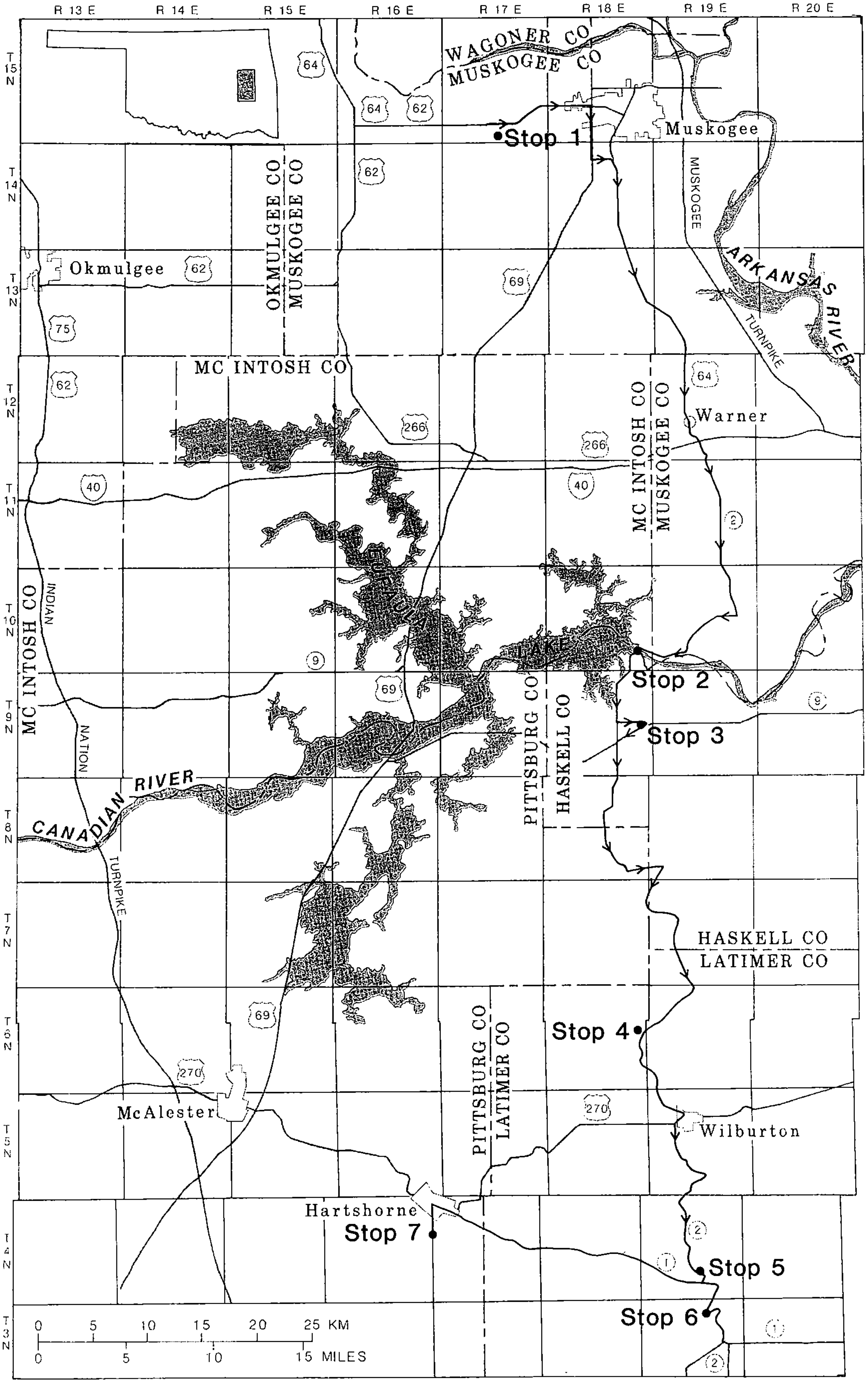
Shelf-to-Basin Geology and Resources of Pennsylvanian Strata in the Arkoma Basin and Frontal Ouachita Mountains of Oklahoma

Oklahoma Geological Survey
Guidebook 25

1988



FIELD-TRIP ROUTE

Oklahoma Geological Survey
Charles J. Mankin, *Director*

ISSN 0078-4400

Guidebook 25

SHELF-TO-BASIN GEOLOGY AND RESOURCES OF PENNSYLVANIAN STRATA IN THE ARKOMA BASIN AND FRONTAL OUACHITA MOUNTAINS OF OKLAHOMA

Kenneth S. Johnson, *Editor*
Oklahoma Geological Survey
University of Oklahoma
Norman, Oklahoma

Contributors:

Charles A. Ferguson, Oklahoma Geological Survey
LeRoy A. Hemish, Oklahoma Geological Survey
Kenneth S. Johnson, Oklahoma Geological Survey
Neil H. Suneson, Oklahoma Geological Survey
Glenn S. Visher, Geological Services and Ventures, Tulsa
W. David Wylie, Consulting Geologist, Oklahoma City

Guidebook for field trip held October 1, 1988, following the 25th annual national meeting of the American Institute of Professional Geologists, September 28–30, 1988, Tulsa, Oklahoma.

The University of Oklahoma
Norman, Oklahoma
1988

Front Cover

Snow-clad slopes of the Potato Hills, looking south from the crest of Buffalo Mountain in the northern Ouachita Mountains of Oklahoma. The hills are composed of Ordovician through Lower Mississippian bedded cherts and black shales. These rocks are deformed into tight to isoclinal, upright folds, the resistant cherts forming ridges and the shales underlying the valleys. On a larger scale, the Potato Hills form the core of an anticlinorium flanked to the north and south by a pair of wide valleys and prominent ridges. The valleys are underlain by Mississippian volcaniclastic flysch (Stanley Shale), and the ridges, one of which provides the vantage point for this photograph, by sandstone-rich Pennsylvanian flysch (Jackfork Group).

Back Cover

Fossilized compression of a Middle Pennsylvanian lycopod tree, *Lepidodendron obovatum*, exposed on the surface of the Bluejacket Sandstone Member of the Boggy Formation in the SW¼NE¼SE¼NW¼ sec. 13, T. 6 N., R. 18 E., Robbers Cave State Park (Stop 4).

This publication, printed by the Transcript Press, Norman, Oklahoma, is issued by the Oklahoma Geological Survey as authorized by Title 70, Oklahoma Statutes, 1981, Section 3310, and Title 74, Oklahoma Statutes, 1981, Sections 231–238. 500 copies have been prepared for distribution at a cost of $5,078.55 to the taxpayers of the State of Oklahoma. Copies have been deposited with the Publications Clearinghouse of the Oklahoma Department of Libraries.

PREFACE

This guidebook presents new and recently updated information on the geology and mineral resources of Pennsylvanian strata in parts of eastern Oklahoma. The data are presented along a north-south transect that starts in the Muskogee area on the north, and extends southward through the Arkoma basin and into the frontal belt of the Ouachita Mountains south of Wilburton and Hartshorne. Emphasis of the trip will be on the shelf-to-basin changes that occur within selected Pennsylvanian units, on energy resources in the region, and on the sedimentation, stratigraphy, and tectonics of the Ouachita frontal belt.

Most of the data presented here by Oklahoma Geological Survey staff members (Ferguson, Hemish, Johnson, and Suneson) result from a cooperative geologic mapping (COGEOMAP) program that the OGS has been conducting with the U.S. Geological Survey and the Arkansas Geological Commission. This COGEOMAP program, which began in 1985, has been highly successful in focusing interest and personnel on detailed geologic mapping in the Ouachita Mountains of Oklahoma and Arkansas. The OGS effort in this program involves geologic mapping of 7.5′ quadrangles along the frontal belt of the Ouachitas eastward from Hartshorne to the Arkansas state line, and mapping just north of the Choctaw fault in the Arkoma basin, where structural features are related to Ouachita Mountains tectonics. The data presented by Visher and Wylie represent an overview of the geologic setting that partly controls hydrocarbon occurrences in the Arkoma basin—a major gas province.

Stratigraphic nomenclature used by contributors to this guidebook shows some variation, reflecting different personal perspectives and some unresolved stratigraphic problems; no attempt has been made to require that nomenclature be entirely consistent from article to article.

Thanks are extended to Larry N. Stout, geologist/editor for the OGS, and his staff, for their efforts in editing the material for this guidebook and for guiding it through to final publication. Thanks also are extended to T. Wayne Furr, OGS manager of cartography, and his staff, for preparation of illustrations for the guidebook.

KENNETH S. JOHNSON
Field Trip Chairman
and Editor

CONTENTS

GENERAL GEOLOGIC FRAMEWORK OF THE FIELD-TRIP AREA

Kenneth S. Johnson

INTRODUCTION

The Arkoma basin and Ouachita Mountains are elongate tectonic provinces that extend about 250 mi across parts of eastern Oklahoma and western and central Arkansas (Fig. 1). The Arkoma basin is bounded on the north by the Ozark uplift, and it grades northwestward onto a northern shelf area (also called the Cherokee platform). To the south the Arkoma basin is bounded mainly by the frontal belt of the Ouachita Mountains, and on the southwest it abuts the Arbuckle Mountains. Both the Arkoma and Ouachita provinces are characterized by great thicknesses of sedimentary rocks in eastern Oklahoma: about 5,000–20,000 ft of strata in the Arkoma basin, and about 20,000–40,000 ft of strata in the Ouachita Mountains.

These two provinces are an intriguing pair of tectonic features that were most active during the Early and Middle Pennsylvanian. In pre-Pennsylvanian times the Arkoma basin area was part of a broad epicontinental shelf that received a moderately thick sequence of Cambrian through Mississippian shallow-water carbonates and clastics; this area was part of the early and middle Paleozoic Oklahoma basin that covered much of the southern Midcontinent north of the Ouachita geosyncline (Johnson and others, in press). Pre-Pennsylvanian shelf sediments of the Oklahoma basin grade southward into a deep-water facies consisting mainly of black shales and cherts that accumulated in the Ouachita trough. Similar patterns of shelf and deep-water sedimentation also persisted through Morrowan and early Atokan time. During middle Atokan

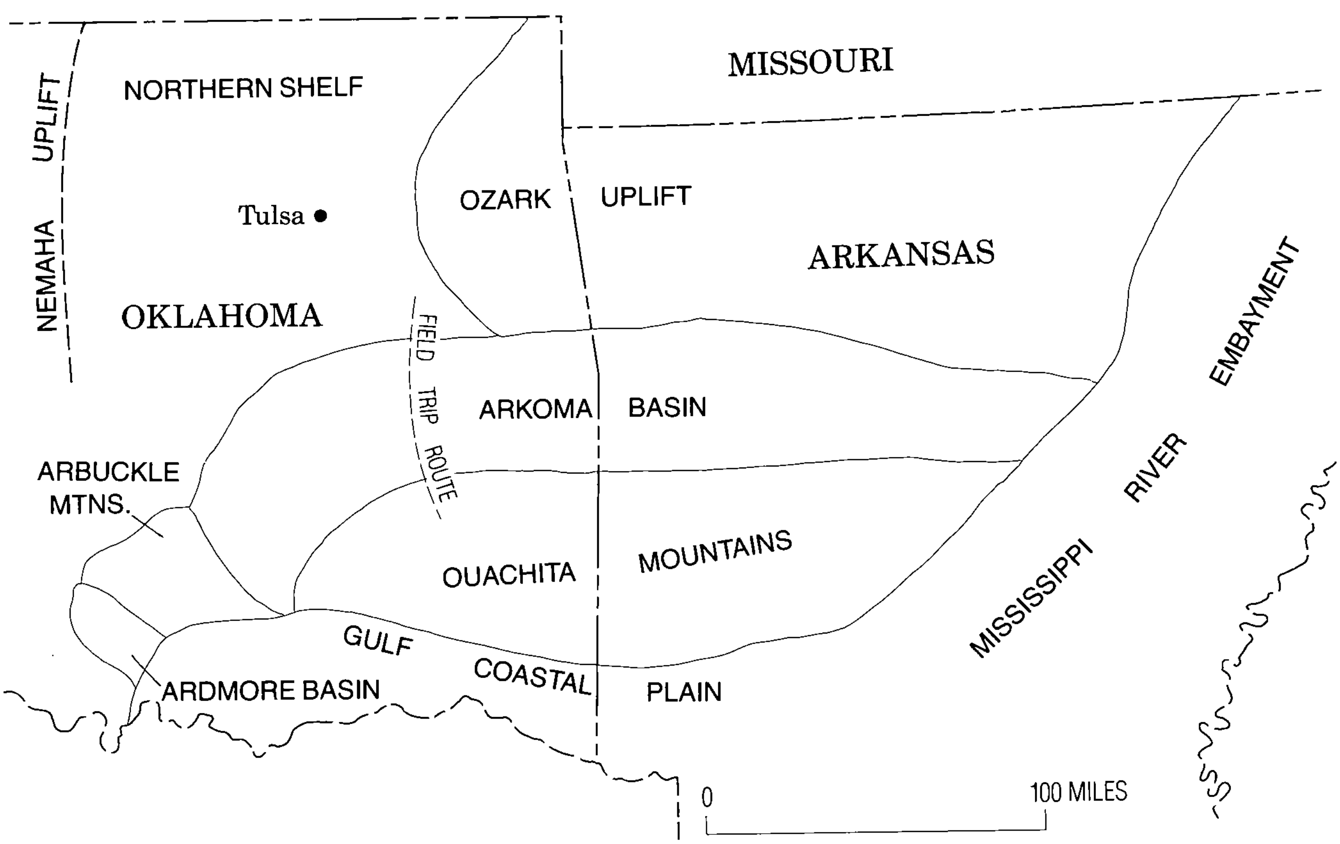


Figure 1. Major geologic provinces of eastern Oklahoma and Arkansas, and approximate field-trip route. Modified from Johnson and others (in press).

through Desmoinesian time, the Arkoma basin developed as a major foreland basin, while the Ouachita trough initially continued to sink and then, in late Atokan time, was subjected to orogenic activity that included folding, thrusting, and marked uplift of the Ouachita facies. With destruction of the Ouachita trough, the depocenter was shifted northward into the Arkoma basin, and sedimentation continued until the entire eastern Oklahoma region was apparently raised above sea level in late Desmoinesian time (Haley, 1982).

Among the more comprehensive studies done on the field-trip area are those of Branson (1961), Flawn and others (1961), Swanson and others (1961), Frezon (1962), Branan (1966), Amsden (1980), Haley (1982), Sutherland and Manger (1984a), Johnson and others (in press), and Sutherland (in press).

The field trip begins on the north, near Muskogee (~50 mi southeast of Tulsa), and extends southward on a transect that crosses the Arkoma basin and enters the frontal zone of the Ouachita Mountains (Fig. 1). In taking this 100-mi-long trip, we have an opportunity to examine the sedimentology, stratigraphy, structure, geologic history, and mineral resources of Early and Middle Pennsylvanian strata in two of the major tectonic provinces along the south margin of the North American craton.

PRE-PENNSYLVANIAN STRATA

In early and middle Paleozoic time, eastern Oklahoma was astride two major tectonic provinces. On the north was the Oklahoma basin, and on the south was the Ouachita trough, a great geosyncline that bordered the southern margin of the craton. The Oklahoma basin in eastern Oklahoma received a sequence of moderately thick and extensive marine limestones and dolomites that are interbedded with thinner marine shales and sandstones. They were laid down upon a modestly rugged, eroded surface of Precambrian granite and rhyolite ~1,285 m.y. old. A time-transgressive basal sandstone, the Cambrian Reagan Sandstone, was deposited across all areas except local topographic highs. This was followed by a series of shallow-water carbonates and thin clastic units (Fig. 2). Total thickness of pre-Pennsylvanian strata in the field-trip area ranges from ~2,000 ft on the northern shelf to ~6,000 ft on the south (Fig. 3).

Deposition in the region apparently was fairly continuous during most of pre-Pennsylvanian time, with the exception of two major epeirogenic uplifts that occurred in Early Devonian (pre-Frisco–Sallisaw) and Late Devonian (pre-Woodford–Chattanooga) times (Amsden, 1980). Little if any folding or faulting accompanied these broad uplifts.

Following the second of these epeirogenic uplifts of the region, the Ouachita trough subsided more rapidly than before and received a much thicker sequence of sandstone and shale (Stanley Shale) than equivalent Mississippian strata in the Oklahoma basin farther north. The marked subsidence of the Ouachita trough during the Mississippian was accompanied by minor southward thickening of equivalent strata in the developing Arkoma basin.

PENNSYLVANIAN STRATA

Morrowan strata in the Arkoma basin represent a continuation of the shelf-like sediments that underlie them, although they contain a significant amount of sand. These strata range from ~300 ft thick on the north to ~1,000 ft thick on the southern margin of the basin, and they grade into 3,000–6,000 ft of deeper-water marine (flysch) sediments (Jackfork Group and Johns Valley Shale) in the Ouachita trough (Fig. 4). Growth faulting continued through Morrowan time, with a major fault in the north part of the trough having shed blocks of pre-Morrowan rock that now are exotics in the Johns Valley Shale (Haley, 1982).

Atokan Series strata consist entirely of the Atoka Formation in the region. The Atoka Formation is about 70% shale in the central and southern parts of the Arkoma basin; it also contains lenses and tongues of sandstone and siltstone, along with several thin coal beds (Cardott and others, 1986). Atokan deposition in the Arkoma basin was characterized by a series of meandering fluvial systems and deltas that had their source to the north and northwest (Sutherland, in press). Thickness of the Atoka ranges from several hundred feet in the north part of the field-trip region to as much as 10,000 ft at the south margin of the Arkoma basin (Figs. 3,4). The thickness of the Atoka increases markedly on the downthrown (south) side of each of a series of syndepositional faults that were most active during deposition of the middle Atoka. In the frontal belt of the Ouachitas, south of the Choctaw fault, the Atoka is a flysch sequence of sandstones and shales as much as 12,000 ft thick (top is eroded).

Early stages of uplift in the Ouachita trough began in late Atokan time. This uplift, called the Ouachita orogeny, resulted from collision of the South American plate into the North American craton and resultant destruction of the Ouachita trough. A series of complex thrust faults and folds brought the flysch sequence (Stanley, Jackfork, Johns Valley, and Atoka) northward out of the trough into what is now the frontal belt of the Ouachita Mountains (Fig. 3).

Subsidence of the Arkoma basin continued intermittently through Desmoinesian time. Although the Ouachita Mountains of Oklahoma were being uplifted, almost all sediments deposited in the Arkoma basin at this time were derived from northern sources. Cyclic sedimentation in the basin resulted from a series of marine transgressions from the west, each of which was followed by fluvial–deltaic sedimentation and development of widespread coal swamps. Individual formations within the Desmoinesian thicken markedly toward the south in the Arkoma basin. Shales make up most of the Desmoinesian strata in the south half of the basin,

	SERIES	ARKOMA BASIN			OUACHITA MOUNTAINS
PENNSYLVANIAN	Desmoinesian	Krebs Gp.	Boggy Fm.	₱bg	
			Savanna Fm.	₱sv	
			McAlester Fm.	₱ma	
			Hartshorne Fm. Upper / Lower	₱hs	
	Atokan	Atoka Fm.		₱a	Atoka Formation
	Morrowan	Wapanucka Ls.		₱m	Johns Valley Shale
		Union Valley Ls. Cromwell ss.			Jackfork Group
MISSISSIPPIAN	Chesterian	"Caney" Sh.		MD	Stanley Shale
	Meramecian				
	Osagean				Arkansas Novaculite
	Kinderhookian				
DEVONIAN	Upper	Woodford Sh.			
	Lower	Hunton Gp.	Frisco Ls. Bois d'Arc Ls. Haragan Ls.	DSOhs	Pinetop Chert
SILURIAN	Upper		Henryhouse Fm.		Missouri Mountain Shale
	Lower		Chimneyhill Subgroup		Blaylock Sandstone
ORDOVICIAN	Upper	Sylvan Sh.			Polk Creek Shale
		Viola Gp.	Welling Fm. Viola Springs Fm.	Ovs	Bigfork Chert
	Middle	Simpson Gp.	Bromide Fm. Tulip Creek Fm. McLish Fm. Oil Creek Fm. Joins Fm.		Womble Shale
					Blakely Sandstone
	Lower	Arbuckle Gp.	West Spring Creek Fm. Kindblade Fm. Cool Creek Fm. McKenzie Hill Fm. Butterly Dol.	O€a	Mazarn Shale
					Crystal Mountain Ss.
					Collier Shale
CAMBRIAN	Upper		Signal Mountain Ls. Royer Dol. Fort Sill Ls.		? ?
		Timbered Hills Gp.	Honey Creek Ls. Reagan Ss.		
PRECAMBRIAN		Granite and rhyolite		p€	

Figure 2. Stratigraphic chart for the Arkoma basin in the field-trip area (modified from Cardott and others, 1986) and for the Ouachita Mountains (modified from Mankin, 1987).

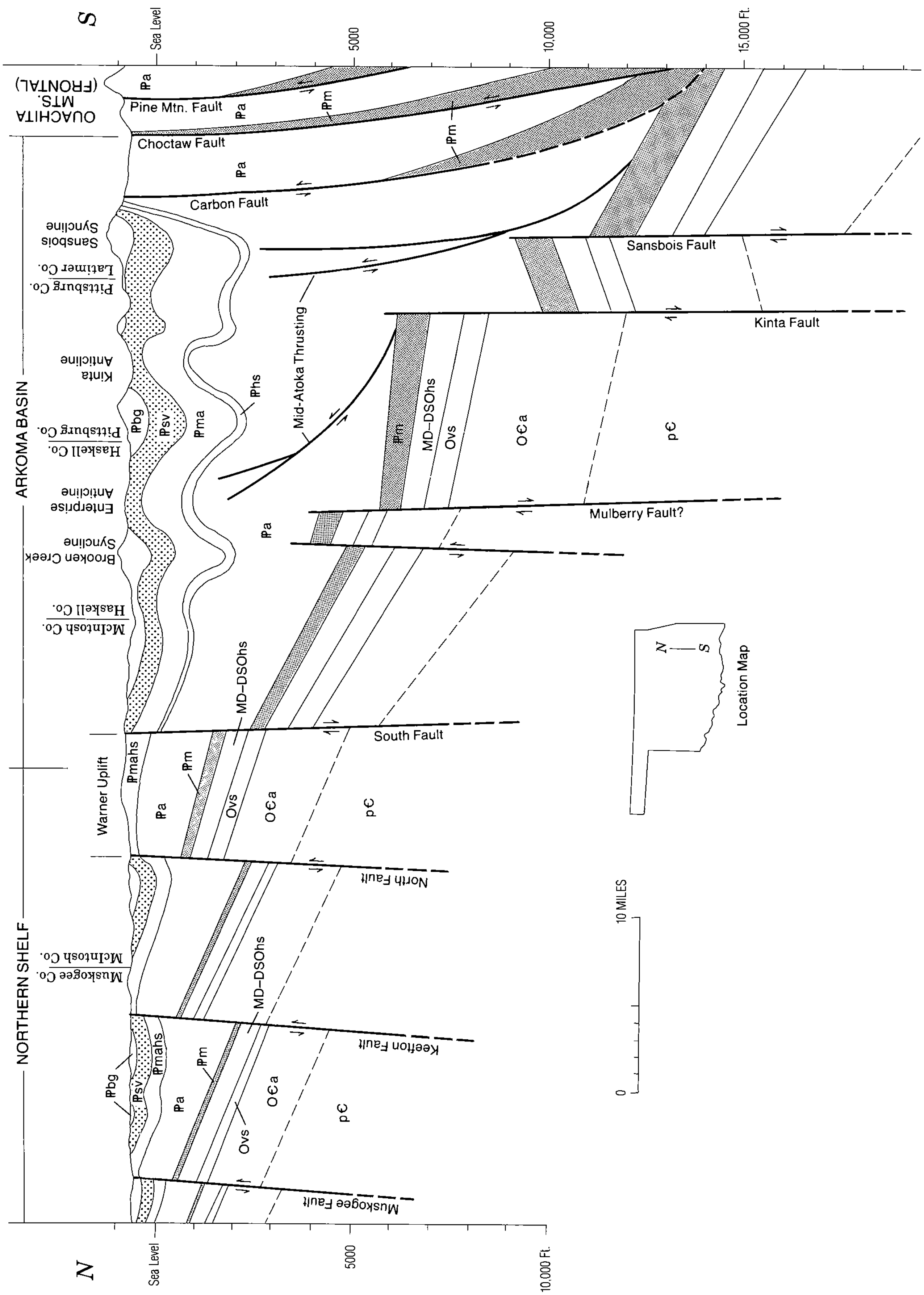


Figure 3. Generalized north–south structural cross section approximately along the field-trip route through the Arkoma basin and frontal zone of the Ouachita Mountains. From Cardott and others (1986). See Figure 2 for explanation of symbols.

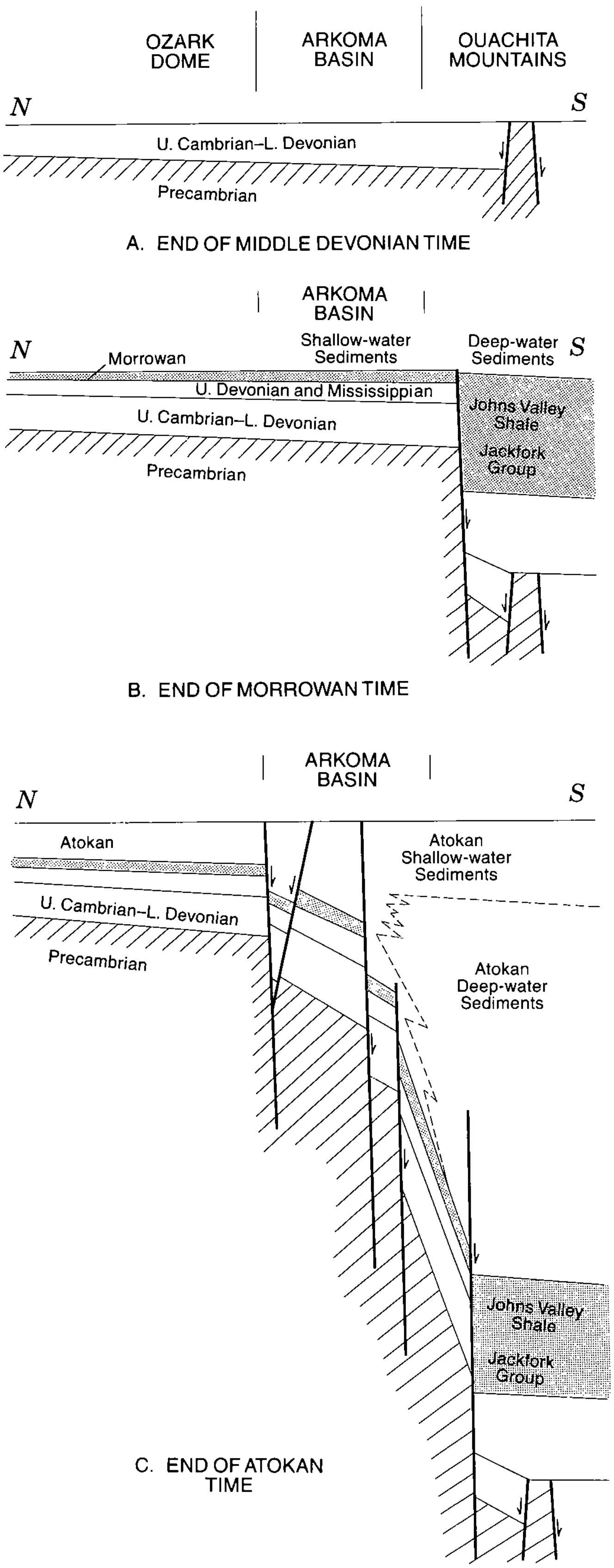


Figure 4. Generalized north–south cross sections showing sedimentation in the Arkoma basin–Ouachita Mountains area from Late Cambrian through Atokan time. Modified from Haley (1982) and Cardott and others (1986).

although several deltaic complexes, such as the Bartlesville–Bluejacket Sandstone Member at the base of the Boggy Formation, constitute major sandstone units.

The early Desmoinesian Krebs Group (Fig. 2) is the only part of the Desmoinesian Series deposited during major subsidence of the Arkoma basin before initial folding of strata in the basin (Sutherland, in press). The next overlying unit, the Thurman Sandstone, contains chert-pebble conglomerates documenting the first significant uplift and erosion of Ordovician and Devonian cherts from the central parts of the Ouachita Mountains (the Thurman and younger strata have been eroded from the field-trip area and now are exposed only farther west).

The youngest formation of the Krebs Group exposed along the field-trip route is the Boggy Formation. A commercially important coal bed, the Secor coal, is present just above the top of the Bluejacket–Bartlesville Sandstone. The Secor coal originated in a deltaic environment in peat swamps that developed during regressive episodes in the cyclic history of Desmoinesian sedimentation (Hemish, 1988b). The first stop on the field trip is at a strip mine where the Secor coal and associated strata of the Boggy Formation are well exposed, and the next three stops focus on depositional features of the Bluejacket–Bartlesville Sandstone. The remaining stops examine structural and depositional features of flysch units in the Ouachita Mountains frontal belt.

COAL GEOLOGY OF THE LOWER BOGGY FORMATION IN THE SHELF-TO-BASIN TRANSITION AREA, EASTERN OKLAHOMA

LeRoy A. Hemish

INTRODUCTION

The purpose of this paper is to provide insight into the stratigraphic relationship of the coal beds in the lower part of the Boggy Formation in the transition area from the northeastern Oklahoma shelf to the Arkoma basin. Identification and correlation of several coal beds in this area, where little was previously known, has been made possible through field investigations and core-drilling conducted by the writer for the Oklahoma Geological Survey (OGS).

COAL STRATIGRAPHY

Coal beds of economic importance occur in area of ~8,000 mi^2 in eastern Oklahoma. These coal resources are bituminous in rank and are in beds of Middle and Late Pennsylvanian age, 270–300 m.y. old (Friedman, 1974, p. 5–6).

Within this framework, coal beds of commercial importance occur discontinuously in the lower part of the Boggy Formation throughout the Arkoma basin and in the southeastern part of the northeastern Oklahoma shelf area. Figure 1 shows the eastern Oklahoma commercial coal belt, and delineates its two general regions: (1) the northeastern Oklahoma shelf area, and (2) the Arkoma basin.

There are marked differences in the coal-bearing strata between these two general areas. Figure 2 is a generalized geologic column showing the coal-bearing strata of the northeastern Oklahoma shelf area, and Figure 3 is a generalized geologic column showing the coal-bearing strata of the Arkoma basin (Hemish, 1988a). The main differences between the two areas are:

1) Coal-bearing rocks present above the Senora Formation in the shelf area are absent in the Arkoma basin;

2) Stratigraphic units are generally much thicker in the Arkoma basin;

3) Commercial coal beds in the northern shelf area pinch out to the south and are absent in the basin; conversely, certain well-developed commercial coals in the Arkoma basin, such as the Hartshorne coal, pinch out to the north, or have no commercial value in the shelf area, owing to thinness;

4) Quality of the same coal in the two regions often varies because of different depositional environments. For example, the Secor coal, the most important commercial coal in the study area, is a high-quality coal with a low ash content and a sulfur content averaging <1% in southwestern Wagoner County and northwestern Muskogee County (Hemish, 1986a, p. 178). Friedman (1978, p. 24) believed that the peat which formed the Secor coal in this area was deposited not far from an alluvial environment, possibly in an upper-deltaic-plain setting. By contrast, the Secor coal in the Arkoma basin has a higher ash content and a much higher sulfur content, averaging 4.4% in an area interpreted as pro-delta, and 5.7% in an area interpreted as a large, lower deltaic plain (Friedman, 1978, p. 24). It seems reasonable to assume that the peat that formed the Secor coal in the Arkoma basin was closer to the sea, and was thus influenced by marine waters, which would account for its higher sulfur content. Marine and brackish-water conditions in peat swamps are known to have a strong influence on the sulfur content of coals (White and Thiessen, 1913, p. 58; Stach and others, 1975, p. 29–31; Reidenouer and others, 1979, p. 88; Williams and Keith, 1979, p. 109; Ward, 1984, p. 175).

General Stratigraphy of the Lower Boggy Formation

For purposes of this report, the lower Boggy Formation is defined as the interval from the base of the Bluejacket Sandstone Member upward to an arbitrary cutoff in the shale unit above the Inola Limestone Member (Fig. 4). Figure 5 is an oil-well log showing electric-log characteristics of the lower Boggy Formation and associated strata.

Three coal beds—an unnamed coal, the Secor coal, and the Secor rider coal—have been identified with confidence in the lower Boggy Formation in that part of the Arkoma basin south of central McIntosh County. Two of these beds, the Secor and Secor rider, persist into the shelf area. Two other stratigraphically higher coal beds are present between the Secor rider and the Inola Limestone. They are the Peters Chapel coal and the Bluejacket coal. These two coals are known to be present in the study area only from southwestern Wagoner County southward to approximately central McIntosh County (Fig. 6).

The stratigraphic relationship of the coal beds present in the lower part of the Boggy Formation in

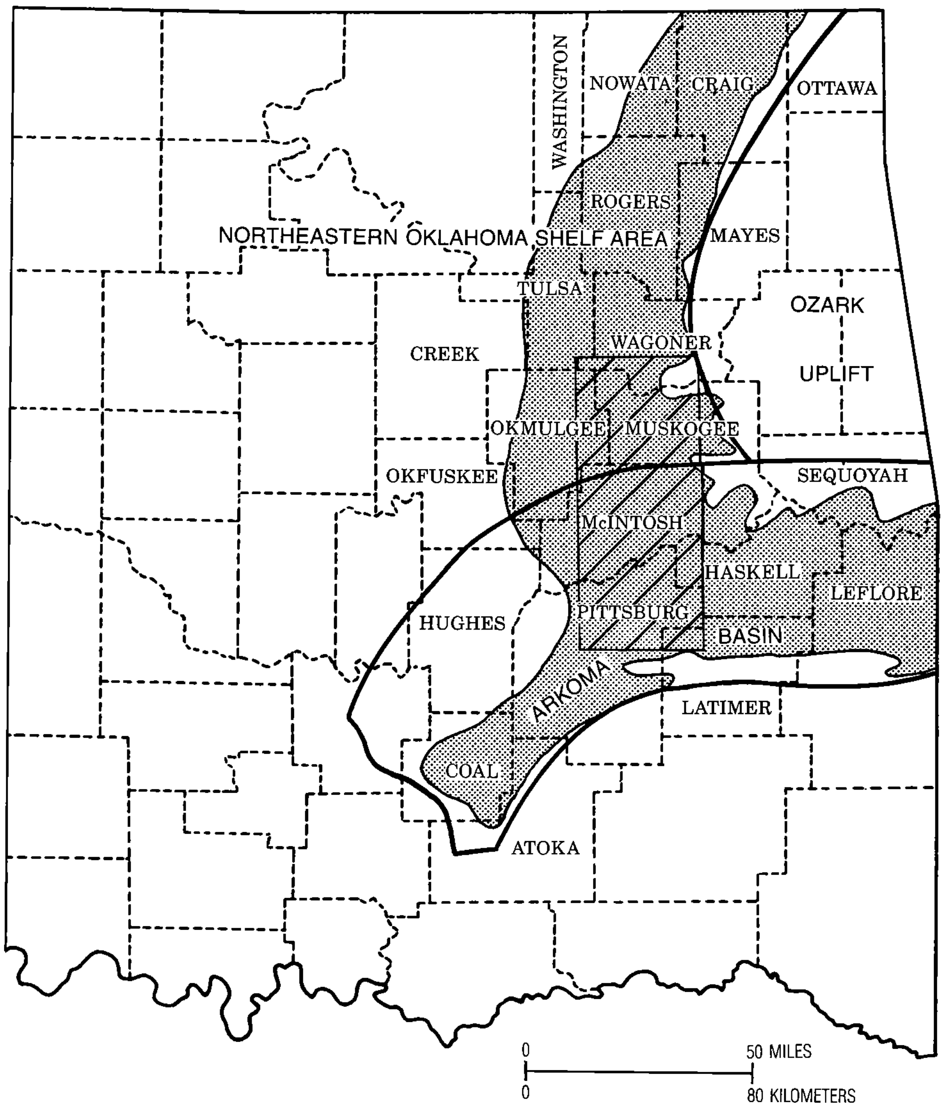


Figure 1. Map of eastern Oklahoma, showing the commercial coal belt (shaded), the Arkoma basin, and the northeastern Oklahoma shelf. Study area shown by diagonal lines. Modified in part from Trumbull (1960) and Friedman (1974).

northwestern Muskogee and southwestern Wagoner Counties was well documented by Hemish (1986a, p. 168–187). However, structural complexities hindered identification and correlation of these coal beds to the south in the shelf-to-basin transition area. Mapping complications also resulted from facies changes related to different depositional environments.

Although none of the coal beds can be traced continuously across the area, sufficient stratigraphic data have been gathered to make correlation possible in the five-county area of this report. More than 1,000 data points in the form of measured sections by the writer, OGS core logs, and exploration drill logs from coal companies were used in the investigation (Hemish, in preparation). Forty core-holes were drilled by OGS in the study area to examine rocks of the lower Boggy Formation and to collect cores for chemical analysis.

The Inola Limestone proved to be one of the most persistent and reliable marker beds in the study area north of the Canadian River. It is generally a 0.5- to 2-ft-thick, medium-dark gray, very fine-grained, impure, silty, highly fossiliferous, brown-weathering limestone containing abundant crinoids and brachiopods.

It is invariably associated with an overlying black, carbonaceous shale containing ironstone concretions, and a persistent, 1- to 2-in.-thick coal bed that occurs immediately under, or within a few inches below, the limestone. This coal bed is the Bluejacket coal. It is known to be present from McIntosh County northward to the Kansas–Oklahoma line (Hemish, 1987, p. 104). It is too thin to have commercial value, except in western Mayes County and locally in Wagoner County.

The Inola Limestone has not been positively identified in the area south of the Canadian River, but limestone associated with black shale and with a thin underlying coal is present in Haskell County (Fig. 6, measured section 11). These units are herein tentatively correlated with the Inola Limestone and Bluejacket coal. A 1-ft-thick, fossiliferous limestone in a similar stratigraphic position, associated with an underlying 10-in.-thick coal bed, has been observed by the writer in the southwest slope of Scaffold Mountain northwest of Blocker in Pittsburg County. These units also may prove to be correlatable with the Inola Limestone and Bluejacket coal.

A second, stratigraphically lower limestone also proved to be a useful marker bed. This unnamed limestone persists from southwestern Wagoner County southward into the Arkoma basin area (Fig. 6). It occurs just above the Secor rider coal. The bed varies from 0.3 to 2 ft in thickness, and is generally dark gray, impure, shaly, thin-bedded, and highly fossiliferous. Brachiopods are the predominant fossils.

Stratigraphic units thicken considerably from north to south, but sequences of beds are still recognizable. For example, the interval from the base of the Secor coal to the black shale above the Inola Limestone is $<$40 ft in Wagoner County, but the same interval some 50 mi to the south, in McIntosh County, is $\sim$100 ft (Fig. 6).

Coal Beds of the Lower Boggy Formation

Five coal beds are known in the lower part of the Boggy Formation in the shelf-to-basin transition area (Fig. 4). Within this area, all but the Bluejacket coal are locally of minable thickness, and have been mined in the past. Only the Secor coal is currently being mined.

Of the five coals, the stratigraphically lowest bed occurs about 30–50 ft or more below the Secor coal bed in a shale or siltstone interval in the upper part of the Bluejacket Sandstone. In this report, the coal bed will remain unnamed, because further investigations are needed in the Arkoma basin to determine if the coal is correlatable with an already-named bed (Lower Witteville), or if it should be assigned a new name. Maximum known thickness (measured by the writer in an outcrop adjacent to an abandoned mine in Pittsburg County, southwest of Blocker) is 15 in. In McIntosh County the thickness of the unnamed bed varies from 12 in. near Onapa to less than 1 in. just south of the Muskogee County line north of Checotah; in that area, this unnamed coal apparently pinches out, and it is not known to have been deposited to the north in the shelf area.

The Secor coal occurs almost immediately above the Bluejacket Sandstone in the study area. It is generally overlain by a shale or siltstone unit that yields an abundance of well-preserved fossil plant specimens. Locally, the upper part of the Bluejacket Sandstone also yields abundant fossil plant material. An unnamed sandstone commonly occurs above the Secor coal in the interval between the Secor and Secor rider coal beds.

Thickness of the Secor coal ranges from 0.1 to 4.3 ft in the Arkoma basin. In general, it is $\leq$1 ft thick in the shelf area, but it is as much as 1.5 ft thick in the vicinity of Porter, Wagoner County. The Secor coal has not been identified north of T. 16 N. in southwestern Wagoner County; possibly it was never deposited there, but more likely it has been removed by erosion. A complex system of NE-trending faults extends through southwestern Wagoner County in the area where the Secor coal is absent (Hemish, in preparation a).

The Secor coal was first mined and named in Pittsburg County (Oklahoma Geological Survey, 1954, p. 129) in the late 1800s. It has been strip-mined in that area—as well as in McIntosh, Muskogee, and Wagoner Counties—continuously during the past 15 years. Three strip mines are currently operating in the Secor coal: (1) the Inter-Chem #2 Mine in McIntosh County, where the coal is 1.0 ft thick; (2) the Pollyanna No. 5 Mine in Muskogee County, where the coal is 1.0 ft thick; and (3) the Hobbs-O'Berg Mine in Wagoner County, where the coal is 1.1 ft thick. Chemical analyses of coal samples collected at the three mines by the writer show that the coal is of high quality (Table 1). It is hvA bituminous in rank, with an average ash content of 2.7%, an average sulfur content of 0.8%, and an average heat value of 14,426 Btu/lb.

The next stratigraphically higher coal bed is the Secor rider coal. It occurs a few inches to 40 ft above the Secor coal and is generally associated with an overlying thin marine limestone and black shale. The Secor rider coal has been mined in Pittsburg County in a multiple-seam operation as recently as 1985. A thickness of 12 in. was measured by the writer at the mine. The maximum known thickness of the Secor rider coal is 18 in., from the log of a coal-exploration hole drilled $\sim$1 mi southwest of Blocker, northern Pittsburg County. The coal bed thins northward and pinches out at the Muskogee–Wagoner County line (Hemish, 1986a, fig. 2). However, the overlying thin limestone and black shale units persist well into southwestern Wagoner County, where they are useful as markers for identifying other coal beds in the lower Boggy Formation (Fig. 6).

The next stratigraphically higher coal occurs in the interval between the unnamed limestone above the Secor rider coal and the Bluejacket coal; it was named the Peters Chapel coal by Hemish (1986a, p. 171), following detailed investigations in northwestern Muskogee County. The Peters Chapel coal bed had been a source of stratigraphic confusion, and had been previously mapped and variously referred to verbally and in published and unpublished documents as the "Bluejacket coal," "Upper Secor coal," "Secor rider coal," and "Secor coal" (Hemish, 1986a, p. 171).

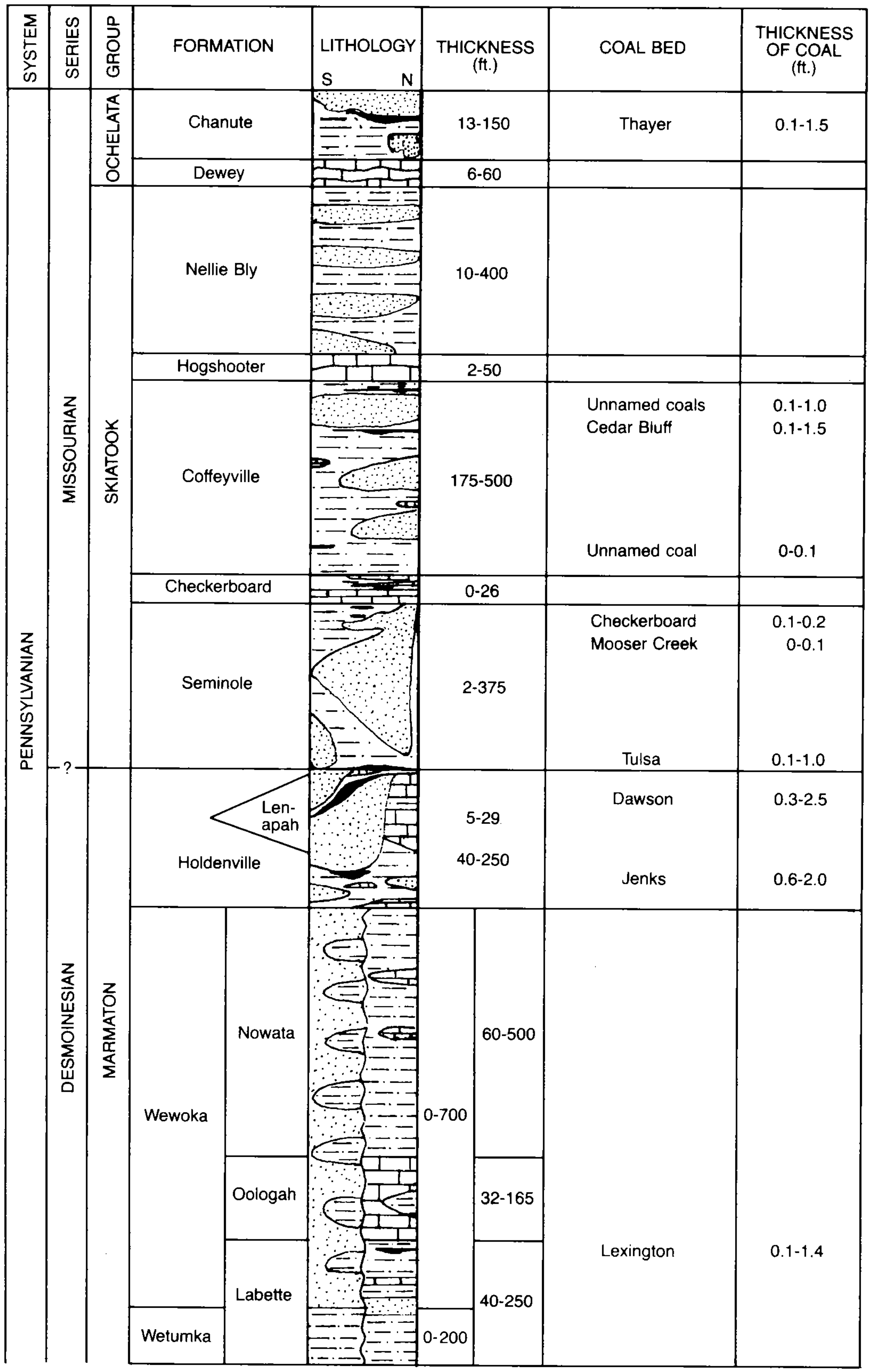


Figure 2. Generalized geologic column of coal-bearing strata, northeastern Oklahoma shelf area. From Hemish (1988a, fig. 6).

PENNSYLVANIAN	DESMOINESIAN	MARMATON	Calvin	Fort Scott	1-90		
					0-400	Mulky	0.5-0.8
						Iron Post	0.3-1.6
						Bevier	0.3-1.0
		CABANISS	Senora		160-500	Unnamed coal	0.1-0.2
						Croweburg	0.2-3.4
						Fleming	0.1-1.5
						Mineral (Morris)	0.1-2.7
						Scammon (?)	0.1-0.5
						Tebo	0.1-0.8
						RC	0.1-0.5
						Weir-Pittsburg	0-6.2
		KREBS	Boggy		35-700	Wainwright (Taft)	0.3-2.3
						Bluejacket	0.1-1.5
						Peters Chapel	0.1-2.0
						Secor rider	0-0.1
						Secor	0.1-1-8
			Savanna		150-200	Drywood	0.1-3.0
						Rowe	0.2-2.5
						Unnamed coal	0.1-0.3
						Unnamed coal	0.1-0.2
						Unnamed coal	0.1-0.6
						Sam Creek	0.1-0.2
						Tullahassee	0.1-0.9
			McAlester		100-400	Spaniard	0.1-1.1
						Keota	0.1-1.0
						Tamaha	0.1-0.3
						McAlester (Stigler)	0.1-1.1
						Keefton (Warner)	0.1-1.0
						Riverton	0.1-0.3
			Hartshorne		0-50	Hartshorne	0.1-0.4
			Atoka		0-975	Unnamed coal	0.1-0.6

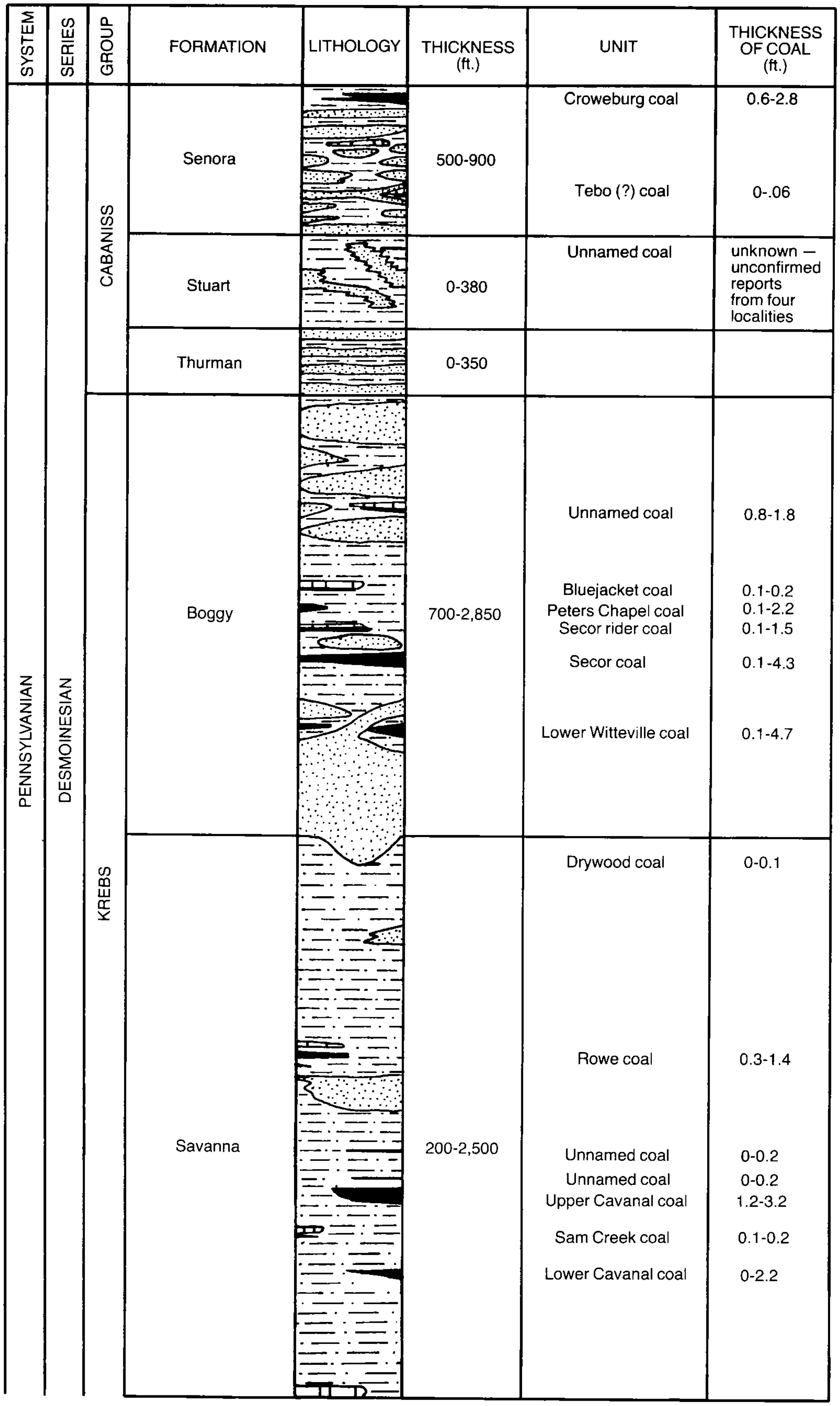
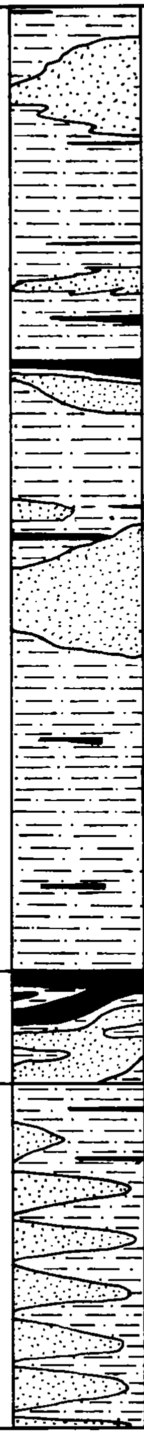


Figure 3. Generalized geologic column of coal-bearing strata, Arkoma basin. From Hemish (1988a, fig. 5).



PENNSYLVANIAN	DESMOINESIAN	KREBS	McAlester	400-2,830	Spaniard coal	0-0.1
					Keota coal	0.1-0.4
					Tamaha coal	0.1-0.3
					Upper McAlester (Stigler rider) coal	0.2-1.7
					McAlester (Stigler) coal	1.0-5.0
					Unnamed coal	0.1-0.2
					Keefton coal	0.1-1.6
					Unnamed coal	0.3-1.0
					Unnamed coal	0.2-0.8
			Hartshorne	50-316	Upper Hartshorne coal	0.2-4.5
					Lower Hartshorne coal	0.7-7.0
			Atoka	0-15,000	Unnamed coal	0-0.5
					Unnamed coal	0-0.5

This report documents the occurrence of the Peters Chapel coal as far south as central McIntosh County (Fig. 6). A thickness of 26 in. was measured in a core near the center of sec. 8, T. 11 N., R. 17 E. (Fig. 6, column 9; Hemish, 1988a, p. 54). However, coal company drill logs show that the coal bed becomes shaly and thins to 2 in. within 0.5 mi to the south. The Peters Chapel coal has not been identified south of sec. 8, T. 11 N., R. 17 E. It also thins rapidly north of the Muskogee–Wagoner County line, and pinches out in the vicinity of Porter, southwestern Wagoner County.

