HOW DO THE STRUCTURES OF THE LATE PALEOZOIC OUACHITA THRUST BELT RELATE TO THE STRUCTURES OF THE SOUTHERN OKLAHOMA AULACOGEN

Steven John Jusczuk
University of Kentucky

2002

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ABSTRACT OF DISSERTATION

A dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy in the College of Arts and Sciences at the University of Kentucky

By

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West Sayville, New York

Director: Dr. William A. Thomas, Professor of Geology
Lexington, Kentucky
2002
ABSTRACT OF DISSERTATION

HOW DO THE STRUCTURES OF THE LATE PALEOZOIC OUACHITA THRUST BELT RELATE TO THE STRUCTURES OF THE SOUTHERN OKLAHOMA AULACOGEN

The thin-skinned structures of the late Paleozoic Ouachita thrust belt intersect the basement structures of the Southern Oklahoma aulacogen beneath the Mesozoic strata of the Gulf Coastal Plain in southeastern Oklahoma. The Ouachita thrust belt forms a large northwest-directed salient which extends primarily in the subsurface from central Mississippi northwestward to Arkansas and eastern Oklahoma, and from there, southwestward toward central Texas.

Kinematics are complicated in the center of the Ouachita salient, where the average southwesterly strike of thrust faults is nearly perpendicular to average trend of compressional basement structures in the Southern Oklahoma aulacogen (Arbuckle uplift) and Muenster arch. Furthermore, the frontal fault of the Ouachita thrust belt curves sharply eastward around the southeastern end of the Arbuckle uplift, and bends sharply to the west between the Arbuckle uplift and the Muenster arch farther south in Texas. Nine new interpreted structural cross sections show the structural complexity of the area where the Ouachita thrust belt intersects the
Arbuckle uplift and Muenster arch.

Detailed study of the structural geology of the Ouachita Mountains and Arkoma basin indicates that along-strike changes in structural style evidently are related to along-strike changes in mechanical stratigraphy (relative thicknesses of weak units, in contrast to stiff units). The middle part of the Stanley Group (Formation) evidently serves as a wavelength transition and/or volume compensation zone. Along-strike change in stratigraphic level of detachments and abrupt eastward thickening of the Atoka Formation along the Ouachita thrust front strongly affected the structural style of the Ouachita thrust belt.

Regional stratigraphy, palinspastic restorations of the footwall cutoff of the Ti Valley fault, and an abrupt change in character of seismic reflectors indicate an abrupt facies transition in the Middle Ordovician-Mississippian succession along the southeastern flank of the Arbuckle uplift and southwestward toward the deep southeastern part of the Ardmore basin. Out-of-syncline structures in the Bryan small-scale salient, distinct sub-thrust angular unconformities imaged on seismic profiles, and sediment dispersal patterns in the early Atokan-Desmoinesian strata of the northern Fort Worth basin (south of the Muenster arch) all indicate that the Tishomingo-Belton and Muenster structures were pre-thrust structural highs.

KEYWORDS: Ouachita thrust belt, Arbuckle uplift, Muenster arch, structure

Steven John Jusczuk

April 29, 2002
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This dissertation is dedicated in memory of my mother.
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Chapter One
Introduction and location

One purpose of this dissertation is to examine the interaction of the autochthonous structures of the Arbuckle uplift with the allochthonous structures of the late Paleozoic Ouachita thrust belt where they intersect beneath Mesozoic Gulf Coastal Plain strata in southeastern Oklahoma. Within a relatively small-scale study area, the frontal fault of the Ouachita thrust belt bends sharply to the southeast around the southeastern end of the Arbuckle uplift, and sharply to the northwest between the Arbuckle uplift and the Muenster arch on the south (Figure 1.1). The sharp bends in the Ouachita thrust belt at the eastern ends of the Arbuckle uplift and Muenster arch are small-scale features of a much longer southwest-striking thrust front that extends from eastern Oklahoma to central Texas, marking the western margin of a large sharp northwest-pointing salient in the Ouachita orogen (the Ouachita salient) (Figure 1.1). The eastern margin of the Ouachita salient has a southeast-strike and extends in the subsurface from Arkansas (Figure 1.1) to central Mississippi. West of the Ouachita thrust front, the predominant northwest-trending folds and northwest-striking faults of the Arbuckle uplift are parallel with other regional faults and folds within the northwest-trending cratonic faulted fold belt called the Southern Oklahoma aulacogen (Figure 1.1).

Another purpose of this dissertation is to examine the kinematics of larger scale sharply curved thrust belts, of which the Ouachita salient is an example. For this purpose, the structures of the Ouachita Mountains, and those of the Arkoma basin to the north and west, are examined in detail. The Ouachita Mountains are located in the center of the Ouachita salient, and strike of the frontal thrust belt bends nearly 90° from east to west. Detailed study of the Ouachita Mountains and adjacent Arkoma basin produces an understanding of regional kinematics and three-dimensional geometry of the allochthonous structures of the Ouachita Mountains that is necessary to constrain interpretations of structure for areas located farther south in the subsurface beneath the Gulf Coastal Plain.
Kinematic problems of sharply curved thrust belts

Palinspastic restorations of traditional cross sections constructed perpendicular to thrust-belt strike of a sharply curved thrust belt, such as across the Ouachita Mountains in the center of the Ouachita salient, produce a large apparent volume surplus (part A in Figure 1.2). Radial or divergent flow can form a curve in a wavefront (thrust belt); however, such a mechanism requires substantial along-strike extension of allochthonous rocks during overthrusting to produce a sharply curved thrust belt. Limited study of strain indicators taken from samples collected in the eastern part of the Ouachita Mountains (Weber and Zimmerman, 1986; Yang and Nielsen, 1995) shows maximum elongation directions perpendicular to strike of thrust faults in the sample localities in the Broken Bow and Benton uplifts (part A in Figure 1.2).

The problem caused by application of a radial-flow model to thrust-belt kinematics is made clear when two cross sections are constructed perpendicular to thrust-belt strike across a sharply curved thrust belt. In the example of a structural salient, the lines formed by the intersection of each cross section with the mapped surface intersect at a single point within the allochthon (part A in Figure 1.2). Separate palinspastic restorations of each individual cross section (assuming plane strain) would restore that single point into two different locations (part A in Figure 1.2). Lengths of thrust faults (wave-fronts) increase during emplacement of the allochthon which requires significant along-strike extension if translation direction is perpendicular to thrust-belt strike (part A in Figure 1.2). In the example of a structural recess, the lines formed by the intersection of each cross section with the mapped surface intersect at a single point in the foreland (part A in Figure 1.2). In this example, lengths of thrust faults decrease in length during emplacement of the allochthon which requires a great amount of along-strike compression if translation direction is perpendicular to thrust-belt strike (part A in Figure 1.2).

Problems specific to “Ouachita” geology

One problem in “Ouachita geology” is that published large-scale regional cross sections of the Ouachita orogen (for example, Arbenz, 1989a; Nicholas, 1989; Viele,
do not represent structures with enough precision to construct a balanced
kineumatic model. For purpose of scale, thin formations are combined with other
thicker formations, and where structures are complicated, faults and stratigraphic
contacts are shown schematically. Precision of regional cross sections decreases with
depth where geometry of structures is more poorly constrained. Precise locations of
hanging-wall and footwall cutoffs must be known to produce a kinematically
balanceable cross section. This problem is most extreme in cross sections of the
southwestern Ouachita Mountains where the frontal thrust fault at Black Knob Ridge
(Figure 1.1) juxtaposes Ordovician-Pennsylvanian deep-water facies strata on the east
against upper Mississippian-Pennsylvanian strata on the west with no intervening
imbricates of Ordovician-lower Mississippian strata. In this area, the footwall cutoff
for the Ordovician-Lower Mississippian strata in the hanging wall of the Ouachita
frontal fault is unknown but is traditionally placed southeast of the southeastern
margin of the Broken Bow uplift (Figure 1.1), which on the basis of two deep wells, a
gravity anomaly, and seismic data is considered the location of the Paleozoic shelf
edge (Lillie and others, 1983; Kruger and Keller, 1986; Leander and Legg, 1988;
Denison, 1989; Mickus and Keller, 1992). Because there is no direct evidence that the
Black Knob Ridge hanging wall once extended northwest of the present day surface
trace, restoring the hanging-wall cutoff of the Ouachita frontal thrust fault at Black
Knob Ridge east of Broken Bow uplift produces a kinematic problem because the
interpreted restored position of the hanging-wall cutoff is much farther southeast than
the corresponding footwall cutoff.

Another problem of “Ouachita geology” is that stratigraphy shown in regional
cross sections of the Ouachita Mountains and adjacent Arkoma basin to the north is
broadly subdivided into shallow-water “foreland facies” and deep-water “Ouachita
facies” rocks with no defined transitional facies (for example, Arbenz, 1989a,b;
Nicholas, 1989; Viele, 1989). Either the transitional facies is in the deep subsurface
across the entire Ouachita region or transitional facies strata are subjectively lumped
together with overlying or underlying “foreland” or “Ouachita” facies according to an
individual author’s interpretation, or the transition zone is very narrow. The
placement of a stratigraphic unit within a particular broadly defined facies disproportionately affects restoration position of hanging wall cut offs in cross sections where footwall cutoffs are poorly constrained or unknown.

**Approach to problem**

In regard to the large-scale purpose of this dissertation, because of the imprecision of the regional cross sections, the lack of precise facies boundaries, and the difficult kinematics of sharply curved thrust belts, a quantitative approach to the objectives of this dissertation is at this time determined inappropriate. Quantitative kinematic modeling based on palinspastically restorated regional cross sections (which were done during the early stage of research) has been exchanged for an examination of along-strike changes in structures related to along-strike changes in mechanical stratigraphy. Variations in mechanical stratigraphy, or variations in relative thicknesses of weak (ductile) layers and stiff layers within a stratigraphic succession, determine the structural style of a thrust belt (Hatcher, 1990).

In regard to the small-scale purpose of this dissertation, a set of nine new interpreted structural cross sections shows the structural complexity of the region where the Arbuckle uplift and Muenster arch intersect the Ouachita orogen (Figure 1.1). Donated paper copies of seismic reflection profiles are used as templates on which the geometry of interpreted structures is delineated. Interpretations are constrained by geologic maps of the southwestern part of the Ouachita Mountains (Hardie, 1988, 1990; Arbenz, 1989b), published cross sections (Huffman and others, 1978, 1987; Arbenz, 1989a; Hardie, 1990), published top-of-Arbuckle-Group structure contour maps (Ewing, 1991; Gatewood and Fay, 1991), published formation-top data (Bradfield, 1957a-c; Flawn and others, 1961), formation-top data supplied in data base form by the Oklahoma Geological Survey, and formation-top data for selected wells supplied by Neil Suneson and Robert O. Fay of the Oklahoma Geological Survey. Part of the pre-Mesozoic outcrop/subcrop of the Ouachita Mountains and adjacent area included with this dissertation is constrained by new cross sections.
Dissertation chapters

Chapter Two (Regional Stratigraphy) is a synthesis of regional stratigraphy and defines the Paleozoic stratigraphy of the Ouachita orogen and its foreland for the length of the Ouachita salient, extending from its intersection with the Appalachian orogen in central Mississippi, westward to the Llano uplift of Texas. For purpose of discussion, stratigraphy is subdivided into shallow-water facies and deep-water facies and further subdivided into specified time intervals. In a chronostratigraphic correlation chart created for this chapter, shallow-water-facies formations are correlated across the foreland of the Ouachita orogen. Another chronostratigraphic correlation chart shows correlations between shallow-water and deep-water facies successions. Both charts also correlate missing records of geologic time (time-gaps) represented by unconformities. Stratigraphic evidence is given for a Middle or Late Ordovician transgression towards the Arbuckle uplift and the adjacent Ardmore basin.

Chapter Three (Regional Structure) mainly defines key tectonic domains and structures of the foreland north and east of the Ouachita orogen and summarizes the structural geology of the Ouachita Mountains; however, several important basement uplifts beneath the interior of the Ouachita allochthon are also defined. Where necessary, this chapter discusses stratigraphy in the context of its structural implications. For purposes of discussion, the Ouachita Mountains are subdivided into three thrust belts: Frontal Ouachita, Central Ouachita, and Southern Ouachita. Along-strike variation in structures within each of the thrust belts is summarized.

Chapter Four (Structural Geology of the Ouachita Mountains and Arkoma basin) gives a detailed description of the along-strike changes in structures within the Frontal Ouachita, Central Ouachita, and Southern Ouachita thrust belts and the adjacent Arkoma basin to the north. The mechanical stratigraphy of the Ouachita Mountains and adjacent foreland stratigraphic succession is defined. Evidence is given for several detachment levels within the shallow-water facies and deep-water facies successions. Along-strike variation in the stratigraphic levels of detachments evidently affects structural style along the Ouachita thrust front on either side of the
frontal thrust fault. The along-strike change in structural style along the Ouachita thrust front is also evidently affected by an abrupt eastward increase in thickness of the Pennsylvanian Atoka Formation.

Chapter Five (Sub-Mesozoic structural cross sections showing intersection of the autochthonous structures of the Arbuckle uplift, Ardmore basin, and Muenster arch and the structures of the Ouachita allochthon in southeastern Oklahoma and northern Texas) examines the structures within a much smaller defined area. Chapter Five presents nine newly interpreted structural cross sections which are based upon seismic reflection profiles. For each cross section, structures are defined as within either the autochthon or allochthon, and key seismic reflectors are identified. Evidence is given for abrupt facies change, or at least a facies transition zone, for the Upper Ordovician-lower Mississippian succession near the Arbuckle uplift. In some locations in several cross sections, the basal décollement of the allochthon appears to coincide with a pre-thrust unconformity.

Chapter Six (Conclusions and Suggestions for Future Research) gives a brief summary of conclusions derived from this dissertation and research experience. Also given are some suggestions for future research.
Figure 1.1: Simplified tectonic map of the center of the Ouachita salient and adjacent foreland to the northwest. The lighter gray shading represents areas covered by Mesozoic strata of the Gulf Coastal Plain. Pre-Mississippian outcrop areas shown in the darker gray shading. Selected regional structures are shown.

Figure 1.2: Apparent volume problems evident from palinspastic restoration of cross sections constructed perpendicular to thrust-belt strike.

Parts A and B show the hanging-wall and footwall cutoffs of a sharply curved thrust fault. In Part A, restorations of cross sections (thick lines) constructed perpendicular to strike of the hanging-wall cutoff produce an apparent volume surplus at salients and an apparent volume deficit at recesses. Part B shows how apparent volume problems are removed if cross sections are restored parallel with translation direction. Restoration perpendicular to strike is parallel with translation direction in the centers of salients and recesses, but elsewhere gives only dip-slip component of translation. For the model shown in part B to be valid, the curvature of the hanging-wall cutoff (thrust fault) must be formed prior to translation.
Chapter Two
Regional Stratigraphy

Introduction

In the broadest sense, the stratigraphy of the Ouachita thrust belt and the adjacent foreland basins can be subdivided into Precambrian through Cambrian igneous basement, Cambrian through early Mississippian pre-orogenic rocks, early Mississippian through Pennsylvanian syn-orogenic clastic wedge, Pennsylvanian-Permian post-Ouachita orogenic cover, early Mesozoic rift-fill and intrusive igneous rocks of the Mississippi embayment and East Texas basin, and the Cretaceous to Recent Gulf Coastal Plain (Plate 2.1).

The following paragraphs discuss only the pre-Mesozoic stratigraphy along the Appalachian-Ouachita-Marathon orogen from Alabama to central Texas. The Paleozoic rocks generally include two significantly different facies, and where possible, lithologic and biostratigraphic correlations are made between shallow-water “shelf” facies and deep-water “off-shelf” equivalents. Descriptions of predominant lithologies are given, along with distinctive minor components. Brief descriptions of fossil faunas, age constraints, and controversies are also included in the following paragraphs. Igneous and metamorphic rocks are discussed in the section of this chapter entitled “basement rocks.” Isotopic ages are listed for selected igneous and metamorphic rocks.

Basement rocks of the craton

Southern Oklahoma aulacogen

Crystalline basement rocks crop out west of the Ouachita Mountains in the Southern Oklahoma aulacogen. The aulacogen (in the sense defined by Schatski, 1946) is a linear arrangement of fault-bounded fold belts which intersects the Ouachita orogenic belt at a high angle. The oldest documented igneous crystalline basement rocks are a Precambrian (1.1 - 1.4 Ga) succession comprised of the Tishomingo and Troy Granites and unnamed granodiorite and granitic gneiss (orthogneiss) (Ham and others, 1954, 1963). Tishomingo/Troy Granites, granodiorite, and orthogneiss are exposed on the crest of the Tishomingo-Belton anticline in the Arbuckle uplift (Plate
2.1) and likely belong to the 1.2-1.4 Ga granite-rhyolite province of the Mid-Continent region of the United States (Ham and others, 1963; and Van Schmus and Bickford, 1993).

Besides the Tishomingo/Troy granites, granodiorite, and orthogneiss that crop out in the Arbuckle uplift (Plate 2.1), younger mafic and felsic igneous rocks also crop out in the southern Oklahoma aulacogen. These igneous rocks are Early Cambrian and estimates of absolute age range from 530 Ma to 539 Ma (Hogan and Gilbert, 1998; Thomas and others, 2000, 2001). Geochemistry and mineralogy suggest syn-rift emplacement at shallow depths and magma derived from continental crust (Gilbert, 1983). The complete succession of Cambrian syn-rift igneous rocks is exposed in the Wichita uplift (Plate 2.1) of southwestern Oklahoma (Gilbert, 1982).

At the base of the syn-rift series of igneous rocks in the Wichita uplift is the Raggedy Mountain Gabbro Group. This basal set of mafic igneous rocks consists of a layered mafic intrusive body (Glen Mountains Layered Complex) cross-cut by the Roosevelt Gabbro plutons (some of which are biotite-bearing) (Gilbert, 1982). Five separate mineralogically and texturally distinct members or “zones” of the Glen Mountains Layered Complex have been mapped in the Wichita Mountains (Gilbert, 1982). Age estimates for the Glen Mountains Layered Complex vary between 1.4 Ga and 500 Ma (Powell and others, 1980); however, a more recent estimate based on Rb-Sr and Sm-Nd isotopic data places the age at 577 Ma (Lambert and Unruh, 1986). The Mount Sheridan Gabbro member of the Roosevelt Gabbros has a more precise and definitive age of 552±7 Ma (U-Pb zircon, Bowring and Hoppe, 1982).

Above the Raggedy Mountain Gabbro Group, separated by an apparent erosional unconformity, is a succession of rhyolites that are cross-cut by epizonal granites (Gilbert, 1982). These granitic rocks crop out at several locations within the Southern Oklahoma aulacogen. The Carlton Rhyolite group (530-539 Ma, Hogan and Gilbert, 1998) is found at four principal locations; three of which are in the Wichita Mountains and Slick Hills of southwestern Oklahoma, and one in the East and West Timbered Hills part of the Arbuckle Mountains (Plate 2.1) (where it is called the Colbert Rhyolite Porphyry) (Ham and others, 1954; Gilbert, 1982).
The lower part of the Carlton Rhyolite Group is intruded by the co-genetic type-A granite plutons of the Wichita Granite Group (consisting of at least 10 recognized members) (Gilbert, 1982). The three main intrusive bodies listed in inferred chronological order are the Mount Scott Granite sill, the Quanah sill, and the Saddle Mountain rhyolite. Cross-cutting relationships indicate that the Mount Scott Granite is the oldest of the epizonal granites that comprise the Wichita Granite Group (Gilbert, 1982). Both the Quanah sill and the Saddle Mountain rhyolite cross cut the Mount Scott Granite; however, age relationships between the Quanah and Saddle Mountain are unclear (Gilbert, 1982).

The Mount Scott Granite consists of two members: a microgranite and a variably granophyric, plagioclase, hornblende porphyry (Gilbert, 1982). The Quanah sill consists of three members: an arfvedsonite-bearing granite and an equivalent fine-grained aplitic phase, and a medium-grained, biotite-bearing granophyre (Gilbert, 1982). The Saddle Mountain rhyolite is hornblende-biotite bearing, and has a spherulitic-porphyritic to granophyric-porphyritic texture (Gilbert, 1982). Isotopic age estimates for both the extrusive and intrusive phases of the Carlton Rhyolite Group and the Wichita Granite Group are nearly equivalent and are Early Cambrian (539 ± 5 Ma, Thomas and others, 2001).

Precambrian to Middle Cambrian metasedimentary rocks are rare at the surface within the Southern Oklahoma aulacogen. Scattered xenoliths of metamorphic rocks and the thin, discontinuous Meers Quartzite are the only pre-Late Cambrian metasedimentary rocks which crop out within the aulacogen. All these rarities are found in the Wichita Mountains of southwest Oklahoma.

Constraints on the age of the Meers Quartzite are not precise. Older publications (such as Ham and others, 1964) suggest that rocks included in the Meers Quartzite are fragments of Precambrian sedimentary rocks found south of the Wichita Mountains in the subsurface. More recent studies suggest that the Meers Quartzite is a remnant of a thin rock layer deposited above the Raggedy Mountain Gabbro Group prior to emplacement of the Cambrian rhyolite (Gilbert, 1982).

Numerous wells located along the southern flank of the Wichita Mountains
(Wichita uplift) penetrate “generally low-grade metamorphosed graywackes, shales, siltstones, sandstones, arkoses, and bedded cherts” (Sides and Miller, 1982). These rocks, referred to as the Tillman Metasedimentary Group, are known only in the subsurface, and are interpreted to be Precambrian in age (Ham and others, 1964). Seismic reflection profiles across the Wichita uplift show 7 to 10 km of layered reflectors (interpreted as the Tillman Metasedimentary Group) south of the uplift, but the layered reflectors are lacking north of the uplift (Brewer and others, 1981, Brewer, 1982). The northwest-trending southern margin of the Wichita uplift may mark the position of a Precambrian fault (Brewer and others, 1983) where equivalents of the Tillman Metasedimentary Group to the north of the Wichitas were uplifted and eroded prior to Cambrian rifting.

**Llano uplift**

The Llano uplift is a circular foreland basement uplift in central Texas (Barnes, 1988) (Plate 2.1). Basement rocks within the Llano uplift are cross cut by numerous faults. The predominant trend of basement faults in the area is northeast-southwest (Barnes, 1988). Precambrian rocks in the center of the Llano uplift are in a low-relief topographical basin surrounded by flat-topped, discontinuous limestone hills (Barnes, 1988). The Precambrian rocks can be combined into three broad subdivisions in ascending chronological order: the metasedimentary Llano Group (Supergroup); metamorphosed gabbroic and granitic igneous rocks; and post-metamorphic igneous rocks.

The oldest formation of the Llano Supergroup is the Valley Springs Gneiss, dated at 1120±25 Ma (Rb-Sr and K-Ar) (Barnes, 1988). The Valley Springs is microcline-quartz gneiss with subsidiary biotite and hornblende. Stratigraphically above the Valley Springs Gneiss are the Lost Creek Gneiss Formation and Packsaddle Schist Group. The Lost Creek Gneiss has been inferred to be both metasedimentary (Ragland, 1960) and metamorphosed rhyolite (Garrison and others, 1979).

