sediments grade into each other within a short distance, the facies change being possibly aided by a barrier of unknown character and position, in which case thrusting is not needed.

It is now known that large-scale thrusting in the Ouachita Mountains has occurred, although the exact magni-

tude is unknown. Arbenz's (1989, p. 631) suggestion that the frontal belt of the Oklahoma Ouachita Mountains is "a highly telescoped unit shortened to 25% of its original width within which the main facies changes take place" is probably correct and eliminates the need for Bengalia.

There are many specific arguments against evidence used to support the idea of a long- or short-lived depositional barrier separating the area in which foreland-facies strata were deposited from that in which Ouachita-facies strata were deposited. (1) Many of the outcrops of "Arkansas Novaculite" used by Kramer (1933, fig. 2) to outline Bengalia are, in fact, Woodford Chert present at the base of thrust sheets (Hendricks and others, 1947) or as large olistoliths in the Johns Valley Shale (Suneson and others, 1990, p. 50). It is possible that some, if not all, of the Woodford–Caney blocks are exotic (see Stop 12). If this is true, then the Johns Valley Shale extends to the Pine Mountain fault and would include the Woodford–Caney–"Springer" sequence of Hendricks and others (1947). This would explain the apparent absence of a Wapanucka- or Chickachoc-equivalent between the Pine Mountain and Ti Valley faults, used by Fellows (1964, p. 77) as evidence for a bathymetric (or topographic) high. (2) The strata underlying the Potato Hills, also presumed to be a Bengalia inlier, are allochthonous (Arbenz, 1968) (see Stop 23). (3) The differences in post-Simpson/Womble foreland- and Ouachita-facies strata are more apparent than real (Ham, 1959, quoted above). The two facies probably grade into each other in the subsurface, and the transition from foreland- to Ouachita-facies probably occurs at least 50 mi south of "Bengalia" (Arbenz, 1989, pl. 11). (4) Boulders of foreland-facies strata, some extremely large, were deposited as part of the Johns Valley Shale. The shallow-water limestone lenses noted by Fellows (1964, p. 77) in the upper part of the Johns Valley Shale are probably olistoliths and are not evidence of a bathymetric high separating areas of Wapanucka–Chickachoc and Johns Valley deposition.

In summary, Bengalia never existed. Recent studies and mapping clearly demonstrate that the Johns Valley Shale contains foreland-facies olistoliths, some extremely large, that range in age from Ordovician to Pennsylvanian. In addition, modern concepts of thrust belts, including the Ouachita fold and thrust belt, allow for large-scale thrusting. In the Ouachita Mountains, Ouachita-facies deep-water strata have been juxtaposed against foreland-facies shallow-water strata along thrust faults with a cumulative offset on the order of 50 mi or more.

63.7 0.7 Stop 12. Behind farm house and barn in shallow creekbed. This exposure is on private land—please leave all gates as you find them. If you would like to visit this outcrop in the future, please write to: Jones Residence, Route 1, Box 970, Wister, OK 74966.

STOP 12

Woodford Chert–Caney Shale Olistolith in the Johns Valley Shale

Location: NE¼NW¼SW¼ sec. 11, T. 4 N., R. 21 E., Latimer County

This outcrop (Fig. 52) was first described by Kramer (1933), who considered it to be Arkansas Novaculite. It did not appear on the 1926 geologic map of Oklahoma (Miser, 1926), but did appear on the 1954 map (Miser, 1954). Fellows (1964) described and mapped the outcrop in detail; his measured section is reproduced here as Measured Section, Stop 12. The bedrock outcrop is almost 400 ft long. Its eastern end is covered by alluvium, but its western end is difficult to locate precisely because bedrock exposures gradually pass into float blocks. However, local accumulations of Woodford float suggest that the outcrop is at least 2,000 ft long. Based on detailed mapping in the Talihina 7.5′ quadrangle, Suneson and



Figure 52. Woodford Chert at Stop 12. This large outcrop of Woodford Chert and Caney Shale is probably a submarine-slide block in the Johns Valley Shale. Photograph was taken looking down at top of outcrop. Dip of beds is nearly vertical.

Measured Section, Stop 12

Woodford Chert–Caney Shale in the Johns Valley Shale

NE¼NW¼SW¼ sec. 11, T. 4 N., R. 21 E., Latimer County. Section measured along tributary of Long Creek west of Bengal (from Fellows, 1964, p. 97).

	Thickness (feet)
CANEY SHALE:	
9. Shale, black, fissile; with dense, black petroliferous, limestone concretions	30.0
8. Siltstone, gray, thin-bedded; with plant-fragment impressions	12.0
7. Shale, green, silty	2.0
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 0.5–5 in. thick; and shale, black, fissile, with common, but not abundant, phosphatic nodules	11.0
3. Shale, olive green, silty, easily eroded; resembles some shales of the Stanley Group	3.3
2. Chert, black; in beds 0.5–6 in. thick; and shale, nodules are present in both chert and shale	14.0
1. Shale, black, fissile	7.0
Total Woodford exposed	35.8

Ferguson (1990) interpreted this outcrop as a large olistolith in the Johns Valley Shale and not a thrust slice surrounded on all sides by faults (Fellows, 1964).

This exposure is the easternmost of several large outcrops of Woodford Chert–Caney Shale that extend from here to near Wesley, Oklahoma (sec. 1, T. 1 N., R. 13 E.). Hendricks and others (1947) believed these large outcrops marked the base of the hanging walls of major thrust faults and mapped some thrusts accordingly. It seems possible that the many large outcrops of (Pinetop)–Woodford–Caney–"Springer" in the frontal belt north of the Ti Valley fault are stratigraphically allochthonous, that is, are large, stratigraphically coherent, submarine-slide or debris-flow blocks that are part of the Lower Pennsylvanian section (Johns Valley) below the Atoka Formation. These outcrops may represent slide blocks initially derived from the north that subsequently were moved to the north as parts of thrust sheets. The smaller fragments of Woodford Chert and Caney Shale that are common in the more typical Johns Valley outcrops to the south may represent the more broken, basinward equivalent of the large, coherent blocks similar to the one at this stop.

The Woodford Chert and its equivalent units have long been recognized as petroleum source rocks in the Midcontinent region. Therefore, the different interpretations of the Woodford–Caney outcrops here and in other parts of the Ouachita Mountains frontal belt have significant implications for hydrocarbon exploration. If the interpretation that the Woodford–Caney outcrops are large olistoliths in the Johns Valley Shale is correct, then the extent of presumably autochthonous Woodford beneath the thrust sheets is unknown and would depend on where the Woodford olistoliths were derived. To determine this, it would have to be known how far to the south the olistoliths were carried as parts of slides and debris flows and how far to the north they were subsequently thrusted. Alternatively, if the Woodford–Caney outcrops are stratigraphically autochthonous, then sub-thrust Woodford is present in the subsurface south of the frontal belt—how far to the south depends on the cumulative displacement on the thrust faults.

TABLE 4. — THERMAL MATURITY OF THE WOODFORD CHERT, SOUTHERN LATIMER COUNTY, OKLAHOMA

Sample no.	Location	Vitrinite reflectance[a] %R_o Range	Mean
OPL679	NE¼NW¼SW¼ sec. 11, T. 4 N., R. 21 E.	0.51–0.79 (40 values)	0.63

[a]Unpublished data, Brian J. Cardott, OGS.

The thermal maturity of the Woodford chert in the exposure is given in Table 4. This value is almost identical to that of smaller Woodford chert outcrops about 5 mi to the west (B. J. Cardott, personal communication). These outcrops are clearly olistoliths in the Johns Valley Shale. The similar maturity values of small Woodford chert blocks in the Johns Valley and that of large blocks of less-clear origin support the interpretation that the large blocks are olistoliths.

Two recently drilled wells are located just to the west of Stop 12. The ARCO (formerly Exxon) No. 1 Ellis (SE¼ SW¼SE¼ sec. 4, T. 4 N., R. 21 E.) spudded on November 2, 1990, and reached a total depth of 15,900 ft on March 9, 1991. The bottom-hole location of the well is ~2,500 ft north-northwest of the surface location. Reported log tops are: Pine Mountain thrust—7,520 ft; Spiro—11,340 ft; Choctaw thrust—12,220 ft; Cecil B—13,510 ft; Spiro—15,380 ft; Wapanucka—15,623 ft; Union Valley—15,808 ft. The well tested noncommercial gas in two Spiro zones (15,450–15,514 ft and 11,334–11,344 ft); higher Atoka sandstones also were tested but the results were not reported. The Mobil No. 1-8 Peachland Creek (SE¼NW¼ SE¼ sec. 8, T. 4 N., R. 21 E.) spudded on June 24, 1991, and reached 17,798 ft total depth on October 30, 1991. Reported log tops are: Atoka—3,500 ft; Wapanucka—15,525 ft. The Wapanucka was perforated from 16,538 ft to 16,572 ft and swabbed dry. Both wells were abandoned.

Retrace route to Bengal.

64.4	0.7	Bengal. Turn right (south) on S.H. 82.
64.5	0.1	Cross trace of "Ti Valley" fault. Objections to naming this (or any) fault the "Ti Valley" have been published elsewhere (Suneson, 1988; Suneson and others, 1990, p. 32). In this area, the Johns Valley Shale is restricted in outcrop to south of a single fault. To the west (Higgins 7.5′ quadrangle) and south-

THE ST. LOUIS-SAN FRANCISCO RAILWAY

The St. Louis–San Francisco (or "Frisco") Railway is the largest railroad in Oklahoma. Its period of greatest growth in Oklahoma was 1900–1904 (Gardner, 1958, p. 11), but in 1886–87, it constructed one of its earliest and longest (144.4 miles) lines, between the eastern border of Oklahoma just west of Fort Smith and the Texas border south of Hugo. Part of this line crossed Winding Stair Mountain between Bengal and Talihina. The history of this railroad is particularly interesting because it reveals the changing attitudes of the Indians in the Choctaw Nation toward white settlers (as represented by the U.S. Government) and vice versa, following the Civil War. The Indians did not view the railroads as beneficial to them, as they had the early military roads and stage lines; rather, they realized that railroads were the first part of an invasion of Indian land by white settlers.

In 1866, the U.S. Government and the Five Civilized Tribes signed the Peace Treaties of 1866, which stated that past treaties between them were void because the Indian Nations had sided with the Confederacy during the Civil War and that two railroads (one north-south, one east-west) would be built across the Indian Territory. The north-south route was to be awarded to the first company to reach a particular point on the south border of Kansas near Chetopah. In the spring of 1870, the Missouri, Kansas, and Texas Railway Company ("Katy") reached this point and was granted permission by the U.S. Government to continue to build to Denison, Texas. The line was completed in late 1872. The Government granted the Katy the right-of-way across the Indian lands, but it did not allow land grants.

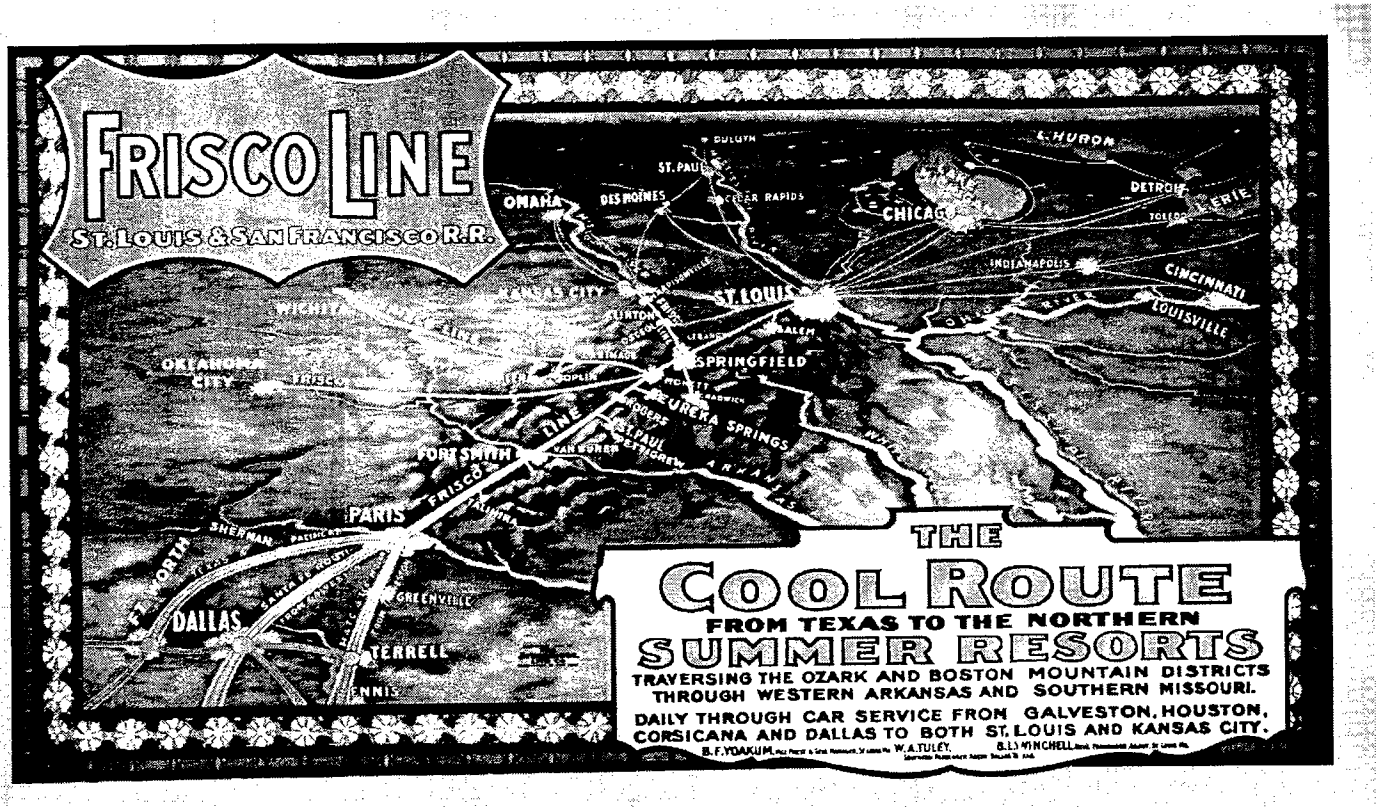
In July 1866, the U.S. Congress chartered the Atlantic and Pacific Railway Company to build a railroad across the Territory from east to west. John C. Fremont, the noted explorer of the American West, was a principal organizer of investors for the new company. Congressional debate on the railroad "centered almost entirely on clerical detail" (Miner, 1972, p. 47). In March 1866, however, the issue of Indian land rights was raised in the U.S. Senate. Congress appeared prepared to issue land grants to the A. & P. through Indian territory even though those lands were not in the public domain. "The axiom that the interest of the United States and railroads it aided was superior to any so-called rights claimed by pockets of alien people standing in the way of commerce" (Miner, 1972, p. 47) held sway. "Congress granted the company 49,500,000 acres in public lands from Missouri to the Pacific, including a conditional grant of twice the usual amount in the Indian Territory. The Indian lands were not public domain, but the Congress had strongly suggested that the United States would take care of the matter of Indian title on behalf of the...company" (Miner, 1969, p. 561).

The railroad relied heavily on land grants, and its ability to sell those lands to settlers, to remain financially solvent. In 1870, the Cherokees passed an act allowing the A. & P. right-of-way privileges only; despite this, in 1871, the A. & P line was built from Seneca to Vinita in the Cherokee Nation. Both towns confirmed Indian suspicions that they wanted little to do with the white man's culture. "The population, whiskey, and morals of Seneca...were all described by the same phrase, 'active but mixed.' It was jokingly said that the only inhabitants of Vinita were U.S. Marshals, there to regulate the houses of ill-repute" (Miner, 1972, p. 79). Without the land grants, the company's financial picture deteriorated because the cattle-shipping business was seasonal and unprofitable. The stage was set for the struggle between the laws of the Five Civilized Tribes and the interests of U.S. companies, and the U.S. Government was caught in between.

In 1871, C. J. Hillyer, company attorney for the A.& P., wrote "one of the most remarkable and important pamphlets in the history of American Indian relations" (Miner, 1969, p. 565). The booklet was entitled *The Atlantic and Pacific Railway and the Indian Territory*, and it "set forward a defense of the 'Manifest Destiny' of white industrial civilization which was so representative of a large portion of 'the climate of opinion' of the times, that it eventually became the essential position of the United States Congress" (Miner, 1969, p. 565). The railway system visualized by Hillyer, centered in Indian Territory, was necessary for the economic well-being of the country; the cost of the system could be paid for with land. The Indians, whose "most prominent and universal characteristic is indolence," should be made wards of the federal government, and any resistance to this policy should be dealt with severely "whether one Indian or five thousand be killed in the operation" (Miner, 1969, p. 565, 566).

In an attempt to thwart A. & P.'s efforts to continue building the railroad across their lands, the Indians demanded exorbitant prices for railroad ties, telegraph poles, roadbed material, and station sites. These high prices, poor business practices, the denial of the land grants, and the panic of 1873 forced A. & P. into receivership in 1875. The A. & P. lines were bought by the Frisco Railway in 1876. The new company made an honest effort to correct some of the problems that plagued the A. & P. In particular, the Frisco attempted to resolve some of the Indian claims against the A. & P.

While this company reorganization was occurring, there was continued pressure on Congress to organize the Indian Territory as a U.S. territory. If this occurred, the Indians realized that the federal government could grant land to the railroad companies and that Indian sovereignty would be compromised. At the same time, there was pressure on Congress to open a new railroad from Fort Smith to Texas, through the Choctaw Nation. In 1877, the question of land grants was ended by a decision from the Federal Land Office Commissioner: "the Frisco–A. & P. combination had no land grant in the Indian Territory 'except as such grant might be acquired from the Indians by the company'" (Miner, 1969, p. 572). Because of this apparent setback, the Frisco abandoned its efforts to build a transcontinental railroad. Instead, it decided to focus its efforts, including a new railway to Texas, nearer to home.



Advertisement for the Frisco Line. Talihina is shown on the same map as Dallas, St. Louis, and Chicago, but the Ouachita Mountains do not rate mention as do the Ozarks and Boston Mountains. (Photograph courtesy of the Frisco Railroad Museum, Springfield, Missouri.)

In 1879, Jay Gould, the Texas railroad baron, became interested in the Frisco because its plans for a Fort Smith to Texas route "threatened his investments [in Texas] with unwanted competition" (Miner, 1972, p. 120). He was too late. The Santa Fe and Frisco railway companies had already agreed to trade directors, set rates, and limit competition between themselves. The marriage went smoothly—for awhile. By 1881, the attitude of the Santa Fe was to attack Gould's Texas monopoly using the Frisco's plans for a route through the Choctaw Nation to Texas.

In October, 1881, Frisco negotiators visited the Choctaw Council in the company of Uri J. Baxter, a law clerk from the Department of the Interior. Apparently, the Council was confused about whether Baxter was representing the U.S. Government or the railroad (Miner, 1969). Baxter did little to resolve the confusion despite government requests that he remain a referee only. Baxter asked the Council for the concession to build the railway through their land. Although a bill granting the concession passed in the Choctaw House of Representatives, the printed version of the bill may not have been the same as the one discussed and there appear to have been certain irregularities in the voting.

The financial concessions offered by the Frisco to the Choctaws were generous; despite this, many Indians were less concerned with material gain than with the principle of national self-determination. In early 1882, delegations from several of the Indian Nations went to Washington, D.C., to attempt to block Senate confirmation of the bill. Supporters of the bill, who obviously would not admit that greed and racism motivated the push to have the Frisco granted the right-of-way across Choctaw land, claimed that those who resisted the development of Indian land were in the pay of the monopolist Gould. Some supporters of the bill dismissed the question of Indian sovereignty altogether. In the Senate debate in April, Senator Augustus Garland (Arkansas) stated, "It is unnecessary to talk about their consent....They are pupils of the Government to be dealt with as the Government sees proper. We deal with them in every conceivable way, from the shirt they wear to the grave we dig for them, and frequently put them in" (Miner, 1969, p. 577).

While the debate continued, Gould acted with characteristic swiftness and unpredictability. In January 1882, he bought controlling interest in the Frisco. Perhaps because the Choctaw lobby to Congress believed the bill could not possibly pass now that Gould owned the railway, they left Washington. Some Senators continued to oppose the bill, but Senate Bill No. 60 passed in July, 1882. The issue of Indian sovereignty over their land died. "The theory of self-determination for the Indians, if it had ever been operative during the treaty period, was officially laid to rest. After 1882, power, not possession, became the test of legitimacy in the expanding industrial nation" (Miner, 1972, p. 128).

In 1885, Congress approved the Frisco to build its railway from Fort Smith to near Paris, Texas. In 1886, the Frisco began construction of its new line across the Choctaw Nation. The line was completed in Paris, Texas, the following year.

		west (secs. 31 and 32, T. 2 N., R. 14 E., as reported by Cline, 1960, p. 63), however, the Johns Valley Shale crops out north of the "Ti Valley" fault. The anastomosing character of the thrust faults in the Ouachita Mountains frontal belt and the probable translation of significant amounts of displacement from one splay to another makes the identification of a single, lengthy, continuous fault difficult, at best, and misleading, at worst.
65.2	0.7	Cross West Fork Rock Creek.
65.3	0.1	Dirt road to left (east) and Stop 13. Note: The route follows the old St. Louis–San Francisco railroad grade and is passable for most passenger cars and vans. It is not passable for buses.
0.0	0.0	Start mileage to Stop 13. Turn left (east) on dirt road.
0.1	0.1	Turn right (south) on St. Louis–San Francisco railroad grade.
2.3	2.2	Stop 13. Large cut on both sides of railroad grade.

STOP 13

Johns Valley Shale at Compton Cut

Location: NW¼NW¼ sec. 20, T. 4 N., R. 22 E., Le Flore County

Note: The location of this outcrop (sec. 18) as given by Harlton (1934, p. 1039) is incorrect.

This exposure of the Johns Valley Shale at Compton Cut (Fig. 53) probably has been visited by most, if not all, of the early geologists who have worked in the Oklahoma Ouachita Mountains. The strata exposed here have fueled arguments about the age and origin of the Johns Valley.

Perhaps the first published report on the age of the fauna in the Johns Valley at this locality is that of Girty (1909, locality 2057, p. 73). He considered these strata to be the Mississippian Caney Shale and recognized that this age assignment required a high sedimentation rate in the Ouachita area:

> If, then, the Caney correlates with the Moorefield, if the Moorefield is of Mississippian age, if the Caney rests on the Jackfork sandstone [and the Jackfork on the Stanley], and if the evidence furnished by the Standley [*sic*] flora halts between upper Mississippian and Pottsville, the conclusion that the Standley is upper Mississippian and that the sedimentation of that period was peculiar and excessive in the Ouachita region seems to involve the fewest inconsistencies. This conclusion is tentative and unsatisfactory, but seems to accord best with the evidence known to me. Future discoveries may prove it wrong. (Girty, 1909, p. 10)

Girty (1909, p. 11) seemed to hope that the Stanley eventually would prove to be Early Pennsylvanian (Pottsville); this would have allowed him to date the Ouachita Caney (Johns Valley) as Early Pennsylvanian. Girty (1909) did not consider the possibility that the fossils were transported and that they did not represent the age of deposition of the enclosing sediment. Later work showed that Girty's suspicions were correct: sedimentation rates in the Ouachita area were extremely high in the Early Pennsylvanian and the "Ouachita Caney" was, in fact, Early Pennsylvanian (but for a reason different from that suggested by Girty).

Ulrich (1927) was the first to recognize the allochthonous nature of Girty's (1909) Mississippian fauna and established that some of the olistoliths in the Johns Valley (including some at Compton Cut) were Early Pennsylvanian. Ulrich (1927) reviewed the history of the name "Caney Shale" and suggested that the olistolith-bearing shale overlying the Jackfork Sandstone and underlying the Atoka Formation be renamed the Johns Valley Shale because (1) the little-studied "Ouachita Caney" (Johns Valley) was Early Pennsylvanian, whereas many published references referred to the Caney as Mississippian; and (2) the "Ouachita Caney" was named by Joseph Taff (1902a) after a small, but "officially" unnamed settlement on the southeast flank of the Tuskahoma syncline, the center of which contained the type section. The settlement was *locally* referred to as Caney, but the name never appeared on any published maps, and the settlement was later renamed Johns Valley.

Ulrich (1927) also commented on the striated olistoliths in the Johns Valley and published photographs of several "scored erratic boulders" from Compton Cut. He maintained that the olistoliths were derived from the bottoms of floating icebergs and suggested that the striations originated as ice-bound boulders ground against boulders that had melted out of previous icebergs. Earlier geologists gave support, albeit tentative, to the ice-rafting hypothesis. Taff (1910, p. 702) was the first to suggest that the striations were caused "by the agencies of ice," but admitted that the evidence was "not conclusive." Woodworth (1912, p. 459) saw no evidence for scratching by ice at Compton Cut and suggested that the striations were caused by movement of the shale and boulders after they were deposited "under conditions of great pressure, as in deep faulting with slickensides." However, Woodworth (1912) seemed to have no explanation for the striated boulders in the little-deformed outcrops of the Johns Valley (e.g., Tuskahoma syncline) other than by ice. In contrast, Van der Gracht (1931) concluded that all the boulders in the Johns Valley are tectonic in origin.

Moore (1934) also used relations observed at Compton Cut to suggest that (1) the Johns Valley is Morrowan or younger and (2) the striations on the olistoliths formed in response to tectonic deformation and not ice-rafting. However, Moore (1934, p. 443) carefully distinguished between the origin of the striations and the emplacement of the boulders in the shale:

> The scratches and slickensides on hard rock boulders and the locally contorted structure in shale, as seen in such exposures as that at Compton Cut, are

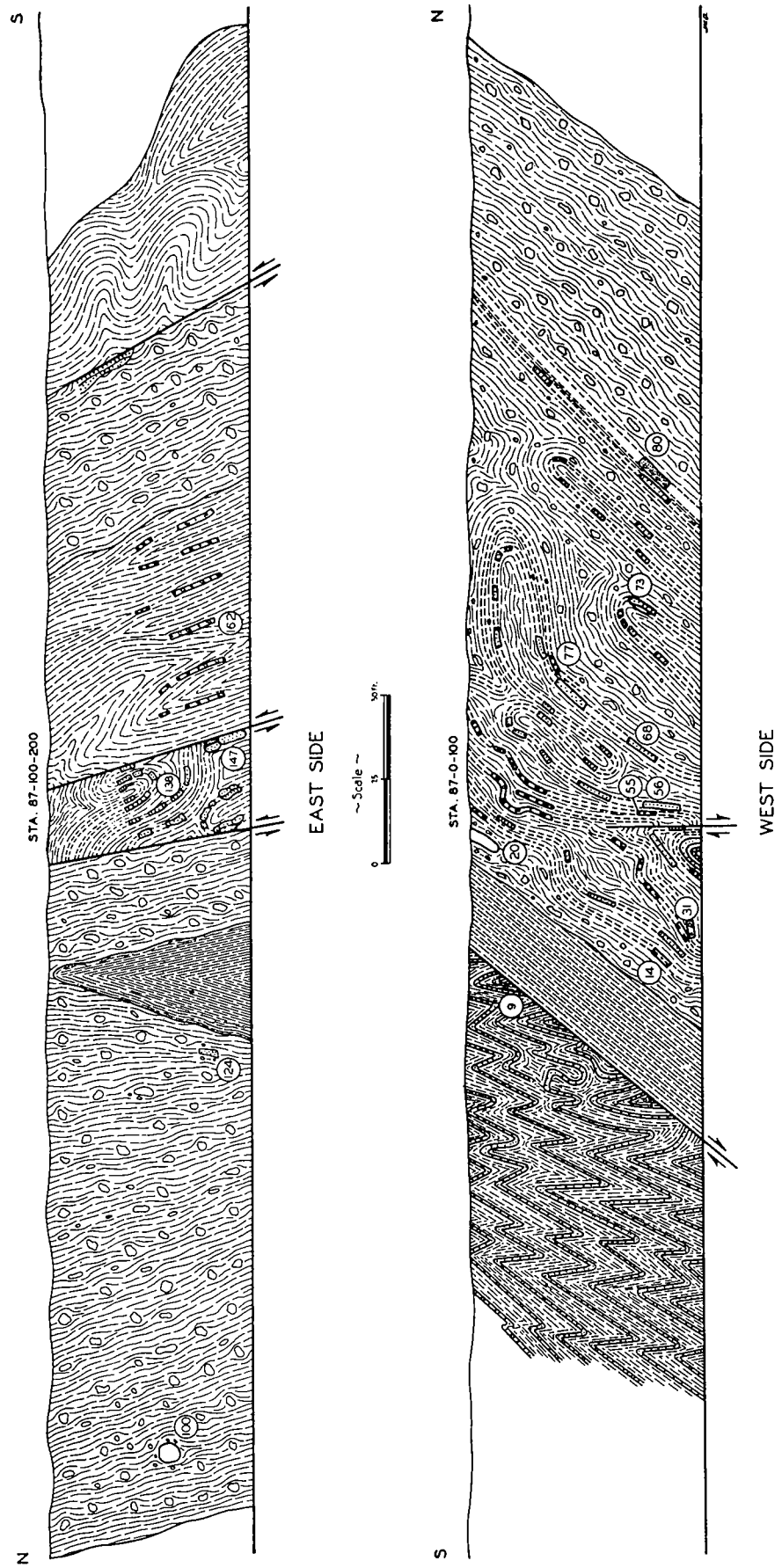


Figure 53. Johns Valley Shale at Stop 13: "Detailed geologic cross section of Compton Cut. Incompetent beds show evidence of flowage" (from Harlton, 1934, fig. 4).

generally accepted as due to folding, but this is regarded as an altogether local matter that is probably quite unrelated to conditions of forming the boulder-bearing shale.

To explain the presence of foreland-facies olistoliths in the Johns Valley Shale, Moore (1934), like many geologists before him, appealed to an ancient highland as a source terrane (see discussion of "Bengalia" at Mile 63.0, Day One, p. 73). This emergent highland, which he called the Ancestral Ouachita Mountains, was located in nearly the same place as Kramer's (1933) "Bengalia"; it differed from Bengalia, however, in that it formed at the same time as deposition of the Johns Valley sediments. The Ancestral Ouachita Mountains were not a long-lived feature, and Moore (1934, fig. 4) did not use them to separate areas of foreland-facies and Ouachita-facies deposition. Moore's (1934, fig. 4) model did require a pre-Atoka unconformity and a 12,000-ft (minimum) fault scarp, neither of which has been observed.

Moore (1934) contributed two important pieces of information to studies of the origin of the Johns Valley. He accepted significant amounts of thrusting (50–70 mi), a range remarkably close to that recently suggested by Arbenz (1989). In addition, Moore (1934, p. 448) recognized that the boulders in the Johns Valley could have been transported a great distance under an extremely low gradient:

> However separated from their mooring, when the "boulders" slid into the oozy mud of the basin it is probable that they found conditions very favorable for lateral creep. Bailey, Collet and Field...have called attention to examples of large exotic rock masses derived from fault cliffs, some of them probably by submarine landslips, and they note evidence of the extremely small gradient necessary to permit considerable lateral movement of the masses embedded in soft muds. We do not know from actual observation how far large boulders resting on or partly embedded in soft mud may be transported through agency of mud flow or creep, nor what minimum gradient is necessary.

Harlton (1933) identified a single species each of a Pennsylvanian foraminifera and ostracod from the Compton Cut outcrop. In 1934, Harlton (1934, fig. 4) published a detailed cross section (reproduced here as Fig. 53) of the strata exposed at Compton Cut. Except for the single(?) south-dipping reverse fault, no information about structural history can be deduced from his figure. Harlton (1934, p. 1039) summarized the structure of this outcrop as follows:

> The excellent exposures at the Compton railway cut... are so intricately disturbed by folding and faulting that a complicated process of unravelling the structural pattern is needed to restore the picture of the original sedimentary sequence. The thickness is almost impossible to estimate, but might approximate 500 feet.

Harlton (1934, p. 1039–1045) described in detail the stratigraphy of the Johns Valley Shale at Compton Cut and the rock types present. His description is too lengthy to review in detail here, but several features of the Johns Valley that he noted are significant: (1) Many of the sandstone beds interbedded in the more abundant shale are dismembered and resemble olistoliths. Limestone beds also are interbedded in the Johns Valley (Harlton, 1934, p. 1939); in fact, these limestone "beds" are olistoliths and illustrate the difficulty in distinguishing between "indigenous" and "exotic" blocks in the shale. (2) Many of the boulders in the Johns Valley are surrounded by "fringes" of pebbles ~1 ft thick. Harlton (1934,) had no explanation for the "fringes." (Fellows [1964, p. 47–48] did not observe these "fringes.") (3) Some of the olistoliths are conglomerates containing pebbles of a wide variety of rock types and ages. Harlton (1934) recognized the significance of the conglomerate boulders, but few recent studies have addressed the implications of more than one period of erosion and deposition of these coarse-grained sediments.

> Of great importance is the fact that the conglomerate boulder represents two periods of generation; first, the deposition of the conglomerate itself and, later, the time of being torn off its exposure and carried away in the late Johns Valley depositing waters. (Harlton, 1934, p. 1044)

(4) Harlton (1934, p. 1041) also noted the presence of gilsonite (grahamite?) in the Compton Cut exposure.

The latest detailed work on Compton Cut is that of Shideler (1968), who identified chert and Fite (Ordovician), Jasper (Ordovician), Viola (Ordovician), and Wapanucka (Pennsylvanian) limestone olistoliths in the Johns Valley here.

In summary, this exposure of the Johns Valley Shale at Compton Cut is one of several that have fueled many arguments concerning Ouachita tectonics, stratigraphy, and sedimentation. Many early ideas proved incorrect as new evidence was uncovered and new geological concepts developed. However, some ideas proposed by early geologists have gained acceptance. Among them are: (1) Mississippian and Early Pennsylvanian sedimentation rates in the Ouachita Mountains area were extremely high (Girty, 1909); (2) the Johns Valley Shale contains Morrowan olistoliths and itself is Morrowan (Ulrich, 1927); (3) displacement on thrust faults is on the order of many tens of miles (Moore, 1934); (4) the transport mechanism for the boulders does not require a steep gradient (Moore, 1934); and (5) locally, the Johns Valley is so completely folded and faulted that exotic olistoliths are difficult to distinguish from dismembered indigenous beds (Harlton, 1934). Two ideas have been less successful: (1) ice-rafting, and (2) Bengalia or the Ancestral Ouachita Mountains. Other ideas have not been investigated: (1) could the striations on the boulders have originated as they were being transported, and not during deformation after they were emplaced in the shale? (2) what are the implications of a multi-cycle origin of the conglomerate boulders in the Johns Valley Formation (see description of Johns Valley Shale near Stapp, Stop 19).

		Retrace route to S.H. 82.
4.6	2.3	S.H. 82. Turn left (south). Reset mileage to 65.3.

66.0 0.7 Stop 14. Large unvegetated outcrop on uphill (south) side of highway.

STOP 14

Johns Valley Shale

Location: NW¼SE¼ sec. 13, T. 4 N., R. 21 E., Le Flore County

Sutherland (1979, p. 62) briefly described the Johns Valley Shale at this stop:

> The upper Johns Valley Shale seen at this locality... differs from the exposure at Stapp [see Stop 19]...in that the erratic limestone and sandstone boulders are much smaller in size and the shale shows no distortion either by sediment transport or by subsequent structural movement. One can see the depositional relationship of the erratics to the shale. At this locality the shale is very evenly bedded, and the small erratics appear as suspended "plums in a pudding." There are numerous small ironstone concretions.

Fellows (1964) estimated the thickness of the Johns Valley in this region to be 500–700 ft. About 70 ft is exposed in this road cut, and these strata are apparently in the upper part of the formation.

Most of the Johns Valley Shale at this outcrop consists of widely separated limestone, siliceous-shale (Woodford?), and black-shale (Caney?) olistoliths supported in a shale matrix. Most of the limestone olistoliths are subrounded, whereas the shale occurs as wisps, streaks, equidimensional blocks, and slabs (as large as 2 ft thick and 30 ft long). The limestone blocks were well indurated when they were incorporated into the Johns Valley; in contrast, the shale olistoliths appear to have varied from poorly to well consolidated. In addition to the olistoliths, the Johns Valley at this location contains: (1) lenses and discontinuous beds of sandstone; (2) ellipsoidal limonite-stained concretions as large as 2.5 ft in diameter; (3) a discontinuous (5-ft-long), 1-ft-thick pebble-conglomerate bed (angular, pebble-sized limestone clasts are supported in a matrix of clean quartz sandstone); (4) relatively continuous sandstone beds; and (5) disharmonic folds.

Of particular interest at this outcrop is a 13-ft-long, 3-ft-thick lens or "pod" of a matrix-supported boulder conglomerate in the Johns Valley Shale (Fig. 54). Most of the clasts are limestone and range from pebble-size to ~2 ft in diameter; some Woodford(?) shale also is present. The matrix consists of silty to sandy shale that locally appears highly fractured. Anastomosing shear zones extend from the base to the top of the lens. The base of the lens also appears sheared. Complex small-scale folds are present in the enclosing shale at the lower end of the lens.

My (NHS) interpretation of the lens is that it is a large fragment of a clast-rich debris flow that subsequently was incorporated into a matrix-rich debris flow (represented by most of this outcrop). The through-going shears are tectonic and are not related to the deposition of the Johns Valley Shale. The small folds at the "tail" of the lens may be depositional, a compaction shadow around the relatively competent (compared to the enclosing shale) lens, or the folds may be related to the later tectonic shearing.

About 0.1 mi north of this outcrop there is a new road cut, already highly vegetated, that exposes the Johns Valley Shale. That outcrop contains: a 3-ft-diameter limestone-breccia block; a 6-ft-long sandy-limestone slab; and abundant phosphatic nodules.

66.9 0.9 South-dipping (35°), south-facing strata of lowermost Atoka Formation on left (east) side of road.

67.0 0.1 Cross axis of "Gobblers Knob syncline." Gobblers Knob is underlain by relatively resistant sandstone beds in the lower part of the Atoka Formation. The syncline does not appear to extend more than 2 mi east–west and probably is surrounded on all sides by thrust faults.

67.1 0.1 North-dipping (60°), north-facing strata of lower part of Atoka Formation on south flank of syncline.

67.5 0.4 Massive sandstone at base of Atoka Formation on left (east) side of road. Sutherland (1979, p. 61–62) described the Johns Valley/Atoka contact at this locality and noted the following general stratigraphy:



Figure 54. "Pod" of matrix-supported boulder conglomerate in the Johns Valley Shale at Stop 14. The "pod" is probably a large clast (olistolith) in a shale-rich debris-flow deposit.

Atoka Formation:

Mostly sandstone (90%), consists of about 10 modified classic turbidite flows. Turbidites have well-developed bottom-transport features indicating paleo-flow direction varying from S. 5° W. to S. 72° W. (Suneson [unpublished field notes] measured paleo-flow directions ranging from S. 20° W. to S. 45° W.) Possibly represents submarine-fan deposit	25 ft
Massive, structureless, fine-grained sandstone	25 ft
Mostly shale with classic turbidite layers; excellent sole marks	70 ft
Johns Valley Shale with limestone olistoliths as much much as 0.7 ft in diameter	20 ft

Sutherland (1979) noted that the sandstone beds show little or no thickness variation across the outcrop.

Poole (1985, p. 49) also measured the strata exposed here. She did not note the Johns Valley Shale (her Facies F) at the base of the section.

Atoka Formation:

Fissile shale, thin-bedded siltstones, thin- and thick-bedded sandstones	50 ft
Thick-bedded (2–3 ft thick) sandstones with thin shale partings	20 ft
Massive, thick-bedded sandstone with 3-ft-deep channel at base	30 ft
Fissile shale, thin-bedded siltstones, thin sandstones with abundant trace fossils and scour marks	90 ft

Poole (1985, p. 52) suggested that the basal sandstones of the Atoka Formation (her Lynn Mountain Formation) probably represent mid-fan environments and possibly the "transition from deposition in shallow mid-fan channels to unconfined sand lobes."

67.9	0.4	Top of Winding Stair Mountain, elevation ~1,400 ft. For the next 1.4 mi, the field route descends the south side of Winding Stair Mountain and ascends stratigraphically through ~4,000 ft of the Atoka Formation. The strata dip ~45° south and face south. At the southern base of Winding Stair Mountain ~2.5 mi west-southwest of here, H & H Star Energy recently drilled the No. 1-28 Lady Luck (SE¼NW¼SE¼ sec. 28, T. 4 N., R. 21 E.). The well spudded on December 30, 1991, and reached 9,976 ft total depth in May, 1992. No log tops were reported. The well tested 4.6 MMcf gas per day from the Atoka at 8,946 ft to 9,032 ft.
69.3	1.4	Cross trace of Windingstair fault.

WINDINGSTAIR FAULT

The Windingstair fault is the southern boundary of the Ouachita Mountains frontal belt, and the general structure changes south of the fault. North of the fault, Lower and Middle Pennsylvanian strata are tightly folded and imbricately thrust-faulted; south of the fault, broad, open synclines are separated by relatively tight, thrust-cored anticlines typical of central-belt structure. Mississippian Stanley Group strata in this part of the mountains are restricted to the hanging wall of the Windingstair fault in contrast to the western frontal belt, where Stanley Group and Ordovician strata are exposed in the footwall of the fault. To the east, Hart (1963) and Seely (1963) also mapped Stanley Group strata in the footwall of the Windingstair fault, but in the hanging wall of the Briery fault.

If the juxtaposition of the Stanley over younger strata indicates major displacement, such displacement occurred on the Windingstar fault in the central part of the Oklahoma Ouachita Mountains. To the west (near the town of Atoka), the Stanley is juxtaposed against younger strata along other faults, suggesting that major displacement was on the Choctaw, Pine Mountain, and Ti Valley faults as mapped by Hendricks and others (1947). To the east, the Briery fault appears to be the locus of major displacement.

69.4	0.1	Entrance (on right) to Talihina city dump. The trace of the Windingstair fault as exposed in the dump is not a single fault plane; rather, it is distributive (Cline, 1968, p. 22). The strata in the dump are extremely disturbed due to natural (thrust faulting) and man-made (bulldozing) processes.
70.3	0.9	Turn right (west) on gravel road toward Lake Carl Albert.
70.5	0.2	Bear left to go over dam.
70.7	0.2	Cross Lake Carl Albert dam.
71.0	0.3	Stop 15. Large outcrop on east side of spillway of Lake Carl Albert dam. Walk up the low hill on the east side of the spillway and descend into the spillway near the east abutment of the low dam. Be careful—much of the floor of the spillway typically is muddy.

STOP 15

Stanley Group Strata

Location: C N½NW¼ sec. 2, T. 3 N., R. 21 E., Latimer County

INTRODUCTION

This outcrop is one of the best exposures of Stanley Group strata in this part of the Ouachita Mountains (Fig. 55). Our reasons for stopping here are (1) to compare the Stanley sandstones and shales at this outcrop with those of the Atoka Formation (see Stop 11 and Stop 20) and with those of the Jackfork Group (see Stop 22) and (2) to investigate whether or not these turbidite-appearing beds are, in fact, turbidites.

Many features of the Stanley strata at this outcrop are similar to those observed in the turbidite strata of the Atoka Formation and Jackfork Group, including sole marks, dewatering structures, and convolute stratification (Fig. 56; Measured Section, Stop 15). Other features, however, are distinctive of the Stanley: (1) many of the sandstone beds are graded and contain matrix-supported, subrounded, coarse-sand-size grains in a muddy, fine-sand-size matrix at the base; (2) shale rip-up clasts are common locally, and typically are concentrated in bedding-parallel zones within sandstone beds; (3) trace fossils are rare; (4) some beds are cemented with calcite, and pyrite occurs locally in trace amounts; (5) shales typically contain discoidal siltstone masses and show pencil structure; (6) cone-in-cone structures and sandstone and siltstone dikes characterize some strata.

Many of the sandstone beds in this outcrop of the Stanley Group do not have typical Bouma-sequence structures (Fig. 46). It is possible that some may not have been deposited by turbidity currents, but by mass-flow mechanisms. Are there any diagnostic features that absolutely distinguish turbidite sandstones from "mass-flow" sandstones?

HISTORY

Age

The "Standley" shale was named by Taff (1902a) for exposures east of the Choctaw fault, east and northeast of the town of Atoka. He believed the formation was Silurian and described it as follows:

> Siliceous chert, black shale, and greenish clay shales occur in alternate members for nearly 800 feet above the base. There are two cherty members, 230 and 40 feet in thickness...above the base of the formation.... Continuing upward in the section, there are greenish and dark shales alternating with drab or brown and moderately hard sandstones until the formation reaches an estimated thickness of 6000 feet. Distinct sandstone members range in thickness from 20 to 100 feet and are separated by shales or shales interstratified with sandy beds 140 to 2000 feet in thickness....The sandstones are so soft that they produce little more effect upon the surface configuration than do the clay shales in which they are embedded. The whole formation invariably crops in nearly level plains or valleys and in such steep slopes at the borders of these valleys as are protected by the harder overlying formation....The formation takes its name from the village of Standley, in the Kiamitia Valley, where it is extensively exposed. (Taff, 1902a, p. 4)

Purdue (1909) changed the spelling slightly (because the name of the village had changed) and applied the name Stanley to the same strata in Arkansas, which he considered to be Carboniferous.

After these initial papers on the Stanley, attention was focused on the question of the age of the unit. Some of the controversy was based on the age of the overlying Jackfork Sandstone and the age and origin of the Johns Valley Shale (see Stop 13). Miser (1921) stated that the Stanley sediments were deposited in the Ouachita geosyncline during the Mississippian. Honess (1923) suggested a Chesterian age based on floral evidence and an Early Pennsylvanian age based on invertebrates. The 1926 geologic map of Oklahoma showed the Stanley as Mississippian (Miser, 1926). Miser and Honess (1927) and Miser and Purdue (1929) maintained that the Stanley was Mississippian. In 1934, Miser suggested that the age was earliest Pennsylvanian, based on the work of White (1934), who studied the flora of the upper part of the Stanley.

Harlton (1934,1938,1959) greatly confused the question of the age of the entire pre-Atokan, post-Devonian sequence by attempting to introduce new system ("Bendian"), series ("Pushmataha"), and formation (e.g., "Barnett Hill") names (Fig. 57). Fortunately, his ideas were not adopted. Harlton (1938), however, raised the Stanley to group status and established three formations within it: (from oldest to youngest) Ten Mile Creek, Moyers, and Chickasaw Creek. This division has been used by many subsequent workers in the Ouachita Mountains.

Cooper (1945) reviewed previous age assignments of the Stanley and concluded, based on his own study of the flora, that the Stanley was pre-Pottsville (Morrowan) and post-Chesterian. Hendricks and others (1947) believed the Stanley was Pennsylvanian based on White (1934,1937) and Harlton (1934,1938). Based on his study of conodonts, Hass (1950,1956) stated that the age of the lower part of the Stanley was Meramecian and that the upper part could be Early Pennsylvanian. The 1954 geologic map of Oklahoma (Miser, 1954) showed the Stanley to be Mississippian and Pennsylvanian. Cline and Shelburne (1959) stated that the Stanley was Meramecian, based partly on their interpretation that the base of the Johns Valley contained autochthonous Mississippian Caney Shale. Branson (1959) agreed with this interpretation and Cline (1960) elaborated upon it.

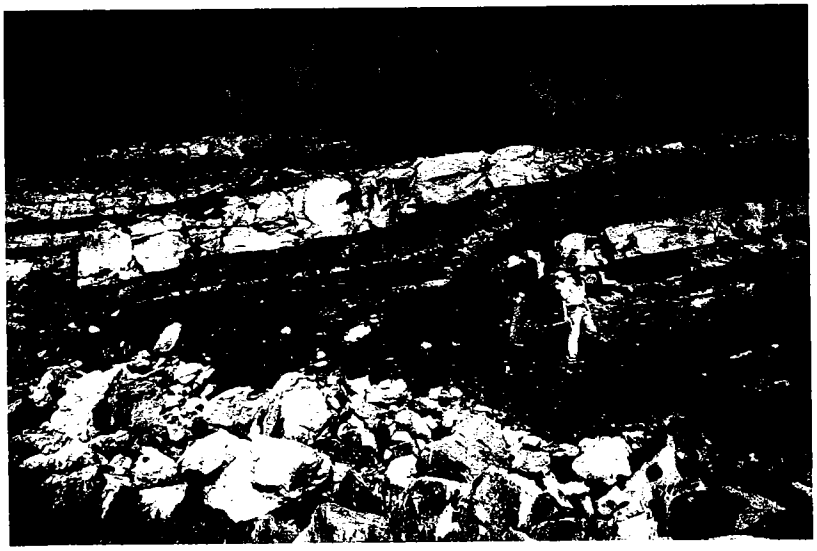


Figure 55. Sandstones and shales of the Stanley Group (probably Moyers Formation) exposed in the spillway of Lake Carl Albert dam at Stop 15.

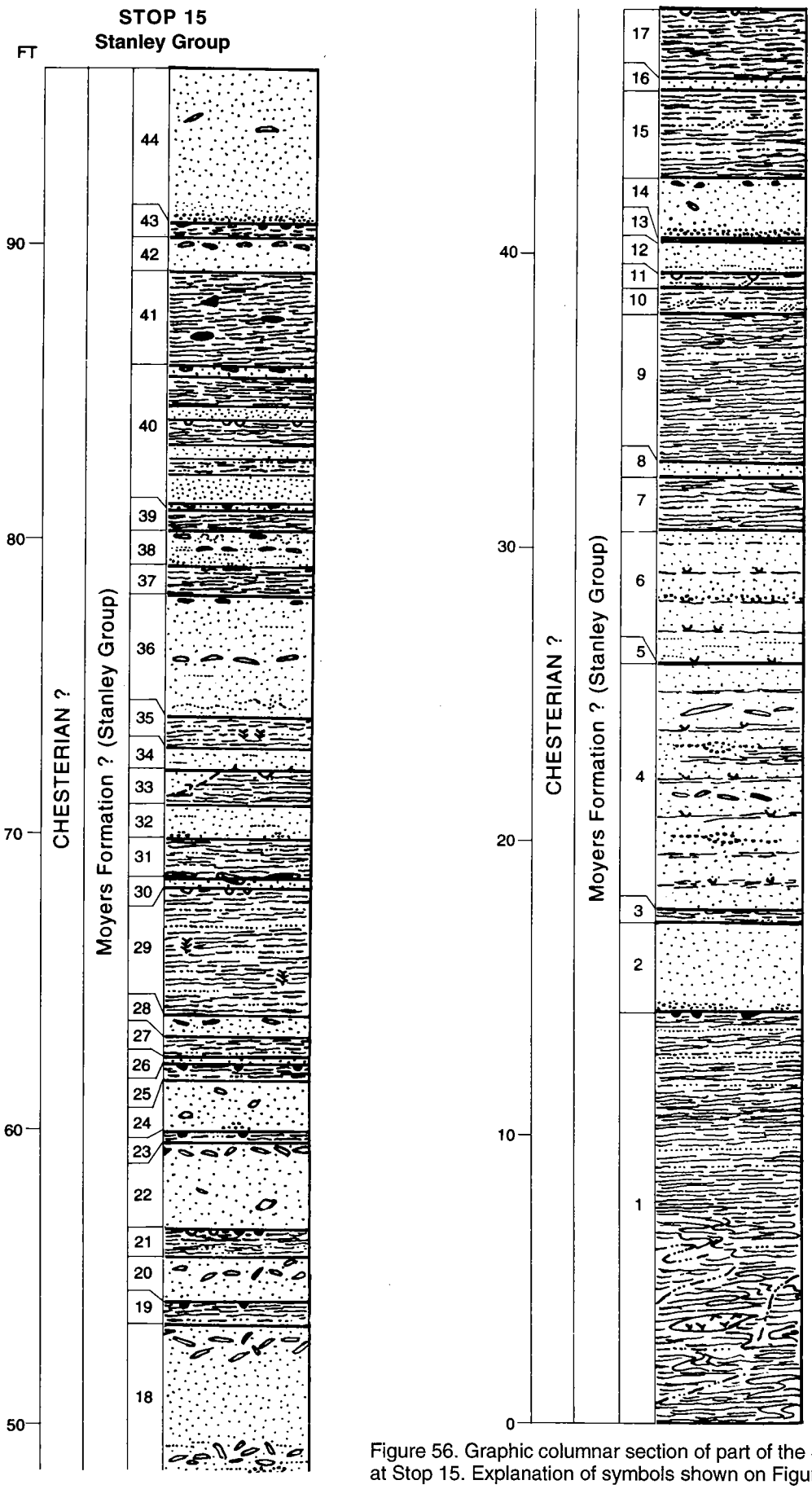


Figure 56. Graphic columnar section of part of the Stanley Group at Stop 15. Explanation of symbols shown on Figure 51A.

Measured Section, Stop 15

Stanley Group

C N½NW¼ sec. 2, T. 3 N., R. 21 E., Latimer County. Outcrop on east side of spillway of Lake Carl Albert dam.

Thickness (feet)

44. Sandstone. Unstratified, shows slight normal grading. Base is sharp, contains abundant scour and drag marks, but locally is very planar. Mostly medium to fine grained, poorly sorted, with subrounded grains in a mud matrix. Contains rare shale rip-up clasts as much as 6 in. long. Trace calcite cement. Base weathers spheroidally 5.3
43. Shale. Very well laminated with well-developed pencil structure 0.5
42. Sandstone. Medium gray (N5). Unstratified. Top and base are sharp. Mostly medium to fine grained, poorly sorted, with subrounded grains in a mud matrix. Shale rip-up clasts as much as 4 in. long concentrated at top of bed 1.2
41. Shale. Fissile, very well laminated with well-developed pencil structure. Cleavage at high angle to bedding planes not present in adjacent sandstone beds. Contains rare 0.5–1-ft-long, blocky to discoidal siltstone masses 3.2
40. Sandstone, shale and siltstone. Includes four sandstone beds (from lowest to highest, Nos. 1, 2, 3, 4). All sandstones medium- to fine-grained, poorly sorted, with abundant mud in matrix. "Rust" spots common, possibly weathered pyrite. Trace of calcite cement throughout. Sandstone No. 1: Greenish gray (5GY6/1) to medium gray (N5). Varies from 1 ft to 1.7 ft thick, may be amalgamated with 1-ft-thick, unstratified, upper bed with planar base and 0.7-ft-thick, poorly stratified, blocky weathering lower bed with shale rip-up clasts concentrated at top. Both have streaks of organic material throughout. Sandstone No. 2: Unstratified, has planar base. Sandstone No. 3: Unstratified, load structures common at base of bed. Sandstone No. 4: Varies from 0.3 ft to 0.6 ft thick, unstratified. Base is planar, top is irregular—appears to be cross-stratified sandstone and shale at high angle to base. Contains very rare 0.25-in. chert fragments throughout and common shale rip-up clasts at top. Locally contains small pieces of plant debris 5.0
39. Shale. Fissile, very well laminated with well-developed pencil structure 0.7
38. Sandstone. Light gray (N7). Poorly parallel stratified, top 0.5 ft shows disrupted bedding, probably dewatering structures. Base and top are sharp. Medium- to fine-grained, poorly sorted, subrounded grains in matrix with abundant mud. Shale rip-up clasts common, most 0.25–2 in. long, but can be as long as 2 ft, concentrated in middle and at top of bed. Calcite cement 1.3
37. Shale. Fissile, poorly developed pencil structure 1.0
36. Sandstone. Light gray (N7). Unstratified to poorly parallel stratified, except for basal 1 ft, which is faintly wavy bedded to locally contorted. Base is sharp, top grades into overlying shale. Medium- to fine-grained, poorly sorted, abundant mud in matrix. Basal 1 ft contains sandstone rip-up clasts; shale rip-up clasts as long as 8 in. concentrated near middle of bed and smaller clasts at top. Calcite cement. Open fracture containing calcite locally coats face of bed 4.2
35. Shale and siltstone. Siltstone beds 0.25–1 in. thick, locally show cone-in-cone structure 1.0
34. Sandstone. Greenish gray (5GY6/1). Possible 1-ft-deep channel at dam abutment. Poorly parallel stratified to wavy-bedded at base. Base and top sharp, poorly developed load structures and well-developed shale flame structures at base. Medium- to fine-grained; subrounded grains are mostly translucent quartz and possibly cryptocrystalline quartz and chert. Calcite cement and in open fractures on face of outcrop. Trace organic debris throughout 0.3–1.0
33. Shale and siltstone 1.2
32. Sandstone. Medium light gray (N6). Poorly parallel stratified. Base sharp, top gradational to overlying shale/siltstone. Base is slightly wavy, possibly caused by incipient loading. Medium- to fine-grained, poorly sorted, subrounded quartz and chert(?) grains in matrix with abundant mud. Calcite cement. Trace organic debris throughout. Attitude: N. 80° E. 13° N. 1.2–1.6
31. Siltstone and shale. Well-stratified. Shale shows pencil structure, siltstone does not. Contains discoidal masses of shale as long as 3 ft at base. One mass folded, one oblique to basal contact and largely embedded in underlying sandstone, one parallel to contact and immediately over sandstone. Shale in discoidal masses has cone-in-cone structure 1.3
30. Sandstone. Unstratified, normally graded from medium- to fine-grained. Base sharp, top gradational to overlying siltstone/shale. Base with poorly developed load structures. Poorly sorted, rounded grains of quartz, chert, and rock fragments in matrix with abundant mud. Shale rip-up clasts in top 0.3 ft. Calcite cement 0–0.7
29. Siltstone and shale. Well-stratified to fissile. Siltstone: Locally has rusty/yellow stain. Grades to fine-grained sandstone with small pieces of organic debris, including plant stems. Locally weathers to 2–3 in. spheroidal masses. Small areas with cone-in-cone structure throughout 4.3
28. Sandstone. Medium gray (N5). Unstratified. Base sharp and extremely planar, top gradational into overlying siltstone/shale. Fine-grained with trace of medium-sized, subrounded quartz grains. Shale rip-up clasts as long as 3 in. concentrated in upper 0.3 ft of bed. Trace calcite cement. Rust spots throughout and along fractures 0.5–1.0
27. Siltstone and shale. Dark gray (N3). Stratified. Siltstone slightly sandy. Top 0.2 ft is shale. Shows fine pencil structure 0.8
26. Sandstone. Medium dark gray (N4). Poorly parallel stratified. Base sharp and mostly planar, top grades into overlying siltstone/shale. Base locally slightly wavy, possibly caused by incipient loading. Where sandstone is thickest, base has grooves that trend about N. 50° E., suggests paleocurrent azimuth N. 50° E.-S. 50° W. Fine-grained, poorly sorted, conspicuous mica parallel to bedding planes, abundant mud in matrix. Trace calcite cement. Plant stems as long as 0.5 in. concentrated on bedding planes 0.1–0.6
25. Siltstone and shale. Stratified. Siltstone grades up into shale 0.8
24. Sandstone. Medium light gray (N6). Unstratified, poorly normally graded with slight concentration of medium-sized grains at base. Base and top are sharp. Base has grooves that trend N. 80° E., suggests paleocurrent azimuth N. 80° E.-S. 80° W. Medium- to fine-grained, poorly sorted, subrounded grains in matrix with abundant mud. Rare shale rip-up clasts as long as 2 in. throughout. Trace calcite cement. Trace small disseminated pyrite and

plant debris throughout. Fracture parallel to bedding cuts bed in half 1.7

23. Siltstone and shale 0.4
22. Sandstone. Medium dark gray (N4). Unstratified. Base sharp, top moderately gradational to overlying siltstone/shale. Base generally planar, locally with load casts and large and small grooves. Large grooves locally give base a rippled appearance. Azimuth of grooves varies slightly—averages N. 85° E., suggests paleocurrent N. 85° E.-S. 85° W. Medium-grained with rare, slightly larger subrounded grains. Mica is conspicuous . Shale rip-up clasts as long as 10 in. concentrated in upper 0.5 ft of bed; rare, smaller clasts throughout. Long axis of clasts subparallel to bedding. Rare, small plant stems parallel to bedding. Mineralized fault plane with N. 25° W. 68° SW. attitude contains slickensides that rake 23° NW., suggests oblique-slip movement 2.9
21. Siltstone and shale. Siltstone locally weathers to 2-in.-diameter spheroidal masses or 1-in.-long angular blocks. Upper 0.3 ft is very fissile shale with well-developed pencil structure 1.0
20. Sandstone. Medium gray (N5) to medium bluish gray (5B5/1). Base sharp, top gradational to overlying siltstone/shale. Base locally with small grooves that trend N. 50° E., suggests paleocurrent azimuth N. 50° E.-S. 50° W. Fine-grained with trace rounded, medium-sized grains—mostly quartz, possibly 20% rock fragments. Poorly sorted, abundant mud in matrix. Shale rip-up clasts as long as 10 in., typically 2 in., concentrated in upper one-half of bed. Trace calcite cement. Joint surface has subhorizontal plumose markings 1.5
19. Shale and siltstone. Shale dark gray (N3). Siltstone 0.3 ft thick, stratified, grades up into shale, shows poor spheroidal weathering. Shale 0.5 ft thick, fissile, shows well-developed pencil structure. Silica cement 0.8
18. Sandstone. Light gray (N7). Mostly unstratified, except for basal 0.8 ft which locally shows faint, parallel laminations. Poorly normally graded with rare, medium-sized, subrounded quartz grains concentrated at base. Base is planar, undulates slightly, has rare, small, load casts. Top grades into overlying shale/siltstone. Fine-grained, poorly sorted, abundant mud in matrix. Shale rip-up clasts as long as 2 ft, concentrated in layers from 0.3 ft to 1.8 ft above base and in upper 1.5 ft of bed. Top of bed is contorted, shows complex mixing of sandstone and shale rip-up clasts. Calcite cement. Face of outcrop shows areas ~5 in. in diameter of brown staining 5.2
17. Shale and siltstone. Shale fissile, slightly silty, shows well-developed pencil structure. Siltstone, slightly sandy, grades up and down into shale. Both rock types locally coated with white clay 2.3
16. Sandstone. Olive black (5Y2/1). Unstratified. Base sharp, top grades into overlying shale/siltstone. Very fine grained. Silica cement. Bottom part of bed weathers spheroidally. Trace organic debris, mostly plant stems, throughout 0.1–0.4
15. Shale and siltstone. Siltstone: Minimum six individual beds, 0.1–0.2 ft thick. Parallel- and cross-laminated. Cross-laminations suggest paleocurrent azimuth about east-west. Shale: Well-laminated, locally silty, well-developed pencil structure typical. 2.9
14. Sandstone. Unstratified, normally graded with subrounded grains as large as 0.07 in. at base, fining upward. Base sharp, contains "rills" that trend N. 60° W., relation to paleocurrent direction unknown. 1-in.-long shale rip-up clasts concentrated in upper 0.2 ft of bed, trace throughout 2.2
13. Shale. 0.1
12. Sandstone. Medium gray (N5). Unstratified except for parallel-stratified basal 0.3 ft of bed. Base and top sharp, small load casts on base. No inclusions. Calcite cement. Attitude: N. 80° E. 13° N. 1.1
11. Siltstone and shale. Laminated, including very small-scale cross-laminations 0.5
10. Siltstone and shale. Siltstone: Dark gray (N3). Parallel- and cross-laminated. Silica cement 0.8
9. Shale. Grayish black (N2). Fissile, shows well-developed pencil structure. Contains minor, thin, siltstone beds 5.2
8. Sandstone. Dark gray (N3). Unstratified, sharp base and top. Very fine grained. Silica cement. Poorly exposed 0.3–0.5
7. Shale. Fissile, shows well-developed pencil structure. Contains minor, thin siltstone beds 2.0
6. Sandstone and minor shale. Includes four sandstone beds (from lowest to highest, Nos. 1, 2, 3, 4), 0.6–1.2 ft thick, separated by 0.1–0.2-ft-thick shale beds. All sandstone beds have sharp bases with small load casts. Sandstone No. 1: Poorly parallel stratified, all others are unstratified. Sandstone No. 3: Slightly graded, also contains 0.1-ft-thick brownish band ~0.3 ft from base of bed. Sandstone No. 4: Contains 0.3-ft-thick siltstone bed near top, also possible shallow channel in sandstone below siltstone. Channel may be low-angle fracture 4.6
5. Shale. 0.1
4. Sandstone, siltstone, and shale. Complexly interbedded. Sandstones: Medium dark gray (N4), mostly 0.7–1.0 ft thick, unstratified to very poorly stratified, locally with dish-and-pillar structures. Sharp bases. Typically fine-grained with rare, medium- to fine-grained, subrounded quartz grains throughout, matrix with abundant mud. Shale rip-up clasts locally common. Trace calcite cement. One sandstone bed contains abundant calcite cement, weathers reddish brown. In places, shale beds appear to pass from coherent, stratiform bodies to large shale rip-up clasts as long as 8 ft. Sandstone locally squeezed into shale rip-up clasts and load casts present at top of some larger clasts. Much pinch and swell of sandstone beds 8.3
3. Silty shale and siltstone. Base gradational 0.4
2. Sandstone. Brownish gray (5YR4/1), unstratified, poorly graded from medium- to fine-grained at base upward to fine-grained. Base sharp and mostly extremely planar, locally with scour-and-fill structures (paleocurrent approximately northeast to southwest and grooves 3 in. deep and 4 in. long oriented N. 60° E.). Poorly sorted, abundant mud in matrix. Grain composition possibly 30% rock fragments. Trace calcite cement. Herringbone fracture pattern on fracture with attitude of N. 45° W. 81° SW. Attitude of bed is N. 85° E. 14° N. 3.1
1. Shale, minor siltstone and sandstone. Shale with well-developed pencil structure. Rare 0.1–0.2-ft-thick siltstone and very fine grained sandstone beds in upper 7 ft. Lower 7 ft: Bedding in shale very disturbed, abundant 0.1–0.2-ft-thick siltstone dikes and sills, some appear folded, others oblique to bedding. One 16-ft-long calcareous shale phacoid with cone-in-cone structure near base 14.0

Total thickness of section 95.4–98.7

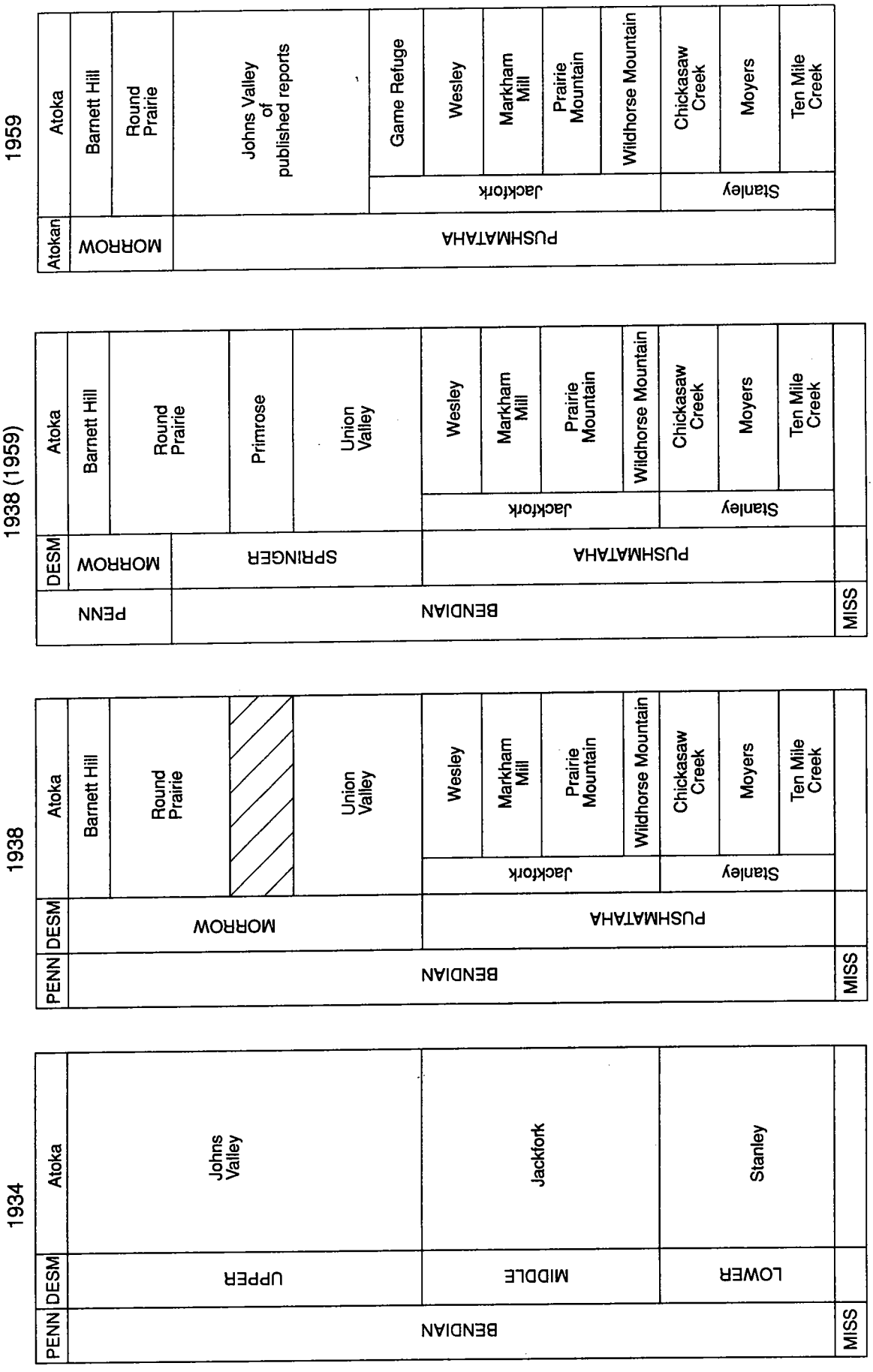


Figure 57. Upper Mississippian–Lower Pennsylvanian stratigraphy of Ouachita Mountains according to Harlton (1934,1938,1959). The column labeled 1938 (1959) is what Harlton (1959) claimed he said in 1938.

In 1977, Gordon and Stone published a paper that clearly established the age of the Stanley Group, as well as that of the overlying Early Pennsylvanian turbidite strata. Upper Chesterian Pitkin Limestone boulders in the Chickasaw Creek Formation proved that the top of the Stanley Group was no older than late Chesterian. The age of the overlying Jackfork Group is Chesterian (based on floral assemblage) to early Morrowan (ammonoids).

Depositional Environment

Miser (1921) was the first geologist to attempt to describe the depositional environment of the Stanley Group. He suggested that the Stanley sediments were derived from an ancient landmass to the south (which he called "Llanoria") and were deposited in the Ouachita geosyncline. The location of the geosyncline is approximately the same as the present location of Stanley outcrops. Miser and Purdue (1929) supported this idea and suggested that the Stanley was deposited as large, freshwater deltas on the southern margin of the geosyncline. Evidence for the relatively shallow-water environment of deposition included ripple marks and locally abundant plant debris in some of the Stanley sediments. Harlton (1934) supported the delta model for the Stanley.

In 1959, Cline and Shelburne proposed a significant, new interpretation of the depositional environment of the Stanley. They maintained that the Stanley, Jackfork, Johns Valley, and Atoka Formations were similar to the black-shale flysch and wildflysch facies (turbidites) of Europe and were deposited in deep water. Van der Gracht (1931) originally used the term "flysch" to describe the Stanley and overlying strata, but did not elaborate on the idea nor clearly state that the sediments were deposited in deep water. Based on sandstone petrography, Cline (1960, p. 100) suggested that the source of the sediments was to the south or southeast. The sediments accumulated in a trough in which turbidity currents, flowing about S. 78° W., were the dominant transport mechanism. Subsequent paleocurrent studies of Cline (1970) and Morris (1974) confirmed the north- to northwest-directed paleocurrents for much of the Stanley in the southern part of the Ouachita Mountains, and west-directed paleocurrents in the northern part of the Stanley outcrop belt. Niem (1976) divided the lower part of the Stanley into a southern/proximal and a northern/distal flysch facies and suggested each accumulated on different parts of a deep-sea fan-complex and adjacent basin plain. He also showed that the more sandstone-rich proximal facies prograded over the distal facies in later Stanley time in the central Oklahoma Ouachitas.

Hatton and Associated Tuffs

Miser (1921, p. 71–72) was the first geologist to comment on tuff beds in the lower part of the Stanley Group:

> Tuffs occur near the base of the Stanley shale in Polk County, Arkansas, and McCurtain County, Oklahoma, in three and possibly four or five beds, which range in thickness from 6 to 85 feet, the lowest bed being the thickest and the most widely distributed....[As] they occur only on the south side of the Ouachita Mountains, and thin out toward the north, and as the size of grains of their component materials apparently decreases towards the north, it would seem that the volcanic materials they contain were ejected from some vent or vents on or near [an] old land area to the south.

Honess (1923, p. 150) doubted that there was more than one tuff bed in McCurtain County, and he suspected that unmapped folds and faults repeated the tuff in some areas. Miser and Purdue (1929, p. 62–63) maintained that there were at least three, and possibly five, tuff beds in the lower part of the Stanley, and they named the lowest and thickest tuff bed the "Hatton tuff lentil." Niem (1976) used differences in lithologic zonation, thickness, and stratigraphic position to identify four main tuff units in the lower part of the Stanley: (from oldest to youngest) Beavers Bend tuff, Hatton tuff, and Lower and Upper Mud Creek tuffs. In addition, Niem (1971) described several thin tuff beds immediately above the Upper Mud Creek tuff, as well as a tuff associated with the Chickasaw Creek Siliceous Shale, the uppermost unit in the Stanley Group.

To date, correlation of the tuff beds in the Stanley, particularly from the Broken Bow uplift to near the Windingstair fault, has proved difficult. Pitt and others (1982, p. 20) identified an 18-ft-thick tuff bed in the lower part of the Stanley on the south side of the Potato Hills. Loomis (1992, p. 159) identified at least two, and possibly three, tuff beds in the same area and correlated them with the Beavers Bend, Hatton, and Mud Creek tuffs. Based on major-, trace-, and rare-earth-element abundances, Loomis (1992) suggested that the source of all the Stanley tuffs was a continental island-arc, although some of the tuffs have a back-arc basin component.

HYDROCARBONS

Five types of hydrocarbons are found in the Stanley Group: (1) solid hydrocarbons, principally grahamite and impsonite; (2) tar sands and oil seeps; (3) shallow oil, mostly from reservoirs tens to hundreds of feet deep; (4) deep oil from reservoirs thousands of feet deep; and (5) natural gas. Most hydrocarbon occurrences in the Stanley occur west and southwest of the Potato Hills and a comprehensive review of them is beyond the scope of this discussion. Taff (1909), Beach (1945), Ham (1956), Howell and Lyons (1959), Jordan (1964), Fay (1976,1982), Curiale (1983), Chenoweth (1989), Suneson and Grasmick (1989), Suneson and Campbell (1990, p. 62–65, 67–70), and Weber (1990) have summarized the location, history, and development of hydrocarbons not only in the Stanley, but also throughout the Oklahoma Ouachita Mountains. The following discussion will focus on hydrocarbon occurrences east of the Potato Hills.

The best-known solid-hydrocarbon deposit in the eastern part of the Oklahoma Ouachitas is the Page impsonite mine. It is located in Jackfork Group sandstones ~3,000 ft above the top of the Stanley (Cardott and others, 1989). Another, much smaller, impsonite deposit (briefly described by Beach [1945, p. 45, field sheet no. 53]) is also located in Jackfork Group strata about 150 ft FEL and 600 ft FNL sec. 20, T. 3 N., R. 25 E. (see Mile 44.5, Day Two). (The previously published location in the

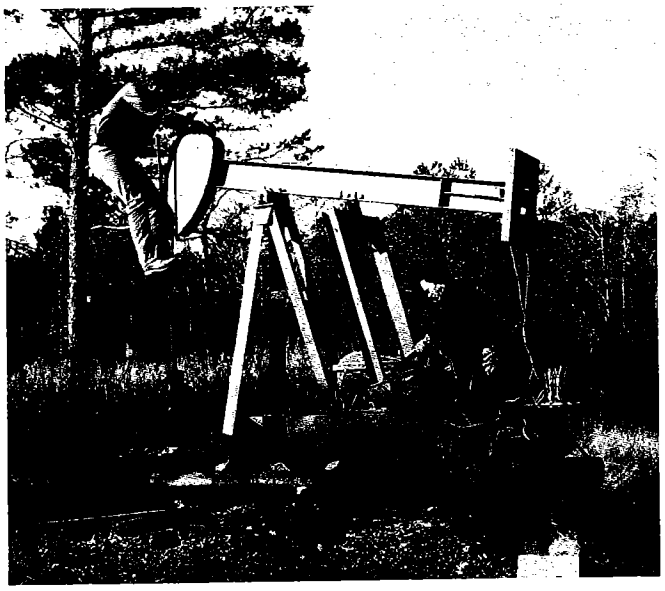


Figure 58. Jock Campbell and Neil Suneson hand-operating the pumpjack on the Select No. 1 Warren, December 5, 1988. After 45 minutes of back-breaking labor, oil was recovered. (Photograph courtesy Jane L. Weber, OGS.)

NW¼NW¼ sec. 21 is incorrect.) Solid hydrocarbons impregnating Jackfork sandstones are also present in Hells Hollow south of Buffalo Wallow Mountain, ~1,400 ft FEL and 800 ft FNL sec. 5, T. 3 N., R. 23 E. The only known deposit in the Stanley east of the Potato Hills is ~2.5 mi northwest of Whitesboro (Cardott and others, 1993), ~150 ft FEL and 1,900 ft FNL sec. 19, T. 3 N., R. 23 E.; the only remaining evidence of solid hydrocarbons are small chips mixed in with the soil.

There are no known oil seeps or tar sands in the Oklahoma Ouachita Mountains east of the Potato Hills.

The easternmost occurrence of shallow oil produced from the Stanley has been called the "Talihina" (informal) field by Suneson and others (1990, p. 4). The "Talihina" field consists of a single well—the Select No. 1 Warren, located in the SW¼NW¼SW¼ sec. 2, T. 3 N., R. 21 E. According to scout-ticket information, the original objective of the well was the Hartshorne Sandstone [*sic*] at 1,450 ft. The well spudded on March 13, 1985, and was completed on September 6, 1985. The total depth of the well is 225 ft and the perforated interval is 156–164 ft, reportedly in the Pennsylvanian. The interval was determined to be wet and the well was turned over to the landowner as a water well. Later, small amounts of oil were found in the well and a stock tank and pumpjack were installed (Fig. 58). The well never produced commercial quantities of oil, but the oil was used locally as a lubricant and preservative. The gravity of the Select No. 1 Warren oil is 37° API (Weber, 1990, p. 106). Weber (1990, 1994) studied the geochemistry of oils from throughout the Ouachita Mountains in Oklahoma. She concluded that all but one of the oils: (1) are of one genetic type; (2) are moderately mature to mature; and (3) were derived from source rocks deposited in a marine environment under mildly reducing conditions. Weber (1994) also suggested that deeply buried foreland-facies or Ouachita-facies strata could be the source of Ouachita oils.

Chenoweth (1989, p. 741) reported a dry hole with a show of oil ~25 mi east of the Potato Hills. This well may be the Muskogee Well Servicing Co. No. 1 Workman, located in the NW¼NW¼ sec. 3, T. 2 N., R. 24 E., ~1 mi northwest of Muse. The well was drilled in 1954 to 343 ft total depth and the scout ticket reported an odor of oil near T.D.

Deep Stanley oil is produced only in the SE Daisy and W Daisy fields. There are no known occurrences of deep Stanley oil east of the Potato Hills.

Natural gas has been produced from Stanley Group strata ~3 mi southeast of Talihina. As many as four wells may have been productive at one time, but records for this part of extreme southwestern Le Flore County are poor. Most of the gas—if, in fact, any ever was produced—probably was used locally. The first reported production in this area was from the Kiamichi Valley No. 1 Allen located in the SW¼SW¼NW¼ sec. 16, T. 3 N., R. 24 E. The well spudded on January 12, 1921, and drilled to 1,407 ft total depth. The well initially flowed 1.25 MMcf gas per day, but production probably declined rapidly (Chenoweth, 1959, p. 206–207). This well was later twinned by the D & J Production No. 1 Taylor, which spudded on March 30, 1963, and was completed as a gas well in January 1965 (Pitt and others, 1982, pl. I; scout ticket). Herndon Map Service shows a gas well in sec. 15 and another in sec. 16, T. 3 N., R. 22 E., but no other information on them is available.

		Return to S.H. 82.
71.7	0.7	Turn right (south) on S.H. 82.
71.9	0.2	Excellent outcrops of north-dipping, north-facing Stanley Group strata in bar ditch on right (west) side of road.
72.8	0.9	End of S.H. 82, intersection with State Highway 1. Turn left (east) toward Talihina.
73.3	0.5	Intersection of S.H. 1 and U.S. 271. End of road log for Day One.

DAY TWO

ROAD LOG

Cumulative mileage	*Interval*	
0.0	0.0	Start trip from Kerr Conference Center parking lot. Drive north on blacktop road to first intersection.
0.2	0.2	Turn right (east) and follow road up dip slope of Savanna sandstone #2. (The Savanna Formation is discussed in detail at Stop 17A.)
0.4	0.2	Crest of ridge; road curves to northeast.
1.3	0.9	Intersection with section-line road to west. Leave Wister 7.5′ quadrangle, enter Poteau West 7.5′ quadrangle. Continue straight on top of Savanna sandstone #2.
1.7	0.4	Road turns to north.
2.4	0.7	Second intersection. Turn right (east) and drive to intersection with U.S. Highway 59.
2.5	0.1	Turn right (south) onto U.S. 59.
2.9	0.4	Cross crest of ridge (Savanna sandstone #2).
3.2	0.3	Descend to flood plain of Poteau River on shale in the McAlester Formation; drive on Quaternary alluvial deposits.
3.5	0.3	Bridge over Long Lake, an abandoned channel segment of the Poteau River.
3.6	0.1	Leave Poteau West 7.5′ quadrangle, enter Wister 7.5′ quadrangle.
4.3	0.7	Bridge over Poteau River. Continue due south on Quaternary terrace and floodplain deposits.
5.6	1.3	Contact between Quaternary terrace deposits and McAlester Formation. Ascend low ridge formed by poorly developed Cameron Sandstone Member of McAlester Formation.
5.7	0.1	Intersection with county road; subtle ridge just south of intersection formed by thin Lequire Sandstone. Dip is 10°–12° northwest here.
5.8	0.1	Descend onto flood plain of Morris Creek.
6.2	0.4	Bridge over Morris Creek. Begin ascent of ridge formed by upper Warner Sandstone.
6.4	0.2	Drive onto Quaternary flood-plain deposits of Morris Creek. Ridge to the right (southwest) formed by lower Warner Sandstone.
6.6	0.2	Turn right (west) onto road leading to Howe.

HOWE, OKLAHOMA

The town of Howe, originally known as Klondike, has always been a coal-mining center. A post office was established here on May 5, 1889. At the turn of the century, about 2,000 people lived in Howe, then the largest town in Le Flore County. The population had grown so quickly that "...one found only rocky, dusty streets with no drainage; winding roads with stumps, ruts, and rocks leading from it into the country. There were no bridges, few fences, with the principal building consisting of miner's [*sic*] shacks, and of course, lots of children, with no schools or churches...." (Peck, 1963, p. 305).

The major industries were the coke ovens, a brick plant, and the nearby coal mines. Coal has been produced commercially from the Lower Hartshorne coal in the Howe area since about 1890. Most of the miners worked for Degnan and McConnel, which was the largest mining interest in the Indian Territory at that time. Production records have been kept by counties only since 1907, when Oklahoma became a state, so total coal production is unknown (Hendricks, 1939, p. 279).

Most of the coal was mined underground. Between 1900 and 1905, a battery of 40 coke ovens at Howe manufactured coke (Hendricks, 1939, p. 281). The remains of the beehive ovens are south of town. The coke produced was of good quality, but the coking was abandoned because of the distance to an adequate market (Hendricks, 1939, p. 281).

On May 5, 1961, a tornado destroyed a 36-block area in the residential part of Howe. At the time, 360 people were living in Howe; 13 people were killed and 56 were injured by the tornado. Only two weeks earlier, the nearby town of Wister was severely damaged by a major flood.

6.7	0.1	Stop 16. Shale pit northwest of road.

STOP 16

McCurtain Shale and Warner Sandstone Members of the McAlester Formation

Location: N½SE½SW¼SE¼ sec. 26, T. 6 N., R. 26 E., Le Flore County

The McCurtain Shale Member (basal member) of the McAlester Formation is well exposed in the shale pit. The

Measured Section, Stop 16

McCurtain Shale and Warner Sandstone Members of the McAlester Formation

N½SE¼SW¼SE¼ sec. 26, T. 6 N., R. 25 E., Le Flore County. Measured from crest of ridge down through shale pit to gully at entrance gate.

	Thickness (feet)
McAlester Formation:	
Warner Sandstone Member:	
4. Sandstone, grayish orange (10YR7/4), grayish red (10R4/2), moderate reddish orange (10R6/6), and moderate reddish brown (10R4/6), very fine to fine-grained, noncalcareous, ferruginous, thick-bedded, massive, breaks into large, irregular blocks; contains obscure evidence for soft-sediment deformation; includes ironstone-pebble conglomerate at base (best observed in float boulders scattered at base of shale pit). About 10 ft of weathered, jointed rock exposed at crest of ridge; top eroded (lower Warner Sandstone)	9.0
McCurtain Shale Member:	
3. Covered interval (probably McCurtain Shale Member)	22.5
2. Shale, moderate yellowish brown (10YR5/4), includes shades of light brown (5YR5/6) and moderate brown (5YR4/4; 5YR3/4); very weathered; crumbly; contains large (3 ft in diameter) concentric-ringed concretionary structures that are ferruginous in part; base gradational; top covered .	16.0
1. Shale, dark gray (N3) to olive gray (5Y4/1), weathers moderate brown (5YR3/4; 5YR4/4), noncalcareous, brittle, fissile; contains ovate concretionary structures ranging from inches to feet in diameter; base covered	28.8
Total thickness of section	76.3

overlying Warner Sandstone Member crops out at the crest of the ridge, but the contact is not exposed. The character of the sandstone can be observed easily by examining the large sandstone float blocks scattered about at the base of the shale pit. Measured Section, Stop 16 gives a description of the strata exposed in the shale pit and on the ridge crest (shown in Figure 59). At this locality, strata dip about 8°–10° northwest toward the axis of the Cavanal syncline.

The McCurtain Shale Member was named by Wilson (1935, p. 508) for the town of McCurtain, Haskell County, Oklahoma, where it crops out extensively. The Upper Hartshorne coal was included in the McCurtain Shale until Oakes and Knechtel (1948, p. 31) redefined the boundaries of the member. As currently recognized by the OGS, the McCurtain Shale Member includes the strata above the top of the Upper Hartshorne coal bed and below the base of the Warner Sandstone Member.

The McCurtain Shale is ~700 ft thick in the field-trip area. It includes an unnamed, reddish brown, very fine grained, discontinuous sandstone (1–10 ft thick or more) about midway in the shale interval (Hemish and Suneson, 1993).

STOP 16
McCurtain Shale and Warner Sandstone Members of the McAlester Formation

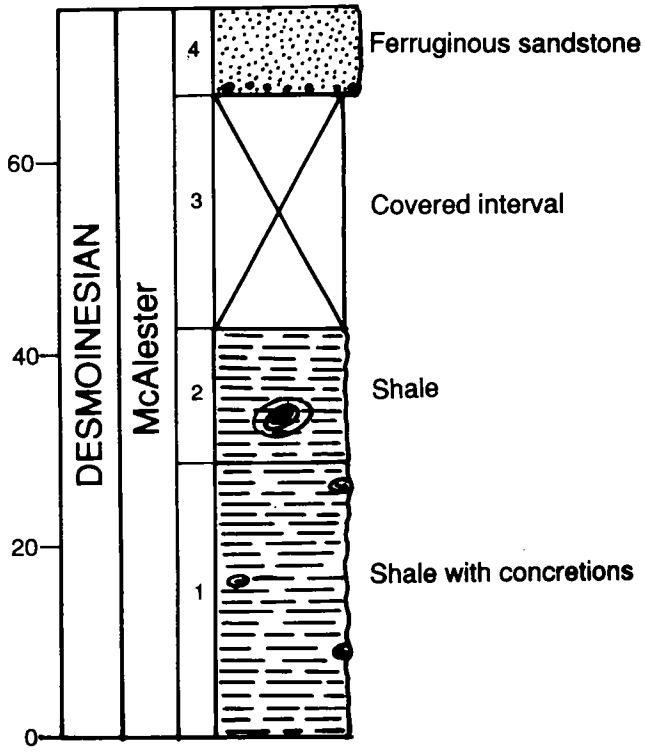


Figure 59. Graphic columnar section of part of the McCurtain Shale and Warner Sandstone Members of the McAlester Formation at Stop 16.

		Return to vehicle, turn around and drive back to U.S. 59.
6.8	0.1	Turn right (south). For the next 1.2 mi, the road crosses Quaternary flood-plain and terrace deposits.
7.3	0.5	Cross concealed axis of the Hartford anticline.
7.9	0.6	Start to ascend subtle ridge formed by the unnamed sandstone in the McCurtain Shale Member. The rocks dip 3°–6° northeast here, away from the axis of the Heavener anticline (see Stop 17B).
8.3	0.4	Cross ridge and continue due south on McCurtain Shale and Quaternary alluvial deposits.
9.2	0.9	Drive onto Hartshorne Formation. Mountain on the left (east) side of highway, just ahead, is Lost Mountain; at the top is a thick, channel-fill facies of the upper Hartshorne Sandstone. The mountain is honeycombed by abandoned underground mines (Fig. 60). An exposure of the Lower Hartshorne coal is visible from the highway at the southeast end of Lost Mountain. The coal is

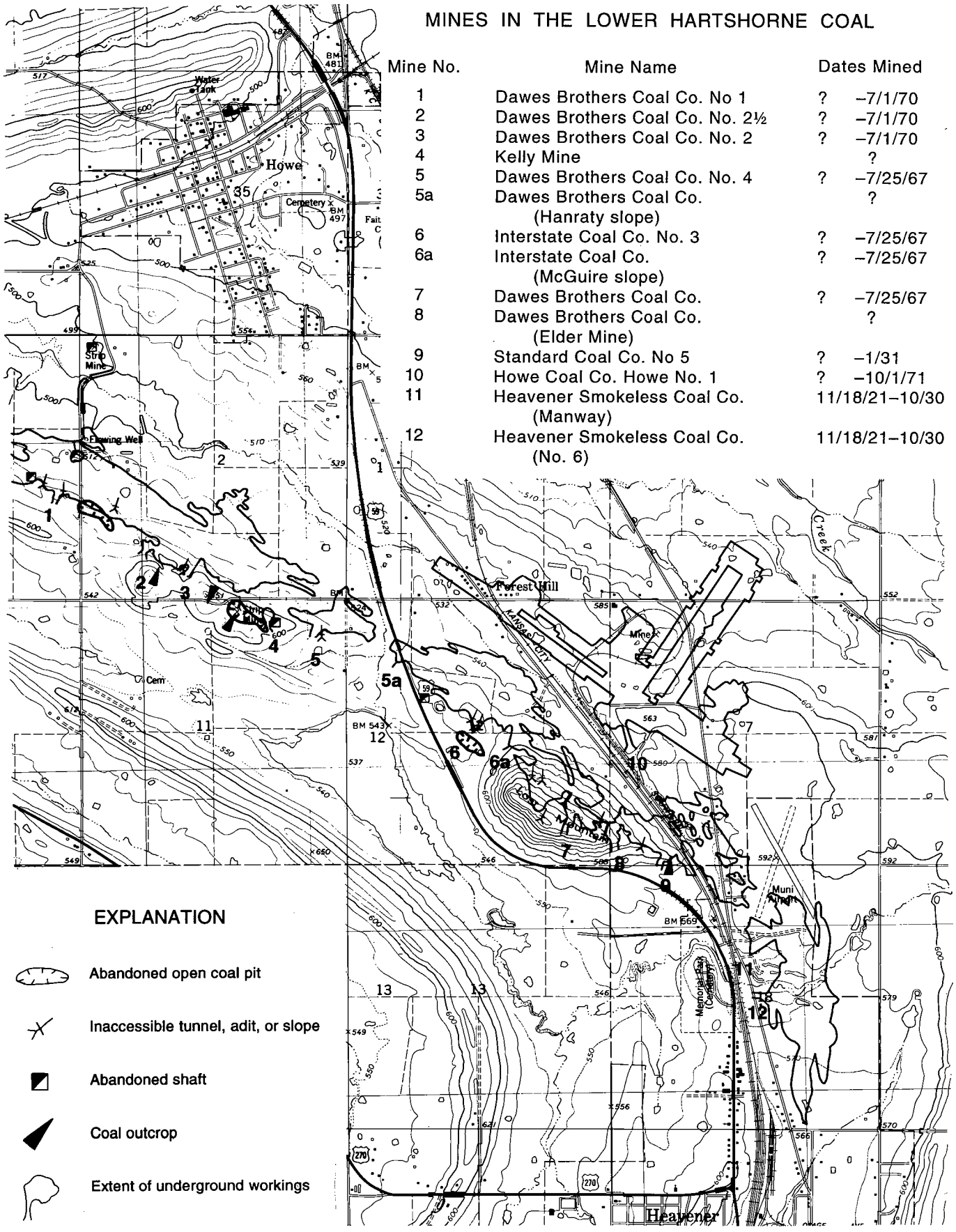


Figure 60. Locations of shafts, adits, and abandoned underground mine workings in the Lost Mountain area, north of Heavener (modified from Donica, 1978, pl. 1).

		~4 ft thick in the area. Follow curving road into the outskirts of Heavener.
9.3	0.1	Leave Wister 7.5′ quadrangle, enter Heavener 7.5′ quadrangle.
11.0	1.7	Turn left (east) and cross Kansas City Southern railroad tracks.
11.1	0.1	Turn right (south) at first intersection beyond railroad tracks; abandoned coal plant to the left.
11.8	0.7	Turn left (east) and follow road to Poteau Mountain, straight ahead.
12.0	0.2	Cross north-south section-line road (Seventh Street). Continue straight (east).
12.3	0.3	Heavener water tower, on north side of road, sits on resistant ridge of Warner Sandstone. The Warner Sandstone Member can be mapped as only a single unit from sec. 17, T. 5 N., R. 26 E., southward and eastward around the foot of Poteau Mountain to the Arkansas state line. The Warner sandstone consists of two sandstone beds separated by shale on the north side of Poteau Mountain and to the west of here (near Wister Lake, Mile 5.9, Day One; Stop 6).
13.1	0.8	Stop 17A, 17B. Headquarters for Heavener Runestone State Park. Park at trailhead to the runestone and walk down trail to excellent exposure of Savanna sandstone #2 (Stop 17A). After looking at the exposure of the Savanna sandstone and at the runestone, return to the parking area and walk up the road to the overlook for a view of the Heavener anticline (Stop 17B).

STOP 17A

Savanna Sandstone and Heavener Runestone

Location: SW¼NE¼SE¼SE¼ sec. 17, T. 5 N., R. 26 E., Le Flore County

The Savanna Formation includes all strata from the top of the youngest unnamed shale unit in the McAlester Formation to the base of the Bluejacket Sandstone Member of the Boggy Formation. The Savanna Sandstone (Savanna Formation) was defined originally by Taff (1899) in the vicinity of Savanna, Pittsburg County, Oklahoma. He described the Savanna as "a series of sandstones and shales about 1,150 feet thick" (Taff, 1899, p. 437). The boundaries of the Savanna have been redefined at various times in the past (Fig. 5), which makes it confusing to read about certain beds in the older literature.

In the field-trip area the Savanna Formation is ~1,700 ft thick and consists predominantly of pale yellowish brown, to olive gray, to medium dark gray shales and

Measured Section, Stop 17A

Savanna Formation (Savanna sandstone #2)

SW¼NE¼SE¼SE¼ sec. 17, T. 5 N., R. 26 E., Le Flore County. Measured ~60 ft northeast from foot bridge above waterfall, just above runestone shelter house, and in bluff west of waterfall in Heavener Runestone State Park.

	Thickness (feet)
KREBS GROUP:	
SAVANNA FORMATION:	
5. Shale, dark gray (N3) to grayish black (N2), weathers light brown (5YR5/6) to moderate yellowish brown (10YR5/4), brittle, breaks into small flakes on the outcrop, contains dark yellowish orange (10YR6/6) clay-ironstone concretions, top covered, basal contact sharp	3.0
4. Sandstone, grayish orange (10YR7/4) to moderate yellowish brown (10YR5/4), very fine grained, shaly, very thin to thin-bedded, wavy-, parallel-bedded; contains small-scale scour features; ripple-marked; includes abundant dusky brown (5YR2/2) carbonized plant fragments on stratification surfaces	1.5
3. Sandstone, light olive gray (5Y5/2) with moderate reddish brown (10R4/6) and moderate yellowish brown (10YR5/4) weathered surfaces, very fine grained, noncalcareous; mostly thin bedded and parallel-, wavy- to flat-bedded; parting surfaces commonly display interference ripple marks; includes a 10-in.-thick deformed bed containing flow rolls near the middle of the unit; base juts out over less resistant underlying strata	13.4
2. Siltstone, interstratified with very fine grained sandstone and shale, medium dark gray (N4), weathers brownish gray (5YR4/1), very thin bedded to laminated, wavy- to lenticular-bedded, obscurely ripple marked, trace fossils and black comminuted plant material on bedding surfaces, unit coarsens upward, heterolithic throughout, grades into underlying unit	17.0
1. Shale, dark gray (N3) to olive black (5Y2/1), oxidized moderate reddish brown (10R4/6) in part, noncalcareous, concretionary structures abundant; base covered	3.5
Total thickness of section	38.4

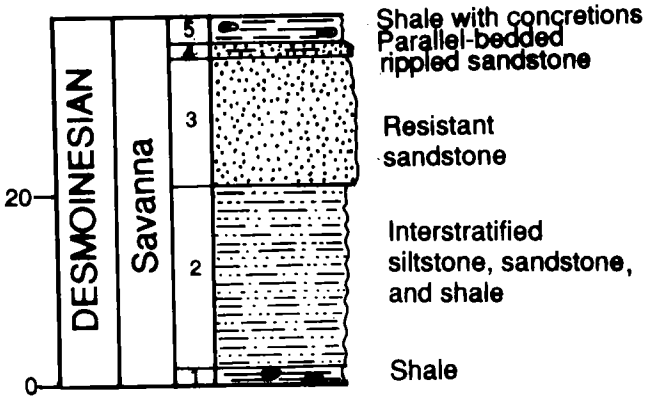


Figure 61. Graphic columnar section of part of the Savanna Formation (Savanna sandstone #2) at Stop 17A.

seven mappable, moderate brown, to grayish orange, to moderate reddish brown, very fine to fine-grained, noncalcareous sandstone units. In this report the seven sandstones are referred to informally as "#1 sandstone of the Savanna Formation" (oldest) through "#7 sandstone of the Savanna Formation" (youngest). The sandstones are massive to thin-bedded and shaly, commonly cross bedded and ripple marked, and in places they contain abundant soft-sediment deformation features. Sole marks (flutes, grooves, brush and prod marks, load casts, and trace fossils) at the base of some sandstone beds are common locally (Hemish, 1991,1992; Hemish and Mazengarb, 1992; Hemish and Suneson, 1993,1994; Hemish and others, 1990a,b,c).

The Savanna Formation includes two thin marine limestones. One is at the base (see Stops 1A and 1B) and the other is ~170 ft higher stratigraphically. They have been correlated tentatively with the Spaniard and Sam Creek Limestones, respectively, of the shelf area to the northwest (Hemish, 1993b).



Figure 62. View of the #2 sandstone of the Savanna Formation in Heavener Runestone State Park. Note the thin, parallel-bedded character of the rock in this vicinity. Many of the characteristics of the sandstone (thin, wavy beds; interference ripples; abundant comminuted plant material; soft-sediment slump features) suggest deposition in a shallow-water, deltaic environment.

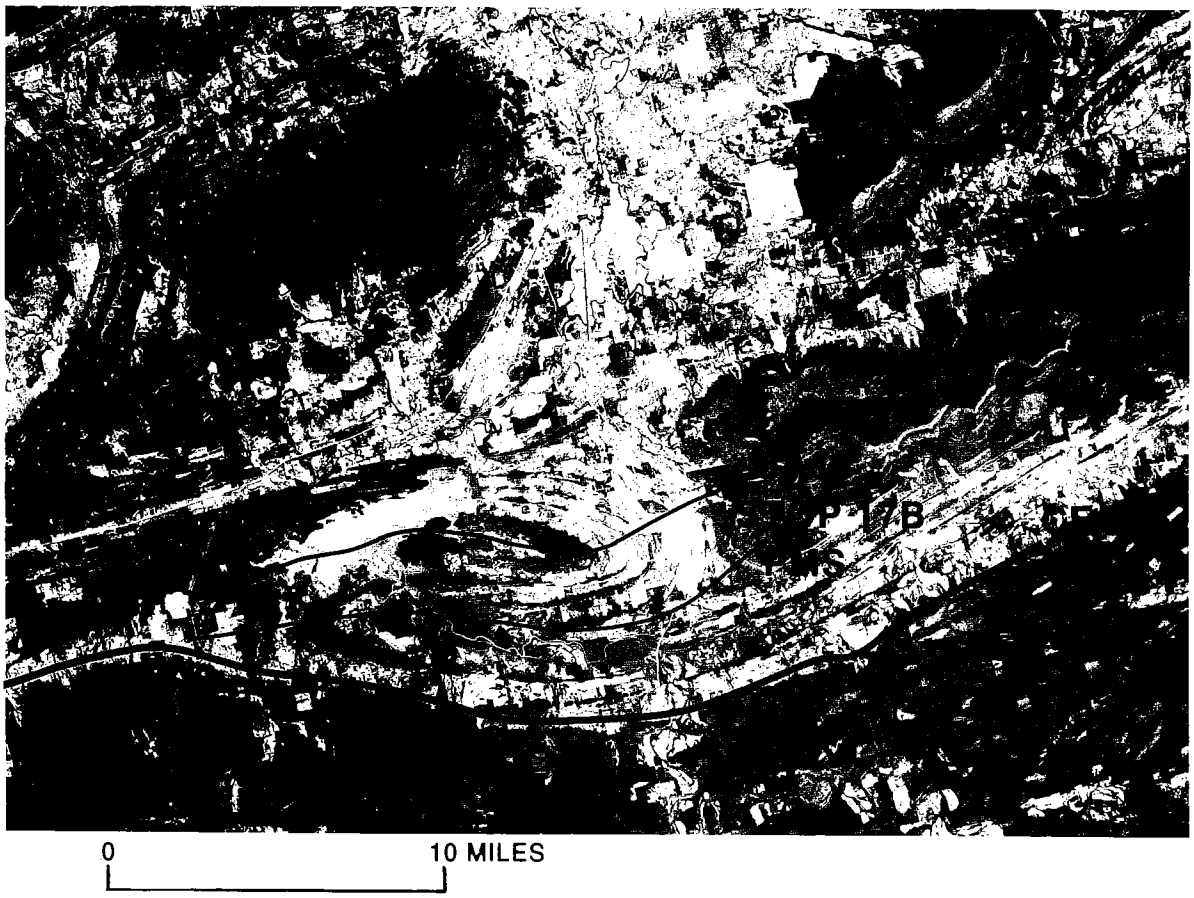


Figure 63. SPOT image of Heavener anticline (HA) and Pine Mountain syncline (PMS). Trace of Choctaw fault (CF) near boundary between mostly cultivated fields to north and mostly forest to south. Poteau Mountain on right side of image, Sugarloaf Mountain in upper right, and Cavanal Mountain in upper left. (Image by SPOT Image Corp. courtesy Dr. P. Jan Cannon, Planetary Data, Tecumseh, Oklahoma.)

The Savanna Formation contains one minable coal (Cavanal coal) in the field-trip area. A thick shale section at the top of the formation underlies the Bluejacket Sandstone Member of the Boggy Formation. The base of the Bluejacket marks the contact between the Savanna and Boggy Formations.

The #1 sandstone of the Savanna Formation crosses the steep, curving road to the park just west of the park boundary. It is a dark yellowish brown, very fine grained, blocky sandstone ~5 ft thick. In the field-trip area, its base marks the contact between the underlying McAlester Formation and the overlying Savanna Formation in the absence of the Spaniard(?) Limestone (see Stop 1A) (Hemish and Suneson, 1993).

The #2 sandstone of the Savanna Formation is well exposed throughout the park. Measured Section, Stop 17A gives a detailed description of the #2 sandstone of the Savanna Formation near the runestone shelter house (Fig. 61). Figure 62 shows the bedding characteristics of the sandstone in the vicinity of the foot trail leading down to the runestone.

Walk up road to the overlook.

STOP 17B

View of Heavener Anticline

On a clear day, the Heavener anticline is visible to the west from the overlook parking area. The anticline is marked by folded, tree-covered ridges of sandstones in the upper part of the Atoka Formation and in the Hartshorne Formation (Fig. 63). The shale intervals that separate the sandstones underlie the mostly cultivated valleys between the ridges.

The relatively high, irregular topography in the core of the anticline is formed by the informally named sandstone of Glendale (Hemish and Suneson, 1993). This unit is the oldest Atoka sandstone exposed in the southern part of the Arkoma Basin. It consists of as many as five mappable sandstones, the highest of which is ~2,400 ft below the top of the Atoka Formation. The sandstone of Glendale includes those identified as "Atoka C, E, and F" in the ARCO No. 1 Runestone (sec. 9, T. 6 N., R. 25 E.), as reported to the Oklahoma Corporation Commission (Fig. 64). In general, the strata in this unit dip moderately to steeply away from the axis of the Heavener anticline; locally, however, the beds are overturned.

The clearly curved ridge surrounding the core of the anticline is underlain by the sandstone of Potts Mountain (informal) (Hemish and Suneson, 1993). The top of this unit is ~1,000 ft below the top of the Atoka Formation and is called "Atoka B" in the ARCO No. 1 Runestone well. There are excellent exposures of this sandstone at the Wister Dam spillway (see Stop 3).

The next curved, tree-covered ridge marks the sandstone of Horseshoe Ridge (informal) (Hemish and Suneson, 1993). It is ~500 ft below the top of the Atoka Formation and correlates with the "Atoka A" in the ARCO No. 1 Runestone well. Although exposures are common along the top of the ridge, they are typically poor. Good exposures (road cut along U.S. 270 ~1 mi west of Heavener; outcrops west of dirt road in NE¼ sec. 23, T. 5 N., R. 24 E.) show clear evidence of channeling.

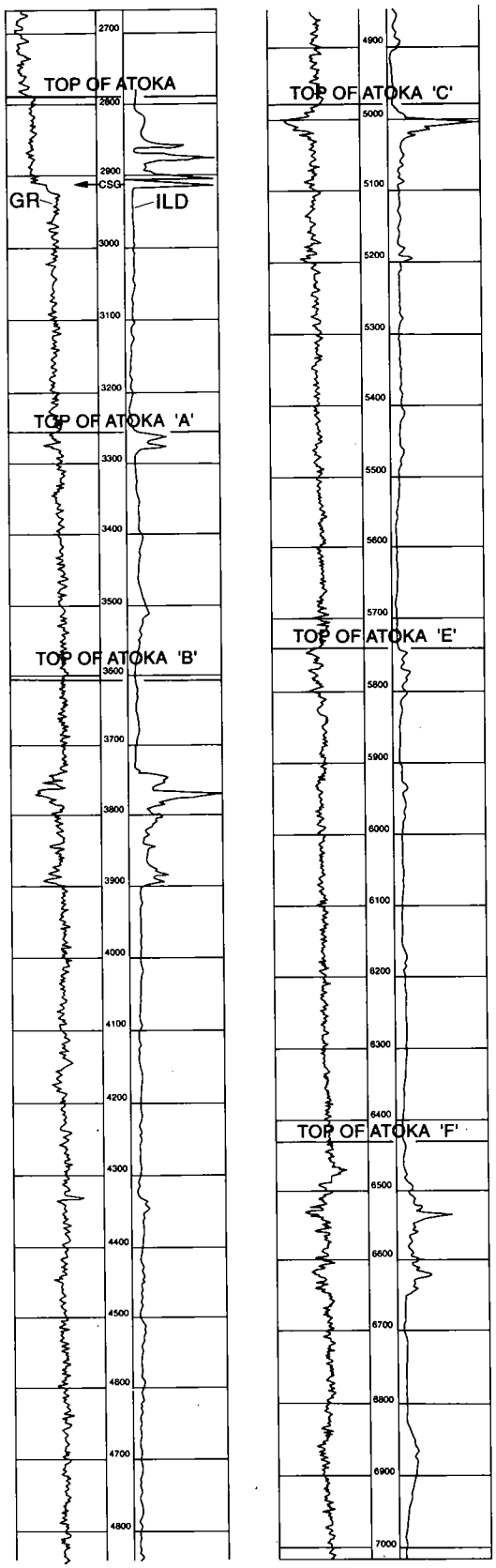


Figure 64. Gamma-ray and resistivity logs of ARCO No. 1 Runestone showing depth, thickness, and log character of sandstones in upper part of Atoka Formation. (Note: Strata are subhorizontal in this well.)



The Heavener Runestone



The Heavener runestone is one of the most visited tourist attractions in southeastern Oklahoma. It is also, perhaps, one of the most controversial. The controversy centers on when the runes were carved into the piece of Savanna sandstone as well as on their translation. The best discussion of what runes are is that of Shirk (1959, p. 365):

> Runes are characters used by Teutonic tribes of northwestern Europe. There are three classes: Anglo-Saxon, German, and Scandinavian. There are not great differences in the form of the characters as used by the three, just as today many languages use the identical alphabet form. In Anglo-Saxon the word *run* means secret and the word *runa* means magician. The use of runes was limited to a small class or profession of priests and magicians. With the introduction of Christianity in northwestern Europe, the use of runes was condemned; and the Church insisted upon the use of the Latin alphabet....Their earliest use was in the 4th century and the latest use was in the 13th in Scandinavia. Their use in England was limited to the period from the 6th to the middle of the 10th centuries.

According to local oral history, the inscriptions were first discovered in the 1830s by a Choctaw hunting party (Farley, 1973, 1990). The stone was rediscovered by early white settlers several times in the late 1800s and early 1900s, when it was known locally as "Indian Rock." In 1923, Carl F. Kemmerer sent a copy of the inscriptions to the Smithsonian Institution, which responded that the characters are runic, but that "whoever made the inscription had a Scandinavian grammar as a guide" (Farley, 1990). In subsequent years, the stone was apparently "lost" until Gloria Farley, a longtime resident of Heavener who had visited the stone with Kemmerer in 1928, rediscovered it once again in 1951. It is through her persistent efforts that foreign experts in runic inscriptions have studied what is now known as the Heavener runestone and that a state park was established in October 1970.

Four ideas have been proposed to explain the origin of the inscriptions. Two support the claim that the runes document Viking exploration of southeastern Oklahoma in about 1000 A.D. One suggests that the runes were carved in the early 1700s, and another, that the runes are of even more recent (1800s or 1900s) origin.

Shirk (1959) was the first to suggest that the Heavener runes are quite ancient; his description is based largely on the interpretations of Mr. Frederick J. Pohl, "a recognized scholar and student of Norse and Viking matters" (p. 364). The eight characters are from an established alphabet of runes: the second and last are "Medieval runes" that were developed and used in Scandinavia after about 900 A.D.; the other characters are "Ancient runes" that are part of an alphabet that was fully developed by about 500 A.D. (Landsverk, 1970). Pohl attached little significance to the different alphabets. The closest translation of the characters into Latin alphabet would be: GNOMEDAL. GNOME has been translated as "sun dial," "monument," or "boundary marker." DAL translates as "valley." GNOMEDAL could therefore mean "monument valley" (not to be confused with the more famous Monument Valley in Arizona) or "valley of the boundary marker" (Shirk, 1959, p. 366).

Alternatively, each character could have a "rune meaning" (an individual word or thought), in which case the Heavener characters could translate as: GIVE SUPPLICATION GOD MAN BEFORE DAY HAS SET (Shirk, 1959, p. 366).

Shirk (1959, p. 367) concluded that the Heavener characters are runic and that "the carving must have been done prior to Columbus; and undoubtedly prior to A.D. 1000." However, he left open the possibility, albeit slight, that the inscriptions were carved during the French period of settlement in the Arkansas River Valley.

McRill (1966) offered a substantially different explanation for the origin of the Heavener runes based on the work of Mr. Kaj Monrad, an archivist at the National Museum at Copenhagen, Denmark. The translation GNOMEDAL can be separated into two words. GNOM refers to "an elemental earth spirit, subterranean goblin, a mountain spirit" (McRill, 1966, p. 125). E is a possessive suffix and DAL is "dale" or "valley." GNOMEDAL can therefore be translated as: EARTH SPIRIT'S DALE. Furthermore, Mr. Monrad suggests that the word "gnome" was derived from the Latin word "gnomus," which means "earth-dweller" (McRill, 1966, 126). Such a derivation is significant, he believes, because southeastern Oklahoma was inhabited by a flourishing mound-building culture prior to European exploration.

Monrad suggests that the "inscription was cut in the 16th, 17th, or 18th century by a not wholly unlearned man with good knowledge of the eldest runes, as the interest for antiquity had a flowering in the last two mentioned centuries here in Scandinavia" (McRill, 1966, p. 125–126). The first French expeditions came into the Arkansas River Valley about 1718–1720; some of these early colonists were German. Captain Jean Bossu of the French Marines reported (English translation published in 1771) "two German villages...under the supervision of a Swedish captain, Mr. Arntsbourg" (McRill, 1966, p. 127) on the Arkansas River. Furthermore, the idea that the Heavener runestone may be relatively modern is supported by the word *poteau*, the French word for "post." "Could it be that the huge 'post' was carved by Germans in a party under the direction of their learned Swedish captain, describing the valley below and marking their

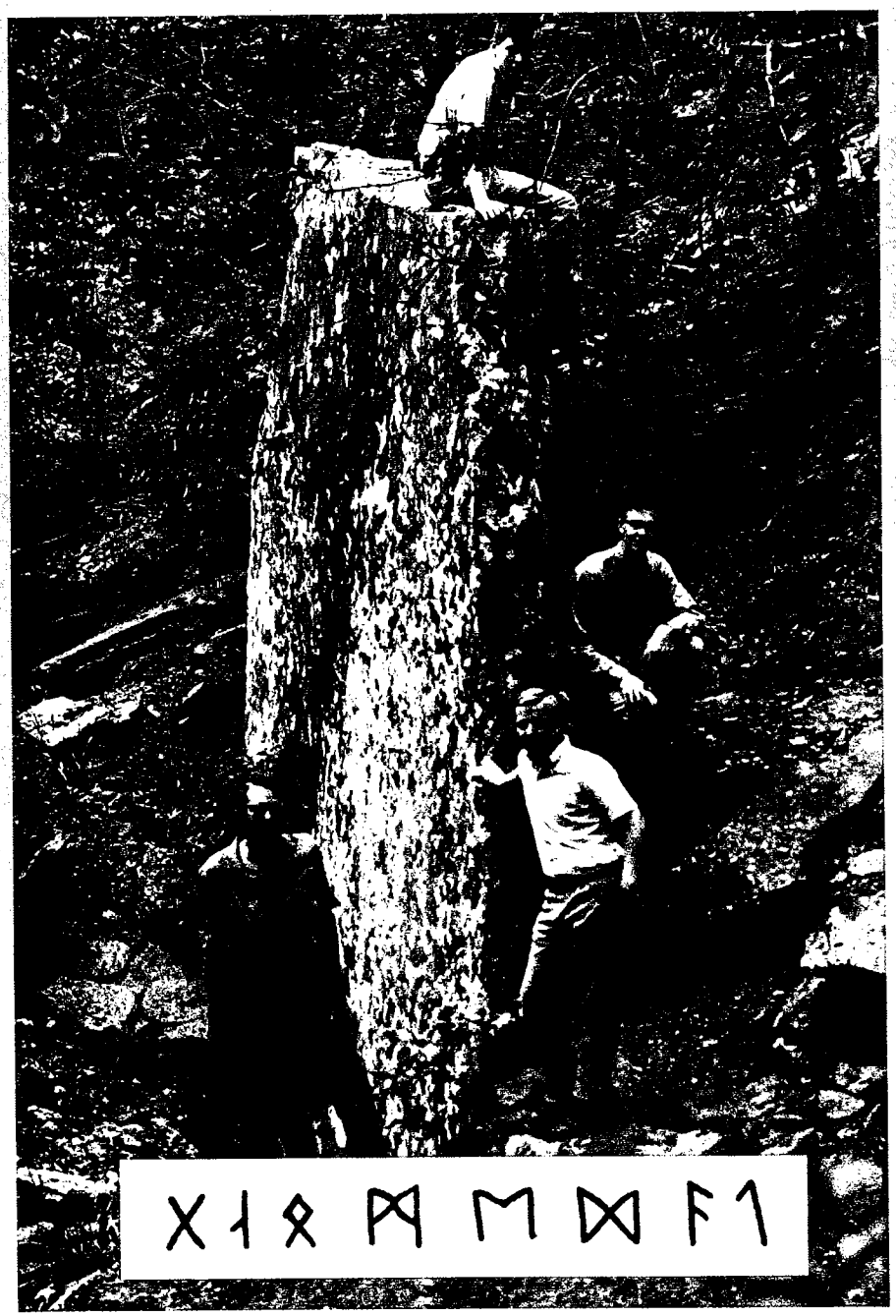
visit to the mountain?" (McRill, 1966, p. 123–124).