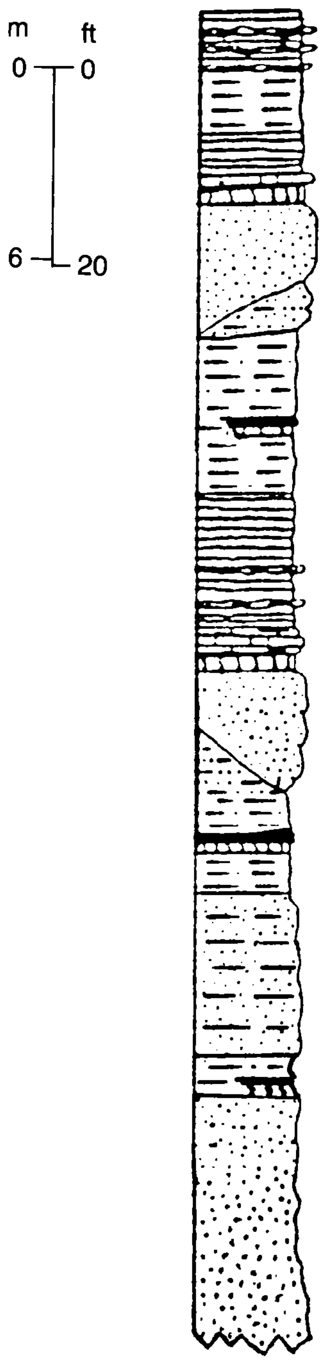


Figure 4. Generalized geologic column showing strata of the lower Boggy Formation in the study area. Modified from Hemish (1986a, fig. 4).

The Peters Chapel coal has been mined at several locations (mostly underground) in northwestern Muskogee County, as well as at a few locations near Checotah in McIntosh County (Hemish, in preparation b). In the Checotah area the Peters Chapel coal had not been named prior to this report. It may have been one of several coals erroneously called the "Secor" coal by Oakes and Koontz (1967, p. 47).

The uppermost coal bed in the lower Boggy Formation is the Bluejacket coal, discussed earlier in connection with the Inola Limestone. It is the only coal bed in the lower Boggy Formation that persists northward beyond the study area. It has been identified as far north as northern Craig County, just south of the Kansas–Oklahoma state line (Hemish, 1986b, p. 13).

COAL QUALITY

Table 2 shows typical analyses of the five coal beds present in the study area. In accordance with guidelines set by the American Society for Testing and Materials (ASTM), only samples collected from active coal mines or from fresh cores were used for analytical purposes (ASTM, 1987, p. 226, 7.1). Seventy-one coal samples were analyzed by the OGS Inorganic Chemistry Laboratory. The coals have been ranked in accordance with standard ASTM methods (ASTM, 1987, p. 225–228). Analyses indicate that all the tested coals are in the bituminous class, in the high-volatile A group.

In Table 2, analytical results for the Secor coal are reported by specific areas within the study area because of marked differences in coal quality. The writer believes that these differences (primarily in ash and sulfur content) are a result of different depositional environments. In northeastern McIntosh County (area 5) the Secor and Secor rider coal beds are coalesced to form one bed, called the Secor coal bed. In this area the Secor coal has analytical properties similar to the Secor rider coal. For a complete discussion on these coals in areas 4 and 5, see Hemish (1988b).

Data reported in Table 2 indicate that the coals are generally very low in moisture content, averaging <2%. Ash content is varied. For example, analyses of the Bluejacket coal show an average ash content of >27%; analyses of the Secor rider coal, almost 18%; and analyses of the Peters Chapel coal, almost 16%—all in the high-ash range. Conversely, analyses of the Secor coal from areas 3 and 4 show an average ash content of ~5%, well into the low-ash range. With the exception of the Secor coal in areas 3 and 4, the sulfur content of the coals is generally high, averaging 5.6%.

The sulfur content of the Secor coal from areas 3 and 4 is in the low- to medium-sulfur range, averaging 1.2%. Coal that averages <1% sulfur has been

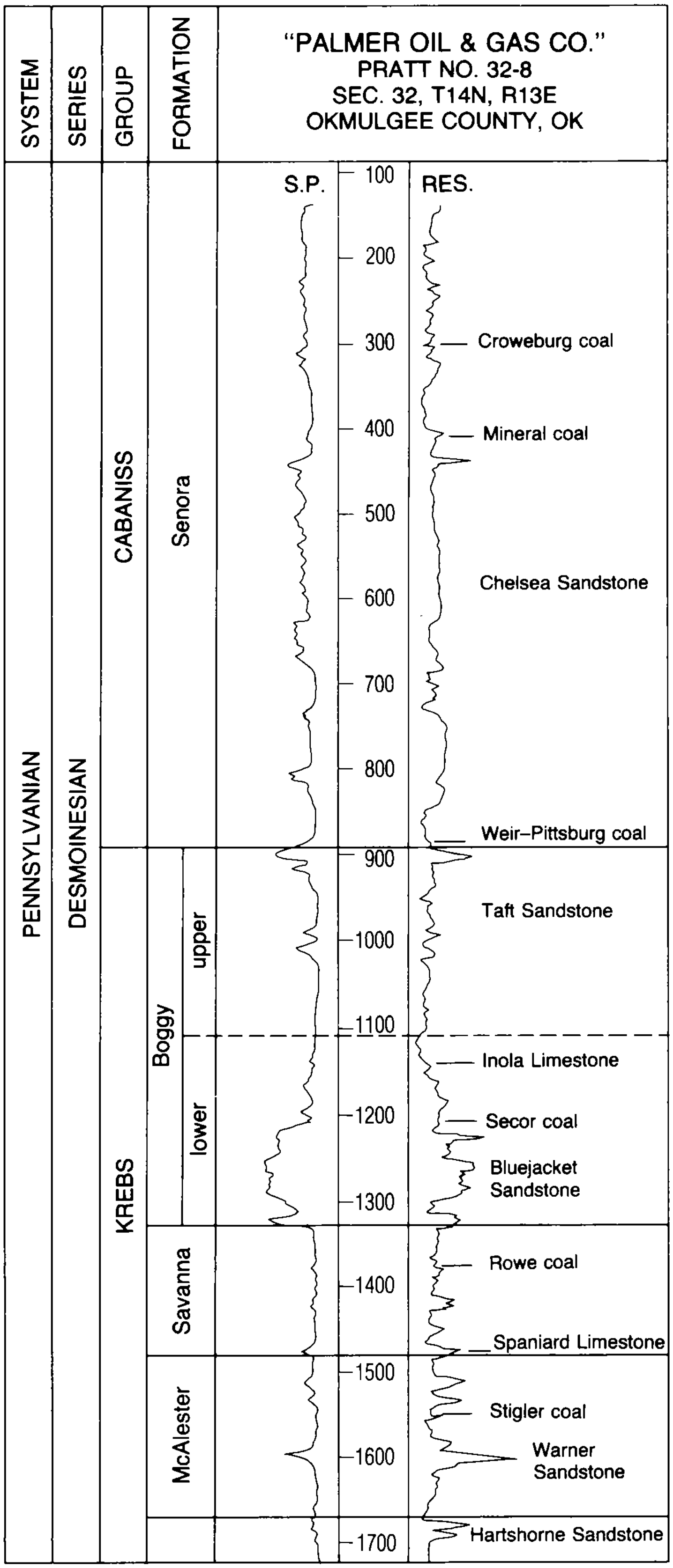


Figure 5. Electric log from a drill hole west of the study area, showing sequence and characteristics of the lower Boggy Formation and associated strata. From Hemish (1986a, fig. 3).

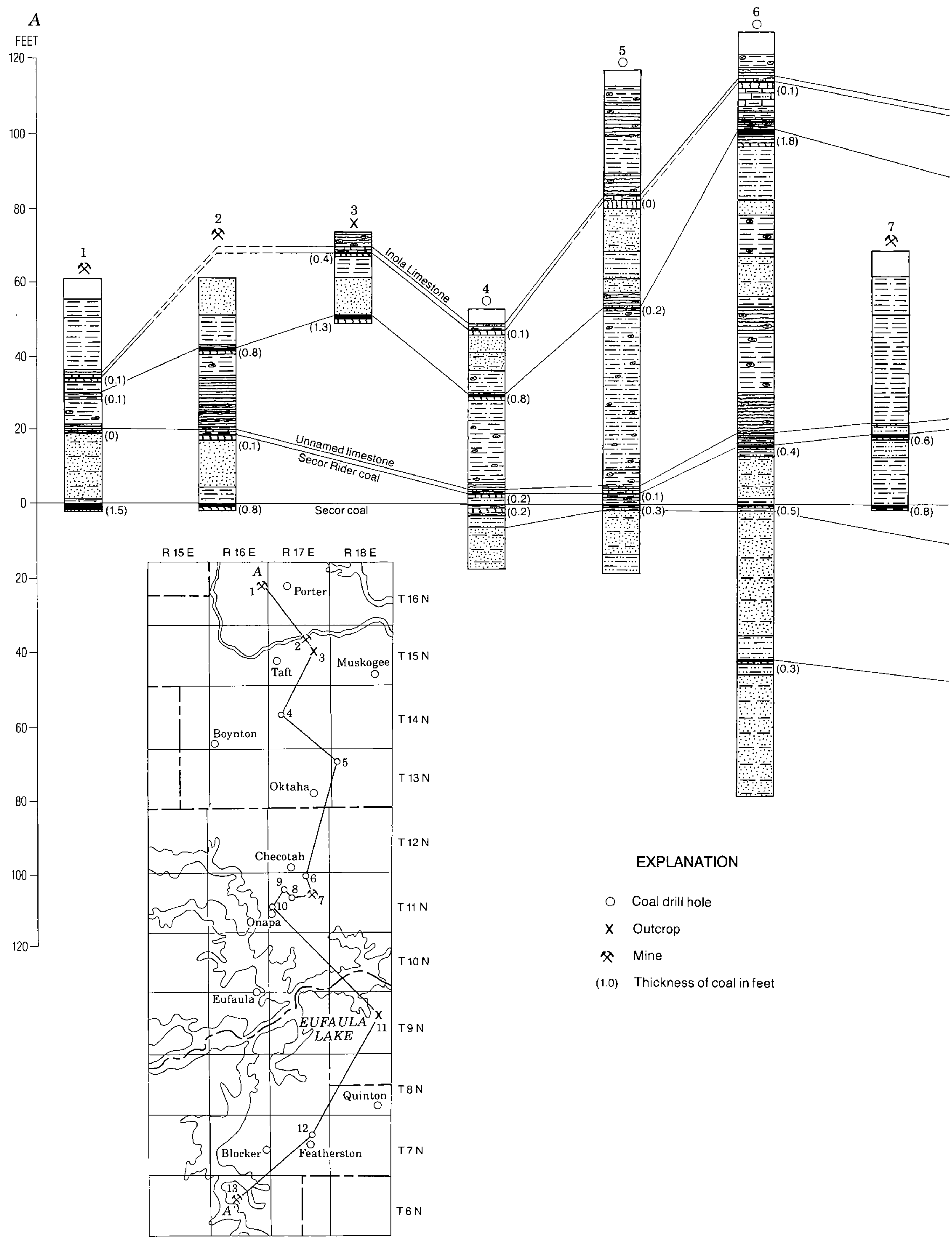


Figure 6. Correlations of key stratigraphic units in the lower part of the Boggy Formation in the study area. No horizontal scale.

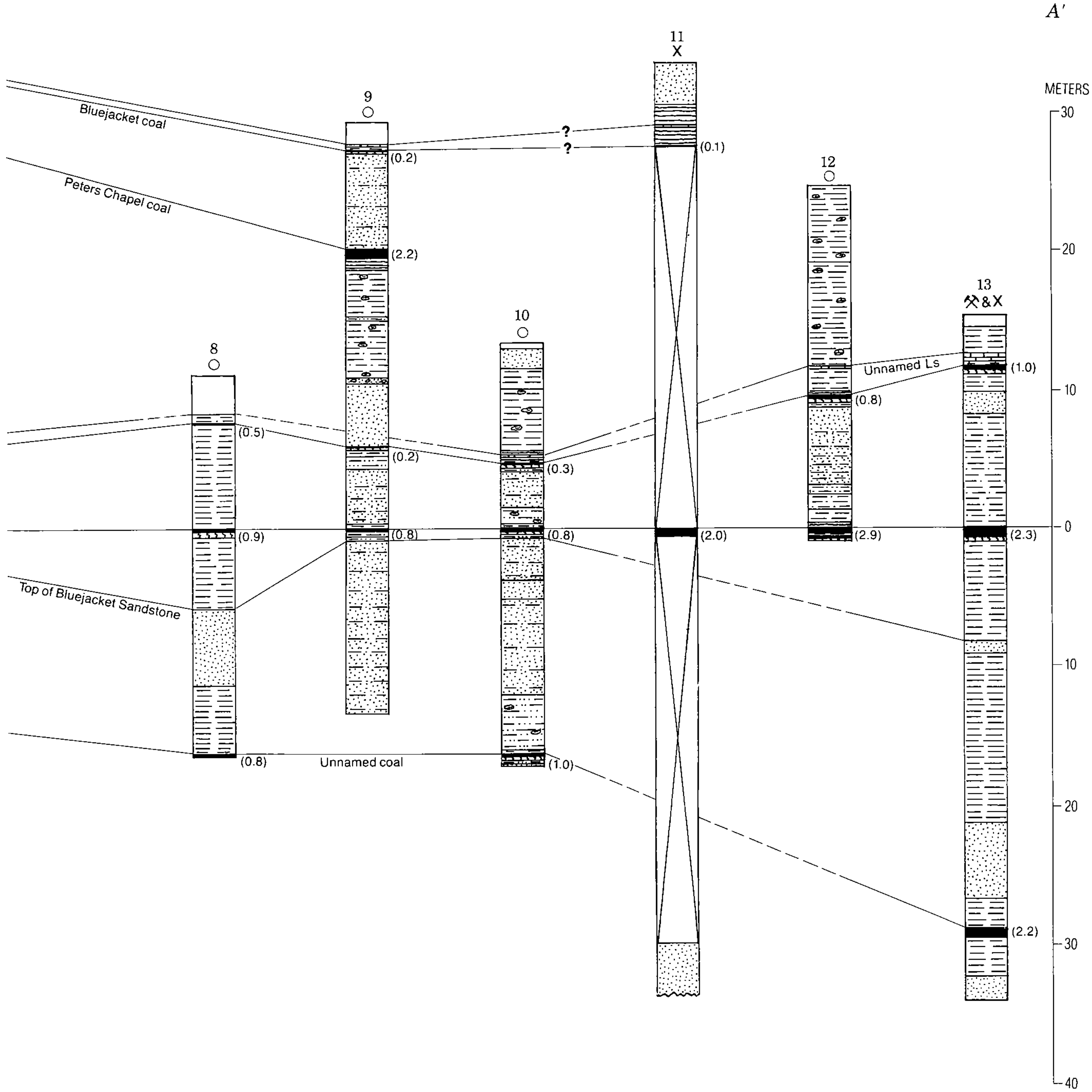

A′
METERS
30
20
10
0
10
20
30
40
8
9
10
11
12
13
Bluejacket coal
Peters Chapel coal
Top of Bluejacket Sandstone
Unnamed Ls
Unnamed coal
?
?
(0.2)
(0.1)
(2.2)
(1.0)
(0.8)
(0.5)
(0.2)
(0.3)
(0.9)
(0.8)
(0.8)
(2.0)
(2.9)
(2.3)
(0.8)
(1.0)
(2.2)

TABLE 1.—ANALYSES OF SECOR COAL SAMPLES COLLECTED IN ACTIVE MINES IN MCINTOSH, MUSKOGEE, AND WAGONER COUNTIES BY HEMISH IN 1988

Mine and operator	Location	Proximate analysis[a] (%) Moisture	Volatile matter	Fixed carbon	Ash	Sulfur (%)	Btu/lb	Rank[b]
Inter-Chem #2 Inter-Chem Co.	SW¼SE¼NE¼SW¼ sec. 6, T. 11 N., R. 18 E., McIntosh County	2.4	33.9	60.4	3.4	0.5	14,569	hvAb
Pollyanna No. 5 P & K Co., Ltd.	NW¼SE¼NE¼SW¼ sec. 34, T. 15 N., R. 17 E., Muskogee County	2.9	36.4	58.6	2.0	0.5	14,530	hvAb
Hobbs-O'Berg Mine Hobbs-O'Berg Mining Co.	NE¼SW¼NE¼SW¼ sec. 9, T. 16 N., R. 17 E., Wagoner County	4.3	36.7	56.4	2.6	1.3	14,178	hvAb

Analyses by OGS Inorganic Chemistry Laboratory.
[a]As-received basis.
[b]hvAb = high-volatile A bituminous (ASTM, 1987, p. 226, **6.2**).

TABLE 2.—AVERAGE ANALYSES OF COALS IN THE LOWER BOGGY FORMATION, MCINTOSH, MUSKOGEE, PITTSBURG, AND WAGONER COUNTIES

Coal bed name and area sampled[a]	Proximate analysis[b] (%) Moisture	Volatile matter	Fixed carbon	Ash	Sulfur (%)	Btu/lb	Number of analyses (whole seam)	Rank[c]
Bluejacket coal Area 1	1.4	33.0	38.4	27.2	5.3	10,544	2	hvAb
Peters Chapel coal Area 1	1.9	35.5	46.8	15.8	6.8	12,336	13	hvAb
Secor Rider coal Area 2	1.8	35.0	45.4	17.8	7.6	11,932	12	hvAb
Secor coal Area 3	2.8	36.2	56.4	4.6	1.4	14,013	20	hvAb
Secor coal Area 4	1.6	34.3	58.9	5.2	1.0	14,268	13	hvAb
Secor coal Area 5	1.1	34.3	51.8	12.8	5.6	12,736	2	hvAb
Secor coal Area 6	2.5	40.2	47.0	10.3	4.5	13,053	4	hvAb
Unnamed coal Area 7	1.7	34.9	55.2	8.2	4.0	13,713	5	hvAb

Analyses by OGS Inorganic Chemistry Laboratory.
[a]*Area 1* = McIntosh and Muskogee Counties; *Area 2* = McIntosh and Muskogee Counties and northern Pittsburg County; *Area 3* = Muskogee and Wagoner Counties; *Area 4* = McIntosh County (area where Secor and Secor rider coal beds are split); *Area 5* = northeastern McIntosh County (area where Secor and Secor rider coal beds are coalesced); *Area 6* = northern Pittsburg County; *Area 7* = McIntosh County and northern Pittsburg County.
[b]As-received basis.
[c]hvAb = high-volatile A bituminous (ASTM, 1987, p. 226, **6.2**).

produced from several mines in McIntosh County and northern Muskogee County. The Secor coal generally has the highest calorific value, which, along with its other desirable analytical properties, makes it the most important commercial coal in the area.

SUMMARY

Five coal beds occur in the lower Boggy Formation in the transition area from the northeastern Oklahoma shelf to the Arkoma basin. The oldest coal occurs in the upper part of the Bluejacket Sandstone Member, and is unnamed in the area of this report. This coal bed ranges from 0.1 to 1.3 ft in thickness, has a high sulfur content, a medium ash content, and an average calorific value of 13,713 Btu/lb. It is not currently mined.

The Secor coal occurs just above the Bluejacket Sandstone Member. Thickness ranges from 0.1 to 4.3 ft. Prior to geologic investigations by the writer, the coal stratigraphy of the shelf-to-basin transition area was poorly understood. The name "Secor" was applied to several coal beds in McIntosh County that occur from a few feet to as much as 100 ft or more above the Bluejacket Sandstone (Oakes and Koontz, 1967, p. 47). Apparently, subsurface data were unavailable at the time of Oakes and Koontz's work, as they did not attempt to differentiate the various coals exposed in the area. It seems likely that any of the four coals occurring between the base of the Boggy Formation and the base of the Bluejacket coal might have been called "Secor." Oakes and Koontz (1967, p. 48) did observe the coal here correlated with the Bluejacket coal. They wrote, "A coal seam from 0.5 to 2 inches thick lies from 0.5 to 3 inches below the Inola Limestone at most exposures of the Inola, as mapped in McIntosh County." Some confusion in mapping also occurred in Muskogee County, where the Peters Chapel coal was misidentified as the Secor coal. In some places, the outcrop boundary of the Peters Chapel coal, marked by numerous slope mines, was mapped as Secor by Oakes (1977, pl. 1). This error was explained and corrected by Hemish (1986a).

The Secor rider coal bed occurs from a few inches to 40 ft above the Secor coal. It coalesces with the underlying Secor in east-central McIntosh County. It is of minable thickness in northern Pittsburg County, and locally in McIntosh County, but it is not currently exploited, because of its high-ash, high-sulfur content.

The Peters Chapel coal occurs from ~30 ft above the Secor coal in Wagoner County, to ~50 ft above in Muskogee County, to ~100 ft above in McIntosh County. It, too, is a high-ash, high-sulfur coal, and, although it is of minable thickness, it is not currently exploited.

The uppermost coal in the lower Boggy Formation is the Bluejacket coal, which is not known to be more than 2 in. thick where identified with certainty in the study area. Although it has no commercial value, it is a useful stratigraphic marker.

Correlations in the shelf-to-basin transition area were hindered by the apparent lack of deposition of any of the five coal beds discussed herein in an area several miles wide just north of the McIntosh–Muskogee County line. The writer believes that the area was probably structurally high during deposition of the lower Boggy sediments, and coal swamps did not exist in the area. However, Figure 6 shows that stratigraphic sequences on either side of this area are sufficiently similar to permit correlation of the various units. Future exploration for coal will be enhanced by the new stratigraphic knowledge gained from this study.

DELTA PATTERNS AND PETROLEUM OCCURRENCES IN THE PENNSYLVANIAN BARTLESVILLE SANDSTONE OF EASTERN OKLAHOMA*

Glenn S. Visher

ABSTRACT.—Data from modern deltas have made it possible to interpret the depositional origin of a number of Pennsylvanian sandstones in northeastern Oklahoma. Objective criteria from cores and outcrops used to define specific areal and vertical depositional patterns include (1) vertical patterns of sedimentary structures, bedding, and grain size; (2) clay mineralogy and detrital clasts; (3) trace fossils; and (4) detailed analysis of textures.

The processes of progradation and maturation are used to develop a four-dimensional deltaic model. Six subdivisions are distinguishable: (1) lower alluvial plain; (2) upper deltaic plain; (3) lower deltaic plain; (4) subaqueous sand sheet; (5) marginal basin; and (6) marginal depositional plain.

Lower and Middle Pennsylvanian sediments were deposited during an overall transgression, but the transgression was marked by extensive regressions. The sandstones are distributed over thousands of square miles, but locally they are lenticular. Each regression is in response to the outbuilding of the shoreline under static sea-level conditions.

INTRODUCTION

Interpretation of the genesis of oil-bearing sandstone in the Midcontinent has been of interest for many decades. Of importance is the recognition of specific vertical profiles of lithologic units, sedimentary structures, textures, and the many other specific physical and biologic aspects of a single genetic time-rock unit.

The process-response model has become the "touchstone" of sedimentologists. Study of the fourth dimension—time—has provided insight into the interpretation of particular areal and vertical sedimentary patterns. By comparing sequences of depositional units from the Holocene with those in ancient sediments, it is possible to determine the basic physical response to a series of events.

STRATIGRAPHIC FRAMEWORK

Figure 1 is a pre-Pennsylvanian paleogeologic map (Jordan, 1962). This map illustrates the tectonic framework at the beginning of the Pennsylvanian, indicating faulted uplifts along the Nemaha ridge and the positions of the Arbuckle Mountains and the Oklahoma City highland. The edge of the continental shelf is well defined by the successively younger stratigraphic units underlying the unconformity to the south. The Ozark dome may have been covered by Chesterian strata (and possibly also by Lower or Middle Pennsylvanian strata).

An isopach map of an interval above the pre-Pennsylvanian unconformity illustrates the continuing effects of these features on deposition (Fig. 2). The interval includes the uppermost Mississippian stratigraphic units and Pennsylvanian strata including Morrowan, Atokan, and most of the Desmoinesian Series. Of major importance is the total thickness indicated by this isopach. The 5,000 ft of shallow marine strata suggests that this period was a major time of subsidence on the continental margin. It also indicates that it was a major period of onlap, as indicated by the progressive thinning toward the areas of truncated Mississippian strata. The thicknesses of sediment indicated by this isopach map are minor compared with those deposited farther south and east in the Arkoma basin.

Transgressive Character of the Lower Absaroka Sequence

The transgressive character of the lower part of the Absaroka sequence is well documented. The cross section and map (Fig. 3) illustrate the successive onlap of younger time-stratigraphic sedimentary units. The onlap suggests a relative rise in sea level, and may represent in part a topographic surface that was inundated as sea level rose.

*Extracted from "Pennsylvanian Delta Patterns and Petroleum Occurrences in Eastern Oklahoma," by G. S. Visher, S. Saitta, and R. S. Phares, American Association of Petroleum Geologists *Bulletin*, v. 55, no. 8, 1971.

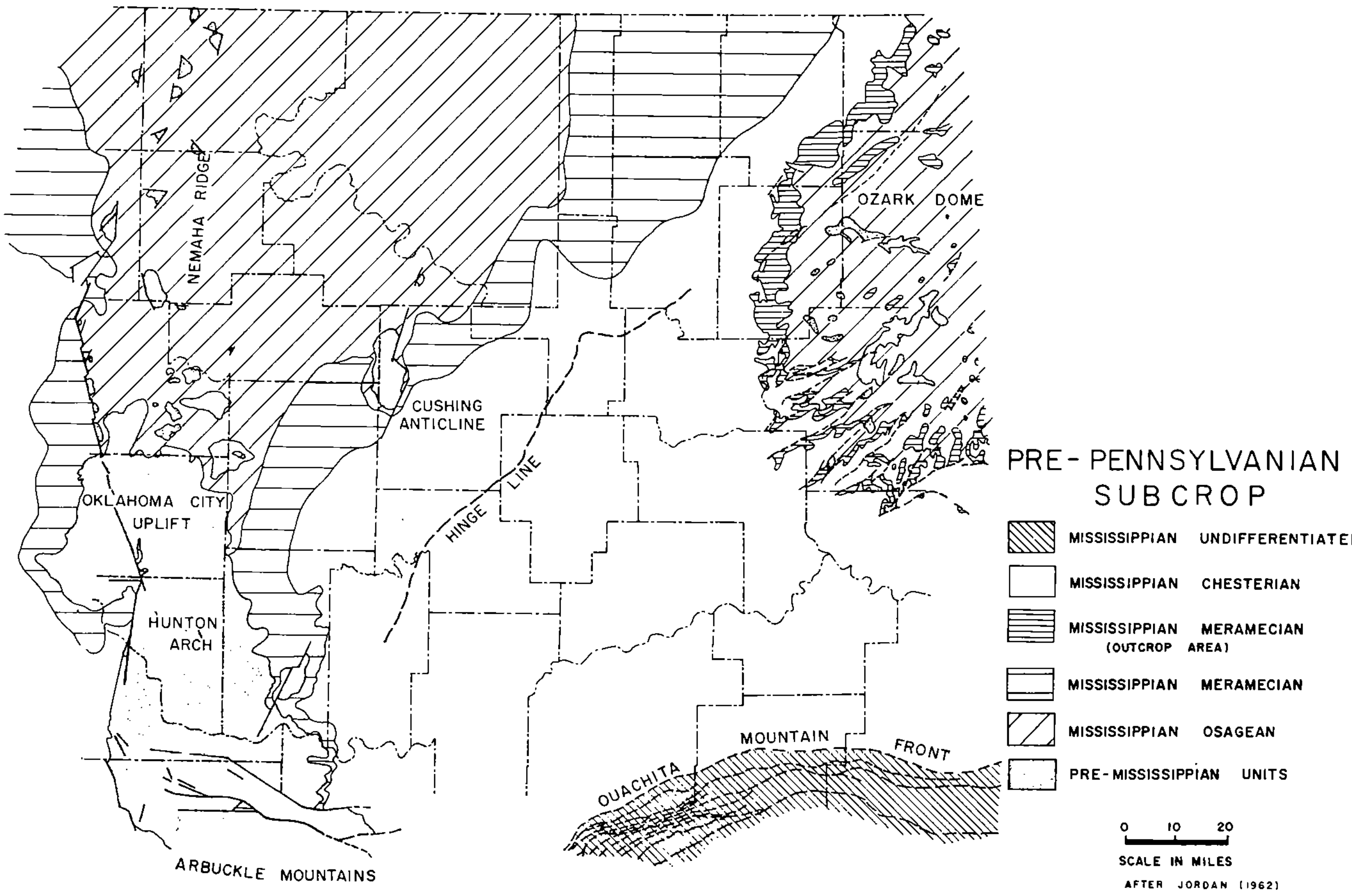


Figure 1. Subcrop map below pre-Pennsylvanian unconformity. Map shows Nemaha ridge and uplifts south of Oklahoma City. Ozark dome probably reflects post-Pennsylvanian truncation. Ouachita Mountains area shows no pre-Pennsylvanian truncation. After Jordan (1962).

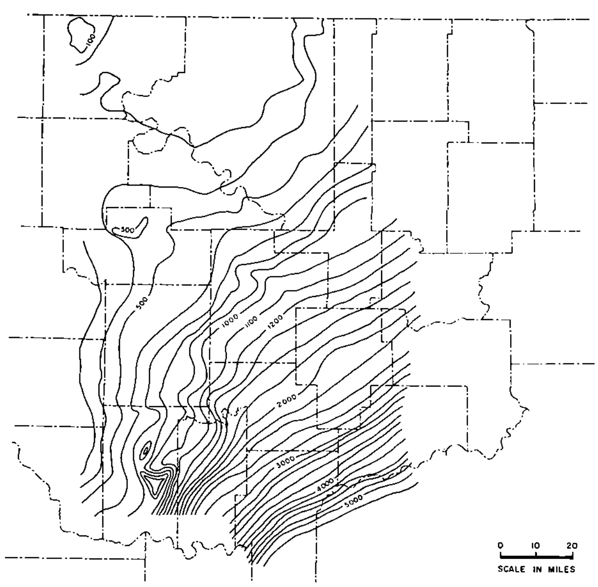


Figure 2. Isopach map from Mayes (Meremecian) to base of Oswego (upper Desmoinesian). This interval includes all sediments deposited during transgression from basin across Oklahoma shelf.

Onlap from South to North

The detailed correlation of thin shelf limestone and other marker-defined units indicates the time sequence of the onlap (Fig. 3). The successive pattern of younger time-stratigraphic units extending farther onto the shelf is well illustrated. The Verdigris Limestone of late Desmoinesian age transgresses the tectonic features outlined by the subcrop map (Fig. 1). The transgression continued for possibly 15 million years and produced a sequence of sediments in excess of 5,000 ft thick. Such a transgression is similar in scope to the basal Cambrian, early Middle Ordovician (St. Peter), or the late Early Devonian (Oriskany) transgressions, which represent the beginnings of other sequences (Sloss, 1963).

DEPOSITIONAL HISTORY

Regressive Sand Units of the Lower Absaroka Sequence

The depositional pattern is one of major regressive progradations interrupting a continuous transgressive depositional period. Each time-stratigraphic

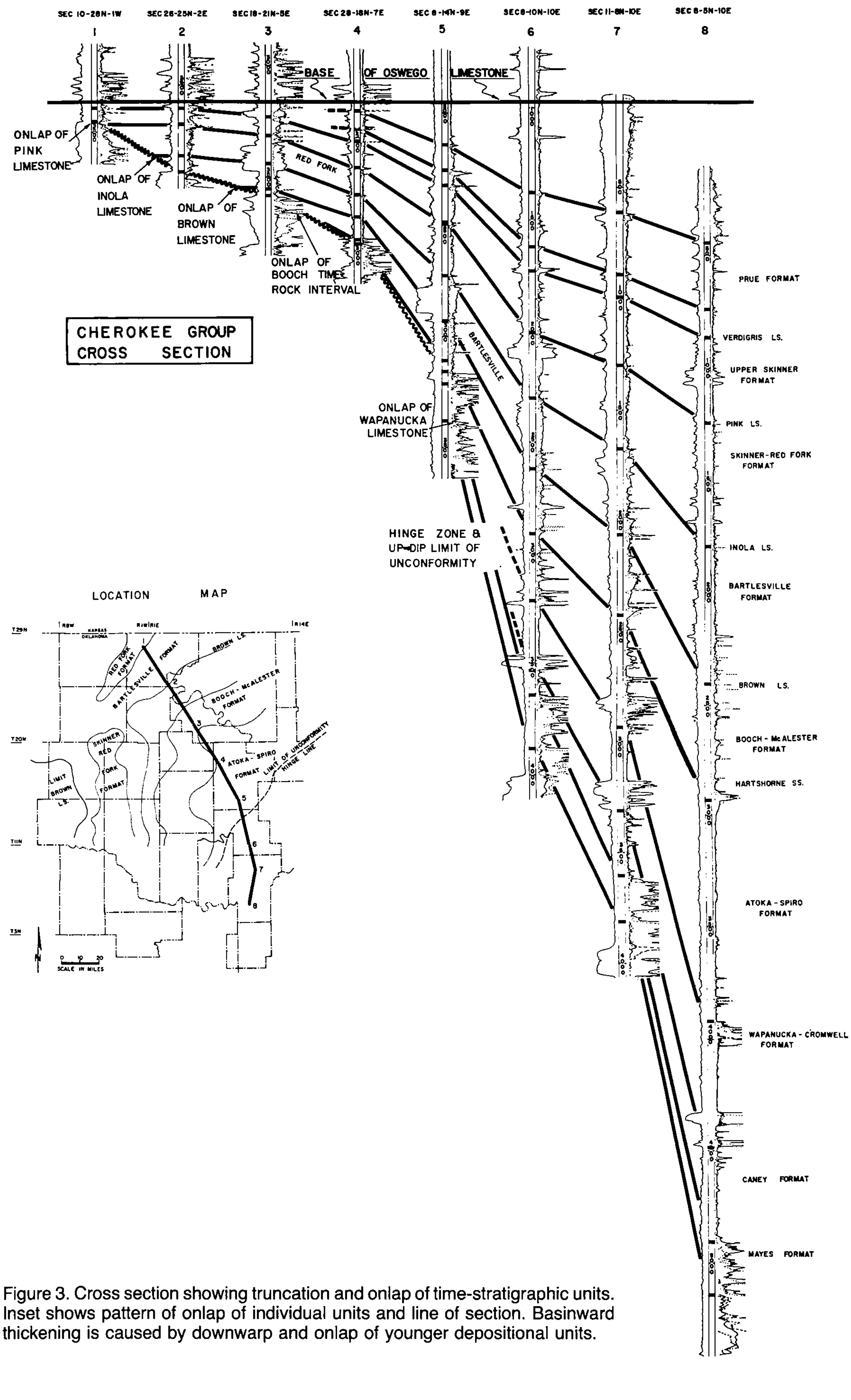


Figure 3. Cross section showing truncation and onlap of time-stratigraphic units. Inset shows pattern of onlap of individual units and line of section. Basinward thickening is caused by downwarp and onlap of younger depositional units.

unit represents a period of regression with a common vertical pattern of depositional environments. This is well illustrated by the pattern of sandstone development within the interval (Fig. 4). The map shows an overlapping pattern of sandstone sheets branching southward and a single narrow sandstone body on the north. Also, there is evidence that the depocenter for younger sandstone bodies developed farther north than for older units. This would suggest a subsiding continental-shelf margin.

Each sandstone has the general shape of a large deltaic unit, and they all show a depositional pattern closely related to the isopach map (Fig. 2).

The continental interior was the probable source of these sands, and their thickness suggests they are the product of a single large, integrated drainage system. As the transgression covered more of the craton, the source was diminished, and finally, after the deposition of the Prue sandstone, the northern source did not furnish sufficient sand for continued regressive progradations.

Shelf Sedimentation

The pattern of deposition indicated by these sandstones is one of progradation through short intervals of time. The interval from the depositional interface to the surface of the sea was filled with sediment. When the sediment–water interface reached sea level, sediment was bypassed to deeper water and the area was subsequently filled to sea level, producing a regression of the shoreline. Limestone, coal, bay sediments, and strandline sandstone deposits are associated with each of the units. Each is separated by a period of transgression produced by subsidence or by an equilibrium between rates of sediment supply, compaction, and the volume of the receptor area (Sloss, 1962). As the deltaic units prograded into deeper water, the rate of regression decreased, and the areal extent of the deltaic unit was limited. When this happened, the relative rise of sea level and compaction could overtake the regression of the strandline, and the overall transgression could begin again.

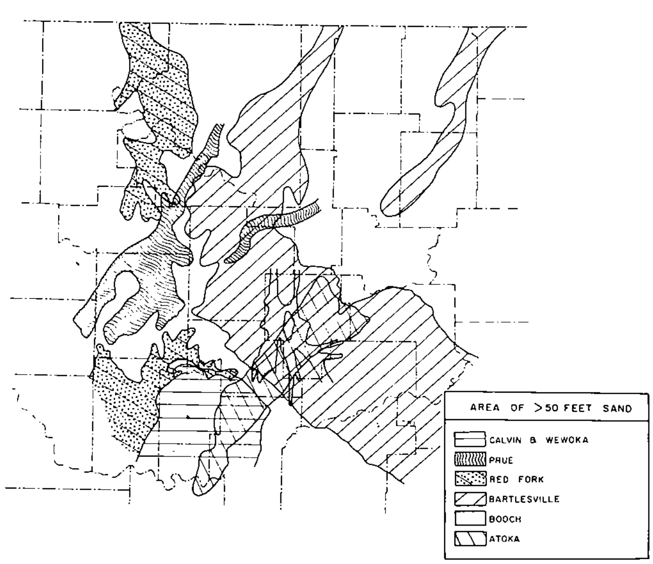


Figure 4. Delta patterns indicated by areas with more than 50 ft of sand deposition in individual units. Areal patterns indicate deepening water on the south during transgression and compaction of previously deposited interdeltaic shale units.

DELTAIC ORIGIN OF THE BARTLESVILLE SANDSTONE

The history of deposition of the lower Absaroka sequence in northeastern Oklahoma has been related to progradation of large deltaic units. Deltas are developed over tens of thousands of years and are composed of a complex of depositional environments. Their interpretation necessitates bringing together a large amount of data over an area of several thousand square miles. Most studies of sedimentary rocks do not cover the range of stratigraphic information needed, or the area necessary to determine the depositional history. Details of depositional history and the determination of local environments require close control not commonly available. The Bartlesville sandstone was chosen as the basis for developing a model of deltaic sedimentation because of the availability of many thousands of well logs, a nearly continuous outcrop along the eastern and northern edges of the delta, numerous cores, and because of its economic importance as a prolific oil-bearing sandstone interval.

Sand Distribution, Paleocurrents, and Isopach Maps

The gross sandstone map was based on thickness measurements from over 5,000 well logs (Fig. 5). These logs also were used as control points for the isopach map (Fig. 6). Additional logs were examined to determine the range of variability within each section (Phares, 1969; Saitta, 1968). Over much of the area, control density was approximately one well per square mile. This provided the detail to show the clearly developed distributary pattern that is indicated.

Paleocurrent measurements made along the outcrop support the pattern shown by the distributaries (Fig. 5). Measurements were made on trough axes of festoon cross-bedded sandstones, rib-and-furrow structures formed by migrating current ripples, current lineation, aligned wood fragments, bottom scour marks and, in a very few places, by the orientation of wave ripples. Several hundred measurements were made in the field, and average azimuths were computed for each outcrop. In nearly every outcrop the dispersion of directional measurements rarely exceed 90° and usually falls within a 45° spread. The types of measured sedimentary structures and the coherence of transport vectors indicate that unidirectional currents were the dominant process producing the deposition of the Bartlesville sandstone.

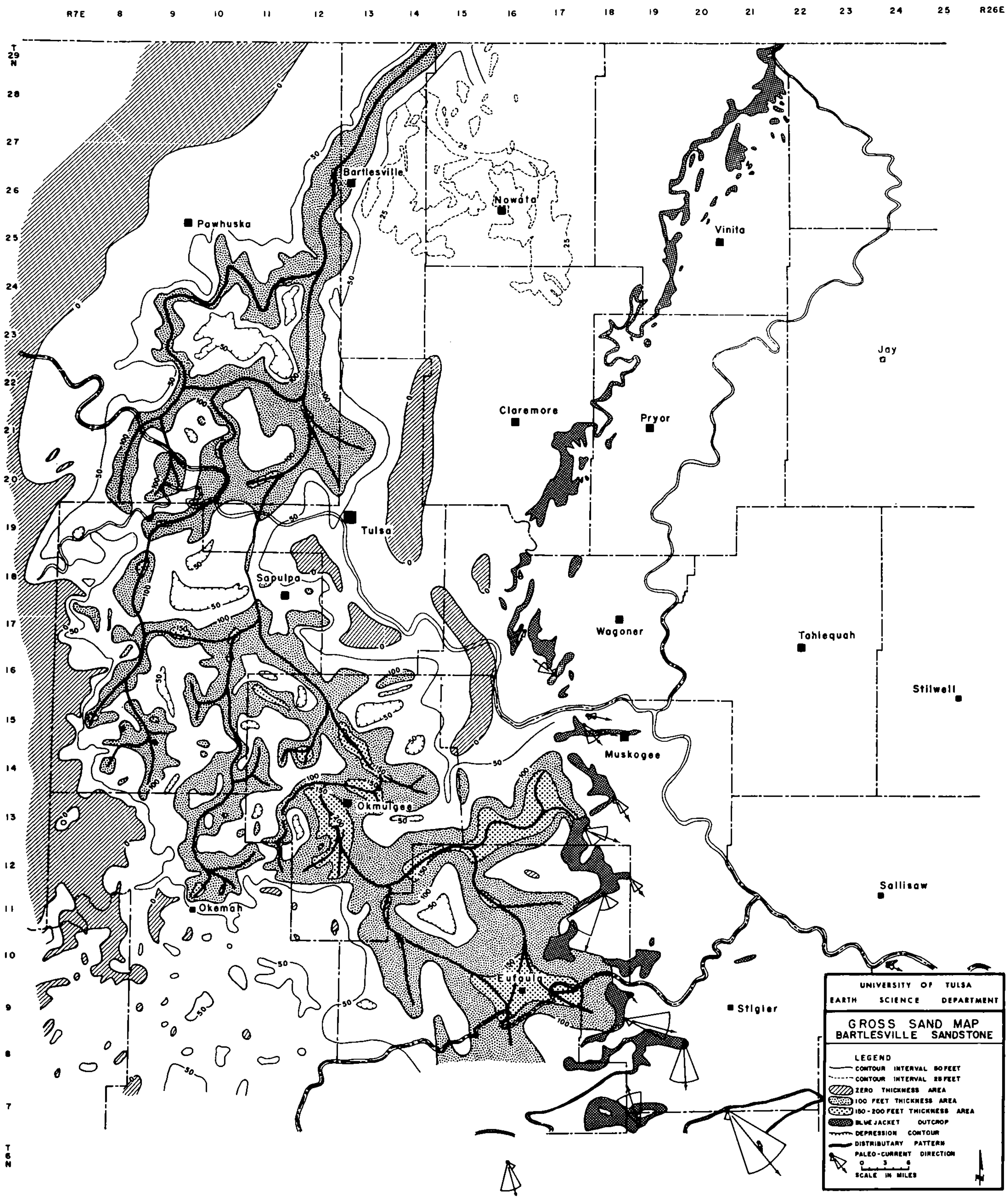


Figure 5. Sand distribution of Bartlesville sandstone. Map shows suggested distributary pattern and paleocurrent directions from outcrop studies.

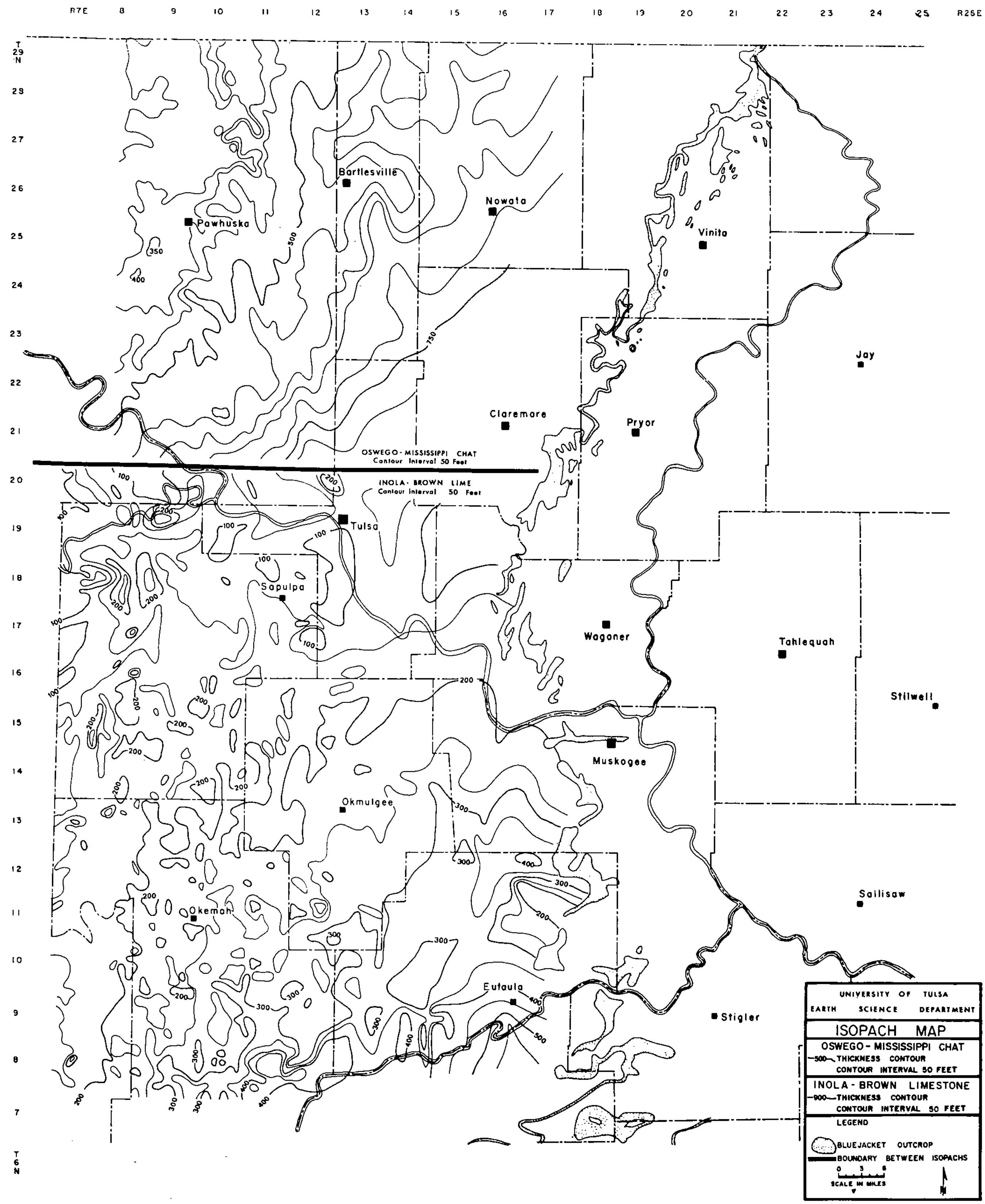


Figure 6. Detailed isopach map of interval including Bartlesville sandstone.

Study of the sand distribution patterns and the paleocurrent indicators alone provided much information on depositional environments within specific areas. A possible tidal inlet occurs on the south, where paleocurrents indicate a northern transport direction. The area northwest of Muskogee shows a north-northwest transport direction, which suggests crevasse deposits. The area south of Muskogee shows dispersion of the paleocurrent measurements, and examination of outcrops suggests that these may be related to bends in a distributary. The value of the sandstone-distribution and paleocurrent map is that it provides a framework for interpreting detailed information from specific areas.

The isopach map (Fig. 6) shows the topographic framework during the deposition of the Bartlesville. A marginal highland was present on the northwest and a basin on the southeast. Local highs occur on the Cushing structure and southwest of Tulsa. Thickness of the interval increases progressively southeastward.

Sedimentary Structures and Textures

Twenty outcrops and 10 cored sections were examined in detail. The nature of the sedimentary structures was recorded, and sandstone units were sampled for textural analysis. Over 200 grain-size distributions were plotted. These data were used to develop depositional sequences to relate to electric logs and to textural plots collected from modern depositional environments.

Direct comparisons were made between the types of sedimentary structures developed within the cores and the grain-size distribution of the sand. Representative units show festoon cross-bedding, current ripples, flaser bedding, and a bioturbated sandstone. The shapes of the log-probability plots are similar to those described for the Mississippi delta (Visher, 1969). The types of sedimentary structures developed are the same as those described by Coleman and Gagliano (1965) for the Mississippi delta. They suggest a range in current velocities from very weak to strong. Trace fossils are not common in either cores or outcrop exposures. Some sandstones and many of the shales show the effects of bioturbation, but thick sandstone units contain only current structures.

Vertical Sedimentary Patterns

Much of the control is based on old electric logs. The vertical patterns interpreted from the electric logs are easily related to the sequence of textures and lithologies. An examination of several thousand logs indicates that six fundamental patterns are developed within the Bartlesville deltaic plain (Fig. 7).

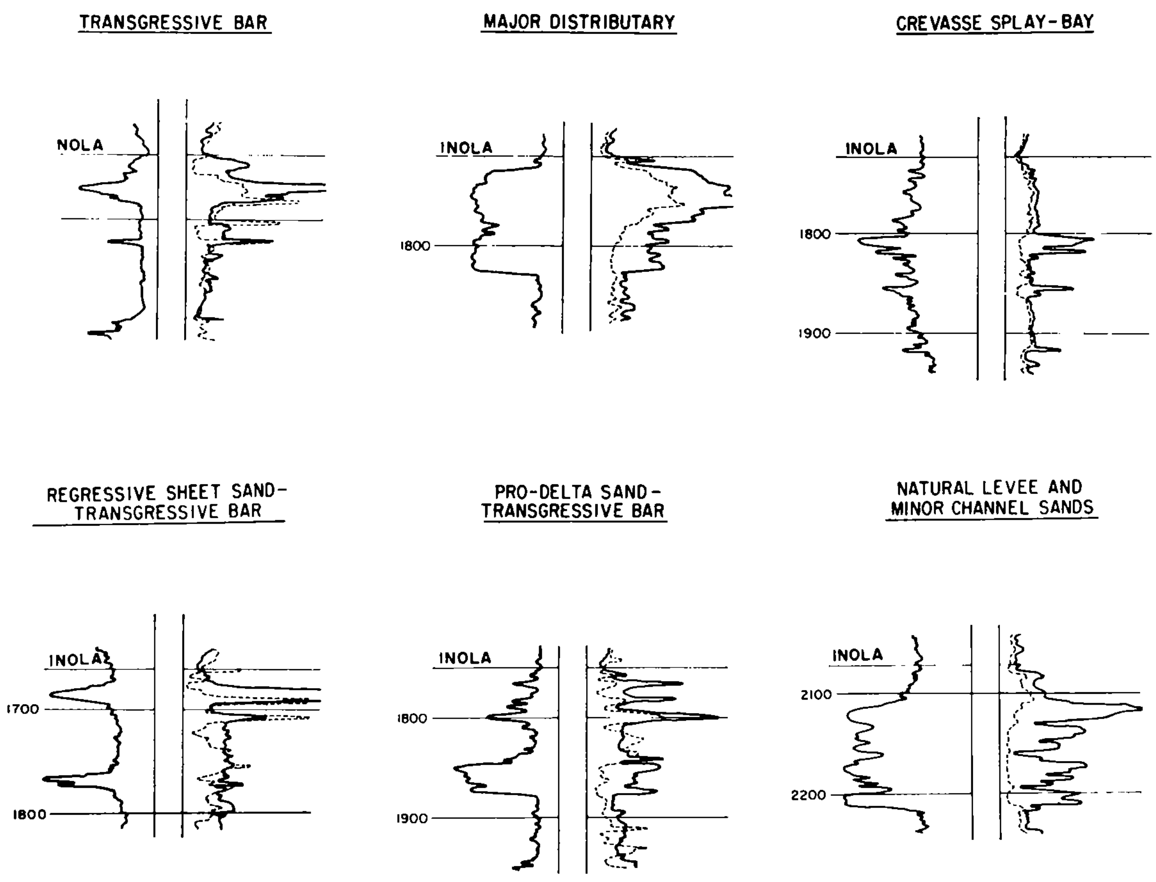


Figure 7. Fundamental types of SP log patterns. Individual environmental sequences are illustrated. Sequences based on lithologic and textural comparisons with cored and outcrop sections.

Each pattern is characteristic of a particular part or subenvironment in the delta. The environmental interpretation is based on similarity between logs from cored sections and studied outcrops. These patterns are typical and are believed to represent the sequence of depositional events that normally occur within a particular subenvironment of the delta. It is believed that they would be directly comparable to logs from other deltas of very different age and geographic distribution.

