The Geology of North America
Vol. F-2, The Appalachian-Ouachita Orogen in the United States
The Geological Society of America, 1989

# Chapter 27

# *Tectonic synthesis of the Ouachita orogenic belt*

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## INTRODUCTION

The tectonic history of the Paleozoic Ouachita orogenic belt composes a complete Wilson cycle: the early history records the rifting of the southern margin of the North American craton and the opening of an ocean basin in which marginal and basinal pre-orogenic sediments were deposited; the later history records the closing of that ocean basin by south-directed subduction and the accompanying deposition and deformation of synorogenic clastic sediments. The off-shelf, pre-orogenic rocks and the deep-water, synorogenic rocks compose the "Ouachita facies" (Fig. 1; Plates 9, 11), deposited beside but not on the North American craton, although the youngest synorogenic sediments lapped onto the southern margin of the craton. Rocks of the Ouachita facies were thrust onto the southern margin of the North American craton during the final stages of ocean closing and constitute the Ouachita orogenic belt. Geologic structures related to the Ouachita orogeny extend beyond the boundaries of the orogenic belt into the cratonal interior.

In general, this chapter will trace the geologic events composing the Wilson cycle of the Ouachitas, departing in places to elaborate on selected topics. Early in the preparation of this volume, the decision was made to impose no single view of Ouachita tectonic history upon the many authors. Therefore, some tectonic scenarios presented in preceding chapters differ from our own. At points of contention and disagreement, the opposing arguments are noted. For readers unfamiliar with the Ouachitas, some confusion may result, but that is preferable to the misconception that all problems of Ouachita geology are solved.

Paleomagnetic reconstructions (Scotese, 1984) place the southern margin of North America in low equatorial latitudes throughout the Paleozoic, generally facing as much as 30° of open ocean toward the south (directions refer to the present orientation of North America). Within this paleogeographic framework, early to middle Paleozoic sediments were deposited on the southern part of the North American plate, some on the craton and some on oceanic crust downslope from it. The major plates converged during the Carboniferous and joined by the Early Permian to form Pangea. The Ouachita orogenic belt records the collision along the southern margin of the North American craton.

The Ouachita orogenic belt comprises several tectonic provinces (Fig. 1). They are located, briefly described, and illustrated in Chapter 16 of this book. Readers will find it helpful as well to refer to Plates 8, 9, 10, and 11 while reading this synthesis chapter.

## THE RIFTING PHASE

In the late Precambrian to Cambrian, the North American craton (Laurentia) rifted along trends later followed by the Appalachian-Ouachita orogenic belt, and oceans opened along the newly formed continental margins. The tectonic morphology of the new continental margins included rifted-margin sediment prisms flanking areas of shelf deposition, and grabens and basins opening toward the ocean (Fig. 2). This morphology was germane to the early history of the Ouachita orogenic belt and is described here, albeit briefly, to set the early Paleozoic tectonic framework.

### *Paleozoic Shelf Deposition*

During the early to middle Paleozoic, from the regions of the southern Appalachians to western Texas, excepting some basins and grabens, shallow-water marine strata were deposited in shelf environments on the North American craton (Thomas, this volume, Fig. 3B). Craton-wide unconformities separate these rocks into sequences (Sloss, 1963) that record numerous transgressions separated from one another by relatively abrupt retreats of sea water and periods of erosion (Sloss, 1982). The history of the shelf strata began with the deposition of Upper Cambrian clastic rocks forming the basal formations of the Sauk sequence (Fig. 3), and continued with the deposition of predominately carbonate rocks forming the strata of the upper Sauk, Tippecanoe, and Kaskaskia sequences. In the shelf areas, the Sauk sequence is widespread and as much as 1.5 km thick,

Viele, G. W., and Thomas, W. A., 1989, Tectonic synthesis of the Ouachita orogenic belt, *in* Hatcher, R. D., Jr., Thomas, W. A., and Viele, G. W., eds., The Appalachian–Ouachita Orogen in the United States: Boulder, Colorado, Geological Society of America, The Geology of North America, v. F-2.

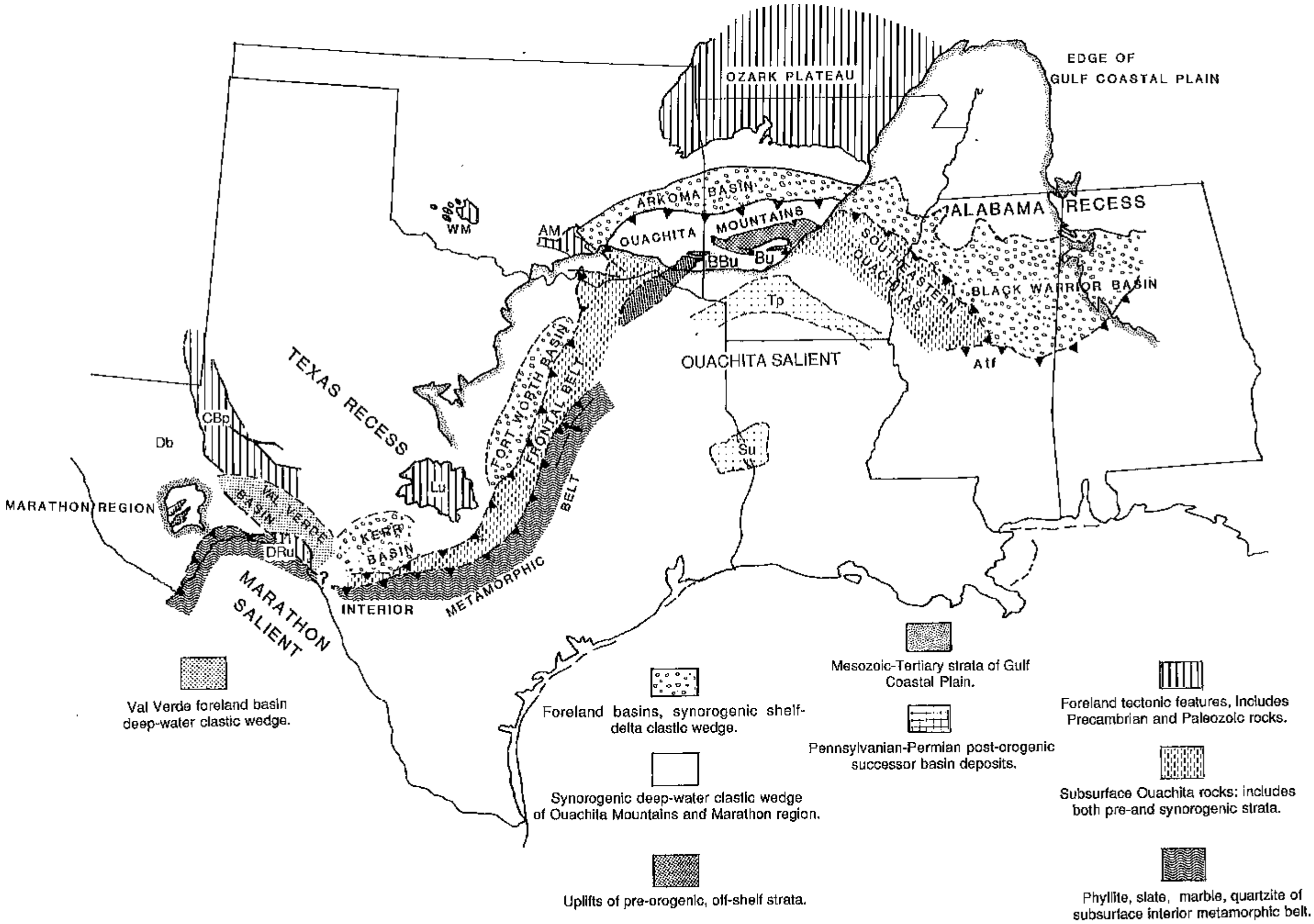


Figure 1. Index map of tectonic elements of the Ouachita orogenic belt and adjacent areas. Abbreviation code: AM = Arbuckle Mountains; Atf = Appalachian tectonic front; BBu = Broken Bow uplift; Bu = Benton uplift; CBp = Central Basin platform; Db = Delaware basin; DRu = Devils River uplift; Tp = Texarkana platform; Su = Sabine uplift; WM = Wichita Mountains.

whereas the Tippecanoe and Kaskaskia sequences have a patchy distribution and are relatively thin (generally less than 250 m). Faunal assemblages suggest that the Sauk, Tippecanoe, and Kaskaskia sequences were deposited well inboard of the continental margins. Strata of the Sauk sequence record the fastest rates of deposition (Fig. 3), but these rates slowed during deposition of the Tippecanoe and Kaskaskia sequences. Probably, the water deepened over the cratonal margin during the deposition of the Tippecanoe and Kaskaskia sequences (Lowe, 1975), for cherty carbonate beds of Devonian and Early Mississippian age are notably abundant along the southern margin of the North American craton, but nowhere in these rocks have continental slope and rise environments been recognized.

### *Paleozoic Graben and Basin Areas*

Indenting the southern margin of the North American craton are three areas of subsidence: the Mississippi Valley graben (Thomas, 1985), the Southern Oklahoma basin, and the Tobosa basin (Fig. 2). All have been interpreted as aulacogens or failed rift arms, and to all a similar tectonic history has been assigned (Burke and Dewey, 1973; Hoffman and others, 1974; Walper, 1977). As tectonic units, the grabens and basins formed during the "early rifting phase," but in our view, the tectonic histories of these areas are dissimilar and constrain the direction of opening of the early Paleozoic ocean.

Geophysical surveys (Ginzburg and others, 1983; Hildenbrand, 1985) of the northern Mississippi Embayment of the Mesozoic-Cenozoic Gulf Coastal Plain, the Rough Creek graben (Soderberg and Keller, 1981), and the Rome trough (Ammerman and Keller, 1979) reveal a system of late Precambrian–early Paleozoic horst and graben structures in the subsurface (Fig. 2) (Thomas, this volume, Fig. 3A). The sedimentary fills of the grabens range from late Precambrian(?) to Late Cambrian and are several kilometers thick—as much as 5.5 km in the Rough Creek graben (Soderberg and Keller, 1981). Generally, the Cambrian strata grade upward from shale and coarse-grained sandstone of feldspathic composition to younger beds of shale and carbonate rock (Houseknecht and Weaverling, 1983). Upper

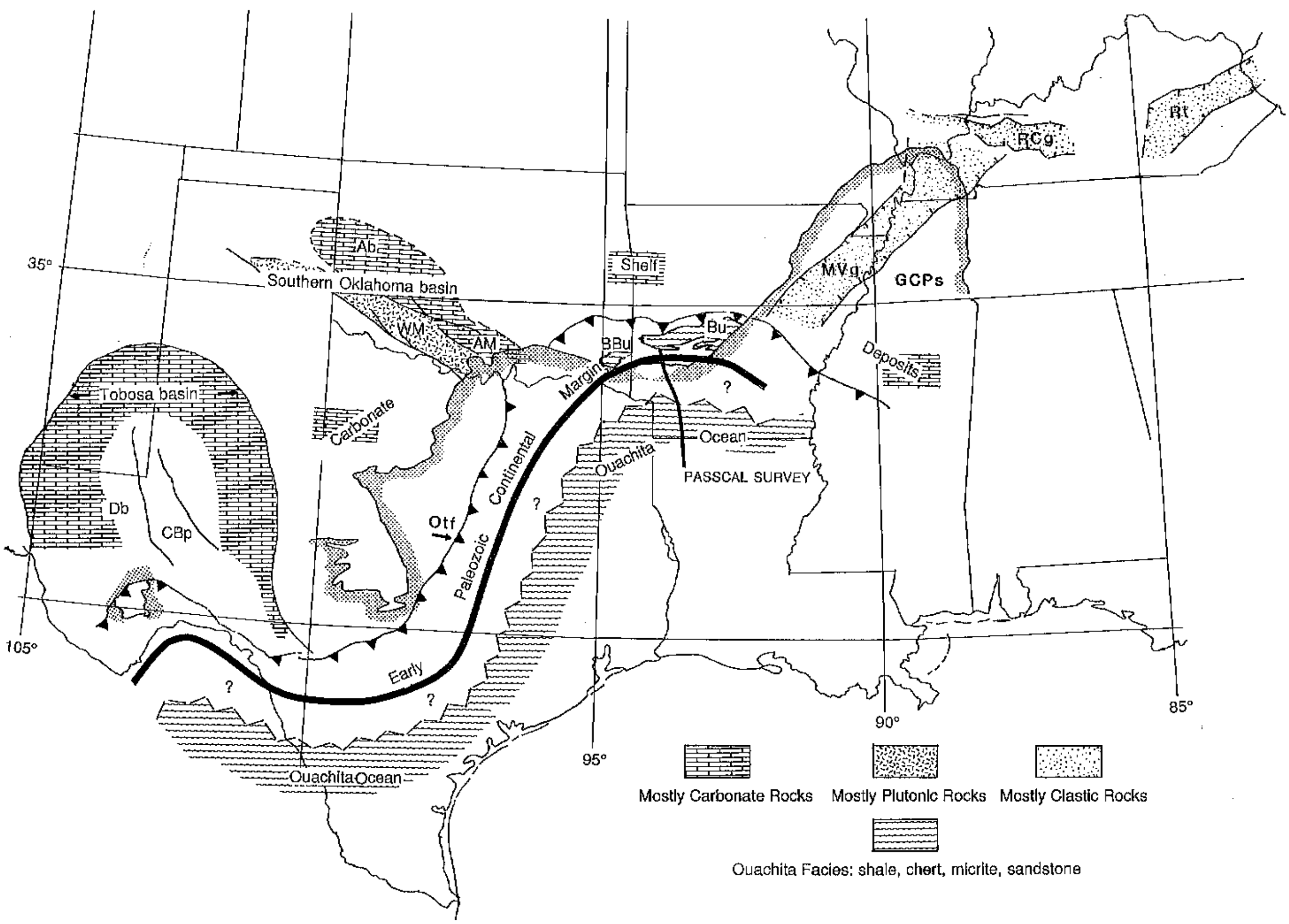


Figure 2. Index map of early Paleozoic tectonic elements and Carboniferous uplifts that are mentioned in the text. Abbreviation code: Ab = Anadarko basin; BBu = present-day Broken Bow uplift; Bu = present-day Benton uplift; CBp = Central Basin platform; Db = Delaware basin; GCPs = Gulf Coastal Plain strata; MVg = Mississippi Valley graben; Otf = present-day Ouachita tectonic front; Rt = Rome trough; RCg = Rough Creek graben; WM = Wichita Mountains. Width of early Paleozoic Ouachita ocean is unknown.

Cambrian carbonate strata were deposited without offset across the boundary faults of the grabens.

Positive Bouguer gravity anomalies associated with the Mississippi Valley–Rough Creek–Rome grabens bespeak dense mafic rocks at depth (Ammerman and Keller, 1979; Braile and others, 1982; Hildenbrand, 1985). The inferred mafic rocks are probably Precambrian (Keweenawan?), although a swarm of ultramafic dikes was intruded in the region of the northern Mississippi Valley graben during the latest Paleozoic (Zartman, 1977). No known volcanic rocks were deposited within the grabens during Early to Middle Cambrian extensional faulting.

In contrast, sedimentary graben fills of Early to Middle Cambrian age are unknown in the Southern Oklahoma basin; instead, volcanic rocks that once floored the Southern Oklahoma basin crop out widely in the Arbuckle and Wichita Mountains, Carboniferous uplifts cored by Precambrian and pre-Morrowan Paleozoic rocks (Fig. 2). In the Wichita Mountains, the volcanic and plutonic rocks record a first phase of generally basaltic-gabbroic igneous activity, extending from the late Proterozoic to the Middle Cambrian, and after some erosion, a second phase of rhyolitic and granitic activity, approximately 525 Ma (Ham and others, 1964; Gilbert, 1982). The distribution of the igneous rocks is restricted, probably because the zone of igneous activity was bounded by high-angle faults.

In the Arbuckle Mountains, rhyolitic rocks compose the upper part of the basement and underlie Upper Cambrian sandstone of the basal Sauk sequence, underlying in turn, some 4 km of mostly carbonate rocks of the Sauk and Tippecanoe sequences (Ham and others, 1964). These sedimentary rocks were deposited in the broad Southern Oklahoma basin, which widened and deepened southeastward toward the ocean that was receiving sediments of the pre-orogenic Ouachita facies.

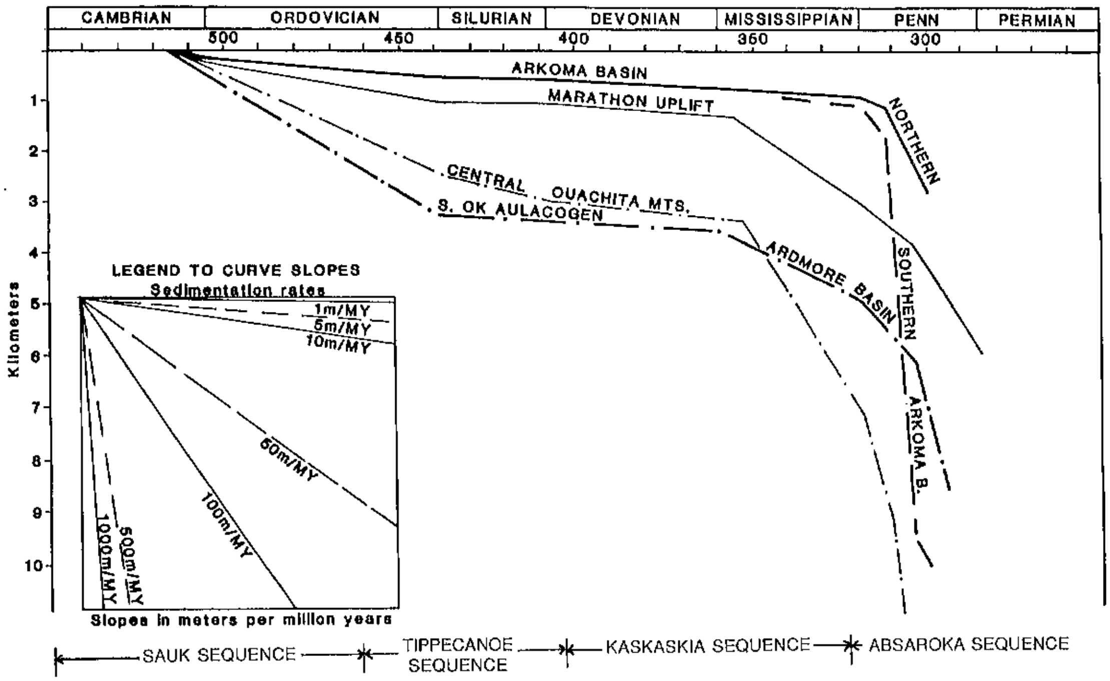


Figure 3. Comparison of depositional rates of Paleozoic strata in: Arkoma basin, Marathon uplift, central Ouachita Mountains, and Southern Oklahoma basin (aulacogen). Slightly modified from Arbenz (1989, Fig. 7).

Still farther west, the Tobosa basin of western Texas (Fig. 2) extended northward from the later site of the Marathon region (Adams, 1965). Bouguer gravity maps (Keller and others, 1981) show a gradient of as much as 120 mgals across the boundary of the Delaware basin and the Central Basin platform, later tectonic subdivisions of the older Tobosa basin (Fig. 2) (Denison, this volume). Crustal models indicate dense rocks beneath the old Tobosa basin, and indeed, wells drilled on the Central Basin platform have penetrated gabbroic rocks beneath the Paleozoic strata (G. R. Keller, personal communication, 1987). Late Paleozoic faults that uplifted the Central Basin platform and separated it from the Delaware basin apparently follow the edges of the gabbroic body.

The basal strata of the Tobosa basin are not well known. Numerous wells on the Central Basin platform have penetrated Upper Cambrian sandstones resting on basement rock, but few wells have been drilled to the equivalent stratigraphic level in the Delaware basin. Stratigraphic cross sections of the Tobosa basin show little variation of thickness of the Cambrian strata, but Lower Ordovician through Mississippian strata, mostly carbonate rocks, show an increase in thickness toward the center of the basin (Adams, 1965). Relative to the margins, deeper water covered the center of the Tobosa basin where subsidence apparently exceeded deposition (Adams, 1965).

These three areas of subsidence—the Mississippi Valley graben, the Southern Oklahoma basin, and the Tobosa basin—are similar and dissimilar. Strongly positive gravity anomalies and high magnetic intensities copy the trends of all three basins and bespeak the presence of dense, mafic rocks at depth. From the Early Ordovician, at least, all three were sites of deposition (Sloss, 1982). The Sauk sequence is much thicker within the basins than on the adjacent shelf areas, as are the Tippecanoe and Kaskaskia sequences in the Southern Oklahoma and Tobosa basins. Through the Middle Cambrian, at least, high-angle extensional faults bounded the Rome–Rough Creek–Mississippi Valley graben system and probably the Southern Oklahoma basin. A lack of data prohibits comment about faults of like age beneath the Tobosa basin.

There are some dissimilarities, however. For example, the abundant late Proterozoic to Middle Cambrian igneous rocks of the Southern Oklahoma basin have no known counterparts in the Rome–Rough Creek–Mississippi Valley graben. Furthermore, the clastic rocks of Early(?) and Middle Cambrian age in the Rome–Rough Creek–Mississippi Valley graben have no known counterparts in the Southern Oklahoma and Tobosa basins.

### *Early Paleozoic Continental Margin*

In the late Precambrian to Cambrian, in the area of the southern Appalachians, rifting opened an ocean, and deposition followed of rifted-margin prisms of strata along the newly formed cratonal margin (Thomas, this volume). In the subsurface of eastern Arkansas and western Mississippi, no equivalent rifted-margin prism is known, but instead, well samples and logs indicate an abrupt transition from the autochthonous Cambrian-Ordovician carbonate shelf into allochthonous deep-water mudstones and cherts of the Ouachita facies (Thomas, 1972). The exact position and breadth of the facies change are unknown, for it lies beneath the younger thrust sheets of the Ouachita orogenic belt.

Farther west, only geophysical signatures provide evidence of the southern margin of the North American craton (Plate 10) (Keller and others, this volume). On aeromagnetic (Zietz, 1982) and gravity maps alike, closely spaced contours define steep gradients of the potential fields along transects of the Ouachita orogenic belt. Bouguer gravity maps (Keller and others, this volume) show a change from negative to positive values on traverses from the craton across the Ouachitas to the Gulf Coastal Plain. Indeed, from the Ouachita Mountains to the Marathon region, a ridge of positive Bouguer anomalies is essentially congruent with the interior or seaward side of the Ouachita orogenic belt and is known as the "interior zone maximum." Moreover, a recent wide-angle reflection/refraction seismic survey in southwestern Arkansas (Fig. 4) demonstrates that the gravity gradient coincides with a southward jump in crustal velocities from the range of 6.1 to 6.5 km/s to 6.7 to 7.1 km/s (Keller and others, 1989). The change in seismic velocities is modeled as an interwedging of granitic and basaltic crust, the zone of transition marking the passage from deeply buried early Paleozoic continental to early Paleozoic oceanic or intermediate crust (Keller and others, 1989). We view the interior zone maximum as the gravity signature of the early Paleozoic margin of the North American craton from the Ouachita Mountains to the Marathon region of Texas (Fig. 2; Plate 10).

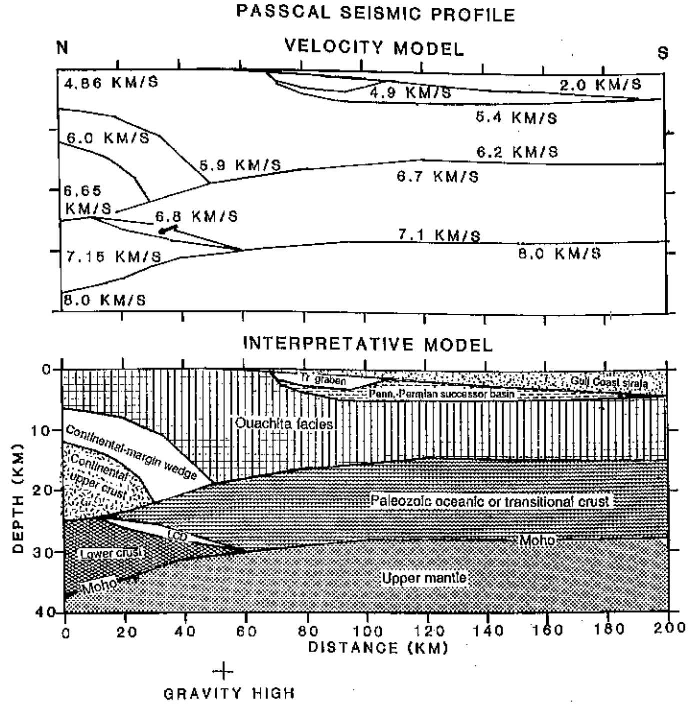


Figure 4. Velocity and crustal models based on PASSCAL reflection/refraction seismic survey. Line of survey shown on Figure 2. Modified from Keller and others (1989).