The third interpretation of the Heavener runestone is based on the work of Alf Monge, a former U.S. Army cryptographer who was born in Norway (Monge and Landsverk, 1967; Landsverk, 1970) . Monge interpreted the runic inscriptions not as letters, but as numbers that were "hidden" in a runic code. He believed the numbers represented a date from the Roman Catholic perpetual calendar, which was in wide use prior to about 1500 A.D., when calendars were first printed annually. Using an auxiliary table known as the "Easter Table," Monge and Landsverk (1967) deduced the date that the Heavener inscription was carved: November 11, 1012 A.D.

The plausibility of such a solution of the runic cryptograph is supported by six other inscriptions found in Oklahoma (Pohl, 1972): Heavener no. 2 (1015 A.D.), Poteau (1017 A.D.), Heavener no. 3 (1022/23 A.D.), Tulsa (1022 A.D.), and Shawnee (1024 A.D.). Landsverk (1970, p. 34) suggests that some of these "dated inscriptions... are so similar in construction that they are most certainly the work of the same runemaster."

Wycoff (1973) (present State Archaeologist) reviewed evidence for the antiquity of the six runestones found in Oklahoma (including the Heavener no. 1 runestone). He suggests that two were probably carved by prehistoric or early historic Native Americans. Based on the character of the inscriptions and hardness of the stone into which they were carved, Wycoff (1973) believes two others were probably cut about ten years before they were found. Another was not present in the 1920s, as reported by a man who grew up near where it was found. Therefore, using other Oklahoma runestones to support the view that the Heavener runestone was carved by Vikings is inadequate. In addition, Wycoff cites several reports that criticize the methodology of the Monge-Landsverk cryptological approach and states that the only people who use it are "those people promoting the cause for Viking explorations in the interior of the United States.... They accept its results because the results match their beliefs" (Wycoff, 1973, p. 28).

Wycoff admits that there is no evidence against a Viking origin for the Heavener runestone, but suggests that it was carved in the 1800s. Fort Smith, 35 miles to the north, was an active trading center and military post visited by people with a European education that "included the study of runic alphabets" (Wycoff, 1973, p. 31). In the 1890s, a resort near Heavener catered to wealthy German and Dutch families. It is possible that the Heavener runestone was carved during an outing and climb of the hill just east of Heavener.



Photograph of the Heavener runestone taken in early 1960s before protective housing was built around it. Runic inscription (shown on inset) is just barely visible in the photo, above and to right of Gloria Farley. Mark Farley on top, Scott Farley behind, and Jim Johnson in front of runestone. (Photograph courtesy of Fran Johnson,* The Heavener Ledger, *Heavener, Oklahoma.)

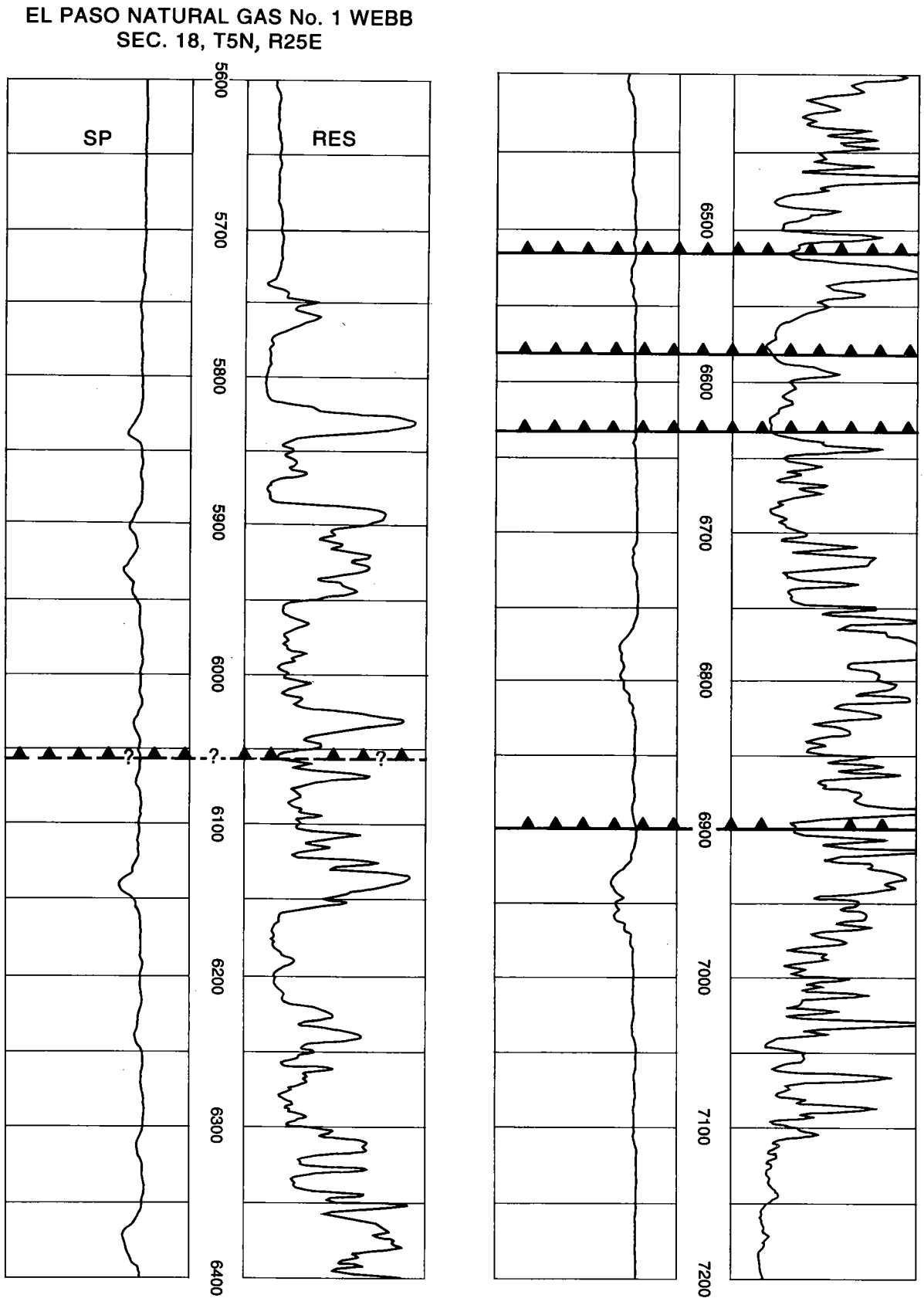


Figure 65. Part of El Paso Natural Gas No. 1 Webb well log showing multiple repetition of beds and interpreted positions of thrust faults.

The outermost curved ridge is formed by the Hartshorne Formation. From this viewpoint, our next stop, the "Heavener road cut" (Stop 18) can be seen. In addition, the strip pit of the Pine Mountain Coal Mine and the reclaimed land adjacent to the pit can be seen.

The Heavener anticline is slightly asymmetric—the sandstones of Glendale and Potts Mountain dip slightly more on the south flank than on the north flank. The sandstone of Horseshoe Ridge is too close to the axis of the Pine Mountain syncline (~2 mi south of the axis of the Heavener anticline) to be useful for determining asymmetry. A brief inspection of two seismic lines across the anticline indicates that the subsurface Spiro sandstone is not folded in a manner similar to that of the strata on the surface.

Several wells have drilled the core of the Heavener anticline (listed in Hemish and Suneson, 1993); the most important for determining the subsurface structure is the El Paso Natural Gas No. 1 Webb (NW¼NE¼ sec. 18, T. 5 N., R. 25 E.). The interval between ~5,700 ft and 7,100 ft shows multiple repetitions of beds (Fig. 65), which suggests thrusting. Tectonic thickening in this interval, evidence for thrust faults elsewhere in the well, and the apparent absence of any major thrust faults at the surface led Hemish and Suneson (1993) to interpret a duplex structure within the Atoka Formation beneath the surface expression of the anticline. They suggest the asymmetry is caused by a south-directed back thrust that is exposed locally at the surface.

		Retrace route to Seventh Street (north-south section-line road).
14.3	~1.2	Turn left (south) on Seventh Street.
15.0	0.7	Turn right (west) on C Street. Enter Heavener, Oklahoma.

HEAVENER, OKLAHOMA

The area near Heavener, Oklahoma, was known to early Native Americans as the "Prairie of the Tall Grass." With the arrival of the Choctaws in the mid-1800s, the town became a missionary center known locally as "Choctaw City" (not to be confused with the Choctaw City [Chahta Tamaha] near Durant, the capital of the Choctaw Nation from 1863 to 1882). With the establishment of the Kansas City Southern railroad in 1896, the town was renamed after Joseph Heavener, a local merchant who had intermarried into the Choctaw tribe and who owned the land on which the town was built.

In 1910, the railroad built a roundhouse in Heavener. The town thrived as railroad men established their homes there.

15.5	0.5	Turn left (south) on Main Street.
15.55	0.05	Turn right (west) and cross Kansas City Southern railroad tracks.
15.6	0.05	Intersection with U.S. 59/270. Turn left (south).
16.6	1.0	Leave Heavener 7.5′ quadrangle, enter Hontubby 7.5′ quadrangle. Road to Pine Mountain Coal Mine on right.

PINE MOUNTAIN COAL MINE

The Pine Mountain Mine has been operated by the Farrel-Cooper Mining Co. since 1990. The Lower Hartshorne coal is mined here by stripping methods (Fig. 66A,



Figure 66. *A*—View of cut and 70-ft highwall in the Pine Mountain Mine. Mine development is progressing toward the treed area. LeRoy Hemish and Bob Cooper (white shirt), Vice President of Farrell-Cooper Mining Co., are standing on spoils that are used for recontouring in Farrell-Cooper's ongoing reclamation program. *B*—Heavy equipment used to load and haul blasted overburden from top of the Lower Hartshorne coal in the Pine Mountain Mine. Overburden consists of sandstones, siltstones, and shales. Casts of tree trunks standing in growth position are common in the unit overlying the coal bed, but none are visible in this photograph.

B). Overburden is removed primarily by two 8-ft^3-bucket draglines to reach the 2.6–3.0-ft-thick coal bed. Maximum thickness of overburden that is removed is ~90 ft.

Monthly coal production from the Pine Mountain Mine was ~23,000 short tons as of May 1994. Data provided by the Farrell-Cooper Mining Co. indicate that the coal has a sulfur content of ~1.0%, an ash content of ~11%, and a heat value of ~12,500 Btu. The coal is trucked to the AES Shady Point co-generation plant, where it is used to generate electricity and CO_2. The Lower Hartshorne coal bed can be seen in the Heavener road cut (Stop 18), where it is 1.9 ft thick.

17.7 1.1 Stop 18.

STOP 18

Upper Atoka Formation and Hartshorne Formation

Location: Heavener road cut, NW¼SE¼NW¼ sec. 36, T. 5 N., R. 25 E., Le Flore County, along U.S. 59/270

The Atoka and Hartshorne Formations at the Heavener road cut have been described in detail by Donica (1978). (His description is reproduced here as Measured Section, Stop 18.) He placed the contact between the two formations at the base of the lowest sandstone (Fig. 67). This sandstone is 10–40 ft above two thin (0.5-ft-thick) coal beds. Hendricks' (1939, p. 267) measured section of the Hartshorne Formation at Petry's Cut along the railroad ~0.9 mi to the east showed the Atoka/Hartshorne contact at the base of a 1.5-ft-thick sandstone that is immediately below the lower of the two thin coals at Petry's Cut. Donica (1978, p. 16) suggested that the sandstone used by Hendricks (1939) to locate the base of the Hartshorne is discontinuous and is not the more extensive Hartshorne sandstone that typically is used to mark the base of the Hartshorne Formation (see discussion at Stop 8). A more likely explanation is that Hendricks (1939, p. 264) believed that the Atoka Formation did not contain any coal beds; therefore, he placed the two thin coals in the Hartshorne Formation. Briggs and others (1975, p. 93) briefly described the strata at the Heavener road cut and appear to include the entire exposed sequence in the Hartshorne Formation.

The strata exposed in the Heavener road cut (Fig. 68) were deposited in interdistributary marshes and swamps in a delta-plain environment (Donica, 1978; Briggs and others, 1975). Coal beds represent periods of peat accumulation with little or no sediment influx; shale intervals represent periods of slightly greater clastic sedimentation; and the sandstones are overbank and/or crevasse-splay deposits that probably represent periods of flooding. Unlike the Hartshorne exposure at Stop 8, the Heavener road cut exposes no distributary-channel deposits. The "alluvial channel" in the lower Hartshorne Formation shown by McDaniel (1961, fig. 8) may refer to the "channel sandstone" pictured by Hendricks (1939, pl. 29A); however, the sandstone pictured by Hendricks (1939) is in the upper part of the Hartshorne and is more likely a distributary channel. Similar, but thicker, dis-

Measured Section, Stop 18

Upper part of Atoka Formation and lower part of Hartshorne Formation

NW¼SE¼NW¼ sec. 36, T. 5 N., R. 25 E., Le Flore County. Measured on west side of the road cut along U.S. Highway 59 [now U.S. 59/270], approximately 1.5 mi south of Heavener, Oklahoma (uppermost 20 ft not included) (from Donica, 1978).

	Thickness (feet)
HARTSHORNE FORMATION:	
32. Sandstone, gray, weathers buff (top of section)	2.0
31. Shale, interlaminated gray and brown, containing siderite nodules and upright *Calamites*	14.0
30. Sandstone, gray, fine-grained, containing upright *Calamites*	9.0
29. Shale, gray, interbedded with cross-bedded sandstone.	6.8
28. Shale, gray, thinly laminated, containing siderite nodules	6.8
27. Shale, black, carbonaceous	1.2
26. Coal (Lower Hartshorne)	1.9
25. Underclay, rooted	0.5
24. Shale, gray, fissile, containing siderite nodules and streaks of coal	9.4
23. Sandstone, gray, cross-bedded containing upright *Calamites*, interbedded with gray shale containing siderite nodules	12.8
22. Shale, gray, containing siderite nodules	5.8
21. Sandstone, gray, fine-grained, containing upright *Calamites* one of which is 1.8 feet tall. The sandstone is variable in thickness	1.7
20. Shale, gray, finely laminated with siderite nodules with which [*sic*] contain plant impressions	4.2
19. Sandstone, gray, finely-grained [*sic*], cross-bedded	1.9
18. Shale, gray, thinly laminated, containing siderite nodules	1.6
17. Sandstone, gray, very fine-grained, cross-bedded	0.5
16. Shale, gray, thinly laminated, containing siderite nodules	2.5
15. Sandstone, gray, very fine-grained, containing upright *Calamites*	5.2
14. Shale, gray finely laminated, with siderite nodules which contain *Calamites* and numerous leaf impressions	5.7
13. Sandstone, gray, fine-grained containing *Calamites*	7.5
ATOKA FORMATION:	
12. Shale, gray, thinly laminated, containing siderite nodules	5.7
11. Shale, carbonaceous, containing thin coal streaks and siderite nodules	3.7
10. Coal	0.5
9. Shale, black, carbonaceous, with numerous coal streaks	11.3
8. Shale, gray, containing siderite nodules	6.5
7. Shale, gray	11.5
6. Shale, sideritic, [overlying] shale, black, carbonaceous, containing thin coal streaks and numerous plant impressions	1.2
5. Coal	0.5
4. Shale, black with coal streaks	0.9
3. Shale, black, containing siderite nodules	0.9
2. Shale, gray, containing siderite nodules	2.0
1. Shale, gray, thinly laminated, with interbedded sideritic shale (base of section)	31.8
[*Total thickness of section*	177.5]

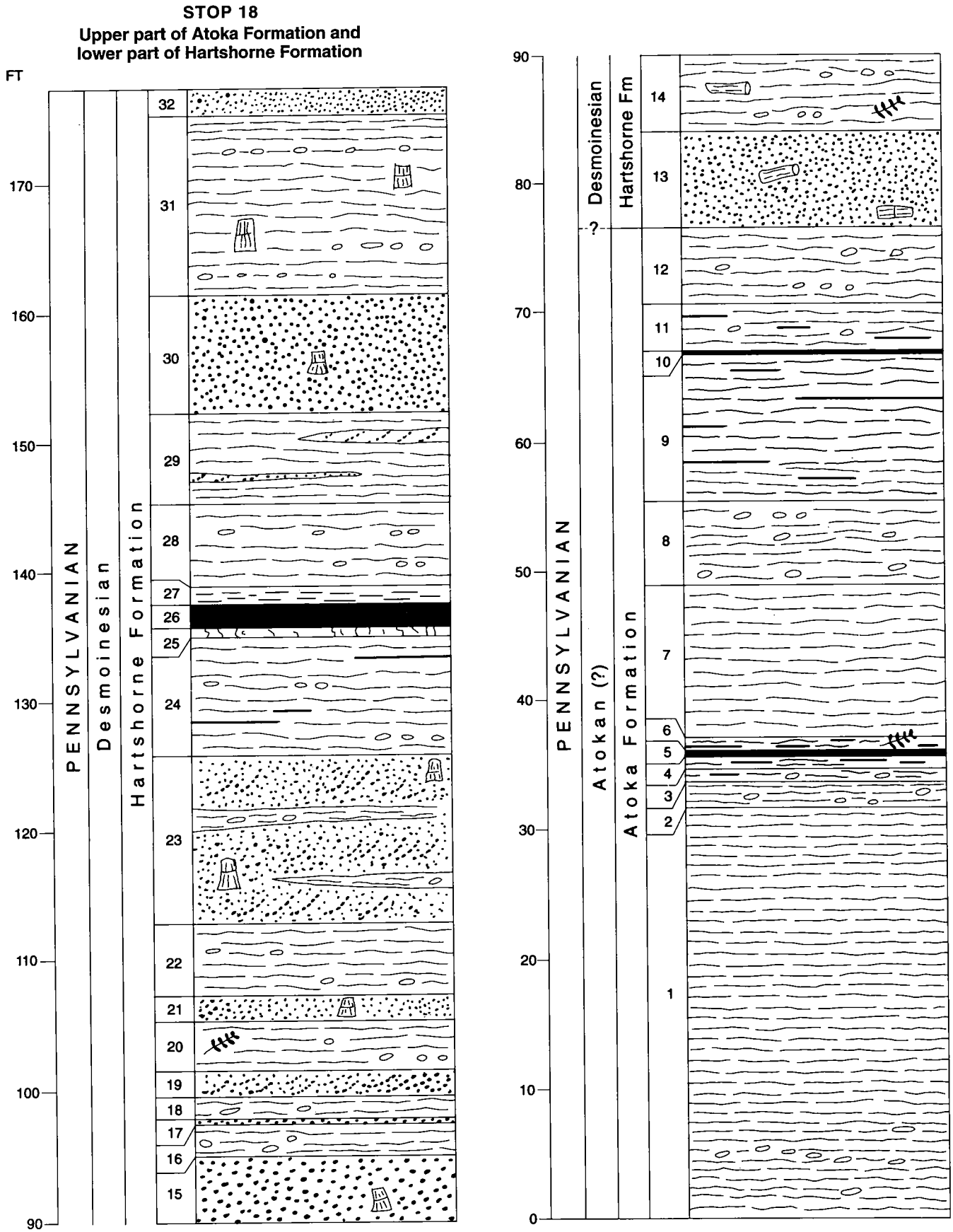


Figure 67. Graphic columnar section of upper part of Atoka Formation and lower part of Hartshorne Formation at Stop 18 (modified from Donica, 1978).



Figure 68. Heavener road cut, Stop 18. Dark-colored strata on left are mostly shale in the Atoka Formation; lighter-colored strata above are the lower part of Hartshorne Formation. Upper Hartshorne sandstone exposed at top of road cut, underlain by conspicuous black band of lower Hartshorne coal.

tributary channels in the upper part of the Hartshorne Formation near Heavener are on Lost Mountain (SE¼ sec. 12, T. 5 N., R. 25 E.) (see Mile 9.2, Day Two) and on the hill in the NE¼ sec. 5, T. 5 N., R. 25 E. (unpublished field observations, L. A. Hemish).

17.8	0.1	Leave Hontubby 7.5′ quadrangle, enter Hodgen 7.5′ quadrangle.
18.2	0.4	Cross Poteau River.
19.3	1.1	Intersection with paved road to Summerfield (west) and Loving (east).
19.4	0.1	Enter Hodgen.

HODGEN, OKLAHOMA

The town was founded in 1896 and named Houston after the famous Texan, Sam Houston. The name was changed in 1910 to Hodgen, a misspelling of the name of J. W. Hodgens, a timber buyer for the Kansas City Southern Railway.

20.2	0.8	Cross trace of Choctaw fault. The exact location of the trace of the Choctaw fault cannot be determined here. It can be located to within ~2,000 ft 1.5 mi to the east and 2 mi to the west. At these localities, strata in the upper part of the Atoka Formation north of the fault dip steeply and face north. Strata in the middle(?) part of the Atoka Formation south of the fault are turbidites, face north, and dip steeply (upright to overturned).
20.4	0.2	Highway bends right (west).

Outcrops on the right (north) are turbidites in the Atoka Formation. This section was measured by McDonald (1986, fig. 77) who described it as a shale-rich deep-water sequence dominated by lithofacies D (pelitic-arenaceous) and G (hemipelagic and pelagic) of Mutti and Ricci Lucchi (1978). Sedimentary structures such as sole marks (flutes, grooves, prod marks, and bounce casts), incomplete Bouma sequences (Fig. 46), and convolute stratification characterize many of the sandstone beds.

		For the next 3 mi, the road crosses steeply dipping, imbricately thrust-faulted Atoka Formation turbidites.
24.4	4.0	Cross Cedar Creek.
25.1	0.7	Intersection with Holson Valley Road to right (west). Continue south on U.S. 59/270.
26.5	1.4	Leave Hodgen 7.5′ quadrangle, enter Hontubby 7.5′ quadrangle.
26.9	0.4	Village of Stapp. Harlton (1938, p. 893) named the Stapp conglomerate member of the Union Valley Formation (Morrowan) for exposures along the Kansas City Southern railroad 1 mi south of Stapp. These strata are now included in the Johns Valley Shale (Hart, 1963).

STAPP, OKLAHOMA

The village of Stapp was founded as Thomasville in 1897; a post office was maintained there from 1918 to 1944. The town was originally a thriving lumber town with a population of about 1,000. Nothing remains of the old town; U.S. 59/270 passes through what used to be the lumber yards and single-bandsaw mill of the Buschow Lumber Co. In contrast to the Dierks Lumber Companies, which were some of the first in the timber industry to practice conservation (see History of Logging in the Oklahoma Ouachitas, p. 116), the Buschow operation was seen as "...the ultimate in devastation, ...one of the bad ones, ...a cut-out-and-get-out operation" (Smith, 1986, p. 120).

27.1	0.2	Leave Hontubby 7.5′ quadrangle, enter Hodgen 7.5′ quadrangle.
27.4	0.3	Leave Hodgen 7.5′ quadrangle, enter Big Cedar 7.5′ quadrangle.
27.5	0.1	Cross Shawnee Creek.
27.6	0.1	Road to Ouachita Vocational Technical Camp to right (west).
27.7	0.1	Stop 19.

STOP 19

Johns Valley Shale at Stapp

Location: C E½ sec. 12, T. 3 N., R. 25 E., along U.S. 59/270, Le Flore County

PREVIOUS STUDIES

Like the outcrop at Compton Cut (Stop 13) and that at the Hairpin Curve (SE¼ sec. 3, T. 3 N., R. 19 E.) (Suneson and others, 1990, p. 19–20), this outcrop and the one along the Kansas City Southern railroad ~0.25 mi to the southeast have fueled many of the arguments about the origin of the Johns Valley Shale. Taff (1910) may have been the first geologist to recognize what he mapped as Caney Formation this far to the east from its type section in the Tuskahoma syncline, Pushmataha County. Miser (1926) mapped the Stapp outcrops as Caney on his geologic map of Oklahoma. Ulrich (1927, p. 22) proposed the name Johns Valley Shale "for the Pennsylvanian black shale" that "rests on the Jackfork Sandstone and is overlain by ... the Atoka Formation." He recognized that the Johns Valley contained erratics of older formations; boulders from three outcrops near Stapp (nos. 23, 24, and 25 of Ulrich, 1927, p. 11) contained fossils similar to those in the Bromide, Viola, and Wapanucka Formations. Powers (1928) apparently rejected using Johns Valley Shale and instead suggested that the boulders occurred in the Caney, Wapanucka, and Atoka Formations. The sandy matrix in the boulder conglomerate along the railroad southeast of Stapp led Powers (1928, p. 1041) to map it as basal Atoka, but he also claimed that boulder-bearing Caney Shale was present nearby. Like Ulrich (1927), Powers (1928) believed that the boulders were glacial dropstones; however, Powers (1928, p. 1044) added that the mountainous region that supplied the glaciers may have been a pre-Ouachita uplift near the present-day Ouachita Mountains (see Stop 13, and discussion of "Bengalia," p. 73).

Studies of the Stapp (and other) outcrops by Harlton clarified (1933), then confused (1934,1938) the relation of the Johns Valley to surrounding strata (Fig. 57). Harlton (1933) identified this highway road cut as Johns Valley and identified Early Pennsylvanian ostracods and foraminifers in the shales. In 1934, Harlton applied the term "Bendian" to strata deposited between the Chesterian and Desmoinesian and included three of the four principal turbidite units in the Ouachitas (Stanley, Jackfork, Johns Valley) within it. He divided the Johns Valley into upper and lower parts and suggested that the railroad cut at Lamberson Spur (near Compton Cut, Stop 13) was in the upper part (Harlton, 1934, p. 1039). He also suggested that the Lamberson Spur outcrop, as well as that exposed along the railroad southeast of Stapp, were in the basal part of the Atoka Formation (Desmoinesian) (Harlton, 1934, p. 1025,1028,1047). Despite his confusing stratigraphy, Harlton (1934, p. 1028)—in agreement with some present-day interpretations—mentioned "submarine landslides and mud flows" as possible mechanisms for depositing the boulders, but he wasn't entirely pleased with the interpretation.

Harlton's 1938 paper is even more confusing than his 1934 paper. The Bendian period is divided into the Pushmataha (lower) and Morrow (upper) series and the name Johns Valley is abandoned (Fig. 57). Outcrops referred to as the Johns Valley in 1934 are either Round Prairie (including this outcrop) (Fig. 69A), Wesley, or Markham Mill. The Stapp Conglomerate (previously at the base of the "Desmoinesian" [*sic*] Atoka Formation) is elevated to a member within the Morrowan Union Valley Formation. Harlton (1938, fig. 18) sketched this outcrop in detail (Fig. 69A), although some of his structural interpretations are questionable given the present condition of the

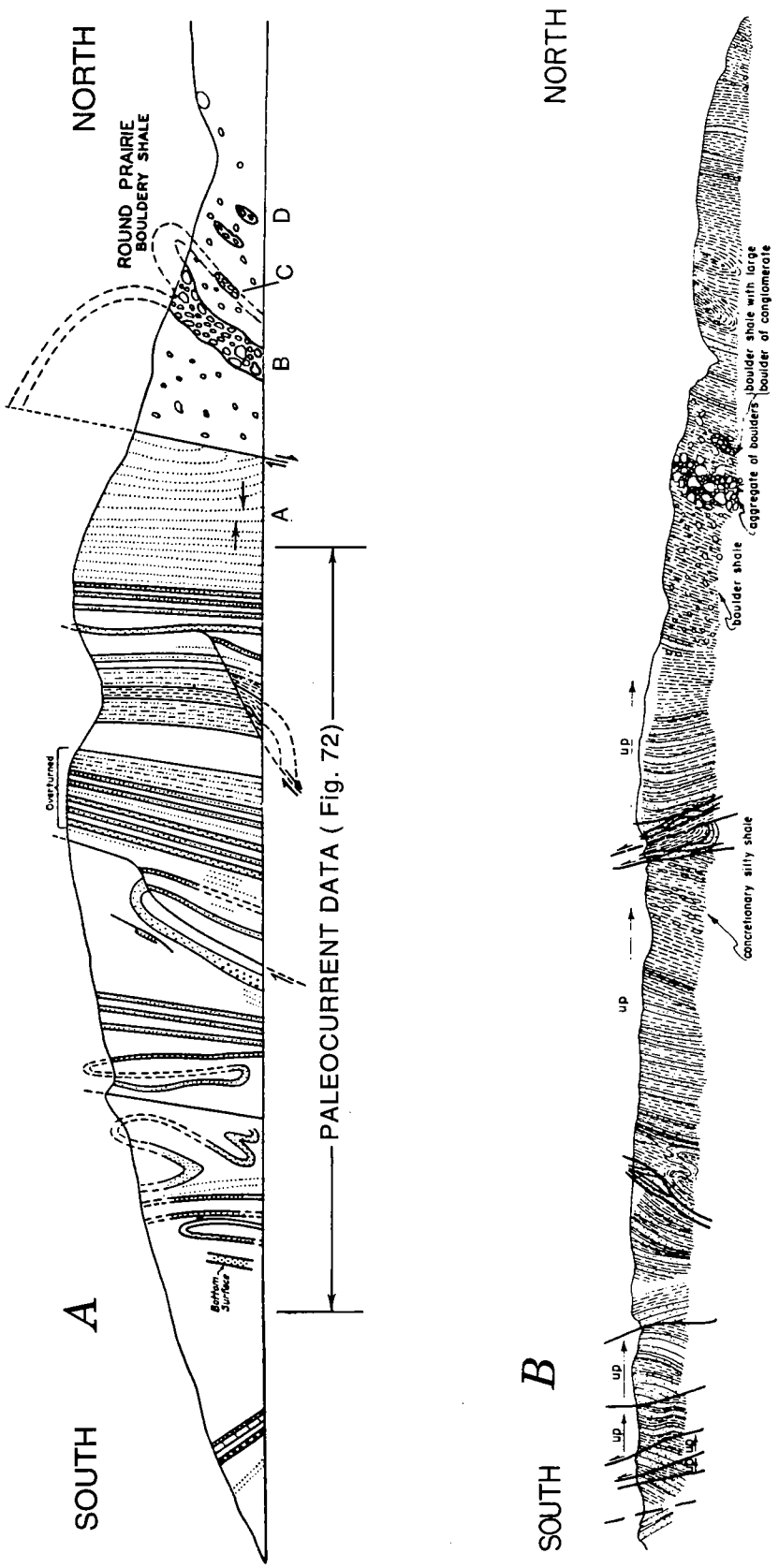


Figure 69. *A*—"Diagram of beds exposed in road cut south of Stapp, showing position of boulders in Round Prairie Shale referred to in text" (from Harlton, 1938, fig. 19). Capital letters refer to observations by NHS (see text): A = opposing facing directions indicate isoclinal folding (arrows added by NHS); B = boulder bed extends to base of exposure; C = part of boulder bed not observed; D = structurally dismembered sandstone beds. *B*—Sketch of same road cut by Misch and Oles (1956).

outcrop. (A sketch of the same outcrop [Fig. 69B] by Misch and Oles [1956] is included for comparison.) In addition, Harlton (1938, p. 897–899) described in detail a boulder conglomerate in this outcrop:

> A striking exposure of the Round Prairie bouldery shale may be observed in Sec. 25 [*sic*], T. 3 S. [*sic*], R. 25 E., just south of the village of Stapp, Oklahoma, on U.S. Highway 270 (Fig. 19). These exposures are comparable to the complicated folded and faulted beds of the excellent exposures at Compton Railway cut.
>
> Additional interesting evidence concerning the history of the exotics may be observed at this exposure. Some of the boulder arrangements (Fig. 19, Nos. 30–38) introduce a fascinating elucidation in this regard. At these stations a thick boulder zone is found, which is characterized by small pebbles approximately 0.25 inch in diameter, varying to cobbles and boulders some of them several feet in diameter. Many of the boulders are well rounded to subrounded and show interesting pitted surface phenomena, which undoubtedly occurred before they dropped into the matrix composing this zone. Most of the pebbles and cobbles also are rounded to subrounded and well worn, and scores of them show the pitted surfaces. All exotics in this zone are very closely bounded in a terrigenous clayey deposit.
>
> The bonding material of the boulder zone at this locality consists of red, greenish gray, yellowish gray, and slightly carbonaceous dark gray to black clay and shale. The colors are not distributed evenly and there is no suggestion of bedding planes. Splotches of clay of one color may lie adjacent to clay or shale of any other of the aforementioned colors. Everywhere within the matrix may be found sandstone or limestone grains, ranging from ⅛ millimeter to several millimeters. These do not show any tendency toward being sorted, either by size or by type of material. Some surfaces show flat flakes of hematite, now partly weathered to limonite. The orientation of these hematite flakes suggests that they have been formed along planes of differential movements. The entire bonding composition shows no indication of normal bedding or sorting, commonly found in sediments deposited by water.
>
> The boulders, cobbles, and pebbles within this zone are made up predominantly of limestone concretions and phosphatic nodules of Caney (Chester) age and different varieties of Simpson (Ordovician) sediments. Many cobbles and pebbles of black chert of Pushmataha (Bendian) age are found.
>
> Several boulders and cobbles of conglomerate also are present in this zone, and some of the conglomeratic boulders are made up of the following constituents.
>
> Drab arenaceous dolomite, matrix dense silty dolomite; tan dense limestone; greenish flaky shale in thin laminae; brown arenaceous dolomite. The bond is firm, limonitic. Tan dense limestone; coarse rounded, frosted sand; dark gray shale; whitish gray bird's-eye-like limestone; brownish finely crystalline dolomite; brown arenaceous dolomite.
>
> Buff dense limestone; brown highly dolomitic, rounded frosted sand; brown arenaceous dolomite; gray finely crystalline arenaceous dolomite, sand grains rounded, frosted, and pitted—gilsonitic in part, much calcite veining; dark gray limestone concretions from the Caney of Chester age; black shale; clay ironstone concretions.
>
> Some cobbles of conglomerate are described as follows.
>
> Tan dense limestone; buff finely crystalline limestone; brown arenaceous dolomite; light gray coarsely crystalline dolomite, siliceous residue consists of silicified brachiopod fragments probably of Trenton age.
>
> Rounded pebbles of gray coarsely crystalline fossiliferous limestone; dark gray Mississippian limestone concretions and brown arenaceous dolomite.
>
> Tan dense limestone matrix; grey arenaceous saccharoidal dolomite, grains rounded and frosted; black shale.

Clearly, the Johns Valley Shale at this and other localities in the Ouachita Mountains was enigmatic to many early stratigraphers, apparently more so to Harlton than to others.

Shideler (1968) identified boulders from Ozark and Arbuckle foreland facies strata in this outcrop, including the Bromide (Ordovician), Chimneyhill (Silurian), Fite (Ordovician), Jasper (Ordovician), Kindblade (Ordovician), Wapanucka (Pennsylvanian), and West Spring Creek (Ordovician) Formations. Visher (1978) suggested that the clasts were derived entirely from Ouachita facies strata (Bigfork limestone and chert, Arkansas Novaculite, Caney Shale, and Jackfork sandstone), but this is incorrect. His interpretation of this outcrop—that it is a tectonic breccia and not a stratigraphic unit—differs substantially from most interpretations of the Johns Valley at this location (e.g., Briggs and others, 1975; Stone, 1978).

RECENT STUDIES

The most recent study of this outcrop of Johns Valley Shale is by McDonald (1986), who measured the section here and interpreted it in terms of the Mutti and Ricci Lucchi (1978) turbidite facies classification scheme. McDonald (1986, p. 121) suggested the boulder conglomerate is lithofacies A (arenaceous-conglomeratic) surrounded above and below by lithofacies F (chaotic) debris-flow deposits. Most of the remainder of the outcrop consists of lithofacies C (arenaceous-pelitic) and D (pelitic-arenaceous) turbidites. The presence of exotic boulders in lithofacies A and F suggests they are related; both probably are mass-flow deposits, distinguished pri-



Figure 70. "Boulder bed" (of Harlton, 1938, fig. 19) in Johns Valley Shale at Stop 19. Map case on left side of outcrop is ~1 ft^2.

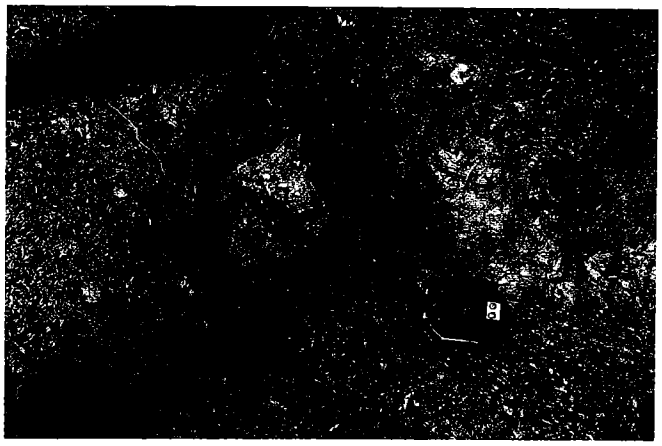


Figure 71. Limestone cobble in "cobbly shale" at Stop 19. Shale adjacent to cobble appears deformed around cobble; this probably was caused by differential compaction of soft mud around hard clast. Brunton compass for scale.

marily by their sediment-support mechanism during transport (Nardin and others, 1979). The bouldery shale is a debris-flow deposit and the boulder conglomerate is a grain-flow deposit.

In preparation for this field trip, I (NHS) examined this outcrop (yet again) and several of my observations and interpretations are presented below. For simplicity, "boulder bed" (Fig. 70) refers to that shown at "B" on Figure 69A. The adjacent shales are called "cobbly shales" and the interbedded sandstones and shales to the south are referred to as turbidites.

1. There is no evidence that the boulder bed is folded as pictured by Harlton (1938, fig. 19) (Fig. 69A). Similarly, there is no evidence for the isoclinal folding near the south end of the outcrop where shown on Figure 69A. However, ~70 ft south of the boulder bed, adjacent and steeply dipping sandstone beds in the turbidites face in opposite directions (Fig. 69A, at A) and indicate an isoclinal syncline is present.

2. There is no evidence for the faults shown by Harlton (1938, fig. 19) (Fig. 69A). (However, it is possible that the exposure has degraded since Harlton's visit.) There are some faults subparallel to bedding as shown by dismembered (sheared) sandstone beds ~40 ft north of the boulder bed (Fig. 69A, at D).

3. Boulder bed no. 30 of Harlton (1938, fig. 19) (Fig. 69A, at C) was not observed.

4. Shale immediately adjacent to the boulder bed appears "squeezed" into the spaces between adjacent boulders. Shale has also been "squeezed" into the cracks in some large boulders ~170 ft north of the boulder bed. Shale adjacent to cobbles in the cobbly shale appears deformed around the cobbles (Fig. 71). These features probably were caused by burial and differential compaction of the incompetent shale around the competent cobbles and boulders.

5. The matrix between the boulders is shale, but the boulders are clast supported (Harlton, 1938; McDonald, 1986).

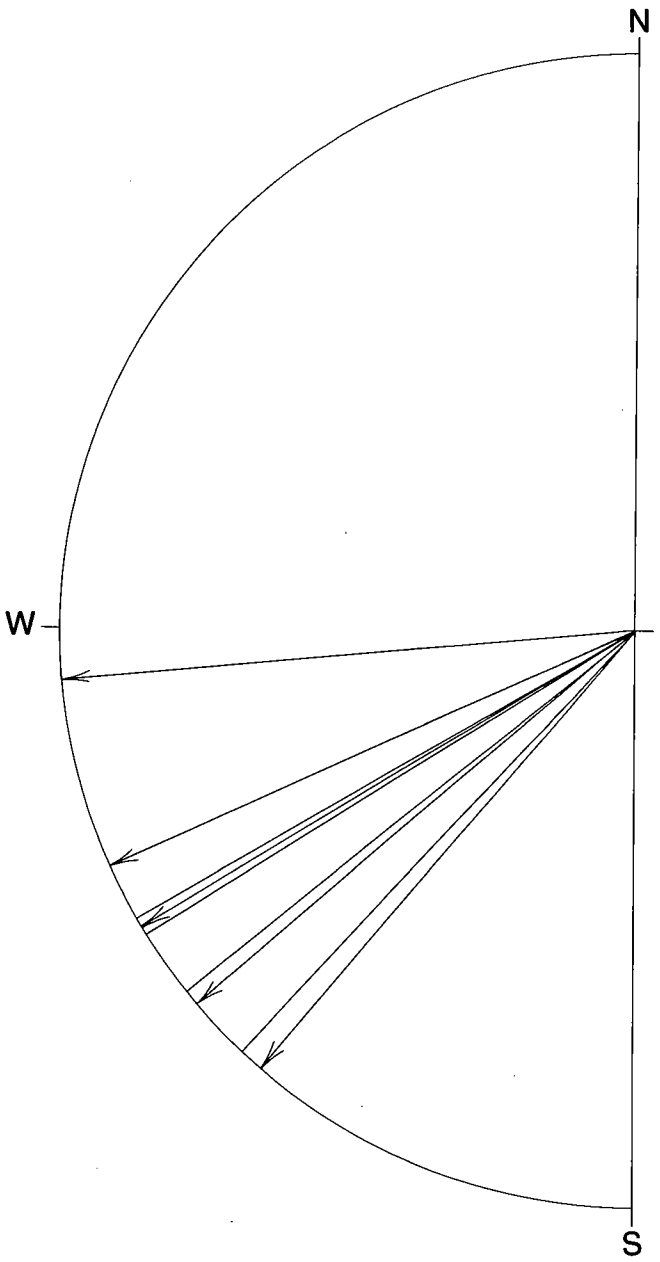


Figure 72. Paleocurrent directions in Johns Valley Shale at south end of exposure (Stop 19) (Fig. 69A). Lines with arrowheads are bearings of flute or bounce casts; lines without arrowheads are azimuths of groove casts.

6. The "pitted surfaces" noted by Harlton (1938) were not observed. Similarly, none of the boulders or cobbles in the boulder bed or the adjacent shales are striated, as has been reported elsewhere (see discussion at Stop 13).

7. Most of the boulders are rounded. The planar surfaces on some are the result of breakage along bedding planes and/or fractures within the boulders.

8. Most of the turbidite strata face north and are overturned. Paleocurrent directions are mostly to the west-southwest or southwest (Fig. 72).

28.0	0.3	Leave Big Cedar 7.5′ quadrangle, enter Page 7.5′ quadrangle.
30.1	2.1	Intersection of U.S. Highways 259 to right (south) and 59/270 straight ahead (east). Turn right (south) on U.S. 259.
30.7	0.6	Stop 20. Park on right side of road and walk back to outcrop.

STOP 20

Atoka Formation

Location: SE¼NW¼SE¼ sec. 17, T. 3 N., R. 26 E., along U.S. 259, Le Flore County

The strata exposed here (Fig. 73) are probably typical of most of the Atoka Formation in the frontal belt of the Ouachita Mountains. (Measured Section, Stop 20 gives a description of the strata.) The shales and mudstones that constitute ~80% of the Atoka are rarely so well exposed. Many features characteristic of turbidites are exposed in this outcrop: graded bedding, sole marks, and complete and incomplete Bouma sequences (Fig. 46). This outcrop is lithofacies D (pelitic-arenaceous) of Mutti and Ricci Lucchi (1978) and probably was deposited in an outer-fan environment. Paleocurrent directions are mostly to the west-northwest (Fig. 74).

Measured Section, Stop 20

Part of Atoka Formation

SE¼NW¼SE¼ sec. 17, T. 3 N., R. 26 E., along U.S. 259, Le Flore County.

Thickness (feet)

ATOKA FORMATION:

23. Shale, olive black (5Y2/1); silty 3.0
22. Sandstone, medium dark gray (N4); fine-grained. 4-in.-thick unstratified lower part under 2-in.-thick wavy-bedded Tc. Blocky. Base with very abundant trace fossils. Sandstone very quartzose, trace pyrite, white mica parallel to laminations 0.5
21. Sandstone, medium gray (N5); ungraded at base. Bottom part of sandstone blocky; upper part flaggy to platy. 9-in.-thick unstratified Ta(?) under 7-in.-thick very poorly stratified Tb under minimum 6-in.-thick Tc under Td (all thicknesses approximate). Base with large, bulbous load casts. Sandstone fine-grained, quartzose, trace pyrite. Overlain by 4.8 ft of shale 8.3
20. Shale, grayish black (N2); silty. Hackly 3.3
19. Sandstone. 8-in.-thick parallel-stratified Tb grades up to siltstone and shale 0.7
18. Sandstone, medium dark gray (N4); fine-grained. 6-in.-thick Tb under 4-in.-thick poorly exposed fissile siltstone (Td?). Tc absent. Very quartzose. Overlain by 3.1 ft of shale 3.9
17. Sandstone. 6-in.-thick parallel-stratified Tb. Load casts on base. Overlain by 4.6 ft of shale 5.1
16. Sandstone. 4-in.-thick Tc with very low inclination cross-stratification. Trace fossils on base. Overlain by 0.8 ft of shale 1.2
15. Shale, olive black (5Y2/1); silty. Weathered to pencil structure on surface; spheroidally weathered below surface 8.8

(Measured section offset ~40 ft to east.)

14. Sandstone, medium gray (N5); ungraded. 2-ft-8-in.-thick faintly parallel stratified Tb under 5-in.-thick wavy-bedded to low-inclination cross-stratified Tc under 14-in.-thick Td siltstone that grades up to shale. Base with groove (and rare flute) casts; most rake about –8°E. (paleocurrent S. 72° E. to N. 72° W.), but rake of other groove casts varies from 7° W. to 40° E. (paleocurrent directions range from S. 87° E.-N. 87° W. to S. 40° E.-N. 40° W.). Sandstone fine-grained; very quartzose, with trace pyrite 4.3
13. Shale with four 2–4-in.-thick blocky sandstones. Base of sandstones locally with load casts; tops poorly exposed though appear to grade into shale 6.6
12. Shale, olive black (5Y2/1), very silty. Well-developed pencil structure—small if weathered, larger if fresh.. 4.8
11. Sandstone, medium dark gray (N4); fine-grained. May be very slightly graded. Unstratified or very poorly parallel stratified lower part (Tb) under 8-in.-thick Tc under very thin Td. Base with large trace fossils (*Chondrites*) and groove casts (paleocurrent S. 59° E.-N. 59° W.). (Small fault with ~1.5-ft separation (west side down) offsets this bed just west of measured section) 2.4
10. Sandstone, medium gray (N5); very fine grained. 4-in.-thick parallel-stratified Tb under 2-in.-thick siltstone (Td?). Laminations marked by dark bands (organic material?) with trace pyrite. Quartzose. Overlain by 2.0 ft of shale 2.5
9. Sandstone, medium dark gray (N4); very fine grained. Cross-stratified Tc under very thin Td. Quartzose. Overlain by 1.1 ft of shale 1.3
8. Sandstone, medium gray (N5); very fine grained. 3-in.-thick cross-stratified Tc grades up to very thin Td that grades up to shale. Slightly irregular base reflects cross-stratification. Quartzose; white mica conspicuous on laminations. Overlain by 1.1 ft of shale 1.3
7. Sandstone. 5-in.-thick parallel-laminated Tb grades up to about 2-in.-thick cross-laminated Tc. Very irregular base caused by load casts. Overlain by 2.5 ft of shale 3.1
6. Shale, grayish black (N2); silty; "hackly" 3.3
5. Sandstone, medium dark gray (N4); very fine grained. Faintly parallel laminated (Tb); may grade up to unexposed and weathered Tc and/or Td and/or shale. Base with trace fossils and groove casts (paleocurrent S. 69° E.-N. 69° W.). Sandstone very quartzose 0.7
4. Shale, olive black (5Y2/1); very silty. Well-developed pencil structure 2.4
3. Sandstone, medium dark gray (N4); poorly exposed. Very fine grained. Faintly parallel stratified. Grades up to siltstone and shale. Quartzose, hard, abundant white mica parallel to laminations 0.3
2. Shale, grayish black (N2); very silty. Not very fissile; more typically "hackly" 14.0
1. Sandstone, medium gray (N5); slightly graded from medium-fine-grained at base to fine-grained at top. All but top 3.5 ft is poorly parallel-stratified (Tb); top cross-stratified and locally with dish-and-pillar structures (Tc). Base slightly irregular but largely covered; top covered. Sandstone quartzose; little matrix 2.3

Total thickness of section 84.1

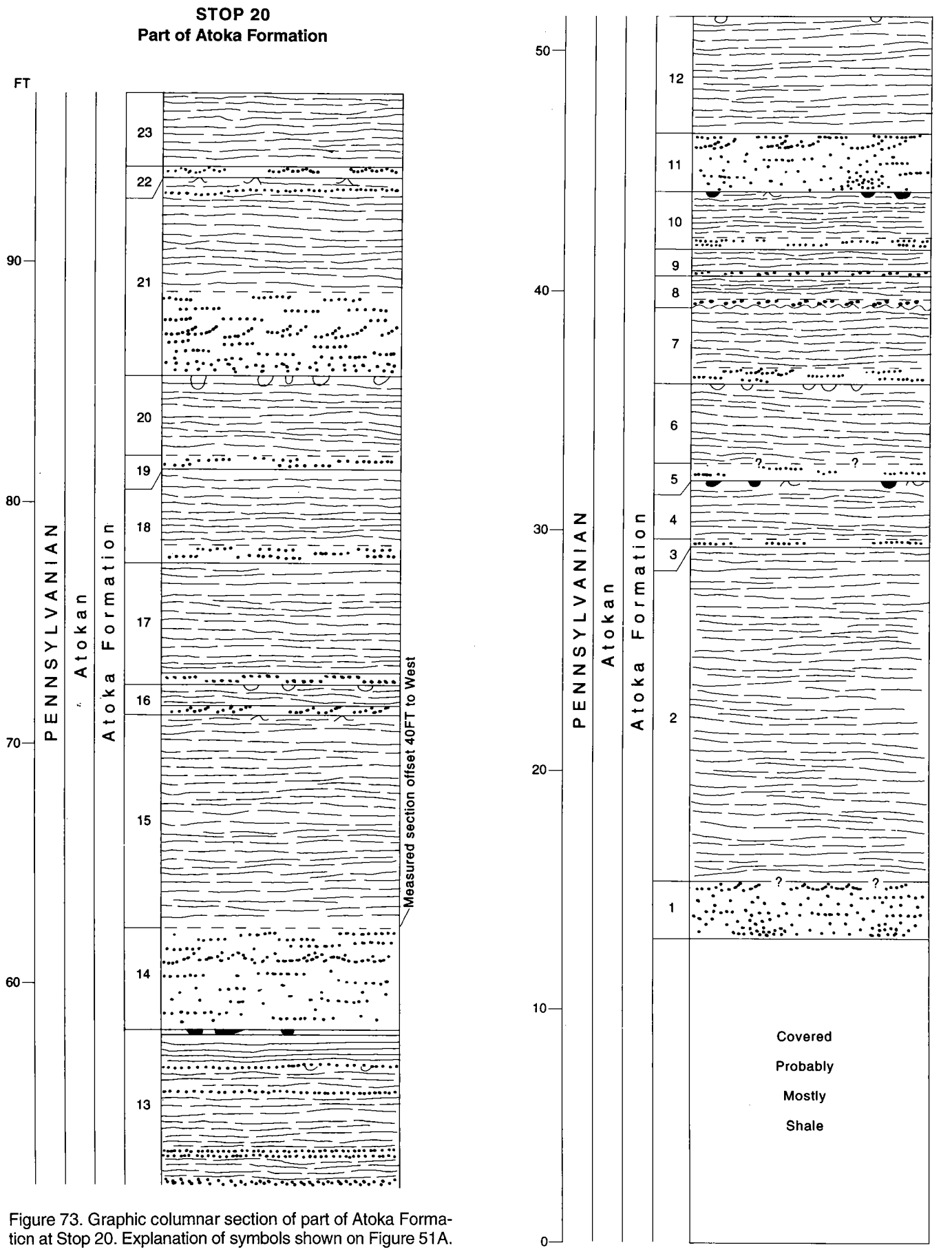


Figure 73. Graphic columnar section of part of Atoka Formation at Stop 20. Explanation of symbols shown on Figure 51A.

Seely (1963) places these strata ~6,500 ft above the base of the Atoka Formation on the north flank of the Spring Mountain syncline. The axis of the syncline approximately follows the creek on the right (north) side of the road ~1.5 mi west of here.

Another sequence of Atoka turbidites is well exposed in the stream bed to the north, just on the other side of the road. Approximately 120 ft of section is exposed; 20% is sandstone and the remainder is mostly shale. Paleocurrent directions based on sole marks are to the west–northwest, whereas current directions based on ripple cross-stratification (Bouma sequence Tc) are to the north-northwest (Fig. 74).

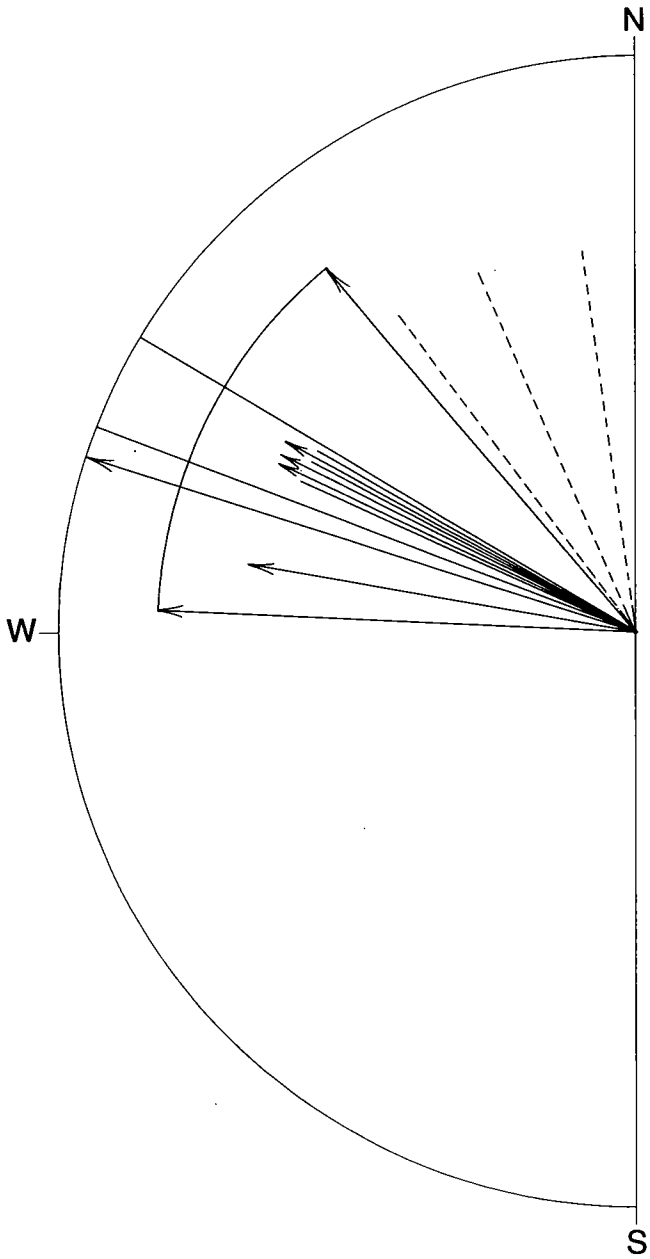


Figure 74. Paleocurrent directions in Atoka Formation at and near Stop 20. Lines with arrowheads are bearings and lines without arrowheads are azimuths of directional sole marks. Long lines are from Measured Section, Stop 20. Intermediate lines with arc show range of paleocurrent directions from Unit 14 of measured section. Short lines are from sandstone beds in stream north of road cut; solid lines are based on sole marks and dashed lines are based on ripples in Tc units.

32.2	1.5	Creek on north side of road marks the axis of the Spring Mountain syncline.
32.4	0.2	Approximate top of measured section of Atoka Formation (Hart, 1963, p. 71–76).

ATOKA FORMATION, SPRING MOUNTAIN SYNCLINE

Beginning here, Hart (1963) measured ~5,600 ft of Atoka Formation turbidites on the south flank of the Spring Mountain syncline. Of particular interest are the following observations made by Hart (1963, p. 71–76):

1) Fossil molds in intervals 220, 218, 206, 211, 149, and 147. (One of us [NHS] attempted to relocate Hart's fossil-mold sandstones, but was able to identify only nos. 220 and 147.) Our experience elsewhere in the frontal belt is that fossil molds are present only at the base of the Atoka Formation (Hart nos. 149 and 147).

2) A siliceous shale(?) (Hart's interval no. 167) ~900 ft above the base of the formation. Hendricks and Averitt (Tulsa Geological Society, 1947, p. 32) described a siliceous shale ~160 ft above the base of the Atoka Formation at the Hairpin Curve (SE¼ sec. 3, T. 3 N., R. 19 E.). A siliceous shale is present near the base of the Atoka Formation in the C sec. 5, T. 3 N., R. 18 E., in the Higgins 7.5′ quadrangle. Possibly due to the degraded condition of the outcrop compared to when Hart described it, Hart's siliceous shale could not be identified.

3) A general fining-upward of the sandstones in the Atoka Formation. Nearly all the sandstones above Hart's interval 163 are fine grained or very fine grained, whereas many of the sandstones below that interval are fine to medium grained. This feature was noted also by Suneson (1991) and Hemish and Suneson (1991) in the Blackjack Ridge and Leflore Southeast 7.5′ quadrangles to the west.

32.9	0.5	Leave Page 7.5′ quadrangle, enter Big Cedar 7.5′ quadrangle.
33.8	0.9	McDonald (1986) also measured several sections of Atoka Formation on the left (south) side of the highway. Mutti and Ricci Lucchi (1978) lithofacies B (arenaceous) and E (pelitic-arenaceous with percent sandstone > shale) are most common and G (hemipelagic and pelagic) less so. Overall, McDonald (1986) considered this to be a sandstone-rich part of the Atoka Formation.
34.3	0.5	Intersection U.S. 259 and State Highway 1 (Talimena Drive). Turn right onto on-ramp to Talimena Drive.
34.5	0.2	Talimena Drive. Turn left (east).
35.2	0.7	Big Cedar Vista. Good view of Kiamichi Mountain (north flank of Lynn Mountain syncline), Rich Mountain, and Simmons Mountain. The plaque notes that President John F. Kennedy was in Big

		Cedar to dedicate U.S. 259 across the Ouachita Mountains.
35.3	0.1	Forest Service Road 6007 to left (east).
35.7	0.4	Excellent outcrops of northeast-facing Atoka Formation sandstones on south flank of Spring Mountain syncline. A wide variety of sole marks and soft-sediment-deformation features occur in these strata. Paleocurrent directions based on sole marks are mostly south-southeast to north-northwest; some indicate a current direction to the west-northwest.
36.0	0.3	Beautiful view of Rich Mountain directly in front.
36.2	0.2	Forest Service Road 6029-2 to right (west).
36.8	0.6	Leave Big Cedar 7.5′ quadrangle, enter Page 7.5′ quadrangle. Entrance to Robert S. Kerr Arboretum. This valley is underlain by Johns Valley Shale.
37.3	0.5	Leave Page 7.5′ quadrangle, enter Big Cedar 7.5′ quadrangle.
37.4	0.1	"Rock glacier" on left.
37.7	0.3	Several large outcrops of Stanley Group strata (Moyers Formation) on left. The following features are worth noting: (1) most of the outcrops consist of extremely fissile, "flaky" shale; (2) the less-abundant sandstone occurs in beds ranging from 1 in. to 3 ft thick, as thin discontinuous pods and lenses, and as large (1.5-ft diameter) isolated boulders; rounded grains are conspicuous in some of the sandstones; (3) sole marks on some of the sandstone beds indicate a N. 85° E. to S. 85° W. paleocurrent direction.
38.0	0.3	Stop 21. Just north of hairpin curve on Talimena Drive.

STOP 21

Stanley/Jackfork Contact and Chickasaw Creek Siliceous Shale

Location: NW¼SE¼SW¼ sec. 36, T. 3 N., R. 25 E., Le Flore County

The Chickasaw Creek Siliceous Shale was defined as the youngest formation in the Stanley Group by Harlton (1938) based on his studies in the western Ouachita Mountains in Oklahoma. He identified two type localities in Atoka County, ~75 mi west-southwest of this outcrop. Earlier workers had noted siliceous shale and chert beds in the Stanley, but none had recognized their apparent widespread character or usefulness as strati-

Measured Section, Stop 21

Stanley/Jackfork contact and Chickasaw Creek Siliceous Shale

NW¼SE¼SW¼ sec. 36, T. 3 N., R. 25 E.

Thickness (feet)

JACKFORK GROUP:

WILDHORSE MOUNTAIN FORMATION:

32. Sandstone, dark gray (N3), fine-grained, 7.5 in. thick. Tbcd, grades into overlying shale. Well-developed cross-stratification that is sheared locally (soft-sediment deformation) in general direction of paleocurrent. Base with groove casts (paleocurrent N. 65° E.-S. 65° W.). Hard, quartzose, conspicuous mica parallel to laminations. Overlain by interbedded shale and 1-in.-thick siltstone and sandstone beds similar to 31 ..... 3.1
31. Shale, poorly exposed, olive black (5Y2/1). Interbedded with ~20% 0.5- to 3-in.-thick siltstone (Td) and sandstone (Tc) beds 8.4
30. Interbedded sandstone and shale. Four sandstones. Lowest sandstone—dark gray (N3), fine-grained, parallel-stratified (Tb) under thin cross-stratified (Tc) sandstone, 13 in. thick. Thin cross-stratified (Tc) interval locally in middle of Tb. Base with groove and rare flute casts (paleocurrent N. 63° E. to S. 63° W.). Hard, blocky weathering, very continuous over entire outcrop. Next highest sandstone—Tc with abundant dish-and-pillar structures, 3 in. thick. Next highest sandstone—medium dark gray (N4), fine-grained, very faintly parallel stratified (Tb), 7 in. thick. Base with groove casts (paleocurrent N. 72° E.-S. 72° W.). Quartzose, rare feldspar. Upper sandstone—unstratified, blocky weathering, hard, 6 in. thick 3.2
29. Shale, mostly covered, olive black (5Y2/1), silty. Possibly some thin weathered sandstone or siltstone beds ... 7.4
28. Interbedded sandstone and shale. Three sandstones. Lowest sandstone—unstratified, base with groove casts (paleocurrent N. 65° E.-S. 65° W.), blocky weathering, 6 in. thick. Overlain by shale. Middle sandstone—faintly parallel stratified bottom part (Tb) grades upward to well-stratified upper part (Td), top grades into overlying shale; base with well-developed groove casts (paleocurrent N. 67° E.-S. 67° W.). 7 in. thick. Upper sandstone—dark gray (N3), fine-grained, Tbcd, not graded at base, 13 in. thick. Base planar with abundant prod, bounce, and rare flute casts (paleocurrent N. 60°–71° E. to S. 60°–71° W.). Conspicuous feldspar 2.4
27. Sandstone, medium dark gray (N4). Very faintly parallel stratified, ungraded, 11 in. thick. Base with abundant load and groove casts (paleocurrent N. 71° E.-S. 71° W.). Hard, relatively clean. Resembles typical Jackfork sandstone. Overlain by shale that is mostly covered 3.6
26. Sandstone. Two beds separated by 1-in.-thick shale. Lower sandstone—medium dark gray (N4), fine-grained, 2 ft 4 in. thick. Massive, unstratified lower part (Ta?) under faintly cross stratified middle part (Tc) under blocky, unstratified upper part. Ungraded. Base very planar with groove casts (paleocurrent N. 75° E.-S. 75° W.). Little matrix. Upper sandstone—massive, unstratified, blocky-weathering, soft, 3 ft 1 in. thick. Overlain by poorly exposed shale 6.0
25. Convoluted sandstone and shale. Sandstone beds as thick as 4 ft terminated abruptly (fault?), pinch out, show much thickening and thinning. Some beds

with strike oblique to overall strike of sequence. Locally, interbedded sandstone and shale slump-folded and truncated. All sandstone beds with very irregular bases caused by loading. All structures suggest syndepositional and immediately post-depositional slumping 15.0

24. Sandstone, medium gray (N5), medium- to fine-grained. Thickness variable due to soft-sediment deformation and very large load casts on base. Trace small carbonized plant debris 2.6
23. Sandstone, medium dark gray (N4), fine-grained. Faintly parallel stratified (Tb), blocky weathering, 3 ft 3 in. thick. Grades up to very silty, fissile sandstone (Td). Mica parallel to laminations. Here, overlain by about 1-ft-thick shale; immediately to east, shale absent and 24 directly overlies 23 ..... 4.3
22. Sandstone, medium dark gray (N4), very fine grained, 7 in. thick. Probably Tbcd, but stratification very faint. Abundant white mica parallel to laminations. Overlain by shale with thin sandstones 6.3
21. Shale. With one 2-in.-thick Tc sandstone, locally showing soft-sediment deformation 2.7
20. Sandstone, olive gray (5Y4/1), fine-grained. Very faintly cross stratified (Tc); upper part parallel-stratified (Td). Base very planar. Grades into overlying shale. White mica parallel to laminations 0.8
19. Sandstone. Two beds, separated by a 1-in.-thick shale, 1 ft 7 in. thick. Lower part of lower sandstone faintly parallel stratified (Tb); upper part cross-stratified (Tc). Very irregular base caused by loading; top of lower sandstone wavy, shale draped over waves, and thin upper sandstone also draped. Overlain by shale.. 2.1
18. Shale, olive black (5Y2/1), no silt. Fissile. With ~5% 0.5- to 2-in.-thick Tc and Td sandstones, locally with small trace fossils on base 4.6

(Measured section offset 225 ft to west-southwest.)

17. Sandstone, dark gray (N3). Faintly parallel stratified throughout with common dish-and-pillar structures. Load casts common on base. Irregular blocky weathering. Abundant feldspar. Divides into two beds separated by 5-in.-thick shale at road level 2.8
16. Shale, olive gray (5Y3/2), slightly silty. Locally fissile and with pencil structure. Includes rare sandstones 0.5 to 3 in. thick, cross-stratified (Tc) and parallel-stratified (Td) 4.3
15. Sandstone, medium dark gray (N4), fine-grained. Faintly parallel stratified (Tb). Base planar, no features. Top grades into overlying shale. Thinner at road level 2.3

STANLEY GROUP:

CHICKASAW CREEK SILICEOUS SHALE:

14. Interbedded siliceous shale and shale. Siliceous shale mostly grayish black (N2) to dark gray (N3). Extremely well developed parallel laminations expressed as slightly different colors. Pods or areas of white or pinkish rounded specks. Blocky weathering, limonite-stained fractures, locally subconchoidal fractures. Siliceous shale locally grades into shale. Shale olive black (5Y2/1), fissile, not silty 3.1

STANLEY GROUP:

MOYERS FORMATION:

13. Shale. With ~15% 0.5- to 1-in.-thick siltstones and sandstones, probably Td 3.7
12. Sandstone, medium dark gray (N4), fine-grained. Unstratified. Sharp, planar base, equally sharp top. Blocky weathering. Conspicuous feldspar and rock fragments 2.0
11. Shale, grayish black (N2). Fissile, locally with good pencil structure. Rare 0.5- to 1-in.-thick siltstones and sandstones 5.7
10. Sandstone, medium gray (N5). Unstratified, ungraded. Extremely planar base, sharp top. Thicker at road level 1.3
9. Sandstone, medium gray (N5), fine-grained, 3 in. thick. Well-developed cross-stratification (Tc), possible 0.5-in.-thick Td at top; does not grade into overlying shale. Very continuous. Blocky weathering, limonite stains on fractures; weathers similar to siliceous shales. Quartzose, little matrix. Overlain by shale 0.8
8. Shale with ~30% sandstone and siltstone. Sandstone and siltstone beds 1–6 in. thick, parallel-stratified (Td), very continuous. Locally with shale rip-up clasts 8.3
7. Sandstone, dark gray (N3), fine-grained. Unstratified, not graded. Base slightly irregular due to loading or scouring. Faint groove casts indicate N. 59° E.-S. 59° W. paleocurrent. Blocky weathering. Bed thins to 2 ft 6 in. at road level 3.3
6. Shale, olive black (5Y2/1), not silty. With ~20% 0.5- to 1-in.-thick extremely fine grained sandstones (Td) that pinch and swell and locally lens out 3.1
5. Interbedded sandstone and shale. Highest sandstone—fine- to medium-grained, light olive gray (5Y6/1), 13 in. thick. Graded, unstratified lower part (Ta), upper 6 in. very convoluted (dewatering structures) (Tb). Base with small load casts, rare small trace fossils. Sharp upper contact. Blocky weathering. Next lowest sandstone—medium light gray (N6), fine-grained, 13 in. thick. Faintly parallel stratified (Tb); upper 3 in. cross-stratified (Tc) with slight soft-sediment deformation features. Base slightly graded with rare but conspicuous larger grains; also 1 in. load casts. Sharp upper contact. Next lowest sandstone—medium bluish gray (5B5/1), fine-grained, 8 in. thick. Faintly parallel stratified with slightly convoluted upper 3 in. (Tb). Not graded. Small sole marks (load casts or trace fossils). Sharp upper contact. Feldspar. Lowest sandstone—medium gray (N5), fine-grained, silty, 9 in. thick. Mostly unstratified (Ta?), though top 4 in. slightly convoluted. Not graded. Base with groove casts (paleocurrent N. 85° E.-S. 85° W.); sharp upper contact. Feldspar. 5.0
4. Shale, olive gray (5Y4/1); not silty. Rare 0.5- to 1-in.-thick, continuous, Td sandstones; medium dark gray (N4), very fine grained. Some pinch-and-swell 7.3
3. Sandstone, moderate olive brown (5Y4/4), fine-grained with silty to muddy matrix. Wavy stratification. Very weathered; upper and lower parts show platy weathering; middle flaggy. Possibly Tb 2.5
2. Shale with minor sandstone. Sandstone occurs as 1–2-in.-thick beds and discontinuous lenses that are folded locally about axial surfaces parallel to bedding. Bivergent sense of shear 6.0
1. Shale with minor sandstone, 7 ft 7 in. thick. Shale locally silty, varies from very to slightly fissile. Locally moderately convoluted. Lower 4 ft with several 1-in.-thick contorted sandstone dikes. Uncommon 1–3-in.-thick discontinuous Tc lenses. Some sandstone beds continuous for 1.5 ft. Overlain by ~8.8 ft of covered interval, probably shale 16.4

Covered interval at base 2.7

Total thickness of section 153.1

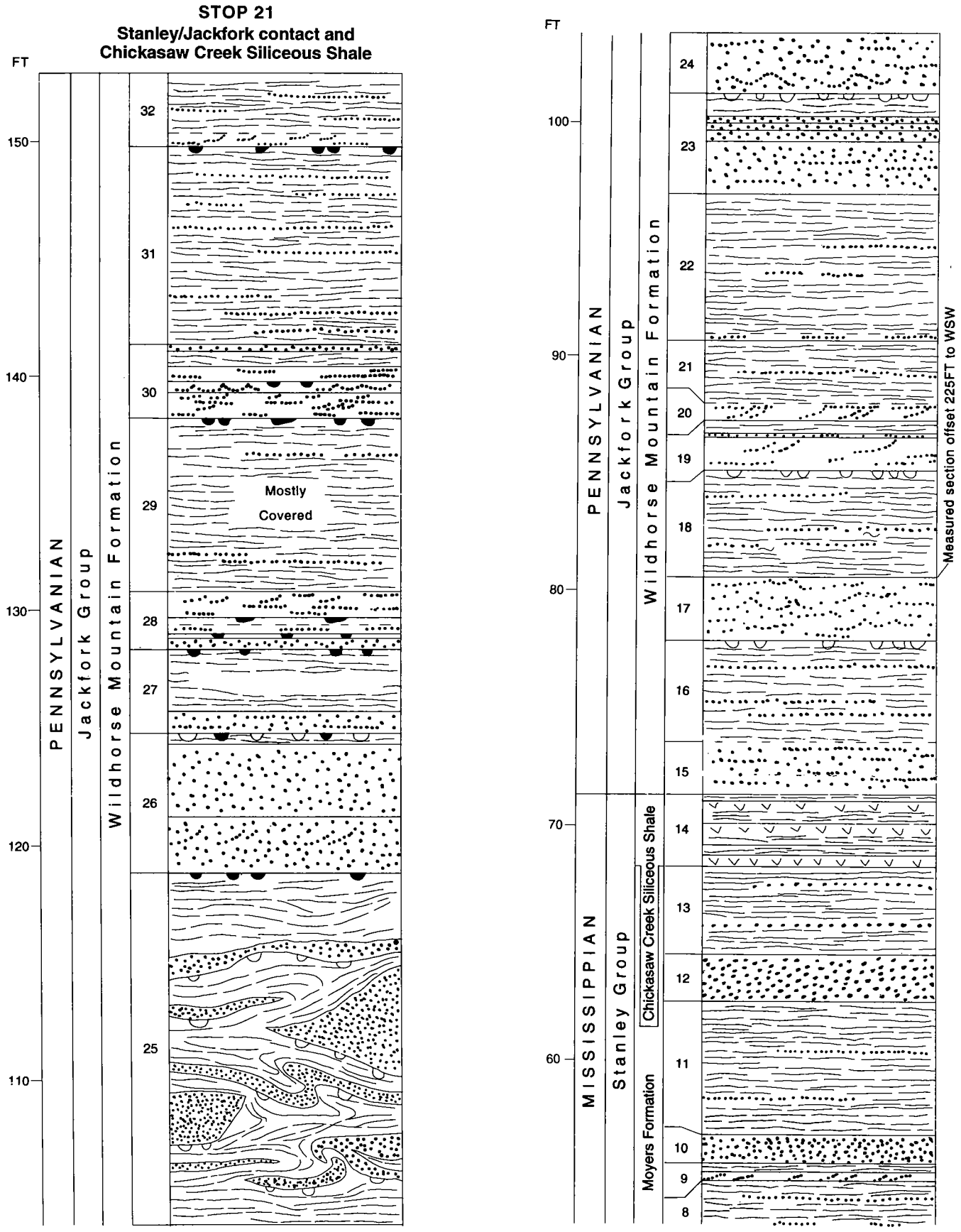


Figure 75 *(this page and opposite)*. Graphic columnar section of Stanley Group/ Jackfork Group contact and Chickasaw Creek Siliceous Shale at Stop 21. Explanation of symbols shown on Figure 51A.

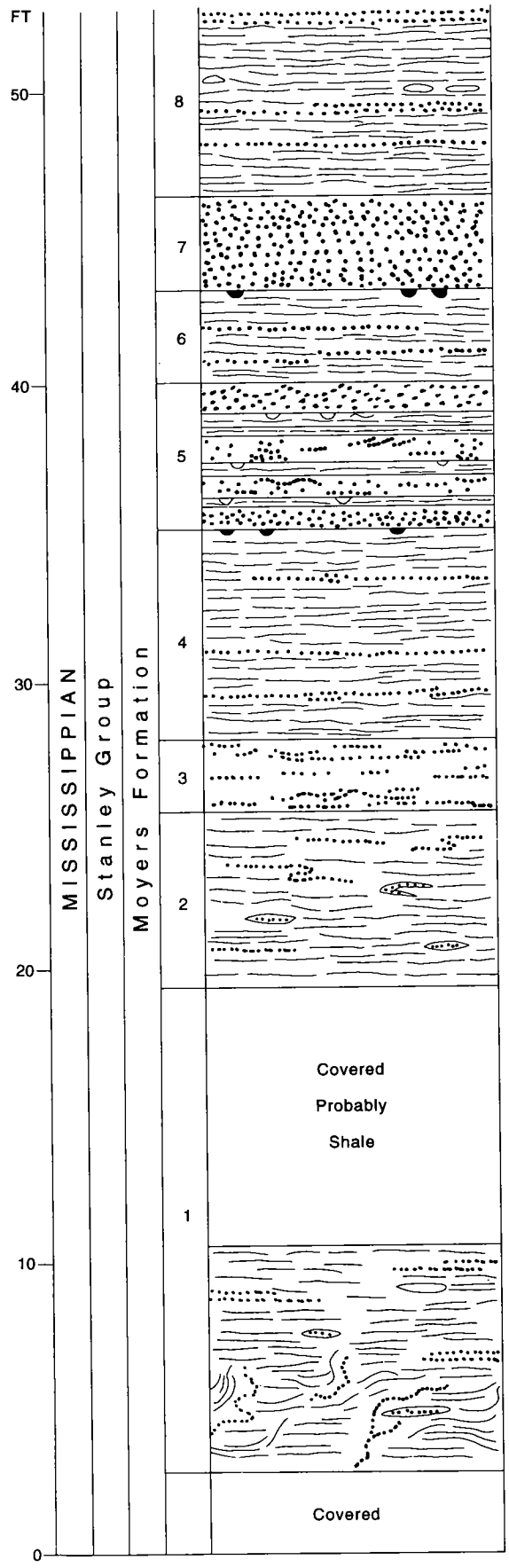


Figure 75 *(continued).*

graphic markers. Pitt and others (1982) mapped two siliceous shales (Battiest Chert Member of the Tenmile Creek Formation, Schoolhouse Chert Member of the Moyers Formation) throughout Pushmataha County. Our limited mapping of the Stanley suggests that many of the Stanley siliceous shales are more discontinuous than shown by Pitt and others (1982). This agrees with Piatt (1962), who mapped southeastern Pushmataha County: he did not find the Battiest Chert (p. 16) and noted that the Schoolhouse Chert was absent to (at best) discontinuous (p. 21).

Unlike the siliceous shales in the Tenmile Creek and Moyers Formations, those in the Chickasaw Creek appear to be present throughout most of the Ouachita Mountains in Oklahoma and Arkansas. Hart (1963) and Pitt and others (1982) reviewed where the Chickasaw Creek has been recognized. Of particular interest is the wide range of thicknesses reported for the formation—from ~400 ft in Atoka County (Ray, 1960) to 34(?) ft in Polk County, Arkansas (Seely, 1963). This variation is probably due to inconsistent determinations of the top and bottom contacts of the formation (Hart, 1963; Seely, 1963). The Chickasaw Creek Siliceous Shale shown on Figure 75 is ~3 ft thick and includes only chert and siliceous-shale beds. (Measured Section, Stop 21 describes the units in Fig. 75.) The blocky-weathering beds at this outcrop are similar to the highest siliceous shale or chert in the Chickasaw Creek recognized elsewhere in the Ouachita Mountains.

At this locality (Fig. 76A,B), the ~3-ft-thick siliceous shale (Unit 14, Measured Section, Stop 21) is the only siliceous shale separating the Stanley Group strata described at Mile 37.7, Day Two, and Jackfork Group sandstones just up the road from this stop. Hart (1963, p. 18–19) identified two siliceous shale beds ~75 ft apart, with an intervening 8- to 10-ft-thick tuffaceous sandstone, on the south flank of the Simmons Mountain syncline ~2 mi to the southwest; the second siliceous shale and the tuffaceous sandstone are not present here. Similarly, chert-pebble conglomerate beds, noted elsewhere in the Chickasaw Creek Siliceous Shale (e.g., Harlton, 1938; Pitt and others, 1982), are not present here.

The age of the Chickasaw Creek Siliceous Shale has not been determined directly, despite abundant radiolarians and conodonts (Harlton, 1938). There is no evidence for unconformities within the Stanley Group (Morris, 1989, p. 592) and the Chickasaw Creek is conformably overlain by the Jackfork Group (Hart, 1963; Seely, 1963). The upper part of the Stanley below the Chickasaw Creek is Chesterian (Gordon and Stone, 1977; Ethington and others, 1989). Pitkin Limestone (Chesterian) and Fayetteville Shale (Chesterian) boulders in the Chickasaw Creek ~35 mi west of Little Rock indicate it is, at the oldest, upper Chesterian (Gordon and Stone, 1977). Plant fossils from the lower part of the Jackfork Group in the Buffalo Mountain syncline just west of Talihina are Chesterian (White, 1937; Gordon and Stone, 1977), and marine invertebrates from several localities in Arkansas are early Morrowan (Gordon and Stone, 1977). In summary, the age of the Chickasaw Creek Siliceous Shale is equivocal because most, if not all, of the fossils in it and in the immediately overlying Jackfork are allochthonous and establish maximum ages only. The top of the Chicka-

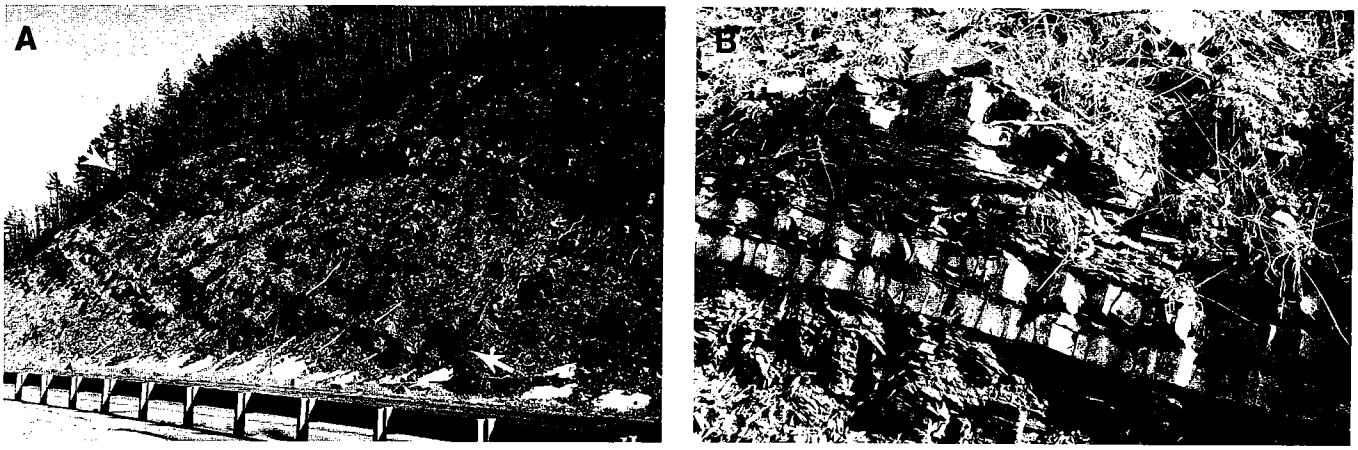


Figure 76. *A*—Road cut along Talimena Drive exposing Chickasaw Creek Siliceous Shale (arrows). *B*—Blocky-weathering siliceous shale beds in Chickasaw Creek Siliceous Shale. Pencil in center-right is 6 in. long.

saw Creek in Figure 75 marks the Mississippian/Pennsylvanian boundary, but it is possible that the boundary lies somewhere in the lowermost Jackfork.

41.5 3.5 Retrace route to intersection of U.S. 259 and Talimena Drive.

For most of its length between here and its western end (Mile 61.9, Day Two), Talimena Drive follows the top of Winding Stair Mountain. It is largely within strata of the Wildhorse Mountain Formation of the Jackfork Group (Hart, 1963, pl. 1).

42.6 1.1 Good outcrop of stacked sandstones in Jackfork Group on left (south) (Fig. 77).

42.9 0.3 Good outcrop of shale in Jackfork Group strata.

43.3 0.4 Entrance to Emerald Vista Campground and Picnic Area.

43.4 0.1 Forest Service road to left (south).

44.5 1.1 Cross Ouachita Mountains Trail. The trail is 186 mi long and extends from near Little Rock, Arkansas, to Talimena State Park in Oklahoma. On the west side of the creek ~300 ft below this saddle on the north side of Winding Stair Mountain (100 ft FEL, 600 ft FNL, sec. 20, T. 3 N., R. 25 E.) are two small asphaltite pits. Cardott and others (1993) determined that the material is epiimpsonite, and Finkelman and Cardott (1993) report an unusually high vanadium content. The asphaltite was mined and used in blacksmiths' forges prior to 1936 (Beach, 1945).

45.1 0.6 Forest Service Road 6014 to Cedar Lake on right (north).

45.6 0.5 Shawnee Vista. Good view to the north of Cedar Lake, Cavanal Mountain, Heavener and the Heavener road cut (Stop 18), Sugarloaf Mountain, Poteau Mountain, Walker Mountain, and Shawnee Ridges. Good view also of areas of planted pine trees.

46.7 1.1 Entrance to Horse Thief Spring Picnic Area. The spring was frequented in the late 1800s by a number of horse thieves and bank and train robbers, including the illustrious (infamous?) Belle Starr.

47.1 0.4 Cedar Lake Vista. Good view to the north of Cavanal Mountain, Heavener, Cedar Lake, Sugarloaf Mountain, Poteau Mountain, Walker Mountain, and Shawnee Ridges.



Figure 77. Outcrop of stacked sandstones in Jackfork Group at Mile 42.6, Day Two, Talimena Drive.

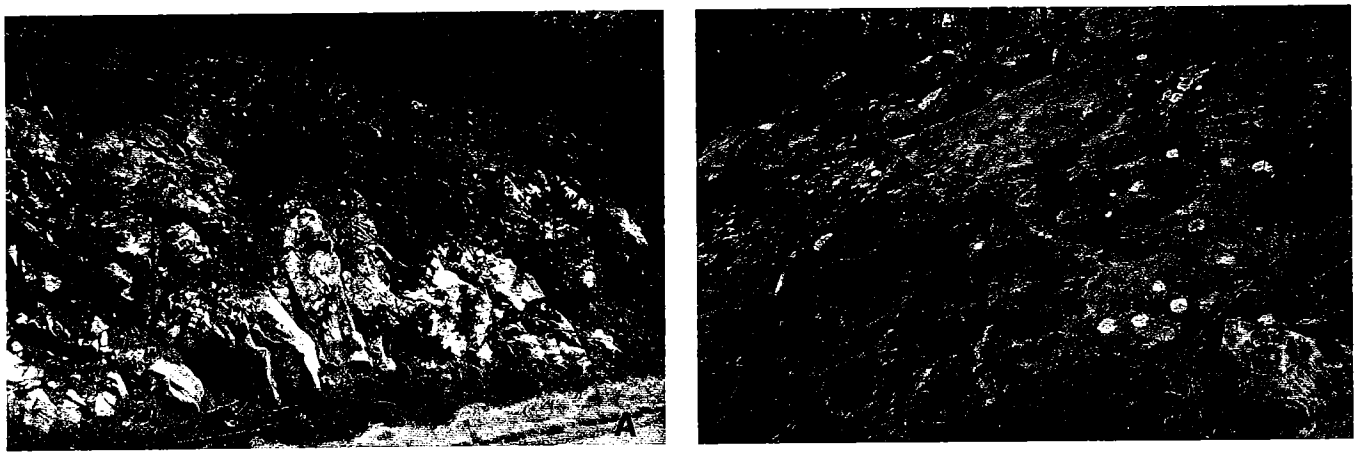


Figure 78. *A*—Blocky, unstratified, jointed sandstones in Jackfork Group at Mile 47.5, Day Two, Talimena Drive. *B*—Top of sandstone bed in Jackfork Group showing interference ripples in Tc unit (Mile 47.7, Day Two, Talimena Drive).

47.7	0.6	Good outcrop of Jackfork sandstones (Figs. 78A,B).
48.3	0.6	Leave Big Cedar 7.5′ quadrangle, enter Muse 7.5′ quadrangle.
48.9	0.6	Mostly grassed-over outcrop of Jackfork sandstones on right (north). Blocks of sandstone breccia (probably fault breccia) are conspicuous.
50.2	1.3	Cross Ouachita Mountains Trail.
51.0	0.8	Good outcrop of Jackfork shales on right (north). The shale is light olive gray (5Y 5/2), not silty, flaky weathering, and it is interbedded with very thin siltstone beds.
51.3	0.3	Forest Service Road 6022 to Billy Creek Campground and Muse on left (south).
51.5	0.2	Lenox Vista. Good view to south of Lenox Ridge, Kiamichi River Valley, Kiamichi Mountain, and Lenox Townsite. Lenox Community and Mission Church were established in 1852 by Pastor S. L. Hobbs. Lenox was named after the New Hampshire hometown of Hobbs' wife, Cornelia Sears. Hobbs graduated from Phillips Academy and the Medical Institute of New Hampshire and was sent to the Choctaw Nation by the American Board of Foreign Missions.
51.6	0.1	Good outcrop of thickening-upward amalgamated sandstone beds (at base) and interbedded sandstone and shale (towards top). Many of the sandstone beds are characterized by excellent trace fossils and sole marks on the base and wavy-bedded tops. The wavy tops represent the Tc (Bouma sequence) interval (Fig. 46); the Td interval is thin, absent, or eroded.
51.7	0.1	Leave Muse 7.5′ quadrangle, enter Leflore Southeast 7.5′ quadrangle.
52.9	1.2	Stop 22.

STOP 22

Sugarloaf Vista

Location: C SL SE¼SW¼ sec. 6, T. 3 N., R. 24 E., Le Flore County

This stop is primarily a camera stop. If the weather is clear, the overlook provides a good view to the north of the San Bois Mountains, Cavanal Mountain, Blue Mountain, Sugarloaf Mountain, Poteau Mountain, and Holson Valley.

53.1	0.2	Excellent outcrop of shale in Jackfork on left (south). The shale varies in color from light greenish gray (5GY8/1) to medium dark gray (N4) to light olive gray (5Y5/2). Interbedded sandstone and siltstone beds vary from soft and friable to hard. Flute casts indicate a N. 85° E. to S. 85° W. paleocurrent direction.
53.6	0.5	Large outcrop of thin, mostly Tc Jackfork sandstones on left (south). Trace fossils, tool marks, and groove casts are well preserved on the base of many beds. Some beds are wavy-bedded and have a very thin Td interval draped over a wavy Tc interval.
54.7	1.1	Excellent outcrop of shale (Fig. 79) with cobble-size, elliptical, siderite-cemented concretions.
55.9	1.2	Deadman Gap. Cross Ouachita Mountains Trail.

HISTORY OF LOGGING IN THE OKLAHOMA OUACHITAS

By the end of the 19th century, most of the great forests east of the Rocky Mountains had been cut for their timber. Before the Civil War, New England had been logged out; after the Civil War, sawmillers moved into the Great Lakes states and the southeastern U.S. There the industry developed its "cut out and get out" philosophy of timber harvesting—moving men, towns, and equipment about every 20 years as areas became logged out.

At the turn of the century, the last remaining virgin shortleaf-pine forest east of the Rockies available to land buyers and speculators was in the Ouachita Mountains of west-central Arkansas and southeastern Oklahoma. Up to this time, the area had been largely overlooked because the steep topography made logging more difficult than in the flatlands to the south. Three factors, however, contributed to the opening of the Ouachitas to logging operations: (1) the area was poor farming land and the sparse rural population, barely able to eke out a living in most cases, welcomed the timber companies; (2) trunk-line railroads had been extended into the area (e.g., Kansas City Southern between Heavener, Oklahoma, and Mena, Arkansas, in 1896); and (3) the nation needed timber.

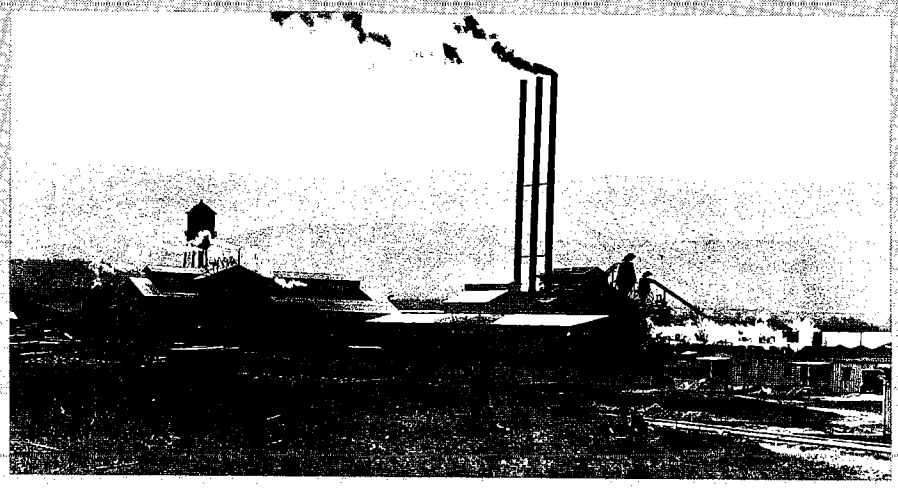
The two major lumber manufacturers in the Ouachitas in the first half of the 20th century were the Caddo River Lumber Company and the Dierks Lumber and Coal Company. In addition, there were several much smaller companies (for example, the Buschow Lumber Company) (see Stapp, Oklahoma, p. 103). Caddo River, run by Thomas W. Rosborough, operated entirely in Arkansas. Of particular interest to this guidebook is Dierks, which built and operated the Pine Valley mill near the present village of Muse along the north side of the Kiamichi Valley.

The Dierks company grew out of a string of retail lumber and coal yards in Iowa and Nebraska developed by Herman Dierks and his three brothers. In 1895, the brothers incorporated the Dierks Lumber and Coal Company and, in 1896, moved the company's headquarters to Kansas City. In 1897, the brothers bought a small planing mill at Petros, near present-day Heavener. They soon realized that the mill was at the northern edge of the pine belt. Timber was scarce, so in 1900 they decided to buy some land and a larger sawmill at De Queen, Arkansas. In 1902, Herman Dierks moved to De Queen to manage the sawmill and nearby logging operations. He soon began eyeing the vast timber resources of the Indian Territory. In 1910, Dierks formed the Choctaw Lumber Company and opened a double-band mill at the new Dierks town of Bismark (now Wright City), Oklahoma. At the same time, the Dierks brothers began building logging railroads to the north into the Ouachitas and to the east where, in 1912, they opened two mills (one for pine, the other for hardwood) at Broken Bow. By 1918, "there were three big pine mills to cut the timber that Dierks had begun to buy in Arkansas and Oklahoma soon after 1900" (p. 75). (The De Queen sawmill, which burned down in 1909, had been replaced by a large new mill at Dierks, Arkansas.)

Throughout the 1920s, the Dierks companies continued to profit and to buy timberland in the Ouachitas. Plans were made for two new mill towns, each with its own double-band sawmill and logging railroads. Pine Valley Lumber Company in the upper Kiamichi Valley, Oklahoma, started production in 1927 and Mountain Pine, near Hot Springs, Arkansas, began in 1928. The mill at Pine Valley was originally the idea of Herman's son, Fred H. Dierks; he wanted a relatively modest-sized mill similar to that operated by the Buschow brothers near Stapp. The project, however, soon became a family enterprise, influenced by the general optimism of the mid-1920s and the profitable Dierks businesses. "The original plan developed into an elaborate design for a much larger mill, a twin to Mountain Pine. [But] everybody seemed to forget that Mountain Pine had access to more than a hundred thousand acres of timber including some of the finest pine in the Ouachitas...[and] Pine Valley had much less....Even so, the Pine Valley project was pushed along in tandem with Mountain Pine" (p. 78).

Sawmill describes the mill town and its people (p. 78–79):

> Pine Valley, at the end of a fifteen-mile railroad branching off from the Kansas City Southern, was developed as a self-contained town with commissary, drugstore, post office, barbershop, doctor's office, and school. A hotel, or boardinghouse, offered sixty-eight rooms, a spacious lobby, and a dining room—"the finest mill town hotel in the South," proclaimed a reporter for *The Dierks Industries Co-Operator*, a monthly newsletter for employees. Over one hundred houses, some of them architect-designed, comprised "the finest



The Pine Valley sawmill, built as "the finest of its kind in the world" (Smith, 1986, p. 81). (Frank J. Gibbs photograph courtesy of Forest Heritage Center, Broken Bow, Oklahoma.)

Editors' note: This brief summary of logging in the northern Ouachitas of Oklahoma is based on *Sawmill: The Story of Cutting the Last Great Virgin Forest East of the Rockies* by Kenneth L. Smith (1986).

company houses which we have seen in any of the Dierks Industries communities." A movie house, which also served as an auditorium and church, could seat five hundred, and had a stage with dressing rooms and an orchestra pit. The commissary offered the services of a tailor and would deliver ice made in Pine Valley's own ice plant. Pine Valley even had a jail, a concrete pillbox with two cells, one for whites and the other for blacks.

The sawmill was the most modern of any built during that time. All the machines were driven by electricity supplied by a 1,500-kilowatt turbine. The planing mill was floored with an acre of concrete instead of wood. The newsletter reporter wrote that "this plant is said to be the finest of its kind in the world. From this standpoint of efficient lay-out, there is no counterpart for Pine Valley."

Still, Pine Valley was a sawmill town: Silk Stocking Hill, a white neighborhood below the hill, and the Colored Quarters. Black people lived in a community north of the sawmill with their own school, boardinghouse, Barrel House, and tiny Baptist church. For blacks and whites too, the largest number of houses were three-room shotgun dwellings, with outside hydrants and privies, that rented for about three dollars a month. Compared with older sawmill towns, though, Pine Valley was a little better; even the smallest house had a coat of white paint. The company sponsored a baseball team, the Lumberjacks, and not long after the mill started up a high school was opened, with a football team called the Pine Valley Panthers.

Three prominent smoke stacks rose two hundred feet high and could be seen far up and down the Kiamichi. The mill's whistle could be heard even farther, sounding like a summons. Valley farmers quit farming and came to work at the mill; before long it seemed that everybody in the upper Kiamichi worked for Pine Valley Lumber Company. As Bill McBride, a longtime resident, says: "The wages were cheap; the work was hard. But actually it was about the only thing there was in this country to do."

In the 1920s, southern lumber companies faced stiff competition from the west coast. They were also burdened with an increasing amount of cutover land. In 1917, it was estimated that southern lumber companies owned an area of cutover land half the size of Texas. Some companies, including Dierks, experimented with converting cutover land to cattle ranches, peach orchards, strawberry farms, etc. They were unsuccessful, mostly because the soil was too poor to support sustained agriculture. In 1923, the Dierks family began to consider harvesting second-growth timber and hired William L. Hall, an experienced forester "whose survey of the public domain in the Ouachitas had led to the creation of the Arkansas National Forest in 1907" (p. 116). Hall reported that there was enough virgin and second-growth timber to allow Dierks "to practice sustained yield forestry and cut trees forever" (p. 116).

Based on Hall's recommendations, Dierks began practicing the then-radical idea of forest conservation, in particular, uneven-age management. Forests were selectively cut with smaller trees left standing; trees damaged by wind, fire, or disease were cut and, where possible, sold; slash was collected; stands with too many small trees were thinned. Most importantly, Hall recommended fire protection, so Dierks hired two fire-prevention supervisors, who in turn supervised 62 local fire wardens, each of whom oversaw 15,000 to 20,000 acres of forest land and educated their neighbors about the damage caused by fires.

The Great Depression of 1929 was particularly hard on the timber companies and people in the Ouachitas. Construction across the country declined by two-thirds; "in the south, pine lumber production plummeted from 13 billion board feet in 1925 to 3 billion...in 1932" (p. 124). Mill employment in Arkansas dropped 60% and the wages of those with jobs dropped drastically. In 1932, the Buschow brothers' mill at Stapp, Oklahoma, closed. Beginning in 1931, Dierks shut down its mills at Mountain Pine, Dierks, Broken Bow, and Wright City; they would have to remain closed until better times. Only Pine Valley remained open, but lost money. In October 1932, the Pine Valley Lumber Company, the two other Dierks companies, and the Dierks timberlands, "were placed in receivership under... federal bankruptcy law" (p. 132).

During the 1930s, Dierks sold much of its cutover land to the U.S. Forest Service. The U.S. Forest Service had been established in 1891 and the Arkansas (renamed Ouachita in 1926) National Forest, in 1907. The Arkansas National Forest was the first national forest in the south; it was created by President Theodore Roosevelt on the advice of Forest Service chief Gifford Pinchot, who was following the recommendations of forester William L. Hall (who later advised Dierks). It was not until the 1930s, however, that the national forest became an integral part of the area's economy. The Civilian Conservation Corps employed hundreds and completed a number of construction projects and fought wildfires. In addition, the Forest Service began to buy cutover land and worn-out farms, all destined for watershed protection, wildlife habitat, recreation sites, livestock grazing, and production of second-growth timber. Although the price the Forest Service paid for the land (mostly less than $4.00 an acre) seems cheap, forest-service money was the only money around—there were no other buyers.

As a result of the company's reorganization and holding expenses down, as well as the repeal of Prohibition (oak staves could be used once again for making whiskey barrels) and a slightly improved economy, Dierks began to recover. The company reopened its idle mills. In 1936, even Pine Valley showed a profit, its first since opening nine years earlier. Despite this, Pine Valley was doomed.

> In the fall of 1941, the Dierks mill at Pine Valley, Oklahoma, cut out. Fourteen years after the mill had opened, its timber was gone. Dierks had cut everything in the upper Kiamichi Valley, and even had built a truck road over Kiamichi Mountain in order to cut timber that earlier had been reserved for the mill at Wright City. First the Pine Valley sawmill shut down, and after some months the planing mill did also. When the planer closed, a power plant operator tied down the whistle. A Pine Valley resident recalls: "When that whistle blew the last time, it was so lonesome. It just blew and blew and blew. It was so sad." People moved away and Dierks dismantled and salvaged....Nearly all the buildings were moved or torn down. (p. 182)

In 1969, Weyerhaeuser Company bought Dierks Forests, Inc. for $317 million and stock. Included in the sale were Dierks' mills, plants, railroads, and timberlands. The last and largest of the family-owned timber firms (p. 206) in the Ouachitas ceased to exist.

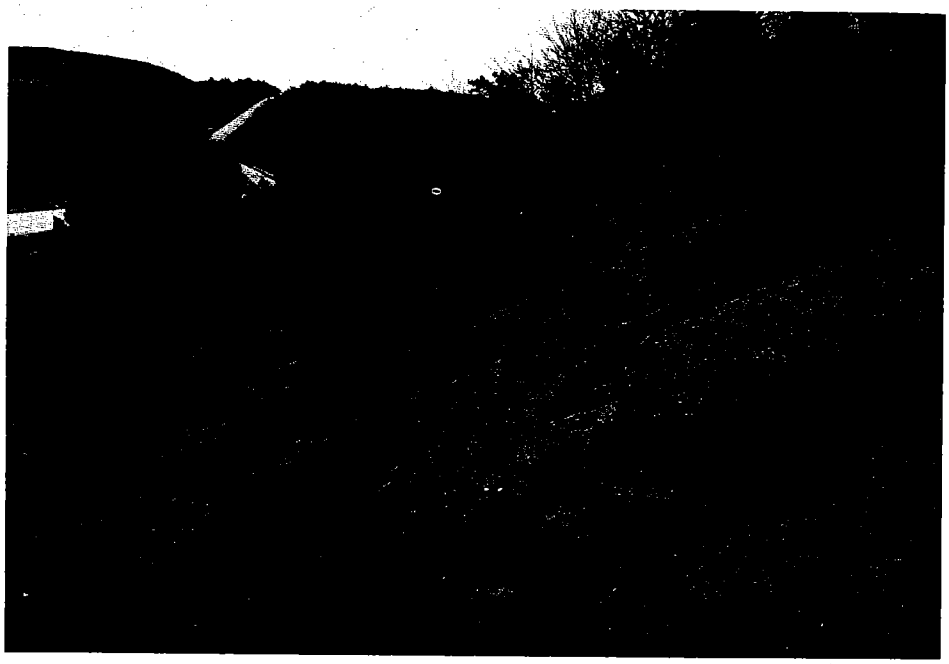


Figure 79. Shale in Jackfork Group, Mile 54.7, Day Two, Talimena Drive.

56.1	0.2	Dead Man Vista. Good view to north of Cavanal Mountain, San Bois Mountains, Mobil No. 1 Green Bay Packaging well site (sec. 8, T. 4 N., R. 23 E., 21,935 ft T. D. [Arbuckle], dry).
		Leave Leflore southeast 7.5′ quadrangle, enter Blackjack Ridge 7.5′ quadrangle.
56.2	0.1	Forest Service Road 6010 to Holson Valley on right (north).
56.6	0.4	Large outcrop of Jackfork sandstones and shales on right (north). Some shale here is very dark olive black (5Y2/1) to grayish black (N2) and may be organic-rich. Some sandstone beds at the bottom of the outcrop are folded; the folds do not look systematic and the plunge of the fold axes varies from very steep to subhorizontal. Are these folds tectonic or the result of near-surface creep on the soft shales?
57.0	0.4	Forest Service Road 6010 to Whitesboro on left (south). Deeply weathered friable sandstones on right.
57.5	0.5	Radio Tower and Sycamore Lookout Tower. Jackfork sandstones are folded to form a moderately open anticline that plunges about 25° to S. 52° W. (π-diagram).
57.8	0.3	Panorama Vista. Good view to north of San Bois Mountains, Cavanal Mountain, Poteau Mountain, Walker Mountain, Blue Mountain, Wister Lake, Summerfield, Holson Valley, and the Mobil No. 1 Green Bay Packaging well site. Along the highway is a cleared area used as a hang-glider launch site.
59.1	1.3	Road bends to left, small pull-off area on right. Short walk to Bear Den Cave.
59.9	0.8	Holson Valley Vista. Good view to north. Similar to Panorama Vista (Mile 57.8, Day Two).
60.1	0.2	Potato Hills Vista. Good view to south of Kiamichi Mountain, State Highway 63 and U.S. 271, the Potato Hills, and Talihina. The sign says that the Potato Hills are the westernmost occurrence of the Arkansas Novaculite. In fact, the novaculite crops out on Black Knob Ridge in Atoka County (~60 mi west-southwest of here) and is the principal reservoir in the Isom Springs Oil Field in Marshall County (~110 mi west-southwest of here).
60.2	0.1	Dirt road on left (south).
60.6	0.4	Good outcrop of shale in Jackfork Group on left (south). Shale appears greenish, but is light olive gray (5Y5/2) where fresh and yellowish gray (5Y7/2) where weathered. Locally, the shale appears reddish (brownish gray [5YR4/1] where fresh and grayish brown [5YR3/2] where weathered.) The shale contains elliptical concretions as long as 8 in.

60.8 0.2 Dirt road on left (south).

61.3 0.5 Old Military Road Picnic Area.

61.5 0.2 Choctaw Vista. Good view to south of Kiamichi Mountain, State Highway 63, U.S. 271, and Talihina. The sign describes the Kiamichi Valley as the heart of the Choctaw Nation and discusses the "Trail of Tears."

61.8 0.3 The grassy slope on the left (south) is probably the Johns Valley Shale, although olistoliths are absent.

There is an excellent outcrop of the Johns Valley on the west side of U.S. 271, 0.3 mi north of the intersection with Talimena Drive (Mile 61.9, Day Two). Limestone and chert cobbles and boulders are abundant. There are also granule- and pebble-conglomerate beds similar to those along the railroad near Stapp (see Stop 19).

61.9 0.1 End of Talimena Drive. Intersection of S.H. 1 (Talimena Drive) and U.S. 271. Turn left (south) on S.H. 1/U.S. 271 toward Talihina.

For the next 2 mi, the road crosses a series of thrust faults in the footwall of the Windingstair thrust fault that are arcuate in map view and probably spoon-shaped in three dimensions. The glide plane for each is the Johns Valley Shale.

62.8 0.9 Small, north-vergent flexural-slip(?) fold in Jackfork sandstone is exposed on the left (east) side of road. The fold plunges 19° to S. 80° W. and has an overturned (south-dipping) north limb.

62.9 0.1 Cross thrust fault. Johns Valley in the hanging wall is characterized by coarse-grained sandstone, granule conglomerate, and chert-breccia beds.

63.3 0.4 Entrance to Talimena State Park.

63.7 0.4 Cross higher thrust fault. Johns Valley is in the hanging wall and Atoka in the footwall. Coarse-grained sandstones to granule conglomerates in the Johns Valley are exposed in road cuts on the left (east) side of the road. Excellent outcrops of Johns Valley (limestone and chert cobbles, limestone conglomerate fragments, phosphate nodules, Caney Shale fragments) are exposed in the stream on the right (west) side of the road immediately south of here. Johns Valley was identified also by Ulrich (1927) (his locality 19) immediately to the east (SE¼ sec. 36, T. 4 N., R. 22 E.).

64.0 0.3 Small, north-vergent anticline with an overturned north limb is exposed on the left (east) side of the road. The fold plunges 20° to S. 80° W. Strata resemble those in the Atoka Formation but are, in fact, Johns Valley. Poole (1985, p. 110) mapped a series of small, north-vergent folds at this outcrop and suggested that the attitude of the axial surface of the folds is N. 75° W. 40° S.

64.3 0.3 Road cut through Atoka Formation.

64.5 0.2 Cross trace of Windingstair fault. Stanley Group strata are in the hanging wall and Atoka in the footwall. For the next 19 mi, the field-trip route is entirely in Stanley Group sandstones and shales.

65.3 0.8 Road to left (east) to rifle range.

67.1 1.8 Leave Blackjack Ridge 7.5′ quadrangle, enter Talihina 7.5′ quadrangle.

68.9 1.8 Enter town of Talihina. Road swings right.

69.7 0.8 Church on right (north) is faced with pieces of Atoka sandstone that show excellent examples of many types of sole marks. The slabs are similar to those that were used to face the Buffalo Valley School described by Suneson and others (1990, p. 17) (Fig. 80).

70.1 0.4 Intersection. U.S. 271 turns left (south); S.H. 1 turns right (north). Turn right.

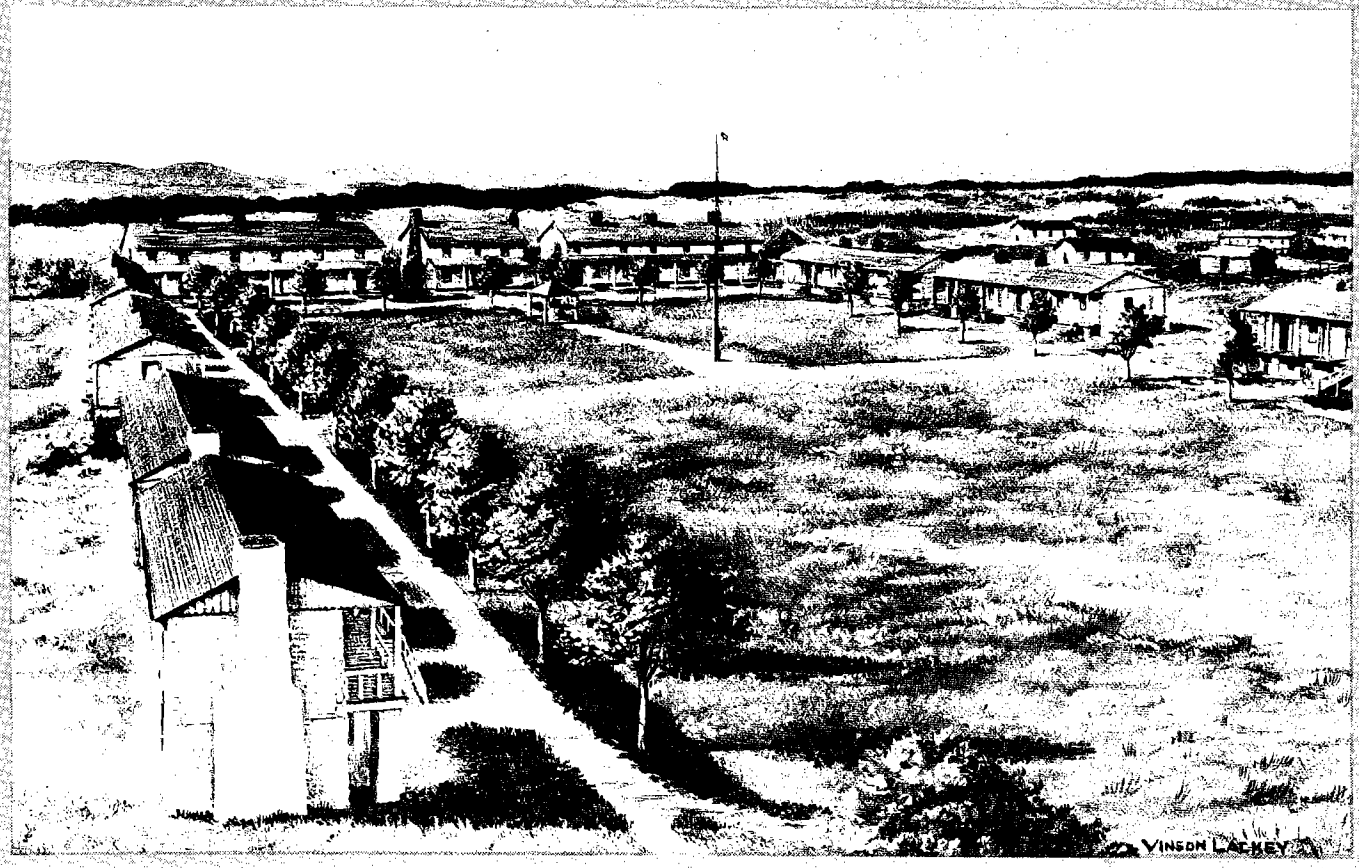


Figure 80. Excellent sole marks preserved on sandstone slabs used as facing on Buffalo Valley School (Suneson and others, 1990). Similar slabs have been used to face several buildings in Talihina.

History of the Old Military Road

During the first part of the 19th century, the U.S. Army built a number of roads across what was then the Indian Territory. The purpose of these roads was to "insure an effective means of communication and transportation between the rear echelon and advanced [military] posts, ...move men, material, and the mail, ...to extend and to protect the frontier, and to implement federal plans regarding Native Americans" (Corbett, 1982, p. 294). After they were built, the roads served as trade and emigration routes for the Indians that lived nearby and, later, for white settlers.

The necessity of the road across Winding Stair Mountain (then called "Ozark Mountain") can be traced to the Treaty of Dancing Rabbit Creek (signed in September 1830), which called for the resettlement of the Choctaw Indians from their homelands in Mississippi to what is now southeastern Oklahoma. Cantonment Towson, located near present-day Fort Towson just north of the Red River, was reactivated to serve as the center for distributing supplies to the Choctaws who would be living near there. Francis W. Armstrong, federal agent for the Choctaws, established his agency (later known as Skullyville and located near present-day Spiro) about 13 miles west of Fort Smith, with the intention of moving the In-



Fort Towson, Indian Territory, is depicted in this painting entitled "Ft. Towson, About 1845," by Vinson Lackey. This view of Fort Towson is looking north; the buildings include barracks, a schoolroom, offices and storerooms, officers' quarters, and a guardhouse. (From the collection of Gilcrease Museum, Tulsa, Oklahoma, accession no. 0127.1398.)

dians up the Arkansas River to the agency and allowing them to disperse across their new land toward Cantonment Towson. In 1831, Armstrong received permission from the Secretary of War to use soldiers from nearby Fort Gibson to build a road from Fort Smith to Cantonment Towson.