The Packsaddle Schist Group is separated into four formations: Click, Rough Ridge, Sandy, and Honey. The dominant lithologies contained within the Packsaddle Schist are graphite schist, fine-grained metamorphosed rhyolite/felsite, marble, and
amphibolite schist.

Estimated maximum surface and subsurface thickness of the Packsaddle Schist within the Llano uplift is 6.8 km (Fay, 1986a). The thickness of the Packsaddle Schist is similar to the 7-10 km thickness of the Tillman Metasedimentary Group of the Texas-Oklahoma subsurface (south of the Wichita uplift). The Tillman Metasedimentary Group and Packsaddle Schist Group may be the stratigraphically equivalent parts of a Precambrian sedimentary sequence that covers much of the central Texas craton.

Cross-cutting the Llano Supergroup are various metamorphosed gabbros, granites, and diorites. On the basis of cross-cutting relationships, the oldest of these meta-igneous rocks is the Coal Creek Serpentinite (Barnes, 1988). This serpentinite contains inclusions of talc-schist and amphibolite-schist (apparently derived from the older Packsaddle Group) and is cut by aplite, pegmatite dikes, and “quartz bodies” (Barnes, 1988). The Coal Creek is also cross-cut by the Big Branch Gneiss (metamorphosed quartz-diorite). The youngest of the meta-igneous rocks is the metamorphosed granite Red Mountain Gneiss.

Intruding both the metasedimentary Llano Supergroup and the meta-igneous suite of rocks are several younger, unmetamorphosed granitic igneous rocks. The youngest of these is a quartz-granite-rhyolite porphyry called Llanite, formed as numerous cross-cutting dikes (Barnes, 1988). Older unmetamorphosed granites include: the Sixmile Granite, Gatman Creek Granite, and the Town Mountain Granite. The “quartz bodies” within the Coal Creek Serpentinite are inferred to be derived from the Town Mountain Granite (Barnes, 1988). Most of the larger granitic formations within the Precambrian strata of the Llano uplift give isotopic ages of 1030±20 Ma (Rb-Sr and K-Ar) (Barnes, 1988). In general, most of the igneous and meta-igneous rocks of the Llano uplift correspond chronologically to the Tishomingo/Troy Granites, granodiorite and orthogneiss of southeastern Oklahoma.

**Basement rocks beneath the Ouachita orogen**

**Ouachita Mountains**

The exact nature of the Cambrian and Precambrian basement rocks beneath the
Ouachita thrust belt is unclear, because of sparse deep wells that penetrate basement. No wells have penetrated igneous basement within the Ouachita Mountains. Because well data are lacking, interpretation of basement lithology and structure are based upon seismic and gravity data. Seismic and gravity data (Kruger and Keller, 1986; Keller and others, 1989; Mickus and Keller, 1992) for the Ouachita Mountains region indicate that continental crust (metamorphic/igneous) type basement rocks extends at least as far as the southeastern margin of the Broken Bow uplift and the southern margin of the Benton uplift (Plate 2.1). Seismic and gravity data indicate that the allochthonous strata of both the Broken Bow and Benton uplifts rest above faulted basement arches (Kruger and Keller, 1986; Mickus and Keller, 1992). Several other basement arches, most notable of which is the Waco uplift (Plate 2.1), are located along the Ouachita orogen in Texas. Attenuated continental basement or oceanic basement is postulated for areas farther southeast (Kruger and Keller, 1986; Mickus and Keller, 1992).

**Waco uplift**

The Shell No.1 Barrett well (Hill County, Texas) is one of the few deep wells that penetrated igneous basement rocks beneath overlying allochthonous rocks of the Ouachita orogen. Seismic reflection profiles across Hill County, Texas, show that the Shell No. 1 Barrett well is located in an arched structure (considered to be basement cored) called the Waco uplift (Nicholas and Rozendal, 1975) (Plate 2.1). The Shell No.1 Barrett well bottomed in 162 m of 1000-Ma quartz-diorite (Nicholas and Rozendal, 1975).

**Cambrian to Early Mississippian pre-orogenic strata of the Ouachita orogen (overview)**

The following chapter is a synthesis of regional stratigraphy of the Ouachita orogen and adjacent foreland. Some regional structures are discussed to show variations in stratigraphy across basins and uplifts. Stratigraphy for each of a set of time periods (listed below) is discussed for the Ouachita orogen and adjacent foreland from east (in Alabama) to west (Texas). For each time period, foreland facies stratigraphy is discussed first, followed by a discussion of deep-water facies. Plate 2.2 shows
correlation of stratigraphic units of the shallow-water foreland-facies Paleozoic succession of the foreland north and west of the Ouachita orogen from Mississippi in the east to central Texas in the west. Plate 2.3 shows the correlation between foreland facies and deep-water facies strata. North American series names are given within the text of this chapter and corresponding European series names are shown in Plates 2.2 and 2.3. Both charts use the timelines in the Mankin (1986) correlation of stratigraphic units in North America—Texas-Oklahoma tectonic region correlation chart as a template. Stratigraphic correlations in this discussion are based upon the following sources: Grohskopf (1955), Palmer (1962), Ham and others (1964), Bush and others (1977), Copeland (1986), Fay (1986a-i), Fay and others (1986), Haley and Stone (1986), Thomas and Osborne (1987a-b), Weaverling (1987), Ethington and others (1989), Gatewood and Fay (1991), Kier (1988), Thomas (1988), VanArsdale and Schweig (1990), and Thomas (1991).

This section is subdivided as follows: a) Late Cambrian-early Mississippian shallow-water facies pre-Ouachita orogenic strata, b) Late Cambrian-early Mississippian deep-water facies pre-Ouachita orogenic strata, pre-Atoka Formation (early Mississippian [Meramecian] to Pennsylvanian [latest Morrowan-earliest Atokan] shallow-water facies, pre-Atoka Formation (early Mississippian [Meramecian] to Pennsylvanian [latest Morrowan-earliest Atokan] deep-water facies, e) rapid-deposition phase, shallow-water facies (Morrowan-Desmoinesian)–Pottsville Formation, Atoka Formation, and Desmoinesian strata, f) rapid-deposition phase, deep-water facies Johns Valley Shale-Atoka Formation, g) post-orogenic strata of Ouachita foreland (latest Pennsylvanian [Virgilian]-Permian), h) post-orogenic strata of the Texarkana platform (Pennsylvanian [Desmoinesian]-Permian) (Plate 2.1).

A) Cambrian to Early Mississippian pre-orogenic strata of the Ouachita orogen and adjacent foreland—shallow-water, foreland facies

Basal transgressive sequence (Early to Late Cambrian)

_**Birmingham graben** (beneath Appalachian allochthon)_

The Birmingham graben is located in the subsurface beneath the Appalachian
thrust belt of central Alabama (Plate 2.1). Although it is not within the Ouachita structural salient, Cambrian stratigraphy within the Birmingham graben is briefly discussed to show comparison to Cambrian stratigraphy of the Ouachita foreland.

Well and seismic data show that the sedimentary rocks of the Birmingham graben are bounded by northeast-trending basement faults (Thomas, 1991). The trends of the basement faults are nearly parallel with the trends of basement faults in the Mississippi Valley graben (Plate 2.1). The sedimentary succession in the Appalachian thrust belt in Alabama begins with a 700+ m thick layer of Early Cambrian clastic rocks. These rocks are equivalents of Chilhowee Group clastic rocks that crop out in the Blue Ridge part of the Appalachian orogen (Plates 2.1 and 2.2) (Thomas, 1991). The Chilhowee Group is considered to mark the base of the Cambrian post-rift passive-margin transition (Thomas, 1991). A palinspastically restored cross section of the southern Appalachians shows a much thicker succession of Early to Middle Cambrian (pre-Knox) strata within the Birmingham graben (Figure 4 in Thomas, 1991).

The base of the Chilhowee Group clastic rocks is considered the basal transgressive unconformity surface (Plates 2.1 and 2.2). The Early Cambrian Chilhowee Group occupies the same lithostratigraphic position as the Late (possibly latest Middle) Cambrian Reagan Sandstone of the Arbuckle Mountains in Oklahoma and the Riley Formation of the Llano uplift of Texas (Plates 2.1 and 2.2). The difference in age of the basal transgressive units indicates that post-rift subsidence began earlier along the southern Appalachian foreland than farther west along the Ouachita foreland.

Above the basal Chilhowee post-rift clastic succession is an Early to Late Cambrian transgressive succession of non-marine to marine sandstones and shales that grade upward and eastward into limestones and dolostones (upper Chilhowee Group through Shady Dolostone), Rome Formation, and Conasauga Formation (interbedded carbonates and clastic rocks) (Plate 2.1) (Thomas, 1988, 1991; Rankin and others, 1989). The Shady Dolostone pinches out, and is absent in the subsurface along a basement arch west of the Birmingham graben (Rankin and others, 1989). The
maximum restored thickness of the Cambrian clastic and carbonate sequence (Chilhowee Group through Conasauga Formation) above the deepest part of the Birmingham graben is approximately 2 km (Thomas, 1991). Palinspastically restored cross sections constructed across the Birmingham graben show that the lower parts of the Rome-Conasauga succession are offset by normal faults (Thomas, 1991). Rocks within the faulted margins of the Birmingham graben are stratigraphically equivalent to mudstone facies farther west in the Mississippi Valley graben (Thomas, 1991).

**Black Warrior basin**

Two sedimentary basins, the Black Warrior basin and Mississippi Valley graben are located along the northeastern limb of the Ouachita structural salient (Plate 2.1). The Black Warrior basin is centered along the border between northeast Mississippi and northwest Alabama (Plate 2.1). Part of the Black Warrior basin is exposed along the southwestern foreland of the Appalachians; however, most of it lies within the Appalachian and Ouachita foreland beneath the Cretaceous Gulf Coastal Plain (Plate 2.1). Rocks along the southwestern and southeastern margins of the Black Warrior basin are imbricated within the Ouachita and Appalachian thrust belts, respectively.

Seismic and well data show that a maximum of <700 m of pre-Knox Group (pre-Franconian) Cambrian sandstones, shales, and carbonates rest above Precambrian basement within the Black Warrior basin (Thomas, 1988). Within the Black Warrior basin, the pre-Knox Group sedimentary sequence thins northward from a maximum near the Appalachian and Ouachita thrust fronts toward the Nashville dome (Thomas, 1988). The pre-Knox Group sequence is 300 m thick on the southwestern end of the Nashville dome in south-central Tennessee (Thomas, 1991). The proportion of clastic rocks in pre-Knox Group sequence increases eastward towards the Appalachian orogen (Thomas, 1988).

Well data for the northern and western Black Warrior basin show a thin (<80 m) basal sandstone (Weisner Sandstone in Rankin and others, 1989) and overlying sandy dolostone resting unconformably upon Precambrian basement (Figure 4 in Thomas, 1988). Because fossil data are lacking, no precise age is given for the
sandstone and lower part of the dolostone. Thin green shales interbedded within younger dolostones are lithostratigraphically correlated with green shales in the Conasauga Formation farther east in the Appalachian orogen (Plates 2.1 and 2.2). The basal sandstone-dolostone pair in the subsurface of the northern and western Black Warrior basin may correlate with the Chilhowee Group-Weisner Sandstone-Shady Dolomite (Early Cambrian) succession of the Appalachian orogen (as suggested by Figure 14 in Rankin and others, 1989). The basal sandstone-dolostone pair also lithostratigraphically correlates with Lamotte Sandstone-Bonneterre Dolomite succession farther west across the Ozark dome and subsurface of northern Arkansas (Plates 2.1 and 2.2) (Bush and others, 1977; Thomas, 1988; Van Arsdale and Schweig, 1990). On the basis of biostratigraphy, the age of the Lamotte-Bonneterre succession in the Ozark region (Plates 2.1 and 2.2) is early Late Cambrian (Dresbachian) (Haley and Stone, 1986). The Lamotte-Bonneterre succession rests unconformably upon Precambrian basement rocks in the subsurface of southern Missouri and northern Arkansas. The Late Cambrian Lamotte Sandstone (Arkansas and Missouri) may be a time transgressive basal sandstone equivalent of the Early Cambrian Chilhowee Group clastic rocks (Appalachians) (Plates 2.1 and 2.2). Depending on location within the Black Warrior basin, the age of the basal sandstone unit is likely intermediate in age between Early and Late Cambrian.

**Mississippi Valley graben/Reelfoot rift**

The Mississippi Valley graben is a northeast-trending, normal-fault-bounded basin that extends southward beneath the Ouachita thrust belt in the subsurface east of the Ouachita Mountains (Plate 2.1). The graben is buried by Mesozoic strata of the Mississippi embayment of the Gulf Coastal Plain. Wells in the region show that >1 km of pre-Knox Group Cambrian sedimentary rocks overlie crystalline basement in the Mississippi Valley graben (Thomas, 1991). According to regional well data, the pre-Knox Group Cambrian sedimentary succession is much thinner and locally absent outside the Mississippi Valley graben (Thomas, 1991).

The lithologic succession along the western margin of the graben varies (up-section) from relatively thin arkosic-quartzose sandstone, oolitic limestone, and
dolostone, upward to a thick succession of calcareous mudstone (Thomas, 1991). The lower rock units are similar to the Timbered Hills Group of the Arbuckle Mountains farther west (Plates 2.1 and 2.2). Lithologies along the eastern margin of the graben are apparently more randomly distributed, and include siltstone, mudstone, fine-grained sandstone, and silty to argillaceous limestone (Thomas, 1991). Trilobites indicate early Late Cambrian age for the upper part of clastic (syn-rift) succession (Grohskopf, 1955; Palmer, 1962; Weaverling, 1987). The maximum drilled thickness of pre-Knox Group rocks within the Mississippi Valley graben is approximately 1200 m; however, an interpreted cross section of northern part of the graben (Figure 4 in Thomas, 1991) suggests that the maximum thickness is nearly 2500 m.

**Arkoma basin-Ozark dome (Arkansas)**

The Arkoma basin is convex-to-northwest arcuate foreland basin which extends from western Oklahoma into Arkansas (Plate 2.1). The Ozark dome is located north of the Arkoma basin in northern Arkansas and southern Missouri (Plate 2.1). The oldest foreland-facies sedimentary rock exposed within Arkansas is the Early Ordovician Jefferson City Dolomite that crops out along the southern flank of the Ozark dome (Plates 2.1 and 2.2) (Bush and others, 1977). The basal transgressive succession in Arkansas is recognized in the subsurface as the Lamotte Sandstone and Bonneterre Dolomite (Plates 2.1 and 2.2). The oldest possible age for the Lamotte Sandstone is earliest Late Cambrian (Dresbachian) (Haley and Stone, 1986). The age of the basal transgressive formation in the subsurface of the Ozark dome and Arkoma basin region of Arkansas is younger than the Early Cambrian basal Chilhowee Group sandstone (Weisner Sandstone, Figure 14 in Rankin and others, 1989) of the Birmingham graben and southern Appalachian orogen (Plates 2.1 and 2.2).

No Paleozoic sedimentary rocks older than Dresbachian (early Late Cambrian) crop out within the Ouachita foreland of Arkansas, Oklahoma, and Texas (Ham, 1973; Bush and others, 1977; Mankin, 1986; Kier, 1988). Deep well data south and east of the Ouachita thrust-front are scarce and have not penetrated sedimentary rocks older than pre-Lamotte (Reagan) Sandstone (Morris, 1974; Denison, 1989). However, an interpreted structural cross section constructed across the eastern Arkoma basin
indicates abrupt southward thickening of pre-Lamotte strata near the Ouachita thrust front (Van Arsdale and Schweig, 1990). The eastern Arkoma basin is near the western boundary of the Mississippi Valley graben (Plate 2.1); however, the Van Arsdale and Schweig (1990) cross section is located west of the southern projection of the western boundary fault of the Mississippi Valley graben. It is unclear whether the pre-Lamotte strata shown in the Van Arsdale and Schweig (1990) cross section thickens southward toward a deep basin located to the south, or southeast towards the Mississippi Valley graben (Thomas, 1991). The estimated maximum thickness of the pre-Lamotte strata in the Van Arsdale and Schweig (1990) cross section is approximately 2400 m (1 second 2-way travel time at an estimated interval seismic velocity of 16000 ft/sec). The estimated thickness of 2400 m is approximately equivalent to maximum interpreted thickness of pre-Knox Cambrian strata in the deep central part of the northern Mississippi Valley graben (Plates 2.1) (Figure 4 in Thomas, 1991).

**Southern Oklahoma aulacogen (Wichita and Arbuckle uplifts)**

Exposures in the Southern Oklahoma aulacogen show that an erosional nonconformity separates the Precambrian and Early Cambrian igneous terrain from a Late Cambrian (Franconian) transgressive succession of sandstones and dolomitized limestones. This transgressive sequence locally onlaps onto paleotopographic highs composed of rhyolite, and basal units are quartz-rich and contain glauconite (Figure 2.1) (Donovan, 1986). The basal formation exhibits a decrease in glauconite and quartz content up-section and a corresponding increase in limestone content (Figure 2.1) (Donovan, 1986).

In Oklahoma, these rocks are called the Timbered Hills Group (Reagan Sandstone and Honey Creek Formation) named for the type localities in the western Arbuckle Mountains (Plates 2.1 and 2.2). The Reagan Sandstone is arkosic and glauconitic, whereas the overlying Honey Creek is trilobite-pelmatozoan limestone (Ham, 1973). North of the Arbuckle Mountains, the Honey Creek Formation grades into fossiliferous, sandy limestones (Ham, 1973). Both formations of the Timbered Hills Group are thickest in the western parts of the Arbuckle Mountains (Reagan = 137 m, Honey Creek = 33 m) (Ham, 1973).
**Llano uplift**

Surface geology in the Llano uplift region of central Texas also includes a transgressive succession of rocks that sit unconformably on basement (Kier, 1988). Here the onlap surface is above tightly folded Precambrian metasedimentary and meta-igneous rocks. In the Llano uplift, Middle to Late Cambrian (possibly Early Ordovician) sandstones, limestones, siltstones and dolomites rest unconformably above Precambrian basement rocks (Kier, 1988).

Stratigraphy of the Llano region indicates that the basal transgressive phase of Paleozoic deposition began earlier (Dresbachian rather than Franconian) and continued longer here than farther north in the Southern Oklahoma aulacogen (Late Cambrian/Early Ordovician rather than Late Cambrian) (Plates 2.1 and 2.2). Age of the onlapping basal transgressive formations likely varies as a function of differences in paleotopography and regional thermal and isostatic subsidence histories.

The basal transgressive sequence of rocks in the Llano region is assigned to the Moore Hollow Group which includes the Riley and Wilberns formations (Figure 2.1 and Plate 2.2) (Kier, 1988). The oldest unit, the Riley Formation is subdivided into three members. The older two have roughly equal maximum thicknesses (Hickory Mountain Sandstone = 143 m, and Cap Mountain Limestone = 125 m) and the youngest is much thinner (Lion Mountain Sandstone = 21 m) (Kier, 1988).

From bottom to top, the Hickory Mountain Sandstone Member ranges from siltstone (locally arkosic), local feldspathic pebble conglomerate, argillaceous and micaceous siltstone, and coarse-grained iron-cemented sandstone (Kier, 1988). The variable lithology of the Hickory Mountain Sandstone Member results from paleotopography of the basement onlap surface (Figure 2.1). Local relief of this nonconformity surface is estimated to be 200 m in the Llano uplift region (Kier, 1988). The relief of the nonconformity surface in the Llano region is double the relief of the basal transgressive nonconformity surface in the Slick Hills region of the Southern Oklahoma aulacogen (110 m) (Figure 2.1) (Donovan, ed., 1986).
The Cap Mountain Limestone Member changes upsection from sandy limestone, to a silty limestone, to glauconitic limestone. The Lion Mountain Sandstone Member contains coarse-grained, glauconitic, crossbedded sandstone, and glauconitic trilobite-bearing coquina with phosphatic brachiopods. Thicknesses of all members of the Riley Formation increase eastward towards the Ouachita thrust-front (Kier, 1988).

Within the Llano uplift, the Wilberns Formation rests unconformably above the Riley Formation. The Wilberns Formation is subdivided into four members. The basal Welge Sandstone Member is a sparsely fossiliferous, non-glauconitic, medium-to coarse-grained well-sorted quartz sandstone (Kier, 1988). Glauconite content increases to the southeast, and contact between the Welge Member and the underlying Lion Mountain Sandstone Member is conformable to the east in the subsurface beneath the Ouachita thrust-front (Kier, 1988). The Weldge Sandstone Member is relatively thin (9 m). Above the Welge Sandstone Member, the Morgan Creek Limestone Member commonly contains ooids, and varies upsection from a glauconitic limestone, to a silty and argillaceous limestone, to interbedded glauconitic and silty limestone (Kier, 1988). The Morgan Creek Member grades westward into a sandstone. The Point Peak Member, above the Morgan Creek Member, consists of thin-bedded, argillaceous, glauconitic, calcareous siltsone and argillaceous silty limestone. The upper part of the Point Peak contains intraformational conglomerates and stromatolites (Kier, 1988). Capping the Wilberns Formation of the Moore Hollow Group is the San Saba Member. The San Saba contains fine- to medium-grained glauconitic limestone, coarse-grained limestone, chert, and dolostone. Limestone and stromatolites are more prevalent in western (older) parts of the San Saba Member, whereas dolomite is predominant in eastern (younger) parts. The boundary between the Moore Hollow Group and the overlying Ellenburger Group is "time-transgressive," placed in the upper Cambrian in the east and in the subsurface, and in the lower Ordovician to the west. All formations in the Moore Hollow Group thicken eastward.

In general, the lithology of the Moore Hollow Group shows an upward decrease in quartz and feldspar content and increase in limestone content. Glauconite
is concentrated in the middle and upper parts of the Riley Formation, and is found in all the Wilberns Formation lithologies. Chert is concentrated in the uppermost part of the Moore Hollow Group. The Moore Hollow Group is similar lithologically to the Lamotte Sandstone-Bonneterre Dolomite succession of Arkansas and the Timbered Hills Group (Reagan Sandstone, Honey Creek Formation) of southern Oklahoma; however, chrono-stratigraphically, the top of Moore Hollow Group corresponds with the top of the Signal Mountain Formation limestone of the lower Arbuckle Group (Cambrian-Ordovician boundary) in the Southern Oklahoma aulacogen (Plates 2.1 and 2.2) (Ham, 1973; Fay 1986a and h). Age of the top of the Moore Hollow Group decreases in the western parts of the Llano uplift where it is Early Ordovician (Kier, 1988).

Knox-Ar buckle-Ellenburger time (Late Cambrian to Early Ordovician)

Black Warrior basin and Appalachian orogen

Restoration of a cross section to remove Appalachian thrust translation across the Black Warrior basin and Birmingham graben (southeastern Appalachian orogen) shows no offset of Knox Group strata across the Birmingham graben (Thomas, 1991). Therefore, by Late Cambrian time, subsidence within the Birmingham graben ceased, and the graben was completely filled. Therefore, for discussion of Knox Group and younger rocks, the Birmingham graben area is considered the intersection of the Black Warrior basin with the southern Appalachian orogen (Plates 2.1 and 2.2).