Environmental Reconstruction

It is difficult to reconstruct a detailed depositional environmental pattern that spreads over an area of thousands of square miles and represents deposition during tens of thousands of years. The primary basis for the reconstruction is the vertical sequence of depositional units indicated by the electric logs (Fig. 8). The electric-log patterns suggest that there is a developmental sequence represented by the vertical profile of textures, structures, and lithologic units. By using vertical sedimentary patterns and correlating them with geometry, paleocurrents, sedimentary structures, mineralogy, textures, and petrography, it is possible to reconstruct the areal pattern of particular subenvironments. Work by Coleman and Gagliano (1964) suggests that a major sixfold classification of deltaic elements can be developed for most modern deltas. These include (1) alluvial valley; (2) upper deltaic plain; (3) lower deltaic plain; (4) subaqueous sand sheet (distributary mouth bar); (5) marginal basin; and (6) marginal depositional plain. These six units have been defined for the Bartlesville sandstone. Each area has a characteristic log shape, each may be related to a specific geomorphic area, and each is a product of specific depositional processes.

More-detailed environmental interpretations have been suggested by the sedimentary structure and sandstone textures for parts of the deltaic complex. On the northeast a large area of sandstone has been interpreted to be a product of crevasse splay deposition. Cores from interdistributary areas show silt and fine sand deposited from suspension, and flaser bedding with burrows; this suggests deposition in a bay environment. Paleocurrents, and sedimentary structures and textures, indicate that certain outcrop areas are tidally deposited. On the northwest, deposition appears to be related to marine bars, possibly as a result of a later transgression. These details cannot be verified by use of electric-log shapes, but they are consistent with the pattern of deposition.

DELTAIC MODEL

A generalized deltaic model must incorporate several different features. It should include the elements of a delta described by Coleman and Gagliano (1964), the processes important in the building of a delta, and, most important, the developmental history of a deltaic sandstone. These aspects taken together can be used to explain a single depositional sequence and a particular pattern of depositional environments, and used to define the geomorphic elements within the large progradational deltaic units developed during the deposition of the lower Absaroka sequence.

Regressive and Transgressive Patterns

Progradation does not end the formation of a deltaic complex. In the upper parts of a delta, deposition continues with the filling of bays and marshes with sediment, and finally natural levees and overbank deposits fill the area to sea level or slightly above. The pattern can be shown by a cross section from the upper part of the delta (Fig. 9). There, crevasse deposits are illustrated on the northeast, and bay, marsh, and restricted marine deposits on the southwest.

Also illustrated is the presence of a sandstone unit across the top of the section, overlain by a limestone. This represents the closing episode of the abandonment of a delta and transgression of the sea across the delta complex. The same upper sandstone unit can be seen in logs illustrated in the south and southwest of the mapped area (Fig. 8). This transgressive sandstone covers the entire delta complex, but locally it formed a thicker marine bar sandstone separated by shale from the lower, more-massive sandstone unit. In other places the sandstone was produced by reworking of the natural levee and is capped by limestone.

The vertical sequence of lithologies, textures, and sedimentary structures depends on the position within the delta, the relation to waves, and the amount of compaction that may have taken place.

Deltaic Sequences

Three deltaic sequences can be used to characterize most sections within the Bartlesville sandstone (Fig. 10). These are the products of the historical development of the delta and should be similar to other lower Absaroka deltaic units.

The interdistributary sequence is similar to the marginal depositional basin element of the delta. It is gradational into the distributary type sequence. Variations in the thickness of the sandstones are related to proximity to a distributary. The basal marginal marine sandstone is related to early deposition on the floor of the bay when the distributary was discharging nearby. The areal distribution of this unit may be very widespread and depends upon depth of water, marine currents, and wave action. The amount of bioturbation and the types of sedimentary structures in the overlying thin sandstone depend on how open the bay was to marine waters and processes. The upper units also may be variable, but a transgressive unit of sandstone and limestone is nearly universal.

The distributary sequence depends on the position within the deltaic complex. If it has matured into an alluvial-plain-type channel, it may show a marked upward decrease in grain size. Also, if it is a major channel near the lower part of the delta, it may be

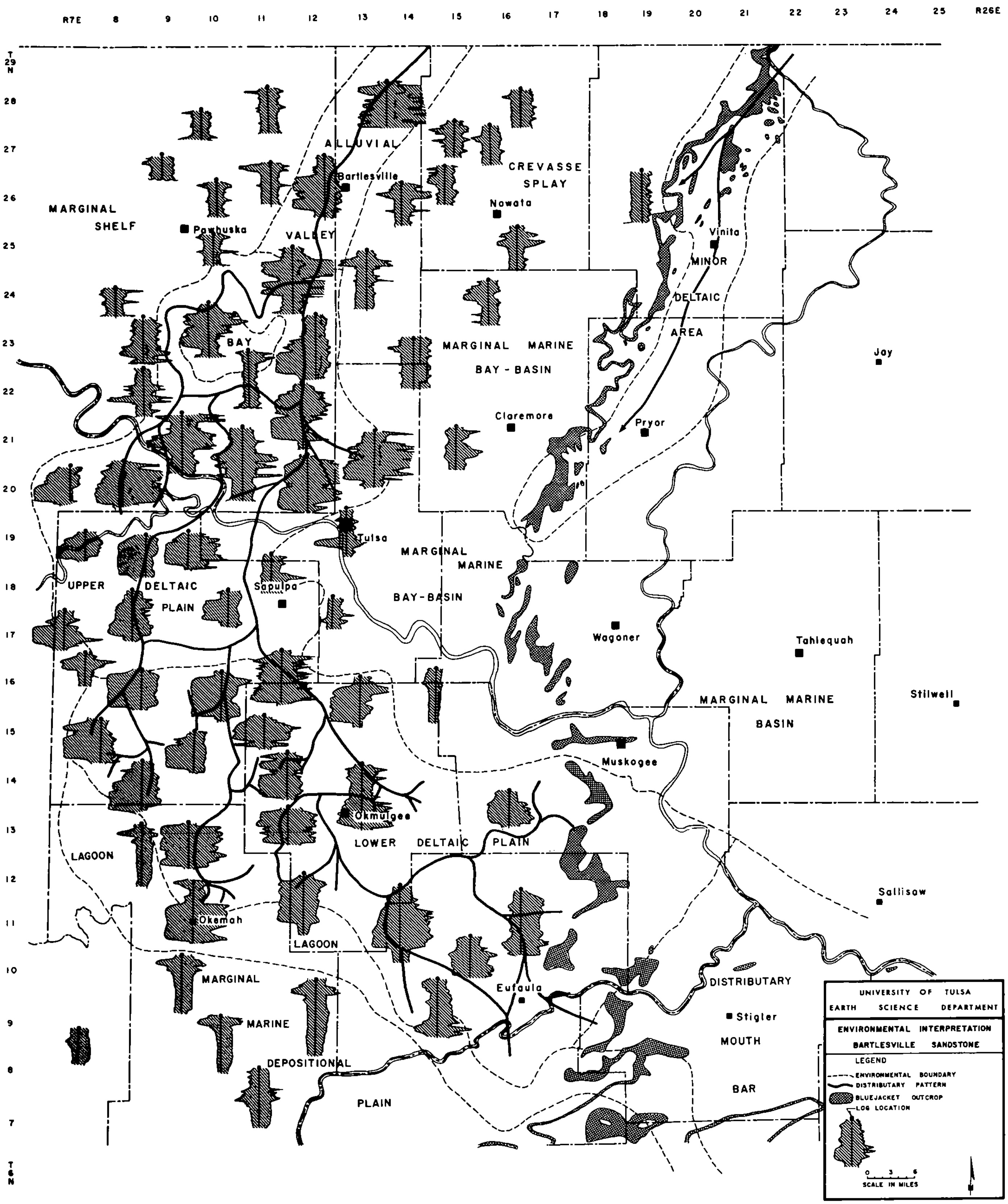


Figure 8. E-log patterns and environmental reconstruction of deltaic elements. Environmental interpretations based on sandstone geometry, vertical sequences, sedimentary structures, textures, and clay mineralogy.

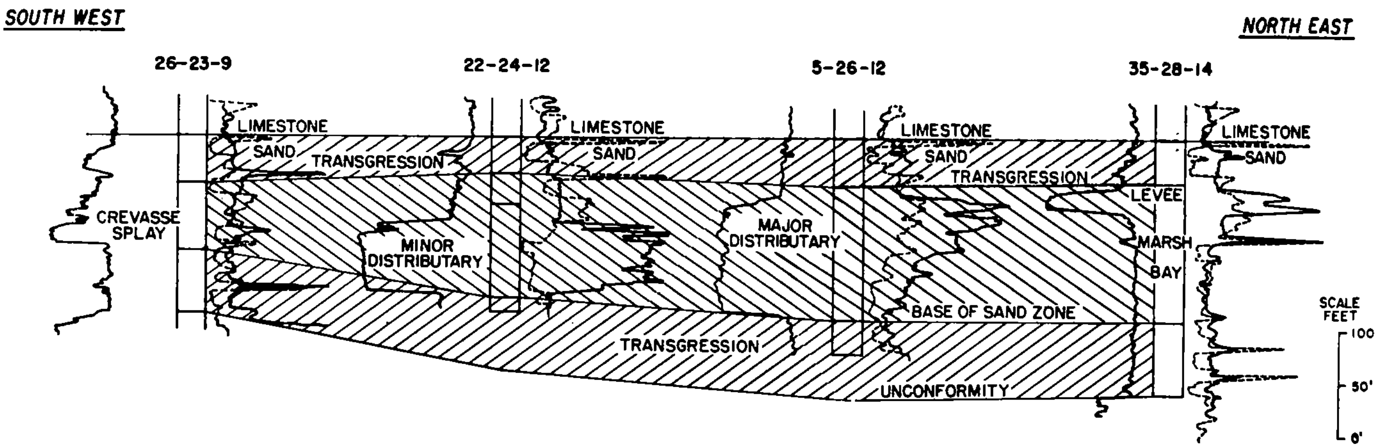


Figure 9. E-log section indicating development sequence within upper part of Bartlesville delta. Regressive and transgressive intervals are illustrated. Lateral variation in depositional units can be seen.

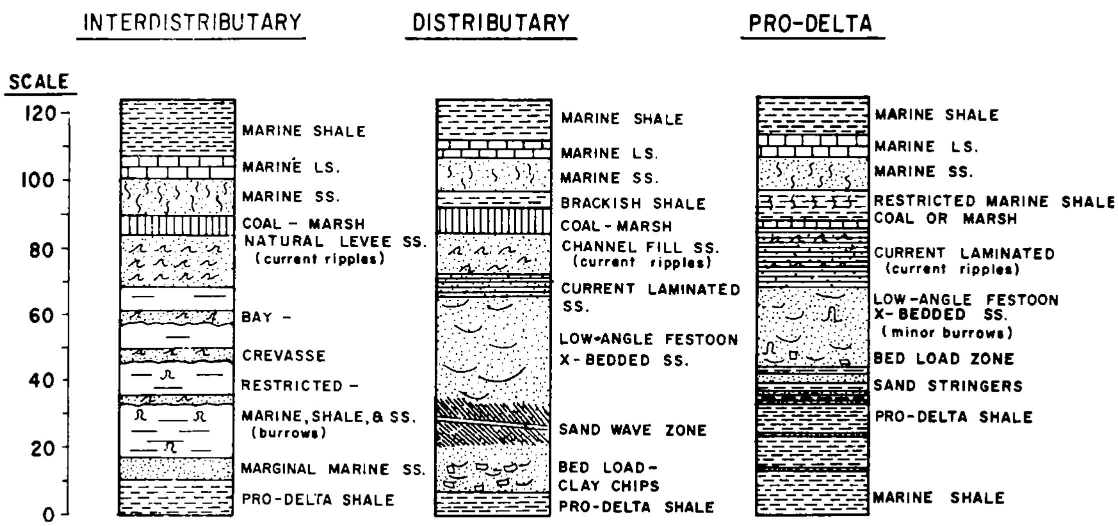


Figure 10. Genetic sequences of lithologic units. These patterns characterize process-response sequences within individual areas of delta.

interbedded with thin siltstones and shales, have some minor bioturbation, and also show considerable scour into the underlying prodelta-type sandstone and shale. In unusually thick sandstone units the transition upward into marine sandstone and limestone may be very abrupt, without the deposition or preservation of coal or marsh units.

The prodelta sequence grades laterally into the distributary sequence and also into the marginal depositional plain. It may show either gradual or abrupt transition from marine sedimentary units into current-deposited clastics. The sequence represents a slow shallowing of water during deposition and decreasing marine influence. This entire sequence may be removed locally or replaced by the distributary sequence. Bioturbation may be associated with current features at the base of the sandstone.

Each of these three units can be recognized in the cores and outcrop sections, and commonly are suggested by the electric-log shapes (Fig. 7). The Bluejacket delta probably is an intermediate type between highly elongate and tidal delta types (Fisher, 1969). It is primarily a sandstone sheet, but the marine units are not important except in the lower part of the delta. This type of delta appears to be typical of the Early Pennsylvanian shelf. Probably only in the distal parts are water depth, and wave and tidal energy, sufficient to strongly modify the depositional pattern.

OIL OCCURRENCE IN DELTAIC SANDSTONES

Sandstones of the lower Absaroka sequence are extremely important oil reservoirs. More than 2 billion bbl of oil has been produced from these sandstones. Most of the oil is in stratigraphic traps. It is of particular economic importance to understand the reasons for these occurrences.

Environmental Control

The shelf edge and marginal basin is the habitat of oil for many reasons. Rivers provide nutrients to the sea, and life grows in abundance in the waters surrounding deltas. Deltaic areas are commonly covered with a dense cover of vegetation, and organic material is one of the most characteristic detrital constituents of deltaic sandstones and shales. In addition, the rapid rates of sedimentation related to deltas produce depocenters where rapid burial of organic materials prevents oxidation and loss of carbon compounds to the sea or atmosphere. Deltaic areas contain a varied complex of depositional environments. They are characterized by abrupt vertical and lateral variations in sedimentary facies, which provide excellent stratigraphic traps. Also, the association of channel and wave processes produces winnowing and the formation of relatively permeable sandstone reservoirs. Another aspect favorable for oil accumulation in deltas is the tectonic and structural instability that commonly is associated with high depositional rates. Growth faults, mud diapirs, and other early sedimentary adjustments may provide traps. A final factor may be associated with clay deposition. Poorly crystallized aluminosilicates are eroded off the surface of the continent, and these "clays" are transported to the delta. Their degraded character makes them scavengers for organic colloids and other charged particles. These "clays" are deposited with the sands in the marine environment. Shortly after burial the "clays" recrystallize, increase in grain size, and form clay minerals. They do not contain as much water or organic complexes as they did when deposited. Water loss from the shales may provide the mechanism for transporting "organic complexes" into the reservoirs. These source beds connected to reservoirs provide the hydrocarbons to fill the porous and permeable deltaic sandstones.

Bartlesville Sandstone Oil Fields

The production map from the Bartlesville sandstone explains why it has become so well known (Fig. 11). Probably 1.5 billion bbl of oil has been removed from this single sandstone reservoir (Weirich, 1968). Prospecting for oil is still proceeding in some areas, and commercial reserves are being discovered. The pattern of petroleum accumulation has been dictated by the permeability trends and later structural tilting of the area. If some idea of the tectonic and depositional patterns had been known when oil was first discovered in this sandstone, exploration could have proceeded in a systematic manner. The drilling of thousands of dry holes, the development on 5-acre spacing, and other wasteful methods could have been prevented.

CONCLUSIONS

It has been shown that the history of the Early and Middle Pennsylvanian of eastern Oklahoma was one of delta progradation. The theme is one of transgression coupled with alternating regressive periods. Once the framework and environmental control were set, the pattern was repeated again and again. The depositional cycles of coal, thin limestone, thick sandstone, and marine shale were repeated in a systematic manner across the shelf.

Sedimentation was nearly continuous, but depositional rates varied widely depending on time and position on the shelf. Individual limestones may represent as much time as the progradation of a delta lobe. Erosional scour surfaces and disconformities were local and did not affect the overall sedimentation history.

It is possible to determine "ideal" cycles or patterns within the deltaic sequence. These may provide the basis for interpreting the origin of individual stratigraphic units.

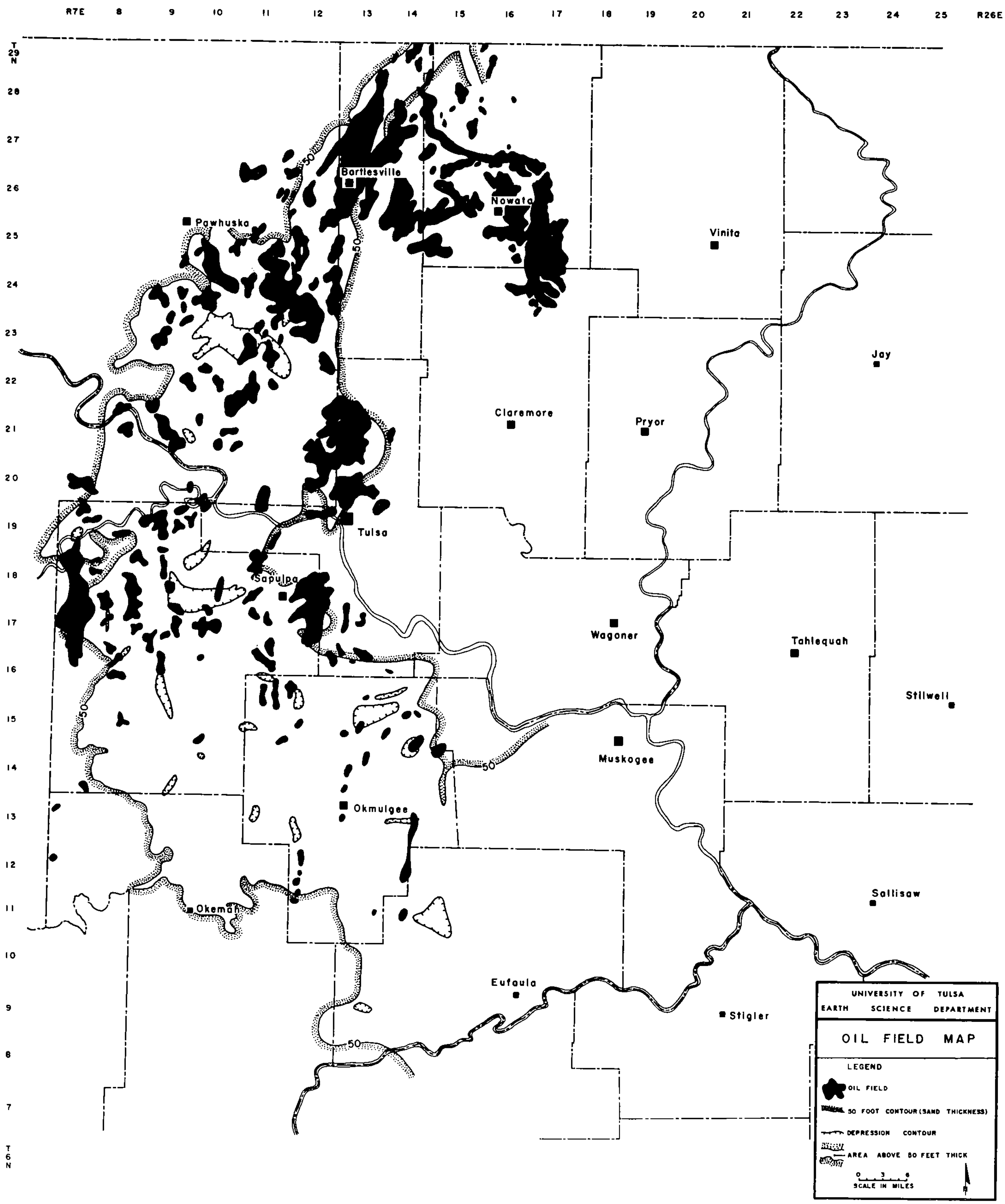


Figure 11. Distribution of oil fields within Bartlesville sandstone. Pattern shows concentration of oil on updip pinchout of sandstone. Combination of structural dip and sandstone distribution is the basis of individual accumulations.

THE GEOLOGY OF THE TI VALLEY FAULT IN THE OKLAHOMA OUACHITA MOUNTAINS

Neil H. Suneson

INTRODUCTION

The Ti Valley fault is generally believed to be the largest thrust fault in the Ouachita Mountains frontal belt in southeastern Oklahoma. Although the Choctaw fault forms the geomorphic northern boundary of the Ouachita Mountains and is the leading thrust fault of the Ouachita fold and thrust belt at its present level of exposure, the Ti Valley fault has a larger displacement and can be thought of as being the geologic northern boundary of the Ouachita Mountains, for reasons described below. The trace of the Ti Valley fault is ~240 mi long, extending from near Atoka, Oklahoma, to near Jacksonville, Arkansas. (In Arkansas, parts of the Ti Valley fault are called the Y City fault.) Hendricks (1959, p. 50) estimated the "minimum displacement" on the fault to be 20 mi. The Ti Valley fault is covered by the Cretaceous Antlers Formation near Atoka and by Quaternary sediments of the Mississippi Embayment in Arkansas. The fault possibly extends far to the southwest and east in the subsurface (Flawn and others, 1961, pls. 2,3).

The Ti Valley fault was first named by Powers (1928), and the significance of the fault as a boundary between different geologic provinces was first recognized by Ulrich (1927). Perhaps the most poetic description of the fault can be ascribed to Cline (1960, p. 16), who likened it to "a sort of Chinese wall for Ouachita stratigraphers." Ulrich (1927, p. 26) suggested that "a great overthrust fault . . . more or less indefinitely in the valley north of the Winding Stair Mountain" (near the village of Ti as shown on his fig. 1, p. 13) formed the northern edge of the Ouachita depositional basin, represented by Johns Valley, Jackfork, and Stanley strata. Miser (1929) emphasized the importance of the Ti Valley fault by recognizing that strata of the Arbuckle Mountains facies are present north of the fault and strata of the Ouachita Mountains facies are present to the south. Ham (1959, p. 73) summarized the pre-Stanley correlation of Arbuckle-facies and Ouachita-facies strata, and Hendricks and others (1947) listed the post-Stanley correlations (Fig. 1). Miser (1929) suggested that the Woodford chert, Caney shale, and Atoka Formation are the only strata common to both facies; subsequent workers mapped Devonian strata south of the Ti Valley fault as Arkansas Novaculite and have shown that the large masses of Caney shale south of the fault are probably allochthonous slide masses within the Morrowan Johns Valley Formation. Devonian and Mississippian strata north of the fault typically are interpreted as relatively thin slices of Woodford chert and Caney shale at the bases of thrust sheets. The strata are "stratigraphically autochthonous" (i.e., they are not large blocks in younger mud slides), but "structurally allochthonous" (i.e., they are within thrust sheets). The only unit common to both sides of the Ti Valley fault is the Atoka Formation. Hendricks and others (1947) showed pre-Atoka strata immediately north of the Ti Valley fault to include, in descending stratigraphic order, Pennsylvanian Springer Formation, Mississippian Caney shale, Devonian(?) Woodford chert, and Devonian Pinetop chert and an unnamed limestone. Marcher and Bergman (1983) remapped these strata as follows: Morrowan Limestone Gap shale and Chesterian Goddard shale (both mapped as Springer by Hendricks and others, 1947), Mississippian Delaware Creek shale (Caney), Kinderhookian and Upper Devonian Woodford shale, and Lower Devonian Pinetop chert. South of the Ti Valley fault, the strata are: Atoka Formation, Morrowan Johns Valley Formation, Morrowan Jackfork Group, and Mississippian Stanley Group.

It is clear that the geology and location of the Ti Valley fault are related to Mississippian and Morrowan stratigraphic problems, many of which have not been resolved. For example, field recognition of the Johns Valley Formation is difficult because boulder-bearing beds are present throughout the Morrowan Jackfork Group (e.g., Hendricks and others, 1947). A stratigraphic problem north of the Ti Valley fault is the recognition and correlation of Caney, Springer, and basal Atoka shales throughout the different thrust sheets. Correlation of different formations and recognition of facies changes within each formation across the Ti Valley fault and thrust faults north of it are, as yet, unresolved. Measurement of displacement on the Ti Valley fault and other thrust faults in the Ouachita Mountains frontal belt depends, in large part, on the resolution of these stratigraphic problems.

REGIONAL GEOLOGIC SETTING

The exposed part of the Ouachita Mountains in southeastern Oklahoma and west-central Arkansas constitutes the eroded remnants of a large fold and thrust belt that originally extended from western Alabama (e.g., Thomas, 1985) to northeastern Mex-

	SERIES		ARBUCKLE MOUNTAINS	OUACHITA MOUNTAINS
PENN.	Atokan		Atoka Fm.	Atoka Formation
PENN.	Morrowan		Wapanucka Ls.	Johns Valley Shale
PENN.	Morrowan		Springer Fm.	Jackfork Group
MISSISSIPPIAN	Chesterian		Sycamore Ls. — Caney Sh.	Stanley Shale
MISSISSIPPIAN	Meramecian		Sycamore Ls. — Caney Sh.	Stanley Shale
MISSISSIPPIAN	Osagean		Welden Ls.	Arkansas Novaculite
MISSISSIPPIAN	Kinderhookian		Welden Ls.	Arkansas Novaculite
DEVONIAN	Upper		Woodford Sh.	Arkansas Novaculite
DEVONIAN	Lower	Hunton Gp.	Frisco Ls. Bois d'Arc Ls. Haragan Ls.	Arkansas Novaculite Pinetop Chert
SILURIAN	Upper	Hunton Gp.	Henryhouse Fm.	Missouri Mountain Shale
SILURIAN	Lower	Hunton Gp.	Chimneyhill Subgroup	Blaylock Sandstone
ORDOVICIAN	Upper		Sylvan Sh.	Polk Creek Shale
ORDOVICIAN	Upper	Viola Gp.	Welling Fm. Viola Springs Fm.	Bigfork Chert
ORDOVICIAN	Middle	Simpson Gp.	Bromide Fm. Tulip Creek Fm. McLish Fm. Oil Creek Fm.	Womble Shale
ORDOVICIAN	Middle	Simpson Gp.	Joins Fm.	Blakely Sandstone
ORDOVICIAN	Lower	Arbuckle Gp.	West Spring Creek Fm. Kindblade Fm.	Mazarn Shale
ORDOVICIAN	Lower	Arbuckle Gp.	Cool Creek Fm. McKenzie Hill Fm.	Crystal Mountain Ss.
ORDOVICIAN	Lower	Arbuckle Gp.	Butterly Dol.	Collier Shale
CAMBRIAN	Upper	Arbuckle Gp.	Signal Mountain Ls. Royer Dol. Fort Sill Ls.	Collier Shale ? ——— ?
CAMBRIAN	Upper	Timbered Hills Gp.	Honey Creek Ls. Reagan Ss.	

Figure 1. Correlation of Arbuckle- and Ouachita-facies strata. Modified from Hendricks and others (1947), Ham (1959), and Mankin (1987).

ico (Flawn and others, 1961, p. 99). The tectonic belt marks the site of a long-lived depositional basin called the Ouachita trough. Throughout the early and middle Paleozoic, 3,800–6,300 ft (Ham, 1959) of deep-ocean sediments (mostly dark shale and chert, with minor limestone and coarse clastics) accumulated in the trough. These deep-water sediments are now exposed in structurally allochthonous positions at Black Knob Ridge near Atoka, Oklahoma, in the Potato Hills near Talihina, Oklahoma, and in the Broken Bow–Benton uplift between Broken Bow, Oklahoma, and Benton, Arkansas (Fig. 2). In contrast, 6,800–11,500 ft of equivalent, relatively shallow-water carbonates, sandstone, and shale

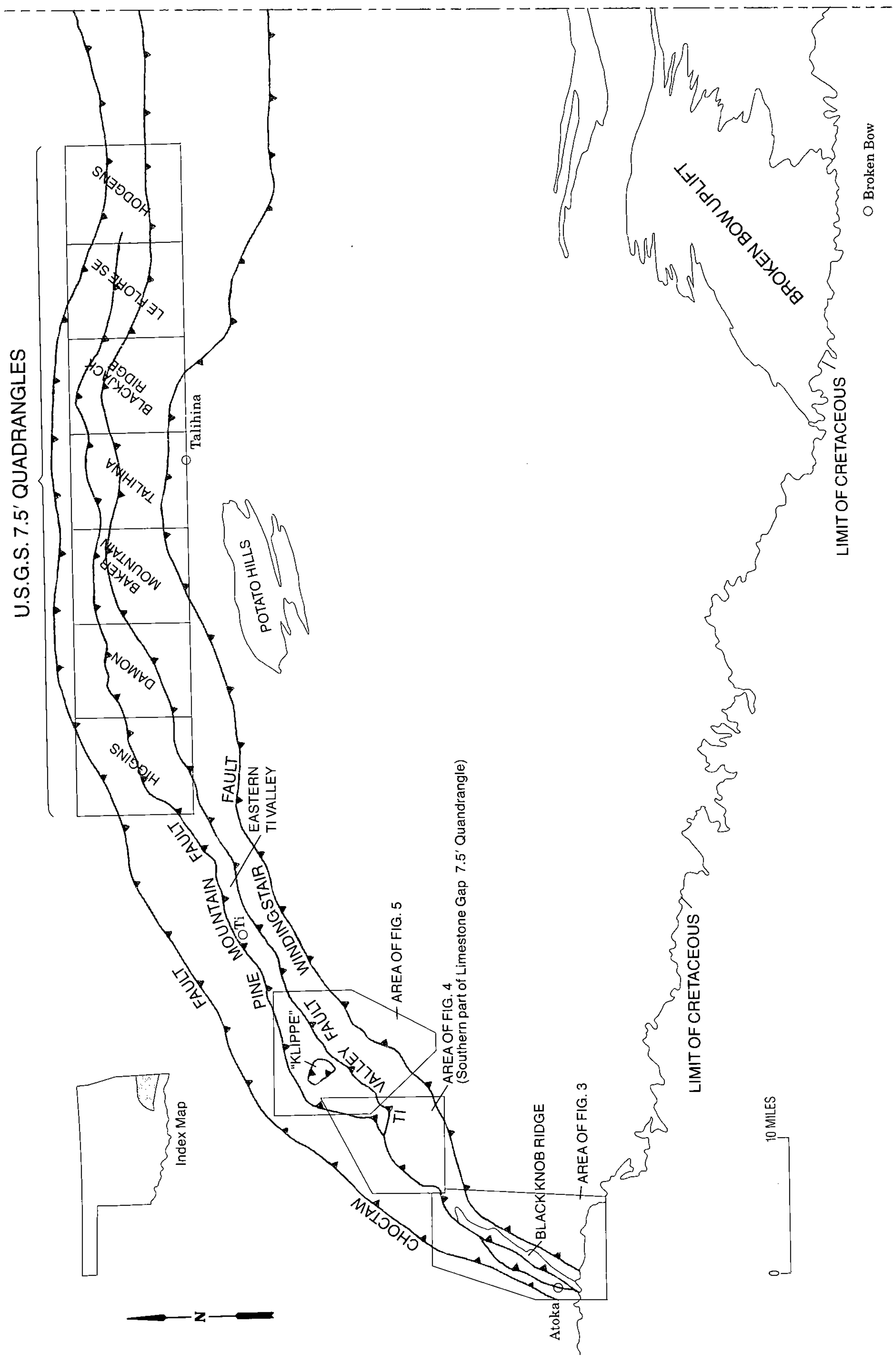


Figure 2. Major faults in the Ouachita Mountains frontal belt (from Miser, 1954), and other features referred to in text.

accumulated in shelf areas west and north of the present Ouachita Mountains. These strata of the "Arbuckle Mountains facies" extend beneath the Ouachita Mountains frontal belt (e.g., see discussion of Exxon No. 1 Retherford well, below), and they may extend beneath the Benton uplift in Arkansas (Lillie and others, 1983, fig. 5B) or the Broken Bow uplift in Oklahoma (Leander and Legg, 1988).

In Early Mississippian time, the sedimentary history of the Ouachita trough changed abruptly, and 25,000–43,500 ft of flysch accumulated in an approximately E–W-trending basin (Morris, 1974b; Briggs and Roeder, 1975) from the Early Mississippian to Early Pennsylvanian. The axis of the depositional basin migrated northward with time (Briggs and Roeder, 1975, p. 7), so that the youngest flysch unit (Atoka Formation) reaches a maximum thickness in the northern part of the Ouachita tectonic belt just south of the Arkoma basin. Tectonic shortening of the flysch basin probably began in the Early Mississippian, peaked in the Early Pennsylvanian, and lasted to the Early Permian, forming a classic fold-and-thrust belt.

In Oklahoma, the Ouachita tectonic province is divided structurally into three belts. From north to south, these are the frontal belt, consisting of closely spaced thrust faults and relatively tight folds; the central belt, consisting of widely spaced thrust faults and mostly broad, open synclines; and the Broken Bow uplift, consisting of Ordovician through Lower Mississippian strata deformed into isoclinal and isoclinal recumbent folds. The Windingstair fault forms the boundary between the frontal and central belts, and the outcrop pattern of lower and middle Paleozoic strata approximates the area of the Broken Bow uplift (Fig. 2). Strata within the frontal belt are mostly flysch of the Atoka Formation. In the northern part of the frontal belt, sub-Atoka shelf deposits of the Springer shale and Wapanucka limestone typically mark the base of as many as five individual imbricate thrust sheets. In the southern part of the frontal belt, equivalent strata (Jackfork Group, Johns Valley Formation) are more basinal and typically mark the bases of imbricate thrust sheets at the present level of exposure. Therefore, within the now tightly compressed frontal belt, the Morrowan shelf-to-basin transition is preserved, and a palinspastic restoration of the frontal belt should show the configuration and dimensions of the Morrowan flysch basin. Hendricks and others (1947) mapped a large part of the western Ouachita Mountains frontal belt, and Hendricks (1959, p. 50) gave minimum displacements on the faults of the frontal belt; he suggested that the Ti Valley fault has the greatest displacement in the frontal belt, based partly on facies changes across the fault and a klippe.

PURPOSE AND DEFINITIONS

Since Hendricks's (1959) classic paper on the structural geology of the Ouachita Mountains frontal belt, information from new geologic maps and exploratory hydrocarbon wells requires that some of the conclusions reached in that paper be reexamined. This paper reviews some of the features of the Ti Valley fault as it was mapped in 1947 by Hendricks, Gardner, Knechtel, and Averitt, and subsequently by others, in the western part of the Ouachita Mountains. This paper also brings to bear some new well information, and some previously published but uninterpreted or forgotten surface information related to the geology of the fault. To the east, new mapping by Suneson and Ferguson (unpublished), Fellows (1964), and Hart (1963) is also used to reevaluate certain prevailing assumptions regarding the geology of the Ti Valley fault. Those features of the fault discussed by Hendricks (1959) but requiring reexamination are the dip of the fault, the amount of movement on the fault, and the glide plane(s) for the fault. In addition, I will describe footwall and hanging-wall structures.

Throughout this paper, the term *thrust fault* is used instead of reverse fault, despite my recognition that the dips of many, if not most, of the faults in the frontal Ouachita Mountains are >45°. The measurable or inferred angle of dip is largely a function of the present level of exposure. I believe, as do many other workers in the Ouachita Mountains, that the dip of most of the faults exposed at the surface decreases downward in the subsurface, and most faults probably merge with one or more subhorizontal sole thrusts. I prefer the term *thrust fault* because the Ouachita Mountains, particularly the frontal belt, are part of what has been classically described as a large fold-and-thrust mountain range extending (in the subsurface, in part) from the extreme southeastern United States to northern Mexico. The general map pattern of thrust faults in the frontal belt resembles an imbricate thrust system (terminology of Boyer and Elliot, 1982, p. 1198–1199), except locally where Hendricks and others (1947) map crosscutting thrust faults. In these rare cases, duplexes may be present.

It should also be noted that I use the terms *formation* and *group* informally, reflecting common local usage.

I describe the geology of the Ti Valley fault in Oklahoma, from the southwest part of the Ouachita Mountains east to the Arkansas state line. I divide the fault into segments based on the geology of the hanging wall, features associated with the fault itself, and the source of data on the fault (Fig. 2). The Black Knob Ridge area near Atoka, Oklahoma, is characterized by Ordovician rocks in the hanging wall of the fault. The Limestone Gap Quadrangle is discussed separately because the fault forms a large salient to the northwest and contains folds in the hanging wall. The fault in southwest Ti Valley has been the subject of controversy because Hendricks and others (1947) mapped a large klippe there, clearly indicating that the fault has a very low dip. In the eastern part of Ti Valley, Morrowan strata are in the hanging wall, and the fault trace is relatively straight. The Higgins, Damon, and Baker Mountain

Quadrangles are considered separately because a new, detailed surface map (scale 1:24,000; Suneson and Ferguson, unpublished) suggests there is no evidence that the facies changes across the Ti Valley fault is as abrupt as it may be to the southwest near Black Knob Ridge. In the Talihina Quadrangle, footwall strata locally are the Caney shale and Woodford chert. To the east, the Ti Valley fault juxtaposes Atoka against Atoka and much of the area has not been mapped in detail.

BLACK KNOB RIDGE AREA

Taff (1902) first mapped the geology of the Black Knob Ridge area and showed a major fault juxtaposing Carboniferous Atoka Formation on the west and Silurian Stringtown shale, Talihina chert, and Standley (original spelling) shale (oldest to youngest) on the east. He recognized this major fault as a thrust fault and named it the Choctaw fault. Hendricks and others (1937) mapped Black Knob Ridge at a scale of 1:24,000 and separated the hanging-wall strata exposed there into the Ordovician Womble shale and Bigfork chert, Silurian Polk Creek shale and Missouri Mountain shale, Devonian Arkansas Novaculite, and Pennsylvanian Stanley (new spelling) shale. They separated the footwall strata into the Mississippian Caney shale and Pennsylvanian Springer and Atoka Formations. For most of the length of Black Knob Ridge, the fault separating the two sequences was named the Ti Valley–Choctaw fault; according to their mapping, these faults diverge in the northern part of their map area, the trace of the Choctaw fault trending almost due N and the trace of the Ti Valley trending NE. Hendricks and others (1937, fig. 2) showed a block of undifferentiated Carboniferous strata between the Choctaw and Ti Valley faults, and suggested (p. 14) that they may include Stanley and Jackfork strata. Subsequently, Hendricks and others (1947) showed those strata to be Atoka Formation.

The most recent detailed geologic maps of the Ti Valley fault in the Black Knob Ridge area are those by Hendricks and others (1947) and Leonhardt (1983). Hendricks and others (1947) refined the structural geology of the hanging-wall and footwall strata. They recognized two major thrust faults in the footwall of the Ti Valley fault (Choctaw and Pine Mountain), two large synclines in Atoka strata west of the Ti Valley fault, two large anticline–syncline pairs and many small thrusts in the lower Paleozoic strata of Black Knob Ridge east of the fault, and the major Windingstair thrust fault in the Stanley shale east of Black Knob Ridge (Fig. 3). Leonhardt (1983) and Leonhardt and Aiken (1984) focused on both large-scale and small-scale structures in the lower Paleozoic strata. Leonhardt (1983) combined the Pine Mountain and Ti Valley faults into a single fault system separating Ordovician Womble shale from Pennsylvanian Atoka Formation, locally mapped considerably more Womble shale in the hanging wall of the Ti Valley fault than did Hendricks and others (1947), and suggested that the trace of the Ti Valley fault curves more (particularly near the anticline–syncline pairs on Black Knob Ridge) than previous mapping showed. In addition, Leonhardt (1983) mapped a number of small faults with apparent left-lateral displacement along Black Knob Ridge; several of these were mapped as small thrust faults by Hendricks and others (1947).

The Womble shale forms the glide plane for the Ti Valley fault throughout most of its length west of Black Knob Ridge. Comparison of mapped attitudes and map patterns suggests that the fault is generally parallel to bedding in both footwall and hanging-wall strata except for those areas near the anticline–syncline pairs, where the Womble is probably complexly folded into the cores of the folds. The Bigfork chert conformably overlies the Womble (Hendricks and others, 1937, p. 6); as a result, attitudes of the Bigfork probably approximate those in the poorly exposed Womble. The Bigfork chert on Black Knob Ridge generally dips SE at angles ranging from 40° to vertical and averaging ~63°. West of the fault trace, the Atoka Formation is poorly exposed, but it dips SE at an angle of ~70°. Assuming that the Ti Valley fault is parallel to bedding, it probably dips SE at ~65° beneath Black Knob Ridge at the present level of exposure. Taff (1902) showed the fault to dip SE at ~60°, and Hendricks and others (1947) showed it to dip SE at ~75° near Stringtown.

Two "deep" exploration wells have been drilled into the hanging wall of the Ti Valley fault near Black Knob Ridge, and a third well has been drilled farther southeast (Fig. 3). The Putman No. 1 Culbertson (NE¼NE¼SW¼SW¼ sec. 6, T. 2 S., R. 12 E.) was spudded in November 1972 near the Womble–Bigfork contact on the axis of the southern anticline in Black Knob Ridge strata. It was drilled to 5,210 ft and abandoned as a dry hole in December 1972. The Texinia No. 1 Lucy (NW¼NE¼SW¼ sec. 8, T. 2 S., R. 12 E.), ~1.5 mi east of Black Knob Ridge and southeast of the southern anticline–syncline pair, was spudded by Texaco in the Stanley Group immediately east of the poorly exposed Windingstair fault on 31 December 1984. Drilling was completed on 10 June 1985 at 16,855 ft T.D. The well was plugged back to 12,080 ft and turned over to Texinia Corp. in September 1985. The well was completed as a producer in the Morrowan Cromwell sandstone, a subsurface unit in the Springer. The third test is the TXO No. 1 Anson (S½NW¼ sec. 35, T. 2 S., R. 12 E.), 6 mi southeast of the trace of the Ti Valley fault and 7 mi southeast of the town of Atoka. The well is east of the Windingstair fault trace, on the southeast flank of the broad Round Prairie syncline. It was spudded in Cretaceous Antlers Formation, which in that area forms a thin cover on uppermost Stanley Group. Drilling was started on 3 April 1981 and finished on 6 June 1981 at 8,800 ft T.D. The well was completed as a gas producer in the Arkansas Novaculite.

No data have been released on the No. 1 Culbertson, and very little information on the No. 1 Lucy is available. Reports filed with the Oklahoma Corporation Commission state that the No. 1 Lucy (G.L. 562 ft) drilled Arbuckle-facies strata, clearly indicating that the well penetrated the Ti Valley fault. The fault is shallower than 11,653 ft, the top of the perforated Cromwell interval. Some data are available for the No. 1 Anson; formation tops in this well (D.F. 605) are Arkansas Novaculite 6,822 ft; Missouri Mountain shale 7,521 ft; Bigfork chert 7,612 ft; Womble shale 8,507 ft.

The dip of strata in the No. 1 Anson averages ~20° E, although the strike varies roughly from WNW to NNE. The middle and lower Paleozoic strata strike roughly N and dip 15–20° E. If a fault thrusts Womble over younger strata just below the bottom of the well, it would project westward to the mapped trace of the Windingstair fault, assuming that it is parallel to bedding. This suggests that (1) the well drilled hanging-wall strata of the Windingstair fault, (2) the Windingstair fault is a low-angle fault (20°) between its trace and the well, and (3) the Ti Valley–Windingstair branch line is east of the well. Alternatively, the well may have drilled the hanging wall of a listric thrust fault east of the Windingstair or a fold in the hanging wall of a subhorizontal Windingstair fault; such a fold would be similar to the folds in the hanging wall of the Ti Valley fault along Black Knob Ridge.

Hanging-wall and footwall structures on either side of the Ti Valley fault are different. The average strike of the Ti Valley fault near Black Knob Ridge is about N. 30 E. Two large-scale folds in the footwall trend N. 35–45° E. The large-scale folds along Black Knob Ridge trend N. 57–60° E. (Leonhardt, 1983). Faults in the footwall are thrust faults that approximately parallel the trace of the Ti Valley fault. The major hanging-wall thrust (Windingstair) parallels the Ti Valley fault, but numerous small faults that strike approximately E–W offset strata of Black Knob Ridge. Hendricks and others (1947) mapped these small faults as thrusts, whereas Leonhardt (1983) mapped them as left-lateral strike-slip faults.

About 2 mi northeast of Stringtown, a small block of hanging-wall strata strikes roughly N–S, and the Ti Valley fault juxtaposes footwall Atoka against Womble on the southwest to Stanley on the northeast (Fig. 3).

LIMESTONE GAP QUADRANGLE

Hendricks and others (1947) mapped the geology of the Limestone Gap 7.5′ Quadrangle (Fig. 4). There, the Ti Valley fault bends north and presumably crosscuts the NE-striking Pine Mountain fault; curves eastward and then southeastward, uncovering the Pine Mountain fault; and then continues with a NE strike (Fig. 2). The map pattern resembles a large salient in the trace of the Ti Valley fault. Footwall strata are entirely within the Atoka Formation, although the exact position within the Atoka Formation is unknown. Hanging-wall strata adjacent to this fault range from Stanley to Atoka.

Except at the edges of the fault salient where the strike of hanging-wall strata is oblique to the trace of the Ti Valley fault, the fault appears to parallel bedding in the hanging wall. The dip of hanging-wall strata (Atoka to the southwest, Stanley to the northeast) averages ~70° SE; if the Ti Valley fault is a bedding-plane thrust, it probably dips SE at ~70°—about the same as the dip of the fault near Black Knob Ridge. At the ends of the salient, where the fault crosscuts steeply dipping hanging-wall strata, the map pattern suggests that the fault has a dip <70°. Alternative explanations to the shallow dip are (1) that the fault is offset by two unmapped cross-faults with a significant component of strike-slip displacement, or (2) that the fault is folded. Hendricks and others (1947) mapped the margin of the salient as the Ti Valley fault, presumably because Stanley shale is present along the northeast edge of the hanging wall. Except for a thin fault sliver of Johns Valley beneath Atoka along the southeast margin of the salient, there is no Johns Valley in the salient; in fact, along the northwest flank of the syncline within the eastern part of the salient, Johns Valley is absent and Atoka directly overlies Jackfork (Fig. 4).

No oil-and-gas tests have been drilled southeast of the Ti Valley fault near the salient within the Limestone Gap Quadrangle.

Structures north of the trace of the Ti Valley fault mapped by Hendricks and others (1947) consist entirely of thrust faults subparallel to the Ti Valley fault. Structures within the salient south of the fault consist of folds parallel to the overall trace of the fault but offset by the fault at the southwest and northeast ends of the thrust salient. A thrust fault parallel to the overall trend of the fault (NE) is present to the southeast and could be considered to form the southeast margin of the salient (Fig. 4). Hendricks and others (1947) probably did not believe that this is the Ti Valley fault, because they mapped Stanley shale to the north and Johns Valley shale immediately northwest of its trace. It is possible that the fault mapped as Ti Valley is a rejoining splay (terminology of Boyer and Elliot, 1982) of this presumably secondary fault to the southeast.

"KLIPPE" AREA

The first detailed map of the Ti Valley fault in the southwest end of Ti Valley (Hendricks and others, 1947) showed a klippe of the Ti Valley fault extending nearly 2.5 mi northwest of the main trace of the fault (Fig. 5). Hendricks and others (1947) showed the west and north sides of the klippe to be clearly defined, whereas the east and south sides were inferred. Misch and Oles (1957) suggested that the Ti Valley "klippe" consists of a zone of cross-faults on its western side, and that beds within the "klippe" could be traced continuously eastward into rocks of the underlying Pine Mountain thrust sheet. In the apparently bitter public dispute that followed, Hendricks (1958, p. 2759–2760) defended his mapping of the klippe, but conceded that the southeast part of the klippe was poorly defined and that the klippe could

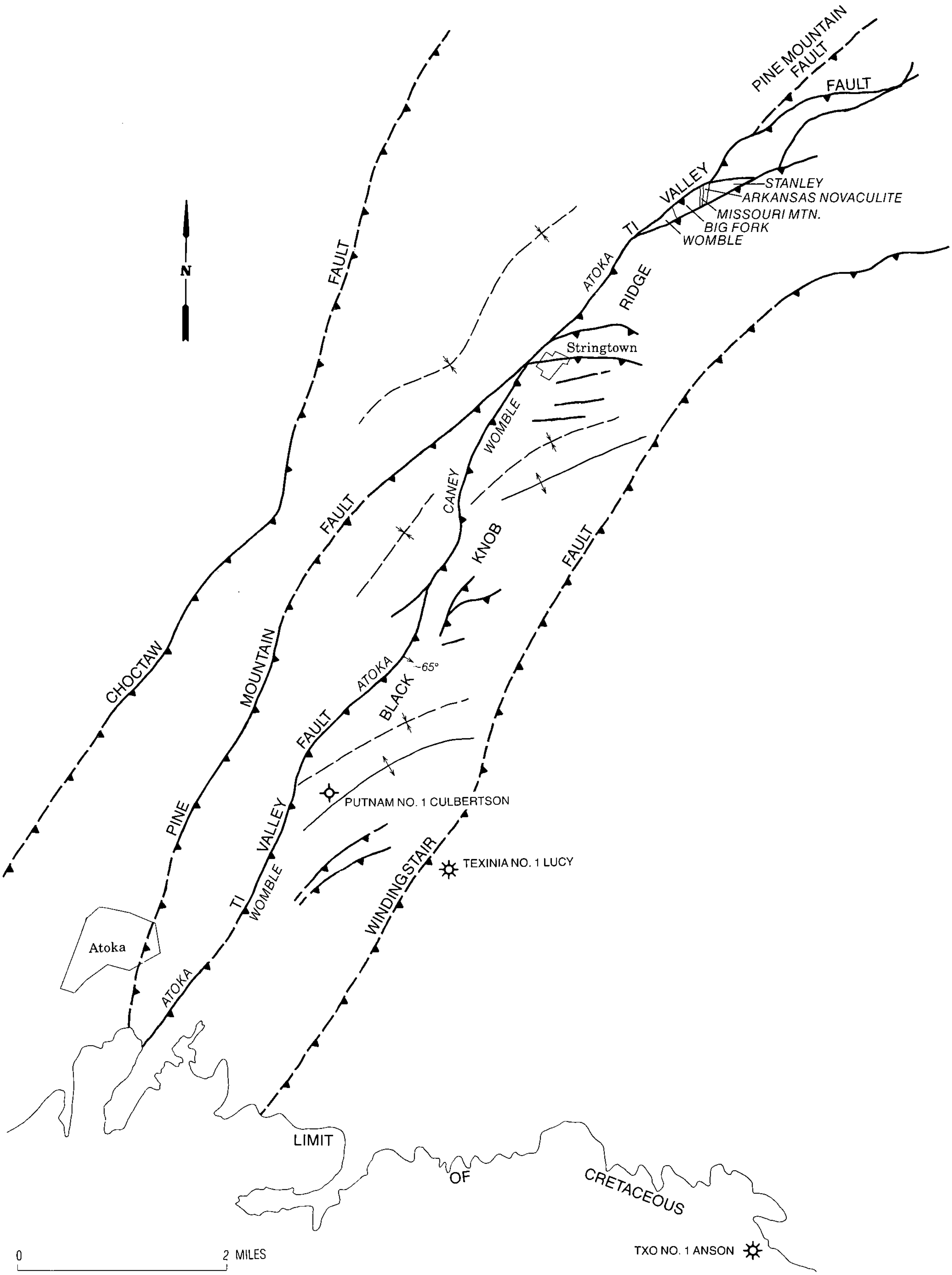


Figure 3. Features in Black Knob Ridge area. Geology from Hendricks and others, 1947.