In February 1832, Armstrong hired Col. Robert Bean, a well-known frontiersman, and Jesse Chisholm, a local mixed-blood Indian trader, to survey the road. In March, Capt. John Stuart, an infantry officer from Fort Gibson, was ordered to select a detachment of soldiers and build the road to the recently redesignated Fort Towson. On March 28, the soldiers began their work. On April 15, they reached Skullyville, and on the 29th, they arrived at the base of Cavanal ("Cavinole") Mountain. Despite the heavy rains and lack of equipment (particularly wagons and oxen), Capt. Stuart continued his road-building. (One problem for this endeavor was that the road was being constructed under the direction of the Indian Department, but it was being supplied by the War Department. The lack of communication between the two departments [and political infighting?] caused many unnecessary hardships for the soldiers.)

Shortly after leaving Cavanal Mountain, Capt. Stuart was told by some Choctaws

> "...that it would be necessary for me to use great circumspecting in all of my future movements in order to avoid being surprised by the Western Indians, who it was said were committing depredations on the South side of the Ridge, consequently it became necessary to keep the arms of the working men constantly near them, and having no cart with me, or wagon to spare, I was compelled to construct a sled with a suitable gun rack in it, in which about 10 stand of arms were conveyed and kept constantly up near the front of the party when moving" (Foreman, 1927, p. 344).

About May 15th, he met up with Col. Bean, who had finished blazing the trail to Horse Prairie, about 20 miles north of Fort Towson. Bean continued on to Fort Smith and later returned to join Stuart's party

> "...on the South side of the mountain; he informed me...that agreeably to his opinion, I was then at least 90 or a 100 miles from the point of destination, a much greater distance than I had any idea of; here my apprehensions of a scarcity of provisions began to increase, in consequence of which I came to the conclusion that I would be compelled to make the road which had thus far opened from 16 to 20 feet wide, some what narrower and in a few days more I reached the commencement of the pass of the Ozark Mountain where I pitched my camp and proceeded to make a minute examination of the rout the whole way over the mountain, and found it to be extremely rough and rocky for a great portion of the way, and in some places it was almost impossible, and had I followed closely the way blazed over the main Peak of the Mountain I would not have been able to have made a wagon road over it in less than two weeks but I fortunately found a pass over it which was in my opinion decidedly a better rout than that blazed, and not more than two or three hundred yards farther, and so much more easily opened that I succeeded in making a good wagon road the whole way over in about 7 or 8 days.
>
> It was now near the last of May" (Foreman, 1927, p. 345).

Despite being low on provisions, the threat of Pawnees ("who were in considerable force immediately on the west side of the Boggy" [Foreman, 1927, p. 346]), and a Lieut. Northrop's observation that the worst roadbuilding was yet to come, Capt. Stuart continued. He considered abandoning the project, "but the rout had been blazed, the Indians and others had began to travel on it, public expectation was awakened, and it was very important that a road should then be made through" (Foreman, 1927, p. 346).

Stuart's work party arrived at Horse Prairie on June 16, 1832, about 3½ months after starting the road construction at Fort Smith, but they were unable to be reprovisioned from Fort Towson because the supply boat was grounded in the Red River a hundred miles below the fort. Capt. Stuart put his men on half-rations and started the return trip to Fort Smith on June 19th; along the way, he completed a rough survey and made what small improvements in the road he had time for. The party arrived in Fort Smith on June 28th.

After returning to Fort Gibson, Capt. Stuart prepared this description of the road over Winding Stair (Ozark) Mountain, based on the survey his men completed during their return trip:

> "...this mountain is upwards of 4 miles over, and the ascent on the South side is very gradual but is very Rocky, after reaching the Summit the road runs on the narrow top of a spur of this mountain about a mile, when it turns down the North side which is much more steep than the South side, and is for a part of the way down, very rocky, here great labour was required to make a good road, but it was not half so difficult as several other places on the Road" (Foreman, 1927, p. 337).

The Military Road from Fort Smith to Horse Prairie enabled the Choctaws to emigrate more easily to the Kiamichi Valley and to receive rations, and it provided a communications link for the military. The road, however, was not used after 1838, when a new one extending all the way to Fort Towson was constructed farther to the east. This road was designed to "promote frontier protection and provide a more convenient route to emigrants travelling from Illinois and Missouri to Texas" (Shoemaker, 1982, p. 25). In addition, the military could march from Fort Smith to Fort Towson in 8½ days along the new road, considerably faster than along the original road built by Capt. Stuart.

TALIHINA, OKLAHOMA

Talihina is a Choctaw word meaning "iron road" and it refers to the 1888 arrival of the Frisco railroad in this part of Oklahoma (see "The St. Louis–San Francisco Railway," p. 76). Six passenger trains, one local freight, and one fast freight stopped in Talihina daily. Prior to the railroad, the village was a small missionary settlement.

Access to Talihina by other than rail remained difficult until 1919, when the road from Red Oak over Winding Stair Mountain (S.H. 82) was built by convict labor.

70.6	0.5	Intersection with S.H. 82 north to Red Oak. Continue straight (west) on S.H. 1.
71.8	1.2	Intersection with State Highway 63A to State Veterans Hospital and Lake Carl Albert to north. Continue straight (west) on S.H. 1.
72.0	0.2	Intersection with road to Buffalo Mountain Lookout. Bear left on main road (S.H. 1).
72.5	0.5	Leave Talihina 7.5′ quadrangle, enter Albion 7.5′ quadrangle.
74.9	2.4	Leave Albion 7.5′ quadrangle, enter Kiamichi 7.5′ quadrangle.

For the next several miles, the road continues to cross poorly exposed shales and sandstones of the Stanley Group. On the right (north) side of the road is Buffalo Mountain, which is capped by sandstones in the lower part of the Jackfork Group. These are probably within the Wildhorse Mountain Formation (Suneson and Ferguson, 1989). The contact between the Stanley and overlying Jackfork is placed about two-thirds of the way up the side of the mountain, based on the presence of the Chickasaw Creek Siliceous Shale there (Roe, 1955).

On the left (south) side of the road are the Potato Hills, which are underlain by (oldest to youngest) Ordovician Womble Shale, Bigfork Chert, and Polk Creek Shale; Silurian Missouri Mountain Shale; Devonian–Mississippian Arkansas Novaculite; and Mississippian Stanley Group (see Stop 23). The major structural features in the valley between the Potato Hills and Buffalo Mountain are two north-vergent thrust faults (Jackfork Mountain thrust, Buffalo Creek thrust) (Allen, 1994).

79.3	4.4	Cross section A–A′ (Fig. 81) crosses S.H. 1 at this location. The traces of the Jackfork Mountain and Buffalo Creek thrust faults, each of which juxtaposes Stanley in the footwall against Stanley in the hanging wall, are somewhere in the flat valley floor just north of the road.
80.2	0.9	Turn left (south) off of S.H. 1 onto section-line road (between secs. 16 and 17, T. 3 N., R. 20 E.).
80.5	0.3	Cross Little Buffalo Creek.
81.2	0.7	Turn right (west) off of north-south section-line road onto east-west road (through middle of sec. 20).
82.2	1.0	Road turns left (south).
82.7	0.5	Road ends. Turn right (west) onto east-west section-line road between sections 19 and 30.
83.1	0.4	Road forks. Turn left (south).
83.5	0.4	Cross Cedar Creek. Complexly deformed Stanley Group strata are present in creek banks upstream (to the left).
83.6	0.1	Cross trace of North Potato Hills thrust fault. Allen (1993) interprets this fault as a back thrust that developed at a lower hinge area and attained its present position by being transported up a ramp during subsequent displacement.
84.3	0.7	Stop 23. Exposures and outcrops along Cedar Creek.

STOP 23

Lower Paleozoic Section Exposed by Thrust and Compartmental Faulting, North-Central Potato Hills, Ouachita Mountains, Oklahoma

Mark W. Allen
EnecoTech, Inc.
Tulsa, Oklahoma

Location: SW¼SE¼SE¼ sec. 30 and N½NE¼ sec. 31, T. 3 N., R. 20 E., Latimer County

INTRODUCTION

One of the best exposed, most accessible, and most complete sections of lower and middle Paleozoic strata is here in the central Potato Hills (Fig. 82). At this stop, Middle Ordovician Womble Shale through Mississippian Stanley Group are folded into an overturned anticline created by drag on the Cedar Creek fault. The strata are exposed on the south limb of the anticline.

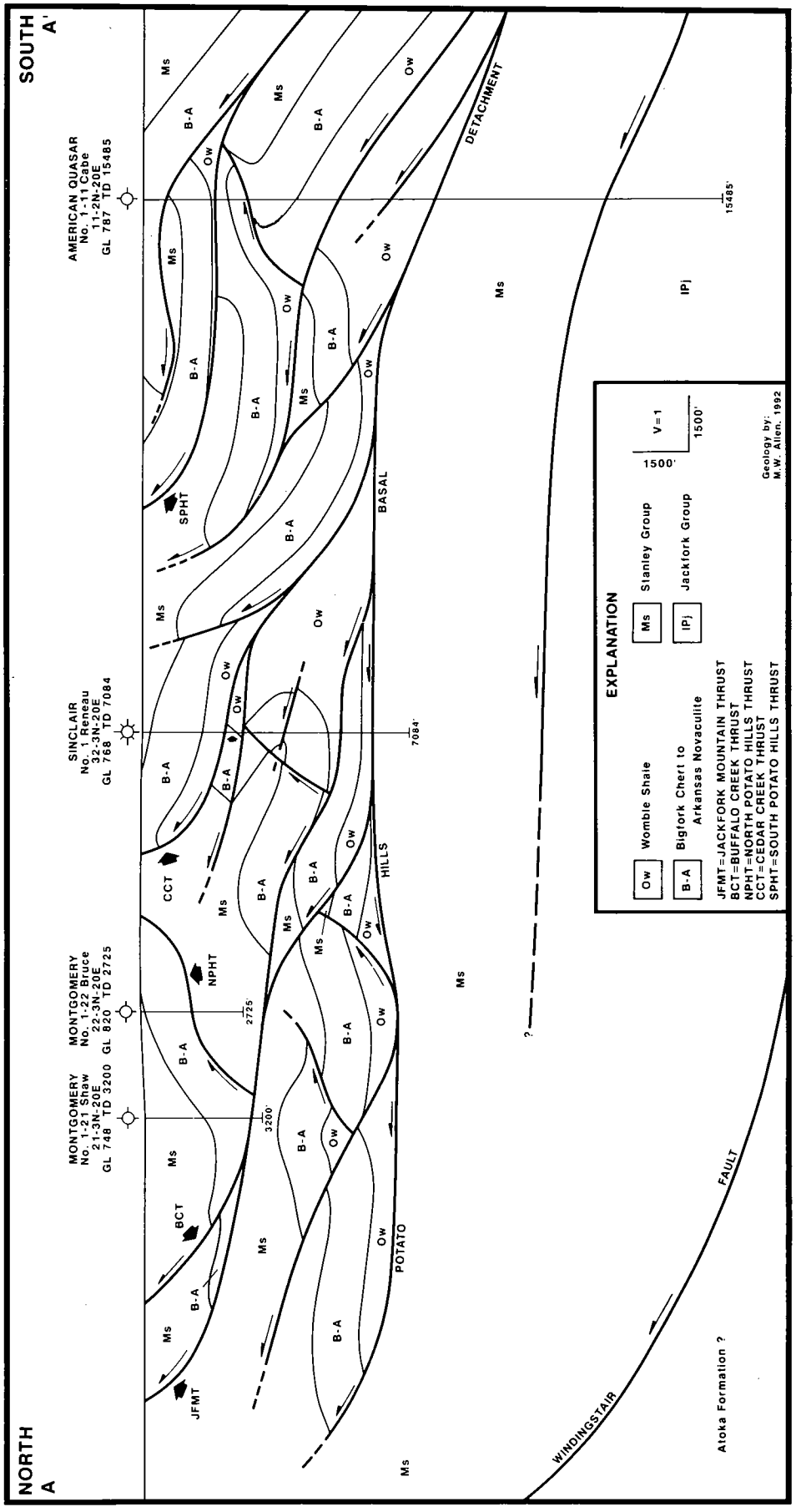


Figure 81. North-south cross section across the Potato Hills showing relation of Jackfork Mountain and Buffalo Creek thrusts to Potato Hills basal detachment and Windingstair fault (from Allen, 1994, fig. 3.).

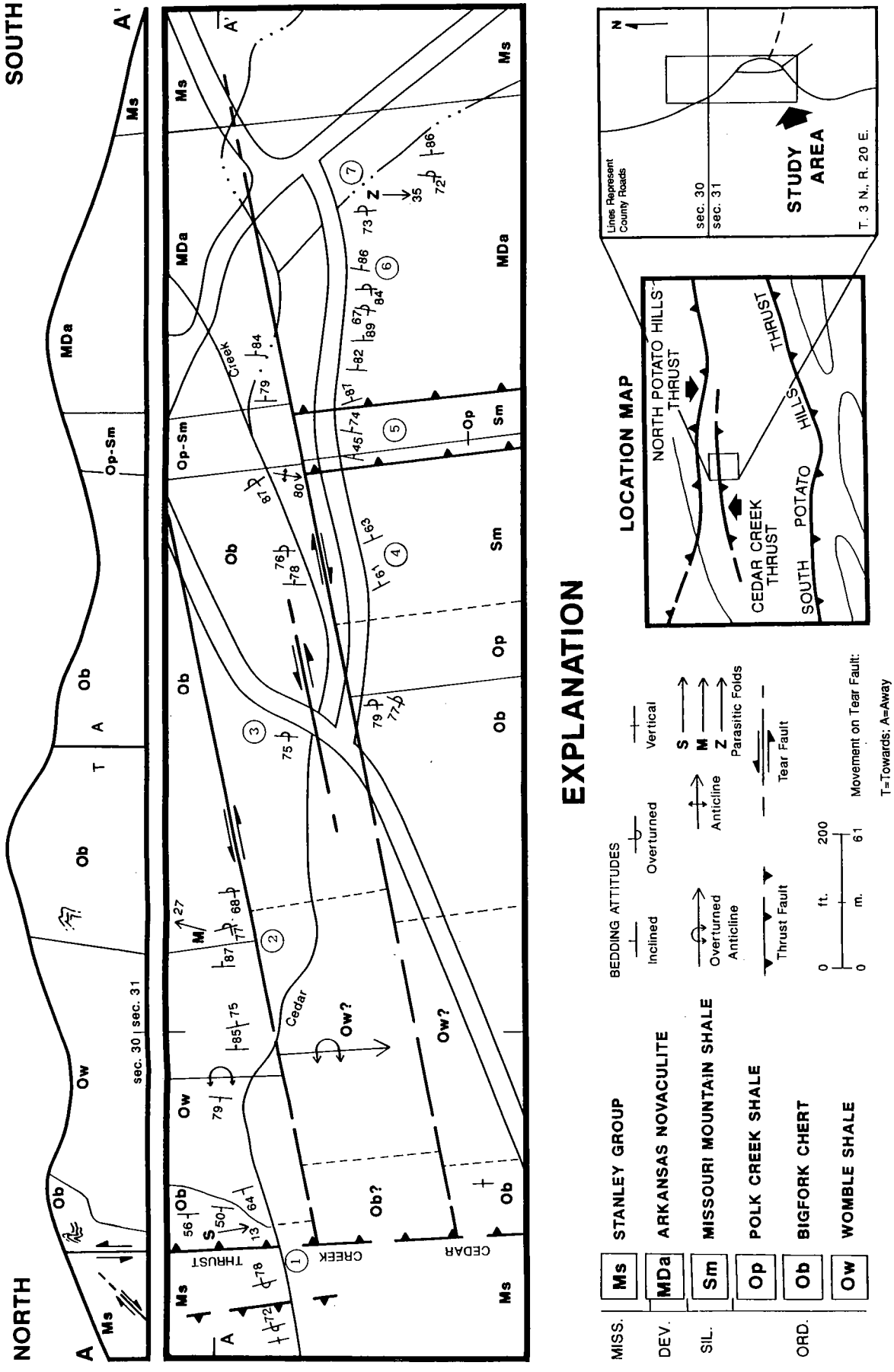


Figure 82. Geologic map and cross section of area near Stop 23. Circled numbers refer to detailed descriptions in text.

STRATIGRAPHY

The Womble Shale is the oldest unit exposed in the Potato Hills. In the stop area, the Womble is a dark gray (N3) to grayish black (N2), fissile, graptolitic shale. The Upper Ordovician Bigfork Chert conformably overlies the Womble Shale. The Bigfork Chert is a dark gray (N3) to grayish black (N2), well-stratified spicular chert with interbedded medium gray (N5) to medium dark gray (N4) shale. The poorly exposed Upper Ordovician Polk Creek Shale, a dark gray (N3) shale with interbedded grayish black (N2) chert, conformably overlies the Bigfork Chert. The Lower Silurian Missouri Mountain Shale is a light olive gray (5Y5/2) to dark greenish gray (5G4/1) siliceous, laminated, platy shale. An unconformity separates the Missouri Mountain and Polk Creek Shales.

The Arkansas Novaculite ranges in age from Early Silurian to Early Mississippian. This unit has been divided into three members (e.g., Lowe, 1989), all of which occur at this stop. The lower member is mostly a grayish green (5G5/2) and grayish green (10GY5/2) chert with interbedded light gray to medium light gray (N7 to N6) siliceous shale. The middle member is a light olive (10Y 5/4) to moderate yellowish green (10GY6/4) shale with interbedded grayish black (N2) to light olive gray (5Y5/2) chert. This member contains some black to grayish black (N1 to N2) carbonaceous shale in the lower part. The upper member is a dark yellowish green (10GY4/4) to light greenish gray (5GY8/1) chert with interbedded dark yellowish green (10GY4/4) shale.

The Stanley Group is the youngest unit exposed in the stop area. The Stanley is mostly a grayish green (10G4/2) to dark greenish gray (5G4/1) shale. No sandstone is present at this stop.

PREVIOUS WORK

The lower and middle Paleozoic section exposed here has been measured by Miller (1955) and Bennison and Johnson (1959). Miller (1955) measured ~1,030 ft of section between the base of the Bigfork and the top of the Arkansas Novaculite. He interpreted a continuous stratigraphic sequence on the south limb of the anticline. He did not recognize any faults or minor folds in the section.

Bennison and Johnson (1959) measured ~1,680 ft, which included parts of the Womble and the Stanley, and suggested a continuous section from the Womble to the Stanley. They indicated the Womble was in fault contact with the Stanley at the Cedar Creek thrust and that "Bigfork and novaculite blocks can be observed in the fault zone" (Bennison and Johnson, 1959, p. 67). They did not identify any faults (other than the Cedar Creek fault) or folds. They interpreted the Cedar Creek thrust fault as the result of "overstretching and rupturing" along the north limb of the anticline (Bennison and Johnson, 1959, p. 67).

CROSS-SECTION DESCRIPTION

Cross section A–A′ (Fig. 82) illustrates the stratigraphic sequence and structural style at this stop. At the north end of the stop area, a fault zone juxtaposes shale of the Stanley Group and Bigfork Chert (location 1, Fig. 82). This fault zone corresponds with the location of the Cedar Creek fault, which has been mapped throughout this part of the Potato Hills. The Cedar Creek fault is approximately vertical and strikes N. 88° E. At creek level, 27 ft of Bigfork Chert is exposed (Fig. 82). Bennison and Johnson (1959, p. 67) suggested that this Bigfork interval corresponded to an almost vertical fault zone separating the Stanley and Womble Formations and "horse blocks of Big Fork and novaculite can be observed in the fault zone." In fact, this interval is entirely Bigfork Chert. North-facing beds and a dextral (or "S") fold (Fig. 82) at this outcrop indicate a major anticlinal axis to the south. Near the road, in the pasture to the west of Cedar Creek, is a small outcrop of Bigfork Chert, the apparent continuation of the Bigfork east of Cedar Creek. The strata in this outcrop have the same strike as the Bigfork across the creek, but are vertical.

The Womble Shale, complexly folded and faulted near the Cedar Creek fault, has an apparent thickness of 418 ft. South of the fault, it is mostly covered. The anticlinal axis is in the covered interval and facing directions could not be determined to locate the axis exactly. The southern Bigfork/Womble contact is exposed on the south limb of the anticline, east of Cedar Creek (location 2, Fig. 82). At the contact, the Womble appears to have been dragged southward along a left-lateral tear fault.

The Bigfork Chert is 713 ft thick from the southern Womble/Bigfork contact south. Along the creek, lower to middle Bigfork Chert dips 68°–77° north. Farther south where the road crosses Cedar Creek (location 3, Fig. 82), beds dip north and show slight buckling and possible bedding-plane slip. There is a small left-lateral tear fault on the west side of the stream crossing. At this point, walk up the road on the west side of Cedar Creek and examine the upper Bigfork through the upper Arkansas Novaculite.

The Bigfork Chert/Polk Creek Shale contact is tentatively placed at the southernmost exposure of chert which dips 77°–79° north. The Polk Creek interval through the lowermost Missouri Mountain Shale is covered by vegetation. The middle and upper parts of the Missouri Mountain Shale are exposed in the road cut and can be seen best in the drainage ditch next to the road (location 4, Fig. 82). The Polk Creek Shale and the Missouri Mountain Shale are estimated to be 110 ft and 230 ft thick, respectively. Exact thicknesses could not be obtained due to the inexact location of the unconformable contact between the Polk Creek and Missouri Mountains Shales. A repeated, faulted interval of Polk Creek/Missouri Mountain shale (16 ft and 68 ft thick, respectively) occurs near the top of the road cut (location 5, Fig. 82) and is marked by a distinct color change from greenish Missouri Mountain Shale to dark gray shale and grayish black chert (Polk Creek Shale). The fault zone is indicated by folding and fracturing. Another fault zone to the south is indicated by an oxidized zone at the Missouri Mountain/Arkansas Novaculite contact. The unconformity between the Polk Creek Shale and the Missouri Mountain Shale between the fault zones is not obvious.

The Arkansas Novaculite is 440 ft thick in this area and its three-part division can be identified. The exposures along the road cut (location 6, Fig. 82) comprise the lower chert and the middle shale members. The upper chert member is not present in the road cut but can be seen to the south in the vegetation. A sinistral (or "Z") fold (loca-

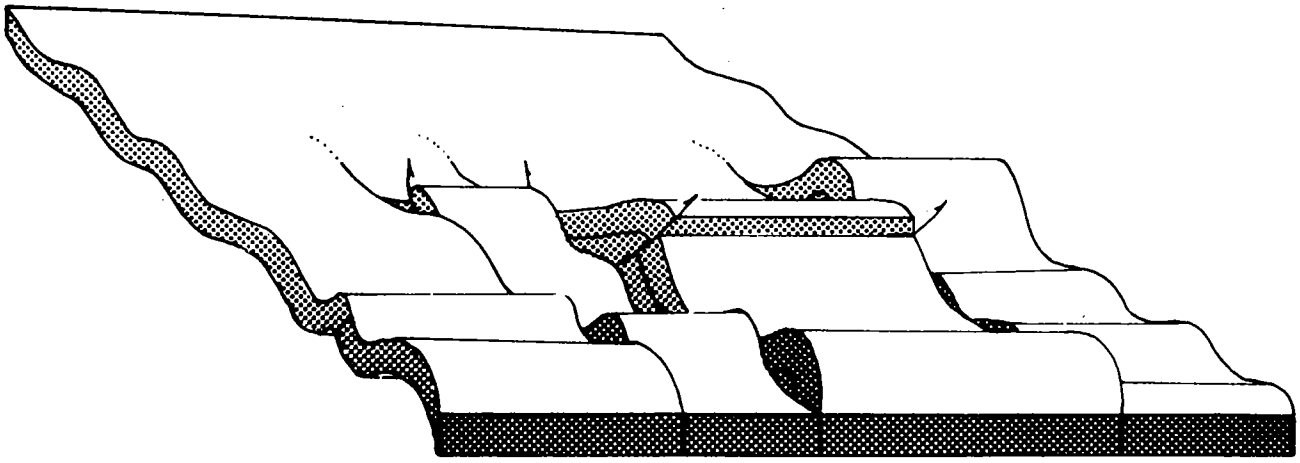


Figure 83. Block diagram illustrating compartmental faulting (from Davis, 1984; original artwork by R.W. Krantz).

tion 7, Fig. 82) in the road cut indicates that the direction of the major anticlinal axis is to the north. The contact between the Arkansas Novaculite and Stanley Group is mapped at the last occurrence of chert.

DISCUSSION

The structural geology of the stop area as shown on Figure 82 is more complex than what can be observed along the creek and road. The important structural features are the parasitic folds, tear faults and minor thrust/reverse faults, most of which have not been mapped previously. The parasitic folds indicate that a major anticlinal axis occurs somewhere within the Womble Shale. Bedding attitudes in the Bigfork Chert on either side of the Womble suggest that the axial plane dips north and is, therefore, overturned. Evidence for two major tear faults (designated east and west) with mappable horizontal translation includes offset contacts (e.g., upper Bigfork contact along the west fault) and drag along the fault plane (e.g., at the southern Womble/Bigfork contact along the east fault). The tear faults also show apparent vertical translation. The entire area between the two faults is lower in elevation than the strata on the other sides of the faults. The apparent vertical offset on the east fault is down to the west, whereas that of the west fault is down to the east.

The variable separations across the tear faults indicate that they are probably compartmental faults (Fig. 83). Compartmental faults serve as partitions between domains of strata in which the same shortening has been achieved in different ways. Compartmental faults are unlike strike-slip faults in that the sense of slip need not be uniform along the length of the fault. Also, they are surprisingly short when compared to the magnitude of slip they accommodate (Davis, 1984). Compartmentalization allows different structural profiles to develop simultaneously and independently while the overall structural balance is maintained (Brown, 1988). Most compartmental faults have finite ends that limit the amount of strike-slip movement; in this case, the northern end is probably at the Cedar Creek fault. The southern ends of the faults juxtapose Stanley against Stanley and could not be located.

Compartmental and strike/slip faults occur throughout the Potato Hills and have been described as "cross faults having a predominant northwest trend and showing dip slip and right-lateral strike slip" (Arbenz, 1968, p. 113). The cross faults were thought to have been the result of a "general late orogenic stretching" (Arbenz, 1968, p. 113); however, extension is not consistent with the tectonic processes observed on the surface and in the subsurface in the Potato Hills (Allen, 1994).

End of field trip.

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Part II



Contributed Papers

Geologic Relations Along a Regional Cross Section from Spavinaw to Broken Bow, Eastern Oklahoma

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Houston, Texas

ABSTRACT.—A north-south geologic cross section constructed from map analysis, well control, and seismic-reflection data illustrates the regional structural and stratigraphic relations along a 170-mi traverse in eastern Oklahoma through the Ozark uplift, Arkoma Basin, and Ouachita thrust belt, which are related Paleozoic tectono-stratigraphic elements on the North American side of a collisional orogenic system. Outcrops in the western Ozarks show the gentle deformation and thin, unconformity-riddled strata of a long-lived carbonate shelf. Basement block faults along the southern edge of the Ozarks, including the large, down-to-the-basin Mulberry fault, are high-angle normal faults near the surface and terminate at a low-angle intrabasement detachment. The detachment surface dips southward from ~9,000 ft deep near Brushy to 30,000 ft deep at the Mulberry fault near Muldrow, ~11 mi to the south. This deep detachment is interpreted to underlie all of the Arkoma Basin and Ouachita thrust belt. Basement horsts and grabens under the Arkoma Basin and frontal Ouachitas are interpreted to terminate against this deep low-angle fault, and together these features suggest a Late Mississippian–Early Pennsylvanian episode of north-south crustal extension of the southern edge of the North American craton penecontemporaneous with convergence and early stages of collision with a southern continent (Yucatan?).

Above the basement-fault complex, Carboniferous rocks of the Arkoma Basin are detached along listric thrusts, some of which are reactivated listric normal faults. These rocks are deep- to shallow-water clastic strata deposited in an evolving foredeep basin. Along the Ouachita front these units are involved in deep duplex structures. At the Choctaw fault the shelfal rocks seen in the Ozarks lie 20,000 ft below the surface. The Ouachita thrust-belt exposures are of basinal facies coeval with the shelf units of the Ozarks. They have been thrust >45 mi to the north of their original site of deposition, on the basis of interpretation of data from the Sohio No. 1-22 Weyerhaeuser well near Broken Bow. Surface thrusts and folds in the Carboniferous units between the Choctaw and Octavia faults are interpreted to flatten above a duplex structure at 9,000 ft (north) to 25,000 ft (south) below the surface. The duplex may contain Carboniferous units and is detached above basement and lower Paleozoic shelf units at 20,000 ft (north) to 30,000 ft (south) below the surface.

The low-grade metasedimentary strata at the surface of the Broken Bow uplift—basinal lower Paleozoic clastic rocks, chert, and carbonate rocks—exhibit small-scale folds with south vergence and probable duplex structures that have been arched above a basement duplex produced in the final stages of collision. The detachment between the surface basinal-facies structures and underlying shelf facies and basement structures was penetrated in the deep 1-22 well at ~11,700 ft below the surface. The lower Paleozoic basinal facies must have been deposited south of this well, probably south of the town of Broken Bow. Overall vergence of the Ouachita thrusts is northward; south-verging structures south of the Boktukola fault may be due in part to back thrusting in the middle of the shale-rich lower Paleozoic section, sandwiched between the thick sandstone-rich upper Paleozoic section and basement as it ramped over the paleoshelf edge.

Dramatic thickness increases and facies changes are observed between shelf (foredeep) and basin (thrust belt) stratigraphic units, but the transition between shelf and basin facies, including paleoshelf edges, is hidden beneath the surface. The early Paleozoic shelf edge and shelf were broken by normal faults accompanied by subsidence, causing Pennsylvanian shelf edges to be shifted at least 60 mi to the north of Broken Bow by Atokan time. Deep-water Lower Pennsylvanian clastic strata were deposited on top of the lower Paleozoic shallow-water shelf section despite a worldwide lowstand of the sea because of this tectonic subsidence. The large influx of clastic sediments and progressive tectonic constriction of the foredeep caused shallow-water facies to accumulate during the rest of the Pennsylvanian worldwide highstand conditions.

Roberts, M. T., 1994, Geologic relations along a regional cross section from Spavinaw to Broken Bow, eastern Oklahoma, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 137–160.

INTRODUCTION

The eastern boundary of Oklahoma traverses related tectonic elements of the Ouachita orogenic system: from north to south, the Ozark uplift, Arkoma Basin, frontal zone of the Ouachita thrust belt, and the "core zone" or Broken Bow uplift of the thrust belt. A geographic strip of eastern Oklahoma through these elements has been the destination of many geologic field trips to view excellent examples of shelf-carbonate rocks (Ozarks), fluvial-deltaic clastic rocks (Arkoma Basin), submarine-fan clastic rocks, and the somewhat enigmatic novaculite deposits (Ouachitas). These Paleozoic shelf and basinal facies crop out in a reasonably compact area, juxtaposed by thrusting, and outcrops are both numerous and accessible via highway systems such as Oklahoma Highway 10 and U.S. Highway 59/259 (Fig. 1).

The purpose of this paper is to present and discuss a geologic cross section (Pl. 1, p. 160) that transects this eastern Oklahoma strip from near Spavinaw Dam in the north to Broken Bow in the south, a distance of ~170 mi. The cross section incorporates well data, surface geology, and seismic data into a synthesis of the crustal structure. It was undertaken originally to be used as a framework in which to view the surface outcrops visited on so many field trips, especially for those trips intending to illustrate shelf-to-basin "transitions" as the transitions are hidden beneath the surface and obscured by tectonic shortening.

Previous cross sections have been published both east and west of the one presented here. Wickham (1977), Arbenz (1989), and Briggs and Roeder (1975) drew cross sections along a line from Broken Bow northwest through the Potato Hills and into the Arkoma Basin. In Arkansas, work by Arbenz (1984), Viele (1979), and Zimmerman and others (1982) and a COCORP line (Lillie and others, 1983) illustrate interpretations of the Ouachita system to the east of the cross section shown here. The interpreted styles of deformation in the Arkoma Basin, frontal Ouachitas, and Broken Bow (Benton) uplift differ considerably among these workers. Both the structural styles of the major tectonic elements and the structural relations between the elements are poorly understood. No completely satisfactory synthesis exists for this region that unifies the structural and stratigraphic development of the shelf and basin, foredeep, and thrust belt.

CONSTRUCTION OF THE CROSS SECTION AND GENERAL OBSERVATIONS

The cross section (Pl. 1) was placed so as to tie as closely as possible in a reasonably linear traverse many of the popular field-trip stops of eastern Oklahoma, linking the Ozark dome, Arkoma Basin, and Ouachita thrust belt. The northern third of the section was constructed from surface observations. The central part of the section (about half of the total length) was drawn from an integration of surface data, well data, and industry seismic-reflection data. The southern end of the section was interpreted from surface observations, constraints imparted by the section to the north, and a key deep exploration well. Sources for surface geology include Huffman (1958), Briggs (1973), Hart (1963), Honess (1923), Marcher and Bergman (1983), Miser (1954), and Shelburne (1960).

Most of the major stratigraphic units are exposed along or near the cross section, but many important structural features are hidden from the surface view and revealed only by the subsurface control. The deep struc-

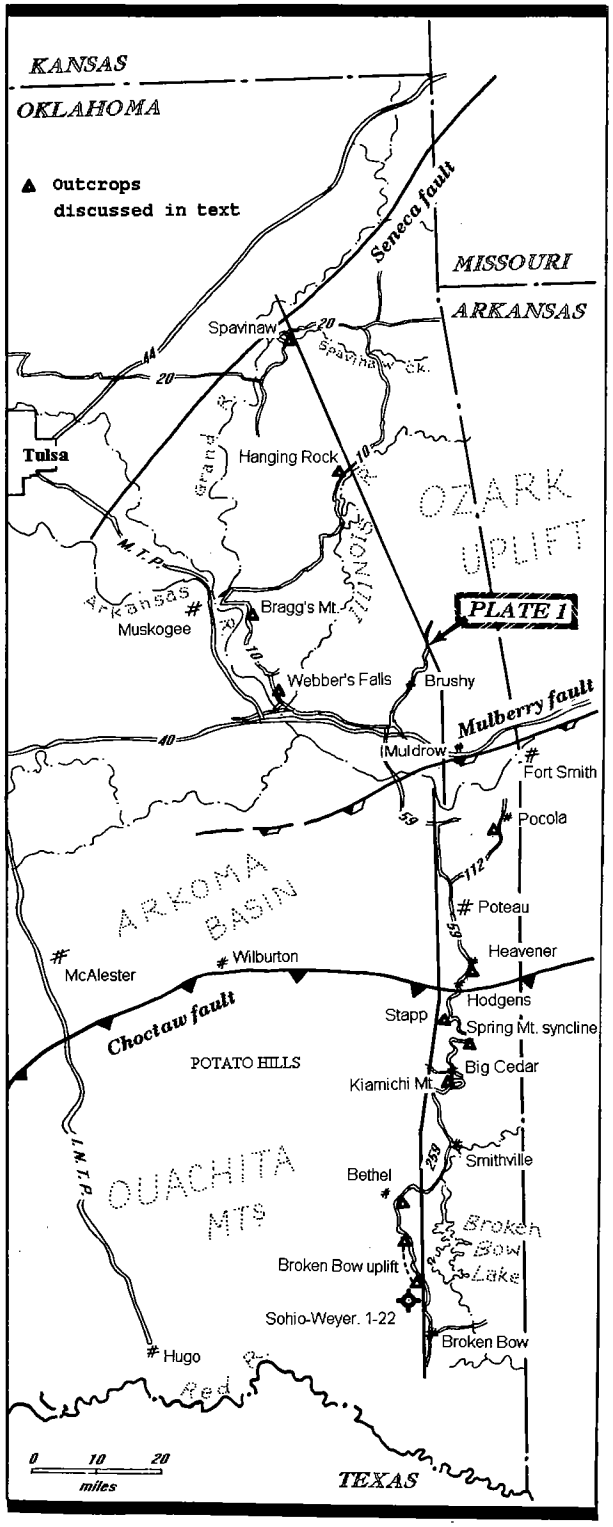


Figure 1. Area map showing location of cross section (Pl. 1) and localities discussed in text.

ture south of the Choctaw fault (Pl. 1) is only partially imaged by seismic data and contains few well penetrations. Also hidden are major tectono-stratigraphic boundaries such as the early Paleozoic shelf edge and the transition from shallow- to deep-water facies of the Morrowan interval, including the hypothetical source terrane for the Johns Valley boulder beds.

North of the Mulberry fault (Pl. 1), surface structure reflects basement structure. This part of the cross section traverses the southwestern flank of the Ozark uplift, a large low-relief basement dome cut by high-angle normal and, perhaps, strike-slip faults. The Precambrian basement crops out in Spavinaw Creek just downstream of Spavinaw Dam (Fig. 1). Here it is a pink granite dated at ~1.3 Ga (Denison, 1981). It is embedded in a rhyolitic to andesitic volcanic group of similar age known from well control. This basement complex appears well-layered on seismic lines that cross the southern flank of the Ozarks and northern edge of the Arkoma Basin. Intrabasement seismic reflectors appear to bound large lensoid to tabular rock units. More seismically stratified units may be volcanic rocks or metasedimentary rocks, and the lensoid bodies may be intrusive rocks. Also imaged in industry seismic data are gently south-dipping reflections that define a plane that crosscuts these seismic-stratigraphic rock units of the basement. The high-angle faults seen at the surface and on seismic data appear to terminate at the plane defined by these low-angle reflections. The low-angle events are interpreted as reflections from an intrabasement décollement (Pl. 1). In Plate 1, that part of the low-angle basement fault north of the Mulberry fault is visible on seismic data along the line of section. The detachment (reflector) is ~9,000 ft deep near Brushy and ~30,000 ft deep near Muldrow, 11 mi to the south. Similar relations are seen on seismic data in central Arkansas.

South of the Mulberry fault, surface structures are detached from basement structures. The basement and lower Paleozoic shelfal rocks under the Arkoma Basin are broken into a complex of horsts and grabens; individual faults locally have >4,000 ft of vertical displacement. These structures are documented with seismic and well data. The basement décollement visible on the north flank of the Arkoma Basin is interpreted to extend under the entire basin and thrust belt, although in the south it lies too deep to be seen on the industry seismic lines. If present, this master low-angle fault and its overlying fault complex are not explainable by the simple loading models applied to foredeep development and suggest an episode of north-south crustal extension of the North American plate penecontemporaneous with the advance of the thrust belt from the south. This extension lasted from Late Mississippian to Early Pennsylvanian, on the basis of interpretation of seismic data tied to wells. Some of the normal faults above the décollement appear to have been reactivated as reverse faults near the frontal zone of the thrust belt off the line of this section. The core of the Broken Bow uplift is interpreted to contain basement thrusts reactivated from the earlier extensional complex during the final stages of collision.

Overlying the basement rocks are Cambrian to Pennsylvanian strata (Fig. 2). In the north, on the Ozark dome, these strata are dominantly shelf-carbonate rocks and form a relatively thin cover on the basement. To the south, in the Arkoma Basin, the Cambrian through Devonian rocks maintain their shelfal nature. These units can be followed on seismic data under the frontal Ouachita thrusts and are interpreted to extend at least as far south as the Broken Bow uplift on the basis of the interpretation of data from the Sohio No. 1-22 Weyerhaeuser well (Pl. 1). Coeval basinal clastic strata, chert, novaculite, and minor carbonate units exposed in the Broken Bow uplift are interpreted to have been tectonically transported at least 47 mi to the north to their present position overlying the shelf facies. Outcrops of these basinal facies in the Potato Hills support an overthrust distance of at least 50 mi.

Regional relations in the Carboniferous units are complex because they were deposited syntectonically with the Appalachian-Ouachita orogeny. The carbonate platform survived into the Early Pennsylvanian in northeast Oklahoma (into Morrowan time). It was overwhelmed with clastic sediments progressively from east to west, beginning in the Mississippian in the Black Warrior Basin. Thick Upper Mississippian and Morrowan clastic rocks are present above the shelf carbonate rocks in north-central Arkansas. In northeast Oklahoma, Morrowan carbonate strata are disconformably overlain by Atokan fluvial and shoreline clastic strata. To the south in the Arkoma Basin, Carboniferous units thicken dramatically, especially the Atokan strata (Fig. 2), and are deformed by listric growth faults and later thrust faults. Seismic geometries strongly suggest that many of the thrusts are reactivated listric normal faults. This structural style is especially clear on seismic lines in the northern parts of the basin, in central Arkansas, which show listric normal faults in Atokan, Morrowan, and Chesterian strata that flatten above the earlier Mississippian basement block faults. In more southern areas, listric thrusts show clearly compressional folds and overthrust geometries in Atokan units, but have remnant normal displacements in Morrowan units (Fig. 3).

Carboniferous units are fluvial deltaic throughout northern areas (except for the remnant Morrowan shelf in eastern Oklahoma), but in southern areas, these sediments are deep-water clastic rocks that include thick submarine-fan deposits. The transition from shallow- to deep-water facies has proved difficult to locate and is buried deeply under the thrust belt. Seismic evidence suggests a Morrowan shelf-slope break ~25 mi north of Little Rock, Arkansas. It trends west through central Arkansas, but by latest Morrowan time it had prograded south of the Choctaw fault in eastern Oklahoma (Wapanucka shelf). The transition from shallow to deep water in Atokan time migrated northward as the old shelf was broken down by normal faulting and loading from the advancing Ouachita thrust sheets. This transition lies under the central part of the Arkoma Basin. Atokan units in the Ozarks and northern Arkoma Basin are fluvial or shallow-marine clastic rocks; however, lower Atoka rocks in the subsurface of more southern areas of the basin and exposed in frontal thrust sheets are submarine-fan and associated deep-water clastic rocks. Younger Atokan and Desmoinesian clastic rocks are fluvial to shallow marine and represent the final fill of the Arkoma foredeep basin.

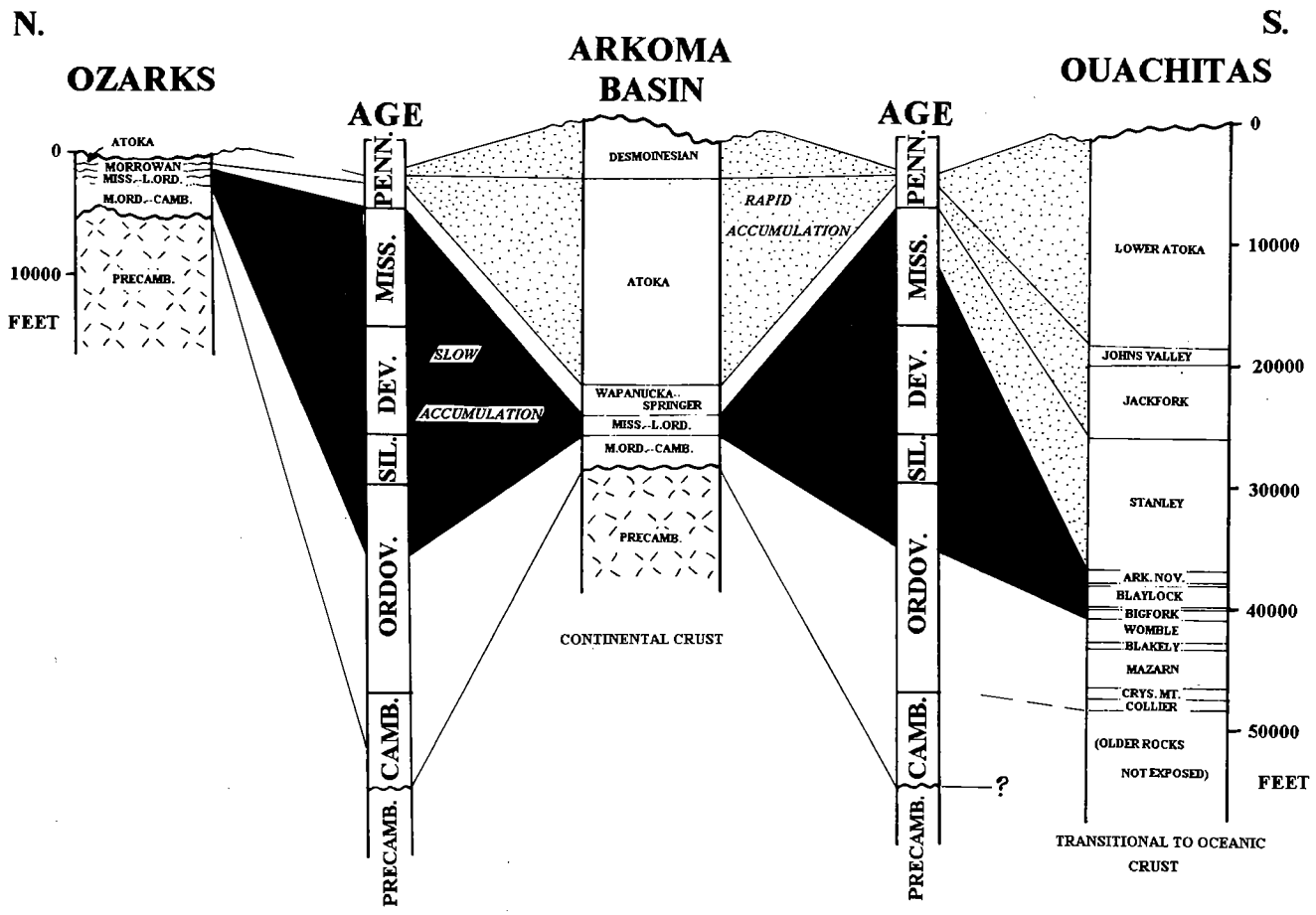


Figure 2. Time-thickness diagram, Paleozoic units of the Ozark uplift, Arkoma Basin, and Ouachita Mountains.

Plate 1 shows a deep duplex structure involving Mississippian(?) through lower Atokan rocks beneath the southern part of the Arkoma Basin and extending southward at least to the Kiamichi Mountain area. In the Heavener area the evidence for this duplex includes seismic and well data. In the nearby Red Oak gas field, this duplex is entirely within the Atoka (Fig. 4). South of the Choctaw fault the evidence for this duplex is based on seismic data. In the area of the Simmons Mountain syncline, between the Briery and Windingstair thrusts, seismic data clearly show that the surface structures are detached from a deeper complex (at ~2 s = 15,000 ft?), which is in turn detached from probable basement and cover rocks (at ~4 s = 25,000–30,000 ft?). The age of the rocks in this duplex is uncertain, but is believed to be Carboniferous. The deep structure south of Kiamichi Mountain is speculative, and that shown on Plate 1 is diagrammatic, as is the surface structure of the Broken Bow uplift. Map analysis and interpretation of the Sohio No. 1-22 Weyerhaeuser well data suggest that the Broken Bow uplift contains both cover and basement duplexes and that the rocks there have undergone multiple deformations. At the surface of the Broken Bow uplift are exposures of basinal facies that are coeval with the early Paleozoic shelf facies exposed in the Ozarks on the northern end of the cross section and with the shelf carbonates shown by the deep well to lie beneath the surface complex at ~11,700 ft. The early Paleozoic shelf edge must lie to the south of this well, perhaps south of the town of Broken Bow. The early Paleozoic Ouachita basin, therefore, was also south of Broken Bow.

SURFACE GEOLOGY ALONG THE LINE OF CROSS SECTION

Ozark Uplift

The Grand and Arkansas Rivers have cut a semicircular path around the western flank of the Ozark dome. Their tributaries, some with steep canyon walls, and road cuts provide many outcrops of the thin stratigraphic section of the Paleozoic shelf. Four outcrop areas illustrate the geology of the Ozark uplift shelf facies: Spavinaw Creek, Oklahoma Highway 10 road cuts along the Illinois River and at Braggs Mountain, and the cliff face at lock and dam 16 near Webbers Falls (Fig. 1). Paleozoic carbonate and shale units in this part of the Ozarks are generally <2,000 ft thick (compare with >40,000 ft of equivalent-age strata in the Ouachita Mountains). The shelf units contain numerous unconformities, notably a Middle Devonian angular one and an erosional one at the Chesterian/Morrowan boundary.

Just below Spavinaw Dam, on Spavinaw Creek, the Precambrian basement is exposed, one of numerous basement knobs in northeast Oklahoma (others are known only from subsurface control). Ease of access and

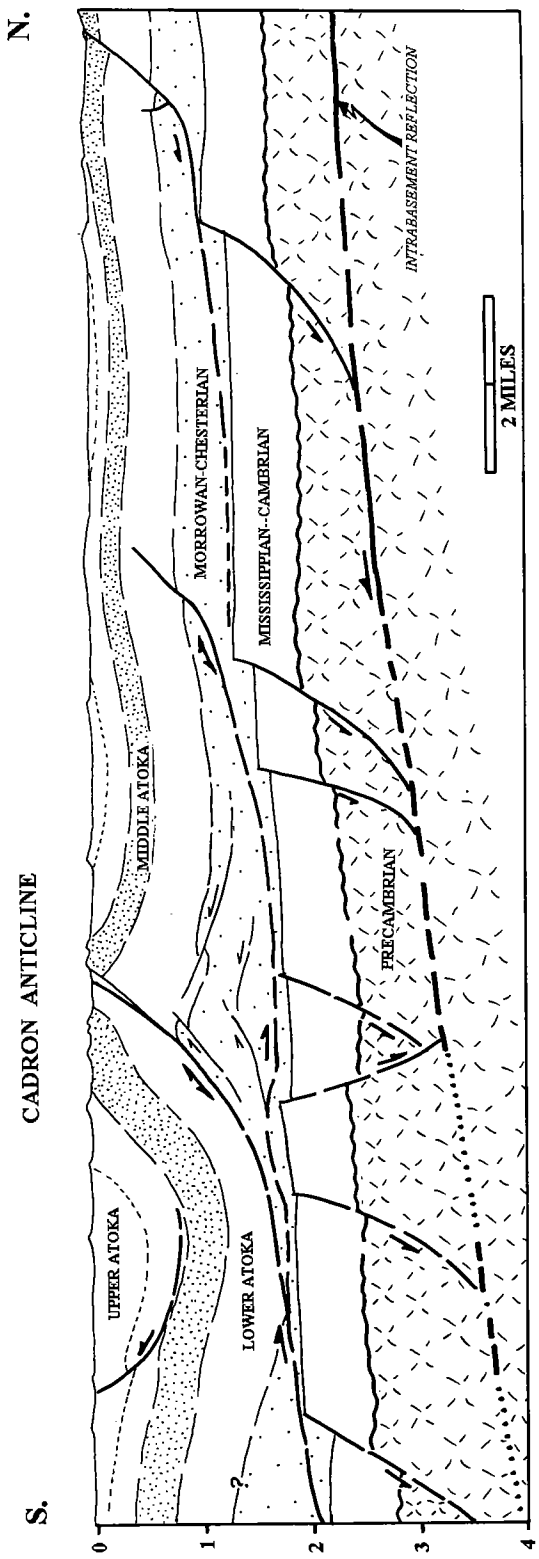


Figure 3. Line drawing of seismic reflection line from Faulkner County, Arkansas, illustrating dual nature of some listric faults: remnant normal offset of deep units and reverse offset of shallower Atokan rocks. Note also the intrabasement reflector against which high-angle basement faults terminate.

natural beauty have made this outcrop a popular field-trip stop. The Precambrian surface in northeast Oklahoma is dotted with erosional hills that rise several hundred to >1,000 ft above the surrounding surface (Reeder, 1974). The Spavinaw Granite outcrop is the exhumed top of one of these hills and affords a rare view of the basement in this region. Basement and cover rocks are, in general, gently deformed in this part of the uplift, but an exception lies ~4 mi north of Spavinaw (Pl. 1) in the Seneca fault zone. This feature is a northeast-trending, narrow linear graben whose structure and trend are suggestive of an extensional strike-slip fault. It may be analogous to the "crustal escape" faults hypothesized for the foreland of other collisional systems.

At low-river stage, the Ordovician Cotter Dolomite contact with Precambrian basement is exposed (Fig. 5). The Cotter is equivalent to upper Arbuckle Group carbonate rocks of the southern Oklahoma area. The south canyon wall of the creek exposes Cotter Dolomite overlain unconformably by Devonian Chattanooga Shale and Lower Mississippian St. Joe, Reeds Spring, and Keokuk Formations (Huffman, 1964). Ordovician "Simpson" (Burgen) through Devonian "Hunton" (Sallisaw) rocks are missing at the unconformity because of Middle Devonian tilting and erosion over a wide area of the Ozarks. About 25 mi to the south, exposures along the Illinois River show the Chattanooga Shale unconformably overlying Simpson-age rocks (Fite, Tyner, and Burgen Formations). The Chattanooga Shale and its basal Sylamore Sandstone Member form a relatively uniform blanket over the region and represent a transgressive-systems tract. Oklahoma Highway 20 road cuts near Spavinaw and Oklahoma Highway 10 exposures along the Illinois River offer good outcrops of this shale. At Spavinaw and at Hanging Rock on the Illinois, the Chattanooga Shale has yielded T.O.C. values of 4.5–5.1% (Cities Service Co., unpublished data). The shale is ~60 ft thick here; equivalent Arkansas Novaculite shales and cherts in the Ouachitas are 1,000 ft thick. Both units were deposited during a Late Devonian to Early Mississippian highstand of the sea.

Arbuckle Group–age rocks older than the Cotter Dolomite are not exposed in the western Ozarks but have been penetrated by wells in the southern flank of the Ozark uplift. These Cambrian to Ordovician rocks are dominantly shelf carbonates and are as thick as 2,000 ft. More than 7,000 ft of shallow-water Arbuckle Group carbonate rocks are present in the Ardmore Basin area. In the Ouachita basin, originally south of Broken Bow, Arbuckle equivalents were chiefly fine-grained clastic strata now exposed in the Broken Bow uplift. Their thickness is obscured by complex structure but may be as much as 10,000 ft. Seismic data and deep wells show that the lower Paleozoic platform rocks seen in the Ozarks extend under the Arkoma Basin and Ouachita thrust belt.

At Hanging Rock on the Illinois River, unconformably above the Chattanooga Shale, is an extremely thin section of the St. Joe Limestone (Fig. 6). This unit contains bioherms as thick as 100 ft near Kenwood, ~20 mi to the north. Handford and Manger (1993) placed the St. Joe Limestone in a transgressive-systems tract. It is overlain by the Reeds Spring and Keokuk Formations (equivalent

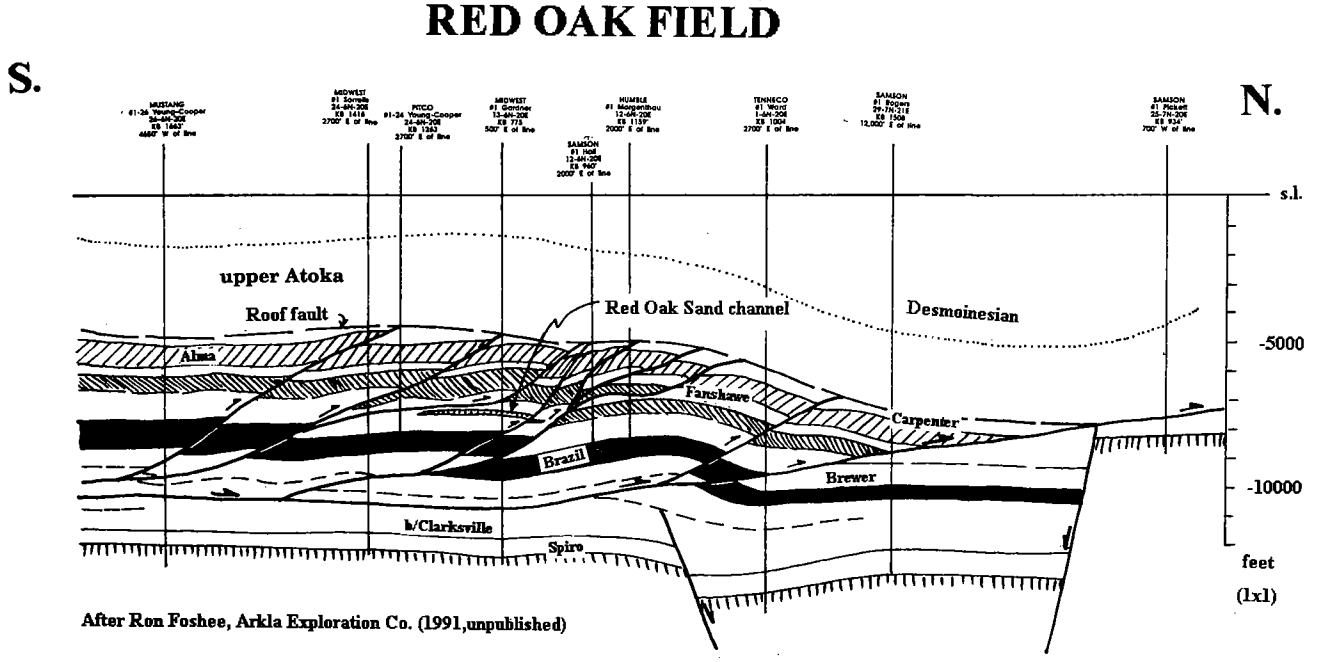


Figure 4. Cross section through Red Oak gas field based on well-log correlations guided by seismic-reflection data. The field is in a duplex whose structure contrasts markedly with the simple surficial folds.

to the Boone of Arkansas), which are cherty limestones placed in a conformable highstand tract by Handford and Manger (1993), but shown to be unconformable with the St. Joe by Huffman (1958). The top of these Boone units is a regional unconformity and is the middle Mississippian sequence boundary. These Lower Mississippian units are only ~250 ft thick here. On seismic lines to the south, the Boone limestones are a strong reflector marking the top of the carbonate shelf. Farther to the south, the reflector shifts down to the Hunton Group as the Mississippian interval changes to shale in the Arkoma Basin.

Younger Carboniferous units are exposed along Oklahoma Highway 10 at Braggs Mountain and near Webbers Falls. At Braggs Mountain, Chesterian Fayetteville Shale and Pitkin Limestone are unconformably overlain by Morrowan Hale and Bloyd Formation carbonates and shale, locally called the Sausbee and McCully Formations (Sutherland, 1979). The Atoka is poorly exposed and unconformable above the Morrowan strata. The Chesterian–Morrowan interval here is <250 ft thick, whereas in the Ouachitas it is >12,000 ft thick. The shallow-water sandy limestones and shales of Morrowan age are 142 ft thick here and are the same age as nearly 3,000 ft of fluvial-deltaic clastic rocks in central Arkansas (Fig. 7) and as much as 7,500 ft of turbidite sandstones, shales, and olistostromal deposits in the Ouachitas (Jackfork and Johns Valley Formations). In the southern Arkoma Basin, equivalent rocks are ~2,000 ft thick and include the Wapanucka Limestone, Cromwell sandstone, and Springer Shale, which are involved in the frontal thrusts of the Ouachitas. Figure 2 shows the large thickness variations of the Carboniferous units along the transect.

The Fayetteville Shale may contain a maximum-flooding interval, followed by a shoaling-upward section that includes the Pitkin Limestone, deposited during a highstand period (Handford and Manger, 1993). The Pitkin Limestone is the best-exposed unit at Braggs



Figure 5. Onlap of Ordovician Cotter Dolomite (C) against 1.3 Ga Spavinaw Granite (PC) knob at Spavinaw Creek below Spavinaw Dam. Bluff in the background exposes Devonian Chattanooga Shale unconformably above Cotter. Mississippian Boone Limestone equivalents cap the sequence (B).

Mountain and contains cross-bedded oolites and crinozoan grainstones. Its top is an erosional unconformity and is the sequence boundary between Mississippian and Pennsylvanian units (Fig. 8). The appearance of sand in the overlying Sausbee limestones is the first hint locally of the progressive east-to-west migration of clastic sediments that overwhelmed the carbonate platform from Mississippian time (Black Warrior Basin) into Morrowan time (central Arkansas) and, finally, Atokan time (fluvial-deltaic clastic rocks in the eastern Oklahoma Ozarks). The basal Atoka Formation a few miles east of Braggs in the Cherokee public hunting area is composed of cross-bedded sandstones interpreted as braided-stream deposits. A few miles to the south at Tenkiller Dam, ~270 ft of fluvial-deltaic clastic rocks of the Atoka Formation are exposed (Visher, 1978).



Figure 6. Bluff along the Illinois River at Hanging Rock exposes Devonian Chattanooga Shale (C) and overlying Mississippian St. Joe (St.J), Reeds Spring (R), and Keokuk limestones (top of hill out of view). Here the Chattanooga unconformably overlies Simpson-age strata that crop out in a road cut south of Hanging Rock. Note pinchout of the St. Joe to left of arrow. View is to north.



Figure 7. Braided-stream facies of the Morrowan section near Almond, Arkansas. This and the thick deltaic facies south of Almond contrast sharply with the thin Morrowan carbonate facies of the eastern Oklahoma Ozarks.

A final locality in the Ozark uplift can also be viewed as an illustration of the geology of the northern edge of the Arkoma Basin: the cliff face at lock and dam 16 near Webbers Falls (Fig. 9). It is a type section for the Sausbee and McCully Formations (~180 ft exposed here; Sutherland, 1979). On the southeast end of the exposure, down-to-the-south normal faults bring deltaic clastic strata of the lower Atoka Formation down against the Morrowan shelfal carbonate rocks. The Atoka Formation here resembles the lower "Cecil-Spiro series" of the subsurface of the Arkoma Basin to the south and east.

The Sausbee Formation is composed of shallow subtidal to intertidal grainstones, algal wackestones, and shales, with channels, bryozoan mounds, cross-bedding, and algal-mound features exhibited on the outcrop. The McCully Formation is a shaly unit bounded by limestones. The Atoka/McCully contact is disconformable, and karst development is present at the top of the McCully.

The basal unit of the Atoka Formation here is a transgressive marine sandstone with limestone pebbles. It is overlain by a marine to prodelta shale, followed by a coarsening-upward deltaic sequence (Fig. 10). The delta-front sandstones are capped by a thin transgressive sandstone. Several other similar deltaic cycles follow up-section. The cyclical, deltaic Cecil-Spiro series of the lower Atoka Formation underlies much of the northern Arkoma Basin. A stratigraphic cross section by Haley (1982), following north-south well control in western Arkansas, indicates that the Morrowan shelf facies and lower Atokan Cecil-Spiro series can be traced 30–35 mi south of the Mulberry fault. Farther south, these units are deep-water facies consisting of slope to basinal shales with turbidite sandstones (Fig. 11). Turbidite sandstones and shales of the lower Atoka Formation in the southern Arkoma Basin and exposed in frontal thrusts of the Ouachitas are as much as 18,000 ft thick (Stone and Haley, 1978). Lower Atoka Formation rocks on the north flank of the Arkoma Basin are <1,000 ft thick.

The normal faults seen at this site are typical of the many small faults that cut the western flank of the Ozarks. They may, in part, be related to the flexure of the crust

due to the tectonic load created by thrusting in the thrust belt to the south. Apparent low-angle detachments seen on seismic data in this and other areas of the southern Ozarks and northern Arkoma Basin, however, are strong arguments for Early Pennsylvanian extension of the crust. It could be thought of as a "pulling" of the edge of North American crust into the subduction zone that must have been to the south at an as-yet-undetermined position. A large down-to-the-basin fault, the Mulberry fault (Pl. 1), ~20 mi to the south, is sometimes used to mark the north edge of the Arkoma Basin.



Figure 8. Unconformable contact between the Chesterian Pitkin Limestone (P) and the Morrowan Sausbee Formation (S) on Braggs Mountain. This important sequence boundary is conformable in Ouachita basinal facies (Fig. 18). The cross-bedded, oolitic Pitkin here is overlain by basal Sausbee limestone with quartz sand grains. The contact is erosional.



Figure 9. Cliff face at Webbers Falls lock and dam 16. Down-to-the-south normal faults bring deltaic facies rocks of the Atoka Formation (A) down against Morrowan Sausbee (S) and McCully (M) Formations, the last of the Paleozoic carbonate-platform units. McCully overlies the Sausbee at top of cliff at right and is downdropped against it on the left.

Arkoma Basin

Surface exposures in the Arkoma Basin are of fluvial-deltaic upper Atokan and Desmoinesian formations. Deeper geologic relations are known chiefly from well and seismic data. Early Atokan and Morrowan facies equivalent to those in the deep subsurface of the southern Arkoma Basin crop out in the frontal zone of the Ouachita thrust belt. The Arkoma Basin is a Pennsylvanian foredeep basin developed over the block-faulted and sunken, earlier Paleozoic shelf. In the north, fluvial-deltaic clastic rocks of chiefly Atokan age overlie the shelf facies. In the south, lower Atokan strata are deep-water clastic sedimentary rocks. Overlying middle Atokan through Desmoinesian strata record a generally shallowing-upward succession that is capped by fluvial-deltaic clastic sedimentary units and coals. Road cuts along Oklahoma Highway 112 and U.S. 59 between Pocola and Heavener expose examples of these shallow-water facies (Fig. 1).

The northernmost anticlines in the Arkoma Basin contain excellent exposures of Atokan deltaic sequences. Near Pocola on Oklahoma Highway 112, on the south flank of Backbone anticline, middle to upper Atokan coarsening-upward clastic sequences are exposed (Fig. 12A,B). These strata have been interpreted variously as intertidal (Chamberlain and Basan, 1978), deep-water outer-shelf edge (Visher, 1978), and deltaic (Stone and others, 1979). They are best interpreted as a river-dominated deltaic sequence prograding over a tidally influenced prodelta section. A coal is found locally at the top of the sequence. The sandstones, here interpreted as river-mouth bar deposits, are correlated with the Upper Alma gas sand of the subsurface, which produced the first gas field in Arkansas (Boyd Haley, personal communication). Upper Atokan deltas viewed in several outcrops in Arkansas show a stronger tidal influence in bed forms of the upper sandstones, perhaps due to the tectonically constricted form of the basin by that time.

In the subsurface of the Backbone anticline, block-faulted Morrowan and older units are structurally detached from shallower Atokan and Desmoinesian rocks by the north-directed Backbone thrust. The fault is interpreted to have had Atokan listric growth-fault movement prior to thrusting (Fig. 13). The Bonanza gas field lies just to the east of this site under this anticline. Production is from the subthrust shelf units in the deeper block-fault terrane. An important deep well here, the Shell No. 1

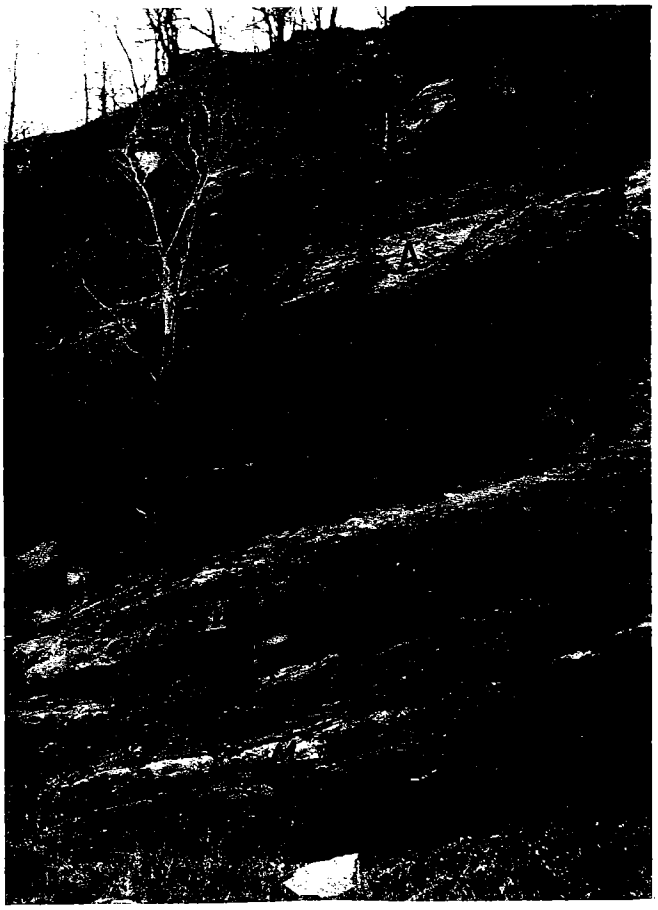


Figure 10. Unconformable contact between McCully Formation (uppermost Morrowan) and basal Atoka Formation clastic rocks at lock and dam 16. Upper McCully shows evidence of karst development. Basal transgressive sandstone (bts) separates McCully (M) limestones and shales from Atoka (A) deltaic cycles. At least three delta cycles are exposed and are correlated with the Cecil-Spiro series of the subsurface Arkoma Basin to the south (left).

Western Coal and Mining (sec. 36, T. 7 N., R. 32 W.), penetrated to basement and passed through ~2,000 ft of Arbuckle Group strata that showed local development of vuggy porosity (third well from the north end, Fig. 13).

In the Poteau-Heavener area are large surface synclines that preserve the youngest foredeep fill: Desmoinesian fluvial-deltaic clastic rocks and coals. Thick delta-front and channel sandstones cap the highest hills in the basin. The synclines are detached above block-faulted Morrowan and older units deep in the basin. The detachment thrust is 14,000 ft deep under Cavanal Hill near Poteau (Pl. 1). In places, especially near the Choctaw fault, surface synclines overlie anticlinal duplex structures formed in Morrowan and Atokan units, which in turn overlie the deep block-faulted shelf terrane. At least one duplex lies 7,000–10,000 ft below the Heavener surface anticline. The Spiro-Wapanucka gas play near Wilburton is targeting such duplexes, and the Wilburton deep Arbuckle field is in the block-faulted subthrust terrane. Near Poteau are two deep wells, the Arkla Downes and Arco Runestone, which tested the deep block-faulted units (Pl. 1). Asphaltic residue was reported in the Woodford Shale through Arbuckle Group strata in these wells (Ron Foshee, personal communication). Giant gas fields occur in this area in Atoka Formation sandstones, for example, the Kinta and Red Oak fields. Both deltaic and turbidite sandstones produce in the basin, and although structure is a factor, stratigraphy plays a major role in entrapment. Evolving Atokan foredeep paleogeography, growth faulting, and sand supply controlled reservoir distribution. The platform rocks that crop out in the Ozarks are 15,000–20,000 ft below the surface in the southern part of the basin.

The Atoka Formation has been estimated to be as much as 27,000 ft thick in Arkansas (Stone and Haley, 1978). Along the line of Plate 1, however, the Atoka Formation is only ~15,000 ft thick, although a small amount has been eroded away. This is about a 50-fold increase in thickness over Atokan units in the Ozarks. A similar ratio holds for Morrowan units that thicken from <200 ft in the Ozarks of eastern Oklahoma to 7,500 ft in the Ouachita thrust belt (Fig. 2).

Ouachita Thrust Belt

Numerous road cuts on U.S. 59/259 between Heavener and Broken Bow (Fig. 1) allow views of Ouachita thrust-belt geology, although the major faults are not well exposed. The trace of the Choctaw thrust can be inferred by contrasting structures in the Heavener and Hodgens areas. About 1 mi south of Heavener, a road cut exposes gently deformed Desmoinesian Hartshorne delta-plain deposits and coal (Fig. 14). About 3 mi farther south (just south of Hodgens), U.S. 59 road cuts exhibit more strongly deformed deep-water clastic rocks of distal submarine-fan and basin-plain facies of the lower Atoka Formation (Fig. 15). The Choctaw thrust, generally considered the frontal Ouachita thrust, has brought to the surface some examples of the Atoka Formation rocks that underlie the shallow-water clastic rocks exposed in the Arkoma Basin just to the north.

The structure of the frontal fan of thrusts of the Ouachitas is complex and generalized in Plate 1. The Choctaw fault trace terminates ~30 mi to the east of this line of section. Through central Arkansas its equivalent is the Ross Creek fault, en echelon to the south of the Choctaw. The lower Atoka Formation clastic strata are part of a large submarine-fan system deposited chiefly by east-to-west paleocurrents. Near Little Rock, Arkansas, lower Atoka Formation fan-channel complexes crop out. From near Perryville to near Danville, sand-rich middle-fan lobes crop out. In eastern Oklahoma south of Hodgens, the distal-fan to basin-plain shales with thin turbidites are exposed. This pattern is clearly visible on Landsat images and is reflected in deformation styles. For example, the relatively continuous sand-rich middle-fan lobes form continuous surface ridges in relatively simple thrust sheets in west-central Arkansas, whereas in eastern Oklahoma, the shaly outer-fan deposits are more complexly deformed and surface topography is of lower relief and more irregular.

About 3.5 mi south of the Choctaw fault, U.S. 59 crosses the Ti Valley thrust, apparently a major overthrust fault that, on a regional scale, brings basinal Ouachita facies far to the north over platform and fore-

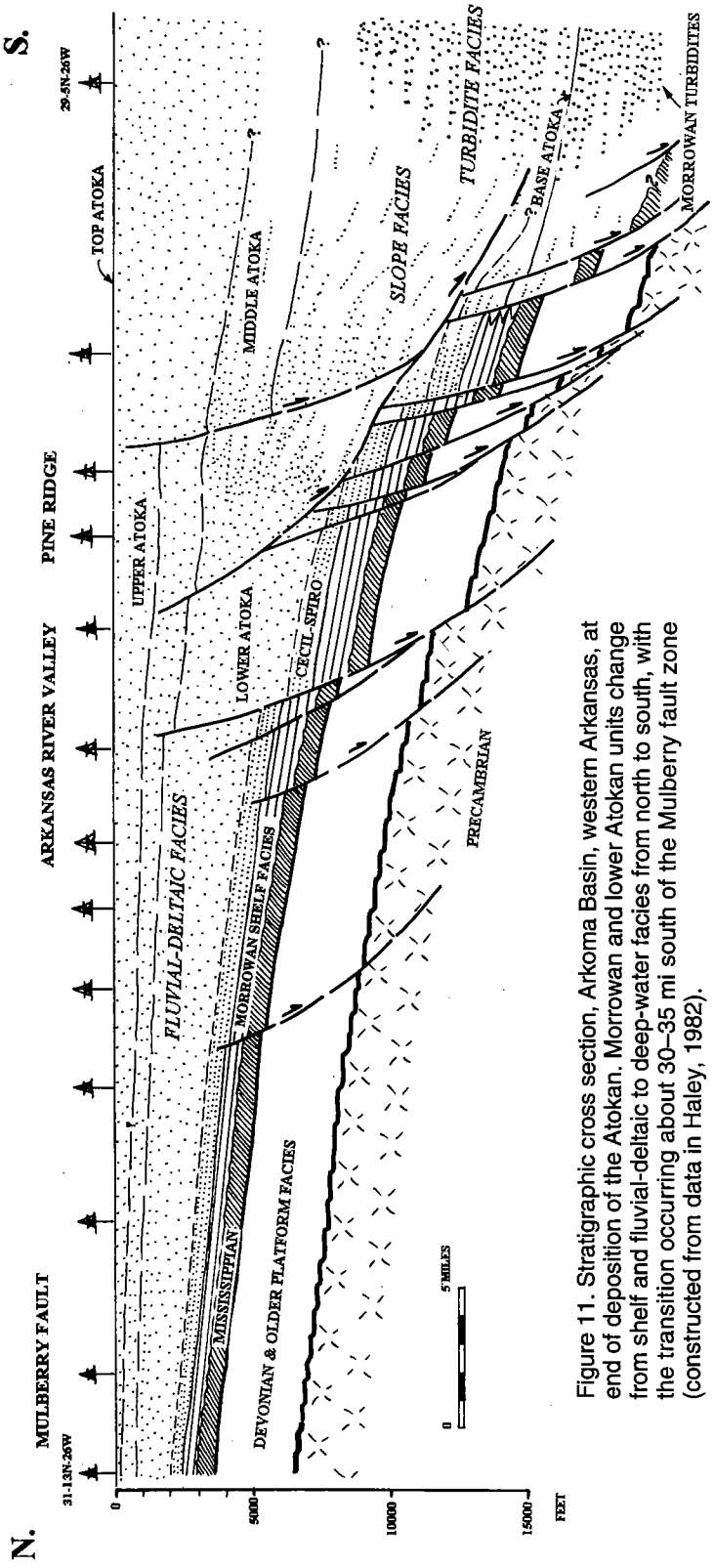


Figure 11. Stratigraphic cross section, Arkoma Basin, western Arkansas, at end of deposition of the Atokan. Morrowan and lower Atokan units change from shelf and fluvial-deltaic to deep-water facies from north to south, with the transition occurring about 30–35 mi south of the Mulberry fault zone (constructed from data in Haley, 1982).

deep units. Upper-plate strata south of the Ti Valley fault may have traveled >45 mi to the north (Pl. 1). In Arkansas, the correlative fault is the Y City fault, separating Arkoma Basin from Ouachita facies. Typical Carboniferous Ouachita facies are exposed in road cuts along U.S. 59/259 between Hodgens and Smithville. The youngest units south of the Ti Valley fault are lower Atokan distal submarine-fan and basin-plain clastic rocks similar to those in the frontal thrust fan. Exposures in the Spring Mountain syncline south of Stapp exhibit fan-fringe lobe sandstones and basinal shales (Fig. 16) with well-developed Bouma sequences as much as 3 ft thick. Paleocurrents flowed generally westward, but some opposed directions are present. The basin may have had a very flat floor at that time. Foraminifera and radiolaria of "deep bathyal" types have been found here (Stone and Haley, 1978). These units are roughly age-equivalent with the Atokan deltaic facies exposed at Webbers Falls (described above). On the basis of seismic data, the slightly overturned syncline exposed here is structurally detached from deeper foredeep and platform units along a low-angle thrust ~16,000 ft below the surface. The platform reflector (Hunton Group?) is ~26,000 ft below the surface. Duplex(es) of Mississippian, Morrowan, and lower Atokan(?) rocks appear to lie between the detachment and platform (Pl. 1). Atoka Formation strata in the Spring Mountain syncline overlie the Jackfork and Johns Valley Formations and Stanley Shale basinal facies above the detachment. The Johns Valley Shale, immediately under the Atoka Formation and of latest Morrowan age, crops out just to the north in the Stapp syncline. Although this classic olistostrome deposit with large exotic blocks has been considered by some to be a tectonic unit, it appears to be a stratigraphic layer. The boulder facies of the Johns Valley Shale occurs mainly in western, northern, and central parts of the thrust belt and contains blocks derived from shelf-facies strata exposed to the west and north and ranging in age from Cambrian to Morrowan (Shideler, 1970). A slump mass of "Caney" Shale 3,000 ft long occurs in the Johns Valley Shale west of here. Near Stapp, the shale contains blocks of carbonate-platform facies (Fig. 17) and a possible disrupted Morrowan slope-channel conglomerate mass. The Johns Valley Shale has been hypothesized to be derived from fault scarps or a now-buried horstlike terrane called "Bengalia" (e.g., Stone and Haley, 1978). A scarp or series of scarps formed during the extension of the crust in Late Mississippian to Early Pennsylvanian time could have supplied the detritus. In a simplistic way, viewed as a simple scarp, to obtain the stratigraphic range observed, the scarp would have to have exposed 5,000–8,000 ft of section. Because the contemporaneous Wapanucka Limestone was deposited near sea level, the Johns Valley basin may have been, therefore, 5,000–8,000 ft deep.

A much-studied thick section of Mississippian Stanley Shale through Morrowan Jackfork Sandstone is exposed on the north flank and crest of Kiamichi Mountain, the north flank of the Lynn Moun-

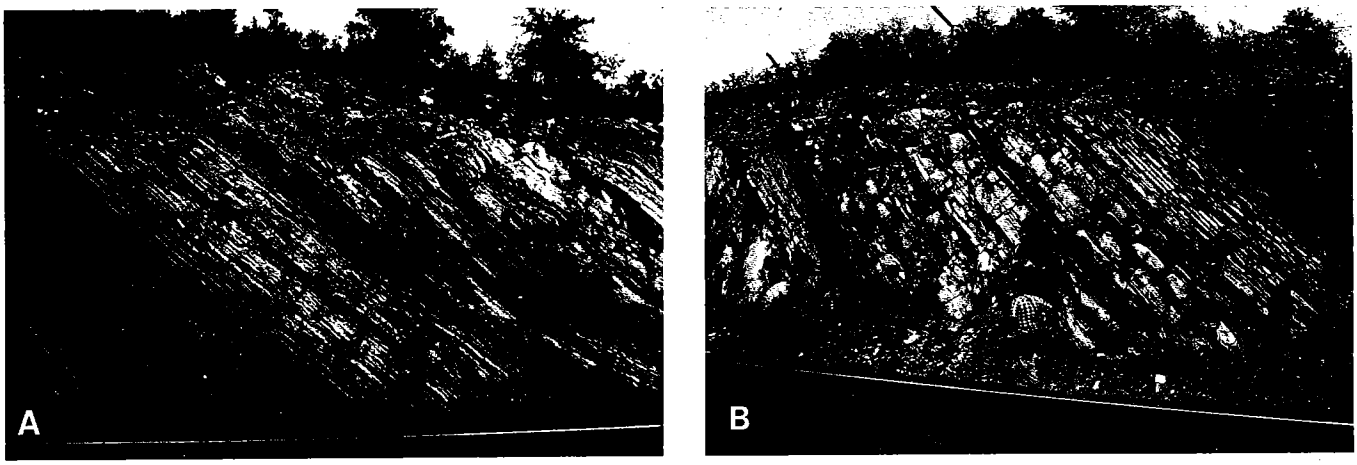


Figure 12. Road cut south of Pocola. (*A*) Coarsening-upward sequences of Atokan clastic rocks are exposed. These strata are delta-fringe and prodelta deposits just below the Upper Alma sandstone *(photo B)*. The rocks contain abundant ripple bedding. (*B*) River-mouth bar and other delta-front sandstones of the Upper Alma (between dashed lines). These units are typical upper Atokan rocks.

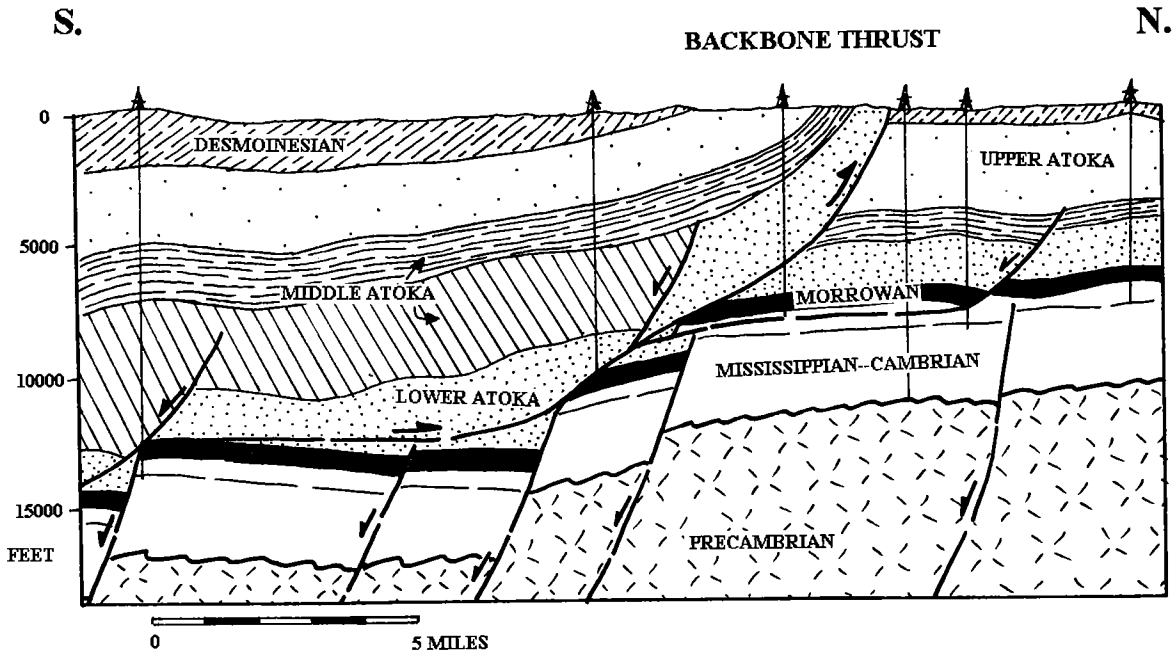


Figure 13. Interpretation of well data across Backbone anticline showing Atokan growth faults reactivated by later Backbone thrust (modified from Buchanan and Johnson, 1968).

tain syncline. This locality is south of the Stapp–Spring Mountain syncline exposures and separated from them by the Simmons Mountain–Rich Mountain syncline and the Windingstair thrust, which crosses U.S. 259 near Big Cedar. Thick, turbidite sandstone–rich Jackfork and Atoka Formation units are preserved in the large synclines in this part of the thrust belt; these "slabs" are simply deformed into broad folds that appear to "float" on more highly deformed shale-rich clastic rocks of the Stanley Shale (Pl. 1). The small-wavelength folds of pre-Stanley units shown in Plate 1 may be separated from the broad folds by duplexlike roof faults (not shown on the cross section).

About 2 mi south of Big Cedar, the Stanley/Jackfork contact is exposed in a road cut on U.S. 259 (Fig. 18). This Chesterian/Morrowan boundary is conformable here, in contrast to being an erosional unconformity on the shelf, as seen at Braggs Mountain (Fig. 8). In fact, the Mississippian/Pennsylvanian boundary is an unconformity over much of North America and is correlated with a worldwide sea-level fall at ~320 Ma. (Ross and Ross, 1988). Handford and Manger (1993) have used a 330 Ma date for this boundary in the Ozarks. In the Ouachita basin, this boundary is a sequence boundary. Stone and Haley (1978) placed the Stanley-Jackfork contact at the first appearance of orthoquartzitic sandstone (compared



Figure 14. Desmoinesian delta-plain facies with coal south of Heavener are representative of the final foredeep fill of the Arkoma Basin. A few miles south of this outcrop the Choctaw fault brings turbidite facies of the lower Atoka Formation to the surface.



Figure 15. Overturned Bouma sequence (top to left) in lower Atoka Formation south of Choctaw fault, Ouachita Mountains.

with the more lithic, feldspathic Stanley sandstones). Below this quartzose sandstone, and separated from it by uppermost Stanley turbidites, is the Chickasaw Creek Shale of the Stanley Group, which is a siliceous shale rich in microfauna (Fig. 19). It is here suggested that the Chickasaw Creek Shale is a condensed section, containing a maximum-flooding interval, and that the "sudden" appearance of quartzose basal Jackfork turbidite sandstones is a response to the sea-level fall at 320–330 Ma. The basal Jackfork represents a "basin-floor fan" on the sequence boundary. The Morrowan was, on the average, a lowstand of the sea (Ross and Ross, 1988) and in the basin is represented by 7,500 ft of turbidites and olistostromal deposits.

From the contact through the crest of Kiamichi Mountain and into the Lynn Mountain syncline, >6,000 ft of Jackfork Sandstone is exposed (Cline and Moretti, 1956). The underlying Stanley Shale is as much as 11,000 ft thick. The Jackfork here is ~60% turbidite sandstone and 40% shale (Fig. 20). The lower Jackfork appears to be a prograding submarine-fan system with shaly lobes near the base of the section and slumped beds and channel sandstones near the crest of the hill, upsection. The turbidite sands were deposited from east to west. Major feeder-channel deposits are exposed near Arkadelphia and at North Little Rock, 100–140 mi east of Kiamichi Mountain. Slope-related slumps entered the basin from the north and are best exposed near Maumelle Lake west of Little Rock. These thick, sand-rich deposits probably resulted from orogenesis in the southern Appalachians. Sedimentation rates began a rapid increase in the basin in Stanley and Jackfork time, culminating in very high Atokan rates (Fig. 2). The preceding Devonian–Lower Mississippian Arkansas Novaculite accumulated at about 14–15 ft/m.y. The Stanley rates were ~500 ft/m.y., and the Jackfork rates were ~1,000 ft/m.y. The Atoka rates may have reached 9,000 ft/m.y. Elsewhere in Jackfork time, thin shelf deposits and unconformities were forming in the Ozarks, thick delta deposits accumulated in north-central Arkansas, Springer shales accumulated on a gentle slope in south-central Oklahoma, and coarse clastic sediments were being shed from the newly emergent Criner Hills in the Ardmore Basin to the west.

Broken Bow Uplift

South of the Boktukola fault (Pl. 1), the surface geology changes dramatically from the broad, open folds in Carboniferous rocks exposed to the north. Outcrops in the Broken Bow uplift reveal small-scale, complex folds and faults, commonly overturned, in places isoclinal, with apparent south vergence in Mississippian to Ordovician basinal strata that are generally metamorphosed to lower-greenschist facies. These strata also are cut by late-stage quartz veins that, in places, reach spectacular proportions. In Arkansas, the equivalent feature is called the Benton uplift; it covers a much larger area and lies farther north with respect to the frontal zone of the Ouachitas, reaching the latitude of Little Rock before disappearing under the younger cover of the Mississippi embayment. Coastal-plain sedimentary units cover the rocks of the uplift south of Broken Bow. Several road cuts along U.S. 259 west of Broken Bow Lake (Fig. 1) allow views of this complex geology, but study of the geologic maps of the region combined with key seismic and well data gives the best insight into the overall structure of the uplift (Fig. 21).

On the south limb of the Linson Creek synclinorium, ~5.5 mi south of Bethel, a U.S. 259 road cut has been a



Figure 16. Distal submarine-fan facies, "fan-fringe lobes" (E. Mutti, personal communication) in lower Atoka Formation near axis of the Spring Mountain syncline, Ouachita thrust belt. Thick sandstones of middle-fan lobes occur in this interval to the east, for example, near Perryville, Arkansas. Paleocurrents flowed to the west.



Figure 17. Outcrop of upper Morrowan Johns Valley Shale near Stapp, Ouachita thrust belt. Boulder is limestone with algal laminae derived from the lower Paleozoic shelf section to the north and deposited as part of submarine flows into the deep Ouachita basin.

popular field-trip stop (e.g., Stone and Haley, 1978; Nielsen and others 1989). Thrusts and disharmonic folds in the lower Stanley Shale are exposed (Fig. 22). Slaty cleavage is present. Tuffaceous sandstones also are present, derived from volcanic sources to the south and deposited by turbidity currents into the basin. The synclinorium is an east-trending, faulted, and complexly folded synform with south-verging small-scale folds superimposed on a north-directed thrust sheet of Ordovician through Mississippian clastic rocks and cherts. Cleavage dips to the north. Coeval carbonate-platform units are interpreted to be beneath a major décollement at depths of >30,000 ft (Pl. 1).

For the next 10 mi to the south, U.S. 259 road cuts provide numerous exposures of the intricate, south-verging faulted folds of the Broken Bow anticlinorium (Figs. 23, 24). Mississippian to Ordovician (Stanley Shale to Collier Shale) basinal facies exhibit polyphase deformation and offset, folded slaty cleavage. Many faults with apparent "normal" displacement are present; these are interpreted as "back-folded" thrusts. Quartz veins cutting these rocks have been dated (by using adularia crystals within the veins) as Late Pennsylvanian–Early Permian (Denison and others, 1977). Overturned folds are common, and isoclinal and recumbent folds are present (Fig. 25). Devonian metamorphic ages have been reported from the Mazarn Shale near here, and Collier and Mazarn samples from a nearby well yielded Pennsylvanian to Early Permian metamorphic dates (Denison and others, 1977).

The older Ordovician basinal clastic rocks (and minor carbonate rocks) are deep-water facies of the shelfal Arbuckle Group rocks exposed in the Ozarks and penetrated by the Sohio No. 1-22 Weyerhaeuser well beneath the Broken Bow uplift. In 1987, Sohio drilled the nearly 19,000-ft well through the slates, phyllites, and calcareous quartzites of the Cambrian–Ordovician basinal Ouachita facies exposed at the surface. At 11,714 ft, the well passed through a thrust and into shelfal Simpson Group and Arbuckle Group facies (Leander and Legg, 1988). This well is interpreted to have

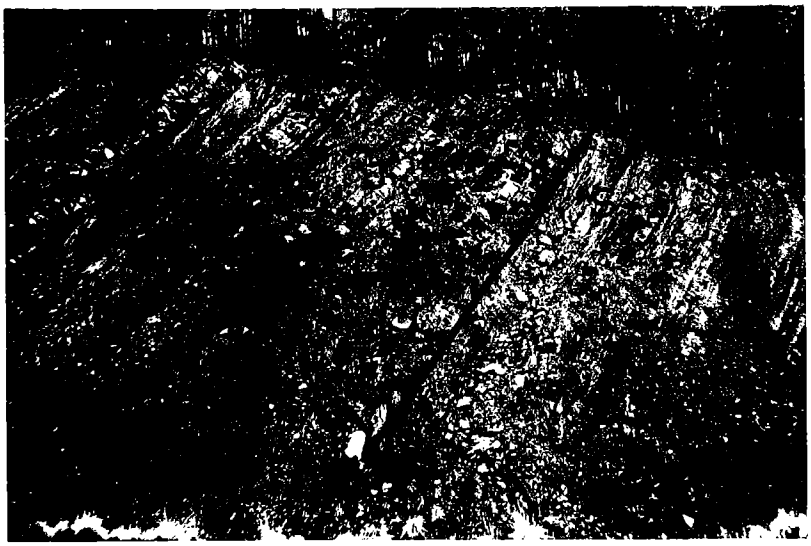


Figure 18. Contact between the Stanley Shale (S) and the Jackfork Sandstone (J) on Kiamichi Mountain south of Big Cedar. This conformable contact is placed at the first appearance of orthoquartzitic sandstones above the more lithic Stanley (C. Stone, personal communication). The basal Jackfork submarine-fan lobes are interpreted to sit on the 330 Ma sequence boundary that marks the Chesterian/Morrowan boundary. The Jackfork is a lowstand deposit, and the underlying Stanley is a highstand deposit.

penetrated the sedimentary upper part of a basement duplex below the thrust (Pl. 1). The Ouachita basinal facies must have been deposited south of the well (which is ~6 mi north-northwest of the town of Broken Bow). This point of control is suggestive of >45 mi of overthrusting of the Ti Valley sheet, which is interpreted to carry basinal facies to the north. Seismic data can be interpreted to indicate pre–Arbuckle Group sedimentary rocks in the uplift; perhaps Cambrian to late Precambrian rift and/or passive-margin fill. The layering may also be due to thrusting or layered basement, however. Similar basement duplexes are present along strike, for example, as revealed by seismic data under the Benton uplift in west-central Arkansas, under the Mississippi embayment in eastern Arkansas, and (as also documented by drilling) in Hill and Lamar Counties, East Texas.

Map analysis also indicates that the Benton uplift in Arkansas contains cover duplexes as well as the deep basement duplex. Such complex deformation probably accounts for the great apparent thickness of the early Paleozoic shaly units. Some of the south vergence of the small-scale folds may be due to "back-folding" of higher sheets by the deeper cover and basement duplexes (Fig. 26). South-verging folds can form early within north-directed larger-scale thrust sheets, however (Fig. 27). At least some of the south-verging folds may have resulted from early deformation in the middle of the thick shale-rich section that is overlain by thick sandstones. The intricate folding may, in part, be due to ductility contrasts. The relatively "stiff" Arkansas Novaculite to Bigfork Chert interval was sandwiched between thick shale packages and formed smaller-wavelength folds during initial movement of larger north-directed thrust sheets.



Figure 19. Chickasaw Creek Shale of the upper Stanley Group on Kiamichi Mountain just below section shown in Figure 18. This siliceous shale, rich in microfossils, is interpreted as a condensed section deposited during maximum-flooding interval in the basin prior to the 330 Ma sea-level fall. Siliceous beds stand out in relief on outcrop.



Figure 20. Jackfork Sandstone, part of a 6,000-ft-thick section on Kiamichi Mountain. Well-exposed submarine-fan facies of the lower Jackfork contrast here with the thin Morrowan shelf facies of the western Ozarks (Fig. 9).

The Broken Bow uplift was deformed in at least three phases (Feenstra and Wickham, 1975). The first two were associated with northward-moving thrust sheets, and the last was related to formation of the basement duplex under the thrust complex, folding it into an anticlinorium. This late-stage basement involvement is interpreted to have been due to final stages of continental collision between the basements of North America (which was driven back on itself) and the southern continent (Yucatan?). The older deep-basement extensional detachment and the normal-fault complex above it are interpreted to have been utilized during collision to form the duplex (Pl. 1).

DISCUSSION

In many ways the Ouachita orogenic system has been enigmatic. As late as 1957, at least some geologists thought the belt lacking in major thrusts (Misch and Oles, 1957), although Miser (1929) illustrated major thrusts earlier. Considerable differences exist in the literature in interpretations of structures and structural styles of both surficial and deep rocks. The purpose of this paper is to present a regional cross section that ties the major elements of the Ouachita system (Ozark uplift, Arkoma Basin, and Ouachita thrust belt, including the Broken Bow uplift) together. This cross section is based on industry wells and seis-

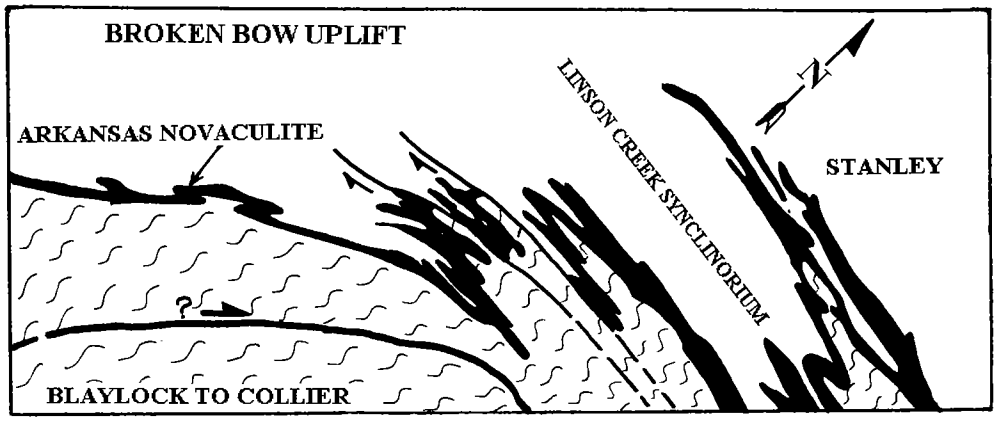


Figure 21. Down-plunge view of northwest flank of the Broken Bow uplift reveals its complex structure even when viewed from a generalized structure map (map redrawn from Nielsen and others, 1989).



Figure 22. Disharmonic folds in the Stanley Shale on south flank of the Linson Creek synclinorium (north flank of Broken Bow uplift). North is to right. Faults are directed northward; folds are slightly overturned to south. Outcrop is just south of Bethel.

mic-reflection data in critical areas and those surface exposures along the section that illustrate the major stratigraphic units of the eastern Oklahoma Paleozoic succession. Certain elements of the cross section (Pl. 1) are as accurately represented as possible at the scale used; for example, the basement block faults under the Arkoma Basin are based on seismic and well data. Other parts of the cross section are more diagrammatic, for example, the fold-and-fault pattern of small-scale features in the Broken Bow uplift. An important difference between this cross section and previous ones published for the region is the interpretation that a deep basement detachment exists beneath the region extending from the southern flank of the Ozarks under the Arkoma Basin and Ouachita thrust belt to the core of the Broken Bow uplift. The presence of a detachment is supported by intrabasement reflections on seismic data on the north flank of the Arkoma Basin in Oklahoma and Arkansas against which high-angle normal faults terminate. Cover duplex structures have been targets of gas exploration for several years along the frontal part of the Ouachitas. Plate 1 was drawn to carry the duplexes far to the south under the thrust belt; this interpretation is supported by seismic data at least as far south as Kiamichi Mountain. Duplexes are inferred to be present in cover rocks of the Broken Bow uplift as well as in the basement in the core of the uplift.

The cross section was originally constructed to provide a regional tectonic setting for outcrops along it that I have visited on many field trips. By zigzagging across the countryside, it is possible to put together a paleogeographic scenario of certain time slices, e.g., the Morrowan, where dramatic thickness and facies changes can be seen from the thin carbonate shelf in northeastern



Figure 23 *(left)*. Overturned folds in the shaly Blaylock Sandstone, Broken Bow uplift, west of Broken Bow Lake. South is to right.



Figure 24. Overturned fold in the Arkansas Novaculite, Broken Bow uplift, west of Broken Bow Lake. Fold is overturned to south.

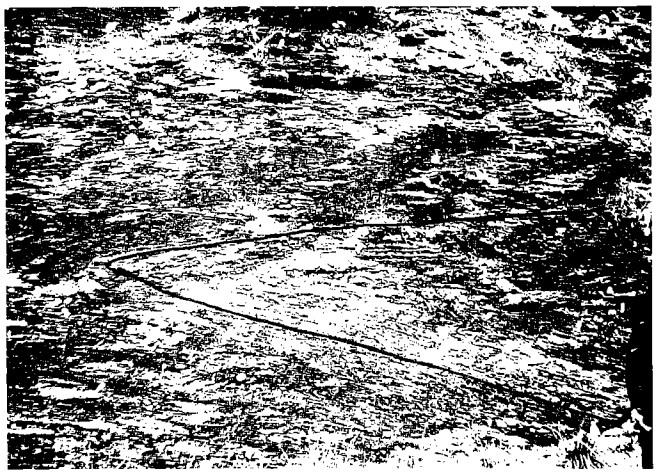


Figure 25. Recumbent fold in the Mazarn Shale, Broken Bow uplift, west of Broken Bow Lake. Cleavage is well-developed. This Ordovician basinal clastic unit is equivalent to shelfal limestones of the Ozarks represented by the Cotter (Fig. 5). The lower Paleozoic shelfal strata were penetrated by a deep well near this outcrop; they are separated from these highly deformed coeval basinal facies by a thrust fault.

Oklahoma to the braided-stream sandstones and thick deltaic sections of central Arkansas and the very thick submarine-fan sequences of the Ouachitas. But the deep subsurface relations and many facies transitions cannot be illustrated in outcrop. This fact is especially true for the lower Paleozoic shelf units exposed in the Ozarks, which plunge under the Arkoma Basin and Ouachita thrusts and are involved in a basement duplex under the surficial Broken Bow uplift rocks; basinal strata coeval with the shelf units have been thrust over the shelf units by tens of miles. The early Paleozoic shelf extended at least as far south as Broken Bow and has been buried by late Paleozoic foredeep fill and thrust sheets for a distance of at least 100 mi along this transect.

This 170-mi-long transect illustrates only "half" of the continental collisional orogenic system. The other part lies beneath the coastal plain and under land rifted away (Yucatan?) when the Gulf of Mexico was created. On the North American foreland side of the collision, cratonal arches formed; these are continental analogues of the oceanic crustal bulges observed seaward of trench-subduction zones, including the Ozark uplift and its sister arches, the Pascola, Nashville, and Cincinnati. The Arkoma Basin and similar features, the Black Warrior, Appalachian, and Fort Worth Basins, are foredeeps—the continental analogue of trenches.

The Arkoma Basin, however, may not be a classic foredeep in the sense of its being formed chiefly by loading of the crust by the advancing thrust belt. The extensive basement block faulting and inferred deep, low-angle basement detachment zone suggest Late Mississippian–Early Pennsylvanian foreland extension toward the subduction complex that must have separated the North American plate from the southern plate. By middle Atokan time it was a more classic foredeep, but Late Pennsylvanian–Early Permian compression detached the fill from the underlying shelf-basement complex. The extensional deformation of the early Paleozoic shelf and basement in the Late Mississippian and Early Pennsylvanian caused the subsidence that forced the translation of the shelf edge far to the north and created the deep basin that received the thick upper Stanley, Jackfork, and lower Atoka turbidites. This tectonism allowed accumulation of thick, deep-water deposits of Morrowan and Atokan age during a period of worldwide lowstand of the sea on top of what had been a shallow-water shelf for most of the Paleozoic. The voluminous clastic-sediment influx in Carboniferous time following what had been a long, sediment-starved interval (Fig. 2) was a result of Appalachian orogenesis to the east. Despite rising sea levels throughout the rest of the Pennsylvanian and the worldwide highstand in the Permian (Ross and Ross, 1988), a generally shallowing-upward sequence followed from middle Atokan through Desmoinesian time until emergence of the area in the Permian(?). This depositional pattern was caused by (1) voluminous clastic-sediment

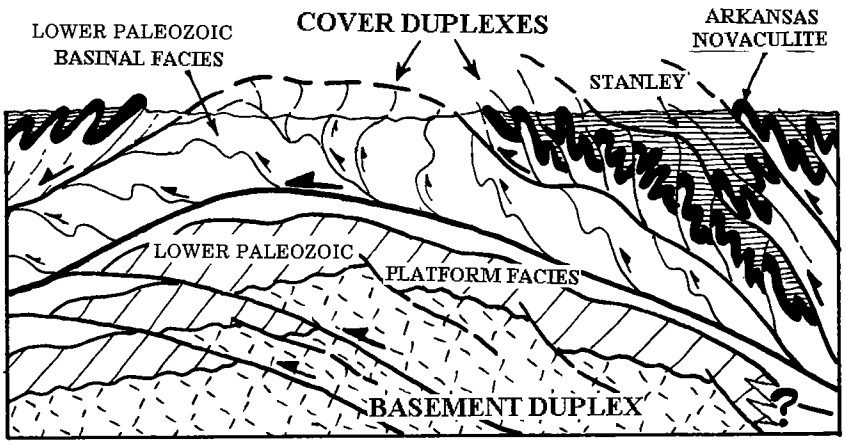


Figure 26. Diagrammatic model for stacked-duplex interpretation of the Broken Bow uplift, based on map analysis and an interpretation of the Sohio No. 1-22 Weyerhaeuser well. North is to left.

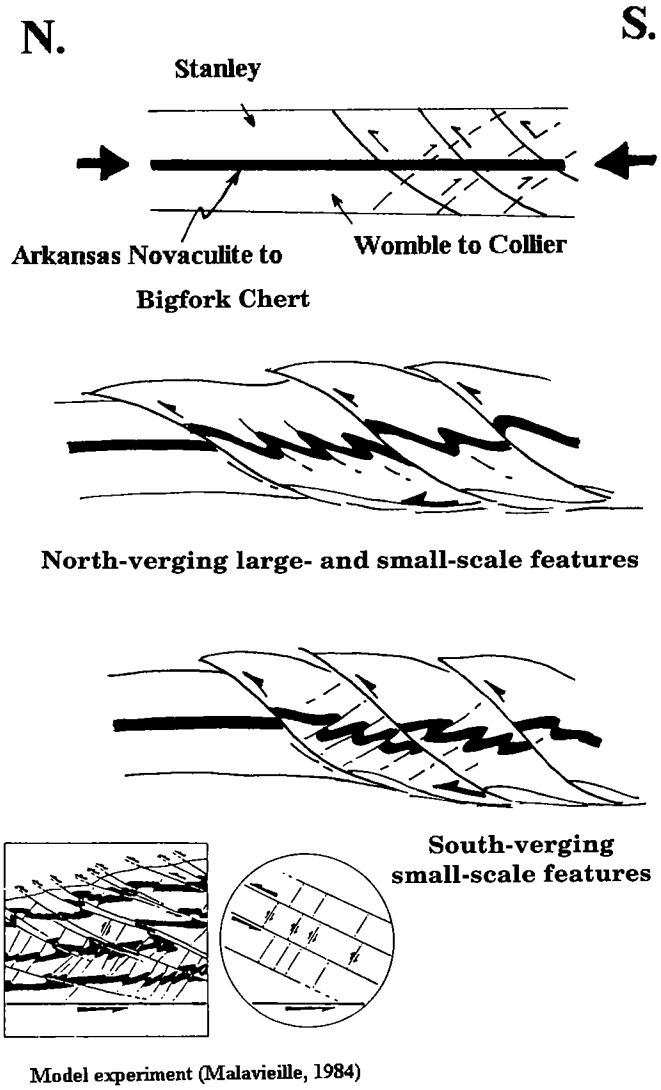


Figure 27. Model for the development of south-verging small-scale structures in the Broken Bow uplift. Stiff chert layers fold with smaller wavelength in middle of shale pile. "Back thrusts" are as likely as forward ones.

influx, (2) tectonic constriction of the basin by the advancing thrust sheets, and (3) late uplift related to the final stages of collision.

Structures of the Ouachitas are complex and incompletely known. The thrust belt has characteristics of a classic Alpine-type collisional system and some features analogous to the accretionary wedges of oceanic subduction zones. The Makran foldbelt, which includes a submarine thrust belt forming today in the Arabian Sea (offshore Pakistan) (illustrated by Lehner and others, 1983) may be a modern analogue for the early stages of development of the Ouachita thrust belt, which was submarine until very late in its history. The relatively simple, submarine, imbricate Ouachita thrust belt was modified by a late-stage continental collision that caused the emplacement of a basement duplex obliquely under the belt. Some of the Ouachita structural-style departures from the Appalachians and younger Rocky Mountain thrust belts are the result of lithologic differences. Instead of ductile clastic sedimentary rocks overlying thick, thrust-faulted carbonate units, the Ouachitas contain thick sandstone-rich units overlying more ductile, thick, shale-rich units. The Late Pennsylvanian to Early Permian basement-duplex creation arched the thrust belt and gave it a "tectonic head" for renewed northward movement, compressing the foredeep fill and detaching it from the underlying shelf and basement blocks (in places utilizing preexisting listric normal-fault planes). The basement compression also was transmitted via the preexisting block fault–detachment system to locally reactivate these older normal faults as basement reverse faults as far north as the Wilburton gas field near the Choctaw fault.

Figure 28 summarizes the Paleozoic tectono-stratigraphic history of this southern edge of the North American craton. The southern margin was a passive margin developed in the late Precambrian to Cambrian and relatively stable into the Devonian (Fig. 28A). Thin shelf-carbonate rocks seen in the Ozarks thicken slightly toward the shelf edge, which was at least as far south as Broken Bow. To the south, thick slope and basinal clastic sediments accumulated. Perhaps deformation began far to the south in the Devonian, on the basis of scattered metamorphic dates. Deformation is more strongly hinted at by the appearance of tuffs, lithic sandstone tur-

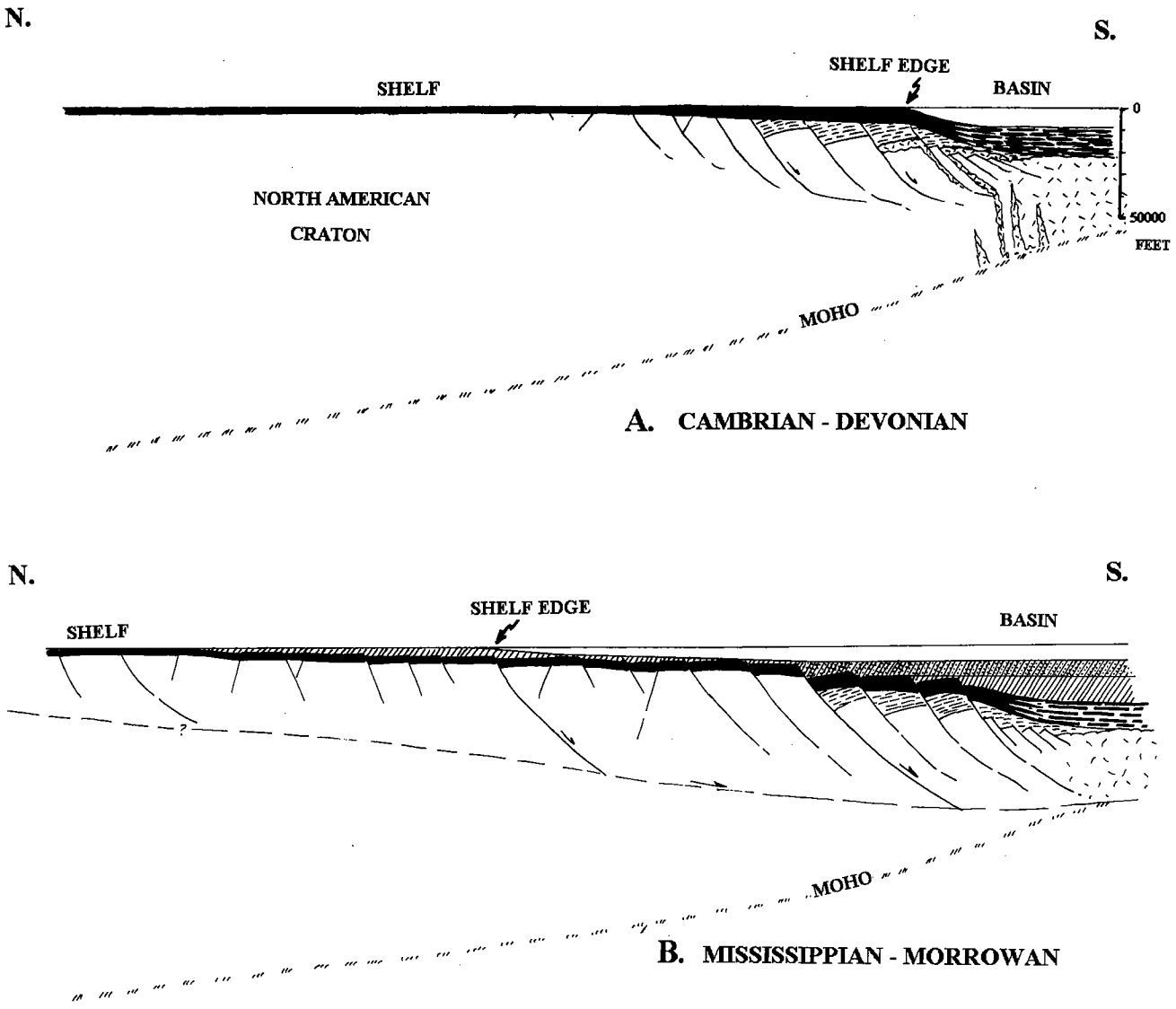
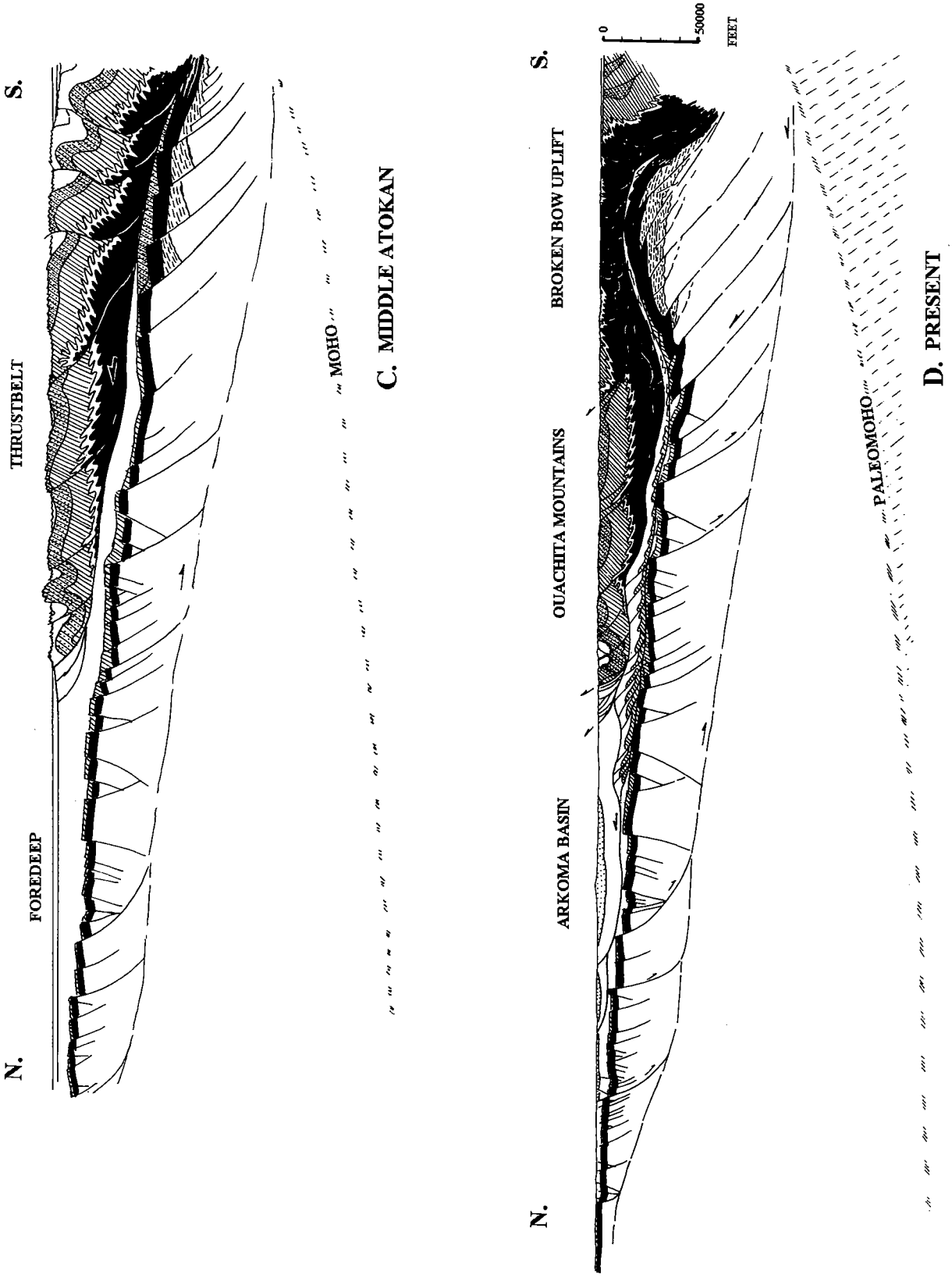


Figure 28 *(above and opposite page).* Historical development of the Ouachita region: (*A*) Cambrian–Devonian stable margin with moderate to slow sediment-accumulation rates. Areas now in Ozarks, Arkoma Basin, and exposed Ouachita thrust belt were part of a broad, shallow-water shelf. (*B*) Mississippian–Morrowan extensional block faulting of the shelf caused subsidence and translated the shelf edge far to the north. Turbidites shed from rising Appalachians to east accumulated in the basin. (*C*) Middle Atokan thrust sheets overrode block-faulted platform and constricted the Atokan foredeep basin. (*D*) Late Pennsylvanian to Early Permian basement-duplex formation redeformed the thrust belt and deformed the foredeep. The present picture results from erosion and subsidence related to the formation of the Gulf of Mexico.