Along the interior zone maximum, only a single geophysical profile known to us provides a hint of a buried rifted-margin prism (Lillie, 1984). In west-central Arkansas, a deep, wedge-shaped packet of reflectors, recorded on a COCORP seismic reflection profile (Lillie and others, 1983), may indicate a rifted-margin prism; possibly, the reflections come from early Paleozoic sedimentary and volcanic rocks deposited along the continent-ocean boundary (Plate 11, cross section C-C′) (Lillie, 1985).

The gravity and aeromagnetic gradients, and the inferred transition from continental to oceanic crust, might be associated with Mesozoic rifting of the southern margin of North America and the Late Jurassic–middle Cretaceous opening of the Gulf of Mexico (Winker and Buffler, 1988). The Mesozoic extension along the trend of the Ouachitas, however, appears to be too small to account for the dimensions of the potential field gradients; therefore, they are better interpreted as the geophysical signature of the change from early Paleozoic continental to early Paleozoic oceanic crust (Keller and others, 1989).

### *Pre-Orogenic Ouachita Strata*

Areas of outcrop of lower to middle Paleozoic Ouachita facies in the Benton and Broken Bow uplifts of the Ouachita Mountains of Arkansas and Oklahoma (Fig. 2; Plate 8) and in the Marathon region of Texas (Plate 8) coincide with or lie a short distance inboard of the trend of the interior zone gravity maximum. The outcrops of Ouachita rocks lie close to the seaward ends of the Mississippi Valley graben, the Southern Oklahoma basin, and the Tobosa basin. Furthermore, if the interior zone maximum marks the seaward edge of the North American craton, the outcrops of Ouachita rocks should lie above carbonate-shelf strata of equivalent or somewhat older age.

The present-day juxtaposition of the Ouachita and carbonate-shelf facies suggests that transitions or a blending of facies with one another should exist. But the lower to middle Paleozoic strata, exposed in the Ouachita Mountains and in the Marathon region, are strikingly different from strata of equivalent age on the North American craton (Fig. 5). For one thing, the Ouachita strata are mostly shales, sandstones, cherts, and micrites deposited in deep water, whereas the shelf rocks are mostly limestones and dolomites deposited in shallow water. Strata in the cratonic grabens and basins thicken toward the Ouachita Mountains and Marathon region, but lateral equivalents of the basin-graben fills are not known in the Ouachita rocks. Only scattered boulders of granite and carbonate rock in the Ordovician formations of the Ouachita Mountains and Marathon region provide some hint of a continental shelf upslope to the north or west. At the Ouachita tectonic front, isopachous contours of the basin fills are sharply truncated, as are gravity and magnetic contour lines of similar trend (Plate 10). The widespread unconformities bounding the stratigraphic sequences (Sloss, 1963) of the craton are not recognized in either the Ouachita Mountains or the Marathon

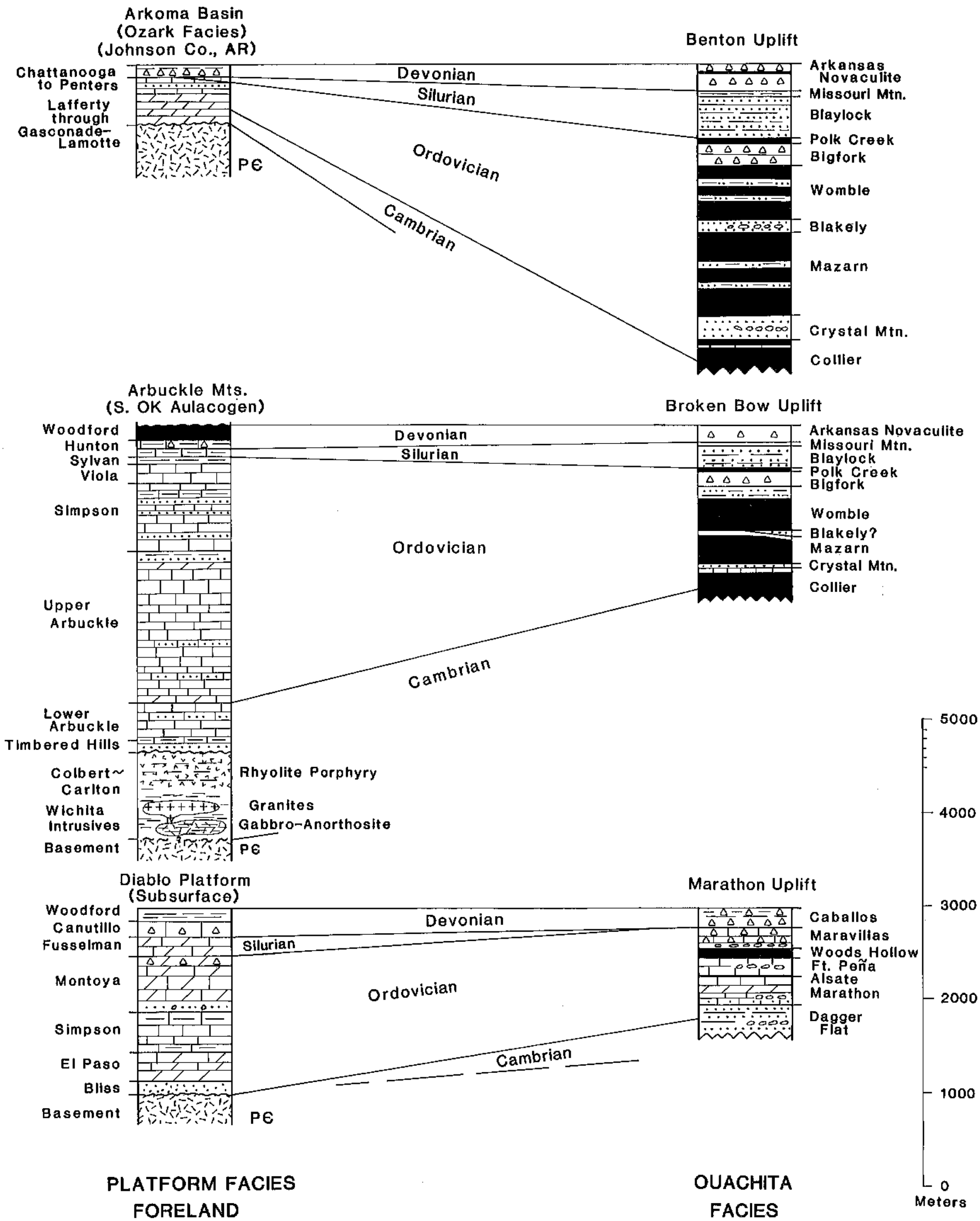


Figure 5. Generalized stratigraphic sections of pre-Chesterian strata in the Benton and Broken Bow uplifts and in the Marathon region compared to stratigraphic sections from nearby areas on the North American shelf. From Arbenz (1989, Fig. 5).

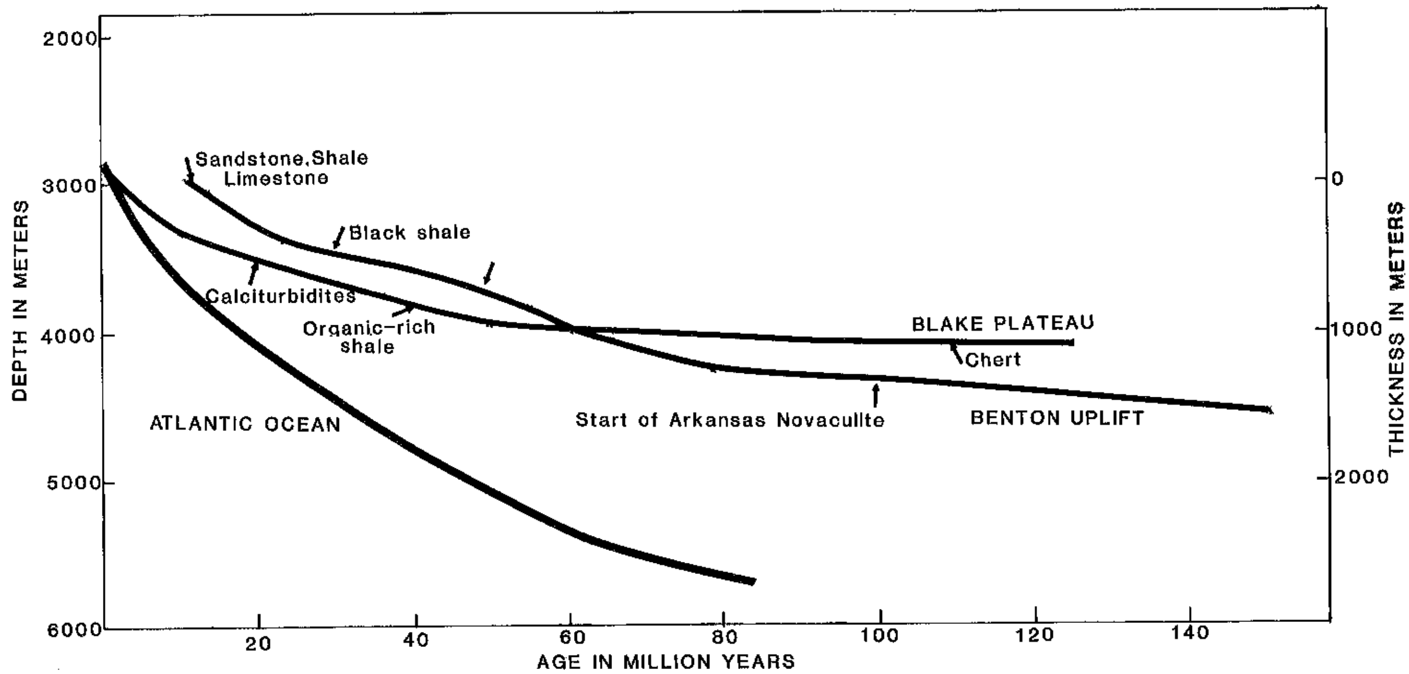


Figure 6. Comparison of subsidence rates of the floor of the Atlantic Ocean (Sclater and Francheteau, 1970), the Blake Plateau (Sheridan and Gradstein, 1982), and the pre-orogenic strata of the Benton uplift. The Atlantic Ocean curve relates depth to age of the oceanic crust. The Blake Plateau and Benton uplift curves relate stratigraphic thickness, uncorrected for compaction, to the age of the strata.

region. Furthermore, a pronounced change of biofacies distinguishes the rocks of the Ouachitas from those of the craton (Wilson, 1954; Ethington and others, this volume). Yet, Upper Cambrian trilobites in the lowermost formation (Collier Formation; Fig. 5) of the Benton uplift of the Ouachita Mountains clearly have North American affinities and lived in outer-shelf to upper-slope environments (Hart and others, 1987). Conodonts from the Ordovician Ouachita strata include representatives of both North American provinces (shallow and/or warm water) and the North Atlantic province (deep and/or cold water) (Repetski and Ethington, 1977; Ethington and others, this volume).

On the Benton uplift, the aggregate thickness of the pre-orogenic strata is about 3.5 km, but this figure is tenuous because of the intense deformation. The pre-orogenic Ouachita strata are about the same thickness as equivalent-age strata of the Southern Oklahoma basin (Fig. 5) (Ham, 1959).

The pre-orogenic strata of the Benton and Broken Bow uplifts may be divided roughly into two parts: a lower succession comprising shale, sandstone, and micrite; and an upper succession comprising black shale, siliceous shale, ribbon chert, and thick-bedded novaculite (Lowe, this volume). Quartzose sandstones of the lower succession are made up of well-sorted, rounded grains, indicating initial sorting by waves and currents in shallow water, but graded beds and sole marks suggest transport of the sand by inertial flows from shallower to deeper water. Interstratified debris-flow deposits (Haley and Stone, 1977) contain boulders of limestone, chert, and granitic rocks that yield zircons with ages ranging from 1284 ± 12 Ma to 1407 ± 13 Ma (Bowring, 1984). The lower clastic succession fills an interval of time of about 50 m.y., Late Cambrian to Middle Ordovician, and relative to the upper succession, it was deposited rapidly (Fig. 3). The paleotectonic site of deposition of the lower succession was probably on the continental slope and rise of early Paleozoic North America, and possibly the bouldery debris came from the walls of submarine canyons cutting the continental margin.

Succeeding the lower clastic strata are black shales and, locally, phosphatic sandstones (Stone and Sterling, 1962) that grade upward into a siliceous succession composed primarily of bedded cherts, siliceous shales, and novaculites, although some shales and sandstones are interstratified. In the southwestern part of the Benton uplift, the novaculite contains extensive deposits of manganese (Miser, 1917). Fossils of the siliceous succession are graptolites, radiolaria, sponge spicules, and conodonts, suggestive of deep-water environments, whereas cherty formations of equivalent age on the craton contain a shallow-water biofacies. The siliceous strata, which are slightly more than 1 km thick, were deposited from the Late Ordovician to the Early Mississippian or for about 100 m.y., a slow rate of sedimentation indeed. The composition and fauna of the siliceous succession suggest deposition in deep water in a deeply foundered if not oceanic basin. In the changing rates of deposition and composition through time, the strata of the Benton uplift mimic the deposition of strata on the Blake Outer Ridge, which record the opening and spreading of the present Atlantic Ocean (Fig. 6) (Sheridan and Gradstein, 1982).

The same general patterns of deposition are recognizable in the Marathon region (Fig. 5) (McBride, this volume): a lower succession of argillaceous to sandy rocks, interstratified with limestones and containing debris flows, passes upward into bedded chert and novaculite. Clasts of Late Cambrian limestones contained within Ordovician shales of the lower succession contain faunas typical of the seaward margin of the shelf (Palmer and others, 1984); therefore, it seems probable that the lower succession was deposited seaward of the continental shelf on the continental slope or rise. Several formations of the lower clastic succession contain sparsely distributed clasts of igneous and metamorphic rocks. With McBride (this volume), we view the overlying siliceous succession of the Marathon region as being deposited in deep water in an oceanic basin. The total thickness of the pre-orogenic stratigraphic column is about 950 m.

In general in the Marathon region, the upper siliceous succession is more calcareous and less siliceous, particularly along the Marathon anticlinorium (Fig. 7) (Wilson, 1954), than is the siliceous succession in the Benton and Broken Bow uplifts. The Caballos Novaculite is more reminiscent of the Arkansas Novaculite at Black Knob Ridge at the southwestern end of the exposed Ouachita thrust belt in Oklahoma (Plate 9) than it is of the Arkansas Novaculite within the Benton uplift. Wilson (1954) noted that the rocks of the Solitario uplift (Fig. 7), southwest of the Marathon region, were more akin to the rocks of the Broken Bow uplift than to the rocks of the Marathon outcrops; he suggested that the Solitario rocks were deposited farther eastward in a more interior position of the Ouachita orogenic belt than the rocks of the Marathon region. Personal visits to the Marathon region and descriptions in the literature suggest to us that the pre-orogenic strata of the Marathons were deposited closer to North America than equivalent strata of the Benton and Broken Bow uplifts.

We disagree with interpretations by Lowe (this volume) of a shallow-water setting for parts of the Arkansas Novaculite. Limitations of space preclude a full discussion, but many of the features he describes in the Arkansas Novaculite (brecciation, dissolution, fracture fillings, gas bubbles, oxide cements) may be attributed, in our view, to a loss of water and volume from siliceous oozes, transforming diagenetically to microcrystalline cherts, and to hydrothermal activity on the ocean bottom. Abundant manganese, sedimented with the novaculite (Miser, 1917), also suggests a deep-marine environment of deposition. Moreover, deposits of barite immediately above the Arkansas Novaculite in the Mississippian shales of the Stanley Shale indicate strong hydrothermal activity in a deep-water environment (Hanor and Baria, 1977; Shelton, this volume). Lowe (this volume) assigns the northern conglomeratic facies of the Arkansas Novaculite in the Benton uplift to deep water and the southern, massive facies to shallow water. Our view is exactly opposite: the northern facies was deposited closer to North America than the southern. Admittedly, the areal distribution of these facies within the Benton uplift is, in part, a result of thrust faulting, and admittedly, an accurate palinspastic restoration may not be possible, but nevertheless, a transition from deeper to shallower water seems evident in outcrops of Arkansas Novaculite on a traverse from the Broken Bow uplift to Black Knob Ridge to outcrops of Woodford Chert in the Arbuckle Mountains (Plate 9).

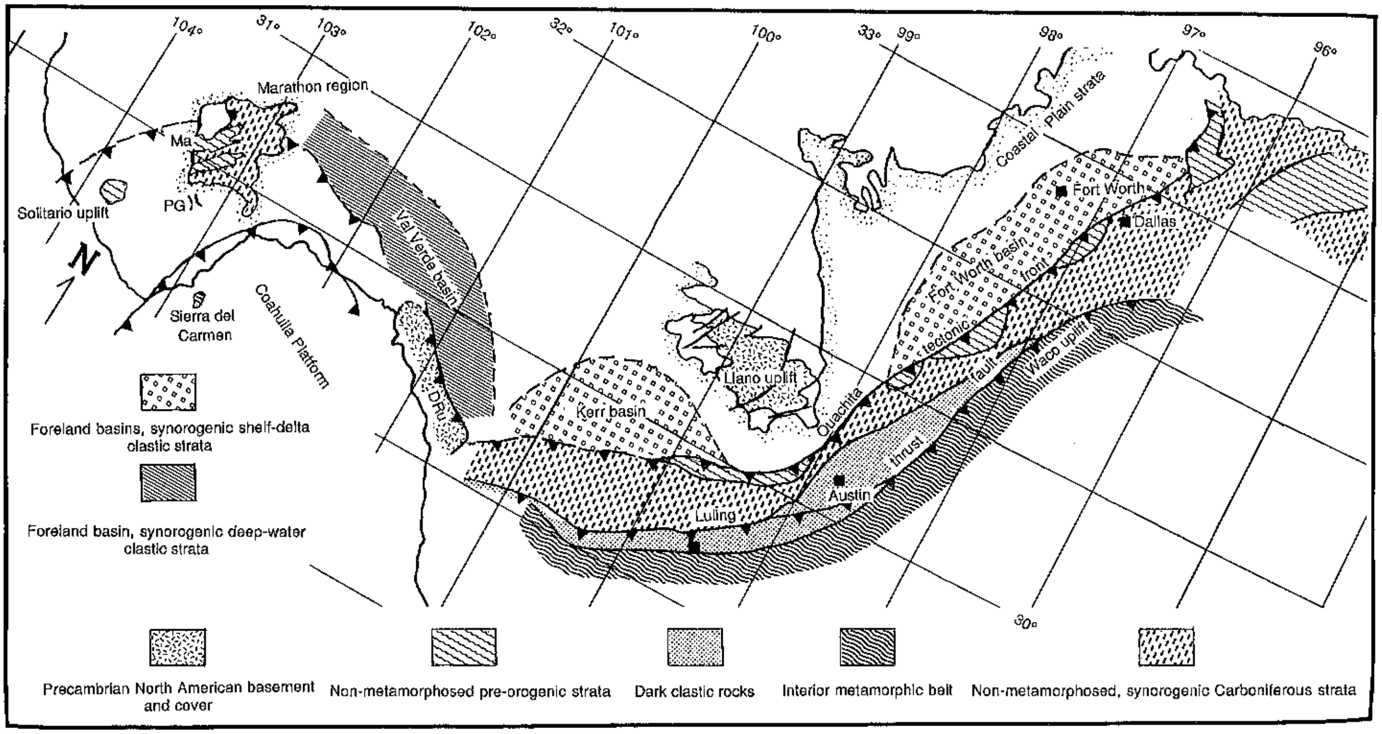


Figure 7. Geologic map of the Ouachita subsurface belts of eastern Texas. Modified from Flawn and others (1961). Abbreviation code: Ma = Marathon anticlinorium; PG = Persimmon Gap.

Lowe (1985) and Arbenz (1989) proposed that the pre-orogenic strata of the Ouachitas were deposited in narrow intra-cratonal basins analogous to those of the Mississippi Valley graben and the Southern Oklahoma basin. The differences between the shelf and Ouachita successions, however, in stratigraphy, composition and associated mineralization, fossil fauna, and geophysical signature seem to us to preclude Lowe's model. Evidence is tenuous for a proximal landmass outboard of the Ouachitas during the early and middle Paleozoic; clastic units, such as the Blaylock Sandstone (Fig. 5) (Lowe, this volume), attributed to southern sources, may have been introduced from numerous points along the southern margin of North America and transported by contour currents along the continental slope and rise (Satterfield, 1982). Only in the synorogenic Carboniferous strata is there clear evidence from sedimentary structures and sandstone mineralogy for an outboard source of sediment.

The pre-orogenic sediments of the Ouachita orogenic belt were deposited adjacent to but not on the North American continent, as was recognized by King (1937, 1975). They were deposited on the North American continental slope and rise and probably on the abyssal floor of the bordering ocean. It is not possible to say how far seaward the pre-orogenic sediments extended.

### *Subsurface Pre-Orogenic Ouachita Strata, Eastern Texas*

In the subsurface of eastern Texas, numerous wells have penetrated rocks of the Ouachita orogen arranged in north-south–trending belts (Fig. 7) (Flawn and others, 1961; Nicholas and Waddell, this volume). The westernmost of the subcrop belts includes a non-metamorphosed pre-orogenic sequence containing rocks at least as old as the Ordovician black shales that lie beneath the upper siliceous succession of the Ouachita Mountains. The most distinctive subsurface samples, however, are from the upper siliceous succession. The Arkansas Novaculite is recognized, but generally as a tan or greenish chert containing radiolaria. The prominent white cherts and massive novaculites of the Benton and Broken Bow uplifts are only locally known.

Flanking the non-metamorphosed belt on the east is a belt of dark clastic rocks, abundantly carbonaceous to graphitic and containing mica and dolomite (Flawn and others, 1961). The dark clastic rocks lie in two metamorphic zones, a western zone of incipient metamorphism, and an eastern zone of greater recrystallization that contains carbonaceous matter converted to graphite and argillaceous material reconstituted to new sericite and chlorite. The age of the dark clastic rocks is uncertain, but Flawn and others (1961) argued that they were older than Middle Pennsylvanian and perhaps considerably older. Viele has examined some thin sections from this unit and has found them to be remindful of the lower clastic rocks of the Benton uplift, but until better evidence appears, they are best dated as pre-Pennsylvanian.

### *Subsurface Interior Metamorphic Belt, Eastern Texas*

Lying still farther east in the subsurface is the interior metamorphic belt (Fig. 7) made up of phyllite, slate, marble, metaquartzite, and schist (Flawn and others, 1961; Nicholas and Waddell, this volume). Relative to all other units in the Ouachitas, the interior metamorphic belt is the most strongly metamorphosed. In thin sections of the quartzites and marbles, the original rounded outlines of the quartz grains are still visible, but in thin sections of marbles, recrystallization has blurred the original sedimentary fabric. Chlorite and sericite are the predominant phyllosilicates. Flawn and others (1961) reported the presence of garnet in the southern part of this zone and suggested that metamorphism increases southward. Throughout the belt, shearing predominates over recrystallization.

The interior metamorphic belt appears to be a distinct petrologic province. Citing petrologic similarities, Flawn and Maxwell (1958) tentatively correlated the only outcrop of the interior metamorphic belt, at the base of the Sierra del Carmen of Mexico (Fig. 7), with the Marathon Limestone (Fig. 5) of the Marathon region. Direct evidence for the age of the rocks of the interior metamorphic belt is unknown. They are described here with the pre-orogenic rocks, primarily because they are so dissimilar to Ouachita rocks of Carboniferous age.

