GUIDEBOOK 86-3

STATE OF ARKANSAS

ARKANSAS GEOLOGICAL COMMISSION

Norman F. Williams, State Geologist

SEDIMENTARY AND IGNEOUS ROCKS OF THE OUACHITA MOUNTAINS OF ARKANSAS

A Guidebook with Contributed Papers

Part 2

BY
CHARLES G. STONE and BOYD R. HALEY

With Contributions By
Timothy L. Cox, William D. Hart, James E. Jennings, Martin H. Link,
Ellen Mullen Morris, Michael T. Roberts, James Stitt, and Jay Zimmerman

Prepared for
THE GEOLOGICAL SOCIETY OF AMERICA
Annual Meeting
San Antonio, Texas
October, 1986

Little Rock, Arkansas
November, 1986

CORRELATION OF PALEOZOIC ROCKS IN THE OZARK, ARKANSAS VALLEY, AND OUACHITA MOUNTAIN REGIONS, ARK.

AGE			OZARK – ARKANSAS VALLEY SECTION		MAP SYM.		OUACHITA MTN. SECTION		MAP SYM.	
CARBONIFEROUS SYSTEM	PENNSYLVANIAN	DES MOINES	Boggy Fm.		ℙby		Missing			
			Savanna Fm.		ℙsv					
			McAlester Fm.		ℙma					
			Hartshorne Sandstone		ℙhs					
		ATOKA	Atoka Fm.		ℙa		Atoka Fm.		ℙa	
		MORROW	Bloyd Shale.	Kessler Ls Mbr.	ℙbh	ℙpk	Johns Valley Shale		ℙjv	
				Woolsey Mbr		ℙpw				
				Brentwood Ls Mbr		ℙpb				
			Hale Fm.	Prairie Grove Mbr		ℙhp	Jackfork Fm.		ℙj	
				Cane Hill Mbr	ℙhc					
	MISSISSIPPIAN	UPPER	Pitkin Limestone		Mpfb	Mp	Stanley Shale	Chickasaw Creek Mbr.	Ms	
			Fayetteville Shale	Wedington SS Mbr.		Mf				
			Batesville Sandstone	Hindsville Ls. Mbr.		Mbh				
			Ruddell Shale		Mr			Hatton Tuff		
			Moorefield Fm.		Mm			Hot Springs SS Mbr.		
		LOWER	Boone Fm.	Short Creek Oolite Mbr.	Mb		Arkansas Novaculite	Upper Div.	MDa	
				St Joe Ls. Mbr.						
			Chattanooga Shale		MDcp					
DEVONIAN		UPPER		Sylamore SS				Middle Div.		
		MIDDLE	Clifty Limestone					Lower Div.		
		LOWER	Penters Chert							
SILURIAN		UPPER	Missing		Slsb		Missouri Mountain Shale		SmOpc	Smb
			Lafferty Limestone				Blaylock Sandstone			
			St. Clair Limestone							
		LOWER	Brassfield Limestone							
ORDOVICIAN		UPPER	Cason Shale		Ocj		Polk Creek Shale			Opc
			Fernvale Limestone		Of		Bigfork Chert		Obf	
		MIDDLE	Kimmswick Limestone							
			Plattin Limestone				Womble Shale		Ow	
			Joachim Dolomite							
			St. Peter Sandstone		Ose					
			Everton Fm.	Jasper Ls Mbr.			Blakely Sandstone		Ob	
				Newton SS Mbr						
				King River SS Mbr.						
		LOWER	Powell Dolomite		Op		Mazarn Shale		Om	
			Cotter Dolomite		Ocjc					
			Jefferson City Dolomite				Crystal Mountain Sandstone		Ocm	
			Roubidoux Fm.		Not exposed		Collier Shale		Oc	
			Gasconade-VanBuren Fm	Gunter Mbr.						
CAMBRIAN		UPPER	Eminence Dolomite							
			Potosi Dolomite							
			Derby-Doerun-Davis Fm							
			Bonneterre Dolomite				Basal Collier and Older rocks not exposed			
			Lamotte Sandstone							
PRE-C			Igneous Rocks							

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FOREWORD

The primary purpose of this two-volume guidebook is to demonstrate the complex geologic environment of the rocks forming the central and eastern parts of the Ouachita Mountains of Arkansas. The sites were selected so as to emphasize (1) the various depositional environments of the formations composing this vast wedge of Upper Cambrian to Middle Pennsylvanian sedimentary rocks and (2) the deformational history of the rocks during the late Paleozoic Ouachita orogeny. Commensurate with their very small volume and limited outcrops, much less emphasis is placed on the allochthonous serpentinite and metagabbro masses of probable Precambrian age and the post-orogenic, alkalic and associated intrusive bodies of Cretaceous age.

Part 1, the main volume of the set, was rapidly assembled for a field trip planned in 1985 by another Society, but subsequently canceled. Most of the stops included are appropriate for the Geological Society of America's field excursion. However, altered goals and different logistics require that some stops not be visited. To avoid reformatting and reprinting Part 1, we have retained the original order and the original stop numbers, although the order in which they will now be visited will be different. A listing comparing the original and revised orders of the stops will be distributed to attendees before the trip begins. For the present field trip, additional stops in the easternmost part of the Ouachita Mountains have been included. They are described in Part 2 of this set.

Both volumes contain short papers contributed by other workers active in the area. The papers provide a sampling of recent types of studies being conducted in the Ouachita Mountains and also a broader interpretive context for the stop descriptions. Our sincere thanks to those who have taken the effort to prepare these papers. Their names, which appear on the covers and title pages, as well as on the papers, will not be repeated here.

We are indebted to George W. Colton, who assisted with many of the editorial chores, and to John David McFarland, III, who provided most of the photographs that accompany the stop descriptions. Special thanks are also expressed to Charles B. Germer, Rufus J. LeBlanc, John Long and Buster and Johnny Warner for their kind assistance. We thank Adrian Hunter, Virginia Snyder, Susan Young and other personnel of the Arkansas Geological Commission for their diligent efforts in the compilation of this volume.

Charles G. Stone
Boyd R. Haley,
November 1, 1987

CONTENTS

STOP DESCRIPTIONS

CONTRIBUTED PAPERS

STOP DESCRIPTIONS
EASTERN OUACHITA MOUNTAINS, ARKANSAS

STOP 1 -- JACKFORK SANDSTONE AT THE ABANDONED BIG ROCK QUARRY, NORTH LITTLE ROCK

By
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This abandoned quarry (Fig. 1) is in the lower part of the upper Jackfork Sandstone and was active in the 1950's for rock aggregate (Stone and McFarland, 1981). At the present time it is being used as a storage area for sand dredged from the bed of the Arkansas River.

This is one of the finest exposures of the Jackfork Sandstone in Arkansas. More than 200 feet of quartzose sandstone, siltstone, conglomerate, and shale are exposed along the walls of the quarry (Figs. 2 and 3). These rocks are interpreted to be channel-fill and levee deposits in a submarine canyon or in the upper part of a submarine fan complex. At least 14 channels and associated levees and overbank sequences are visible. The dimensions of the submarine fan-channel system are not known precisely, but it is at least 6 miles wide and 10 to 15 miles long. Sole marks and channel geometries suggest paleocurrents to the southwest and west. Load casts, pebbly mudstones, shale intraclasts, and slump features are common sedimentary features. Fragments of plants and invertebrate fossils of Early Pennsylvanian age occur in some of the siltstone and sandstone beds.

CHANNEL-FILL DEPOSITS

The channel-fill deposits consist of both massive sandstone beds which are amalgamated and locally thin and fine upward and some mud-filled channels (shale drapes) which thicken and coarsen upward into sandstone. Associated levee and interchannel deposits occur laterally adjacent to and between the channels in the upper part of the sequence. The levee sequences are inclined away from the channels, commonly slump folded, and pinch out within the outcrops. The inferred turbidite fill consists of lag deposits (stage I, Fig. 4) along the channel base (not exposed in the quarry), sandstone-filled channels (stage II) exposed at the base of the quarry, and channel/levee complexes (stage III) seen in the upper quarry cuts. The entire sequence represents the gradual filling of a canyon or upper fan channel system and its final abandonment.

THE SHELL CORE

Shell Oil Company drilled a corehole on top of Big Rock Mountain just above the middle of the quarry face. One hundred feet were cored of which only the lower 80 feet were recovered. The upper 35 feet of core (20 to 55 foot depth) were partially weathered. Figure 5 is a lithologic summary of the core hole, emphasizing the sedimentary features and channel-fill sequences encountered. Some close-up photos of the core are illustrated in Figure 6 and the stratigraphic positions of the parts photographed are shown in Figure 5.

The core consists of alternating sandstone, shale intraclast conglomerate, and shale beds from the upper 100 feet of the quarry in the upper channel/levee complex (stage III, Fig. 4). Twenty-six sandstone beds occur in the core and make up 63 percent of the cored interval. Individual beds range from 0.2 to 24.7 feet in thickness, averaging 1.9 feet. The sandstone is very fine to medium grained, averag-

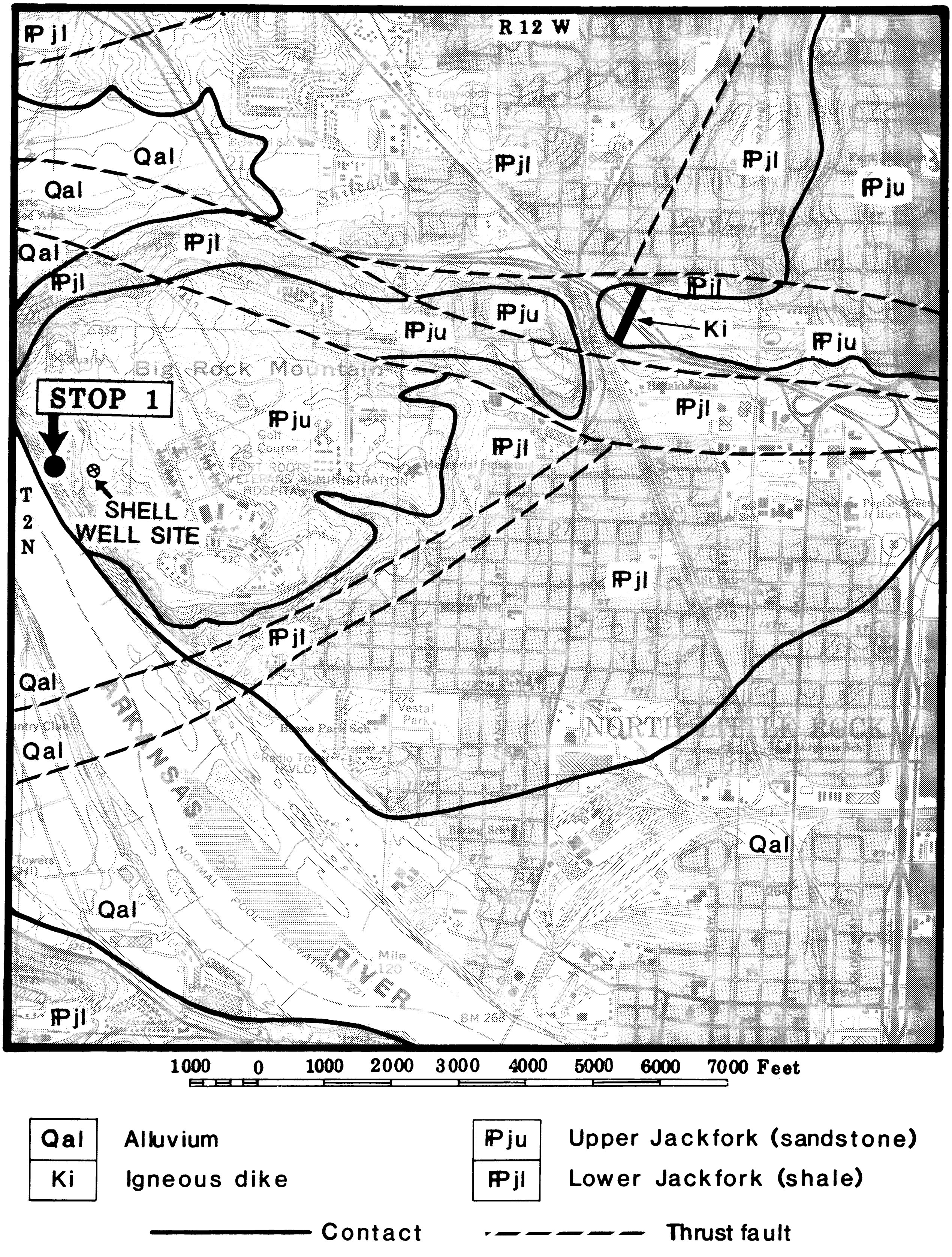
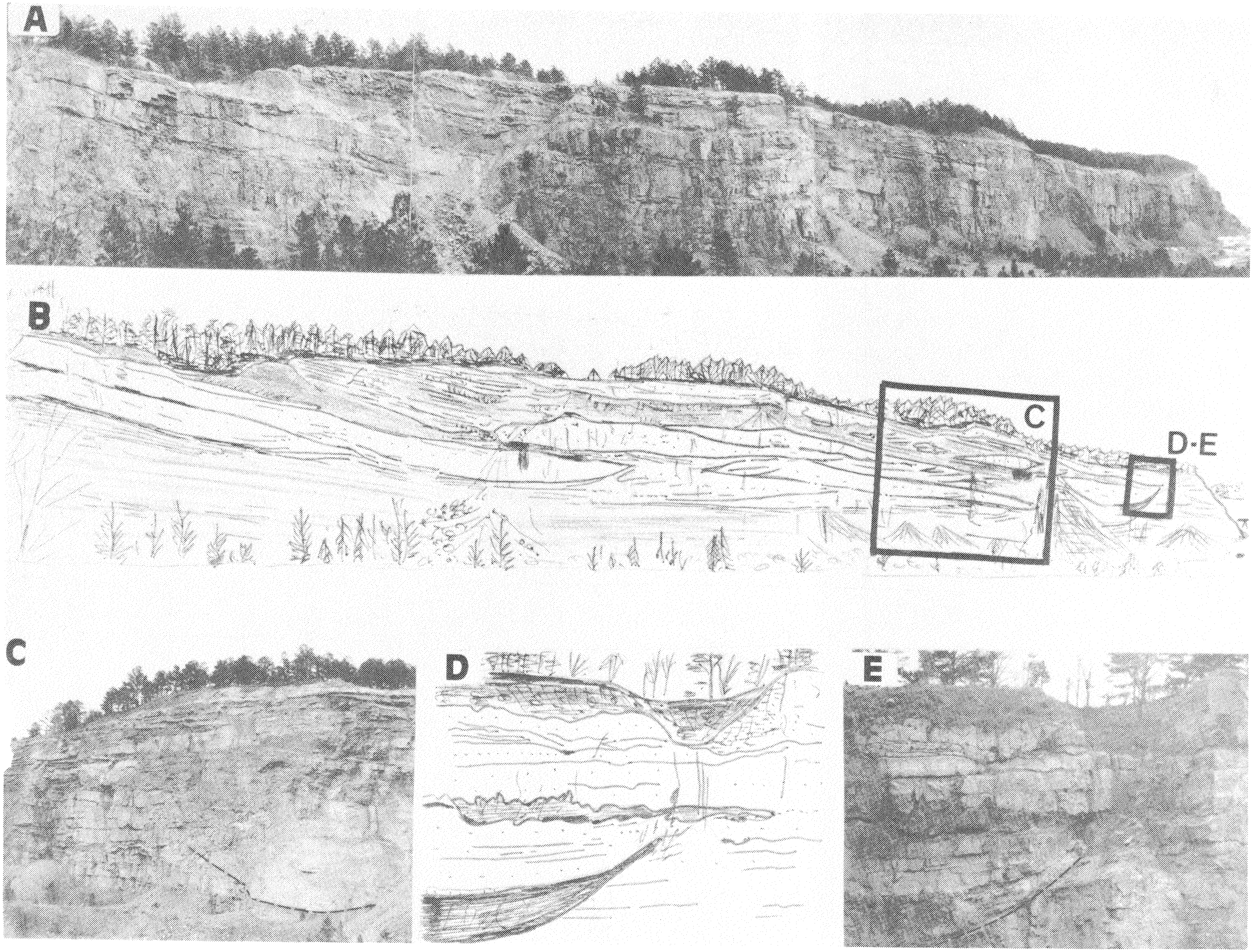


Figure 1. Geologic map of the North Little Rock area showing location of the abandoned Big Rock Quarry (Stop 1).

Figure 2. Stop 1. Lower part of the upper Jackfork Sandstone in the face of the Big Rock Quarry in North Little Rock. A. Full height of the quarry face showing 200-foot sequence of channel-fill, levee, and overbank deposits of a submarine fan. B. Close-up view at north end of quarry showing several submarine fan-channel/levee sequences in more detail.


A
B
C
D-E
C
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E

Figure 3. Stop 1. Photographs and sketches of Morrowan submarine canyon or upper fan channel complex, Big Rock Quarry, North Little Rock. A. North to south view of the east face of the Big Rock Quarry. B. Sketch of Figure 3A, noting the lenticular, laterally aggrading channel-fill sandstone deposits and the more shaly upper part. Boxes denote position of Figures 3C-E. C. Photograph showing some of the major angular discordances between channel and the channel-margin deposits where levees are inclined away from the channels. D. Sketch of Figure 3E showing the side of a channel-fill where strata are onlapping the side of channel. E. Photograph of Figure 3D showing the side of the channel outlined.

ing a fine-grain size. The beds are massive-appearing and locally contain parallel laminae, dish structures, wavy laminae, shale interclasts, cross beds, and slump structures (Figs. 5 and 6). Bouma (1962) sequences are rare. The sandstone beds are cut by vertical to 45-degree-angle calcite-filled veins.

Shale intraclast conglomerate and the highly slumped zones make up 29 percent of the cored interval and average 1.0 feet in thickness. They consist of angular to subrounded clasts of shale, shale and siltstone, and fine-grained sandstone fragments (Fig. 6). Shale interbeds make up the remaining 9 percent of the cored interval and average 0.3 feet in thickness. The shale is finely laminated and contains some plant fragments and abundant mica.

The cored interval is interpreted to be channel-fill deposits in the upper part of a submarine canyon or upper fan channel system. At least two, and probably part of a third channel-fill sequence, are recognized in the core. These channels can be seen in the upper quarry outcrops where each consists of a basal zone of shale intraclasts overlain by massive-appearing sandstone beds. The overall sequence in these deposits is thickening and coarsening upwards.

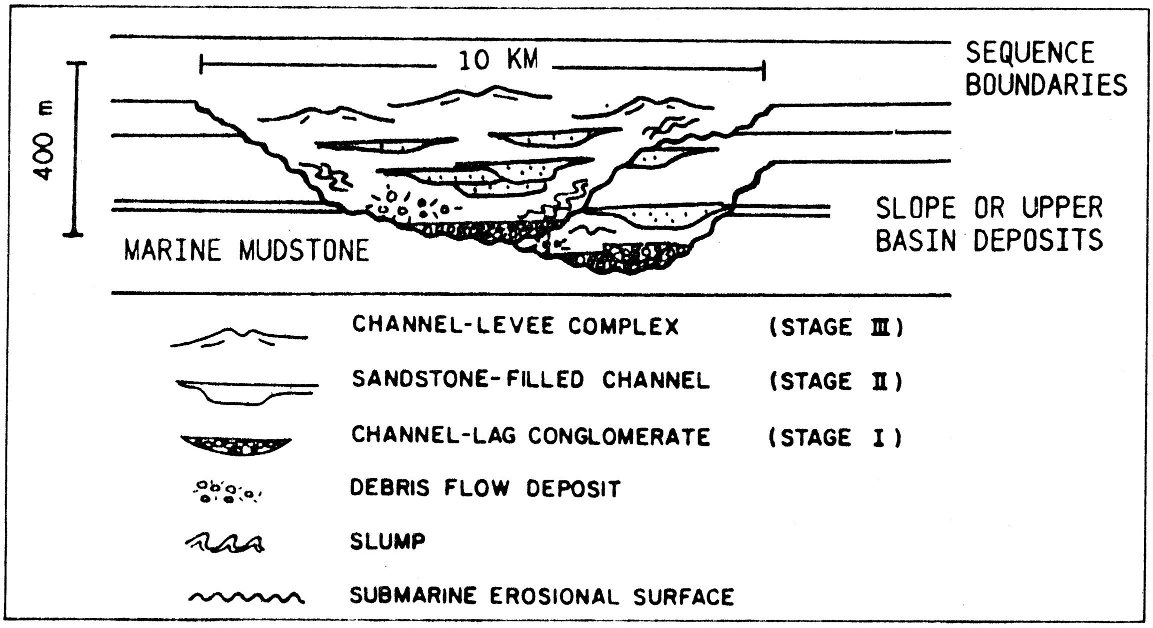
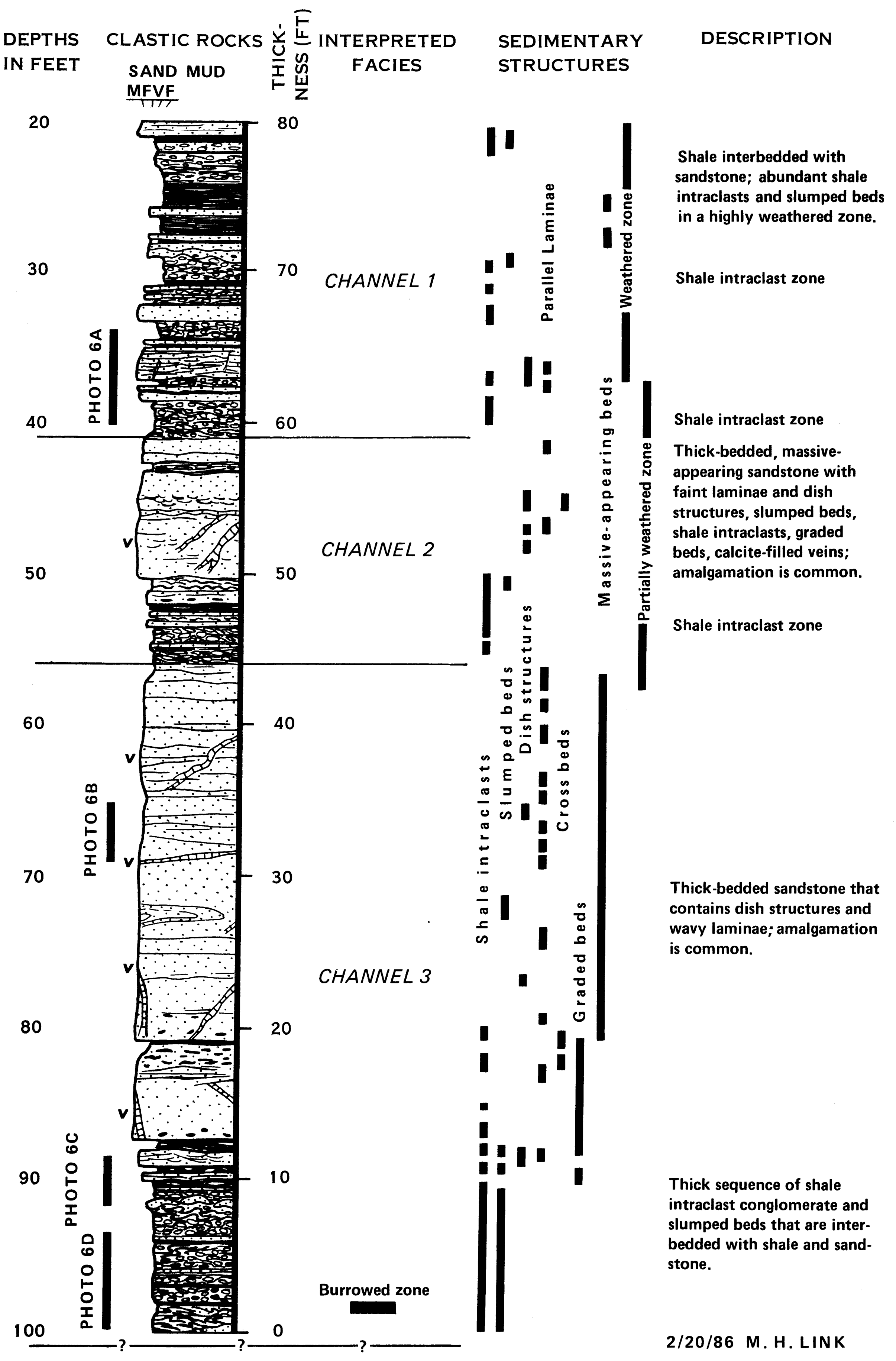


Figure 4. Stop 1. Turbidite fill of large-scale submarine erosional features. A hypothetical example. Note different stages in the sedimentary fill (modified after Mutti, 1985).

Figure 5. Stop 1. Lithologic summary of the Shell Oil Company Big Rock Quarry corehole, with emphasis on the sedimentary features. Location of cored sections illustrated in Figure 6 are identified to left of column.



Shale interbedded with sandstone; abundant shale intraclasts and slumped beds in a highly weathered zone.

Shale intraclast zone

Shale intraclast zone

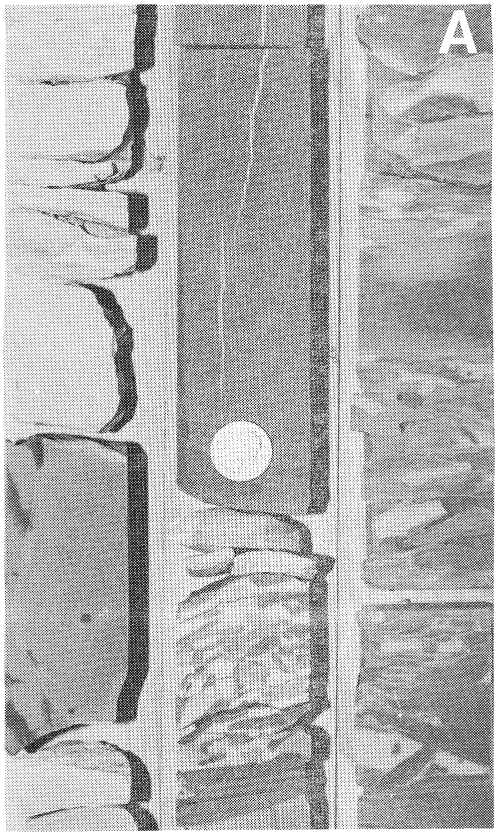
Thick-bedded, massive-appearing sandstone with faint laminae and dish structures, slumped beds, shale intraclasts, graded beds, calcite-filled veins; amalgamation is common.

Shale intraclast zone

Thick-bedded sandstone that contains dish structures and wavy laminae; amalgamation is common.

Thick sequence of shale intraclast conglomerate and slumped beds that are interbedded with shale and sandstone.

2/20/86 M. H. LINK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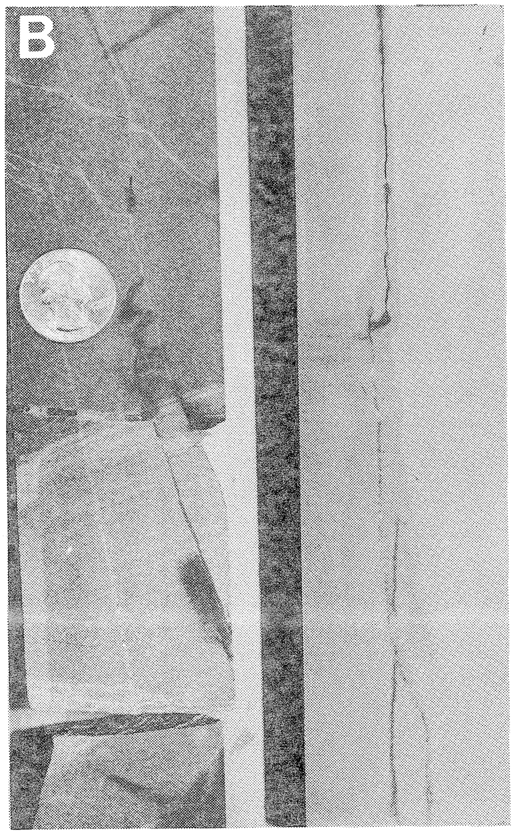
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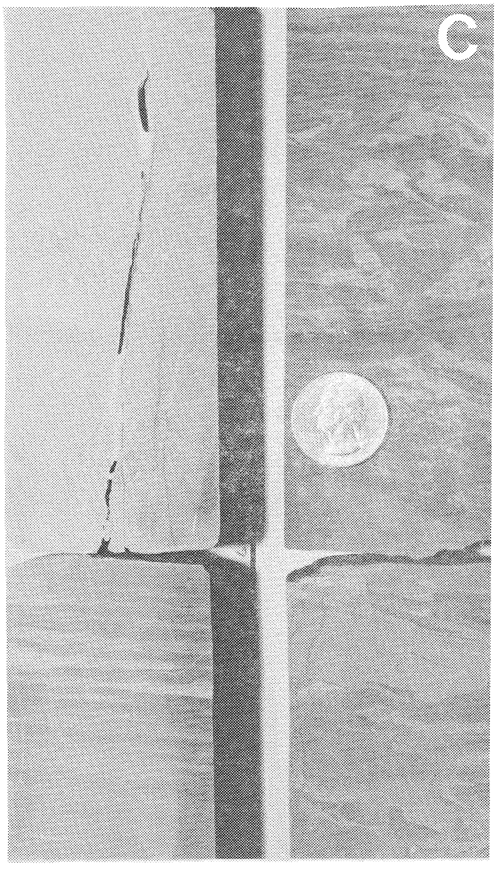
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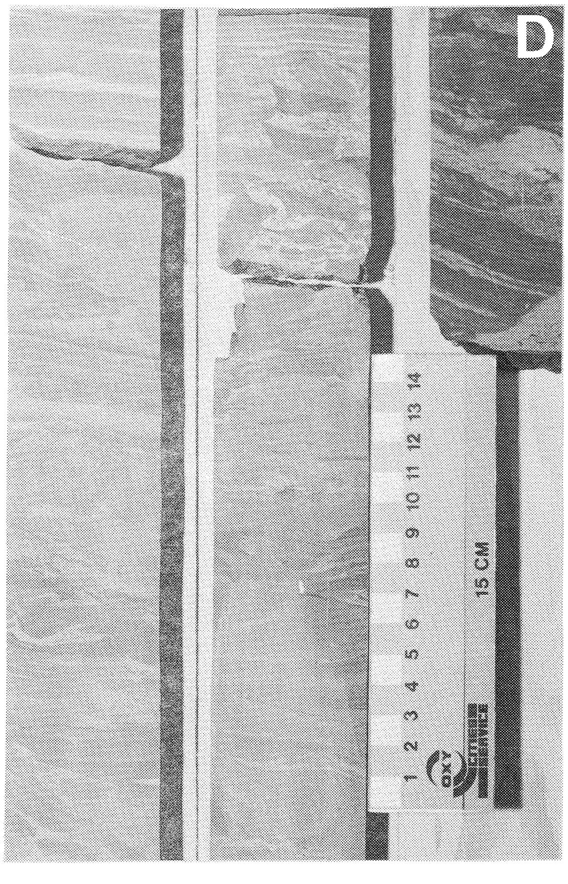
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PETROGRAPHY

Thin section examination by Morris (1985) revealed that the Jackfork sandstones in the quarry are poorly sorted, micaceous, mostly quartzose, non-feldspathic, and contain assorted shale clasts. The sediments probably were derived from a source area to the northeast. He states that these clastics have little similarity to most of those found in the Jackfork in the southern Ouachita Mountains, which are feldspathic and contain schistose rock fragments. Morris postulated a southeastern source area for the Jackfork in the southern Ouachita Mountains.

ACKNOWLEDGEMENTS

We express our sincere appreciation to Rufus J. LeBlanc of Shell Oil Company for his kind assistance in evaluating the strata in the quarry and for making the well cores and other information available to us. We have visited this quarry with many other people and discussed the facies associations to great lengths. People who have influenced our interpretations include R. J. LeBlanc, E. Mutti, M. T. Roberts, R. W. Tillman, Alan Thomson, R. C. Morris, B. R. Haley, and others.

For additional discussion on the Pennsylvanian paleogeography in central Arkansas, including the Big Rock quarry area, see the paper by M. H. Link and T. R. Roberts later in this volume.

Figure 6. Stop 1. Core photographs of Shell's Big Rock Quarry core hole. See Figure 5 for location. A. Core depth 33.7-40.6 feet, interbedded sandstone and shale intraclast zones. The sandstones are laminated and contain dish structures. Note the nearly vertical calcite-filled veins. B. Core depth 66-69.2 feet, massive-appearing and laminated sandstones. C. Core depth 88.6-92 feet, interbedded sandstone with ripple-laminated shale and shale intraclast zones. D. Core depth 93.8-100 feet, the basal part of the core showing alternating shale intraclasts and laminated to rippled shale and siltstone beds.

←

STOP 2 -- JACKFORK SANDSTONE AT I-430 ROADCUT IN NORTHWEST LITTLE ROCK

Large exposures of the middle and upper parts of the Jackfork Sandstone are present on both sides of Interstate 430 in northwestern Little Rock immediately north of Arkansas Highway 10 (Fig. 7). The locality is on the south flank of the Big Rock syncline in the eastern Aly belt (Haley and Stone, 1981; Fig. 8, this report). The rocks are quite intensely sheared and several thrust faults are visible.

The middle part of the Jackfork (Fig. 9A) consists mostly of black shale and the upper part (Fig. 9B) consists of interlayered thin- to massive-bedded, gray, quartzitic sandstone and dark-gray shale that weathers brown or maroon. Minor intraformational sandstone-shale conglomerates are also present. The sandstone sequences, which become finer grained and thinner bedded upward, represent submarine fan-channel deposits. The beds here were probably deposited farther down the slope of the submarine fan than the equivalent strata at Big Rock Quarry a few miles to the east. Some contorted features characteristic of soft-sediment slumping, but later deformed by tectonic processes, are visible.

Nearly all of the rocks dip to the north, as do the cleavage and most fault planes. Minor folds are upright or overturned to the southwest and have nearly horizontal hinge lines. Deformation apparently involved a succession of events: (1) southward slumping down the continental slope; (2) northward stacking of several thrust-fault slices; (3) folding during several episodes; and (4) backfolding and

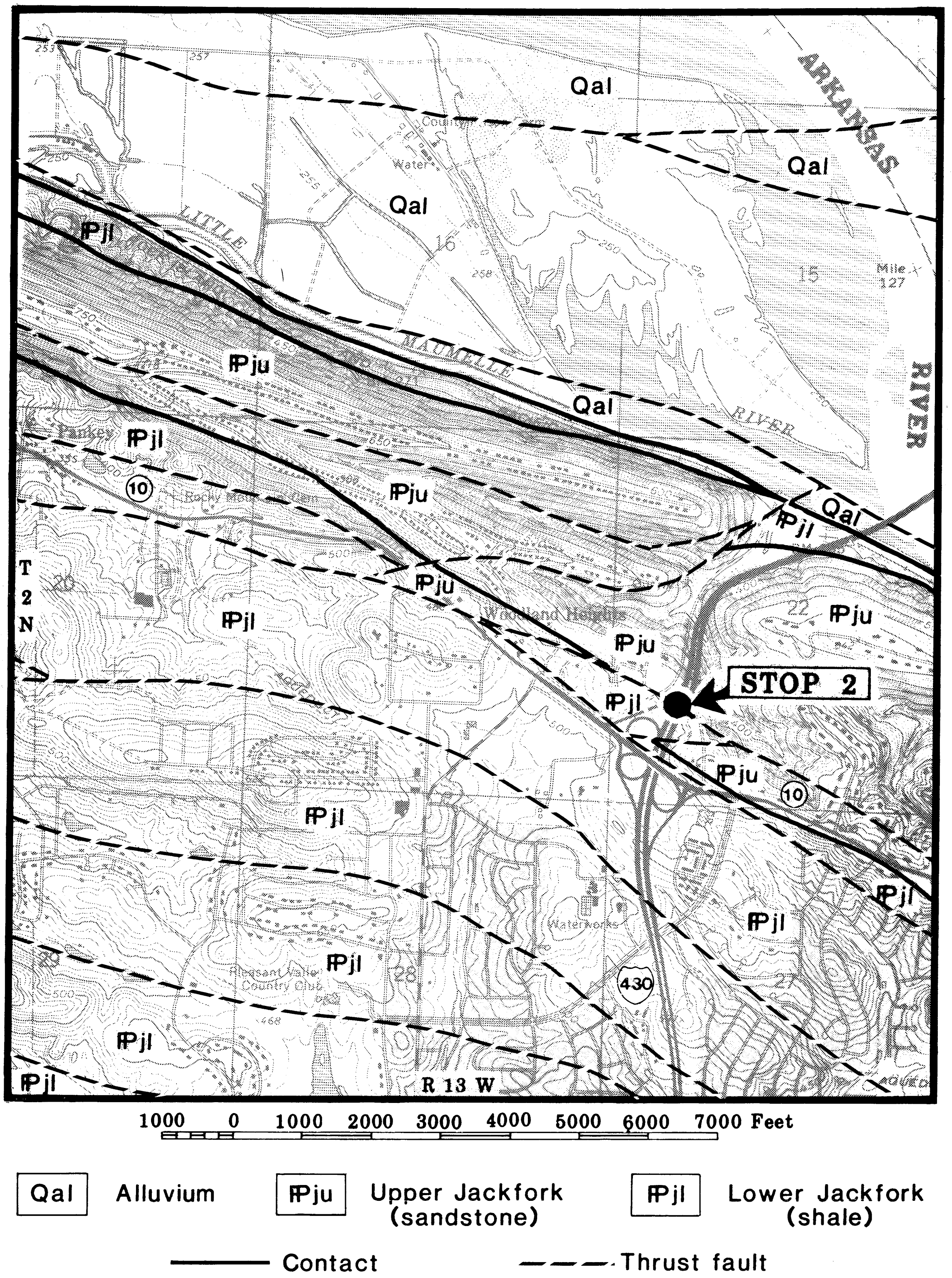
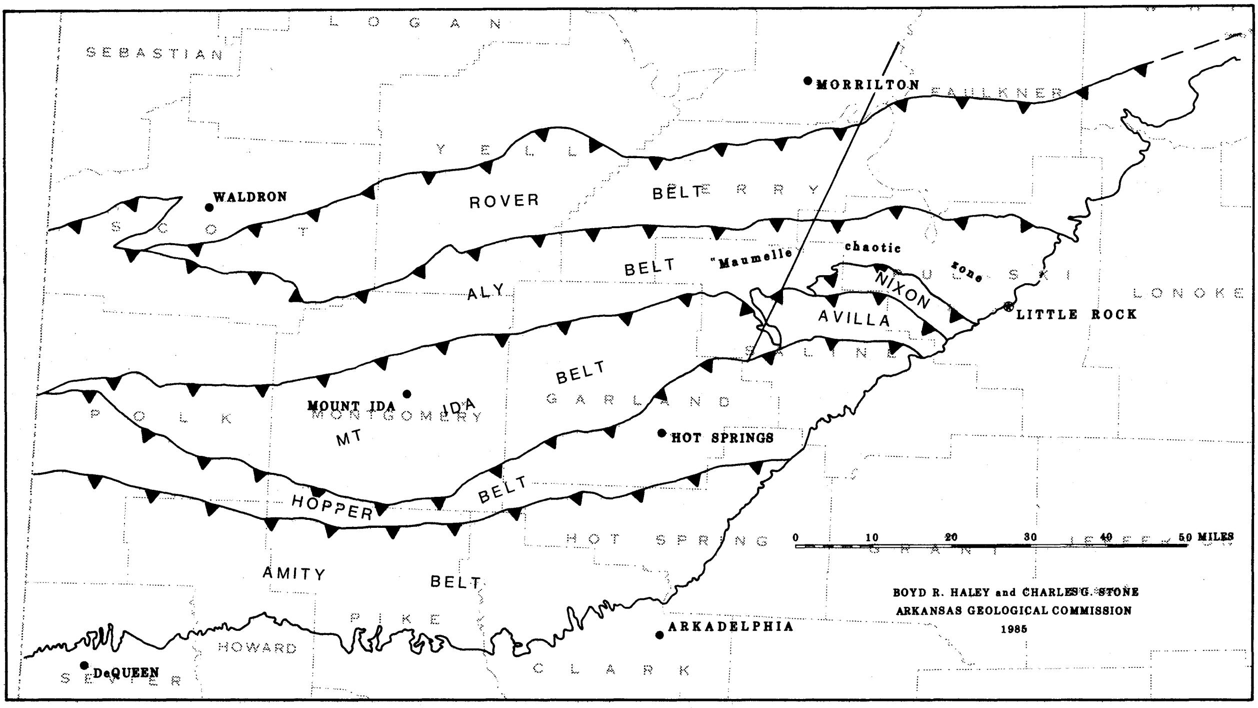


Figure 7. Geologic map of northwestern Little Rock in vicinity of Stop 2 along Interstate 430.



BOYD R. HALEY and CHARLES G. STONE
ARKANSAS GEOLOGICAL COMMISSION
1985

Figure 8. Sketch map showing the structural belts recognized by Haley and Stone (1981, 1982) in the Ouachita Mountains of Arkansas. Only those thrust faults separating structural belts are shown. Map also shows the approximate extent of the Maumelle zone or Maumelle chaotic zone of Viele (1973, 1979).

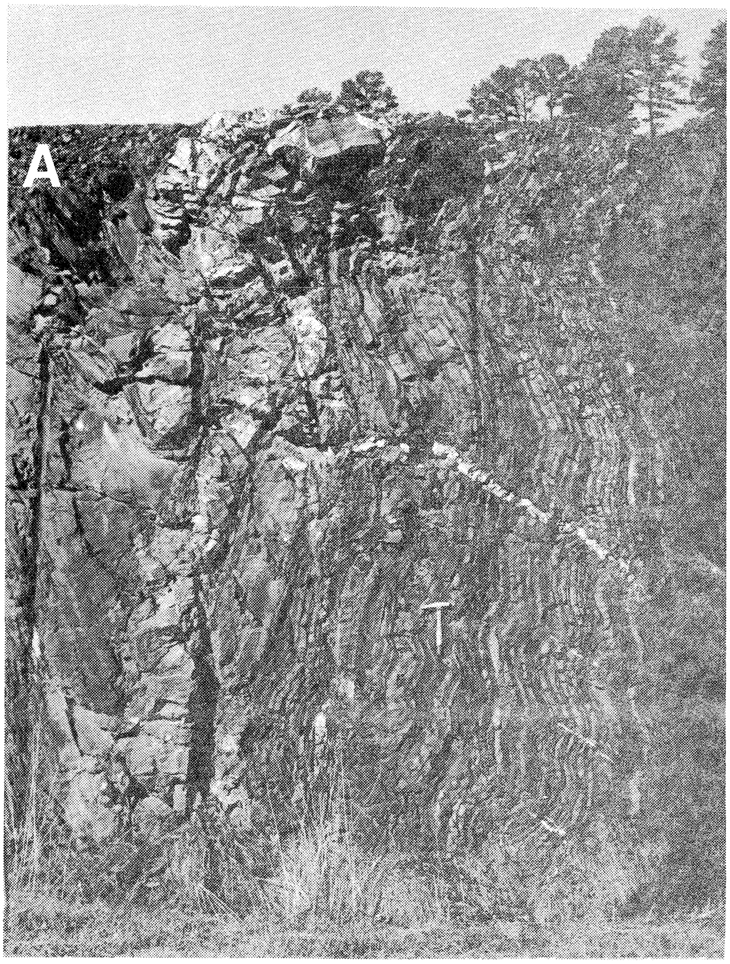
faulting caused, in part at least, by the piling up and crowding at the toe of the larger thrust sheets. The last 3 events took place in late Paleozoic time as the Ouachita Mountains were being formed.

Hypogene quartz veins of late Paleozoic age are common in the strata and some of them contain rectorite, cookeite, pyrite and other minerals. Dickite is abundant on slickenside surfaces in the fault zones.

In much of the area south of here, the Jackfork and older strata are more complexly folded and faulted.

STOP 3 -- DEFORMED MIDDLE AND UPPER ORDOVICIAN ROCKS WEST OF LITTLE ROCK

Ordovician rocks of the Womble Shale, Bigfork Chert, and Polk Creek Shale of the Nixon belt (Haley and Stone, 1982) are exposed on both sides of Interstate 430 immediately north of Colonel Glenn Road (Fig. 10). The rocks, which appear to dip uniformly to the north, actually consist largely of stacked isoclinal folds (Figs. 11A, 11B) with southward vergence. This arrangement is most discernible in the Bigfork. Cleavage is well developed and typically is parallel to bedding. The exposures are cut by northward dipping thrust faults which are essentially parallel to bedding. The faults are marked by thin quartz veins and slickensides. At the southwest end of the exposed area, a large block of the Lower Division of the Arkansas Novaculite has been thrust over the Womble Shale. On the east side of the roadcut two dikes of alkalic igneous rocks, weathered entirely to clay, are present in the Bigfork. Similar intrusive rocks in central Arkansas have been dated as early Late Cretaceous in age (Howard, 1986).



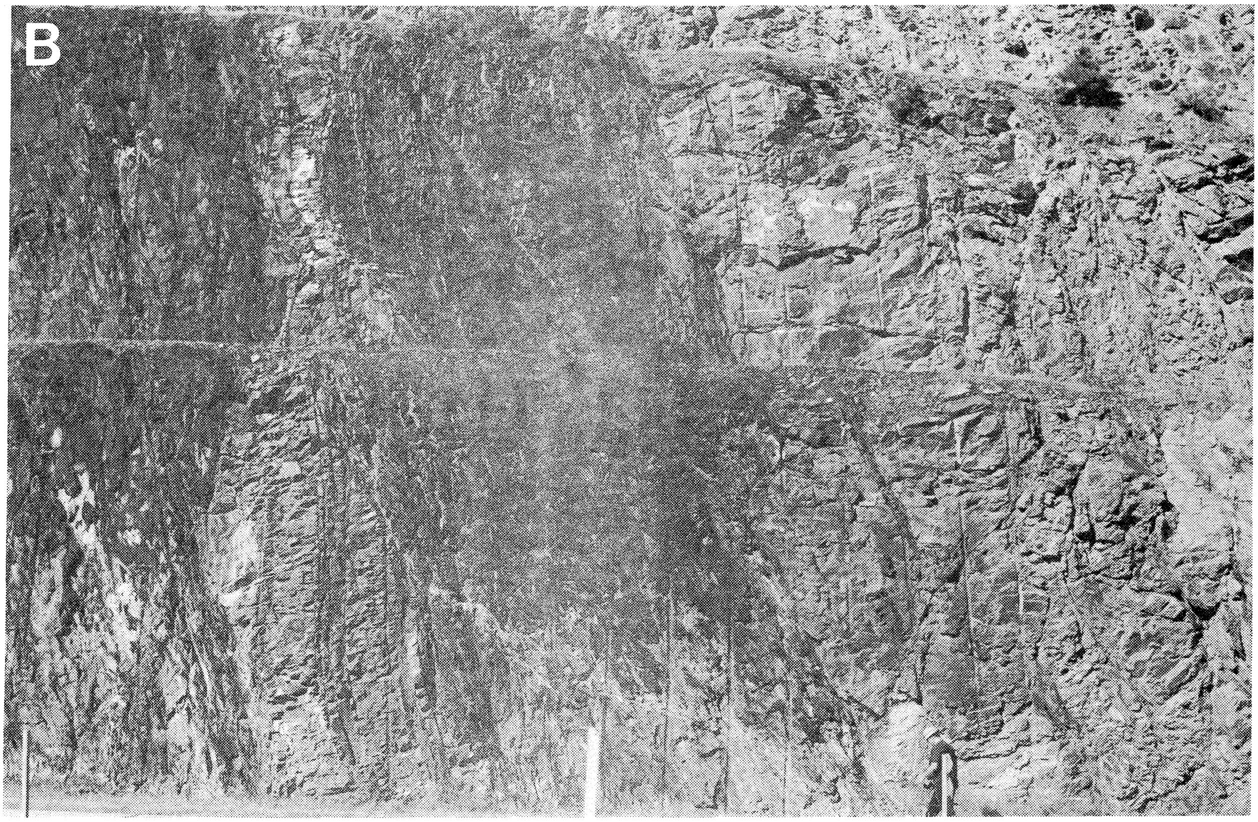
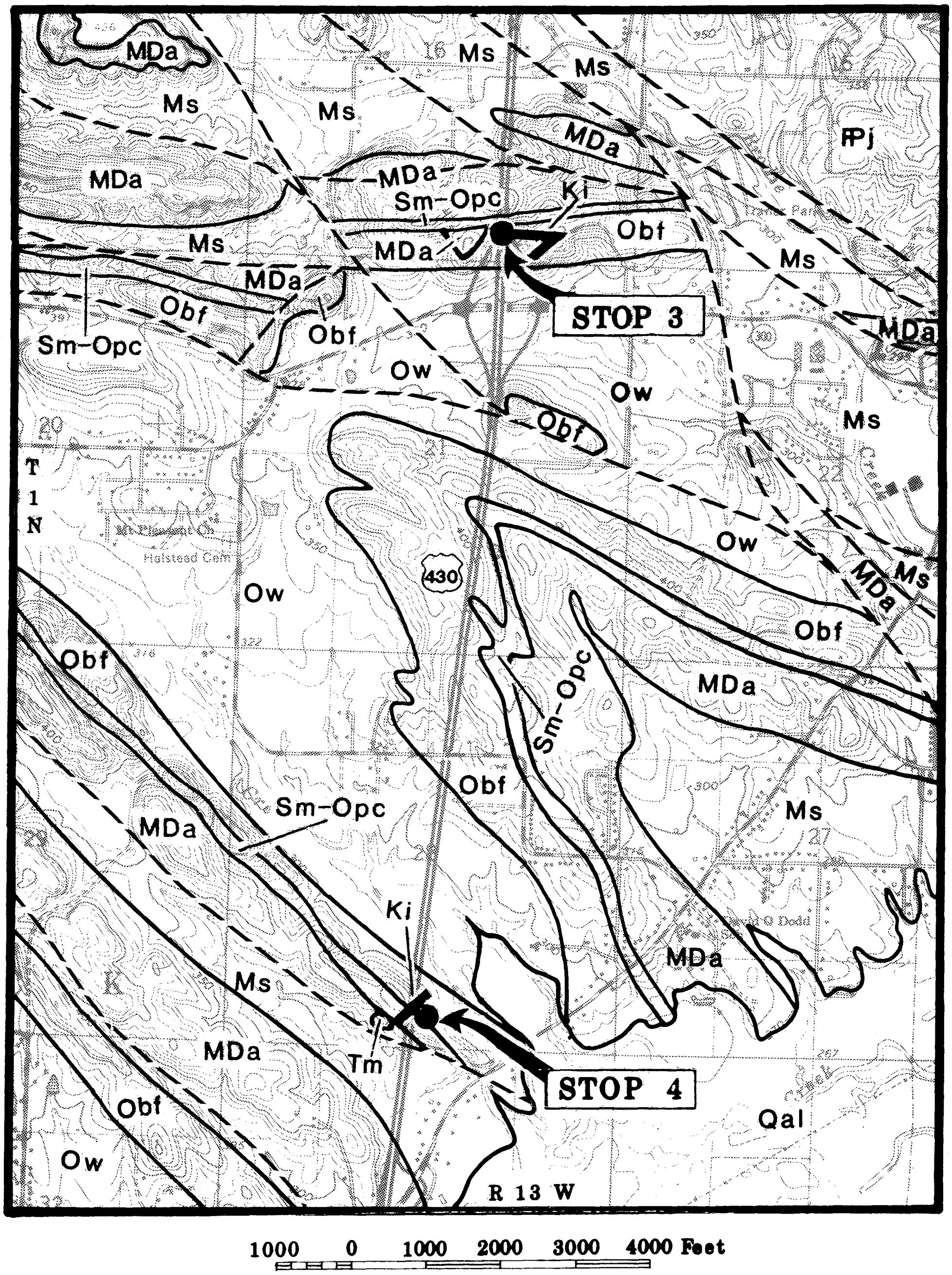


Figure 9. Stop 2. Jackfork Sandstone along Interstate Highway 430. A. Middle part of Jackfork on west side of highway. Interbedded sandstone and shale dissected by northward dipping cleavage and veins of milky quartz. Youngest beds to the right (north). B. West wall of roadcut along Arkansas Highway 10 access road. Highly fractured quartzitic sandstone of the upper Jackfork (right) overlying intensely sheared middle Jackfork black shale (left).