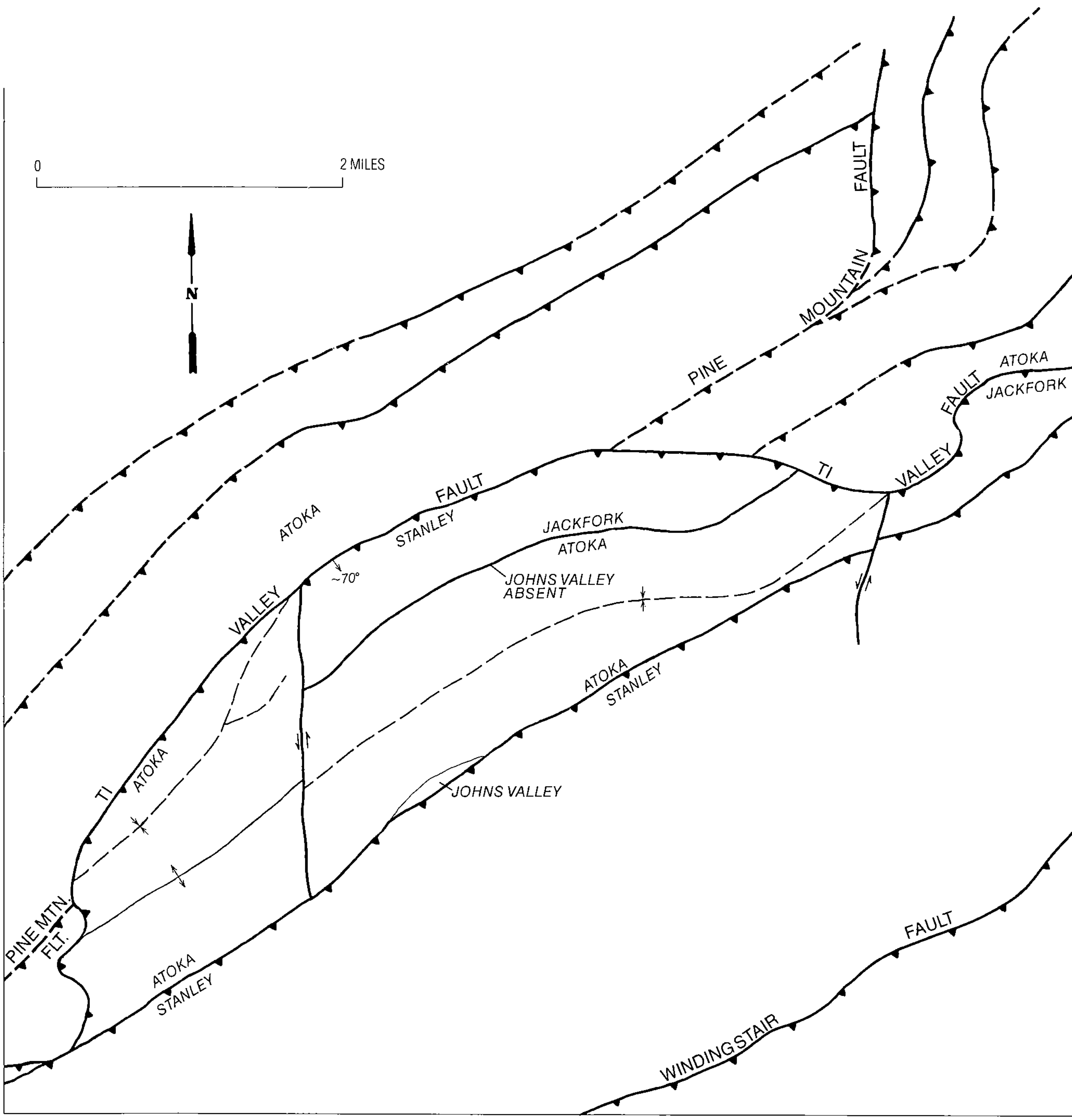


Figure 4. Features near the Ti Valley fault, Limestone Gap 7.5′ Quadrangle. Geology from Hendricks and others (1947).

actually be a large, NW-trending salient in the fault. Misch and Oles (1958, p. 2776) defended their original criticisms, based partly on their mapping of continuous beds across the northeast margin of the klippe. They also concluded that a well, the Southwest Exploration No. 1 Hoehman (NE¼SE¼ sec. 16, T. 2 N., R. 14 E.; Fig. 5) confirmed their structural interpretations. Fay (1984) mapped the distribution of units near the klippe in much the same manner as Hendricks and others (1947), but mapped the fault as a salient over strata of the Pine Mountain sheet. The most recent map of the Ti Valley klippe area is by Hardie (1986), who largely agreed with the distribution of rock types as shown by Hendricks and others (1947), but had a significantly different interpretation (Hardie, 1986,1987). Mapping by Hendricks and others (1947) and Hardie (1986) showed Jackfork Group, Johns Valley Formation, and Atoka Forma-

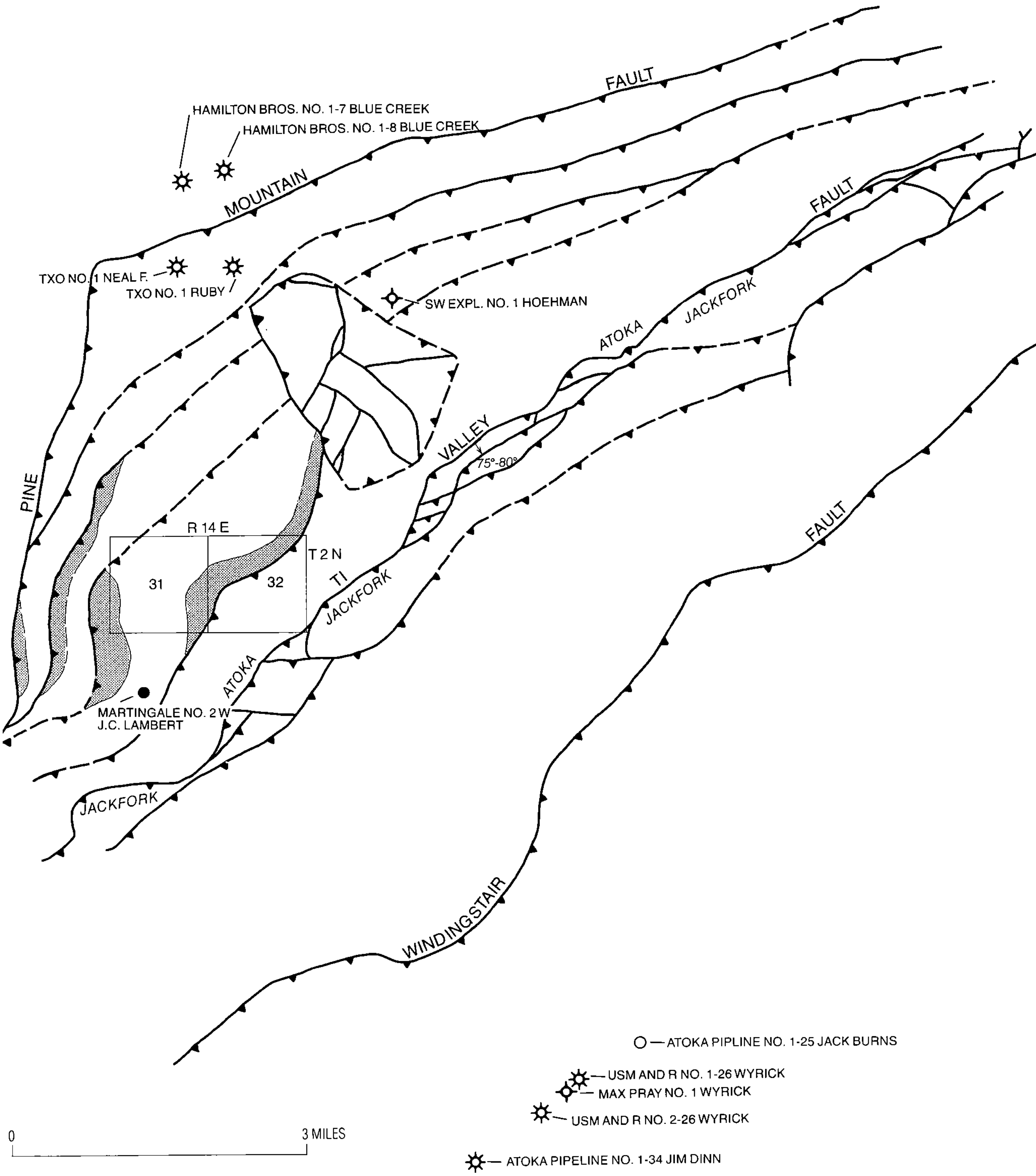


Figure 5. Features in "klippe" area. Geology from Hendricks and others (1947). Pattern represents Springer, Caney, and/or Woodford outcrops.

tion complexly faulted and folded in the Ti Valley klippe, whereas structural relations are relatively straightforward immediately south of the main trace of the fault.

The public argument between Hendricks (1958) and Misch and Oles (1958) regarding the possible existence of the "klippe" was based on interpretations of the local structural geology. Neither mentioned the stratigraphic evidence that basinal Stanley, Jackfork, and Johns Valley strata were mapped north of the main trace of the Ti Valley fault, except for the implication by Misch and Oles (1958) that those strata were misidentified in the original mapping by Hendricks and others (1947). The presence of any basinal strata supports the existence of a klippe or salient. However, the importance of the Johns Valley is questionable, considering Cline's (1960, p. 63) observation that limestone exotics are present (secs. 31 and 32, T. 2 N., R. 14 E.; Fig. 5) in strata mapped as Springer by Hendricks and others (1947). The implications of exotics in sub-Atoka strata north of the Ti Valley fault are more fully explored below in the section describing the fault in the Higgins, Damon, and Baker Mountain Quadrangles.

For much of its length in the "klippe" area, the Ti Valley fault trace approximately parallels the strike of footwall and hanging-wall strata, although the strike in some small hanging-wall blocks surrounded by faults is oblique to the trace of the fault. Footwall Atoka strata are generally vertical, but locally dip SE at angles as low as ~55°. The average dip of footwall strata is ~80° SE. The dip of hanging-wall strata varies widely, but averages ~75°. If the Ti Valley fault is parallel to bedding, as the map pattern suggests, its dip at the present level of exposure is about 75–80°. This is near or slightly greater than the dip farther southwest. Most other workers show the fault dipping relatively steeply SE: 25° (Hendricks and others, 1947, sheet 3), 60° (Rippee, 1981), 75° (Fay, 1984), and 65–80° (Hardie, 1986, pls. IV–VII). This suggests that if the "klippe" is, in fact, a klippe or salient, the fault changes dip significantly just above its present level of exposure and "ramps" down-section through footwall Atoka strata. (This observation—coupled with Misch and Oles's [1958] criticisms and Cline's [1960] mapping of limestone exotics north of the Ti Valley fault, approximately on strike with the "klippe"—casts considerable doubt on Hendricks and others' [1947] interpretation of a klippe or salient in this area.)

The Ti Valley fault and the "klippe" virtually mark the southern edge of the large Pittsburg–Pittsburg South–Blanco South gas field. Two wells ~1 mi northwest of the northern end of the Ti Valley klippe drilled an autochthonous subthrust sequence of Arbuckle-facies strata. The Hamilton Brothers 1-7 Blue Creek (S½S½S½NE¼ sec. 7, T. 2 N., R. 14 E.) (D.F. 1,187 ft) recorded the following tops: Wapanucka 3,776 ft; Wapanucka repeat 5,502 ft; Wapanucka 10,476 ft; Cromwell 11,397 ft; Caney 12,126 ft; Sycamore 12,461 ft; Woodford 12,565 ft; Sylvan 12,738 ft; Viola 12,784 ft; Bromide 12,942 ft; McLish 13,482 ft. The Hamilton Brothers 1-8 Blue Creek (CNW¼ sec. 8, T. 2 N., R. 14 E.; discovery well for the South Pittsburg field) was nearly identical. Hardie (1986, pl. II) showed two other wells (TXO No. 1 Ruby and TXO No. 1 Neal F; Fig. 5) ~0.5 mi northwest of the klippe penetrating sub-Choctaw-fault Wapanucka limestone. Arbuckle-facies strata (Woodford chert, Caney shale, Springer shale) are exposed at the surface southwest of the klippe in three splays (Fig. 5) associated with the Pine Mountain fault (Hendricks and others, 1947) and are present in the subsurface in the Martingale No. 2W J. C. Lambert (NW¼NE¼SW¼ sec. 6, T. 1 N., R. 14 E.; Fig. 5).

Misch and Oles (1958, p. 2776) claimed that the Southwest Exploration No. 1 Hoehman (G.L. 775 ft) confirms their structural interpretation that a klippe does not exist, but they did not present any data. The well was spudded 16 April 1954 in Atoka Formation just northeast of the klippe mapped by Hendricks and others (1947), within the salient mapped by Fay (1984), and within the trace of the gravity-slide block mapped and interpreted by Hardie (1986,1987); the well was drilled to 8,744 ft T.D. Scout tickets report Stanley from the surface to T.D. Scout tickets also show that dips from cores average ~39° between 5,773 and 6,614 ft, and ~25° between 8,551 and 8,663 ft. These dips are significantly less than those on the surface, but may deviate from true dips because of unknown hole deviation. My analysis of logs from this well indicates interbedded sandstone, siltstone, and shale to T.D., and the existence of a klippe or salient at shallow depths could not be confirmed or denied.

Other wells of importance to any new evaluation of the Ti Valley fault are the deep wells drilled immediately south of the West Daisy oil field in the southeast corner of T. 1 N., R. 14 E., ~2 mi southeast of the trace of the Windingstair fault and 5 mi southeast of the trace of the Ti Valley fault (Fig. 5). In order of spud date, these wells are Max Pray No. 1 Wyrick (drilled 6 July 1957 to 9 August 1958, 12,088 ft T.D.); U.S. Mineral and Royalty No. 1-26 Wyrick (drilled 24 May 1978 to 29 August 1978, 9,482 ft T.D.); U.S. Mineral and Royalty No. 2-26 Wyrick (drilled 31 January 1981 to 8 July 1981, 9,105 ft T.D.); Atoka Pipeline No. 1-25 Jack Burns (drilled 23 July 1985 to 5 November 1985, 11,702 ft T.D.); Atoka Pipeline No. 1-34 Jim Dinn (drilled 10 November 1986 to 25 April 1987, 12,456 ft T.D.).

Most of the wells penetrated below the Mississippian Stanley Group. From a geologic perspective, the most important well is the Max Pray No. 1 Wyrick (D.F. 764 ft). Branan and Jordan (1960) reported the following tops from this well: Arkansas Novaculite 8,335 ft; Polk Creek 8,400 ft; Big Fork cherty limestone 8,420 ft; Womble shale 9,098 ft (to T.D.). Scout tickets reported that the dip from two cores taken between 11,912 and 12,029 ft averaged ~68°, suggesting that the wells penetrated strata in the hanging wall of the Windingstair fault or a splay of the fault, and that the Ti Valley fault is considerably deeper than 11,500 ft beneath the West Daisy field. However, it appears that the glide plane for the Windingstair fault or its splay at the well location is the Womble shale, as it is for the Ti Valley fault at Black Knob Ridge.

Surface structures north of the main trace of the Ti Valley fault are a series of thrust faults subparallel to the trace of the fault. Surface structures south of the fault are mostly subparallel thrusts, but also include several small blocks bounded by faults both oblique to and parallel to the fault. There are no mapped folds in the footwall or hanging wall near this part of the Ti Valley fault.

EASTERN TI VALLEY

Hendricks and others (1947) mapped Ti Valley fault trace trending ENE across the southern part of Ti Valley (Fig. 2); they showed mostly Atoka Formation in the footwall, but locally mapped Woodford, Caney, and Springer above minor(?) thrusts in the footwall. Hanging-wall strata adjacent to the fault are mostly Atoka Formation; locally, thin outcrops of Johns Valley Formation were mapped. The mapping suggests that the glide plane for the Ti Valley fault is in the Johns Valley or a shale(?) in the lowermost Atoka. The hanging wall contains closely spaced imbricate thrust faults and consists of Jackfork, Johns Valley, and Atoka strata.

The strike of footwall strata generally parallels the trace of the Ti Valley fault, and dips are mostly vertical. Facing direction is unknown. The attitude of hanging-wall strata is similar to that in the footwall. The map pattern suggests that the Ti Valley fault is parallel to bedding in this area; therefore, the fault apparently dips somewhat more steeply here (80–90° SSE) than farther southwest.

No exploratory wells have been drilled in either the footwall or hanging wall of the Ti Valley fault near the east end of Ti Valley.

Hendricks and others (1947) did not map any folds immediately north or south of the trace of the Ti Valley fault in the eastern end of Ti Valley. Structures in the footwall of the fault consist entirely of thrust faults spaced about a mile apart at the present level of exposure. Structures in the hanging wall consist entirely of closely spaced, locally anastomosing thrust faults—many with very small apparent displacement. The thrusts in the hanging wall are spaced about 500–2,500 ft apart.

HIGGINS, DAMON, AND BAKER MOUNTAIN QUADRANGLES

The first detailed geologic map of the Ti Valley fault area in the area of the Higgins, Damon, and Baker Mountain Quadrangles was made by Hendricks and others (1947), who mapped the Ti Valley fault in the southwest corner of the Higgins Quadrangle (Fig. 6); they showed footwall Atoka Formation juxtaposed against Johns Valley strata overlain by Atoka in the hanging wall. East of the southwest corner of the Higgins Quadrangle, mapping by Hendricks and others (1947) was entirely within the hanging wall of the Ti Valley fault. Fellows (1964) mapped the southeast quarter of the Damon Quadrangle and most of the Baker Mountain Quadrangle (Fig. 2). He showed the Ti Valley fault juxtaposing Atoka Formation in the footwall against Atoka Formation in the hanging wall. His map and cross sections indicate that Johns Valley strata occur south of at least one, and locally two, thrust faults in the hanging wall of the Ti Valley fault. Bowsher and Johnson (1968) mapped the geology along Highway 2 in the middle of the Damon Quadrangle, locating the Ti Valley fault in the same place as Fellows (1964), and showing nearly vertical isoclinal folding in footwall and hanging-wall Atoka strata on both sides of the fault.

Recent detailed geologic mapping at a scale of 1:24,000 by Suneson and Ferguson (unpublished) has clarified relations on either side of the Ti Valley fault in the Higgins, Damon, and Baker Mountain 7.5′ Quadrangles. An olistostrome lithostratigraphically similar to the Johns Valley Formation is present ~2 mi north of where Hendricks and others (1947) mapped the Ti Valley fault in the Higgins Quad-

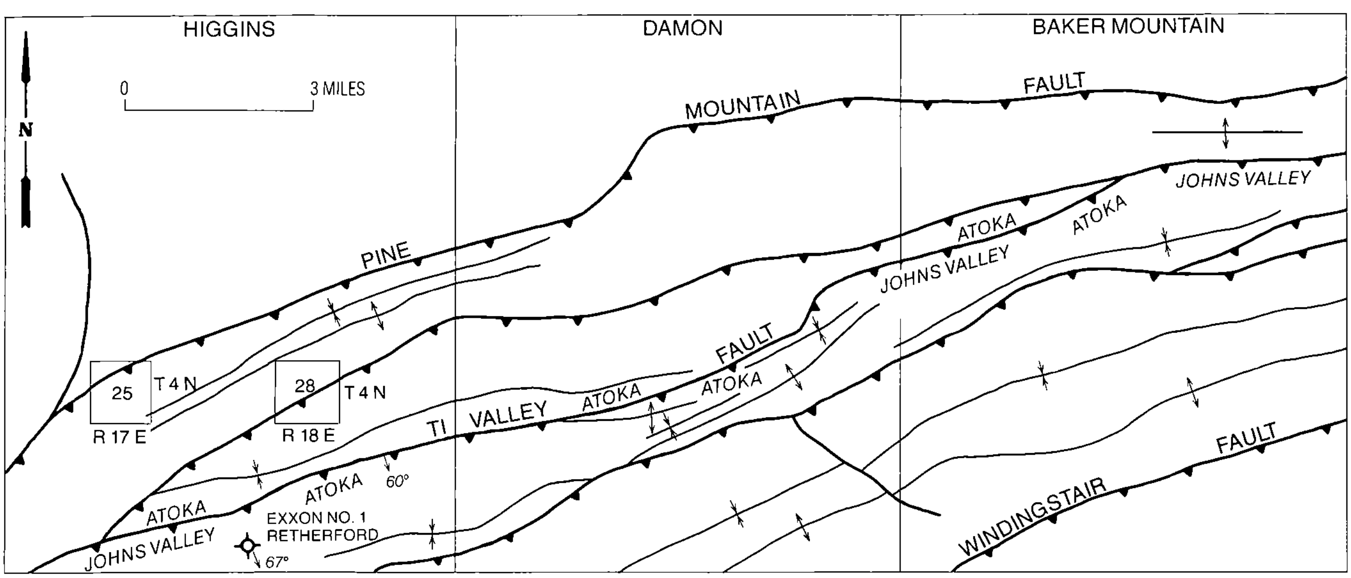


Figure 6. Features in Higgins, Damon, and Baker Mountain 7.5′ Quadrangles. Geology is Suneson's interpretation based on mapping by Suneson and Ferguson (unpublished).

rangle. This unit was briefly mentioned by Powers (1928, p. 1041), who described an outcrop of Caney shale containing "glacial" boulders (Power's quotation marks) in sec. 25, T. 4 N., R. 17 E. (Fig. 6). The olistostrome is "exposed" for ~1,000 ft along a north-flowing stream; it contains cobbles of chert and large blocks of a microcrystalline, platy yellowish-gray limestone (Suneson and Ferguson, unpublished guidebook). Robert C. Grayson, Jr., (personal communication) identified conodonts from a poorly exposed limestone near the base of the olistostrome as Middle or Late Ordovician. The exposed part of this olistostrome is small (about 1,000 by 1,500 ft), but the unit may be largely unexposed (large limestone blocks have been recognized only along the stream). The stratigraphic position of the olistostrome is near the base of the Atoka Formation. It may be similar in origin and position to the exotic-block-bearing "Springer" noted by Cline (1960) in the west end of Ti Valley (Fig. 5); however, the presence of Wapanucka limestone in normal stratigraphic position at 7,880 ft in the Exxon No. 1 Retherford well (Fig. 6; see below)—clearly in a hanging-wall position relative to the olistostrome—suggests that the olistostrome is not equivalent to the Springer, unless the Wapanucka was locally eroded at that location. (It should be noted that the olistostrome is located in the Pine Mountain sheet, and Hendricks and others (1947) report that the Wapanucka is absent throughout this sheet and Atoka overlies Springer.)

Two and one-half miles to the east, a large (about 1,000 by 800 ft) block of interbedded siliceous siltstone, quartzite, chert, and chert breccia is present in the SW¼ sec. 28, T. 4 N., R. 18 E. (Fig. 6). Contacts with surrounding strata are covered, but the chert does not appear to have the lenticular outcrop pattern of cherts mapped by Hendricks and others (1947) north of the Ti Valley fault. The chert is ~3,000 ft west of a small outcrop of micritic limestone and calcareous shale. Grayson (personal communication) suggests a Meremecian or early to middle Chesterian age for the limestone. In age and lithology, the outcrop is similar to outcrops of Caney shale elsewhere in the frontal belt.

The importance of these three outcrops relative to the Ti Valley fault is significant. If the olistostrome is within the Johns Valley Formation, and the Johns Valley Formation is restricted to the hanging wall of the Ti Valley fault, the fault must pass north of the outcrop; this is considerably farther north than where Hendricks and others (1947) mapped it. However, the chert appears to be more similar to the Pinetop than any other chert exposed in the frontal belt of the Ouachita Mountains. If the Pinetop is stratigraphically autochthonous, it "should" be north of the Ti Valley fault. The outcrop of Caney(?) may be north of the fault and stratigraphically autochthonous, or it may be an olistolith within the Johns Valley Formation, in which case it is south of the fault. Field relations clearly indicate that the fault cannot pass north of the olistostrome as a large salient and south of the chert. This suggests that the olistostrome is, in fact, north of the Ti Valley fault, and that it may represent the marginward equivalent of the largely basinal Johns Valley Formation. The outcrop patterns of the chert (Pinetop?) and Caney(?) suggest that they are also allochthonous and may represent slide material derived from a source terrane north of the Ouachita trough.

The presence of Johns Valley-like strata north of the Ti Valley fault suggests that using this "Ouachita facies" unit to locate the fault is unacceptable. Similarly, in the Higgins Quadrangle and to the east, the northernmost Jackfork strata (also "Ouachita facies") typically crop out in the hanging wall of thrust sheets south of the Ti Valley sheet. In the Baker Mountain and Talihina Quadrangles, Stanley Group strata are restricted to the Windingstair sheet. These observations suggest that the practice of naming different faults in the frontal Ouachita Mountains, particularly the Ti Valley, as has been done to the west, may not be appropriate in the Higgins Quadrangle and to the east, because the northernmost outcrops of Carboniferous Ouachita-facies strata appear to be distributed over a broad zone of thrust faults with progressively older strata exposed progressively southward. "Johns Valley-like" strata first crop out north of the traditionally mapped Ti Valley fault, Jackfork strata well south of the fault, and Stanley in the hanging wall of the Windingstair. In summary, at the present level of exposure, there is no evidence that the juxtaposition of Ouachita- and Arbuckle-facies strata occurs along one fault. In fact, the juxtaposition of contrasting facies appears to be distributed along several faults. To emphasize my reluctance to name individual faults—yet recognizing that the Ti Valley fault is firmly established in Ouachita literature—I will use the term in quotation marks ("Ti Valley fault") throughout the remainder of this section.

The average dip of strata near the "Ti Valley fault" is ~60°. The trace of the thrust fault in the three quadrangles generally parallels the strike of bedding, but in several places clearly crosscuts folds and is oblique to strike. If the fault generally parallels bedding, it probably dips at an angle of ~60° at the present level of exposure; this is significantly less than the dip inferred to the west. Fellows (1964, pl. I) showed the "Ti Valley fault" dipping 75–80° S.

The only well that clearly penetrated the "Ti Valley fault" is the Exxon No. 1 Retherford (D.F. 1,253 ft) (NW¼SW¼SE¼ sec. 5, T. 3 N., R. 18 E.; Fig. 6). The well was spudded in lowermost Atoka Formation on 7 July 1984 and finished drilling on 17 January 1985 at 19,046 ft T.D. The well was completed as a dry hole. The upper 1,943 ft of the well was not logged; as a result, a small thrust juxtaposing Johns Valley Formation over lower Atoka was not recorded, and logging started in Atoka Formation in the hanging wall of the "Ti Valley fault." Log analysis indicated the following formation tops in the well (tops below Springer at 15,550 ft from OCC Form 1002A): Atoka 1,943 ft; "Spiro" sandstone (industry term for medium-grained basal Atoka sandstone, locally complexly interbedded with Wapanucka limestone) 6,200 ft; thrust fault and Atoka 6,651 ft; "Spiro" 7,057 ft; Johns Valley 7,409 ft; thrust fault ("Ti Valley") and "Spiro" 7,752 ft; Wapanucka 7,880 ft; Springer 7,984 ft; thrust fault and Atoka 8,259 ft; thrust fault(?) 12,230 ft; thrust fault and Wapanucka 15,100 ft; Springer 15,550 ft; Cromwell 16,090 ft; Caney 16,930 ft; Caney repeat 17,065 ft; Woodford 17,469 ft; Hunton 17,601 ft; Sylvan 17,630 ft; Viola 17,692 ft; Bromide 17,818 ft; McLish 18,121 ft; Oil Creek upper 18,239 ft; Oil Creek basal 18,336 ft; Joins 18,504 ft; Arbuckle 18,678 ft.

The "Ti Valley fault" was penetrated at a depth of 7,752 ft as indicated by the presence of Johns Valley in the hanging wall of the fault and Wapanucka in the footwall. Below the lowest Atoka, the sequence is typical Arbuckle-facies strata; this suggests that the fault at ~15,100 ft is a major bedding-plane thrust (décollement?) that may be the Choctaw or Carbon fault. A dipmeter survey indicates that the strata dip on the average at an angle of ~55° from ~6,900 ft to 7,300 ft in the hanging wall of the "Ti Valley fault," and at an angle of ~60° at ~7,900 ft in the footwall (dipmeter data closer to the fault were of poor quality). These values for the dip of the fault are similar to the assumed approximate value based on parallelism with strata at the surface. Based on the mapped location of the fault and its depth in the well, the fault dips at 67° S.

Analysis of logs from the Exxon No. 1 Retherford also suggests that the hanging-wall glide planes for the three thrust faults associated with the "Ti Valley" are near the base of the "Spiro" sandstone, within the Johns Valley Formation, and within the Springer Formation. These relations are similar to those observed at the surface elsewhere in the Higgins Quadrangle: The Springer Formation forms the hanging-wall glide plane for a fault (Choctaw?) in the northern part of the quadrangle; basal "Spiro" in the hanging wall is juxtaposed against Atoka along a fault that parallels Cedar Creek; and several areas underlain by Johns Valley Formation in the southern parts of all three quadrangles are bounded by thrust faults along their northern margins.

Footwall and hanging-wall strata are folded (locally isoclinal and/or overturned) near the "Ti Valley fault" and associated faults (Fig. 6). This feature is significantly different from the fault to the west, where Hendricks and others (1947) showed few folds close to the fault. The feature may be an artifact of the mapping (Hendricks and others did not show facing directions on their map), or it may be real, the difference in style occurring where the Pine Mountain fault strikes NE (as mapped by Hendricks and others, 1947) near the western edge of the Higgins Quadrangle and then strikes N, paralleling Gaines Creek (Suneson and Ferguson, unpublished mapping; Fig. 6).

In summary, the "Ti Valley fault" in the Higgins, Damon, and Baker Mountain Quadrangles differs in several respects from the fault to the southwest:

1) Johns Valley-like strata are not restricted to the area south of the fault. This makes recognition of the fault considerably more difficult, particularly as footwall Atoka is commonly juxtaposed against hanging-wall Atoka.

2) The northernmost exposures of lower Atoka or Johns Valley olistostromes are as much as several miles north of the northernmost exposures of Jackfork strata, suggesting that displacment is distributed along several faults rather than one (the "Ti Valley fault"). Similarly, the Stanley Group is restricted to the area south of the Windingstair fault.

3) The "Ti Valley fault" appears to dip at an angle of ~60° and is generally bedding-parallel, but locally crosscuts strata and terminates folds.

4) Folding is common in both the footwall and hanging wall of the "Ti Valley fault."

TALIHINA QUADRANGLE

Fellows (1964) studied the geology of the Ti Valley fault in the Talihina 7.5′ Quadrangle. He showed footwall strata to consist entirely of Atoka Formation, with the exception of a small area south and west of Bengal, where the footwall of the Ti Valley fault consists of Woodford chert overlain by Caney shale (Fig. 7). The block containing the Woodford chert and Caney shale is surrounded by faults and is not related in any simple mappable way to the surrounding Atoka Formation. (The Woodford–Caney outcrop at Bengal was briefly described by Kramer [1933], who mistakenly referred to the outcrop as Arkansas Novaculite and indicated that Atoka Formation was deposited directly on the novaculite.) Fellows (1964, pl. I) also mapped Springer shale along two thrust faults immediately north of the Ti Valley fault. Hanging-wall strata in the Talihina Quadrangle are mostly Johns Valley shale. Fellows's (1964) map indicates that the trace of the Ti Valley fault is slightly oblique to the strike of both footwall and hanging-wall strata. The average dip of footwall strata is ~76°, and the dip of hanging-wall strata is ~68°, suggesting that the fault may have an overall dip of ~70° S. Fellows (1964, pl. I, cross section D–D′) showed the fault dipping 75°.

No wells have been drilled into the hanging wall of the Ti Valley fault in the Talihina Quadrangle.

Fellows (1964) mapped one large anticline and one small syncline in the hanging wall of the Ti Valley fault; he mapped no folds in the footwall, but stated that the fault was the northern margin of his mapped area, and structures north of the fault appear to be the result of aerial-photograph study only.

EASTERN OUACHITA MOUNTAINS FRONTAL BELT, OKLAHOMA

To the east, the Ti Valley fault juxtaposes Atoka Formation in the footwall against Atoka Formation in the hanging wall (Hart, 1963). The exact stratigraphic positions of the juxtaposed Atoka Formation on either side of the fault are unknown. The Johns Valley shale is restricted to the area south of a thrust

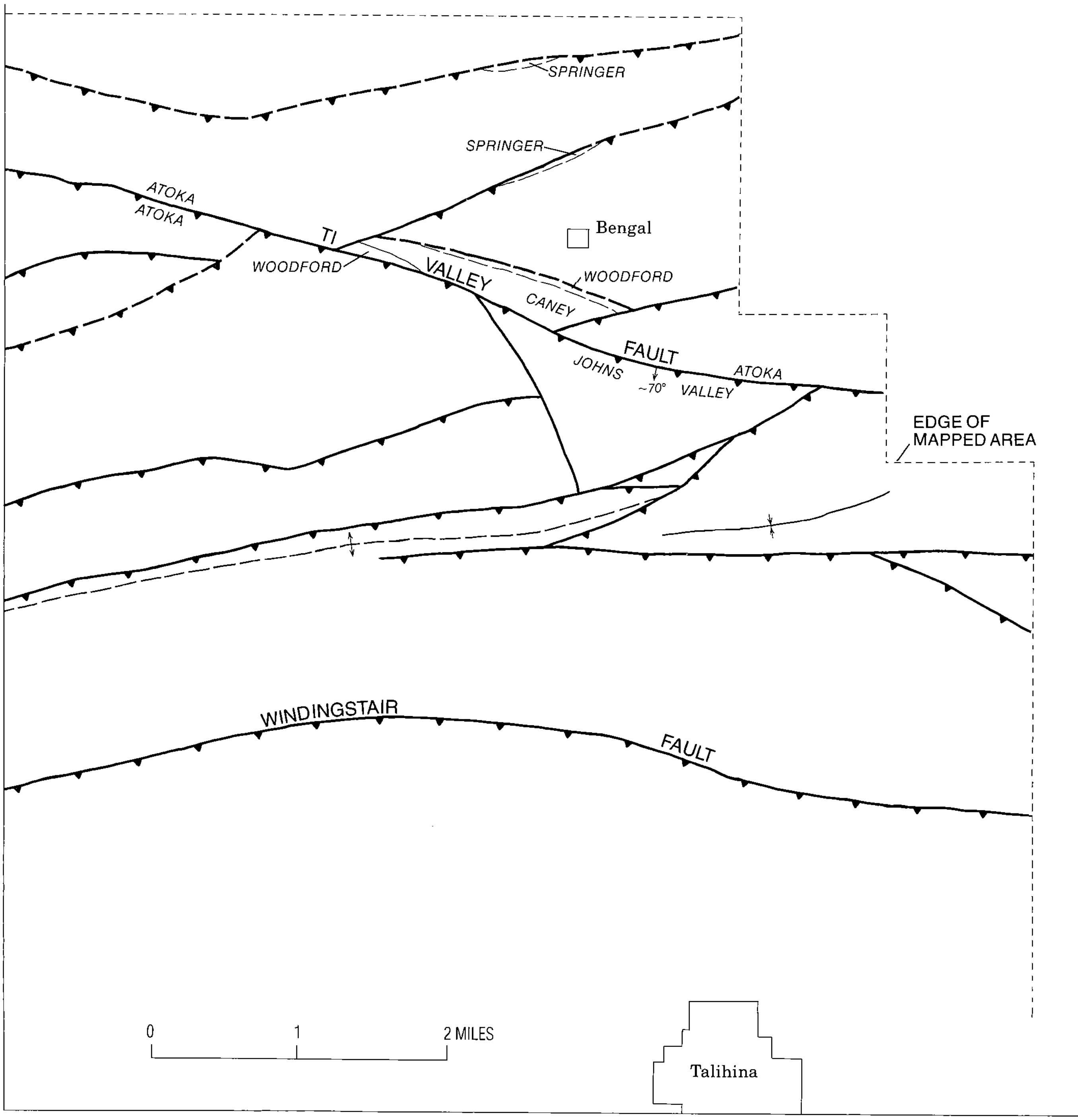


Figure 7. Features in Talihina 7.5′ Quadrangle. Geology from Fellows (1964).

fault south of the Ti Valley fault. Hart (1963, p. 53) suggested that displacement on the fault decreases from west to east across the eastern part of the Blackjack Ridge, Le Flore SE, and Hodgens Quadrangles. This is consistent with my observation that displacement is distributed among several thrust faults in the Higgins, Damon, and Baker Mountain Quadranges, and that displacement along one fault—the Ti Valley fault (Hendricks, 1959, p. 50)—decreases eastward. However, southeast of the Hodgens Quadrangle, Seely (1963) mapped Stanley Group north of the Windingstair fault, suggesting that the zone of thrust faults in which Ouachita- and Arbuckle-facies strata are juxtaposed may be narrower than to the west (near the Damon Quadrangle).

Detailed mapping in the vicinity of the Ti Valley fault in easternmost Oklahoma has not established its exact location or geologic relations. Miser (1954) showed the Ti Valley fault juxtaposing Atoka against Atoka in Le Flore County. In central Scott County, Arkansas, Reinemund and Danilchik (1957) mapped the Ti Valley fault juxtaposing footwall Atoka Formation against hanging-wall Johns Valley. The fault undoubtedly continues eastward, but the geology of the fault in Arkansas is beyond the scope of this report.

SUMMARY

Several aspects of the geology of the Ti Valley fault suggest that it may not be the fault with the largest displacement in the frontal belt of the Ouachita Mountains in Oklahoma, particularly in the central and possibly in the eastern part. The two surface criteria used by Hendricks (1959) to document displacement on the fault (facies change, klippe) are questionable. Exotic-bearing strata similar to the Johns Valley Formation occur north of the fault in at least two places, suggesting that sub-Atoka facies changes might be gradational across the frontal belt. Similarly, the surface appearance of thick Jackfork Group strata generally occurs two or more thrusts south of the Ti Valley fault, particularly in the newly mapped central part of the frontal belt. In Oklahoma, thick Stanley Group strata are generally restricted to the hanging wall of the Windingstair fault, except in the extreme western and eastern parts of the frontal belt. These observations and the field evidence that the faults are generally parallel to bedding suggest that sub-Atoka facies and thickness changes are distributed over several thrust faults and not necessarily a single fault—the Ti Valley. It is possible that the faults are cutting up-section (but are parallel to strike) at the present level of exposure and that a significant thickness of Stanley Group strata, for example, is present in the footwall of the Windingstair fault in the subsurface. This would narrow the zone of facies juxtaposition and increase the amount of displacement on fewer faults. However, there is no field evidence in the central frontal belt to support this hypothesis. The existence of the klippe is controversial (e.g., Misch and Oles, 1958), particularly in light of the fact that Cline (1960) mentioned Johns Valley-like strata northwest of the Ti Valley fault immediately southwest of the klippe.

The stratigraphic throw across the Ti Valley fault is clearly greater in the southwest corner of the frontal belt, where Ordovician Womble shale is juxtaposed against Pennsylvanian Atoka Formation, than it is in the north-central part of the belt, where the fault juxtaposes Atoka against Atoka. It is possible that the total displacement on the fault changes along strike and that tectonic shortening is accomplished on many faults and folds in the north-central part of the frontal belt, whereas folding is rare to the southwest—although Hendricks (1959, p. 49) briefly mentioned shortening by folding. Measurement of the total displacement on the Ti Valley fault awaits the release of detailed seismic data and the drilling of more deep hydrocarbon wells.

ACKNOWLEDGMENTS

Mapping in the Higgins, Damon, and Baker Mountain Quadrangles was supported, in part, by a grant from the U.S. Geological Survey's COGEOMAP Program. Several oil companies generously contributed dipmeter logs for this study and the COGEOMAP Program. The author is especially grateful to Charles Ferguson for many thoughtful discussions, and to J. Kaspar Arbenz and Ferguson for their critical reviews of the manuscript.

TECTONIC IMPLICATIONS OF EARLY PENNSYLVANIAN PALEOCURRENTS FROM FLYSCH IN THE OUACHITA MOUNTAINS FRONTAL BELT, SOUTHEAST OKLAHOMA

Charles A. Ferguson and Neil H. Suneson

ABSTRACT.—New paleocurrent data from the youngest part of the late Paleozoic Ouachita flysch, Atoka Formation (Pennsylvanian), suggest that sediment was distributed by at least two, possibly three, spatially distinct and coeval, oppositely directed axial fan systems. These fan systems are now separated into the three major high-angle reverse fault blocks that constitute the Ouachita Mountains frontal belt.

The largest fan system was W-directed and was part of a long-lived system that dominated the main part of the deep Ouachita basin. A smaller E-directed system (or systems), overlying Morrowan shelf-facies rocks that had foundered abruptly in the early Atokan, developed along the north edge of the Ouachita basin. Foundering, enhanced by south-side-down growth faulting, was caused by flexural subsidence of the southern passive margin of North America in response to tectonic loading of a northward-converging landmass. Growth faulting, well documented during the Atokan in the Arkoma basin, probably also occurred farther south, in what is now the frontal belt. Terrace-like, margin-parallel troughs created by these listric growth-fault blocks channeled Atokan flysch eastward from a western source, possibly the Arbuckle uplift, and isolated the fan system(s) from a larger and deeper trough to the south. Degradation of fault scarps bounding the north sides of the troughs produced widespread olistostromes containing blocks of the older Paleozoic shelf.

During the Pennsylvanian Ouachita orogeny, the growth-fault blocks composed of relatively rigid pre-Mississippian shelf-facies rocks (mostly limestone), were overridden by shale-rich, plastic Carboniferous rocks along a N-stepping décollement. High-angle reverse faults were formed by upward deflection of the décollement at preexisting growth-fault scarps. These faults were later carried northward as the décollement continued to propagate. In this way, flysch originally deposited in isolated growth-fault troughs could be contained within blocks bounded by major high-angle reverse faults of the present day frontal belt.

PURPOSE

This paper presents recently acquired paleocurrent data from Early Pennsylvanian flysch (sandstone turbidites) in the western part of the Ouachita Mountains frontal belt of southeast Oklahoma (west of 95°7.5′ W.). Previous workers (Briggs, 1962; Briggs and Cline, 1967) determined that these sands were distributed from east to west, with a minor component of southerly paleoflow. New data show that, although W-directed paleocurrents are abundant, there is a significant, spatially distinct, population of E-directed paleocurrents. These paleocurrents, unrecognized until recently (Ferguson and Suneson, 1988), occur in the youngest part of the Ouachita flysch, Atoka Formation (Taff and Adams, 1900), and are restricted to the north edge of the orogenic belt.

Two groups of paleocurrent measurements are presented here. The largest group, mostly from Atoka Formation, but also from older flysch, was collected concurrently with detailed geologic mapping of three 7.5′ quadrangles in the frontal belt (east of 95°30′ W.). A smaller group was collected from nine widely distributed exposures of Atoka Formation farther west in the frontal belt.

GEOLOGIC SETTING

Ouachita Basin, Arkoma Basin System

The Ouachita basin of southeast Oklahoma and west-central Arkansas originated along the southern edge of North America during the late Proterozoic to Cambrian rifting event that led to the opening of the Iapetus Ocean (Keller and Cebull, 1973; Walper, 1977). In the basin, Cambrian through Early Mississippian strata are starved-basin (Cline, 1970; Morris, 1974a), deep-ocean-facies rock (black shale, chert, minor sandstone). Equivalent strata to the north, along the early Paleozoic southern edge of the North American continent, are shallow-water shelf rocks (Arbuckle Group through Woodford Shale).

The intervening continental-shelf break, which persisted as a passive margin until the Early Pennsylvanian, is now an allochthonous feature somewhere in the Ouachita frontal belt.

In Early Mississippian time, the deep basin had evolved into an E–W-trending flysch trough bounded to the south by a northward-converging landmass christened Llanoria by E. T. Dumble and Sidney Powers in 1920 (see discussion by Miser, 1921, p. 62). The passive margin to the north, mantled by the Morrowan–Atokan shallow-water-shelf Wapanucka Formation (Grayson, 1979) foundered abruptly during the early Atokan. Lithospheric loading of a tectonically thickened package north of the Llanoria landmass is thought to be the cause of flexural subsidence that led to the foundering event (Houseknecht, 1986). The subsidence was enhanced by short-lived (early to middle Atokan), margin-parallel, south-side-down growth faulting of the passive margin (Koinm and Dickey, 1967; Buchanan and Johnson, 1968). The resulting Arkoma foreland basin filled rapidly with a north-thinning wedge of mostly deep-water-facies Atokan flysch. Eventually (by at least the Desmoinesian) the foreland evolved into a north-directed fluvial molasse basin.

Ouachita Frontal Belt

The approximately 15-km-wide Ouachita Mountains frontal belt consists of as much as 4.0 km of complexly folded Pennsylvanian strata in steeply S-dipping, imbricate high-angle reverse fault blocks. Three major fault blocks, named after the northward bounding fault, constitute the frontal belt between 95° and 95°30′ W. From north to south, these are the Choctaw, Pine Mountain, and Ti Valley blocks. The pre-Atokan stratigraphy of each block, summarized in Figure 1, reflects southward facies changes of time-equivalent strata from shallow-water shelf to deep basin. For the pre-Pennsylvanian sequence, the major facies change from shallow shelf to deep basin occurs across the Ti Valley fault (Hendricks, 1959; Cline, 1960). This facies change migrated north during the Morrowan; its allochthonous position, defined by the southern limit of Wapanucka Formation, is now somewhere between the Choctaw and Pine Mountain faults. The basinward equivalent of the Wapanucka Formation is the Johns Valley Shale (Taff, 1902), a shale-rich flysch sequence with olistostromes containing exotic shelf-facies blocks that range in age from Cambrian to Pennsylvanian (Ulrich, 1927; Shideler, 1970). In the Pine Mountain block, pre-Wapanucka shelf-facies rocks (Springer shale and Caney shale) are overlain by Atoka Formation flysch with rare olistostromes near the base that are lithostratigraphically correlated with the Johns Valley Shale (Fig. 1). Farther south, in the Ti Valley block, Morrowan Johns Valley Shale and Jackfork Group olistostromes are widespread, and older strata are deep-basin-facies rocks.

The only lithologic unit common to each fault block is the Atoka Formation (Fig. 1), which consists of

TIME		FAULT BLOCKS: CHOCTAW	FAULT BLOCKS: PINE MOUNTAIN	FAULT BLOCKS: TI VALLEY
PENNSYLVANIAN	DESMOINESIAN	Krebs Group: fluvial molasse, sandstone and shale	?	?
PENNSYLVANIAN	ATOKAN	Atoka Formation: marine flysch, shale, and sandstone		
PENNSYLVANIAN	MORROWAN	Wapanucka Formation: shelf-facies limestone, sandstone, and minor shale	Johns Valley Shale: marine shale-rich flysch with olistostromes	
PENNSYLVANIAN	MORROWAN	Springer shale: outer-shelf-facies calcareous shale and sandstone		Jackfork Group: marine sand-rich flysch with olistostromes
MISSISSIPPIAN	CHESTERIAN	Caney shale: outer-shelf-facies calcareous shale and pelagic limestone		Stanley Group: marine volcaniclastic shale-rich flysch with tuff beds
CAMBRIAN through EARLY MISSISSIPPIAN		Shelf-facies rocks: Arbuckle Group – Woodford Shale		Marine deep-basin-facies rocks: Arkansas Novaculite and older strata

Figure 1. Generalized stratigraphic framework of the frontal belt of the Ouachita Mountains. Shelf-facies Caney shale and Springer shale are informal names.

dark-gray shales and fine-grained, feldspathic, quartzose sandstones. Two general lithologic associations are recognized in the Atoka Formation. The most common association is thick shale sequences interbedded with thin- to medium-bedded, planar-laminated and ripple-cross-laminated sandstone ("b" and "c" divisions of the Bouma [1962] turbidite sequence). The second association is stacked sequences of thick-bedded, massive sandstone (the Bouma "a" division) with thin shale interbeds. These lithologies are indicative of middle to outer reaches of a deep-sea fan (Walker, 1978), the stacked sequences representing deposits of distributary channels. The Atoka Formation is 3.7–5.0 km thick along the southern edge of the Arkoma foreland basin. To the south, in the Ouachita basin, its estimated thickness is as much as 8.2 km (Morris, 1974a), but, because of structural complications, poorly understood biostratigraphy, and the lack of an exposed top, it is impossible to estimate its greatest thickness here.

PREVIOUS PALEOCURRENT STUDIES

Figure 2 is a typical paleogeographic map of the late Paleozoic Ouachita basin (Houseknecht and Kacena, 1983), showing that sediment, mostly de-

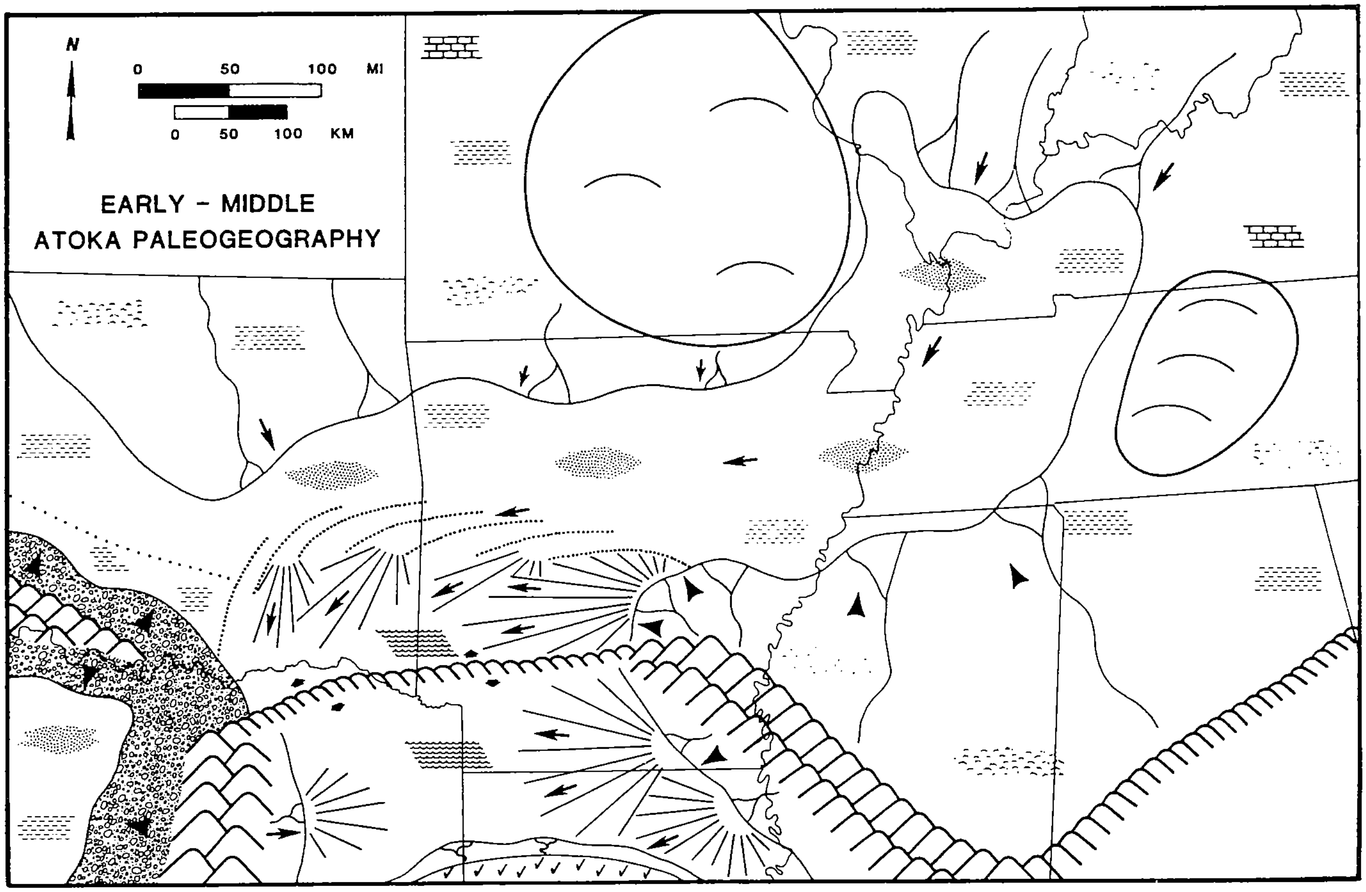


Figure 2. Early Pennsylvanian paleogeography of the Ouachita region proposed by Houseknecht and Kacena (1983, fig. 9). Tectonic features and lithologies indicated by symbols. Arrows indicate sediment transport direction.

rived from erosion of the Appalachian highlands, was channeled westward through the Black Warrior and Illinois basins, and eventually distributed from east to west in a deep-sea axial fan system. This sediment-dispersal pattern is supported primarily by a paleocurrent study of the Ouachita flysch sequence by Briggs and Cline (1967). Paleocurrent and petrologic studies of the older flysch sequence of the Stanley Group and Jackfork Group indicate additional contributions of sediment, in part volcaniclastic, from a southern orogenic source area (Briggs and Cline, 1967; Graham and others, 1976; Mack and others, 1981). Similar studies of the Atoka Formation suggest that a secondary northern cratonic source became important during the later stages of flysch sedimentation (Briggs and Cline, 1967; Vedros and Visher, 1978; Houseknecht, 1986).

Briggs and Cline's (1967) paleocurrent map for the Ouachita flysch sequence (Fig. 3) shows a dominance of W-directed azimuths. Each one of these azimuths represents the average of ~13 separate measurements in an area of 1 mi^2 (Briggs and Cline, 1967, p. 992). From the appendix of Briggs (1962), it should be noted—particularly in the frontal belt—that his average azimuths were derived mostly from non-directional indicators. The study of Briggs and Cline (1967) is still the principal source of paleocurrent data in the Ouachitas, and, although it is an excellent work, it is important to note that it was only a reconnaissance study. Large areas of the Ouachitas, including parts of the frontal belt, are represented by only one or two azimuths. Subsequent field studies in the Ouachita frontal belt have concluded that paleocurrents are similar to those originally reported for that area (Knight, 1985; Nooncaster, 1985), but these conclusions are not supported with new data. Invariably, the only previous report is the work of Briggs and Cline (1967).