## TECTONIC HISTORY, RIFTING PHASE OF WILSON CYCLE

Burke and Dewey (1973) noted that many fault-bounded basins extend inland into cratons from triple junctions. They described two triple junctions in the region of the Ouachitas: one of Mesozoic age, the Jackson triple junction lying within the mouth of the Mississippi Embayment of the Gulf Coastal Plain; and the other of Paleozoic age, the Dallas triple junction lying toward the mouth of the Southern Oklahoma basin. Later authors (Hoffman and others, 1974; Ervin and McGinnis, 1975; Walper, 1977; Viele, 1979a; Kruger and Keller, 1986) accepted this general idea and visualized an ocean opening in early to middle Paleozoic time along the southern margin of North America. The trend of the spreading oceanic rise foretold the trend of the Ouachita orogenic belt. The failed arms of the triple junctions extended inland along the Mississippi Embayment and the Southern Oklahoma basin and were termed aulacogens, following the restricted usage of the term by Burke and Dewey (1973).

Several things are troublesome about this scenario. The rocks of the Rome–Rough Creek–Mississippi Valley graben, the Southern Oklahoma basin, and the Tobosa basin share neither a common composition nor a common history. The differences in geologic history preclude a common tectonic origin. The oldest exposed rocks of the Benton and Broken Bow uplifts exhibit no evidence of deposition in rift valleys. Furthermore, no offsets are recognized of the pre-orogenic Ouachita facies along the multitude of transform faults that are necessitated in the active spreading arms of triple junctions. Finally, no convincing evidence is known to us, geophysical or geological, for a buried rifted-margin prism of early Paleozoic strata along the southern margin of the craton.

An alternative scenario is used in this chapter for the opening of the ocean (Thomas, 1985, this volume, Fig. 3A). By late

Precambrian–Early Cambrian time, rifting and southeastward drift of the split-off plate had opened an ocean, and a rifted-margin prism had formed along the trend of the southern Appalachians. A post-rift unconformity along the Appalachians shows that rifting there was complete by the beginning of the Cambrian Period. By Early to Middle Cambrian time, the locus of extension had shifted westward to the Rome–Rough Creek–Mississippi Valley graben. A large northwest-trending transform truncated the Mississippi Valley graben and marked the approximate trend of the yet-to-be-formed Ouachita Mountains (Thomas, this volume, Fig. 3). A broad rifted-margin prism was not formed along the southern margin of North America, because across the transform fault, the transition from shallow to deep water was abrupt. A necessary correlative of this hypothesis is that rifting and the formation of a rifted-margin prism should have occurred along the present trend of the subsurface Ouachitas of eastern Texas (Thomas, this volume, Fig. 3A), but no evidence available from the subsurface of eastern Texas documents a buried rifted-margin prism.

In the Arbuckle and western Ouachita Mountains of Oklahoma, stratigraphy and rock composition do indicate a transition zone, tectonically shortened, from the shallow-water shelf carbonates of the Arbuckle Mountains to the deep-water, pre-orogenic clastic and siliceous strata of Black Knob Ridge and the Broken Bow uplift (Plate 8). In our view, however, the Southern Oklahoma aulacogen is not the failed arm of a triple junction but, instead, a leaky transform fault (Thomas, 1986). A graben-fill facies like that of the extensional Mississippi Valley graben has not been found, because in the Southern Oklahoma transform system, volcanism and plutonism along the transform replaced the deposition of clastic rocks. After the keel of hot mafic rocks was emplaced, the area subsided via thermal contraction and isostatic adjustment to form the Southern Oklahoma basin that received the sediments of the Sauk and overlying sequences.

The pre-orogenic strata of the Ouachita orogenic belt record the first half of a Wilson cycle: the opening of an ocean and the subsequent widening and deepening of that ocean as the oceanic crust cooled and subsided. The Ouachita ocean first opened during the Early and Middle Cambrian, but the strata associated with the initial rifting are lost to view in the subsurface. During the Late Cambrian and through much of the Ordovician, sediments of the lower clastic succession were deposited on the North American continental slope and rise. From the Late Ordovician through the Early Mississippian, sediments of the upper siliceous succession were deposited on the lower slope and rise and on the abyssal ocean floor. Abyssal conditions prevailed until the onset of subduction and the closing of the ocean.

## SYNOROGENIC CARBONIFEROUS STRATA

The Carboniferous strata of the Ouachita orogenic belt record a major change in tectonic regime: a change from the spreading phase of a Wilson cycle dominated by extensional tectonics and the opening of an ocean to a closing phase dominated by contractional tectonics and the closing of an ocean. Everywhere along the length of the Ouachita orogenic belt, rapid deposition of clastic sediment succeeded the slow deposition of chert and siliceous shale. Taken as a whole, the clastic strata record the filling, from east to west, of a remnant ocean basin (Thomas, this volume, Fig. 3C, D, E), yet along the length of the basin, or across it on any chosen traverse, different sedimentary facies are present primarily because of different degrees of infilling and closing of the basin. These facies will be described from east to west, from Alabama to Texas. The discussion will be general and interpretive, as detailed descriptions of the Carboniferous strata have been presented in preceding chapters of this volume by Morris and McBride.

### *Black Warrior Basin, Alabama and Mississippi*

In the subsurface of the Gulf Coastal Plain, southwest of the surface Black Warrior basin (Fig. 1), the initial stages of infilling of the Ouachita ocean are lost to view. During late Meramecian time, the oldest clastic sediments reached the foreland Black Warrior basin, which lies above continental crust, and prograded over the older Paleozoic shelf succession, forming a wedge of shallow-marine to deltaic deposits that continued to prograde northeastward during Chesterian time (Plate 9) (Thomas, 1974, 1985, this volume, Fig. 3C). Paleogeographic maps constructed from subsurface data (Thomas, 1979, 1988a, 1988b) show a regional change of facies from clastic rocks on the southwest to carbonate rocks on the northeast (Thomas, this volume, Fig. 3C). Similarly, deltaic sands and muds of the overlying, Lower Pennsylvanian Pottsville Formation prograde northeastward, but instead of intertonguing with carbonate rocks, they meet in northeastern Alabama with a similar clastic wedge prograding southwestward from the Appalachians (Hobday, 1974; Thomas, 1974).

Petrographic studies of the Mississippian-Pennsylvanian sandstones indicate a multipart provenance (Mack and others, 1983). Clasts of schist, phyllite, and slate are abundant, as are clasts of sandstone, siltstone, and chert, the last coming from bedded cherts probably deposited in deep water. In addition, the sandstones contain fragments of andesitic-basaltic rock, andesitic-dacitic rock, and devitrified glass. In many samples, the relative percentages of grains fall between the fields defined on the provenance diagrams of Dickinson and Suczek (1979), suggesting that several tectonic terranes composed the provenance. Probably, the bulk of the sediment came from an uplifted orogen containing a volcanic arc and a subduction complex. These tectonic terranes lay in the Ouachita orogenic belt southwest of the Black Warrior basin; all are buried today beneath the Gulf Coastal Plain.

### *Ouachita Mountains and Arkoma Basin, Arkansas and Oklahoma*

Beneath the sediments of the Gulf Coastal Plain, between the subsurface Black Warrior basin and the Ouachita Mountains, the Carboniferous strata change from shallow-water clastic rocks to interstratified sandstone and shale strongly reminiscent of the deep-water flysch of the Alpine and Carpathian mountain belts (Plate 9) (Cline, 1970). The base of the Carboniferous strata appears to be Meramecian, but the bulk of the Mississippian strata in the Ouachita Mountains is Chesterian (Gordon and Stone, 1977) and laterally equivalent to the shallow-water, deltaic Mississippian rocks of the Black Warrior basin.

In Arkansas, the contact of the basal beds of the Mississippian Stanley Shale with the underlying Arkansas Novaculite is conformable and gradational, although local unconformities have been reported along the southern margin of the Benton uplift (Purdue and Miser, 1923; Miser and Purdue, 1929). At Oklahoma localities—Black Knob Ridge, the Potato Hills, and the Broken Bow uplift—the contact is reported to be gradational (Goldstein and Hendricks, 1962).

Deep-water shales and sandstones intermixed with scattered debris flows compose the bulk of the Stanley Shale (the formation has group stratigraphic rank in Oklahoma) (Fig. 8). Continued deposition of similar sandstones and shales formed successively: the Jackfork Formation (also a group in Oklahoma), the Johns Valley Formation, and the Atoka Formation (Fig. 8) (Cline, 1960; Morris, 1974, this volume). Petrographically, the sandstones of the Ouachita Mountains are similar to those of the Black Warrior basin (Graham and others, 1976)—predominantly quartzose—but the Stanley contains a significant percentage of metamorphic rock fragments and a lesser percentage of feldspar, mostly albite (Morris, this volume). Sandstones of the Jackfork are typically quartzose, whereas sandstones of the Atoka are typically micaceous. In outcrops assigned to the Johns Valley Formation, especially, exotic blocks of carbonate rock from Paleozoic formations of the Arbuckle and Ozark regions of the cratonic shelf lie in shales deposited in deep water.

The Carboniferous formations of the Ouachita Mountains are only sparsely fossiliferous, and stratigraphic correlations are uncertain from region to region, being based in part on the ages of contained exotic blocks (Ethington and others, this volume; Morris, this volume). Formation names were assigned to the Carboniferous strata within the different regions of the Ouachita Mountains because of similarities in the stratigraphic succession and rock composition, but the faunal evidence for equivalency is nebulous. The age of the Johns Valley Formation is especially in question, for assignments of stratigraphic position have ranged from upper Chesterian–lower Morrowan (Cline and Shelburne, 1959) to lower Morrowan (Walthall, 1967) to upper Morrowan (Gordon and Stone, 1977).

In the Ouachita Mountains, the aggregate thickness of the Carboniferous strata is as much as 16,000 m and perhaps greater (Figs. 3, 8) (Morris, this volume). Older Carboniferous formations, the Mississippian Stanley and the Pennsylvanian Jackfork, are thicker in the central and southern Ouachita Mountains than in the frontal Ouachitas (Fig. 9). The youngest Carboniferous formation, the Atoka, is thickest along the boundary of the Ouachita thrust belt, as defined by the Choctaw thrust fault (Fig. 9; Plate 8), and the Arkoma basin, and it clearly exhibits a northward shift of the depositional axis. Rates of deposition of the Carboniferous strata were extremely high. Using thicknesses uncorrected for compaction, Houseknecht (1986) estimated rates of deposition per million years of: 106 m for the Stanley, 420 m for the Jackfork, and 1,000 m for the Atoka (Fig. 10).

The synorogenic Carboniferous formations of the Ouachita Mountains, south of the Ti Valley and Y City thrust faults of the frontal imbricate zone (Plate 8), form a complex of generally westward-prograding submarine fans deposited along the axis of the closing ocean basin (Fig. 11) (Moiola and Shanmugam, 1984; Morris, this volume). In addition, many marginal fans entered the main trough from point sources along the flanks. In any single stratigraphic section, proximal and distal facies variously succeed one another, and where separate fans intertongue, different facies of different fans are in vertical succession. An additional complicating factor in the Ouachitas is that the remnant ocean probably included several different types of basins: abyssal sea floor, trench floor, slope basins, and fore-arc basins. Turbidite facies deposited in one basin may be identical to turbidite facies deposited in another, and therefore, tectonic position must be considered in interpreting the environment of deposition. Viele (1979a) suggested the synorogenic strata lying south of the Benton uplift were deposited in a fore-arc basin, whereas those to the north were deposited in trench floor and slope basin environments. Yet even this scenario is too simple, for the flood of clastic sediment into the remnant ocean was so great the youngest of the flysch units lapped onto the southern margin of the North American craton. Modern analogs for the Carboniferous environments of deposition may be those subduction complexes so choked with sediment that the physiographic definition of separate parts is muted.

In our view, the Atoka Formation, the youngest of the Carboniferous flysch units, was deposited in several different basins. The Atokan strata of the southern Ouachitas of Arkansas and the central Ouachitas of Oklahoma compose the youngest exposed filling of a fore-arc basin. The Atokan strata of the frontal Ouachitas of Oklahoma, however, were deposited in an abyssal trough and extended northward with stratigraphic continuity into the Arkoma basin, a peripheral foreland basin lying above continental crust (Figs. 10, 11; Plate 9; Plate 11, cross sections C-C′, D-D′, E-E′). Thrust faulting has foreshortened the transition. Furthermore, within the Arkoma basin, the Atokan strata record several different depositional environments (Fig. 11), ranging from a shallow-water, muddy-slope facies on the north to a deep-water submarine fan facies on the south (Zachry, 1983; Houseknecht, 1986).

Changes of thickness of the Atoka Formation are abrupt and related to deposition across syndepositional high-angle faults, for the most part down-to-the-south (Figs. 11, 12). The faults were active as early as Late Mississippian (D. Houseknecht, personal communication, 1988) in the easternmost subsurface part of the Arkoma basin, which now lies beneath the Mississippi Embayment (Fig. 12; Plate 9), and migrated in space and time to the west and north. In central Arkansas, Morrowan strata thicken southward across the faults, but in eastern Oklahoma, middle

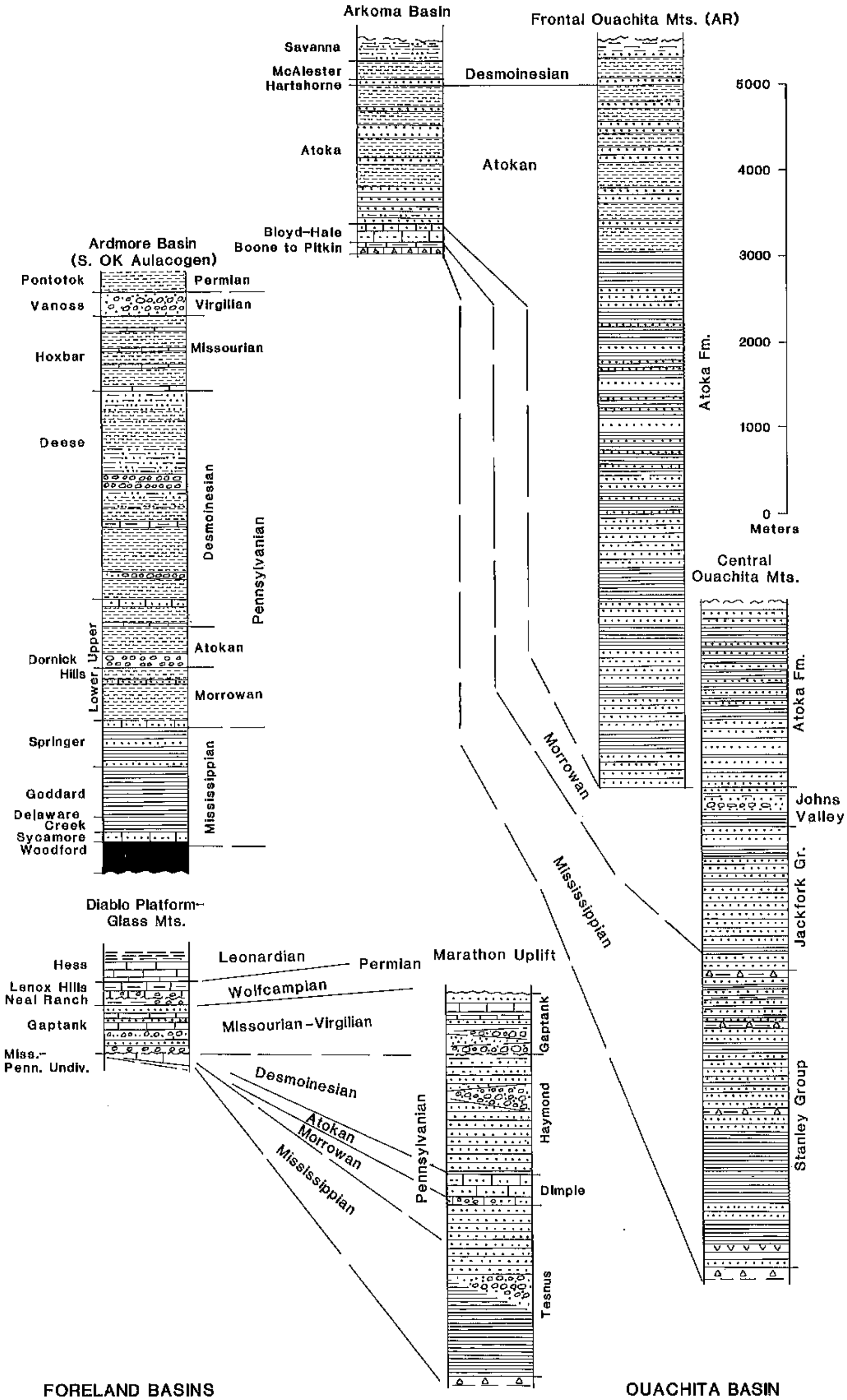


Figure 8. Generalized stratigraphic sections of Carboniferous strata from the Ouachita Mountains and the Marathon region compared to nearby areas on the North American shelf. After Arbenz (1989, Fig. 6).

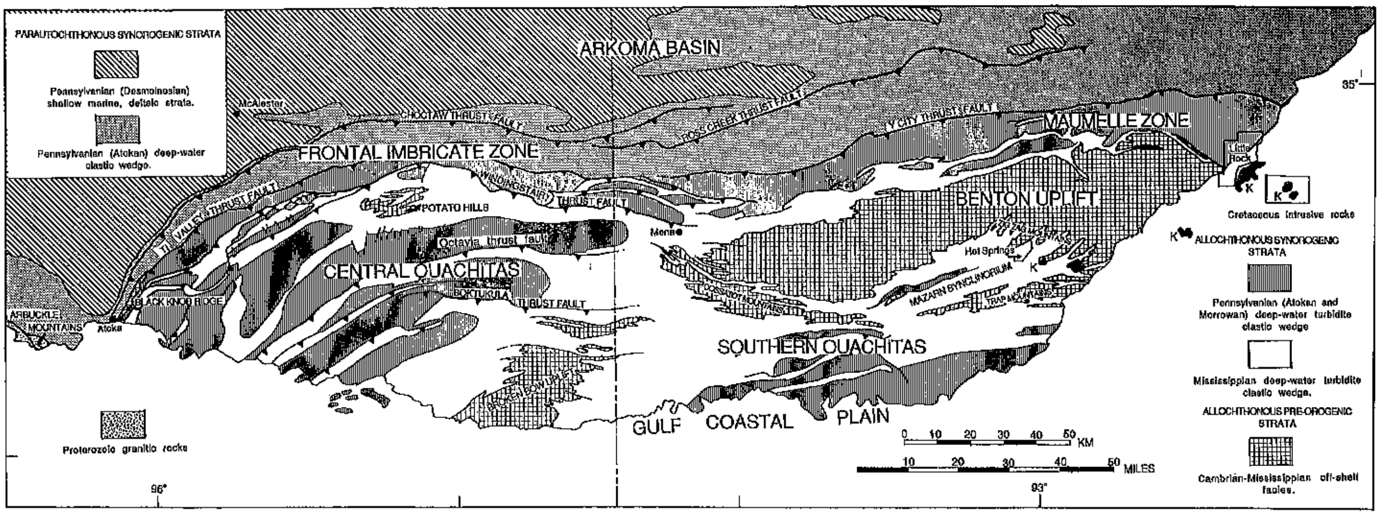


Figure 9. Index map of the Ouachita Mountains of Arkansas and Oklahoma.

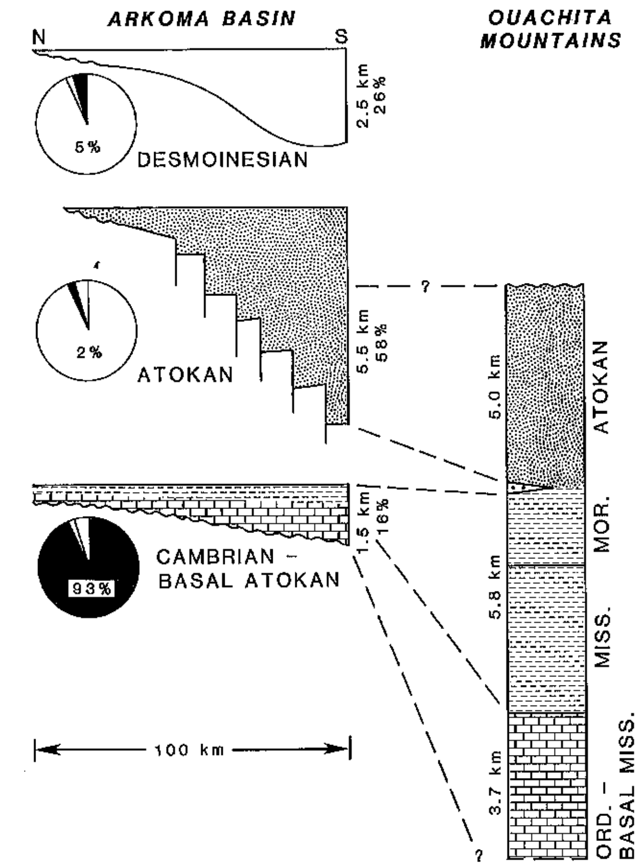


Figure 10. Tectonic stratigraphy of the Arkoma basin and Ouachita Mountains showing the type of basin subsidence within the Arkoma basin. Patterns show genetically related strata and do not imply composition. Pie diagrams show the amount of time represented by each genetic package of strata, expressed as a percentage of total Paleozoic time recorded by strata. Note especially the rapid Mississippian-Morrowan deposition in the Ouachita Mountains and the rapid Atokan deposition in the Arkoma basin. After Houseknecht (1987).

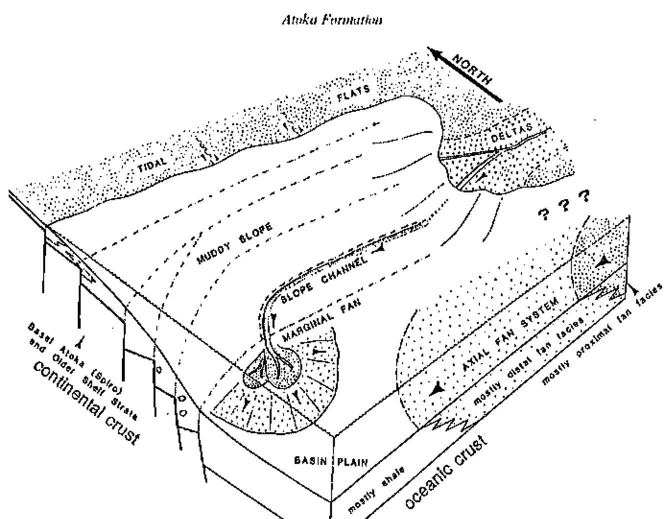


Figure 11. Environments of deposition of Atoka Formation in the Arkoma basin. Slightly modified from Houseknecht (1986).

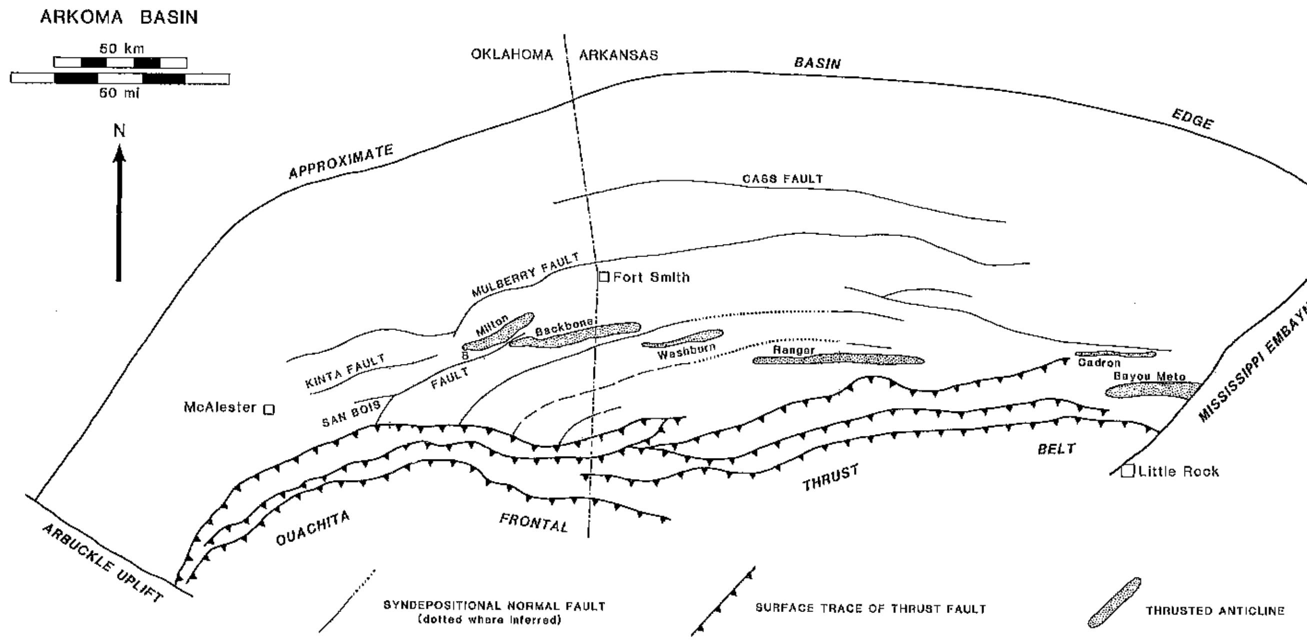


Figure 12. Index map of structures of the Arkoma basin and the frontal thrust fault belt of the Ouachita Mountains. After Houseknecht (1986).