Qal	Alluvium	MDa	Arkansas Novaculite
Tm	Midway Group	Sm-Opc	Missouri Mountain Shale-Polk Creek Shale
Ki	Igneous dikes	Obf	Bigfork Chert
PPj	Jackfork Sandstone	Ow	Womble Shale
Ms	Stanley Shale		

——— Contact – – –Thrust fault

Figure 10. Geologic map in vicinity of western Little Rock showing location of Stops 3 and 4.

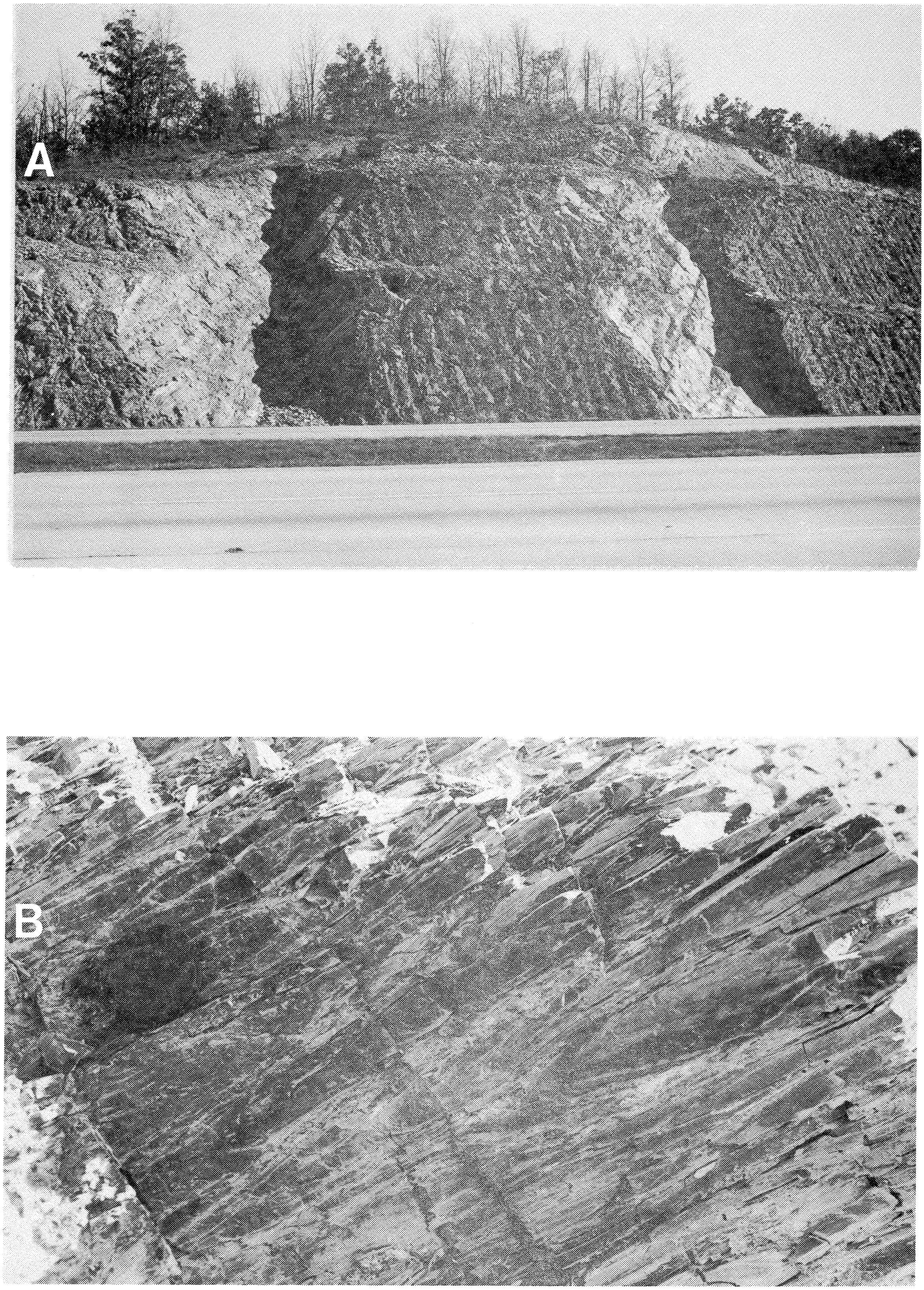


Figure 11. Stop 3. Chert and siliceous shale of the Bigfork Chert exposed on east side of Interstate 430 near Colonel Glenn Road. North is to left. A. Distant view. Two igneous dikes, highly altered, occupy the recessed areas in shadow. B. Closeup of part of above view showing overturned isoclinal folds. Note pervasive shearing and crowding in fold hinges.

In nearby localities, graptolites of Middle Ordovician age have been collected from shale in the upper part of the Womble and graptolites of Late Ordovician age from shale in the Polk Creek.

Two opinions have been suggested to explain the complex structure of the rocks in this belt. Viele (1973) proposed gravitational sliding and cascading of rock units northward from a series of nappes. Stone and McFarland (1981) suggested a series of northward-moving thrust plates with an overall southward increase in the structural complexity of the rocks in each thrust plate. They attributed the southward-overturning of the rocks and fault planes south of Stop 2 to backfolding of the toe of each thrust plate and also to subsequent thrust faulting. Subsequent or yet younger faulting is suggested by the outlier of Arkansas Novaculite overlying Womble at this stop.

STOP 4 -- BIGFORK CHERT TO ARKANSAS NOVACULITE SOUTHWEST OF LITTLE ROCK

A long sequence of intensely sheared and tightly folded Ordovician to Devonian-Mississippian rocks of the Nixon belt is exposed on both sides of Interstate 430 about 0.4 miles north of Arkansas Highway 5 (Figure 10). From north to south the sequence consists of: 1) chert and siliceous shale of the Ordovician Bigfork Chert, 2) black shale and dark gray chert of the Polk Creek Shale (also Ordovician), 3) gray and tan shale with thin chert and novaculite beds of the Silurian Missouri Mountain Shale, 4) massive-bedded very light gray novaculite (tripolitic at the base) of the Lower Division of the Devonian Arkansas Novaculite, and 5) black siliceous shale and chert of the Middle Division of the Devonian-Mississippian Arkansas Novaculite. The sediments from which these rocks formed indicate very slow rates of deposition.

The rocks (Fig. 12, 13) dip to the north and are overturned to the south. The style of folding varies with lithology. Pervasive north-dipping cleavage is evident in the Bigfork. Some northward-dipping thrust faults with quartz veins and minor gouge are also present. Their direction of displacement is as yet unclear. The intense shearing has obliterated many of the lithic characteristics and apparently most microfossils. However some conodonts, sponge spicules, and radiolaria are present in a few beds in the Middle Division of the Arkansas Novaculite.

An alkalic igneous dike which has been altered to clay dissects the Middle Division of the Arkansas Novaculite on the west side of the road. It is assumed to be of the same age, i.e., early Late Cretaceous, as similar intrusive bodies in central Arkansas. Both the novaculite and the dike are overlain unconformably by a thin cover of clay, sand, and some gravel that may represent a remnant of the Midway Group of Paleocene age.

Keller et al. (1985) have shown that the size of the polygonal triple-point grains developed during recrystallization of chert or novaculite at this stop is probably related to low-rank regional (thermal) metamorphism, perhaps enhanced by the implacement of a pluton nearby. The triple-point grains average about 40 microns in diameter and are among the coarsest yet measured from the Ouachita Mountains. Most of the triple-point texture was formed during late Paleozoic deformation.

According to Viele (1973), two major fold trends have been developed in the lower and middle Paleozoic rocks in this general area. The Alexander trend consists of southward-overturned folds with nearly horizontal hinge lines. The Ellis Mountain trend is overturned to the southwest but the fold hinge lines are reclined. Most of the folds at this site belong to the Alexander trend but the cleavage is believed to be associated largely with the Ellis Mountain trend.

Figure 12. Stop 4. Bigfork Chert and Polk Creek Shale exposed on west side of Interstate 430. Shale and chert of the Bigfork (to the right, i.e. north) in fault contact with black shale and some chert of the Polk Creek (to the left, i.e. south).

Figure 13. Stop 4. Massive-bedded novaculite of the Lower Division of the Arkansas Novaculite on the right (north) and thin beds of chert and siliceous shale of the Middle Division of the Arkansas Novaculite on the left (south). The rocks are capped by a thin sandy rubble that may be a remnant of the Paleocene Midway Group. The exposure is on west side of Interstate Hwy. 430.

STOP 5 -- THE WARNER SERPENTINE-TALC (SOAPSTONE) PITS

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Fayetteville, Arkansas 72701

The history of "soapstone" mining in Saline County, Arkansas goes back to 1881, as reported by local residents in Comstock (1888). Mining has been fairly steady since 1953 with numerous pits, trenches, and shafts having been dug in the area. The Warner Pits, a cluster of 3 closely spaced workings (Fig. 14), are about 3 miles north of Avilla. They are currently being worked by Johnny Warner for the Milwhite Co. in Bryant, Arkansas. Older major workings are present in the two adjacent sections to the west. They are the Inman Pit, now full of water, and the Anderson Pit, now abandoned except for the sporadic production of minor amounts of Womble Shale.

The serpentine-talc pits are located in the far eastern end of the Benton-Broken Bow uplift in the Avilla belt, one of the two metamorphic maxima of the Ouachitas. The other significantly metamorphosed center is the Broken Bow uplift in Oklahoma, where, in McCurtain County, a single body of metagabbro is present. The igneous bodies in Saline County are enclosed by the Ordovician Womble Shale and Bigfork Chert and trend east-west, parallel to the regional strike. These are the only Paleozoic igneous rocks now known in the Ouachitas of Arkansas except for the Hominy Hill metagabbro, which is located along strike 18 km (11 mi) to the southeast, and is believed to be related to the serpentine-talc bodies of Saline County (Morris and Stone, 1986). Most of the other exposed igneous rocks in Arkansas are considered to be Cretaceous in age.

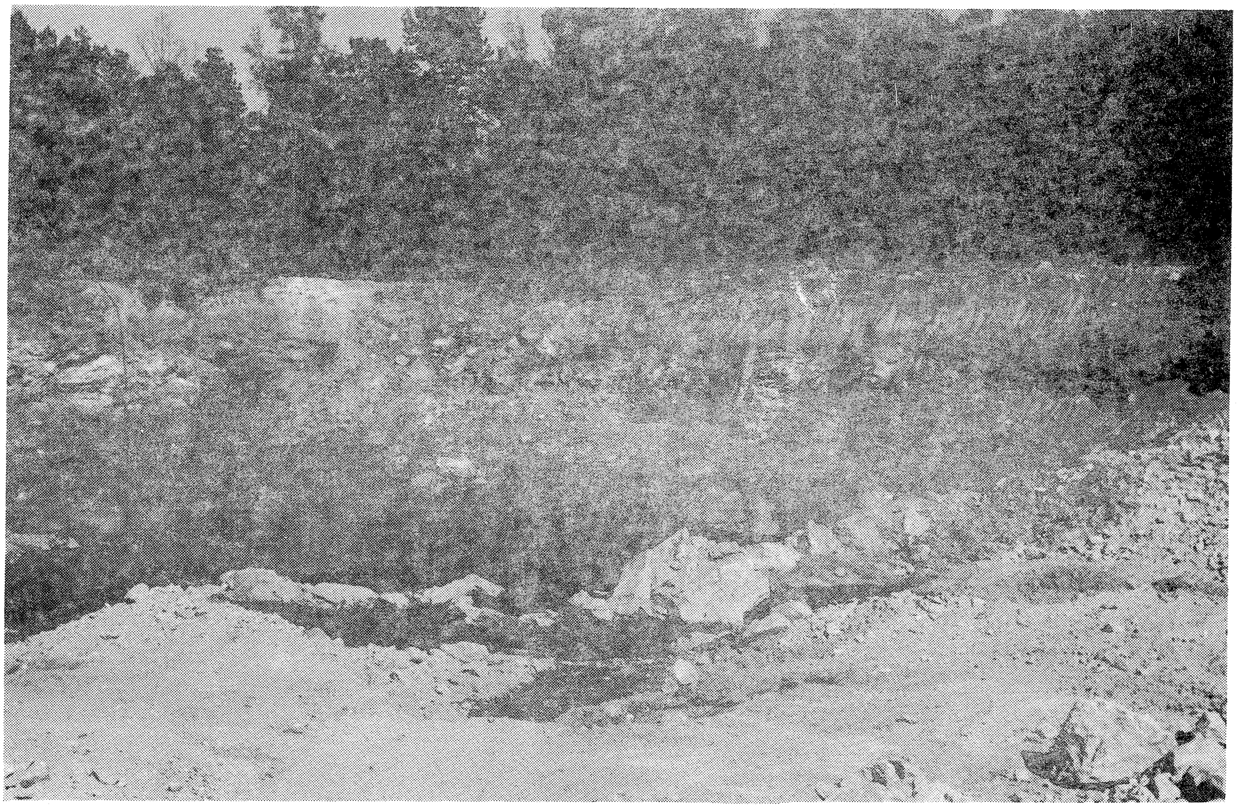


Figure 14. Stop 5. View of the New Warner Pit operated by the Milwhite Company in northern Saline County.

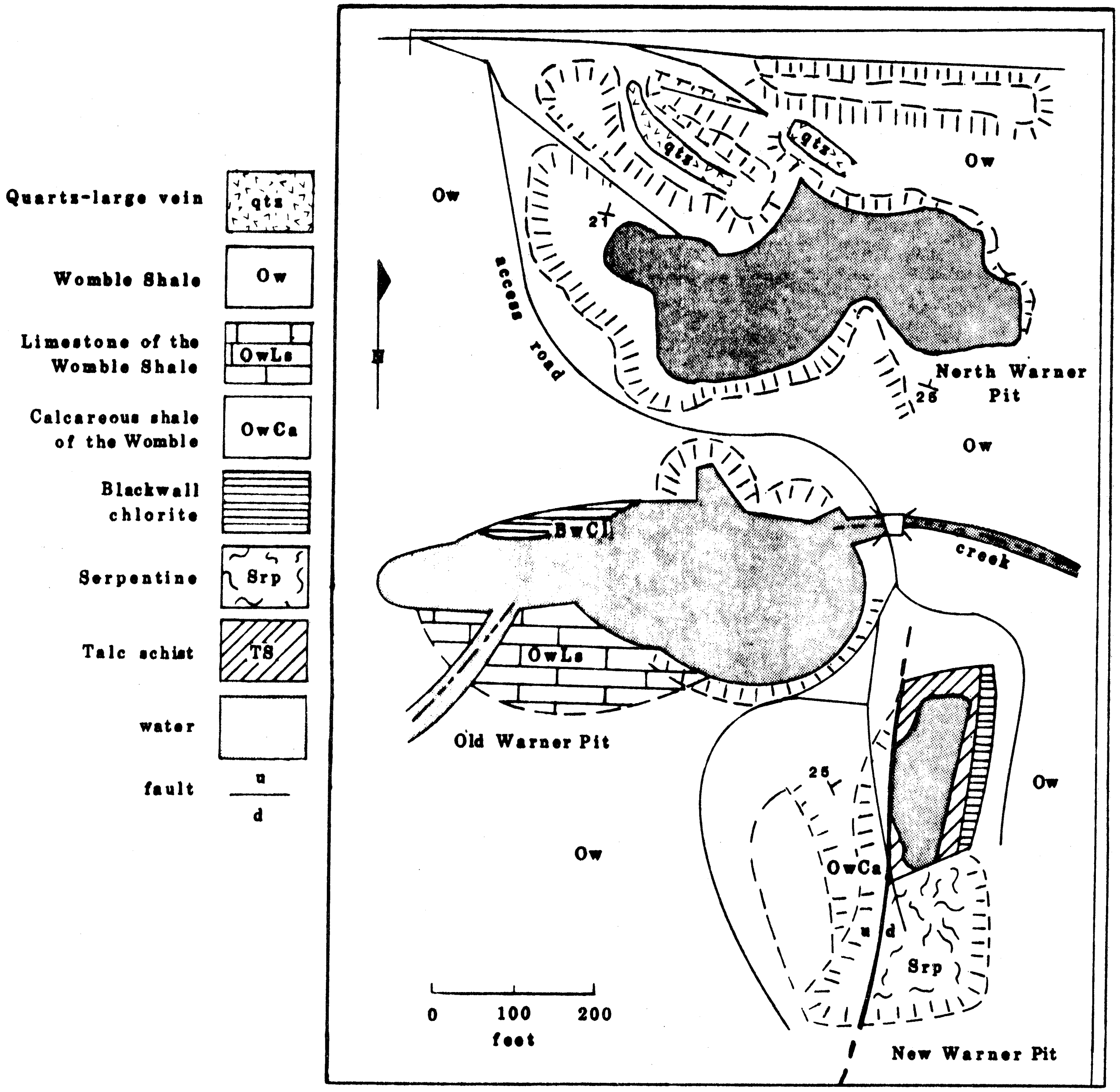


Figure 14. Stop 5. Geologic map of the Warner soapstone pits, Saline County, Arkansas.

These highly sheared ultramafic rocks in the Warner pits are derived from peridotite with the major serpentine being antigorite, the high-temperature end-member of the serpentine group. The stability range of antigorite is 460°-500°C, which is considerably higher than the temperature of regional metamorphism, --low greenschist facies (230°-275°C). The temperature difference proves that the serpentinization of the peridotite took place prior to emplacement.

The bodies show a zonation from the serpentine-talc core outward to a talc schist containing talc, carbonates, pyrophyllite veins, and minor sulfides (Fig. 15). Outside of the talc schist on the north side of the Old Warner Pit and the east side of the New Warner Pit is a well developed blackwall chlorite zone. The blackwall chlorite has numerous light-green talcose veins which thin gradually outward from the core. Minerals of the blackwall chlorite include chlorite, talc, Al-montmorillonite, penninite, kaolinite, and minor rutile. Minor, uneconomical amounts of nickel have also been reported, as has amethyst (Stone and Sterling, 1964).

The talc, talc schist, and the blackwall chlorite are the result of later metasomatic activities, while the clay minerals and the sulfides are believed to be hydrothermal in origin.

All foliations in the enclosing Womble Shale follow the regional trend, striking northwest and dipping northeast, as do the serpentine, talc schist, and the blackwall chlorite. The Bigfork in the Anderson area, on the other hand, strikes northeast and dips southeast. This discordance has been shown to be related to local faulting and folding as some of the foliations are parallel to bedding on the limbs of folds overturned to the southeast (Cox, 1986).

The temperature differences, concordant structure, lack of contact metamorphism, and the geochemistry of the serpentine-talc occurrences and the metagabbro at the related Hominy Hill locality are compatible with either mid-ocean ridge or transform fault origins (Morris, personal communication, 1986). This relationship suggests that the serpentine-talc rocks are a fossil peridotite which rose as a diapir into wet sediments before lithification and deformation. The diapir could have been a textbook example of subduction zone igneous activity, or the result of transform faulting, or of emplacement along a rift zone as proposed by Thomas (1986). Thomas (ibid.) places the transform fault just south of the present location of the Ouachitas. After the diapir rose into the pre-orogenic Ouachita trough, it was brought to its present location and folded and faulted into its present shape by the Ouachita orogeny.

Points of interest in the Warner pit area are the massive quartz veins located on the north side of the North Warner Pit and the massive homogenous limestone along the south "shore" of the Old Warner Pit. Note the two episodes of calcite veining as the younger white calcite veins cross-cut the older, folded black calcite veins. In the New Warner pit along the south wall, is a sequence of calcareous black shale and micritic black limestone which is highly folded and contains laminations of graphite-like dust and disseminated and veined pyrite.

The serpentine and the talc schist of the New Warner Pit itself contain highly convoluted and anastomosing veins of talc and pyrophyllite. Some of these veins are 20 cm (8 in) wide and are traceable for meters. The talc and pyrophyllite veins have a very distinct cockscomb texture.

STOP 6 -- OLISTOLITHIC INTERVAL IN THE MIDDLE JACKFORK SANDSTONE AT LAKE MAUMELLE

The rocks exposed in this roadcut on Arkansas Highway 10 (Fig. 16) consist of black and reddish-brown shale with angular to rounded olistoliths of light-gray quartzose sandstone, very similar lithologically to the typical sandstone of the Jackfork. When comparable olistoliths are partially or

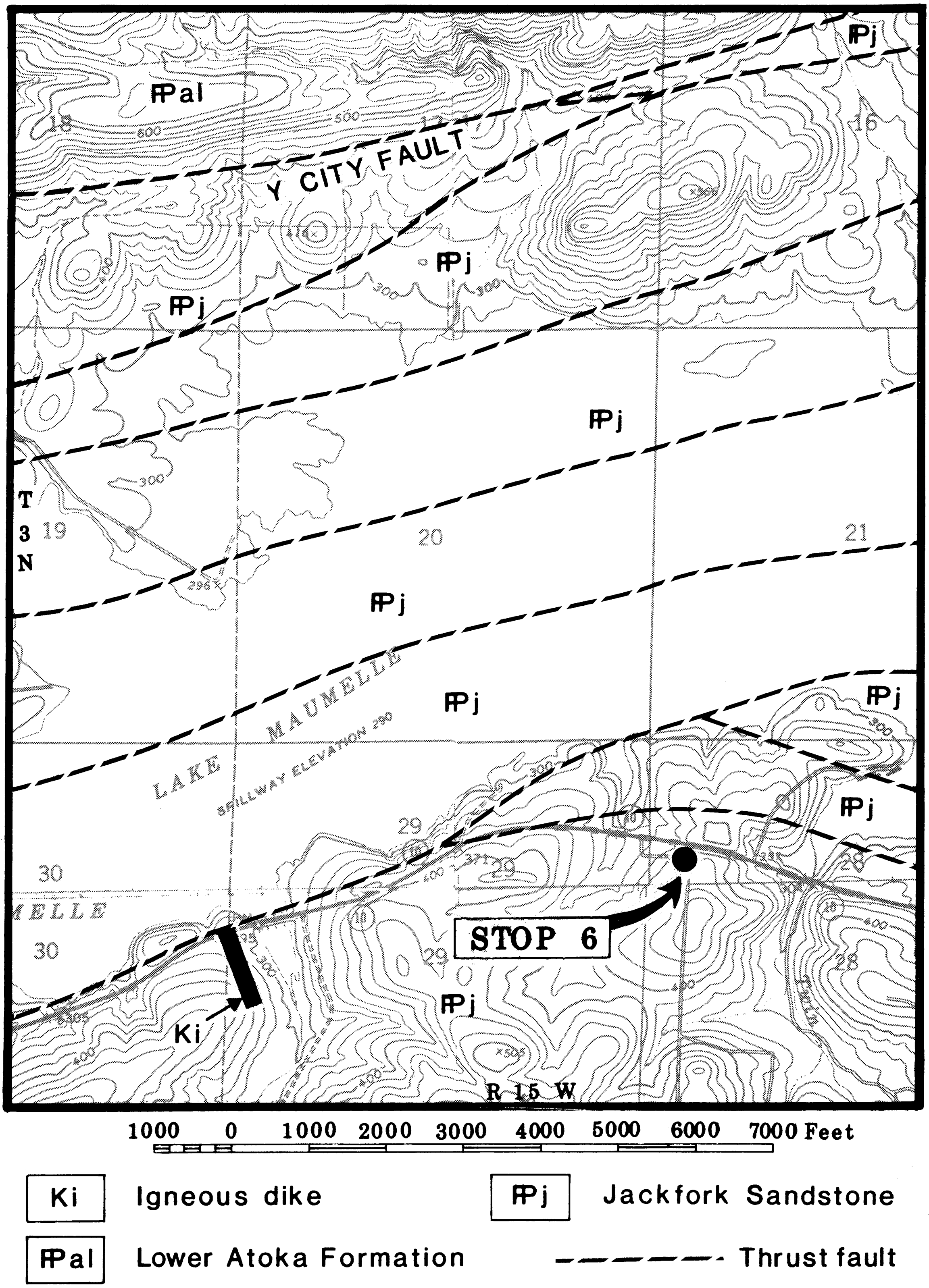


Figure 16. Geologic map of part of Lake Maumelle area showing location of Stop 6, which is approximately 2.5 miles east of bridge where Arkansas Highway 10 crosses the lake.

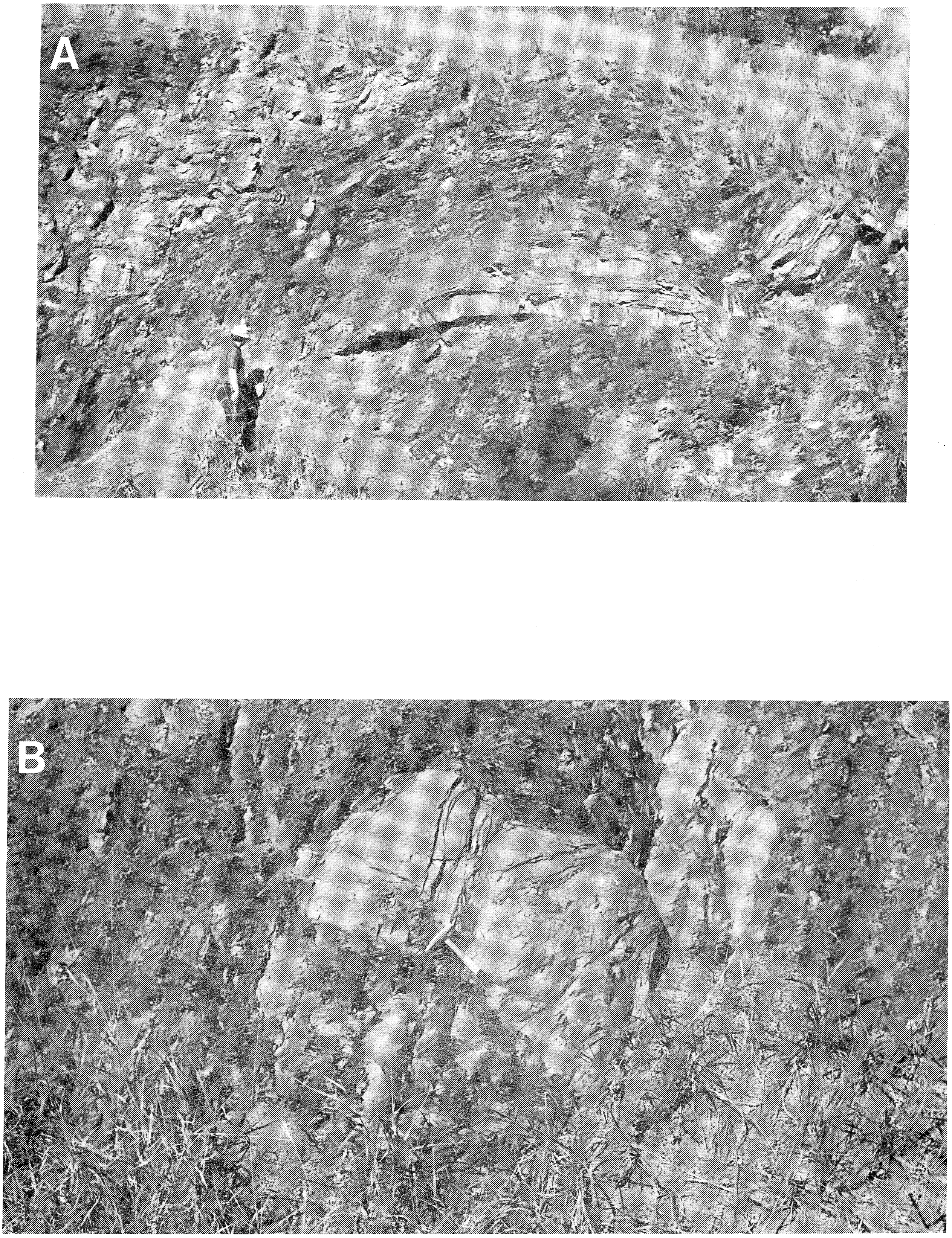


Figure 17. Stop 6. Olistoliths in the middle part of the Jackfork Sandstone on Arkansas Highway 10 near Lake Maumelle. A. Quartzitic sandstone olistoliths and iron carbonate concretions in black shale. B. Close-up showing healed fractures in a sandstone olistolith.

completely weathered from the shale and are exposed on the surface, they resemble the "knockers" described by Alpine geologists. Fossil-bearing sandstone olistoliths have not been found at this outcrop, but are present in nearby outcrops. This locality is in the Aly belt of Haley and Stone, 1981 (Fig. 8). Thrust faults are present and are marked by slickensides and quartz veins. Some of the quartz veins contain minor amounts of lead, zinc, copper, and silver minerals.

Olistolithic zones of the type observed in this outcrop, with their characteristic rubbly bedding, are common in the middle part of the Jackfork Sandstone in Pulaski County and are present, but less abundantly, for a distance of about 130 miles to the west. Similar zones are also present in the lower part of the Atoka Formation, are very common in the Johns Valley Shale, and common in the Stanley Shale. Tectonic deformational indicators such as the faults, slickensides, incrustations of dickite, quartz veins, and cleavage present here are not unique to this outcrop, but are present in many of the other olistolithic zones.

Invertebrate fossils have been collected from sandstone olistoliths at two localities in the middle Jackfork east of this stop. Fossil-bearing sandstone olistoliths appear to have been limy and are thought to have slid southward from an area of shallow-water deposition. The following ammonoid cephalopods of early Morrowan (Hale) age have been identified: *Syngastrioceras globusum* (Easton), *Bisatoceras (Schartymites) paynei* (Gordon), *Reticuloceras tiro* (Gordon), *Retites semiretia* (McCaleb), *Cymoceras adonis* (Gordon), and *Stenopronorites quinni* (Gordon) (Gordon, 1968 and Gordon and Stone, 1977).

Viele (1973, 1979) named the belt in the vicinity of this stop the "Maumelle chaotic zone". He suggested that it formed as part of a subduction complex upon an overriding plate to the south when the plate was widely separated from the shelf-slope zone of North America.

Voicing another opinion, Morris (1985, p. 29) noted that from all appearances the rubble bedding in this area is indistinguishable from that in the Jackfork in other parts of the frontal Ouachita Mountains. He further offered the following in rebuttal:

> "1) The rocks of the "Maumelle Zone" are an integral part of the Jackfork Group. 2) The Jackfork had a clear connection with the shelf/slope area of North America, having received from there the limestone, chert, and non-feldspathic sandstone clasts in rubbly flows and the rounded, mature mineral suite for the sands forming the slope fans. 3) Point 2 may be repeated for the underlying Stanley as well as the overlying Johns Valley. 4) The rubbly zones, many in number, can be mapped, although little of this has been done."

STOP 7 -- WINDOWS AND KLIPPEN IN ORDOVICIAN ROCKS NEAR PARON

Near the bridge 0.2 miles southwest of Paron on Arkansas Highway 9, a major thrust fault separates the covered lower Stanley Shale from the Bigfork Chert which is exposed to the south (Fig. 18). This fault marks the boundary between the Aly structural belt on the north and the Avilla belt on the south. The cherts and siliceous limestones of the Bigfork are intensely sheared. Traces of petroliferous minerals were noted in quartz-calcite veinlets cutting the unit. Continuing southwestward, thin-bedded, dense, gray limestone and shale of the upper Womble Shale are exposed. Next, along the north side (notice abundant cedar trees), a very large, low-angle, folded fault plane (thin white gouge zone) separates the Womble from banded slate, limestone and siltstone of the Mazarn Shale. Some of the banded slates of the Mazarn contain a significant graptolite fauna which is currently being studied by Stan Finney of Oklahoma State University. Several generations of horizontal to steeply plunging folds are present. Intense northward-dipping cleavage and large milky quartz veins are characteristic of this outcrop.

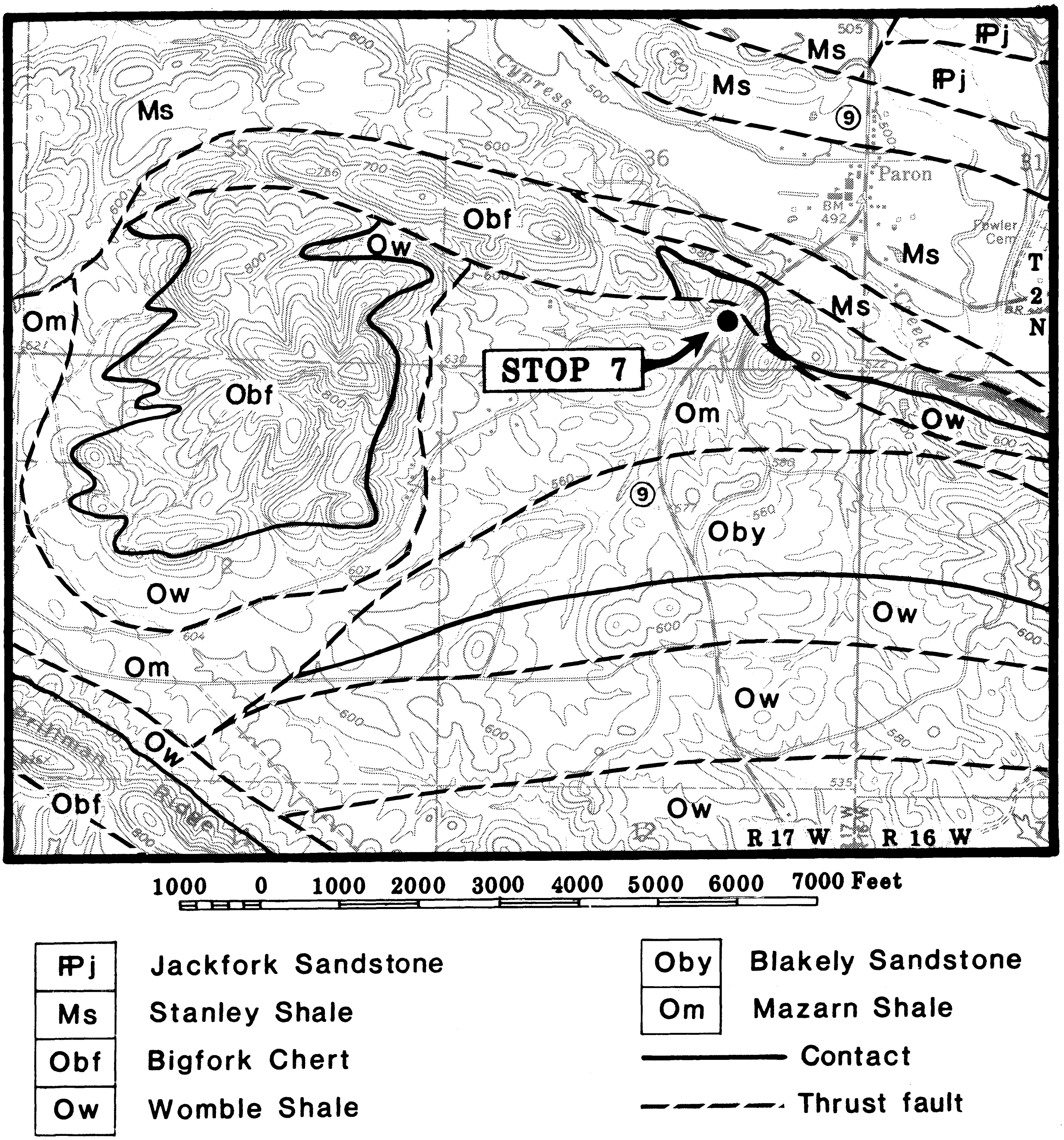


Figure 18. Geologic map of the Paron area showing location of Stop 7. Arrow marks south end of exposures.

The structure in this area is complex and it goes without saying that interpretations may change. First, along Highway 9, the map trends indicate east-west or northwest-southeast striking beds. Dips are consistently to the north and the succession of the formations as a whole indicates a stratigraphic section that becomes progressively younger northward. Axial surfaces and cleavages are moderately inclined and dip north, but fold hinges are consistently of high rake; the folds are either raking folds or reclined folds.

Toward the west (Fig. 18), in an area unfortunately inaccessible to buses, there are two isolated areas of Bigfork Chert and Womble Shale lying across the Mazarn Shale. As they cannot be connected to any nearby outcrops of Bigfork, they do not appear to be the result of cross folding but are more probably klippen. Viele thinks they may be part of the nappe composing the Ellis Mountain-Ferndale trend. On the west side of the southernmost klippe, recumbent isoclinal folds having northward rotations are backfolded to Z patterns when viewed down plunge. Z folds on the lower limbs are accented; S folds on the upper limbs are damped.

These relations along the highway and at the klippen suggest superposed folding, -- a northward direction of tectonic transport followed by backfolding to the south.

STOP 8 -- THRUST FAULT BETWEEN THE WOMBLE (ORDOVICIAN) AND STANLEY (MISSISSIPPIAN) FORMATIONS NEAR STEINER MOUNTAIN

The incomplete series of exposures at the northeast end of Steiner Mountain in Saline County illustrates a significant low-angle thrust fault near the eastern terminus of the Mt. Ida structural belt (Fig. 19). Shales and conglomeratic limestones in the lower part of the Middle Ordovician Womble Shale (of the Mt. Ida belt) are thrust over shales and minor siltstones of the Mississippian Stanley Shale (of the Aly belt). Recently, Ray Ethington, University of Missouri, obtained conodonts from limestones submitted by Stone from this locality. The conodonts corroborate the earlier assignment of these rocks to the lower part of the Womble Shale.

The prominent structural indentation (Fig. 19) consisting of the Stanley Shale and some older rocks of the Aly belt along the Alum Fork of the Saline River is named the Alum Fork re-entrant. Immediately to the east and southeast of this area, the rocks of both the Aly and Mt. Ida belts (Fig. 20) are overthrust by the Mazarn and other Ordovician rocks belonging to the Hopper and Avilla belts and this major thrust fault has been named the Alum Fork décollement. Along the surface trace of the Alum Fork décollement, displacement of at least 8 miles can be measured.

The Mt. Ida belt in the vicinity of Steiner Mountain also contains several other structural plates, thrust faults, and probable duplexes that, through erosion, have left a series of windows, klippen and other features. Several of the thrust faults are estimated to have several miles of displacement. Some typically small milky quartz veins of hydrothermal origin are present in the rocks comprising the Mt. Ida and Aly belts in this area, but they are significantly more abundant and larger in the Hopper and Avilla belts.

There have been numerous investigations in recent years to unravel the complex geology in the eastern part of the Benton uplift of the Ouachita Mountains. An incomplete summary of this work includes: the basic field mapping in the late 1950's and early 1960's by Sterling, Stone and Holbrook; somewhat later, the initial quantitative structural analysis mostly by Viele; later the regional and detailed mapping by Haley and Stone; next the studies by Viele and students, concurrent with more work by Haley, Stone, Ethington and others. Several publications by Viele have provided an explanation for the origin of the complex structure in these rocks. Haley and Stone have offered alternative proposals but until recently have published relatively little on this subject. Figure 20 is a nearly north-south cross section through this region. We believe that this represents a logical interpretation of the structure of these rocks. Major décollements are shown as flooring the strata in the Avilla and Hopper belts and overriding the more northerly facies of the Mt. Ida and parts of the Aly belts. We also show the Maumelle chaotic zone of Viele (1973) as a tectonically imprinted, mostly slope facies of the Jackfork Sandstone and other formations.

Time, weather, and road conditions permitting, the trip will proceed northward on Weyerhaeuser

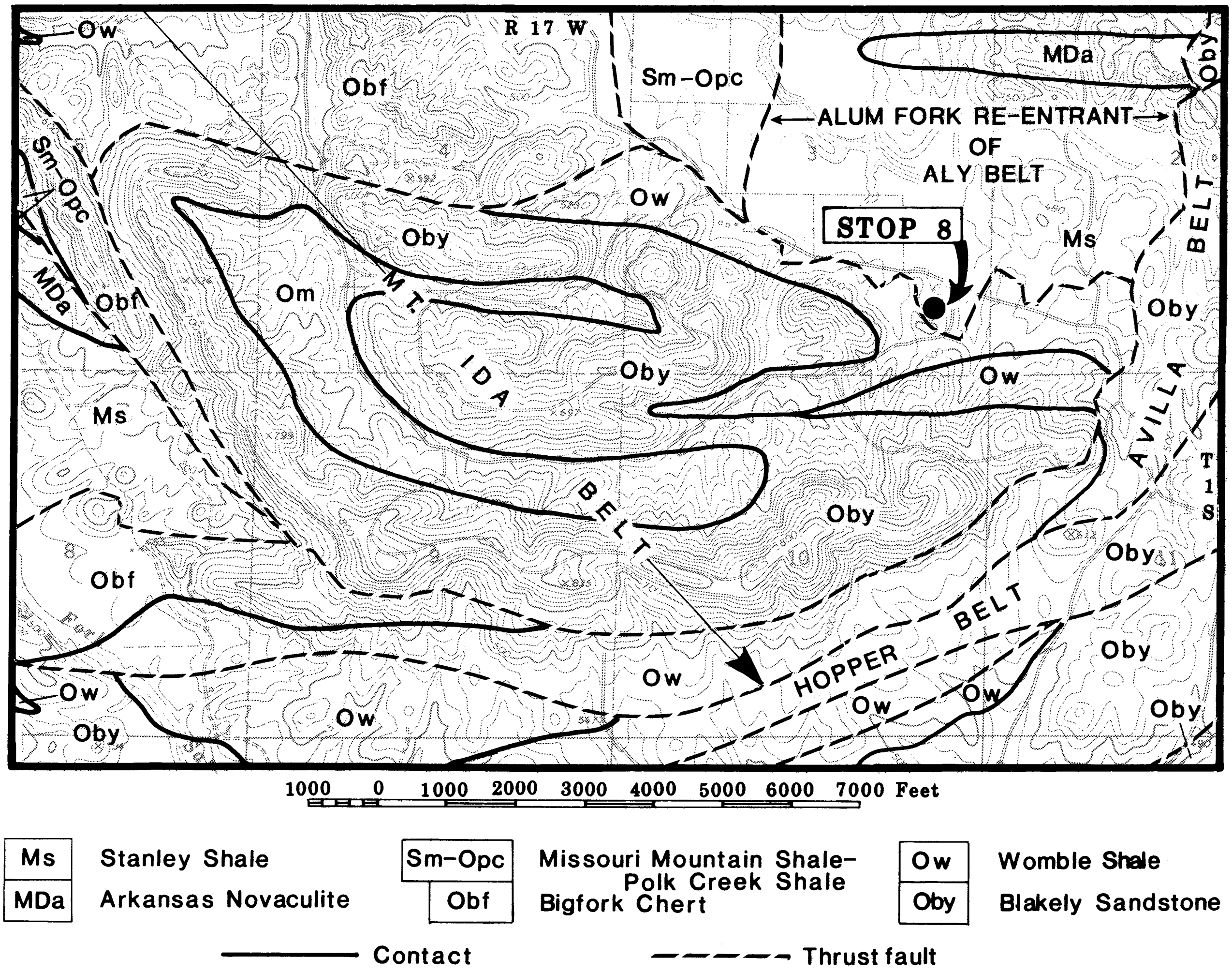


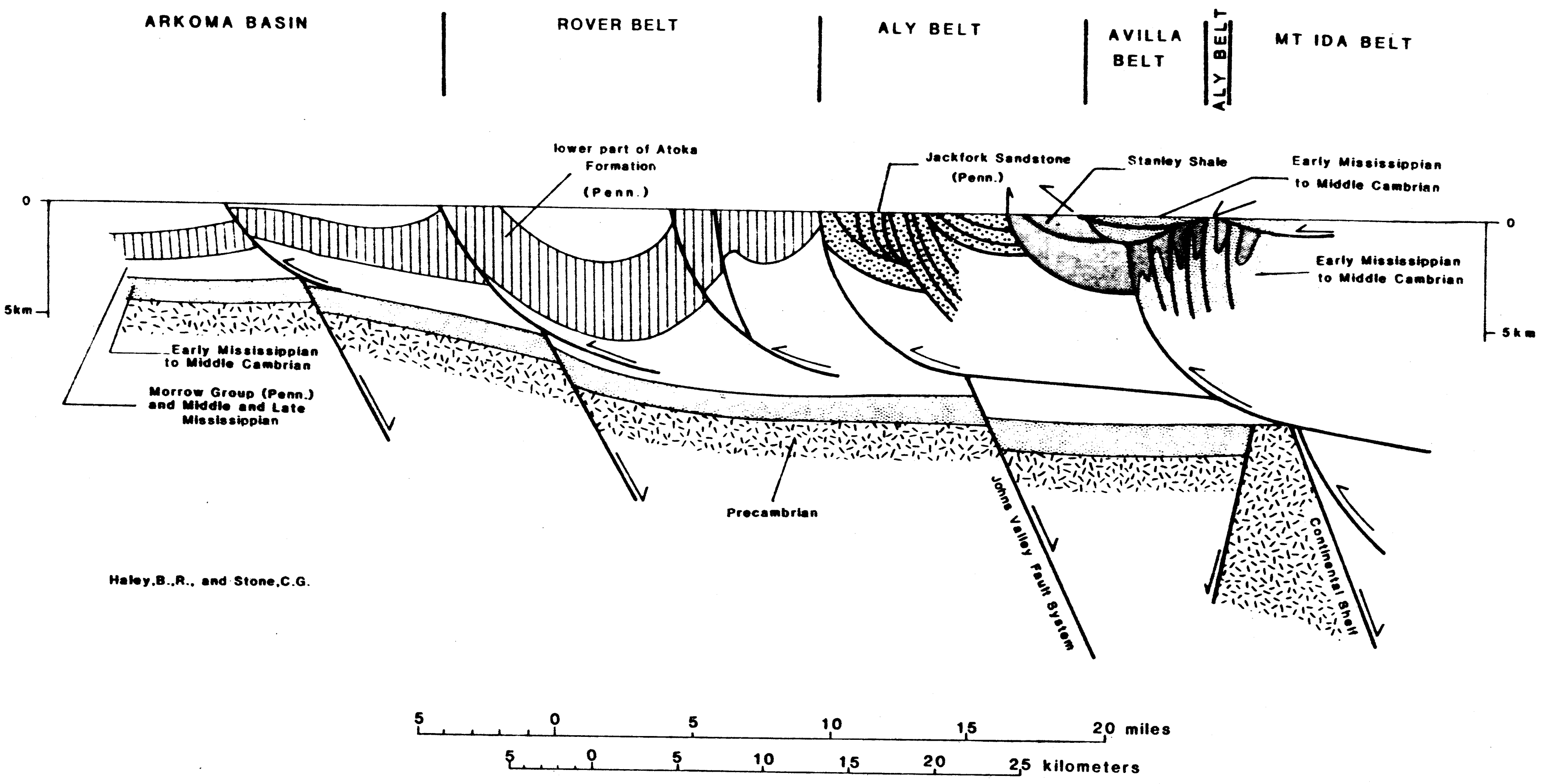
Figure 19. Geologic map of Steiner Mountain area showing location of Stop 8. Geology from Haley and others, 1976.

Company road No. 24530 for about 1/2 mile to the Alum Fork of the Saline River. At this locality the lower Stanley Shale can be examined in the Alum Fork re-entrant. A side trip may also be made into another drainage about 1/4 mile to the northeast where there is an exposure of a 5-to 15-foot-thick calcite-bearing, mylonized breccia that represents the Alum Fork décollement zone. At this locality micritic, silty limestones and black shales of the Lower Ordovician Mazarn Shale overlie shales with some very thin graywackes and cone-in-cone concretions of the lower Stanley Shale (Mississippian).

For about the next 20 miles, travel will be primarily on Weyerhaeuser Company roads. We wish to extend thanks to Mr. Bob Bearden and Mr. William Willis of Weyerhaeuser for their assistance and for obtaining permission to use the roads.