METHODS

Detailed geologic mapping of the frontal belt began in October 1986. During the mapping, paleocurrent azimuths were measured in all sandstones with well-preserved primary sedimentary structures that could be found. The most reliable indicators of paleoflow in turbidite sandstones are sole markings, flute molds in particular. Flute molds (commonly yet technically incorrectly termed flute "casts") originate as asymmetric erosional scours developed in a substrate of cohesive, fine-grained sediment. The scoured de-

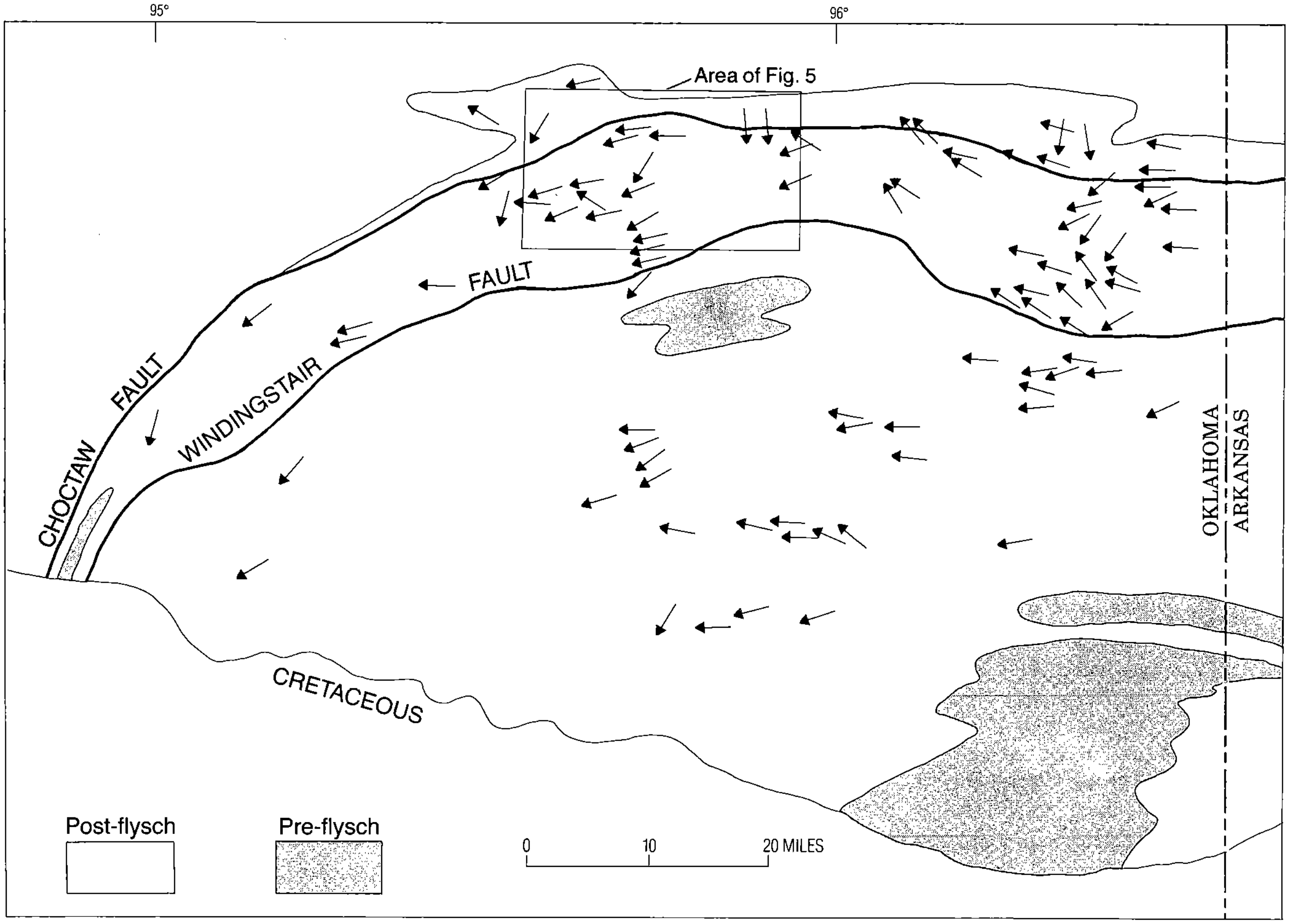


Figure 3. Directional paleocurrents from the Ouachita flysch sequence reported by some previous workers (modified from Briggs and Cline, 1967, fig. 4).

pressions are commonly filled with non-cohesive sediment, typically sand transported by an eroding current. After burial and lithification, the scours or flutes are preserved as molds on the sole of the more-resistant sandstone bed. About 90% of the paleocurrent directions reported here were measured from flute molds. Other paleocurrent indicators, such as climbing-ripple bedforms (the "c" division of the Bouma sequence), can be used, but they are difficult to measure because they require rare three-dimensional exposures. In (~10% of the measurements reported here, the azimuths of nondirectional sole marks (poorly developed flute molds, dragmarks, etc.) were assigned a directional sense from the direction of ripple-bedform migration within the same Bouma sequence.

Flow reversal in turbidity currents (Pickering and Hiscott, 1985; Pantin and Leeder, 1987) can produce ripple bedforms that climb in a direction opposite to the original turbidity current. These reversal bedforms occur in sequences usually separated by shale partings and are marked by a renewed vertical progression of the Bouma sequence. Reversal bedforms occur in the study area, but paleocurrent measurements from them are not reported in this paper (see, for example, an exposure in a creek bed on the west side of a county road in the NE¼NE¼ sec. 35, T. 5 N., R. 20 E.).

Paleocurrent azimuths reported in this paper were measured and rotated back to horizontal using the strike of bedding as the rotational axis. This assumes that structural tilting did not involve any significant rotation around nonhorizontal axes, which is probably a good assumption, given the style of coaxial deformation evident in the area of the detailed study (east of 95°30′ W.). However, reconnaissance paleocurrent measurements farther west in the frontal belt could be more susceptible to counterclockwise rotation.

DATA

Paleocurrents East of 95°30′ W.

A total of 379 directional paleocurrent azimuths were measured from the Atoka Formation during mapping of the Higgins, Damon, and Baker Mountain 7.5′ Quadrangles, and southern parts of the Wilburton and Panola 7.5′ Quadrangles. All of the azimuths are collectively plotted on rose diagrams, which show a pronounced bimodal distribution clustered about means of 260° and 65° (Fig. 4). The distribution of individual measurements and/or clusters of measurements is shown in conjunction with a simplified geologic map of the aforementioned quadrangles (Figs. 5,6). It should be noted here, that the geology in Figure 5 is Ferguson's interpretation of unpublished mapping by Suneson and Ferguson.

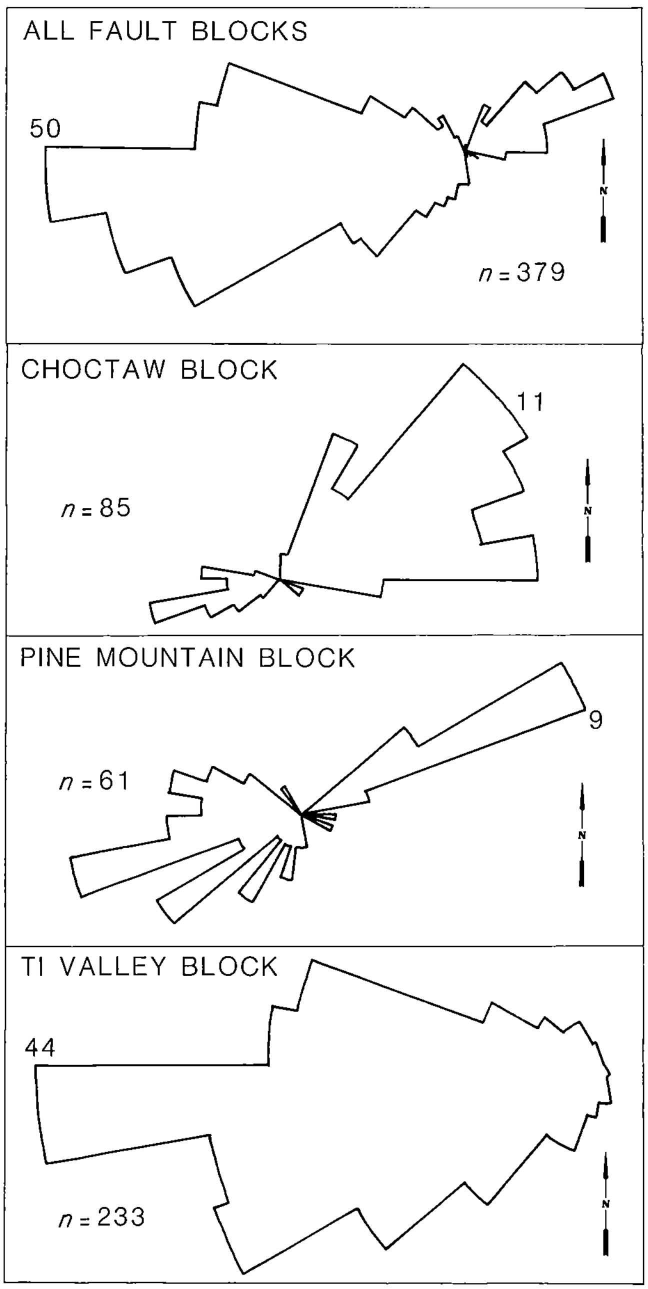


Figure 4. Rose diagrams of directional paleocurrents from the Atoka Formation of the Ouachita frontal belt between 95°7.5′ and 95°30′ W. Top diagram combines measurements from all three structural blocks. Diagrams at different scales.

A large number of rose diagrams were constructed, combining various populations of paleocurrents. The populations were partitioned according to lithologic unit, fault block, and combinations of fault blocks. From consideration of these rose diagrams, three natural populations of paleocurrents are identified, each genetically related to a major fault block (Ti Valley, Pine Mountain, or Choctaw). The southern limit of E-directed paleocurrents is the Ti Valley fault.

Choctaw Block

The maximum thickness of Atoka Formation in the Choctaw block is ~3.5 km. Here the base is well constrained by slivers of Wapanucka Formation along the hanging wall of high-angle reverse faults. Paleocurrents are more difficult to measure in this block, partly because of the abundance of bedding-parallel drainages, but also because sole markings in the sandstones are not often well developed or have been obscured by intense bioturbation. However, a good number (85) were measured, and it is clear that E-directed paleoflow dominated (Fig. 4). A small W-directed population was obtained west of 95°15′ W. representing individual beds intimately interleaved with beds of the opposite sense (for an example, see exposure in a major creek bed, SE¼ sec. 20, T. 5 N., R. 19 E.). Farther east, the W-directed paleocurrents become increasingly abundant in the Baker Mountain Quadrangle, and E-directed paleocurrents are very rare (Fig. 6).

Pine Mountain Block

The thickness of the Atoka Formation in the Pine Mountain block might be as little as 1.5 km or as much as 3.0 km. Most of the paleocurrents are W-directed (Fig. 4), but there is a small E-directed population. The impression from field observation is that the E-directed sandstones are intimately interleaved with the more abundant W-directed sandstones.

Ti Valley Block

The greatest thickness of Atoka Formation in the Ti Valley block is about 2.0–2.5 km. Paleocurrents from this block (Fig. 4) are distributed evenly throughout the Atoka and are dominantly W-directed. There are a couple of areas where the dominant paleocurrent trend deviates notably from the mean of ~260° (Fig. 6). In the western Damon Quadrangle there is a prevailing shift towards NW-directed currents, and in the east-central Baker Mountain Quadrangle there is a zone where S- to SW-directed currents prevail.

Paleocurrents from Morrowan flysch (Jackfork Group and Johns Valley Shale) are restricted to the southern parts of the Damon and Baker Mountain Quadrangles (Fig. 6). A total of 15 measurements from this sequence show a bimodal westerly and southerly distribution of paleoflow (Fig. 7). The S-directed currents are from the Johns Valley Shale, and the W-directed currents are from Jackfork Group sandstones.

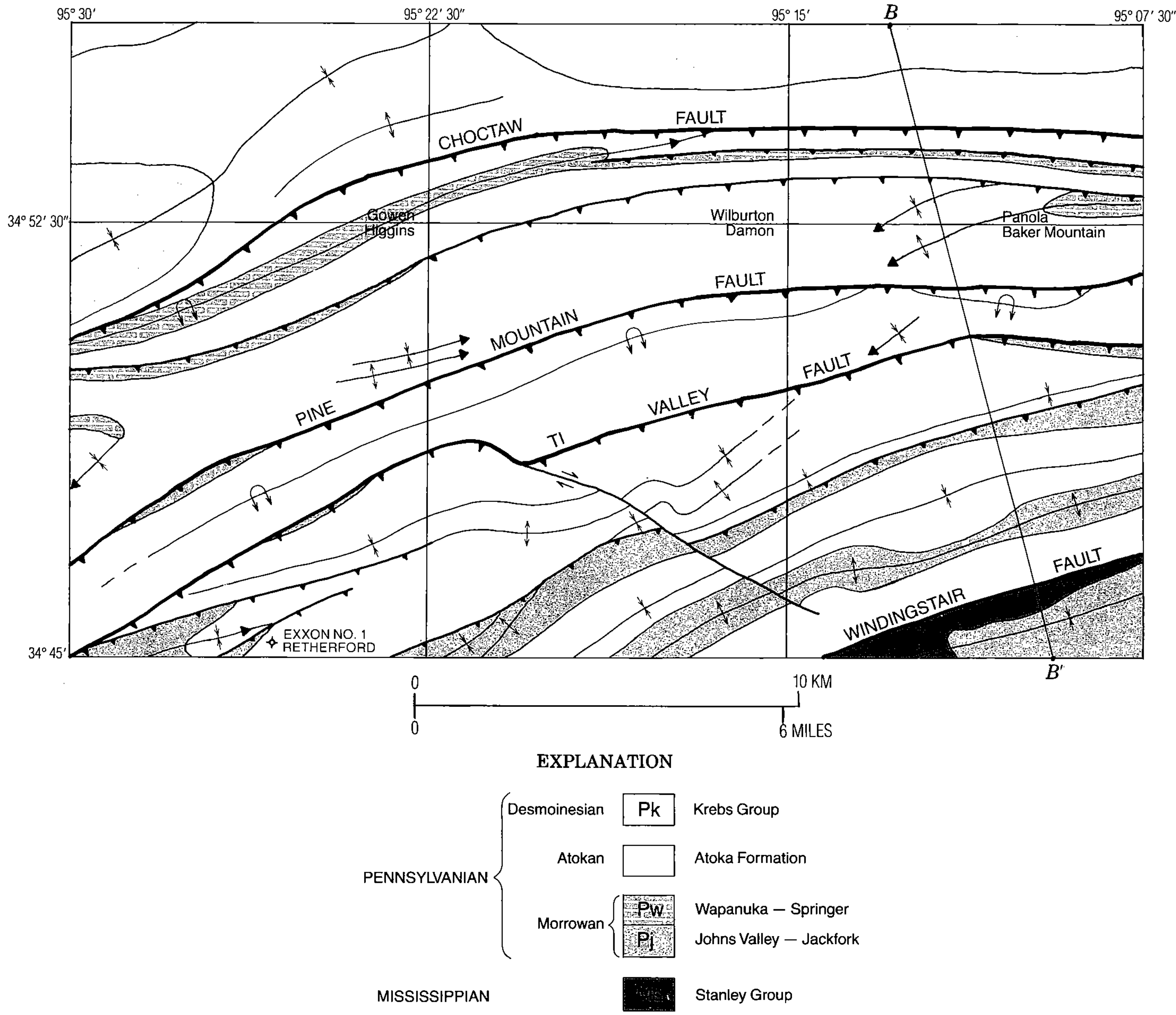


Figure 5. Simplified geology from the central Ouachita frontal belt, Latimer County, Oklahoma. Line B–B′ refers to cross section in Figure 10.

Paleocurrents from the Western Frontal Belt

A total of 76 paleocurrents (Fig. 6) were measured from nine widely distributed localities in the western part of the frontal belt (between 95°30′ and 96° W.). This reconnaissance study was undertaken to see if the paleocurrent patterns recognized east of 95°30′ W. continued to the southwest. Six of the localities are in the Choctaw block, and three in the Ti Valley block. Paleocurrents in the Pine Mountain block, more recondite because of the subdued topography and poor exposure, were not measured. All of the localities were chosen from the detailed geologic map of the western Ouachita Mountains prepared by Hendricks and others (1947). The localities represent sequences 15–600 m thick in the lower 125–1,000 m of the Atoka Formation. Detailed descriptions of each locality can be found in the appendix. A rose diagram from each locality is plotted on a map of the southwest frontal belt (Fig. 8).

Choctaw Block

Fifty-seven paleocurrents from six localities in the Choctaw block were measured. The data show a dominance of NE-directed paleoflow. Paleocurrents in the Choctaw block tend to parallel the structural grain of the frontal belt, becoming more N-directed to the southwest (Fig. 8). This may reflect an increase in counterclockwise rotation related to non-coaxial fold-

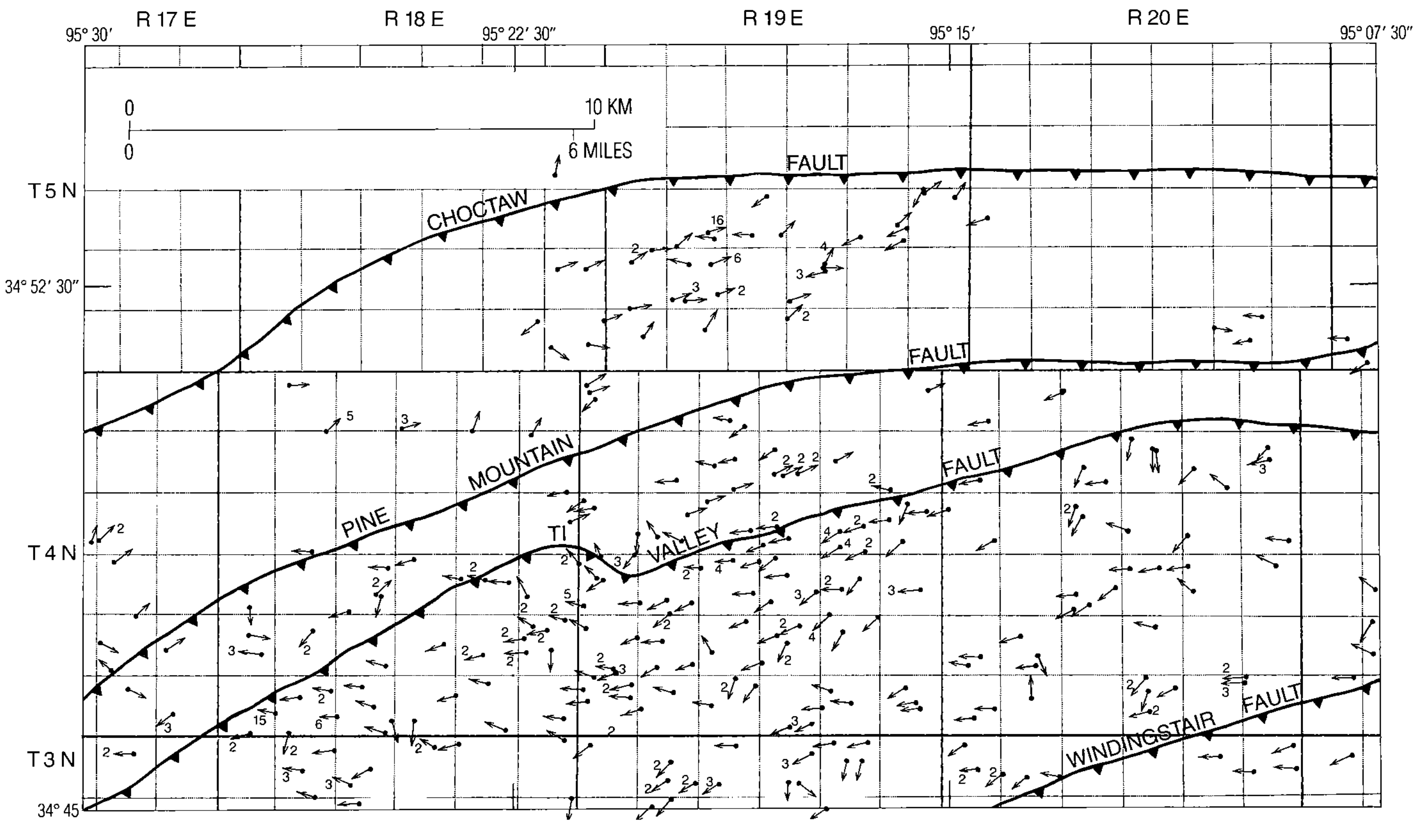


Figure 6. Directional paleocurrent azimuths from the area shown in Figure 5. Most arrows are individual paleocurrent measurements. Numbers indicate number of measurements for averaged values.

ing (Leonhardt, 1983) in this fault block. The important thing to note here is that very large rotations would not change the general easterly paleoflow in this block.

Ti Valley Block

Nineteen paleocurrents were measured from three localities of Atoka Formation in the Ti Valley block (Fig. 8). These azimuths are very similar to those observed to the east, and there is no apparent change of orientation from east to west in this part of the frontal belt.

DISCUSSION

Axial Sediment Distribution in the Ouachita Trough

New paleocurrent data strongly support the interpretation that margin-parallel axial fan systems distributed Early Pennsylvanian flysch in the northern part of the Ouachita basin (Briggs and Cline, 1967; Cline, 1970; Morris, 1974a; Graham and others, 1976; Moiola and Chanmugam, 1984). In addition, the data indicate that two opposing coeval depositional systems were active in Atokan time.

Two Opposing Systems

The lower 2.0–2.5 km of Atoka Formation in the Ti Valley block was clearly deposited by west-flowing turbidity currents. In the Ti Valley block, Atoka Formation overlies a thick sequence of older flysch (Jackfork Group and Stanley Group) which was also deposited by west-flowing currents (Briggs and Cline, 1967; Briggs, 1974; Morris, 1974b). The sediment-dispersal system in this block was consistently W-directed throughout the Carboniferous, and apparently extended into the western part of the Ouachita basin (Fig. 8).

Paleocurrents in the Choctaw block are almost exclusively NE-directed in most of the frontal belt. This unimodal distribution is interpreted as evidence for a NE-directed axial fan system that was somehow isolated from an apparently coeval, yet longer-lived, W-directed axial system in the larger and presumably deeper Ouachita basin to the south. The predominant E-directed paleoflow in flysch overlying the Wapanucka Formation suggests that the foundered Wapanucka shelf edge was the divide between two sediment-dispersal systems. Outboard from the shelf edge, which in allochthonous position coincides roughly with the trace of the Pine Mountain fault, E-directed paleocurrents are rare or absent.

The Importance of Growth Faulting

Down-to-the-south, margin-parallel growth faulting, documented during deposition of the lower Atoka Formation in the Arkoma basin (Koinm and Dickey, 1967; Houseknecht, 1986), probably also occurred in what is now the northern part of the frontal belt (Arbenz, 1984). Figure 9 includes a structural map and cross section of major growth faults in the Arkoma basin. A revised version of Early Pennsylvanian paleogeography (Fig. 10) suggests that sea-floor topography along the north edge of the Ouachita basin consisted of terrace-like, margin-parallel ridges and valleys formed by growth-fault blocks tilted toward the continent. The ridges would have acted as divides isolating individual axial sediment-dispersal systems in the intervening valleys. Houseknecht (1986) offered a similar interpretation for the origin of elongate, margin-parallel sand bodies in the Atoka Formation for the subsurface of the Arkoma basin to the east.

The relatively steep, basin-facing fault scarps would have been an ideal source for boulders of older shelf-facies rocks that occur in olistostromes of the Morrowan Johns Valley Shale and Jackfork Group. In order to sample rocks as old as the Arbuckle Group, these faults would have had to expose ~1.0 km of shelf strata below the Wapanucka Formation. This thickness is from a summary by Suneson (this volume) of the 5.7-km-deep Exxon No. 1 Retherford well, which was spudded ~3 km south of the Ti Valley fault in the southern part of the Higgins 7.5′ Quadrangle (Fig. 5).

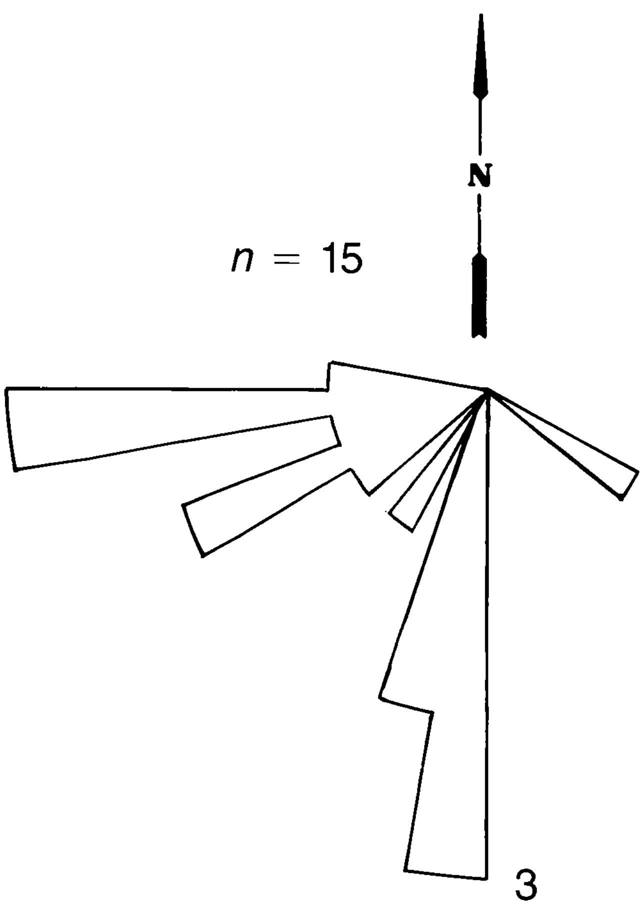


Figure 7. Rose diagram of directional paleocurrents from sandstones of the Jackfork Group and Johns Valley Shale in the southern Damon and Baker Mountain 7.5′ Quadrangles.

Mixing of Paleocurrents

Mixing of E- and W-directed paleocurrents occurs in two places in the area of detailed study (Fig. 6): the eastern part of the Choctaw block, and throughout the Pine Mountain block. The paleogeographic map (Fig. 10) shows two opposing depositional systems interfingering in these areas. In the eastern Choctaw block, the distal part of an E-directed axial system interfingers with W-directed turbidites that increase in abundance to the east. This change occurs to the south of the Red Oak sandstone, a SW-directed marginal fan in the subsurface of the Arkoma basin (Vedros and Visher, 1978; Houseknecht, 1986). The prevalence of S-directed paleocurrents farther south, in the Ti Valley block of the central Baker Mountain Quadrangle (Fig. 6), could be construed as evidence for this marginal fan system.

The paleocurrent pattern in the Pine Mountain block differs from that of the eastern Choctaw block in that mixing occurs throughout the studied area (Fig. 6). The dominance of W-directed paleoflow in this depositional system suggests that it was isolated from the Choctaw block, probably by a topographic barrier such as the Wapanucka shelf edge (Fig. 11B). E-directed turbidites in the Pine Mountain block are interpreted to be distal equivalents of a system to the west (Fig. 10). Alternatively, they could have spilled over from the Choctaw system to the north.

It is difficult to say how turbidites in the Pine Mountain block related to the large fan system in the Ti Valley block farther south. However, the complete absence of easterly paleoflow in the Ti Valley block, suggests that there was a topographic barrier along the southern edge of the Pine Mountain block, similar to the one south of the Choctaw block.

A Western Source for Part of the Atoka Formation

The unidirectional paleoflow of Atoka Formation turbidites in the northern part of the frontal belt suggests a sediment source area to the west of the present Ouachita uplift during the Early Pennsylvanian. Coarse conglomerates have been described from the Atoka Formation in the western frontal belt, near the town of Atoka, and were interpreted by Hendricks and others (1937) and Hendricks (1959) as evidence for a nearby uplift. There is also good stratigraphic evidence of Early Pennsylvanian uplift of the northeastern Arbuckle Mountains. The Hunton arch, in western Coal County, is unconformably overlain by a W-thinning, transgressive wedge of middle Atoka Formation (Grayson, 1981; Sutherland and others, 1982; Sutherland and Manger, 1984b;

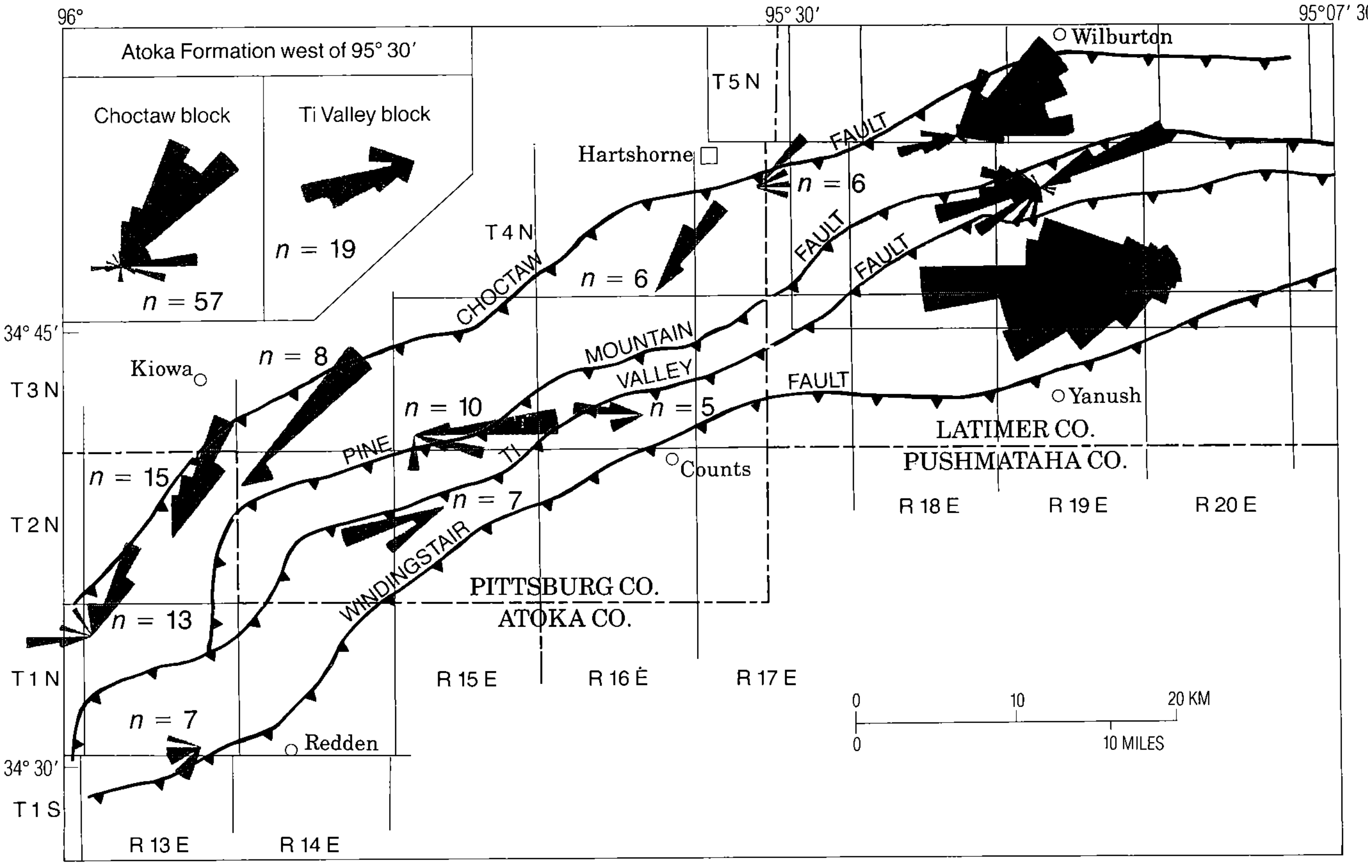


Figure 8. Rose diagrams of directional paleocurrents from nine exposures of Atoka Formation in the western part of the frontal belt (west of 95°30′ W.). Also shown is the detailed study area (shaded rectangular area) with rose diagrams from Figure 4.

Zachry and Sutherland, 1984). In addition, Archinal (1979) described Atokan syndepositional faulting in this area. New paleocurrent data indicate that sediment was transported from the Arbuckle region, but these data alone cannot locate a specific source area (or areas). Petrologic and geochemical analyses of E-directed sandstones and associated shales in the Choctaw block are required for provenance studies.

Structural Model for the Frontal Belt

A two-stage structural model for the formation of the frontal belt is shown in Figure 11. The main purpose of this model is to show how Atoka strata, deposited by opposing axial-fan systems in growth-fault-controlled troughs, could later be contained within high-angle-reverse-fault blocks whose origins are related to the original growth-fault topography. The structural cross section (Fig. 11A) was constructed from surface information (Fig. 5), and should be considered very hypothetical below 2 or 3 km.

Figure 11B is a cross section of the frontal belt as it might have been in Atokan time. It shows relatively resistant, approximately 1-km-high buttresses of early Paleozoic shelf strata buried under the north edge of the Ouachita basin.

Figure 11A depicts the present-day result of the N-directed Ouachita orogeny (note that a great deal of the original rock is now eroded away). In the model, growth-fault buttresses act as ramps deflecting a basal décollement upward, forming high-angle reverse faults. The high-angle reverse fault formed at each buttress is the northern boundary of strata deposited in the original growth-fault block (e.g., the Ti Valley reverse fault forms at the buttress along the north edge of the Ti Valley growth-fault block). As the décollement propagated northward, it carried each high-angle reverse fault block with it.

Certain features of the structural cross section (Fig. 11A) should be explained. Although the model implies that faults become younger toward the foreland, hindward younging is also possible (Suneson and Ferguson, 1987). If forward motion of the décollement is stalled, hindward propagation of younger high-angle reverse faults, such as the Windingstair or Eightmile (Fig. 10A), can occur at overridden buttresses (note that the model shows these two faults carried slightly northward by later motion of the décollement). A second feature of the model involves secondary folding of the high-angle reverse faults, which has been proposed for the Ti Valley fault at Black Knob Ridge in the western Ouachita Mountains (Leonhardt, 1983; Nielson and Leonhardt,

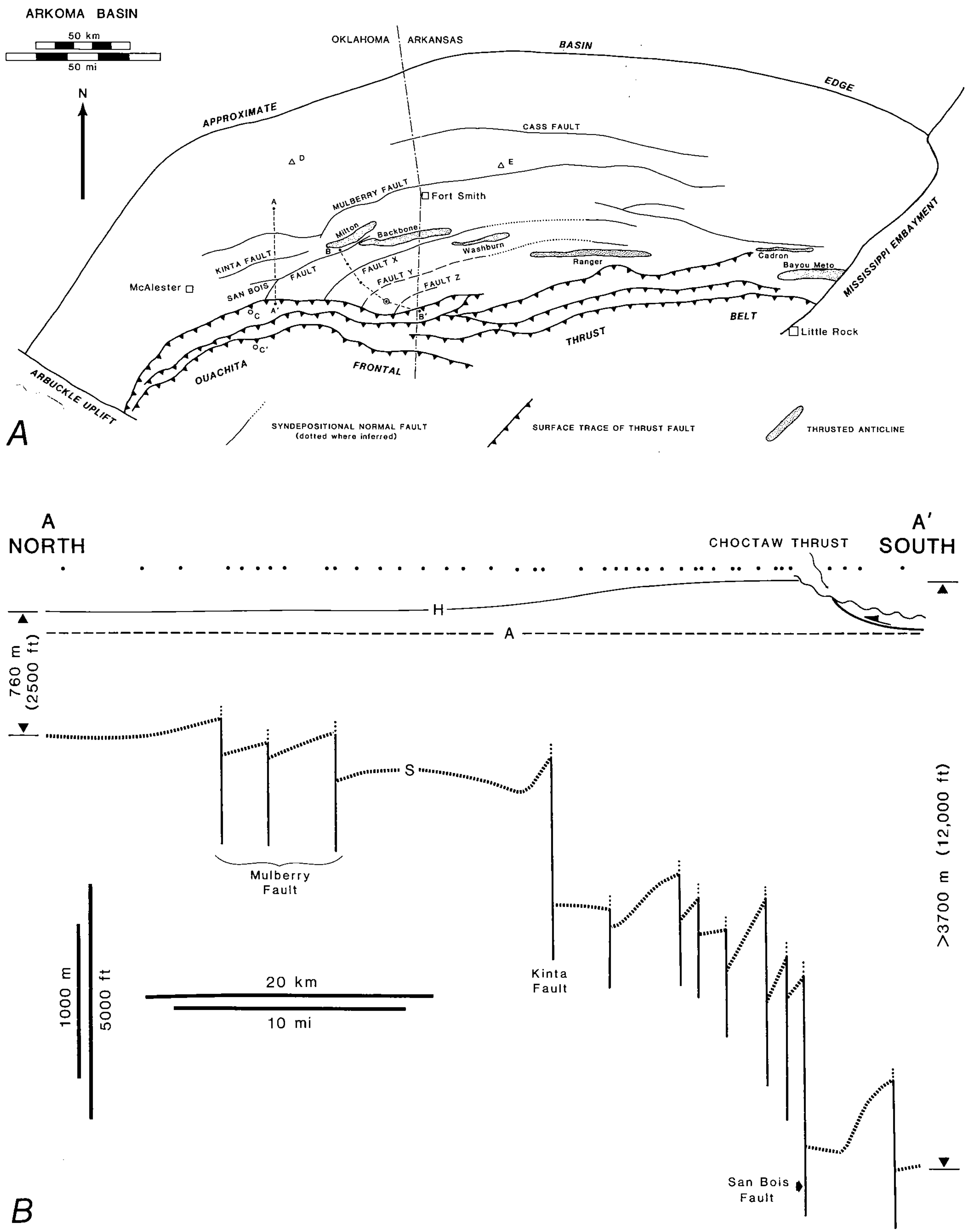


Figure 9. *A*, south-side-down, syndepositional normal faults in the subsurface of the Arkoma basin and their relationship to major thrust faults of the Ouachita frontal belt (from Houseknecht, 1986, fig. 5). *B*, cross section A–A′ (from Houseknecht, 1986, fig. 6), constructed from wire-line logs of 41 wells (locations indicated by dots). S = Spiro Sandstone (Wapanucka Formation equivalent), A = widespread upper Atoka key bed that serves as a datum for the cross section, H = Hartshorne Sandstone (base of the Desmoinesian Krebs Group molasse sequence).

1983). Fault-propagation folding of the original high-angle reverse faults occurred as the basal décollement overrode additional buttresses (Fig. 10A). An apparent klippe of the Ti Valley fault mapped by Hendricks and others (1947) in the western frontal belt could be explained in this way (imagine a slightly deeper erosion level of the Ti Valley fault in Figure 11A). There is also good evidence of secondary folding of folds related to the Ti Valley fault in the area of detailed study (Fig. 5).

CONCLUSIONS

Early Pennsylvanian flysch sediment was distributed by two, or possibly three, contemporaneous, margin-parallel axial fan systems in the frontal belt of the Ouachita Mountains. Each system is contained within one of three major high-angle-reverse-fault blocks that make up the frontal belt: Choctaw, Pine Mountain, and Ti Valley, from north to south. The two northern systems were partitioned into terrace-like troughs formed by south-side-down, listric normal faulting of a passive margin along the northern edge of the Ouachita basin. The northernmost block (Choctaw) hosted an E-directed system that overlies shelf-facies Wapanucka Formation, and appears to die out to the east (near 95° W.). In the southernmost block (Ti Valley), a W-directed axial system is the upper part of an older system that dominated the deep Ouachita basin during the Carboniferous. Turbidites in the Pine Mountain block, overlying slope-facies rocks and mostly W-directed, were possibly part of the large fan to the south, but the possibil-

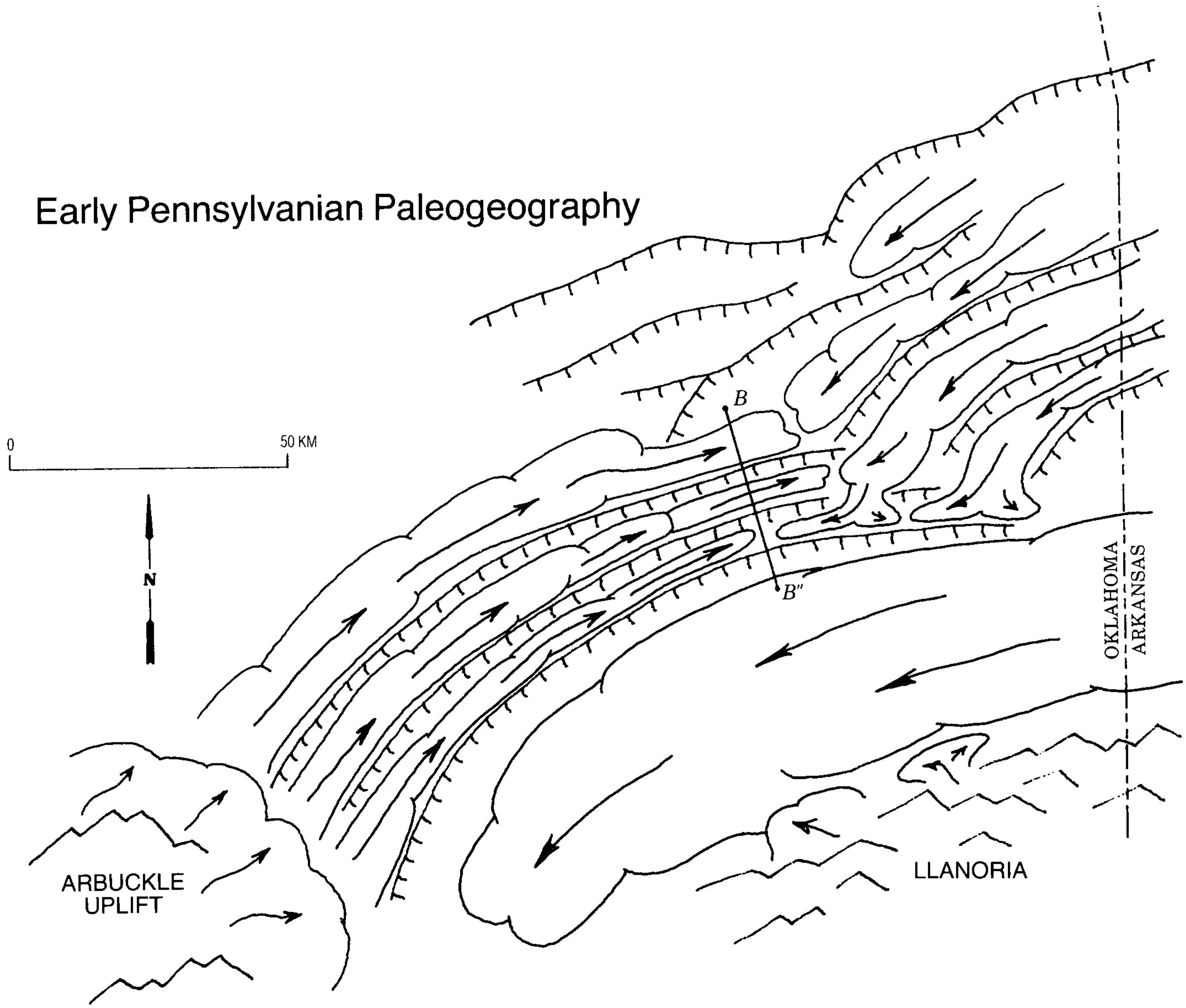


Figure 10. Early Pennsylvanian paleogeography of the Ouachita region, showing two opposing axial fan systems and their relationships to growth-fault troughs along the north edge of the Ouachita basin. The three southernmost faults are, from south to north, the northern boundaries of growth-fault blocks depicted in Figure 11: Ti Valley block, Pine Mountain block, Choctaw block. The other faults are those previously recognized in the subsurface of the Arkoma basin (see Fig. 9). Line B–B'' represents cross sections in Figure 11.

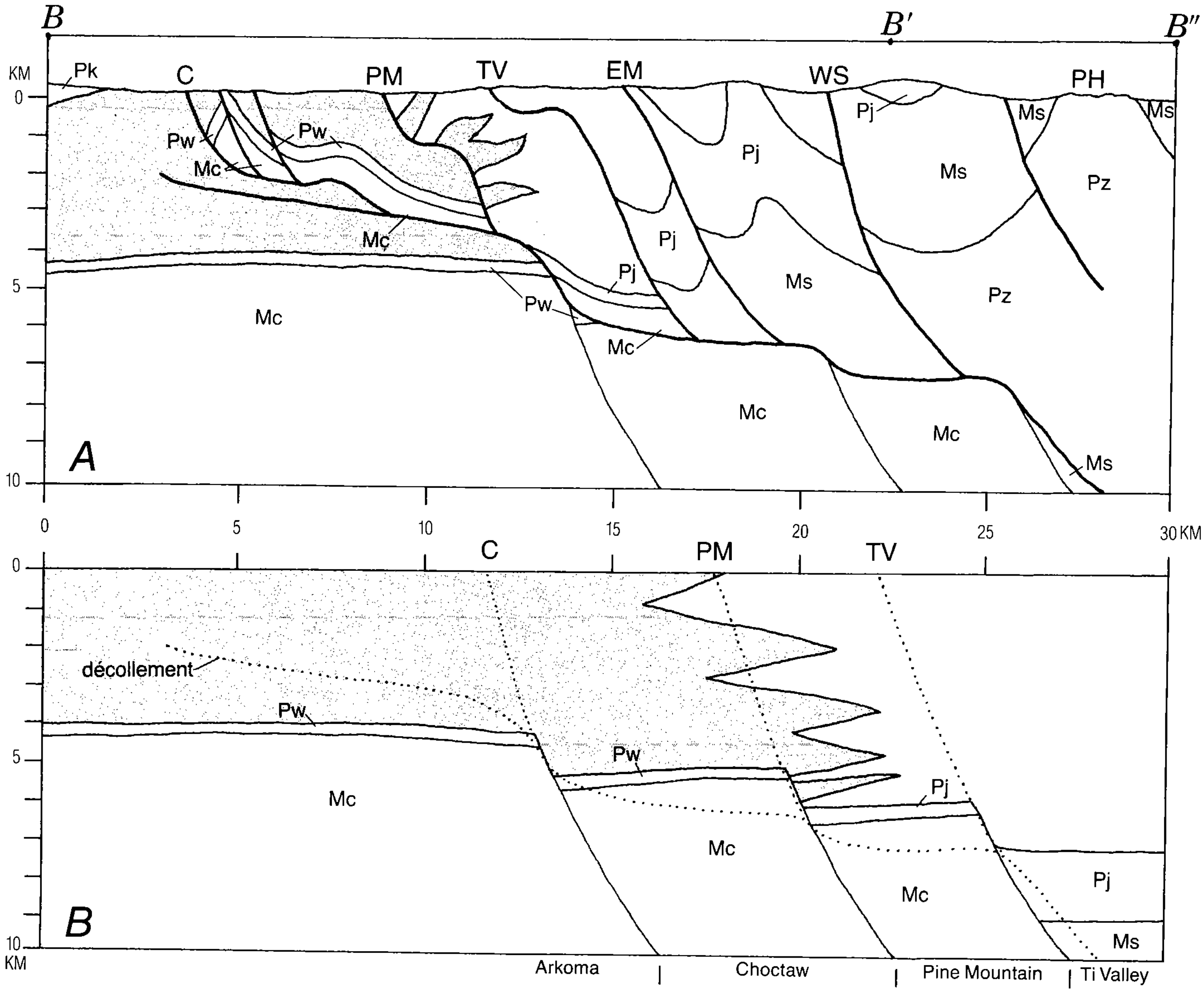


Figure 11. Two generalized geologic cross sections, Ouachita Mountains frontal belt. Points B′ and B″, anchored to basement in both sections, refer to the south end of cross-section lines in Figures 5 and 10. *A*, present configuration of the frontal belt (from Fig. 5), *B*, Atokan reconstruction of the same region (from Fig. 10) showing future décollement surface and original location of some high-angle reverse faults (dotted lines). E-directed and W-directed Atoka Formation strata are shown with heavy and light stipples, respectively. Mc = all shelf-facies rocks below and including the Caney shale. Pz = all pre-Stanley deep-basin-facies rocks. C = Choctaw fault. PM = Pine Mountain fault. TV = Ti Valley fault. EM = Eightmile Mountain fault. WS = Windingstair fault. PH = Potato Hills. Other symbols explained in Figure 5.

ity of a discrete E-directed system farther to the west cannot be excluded.

New paleocurrent data indicate that a source for flysch sediment in the northernmost frontal belt (Choctaw block) lay somewhere to the west of the present Ouachita orogenic belt. The most probable source terrain was an Arbuckle highland. This interpretation is supported by biostratigraphic and structural relationships established along the northeast flank of the Arbuckle uplift.

A structural model, based primarily on interpretations of new paleocurrent data, suggests a relationship between high-angle-reverse-fault blocks of the frontal belt and growth-fault blocks that formed along the north edge of the Ouachita basin in the Pennsylvanian.

ACKNOWLEDGMENTS

Field work for this study was partially supported by the U.S. Geological Survey COGEOMAP Program. Thanks to Gerald Ross, Jim Sears, and Don Winston for helpful discussions.

APPENDIX

LOCATIONS OF PALEOCURRENT MEASUREMENTS IN THE SOUTHWEST FRONTAL BELT

Paleocurrent measurements in the area of detailed study are precisely located on mylar overlays that accompany the detailed geologic maps of the area. These maps are available in open-file form at the Oklahoma Geological Survey.

Paleocurrent rose diagrams from nine exposures of Atoka Formation in the southwest Ouachita Mountains frontal belt (west of the detailed study area) are presented in the main part of this paper (Fig. 8). The location of each exposure and an estimate of the stratigraphic position (relative to the Wapanucka Formation or Johns Valley Shale) are provided here. The exposures are listed from east to west, first from the Choctaw block and then from the Ti Valley block. Most of the paleocurrents were measured from directional flute molds, but some nondirectional sole markings were assigned a directional sense from ripple bedforms in the same bed. These azimuths are designated with an "r" in this appendix.

1) Complexly folded, thin- to medium-bedded sandstones exposed in first road cut east of a distinctive ridge of Hartshorne Sandstone (Desmoinesian) along State Highway 1, ~2 mi east of Hartshorne in the Hartshorne SE 7.5′ Quadrangle, Pittsburg County (SW¼NE¼ sec. 19, T. 4 N., R. 17 E.). Five paleocurrents were measured from a 30-m-thick interval along the north side of the road. This is just south of the Choctaw fault and is probably <600 m above the Wapanucka Formation. Azimuths: 50,50,64,90,108.

2) S-dipping sequence of thin- to medium-bedded sandstones exposed along gravel road in the Hartshorne SE 7.5′ Quadrangle, Pittsburg County (SE¼ sec. 35, T. 4 N., R. 16 E.). Six measurements from a 30-m-interval near the base of the hill in the borrow ditch on the east side of the road. This sequence starts ~125 m above the Wapanucka Formation. Azimuths: 40,40,45,45,47r,57r.

3) S-dipping sequence in large road cut along west side of a frontage road for the Indian Nation Turnpike in the Pittsburg 7.5′ Quadrangle, Pittsburg County (NE¼NW¼NE¼SW¼ sec. 32, T. 3 N., R. 15 E.). Ten measurements from a 150-m-thick sequence of mostly medium-bedded sandstones at the south end of the road cut. Probably 600–1,000 m above the Wapanucka Formation. Azimuths: 80r,82,86r,89, 89r,103r,105r,115,175r,186.

4) S-dipping sequence exposed on the south side of a drill pad cut on top of Pine Mountain in the Pittsburg 7.5′ Quadrangle, Pittsburg County (SE¼SE¼NW¼ sec. 7, T. 2 N., R. 14 E.). Eight measurements from a 15-m-thick sequence of medium-bedded sandstones, 350–600 m above the Wapanucka Formation. Azimuths: 44r,44r,47r,50r,50r,54,65r,71r.

5) S-dipping sequence exposed in major canyon in the south-central part of sec. 22, T. 2 N., R. 13 E. in the Kiowa 7.5′ Quadrangle, Atoka County. Ten measurements from a 200-m-thick sequence of medium-bedded sandstones that starts at the 900-ft-contour level in the canyon, ~400 m above the Wapanucka Formation. Azimuths going upstream from the 900-ft contour: 28r,15,35,30,7,37,27,15,9,22. Five more azimuths from a correlative sequence along the ridge northeast of the canyon: 22,28,32,31,20.

6) A predominantly S-dipping but complexly folded sequence in the major northwest-flowing canyon in sec. 7, T. 1 N., R. 13 E., Limestone Gap 7.5′ Quadrangle, Atoka County. It is difficult to estimate the thickness of the measured interval, because of the folding, which appears to conform to the course of the canyon. It could be as thin as 300 m, but it could be >700 m. The interval starts ~450 m above the Wapanucka Formation. Thirteen measurements are from medium- to thick-bedded sandstones, starting at the 840-ft-contour level and going upstream: 40,357, 7r,330,22r,32,24,220,260,25r,15,290.

7) A sequence exposed in the core of a nearly isoclinal E–W-trending syncline in the Counts 7.5′ Quadrangle, Pittsburg County (SW¼ sec. 26, T. 3 N., R. 16 E.). Five measurements from a 600-m-thick sequence of medium-bedded sandstones exposed in a major south-flowing canyon. The lower end of the sequence is ~350 m above the Johns Valley Shale. Azimuths from north to south: 272,274,284,249,252.

8) A S-dipping sequence (N½ sec. 16, T. 2 N., R. 15 E.) from the north limb of the same syncline described for location 7. Five measurements are from a 120-m-thick sequence of medium-bedded sandstone ~150 m above the Johns Valley Shale. Azimuths: 250,245,251r,232r,256. The first two of these measurements are from a distinct meander bend in a major north-flowing stream. The other three are from the west road cut of the Indian Nation Turnpike, ~0.5 km east and along strike from the meander bend. Two more measurements (azimuths 232,258) are from the east side of the next road cut to the south, which is probably 450 m above the Johns Valley Shale.

9) This is from a S-dipping overturned syncline exposed in Chickasaw Creek in the Limestone Gap 7.5′ Quadrangle, Atoka County (N½ sec. 2 and N½ sec. 3, T. 1 S., R. 13 E.; and SW¼ sec. 35, T. 1 N., R. 13 E.). Seven measurements are from a 400-m-thick sequence of medium- to thick-bedded sandstones that starts ~200 m above the Johns Valley Shale. Azimuths: 204,263,228,252,246,220,283.