Atokan strata thicken southward across the faults. As the faulting migrated westward and northward, so did deposition of the deep-water facies of the Atoka. In central Arkansas, in the area of the Bayou Meto anticline, sediments of the lower Atoka were deposited in deep water, but in eastern Oklahoma, in the area of the San Bois fault (Fig. 12), sediments of the lower Atoka were deposited in shallow water. Not until middle Atokan time, in eastern Oklahoma, did the basin floor drop and deep-water sediments accumulate (Houseknecht, 1986).

Middle Atokan strata, especially, increase abruptly in thickness by as much as 1,000 m across the high-angle faults. Nevertheless, neither well-log correlations nor seismic reflection profiles provide any evidence of erosion of either the fault scarp or the hanging wall block. Rates of deposition were at least equal to rates of tectonic displacement, and the faults were never high-relief physiographic features (Houseknecht, 1986).

Shallow-water deposits of the upper Atoka rapidly infilled the ancestral Arkoma basin and grade without intervening unconformities into overlying Desmoinesian sandstones. In these Desmoinesian sandstones, patterns of deposition begun during middle Atokan time were maintained: the sediment came primarily from point sources along the flanks of the basin, but once within the basin, the main direction of transport was westward down the basinal axis. The Desmoinesian strata thicken gradually southward without abrupt changes of thickness across high-angle faults now apparently inactive. The primary environments of deposition were tidally dominated deltas and fluviatile channels (Houseknecht and others, 1983). Clasts of slate in the Desmoinesian sandstones possibly came from the region of the Ouachita Mountains, the first indication of uplift and erosion of the Ouachitas.

South of the Ouachitas, in the subsurface of Arkansas and Louisiana, the Desmoinesian strata of the Texarkana platform (Fig. 1; Plate 9) (Meyerhoff, 1973) are mostly shallow-water fossiliferous limestones and fine-grained clastic rocks (Plate 9) (Vernon, 1971; Nicholas and Waddell, this volume) resting with angular unconformity across broad folds of the Atoka Formation.

In our view and in the view of Houseknecht (1986), which are opposite to those of Morris (this volume), the composition of the framework grains indicates primarily a southern source for the Carboniferous sandstones of the Ouachita Mountains and the Arkoma basin. The provenance was made up of multiple terranes: recycled orogen, accretionary wedge, and volcanic arc. In late Meramecian time, sediment dispersed from this provenance first arrived in the Ouachita remnant ocean, and in Chesterian time, it arrived in great quantity. By Atokan time, the southward-derived sediment was spilling northward into the Arkoma basin. In the Ozark region of the cratonic shelf, however, thin limestones and shales were being deposited during the Meramecian (Sutherland and Manger, 1979). Sediment was introduced from the north to the Ouachita and Arkoma basins during the Pennsylvanian Period, primarily through the ancestral Illinois basin, but probably in lesser quantities than from the south and east (Plate 9) (Houseknecht, 1986). Once in the Ouachita trough, whatever the source, the sediments prograded westward down the axis of the remnant ocean.

### *Carboniferous Strata of Ouachitas in Texas*

In southeastern Oklahoma, the Carboniferous strata of the Ouachita Mountains pass southward beneath the Mesozoic and Tertiary beds of the Gulf Coastal Plain. Ouachita rocks do not appear at the surface between southern Oklahoma and the Marathon uplift, but they are present in many wells in eastern Texas (Fig. 7) (Nicholas and Waddell, this volume, Fig. 1). Wells drilled in the northern and central segments of this belt penetrated mostly Stanley Shale in the subsurface frontal zone of the Ouachita orogen (Plate 9) (Flawn and others, 1961). The Stanley contains tuffaceous material and is otherwise similar to the Stanley of the western Ouachita Mountains. The Atokan sandstones, however, are somewhat different from their counterparts in the Ouachita Mountains in that, close to the subsurface tectonic front, they contain abundant grains of slate, phyllite, and metaquartzite. Present as well are clasts of chert, similar to the chert of the Big Fork Chert and Arkansas Novaculite, and clasts of vein quartz (Flawn and others, 1961). In general, toward the south in the subcrop belt, the feldspar content in the Carboniferous strata increases, and many of the sandstones are true arkoses. Potassium feldspar appears, as do faded and bleached biotite and fragments of a microgranular, feldspathic igneous rock (Flawn and others, 1961).

The age of closing or filling of the Ouachita ocean in the subsurface belt of eastern Texas appears to be somewhat earlier than in the Ouachita Mountains of Arkansas and Oklahoma. In the Fort Worth or Strawn basin (Fig. 1), a peripheral foreland basin described by Crosby and Mapel (1975) and Lovick and others (1982), Atokan strata are predominantly fluviatile and deltaic. Desmoinesian strata contain coals, red beds, and conglomerates containing clasts of chert possibly eroded from uplifted Ordovician-Devonian strata of the Ouachita orogenic belt. Relative to the Arkoma basin, post-orogenic coarse clastic sediments from the Ouachitas entered the Fort Worth basin at an earlier date and were substantially coarser than those in the Arkoma basin.

Farther to the south, the peripheral Kerr basin (Fig. 1) contains 2,100 to 2,500 m of Pennsylvanian rocks (Crosby and Mapel, 1975). The oldest rocks of the basin, probably Atokan, record a continuous progradation from a muddy to a sandy sea floor. Carbonate rocks rimmed the western and northern margins of the basin in Desmoinesian time until muds and sands covered them in the late Desmoinesian. The clastic sediment came from the area of the Ouachita folded belt, but it is not as coarse grained as that of the Fort Worth basin.

Northwest of the Kerr basin, the Val Verde basin (Fig. 1), a narrow peripheral basin lying northeast of the Devils River uplift, contains about 4,300 m of Upper Pennsylvanian–lower Wolfcampian deep-water sandstone and shale (King, 1975), similar to the middle Atokan rocks of the Arkoma basin. The rocks of the Val Verde basin lie above continental crust on the northeast side of the Devils River uplift.

In the Marathon region, as in the Ouachita Mountains, rapid deposition of Carboniferous turbidites followed the slow accumulation of the middle Paleozoic siliceous succession (McBride, this volume). The lower contact of the turbidites with the Caballos Novaculite is conformable and locally gradational. The Carboniferous strata of the Ouachita Mountains and the Marathon region are essentially stratigraphic equivalents, but some differences are apparent (Fig. 8). First, within the Marathon uplift, the turbidites are much thinner, having a maximum thickness of about 4,500 m, or only about one-fourth that of equivalent strata in the Ouachita Mountains. As in the Ouachita Mountains, however, the strata thicken away from the craton. The Morrowan Dimple Limestone of the Marathon succession (McBride, this volume) contains a carbonate-shelf facies along the northern margin of the Marathon region, a slope facies to the south of the shelf-carbonate facies, and a basinal facies still farther south, all transitional with one another (Thomson and Thomasson, 1964; Ross, 1986; McBride, this volume, Fig. 11). A similar transition is not known in the Lower Pennsylvanian strata of the Ouachita Mountains, probably because of foreshortening by thrust faults. A point of similarity, however, with the Carboniferous strata of the Ouachita Mountains is that exotic clasts from shelf and platform rocks to the north and northwest are present in both the Mississippian and Lower Pennsylvanian flysch of the Marathon region.

Quite different are some boulder beds in the Pennsylvanian Haymond Formation, for they provide evidence for an orogenic highland on the outboard side of the Marathon basin of deposition. Large clasts of novaculite lie in the boulder beds in the company of clasts of Devonian metamorphic rocks (Denison and others, 1969). No metamorphic rocks of a similar nature are known in the Carboniferous strata of the Ouachita Mountains.

Petrographically, the Carboniferous rocks of the Marathons and Ouachitas are similar. In order of decreasing percentage, the framework grains of the Tesnus are quartz, feldspar, and metamorphic rock fragments; the framework grains of the Haymond are the same, although the rocks are somewhat more feldspathic and less quartzóse (McBride, this volume). Ross (1986) noted that grain size increases markedly in the upper Tesnus and that metamorphic and granitic rock fragments increase in abundance. In general the grain size of the Carboniferous sandstones of the Marathon region increases upward, probably signifying the closer approach on the south and southeast of an orogenic terrane.

## VOLCANIC AND IGNEOUS ROCKS

The oldest known igneous rocks in the Ouachita orogenic belt are serpentinites (Sterling and Stone, 1961) found in the Benton and Broken Bow uplifts. Because they are described in another chapter (Nielsen and others, this volume), suffice it to say here that in the eastern end of the Benton uplift, several isolated pods of highly altered, nickel-bearing serpentinite lie at or near the base of the Paron nappe (Fig. 13) (Nielsen and others, this volume, Fig. 1). The pods are small, averaging about 100 m in length (Mullen, 1984), and float as unattached masses in Ordovician cherts and black slates. The serpentinites and surrounding rocks alike show the impression of later folding and faulting. Along strike, some 10 km southeast of the serpentinite pods, are outcrops of alkalic metagabbro (Morris and Stone, 1986; Nielsen and others, this volume). Similar rocks, first noted by Honess (1923), occur in the Broken Bow uplift. Contact metamorphic zones are lacking, and the gabbros show the impression of later deformation. Morris and Stone (1986) reported a whole-rock K-Ar date of 1025 ± 10 Ma and suggested the metagabbros may be transform-related fragments of oceanic crust. Excepting igneous fragments in debris flows, no other pre-Carboniferous igneous rocks are known in outcrops of the Ouachita orogenic belt, although Wilson (1954) mentioned a pre-Cretaceous basaltic intrusion in Ordovician strata in the southeastern part of the Marathon region.

In southern Texas, in the northwest-trending segment of the subcrop belt of Ouachita rocks and the immediately adjacent foreland (Figs. 1, 7), nine wells have encountered andesitic, basaltic, and granitic rocks (Flawn and others, 1961). Some of the basaltic rocks have been described as greenstones. All samples have been sheared and partially sericitized and chloritized. The only certainty about the age of these rocks is that they pre-date late Paleozoic shearing and metamorphism.

In the basal part of the Mississippian Stanley Shale of western Arkansas and eastern Oklahoma, several beds of rhyodacitic tuff contain textures and structures indicating deposition by sub-sea ash flows and falls (Fig. 8) (Niem, 1977). The tuff beds increase in thickness and number toward the south. In the stratigraphically equivalent Tesnus Formation of the Marathon uplift, beds of tuff are lacking, although some tuffaceous material is found in the upper third of the formation (McBride, this volume). In addition, Decatur and Rosenfeld (1982) have described three thin, radioactive markers widespread in the Mississippian shales of western Texas and have attributed the markers to explosive volcanic events in the Ouachita orogen. Pennsylvanian formations of the Ouachita orogen are generally feldspathic in varying degree but contain no known beds specifically designated as pyroclastic.

In northern Louisiana, in the subsurface of the Sabine uplift, three wells have penetrated rhyolite porphyry lying beneath fossiliferous Desmoinesian carbonate rocks (Fig. 1) (Nicholas and Waddell, 1982, this volume). The stratigraphic position and composition suggest that they lie at or near the source vents of the tuffs in the basal Stanley Shale. They provide direct evidence for Carboniferous volcanism to the south of the Ouachita orogenic belt.

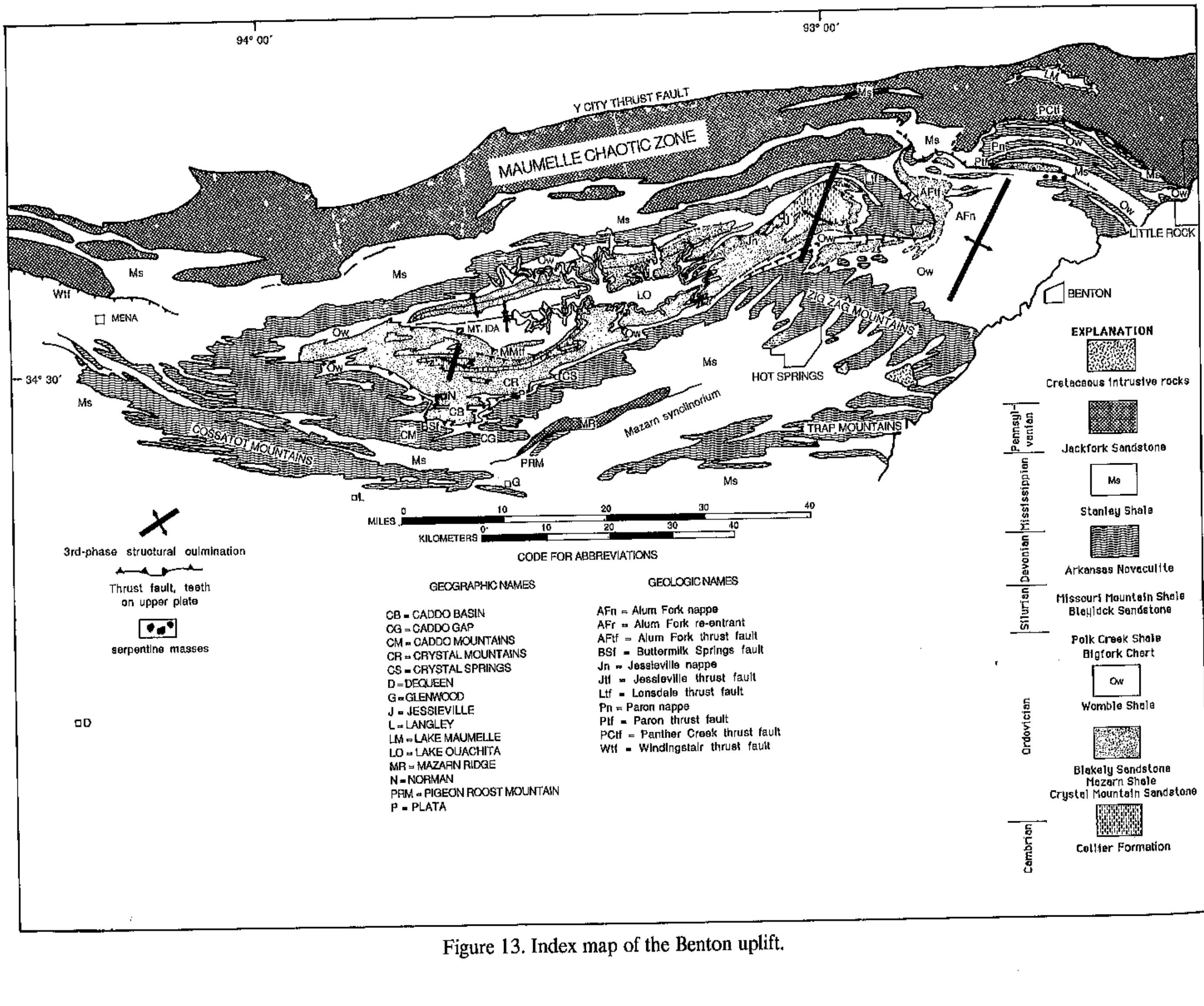


Figure 13. Index map of the Benton uplift.

## OUACHITA STRUCTURAL PROVINCES

Our review of the structural provinces of the Ouachita orogenic belt follows the same plan as that for the Carboniferous stratigraphy, from east to west along the orogenic belt from the Black Warrior basin through the Ouachita Mountains to the Marathon region of Texas. Along this trend the Ouachitas exhibit many differences in tectonic style, which are in part a theme of this chapter.

### *Black Warrior Basin of Alabama and Mississippi*

In the subsurface Black Warrior basin, the strata dip homoclinally southwestward at less than two degrees (Plate 9) (Thomas, 1985, 1988a, 1988b) and are broken by a system of northwest-trending normal faults, dropping the homocline down to the southwest by more than 2 km. From central Mississippi westward, the belt of normal faults is bordered successively on the south by a frontal thrust belt of shaly clastic rocks and farther south by slates bearing quartz veins. Subsurface control does not permit exact definition of the tectonic front, but by analogy with outcrops in Arkansas, it is shown as a thrust front of irregular trace, and the rocks south of it are interpreted to be part of the Ouachita orogenic belt (Plate 9). Southeastward from central Mississippi, in the subsurface, the "Ouachita" slates pass beneath thrust sheets containing a stratigraphic sequence like that in the Appalachian outcrops in Alabama (Thomas, this volume). These thrust sheets, in turn, dip beneath subsurface metamorphic rocks of the Talladega slate belt (Plate 9) (Thomas, 1973, 1985). The cross-cutting relations show that the frontal thrust faults of the Ouachitas pre-date the Alleghanian thrust faults of the southern Appalachians.

### *Arkoma Foreland Basin*

Unlike the subsurface Black Warrior basin, where only high-angle faults are known, both high-angle and thrust faults are known in the Arkoma basin (Fig. 12; Plate 8). Syndepositional high-angle faults and associated drape folds are the predominant structures in the northern part of the basin and are known in the subsurface of the central and southern parts of the basin through drilling and geophysical exploration (Koinm and Dickey, 1967; Berry and Trumbly, 1968; Buchanan and Johnson, 1968; Houseknecht, 1986). In the central and southern parts of the basin, the high-angle faults lie below the regional detachment surface of a broad belt of north-verging thrust faults (Plate 11, cross sections C-C′, D-D′, E-E′).

In map view, the trace of the high-angle faults generally parallels the trend of the Ouachita thrust belt on the south, but some high-angle faults swing to northeasterly trends near the Mississippi Embayment, and many turn southward in Oklahoma, intersecting the front of the thrust belt at near right angles (Fig. 12). In the eastern end of the Arkoma basin, other high-angle faults trend southeastward, evidently connecting with the high-angle faults of the Black Warrior basin (Plate 9).

In the Arkoma basin, most high-angle faults are down toward the south, but some face north, forming grabens overlain in several places by broad synclines. The faults offset the underlying basement (Buchanan and Johnson, 1968) and, within the limits of resolution of seismic reflection surveys (Lillie and others, 1983), are not visibly listric. Although the high-angle faults of the Arkoma basin were syndepositional, they are clearly unlike the rotational slump faults of the northern Gulf Coast. The high-angle faults of the Arkoma basin drop the North American basement from depths of approximately 1 km in the northern part of the basin to depths below 10 km at the northern margin of the Ouachita thrust belt (Lillie and others, 1983). High-angle faults are inferred to be present to the south beneath the Ouachita thrust sheets. It seems likely that the upper parts of some of the high-angle faults were cut off and translated northward by thrust faults, but allochthonous sheets containing rootless high-angle faults have not been specifically identified.

In the central and southern Arkoma basin, numerous thrust faults reach the surface in anticlines such as the Backbone and Washburn (Fig. 12; Plate 8; Plate 11, cross sections C-C′, D-D′, E-E′), and blind thrust faults lie beneath other surface anticlines. Historically, the Choctaw and Ross Creek thrust faults have been used to define the northern margin of the Ouachita thrust belt, but other thrust faults extend north of these faults. Within the broad zone of compressional surface structures in the Arkoma basin, narrow, upright (though slightly north-vergent) anticlines lying en echelon to one another (Fig. 12) separate broad, box-shaped synclines. The thrust faults are listric; they dip steeply southward beneath the anticlines and flatten toward the south. The direction of translation was predominantly northward, although local "triangle zones" contain south-directed thrusts. The major detachment surface of the frontal thrust faults is in the Atoka Formation, for only locally are Morrowan beds present in the hanging walls. In the Arkoma basin, the time of last thrust fauting post-dates the deposition of the Desmoinesian Boggy Formation (Plate 8), which is involved in the related folding. Within the limits of the basin, rocks post-dating the faulting are unknown.

### *Ouachita Thrust Belt*

The traditional markers of the northern edge of the Ouachita thrust belt—the Ross Creek and Choctaw faults—repeat parts of the Atoka Formation and are associated with reversals of dip and "triangle zones" (Arbenz, this volume). In addition, they mark the northern margin of a zone of closely spaced imbricate faults and duplexes shown as the frontal imbricate zone (Fig. 9; Plate 8; Plate 11, cross sections C-C′, D-D′, E-E′). In Oklahoma, the Windingstair thrust fault marks the interior margin of the frontal imbricate zone, separating it from the northern central thrust belt on the south. As the Windingstair fault disappears to the east, the northern margin of the Benton uplift is taken as the southern margin of the frontal imbricate zone in Arkansas.

The thrust faults of the Ouachitas exhibit typical "sled runner" profiles (Arbenz, this volume). In the Oklahoma segment of the frontal imbricate zone (Plate 8; Plate 11, cross sections D-D′, E-E′), the thrusts are stacked in multilevel duplexes, and the strata within the thrust sheets are complexly folded. Numerous imbricate thrust sheets repeat Morrowan strata between the Choctaw and Ti Valley thrust faults. At Black Knob Ridge and in the Potato Hills, outcrops of Ordovician Womble Shale in the allochthonous sheets indicate upward ramping of the thrust faults from stratigraphic levels at least as deep as the Womble Shale (Arbenz, 1968).

In the central zone of the Ouachita Mountains, south of the Potato Hills (Fig. 9), there is less evidence for ramping and duplexing of thrust faults. The major structures are broad synclines outlined on the surface by massive sandstones in the Pennsylvanian formations. The synclines plunge gently to the west or southwest, and some are truncated on the southern limbs by thrust faults of apparently small displacement. The thrusts emplace the Mississippian Stanley Shale over Morrowan to Atokan strata. Relative to the frontal imbricate zone, the faults are widely spaced.

The same pattern persists south of the Benton uplift in the southern Ouachita Mountains (Fig. 9, Plate 8). In this region, Carboniferous strata dip homoclinally southward, excepting local small folds, and are broken by thrust faults that repeat partial stratigraphic sections of the Stanley Shale or thrust the Stanley over Morrowan beds, although some faults thrust Morrowan beds over Atokan strata.

### *The Mélange-Olistostrome Problem*

In the Carboniferous strata of the Ouachita Mountains, the most complex zones of deformation lie in the frontal imbricate zone of the Ouachita thrust belt. The tectonic setting of this zone has been the subject of a long and heated geological discussion, mostly about the style of deformation of the rocks in the Maumelle chaotic zone (Viele, 1973, 1979a) and the origin of the Johns Valley Formation. As this argument is germane to the origin of the entire Ouachita orogenic belt, the rocks of the Maumelle zone and Johns Valley Formation will be discussed here in some detail, although they are described in other chapters of this volume (Morris, this volume; Nielsen and others, this volume).

The Maumelle chaotic zone lies between the Y City thrust fault and the northern margin of the Benton uplift of Arkansas (Fig. 9; Plate 8). Recent mapping in the eastern part of the frontal Ouachitas of Oklahoma shows that the Maumelle zone extends at least that far west (Underwood and Viele, 1985; Poole, 1985; McDonald, 1986). In Arkansas, the zone, which is easily visible on radar and Landsat mosaics, is one of broken, irregular topography, standing in sharp contrast to the linear ridges and valleys of the region north of the Y City fault. The Maumelle zone is developed entirely in Carboniferous sandstone and shale, mostly in the Jackfork Formation near Little Rock, but also in Mississippian through Atokan strata in western Arkansas, and in the Atoka Formation in the eastern part of the frontal imbricate zone of Oklahoma.