N.
S.
FOREDEEP
THRUSTBELT
MOHO
C. MIDDLE ATOKAN
N.
S.
ARKOMA BASIN
OUACHITA MOUNTAINS
BROKEN BOW UPLIFT
0
50000
FEET
PALEOMOHO
D. PRESENT

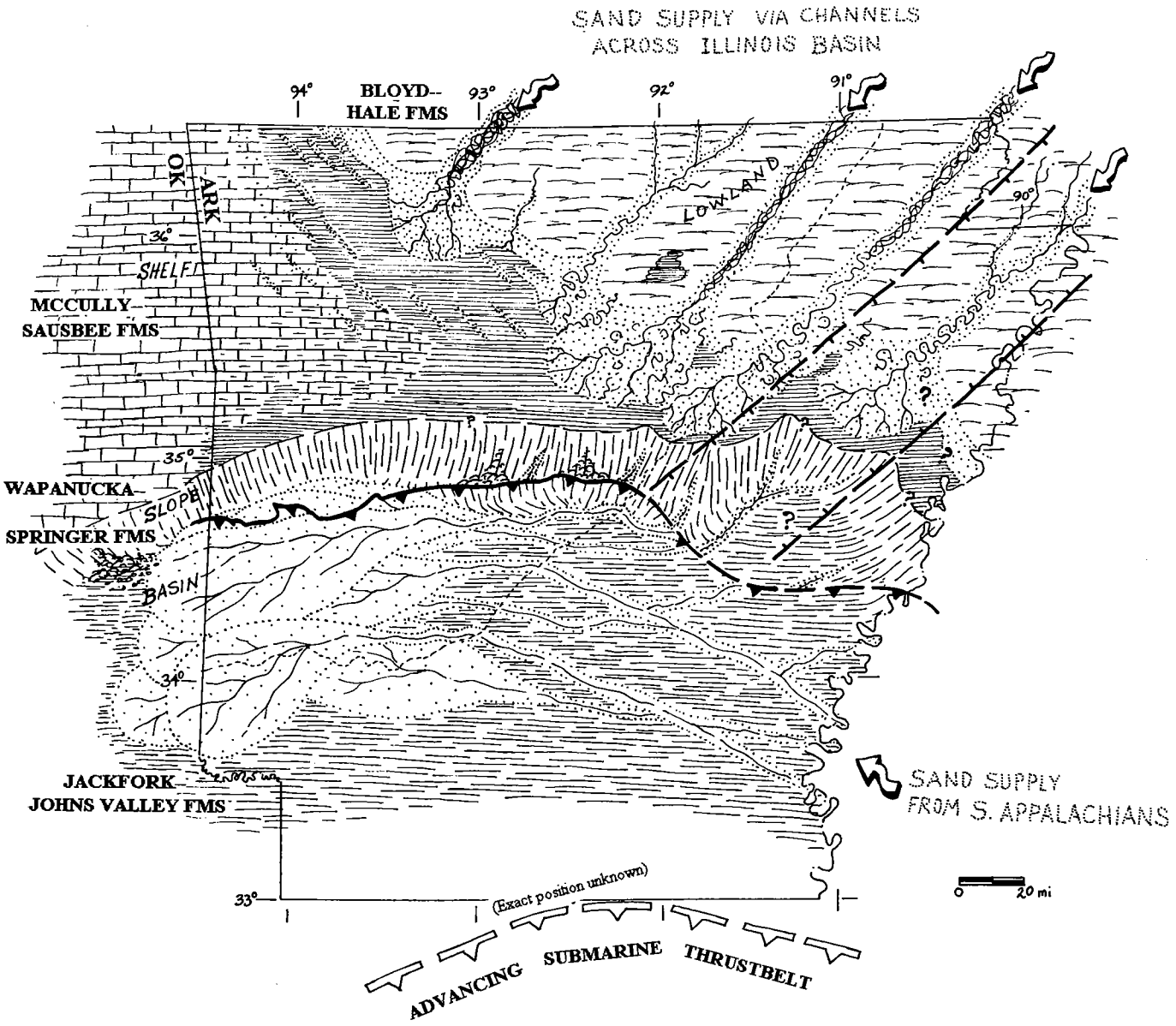


Figure 29. Paleogeographic map of Morrowan time. Not all facies are exact contemporaries, but map is representative of the interval. Shelf edge in central Arkansas is interpreted from seismic data. Units south of Y City fault (solid barbed line) have been transported at least 45 mi north of original site of deposition.

bidites, and an increased sedimentation rate in the Meramecian–Chesterian Stanley Shale. In the Chesterian–Morrowan, the shelf was broken by extensional faulting and subsided (Fig. 28B). The shelf edge shifted far northward, and by the end of the Morrowan, the basin had accumulations of thick, sandstone-rich turbidite submarine fans, capped by olistostromes shed from still active fault scarps perhaps under what is now the frontal thrust zone. In the western Ozarks, an unconformity separates thin Chesterian and Morrowan carbonate strata, and in central Arkansas, Morrowan fluvial-deltaic sequences buried the carbonate platform (Fig. 29). The early Atokan resembled the Morrowan (Fig. 30), but apparently in a short time interval between the early and middle Atokan(?), the basinal section was thrust northward over the earlier Paleozoic shelf for a distance of >45 mi (Fig. 28C). Another way of describing this event is as a rapid subduction episode where the attenuated cratonal margin was subducted under the accretionary wedge. Rates of collision may have been in the range of 6–10 cm/yr. The tectonically restricted foredeep basin filled with clastic sediments into the Late Pennsylvanian. The resistance of buoyant continental crust to subduction may have caused the deep duplex creation, forming the Broken Bow uplift near the end of the Pennsylvanian or in the Early Permian. Erosion of the system, perhaps not emergent until the uplift of the Broken Bow anticlinorium, and subsidence related to the Mesozoic creation of the Gulf of Mexico, led to the present conditions (Pl. 1; Figs. 28D,31).

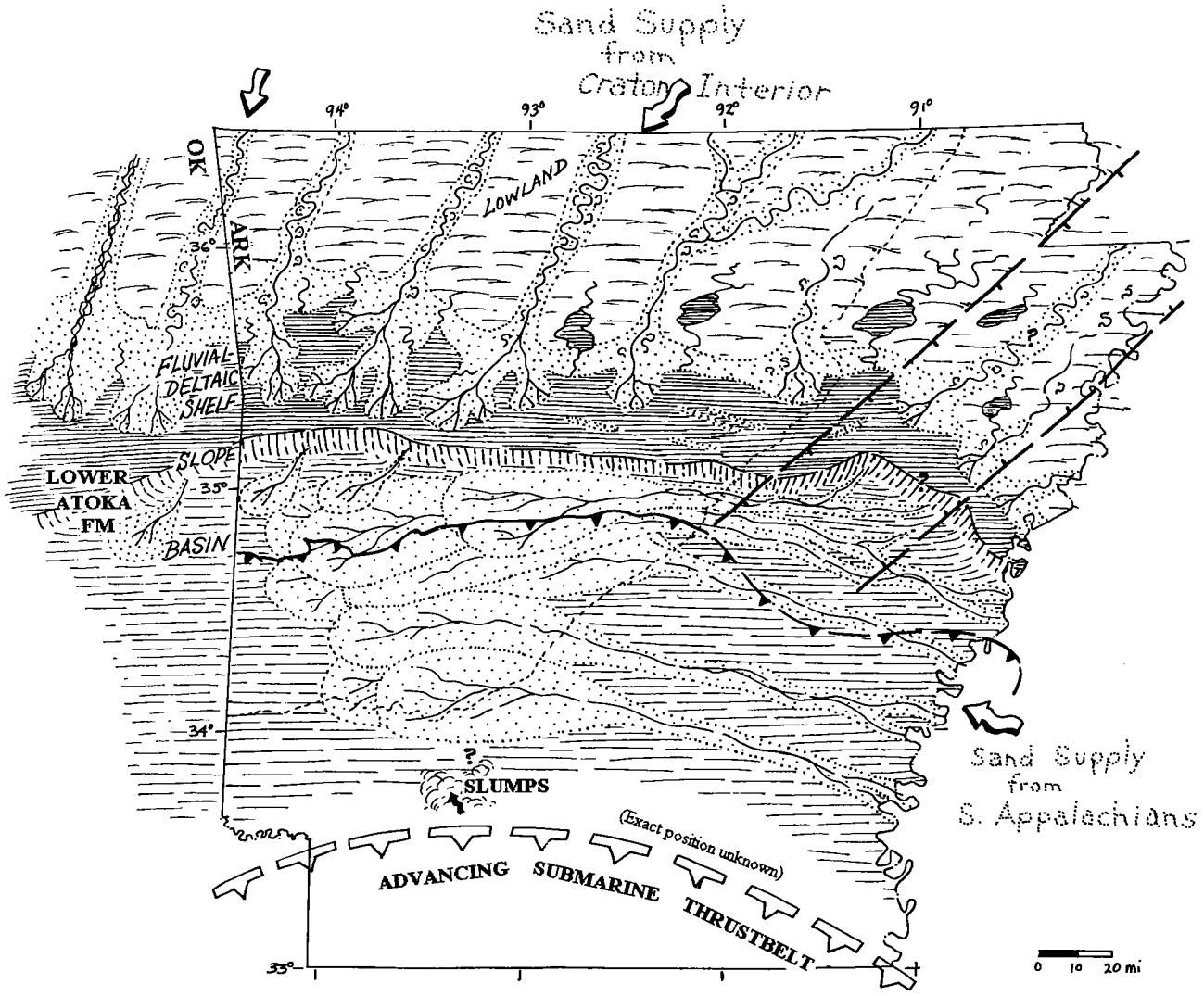


Figure 30. Paleogeographic map of the early Atokan. In the later Atokan the deltaic facies prograded southward and covered the deep-water deposits from the north while thrust sheets to the south deformed them.

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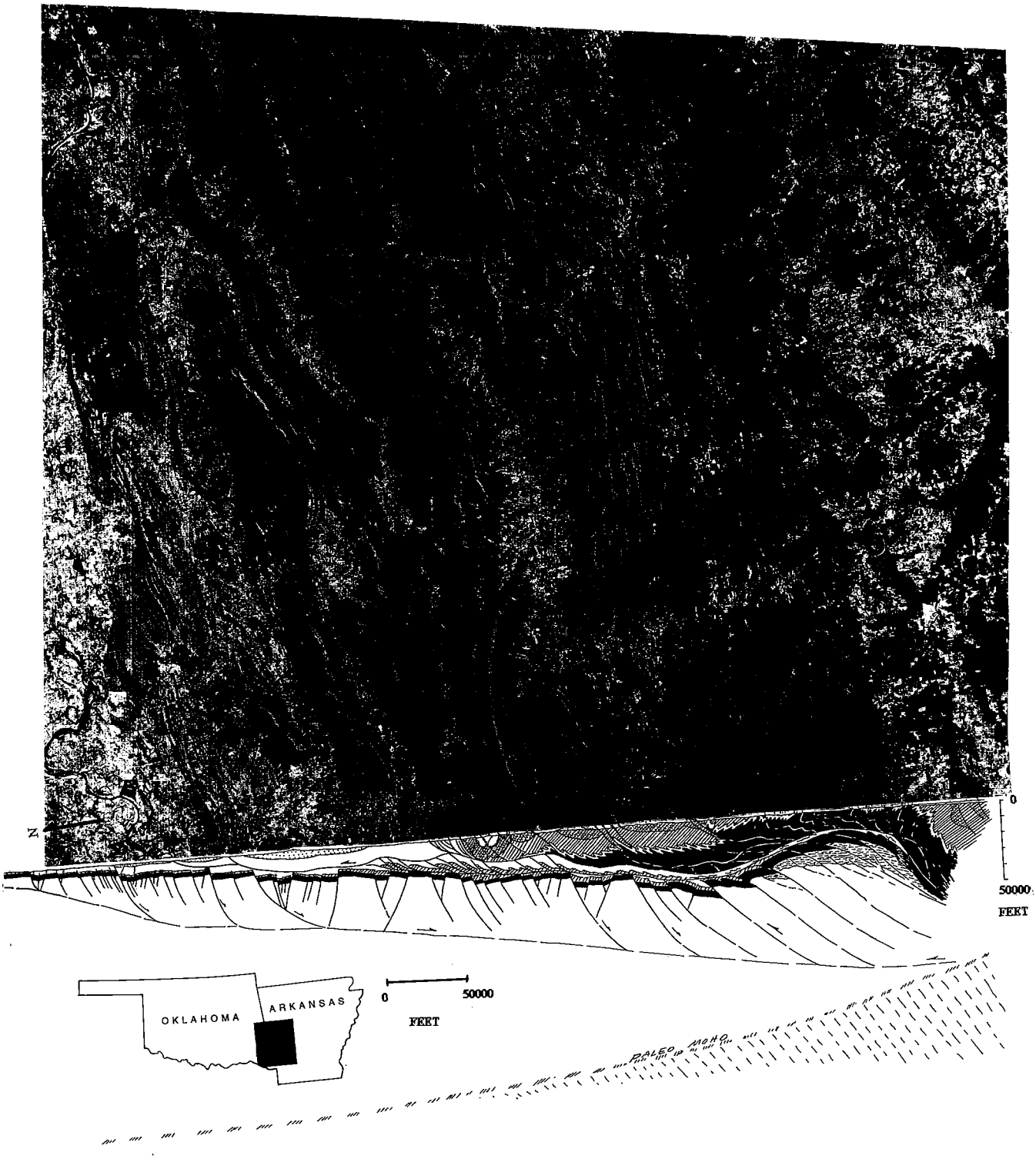


Figure 31. Cross section with Landsat image of the surface of the Arkoma Basin and Ouachita Mountains at the same scale. View shows surface expression of the more resistant sandstone, chert, and novaculite units from the line of section in eastern Oklahoma into western Arkansas. Thin ridges on the north end are Atoka Formation deltaic beds. Ridges along the Ouachita frontal zone include Atoka submarine-fan facies. Jackfork fan facies form the synclinal ridges in the middle part of the image. Intricate topography on the south side reflects the Broken Bow–Benton uplift complex. Location of cross section is shown on inset map and Figure 1.

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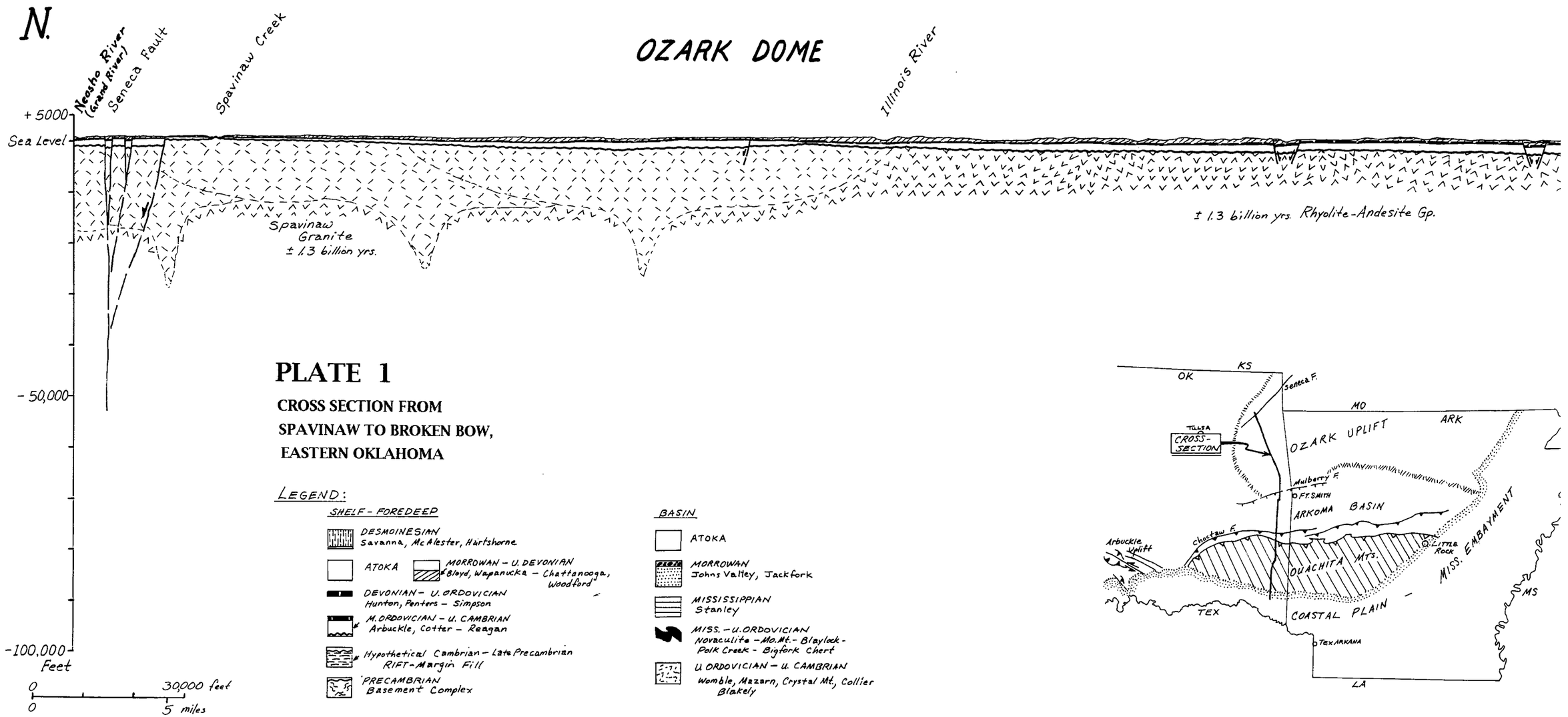


PLATE 1

CROSS SECTION FROM SPAVINAW TO BROKEN BOW, EASTERN OKLAHOMA

LEGEND:

SHELF - FOREDEEP

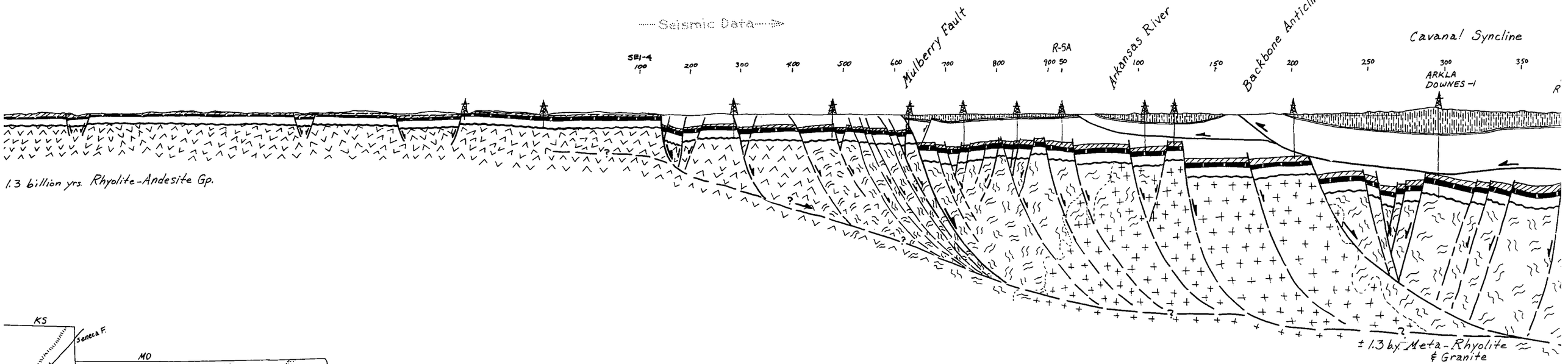
- DESMOINESIAN — Savanna, Mc Alester, Hartshorne
- ATOKA
- MORROWAN – U. DEVONIAN — Boyd, Wapanucka – Chattanooga, Woodford
- DEVONIAN – U. ORDOVICIAN — Hunton, Penters – Simpson
- M. ORDOVICIAN – u. CAMBRIAN — Arbuckle, Cotter – Reagan
- Hypothetical Cambrian – Late Precambrian RIFT-Margin Fill
- PRECAMBRIAN — Basement Complex

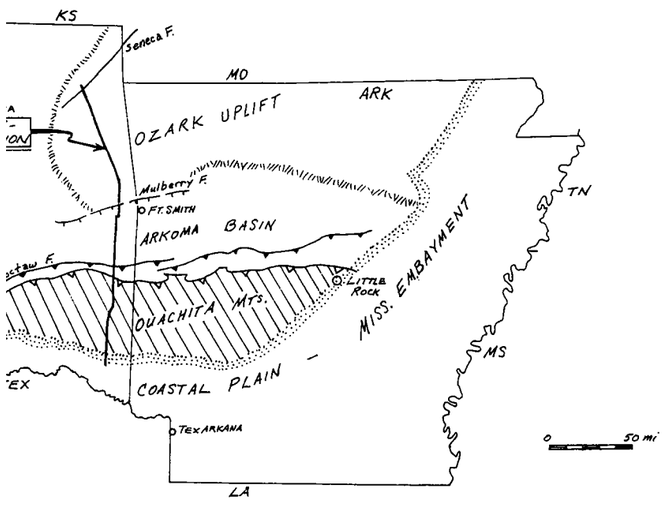
BASIN

- ATOKA
- MORROWAN — Johns Valley, Jackfork
- MISSISSIPPIAN — Stanley
- MISS. – U. ORDOVICIAN — Novaculite – Mo. Mt. - Blaylock - Polk Creek - Bigfork Chert
- U. ORDOVICIAN – u. CAMBRIAN — Womble, Mazarn, Crystal Mt., Collier Blakely

Michael T. Roberts, 1992

ARKOMA BASIN





BASIN

OUACHITA MTS

a.l Syncline

Heavener Anticline

Choctaw Fault

Ti Valley Fault

Seismic Data

Windingstair Fault

Briery Fault

Big Cedar

Kiamichi Mt.

Octavia Fault

Boktukola Fault

BROKE

ES-1

ARCO RUNESTONE-1

CGG-103

350 400 450 500 550 600 900 600 300

Rhyolite

anite

"PALEO-MOHO"

OUACHITA MTS

S.

nic Data

300

Windingstair Fault

Big Cedar

Klamichi Mt.

Octavia Fault

Boktukola Fault

BROKEN BOW UPLIFT

SOHIO - WEYERHAEUSER 1-22

Broken Bow

COASTAL PLAIN

K

Sea Level

-50,000

-100,000

Feet

0 30,000 feet

0 5 miles

"PALEO - MOHO"

Conodont Biostratigraphy of the Missouri Mountain Shale (Silurian–Early Devonian?) and the Arkansas Novaculite (Devonian), Black Knob Ridge, Atoka County, Oklahoma

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Texas Tech University
Lubbock, Texas

Jill N. Haywa-Branch
U.S. Department of Agriculture
Shelby, Montana

ABSTRACT.—Conodonts permit dating of the Missouri Mountain Shale and Arkansas Novaculite of the Ouachita facies in the Black Knob Ridge area and correlations to be effected to North America shelf successions. The lower half of the Missouri Mountain Shale is late Llandoverian, Telychian (Silurian), in age, on the basis of *celloni* and *amorphognathoides* Zone conodont faunas. A major hiatus encompassing most of the Llandoverian separates the Missouri Mountain from underlying Ashgillian (Ordovician) strata. The association of conodonts with terrigenous sandstones suggests that most conodonts were derived from an adjacent shelf area with the coarse clastic debris during a lowstand in sea level that produced a Telychian unconformity on the shelf. Sparse conodont faunas from the upper half of the Missouri Mountain Shale range in age from the Wenlockian, through the Ludlovian, to near the Silurian/Devonian boundary.

The green-gray spiculitic cherts that comprise the lowest unit of the Arkansas Novaculite (Unit 1) are interpreted to be Emsian to Eifelian (Early–Middle Devonian) in age, on the basis of a few elements of a distinctive *Dvorakia* species. The hiatus at the base of the Arkansas Novaculite spans most of the Early Devonian. The Penters Chert (Arkansas) and Sallisaw Formation (Oklahoma) are the most likely shelf correlatives. Interbedded dark cherts and shales that compose the middle of the Arkansas Novaculite (Unit 2) range in age from the Givetian (Middle Devonian) into the Frasnian (Late Devonian). The most diagnostic conodont faunas belong to the late Givetian *varcus* and *hermanni* Zones and occur in the middle of Unit 2. At the Grants Gap section, Unit 2 rests directly on the Missouri Mountain Shale. Here, Unit 1 of the Arkansas Novaculite was removed by a mid-Devonian episode of erosion or nondeposition; the Unit 1/Unit 2 contact is also interpreted to be an unconformity. The lower few meters of the green and red siliceous shales that form the upper part of the Arkansas Novaculite (Unit 3) contain early Fammenian (Late Devonian) conodonts. Units 2 and 3 of the Arkansas Novaculite correlate with the Clifty Formation and the overlying Sylamore Sandstone–Chattanooga Shale section of the Arkansas Ozarks.

The pattern of discrete lithologic units separated by regional unconformities observed in Late Ordovician through Devonian units at Black Knob Ridge corresponds closely with that known from the shelf successions, suggesting a common response to regional tectonic or eustatic events. We propose that the Late Ordovician through Devonian depositional history preserved in the Black Knob Ridge area may also apply to coeval Ouachita facies strata in the core area of the Ouachita Mountains.

INTRODUCTION

The combination of limited surface exposure, structural complexity, and low-grade metamorphic overprints, exacerbated by inadequate chronologic control on depositional and tectonic events, has led to contradictory models for the origin and early evolution of the Ouachita orogenic belt. The most enigmatic portion of the Ouachita rock succession is the relatively thin interval, ~1,000 m thick, of Upper Ordovician through Lower Mississippian siliceous shales, cherts, novaculites, and lesser sandstones and limestones that compose the upper part of the offshelf, preorogenic suite of the Ouachita facies (Fig. 1; Lowe, 1989; Viele and Thomas, 1989). In the Ouachita Mountain region, relatively little is known about the age relations of younger preorogenic strata.

Barrick, J. E.; and Haywa-Branch, J. N., 1994, Conodont biostratigraphy of the Missouri Mountain Shale (Silurian–Early Devonian?) and the Arkansas Novaculite (Devonian), Black Knob Ridge, Atoka County, Oklahoma, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 161–177.

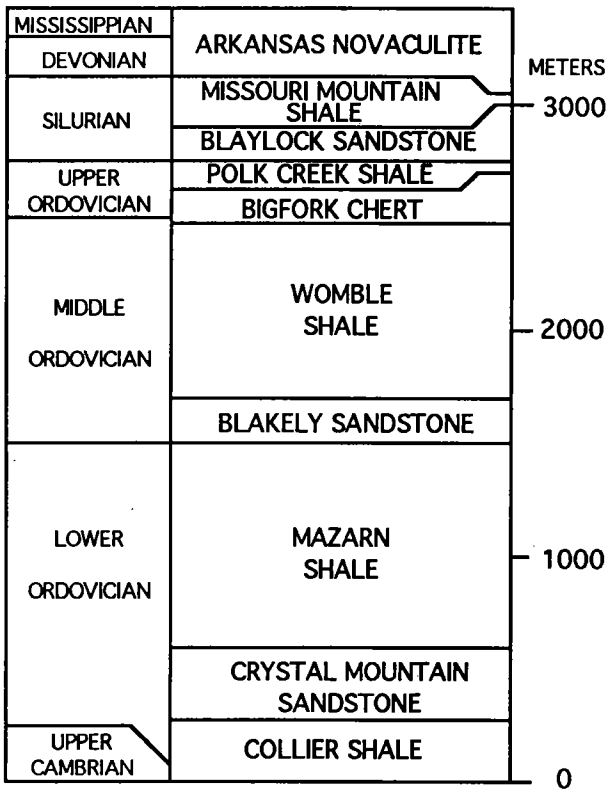


Figure 1. General stratigraphic column of preorogenic strata of the Ouachita facies in the core area of the Ouachita Mountains. Formations are scaled relative to stratigraphic thickness in the southern part of the Benton uplift, Arkansas. Modified from Lowe (1989).

Although recent work on graptolites (Finney, 1986) and conodonts (Repetski and Ethington, 1977; Ethington and others, 1989; Krueger and Ethington, 1991) has begun to clarify ages of Ordovician units, no comparable progress has been made in Silurian and Devonian strata. Most existing biostratigraphic information on the Silurian–Devonian section consists of isolated reports of fossils, some more than 60 years old, for which there has been little recent confirmation. In this paper, we present results of our work on conodont faunas from Silurian and Devonian strata of the Ouachita facies in the Black Knob Ridge area of Oklahoma. Although these data were obtained from only a small part of the Ouachita Mountain region, they have important implications for understanding the sequence of Silurian and Devonian depositional events in the Ouachita orogenic belt and their relation to events on the North American shelf.

PREOROGENIC STRATA IN THE OUACHITA FACIES IN THE OUACHITA MOUNTAINS

Depending on how one interprets the sedimentology and facies relations, preorogenic strata of the Ouachita facies were deposited either in a setting seaward of the continental shelf extending into deep-water oceanic settings (Viele and Thomas, 1989) or in deep, narrow intracratonic basins analogous to those of the Mississippi Valley graben and the Southern Oklahoma basin (Lowe, 1985,1989; Arbenz, 1989). The lower interval of Upper Cambrian to Middle Ordovician strata includes shales, sandstones, and subordinate carbonates that contain strong evidence for deposition in deeper-water slope to rise environments. Sedimentary features indicative of gravity transport and deposition dominate (Lowe, 1989), and the units contain diverse graptolite faunas and conodonts characteristic of the cool water, deeper water North Atlantic Province (Repetski and Ethington, 1977; Ethington and others, 1989). The existence of boulders of limestone, chert, and granitic rocks in debris-flow deposits suggest a source area proximal to the early Paleozoic shelf (Viele and Thomas, 1989), as does the rare occurrence of shallow-water, warm-water conodonts of the cratonic North American Midcontinent Province (Repetski and Ethington, 1977; Ethington and others, 1989). Lowe (1985,1989), though, argued that some of the material was derived from a block bounding the southern edge of the riftlike Ouachita trough. The clastic sediment–dominated Upper Cambrian through Middle Ordovician sequence of the Ouachita facies contrasts strongly with the thicker carbonate-dominated section of the southern margin of North America. Viele and Thomas (1989) have pointed out that the widespread unconformities bounding stratigraphic sequences of the shelf and craton have not been recognized in the Ouachita Mountains.

During the Middle Ordovician, the pattern of deposition in the Ouachita facies began to change. The supply of terrigenous clastic detritus gradually decreased, and the relative proportion of organic components, silica, and carbonaceous matter increased (Lowe, 1989). In the Ouachita Mountains, the Middle to Upper Ordovician Bigfork Chert consists of a mixture of siliceous, carbonaceous, and fine carbonate material deposited in deep-water settings (Finney, 1986), and the overlying Upper Ordovician Polk Creek Shale is a dark, fissile, carbonaceous, graptolitic shale. During the Silurian, coarse, terrigenous clastic sediment flooded into the deep-water basin from source areas to the south or east, depositing the Blaylock Sandstone, before a return to shale-dominated deposition as represented by the Missouri Mountain Shale (Lowe, 1989). Deposition of siliceous sediment may have started as early as the Silurian and dominated from the Devonian into the Early Mississippian, forming the complex series of white novaculites and varicolored radiolarian-bearing cherts and shales that compose the Arkansas Novaculite. Because of the unusual nature of these strata and poor biostratigraphic information, the depositional setting, origin of siliceous sediment, facies relations, age of constituent members, and paleogeographic significance of this formation remain highly controversial, as is true for the lithologically similar unit in the Marathon region, the Caballos Novaculite (Arkansas Novaculite—Sholes, 1977; Lowe, 1985,1989; Viele and Thomas, 1989. Caballos Novaculite—McBride and Thomson, 1970; Folk, 1973; Folk and McBride, 1976; McBride and Folk, 1977; McBride, 1989). The influx of voluminous quantities of coarse, terrigenous clastic sediment along the length of the Ouachita orogenic belt during the Mississippian heralded the onset of orogenic activity and the filling, from east to west, of a remnant ocean basin (Thomas, 1989).

Unlike the underlying strata, the general sequence of Upper Ordovician to Lower Mississippian lithologic units in the Ouachita facies in the Ouachita Mountains has invited comparison and correlation on lithologic grounds with the shelf successions of Oklahoma and Arkansas. What biostratigraphic information that is available thus far has tended to confirm these correlations.

Graptolites have been reported from numerous sections of the Middle to Upper Ordovician Bigfork Chert and Upper Ordovician Polk Creek Shale. Ruedemann and Decker (1934) noted the similarity between graptolite faunas of the Bigfork Chert and the carbonates of the Viola Group in the Arbuckle Mountains. Finney (1986, 1988) recently studied the graptolite succession in the Bigfork Chert in the Black Knob Ridge area and the Viola Group, confirming and improving these correlations. Decker (1935) indicated that the graptolite fauna of the Polk Creek Shale is like that of the Sylvan Shale in the Arbuckle Mountains, a fauna that has been assigned to the Ashgillian *Dicellograptus complanatus* Zone of Berry (Ethington and others, 1989).

Miser and Purdue (1929) discussed the graptolite collections of Ulrich from the Blaylock Sandstone in the Caddo Gap Quadrangle, Arkansas. As currently interpreted (Ethington and others, 1989), this fauna suggests correlation with the middle Llandoverian *Monograptus gregarius* Zone. Pitt and Spradlin (1964) reported that in McCurtain County, Oklahoma, Late Ordovician graptolites, including *Dicellograptus complanatus*, occur in the lower part of the Blaylock.

The Blaylock is absent in several areas along the western and northern edges of the Ouachita area, where the Missouri Mountain Shale rests on Ordovician strata (Potato Hills, Oklahoma—Pitt and others, 1982; Black Knob Ridge, Oklahoma—Hendricks and others, 1937; Board Creek, Arkansas—Sellars, 1967; and northeastern sections in Arkansas—Sterling and others, 1966). In some areas where it does rest on the Blaylock Sandstone, the base of the Missouri Mountain Shale has been interpreted locally as being an unconformable contact, on the basis of the presence of conglomeratic beds at its base (Sellars, 1967). In many other sections, though, the Blaylock/Missouri Mountain Shale contact is indistinct owing to a complete lithologic intergradation between the two units.

The Missouri Mountain Shale has yielded only a few fossils, all described from the Black Knob Ridge area. Hendricks and others (1937, p. 10) reported that "a few conodonts, spicules, and fragments of brachiopods together with one bryozoan" were found in the upper part of the formation at Grants Gap and stated that this fauna appeared to be Silurian in age. Harlton (1953, p. 788) indicated that he and J. V. Howell collected a few conodonts and bryozoans at the same locality, one bryozoan of which had close Henryhouse affinities. Because of this collection, Harlton (1953) correlated the Missouri Mountain Shale with the Upper Silurian Henryhouse Formation, an argillaceous carbonate unit in the Arbuckle Mountains. Honess (1923) had earlier reported two beds of carbonates bearing fossil debris and pyroclastic material from near the top of the Missouri Mountain Shale in McCurtain County, Oklahoma, but did not identify any fossils. On the basis of interpreted facies relations, some authors have suggested that the top of the Missouri Mountain Shale may be a time equivalent of the lower part of the overlying Arkansas Novaculite (Sellars, 1967; Lowe, 1989). This suggestion would make the top of the Missouri Mountain Early Devonian in age, or the base of the Arkansas Novaculite would be Late Silurian in age.

Miser and Purdue (1929) and subsequent authors have discussed the lack of equivalent clastic strata on the carbonate-dominated shelf areas, but emphasized the presence of fine argillaceous material in Silurian carbonates in Tennessee and Oklahoma. Only the Upper Ordovician–Lower Silurian Cason Shale of Arkansas, excluded from consideration by these early workers because of its assumed Ordovician age, possesses any significant quantities of clastic material on the shelf succession (Craig, 1993). More recent correlation charts of Arkansas strata (e.g., Stone and others, 1986) often represent the Blaylock Sandstone as being equivalent to the upper Cason Shale through Lafferty Limestone interval and the Missouri Mountain Shale as lacking correlative strata in the Ozark–Arkansas Valley section. Similarly, correlation charts for the Oklahoma shelf region show the Blaylock and Missouri Mountain Shale as being the time equivalents of part of the Hunton Group (e.g., Decker, 1959; Ham, 1959).

The Arkansas Novaculite has traditionally been subdivided into three members or divisions, a lower novaculite division, a middle chert and shale division, and an upper novaculite division. Sholes (1977) subdivided the formation into five members: lower chert and shale member, lower novaculite member, middle chert and shale member, upper novaculite member, and upper chert and shale member. The distinct changes in lithology that occur when traced northward from Hot Springs, Arkansas, have been mentioned by many workers, including Purdue and Miser (1923), Lowe (1977,1989), and Sholes (1977). In these more northern exposures, the recognition of divisions or members is difficult because of the absence of the white novaculite units. This northern facies (Lowe, 1977) is dominated by black chert, shale, sandstone, siltstone, and chert- and shale-pebble conglomerates.

Because the lower novaculite member of the Arkansas Novaculite has yielded only sparse fossil material, its age and correlation to the shelf sequence have been based on stratigraphic position and lithologic similarities. The only fossil identified from the novaculite is a single brachiopod illustrated by Honess (1923) from near the top of the lower member novaculite, which was identified as *Leptocoelia flabellites* (Conrad) by Schuchert, and apparently confirmed by Ulrich, in C. L. Cooper (1935). G. A. Cooper, in Amsden (1983), reexamined this specimen and considered it to be indeterminate to the generic or specific level. Supported in part by the reported Early Devonian age of this brachiopod, most workers considered the lower novaculite to be Early Devonian or Early–Middle Devonian in age. Lithologic correlations were suggested, most commonly to chert-rich Lower–Middle Devonian units on the shelf areas: Camden Chert of Tennessee (Miser, 1917; Miser and Purdue, 1929; Cooper and others, 1942), Bois d'Arc Formation in the Arbuckle Mountains of Oklahoma (Hendricks and others, 1947), Sallisaw Formation in northeastern Oklahoma (Miser,

1944), Penters Chert of northern Arkansas (Giles, 1935; Miser, 1944; Kinney, 1946; Thomas, 1989), and Pinetop Chert of the frontal zone of the Ouachita Mountains (Cooper and others, 1942; Miser, 1944; Honess, 1923; Amsden, 1983).

Amsden (1983) reviewed the brachiopod fauna of the Pinetop Chert and listed a small ostracode fauna identified by R. F. Lundin. He concluded that the Pinetop could be correlated with the lower Lochkovian Haragan–Bois d'Arc sequence in the Arbuckle Mountains. In contrast to the Devonian age proposed by most workers, Pitt and others (1982) reported that L. R. Wilson had obtained palynomorphs from the lower Arkansas Novaculite in the Potato Hills that Wilson correlated with species from the middle Silurian (late Llandoverian–Wenlockian) Prices Falls Shale of the Arbuckle Mountains.

The best age information for the Arkansas Novaculite is based on conodont faunas obtained from the middle chert and shale member (Cooper, 1931,1935; Hass, 1951). Hass (1951) recovered Late Devonian conodonts from between 56 and 14 m below the top of the middle chert and shale member (Middle Division) at Caddo Gap; the lower 50 m of the member did not yield any conodonts. The species of *Palmatolepis* listed from this interval indicate an age range from the middle Frasnian into the middle Fammenian, with the boundary between the two epochs lying somewhere between the collections taken at 56 and 42 m below the top of the unit. Kinderhookian (Mississippian) taxa (*Siphonodella* species) appear at 8.5 m and extend up to 3.5 m below the top of the middle chert and shale member. The presence of *Gnathodus punctatus* indicates that this fauna may lie near the top of the Kinderhookian. Additional collections of conodonts from the middle chert and shale member have been reported by other workers, each supporting the general Late Devonian–Early Mississippian age assignment proposed by Hass (1951), but not providing any further constraints on the age of the unit (e.g., Walliser, in Stone, 1978, p. 115). These conodont faunas supported the original lithologic correlations, inferred by earlier workers, showing that Arkansas Novaculite above the lower novaculite member was equivalent to the Misener Sandstone–Woodford Shale of the Arbuckle Mountains and Sylamore Sandstone–Chattanooga Shale of the Ozark region (Hass, 1951).

The age of the lower part of the middle chert and shale member and its equivalents in the shelf succession remains less clear. The possibility that Middle Devonian strata may be present has never been excluded (Hass, 1951), nor demonstrated. Arkansas correlation charts show that the Clifty Formation, of probable late Middle Devonian age (Amsden, 1988), is a possible correlative of the lower part of the middle chert and shale member. Sholes (1978) recovered palynomorphs from a level 10 m above the top of the middle chert and shale member at Black Knob Ridge. The identified taxa, *Desmochitina* sp. and *Maranhites* sp., suggest a possible late Middle Devonian age.

Primarily by stratigraphic position, the upper novaculite has been correlated with the Boone Limestone of the Ozark region and the Welden Limestone of the Arbuckle Mountains, both of which are essentially Osagean in age. Hass (1951) reported that a small conodont fauna from a shale interbed in the upper novaculite near Hatton, Arkansas, included species of *Gnathodus* without *Siphonodella*, suggesting an Osagean or younger Mississippian age for the novaculite.

Conodonts (Hass, 1950; Elias, 1959) occurring as low as 23 m above the base of the overlying Stanley Shale are interpreted to be Chesterian (Late Mississippian) in age (Ethington and others, 1989). Other faunas from the Stanley Shale give ages that are no older than the Meramecian (Gordon and Stone, 1977).

BLACK KNOB RIDGE SUCCESSION, ATOKA COUNTY, OKLAHOMA

Hendricks and others (1937) provided a detailed description of the structure and stratigraphy of the Ouachita facies at Black Knob Ridge, to which only minor additions need to be made. The Upper Ordovician through Lower Mississippian lithologic sequence at Black Knob Ridge differs somewhat from what is described as the typical Ouachita succession and has been interpreted as representing a transitional area between the Ouachita facies and the shelf facies. Locations of the three sections studied in the Black Knob Ridge area are shown in Figure 2.

Polk Creek Shale

Finney (1986,1988) has recently redescribed the Bigfork Chert at Black Knob Ridge and discussed the significance of the graptolite faunas. The overlying Polk Creek Shale is variable in thickness, ranging up to a maximum of 43 m in southern sections. North of Stringtown, the Polk Creek is absent, such as at the Grants Gap section (Fig. 2), except for one locality where 1 m of shale remains (Hendricks and others, 1937). The upper part of the Polk Creek Shale is a a soft, brown to gray, clayey shale at the sections examined in this study. In addition to the distinctive brown to gray color, the presence of layers of Ordovician graptolites distinguishes it from the overlying units.

Missouri Mountain Shale

The Blaylock Sandstone of the core region of the Ouachita Mountains is absent, and the Missouri Mountain Shale lies directly on Upper Ordovician strata. The Missouri Mountain Shale is primarily a greenish clay shale, although siliceous shale, chert, sandy shale, and sandstone occur at certain levels at different sections. Like the underlying Polk Creek Shale, the Missouri Mountain is generally poorly exposed, and the contact between the units is difficult to examine.

During this research the opening of a shallow pit just south of the East Atoka Road Cut section (Fig. 2) partly exposed the two units in the floor of the pit. Here, the brown graptolitic shale of the upper Polk Creek is overlain by 2 m of gray and brown clay shale lacking graptolites. Above this interval, green-gray clay shale with numerous reddish oxidized zones occurs, which was interpreted to be the Missouri Mountain Shale. The pit reveals ~30 m of green-gray shale; the true thickness is difficult to determine because of minor structural deformation. Although the exposure is incomplete, no resistant sandstone or chert beds were observed in the pit.

At the Asphalt Pit section (Fig. 3), 7 m of brown graptolitic clay shale of the Polk Creek crops out in the bank of a small stream and is overlain by 3.5 m of gray shale. Above a 2.4-m-thick covered interval lies a ledge-forming, 40-cm-thick chert-granule conglomerate that marks the base of the Missouri Mountain Shale. Only a little more than 6 m of the formation is exposed above the conglomerate. The lower 3 m consists of greenish-gray clay shale with zones of interbedded thin (2–10 cm) sandstones. At a level just over 3 m above the conglomerate, red clay shale appears, and the red color dominates over green up to the top of the exposed section, which is marked by a 10-cm-thick sandstone, colored black by asphalt.

The Polk Creek Shale is absent at the Grants Gap section, and the Missouri Mountain Shale rests on the Bigfork Chert (Fig. 4). Hendricks and others (1937) and Sholes (1978) gave descriptions of this section, which

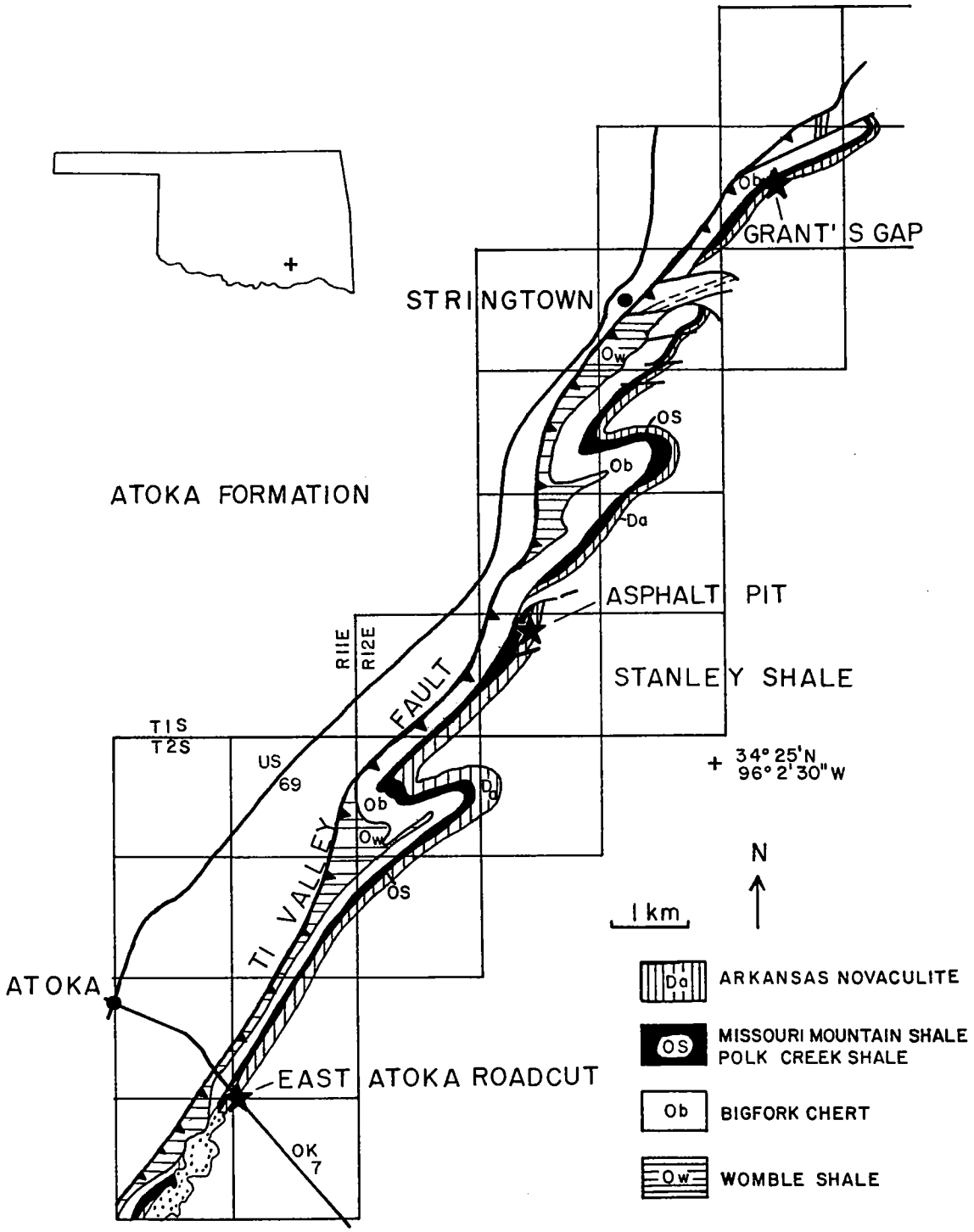


Figure 2. Geologic map of the Black Knob Ridge area, Atoka County, Oklahoma, showing locations of measured sections. Modified from the geologic map of Hendricks and others (1937, fig. 2).

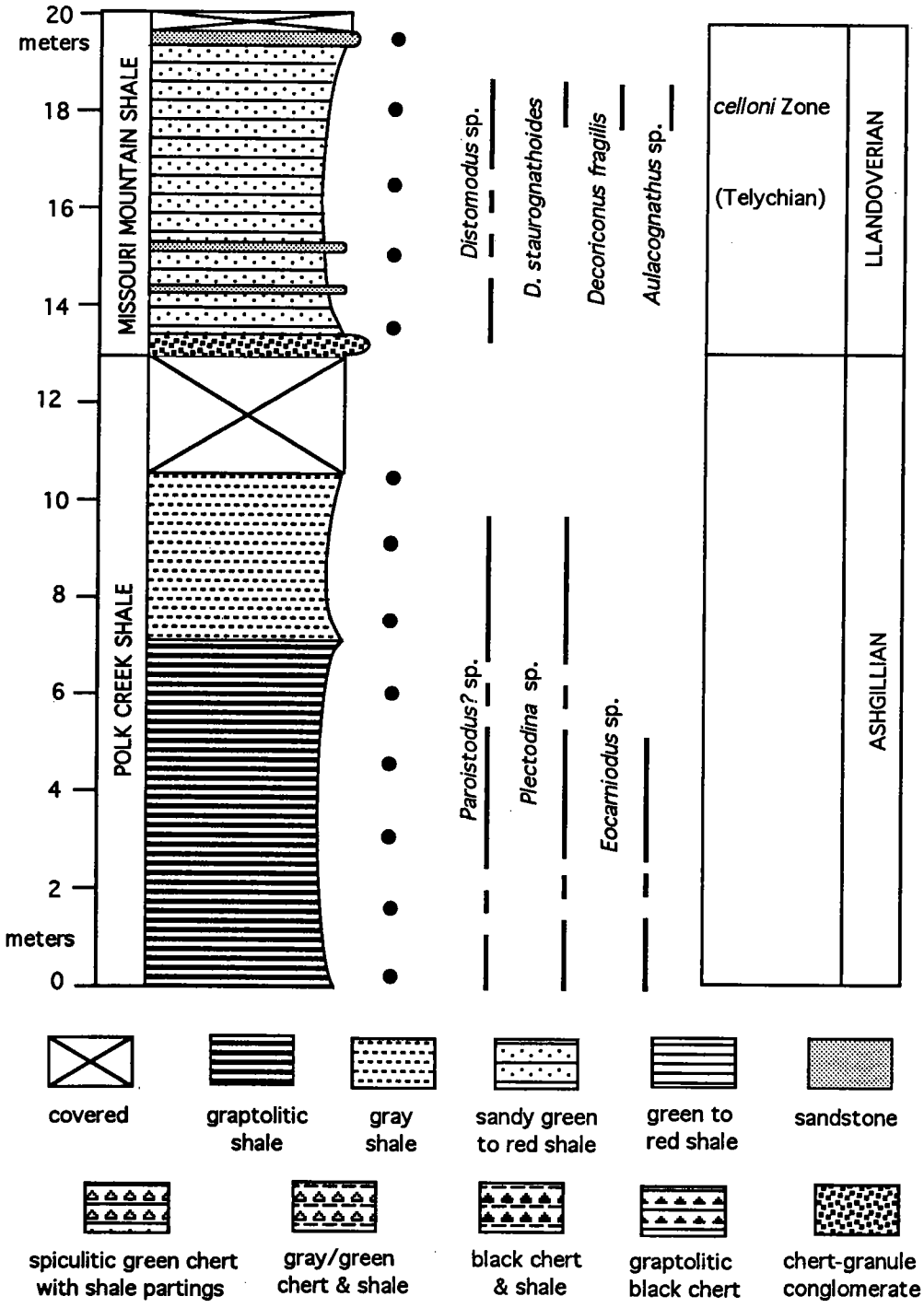
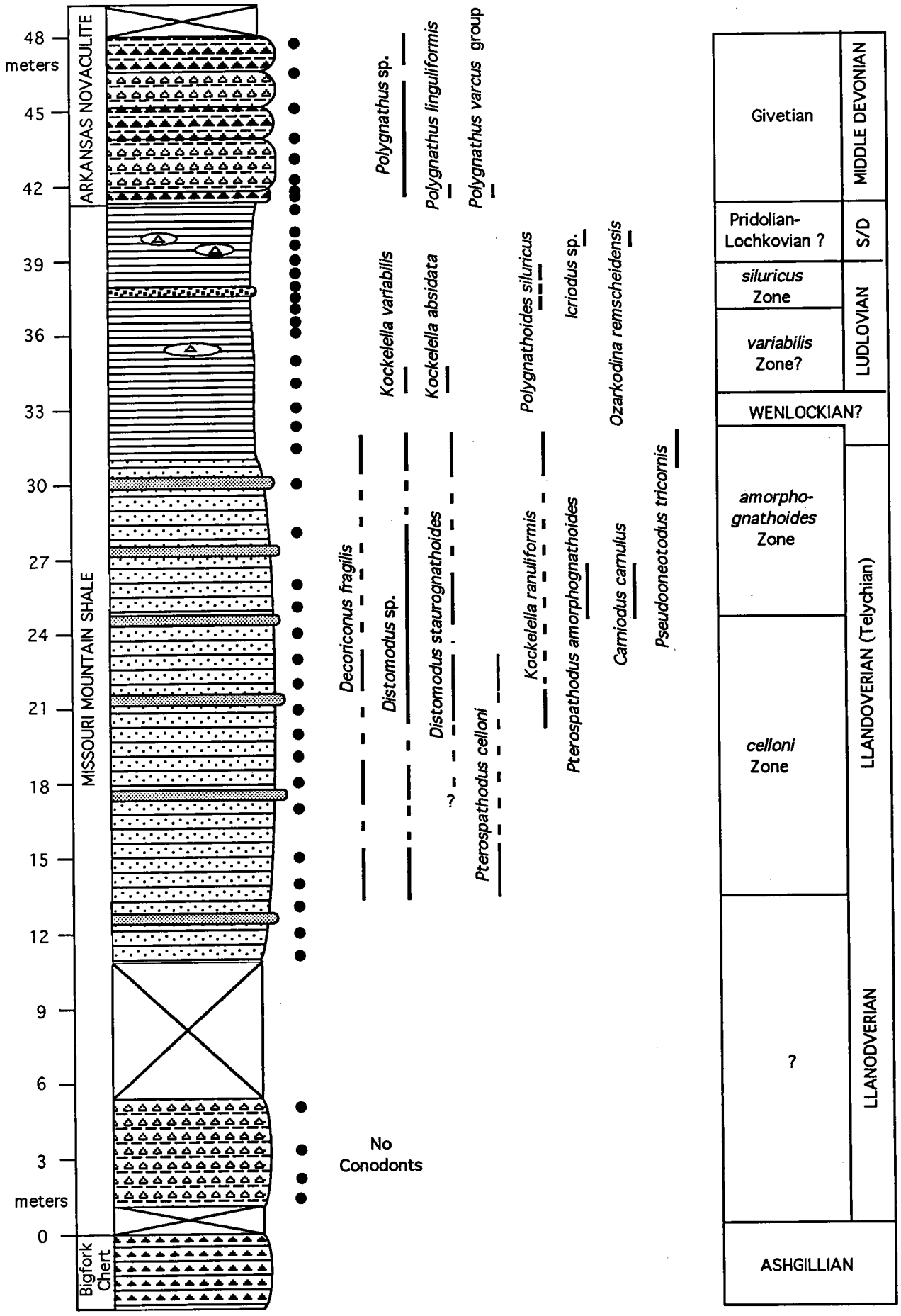


Figure 3. Stratigraphic column of the upper Polk Creek Shale and lower Missouri Mountain Shale at the Asphalt Pit section (NE¼NW¼ sec. 32, T. 1 S., R. 12 E.). Black dots along column represent conodont samples. Ranges of selected conodonts, conodont zones, and age interpretations are shown to the right.

runs along the bed and north side of a small stream. This is the locality from which the only identified fossils from the Missouri Mountain Shale were reported. The base of the formation and the contact with the Bigfork Chert were covered, but Hendricks and others (1937) reported that the lowermost 1 m consists of green siliceous shale, containing chert fragments. Above this shale, Hendricks and others (1937) described a 6.2-m interval of white to green cherty shale grading into chert, 4.5 m of which was exposed when we sampled. Above the cherty shale interval, greenish-gray sandy clay shales dominate the section, but reddish-gray intervals are common. The lower part of these shales was poorly exposed in the stream

Figure 4 *(opposite page)*. Stratigraphic column of the Missouri Mountain Shale and lower Arkansas Novaculite at the Grants Gap section (SE¼NW¼, sec. 10, T. 1 S., R. 12 E.). Explanation of lithologic symbols is given in Figure 3.


48
meters
45
42
39
36
33
30
27
24
21
18
15
12
9
6
3
meters
0
ARKANSAS NOVACULITE
MISSOURI MOUNTAIN SHALE
Bigfork Chert
Decoriconus fragilis
Distomodus sp.
Distomodus staurognathoides
?
Pterospathodus celloni
Kockelella variabilis
Kockelella absidata
Kockelella ranuliformis
Polygnathus sp.
Polygnathus linguliformis
Polygnathus varcus group
Polygnathoides siluricus
Icriodus sp.
Pterospathodus amorphognathoides
Ozarkodina remscheidensis
Camiodus camulus
Pseudooneotodus tricornis
No Conodonts
Givetian
MIDDLE DEVONIAN
Pridolian-Lochkovian ?
S/D
siluricus Zone
variabilis Zone?
LUDLOVIAN
WENLOCKIAN?
amorpho-gnathoides Zone
celloni Zone
?
LLANDOVERIAN (Telychian)
LLANODVERIAN
ASHGILLIAN

bed, but 11 m above the base, the shale was well exposed in the stream cut bank.

Isolated thin sandstone beds appear as low as 16 m above the base, but do not become common until above 22 m. Over the next 9 m, the section is dominated by brownish sandstone beds with interbedded greenish-gray sandy clay shale. The sandstone beds are as much as 10 cm thick and display grading suggestive of deposition by gravity flows. Hendricks and others (1937) reported these sandstones as being fossiliferous, but we did not observe any fossil material on the outcrop. Certain sandstones possess a black matrix, suggestive of a high asphalt content. The sandstone beds disappear at 31 m, and the next 6.5-m interval consists of reddish- and greenish-gray shale, bearing minor thin layers (<5 cm) of dark green to black chert. Near the top of this interval is a lenticular chert-granule conglomerate, 20 cm thick. This conglomerate appears to lie at about the same level as the 2-ft conglomerate described by Hendricks and others (1937), from which one fossil was obtained. The uppermost 3.5 m of the section (37.5–41 m) consists of green clay shale bearing 5- to 20-cm-thick lenticular chert beds.

Arkansas Novaculite

The best section of the Arkansas Novaculite in the Black Knob Ridge area occurs along Oklahoma Highway 3 east of Atoka and in shale pits north and south of the highway (East Atoka Road Cut, Fig. 5). Both Hendricks and others (1937) and Sholes (1978) have described this section. The contact between the top of the Missouri Mountain Shale and the base of the Arkansas Novaculite was exposed in the floor of the shale pit just south of the highway. The uppermost 2.5 m of the Missouri Mountain Shale consists of clayey, green to greenish-gray, somewhat flaky shale; below this level it was poorly exposed. Thinly bedded, green-gray chert of the lowermost unit of the Arkansas Novaculite rests on the green shale with a sharp contact. No interbedding of the chert and shale was observed.

Along the highway, a distinctive ridge-forming unit of green-gray chert and interbedded shale forms the lower 34 m of the formation (Unit 1). The basal 20 m consists of thinly bedded green-gray chert with partings of siliceous shale. The chert beds gradually thicken up-section, and the shale partings nearly disappear. An interval of green blocky shale with little chert forms a major break in the chert section from 20.0 to 21.1 m. Above the shale interval, the green-gray chert beds are thicker, possess irregular to knobby bedding surfaces, and are separated by partings of siliceous shale. Near the top of the unit, the beds change into a softer, lighter-colored chert. The uppermost bed is heavily stained yellow-brown and may represent an exposure surface. Sholes (1977,1978) described the cherts as spiculitic with a pelloidal texture. HF-insoluble residues obtained from these cherts during processing for conodonts consisted of masses of monaxon spicules and fewer tetraxon spicules.

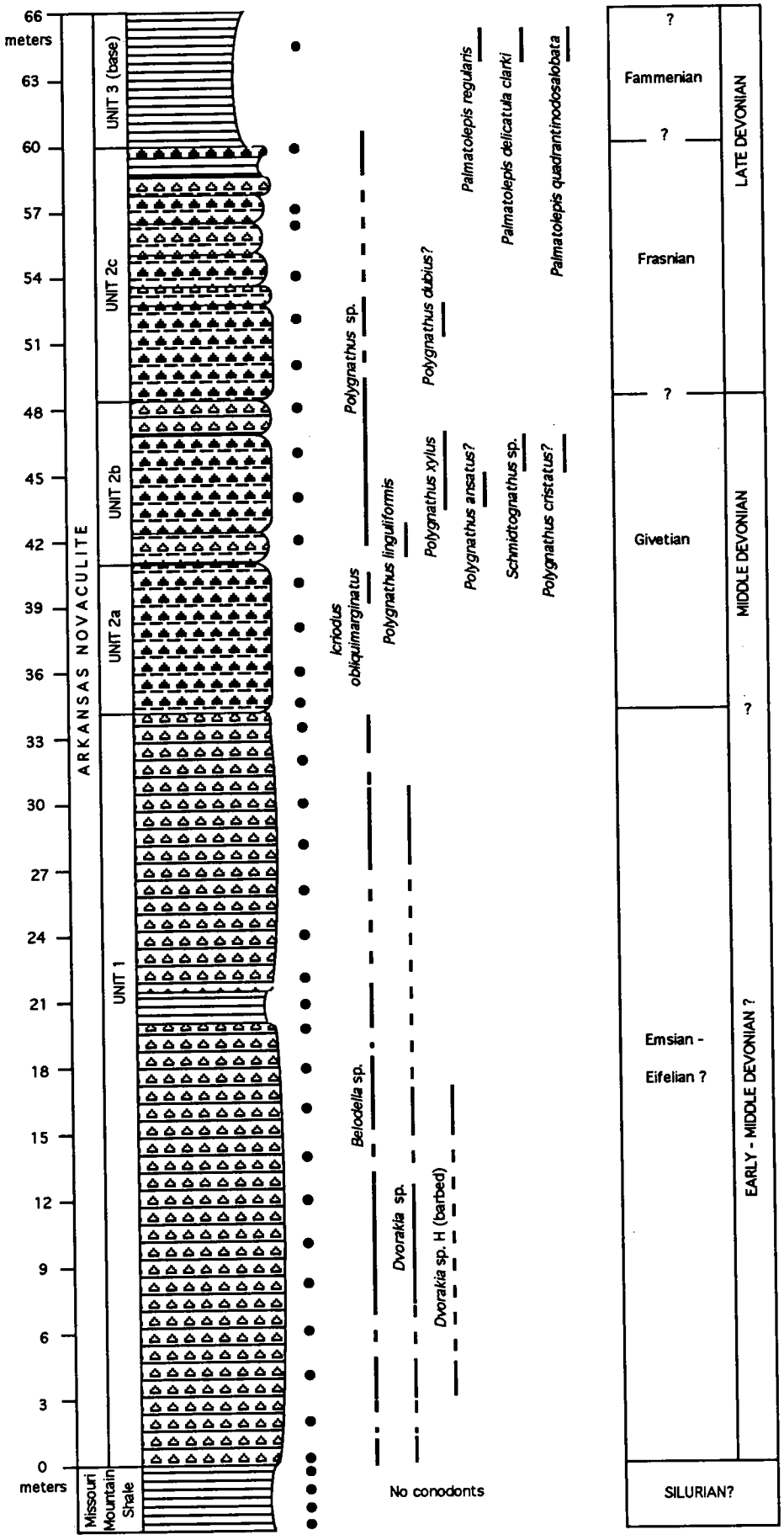
A sharp lithologic break separates the green-gray cherts of Unit 1 from the cherts and shales of Unit 2. This 26-m-thick unit is characterized by the dominance of black fissile shale and dark gray to black chert. Less common beds of greenish-gray chert and shale and rare thin sandstones and chert-granule conglomerates are present. Three gradational subunits can be recognized within Unit 2: 2a—a 7.5-m sequence of soft black chert interbedded with black shales; 2b—a 7.0-m sequence with subequal amounts of gray-green and black chert and interbedded shale; and 2c—an 11.5-m sequence dominated by black chert and fissile black shale in the lower part, the top of which grades into Unit 3. The cherts and shales of Unit 2 contain radiolarians, palynomorphs, including tasmanitids, and conodonts as rare molds in bedding surfaces of black shales. Sholes (1977,1978) described the cherts in this unit as nonspiculitic. He emphasized the presence of clastic laminae in many chert beds, which he interpreted to represent basal units of fine-grained turbidites.

The base of Unit 3 is placed where black chert and shale disappear from the section. Through the top 8 m of Unit 2c, green-gray shale appears in increasing amounts, but black chert and black fissile shale persist as lithologic components. About 30 m of green-gray and red siliceous shale with thin lenticular green to brown chert of Unit 3 were partially exposed in shale pits north of the highway. Only the lower 6 m were well exposed and sampled for this study. Hendricks and others (1937) reported that 9 m of brown novaculite occurs at the top of the section. At the time of our work, only a low ridge of 3 m of brown chert was poorly exposed in the same stratigraphic position. The contact between the Arkansas Novaculite and overlying Stanley Shale was covered. Sholes (1978) incorrectly placed the base of the Stanley Shale at the base of our Unit 3, making his Black Knob Ridge section (Sholes and McBride, 1975, fig. 3; Sholes, 1977, fig. 4) at least 30–40 m too thin.

At the Grants Gap section, a different lithologic sequence occurs at the base of the Arkansas Novaculite (Fig 4). Sholes (1978) included 14 m of shale—an interval that we place in the upper part of the Missouri Mountain Shale—in his lower chert and shale member of the Arkansas Novaculite, because of the presence of a few chert beds in this interval. We, however, place the base of the Arkansas Novaculite where the shales of the upper Missouri Mountain Shale are abruptly overlain by a chert-dominated interval, as did Hendricks and others (1937). The base of the Arkansas Novaculite is a thin (10 cm) black chert that rests with a sharp contact on green shale of the Missouri Mountain Shale; there is no interbedding of the two lithologies. The basal chert is followed by 20 cm of black fissile shale with lenticular black chert, and a 15-cm-thick bed of chert breccia lies above it. Only about 7 m more of section is well exposed, which consists of major intervals (1–2 m) of green to brown chert with siliceous shale partings separated by smaller units (≤0.5 m) of black chert and black to dark brown fissile shale. A few radiolarians occur in the black chert beds.

The basal 7 m of the Arkansas Novaculite at Grants Gap does not resemble the spiculitic green-gray cherts of

Figure 5 *(opposite page)*. Stratigraphic column of the upper Missouri Mountain Shale and Arkansas Novaculite at the East Atoka Road Cut (in Scratch Hill along Oklahoma Highway 7; SE corner, sec. 14, T. 2 S., R. 11 E.). Explanation of lithologic symbols is given in Figure 3.


66
meters
63
60
57
54
51
48
45
42
39
36
33
30
27
24
21
18
15
12
9
6
3
0
meters
ARKANSAS NOVACULITE
UNIT 3 (base)
UNIT 2c
UNIT 2b
UNIT 2a
UNIT 1
Missouri Mountain Shale
Icriodus obliquimarginatus
Polygnathus sp.
Polygnathus linguliformis
Polygnathus xylus
Polygnathus dubius?
Polygnathus ansatus?
Schmidtognathus sp.
Polygnathus cristatus?
Palmatolepis regularis
Palmatolepis delicatula clarki
Palmatolepis quadrantinodosalobata
Belodella sp.
Dvorakia sp.
Dvorakia sp. H (barbed)
No conodonts
?
Fammenian
?
Frasnian
?
Givetian
?
Emsian -
Eifelian ?
LATE DEVONIAN
MIDDLE DEVONIAN
EARLY - MIDDLE DEVONIAN ?
SILURIAN?

Unit 1 at the East Atoka Road Cut, but instead is similar to the cherts and shales of Unit 2, especially subunit 2b. For this reason, we assign the base of the Arkansas Novaculite at Grants Gap to our Unit 2 and infer that the base of the section is unconformable, with strata equivalent to our Unit 1 at the East Atoka Road Cut missing.

CONODONT FAUNAS

Conodonts were recovered from the upper Polk Creek Shale, Missouri Mountain Shale, and Arkansas Novaculite at the sections described above. The clay shales of the Polk Creek and Missouri Mountain disaggregated readily in white kerosene and water. Conodonts from siliceous black, green, and red shales of the Arkansas Novaculite were preserved as molds.

Cherts of the Arkansas Novaculite were dissolved by using two different techniques. Samples processed early in this research were treated with dilute HF according to the method outlined by Barrick (1987). This process yielded rather large insoluble residues consisting of masses of spicules from the spiculitic cherts and organic aggregates from black cherts. Later, on the recommendation of Glen Merrill (University of Houston–Downtown), we used a modified version of the process of Magné and Dufaure (1964), described in Stone (1987). In a large (2 L) plastic beaker, 200 g of crushed sample was placed, and a mixture of 0.4 L of 35% H_2O_2 and 0.6 L of 52% HF was poured on. The dissolution process is rapid, less than 1 h, but strongly exothermic. The caution of Merrill (in Stone, 1987) to place the beakers in a large plastic tray should be heeded, especially in the case of black, organic-rich cherts, where the solution often boiled over the top of the beaker. The quantity of chert dissolved by one treatment is comparable to that obtained by four successive days of treatment in dilute HF (40–60% of the sample). The residues obtained with this method were smaller and cleaner, and the larger conodont elements tended to be better preserved than those obtained by slow dissolution in dilute HF. However, the process did appear to destroy siliceous microfossils, such as radiolarians, which are often recovered by the other method.

Polk Creek Shale

Although no conodonts had been reported previously from the Polk Creek Shale, because of the good exposures of the brown and gray clay shale at two localities, samples were collected and processed. No conodont elements were recovered from seven 2-kg samples in the shale pit south of the East Atoka Road Cut. At the Asphalt Pit locality (Fig. 3), however, six of eight 2-kg samples yielded conodont elements. The conodont elements are small in size, even for Ordovician taxa, and the greatest number recovered per kilogram was 100. The most common elements are oistodiform and distacodiform elements that belong to a species of either *Besselodus* or *Paroistodus?* (see Nowlan and others, 1988). *Plectodina* elements occur in each sample, and a few elements of *Panderodus* and *Eocarniodus* are present.

Although the Polk Creek conodonts need to be more thoroughly studied, the fauna clearly includes Late Ordovician forms. *Eocarniodus* ranges through the Ashgillian Series in Great Britain, including the uppermost stage, the Hirnantian (Orchard, 1980). The Polk Creek fauna also bears some similarity to conodonts from the Keel Limestone of Oklahoma and the upper oolite of the Cason Shale of Arkansas, which are considered to be Hirnantian in age on the basis of brachiopods (Amsden and Barrick, 1986). However, no conodonts diagnostic of the Hirnantian were obtained from the Polk Creek.

Missouri Mountain Shale

No conodonts were recovered from the lower 3 m of green clay shale of the Missouri Mountain Shale at the shale pit south of the East Atoka Road Cut, nor from the green cherty shale and chert exposed in the lower 6 m of the formation at the Grants Gap section. At the Asphalt Pit section (Fig. 3), a small conodont fauna consisting of broken elements was obtained from the sandy shales above the chert-granule conglomerate marking the base of the unit. Elements of the coniform taxon *Dapsilodus praecipuus* dominate the fauna, and *Dapsilodus sparsus, Distomodus staurognathoides, Decoriconus fragilis, Pseudooneotodus tricornis, Panderodus unicostatus,* and indeterminate species of *Aulacognathus, Pterospathodus,* and *Oulodus* are present. This fauna is late Llandoverian in age and appears to be no older than the Telychian *celloni* Zone, because of the presence of *Aulacognathus* and *Decoriconus fragilis.* A few elements of Ordovician taxa like those found in the Polk Creek Shale were recovered from one sample, suggesting some reworking of material from older strata.

A similar conodont fauna appears at the Grants Gap section (Fig. 4) in the sandstone-bearing shales extending from 14 to 24 m above the base. As at the Asphalt Pit, most elements are badly broken, and the fauna is strongly dominated (80%) by elements of *Dapsilodus praecipuus.* In addition to the species occurring at the Asphalt Pit, *Pterospathodus celloni, P. pennatus,* and *Kockelella ranuliformis* are present, confirming the *celloni* Zone age of the fauna. A few reworked Ordovician elements also occur. The abundance of conodont elements is related to the sand content of the interval sampled. Intervals with relatively few sand grains in washed residues yielded few conodont elements, typically specimens of *D. praecipuus,* whereas those with abundant sand produced large faunas including elements of pectiniform- and ramiform-bearing taxa. If the coarser clastic material was being introduced into the depositional site by gravity flows, as the nature of the sandstone beds suggest, then the majority of conodonts were also.

Pterospathodus amorphognathoides and *Carniodus carnulus* appear, and *P. celloni* disappears, 25 m above the base, still within the interval of sandstone and shale. Except for these changes, the fauna is identical with that of the underlying beds, including the presence of reworked Ordovician elements. The appearance of *P. amorphognathoides* marks the base of the *amorphognathoides* Zone, which spans from the latest Llandoverian into the base of the Wenlockian. The top of the *amorphognathoides* Zone at the Grants Gap section is difficult to determine because the diagnostic species is present only near the base. Samples extending up to 31 m above the base, just above the top of the sandstone-bearing interval, include elements of *Distomodus staurognathoides* and *Pseudooneotodus tricornis* (species that

are not known to range significantly above the *amorphognathoides* Zone) and thus are included in the zone.

Many fewer conodont elements were recovered from the shales lying above the sandstone-bearing interval, from 32 to 38 m above the base of the Missouri Mountain Shale (up to 3 m below the top of the formation). Elements of *Dapsilodus obliquicostatus, Panderodus unicostatus, Decoriconus fragilis, Ozarkodina excavata?*, and *Kockelella* or *Oulodus* ramiforms occur in small numbers. At 34 m, a small fauna characterized by the presence of the *Kockelella* species, *K. variabilis, K. absidata,* and *K. stauros?* was obtained. This association of species occurs in lower to middle Ludlovian strata (*variabilis* Zone?). At 36.7 m, *Polygnathoides siluricus* occurs, and at 38 m, *P. emarginatus* (probably the Pb element of *P. siluricus*) is present. These forms are indicative of the middle Ludlovian *siluricus* Zone.

Only a few conodonts could be extracted from the top 3 m of the Missouri Mountain Shale at Grants Gap. *Ozarkodina remscheidensis eosteinhornensis* and some fragments of an *Icriodus* species were obtained from the shale 1.5 m below the top. This association suggests an age near the Silurian-Devonian boundary.

No conodonts were recovered from the green shale in the upper 3 m of the Missouri Mountain Shale in the shale pit south of the East Atoka Road Cut.

Arkansas Novaculite

Few conodonts were obtained from Unit 1, the lower green chert unit, of the Arkansas Novaculite at the East Atoka Road Cut section (Fig. 5). Chert samples yielded small delicate coniform elements belonging to either of the long-ranging genera, *Belodella* or *Dvorakia.* A few poorly preserved and undiagnostic pectiniform and ramiform elements were obtained, as were several coniforms of the type that occur in the *Icriodus* apparatus. Disaggregation of the siliceous shale occurring in the middle of the unit contributed indeterminate fragments of *Icriodus* Pa elements.

The presence of the *Icriodus* fragments restricts the age of the lower green chert and shale unit to the Devonian. The few fragments of spindles obtained possess three well-developed denticle rows, excluding the oldest species of *Icriodus, I. woschmidti.* The only form that permits a more narrow age range to be defined is an unusual *Dvorakia* element, *Dvorakia* sp. H. In two samples below the siliceous shale interval, there are elements of *Dvorakia* with a subtriangular cross section; on the inner lateral margin of these elements are a few robust, barb-like denticles that point apically. This element lacks a name, but has been reported and illustrated from several faunas in the Midcontinent of North America, most often in association with faunas bearing *Icriodus latericrescens robustus* (Illinois Basin—Orr, 1964; northern Kentucky—Klug, 1983; Ontario—Uyeno and others, 1982), which ranges from the late Emsian into the Eifelian (Uyeno and others, 1982; Klug, 1983). This type of *Dvorakia* element is not known from older Devonian units, but similar elements range higher, into Givetian beds (Middle *varcus* Subzone) in northern Iowa and southern Minnesota (Klapper and Barrick, 1983).

The dark cherts of Unit 2 of the Arkansas Novaculite at the East Atoka Road Cut section (Fig. 5) yielded radiolarians in sparse numbers and variable numbers of conodont elements. No conodonts were obtained from the base of the unit, but at the top of the lower subunit of black chert and shale (2a), a single specimen of *Icriodus obliquimarginatus* was obtained. The middle subunit of gray-green and black chert and interbedded shale (2b) produced a few identifiable conodonts from three successive chert samples. The lowest sample at 42 m is characterized by *Polygnathus linguliformis*, gamma morphotype, and fragments of other species. At 44 m, the fauna is dominated by members of the *Polygnathus varcus* group, including *P. xylus* and forms resembling *P. ansatus. Polygnathus xylus* is present at 46 m, where fragmentary Pa elements of a *Schmidtognathus* species (*S. hermanni?*) and a member of the *Polygnathus cristatus* group occur. Another chert sample near the top of subunit 2b (48 m) contained only indeterminate *Polygnathus* elements. From the upper subunit of black chert and fissile black shale (2c), few conodonts were obtained. Fragmentary Pa elements that belong to a species of the *Polygnathus dubius* group occur in a black shale at 52 m, 3.5 m above the base of the subunit. Otherwise, only a few indeterminate *Polygnathus* fragments are present up through the top sample at 60 m.

Icriodus obliquimarginatus ranges from the top of the Eifelian into the Givetian (Middle *varcus* Subzone) (Uyeno and others, 1982), which places the top of subunit 2a firmly in the Middle Devonian. The association of species in the lower two samples from subunit 2b suggests a *varcus* Zone assignment for the fauna. The presence of *Polygnathus ansatus*–like morphotypes in the fauna at 44 m may restrict the age of this sample to the Middle or Upper *varcus* Subzones, but better-preserved specimens are needed to confirm this assignment. The presence of a *Polygnathus cristatus* group species and *Schmidtognathus* in the sample at 46 m permits an assignment to the *hermanni* Zone of late Givetian age. *Polygnathus* elements resembling *P. dubius* have a long range, from the late Givetian through the Frasnian. Subunit 2c appears to be Frasnian in age.

In Unit 3, conodonts are preserved as molds in the green and red siliceous shale at several levels. A good fauna 5 m above the base of the unit includes the following species: *Palmatolepis delicatula clarki, Pa. quadrantinodosalobata,* and *Pa. regularis.* The first two of these species occur together in the Lower and Middle *crepida* Zones (Klapper and Ziegler, 1979), which indicates an early Fammenian age for the collection. Additional study of material from Unit 3 should yield a more complete succession of conodonts. Samples from the "upper novaculite" in the shale pit north of the road cut failed to provide any conodont elements.

Just as the lithology of the base of the Arkansas Novaculite at the Grants Gap section differs from that of the base of the formation at the East Atoka Road Cut, so do the conodont faunas. Conodont elements are relatively common in cherts at the base of the formation at Grants Gap (Fig. 4), but diagnostic elements are rare. *Polygnathus linguliformis,* broken specimens of the *Polygnathus varcus* group, and *Icriodus* fragments occur in the black shale 10–20 cm above the base. Higher samples yielded juvenile and fragmented *Polygnathus* elements, many of which could be representatives of *P. linguli-*

formis and the *P. varcus* group. This fauna appears to be identical with that recovered from the middle subunit of Unit 2 at the road-cut section and is probably also Givetian in age.

INTERPRETATION AND CORRELATION

The conodont faunas obtained from the Missouri Mountain Shale and Arkansas Novaculite at Black Knob Ridge suggest a complex local stratigraphy that includes several unconformities between and perhaps within these formations. The faunas are sufficiently good to provide some constraints on correlations to adjacent shelf sequences in the Arbuckle Mountains of southern Oklahoma and sections in the Ozark region of northern Arkansas (Fig. 6).

Missouri Mountain Shale

The oldest Silurian conodont fauna obtained from the Missouri Mountain Shale is the *celloni* Zone fauna that occurs at the base of the formation at the Asphalt Pit section and in the middle of the unit at Grants Gap. Because the *celloni* Zone is late Llandoverian (Telychian) in age, a hiatus encompassing most of the Llandoverian separates the Ordovician units from the Missouri Mountain. The absence of the Polk Creek Shale at Grants Gap and the basal sandstone of the Missouri Montana at the Asphalt Pit section suggest that an erosional unconformity is present. This stratigraphic relation is similar to that observed on the shelf regions. In Oklahoma, the Cochrane bears a *celloni* Zone conodont fauna near the top, and it rests unconformably on the Hirnantian (Upper Ordovician) Keel Formation (Barrick and others, 1990). In a similar manner, the Cason limestone ("Brassfield") of Craig (1993), which also contains a *celloni* Zone conodont fauna, has been observed resting directly on the Hirnantian Cason oolite (Craig, 1993).

The *celloni* to *amorphognathoides* Zone interval observed within the sandy interval of the Missouri Mountain Shale, however, does not obviously correspond to the shelf sequences. In Oklahoma, the Prices Falls Shale Member of the Clarita Formation rests unconformably on the Cochrane (Amsden, 1960). The Prices Falls is a basal lag deposit that incorporates glauconite and conodonts reworked from the underlying Cochrane. The Prices Falls, like the base of the overlying Fitzhugh Member at some sections, contains an *amorphognathoides* Zone fauna (Barrick and Klapper, 1976). In northern Arkansas, the "button shale" of the upper Cason Shale overlies the Cason limestone unconformably and represents the basal unit of a transgres-

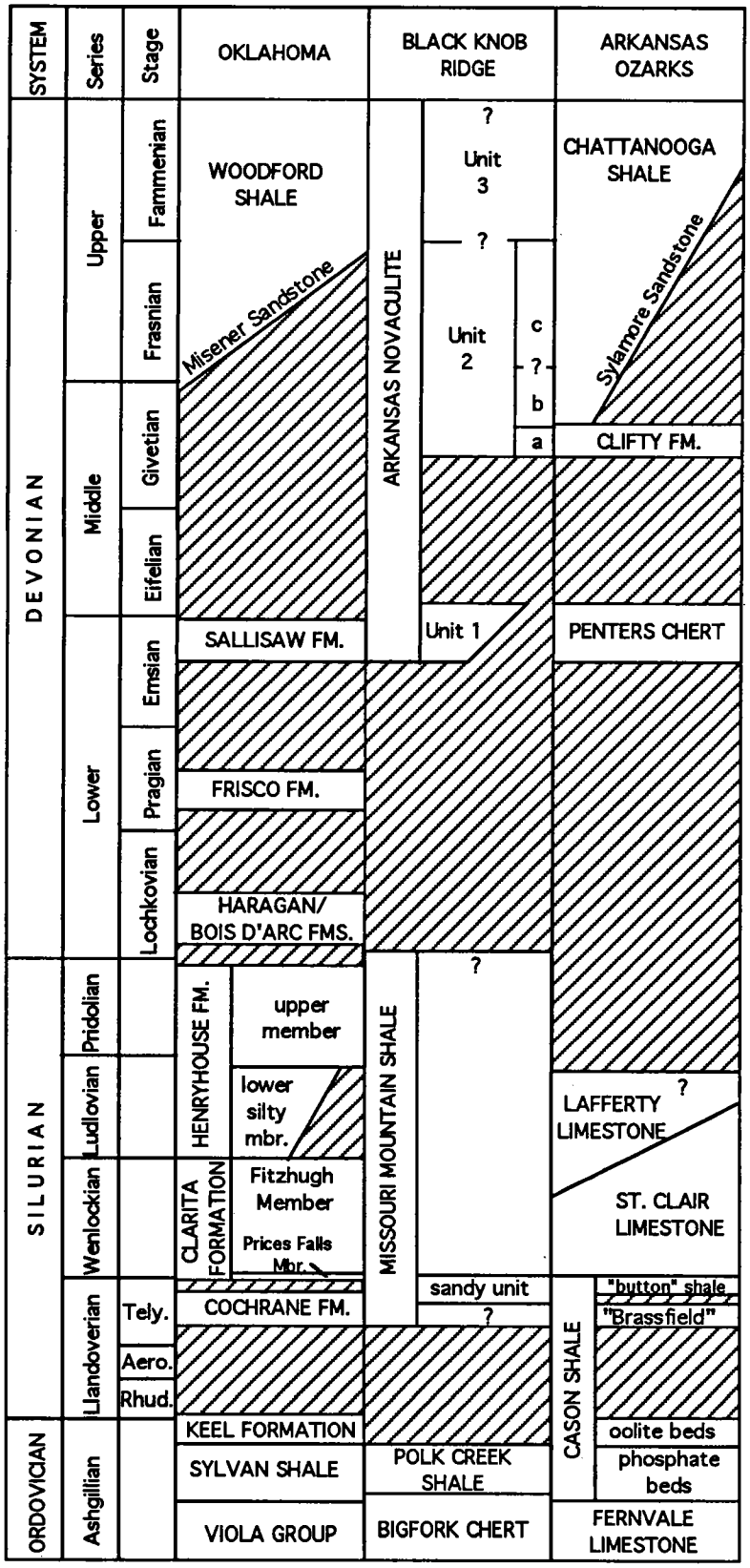


Figure 6. Interpreted ages of the Missouri Mountain Shale and units of the Arkansas Novaculite at Black Knob Ridge. Inferred correlations to the Oklahoma shelf succession and the Arkansas shelf succession in the Ozarks are shown. Rhud. = Rhuddinian; Aero. = Aeronian; Tely. = Telychian. See text for further discussion.

sive interval succeeding an interval of subaerial erosion (Craig, 1993). The thickness of the "button shale" varies considerably, and locally the unit is absent. It grades upward into the limestones at the base of the St. Clair Limestone, which bear an *amorphognathoides* Zone conodont fauna (Craig, 1969,1975).

There is no physical or biostratigraphic evidence for a comparable unconformity within the Missouri Mountain Shale at the Grants Gap section. Instead, we suggest that the sand-bearing interval of the formation may correspond to the late Llandoverian unconformity in the shelf sections. During the time that the Cason limestone and Cochrane were exposed to subaerial erosion, clastic debris and conodont elements may have been carried offshore and deposited by gravity flows into the deeper-water setting inferred for the Missouri Mountain Shale. Although there is no source of sand-sized material in the Oklahoma area, such detritus would have been available in the Arkansas region from the lower shale unit of the Cason Shale, which also provided clastic material for the "button shale" (Craig, 1993). Erosion of this Upper Ordovician unit would have also contributed the reworked Ordovician conodonts that appear sparingly in the Missouri Mountain sandstone beds. In this interpretation, the upper Llandoverian carbonates of the shelf would have no lithic or time equivalent in the Missouri Mountain Shale, unless the siliceous shale and chert at the base of the formation at Grants Gap, which remains undated, is the deeper-water equivalent.

If Wenlockian strata are present in the Missouri Mountain Shale at Grants Gap, they must comprise the 3-m interval below the appearance of *Kockelella variabilis*. If so, this thin interval would correspond to the period of carbonate accumulation that is represented by the St. Clair Limestone and Lafferty Limestone in Arkansas and the Fitzhugh Member of the Clarita Formation in Oklahoma. The occurrence of Ludlovian conodonts coincides with the presence of at least one conglomeratic bed in the shale section. This suggests that, as was the case in the lower Missouri Mountain Shale, an influx of debris bearing conodonts may have been introduced into the depositional site during an interval of erosion on the shelf. Although the St. Clair and Lafferty package ranges at least as high as the Ludlovian (*siluricus* Zone) without interruption (Craig, 1993), throughout the Arbuckle Mountains, lower Ludlovian strata (*variabilis* through *siluricus* Zones) have been removed at most sections (Barrick and others, 1990). Even where lower Ludlovian strata are preserved in the central Arbuckles, the beds are highly argillaceous, silty carbonates in which conodonts have been concentrated and fragmented by sedimentary processes.

The occurrence of a conodont fauna near the top of the Missouri Mountain Shale that indicates an age near the Silurian/Devonian boundary may be accommodated by similar reasoning. Both physical and biostratigraphic evidence exists for an unconformity at the Henryhouse/Haragan boundary in the Arbuckle Mountains (Amsden, 1988; Barrick and Klapper, 1992). The conodont fauna in the Missouri Mountain Shale could have been derived during the erosional event that formed the shelf unconformity at the Silurian/Devonian boundary. If deposition was extremely slow and more or less continuous in the Missouri Mountain Shale, then the equivalents of the upper Ludlovian to Pridolian strata of the upper Henryhouse Formation would consist of <3 m of green shale and chert. An alternative interpretation would have a major unconformity lying above the lower Ludlovian fauna in the shale, a situation that would correspond with the absence of upper Ludlovian and Pridolian units in the Arkansas region (Craig, 1993).

Arkansas Novaculite

The sharp contact at the base of the Arkansas Novaculite and the late Early Devonian maximum age for the basal green chert unit (Unit 1) indicate a major break in deposition between it and the underlying Missouri Mountain Shale. The shelf unit of comparable age is the late Emsian Sallisaw Limestone (Klapper, in Amsden, 1985; Amsden, 1988). The Sallisaw is the most western formation in a widespread series of upper Emsian to Eifelian cherty carbonates that early workers correlated with the lower novaculite division of the Arkansas Novaculite (Penters Chert of Arkansas, Clear Creek Chert of the Illinois Basin, Camden Chert of Tennessee; see Amsden, 1988, for a summary). We assume that the lower green chert unit at Black Knob Ridge is a lithic and time equivalent of the lower novaculite member, but lack any biostratigraphic information to confirm it. The shelf units compose the lower part of the upper Emsian–lower Eifelian transgressive-regressive (T-R) cycle Ic (Johnson and Klapper, 1992), and the basal part of the Arkansas Novaculite also appears to be part of the same T-R cycle.

The Pinetop Chert of the frontal Ouachitas does not correlate with the basal unit of the Arkansas Novaculite at Black Knob Ridge because it is early Lochkovian in age (Amsden, 1983). However, strata equivalent to the base of the Arkansas Novaculite probably are present at the Pinetop section. Hendricks and others (1947) reported just over 2 m of white chert breccia with some lenses of fine-grained crystalline limestone lying just above the Pinetop Chert, which they included as a basal unit of the Woodford Shale. Preliminary work on the conodonts from this white chert suggests that it may be a Sallisaw equivalent and could be the same age as the base of the Arkansas Novaculite.

The physical and biostratigraphic evidence at Black Knob Ridge entails the conclusion that a major unconformity also exists between Units 1 and 2 of the Arkansas Novaculite. Although the biostratigraphic evidence is meager, we conclude that the base of our Unit 2 is no older than Givetian. The conodont fauna of subunit 2b spans the Middle *varcus* Subzone–*hermanni* Zone interval, making it a correlative of the shelf T-R Cycle IIa, the basal part of the Taghanic onlap (Johnson, 1970). Both the Misener Sandstone in the Oklahoma subsurface (Amsden and Klapper, 1972) and the Sylamore Sandstone in northeastern Oklahoma and northern Arkansas (Freeman and Shumacher, 1969) are products of the Taghanic onlap (Johnson and Klapper, 1992). In Arkansas, the Clifty Formation, a calcareous sandstone, underlies the Sylamore Sandstone. Although Hall and Manger (1978) reported a conodont fauna containing fragments similar to *Polygnathus cristatus* (*hermanni* Zone), Amsden (1988) reported that the brachiopod fauna of the Clifty was somewhat older than the *varcus* Zone. The

Clifty Formation could then be considered to be correlative to the basal subunit 2a of our Unit 2. The absence of Unit 1 at Grants Gap can be best explained by its removal during a mid-Devonian episode of erosion.

The lack of diagnostic conodonts in the upper subunit 2c of Unit 2 and the base of Unit 3 prevents precise determination of the Middle/Upper Devonian and Frasnian/Fammenian boundaries in the Arkansas Novaculite. By stratigraphic position, and assuming no major unconformities, the base of the Upper Devonian should lie at the base of or within the upper subunit 2c. The conodont fauna 5 m above the base of Unit 3 belongs to the second conodont zone above the base of the Fammenian, which should lie a short interval below our sample in Unit 3. The age of the base of the Woodford Shale in Oklahoma is not well documented and may vary with the position of sections on the shelf. Over (1990) has studied the basal beds of the Woodford in the Arbuckle Mountains and determined that the oldest black shale appeared during the late Frasnian (Montagne Noire Frasnian Zone 9; Johnson and Klapper, 1992). It is likely then, that the appearance of the black-shale facies on the shelf (Woodford/Chattanooga), interpreted to have formed under anoxic conditions (e.g., Heckel and Witzke, 1979), coincides with the disappearance of black chert and shale and the appearance of green and red siliceous shales at Black Knob Ridge.

IMPLICATIONS FOR OUACHITA STRATIGRAPHY

Major and probably minor unconformities exist between many of the lithologic units at Black Knob Ridge. Even across the relatively short outcrop belt, <15 km, the Polk Creek Shale and the lower part of the Arkansas Novaculite disappear, probably cut out by erosion. Barring the possible juxtaposition of unrelated sections during orogenic deformation, this finding suggests that sufficient relief existed in the depositional site to control the deposition and preservation of major stratigraphic units. Although many sedimentological features indicate deep-water deposition (Lowe, 1989), the presence of multiple unconformities appearing over short geographic distances is more like that found in shelf settings than in basinal areas.

The strong correlation of stratigraphic events in the Black Knob sequence with shelf unconformities and transgressive-regressive cycles is surprising for an inferred basinal setting. Finney (1986) determined that a marked subsidence event at the base of the Viola Group in the Southern Oklahoma basin correlated with an eustatic drop in sea level at the base of the Bigfork Chert in the Ouachita facies. We recognize the basal Silurian unconformity of the shelf region occurring at the base of the Missouri Mountain Shale. Silurian influxes of clastic detritus appear to correspond to erosional events on the shelf. The major subdivisions of the Arkansas Novaculite correlate directly with Devonian transgressive-regression cycles that can be recognized in North American cratonic and shelf sequences. The reasons for the similarity in depositional history between the shelf region and the Black Knob Ridge succession are not clear. Finney (1986) noted that the Bigfork-Viola event coincided with the emplacement of the Taconic allochthon and subsidence of the Taconic foreland in the Appalachian region, implying the possibility of tectonic control. On the other hand, the Devonian transgression-regression events that correspond to the units of the Arkansas Novaculite have been interpreted as eustatic sea-level changes (Johnson and Klapper, 1992). Further detailed biostratigraphic work in conjunction with facies analysis in the Ouachita facies will be required to resolve the relative influence of eustatic events and tectonic events in preorogenic strata.

Because the strata at Black Knob Ridge are considered to represent a transitional setting between the shelf succession and the more typical Ouachita facies, the location of the shelf area that was proximal is important. We are struck by the strong similarity of the stratigraphic succession at Black Knob Ridge to the Arkansas Ozark shelf sections. The pattern of lithologic units and the spacing of unconformities in the Black Knob Ridge area is most like that found in the Ozarks (Fig. 6), and similarities exist with subsurface sections from the Arkoma Basin as far east as the Black Warrior Basin (Thomas, 1989). Comparison between the Arbuckle facies of south-central Oklahoma and the Ouachita facies at Black Knob Ridge is less obvious, and apparent similarities may represent only a response to common eustatic or tectonic events, not geographic proximity.

Although the upper preorogenic strata at Black Knob Ridge do not represent the most typical Ouachita succession, our results should have some bearing on how the stratigraphy of the all parts of the preorogenic section should be interpreted. We propose that the erosional unconformity between the lower novaculite and the middle chert and shale member identified in the Arkansas Novaculite at Black Knob Ridge occurs in most sections. Lowe (1977,1989) has suggested an unconformity at this position for the more southern sections. The disappearance of the lower novaculite in south to north transects (Caddo Gap to Washita, Arkansas, Lowe, 1989, fig. 3) and southwest to northeast transects (Polk County, Arkansas, Sellars, 1966) is probably not due to facies changes, but to erosion of the lower novaculite before deposition of the significantly younger middle chert and shale member, just as observed at Black Knob Ridge. We cannot exclude this possibility on the basis of the general argument that most of the Ouachita facies was deposited in a deep-water setting, making major unconformities unlikely. However, this question and possibility of other widespread unconformities in the Ouachita facies cannot be settled until sufficient biostratigraphic information is obtained from all lithologic units throughout the outcrop area.

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Friable Submarine Channel Sandstones in the Jackfork Group, Lynn Mountain Syncline, Pushmataha and Le Flore Counties, Oklahoma

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ABSTRACT.—More than 6,500 ft of Jackfork Group strata are exposed along U.S. Highway 259 south of Big Cedar, Oklahoma, on the north limb of the Lynn Mountain syncline. These strata exhibit many features typical of turbidites, including Bouma cycles, sole markings, slump and debris-flow deposits, and trace fossils assigned to the *Nereites* facies. Paleocurrent measurements for much of the Jackfork indicate an eastern source. Sandstones are typically fine-grained orthoquartzites and "ring" when struck with a hammer. The U.S. 259 Jackfork sedimentary sequence has been interpreted as representing deposition in the lower part of a submarine-fan lobe. Less well-documented is the presence of friable, medium-grained sandstones containing quartz pebbles and fossils. These sandstones are well exposed in the upper Jackfork Group along U.S. 259 and in the western part of the Lynn Mountain syncline where they exceed 800 ft in thickness. Sedimentary features include both tabular-planar and troughlike cross-bedding, scoured bedding contacts, and southward-directed paleocurrents. Sole markings and Bouma cycles are absent. The sandstones are quartz arenites to subarkoses and are poorly cemented. The depositional environment envisioned for the friable sandstones is a series of coalescing middle-fan submarine channels derived from channels incising a normal-marine shelf environment located to the north.

INTRODUCTION

In 1985, SOHIO drilled the No. 1-7 Weyerhaeuser well on the north flank of the Lynn Mountain syncline (sec. 7, T. 1 N., R. 23 E.). This well was drilled to test a stratigraphic trap consisting of a submarine channel in the Johns Valley Shale. The channel was imaged on high-resolution seismic data, and the presence of channel sandstones in the Johns Valley Shale was confirmed in outcrop north of the well location. The well encountered channel sandstones in the Johns Valley interval as predicted; however, despite numerous small gas shows in the Johns Valley and Jackfork intervals, the well was determined to be noncommercial and was plugged.

Field studies conducted in the Lynn Mountain syncline area since 1991 have subsequently identified submarine channels deeper in the stratigraphic section that are much larger than the Johns Valley channel sandstones. The larger channel sandstones occur in the upper Jackfork Group and have been studied at four outcrop locations. Channel sandstones >800 ft in thickness and >4 mi in width have been identified. These sandstones are highly friable and have sedimentary characteristics unlike the "classic" Jackfork flysch sequence exposed along U.S. Highway 259 south of Big Cedar, Oklahoma.

The sedimentologic and lithologic characteristics of the upper Jackfork channel sandstones in the Lynn Mountain syncline indicate that they are proximal turbidites deposited in the upper part of a middle-fan environment. Paleocurrent data indicate that the channel sandstones were derived from the north, probably across a normal-marine shelf. Previous studies have interpreted an easterly source for the Jackfork Group with proximal turbidites being largely confined to Arkansas. Field studies from the Lynn Mountain syncline suggest that this depositional model for the Jackfork Group is too simplistic and does not accurately portray the depositional setting for the upper Jackfork Group in Oklahoma.

In the Lynn Mountain syncline, the channel sandstones appear to have the best hydrocarbon reservoir potential. Lithologically, they are similar to porous Jackfork sandstones that have tested, or are currently producing, gas south of the Ti Valley thrust fault in Latimer County, Oklahoma.

Pauli, David, 1994, Friable submarine channel sandstones in the Jackfork Group, Lynn Mountain syncline, Pushmataha and Le Flore Counties, Oklahoma, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 179–202.

PREVIOUS STUDIES

The most continuous exposures of the Jackfork Group in Oklahoma occur in the Lynn Mountain syncline along U.S. 259 south of Big Cedar, Le Flore County, and along the Indian Service Road southeast of Albion in Pushmataha County. Cline and Moretti (1956) described and correlated 5,600 and 6,000 ft, respectively, of Jackfork sedimentary rocks at these two locations, which are separated by a distance of 22 mi.

The geologic map of the Lynn Mountain syncline has evolved through time from the contributions of various workers who independently mapped smaller areas within the syncline. Honess (1923,1924) provided the earliest significant description of the geology of the Lynn Mountain syncline. In 1960, Cline published a geologic map of the southwestern part of the syncline west of U.S. Highway 271. Briggs (1973) mapped the eastern part of the Lynn Mountain syncline and expanded on earlier work done by Smith (1967). His map extended from the Oklahoma/Arkansas border westward to the Pushmataha County/Le Flore County, Oklahoma line. The central part of the syncline, in Pushmataha County, was mapped by Schroeder (1975) and Pitt and others (1982). This area of the syncline is noteworthy for its greater structural complexity than the eastern and southwestern areas.

Many workers have described the lithologic and sedimentologic features of the Jackfork Group formations. An overview of the Jackfork Group stratigraphic nomenclature is provided in Pitt and others (1982). Lithologic descriptions of the Jackfork Group formations are provided in Cline (1960) and Shelburne (1960). The sedimentary features characteristic of the Ouachita flysch sequence in general are well illustrated in Cline (1970). Briggs (1973) discussed the sedimentary features commonly found in the Jackfork flysch in the Lynn Mountain syncline and briefly discussed the provenance of the sandstones. Chamberlain (1971) interpreted the paleobathymetry of the flysch from trace fossils. Regional summaries for the Jackfork depositional setting can be found in Morris (1971; 1974a,b), Moiola and Shanmugam (1984), and Coleman and others (1994).

Jackfork Stratigraphic Sequence, Lynn Mountain Syncline

The formations of the Jackfork Group in the Lynn Mountain syncline are illustrated in Figure 1. Differences in the Jackfork Group formation designations and age assignments are largely due to difficulty in identifying and correlating key stratigraphic markers throughout the syncline and the lack of diagnostic fossils within the sequence.

The base of the Jackfork Group is approximately defined by a siliceous shale unit called the Chickasaw Creek Formation. This formation is one of the most widespread stratigraphic intervals in the Ouachita Mountains and has been mapped throughout most of Oklahoma and Arkansas. Cline (1960) placed the Chickasaw Creek in the uppermost Stanley Group. Briggs (1973), however, included it in the Jackfork Group based on a recommendation to him by the Oklahoma Geological Survey.

In the western part of the syncline, Cline and Moretti (1956) were able to recognize most of the Jackfork formation designations proposed by Harlton (1938). Harlton originally subdivided the Jackfork into four formations based on the presence of widespread siliceous shale units that he identified in the Round Prairie and Tuskahoma synclines. His formations, in ascending order, are the Wildhorse Mountain, Prairie Mountain, Markham Mill, and Wesley. He later added (from a suggestion by Cline) the Game Refuge Formation to the Jackfork Group to distinguish a widespread fossiliferous sandstone unit found between the Wesley and Johns Valley Formations.

In his mapping of the eastern Lynn Mountain syncline, Briggs (1973) combined three of the Jackfork formations (Prairie Mountain, Markham Mill, and Wesley) into a single mappable unit. He was unable to differentiate these formations because he could not trace some of the siliceous shale intervals used by Cline (1960) to map the Jackfork formations in the western part of the syncline. He also assigned all of the strata above the Wesley to the Game Refuge Formation because he could not identify the Johns Valley Shale in his mapped area. Limestone olistoliths, a feature unique to the Johns Valley interval, are present in the western Lynn Mountain syncline (see Cline, 1960), but they have not been found in the eastern part of the syncline. Cline's Game Refuge, Johns Valley, and Atoka Formations were also combined by Briggs into a single map unit that Briggs designated the Game Refuge Formation.

Legg and others (1990) used a modified version of the nomenclature provided by Cline (1960) for the Lynn Mountain syncline stratigraphic section. They interpreted the Game Refuge Formation to be basal Atokan in age and to represent channel-fill sandstones incised into the Johns Valley Shale. Consequently, the Wesley Formation rather than the Game Refuge Formation was designated by SOHIO as the youngest Jackfork Group formation.

To date, fossils useful for determining the age of the Jackfork Group in the Lynn Mountain syncline have not been found. However, by age relations established for the Jackfork Group from other areas, it appears that the Jackfork interval in the Lynn Mountain syncline is mostly Morrowan in age. Morrowan fossils have been identified from the Game Refuge Formation (Cline terminology) in the Boktukola syncline, the next large syncline immediately south of the Lynn Mountain syncline. The conodont species *Idiognathoides ouachitensis* has been identified from the Johns Valley interval in the Lynn Mountain syncline, indicating a late Morrowan–early Atokan age (Bob Grayson, personal communication). The presence of Morrowan fossils in the Jackfork Group indicates that it is, at least in part, Pennsylvanian in age and not entirely Mississippian as Cline (1960) believed. The presence of Chesterian Pitkin limestone clasts in the Chickasaw Creek Formation in Arkansas (Gordon and Stone, 1977) also indicates that the Jackfork Group is Pennsylvanian.

SOHIO Field Studies and Exploration in the Lynn Mountain Syncline

Prior to drilling the No. 1-7 Weyerhaeuser well in 1985 in the Lynn Mountain syncline, SOHIO geologists conducted field studies in the area to identify potential reservoir sandstones. Their field work identified channel sandstones in the stratigraphic interval that they be-

lieved was equivalent to the Johns Valley Shale although it was not mapped as such on existing geologic maps. SOHIO noted the lithologic similarity of the Johns Valley channel sandstones in the syncline to the basal Atoka Spiro sandstones in the frontal Ouachitas and hence applied the name "Spiro" to these sandstones (Figs. 1,2).

On the basis of SOHIO's field studies, the channel sandstones in the Johns Valley Shale became the primary objective of the SOHIO No. 1-7 Weyerhaeuser well (Legg and others, 1990). The well was drilled to test a submarine channel in the Johns Valley Shale that SOHIO had mapped by using high-resolution seismic data shot spe-

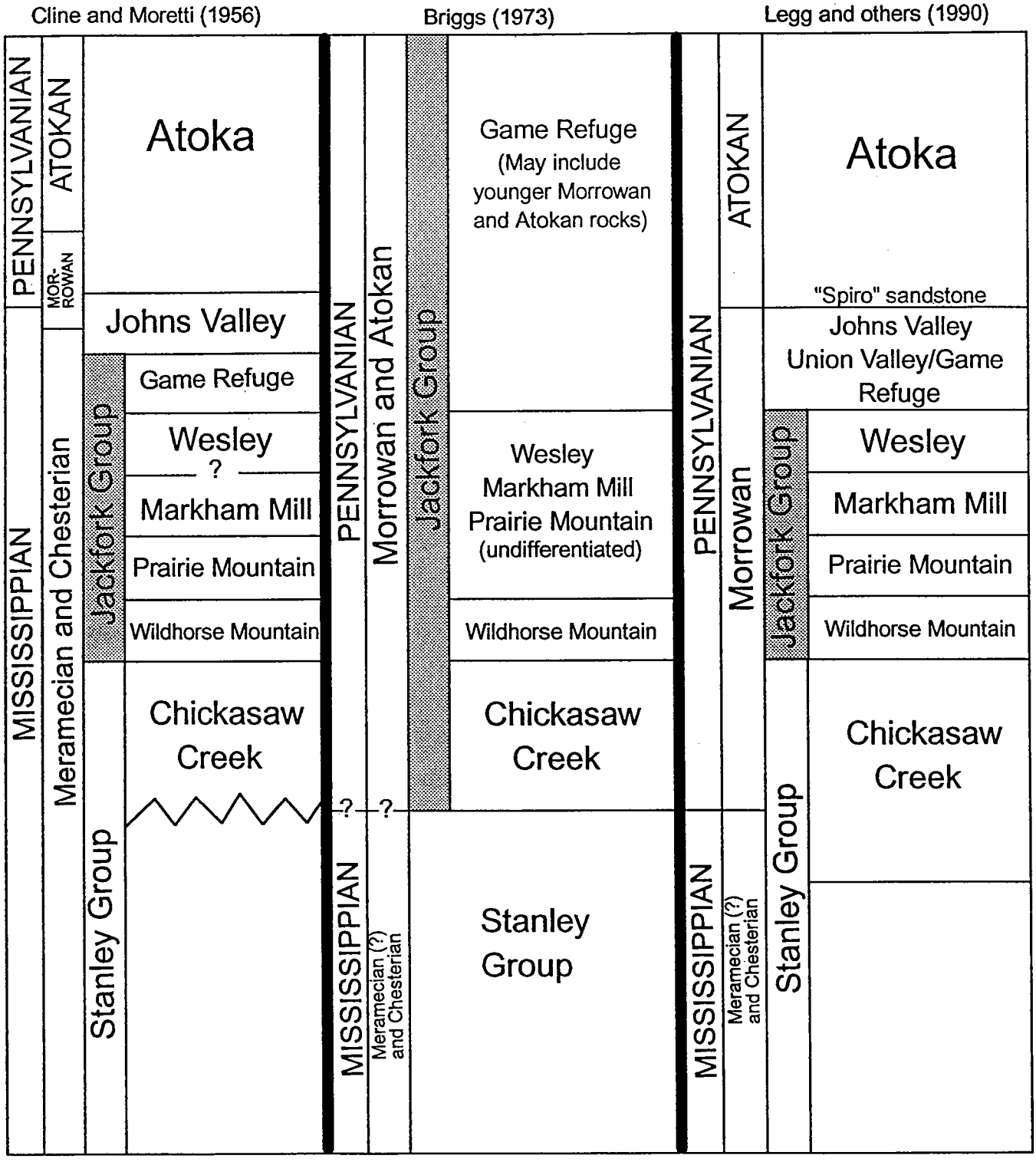


Figure 1. Jackfork stratigraphic sequence, Lynn Mountain syncline (adapted from Cline and Moretti, 1956; Briggs, 1973; and Legg and others, 1990).

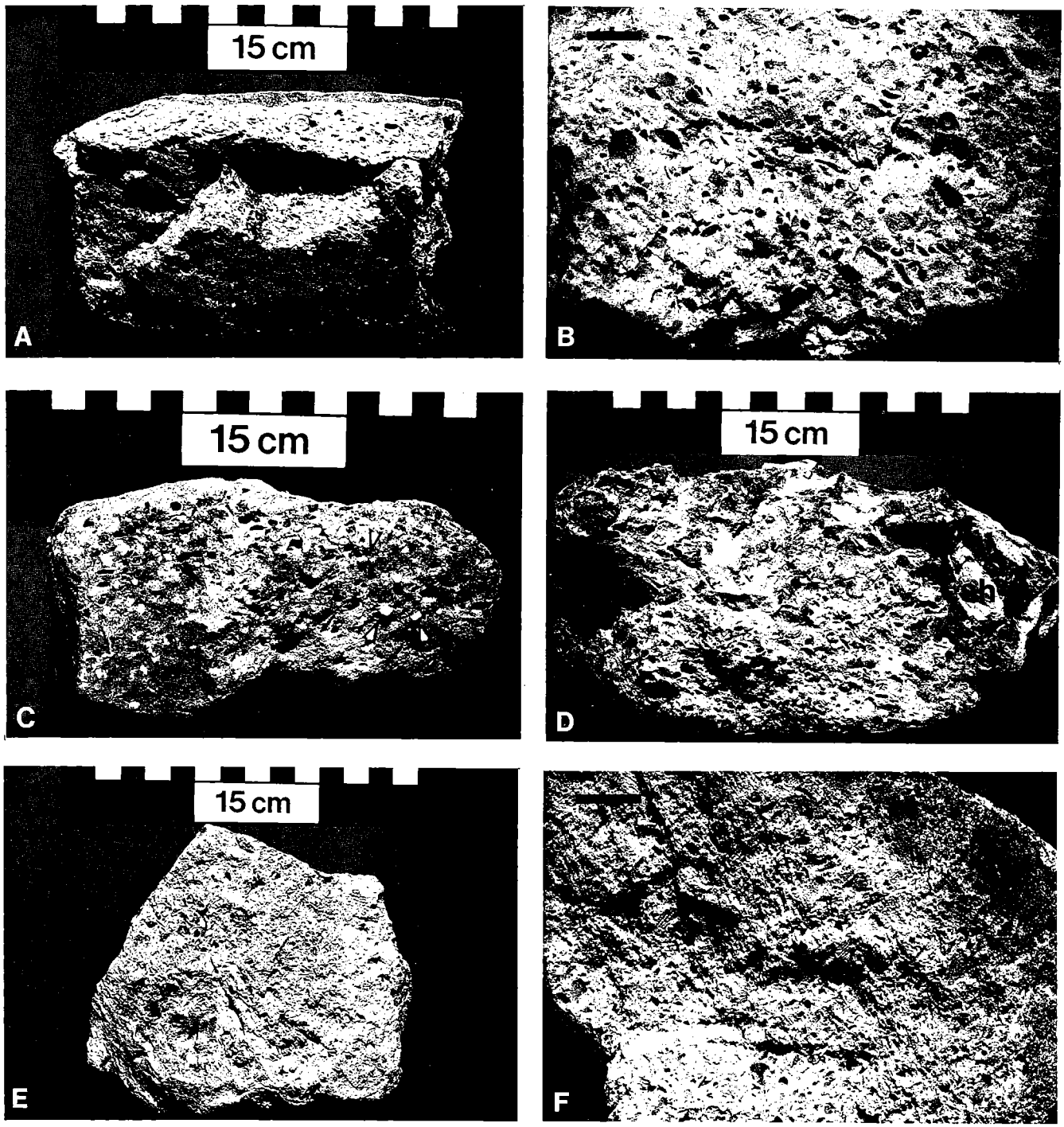


Figure 2. Comparison of "Spiro" channel sandstones (2A–D) from the Lynn Mountain syncline to Spiro sandstones collected in the frontal Ouachitas (2E,F). Similarities include abundant fossil molds and iron staining. The Spiro sandstones, however, lack the quartz pebbles that are common in the Lynn Mountain syncline "Spiro" channel sandstones. *A*—Ferruginous sandstone with abundant fossil molds and quartz granules. *B*—Fossiliferous quartz-pebble sandstones from lower portion of a "Spiro" channel. Molds of crinoid columnals (c) are the dominant fossil constituent in this sandstone. Bar represents 1 cm. *C*—Sandstone containing abundant, well-rounded white quartz granules and pebbles (arrows). *D*—Fossiliferous sandstone with quartz pebbles and green shale rip-up clasts (sh) collected from lower portion of "Spiro" channel sandstone. *E*—Spiro sampled north of Limestone Ridge in sec. 24, T. 5 N., R. 19 E., Latimer County, Oklahoma. The Spiro sandstone at this locality is iron stained and contains numerous molds of fragmented fossils. *F*—Basal Atokan sandstone overlying olistolith-bearing shale of Johns Valley Shale Formation in the NW¼NW¼NW¼ sec. 35, T. 4 N., R. 19 E., Latimer County, Oklahoma. This locality is ~1.5 mi south of the Ti Valley fault. Sandstones consist of abraded fossil material (molds) and are deeply iron stained. Bar represents 1 cm.