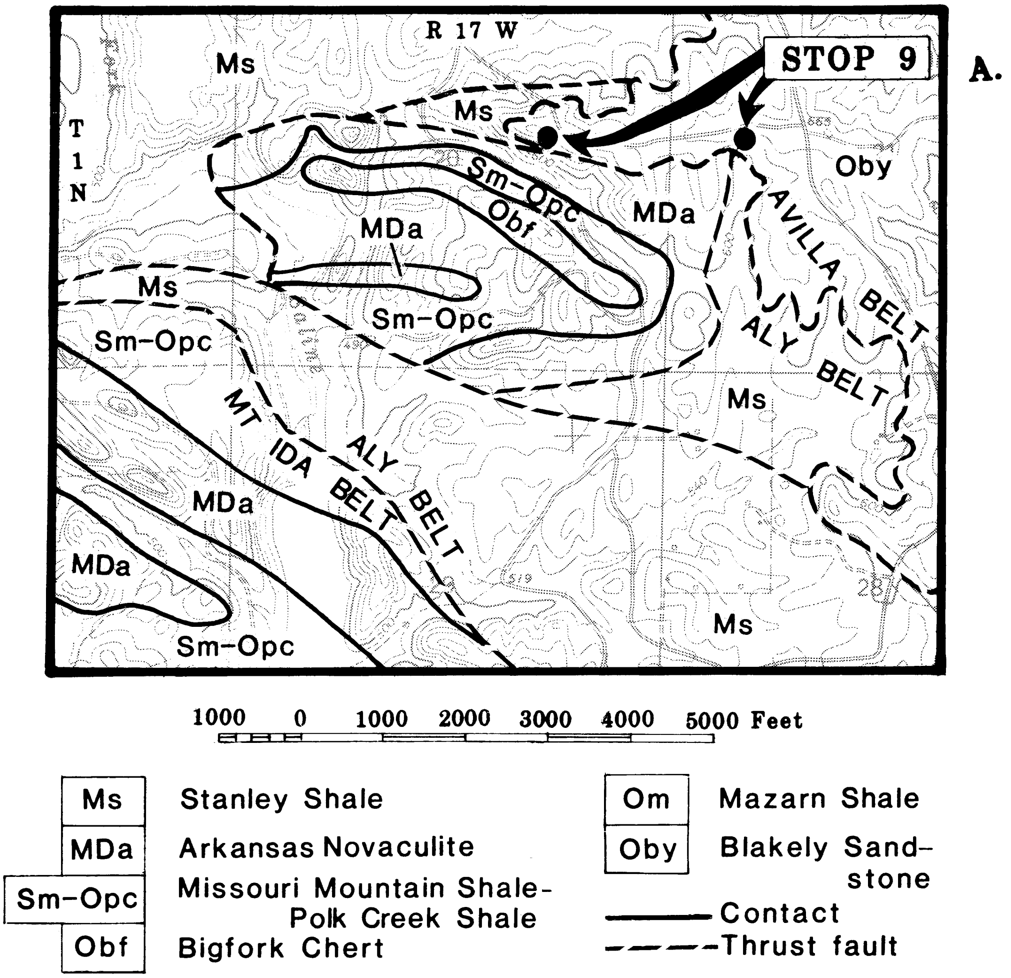
STOP 9 -- ALUM FORK DÉCOLLEMENT WITH BLAKELY SANDSTONE AND MAZARN SHALE IN FAULT CONTACT WITH STANLEY SHALE AND ARKANSAS NOVACULITE

This rather poor exposure is in the Avilla structural belt (Fig. 21) and the exposed sequence of shale, siltstone, and scattered beds of clean, quartzose sandstone belongs to the Blakely Sandstone.



Haley,B.,R., and Stone,C.G.

Figure 20. Schematic north-south cross section showing the succession of structural belts recognized across the Ouachita Mountain fold belt. For location of section see Figure 8.



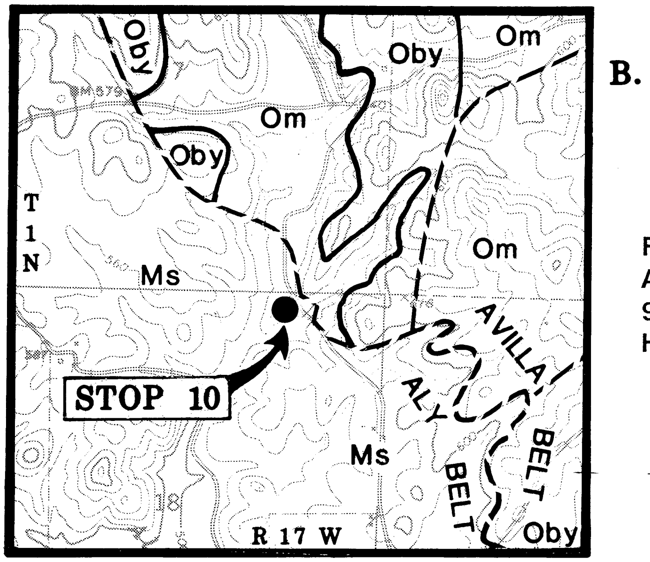


Figure 21. Geologic maps of parts of the Alum Fork décollement. A. Vicinity of Stop 9, and B. vicinity of Stop 10. Geology from Haley and others, 1976.

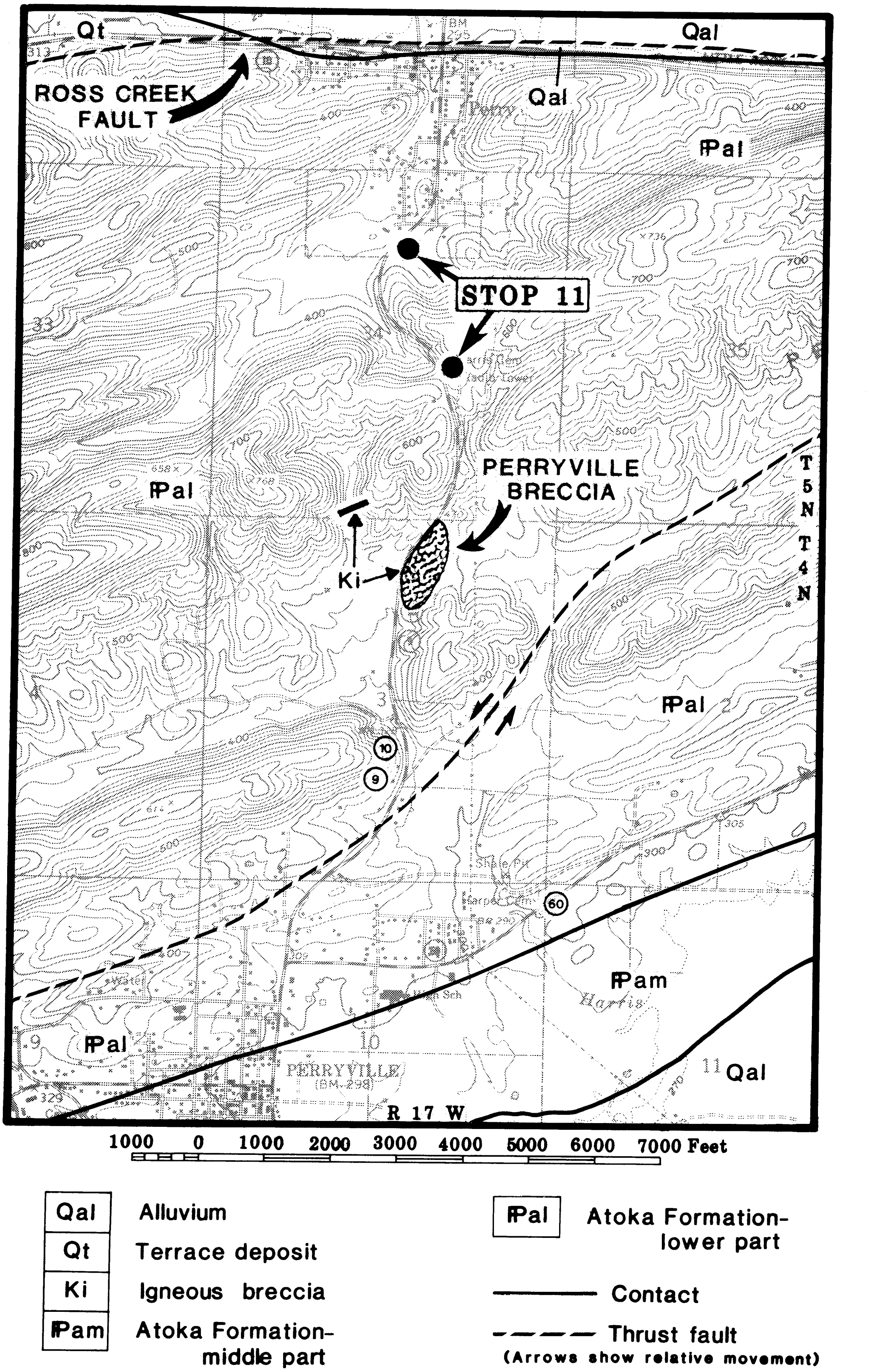


Figure 22. Geologic map of the Perry-Perryville area showing location of Stop 11. Geology by Haley and others, 1976.

Locally there are some erratic granite and meta-arkose boulders and cobbles in the formation. Proceeding down the hill to the west, there are sequences of interbedded weathered limestone, calcareous siltstone, and shale of the Mazarn Shale. Near the base of the hill is a thin sheared interval representing the fault plane of the Alum Fork décollement. Below the fault are some exposures of the lower Stanley Shale of the Aly belt. The northern facies of the Arkansas Novaculite is present immediately to the south and is partially exposed in an adjoining roadcut.

Large milky quartz veins with accumulations of residuum are present in the older overthrust rocks, but are significantly less common in the underlying younger strata. As previously stated, a minimum of at least 8 miles of displacement can be measured along the fault in this area. In all likelihood however, this fault has a displacement of 20 to 40 miles. The true magnitude of the displacement is indicated by the presence of the southern facies of the Arkansas Novaculite in the Avilla belt a few miles northeast of this area, whereas here, the northern facies of the novaculite is present. It is interesting to note that most, if not all, of the numerous thrust faults and related splays of the underlying plate (Aly belt) are also covered by the overlying plate (Avilla belt).

STOP 10 -- ALUM FORK DÉCOLLEMENT WITH MAZARN LIMESTONE (LOWER ORDOVICIAN) IN FAULT CONTACT WITH STANLEY SHALE (UPPER MISSISSIPPIAN)

This is another opportunity to view the rocks on both sides of the Alum Fork décollement (Fig. 21). Along the road, tightly folded, crinkled and cleaved shales of the lower Stanley Shale in the Aly belt are poorly exposed. In the small creek some 25 yards to the north is a largely covered interval containing a thin brecciated, mylonized sequence that represents the fault zone. Directly above this is a very thin sequence of dense, tightly buckled, often recumbent, micritic limestones and some black shales belonging to the Mazarn Shale of the Avilla belt. Samples of limestone from this location submitted to Ray Ethington of the University of Missouri did not yield conodonts, but at Hester Chapel about 1 1/2 miles to the north, a conodont of Lower Ordovician Mazarn affinities was obtained. As stated in the description of the last stop, the structural displacement of the Alum Fork décollement is probably 20 to 40 miles. We have also mapped other major thrust faults that have sizable but unknown displacements in the Avilla belt structurally above the Alum Fork décollement.

In the late 1950's when the work in this region was instigated, in part at the suggestion of the late Hugh D. Miser of the U.S. Geological Survey, it was the prevailing opinion of several geologists working primarily in Oklahoma that the rocks in the Ouachita foldbelt were, in general, very simply folded and not appreciably thrust faulted. One notable investigation described all the Ouachita strata as being autochthonous. This thesis was strongly contradicted by a number of geologists. There were many long and intense (but generally friendly) debates about this complexly rucked stack of Paleozoic rocks. However now, most investigators seem to be in general agreement on several critical observations: 1) the rocks are mostly of deep water origin, 2) they are allochthonous, and 3) there are major thrust faults and there are recognizable structural belts. The real "burning" questions that remain are: Where did these rocks come from? How did they get here? And finally, why are they deformed as they are?

STOP 11 -- LOWER PART OF THE ATOKA FORMATION ON ARKANSAS HIGHWAY 10 NEAR PERRY

Begin walking north along the highway to the bottom of the hill (Fig. 22). *Be careful -- there is much traffic on this road!* This locality is in the Rover belt and the rocks are in the lower part of the Atoka Formation. They exemplify many of the characteristics of flysch deposits. The sequence (Fig. 23) is a

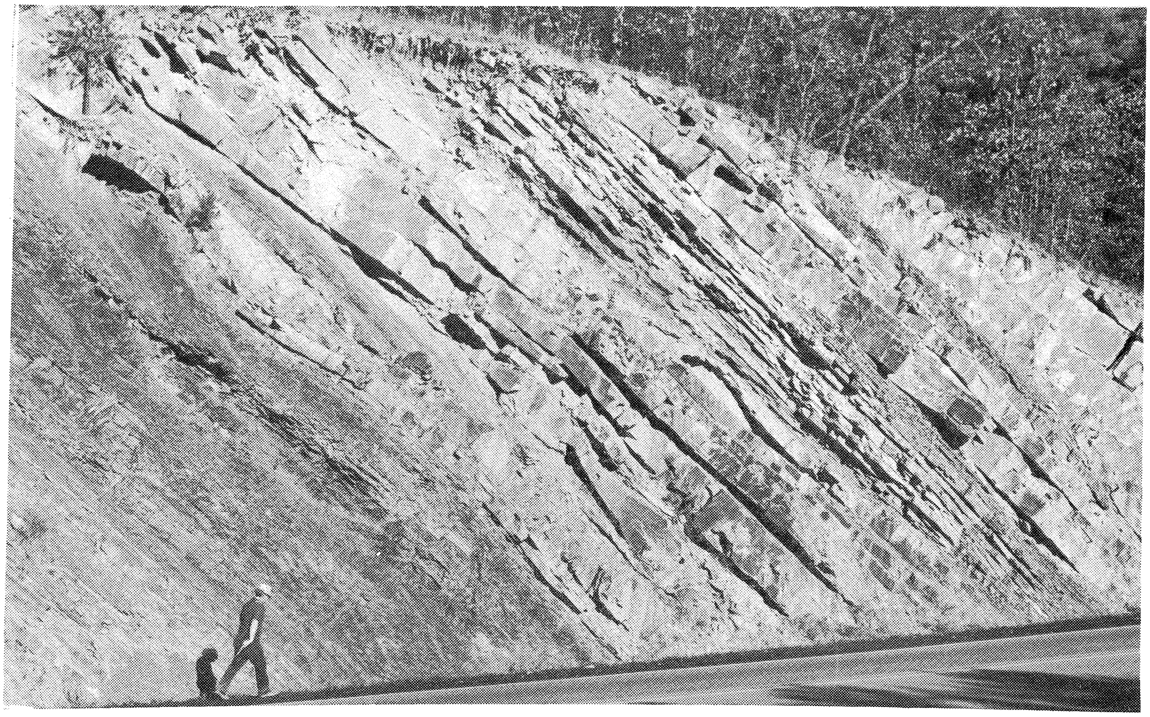


Figure 23. Stop 11. **Rocks in the lower part of the Atoka Formation exposed on east side of** Arkansas Highway 10 between 0.6 and 1.1 miles south of Perry. Interbedded shale and sandstone, the latter in thinning-and fining-upward units, probably represent deep-water submarine fan deposits.

repetitious alteration of fissile black shale with some siderite concretions, micaceous siltstone containing coalified plant debris ("blue beds"), and subgraywacke sandstone. The latter is thin to thick bedded, graded bedded, and many sandstone beds bear bottom markings and may be convolute bedded. Trace fossils indicative of deep water are abundant. Many of these beds represent submarine fan lobes and others represent channel deposits. The sequence is believed to have accumulated in a mid-fan environment.

The rocks here were placed in the Atoka Formation by Croneis (1930). He considered the formation to be about 10,000 feet thick. They were later placed by Stone (1968) in the lower part of the Atoka. We now believe that the lower part of the Atoka is closer to 13,000 feet in thickness and that the total Atoka Formation is about 20,000 feet thick. Based on spore studies by Quinn (1986), the lower part of the lower Atoka is regarded as Morrowan in age.

Near the railroad tracks at Perry, Arkansas some 0.5 miles to the north, the Ross Creek fault has thrust lower Atoka rocks northward over upper Atoka rocks. However in the immediate vicinity of Perry, the upper Atoka does not crop out from beneath the alluvial cover, hence its apparent absence in Figure 22. The stratigraphic displacement across the fault is more than 15,000 feet.

The Perryville igneous breccia (a lamprophyre-carbonatite intrusive), which is probable early Late Cretaceous in age, is exposed shortly south of here near the lone house on the east side of the highway.

STOP 12 -- LOWER JACKFORK SANDSTONE AT THE SPILLWAY ON LITTLE BEAR CREEK LAKE NEAR HOLLIS

On several occasions black bears have been seen near this locality.

More than 100 feet of rather intensely fractured and quartz-veined quartzose sandstone, siltstone and shale of the lower Jackfork Sandstone are exposed along the east wall of the spillway (Fig. 24). This section is about 900-1000 feet above the base of the formation. The rocks are gently dipping and occur near the crest of a slightly vergent (southward) anticline in the Aly belt of the frontal Ouachita Mountains. Most outcrops elsewhere in the area are very steeply dipping. A major east-west thrust fault is present in the lake and dam area to the south. These rocks (Fig. 25) probably represent slightly incised channel-fill sequences on the upper part of a submarine fan. At least five channeled sequences showing signs of migration and reactivation are exposed. The base of the sequence contains sedimentary slump and large pull-a-part features that are further disturbed by structural flowage and small bedding plane and other thrust faults.

Paleocurrent data indicate a source to the northeast and east. We believe that the major source area for the Jackfork Sandstone in this region was to the east or northeast. However, as indicated by Morris (1985) and others, much of the Jackfork in the southern Ouachita Mountains was probably derived from a metasedimentary terrane to the southeast.

The rocks in this portion of the Aly belt are intensely thrust faulted and display prominent shearing. Small fracture-filling quartz veins (Fig. 25) and slickensides coated with yellowish-white splotches or groove-fillings of dickite are abundant. Two major generations of quartz veins are present in this portion of the frontal Ouachita Mountains (Stone and Milton, 1976). The first contains dickite and the latter rectorite and cookeite, both of which are associated with minor ore minerals and other minerals. Some of the quartz veins are crinkled or rotated, indicating either a concurrent component of shear or possibly suggesting a change in the axis of compression during late episodes of Ouachita tectonism.

The major stratigraphic and structural boundary in the frontal Ouachita Mountains is along the northern limits of the Aly belt near the community of Hollis, about two miles north of here. The major thrust faults present in this area (by some interpretations) may represent the Winding Stair, Honess, Briery, Y City, and Ti Valley faults of western Arkansas and eastern Oklahoma. The various sequences of Stanley Shale, Jackfork Sandstone, Johns Valley Shale and Atoka Formation are significantly displaced. Interestingly, some of these units locally contain olistoliths derived from foreland facies to the north (e.g., Gordon and Stone, 1977). This belt of structurally complex rocks has been identified by Viele (1973, 1979) as the Maumelle chaotic zone and as the suture between the North American and South American plates. However, Morris (1985) states that these flysch sequences contain some beds of rounded non-feldspathic quartz sandstone with stable heavy minerals that had a cratonic source to the north and were developed upon the North American plate. We presently concur with Morris' observations and further propose that these structurally complex rocks have resulted from several periods of thrust faulting through a sequence containing intervals of sedimentary mélange -- notably in the middle part of the Jackfork Formation.

Northward from the Aly belt is the Rover belt. It is about 20 miles wide and consists of more gently deformed and less imbricately thrust-faulted sequences of flysch deposits in the lower Atoka Formation. At Danville and Ola the Ross Creek thrust fault is recognized as the north edge of the Rover belt as well as the boundary between the frontal Ouachita Mountains and the Arkoma basin.

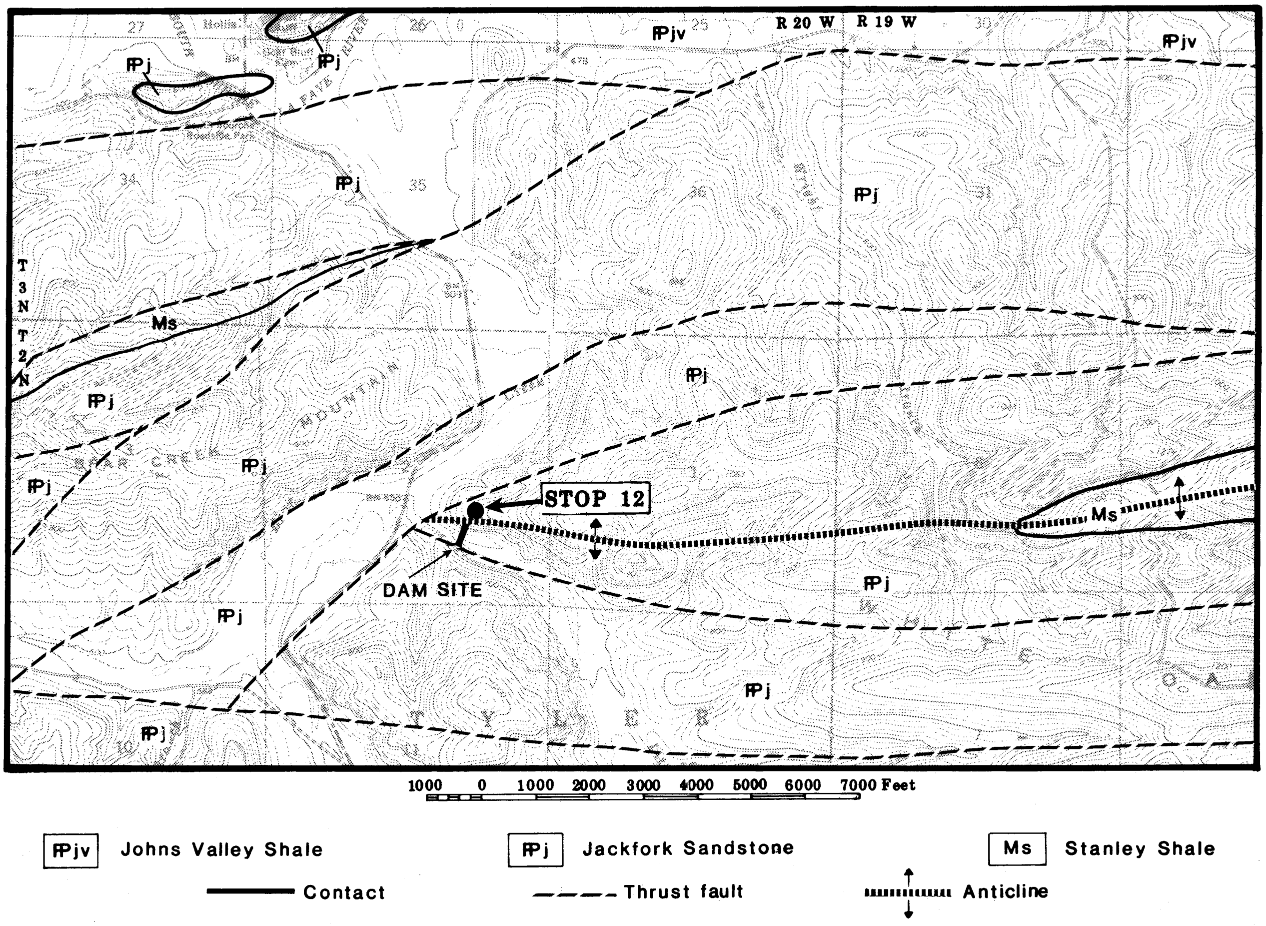


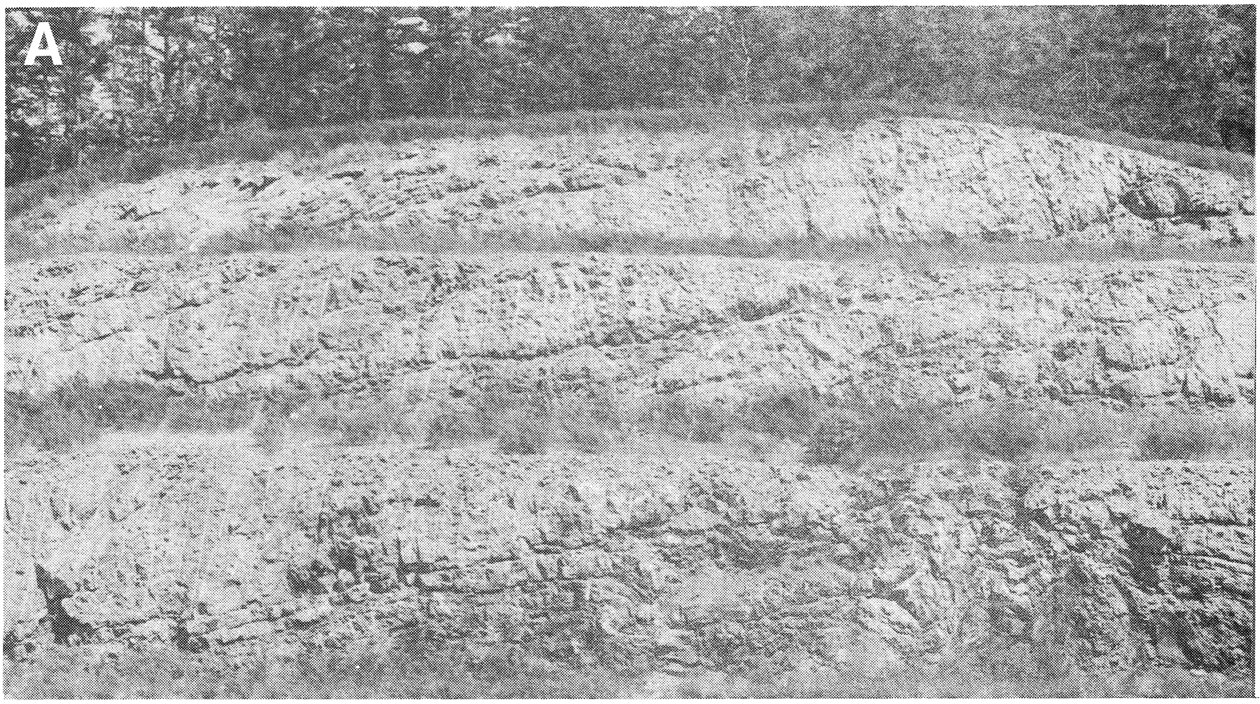
Figure 24. Geologic map of the Bear Creek area showing the location of Stop 12. Geology by Haley and others (1976).

STOP 13 -- GEOMEX QUARTZ CRYSTAL MINE

We wish to express our gratitude to Mr. John Long, Superintendent, Geomex Mine Services, Inc., for providing access to their quartz crystal mine and for his invaluable assistance in the examination of these classic deposits.

Mining of quartz crystals in the Ouachita Mountains of Arkansas has been going on for many years, the first miners probably being the Indians who shaped them into arrowheads. Because of the clarity and perfect shape of many of the individual crystals and crystal clusters, the principal market over the years has been as specimens in both individual and institutional mineral collections. During World War II about five tons of clear quartz crystals from Arkansas were used in the manufacture of radio oscillators to supplement the production from Brazil. Currently quartz crystals are being used for: manufacturing fusing quartz, which has many chemical, thermal and electrical applications; for seed crystals (lasca) for growing synthetic quartz crystals; and, of course, for mineral specimens. It should be noted that the "Hot Springs Diamonds" for sale in the local rock shops and jewelry stores are cut from Arkansas quartz crystals.

Quartz veins are numerous and are found in a wide belt extending from Little Rock, Arkansas to Broken Bow, Oklahoma in the central core area of the Ouachita Mountains. These veins, up to sixty feet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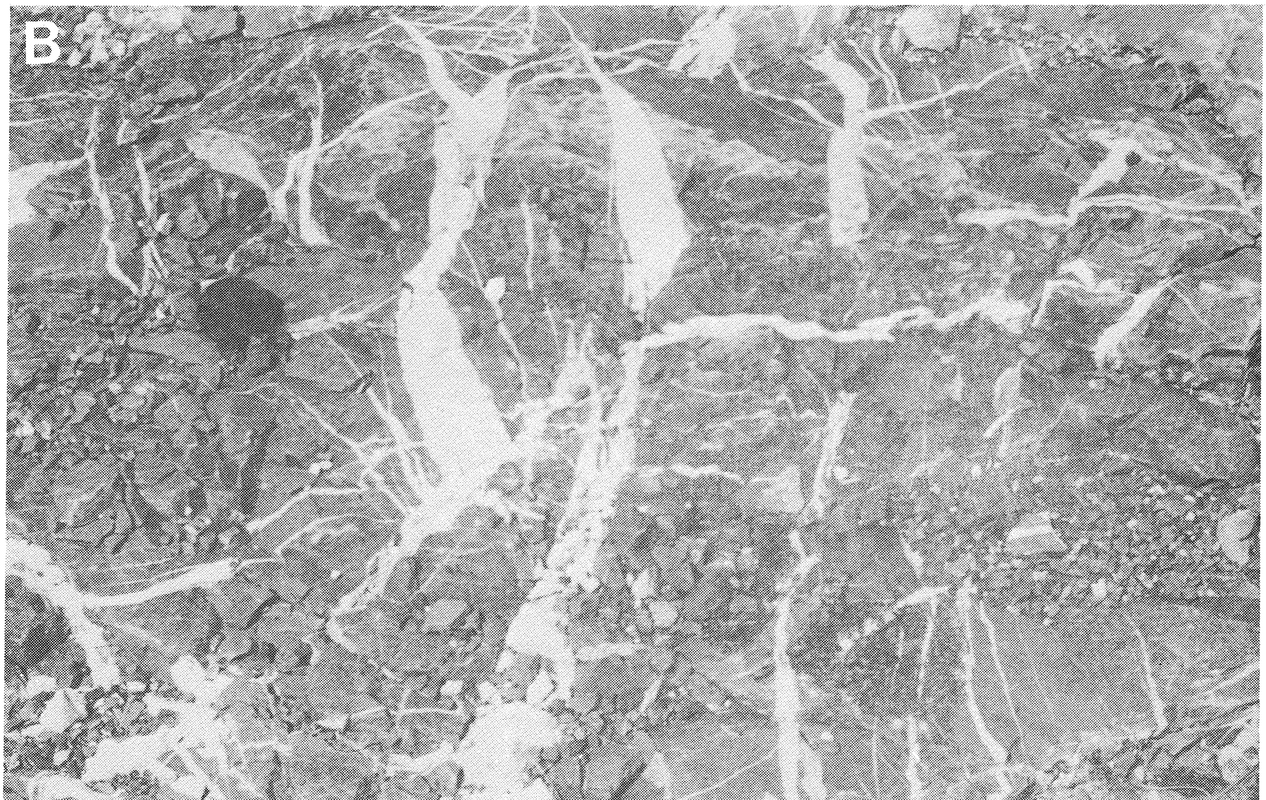
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in width, commonly contain traces of adularia, chlorite, calcite and dickite. In a few places lead, zinc, copper, antimony and mercury minerals are associated with the quartz veins. At relatively few localities however, do individual quartz crystals and crystal clusters attain the size and clarity requisite for mining.

In the Ouachita Mountains there is a close association of quartz veins with some fault zones. It is believed that the quartz veins represent, in part, dewatering processes that took place along the fault zones. The increase in pore fluids may well have contributed to overpressuring and related conditions and enhanced the overall faulting and folding processes. The quartz veins with their associated minerals are presumed to be hydrothermal deposits of tectonic origin formed during the closing stages of the Late Pennsylvanian-Early Permian orogeny in the Ouachita Mountains.

The Geomex Mine is also known as the Coleman Mine, the West Chance Area, Dierks No. 4 Mine, and The Blocher Lead (Fig. 26). It is located just north of a county road about 1.8 miles west of Blue Springs. The quartz crystals occur in veins in limy sandstone and conglomeratic sandstone beds of the Blakely Sandstone (Ordovician). Beds of conglomeratic sandstone exposed in the pit contain abundant weathered meta-arkose and granitic boulders, cobbles, and pebbles, and some clasts of limestone, chert and shale. It is likely that these sediments were deposited in submarine fan channels and were derived from a granite-rich terrane to the north-northeast. It has been postulated by Stone and Haley (1977) and a number of other workers that these exotic boulders are probably Precambrian in age. Some have expressed the opinion that they represent early Cambrian accumulations. Recent work by Bowring (1984) indicates a middle Proterozoic age. This area includes many thrust-faulted sequences with at least two major periods of folding resulting in differing attitudes in fold hinge lines and axial planes. The mine itself is situated on the nose of a large, complexly deformed syncline.

The quartz crystal veins are fracture fillings with the larger and more productive cavities being located at the intersection of two veins. Mining operations are relatively simple, consisting initially of removing overburden and loose rock with a bulldozer to expose the crystal-filled cavities, and then removing the quartz crystals with hand tools. Individual quartz crystals up to five feet in length weighing as much as 400 pounds and clusters 15 feet in length weighing over five tons have been produced from these mines.

Figure 25. Stop 12. Lower part of the Jackfork Sandstone exposed at Little Bear Creek Lake spillway near Hollis, Arkansas. A. Several entrenched and locally slumped sandstone sequences interpreted as upper submarine fan-channel deposits. B. Veins of milky quartz filling fractures along a small fault zone. Some veins are folded, indicating a concurrent component of shear or a later stage of deformation.

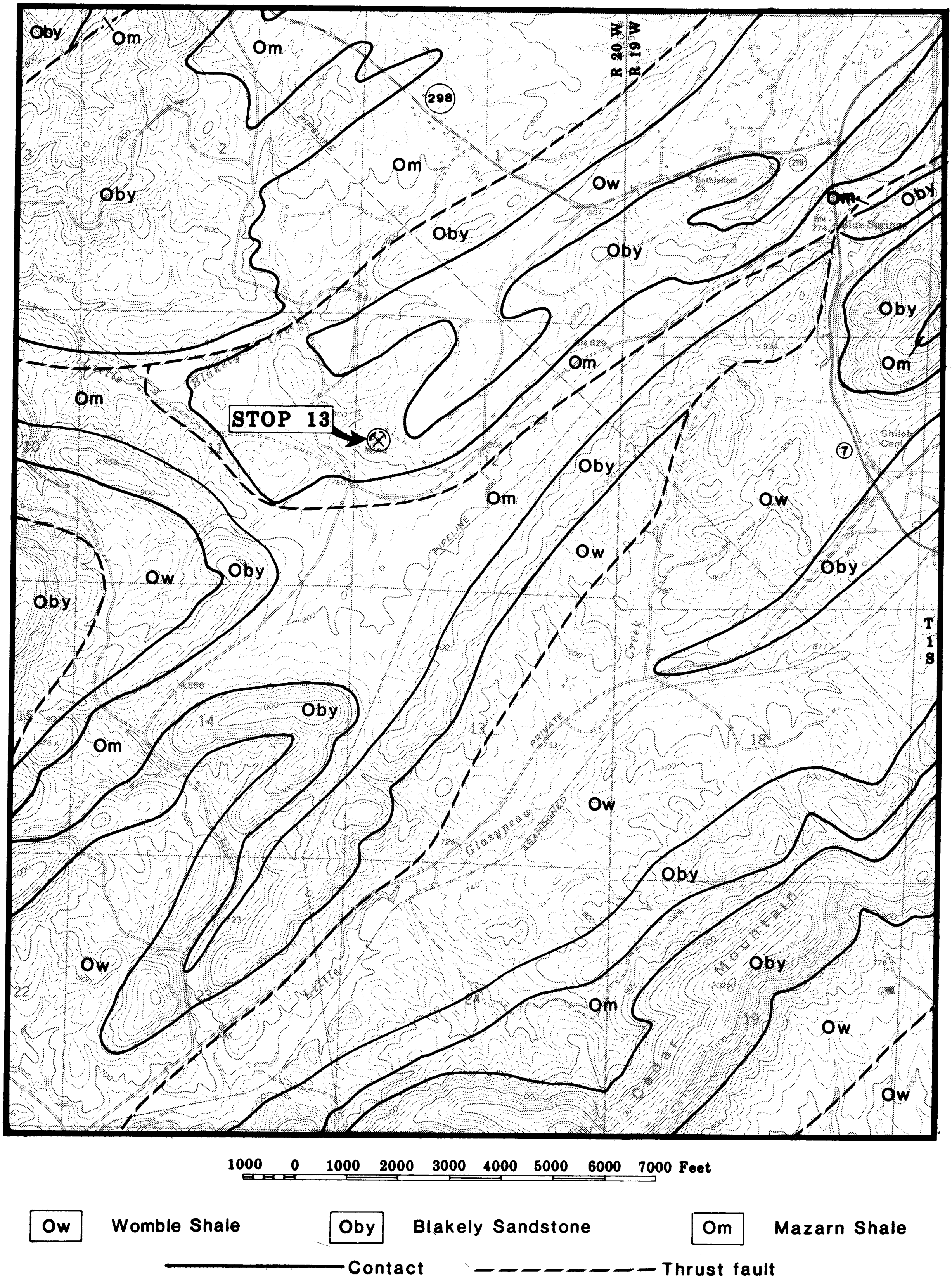
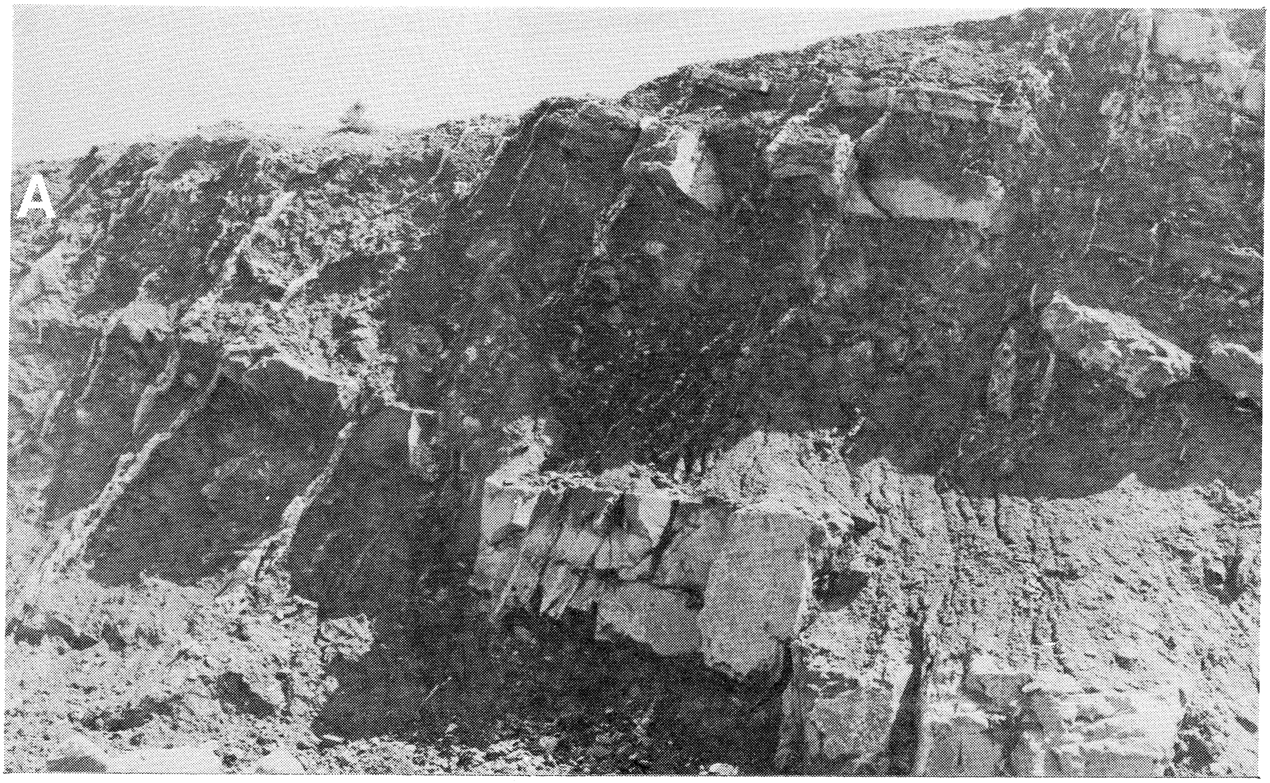


Figure 26. Geologic map of the Blue Springs area, Arkansas showing the location of the Geomex Mine (Stop 13).


A


B

Figure 27. Stop 13. Two views of the Geomex quartz crystal mine. A. Thick olistostromal sequences of Blakely Sandstone with various granite, meta-arkose and limy siltstone erratics in a decalcified sandstone with some lenses of orthoquartzitic sandstone. Small faults visibly offset some of the rocks but most of the "deformation" is considered sedimentary in origin. At least two prominent directions of fracture-filling milky to clear quartz veins are present. B. Mr. Paul Thompson, a geologist formerly with the Geomex Company, stands by large quartz cavities lined with clear and milky quartz crystals at the east end of the mine.

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Pennsylvanian paleogeography for the Ozarks, Arkoma, and Ouachita basins in east-central Arkansas

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ABSTRACT

In early Pennsylvanian time, east-central Arkansas was part of a foreland basin developing over a downwarped Cambrian to Mississippian carbonate shelf. The deep-marine Ouachita basin lay to the south and was tectonically closing by northward advancing thrust sheets. Over 30,000 feet of clastic rocks were deposited in the Ouachita foredeep in the early Pennsylvanian.

In Morrowan time, 200 to 300 feet of carbonate and shale shelf facies accumulated in northeastern Oklahoma. Eastward in north and central Arkansas, the Morrowan facies are mainly fluvial, deltaic, and shallow-marine, quartzose clastic rocks up to 2,000 feet thick. In the Ouachita basin to the south, 7,500 feet of Morrowan quartzose sandstones and shales were deposited as submarine fan, basin plain, and slope facies. Morrowan clastics were derived from the northeast via fluvial channels that cut across the Illinois basin and from the rising Appalachians to the east and southeast. In the Ouachita basin, sediment transport directions were dominantly to the west. In Atokan time, 600 to 6,000 feet of north-derived fluvial, fluvial-deltaic, and shallow-marine clastic rocks were deposited in the Ozarks and northern Arkoma basin areas. To the south, over 20,000 feet of shallow- to deep-marine clastic rocks accumulated in the center of the closing foreland basin. South of the Atokan shelf, over 10,000 feet of north- and east-derived turbidite sandstones and shales were deposited as submarine fans, basin plain and slope facies in early Atokan time. The foreland basin shallowed through Atokan time with the upper Atokan clastics being solely of fluvial, deltaic, and shoreline facies.

The foreland basin gradually closed due to continental collision and was filled primarily from east to west. A few hundred feet of Desmoinesian fluvial and deltaic facies, originally deposited by westward flowing streams, are the last deposits preserved in central Arkansas. Afterwards, continental collision culminated in the middle Pennsylvanian to early Permian and uplift and denudation of much of the basin followed.

INTRODUCTION

Mississippian and Pennsylvanian strata form a nearly continuous belt of rocks deposited in a late Paleozoic foreland or foredeep basin(s) that extend from the southern Appalachians to the Mexican border in southwest Texas (Flawn et al., 1961). Intersecting this foreland belt are the Anadarko-Ardmore aulocogen in Oklahoma and Texas, and the Reelfoot rift in Arkansas. This foreland belt is approximately 1,240 miles long and is largely buried except for outcrops in the Appalachian Valley and Ridge province, Oua-

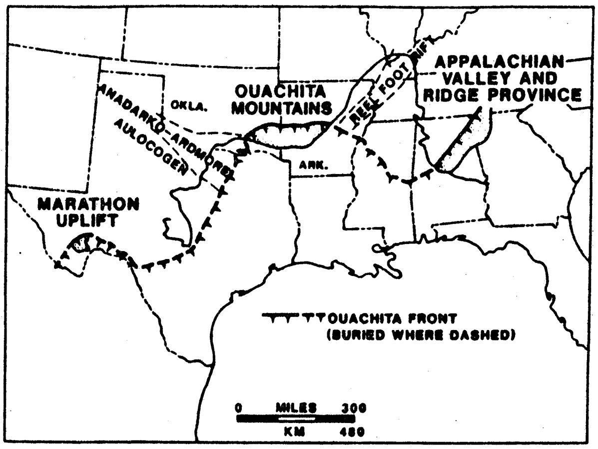


Figure 1. Map of southeastern United States showing outcrop distribution of late Paleozoic foreland basin deposits, trace of Ouachita front, and location of related aulocogens. Modified from Moiola and Shanmugam (1984).

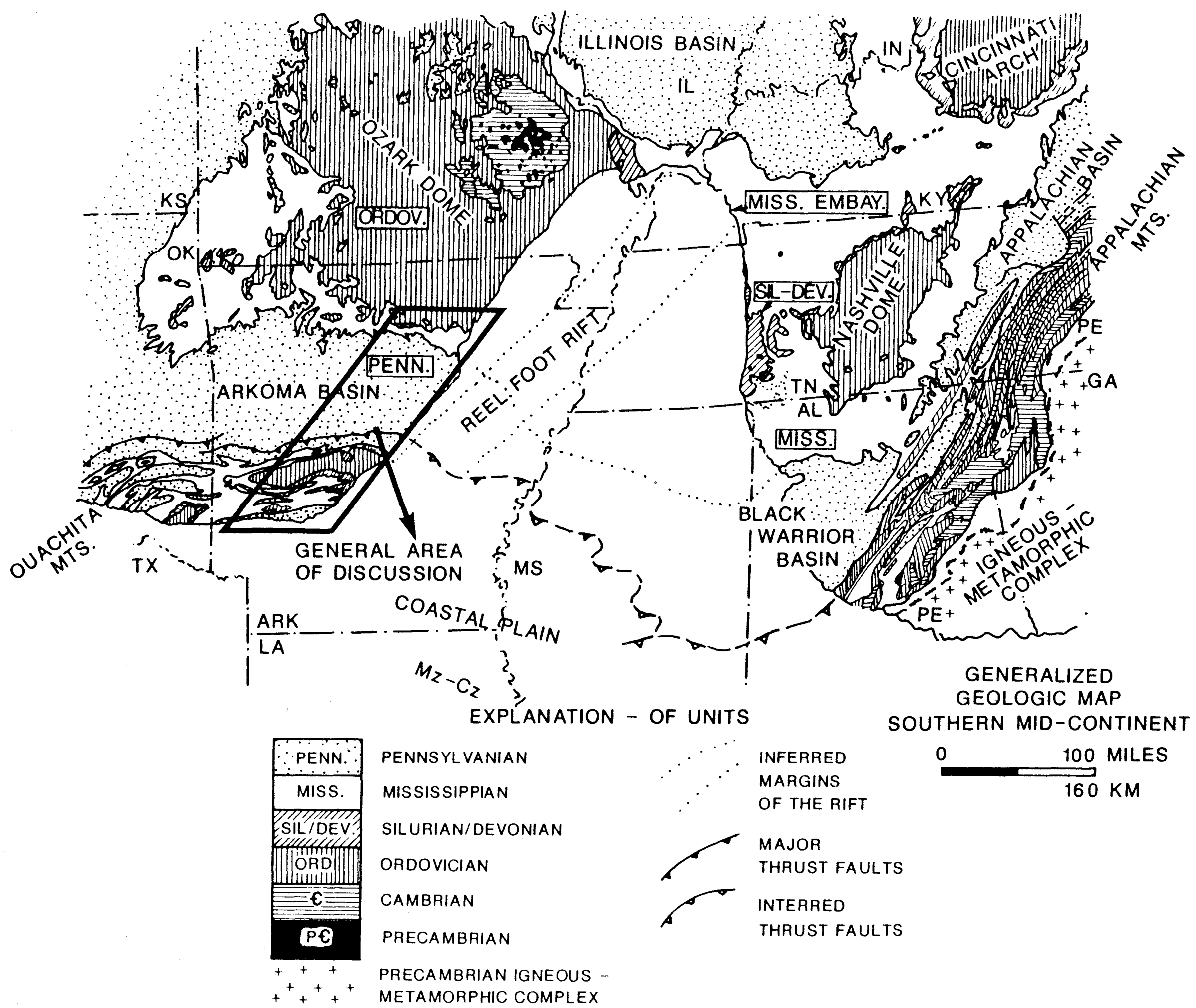
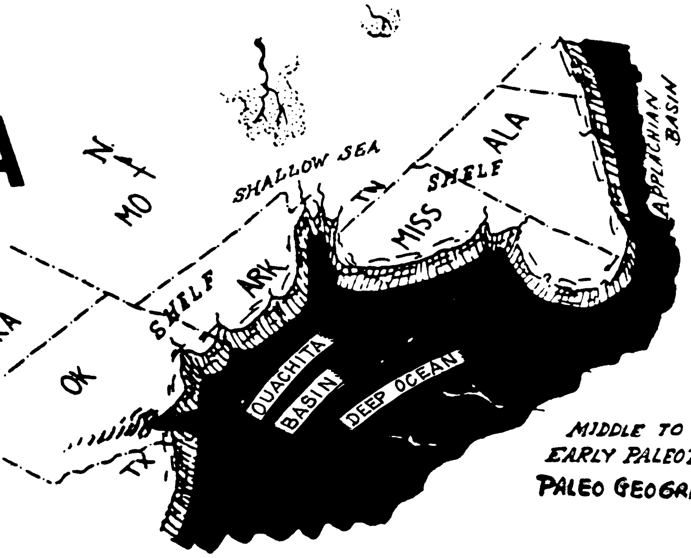


Figure 2. Generalized geologic map of part of the southeastern United States showing age of units, major structural features, and area discussed in this paper.

chita Mountains/Arkoma basin/Ozark region, and the Marathon uplift (Fig. 1). The rocks of the Arkansas portion of the foreland basin are locally over 45,000 feet thick and conformably overlie Cambrian to early Mississippian rocks (Fig. 2). These older rocks were deposited along an Atlantic-type margin during which a period of reduced sedimentation (a starved basin facies) occurred from the latest Ordovician to earliest Mississippian. Sedimentation rates increased in the late Mississippian and very rapid rates of sedimentation occurred in the early Pennsylvanian (Atokan time). During the Atokan, large volumes of sediment derived from the east and north filled the subsiding basin. The high rate of subsidence was due to thrust-sheet loading during the closing of the Ouachita basin from south to north. At the close of the Paleozoic, continental collision culminated with the emplacement of a basement duplex under the core of the thrust-belt (Broken Bow-Benton uplift). The foreland basin was in part uplifted and its uplands subjected to denudation (Fig. 3). Discussions of the plate tectonics and geologic history of the region can be found in Viele, 1973; Graham et al., 1975; Wickham et al., 1976; Briggs and Roeder, 1978; and Arbenz, 1984.