In many outcrops within the Maumelle zone, along the northern side of the Benton uplift, the structural fabric is best described as chaotic (Nielsen and others, this volume, Fig. 2). Shears are ubiquitous (Haley and others, 1976), and most seen in outcrop dip northward. Clasts of sandstone are studded throughout a matrix of shale that is swirled and folded. The shales are scaly and polished. Tight to isoclinal folds in thin beds of sandstone exhibit curvilinear hinges and boudinaged limbs that either pinch out or terminate against shears. Almost all clasts of sandstone exhibit a tracery of web structures (Cowan, 1982) seen in thin section to be cataclasite zones. Tongues of shale intrude fractures in the clasts of sandstone. Some large clasts of stratified sandstone and shale end against shear surfaces, and many beds of sandstone lie in fault contact with the underlying shale (Nielsen and others, this volume, Fig. 3). The evidence is strong that the sandstones were at least partially lithified at the time of deformation. The structures of the Maumelle zone are remindful of the structures described in many accretionary prisms (Carson and others, 1982; Karig, 1983; Byrne, 1984; Raymond, 1984; Sample and Moore, 1987).

On a regional scale, the Maumelle zone grades along strike into belts of tight folding and well-developed cleavage. It lies above a North American basement that has dropped at least 9 km from the northern part of the Arkoma basin (Nelson and others, 1982) to an estimated depth of 10 km beneath the frontal imbricate zone. In western Arkansas, along the trace of the COCORP profile (Lillie and others, 1983), the Maumelle zone coincides with a zone of poor seismic reflections, and even farther west in Oklahoma, it lies along the trend of a broader gravity minimum (Plate 10).

Several workers (Haley and Stone, 1982, 1985; Morris, this volume) have viewed the rocks and structures of the Maumelle zone as a mappable stratigraphic unit made up of olistostromes deposited as surficial slides from a belt of south-facing fault scarps. These authors acknowledge that the beds have undergone later translation along thrust faults, but they argue that the structures in outcrop formed primarily by gravity sliding of unconsolidated muds and sands.

While acknowledging the probable existence of local olistostromes in these rocks, Viele (1979a; Nielsen and others, this volume) views the Maumelle zone primarily as a tectonic mélange formed at the toe of an accretionary prism. He disagrees with the statement that individual beds or parts of a stratigraphic section can be mapped for any distance. Underwood (1984) has noted that although submarine slides are common on the slopes of accretionary prisms, most such slides involve only the slope-covering mud; sands may accumulate locally in trench-slope basins but, viewed on a regional scale, they flow down to the trench floors where they have no potential for deformation by gravity sliding. The aggregate of the evidence, ranging from outcrop-scale structures to regional tectonic relations, bespeaks tectonic deformation of confined, partially lithified to non-lithified sediments. I

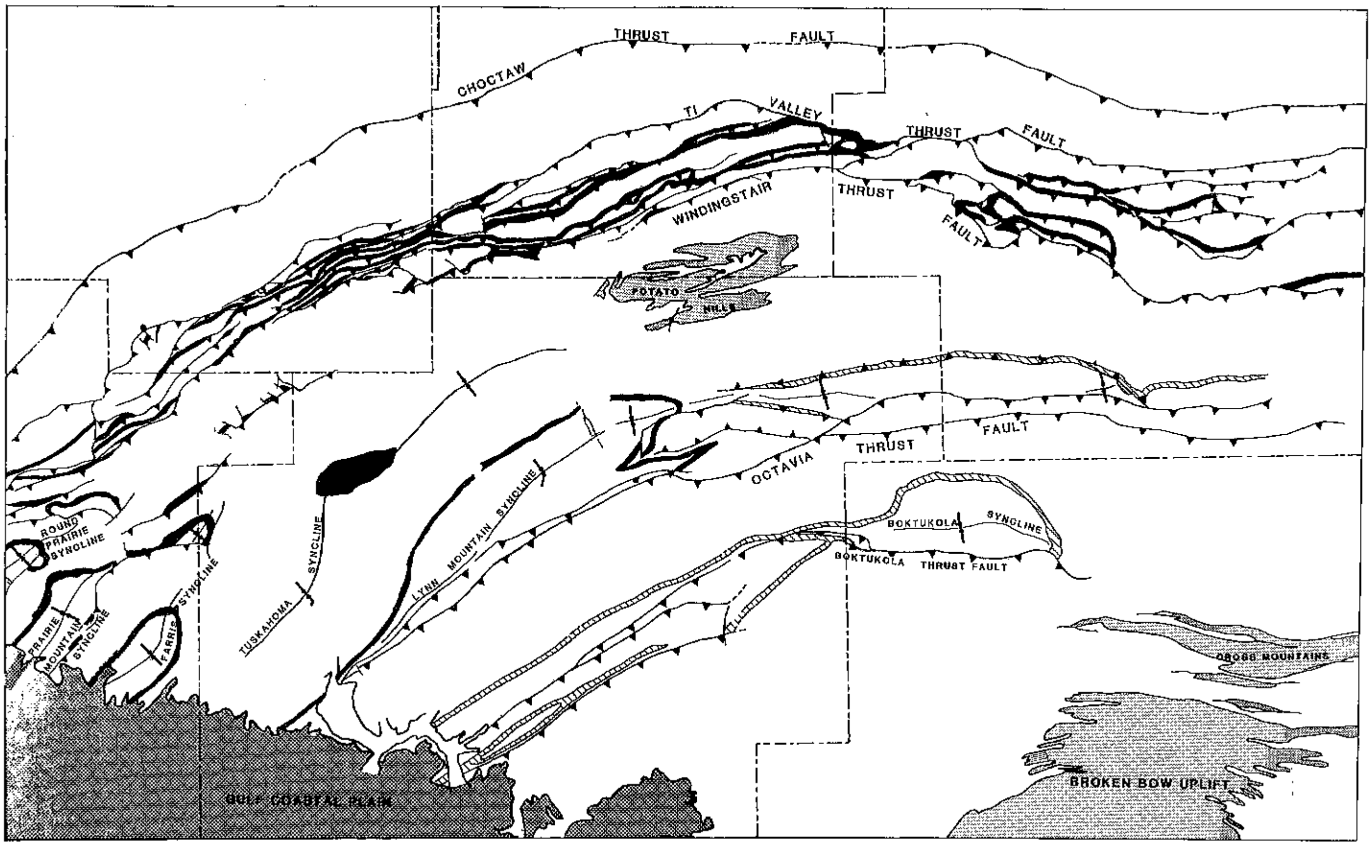


Figure 14. Distribution in Oklahoma of boulder-bearing (solid black) and non-boulder-bearing (diagonal lines) Johns Valley Formation. Most outcrops containing clasts of shelf carbonate rock lie north and west of the Windingstair thrust fault, but a few have been seen to the south and east of the Windingstair. The type locality of the Johns Valley Formation is in the Tuskahoma syncline.

is not necessary to posit an earlier period of general gravity sliding to account for the structures.

Accretionary structures and fabrics are absent from the many olistostromes in the synorogenic Carboniferous strata outside the frontal imbricate belt. Olistostromes are common in these strata in the central and southern parts of the Ouachita Mountains, but they are free of scaly shales, clasts of sandstone with web structures, etc. The mélange belts of the frontal imbricate zone should not be equated to the olistostromes in the Carboniferous strata of the central and southern Ouachita Mountains.

The origin of the Johns Valley Formation and especially of the exotic clasts within it has been the subject of similar discussion. In a belt bounded by the Octavia and Ti Valley thrust faults (Fig. 14), the Johns Valley contains clasts of shelf-carbonate rock, ranging in age from Late Cambrian to Early Pennsylvanian, derived from the area of the Arbuckle Mountains and Ozark plateau (Shideler, 1970). Indeed, during field mapping, the very presence of such clasts may have resulted in the assignment of strata to the Johns Valley Formation (Moiola and Shanmugam, 1984). South of the boulder-bearing belt, the Johns Valley is bounded above and below by the Atoka and Jackfork Formations, respectively, but it is largely free of boulders (Fig. 14) (Cline, 1960; Shideler, 1970). An additional complicating factor is that exotic clasts of carbonate rock are scattered throughout the Carboniferous formations, from the very base of the Stanley in the Mazarn synclinorium (Fig. 13) (Gordon and Stone, 1977) to well up in the Atoka Formation north and west of the Rich Mountain syncline (Plate 8) (McDonald, 1986).

The clasts themselves exhibit a wide range of characteristics. In some outcrops the carbonate boulders are rounded, roughly graded, and lie in channels; they are excellent examples of olistostromes. In other outcrops, the exotic blocks exhibit several directions of slickenlines (Ulrich, 1927), are truncated by shear surfaces, and are laced with web structures; the enclosing shales are scaly and locally exhibit an incipient cleavage. According to Shideler (1970), most of the clasts came from the Arbuckle facies; lesser numbers came from the Ozark facies; and surprisingly, some came from the Arkansas Novaculite and Big Fork Chert of the Ouachita facies. Hendricks (1971) challenged these latter identifications, but Shideler (1971) strongly defended them, and more recent lists include them in the family of Johns Valley clasts (Stone and others, 1979).

Representative of the various views concerning the origin of the clasts are those of Cline and Shelburne (1959), who suggested ice-rafting as the mechanism of transportation; Van der Gracht (1931b) suggested they were a tectonic carpet formed beneath advancing nappes; and numerous authors (Powers, 1928; Miser, 1934; Moore, 1934; Goldstein and Hendricks, 1962; Shideler, 1970; Gordon and Stone, 1977) have suggested the boulders were deposited from submarine debris flows. Several authors (Shideler, 1970; Gordon and Stone, 1977) who favor debris flows suggest they came from a submarine ridge and scarp system, the Bengal high, that lay to the south of the outcrops of boulder-bearing beds beneath the Ouachita thrust sheets. Northward thrusting brought the boulder beds to their present position.

Viele favors the views of Van der Gracht (1931b). The fault-bounded outcrops of boulder-bearing Johns Valley exhibit structures typical of tectonic mélange. No known geophysical evidence has revealed the "Bengal high" beneath the thrust sheets. Moreover, if clasts of Arkansas Novaculite and Big Fork Chert were truly deposited by debris flows in the Johns Valley Formation, somewhere along the trend of the orogenic belt more than 6 km of Carboniferous strata lying between the Johns Valley and the Arkansas Novaculite must have been removed. Nowhere in the Ouachita Mountains of Arkansas and Oklahoma is there evidence of this erosion. Instead, during the Carboniferous, subsidence and deposition were rapid, and ponding sediments would have probably reduced the physiographic expression of any scarps in the remnant ocean.

Viewed from the modern perception of subduction zones, accretionary prisms, and associated mélanges, Van der Gracht's hypotheses seem straightforward. The boulders were tectonically ripped off the North American platform, which was being subducted under the advancing Ouachita thrusts. They were incorporated into the allochthonous sheets, which included rocks of Ouachita facies, and thrust northward to their present position in the frontal imbricate belt. They are similar in origin to the slivers of foreland carbonate rock contained in the mélanges of the Taconic Mountains of eastern New York (Vollmer and Bosworth, 1984).

### *Uplifts of Pre-Orogenic Strata*

Rocks older than the Mississippian Stanley Shale crop out in four areas of the Ouachita Mountains: at Black Knob Ridge, in the Potato Hills, in the Benton uplift, and in the Broken Bow uplift (Plate 8). At Black Knob Ridge, the pattern of folding is essentially harmonic from the lowest to the highest formations, although the more ductile shales exhibit thinner fold limbs and thicker hinge areas. The folds appear to be "fault-propagation" folds (Suppe, 1985) associated with imbricates off the Ti Valley thrust fault (Hendricks and others, 1937; Arbenz, this volume, Fig. 4).

The Potato Hills are a folded thrust sheet (Plate 11, cross section D-D′) of mostly pre-orogenic strata, although the lower part of the Stanley Shale is involved (Arbenz, 1968). The folding may be related to piggy-back transport of a higher thrust sheet on the deeper and younger Windingstair thrust fault. The folded thrust sheet forms an essentially harmonic stack (Suppe, 1985), broken and offset locally by faulting. The folds are mostly tight chevrons and are nearly upright, verging slightly to the north. At Black Knob Ridge and in the Potato Hills, the rocks lack cleavage and in thin section reveal only hints of recrystallization of illitic clay minerals.

The oldest rocks in the Ouachita Mountains are in the Benton and Broken Bow uplifts of Arkansas and Oklahoma, respectively (Figs. 10, 15; Plate 8) (Nielsen and others, this volume). The boundaries of the uplifts, as traditionally mapped, are largely a matter of cartographic convenience, being drawn at the contact of the Arkansas Novaculite and the Stanley Shale (Plate 8). Folds and zones of cleaved rock consistently cross this contact and extend out into the carapace of Carboniferous rocks. The plunge of the Benton uplift is westward in the general direction of the Potato Hills, whereas the eastern flank of the Broken Bow uplift plunges eastward to the south of the Benton uplift, and the western flank plunges westward and southwestward (Plate 8). Fold hinges within the two uplifts do not trend toward one another. The Benton and Broken Bow uplifts do not define a linear structural belt but are instead separate culminations of pre-orogenic strata underlying the entire area of the Ouachita Mountains.

Relative to the overlying Carboniferous strata, the rocks of the Benton and Broken Bow uplifts reveal a complex structural history comprising three and possibly four phases of deformation (Fig. 15) (Nielsen and others, this volume). Within the eastern part of the Benton uplift, the first-phase structures consist of several northward-directed fold and thrust nappes stacked on top of one another (Fig. 15) (Viele, 1966, 1973, 1979b). Thrust faults bounding the nappes only locally extend beyond the boundaries of the Benton uplift out into the Carboniferous strata. The displacement of the nappes relative to one another is not known. The strata in the nappes are primarily the pre-orogenic deep-water rocks deposited off the North American craton. The higher nappes contain the southern facies of the Arkansas Novaculite (Lowe, this volume) and probably rooted farthest to the south. Near the base of the highest nappe (Fig. 13) (Nielsen and others, this volume), several small pods of altered serpentinite and slices of metagabbro lie within the pre-orogenic strata. The serpentinites and gabbros probably were torn from underlying oceanic crust and incorporated into the nappe as it moved northward. The first-phase, northward-directed fold and thrust nappes probably formed above oceanic crust off the southern margin of the North American craton.

In the second phase of deformation, the stack of first-phase nappes was backfolded (Fig. 15). Upright and overturned fold limbs, axial surfaces, and the bounding thrust faults of the nappes were all folded and overturned back toward the south. These second-phase folds constitute the predominant structure seen in outcrop throughout the Benton uplift. Associated with the second-phase folds are numerous south-directed thrusts of small displacement and a pervasive northerly dipping cleavage. Along the northern margin of the Benton uplift, during the second-phase folding, flattening and south-directed shear formed sheath folds several kilometers in length (Sturgess and Viele, 1986).

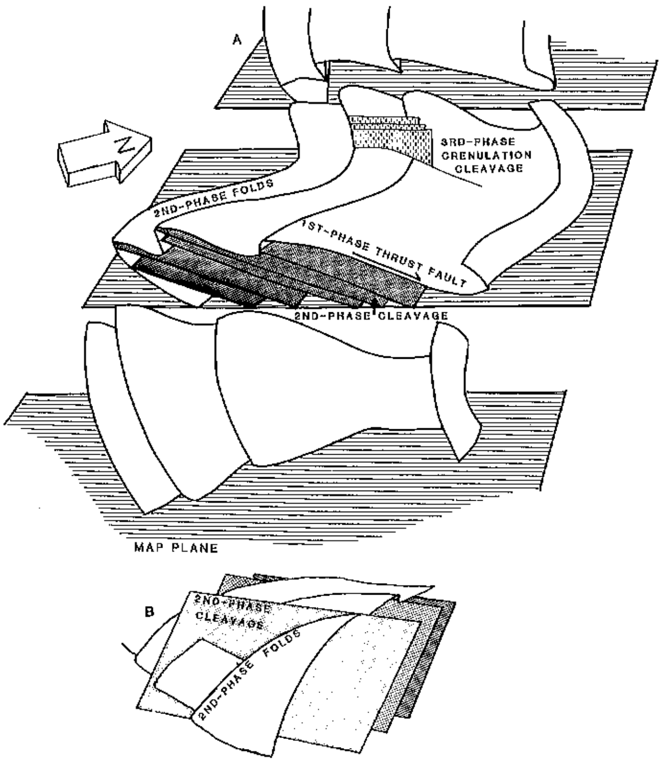


Figure 15. Diagrammatic sketch of phases of deformation in the eastern part of the Benton uplift. First-phase thrust sheets are folded in second-phase folds that porpoise in and out of the map plane forming sheath folds several kilometers in length. The second-phase folds are flattened across the second-phase cleavage planes, which transect the second-phase axial planes at low angle. Based on mapping by Sturgess (1986) and unpublished mapping by Viele.

In the third phase of deformation within the Benton uplift, a crenulation cleavage, seen primarily in thin section, was formed that transects the limbs and cleavage of the second-phase folds (Fig. 15). The third-phase structures also include widely spaced, northeast-trending folds with gently dipping limbs. Second-phase planar elements, bedding, cleavage, and axial surfaces, were broadly arched by the third-phase folds. Culminations within the Benton uplift, marked by outcrops of Collier Formation (Fig. 13), lie along the crests of the third-phase folds (Nielsen and others, this volume).

Three phases of deformation have also been recognized in the Broken Bow uplift, but different researchers have presented different structural histories. Feenstra and Wickham (1975) did not recognize nappes or thrust sheets in the Silurian and Devonian strata of the Broken Bow uplift but thought these strata might have been thrust northward over underlying Cambrian and Ordovician slates. Open and symmetric folds were formed during the first phase of folding. A second phase of deformation produced tightening of the first-phase folds, and southward overturning of some anticlines produced north-dipping axial planes and superposition of a slaty cleavage on the first-phase folds. The third and final stage is represented by gentle folding across a northeast axis and normal faulting that accompanied uplift.

In the view of Nielsen (Nielsen and others, this volume), first-phase folds of the pre-orogenic deep-water strata were south verging and associated with the formation of a penetrative cleavage. A set of second-phase folds is essentially coaxial and south verging and folds the first-phase cleavage. A fan of later(?) faults cuts and rotates the earlier folds and the cleavage. Third-phase folds are open and associated with a crenulation cleavage. They trend northeastward, accenting the eastward and westward plunges of the older folds. The third-phase folds also account for the broad bends in the traces of thrust faults around the northwestern and northern flanks of the Broken Bow uplift (Plate 8).

It is difficult to correlate the different phases of deformation within the Benton and Broken Bow uplifts with the north-directed thrust faulting in the frontal imbricate and central belts of the Ouachita Mountains (Fig. 9; Plate 8). In fact, the phases within the uplifts probably represent a continuum of deformation that partially coincides with the thrust faulting. Yet the thrust faults outside the uplifts are only broadly folded, so these faults must mostly post-date the polyphase faulting and folding within the uplifts. Possibly, thrust faults, such as the Y City and Ti Valley (Fig. 9; Plate 8), rooted deep beneath the already-deformed Benton and Broken Bow uplifts and carried them north and northwestward to their present position above the southern margin of the North American craton (Underwood and Viele, 1985).

A possible fourth phase of deformation involved the rocks of the Ouachita succession and the underlying continental crust of North America as well. Adularia from the Ordovician Womble Shale and quartz veins of the Benton uplift gave a K-Ar date of $262 \pm 10$ Ma, consistent with a Permian age (Shelton and others, 1986). This hydrothermal event, which occurred along much of the Ouachita orogenic belt, may have been associated with essentially vertical arching of the Benton and Broken Bow uplifts. Arbenz (1984), however, views this last phase of deformation as being associated with blind thrust faulting and imbrication of the North American basement, and to this deformation he attributes the southward-overturned folds and metamorphism of the Benton and Broken Bow uplifts. As structural details at this depth are not available from either deep wells or geophysical surveys, it is not possible to comment on the attitude of faults, if any, in the North American basement.

The tight polyphase folding of the Benton and Broken Bow uplifts stands in sharp contrast to the broad synclinal folding of the central province of the Ouachita Mountains (Fig. 9; Plate 8) (Miser, 1929, 1954). This difference may be more apparent than real, for the great synclines of Carboniferous strata in the central province may once have extended over the uplifts of pre-orogenic strata later to be eroded away. Tight folds in the formations of the Benton uplift, including the lower part of the Stanley Shale, plunge westward beneath the broad Lynn Mountain and Boktukola synclines; tight folds of the same formations in the Potato Hills plunge southwestward beneath the broad Tuskahoma syncline (Plate 8). The broad folding of the synclines, probably above a detachment in the Stanley Shale, differs strikingly from the tight folding in the lower Stanley and older formations. (This difference may be seen easily on the 1:250,000 McAlester, Oklahoma, mosaic of radar imagery published in 1984 by the U.S. Geological Survey.) If broad folds of Carboniferous strata once extended over the Benton uplift, a regional two-deck structure, as Miser (1929) observed, would be characteristic of the Ouachita Mountains: an upper deck of broad folds cut by northward-directed thrust faults overlies a lower deck of tight folds generally overturned in the opposite direction toward the south.

### *Subsurface Ouachita Orogenic Belt, Texas*

From the Broken Bow uplift to the Marathon region, the Ouachita orogenic belt lies in the subsurface of southern Oklahoma and Texas and is known through drilling and seismic reflection profiling (Nicholas and Waddell, this volume). Three cross sections (Plate 11, cross sections F-F′, G-G′, H-H′) portray the structure. On sections F-F′ and G-G′, the structural style is similar to that of the Arkoma basin. The deepest structures are Atokan syndepositional high-angle faults that dropped the basement and the cover of shelf strata down toward the Ouachita tectonic front. Overlying the high-angle faults are nearly flat-lying thrust faults that transported Ouachita rocks from east to west. In the westernmost of the thrust sheets, several typical Ouachita formations, such as the Arkansas Novaculite, are recognized in the drill samples.

On cross sections F-F′ and G-G′ (Plate 11), the Luling front or thrust marks an abrupt change from non- or slightly metamorphosed Ouachita rocks to metamorphosed rocks forming the "interior metamorphic belt." As shown on cross section F-F′, the Luling thrust sheet comprises, from bottom to top: slices of granitic basement overlain successively by a thin sandstone; a thick unit of marble; and the contorted beds of the interior metamorphic belt, which lies with thrust contact on the marbles. To the west, the Luling thrust ramps upward and cuts off this sequence. The culmination of granitic rock and the cover of sandstone and marble form the Waco uplift (Fig. 7; Plate 11, cross section F-F′), which in the view of Nicholas and Rozendal (1975) and Nicholas and Waddell (this volume) is made up of a core of North American basement and cover of lower to middle Paleozoic strata. The Waco uplift was elevated by motion on the deeper Luling thrust after the interior metamorphic belt and the Ouachita thrust sheets were emplaced.

Although cross sections of the subcrop belt of eastern Texas are generally similar to cross sections of the Ouachita Mountains, they differ in some ways. They are alike in showing thrust sheets of Ouachita rocks above a down-faulted North American basement, but in the subsurface of eastern Texas, the westernmost thrust faults, forming the Ouachita tectonic front, carry pre-orogenic Ouachita strata in the hanging walls. In Oklahoma, only at Black Knob Ridge (Plate 8) do the thrust sheets of the Ouachita Mountains bring pre-orogenic rocks to the tectonic front. Frontal thrust sheets of Carboniferous strata, like those of the Arkoma basin and the frontal imbricate belt, are apparently absent from the subsurface of eastern Texas. The equivalent of the Luling thrust is not known in the Ouachita Mountains; no visible tectonic break divides the pre-orogenic rocks of the Benton and Broken Bow uplifts into distinct metamorphic provinces.

The quality of seismic reflections just west of the Waco uplift is poor, and well control is unavailable, but it appears that a deep wedge of non-metamorphosed Ouachita(?) rocks separates the Waco uplift and the interior metamorphic belt from the down-faulted North American shelf. Conspicuously absent from published seismic profiles (Rozendal and Erskine, 1971) and cross sections of eastern Texas is any hint of the large basin of stratified reflectors lying on the Gulf side of the Benton uplift (Plate 11, cross section C-C′) (Lillie and others, 1983).

Cross section H-H′ (Plate 11) across the Val Verde basin differs from cross sections to the north and east in that high-angle faults dropping basement rocks down toward the Ouachita front are absent, although south-dipping thrust faults imbricate the Carboniferous strata. A thrust fault marks the northeastern edge of the Devils River uplift, but a published seismic reflection profile shows an unfaulted homocline of Cambrian through Pennsylvanian strata on the north flank of the uplift (Nicholas, 1983).