In eastern parts of the Black Warrior basin along the Appalachian thrust front, the latest Middle to Upper Cambrian Ketona Dolomite rests upon the Conasauga Formation (Plate 2.2) (Thomas and Osborne, 1987a,b; Thomas, 1988). The Ketona Dolomite is recognized in central and eastern Alabama, and is covered by the Copper Ridge Dolomite (lower Knox Group) (Rankin and others, 1989). The Knox Group is a Late Cambrian (Franconian) to Early Ordovician (Canadian) sequence of shallow-water platform dolostones and limestones (Thomas, 1988). Proportion of limestone in the Knox Group increases upsection and southward towards the Appalachian-Ouachita thrust fronts (Thomas, 1988). Lesser lithologies within the Knox Group are quartz sandstone, sandy-limestone/dolostone, and cherty limestone/dolostone (Thomas,
The maximum restored thickness of the Knox Group around the Birmingham graben is 1300 m (Thomas, 1991). The Knox Group is both lithologic and chronologic equivalent of the Arbuckle Group of Arkansas, Oklahoma, and northern Texas and the upper Moore Hollow through Ellenburger groups of Texas.

Mississippi Valley graben

A palinspastically restored cross section constructed across the northern Mississippi Valley graben shows a 200 m down-to-east normal offset of the base of the Knox Group along the western boundary fault (Figure 4 in Thomas, 1991). The Thomas (1991) restored cross section also shows the base of the Knox Group on the western side of the Mississippi Valley graben 200 m higher than on the eastern side, and that the base of the Knox Group is homoclinally tilted towards the eastern margin of the Mississippi Valley graben (Thomas, 1991).

The Knox Group rocks within the Mississippi Valley graben are predominantly limestones and dolostones. The stratigraphy of the Knox Group differs little from the Mississippi Valley graben eastward into the Black Warrior basin. Local dark-colored, fine-grained limestones within Knox Group strata in the southern Mississippi Valley graben suggest southward increase in water depths (Thomas, 1988). The Knox Group thickens to a maximum of 1500 m along the eastern margin of the Mississippi Valley graben (Figure 4 in Thomas, 1991).

Arkoma basin-Ozark dome (Arkansas)

West of the Mississippi Valley graben, the Late Cambrian-Early Ordovician carbonate succession is called Arbuckle Group (Plates 2.1 and 2.2). The Arbuckle Group is the recognized term used for much of the southern midcontinent of North America. Across the Ozark dome region of northern Arkansas and Missouri (Plate 2.1), the Arbuckle Group is subdivided into three Late Cambrian and six Early Ordovician Formations (Bush and others, 1977; Van Arsdale and Schweig, 1990; Suhm, 1997). The basal formation (Franconian) of the Arbuckle Group is called by several names (Derby, Doerun, or Davis Formation), depending on location, and rests conformably upon the Bonneterre Dolomite (Plate 2.2) (Bush and others, 1977; Van Arsdale and Schweig, 1990). The Derby-Doerun-Davis Formation is sequentially
overlain by Potosi and Eminence Dolomites (Plate 2.2). The top of the Eminence Dolomite marks the top of the Cambrian in the vicinity of the southern Ozark dome (Plate 2.2) (Bush and others, 1977; Van Arsdale and Schweig, 1990). The Derby-Doerun-Davis-Eminence succession of Arkansas correlates with the Fort Sill Formation-Bally/Royer Dolomites of the Hunton arch, and with the Fort Sill-Signal Mountain Formation limestones of the Ardmore basin (Plates 2.1 and 2.2) (Ham, 1973).

Resting above the Eminence Dolomite is the Gasconade-Van Buren Formation (Bush and others, 1977; Van Arsdale and Schweig III, 1990) (Plate 2.2). The Gasconade-Van Buren Formation is predominantly dolostone, but lower beds contain local, thin sandstones (Gunter Member) (Gatewood and Fay, 1991). The Roubidoux Formation dolostone rests upon the Gasconade-Van Buren Formation (Bush and others, 1977; Van Arsdale and Schweig, 1990). The Gasconade-Van Buren-Roubidoux Formations sequence correlates with the McKenzie Hill-Cool Creek Formation sequence of Oklahoma (Plates 2.1 and 2.2) (Gatewood and Fay, 1991).

The complete Arbuckle Group is preserved south of the Ozark dome in the subsurface of the southeastern Arkoma basin and farther east towards the southern Mississippi Valley graben (Plate 2.1). Here, a succession of four dolostones (Jefferson City-Cotter-Powell-Smithville Dolomites) overlies the underlying Roubidoux Formation (Plate 2.2) (Bush and others, 1977; Van Arsdale and Schweig, 1990; Suhm, 1997). Westward, towards the South Ozark arch (Plate 2.1), the Smithville and upper parts of the Potter Dolomite are truncated by an unconformity. Farther north, toward the center of the Ozark dome, the Powell and Cotter Dolomites are completely removed by erosion, and the Jefferson City Dolomite crops out (Haley and others, 1993). The Jefferson City-Smithville Dolomite succession correlates with the Kinblade-West Spring Creek Formation succession of Oklahoma (Plate 2.2) (Ham, 1973).

The total thickness of Arbuckle Group carbonates north of the Ouachita thrust front in Arkansas ranges from less than 610 m on the southern flank of the Ozark dome to more than 1360 m in the southeastern Arkoma basin (Gatewood and Fay,
1991; Van Arsdale and Schweig, 1990). Because of sparse well data and great depths, thickness of Arbuckle Group rocks south of the Ouachita thrust-front is poorly known.

**Southern Oklahoma aulacogen (Arbuckle and Wichita uplifts)**

Conformably on top of Honey Creek Formation dolostone is an Upper Cambrian (Franconian) to basal Middle Ordovician (earliest Whiterockian) (Ham, 1973, Ethington, and others, 1989) sequence of alternate mostly dolomitized limestones and thin micrite/shale partings known as the Arbuckle Group in Oklahoma, Arkansas, and Texas (northeast of the Muenster arch) (Plates 2.1 and 2.2). The Arbuckle Group is subdivided into six limestone formations which are listed in ascending order: Fort Sill and Signal Mountain (Franconian-earliest Ibexian), McKenzie Hill, Cool Creek, Kinblade, and West Spring Creek (Ibexian -earliest Whiterockian) (Ham, 1973, Ethington, and others, 1989). Only within the Ardmore basin (Plate 2.1) is the complete West Spring Creek preserved (Ham, 1973). Elsewhere in the Arbuckle uplift, Middle Ordovician Simpson Group strata rest unconformably upon the Lower Ordovician part of West Spring Creek Formation (Ham, 1973). Locally, part or all of the Fort Sill and Signal Mountain limestones are replaced by thick "brown" dolomites (Ham, 1973). These formations are called Butterfly-Royer Dolomite in the Arbuckle Mountains (Plate 2.1) of southeastern Oklahoma (Ham, 1973) and Bally Mountain Dolomite in the Slick Hills (Plate 2.1) of southwestern Oklahoma (Donovan and Ragland, 1986). Dolomite replacement increases towards the east in the Arbuckle Mountains. The dominant Arbuckle Group lithologies include: micrites, intraclast calcarenites, stromatolites, laminated dolomites and dolomitized limestones (Ham, 1973). However, thin, quartz sandstone members are found at the bases of the Cool Creek Formation (Thatcher Creek Member) and West Spring Creek Formation in the Slick Hills (Donovan and Ragland, 1986).

Except for the Cool Creek Formation, all the Arbuckle Group limestones that crop out in the Arbuckle Mountains are fossiliferous, containing trilobites, pelmatozoans, sponges, brachiopods, mollusks, and graptolites in the uppermost part of the West Spring Creek Formation (Ham, 1973). Farther north, within the interior craton, where the entire Arbuckle Group is dolomitized, key "silicified fauna" are used
for stratigraphic control (Ham, 1973). Measured thickness of the Arbuckle Group in the Arbuckle Mountains decreases eastward from 2.0 km to 1.2 km where the Cretaceous Gulf Coastal Plain covers all the older Paleozoic strata. However, proprietary well and seismic data suggest that Arbuckle Group (possibly including Timberred Hills Group) limestones may thicken to as much as 2.4 km towards the northern flank of the Ardmore basin and beneath the Ouachita thrust belt in eastern Oklahoma (Gatewood and Fay, 1991). The Arbuckle Group thins abruptly to the north of the Arbuckle Mountains and is partly absent along the Nemaha ridge (Plate 2.1).

**Ardmore basin and Muenster arch**

Well data shows that the Arbuckle Group thins abruptly toward Texas south of the central axis of the Southern Oklahoma aulacogen. Thickness decreases southwestward from the northern boundary of the Ardmore basin perpendicular to the basin axis (Plate 2.1). Wells in the southern part of Ardmore basin show the maximum thickness of a complete Arbuckle Group sequence to be 1.2 km (Cooper, 1997). Cross sections of part of the Ardmore basin show progressive southwestward truncation of pre-Pennsylvanian strata (Cooper, 1997). Therefore, thinning of the Arbuckle Group south of the Southern Oklahoma aulacogen is likely a result of erosion rather than non-deposition.

Thickness also decreases along the crest of the Muenster arch (Plate 2.1) from an estimated maximum thickness of 1.5 to 1.8 km beneath the Ouachita thrust front northwestward to less than 304 m in north-central Texas (Ewing, 1991; and proprietary seismic data). In some locations along the crest of the Muenster uplift, the entire Arbuckle Group, and all overlying pre-orogenic rocks, are absent because of erosion (Bradfield, 1957a-c; Ewing, 1991). The cross sections of Denison (1982) and Cooper (1997) suggest that as much as 1.2 km of Arbuckle Group has been eroded from the central axis of the Muenster uplift (southwest of the Ardmore basin) prior to deposition of the overlying Pennsylvanian (Desmoinesian) Strawn Group shales, sandstones, and conglomerates.
**Llano uplift**

South of the Muenster uplift, in the Llano uplift (Plate 2.1) of central Texas, the late Cambrian-early Ordovician carbonate-dominated sequence of rocks is much thinner than the equivalent strata to the north in the Arbuckle uplift. The entire carbonate sequence thickens away from the center of the Llano uplift (where Precambrian rocks are exposed). The chronostratigraphic equivalent of the Arbuckle Group of north Texas, Oklahoma, and Arkansas is called the Ellenburger Group in the Llano uplift region (Plate 2.2). The Ellenburger Group is subdivided into three formation listed in ascending order: Tanyard, Gorman, and Honeycut. The Tanyard Formation consists of a lower member (Threadgill) which varies west to east (upsection) from argillaceous, silty limestone to a medium- to coarse-grained dolomite; and an upper member (Staendebach) which varies west to east (upsection) from dolomite to "dolomoldic and oolitic" cherty dolomite (Kier, 1988). Above the Tanyard, the Gorman Formation varies (upsection) from microgranular dolomite to micrite containing lenses of quartz sandstone (Kier, 1988). The Ellenburger Group is capped by the Honeycut Formation, which consists of limestone, dolomite, and microgranular dolomite (Kier, 1988). The maximum measured thickness of the Ellenburger Group on the flanks of the Llano uplift is greater than 447 m (Kier, 1988). Wells to the southeast show that the Ellenburger thickens beneath the Ouachita thrust-front (Flawn and others, 1961; Ewing, 1991).

The Ellenburger Group is considered the stratigraphic equivalent of the Arbuckle Group; however, the upper part of the Moore Hollow Group is chronostratigraphically equivalent to the lower part of the Arbuckle Group in southern Oklahoma. For regional comparison, it is best to combine all the Dresbachian-early Whiterockian strata, and the maximum thickness of this succession of rocks is 933 m in the Llano uplift (Kier, 1988) and 2.1 to 2.4 km in southeastern Oklahoma (Ham, 1973; Gatewood and Fay, 1991).
Post Sauk unconformity (Simpson Group time)—onset of episodic influx of terrigenous sediment from interior North American craton (Early to Late Ordovician)

Black Warrior basin and Appalachian orogen

In the southern Appalachian thrust belt, a Middle to Late Ordovician succession of clastic rocks rests unconformably above the Knox Group carbonates. The dense Knox Group dolostones beneath the unconformity contrast with the overlying thin Lenoir Limestone-Athens Shale strata (lower Whiterockian) (Plate 2.2) (Mankin, 1986; Thomas, 1988; Drake and others, 1989; Finney, 1997). The Lenoir Limestone and Athens Shale strata correlate with the lower part of the Chickamauga Group carbonates farther west within the subsurface of the Black Warrior basin (Plate 2.2) (Thomas, 1988). Thickness of the Chickamauga Group (approximately 350 m) varies little within the Black Warrior basin (Figure 4 in Thomas, 1988). The pre-Trentonian beds of the Chickamauga Group correlate with the Simpson Group of the southern Midcontinent region of Oklahoma, Arkansas, and Texas (Plate 2.2).

Rare, thin clastic beds are found within lower and middle parts of the Chickamauga Group carbonates. In northern and eastern parts of the Black Warrior basin, the lower Chickamauga Group includes dolomitic green shales, limestones, and sandy limestones (Thomas, 1988). Sandstones are not common within the Chickamauga Group; however, the correlative Simpson Group of the southern midcontinent of North America contains numerous sandstone beds (Suhm, 1997). The very coarse-grained (pebble-boulder) Attala Chert Conglomerate in the basal Chickamauga Group in the Appalachian thrust belt contains chert derived from the underlying Knox Group dolostones (Thomas and Drahovzal, 1973). The Attala Chert Conglomerate is a basal conglomerate analagous to the laterally discontinuous basal Joins Formation (basal Simpson Group) conglomerate in the subsurface of the eastern end of the Southern Oklahoma aulacogen (Plates 2.1 and 2.2) (Suhm, 1997). However, the Joins conglomerate contains clasts of limestone and dolostone rather than chert (Suhm 1997).
The unconformity at the top of the Knox Group in the southern Appalachians is the Sauk-Tippecanoe unconformity (Thomas, 1988; Finney, 1997). The Sauk-Tippecanoe unconformity is more prominent farther west in the southern and central part of North America where Middle Ordovician sandstones rest upon Early Ordovician carbonates (Finney, 1997; Suhm, 1997). Farther north within the Appalachians, Middle Ordovician limestones locally rest conformably upon Early Ordovician dolostones (Ross and others, 1982; Finney, 1986; Finney and others, 1996; Finney, 1997). The Sauk-Tippecanoe unconformity is not clearly defined within the Black Warrior basin (Plate 2.1) (Figure 4 in Thomas, 1988). The lack of laterally continuous sandstone beds (i.e. Simpson-type sandstones) above the Knox Group suggests that the most (if not all) of the Black Warrior basin remained below sea-level during the Early and Middle Ordovician. Horizontal and vertical changes between dolostone and limestone illustrated on a regional stratigraphic cross section of the Black Warrior basin indicate temporal and geographical changes between shallower subtidal and supratidal (dolostone) and deeper subtidal (limestone) deposition (Figure 2.2) (Thomas, 1988).

**Arkoma basin-Ozark dome (Arkansas)**

The Middle to basal Upper Ordovician Simpson Group rocks in the subsurface of the Arkoma basin in Arkansas are the Everton Formation, St. Peter Sandstone, Joachim Dolostone, and the dense Plattin Limestone (Plate 2.2) (Ethington and others, 1989; Van Arsdale and Schweig, 1990; Suhm, 1997). The base of the Everton Formation rests unconformably from east to west upon: Smithville Formation (dolostone), Powell Dolomite, and Cotter Dolomite (Suhm, 1997). The Smithville, Powell, and Cotter are all dense limestones and dolostones that correlate with the Arbuckle Group of Oklahoma and northern Texas (Plate 2.2). The unconformity at the base of the Everton Formation is the Sauk-Tippecanoe unconformity that marks the base of a succession of formations with up-section, episodic increase in quartz and clastic detritus derived from the North American craton (Suhm, 1997).

In the eastern part of the Arkoma basin, the Everton Formation is separated into four members (Member A limestone and dolostone, Calico Rock Member (sandy
limestone and sandy dolostone), and Members B and C (limestones and dolostones) (Figure 2.3, and Plate 2.2) (Suhm, 1997). The Calico Rock Member pinches out westward where Member A and the Calico Rock Member are replaced by the Sneeds Member (limestone and dolostone) (Figure 2.3). Along the southern flank of the Ozark dome (Plate 2.1) in northern Arkansas, a conglomerate at the base of the Sneeds Member contains clasts of “detrital chert and dolomite in a sandy dolomite matrix” (Suhm, 1997). The chert and dolomite clasts are derived from the underlying Powell and Cotter dolostones beneath the Sauk-Tippecanoe unconformity (Suhm, 1997). The basal Sneeds conglomerate likely correlates with the Attala Chert Conglomerate of the Black Warrior basin (Thomas and Drahovzal, 1973) (Plate 2.2).

Farther west towards the northeastern flank of the Arbuckle uplift (Plate 2.1), both the Sneeds Member and Member B limestones and dolostones interfinger with and grade into the Newton Member of the Everton Formation (Figure 2.3) (Suhm, 1997). The Newton is a sandy limestone/dolostone which correlates with, and extends into the Burgen Sandstone towards northeast Oklahoma (Suhm, 1997). The Newton Member is also equivalent to the sandy facies at the basal part of the Oil Creek Formation within the Arbuckle uplift (Figure 2.3, and Plates 2.1 and 2.2) (Suhm, 1997).

In the northeastern Arkoma basin, the Member C limestones and dolostones of the Everton Formation rest above the Member B limestones and dolostones (Figure 2.3). Farther west, the limestones and dolostones of Member C grade into the Jasper Member (Figure 2.3) (Suhm, 1997). The lithology of the Jasper varies locally from sandstone to fossiliferous limestone to dolostone (Suhm, 1997). The sandstone facies of the Jasper correlates with the lower part of the middle Tyner Formation in the Ozark region and the sandstone at the base of the McLish Formation in the Arbuckle uplift (Figure 2.3, and Plates 2.1 and 2.2). In western Arkansas, Member C of the Everton Formation is a quartz-enriched sandy limestone/dolostone (Suhm, 1997). Member C correlates with the lower Tyner Formation of the Ozark dome region (Plates 2.1 and 2.2) (Suhm, 1997). Southwest toward the Arbuckle uplift in Oklahoma, the sandy-facies Member C of the Everton Formation grades into “shaley”
limestones of the Oil Creek Formation (Figure 2.3, and Plates 2.1 and 2.2) (Suhm, 1997). Shale layers in the Oil Creek are thin and make up a small fraction of the Oil Creek Formation; however, they indicate initial stages of terrestrial sediment influx onto the carbonate shelf in the southern Midcontinent (Suhm, 1997).

Resting above the Everton Formation in eastern Arkansas is the Joachim Formation (Plate 2.2). The Joachim Formation varies from quartz-rich sandy limestones at the base to micrites and dolomicrites with “rip-up clasts, mud cracks, and stromatolites” (Craig and others, 1986). In central Arkansas, much of the Joachim Formation is replaced by the St. Peter Sandstone (Plate 2.2). The St. Peter Sandstone is a quartz-rich, reworked, eolian sandstone that extends across a northeast-trending area in central Arkansas (Suhm, 1997). St. Peter-type sandstones are locally found farther southwest in the Arbuckle uplift of Oklahoma within the McLish and Tulip Creek Formations of the Simpson Group (Plates 2.1 and 2.2) (Suhm, 1997).

Above the Joachim Formation limestones and dolostones in the northern Arkoma basin rests the Plattin Formation (Plate 2.2). The Plattin limestone is “sublithographic to lithographic” and contains “several very thin green-shale beds” (Suhm, 1997). The base of the Plattin is a disconformity in western parts of the Arkoma basin, but appears to be conformable in eastern parts (Suhm, 1997). In the northeastern Arkoma basin, the Plattin limestones interfinger with the Joachim Formation. Farther northeast, in the subsurface of Missouri, the Joachim and Plattin limestones interfinger with the St. Peter Sandstone and Starved Rock Sandstones (Plate 2.2) (Suhm, 1997).

The Plattin limestone correlates with the Bromide Formation in in the Arbuckle uplift of Oklahoma (Plates 2.1 and 2.2). The upper part of the Bromide Formation in Oklahoma contains thin green-colored shales similar to those found in the Plattin limestones (Suhm, 1997). Across the Arbuckle uplift and along the southern flank of the Ozark dome (Plate 2.1), the Bromide Formation is predominantly fine- and medium-grained limestones and dolostones. The limestone/dolostone facies of the Bromide interfingers with a sandy, lime-mudstone facies along the southern flank of the Tishomingo-Belton anticline ( Figure 2.3, and Plate 2.1) (Suhm, 1997).
Across the Arbuckle uplift outcrop belt, and in the subsurface along the southern flank of the Ozark dome (Plate 2.1), an unconformity separates the top of the Bromide Formation from the base of the overlying Viola Group (Plate 2.2) (Suhm, 1997). Farther north toward the center of the Ozark dome, more of the upper Simpson Group rocks are missing beneath the base of the Viola Group. Here, the “Viola Group” type rocks (Trentonian stage) rest unconformably upon the middle Tyner Formation (Upper Chazy stage, correlative with upper Tulip Creek Formation of Oklahoma and Joachim Formation of Arkansas) (Plate 2.2) (Suhm, 1997).

Overall, the total thickness of the Simpson Group rocks is minimum along the southern flank of the Ozark dome where the upper part is truncated by a pre-Viola Group unconformity (Suhm, 1997). Here the maximum thickness of the Simpson Group is less than 100 m. The thickness of the Simpson Group increases both to the southwest towards the Ardmore basin (Plate 2.1) and eastward towards Mississippi. Thickness of the Simpson Group in Arkansas is at a maximum in the eastern part of the Arkoma basin (Plate 2.1). Here, well data indicate that the maximum subsurface thickness of Simpson Group rocks is approximately 450 m (Suhm, 1997). Most of the eastward thickening results from increased thickness of the Everton Formation (Figure 2.3) (Suhm, 1997).

**Southern Oklahoma aulacogen (Wichita and Arbuckle uplifts)**

Above the Arbuckle Group carbonates of Oklahoma, Arkansas, and north Texas is a Middle Ordovician (Whiterockian) to basal Upper Ordovician (Mohawkian [Black Riverian]) succession of interbedded shales, limestones and sandstones called the Simpson Group (Plate 2.2) (Ham, 1973, Ethington, and others, 1989). In contrast to the Arbuckle Group, which consists mostly of dense limestones and dolostones, the lithologies of the Simpson Group are more variable. The combination of mapping of surface geology in the Arbuckle and Wichita uplifts of Oklahoma and detailed mapping of subsurface Simpson Group strata based upon numerous regional wells indicate lateral interfingering of Simpson Group lithologies (Figure 2.3) (Suhm, 1997).

One distinct lithology in the Simpson Group is quartz arenites ("cleanly washed sands") (Ham, 1973). The clean quartz arenites represent fossil beach sands
deposited along both the shoreline and in barrier beaches during “Simpson-time” (Suhm, 1997). Provenance studies show that the source area for most of the quartz grains and clasts in the Simpson Group sandstones is the interior craton north of the Ouachita Mountains (Suhm, 1997). Paleocurrent analyses for allochthonous deep-water facies equivalents of the Simpson Group (see Plate 2.3) that crop out in the Ouachita Mountains also indicate a northerly source area for mineral and rock clasts (Lowe, 1989).