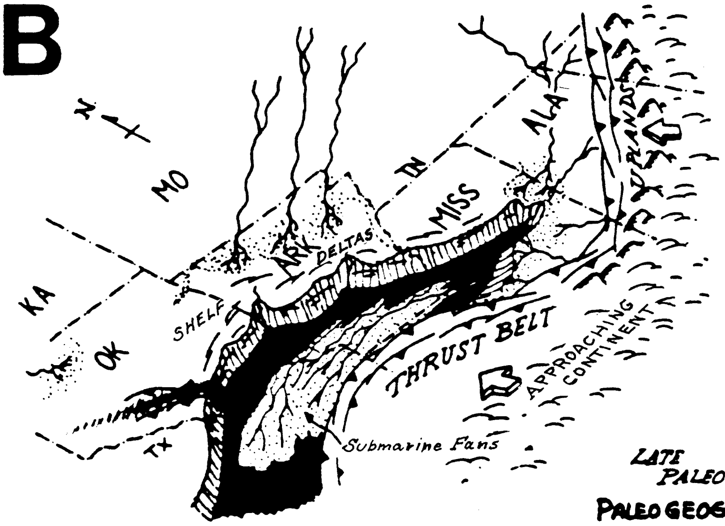
The purposes of this paper are: 1) to briefly describe the Pennsylvanian stratigraphic and tectonic relationships in east-central Arkansas; and 2) to present some interpretative diagrams for Pennsylvanian paleogeography for this area. The area of discussion extends from Friendship, Hot Spring County, to Batesville, Indepen-

A



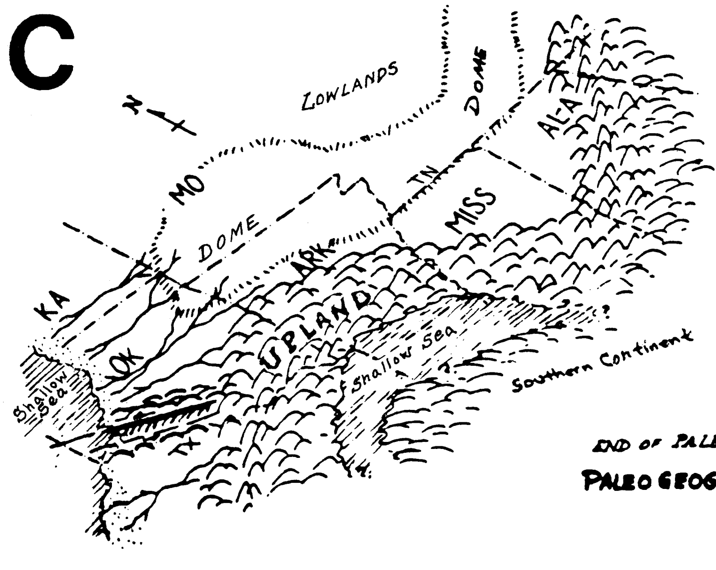
MIDDLE TO EARLY PALEOZOIC PALEO GEOGRAPHY

B



LATE PALEOZOIC PALEO GEOGRAPHY

C



END OF PALEOZOIC PALEO GEOGRAPHY

Figure 3. Paleogeographic maps in the Appalachian and Ouachita basins for the early to late Paleozoic. A) Cambrian to Mississippian interval representing a dominantly passive Atlantic-type continental margin. B) Mississippian to early Pennsylvanian interval, showing the closure and uplift of the southern Appalachian basin and narrowing of the Ouachita basin by northward advancing thrust sheets. C) Middle Pennsylvanian to Permian interval depicting the ultimate closing of the Ouachita basin, reactivation of the Ozark dome, and uplift and erosion of the foreland basin. Figure A modified from Chamberlain, 1978.

dence County, Arkansas, and includes the easternmost exposures of Pennsylvanian rocks in the Ouachita Mountains, Arkoma basin, and the Ozarks (Fig. 2). Pennsylvanian deep-water facies have been described in some detail in this area by Cline, 1966, 1970; Briggs and Cline, 1967; Morris, 1971, 1974; Moiola and McBride, 1978; Thomson and LeBlanc, 1975; Stone and McFarland, 1981; Stone and Bush, 1984; and Moiola and Shanmugam, 1984, among others. The nonmarine and shallow-marine facies in this area, in contrast, have not received as much attention. Equivalent rocks to the west in western and northern Arkansas and eastern Oklahoma have been described in some detail by Stone, 1968; Sutherland and Manger, 1977, 1979; Zachry, 1975, 1977; Zachry and Haley, 1975; and Stone and McFarland, 1981.

GEOLOGIC SETTING

Between 1,000 and 30,000 feet of Pennsylvanian strata accumulated in parts of the Ozarks, Arkoma basin, and Ouachita basin in east-central Arkansas (Fig. 4). In the Ozarks and Arkoma basin, these rocks overlie about 5,000 feet of shelfal lower Cambrian to Mississippian rocks that in turn rest on Precambrian granitic basement. The Pennsylvanian section is 1,000 feet thick in the northern Arkoma basin and ranges from Morrowan to Desmoinesian (lower to middle Pennsylvanian) in age. The equivalent section in the Ouachita basin to the south (Ouachita facies) is over 30,000 feet thick and overlies about 20,000 feet of Ordovician-Mississippian basinal strata. There are marked differences in the lithologic character, thickness, and depositional patterns of age-equivalent units between these two areas.

The Ozarks, Arkoma basin, and Ouachita Mountains in Arkansas are, respectively, a genetically related late Paleozoic cratonic dome (Ozarks), a foreland (foredeep) basin (Arkoma basin), and a thrust-belt system (Ouachita Mountains) (Fig. 3). In the earlier Paleozoic, the dome and foreland basin areas were a broad continental shelf separated by a narrow slope from a deep basin to the south (Ouachita trough). A thrust-belt sequence developed late in Mississippian (?) time to the south in the basinal area, and in Pennsylvanian time the basinal facies were thrust over the shelf and slope facies (Fig. 3B). In the early stages of basin closure (by Atokan time), a successor foreland basin had developed on the older shelf and slope of the tectonically foundered margin. This area was the site of deep-water sedimentation until mid-Atokan time. The foreland basinal axis trended east-west and followed the older Paleozoic slope and continental rise. Paleocurrents were to the west in the deep-water facies (Briggs and Cline, 1967; Morris, 1974, among others). The axis of the basin shifted northward through the Pennsylvanian as the basin closed.

The nomenclature for the upper Mississippian and lower Pennsylvanian strata in Oklahoma and Arkansas is complex (Fig. 5). It is related to the dramatic facies changes from shelf to basin deposits and the thickening of shelfal units toward eastern Arkansas (Mississippi embayment) and toward the south (basinward). Facies changes include changes in lithologies from thin carbonates in the west to thick clastic sequences in the east and south in the Morrowan. The upper Mississippian (Chesterian) in Arkansas consists of, in ascending order, Hindsville (Batesville), Fayetteville, Pitkin, and Imo formations on the Ozark shelf, and most of the Stanley Group in the basin to the south. The shelf units are dominantly carbonates with shale interbeds (Handford, 1986), and the deep-water rocks are mainly shale-rich clastics with some interbedded volcanic tuffs and tuffaceous sandstones (Niem, 1976).

The youngest Paleozoic strata that crop out in Arkansas are early to middle Pennsylvanian (Morrowan, Atokan, and Desmoinesian) in age. On the Ozark shelf in northeast Oklahoma, the

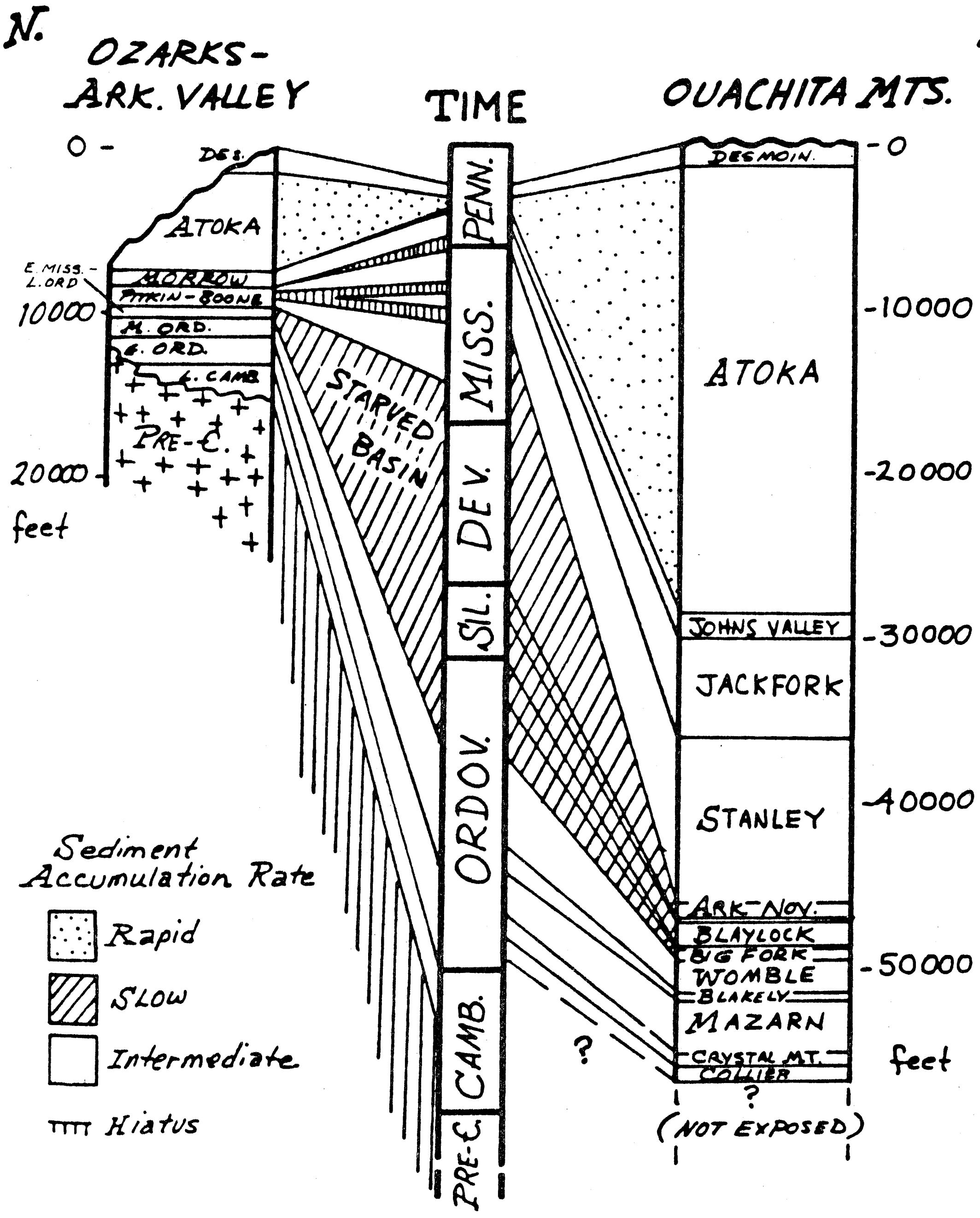


Figure 4. Generalized north to south stratigraphic sections between the Ozarks/Arkansas Valley (the shelf) and the Ouachita Mountains (the basin). Relative thickness and sediment accumulation rate are shown to depict (1) a period of starved basin sedimentation from the Late Ordovician to Early Mississippian, and (2) a period of rapid sedimentation in the Pennsylvanian.

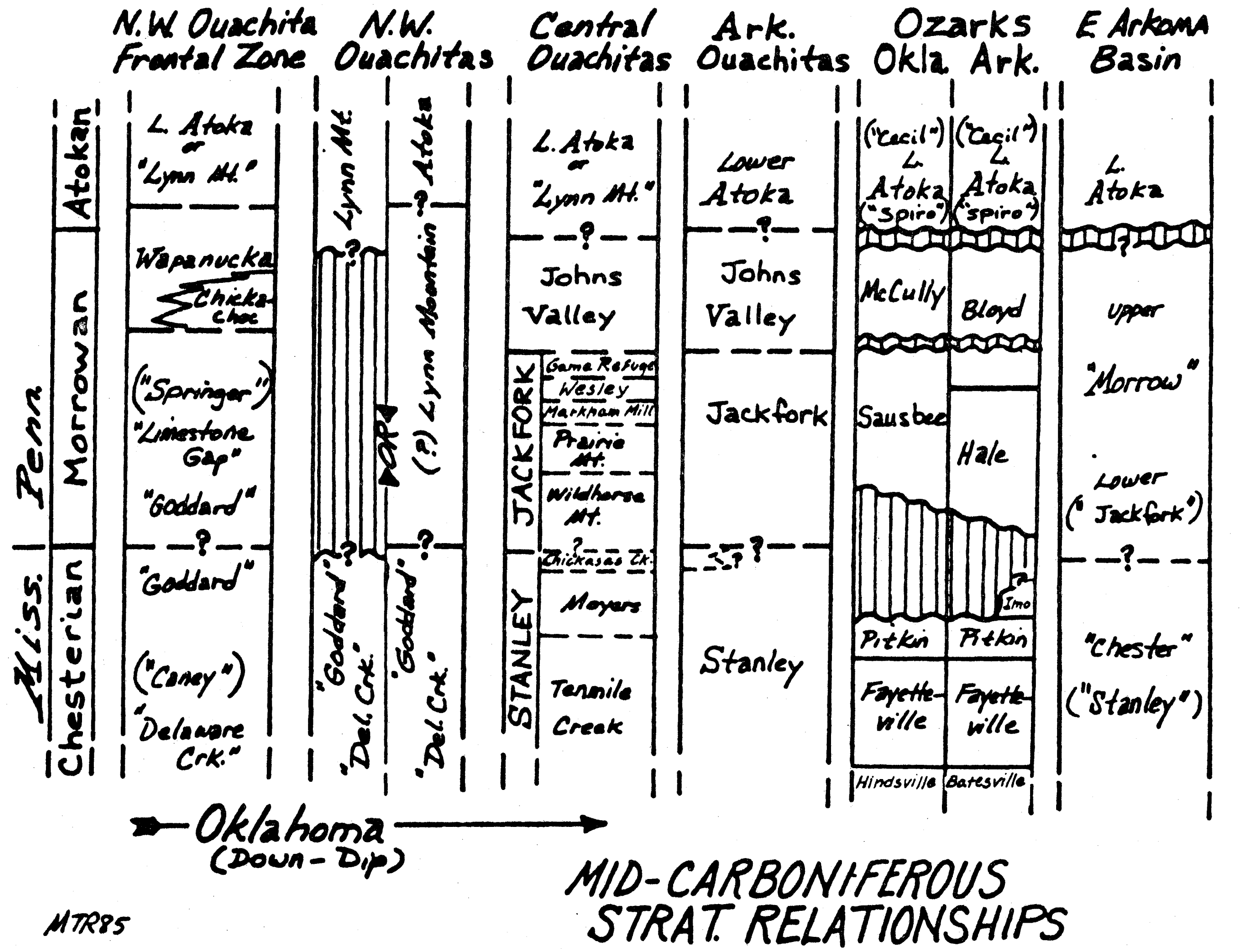
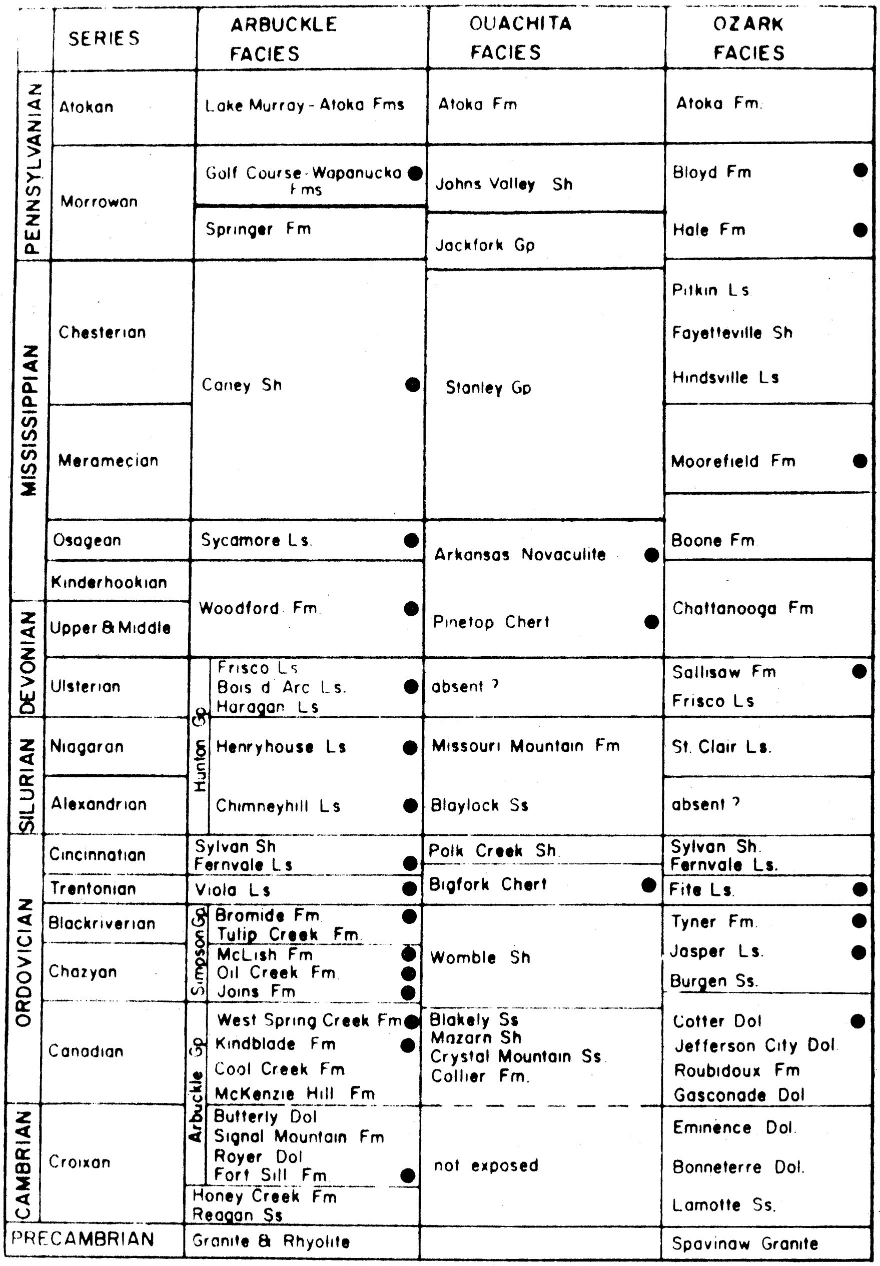


Figure 5. Stratigraphic nomenclature for mid-Carboniferous strata in the Ozarks, Arkoma basin, and Ouachita Mountains. Vertical line pattern denotes erosional gap or depositional hiatus, taken in part from Sutherland and Manger, 1979.

Morrowan units are thin, dominantly carbonate or mixed carbonate/clastic sequences, and are known as the Sausbee and McCully formations. They are unconformably overlain by fluvial-deltaic clastics of the Atoka Formation. The Oklahoma Morrowan units grade laterally eastward into the Hale and Bloyd formations in northwest Arkansas. Morrowan facies thicken toward east-central Arkansas and to the south, where correlation of upper and lower Morrowan units with the Hale and Bloyd subdivisions or units becomes difficult. The Morrowan strata on the northern shelf range in thickness from less than 200 feet in northeastern Oklahoma to over 2,000 feet in east-central Arkansas. Basinward to the south, these units are known as the Jackfork Formation, Sandstone, or Group and the Johns Valley Shale. The Jackfork is an alternating quartzose sandstone and shale sequence about 6,000 feet thick, which has been informally subdivided into five units in the central Ouachitas (Wildhorse Mountain, Prairie Mountain, Markham Mill, Wesley, and Game Refuge, in ascending stratigraphic order) (modified after Briggs and Roeder, 1978). The Johns Valley Shale is a shale sequence up to 1,500 feet thick that contains exotic clasts from many older Paleozoic units representing three major facies (Fig. 6, 7). Some sandstone interbeds are also present. It is found only in the basinal facies overlying the Jackfork. In southern Oklahoma, Jackfork-equivalent shelf and slope units are the upper Goddard(?), Limestone Gap, or "Springer" sequences, which are dominantly shales. Johns Valley Shale is equivalent to the Wapanucka Formation. The



	SERIES	ARBUCKLE FACIES	OUACHITA FACIES	OZARK FACIES
PENNSYLVANIAN	Atokan	Lake Murray - Atoka Fms	Atoka Fm	Atoka Fm.
	Morrowan	Golf Course-Wapanucka Fms ● Springer Fm	Johns Valley Sh Jackfork Gp	Bloyd Fm ● Hale Fm ●
MISSISSIPPIAN	Chesterian	Caney Sh ●	Stanley Gp	Pitkin Ls Fayetteville Sh Hindsville Ls
	Meramecian			Moorefield Fm ●
	Osagean	Sycamore Ls. ●	Arkansas Novaculite ●	Boone Fm
	Kinderhookian	Woodford Fm ●	Pinetop Chert ●	Chattanooga Fm
DEVONIAN	Upper & Middle			
	Ulsterian	Hunton Gp Frisco Ls Bois d Arc Ls. ● Haragan Ls	absent ?	Sallisaw Fm ● Frisco Ls
SILURIAN	Niagaran	Henryhouse Ls ●	Missouri Mountain Fm	St. Clair Ls.
	Alexandrian	Chimneyhill Ls ●	Blaylock Ss	absent ?
ORDOVICIAN	Cincinnatian	Sylvan Sh Fernvale Ls ●	Polk Creek Sh	Sylvan Sh. Fernvale Ls.
	Trentonian	Viola Ls ●	Bigfork Chert ●	Fite Ls. ●
	Blackriverian	Simpson Gp Bromide Fm ● Tulip Creek Fm.	Womble Sh	Tyner Fm. ●
	Chazyan	McLish Fm ● Oil Creek Fm. ● Joins Fm ●		Jasper Ls. ● Burgen Ss.
	Canadian	Arbuckle Gp West Spring Creek Fm ● Kindblade Fm ● Cool Creek Fm McKenzie Hill Fm	Blakely Ss Mazarn Sh Crystal Mountain Ss Collier Fm.	Cotter Dol ● Jefferson City Dol Roubidoux Fm Gasconade Dol
CAMBRIAN	Croixan	Butterly Dol Signal Mountain Fm Royer Dol Fort Sill Fm ● Honey Creek Fm Reagan Ss	not exposed	Eminence Dol. Bonneterre Dol. Lamotte Ss.
PRECAMBRIAN		Granite & Rhyolite		Spavinaw Granite

Figure 6. Stratigraphic chart showing formations from which exotic clasts (solid circles) have been found in the Johns Valley Shale. From Stone et al. (1979) after Shideler (1970).

Wapanucka is a shelf-edge limestone deposit and its Chickachoc chert member is more of a shaly slope sequence.

The lower Atoka Formation overlies Morrowan units in Oklahoma and northwest Arkansas. Lower Atoka units are variously called the Spiro Sand, Lynn Mountain Formation, and Trace Creek Shale; in most places, the Atoka is subdivided into lower, middle, and upper members (Stone, 1968). The lower Atoka consists of thin fluvial-deltaic facies in the north and thick turbidite facies in the south. By upper Atoka time, fluvial-deltaic deposition dominated all areas of Atoka accumulation. In Oklahoma and Arkansas, the subsurface nomenclature for the various units of the Atoka Formation is very complex and changes from one township to another (see Branan, 1968, for example). The Atoka ranges from a few hundred to over 20,000

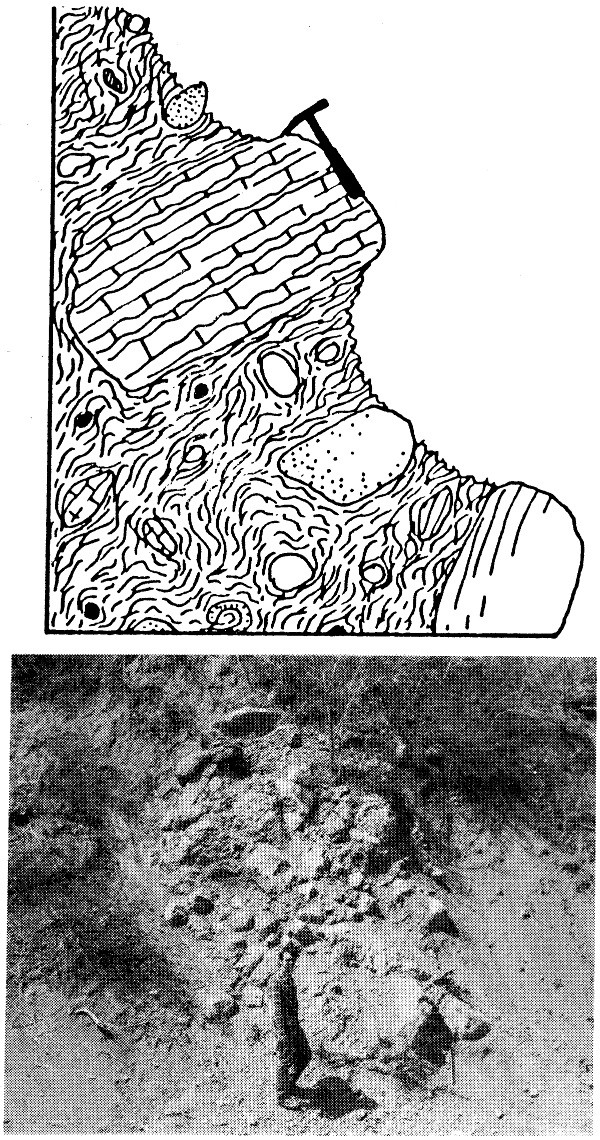


Figure 7. Sketch and photograph of two outcrops of the Johns Valley Shale near Stapp, Oklahoma. Disorganized shale fabric and matrix-supported exotic clasts are typical of this olistostromal facies.

feet in thickness. Desmoinesian strata overlie the Atoka in Oklahoma and in the western part of Arkansas, and include from older to younger, the Hartshorne, McAlester, Savanna, and Boggy formations. These are coal-bearing fluvial, coastal plain, and marginal marine deposits totaling about 5,000 feet in thickness.

PENNSYLVANIAN PALEOGEOGRAPHY

The Pennsylvanian paleogeographic reconstructions for east-central Arkansas presented here involve mainly three time slices: the Morrowan, Atokan, and Desmoinesian intervals. The Desmoinesian section, for the most part, is poorly preserved here. It is better preserved in synclines in parts of western Arkansas and eastern Oklahoma. The Morrowan and Atokan intervals are similar in their facies associations and both show the evolution of the gradual closing of a foreland basin, culminating in the uplift of this area in late or post-Desmoinesian time.

Morrowan

During the Morrowan on the Ozark shelf, shallow-marine carbonates were deposited in eastern Oklahoma and quartzose fluvial-deltaic clastic facies were deposited to the east in northern Arkansas. Coeval slope and deep-water quartzose clastic facies accumulated in the Ouachita basin to the south (Fig. 8). The Morrowan sediments expanded from a few hundred to thousands of feet in thickness from north to south and northwest to southeast(?), aided by growth faults along the shelf margin and slope. Seismic reflection data, as interpreted by the authors, show that these growth faults were localized over older basement-involved normal fault blocks and flatten above the early Mississippian or older platform units. Submarine landslides and related debris flows occurred on the slope and the resultant olistostrome deposits accumulated in the basin and the lower part of the slope to form parts of the Jackfork and Johns Valley formations.

Figures 9A and 9B illustrate the relatively thin (under 200 feet thick) Morrowan carbonate and mixed carbonate/clastic sequences (McCully and Sausbee formations) from the northeast Oklahoma shelf. Time-equivalent clastic units (Bloyd and Hale formations in northwest Arkansas) interfinger with the carbonates to the west. The carbonate rocks represent a shallow-marine, perhaps Bahama-like carbonate shelf facies, and their distribution and facies relationships are discussed by Sutherland and Henry (1977) and Sutherland and Manger (1979). To the south and southwest, the Oklahoma equivalents are the "Springer" slope facies and overlying Wapanucka shelfal limestone facies, which crop out in the Arbuckle Mountains and frontal Ouachita thrust-belt.

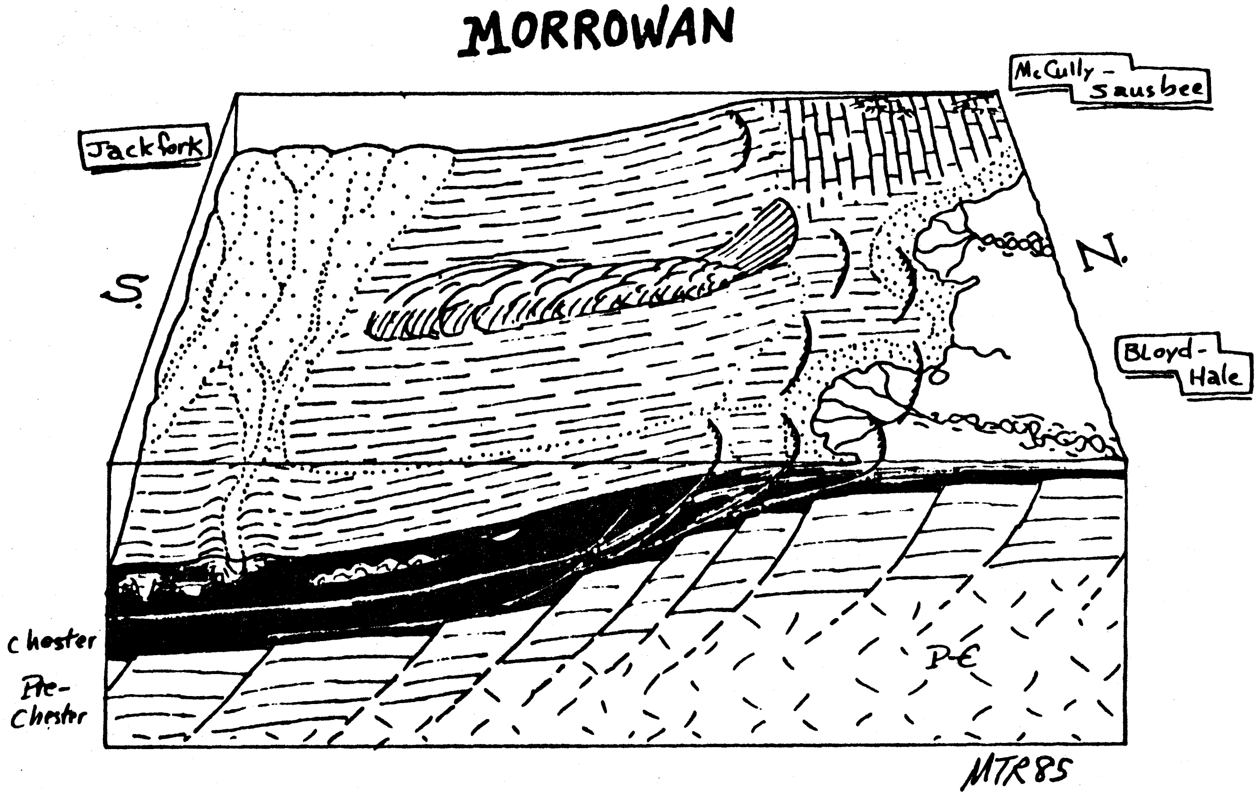


Figure 8. Morrowan paleogeographic sketch showing fluvial, deltaic, shallow-marine, and shelf facies to the north and slope, submarine fan, and basin facies to the south. Note the position of growth or listric faults localized at the outer shelf over pre-existing rift features. On the shelf, the western carbonate units in Oklahoma are the McCully and Sausbee formations in Arkansas, which are equivalent in age to the eastern clastic facies of the Bloyd and Hale formations. In the basin, the coeval units are the Jackfork and Johns Valley formations (not shown).

Eastward, in Arkansas, the dominantly clastic Bloyd and Hale formations constitute the Morrowan (Figs. 9C-F). They expand in thickness from less than 200 feet up to 2,000 feet (Fig. 10) from west to east. The Bloyd Formation in northwestern Arkansas is interpreted to consist of paralic coastal plain, fluvial-deltaic, and transgressive marine units (Zachry, 1977). Paleocurrent measurements from the cross-bedded fluvial units suggest rivers flowing to the south-southwest. In east-central Arkansas, the Bloyd and Hale sections are much thicker and have not been differentiated completely. The facies are similar to those of western Arkansas but are thicker and contain better developed deltaic, delta plain, and sandy fluvial deposits (Figs. 9C and 9D). Major braided stream deposits are transitional southwestward into smaller and more numerous meandering streams, and eventually into distributary channels that fed prograding deltas. Major braided systems are recognized near Gaither Mountain and in the Concord-Almond area (Fig. 9C, E) in northern Arkansas. The deltas are shelf-depth, relatively thin (ca. 40 feet), and widespread in their distribution (Figs. 9D, 9F). For the most part, they are fluvial- to wave-dominated systems and are characterized by highly bioturbated prodelta facies and thin to poorly preserved foreshore facies, locally capped or cut into by river-mouth bar and/or distributary channel facies. Morrowan deltaic deposits are well exposed in quarries in White County, especially near Bald Knob and Judsonia (Fig. 9D).

The Morrowan slope facies, for the most part, is not well exposed in the area, although the Jackfork units of the frontal thrusts in the Ouachitas may contain lower slope facies (for example, near Lake Maumelle). Possible slope facies are poorly exposed in the core of Bayou Meto anticline (C. G. Stone, 1985, personal commun.). The deep-marine Jackfork Formation and Johns Valley Shale have been more

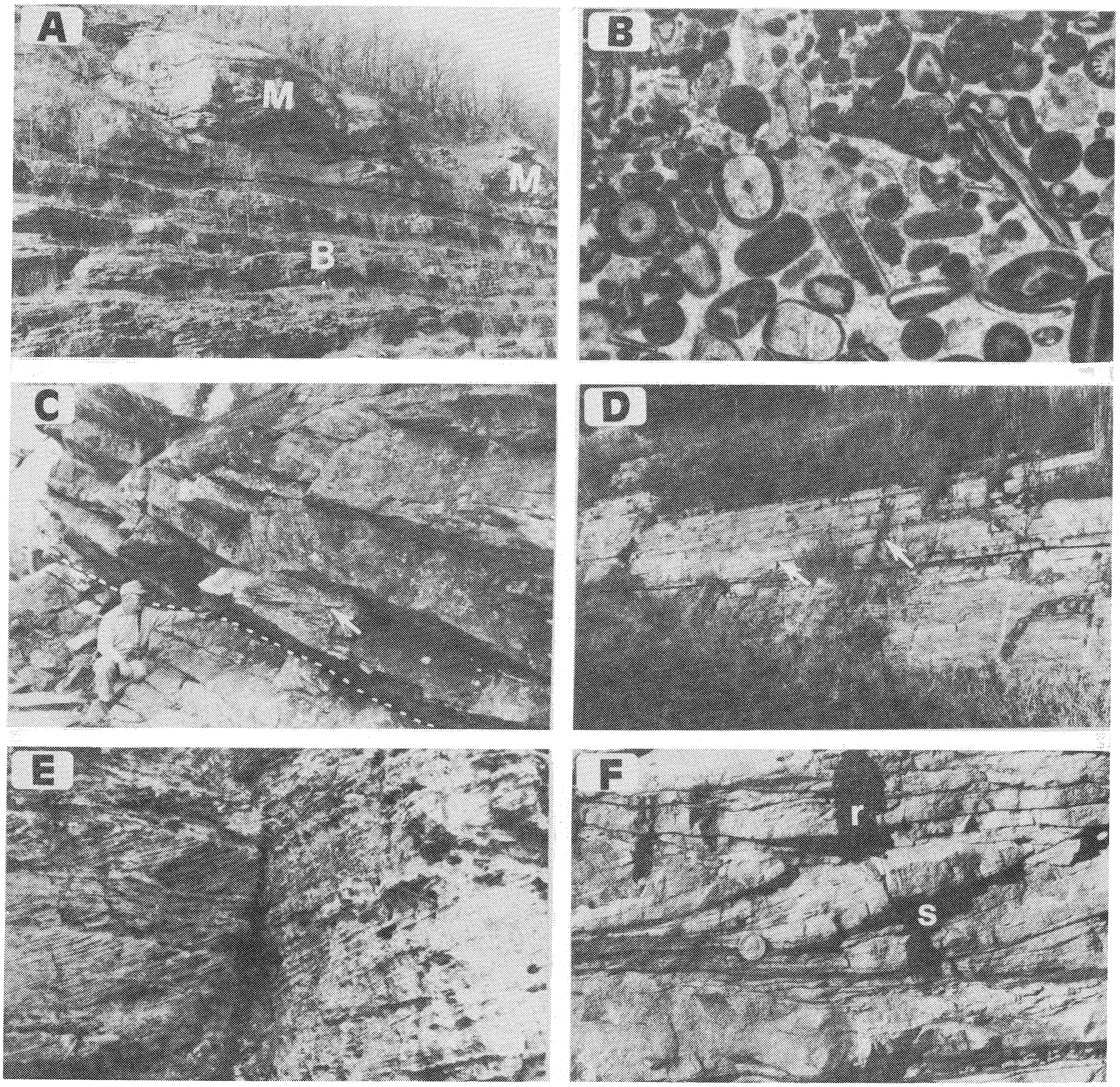


Figure 9. Photographs of Morrowan shelfal carbonates, fluvial and shallow-marine clastic rocks. A) Sausbee Formation at Webber Falls, Oklahoma; B--Briggs, and M--Brewer's Bend members. Note the large carbonate mounds (M) in the Brewer's Bend member. B) Photomicrograph of a carbonate-cemented oolite grainstone with abundant bryozoan and crinoid fragments; Sausbee Formation at Bragg's Mountain. C) Morrowan meandering fluvial deposits incised into marine units at Salado Creek, Arkansas. Note laterally accreting sigmoidal cross-bedded units (arrows) at the base of the channel fill. D) Morrowan foreshore (beach?) accretionary bedding (arrows) along Highway 67 near Judsonia, Arkansas. E) Unimodal, directional, planar-laminated cross-bedding in Morrowan braided stream deposit (view about 5 feet high), near Almond, Arkansas. F) Shallow-marine, tidal, sigmoidal (s) bedforms overlain by ripple-laminated beds (r); along Highway 67 near Judsonia.

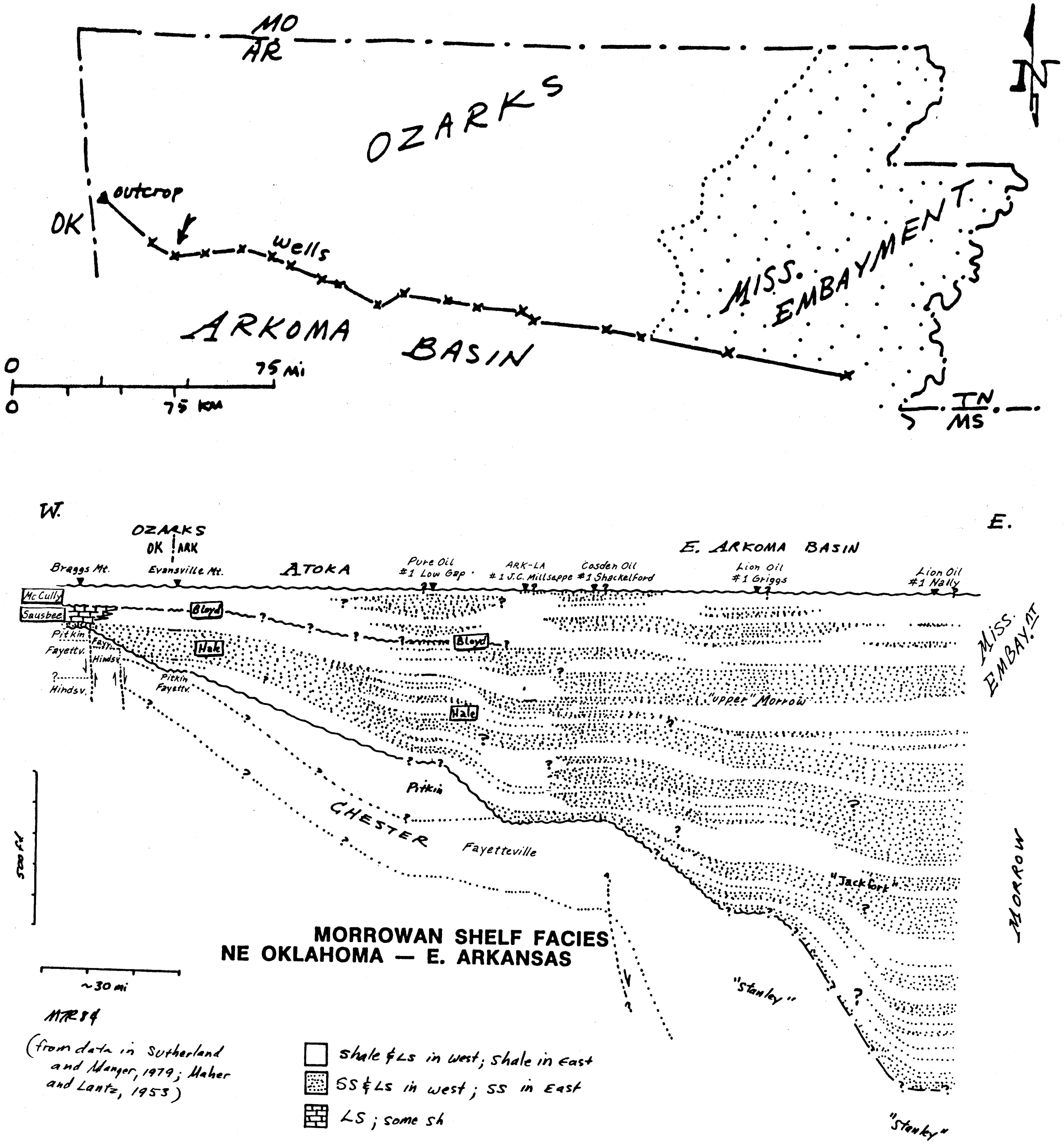
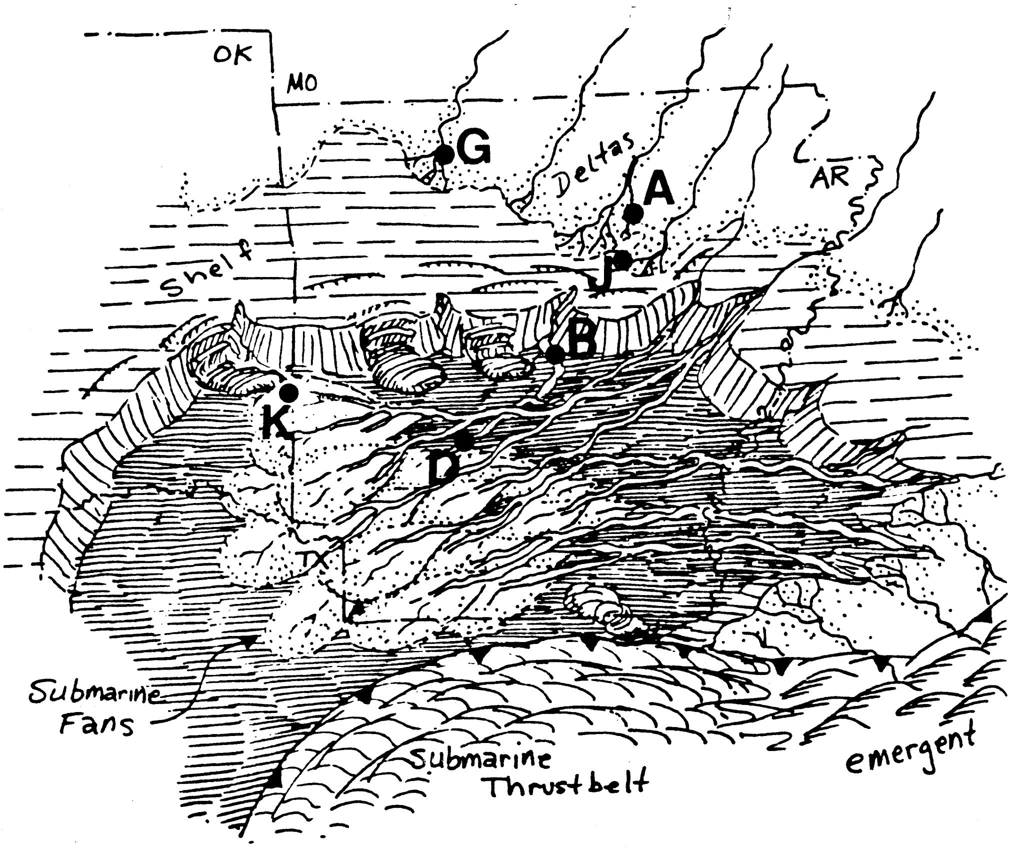


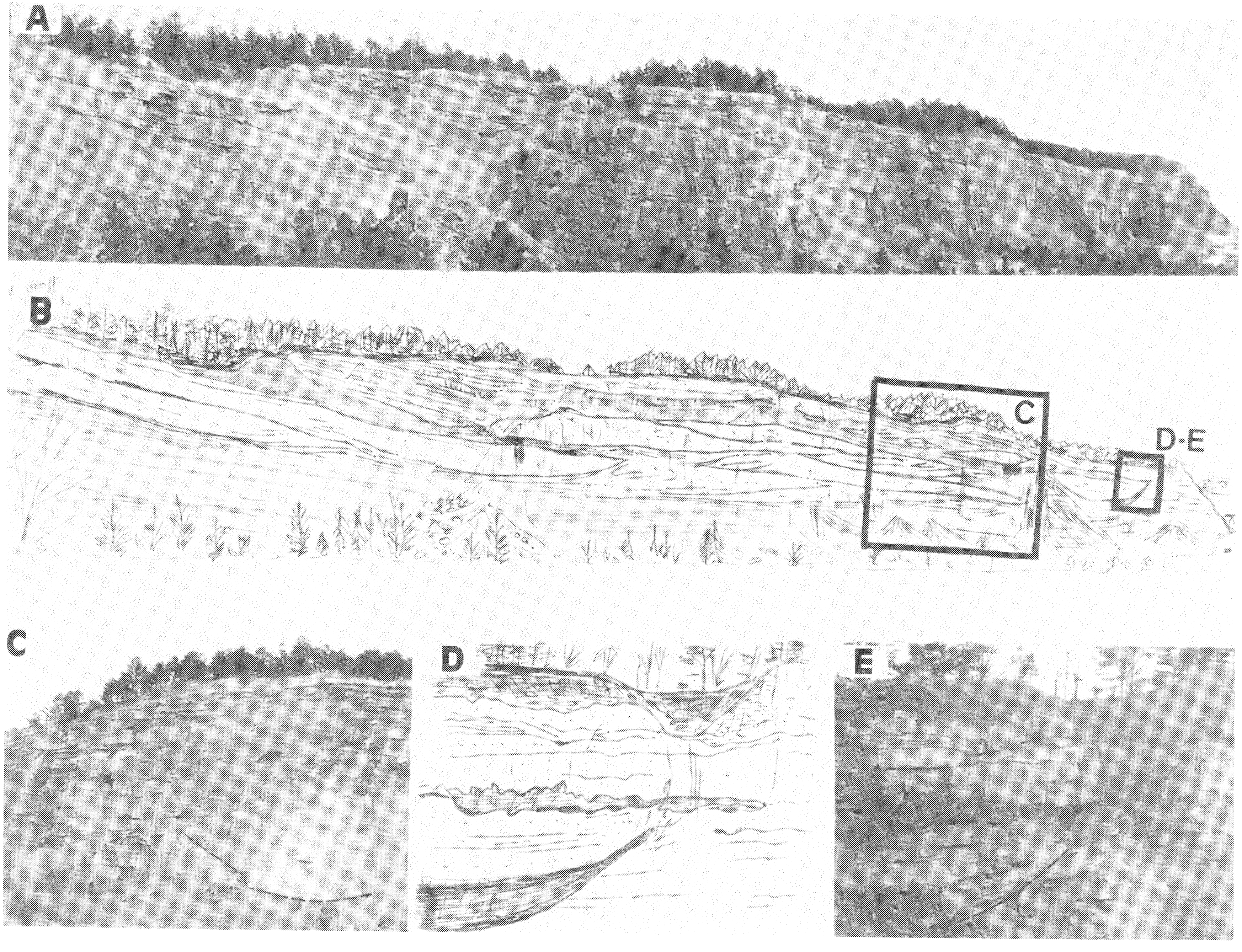
Figure 10. An east-west section showing the stratigraphic relations for the Morrowan. Note the facies change from carbonates in the west to thick clastic sequences to the east. Constructed from data in Sutherland and Manger (1979), and Maher and Lantz (1953). Line of section shown on sketch map above.



DIAGRAMMATIC PALEOGEOGRAPHY—JACKFORK

Figure 11. Diagrammatic map showing paleogeography during deposition of the Jackfork Formation. Abbreviations: B--Big Rock Quarry in North Little Rock; D--DeGray Dam/Friendship area; G--Gather Mountain; J--Judsonia; and A--Almond. Note that the Jackfork fan system consists of submarine canyons and inner fan channels to the northeast near Little Rock, middle fan-channel complexes in the DeGray Dam-Friendship area, and outer fan lobes at Kiamichi Mountain in Oklahoma.