On the Devils River uplift, several wells drilled to basement have penetrated a stratigraphic section ranging from Atokan to Late Cambrian. The rocks have undergone low-grade metamorphism, but the section is stratigraphically and compositionally comparable to the foreland section to the north and east (Nicholas, 1983). No penetrations of pre-Atokan rocks of the subsurface Ouachita frontal belt are known to the north of the Devils River uplift or upon it, but on the southern flank of the uplift, a thin slice of frontal-belt rocks may be sandwiched between the interior metamorphic belt and foreland carbonate strata, which rest on basement (Flawn and others, 1961). Along much of the southern and southwestern flank of the Devils River uplift, rocks of the interior metamorphic belt are faulted against the cover of metamorphosed foreland carbonate strata.

The basement rocks of the Devils River uplift, as revealed by wells, consist of Cambrian(?) to Precambrian, metamorphosed sedimentary and igneous rocks. Cross section H-H′ (Plate 11) shows a carapace of metasedimentary rocks interstratified with some metavolcanic rocks over a dense epidote-tremolite schist, both enveloping a deeper core of meta-igneous and metavolcanic rocks. The epidote-tremolite schist forms a prominent seismic reflector. Flawn and others (1961) noted the similarity of the metavolcanic rocks of the Devils River uplift to the Precambrian metarhyolite in the Van Horn area of western Texas and commented on the dissimilarity of the Devils River rocks to the metamorphic rocks of the Ouachita belt.

The patterns on Bouguer gravity maps of the Val Verde basin and the Devils River uplift differ from those of other parts of the subsurface Ouachita belt and from those of the Arkoma basin (Plate 10) (Keller and others, this volume). Unlike the Arkoma, Fort Worth, and San Antonio basins (Keller and others, this volume, Fig. 9), the Val Verde basin does not coincide with a gravity minimum; instead, it lies on a gradient of about 50 mgal, extending from negative values on the north to positive values on the south. The Devils River uplift lies on the trend of the interior zone gravity maximum extending from the Benton uplift to south of the Marathon region (Plate 10) (Keller and others, this volume, Fig. 9), but unlike the Waco, Benton, and Broken Bow uplifts, it shows a strongly positive, local Bouguer anomaly superimposed on the regional trend.

Is the Devils River uplift fundamentally autochthonous or allochthonous? Nicholas (1983) noted that the stratigraphic continuity of the Paleozoic strata on the uplift with the Paleozoic cover of the shelf precluded any major translation, but he did not rule out movements of a few tens of kilometers. Viele, noting the cover of Paleozoic shelf strata on the Devils River uplift, the core of Precambrian metavolcanic rock, and the apparent absence of pre-Atokan Ouachita rocks, regards the Devils River uplift as a foreland tectonic element. Thomas views it as similar to the external basement massifs of the Appalachians.

The thrust faults within the Val Verde basin are akin to the thrust faults of the Arkoma basin. They imbricate deep-water synorogenic sandstones and shales that were deposited in foreland basins in front of but not within the Ouachita orogenic belt.

### *The Marathon Region*

Cross sections through the Marathon region by Muehlberger and Tauvers (Plate 11, cross sections I-I′, J-J′) represent a more "typical" style of Ouachita deformation. In the absence of subsurface data, the cross sections have not been extended below the major décollement, and the presence or absence of high-angle faults breaking the basement has not been determined. In the northern part of the Marathon uplift, the Dugout Creek thrust forms the basal décollement beneath two northeast-trending anticlinoria, the Marathon and the Dagger Flat, although wells indicate the presence of a deeper blind thrust. The Dugout Creek thrust fault cuts across the limbs of the anticlinoria and post-dates folding (King, 1975). As in the Ouachita Mountains, a two-decked structure is evident: tight folds in the pre-orogenic strata underlie broad folds in the synorogenic Carboniferous strata. The Dagger Flat and Marathon anticlinoria contain folded thrust sheets similar in map and cross sectional views to those of the Potato Hills of Oklahoma, which are smaller in scale. One cannot help but observe that if the Dagger Flat and Marathon anticlinoria were backfolded toward the south, the structure would resemble that of the Benton and Broken Bow uplifts. Cleavage and metamorphic recrystallization are lacking from the rocks of the Marathon region except within a small inlier of Paleozoic strata at Persimmon Gap, which lies to the south of the Marathon region (Fig. 7).

Little is known about the structure of the southern interior parts of the Marathon region. In the southeastern part, three large east- to northeast-plunging synclines, containing the Pennsylvanian Haymond Formation in the troughs, dominate the surface structures (Fig. 16; Plate 8). Thrust faults overrun the southern

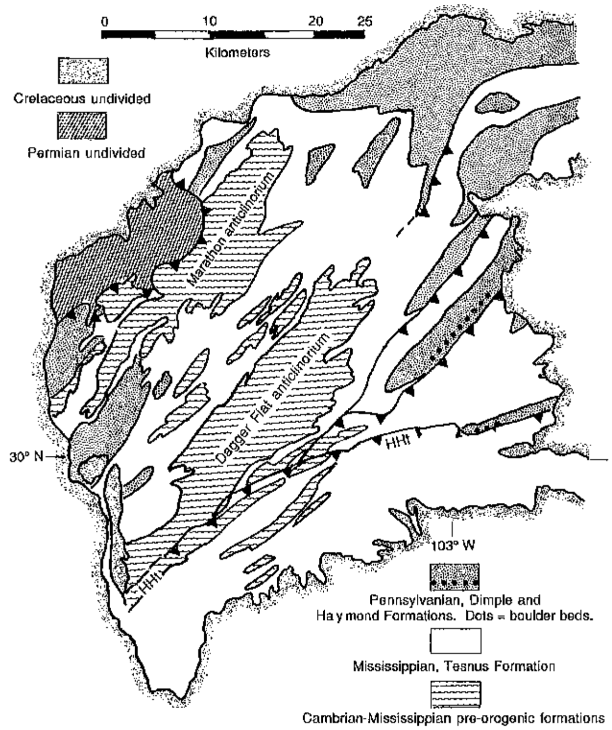


Figure 16. Index map of the Marathon region showing features mentioned in text. Abbreviation code: HHt = Hells Half Acre thrust. Further explanation in text.

flanks of each of these synclines, bringing the Mississippian Tesnus Formation over the Haymond. Toward the southwest, rising along a lateral ramp (Muehlberger and Tauvers, this volume), the thrusts bring sheets of early to middle Paleozoic pre-orogenic strata to the surface (Plate 8, inset map of Marathon region).

The southern flanks of the two southeasternmost synclines contain significant outcrops of bouldery beds in the Haymond Formation. One such bed lies near the eastern end of the Hells Half Acre thrust fault (Fig. 16) (Palmer and others, 1984), where the Tesnus Formation crops out in the hanging wall of the thrust immediately above a dip slope in the Haymond Formation in the footwall. Forming part of the dip slope, a bed of conglomerate, stratigraphically high in the Haymond, contains clasts of mostly granule-sized (A. R. Palmer, written communication, 1988) chert in a sandstone-shale matrix, but also present are rounded boulder-size clasts of limestone. These contain several North American taxa of Middle Cambrian trilobites and brachiopods. Thus, the boulders are older than any known strata in the Ouachita orogenic belt.

About 10 km to the north, in a lower, more northwesterly thrust sheet, stratigraphically bounded beds in the lower part of the Haymond Formation contain a variety of boulders, including some of metamorphic rocks that yield middle Paleozoic whole-rock and mica Rb-Sr ages (Denison and others, 1969). These beds also lie just below a thrust fault. Probably, the boulder beds in both synclines were olistostromal and derived from the uplifted interior of the Ouachita orogen, but similar boulder beds have not been reported to our knowledge from the unfaulted northern flanks of the Haymond-filled synclines.

Still farther to the south, across the Rio Grande in Mexico, rocks of the interior metamorphic belt crop out in the Sierra del Carmen (Fig. 7) (Flawn and others, 1961). Thus, in the Marathon region, as in the subcrop of east Texas, a broad belt of non-metamorphosed Ouachita strata lies between the interior metamorphic belt and the North American foreland rocks. In the Marathon region, no evidence is known of a large basin lying seaward of the uplifted region as on the south of the Benton uplift of Arkansas.

It is noteworthy that, unlike the Benton and Broken Bow uplifts, the Marathon region is associated with a large negative Bouguer gravity anomaly (Handschy and others, 1987). The interior zone gravity maximum swings a tighter arc from the Devils River uplift to the region of the Sierra del Carmen (Keller and others, this volume), and the Coahuila block of northern Mexico lies on the inside of this arc (Fig. 7) (Keller and others, this volume, Fig. 7).

## METAMORPHISM AND ISOTOPIC DATING

Outcrops within the Benton and Broken Bow uplifts and in the surrounding Carboniferous strata exhibit very low-grade metamorphism, generally within the zeolite facies but locally extending into the lower greenschist facies. Within the slates, the predominant phyllosilicate is illite, much of it having a sharpness ratio of less than 5 (Weaver, 1960, 1961; Guthrie and others, 1986), although in some areas the slates contain patches of chlorite. Quartzose sandstones, viewed in thin section, are made up of sutured and fractured grains cemented by large amounts of secondary silica. In many samples, original rounded outlines of the sedimentary grains are readily visible. Sericite and chlorite occur in pore spaces. Micrites, generally containing abundant silt-sized grains, are not recrystallized (Keller and others, 1985). The Arkansas Novaculite consists of nearly amorphous chert in the western part of the Benton uplift and of microcrystalline quartz in the eastern part (Fig. 17). Quartz veins are ubiquitous in the pre-orogenic rocks, and in areas of intense deformation are also present in the Carboniferous sandstones and shales. Quartz-bearing graywackes from the Stanley Shale contain analcite and thompsonite, indicating a peak temperature of metamorphism of less than 300° C (Jackson, 1968). Throughout the Benton uplift, conodonts from Ordovician micrites are generally black (R. Ethington, personal communication, 1987) and indicate somewhat higher temperatures, approaching 300° C, than would be inferred from the mineralogy of the rocks.

The grade of metamorphism varies throughout the Benton and Broken Bow uplifts (Fig. 17). Internal strain and metamorphic recrystallization, associated with the second-phase folding, were greatest in the northern part of the Benton uplift and in the adjacent Maumelle zone. In the eastern end of the Benton uplift, heat from Mesozoic intrusive rocks (Fig. 17) has probably accented recrystallization; the gravity signatures of the Mesozoic intrusives extend well beyond the areas of outcrop and indicate that igneous rocks may underlie much of the eastern Benton uplift. Internal strain and metamorphism both decrease toward the south and west, for strata in the southwestern part of the Benton uplift and in the Trap and Cossatot Mountains (Fig. 13; Plate 8) are neither cleaved nor greatly recrystallized.

Within the exposed part of the Broken Bow uplift, the inten-

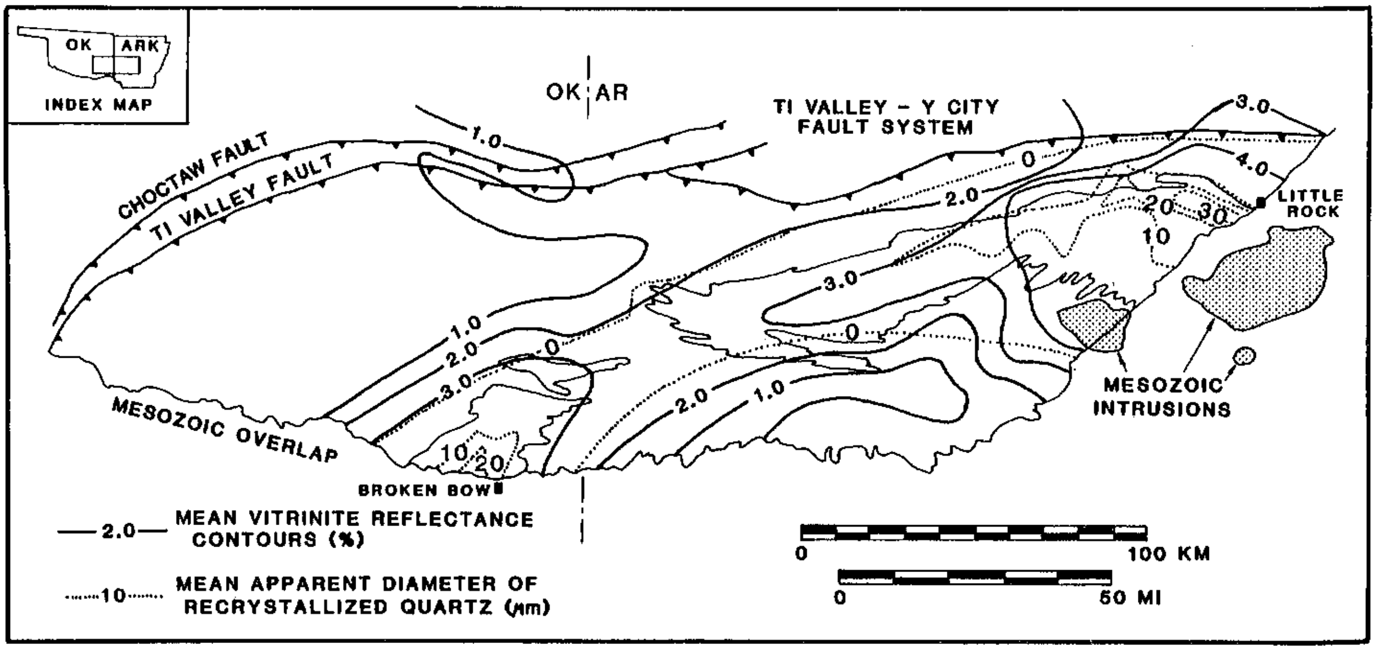


Figure 17. Map showing thermal maturity of exposed strata of Ouachita Mountains. Based on Houseknecht and Matthews (1985) and Keller and others (1985).

sity of deformation and metamorphic recrystallization increases generally southward toward the core of oldest rocks (Miser, 1959). The southward increase in deformation and recrystallization within the Broken Bow uplift is opposite to the gradient of deformation and recrystallization within the Benton uplift. Relative to the subsurface interior metamorphic belt of eastern Texas, metamorphic grade in the Benton and Broken Bow uplifts is lower and much more variable. Exposed rocks of the Marathon uplift are not metamorphosed.

In the core of the Broken Bow uplift, Viersen and Cochran's 25-1 Weyerhaeuser well penetrated 3,048 m of metamorphic rocks all lying within the greenschist facies (Goldstein, 1975; Denison and others, 1977). The rocks range from graphitic marbles with sandy and slaty phyllite intervals near the surface to quartzo-feldspathic schists with graphitic phyllite intervals at depth. These rocks were correlated with the rocks of the subsurface interior metamorphic belt of eastern Texas (Denison and others, 1977).

Denison and others (1977) reported on 38 isotopic ages determined from a variety of igneous and metamorphic rocks sampled along the Ouachita orogenic belt from Arkansas to Texas and also reviewed the results of earlier isotopic age dating. The oldest isotopic dates from the Ouachitas are Devonian and come from the igneous and metamorphic boulders in the Pennsylvanian Haymond Formation of the Marathon region. A second set of Devonian dates comes from the Ordovician Mazarn Shale of the Broken Bow uplift. They are K-Ar ages determined from sericite-muscovite concentrates, and fall into two sets: one ranging from 313 to 324 Ma and the other from 358 to 378 Ma. No difference was noted in the degree of metamorphic recrystallization of the two sets, and for this reason the possibility was discounted that the older dates were inherited from detrital micas. Yet it seems unlikely that the Mazarn Shale was being metamorphosed while the Arkansas Novaculite was being deposited above it, a difficulty that Denison and others (1977) recognized. Structural fabrics in the Broken Bow uplift are penetrative from Upper Cambrian to Mississippian rocks, and no structures have been recognized in the uplift separating the Mazarn from the younger formations. The Devonian isotopic dates from the Mazarn Shale in the Broken Bow uplift perforce remain unexplained, although Denison and others (1977) suggest the possibility of heretofore unmapped structural boundaries.

Throughout the Ouachita Mountains of Arkansas and Oklahoma, field relations indicate a post-Atokan time of deformation affecting the entire stratigraphic section. Nevertheless, the Mississippian strata of the foreland Black Warrior basin and the entire length of the Ouachita orogenic belt contain metamorphic clasts derived from the south. Certainly, orogenic highlands were present by Mississippian time.

K-Ar ages of scattered samples from the interior metamorphic belt of eastern Texas range from 248 to 320 Ma. On the Waco uplift (Fig. 7; Plate 11, cross section F-F′), the Shell No. 1 Barrett well penetrated more than 2,987 m of slate and phyllite and more than 1,828 m of marble before entering a granitoid basement. K-Ar ages from the pre-Mesozoic units range from 257 to 380 Ma and show no relationship to either depth or rock composition (Nicholas and Rozendal, 1975). Nor do isotopic dates from the Devils River uplift show any correlation to depth or rock composition (Denison and others, 1977; Nicholas, 1983).

These dates indicate that cooling and crystallization of potassium-bearing minerals took place primarily during a time interval ranging from Pennsylvanian to mid-Permian. Along the entire length of the Ouachitas, orogenic events reset the Precambrian and early Paleozoic rocks to Carboniferous isotopic ages. Unfortunately, the orogenic heating so thoroughly homogenized isotopic ages from different structural provinces that they are of little use in correlating or differentiating one province from another.

## TECTONIC SYNTHESIS

The earliest compilers (Powers, 1928; Miser, 1929) of Ouachita geology recognized the allochthonous nature of the orogen and the emplacement of it by northward-moving thrust faults. Most later compilations have followed the same general scheme but have differed greatly in assigning a tectonic mechanism to explain the orogenesis. All compilers have confronted the same problems: how to account for the abrupt change in style and rate of deposition between the lower and upper parts of the Ouachita stratigraphic section; how to account for the presence of exotic clasts of platform rocks in strata of the Ouachita facies; how to account for the reversal of vergence of folds between the higher and lower parts of the Ouachita stratigraphic section; and how to relate the structures of the Ouachita folded belt to those of the foreland.

Not all regional compilations, of course, have addressed all the problems; some have focused instead on specific topics. Powers (1928), for example, recognized the need to bring foreland rocks close to the trend of the outcrops in the Johns Valley Formation and, therefore, postulated a deep, southward-directed thrust fault that transported foreland rocks, as high as the Atoka Formation, far to the south. Soon after, as shown on the cross sections of Hendricks and others (1971), the northward-moving Choctaw and Ti Valley thrust faults sliced off parts of the Powers thrust sheet and transported them back to the north, thereby incorporating blocks of foreland carbonate rocks in the Ouachita thrust sheets. These ideas were largely rejected by later workers, and the Johns Valley boulders were viewed as clasts in debris flows moving downslope from an island chain "Bengalia" (Kramer, 1933) or fault scarp (Shideler, 1970; Gordon and Stone, 1977). The strongest advocates of gravity sliding (Haley and Stone, 1985; Stone and Haley, 1985; Morris, 1971, 1982, this volume) attributed not only the boulders of the Johns Valley Formation to downslope movements but also assigned to gravity sliding the structures in the Maumelle zone, as Viele (1973) initially did.

A second area of discussion has concerned the structural arrangement of the Ouachita Mountains as a whole. Miser

(1929) contrasted the tight folding and minor thrust faulting of the Ouachita Mountains of Arkansas to the long, parallel thrust faults and broad folds of the Ouachita Mountains of Oklahoma. In addition, his cross section of the Ouachita Mountains of Oklahoma contrasted the tight folding of the Arkansas Novaculite and older formations to the open folding of the Carboniferous formations. In his text, he commented explicitly on the difficulty of folding the great mass of sandstone of the Jackfork and Atoka Formations into isoclinal folds. Implicit in his discussion and cross sections is a portrayal of the Ouachitas as a folded belt with an upper and a lower deck. Haley and Stone (1981, 1982), on the other hand, divided the Ouachita Mountains of Arkansas into six east-west–trending belts separated by major fault zones. Each belt is said to have a characteristic set of structures. Their cross section of the Benton uplift shows it as a belt of imbricate thrust sheets floored by southward-dipping listric thrust faults rising off a basal décollement (Haley and Stone, 1984). Their portrayal of the Ouachita Mountains of Arkansas as an intensely faulted terrane is opposed to that of Miser (1929), who commented on the lack of faulting and emphasized the tight folding.

Recently, Zimmerman (1986) has reviewed ideas concerning the southward-verging folds within the Benton and Broken Bow uplifts. Some authors (Feenstra and Wickham, 1975) attributed them to northward underthrusting of the basement; Arbenz (1984) suggested that the southward-verging structures were formed by thrusting the pre-orogenic strata northward beneath a rigid cap of synorogenic Carboniferous sandstones, thereby generating a southward-rotating simple shear. Still other authors (Cambray and Welland, 1985) attributed the southward-verging folds to south-moving thrust faults. None of these scenarios is satisfactory, for none relates the style of deformation to the larger tectonic setting and history of the Ouachita orogenic belt.

On the craton in the foreland of the Ouachita orogenic belt, faults of late Paleozoic age outline many structures that offset Precambrian basement. Some authors (Kluth and Coney, 1981; Thomas, 1983; Viele, 1983, 1986) have related these structures to the Ouachita orogeny, but others (Denison, this volume) have seen no distinct simple relationship. In these latter models, the foreland structures are attributed to vertical movements, or to northeast-southwest compression, but fundamentally, they are viewed as cratonal structures loosely linked, if at all, to the late Paleozoic folded belts flanking the North American craton.

Many of the problems of Ouachita tectonics are resolved when they are viewed wholeheartedly in the context of plate tectonics (Fig. 18). This entails not just a vague endorsement of the mechanism of plate tectonics in which subduction zones lie somewhere off to the south, but a specific comparison of parts of the Ouachita orogenic belt to parts of modern subduction zones. The Ouachitas are not a foreland fold-thrust belt analogous to the Valley and Ridge province of the Appalachians or to the Sevier belt of the western United States. They are better understood in terms of the kinematics of oceanic trenches and especially of collision zones where a continental plate is entering the trench. In brief, our tectonic model, which is an enlargement of earlier

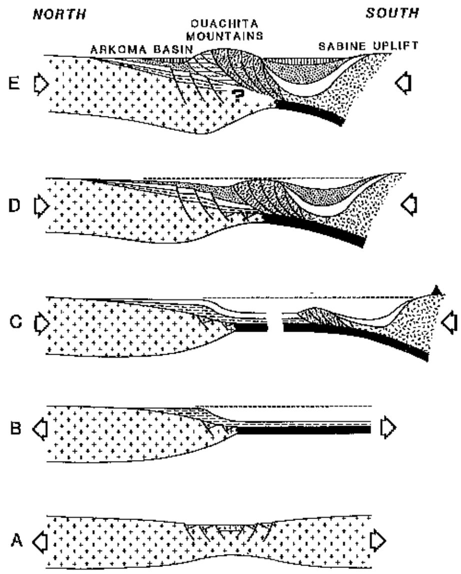


Figure 18. Diagrammatic cross sections showing the tectonic development of the southern margin of North America and the Ouachita orogenic belt, much shortened from north to south. (A) late Precambrian–earliest Paleozoic, (B) Late Cambrian–earliest Mississippian—the Ouachita ocean at its widest and deepest, (C) Early Mississippian–earliest Atokan—growth of an accretionary wedge offshore accompanied by flysch deposition in a fore-arc basin on the south and a trench on the north, (D) early-middle Atokan—thrusting of the accretionary wedge onto the southern margin of North America, which breaks down by faulting. Continued deposition of flysch in fore-arc basin and trough environments, (E) late Atokan–Desmoinesian—continued thrusting of the Ouachita orogen onto North America; possible faulting beneath the Benton and Broken Bow uplifts. Pattern key: crosses = continental crust (undifferentiated in A, North American in B–E); straw hachures = exotic continental or transitional crust beneath a microcontinent or arc complex; black = early Paleozoic oceanic crust; heavy dots = basal Paleozoic strata in half grabens; horizontal dashes = upper Cambrian to basal Mississippian strata including shelf and off-shelf facies; white = basal Mississippian to lower Atokan strata; sand stippling = lower to middle Atokan strata; vertical lines = upper Atokan to Desmoinesian strata; mottled = Ouachita accretionary prism composed of pre-orogenic, off-shelf strata and Carboniferous synorogenic strata; wavy lines = Ouachita frontal imbricate belt and Maumelle zone; black triangle = magmatic arc volcanoes. Figure modified from Houseknecht (1986).

models (Briggs and Roeder, 1975; Graham, and others, 1975; Wickham and others, 1976; Thomas, 1976, 1977; Viele, 1977, 1979a, 1979b; Nelson and others, 1982; Lillie and others, 1983; Lillie, 1985), involves the closing of an ocean by subduction of the North American plate beneath a growing accretionary prism and an attendant island arc that lay to the south. Strata deposited on transitional and oceanic crust of the southern part of the North American plate were transferred to the upper plate by thrust faults propagating northward from the toe of the accretionary prism. During the later stages of the orogeny, the accretionary prism was thrust over the southern margin of North America, loading the continental crust and forming the structures of the foreland.