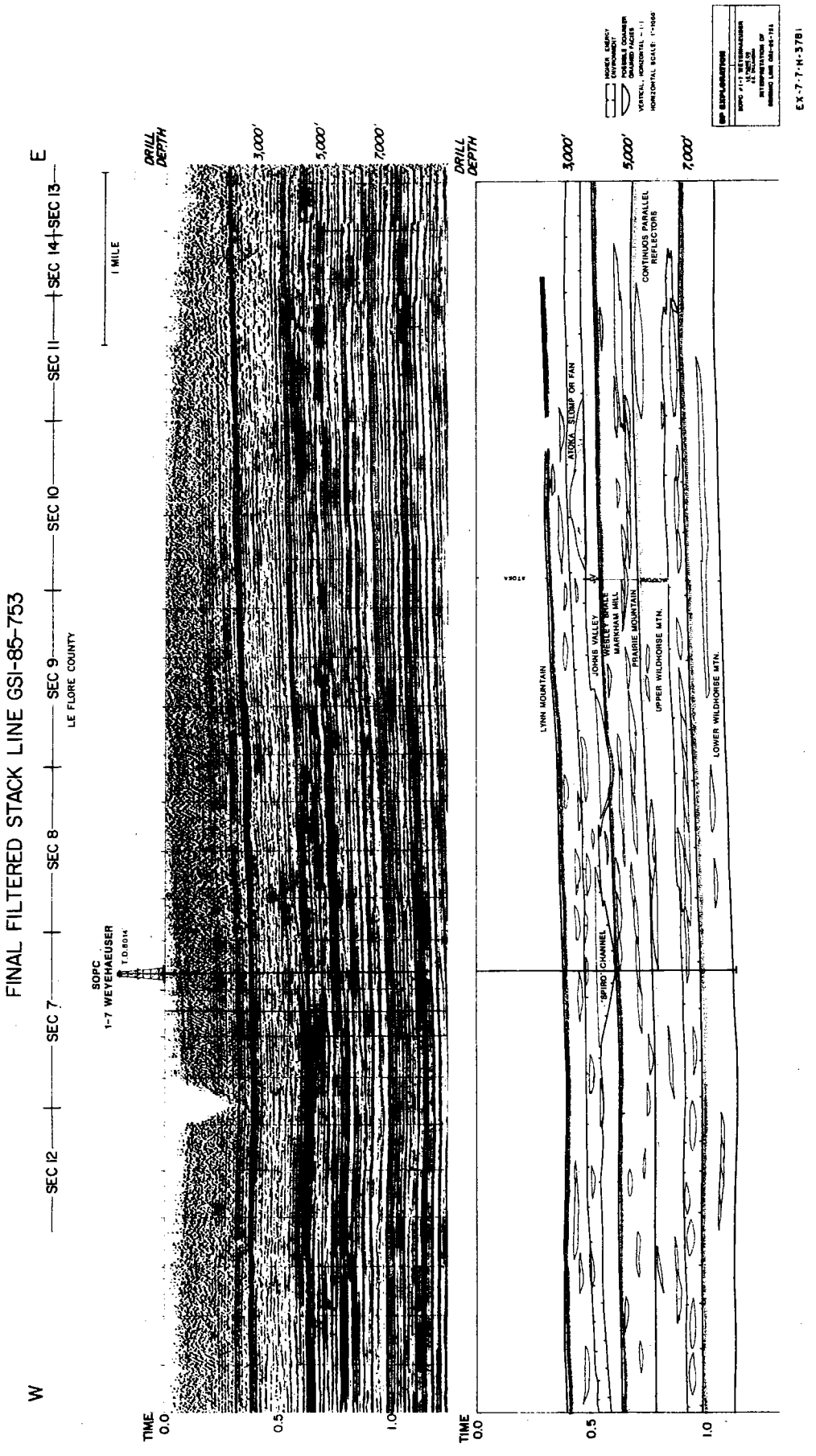


Figure 3. "Spiro" channel drilled by the SOHIO No. 1-7 Weyerhaeuser well (sec. 7, T. 1 N., R. 23 E.) on the north flank of the Lynn Mountain syncline. Gas shows were encountered during the drilling of the Johns Valley and Jackfork intervals. The interpretation also shows possible channels located deeper in the Jackfork sequence. (From Legg and others 1990.)

cifically for identifying stratigraphic targets (Fig. 3). Secondary targets included smaller fan-lobe and channel-like features in the Jackfork Group. The well confirmed the presence of Johns Valley channel sandstones in the subsurface and the viability of SOHIO's play concept by penetrating a channel-fill sequence interpreted by SOHIO to be nearly 700 ft thick. The channel-fill sequence consisted largely of shale, but it contained two clean channel sandstones that were 35 and 60 ft thick, respectively. A core from the lower channel sandstone consisted of fossiliferous quartz sandstones similar to the channel sandstones identified in the Johns Valley Shale outcrop. Gas was encountered in the lower channel sandstone and was flared on all subsequent bit trips to total depth. Despite numerous gas shows, the well was determined to be noncommercial, and it was plugged and abandoned.

WEYERHAEUSER LYNN MOUNTAIN SYNCLINE FIELD STUDIES

Weyerhaeuser initiated field studies in the Lynn Mountain syncline beginning in early 1991 in an effort to document the distribution and reservoir quality of the basal Atokan "Spiro" channel sandstones previously recognized by SOHIO. While studying the "Spiro" channel sandstones for Weyerhaeuser, Grayson (1991) identified the presence of channel sandstones in the upper Jackfork. These channel sandstones appeared to be much thicker and have a much broader distribution than the overlying "Spiro" sandstones. Grayson recommended additional studies of the upper Jackfork sandstones because their lithologic characteristics were similar to the "Spiro" channel sandstones.

Follow-up studies focusing on the upper Jackfork channel sandstones by Cuyler (1992), Tillman (1992), and myself have confirmed Grayson's initial observations. These studies have documented the existence of large channel sandstone complexes at four locations in the Lynn Mountain syncline in the upper Jackfork Group formations extending from sec. 29, T. 1 N., R. 18 E. to sec. 34, T. 2 N., R. 25 E., over a distance of ~48 mi (Fig. 4).

Measured sections, surface mapping of the channel sandstones and their associated facies, and petrographic examination of the upper Jackfork channel sandstones, including a comparison to the producing Jackfork sandstones in Latimer County, Oklahoma, have provided a greater understanding of the geology and reservoir potential of these sandstones in the Lynn Mountain syncline. This paper presents the results of the geologic studies conducted over the past three years in the channel-bearing upper Jackfork Group formations in the Lynn Mountain syncline.

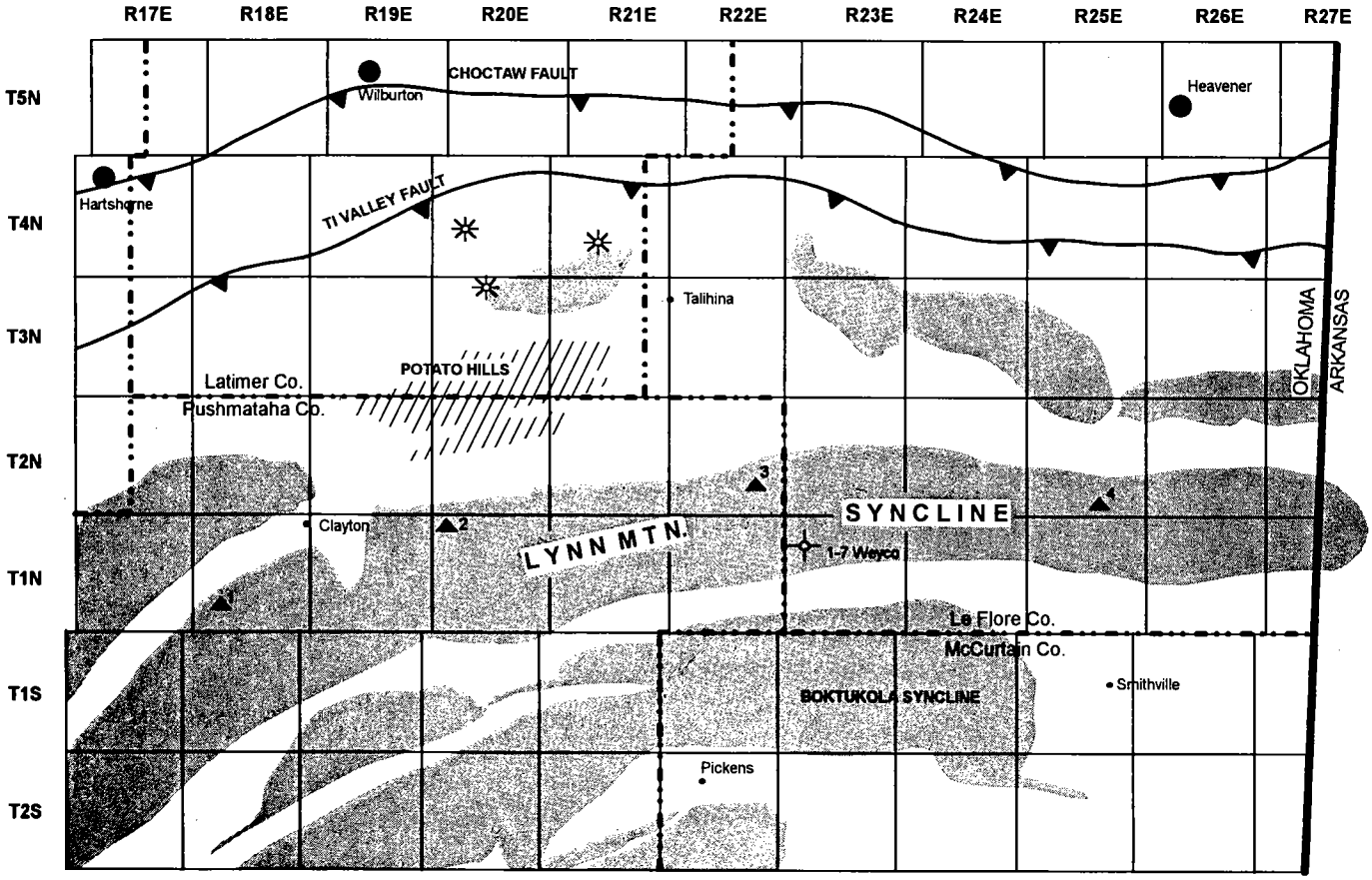


Figure 4. Location of channel sandstones in the upper Jackfork Group formations in the Lynn Mountain syncline. Numbered triangles indicate localities where Weyerhaeuser has conducted field studies. 1 = Prairie Mountain, 2 = McKinley Rocks, 3 = Holly Creek, and 4 = U.S. Highway 259. Gas symbols mark the locations of wells currently producing from the Jackfork in Latimer County, Oklahoma.

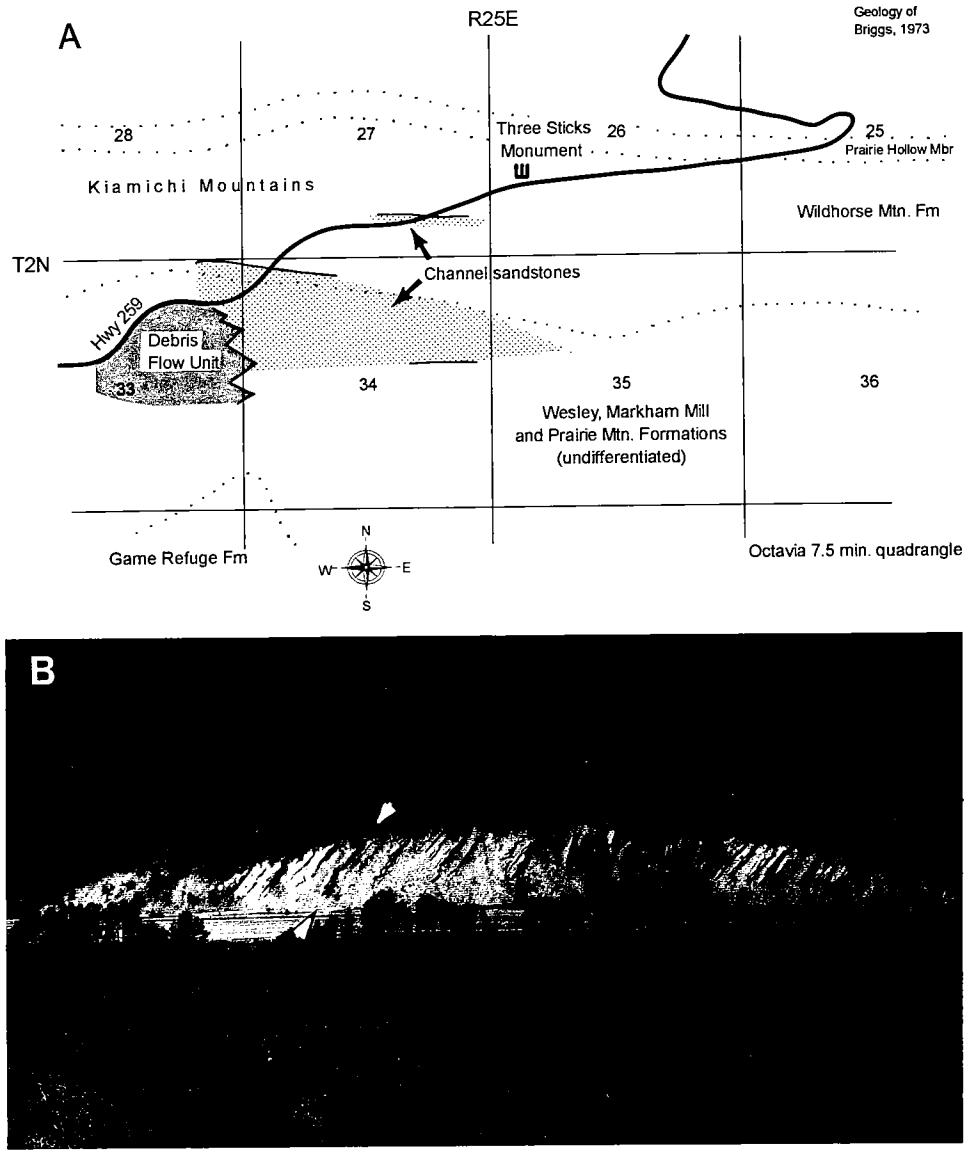


Figure 5. *A*—Location of friable channel sandstones in the upper Wildhorse Mountain Formation (Briggs, 1973) exposed along U.S. Highway 259 south of Three Sticks Monument. A debris-flow unit occurs within the channel sandstones along the highway in the NE¼ sec. 33. *B*—Aerial view (looking west) of friable channel sandstones overlying deep-water thin-bedded turbidites in sec. 34, T. 2 N., R. 25 E., Le Flore County, Oklahoma. Arrows show contact.

Characteristics of Upper Jackfork Channel Sandstones

Upper Jackfork channel sandstones in the Lynn Mountain syncline are well exposed along U.S. 259 in secs. 27 and 34, T. 2 N., R. 25 E., Le Flore County, and at McKinley Rocks in sec. 6, T. 1 N., R. 20 E., Pushmataha County. The following discussion summarizes some of the sedimentologic features of the channel sandstones at these locations.

Jackfork Channel Sandstones along U.S. 259 (secs. 27 and 34, T. 2 N., R. 25 E.)

Two Jackfork channel sandstones are exposed along U.S. 259 near the top of the Wildhorse Mountain Formation (Fig. 5), as mapped by Briggs (1973). Morris and others (1979) illustrated, but did not describe in detail, the upper (sec. 34) channel exposed at this locality. Cline and Moretti (1956) measured 40 and 50 ft for the upper and lower (sec. 27) channel sandstones, respectively. However, they only measured the lowermost part of the upper channel sandstone located in the NW¼NW¼NW¼ sec. 34, T. 2 N., R. 25 E. The upper channel sandstone has been traced farther south to the center of sec. 34. It is abruptly overlain by hard, nonchannel turbidites in the E½ sec. 34, and it has a calculated thickness of almost 1,000 ft.

The Jackfork channel sandstones along U.S. 259 are medium-grained quartz arenites. They are extremely friable (Fig. 6A,C), in marked contrast to most of the Jackfork sandstones exposed along the highway, which are predominantly very hard, tightly cemented orthoquartzites. Green shales overlie and are interbedded with the channel sandstones. These shales are noticeably lighter in color than the dark gray and black shales that are found throughout most of the Jackfork flysch sequence exposed along U.S. 259.

The basal contacts of the U.S. 259 channel sandstones are very sharp. The basal contact of the sec. 27 channel is exposed on a broad dip-face just south of Three Sticks Monument (Fig. 5) where friable sandstones overlie medium-bedded orthoquartzites along a sharp planar contact (Fig. 6D). At the sec. 34 channel locality, the basal contact is an undulatory scour surface (Fig. 6B) along which the channel sandstones are incised into thin-bedded turbidites that contain a deep-water trace-fossil assemblage (Fig. 6A). Small scours also occur between beds within the channel sandstones.

Jackfork Channel Sandstones at McKinley Rocks (sec. 6, T. 1 N., R. 20 E.)

The most complete exposure of a Jackfork channel complex found to date in the Lynn Mountain syncline occurs at McKinley Rocks (Fig. 7). At least three stacked channel sequences occur within a stratigraphic interval of 820 ft in the upper Jackfork Group formations at this

location. The thickest individual channel sequence at this locality is ~144 ft thick (Fig. 8).

Fossiliferous, medium- to coarse-grained, poorly sorted quartz arenites occurring in beds that are 2–5 ft thick are found near the base of the channel sequence (Fig. 9). These sandstones are characterized by the presence of abundant fossils (preserved as molds), plant fragments, and numerous shale rip-up clasts. They also contain many well-rounded granule- to pebble-sized white quartz grains that typically "float" in a fine- to medium-grained sandstone matrix.

The coarsest sandstones overlie well-sorted, medium-grained quartz arenites along scoured contacts having as much as 2 ft of relief (Fig. 10A,B). Overall, the channel sequence fines upward with medium-grained and fine- to medium-grained sandstones composing the middle and upper parts of the sequence, respectively. The medium-grained sandstones occur in discontinuous beds that have lenticular (Fig. 10C) and tabular-planar (Fig. 10D) geometries. Scoured surfaces may also occur within these sandstones where individual beds terminate abruptly along discrete high-angle surfaces (Fig. 10D).

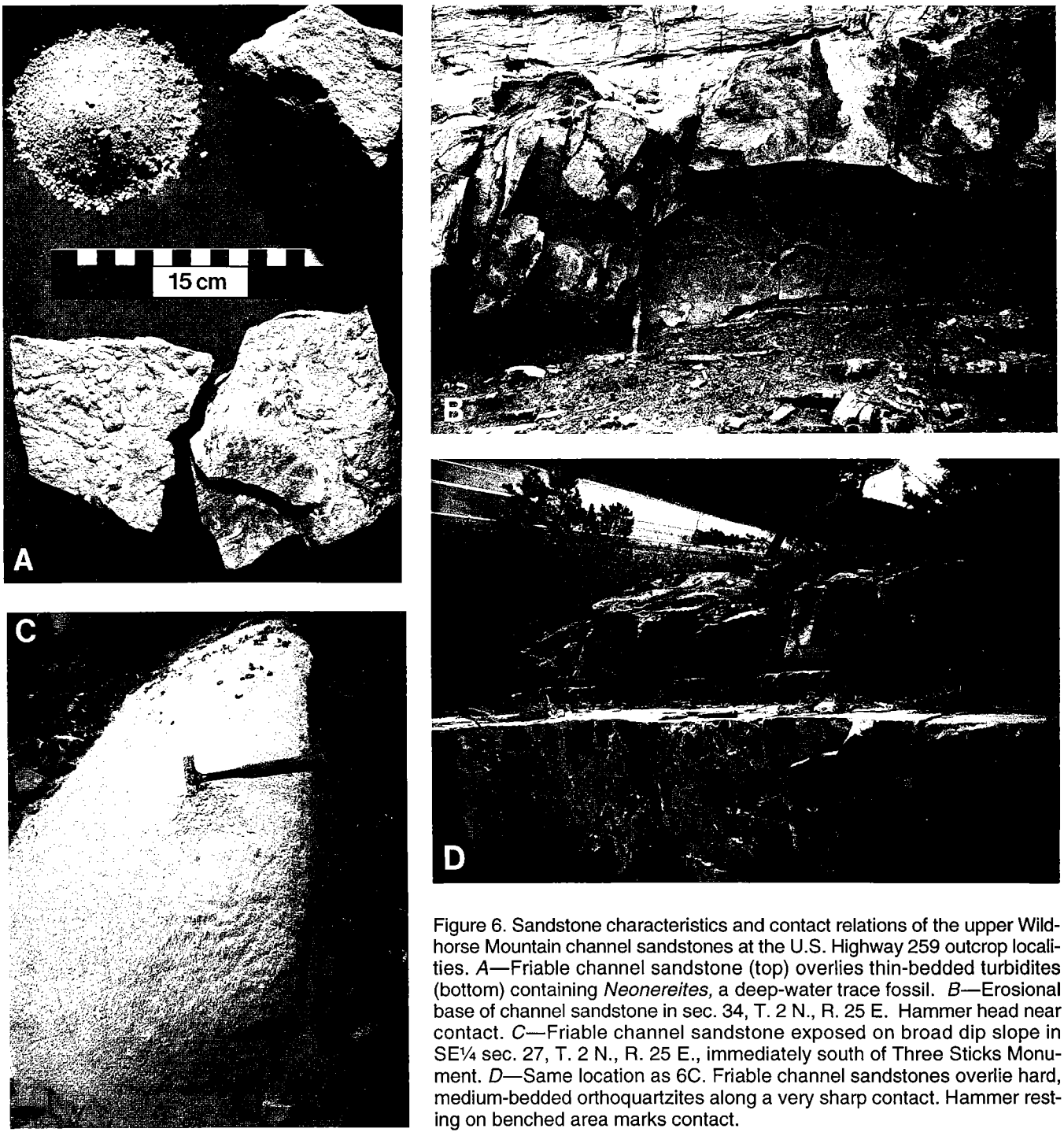


Figure 6. Sandstone characteristics and contact relations of the upper Wildhorse Mountain channel sandstones at the U.S. Highway 259 outcrop localities. *A*—Friable channel sandstone (top) overlies thin-bedded turbidites (bottom) containing *Neonereites,* a deep-water trace fossil. *B*—Erosional base of channel sandstone in sec. 34, T. 2 N., R. 25 E. Hammer head near contact. *C*—Friable channel sandstone exposed on broad dip slope in SE¼ sec. 27, T. 2 N., R. 25 E., immediately south of Three Sticks Monument. *D*—Same location as 6C. Friable channel sandstones overlie hard, medium-bedded orthoquartzites along a very sharp contact. Hammer resting on benched area marks contact.

Although scoured surfaces are common within the channel-sandstone sequence, sole marks are absent on the undersides of individual sandstone beds (Fig. 10E).

Paleocurrent indicators are typically difficult to find in the Jackfork channel sandstones. Even at the McKinley Rocks locality, where internal stratification within sandstone beds can be observed, interpreting paleocurrents from bedding features is difficult. Figure 11A shows tabular-planar bedding in medium-grained sandstones that are sharply truncated along a scoured surface by fossiliferous sandstones containing quartz granules. When corrected for structural dip (31°S), the scoured surface and the foreset laminations in the tabular-planar beds are nearly flat, and the tabular-planar beds dip ~35°N. The interpretation of paleocurrents from this outcrop is further complicated by the fact that the entire sequence of tabular-planar–bedded sandstones below the scour surface appears to be contained within a large trough-shaped scour whose axis is oriented north-south. A north-south orientation is more consistent with paleocurrent measurements taken on current ripples within Jackfork channel sandstones at other localities in the Lynn Mountain syncline.

The upper part of the lowest channel sequence at McKinley Rocks is not well exposed. Sandstone units above 240 ft on the measured section have sedimentary features indicative of both channel and nonchannel sandstones. The poorer exposure suggests that interbedded shales are present in the interval. Shale is much more common in the nonchannel facies. These relations are interpreted to mark the termination of sand deposition in the lowest channel and the onset of sedimentation in a fan-lobe or channel-levee setting.

Table 1 summarizes the characteristics of the channel and nonchannel sandstones observed in upper Jackfork strata in the Lynn Mountain syncline. The sedimentary features associated with the nonchannel sandstones are identical to those typically described from the "classic" Jackfork flysch (Cline, 1970).

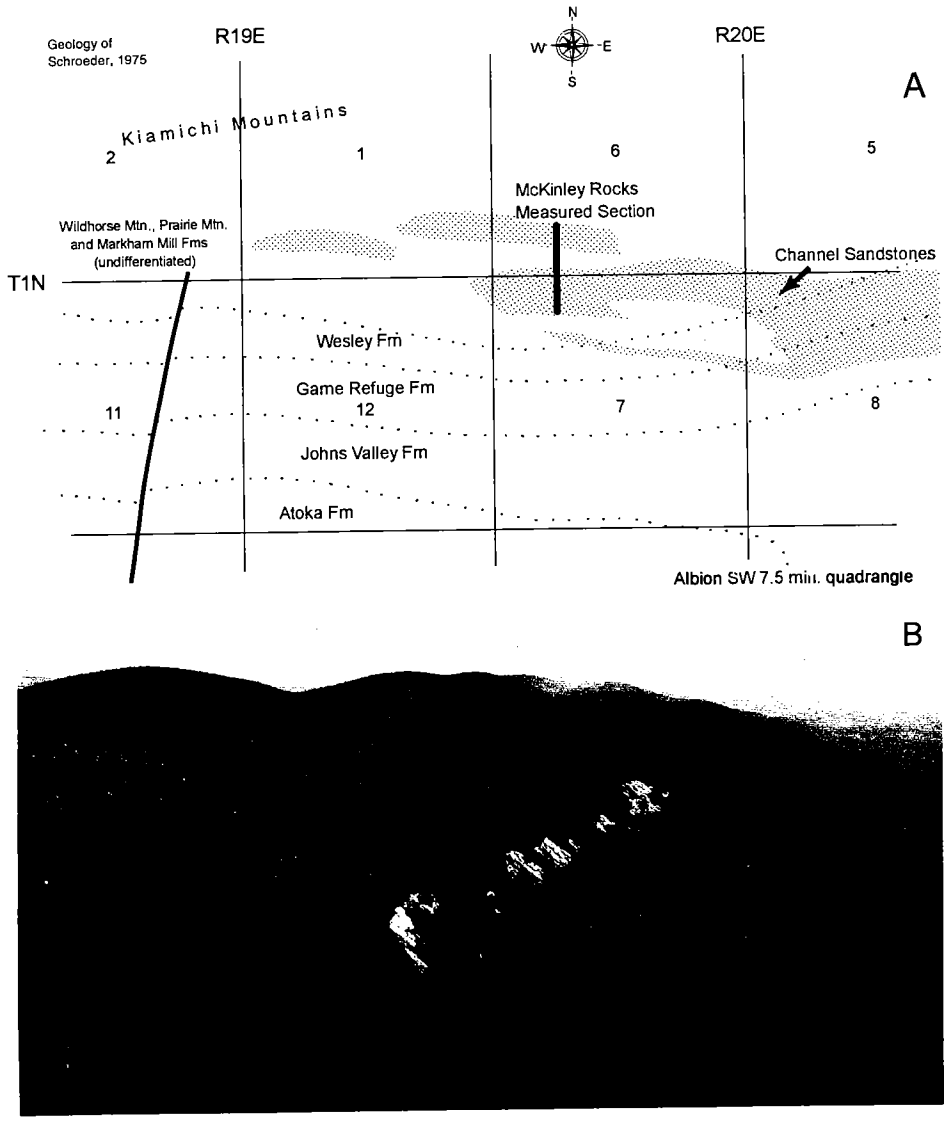


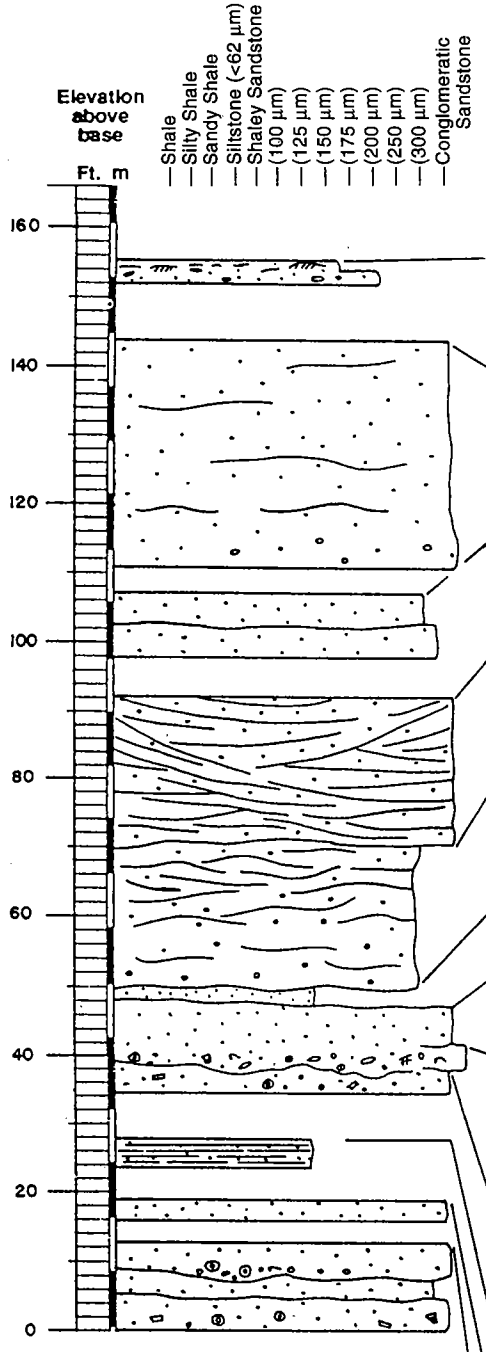
Figure 7. *A*—Location of exhumed channel sandstone in the upper Jackfork Group formations at the McKinley Rocks locality in sec. 6, T. 1 N., R. 20 E. *B*—Aerial view (looking east) of McKinley Rocks outcrops. These sandstone outcrops constitute the stratigraphic interval from 0 to 92 ft illustrated in Figure 8.

Areal Extent of Channel Sandstones and Associated Nonchannel Facies

Figure 12 shows the areal extent of channel and nonchannel sandstones based on field mapping in the area of Holly Mountain and along Honobia River. The main channel-bearing interval occurs along Holly Mountain in the upper Jackfork Group. It appears that all of Holly Mountain consists of a complex of stacked channel sandstones. Thinner channel sandstones (30–110 ft thick) are also present more locally in the upper Jackfork and in the Johns Valley Shale near the Atoka contact (i.e., "Spiro" channels).

Clean, fine- to medium-grained friable sandstones that locally contain quartz pebbles (Fig. 13A) dominate outcrops within the Holly Mountain channel complex. Channel sandstones at the base of the channel complex contain oriented plant fragments (Fig. 13B). The overlying Johns Valley Shale consists mostly of green shale, but it also contains thin- to medium-bedded, fine-grained orthoquartzites (i.e., nonchannel sandstones). Orthoquartzites in the Johns Valley Shale contain abundant sole marks and trace fossils (Fig. 13C,D). Nonchannel sandstones are also found below the Holly Mountain channel com-

Described by D. A. Pauli 11/8–9/93
Albion SW Quadrangle (7.5')
Access is Weyerhaeuser logging road 95500, which is 2.8 mi S of the Clayton Lake State Park entrance on Hwy 271. Follow 95500 to K Trail. Base of section is immediately N of the highest outcrop (OTC 3) at McKinley Rocks. Measured 0–160 ft on N-S traverse across OTC 3. 160–835 ft measured along K Trail. Sedimentary features are not readily observable in most units between 92 and 835 ft, and facies interpretations are based primarily on lithologic similarities. Structural strike, N. 80° W.; dip, 31° S.



Figures:
10D
10C
10A,B
11A
10E
9A,B

Unit No.	Thickness		
19	25'	155–180 ft	Covered.
18	3'	152–155 ft	Sandstone (150–200 µm). Fine-grained, well-sorted. Thin-bedded with ripple laminations and abundant randomly oriented plant detritus along laminae. Black shale rip-up clasts up to 1 cm in diameter. Hard orthoquartzite.
17	8'	144–152 ft	Covered.
16	33'	111–144 ft	Sandstone (350–500 µm, 1000 µm maximum). Medium- to upper-medium-grained, moderately sorted, subrounded to rounded grains with some floating quartz granules at 112 ft. Very friable. Fe-stained. Trace mica. Clean quartz arenite. CHANNEL.
15	4'	107–111 ft	Covered.
14	9'	98–107 ft	Sandstone (350–500 µm). Medium- to upper-medium-grained, moderately sorted, subrounded to rounded grains. Very friable. Fe-stained. Trace mica. Clean quartz arenite. CHANNEL.
13	6'	92–98 ft	Covered. Channel-sandstone float.
12	22'	70–92 ft	Sandstone (350–750 µm, 1000 µm maximum). Medium- to lower-coarse-grained, moderately sorted, rounded to well-rounded grains. Very friable. White micas (1–2%). Fe rind on outer surface, which when broken reveals friable sands. Clean quartz arenites. Irregular tabular-planar (beds pinch and swell) bedding. Broad dip face at 92 ft weathers into nodular blocks. CHANNEL.
11	20'	50–70 ft	Sandstone (250–350 µm, 500 µm maximum). Upper-fine- to medium-grained, moderately to well-sorted, rounded to well-rounded grains. Fe-stained. Trace mica. More cemented than sand units above and below. Beds pinch and swell. Sand at 50 ft is poorly sorted and dirtier, and contains more black carbonaceous grains than rest of unit. CHANNEL.
10	2'	48–50 ft	Sandstone (100–150 µm). Fine-grained, well-sorted. Hard orthoquartzite. No mica. Very clean arenite. CHANNEL.
9	8'	40–48 ft	Sandstone (250–500 µm, 1000 µm maximum). Upper-fine- to upper-medium-grained, moderately to poorly sorted, sub-rounded to well-rounded grains. Common black carbonaceous grains. Moderately friable. Trace mica. Fe-stained. CHANNEL.
8	2'	38–40 ft	Sandstone (350–750 µm). Medium- to lower-coarse-grained, poorly sorted, rounded to well-rounded. Fossiliferous (molds of crinoid columns, whole thin-walled brachiopod valves, fenestrate bryozoan fragments, gastropods). Abundant gray and green shale rip-up clasts (0.5 cm diameter). Common white quartz granules (2000–4000 µm) and pebbles (>4000 µm) floating in medium-sand matrix. Deeply Fe-stained. Similar sandstones at other outcrops along McKinley Ridge overlie scoured surfaces and are 1–5 ft thick. CHANNEL.
7	3'	35–38 ft	Sandstone (250–500 µm, 1000 µm maximum). Medium- to upper-medium-grained with some coarser grains, well-sorted, rounded to well-rounded grains. Crinoid columnals (molds). Common dark organic(?) grains. Friable. Fe-stained. No mica. CHANNEL.
6	7'	28–35 ft	Covered.
5	4'	24–28 ft	Sandstone (100–150 µm). Fine-grained, well-sorted. Hard orthoquartzite. Fine parallel laminations. No mica. Very clean arenite. CHANNEL.
4	5'	19–24 ft	Covered.
3	3'	16–19 ft	Sandstone (250–500 µm, 1000 µm maximum). Medium- to upper-medium-grained, well-sorted, rounded to well-rounded grains. Some black organic(?) grains. Clean arenite. Fe-stained. No mica. CHANNEL.
2	3'	13–16 ft	Covered.
1	13'	0–13 ft	Sandstone (350–500 µm, 2000 µm maximum). Medium- to upper-medium-grained with quartz granules locally, moderately to well-sorted, rounded to well-rounded grains. Fossil molds (crinoid columnals, brachiopods). Scoured surface at 9 ft. Small green shale rip-up clasts. Mica (<1%). Moderately friable. Fe-stained. CHANNEL.

Figure 8. Partial measured section of Jackfork Sandstone from the McKinley Rocks locality (NW¼SE¼SW¼ sec. 6, T. 1 N., R. 20 E., to NE¼NW¼NW¼ sec. 7, T. 1 N., R. 20 E.). The stratigraphic interval illustrated is interpreted to represent the lowest of at least three separate channel sequences that have been measured at this locality. Location of photographs shown in Figures 9, 10, and 11 are indicated along the left margin.

TABLE 1. — SEDIMENTARY AND OUTCROP CHARACTERISTICS OF CHANNEL AND NONCHANNEL SANDSTONES FROM THE UPPER JACKFORK GROUP, LYNN MOUNTAIN SYNCLINE

Channel Sandstones	Nonchannel Sandstones* (i.e., fan lobe or channel levee)
Fine- to coarse-grained. Coarsest sands contain floating quartz granules and pebbles and are poorly sorted.	Uniformly very fine grained to fine-grained.
Friable to entirely uncemented in outcrop.	Hard, well-cemented (i.e., orthoquartzites) in outcrop.
Fossiliferous. Molds of well-preserved fossils are common.	Rare fossils. Local occurrences of highly abraded fossils (i.e., molds).
Bouma sequences absent.	Partial to complete Bouma sequences.
Sole marks rare or absent. Load structures present.	Common sole marks (i.e., flutes and grooves).
Scoured contacts. Scoured surfaces have up to 2 ft of relief.	Rare small-scale scours.
Discontinuous beds with tabular-planar, lenticular, and trough cross-bedding. Current-ripple laminations locally.	Planar beds that are laterally continuous. Typically parallel-laminated. Ripple lamination and dish structures also present.
Sandstones fine upward.	Homogeneous grain size.
Minor interbedded shale. Shales typically light green and gray in color.	High percentage of interbedded shales. Shales typically dark gray and black in color. Shales may be siliceous.
Trace fossils absent or rare.	Common bedding-plane trace fossils.
Small clay rip-ups and plant fragments associated with coarsest sandstones found in lower part of channel sandstones.	Rip-up clasts present; often concentrated at top of beds.
Nodular float blocks.	Rhombic float blocks.
Deeply iron-stained. Typically have hard surface rind of iron-rich cements.	Iron staining limited to surface only; does not penetrate tightly cemented sandstones.

*Sedimentary features commonly found in the "classic" Jackfork flysch.

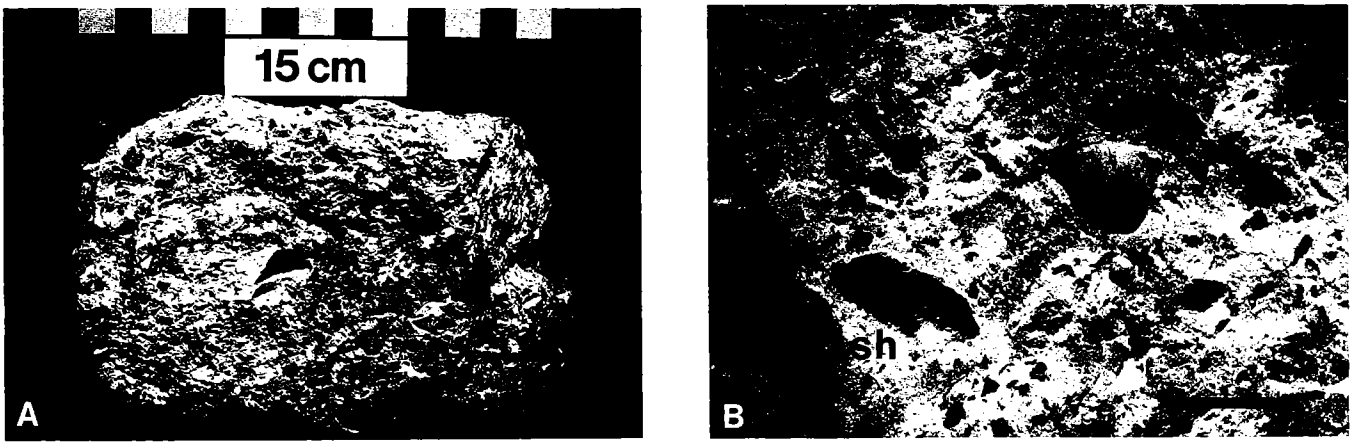


Figure 9. Sandstones sampled near the base of the McKinley Rocks section. *A*—Fossiliferous quartz arenite containing shale rip-up clasts, plant fragments, and "floating" quartz pebbles (arrow). *B*—Same as 9A showing an internal mold of a whole brachiopod valve (b) and mold of green shale rip-up clast (sh). Bar represents 1 cm.

plex. This sandstone-rich interval consists predominantly of fine-grained orthoquartzites containing incomplete Bouma sequences (mostly B and C divisions, Fig. 13E), dish structures (Fig. 13F), and sole marks.

The upper contact of the Holly Mountain channel complex is clearly defined by an east-trending strike valley in the S½ secs. 28, 29, and 30, T. 2 N., R. 23 E. The base of the channel complex is well exposed in sec. 28, T. 2 N., R. 23 E., where it abruptly overlies very hard nonchannel turbidites. Also, an apparent facies change occurs in sec. 28; this defines the eastern limit of the channel complex. In the east half of this section, clean, medium-grained quartz arenites are rare, and the sandstones are fine-grained sublitharenites.

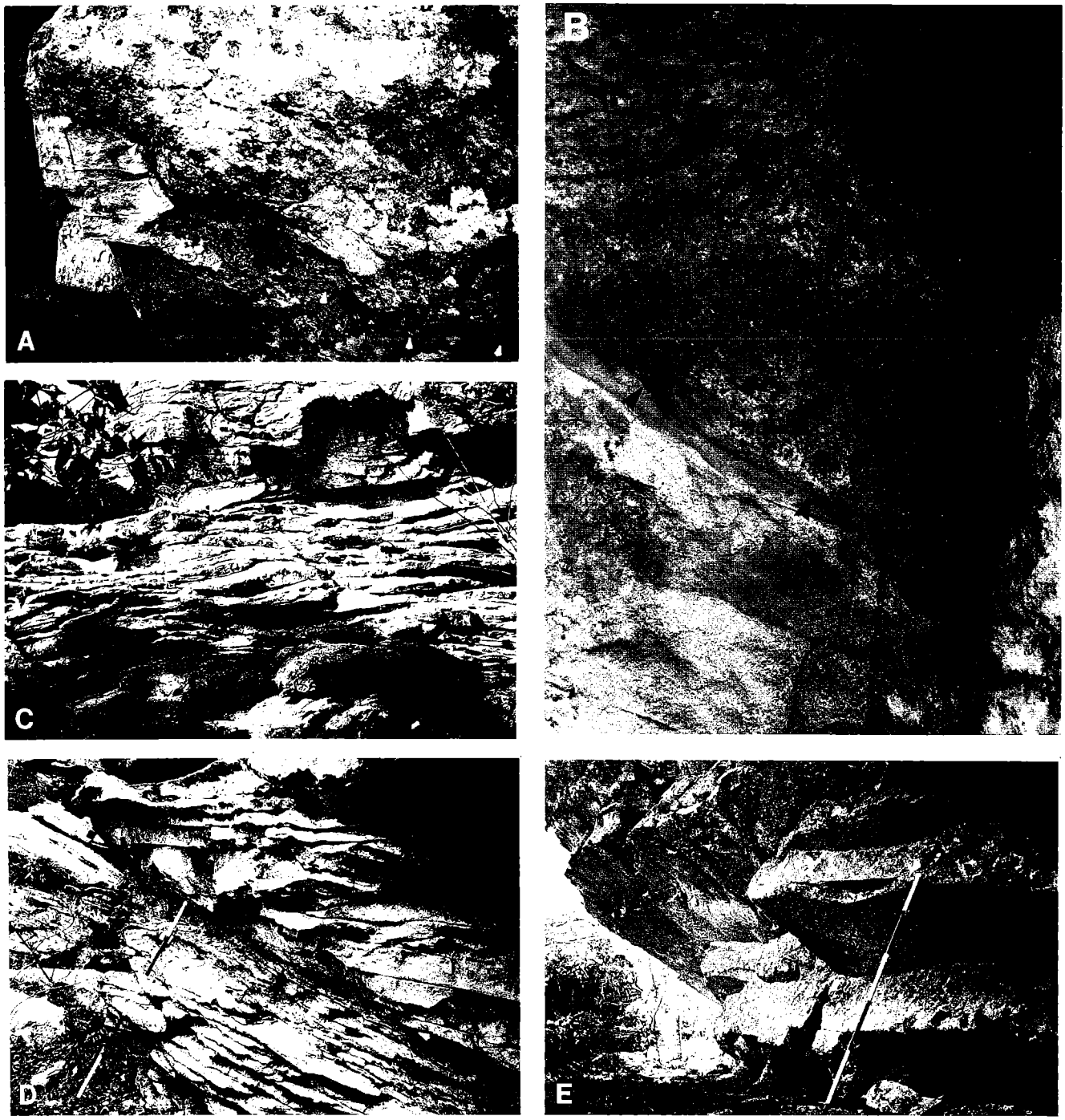


Figure 10. Bedding features and contact relations from outcrops at the McKinley Rocks locality. *A*—Fossiliferous quartz-pebble sandstone scoured into medium-grained sandstone. Maximum relief along this scour surface is ~2 ft. Pen cap is 2 in. long. *B*—Same as 10A. Close-up of scour surface. *C*—Lenticular bedding in middle part of channel sequence. *D*—Tabular-planar beds in medium-grained sandstones that truncate abruptly along scour(?) surface (arrows). *E*—Base of channel sandstone beds illustrating the absence of sole marks.

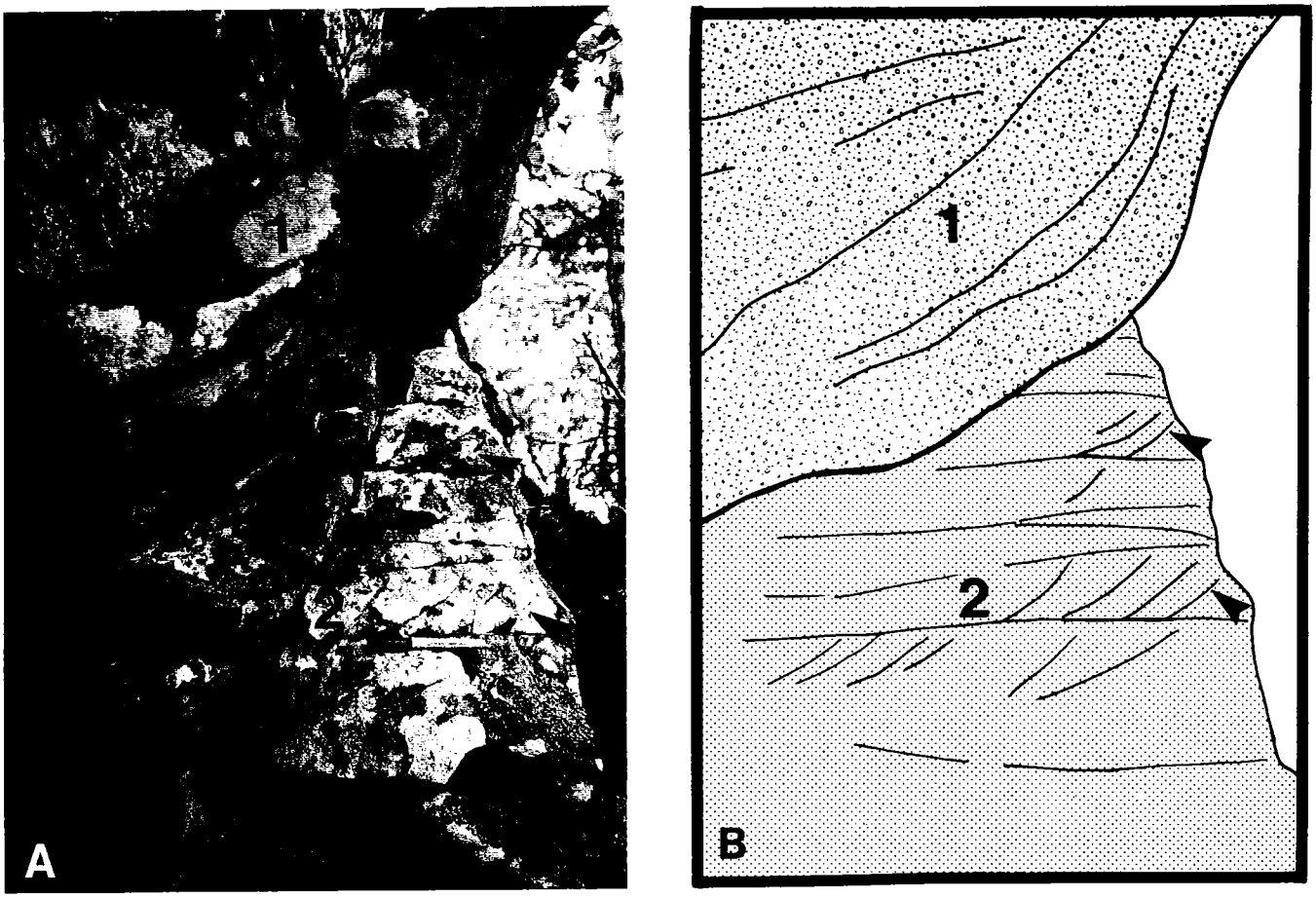


Figure 11. *A*—Scour contact between (1) fossiliferous quartz-pebble sandstone above and (2) medium-grained sandstone below. Highest pen in photo marks the contact. Foresets (arrows) in the lower tabular-planar–bedded sandstones dip toward the south (left in photo) but are nearly flat when corrected for structural dip. *B*—Schematic of 11A.

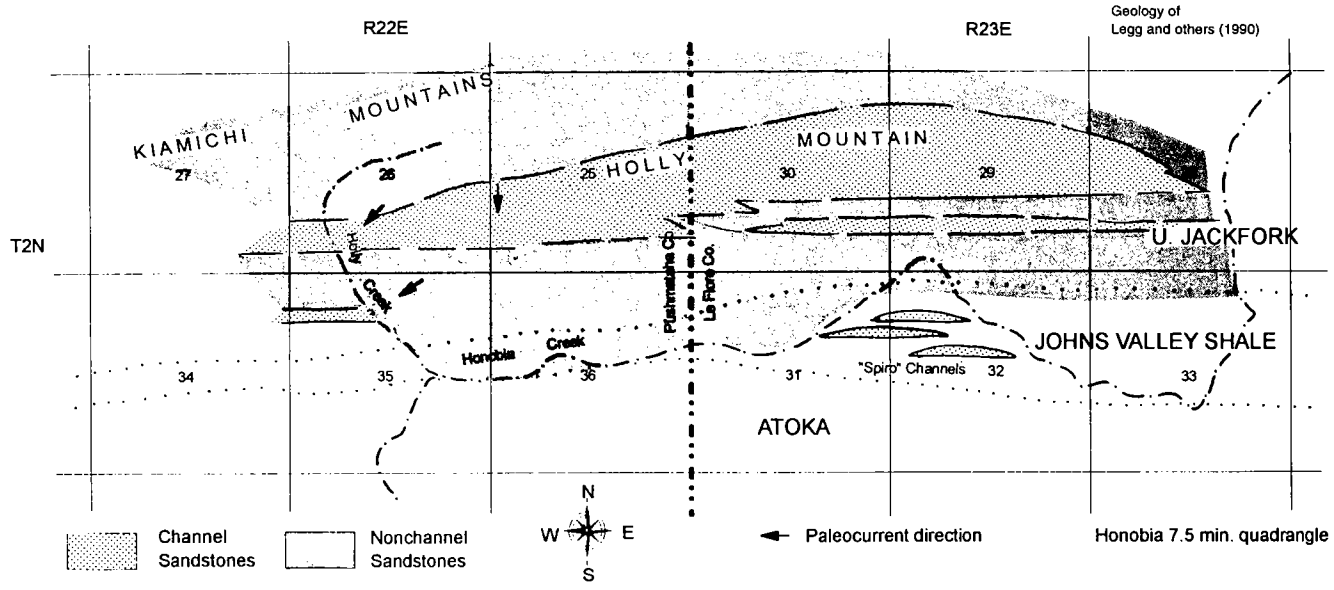


Figure 12. Map of Jackfork channel sandstones at the Holly Creek locality. Holly Mountain consists of a series of coalescing channel sandstones (stippled pattern) found within the upper Jackfork. The main channel complex illustrated is >4.5 mi wide. Smaller channel sandstones occur more locally in the upper Jackfork and in the overlying Johns Valley Shale. (Terminology of Legg and others, 1990.)

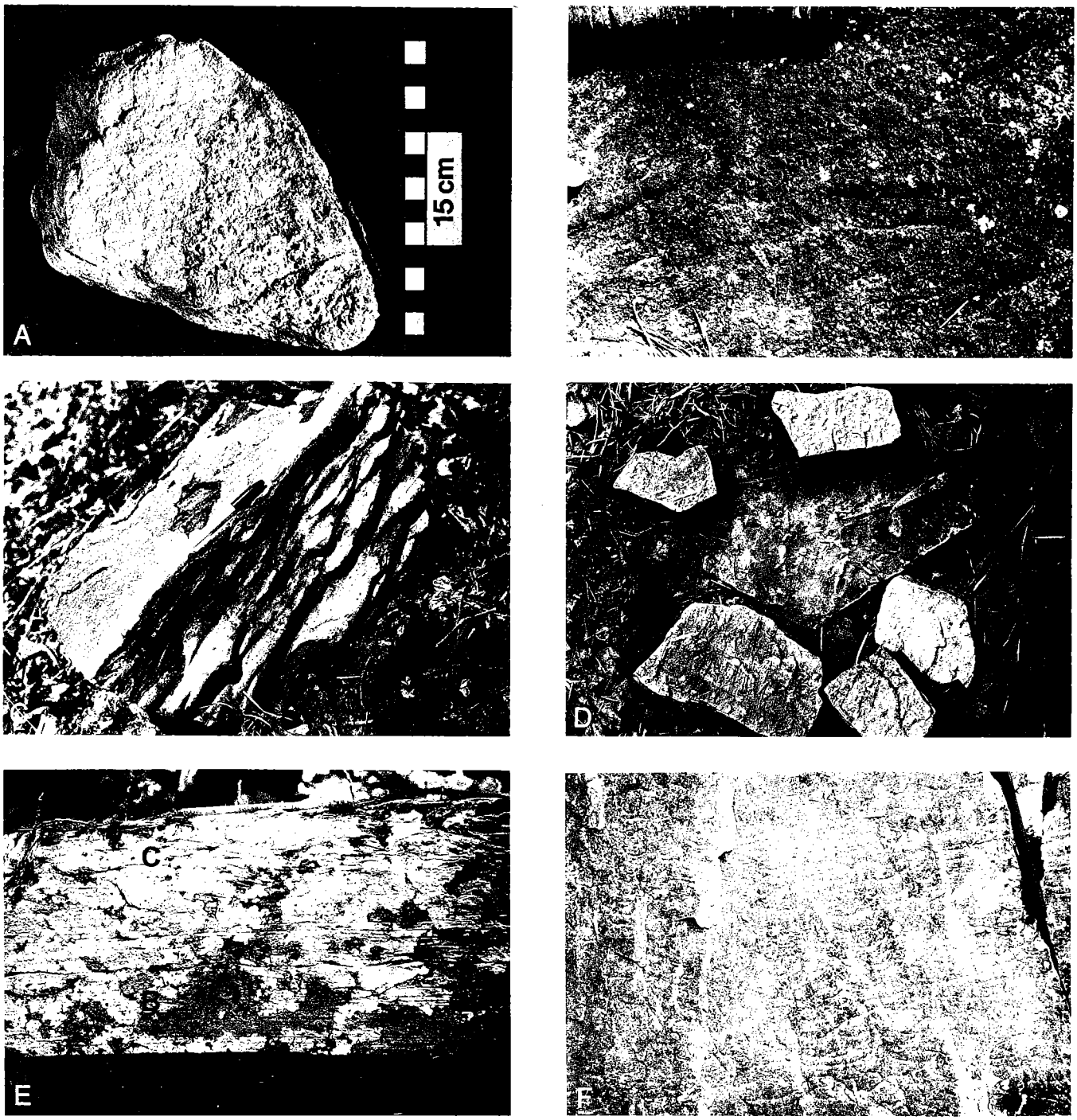


Figure 13. Sedimentary features associated with channel and nonchannel sandstones from the Holly Mountain channel complex. Features associated with channel sandstones (13A, B), shale-rich nonchannel turbidites (13C,D), and sand-rich nonchannel turbidites (13E,F) are illustrated. *A*—Medium- to coarse-grained sandstones in lower part of bed (right side of photo) containing quartz granules and molds of crinoid columnals interbedded with friable medium-grained sandstones in upper part of bed (left). *B*—Oriented plant impressions in float block near base of channel complex. Finger points to cast of *Calamites*. *C*—Fluted turbidite sandstone (i.e., orthoquartzite) from the upper Jackfork just above the top of the channel complex in W½ sec. 29, T. 2 N., R. 23 E. Thin- to medium-bedded turbidite sandstones and interbedded green shale comprise most of the upper Jackfork and Johns Valley Shale intervals above the Holly Mountain channel complex and along Honobia Creek. *D*—Trace fossils from the upper Jackfork immediately above the Holly Mountain channel complex. Both horizontal and vertical forms are present. *E*—Hard, sand-rich turbidites showing Bouma B and C units. *F*—Dish structures in hard, sand-rich turbidites below the Holly Mountain channel complex.

To date, the western limit of the Holly Mountain Jackfork channel complex has not been well defined. However, channel sandstones have been observed in the SE¼ sec. 27, T. 2 N., R. 22 E. From the field relations described above and a measured section in sec. 30, T. 2 N., R. 23 E., the Holly Mountain channel complex is at least 4.5 mi wide and 900 ft thick.

DEPOSITIONAL MODEL

Existing regional depositional models for the Jackfork Group call upon east-to-west transport of sediments via a system of submarine fans with proximal facies (i.e., upper-fan and upper middle-fan facies) mostly confined to Arkansas. These models have been developed from the recognition of proximal turbidites at several classic Jackfork localities in Arkansas (e.g., Moiola and Shanmugam, 1984; Jordan and others, 1991) and a preponderance of paleocurrent data indicating westward transport (Morris, 1974a,b).

Moiola and Shanmugam (1984) developed a depositional model for the Jackfork Group from detailed facies analysis of well-known exposures in Arkansas and Oklahoma. The depositional environment they envisioned for the Jackfork consisted of a series of westward-prograding submarine-fan complexes forming in an elongate deep-water basin. Upper-fan facies were deposited near Little Rock, Arkansas, middle-fan channel and interchannel facies farther south and east in Arkansas (e.g., DeGray Spillway), and lower fan-lobe deposits in southeastern Oklahoma (e.g., Kiamichi Mountain, U.S. 259). They believed that a similar east-to-west, proximal-to-distal model was equally applicable to describe the depositional setting of the Stanley, Johns Valley, and Atoka flysch sequences as well. More recently, Shanmugam and Moiola (1994) reinterpreted the sediments that now form the Jackfork Group as having originated predominantly by debris flows or slumps in a slope-dominated environment rather than a true submarine-fan setting. Their new interpretation was based on a detailed examination of internal bedding features.

Evoy (1989,1990) also favored a westward-prograding turbidite depositional model for the Jackfork Group. He constructed regional lithofacies maps that showed an overall decrease in depositional energy and sandstone content in the Jackfork from southwestern Arkansas into southeastern Oklahoma. Using the fan model of Mutti and Ricci Lucchi (1978), he placed the Jackfork of the Lynn Mountain syncline in a middle- to outer-fan depositional setting.

Coleman and others (1994) provided a depositional model for the Jackfork that differed little from earlier models. The Jackfork in central Arkansas, they stated, is dominated by massive channel-levee sandstone complexes that were deposited in an upper-fan environment of a deep-water submarine-fan complex. The thin- to medium-bedded sandstones and shales that dominate Jackfork outcrops in western Arkansas and Oklahoma were interpreted as a more distal submarine-fan facies. Their model, shown in Figure 14, supports a lower middle-fan to lower-fan environment for the Jackfork in eastern Oklahoma, including the Lynn Mountain syncline.

It is not clear whether the friable, medium-grained channel sandstones in the upper Jackfork of the Lynn Mountain syncline were recognized by the authors mentioned above. To date, the thickness and geometry of the channel sandstones in Oklahoma have not been well documented. Evoy (1989) acknowledged the problematic occurrence of quartz-pebble conglomerates at some of his distal-facies localities (i.e., Tuskahoma syncline). He attributed deposition of the conglomerates to local or rare flow conditions and suggested that they may have been derived from local submarine canyons that incised a northern or southern continental shelf. Moiola and Shanmugam (1984) based their depositional model for the Jackfork in southeastern Oklahoma on their facies interpretations for the dominantly nonchannel lower Jackfork turbidites along U.S. 259. Coleman and others' (1994) depositional model seems to stem mostly from earlier studies and their own interpretations of the proximal turbidite facies exposed at several classic Jackfork localities in Arkansas. In either case, it is apparent that the existing models developed for the Jackfork may be insufficient for explaining the depositional regime of the group as a whole.

Morris (1989) recognized coarse-grained, fossiliferous, upper Jackfork (Brushy Knob Formation) sandstones in Arkansas that overlie thick olistostromes. He interpreted these massive cliff-forming sandstones as transverse channel deposits that fed marginal submarine fans particularly along the northern margin of the Ouachita trough. These channels, he stated, had far less areal extent than the major axial channels that fed the Jackfork from the east. He did not describe comparable sandstones in the upper Jackfork in Oklahoma.

A new depositional model for the upper Jackfork is proposed in Figure 15. The model differs from the aforementioned Jackfork models in two ways: (1) it includes a northern sand source for the upper Jackfork in Oklahoma, and (2) it indicates that sedimentation occurred in a proximal, rather than distal, submarine-fan environment. The model shows upper Jackfork channels incising a normal-marine shelf along the northern margin of the Ouachita trough. Sand was carried southward and deposited in an upper middle-fan environment in the area of the Lynn Mountain syncline. Debris-flow deposits, derived from slope failures, were deposited contemporaneously with the channel sandstones in a variety of submarine-fan subenvironments.

Southerly directed paleocurrents have been recognized from the channel and channel-levee sandstones in the upper Jackfork interval from the Lynn Mountain syncline. A northern provenance is also indicated by the composition of the channel sandstones. Fossil molds are abundant in the channel sandstones and are sometimes well preserved (Fig. 9B). They were probably derived from a normal-marine shelf located to the north and were carried basinward within submarine channels. It is unlikely that these fossils could have survived the long-distance transport required in the depositional models discussed earlier that call upon an eastern source for the sandstones.

The presence of a coarse-grained fraction in the Lynn Mountain syncline channel sandstones also suggests deposition in a more proximal submarine-fan environ-

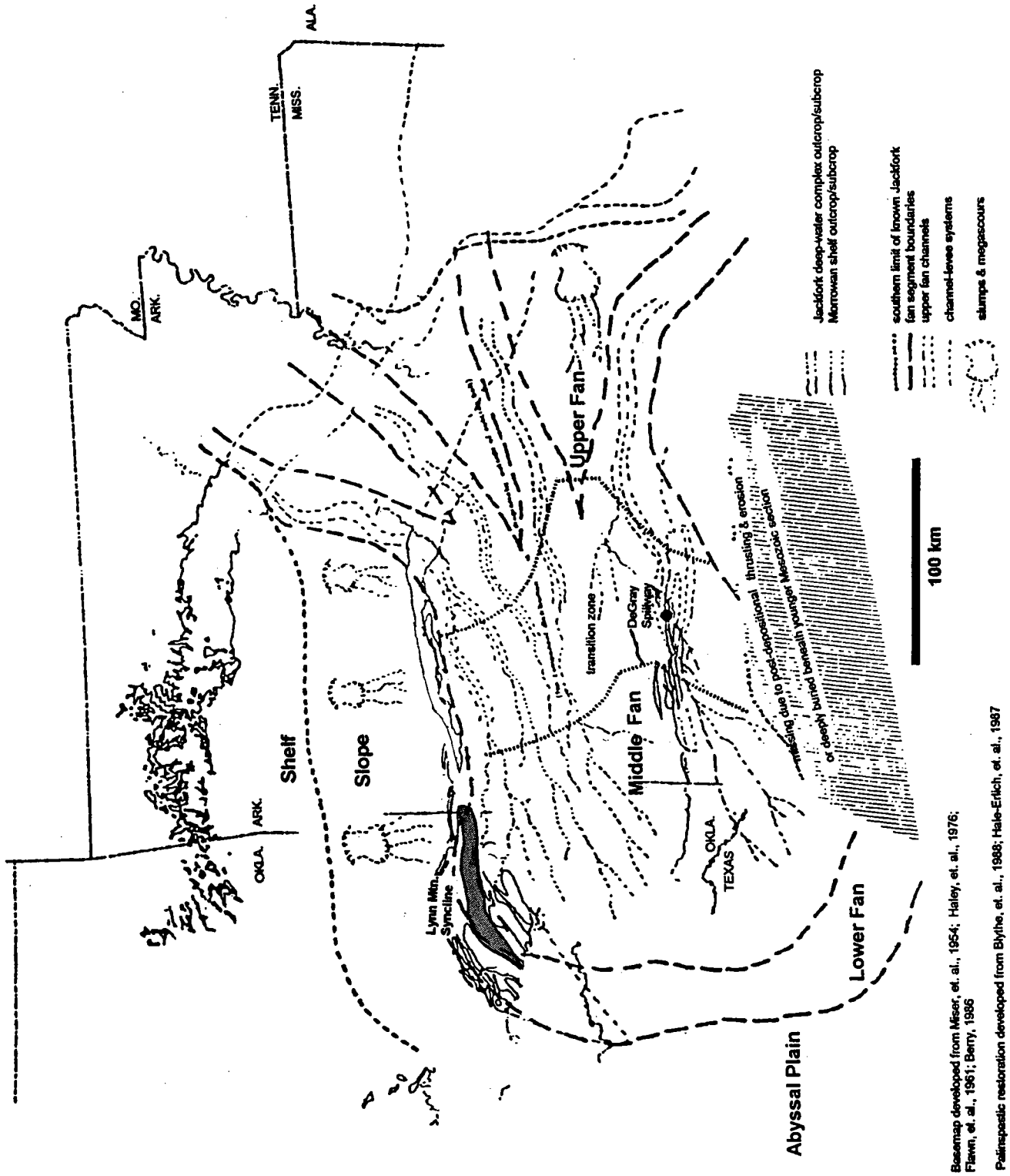


Figure 14. Jackfork Group depositional model from Coleman and others (1994). In this model, major submarine-fan channels are confined to the upper and middle fan in Arkansas. The Lynn Mountain syncline is located on the lower reaches of the fan in a lower middle-fan to lower-fan setting.

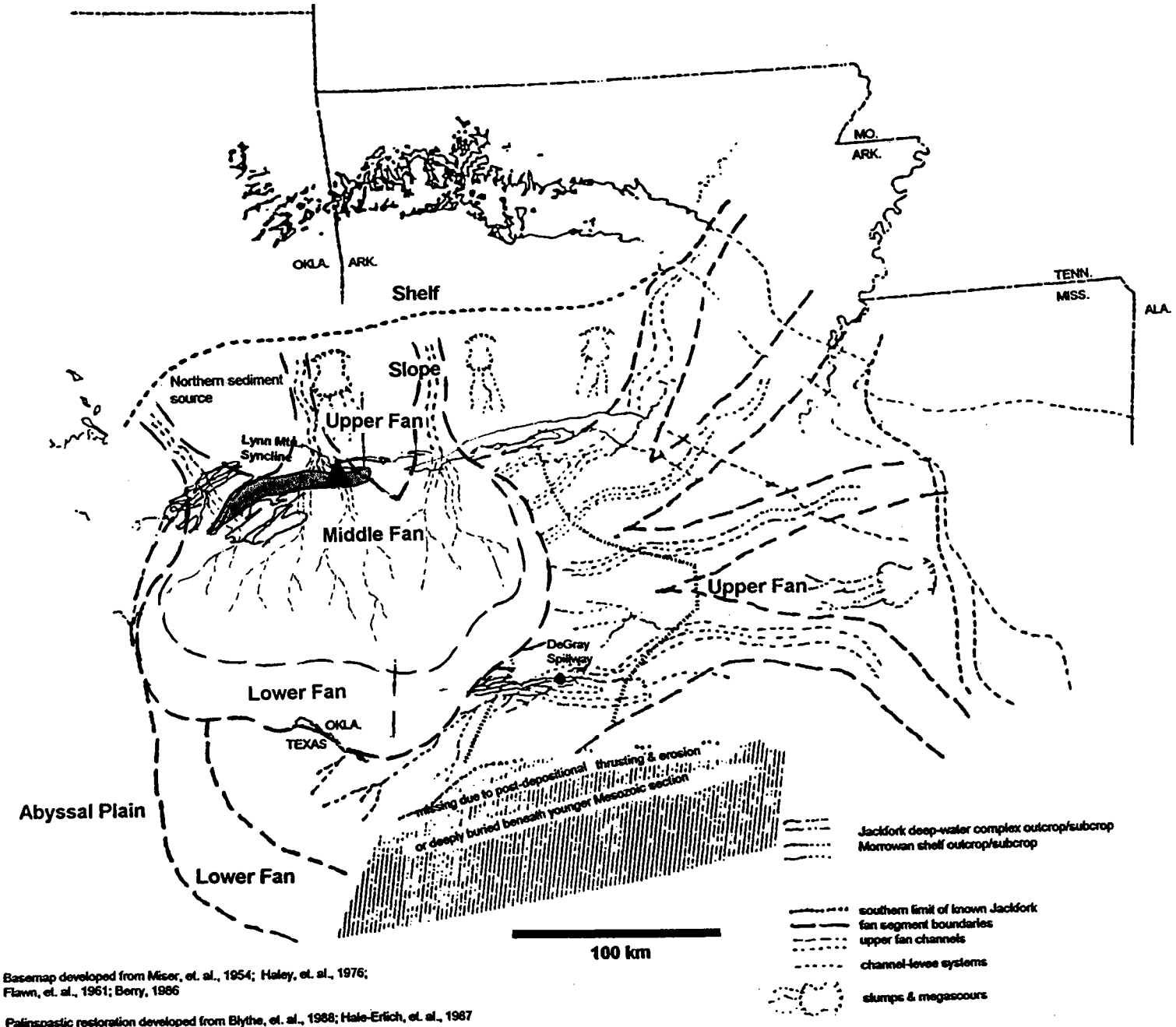


Figure 15. Generalized depositional model for the upper Jackfork Group formations. Unlike previous depositional models for the Jackfork, this model recognizes a significant northern sand source in Oklahoma and possibly Arkansas and the presence of proximal submarine-fan facies (i.e., middle to upper-middle fan) in the Lynn Mountain syncline. The triangle marks the location along U.S. Highway 259 (sec. 34, T. 2 N., R. 25 E.) where channels in the upper Jackfork are incised into lower-fan turbidites (Fig. 5B). The areal extent of the upper Jackfork fan in eastern Oklahoma and western Arkansas, as well as the fan-segment boundaries, is speculative in this model. (Modified from Coleman and others, 1994.)

ment than has been previously recognized. Quartz granules and pebbles are common in the channel sandstones and bear a resemblance to the "grit" facies of the Jackfork found in Arkansas (Fig. 16). This facies has been interpreted by various authors (Jordan and others, 1991, table 5; Coleman and others, 1994) as representative of an upper- to middle-fan submarine-fan environment. Coarse sand–sized and granule-sized sediment is typically deposited in upper- and middle-fan channels or the upper parts of middle-fan depositional lobes (Jordan and others, 1991). In the Mutti and Ricci Lucchi (1978) fan model, the coarsest sandstones (their facies A and B) are found in upper- and middle-fan channels.

The sedimentary features present in the upper Jackfork channel sandstones in the Lynn Mountain syncline are indicative of sediments deposited in a middle-fan, and possibly upper-fan, setting. Irregular and lenticular beds that lack sole marks characterize these sandstones. Scoured surfaces and the presence of abundant shale rip-up clasts in the channel sandstones were probably the product of highly erosive high-density turbidity currents. Debris flows containing matrix-supported clasts occur within the upper Jackfork channel complex exposed along U.S. 259 (Fig. 5A). Debris flows are common in slope environments. The combination of these sedimentary features favors the interpretation of a proximal submarine-fan channel setting for the friable channel sandstones found in the Lynn Mountain syncline.

Future studies should focus on the same channel-bearing upper Jackfork and Johns Valley stratigraphic interval in other synclines located farther south to determine whether channel sandstones exist there as well. Additional studies are needed to determine whether the proposed depositional model is equally applicable to other areas in Oklahoma. Of particular interest are the "Morrowan" fossil localities described from the Boktukola syncline (Honess, 1924; Shelburne, 1960).

SANDSTONE PETROGRAPHY AND RESERVOIR POTENTIAL

Petrographic examination of the upper Jackfork channel sandstones from the Lynn Mountain syncline indicates that they are predominantly quartz arenites and

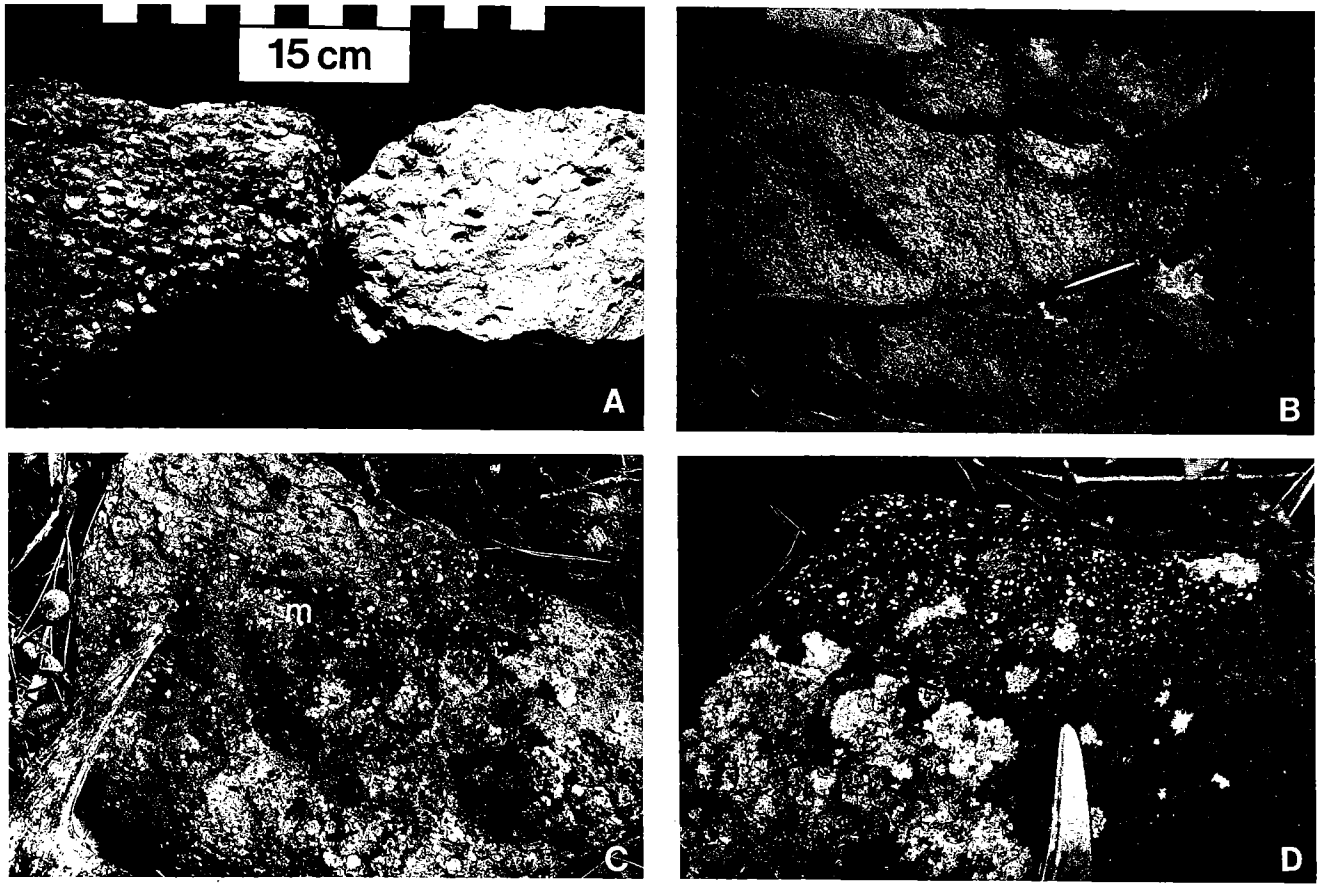


Figure 16. Comparison of Jackfork sandstones from proximal submarine-fan facies from the DeGray Spillway in Arkansas and from the Lynn Mountain syncline. *A*—Quartz-pebble conglomerate (left) and medium-grained sandstone containing "floating" quartz pebbles (right) from the upper part of the DeGray Spillway Jackfork stratigraphic section in sec. 13, T. 6 S., R. 20 W., Clark County, Arkansas. The sample on the left is representative of the Jackfork "grit" facies. *B*—Quartz-granule conglomerate infilling small scour at DeGray Spillway. Unlike the proximal submarine-fan sandstones in the Lynn Mountain syncline, the Arkansas sandstones are well cemented in outcrop. *C*—Channel sandstone from Lynn Mountain syncline containing white quartz granules and pebbles and molds (m) of shale rip-up clasts. *D*—Quartz-granule conglomerate infilling small scour in channel sandstone. Sandstones shown in 16C,D are from the Holly Mountain channel complex illustrated in Figure 12.

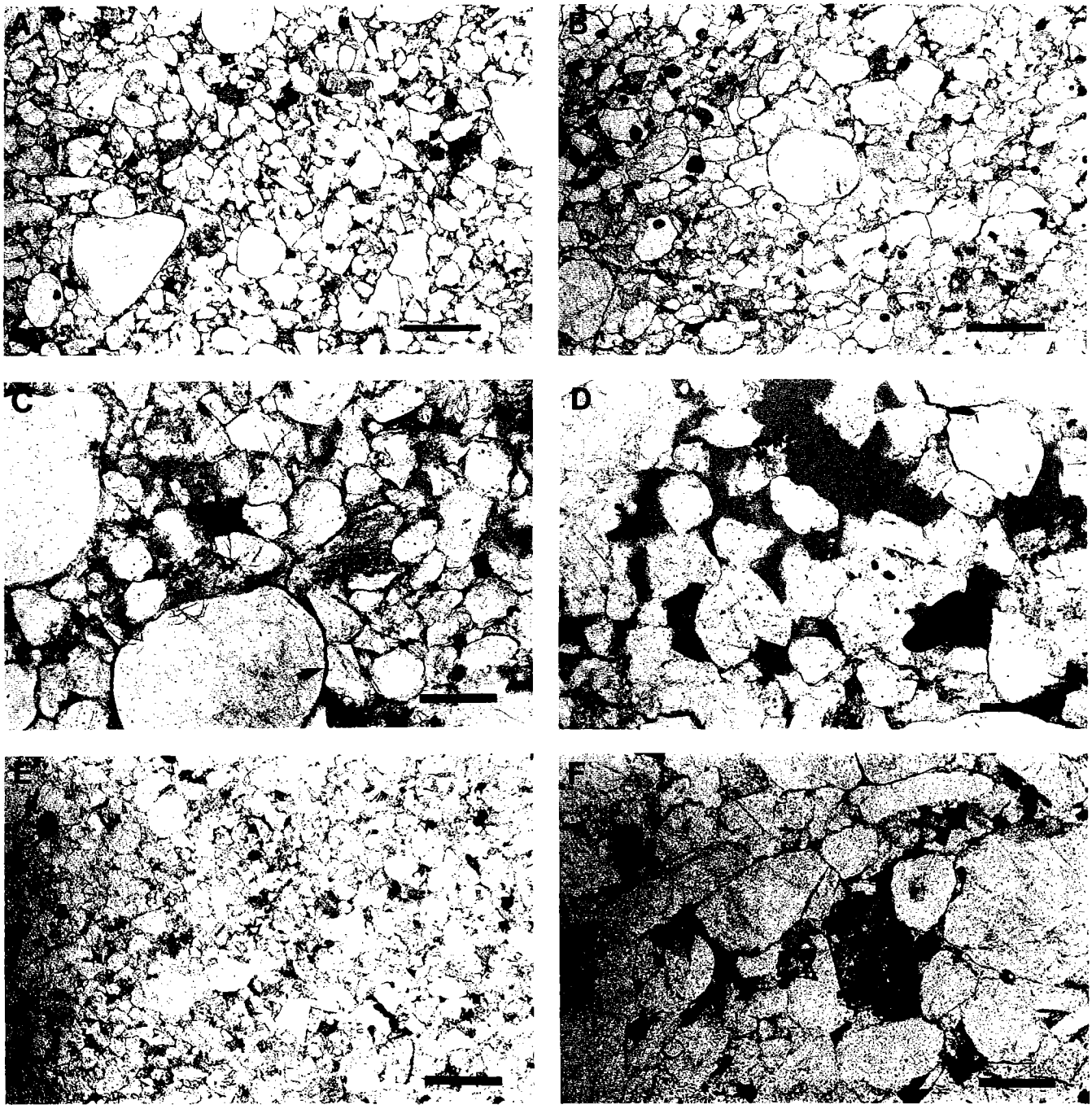


Figure 17. Photomicrographs of the Jackfork sandstone from the Lynn Mountain syncline. *A*—Poorly sorted, bimodal, fine- to coarse-grained quartz arenite with interparticle and moldic porosity. Detrital clay minerals (5%) have inhibited the development of quartz overgrowths. This friable channel sandstone was sampled at the Prairie Mountain locality in the S½NE¼ sec. 29, T. 1 N., R. 18 E. Bar represents 0.5 mm. *B*—Moderately sorted, medium-grained subarkose from channel sandstone in NE¼SW¼SE¼ sec. 27, T. 2 N., R. 25 E. Most of the porosity in this sandstone is secondary moldic and fracture porosity. Hairline fractures surround most of the grains, giving the sample an "exploded" appearance. Bar represents 0.5 mm. *C*—Same as 17A showing vermiform pore (arrow) along grain boundary, a textural feature that may indicate a precursor carbonate cement. Bar represents 0.2 mm. *D*—Oversized pore in channel sandstone. This type of porosity may have been formed by the dissolution of poikilotopic calcite cement. Bar represents 0.2 mm. *E*—Fine-grained quartz arenite from nonchannel turbidites located at Three Sticks Monument (SW¼NW¼SW¼ sec. 26, T. 2 N., R. 25 E). This sandstone is pervasively cemented by quartz (i.e., orthoquartzite) and is representative of many of the Jackfork sandstones exposed along U.S. Highway 259 north of Kiamichi Mountain. Grain boundaries are extensively sutured. Small isolated pores make up 1.3% of this sample and are probably due to dissolved feldspars. Bar represents 0.5 mm. *F*—Partially dissolved feldspar grain in fine- to medium-grained nonchannel sandstone from Holly Creek locality (SW¼NW¼SW¼ sec. 26, T. 2 N., R. 22 E.). Bar represents 0.2 mm.

subarkoses. Sublitharenites and quartz arenites occur in the interchannel and nonchannel sandstones. The percentage of feldspars in the sandstone samples may be slightly underestimated due to leaching.

The photomicrographs shown in Figure 17 illustrate the textural features that are typical of the various sandstone types. Channel sandstones are typically moderately to poorly sorted and medium to coarse grained (Fig. 17A,B). They commonly have bimodal grain-size distributions with the coarsest grains having the highest degree of rounding. Interparticle and intraparticle porosity is present, the later usually caused by the dissolution of feldspars or fossils.

The presence of detrital clay minerals in the channel sandstones has inhibited the formation of syntaxial quartz cement. Microcrystalline quartz is the dominant cement in the channel sandstones. The friable nature of the channel sandstones is, in part, due to incomplete cementation by this cement type. The friability also appears to be due to the presence of hairline fractures that follow grain boundaries (Fig. 17C). The fractures may have originated by pressure release associated with overburden removal and/or dissolution of a former carbonate cement. The vermiform nature of some of these fractures may have originated by etching of grains during the precipitation of carbonate cements (Fischer, 1993). Isolated porous "patches" having rhombic pores may also indicate precursor poikilotopic calcite cements (Fig. 17D).

Nonchannel sandstones are quartz arenites that are typically very fine to fine grained and moderately to well sorted (Fig. 17E). Sutured grain boundaries and stylolitization indicate that these sandstones have undergone extensive pressure solution. These sandstones are pervasively cemented by quartz (i.e., orthoquartzites), which accounts for their extremely hard character in outcrop. Porosity, when present, is usually moldic porosity related to feldspar dissolution (Fig. 17F) or the leaching of fossil debris. Mineralogically, the nonchannel sandstones are similar to the channel sandstones (Fischer, 1993).

Petrographic study of the Jackfork sandstones indicates that the sandstones having the most favorable reservoir characteristics are quartz arenites associated with submarine-fan channels (i.e., channel sandstones). Morris and others (1979) recognized the reservoir potential of the Jackfork channel sandstones in the upper Wildhorse Mountain Formation at the U.S. 259 locality. They estimated an effective porosity for these sandstones of 13.6% and a permeability of 7.28 millidarcies (md). Their study, which indicated that proximal turbidites contained sandstones with the best reservoir properties, established a relation between turbidite facies and reservoir potential.

Jackfork channel sandstones collected by Weyerhaeuser from the Lynn Mountain syncline have average porosities of 8–11.0% with values as high as 15.3% (Fischer, 1993; based on point counts, $n = 300$). These porosity estimates do not include porosity contributed by hairline fractures surrounding grains. The contribution of this pore type to the reservoir quality of the Jackfork, if it is preserved in the subsurface, would be significant because it would greatly enhance permeability. Secondary porosity, usually caused by the leaching of feldspars, is common in the Lynn Mountain sandstone samples. It is present in both the channel and nonchannel facies. Some interchannel (i.e., levee) sandstones also have good porosities; however, their permeability may be reduced by a higher percentage of detrital clay.

The upper Jackfork channel sandstones from the Lynn Mountain syncline are similar to porous Jackfork sandstones from wells drilled in Latimer County, Oklahoma. Photomicrographs of porous lower Jackfork sandstones taken from the H&H Star No. 1-4 Hope (sec. 4, T. 3 N., R. 20 E.) and ARCO No. 1-12 Borne (sec. 12, T. 4 N., R. 20 E.) wells are shown in Figure 18. The sandstones are

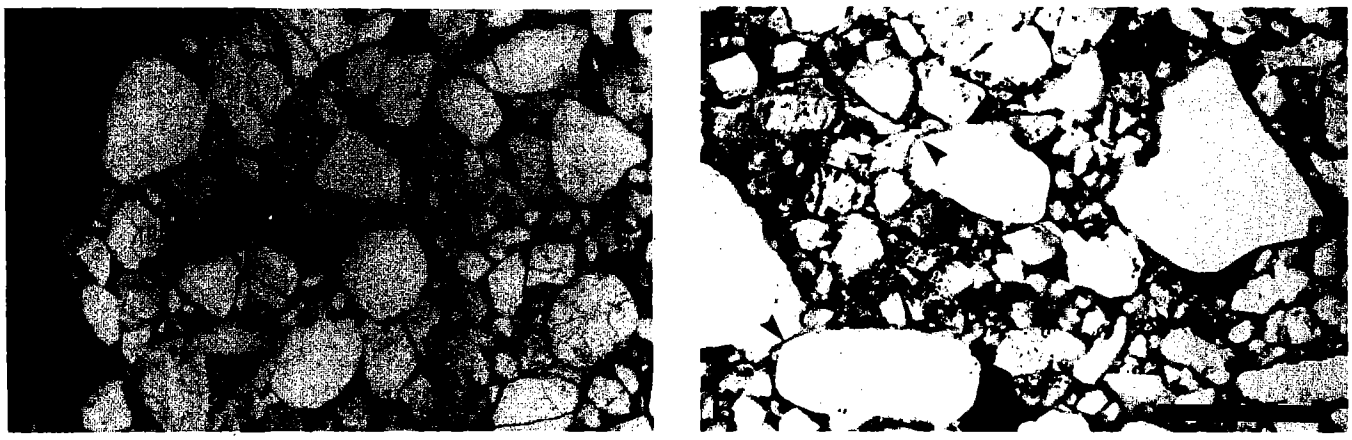


Figure 18. Photomicrographs of the Jackfork sandstone sampled from two Latimer County wells. *A*—Producing Jackfork interval in the H&H Star No. 1-4 Hope well (sec. 4, T. 3 N., R. 20 E.) at 18,800 ft. Sandstone has bimodal texture with coarser grains having a higher degree of rounding. Large pore (arrow) was probably produced by dissolution of a feldspar grain. Log porosities from the perforated interval in this well are 8–10%. Magnification 100×. *B*—Porous zone from 8,500 ft in the ARCO No. 1-12 Borne well (sec. 12, T. 4 N., R. 20 E.). Sandstone is a bimodal, poorly sorted, fine- to coarse-grained feldspathic arenite with interparticle and secondary dissolution porosity. Some of the porosity in this sample appears to be fracture porosity surrounding grains (arrows). Permeability may be reduced by pore-lining clay minerals. Porosity = 9.9%, permeability = 0.05 md. Bar represents 0.25 mm.

poorly sorted, medium-grained bimodal quartz arenites and feldspathic arenites with intergranular and secondary moldic porosity (caused by feldspar dissolution). Larger grains exhibit a higher degree of rounding. Some of the porosity in the sample from the No.1-12 Borne well is fracture porosity surrounding grains (Fig. 18B) and may be analogous to a common pore type observed in the Jackfork channel sandstones from the Lynn Mountain syncline.

JACKFORK GAS PRODUCTION— CURRENT AND FUTURE POTENTIAL

Jackfork Group sandstones currently produce natural gas from three wells in Latimer County, Oklahoma (Fig. 19). Productive intervals are located in both the "upper" and "lower" Jackfork, as designated in Figure 20. Initial potentials for these wells range from 1.9 to 5.7 MMcfgpd.

Porous (i.e., porosity ≥10%) Jackfork sandstones have also been encountered in other Latimer County wells drilled in T. 4 N., R. 20 E. Some of these zones have tested noncommercial gas rates. Figure 21 shows the porosity logs from two of these wells. The H&H Star No. 1-16 Coopers Hollow well (sec. 16, T. 4 N., R. 20 E.) drilled >77 ft of porous "lower" Jackfork. This well is currently producing from the Spiro sandstone, and the porous Jackfork interval remains untested behind pipe. ARCO tested noncommercial gas rates from a porous interval in the "lower" Jackfork in their No. 1-12 Borne well.

The Atoka through Jackfork stratigraphic sequence in the producing wells from Latimer County is remarkably similar to the stratigraphic sequence penetrated in the SOHIO No. 1-7 Weyerhaeuser well that was drilled on the north flank of the Lynn Mountain syncline. A correlation from the No. 1-7 Weyerhaeuser well to the H&H Star No. 1-4 Hope and No. 1-28 Lady Luck (sec. 28, T. 4 N., R. 21 E.) wells is shown in Figure 20. The basal Atoka and upper Jackfork contacts are very sharp in all three wells. Also, the Johns Valley Shale contains isolated channel sandstones and is similar in thickness in each of the wells. The correlation indicates that the "lower" Jackfork stratigraphic interval containing the productive sandstones in the Latimer County wells was not penetrated in the No. 1-7 Weyerhaeuser well.

Current Jackfork production in Latimer County, Oklahoma, is from thrust sheets carried in the hanging wall of the Ti Valley fault. Production occurs on the south-dipping limb of large duplex structures, and one of the wells

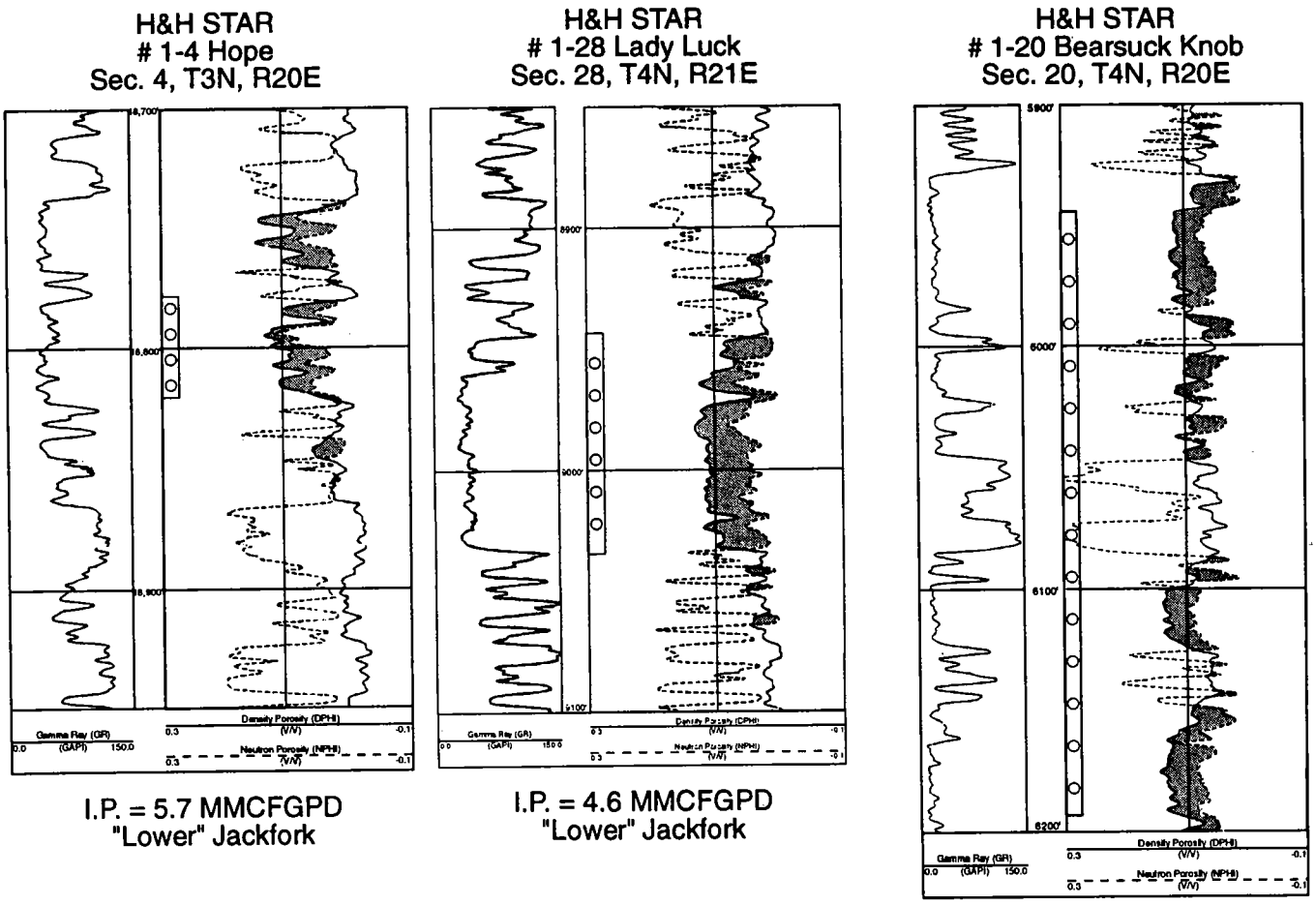


Figure 19. Porosity logs for three wells currently producing from the Jackfork in Latimer County, Oklahoma. Gross perforations are shown in depth track.

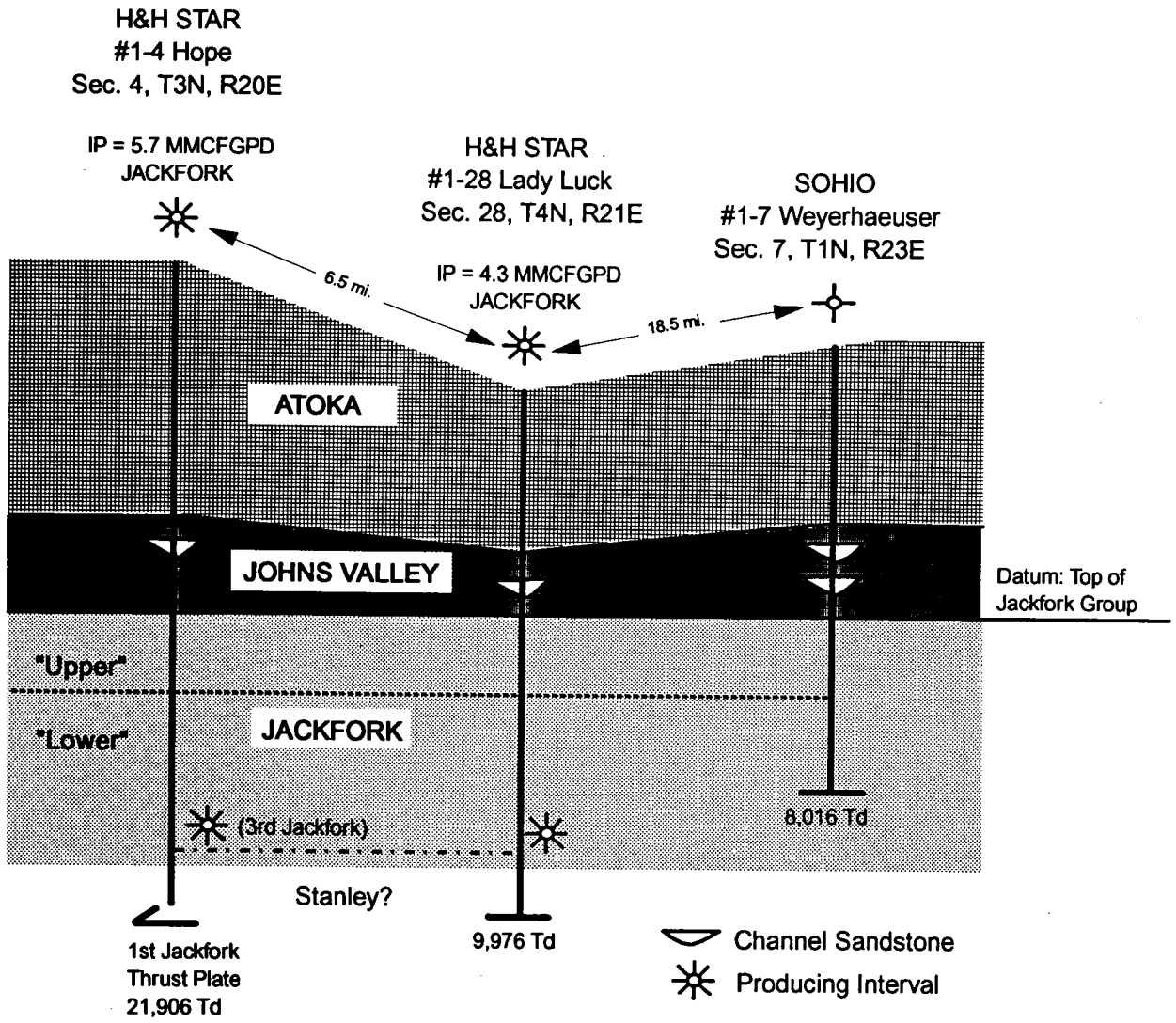


Figure 20. Correlation of the Atoka, Johns Valley, and Jackfork intervals penetrated in two Latimer County wells to the SOHIO No. 1-7 Weyerhaeuser well that was drilled on the north flank of the Lynn Mountain syncline. Similarities between the wells include (1) sharp contacts at the base of the Atoka and top of the Jackfork Group, and (2) isolated channel sandstones in the Johns Valley Shale with similar thicknesses in each well. The producing intervals in the Latimer County wells are shown by gas symbols; however, production in the No. 1-4 Hope well is from the third Jackfork thrust plate in approximately the same stratigraphic interval shown for the first Jackfork thrust plate in the figure. The correlation indicates that the producing interval in the "lower" Jackfork was not penetrated in the No. 1-7 Weyerhaeuser well.

(H&H Star No. 1-20 Bearsuck Knob, sec. 20, T. 4 N., R. 20 E.) produces from the "upper" Jackfork only 3,242 ft downdip from the Jackfork outcrop. At least two of the wells currently producing from the Jackfork were targeted to deeper horizons and may not have encountered the Jackfork reservoir sandstones in a favorable structural position. It seems likely that a stratigraphic component is an integral part of trapping natural gas in the Jackfork on the basis of these wells that are producing downdip from fold axes. Seismic data suggests that the trapping mechanism in these Jackfork reservoirs is the termination of porous zones against thrust faults.

Future Jackfork production may be found in anticlines associated with Jackfork duplexes, stratigraphic traps on the limbs of these folds, or in combination traps. In the Lynn Mountain syncline, all of these trap types may occur. Anticlinal folds occur in the central part of the syncline along the Hardy Creek anticline. Stratigraphic traps may occur where channels cut across structural dip on the north and south limb of the syncline. Combination traps may be encountered along the south limb of the syncline where it is truncated by the Octavia thrust fault. This relation may juxtapose the Stanley, a potential source rock, against Jackfork channel reservoirs. The risks and rewards associated with each of these trap types in the Jackfork will only be established through future drilling.

CONCLUSIONS

Friable channel sandstones occur in the upper Jackfork Group in the Lynn Mountain syncline. Sedimentary

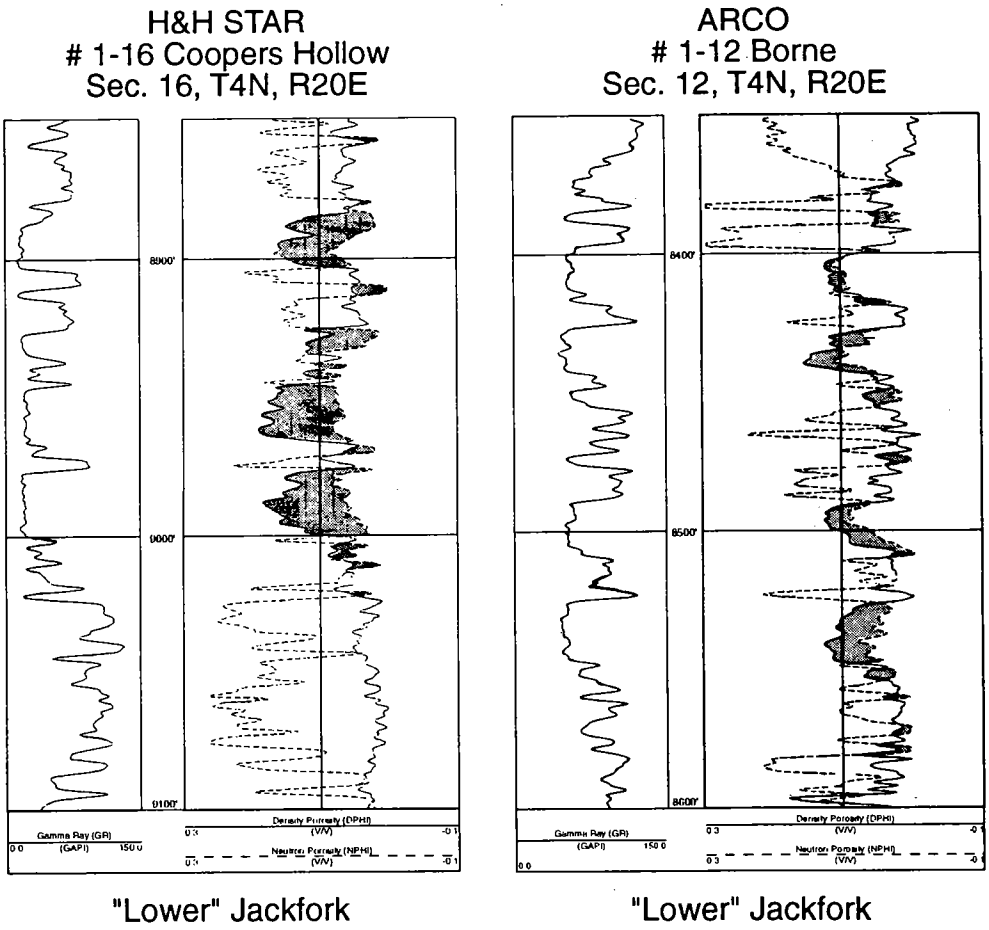


Figure 21. Porosity logs from the Jackfork interval in the H&H Star No. 1-16 Coopers Hollow and ARCO No. 1-12 Borne wells. Log porosities ≥10% are present in both wells. The No. 1-16 Coopers Hollow well penetrated 77 ft of ≥10% porosity in the Jackfork but is currently producing gas from the Spiro sandstone. The Jackfork has not been tested in this well. The No. 1-12 Borne tested the Jackfork but was also completed as a Spiro producer. A photomicrograph from the porous interval at 8,500 ft in the No. 1-12 Borne well is shown in Figure 18B.

features present in the channel sandstones indicate that they had a northerly source and were deposited in an upper middle-fan to upper-fan environment on a submarine fan. Associated with the channel sandstones are nonchannel turbidites that have many of the sedimentary features typically associated with the classic Jackfork flysch. Petrographic examination of the channel sandstones indicates that they could act as viable gas reservoirs. They are texturally similar to productive Jackfork reservoirs 15 mi to the north in Latimer County, Oklahoma.

ACKNOWLEDGMENTS

I would like to thank Russ Cunningham of GHK and Vastar Resources for allowing me to use the photomicrographs from the H&H Star No. 1-4 Hope and ARCO No. 1-12 Borne wells, respectively. I would also like to thank my reviewers for their helpful thoughts and suggestions.

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Sedimentology and Sequence Stratigraphy of Jackfork Group, U.S. Highway 259, Le Flore County, Oklahoma

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ABSTRACT.—The Morrowan Jackfork Group, where it crops out in the Ouachitas, has long been regarded as a deep-water, primarily turbidite, deposit. Recent work in southeastern Oklahoma indicates that at least two or more periods of significant shallowing occurred during the deposition of the Jackfork. I have revisited and redescribed outcrops along U.S. Highway 259, originally described by Lewis Cline and Frank Moretti in 1956, and have interpreted these outcrops to include shelf, slope, and deltaic deposits. These relatively shallow water deposits occur in the Markham Mill and Prairie Mountain Formations of the Jackfork Group.

Several significant sequence-stratigraphic surfaces occur in the Jackfork Group. Relatively abrupt shallowing is evident at what are interpreted to be sequence boundaries at the base of the Prairie Mountain Formation and at the base of the shelf and deltaic deposits that form the lower part of the Markham Mill Formation. Significant flooding surfaces are observed between the slope and overlying shelf deposits in the upper Prairie Mountain Formation and either in the upper part of the Markham Mill Formation or at the base of the Wesley Formation. A third flooding surface may occur at or near the base of the Game Refuge Formation.

Slope, shelf, and deltaic deposits in the Jackfork Group are characterized by features distinctly different from those of the turbidites in the Wildhorse Mountain Formation, which forms much of the Jackfork along U.S. 259.

The slope deposits are characterized by the presence of locally abundant, rounded and transported, primarily sandstone blocks supported in a fine-grained matrix of mudstone and shale. The clasts range in diameter from several centimeters to several meters and occur at numerous positions within the upper-middle part of the Jackfork. Shelf deposits in the Prairie Mountain Formation are characterized by thin, generally undeformed, subhorizontally laminated 1- to 3-ft-thick sandstones interbedded with shales. The upper surfaces of some of the thinner shelf sandstones have burrows including *Zoophycos*.

Deltaic deposits are primarily thinly bedded, current-rippled, and subhorizontally laminated lenticular splay sandstones interbedded with shales. In the deltaic deposits, local channeling and thicker sandstone beds occur, as does localized soft-sediment deformation.

INTRODUCTION

The outcrops along U.S. 259, south of Big Cedar, Oklahoma, were the subject of extensive work by Lewis Cline and his University of Wisconsin students in the mid-1950s. One of the key measured sections that grew out of their work was published by Cline and Moretti in 1956. Their section extends from below the base of the Jackfork Group (Jackfork Group of this paper) to what they defined as the Atoka Formation at the top. Their "Atoka top" was included by Briggs (1973) in the Game Refuge Formation (Game Refuge Formation of this paper) of the Jackfork Group. This paper focuses on the Jackfork Group outcrops along U.S. 259, one of the best exposed and most complete Jackfork sections in eastern Oklahoma.

At the time Cline and Moretti measured the section in this area, U.S. 259 had just been finished, and there were numerous well-exposed outcrops along the right-of-way. Since their work was completed, many of the outcrops have deteriorated, and a few new ones have been excavated. During the course of my work, I visited most of the currently well-exposed outcrops along the highway and, for the most part, was able to consistently locate myself on Cline and Moretti's measured section. Suggested revisions and additions to their section on a unit by unit basis are included in the slightly revised version of their measured section (Table 1). A more detailed description of part of their measured section is the subject of this paper and should be referred to often while reading this paper (see Fig. 5).

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TABLE 1. — U.S. HIGHWAY 259 MEASURED SECTION (UNITS 1–63)*

	Thickness (feet)
Atoka Formation — *Note: What is designated here as Atoka is actually Game Refuge Formation, the uppermost formation in the Jackfork Group according to Briggs (1973, p. 29).*	
63. Poorly exposed sandy shale with rectangular blocks of hard quartzitic sandstone. Sandstone float containing a sandstone cast fauna was found in this interval. The fauna may be Honess' "Morrow fauna." *(Crinoids.)*	
62. Poorly exposed blue-gray shale	90
61. Shale and thin-bedded sandstone. Shale predominates near the base but at the top the well-bedded, thin, white sandstone may comprise 40% of the rock	245
60. Sandstone and sandy shale; hard, white, quartzitic sandstone comprises 60% of the interval	165
JACKFORK GROUP	
Wesley Formation	
59. Shale; weathers light blue-gray; contains at least one 5-inch bed of black green-gray weathering chert	150
Markham Mill Formation	
58. Sandstone; soft, green-gray, a small amount of interbedded lignitic sandy shale near top	38
57. Shale; sandy; exposed at sharp curve in road	27
Prairie Mountain Formation — *Top of formation occurs near the base of Unit 56. Top of formation marked by sequence boundary.*	
56. Sandstone; some very hard white massive rock but toward the top becomes green and friable	148
55. Light blue-gray sandy shale with rounded boulder-like masses of brown-gray quartzitic sandstone *(boulder shale)*	105
54. Poorly exposed, massive, yellow-weathering sandstone	90
53. Shale and sandstone	195
52. Massive to medium bedded, fine-grained, firm white sandstone; weathers yellow; thickness of interval hard to calculate *(includes flooding surface)*	230
51. Poorly exposed light blue-gray sandy shale with some intercalated hard, thinly bedded, quartzitic sandstone. *(Measured Section 6A-93 starts about 450 feet above the base of this unit.)*	650
50. Light gray yellow-weathering sandstone and sandy shale in about equal proportions	278
49. Shale; soft, slightly sandy at base; weathers light gray	52
48. Massive yellow sandstone	20
47. Covered; some poorly exposed yellow-weathering sandstone	50
46. Poorly exposed sandy blue-gray shale and soft sandstone; exposed in the saddle where the new highway intersects the old road to Big Cedar. *(Now at this location there is only a short north-south road linking Kiamichi Mountain crestal road and U.S. Highway 259.)*	327
Wildhorse Mountain Formation, *Upper Part*	
45. Sandstone; soft, *friable* medium-grained, *(200 μm)*, light gray, weathers yellow. *(Sharp lower contact with 1-foot erosional relief. This unit probably belongs in Prairie Mountain Formation.)*	40
Sequence Boundary	
44. Shale with subordinate sandstone; dark blue-gray laminated shale with interlaminated gray siltstone and sandstone with evenly bedded white quartzitic sandstone; sandstone comprising 35% of upper portion of interval, relatively unimportant in lower portion	120
43. Sandstone and shale; 80% of interval comprised of fine-grained, hard, well bedded, quartzitic sandstone in beds up to 1½ feet thick; dark gray sandy shale intercalated, becoming important toward the base	55
42. Shale and sandstone; soft white sandstone at base grading upward into blue-gray laminated silty shale. *("Ouachita National Forest" sign near base of unit.)*	35
41. Covered; poor exposures indicate sandy shale	310
40. Sandstone; soft, yellow-weathering; thickness hard to calculate because road parallels strike	50
39. Sandstone; hard, white to light gray, well bedded in even beds two feet thick	75
38. Lower one-half of interval covered; upper one-half appears to be badly weathered gray sandy shale with interbedded fine-grained quartzitic sandstone	155
Big Stick Monument (at roadside, 5.4 miles south of Big Cedar)	
37. Sandstone; hard, white to pink, fine-grained, quartzitic sandstone; forms crest of ridge where highway cuts through	170
36. Shale and sandstone; dark blue-gray sandy shale and hard to firm gray sandstone; sandstone-shale ratio 60:40	135
35. Sandstone and shale; sandstone comprises 65–70% of interval; remainder composed of dark blue-black shale with alternating laminae of gray siltstone and some sandstone beds up to a foot in thickness	117
34. Shale; black with interlaminated gray siltstone (35%); one bed of hard, quartzitic sandstone	9.5
33. Sandstone; predominantly hard, medium blue-gray, subquartzitic sandstone in beds averaging from 2 to 6 feet; rarely, a well bedded silty shale intervenes *(includes thickening-upward sandstone)*; includes a fault with small displacement down to west	265
32. Shale and sandstone; dark blue-gray shale with interlaminated 1 to 2 inch beds of hard quartzitic gray sandstone occurring in 30% of interval; total sandstone content estimated at 45% of interval	52
31. Predominantly dark gray laminated shale interbedded with thin 1 to 2 inch beds of gray sandstone; occasional 2 to 5 foot beds of soft, dirty, poorly sorted, yellow-weathering sandstone; grades into zone 32 above where the sandstone merely becomes more quartzitic and better jointed; about 60% shale and 40% sandstone	105

TABLE 1 *(continued)*. — U.S. HIGHWAY 259 MEASURED SECTION (UNITS 1–63)

	Thickness (feet)
PRAIRIE HOLLOW MAROON SHALE MEMBER	
30. Shale and sandstone; mostly *pinkish to* gray shale and green-gray, silty, massive sandstone that weathers yellow; some brownish-red shale and silty shale at intervals	130
Wildhorse Mountain Formation, Lower Part	
29. Sandstone; massive below becoming bedded above and containing a small amount of shale at top; hard and quartzitic below, becoming poorly sorted and brown-gray above	30
28. Shale; dark blue-gray laminated shale with some silty and sandy beds included	39
27. Sandstone; fine-grained, blue-gray, hard, quartzitic, weathers brown; well defined beds about 2 feet thick	50
26. Shale; dark blue-gray to black, blocky clay-shale; some rounded hard quartzitic masses near the top that may be concretionary *"glupes" and "globs," a nontectonic, sedimentary, slurried unit. Above deformed unit is a generally thinning-upward unit overlain by shale.*	52
25. Sandstone; gray, weathers brown-gray	73
24. Dark laminated shale and interlaminated hard, white, quartzitic sandstone in about equal amounts	135
23. Sandstone; soft, white, weathers yellowish; beds up to 2 and 3 feet thick	45
22. Shale and sandstone; dark blue-gray laminated shale interbedded with white, well bedded, quartzitic sandstone occurring as thin laminae in the shale; sand-shale ratio 4:6	173
21. Sandstone; soft, dirty, massive, weathers yellowish	40
20. Shale; blue-gray with beds of dirty gray sandstone	22
19. Covered	20
18. Interval 65% sandstone, 35% sandy shale; silty, poorly sorted, yellow-weathering sandstone; blue gray shale with interlaminated light gray siltstone or fine-grained sandstone; center of cut faulted; displacement unknown but not great	115
17. Predominantly laminated blue-gray shale with laminae of ash-gray siltstone; one 8-foot bed of yellow-weathering friable sandstone one-third the distance above the base	60
16. Covered; estimated	50
15. Sandstone; soft, friable, gray, weathers yellowish, in beds averaging about a foot in thickness; 95% sandstone in lower part of interval decreasing to 80% above	40
14. Shale; dark gray, well-bedded shale with some light gray, hard, quartzitic sandstone in beds up to 6 inches; soft, friable, yellow sandstone begins to be prominent toward top	100
13. Sandstone and sandy shale; firm to friable yellow-weathering gray sandstone and separating sandy gray shales (30%)	80
12. Covered	125
11. Sandstone with perhaps 30% purple-gray to brown-gray, sandy, lignitic shale and some laminated dark blue-gray shale; firm to friable, fine to medium-grained, ash-gray, yellow-weathering sandstone in beds averaging 1½ feet thick; many beds have ripple-marked upper surfaces	140
10. Shale and sandstone; predominantly dark blue-gray, laminated shale with thin even-bedded sandstone layers; at top of zone about 5 feet of white, hard, ripple-marked sandstone	25
9. Alternating sandstone and shale in subequal proportions (ratio 6:4); sandstone is light gray, medium to fine-grained, poorly sorted, in well developed beds averaging less than 2½ feet and with some poorly developed ripple marks; dark gray to brown shale, the brown layers being sandy, lignitic, and containing many plant remains	135
8. Covered	35
7. About 50% of interval medium-gray, poorly sorted sandstone in beds up to 1½ feet, the units averaging 20 feet; separated by subequal amounts of dark blue-gray sandy shale; shale laminated with 1 to 3 inch beds of brown-gray lignitic sandstone	95
6. Shale and sandstone; blue-gray shales with thin laminae of silty, poorly sorted, fine-grained sandstone with intercalated medium-grained, brown-weathering sandstone; sandstone-shale ratio 6:4	37
STANLEY SHALE GROUP	
Chickasaw Creek Siliceous Shale Formation	
5. Shale; soft, dark gray to black when fresh, weathers light blue-gray; a few 1 to 4 inch beds of white siltstone and brown-gray poorly sorted sandstone *(Briggs moved unit 5 into Jackfork Group, 1973, p. 39; location is NE¼NE¼ sec. 26, R. 25 E., T. 1 N.; 2.35 miles south of Big Cedar, intersection of Highways 254 and 63.)*	205
4. Siliceous shale, black laminated siliceous shale containing several thin beds of blue-black chert mottled with white almond-shaped siliceous areas	15
3. Shale; poorly exposed, dark gray to black, laminated, weathers light blue-gray; includes some siliceous shale and some minor amounts of fine-grained ash-gray siltstone and sandstone	55
2. Siliceous shale; dark gray to black laminated siliceous shale with thin 1 to 3 inch hard, ash-gray siltstones comprising about 20% of the interval; dark blue–black chert in 1 to 3 inch bands, speckled with white siliceous areas	15
1. Dark blue gray to black laminated shale predominating (60–65%); fine-grained ash-gray sandstone in even beds up to 3 feet in thickness second in importance; sandy shale, 5%	55

*Originally measured by Cline and Moretti, 1956; modified, refined, and interpreted by R. W. Tillman in 1993 and 1994. Italicized parts are changes and additions made to section by Tillman.