Figure 12. Photographs and sketches of submarine canyon or inner fan-channel complex in the Jackfork Formation, Big Rock Quarry, North Little Rock. A) North to south view of the east face of the Big Rock Quarry. B) Sketch of Figure 12A, noting the lenticular, laterally aggrading channel-fill sandstone deposits and the more shaly upper part. Boxes denote position of Figures 10C-E. C) Photograph showing some of the major angular discordances between channel and the channel-margin deposits where levees are inclined away from the channels. D) Sketch of Figure 12E showing the side of a channel-fill where strata are onlapping the side of channel. E) Photograph of Figure 12D showing the side of the channel outlined.


A
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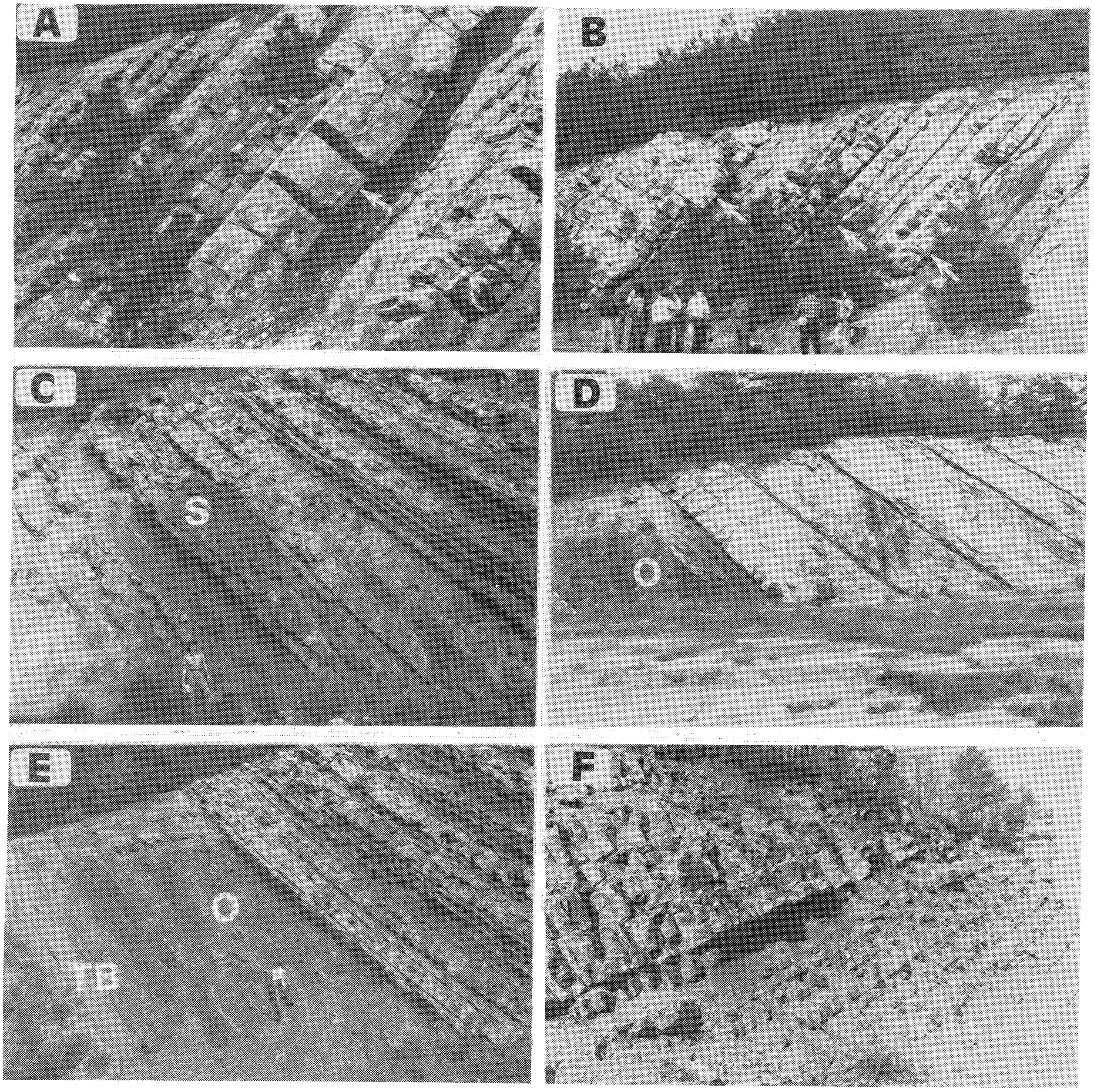


Figure 13. Photographs of submarine fan facies in the Morrowan Jackfork Formation. A) and B) Channel-fill or channel-lobe transition deposits which thin and fine upward to left; note the erosive basal contacts (arrows) of successive packages; Friendship area along U.S. Interstate Highway 30. C) Channel-fill deposits (C) overlain by a slumped unit (S), which is in turn onlapped and overlain by laterally continuous cravasse splay lobe(?) deposits; DeGray Dam. D) Uppermost major channel-fill deposit at DeGray Dam. Note the olistostrome (O) unit at the base (--) of the channel whose top is to the right. E) Thin-bedded (TB) turbidites overlain by an olistostrome (O) unit which is in turn overlain by alternating lobe(?) or crevasse splay beds; DeGray Dam. F) Depositional lobe which thickens and coarsens upward; Kiamichi Mountain along Highway 259.

extensively studied. The Jackfork Formation reaches over 6,000 feet in thickness and consists of alternating sandstone and shale packages interpreted to be submarine fan, basin plain, and slope facies (Fig. 11). The Jackfork contains about 30 to 60 percent sandstone with individual sand-rich packages being up to 1,000 feet thick and composed of 60 to 80 percent sandstone. Slope, submarine canyon, and inner fan deposits are recognized in the northeasternmost Jackfork outcrops in the Little Rock area. An excellent exposure of a submarine canyon fill or an inner fan-valley system occurs in Big Rock Quarry in North Little Rock (Fig. 12). Middle fan-channel complexes occur in central Arkansas in the Friendship-DeGray and Dierks Dam areas, and middle to outer fan lobes occur in western Arkansas and eastern Oklahoma, from near Mena (Rich Mountain) to Kiamichi Mountain (Thomson and LeBlanc, 1975; Moiola and Shanmugam, 1984) (Fig. 13). Paleocurrents are mainly to the west in these rocks. We envision a large submarine fan system with elongate lobes, a complex channel/lobe transition zone, and major submarine canyon/inner fan/slope. The sediments were deposited from east to west for the Jackfork Formation (Fig. 14). Similar elongate, deep-water, foreland basin assemblages have been described by Mutti (1985) in Spain.

The Johns Valley Shale is up to 1,500 feet thick and overlies the Jackfork Formation. It consists of shale with exotic clasts up to boulder size (olistoliths) and large slump masses that range from Cambrian to Morrowan in age (Figs. 6, 7) (Ulrich, 1927; Shideler, 1970; Stone et al., 1973; Stone and McFarland, 1981). Turbidite sandstone beds also occur in the Johns Valley Shale. The Johns Valley has been interpreted by many authors to represent submarine slump and landslide deposits derived mainly from northern and northwestern sources in Arbuckle and Ozark shelf facies. The source terrane for these clasts and slide masses, some of which are thousands of feet across, may have been large normal fault scarps along the northern and western basin boundaries produced during tectonic foundering of the basin margin by thrust-sheet loading to the south (Fig. 15). Seismic data in Oklahoma show huge normal fault scarps in the shelf facies, overidden by Ouachita facies thrusts. The facts that the coeval shelf-carbonate Wapanucka facies was deposited near sea level and the the Johns Valley Shale contains exotic clasts from the Wapanucka through the Cambrian section, strongly suggest that the Johns Valley was deposited near the base of a fault scarp or series of scarps with at least a mile of stratigraphic throw. Water depths are estimated to be over

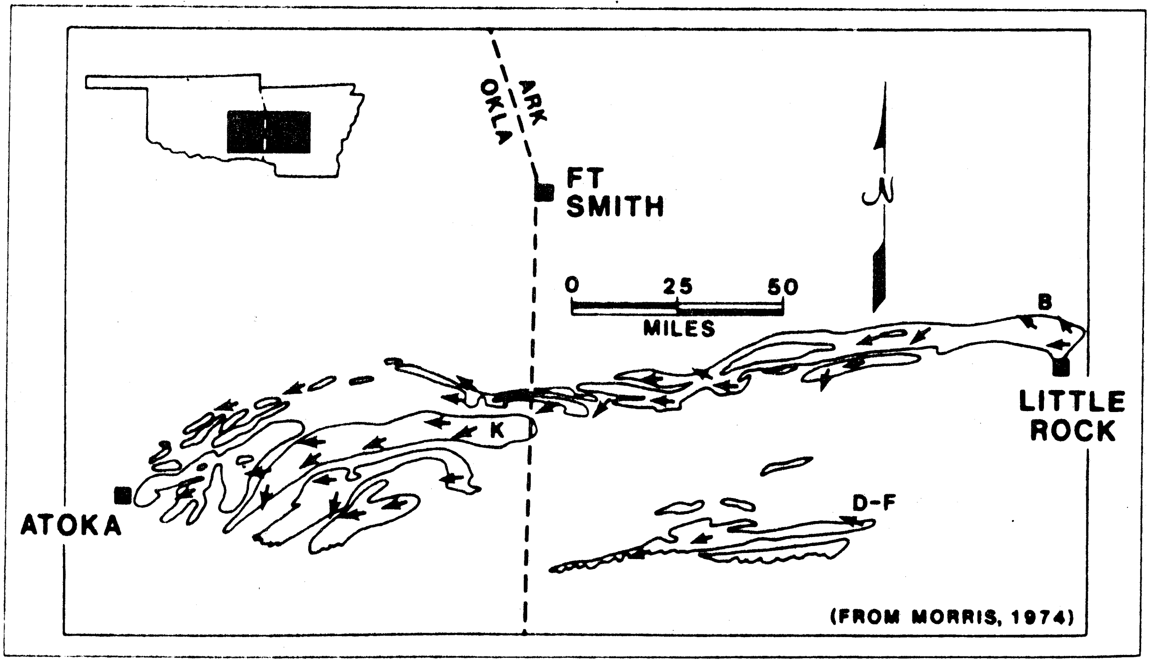


Figure 14. Map of paleocurrent directions in the Jackfork Formation. Note the pronounced westward trend.

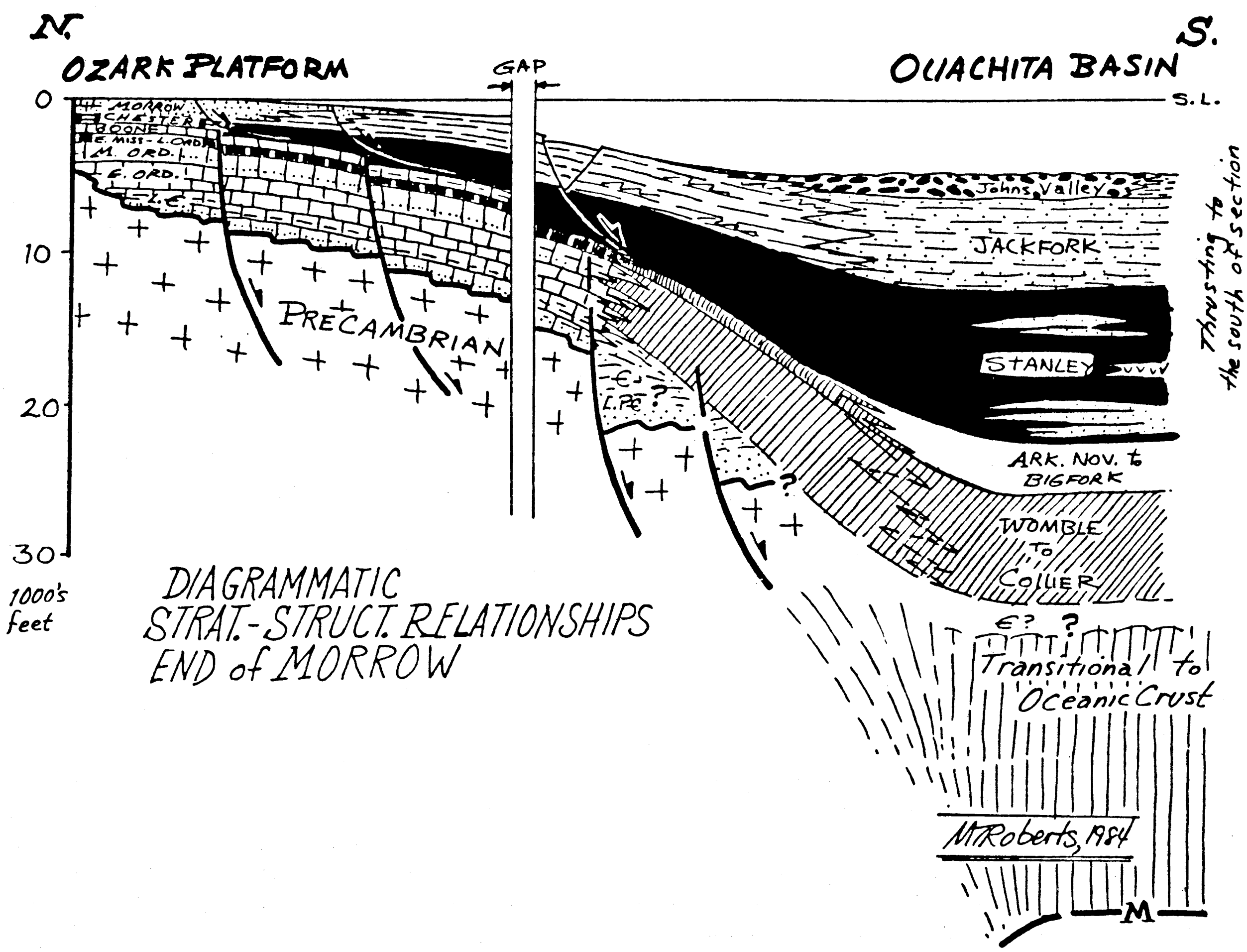


Figure 15. Diagrammatic north-south stratigraphic and structural cross section at the end of the Morrowan. Note the positions of the Ozark platform in relation to the Ouachita basin and the facies changes that occur in the shelf/slope/basin transition. The growth faults are localized over older rift structures and the Morrowan sections thicken on the down-thrown sides of the listric faults. The Cambrian to Mississippian shelfal units must have been exposed locally to erosion, perhaps in the gap indicated, because lower Paleozoic carbonates are found in the Johns Valley Shale.

5,000 feet, perhaps 7,000 feet, during Johns Valley Shale deposition.

Atokan

During the Atokan interval, the foreland basin narrowed and the axis migrated northward markedly due to the advancing thrust sheets from the south (Fig. 16). The basin shoaled through Atokan time due to sedimentary infilling. Fluvial-deltaic and shallow-marine facies were deposited on the shelf to the north, and slope and deep-water facies accumulated in the basin to the south in early to middle Atokan time. By late Atokan time, only shallow-marine and fluvial-deltaic facies were deposited in the basin. Relatively high angle growth faults, which become flatter above the Cambrian-Mississippian platform rocks, are common near the shelf-slope break. In post-Atokan time, some of these listric faults were reutilized as thrusts during the culmination of the Ouachita orogeny.

The Atoka Formation is about 27,000 feet thick and consists of shale, sandstone, siltstone, and some coal. The Atoka is informally subdivided into three mappable (lower, middle and upper) members (Stone, 1968). On the shelf in the northern Arkoma basin, Atokan fluvial-deltaic and shallow-marine facies include wave-, river-, and tidally-influenced deltas, distributary channel, and river-mouth bar deposits, transgressive

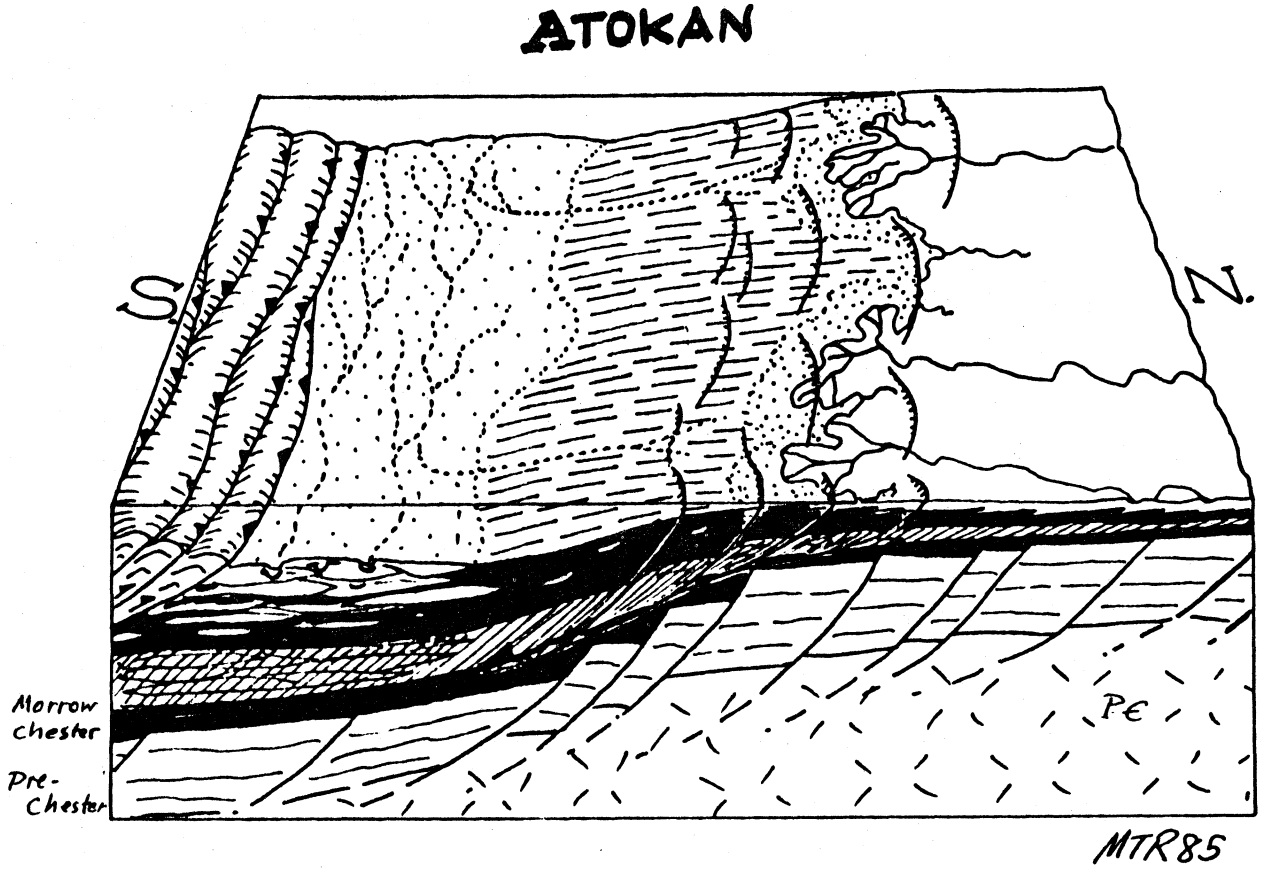


Figure 16. Paleogeographic sketch for the Atokan showing fluvial, deltaic, shallow-marine, and shelf facies to the north, and slope, submarine fan, and basin facies to the south. The advancing thrust sheets from the south can be seen restricting and narrowing the elongate basin. Growth faults are depicted at the outer shelf edge with local development of shelf edge deltas. Sediment transport is to the south/southwest on the shelf and to the west in the basin.

marine shelfal units, deltaic coastal-plain deposits, coal, and meandering stream sequences (Stone, 1968; O'Donnell, 1983) (Figs. 17A-17F). The coal and transgressive marine units (Fig. 17A) are relatively thin and locally widespread. The tidal influence on the deltas apparently became more pronounced as the basin constricted in late Atokan time, as tidal bedforms are more abundant in these deposits (Fig. 17E).

To the south (basinward), the lower Atoka and part of the middle Atoka are composed of sandstone-rich, deep-water deposits, similar to the Jackfork facies, but more than twice as thick. The deep-water facies were confined largely to an elongate east-trending trough that formed the southern part of the Arkoma basin. Paleocurrents were mainly to the west (Morris, 1974) in this basin. Deposition occurred during active closing of the basin by thrusts advancing from the south.

The Atokan interval represents a period of very rapid sedimentation in the foreland basin (Fig. 18). This high rate of sediment accumulation is interpreted to be due to the closing and narrowing of the basin that confined the sediment and increased the volumes of clastics contributed from rising source areas to the east in the Appalachian region.

Desmoinesian

In Desmoinesian time, the northward advancing thrust sheets and increasing influx of sediments had nearly closed the foreland basin, shifting the depo-center to western Arkansas and Oklahoma. The Ozark dome was reactivated as a positive lowlands feature to the north (Fig. 19).

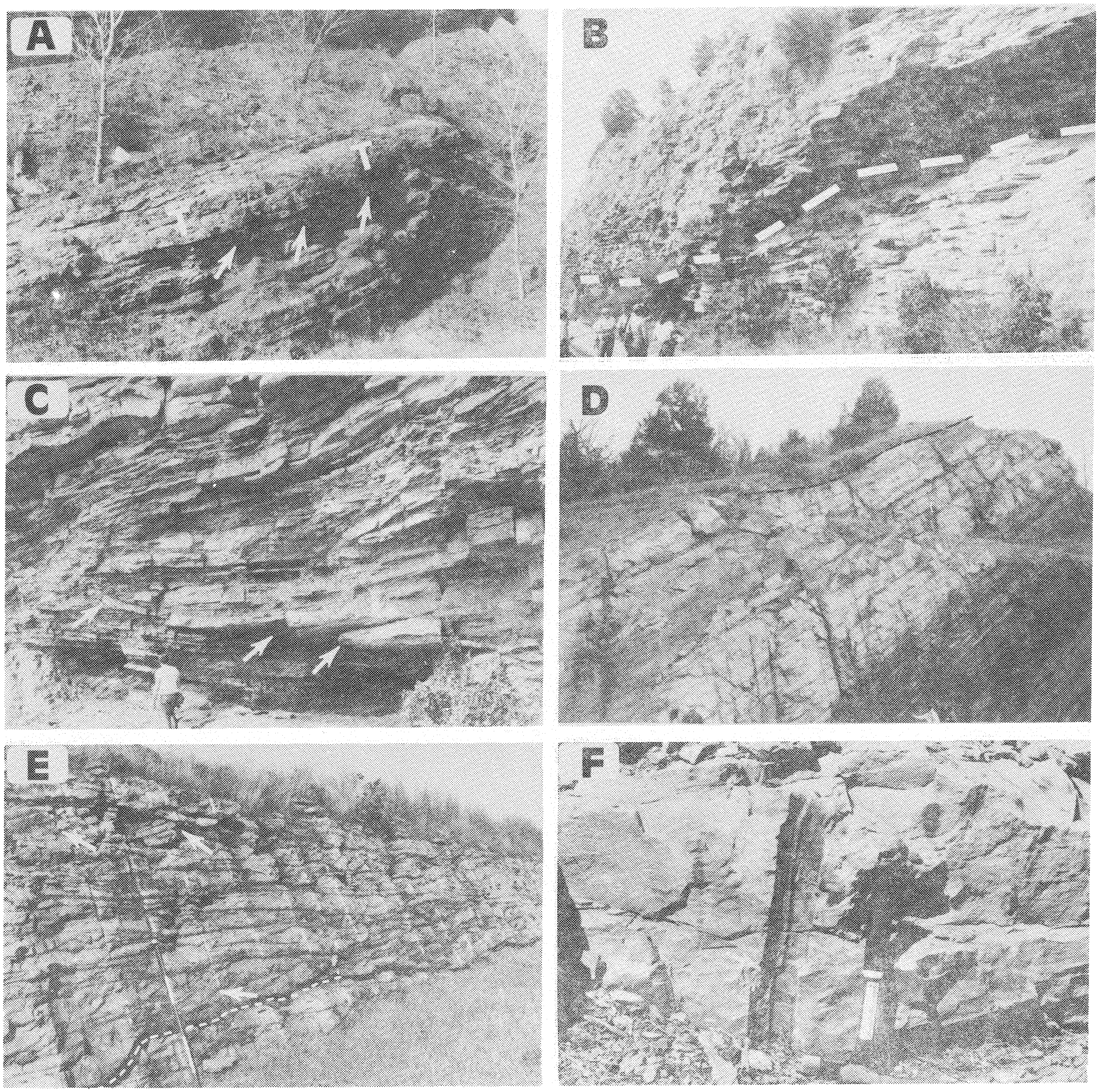
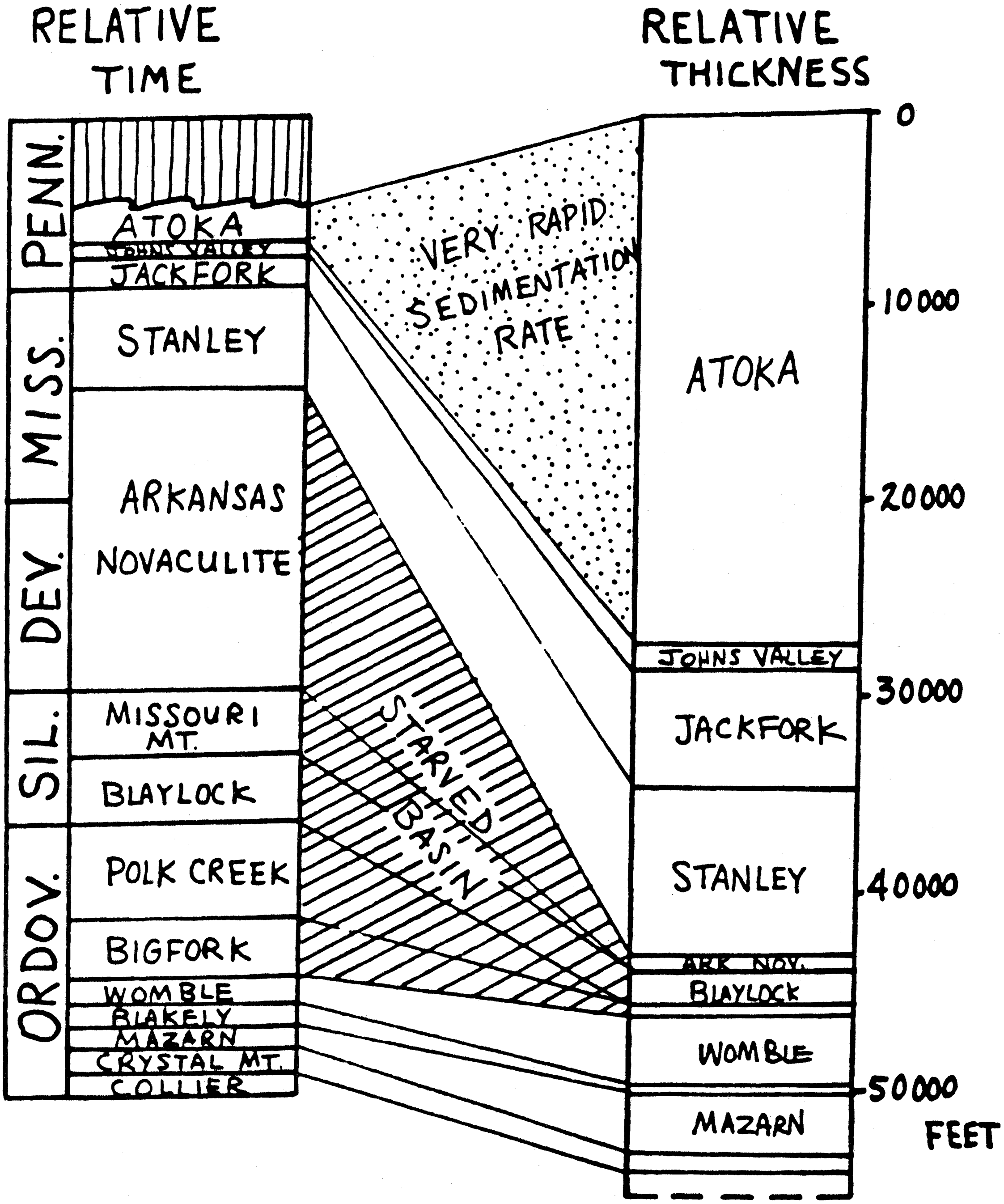


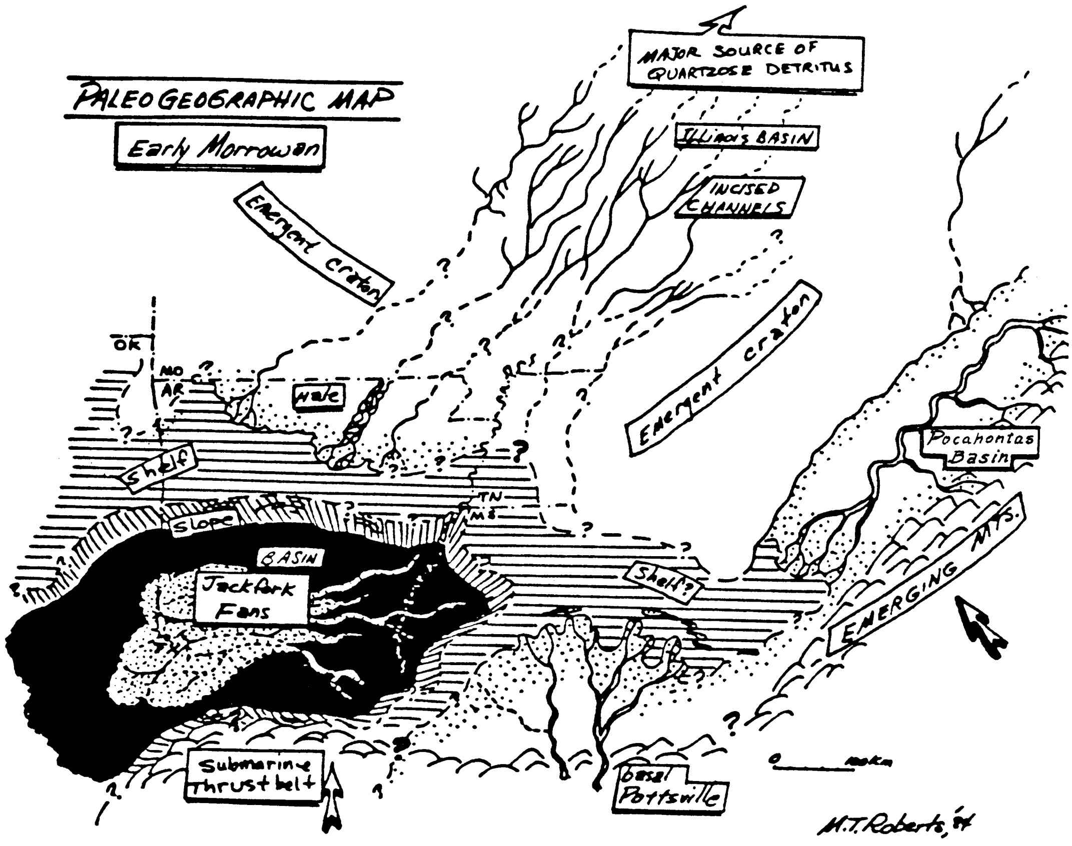
Figure 17. Photographs of Atokan and Desmoinesian outcrops. A) Atokan basal transgressive marine unit (T) overlain by pro-delta shales and underlain by well-developed karst topography (arrows) on Morrowan carbonates; Webber Falls, Oklahoma. B) and C) Large, delta distributary channel complex in the upper Atoka at Ozark, Arkansas. Note the erosional downcutting (outline) in Figure 17B and laterally accreting point bar surfaces (arrows) in Figure 17C. D) Thickening- and coarsening-upward deltaic sequence in the middle Atoka along Highway 9 near Morrilton, Arkansas. Note shale plug (abandoned shale-filled channel) at the top of the sequences (highlighted area). E) Tidally-influenced channel deposit along Highway 35 near Ozark, from the upper Atoka (scale = 1.5m). Note the scour base (highlighted) and sigmoidal bed forms (arrows). F) Mold of upright tree (<u>Calamites</u>) (arrows) in the Desmoinesian McAlester Formation along Highway 59 near Heavener, Arkansas. Several well-developed coal seams are present at this delta-plain outcrop.



PALEOZOIC UNITS
OUACHITA MTS., ARKANSAS

(MTR 80 from data in Stone et al., 1973; Haley, 1976)

Figure 18. Paleozoic stratigraphic section showing the relative thickness of units in the basinal part of the Ouachita basin as related to time. Note the starved-basin (reduced sedimentation) phase for the Ordovician Bigfork and Polk Creek, Silurian Blaylock and Missouri Mountain, and Devonian-Mississippian Arkansas Novaculite formations. Note also the period of very rapid sedimentation in Atoka time.


PALEO GEOGRAPHIC MAP
Early Morrowan
Emergent craton
MAJOR SOURCE OF QUARTZOSE DETRITUS
ILLINOIS BASIN
INCISED CHANNELS
Emergent craton
OK
MO
AR
Wale
Shelf
Slope
TN
MS
BASIN
Jackfork Fans
Shelf?
Pocahontas Basin
EMERGING MTS.
Submarine Thrust belt
basal Pottsville
0 100km
M.T.Roberts, '84

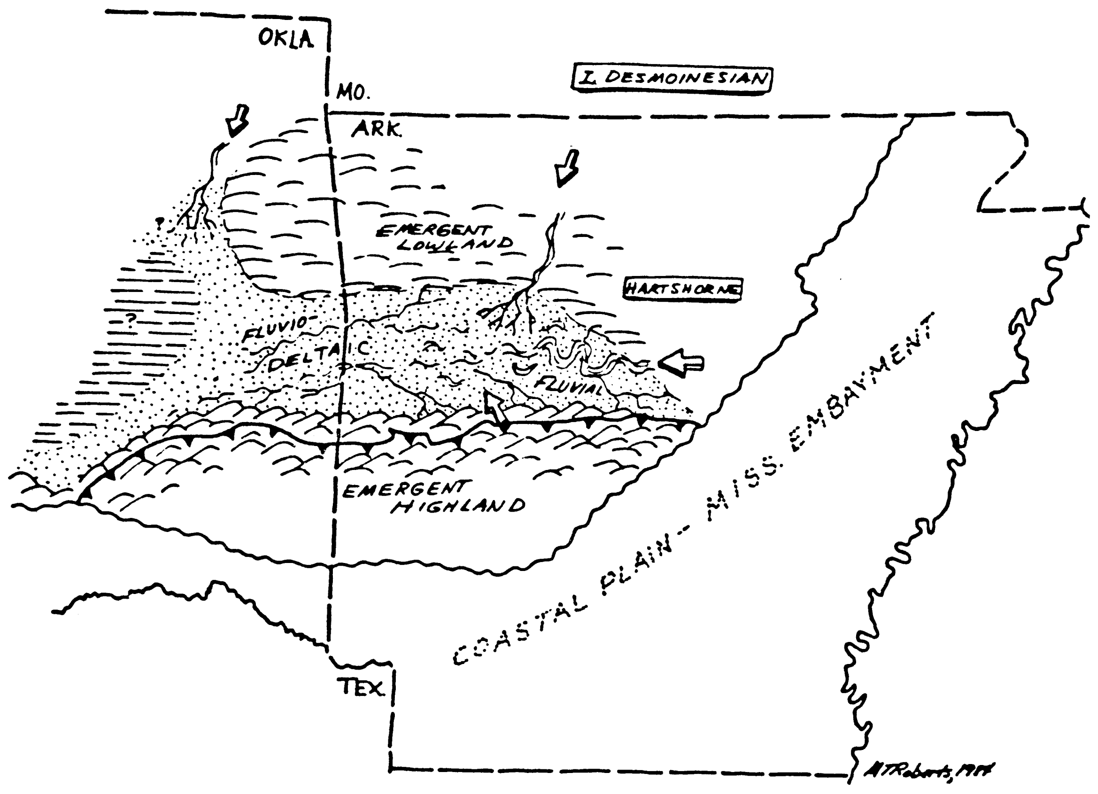

OKLA
MO.
ARK
I. DESMOINESIAN
EMERGENT LOWLAND
HARTSHORNE
FLUVIO-DELTAIC
FLUVIAL
EMERGENT HIGHLAND
COASTAL PLAIN - MISS. EMBAYMENT
TEX
M.T.Roberts, 1984

Figure 19. Paleogeographic maps showing probable source areas for the early Morrowan and late Desmoinesian rocks. The major source for quartzose sediment is inferred to have been to the northeast and derived from the Canadian shield and interior drainage basin of the Appalachian orogenic belt. In the Morrowan, sediments for the most part by-passed the Illinois basin and Reelfoot rift(?) and were deposited to the south in the Ouachita basin. Sources to the south (Llanoria) and east (Black Warrior basin) probably also contributed sediment to the basin. The strongest evidence for a southern source is in the volcanic-bearing sediments for the Stanley Formation. In late Desmoinesian, the Ouachita basin gradually closed. Uplands to the south and lowlands to the north caused the basin to be restricted and sediments to be funneled to the west.

The foreland basin closed diachronously, starting in the southern Appalachians and proceeding westward into Oklahoma. By the end of Desmoinesian time, the eastern part of the central Arkansas foredeep basin had been filled. The sea retreated westward into Oklahoma and west-flowing rivers across Arkansas fed deltas near the Oklahoma border (Houseknecht and Kacena, 1983). These rivers were confined by the Ozark dome (north) and the frontal Ouachitas (south). Desmoinesian sediments were folded during the middle Pennsylvanian-early Permian culmination of the Ouachita orogeny. The Permian record is partly preserved to the west in Oklahoma.

The preserved Desmoinesian (Hartshorne) section in central Arkansas consists mainly of fluvial deposits. It is transitional into coal-bearing fluvial-deltaic and shallow-marine units in western Arkansas and eastern Oklahoma (Houseknecht and Kacena, 1984; Houseknecht and Matthews, 1985) (Fig. 17F). Middle Pennsylvanian to Permian rocks are not preserved in east-central Arkansas and, presumably, were never deposited or were eroded away during the orogenic mountain-building events caused by continental collision. These same strata are locally well preserved in Oklahoma and Kansas, where these areas contain cratonal, shallow subsiding basins that are influenced by wrench fault structures such as the Wichita-Arbuckle, Criner and Nemaha trends.

Provenance

The foreland basin developed over a downwarped shelf or platform sequence of Cambrian to Mississippian rocks dominated by carbonate deposition. In contrast, the clastic sediments of the foreland basin are mainly fine- to medium-grained quartzose (>90% quartz) sandstone and shale derived externally. Paleocurrent and petrographic data show that during the basin's history, quartzose sediment was derived from the north, northeast, and east across what are now the Illinois basin, Ozark dome, Reelfoot rift, Nashville dome, and Black Warrior basin area. The major source areas for this sediment were the Canadian Shield and the eastern U.S. interior drainage basins of the Appalachian orogenic belt (Fig. 19). Clastic sediments largely bypassed the interior cratonal areas, normally shallow seas, during low sea level stands and were deposited in the foreland basin from the southern Appalachian Mountain area. The Ozark and Nashville domes rose during final stages of the orogeny, isolating the Illinois basin from the Arkoma foreland basin.

With continental collision, the various foreland basins that flanked the late Paleozoic "Atlantic" and "Gulf Coast" margins began to close from the east and south. Beginning in late Mississippian (Stanley time), the Ouachita basin began to close south of east-central Arkansas and some south-derived clastic and volcanic material occur in the Stanley Shale (Niem, 1976). Some of the Pennsylvanian sediments were also likely derived from the east from the rising southern Appalachians via the Black Warrior basin and perhaps from the southeast from the postulated land mass of Llanoria (Graham et al., 1975; Mack et al., 1981; Mack et al., 1983; Owen, 1984; Owen and Carozzi, 1986).

Starting in the middle Pennsylvanian to early Permian, the southern margin of the foreland basin was folded, uplifted, and subjected to denudation. Reworked older Paleozoic strata, mainly chert conglomerates and lithic intraclasts, were shed to the north and northwest (Houseknecht and Kacena, 1983; Sutherland, in preparation) into the gradually closing basin.

The Ouachita thrust-belt segment of central Arkansas and southeastern Oklahoma was probably submarine until the late stages of the orogeny.

SUMMARY AND CONCLUSIONS

1. The Pennsylvanian strata in east-central Arkansas are part of a late Paleozoic foreland basin system that originally extended from the southern Appalachians to west Texas.

2. Early Pennsylvanian Morrowan and Atokan rocks of east-central Arkansas are composed of quartzose sandstone and shale derived mainly from the northeast from the North American craton and Appalachian orogenic belt. These clastics buried the Cambrian to Mississippian carbonate-dominated platform rocks, the last remnant of which were the Morrowan carbonate rocks of northeast Oklahoma.

3. The clastic sedimentary facies that make up the Morrowan and Atokan intervals are similar in type and paleogeographic distribution. Fluvial, deltaic, shallow-marine, and shelf facies occur on the northern flank of the basin, while slope, submarine fan, and basin plain facies are present to the south. Sediment transport on the shelf was mainly to the southwest, and in the basinal area it was to the west. Thicknesses increase drastically from north to south, aided by listric growth faults along the shelf edge and slope.

4. In Atokan through early Desmoinesian times, the foreland basin gradually closed by advancing thrust sheets from the south. Sediments accumulated in the basin at extremely high rates. By late Atokan time, fluvial, deltaic, shallow-marine, paralic, and transgressive units were deposited throughout this shoaling basin. Ouachita uplands to the south and the Ozark dome to the north restricted sedimentation in the narrowing basin by Desmoinesian time and funneled sediments to the west.

5. Continental collision culminated in the post-Desmoinesian (middle Pennsylvanian to early Permian) uplift of the area, followed by erosion and/or nondeposition.

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REFERENCES

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

Branan, C. B., Jr., 1968, Natural gas in Arkoma Basin of Oklahoma and Arkansas, in Beebe, B. W., ed., Natural gas of North America, American Association of Petrol. Geol. Memoir 9, p. 1616-1635.

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

Briggs, G., and D. Roeder, 1978, Sedimentation and plate tectonics, Ouachita Mountains and Arkoma Basin, in Briggs, G. E., E. F. McBride, and R. J. Moiola, eds., Field trip guidebook to the sedimentology of Paleozoic flysch and associated deposits, Ouachita Mountains - Arkoma basin, Oklahoma: Dallas Geol. Soc. Guidebook, p. 1-22.

Chamberlain, C. D., ed., 1978, A guidebook to the trace fossils and paleoecology of the Ouachita geosyncline: SEPM, Tulsa, Oklahoma, 68 p.

Cline, L. M., 1966, Late Paleozoic rocks of the Ouachita Mountains, a flysch facies, in Flysch facies and structure of the Ouachita Mountains: Kansas Geol. Soc. 29th Field Conf. Guidebook, p. 91-111.

___________, 1970, Sedimentary features of late Paleozoic flysch, Ouachita Mountains, Oklahoma, in Lajoie, J., ed., Flysch sedimentology in North America: Geol. Assoc. Canada Spec. Paper 7, p. 85-101.

Flawn, P. T., A. Goldstein, Jr., P. B. King, and C. E. Weaver, 1961, The Ouachita System: Texas Univ. Bur. Econ. Geology Pub. 6120, 401 p.

Graham, S. A., W. R. Dickinson, and R. V. Ingersoll, 1975, Himalayan-Bengal model for flysch dispersal in the Appalachian-Ouachita system: Geol. Soc. Amer. Bull., v. 86, p. 273-286.

Haley, B. R., et al., 1967, Geologic map of Arkansas: U.S. Geological Survey and Arkansas Geological Commission, scale 1-500,000.

Handford, C. R., 1986, Facies and bedding sequences in shelf-storm-deposited carbonates--Fayetteville Shale and Pitkin Limestone (Mississippian), Arkansas: Jour. Sed. Petrology, v. 56, p. 123-137.

Houseknecht, D. W., and J. A. Kacena, 1983, Tectonic and sedimentary evolution of the Arkoma foreland basin, in Tectonic-sedimentary evolution of the Arkoma basin: Society of Econ. Paleontologists and Mineralogists Midcontinent Sect., v. 1, p. 3-33.

Houseknecht, D. W., and S. M. Matthews, 1985, Thermal maturity of Carboniferous strata, Ouachita Mountains: American Assoc. Petrol. Geologists Bull., v. 69, p. 335-345.

Mack, G. H., W. C. James, and W. A. Thomas, 1981, Orogenic provenance of Mississippian sandstones associated with southern Appalachian-Ouachita orogen: American Assoc. Petrol. Geologists Bull., v. 65, p. 1444-1456.

Mack, G. H., W. A. Thomas, and C. A. Horsey, 1983, Composition of Carboniferous sandstones and tectonic framework of southern Appalachian-Ouachita orogen: Jour. Sed. Petrology, v. 53, p. 931-946.

Maher, J. C., and R. J. Lantz, 1953, Correlation of pre-Atoka rocks in the Arkansas Valley, Arkansas: U.S. Geol. Survey Oil and Gas Inv. Chart OC-51.

Moiola, R. J., and E. F. McBride, 1978, Sedimentology of Ouachita turbidites, southeastern Oklahoma--A summary, in Briggs, E. F., E. F. McBride, and R. J. Moiola, eds., Sedimentology of Paleozoic flysch and associated deposits, Ouachita Mountains-Arkoma basin, Oklahoma: Dallas Geol. Soc. Guidebook, p. 42-50.

Moiola, R. J., and G. Shanmugam, 1984, Submarine fan sedimentation, Ouachita Mountains, Arkansas and Oklahoma: Trans. Gulf Coast Assoc. Geol. Soc., v. XXXIV, p. 175-182.

Morris, R. C., 1971, Stratigraphy and sedimentology of the Jackfork Group, Arkansas: American Assoc. Petrol. Geol. Bull., v. 55, p. 387-402.

1974, Sedimentary and tectonic history of the Ouachita Mountains, in W. R. Dickinson, ed., Tectonics and sedimentation: Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 22, p. 120-147.

Mutti, E., 1985, Turbidite systems and their relations to depositional sequences, in Zuffa, G. G., ed., Provenance of arenites, D. Riedel Publ. Co., p. 65-93.

Niem, A. R., 1967, Patterns of flysch deposition and deep-sea fans in the Ouachita Mountains, Oklahoma and Arkansas: Jour. Sed. Petrology, v. 46, p. 633-646.

O'Donnell, M. R., 1983, Regressive shelf deposits in the Pennsylvanian Arkoma Basin, Oklahoma and Arkansas: Shale Shaker: Jour. Oklahoma City Geol. Soc., v. 34, p. 23-37.

Owen, M. R., 1984, Southern source for upper Jackfork Sandstone, Ouachita Mountains, Arkansas, in Stone, C. G., and B. R. Haley, eds., A Guidebook to the geology of the central and southern Ouachita Mountains, Arkansas: Arkansas Geol. Comm. Guidebook 84-2, p. 116-122.

Owen, M. R., and A. V. Carozzi, 1986, Southern provenance of upper Jackfork Sandstone, southern Ouachita Mountains: Cathodoluminescence petrology: Geol. Soc. America Bull., v. 97, p. 110-115.

Shideler, G. L., 1970, Provenance of Johns Valley boulders in late Paleozoic Ouachita facies, southeastern Oklahoma and southwestern Arkansas: American Assoc. Petrol. Geol. Bull., v. 54, pp. 789-806.

Stone, C. G., 1968, The Atoka Formation in north-central Arkansas: Arkansas Geol. Comm., 25 p.

Stone, C. G., et al., 1979, Stop descriptions--Third 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. 39-53.

Stone, C. G., and W. V. Bush, 1984, Summary of the geology of the central and southern Ouachita Mountains, Arkansas, in Stone, C. G., and B. R. Haley, eds., A Guidebook to the geology of the central and southern Ouachita Mountains, Arkansas: Arkansas Geol. Comm. Guidebook 84-2, p. 65-75.

Stone, C. G., B. R. Haley, and G. W. Viele, 1973, A guidebook to the geology of the Ouachita Mountains, Arkansas: Arkansas Geol. Comm. Guidebook 73-1, 113 p.

Stone, C. G., and J. D. McFarland, III, 1981, Field guide to the Paleozoic rocks of the Ouachita Mountains and Arkansas Valley provinces, Arkansas: Arkansas Geol. Comm. Guidebook 81-1, 140 p.

Sutherland, P. K., in preparation, Late Mississippian and Pennsylvanian depositional history of the Arkoma basin area, Oklahoma and Arkansas: Geol. Soc. America Bull.

Sutherland, P. K. and T. W. Henry, 1977, Carbonate platform facies and new stratigraphic nomenclature of the Morrowan Series (Lower and Middle Pennsylvanian), northeastern Oklahoma: Geol. Soc. America Bull., v. 88, p. 425-440.

Sutherland, P. K. and W. L. Manger, 1977, Upper Chesterian-Morrowan stratigraphy and the Mississippian-Pennsylvanian boundary in northeastern Oklahoma and northwestern Arkansas: Oklahoma Geol. Survey Guidebook 18, 183 p.

__________, 1979, Mississippian-Pennsylvanian shelf-to-basin transition, Ozark and Ouachita regions, Oklahoma and Arkansas: Oklahoma Geol. Surv. Guidebook 19, 81 p.

Thomas, A. and R. J. LeBlanc, 1975, Carboniferous deep-sea fan facies of Arkansas and Oklahoma (Abstr.): Geol. Soc. America Abstracts, v. 7, p. 1298-1299.

Ulrich, E. O., 1927, Fossiliferous boulders in the Ouachita "Caney" Shale and the age of the shale containing them: Oklahoma Geol. Surv. Bull., v. 45, 48 p.

Viele, G. W., 1973, Structure and tectonic history of the Ouachita Mountains, Arkansas, in DeJong, K. A., and R. Scholten, eds., Gravity and tectonics: New York, John Wiley and Sons, p. 361-377.

Wickham, J. S., D. R. Roeder, and G. Briggs, 1976, Plate tectonics model for the Ouachita foldbelt: Geology, v. 4, p. 173-176.

Zachry, D. L., Jr., 1975, Early Pennsylvanian fluvial sedimentation in northwestern Arkansas (Abstract): American Assoc. Petroleum Geologists and Soc. Econ. Paleontologists and Mineralogists Annual Meetings Abstracts, v. 2, p. 84-85.