ARKOMA BASIN OVERVIEW

W. David Wylie

INTRODUCTION

This overview has been adapted primarily from Branan (1966); in my opinion, his paper provides the best overview of geology and petroleum development for the Arkoma basin novice, although more-recent drilling activity and published regional studies have better defined these trends.

The Arkoma basin is situated in southeastern Oklahoma and west-central Arkansas. It is essentially a dry-gas province (Fig. 1) where more than 25 zones, ranging in age from Middle Pennsylvanian (Krebs Group) to Early Ordovician or Late Cambrian (Arbuckle Group), are commercially productive. Production comes from depths of 1,500–15,000 ft. In-place reserves range from several hundred million cubic feet of gas to 60 billion cubic feet of gas per well.

The arcuate basin measures ~250 mi in an east–west direction; it was compressed during the Ouachita orogeny to its present width of 20–50 mi. The basin sediments are thickest along its southern margin, near the Ouachita Mountains system, where as much as 30,000 ft of sedimentary rocks may exist as a result of rapid deposition during basin subsidence and repetition of strata by thrust faulting. The cross section of Wylie and others (this volume) shows the major structures and stratigraphic units in the Arkoma basin along the transect of the field trip.

GEOLOGICAL HISTORY

Prior to the post-Hunton orogeny, the present sites of the Arkoma basin and Ouachita Mountains were part of the eastern flank of the Oklahoma embayment (the Oklahoma basin, as described by Johnson in this volume). During much of the Late Cambrian and Early Ordovician, an extensive transgression resulted in the deposition of shallow marine carbonates and minor sandstone units of the Arbuckle Group throughout the Midcontinent.

Following the emergence of the craton during the Early Ordovician and a corresponding influx of fine-grained clastics from the north, thick sandstone units of the Simpson Group (Middle Ordovician) were deposited throughout the Arkoma region. Holden (1965, p. 137) reported a northeastern source for the Simpson sand and a shoreface depositional environment "bordering the emergent shelf of northern Oklahoma, Kansas and western Missouri . . . around and over a subsiding shelf north of the Ouachita geosyncline." Isopach work by Holden in the same report indicates that the Simpson Group increases in thickness "southward and ranges from 250 ft in east-central Oklahoma to more than 1,500 feet in White County, Arkansas" (Holden, 1965, p. 134–135, fig. 1). Suhm (1979) outlined the paleogeographic environments of the Simpson units in Arkansas and discussed their hydrocarbon potential in stratigraphic and fracture-induced traps.

Following deposition of the Simpson Group, marine carbonate and shale characterized sedimentation in the Arkoma region. Rock units included in this sequence include the Viola (carbonate), Sylvan (shale), and Hunton (carbonate). An excellent, detailed study of the Hunton Group in eastern Oklahoma has been completed by Amsden (1980).

Tectonic uplift associated with the post-Hunton orogeny or Acadian disturbance transformed the Arkoma region from a marine platform or shelf into a large positive arch during Late Devonian time, resulting in varying degrees of erosion of the Hunton Group. The Hunton was completely removed throughout much of Pittsburg, Latimer, and Le Flore Counties in eastern Oklahoma, as documented in papers by England (1961) and Amsden (1980).

Following the episode of post-Hunton erosion, a shallow Late Devonian–Early Mississippian sea with euxinic conditions inundated most of the Midcontinent and Arkoma region. Black, pyritic Woodford shale succeeded a paleotopographically controlled basal Misener (Sylamore) sandstone. Throughout the remainder of the Mississippian, warm, shallow seas occupied the Arkoma region, and carbonate sedimentation prevailed. At that same time, deep-water flysch sands and shales of the Stanley Shale were being deposited south of the Arkoma in the Ouachita geosyncline (Wickham, 1978).

Sedimentation was once again interrupted by gentle, widespread emergence and erosion in the Midcontinent and Arkoma region prior to deposition of the Pennsylvanian System. Rascoe and Adler (1983) have summarized the events and association between the pre-Morrowan cratonic epeirogeny, which was centered around the Transcontinental arch, and the subsequent Pennsylvanian sedimentation patterns. Shallow marine, deltaic, and non-marine depositional environments coexisted during the Morrowan in the Arkoma region. Zachry (1979) has documented several south-flowing braided-stream systems in the middle Bloyd sandstone (Morrowan) of northwest Arkansas. This interpretation was derived from detailed outcrop studies of sedimentary structures and paleocurrent measurements of tabular cross-strata.

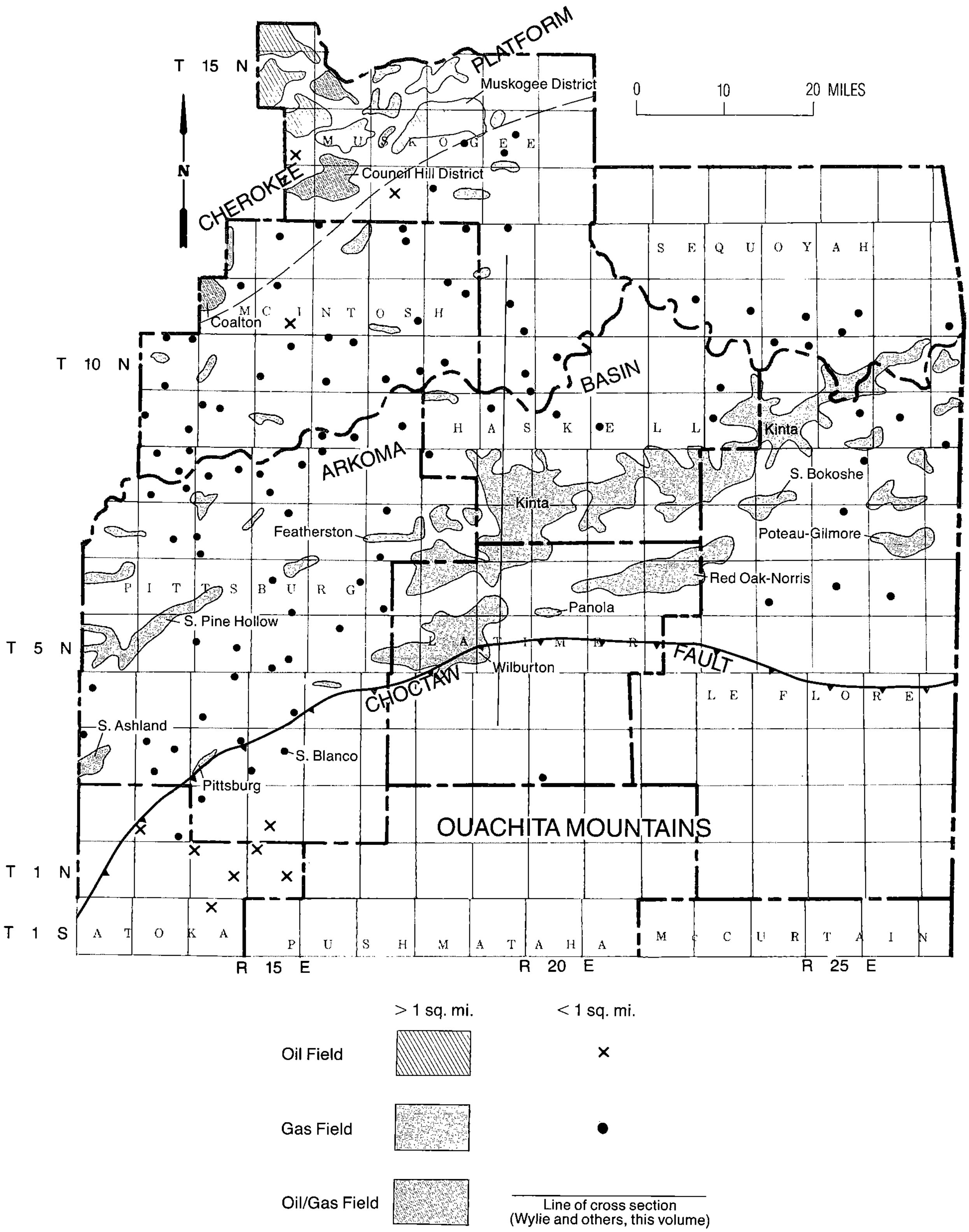


Figure 1. Oil and gas fields in parts of the Arkoma basin and Ouachita Mountains of eastern Oklahoma (from Burchfield, 1985). Some of the major fields are identified.

Rascoe and Adler (1983, p. 980) reported that significant tectonic activity recurred during the Wichita orogeny "as a result of collision between North America and South America plates from late Morrowan through Atokan, and into early Desmoinesian time." They went on to report that the Arkoma basin actually "originated in Atokan time as the southern margin of the North American craton" and was downwarped "progressively from south to north by the overriding South American plate . . . The strong subsidence of the Arkoma basin in middle Pennsylvanian time was accompanied by down-to-the-basin faulting in Southeastern Oklahoma" (Rascoe and Adler, 1983, p. 982).

An unconformity exists along the north-central margin of the basin between the basal Atokan Spiro sand and the underlying Morrowan Wapanucka limestone. Spiro sandstone trends represent deposition in the broad delta complex with distributary channels, tidal channels, and shallow-water, interfingering marine sandstone bars and carbonates. The Spiro is a favorable exploration target, because production has been established in both stratigraphic and structural traps.

Regional studies also indicate that the Spiro was deposited "prior to the major basin downwarping and subsequent thickening of the Atoka section" (Branan, 1966, p. 1627). Consequently, the Spiro is an important reference datum for measuring middle Atokan growth faulting, regional mapping of the total thickness of Atokan sediments, and constructing a post-Atokan and post-Ouachita structural setting of the basin. This latter point is enhanced by the fact that the Spiro is underlain by an excellent seismic reflector, the Wapanucka limestone.

Most of the Atokan clastics in the eastern part of the basin appear to have been deposited in deltaic fashion. Haines (1981) described six distinct middle Atokan deltaic sequences from the subsurface of north-central Arkansas. "They are all variable in lateral extent, thickness and depocenter location, but all have a north–east source and broad distribution along the shelf" (Haines, 1981, p. 42).

Middle Atokan sandstone units are also developed in the eastern Oklahoma portion of the basin. These units are generally developed parallel to the basin axis on the downthrown side of the major middle Atokan growth faults and may be deep-water-channel, fan, or turbidite deposits correlative with the Arkansas deltaic sequences (Rascoe and Adler, 1983, p. 988). Subsurface mapping by the present author indicates that some of these units may have also received some sediments from several point sources along the E–W-trending growth faults.

Recent studies of Atokan turbidite outcrops south of Wilburton (Ferguson and Suneson, this volume) indicate mixed paleocurrent directions on successive thrust sheets within the Ouachita frontal belt.

An accurate measurement of the maximum thickness of the Atokan has been greatly complicated or masked by the folding and/or thrust faulting of the interval adjacent to the Ouachita Mountains system. Estimates ranging from 10,000 ft (Rascoe and Adler, 1983, p. 988) to >20,000 ft of Atokan rocks (Branan, 1966, p. 1619) have been reported.

Significant structural deformation and/or uplift associated with the Ouachita orogeny began in late Atokan time and continued more dramatically into the Permian. The paleogeography and paleogeology of the Midcontinent at the close of the Atokan have been diagrammatically depicted by Rascoe and Adler (1983, p. 990, fig. 6). Rapid subsidence and growth faulting concluded at the end of the Atokan, and the Arkoma basin took on a semi-continental setting with vast swampy lowland areas transected by several east–west fluvial systems during the Desmoinesian. These systems are represented by the thick channel sandstones and coal measures that are found in the Krebs Group.

Uplift of the Ouachita Mountains during the Permian transformed much of the basin area into a positive land area subject to erosion, and the Ouachita fold belt became the principal source of clastics in the southern Midcontinent.

Branan (1966, p. 1623) reported that the Arkoma basin was later tilted westward, "presumably by the later Appalachian movement," and also during "late Permian time and during the Cretaceous Laramide orogeny which had minor influence." These final tectonic adjustments concluded significant geological activity in the Arkoma basin and left the area physiographically similar to its present state.

HISTORY OF PETROLEUM DEVELOPMENT

Bartlett (1966, p. 126) reported that the first known test well in the Arkoma basin was drilled to 1,400 ft in 1887 by Choctaw Oil & Gas Co. near downtown Fort Smith, Arkansas. Bartlett also reported that several more drilling attempts were unsuccessful, and it was not until 1902 that commercial quantities of natural gas were discovered in the basin at Mansfield, Arkansas, 25 mi south of Fort Smith. The Choctaw Oil & Gas Co. #2 Duncan tested a well-expressed anticline and was completed at 1,125 ft, flowing 960 MCFG/D from an upper Atoka sandstone (Bartlett, 1966, p. 126).

Branan (1966, p. 1619) noted that "the surface exploration for coal seams" (Krebs Group) provided surface-structure maps of anticlines and encouraged early wildcatters to drill for oil and gas.

Discoveries in eastern Oklahoma began with the Poteau-Gilmore Field on such a structure in Le Flore County (Fig. 1). The 1910 discovery was the Le Flore Gas and Electric Co. #1 Hill. It was drilled to 1,687 ft and completed in the Hartshorne sandstone (basal Krebs Group) for 4,500 MCFG/D (Bartlett, 1966, p. 126).

Shallow drilling at 2,500–5,500 ft on well-defined surface structures for Hartshorne and Atoka gas zones continued until 1950. Bartlett (1966, p. 126) stated that "in 1950 only 37 wells were drilled in the entire basin. By 1960, this had risen to 85 and in 1964, 115 wells were drilled in eastern Oklahoma, 63 in northwestern Arkansas."

Deeper drilling at 10,000–13,500 ft in the basin was encouraged during the 1960s when Midwest Oil Corp. and Frankfort Oil Corp. discovered and delineated the Red Oak-Norris Field in portions of T. 6–7 N., R. 20–23 E., Latimer County, Oklahoma (Fig. 1). Here the middle Atoka Red Oak and basal Atoka Spiro sandstones are productive and were penetrated at depths of 7,000–9,000 ft and 11,000–13,200 ft, respectively. This area previously had been developed since 1912 along the surface-expressed Brazil anticline for shallow Hartshorne gas (Six, 1966).

Delineation of the deeper production revealed a number of subsurface thrust faults which caused repetition of the Red Oak interval (Six, 1966, p. 1648).

Deeper drilling was also encouraged by the practice of air/gas drilling (Branan, 1966, p. 1630). This technique reduced drilling costs greatly by eliminating excessive mud expenditures and increasing penetration rates. It also provided a continuous drill-stem test by allowing productive zones to flare while drilling.

Amoco Production Co. is currently completing a 42-well infill drilling program in the Red Oak-Norris Field (Oil and Gas Journal, 1988, p. 30). The 96-well field has produced more than 550 BCFG since its discovery in 1959, as Red Oak and Spiro combined reserve potential ranges from 30–60 BCFG per 640-acre spacing unit. Amoco also reduced drilling time and costs for the 12,000-ft wells by drilling with air to 8,000 ft, then mudding up with an oil-base mud to control Red Oak gas flow while drilling onto the Spiro. These two steps allowed initial rapid penetration rate and eliminated the necessity of running intermediate casing to control the Red Oak—a savings of $600,000 (Oil and Gas Journal, 1988, p. 30).

Deep drilling also meant success in the Wilburton area during the 1960s. Here, shallow production had been established along the Adamson and Wilburton anticlines. However, in 1960 Ambassador Oil Corp. completed the #1 Williams (sec. 23, T. 5 N., R. 18 E.) for CAOF 8,300 MCFG/D, SIBHP 4629#, as a Spiro discovery. The well had a total depth of 9,704 ft (Berry and Trumbly, 1968, p. 93). Following this discovery the field was extended to 65 gas wells by 1968.

As at Red Oak-Norris, delineation of the Wilburton Field revealed a complex subsurface of folding and imbricate thrust faulting of the middle and lower Atoka strata. Most significant are the Carbon and Choctaw thrust faults, which are exposed at the surface (Wylie and others, this volume).

The Red Oak and Spiro are also the primary producing zones at Wilburton. In portions of the field, they are dually completed for reserves which range from 5 to 40 BCFG per well per zone or individually in repeated thrust sections. While the folding and thrusting cause correlation problems, reserve potential is greatly enhanced by the repetition of objectives and development of secondary porosity and permeability through micro-fracturing.

The Wilburton Field has undergone infill drilling by various operators in recent years. Here too, the activity is encouraged by increased deliverability and reserve potential from better Red Oak and Spiro structural or stratigraphic positions, as well as better sales contracts.

Exploration excitement has also been generated this past year in the Wilburton Field (Fig. 1). Arco Oil and Gas has released details of an Arbuckle discovery at the #2 Yourman (sec. 15, T. 5 N., R. 18 E.). The 15,391-ft deeper pool discovery established a potential test of 9,309 MCFG/D, 20/64″ ck, FTP 3140#, from the Arbuckle interval, 14,259–14,500 ft (Petroleum Information, 1988). Nearest Arbuckle production lies 30 mi to the northeast at Red Oak-Norris, where Amoco flowed 5,451 MCFG/D from 13,302–13,310 ft at its #2 Roy Reed "C" unit (sec. 4, T. 6 N., R. 23 E.) discovery (Petroleum Information, 1988). Since these news releases, Arkoma explorationists have been eagerly reviewing seismic and well data in search of more Arbuckle structural features and porosity development.

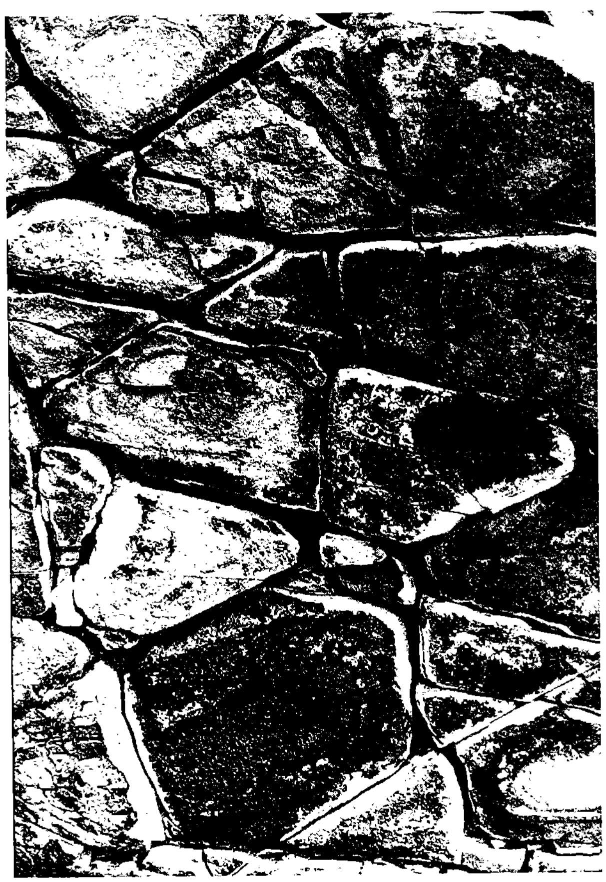
The drilling boom of the late 1970s and early 1980s broadened exploration in both the shallow and deeper portions of the basin. A number of shallow Spiro and Hunton discoveries were made along the northern margin of the basin (McCaslin, 1982). This area had previously been thought to be beyond the natural-gas migration or trapping limits of the basin.

In contrast to this shallow activity, production trends at Wilburton were extended south of the historical structural boundary of the basin, the Choctaw fault. The Pittsburg Field complex (T. 2–3 N., R. 13–15 E.) produced discoveries in the Cromwell, 2,200 MCFG/D (1979); Woodford–Hunton, 986 MCFG/D (1981); and Spiro–Wapanucka (1982) (Richardson, 1987).

As at Wilburton and Red Oak-Norris, reserve potential exists in complexly folded and thrust-faulted strata, particularly repeated Spiro–Wapanucka sheets, at depths to 13,000 ft.

In the future, the Arkoma basin should continue to provide success for the explorationist. The northern margin of the basin is relatively unevaluated, particularly in the Hunton, Viola, Simpson, and Arbuckle strata. Many structures that are currently producing from Morrowan, Atokan, and Desmoinesian zones have not been sufficiently drilled deeper to evaluate the lower Paleozoic section. Finally, large expanses of the basin in east-central Arkansas have not been drilled. Density of wildcat wells averages less than 1 well per township (36 mi^2) east of the Jerusalem Field in Conway County, Arkansas. However, explorationists will have to continuously risk the rewards of repeated-zone or multiple-zone potential against the secrets of the basin's structural and stratigraphic complexities.

FIELD-TRIP
STOP DESCRIPTIONS



Conjugate joints in Atoka Formation sandstones in the Ouachita Mountains frontal belt, Oklahoma.

STOP 1
SECOR COAL IN POLLYANNA NO. 5 STRIP MINE, WEST OF MUSKOGEE

LeRoy A. Hemish

LOCATION

The Pollyanna No. 5 Mine is operated by P&K Company, Ltd., in sec. 34, T. 15 N., R. 17 E., Muskogee County, Oklahoma (Fig. 1). It is ~5.5 mi west of the junction of U.S. Highway 69 with U.S. Highways 62 and 64 and State Highway 16 in Muskogee.

DESCRIPTION

The first stop on the field trip features an operating strip pit where the Secor coal is being mined. A maximum of 50 ft of overburden is being removed. Strata of the lower part of the Boggy Formation, overlain by ~4 ft of Quaternary deposits, are exposed in the highwall of the mine (Fig. 2). A thin coal seam, ~2 in. thick, can be observed in the highwall ~24 ft above the Secor coal. This thin coal is the Secor rider. It is overlain by a persistent, 2-in.-thick, impure, fossiliferous, marine limestone, which in turn is overlain by a thick section of grayish-black, noncalcareous, brittle shale containing rare fossils and ironstone concretions. The interval between the Secor and Secor rider coal beds is composed mostly of silty shale interlaminated with very fine-grained sandstone. This interval is particularly well known for its assemblage of abundant, well-preserved plant fossils. The abundance of the fossils and the genera

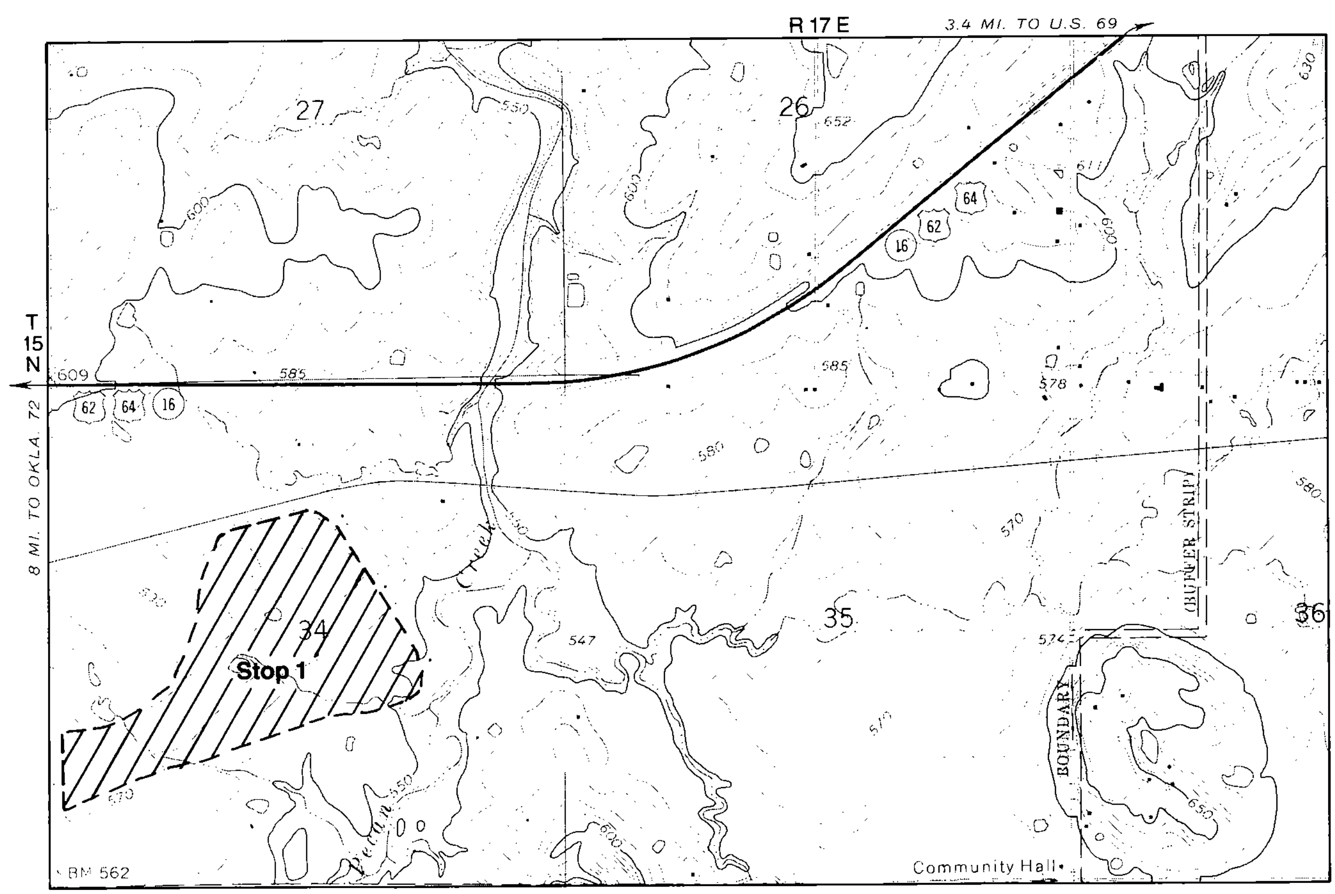


Figure 1. Location of Stop 1 at the Pollyanna No. 5 Mine, 5.5 mi west of Muskogee, in Muskogee County. Base from Southwest Muskogee 7.5′ Quadrangle.

SYSTEM	SERIES	GROUP	FORMATION	THICKNESS (ft)	DESCRIPTION
					Note: Beds dip S. 40 E. at 4°
QUAT.	UNDIF.			1.3	Silt, dusky-brown, unconsolidated; contains organic material (topsoil).
				2.7	Clay, dark-yellowish-brown to light-brown with moderate-reddish-brown and blackish-red mottling, pisolitic, sticky, gravelly, weathered.
PENNSYLVANIAN	DESMOINESIAN	KREBS	Boggy	4.0	Shale, brownish-black with light-brown iron oxide staining on stratification surfaces, noncalcareous; includes brachiopod fossils on some bedding planes; contains some dark-yellowish-orange ironstone concretions.
				7.0	Shale, grayish-black, noncalcareous, brittle; includes trace fossils and rare brachiopod valves on stratification surfaces; contains scattered layers of brownish-gray ironstone 1 – 2 in. thick.
				0.2	Limestone, medium-dark-gray, silty, fossiliferous; contains fossil hash and brachiopods.
				0.2	Shale, black, carbonaceous, pyritic.
				0.2	Coal, black, bright, moderately friable; cleats closely spaced (Secor rider coal bed).
				0.7	Shale, black, coaly.
				0.5	Underclay, medium-dark-gray; contains some very thin laminae of bright coal.
				22.3	Shale, medium-light gray, very silty, noncalcareous; shaly siltstone interlaminated with very fine-grained sandstone in part; contains abundant well-preserved black carbonized plant compressions.
				1.0	Coal, black, bright, moderately friable; contains minor pyrite on cleat surfaces (Secor coal bed).
				1.3	Underclay, medium-dark-gray; contains abundant black carbonized plant compressions; slickensided.

Figure 2. Stratigraphic section exposed in highwall at the Pollyanna No. 5 Mine (OGS measured section MM-2-88-H, NW¼SE¼NE¼SW¼ sec. 34, T. 15 N., R. 17 E.).

present indicate a tropical, swamp-forest environment. Carbonized compressions of *Calamites, Lepidodendron, Stigmaria, Sigillaria, Annularia, Pecopteris, Alethopteris*, and *Neuropteris* have been collected from this interval in strip mines around Porter, Oklahoma, ~10 mi north of the Pollyanna No. 5 Mine. Similar assemblages can be collected from the strata above the Secor coal at this mine.

Both the Secor rider and Secor coal beds were sampled by the writer in the Pollyanna No. 5 Mine in January 1988. Analysis of the coal samples was done by the Oklahoma Geological Survey Inorganic Chemistry Laboratory. Table 1 shows analytical data for the Secor rider and Secor coals. The Secor rider has a higher ash content (11.2%), and a higher sulfur content (3.9%) than the Secor coal, which has an ash content of only 2.0%, and a sulfur content of only 0.5%. The Secor rider coal is not of minable thickness in Muskogee County, so the Secor coal is the only bed mined at this site. Because of its superior quality (note the heat value of 14,530 Btu/lb), the Secor coal is a much sought-after product.

The first cut of the Pollyanna No. 5 Mine was opened in January 1987 in the SW¼ sec. 34, T. 15 N., R. 17 E., Muskogee County. Strata dip S. 40° E. at ~4°, but the dip is not uniform and generally steepens near the south end of the cut. The cut extends from the outcrop boundary of the Secor coal on the north in a southerly direction to a point where the thickness of the overburden exceeds 50 ft; excavation ceases at that point for economic reasons.

Most of the overburden is removed with bulldozers, but a dragline is also brought into use at the deep end of the cut. The overburden is "shot" prior to removal to break the consolidated rock and to facilitate handling. Figure 3 shows some of the equipment used in the mine.

TABLE 1.—COAL ANALYSES

	Secor	Secor rider
Proximate		
Ash	2.0	11.2
Volatile matter	36.4	40.2
Fixed carbon	58.6	45.7
Sulfur		
Total	0.5	3.9
Pyritic	0.07	3.20
Sulfate	<0.01	0.01
Organic	0.5	0.7
Gross calorific value (Btu/lb)	14530	13026
Chlorine	0.28	0.13

Analyses on as-received basis.
Values other than calorific value in wt.%.
Determinations by K. Catto, J. Crawford, P. Hanley, S. Weber.

Figure 3. View of the Pollyanna No. 5 Mine, showing part of the strip-mine operation as it appeared in January 1988. At left is a bulldozer pushing aside the unconsolidated surface material; at right is a shot-hole rig drilling blast holes in consolidated overburden; and in the center is a dragline removing rock overburden that has been blasted.

Production from the mine has averaged ~6,500 ton/month. The mine is in operation six 10-hr days each week, and generally provides employment for 15 men. At the time the writer visited the mine early in 1988, the coal was being shipped from the mine in trucks. The coal was used for generating electric power when the mine originally opened, but has since gone for other purposes. It is often blended with an inferior coal to enhance the overall quality.

The first step in physically opening a new mine is the removal and stockpiling of topsoil from the permitted area. In the P&K 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. Anticipated completion of the mining operation at this site is early in 1989.

STOP 2
BARTLESVILLE–BLUEJACKET SANDSTONE AT EUFAULA DAM

Glenn S. Visher

LOCATION

This stop includes three sites in the vicinity of Eufaula Dam on the Canadian River (Fig. 1). The three sites are in secs. 25 and 35, T. 10 N., R. 18 E., in McIntosh and Haskell Counties. Eufaula Dam is crossed by State Highway 71 ~5 mi north of the town of Enterprise.

DESCRIPTION

Site 2A is 0.4 mi northwest of the dam where the top of a hill has been quarried. The stratigraphic interval is >125 ft thick, and is developed on the margin of the principal southeast distributary of the Bartlesville–Bluejacket delta. Depositional facies include pro-delta carbonaceous siltstone and fine-grained sandstone laminae and thin bedding units. Channel-margin and basal units represented by shale-clast conglomerates are illustrated in Figure 2. In other areas of the exposure the channel transition is gradational, with thinly laminated siltstones and carbonaceous shales (Fig. 3). The principal section is a thin-bedded to macro-bedded channel-sandstone interval with high lateral heterogeneity (Fig. 4). Bedding units appear to be lenticular, suggesting that this section is nearly perpendicular to the channel axis.

This exposure represents a complex history of channel fill. Subsurface log responses to the northwest, and another channel outcrop, more than 20 mi southeast, reflect the extent of channel extension and progradation of the delta platform. This interval was deposited below sea level, and represents channeling to a depth >125 ft into the delta platform. Fluid-flow units within producing reservoirs would be lenticular and highly variable in shape, with a geometry of a few tens of feet wide and a few hundred feet long.

Site 2B, at the north abutment of the dam, appears to be a more nearly longitudinal section within the distributary channel. It may also represent a section nearer to the channel axis. Bedding units are more continuous, and contain fewer major shale–siltstone breaks.

Site 2C, 0.2 mi southwest of the dam, comprises depositional units adjacent to the major distributary seen in Sites 2A and 2B. The basal unit shows a fining-upward interval, and probably represents a crevasse splay deposit. The ripple-laminated, flaser-bedded, middle interval is typical of bay-fill sequences. The uppermost unit may be a complex tidal-flat–levee depositional sequence. Mapping of the geometry of these facies would be essential to paleogeographic interpretation of the delta platform in this area.

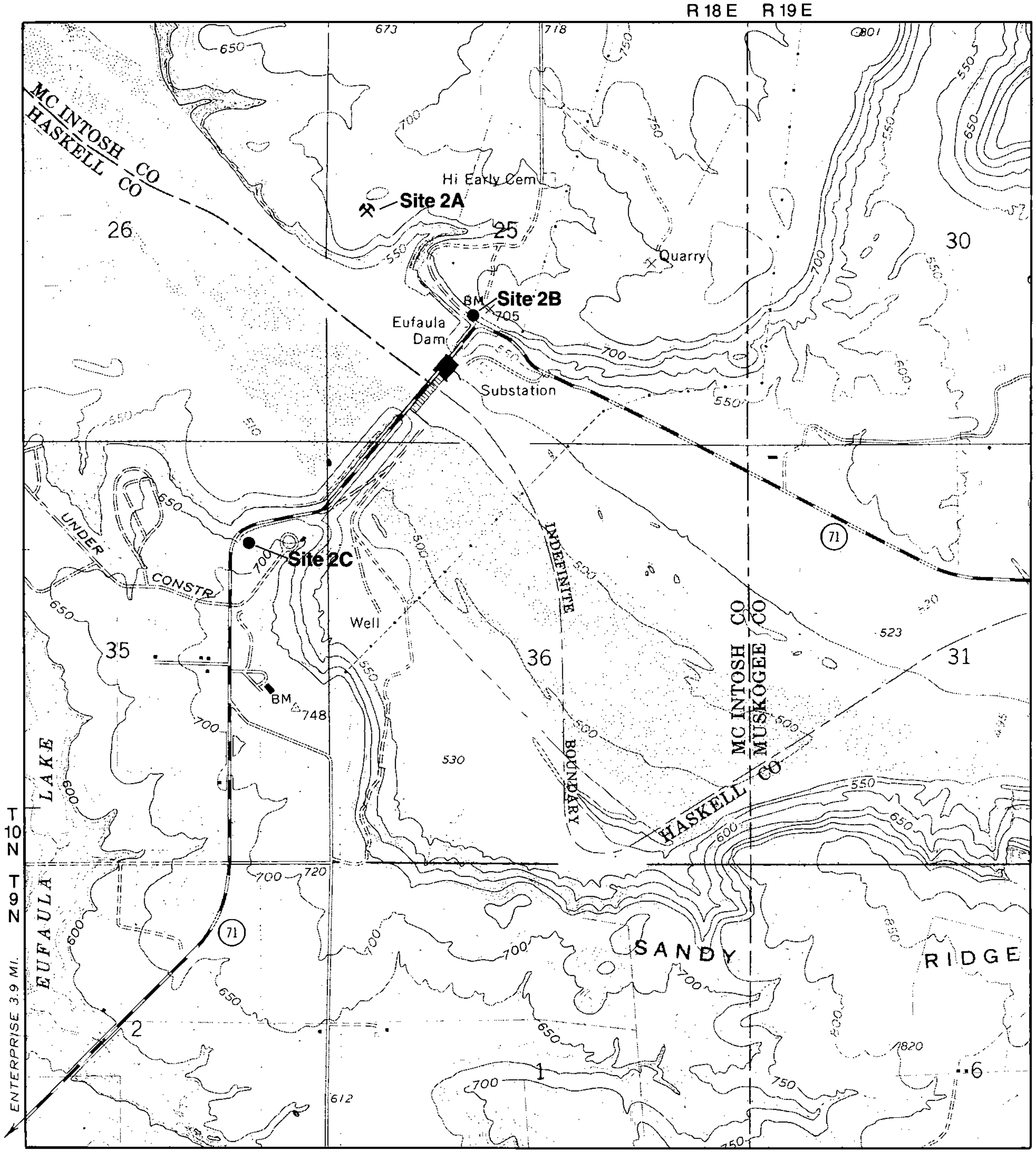


Figure 1. Location of Stop 2 (Sites 2A–C) at Eufaula Dam, ~5 mi north of Enterprise, McIntosh and Haskell Counties. Base from Porum 7.5′ Quadrangle.

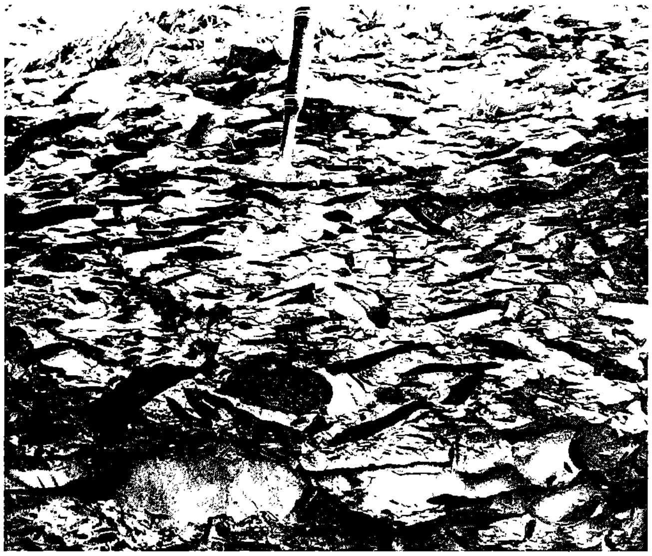


Figure 2. Shale-clast conglomerate along the margin and base of some channels of Bartlesville–Bluejacket distributary at Site 2A.

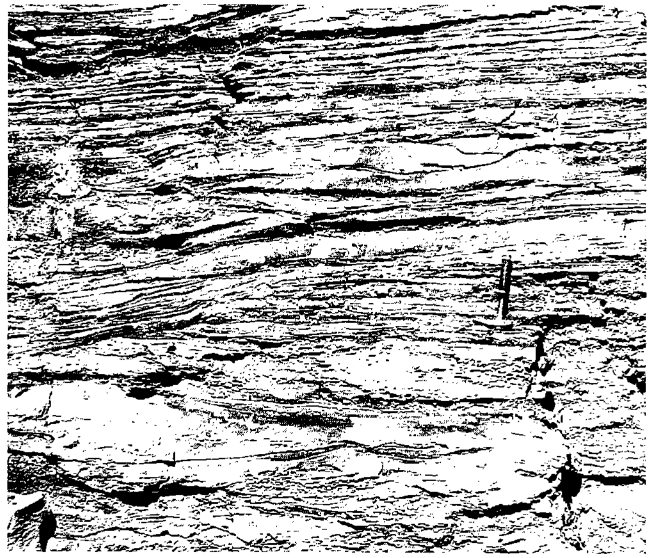


Figure 3. Thin-bedded siltstones and carbonaceous shales in some channels of distributary at Site 2A.

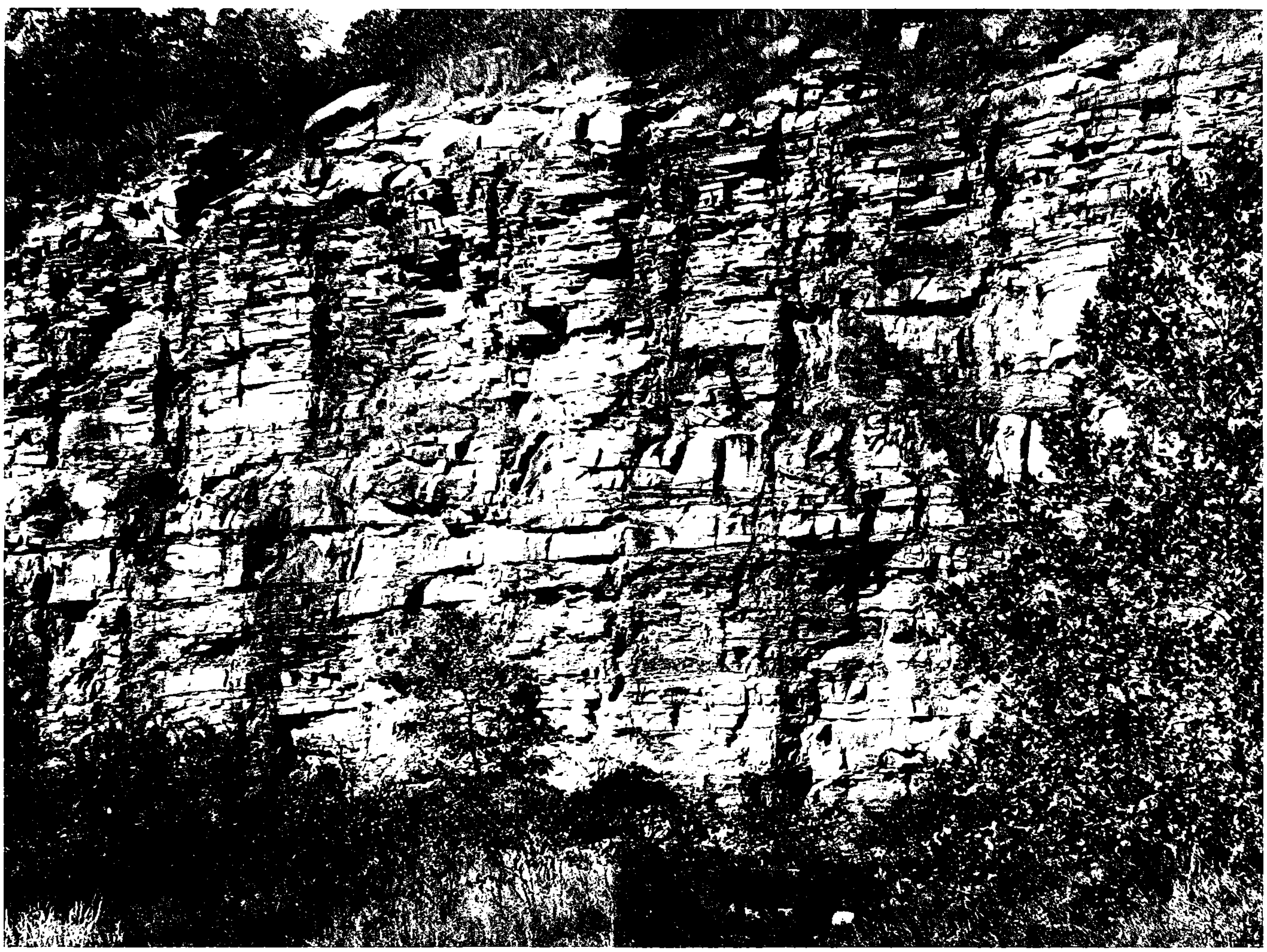


Figure 4. Major exposure at Site 2A comprises thin- to thick-bedded channel sandstones that have a high degree of lateral variability.

STOP 3
STIGLER STONE COMPANY QUARRY AT ENTERPRISE

Glenn S. Visher

LOCATION

The Stigler Stone Co. operates a quarry 2 mi northeast of the town of Enterprise, in sec. 24, T. 9 N., R. 18 E., of Haskell County (Fig. 1). This stop is on the north side of State Highway 9.

DESCRIPTION

This exposure represents the upper portion of channel fill in the Bartlesville–Bluejacket Sandstone. Outcrop geometry suggests that this sequence is part of a distributary-mouth bar. Sedimentary features reflect parallel-bedded, current-laminated and -lineated, fine-grained sandstone (Fig. 2). Internal structures indicate shallow-water, upper-flow-regime sedimentary structures (Fig. 3). Ripple lamination suggests antidune flow conditions (Fig. 4), consistent with a shallow channel cutting into the distributary-mouth bar. A few of the bedding surfaces contain concentrations of carbonaceous fragments and shale clasts (Fig. 5). Such a complex sug-

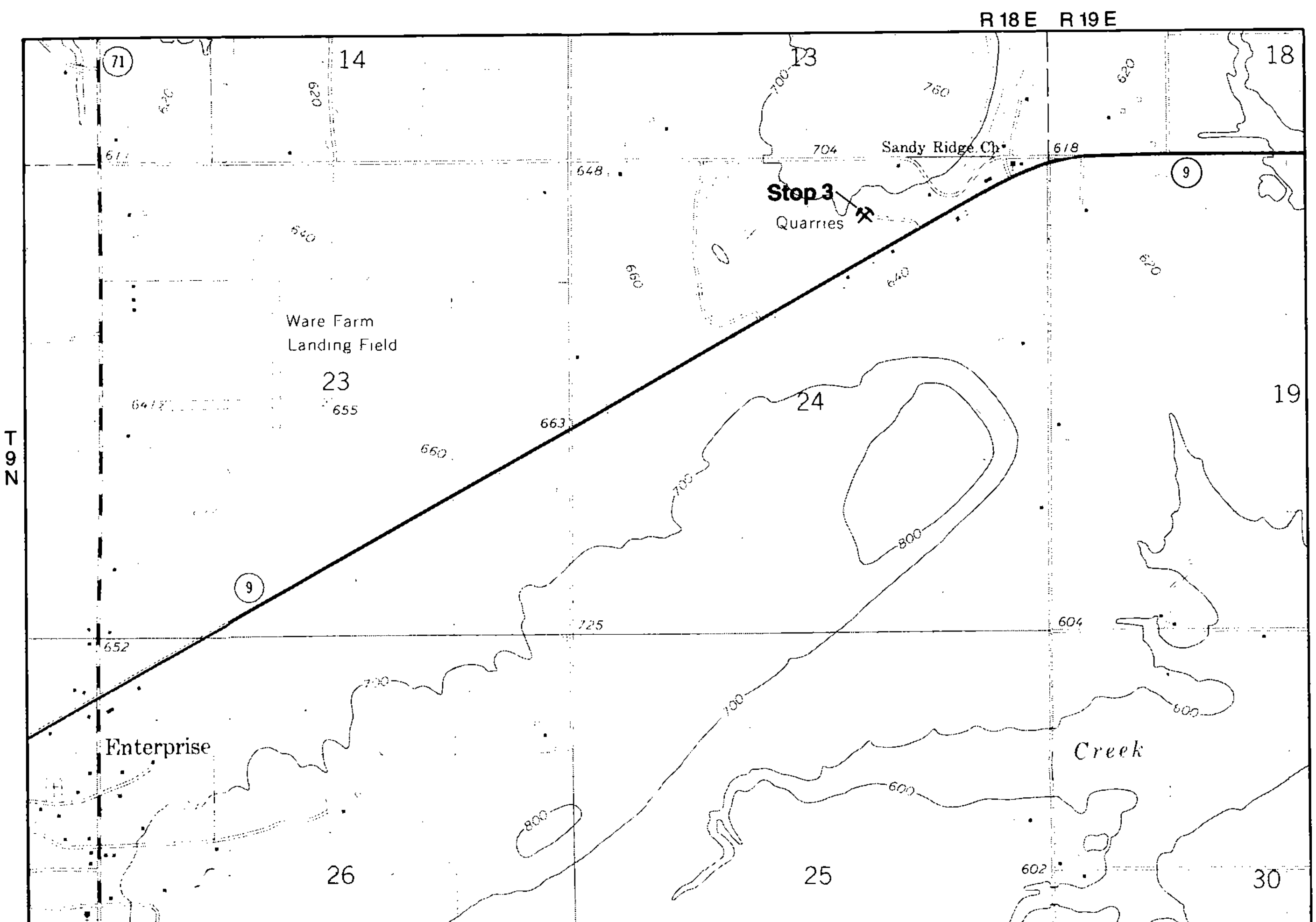


Figure 1. Location of Stop 3 at the Enterprise quarry of Stigler Stone Co., 2 mi northeast of Enterprise, Haskell County. Base from Quinton North and Enterprise 7.5′ Quadrangles.

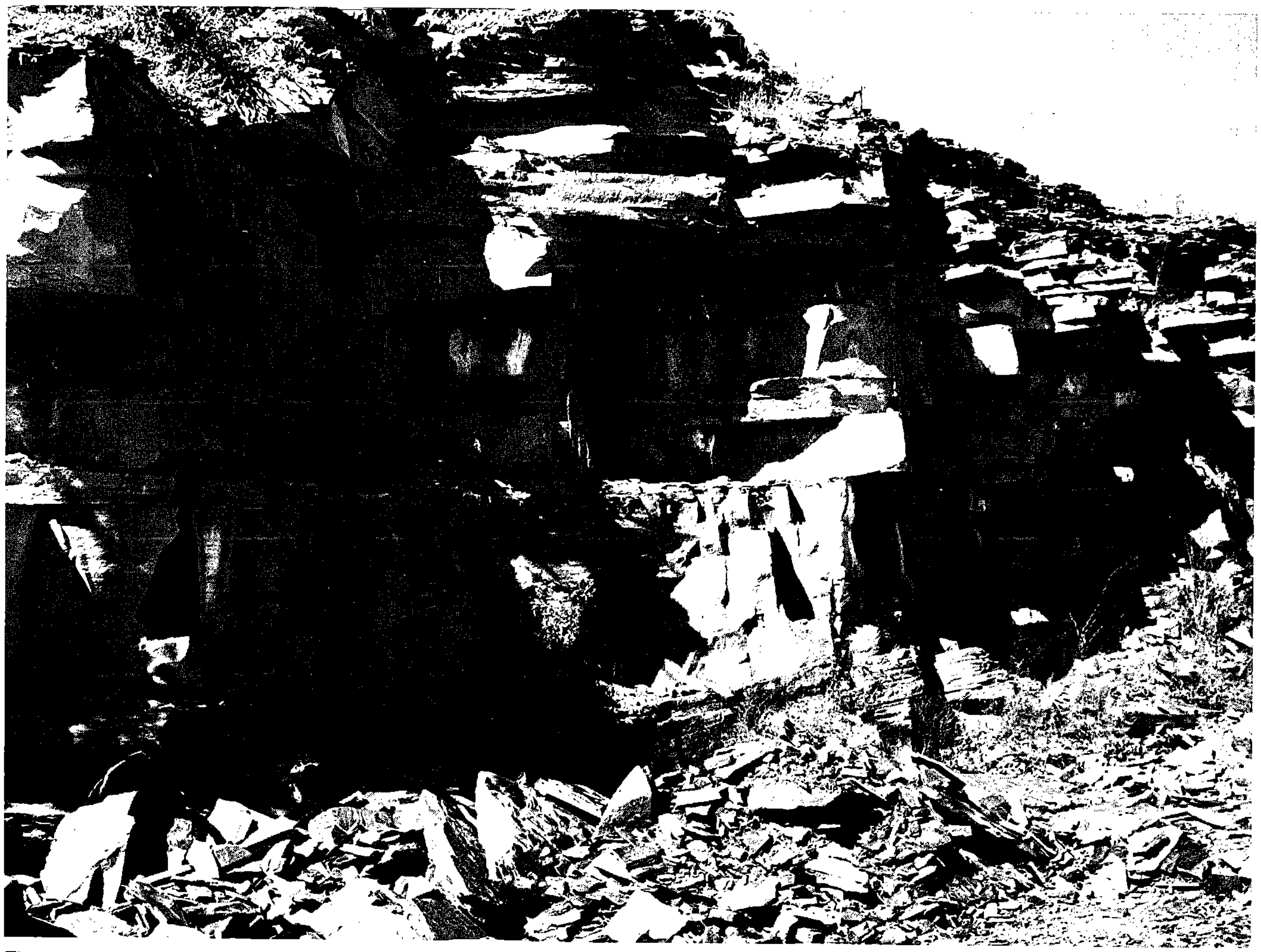


Figure 2. Fine-grained sandstone in quarry is parallel-bedded, current-laminated, and current-lineated.

Figure 3. Sandstones show shallow-water, upper-flow-regime sedimentary structures.

gests confined channel development after deposition of shallow-water sandstone.

This depositional history indicates rapid sedimentation, and then a longer period of maturation and superposition of depositional environments. The long history of deposition yields a complex reservoir pattern. Paleogeographic reconstruction of individual reservoir units requires an understanding of this history and the reconstruction of detailed time-stratigraphic correlations. Simplistic interpretation of depositional sequences from well logs cannot always be utilized to interpret depositional environments, nor to predict reservoir production.

West of this stone quarry is another quarry, presently filled with water, where a 125-ft-thick distributary channel of the Bartlesville–Bluejacket Sandstone was formerly exposed. Figure 6 shows core representing this interval, which reflects the SE-trending distributary that prograded 40 mi into the basin near Wister, Oklahoma.

Figure 4. Ripple laminations in sandstone suggest antidune flow conditions, consistent with a shallow channel cutting into a distributary-mouth bar.