North and east of the southern flank of the Tishomingo-Belton anticline (Plate 2.1), several unconformity surfaces are recognized within the Arbuckle and Simpson Groups (Ham, 1973; Denison, 1997). Only the deep central part of the Ardmore basin is free of recognized unconformities within Ordovician strata (Suhm, 1997; Ham, 1973). The unconformity surface located below the base of the Simpson Group is the Sauk-Tippecanoe unconformity. The exact placement of the unconformity surface within the Ordovician strata of the Southern Oklahoma aulacogen varies among researchers. Stratigraphic columns of Ham (1973) show the major unconformity at the base of the Simpson Group (either base of Joins or Oil Creek Formations). Later publications place the Sauk-Tippecanoe unconformity within the upper part of the West Spring Creek Formation (uppermost Arbuckle Group) at the top of the Canadian Series rocks (Ross and others, 1982; Finney, 1986; Derby and others, 1992; Finney, 1997).

The Sauk-Tippecanoe unconformity is recognized throughout much of the interior craton of North America (Finney, 1997). The region-wide unconformity corresponds to a time of relatively low sea level. Biostratigraphy indicates a eustatic drop in sea level within the upper Arbuckle Group (top of Canadian Series) (Ethington and Dresbach, 1990; Derby and others, 1991). However, because the length of unrecorded time represented by the unconformity is at a maximum in the central part of the North American craton, the relative drop in sea level also had an important isostatic component. The influx of craton-derived quartz and rock clasts into the Simpson Group (and into deep-water off-shelf facies equivalents farther south) is the result of erosion within the midcontinent.
Besides the quartz arenites, another important lithology recognized in the Simpson Group is green-colored shales (Ham, 1973). Green-colored shales are not found below the Simpson Group but are recognized at several horizons above the Simpson Group. The green shales serve as key marker beds throughout southern Oklahoma outcrop region. Of key importance is the recognition of green shales in lower parts of the deep-water facies equivalent of the upper Simpson Group (Womble Shale) at Black Knob Ridge (Figure 2.4, and Plates 2.1 and 2.2) (Morris, 1974). The green-colored shales may represent transitional facies intertongues between the lighter-colored (more oxidized) shallow-water facies Simpson Group and the dark gray- to black-colored (more reduced) deep-water facies Womble Shale (Figure 2.4, and Plates 2.1 and 2.2).

The Simpson Group of Oklahoma and north Texas is subdivided into five formations listed in ascending order: Joins Limestone (thin, absent in eastern part of Arbuckle uplift), Oil Creek, McLish, Tulip Creek, and Bromide (Plate 2.2) (Ham, 1973). Along the southern flank of the Tishomingo-Belton anticline (Plate 2.1), the Joins Limestone is a carbonate conglomerate which rests unconformably upon Arbuckle Group limestones (Suhm, 1997). Farther north, the Joins Limestone interfingers with quartz-rich sandstones of the basal Oil Creek Formation (Figure 2.3) (Suhm, 1997). The basal Oil Creek Formation rests unconformably upon Arbuckle Group along the crest of the Tishomingo-Belton and Hunton anticlines (Figure 2.3 and Plate 2.1). The overlying middle Oil Creek Formation also grades from limestone facies to sandstone facies (arenites and calcarenites) from southwest to northeast over the crest of the Tishomingo-Belton anticline (Figure 2.3 and Plate 2.1) (Suhm, 1997). However, the limestone facies of the middle Oil Creek Formation extends farther north (Figure 2.3) (Suhm, 1997). Farther east, the limestone facies of the middle Oil Creek Formation interfingers with the sandy limestone/dolostone facies Member C of the Everton Formation (Figure 2.3). In contrast to the Oil Creek Formation, the overlying McLish and Tulip Creek and Bromide Formations switch southwestward from limestone facies along the crest of the Tishomingo-Belton and Hunton anticlines.
to sandstone facies along the southern flank of the Tishomingo-Belton anticline (Figure 2.3 and Plate 2.1).

In the southwestern part of the Arbuckle Mountains, the Simpson Group has a maximum exposed thickness of 700 m and consists of roughly equal proportions of skeletal calcarenites, skeletal calcilutites, shale, and sandstone (Ham, 1973). Subcrop maps based upon regional well data show that the Simpson Group is also thickest in the subsurface around the southeast-plunging nose of the Tishomingo-Belton anticline (Suhm, 1997) (Figure 2.3 and Plate 2.1). The maximum exposed thickness of the Simpson Group along the northeastern flank of the Hunton anticline in eastern Oklahoma is approximately 300 m (Ham, 1973; Suhm, 1997).

Fossil fauna of the upper Simpson Group differs from that of older Paleozoic carbonates. Most notably, bryozoans, cystoids, crinoids, and algal-mats not found in older rocks distinguish Simpson Group carbonates (Ham, 1973). Mollusks, sponges, trilobites, and a diverse set of brachiopods are also found in the Simpson Group (Ham, 1973). Hemispherical stromatolites (found in older Ordovician carbonates) and deep-water graptolites (found in overlying Viola Group limestones) are noticably lacking (Ham, 1973).

**Ardmore basin and Muenster arch**

Well data show that the Simpson Group has a maximum thickness more than 760 m in the deepest part of the Ardmore basin. The stratigraphic succession varies from: the basal Joins Formation coarse-grained limestone; upsection to layered fine- to medium-grained limestones and shales (Oil Creek-McLish-Tulip Creek Formations); to a shaley lime-mudstone (lower Bromide Formation); to a quartz rich sandy limestone (middle Bromide Formation); and to the dense limestone of the upper Bromide Formation at the top (Suhm, 1997). Overall, the stratigraphic succession shows a deepening of the Ardmore basin (and Anadarko basin, Plate 2.1) during Simpson time (Chazy-Blackriverian stages of the Ordovician) (Figure 2.3). The sandy facies, middle Bromide Formation is elongated northwest-southeast (parallel with the trend of the Ardmore basin) and has the subsurface geometry of a shallow-water, submarine delta-fan (Plate 2.4) (Suhm, 1997).
Because of erosional truncation, the Simpson Group is partially or completely absent from some wells in the southern parts of the Ardmore basin-Criner Hills arch (Cooper, 1995), and along the Muenster uplift (Plate 2.1) (Bradfield, 1957a-c). Wells located on the northeastern flanks of the Muenster uplift in Cooke and Grayson counties Texas show Morrowan (Dornick Hills Group/Atoka/grey Strawn Group) and Desmoinesian (red Strawn Group) sandstones, conglomerates, and shales resting unconformably upon lower Simpson rocks (McLish and Oil Creek Formations). The lower Simpson rocks on the northern and eastern flanks of the Muenster arch are sandy lime-mudstones and coarser-grained limestones (Suhm, 1997). In some areas along the crest of the Muenster uplift, Morrowan and/or Desmoinesian and/or Permian rocks rest directly upon crystalline igneous basement (Bradfield, 1957a-c; Ewing, 1991).

The base of the Simpson Group rocks along the Muenster arch is placed beneath a quartz-rich sandy oolitic limestone (Suhm, 1997). This sandy unit has also been named Ellenburger “A” by Whiteside and McCommons, 1991). The base of the sandy oolitic limestone is considered to be the Sauk-Tippecanoe unconformity along the Muenster arch (Suhm, 1997). Placement of the Sauk-Tippecanoe unconformity at the base of the Simpson Group is consistent with mapping of stratigraphy in the Arkoma basin. However, placement of the Sauk-Tippecanoe unconformity within the upper part of the Ellenburger Group is consistent with placement of the unconformity within the upper part of the West Spring Creek Formation (upper Arbuckle Group) in parts of the Southern Oklahoma aulacogen (Plate 2.2) (Ross and others, 1982; Finney, 1986; Derby and others, 1991; Finney, 1997).

Llano uplift

South of the Muenster uplift, on the exposed flanks of the Llano uplift (Plate 2.1), Middle Ordovician strata are very thin and discontinuous. Middle Ordovician through early Devonian (and earliest Mississippian) strata are visible only as crack-fills in older rock units, or within closed depressions and collapse structures (Kier, 1988). At most, roughly 20 m of Middle Ordovician (Whiterockian) to lower Mississippian (Kinderhookian) rocks are found across the Llano uplift (Fay, 1986a).
For this reason, middle Ordovician through Devonian foreland facies rocks for the Llano region are not discussed in subsequent paragraphs.

**Viola Group-Sylvan Shale time (Late Ordovician)**

**Black Warrior basin and Appalachian orogen**

Across much of the southern Appalachian thrust belt, Ordovician rocks younger than the Athens Shale and Silurian to lower Mississippian rocks are absent (Plates 2.1 and 2.2) (Drake and others, 1989). Farther northwest toward the Black Warrior basin, the Athens Shale grades into the lower part of the Chickamauga Group (Plates 2.1 and 2.2). An upper Middle to Upper Ordovician clastic succession (Greensport-Colvin Mountain-Sequatchie Formations) is preserved in part of the southern Appalachian frontal thrust belt (Plate 2.2) (Thomas, 1988). The Greensport-Colvin Mountain-Sequatchie Formations clastic wedge has a maximum total thickness of 130 m in Alabama, and thins westward, interfingers with, and progrades over Middle to Upper Ordovician upper Chickamauga Group (or Maysville and Nashville Groups-Trentonian through Maysvillian) limestones (Mankin, 1986; Thomas and Osborne, 1987 a,b; Thomas, 1988).

The Trentonian through Maysvillian beds of the upper Chickamauga Group correlate with the Viola Group west of the Mississippi embayment (Plates 2.1 and 2.2). Scattered, thin (< 35 m) sandstones, sandy limestones, and black shales located above Upper Ordovician limestones may represent the westernmost discontinuous extent of the Sequatchie Formation (Thomas, 1988). Black shales in the upper part of the Ordovician succession represent an eastern stratigraphic equivalent of the green- to black-colored Sylvan-Cason Shale found farther west in Arkansas, Oklahoma, and Missouri (Plate 2.2) (Mankin, 1986; Thomas and Osborne, 1987 a,b; Ethington and others, 1989). The combined maximum thickness of upper Chickamauga Group (Trentonian-Maysvillian) through Sequatchie Formation (Richmondian) rocks within the Black Warrior basin is approximately 150 m (Figure 2.2) (Thomas, 1988).

**Arkoma basin-Ozark dome (Oklahoma-Arkansas-Missouri)**

The Viola Group in the Arkoma basin includes the Kimmswick and Fernvale limestones in the subsurface of central Arkansas (Plates 2.1 and 2.2) (Van Arsdale and
Schweig, 1990; Suhm, 1997). The maximum total thickness of the Viola Group in the Arkoma basin is approximately 120 m (Ham, 1973). The Kimmswick and Fernvale Formations extend northeast into Missouri. Across the Ozark dome (Plate 2.1) of Oklahoma, Arkansas, and Missouri, the Fernvale Limestone is a recognized formation; however, the Kimmswick Formation limestone of the Arkoma basin correlates with the upper Tyner and Fite Formations of the southern flank of the Ozark dome (Suhm, 1997) (Plate 2.2).

In the eastern Arkoma basin, the lower part of the Kimmswick limestone rests conformably above the Plattin Formation limestone of the upper Simpson Group. In the subsurface, clear distinction cannot be made between the Kimmswick and Plattin Formations (Suhm, 1997). In the western Arkoma basin, and along the southern flank of the Ozark dome, the base of the Viola Group limestones rest unconformably upon Simpson Group rocks (Figure 2.3) (Suhm, 1997). A progressively greater thickness of Simpson Group rocks is truncated beneath the base of the Viola Group northward towards the southern margin of the Ozark dome in northeastern Oklahoma and northern Arkansas (Plates 2.1 and 2.2) (Suhm, 1997). Both Simpson and Viola strata are absent farther north in the central core of the Ozark dome where Arbuckle Group and Precambrian basement are exposed (Bush and others, 1977; Thomas and others, 1989).

The Viola Group is a shallowing-upward sequence of fine- to medium-grained limestones (and lesser shales) capped by skeletal grainstones (Denison, 1997). The basal rocks of the Viola Group are very dense limestones similar to the underlying Bromide (upper Simpson Group) limestones. However, in contrast to the Simpson Group carbonates, the Viola Group carbonates are carbonaceous and graptolitic. Organic carbon, graptolite, and shale content increase southwestward with increased thickness of the Viola Group towards the Ardmore basin (Plate 2.1) (Denison, 1997). The abrupt decrease in grainsize of the basal Viola Group limestones and increased organic carbon content suggest an abrupt relative sea-level rise prior to deposition of the basal Viola Group rocks (Denison, 1997).
Within the eastern Arkoma basin, and farther north and east into Missouri and Mississippi, the Cason Shale rests above the Fernvale grainstones (Plate 2.2). The Cason Shale is a dark gray, to greenish-gray laminated shale equivalent to the Sylvan Shale recognized in the Southern Oklahoma aulacogen (Plates 2.1 and 2.2) (Ham, 1973, Thomas, 1975). The abrupt change from skeletal limestones of the upper Viola Group to laminated shales suggests abrupt deepening of the Arkoma basin region prior to Cason Shale deposition. However, there is no evidence of a regional unconformity separating the upper Viola Group from the Cason Shale (Denison, 1997). Amsden (in Johnson and others, 1988) suggests that the Sylvan/Cason Shale horizon represents an abrupt increase in terrigenous sedimentation which “choked-off” limestone precipitation regardless of depth.

In the Ozark Mountains, north of the Arkoma basin, the Cason Formation includes members that are lithologic and stratigraphic equivalents of the lower part of the Hunton Group (Keel oolite and Cochrane limestone) in the Southern Oklahoma aulacogen (Plates 2.1 and 2.2). These members are late Ordovician-Silurian (Cincinnatian) in age (Ethington, and others, 1989). In the Ozark region, the oolite member of the Cason Formation rests unconformably upon upper Viola Group (Fernvale) grainstones (Ethington, and others, 1989). The basal shale member of the Cason Formation (lithostratigraphic equivalent of the Sylvan Shale) is absent across the Ozark region (Figure 2.5 and Plate 2.1). For purpose of discussion of regional stratigraphy, the oolite and limestone members of the Cason Formation in the Ozark region should be considered part of the Hunton Group rocks. The term “Cason Shale” is used for the upper Ordovician laminated dark-colored shale (Sylvan-type shale) found beneath the oolite bed across the Arkoma basin and farther southwest in the Southern Oklahoma aulacogen (Figure 2.5, and Plates 2.1 and 2.2).

**Southern Oklahoma aulacogen (Wichita and Arbuckle uplifts)**

Disconformably above the Simpson Group within the Southern Oklahoma aulacogen is a relatively thin succession of Middle to Upper Ordovician (Cincinnatian)Viola Group carbonates (Plates 2.1 and 2.2) (Ham, 1973). The Viola varies (upsection) from laminated siliceous carbonate, to burrowed skeletal mudstone,
to pelmatozoan calcarenite (Ham, 1973). Proportion of calcarenite increases northeastward in the Arbuckle Mountains, whereas measured thickness decreases from a maximum of 274 m to 122 m. Graptolites and trilobites are dominant in the lower parts of the Viola Group micrites, but pelmatozoans, brachiopods, bryozoans, and mollusks are concentrated in the calcarenites of the upper Viola Group.

Across the Hunton arch of southeastern Oklahoma, the Viola Group is separated into the Viola Springs Formation skeletal micrite capped by the Welling Formation skeletal calcarenite (Denison, 1997). The Viola Springs Formation makes up 80 percent of the total thickness of the Viola Group across the Hunton arch (Denison, 1997). Farther southwest towards the Arbuckle anticline (Plate 2.2), the proportional thickness of the Viola Springs Formation increases to 95 percent (Denison, 1997).

Across the Arbuckle anticline and Hunton arch (Plate 2.1), unconformity surfaces separate the basal part of the Viola Springs Formation (Seminole Sandstone Member, Suhm, 1997) from underlying Simpson Group rocks and the overlying remainder of the Viola Springs Formation (Ireland, 1965; Amsden, 1980; Amsden and Sweet, 1983; Carlson and Newell, 1997). This basal dense limestone is similar to Simpson Group type limestones and has traditionally been considered the uppermost Corbin Ranch submember of the Bromide Formation (uppermost Simpson Group) (Finney, 1988; Suhm, 1997). However, detailed study of fossil ostracodes, conodonts, and other fauna indicate that the Corbin Ranch limestone is Trentonian in age (correlative with Viola Group) rather than Blackriverian (such as the upper Simpson Group) (Harris, 1957; Amsden and Sweet, 1983; Sweet, 1992). There is also a regional discordance between the Corbin Ranch and underlying Bromide Formation (uppermost Simpson Group) (Amsden and Sweet, 1983).

Along the northern flank of the Arbuckle and Tishomingo-Belton anticlines (Plate 2.1), the “Corbin Ranch” basal limestone of the Viola Springs Formation grades into a coarse-grained dolostone (Suhm, 1997). Still farther north, in central Oklahoma, the dolostone grades into a quartz-rich, dolomitic, limey sandstone. This limey sandstone called the Seminole Sandstone (Figure 2.3) is lithologically similar to
sandy formations in the underlying Simpson Group (Suhm, 1997). In contrast to the northeast-southwest elongation of most of the Simpson Group sandstones, the northwest-southeast orientation of the regional isopach map of the Seminole Sandstone suggests a different transport direction (Suhm, 1997). The different orientation of the sandstone body, combined with the observation that the Seminole Sandstone interfingers with the Corbin Ranch limestone (clearly of Viola Group age), indicates that the Seminole Sandstone belongs to the basal part of the Viola Group rather than the upper part of the Simpson Group (Suhm, 1997). Northeast of central Oklahoma, the Seminole Sandstone facies of the basal Viola Group grades into sandy limestones of the upper Tyner and Fite Formations (Plate 2.2) (Suhm, 1997).

Resting above the skeletal calcarenite at the top of the Viola Group is the late Ordovician (Cincinnatian) Sylvan Shale (Ham, 1973). The Sylvan is a laminated, dark-gray to greenish-gray shale that contains graptolites and a "rich chitinozoan fauna," but lacks significant amounts of other fossils (Ham, 1973; Thomas, 1977). Although thin and laterally discontinuous shale layers are found in the Simpson Group strata, the Sylvan Shale is the oldest laterally continuous shale formation across the foreland of the Ouachita structural salient (northeast of the Muenster arch). The Sylvan Shale correlates with the Cason Shale of the eastern Arkoma basin, and the Sequatchie Formation within the Black Warrior basin and along the southwestern edge of the Appalachian thrust front (Plate 2.2) (Thomas, 1977; Thomas and Osborne, 1986a). Thomas (1977) interprets the Sylvan/Cason shale as an intertongue of the deep-water, shale-dominated, off-shelf Ouachita facies rocks.

Because of the stark contrast in lithologies between the Viola Group limestones and the Sylvan Shale, a disconformity is traditionally placed at the base of the Sylvan Shale (Ham, 1973). The lower part of the Sylvan Shale contains graptolites and is carbonaceous, both of which imply deposition in deeper water (below wave base) (Finney, 1997). Also, the Viola Group/Sylvan Shale contact corresponds an early Cincinnatian regional eustatic highstand (Finney, 1997). However, across the Southern Oklahoma aulacogen and Arkoma basin (Plate 2.1) the contact between the top of the Viola Group and the base of the Sylvan appears
conformable (Denison, 1997). An alternative interpretation suggests that the abrupt change in lithology is the result of an abrupt increase in terrigenous sediment from the north, rather than solely an abrupt rise in relative sea-level prior to deposition of the Sylvan Shale (Amsden, in Johnson and others, 1988).

Like all older Ordovician sedimentary rocks in the Southern Oklahoma aulacogen (Plate 2.1), the Sylvan Shale thins eastward away from the Ardmore basin. The Sylvan Shale thins from a maximum thickness of more than 100 m in the deep Ardmore basin to 53 m on the Hunton arch (Ham, 1973, Denison, 1997) (Plate 2.1).

**Ardmore basin and Muenster arch**

The Viola Group is thickest in the subsurface of the deep Ardmore basin and along the southern and eastern margins of the subsurface Tishomingo-Belton anticline (Plate 2.1). Proportional content of organic carbon, graptolites, and fine-grained laminated lime-mudstones is also greatest in the deep Ardmore basin (Denison, 1997; Finney, 1997; Suhm, 1997). The base of the Viola Group corresponds to a regional eustatic lowstand (and erosional unconformity surfaces) farther northeast in the North American craton (Finney, 1997). However, lithology of the basal Viola Group in the deep Ardmore basin and southern flank of the Arbuckle and Tishomingo-Belton anticline (Plate 2.1) indicates deposition in a deeper, subtidal environment (Finney, 1997). Lithology and fossil fauna suggest that the basal Viola Group limestones along the Arbuckle and Tishomingo-Belton anticlines and in deeper parts of the Ardmore basin were deposited in a water depth range of 180 to 450+ meters (Finney, 1988). The coincidence of a local relative sea-level highstand in the Ardmore basin and other parts of the Southern Oklahoma aulacogen (Plate 2.1) with a regional relative sea-level lowstand indicates that Southern Oklahoma aulacogen was isostatically subsiding faster than the surrounding platform during the Late Ordovician (Mohawkian) eustatic decrease in sea level (Finney, 1986, 1988, 1997).

Because of increased erosional truncation, progressively younger Late Ordovician rocks of the Viola Group and overlying Sylvan Shale are sequentially absent south of the northwest-southeast-trending Criner Hills uplift fault trend of southern Oklahoma (Plate 2.1) (Cooper, 1995). The Criner Hills uplift fault trend
separates the deep Ardmore basin from the shallower Marietta basin to the south (Plate 2.1). South of the Criner Hills uplift fault trend, in deeper parts of the Marietta basin and around the southeastern flank of the Muenster arch (Plate 2.1), Early Pennsylvanian (Morrowan) clastic rocks of the Dornick Hills Group rest unconformably upon Middle Ordovician Simpson Group limestones, and sandy/shaley-limestones (Bradfield, 1957a-c; Cooper, 1995). Along the crest of the Munster arch, Pennsylvanian (Desmoinesian) to Permian rocks rest unconformably upon either Precambrian igneous basement or Cambrian-Early Ordovician shelf carbonates of the Arbuckle Group (Ewing, 1991; Cooper, 1995).

**Hunton Group through Welden Formation time; slow sedimentation and numerous relative sea-level rises and falls (numerous unconformities) (latest Ordovician to early Mississippian)**

**Black Warrior basin and southern Appalachian orogen**

An unconformity separates a thin sequence of Silurian rocks from a thicker succession of Ordovician rocks (Thomas and Osborne, 1987a,b; Thomas, 1988). In the northern and southern parts of the Black Warrior basin, grey shales, sandy limestones, and sandstones of the upper Chickamauga Group and Sequatchie Formation are truncated beneath the basal Silurian unconformity (Figure 2.2). The Silurian succession of the Black Warrior basin is predominantly limestones and dolostones, and ranges from 20 to 125 m thick (Figure 2.2). The Silurian carbonate rocks of the Black Warrior basin correlate with the Silurian part of the Hunton Group west of the Mississippi embayment in Arkansas, Oklahoma, and northern Texas (Plates 2.1 and 2.2).