__________, 1977, Stratigraphy of middle and upper Bloyd strata (Pennsylvanian, Morrowan), northwestern Arkansas, in Sutherland, P. K., and W. L. Manger, eds., Upper Chesterian-Morrowan stratigraphy and the Mississippian-Pennsylvanian boundary in northeastern Oklahoma and northwestern Arkansas: Oklahoma Geol. Surv. Guidebook 18, p. 61-66.

Zachry, D. L., Jr. and B. R. Haley, 1975, Stratigraphic relationships between the Bloyd and Atoka Formations (Pennsylvanian) of northern Arkansas, in Contributions to the geology of the Arkansas Ozarks: Arkansas Geol. Comm., p. 96-106.

Fossil plants from the Jackfork Sandstone in the Ouachita Mountains of Arkansas

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ABSTRACT

A variety of fossil plants, including diverse forms of foliage, have been found in the Jackfork Sandstone in the southern Ouachita Mountains of Arkansas. The flora contains *Lepidodendron aculeatum*, lycopod leaves, *Lepidostrobus cf. L. peniculus*, *Lepidostrobophyllum* sp., *Stigmaria ficoides*, *S. stellata*, *Calamites cistiiformis*, *C. roemeri*, *Asterophyllites grandis*, *Adiantites* sp., *Alloiopteris cf. A. Arkansana*, cf. *"Mariopteris" renieri*, *Neuropteris antecedans*, *Sphenopteris cf. S. ettingshauseni*, *S. cf. S. hollandica*, *S. ("Rhodea") moravica*, *S. cf. S. stangeri*, *S. trifoliatoides*, seed fern petioles, *Telangiopsis* sp., *Scheutzia* sp., *Lagenospermum* sp., *Rhynchogonium choctavense*, *Trigonocarpus gillhami*, *Artisia transversa*, *Cardiocarpon (Cordaicarpus) minimus*, *C. (C.) cf. C. elongata*, *Samaropsis florini*, and *S. parvefluitans*. The fossil plants of the Jackfork Sandstone indicate that the unit is younger than the youngest beds of the type Mississippian, but older than beds recognized as Pennsylvanian.

Identifiable fossils, especially invertebrate macrofossils, are quite rare in the Jackfork Sandstone of the Ouachita Mountain region in Arkansas and Oklahoma. In addition, the general scarcity of reliable stratigraphic marker beds, the complexity of the regional structure, and the frequency of reworking among fossils pose major difficulties for the accurate dating of the unit. As a result, although the strata have for many years been recognized as Carboniferous (e. g. Ulrich, 1911), more precise chronostratigraphic placement has proved elusive. Some authors have regarded the Jackfork Sandstone as Mississippian (Cline, 1956, 1960; Cline and Shelburne, 1959; Miser and Hendricks, 1960), while others have considered it Pennsylvanian (White, 1934, 1937b; Gordon and Stone, 1977; Stone and McFarland, 1981) or perhaps something in between (White, 1937b; Harlton, 1938; Cooper, 1945; Morris, 1971).

Plant megafossils are particularly valuable as biostratigraphic indicators in the Jackfork Sandstone because they are the most common type of fossil and because they are less subject to reworking than most other kinds of fossils. To date, however, the work of White (1937b) has provided the only systematic treatment of the plant fossils from these strata. Unfortunately, as noted by Mamay (1960), the Jackfork plants described by White (1937b) are exceedingly fragmentary and the generally coarse-grained nature of the enclosing matrix severely limits the quality of preservation. Furthermore, many refinements of the taxonomy and the stratigraphic ranges of fossil plants have been made since White (1937b) completed his investigation. A few taxonomic revisions have been suggested by Mamay (1960), and based on the published illustrations (White, 1937b), additional changes can be made. After the essential taxonomic revisions have been made, the small size of White's (1937b) flora is apparent. Indeed, White (1937b) reported only *Lepidodendron aculeatum*, *Lepidostrobus peniculus cf. Neuropteris antecedans*, and indeterminable foliage from the Jackfork Sandstone in Arkansas. Another specimen of *Lepidodendron* has been figured by Stone and McFarland (1981) from the Jackfork Sandstone in Arkansas and probably belongs to the same species as the material figured by White (1937b). The list of fossil plants from the Jackfork Sandstone in Oklahoma (White, 1937b) reflects only slightly more diversity than the list for Arkansas. White's (1937b) flora from Oklahoma includes *Lepidodendron aculeatum*, *Lepidostrobus peniculus*, *Stigmaria ficoides*, *Calamites roemeri, C.* sp., *?Rhabdocarpus* sp., *Rhynchogonium choc-*

tavense, *Trigonocarpus gillhami*, and *T. vallisjohnni*.

Recent field work in the Jackfork Sandstone in the southern Ouachita Mountains of Arkansas (Fig. 1) has uncovered a number of new fossil plant localities (Table 1). A variety of fossil plants has been found, including examples of lycopods, arthrophytes, ferns, seed ferns, and cordaiteans. Calamitean pith casts and impressions are the most conspicuous element in the flora, although lycopod stem impressions and various seed casts are also widespread. On the other hand, identifiable foliage is much less common. Nevertheless, in spite of this and the fragmentary nature of much of the material, a surprisingly diverse assemblage of foliar remains is present. The flora of the Jackfork Sandstone in Arkansas now includes *Lepidodendron aculeatum* (Pl. 1, fig. 1), lycopod leaves (Pl. 1, fig. 2), *Lepidostrobus cf. L. peniculus* (Pl. 1, fig. 4), *Lepidostrobophyllum* sp. (Pl. 1, fig. 5), *Stigmaria ficoides* (Pl. 1, fig. 6), *S. stellata* (Pl. 1, fig. 7), *Calamites cistiiformis* (Pl. 1, fig. 8-12), *C. roemeri* (Pl. 1, fig. 13; Pl. 2, fig. 1,2), *Asterophyllites grandis* (Pl. 2, fig. 3,4), *Adiantites* sp. (Pl. 2, fig. 5), *Alloiopteris cf. A. arkansana* (Pl. 2, fig. 6,7), cf. *"Mariopteris" renieri* (Pl. 2, fig. 8), *Neuropteris antecedans* (Pl. 2, fig. 9,10), *cf. Neuropteris* sp. (Pl. 3, fig. 1), *Sphenopteris cf. S. ettingshauseni* (Pl. 3, fig. 2,3), *S. cf. S. hollandica* (Pl. 3, fig. 4-6), *S. ("Rhodea") moravica* (Pl. 3, fig. 8-11), *S. cf. S. stangeri* (Pl. 3, fig, 12, 13), *S. trifoliatoides* (Pl. 3, fig. 14), seed fern petioles (Pl. 4, fig. 1), *Telangiopsis* sp. (Pl. 4, fig. 2), *Scheutzia* sp. (Pl. 4, fig. 3), *Lagenospermum* sp. (Pl. 4, fig. 4), *Rhynchogonium choctavense* (Pl. 4, fig. 5,6), *Trigonocarpus gillhami* (Pl. 4, fig. 7), *Artisia transversa* (Pl. 4, fig. 8), *Cardiocarpon (Cordaicarpus) minimus* (Pl. 4, fig. 9), *C. cf. C. elongata* (Pl. 4, fig. 10), *Samaropsis florini* (Pl. 4, fig. 11), and *S. parvefluitans* (Pl. 4, fig. 12, 13). The occurrence of these species at each of the different collecting localities in the Jackfork

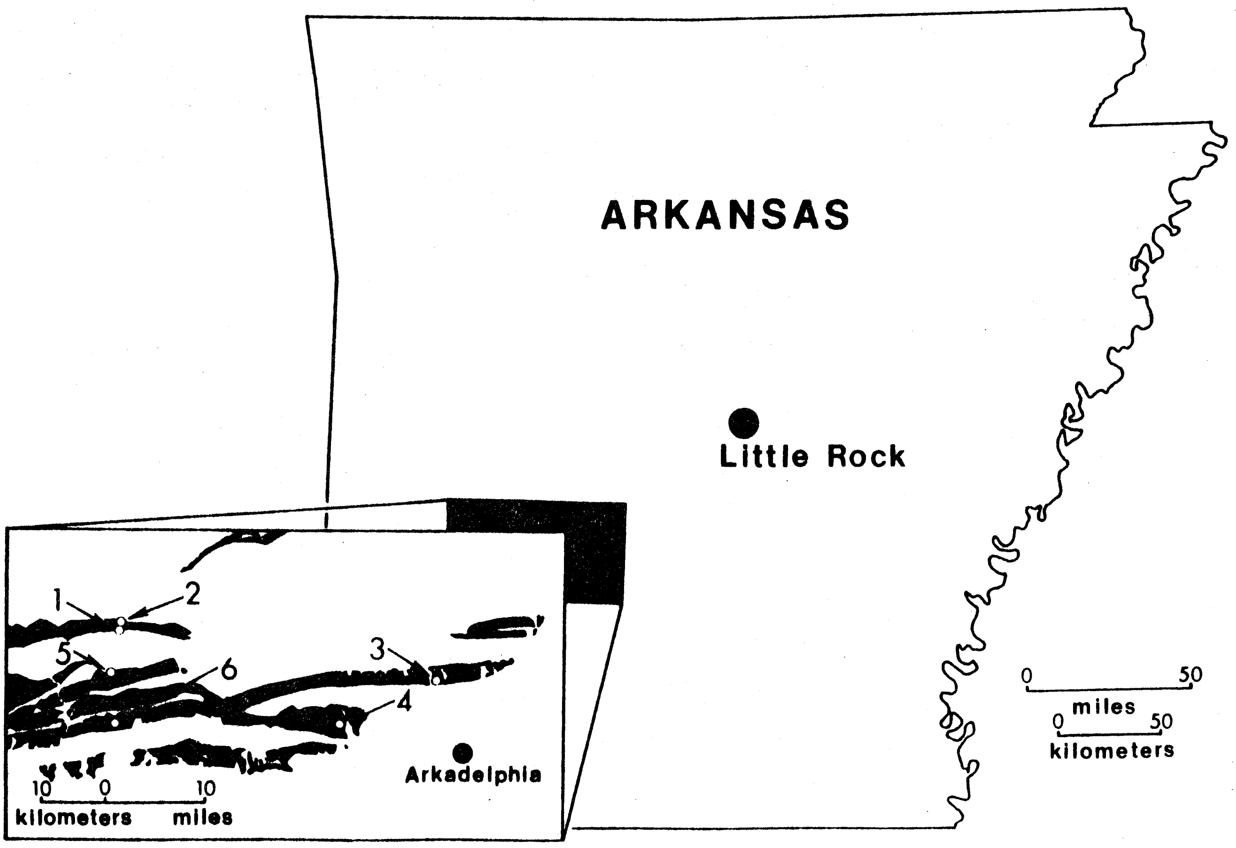

Figure 1. Outline map of Arkansas with inset of the area from which the fossil plants were collected. On the inset, the outcrop of the Jackfork Sandstone is blackened and the white dots indicate collecting sites.

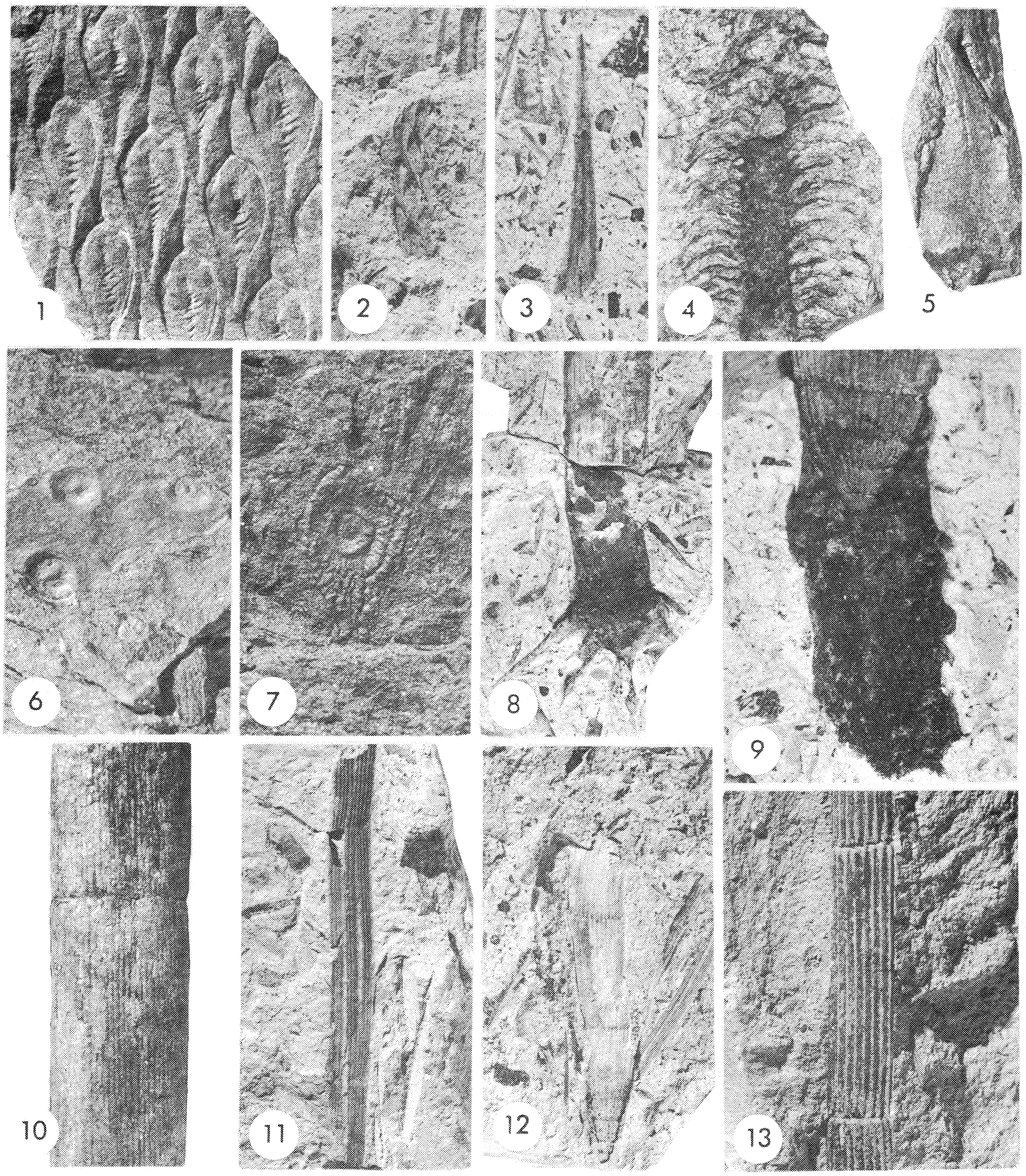

Plate 1. Jackfork plants. 1. *Lepidodendron aculeatum*. Locality 3-6. X1. 2. *L*. sp. Locality 1-2. X2. 3. Lycopod leaf. Locality 1-2. X2. 4. *Lepidostrobus (sensu lato) cf. L. peniculus*. Locality 3-1. X1. 5. *Lepidostrobophyllum* sp. Locality 3-2. X1. 6. *Stigmaria ficoides*. Locality 3-3. X2. 7. *S. stellata*. Locality 3-3. X2. 8. *Calamites cf. C. cistiiformis* with roots at base. Locality 1-2. X2. 9. *C. cf. C. cistiiformis*. Enlarged view of specimen with basal pith cast and roots. Locality 1-2. X2. 10. *C. cistiiformis*. Locality 5. X1. 11. *C. cistiiformis*. Locality 1-2. X2. 12. *C. cistiiformis*. Pith impression from base of small branch. Locality 1-2. X2. 13. *C. roemeri*. Locality 6. X1.

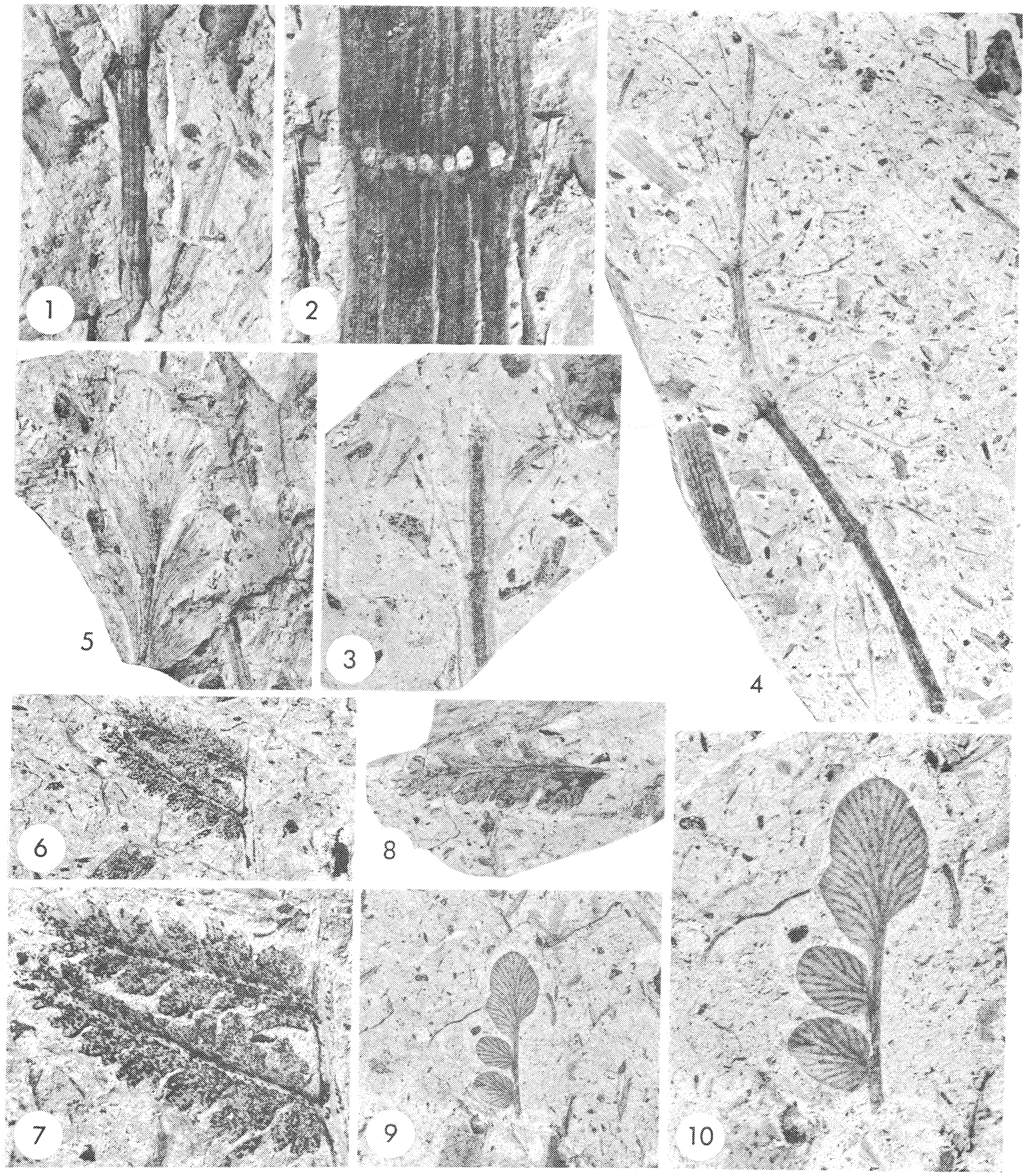

Plate 2. Jackfork Plants. 1. *Calamites roemeri*. Locality 1-2. X2. 2. *C. roemeri*. Enlargement of nodal region. Locality 1-2. X2. 3. *Asterophyllites grandis*. Locality 1-2. X2. 4. *A. grandis*. Locality 1-2. X2. 5. *Adiantites* sp. Locality 1-2. X2. 6. *Alloiopteris cf. A. arkansana*. Locality 1-2. X2. 7. *A. cf. A. arkansana*. Locality 1-2. X4. 8. *Cf. Mariopteris renieri*. Locality 1-2. X2. 9. *Neuropteris antecedans*. Locality 1-2. X2. 10. *N. antecedans*. Locality 1-2. X4.

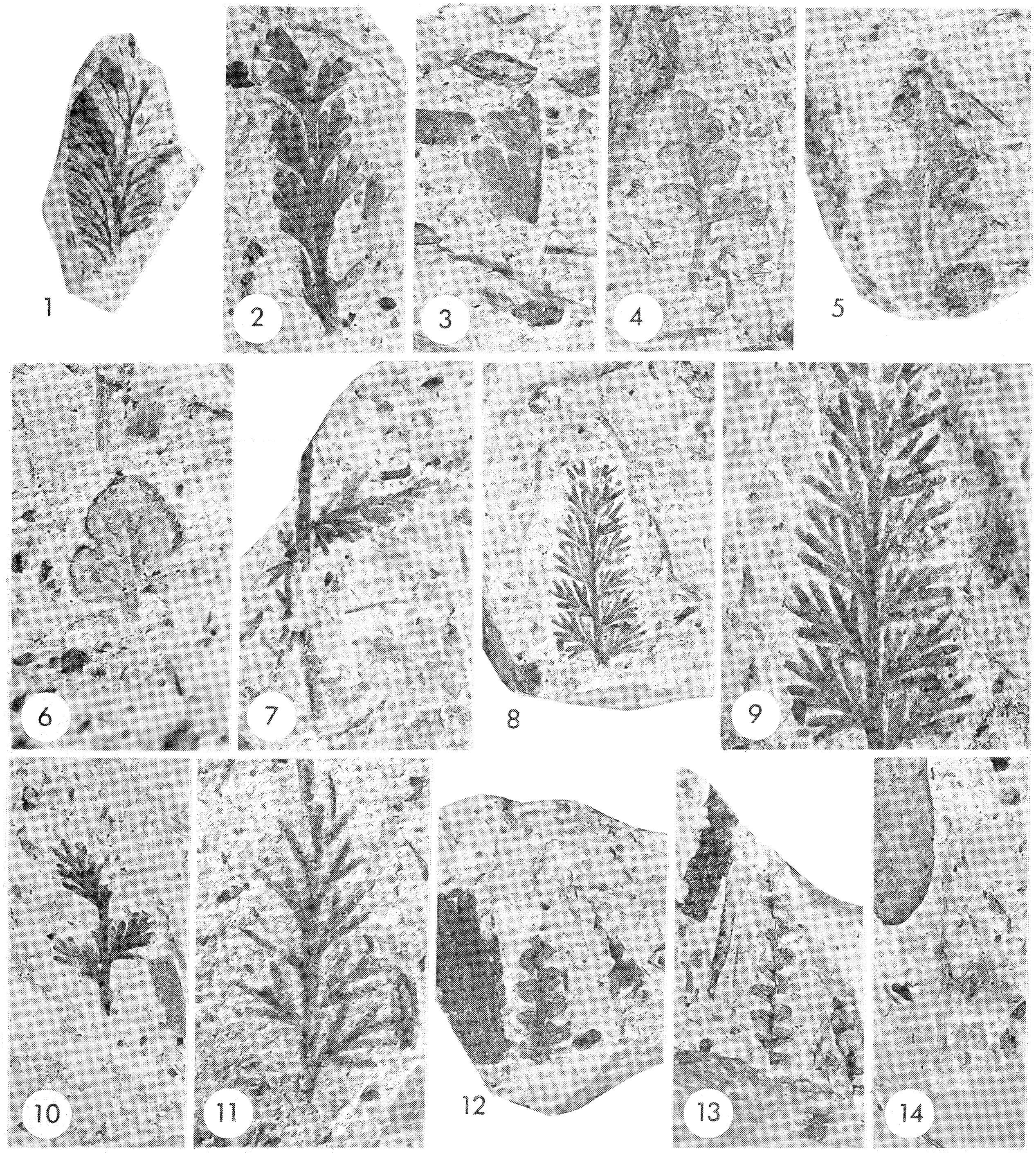

Plate 3. Jackfork Plants. 1. *Cf. Neuropteris* sp. Pinnule with anastomosing veins. Locality 1-2. X4. 2. *Sphenopteris cf. S. ettingshauseni*. Locality 1-2. X2. 3. *S. cf. S. ettingshauseni*. Locality 1-2. X2. 4. *S. cf. S. hollandica*. Locality 1-2. X2. 5. *S. cf. S. hollandica*. Locality 1-2. X4. 6. *S. cf. S. hollandica*. Locality 1-2. X4. 7. *S. ("Rhodea") moravica*. Locality 1-2. X2. 8. *S. moravica*. Locality 1-2. X2. 9. *S. moravica*. Locality 1-2. X4. 10. *S. moravica*. Locality 1-2. X2. 11. *S. moravica*. Locality 1-2. X4. 12. *S. cf. S. stangeri*. Locality 1-2. X2. 13. *S. cf. S. stangeri*. Locality 1-2. X2. 14. *S. trifoliatoides*. Locality 1-2. X2.

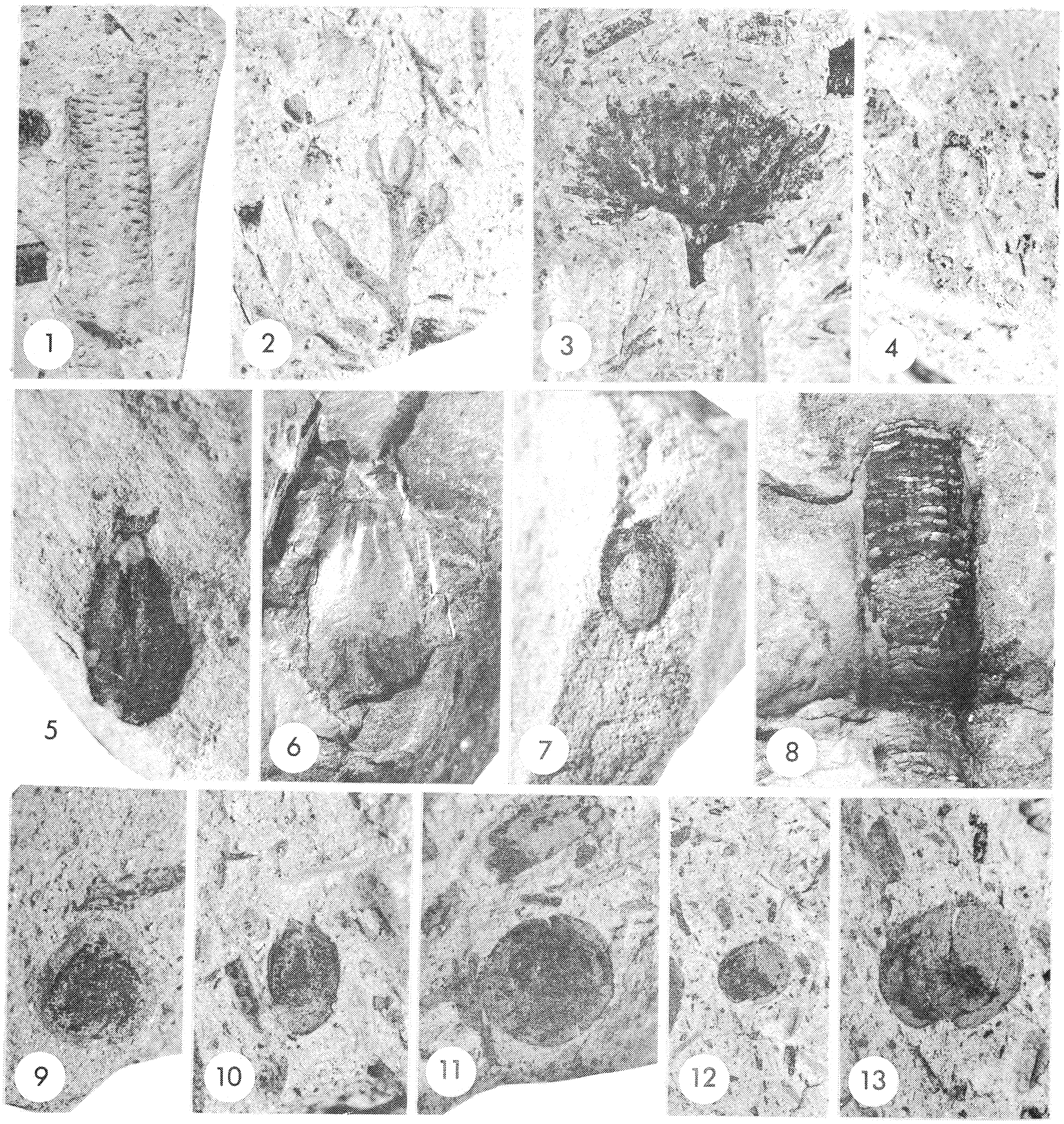

Plate 4. Jackfork Plants. 1. Seed fern petiole. Locality 1-2. X2. 2. *Telangiopsis* sp. Locality 1-2. X4. 3. *Scheutzia* sp. Locality 1-2. X2. 4. *Lagenospermum* sp. Locality 1-2. X4. 5. *Rhynchogonium choctavense*. Locality 2. X2. 6. *R. choctavense*. Locality 3-2. X2. 7. *Trigonocarpus gillhami*. Locality 2. X2. 8. *Artisia transversa*. Locality 4. X2. 9. *Cardiocarpon (Cordaicarpus) minimus*. Locality 3-3. X4. 10. *C. (C.) cf. C. elongata*. Locality 1-2. X4. 11. *Samaropsis florini*. Locality 1-2. X4. 12. *S. parvefluitans*. Locality 1-2. X2. 13. *S. parvefluitans*. Locality 1-2. X4.

Table I. List of localities in the Jackfork Sandstone in the southern Ouachita Mountains of Arkansas from which fossil plants were collected during this study.

Locality 1: Glen South

SE 1/4, SE 1/4, NE 1/4, Sec. 24, T. 5 S., R. 25 W., Pike Co., Arkansas, in shale and siltstone exposed in roadcut along Rt. 27 south of Salem near Glen Lookout. 1-1, lower bed. 1-2, upper bed.

Locality 2: Glen North

SE 1/4, NE 1/4, NW 1/4, Sec. 24, T. 5 S., R. 25 W., Pike Co., Arkansas, in siltstone and sandstone exposed in cut for Rt. 27 south of Salem near Glen Lookout.

Locality 3: De Gray Lake

W 1/2, Sec. 13, T. 6 S., R. 20 W., Clark Co., Arkansas, in strata exposed in overflow spillway for De Gray Lake. 3-1, NE 1/4, SW 1/4, NW 1/4, Sec. 13, in siltstone and sandstone exposed at 70' on section by Morris (1977). 3-2, NE 1/4, SW 1/4, NW 1/4, Sec. 13, in shale exposed at 128' in section by Morris (1977). 3-3, NE 1/4, SW 1/4, NW 1/4, Sec. 13, in shale and siltstone exposed at 202' on section by Morris (1977). 3-4, SE 1/4, SW 1/4, NW 1/4, Sec. 13, in sandstone exposed at 560' on section by Morris (1977). 3-5, SE 1/4, SW 1/4, NW 1/4, Sec. 13, in sandstone and siltstone exposed at 635' on section by Morris (1977). 3-6, SE 1/4, SW 1/4, NW 1/4, Sec. 13, in shale, siltstone, and sandstone exposed at 840' on section by Morris (1977).

Locality 4: Hollywood Quarry

SE 1/4, NE 1/4, SE 1/4, Sec. 8, T. 7 S., R. 21 W., Clark Co., Arkansas, in siltstone and sandstone exposed in upper bench of quarry.

Locality 5: Kirby Mountain

SW 1/4, SW 1/4, NE 1/4, Sec. 14, T. 6 S., R. 25 W., Pike Co., Arkansas, in shale siltstone, and sandstone exposed along Rt. 27 near the base of the Jackfork section measured by Morris (1971).

Locality 6: Kirby Lookout

NW 1/4, SE 1/4, NE 1/4, Sec. 11, T. 7 S., R. 25 W., Pike Co., Arkansas, in siltstone exposed in roadcut along Rt. 27 adjacent to Kirby Lookout.

Sandstone of Arkansas is shown in Table II.

Because the flora of the Jackfork Sandstone has been greatly enlarged as a result of this investigation, there is now sufficient information for reasonably accurate age comparisons to be made between the Jackfork flora and other floras of Euramerica. In this regard, the diverse forms of foliage encountered are especially important. Foliage is widely utilized for biostratigraphic purposes because many forms have short stratigraphic ranges, and because the ranges are better known than those of some other types of plant fossils. The type Mississippian, like the Jackfork, is floristically much better known now (Jennings, 1984; Jennings and Fraunfelter, in press) than it was at the time that White (1937b) considered the question of age relationships. The top of the type Mississippian in the Illinois basin is unconformable, but there is a good sequence of floras below the boundary. The Grove Church Formation is the uppermost unit of the Mississippian sequence, and has a flora containing abundant arthrophytes and seed ferns. The arthrophytes present in it include *Calamites cistiiformis*, *C. roemeri*, and *Asterophyllites grandis*. These species are somewhat long-ranging, and all of them occur in both the Grove Church Formation and the Jackfork Sandstone. On the other hand, the seed fern foliage in the Grove Church Formation is distinct from that in the Jackfork Sandstone. The Grove Church has abundant representatives of *Sphenopteris affinis*, *S. macconochii*, *S. crassa*, and *S. fragilis*, whereas these species are absent from the Jackfork.

Table II. Chart showing the localities in the Jackfork Sandstone of Arkansas from which each type of fossil plant was obtained during this study.

FOSSIL PLANT	LOCALITY											
	1-1	1-2	2	3-1	3-2	3-3	3-4	3-5	3-6	4	5	6
Lepidodendron sp.		X										
Lepidodendron aculeatum						X	X		X			
lycopod leaves		X		X		X						X
Lepidostrobus cf. L. peniculus				X		X						
Lepidostrobophyllum sp.					X							X
Stigmaria ficoides		X		X		X				X	X	X
S. stellata		X				X						
Calamites sp.	X	X	X	X	X	X		X		X	X	X
C. cistiiformis		X	X	X		X					X	X
C. roemeri		X		X		X				X		X
Asterophyllites grandis		X										
Adiantites sp.		X										
Alloiopteris cf. A. arkansana		X										
cf. "Mariopteris" renieri		X										
Neuropteris sp.		X			X						X	X
N. antecedens		X	X									
Sphenopteris cf. S. ettingshauseni		X										
S. cf. S. hollandica		X										
S. ("Rhodea") sp.	X	X	X		X	X					X	X
S. ("R.") moravica	X	X										
S. cf. S. stangeri		X			X							
S. trifoliatoides		X										
seed fern petioles		X										
Telangiopsis sp.		X										
Scheutzia sp.		X										
Lagenospermum sp.		X										
Rhynchogonium choctavense		X	X			X						
Trigonocarpus sp.		X	X			X						X
T. gillhami			X			X						
Artisia transversa										X		
Cardiocarpon (Cordaicarpus) sp.		X				X					X	X
C. (C.) minimus						X						
C. (C.) cf. C. elongata		X										
Samaropsis florini		X										
S. parvefluitans		X										

Alloiopteris cf. A. arkansana, *Sphenopteris cf. S. hollandica*, and *S. ("Rhodea") moravica*, which are present in the Jackfork Sandstone, are absent from the type Mississippian. *Neuropteris antecedans* is the only form of seed fern foliage present in both floras. This species, which is well known in European deposits (e. g. Stur, 1875; Stockmans and Willière, 1952-53; Crookall, 1959; Purkyňová, 1975), is known to exist throughout most of the Upper Mississippian, as well as in younger strata. The cordaiteans constitute perhaps the most conspicuous element that is present in the Jackfork flora, but absent from the type Mississippian. Other known Mississippian floras (e. g. Gillespie and Pfefferkorn, 1970; Read and Mamay, 1964; White 1937a; Arnold and Sadlick, 1962) of North America differ from the Jackfork flora in the same manner as the flora of the upper part of the type Mississippian. The differences between the flora of Upper Mississippian strata and the flora of the Jackfork Sandstone indicate that the Jackfork was deposited slightly later.

Comparison with recognized Pennsylvanian floras of North America reveals that the Jackfork flora is different from these also. Although the presence of *Lepidodendron aculeatum*, *Alloiopteris*, and cordaiteans shows affinity with recognized Pennsylvanian-age floras, *Sphenopteris cf. S. ettingshauseni*, *S. cf. S. hollandica*, *S. ("Rhodea") moravica*, *S. cf. S. stangeri*, and *S. trifoliatoides*, for example, are unknown in strata that can unequivocally be referred to the Pennsylvanian. Indeed, *Stigmaria stellata*, *Sphenopteris ("Rhodea") moravica*, and *Neuropteris antecedans* give the Jackfork flora an appearance that is more like latest Mississippian floras, even though these species are not restricted to the Mississippian. The large neuropteroid, alethopteroid, and mariopterioid pinnules that are characteristic of Pennsylvanian floras (e. g. Read and Mamay, 1964) are conspicuously absent from the Jackfork Sandstone.

Among North American floras, the two that are the closest time equivalents to the Jackfork flora appear to be those from the Manning Canyon Shale of Utah (Tidwell, 1967; Tidwell et al., 1974) and from the lower-middle Parkwood Formation of Alabama (Jennings and Thomas, in press). Like the Jackfork flora, these floras both contain a combination of floral elements that are characteristic of Late Mississippian strata with elements widespread in Pennsylvanian strata. On the other hand, comparable floras have yet to be reported from the central Appalachians, and lower Pennsylvanian strata there lie directly on strata equivalent to the uppermost type Mississippian (Wagner, 1982; Jennings and Fraunfelter, in press).

In Europe, floras of Namurian age are strikingly similar to the flora of the Jackfork Sandstone of Arkansas. The type Namurian has been investigated by Stockmans and Willière (1952-53, 1955), and contains a large flora. This flora shares *Lepidodendron aculeatum*, *Calamites cistiiformis*, *C. roemeri*, *Asterophyllites grandis*, *Adiantites* sp., *Alloiopteris* sp., *"Mariopteris" renieri*, *Neuropteris antecedans*, *Sphenopteris hollandica*, *S. ("Rhodea")* sp., *S. cf. S. stangeri*, *S. trifoliatoides*, *Telangiopsis* sp., *Artisia transversa*, *Samaropsis florini*, and *S. parvefluitans* with the flora of the Jackfork Sandstone. Unfortunately, there is an unconformity at the base of the type Namurian (Stockmans and Willière, 1955), and strata from other areas have been added to its base in an attempt to fill the chronostratigraphic gap. The current interpretation of the sequence of these "lower Namurian" strata is not, however, beyond question. White (1937b), in fact, suggested that some of these strata perhaps should not have been placed in the Namurian. In any event, the flora of even the basal strata of the type Namurian indicates clearly an age that is younger than the age of the youngest strata of the type Mississippian (Jennings and Fraunfelter, in press).

Since the fossil floras indicate that strata equivalent in age to the Jackfork Sandstone are not present in either the type Mississippian or the type Pennsylvanian, there is no objective basis for assigning the unit to either system. It has not, thus far, been possible to trace the floral sequence in the southern Ouachitas upward into strata of definite Pennsylvanian age. While the internal stratigraphy of the Jackfork Sandstone presents problems, floral material has been obtained from beds regarded as lower (e.g. locality 5), middle (e.g. locality 3), and upper (e. g. locality 3-6, 4) Jackfork (see Morris, 1971; Stone and McFarland, 1981). With available data, it has not proved possible to distinguish differences in age at different levels within the Jackfork based on the flora. Interestingly, comparison with the flora of the Stanley Shale described by White (1937b) shows great similarity between this flora and that of the Jackfork. The best of White's (1937b) material, particularly the foliage, came from a single outcrop. That site is identified as upper Stanley (Miser and Purdue, 1929), and it may not be surprising that its flora is very similar to the Jackfork flora. *Lepidodendron aculeatum*, *Stigmaria ficoides*, *Lepidostrobus peniculus*, *Calamites cistiiformis*, *C. roemeri*, *Alloiopteris arkansana*, *cf. Neuropteris*, *Sphenopteris trifoliatoides*, *Trigonocarpus gillhami*, and *T. vallisjohnni* are present in both the Stanley and the Jackfork. In addition, specimens of *Sphenopteris ("Rhodea")*, that White (1937b) described from the Stanley are similar to *S. ("Rhodea") moravica*, which occurs in the Jackfork. The absence of cordaiteans from the Stanley is the most obvious difference between the two floras at present. Although it is not possible at present to establish the position corresponding to the upper boundary of the type Mississippian or the lower boundary of the type Pennsylvanian in the southern Ouachitas, it appears that at least half of the Jackfork and the upper part of the Stanley indeed represent deposits that formed during a time when erosion was active over much of North America, as suggested by various authors (e. g. White, 1937b; Harlton, 1938; Cooper, 1945; Morris, 1971). The fact that fossil plant material sufficient for age determination is present in the Jackfork Sandstone suggests the possibility that further investigation of fossil plants may clarify the age relationships of other upper Paleozoic units in the Ouachita Mountains.

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REFERENCES

Arnold, C. A., and Sadlick, W., 1962, A Mississippian flora from northeastern Utah and its faunal and stratigraphic relations: University of Michigan Museum Paleontological Contributions, v. 17; p. 241-263.

Cline, L. M., 1956, Some stratigraphic studies of the Mississippian and Pennsylvanian rocks of the Ouachita Mountains, Oklahoma: Tulsa Geological Society Digest, v. 24; p. 100-106.

__________, 1960, Stratigraphy of the Late Paleozoic rocks of the Ouachita Mountains, Oklahoma: Oklahoma Geological Survey Bulletin 85, 113 p.

Cline, L. M., and Shelburne, O. B., 1959. Late Mississippian-Early Pennsylvanian stratigraphy of the Ouachita Mountains, Oklahoma, in The geology of the Ouachita Mountains - A symposium: Dallas Geological Society and Ardmore Geological Society, p. 175-208.

Cooper, C. L., 1945, Age relationships of Stanley and Jackfork Formations of Oklahoma and Arkansas: Journal of Geology, v. 53; p. 390-397.

Crookall, R., 1959, Fossil plants of the Carboniferous rocks of Great Britain: Memoirs of the Geological Survey of Great Britain, Palaeontology, v. 4(2); p. 85-216.

Gillespie, W. H., and Pfefferkorn, H. W., 1979, Distribution of commonly occurring plant megafossils in the proposed Pennsylvanian System stratotype, in Englund, K. J., Arndt, H. H., Henry, T. W., eds., Proposed Pennsylvanian System strototype, Virginia and West Virginia: American Geol. Inst. Selected Guidebook Series No. 1, p. 87-96.

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, Vol. 1: Arkansas Geological Commission, p. 70-91.

Harlton, B. H., 1938, Stratigraphy of the Bendian of the Oklahoma salient of the Ouachita Mountains: American Association of Petroleum Geologists Bulletin, v. 22; p. 852-914.

Jennings, J. R., 1984, Distribution of fossil plant taxa in the Upper Mississippian and Lower Pennsylvanian of the Illinois Basin: Neuvième Congrés International de Stratigraphy et de Géologie du Carbonifère. Compte Rendu, Vol. 2: Carbondale, Ill., Southern Illinois University Press, p. 301-312.

Jennings, J. R., and Fraunfelter, G. H., in press, Preliminary report on macropaleontology of strata above and below the upper boundary of the type Mississippian: Illinois State Academy of Science Transactions.

Jennings, J. R., and Thomas, W. A., in press, Fossil plants from Mississippian/Pensylvanian transition beds in the southern Appalachians: Southeastern Geology.

Mamay, S. H., 1960, In Miser, H. D., and Hendricks, T. A., Age of Johns Valley Shale, Jackfork Sandstone, and Stanley Shale. American Association of Petroleum Geologists Bulletin, v. 44; p. 1831.

Miser, H. D., and Hendricks, T. A., 1960, Age of Johns Valley Shale, Jackfork Sandstone, and Stanley Shale: American Association of Petroleum Geologists Bulletin, v. 44; p. 1829-1834.

Miser, H. D., and Purdue, A. H., 1929, Geology of the DeQueen and Caddo Gap quadrangles, Arkansas: United States Geological Survey Bulletin, 808, 195 p.

Morris, R. C., 1971, Stratigraphy and sedimentology of Jackfork Group, Arkansas: American Association of Petroleum Geologists Bulletin, v. 55; p. 387-402.

___________, Flysch facies of the Ouachita trough - with examples from the Spillway at De Gray Dam, Arkansas, In Stone, C. G., ed., Symposium on the geology of the Ouachita Mountains, Vol. 1: Arkansas Geological Commission, p. 158-168.

Purkyňová, E., 1975, Die Unternamurflora des beckens von Horní Slezsko (ČSSR): Paläontologishe Abhandlungen, Abteilung B., Band III (Heft 2); p. 129-168.

Read, C. G., and Mamay, S. H., 1964, Upper Paleozoic floral zones and floral provinces of the United States: United States Geological Survey Professional Paper 454K; p. 1-35.

Stockmans, F., and Willière, W., 1952-53, Végétaux Namuriens de la Belgique: Association pour l'Étude de la Paléontologie et de la Stratigraphie Houillères, Pub. no. 13, 382 p.

___________, 1955, Vegetaux Namuriens de la Belgique, II, Assise de Chokier, Zone de Bioul: Association pour l'Etude de la Paleontologie et de la Stratigraphie Houilleres, Pub. no. 23, 35 p.

Stone, C. G., and McFarland, J. D., III, 1981, Field guide to the Ouachita Mountain and Arkansas Valley provinces, Arkansas: Arkansas Geological Commission Guidebook 81-1, 140 p.

Stur, D., 1875, Beiträge zur Kenntnis der Flora der Vorwelt, Band I, Die Culm-Flora des mährisch-schlesischen Dachschiefers: Abhandlungen der kaiserlich-könglichen Geologisches Reichsanstalt, v. 8; p. 1-106.

Tidwell, W. D., 1967, Flora of Manning Canyon Shale, Part I: A lowermost Pennsylvanian flora from the Manning Canyon Shale, Utah, and its stratigraphic significance: Brigham Young University Geology Studies, v. 14; p. 3-66.

Tidwell, W. D., Medlyn, D. A., and Simper, A. D., 1974, Flora of the Manning Canyon Shale, Part II: Lepidodendrales: Brigham Young University Geology Studies, v. 21; p. 119-146.

Ulrich, E. O., 1911, Revision of the Paleozoic systems: Geological Society of America Bulletin, v. 22; p. 281-680.

Wagner, R. H., 1982, Floral changes near the Mississippian-Pennsylvanian boundary; an appraisal, in Ramsbottom, W. H. C., Saunders, W. B., and Owens, B., eds., Biostratigraphic data for a mid-Carboniferous boundary: Palaeobotanical Contributions, Leeds, p. 120-127.

White, D., 1934, Age of Jackfork and Stanley Formations of Ouachita Geosyncline, Arkansas and Oklahoma, as indicated by plants: American Association of Petroleum Geologists Bulletin, v. 18, p. 1010-1017.

___________, 1937a, Fossil flora of the Wedington Sandstone Member of the Fayetteville Shale: United States Geological Survey Professional Paper 186-B, p. 13-41.

___________, 1937b, Fossil plants from the Stanley Shale and Jackfork Sandstone in southeastern Oklahoma and western Arkansas: United States Geological Survey Professional Paper 186-C, p. 43-67.

Late Cambrian North American trilobites and the structural geology of the Jessieville area in Garland County, Arkansas

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INTRODUCTION

Earlier structural mapping by Hart (1985) demonstrated the existence of previously unknown exposures of the Collier Shale in the northwestern portion of the Goosepond Mountain 7.5' quadrangle in Saline and Garland Counties, Arkansas. Limestones within this area were found to contain conodonts in relatively low abundance, but those recovered were diagnostic of the Lower Ordovician part of the Collier Shale as shown by previous studies within the Benton uplift (Repetski and Ethington, 1977; Ethington, 1984; and Ethington, personal commun.). In addition, limestone samples collected by Hart (1985) from the same area (Fig. 1, locality A) yielded several species of trilobites which indicated to one of us (Stitt) Late Cambrian ages and North American affinities.

These findings prompted mapping and sampling of the rocks immediately to the west, in the area surrounding Jessieville, Garland County. The results of this project are presented here as a brief overview which will be expanded in more detail when paleontological analyses of the limestone samples are completed.

TRILOBITES

Trilobites were recovered from five localities during recent structural mapping by the first author and, in part, the third author. These localities are in the area of Jessieville, Arkansas as shown on the map (Fig. 1, localities C-G). The trilobites are locally abundant and occur in dark-gray and black limestones of the Collier Shale.

Species of trilobites identified to date include:

Cliffia lataegenae (Wilson)
Cliffia wilsoni Lochman
Comanchia amplooculata (Frederickson)
Dellea suada (Walcott)
Housia vacuna (Walcott)
Irvingella major Ulrich and Resser
Kindbladia wichitaensis (Resser)
Pseudagnostus communis (Hall and Whitfield)
Pseudokingstonia exotica Palmer
Pterocephalia sanctisabae Roemer

In addition to the trilobites on this list, another five to ten species are present that have not been identified or have been only tentatively identified at this time.