APPROACHES TO DEFINITION OF JACKFORK STRATIGRAPHY

The area of this study is in the interval between the Windingstair fault and the Octavia fault in southeastern Oklahoma, near the Arkansas state line. A number of workers have mapped and interpreted the stratigraphy of parts of this and adjoining thrust-fault blocks (Fig. 1). Garrett Briggs has done the most detailed work to date in the area of the Cline and Moretti section on U.S. 259.

The stratigraphy of the Jackfork Group in southeastern Oklahoma has been in flux almost from the first day a geologist began to study the unit. The name "Jackfork Formation" was first used by Taff (1902) to designate a thick sequence of sandstones and shales he encountered on Jackfork Mountain in Atoka County, Oklahoma.

Harlton (1938) elevated the Jackfork Formation to the Jackfork Group and subdivided it into four formations on the basis of laterally persistent shales present in the Round Prairie syncline east of the town of Atoka. The Jackfork Group was further redefined 35 years later by Briggs (1973) to include the following formations, from bottom to top: Chickasaw Creek, Wildhorse Mountain, Prairie Mountain, Markham Mill, Wesley, and Game Refuge (Fig. 2). Briggs did not recognize all six formations in his mapping of the area in Le Flore County, Oklahoma (Fig. 3). He grouped the Prairie Mountain, Markham Mill, and Wesley into one mappable unit. I believe that subdivision of these three formations is possible for purposes of mapping and to better understand the sequence stratigraphy of the Jackfork. I follow Cline and Moretti's (1956) units and Briggs's (1973) reassignment of ages in the following discussion of the geology of the study area. The reader is referred to Briggs (1973) for a detailed discussion of the stratigraphy and regional aspects of the formations in the Jackfork Group in southern Le Flore County (Fig. 1).

A generalized north-south cross section through the Mississippian and Pennsylvanian strata (including the Jackfork Group) in the area of Cline and Moretti's (1956) measured section shows the relation between the stratigraphic units and the thrust faulting (Fig. 4). In an effort to define the environments of deposition and sequence stratigraphy of the Jackfork Group, I measured a 1,260-ft section through most of the upper part of the Jackfork (Fig. 5). For individual units, lithology, mean grain size, sorting, sedimentary structures, and inferred sedimentary environments are indicated. Variation in mean grain size is commonly one of the keys to recognizing unit boundaries. Table 2 is a unit by unit comparison of units described in Figure 5 with the units described by Cline and Moretti.

WILDHORSE MOUNTAIN FORMATION

The Wildhorse Mountain Formation, which probably forms 70% of the Jackfork Group along U.S. 259, is like most other Jackfork turbidites described in the literature (Morris, 1974,1975). However, the upper part of the Jackfork Group (Prairie Mountain Formation and higher) is poorly documented and significantly underinterpreted in the literature.

Thick Wildhorse Mountain Formation turbidites crop out east and southwest of Big Stick Monument (Fig. 3).

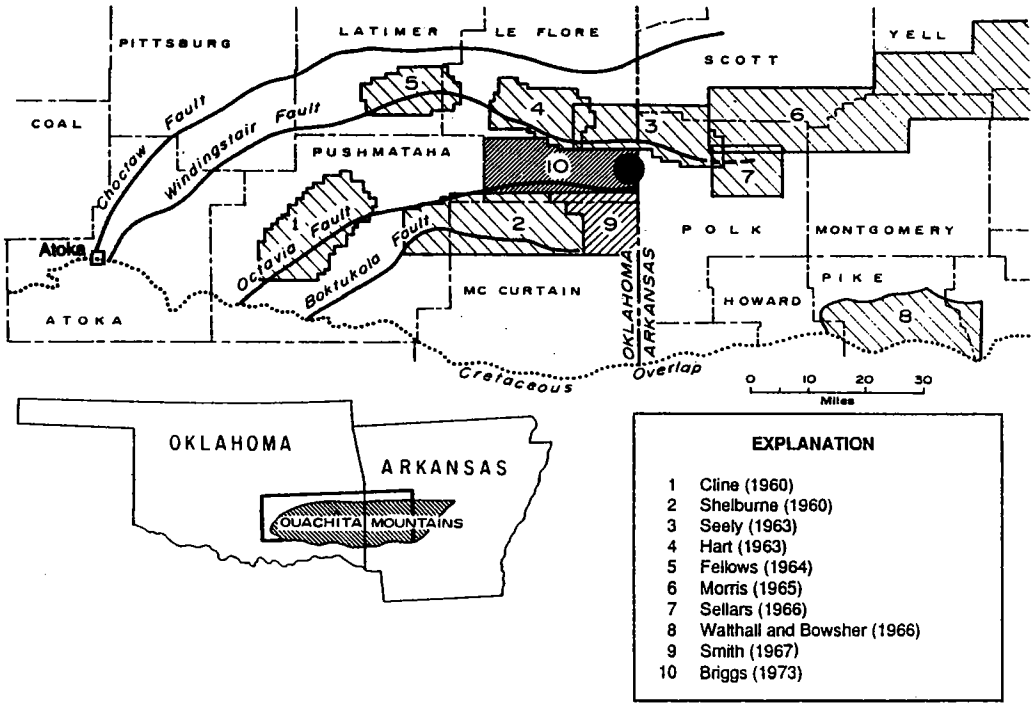


Figure 1. Location map for Ouachita Mountains in southeastern Oklahoma and western Arkansas. Black circle is location of U.S. Highway 259 measured section. (Modified from Briggs and Smith, 1973.)

Most of the turbidite sandstones are thick and well cemented; they were deposited in deep water. Near and southwest of Three Sticks Monument along U.S. 259, the upper part of the Wildhorse Mountain Formation becomes slightly shalier, and distinct packages of turbidite sandstones occur between shales. Several of the sandstones contain accumulations of coalified plant fragments and carbonaceous debris. These accumulations of shallow-water carbonaceous detritus occur in the turbidite sandstones in several forms including (1) carbonaceous fragments disseminated throughout the quartz sandstones, (2) as accumulations on bedding or lamination surfaces, and (3) as the dominant fraction in "slurried beds" (Fig. 6).

Erosion on an outcrop scale at the base of the turbidite packages is generally minimal. In the outcrops along U.S. 259, erosion within the packages at the base of individual beds is usually also minor and has a relief of less than 1 ft (Fig. 7). Rarely, erosion at the base of some packets of beds has resulted in a relief of up to 1.5 ft. Characteristics of Jackfork turbidites are summarized in Table 3.

TABLE 2. — COMPARISON OF UNITS AND THICKNESSES IN JACKFORK GROUP

Formations	Units: This paper	Units: Cline and Moretti (1956)	Thicknesses: This paper	Thicknesses: Cline and Moretti (1956)
Wesley	30–32	59	66 ft (part)	150 ft
Markham Mill[a]	29	58	29 ft	38 ft
	28	56 (part), 57	156 ft	175 ft (part)
	19–27[b]	56 (part)	58 ft	148 ft (part)
Prairie Mountain				
	18	55 and 56 (part)	105 ft	105 ft and 148 ft (part)
	9–17	53, 54	330 ft	285 ft
	4–8[b]	52	240 ft	230 ft
	1–3	51	200 ft (part)	650 ft

[a]Base of Markham Mill Formation as designated at this location is a sequence boundary.
[b]Flooding surfaces above Unit 8 (this paper) and above Unit 27 (this paper).

PRAIRIE MOUNTAIN FORMATION

I interpret the top of the Wildhorse Mountain Formation to be at the base of Unit 45 of Cline and Moretti (1956). Cline and Moretti picked the base of the Prairie Mountain Formation just above Unit 45, rather than at the base of the unit. I interpret the contact at the base of Unit 45 to be a sequence boundary that represents significant shallowing.

The basal sandstone of the formation is friable and generally massive appearing; it forms 2- to 3-ft-thick lenses that thin and thicken over the length of the outcrop (Fig. 8). Horizontal bedding or lamination is absent. The sandstone may include large troughs and white to

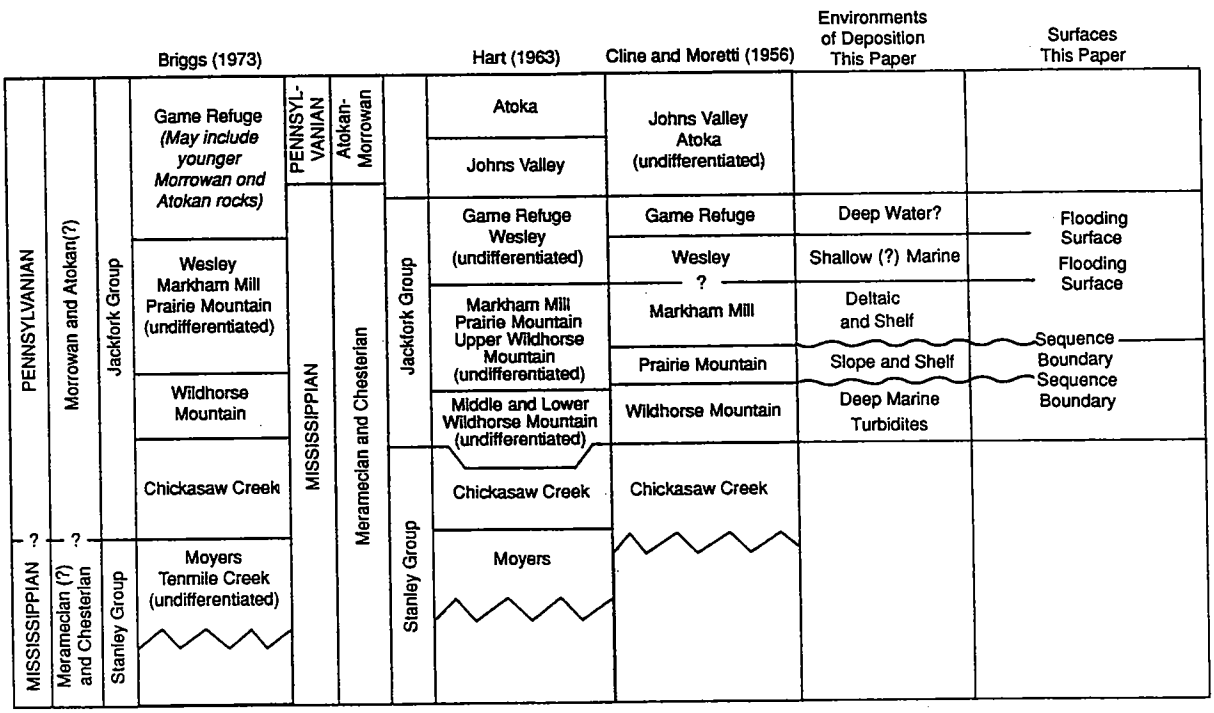


Figure 2. Stratigraphic reference sections of previous workers in southeastern Oklahoma. Cline and Moretti's (1956) units are used in this paper. Sequence stratigraphic surfaces are indicated. (Modified from Briggs, 1973.)

pink shale drapes on some 3- to 6-ft-wide depressions. The total volume of shale in outcrop is <2%. Laterally, shales are truncated (by troughs?). Erosional relief at the contact of the units is as much as 1.5 ft. Only a trace of burrowing is observed. Grain size ranges from 125 to 250 mm; mean grain sizes 175–200 mm. Sorting is moderate to poor. There is a strong grain-size contrast between the basal beds of this unit and the underlying, thinly bedded turbidites (mean grain size 100–125 mm). In hand sample, Cline and Moretti's (1956) unit 45 appears to be relatively porous and permeable. This sandstone (channel?) probably was deposited entirely by traction currents.

Greenish-pink clayey shale occurs just above the porous sandstone. The clayey shale is typical of that observed in the Prairie Mountain and some parts of the Atoka Formations in other areas in Le Flore County and is interpreted to have been deposited by processes other than turbid flow. This type of clayey shale may have been deposited in relatively shallow water and is quite different from the mudstones interbedded with or associated with the turbidites of the underlying Wildhorse Mountain Formation.

Boulder beds occur at several locations along U.S. 259. The lowest "boulder bed" is Unit 3 (123–200 ft, Fig. 5). The boulders in these beds are rounded blocks of sandstones that, for the most part, are not in contact with one another; rather, individual boulders are separated by a silty and muddy shale matrix. Units such as these are commonly associated with slope deposits (Table 4). These beds (Units 3, 9, and 18; Fig. 5) were not singled out in Cline and Moretti's (1956) measured section but were noted by Briggs (1973) and Black (1969). They occur in several units in the upper part of the Prairie Mountain Formation.

Measured section 6A-93 (Fig. 5) extends from the middle of the Prairie Mountain Formation to near the top of the Wesley Formation, Cline and Moretti's Unit 59. The newly measured section, which is the subject of this paper, does not include units below Cline and Moretti's Unit 51. The section will be described generally from the base upwards.

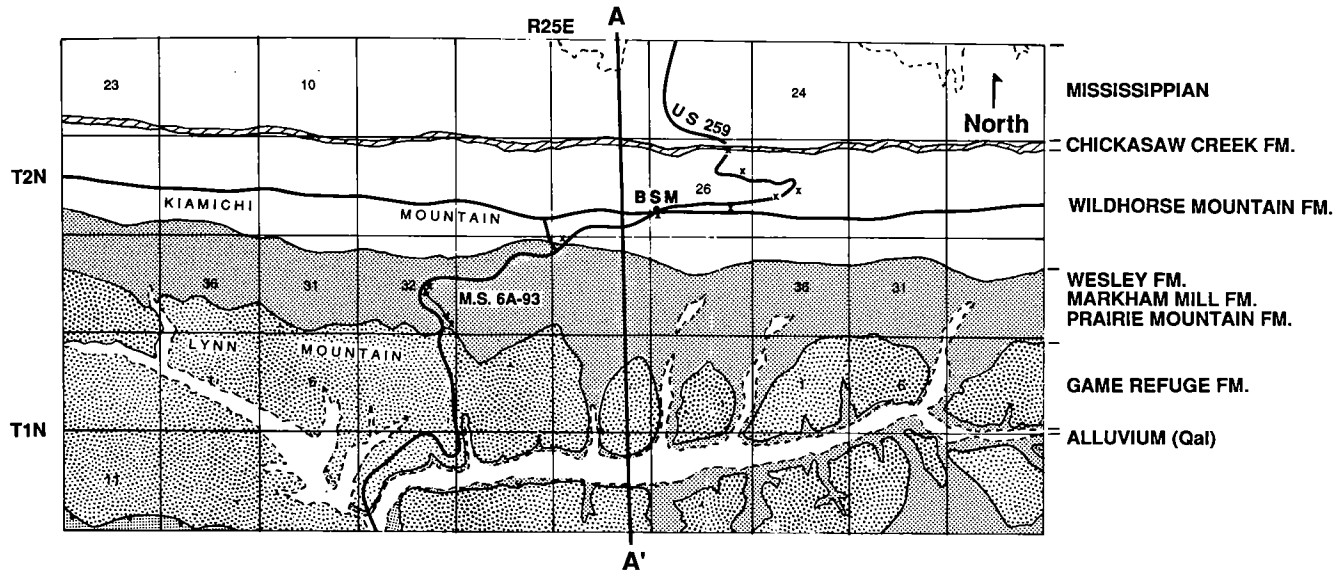


Figure 3. Geologic map of part of Lynn Mountain syncline in study area. Locations of measured section 6A-93 and stratigraphic cross section A–A′ are indicated. Base of Chickasaw Creek Formation is 2.4 mi south of Big Cedar. BSM = Big Stick Monument; x = diagnostic Jackfork Group outcrops. (Modified from Briggs and Smith, 1973.)

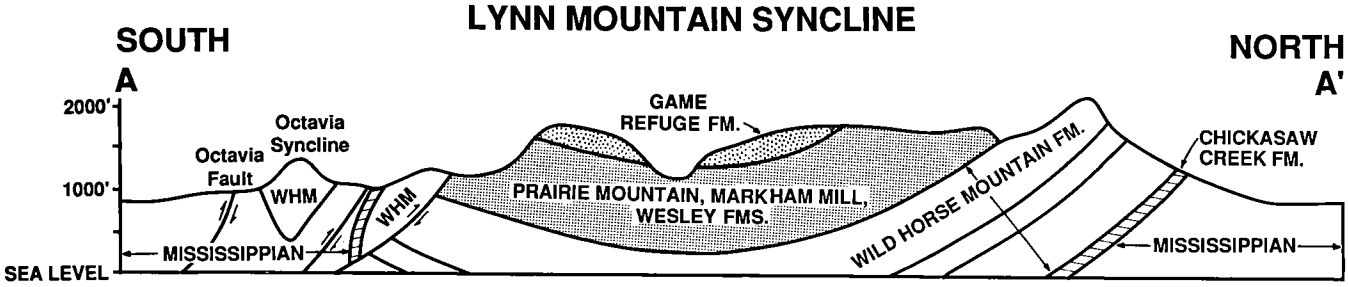


Figure 4. Stratigraphic cross section A–A′. WHM is Wildhorse Mountain Formation. Location is ~1.7 mi east of measured section 6A-93 along U.S. Highway 259. (Modified from Briggs and Smith, 1973.)

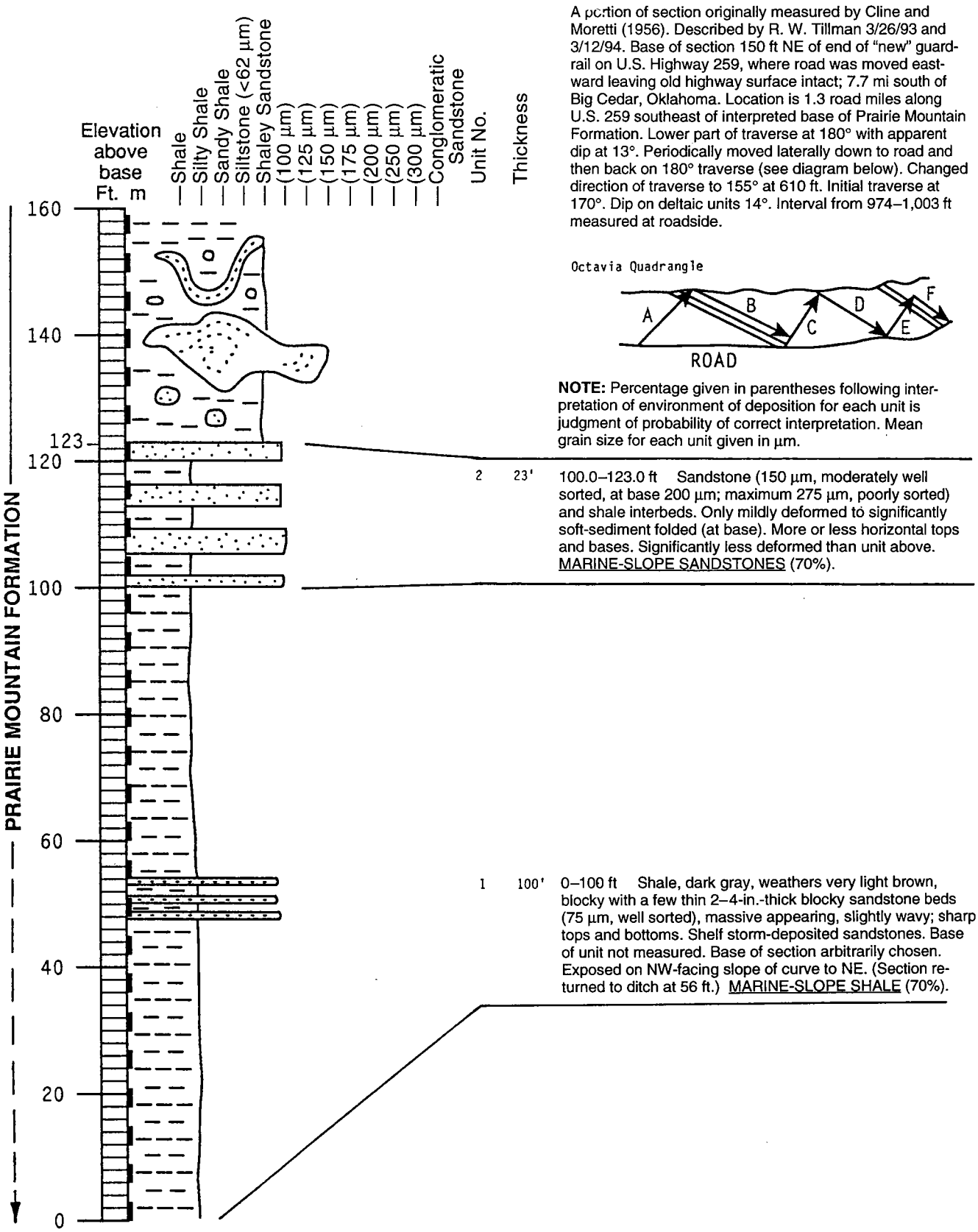
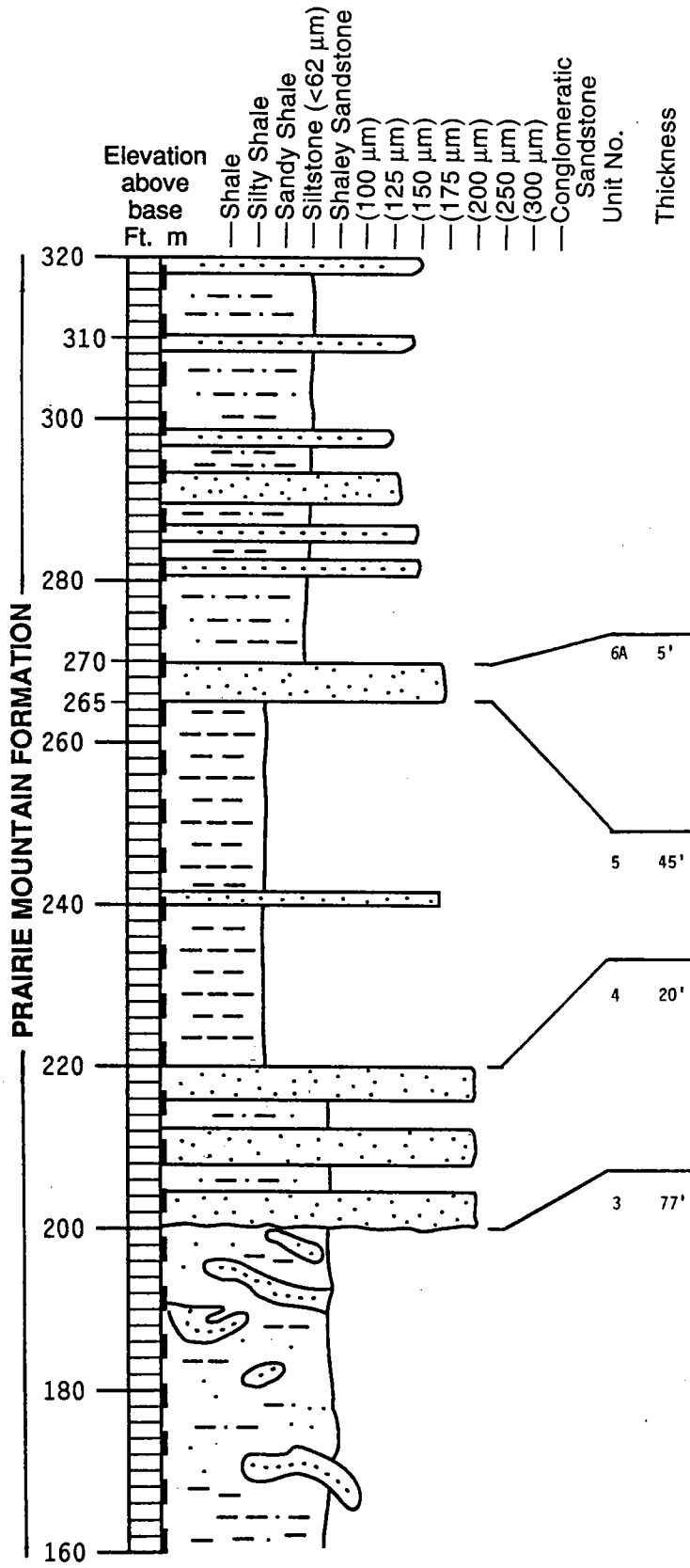


Figure 5 *(p. 209–215)*. Jackfork Group, Measured Section 6A-93, along U.S. Highway 259, NW¼SE¼ to SE¼SE¼ sec. 32, T. 2 N., R. 25 E., Le Flore County, Oklahoma.



Unit No.	Thickness	
6A	5'	265.0–270.0 ft Sandstone (175 µm, moderately well sorted), blocky. Includes 3 beds ≤2 ft thick, sharp tops and bases. Laminations, 0.5 cm thick, slightly wavy, with 0.5-in. amplitude. Upper part of lowest bed deformed and eroded. Erosional concave-down bedding surface with amplitude of 0.8 ft. Burrows essentially absent internally and from bedding surfaces, except on some thin beds where surfaces highly burrowed. MARINE-SHELF SANDSTONE (75%).
5	45'	200.0–265.0 ft Shale, light gray, clayey non-fissile(?). Non-bedded shale except in 5 ft above sandstone. Less than 10% thin (<1 in. thick) 150–175 µm, moderately sorted sandstone lenses or beds. SHALLOW(?)-MARINE SHALE (90%).
4	20'	200.0–220.0 ft Sandstone (125 µm, moderately well sorted; 150 µm at top), beds 1–3 in. thick. Sharp upper and lower contacts. Only slight deformation; massive appearing. Apparent dip 33° to south. A few nondescript burrows on some of thinner (0.1 ft thick) bed bedding surfaces. Base of unit is at top of any significant deformation. Base disconformable. 20% ≤1-ft-thick shale beds. (Returned to roadside ditch at 220 ft.) SHELF SANDSTONE (75%).
3	77'	123.0–200.0 ft Shale (70%) with deformed blocks of sandstone (30%, 150 µm, moderately well sorted) interspersed. 30% blocks of sharply bent and deformed beds of sandstone and composite blocks (consisting of many cemented smaller blocks) ≤20 ft in diameter. Locally shale "squeezed" out from between sandstone beds. Larger blocks appear to have been deposited locally and deformed more or less in place. Many (15%) smaller rounded "glupes" and "globs." Near top are elongate deformed 1-ft-thick beds that are bent and have curved surfaces. Some thin beds distinctly horizontally laminated on millimeter scale and burrows on bedding surfaces. SLOPE DEPOSIT (85%).

Figure 5. *(Continued.)*

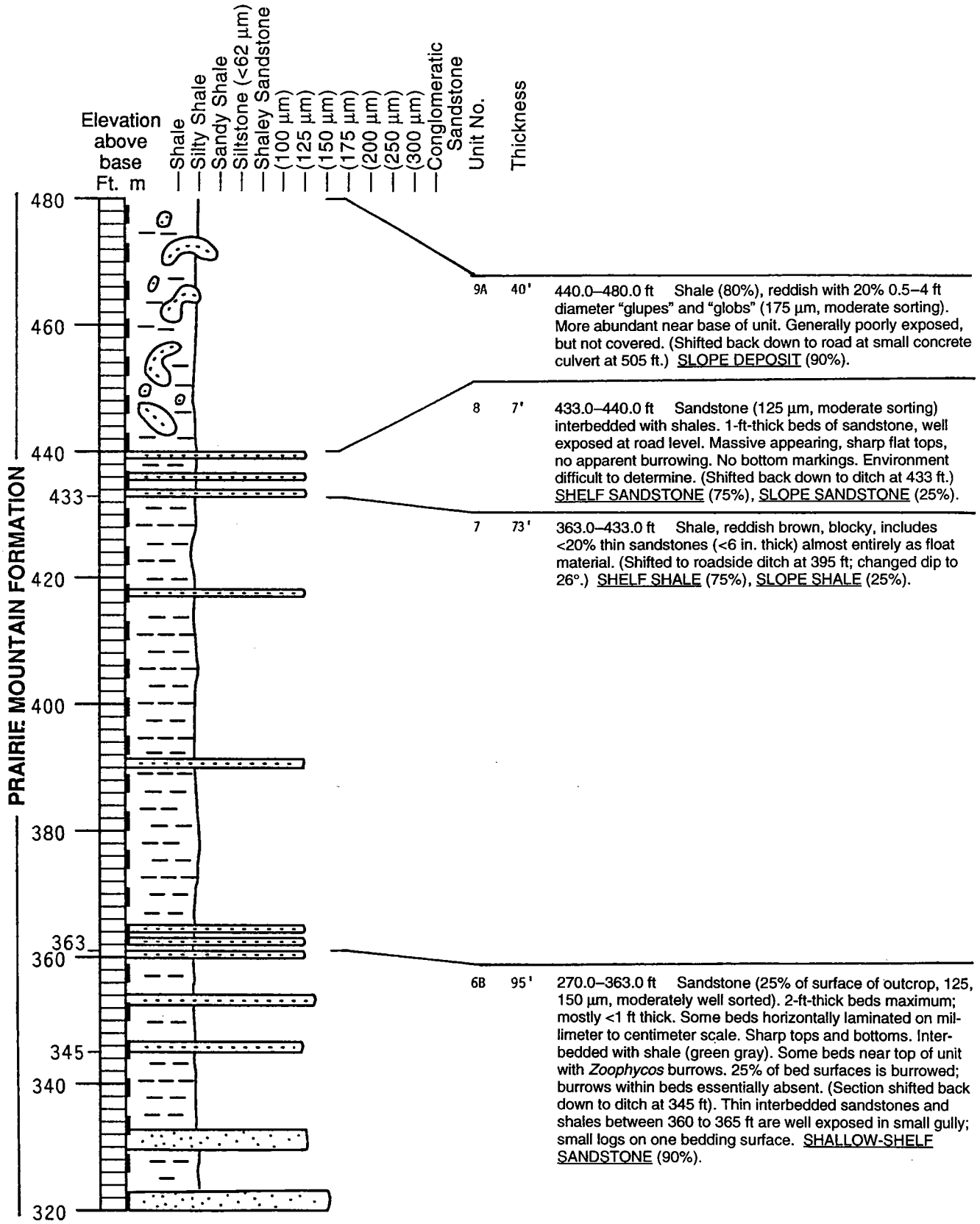


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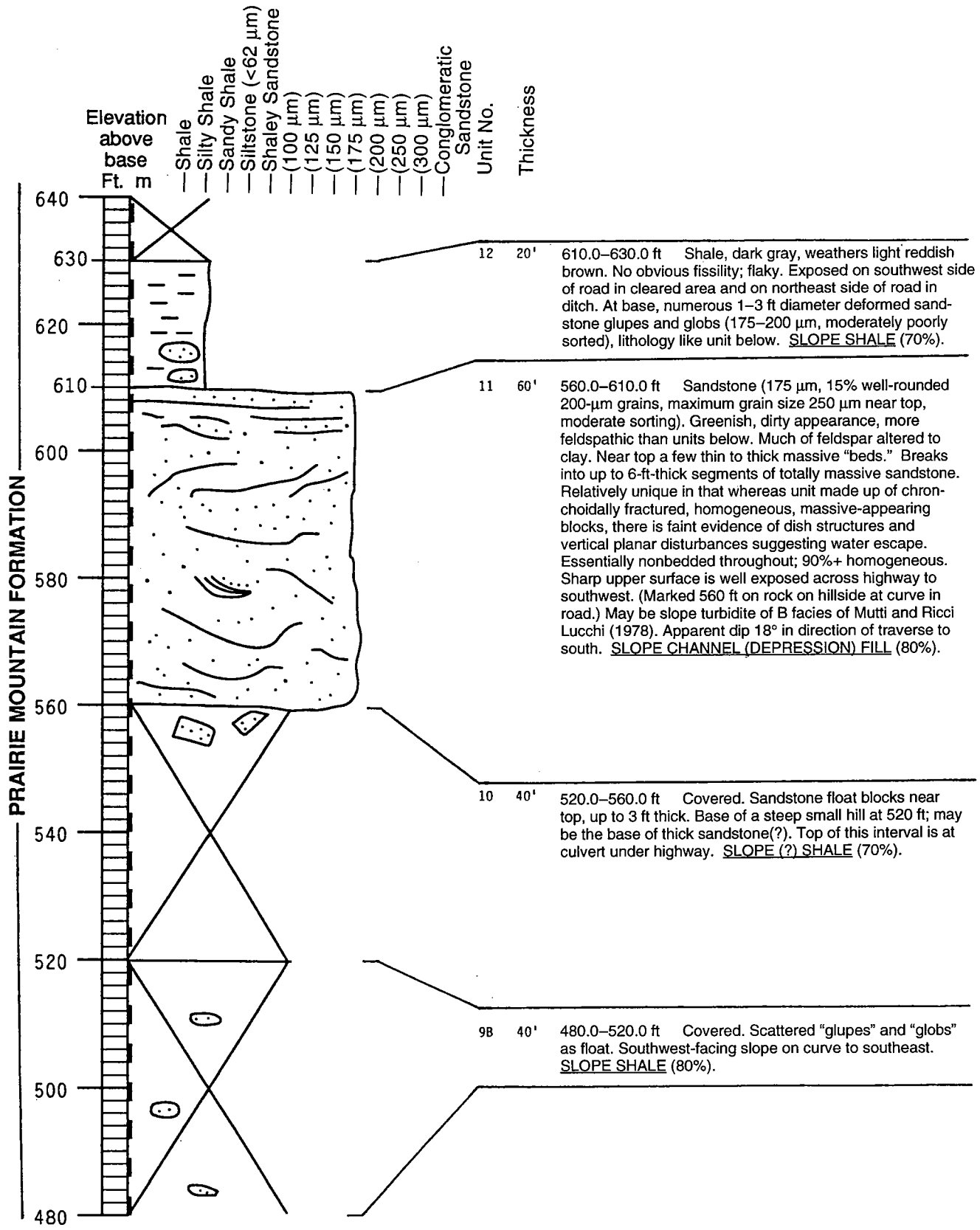


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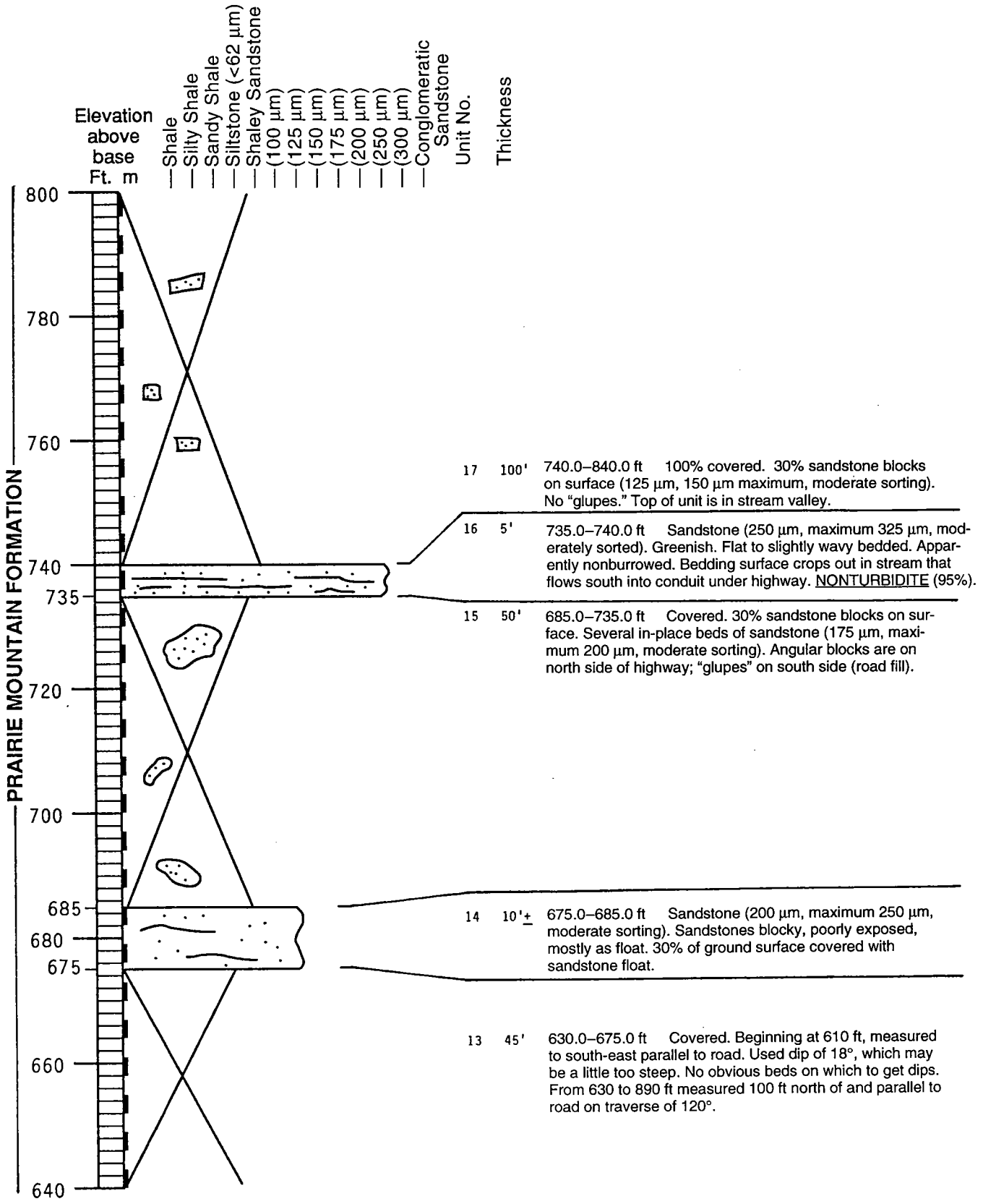


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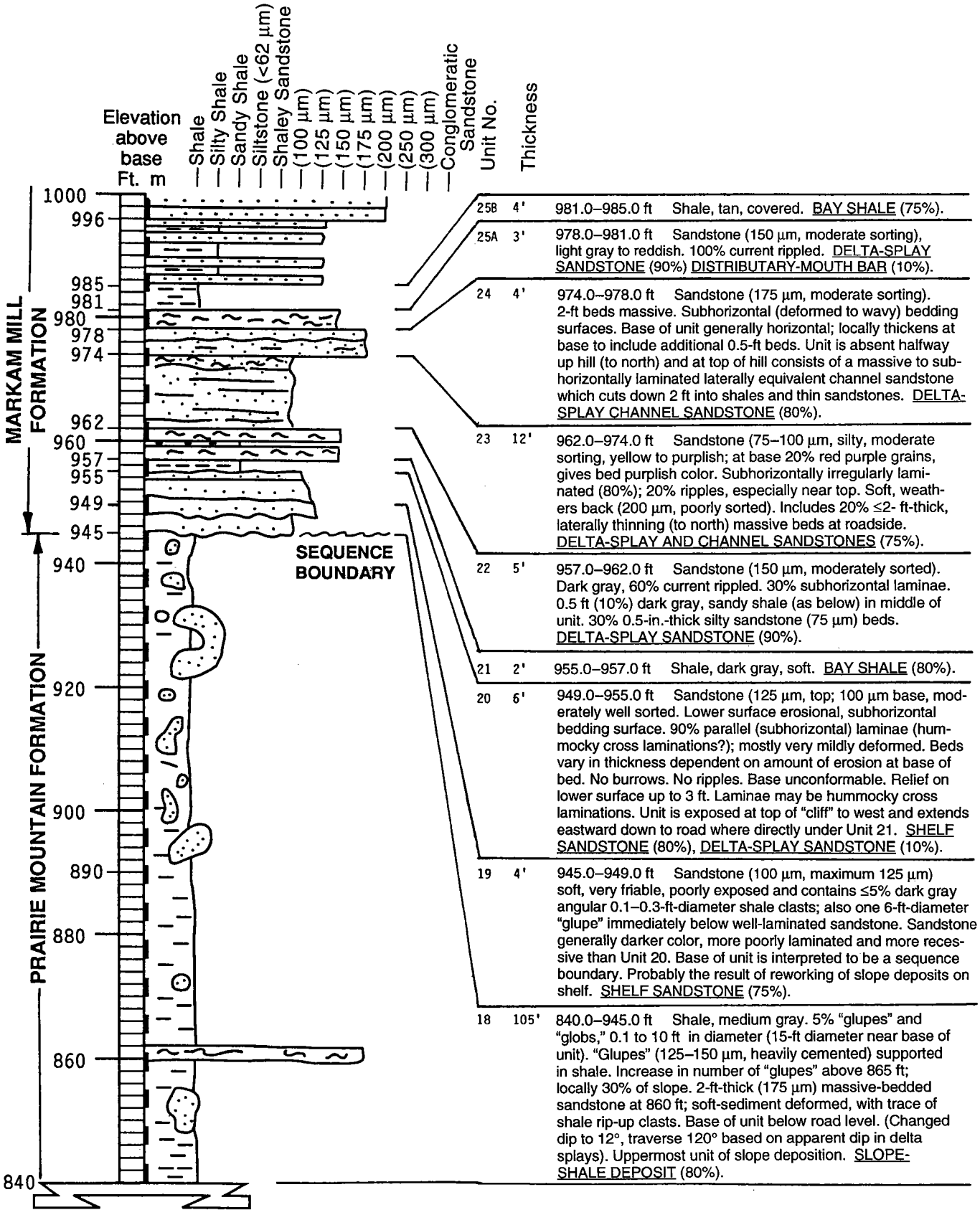


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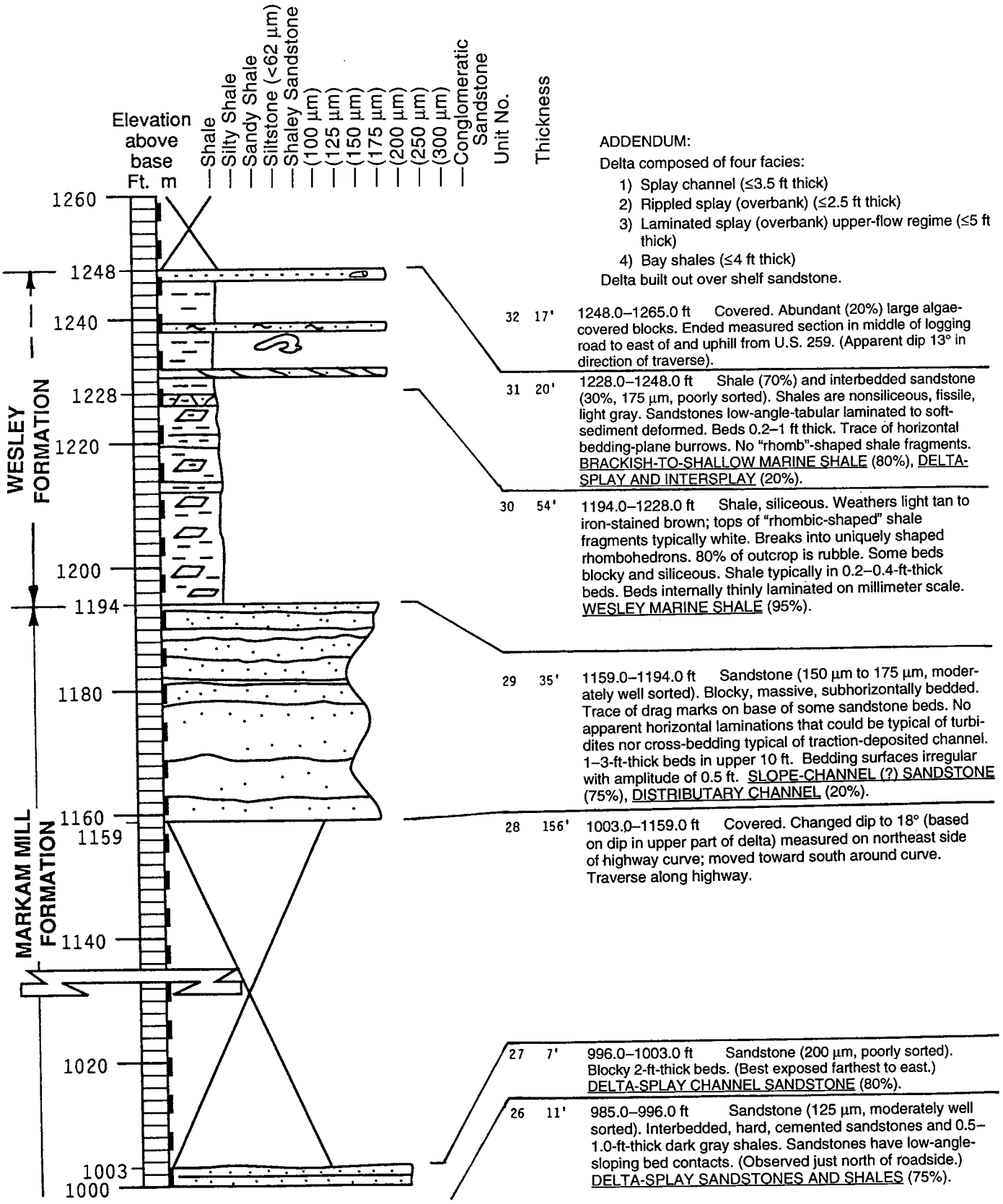


Figure 5. *(Continued.)*

TABLE 3. — SUITE OF TURBIDITE FEATURES OBSERVED IN WILDHORSE MOUNTAIN FORMATION OF JACKFORK GROUP

Horizontal bedding
Sharp tops and bases of beds
Bouma sequences
Partial Bouma sequences
Asymmetrical current ripples (as opposed to symmetrical wave ripples)
Absence of traction features larger than ripples
Horizontal laminations
Normal grading
Inverse grading
Bottom marks Flutes (directional) Grooves (bidirectional)
Deep-water trace fossils
Water-escape features Dish structures Pipes
Rip-up clasts concentrated at top of beds
Local interbedding of shale and sandstone on fine scale Levee deposits Overbank deposits

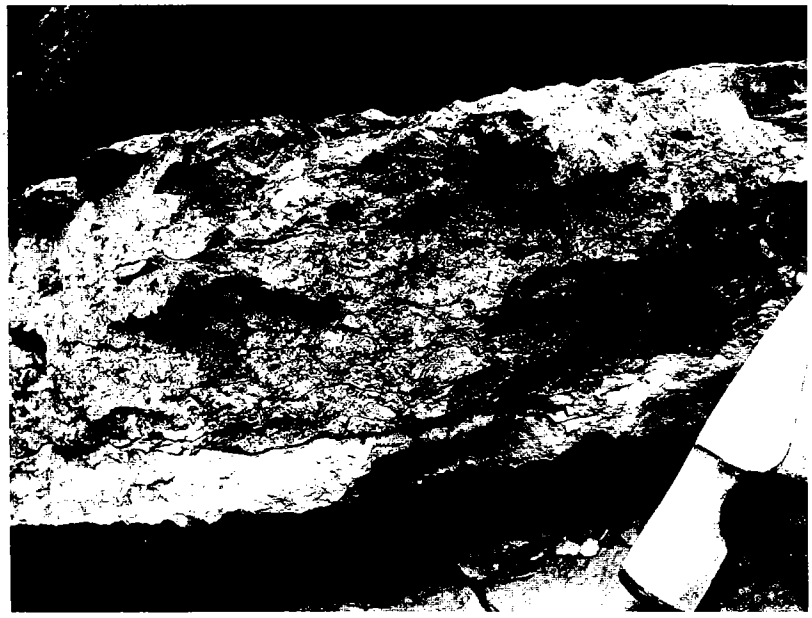


Figure 6. Slurried "turbidite" (blue-bed) containing abundant plant and wood fragments. Plant fragments are jumbled and have no preferred orientation. Thin turbidite sandstones are interbedded (hammer head for scale). Bed is just below strata shown in Figure 7. Wildhorse Mountain Formation of Jackfork Group.

Shelf Sandstones and Shales

A thick shale interval from 200 to 440 ft in the upper part of Prairie Mountain Formation contains <25% sandstone and is interpreted to consist of shelf deposits. A shelf interpretation is supported by the presence of thinly bedded, horizontally laminated sandstones with burrows of moderately shallow water type (including *Zoophycos*). Wood fragments occur on some bedding surfaces (Fig. 9). Burrows and trails are generally limited to bedding surfaces (Figs. 10,11). The tops and bases of some of the sandstones are gradational; others are sharp. Slightly wavy laminations also occur. Interbedded shales are pastel colors (light gray, pink, and light green). Neither fissility nor bedding is observed in the shales except where sandy lenses occur, and there bedding is horizontal. Deformation and transported clasts typical of the slope deposits are absent, as are packages of sharp-based and sharp-topped turbidite sandstones. The absence of typical slope and turbidite characteristics adds further to a shelf interpretation for this interval.

The characteristics of shelf sandstones are listed in Table 5. Many of the characteristics listed under "Nearshore and Shoreface Attached Shelf-Ridge" are observed in those Prairie Mountain Formation strata interpreted as shelf deposits. The characteristics of offshore-shelf sandstones are also listed in Table 5 to contrast with the characteristics observed in the Jackfork.

Slope Deposits

Overlying the lowest shelf deposit is one of at least three thick "boulder shale" intervals (Unit 9, 440–520 ft). Here, as in the "boulder beds" below (Fig. 12; Unit 3 and Unit 9, Fig. 5), the clasts occur as "raisins in a pudding" rather than being clast-supported. These boulders are sedimentary and as much as 8 ft in diameter (Fig. 13). The exact composition of the boulders was not documented by me; however, Briggs (1973) noted the following:

> In Unit 55 [of Cline and Moretti, 1956] the "boulders are not concentrated at any particular horizon but are randomly distributed throughout the shale. Most of the boulders are composed of sandstone but the sandstone boulders vary considerably in size, color, matrix content and induration. Ironstone concretions and a few small chert cobbles were also observed. No limestone boulders were found."

As with the "boulder beds" in Unit 3 (Fig. 5), these deposits suggest at least significant local relief, possibly on a slope. The slope may have been the margin of a continent or the result of faulting.

MARKHAM MILL FORMATION

The top of the "boulder shale" (945 ft) is interpreted to be a sequence boundary separating moderately deep water slope deposits below from the overlying shallow-marine shelf and deltaic sandstones above (Figs. 14,15).

Shelf Deposits

The interval from 945 to 955 ft is interpreted as a shelf sandstone deposit and includes two distinct units (19 and 20). These units lie between what are interpreted as slope deposits below and deltaic deposits above. A sequence boundary probably exists at the base of these shelf deposits, and parasequence boundaries occur within (at the unit boundary, Fig. 5) and at the top of the deposits. The lower shelf unit (19) is interpreted as a reworked sandstone that may have resulted from reworking of underlying slope sandstones. Because of the reworking, the lower contact of this unit appears to be slightly transitional. The lower shelf sandstones (Unit 19) are darker colored, more poorly laminated, and less well cemented than the overlying shelf sandstones (Unit 20).

Unit 20 (949–955 ft) is a very distinct and resistant sandstone that is well exposed in outcrop (Figs. 14,15). The internal laminations are 90% subhorizontal with a suggestion of hummocky cross-stratification (HCS). The unit, as a whole, is mildly soft-sediment deformed. Beds vary in thickness depending on the amount of erosion at the base of individual beds and at the base of the unit (3 ft). The absence of ripples in this unit probably eliminates any possibility that this is a distributary mouth–bar sandstone.



Figure 7. Turbidites in Wildhorse Mountain Formation of Jackfork Group. Note abrupt horizontal upper and lower bed contacts typical of turbidites. Beds are 0.5–3 ft thick and are located along U.S. Highway 259, 1.3 mi northeast of measured section 6A-93.



Figure 8. Base of Prairie Mountain Formation. Shallow-water, yellowish white, friable, 50-ft-thick channel sandstone. Sandstone is in sharp contact (below photo) with Wildhorse Mountain sandstone turbidites. Outcrop is stratigraphically above and a short distance west of rocks shown in Figure 7 (location: NW¼NW¼NW¼ sec. 34, T. 2 N., R. 25 E).



Figure 9. Wood fragments on bedding surface of shelf sandstone at top of Unit 6B at 364 ft in measured section 6A-93 (Fig. 5). Fragments are oriented parallel to bedding.

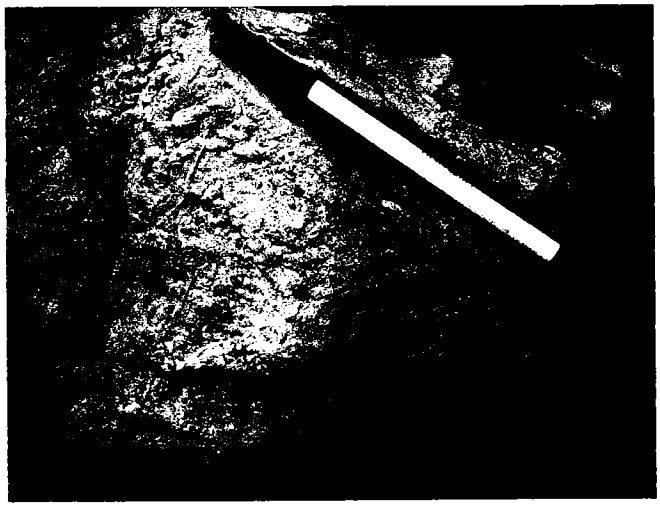


Figure 10. Tracks and trails on thin-bed bedding surface in shelf sandstone. Fine grazing pattern visible below and to the right of end of pen. Unit 4 in measured section 6A-93 (Fig. 5). (Pen for scale.)

TABLE 4. — CHARACTERISTICS OF SLOPE DEPOSITS

Primarily fine grained (shale, mudstone, and silt)

Thin mudstones (1–5 cm) may show grading

Variable size clasts may be common
- (a) Deposited primarily by gravity
- (b) Most had intrabasinal source
- (c) May contain a relatively narrow range of rock types, mostly sandstones
- (d) Commonly rounded and/or includes "bottom marks" on all sides ("glupe") or on one side ("glob")
- (e) Very variable in size; 1-cm to several meters diameter
- (f) Commonly matrix-supported

May include slope-channel sandstone filling

May form very thick accumulations
- (a) Where relative sea level rises
- (b) Where down-to-basin faulting occurs
- (c) Seaward of delta

May contain both "shallow-water" and "deep-water" burrows, including *Zoophycos*



Figure 11. Trails of pellets on bedding surface of shelf sandstone. Unit 4 in measured section 6A-93 (Fig. 5). (Pen for scale.)

Deltaic Deposits

The sandstones in the interval above the proposed sequence boundary (near the base of Cline and Moretti's [1956] Unit 56, Table 1) are interpreted to include a variety of shallow-marine, brackish, and possibly continental deposits including deltaic sandstones (Figs. 16–19). The interval from 949 to 996 ft (Units 22–27; 47 ft thick, Fig. 5) is interpreted to be deltaic, and the deposits are characterized primarily by thick sets of thinly bedded current ripples varying from 20% (Unit 23) to 100% (Unit 25A) ripples. Subhorizontal lamination is also prominent in some units, varying from 30% (Unit 22) to 80% (Unit 23). The mean grain size varies from 100 to 150 μm in the rippled and laminated beds.

Massive to laminated channeled sandstones that are as much as 4 ft thick, but thin laterally to 0 ft, are also observed (Unit 24, Fig. 5; Figs. 20,21). Massive to laminated beds 2 ft thick and interbedded with thinner-bedded sandstones form as much as 20% of Unit 23 and the entirety of Unit 27 (Fig. 5). Lateral variation of bedding types over short distances (<10 ft) is common. No burrowing was observed. All the deposits in the deltaic interval were emplaced by traction and are interpreted to represent delta splays and delta-splay channels. A relatively thick distributary channel, near the measured section, is postulated as a source for the splay sandstones. A summary of the features observed in the deltaic deposits of the Jackfork Group is given in Table 6. The shelf and deltaic deposits (Units 19–26 and the shelf deposits in the interval from 200 to 440 ft; Fig. 5) were deposited in a much shallower environment than has been previously interpreted for the Jackfork Group. The fluvial channel (Table 7) or distributary channel, which must be the source of the delta splays, is probably not present in the U.S. 259 outcrop unless the sandstone in the interval from 1,159 to 1,194 ft (Fig. 5) is a distributary channel. Sedimentary features are sparse in this unit (29), and a precise interpretation of environment of deposition cannot be made.

Slope and/or Deltaic Channel Sandstones

Approximately 56 ft are covered above the top of the well-exposed deltaic deposits. A thick sandstone in the interval from 1,159 to 1,194 ft (Unit 29, Fig. 5) is superfi-



Figure 12. Aerial photograph showing "boulder bed" near base of measured section 6A-93, Unit 3 (Fig. 5). "Boulders" are rounded, transported fragments of sandstone beds that may have "bottom marks" on all sides, informally designated as "glupes," or only on one side, "globs." May represent a slope deposit.

cially like Unit 11, which is a 60-ft-thick sandstone. Both sandstones are generally massive appearing and subhorizontally bedded. Bedding surfaces are slightly irregular with an amplitude of 0.5 ft. Angle-of-repose traction cross-lamination is absent. Both units (11 and 29) have mean grain sizes slightly coarser (175 μm) than most of the deposits interpreted as deltaic and shelf. Unit 11 contains grains as large as 250 μm and appears to be less quartzose than most other units.

Some beds in Unit 11 are 6 ft thick and break into conchoidally fractured, homogeneous blocks having faint dish structures and vertical planar disturbances that



Figure 13. Large, transported, rounded clasts "floating" in shale matrix. Maximum diameter of "glupes" and "globs" in this unit is 10 ft. Location of photograph is near top of Unit 18 in measured section 6A-93.



Figure 14. Sequence boundary (arrow) between slope deposits (below) and shelf sandstone (above). Contact is between Units 18 and 19 at 945 ft in measured section 6A-93 (Fig. 5). "Glupes" near right side of photo are in slope deposit at the very top of Unit 18. Unit 19 is a sandstone (reworked on shelf?) that is more friable and more recessive than Unit 20. Markham Mill Formation of Jackfork Group.



Figure 15. Planar to hummocky laminae in sandstone (Unit 20 of measured section 6A-93, Fig. 5) just above sequence boundary. Upward curving of laminae may be hummocky cross-stratification (HCS), which is indicative of wave or combined flow deposition. Environment probably inner-shelf deposition, possibly delta fringe.



Figure 16. Interbedded, thin, rippled delta-splay sandstones and "lagoonal" shales and mudstones. Photograph taken 50 ft north of U.S. Highway 259 and location of measured section 6A-93 (Fig. 5). Equivalent to Units 23 and 24.

TABLE 5. — CHARACTERISTICS OF SHELF SANDSTONES

Nearshore and Shoreface Attached Shelf Ridge

1. Primarily wave deposited
   (a) HCS (hummocky cross-stratification), lower shoreface and offshore transition
   (b) SCS (swaley cross-stratification), middle shoreface
   (c) Subhorizontally to horizontally laminated
   (d) "Smeared-out troughs"; formed by combined flow
   (e) Commonly fine-grained; moderately well sorted
2. May contain some tidal and storm deposits
3. Suite of trace fossils will be those from nearshore zone

Offshore Shelf Ridge (Not present in Jackfork)

1. Primarily storm-current deposits
2. Marine sand body surrounded by shale; sides, bottom, and top
3. Coarsens upward, facies by facies, from silt to medium-grained sand
4. Facies may climb stratigraphically in oblique fashion
5. Glauconitic (trace to 30% locally)
6. Predominently angle of repose or combined flow cross-lamination; upper portion. Current rippled; lower and lateral parts of ridge
7. Sand body asymmetrical
   (a) In form
   (b) In distribution and types of lithofacies
   (c) In occurrence of siderite and clay rip-up clasts
8. Relatively uniform flow directions (oriented subparallel to regional bottom contours)
9. Absence of shoreline-type burrows such as *Asterosoma*
10. Absence of littoral-zone foraminifera suite
11. Absence of evidence of subaerial exposure
12. May contain wave-generated deposits near top of shallowing-up sequence

probably were avenues for water escape from the sand soon after deposition. Unit 11 is interpreted to be a slope channel-fill deposit and may be a "B" facies turbidite deposit (terminology of Mutti and Ricci Lucchi, 1978). Although these two units appear to be similar, the mechanism of deposition of Unit 29 is less interpretable; however, it may also be a slope channel-fill sandstone.

WESLEY FORMATION

The Wesley Formation is the highest stratigraphic unit redescribed and interpreted along U.S. 259 (Fig. 5). This cherty black shale has a very distinctive appearance: it breaks into numerous, small, well-cemented rhombohedrons that typically have "chalky"-white bedding surfaces (Fig. 22). The cherty shale is interbedded with very thin friable sandstones (Fig. 23). The Wesley Formation (Units 30 and 31, Fig. 5) is interpreted to be marine.

The Game Refuge Formation was not included in the newly measured section, but was observed in the course of measuring and describing several unpublished sections.

SEQUENCE STRATIGRAPHY

Several important sequence-stratigraphic surfaces occur in the series of outcrops along U.S. 259. The base of the Prairie Mountain Formation, which is well below the newly measured section, is interpreted to be a sequence boundary (Fig. 2). Near the top of the Prairie Mountain, the water in which sands were deposited shallowed. Just above the "boulder shale" at the top of the Prairie Mountain For-



Figure 17. Relatively thick sets of rippled, very fine grained (100 μm) sandstone interbedded with irregular, subhorizontally bedded and laminated sandstone beds. Delta-splay sandstone, Unit 23 in measured section 6A-93 (Fig. 5).



Figure 18. Ripples typical of very fine grained (100–125 μm) sandstones in delta splays. Markham Mill Formation of Jackfork Group. Unit 23 in measured section 6A-93 (Fig. 5).

TABLE 6. — CHARACTERISTICS OF DELTAIC DEPOSITS (MARKHAM MILL FORMATION)

1. All deposits show evidence of traction current deposition
2. Interbedded sandstones and shales
3. Sandstones are current rippled (most common) and horizontally laminated (high-flow regime)
4. Distributary and fluvial channels must be present locally(?) but not in measured section outcrop
5. Splay and overbank sandstones and lagoonal shales predominate
   (a) Locally silty
   (b) Locally lenticular
   (c) Interbedded on variety of scales
   (d) Shales are pastel in color and generally lack fissility
6. Sandstones commonly have sharper bases than tops
7. Burrowing sparce to absent
8. Locally carbonaceous; more carbonaceous than most of deeper-water facies
9. Base of delta only a few feet above sequence boundary that separates shelf sandstones from underlying slope deposits
10. Absence of shoreface, slope, and turbidite features

TABLE 7. — CHARACTERISTICS OF FLUVIAL SANDSTONES

Scoured base of channel

Beds show cut and fill (1–3 ft minimum)

Laterally variable

Cross-laminated
   (a) Troughs
   (b) Planar-tabular (braided)

Traction deposits (absence of turbidity deposition features)

Current ripples

Planar-horizontal (upper flow regime) laminations

Locally fining upward (point bar)
   (a) Decrease in mean grain size upward
   (b) Increase in amount of clay upward

May have lateral accretion beds

May be poorly sorted

May contain wood and carbonaceous fragments

Interbedded shales are light colored

Clay rip-up clasts may be common locally

Interbedded relatively thin sandstones and shales
   (a) Interbedded overbank sandstones
   (b) Rippled splay sandstones

mation, a sequence boundary apparently separates "slope" deposits below from shallow-marine shelf and deltaic deposits of the Markham Mill Formation above. At least one flooding surface may occur near the top of the Markham Mill Formation or at the base of the Wesley Formation. A flooding surface probably also occurs at or near the base of the Game Refuge Formation. Several parasequence boundaries may be present in the Prairie Mountain Formation and the upper part of the Markham Mill Formation.

CONCLUSIONS

Relatively thick sections of the Jackfork Group are interpreted as significantly shallower water deposits than many previous workers have indicated. A 240-ft-thick interval (Units 2–8, Fig. 5) and a 10-ft-thick interval (Unit 20, Fig. 5) interpreted as shelf deposits and a 46-ft-thick deltaic sequence (Units 22–27, Fig. 5) are the shallowest water deposits observed. The shelf deposits are characterized by thin, predominantly subhorizontally laminated sandstones; some have burrowed upper surfaces, and a few have internal burrowing. These sandstones are interbedded on a variety of scales with shale and mudstone.

Deltaic units are characterized throughout by traction deposition in the form of thick rippled beds; thin, laterally discontinuous channels; local soft-sediment deformation; and several "massive-appearing," relatively thicker, more resistant sandstones.

Several thick intervals, which may have been deposited on slopes (either fault-related or on the edge of a continent), are present. These slope deposits are marked by numerous small to large deformed blocks of sandstone in a mudstone or shale matrix.



Figure 19. Soft-sediment–deformed, thin-bedded sandstones in delta splay (Unit 24 in measured section 6A-93, Fig. 5). Located 3 ft above small channel in Figure 20. (Hammer for scale.)



Figure 20. Lens-shaped delta-splay channel sandstone equivalent to Unit 24 in measured section 6A-93 (Fig. 5). Sharp lower base. Eroded into thinly bedded mudstones and sandstones. Thins toward both edges of photo. Width of channel is ~15 ft. (Hammer on left for scale.)



Figure 21. Close-up of Unit 24 in Figure 20. Note sharp base of channel sandstone. Internally massive to faintly horizontally laminated. (Hammer for scale.)



Figure 22. Wesley Formation of Jackfork Group. Distinct rhombohedron shapes and white bedding surfaces typical of cherty, thin-bedded marine mudstones that make up most of this unit. Unit 30 in measured section 6A-93 (Fig. 5).



Figure 23. Wesley Formation. Typical black, blocky, siliceous mudstone interbedded with shale. (Hammer for scale.) Unit 30, near top of measured section 6A-93 (Fig. 5).

Sequence boundaries are interpreted at the base and the top of the Prairie Mountain Formation. A significant flooding surface occurs at or near the top of the Markham Mill Formation. Additional flooding surfaces and parasequence boundaries may be interpreted between several units of the measured section.

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Three-Dimensional Seismic Interpretation of the Triangle Zone of the Frontal Ouachita Mountains and Arkoma Basin, Pittsburg County, Oklahoma

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ABSTRACT.—The boundary between the Ouachita fold-and-thrust belt and the foreland Arkoma Basin has been interpreted as a triangle zone. Three-dimensional seismic-reflection data from this triangle zone show details not recognized in conventional two-dimensional seismic data. The data indicate that the Kiowa syncline in the Arkoma Basin has been passively uplifted by blind thrusting at deeper levels. Offset is most apparent in two strong reflectors: the Morrowan Wapanucka Limestone and the Silurian–Devonian Hunton Group. Detachments for the thrusts in the Wapanucka Limestone are in the Springer Shale and the Woodford Shale. The thrusts offsetting the Hunton Group have detachments in the Simpson Group and possibly deeper in the Arbuckle Group. Horizontal shortening in the Wapanucka Limestone ranges from tens of meters to a few kilometers; horizontal shortening in the Hunton Group ranges from tens of meters to a few hundred meters.

Four reflectors were chosen for analysis: a reflector in the lower Atoka Formation, two repeated Wapanucka Limestone sections and the Hunton Group. All of these surfaces exhibit the same geometry, and their fold axes plunge to the southwest. The surface geometry and fold-axis data suggest that thrusting took place first in the Wapanucka Limestone and later in the Hunton Group. Faulting in the Hunton Group has been traditionally interpreted as normal, although basement-involved reverse faults, such as the Carbon fault, have been mapped. The interpretation of some individual in-line or cross-line profiles may also show normal faults, but the interpretation of the whole 3-D survey supports the proposed compression in the older sequence.

INTRODUCTION

The Ouachita system is a late Paleozoic orogen that extends from the southern end of the Appalachian Mountains to West Texas. The Ouachita Mountains and the southern part of the Arkoma Basin are the main outcrop regions of the late Paleozoic collisional orogen along the North American craton (Arbenz, 1989). The boundary between the frontal zone of the Ouachita Mountains and the foreland Arkoma Basin is drawn along the trace of the Choctaw fault. The Arkoma Basin is a mildly compressed fold belt characterized by open folds with minor faulting, whereas the frontal zone presents imbricated thrust faults with tight to overturned folds. The transition from fold-and-thrust belt to foreland basin is marked by a classic triangle zone (Hardie, 1988; Bertagne and Leising, 1989; Perry and Suneson, 1990; Wilkerson and Wellman, 1993).

This study uses a 28.45 km^2 3-D seismic survey from the triangle zone between the frontal Ouachita Mountains and southern Arkoma Basin to produce a detailed 3-D picture of the blind thrusting under the frontal syncline of the triangle zone. The 3-D survey is located in southeastern Oklahoma, in Pittsburg County, 10 km southeast of McAlester (Fig. 1). The survey is located 2 km north of the trace of the Choctaw thrust. It has a rectangular shape, with its long side parallel to the southwest-trending structural grain.

With the development of 3-D seismic technology, a more accurate image of the subsurface geology is possible. When the geology is complex or high resolution is required, 3-D seismic surveys offer the most detailed picture of the subsurface. The data density in 3-D surveys is adequate to satisfy the sampling theorem in space as well as in time. Migration of 3-D data eliminates sideswipe and places dipping and curvilinear reflectors in

Valderrama, M. H.; Nielsen, K. C.; McMechan, G. A.; and Hunter, Holly, 1994, Three-dimensional seismic interpretation of the triangle zone of the frontal Ouachita Mountains and Arkoma Basin, Pittsburg County, Oklahoma, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 225–241.

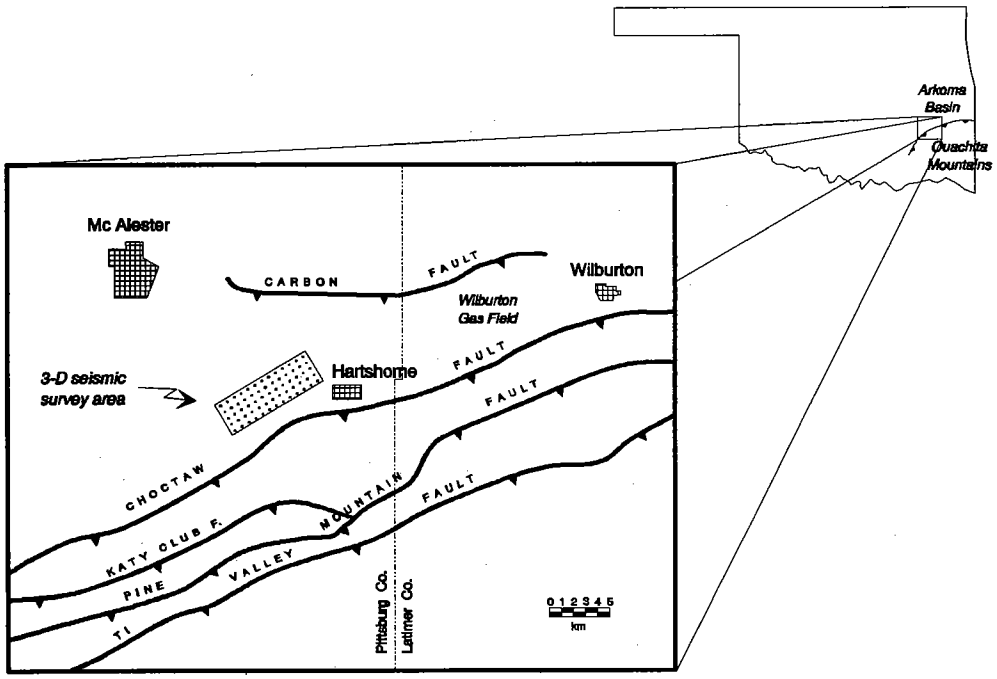


Figure 1. Location map showing the study area in Oklahoma. The surface trace of the main faults of the frontal zone are shown for reference. The Choctaw thrust separates the Arkoma Basin and the Ouachita Mountains in Oklahoma.

their proper vertical and horizontal positions. One of the most important advantages of a 3-D seismic survey is that the data reduce to a volume that can be examined in various ways: vertical slices in any arbitrary direction, time slices, chair displays, and isometric views. This facility is important because a feature that may not be clear in one type of a section may be evident in another. The time slice is very useful for structural interpretation as it shows the structural position of the reflectors, very much like a geologic map.

Regionally, the stratigraphy of the southern part of the Arkoma Basin and the northern part of the Ouachita Mountains frontal zone is indistinguishable, with the exception that north of the Choctaw thrust, foreland sedimentary rocks of the Krebs Group crop out (Suneson and Campbell, 1990). The stratigraphy of the area is summarized in Figure 2. South of the Choctaw thrust, in the frontal zone, the northwest-vergent imbricate thrusts involve mainly the Morrowan Wapanucka Limestone and the Atoka Formation. North of the Choctaw thrust, in the Arkoma Basin, the Atoka Formation and the Krebs Group are exposed.

The main structural features in the study area are the northwest-verging Choctaw thrust and the frontal Kiowa syncline (Fig. 3). In the southeastern part of the 3-D survey area, the Craig anticline and the Haileyville-Hartshorne syncline form an anticline-syncline pair parallel to the Choctaw thrust. These two folds are traced for ~15 km, whereas the Kiowa syncline is a major fold that can be traced for >40 km. The study area is a few kilometers southwest of the Wilburton gas field (Berry and Trumbly, 1968) which produces from Atoka sandstones, the Wapanucka Limestone, and Arbuckle Group carbonates (Tilford, 1990).

The Triangle Zone

Triangle zones are common structural elements at the leading edge of some fold-and-thrust belts. They are located between the last thrust and a frontal syncline in the foreland. This type of structure has been described in the Alberta foothills of the Rocky Mountains (Jones, 1982); Pakistan (Banks and Warburton, 1986); the west Mackenzie Mountains (Canada), Peru, and Alaska (Vann and others, 1986); and the Ouachita Mountains (Hardie, 1988). Jones (1982) was the first to thoroughly describe the main elements of a triangle zone.

Arbenz (1984) first illustrated this style of deformation in the frontal Ouachita thrust belt, in a schematic section across the thrust belt in western Arkansas, showing foreland-directed blind thrusting north of the Choctaw thrust. Hardie (1988) described in detail the transition of the frontal Ouachitas fold-and-thrust belt to the foreland Arkoma Basin and was the first to recognize the structural setting as a triangle zone. He studied the Pittsburg 7.5′ quadrangle, 25 km southwest of Hartshorne, in Pittsburg and Kiowa Counties. On the basis of surface and subsurface data, he concluded that the structural transition from the Ouachita Mountains into the Arkoma Basin is an incipient triangle zone and that a more mature and deeply eroded triangle zone exists farther south, within the frontal Ouachitas. He proposed two regional detachments: one at the base of the Springer Shale and the other at the base of the Woodford Shale.

Arbenz (1989) used seismic sections across the structural front of the Ouachita Mountains in Pittsburg County to illustrate a "well-developed" triangle zone with underthrust and upturned Atokan and Desmoinesian strata. Bertagne and Leising (1989) identified this

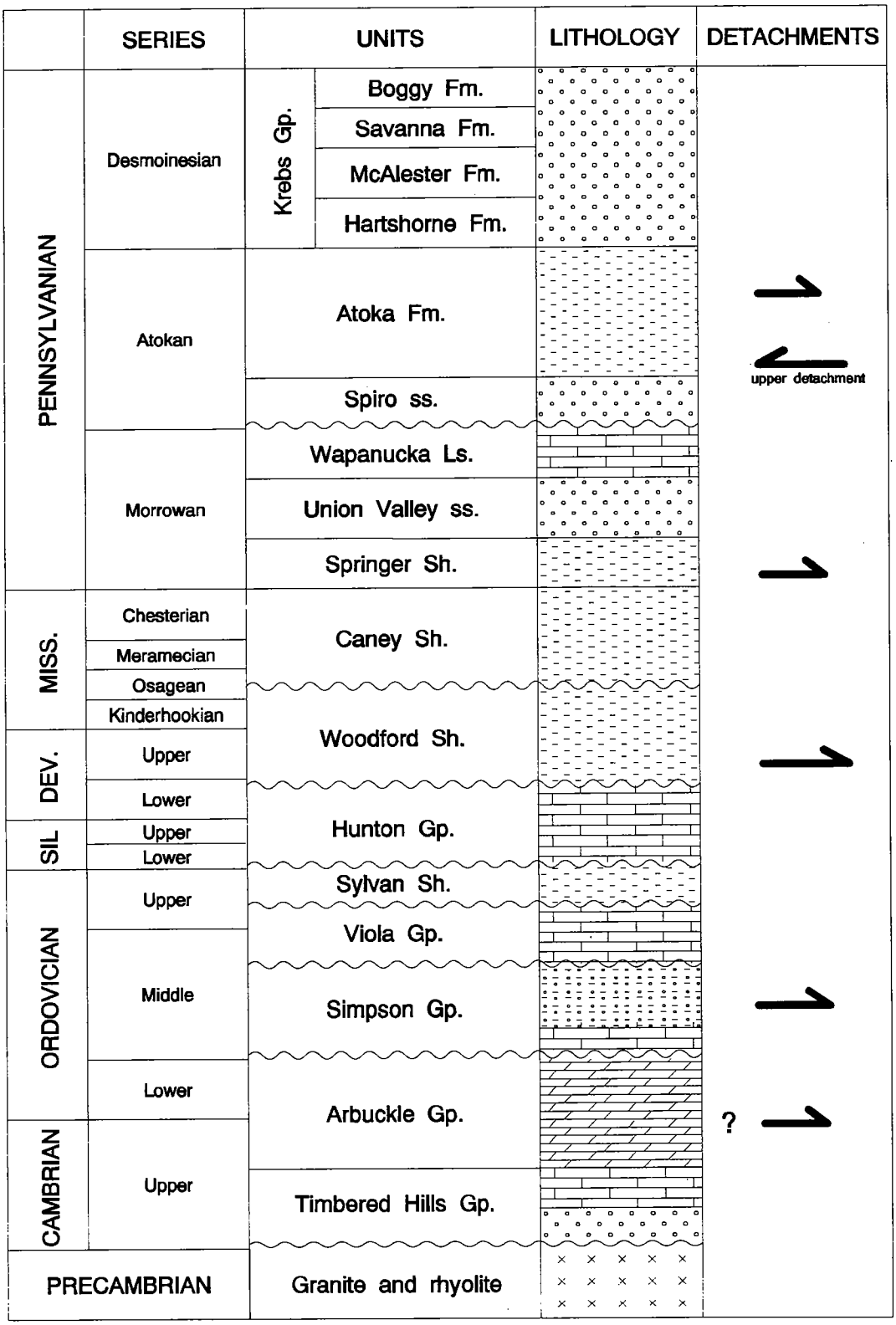


Figure 2. Stratigraphic column for the triangle zone of the Ouachita Mountains and Arkoma Basin (modified from Suneson and others, 1990) showing major detachments. The upper detachment or passive-roof thrust of the triangle zone is in the lower Atoka Formation and has opposite vergence.

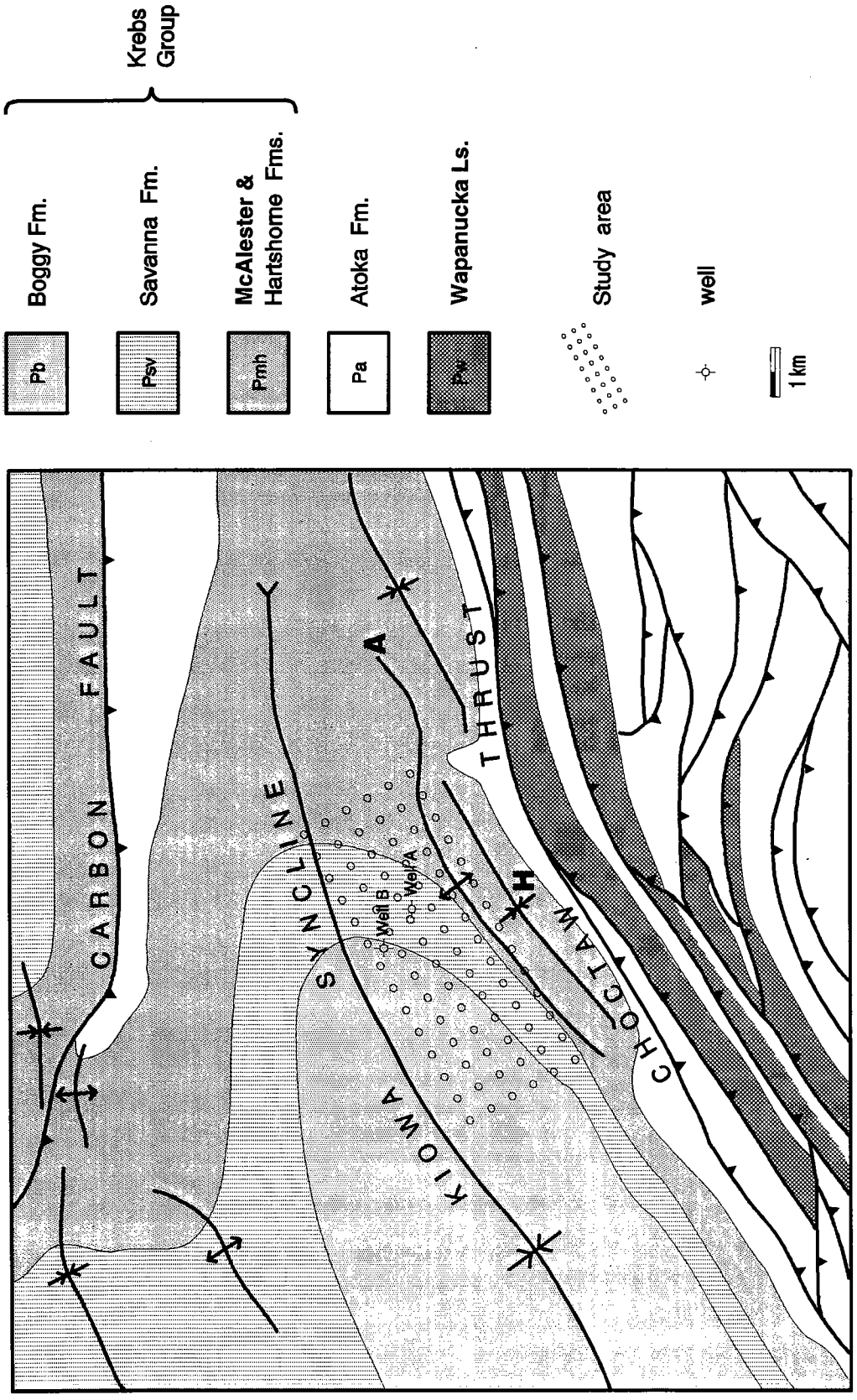


Figure 3. Geologic map of the study area showing the location of the 3-D seismic survey. A = Craig anticline, H = Haileyville syncline.

area as the "Ouachita frontal fairway" and described a structurally complex zone extending ~19 km north and south of the surface trace of the Choctaw thrust. They examined conventional 2-D seismic data and indicated that sideswipe is a common problem encountered in the "frontal fairway."

Tilford (1990) also used seismic and well data to describe the "wedge zone" as allochthonous, sometimes overturned, thrust sheets of Spiro sandstone and Wapanucka Limestone, bounded by Atokan and Morrowan shales. The "wedge zone" occurs in the interval between the overlying rocks of the Ouachita overthrust and the underlying autochthonous units of the Arkoma Basin. Tilford (1990) described the faulting in the wedge zone as imbricate, forward-directed sole thrust faults, additional southward-directed backthrusts, and antithetic relaxation faults. He also reported that some wells in the wedge zone exhibit overturned bedding, recognized by inversion of the Spiro-Wapanucka units. Suneson and Campbell (1990) recognized the Spiro-Wapanucka-Cromwell blind and imbricate thrust faults north of the Choctaw as a possible exploration play.

Perry and Suneson (1990) also examined a seismic-reflection profile across the frontal zone of the Ouachita thrust belt near Hartshorne and confirmed that it is a triangle zone. They concluded that the frontal thrusting is younger than the Hartshorne Formation of late Middle Pennsylvanian (Desmoinesian) age, the youngest Paleozoic unit preserved along the seismic profile. Perry and Agena (1992) indicated that the triangle zone and associated imbrication represent as much as 10 km of shortening.

Wilkerson and Wellman (1993) relied on seismic and well data to construct nine cross sections through the frontal zone in Latimer and Pittsburg Counties. They identified two blind thrust systems below the Choctaw thrust and named them the Gale and Buckeye systems. In addition, they reported transverse structures: a left-stepping zone of oblique ramps and a tear fault within the Gale thrust sheet. These transverse structures are spatially coincident with similar structural trends in lower Paleozoic strata beneath the thrust system as well as surface structural trends.

Middle Atokan Normal Faulting

Faulting of the Wapanucka Limestone has been traditionally interpreted as normal in 2-D seismic profiles (Wilkerson and Wellman, 1993; Reeves and others, 1990; Perry and Suneson, 1990; Bertagne and Leising, 1989). These faults were active during the middle Atokan and apparently were caused by vertical loading and flexural bending of the foreland crust related to a subduction zone to the south (Houseknecht, 1986). Tilford (1990) suggested that some of these faults were reactivated as reverse faults during thrust emplacement. Camp and Ratliff (1988) interpreted a gas-producing structure in the Arbuckle Group as the hanging wall of a high-angle, south-dipping reverse fault. On the basis of thickness changes of Pennsylvanian–lower Atokan sediments and by analogy to comparable extensional faults to the north, they suggested that this fault may be a reactivated normal fault. They also suggested that at least some of the movement on the Arbuckle reverse fault appears to postdate the overlying thrust-faulted structures.

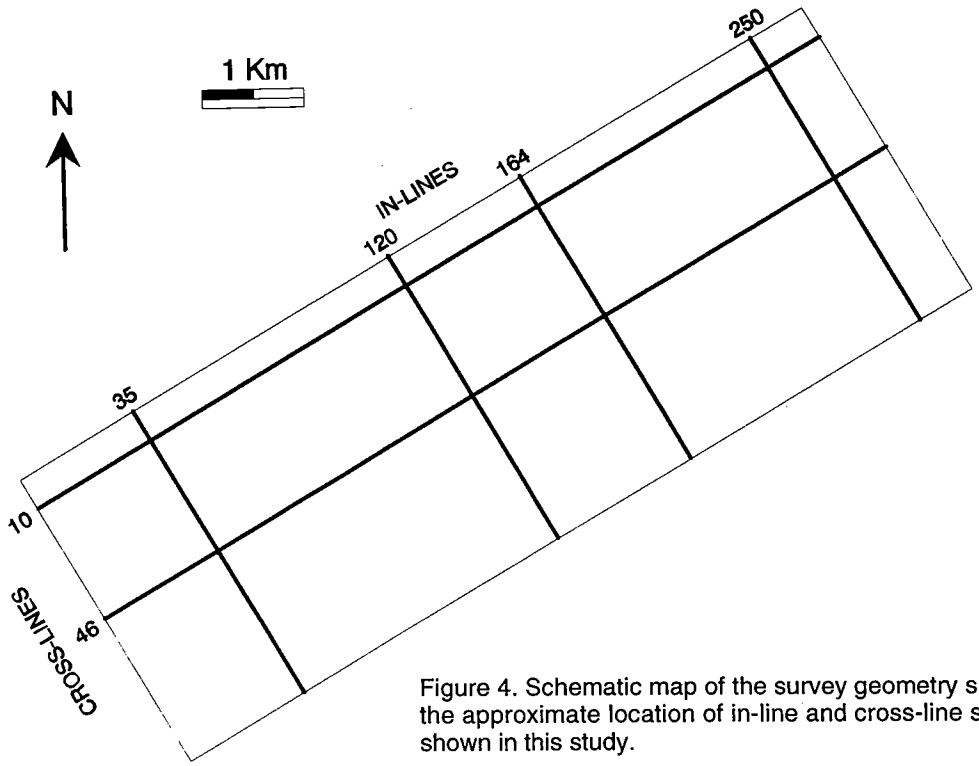


Figure 4. Schematic map of the survey geometry showing the approximate location of in-line and cross-line surveys shown in this study.

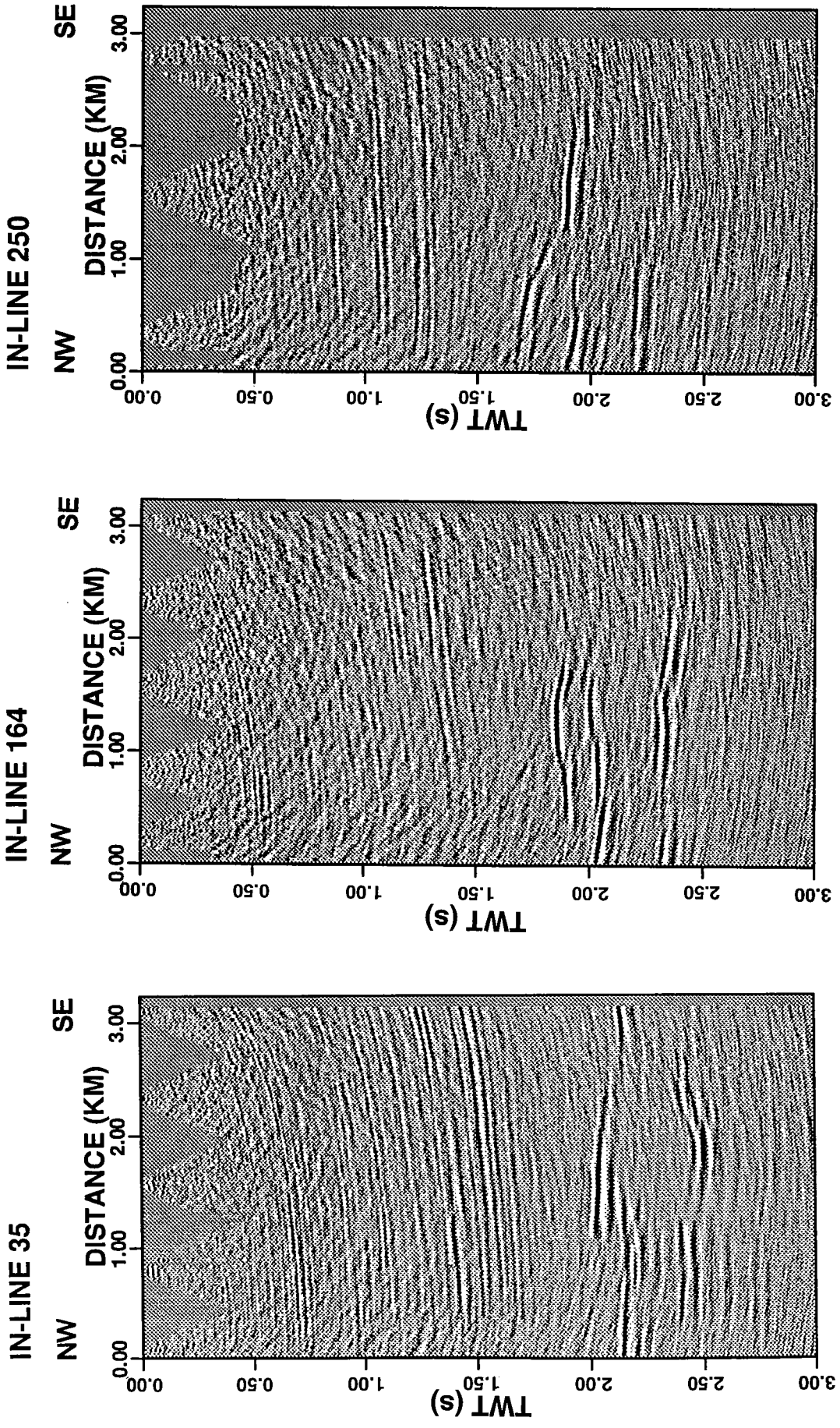


Figure 5. Variable-density displays of in-line profiles 35, 164, and 250. Approximately 1:1 horizontal to vertical scale.

THE 3-D SEISMIC DATA SET

The final migrated data set has 268 in-line profiles in the southwest-northeast direction (along the receiver-line direction) and 102 cross-line profiles in the southeast-northwest (along the shot-line direction), for a total of 27,336 bins (common cell gathers). The spacing between each bin is 32 m. The length in the cross-line direction is 8.54 km and in the in-line direction 3.33 km, for a total of 28.45 km^2 (Fig. 4). The data were acquired in eight-line swaths, each consisting of eight receiver lines, 256 m apart. Each swath has a three-line overlap with those on either side, for a total of 33 receiver lines. Each receiver line had 50 receiver groups; one every 64 m. Five source lines spaced an average of 800 m apart run approximately perpendicular to receiver lines; sources along each line are an average of 64 m apart (Brzostowski and McMechan, 1991).

The quality of the data varies according to the fold coverage. The maximum fold reached in the central part of the data was 21 and decreases toward the edges of the survey. There is an artifact produced by migration near the edges of the data. This is more evident in the in-line profile and may encourage spurious interpretations.

One of the main advantages of using a 3-D data volume is that the interpreter can select the slice that best exhibits a given feature. For example, most of the faults trend southwest-northeast and hence are difficult to pick on the cross-line direction, but are clearly seen on the in-line profiles. Thus, most of the faults were picked on the in-line profiles and later correlated on the cross-line profiles. Often these faults do not have an evident seismic expression on the cross-line profiles. There are several examples where some features are only clearly seen on a specific type of slice.

STRUCTURAL INTERPRETATION

Figure 5 presents three representative southeast-northwest in-line profiles. In-line profile 35 is to the southwest end of the survey area, and profile 250 is to the northeast end (Fig. 4). The upper 1.5 s two-way traveltime (TWT) of the sections present northwest-dipping parallel reflections from the southern limb of the Kiowa syncline. Most of these reflections correspond to the Atoka Formation and the lower part of the Krebs Group. The strong reflections between 2.0 and 2.2 s TWT in in-line profile 35 (Fig. 5) correspond to repeated Spiro-Wapanucka sections. In in-line profiles 164 and 250 (Fig. 5), the Wapanucka thrust sheet has more displacement and is easily distinguished from the footwall section. The other high-amplitude reflections between 2.2 and 2.5 s TWT correspond to the Hunton Group. The main structure seen in the data is the northwest-vergent fault-bend fold below the southern limb of the Kiowa syncline that repeats the Wapanucka Limestone.

In-line profile 164 (Fig. 6) crosses the site of well A (location shown in Fig. 3) that reached a first Spiro-Wapanucka contact at 3,380 m at 1.86 s TWT. A thrust fault was found at 3,654 m, and a second Spiro-Wapanucka contact was found at 3,663 m, which is the strong reflector at 2.0 s TWT. A second well (B), northwest of well A and toward the edge of the data (Fig. 3), was drilled to a total depth of 3,650 m. It remained in the Atoka Formation and never reached the Spiro-Wapanucka contact.

Reflector A (Fig. 6) is the the last clear reflection of the lower Atoka Formation that could be traced throughout the data set. Reflector A separates the undeformed limb of the Kiowa syncline from the more structurally complex zone underneath. The deformation in the Wapanucka Limestone exhibits the typical northwest-vergent thrusting of the frontal Ouachita Mountains. Three northwest-vergent thrusts offset the Wapanucka Limestone; they have been labeled, from top to bottom, w1, w2, and w3 (Fig. 6). The Wapanucka thrust sheet above thrust w1 has been displaced at least 1.6 km in this pro-

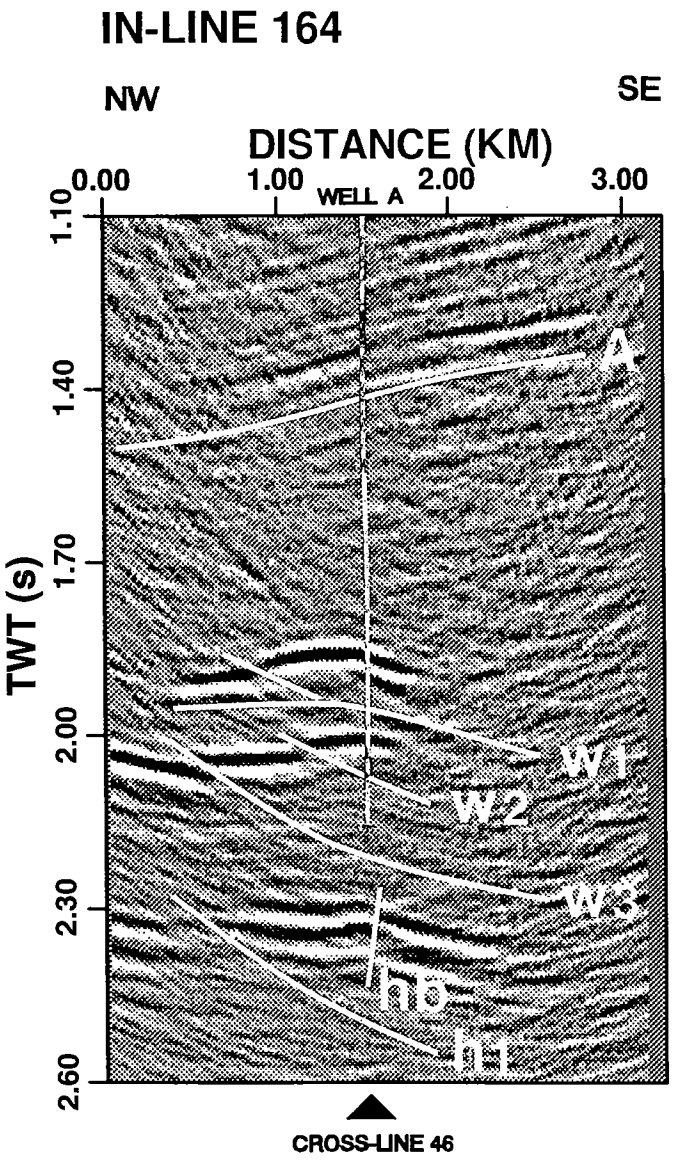


Figure 6. In-line profile 164 showing the main structural features of the study area. The two reflectors around 2.0 s TWT are two sections of Wapanucka Limestone. The strong reflector around 2.3 s TWT is the Hunton Group. Labeled features are discussed in the text.

file, whereas thrusts w2 and w3 record a limited shortening of a few tens of meters.

Neither of the wells in the area reached the strong reflector at 2.3 s TWT. The stratigraphic position and a good correlation with a structure map to the top of the Hunton Group (Wilkerson and Wellman, 1993) in an area adjacent to the survey area support the interpretation that this reflector is the Hunton. The Hunton Group shows a more complex faulting pattern including northwest-vergent thrusts (h1, h2, and h3) and a backthrust (hb; see Fig. 6).

Cross-line profile 46 (Fig. 7), from southwest to northeast, intersects in-line profile 164 at well A and illustrates the structural features in the strike direction. The axis of the Kiowa syncline plunges and flattens to the southwest. Above reflector A, parallel, continuous, and rare high-amplitude reflections are seen in the lower Atoka Formation. The area between the repeated Wapanucka section and reflector A lacks strong reflections, but some weak reflections parallel to A are observed.

The two sheets of the Wapanucka Limestone in Figure 7 are dipping to the southwest, similar to the strata of the overlying syncline; w1 is the main thrust that repeats the Wapanucka Limestone throughout most of the cross-line. Thrust w2, which shows minor displacement in the in-line profiles, has a very limited extent and dies out laterally. Thrust w3 is interpreted to have significant lateral extent but does not exhibit any apparent displacement in the cross-line profiles. The interpretation of w3 was done primarily on the in-line profiles and then correlated to the cross-line sections.

Other thrusts that affect the Hunton Group are not apparent in in-line profile 164, but are clearly seen in cross-line profile 46 (labeled from top to bottom h1, h2, and h3 in Fig 7). Thrust h1 does not show any apparent displacement, whereas thrust h2 not only affects the Hunton Group, but also the lower Wapanucka Limestone. In cross-line profile 46, these structures are also plunging to the southwest, and the two sections of the Wapanucka Limestone and Hunton Group appear to be folded in the same way.

In-line profile 120 (Fig. 8) is 1,450 m southwest of in-line profile 164 and displays the same structural features. It shows the faults affecting the Hunton Group: three northwest-vergent thrusts (h1, h2, and h3) and a backthrust (hb). The thrust faults and backthrust are not very apparent in this section, but the interpretation of the whole survey shows that the Hunton Group is repeated; therefore, the only plausible explanation is that these faults are reverse. The faulting in the two sections of

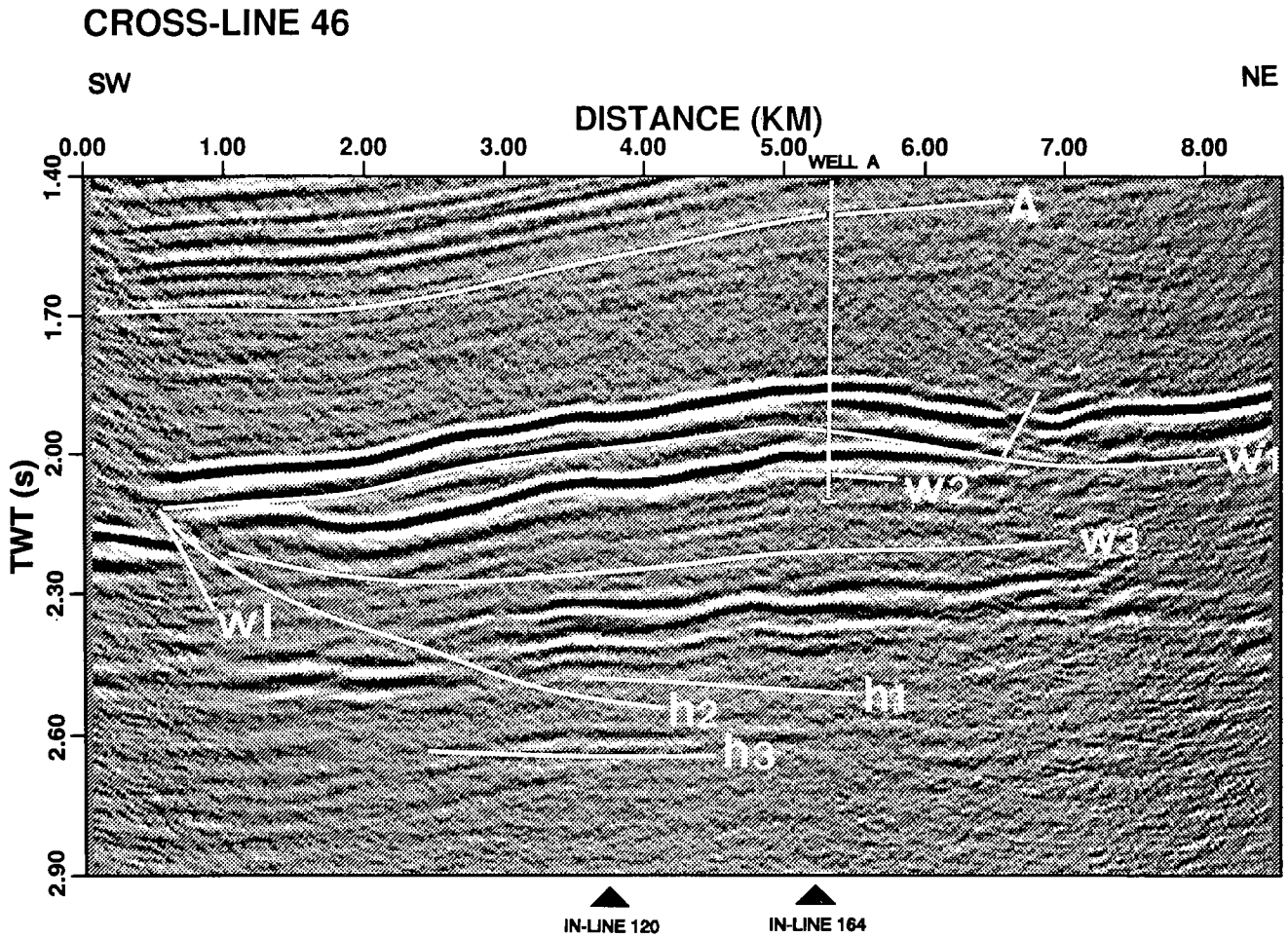


Figure 7. Cross-line profile 46 illustrating the main reflectors in a strike view. Labeled features are discussed in the text.

Wapanucka is similar to in-line profile 164, except that thrust w2 is not present. Note that thrust w3 folds the overlying section, including the Wapanucka thrust sheet and w1.

One of the advantages of 3-D seismic data is that time slices provide a view very much like a geologic map. The principal reflectors can be mapped in sequential sections to outline structures in three dimensions. The position of a reflector in successive time slices is related to its dip. The higher the dip, the narrower the reflector will appear; conversely, the lower the dip, the thicker it will appear. Variations in reflection amplitude can, however, have the same effect as dip changes in a constant-amplitude reflector (Brown, 1991).

Three time slices (Fig. 9) illustrate these structures in map view. The time slice at 1,860 ms TWT shows the Wapanucka thrust sheet forming an anticline bearing S. 40° W. The time slice at 2,000 ms TWT shows that the Wapanucka thrust sheet has moved to the southwest, and the footwall Wapanucka emerges in the core of the anticline. The thrust that separates the reflections is also folded, suggesting that the formation of this anticline postdates the thrusting that repeats the Wapanucka Limestone.

At 2,000 ms TWT, the azimuth of the anticline is S. 65° W., more westerly directed. This change in azimuth may be the result of differential transport of the thrust sheet, which can be observed in the three in-line profiles of Figure 5. Thrust w3 offsets the footwall Wapanucka and has a relatively straight southwest-northeast trend suggesting that it is younger and responsible, in part, for the folding of the overlying anticline. The time slice at 2,400 ms TWT at the Hunton level is not as clear as the shallower slices, but serves to illustrate the trend of the northwest-vergent thrusts and the backthrust.

The reflection from the backlimb of the fault-bend fold has very low amplitudes or is absent, probably owing to the increased convexity that defocuses the energy or to the steeper dip that reflected the energy out of the geophone array.

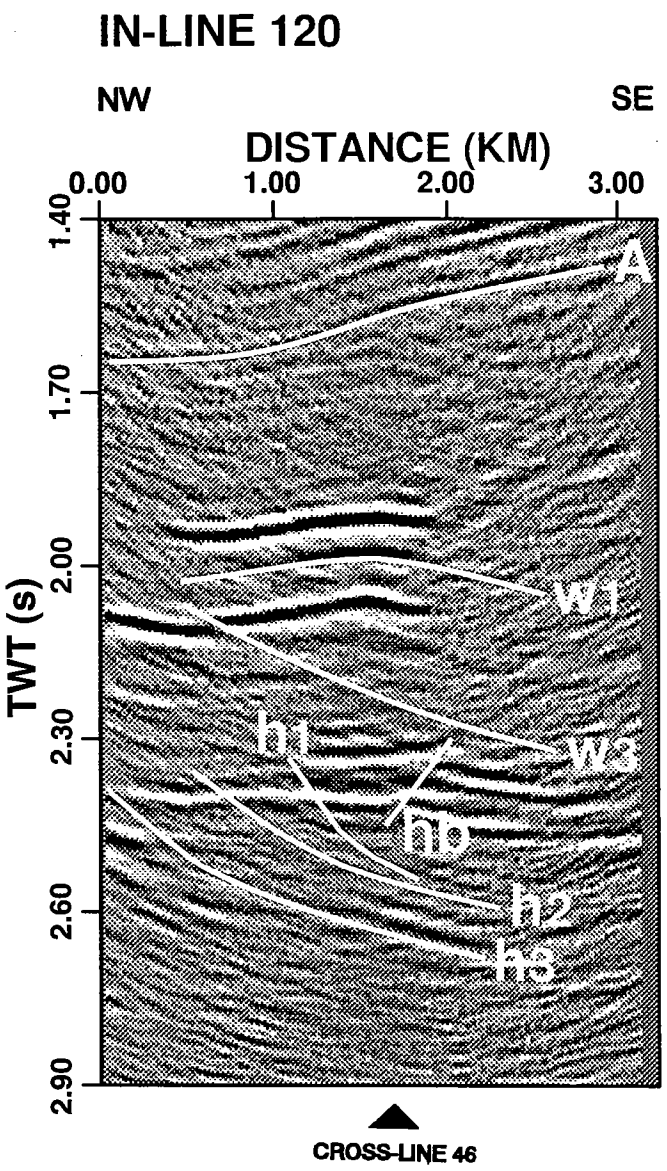


Figure 8. In-line profile 120 illustrating the interpretation of the faulting in the Hunton Group. Labeled features are discussed in the text.

Upper Detachment

The triangle-zone model (Jones, 1982) requires the presence of an upper detachment or "passive roof thrust" (Banks and Warburton, 1986) that separates strata of the frontal syncline from the underlying blind thrusts. The upper detachment is the zone where the duplexes of the triangle zone merge with the overlying syncline. This upper detachment must be present in the area, but does not have a clear seismic expression, probably because it is parallel to the bedding of the syncline and is localized within shales that do not have any significant acoustic impedance contrasts that would generate a reflection. The upper detachment of the triangle zone should be below this reflection and parallel to the bedding of the overlying syncline.

Thrusting in the Wapanucka Limestone

The most prominent structure in the data is the fault-bend fold (Suppe, 1983) that repeats the Wapanucka Limestone. The displacement on the underlying thrust w1 progressively increases from southwest to northeast; it varies from ~700 m in in-line profile 35 to ~1.6 km in in-line profile 164, and it is >1.6 km in in-line profile 250 since the leading edge is out of the data (Fig. 5). The increase in displacement along strike is also accompanied by an increase in the ramp angle to the northeast.

The detachment for thrust w1 is located ~60 ms TWT below the hanging wall Wapanucka Limestone in the Springer Shale. The calculated dip angles for w1 vary from ~11° in in-line profile 164 to 20° in in-line profile 250. The detachment for w2 is also in the Springer Shale, whereas w3 has a detachment above the Hunton, in the Woodford Shale (see in-line profile 120, Fig. 8).

Thrusts w2 and w3 have much less displacement; the vertical throw of these faults rarely exceeds 20 ms TWT (~40 m). Their average dip angle measured along the in-line profiles is ~15°. Thrust w3 is fairly continuous, but

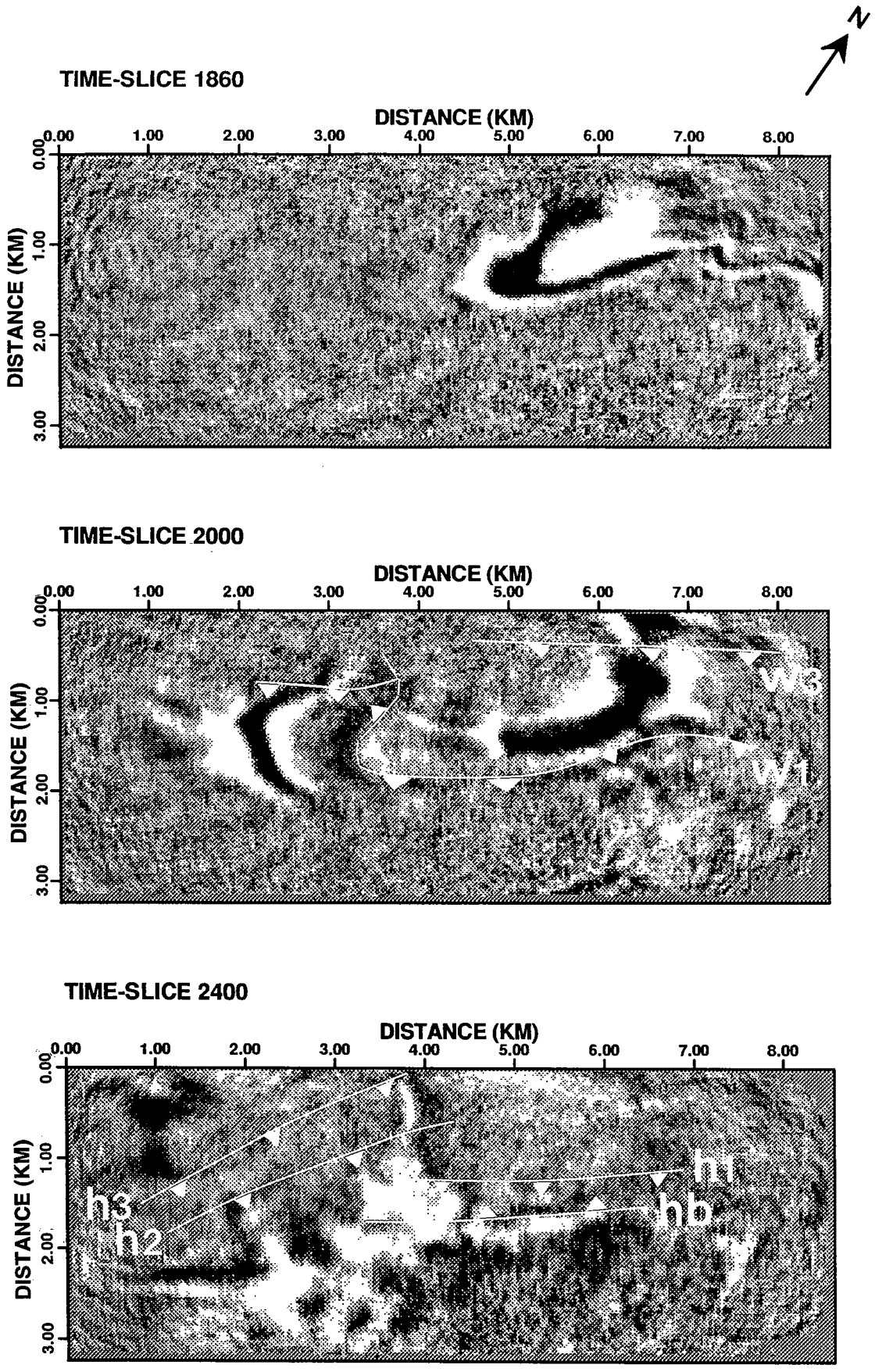


Figure 9. Time slices at 1,860, 2,000, and 2,160 s TWT. See text for discussion.

w2 is of very limited extent and dies out laterally. It is likely that there are more of these small faults in the lower Wapanucka Limestone, but they were not mapped. The number of smaller faults increases exponentially according to the fractal (scale-invariant) nature of faulting (Yielding and others, 1992).