In the Black Warrior basin, a thicker succession of Middle to Upper Devonian cherty limestones and novaculites rests unconformably above Silurian carbonate rocks (Figure 2.2) (Thomas and Osborne, 1986a,b). The Devonian succession is thickest in the center of the Black Warrior basin (> 300 m) and is completely truncated by an unconformity in the northeastern parts of the basin (Figure 2.2). The Devonian sequence of shallow-water to deep-water transitional cherty carbonate rocks of the
Black Warrior basin correlates with the Penters Chert of the Arkoma basin and Ozark dome regions (Figure 2.2, and Plates 2.1 and 2.2).

A thin (< 5 m) and discontinuous upper Devonian black shale covers the middle to upper Devonian cherty carbonate sequence across the eastern part of the Black Warrior basin (Figure 2.2 and Plates 2.1 and 2.2). In the northern Black Warrior basin, toward the Nashville dome, the unconformity at the base of the black shale truncates Upper Ordovician through Devonian rocks (Figure 2.2). In the northern Black Warrior basin, the thin upper Devonian black shale is referred to as Chattanooga Shale (Figure 2.2 and Plate 2.2). A thicker, and darker colored, Chattanooga Shale is found farther west in surface exposure along the southern flank of the Ozark dome and in the subsurface of the Arkoma basin (Figure 2.2 and Plate 2.2).

The thin Chattanooga Shale and a very thin lower Mississippian green shale (Maury Shale) in the eastern part of the Black Warrior basin are covered by lower Mississippian limestones and cherts (Figure 2.2). These lower Mississippian limestones and cherts thicken northward to more than 150 m toward the Nashville dome (Figure 2.2). In the northern and eastern Black Warrior basin and southern Appalachian orogen, the lower Mississippian succession of cherty limestones is classified as the Fort Payne Chert and Tuscumbia Limestone (Figure 2.2 and Plate 2.2). The Fort Payne Chert correlates with the lower Mississippian Boone Formation farther west in the Ozark dome and in the subsurface of the eastern Arkoma basin (Figure 2.2 and Plates 2.1 and 2.2). Rocks of Tuscumbia age are unrecorded in the Ozark dome, Arkoma basin, and Hunton arch areas of Missouri, Arkansas, and northeastern Oklahoma; however, the lower part of the Caney Shale in the Arbuckle anticline and Ardmore basin areas of Oklahoma is the chronostratigraphic equivalent (Plates 2.1 and 2.2).

Arkoma basin-Ozark dome (Oklahoma, Arkansas, and Missouri)

The Hunton Group in the Arkoma basin (Plate 2.1) of Arkansas is subdivided into four formations. Listed from oldest to youngest, the formations are: Brassfield, St. Clair, Lafferty, and Penters (Plate 2.2) (Van Arsdale and Schweig, 1990). Farther
north in the Ozark region, the Brassfield Formation of the lower Hunton Group corresponds stratigraphically with limestone facies of the Cason Formation (Plate 2.2) (Ethington and others, 1989).

A thin (< 5 m thick) oolite bed at the base of the Hunton Group is a regional enigma. Except in the Ozark dome region where the lower shale bed of the Cason Formation is missing and replaced by an unconformity surface, the oolite bed overlies laminated dark-colored shales (Cason/Sylvan Formations) with no visible evidence of unconformity (Denison, 1997). The enigma of the oolite bed, called the Keel oolite in the Arbuckle uplift (Plate 2.1) of Oklahoma, is as follows: How can a clear-water, high-wave-energy formation such as an oolite (Newell and others, 1960) rest above a fine-grained shale without incorporation of shale detritus within the oolite? A possible explanation requires subaerial lithification of the Cason/Sylvan Shale prior to deposition of the oolite bed; however, region-wide, there is no direct visible evidence of an erosional surface (Amsden, in Amsden and Barrick, 1986; Denison, 1997).

On the southern flank of the Ozark dome (Plate 2.1), an interpreted unconformity separates the top of the oolite member of the Cason Formation from an upper limestone member (Suhm, 1997). On the basis of biostratigraphy, approximately 8 m.y. of time is unrecorded at the unconformity. A correlative unconformity with the same time span separates the oolitic Keel Formation from the Cochrane Formation limestone in the Arbuckle uplift of southern Oklahoma (Plate 2.1) (Ethington and others, 1989).

Deposited conformably upon the Cason/Brassfield Formations of the southern Ozark dome and Arkoma basin are the limestones of the St. Clair and Lafferty Formations (Plates 2.1 and 2.2). Biostratigraphy indicates that the St. Clair and Lafferty Formations comprise a complete chronological sequence of rocks extending to the end of the Silurian (Ethington and others, 1989). The total maximum thickness of Hunton Group strata throughout the Ozark dome-Arkoma basin region is approximately 180 m (Ham, 1973). The Hunton Group thins north and west and is truncated in the center of the Ozark dome and along the crest of the Nemaha ridge in northern Oklahoma (Plate 2.1) (Amsden, in Ham, 1973, p. 39-53).
An unconformity (20 m.y. time span) at the top of the Silurian separates the top of the Lafferty Formation limestone from the Penters Chert (Plate 2.2) (Ethington and others, 1989). The Penters Chert contains both chert beds and dolomitic limestones, and has a maximum exposed thickness of 79 m along the southern flank of the Ozark dome in northern Arkansas (Plate 2.1) (Bush and others, 1977). The Penters Chert thins northward toward the center of the Ozark dome. Because of erosional truncation, correlative middle Devonian formations are not found within the Arbuckle uplift of Oklahoma (Plates 2.1 and 2.2) (Ethington, and others, 1989). However, a complete succession of Devonian-age deep-water facies cherty formations (Arkansas Novaculite) are found farther southeast within the allochthonous terrain of the Ouachita Mountains (Plates 2.1 and 2.3). The combination of chert and dolomitic limestones contained within the Penters Chert suggests that the Penters Chert is a transitional facies intertongue of the Devonian-age deep-water facies Arkansas Novaculite (Plate 2.3).

In the Ozark region, unconformably (10 Ma of unrecorded time) above the Penters Chert is the thin Clifty Formation limestone (Plate 2.2) (Ethington and others, 1989). An unconformity of lesser time span separates the Clifty Formation limestone from clean sandstone at the base of a succession of laminated black shale (Bush and others, 1977; Ethington and others, 1989). The clean sandstone and black shale are called respectively the Sylamore Sandstone and Chattanooga Shale. The Sylamore Sandstone is a basal sandstone member of the Chattanooga Shale that is absent in deeper parts of the subsurface Arkoma basin (Plate 2.1) (Van Arsdale and Schweig, 1990).

The Sylamore Sandstone-Chattanooga Shale strata thicken southward from the southern part of the Ozark dome toward the Arkoma basin in the subsurface. Thickness increases southward from 21 m (surface measurement where upper beds are truncated by a pre-Boone unconformity) (Bush and others, 1977) to more than 38 m in the subsurface in the eastern Arkoma basin (Van Arsdale, and Schweig, 1990). The laminated dark-colored Chattanooga Shale extends eastward discontinuously into the subsurface Black Warrior basin and is absent west of Alabama (Figure 2.3) (Thomas,
Southeast of the Ozark region, toward eastern Arkansas and Missouri, the Chattanooga Shale is truncated by an unconformity at the base of the Early Mississippian Boone Formation (Plate 2.1) (Bush and others, 1977). Here, the Boone Formation rests unconformably upon Potters Chert (Bush and others, 1977). The southeastward increase in chert beds suggests increased water depth (basin subsidence) towards the Arkansas-Mississippi border region in the Late Devonian.

Along the southern flank of the Ozark dome and in the Arkoma basin (Plate 2.1), an unconformity separates the top of Chattanooga Shale from the base of the Boone Formation (Ethington and others, 1989). In the Ozark region, lower Mississippian (latest Kinderhookian) Boone Formation beds rest upon uppermost Devonian) beds of the Chattanooga Shale (< 10 m.y. of unrecorded time) (Ethington and others, 1989).

Along the southern margin of the Ozark dome in Arkansas, the Boone Formation is latest Kinderhookian to early Meramecian in age, and the dominant lithologies comprising the Boone Formation are limestone, chert, and oolite (Bush and others, 1977; Ethington and others, 1989). Within the area comprising the Ozark dome and Arkoma basin, the Boone Formation is thickest along the South Ozark arch (Plate 2.1), in contrast to all older formations that thicken southeastward and southwestward toward the Arkoma basin (Figure 2.5) (Bush, and others, 1977; from Suhm, 1997). To the east and west of the South Ozark arch, upper beds of the Boone are progressively truncated by an erosional unconformity (Figure 2.5). This erosional truncation indicates that the eastern and western flanks of the South Ozark arch (Plate 2.1) underwent relative uplift (with respect to the center of the arch) prior to deposition of post-Boone strata.

Along the eastern flank of the South Ozark arch (Plate 2.1), in eastern Arkansas, the upper Mississippian beds of the upper Boone Formation are unconformably overlain by the Moorefield Formation, and the finer grained Ruddell Shale rests conformably upon the Moorefield (Figure 2.5 and Plate 2.2). Along the western flank of the South Ozark arch (Plate 2.1), in western Arkansas and northeastern Oklahoma, the Moorefield Formation rests unconformably upon the
Boone Formation and the Ruddel Shale is absent (Figure 2.5). Along the crest of the South Ozark arch (Plate 2.1), the upper Mississippian (lower Chesterian) Hindsville Limestone member of the Batesville Formation rests unconformably on of the lower Mississippian (lower Meramecian) Boone Formation (Figure 2.5 and Plate 2.2) (Bush and others, 1977; Ethington and others, 1989).

The unconformity at the base of the Boone Formation is laterally continuous and important because it separates formations folded over the South Ozark arch, below the unconformity, from relatively horizontal strata above the unconformity (Figure 2.5). Southwest of the Ozark dome, in northeast Oklahoma, the unconformity at the base of the Chattanooga Shale separates South Ozark arch strata (syncline) from an overlying broad anticline (Figure 2.5). Farther southwest, throughout much of Oklahoma, the unconformity at the base of the Woodford Formation/Chattanooga Shale is a region-wide unconformity (Figure 2.5) (Ham, 1973, Cooper, 1995).

The unconformity at the base of the Batesville Formation is also regionally significant. All formations below the latest Meramacian-earliest Chesterian (pre-Batesville) unconformity surface show some degree of discordance in comparison to post-Batesville formations (Figure 2.5). All the formations above the basal Batesville Formation unconformity are parallel to the unconformity. Therefore, the pre-Batesville unconformity marks the beginning of rock strata that conform to the late Paleozoic Ouachita orogenic subsidence history.

**Southern Oklahoma aulacogen (Wichita and Arbuckle uplifts)**

The basal oolitic Keel Formation of the Hunton Group rests directly upon the fine-grained, dark-colored Sylvan Shale. Although the fact that no fine-grained shale is included in the Keel oolite (suggesting a hardground at the top of Sylvan Shale prior to Keel deposition), the traditional unconformity placed at the base of Hunton Group strata (Keel oolite) is questioned (see argument in previous paragraphs). In addition to this questionable basal disconformity, there are several other clearly defined unconformities within the Hunton Group. The Hunton Group is the thinnest succession of carbonates that crop out in the Arbuckle Mountains (Plate 2.1). Ham (1973) suggests that almost half of the Hunton Group rocks were removed by erosion.
The basal layer of the Hunton Group is the thin uppermost Ordovician (Cincinnatian) Keel oolite which is exposed throughout the entire Arbuckle Mountains outcrop belt (Ham, 1973, Ethington and others, 1989). The Keel oolite correlates with oolite beds in the Cason/Brassfield Formations farther northeast in Arkansas (Plate 2.2) (Bush and others, 1977; Ethington and others, 1989; Van Arsdale and Schweig, 1990).

Although thickness of the Hunton Group is relatively uniform from southwest to northeast across the Arbuckle uplift, much of the lower succession is missing because of erosional truncation across the Hunton arch (Plate 2.1) (Ham, 1973). Across the entire Arbuckle uplift, the basal formation of a Silurian sequence of rocks rests unconformably above the Keel oolite (10 m.y. of unrecorded time) (Ham, 1973; Ethington and others, 1989). This basal formation is the lower Silurian (Alexandrian-early Niagaran) glauconitic, pelmatozoan calcarenite Cochrane Formation (Plate 2.2) (Ham, 1973; Ethington and others, 1989). Conformably above is the lower Silurian (Niagaran) pelmatozoan calcarenitic and calcilutitic Clarita Formation (Plate 2.2) (Ham, 1973; Ethington and others, 1989). The Keel, Cochrane, and Clarita Formations are included in the Chimneyhill Subgroup (Amsden, in Ham, 1973). The Chimneyhill Subgroup correlates with Cason/Brassfield through St. Clair Formations farther northeast in Arkansas (Plate 2.2) (Ethington and others, 1989; Van Arsdale and Schweig, 1990).

In southwestern parts of the Arbuckle uplift, along the northern margin of the Ardmore basin (Plate 2.1), the argillaceous skeletal calcilutitic and calcarenitic upper Silurian (Niagaran-Cayugan) Henryhouse Formation rests upon the top of the Chimneyhill Subgroup (Plate 2.1) (Ham, 1973, Ethington and others, 1989). The base of the Henryhouse is either conformable (Ethington and others, 1989), or unconformable with a small time gap (Amsden, in Ham, 1973). Farther northeast, across the Hunton arch (Plate 2.1), the entire Henryhouse Formation, and upper Chimneyhill Subgroup beds are erosionally truncated and covered with overlying Devonian Hunton Group rocks (Ham, 1973). In Arkansas, the Lafferty Formation correlates with the Henryhouse Formation (Plate 2.2) (Ethington and others, 1989).
Across the entire Arbuckle uplift (Plate 2.1), lower Devonian (Ulsterian) argillaceous skeletal calcilutitites (Haragan-Bois d'Arc Formations) unconformably rest upon older Silurian Hunton Group rocks (Plate 2.2) (Ham, 1973). The Haragan-Bois d’Arc Formations are thickest across the Hunton arch (Plate 2.1) where they rest unconformably upon the Chimneyhill Subgroup (Ham, 1973). Farther southwest, in the Arbuckle anticline (Plate 2.1), the upper part of the Haragan-Bois d’Arc Formations is truncated by an unconformity at the base of the Woodford Formation (upper Devonian) (Plate 2.2).

The youngest Hunton Group rock is found in the Hunton arch (Plate 2.1) of Oklahoma (Ham, 1973). Here, the lower Devonian (Ulsterian) skeletal calcarenitic Frisco Formation rests unconformably (< 10 Ma time gap) upon the Haragan-Bois d’Arc Formations (Plate 2.2) (Ham, 1973, Ethington and others, 1989). Southwest of the Hunton arch, the Frisco Formation is absent, likely as a result of complete erosional removal prior to Woodford Formation deposition (Ham, 1973). The southwestward erosional truncation of the Frisco Formation, and underlying upper part of the Haragan-Bois d’Arc Formations, indicates a relative subsidence of the Hunton arch (in comparison to the Ardmore basin (Plate 2.1), prior to Woodford Formation deposition. Because of a pre-Penters Chert erosional unconformity, the Haragan-Bois d’Arc and Frisco Formations have no correlative rocks farther northeast in Arkansas (Plate 2.2). The maximum measured thickness of the Hunton Group within the Arbuckle Mountains is 108 m; however, toward the northwest in the subsurface, the Hunton Group increases in thickness to 243 m (Amsden and Rowland, 1967).

Unconformably above the Devonian beds of the Hunton Group in the Arbuckle uplift (above Haragan-Bois d'Arc Formation towards the southwest, and above Frisco Formation towards the northeast) is a distinctly different rock unit called the Woodford Formation. The Woodford is an upper Devonian (Senecan) to lower Mississippian (Kinderhookian) dark, fissile, carbonaceous shale (Plate 2.2) (Ham, 1973; Ethington and others, 1989). Included within the Woodford Formation are vitreous bedded chert and siliceous shale, and the upper beds contain nodules and plates of phosphate (Ham, 1973). Chert content increases towards the east, especially
in the subsurface (Gatewood and Fay, 1991). Basal beds of the Woodford contain "[silicified] spores of forest trees and silicified fragments (Callixylon)" (Ham, 1973). On the basis of conodont biostratigraphy, all but the uppermost beds (< 3 m) of the Woodford Formation is late Devonian (Senecan) (Hass and Huddle, 1965). The Woodford Formation correlates with Sylamore Sandstone-Chattanooga Shale of Arkansas (Plate 2.2) (Bush, and others, 1977; Ethington and others, 1989).

Within the Arbuckle uplift (Plate 2.1), the Woodford Formation increases in thickness eastward towards the Ouachita thrust belt in southeastern Oklahoma. Thickness of the Woodford increases from 106 to 122 m in the western part of the Arbuckle uplift, to 170 m in the eastern part if the Arbuckle uplift outcrop belt (Ham, 1973). In contrast to pre-Hunton strata that increase in thickness toward the Ardmore basin, the Woodford increases in thickness toward the Ouachita structural salient (Plate 2.1). This switch indicates that relative subsidence increased east of the Arbuckle uplift during the Late Devonian. The lack of significant thickness of Silurian-Devonian rocks (in comparison to Cambrian through Ordovician sequence) beneath the basal Woodford unconformity suggests a dramatic decrease in sedimentation rate during the Silurian and Devonian. The decrease in sedimentation rate indicates that the Ardmore basin and Arbuckle uplift regions of the Southern Oklahoma aulagogen (Plate 2.1) reached isostatic equilibrium prior to Woodford deposition. Therefore, throughout much of Oklahoma, the base of the Woodford Formation (Late Devonian) marks the onset of a basin subsidence history that post-dates isostatic adjustment related to Late Cambrian rifting and platform carbonate deposition. To the northeast of the Arbuckle uplift in northeast Oklahoma and Arkansas, along the crest of the South Ozark arch, the Woodford-Chattanooga Shale is truncated beneath the basal Boone Formation unconformity (Figure 2.5 and Plate 2.1) (Bush and others, 1977).

An unconformity above the Woodford/Chattanooga formations, known from surface exposure (except in southwestern Arbuckle Mountains, Ham, 1973) and known from numerous wells throughout Oklahoma and Arkansas (Arbenz, 1989c,d), separates Woodford/Chattanooga carbonaceous shales from thin lower and middle
Mississippian limestones (Sycamore-Weldon-Boone Formations). The Sycamore Formation (Osagean-Meramacean) is recognized only on the surface of the southern and western Arbuckle Mountains where it reaches a maximum thickness of 106 m (Plate 2.2) (Ham, 1973; Ethington and others, 1989). The Sycamore is interbedded sparsely fossiliferous calcisiltite and dark-gray shale (Ham, 1973). A much thinner (1.5 m) equivalent, the Welden Limestone (Osagean), is exposed in northeastern parts of the Arbuckle Mountains (Plate 2.2) (Ham, 1973; Ethington and others, 1989).

**Ardmore basin and Muenster arch**

In general, all the Upper Ordovician (Cincinnatian) through lower Mississippian (Osagean) rocks thicken towards the center of the Ardmore basin (Ham, 1973) (Plate 2.1). Although stratigraphy of the deepest part of the southeastern Ardmore basin, and its southeastward projection beneath strata of both the Ouachita orogen and the Cretaceous Gulf Coastal Plain (Plate 2.2) is conjectural, well data for the northwestern part of the Ardmore basin indicates that the Hunton strata are locally truncated beneath a laterally continuous unconformity at the base of the Woodford Formation (Cooper, 1995). In the northwestern Ardmore basin (Plate 2.1), both the base and top of the Sycamore Formation appears conformable and gradational with the top of the underlying Woodford Formation, and the base of the overlying Caney Shale, respectively (Ham, 1973; Cooper, 1995). The limestone content of the Sycamore Formation decreases and shale content increases toward the center of the Ardmore basin, and the top of the Sycamore Formation is marked by the uppermost appearance of limestone beds (Cooper, 1995). Farther south, toward the Muenster arch, Hunton Group-Sycamore Formation rocks (and overlying Woodford Formation) are truncated beneath an unconformity at the base of the Morrowan Dornick Hills Group clastic rocks (Plate 2.2) (Cooper, 1995).

**Llano uplift**

Upper Ordovician through upper Devonian rocks are sparse; however, more extensive lower Mississippian and younger rocks crop out in the Llano uplift (Plate 2.1). A thin and discontinuous stratigraphic correlative of the Sycamore Formation,
the Chappel Limestone, crops out in the Llano uplift and has been penetrated by wells along the Ouachita thrust front of central Texas (Plate 2.2) (Cheney, 1929; Kier, 1988). The Chappel Limestone varies (upsection) from a basal quartz sandstone to a fine-grained, dense, "crinoidal biosparite and biomicrite" with local glauconite deposits (Kier, 1988). Macroscopic and microscopic fossils are abundant and include trilobites, brachiopods, ostracodes, algae, forams, and conodonts (Kier, 1988). Conodonts are generally Mississippian in age; however, Devonian ("possibly reworked") types are also found (Kier, 1988). The Chappel Limestone reaches a maximum thickness of 15 m, and rests unconformably within depressions on underlying Ordovician Ellenburger Group carbonates (Kier, 1988).

**B) Deep-water facies pre-orogenic strata of the Ouachita orogen**

**Cambrian-early Mississippian (Overview)**

In western Mississippi, a northwest-striking, thrust-faulted subcrop belt of partly carbonaceous clastic rocks (including coal beds) marks the southwestern boundary of the Black Warrior basin (Plate 2.1) (Thomas, 1989; Thomas and others, 1989). Because the faulted succession of clastic rocks is lithologically similar to the Pennsylvanian mudstones and sandstones that are exposed along strike farther east in the northeastern “frontal” part of the Ouachita thrust belt in Arkansas (Thomas, 1989) it is considered the southeastward subsurface extension of the Ouachita “frontal” fold-thrust belt strata (Plate 2.1) (Thomas and others, 1989).

Southwest of the subsurface Ouachita frontal fold-thrust belt in western Mississippi is a northwest-striking subcrop belt of dark-colored slates (including quartz veins) called the Western Mississippi slate belt (Thomas, 1972b). Rocks within the Western Mississippi slate belt are mostly dark-colored pelitic rocks including dark-colored siliceous mudstone, dark-colored chert, and sandstone (Thomas, 1989). In a very broad sense, the stratigraphic succession within the Western Mississippi slate belt is lithogically similar to the Ordovician-Mississippian deep-marine succession that crops out in the “central” uplifted core of the eastern Ouachita Mountains (Benton uplift) (Plate 2.1) (Thomas, 1989). Therefore, the Western Mississippi slate belt is
considered the southeastward subsurface extension of the Ouachita “central” uplift (Thomas and others, 1989)

Predominance of mudstones in surface exposure within the Ouachita Mountains and lack of biostratigraphy within rock units drilled in the subsurface Western Mississippi slate belt makes direct correlation impossible. Siliceous pelitic rocks of the Western Mississippi slate belt are similar to units within the Mississippian mud-rich turbidite succession of the east-central Ouachita Mountains (lower Stanley Group) (Thomas, 1989). Cherts of the Western Mississippi slate belt are similar to units found in the Ordovician, Devonian, and Mississippian strata of the east-central Ouachita Mountains (Thomas, 1989). An important contrast between the stratigraphic successions of the east-central Ouachita Mountains and the Western Mississippi slate belt is the absence of a thick chert unit comparable to Arkansas Novaculite (Devonian-lower Mississippian) within the latter (Thomas, 1989).