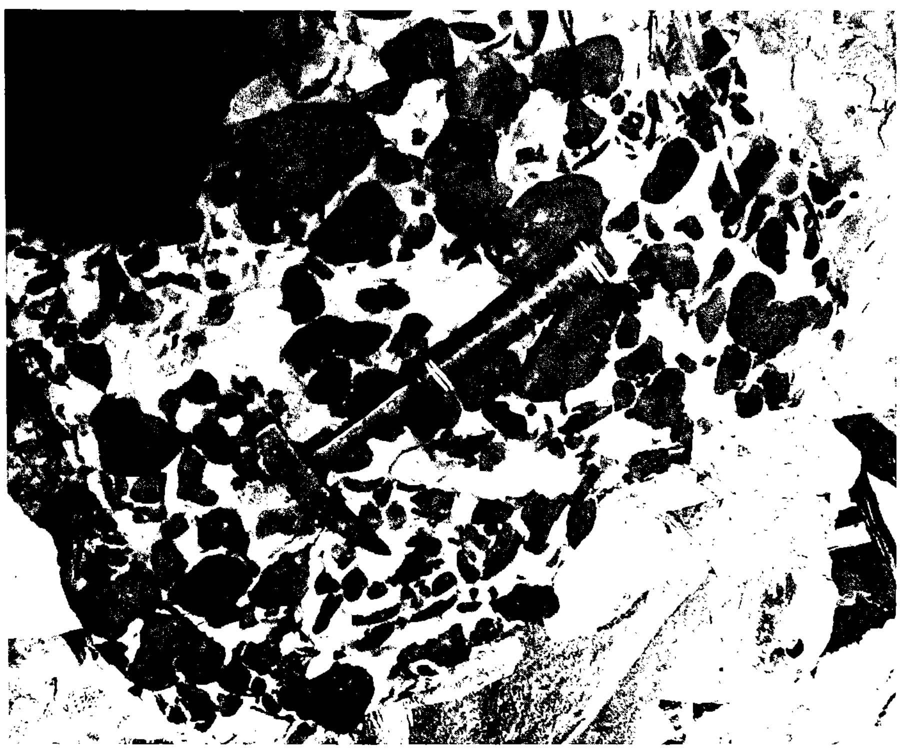
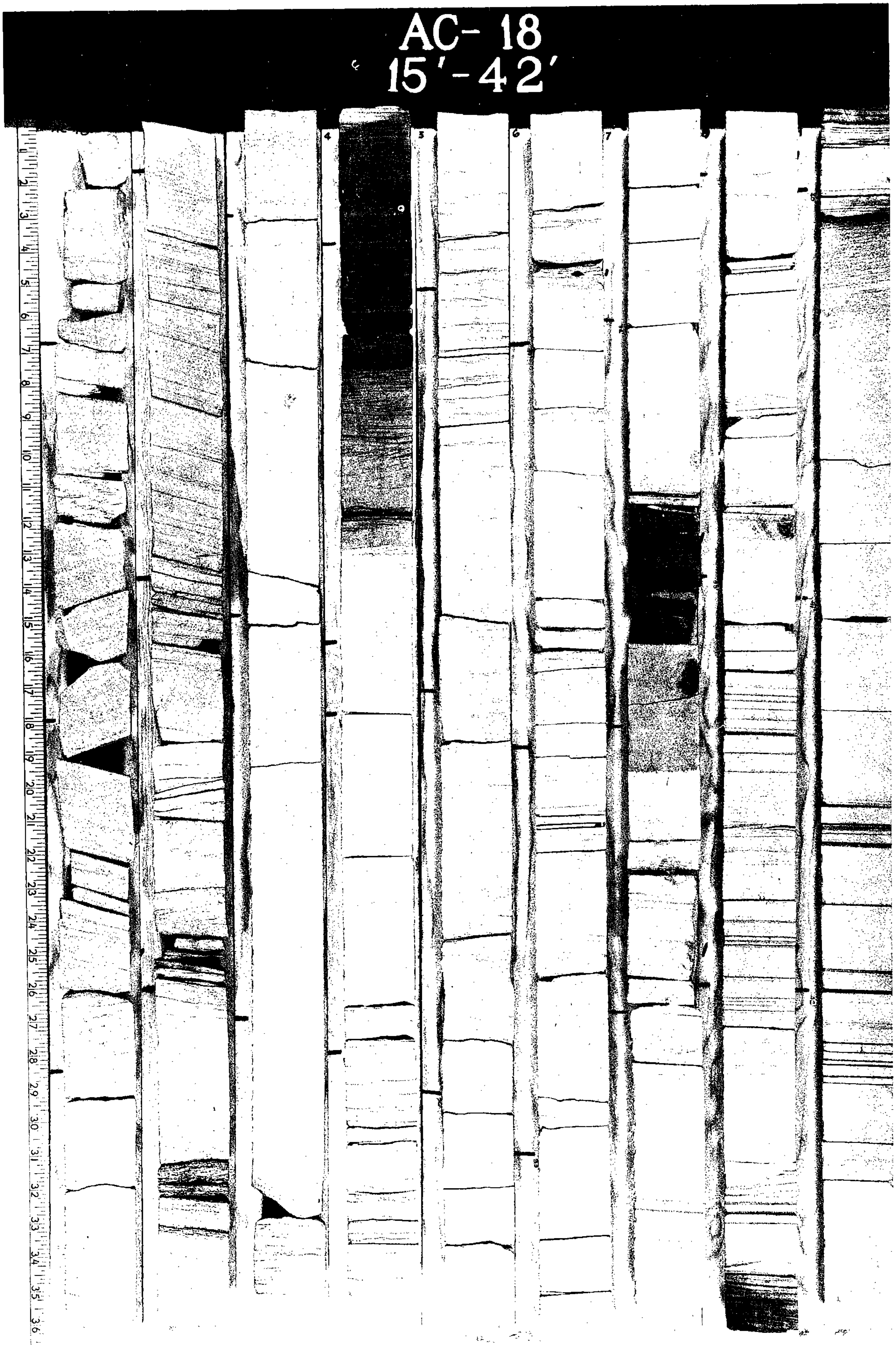
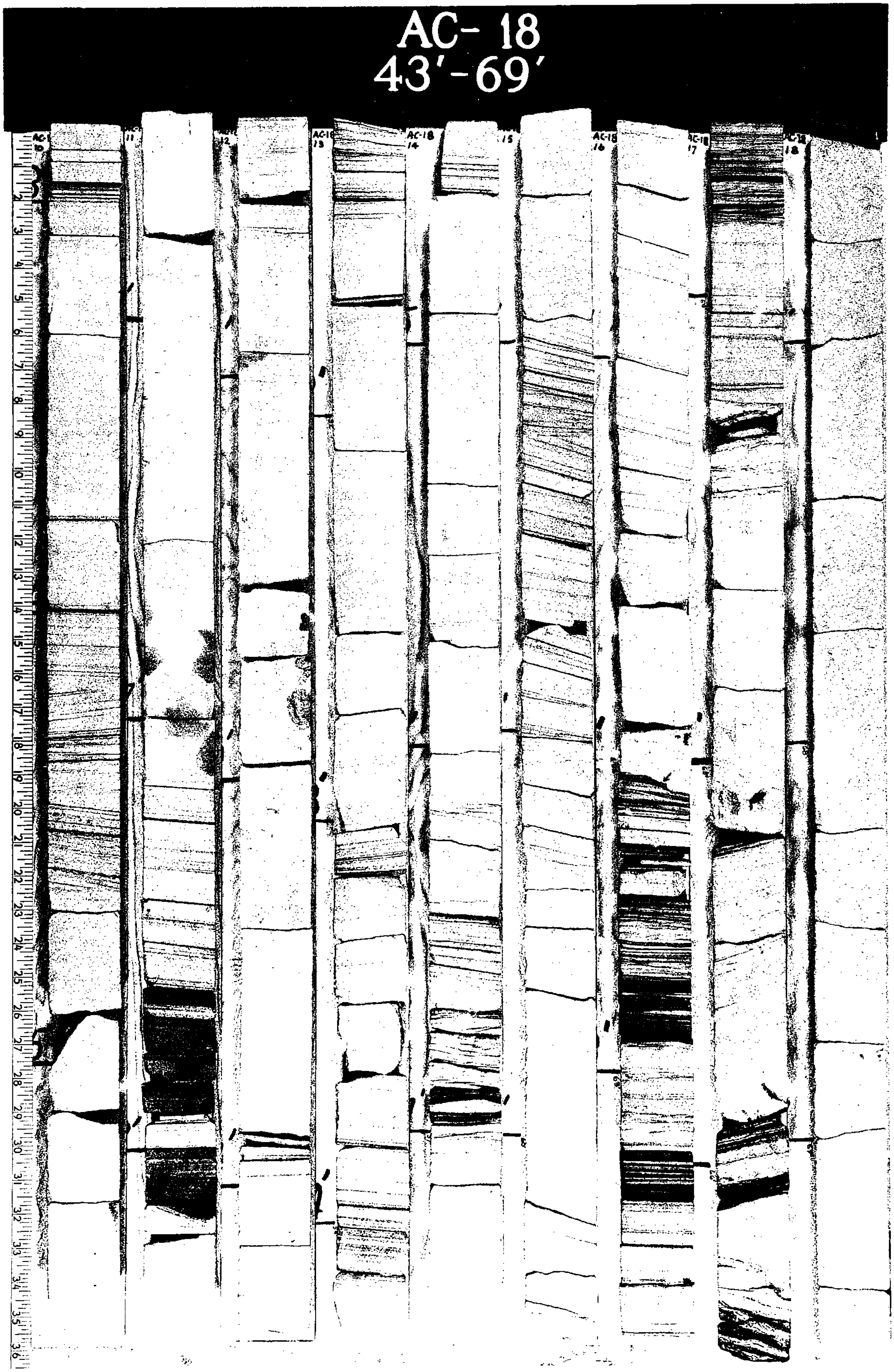
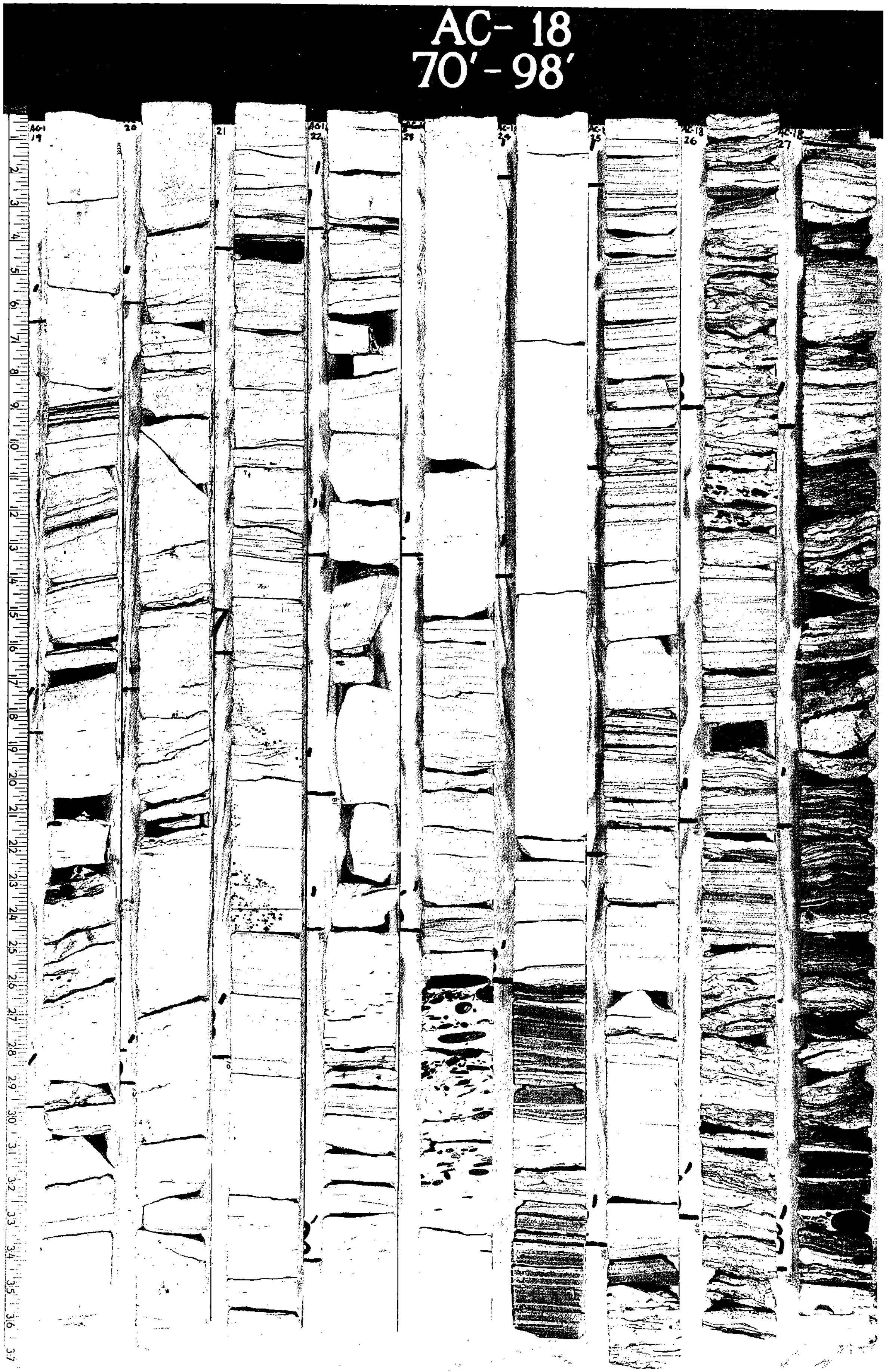


Figure 5. Some bedding planes contain shale clasts and fragments of carbonaceous matter.

Figure 6 (on p. 80–82). Two-inch-diameter core through a thick distributary channel in the Bartlesville–Bluejacket Sandstone just west of the Stigler Stone Co. quarry. Three photos represent depth ranges of 15–42 ft, 43–69 ft, and 70–98 ft, respectively.


AC- 18
15'-42'


AC- 18
43′-69′


AC- 18
70'-98'

STOP 4
LAKE CARLTON AND LAKE WAYNE WALLACE, ROBBERS CAVE STATE PARK

Glenn S. Visher and LeRoy A. Hemish

LOCATION

Bartlesville–Bluejacket Sandstone is exposed in cliff on the west side of Lake Carlton and in a road and cutbank for a gas well near the dam of Lake Wayne Wallace. These sites are in secs. 13 and 24, T. 6 N., R. 18 E., Latimer County (Fig. 1). The measured section (Fig. 2) begins just southwest of Lake Wayne Wallace Dam and continues to the west and north along a road to an area cleared for the gas well near the top of the hill.

DESCRIPTION

Excellent exposures of the Bartlesville–Bluejacket Sandstone can be seen in the cliff along the west side

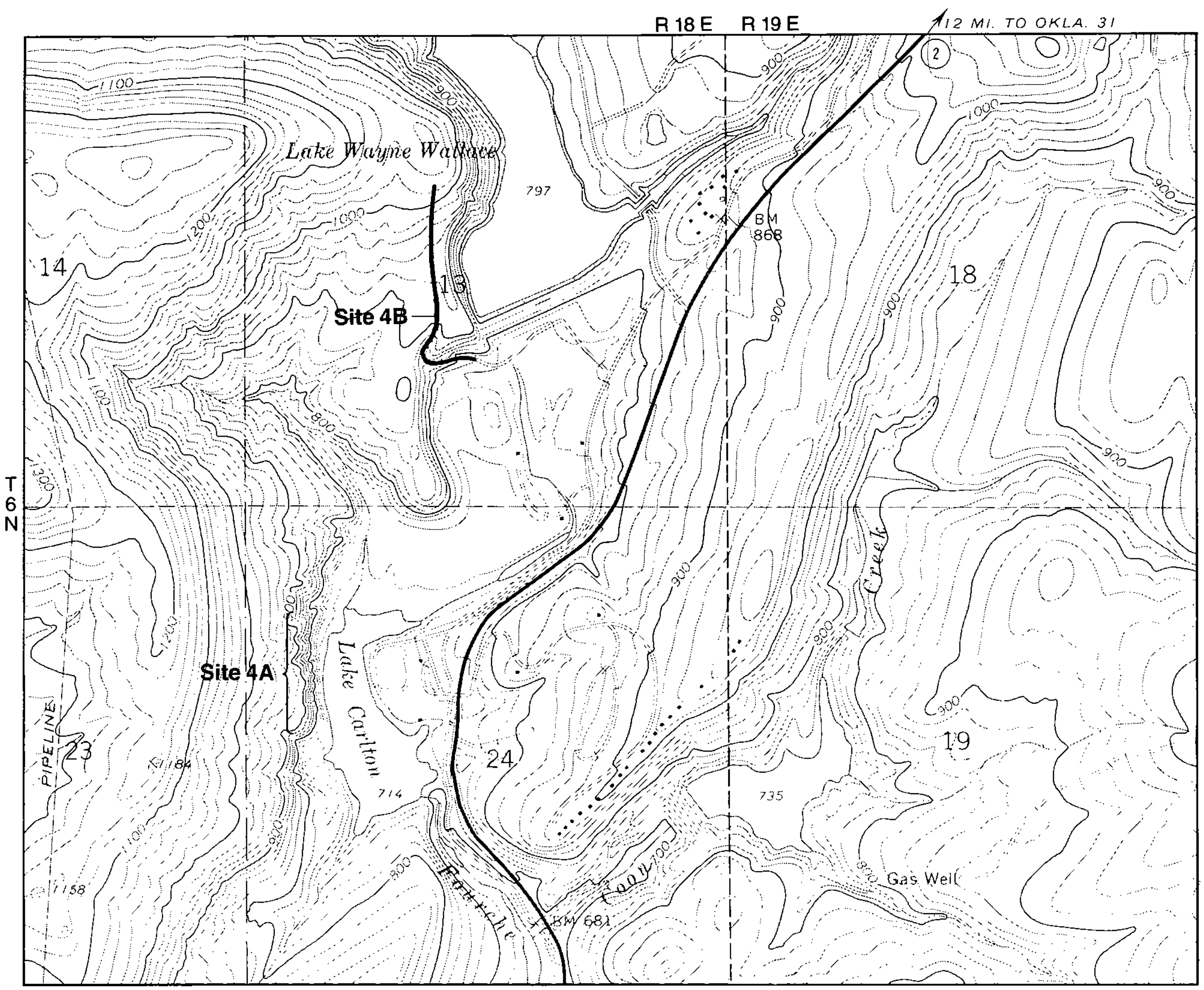


Figure 1. Location of Stop 4 sites at Lake Carlton and Lake Wayne Wallace, Robbers Cave State Park. Sites are ~5 mi north of Wilburton, Latimer County. Base from Wilburton 7.5′ Quadrangle.

SYSTEM	SERIES	GROUP	FORMATION	MEMBER	LITHOLOGY	THICK-NESS (ft)	DESCRIPTION
PENNSYLVANIAN	DESMOINESIAN	KREBS	Boggy	Bartlesville–Bluejacket Sandstone		8.0	Sandstone, yellowish-gray, very fine-grained, dense, noncalcareous, mottled, bioturbated; contains sparse, well-preserved burrows; includes some layers of light-gray shale up to 1 ft thick. **Offshore subaqueous bar.**
						2.0	Shale, dark-yellowish-brown and moderate-reddish-brown, silty, very soft and weathered. **Near-shore lagoon.**
						3.0	Shale, light-brown with moderate-reddish-brown mottling, weathered; contains weathered clay-ironstone concretions. **Transgressive shale.**
						(0.1)	Shale, dusky-brown, very carbonaceous. **Reworked coastal marsh.**
						1.4	Shale, light-brownish-gray with yellowish-gray and grayish-red bands, carbonaceous in part; contains stringers of moderate-reddish-orange clay-ironstone that contain fossil plant material and weather to fragments littering the outcrop. **Reworked coastal marsh.**
						(0.1)	Coal, black, very soft and smutty (Secor coal). **Marsh.**
						3.4	Underclay, grading to shale; yellowish-gray to pale-yellowish-brown, soft; contains abundant fossil plant material, carbonaceous in part.
						45.0	Sandstone, yellowish-gray to grayish-orange to dark-yellowish-orange, very fine-grained, micaceous, noncalcareous; upper contact sharp but irregular; on upper surface, casts of *Stigmaria, Calamites, Lepidodendron,* and other plant fossils abundant; thin-bedded, slightly carbonaceous, moderately friable and silty from 35–42 ft.; ripple-marked on some exposed bedding surfaces. **Bay-fill sandstone, capped by levee.**
						30.5	Covered interval (assumed to be shale). **Open bay.**
						4.5	Sandstone, dark-reddish-brown, fine- to medium-grained, noncalcareous, ferruginous, thin- to medium-bedded. **Subaqueous levee.**
						4.0	Covered interval (assumed to be shale).
						2.5	Sandstone, moderate-yellowish-brown, fine-grained, noncalcareous, thin-bedded; ripple marks abundant.
						1.5	Shale, very pale-orange, noncalcareous, silty, interbedded with thin stringers of micaceous siltstone and very fine-grained sandstone.
						19.5	Sandstone, moderate-yellowish-brown, fine-grained, noncalcareous, thin-bedded; ripple marks abundant, shaly below 110 ft. **Submarine bar.**
						2.0	Shale, grayish-orange, silty, noncalcareous; interbedded with thin stringers of very fine-grained micaceous sandstone; unit is bioturbated.
						46.5	Sandstone, moderate-yellowish-brown, very fine- to fine-grained, noncalcareous, ripple-marked, micaceous, thin-bedded; contains black macerated plant fragments on stratification surfaces; includes medium-gray shale pebbles in places; shaly 144–164 ft.; massive in lower 2 ft.; base of unit covered. **Pro-delta distributary-mouth bar.**

Figure 2. Pennsylvanian strata exposed along gas-well road, Site 4B. Measured section prepared by Hemish; environmental interpretations (**bold**) by Visher.

of Lake Carlton (Site 4A). An exposure along a road (Site 4B; Fig. 2) show a pro-delta sequence (Fig. 3), overlain by a lagoonal-bay shale, a coarsening-upward levee system, a transgressive coastal-marsh deposit, and a coarsening-upward marine coastal sequence.

In deeper water, as indicated by the thickness of the pro-delta sequence, subaqeous gravity transport was the principal transport process; fine-grained sandstone units of Tertiary age may be commercially productive in such a setting, but in highly altered Paleozoic rock units this is rarely possible. This outcrop must be near the shelf edge of the Bartlesville–Bluejacket interval, and in this position growth faulting is often present. Little evidence of synsedimentary listric faulting has been observed in outcrop or wells in this area.

Figure 3. Thin-bedded, shaly sandstone in the lower part of the Bartlesville–Bluejacket Sandstone exposed along road at Site 4B. Depositional environment is interpreted as a pro-delta distributary-mouth bar.

STOP 5
ASYMMETRICAL FOLDS IN THE ATOKA FORMATION

Charles A. Ferguson

LOCATION

This stop is in and adjacent to a road cut along State Highway 2 in sec. 27, T. 4 N., R. 19 E. (Fig. 1). The site is ~12 mi south of Wilburton, and is 1.9 mi south of the Spanish War Veterans Colony, or 0.9 mi north of the intersection with State Highway 1. Pull off the highway on a short dirt road that turns

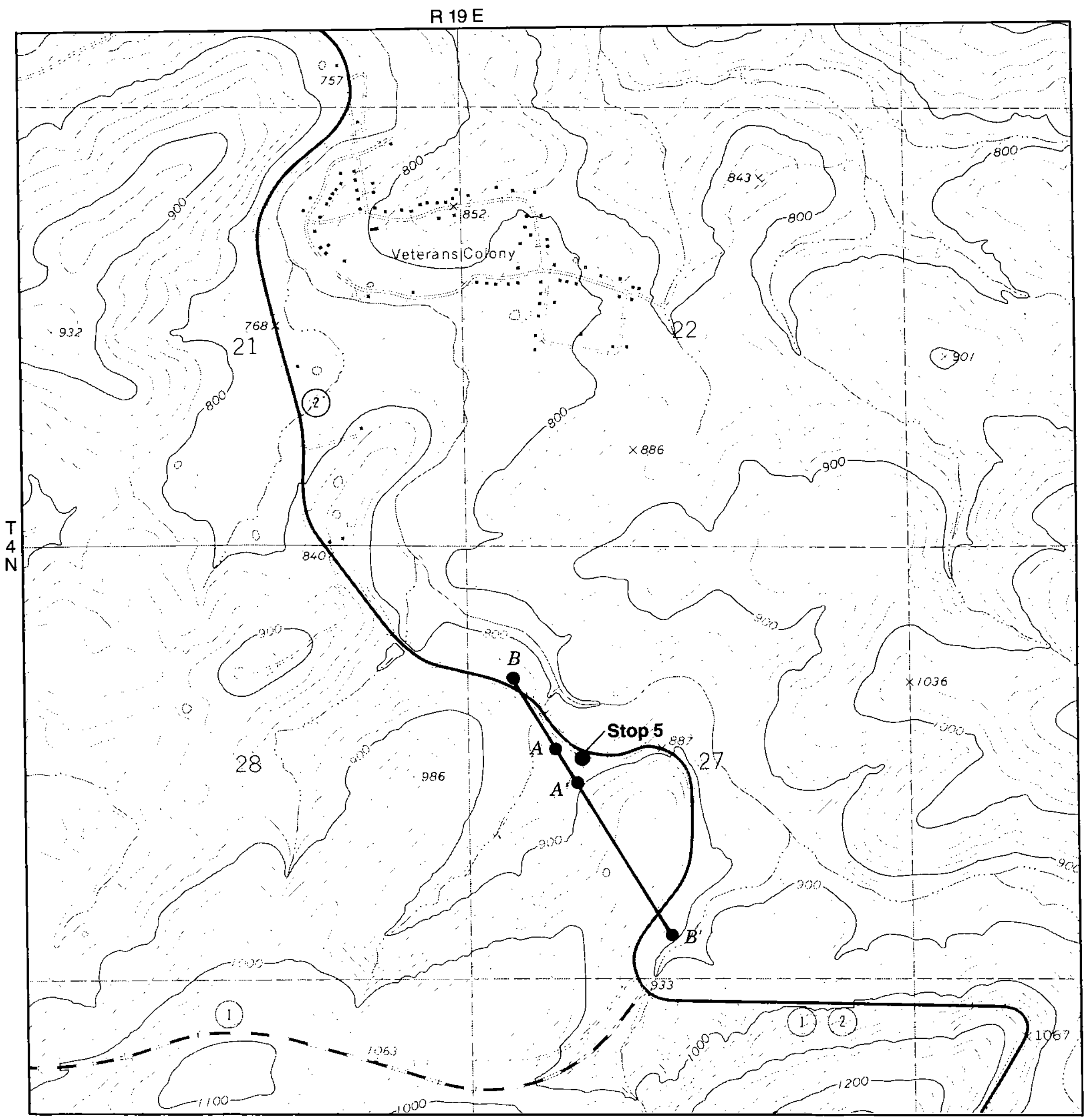


Figure 1. Location of Stop 5 along State Highway 2 in Latimer County. Base from the Damon 7.5′ Quadrangle.

southwest of the highway (note the large yellow real-estate sign here). The exposure to be examined is in the borrow ditch on the south side of the road, behind the first road reflector east of the turnoff.

DESCRIPTION

Strata at this stop are medium-bedded turbidite sandstones and shale of the lower Atoka Formation. Structurally, this is the northern fold of an asymmetric syncline–anticline pair on the north limb of a major overturned anticline. Figure 2 is a cross-sectional sketch of the exposure showing its relationship to local structure. North-facing strata on this limb are vertical to slightly overturned, and south-facing strata dip south at 50–60°. This style of folding is typical of the area, and reflects the overall northward vergence of structures in the Ouachita Mountains frontal belt.

The major anticline, of which the structure examined here is only a part, is one of many folds in the northern part of the Ti Valley fault block, the southernmost of three major blocks that constitute the frontal belt. Each block consists almost entirely of Atoka Formation flysch. Older strata (Morrowan) are different in each block, and reflect southerly changes from shelf facies to deep-basin facies. In the Ti Valley block, Atoka Formation overlies a shale-rich flysch unit with olistostromes called the Johns Valley Shale (Stop 6). To the north, the Atoka Formation overlies shallow-water-facies limestone and sandstones called the Wapanucka Formation (Stop 7).

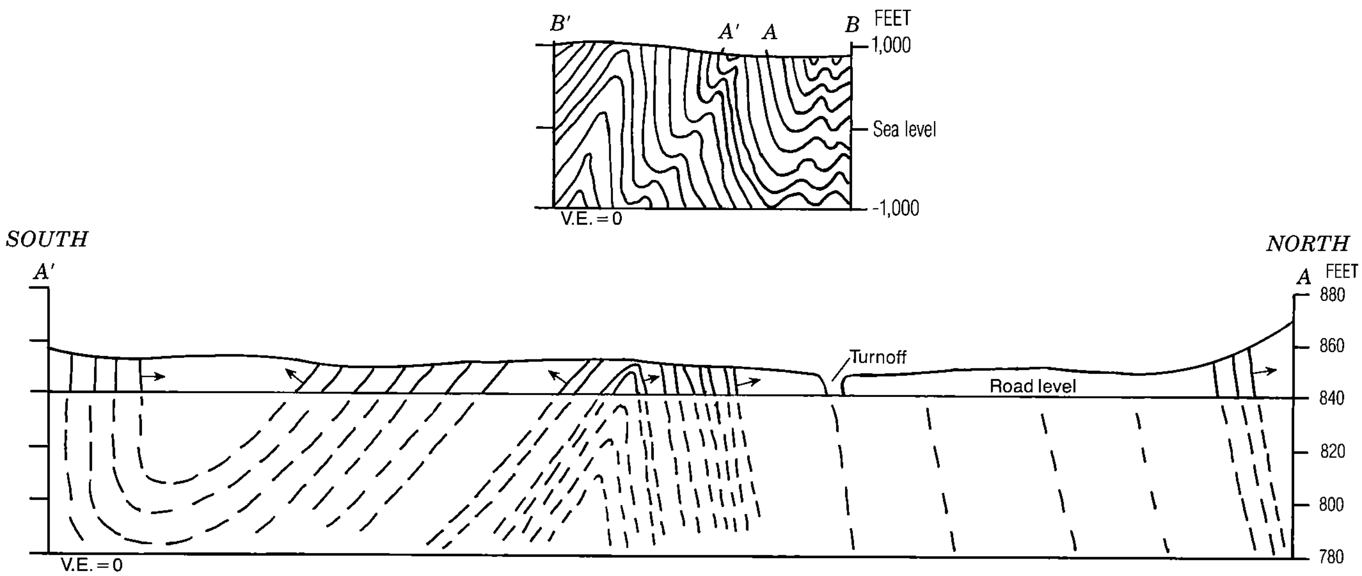


Figure 2. Structural cross sections, Stop 5 (see Fig. 1 for location of cross sections). A–A′ is an enlarged and detailed segment of B–B′. In each cross section the horizontal and vertical scales are equal. Arrows in A–A′ indicate facing direction (top of beds).

STOP 6
ATOKA, JOHNS VALLEY, AND JACKFORK STRATA, HAIRPIN CURVE LOCALITY

Neil H. Suneson

LOCATION

This well-known and much-studied locality is in the vicinity of Hairpin Curve along State Highways 1 and 2, in secs. 2 and 3, T. 3 N., R. 19 E., Latimer County (Fig. 1). The stop is ~3 mi (by road) south of the convergence of State Highways 2 and 3, and ~15 mi south of Wilburton.

DESCRIPTION

The Hairpin Curve locality is perhaps the most famous and frequently visited geologic road cut in the frontal belt of the Ouachita Mountains. It has been described in at least seven published guidebooks of the mountains (Tulsa Geological Society, 1947; Cline and others, 1959; Cline, 1968; Hare, 1973; Briggs and others, 1975; Decker and Black, 1976; Chamberlain, 1978); the site undoubtedly has been visited by many university and industry field trips. In addition to the guidebooks, several studies have focused on specific features of the strata exposed at this locality: Shideler (1968, p. 129; 1970) studied the origin of the boulder assemblage in the Johns Valley Formation; Chamberlain (1970, p. 17f; 1971) studied the trace-fossil assemblage in the sandstones, and Nooncaster (1985) studied the petrography of the sandstones. The strata exposed were measured by Hendricks and Averitt in 1939 (Tulsa Geological Society, 1947, p. 32–34) and Cline and Laudon in 1958 (Cline and others, 1959, p. 35–37,40–41). Despite the intense scrutiny the strata at Hairpin Curve have received, there is no general agreement about the geology of the area—in fact, the interpretation published here is the fifth to be published.

The aforementioned studies of the Hairpin Curve strata can be summarized as follows: Most authors agree that the Atoka/Johns Valley contact is exposed here and is depositional, although the exact position of the contact is debatable, and it is probably gradational. There is a siliceous (spicular) shale zone ~160 ft above the top of the Johns Valley that Hendricks claimed was useful in mapping the base of the Atoka (throughout this discussion, I will not repeatedly refer to sources of information, except where new references are cited). Cline correlated this spiculite zone with the upper Wapanucka and suggested that the base of the Atoka Formation here may be Morrowan. Similarly, Ferguson and I have observed a very sparse fauna from the base of the Atoka at the sharp bend in the road ~1.8 mi to the north, vaguely reminiscent of the fauna that is so profuse in the "Spiro" sandstone. The "Spiro" is complexly interbedded with the Morrowan Wapanucka limestone just south of the Choctaw fault. Based on the trace-fossil assemblage in the Atoka and underlying Johns Valley, Chamberlain suggested that the sandstones exposed here represent proximal turbidites, fluidized flows, and debris flows deposited in a deep basin or lower-slope environment. Paleocurrent measurements (Ferguson and Suneson, this volume) indicate an east-to-west current direction for the Atoka turbidites. Nooncaster reported that the Atoka sandstones at this locality contained about 80–90% quartz, 2–4% metamorphic rock fragments, 2–5% feldspar, and 2–15% matrix.

Historically, the Johns Valley Formation at the Hairpin Curve has been separated into two units: Hendricks called these the upper boulder bed and lower boulder bed (predominantly Caney shale) of the Johns Valley, and Cline called them the upper Johns Valley and lower Johns Valley (Caney facies). This outcrop, among others, undoubtedly served as the focal point for the long-lasting Caney/Johns Valley controversy. The argument (which has been resolved by recent work in Arkansas) was based on interpretations of the depositional environment of the Caney part of the Johns Valley and can be summarized as follows: Cline (and others) felt that the Mississippian Caney shale was stratigraphically autochthonous, that it had been deposited on Jackfork strata south of the Ti Valley fault, that the Jackfork was therefore Mississippian, and that the exotic-boulder-bearing Morrowan Johns Valley was deposited on it. Hendricks (e.g., Hendricks and others, 1947) believed that the Caney part of the Johns Valley is stratigraphically allochthonous, that Mississippian (as well as older) strata are enclosed in a Pennsylvanian matrix, and that the Jackfork is Pennsylvanian. Later, Gordon and Stone (1977) showed that the Jackfork and Johns Valley are Morrowan.

Perhaps the best description of the "upper" Johns Valley at Hairpin Curve is that of Cline and others (1959), because the outcrops were fresh due to road widening:

> "The shales containing the limestone erratics are of special interest. The most conspicuous

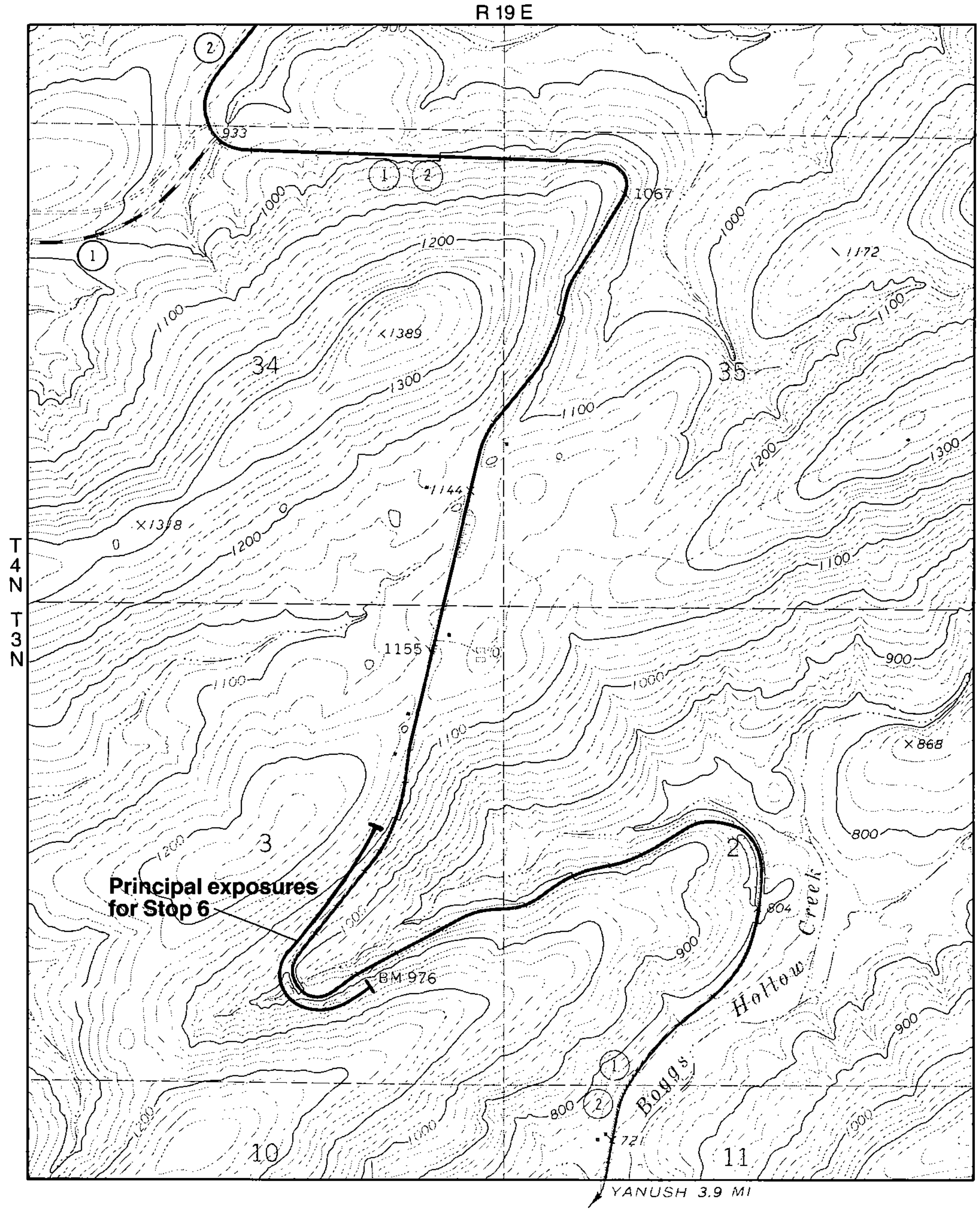


Figure 1. Location of Stop 6 at Hairpin Curve along State Highways 1 and 2, Latimer County. Base from Damon 7.5′ Quadrangle.

boulder bed, described above as zone 4 (about 143′ below the base of the Atoka), has been interpreted by some as a friction carpet at the base of an advancing thrust sheet. This boulder-bearing clay-shale rests in a channel which cuts out at least 11½ feet of zone 3. There is a noticeable decrease in the size of the erratics upward in this deposit, the overall effect being not unlike that of graded bedding, but it is of course on a somewhat larger scale than the usual examples. The erratics in the lower part of the channel fill include well rounded boulders with diameters in excess of a foot, slightly rounded blocks of similar dimensions, the whole being embedded in a clay-shale matrix. Upward the boulders give way to cobbles and they in turn give way to pebbles which are widely separated in the shale and give a plum pudding effect. Throughout this deposit there are rounded masses of hard, brown, quartzitic sandstone. We interpret this particular boulder bed as the product of a single turbidity flow or submarine slide. The flow initially attained a high velocity during which phase it was able to transport boulders and scour previously deposited muds and sands. As the peak of the flow was reached and the velocity trailed off, pebbles began to drop out and were deposited with the muds of the flow and the muds obtained from the reworked bottom. The rolled sand-

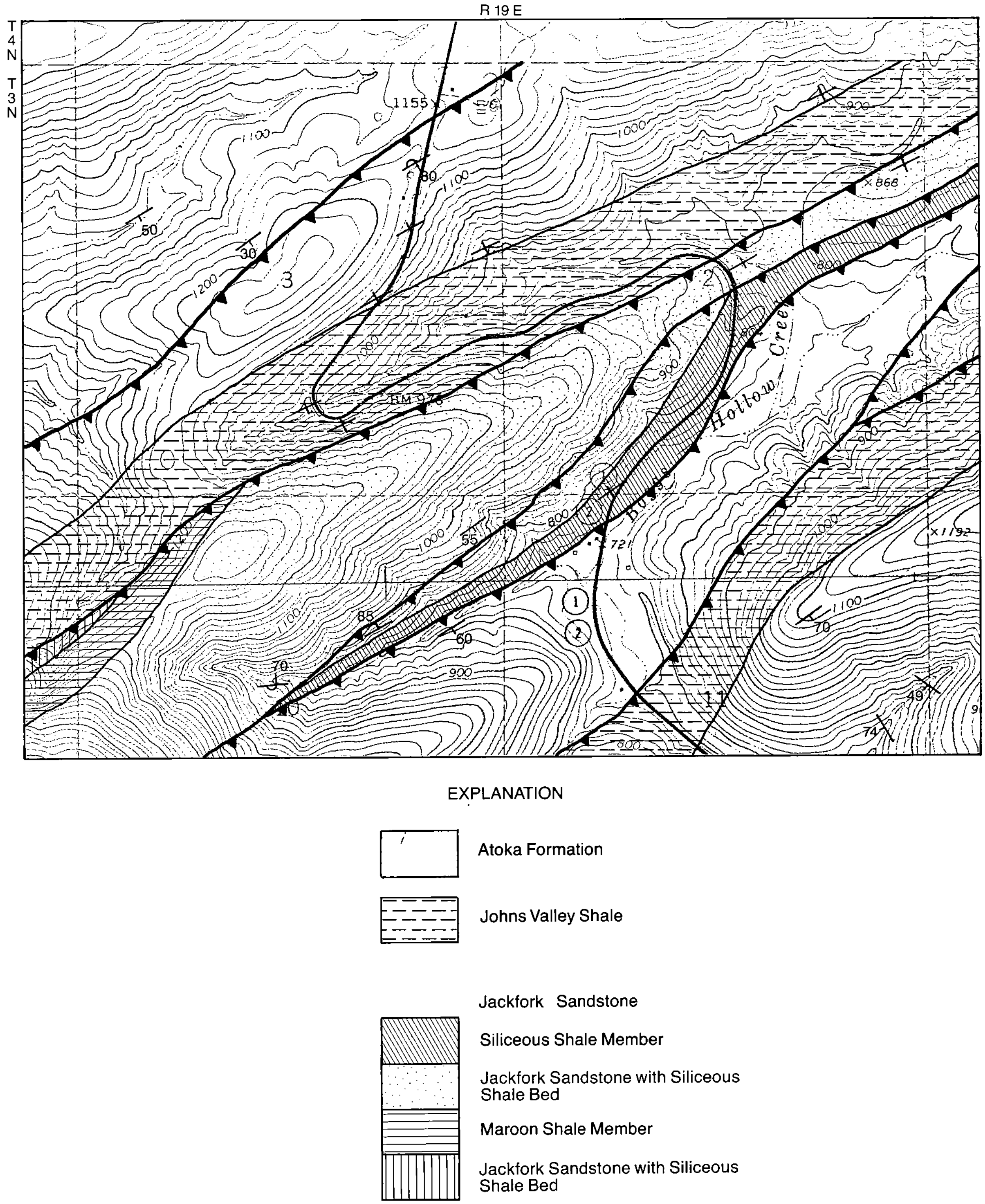
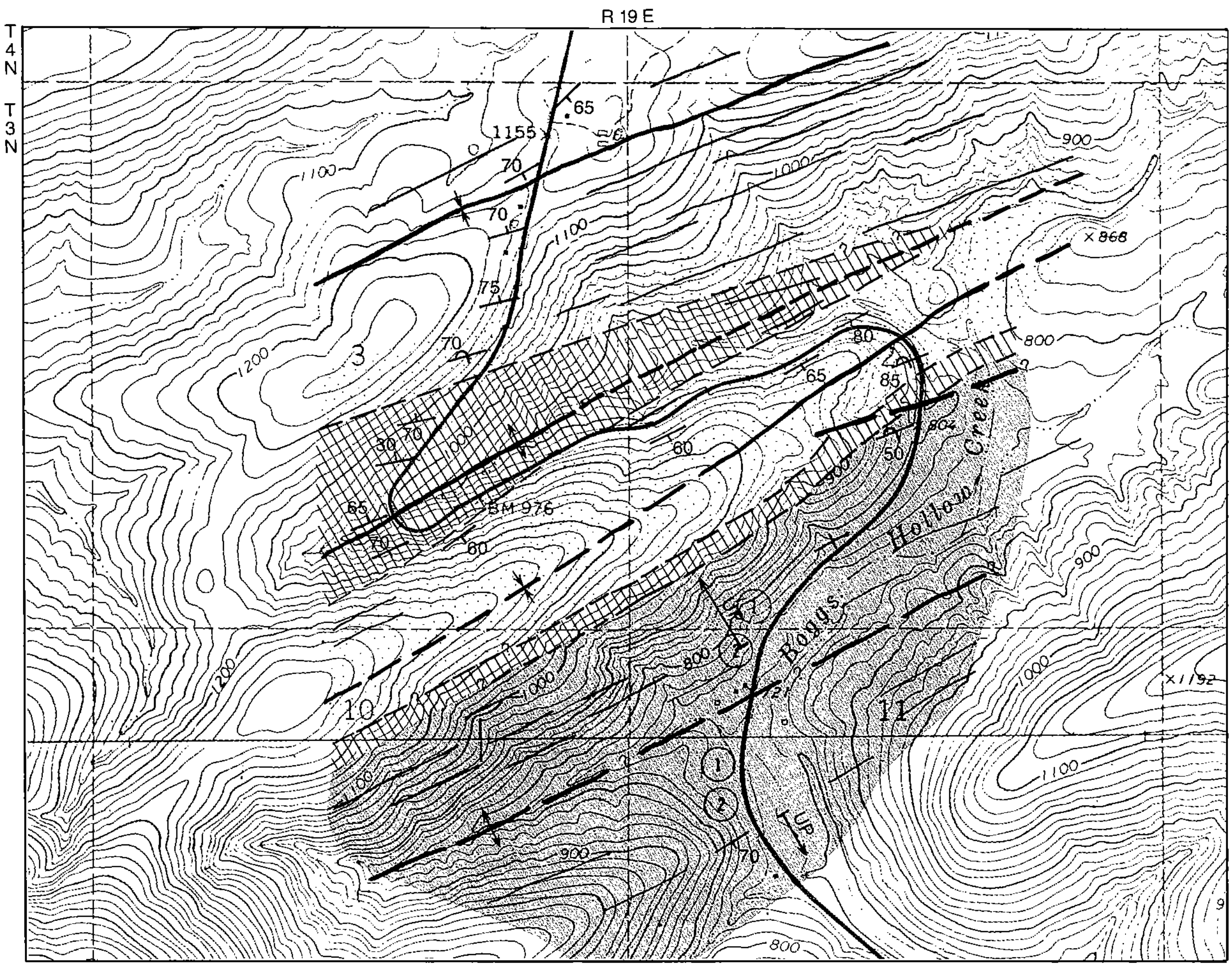


Figure 2. Geologic map of Hairpin Curve locality according to Hendricks and others (1947). Note that except for that part of the Atoka Formation overlying the Johns Valley just north of the curve, all the strata appear to face southeast.

stone masses represent lenses of sand torn from the bottom and rolled along the flow. The convolute bedding and the flute casts on the undersurfaces of some of the sandstones support the general view that turbidity currents were operative during this time."

Shideler recognized boulders and cobbles from the following formations in the Johns Valley on the north side of Hairpin Curve (i.e., "upper" Johns Valley): Boone, Bromide, Chimneyhill, Fite, Fort Sill, Henryhouse, Kindblade, McLish, Oil Creek, Pinetop, Sycamore, Viola, and West Spring Creek. Nooncaster reported that the "upper" Johns Valley sandstones contain ~95% quartz, 1% metamorphic rock fragments, 2% feldspar, and 2% matrix.

To the south, the "lower" Johns Valley or "Caney facies" of the Johns Valley is exposed in now deeply weathered road cuts. Hendricks measured 100 ft of lower Johns Valley containing a 550- × 30-ft coherent Caney shale block. Cline measured 117 ft of lower Johns Valley, in which 27 ft were "typical Mississippian Caney shale." In general, stratification in the Caney is parallel to that in the Johns Valley, but the Caney is typically harder, more fissile, and blacker than the Johns Valley. In many places in the frontal Ouachitas, the Caney contains limestone boulders that are better described as "concretions," with stratification parallel to that of the enclosing shale, than erratics or exotics, in which the limestone boulders are clearly foreign to the enclosing shale. On the south side of Hairpin Curve ("lower Johns Valley"), Shideler found boulders and cobbles of the following units: Bigfork, Bois d'Arc, Fernvale, Fite,



EXPLANATION

Atoka/Upper Jackfork Formations

Upper Jackfork Formation

Lowest upper Jackfork Formation, Contains Boulder-Bearing Shale Member at Base

Middle and Lower Jackfork Formation

Figure 3. Geologic map of Hairpin Curve locality according to Misch and Oles (1958). Note that no thrust faults are mapped near the curve.

Joins, Sycamore, Tyner, Viola, Wapanucka, and West Spring Creek.

In contrast to the attention that has been paid to the Atoka and Johns Valley Formations, the Jackfork at this locality is virtually unstudied, perhaps because it is even more poorly exposed than the Johns Valley, is less interesting than the Johns Valley, and is nearly a mile down the road beyond the big turnout. Most authors agree that the sandstones that immediately underlie the Johns Valley are part of the Game Refuge Formation, the uppermost formation in the Jackfork Group. Underlying the Game Refuge is the Wesley Formation, a unit that many authors called the Wesley siliceous shale member of the Jackfork Formation.

At the Hairpin Curve locality, the Game Refuge

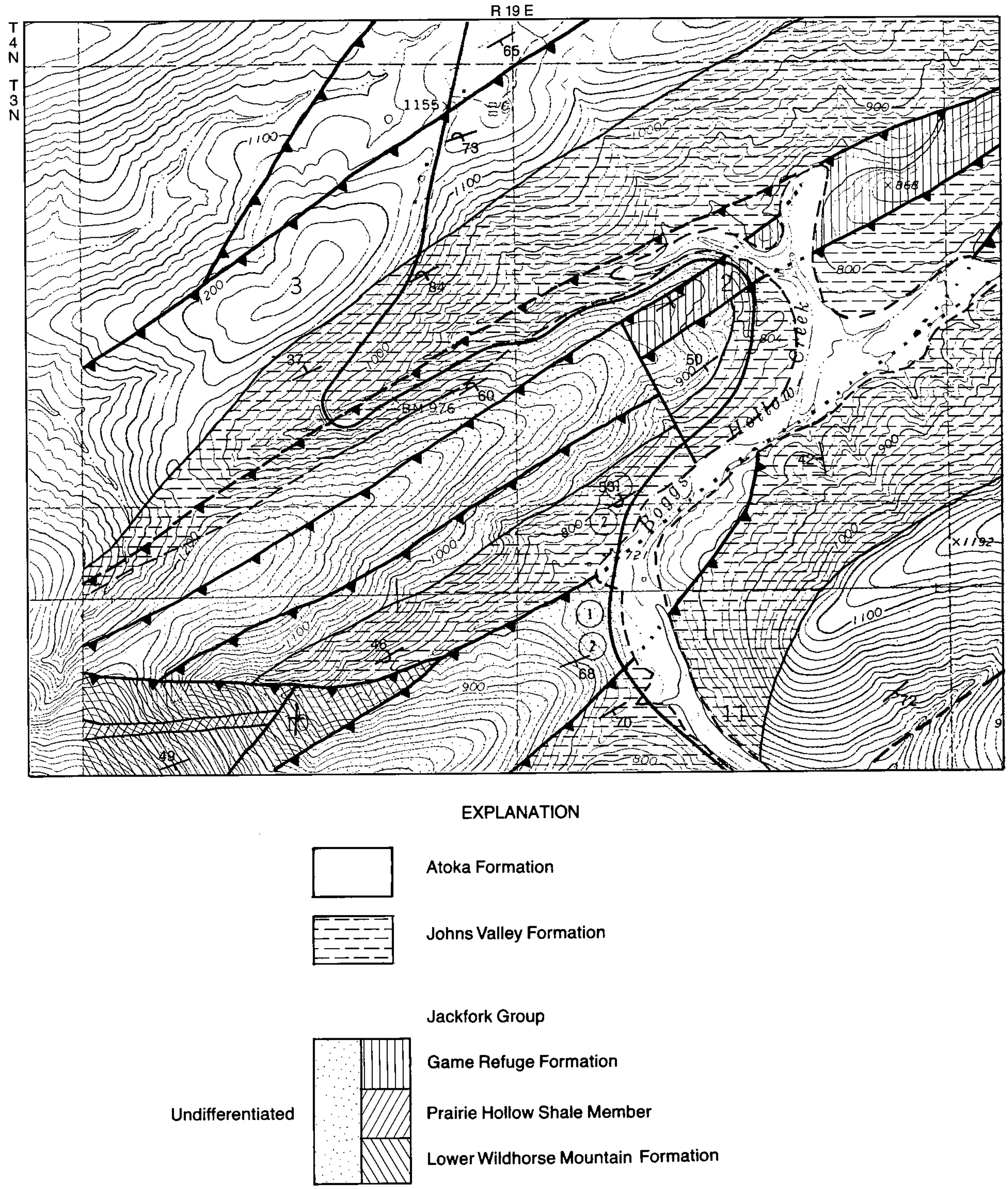


Figure 4. Geologic map of Hairpin Curve locality according to Fellows (1964). Note the complex pattern of thrusts and cross faults and variably facing fault blocks.

sandstone consists of interbedded sandstone, siltstone, and shale (the siltstone and shale are largely covered and only inferred.) The sandstones range from a few inches to possibly as much as 70 ft thick, but average about 2–5 ft thick. Bouma sequences are well developed in the thinner sandstones; the thicker sandstones tend to be unstratified and massive. The sandstones are similar to Atoka sandstones, except that locally they are distinctly bimodal, (1–2-mm grains in a fine-grained matrix), contain fossil fragments (brachiopods, crinoids), and are interbedded with granule conglomerates. The Wesley shale consists of brownish fissile mudstone with rare 1-in. to 1-ft sandstone beds exhibiting a typical T(b)cd Bouma sequence.