Hunton Faults in the Hunton Group

The faults that cut the Hunton Group have been interpreted as reverse. In-line profile 120 (Fig. 8) shows the three main northwest-vergent thrusts and a backthrust. The detachments are probably in the Simpson Group for h1 and h2 and in the Arbuckle Group or deeper for h3. The interpretation of these faults as reverse is not obvious in many in-line profiles, but the interpretation of the whole survey shows that there are repeated sections of the Hunton. The reverse motion of these faults is seen more clearly in some cross-line profiles. Cross-line profile 10, on the northwest side of the survey, shows clearly the reverse motion of faults h3 and h1 (Fig. 10). The offset of these faults is not great; therefore, the higher resolution of 3-D seismic data is required to identify them.

Fault h3 is reverse in cross-line profile 10, but at in-line profile 35 (Fig. 11), 2 km to the southwest, it is interpreted as a normal fault. This evidence suggests that these faults, interpreted as reverse, may have been normal faults reactivated during the Ouachita orogeny, as postulated by Tilford (1990). Fault h3 reveals normal separation at the southwest end of the survey where the overlying Wapanucka thrust w1 has the least displacement, whereas at the northeast end of the survey, the Hunton faults have reverse separation where displacement on the Wapanucka thrust w1 is greatest. These variations suggest some connection between the amount of compression and the reactivation of the normal faults.

Lateral Ramps

At the southwest end of cross-line profile 46 (Fig. 7), the h2 and w3 thrusts merge at w1 and cut the lower Wapanucka. Coincidentally, the hanging-wall Wapanucka seems to end at this point. Southwest of w1, the lower Wapanucka Limestone seems to be undisturbed. Fault w1 appears to have determined the location of the lateral ramps of h2 and w3 (and possibly the emplacement the Wapanucka thrust w1).

Fault w1 is seen in cross-line profile 46 as a reverse fault and in cross-line profile 10 (Fig. 10) as a normal fault. This change in fault-plane dip may have several explanations. One possibility is that it was originally a normal fault, but was reactivated as a reverse fault during

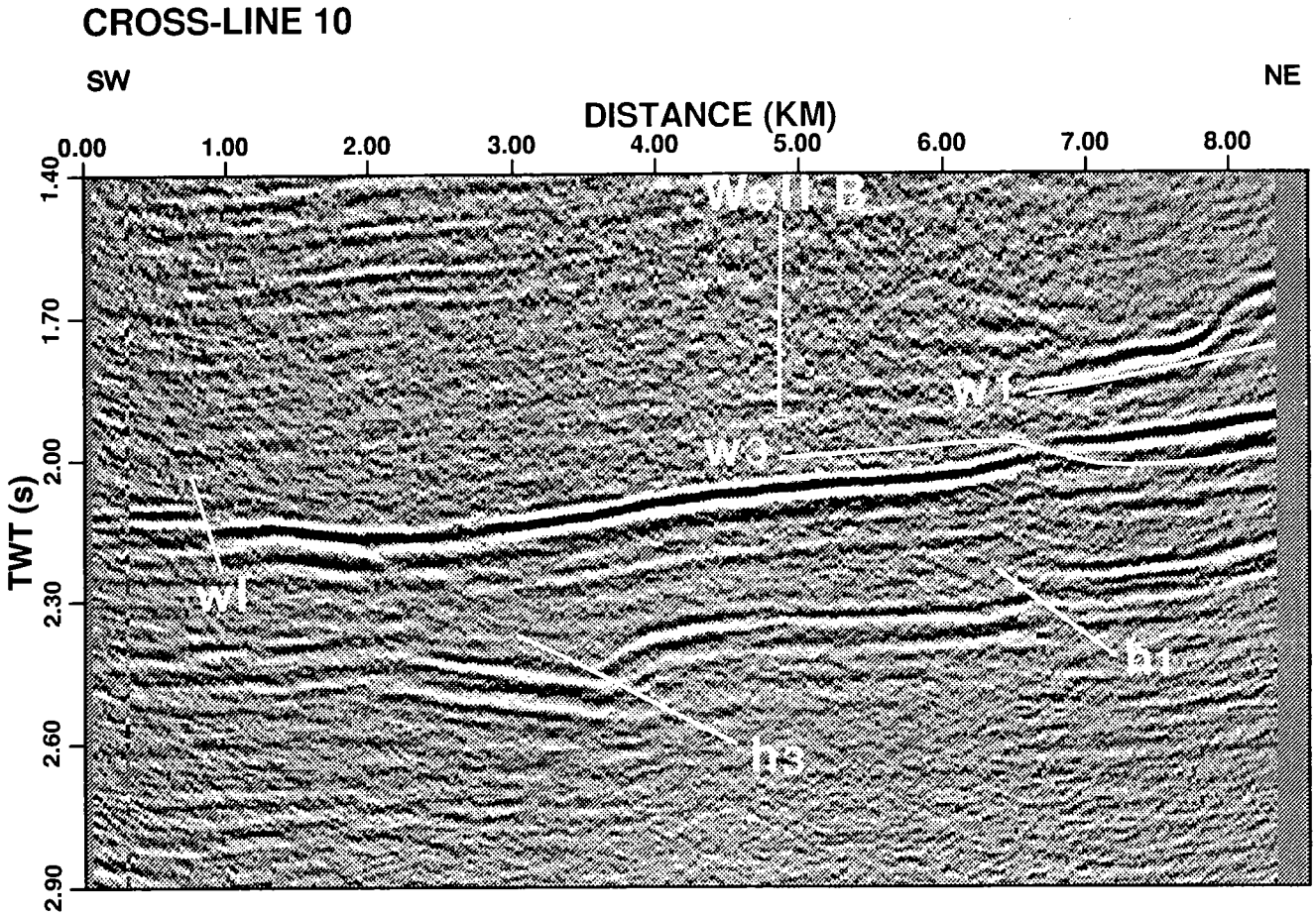


Figure 10. Cross-line profile 10 illustrating the reverse faulting of Hunton Group in strike view.

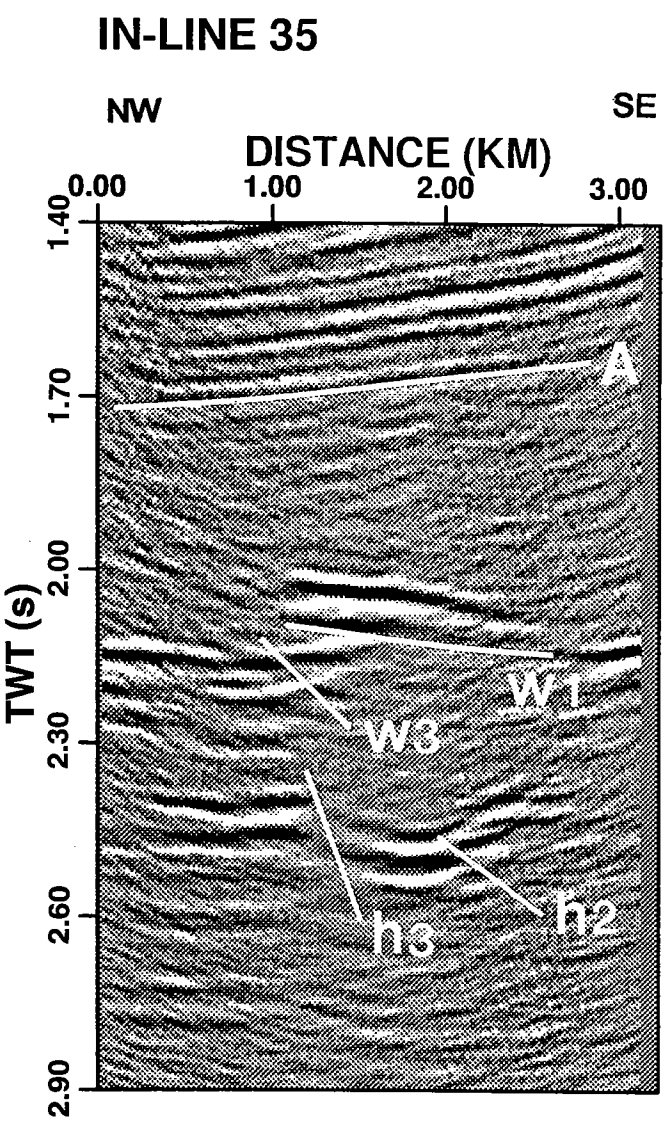


Figure 11. In-line profile 35 showing h3 as a normal fault.

the Ouachita orogeny. Another interpretation is that the change in dip of the fault plane is a product of rotation along a tear fault. In any case, this fault seems to have been formed prior to the thrusts and to have acted as a buttress during compression. It is not clear whether this fault cuts through the underlying Hunton Group.

3-D SUBSURFACE STRUCTURAL GEOMETRIES

Figure 12A is an amplitude and time-contour map for the hanging-wall Wapanucka thrust sheet. The variations in reflector amplitude are attributed mainly to reflector curvature. Higher amplitudes are associated with the axial part of the fold where the dip is gentle. The amplitude of the backlimb of the fold is very low, where the reflector is more convex. Minor variations in amplitude can be related to lithologic variations or fluid content. The structurally higher part of the fold is located in the north part of the area at time 1,640 ms TWT (~3,060 m), and the lower part is at ~2,060 ms TWT (~3,820 m).

The amplitude-horizon map for the footwall Wapanucka Limestone (Fig. 12B) shows a very good correlation between blocks of different amplitudes and the traces of the faults. The higher part of the unit is located in the northeastern part of the area. The highest time contour is 1,925 ms TWT, which corresponds to ~3,580 m, and the lowest is at 2,150 ms TWT (3,980 m). Faults H1 and H2 in the amplitude-horizon map of the Hunton Group (Fig. 12C) separate the southwest-trending, higher-amplitude hanging-wall block from the lower amplitudes to the northwest. The highest part is again in the northeastern corner of the area, from 2,180 ms TWT, which corresponds roughly to 4,040 m, and the lowest is at 2,440 ms TWT, which is ~4,510 m.

Figure 13 shows perspective views of the surfaces of reflector A, the Wapanucka thrust sheet, the lower Wapanucka Limestone and the Hunton Group. This view is from the south-southwest and at an angle of 30°. The similar surface geometries not only suggest that these surfaces were involved in the same folding, but also that their original configurations were more or less parallel. It is clear from this perspective view that the southern limb of the Kiowa syncline (reflector A) has been uplifted by underlying thrusts.

From the time-contour maps and with a depth-conversion table, the bearing and plunge of the fault-bend fold can be calculated. Figure 14 is a stereonet with these fold-axis data for the hanging-wall and footwall Wapanucka Limestone and for the Hunton Group. The trend and plunge of the Kiowa syncline were measured from a geologic map and are shown for reference. The three subsurface features plunge 2°–10° to the southwest. The average trend and plunge of the Kiowa syncline are 243° and 14°, respectively. This similarity indicates that the Kiowa syncline and the three features in the Wapanucka and Hunton were involved in the same folding event. The scatter of the fold-trend data may be related to differential transport along the thrusts.

SEQUENCE OF DEFORMATION

A sequence of deformational events based on the interpretations above is proposed. The deformation began in the foreland basin when the sub-Atoka sedimentary section was involved in normal faulting (Houseknecht, 1986; Sutherland, 1988). The first thrust to develop was w1, which formed as a fault-bend fold with its detachment in the Springer Shale. This deformation initiated the passive uplift of the overlying Atokan and younger section. Thrusts w2 and w3 followed, slightly folding the overlying Wapanucka section. The detachments for these thrusts are in the Springer and lower part of the Woodford Shales, respectively.

The normal faults that offset the Hunton Group and deeper strata were reactivated as reverse faults during the later stages of thrust emplacement. These faults show a greater reverse displacement where the overlying Wapanucka thrust has more displacement. They are still normal, or the reactivation is less to the southwest, where the Wapanucka thrust w1 has less displacement. This observation suggests that there is some direct corre-

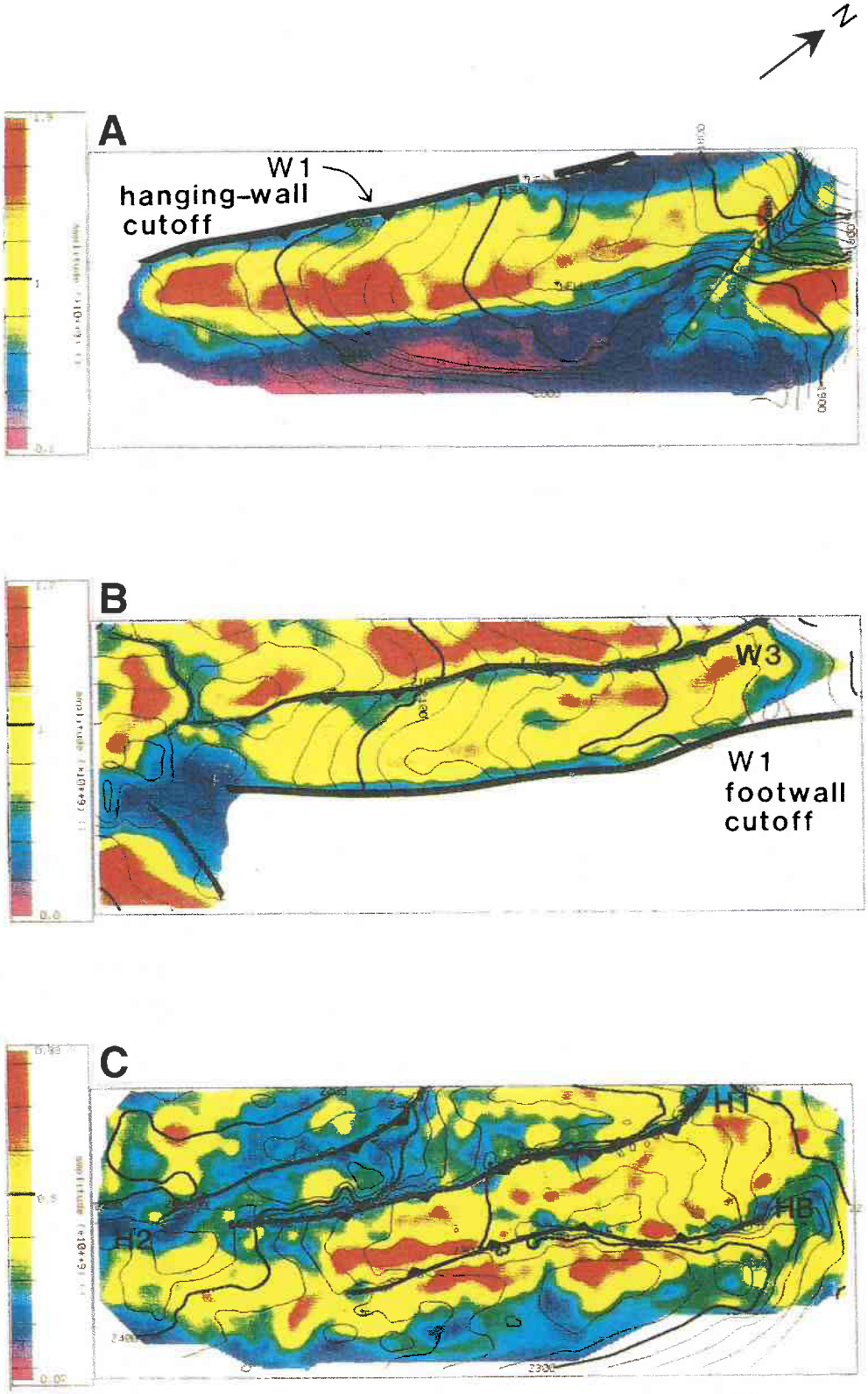
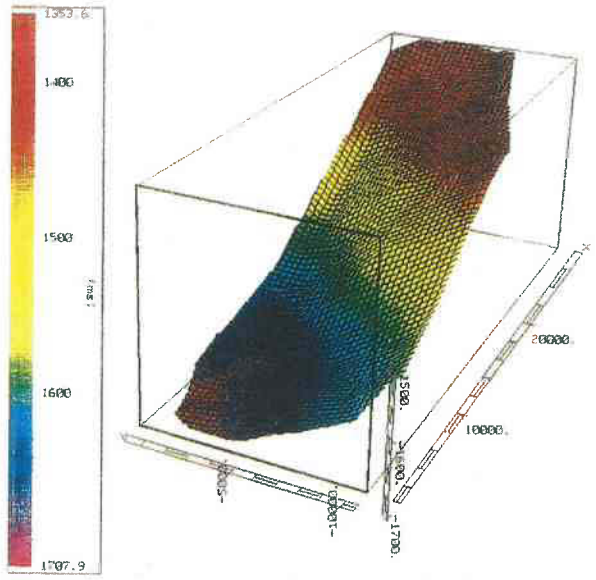
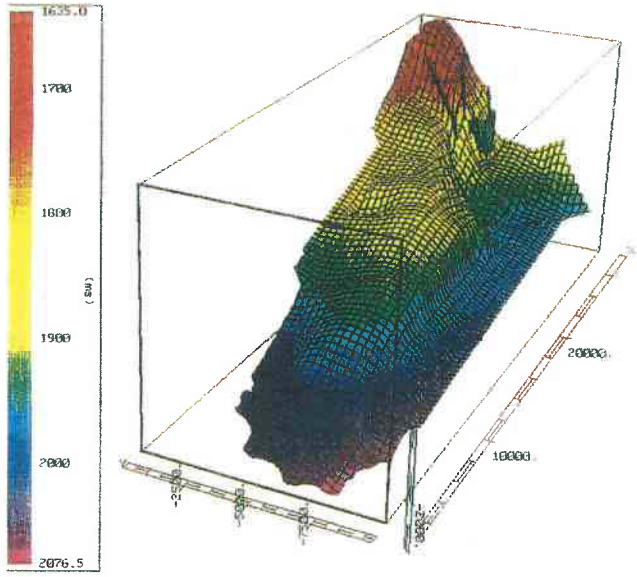


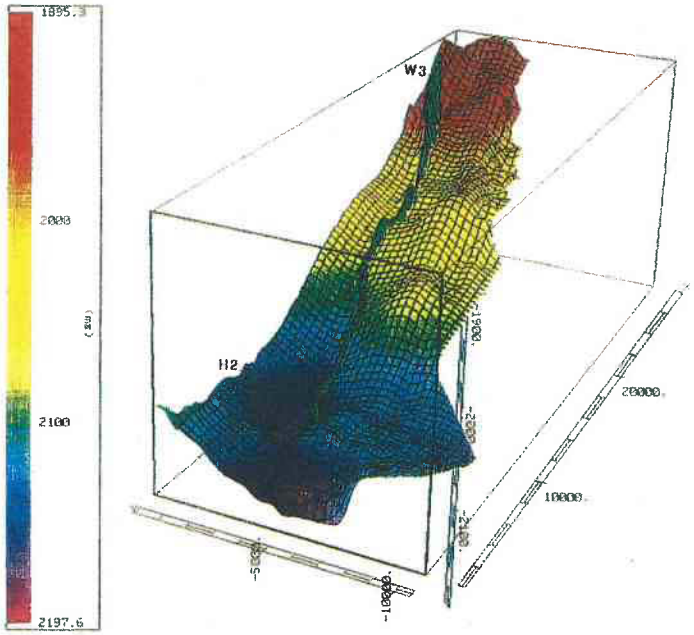
Figure 12. Amplitude and time-contour maps of (*A*) the Wapanucka thrust sheet, (*B*) the footwall Wapanucka, and (*C*) the Hunton Group.



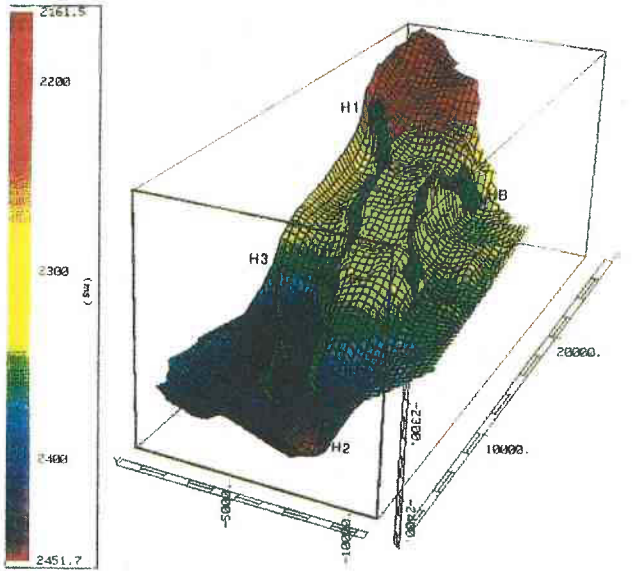
REFLECTOR A



HANGING-WALL WAPANUCKA



FOOTWALL WAPANUCKA



HUNTON GROUP

Figure 13. Three-dimensional perspective views of the four reflectors interpreted from the south at an angle of 30°.

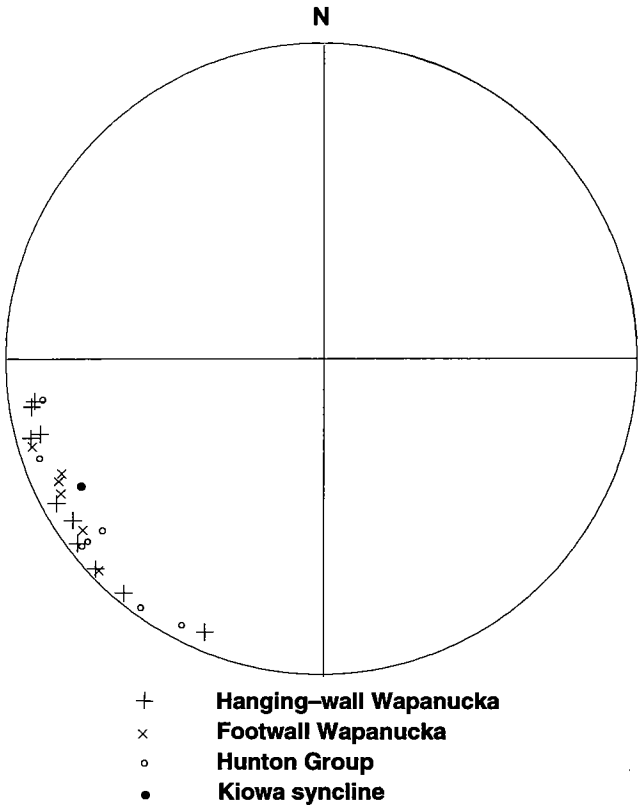


Figure 14. Stereonet of fold-axis data.

lation between the Wapanucka thrusting and the reactivation of the Hunton faults. The Hunton faults appear to have been reactivated when the Wapanucka thrust sheet was emplaced; however, the Hunton faults continued moving and subsequently folded the entire overlying sequence.

Faulting in the Hunton Group and deeper strata has been traditionally interpreted as normal (Wilkerson and Wellman, 1993; Suneson and others, 1990; Bertagne and Leising, 1989; Perry and Suneson, 1990). The conclusions drawn from this study agree with those of Tilford (1990), who proposed that some of the normal faults were reactivated as reverse faults during the Ouachita orogeny. Camp and Ratliff (1988) generated a balanced cross section through the Wilburton field and interpreted a deep gas-productive Arbuckle structure as a high-angle, south-dipping reverse fault. They also suggested that at least some movement on the reverse fault appears to postdate the overlying thrusts in the Wapanucka Limestone. The observations in this study agree with the conclusions of Camp and Ratliff (1988) that some of the deeper faults are reverse and postdate the overlying Wapanucka thrust.

Figure 15 is a southeast-northwest schematic cross section through the triangle zone between the Ouachita Mountains and Arkoma Basin that shows the location of the study area and its relation to the adjacent structures. To the southeast, under the Choctaw thrust, there is evidence for several additional imbrications of the Wapa-

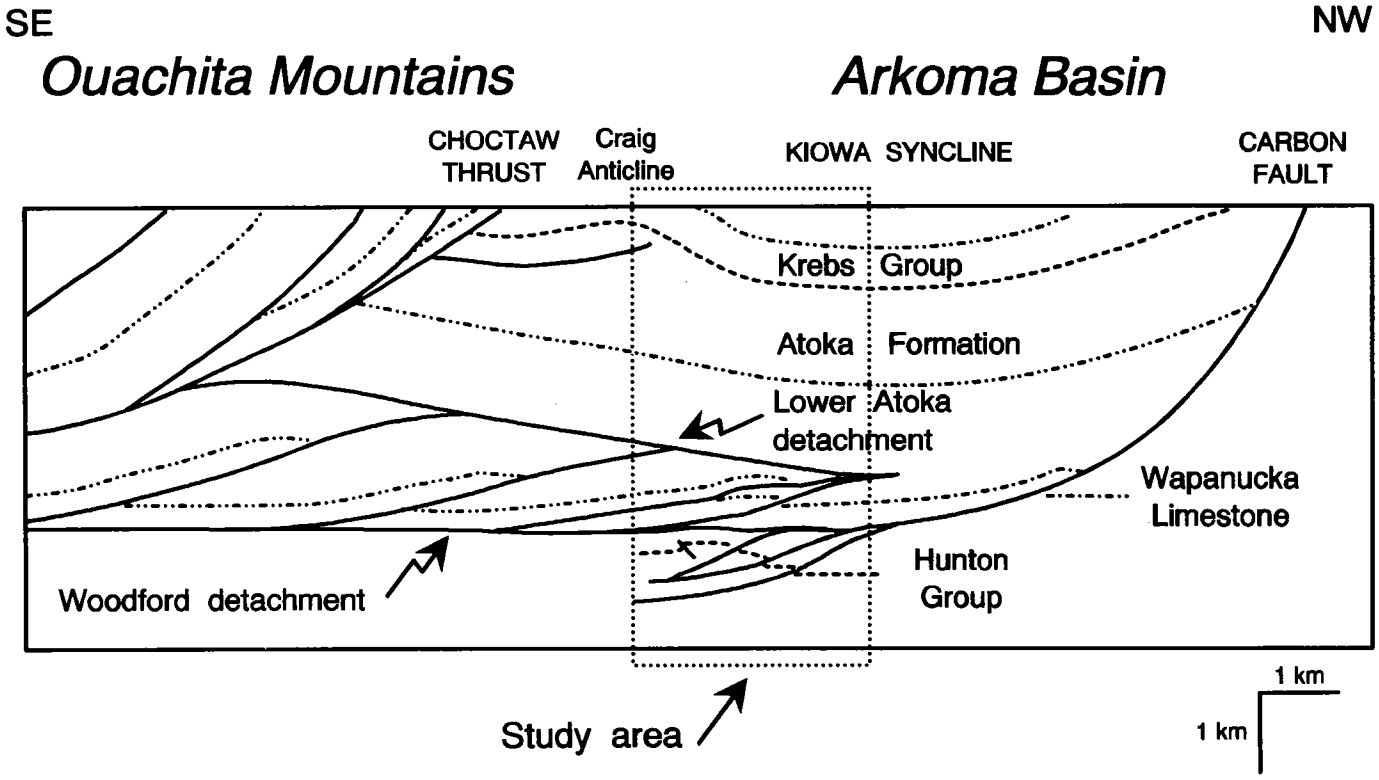


Figure 15. Schematic cross section through the triangle zone of the Ouachita Mountains and Arkoma Basin, showing the study area.

nucka Limestone (Wilkerson and Wellman, 1993). The thrusts in the Wapanucka Limestone in the study area are interpreted to merge with the upper detachment of the triangle zone, and the thrusting in the Hunton Group is interpreted to be related to the Carbon fault to the north.

The triangle-zone model explains the uplift of the limb of the frontal syncline by the emplacement of duplexes below, but in this area, between the Kiowa syncline and the Choctaw thrust, there is an anticline-syncline pair with a wavelength of ~2 km (Fig. 3). The Craig anticline is interpreted to be produced by a reverse fault in the Atoka Formation. This structure is interpreted as an earlier structure that was subsequently folded by the underlying duplexes, forming the Haileyville syncline on the backlimb of the Craig anticline. It is not likely that this anticline is related to the underlying duplexes.

Perry and Suneson (1990) and Reeves and others (1990) published two seismic sections near this area showing small triangle zones with duplexes involving only strata of the Atoka Formation. This study suggests that several triangle zones at different stratigraphic levels with independent lower and upper detachments are present: the main triangle zone that uplifts the frontal syncline occurs at the Wapanucka level, and smaller triangle zones in the upper Atoka Formation are responsible for the smaller-scale folds.

The evidence presented in this study suggests that compression during the Ouachita orogeny affected deeper levels than generally recognized. The high resolution of this 3-D survey has permitted the recognition of reverse faulting below the Wapanucka Limestone and has helped constrain the timing of the events.

CONCLUSIONS

The interpretation of this 3-D survey has provided a detailed structural picture of a part of the triangle zone of the frontal Ouachitas, previously unattainable with 2-D seismic and well data. This area is a triangle zone (Jones, 1982) with northwest-vergent blind thrusting under the frontal Kiowa syncline. The southern limb of the Kiowa syncline has been passively uplifted by the blind thrusting at deeper levels. There are three levels of thrusting involving the Wapanucka Limestone, the Hunton Group, and (with smaller offsets) the Atoka Formation.

There are two different types of thrusting in the Wapanucka Limestone. The first thrust to form, w1, records significant transport with at least 1.5 km of displacement. The later thrusts affecting the Wapanucka Limestone, w2 and w3, have displacements on the order of tens of meters. The detachments for the thrusts offsetting the Wapanucka Limestone are localized in the Springer Shale and in the lower part of the Woodford Shale. The thrusts affecting the Hunton Group have two detachments in the Simpson and Arbuckle Groups.

The 3-D seismic data clearly show that faulting in the Hunton Group is mainly reverse. Three northwest-vergent thrusts and a backthrust were identified in the Hunton Group. Their displacements range from tens of meters to as much as a few hundred meters. The location of these thrusts may have been controlled by prior normal faults. For h1 and h2, the detachment is interpreted to be in the Simpson Group, and for h3 it appears to be in the Arbuckle Group, possibly even deeper.

The amplitude-horizon maps are consistent with the interpretation of the traces of the main faults and may used as an interpretation tool. The time-contour maps can be converted to depth with a time-depth conversion chart and used to calculate the plunge of the folds.

The trend of the Kiowa syncline, and of folds within the hanging-wall and footwall Wapanucka Limestone sections and within the Hunton Group, is southwesterly with an average plunge of 5°. The geometric correspondence of these four features suggests that they were affected by the same compressional events. The youngest structures appear to be at the deeper levels, in the Hunton Group, and they fold the overlying sequence. This reverse faulting in the Hunton Group may be related to the Carbon fault, which crops out a few kilometers north of the area.

Faulting under the Woodford Shale has been traditionally interpreted in 2-D seismic profiles as normal, with minor reverse motion. Although basement-involved reverse faults, like the Carbon fault, have been mapped (Tilford, 1990), they have not been well documented. The 3-D data have a much higher resolution than 2-D seismic data and clearly illustrate that compression during the Ouachita orogeny was widespread and affected levels as deep as the Hunton Group. It is possible that this could have been a mechanism for fold development in other parts of the Arkoma Basin.

ACKNOWLEDGMENTS

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Map-Scale Fold in the Ti Valley Thrust Sheet of the Ouachita Fold-and-Thrust Belt, Latimer and Le Flore Counties, Oklahoma

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Jay Namson
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ABSTRACT.—A map-scale fold is expressed on the surface in the hanging wall of the Ti Valley fault in the Ouachita Mountains of southeast Oklahoma. The Jackfork Sandstone is the principal stratigraphic unit involved in the structure that formed in response to a north-verging thrust fault ramping up-section from a lower detachment. This anticline is informally named the Baker Mountain structure and overlies a blind duplex formed on the hanging wall of the Ti Valley thrust sheet. The geometric components of the fold include a south-dipping backlimb, faulted anticlinal crest, north-dipping forelimb, forelimb syncline, and a hanging-wall cutoff. All these features are distinguishable in the surface geology. Exploration opportunities in the Ti Valley thrust sheet include the Jackfork Sandstone on the Baker Mountain structure and in the underlying Ti Valley duplex. Production has been established in several recently drilled wells in the 5,400-ft-thick Jackfork Sandstone section; however, the exploration potential remains largely untapped.

INTRODUCTION

The area of interest is located in the Ouachita fold-and-thrust belt of southeastern Latimer and westernmost Le Flore Counties, Oklahoma (Fig. 1). This region is in the Ouachita Mountains and is characterized by steep, rugged terrain. A large, map-scale fold is exposed at the surface in the hanging wall of the Ti Valley fault. The fold is generated by a north-verging thrust fault ramping up-section through the Jackfork Sandstone. The geometric components of the structure can be traced in the surface geology along strike for ~20 mi. The fold consists of a steeply south-dipping backlimb, a faulted anticlinal crest, a forelimb syncline, and a hanging-wall cutoff. The stratigraphic units involved in the fold and exposed at the surface include the Jackfork Sandstone, Johns Valley Shale, and undifferentiated Atoka Formation (Fig. 2). A geologic cross section has been constructed utilizing detailed surface geology, seismic-reflection data, and limited subsurface geologic data.

BAKER MOUNTAIN STRUCTURE

The Baker Mountain structure (named here) is located in the footwall of the Windingstair fault and the hanging wall of the Baker Mountain fault (also named here). The Baker Mountain fault is an imbricate thrust of the Ti Valley fault and juxtaposes Mississippian–Pennsylvanian Jackfork Sandstone in the hanging wall opposite Morrowan–Atokan units in the footwall (Fig. 3). This is the northernmost occurrence of the Jackfork Sandstone in the Ouachita fold-and-thrust belt. Early workers (summarized by Suneson, 1988) have considered the Ti Valley fault to be the boundary between Arbuckle facies and Ouachita facies rocks. Although the Ti Valley fault does mark a change in regional stratigraphic correlations (Suneson, 1988, p. 33), recent drilling has indicated that Arbuckle facies rocks of Cambrian through Ordovician age (Viola, Simpson, and Arbuckle Groups) exist in the Ti Valley footwall below a bedding-plane detachment. This low-angle detachment is also the posi-

Cunningham, R. D.; and Namson, Jay, 1994, Map-scale fold in the Ti Valley thrust sheet of the Ouachita fold-and-thrust belt, Latimer and Le Flore Counties, Oklahoma, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 243–248.

tion of the divergence of the Ti Valley fault and the Choctaw fault.

The Baker Mountain structure is interpreted to be a fault-propagation fold formed by the Baker Mountain thrust fault ramping up-section through the Jackfork Sandstone. A fault-propagation fold consumes all the slip in the folding of the strata, and no slip is translated out of the structure. Fault-propagation folds have a minimum amount of shortening and are important hydrocarbon traps in fold-and-thrust belts (Mitra, 1990). The Baker Mountain fault is rooted in the Ti Valley fault at a deeper detachment located to the south of its surface exposure, and it cuts off a duplex structure formed by imbricate faults in the hanging wall of the Ti Valley fault. The Baker Mountain structure consists of a backlimb dipping 70° to the south near the surface expression of the Windingstair fault. The structural crest is cut by minor imbricate faults. The Jackfork Sandstone forms the

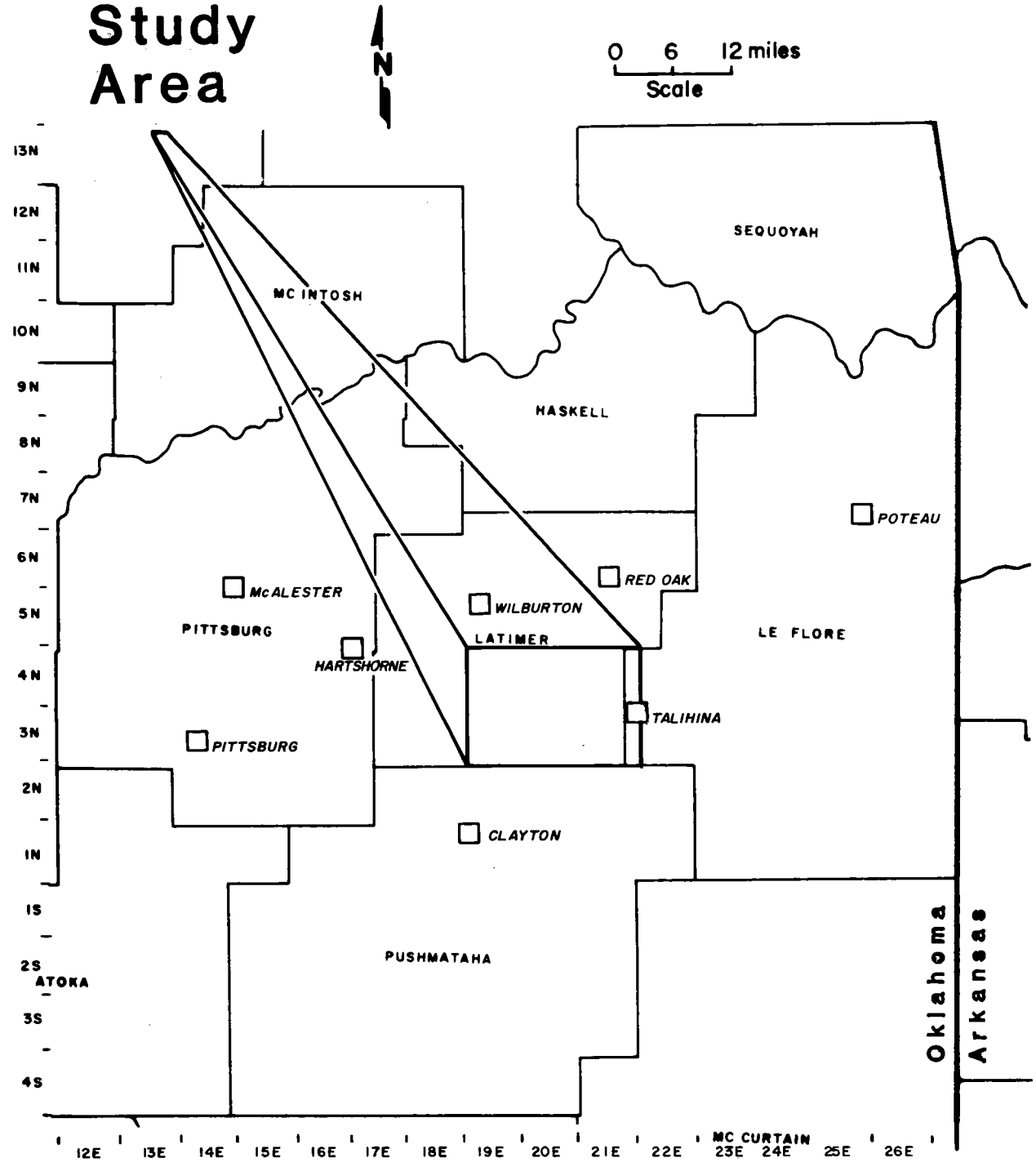


Figure 1. Area of study in the Ouachita fold-and-thrust belt of southeast Oklahoma.

"core" of the structure and is intermittently exposed along strike (Fig. 4). These Jackfork Sandstone outcrops indicate erosion to a lower stratigraphic level on the structure. This deeper erosion is caused by increased structural uplift either on the Baker Mountain fault or on one of the lower detachments in the Ti Valley sheet. The forelimb of the structure dips 75° to the north. A forelimb syncline is created as the Baker Mountain fault translates slip to the north along a Jackfork Sandstone bedding-plane detachment and then cuts up-section to form a hanging-wall cutoff. The Jackfork Sandstone is present in the subsurface in the forelimb syncline as evidenced by the outcrop at Bear Suck Knob and Weber Mountain (secs. 19–20, T. 4 N., R. 20 E.). A hanging-wall cutoff is interpreted because the Jackfork Sandstone is not exposed at the surface everywhere along strike. The fold geometry illustrated in Figure 3 above the ground surface is diagrammatic, and the exact amount of slip is unknown.

EXPLORATION SIGNIFICANCE

The primary exploration target in the area of interest is the Jackfork Sandstone. The Jackfork Sandstone is ~5,400 ft thick and contains regionally correlatable units. The section consists predominantly of sandstone units and relatively thin interbedded shales. Porosity in the sandstones can be as high as 15–18%, with permeabilities in the tens of millidarcies. Natural-gas production has been established in several wells in Tps. 3 and 4 N., Rs. 20 and 21 E. (Fig. 5). The H&H Star Daniel Hope well located in sec. 4, T. 3 N., R. 20 E., was completed in December 1992, flowing 5.7 MMcfgpd. Production is from a lower sandstone unit of the Jackfork at ~18,800 ft that is informally named the Hope sandstone. The correlative stratigraphic interval is also productive at a depth of 9,000 ft in the H&H Star Lady Luck well located in sec. 21, T. 4 N., R. 21 E. This well was completed in April 1993, flowing 4.6 MMcfgpd. The large difference in depth of this correlative unit is due to the structural complexity of the region. The Hope well produces from the lowermost sheet of the Ti Valley duplex whereas the Lady Luck is productive from an upper sheet. This duplex is in the footwall of the Baker Mountain fault. Because the sandstone is porous and productive in the footwall, it makes an excellent exploration target in the hanging wall. The Hope sandstone in the hanging wall of the Baker Mountain fault is cutoff in the subsurface and may represent a regional exploration play that is, as yet, untested. Other porous Jackfork Sandstone units in structurally favorable positions on the hanging wall of the Ti Valley fault may also be prospective. The Ti Valley duplex and overlying Baker Mountain structure afford an opportunity to penetrate the entire Jackfork Sandstone section at least three times with only one wellbore. Additionally, small-displacement imbricate thrusts may create local structural traps within the duplex or the Baker Mountain structure.

SYSTEM	Series	Formation	Abbrv. this paper
PENNSYLVANIAN	Atokan	ATOKA	Pa
	Morrowan	JOHNS VALLEY	Pjv
		JACKFORK(Hope Ss.)	MPjf
MISS.	Chesterian	STANLEY	Ms
CAMBRIAN-DEVONIAN		Undifferentiated	€-D

Figure 2. Stratigraphic chart of units in the area of interest. Adapted from Whiteside and Grayson (1990).

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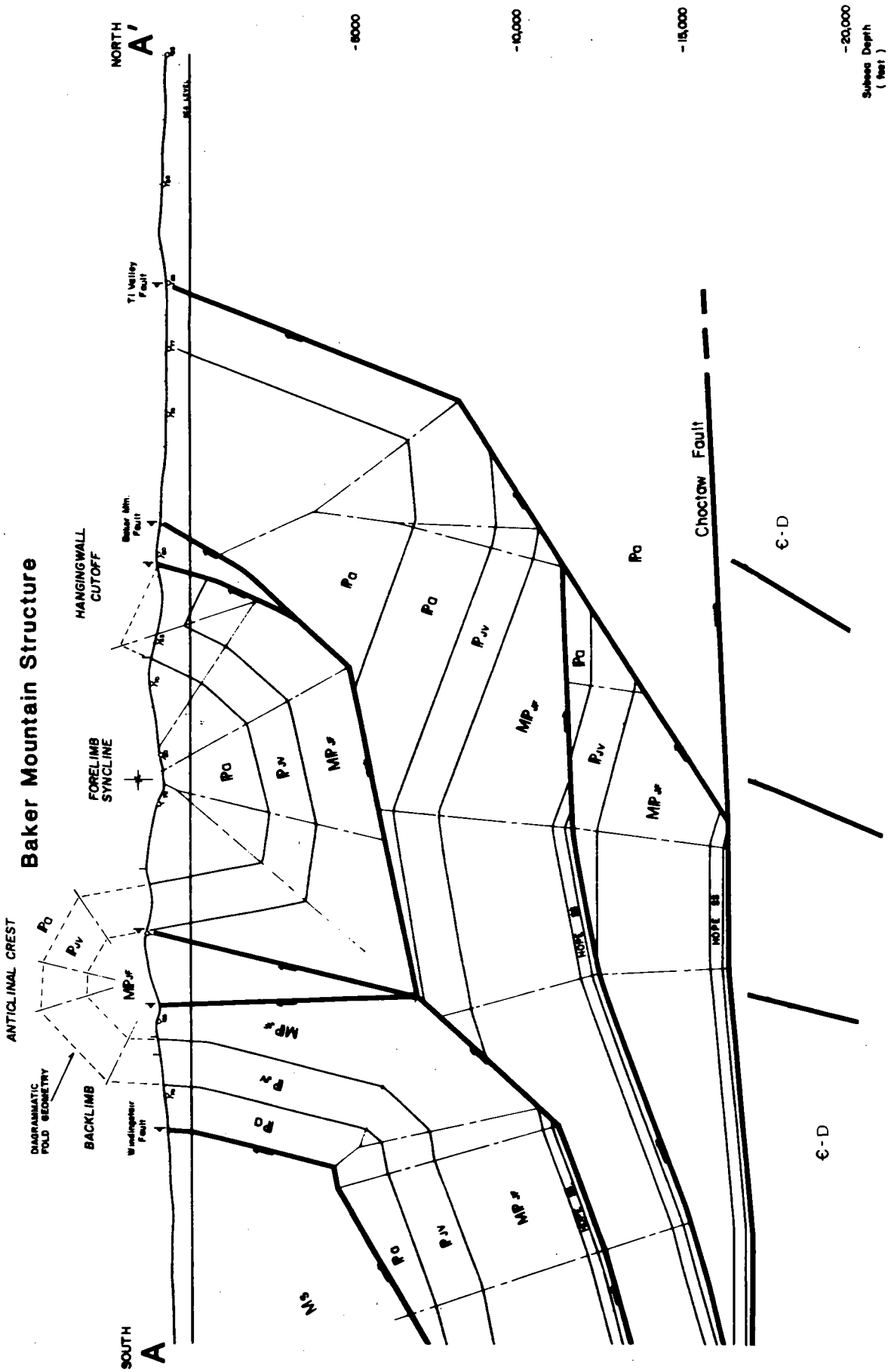


Figure 3. Geologic cross section of the Ti Valley hanging-wall units. The Baker Mountain fault is an imbricate thrust of the Ti Valley fault and creates the Baker Mountain structure. For location of cross section, see Figure 4.

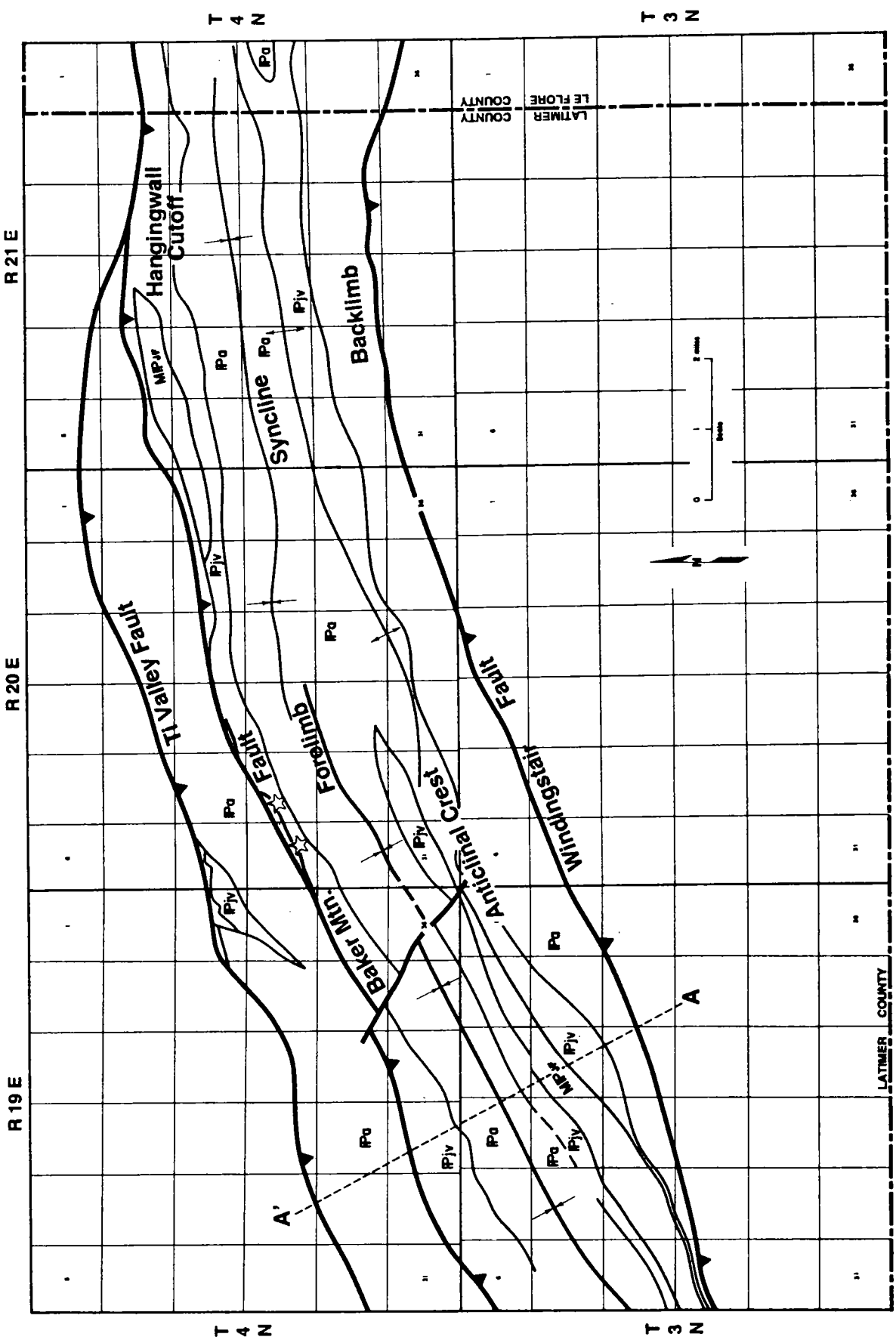


Figure 4. Simplified geologic map illustrating the geometric components of the Baker Mountain structure. The star shapes locate the Bear Suck Knob and Weber Mountain Jackfork Sandstone outcrops referred to in the text. Modified from Suneson and Ferguson (1989,1990) and Hendricks and others (1947).

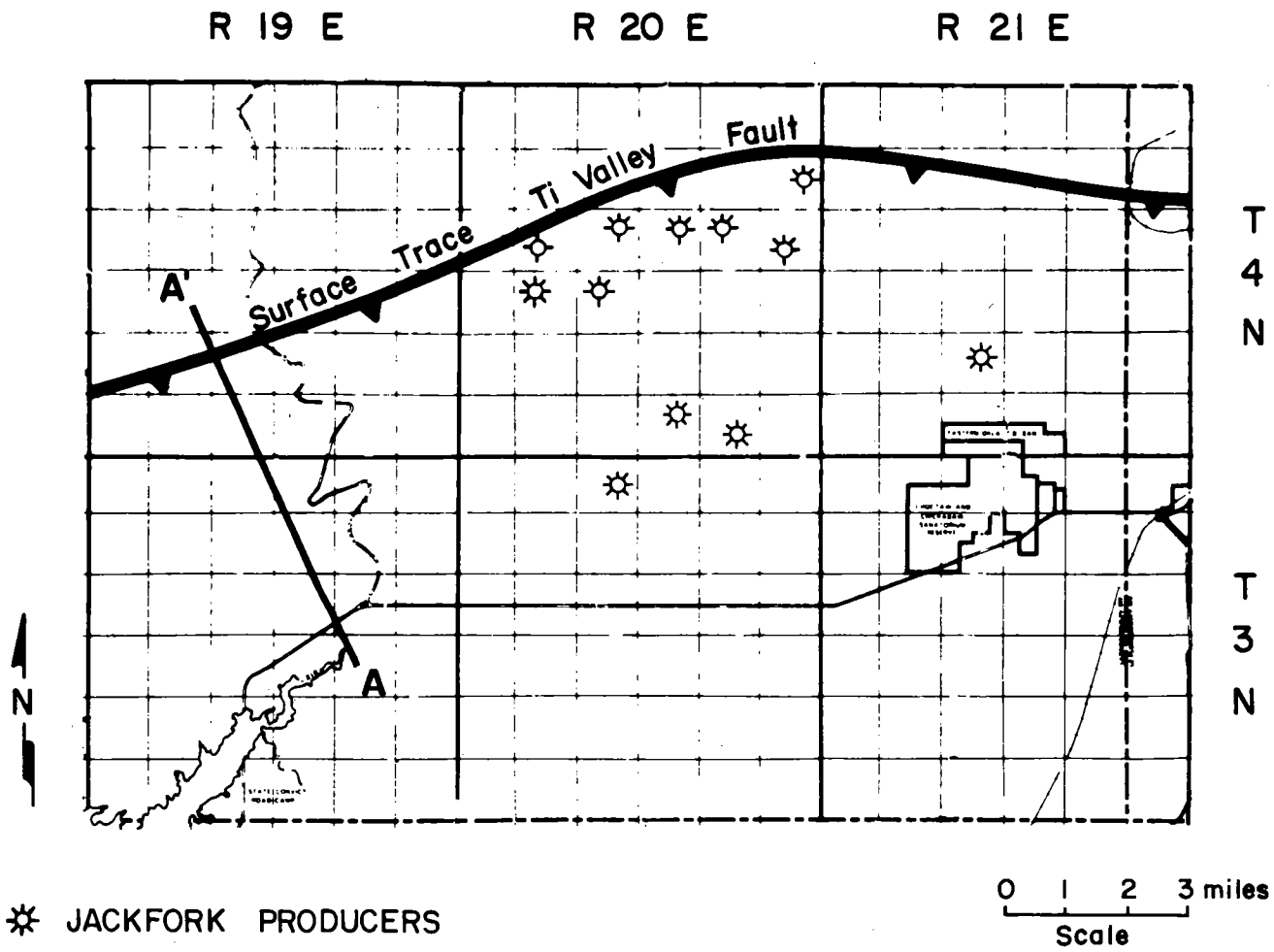


Figure 5. Natural-gas shows and production from the Jackfork Sandstone.

Preliminary Interpretation of a Seismic Profile and the Spiro Reservoir Pressure Data in the Vicinity of the Wilburton Gas Field

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Oil and gas exploration in the frontal belt of the Ouachita Mountains of southeastern Oklahoma was accelerated during the mid-1980s by the discovery of gas in Arbuckle carbonate and Spiro sandstone reservoirs in the South Hartshorne, Southwest Haileyville, and Wilburton fields. Bertagne and Leising (1989), Bertagne and others (1991), Reeves and others (1990), Perry and Suneson (1990), and Wilkerson and Wellman (1993) published seismic profiles and stressed the importance of structural interpretations of the seismic-reflection data in the frontal Ouachita belt.

We have been working on constructing several structural cross sections in the Wilburton field area (Fig. 1) to determine the detailed geometry of thrusting. The cross sections are based on updated surface geology, recent subsurface control, and our interpretation of 27 seismic profiles provided by the Exxon Corp. We have also collected pressure data from the Spiro reservoirs to determine the relation between structural configuration and the pressure gradients of the Spiro reservoirs.

Figure 2 shows our interpretation of a north-south seismic-reflection profile across the eastern end of the Wilburton gas field. In the southern part of the profile, the detachment surface is in the Woodford Shale. Well-developed duplex structures are present on the footwall block of the Choctaw fault zone where the floor thrust (detachment surface) is within the Springer Shale. The footwall block contains hinterland-dipping imbricate thrust faults that are splayed from the Springer detachment. Although the presence of a roof thrust in the footwall block is questionable in this profile, it is clearly observable in other seismic profiles to the west.

There is a rough correlation between the pressure data in the Spiro reservoirs and the duplex structures. The highest pressure gradients found in the Spiro sandstones were brought to structurally higher positions through the thrust faults in the duplex structures (Fig. 2). By contrast, the structurally lower reservoirs exhibit lower pressure gradients on both sides of the structurally higher Spiro sandstone. Therefore, we suggest that the thrusting in the duplex structures in the subsurface has acted as a seal for gas in the Spiro sandstone.

There are many oil seepages in the frontal Ouachita belt, however, and these suggest that not all the thrust faults in the region have acted as seals for oil migration. Our field observations indicate that these seepages are close to the major thrust faults of the region such as the Choctaw and Ti Valley fault zones. Moreover, the Spiro sandstone is not productive in the hanging-wall block (upper block) of any of these faults. Therefore, we tentatively suggest that the overpressured reservoirs in the Spiro sandstone can be found only in the subsurface duplex structures within the frontal Ouachita belt.

The presence of a triangle zone in the frontal Ouachita belt has been proposed, first by Hardie (1988) and later by Reeves and others (1990) and Perry and Suneson (1990). Although our interpretations are only preliminary, there is a possibility that a triangle zone does not exist in the Wilburton area. If the Carbon fault is a south-dipping thrust fault (Neil Suneson, personal communication, 1994), it is possible that the Carbon fault is a splay upward and to the north from either the Springer detachment in the subsurface or the roof thrust to the west.

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Cemen, Ibrahim; Al-Shaieb, Zuhair; Feller, R.; and Akthar, S., 1994, Preliminary interpretation of a seismic profile and the Spiro reservoir pressure data in the vicinity of the Wilburton gas field, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 249–251.

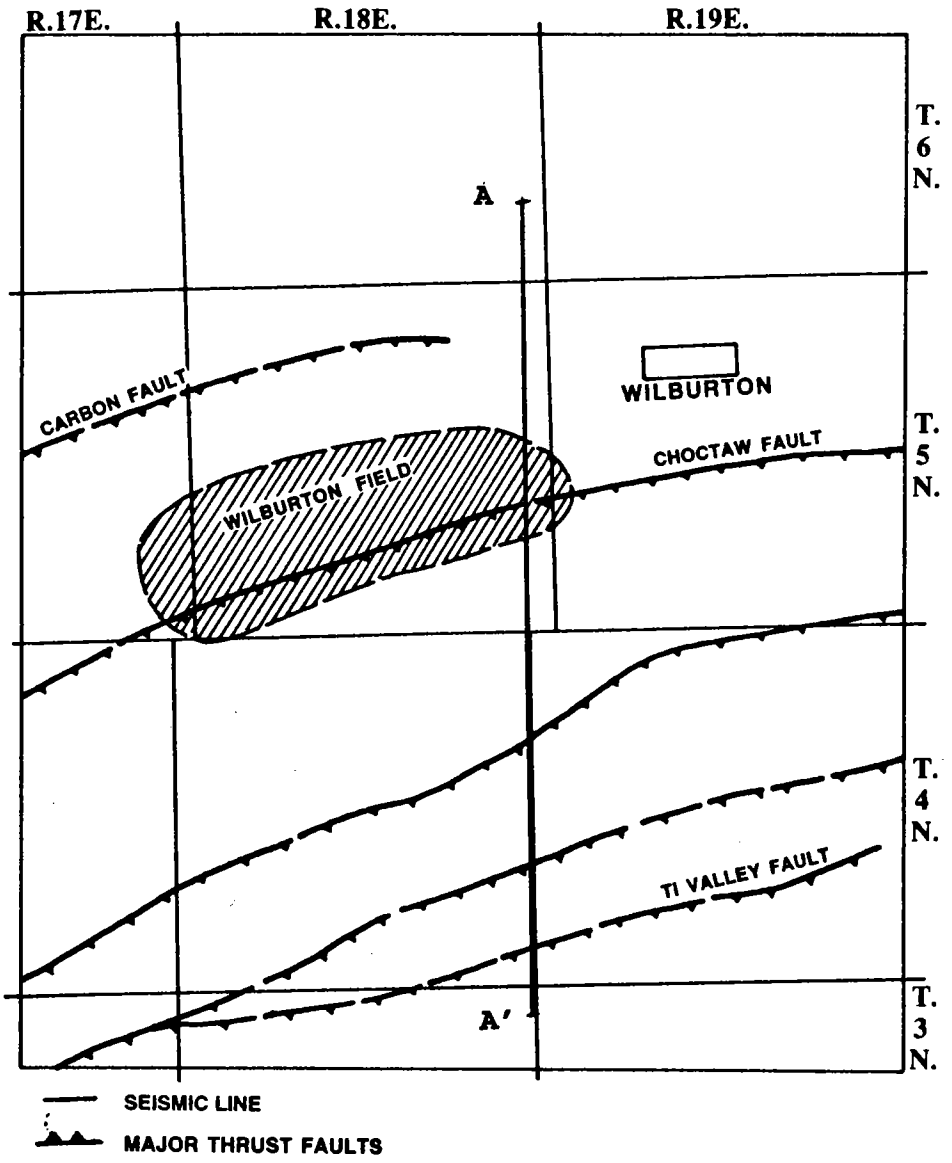


Figure 1. A schematic tectonic map showing the major thrust faults in the vicinity of the Wilburton gas field. Line A–A′ indicates the approximate location of the interpreted north-south seismic-reflection profile.

Reeves, L. D.; Schreiner, W. P.; and Sheffield, T. M., 1990, Stop 6: New State Mountain (Amoco 1-5 Rosso Unit), *in* Suneson, N. H.; Campbell, J. A.; and Tilford, M. J. (eds.), Geology and resources of the frontal belt of the western Ouachita Mountains, Oklahoma: Oklahoma Geological Survey Special Publication 90-1, p. 37–40.

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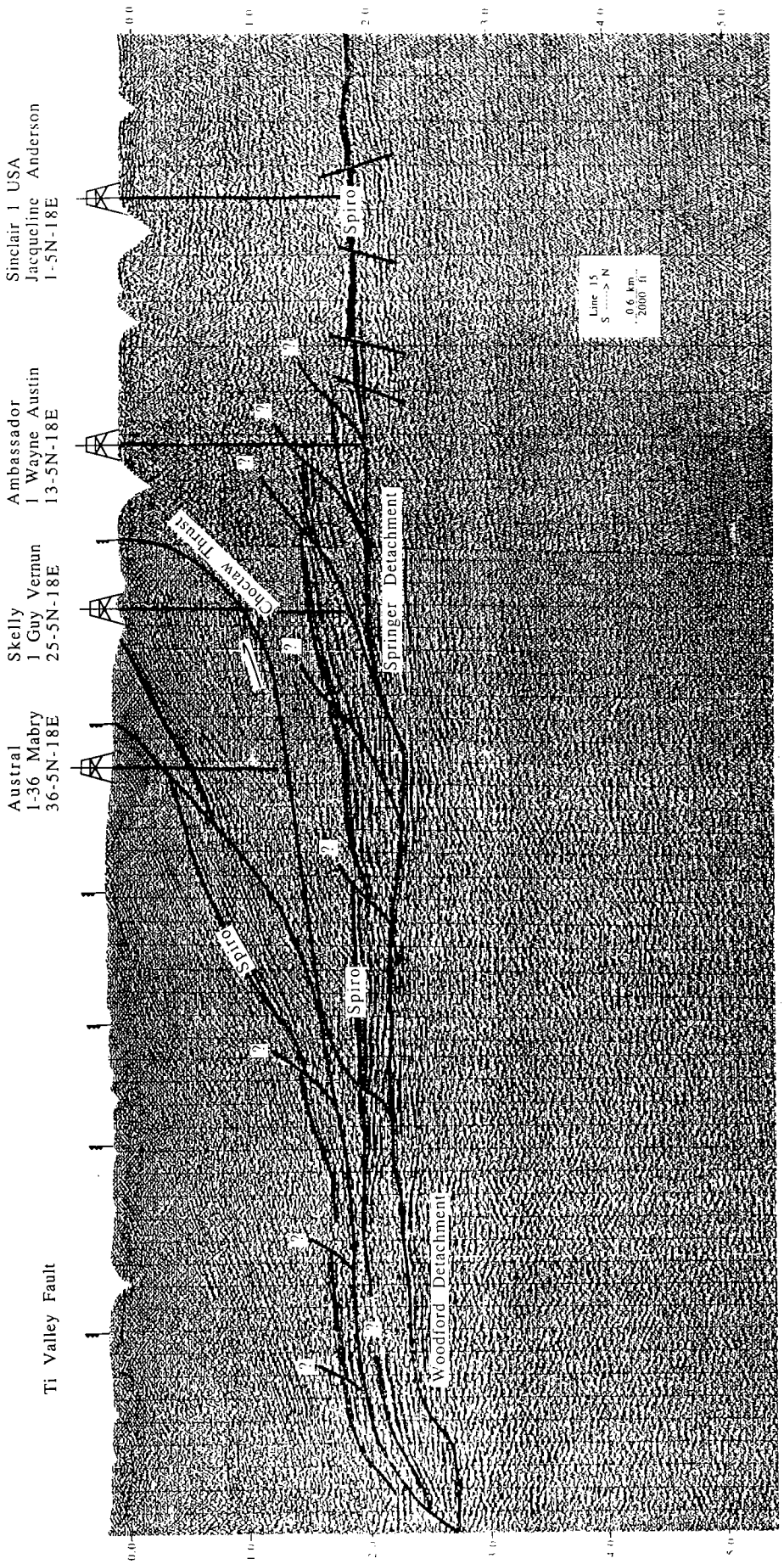


Figure 2. Interpreted north-south seismic-reflection profile in the Wilburton area. See text for explanation.

Structural Analysis and Tectonic History of the Potato Hills, Ouachita Mountains, Oklahoma

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ABSTRACT.—On the basis of thrust-fault intersections and opposing compressional directions, the North and South Potato Hills thrusts are recognized as separate faults; a window through a single fault is nonexistent. The northeastern, northwestern, and western Potato Hills are key areas for interpreting the overall structure of the area. Evidence that the two thrusts are separate faults includes: (1) the intersection of the North Potato Hills thrust with the South Potato Hills thrust in the northeastern Potato Hills, (2) the extension of the North Potato Hills thrust to the northwest in the northwestern Potato Hills, and (3) the extension of the South Potato Hills thrust to the west in the western Potato Hills.

Erosion through a folded thrust (i.e., the Potato Hills thrust) to produce a window does not explain the structural geology of the Potato Hills. Seismic data, electric and sample logs, and surface geology reveal the Potato Hills thrusts to be an out-of-sequence imbricate-fan thrust system. General north-south compression resulted in north-directed, subhorizontal translation on the Potato Hills basal detachment. Imbrications then developed out-of-sequence from north to south. Most thrusts lost displacement and eventually died out, transferring slip to a fold or distributing slip along several splays.

INTRODUCTION

Location

The Potato Hills are located in southern Latimer and northern Pushmataha Counties in the Ouachita Mountains, southeastern Oklahoma (Fig. 1). They lie ~3 mi south of the Windingstair fault and form a series of low, isolated hills rising 200–500 ft above the surrounding valley.

Previous Work

The structural relations and the tectonic evolution of the Potato Hills have been controversial for more than 60 years. The Potato Hills have previously been interpreted as a window through either the Windingstair fault (Miser, 1929) or the Jackfork Mountain thrust (Miller, 1955; Roe, 1955; Arbenz, 1968). Opponents of the window hypothesis include Misch and Oles (1957), Tomlinson (1959), Pitt (1971,1974), Pitt and others (1982), and Allen (1990,1991a,1992,1993,1994).

This paper summarizes my previous work involving detailed field mapping, seismic interpretation, and log analysis covering key areas in the Potato Hills, i.e., the northeastern, northwestern, and western Potato Hills.

STRATIGRAPHY

The Potato Hills are a structurally complex area composed of lower and middle Paleozoic strata. Rocks that crop out range from Middle Ordovician Womble Shale to the Mississippian Stanley Group (Fig. 2). A description of these units is included in Stop 23 of this guidebook and will not be repeated here. The key factor that controls structural style is that the Bigfork Chert and Arkansas Novaculite, both thick chert units, are "sandwiched" by ductile shale units (Womble Shale, Polk Creek–Missouri Mountain Shales, and Stanley Group).

STRUCTURAL GEOLOGY

The deformation style in the Potato Hills is folding that resulted from thrust faulting. Five major thrusts are present in the Potato Hills area and are responsible for the structural features observed. The North Potato Hills thrust, South Potato Hills thrust, and Cedar Creek thrust are easily identified on the surface and in the subsurface because they juxtapose Stanley Group strata against older units. The Jackfork Mountain thrust and the Buffalo Creek thrust are not easily located in the field because they juxtapose Stanley against Stanley, and they

Allen, M. W., 1994, Structural analysis and tectonic history of the Potato Hills, Ouachita Mountains, Oklahoma, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 253–261.

can only be identified inexactly in the subsurface. Steep to vertical tear faults are present in the hanging walls of a few thrusts. Some of these tear faults display horizontal as well as vertical movement.

I will discuss the key areas from east to west, focusing on the structural relations of the North Potato Hills thrust. Proving that this thrust is not the northern boundary fault of the previously proposed window is important for interpreting the structural and tectonic history of the Potato Hills.

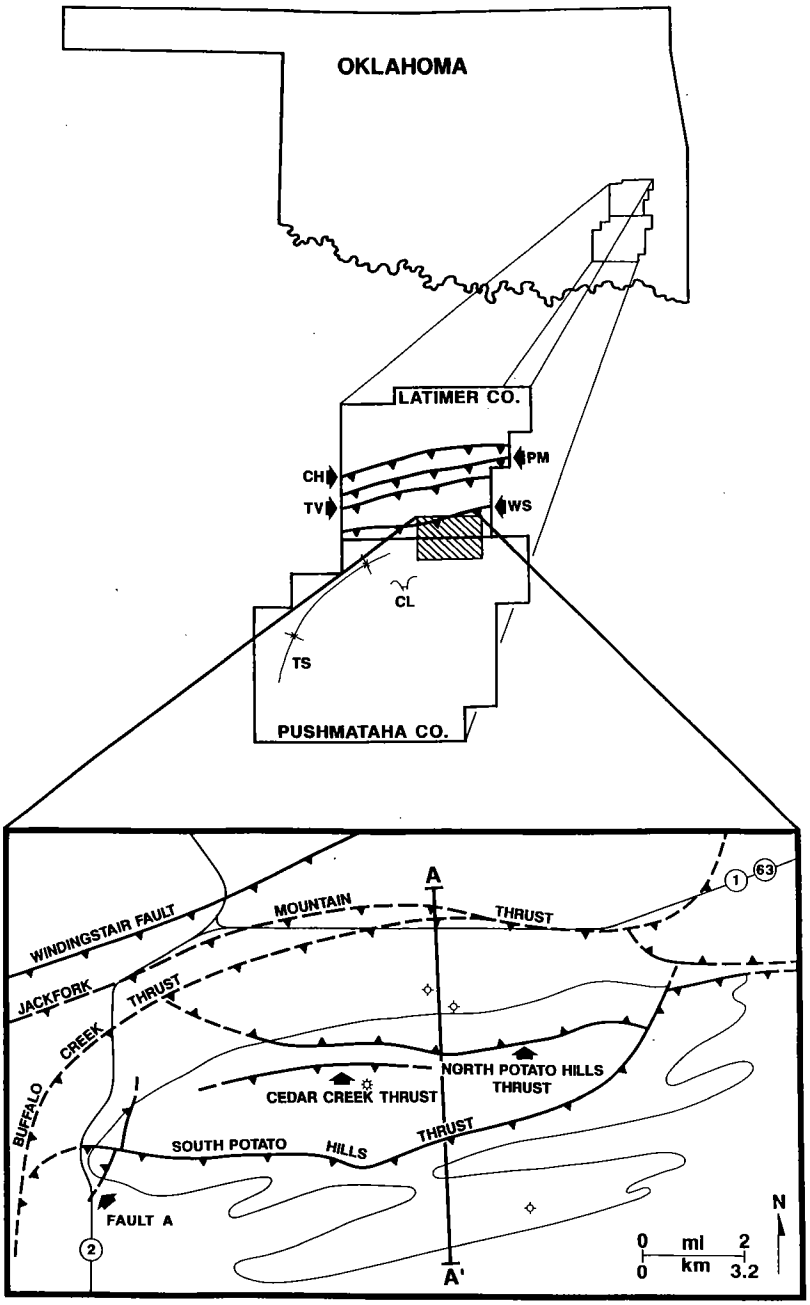


Figure 1. Potato Hills location map, Latimer and Pushmataha Counties, Oklahoma. CH = Choctaw fault, PM = Pine Mountain fault, TV = Ti Valley fault, WS = Windingstair fault, TS + CL = Tuskahoma syncline, CL = Clayton horst. Cross section A–A′ is illustrated in Figure 6.

Northeastern Potato Hills

The North and South Potato Hills thrusts converge in the northeastern Potato Hills (Fig. 3). Surface geology supported by subsurface control indicates that the north thrust and the south thrust are separate faults with different vergences (Allen, 1991a). The Sinclair No. 1 Coussens (SW¼SW¼SE¼ sec. 21, T. 3 N., R. 21 E.) penetrated the South Potato Hills thrust at 1,862 ft and entered the Bigfork Chert in the North Potato Hills thrust hanging wall.

The area contains three major east-plunging folds (Fig. 3) oriented, from north to south, 089°/20°, 084°/20°, and 092°/18° with axial-plane dips of 59° S., 86° S., and 77° S., respectively. The orientations of the folds resulted from north-directed compression and indicate a structural and tectonic relation to the South Potato Hills thrust. This is indicated by the south dip of the axial planes of the folds, their north vergences, and their location in the South Potato Hills thrust hanging wall.

In the northeastern Potato Hills, the Jackfork Mountain thrust changes strike to the northeast (Fig. 3), as determined by bed orientations, and corresponds to the thrust mapped by Suneson and Ferguson (1990) that truncates the eastern end of the Buffalo Mountain syncline to the north. The trace of the South Potato Hills thrust also "curves" to the northeast (Fig. 3), suggesting northwest transport. Joints in the Bigfork Chert, located in the hanging wall of the thrust, indicate a maximum stress direction of N. 67° W. The change in strike of the South Potato Hills thrust indicates either (1) a late orogenic compressional shift from the north to the northwest or (2) the presence of an oblique ramp (Allen, 1991a). The trend of the three major east-plunging folds in the hanging wall of the South Potato Hills thrust (Fig. 3) argues against a change in direction of compression. The tectonic setting, the change in strike of the South Potato Hills and Jackfork Mountain thrusts, and the northeast alignment of parts of the Jackfork Mountain and South Potato Hills thrusts in the northeastern Potato Hills favor the existence of an oblique ramp. The localized northwest compression resulted from movement of the South Potato Hills thrust over the oblique ramp.

Northwestern Potato Hills

Two hypotheses concerning the North Potato Hills thrust are that (1) it is the north boundary fault of a window (Miser, 1929; Miller, 1955; Roe, 1955; Arbenz, 1968) or (2) it is a backthrust in an imbricate-fan thrust system (Allen, 1994). The previously proposed window fault, thought to circle the

PERIOD	SERIES	UNIT
Mississippian	Chesterian	Stanley Group
	Meramecian	
	Osagean	Arkansas Novaculite
	Kinderhookian	
Devonian	Upper	
	Lower	
Silurian	Upper	
	Lower	Missouri Mountain Shale
Ordovician	Upper	Polk Creek Shale
		Bigfork Chert
	Middle	Womble Shale

Figure 2. Stratigraphic sequence, Potato Hills. Modified from Allen (1993).

Potato Hills, was called the Potato Hills thrust (Fig. 4) and is composed of the North and South Potato Hills thrusts. The key to resolving the controversy is locating the trace of the thrust in the northwestern Potato Hills.

The trace of the North Potato Hills thrust, where mappable, is ~6.5 mi long and strikes west-northwest. The western end of the fault juxtaposes poorly exposed shale of the Stanley Group on each side, making it difficult to map. To the east, the thrust juxtaposes Womble Shale through Stanley on the north against Stanley on the south. Displacement decreases to the west. Where the thrust is exposed in a stream cut, it dips ~40° N. and displays overturned beds and north-dipping axial planes of minor folds, indicating south-directed translation. The north dip is supported by subsurface information, including logs from two wells drilled in the North Potato Hills thrust hanging wall and a proprietary seismic line. These indicate a curved, north-dipping fault plane. A north-dipping fault plane is opposite to the regional tectonic transport direction and, by definition, is a backthrust.

Two exposures support extending the North Potato Hills thrust to the west-northwest as shown in Figure 4. A small fault mapped by Miller (1955) (NE¼ sec. 26, T. 3 N., R. 19 E.) offsets a Stanley marker bed and intersects the projected trace of the North Potato Hills thrust. The fault is downthrown to the southwest as is the North Potato Hills thrust. Additionally, Miller (1955) mapped this fault as a normal fault; this interpretation suggests extension, which is inconsistent with the deformational style in the Potato Hills. Another key feature is a small fold (C S½ sec. 23, T. 3 N., R. 19 E.) (Fig. 4). The fold has a trend and plunge of 283°/19° and indicates that compressive forces were either south-southwest or north-northeast. The trend of the fold parallels the projected trace of the thrust. The attitude and vergence of the north Potato Hills thrust are very different from those of the proposed "window" fault.

Western Potato Hills

In the western Potato Hills, the South Potato Hills thrust has been mapped as curving to the northeast (Miller, 1955) or as continuing to the west beyond the Potato Hills (Pitt, 1974; Allen, 1992). Surface geology and interpretation of well logs (multiple repeated Bigfork Chert through Stanley intervals in the Gragg No. 1 Kenman [NE¼NE¼ sec. 8, T. 3 N., R. 19 E.]) indicate that the South Potato Hills thrust does extend westward beyond the Potato Hills.

Two distinct compression directions can be interpreted in the western Potato Hills. The first is north-directed, resulting in the South Potato Hills thrust. The second is southeast-directed, resulting in fault A (Fig. 5). Fault A strikes north-northeast; this orientation indicates east-southeast transport and suggests that fault A is an antithetic backthrust that resulted from volumetric crowding and hanging-wall rotation of the frontal footwall ramp. The orientation and displacement are controlled by the Buffalo Creek thrust from which it originated. Therefore, fault A is analogous to but not structurally associated with the North Potato Hills thrust.

Deformation in the western Potato Hills appears to have been more extreme than elsewhere in the Potato Hills as evidenced by steep to vertical beds and isoclinal folding of the Arkansas Novaculite. The compressional event directed to the southeast (related to fault A) appears to have been stronger than the north-directed compressional event (related to the South Potato Hills thrust) because the more intense folding has a consistent southeast vergence. This intense folding was probably not caused by movement on fault A, which has a small amount of translation. A structure west of fault A must also contribute to the southeast-directed compression. Mapping west of the Potato Hills is not possible owing to the presence of Sardis Reservoir. However, regional anal-

ysis did reveal a possible transverse structure west of the Potato Hills (discussed below) that may have directly or indirectly influenced compression in the western Potato Hills.

Approximately 5 mi southwest of the Potato Hills (T. 1 N., R. 19 E.) is the Clayton horst. This structure is significant because there is >7,000 ft of uplift along near-vertical fault planes (Pitt and others, 1982). On the basis of Cline's (1960) map, the two major faults of the horst strike N. 20° W. and N. 10° E. If the Clayton horst is projected northward, it would cross the area west of the Potato Hills and may represent the transverse structure (Allen, 1992). Evidence for this structure in the surface geology includes a branching of lineaments from the

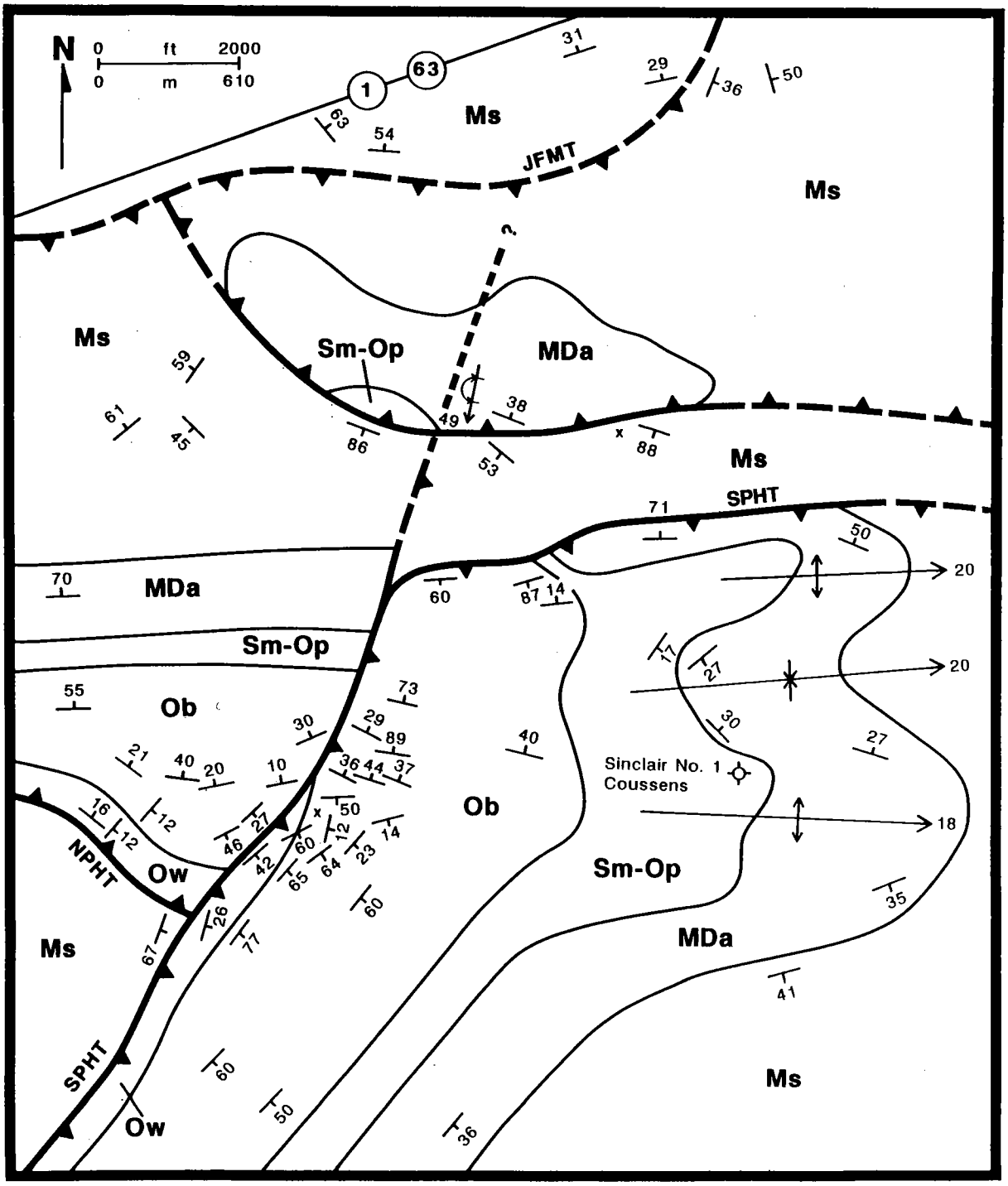


Figure 3. Geologic map, northeastern Potato Hills (from Allen, 1991a). Ow = Womble Shale, Ob = Bigfork Chert, Sm-Op = Polk Creek–Missouri Mountain Shales, MDa = Arkansas Novaculite, Ms = Stanley Group. NPHT = North Potato Hills thrust, SPHT = South Potato Hills thrust, JFMT = Jackfork Mountain thrust. Small x's mark locations of joint data.

northward projection of the horst and an isolated outcrop of Bigfork Chert (S½NE¼ sec. 30, T. 3 N., R. 19 E.) in the Jackfork Mountain thrust sheet (Allen, 1992, fig. 6, p. 181).

The exact nature of this structure is unknown; however, it is possible that the surface geology reflects an approximately north-striking ramp or compartmental fault associated with the Windingstair fault. Compartmentalization is based on the Tuskahoma syncline to the west of the structure juxtaposed against the anticlinal valley containing the Potato Hills to the east. Also, the apparent termination of the major structures and the lack of any direct match-up across the transverse structure suggest compartmentalization.

SUBSURFACE STRUCTURAL ANALYSIS

Subsurface Data

Structural relations in Figure 6 are based on seismic interpretation, multiple well control, and surface geology. The key wells are the Sinclair No. 1 Reneau and the American Quasar No. 1-11 Cabe because they are the only wells to penetrate the Potato Hills basal detachment (Fig. 6). Electric and sample-log analysis of the No. 1 Reneau indicates that three major and two minor faults are present above the basal detachment. The well penetrated the basal detachment at 5,780 ft and encountered strata interpreted as Mississippian Stanley Group. Analysis of electric and sample logs from the No. 1-1 Cabe indicates several repeated Bigfork-Womble intervals above the basal detachment at 7,812 ft that juxtaposes Womble Shale in the hanging wall and Stanley Group strata in the footwall. The well encountered 4,542 ft of Stanley in fault contact with underlying sandstone of the Pennsylvanian Jackfork Group.

A proprietary seismic line indicates several subhorizontal faults that ramp and imbricate as they approach the surface. The chert units and sandstone beds in the Stanley and Jackfork Groups produce reflection markers on the seismic data. The No. 1 Reneau and the No. 1-11 Cabe logs were superimposed on the seismic line, and rock types were compared with reflections.

Interpretation

Electric and sample logs from the No. 1 Reneau and No. 1-11 Cabe indicate multiple thrust faults separating repeated and locally overturned lower and middle Paleozoic strata. Cross section A–A′ (Fig. 6) shows that the

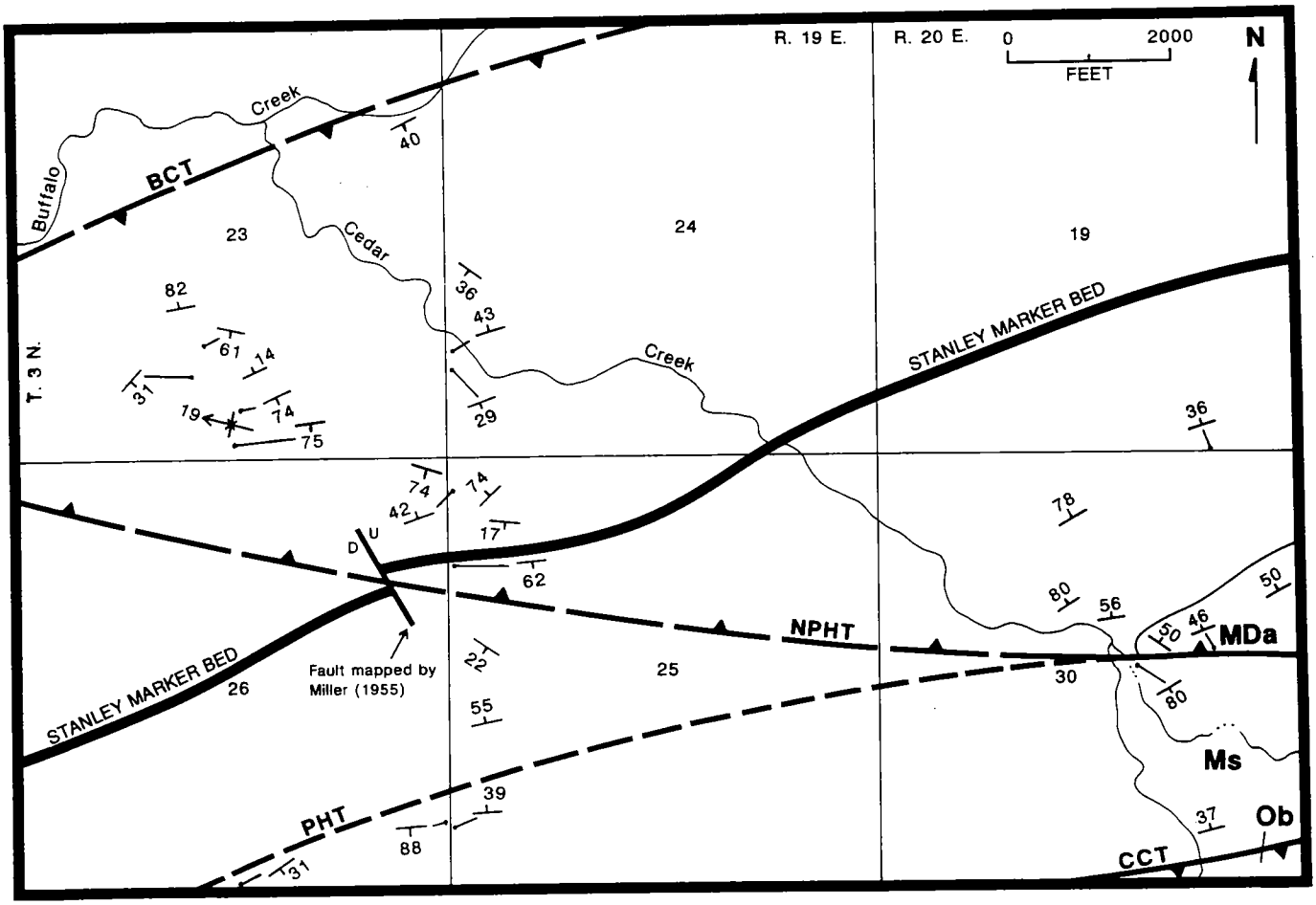


Figure 4. Geologic map, northwestern Potato Hills (from Allen, 1993). CCT = Cedar Creek thrust, BCT = Buffalo Creek thrust, PHT = formerly proposed Potato Hills thrust. Tenmile Creek siliceous shale is the marker bed in Stanley Group. Refer to Figure 3 for other symbols.

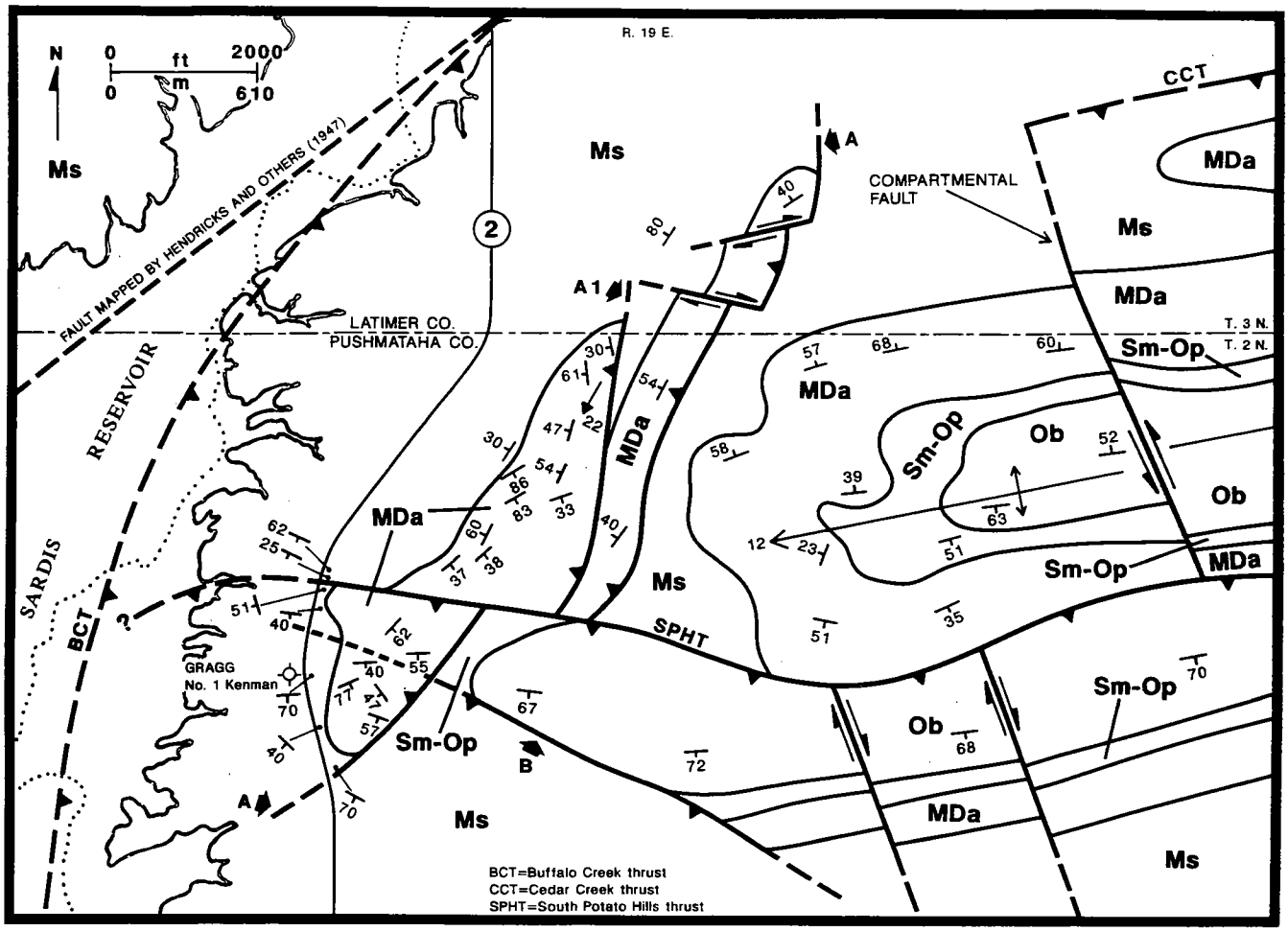


Figure 5. Geologic map, western Potato Hills (from Allen, 1992). A, A1, and B refer to minor thrust faults. Dotted line in Sardis Reservoir indicates former location of Buffalo Creek before the creation of Sardis Reservoir. Refer to Figures 3 and 4 for other symbols.

major thrusts in the Potato Hills cut up-section from a basal detachment but do not appear to rejoin at a higher level. This interpretation is supported by (1) anticlinal folds over most of the major thrusts and (2) the tendency of the thrusts to splay after moving over a ramp.

Figure 6 shows several of the major subhorizontal thrusts' being terminated by dashed lines at their northern ends or where they approach the surface. This approach was taken because (1) some may be blind thrusts that do not reach the surface and (2) some juxtapose Stanley against Stanley and their trace cannot be identified in the field. The thrust faults shown with a solid line on Figure 6 can be identified on the seismic data, and their traces can be mapped in the field. No attempt was made to extend the faults beyond where the data allowed.

THRUST-BELT MODELS

An imbricate fan is a thrust system in which faults cut up-section from a basal detachment but do not rejoin at a higher stratigraphic level (Marshak and Woodward, 1988). Each thrust in an imbricate fan loses displacement up-section and eventually dies out by transferring slip to a fold at its tip or by distributing it along several splays (Mitra, 1986). A duplex is a thrust system in which faults cut up-section from a basal detachment and merge at a higher stratigraphic level to form another continuous detachment (Marshak and Woodward, 1988).

One faulting sequence in thrust belts is called a break-forward sequence in which structurally lower faults (faults at greater depths) are younger than overlying faults (Marshak and Woodward, 1988). In a thrust belt of this type, the more external faults (farthest toward the foreland) are generally the youngest, whereas the more internal faults are the oldest. Break-forward sequencing would suggest that the faults are progressively younger toward the foreland; this is considered a "normal" sequence of thrusting (Marshak and Woodward, 1988). The spatial arrangement of these faults implies that older faults can be folded by younger faults.

A different faulting sequence in thrust belts is termed out-of-sequence faulting in which structurally lower faults are older than overlying faults (Marshak and Woodward, 1988). Minor thrusts rooted in a major thrust are sometimes interpreted as imbrications in the hanging wall of an existing thrust sheet and are believed to develop out-of-sequence (Jones, 1987). Boyer and Elliott

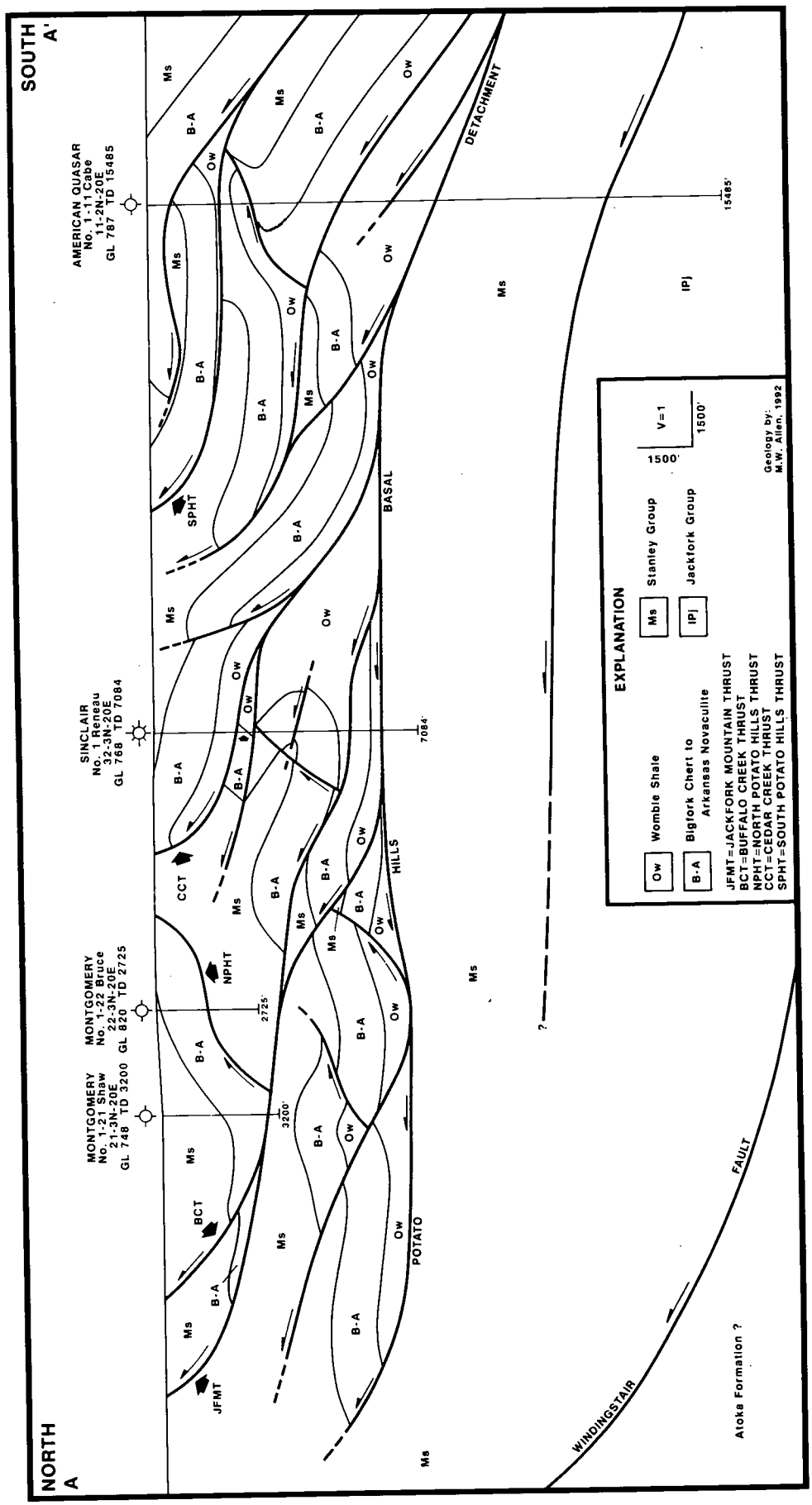


Figure 6. Structural cross section (A–A'), central Potato Hills (from Allen, 1994). Refer to Figure 1 for cross section location.

(1982) stated that an imbricate-fan thrust system is a structure in which out-of-sequence thrusting could occur.

Field relations show only the sequence of terminations of events and not the sequence in which those events started (Jones, 1971). This limitation means that the position and orientation of thrust faults have little bearing on when each thrust developed and cannot be used as sequence indicators. A timing indicator that may be used is whether the thrusts have been folded by subthrust stacking. Folded faults may be expected to occur in any thrust belt if the stratigraphic section has contrasting competencies and the sequence of thrusting is normal (Jones, 1971).

Interpretation of subsurface data (Fig. 6) indicates that the faults cut up-section but do not rejoin at a higher stratigraphic level as indicated by the folded strata above the fault planes. Each thrust appears to have lost displacement up-section and transferred slip to a fold or distributed slip along several splays. The major thrusts are not folded by subthrust stacking. These interpretations support an imbricate-fan thrust system in the area.

The tectonic implications of this interpretation are that thrusting in the Potato Hills was north-directed translation on a subhorizontal thrust fault defined as the Potato Hills basal detachment. Imbrications developed from the basal detachment sequentially from north to south. Evidence for this out-of-sequence thrusting is the absence of folded faults in the southern part of the Potato Hills (Fig. 6).

TECTONIC HISTORY

The compressional event for which there is the most evidence in the Potato Hills was directed N. 9° W.; this conclusion is indicated by fault and fold orientation and is based on the fact that minor-fold axes (Allen, 1990) are perpendicular to the primary force in an orogenic belt (Billings, 1972). A subtle compressional event directed S. 6° W. is indicated in the northern and central Potato Hills. Evidence of south-directed compression diminishes southward and was not observed in the southern Potato Hills. South-directed compression resulted from backthrusting on the North Potato Hills thrust. Tomlinson (1959), Pitt (1974), and Pitt and others (1982) indicated multiple deformational events in the Potato Hills. Interpretation of mesoscopic structures indicates that this area was subjected to a progressive deformational event (Allen, 1990). The previous interpretations of multiple deformational events were probably based on structures created by backthrusting.

The anticline associated with the Cedar Creek thrust (Fig. 5) contains parasitic folds on both limbs and near the fold hinge (Allen, 1991b). Three common settings for parasitic folds are (1) on the flanks of major folds, (2) in regions of thrust tectonics, and (3) where the gliding of an approximately horizontally stratified rock mass has taken place (Wilson, 1982). The first two settings apply to the Potato Hills, and the third setting is inferred on the basis of the cross section in Figure 6. The low-angle basal detachment beneath the Potato Hills illustrated in the cross section (Fig. 6) suggests that the lower Paleozoic strata were initially transported horizontally before faulting and folding.

Parasitic folds are probably best accounted for by layer-parallel shortening (Ramsay and Huber, 1987) that can create minor folds in a layered sequence before the onset of significant folding (Davis, 1984). Parasitic folds in the Potato Hills suggest that layer-parallel shortening (or "gliding" [Wilson, 1982]) was the tectonic mechanism responsible for deformation in the Potato Hills area.

HYDROCARBON POTENTIAL

Shallow Production

Shallow production from the abandoned two-well Potato Hills field was established from fractured, vuggy Bigfork Chert (with Womble Shale as the source) in the central Potato Hills. On the basis of log analysis, the field is a structural trap below the Cedar Creek thrust. The structural style suggests a footwall trap with subthrust closure produced by frictional drag.

The discovery well, the Sinclair No. 1 Reneau, was spudded on May 9, 1959, and completed on March 17, 1960. The perforated interval was from 2,340 ft to 2,480 ft, and the IPF was 1.85 MMcfgpd. The Wyoming Oil and Gas No. 1 Lonnie Allen (C SE¼NW¼, sec. 31, T. 3 N., R. 20 E.) was spudded on March 18, 1974, and completed on May 8, 1974, extending the field to the west. The well was perforated between 2,787 ft and 3,195 ft, and the IPF was 2.2 MMcfgpd. The perforated intervals vary by several hundred feet; however, they are the same stratigraphic and structural interval; a compartmental fault separates the two wells.

Deep Potential

Duplexes and imbricate thrust systems are recognized as important components of most thrust belts and form important structural hydrocarbon traps (Mitra, 1986). One of the most common structural positions of duplexes is in the footwall of major ramp anticlines. Advantages deep subthrust duplexes have over shallow structures are higher confining pressures and lower susceptibility to invasion and flushing by surface water.

Interpretation of seismic data indicates that the gas-productive Spiro sandstone–Wapanucka Limestone that has been extensively developed to the north is present beneath the Potato Hills at ~20,000 ft (Allen, unpublished). This interval is allochthonous and occurs in a duplexlike geometry below the Windingstair fault. The Spiro-Wapanucka interval also appears to exist in normal-fault blocks at ~28,000 ft. The structural style is similar to that at Wilburton to the north and indicates that the carbonate shelf extended this far south.

SUMMARY

On the basis of thrust-fault intersections and opposing compressional directions, I have shown the North and South Potato Hills thrusts to be separate faults (Allen, 1991a,1992) and that a window is nonexistent. Erosion through a folded thrust (i.e., the Potato Hills thrust) to produce a window does not explain the structural geology of the Potato Hills. Interpretation of seismic data, electric and sample logs, and surface geology leads me to suggest that the Potato Hills are an out-of-se-

quence, imbricate-fan thrust system. General north-south compression resulted in north-directed, subhorizontal translation on the Potato Hills basal detachment. Imbrications then developed out-of-sequence from north to south. Most thrusts lost displacement and eventually died out by transferring slip to a fold or by distributing slip along several splays.

The northeastern, northwestern, and western Potato Hills are key areas for interpreting the overall structure of the area. The intersection of the North Potato Hills thrust with the South Potato Hills thrust in the northeastern Potato Hills, the extension of the North Potato Hills thrust to the northwest in the northwestern Potato Hills, and the extension of the South Potato Hills thrust to the west in the western Potato Hills clearly indicate that the two thrusts are separate faults.

Two major subsurface structures are indicated by surface geology on the eastern and western ends of the Potato Hills. An oblique ramp is inferred to exist at the eastern end of the Potato Hills on the basis of local and regional surface expressions of major thrust faults and fold orientations. A transverse structure is inferred to exist to the west of the Potato Hills on the basis of abrupt changes in structural style and surface expressions of major thrusts in the area. The transverse structure is thought to be structurally related to the Clayton horst, located a few miles southwest of the Potato Hills.

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Regional Variations in Quartz Deformation Mechanisms, Eastern Ouachita Mountains

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INTRODUCTION

The Ouachita Mountains contain outcrops of a continuous late Paleozoic orogenic belt that borders the southern margin of the North American craton. This fold-and-thrust belt is made up of a thick sequence of largely clastic Paleozoic rocks deformed during the Carboniferous Ouachita orogeny. Deformation arising from this event has resulted in a complex but well-known suite of macroscopic and mesoscopic structural features including folds, faults, and cleavage (Stone and Haley, 1984; Arbenz, 1989; Nielsen and others, 1989). Less well known is how deformation has been accommodated at the microscopic scale.

Microstructures and textures form at different stress levels and in different kinematic frameworks, and they serve as a primary means for identifying operative deformation mechanisms, rock rheology, and dynamics during deformation (Groshong, 1988; Knipe, 1989). The integration of microscopic data with mesoscopic and macroscopic observations increases validity of structural interpretations (Schedl and van der Pluijm, 1988). Microstructural evolution is controlled largely by a combination of environmental factors such as temperature, pressure, shear stress, fluid pressure, strain, and strain rate. The purpose of this study is to examine regional variations in quartz deformation microstructures in sandstones of the Ouachita Mountains and the adjacent Arkoma Basin along a north-south transect in western Arkansas. Because microstructures reflect the operative deformation mechanisms in a rock, knowledge of regional variations in microstructural type and abundance provides constraints on the conditions of deformation and provides recognition of rocks deformed under contrasting conditions. Where a change in fabric is readily discernible, a deformation front can be located. Knowledge of deformation conditions and structural boundaries allows testing of various models for the tectonic history of the Ouachita-Arkoma region.

GEOLOGIC SETTING

The Ouachita-Arkoma region in Arkansas is commonly divided into four tectonic provinces (Arbenz, 1989): (1) the Arkoma Basin, (2) the inner and outer frontal imbricate zone, (3) the Benton uplift, and (4) the southern central thrust belt (Fig. 1). This subdivision is based on similar structural style and lithology, but is somewhat arbitrary because structural and lithologic changes are typically gradational. The rocks exposed range in age from the Cambrian Collier Shale to the Pennsylvanian Hartshorne Sandstone (Fig. 2). A brief description of each of the tectonic provinces summarized from Arbenz (1989) is provided below:

The southern part of the *Arkoma Basin* in Arkansas is a compressional foldbelt of Carboniferous rocks interpreted as both a structural and depositional basin. The northern margin of the west-striking basin is the autochthonous cratonic foreland. The southern part of the basin is dominated by thin-skinned thrust sheets forming broad synclines and narrow anticlines detached from lower Atokan and older rocks along a décollement. To the north, high-angle growth faults characterize a gradual passage from compressional to extensional "drape" folds. The southern boundary of the basin is the Ross Creek fault (Fig. 1) (Arbenz, 1984).

The *frontal imbricate zone* is a thrust belt marked by significant shortening of preorogenic and synorogenic turbidites. The thrust faults dip steeply at the surface and flatten with depth to a south-dipping basement décollement. Many closely spaced, north-verging thrust faults characterize the western part of the belt; however, south-verging structures are more common to the east. In the eastern Ouachitas, the frontal imbricate zone is divided into an outer and an inner zone separated by the Y City fault (Fig. 1). This fault separates mostly Atoka Formation rocks to the north from mostly Jackfork Formation rocks to the south. The Maumelle chaotic zone (Viele, 1966) refers to the transition zone between the Benton uplift on the south and the frontal imbricate zone on the north.

The *Benton uplift* is an east-northeast-striking anticlinorium along which pre-Carboniferous rocks are exposed. The anticlinorium crosses the southern central thrust belt obliquely. The uplift exhibits numerous south-verging overturned folds, low-grade metamorphic facies, and a penetrative north-dipping cleavage overprinted by a crenulated cleavage that most workers interpret as evidence for a second deformation phase. Some workers consider the Benton uplift to be a large anticlinorium made up of a series of stacked fold-and-thrust nappes that extends westward into Oklahoma as the Broken Bow uplift (Viele, 1966; Nielsen and others,

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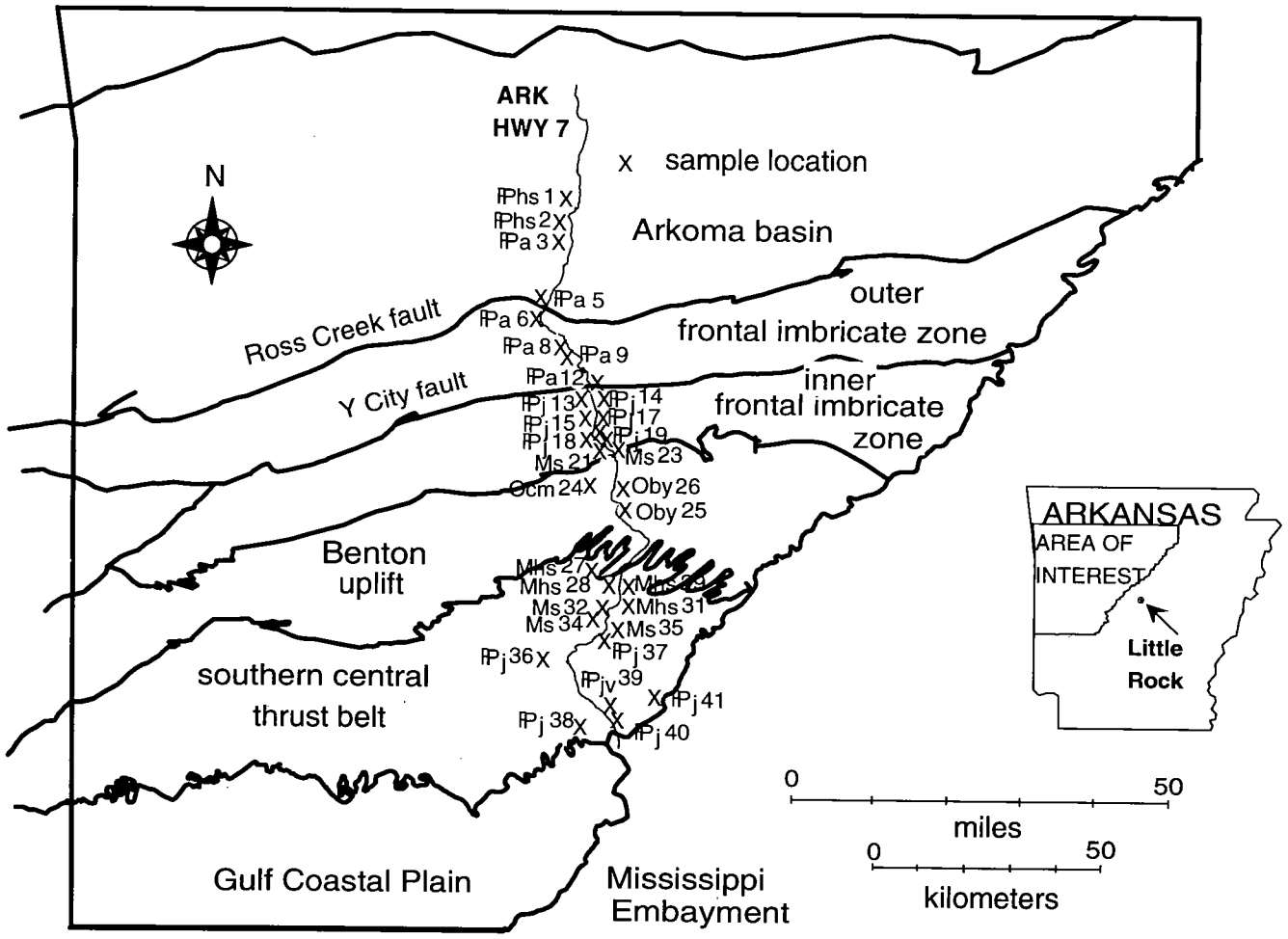


Figure 1. Sample location map with structural province boundaries modified from Arbenz (1984).

1989). The boundary between the Benton uplift and the adjacent tectonic provinces is usually drawn at the contact between the Arkansas Novaculite and the overlying Stanley Shale. The structures of the uplift, however, likely extend out into the surrounding Carboniferous strata.

The *southern central thrust belt* displays mostly northwest verging folded and faulted Carboniferous rocks. This belt is similar to the central belt north of the Broken Bow uplift in Oklahoma in the western Ouachitas and lies south of the Benton uplift in Arkansas in the eastern Ouachitas (Fig. 1).

Four distinct "stories" in the stratigraphic sequence in the *southern central thrust belt* exhibit contrasting mechanical behavior and structural style (Fig. 2). The stratigraphically highest (story 1) includes the succession of massive turbidite fan complexes of the upper part of the Stanley Shale through the Atoka Formation. The succession of fan complexes behaves mechanically as a single unit. Parts of the succession are deformed in flexural folds, and parts are broken by thrust faults. Story 2 (the next lower level) is an intensely deformed detachment zone composed of ductile shales and thin sandstones in the middle and lower parts of the Stanley Shale. Story 2 includes the Hot Springs Sandstone Member of the Stanley Shale. This sandstone is a quartz arenite under the classification of Dott (1964). Alternatively, the top of this ductile interval may represent a submarine angular unconformity (Babaei and Viele, 1992). The pre-Carboniferous sequence from the upper part of the Womble Shale (Ordovician) to the Arkansas Novaculite (Mississippian–Devonian) forms story 3 and displays folds with sharp hinges and planar limbs characteristic of well-layered rocks having high ductility contrasts. Story 4 is also ductile and extends upward from the Collier Shale through the interstratified Crystal Mountain and Blakely Sandstones to the Womble Shale. The sandstones in this story form fold trains, and the shales are intensely folded and sheared.