The first hint of contractive tectonic events associated with the closing of the Ouachita ocean comes from isotopic ages. The isotopic dates from the metamorphic clasts of the Haymond boulder beds of the Marathon region (Fig. 16) indicate Devonian deformation and metamorphism far to the south or southeast of the Marathon region. The deformation and metamorphism must have been seaward of the site of deposition of the Middle Cambrian boulders of limestone in the Haymond Formation, because these boulders are neither deformed nor metamorphosed. Indeed, the Devonian deformation and metamorphism must have been seaward of the site of deposition of the Devonian Caballos Novaculite of the Marathon region (McBride, this volume), because these rocks too are unmetamorphosed. And finally, the deformation and metamorphism pre-dated thrusting of the Ouachita facies onto North America, because the Tesnus-Haymond succession was deposited on the Caballos, which was deposited off the North American continental shelf. As Devonian deformation and metamorphism are not known from the North American side of the Ouachita ocean, the boulders of metamorphic rock must have come from an uplift within the Ouachita ocean or from a terrane on the far shore of the Ouachita ocean.

Two lines of evidence from Mississippian strata record the change from an opening to a closing ocean. First, the framework grains of the Mississippian and Pennsylvanian sandstones prograding northeastward into the foreland Black Warrior basin indicate a provenance of low-grade metamorphic, sedimentary, and volcanic rocks. The source, which was to the southwest, was the accretionary prism and island arc of the upper plate of a closed subduction zone (Mack and others, 1983). The Ouachita remnant ocean closed first at the southeastern end (Thomas, this volume, Fig. 3). The second line of evidence for ocean closing is found in the striking change in the rate and nature of deposition along the trend of the Ouachitas from Arkansas to the Marathon region of Texas. Rapidly deposited turbidites in a thick succession containing stratiform deposits of barite at the base, abruptly succeeded slowly deposited, radiolarian-bearing cherts containing abundant manganese. This sequence is precisely reminiscent of oceanic trenches where thick, trench-fill clastic sediments are deposited on abyssal-plain sediments as they enter the trench (Lash, 1985; Piper and others, 1985). The contact of the Arkansas Novaculite with the Stanley Shale records this change of environment, as does the contact of the Caballos Novaculite with shale of the Tesnus Formation.

Within the Ouachita Mountains of Arkansas and Oklahoma, the oldest structures are thrust faults largely confined within the novaculite carapace of the Benton and Broken Bow uplifts (Fig. 13). In the Benton uplift, the thrusts stacked the pre-orogenic strata into fold and thrust nappes after deposition of the lower part of the Mississippian Stanley Shale, but before the second-phase southward folding. The highest nappe, which probably came from farthest south, sliced off pods of serpentinite and gabbro from the ocean floor, incorporating them in the lower part of the nappe as it moved northward. In addition, during this period of thrusting, folds were nucleated that were continuously tightened during later phases of deformation. These deformations took place as the pre-orogenic sediments were stripped from the subducting ocean floor and added to the base of a growing accretionary prism. These deformations pre-dated thrusting onto the North American shelf and took place above oceanic or attenuated continental crust (Fig. 18).

Yet even as the pre-orogenic strata were stacked into a growing accretionary prism, the upper part of the Stanley Shale and younger Pennsylvanian sandstones and shales were deposited at ever-increasing rates across the top and flanks of the prism. The two-decked structure of the Ouachita Mountains primarily reflects the deposition of Carboniferous strata across the quiescent top and southern flank of the prism. Perhaps the abrupt thinning of the Stanley Shale from south to north in the Ouachita Mountains (Figs. 8, 10) (Cline, 1960) records the prograding of fore-arc basin strata over the prism. Perhaps a time-transgressive unconformity, which should mark the floor of a fore-arc basin, lies hidden in the poorly exposed shales of the Stanley. Thrust faulting along a décollement within the Stanley Shale at about this stratigraphic level (Arbenz, this volume) obscures the relationship between the upper and lower parts of the formation. By Morrowan time, the growth of the accretionary prism probably formed a submarine ridge, an ancestral Benton uplift, imperfectly dividing the closing ocean into two primary basins of deposition, a trench on the north, and a fore-arc basin on the south. As the flood of sediment was great and as the growth of the accretionary prism was irregular, some sediment flows crossed from the fore-arc basin to the trench.

Carboniferous strata deposited on the north of the accretionary prism were subducted and intensely deformed at the toe. Large-scale underplating of Carboniferous strata, similar to the underplating of the modern Makran accretionary prism (Platt and others, 1985), rotated the older pre-orogenic strata within the prism upward and southward. Possibly, the strongly negative gravity anomalies along the trend of the Ouachita orogenic belt (Keller and others, this volume) mark zones of large-scale underplating of Carboniferous sandstone and shale to the base of the Ouachita accretionary prism. The accretionary fabrics of the Maumelle zone and the boulder-bearing, fault-bounded slices of the Johns Valley Formation probably were formed at this time. Thrust faults now slicing into the subducting continental margin of North America transferred shelf-carbonate blocks into the upper plate, where they were mixed with clasts from the siliceous pre-orogenic Ouachita facies and transported northward in thrust sheets of Pennsylvanian shales. Probably, by this time, the oceanic and transitional crust of the North American plate had been subducted, and continental crust lay below the Ouachita accretionary prism and the Arkoma foreland basin. A cross section of the Ouachita collision zone at this time would have resembled the present-day Timor trough (Karig and others, 1987).

South of the Ouachita accretionary prism, the fore-arc basin was not completely free of compressional stresses. The fore-arc basin was not tectonically collapsed, but some shortening was evident as widespread detachment zones formed, separating the tightening folds of the lower deck from the broad folding of the upper deck. Out-of-sequence thrust faults of relatively small displacement formed on the steep southern limbs of the broad synclines.

As the ocean closed, the flood of clastic detritus from the uplifted orogen to the south and east buried the physiographic partitions of the collision zone. The sediments spread across the fore-arc basin, the buried accretionary prism and trough, and onto the North American shelf. During the middle Atokan, the Ouachita accretionary prism of Arkansas and Oklahoma was thrust onto the southern margin of North America (Fig. 18) (Van der Gracht, 1931a; Houseknecht, 1986). Among the structures associated with this deformation were the Ti Valley and Y City thrust faults, which transported the Ouachita rocks over the peripheral margin of the Arkoma foreland basin. Probably, the duplexes of the frontal imbricate zone of Oklahoma were formed at this time. The high-angle faults of the Arkoma basin record the breakdown of the shelf area under the load of the obducting Ouachitas and the downbending of the North American plate as it approached the subduction zone (Houseknecht, 1986). Thrust faults continued to form at the toe of the wedge and advance northward, loading more inboard parts of the Arkoma basin.

In the Marathon region, in a like manner, the Dugout Creek thrust carried the already deformed, pre-orogenic rocks over the southern edge of the foredeep (King, 1975).

Remaining elusive is the tectonic setting of the second-phase south-directed folding and the accompanying formation of cleavage in the Benton and Broken Bow uplifts (Fig. 15). Clearly, the second-phase folding post-dates formation of the first-phase thrust sheets within the Benton uplift, because the faults are folded. Clearly, the second-phase folding post-dates deposition of the Pennsylvanian Jackfork Formation, for along the northern side of the Benton uplift in the Maumelle zone, the Jackfork is tightly folded and intensely cleaved. The Jackfork and Atoka Formations did not act as a rigid lid in the frontal imbricate belt of Oklahoma, for here too these formations are tightly folded and thrust faulted. Possibly, the south-vergent structures formed during subduction by off-scraping of strata from the lower plate onto the upper plate, but major south-directed thrust faults have not been found. Alternatively, the south-verging structures may have formed when the Ouachita accretionary prism ramped upward off the ocean floor onto the down-faulted North American basement and cover of shelf strata. Movement of an allochthonous sheet over a ramp results in rotation and strain of the overriding sheet (Mandle and Crans, 1981). Perhaps movement up a ramp 10 km(?) in height, combined with rotation of the prism by underplating of sediment, accounts for the formation of the second-phase folds and cleavage.

Cross sections across the subsurface Ouachita orogen of eastern Texas are markedly dissimilar to those across the Ouachita Mountains of Arkansas. The Waco uplift (Plate 11, cross section F-F′) (Nicholas and Waddell, this volume) may represent a late uplift of overridden North American basement, or it may represent a suspect terrane accreted onto North America (Hatcher and Viele, 1982). The Waco uplift lies on the interior zone gravity maximum, but the absence of any local gravity anomaly associated with the uplift suggests it is a thin allochthonous slice (Plate 10) (Keller and others, this volume). Numerous authors (Flawn and others, 1961; Denison and others, 1977; Nicholas and Waddell, this volume) have suggested that the rocks of the interior metamorphic belt are subsurface extensions of the Benton and Broken Bow uplifts, but we persist in viewing the interior metamorphic belt as an accreted terrane that overran the pre-orogenic rocks deposited in the Ouachita ocean.

The interior zone gravity maximum provides a reference line for estimating the distance of thrusting of the Ouachita orogenic belt over the North American continental margin. If the gravity maximum marks the transition from early Paleozoic continental to oceanic crust, and if the early Paleozoic Ouachita facies was deposited at the base of the North American continental slope, then the minimum distance of thrusting must be the distance from the interior zone maximum to the northern or westernmost outcrop of Ouachita rocks. Or, in other words, it is the distance from the southeastern side of the Broken Bow uplift to Black Knob Ridge, or about 150 km (Plates 8, 9, 11). In central Arkansas, the distance from the interior zone maximum to the Y City thrust fault is about 65 km. Does this indicate oblique subduction? The distance from the interior zone maximum to the northern edge of the Marathon region is about 75 km. Obviously, the upper plate is much shortened by imbrication of thrust sheets, folding, and internal strain in the rocks. The amount of shortening of the lower plate is not known, but no geophysical evidence suggests to us imbrication of the North American basement.

The closing of the ocean and thrusting of the accretionary prism of Ouachita rocks onto North America, as early as Mississippian time along the southwestern flank of the Black Warrior basin and as late as Wolfcampian time in the Marathon region, signaled the final stages of the Ouachita orogeny. On the North American continent, the final pulse of deformation in the Early Permian may have arched the Benton and Broken Bow uplifts. Heating and dewatering of the accretionary prism resulted in a wave of hydrothermal fluids spreading from the Ouachitas toward the interior of the craton, accounting for many of the late Paleozoic, Mississippi Valley–type ore deposits (Leach and Rowan, 1986).

The Ouachita orogenic belt is made up of rocks deposited on the North American plate and transferred by thrust faulting to the upper plate. The surface trace of the thrust or suture separating the North American plate from the upper plate follows the frontal imbricate belt of the Ouachita Mountains. Down dip, the fault plane follows the arch of the Benton uplift, and from there it dips gently southward. The fundamental crustal suture separating the North American plate from the upper southern plate lies deep in the subsurface south of the Benton and Broken Bow uplifts. The late Paleozoic volcanic rocks of the Sabine uplift are on the upper plate. In the subsurface of eastern Texas, the western edge of the Ouachita frontal belt marks the bounding trace of the

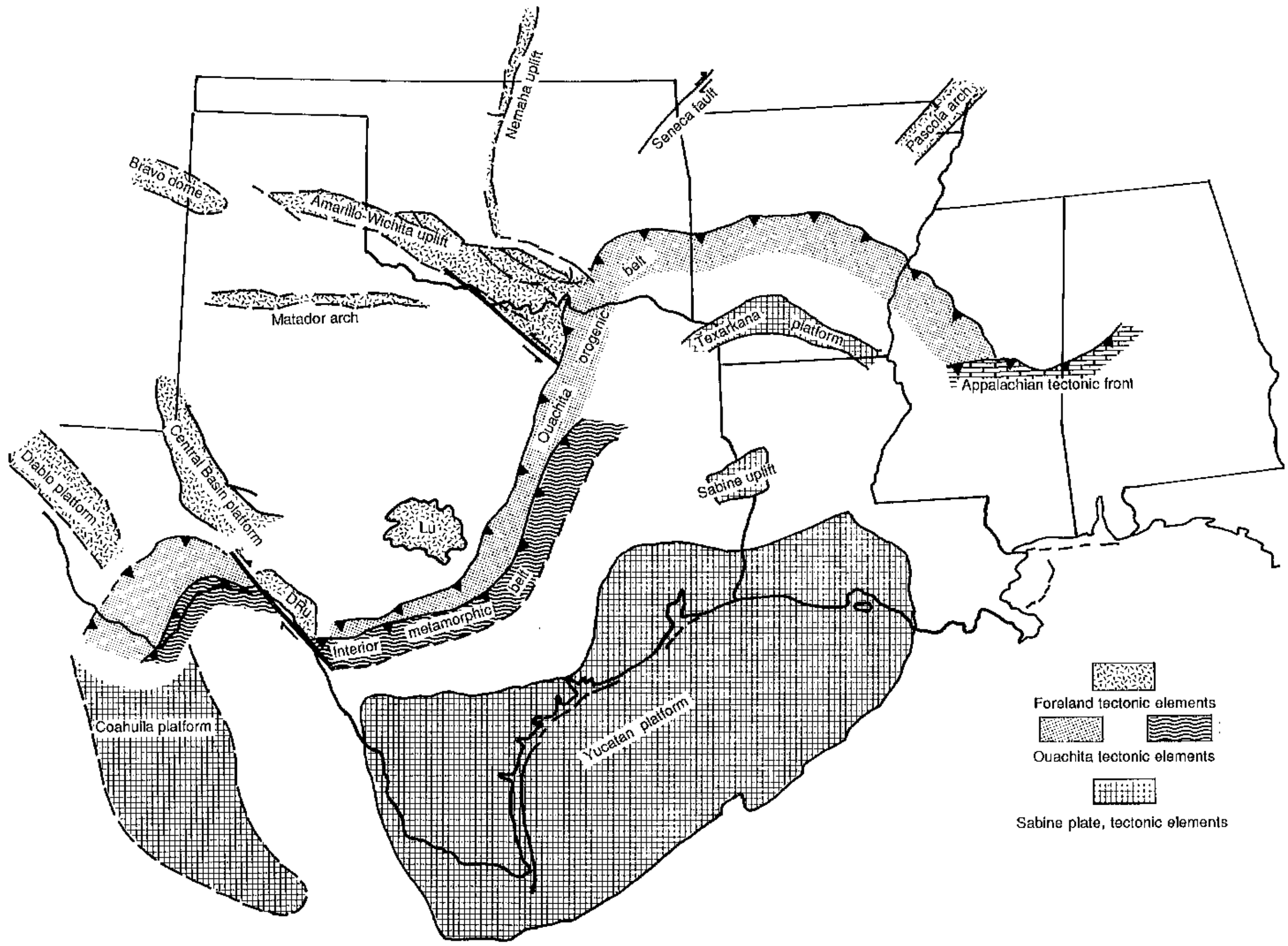


Figure 19. Tectonic elements composing the Sabine plate at the end of the Ouachita orogeny and associated foreland structures. Compiled from King (1975), Pindell (1985), and Handschy and others (1987).

Ouachita orogenic belt, as does the Dugout Creek thrust fault in the Marathon region (Plate 8, inset map of Marathon region).

The outboard or Gulfward side of the Ouachita orogenic belt was filled with a collage of tectonic elements (Pindell, 1985; Dunbar and Sawyer, 1987), designated collectively as the Sabine plate (Fig. 19). These elements, which included the Yucatan block, filled the area of the Gulf of Mexico and separated the North American and Gondwanan plates. The Sabine terranes should be considered in paleomagnetic reconstructions of Pangea, especially those debating the position of the northern margin of Gondwanaland (Van der Voo and others, 1976; Morel and Irving, 1981). By Triassic time, rifting within the Sabine plate marked the onset of the next Wilson cycle, the Mesozoic opening of the Gulf of Mexico.

## REFERENCES CITED

Adams, J. E., 1965, Stratigraphic–tectonic development of Delaware basin: American Association of Petroleum Geologists Bulletin, v. 49, p. 2140–2148.

Ammerman, M. L., and Keller, R. G., 1979, Delineation of Rome trough in eastern Kentucky by gravity and deep drilling data: American Association of Petroleum Geologists Bulletin, v. 63, p. 341–353.

Arbenz, J. K., 1968, Structural geology of the Potato Hills, Oklahoma, *in* Cline, L. M., ed., Guidebook to the geology of the western Arkoma basin and Ouachita Mountains, Oklahoma: Oklahoma City Geological Society Guidebook, p. 109–121.

——, 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. 76–84.

——, 1989, The Ouachita system, *in* Bally, A. W., and Palmer, A. R., eds., The Geology of North America; An Overview: Boulder, Colorado, Geological

Society of America, The Geology of North America, v. A, p. 371–396.

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

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., A guidebook to the geology of the central and southern Ouachita Mountains, Arkansas: Arkansas Geological Commission Guidebook 84-2, p. 123.

Braile, L. W., Keller, G. R., Hinze, W. J., and Lidiak, E. G., 1982, An ancient rift complex and its relation to contemporary seismicity in the New Madrid seismic zone: Tectonics, v. 1, p. 225–237.

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

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

Burke, K., and Dewey, J. F., 1973, Plume-generated triple junctions: Key indicators in applying plate tectonics to old rocks: Journal of Geology, v. 81, p. 406–433.

Byrne, T., 1984, Early deformation in mélange terranes of the Ghost Rocks Formation, Kodiak Islands, Alaska, *in* Raymond, L. A., ed., Mélanges: Their nature, origin, and significance: Geological Society of America Special Paper 198, p. 21–51.

Cambray, F. W., and Welland, M.J.P., 1985, Southward verging structures and co-axial refolding in the Benton uplift, Arkansas: A result of southerly directed thrusting: Geological Society of America Abstracts with Programs, v. 17, p. 538.

Carson, B., von Huene, R., and Arthur, M. A., 1982, Small-scale deformation structures and physical properties related to convergence in Japan trench slope sediments: Tectonics, v. 1, p. 277–302.

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

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

Cline, L. M., and Shelburne, O. B., 1959, Late Mississippian–Early Pennsylvanian stratigraphy of the Ouachita Mountains, Oklahoma, *in* Cline, L. M., Hilseweck, W. J., and Feray, D. E., eds., The geology of the Ouachita Mountains, a symposium: Dallas and Ardmore Geological Societies, p. 175–208.

Cowan, D. S., 1982, Deformation of partly dewatered and consolidated Franciscan sediments near Piedras Blancas Point, California, *in* Legget, J. K., ed., Trench–forearc geology: Sedimentation and tectonics on modern and ancient active plate margins: Geological Society of London Special Publication 10, p. 439–457.

Crosby, E. J., and Mapel, W. J., 1975, Central and west Texas, *in* McKee, E. D., and Crosby, E. J., coordinators, Paleotectonic investigations of the Pennsylvanian System in the United States: U.S. Geological Survey Professional Paper 853, p. 197–232.

Decatur, S. H., and Rosenfeld, J. H., 1982, Mississippian radioactive marker beds and their possible relationship to the Ouachita orogeny: Geological Society of American Abstracts with Programs, v. 14, p. 109.

Denison, R. E., Kenny, G. S., Burke, W. H., Jr., and Hetherington, E. A., Jr., 1969, Isotopic ages from igneous and metamorphic boulders from the Haymond Formation (Pennsylvanian), Marathon Basin, Texas and their significance: Geological Society of America Bulletin, v. 80, p. 245–256.

Denison, R. E., Burke, W. H., Otto, J. B., and Hetherington, E. A., 1977, Age of igneous and metamorphic activity affecting the Ouachita folded belt, *in* Stone, C. G., ed., Symposium on the geology of the Ouachita Mountains, Volume 1: Arkansas Geological Commission, p. 25–40.

Dickinson, W. R., and Suczek, C., 1979, Plate tectonics and sandstone compositions: American Association of Petroleum Geologists Bulletin, v. 63, p. 2164–2182.

Dunbar, J. A., and Sawyer, D. S., 1987, Implications of continental crust extension for plate reconstruction: An example from the Gulf of Mexico: Tectonics, v. 6, p. 739–755.

Ervin, C. P., and McGinnis, L. D., 1975, Reelfoot rift: Reactivated precursor to the Mississippi Embayment: Geological Society of America Bulletin, v. 86, p. 1287–1295.

Feenstra, R., and Wickham, J., 1975, Evolution of folds around Broken Bow uplift, Ouachita Mountains, southeastern Oklahoma: American Association of Petroleum Geologists Bulletin, v. 59, p. 974–985.

Flawn, P. T., Goldstein, A., Jr., King, P. B., and Weaver, C. E., 1961, The Ouachita system: Austin, University of Texas, Bureau of Economic Geology Publication Number 6120, 401 p.

Flawn, P. T., and Maxwell, R. A., 1958, Metamorphic rocks in the Sierra del Carmen, Coahuila, Mexico: American Association of Petroleum Geologists Bulletin, v. 42, p. 2245-2249.

Gilbert, C. M., 1982, Geologic setting of the eastern Wichita Mountains with a brief discussion of unresolved problems, *in* Gilbert, C. M., and Donovan, R. N., eds., Geology of the eastern Wichita Mountains, southwestern Oklahoma: Oklahoma Geological Survey Guidebook 21, p. 1–30.

Ginzburg, A., Mooney, W. D., Walter, A. W., Lutter, W. J., and Healy, J. H., 1983, Deep structure of northern Mississippi Embayment: American Association of Petroleum Geologists Bulletin, v. 67, p. 2031–2046.

Goldstein, A., Jr., 1975, Geologic interpretation of Viersen and Cochran's 25-1 Weyerhaeuser well, McCurtain County, Oklahoma: Oklahoma Geology Notes, v. 35, p. 167–181.

Goldstein, A., Jr., and Hendricks, T. A., 1962, Late Mississippian and Pennsylvanian sediments of Ouachita facies, Oklahoma, Texas, and Arkansas, *in* Branson, C. C., ed., Pennsylvanian System in the United States: American Association of Petroleum Geologists, p. 385–430.

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, Volume 1: Arkansas Geological Commission, p. 70–91.

Graham, S. A., Dickinson, W. R., and Ingersoll, R. V., 1975, Himalayan–Bengal model for flysch dispersal in the Appalachian–Ouachita system: Geological Society of America Bulletin, v. 86, p. 273–286.

Graham, S. A., Ingersoll, R. V., and Dickinson, W. R., 1976, Common provenance for lithic grains in Carboniferous sandstones from Ouachita Mountains and Black Warrior basin: Journal of Sedimentary Petrology, v. 46, p. 620–632.

Guthrie, J. M., Houseknecht, D. W., and Johns, W. D., 1986, Relationships among vitrinite reflectance, illite crystallinity, and organic geochemistry in Carboniferous strata, Ouachita Mountains, Oklahoma and Arkansas: American Association of Petroleum Geologists Bulletin, v. 70, p. 26–33.

Haley, B. R., and Stone, C. G., 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, Volume 1: Arkansas Geological Commission, p. 107–114.

—— , 1981, Structural framework of the Ouachita Mountains, Arkansas: Geological Society of America Abstracts with Programs, v. 13, p. 239.

—— , 1982, Structural framework of the Ouachita Mountains, Arkansas: Geological Society of America Abstracts with Programs, v. 14, p. 113.

—— , 1984, Structural cross section of the eastern Ouachita Mountains, *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. 31.

—— , 1985, Structural history of the Ouachita Mountains, Arkansas: Part 1—The big picture: Geological Society of America Abstracts with Programs, v. 17, p. 160–161.

Haley, B. R., Glick, E. E., Bush, W. V., Clardy, B. F., Stone, C. G., Woodward, M. B., and Zachry, D. L., 1976, Geologic map of Arkansas: Arkansas

Geological Commission and U.S. Geological Survey, scale 1:500,000.