Although the stratigraphy of the units exposed at Hairpin Curve is relatively well understood, the geology is not. Figures 2–6 are five geologic maps of the

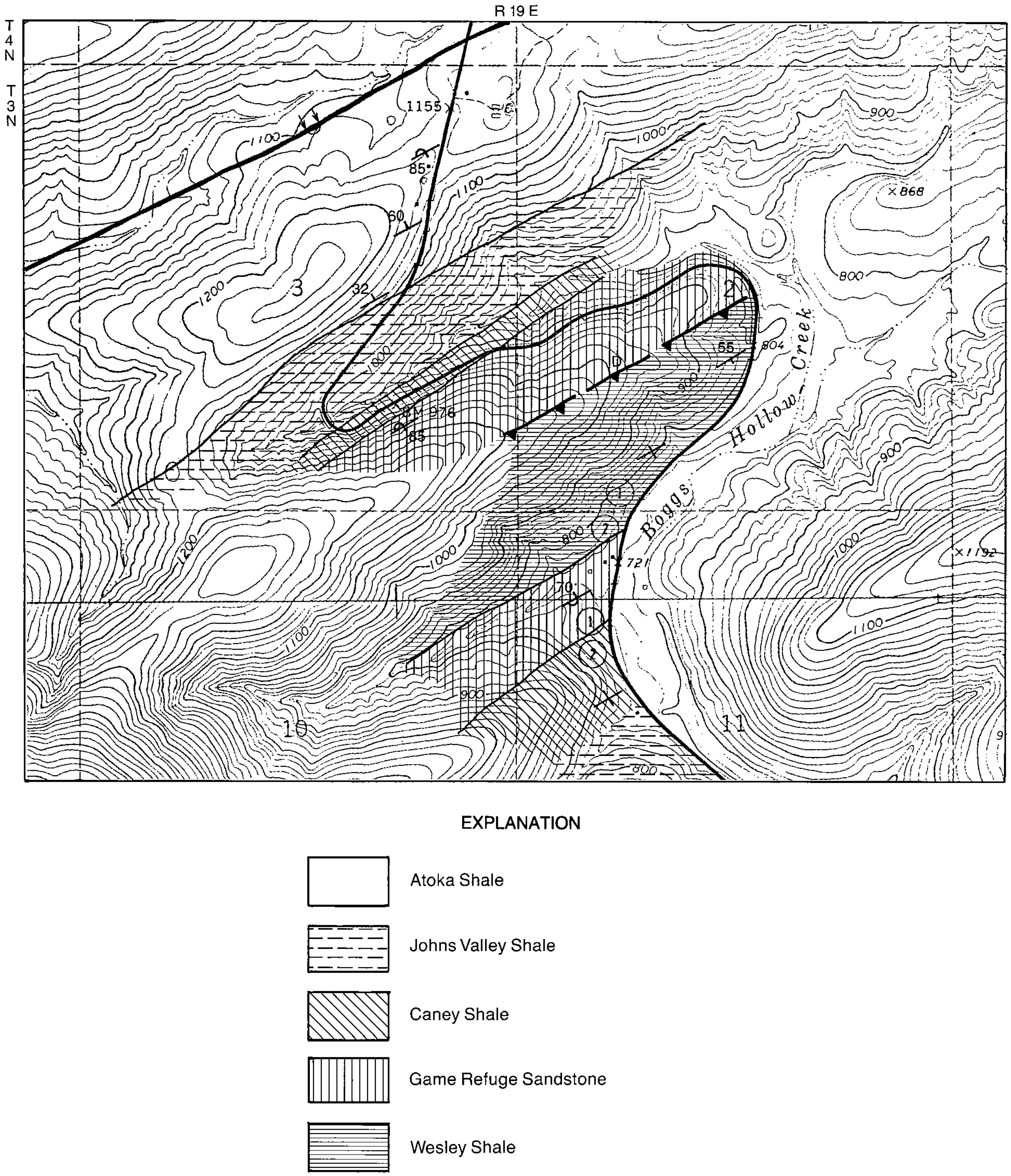


Figure 5. Geologic map of Hairpin Curve locality according to Bowsher and Johnson (1968).

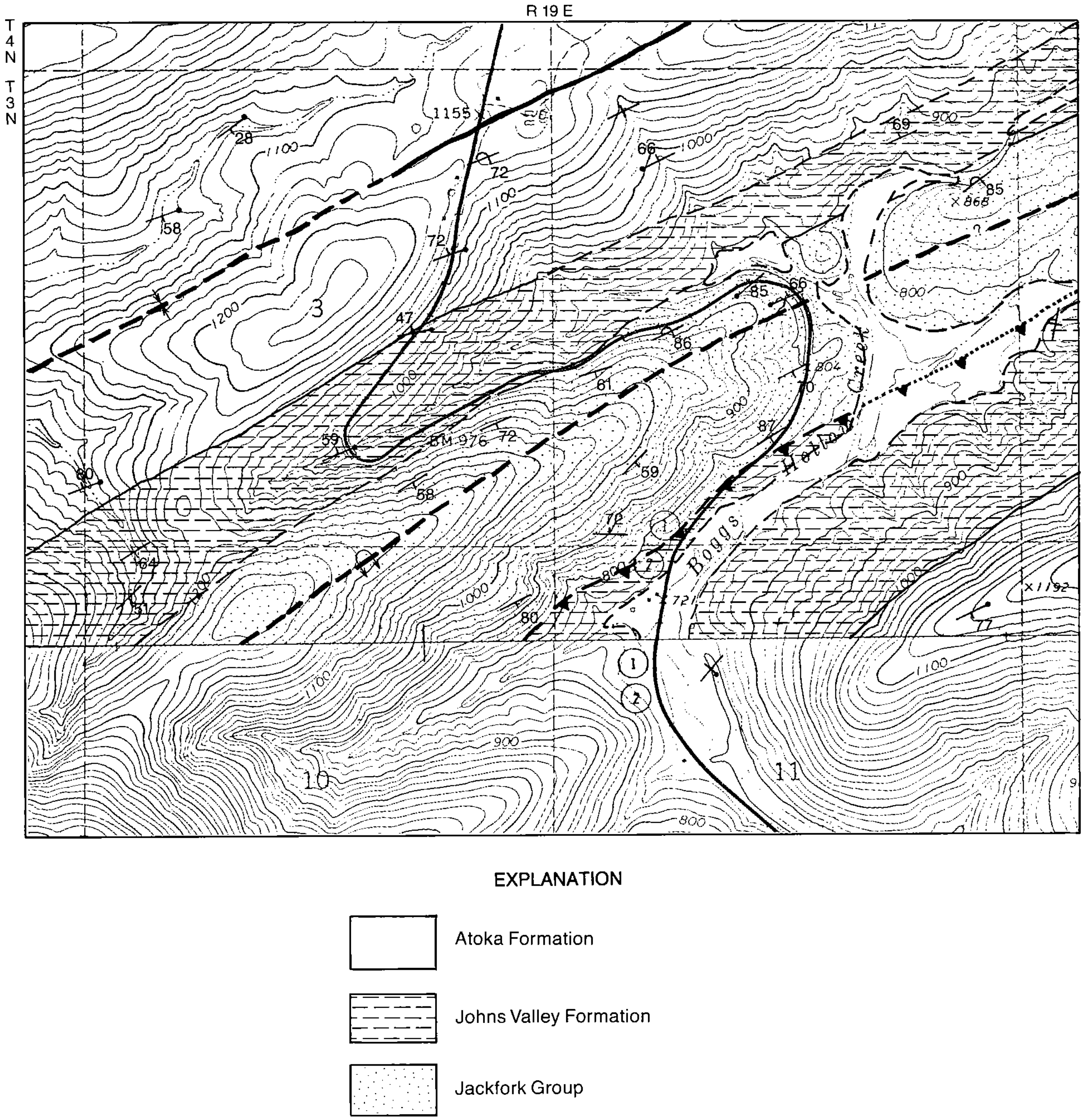


Figure 6. Geologic map of Hairpin Curve locality according to Suneson and Ferguson's mapping (Suneson's interpretation). Note that the map pattern is that of an overturned (to the north) anticline with a thrust-faulted southeast limb.

Hairpin Curve locality. The major features of each of the maps are as follows: Hendricks and others (1947) bound nearly every unit with the thrust fault (except for the Atoka–Johns Valley contacts in the north and southeast) and show a large thrust slice of Atoka Formation over Jackfork to the southeast (Fig. 2). Misch and Oles (1958) redefined the stratigraphy, preferring to call the Johns Valley a "boulder-bearing shale at the base of the upper Jackfork" (Fig. 3); they mapped what appears to be the northwest flank of an anticlinorium with progressively older strata exposed to the southeast. Fellows (1964) mapped a complicated pattern of thrusts, cross faults, and overturned beds, although his map pattern was essentially that of a thrust-faulted anticline with Atoka exposed to the northwest and southeast (Fig. 4). Bowsher and Johnson (1968) mapped a relatively simple anticline with a single thrust fault in the core; however, they mapped the Jackfork below the Caney in the lower part of the Johns Valley (Fig. 5). Suneson and Ferguson (unpub.) mapped a relatively simple overturned (to the northwest) anticline with a thrust fault in the core juxtaposing Johns Valley against Jackfork (Fig. 6; Suneson's interpretation).

STOP 7
WAPANUCKA FORMATION: A MORROWAN–ATOKAN SHALLOW-WATER LIMESTONE AND SANDSTONE

Charles A. Ferguson

LOCATION

Figure 1 shows the location of Stop 7, at a water gap at Hartshorne Lake Dam. Stop 7 is 1.8 mi south of U.S. Highway 270 on 7th Street in Hartshorne, sec. 13, T. 4 N., R. 16 E., Pittsburg County. Pull off at a large turnout that overlooks Hartshorne Lake and Dam.

DESCRIPTION

The Wapanucka Formation is the youngest shallow-water shelf-facies sequence in the Ouachita–Arkoma basin. It overlies at least 1 km of Paleozoic shelf-facies rock (mostly limestone), including strata down to the Cambrian. In the Arkoma basin, the Wapanucka is overlain by 5 km of deep-water flysch of the Atoka Formation. The Wapanucka shelf foundered during the early Atokan, apparently due to flexural subsidence related to tectonic loading of an obducting landmass to the south.

Grayson (1979) described the Wapanucka Formation in detail throughout the frontal belt. He recognized four informal members: the Chickachoc chert member, a lower limestone member, a middle shale member, and an upper sandstone–limestone member. The Chickachoc chert is an outer-shelf-facies time equivalent of the lower limestone member.

The Wapanucka Formation and part of the underlying Carboniferous sequence are exposed at the bases of fault blocks in the northern Ouachita Mountains frontal belt (the exposure at this stop is in the northernmost of these fault blocks). As a consequence, basinward facies changes of the Wapanucka are found in fault blocks progressively to the south (Fig. 2). The most recognizable basinward changes are a dramatic increase in thickness of the middle shale member, and a transition from spicular limestone to spiculite in the lower two members.

A 130-m-thick section of the Wapanucka was measured in the water gap at Stop 7 (Fig. 3). The Wapanucka overlies a Morrowan calcareous-shale interval informally named the Springer shale. The Springer, not exposed here, contains a pelagic fauna (ammonites) suggestive of an outer-shelf environment.

The best exposures of the lower member are along the stream bed, but it can also be examined along the road cut. Downstream from the bridge are three small outcrops of medium-bedded, bioclastic micrites and grainstones (total thickness 17.5 m). From here, there is a 60-m covered interval up to good exposures of spicular limestone, still of the lower member, just south of the bridge. Follow the east bank of the stream up to the base of the dam to examine the upper part of the lower member (25 m). The rest should be examined along the road cut west of the stream. Start with the middle shale member (11 m) that is mostly covered in a recessive interval; it consists of olive-gray shale similar to that in the overlying Atoka Formation. The upper member (13 m) consists of thickening-upward sequences of fine- to medium-grained, tabular-bedded sandstones and rare bioclastic grainstones.

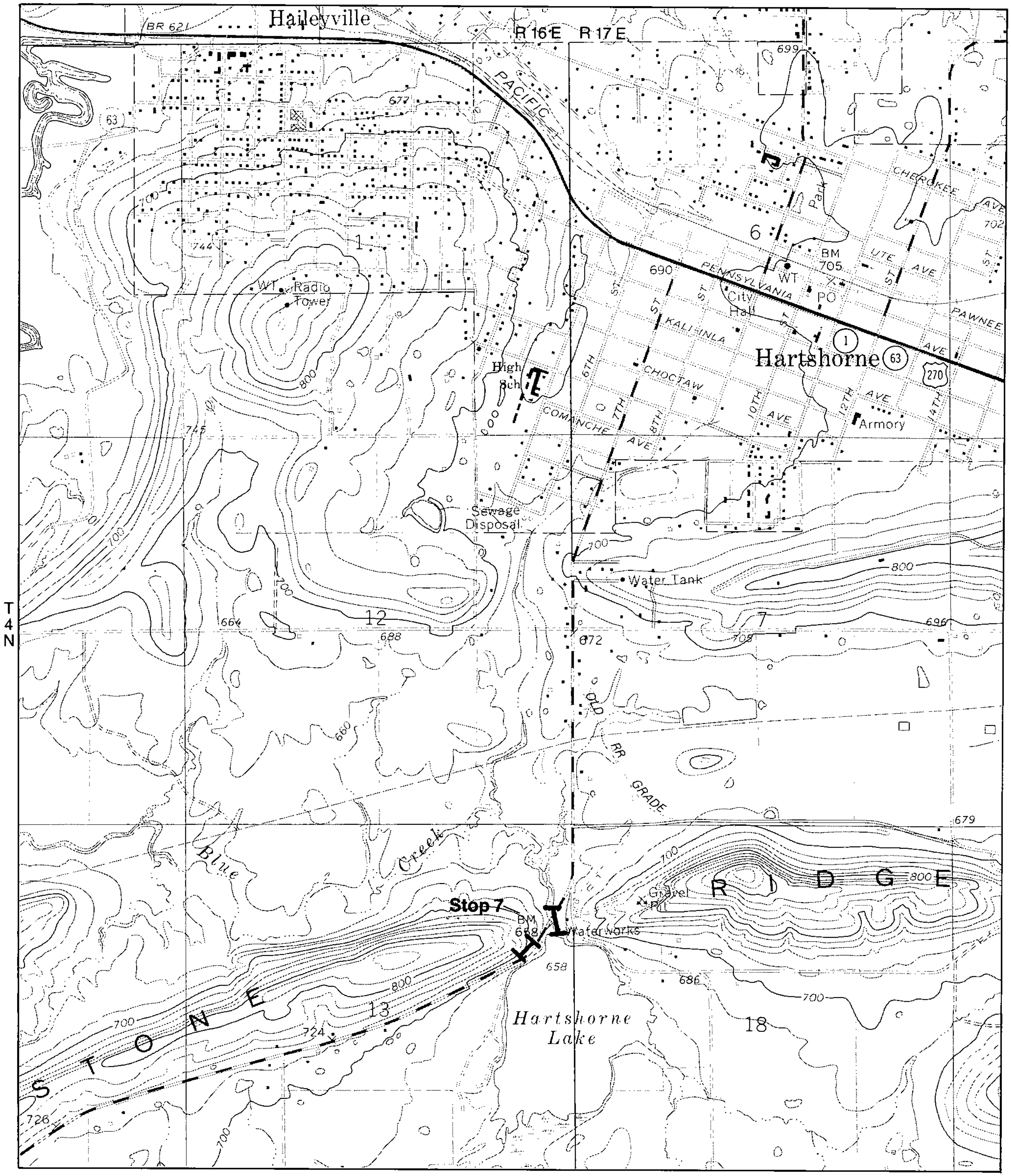


Figure 1. Location of Stop 7, in a water gap just north of Hartshorne Lake, Pittsburg County. Base from Hartshorne 7.5′ Quadrangle.

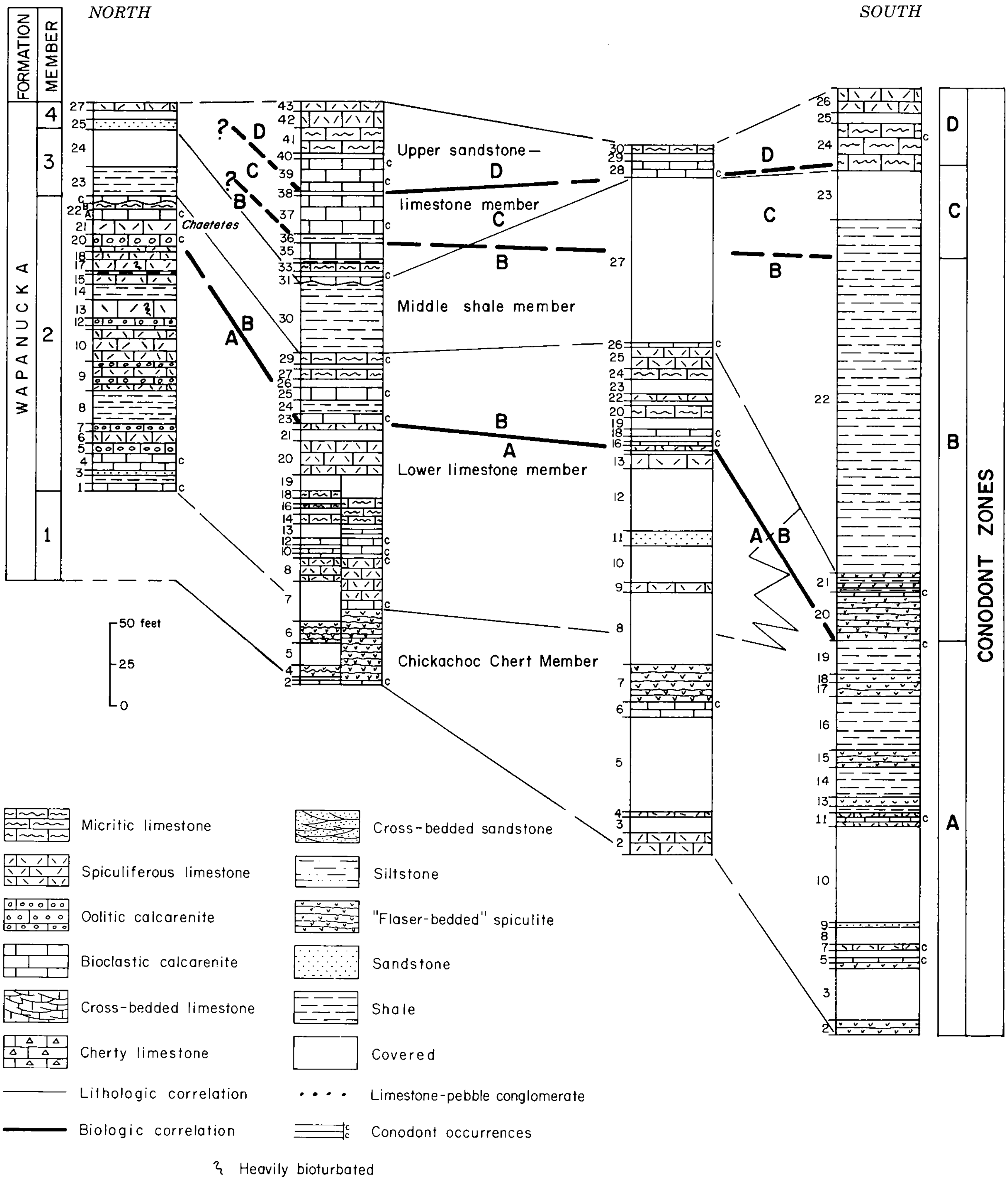


Figure 2. Basinward facies changes in Wapanucka Formation adjacent to Indian Nations Turnpike south of McAlester (reproduced from Grayson, 1979). These measured sections represent an area ~12 mi west–southwest of Stop 7. The sections were measured within ~3 mi of each other; however, the original distance was much greater (tens of miles) between each section, as they are from separate fault blocks that were imbricated during the Ouachita orogeny. Conodont zones are (A) *Idiognathoides convexus*, (B) *Neognathodus kanumai–Idiognathoides ouachitensis*, (C) *Diplognathodus orphanus*, (D) *Streptognathodus elegantulus*. The Morrowan–Atokan boundary is the boundary between zones C and D.

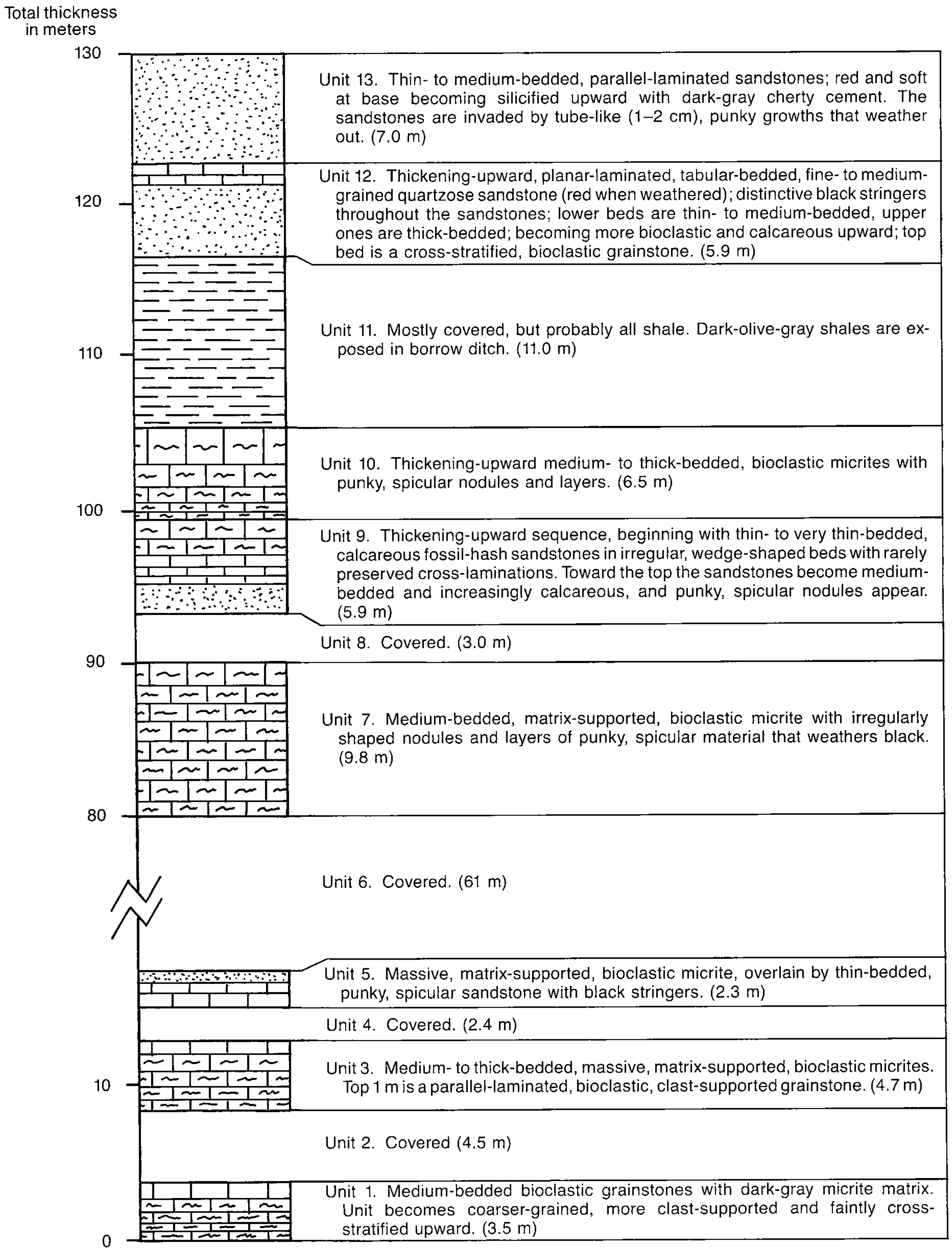


Figure 3. Stratigraphic section of the Wapanucka Formation in the water gap north of Hartshorne Lake, Stop 7. Lithologic symbols as in Figure 2.

COLLECTED REFERENCES

American Society for Testing and Materials, 1987, Gaseous fuels; coal and coke, *in* Annual book of ASTM standards; petroleum products, lubricants, and fossil fuels: American Society for Testing and Materials, v. 05.05, 561 p.

Amsden, T. W., 1980, Hunton Group (Late Ordovician, Silurian, and Early Devonian) in the Arkoma basin of Oklahoma: Oklahoma Geological Survey Bulletin 129, 136 p.

Arbenz, J. K., 1984, A structural cross-section through the Ouachita Mountains of western Arkansas, *in* Stone, C. G.; and Haley, B. R. (eds.), A guidebook to the geology of the central and southern Ouachita Mountains, Arkansas: Arkansas Geological Commission Guidebook 84-2, p. 76–84.

Archinal, B. E., 1979, Atoka Formation (Pennsylvanian), deposition and contemporaneous structural movement, southwestern Arkoma basin, Oklahoma, *in* Hyne, N. S. (ed.), Pennsylvanian sandstones of the mid-continent: Tulsa Geological Society Special Publication 1, p. 259–267.

Bartlett, C. S., Jr., 1966, Arkoma's search for porosity broadens: Oil and Gas Journal, January 24, p. 126–127,129,131–132,144.

Berry, R. M.; and Trumbly, W. D., 1968, Wilburton Gas Field, Arkoma basin, Oklahoma, *in* Cline, L. M. (ed.), Guidebook to the geology of the western Arkoma basin and Ouachita Mountains, Oklahoma: Oklahoma City Geological Society Guidebook, p. 86–103.

Bouma, A. H., 1962, Sedimentology of some flysch deposits: Elsevier, Amsterdam, 168 p.

Bowsher, A. L.; and Johnson, N. L., 1968, Road log for second day, *in* Cline, L. M. (ed.), A guidebook to the geology of the western Arkoma basin and Ouachita Mountains, Oklahoma: Oklahoma City Geological Society Guidebook, p. 24–62.

Boyer, S. E.; and Elliott, D., 1982, Thrust systems: American Association of Petroleum Geologists Bulletin, v. 66, p. 1196–1230.

Branan, C. B., Jr., 1966, Natural gas in Arkoma basin of Oklahoma and Arkansas, *in* Beebe, B. W.; and Curtis, B. F. (eds.), Natural gases of North America: American Association of Petroleum Geologists Memoir 9, v. 2, p. 1616–1635.

Branan, C. B., Jr.; and Jordan, Louise, 1960, Recent exploration in the Arkoma basin and Ouachita Province, southeastern Oklahoma: Oklahoma Geology Notes, v. 20, p. 140–147.

Branson, C. C., 1961, Pennsylvanian system of the Arkoma basin and of the Midcontinent platform: Proceedings of the Seventh Biennial Geological Symposium, University of Oklahoma, Norman, v. 7, p. 179–194.

Briggs, G., 1962, A paleocurrent study of upper Mississippian and lower Pennsylvanian rocks of the Ouachita Mountains and Arkoma basin, southeastern Oklahoma: University of Wisconsin, Madison, unpublished Ph.D. dissertation, 123 p.

——— 1974, Carboniferous depositional environments in the Ouachita–Arkoma basin area of southeastern Oklahoma, *in* Briggs, G. (ed.), Carboniferous of the southeastern United States: Geological Society of America Special Paper 148, p. 225–239.

Briggs, G.; and Cline, L. M., 1967, Paleocurrents and source areas of late Paleozoic sediments of the Ouachita Mountains, southeastern Oklahoma: Journal of Sedimentary Petrology, v. 37, p. 985–1000.

Briggs, G.; and Roeder, Dietrich, 1975, Sedimentation and plate tectonics, Ouachita Mountains and Arkoma basin, *in* Briggs, G.; McBride, E. F.; and Moiola, R. J. (eds.), Sedimentology of Paleozoic flysch and associated deposits, Ouachita Mountains–Arkoma basin, Oklahoma: Dallas Geological Society Guidebook, p. 1–22.

Briggs, G.; McBride, E. F.; and Moiola, R. J. (eds.), 1975, Sedimentology of Paleozoic flysch and associated deposits, Ouachita Mountains–Arkoma basin, Oklahoma: Dallas Geological Society Guidebook, 128 p.

Buchanan, R. S.; and Johnson, F. K., 1968, Bonanza gas field—a model for Arkoma basin growth faulting, *in* Cline, L. M. (ed.), A guidebook to the geology of the western Arkoma basin and Ouachita Mountains, Oklahoma: Oklahoma City Geological Society Guidebook, p. 75–85.

Burchfield, M. R., 1985, Map of Oklahoma oil and gas fields: Oklahoma Geological Survey Map GM-28, scale 1:500,000.

Cardott, B. J.; Hemish, L. A.; Johnson, C. R.; and Luza, K. V., 1986, The relationship between coal rank and present geothermal gradient in the Arkoma basin, Oklahoma: Oklahoma Geological Survey Special Publication 86-4, 65 p.

Chamberlain, C. K., 1970, Trace fossils and paleoecology of the Ouachita Mountains of southeastern Oklahoma: University of Wisconsin, Madison, unpublished Ph.D. dissertation, 106 p.

——— 1971, Bathymetry and paleoecology of Ouachita geosyncline of southeastern Oklahoma as determined from trace fossils: American Association of Petroleum Geologists Bulletin, v. 55, p. 34–50.

——— (ed.), 1978, A guidebook to the trace fossils and paleoecology of the Ouachita Mountains: Society of Economic Paleontologists and Mineralogists Guidebook, 68 p.

Cline, L. M., 1960, Stratigraphy of the late Paleozoic rocks of the Ouachita Mountains, Oklahoma: Oklahoma Geological Survey Bulletin 85, 113 p.

——— (ed.), 1968, A guidebook to the geology of the western Arkoma basin and Ouachita Mountains, Oklahoma: Oklahoma City Geological Society Guidebook, 126 p.

——— 1970, Sedimentary features of late Paleozoic flysch, Ouachita Mountains, Oklahoma: Geological Association of Canada Special Paper 7, p. 85–101.

Cline, L. M.; Hilseweck, W. J.; and Feray, D. E. (eds.), 1959, The geology of the Ouachita Mountains—a symposium: Dallas Geological Society and Ardmore Geological Society Guidebook, 208 p.

Coleman, J. M.; and Gagliano, S. M., 1964, Cyclic sedimentation in the Mississippi River deltaic plain: Transactions of the Gulf Coast Association of Geological Societies, v. 14, p. 67–80.

——— 1965, Sedimentary structures—Mississippi River deltaic plain, *in* Middleton, G. V. (ed.), Primary sedimentary structures and their hydrodynamic interpretation: Society of Economic Paleontologists and

Mineralogists Special Publication 12, p. 133–148.
Decker, R. E.; and Black, B. (eds.), 1976, A study of Paleozoic rocks in Arbuckle and Ouachita Mountains of southern Oklahoma: Gulf Coast Association of Geological Societies Guidebook, 132 p.
Dogan, N., 1969, A subsurface study of Middle Pennsylvanian rocks (from the Brown limestone to the Checkerboard Limestone) in east-central Oklahoma: University of Tulsa unpublished M.S. thesis, 67 p.
England, R. L., 1961, Subsurface study of the Hunton Group (Silurian–Devonian) in the Oklahoma portion of the Arkoma basin: Shale Shaker Digest, v. 12–14, p. 19–35.
Fay, R. O., 1984, Geologic map of western Ouachita Mountains in parts of Atoka, Pittsburg, and Pushmataha Counties, Oklahoma: Oklahoma City Geological Society Discussion Group, Map No. 1, scale 1:63,360.
Fellows, L. D., 1964, Geology of the western part of Winding Stair Range, Latimer and Le Flore Counties, Oklahoma: Oklahoma Geological Survey Circular 65, 102 p.
Ferguson, C. A.; and Suneson, N. H., 1988, Arbuckle source for Atoka Formation flysch, Ouachita Mountains frontal belt, Oklahoma; new evidence from paleocurrents [abstract]: American Association of Petroleum Geologists Bulletin, v. 72, p. 184–185.
Fisher, W. L., 1969, Facies characterization of Gulf Coast basin delta systems, with some Holocene analogues, *in* Geology of the American Mediterranean: Transactions of the Gulf Coast Association of Geological Societies, v. 19, p. 239–261.
Flawn, P. T.; Goldstein, August, Jr.; King, P. B.; and Weaver, C. E., 1961, The Ouachita System: Bureau of Economic Geology, University of Texas Publication 6120, 401 p.
Frezon, S. E., 1962, Correlation of Paleozoic rocks from Coal County, Oklahoma, to Sebastian County, Arkansas: Oklahoma Geological Survey Circular 58, 53 p.
Friedman, S. A., 1974, An investigation of the coal reserves in the Ozarks section of Oklahoma and their potential uses; final report to the Ozarks Regional Commission: Oklahoma Geological Survey Special Publication 74-2, 117 p.
——— 1978, Desmoinesian coal deposits in part of the Arkoma basin, eastern Oklahoma: American Association of Petroleum Geologists Guidebook, field trip 2, Oklahoma City Geological Society, 62 p.
Gordon, M., Jr.; and Stone, C. G., 1977, Correlation of the Carboniferous rocks of the Ouachita trough with those of the adjacent foreland, *in* Stone, C. G. (ed.), Symposium on the geology of the Ouachita Mountains: Arkansas Geological Commission Miscellaneous Publication 13, p. 70–91.
Graham, S. A.; Ingersoll, R. V.; and Dickinson, W. R., 1976, Common provenance of lithic grains in Carboniferous sandstones from Ouachita Mountains and Black Warrior basin: Journal of Sedimentary Petrology, v. 46, p. 620–632.
Grayson, R. C., 1979, Stop descriptions—fifth day, *in* Sutherland, P. K.; and Manger, W. L. (eds.), Mississippian–Pennsylvanian shelf-to-basin transition, Ozark and Ouachita regions, Oklahoma and Arkansas: Oklahoma Geological Survey Guidebook 19, p. 67–76.
——— 1981, Distribution and significance of allochthonous conodonts in early Pennsylvanian rocks of eastern Oklahoma [abstract]: Geological Society of America Abstracts with Programs, v. 13, p. 279.
Haines, R. A., 1981, Middle Atokan delta systems in the Arkoma basin of Arkansas, *in* Technical proceedings of the American Association of Petroleum Geologists Mid-Continent Regional Meeting: Oklahoma City Geological Society, p. 42–49.
Haley, B. R., 1982, Geology and energy resources of the Arkoma basin, Oklahoma and Arkansas, *in* Procter, P. D.; and Koenig, J. W. (eds.), Selected structural basins of the Midcontinent, U.S.A.: University of Missouri–Rolla, UMR Journal, no. 3, p. 43–53.
Ham, W. E., 1959, Correlation of pre-Stanley strata in the Arbuckle–Ouachita Mountain regions, *in* Cline, L. M.; Hilseweck, W. J.; and Feray, D. E. (eds.), The geology of the Ouachita Mountains—a symposium: Dallas Geological Society and Ardmore Geological Society Guidebook, p. 71–86.
Hardie, W. E., 1986, Surface to subsurface structural interpretation of the Pittsburg Quadrangle, Pittsburg County, Oklahoma: Baylor University unpublished B.S. thesis, 93 p.
——— 1987, Surface to subsurface structural interpretation of the Pittsburg Quadrangle, Pittsburg County, Oklahoma [abstract]: Geological Society of America Abstracts with Programs, v. 19, p. 169.
Hare, M. G. (ed.), 1973, A study of Paleozoic rocks in Arbuckle and western Ouachita Mountains of southern Oklahoma: Shreveport Geological Society Guidebook, 77 p.
Hart, O. D., 1963, Geology of the eastern part of Winding Stair Range, Le Flore County, Oklahoma: Oklahoma Geological Survey Bulletin 103, 87 p.
Hemish, L. A., 1986a, Stratigraphy of the lower part of the Boggy Formation (Desmoinesian) in northwestern Muskogee and southwestern Wagoner Counties, Oklahoma: Oklahoma Geology Notes, v. 46, p. 168–187.
——— 1986b, Coal geology of Craig County and eastern Nowata County, Oklahoma: Oklahoma Geological Survey Bulletin 140, 131 p.
——— 1987, Names of coal beds in the northeastern Oklahoma shelf area: Oklahoma Geology Notes, v. 47, p. 96–113.
——— 1988a, Report of core-drilling by the Oklahoma Geological Survey in Pennsylvanian-age rocks of the northeastern Oklahoma coal belt, 1983–86: Oklahoma Geological Survey Special Publication 88-2, 174 p.
——— 1988b, Coalescence of the Secor and Secor rider coal beds in the Shady Grove Creek area, northeastern McIntosh County, Oklahoma, with interpretations concerning depositional environments: Oklahoma Geology Notes, v. 48, p. 100–119.
——— [in preparation a], Coal geology of parts of Creek, Tulsa, Wagoner, and Washington Counties, Oklahoma: Oklahoma Geological Survey.
——— [in preparation b], Coal geology of McIntosh and Muskogee Counties, Oklahoma: Oklahoma Geological Survey.
Hendricks, T. A., 1958, Interpretation of Ouachita Mountains of Oklahoma as autochthonous folded belt: preliminary report—discussion: American Association of Petroleum Geologists Bulletin, v. 42, p. 2757–2765.
——— 1959, Structure of the frontal belt of the Ouachita Mountains, *in* Cline, L. M.; Hilseweck, W. J.; and Feray, D. E. (eds.), The geology of the Ouachita Mountains—a symposium: Dallas Geological Society and Ardmore Geological Society Guidebook, p. 44–56.
Hendricks, T. A.; Knechtel, M. M.; and Bridge, J., 1937, Geology of Black Knob Ridge, Oklahoma: American Association of Petroleum Geologists Bulletin, v. 21, p. 1–29.
Hendricks, T. A.; Gardner, L. S.; Knechtel, M. M.; and Averitt, P., 1947, Geology of the western part of the

Ouachita Mountains in Oklahoma: U.S. Geological Survey Oil and Gas Investigations Series, Preliminary Map 66, scale 1:42,240.

Holden, F. T., 1965, The Simpson Group of the Arkoma basin, Oklahoma and Arkansas: Tulsa Geological Society Digest, v. 33, p. 134–143.

Houseknecht, D. W., 1986, Evolution from passive margin to foreland basin: the Atoka Formation of the Arkoma basin, south-central U.S.A., *in* Allen, P. A.; and Homewood, P. (eds.), Foreland basins: International Association of Sedimentologists Special Publication 8, p. 327–345.

Houseknecht, D. W.; and Kacena, J. A., 1983, Tectonic and sedimentary evolution of the Arkoma foreland basin, *in* Houseknecht, D. W. (ed.), Tectonic-sedimentary evolution of the Arkoma basin: Society of Economic Paleontologists and Mineralogists Midcontinent Section, v. 1, p. 3–33.

Johnson, K. S.; Amsden, T. W.; Denison, R. E.; Dutton, S. P.; Goldstein, A. G.; Rascoe, B., Jr.; Sutherland, P. K.; and Thompson, D. M. [in press], Southern Midcontinent region, *in* Sloss, L. L. (ed.), Sedimentary cover—North American craton, U.S.: The Geology of North America, Geological Society of America, Boulder, v. D-2.

Keller, G. R.; and Cebull, S. E., 1973, Plate tectonics and the Ouachita system in Texas, Oklahoma, and Arkansas: Geological Society of America Bulletin, v. 83, p. 1659–1666.

Knight, K. S., 1985, Lithostratigraphy, facies, and petrology of the Atoka Formation (Pennsylvanian), eastern frontal Ouachita Mountains, southeastern Oklahoma: University of Arkansas unpublished M.S. thesis, 125 p.

Koinm, D. M.; and Dickey, P. A., 1967, Growth faulting in McAlester basin of Oklahoma: American Association of Petroleum Geologists Bulletin, v. 51, p. 710–718.

Kramer, W. B., 1933, Boulders from Bengalia: Journal of Geology, v. 41, p. 590–621.

Leander, M. H.; and Legg, T. E., 1988, Potential for subthrust gas fields: results of recent deep drilling in the Ouachitas [abstract]: American Association of Petroleum Geologists Bulletin, v. 72, p. 211.

Leonhardt, F. H., 1983, Deformational history of Black Knob Ridge, southeastern Oklahoma: University of Texas, Dallas, unpublished M.S. thesis, 83 p.

Leonhardt, F. H.; and Aiken, C., 1984, Gravity study of some shallow and deep structures in the Ouachita tectonic belt, Oklahoma [abstract]: Geophysics, v. 49, p. 618.

Lillie, R. J.; Nelson, K. D.; De Voogd, B.; Brewer, J. A.; Oliver, J. E.; Brown, L. D.; Kaufman, S.; and Viele, G. W., 1983, Crustal structure of Ouachita Mountains, Arkansas: a model based on integration of COCORP reflection profiles and regional geophysical data: American Association of Petroleum Geologists Bulletin, v. 67, p. 907–931.

Mack, G. H.; James, W. C.; and Thomas, W. A., 1981, Orogenic provenance of Mississippian sandstones associated with southern Appalachian–Ouachita Orogen: American Association of Petroleum Geologists Bulletin, v. 65, p. 1444–1456.

Mankin, C. J. (coordinator), 1987, Correlation of stratigraphic units of North America (COSUNA) project; Texas–Oklahoma tectonic region correlation chart: American Association of Petroleum Geologists.

Marcher, M. V.; and Bergman, D. L., 1983, Reconnaissance of the water resources of the McAlester and Texarkana Quadrangles, southeastern Oklahoma: Oklahoma Geological Survey Hydrologic Atlas 9, scale 1:250,000, 4 sheets.

McCaslin, J. C., 1982, Earlsboro gas find to stir interest in Spiro sands of Oklahoma's Arkoma basin: Oil and Gas Journal, June 7, p. 214–216.

Misch, P.; and Oles, K. F., 1957, Interpretation of Ouachita Mountains of Oklahoma as autochthonous folded belt: preliminary report: American Association of Petroleum Geologists Bulletin, v. 41, p. 1899–1905.

——— 1958, Interpretation of Ouachita Mountains of Oklahoma as autochthonous folded belt: preliminary report—reply: American Association of Petroleum Geologists Bulletin, v. 42, p. 2765–2783.

Miser, H. D., 1921, Llanoria, the Paleozoic land area in Louisiana, and eastern Texas: American Journal of Science, v. 2 (series 5), no. 8, p. 61–89.

——— 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas: Oklahoma Geological Survey Bulletin 50, 30 p.

——— 1954, Geologic map of Oklahoma: U.S. Geological Survey and Oklahoma Geological Survey, scale 1:500,000, 2 sheets.

Moiola, R. J.; and Chanmugam, G., 1984, Submarine fan sedimentation, Ouachita Mountains, Arkansas and Oklahoma: Transactions of the Gulf Coast Association Geological Society, v. 34, p. 175–182.

Morris, R. C., 1974a, Sedimentary and tectonic history of the Ouachita Mountains, *in* Dickinson, W. R. (ed.), Tectonics and sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication 22, p. 120–142.

——— 1974b, Carboniferous rocks of the Ouachita Mountains, Arkansas: a study of facies patterns along the unstable slope and axis of a flysch trough: Geological Society of America Special Paper 148, p. 241–279.

Nielsen, K. C.; and Leonhardt, F. H., 1983, Deformational sequence of frontal zone thrust sheets, Ouachita Mountains, Oklahoma [abstract]: Geological Society of America Abstracts with Programs, v. 15, p. 653.

Nooncaster, J. R., 1985, Lithostratigraphy, petrology, and facies, Atoka Formation (Pennsylvanian), western frontal Ouachita Mountains, southeastern Oklahoma: University of Arkansas, Fayetteville, unpublished M.S. thesis, 125 p.

Oakes, M. C., 1977, Geology and mineral resources (exclusive of petroleum) of Muskogee County, Oklahoma: Oklahoma Geological Survey Bulletin 122, 78 p.

Oakes, M. C.; and Koontz, Terry, 1967, Geology and petroleum of McIntosh County, Oklahoma: Oklahoma Geological Survey Bulletin 111, 88 p.

Oil and Gas Journal, 1988, Red Oak-Norris infill program nearly complete: Oil and Gas Journal, February 8, p. 30.

Oklahoma Geological Survey, 1954, Names of Oklahoma coal beds: The Hopper, v. 14, p. 121–132.

Pantin, H. M.; and Leeder, M. R., 1987, Reverse flow in turbidity currents: the role of internal solitons: Sedimentology, v. 34, p. 1143–1155.

Petroleum Information, 1988, Arco releases details at deep zone discovery in Latimer County, Oklahoma: Petroleum Information, Oklahoma Report, v. 35, no. 12, p. 1.

Phares, R. S., 1969, Depositional framework of the Bartlesville sandstone in northeastern Oklahoma: University of Tulsa unpublished M.S. thesis, 59 p.

Pickering, H. M.; and Hiscott, R. N., 1985, Contained (reflected) turbidity currents from the Middle Ordovician Cloridorme Formation, Quebec, Canada: an alternative to the antidune hypothesis: Sedimentology, v. 32, p. 373–394.

Powers, S., 1928, Age of the folding of the Oklahoma Mountains: Geological Society of America Bulletin, v. 39, p.

1031–1072.
Rascoe, B., Jr.; and Adler, F. J., 1983, Permo-Carboniferous hydrocarbon accumulations, Mid-Continent, U.S.A.: American Association of Petroleum Geologists Bulletin, v. 67, p. 979–1001.
Reidenouer, David; Williams, E. G.; and Dutcher, R. R., 1979, The relationship between paleotopography and sulfur distribution in some coals of western Pennsylvania, *in* Ferm, J. C.; and Horne, J. C. (eds.), Carboniferous depositional environments in the Appalachian Region: Carolina Coal Group, Dept. of Geology, University of South Carolina, p. 88–101.
Reinemund, J. A.; and Danilchik, W., 1957, Preliminary geologic map of the Waldron Quadrangle and adjacent areas, Scott County, Arkansas: U.S. Geological Survey Oil and Gas Investigations Series, Map OM 192, scale 1:48,000.
Richardson, E. R., 1987, Pittsburg Field complex offers opportunities for additional well locations: Oil and Gas Journal, June 15, p. 64–66.
Rippee, D. S., 1981, Geology along a cross section through the frontal Ouachita Mountains in Pittsburg, Atoka, and Pushmataha Counties, Oklahoma: University of Oklahoma unpublished M.S. thesis, 90 p.
Saitta, B. S., 1968, Bluejacket Formation—a subsurface study in northeastern Oklahoma: University of Tulsa unpublished M.S. thesis, 142 p.
Seely, D. R., 1963, Structure and stratigraphy of the Rich Mountain area, Oklahoma and Arkansas: Oklahoma Geological Survey Bulletin 101, 173 p.
Shideler, G. L., 1968, Petrography and provenance of the Johns Valley boulders, Ouachita Mountains, southeastern Oklahoma and southwestern Arkansas: University of Wisconsin, Madison, unpublished Ph.D. dissertation, 133 p.
——— 1970, Provenance of Johns Valley boulders in late Paleozoic Ouachita facies, southeastern Oklahoma and southwestern Arkansas: American Association of Petroleum Geologists Bulletin, v. 54, p. 789–806.
Six, D. A., 1966, Red Oak-Norris Gas Field, Brazil anticline, Latimer and Le Flore Counties, Oklahoma, *in* Beebe, B. W.; and Curtis, B. F. (eds.), Natural gases of North America: American Association of Petroleum Geologists Memoir 9, v. 2, p. 1644–1657.
Sloss, L. L., 1962, Stratigraphic models in exploration: American Association of Petroleum Geologists Bulletin, v. 46, p. 1050–1057.
——— 1963, Sequences in the cratonic interior of North America: Geological Society of America Bulletin, v. 74, p. 93–114.
Stach, E.; Mackowsky, M.-Th.; Teichmuller, M.; Taylor, G. H.; Chandra, D.; and Teichmuller, R., 1975, Coal petrology: Gebruder Borntraeger, Berlin, 428 p.
Suhm, R., 1979, Hydrocarbon potential of Simpson equivalents in Arkansas's Arkoma basin: Oil and Gas Journal, February 12, p. 150–153.
Suneson, N. H.; and Ferguson, C. A., 1987, Ouachita Mountains frontal belt, Oklahoma [abstract]: Geological Society of America Abstracts with Programs, v. 19, p. 860.
Sutherland, P. K. [in press], Arkoma basin, *in* Sloss, L. L. (ed.), Sedimentary cover—North American craton, U.S.: The Geology of North America, Geological Society of America, Boulder, v. D-2.
Sutherland, P. K.; and Manger, R. L. (eds.), 1984a, The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, 198 p.
——— 1984b, The Atokan Series: an interval in search of a name, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 9–18.
Sutherland, P. K.; Archinal, B. E.; and Grubbs, R. K., 1982, Morrowan and Atokan (Pennsylvanian) stratigraphy in the Arbuckle Mountains area, Oklahoma, *in* Sutherland, P. K. (ed.), Lower and middle Pennsylvanian stratigraphy in south-central Oklahoma: Oklahoma Geological Survey Guidebook 20, p. 1–18.
Swanson, R.; Goldstein, A., Jr.; Koontz, T.; and Bowsher, A. L., 1961, Symposium on the Arkoma basin: Tulsa Geological Society Digest, v. 29, p. 54–88.
Taff, J. A., 1902, Description of the Atoka Quadrangle [Indian Territory]: U.S. Geological Survey Geologic Atlas, Folio 79, 8 p.
Taff, J. A.; and Adams, G. I., 1900, Geology of the eastern Choctaw coal field, Indian Territory: U.S. Geological Survey 21st Annual Report, p. 257–311.
Thomas, W. A., 1985, The Appalachian–Ouachita connection: Paleozoic orogenic belt at the southern margin of North America: Annual Review of Earth and Planetary Science, v. 13, p. 175–199.
Trumbull, J. V. A., 1960, Coal fields of the United States, exclusive of Alaska: U.S. Geological Survey map, scale 1:5,000,000, sheet 1.
Tulsa Geological Society, 1947, Field conference in western part of the Ouachita Mountains in Oklahoma: Tulsa Geological Society Guidebook, 56 p.
Ulrich, E. O., 1927, Fossiliferous boulders in the Ouachita "Caney" shale and the age of the shale containing them: Oklahoma Geological Survey Bulletin 45, 48 p.
Vedros, S. G.; and Visher, G. S., 1978, The Red Oak Sandstone: a hydrocarbon-producing submarine fan deposit, *in* Stanley, D. J.; and Kelling, G. (eds.), Sedimentation in submarine canyons, fans, and trenches: Dowden, Hutchinson, and Ross, Stroudsburg, Pennsylvania, p. 279–308.
Visher, G. S., 1968, Depositional framework of the Bluejacket–Bartlesville sandstone, *in* Visher, G. S. (ed.), Geology of the Bluejacket–Bartlesville sandstone of Oklahoma: Oklahoma City Geological Society Guidebook, p. 33–51.
——— 1969, Grain size distributions and depositional processes: Journal of Sedimentary Petrology, v. 39, no. 3, p. 1074–1106.
Walker, R. G., 1978, Deep-water sandstone facies and ancient submarine fans: models for exploration for stratigraphic traps: American Association of Petroleum Geologists Bulletin, v. 62, p. 932–966.
Walper, J. L., 1977, Paleozoic tectonics of the southern margin of North America: Transactions of the Gulf Coast Association Geological Society, v. 27, p. 230–241.
Ward, C. R., 1984, Geology of coal, *in* Ward, C. R. (ed.), Coal geology and coal technology: Blackwell Scientific Publications, Boston, p. 151–176.
Weirich, T. E., 1968, History of the Bartlesville oil sand, *in* Visher, G. S. (ed.), Geology of the Bluejacket–Bartlesville sandstone of Oklahoma: Oklahoma City Geological Society Guidebook, p. 69–72.
White, David; and Thiessen, Reinhardt, 1913, The origin of coal: U.S. Bureau of Mines Bulletin 38, 390 p.
Wickham, J., 1978, Ouachita foldbelt: a Paleozoic continental margin, *in* Field guide to structure and stratigraphy of the Ouachita Mountains and the Arkoma basin: American Association of Petroleum Geologists Annual Meeting, Oklahoma City, p. 35–59.
Williams, E. G.; and Keith, M. L., 1979, Relationship between sulfur in coals and the occurrence of marine roof

beds, *in* Ferm, J. C.; and Horne, J. C. (eds.), Carboniferous depositional environments in the Appalachian Region: Carolina Coal Group, Dept. of Geology, University of South Carolina, p. 102–109.

Zachry, D. L., 1979, Early Pennsylvanian braided stream sedimentation, northwest Arkansas, *in* Hyne, N. S. (ed.), Pennsylvanian sandstones of the mid-continent: Tulsa Geological Society Special Publication 1, p. 269–281.

Zachry, D. L.; and Sutherland, P. K., 1984, Stratigraphy and depositional framework of the Atoka Formation (Pennsylvanian) Arkoma basin of Arkansas and Oklahoma, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 9–18.