METHODS

I collected 33 samples along a strike-normal traverse roughly following Arkansas Highway 7 (Fig. 1). Effects of local deformation such as mesoscopic fold hinges or fault zones were avoided in the selection of sample sites. Thin sections from the samples were cut normal to bedding and parallel to dip. Microstructural abundances were determined by using a petrographic microscope and point-counting 250 grains per thin section along parallel, evenly spaced traverses. I counted all intragranular, intergranular, and transgranular structures associated with the grain beneath the crosshairs. Therefore, the

combined microstructural totals for each thin section may be greater than the number of grains counted. Assuming random distribution of the features being counted, the errors at a 95% confidence level associated with 10, 20, and 50 volume percent are 4%, 5%, and 6%, respectively (Van der Plas and Tobi, 1965). If a binomial distribution of features is assumed, the errors are 2.5%, 3%, and 3.7%, respectively (Koch and Link, 1970).

RESULTS

Observed Microstructures

Grain-scale deformation observed in the Ouachita rocks was accommodated by four mechanisms: (1) dislocation flow, (2) pressure solution, (3) recrystallization, and (4) microfracturing. Observed microstructures were identified by using criteria stated by Groshong (1988). Total microstructure abundance for all samples in the study is presented in Figure 3.

Microstructures related to dislocation flow include sweeping undulatory extinction, patchy extinction and subgrains, deformation bands, and deformation lamellae (Fairbairn lamellae). Undulatory extinction is caused by a bending of the crystal lattice of more than a few degrees. Patchy extinction results from multiple extinction domains with sharp boundaries, unlike the sweeping of the extinction domain across the grain in undulatory extinction. Subgrains have crystallographic misorientations of 1°–15°. Deformation lamellae are defined by parallel zones of differing birefringence. Deformation bands are 10–100-mm-wide, tabular, high-strain zones, where the crystal lattice has been rotated as much as 30°. The bands commonly are subnormal to deformation lamellae where they are identified in the same grain. These structures indicate crystal plasticity (Hirth and Tullis, 1992), which is the permanent deformation of single crystals by internal deformation mechanisms without loss of cohesion between crystals.

Microstructures indicative of recrystallization are serrated grain boundaries and mortar texture. Serrated boundaries are caused by thermally activated grain-boundary migration and denote the early stage of new subgrain development. They are transitional to mortar texture, identified on the basis of small grains formed along original grain boundaries. White (1976) identified the texture of well-developed subgrains and recrystallized grains at grain boundaries and relatively deformation-free interiors and called this "core-mantle structure."

Pressure-solution microstructures observed include sutured grain boundaries, overgrowth "beards," and transgranular solution surfaces (stylolites). Pressure solution is the result of diffusive mass transfer at low temperature in the presence of fluids.

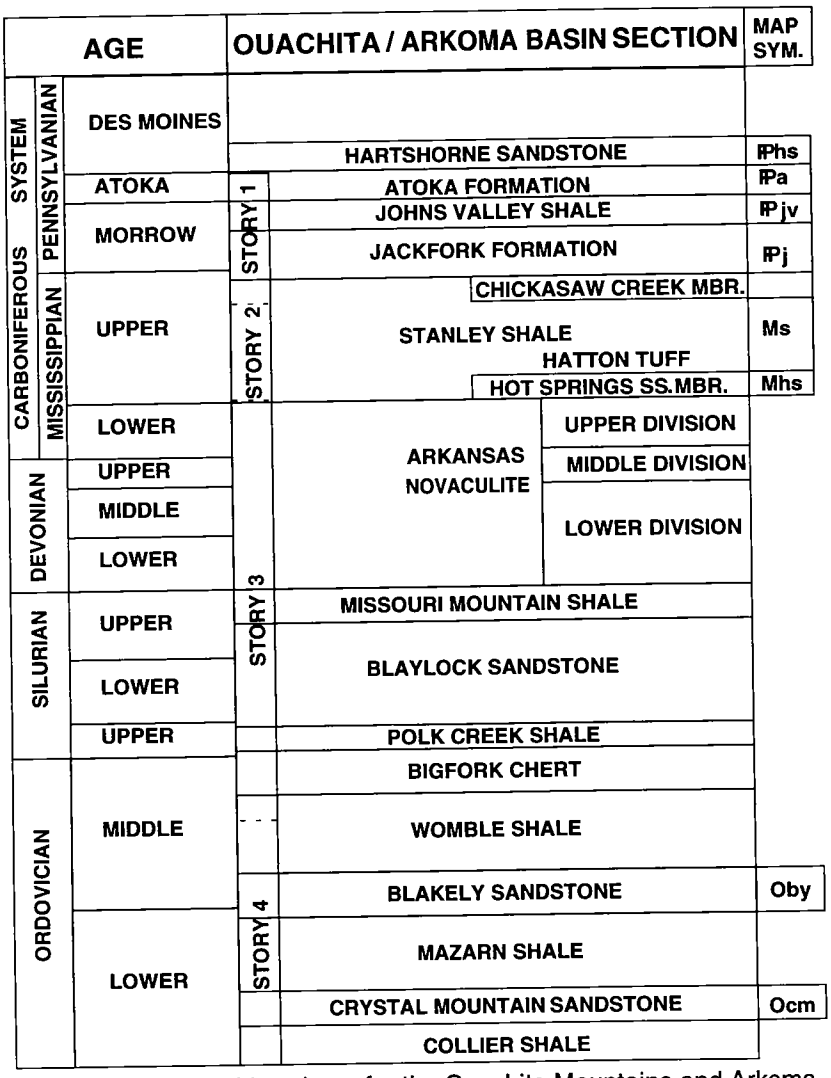


Figure 2. Stratigraphic column for the Ouachita Mountains and Arkoma Basin rocks (Stone and Haley, 1984). Stories 1–4 are the divisions of the southern central thrust belt referred to in the text.

Microstructures diagnostic of microfracturing include fluid-inclusion planes (FIPs), microveins, and cataclastic bands. FIPs, also known as Tuttle lamellae, are healed extensional microfractures decorated by fluid inclusions. They include both intragranular and transgranular types and occur in single or multiple orthogonal sets. Microveins are filled transgranular microfractures. Cataclastic bands are discordant, tabular, subplanar zones of fine-grained quartz.

The most commonly observed microstructures along the transect are those resulting from dislocation flow (undulatory extinction) and pressure solution (sutured grain boundaries) (Fig. 3). Grain-boundary recrystallization is the least important deformation mechanism.

Regional Variation in Microstructural Abundance

Important patterns of regional variation in the relative abundances of microstructures were observed (Fig. 4). Samples from the Arkoma Basin are relatively undeformed and have accommodated strain primarily by dislocation flow. Other mechanisms are relatively unimportant, although microfracturing (shown by FIPs) gradually

increases to the south. Overall microstructural abundance only slightly increases in samples from the outer frontal imbricate zone; however, this increase is particularly evident in sample Pa 8 in which all microstructural abundances—including grain-boundary recrystallization, dislocation flow, and pressure solution—are anomalously high. The microstructural abundance in Pa 8 is likely the result of local deformation associated with a mapped thrust fault where the sample was collected. Microstructural abundance within the frontal imbricate zone increases unevenly. Pressure solution increases the most except in the Stanley Shale samples, which show the highest percentage of undeformed grains. Undeformed-grain percentages decrease abruptly in the older rocks of the Benton uplift. Pressure solution increases greatly as does dislocation flow. These samples, along with samples from the extremely clean quartz arenite called the Hot Springs Sandstone Member, show the greatest relative microstructural abundances of all the samples collected. The samples from the southern central thrust belt, excluding the Hot Springs Sandstone Member, display little pressure solution. The exceptions are samples Pj 38 and Pj 40, which have a relatively constant abundance of dislocation flow microstructures. Samples Pjv 39 and Pj 41 show low percentages of pressure solution, microfracturing, and grain-boundary recrystallization.

DISCUSSION

The regional variations in microstructural abundance compares favorably to the accepted division of the Ouachita Mountains–Arkoma Basin region into tectonic provinces (Figs. 2,4). The Arkoma Basin is well defined by both relative abundance of all microstructures and the southerly increase in microfracturing, especially near the Ross Creek fault, the traditional boundary between the Arkoma Basin and the outer frontal imbricate zone in Arkansas. Variations in the relative abundances in the outer and inner frontal imbricate zone are likely a result of sample proximity to the major thrust faults, as well as how the different lithologies accommodated the deformation. The anomalously high percentage of undeformed grains and low microstructure abundance in the matrix-rich Stanley Shale sandstones as compared to adjacent samples likely reflect accommodation of strain by the matrix rather than deformation of the framework grains. Complex differences and similarities are displayed in the Jackfork Formation sandstone samples from the frontal imbricate zone and the southern central thrust belt. Dislocation flow and grain-boundary recrystallization show similar trends, whereas pressure solution and microfracturing vary greatly, showing a different set of operational deformation mechanisms. The similarities in microstructural abundance of the pre-Carboniferous rocks of the Benton uplift and the extremely clean quartz arenite of the Hot Springs Sandstone Member of the Stanley Shale show the extension of structures into the neighboring Carboniferous strata.

Most of the microstructures counted in the samples are assumed to have formed *in situ*. This assumption is valid for transgranular features, for microstructures caused by impinging grains, and for microstructures that transect contacts between grains and overgrowth cement (Onasch and Dunne, 1993). Many fluid-inclusion planes that cut across grains were noted, as well as deformation bands caused by impinging neighboring grains. Optically discontinuous cementation rims were seen to be deformed by impinging grains. In addition, contrasting abundance was seen to correlate with structural position. Some of the microstructures observed in this study are undoubtedly inherited from deformation in the grains' source terrane. A maximum of 5% of quartz grains affected by pressure solution during diagenesis has been generally assumed (Groshong, 1988; Knipe, 1989). Pressure-solution percentages in this study exceed this number by a significant margin.

The microstructures observed in the eastern Ouachita Mountains and their corresponding deformation mechanisms are typical of those formed under low-temperature conditions of deformation (Groshong, 1988; Lloyd and Knipe, 1992). Groshong (1988) introduced the concept of a deformation-mechanism association as a tool for interpretation of structural provenance and tectonic analysis. A deformation-mechanism association is defined by Groshong (1988) as the deformation mechanisms formed under the same environmental conditions (temperature, confining pressure, differential stress, strain rate, pore-fluid type, and pore pressure) in the same rock type. Five deformation-mechanism associations are recognized and are presented in Table 1 (from Groshong, 1988).

Microscopic deformation as observed in the present study belongs to Associations II (brittle-framework deformation; II P, II F, and II PF) and III (semibrittle framework deformation; III P, III F, and III PF). A few examples of Association IV R mechanisms are noted in the more highly deformed rocks of the Benton uplift and the northern part of the central thrust belt. Recrystalli-

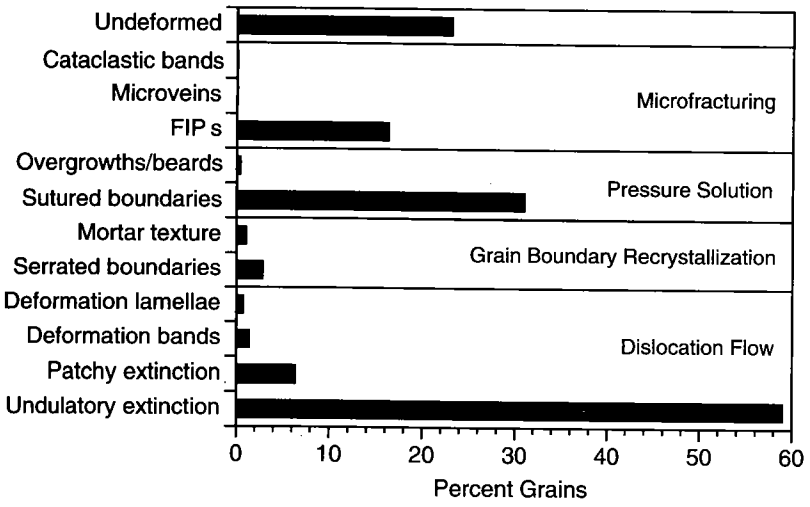


Figure 3. Histogram showing the abundances of microstructures in all sampled sandstones of the eastern Ouachita Mountains.

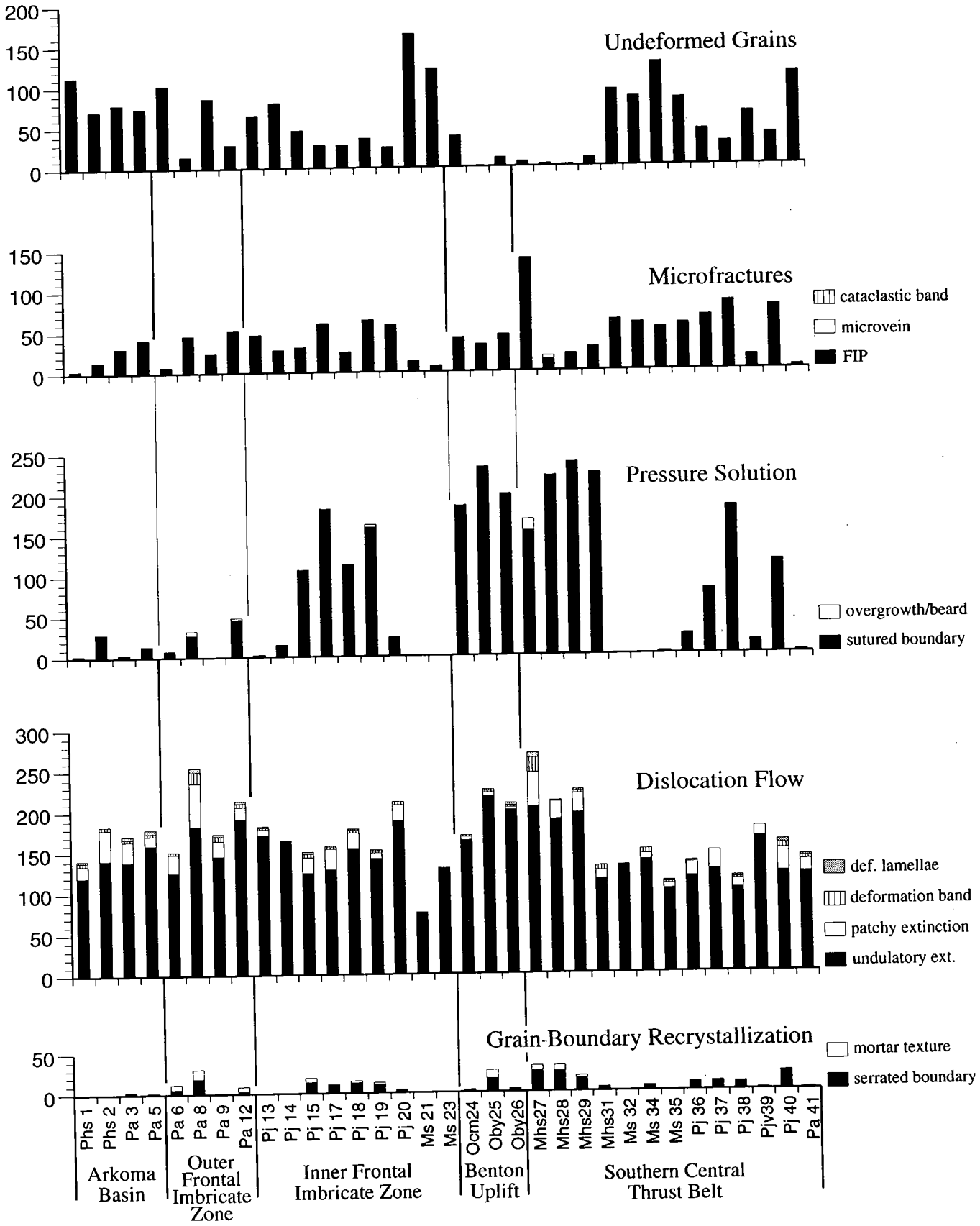


Figure 4. Histograms showing abundances of deformation mechanisms and resulting microstructures in each sample, grouped by structural province. Sample locations in Figure 1.

TABLE 1. — DEFORMATION MECHANISM ASSOCIATIONS
(after Groshong, 1988)

Association	Characteristics
I	Hydroplastic deformation; cohesionless particulate flow without deformation of framework minerals.
II	Brittle-framework deformation; crystal-plastic deformation small with slight undulatory extinction and rare deformation lamellae. Crystal deformation by fracturing or faulting (II F) or pressure solution (II P) or both (II FP).
III	Semibrittle framework deformation; obvious crystal-plastic glide insufficient to produce grain-orientation fabric; obvious undulatory extinction and/or deformation lamellae along with fracture (III F), pressure solution (III P), or both (III FP).
IV	Tectonites; deformation induced crystallographic fabric; glide tectonites (IV G) caused by glide mechanisms and recrystallization tectonites (IV R) in which original grain boundaries are not preserved.
V	Annealed; no more than weak crystallographic fabric and original grain boundaries obliterated.

zation noted in Association IV marks the upper limit of low-temperature deformation. No deformation mechanisms of Association V (annealed grain boundaries formed under static conditions) were recognized. A crystal-deformation front cannot be defined because no Association I (hydroplastic) deformation mechanism was noted. Frictional grain-boundary sliding indicative of hydroplastic deformation is difficult to identify in naturally deformed rocks because grain shape, size, and internal structure may be unaffected by the deformation (Knipe, 1989). A crystal-plastic front exists between the rocks of the frontal imbricate zone and the Benton uplift. This front cannot be defined precisely between samples Ms 23 and Ocm 24 or samples Mhs 31 and Ms 32 because of the change in rock type. All the samples studied exhibit significant Association III F mechanisms; Association III FP is evident in the more arenaceous samples from the inner frontal imbricate zone, the Benton uplift, and the southern central thrust belt. Association IV R is noted in these same rocks.

In a previous study of microstrain features of rocks of the Benton uplift (Karlo, 1977), little evidence of microscopic strain was reported in the sandstones studied. Evidence of plastic flow and tectonically induced shape change were not reported. Microstructures indicating pressure solution were lacking, and undulatory extinction was uncommon. In the shaly sandstones, the grains were reported to be unfractured. Karlo (1977) sought to explain the "ductility" noted in rocks of the Benton uplift by theorizing the presence of pore fluids under pressure. These results differ from the present study, in which intragranular deformation is common in the samples with the greatest relative deformation. Evidence of crystal-plastic deformation was seen in all samples; dislocation flow is a very important mechanism in samples from the Benton uplift and the northern part of the central thrust belt.

CONCLUSIONS

Microstructural abundances observed in this study indicate brittle and semibrittle deformation regimes. Strain caused by crystal-plastic glide (Association II) is taken to be 0–2% (Groshong, 1988). This strain is sufficient to form the mesoscopic structures seen in the rocks in the Arkoma Basin and the frontal imbricate zone and the Carboniferous rocks of the central thrust belt, excluding the Hot Springs Sandstone Member. As deformation-mechanism associations are related to rock type, a deformation front cannot be mapped in formations containing contrasting lithologies. The rocks of the Benton uplift and the northern part of the southern central thrust belt were deformed in a semibrittle deformation regime denoting 2–15% strain (Groshong, 1988). This strain amount is in general agreement with most models of tectonic history of the Ouachita Mountains in Arkansas. Tectonites of Association IV are rare, and low-temperature deformation is therefore suggested.

The opportunity exists for an expanded study to take more samples of appropriate lithology along the Highway 7 transect and transects to the east and west, in rocks in Arkansas and Oklahoma. The information on how deformation is accommodated at the microscopic scale is needed as quantitative data to support current ideas of formation of the Ouachita Mountains. Deformation-mechanism maps could be constructed with sufficient data. These maps would identify regions in stress–temperature and stress–grain size space where different rocks are dominated by different deformation-mechanism associations.

ACKNOWLEDGMENTS

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Thermal Maturity of Surface Samples from the Frontal and Central Belts, Ouachita Mountains, Oklahoma

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ABSTRACT.—The frontal and central belts of the Ouachita Mountains, Oklahoma, contain oil, gas, and solid hydrocarbons (asphaltite and asphaltic pyrobitumens). The presence of hydrocarbons implies available mature hydrocarbon source rocks. This paper documents the level of thermal maturity of potential hydrocarbon source rocks of Ordovician to Pennsylvanian age in the study area.

The level of thermal maturity was determined by measuring the random vitrinite reflectance of 36 surface rock samples of post-Silurian age and the random graptolite reflectance of two surface rock samples of Ordovician age. The mean random vitrinite reflectance ($\bar{R}_o$) ranges from 0.47% to 1.12%. The mean random graptolite reflectance is 0.66% and 0.82%.

All the samples in this study are early mature with respect to the generation of liquid hydrocarbons ("oil window") and thus have the potential of generating and preserving liquid hydrocarbons.

INTRODUCTION

A cooperative geologic mapping program (COGEOMAP) of the Ouachita Mountains was begun in 1985 by the U.S. Geological Survey, Arkansas Geological Commission, and Oklahoma Geological Survey. An evaluation of potential hydrocarbon source rocks was included in the Oklahoma part of the program. Hydrocarbon source rocks are characterized by the quantity and type of organic matter and level of thermal maturation (Tissot and Welte, 1984). The objective of this study was to determine the thermal maturity by organic petrography (vitrinite-reflectance analysis) and the quantity of organic matter by organic geochemistry (total organic carbon and extractable organic-matter analyses) of shale and chert samples collected in conjunction with detailed surface mapping and during reconnaissance. Weber (1992) reported the results of quantitative organic-geochemical analyses of 50 surface samples. This paper presents the results of the vitrinite- and graptolite-reflectance analyses of those samples.

PREVIOUS INVESTIGATIONS

Curiale (1983), Fallin (1985), and Suneson and Campbell (1990) summarized the occurrence of oil, gas, and solid hydrocarbons (asphaltite and asphaltic pyrobitumen) in the Ouachita Mountains of Oklahoma. The presence of hydrocarbons implies available mature hydrocarbon source rocks. The section comprising the Ordovician Womble Shale through Devonian–Lower Mississippian Arkansas Novaculite contains oil-generative, type II kerogen, and the section comprising the Arkansas Novaculite through Middle Pennsylvanian Atoka Formation contains gas-generative, type III kerogen (Curiale, 1983; Curiale and others, 1984; Fowler and Douglas, 1984; Zemmels and others, 1985; Guthrie and others, 1986). On the basis of plots of extractable organic matter (EOM $\geq$60 ppm) vs. total organic carbon (TOC $\geq$0.5 wt%), Weber (1992) concluded that, of the surface samples analyzed in the study, one Woodford Shale (Upper Devonian–Lower Mississippian) sample, two Caney Shale (Mississippian) samples, and one Atoka Formation sample had fair to good oil-source potential. However, these outcrop-sample, organic-richness measurements may be minimum values (Clayton and Swetland, 1978).

Houseknecht and Matthews (1985) illustrated an isoreflectance map of the Ouachita Mountains based on 89 surface samples of primarily Carboniferous strata. Thermal maturity of the Oklahoma segment, based on 32 samples, was shown by $\bar{R}_o$ values ranging from 0.24% in the west to 3.96% in the core area (Matthews, 1982). Vitrinite-reflectance values in the present study area are from 0.24% to 1.16%. Houseknecht and Matthews (1985, p. 344) concluded that "in the Oklahoma portion of the Ouachitas, thermal maturity increases with stratigraphic age, thermal-maturity contours parallel structural grain,

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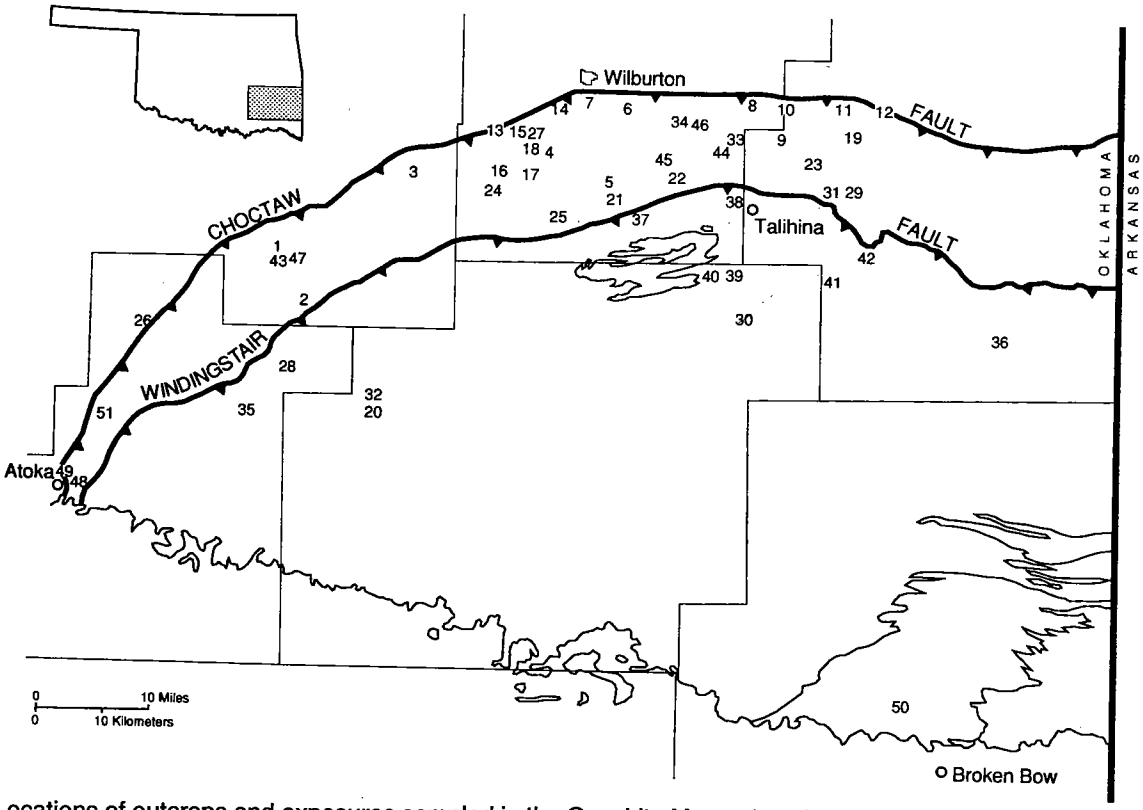


Figure 1. Locations of outcrops and exposures sampled in the Ouachita Mountains, Oklahoma (modified from Weber, 1992).

and older strata on upthrown sides of thrust faults are more thermally mature than younger strata on downthrown sides. Sedimentary burial appears to have exerted a dominant influence on thermal maturation, although tectonic burial was also important." Guthrie and others (1986), using the samples from Houseknecht and Matthews (1985), established the relations among vitrinite reflectance, illite crystallinity, and organic geochemistry as thermal-maturity indicators.

By combining structural mapping with vitrinite-reflectance measurements, Underwood and others (1988) described three types of thermal-maturity anomalies (maturity inversion, shear-heating aureoles, and abnormal maturity gradients) within the frontal belt of the Ouachita Mountains in Arkansas.

SAMPLES AND ANALYTICAL METHODS

Staff members of the Oklahoma Geological Survey collected 51 grab samples of indurated shale and chert from outcrops and exposures of the Collier Shale (Upper Cambrian–Lower Ordovician), Bigfork Chert and Polk Creek Shale (Upper Ordovician), Arkansas Novaculite (Devonian–Lower Mississippian), Woodford Shale (Upper Devonian–Lower Mississippian), Stanley Group and Caney Shale (Mississippian), Jackfork Group (Lower Pennsylvanian), Springer Formation (Upper Mississippian–Lower Pennsylvanian), Johns Valley Shale (Lower Pennsylvanian), and Atoka Formation (Middle Pennsylvanian) in Oklahoma (Fig. 1, Table 1).

Approximately 100 g of each sample was crushed to −20 mesh (<0.85 mm), which is the minimum grain size preferred for petrographic analysis, whereas finer samples (higher-number mesh sieves) are used for complete chemical maceration. Kerogen (insoluble organic matter) was isolated with HCl and HF acids by using standard procedures.The dried kerogen concentrate was mixed with epoxy and prepared as 32-mm-diameter pellets. The sample surface of the pellets was ground with 400- and 600-grit silicon carbide papers and polished with 0.3- and 0.05-μm aluminum oxide slurries. The pellets were stored in a desiccator for at least 15 h before performing the reflectance analysis; the reflectance analysis was completed within two days after polishing to avoid the problem of surface oxidation.

The reflectance analysis measures the percentage of incident white light reflected from a polished surface. Random vitrinite reflectance (plane-polarized light, fixed stage) was measured on all post-Silurian samples with a Vickers M17 research microscope system, with a Gamma Scientific DR-2 digital radiometer and D47A photomultipler detector, at 500× magnification, in monochromatic green light (546 nm), with immersion oil (n_e = 1.5180), and calibrated with glass standards following ASTM (1992, D 2798). Random graptolite reflectance was measured on the Bigfork Chert and Polk Creek Shale samples with the same system and under the same conditions. Cardott and Kidwai (1991) summarized the use of graptolite reflectance as a thermal-maturation indicator for Ordovician samples. The Collier Shale sample was not analyzed petrographically owing to lack of vitrinite or graptolites.

TABLE 1. —VITRINITE REFLECTANCE, GRAPTOLITE REFLECTANCE, AND ORGANIC GEOCHEMICAL DATA

Sample	OPL[a]	Location[b]	$\bar{R}_o$ (%)[c]	R_o range (%)	n[d]	S.D.[e]	TOC (wt%)[f]	EOM (ppm)[g]
				Atoka Formation				
1	637	32-3N-15E	0.65	0.49–0.79	46	0.08	0.95	167
2	639	21-2N-15E	0.81	0.57–0.96	58	0.08	0.66	87
3	669	35-4N-16E	0.70	0.57–0.82	35	0.06	0.22	10
4	670	23-4N-18E	0.91	0.72–1.04	34	0.08	0.41	20
5	671	28-4N-19E	0.94	0.81–1.07	48	0.07	0.51	14
6	672	32-5N-20E	NA[h]	NA	NA	NA	0.32	10
7	673	29-5N-19E	0.76	0.58–0.91	68	0.09	0.89	160
8	680	23-5N-21E	0.93	0.75–1.03	55	0.07	0.62	16
9	681	5-4N-22E	0.90	0.75–1.03	37	0.08	0.59	10
10	682	21-5N-22E	NA	0.78–0.99	10	NA	0.78	10
11	683	21-5N-23E	NA	0.78–1.06	7	NA	1.20	63
12	684	30-5N-24E	0.92	0.80–1.04	34	0.07	0.49	10
13	743	2-4N-17E	0.52	0.41–0.66	110	0.05	1.00	116
14	744	34-5N-18E	0.61	0.56–0.68	39	0.03	0.42	42
15	746	5-4N-18E	0.60	0.49–0.71	37	0.06	0.50	134
16	747	24-4N-17E	0.78	0.65–0.89	37	0.08	1.50	168
17	749	28-4N-18E	NA	0.70–0.92	7	NA	1.20	146
18	757	21-4N-18E	NA	0.65–1.06	9	NA	0.74	30
19	925	4-4N-23E	1.02	0.85–1.16	38	0.08	1.80	1,212
				Johns Valley Shale				
20	648	9-1S-16E	0.86	0.71–1.00	51	0.09	0.34	65
21	674	3-3N-19E	0.96	0.82–1.06	49	0.06	0.30	10
22	676	26-4N-20E	0.92	0.76–1.06	50	0.08	0.77	38
23	686	24-4N-22E	0.97	0.81–1.08	50	0.06	0.45	15
24	748	2-3N-17E	0.71	0.56–0.83	31	0.07	3.00	356
25	755	12-3N-18E	0.82	0.75–0.92	30	0.05	2.40	573
				Springer Formation				
26	636	1-1N-12E	NA	0.54–0.85	8	NA	0.52	50
27	745	5-4N-18E	NA	0.46–0.77	12	NA	0.30	75
				Jackfork Group				
28	641	19-1N-15E	0.65	0.46–0.81	100	0.07	2.00	310
29	685	34-4N-23E	1.11	0.93–1.28	46	0.08	0.82	22
30	689	25-2N-21E	NA	NA	NA	NA	0.29	10
31	924	31-4N-23E	1.04	0.91–1.19	44	0.08	0.49	52
				Caney Shale				
32	649	4-1S-16E	0.53	0.40–0.67	39	0.08	5.40	1,570
33	678	11-4N-21E	NA	0.51–0.64	2	NA	4.60	6,294
34	810	6-4N-21E	NA	0.45–0.75	15	NA	4.40	2,816
				Stanley Group				
35	643	9-1S-14E	NA	0.51–0.79	17	NA	0.39	120
36	647	26-2N-25E	1.12	0.91–1.27	40	0.10	0.94	170
37	675	16-3N-20E	0.64	0.52–0.79	50	0.06	0.32	46
38	677	2-3N-21E	0.74	0.62–0.83	32	0.06	0.19	32
39	687	1-2N-21E	0.73	0.63–0.85	33	0.06	0.24	49
40	688	2-2N-21E	1.01	0.85–1.18	44	0.08	0.17	12
41	690	7-2N-23E	NA	0.90–1.04	9	NA	0.23	41
42	691	33-3N-23E	0.98	0.84–1.08	30	0.06	0.22	44
				Woodford Shale				
43	638	5-2N-15E	0.52	0.39–0.69	81	0.07	3.30	1,153
44	679	11-4N-21E	0.63	0.51–0.79	40	0.09	3.60	2,436
45	809	15-4N-20E	0.64	0.55–0.76	35	0.06	2.00	514
46	811	6-4N-21E	0.62	0.52–0.76	35	0.08	8.50	3,342
47	914	4-2N-15E	0.52	0.45–0.67	34	0.05	5.30	2,248
				Arkansas Novaculite				
48	645	13-2S-11E	0.47	0.37–0.57	50	0.05	12.50	3,300
				Polk Creek Shale				
49	644	14-2S-11E	0.66 (gr)	0.52–0.77	50	0.06	6.10	980
				Bigfork Chert				
51	920	9-1S-12E	0.82 (gr)	0.58–1.00	60	0.10	NA	NA
				Collier Shale				
50	646	17-5S-24E	NA	NA	NA	NA	0.29	50

[a]Oklahoma Geological Survey Organic Petrography Laboratory sample number.
[b]Section-Township-Range.
[c]Mean random vitrinite (gr = graptolite) reflectance.
[d]Number of reflectance measurements.
[e]Standard deviation.
[f]Total organic carbon (weight percent; from Weber, 1992).
[g]Extractable organic matter (parts hydrocarbons per million parts rock; from Weber, 1992).
[h]NA = Not available.

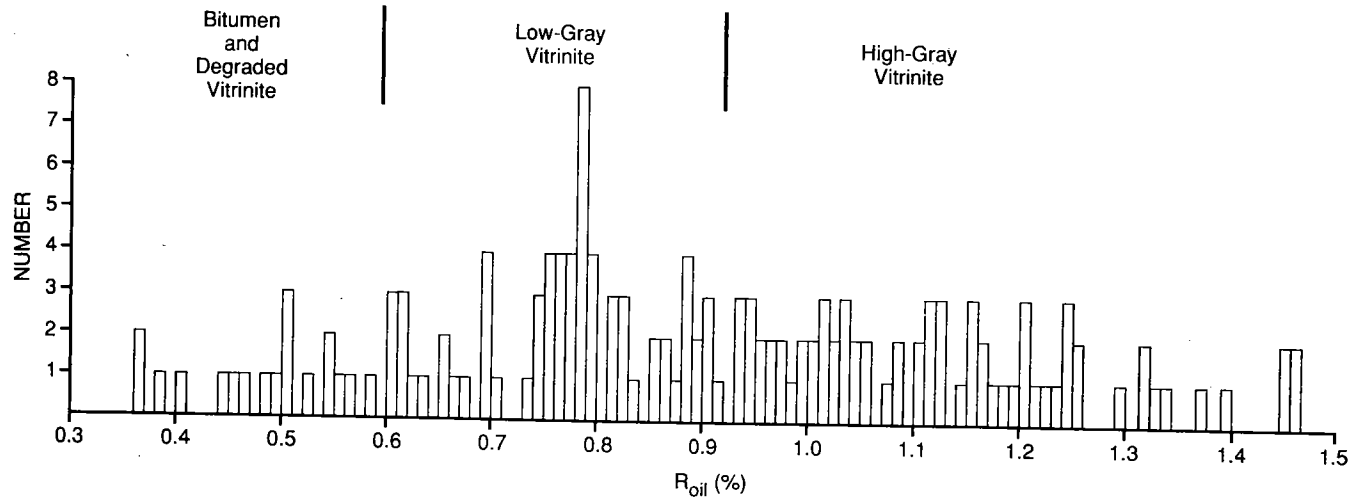


Figure 2. Vitrinite-reflectance histogram of sample 7 (OPL 673; Atoka Formation) showing the distribution of low-gray (0.58–0.91% R_o) and high-gray (0.93–1.46% R_o) vitrinite.

RESULTS

The mean random reflectance value, reflectance range, number of measurements, and standard deviation of low-gray vitrinite or graptolite are presented in Table 1; low-gray vitrinite (described by Bostick, 1979) is the indigenous vitrinite population interpreted as the lowest-reflecting vitrinite on a reflectance histogram (bar plot of reflectance vs. frequency). The vitrinite particles range in size from about 10 to 100 μm and are usually <20 μm. Particles <7 μm (measuring-spot size) were not measured. Barker and Pawlewicz (1993) recommended a minimum of 20–30 reflectance measurements for statistically valid mean vitrinite-reflectance values. A minimum of 30 reflectance measurements from high-quality (nonpitted, nonweathered) particles was used in this study. Several samples (6, 10, 11, 17, 18, 26, 30, 35, 41, and 42) were vitrinite deficient with fewer than 30 reflectance values after four analyses (two pellets polished twice) and were determined to be invalid for thermal-maturity evaluation. Some samples (e.g., numbers 27, 33, and 34) were not analyzed completely because nearby samples were (numbers 15, 44, and 46, respectively).

High-gray vitrinite is recycled from older rocks and has a higher reflectance than low-gray vitrinite (Bostick, 1979). On a reflectance histogram, low-gray vitrinite usually has a Gaussian distribution with reflectance values (R_o) spanning ~0.3%, beginning from the lowest-reflecting, high-quality vitrinite grain, at maturities indicated by R_o values of <1.0%; the distribution increases with increasing maturity. High-gray vitrinite is identified by R_o values in a reflectance histogram that are higher than those in the tail of the low-gray vitrinite distribution. Figure 2 is a reflectance histogram of an Atoka Formation sample showing the distribution of low-gray and high-gray vitrinite. The separation between low-gray and high-gray vitrinite is at 0.92% R_o. Figure 3 is a photomicrograph of an Atoka Formation sample showing the broad reflectance distribution of low-gray and high-gray vitrinite. The Pennsylvanian samples had an abundance of high-gray vitrinite, redeposited from older rocks during the Ouachita orogeny. Houseknecht and Matthews (1985) did not identify and remove recycled vitrinite from their reflectance histograms. They indicated that, for that reason, their mean vitrinite-reflectance values were probably higher than accurate values.

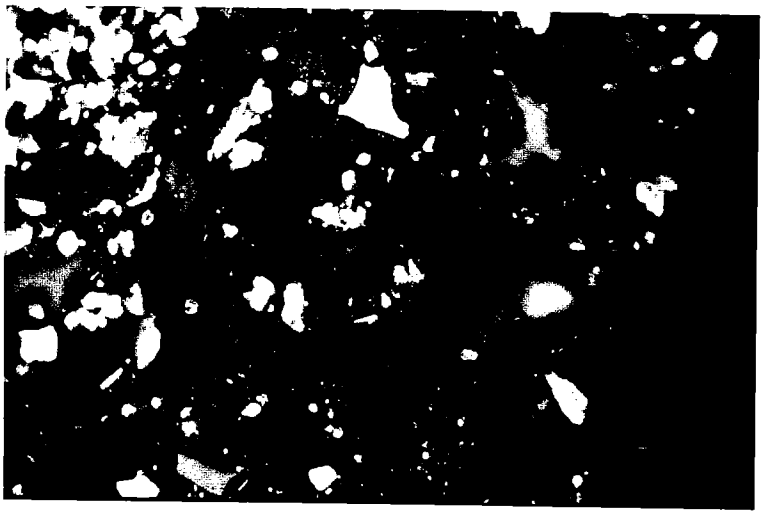


Figure 3. Photomicrograph of sample 16 (Atoka Formation) showing the reflectance variability of low-gray (0.86% R_o) and high-gray vitrinite (1.10% R_o) (reflected white light, 140-μm field width).

Because all the samples are from the surface, the probability that some of the samples are weathered is great. Lo and Cardott (1994) found that vitrinite reflectance decreased by as much as 0.2% R_o (absolute) in naturally weathered surface samples from southern and eastern Oklahoma. Vitrinite grains showing petrographic signs of weathering (microfractures, dark reaction rims; Fig. 4) were not included in the tabulation of mean random reflectance.

The spread of vitrinite-reflectance measure-

ments is from 0.12% to 0.41%; graptolite-reflectance measurements have a spread of 0.25% and 0.42% (Table 1). A standard deviation of 0.07 or less is ideal for mean vitrinite-reflectance values of <1.0% measured on kerogen concentrate samples (Hunt, 1979). The standard deviation of mean vitrinite-reflectance values in this study ranges from 0.03 to 0.10. The spread of reflectance values and standard deviations increase with increasing thermal maturity. Even though high-gray vitrinite values were selectively removed from the reflectance histograms, samples with a standard deviation of >0.08 and a reflectance (R_o) range of >0.3% may have slight overlap with high-gray vitrinite. Graptolites have biaxial optical character with stronger anisotropy, broader reflectance range, and higher standard deviation than vitrinite (Goodarzi, 1990).



Figure 4. Photomicrograph of sample 24 (Johns Valley Shale) showing microfractures and dark reaction rims of weathered vitrinite (reflected white light, 90-μm field width).

DISCUSSION

The mean random vitrinite-reflectance (R_o) values of 36 post-Silurian surface samples from the study area range from 0.47% to 1.12% (Table 1). The mean random graptolite-reflectance values of the Polk Creek Shale and Bigfork Chert samples are 0.66% and 0.82%, respectively. Graptolite reflectance is not equivalent to vitrinite reflectance; however, the correlation is not known. Bertrand (1990) indicated that, at the same maturity level, graptolite reflectance is slightly less than vitrinite reflectance. Goodarzi (1990) and Goodarzi and others (1992) showed that graptolite reflectance is greater than vitrinite reflectance at the same maturity level. Goodarzi and others (1993) indicated that a graptolite reflectance of 0.7–1.7% is equivalent to a vitrinite reflectance of 0.5–1.3%.

The early-mature zone with respect to the generation of liquid hydrocarbons (i.e., the "oil window") correlates with a vitrinite reflectance of 0.5–1.3%. All the samples in this study are within the early-mature zone and have the potential for generating and preserving liquid hydrocarbons if they meet the requirement for organic richness (minimum of 0.5 wt% TOC) and contain oil-generative organic matter (kerogen types I or II). Weber (1992) showed that at least one sample from each formation had a minimum of 0.5 wt% TOC (Table 1).

Figure 5 is a map of vitrinite- and graptolite-reflectance values. The mean reflectance values vary spatially as a result of thrusting. There is a general trend of increasing thermal maturity from west to east in the northeastern part of the study area. However, there is not a general trend of increasing thermal maturity with increasing age, as reported by Houseknecht and Matthews (1985). Samples of younger strata (Atoka Formation, Johns Valley Shale, and Springer Formation) have higher mean vitrinite-reflectance values than nearby samples of older rocks (Caney Shale, Woodford Shale, and Arkansas Novaculite). In addition to attributing these reflectance values to depositional and tectonic burial (preorogenic and synorogenic coalification), the abundance of high-gray vitrinite in the Pennsylvanian samples could have skewed the mean vitrinite reflectance to a higher value (e.g., sample 2 has a reflectance range of 0.39%). Furthermore, Mississippian sample 40 (1.01% $\bar{R}_o$) may contain only high-gray vitrinite, a conclusion based on a comparison with sample 39 (0.73% $\bar{R}_o$), which was collected nearby from the same unit (Stanley Group). Variation of phytoclast size and quantity can also affect the mean value. Within the Stanley Group, sample 39 had easily recognizable low-gray vitrinite, whereas samples 40 and 42 had many small phytoclasts. Moreover, the vitrinite-reflectance values of pre-Mississippian rocks containing type II kerogen could be suppressed (Price and Barker, 1985).

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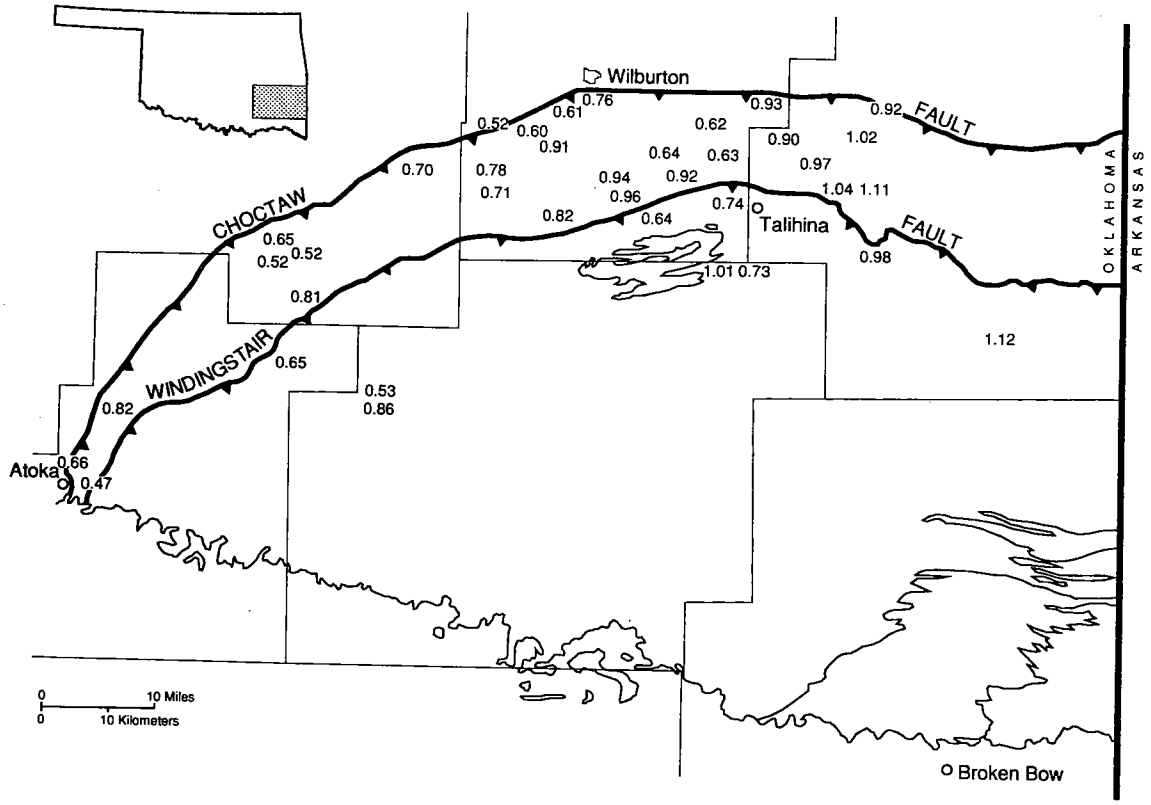


Figure 5. Map of mean random vitrinite- and graptolite-reflectance values.

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Subsurface Temperatures in the Arkoma Basin, Southeastern Oklahoma

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ABSTRACT.—We collected 345 bottom-hole temperatures (BHTs) from log headers of 199 oil and gas wells located on a N-S cross-section through the Arkoma Basin, southeastern Oklahoma. Fifty-six sets of multiple time-temperature measurements were corrected by means of a Horner plot. An average depth-time correction scheme was applied to the remaining 221 BHTs that were not suitable for a Horner plot correction. A total of 277 corrected BHTs was used to estimate equilibrium formation temperatures in the Arkoma Basin.

The average geothermal gradient is 23°C/km. However, there is considerable lateral variation. The average geothermal gradient in the southern part of the basin is 20°C/km, exceeds 30°C/km in the middle part of the basin, and is 24°C/km on the northern part of the basin. This variation in geothermal gradients could be due to the changes in thermal conductivity, changes in radioactive heat generation in basement rocks, or heat transport in the basin arising from ground-water flow.

INTRODUCTION

Oil and gas are produced by the thermal maturation of organic matter in sedimentary basins (Tissot and others, 1987; Mackenzie and Quigley, 1988). Understanding the thermal history of a sedimentary basin is therefore crucial to understanding the timing and extent of hydrocarbon generation. Because subsurface temperature is sensitive to ground-water movement, studies of past and present temperature can also provide insight into the nature of regional ground-water flow systems that may play a crucial role in oil migration (Hubbert, 1953; Hitchon, 1984; Toth, 1988; Garven, 1989). Knowledge of these factors enhances our theoretical understanding of the processes of hydrocarbon generation and migration and provides practical information that can be incorporated into exploration strategies for specific basins.

A logical starting point for a thermal history study is the present-day thermal state. Knowledge of the present-day thermal state, based on direct measurements of temperature, provides a solid starting point for extrapolating back in time. The present-day thermal regime is also indicative of hydrodynamic processes within a basin that affect temperature and redistribute heat. Once the present-day thermal state is understood, the thermal analysis can be extended into the past through the interpretation of paleothermal indicators such as vitrinite reflectance (e.g., Deming and Chapman, 1989).

In this paper we report preliminary results of our study of present-day temperature in the Arkoma Basin in southeastern Oklahoma. The Arkoma Basin is a prolific natural-gas province (Houseknecht and others, 1992), and as of the end of 1988, there were 22 oil and gas fields located in the nearby Ouachita Mountains (Suneson and others, 1990; Tilford, 1990).

GEOLOGIC SETTING

The Arkoma Basin is a peripheral foreland basin associated with the Ouachita fold belt of southern Arkansas and southeastern Oklahoma. It is bounded to the north and northwest by the Ozark uplift and the central Oklahoma platform, respectively, and to the south by the Choctaw fault and in Arkansas by the Ross Creek fault (Johnson and others, 1988).

Figure 1 shows the location of a geologic cross section (A–A′) through the Arkoma Basin; the cross section and a generalized stratigraphic column are shown in Figures 2 and 3, respectively. For detailed summaries of the geologic history and setting of the Arkoma Basin, the reader is referred to Amsden (1980), Houseknecht and others (1983), Zachry and Sutherland (1984), Johnson and others (1988), Sutherland (1988), and Vanarsdale and Schweig (1990).

TEMPERATURE DATA

The type of temperature data that is usually available from sedimentary basins is bottom-hole temperatures (BHTs), temperatures measured at or near the bottom of

Lee, Youngmin; Chen, K. F.; and Deming, David, 1994, Subsurface temperatures in the Arkoma Basin, southeastern Oklahoma, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 277–282.

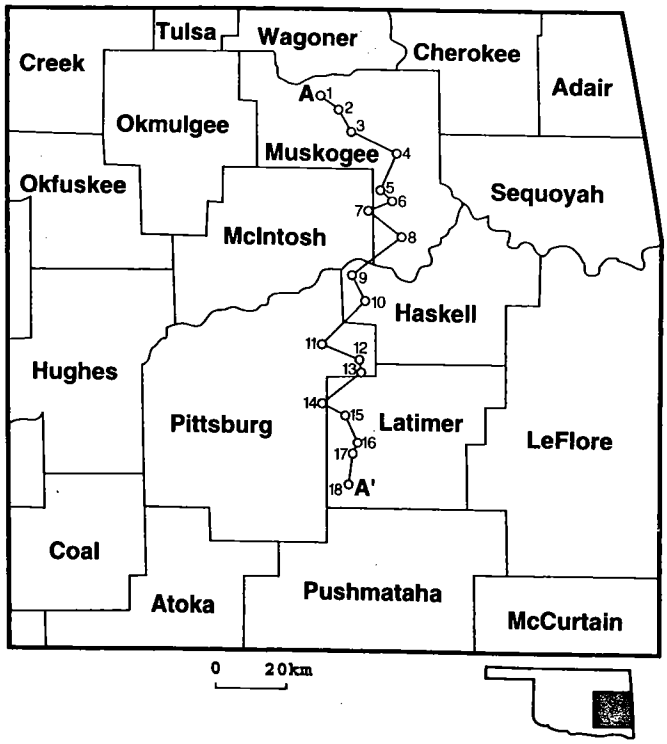


Figure 1. Location map showing cross section A–A' and wells used in construction of cross section (after Johnson, 1986).

a borehole during geophysical logging. Unfortunately, BHTs are of poor quality. They tend to be lower than equilibrium formation temperatures because of the cooling effect of drilling fluid circulating at the bottom of boreholes. Corrections can be made for drilling disturbances, but the information needed to make accurate corrections is usually not available (Deming, 1989). It is therefore necessary to take the presence of error into account when trying to estimate equilibrium formation temperature from raw BHTs.

In general, the process of reconstructing basin temperature from analysis of BHT data requires three steps. First, raw data are extracted from well-log headers and screened. At this stage, temperatures that appear to be abnormally high or low are excluded. Temperatures measured at a shut-in time of less than 4 h after the cessation of drilling fluid circulation are also excluded. Second, raw BHTs are corrected for drilling disturbances. A number of correction schemes are available (e.g., Bullard, 1947; Lachenbruch and Brewer, 1959; Luheshi, 1983; Shen and Beck, 1986); these were reviewed by Deming (1989) who offered practical recommendations. Rarely, however, does sufficient information exist to allow accurate corrections to be made; the best that can usually be hoped for is to reduce the systematic bias introduced by the drilling disturbance. A third stage of interpretation is essential. In the third stage, the corrected data are averaged through some interpretive model that

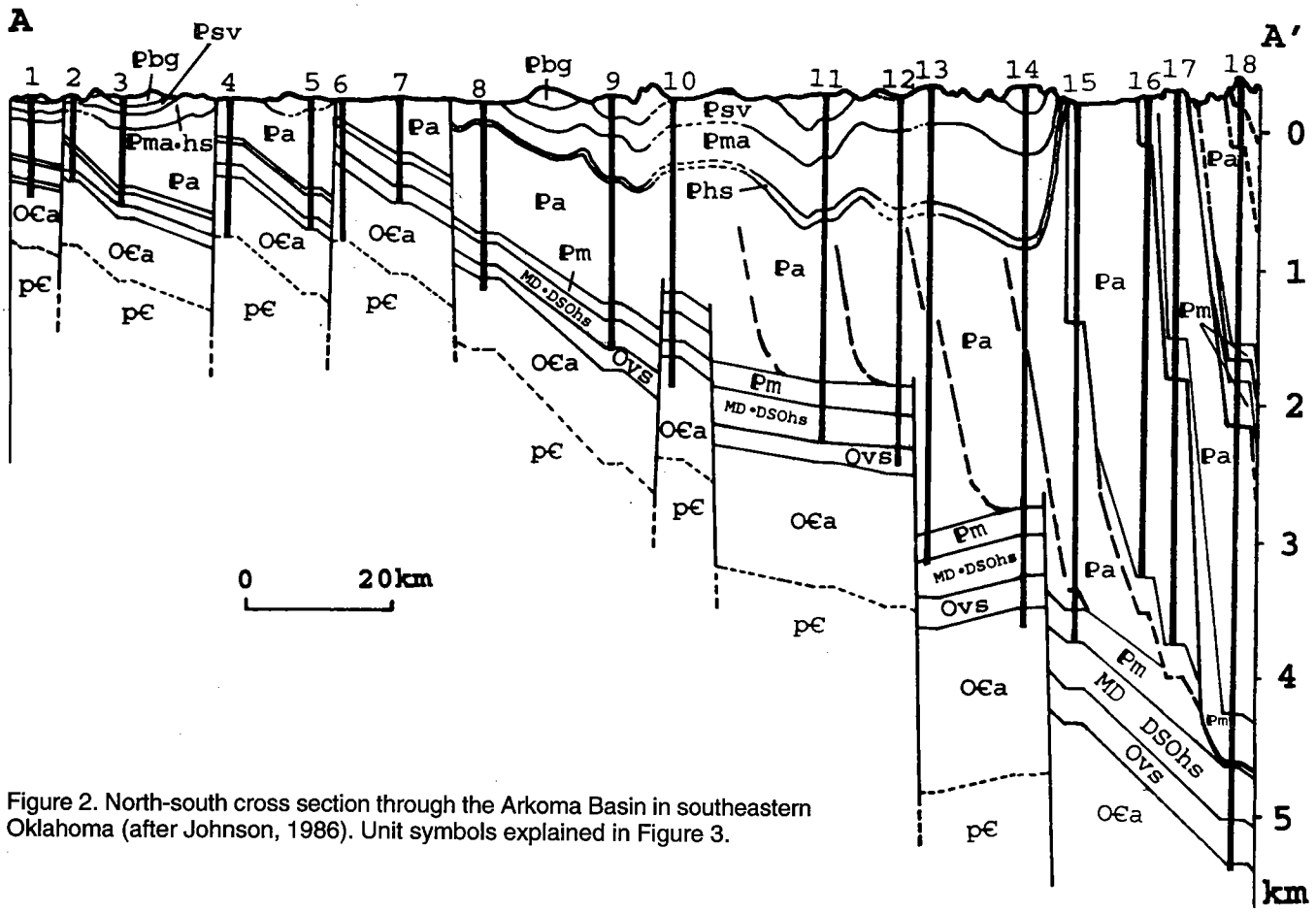


Figure 2. North-south cross section through the Arkoma Basin in southeastern Oklahoma (after Johnson, 1986). Unit symbols explained in Figure 3.

	SERIES	FORMATION		
PENNSYLVANIAN	Desmoinesian	Krebs Gp.	Boggy Fm.	Pbg
			Savanna Fm.	Psv
			McAlester Fm.	Pma
			Hartshorne Upper / Lower	Phs
	Atokan	Atokan Fm.		Pa
	Morrowan	Wapanucka Ls. Limestone Gap Sh. Union Valley Ls. Cromwell ss.		Pm
MISSISSIPPIAN	Chesterian	Goddard Fm.		MD
	Meramecian	Delaware Creek Sh. (="Caney") Welden Ls.		
	Osagean			
	Kinderhookian	Woodford Sh.		
DEVONIAN	Upper			
	Lower	Hunton Gp.	Frisco Ls. Bois d'Arc Ls. Haragan Ls.	DSOhs
SILURIAN	Niagaran		Henryhouse Fm.	
	Alexandrian		Chimneyhill Subgroup	
		Sylvan Sh.		
ORDOVICIAN	Upper	Viola Gp.	Welling Fm. Viola Springs Fm.	Ovs
	Middle	Simpson Gp.	Bromide Fm. Tulip Creek Fm. McLish Fm. Oil Creek Fm. Joins Fm.	
	Lower	Arbuckle Gp.	West Spring Creek Fm. Kindblade Fm. Cool Creek Fm. Mckenzie Hill Fm. Butterly Dol.	O€a
CAMBRIAN	Upper		Singnal Mountain Ls. Royer Dol. Fort Sill Ls.	
		Timbered Hills Gp	Honey Creek Ls. Reagan Ss.	
PRECAMBRIAN		Granite and Rhyolite		p€

Figure 3. Generalized stratigraphic column for the Arkoma Basin (after Cardott and others, 1986, p. 13).

reduces the random error in individual measurements. Of necessity, when the data are averaged, resolution suffers although error is reduced.

From examination of well-log headers on file at the Oklahoma Geological Survey, we collected 345 BHTs measured in Arkoma Basin wells located near north-south cross section A–A′ (Fig. 1). From these 345 values, we were able to extract 56 data sets representing multiple temperature measurements made in the same borehole at the same depth but at different times. These 56 sets of multiple time-temperature measurements were used to estimate equilibrium formation temperatures by applying a Horner plot correction with an assumed circulation time of 5 h. An assumed circulation time of 5 h was used because, on American well-log records, circulation time is usually not given. The remaining 221 BHTs were corrected by an average depth-time correction scheme described by Deming (1989), yielding a total of 277 estimates of formation temperature.

The average geothermal gradient within the Arkoma Basin determined by least-squares from 277 corrected BHTs is 23°C/km (Fig. 4). At a depth of ~4 km, the approximate spread in the data is 25°C. However, our *a priori* estimate of the average magnitude of random error in these corrected temperatures is ±7°C (see discussion in Deming, 1994). It follows that much of the divergence of the temperature estimates from the least-squares line probably represents variation in thermal state throughout the Arkoma Basin, not random error.

One of the limitations of the average-gradient model (Fig. 4) is its inability to reveal spatial changes in temperature along the cross section. Determining one average gradient for the entire basin averages random error, which is desirable, but also limits spatial resolution, which is undesirable. A more optimum trade-off between resolution and error reduction can be obtained with the method of variable bias as described by Deming and others (1990). The method of variable bias involves local determination of an average geothermal gradient from the n closest BHTs (n = number). The n closest BHTs are given higher weightings; the rest of the data set receives a lower weighting. As n becomes very large, all of the data are given nearly equal weighting. Thus, for large values of n, spatial resolution is low, but random error is also low. If $n = 1$, only the closest BHT is given appreciable weighting. In this case, spatial resolution is high, but random error is not averaged or reduced.

Figure 5 shows estimated subsurface temperatures along cross section A–A′ for $n = 1$, $n = 10$, and $n = \infty$. If $n = \infty$ (Fig. 5A), the esti-

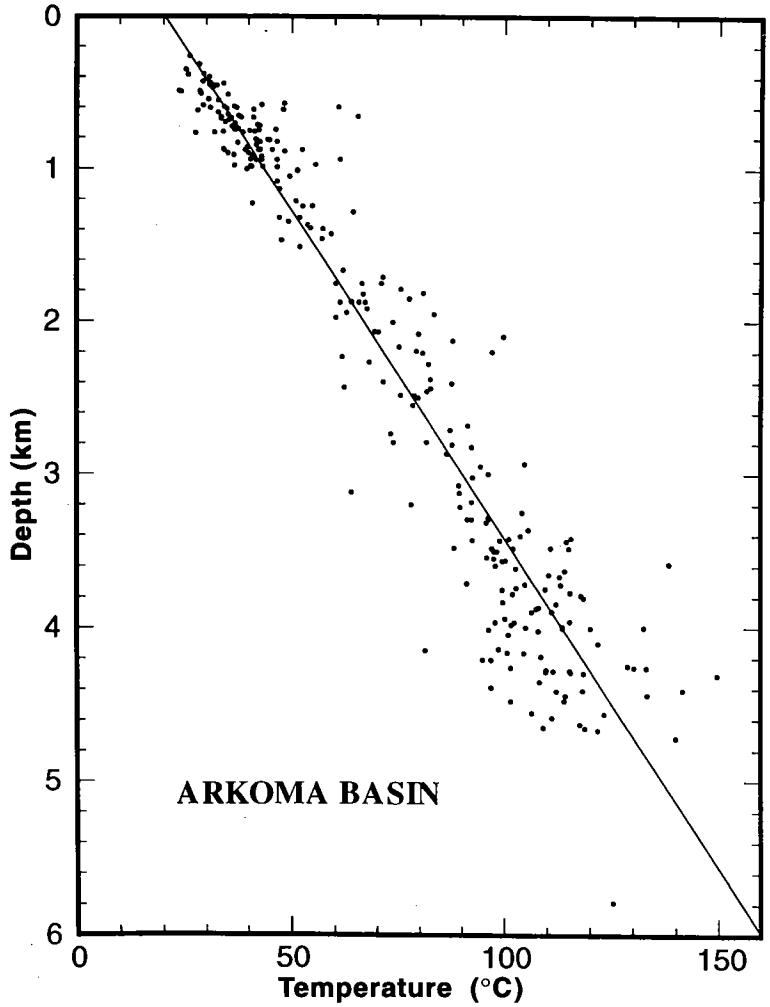


Figure 4. Corrected bottom-hole temperatures from the Arkoma Basin. A least-squares fit yields an average geothermal gradient of 23°C/km.

mated temperatures are the same as those determined by the average-gradient model (Fig. 4), in which case there is no lateral variation across the basin. If $n = 1$ (Fig. 5B), the estimated temperature field shows high-frequency variations in temperature across the basin that are probably unrealistic. If $n = 10$ (Fig. 5C), estimated temperatures and geothermal gradients across the basin show a large lateral variation, but the isotherms are still smooth, indicating that random error has probably been reduced in comparison to the situation shown in Figure 5B.

The estimated temperatures (Fig. 5C) along cross section A–A′ vary from ~20°C near the ground surface to >140°C at the Precambrian boundary in the southern part of the basin. The average geothermal gradient in the southern part of the basin is 20°C/km, exceeds 30°C/km in the middle part of the basin, and is 24°C/km in the northern part of the basin. At a depth of 2 km, the lateral variation in basin temperature is ~20°C, four times the estimated level of random error in the corrected BHTs. Our interpretation is that the estimated temperature field shows a real change in average thermal conditions through the Arkoma Basin and is likely not an artifact of problems related to temperature measurement and correction.

INTERPRETATION

Geothermal gradient is directly proportional to heat flow and inversely proportional to thermal conductivity. The relatively low thermal gradient in the southern part of the basin and the relatively high thermal gradient in the middle part of the basin may simply be due to variations in the thermal conductivity of sedimentary rocks in the Arkoma Basin. We are currently proceeding with thermal conductivity measurements to test this hypothesis and estimate heat flow.

If the results of our thermal conductivity measurements show that heat flow varies along cross section A–A′ in a manner analogous to the geothermal gradient, hypotheses involving heat transport or lateral variations in the concentration of radioactive heat-generating elements must be considered.

Ground-water flow has the potential to be an effective agent for redistributing heat in sedimentary basins. For example, Deming and others (1992) concluded that regional groundwater flow is probably responsible for systematic variations in near-surface heat flow observed in the North Slope Basin, Alaska. In the case of the Arkoma Basin, the topographic slope would be consistent with regional flow from south to north with concomitant heat transport. The magnitude of heat transport by ground-water movement has an exponential dependence upon the depth and velocity of circulation. If the depth of circulation reaches several kilometers, near-surface heat flow can be reduced significantly for relatively low (0.1–1.0 cm/yr) Darcy velocities. High permeability aquifers and conspicuous signs of underground flow (e.g., artesian flow) are not a prerequisite.

Variations in the concentration of heat-producing radioactive isotopes in Precambrian basement rocks (granite and rhyolite) could also be responsible for the observed (Fig. 5C) variation in thermal state. This hypothesis could be tested through measurements on basement rocks, but samples may not be available. The observed variation in thermal state could be due to some deep-seated, unexplained variation in background heat flow. However, the depth at which such a variation arises could probably not be greater than the wavelength of the observed change in the basin (about 60–70 km).

SUMMARY

The average geothermal gradient in the Arkoma Basin estimated from 277 corrected BHTs along a north-south cross section is 23°C/km. However, the geothermal gradient shows considerable horizontal variation. On the southern end of the cross section, the average geothermal gradient is ~20°C/km, exceeds 30°C/km in the middle, and is ~24°C/km on the northern end. The observed

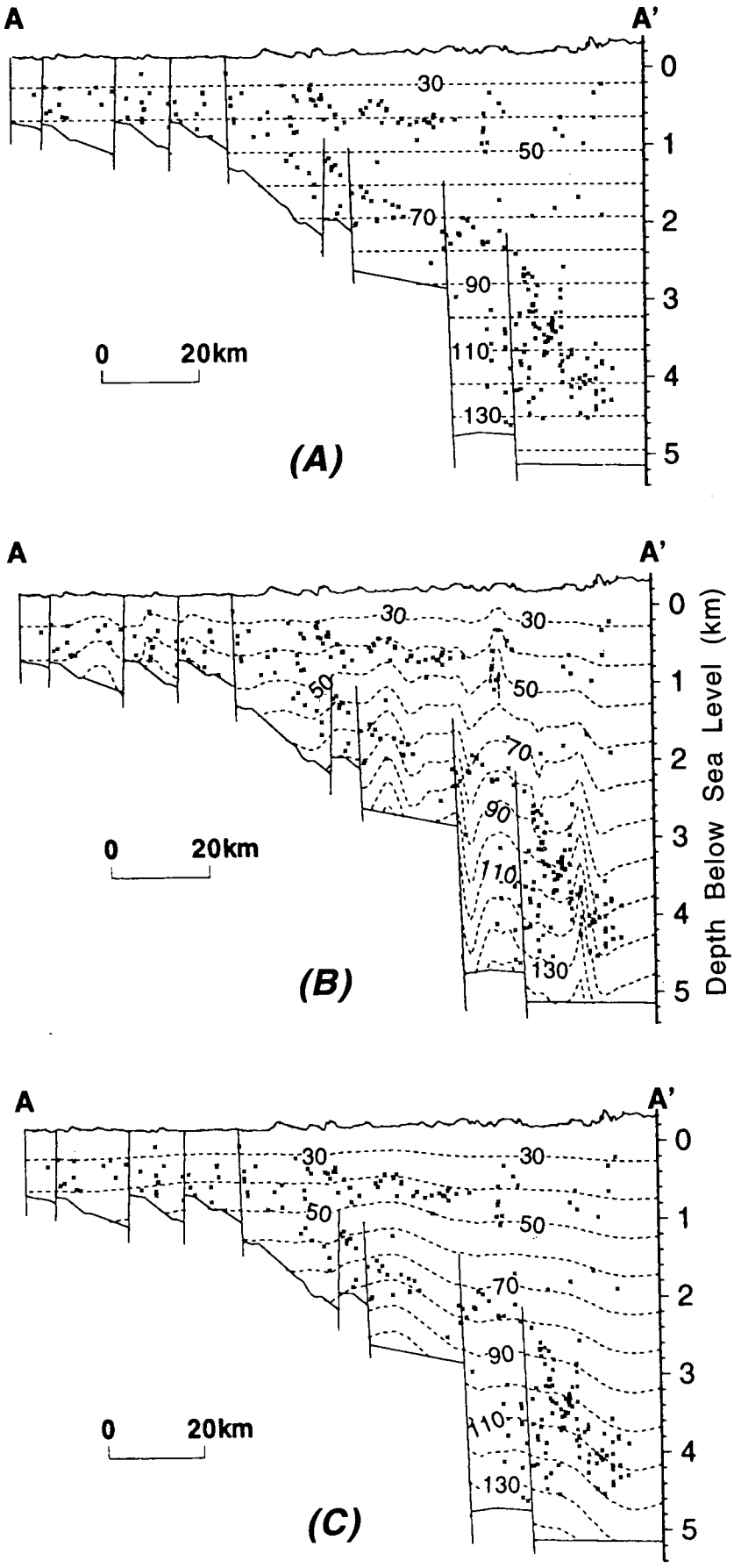


Figure 5 *(left)*. Estimated temperatures (°C) along a north-south cross section through the Arkoma Basin. Three plots illustrate trade-off between resolution and reduction of random error in temperature data. (*A*) All 277 corrected BHTs are given equal weighting; resolution is poor, but random error is relatively low. (*B*) The closest corrected BHT is weighted heavily; resolution is high, but random error is also relatively high. (*C*) The closest 10 corrected BHTs are weighted heavily; resolution is better than for plot *A*, but random error is reduced lower than for plot *B*. Xs show location of BHT measurements projected onto cross section A–A'. Solid lines along the bottom of each cross section show top of Precambrian granite and rhyolite, as shown in Figure 2. Solid lines at top of each cross section show topography.

variation in geothermal gradients could be due to changes in thermal conductivity, changes in radioactive heat generation in basement rocks, or heat transport in the basin arising from ground-water flow. We are proceeding with thermal-conductivity measurements on Arkoma Basin rocks for the purpose of estimating heat flow and evaluating these hypotheses.

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In-Situ Stress Orientation, Western Arkoma Basin and Ouachita Mountains, Oklahoma

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ABSTRACT.—Wellbore-breakout analysis is a common technique for determining in-situ horizontal-stress orientations. Four-arm caliper logs from 24 wells in the Arkoma Basin and Ouachita Mountains of Oklahoma were examined to determine contemporary stress orientations. Twenty wells contained usable data that indicate a N. 20° E. ± 20° breakout orientation. The direction of S_{hmax} and probable direction of induced hydraulic fractures is about N. 70° W. (west-northwest). This direction is oblique to that of the midplate stress province (which includes the Arkoma Basin and Ouachita Mountains) of other workers. The origin of the anomalous stress orientation is unknown, but may be related to the deep gravity minimum centered in the Ouachitas.

INTRODUCTION

Identification

Wellbore breakout is one type of elongation in the cross-sectional shape of a wellbore. It can be identified on four-arm caliper logs that typically are run with dipmeter or formation-microscanner logs. More recently, borehole televiewers (BHTV) have been used to map wellbore breakouts. Ideally, breakout results in a wellbore that is approximately elliptical in cross section; its short axis is in-gauge (equal to the drilled diameter of the hole), but its long axis is enlarged (greater than the drilled diameter) (Fig. 1). In practice, the short axis of a breakout zone typically is slightly larger than the drilled diameter; in rare cases, the short axis is smaller than the drilled diameter. Zones of breakout vary in length; the lower detectable limit is defined by the length of the pad on the four-arm caliper tool, although BHTV can detect smaller zones. The tops and bottoms of breakout zones should be abrupt (Fig. 2); a circular or slightly out-of-gauge hole should abruptly become elliptical. In practice, however, the tops and bottoms of breakout zones vary from abrupt to gradual (Fig. 3).

Wellbore breakout is distinctive from other kinds of hole elongations. Cox (1970) first recognized a uniform direction of borehole elongation over a large area in Alberta, Canada; Cox determined that it was not related to stratal dip and that departures from that direction were, in some cases, associated with deviated wells. Babcock (1978) distinguished breakout from long elongations caused by mechanically generated wear on the sides of deviated wells during drilling. Bell and Gough (1979) showed that breakouts are not related to rock type, age of strata, dip, or hole deviation. Zones of hole enlargement formed by preferential erosion of less competent or more fractured rocks by drilling fluids (one or two axes) are termed "washouts" (Dart and Zoback, 1989). Dart and Zoback (1989) distinguished randomly oriented washouts that form in easily eroded rocks from preferentially oriented washouts (POWs). POWs are probably zones of breakout that form in easily eroded rock types (Fig. 3). Bell and Babcock (1986) suggested that high mud pressures could result in the wellbore's being hydraulically fractured; the fractured rock would be selectively eroded by the drilling fluid, leaving a zone

AUTHOR'S NOTE: I collected the data for this study in 1988 and 1989 and analyzed those data between 1988 and 1991. Since 1989, a large number of new exploratory and development wells have been drilled in the Oklahoma part of the Arkoma Basin and Ouachita Mountains. Four-arm caliper logs were run in many of those wells, making them suitable for wellbore-breakout analysis. I have not attempted to acquire those logs and begin a more thorough study of contemporary in-situ stress in southeastern Oklahoma. Rather, this study is based on 24 wells drilled before 1989. The results are preliminary; however, the consistency of the in-situ stress orientations reported here suggests that additional data are unlikely to change the principle conclusions of this study.

Operators with the necessary data are welcome to submit those data to me for breakout analysis. Four-arm caliper logs are generally run in conjunction with the following logs: dipmeter, high-resolution dipmeter, stratigraphic high-resolution dipmeter, formation microscanner, formation microresistivity scanner, and fracture identification. Additional data for breakout analysis will permit more thorough study of stress orientations in the Arkoma Basin and Ouachita Mountains, as well as identification of possible local perturbations in the regional stress field.

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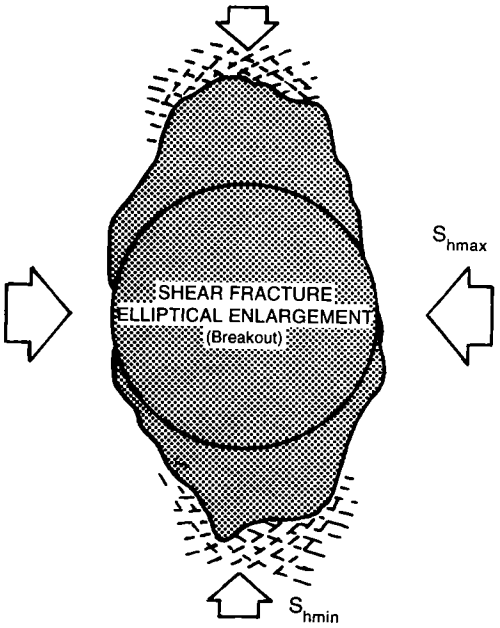


Figure 1. Ideal cross section of wellbore showing relation of wellbore breakout to S_{hmax} and S_{hmin} (from Dart, 1990, p. F8).

of elongation with a long axis perpendicular to that of a breakout (discussed below). Other types of hole elongation may be caused by unequal mud caking on enlarged holes (Bell and Babcock, 1986) and plastic deformation of the borehole (Dart and Zoback, 1989).

The recognition of different types of borehole elongations led several investigators to establish criteria for distinguishing breakout from mechanically generated wear, washouts (erosion) of certain rock types, elongation parallel to induced and natural fractures, and unequal mudcakes. Some of the criteria that have been suggested are (1) caliper tool must rotate (caused by cable torque) above and below the breakout and cease rotation in the zone of breakout; (2) one caliper (small axis) must record the drilled or slightly enlarged hole diameter; if the small-axis caliper is slightly larger than the bit size, it should show less variation than the large-axis caliper; (3) the top and bottom of the zone of breakout must be abrupt—the hole diameter must increase sharply; (4) the borehole must be within 5° of vertical; (5) the length of the breakout zone should be greater than 1 ft; and (6) the direction of elongation should not be parallel to the hole deviation (Plumb and Hickman, 1985; Bell and Babcock, 1986; Bell and others, 1992). Figure 2 illustrates many of these criteria.

In addition to establishing criteria for recognizing breakout in wellbores, Dart (1987), Zoback and Zoback (1989), Zoback and Zoback (1991), and Zoback (1992a) ranked breakout data in terms of quality. Some of these criteria are similar to those used for distinguishing breakout from other types of hole elongations and include (1) degree of tool rotation, (2) abruptness of termination, (3) amount of hole deviation, and (4) amount of washouts. Other criteria are (5) smoothness of wellbore, (6) scatter of data, and (7) frequency, overall length, and consistency of occurrence in an individual well.

In general, the most important criteria for individual breakout recognition and quality were found to be (1) the relation of caliper-size variation, (2) well deviation (for highly deviated wells), and (3) absence of extensive washouts. Tool rotation and abruptness of termination affect the quality, but not the recognition, of individual breakouts. Direction of hole deviation is not important. Number and consistency of breakouts are a general measure of the quality of the data set but are not a measure of individual breakout quality. Although scatter of the data can make "true" breakout orientation difficult to recognize, it is significant and may have geologic implications (e.g., Dart, 1990; Bell and others, 1992).

Origin

Wellbore breakout forms as a result of unequal horizontal compressive stresses around a wellbore; it is directly related to the present-day

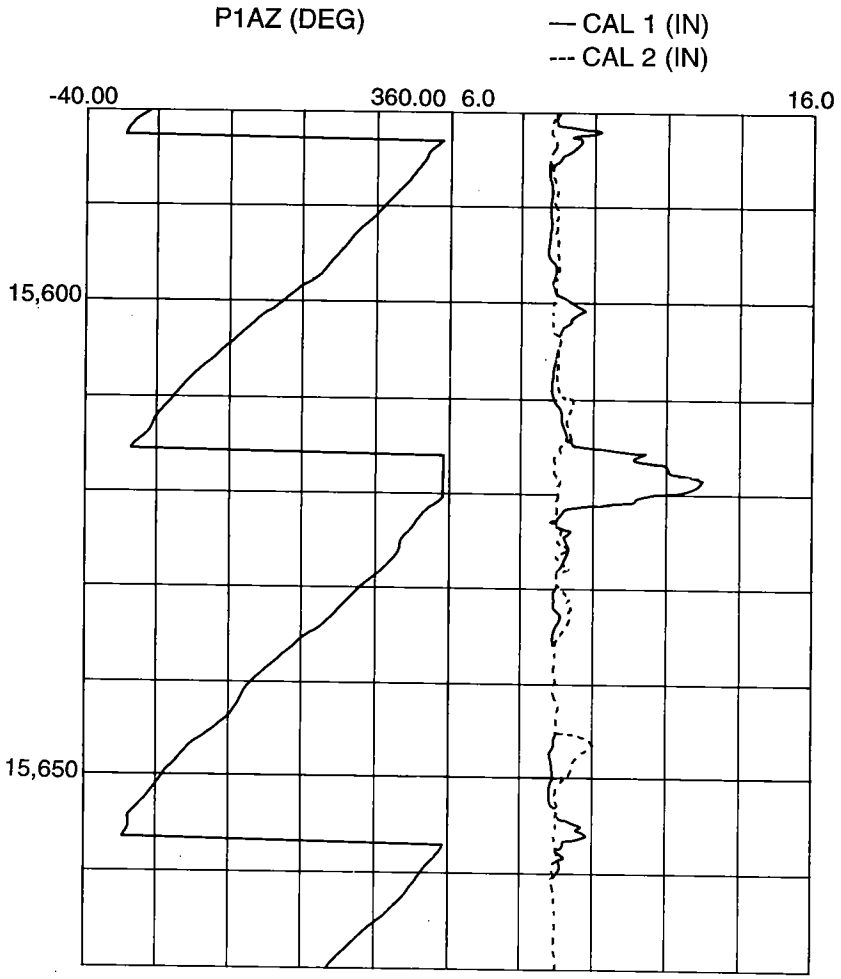


Figure 2. Part of four-arm caliper log from the ARCO No. 1 Hart well illustrating a zone of breakout (15,615–15,621 ft). Note that (1) the top and bottom of the zone of breakout are abrupt, (2) the small axis is in-gauge, and (3) tool rotation (denoted by P1AZ) stopped through the zone of breakout.

state of stress in the Earth's crust. The only serious disagreement with this observation was that of Babcock (1978), who published the first paper using extensive breakout data and suggested that breakout forms as a result of preferential erosion parallel to natural fractures. The natural-fracture origin was quickly challenged (Bell and Gough, 1979; Zoback and Zoback, 1980; Gough and Bell, 1981,1982), and by 1986, Babcock agreed that breakout forms as a result of in-situ stress around a wellbore (Bell and Babcock, 1986). The relation of breakout to in-situ stress was confirmed by Gough and Bell (1981, 1982), who showed that breakout directions are perpendicular to fractures formed by hydraulic-fracturing operations, and by Hickman and others (1985) and Stock and others (1985), who used a BHTV to distinguish natural fractures from zones of breakout. Several other types of data have been used to measure stress orientation and have confirmed that wellbore breakout results from unequal horizontal stresses around a wellbore; these data include (1) overcoring, (2) core-disking, (3) petal-centerline fractures in cores, (4) stress- and strain-gauge grids, (5) tiltmeter studies, (6) fault-plane solutions, (7) slip azimuths of Quaternary faults in surface excavations, and (8) alignment of recent volcanic vents (Zoback and Zoback 1980; Gough and Bell, 1981; Bell and Babcock, 1986; Paillet and Kim, 1987; Plumb and Cox, 1987; Dart and Zoback, 1989).

Much recent research on wellbore breakout has focused on the exact mechanism of borehole elongation and the relation of breakouts to the magnitude of in-situ stresses (e.g., Gough and Bell, 1982; Bell and Gough, 1983; Mastin, 1988; Zheng and others, 1989; Moos and Zoback, 1990; Shamir and Zoback, 1992). For the purposes of this paper, breakouts are considered to be

> ...zones of spalling and fracture on opposite sides of a well bore, which elongate it in cross section from circular to approximately elliptical. Breakouts generally form by compressive failure of the rock at the borehole wall and tend to occur at the azimuth along the well bore where the compressive stress is greatest....Theory predicts that the most compressive stress at the borehole wall exists at the azimuth parallel to the least compressive remote stress that acts perpendicular to the borehole axis. (Mastin, 1988, p. 9187)

All the research that uses breakout to identify the orientation of the relative maximum (S_{hmax}) and minimum (S_{hmin}) horizontal-stress directions assumes that the wellbore is approximately vertical and therefore parallel to the third principal-stress direction (S_v). Comparison of stress directions based on interpretation of breakout data with those determined by other methods (particularly hydraulically induced fractures) suggests that this assumption is valid (Plumb and Cox, 1987; Mastin, 1988).

Additional recent research has (1) treated wellbore-breakout data statistically (Dart, 1987; Plumb and Cox, 1987), (2) investigated breakout data on a regional to worldwide basis and related stress orientations to crustal kinematics (Zoback and Zoback, 1980,1989; Zoback and Zoback, 1991; Zoback, 1992a), (3) attempted to explain anomalies in the predicted stress patterns (Bell and others, 1992; Zoback, 1992a,b), and (4) used in-situ stress orientation for evaluating hazards associated with seismic and aseismic deformation and the potential for slip along known faults (Dart and Zoback, 1989; Dart, 1990). Zoback and Zoback (1991) and Zoback (1992a) have identified what

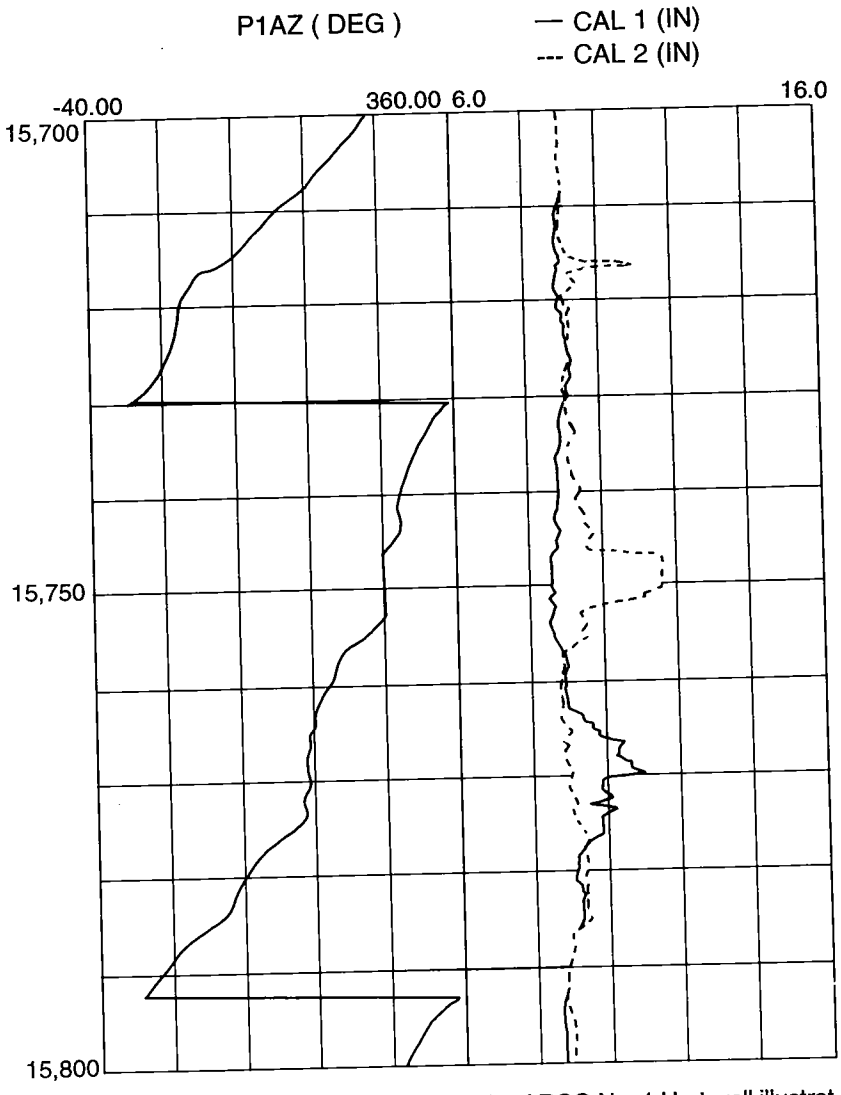


Figure 3. Part of four-arm caliper log from the ARCO No. 1 Hart well illustrating two zones of breakout (15,746–15,752 ft and 15,763–15,774 ft). Note that the lower zone does not show an abrupt top or bottom, but that tool rotation stops. The upper zone shows an abrupt top and bottom, but is enclosed within a zone of probable preferentially oriented washout (POW) in which tool rotation either slows, stops, or rotates normally. The orientation of both zones of breakout is about N. 20° E.

they call the "midplate stress province" of the North American plate. This province extends from near the Rocky Mountain front east to the western Atlantic continental shelf, and from the Gulf Coast in the south to the Arctic Circle, and includes all of Oklahoma. The maximum horizontal-stress direction (S_{hmax}) (compressive) is oriented east-northeast to northeast and is caused by lithosphere-scale "ridge push" from the Mid-Atlantic Ridge.

Hydrocarbon Exploration and Development

Knowledge of contemporary in-situ stress orientation is important for optimal development of oil and gas fields, particularly in areas where hydraulic fracturing is routine. Bell and Babcock (1986), Dart and Zoback (1989), and Dart (1990) have shown that the engineering and design of drilling and hydraulic fracturing operations can be affected by in-situ stress conditions. For example, hydrocarbon reservoirs can be more effectively drained if development wells are more closely spaced perpendicular to the direction of induced fractures (Brown and others, 1980), i.e., parallel to the direction of breakout. If the length of the induced fracture is known, the most appropriate spacing for wells parallel to the direction of the fracture can be determined.

In addition, Bell and Babcock (1986) showed that the ability to predict induced-fracture orientation is important when an exploration well has missed its target; in some cases, it may be possible for an induced fracture to connect with the missed target, e.g., a reef. They also suggested that correct prediction of fracture orientation is an aid to blow-out control because it enables relief wells to be better located.

SOUTH-CENTRAL U.S. STUDIES

Previous Work

The most thorough studies of in-situ stress in Oklahoma are by Dart (1987,1990). The data presented in these reports were used by Zoback and Zoback (1989) to document the tectonic stress field in the conterminous U.S. and by Zoback (1992a) in her World Stress Map project. There is some overlap in the studies by Dart (1987,1990); although not stated, the data presented in Dart (1990) for the eastern parts of the Anadarko and Marietta Basins appear to be a refinement of those presented in Dart (1987). The earlier study, however, covers a much larger area and includes wellbore-breakout data from the Ouachita Mountains and Arkoma Basin.

Breakout data from the eastern part of the Anadarko Basin in Oklahoma yield a consistent east-northeast (N. 78° E.) maximum horizontal principal-stress direction (S_{hmax}) (Dart, 1990). This direction is similar to that of the North American midplate province (Zoback and Zoback, 1991; Zoback, 1992a). The in-situ stress orientation in the eastern part of the Marietta Basin is less well constrained. Wellbore enlargements consist of generally northwest-oriented breakouts and generally northeast-oriented fracture-related enlargements (Dart, 1990). S_{hmax} in the Marietta Basin is N. 41° E. The northeast-oriented fractures consist of induced fractures, natural fractures, and/or joint sets. Dart (1990) suggested that the bimodal data set in the Marietta Basin may be typical of data collected in areas where wells are drilled between subparallel faults that strike oblique to S_{hmax}.

Dart (1987) reported breakout data from one well in the Arkoma Basin and one well in the Ouachita Mountains in Pittsburg County, Oklahoma. The Arkoma Basin well showed a mean breakout direction of about N. 65° E., which suggests a north-northwest direction for S_{hmax}. Dart (1987) assigned a relatively low confidence level to this orientation. The most common breakout direction based on total footage was about N. 45° W., which suggests a northeast direction for S_{hmax}. However, there was significant scatter in the direction of individual breakouts; Dart (1987) considered this data set "no good." Dart (1987) examined 14 wells in the Arkoma Basin in Arkansas. Six wells were shown as having a statistical quality rating of A, B, or C; the directions of S_{hmax} from these wells are (from west to east) N. 33° E. (AR 15 [Dart's well designation]), N. 17° W. (AR 10), N. 65° E. (AR 7), N. 40° W. (AR 12), N. 85° W. (AR 4), and N. 42° E. (AR 9). AR 7 and AR 12 are within a few miles of each other. The report by Dart (1987) contains no reliable breakout data for wells in the Ouachita Mountains, and the data for wells in the Arkoma Basin in Oklahoma and Arkansas show a great deal of scatter.

Present Study

Results

Throughout this section, the term "breakout" refers to features on the four-arm caliper logs that are *interpreted* as wellbore breakouts. In some cases, other types of hole elongations were recorded as breakouts; however, analysis of the data (below) shows that these kinds of elongations can be recognized when a sufficient number of wells have been studied.

The four-arm caliper logs from 24 wells in Oklahoma were examined for this study. The well locations are shown in Figure 4. Ten wells are located in the Ouachita Mountains (surface location south of the trace of the Choctaw fault). Thirteen wells are located in the Arkoma Basin; six in T. 5 N., R. 18 E. One well is located in the Gulf Coastal Plain just south of the western end of the Ouachita Mountains.

Table 1 summarizes the breakout data for the Oklahoma Ouachita Mountains and Arkoma Basin. "Depth logged" refers to that part of the well for which four-arm caliper data were available. "Quality of breakout" is a qualitative ranking of the quality of the breakout data used in constructing the rose diagrams (Fig. 5). The ranking (A is excellent, D is poor) is based on several criteria:

1) *Tool rotation.*—The quality of the breakout data is degraded if the caliper tool was not rotating or did not stop rotating in the zone of breakout.

2) *Shape of breakout zone.*—The quality is degraded if hole elongation was not abrupt or was very slight or if the short axis was significantly enlarged relative to bit size. POWs are included with the breakout data because they grade into each other and reflect unequal horizontal stresses around the wellbore.

3) *Log quality.*—The quality of the breakout data is degraded if log curves overlap and/or the caliper curves are difficult to read. The columns labeled "number of

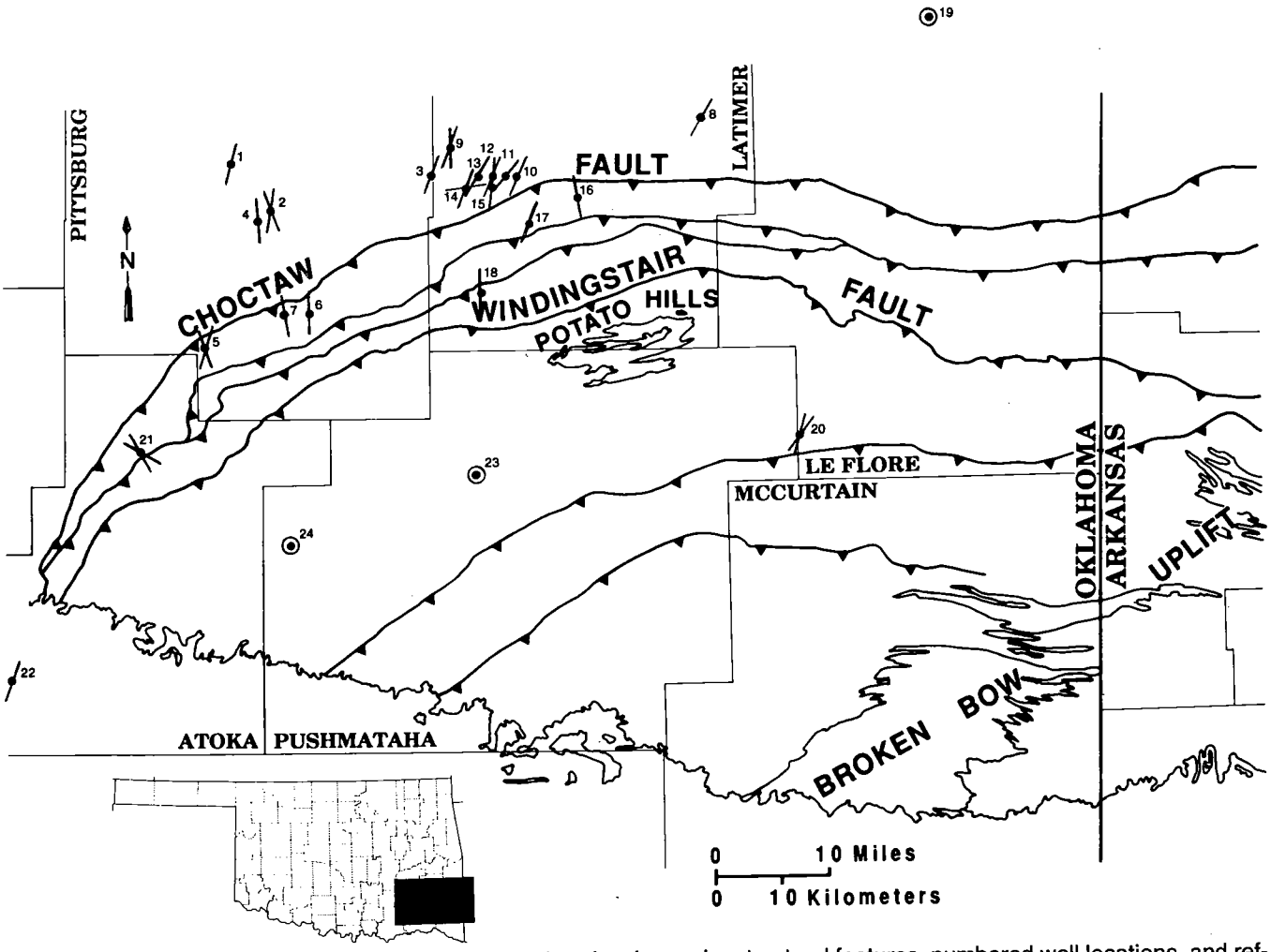


Figure 4. Map of Arkoma Basin and Ouachita Mountains showing major structural features, numbered well locations, and reference line(s) that approximate breakout orientations. Circled well locations did not show a preferred breakout orientation.

breakouts/feet logged" and "feet of breakouts/feet logged" reflect how much breakout is present in a well (over the interval logged) and is another qualitative indicator of data quality. Large numbers in these columns indicate that breakout zones are common; small numbers indicate that breakout zones are rare.

Figure 5 is a series of rose diagrams showing the direction of breakout for the 24 wells used in this study. The upper hemisphere is based on the number of feet of breakout over the logged interval; the lower hemisphere is based on the number of breakouts. The heavy lines plotted on the outside of the rose diagrams (Fig. 5) also are plotted in Figure 4 through the well locations and are used merely as a reference line; they may represent the approximate breakout direction(s) but have no statistical significance. The data for three wells (nos. 19, 23, and 24) were too scattered to determine an approximate breakout direction.

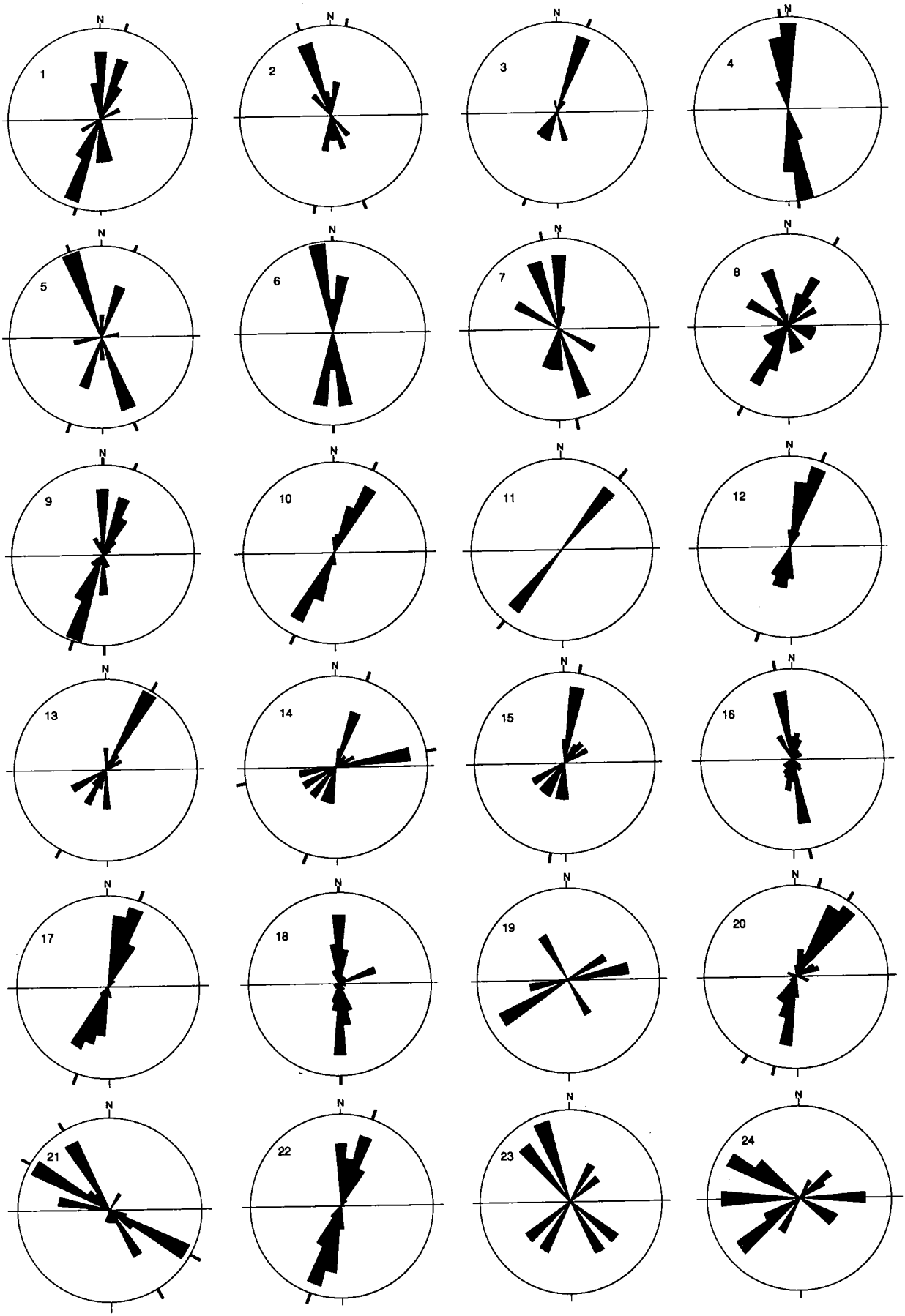
Figure 6 is a histogram of the interpreted approximate breakout directions shown on Figure 5 (rounded to the nearest 10°). Some wells showed more than one breakout direction; both are plotted on Figure 6. The most common breakout direction is N. 20° E., and nearly 70% of the wells had a breakout direction of N. 20° E. ± 20°. The N. 20° E. direction becomes more significant if several apparently anomalous "breakout" directions are considered. The N. 60° W. and N. 30° W. directions occur in well no. 21. This well is highly deviated (maximum 23°) (Table 1), and the borehole was extremely irregular, suggesting that the data may be in error. The two N. 20° W. directions were from wells (nos. 2 and 5) that also showed N. 10° E. and N. 20° E. breakout directions, suggesting that the N. 20° W. directions may represent preexisting fractures or joints. The N. 80° E. direction is also from a well (no. 14) that showed a N. 20° E. breakout direction and may represent preexisting fractures. The N. 20° W. and N. 80° E. directions are perpendicular and parallel, respectively, to the general structural grain of the Ouachita tectonic belt in this area; fractures with these orientations are not unlikely. If the anomalous data from wells nos. 21, 2, 5, and 14 are eliminated, all the breakout directions for the 21 wells for which an approximate direction could be determined are from N. 10° W. to N. 40° E. and 85% are N. 20° E. ± 20°.

TABLE 1. — WELLBORE-BREAKOUT DATA, ARKOMA BASIN AND OUACHITA MOUNTAINS, OKLAHOMA

Map reference	Well (operator, no., farm)	County	Sec./T./R.	Date logged mo/yr	Depth logged (ft)	Maximum hole deviation (°)	Number of zones showing breakout	Total feet of breakout	Quality of breakout	Number of breakouts/ feet logged ×1,000	Feet of breakouts/ feet logged ×100
1	Grace 1 Mitchell	Pittsburg	8/5N/14E	7/80	7,400–9,850	2.5	11	79	C	5	3
2	Beren 1-30 Turner	Pittsburg	30/5N/15E	10/81	1,785–3,615	2	7	97	C	4	5
3	Samson 1 Bobo	Pittsburg	16/5N/17E	7/78	8,200–9,351	5	4	73	C	4	6
4	Beren 1-1 Alexander	Pittsburg	1/4N/14E	4/81	2,483–3,183	3.5	6	72	B	9	10
5	Kerr-McGee 1-31 Lee Scott	Pittsburg	31/3N/14E	6/79	54–2,642	2.5	7	65	C	3	3
6	Getty 1-14 Sweetin	Pittsburg	14/3N/15E	7/82[a]	12,380–13,860	7.5	5	31	B	3	2
7	Texaco 1-16 Isbell	Pittsburg	16/3N/5E	2/85	5,966–13,852	13	6	186	B	1	2
8	Exxon 1 Circle F Ranch	Latimer	22/6N/21E	1/86	9,000–12,875	4	14	118	C	4	3
9	ARCO 1 Sharp	Latimer	2/5N/17E	11/88	13,800–15,509	3	18	226	B	11	13
10	ARCO 3 Costilow	Latimer	14/5N/18E	5/88	13,054–14,140	7	12	73	A	11	7
11	ARCO 2 Yourman	Latimer	15/5N/18E	6/87	10,820–11,246	2	1	16	B	2	4
12	ARCO 2-16 Kilpatrick	Latimer	16/5N/18E	6/88	12,787–13,961	3	6	22	B	5	2
13	ARCO 2 Fazekas	Latimer	17/5N/18E	12/87	12,958–14,990	4	8	57	B	4	3
14	ARCO 2 Bennett State	Latimer	19/5N/18E	9/88	11,500–13,476	6	5	41	C	3	2
15	ARCO 2 Paschall	Latimer	21/5N/18E	10/88	13,154–14,256	11	5	24	B	5	2
16	Shell 32-27 Williams	Latimer	27/5N/19E	12/80[b]	1,190–14,294	7	49	679	B	4	5
17	ARCO 1 Hart	Latimer	6/4N/19E	1/89	15,400–17,764	12	41	759	A	25	46
18	Exxon 1 Retherford	Latimer	5/3N/18E	1/84	1,943–7,352	6	25	167	C	5	3
19	Getty 1 Calhoun	Le Flore	2/7N/24E	3/80	4,202–9,649	4.5	4	211	C	1	4
20	Sohio 1-7 Weyerhaeuser	Le Flore	7/1N/23E	5/87	623–8,015	4.5	23	621	C	3	8
21	ARCO 1-19 Ingersoll	Atoka	19/1N/13E	10/87	12,600–15,060	23	32	971	D	13	39
22	Exxon 1-"A" Mullin	Atoka	32/3S/11E	1/78	9,070–13,188	5.5	48	364	C	12	9
23	Sohio 1-29 Trotter Dees	Pushmataha	29/1N/18E	4/84[b]	650–6,798	8	4	26	C	1	0
24	Jones & Pellow 27-1 Smith	Pushmataha	27/1S/15E	9/81	6,225–10,467	7.5	8	56	C	2	1

[a]Spud date. Log date not given on log. [b]Date of last log.

Figure 5 *(opposite page)*. Rose diagrams showing breakout orientations for wells shown in Figure 4. Upper hemisphere shows total feet of breakout; lower hemisphere shows number of breakouts. Heavy lines just outside diagram are reference lines that visually approximate average breakout orientations, but have no statistical significance. These lines are the same as those plotted on Figure 4.


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Interpretation

Based on 20 wells for which good and consistent borehole-elongation data were available, the approximate average breakout direction in the Arkoma Basin and Ouachita Mountains in Oklahoma is N. 20° E. This finding suggests that S_{hmax} (and the orientation of induced hydraulic fractures) is N. 70° W. (west-northwest). This direction is significantly different from the east-northeast direction found by Zoback and Zoback (1991) and Zoback (1992a) for the "midplate stress province" of the North American plate.

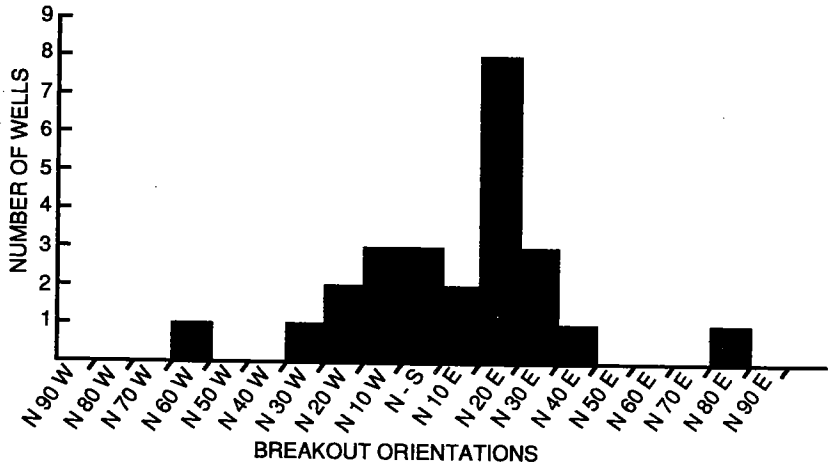


Figure 6. Histogram of approximate breakout orientations.

Anomalies in regional in-situ stress orientations are not uncommon. Bell and others (1992) discussed three alternatives to explain the anomalous stress regime in the southwestern part of the Aquitaine Basin, France: (1) boundary zone between a northern European stress province and a Pyrenean stress province, (2) geomechanical discontinuities in the structurally complex Aquitaine Basin, and (3) weak horizontal-stress anisotropy. None of these explanations satisfactorily explains the anomalous Arkoma Basin–Ouachita Mountain stress direction. Unlike the Pyrenees, the Ouachita Mountains are not actively overriding the Arkoma Basin. Breakout direction in the structurally complex Ouachita Mountains is the same as that in the structurally simple Arkoma Basin. The widespread and relatively consistent west-northwest direction suggests that horizontal stresses are unequal in magnitude but uniform in direction.

Zoback (1992a) discussed several "second-order stress sources" that affect regional stress patterns. Those that are associated with flexural stresses include "sediment loading, particularly along continental margins; glacial rebound; seamount loading; and the upwarping of oceanic lithosphere oceanward of the trench" (p. 11,720). These processes occur in areas of active tectonics and/or sedimentation and are not relevant to the Arkoma Basin–Ouachita Mountains. Zoback (1992a) also suggested that buoyancy forces related to lateral differences in crustal density can affect regional stress patterns; however, all the examples cited are from areas of active tectonism. Lateral crustal-strength contrasts are restricted to transform plate boundaries, such as along the San Andreas fault.

The origin of the west-northwest S_{hmax} direction is unknown. One possibility is that it is related to the –110-mgal east-northeast–trending gravity low centered in the Ouachita Mountains just southeast of the Potato Hills (Kruger and Keller, 1986). This minimum is probably caused by a structurally(?) thickened Carboniferous sedimentary section in conjunction with an upper- and/or lower-crustal downwarp (Kruger and Keller, 1986, p. 683). A possible objection to the gravity minimum as a cause for the anomalous S_{hmax} direction is that well number 20 is located near the center of the minimum, whereas all other wells are on the steep north flank of the minimum. Future work, particularly in areas south of and outside the gravity minimum, may clarify whether it is the cause of the anomalous stress orientation in the Arkoma Basin–Ouachita Mountains area.

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Paleozoic Rocks of the Western Frontal Ouachita Mountains and Arkoma Basin, Arkansas: Summary

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GEOLOGIC SETTING

Paleozoic rocks in the western frontal Ouachita Mountains are intricately folded, complexly thrust-faulted, and mildly to severely sheared. The exposed rocks are mostly Devonian through Middle Pennsylvanian and include the Arkansas Novaculite, Stanley Shale, Jackfork Sandstone, Johns Valley Shale, and Atoka Formation. Most of these rocks are considered to be of deep-marine origin. Their thickness ranges from ~50,000 ft in the south to <30,000 ft along the northern part of the frontal Ouachita Mountains. The effects of the late Paleozoic compressional forces that formed the Ouachita Mountain fold belt also lessen in intensity toward the north. Rock types in decreasing order of abundance are shale, sandstone, siltstone, chert-novaculite, conglomerate, tuff, and limestone.

In the southwestern Arkoma Basin there are exposures of thick, interbedded sequences of deep-marine sandstone and shale of the lower Atoka Formation (Middle Pennsylvanian). These lower Atokan rocks grade and thin toward the north across postulated large, "buried," generally downthrown-to-the-south growth faults into shallow-marine and deltaic facies. Thin to thick intervals of shallow-marine and deltaic sandstone, shale, and some thin coal beds occur in the overlying middle and upper Atoka, Hartshorne, McAlester, and Savanna Formations—all of Middle Pennsylvanian age. In places the stratigraphic succession is disrupted by large, mostly northward-translated thrust plates. The strata are usually inclined rather steeply along the crests of anticlines and gently in the broader synclines.

GAS AND COAL PRODUCTION

Since the discovery of the shallow Mansfield gas field in March 1902, many hundreds of shallow to rather deep wells have been drilled, and considerable dry-gas reserves have been proved in the western Arkoma Basin of Arkansas. Most gas production has been from shallow-marine and deltaic sandstone sequences of the Atoka (Middle Pennsylvanian) and Hale Formations (Early Pennsylvanian). Some gas production, however, has been from proximal deep-marine sandstone of the lower Atoka Formation and older platform limestone and chert sequences. Future natural-gas exploration should be focused primarily on deeper drilling targets. Drilling activities will be strongly aided by seismic evaluations to pinpoint growth faults, stratigraphic traps, unconformities, depositional trends, and other features.

Low-volatile bituminous to semianthracite coal has been mined in the western Arkoma Basin of Arkansas since 1840, although large-scale production did not begin until about 1880. The most productive coal seam—the lower Hartshorne coal—underlies a larger area than any other coal in Arkansas. It averages about 3 ft thick, but locally exceeds 8 ft. Whereas in 1958 there were some 44 active coal mines in or near this area, at present there is but one underground mine near Hartford, Arkansas, and a few small surface ventures. The coal reserves and recoverable reserves are estimated at 2,080 and 1,040 million short tons, respectively.

RESULTS OF RECENT GEOLOGIC MAPPING

Many of the maps and illustrations shown in the poster session accompanying this meeting were generated by an intensive 7½-year field and aerial-photographic mapping project undertaken by Boyd R. Haley and me in central and west-central Arkansas. This program was part of a still larger cooperative venture (COGEOMAP) with the Oklahoma and the U.S. Geological Surveys that was designed to generate geologic maps and reports providing information on the complex structure, stratigraphy, and mineral resources in the Ouachita Mountain orogen. In addition, important biostratigraphic evaluations of the lower Paleozoic strata of the Ouachitas include studies by Ray Ethington and Jim Stitt (University of Missouri–Columbia), John Repetski (U.S. Geological Survey), and others.

In Arkansas, 178 geologic maps have been completed at a scale of 1:24,000 and then reduced and compiled for publication at a scale of 1:100,000. A few of the many geologic features that are newly illustrated in the western frontal Ouachita Mountains and Arkoma Basin on the COGEOMAP maps include (1) the duplex, triangle, or

Stone, C. G., 1994, Paleozoic rocks of the western frontal Ouachita Mountains and Arkoma Basin, Arkansas: summary, *in* Suneson, N. H.; and Hemish, L. A. (eds.), Geology and resources of the eastern Ouachita Mountains frontal belt and southeastern Arkoma Basin, Oklahoma: Oklahoma Geological Survey Guidebook 29, p. 293–294.

"teepee" nature of the structure in Atokan rocks along Dutch Creek in Scott and Yell Counties; (2) revisions of the Choctaw "frontal" thrust-fault zone west of Waldron, Scott County; (3) décollements in the Stanley Shale west of Potter, Polk County; (4) complex backthrusting at a number of sites—east of Acorn, Polk County, and elsewhere; and (5) curious lineaments trending northwest and northeast noted on aerial photographs at several localities in western Polk County.

In addition, the poster session includes several theoretical cross sections depicting the depositional systems and structural origins of the Ouachita Mountains and Arkoma Basin beginning in early and extending through late Paleozoic time. The thin, preflysch, lower Paleozoic strata of the Ouachita trough record the development of the broader early basin from a continental rift into a narrower but deep ocean basin. Extensive late Paleozoic compressional forces along the southern margins of the basin aided in the northward deepening and migration of the Ouachita trough while >45,000 ft of Carboniferous deep-marine turbidites were deposited. Eventually, in very late Paleozoic time, the Ouachita Mountains were further thrust-faulted and uplifted into a complexly deformed mountain range.