Although within the southern part of the Ouachita Mountains the entire Devonian-early Mississippian Arkansas Novaculite succession (all three members) is preserved, in the northern part of the Ouachita Mountains (Plate 2.1) the upper and lower members are absent and replaced by unconformities at the base and top of the middle member (Lowe, 1989). Farther north in the Arkoma basin (Plate 2.1), the relatively thin Pinters Chert (latest Early to Middle Devonian) rests unconformably between underlying Silurian Hunton Group (cherty limestone) and overlying lower Mississippian Boone Formation (limey chert) (Plate 2.2). Farther east in the Black Warrior basin (Plate 2.1) even less of the Silurian-Devonian is recorded in the stratigraphy (Plate 2.2); however, a thick Devonian chert unit pinches out eastward across the basin (Thomas, 1988). The lack of a thick Arkansas-Novaculite-type chert unit within the Western Mississippi slate belt could indicate that much of the Devonian (and possibly part of the Silurian) is unrecorded. Furthermore, the lack of a thick Arkansas-Novaculite-type chert unit suggests that the stratigraphic succession within the Western Mississippi slate belt is intermediate (transitional) between the shallow-marine succession of the Black Warrior and Arkoma basins on the northeast and the deep-marine succession of the east-central Ouachita Mountains (Plate 2.1)
Because of the lack of detailed information about the stratigraphy of the Western Mississippi slate belt, in further paragraphs the description of the deep-marine pre-orogenic (pre-early Mississippian) strata is limited primarily to the Ouachita Mountains outcrop area.

**Late Cambrian through Middle Ordovician (late Ibexian) of the Ouachita Mountains**

On the basis of biostratigraphy, the deep-water facies equivalent of the Upper Cambrian transgressive sandstones, and overlying platform limestones, and dolostones of the foreland (Timbered Hills Group through middle Arbuckle Group) is the Collier Formation (Plate 2.3) (Ethington and others, 1989). The Collier Formation crops out in the Broken Bow and Benton uplifts (Ouachita Mountains) of eastern Oklahoma and Arkansas (Plate 2.1). Because the oldest Collier beds are located above thrust faults within the Ouachita Mountains thrust belt, the maximum age and the lithology of Collier (and pre-Collier) deep-water facies sedimentary rocks are not known. The top of the Collier Formation correlates chronologically with the lower part of the Cool Creek Formation (Ibexian) (Plate 2.3). The Collier Formation contains both limestones and shales, and the proportion of limestone beds decreases upsection (Gatewood and Fay, 1991). Fossils are virtually unrecognized in the Collier Formation; however, a Late Cambrian (Franconian) trilobite and protoconodont are reported in exposures near Jessieville, Arkansas (Hart and others, 1987). Early Ordovician conodonts are also reported (Repetski and Ethington, 1977).

The shales of the upper Collier Formation grade conformably upward into the Lower Ordovician (Ibexian) Crystal Mountain Sandstone (Plate 2.2). The Crystal Mountain Sandstone is predominantly a light-gray, well-sorted, fine- to medium-grained quartz arenite. The sandstone is slightly arkosic, and there is a higher percentage of quartz clasts than lithic fragments (Arbenz, 1989; Lowe, 1989; Morris, 1974). The Crystal Mountain Sandstone contains no reported fossils, and the age is biostratigraphically constrained by rare conodonts in the upper beds of the underlying Collier Formation (Ethington and others, 1989). The Crystal Mountain Sandstone is interpreted as a "proximal turbidite" (Morris, 1974). The shallow-water correlative
unit of the Crystal Mountain Sandstone west of the Ouachita Mountains in Oklahoma is the Cool Creek Formation (Plate 2.3) (Ethington and others, 1989).

Conformably and gradationally above the Crystal Mountain Sandstone is the Mazarn Shale (Plate 2.3). Where the complete Ordovician deep-water facies succession crops out, the Mazarn Shale is recognized in the Broken Bow and Benton uplifts (Plate 2.1) of eastern Oklahoma and Arkansas. A relatively thick quartz sandstone (Blakely Sandstone) separates the dark-colored Mazarn Shale from a younger, thick, dark-colored shale (Womble Shale).

The Mazarn Shale includes many thin layers of quartzose and calcareous sandstones, siltstones, and shales which are considered to represent distal turbidites (Morris, 1974). Similar to the underlying Crystal Mountain, the sandstones of the Mazarn Shale are slightly arkosic, and include volcanic debris, suggesting derivation from exposed granites and rhyolites in the hinterland (northern source area) (Arbenz, 1989; Lowe, 1989). In contrast to the Crystal Mountain Sandstone, the sandstones in the Mazarn Shale are enriched in lithic fragments (50% lithic:50% quartz). Biostratigraphy of conodonts and rare graptolites constrain the age of the base of the Mazarn, and the top of the Mazarn Shale correlates with the top of the West Spring Creek (uppermost Arbuckle Group) shallow-water platform limestone in Oklahoma (Ethington and others, 1989)(Plate 2.2).

The maximum estimated total thickness of the Collier through Mazarn succession in the Ouachita Mountains of Arkansas is greater than 1475 m (Stone and others, 1973). Thicknesses of strata in the Ouachita Mountains are difficult to determine because of structural complexity. The Collier Formation-Mazarn Shale (and Blakely Sandstone) succession thins westward in the Ouachita Mountains, and neither crop out, nor are penetrated by wells west of the Broken Bow uplift (Plate 2.1) (Morris, 1974; Arbenz, 1989a,b).

**Middle to Late Ordovician (late Ibexian to Mohawkian) of the Ouachita Mountains**

Stratigraphically above the Mazarn Shale, in eastern parts of the Ouachita Mountains, are the Middle to Upper Ordovician (Whiterockian-Mohawkian) Blakely
Sandstone and Womble Shale (Plate 2.3) (Ethington and others, 1989). The Blakely Sandstone (Whiterockian) is a mixture of "massive to thin-bedded, well-sorted, fine-grained quartz arenite and varicolored silts and shales" (Morris, 1974). The Blakely Sandstone contains limestone lenses and fossils (including graptolites). Age of the upper part of the formation is well constrained by biostratigraphy. Because fossils are sparse in the lower part of the Blakely Sandstone, the lower boundary with the Mazarn Shale is not well constrained (Ethington and others, 1989). This unit is thickest and best recognized in the Benton uplift of Arkansas. Included within the Blakely Sandstone are several layers of "rubble-bedding" that contain large, rounded, meta-arkosic boulders (Morris, 1974). The quartz to lithic ratio of the Blakely Sandstone is similar to that of the older Crystal Mountain Sandstone (Lowe, 1989).

The Blakely Sandstone is interpreted as a rapidly deposited "proximal turbidite" or submarine slump deposit with a northern source area (Morris, 1974). As does the underlying Mazarn Shale, the Blakely Sandstone thins westward from its maximum estimated thickness of 122 m in the Benton uplift to <15 m in the Broken Bow uplift (Stone and others, 1973; Morris, 1974). The base of the Blakely Sandstone chronologically correlates with the base of the Sauk-Tippecanoe erosional unconformity time gap in the Hunton arch and Arkoma basin/Mississippi embayment areas north and west of the Ouachita Mountains (Plates 2.1, 2.2, and 2.3) (Ethington and others, 1989; Finney, 1997; Suhm, 1997). The age of the top of the Blakely Sandstone is older than the top of the Sauk-Tippecanoe time-gap north of the Ouachita Mountains (Arkoma basin), but younger than the top of the time gap farther west (Hunton arch) (Plates 2.1, 2.2, and 2.3) (Ethington and others, 1989; Finney, 1997; Suhm, 1997). On the basis of chronologic correlation, it is suggested that deposition of the Blakely Sandstone (and perhaps the lower part of the overlying Womble Shale) is a sedimentary response to middle Ordovician erosion of the interior craton (Plates 2.1, 2.2, and 2.3) (Ethington and others, 1989; Finney, 1997; Suhm, 1997). Regional distribution of Ordovician sandstones across the southern midcontinent suggests transport of sediment from the Ozark dome (or “island”), and from areas farther north, toward the eastern part of the Ouachita off-shelf region (Plate 2.4) (Suhm, 1997).
Resting sequentially above the Blakely Sandstone in the Benton uplift is the dark-colored to black Womble Shale which contains abundant graptolites and conodonts, and is biostratigraphically well constrained (Ethington, and others, 1989) (Plate 2.3). Noted in the lower part of the formation are "calcareous distal turbidites" in contrast to proximal-type ("siltstone and pelletal") turbidites and "phosphatic breccias" which comprise the upper part of the formation (Morris, 1974). The Womble is also very carbonaceous, and is considered a hydrocarbon source formation (Arbenz, 1989).

The sandstones in the Womble Shale are similar in petrology to those in the Mazarn Shale, and contrast with those of the Blakely Sandstone which contain more lithic fragments (50% quartz : 50% lithic) (Lowe, 1989). The total estimated maximum thickness of the Womble Shale is 1064 m in the Benton uplift (Stone, and others, 1973). The Womble Shale and underlying Blakely Sandstone and Mazarn Shale contain regionally variable paleocurrent indicators. A northern source area is indicated across the Benton uplift (Plate 2.1); a southern source area is indicated for the Broken Bow uplift (Plate 2.1); and an eastern source area is indicated for areas inbetween (Lowe, 1989). The westward paleocurrents may indicate deep-water submarine contour currents. The northward paleocurrents may indicate a "southern source area" microcontinent (Llanoria of Lowe, 1989), or northward contour currents deflecting sediment derived from the Texas craton (Plate 2.1). The Ouachita Mountains are structurally complex, and much more study is necessary to further constrain interpretations of paleocurrent directions.

The Womble Shale thins westward across the Ouachita Mountains. At Black Knob Ridge, the Middle to Upper Ordovician Womble Shale is approximately 61 m thick (Plates 2.1 and 2.3) (Morris, 1974). Late Ordovician (Mohawkian) age graptolites were collected from samples of Womble Shale derived from the Stringtown Quarry in Black Knob Ridge (Finney, 1988). The Womble Shale exposed along Black Knob Ridge correlates chronologically with the Bromide Formation (upper Simpson Group) limestone that crops out nearby to the west in the Arbuckle Mountains (Plates 2.1, 2.2, and 2.3). In contrast to the Bromide Formation limestone, the Womble Shale
at Black Knob Ridge is primarily siliceous shale and bedded chert (Finney, 1988). Several relatively thin (1-2 m) limestone-clast conglomerate beds are located near the top of the Womble Shale in the Stringtown Quarry (Finney, 1988). The time period of the upper part of the Womble Shale corresponds with an unconformity at the top of the Bromide Formation observed farther west in the Arbuckle Mountains (Plates 2.1, 2.2, and 2.3). The limestone clasts incorporated in the upper part of the Womble Shale in the Stringtown Quarry (Black Knob Ridge) may be derived from eroded upper Simpson Group rocks of the Arbuckle Mountains and eastern Oklahoma shelf regions. The base of the Womble Shale is not exposed along Black Knob Ridge, and well data show that in the Black Knob Ridge and Potato Hills (Plate 2.1) parts of the western Ouachita Mountains the Womble Shale is in thrust-fault contact with younger Mississippian and/or Pennsylvanian rocks (Arbenz, 1989; Allen, 1994). Deep-water facies rocks older than the Upper Ordovician Womble Shale do not crop out west of the Broken Bow uplift (Plate 2.1).

**Late Ordovician ("Trentonian" through Cincinnatian)**

**Ouachita Mountains**

Capping the Ordovician deep-water facies of the Ouachita Mountains are the Bigfork Chert and Polk Creek Shale. The Bigfork is a dark-colored, thin-bedded, fractured chert with lesser "pelagic" interbedded muddy shales and siliceous shales, along with "turbiditic" siliceous calcisiltites and calcarenites (Morris, 1974). The Bigfork Chert is relatively hard, resists weathering, and therefore is a key marker horizon in the lower Paleozoic strata of the Ouachita Mountains. The uppermost Mohawkian ("Trentonian") through middle Cincinnatian (Maysvillian) Bigfork Chert of the Ouachita Mountains correlates with the shallow-water Viola Group and Kimmswick-Fernvale limestones exposed in the Southern Oklahoma aulacogen, Arkoma basin, and Ozark uplift (Plates 2.1, 2.2, and 2.3). The thickness of the Bigfork Chert is relatively constant throughout the Ouachita Mountains; a maximum estimated thickness is 243 m (Stone, and others, 1973).

Although the Bigfork Chert has a relatively constant thickness throughout the Ouachita Mountains, the lithology varies significantly from east to west. The most
noticeable change is the increase in proportion of limestone content in the western Ouachita Mountains. Limestone content defines the Bigfork Chert in the Black Knob Ridge area (Finney, 1988). The lower and upper boundaries of the Bigfork Chert are placed at the lowest and highest continuous limestone beds within the Ordovician stratigraphic sequence of Black Knob Ridge (Finney, 1988). Skeletal (including pelmatozoans and brachiopods) calcarenites comprise 50% of the Bigfork Chert of Black Knob Ridge (Finney, 1988).

Three medium-to-thick limestone clast conglomerate beds are located within the upper part of the Bigfork Chert in the Stringtown Quarry at Black Knob Ridge (Plate 2.1) (Finney, 1988). The limestone clasts are “angular, graded, and self-supporting,” and contain a variety of shallow-water fossils (Finney, 1988). The limestone clasts contain fossil assemblages characteristic of limestones of the Bromide Formation (upper Simpson Group) and Viola Group (Finney, 1988). In addition to fossil fauna incorporated into limestone clasts in the upper part of the formation, radiolarians and sponge spicules are found in all Bigfork Chert lithologies (Finney, 1988). Graptolites and chitinozoans are common in limestones and shale beds. Graded calcareous sandstones and limestone conglomerate “sharply overlie thin beds of siliceous shale and are overlain gradationally by bedded chert” and the sharp contacts at the top of the siliceous shale beds likely mark submarine scour-surfaces (Finney, 1988).

Throughout most of the Ouachita Mountains, the Upper Ordovician (Cincinnatian) Polk Creek Shale rests conformably above the Bigfork Chert (Plate 2.3). In much of the northern part of Black Knob Ridge (Plate 2.1), the Polk Creek Shale is absent because of erosional truncation, and where the Polk Creek Shale is absent, the Silurian Missouri Mountain Shale rests unconformably upon Upper Ordovician Bigfork Chert (Hendricks and others, 1937).

The Polk Creek Shale is a black, fissile, graptolite-bearing, carbonaceous shale (Morris, 1974). Included in the Polk Creek Shale are sparsely exposed thin beds of "calcareous chert" (Morris, 1974). As a whole, the Polk Creek is much thinner than any of the older pre-orogenic formations that crop out in the Ouachita Mountains.
Also, in contrast to the older formations, the Polk Creek Shale thickens westward and has a maximum of 53 m in the southern part of Black Knob Ridge (Plate 2.1) (Hendricks and others, 1937, Morris, 1974). Lack of sufficient number and quality of indicator fossils (graptolites or conodonts) results in inaccurate age determinations for the bases of both the Bigfork Chert and Polk Creek Shale (Ethington and others, 1989). The age of the top of the Polk Creek is also uncertain, but is placed at the Ordovician-Silurian boundary (top of Cincinnatian) (Plate 2.2). The dark-colored Polk Creek Shale roughly correlates with uppermost Viola Group-Fernvale Formation and the overlying green to dark-colored Sylvan-Cason shale of the Arbuckle Mountains, Arkoma basin, and southern flank of the Ozark dome (Plates 2.1, 2.2, and 2.3).

**Between Waco and Llano uplifts**

Several deep wells in east-central Texas have penetrated deep-water “Ouachita facies” shales, sandstones, cherts, and novaculites (Nicholas and Rozendal, 1975; Nicholas, 1989; Nicholas and Waddel, 1989). The westernmost penetration of “Ouachita facies” rocks beneath the Mesozoic Gulf Coastal Plain marks the western boundary of the Ouachita structural salient in Texas (Flawn and others, 1961). Well and seismic data east of the Llano uplift show a relatively thin (< 1000 m estimated stratigraphic thickness), imbricated succession of Ordovician through Devonian shales, cherts, and novaculites (cross sections F-F’ and G-G,’ Nicholas, 1989). The “Ouachita facies” succession of rocks drilled in the subsurface east of the Llano uplift apparently lack the thick “Mazarn-Womble Shale” succession which crops out in the eastern Ouachita Mountains.

Cross sections F-F’ and G-G’ (Figure 2.6) across the Ouachita thrust front east of the Llano uplift (Plate 2.1) show a relatively thin succession of Ordovician strata above the basal décollement and beneath the base of a thin chert-novaculite succession. The base of the thin chert-novaculite succession correlates with the base of the Upper Ordovician Bigfork Chert of the Ouachita Mountains. Cross sections of the subsurface frontal part of the Ouachita thrust belt southeast of the Ardmore basin (Hardie, 1990) suggest that the underlying strata beneath the chert succession in cross
sections F-F’ and G-G’ (Figure 2.6) likely correlate with the Womble Shale. The entire 1000 m of Ordovician-Devonian “Ouachita facies” strata in the subsurface frontal belt of Ouachita orogen east of the Llano uplift and west of the Waco uplift correlates with Womble Shale-Bigfork chert-Arkansas Novaculite. A similar Womble Shale-Bigfork Chert-Arkansas Novaculite succession exposed along strike to the northeast in both the Black Knob Ridge and Potato Hills localities in the Ouachita Mountains (Plate 2.1) (Hendricks and others, 1937; Morris, 1974; Allen, 1994). Because of scarcity of well data along the Ouachita thrust front between the Waco and Llano uplifts, the stratigraphy of the pre-Mississippian part of the “Ouachita facies” rocks is not discussed further.

**Silurian, Devonian, and Early Mississippian (early Meramacian) of the Ouachita Mountains**

The Silurian stratigraphy varies significantly across the Ouachita Mountains. Within the Broken Bow and Benton uplifts (Plate 2.1), the Blaylock Sandstone rests upon the Polk Creek Shale (Morris, 1974; Thomas, 1975). The Blaylock Sandstone is an interbedded series of greenish-gray, laminated, very fine-grained, arkosic, graywacke sandstones, siltstones, and greenish-gray shales and is interpreted as a distal turbidite submarine fan (Morris, 1974; Lowe, 1989). The Blaylock Sandstone is thickest (456 m) across the Broken Bow uplift and southwestern Benton uplift. It thins abruptly east and west along strike and is completely absent in Black Knob Ridge (Plate 2.1) (Hendricks and others 1937; Stone and others, 1973; Morris 1974). The Blaylock Sandstone also thins abruptly northward perpendicular to thrust belt strike and is absent in the northern Ouachitas (Stone and others, 1973; Thomas, 1977).

Sparse graptolites indicate that Blaylock Sandstone is Silurian (Alexandrian-Niagaran), and the time span of the Blaylock Sandstone roughly corresponds to an unconformity that separates the Keel Formation oolite (uppermost Ordovician basal formation of the Hunton Group) from the Cochrane Formation (base of Silurian part of Hunton Group) in the Arbuckle Mountains (Plates 2.1, 2.2, and 2.3) (Ethington and others, 1989; Lowe, 1989). Onset of Blaylock Sandstone deposition corresponds to, and may have been caused by, Late Ordovician-early Silurian erosional uplift (relative
sea level fall) across the region north of the Ouachita Mountains (Ozark dome, for example) (Carlson and Newell, 1997). However, north-directed paleocurrents measured across the southern Ouachita Mountains suggest a southern source (Morris, 1974; Lowe, 1989). The Blaylock Sandstone is the oldest laterally traceable and “significantly” thick formation within the deep-water facies of the Ouachita Mountains with a consistent north-directed paleocurrent (Morris, 1974).

The Blaylock Sandstone may be a north-directed clastic facies intertongue derived from an uplifted (and partially eroded) microcontinent or island arc located to the south (Morris, 1974). If derived from an “approaching” southern source area, the Blaylock Sandstone is the earliest “synorogenic (albeit distal)” formation in the Ouachita Mountains; however, the base of the synorogenic phase of deposition within the Ouachita Mountains is traditionally placed in the lower Mississippian (Meramecian), at the base of a thick succession of interbedded sandstone and shale turbidites (Morris, 1974; Arbenz, 1989a-d). However, the Meramecian marks the onset of more rapid deposition that resulted from close proximity of an “approaching” southern source area.

Above the relatively thick and localized Blaylock Sandstone across the central and southern Ouachita Mountains is the thin Missouri Mountain Shale. Except for a distinct, thin succession of fine- to medium grained quartz arenite "turbidites" at the top of the formation, most of the Missouri Mountain consists of "varicolored shales" and thickens southeastward to a maximum of 76 m in the Benton uplift (Stone, and others, 1973; Morris, 1974). The Missouri Mountain Shale is Early to Late Silurian (Niagaran to Cayugan) age and correlates with the shallow-water facies Hunton Group (upper Clarita through Henryhouse) of the Arbuckle Mountains and St. Clair through Lafferty formations of the Arkoma basin and southern flank of the Ozark dome (Plates 2.1, 2.2, and 2.3).

Across most of the northern and western parts of the Ouachita Mountains, the Silurian Missouri Mountain Shale rests unconformably upon Upper Ordovician Polk Creek Shale (Hendricks and others, 1937; Lowe, 1989). In the northern part of Black
Knob Ridge, the unconformity at the base of the Missouri Mountain Shale cuts farther down section into the Bigfork Chert (Hendricks and others, 1937).

The Devonian-Early Mississippian (Meramecian) Arkansas Novaculite caps the deep-water “pre-orogenic” succession of Paleozoic rocks in the Ouachita Mountains (Plate 2.3). The dominant lithology is a light-colored to white, intensely fractured, microcrystalline quartz-dominant chert. Three distinct members of the Arkansas Novaculite, recognized from variations in lesser lithologies, crop out in the Ouachita Mountains (Morris, 1974; Lowe, 1989). The upper and lower members are characterized by a laminated texture, "intraformational breccias of black chert, phosphate, and coarse quartz grains," whereas, the middle member consists of "graded laminated siliceous beds and black shales" (Morris, 1974; Lowe, 1989).

The complete Arkansas Novaculite succession (all three members) is preserved only in outcrops of the south-central Ouachita Mountains (Lowe, 1989). The maximum thickness of the Arkansas Novaculite where the complete formation is exposed is 289 meters (Stone and others, 1973). On the basis of conodonts, the age ranges of the Arkansas Novaculite are: lower member ranges from early to middle Devonian (Ulsterian-Erian); middle member ranges from middle Devonian-early Mississippian (Erian to Kinderhookian); and the upper member ranges from early to middle Mississippian (Kinderhookian-early Meramecian) (Plate 2.3) (Mankin, 1986; Ethington, and others, 1989). Recognition of local brecciation, "birds-eye structures," and iron-manganese crusts has resulted in a debate concerning water depth. Some argue for shallow-water subaerial conditions (Lowe, 1989, for example); others for deep-water deposition (Arbenz, 1989, for example).