Ham, W. E., 1959, Correlation of pre-Stanley strata in the Arbuckle-Ouachita Mountain region, *in* Cline, L. A., Hilseweck, W. J., and Feray, D. E., eds., The geology of the Ouachita Mountains, a symposium: Dallas and Ardmore Geological Societies, p. 71–86.

Ham, W. E., Denison, R. E., and Merrit, C. A., 1964, Basement rocks and structural evolution of southern Oklahoma: Oklahoma Geological Survey Guidebook 5, 58 p.

Handschy, J. W., Keller, G. R., and Smith, K. J., 1987, The Ouachita system in northern Mexico: Tectonics, v. 6, p. 323–330.

Hanor, J. S., and Baria, L. R., 1977, Controls on distribution of barite deposits in Arkansas, *in* Stone, C. G., ed., Symposium on the geology of the Ouachita Mountains, Volume 2: Arkansas Geological Commission, p. 42–49.

Hart, W., Stitt, J. H., Hohensee, S. R., and Ethington, R. L., 1987, Geological implications of Late Cambrian trilobites from the Collier Shale, Jessieville area, Arkansas: Geology, v. 15, p. 447–450.

Hatcher, R. D., Jr., and Viele, G. W., 1982, The Appalachian/Ouachita orogens: United States and Mexico, *in* Palmer, A. R., ed., Perspectives in regional geological synthesis, planning for the geology of North America: Boulder, Colorado, Geological Society of America, DNAG Special Publication 1, p. 67–75.

Hendricks, T. A., 1971, Provenance of Johns Valley boulders in late Paleozoic Ouachita facies, southeastern Oklahoma and southwestern Arkansas: Discussion: American Association of Petroleum Geologists Bulletin, v. 55, p. 1365–1366.

Hendricks, T. A., Knechtel, M. M., and Bridge, J., 1937, Geology of Black Knob Ridge, Oklahoma: American Association of Petroleum Geologists Bulletin, v. 21, p. 1–26.

Hendricks, T. A., Gardner, L. S., Knechtel, M. M., and Averitt, P., 1971, Geology of the western part of the Ouachita Mountains of Oklahoma: U.S. Geological Survey Oil and Gas Investigation Preliminary Map 66, 3 sheets.

Hildenbrand, T. G., 1985, Rift structure of the northern Mississippi Embayment from the analysis of gravity and magnetic data: Journal of Geophysical Research, v. 90, p. 12607–12622.

Hobday, D. K., 1974, Beach and barrier-island facies in the upper Carboniferous of northern Alabama: Geological Society of America Special Paper 148, p. 209–224.

Hoffman, P., Dewey, J. and Burke, K., 1974, Aulacogens and their genetic relation to geosynclines with a Proterozoic example from Great Slave Lake, Canada, *in* Dott, R. H., Jr., and Shaver, R. H., eds., Modern and ancient geosynclinal sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication 19, p. 38–55.

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

Houseknecht, D. W., 1986, Evolution from passive margin to foreland basin: The Atoka Formation of the Arkoma basin, south-central U.S.A.: International Association of Sedimentologists Special Publication 8, p. 327–345.

——, 1987, The Atoka Formation of the Arkoma basin: Tectonics, sedimentology, thermal maturity, sandstone petrology: Tulsa Geological Society Short Course Notes, 72 p.

Houseknecht, D. W., and Matthews, S. M., 1985, Thermal maturity of Carboniferous strata, Ouachita Mountains: American Association of Petroleum Geologists Bulletin, v. 69, p. 335–345.

Houseknecht, D. W., and Weaverling, P. H., 1983, Early Paleozoic sedimentation in Reelfoot rift [abs.]: American Association of Petroleum Geologists Bulletin, v. 67, p. 1456.

Houseknecht, D. W., Zaengle, J. F., Steyaert, D. J., Matteo, A. P., Jr., and Kuhn, M. A., 1983, Facies and depositional environments of the Desmoinesian Hartshorne Sandstone, Arkoma basin, *in* Houseknecht, D. W., ed., Tectonic-sedimentary evolution of the Arkoma basin: Society of Economic Mineralogists and Paleontologists, Midcontinent Section, v. 1, p. 53–82.

Jackson, K. C., 1968, Temperature indicators of Arkansas Ouachita deformation [abs.]: Geological Society of America Special Paper 115, p. 371.

Karig, D. E., 1983, Deformation in the forearc: Implication for mountain belts, *in* Hsü, K. J., ed., Mountain building processes: New York, Academic Press, p. 59–72.

Karig, D. E., Barber, A. J., Charlton, T. R., Klemperer, S., and Hussong, D. M., 1987, Nature and distribution of deformation across the Banda Arc–Australian collision zone at Timor: Geological Society of America Bulletin, v. 98, p. 18–32.

Keller, G. R., Lance, J. O., Jr., Sarno, R. D., Graham, R. G., Aiken, C.L.V., Schmidt, J. S., and Hinze, W. J., 1981, A regional geophysical study of southern Trans-Pecos Texas: Society of Economic Paleontologists and Mineralogists, Permian Basin Section, Publication 81-20, p. 103–129.

Keller, G. R., Braile, L. W., McMechan, G. A., Thomas, W. A., Harder, S. H., Chang, W.-F., and Jardine, W. G., 1989, Paleozoic continent-ocean transition in the Ouachita Mountains imaged from PASSCAL wide-angle seismic reflection–refraction data: Geology, v. 17, p. 119–122.

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.

King, P. B., 1937, Geology of the Marathon region: U.S. Geological Survey Professional Paper 187, 148 p.

——, 1964, Interpretation of the Garden Springs area, Texas, by the "down structure" method of structural analysis: U.S. Geological Survey Professional Paper 501-B, p. B1–B8.

——, 1975, The Ouachita and Appalachian orogenic belts, *in* Nairn, A.E.M. and Stehli, F. G., eds., The ocean basins and margins, v. 3: New York and London, Plenum Press, p. 201–241.

Kluth, C. F., and Coney, P. J., 1981, Plate tectonics of the ancestral Rocky Mountains: Geology, v. 9, p. 10–15.

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

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

Kruger, J. M., and Keller, G. R., 1986, Interpretation of regional gravity anomalies in the Ouachita Mountains area and adjacent Gulf Coastal Plain: American Association of Petroleum Geologists Bulletin, v. 70, p. 667–689.

Lash, G. G., 1985, Recognition of trench fill in orogenic flysch sequences: Geology, v. 13, p. 867–870.

Leach, D. L., and Rowan, E. L., 1986, Genetic link between Ouachita folded belt tectonism and the Mississippi Valley-type lead-zinc deposits of the Ozarks: Geology, v. 14, p. 931–935.

Lillie, R. J., 1984, Tectonic implications of subthrust structures revealed by seismic profiling of Appalachian–Ouachita orogenic belt: Tectonics, v. 3, p. 619–646.

——, 1985, Tectonically buried continent/ocean boundary, Ouachita Mountains, Arkansas: Geology, v. 13, p. 18–21.

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

Lovick, G. P., Mazzine, C. G., and Kotila, D. A., 1982, Atokan clastics—Depositional environments in a foreland basin: Oil and Gas Journal, v. 80, p. 181–199.

Lowe, D., 1975, Regional controls on silica sedimentation in the Ouachita system: Geological Society of America Bulletin, v. 86, p. 1123–1127.

——, 1985, Ouachita trough: Part of a Cambrian failed rift system: Geology, v. 13, p. 790–793.

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

Mandl, G., and Crans, W., 1981, Gravitational gliding in deltas, *in* McClay, K. R., and Price, N. J., eds., Thrust and nappe tectonics: Geological Society of London Special Publication 9, p. 41–54.

McDonald, K. W., 1986, Structural style, thermal maturity, and sedimentary facies of the frontal Ouachita thrust belt, LeFlore County, eastern Oklahoma [M.S. thesis]: Columbia, University of Missouri, 211 p.

Meyerhoff, A. A., 1973, Late Paleozoic of western Gulf Coastal Plain, *in* A study of Paleozoic rocks in Arbuckle and western Ouachita Mountains of southern Oklahoma: Shreveport Geological Society Guidebook, p. 31–37.

Miser, H. D., 1917, Manganese deposits of the Caddo Gap and DeQueen quadrangles, Arkansas: U.S. Geological Survey Bulletin 660-C, p. 59–122.

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

——, 1934, Carboniferous rocks of Ouachita Mountains: American Association of Petroleum Geologists Bulletin, v. 18, p. 971–1009.

——, 1954, Geologic map of Oklahoma: U.S. Geological Survey in cooperation with Oklahoma Geological Survey, scale 1:500,000.

——, 1959, Structure and vein quartz of the Ouachita Mountains of Oklahoma and Arkansas, *in* Cline, L. M., Hilseweck, W. J., and Feray, D. E., eds., The geology of the Ouachita Mountains, a symposium: Dallas and Ardmore Geological Societies, p. 30–43.

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

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

Moore, R. C., 1934, The origin and age of the boulder-bearing Johns Valley Shale in the Ouachita Mountains of Arkansas and Oklahoma: American Journal of Science, v. 27, p. 432–453.

Morel, P., and Irving, E., 1981, Paleomagnetism and the evolution of Pangea: Journal of Geophysical Research, v. 86, p. 1858–1872.

Morris, E. M., and Stone, C. G., 1986, A preliminary report on the metagabbros of the Ouachita core, *in* Stone, C. G., and Haley, B. R., eds., Sedimentary and igneous rocks of the Ouachita Mountains of Arkansas, a guidebook with contributed papers, Part 2: Arkansas Geological Commission Guidebook 86-3, p. 87–90.

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

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

——, 1982, Depositional setting of Carboniferous flysch: Geological Society of America Abstracts with Programs, v. 14, p. 571.

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

Nelson, K. D., Lillie, R. J., de Voogd, B., Brewer, J. A., Oliver, J. E., Kaufman, S., Brown, L. D., and Viele, G. W., 1982, COCORP seismic reflection profiling in the Ouachita Mountains of western Arkansas: Geometry and geologic interpretation: Tectonics, v. 1, p. 413–430.

Nicholas, R. L., 1983, Devils River uplift, *in* Bolden, G. P., ed., Structure and stratigraphy of the Val Verde basin–Devils River uplift, Texas: West Texas Geological Society Fall 1983 Field Trip Guidebook, West Texas Geological Society Publication 83-77, p. 1207–1246.

Nicholas, R. L., and Rozendal, R. A., 1975, Subsurface positive elements within the Ouachita foldbelt in Texas and their relationship to the Paleozoic cratonic margin: American Association of Petroleum Geologists Bulletin, v. 59, p. 193–216.

Nicholas, R. L., and Waddell, D. E., 1982, New Paleozoic subsurface data from the north-central Gulf Coast: Geological Society of America Abstracts with Programs, v. 14, p. 576.

Niem, A. R., 1977, Mississippian pyroclastic flow and ash-fall deposits in the deep-marine Ouachita flysch basin, Oklahoma and Arkansas: Geological Society of America Bulletin, v. 88, p. 49–61.

Palmer, A. R., DeMis, W. D., Muehlberger, W. R., and Robinson, R. A., 1984, Geological implications of Middle Cambrian boulders from the Haymond Formation (Pennsylvanian) in the Marathon basin, west Texas: Geology, v. 12, p. 91–94.

Pindell, J. L., 1985, Alleghenian reconstruction and subsequent evolution of the Gulf of Mexico, Bahamas, and proto-Caribbean: Tectonics, v. 4, p. 1–39.

Piper, D.J.W., von Huene, R., and Duncan, J. R., 1973, Late Quaternary sedimentation in the active eastern Aleutian trench: Geology, v. 1, p. 19–22.

Platt, J. P., Leggett, J. K., Young, J., Raza, H., and Alam, S., 1985, Large-scale sediment underplating in the Makran accretionary prism, southwest Pakistan: Geology, v. 13, p. 507–511.

Poole, L. A., 1985, Sedimentology, structural style, and thermal maturity of the Lynn Mountain Formation, frontal Ouachitas, Latimer and LeFlore Counties, Oklahoma [M.S. thesis]: Columbia, University of Missouri, 189 p.

Powers, S., 1928, Age of folding of the Oklahoma Mountains: Geological Society of America Bulletin, v. 39, p. 1031–1072.

Purdue, A. H., and Miser, H. D., 1923, Hot Springs folio, Arkansas: Geological Atlas of the United States, U.S. Geological Survey Folio 215, 12 p.

Raymond, L. A., 1984, Classification of mélanges, *in* Raymond, L. A., ed., Mélanges: Their nature, origin, and significance: Geological Society of America Special Paper 198, p. 7–20.

Repetski, J. E., and Ethington, R. L., 1977, Conodonts from graptolite facies in the Ouachita Mountains, Arkansas and Oklahoma, *in* Stone, C. G., ed., Symposium on the geology of the Ouachita Mountains, Volume 1: Arkansas Geological Commission, p. 92–106.

Ross, C. A., 1986, Paleozoic evolution of southern margin of Permian basin: Geological Society of America Bulletin, v. 97, p. 536–554.

Rozendal, R. A., and Erskine, W. S., 1971, Deep test in the Ouachita structural belt of central Texas: American Association of Petroleum Geologists Bulletin, v. 55, p. 2008–2017.

Sample, J. C., and Moore, J. C., 1987, Structural style and kinematics of an underplated slate belt, Kodiak and adjacent islands, Alaska: Geological Society of American Bulletin, v. 99, p. 7–20.

Satterfield, J., 1982, Sedimentology of the Blaylock Formation in the Bismark and De Roche quadrangles, Arkansas: Geological Society of America Abstracts with Programs, v. 14, p. 135.

Sclater, J. G., and Francheteau, J., 1970, The implications of terrestrial heat flow observation on current tectonic and geochemical models of the crust and upper mantle of the earth: Geophysical Journal of the Royal Astronomical Society, v. 20, p. 509–542.

Scotese, C. R., 1984, An introduction to this volume: Paleozoic paleomagnetism and the assembly of Pangea, *in* Van der Voo, R., Scotese, C. R., and Bonhommet, N., eds., Plate reconstruction from Paleozoic paleomagnetism: American Geophysical Union Geodynamics Series, v. 12, p. 1–10.

Shelton, K. C., Reader, J. M., Ross, L. M., Viele, G. W., and Seidemann, D. E., 1986, Ba-rich adularia from the Ouachita Mountains, Arkansas; implications for a post-collisional hydrothermal system: American Mineralogist, v. 71, p. 916–923.

Sheridan, R. E., and Gradstein, F. M., 1982, Early history of the Atlantic Ocean and gas hydrates on the Blake Outer Ridge: Results of the Deep Sea Drilling Project Leg 76: Geological Society of America Bulletin, v. 93, p. 876–885.

Shideler, G. L., 1970, Provenance of Johns Valley boulders in late Paleozoic Ouachita facies, southeastern Oklahoma and southwestern Arkansas: American Association of Petroleum Geologists Bulletin, v. 54, p. 789–806.

——, 1971, Provenance of Johns Valley boulders in late Paleozoic Ouachita facies, southeastern Oklahoma and southwestern Arkansas: Reply: American Association of Petroleum Geologists Bulletin, v. 55, p. 1366–1368.

Sloss, L. L., 1963, Sequences in the cratonic interior of North America: Geological Society of America Bulletin, v. 74, p. 93–114.

——, 1982, The midcontinent project, *in* Palmer, A. R., Perspectives in regional geological synthesis, planning for the geology of North America: Boulder, Colorado, Geological Society of America, DNAG Special Publication 1, p. 27–39.

Soderberg, R. K., and Keller, G. R., 1981, Geophysical evidence for deep basin in western Kentucky: American Association of Petroleum Geologists Bulletin, v. 65, p. 226–234.

Sterling, P. J., and Stone, C. G., 1961, Nickel occurrences in soapstone deposits, Saline County, Arkansas: Economic Geology, v. 56, p. 100–110.

Stone, C. G., and Haley, B. R., 1985, Structural history of the Ouachita Mountains, Arkansas: Part 2—The eastern Ouachita Mountains: Geological Society of America Abstracts with Programs, v. 17, p. 193.

Stone, C. G., and Sterling, P. J., 1962, New lithologic marker horizons in Ordovician rocks, eastern Ouachitas of Arkansas: American Association of Petroleum Geologists Bulletin, v. 46, p. 387–390.

Stone, C. G., LeBlanc, R. J., Sr., and Haley, B. R., 1979, Stop descriptions; fourth day, morning, *in* Sutherland, P., 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. 55–59.

Sturgess, S. W., 1986, Structural geology of the Alum Fork area, Benton uplift, Arkansas [M.S. thesis]: Columbia, University of Missouri, 125 p.

Sturgess, S., and Viele, G. W., 1986, Regional-scale sheath folding, Benton uplift, Arkansas: Geological Society of America Abstracts with Programs, v. 18, p. 766.

Suppe, J., 1985, Principles of structural geology: Englewood Cliffs, New Jersey, Prentice-Hall Inc., 537 p.

Sutherland, P. K., and Manger, W. L., 1979, Comparison of Ozark shelf and Ouachita basin facies for Upper Mississippian and Lower Pennsylvanian series in eastern Oklahoma and western Arkansas, *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. 1–13.

Thomas, W. A., 1972, Regional Paleozoic stratigraphy in Mississippi between Ouachita and Appalachian Mountains: American Association of Petroleum Geologists Bulletin, v. 56, p. 81–106.

——, 1973, Southwestern Appalachian structural system beneath the Gulf Coastal Plain: American Journal of Science, v. 273A, p. 372–390.

——, 1974, Converging clastic wedges in the Mississippian of Alabama, *in* Briggs, G., ed., Carboniferous of the southeastern United States: Geological Society of America Special Paper 148, p. 187–207.

——, 1976, Evolution of Ouachita–Appalachian continental margin: Journal of Geology, v. 84, p. 323–342.

——, 1977, Evolution of Appalachian–Ouachita salients and recesses from reentrants and promontories in the continental margin: American Journal of Science, v. 277, p. 1233–1278.

——, 1979, Mississippian stratigraphy of Alabama, *in* The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Alabama and Mississippi: U.S. Geological Survey Professional Paper 1110-I, p. I1-I22.

——, 1983, Continental margins, orogenic belts, and intracratonic structures: Geology, v. 11, p. 270–272.

——, 1985, The Appalachian–Ouachita connection: Paleozoic orogenic belt at the southern margin of North America: Annual Review of Earth and Planetary Sciences, v. 13, p. 175–199.

——, 1986, Early Paleozoic southern margin of North America: Geological Society of America Abstracts with Programs, v. 18, p. 772.

——, 1988a, The Black Warrior basin, *in* Sloss, L. L., ed., Sedimentary cover—North American craton: U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. D-2, p. 471–492.

——, 1988b, The Black Warrior basin, *in* Sloss, L. L., ed., Sedimentary cover—North American craton: U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. D-2, Plate 8.

Thomson, A., and Thomasson, M. R., 1964, Sedimentology and stratigraphy of Dimple Limestone, Marathon region, Texas: Society of Economic Paleontologists and Mineralogists, Permian Basin Section, Publication 64-9, p. 22–30.

Ulrich, E. O., 1927, Fossiliferous boulders in the Ouachita "Caney" Shale and the age of the shale containing them: Oklahoma Geological Survey Bulletin 45, 48 p.

Underwood, M. B., 1984, A sedimentologic perspective on stratal disruption within sandstone-rich mélange terranes: Journal of Geology, v. 92, p. 369–385.

Underwood, M. B., and Viele, G. W., 1985, Early Pennsylvanian tectonic transition within the frontal Ouachitas of Arkansas: Geological Society of America Abstracts with Programs, v, 17, p. 195.

Van der Gracht, W.A.J.M. van Waterschoot, 1931a, Permo-Carboniferous orogeny in south-central United States: American Association of Petroleum Geologists Bulletin, v. 15, p. 991–1057.

——, 1931b, The pre-Carboniferous exotic boulders in the so-called "Caney shale" in the northwestern front of the Ouachita Mountains of Oklahoma: Journal of Geology, v. 39, p. 697–714.

Van der Voo, R., Mauk, F. J., and French, R. B., 1976, Permian–Triassic continental configurations and the origin of the Gulf of Mexico: Geology, v. 4, p. 177–180.

Vernon, R. C., 1971, Possible future petroleum potential of pre-Jurassic, western Gulf basin, *in* Cram, I. H., ed., Future petroleum provinces of the United States—Their geology and potential: American Association of Petroleum Geologists Memoir 15, v. 2, p. 954–984.

Viele, G. W., 1966, The regional structure of the Ouachita Mountains of Arkansas: A hypothesis, *in* Field conference on flysch facies and structure of the Ouachita Mountains: Kansas Geological Society Guidebook, p. 245–278.

——, 1973, Structure and tectonic history of the Ouachita Mountains, Arkansas, *in* DeJong, K., and Scholten, R., eds., Gravity and tectonics: New York, John Wiley and Sons, p. 361–377.

——, 1977, A plate tectonic model, Ouachita Mountain folded belt: Geological Society of America Abstracts with Programs, v. 9, p. 661–662.

——, 1979a, Geological map and cross section, eastern Ouachita Mountains, Arkansas: Geological Society of America Map and Chart Series MC-28F.

——, 1979b, Geological map and cross section, eastern Ouachita Mountains, Arkansas: Map summary: Geological Society of America Bulletin, v. 90, p. 1096–1099.

——, 1983, Collision effects on craton caused by Ouachita orogeny: Geological Society of America Abstracts with Programs, v. 15, p. 712.

——, 1986, The subduction of Texas: Geological Society of America Abstracts with Programs, v. 18, p. 779.

Vollmer, F. W., and Bosworth, W., 1984, Formation of mélange in a foreland basin overthrust setting: Example from the Taconic orogen, *in* Raymond, L. A., ed., Mélanges: Their nature, origin, and significance: Geological Society of America Special Paper 198, p. 53–70.

Walper, J. L., 1977, Paleozoic tectonics of the southern margin of North America: Gulf Coast Association of Geological Societies Transactions, v. 27, p. 230–241.

Walthall, B. H., 1967, Stratigraphy and structure, part of Athens Plateau, southern Ouachitas, Arkansas: American Association of Petroleum Geologists Bulletin, v. 51, p. 504–528.

Weaver, C. E., 1960, Possible uses of clay minerals in search for oil: American Association of Petroleum Geologists Bulletin, v. 44, p. 1505–1518.

——, 1961, Clay minerals of the Ouachita structural belt and adjacent foreland, *in* Flawn, P. T., Goldstein, A., Jr., King, P. B., and Weaver, C. E., The Ouachita system: Austin, University of Texas, Bureau of Economic Geology Publication 6120, p. 147–162.

Wickham, J., Roeder, D. R., and Briggs, G., 1976, Plate tectonic models for the Ouachita foldbelt: Geology, v. 4, p. 173–176.

Wilson, J. L., 1954, Ordovician stratigraphy in the Marathon folded belt, west Texas: American Association of Petroleum Geologists Bulletin, v. 38, p. 2455–2475.

Winker, C. D., and Buffler, R. T., 1988, Paleogeographic evolution of early deep-water Gulf of Mexico and margins, Jurassic to Middle Cretaceous (Comanchean): American Association of Petroleum Geologists Bulletin, v. 72, p. 318–346.

Zachry, D. L., 1983, Sedimentologic framework of the Atoka Formation, Arkoma basin, Arkansas, *in* Houseknecht, D. W., ed., Tectonic-sedimentary evolution of the Arkoma basin: Society of Economic Paleontologists and Mineralogists, Midcontinent Section, v. 1, p. 34–52.

Zartman, R. E., 1977, Geochronology of some alkalic rock provinces in eastern and central United States: Annual Review of Earth and Planetary Sciences, v. 5, p. 257–286.

Zietz, I., 1982, Composite magnetic anomaly map of the United States, Part A: Conterminous United States: U.S. Geological Survey Geophysical Investigations Map GP-954-A.

Zimmerman, J., 1986, The problem of antivergent structures in the Ouachita thrust belt, *in* Stone, C. G., and Haley, B. R., eds., Sedimentary and igneous rocks of the Ouachita Mountains of Arkansas, a guidebook with contributed papers, Part 2: Arkansas Geological Commission Guidebook 86-3, p. 79–86.

Manuscript Accepted by the Society May 8, 1989

Printed in U.S.A.