These species are all characteristic of the *Elvinia* Zone (Plate 1) of the Franconian Stage of the Upper Cambrian of North America. The same species of trilobites occur abundantly in light-colored, shallow-water shelf limestones in central Texas (Morgan Creek Member of the Wilberns Formation, Wilson, 1949), Oklahoma (Honey Creek Limestone, Stitt, 1971, 1977), and Missouri (Davis Formation, Kurtz, 1975), as well as in similar strata in the Great Basin of Utah and Nevada (Palmer, 1965). In the Jessieville area, these shelf trilobites are preserved as mostly small specimens (many immature individuals are preserved) that are disarticulated, but unabraded and completely unsorted. Agnostid trilobites are very abundant, including several as yet unidentified forms. Many of the trilobites have been slightly to drastically distorted by later structural deformation of the en-

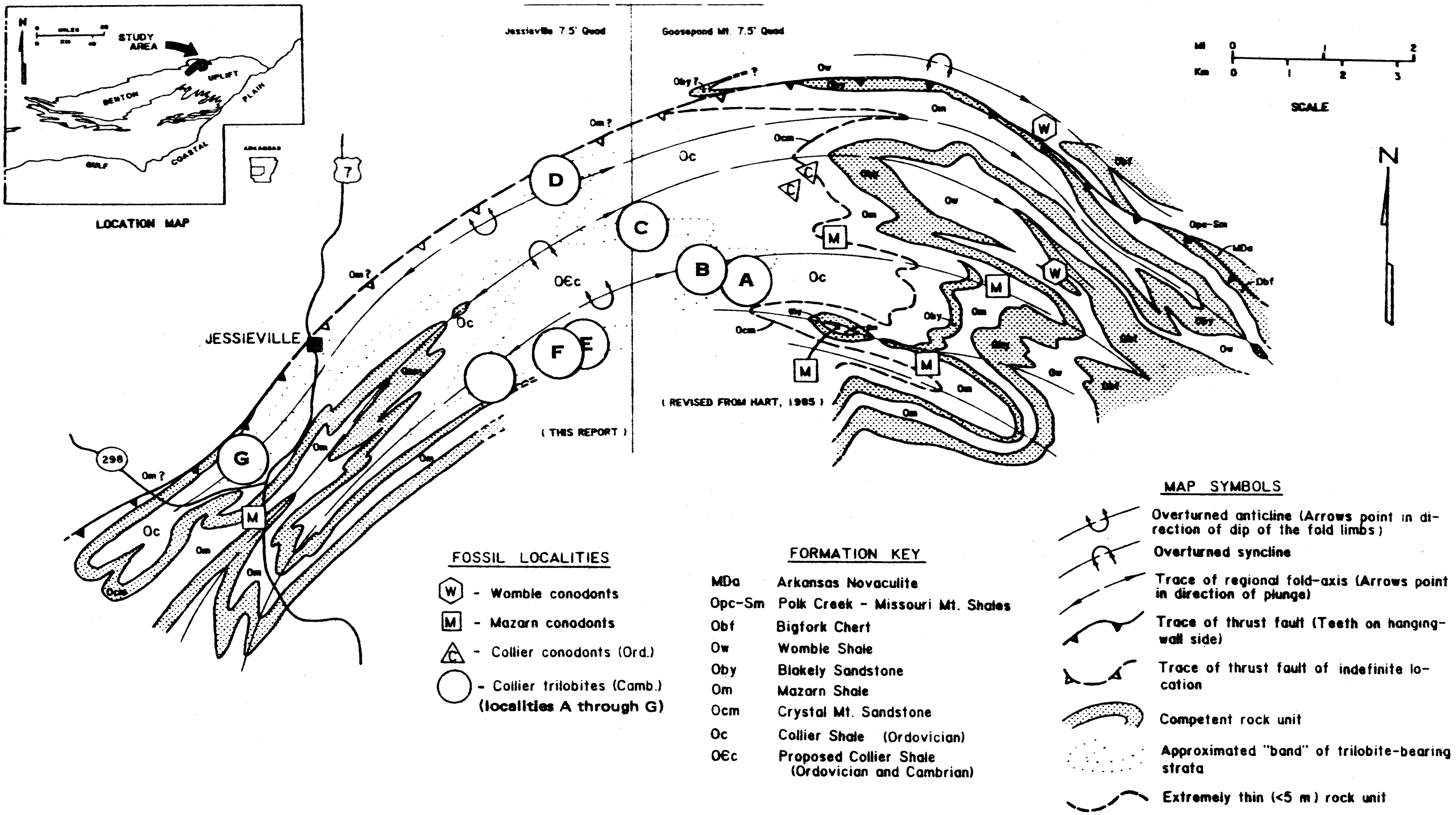


Figure 1. Geologic map of the Jessieville, Arkansas area in Garland County showing fossil localities. Goosepond Mountain quadrangle revised in part from Hart (1985).

closing gray or black limestone.

Trilobites were recovered earlier from two other localities farther east (Fig. 1, localities A and B), from similar limestones, first by Hart (1985) and subsequently by Steven Hohensee (University of Missouri-Columbia). Those trilobites are now thought to be of the same age or slightly younger than the trilobites collected from localities C through G. Inarticulate brachiopods also occur in many of these collections, and a few Late Cambrian conodonts with North American shelf affinities have been identified by Raymond Ethington (University of Missouri-Columbia). These fossils continue to be the subject of a detailed study by Stitt, Hohensee and Ethington.

The trilobites in the Jessieville are were probably deposited in a deep-water, outer-shelf setting, below wave base and close to the margin of the southern edge of the North American craton in Late Cambrian time. Deeper-water trilobites inhabiting slope environments in other areas during the Late Cambrian were predominantly members of the Family Olenidae, but no trilobites belonging to that group have been identified thus far from the Collier Shale.

REGIONAL STRUCTURE

Structural and biostratigraphic data from this and previous studies in the regions to the west (Brown, 1982) and to the east (Hart, 1985) support the existence here of a broadly arched, doubly-plunging anticlinorium that trends southwest-northeast, verges to the southeast and is cored by rocks of the Collier Shale. The core area is bounded on the southwest and northeast by tightly folded, younger rocks of the Crystal Mountain Sandstone. The Crystal Mountain Sandstone to the northeast is a very thin, discontinuous unit shown on the geologic map (Fig. 1) as a single dashed line separating the Collier Shale and the Mazarn Shale. It is possible that the line represents a fault contact, but this cannot be demonstrated at present.

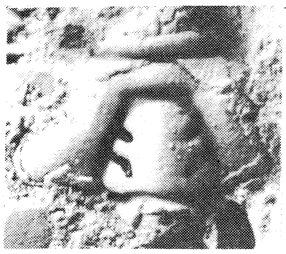
A major thrust fault bounds the Collier rocks of the core area on the northwest and truncates the Crystal Mountain Sandstone and possibly the Mazarn Shale to the southwest. Northeast of the core, the fault enters the Goosepond Mountain quadrangle and is thought to coincide with a thrust fault mapped earlier by Hart (1985). The total area of exposure of the Collier Shale in the core of the anticlinorium exceeds 28

Plate 1. Trilobites of the Elvinia Zone (Late Cambrian) from the Collier Shale, Jessieville area, Arkansas. All specimens from locality D (Fig. 1) except 3 and 6, which are from locality G. All specimens are holaspids unless otherwise noted.

1. *Cliffia lataegenae* (Wilson), cranidium, x6.
2. *Kindbladia wichitaensis* (Resser), cranidium, x6.
3. *Comanchia amplooculata* (Frederickson), two holaspid cranidia, one holaspid pygidium, and one meraspid cranidium, x6.
4. *Dellea suada* (Walcott), cranidium, x12.5.
5. *Housia vacuna* (Walcott), cranidium, and pedicle valve of acrotretid brachiopod, probably *Linnarssonella girtyi* Walcox, x11.
6. *Irvingella major* Ulrich and Resser, slightly compressed cranidium, x6.
7. *Pseudokingstonia exotica* Palmer, cranidium (left center); unidentified agnostid cranidium (right center); tiny unidentified protaspid (right edge), x6.5.
8. *Pseudagnostus communis* (Hall and Whitfield), pygidium; also broken pygidium of *Cliffia* sp., x6.
9. *Cliffia wilsoni* Lochman, cranidium, x6.

10,11. *Pterocephalia sanctisabae* Roemer. 10, cranidium, x1.5; 11, pygidium, x11.5.

12. Piece of limestone showing typical disarticulated preservation of trilobites. Cranidia of *Reaganaspis* n. sp. in upper right and lower center; cranidium of *Pseudagnostus communis* in upper left. Note lack of sorting, and presence of protaspids and many meraspid cranidia and pygidia.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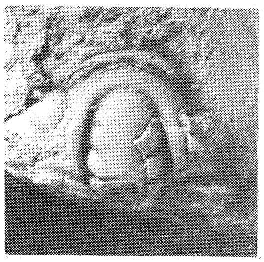
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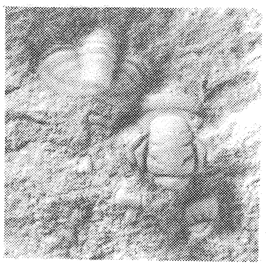
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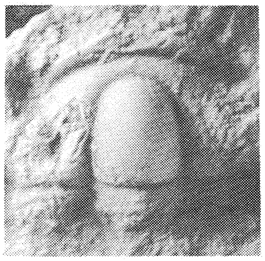
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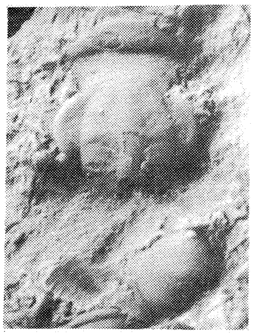
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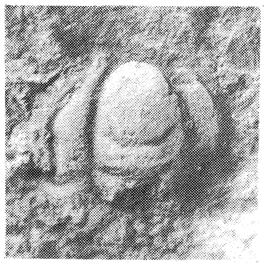
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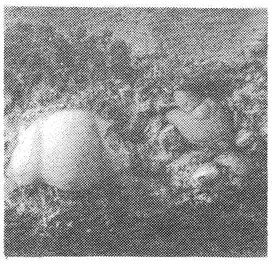
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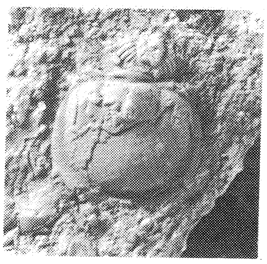
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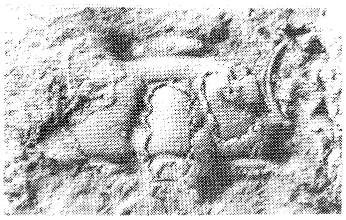
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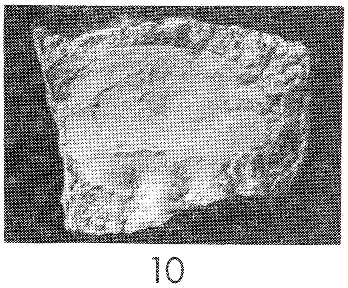
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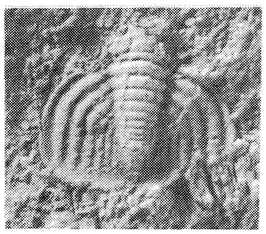
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square kilometers (11 mi^2).

Folds

Folds west of the core plunge gently to the southwest and folds east of the core plunge gently to the northeast. Attitudes of bedding planes indicate that most of the folds are tight to isoclinal and are overturned (verge) to the southeast. Adjacent fold limbs dip at low angles (20 to 35 degrees) to the north and northwest. Attitudes of the prominent "slaty" cleavage surfaces indicate that cleavage is approximately axial-planar to the folds. Fold hinges rake at low angles within the axial planes, indicating a dominance of inclined to near-recumbent fold attitudes. The map traces of the axial planes of the doubly-plunging folds within the anticlinorium are gently bowed to the northwest, suggesting that a broad, regional "upwarp" is superimposed upon the system of tight folds (Fig. 1). At several exposures along the major thrust-fault trace on the northwest, folds were observed within zones of intensely sheared rocks. The folds differ in style and geometry from the regional pattern, and may represent drag folds related to the thrust fault.

Thrust Fault

The major fault along the northwest side of the anticlinorium places the Collier Shale in contact with younger rocks on the northwest side of the fault, possibly with the Mazarn Shale. Placement of the fault trace was based initially upon the truncation of the Crystal Mountain Sandstone southwest of Jessieville (Fig. 1). Here the fault trace roughly follows an incised creek channel that cuts through a prominent ridge of Crystal Mountain Sandstone at a low angle. Continuing to the northeast, the fault trace was extended along the regional trend through several exposures that showed evidence of faulting by the juxtaposition of broadly dissimilar rock types and/or dissimilar structural geometries. Before entering the Goosepond Mountain 7.5' quadrangle to the east, the fault trace is thought to lie approximately 0.5 kilometer north of a Late Cambrian trilobite locality in the Collier Shale (Fig. 1, locality D). In the northeastern portion of the Goosepond Mountain 7.5' quadrangle, Hart (1985) mapped a major thrust fault which now appears to extend westward to coincide with the thrust fault described above. Both to the east and to the west, the thrust fault rises in the stratigraphic section, truncating in sequence the Collier Shale, Crystal Mountain Sandstone, Mazarn Shale and, to the east, the Blakely Sandstone, Womble Shale and the Bigfork Chert along the southern, hanging-wall side of the fault. Younger rocks of the Mazarn Shale through the Arkansas Novaculite constitute the footwall on the north side of the fault.

It is significant that along its entire length, the thrust-fault trace lies along the upper limb of a major south-verging anticline. What this fold/fault geometry suggests in its present configuration is that the younger sequence of rocks to the north of the fault trace are displaced downward and to the north, along a norhtward-dipping fault plane. This constitutes a present "normal" sense of displacement. However, an equally valid interpretation--and the one we favor--is that the older sequence of rocks to the southeast of the fault trace (including the Collier rocks of this report) was initially thrust upwards and to the north over the younger rocks of the footwall, then back-folded to the southeast. A phase of broad, regional "upwarping" in this area would then complete the deformation to produce the present structural configuration.

We emphasize that, at present, the exact position of the trace of the thrust fault remains uncertain along much of its length, especially in the area just northeast of Jessieville. It is anticipated that analyses, now in progress, of several samples of limestone collected across the proposed fault zone will yield microfossil data that will shed more light on the proper position of the fault trace.

DISCUSSION

Of primary importance in the overall structural and tectonic interpretation of the Jessieville area is the distribution of the Late Cambrian trilobite localities in the core area. These localities lie approximately within a "band" which follows the general folding pattern of the surrounding rock units in the area (Fig. 1). The rocks within this band were observed to be of very uniform lithology, usually consisting of lenses and layers of dense, gray or black limestone which are locally folded or boudined. The limestones are interbedded with uniformly light-gray or bluish-bray, calcareous shales and slates. The pale, bluish-gray color of the shales and slates is often rather striking in outcrop. Additionally, initial examinations of the trilobites from the scattered fossil localities indicate that the fossils may all come from the same one or two Late Cambrian trilobite zones (Stitt, personal commun.). Thus, a stratigraphic marker, albeit speculative at this time, may be present within the Collier Shale in this area. This idea may well be of use in mapping other areas in the Ouachitas and in the correlation of drill cores.

We note that the Late Cambrian ages and North American affinities of the trilobites place a significant constraint upon tectonic interpretations of rocks within the Benton uplift. Finally, we note that the rocks exposed nearer the center of the core than the Late Cambrian trilobites localities might well be the oldest strata in the Ouachita Mountains.

ACKNOWLEDGEMENTS

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REFERENCES CITED

Brown, S. H., 1982, Geology of the Mill Creek Mountain area, Garland County, Arkansas: [M.A. thesis] University of Missouri-Columbia, 71 p.

Ethington, R. L., 1984, Conodonts from Ordovician rocks, Ouachita Mountains, 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 GB 84-2, p. 93-98.

Haley, B. R. and others, 1976, Geologic map of Arkansas: U.S. Geological Survey Map, scale 1:500,000.

Hart, W. D., 1985, Structural geology of the north portion of the Goosepond Mountain quadrangle, Ouachita Mountains, Arkansas: [M.S. thesis] University of Missouri-Columbia, 95 p.

Kurtz, V. E., 1975, Franconian (Upper Cambrian) trilobite faunas from the Elvins Group of southeast Missouri: Journal of Paleontology, v. 49, p. 1009-1043.

Palmer, A. R., 1965, Trilobites of the Late Cambrian Pterocephaliid biomere in the Great Basin, United States: U.S. Geological Survey Professional Paper 493, 105 p.

Repetski, J. E., and Ethington, R. L., 1977, Conodonts from graptolite facies in the Ouachita Mountains, Arkansas and Oklahoma, in Stone, C. G., et al., eds., Symposium on the geology of the Ouachita Mountains, v. 1: Arkansas Geological Commission Miscellaneous Publication MP 13, p. 92-106.

Stitt, J. H., 1971, Late Cambrian and earliest Ordovician trilobites, Timbered Hills and lower Arbuckle Groups, western Arbuckle Mountains, Murray County, Oklahoma: Oklahoma Geological Survey Bulletin 110, 83 p.

__________, 1977, Late Cambrian and earliest Ordovician trilobites, Wichita Mountains area, Oklahoma: Oklahoma Geological Survey Bulletin 124, 79 p.

Wilson, J. L., 1949, The trilobite fauna in the Elvinia Zone in the basal Wilberns Limestone of Texas: Journal of Paleontology, v. 23, p. 25-44.

The problem of antivergent structures in the Ouachita thrust belt

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INTRODUCTION

Antivergent structural elements are those with senses of asymmetry opposed to the direction of regional tectonic transport. This paper is about the importance of such structures in interpreting the evolution of the Ouachita Mountains and about some of the difficulties in identifying true antivergent structures of regional and local significance.

Asymmetric folds, overturned in the direction of tectonic (thrust) transport, are important elements in the structural style of many subduction-related mountain belts. Examples can be found on most cross sections through orogens of this type: the Alps (see, for example, compilations of classical and modern cross sections in Rutten, 1969), Himalayas (Gansser, 1966), Appalachians (Rodgers, 1953; Gwinn, 1970; Cloos, 1971; Roeder and others, 1978; Woodward, 1985), Rocky Mountains (Fox, 1959; Bally and others, 1966; Roeder, 1967; Dahlstrom, 1970), and Ouachitas (Miser, 1929; Bush and others, 1977; Viele, 1979; Zimmerman and others, 1982; Arbenz, 1984), to mention only a few.

The directional sense of overturning of asymmetric folds, called *Vergenz* (now, *vergence*) by Stille (1924) has been used routinely, with certain restrictions, to infer thrust transport direction. The general sense of transport in the Ouachitas is south to north. Throughout most parts of the range, macroscopic and a majority of mesoscopic structural elements are north vergent, as expected. But field observations dating from the late nineteenth century (Griswold, 1892) to the present (Cambray and Welland, 1985) have implicitly or specifically cited a significant number of south-vergent folds. Not all such structures are important to regional reconstruction, but if sufficiently common and not readily identified as having been caused locally, antivergence may indicate important aspects of structural development such as retrocharriage or major south-directed thrusting. Such features would have to be incorporated in any model of the overall structural evolution of the range.

MONOCLINIC FOLDS AS INDICATORS OF THRUST TRANSPORT DIRECTION

The asymmetrical folds under discussion are those with monoclinic symmetry (Turner and Weiss, 1963). Their typical shapes in profile consist of relatively long back limbs at shallow angles to an underlying decollement surface and relatively short front (or middle) limbs that are steeply inclined to overturned.

Recent explanations for the development of monoclinic macrofolds in thrust terranes have used the work of Rich (1934) on the Cumberland thrust block as a point of departure. Rich offered an alternative to fold models based entirely on buckling that can accompany layer-parallel crustal shortening by suggesting that layers of rock involved in thrust transport were folded during passage up and over decollement steps (frontal ramps), a process called fault-bend folding by Suppe (1983). In their analytical treatment of Rich's original model, Berger and Johnson (1980) found that fault-bend fold asymmetry increased due to drag of the hanging wall cut-off against the ramp. Fischer and Coward (1982), on the other hand, concluded that folding occurs by layer-parallel shortening and buckling due to drag along the floor thrust prior to translation of the rocks up the frontal ramp. In either case, there is a marked tendency to produce asymmetrical macrofolds overturned in the direction of tectonic transport.

Small, monoclinic mesoscopic folds are also characteristic of thrust faulted terranes and have been used to infer transport direction in much the same way as macrostructures. Although there are numerous ways in which mesofolds can be produced in thrust sheets (see below), and various combinations of layer viscosities and orientations *vis a vis* the strain field that affect final asymmetry (Ghosh, 1966), the folds of immediate interest typically occur in zones dominated by simple shear strain that occur in well-bedded sedimentary rocks deformed at relatively shallow depths (Ramsay and Graham, 1970; Sanderson, 1979, 1982). In using mesofold asymmetry to reconstruct the slip lines and slip planes that describe thrust transport, care must be taken in the choice of structures. Hansen (1971) has established useful criteria for isolating synchronous, single-generation fold sets suitable for analysis, and Wheeler (1978) has illustrated the analytical techniques with practical examples.

A FEW WORDS ABOUT VERGENCE

Stille (1924) introduced the term *Vergenz* to express the directional sense of overturn of monoclinic folds and used their orientations to infer tectonic transport direction (de Sitter, 1964, p. 177). Although the idea of vergence is also applicable to intersecting, planar s-surfaces (bedding-cleavage, cleavage-cleavage), there are currently many more fold observations available from the Ouachitas, and this discussion will be limited to fold vergence. Modifying Stille's original usage, Roberts (1974) defined vergence as "...the horizontal direction, within the plane of the fold profile, towards which the upper component of ...[external] rotation is directed" (Fig. 1).

In his 1981 review of the definitions and current use (as well as misuse) of vergence, Bell underscored the practical value of the concept when correctly applied. Conversely, incorrect use of the term can lead to conceptually disastrous misinformation. Some of the difficulty in assessing the significance of antivergence is semantic rather than purely geological because operational definitions can vary markedly. For example, facing, which is the direction toward the upper surface of a layer (Shrock, 1948) or the direction along the axial surface trace of a fold in which the beds become younger (Shackelton, 1957), has been used as if synonymous with vergence although the two are independent (Fig. 1) and in some cases mutually exclusive (Bell, 1981).

Of the two fold symmetries most often encountered in the Ouachita range, orthorhombic and monoclinic, only the latter can be assigned a vergence direction. Vergence reflects the fundamental symmetry of a monoclinic structure: two-fold rotation axis normal to the single

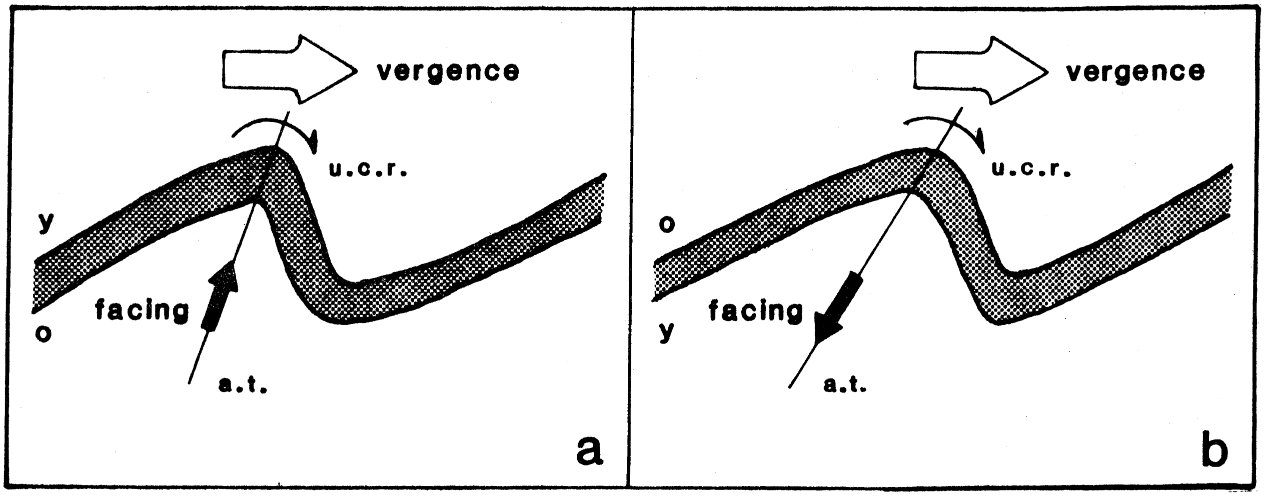

Figure 1. a. An asymmetrical fold set verges to the right (open arrow), following the direction of the upper component of external rotation (u.c.r.). Stratigraphic succession from older (o) to younger (y) beds is upright, and the fold faces upward and to the right, as defined by Shackleton (1957). The solid arrow along the trace of the axial surface (a.t.) shows the facing direction. b. The beds are overturned, and the folded sequence faces downward and to the left. The vergence direction has not been changed by overturning and is still to the right. Modified from Bell (1981).

m-plane. A rotated, north-vergent, monoclinic mesofold is still a north-vergent fold, because it is not possible to change vergence solely by rotation (Fig. 1). Symmetrical folds can be given some designation such as facing direction that reflects spatial orientation, but vergence is meaningless when applied to orthorhombic structures.

Vergence is best characterized by azimuth, which provides unequivocal orientation of the structure, except for vertically plunging folds in which case the term is again meaningless. Using azimuthal direction renders the fold independent of plunge direction which is not true for asymmetrical fold characterizations that use clockwise or counter-clockwise rotation sense alone or alphabetical designations ("S fold" or "Z fold"), both of which vary with plunge.

Although the vergence of a structural element is first and foremost a geometrical property, its cause can often be genetically linked to larger-scale kinematics. There are two common field applications of vergence: its use in a) finding the direction to the hinge of a larger fold from parasitic folds along its limbs, and b) inferring the direction of tectonic transport from the dominant vergence of major and minor folds developed synchronously with thrusting.

OUACHITA TECTONICS - A BRIEF OUTLINE

The main outlines of the tectonic history of the Ouachita Mountains were established during the 1970's and early 1980's and are generally agreed upon. Various developmental models have been presented, discussed or summarized by Viele (1973), Morris (1974), Briggs and Roeder (1975), Wickham, Roeder and Briggs (1976), Stone and Bush (1984), Arbenz (1984), Houseknecht and Matthews (1985), and Kruger and Keller (1986).

The most widely accepted models call for south-directed subduction and the closing of an ocean basin in which pre-flysch Ouachita facies sediments had been deposited in slope, rise, and abyssal environments during early to middle Paleozoic time. Influx of a thick wedge of Carboniferous flysch was followed by collision of the passive, Atlantic-type margin along the southern flank of the North American craton (lower plate) with a subduction complex that had developed south of the subduction zone along the northern margin of the upper plate.

From observations of thrust directions on the inner walls of trenches (Seely and others, 1974), it seems likely that synthetic (Roeder, 1973), non-vergent thrust faults and asymmetric folds had been produced in the accretionary wedge prior to collision. The direction of thrust transport was maintained and its distance substantially increased throughout the middle Pennsylvanian (Atokan) collision events during which the subduction complex was carried north and northwestward across part of the margin of the North American craton. The collision marks that period in the deformation history of the Ouachita range during which the major structural elements were developed. Because the series of events involved northward tectonic transport, the complex sets of structures produced must have been dominated by northward vergence.

Subsequent to their formation, many north-verging structures, including thrust faults, macroscopic folds, and mesoscopic folds were rotated into apparent southward (anti-) vergence. Distinction between these and true antivergent structures is of primary importance in field investigations.

CAUSES OF LOCAL ANTIVERGENCE

Some of the ways in which true antivergent folds can be produced are discussed in this section. The list is by no means exhaustive. Only those causes that have been identified or specifically suggested as having occurred during deformation of the Ouachitas are included.

Prelithification (Slump) Folds

Monoclinic mesofolds can be produced during mass movement of unstable, poorly-consolidated sediment down submarine paleoslopes. Slump folds are common in the Stanley Shale and younger rocks of the Ouachita flysch facies (Morris, 1971). Many of the south-verging folds reported from pre-flysch rocks have also been attributed to slump. Although folds of this type are undoubtedly present, their number, as well as the importance of prelithification mass movement to fold production in rocks older than Carboniferous, may have been overestimated. Slump folds are often difficult to recognize, particularly in areas of limited exposure typical of many localities in the Ouachitas. Criteria such as those proposed by Helwig (1970) and others for their identification should be applied in as strict a manner as possible before attributing a prelithification origin to antivergent folds. It is also worthwhile to note that slump folds in the Ouachitas can verge northward as well as southward as reported by Weber (1986, personal communication) from the Cossatot Mountains.

Conjugate Kink Folds

Several small and at least one large, south-vergent, mesoscopic kink folds are exposed among the more typical rotated, north-vergent kinks in the lower and middle members of the Arkansas Novaculite at Caddo Gap. Both sets of structures are geometrically identical and appear to have been formed synchronously, the antivergent folds being interpreted as conjugate to the north-vergent set (Zimmerman and Ford, in press). Kink folds are not uncommon in other stratigraphic units (the Bigfork Chert, for example), and this cause of antivergent folding may be widely applicable.

Parasitic and Related Mesofolds

Parasitic ("drag") folds can be produced by localized simple shear along the limbs of larger flexural folds. The fact that the sense of asymmetry of parasitic folds is oppositely directed on opposing limbs of the main structure has been used to infer the direction of anticlinal (or synclinal) hinges in areas of limited exposure. Parasitic folds on the northern limbs of north-verging anticlines can be expected to verge southward.

A kinematically equivalent situation arises when rotation of the short forelimb of an asymmetrical macrofold produces both layer-parallel shortening and simple shear that is opposed to the regional shear sense (Ghosh, 1966; Sanderson, 1979), a process that can cause rotated antivergent mesofolds to form in the competent members of a multilayer. Sanderson (1982) has also suggested that the sense of incremental shear in part of a fault-bend fold, deformed by flexural flow, can be reversed into antivergence during thrust transport from a footwall ramp to an upper flat.

Owing to the number of thrust-related macrofolds known to exist in the Ouachitas, classical parasitic folding or kinematically similar variations should be considered early in any investigation of antivergence.

Overturned Limbs of Fold Nappes

The occurrence of fold nappes in the eastern Ouachitas has been proposed by Viele (1973). Although suggestions of nappe-like fold geometry appear on maps of other parts of the range, the presence of large, complex, recumbent structures similar to those of the Helvetic belt of Switzerland has never been firmly established.

In their model of the development of the Morcle nappe, Ramsay and others (1983, Fig. 4) illustrate the formation of strongly attenuated, antivergent folds in the overturned lower limb, caused by rotation of the enveloping surface of a previously formed fold set from the shortening into the extensional field of applied deformation. Rocks folded in this manner typically show evidence of considerable strain which, combined with the inverted stratigraphic sequence of the overturned limb, should permit relatively easy field identification.

Fan-shaped Macrofolds

According to the interpretation of Miser and Purdue (1929), fan-shaped macrofolds with smaller, antivergent folds on their northern (primarily) limbs occur in the central and western Cossatot Mountains. In the eastern Cossatots, however, Weber (1986, personal communication) has found that rotated, north-verging kinks define the macrofold style. There is some evidence that southward (clockwise, looking east) rotation may increase toward the central parts of the range. Additional detailed work in this area is necessary if the fan fold geometry illustrated by Miser and Purdue (1929, Fig. 6 and Pl. 16) is to be confirmed.

Folds Resulting from Southward Thrusting

One of the simplest ways to produce true south-verging folds is by simple shear related to south directed thrust transport. Back-thrusts or antithetic thrusts of varied origin are not uncommon in fold-thrust belts (Gwinn, 1970; Roeder, 1973; Butler, 1982). Such structures tend to be localized, and the distribution of related antivergent folds should not be penetrative on a regional scale.

The possibility of major, as opposed to local or secondary, south-directed thrust transport in the Arkansas Ouachitas seems remote. Minor, southward back-thrusting can be accommodated in terrane with an overall northward vergence, but regional scale tectonic transport from north to south is incompatible with current views concerning the evolution of the orogen.

In the southern part of the Broken Bow uplift, however, both earlier (Honess, 1923) and more modern work (Feenstra and Wickham, 1975; Nielsen, 1982) suggest that the typical southward fold vergence in that area is most compatible with large-scale simple shear. Although various mechanisms such as northward underthrusting of basement (Feenstra and Wickham, 1975), and regional retrocharriage (Roeder, 1973; Briggs and Roeder, 1975) have been proposed, the divergent senses of thrust transport that appear to exist between the southern Broken Bow and the Benton uplifts remains one of the major unsolved problems of Ouachita tectonics.

Misidentified North-Verging Folds

The folds in this category (of the type clearly illustrated in the Arkansas Novaculite at Caddo Gap) are not true antivergent structures but primary, north-verging folds that have been rotated so that their hinge surfaces currently dip to the north rather than to the south. Such folds can be misidentified as antivergent if the operative definition of the term is incorrectly based on hinge surface attitude or facing direction rather than on fold symmetry. Another all-too-common opportunity for misidentification can occur when only the hinge area of a mesofold is exposed. In such case it is probable that neither fold symmetry nor vergence can be specified (Hills, 1972).

APPLICATION TO REGIONAL MODELS

One of the most striking regional characteristics of Ouachita structural geology is the southward rotation of originally north-verging folds, a process generally assigned to a late tectonic stage by most workers. Of the models that have been proposed to account for this phenomenon, two of the most reasonable are regional-scale rotation of most of the Paleozoic section, accompanied by back-thrusting (rétrocharriage), and deformation beneath an "orogenic lid" or *traineau écraseur* (Laubscher, 1983). Careful analysis of fold vergence and mesofold origin should help evaluate the relative merits of the two models.

In the case of rétrocharriage of a sequence of thrust sheets originally transported form south to north, most mesofolds should be rotated, north-vergent structures, and true south-verging folds should fit into one of the categories discussed above. Deformation beneath an orogenic lid, on the other hand, could be expected to produce a large number of relatively late-stage antivergent structures whose distribution is penetrative on a regional scale.

There are difficulties with the application of either model to the Ouachita belt. Rétrocharriage on the required scale has been typically associated with complex interaction between lithospheric plates, such as change of subduction polarity (flip) prior to final continental collision (Roeder, 1973), or simultaneous convergence of plates across an ocean basin flanked by subduction zones with opposed polarity (Roeder, 1976). Neither scenario appears to be entirely compatible with the regional geometry and tectonic event sequence of the Ouachita belt (Wickham and others, 1976), a difficulty shared by the northward underthrusting of basement proposed by Feenstra and Wickham (1975) and strike-slip faulting of basement rocks underlying the Broken Bow uplift (Briggs and Roeder, 1975).

Arbenz (1984) suggested that regional-scale, south-directed simple shear was produced in lower to middle Paleozoic (pre-flysch) rocks, sandwiched between a relatively rigid cap of Carboniferous and younger strata (now eroded) and active, north-verging thrust faults in a basement duplex beneath the Benton uplift. Because this event was to have occurred late in the tectonic evolution of the mountain belt, both rotation of older, north-verging mesofolds and formation of younger, antivergent structures are accounted for. A necessary consequence of the process, however, is superposition of regionally penetrative south-verging folds on the earlier north-verging structures, resulting in polyphase fold systems with triclinic symmetry and the correct sequence of vergence alteration. Complex fold sets of this type have not been widely reported in the increasing number of detailed structural studies that have been completed over much of the Ouachitas during the past decade. In addition, the proposed orogenic lid has been heavily deformed, and its suggested role as a relatively rigid layer appears open to question.

A third alternative, progressive southward rotation of stacked thrust sheets during transport up a succession of frontal ramps (Suppe, 1983) should be considered on an equal footing with other models. The primary purpose of this paper, however, has not been to evaluate current ideas about the late-stage tectonic development of the Ouachitas, but to suggest some ways in which they can be tested by field observation.

CONCLUSIONS

The concepts of vergence and antivergence can be of considerable value to both the field geologist and the tectonic model builder provided that they are correctly and consistently applied. Their additional application to sets of intersecting s-surfaces would provide needed augmentation of the fold data currently on hand.

Every effort should be made to determine the origins of the antivergent structures that are encountered in any detailed structural investigation of the Ouachita Mountains. Those whose formation can be traced to local, restricted strain fields may not be regionally significant, but others, taken in bulk, could provide some of the keys necessary for better evaluation of existing tectonic models.

REFERENCES CITED

Arbenz, J. K., 1984, A structural cross section through the Ouachita Mountains of western Arkansas, in, Stone, C. G., and Haley, B. R., A guidebook to the geology of the central and southern Ouachita Mountains, Arkansas: Arkansas Geological Commission Guidebook 84-2, p. 76-82.

Bally, A. W., Gordy, P. L., and Stewart, G. A., 1966, Structure, seismic data, and orogenic evolution of southern Canadian Rocky Mountains: Bulletin of Canadian Petroleum Geology, v. 14, no. 3, p. 337-381.

Bell, A. M., 1981, Vergence: an evaluation: Journal of Structural Geology, v. 3, no. 3, p. 197-202.

Berger, Philip, and Johnson, A. M., 1980, First-order analysis of deformation of a thrust sheet moving over a ramp: Tectonophysics, v. 70, p. T9-T24.

Briggs, Garrett, and Roeder, D. H., 1975, Sedimentation and plate tectonics, Ouachita Mountains and Arkoma Basin, in, Briggs, G., McBride, E. F., and Moiola, R. J., Sedimentology of Paleozoic flysch and associated deposits, Ouachita Mountains - Arkoma Basin, Oklahoma: Dallas Geological Society, p. 1-22.

Bush, W. V., Haley, B. R., Stone, C. G., Holbrook, D. F., and McFarland, J. D., III, 1977, A guidebook to the geology of the Arkansas Paleozoic area: Little Rock, Arkansas Geological Commission, 79 p.

Butler, R. W. H., 1982, The terminology of structures in thrust belts: Journal of Structural Geology, v. 4, p. 239-245.

Cambray, F. W., and Welland, M. J. P., 1985, Southward verging structures and coaxial refolding in the Benton Uplift, Ouachita Mts., Arkansas: a result of southerly directed thrusting: Geological Society of America Abstracts with Programs, v. 17, no. 7, p. 538.

Cloos, Ernst, 1971, Microtectonics along the western edge of the Blue Ridge, Maryland and Virginia: The Johns Hopkins University Studies in Geology, no. 20, 234 p.

Dahlstrom, C. D. A., 1970, Structural geology in the eastern margin of the Canadian Rocky Mountains: Bull. Canadian Petroleum Geology, v. 18, p. 332-406.

Fischer, M. W., and Coward, M. P., 1982, Strains and folds within thrust sheets: an analysis of the Heilam sheet, northwest Scotland: Tectonophysics, v. 88, p. 291-312.

Feenstra, Roger, and Wickham, John, 1975, Evolution of folds around Broken Bow uplift, Ouachita Mountains, southeastern Oklahoma: American Assoc. Petroleum Geologists Bull., v. 59, p. 974-985.

Fox, F. G., 1959, Structure and accumulation of hydrocarbons in southern foothills, Alberta, Canada: American Assoc. Petroleum Geologists Bull., v. 43/5, p. 992-1025.

Gansser, Augusto, 1964, Geology of the Himalayas: London, Interscience Publishers, 289 p.

Ghosh, S. K., 1966, Experimental tests of buckling folds in relation to strain ellipsoid in simple shear deformations: Tectonophysics, v. 3, p. 169-185.

Griswold, L. S., 1892, Whetstones and the novaculites of Arkansas: Geological Survey of Arkansas, Annual Report for 1890, v. III, 443 p.

Gwinn, V. E., 1970, Kinematic patterns and estimates of lateral shortening, Valley and Ridge and Great Valley provinces, central Appalachians, south-central Pennsylvania, in, Fisher, G. W., and others, eds., Studies of Appalachian geology: central and southern (Cloos Volume): New York, Interscience Publishers, p. 127-146.

Hansen, E., 1971, Strain facies: New York, Springer-Verlag, 207 p.

Helwig, James, 1970, Slump folds and early structures, northeastern Newfoundland Appalachians: Journal of Geology, v. 78, p. 172-187.

Hills, E. S., 1972, Elements of structural geology (2nd ed.): New York, John Wiley & Sons, 502 p.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geological Survey Bull. 32, Pt. 1, 278 p.

Houseknecht, D. W., and Matthews, S. M., 1985, Thermal maturity of Carboniferous strata, Ouachita Mountains: American Assoc. Petroleum Geologists Bull., v. 69/3, p. 335-345.

Kruger, J. M., and keller, G. R., 1986, Interpretation of crustal structure from regional gravity anomalies, Ouachita Mountains area and Gulf Coastal Plain: American Assoc. Petroleum Geologists Bull., v. 70, p. 667-689.

Laubscher, H. P., 1983, Detachment, shear, and compression in the central Alps, in, Hatcher, R. D., Jr., and others, eds., Contributions to the tectonics and geophysics of mountain chains: Geological Society of America Mem. 158, p. 191-211.

Miser, H. D., 1929, Structure of the Ouachita Mountains of Oklahoma and Arkansas: Oklahoma Geological Survey Bull. No. 50, 30 p.

Miser, H. D., and Purdue, A. H., 1929, Geology of the De Queen and Caddo Gap quadrangles, Arkansas: U.S. Geological Survey Bull. 808, 195 p.

Morris, R. C., 1971, Classification and interpretation of disturbed bedding types in Jackfork flysch rocks (Upper Mississippian), Ouachita Mountains, Arkansas: Journal of Sedimentary Petrology, v. 41, p. 410-424.

__________, 1974, Sedimentary and tectonic history of the Ouachita Mountains, in, Dickenson, W. R., ed., Tectonics and sedimentation: S.E.P.M. Special Publication No. 22, p. 120-142.

Nielsen, K. C., ed., 1982, Structural styles of the Ouachita Mountains, southeastern Oklahoma: Geological Society of America South-Central Section Field Trip #3, 124 p.

Ramsay, J. G., and Graham, R. H., 1970, Strain variations in shear belts: Canadian Jour. Earth Science, v. 7, p. 786-813.

Ramsay, J. G., Casey, Martin, and Kligfield, Roy, 1983, Role of shear in development of the Helvetic fold-thrust belt of Switzerland: Geology, v. 11, p. 439-442.

Rich, J. L., 1934, Mechanics of low-angle overthrust faulting as illustrated by Cumberland thrust block, Virginia, Kentucky, and Tennessee: American Assoc. Petroleum Geologists Bull., v. 18, p. 1584-1596.

Roberts, J. L., 1974, The structure of the Dalradian rocks in the SW Highlands of Scotland: Journal of the Geological Society of London, v. 130, pt. 2, p. 93-124.

Rodgers, John, 1953, The folds and faults of the Appalachian Valley and Ridge province: Kentucky Geological Survey Spec. Pub. 1, p. 150-166.

Roeder, D. H., 1967, Rocky Mountains: Berlin, Gebruder Borntraeger, 318 p.

__________, 1973, Subduction and orogeny: Journal of Geophysical Research, v. 78, p. 5005-5024.

__________, 1976, Continental convergence in the Alps: Tectonophysics, v. 40, p. 339-350.

Roeder, Dietrich, Gilbert, O. E., Jr., and Witherspoon, W. D., 1978, Evolution and macroscopic structure of Valley and Ridge thrust belt, Tennessee and Virginia: Univ. of Tennessee Dept. of Geol. Sciences Studies in Geology 2, 25 p.

Rutten, M. G., 1969, The geology of western Europe: Amsterdam, Elsevier Publishing Co., 529 p.

Sanderson, D. J., 1979, The transition from upright to recumbent folding in the Variscan fold belt of southwest England: a model based on the kinematics of simple shear: Journal of Structural Geology, v. 1, p. 171-180.

__________, 1982, Models of strain variation in nappes and thrust sheets: a review: Tectonophysics, v. 88 p. 201-233.

Seely, D. R., Vail, P. R., and Walton, G. G., 1974, Trench slope model, in, Burk, C. A., and Drake, C. L., The geology of continental margins: New York, Springer-Verlag, p. 249-260.

Sitter, L. U. de, 1964, Structural geology, 2nd Ed.: New York, McGraw-Hill Book Co., 551 p.

Stille, Hans, 1924, Grundfragen der vergleichenden Tektonik: Berlin, Gebrüder Borntraeger, 443 p.

Stone, C. G., and Bush, W. V., 1984, Summary of the geology of the central and southern Ouachita Mountains, 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. 65-75.

Suppe, John, 1983, Geometry and kinematics of fault-bend folding: American Journal of Science, v. 283, p. 684-721.

Turner, F. J., and Weiss, L. E., 1963, Structural analysis of metamorphic tectonites: New York, McGraw-Hill

Book Co., 545 p.

Viele, G. W., 1973, Structure and tectonic history of the Ouachita Mountains, Arkansas, in, De Jong, K. A., and Scholten, Robert, eds., Gravity and tectonics: New York, John Wiley & Sons, p. 361-377.

__________, 1979, Geologic map and cross section, eastern Ouachita Mountains, Arkansas: Geological Society of America Map and Chart Series MC-28F, scale, 1:250,000.

Wheeler, R. L., 1978, Slip planes from Devonian Millboro Shale, Appalachian Plateau Province: statistical extensions of discfold analysis: American Journal of Science, v. 278, p. 497-517.

Wickham, John, Roeder, Dietrich, and Briggs, Garrett, 1976, Plate tectonics models for the Ouachita foldbelt: Geology, v. 4, p. 173-176.

Woodward, N. B., ed., 1985, Valley and Ridge thrush belt: balanced structural sections, Pennsylvania to Alabama: Univ. of Tennessee Dept. Geol. Sciences Studies in Geology 12, 64 p.

Zimmerman, Jay, and Ford, J. T., in press, Lower Stanley Shale and Arkansas Novaculite, western Mazarn Basin and Caddo Gap, Ouachita Mountains, Arkansas: Geological Society of America (DNAG) Field Guide to the South Central U.S.

Zimmerman, Jay, Roeder, D. H., Morris, R. C., and Evansin, D. P., 1982, Geologic section across the Ouachita Mountains, western Arkansas: Geological Society of America Map and Chart Series, MC-28Q, scale, 1:250,000.

A preliminary report on the metagabbros of the Ouachita core

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ABSTRACT

Deformed, greenschist-facies metagabbros occur at two widely separated localities in the core of the Ouachita Mountains. Mildly alkalic metagabbro containing relict titanaugite is found at Hominy Hill in the Benton uplift 20 km west of Little Rock, Arkansas. This rock has been dated at 1025 ± 48 Ma by whole-rock K/Ar. Another, even more deformed metagabbro is 215 km to the southwest at Colbert Creek in the Broken Bow uplift in Oklahoma. Both metagabbros appear to be tectonically enclosed within Womble Shale near its contact with overlying Bigfork Chert. The metagabbros of the Ouachita core may be related to early rifting or transform activity in the early Ouachita trough.

INTRODUCTION

Two occurrences of deformed, greenschist-facies metagabbro have been documented in the core of the Ouachita Mountains (Fig. 1). One is at Hominy Hill 20 km west of Little Rock (Williams, 1891), and the other is exposed along Colbert Creek 18 km west of Broken Bow, Oklahoma (Honess, 1923).

HOMINY HILL

The metagabbro of Hominy Hill, discovered by John C. Branner and first reported by Williams (1891), has very limited outcrop. It extends as boudins or 25-meter-wide en enchelon pods for approximately 1.5 km along a NW-SE trend. Its contacts with sedimentary "country rock," (shale and siltstone), observed in an excavation which has since been covered by a lake, commonly appeared sheared and parallel or subparallel to country-rock foliation. The metagabbro is crosscut by an unmetamorphosed Cretaceous lamprophyre dike and is also dissected by essentially undeformed late Paleozoic quartz veins. The

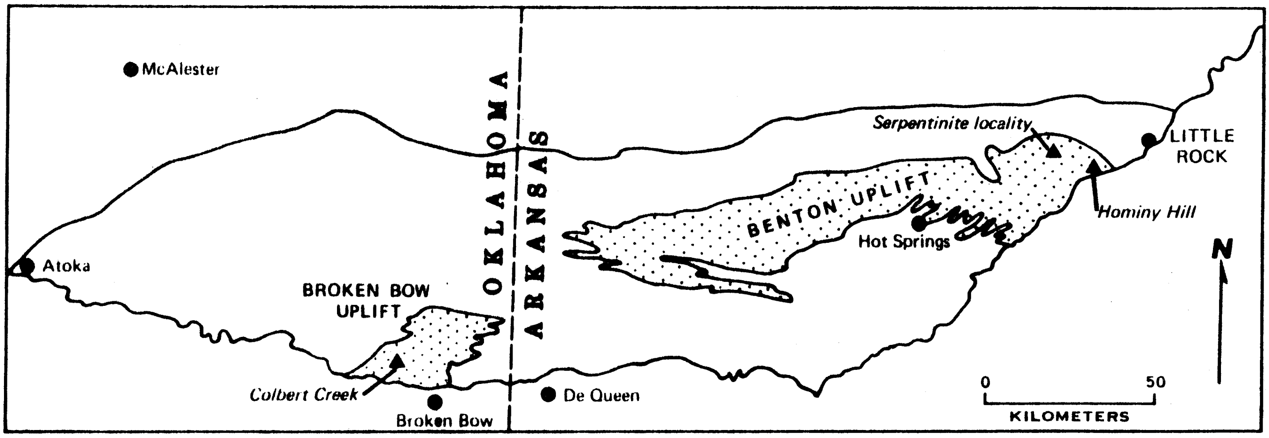

Figure 1. Sketch map showing location (triangles) of metagabbro and serpentinite bodies in the Ouachita Mountains of Arkansas and Oklahoma. Extent of the Ouachita Mountains shown by outline. Benton and Broken Bow uplifts (pre-Carboniferous sedimentary rocks) shown by stippled patterns.

age obtained by K/Ar on a whole-rock sample of fresh metagabbro is 1025 ± 48 Ma (Krueger, Geochron Labs, analyst). The actual age of the rock may be slightly younger due to low K abundances and small amounts of excess Ar (40* Ar/Total Ar = 0.494 - 0.580), but the rock is almost certainly late Precambrian.

The metagabbro of Hominy Hill has been deformed and metamorphosed to greenschist-facies. Its original texture was probably diabasic to granular. Plagioclase is broken and is now albite (Table 1). Relict titanaugite is rimmed by actinolite. Epidote occurs throughout the rock, but is especially well developed as replacement of some plagioclase. Chlorite is ubiquitous. The greenschist minerals are relatively undeformed, so metamorphism post-dated appreciable tectonism.

Mineralogical and geochemical data clearly indicate that the metagabbro of Hominy Hill was an alkalic to mildly alkalic rock. It contains abundant (up to 15 modal percent), well-preserved titanaugite (2.3% TiO_2) which plots in alkalic fields on all pyroxene discriminant diagrams (Figures 2 and 3). Whole rock analyses (Table 2) indicate that the rock contains about 2 percent TiO_2, "gabbroic" abundances of Ni and Cr, and is enriched in Zr. Because of the pervasive metamorphism and spilitization of the metagabbro, whole-rock discriminant diagrams were deemed unreliable, and, in fact, provided a very broad range of "results". However, the elevated abundances of relatively immobile high-charge, small-radius elements such as TiO_2, Nb, and Zr, in concert with the alkalic nature of the relict titanaugite, are good evidence for a mildly alkalic character, perhaps similar to EMORB (Figure 4). Rare earth elements have a flat, 25x chondrite pattern which shows slight La and Eu depletion (Figure 5). This pattern is similar to basalts from transform and other mildly alkalic, oceanic settings.