The lower member of the Arkansas Novaculite is absent in the northern and western parts of the Ouachita Mountains (Hendricks and others, 1937; Lowe, 1989). The lower member either pinches out or is truncated beneath an unconformity at the base of the middle member of the Arkansas Novaculite. Surface geology at Black Knob Ridge (Plate 2.1) indicates a gradational change between the Missouri Mountain Shale and the Arkansas Novaculite (Hendricks and others, 1937); however, a
stratigraphic cross section across the Benton uplift in Lowe (1989) suggests progressive northward erosional truncation of the lower member Arkansas Novaculite.

The upper member of the Arkansas Novaculite crops out only along the southeastern part of the Benton uplift (Plate 2.1). Farther north and west an unconformity at the base of the Stanley Shale (Meramecian) cuts down section into the upper part of the middle member of the Arkansas Novaculite (Lowe, 1989). An unconformity at the base of the Stanley Shale is also argued for out crops along Black Knob Ridge (Hendricks and others, 1937). The total thickness of the Arkansas Novaculite at Black Knob Ridge exceeds the thickness in the Broken Bow uplift, but is less than that in the southeastern Benton uplift (Plate 2.1) (Morris, 1974).

The lower part of the lower member of the Arkansas Novaculite correlates with cherty limestones of the upper Hunton Group in the Arbuckle Mountains and an unconformity in the Ozark dome succession (Plates 2.1, 2.2 and 2.3). The upper part of the lower member of the Arkansas Novaculite correlates with an erosional unconformity in the Arbuckle Mountains, and with the Penters Chert (including time gaps represented by unconformities above and below) in the Arkoma basin and southeastern flank of the Ozark dome (Plates 2.1, 2.2 and 2.3).

The lower part of the middle member of the Arkansas Novaculite correlates with an unconformity beneath the upper Devonian-lowest Mississippian Woodford-Chattanooga shales of Oklahoma and Arkansas (Plates 2.1, 2.2 and 2.3). The upper part of the middle member of the Arkansas Novaculite correlates with Woodford Formation-Sylamore Sandstone-Chattanooga Shale (Plates 2.1, 2.2 and 2.3). Silicified wood is locally found in both the Woodford Formation in the Arbuckle Mountains and in the upper part of the middle member of the Arkansas Novaculite in Black Knob Ridge (Plate 2.1) in the western Ouachita Mountains (Hendricks and others, 1937; Ham, 1973). The clearly recognized unconformity at the base of the Woodford Formation likely extends southward from the Ouachita foreland in the eastern Arkoma basin (in Arkansas) across the Ouachita thrust front where it separates the middle member of the Arkansas Novaculite from underlying Silurian Missouri Mountain Shale (Plate 2.3) (Lowe, 1989).
The upper member of the Arkansas Novaculite roughly correlates with the shallow-water facies Chappel-Sycamore-Weldon-Boone succession of cherty limestones that crop out throughout the southern midcontinent from central Texas to Missouri (Plates 2.2 and 2.3). In the northern Benton uplift part of the Ouachita orogen and in areas north of the Ouachita thrust front in Arkansas, lower Mississippian (Kinderhookian-Osagean) rocks are absent (Plates 2.1 and 2.3). The absence of Kinderhookian-Osagean strata indicates that the restored location of the area between the Ouachita thrust front and the northernmost appearance of the upper member of Arkansas Novaculite was either a submarine shelf-edge (W.A. Thomas, personal communication, 2002) or a subaerially exposed broad arch in the early Mississippian prior to north-directed overthrusting.

C) Pre-Atoka/Pottsville shallow-water foreland facies clastic wedge—Early / Middle Mississippian (Meramecian) to Pennsylvanian (early Morrowan)

Black Warrior basin and Appalachian thrust front

Across the eastern Black Warrior basin (Plate 2.1) and along the Appalachian thrust front, the Tuscumbia Limestone (Meramacian) is conformable above the Fort Payne Chert (Plate 2.2) (Thomas and Osborne, 1987a,b; Thomas, 1988). The Tuscumbia Limestone contains “less chert and more coarse bioclastic limestone” than the underlying Fort Payne Chert (Thomas, 1988). The Tuscumbia Limestone is thickest in the northeastern part of the Black Warrior basin (61 m), and thins both southeastward along the Appalachian thrust front, and southwestward towards the Ouachita thrust front (Thomas, 1977, 1988).

Across the easternmost Black Warrior basin, the Monteagle Limestone rests conformably upon the Tuscumbia Limestone (Thomas, 1988). The Monteagle Limestone is Meramecian to lowermost Chesterian and is conformable beneath the Bangor limestone (Thomas, 1988) (Plate 2.2). Across the entire Black Warrior basin, the Bangor Limestone is Chesterian, and the maximum total thickness of the Monteagle-Bangor Limestones across the easternmost Black Warrior basin is 230 m (Thomas, 1988).
Beds of the Meramecian Tuscumbia Limestone are not penetrated in wells drilled within the deeper southwestern Black Warrior basin (Thomas, 1988). Here, Meramecian or Chesterian sandstones and shales rest upon Osagean cherty limestones (Fort Payne Chert) and mark the basal part of the Floyd Shale (Thomas, 1988). The absence of Tuscumbia Limestone may result from either southwestward facies transition of the Meramecian beds or erosional truncation.

The Meramecian to Chesterian beds of the Monteagle and Bangor Limestones found in the eastern Black Warrior basin interfinger with clastic facies equivalents to the west and southwest (Thomas, 1988). A northwestward-trending belt of sandstones and mudstones (Pride Mountain Formation and Hartselle Sandstone) separates the Monteagle Limestone on the northeast from the Floyd Shale on the southwest (Plate 2.2) (Thomas, 1988). The maximum combined thickness of the Pride Mountain-Hartselle formations is approximately 100 m (Thomas, 1988).

Basal beds of the Chesterian Bangor Limestone extend southwestward over the Pride Mountain and Hartselle Formations (Thomas, 1988). A tongue of the lower Bangor Limestone grades into the upper part of the Floyd Shale in deeper parts of the southwestern Black Warrior basin (Thomas, 1988). The Floyd Shale consists of mostly mudstone and contains some sandstone in the lower part of the formation across the deeper part of the western Black Warrior basin (Thomas, 1977; Thomas, 1988). The Floyd Shale and overlying Parkwood Formation thicken both toward the southwest and southeast away from the Nashville dome (Plate 2.1) (Thomas, 1972). The maximum thickness of the Floyd Shale is 258 m thick in the southwestern Black Warrior basin and 436 m thick within the southern Appalachian thrust belt (Thomas, 1977).

Across the western Black Warrior basin, the upper part of the Floyd Shale grades into the lower part of the Parkwood Formation. The lower and middle Parkwood Formation is predominantly interbedded sandstone and mudstone (Thomas, 1988). The Millerella tongue of the middle Bangor Limestone interfingers with and separates the middle Parkwood from the upper Parkwood. The upper Parkwood
Formation differs from the lower and middle part by addition of limestone beds (Thomas, 1988).

The Parkwood Formation ranges in age from Mississippian (late Chesterian) to Pennsylvanian (Plate 2.2) (early Morrowan). The maximum thickness of the Parkwood Formation in the southwestern part of the Black Warrior basin is 608 m in western parts and 790 m within the southern Appalachian thrust-belt (Thomas, 1977). The maximum thicknesses of the Floyd-Parkwood succession in the southwestern Black Warrior basin (866 m) and southern Appalachian thrust belt (1307 m) far exceed the total thickness of the correlative shallow-marine limestone facies strata (270 m, Tuscumbia-Monteagle-Bangor) in the northeasternmost part of the Black Warrior basin (Thomas, 1977, 1988).

Overall, the Floyd-Parkwood formations are part of a southwest thickening “clastic wedge” that prograded northeastward over shallow-water limestones and cherty limestones (Thomas, 1988). The southeastward thickening of the Floyd-Parkwood formations may be a result of increased subsidence along the Appalachian orogen east of the Black Warrior basin.

**Arkoma basin and Ozark dome (Arkansas and Missouri)**

An apparent unconformity at the base of the Floyd Shale in the western Black Warrior basin is a more pronounced unconformity farther west across the Ozark dome and northern Arkoma basin (Plate 2.1). Here, the Chesterian Batesville Formation sandstone sits unconformably above Osagean-Meramecian Boone Chert (Bush and others, 1977). Boone Chert correlates with the Fort Payne Chert of the Black Warrior basin (Plate 2.2). The sandstone at the base of the Floyd Shale (Meramecian) in the western Black Warrior basin may be the lithologic (or “genetic”) equivalent of the younger Batesville Sandstone (Chesterian) of the Ozark region. The sandstone at the base of the Batesville-Floyd may be a time-transgressive basal formation that separates shallow-water shelf facies below from clastic facies above the unconformity.

On the flank of the Ozark dome, the basal sandstone of the Batesville Formation grades into a limestone which is classified in that area as the Hindsville Limestone member (Plates 2.1 and 2.2) (Bush, and others, 1977). Farther west, in
northeastern Oklahoma, the Hindsville limestone grades into “Grand River” strata (shown in quotes in Bush and others, 1977) (Figure 2.5 and Plate 2.2). The “Grand River”-Hindsville-Batesville rocks correlate with the upper part of the Caney Shale which crops out farther southwest within the Arbuckle uplift (Plates 2.1 and 2.2). Although Chesterian, the age of the top of the “Grand River”-Hindsville-Batesville units do not precisely correlate with the top of the Caney Shale (Plate 2.2). The Batesville Formation thickens southeastward from 23 m in the northern Arkoma basin to 134 m near the Ouachita thrust front (Bush and others, 1977; Van Arsdale and Schweig, 1990).

Conformably above the “Grand River”-Hindsville-Batesville rocks is the Fayetteville Shale (Figure 2.5). Across the Ozark dome and northern Arkoma basin (Plate 2.1), the Fayetteville Shale is primarily black shale with lesser amounts of sandstone and limestone (Bush and others, 1977). A basal sandstone member (Weddington) crops out on the flanks of the Ozark dome (Bush and others, 1977). The Fayetteville Shale is Chesterian in age and 106 m thick across the northern Arkoma basin (Figure 2.5, and Plates 2.1 and 2.2) (Bush and others, 1977). To the east, the Fayetteville Shale correlates with the much thicker Floyd Shale of the Black Warrior basin (Plates 2.1 and 2.2). Southwest, towards the Arbuckle uplift, the Fayetteville Shale grades into upper beds of the Caney Shale and lower beds of the Goddard Shale (Plates 2.1 and 2.2).

Resting conformably above the Fayetteville Shale across the northern Arkoma basin and flank of the Ozark dome is the Pitkin Limestone (Figure 2.5, and Plates 2.1 and 2.2). The Pitkin Limestone consists of massive limestone and shale (Bush and others, 1977). An unconformity marks the top of the Pitkin Limestone and the entire formation is Chesterian (Plate 2.2). The maximum thickness of the Pitkin Limestone across the northern Arkoma basin is 61 m (Bush and others, 1977). Southwestward towards the Hunton arch, the Pitkin Limestone (or correlative unit) is absent (completely truncated by erosion) (Plate 2.2).

The Pitkin Limestone chronostratigraphically correlates with the Parkwood Formation and Bangor Limestone of the Black Warrior basin (Plates 2.1 and 2.2).
Limestone interbeds within the Parkwood Formation are lithologic equivalents of the Pitkin Limestone. The Pitkin Limestone may intertongue with the Parkwood-type clastic rocks in western parts of the Mississippi Valley graben or western Black Warrior basin (Plates 2.1 and 2.2). These intertongues may be similar to intertongues of the Parkwood clastic rocks and Bangor limestones described for the eastern Black Warrior basin (Thomas, 1988).

Across the southern flank of the Ozark dome and the eastern Arkoma basin, the Hale Formation is unconformable above the massive Pitkin Limestone (Plates 2.1 and 2.2). In this region, the Hale Formation is separated into two recognized members (Cane Hill and Prairie Grove) (Bush and others, 1977). The Hale Formation is mostly shale but contains sandstone and calcareous sandstone (Bush and others, 1977). Hale Formation clastic rocks thicken greatly southeastward from 76 m across the northern Arkoma basin to more than 541 m near the Ouachita thrust front (Bush and others, 1977; Van Arsdale and Schweig, 1990).

Within the Arkoma basin, the Hale Formation is Morrowan; however, ages of correlative lithostratigraphic formations farther east and west are not clearly defined. To the east, the Hale Formation chronostratigraphically correlates with the Pottsville Formation of the Black Warrior basin; however, the maximum thickness of the Pottsville Formation is far greater than the Hale Formation (including overlying Bloyd Shale) (Plates 2.1 and 2.2) (Thomas, 1975; Bush and others, 1977). Also, the terrigenous lithologies of the Pottsville Formation (sandstones, conglomerates, coals) differ from the finer-grained and calcareous lithologies of the Hale Formation (Thomas, 1975; Bush and others, 1977). The calcareous shales and sandstones of the Hale Formation (Morrowan) are lithologically more similar to the calcareous clastic rocks that comprise the upper part of the Parkwood Formation (Chesterian) (Bush and others, 1977; Thomas, 1988). East of the Mississippi Valley graben, the sandstone at base of the Hale Formation may be time-transgressive and grade into the beds of the upper Parkwood Formation (Plates 2.1 and 2.2).

The chronostratigraphic equivalent of the Hale Formation farther west toward the Hunton arch is the Limestone Gap Shale (upper “Springer” Group); however,
sandstones in the Hale Formation may lithologically correlate with Rhoda Creek-Union Valley Sandstone (lower “Springer” Group) (Plates 2.1 and 2.2). The Hale-Union Valley-Rhoda Creek Formations may be a laterally continuous time-transgressive unit that extends from the southern flank of the Ozark dome towards the Hunton arch (Plates 2.1 and 2.2). The relationship between the Hale Formation and the lower Springer Group in northeastern Oklahoma is symmetrically similar to the relationship between the Hale and the Parkwood Formation in the Black Warrior basin (Plate 2.2).

Across the southern Ozark dome and eastern Arkoma basin, the Bloyd Shale is conformable above the Hale Formation (Plates 2.1 and 2.2). The Bloyd Shale is primarily shale with lesser amounts of sandstone, limestone, and sparse coal beds (Bush and others, 1977). Southwest of the Ozark dome, the Bloyd Shale interfingers with the Limestone Gap Shale and Wapanucka Limestone (Plate 2.2). Correlation of the Bloyd Shale with formations farther east in the Black Warrior basin is not as clear. The Bloyd Shale chronostratigraphically correlates with the middle part of the Pottsville Formation (Plate 2.2). However, non-calcareous rock types that comprise the Bloyd Shale are found in both the Pottsville and the underlying upper Parkwood (Plate 2.2). Bloyd Shale beds may grade into either upper Parkwood or Pottsville strata. The Bloyd Shale thickens southeastward from 106 m to 260 m near the Ouachita thrust front (Bush and others, 1977; Van Arsdale and Schweig, 1990). An unconformity separates the Bloyd Shale from the overlying basal Spiro Sandstone unit of the Atoka Formation (Plate 2.2).

**Southern Oklahoma aulacogen (Wichita and Arbuckle uplifts)**

At the base of the early Mississippian-late Pennsylvanian foreland synorogenic clastic facies is the Caney Shale. In the Arbuckle Mountains, these "dark-gray, fissile shales" locally contain phosphate and calcareous (septarian) nodules, and cephalopods (Ham, 1973). Along the northern margin of the Ardmore basin, Caney shales are thicker and slightly more siliceous than in outcrops farther north (Ham, 1973). Regional estimates of the maximum thickness of the Caney Shale range from less than 198 m (Ham, 1973) to 304 m (Gatewood and Fay, 1991).
On the basis of cephalopod fauna and graptolites, the age of the Caney Shale ranges from middle Mississippian (Meramacian) to late Mississippian (Chesterian) (Elias, 1956; Ham, 1973; Ethington and others, 1989). In the easternmost part of the Arbuckle uplift, Meramecian age Caney Shale beds sit unconformably upon Kinderhookian age Welden Limestone (Plates 2.1 and 2.2). Southward across the Arbuckle uplift the Weldon Limestone grades into the interbedded limestones and clastic rocks of the Sycamore Formation (Ham, 1973). The unrecorded geologic time represented by the unconformity at the base of the Caney Shale is much smaller along the southern margin of the Arbuckle uplift, where Meramecian age Caney Shale beds rest unconformably upon Osagean limestones in the upper part of the Sycamore Formation (Plates 2.1 and 2.2).

In the southern parts of the Arbuckle uplift, the boundary between the Caney Shale and the overlying Goddard-Springer shale layers is gradational and somewhat arbitrary. The base of the Goddard-Springer sequence is generally placed at the lowest appearance of siderite and "clay-ironstone" beds common to Springer Group rocks (Ham, 1973). These shales contain few fossil invertebrates, but large amounts of spores and pollen (Ham, 1973). Unlike the underlying Caney Shale, the Goddard-Springer Group contains increased proportions of sandstone and limestone (Ham, 1973). Proportion of sandstone beds increase south towards the Ardmore basin where several sandstone formations are recognized (Ham, 1973). The Goddard-Springer shales thicken dramatically southeastward from 106 m, in the northeasterneastern Arbuckle Mountains, to 1368 m in the Ardmore basin (Plate 2.1) (Ham, 1973). The Goddard Shale is entirely Chesterian; however, the Springer Group beds range from Chesterian to Morrowan (Plate 2.2).

Across the Hunton arch (Plate 2.1), an unconformity separates Goddard Shale beds from the overlying lower “Springer” Group (Rhoda Creek-Union Valley Sandstone) (Plate 2.2) (Ethington and others, 1989). Farther northeast towards the Ozark dome (Plate 2.1), the Springer Group rocks grade into the Hale-Bloyd Formations (Plate 2.2).
Across most of the Arbuckle uplift, Hunton arch, and western Arkoma basin the Wapanucka Formation rests conformably above the top of the Springer Group clastic rocks. Along the southern margin of the Arbuckle uplift and within the Ardmore basin, the Primrose Sandstone grades the Wapanucka Formation (Ham, 1973) (Plates 2.1 and 2.2). The Primrose Sandstone is one of the several recognized sandstone formations included within the Springer Group. The Wapanucka Formation contains oolitic limestone, calcareous shale, and quartzose sandstone members (Ham and others, 1954). The Wapanucka Formation is thickest (61 m) across the western part of the frontal zone of the Ouachita thrust belt (Plate 2.2) (Ham and others, 1954). Northeast of the Hunton arch, the Wapanucka Formation interfingers with the Bloyd Shale (Plates 2.1 and 2.2). In the eastern Arkoma basin, the Wapanucka Formation is also truncated by an erosional unconformity at the base of the overlying Spiro Sandstone (Plate 2.2) (Shelton, 1996).

The entire Caney Shale-Goddard-Springer-Wapanucka succession (Meramecian-latest Morrowan) of the Southern Oklahoma aulacogen is lithologically similar, and roughly correlates with the Floyd Shale-Parkwood strata (Meramecian-early Morrowan) of the Black Warrior basin (Plates 2.1 and 2.2). The lower part of the Pottsville Formation (Morrowan) of the Black Warrior basin may be correlative to the upper Springer-Wapanucka rocks of eastern Oklahoma; however, the thickness of the Pottsville (3040 m) is nearly twice as great as the maximum thickness of the entire Caney-Wapanucka strata (1610 m) (Plate 2.2) (Ham, 1973; Thomas, 1977). The Pottsville Formation is much more similar in thickness and lithology to the Atoka Formation (Atokan age-possibly latest Morrowan at base) of the eastern Arkoma basin (Plate 2.2) (Bush and others, 1977; Thomas, 1977). Thickness of the Pottsville suggests that it is the time-transgressive lithologic equivalent of the thick Atoka Formation which crops out west of the Mississippi embayment (Plate 2.2).

**Ardmore basin and Muenster arch**

The Ardmore basin (Plate 2.2) contains the thickest and most continuous Meramecian-Morrowan section of the Southern Oklahoma aulacogen (Ham, 1973). Estimated maximum thickness of the Caney-Springer series exceeds 1600 m (Ham,
1973). No major recognized unconformities are located within the Caney-Springer strata of the deep Ardmore basin. The base of the Caney Shale is apparently unconformable and gradational with the underlying Sycamore Limestone (Ham, 1973). An unconformity that separates the top of the Springer Group from the overlying Dornick Hills Group is progressively more pronounced farther south towards the Muenster arch (Plates 2.1 and 2.2) (Cooper, 1995). Wells in the southeastern end of the Muenster arch, show that Morrowan clastic rocks of the Dornick Hills Group (also classified as “grey” Strawn in Texas) rest unconformably upon Middle Ordovician carbonates and sandstones of the Simpson Group (Bradfield, 1957a-c)(Plates 2.1 and 2.2). Locally along the crest of the Muenster arch the entire Paleozoic section is absent and Mesozoic rocks rest upon Precambrian basement (Ewing, 1991).

**Llano uplift and Fort Worth basin**

Lithostratigraphic (and roughly chronostratigraphic) equivalents of the Caney-Springer-Goddard rocks are found farther south on the eastern flank of the Llano uplift (Plates 2.1 and 2.2). Across the Llano region, resting disconformably above the thin uppermost Devonian (Chautauquan) to lower Mississippian (Kinderhookian) Chappel Limestone is a mixture of dark-colored to black, fissile, very thinly laminated, carbonaceous shale and limestones recognized as the Barnett Shale (Kier, 1988). The base of the Barnett is a thinly-laminated calcisiltite, whereas the top consists of "fine-to coarse-grained goniatite and pellet packestone and phosphatic oomicrite" (Kier, 1988). The upper parts of the Barnett also include microsparite concretions (as much as 2.43 m) that include variably sized coiled and straight cephalopods (Kier, 1988). Glauconite and phosphate content increases westward (Kier, 1988). The maximum thickness of the Barnett Shale in the Llano region is 15 m (Kier, 1988). The Mississippian (Osagean-Chesterian) Barnett Shale thickens eastward away from the Llano uplift towards the Fort Worth basin and Ouachita thrust front (Kier, 1988) (Plates 2.1 and 2.2).

The dark-colored Barnett Shale of the Llano region correlates lithologically and chronologically to the Caney Shale of the Southern Oklahoma aulacogen (Plates
2.1 and 2.2). The much greater thickness of the Caney Shale in the Ardmore basin (198 m) indicates that the Ardmore basin region isostatically subsided to a greater extent than the Llano uplift during deposition of the Caney-Barnett Shales. Restoration of a cross section constructed across the Waco uplift region (Plate 2.1), east of the Llano uplift, shows that the Barnett Shale grades into deep-water sandstone and shale turbidites (Nicholas and Rozendal, 1975; Nicholas, 1989).