Table 1. Mineral compositions, metagabbro of Hominy Hill.

SAMPLE MINERAL	85 100 PYROXENE	85 100 PLAGIOCLASE	85 100 ACTINOLITE	85 102 EPIDOTE
SiO_2 (Wt. %)	48.42	67.90	55.87	38.51
TiO_2	2.52	----	0.00	0.02
Al_2O_3	3.67	20.85	2.16	26.73
FeO*	11.09	0.11	12.45	7.53
MgO	12.35	----	14.42	0.24
CaO	20.85	0.38	11.83	23.17
Na_2O	0.50	11.23	0.96	0.02
K_2O	0.00	0.04	0.02	0.04
TOTAL	99.64	100.53	97.82	96.25
Cr_2O_3 (Wt. %)	0.10	----	----	0.04
NiO	0	----	----	0.07
BeO	----	0.03	----	----
An*	----	0.5	----	----

*An = Anorthite

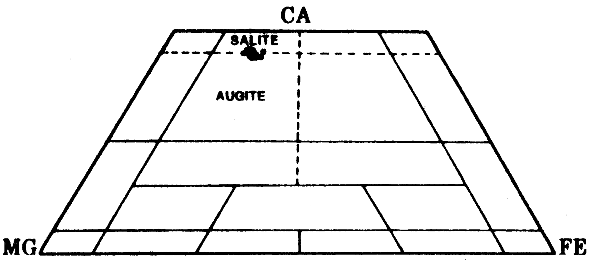

Figure 2. Compositional plot for clinopyroxenes of the gabbro of Hominy Hill. Analyses straddle fields for augite and salite, and are typical of mildly alkaline gabbros or basalts.

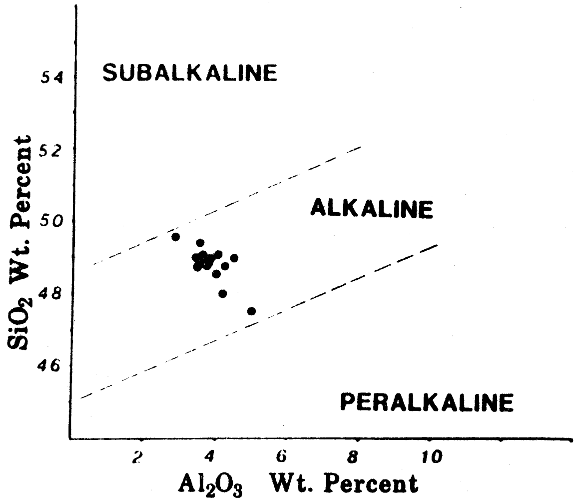

Figure 3. SiO_2 vs Al_2O_3 plot for clinopyroxenes of the Hominy Hill gabbro. Analyses fall within the alkalic field.

COLBERT CREEK

The metagabbro of Colbert Creek, first reported by Honess (1923), is an extremely sheared rock which occurs as isolated pods 3 to 10 m wide along a 1-km-long, NE-SW trend in the southernmost Ouachita core. Some of the adjacent shale has been tectonically enclosed within the meta-igneous rock along the sheared margins

Table 2. Whole rock geochemistry, Ouachita Mountain metagabbros at Hominy Hill, Arkansas and Colbert Creek, Oklahoma. Major-element oxides in weight-percent; trace element abundances in ppm.

	COLBERT CREEK	HOMINY HILL		
SAMPLE	---	E100	E101	E102
SiO_2	47.44	48.70	49.50	49.50
TiO_2	0.89	2.05	2.05	2.03
Al_2O_3	16.59	14.60	14.70	14.70
Fe_2O_3	0.00	10.60	10.50	10.40
FeO	4.87	0.00	0.00	0.00
MnO	0.10	0.19	0.19	0.19
MgO	8.29	6.43	6.32	6.32
CaO	6.02	6.92	7.24	7.19
Na_2O	5.84	4.58	4.69	4.67
K_2O	0.53	0.05	0.06	0.06
P_2O_5	0.13	0.23	0.22	0.23
H_2O+	0.12	3.50	3.60	3.24
H_2O-	0.12	0.10	0.23	0.12
TOTAL	90.94	97.95	98.29	97.65
Rb	----	10	20	10
Ba	----	330	350	340
Sr	----	900	940	950
V	---	260	250	250
Cr	----	180	180	190
Ni	----	88	77	78
Zr	----	190	170	170
F/F+M	0.375	0.602	0.604	0.602
Rb/Sr	0.000	0.011	0.021	0.011
K/Rb	0	42	21	50
K/Ba	0.0	1.3	1.2	1.2
density	2.57	2.55	2.55	2.55

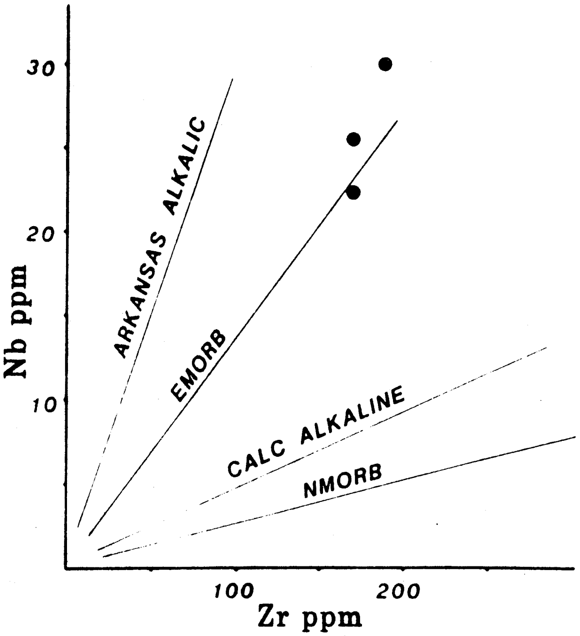

Figure 4. Plot of Zr vs Nb for whole-rock analyses of the Hominy Hill gabbro. Analyses fall on an EMORB (enriched mid-ocean ridge basalt) trend.

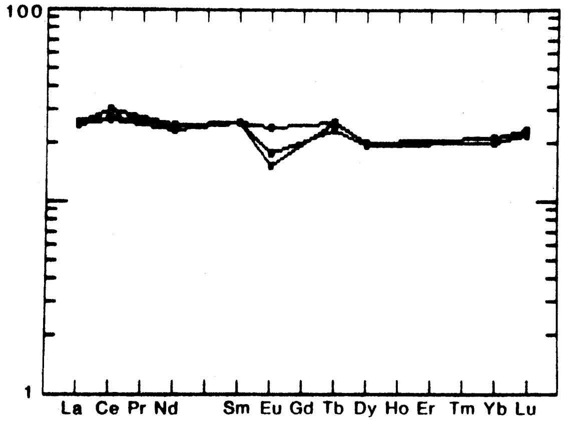

Figure 5. Chondrite-normalized plot of rare-earth elements for whole-rock analyses of the Hominy Hill gabbro.

of these pods.

This metagabbro is also in greenschist-facies. Its original texture was probably granular to hypidiomorphic. Its present fabric is sheared to mylonitic, with rare enclaves of relatively undeformed rock enclosed by zones of intense shearing. Chlorite, epidote, and actinolite are present. Calcite is abundant as a metamorphic and secondary replacement mineral. Microprobe analyses are in progress on plagioclase (albite?) and very small, faintly purple pyroxenes (titanaugite?) in the most undeformed samples.

The samples of the gabbro of Colbert Creek that have been analyzed to date indicate a rock with significantly less TiO_2 than the Hominy Hill gabbro (Table 2). Major and trace element work is in progress on more reliable, representative rocks, however, and a more confident interpretation of the nature of the gabbro of Colbert Creek must await these results.

DISCUSSION

The metagabbros of the Ouachita core are probably related to one another and to the serpentinite pods (Fig. 1) of the eastern Benton uplift discussed by Cox (1986; also this volume). The serpentinite, like the metagabbro of Hominy Hill, may be slightly alkalic in nature (Mullen, 1984). All three lithologies appear to be tectonically emplaced, and all three occur in a similar stratigraphic interval, that is, in the upper Womble Shale near the contact with the overlying Bigfork Chert.

Further work is needed before these rocks can be interpreted with confidence. Their apparently alkalic nature suggests that they are related to transform motion or rifting in the early Ouachita trough, or represent transform-related (EMORB) fragments of oceanic crust which served as basement. Additional geochemistry, dating, and field exploration are in progress to make them a viable piece of the Ouachita puzzle.

ACKNOWLEDGEMENTS

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REFERENCES

Cox, T. L., 1986, A structural study of the serpentinites of the Benton uplift, Saline County, Arkansas [M.S. thesis]: Fayetteville, Arkansas, University of Arkansas, 97 p.

Honess, C. W., 1923, Geology of the southern Ouachita Mountains of Oklahoma. Part I. Stratigraphy, structure, and physiographic history: Oklahoma Geological Survey Bull. 32, 278 pp.

Mullen, Ellen, 1984, Ultramafic pods of the eastern Ouachitas: Ultramafic or alkalic?: Geological Society of America Abstracts with Programs, v. 14, p. 573.

Williams, J. F., 1891, The igneous rocks of Arkansas: Arkansas Geological Survey Annual Rept. for 1890, v. 2, 457 p.

The syenites of Granite Mountain, Arkansas: A progress report

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ABSTRACT

Four distinctive phases of the syenites of Granite Mountain, Pulaski County, Arkansas have been defined on the basis of field mapping, petrography, microprobe studies, geochemistry, and fission-track dating. The oldest rocks are olivine syenites which contain small amounts of olivine (Fo_{70}), plagioclase (An_{40}), and ilmenite. These olivine syenites have an 87/86 Sr ratio of 0.70359, an age of about 100 Ma, and constitute less than 5% of the exposed Granite Mountain pluton. Later phases of the Granite Mountain syenites have ages of 89-85 Ma, significantly higher 87/86 Sr (0.70513 to 0.71980), and exhibit gradational contacts with olivine syenite, although olivine and plagioclase are absent. These rocks seem to represent potassic metasomatism of the original intrusive which may have been an olivine gabbro, as well as later magmatic activity. Xenoliths in the Granite Mountain pluton represent country rock, cumulate, probably immiscible material, and possible clasts of potassic, highly metasomatized mantle source.

INTRODUCTION

The syenites of Granite Mountain, Pulaski County, Arkansas are at the northeastern end of the exposed Cretaceous Arkansas alkalic province, and seem to be the most leucocratic rocks of a sequence which progresses along the West Gulf Coastal Plain-Ouachita Mountain boundary from the diamondiferous lamproites at Prairie Creek (Murfreesboro) to the Magnet Cove ring dike complex, to the Benton County lamprophyres, and finally to syenites in Saline and Pulaski Counties (Fig. 1). The rocks of the Arkansas alkalic province range in age from 106 Ma at Prairie Creek to 85 Ma in Granite Mountain. Isotopic data (Tilton et al., 1987; Morris and Eby, 1986) indicate that the province is mantle-derived with minor input from a crustal component; geochemical modeling indicates that multiple sources were involved in the generation of the magmas (Mullen and Petty, 1985).

The Granite Mountain pluton, as exposed, is an epizonal, ring-shaped intrusion approximately 4 km long and 2 km across, located at the southern edge of Little Rock. It is actively quarried for industrial/commercial use, yielding excellent, fresh exposures for geologic study. This paper will present a summary of work in progress on the Granite Mountain pluton, including field, mineralogical, and geochemical findings. Its conclusions may be further revised in light of new data. This is intended as a progress report rather than a final presentation.

SYENITE UNITS

The syenites of Granite Mountain have been subdivided into four units, based upon field appearance, mineralogy, and chemical and isotopic composition (Table 1). Although these units may be mapped, they are intimately mixed on a centimeter to millimeter scale, have gradational contacts or may grade into and out of one another without ever achieving the "end-member unit". In general they cannot be mapped as simple entities. The Granite Mountain pluton is a complex mixture of these units, rather than a logically zoned intrusion. The four units (olivine syenite, Phase I, Phase II, Phase III) will be discussed in order of decreasing age.

Olivine Syenite

Olivine syenite occurs as isolated, 1- to 2-m patches throughout the Granite Mountain pluton. It is found in the chilled margin as well as in the interior of the body, and appears to be distributed randomly. It is very dark gray to blue-gray, with no hint of white or lighter gray

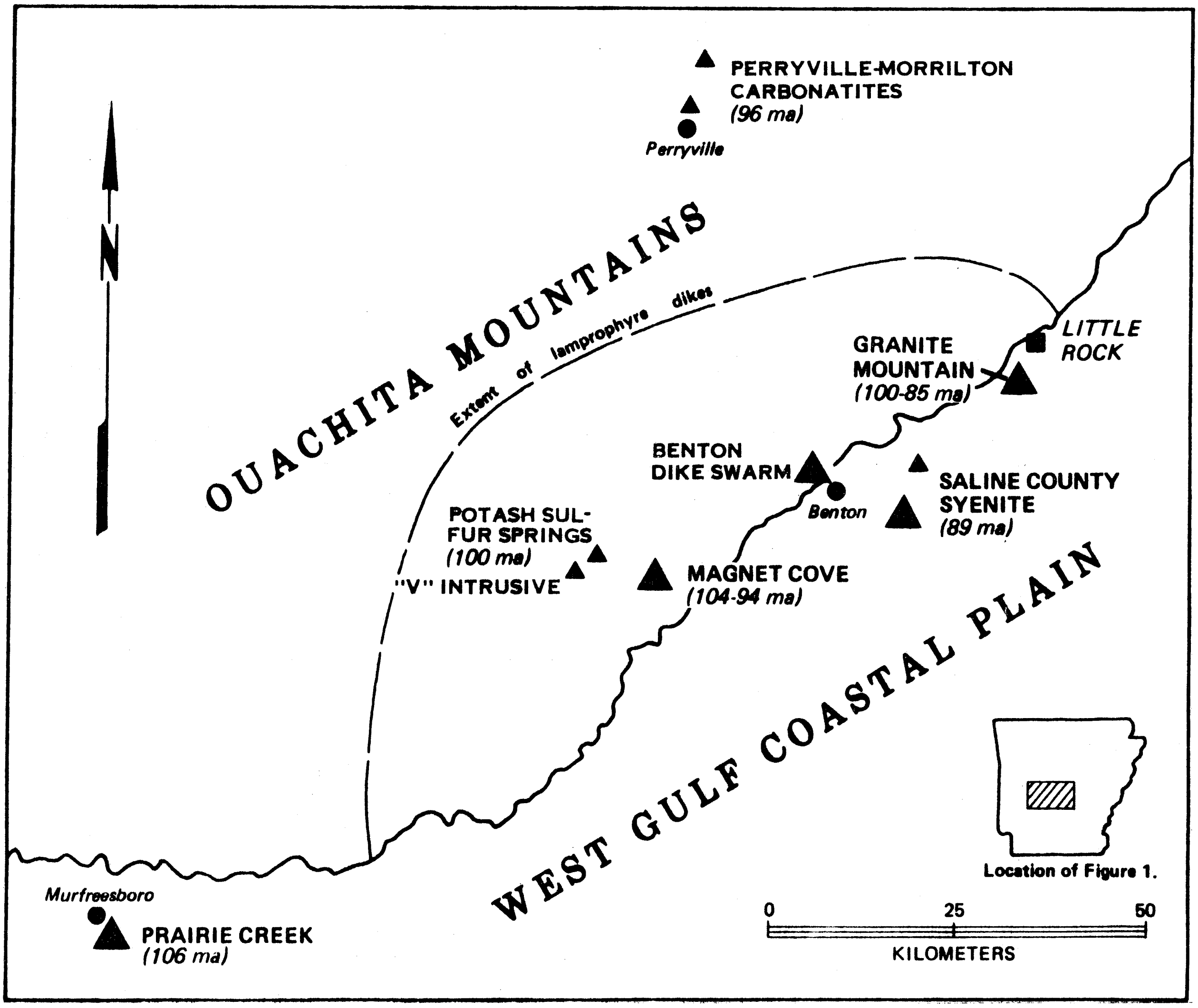

Figure 1. Distribution and fission track ages of principal intrusions of the Arkansas alkalic province.

coloration, commonly has a finer-grained appearance than the surrounding syenite, and grades into the Phase I unit. The olivine syenite comprises less than 5 percent of the Granite Mountain pluton.

The olivine syenite contains small amounts of olivine and plagioclase in addition to minerals more characteristic of syenite and/or pulaskite. Olivine (Fo_{70}) (Fig. 2) constitutes only about one percent of the rock, and occurs as single or multiple crystals less than 1 mm in diameter, commonly rimmed with biotite. The olivine has a relatively high CaO content (0.2%), which may be due to crystallization at low pressure, presence in an alkalic system, or some combination of both. The olivine is not visible in hand sample. Plagioclase (An_{40}) (Fig. 3) also constitutes less than one percent of the rock, and occurs as corroded cores within K-feldspar. Prelimi-

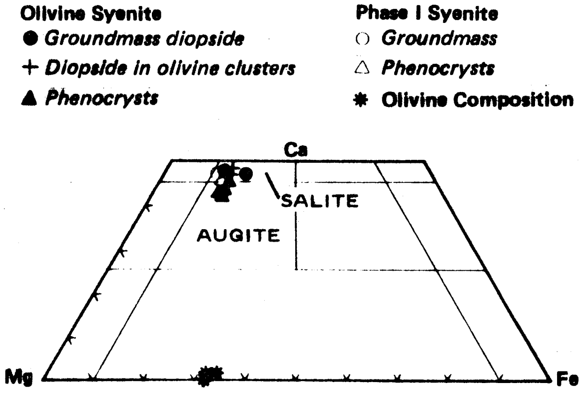

Figure 2. Pyroxene and olivine compositions for olivine syenite and Phase I syenite.

TABLE 1. SYENITE UNITS OF THE GRANITE MOUNTAIN PLUTON.

UNIT & AGE	FIELD APPEARANCE	DISTINGUISHING MINERALOGY	DISTINGUISHING GEOCHEMISTRY
Olivine syenite 95 Ma	Very dark gray to blue-gray. Commonly fine-grained.	olivine (Fo_{70}) diopside plagioclase (An_{40}) anorthoclase large apatite	CaO = 4% P_2O_5 = 0.4% Ba = 4000 ppm Zr = 200 ppm Sr 87/86 = 0.703
Phase I 89 Ma	Gray to dark-gray. Feldspars are large and may be slightly white.	salite K-feldspar ($An_{10}Ab_{40}Or_{50}$) sphene	CaO = 2% Ba = 1500 ppm Zr = 500 ppm Sr 87/86 = 0.7051 P_2O_5
Phase II 87 Ma	White to very light gray. Commonly mottles phase I. May occur in veins or as pegmatite. Associated with miariolitic cavities.	salite with aegirine rims K-feldspar ($An_0Ab_{30}Or_{70}$) large sphene analcime, nepheline	CaO 1% P_2O_5 0.1% Na_2O K_2O Ba = 250 ppm Zr = 1200 ppm Sr 87/86 = 0.7059 to 0.7198
Phase III 85 Ma	Pink, due to pink feldspar. Commonly as pegmatite or in veins and dikes.	X-ray perthite aegirine ziolites, nepheline high-U sphene	CaO = trace Na_2O K_2O Ba 50 ppm La 800x chon.

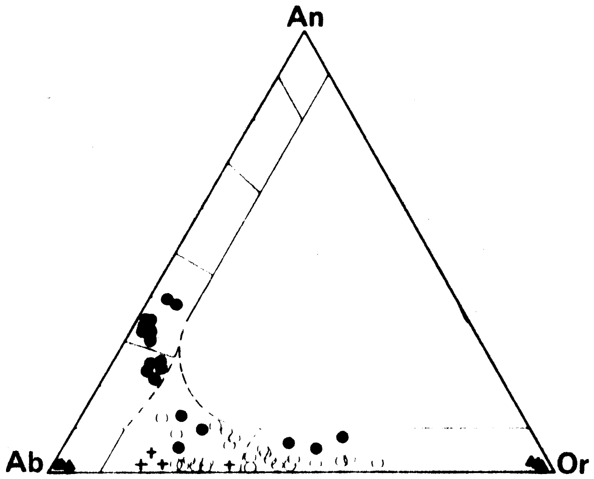

Figure 3. Feldspar compositions for syenites of Granite Mountain. Closed circles = olivine syenite; open circles = Phase I; crosses = Phase II; triangles = Phase III.

nary raman-probe data from this enclosing K-feldspar suggest that it has retained a plagioclase structure and is the result of low-temperature substitution of K for Ca. The K_2O content of plagioclase varies from 0.5% to about 6%.

The most abundant minerals in olivine syenites are K-feldspar, clinopyroxene, and biotite. The K-feldspars are effectively high-Ca anorthoclase ($An_{15}Ab_{65}Or_{20}$), with little alteration and few fluid inclusions. Pyroxene occurs as large, euhedral Mg-augite phenocrysts up to 4 mm in diameter and as lath-shaped, microphenocrysts of salite less than 1 mm long (Fig. 2). Amphiboles in the olivine syenite are kaersutite with 4-6% TiO_2 and MgO/FeO* averaging 1. Biotite is also relatively magnesian; some biotite has rims enriched in TiO_2 and FeO*.

Accessory minerals of olivine syenites are ilmenite, apatite, and nepheline. Apatite is commonly enclosed in or associated with olivine, and is up to 0.5 mm long. Nepheline occurs as replacement patches in corroded plagioclase. Sphene is extremely rare.

Fission-track dates of large apatite enclosed in olivines yield a mean age of 95 Ma (ranging from 108 ± 9 Ma to 92 ± 7 Ma) for olivine syenite (Morris and Eby, 1986). The inclusion of apatite within the early-crystallizing olivine may have shielded it from annealing by subsequent igneous or metasomatic activity. Major- and trace-element abundances of the olivine syen-

TABLE 2. ANALYSES OF REPRESENTATIVE SAMPLES, GRANITE MOUNTAIN PLUTON.

SAMPLE	86190	8614	862	8695	8613	8622
UNIT	OL SY	Ph I	Ph II	Ph III	XEN C	XEN B
SiO_2	58.20	60.30	58.90	56.30	47.50	43.60
TiO_2	1.13	0.93	0.64	0.52	1.41	2.62
Al_2O_3	19.50	18.70	18.70	21.40	5.20	10.80
MgO	1.28	0.76	0.51	0.38	17.60	11.30
MnO	0.25	0.18	0.28	0.23	0.26	0.20
FeO*	3.81	3.12	3.66	2.82	9.03	12.50
CaO	3.52	1.63	0.77	0.94	11.40	13.60
Na_2O	6.52	6.70	7.58	9.67	2.61	1.67
K_2O	4.67	6.71	6.37	6.58	1.59	1.37
P_2O_5	0.37	0.17	0.04	0.04	0.12	0.44
LOI	0.47	0.70	2.54	1.23	3.54	1.00
TOTAL	100.2	100.2	100.3	100.3	100.5	99.4
Ba	3880	1350	250	80	90	600
Rb	100	130	260	210	50	50
Sr	1790	510	240	5	40	500
Cr	1	2	0	0	1540	530
V	62	38	40	40	220	350
Ni	8	5	3	0	380	210
Zr	210	450	1250	360	90	200
Nb	90	150	320	190	60	70
U	2	5	15	30	1	2
Th	7	12	36	52	4	5
Cl	600	1200	50	6300	50	50
S (%)	0	0	0	0	3.44	0
FeO/FeO*	0.60	1.0	0.1	0.05	0.70	0.07

ite serve to distinguish it from later phases (Tables 1 and 2). This unit contains up to four weight percent CaO and is enriched in Sr (1800 ppm) and Ba (4000 ppm). It is enriched in light rare earth elements (LREE), with La = 300x chondrite and La/Yb = 45 (Fig. 4). The olivine syenite has an initial Sr 87/86 ratio of 0.70351 (Kruger, Geochron Labs, analyst).

Phase I Syenite

This rock, correlative to "pulaskite", is a medium-to-dark-gray color, and may be distinguished from olivine syenite in the field by its lighter color and coarser texture. The two lithologies have gradational contacts. Phase I syenite, so-named because it is the first phase of complete metasomatic alteration, comprises 80 to 85 percent of the Granite Mountain pluton.

The principal minerals of the Phase I syenite are K-feldspar, clinopyroxene, biotite, and amphi-

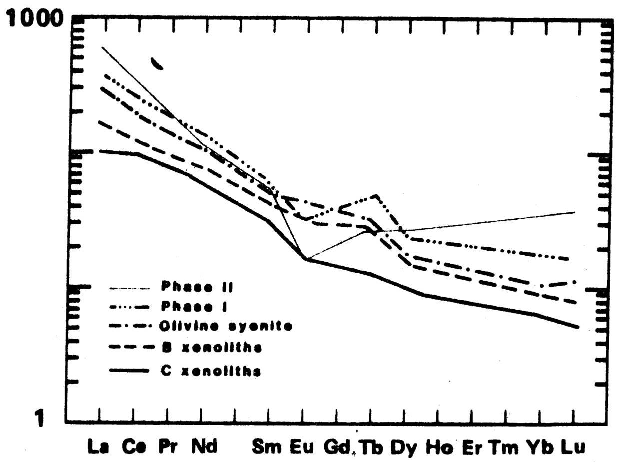

Figure 4. Chondrite-normalized plot of rare earth elements for Granite Mountain syenites and xenoliths.

bole. Feldspars are dominantly sodic orthoclase with a significant content of CaO, ($An_{10}Ab_{40}Or_{50}$), grading into anorthoclase cores (Fig. 3). As in the olivine syenites, raman-probe data indicate retention of some elements of plagioclase structure. Pyroxene compositions are similar to those of the olivine syenite (Fig. 2), although some phenocrysts in Phase I rocks are slightly more calcic than their olivine syenite counterparts. Biotite and amphibole do not change composition from the olivine syenite.

Significantly, sphene appears in the Phase I syenites, and is diagnostic of a Phase I rock. It probably developed from the breakdown of plagioclase and ilmenite, both of which are present in olivine syenite but absent in Phase I. Large apatites are gone in Phase I, and small, 0.01 mm apatite appears instead, usually enveloped in biotite or K-feldspar.

Mean fission-track ages from both apatite and sphene of Phase I syenites are 89 Ma (Morris and Eby, 1986)--significantly younger than the olivine syenite. Because both minerals are newly crystallized in the Phase I rocks, this age should date metasomatism and recrystallization.

Several major and trace element signatures characterize the Phase I syenite (Table 2). CaO, MgO, and P_2O_5 are less abundant than in olivine syenite, whereas K_2O and SiO_2 increase. Ba and Sr decrease markedly in concentration. The rocks are more REE-enriched, with La = 350x chondrite, and La/Yb = 30. The Sr 87/86 ratios of Phase I syenite range from 0.70513 to 0.71980, suggesting varying degrees of crustal input (Tilton et al., 1987).

Phase II Syenite

Phase II Syenites are white rocks which may occur as patches and "blobs" in Phase I and olivine syenite, or may occur as veins which commonly are pegmatitic. They are also associated with miariolitic cavities. The white color is caused by abundant fluid inclusions within feldspars which impart a brown, altered appearance in thin section. The Phase II syenites comprise at least 10% of Granite Mountain, and perhaps 95% of the syenites in Saline County. Their exact proportion is difficult to determine because they are intimately mixed with the other phases and grade into both Phase I and Phase III.

Phase II syenites contain relatively homogeneous feldspars with less than 0.1 weight percent CaO and have compositions of $An_0Ab_{30}Or_{70}$. Pyroxenes are salite with iron-enriched aegirine rims (Fig. 5). Analcime and Na-nepheline are abundant accessories, and sphene is richer in uranium than Phase I rocks.

Fission-track ages from sphene and apatite of Phase II rocks are 87 Ma, close to those of Phase I syenites.

Major- and trace-element data are similar to Phase II rocks, and the trends noted from olivine syenite to Phase I continue. CaO, MgO, P_2O_5, Ba, and Sr decrease. Na_2O and SiO_2 increase. REE are more abundant, with La at least 400x chondrite and La/Yb = 20. The Sr 87/86 isotopic composition of a feldspar separate is 0.70515 (Krueger, Geochron Laboratory, analyst), suggesting crustal influence.

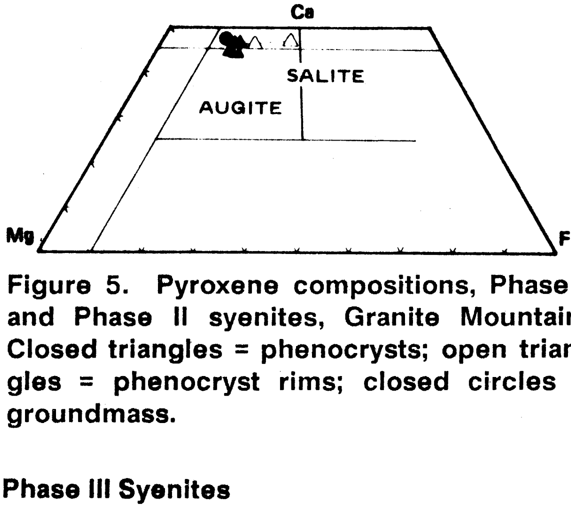

Figure 5. Pyroxene compositions, Phase I and Phase II syenites, Granite Mountain. Closed triangles = phenocrysts; open triangles = phenocryst rims; closed circles = groundmass.

Phase III Syenites

Phase III syenites are faintly to blatantly pink-colored rocks which form dikes or veins and may be pegmatitic. This lithology is gradational to Phase II. Phase III rocks constitute less than 10% of the pluton.

The most characteristic mineral of Phase III is perthitic feldspar (X-ray perthite), in which end-member orthoclase and albite form laminae less than 10 microns wide. the feldspars are also extremely rich in inclusions, and oxidized

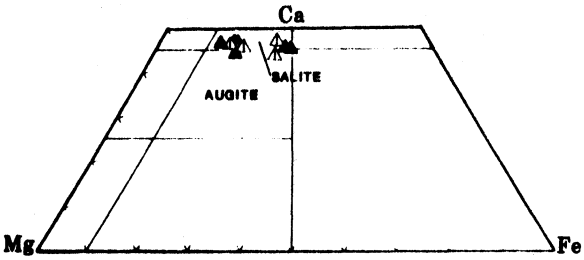

Figure 6. Calcic clinopyroxene compositions, Phase III syenites. Solid circle = groundmass; solid triangle = phenocrysts; open triangle = phenocryst rims.

iron accounts for the diagnostic red or pink color of the rock. The feldspars have no "memory" of plagioclase structure. Salite (Fig. 6) is rare in these rocks, and is replaced by aegirine and a green, sodic amphibole. Biotite is present. Sphene is a high-uranium variety. Analcime and nepheline are common accessories. Apatite is not present. Mean fission-track dates of sphene are 85 Ma (Morris and Eby, 1986).

XENOLITHS

Xenoliths of at least four distinct lithologies are present in the syenite of Granite Mountain. They tend to occur in clusters or swarms which, although uncommon, are distributed throughout the pluton. Xenoliths are also found in isolated occurrences. They are in all phases of syenite. Some swarms are enveloped in Phase II (white) syenite, but swarms occur in Phase I syenite as well, and multiple xenoliths have been recovered from olivine syenite. Only much altered, isolated xenoliths have been observed in the pink Phase III syenite.

Type A Xenoliths

Type A xenoliths are inclusions of country rock. Commonly they are hornfelsed sandstone or shale, and usually have only a narrow reaction rim. They are, unsurprisingly, most abundant near the edge of the pluton, and are usually angular to sub-angular.

Type B Xenoliths

Type B xenoliths are dark-brown porphyritic clasts, commonly angular to sub-angular, with subtly zoned diopside phenocrysts in a very fine groundmass. Pyroxenes in type B xenoliths (Fig. 7) are near-end-member diopside, with an average composition of $Ca_{47}Mg_{45}Fe_8$. They contain up to 1.2 weight percent Cr_2O_3, 1 to 2 weight percent TiO_2, and only traces of Na_2O (0.1-0.2 weight percent). Olivine, feldspars, amphiboles, or biotite are absent form this xenolith type. The only identifiable phases are ilmenomagnetite and chrome spinel. Broad-beam probe analyses of groundmass show that the leucocratic groundmass has $Na_2O/K_2O = 0.10$ whereas the melanocratic groundmass contains about 3.5 weight percent TiO_2, MgO = FeO* (12%), and $Na_2O/K_2O = 1$.

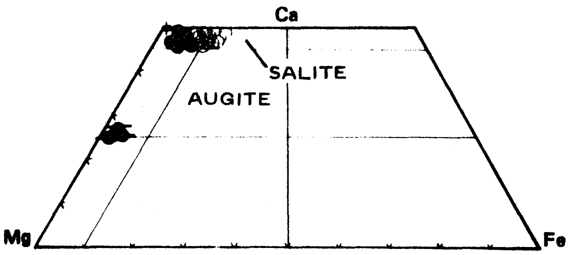

Figure 7. Pyroxene and amphibole compositions, Type B and Type C xenoliths, Granite Mountain. Open circles = type B pyroxene; closed circles = type C pyroxene; closed circle with bar = amphibole reaction rim on pyroxene, type C; crosses = diopside in core of large type B pyroxene.

Major- and trace-element chemistry of type B xenoliths is distinctive (Table 2). They are Ne-normative, with MgO/FeO* about 1, enriched in Ni (220 ppm), Cr (550 ppm), and V (350 ppm) with about 2.5 weight percent total alkalies. They are LREE-enriched, La = 160x chondrite, with a slope to the REE plot which parallels the more REE-enriched syenites (La/Yb = 25).

The type B xenoliths most closely resemble some primitive lamprophyres from the Benton dike swarm (Fig. 1). However, there are some distinctive differences in mineralogy and composition which suggest that these xenoliths are not simply assimilated lamprophyres of the same type. These differences include low REE abundances relative to lamprophyres, low whole-rock and pyroxene TiO_2 compared to lamprophyres, and the absence of phases such as olivine, amphibole and biotite which are present in lamprophyres. The angularity of type B xenoliths, their relatively uniform texture and composition, and the presence of at least one

15-meter wide type B inclusion against which syenite has chilled suggest that the type B xenoliths came from a lamprophyre of distinctive composition which was enveloped by the early syenite.

Type C Xenoliths

The type C xenoliths are commonly rounded and less than 15 cm in diameter. They contain large (2 mm) rounded clinopyroxene within a matrix rich in sulfide and alkalis. The pyroxenes are unzoned diopside (Fig. 7), averaging $Ca_{47}Mg_{48}Fe_5$, with less than 1.0 weight percent TiO_2 and about 0.3 weight percent Na_2O. Olivine and feldspars are absent. Mg-rich amphibole forms reaction rims around diopside (Fig. 7). Biotite is present as a very fine-grained phase. Sulfides--predominantly pyrite with subordinant pyrrhotite--constitute up to 15 percent of the rock. The groundmass is fine grained with Na_2O/K_2O of about 5.

The major- and trace-element chemistry of the type C xenoliths is markedly different than type B (Table 2). They contain Ne and Ac in the norm, Na_2O/K_2O = 1.5 to 2.0, with total alkalis = 4 weight percent, and MgO/FeO* = 1.7. Elemental sulfur constitutes about 3.5 weight percent of type C xenoliths. Compatible elements are enriched, with Cr = 1500 ppm, V = 220 ppm, and Ni = 400 ppm. These xenoliths are also LREE-enriched, with La = 100x chondrite and a REE profile nearly parallel to syenites and the type B xenoliths (La/Yb = 26).

The type C xenoliths, with rounded, end-member diopside in a matrix of sulfides and alkalis most likely represent early cumulates from the Granite Mountain syenites which were caught in an immiscible liquid and consequently segregated from the very early magma. These xenoliths may, in fact, represent early, solidified, liquid blobs, rather than lithic fragments. The absence of olivine is puzzling. It suggests that the blobs formed possibly before olivine began to recrystallize, and that the diopside in the xenoliths is native to the xenolith. The absence of olivine, or any olivine pseudomorphs also indicates that type C xenoliths are not fragments of upper mantle.

Type D Xenoliths

Type D xenoliths are rare. They consist of phlogopite, anorthite, and diopside. Only two of this type have been found to date, and these have been too small for reliable whole-rock chemistry. The mica in the type D xenoliths is, indeed, Mg-rich phlogopite, with MgO/FeO* = 2. These xenoliths have a cumulate fabric, with phlogopite surrounding the other two phases. The type D xenoliths may be bits of a metasomatized mantle source region or may be early cumulates of the syenite magma.

SUMMARY: PETROLOGY OF GRANITE MOUNTAIN

The data presented above indicate that Granite Mountain incorporates at least two major igneous episodes--the first at about 95 Ma with the intrusion of olivine-plagioclase-diopside rocks, and the second at about 89 to 85 Ma which was principally potassium metasomatism and interaction with fluids, transforming much of the earlier rock into pulaskite (Phase I syenite). The initial intrusion was probably gabbroic and mantle-derived. Later syenites cannot be modeled as fractionation products of olivine syenite or as a more ideal gabbroic magma, or as products of fractionation and mixing. Textural, trace-element and microprobe data suggest substantial interaction with late potassic fluids, and isotopic data indicate that later syenite was moderately to strongly affected by a crustal component. Metasomatism is a viable mechanism for developing later phases from the olivine syenite. The olivine, diopside, and plagioclase in the olivine syenite cannot be xenocrysts for a multitude of reasons, including the absence of any olivine in known xenoliths, a radical difference between xenolith and syenite pyroxene compositions, the relatively fractionated compositions of the minerals compared to mantle equivalents, and, finally, their dissimilarity in composition to olivine and pyroxene in likely sources such as lamprophyres.

The Granite Mountain syenites seem to record a complex history of magmatism and metasomatism which is unique. Work is in progress to better answer tantalizing questions such as the source of metasomatic fluids and the relation of this pluton to the rest of the Arkansas alkalic province.

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REFERENCES

Morris, E. M. and Eby, G. N., 1986, Petrologic and age relations in Granite Mountain, Arkansas syenite: Geological Society of America Abstracts with Program, v. 18, p. 256.

Mullen, E. D., and Petty, W. B., 1985, Petrologic relations among syenites and lamprophyric rocks, Arkansas alkalic province: Geological Association of Canada Abstracts with Program, v. 10, p. A42.

Tilton, G. W., Kwon, S. T., and Frost, D. M., 1987, Isotopic relationships in Arkansas alkalic complexes in Morris, E. M. and Pasteris, J. D., eds., Alkalic magmatism and mantle metasomatism: Geological Society of America Special Paper, in press.

REFERENCES CITED IN DESCRIPTIONS OF STOPS -- PAGES 1-35

Bouma, A. H., 1962, Sedimentology of some flysch deposits: A graphic approach to facies interpretation: Amsterdam, Elsevier, 196 p.

Bowring, S. A., 1984, U-Pb zircon ages of granitic boulders in the Ordovician Blakely Sandstone, Arkansas and implications for their provenance, in Stone, C. G., and Haley, B. R., eds., A Guidebook to the central and southern Ouachita Mountains, Arkansas: Arkansas Geol. Comm. Guidebook 84-2, p. 123.

Comstock, T. B., 1888, Report on the preliminary examination of the geology of western central Arkansas: Arkansas Geological Survey Annual Report for 1888, v. 1, pt. II, 320 p.

Cox, T. L., 1986, A structural study of the serpentinites of the Benton Uplift, Saline County, Arkansas: [M.S. thesis], University of Arkansas, 99 p.

Croneis, C. G., 1930, Geology of the Arkansas Paleozoic area with special reference to oil and gas possibilities: Arkansas Geological Survey Bull. 3, 457 p.

Gordon, Mackenzie, Jr., 1968, An early Reticuloceras zone fauna from the Hale Formation in northwestern Arkansas: U.S. Geological Survey Professional Paper 613-A, 19 p.

Gordon, Mackenzie, 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, v. 1: Arkansas Geological Commission Misc. Pub. MP-13, p. 70-91.

Haley, B. R., and others, 1976, Geologic map of Arkansas: U.S. Geological Survey map, scale 1:500,000.

Haley, B. R., and Stone, C. G., 1981, Structural framework of the Ouachita Mountains, Arkansas: Geological Society of America Abstracts with Programs, v. 13, no. 5, p. 239.

___________, 1982, Structural framework of the Ouachita Mountains, Arkansas: Geological Society of America Abstracts with Programs, v. 14, no. 3, p. 113.

Howard, J. M., 1986, Summary of isotopic dates of Cretaceous igneous rocks of Arkansas, in Stone, C. G., and others, Sedimentary and igneous rocks of the Ouachita Mountains of Arkansas, pt. 1: Arkansas Geological Commission Guidebook, GB86-2, p. 107-108.

Keller, W. D., Stone, C. G., and Hoersch, A. L., 1985, Textures of Paleozoic chert and novaculite in the Ouachita Mountains of Arkansas and Oklahoma and their geological significance: Geological Society of America Bulletin, v. 96, p. 1353-1363.

Morris, E. M., and Stone, C. G., 1986, Alkalic metagabbro from the core of the Ouachita Mountains, Arkansas: Geological Society of America Abstracts with Programs, v. 18, no. 3, p. 256.

Morris, R. C., 1985, Slope and axial fan system in Carboniferous rocks, frontal Ouachita Mountains, Oklahoma and Arkansas, in Morris, R. C., and Mullen, E. D., 1985, Alkalic rocks and Carboniferous sandstones, Ouachita Mountains, new perspectives: Guidebook for Geological Society of America Regional Meeting, April 16-18, 1985, p. 1-33.

Mutti, E., 1985, Turbidite systems and their relations to depositional sequences, in Zuffa, G. G., ed., Provenance of arenites, D. Reidel Pub. Co., p. 65-93.

Quinn, H. E., III, 1986, Sedimentary facies, structure and palynology of the lower and middle members of the Atoka Formation, central Arkansas: [M.S. thesis], University of Missouri-Columbia, 157 p.

Stone, C. G., 1968, The Atoka Formation in north-central Arkansas: Arkansas Geological Commission Misc. Pub., 24 p.

Stone, C. G., and Haley, B. R., 1977, The occurrence and origin of the granite-meta-arkose erratics in the Ordovician Blakely Sandstone, Arkansas, in Stone, C. G., ed., Symposium on the geology of the Ouachita Mountains, v. 1: Arkansas Geological Commission, p. 107-111.

Stone, C. G., Haley, B. R., and Viele, G. W., 1973, Guidebook to the geology of the Ouachita Mountains, Arkansas: Arkansas Geological Commission Guidebook GB-73-1, 113 p.

Stone, C. G., and McFarland, J. D., III, 1981, Field guide to the Paleozoic rocks of the Ouachita Mountains and Arkansas Valley provinces, Arkansas: Arkansas Geological Commission, Guidebook 81-1, p. 6-13.

Stone, C. G., and Milton, Charles, 1976, Lithium mineralization in Arkansas, in Vine, J. D., ed., Lithium resources and requirements by the year 2000: U.S. Geological Survey Professional Paper 1005, p. 137-142.

Stone, C. G., and Sterling, P. J., 1964, Relationship of igneous activity to mineral deposits in Arkansas: Arkansas Geological Commission, 22 p.

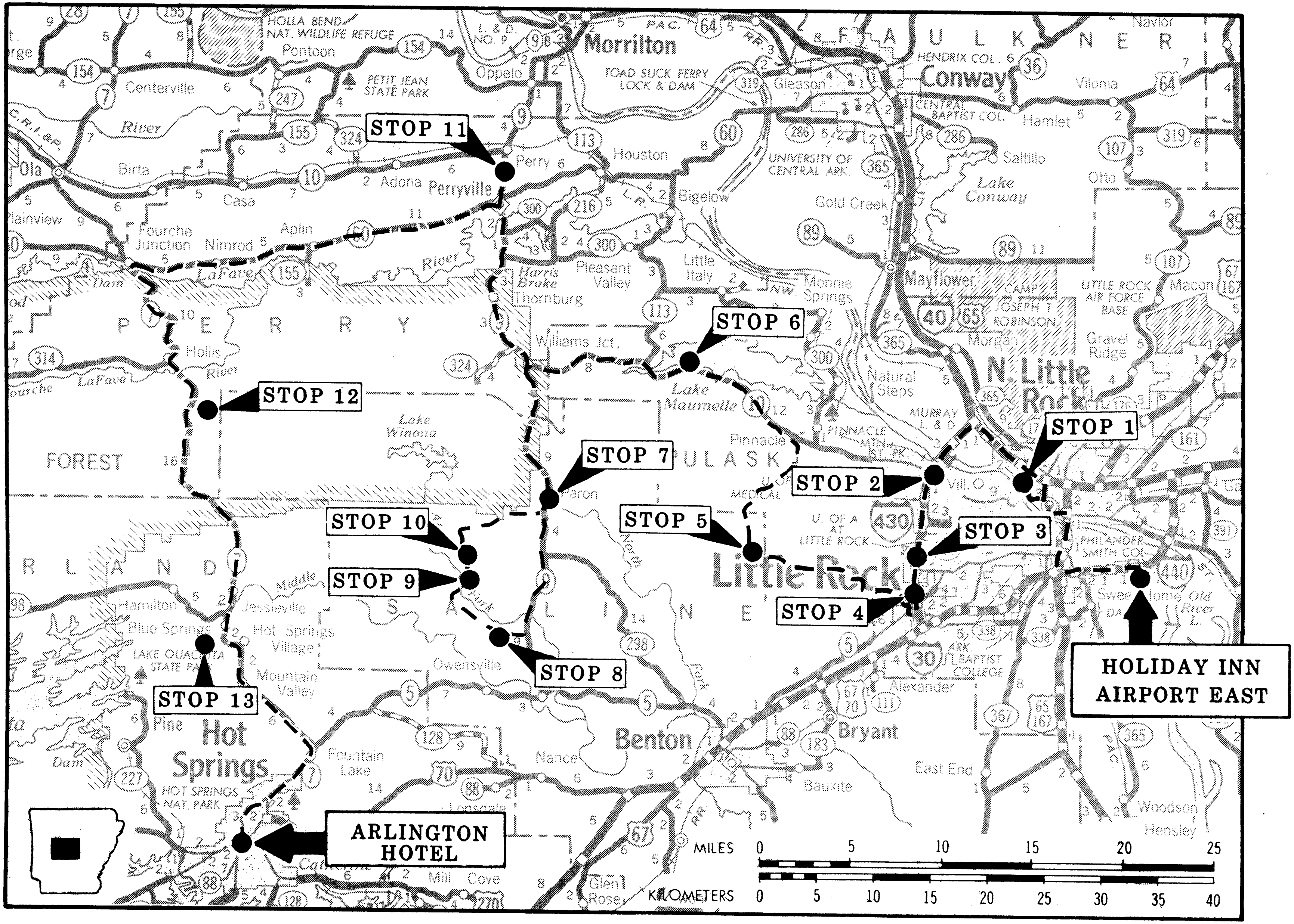
Thomas, W. A., 1985, The Appalachian-Ouachita connection: Paleozoic orogenic belt at the southern margin of North America: Annual Review of Earth and Planetary Sciences, v. 13, pp. 175-199.

Viele, G. W., 1973, Structure and tectonic history of the Ouachita Mountains, Arkansas, in DeJong, K. A., and Sholten, R., eds., Gravity and tectonics: New York, John Wiley & Sons, p. 361-377.

__________, 1979, Geologic map and cross section, eastern Ouachita Mountains, Arkansas: Map summary: Geological Society of America, v. 90, pt. 1, p. 1096-1099.

STRATAGRAPHIC SECTION OF ROCKS EXPOSED IN THE OUACHITA MOUNTAINS AND ARKANSAS VALLEY PROVINCES, ARKANSAS

	MAXIMUM THICKNESS
QUATERNARY	
Alluvium - clay, silt, sand, and gravel	90'
Terrace deposits - gravel, sand, clay	40'
TERTIARY SYSTEM	
Wilcox Group - clay, sand, lignite, bauxite	120'
Midway Group - marl, limestone, shale	75'
CRETACEOUS SYSTEM	
Tokio Formation - gravel, sand, clay	300'
Brownstown Marl - gravel, sand, marl, and clay	250'
Igneous rocks - nepheline syenite, phonolite, ijolite, peridotite,kimberlite	--
Trinity Group - gravel, sand, clay, gypsum, and minor limestone	150-1,000'
PENNSYLVANIAN SYSTEM	
Des Moinesian Series	
Savanna Formation - sandstone and sandy shale	850'
McAlester Formation - shale, sandstone, and coal	1,000'
Hartshorne Sandstone - massive sandstone	325'
Atokan Series	
Atoka Formation - shale and sandstone	27,500+
Morrowan Series	
Johns Valley Shale - shale, minor sandstone and limestone, and erratic boulders	1,500+
Jackfork Sandstone - sandstone and shale	6,000'
MISSISSIPPIAN SYSTEM	
Stanley Shale - shale, sandstone, some chert and minor tuff	11,000'
DEVONIAN AND MISSISSIPPIAN SYSTEMS	
Arkansas Novaculite - novaculite, shale, and conglomerate	950'
SILURIAN SYSTEM	
Missouri Mountain Shale - shale with minor sandstone	250'
Blaylock Sandstone - sandstone, siltstone, and shale	1,500'
ORDOVICIAN SYSTEM	
Polk Creek Shale - shale	175'
Bigfork Chert - chert, limestone, and shale	800'
Womble Shale - shale with some thin limestone and sandstone	1,900'
Blakely Sandstone - shale, sandstone, and erratic boulders	450'
Mazam Shale - shale with some sandstone and limestone	3,000'
Crystal Mountain Sandstone - sandstone, shale and erratic boulders	850'
CAMBRIAN AND ORDOVICIAN SYSTEMS	
Collier Shale - shale and limestone	1,000'



INDEX MAP – FIELD TRIP STOPS