An apparent unconformity separates the Barnett Shale from the overlying Marble Falls Limestone (Plate 2.2) (Kier, 1988). Within the Llano uplift, a resistant and “heavily vegetated limestone ledge” marks the boundary between the Barnett Shale and the Marble Falls Limestone (Kier, 1988). Although field study suggests that the Marble Falls-Barnett contact is conformable, conodont biostratigraphy indicates unconformity (Kier, 1988). Because the boundary between the Barnett and Marble Falls Formations approximates the Mississippian-Pennsylvanian boundary in the Llano region, an unconformity analogous to the one observed at the base of the Hale Formation in the Ozark dome region has traditionally been placed at the base of the Marble Falls Limestone (Plates 2.1 and 2.2) (Kier, 1988; Bush and others, 1977). This controversy is beyond the scope of this dissertation and shall not be examined further.

The Marble Falls Limestone contains “cherty and non-cherty limestones interbedded with shale” (Kier, 1988). The age of the Marble Falls Limestone is primarily Morrowan; however, the basal part may be Chesterian (Plate 2.2) (Fay, 1986a; Kier, 1988). A locally recognized unconformity separates the lower and upper members of the Marble Falls Limestone. The unconformity is within the Morrowan and may correlate to the unconformity between the Dornick Hills Group and underlying Springer Group in the Ardmore basin (Plates 2.1 and 2.2).

The lower member of the Marble Falls Limestone is mostly light- to dark-colored cherty limestone with thin shale beds (Kier, 1988). The limestones vary from coarse to fine grained, are fossiliferous, and contain local coral and algal biolites (Kier, 1988). The upper member consists primarily of light- to dark-colored algal biomicrite, siliceous spiculitic biomicrite, and shale (Kier, 1988). Across the Llano region, thickness of the lower member ranges from 21 m to 45 m (Kier, 1988). The
upper boundary of the Marble Falls Limestone is time-transgressive and gradational with the overlying Smithwick Shale (Kier, 1988) (Plate 2.2).

In stark contrast to the Marble Falls, the overlying Smithwick Shale is a soft, dark-colored to black, sparsely fossiliferous massive shale. Thickness and number of subsidiary siltstone and sandstone beds increase upsection (Kier, 1988). Similar lithologic descriptions of the Smithwick Formation and shallow-water Atoka Formation of the northeastern Arbuckle Mountains suggest stratigraphic correlation. Sparse fossils and siderite ("iron-rich") concretions (Ham, and others, 1954; Kier, 1988) found in both formations also indicate correlation. The Smithwick Shale has a maximum thickness of 114 m in the Llano region (Kier, 1988). The boundary between the Smithwick Shale and overlying Strawn Group sandstones, shales, and limestones appears gradational in outcrop; however, biostratigraphy suggests a small unconformity between the Smithwick Shale and the overlying Strawn Group (Fay, 1986f; Kier, 1988).

The lower Marble Falls Limestone chronostratigraphically correlates with the Springer Group of the Ardmore basin; however, the predominance of limestone facies across the Llano uplift contrasts with the predominance of clastic facies within the Ardmore basin (Plates 2.1 and 2.2). The Marble Falls Limestone is stratigraphically analagous to the slightly older Bangor Limestone of the eastern Black Warrior basin (Plates 2.1 and 2.2). Uplift and erosion marked by a late Morrowan unconformity has likely obliterated intertongues between Marble Falls Limestone and Springer Group sandstones and shales along the crest of the Muenster arch (Plates 2.1 and 2.2). East of the Llano uplift, in the subsurface of the Waco uplift, the Marble Falls Limestone grades into, or interfingers with, undivided moderate- to deep-water sandstone and shale turbidites (Plates 2.1 and 2.2) (Nicholas and Rozendal, 1975).

The upper Marble Falls Limestone and Smithwick Shale correlate with the upper part of the Dornick Hills ("gray" Strawn) rocks of the eastern Muenster arch, Ardmore basin, and southwestern Arbuckle uplift, and with the shallow-water facies Atoka Formation on the northeastern flank of the Arbuckle uplift and Arkoma basin (Plates 2.1 and 2.2). Unconformably above the Smithwick Shale is the Strawn Group.
The Strawn Group rocks that crop out around the Llano uplift correlate with the Deese Group ("red" Strawn) strata of the Muenster arch, Ardmore basin, and Arbuckle uplift (Plates 2.1 and 2.2). The base of the Strawn Group (latest Atokan or Desmoinesian) marks the base of a series of rapidly deposited clastic rocks and limestones (Desmoinesian through Wolfcampian) of the Ouachita foreland of central Texas (Barnes, 1988; Cleaves, 1996).

D) Pre-Atoka Formation deep-water facies syn-orogenic clastic wedge
--Early Mississippian (Meramecian) to Pennsylvanian (early Morrowan in the east to late Morrowan/early Atokan in the west)

Southwestern margin of Black Warrior basin

Correlation of Mississippian-Pennsylvanian aged deep-water facies rocks across the Mississippi Embayment is imprecise because of sparse well data (Thomas, 1972b, 1977, 1988; Thomas and others, 1989). A deformed belt of undivided siliciclastic, carbonaceous, and cherty rocks in western Mississippi (Western Mississippi slate belt) has been drilled along the southwestern flank of the Black Warrior basin (Plate 2.1) (Thomas, 1972b, 1989). Wells along the northeastern side of the Western Mississippi slate belt have drilled partly carbonaceous clastic rocks that may represent generally deeper water correlatives of Meramecian-Morrowan/early Atokan Floyd-Parkwood-Pottsville succession (Thomas, 1988, 1989).

Ouachita Mountains

In southeastern parts of the Ouachita Mountains, the base of the Stanley Group rests conformably upon the upper member of the deep-water pre-orogenic Arkansas Novaculite (Lowe, 1989). North of the Ouachita Mountains, Stanley Group strata grade into shallow-water facies Batesville Sandstone, Fayetteville Shale, and Pitkin Limestone (Plates 2.1 and 2.3). Stratigraphic sections compiled across the northern Benton uplift show that Stanley Group rocks rest unconformably upon upper or middle member Arkansas Novaculite (Plates 2.1 and 2.3) (Lowe, 1989). The amount of unrecorded geologic time represented by the missing strata increases progressively towards the north (Lowe, 1989). In the western part of the Ouachita thrust belt the Stanley Group rests unconformably upon the middle member of the Arkansas
Novaculite (Plate 2.3) (Hendricks and others, 1937; Lowe, 1989). Farther west in the Ouachita Mountains toward the western Arkoma basin, the Stanley Group strata grade into Caney-Goddard-Springer rocks (Plates 2.1 and 2.3).

Stanley Group rocks range in age from early to late Mississippian (Meramacian- Chesterian) (Plate 2.2) (Ethington and others, 1989). As a whole, except for Chesterian-age plant fossils found in middle layers, fossils are rare in the Stanley Group and age constraints are based on fossils derived from foreland-facies carbonate boulders encapsulated in some Stanley Group shale beds (Ethington and others, 1989). Fossil fauna includes: bryozoans, blastoids, brachiopods, mollusks, and trilobites (Ethington, and others, 1989). Incorporation of shallow-water carbonates within the Stanley indicates that part of the Paleozoic carbonate shelf within the Ouachita salient (Plate 2.1) was exposed in the Mississippian.

The Stanley is predominantly a dark, greenish-grey to black-colored shale with subordinate fine-grained, arkosic siltstone and sandstone "proximal and distal turbidites" (Morris, 1974). Sandstone and feldspar content increase southeastward in the Ouachita Mountains (Morris, 1974; Thomas, 1979). Thickness and abundance of "tuffs and "tuffaceous sandstones" also increase toward the southeast (and south) (Morris, 1974; Niem, 1977). Towards the north (and northwest), the Stanley contains more siliceous shale and "impure chert" beds, and contains a thin tuff zone in the upper part (Morris, 1974). The southward increase in proportion of arkosic beds and sandstone, and the southward increase in proportion of tuffs (especially in lower Stanley Group strata) indicate that the source area for both the sediment and volcanic debris was south of the present day Ouachita Mountains (Morris, 1974; Lowe, 1989; Titus and Legg, 1995).

Locally in Arkansas, large blocks of carbonate-platform rocks are found surrounded by "rubble-bedded" Stanley Group shale with sedimentary structures that suggest south-directed, sub-marine slumping and debris flow (Morris, 1974). Sedimentary structures of another distinct part of the Stanley Group, a thin 30 m distinct quartz-arenite known as the Hot Springs Sandstone, suggests a "channel-fill" deposit with a northern provenance (Morris, 1974). Facies and paleocurrent
interpretations suggest a southern provenance for the lower Stanley Group, changing progressively to an eastern provenance for the upper Stanley Group (Thomas, 1979). The localized southward-transported slump deposits and Hot Springs Sandstone, derived from source areas to the north, intersect the predominant westward-transported Stanley strata in the northeastern part of the Ouachita Mountains. The maximum estimated thickness of the Stanley Group in Arkansas is 2584 m (Stone and others, 1973). The estimate of the regional maximum thickness of Stanley Group strata is 3647 m (Thomas, 1977).

Resting above the Stanley Group is a ridge-forming, "rhythmically interbedded", succession of dark-colored to black shales and light-colored, fine-grained, quartzose sandstones referred to as the Jackfork Sandstone (Morris, 1974). The Jackfork ranges in age from latest Mississippian (latest Chesterian) to early Pennsylvanian (middle Morrowan) (Plate 2.3) (Ethington and others, 1989). The age of the base of the Jackfork is uncertain because Chesterian plant fossils may be included in reworked clasts from the underlying Stanley Group (Ethington, and others, 1989). Morrowan age cephalopods, corals, and brachiopods are recognized in the Jackfork strata in Arkansas (Ethington, and others, 1989).

Dominant lithology of the Jackfork Sandstone varies across the Ouachita Mountains. Rubble-bedded, slump-type deposits are most prevalent in the frontal fault zone of Arkansas (Morris, 1974). Thicker, undulose, massive quartzose "scour-and-fill""proximal" turbidites layers predominate in the southeast; whereas, thinner, dark-colored, laminated siliceous "distal" turbidites predominate to the southwest (frontal fault zone of Oklahoma) (Morris, 1974). These distal turbidites locally contain complete Bouma sequences, trace fossils, and "dewatering structures"(LoPiccolo, 1973; Morris, 1974). Also included in the Jackfork of Oklahoma are "several distinct zones" that contain clasts which include: quartz, carbonate, and "shallow-water fossils" (Morris, 1974). The Jackfork Sandstone is generally more quartzose than the underlying Stanley Group, and the Jackfork of the southeastern Ouachita Mountains of Arkansas is richer in both quartz-pebbles and feldspar (Thomas, 1979). Shallow-water facies limestone boulders and quartzose sandstones with interpreted northern
provenance are incorporated into the Jackfork Sandstone in the northeastern Ouachita Mountains; whereas, feldspatic sandstones with interpreted southern (or southeastern) provenance are incorporated into the Jackfork in the southern part of the Ouachita Mountains (Stone and others, 1973; Morris, 1974; Thomas, 1979). Estimates of the maximum thickness of the Jackfork Sandstone range from 1824 m (Stone, and others, 1973) to 2128 m (Thomas, 1977).

Restoration of thrust-imbricated strata within the Ouachita Mountains suggest deep-water facies Jackfork Sandstone grades into shallow-water facies equivalent clastic rocks north and west of the Ouachita Mountains. North of the Ouachita Mountains, the deep-water facies Jackfork Sandstone turbidites grade into the Hale Formation shallow-water facies shale, sandstone, and limestone (Plates 2.1 and 2.3). West of the Ouachita Mountains, Jackfork Sandstone strata grade into shallow-water facies Springer Group clastic rocks (Plates 2.1 and 2.3).

Completing the early Meramecian to latest Morrowan part of the syn-orogenic succession of the Ouachita Mountains is the Johns Valley Shale (Plate 2.3) (Ethington and others, 1989). Although the overlying Atoka Formation is also part of the syn-orogenic succession, it marks an abrupt increase in sedimentation rates from 420 m per m.y. (Jackfork) to 1000 m per m.y. (Houseknecht, 1986; Viele and Thomas, 1989); therefore, it is discussed later in this chapter. Furthermore, a regional unconformity separates latest Morrowan and Atokan age shallow-water foreland facies strata (Dornick Hills Group and “grey” Strawn, for example) from underlying Morrowan and older formations west of the Ouachita thrust belt in Oklahoma and Texas (Plates 2.1, 2.2, and 2.3).

The predominant lithology of the Johns Valley Shale varies upsection (and eastward) from a basal laminated dark-colored to black shale, to interbedded light-grey mudstones and thin, dense, greenish-brown quartzose sandstones (Shideler, 1970; Morris, 1974). The Johns Valley Shale has an estimated maximum thickness of 456 m in the eastern Ouachita Mountains (Stone and others, 1973). The sandstones of the Johns Valley Shale are generally classified as subgraywackes, and include sole markings and convolute bedding (Shideler, 1970). Also scattered throughout the
formation are phosphate, siderite, and limonite nodules, and conglomeratic beds that contain a variety of lithologically distinct clasts (Shideler, 1970). These conglomerates range from fine-grained calcareous sandstones to swirled, contorted boulder beds. Most resistant, coarse-grained "carbonate" formations of the foreland (Upper Cambrian–lower Mississippian strata) are represented as boulders in the Johns Valley Shale. Most of the boulders in the Johns Valley Shale in the western parts of the Ouachita Mountains are Upper Cambrian-Lower Ordovician Arbuckle Group (Shideler, 1970). Transitional and deep-water chert boulders derived from Bigfork Chert (Upper Ordovician) and Arkansas Novaculite (Devonian-lower Mississippian) increase in quantity in the eastern parts of the Ouachita Mountains (Shideler, 1970).

In general, the calcareous detritus within the Johns Valley Shale (and older syn-orogenic strata of the Ouachita Mountains) is considered to be derived from the shelf and craton to the northwest (Shideler, 1970); however, mechanism of emplacement of boulders varies according to interpreter. On one end of the spectrum, Shideler (1970) argues that all of the boulders were deposited by submarine slumps with source area in a shallow-water to deep-water transition zone (shelf-edge). Viele and Thomas (1989) suggest that the Arkansas Novaculite and Bigfork Chert boulders in the Johns Valley Shale are better explained by tectonic emplacement.

Restoration of the Ouachita thrust belt suggests that the Johns Valley Shale grades into shallow-water facies clastic rocks and limestones north and west of the Ouachita Mountains (Plates 2.1 and 2.3). North of the Ouachita Mountains, the Johns Valley Shale grades into the shallow-water facies equivalent Bloyd Shale (Plate 2.3). Northwest and west of the Ouachita Mountains, in Oklahoma, the Johns Valley Shale grades into the clastic rocks of both the uppermost Springer Group and overlying Wapanucka Formation limestones and sandstones (Plate 2.3) (Ethington and others, 1989).

**Waco uplift**

The Stanley Group through Johns Valley Shale succession of the Ouachita Mountains correlates with an undivided succession of deep-water facies sandstones and shale turbidites. Nicholas (1989) uses several wells and to constrain a cross
section constructed through the Waco uplift of Texas (also see Figure 2.6). The Paleozoic rocks of the Waco uplift are completely buried by the Mesozoic Gulf Coastal Plain. A seismic reflection profile provides evidence of a broad subsurface anticline (Nicholas and Rozendal, 1975). West of the core of the Waco uplift, Ordovician-Devonian cherts rest above the basal detachment of the Ouachita allochthon (Nicholas, 1989). The undivided deep-water shales and sandstones resting above these cherts likely correlate to the Stanley Group (and possibly lower parts of the Jackfork Sandstone). The maximum thickness of the undivided deep-water facies clastic rocks (approximately 2750 m) west of the Waco uplift (Nicholas, 1989a) is between the range of reported maximum thickness for the Stanley Group (2584m to 3647m) (Stone and others, 1973; Thomas, 1977).

Across the Waco uplift, the deformed Paleozoic strata above the Cambrian-Devonian carbonates are shown in Nicholas and Rozendal (1975) and in Nicholas (1989) as undivided allochthonous pre-orogenic and syn-orogenic deep-water facies rocks similar to those that crop out in the Ouachita Mountains (Plates 2.1, 2.3, and 2.5). However, cherts drilled in the allochthon to the west, are lacking across the Waco uplift (Figure 2.6) (Nicholas, 1989). The surface above the carbonate rocks (as illustrated in the Shell No. 1 Barrett well, Nicholas and Rozendal, 1975) within the Waco uplift either represents a thrust-fault surface where chert has been displaced westward, or an unconformity surface where the Devonian-Mississippian cherts are absent (Plate 2.5) (Nicholas, 1989). Widespread regional unconformities within the shallow-water facies Devonian-Early Mississippian strata from Texas to Arkansas argue for an unconformity surface above the Waco uplift carbonates (Plates 2.1, 2.2, and 2.3).
E) Rapid deposition phase of the shallow-water facies syn-orogenic clastic wedge--Pennsylvanian to Permian (early Morrowan to Desmoinesian towards the east and late Morrowan to Wolfcampian towards the west)

Black Warrior basin and Appalachian thrust-front

In the eastern Black Warrior basin the stratigraphic base of the phase of rapid deposition is marked by a massive sandstone at the base of the Pottsville Formation (Thomas, 1988) (Plates 2.1 and 2.2). The contrast between the Pottsville and the underlying Bangor Limestone is stark; however, it is uncertain whether the transition is unconformable. A thin set of green and gray shales and sandstones separating the Bangor Limestone from the Pottsville Formation suggest a conformable progression (Thomas, 1972b). Scour-surfaces at the base of the massive sandstone (lowest Pottsville Formation) may indicate a regional unconformity or localized channelization (Thomas, 1988). Unconformities mark the base of the rapid-deposition phase (Strawn Group-Dornick Hills Group-Atoka Formation) farther west within the foreland of the Ouachita orogen (Plates 2.1 and 2.2).

Across western and southwestern parts of the Black Warrior basin, the base of the phase of rapidly deposited clastic sediments is imprecisely known. The massive sandstone at the base of the Pottsville Formation across the eastern part of the Black Warrior basin is replaced in the central part by “mudstone,” and in the western part by “cyclic sandstone and mudstone”(Thomas, 1988). Across the western Black Warrior basin, regional well data suggest that the sandstones and mudstones of the lower Pottsville Formation interfinger with, or grade into, the underlying Parkwood Formation at a diachronous prograding contact (Thomas, 1988).

Across the eastern Black Warrior basin, the basal massive sandstones of the lower Pottsville Formation grade upward into gray mudstones and coal beds (Thomas, 1988). Farther west, the upper Pottsville Formation is a continuous series of cyclic clastic rocks with little difference between lithology of upper and lower parts of the formation (Thomas, 1988). Age of the Pottsville Formation ranges from earliest
Morrowan to earliest Atokan (Plate 2.2). The maximum measured thickness of the Pottsville Formation within the Black Warrior basin is 3040 m (Thomas, 1977).

West of the Black Warrior basin, beds younger than the youngest preserved Pottsville Formation include the shallow-water facies Atoka Formation of eastern Oklahoma and the Arkoma basin (Plates 2.1 and 2.2). The base of the rapidly deposited (thick) succession of clastic rocks (Pottsville-Atoka) is time-transgressive from lower Morrowan (east in the Black Warrior basin) to latest Morrowan (toward the Arkoma basin in the west) (Plate 2.2). The Pottsville Formation thickens southwest across the Black Warrior basin toward the Ouachita thrust front and the uppermost part correlates with, and likely grades into, deep-water facies Atoka Formation strata exposed within Ouachita Mountains) (Thomas, 1988).

**Arkoma basin and Ozark dome (Arkansas and Missouri)**

Across northeastern Oklahoma and the Arkoma basin, the rapid phase of deposition consists of the shallow-water facies Atoka Formation and the overlying Desmoinesian sequence (Hartshorne, McAlester, Savanna, Boggy, Senora, Calvin, Wetumka, and Wewoka Formations) (Krumme, 1981; Visher, 1996) (Plates 2.1 and 2.2). Although the Desmoinesian succession is generally considered post-Ouachita orogenic in the stratigraphic succession of northeastern Oklahoma and Arkansas (marked by an abrupt decrease in sedimentation rate) (Arbenz, 1989c; Viele and Thomas, 1989), it is included in this section because the Desmoinesian succession is much thicker toward the southwest in Texas where the base of the Permian is considered the base of the post-orogenic succession (Muehlberger and Tauvers, 1989, for example). Furthermore, late Pennsylvanian (Virgilian) and Permian strata unconformably overlie all younger strata farther west in the Arbuckle uplift and mark the end of the “Arbuckle” orogeny in the Southern Oklahoma aulacogen (Ham and others, 1954; Ham, 1973; Thomas and others, 1989).

The Hartshorne through Boggy Formations are included in the Krebs Group (Krumme, 1981; Visher, 1996). The Senora and Calvin Formations are included in the Cabaniss Group; and the Wetumka Shale and Wewoka Formation are included in the Marmoton Group (Krumme, 1981; Visher, 1996). Across the central Arkoma
basin and platform region of northeast Oklahoma the Skiatook Group rests above the Marmaton Group (Krumme, 1981; Visher, 1996). The Holdenville Shale at the base of the Skiatook Group previously considered upper Desmoinesian (Ham and others, 1954), is now considered Missourian in age (Krumme, 1981; Visher, 1996). The Skiatook Group completes the late Paleozoic syn-orogenic shallow-water facies succession of rocks of the Ouachita foreland northeast of the Southern Oklahoma aulacogen (Plate 2.1).

The Krebs Group is thickest within cores of synclines near the Ouachita thrust front and thins westwards towards the Nemaha Ridge and Arbuckle uplift (Plate 2.1). The Savanna and Boggy Formations thin slightly from east to west across the western Arkoma basin (Krumme, 1981; Visher, 1996). The underlying Hartshorne and McAlester Formations pinch out westwards (Krumme, 1981; Visher, 1996). A limestone at the base of the Boggy Formation (upper Savanna Formation) is a key marker bed used for regional stratigraphic correlation (Krumme, 1981; Visher, 1996).

In northeastern Oklahoma, the Savanna Formation (Desmoinesian) rests unconformably upon Mississippian limestones (Hale Formation or Springer Group) (Krumme, 1981; Visher, 1996). Like the Krebs Group, the overlying Cabaniss and Marmaton Groups thin westward from the central Arkoma basin towards the Nemaha Ridge (Krumme, 1981; Visher, 1996).

The shallow-water Atoka Formation is primarily sandstone and shale and has a maximum thickness of 1667 m (Bush and others, 1977). The overlying Desmoinesian rocks (Krebs, Cabaniss, Marmaton, and Skiatook Groups) are predominantly shale, with lesser amounts of sandstone and limestone (Krumme, 1981; Visher, 1996). Proportion of sandstone and limestone within the Desmoinesian section increases north and northwestard away from the Ouachita thrust-front (Bush and others, 1977; Krumme, 1981; Visher, 1996). In Arkansas, the Desmoinesian beds have a combined maximum thickness of 617 m (Bush and others, 1977). The Desmoinesian strata of the Arkoma basin correlate with the Deese Group strata that crop out across the southwestern part of the Arbuckle uplift, and farther south in the Criner Hills uplift (Plates 2.1 and 2.2).
The age of the base of the Atoka Formation in the eastern Arkoma basin is controversial, and is placed traditionally at the base of the Atokan; however, some place the age at latest Morrowan (Gatewood and Fay, 1991; C.G. Stone, personal communication, 2000). In the eastern part of the Arkoma basin (and within the western frontal belt of the Ouachita Mountains), the Chickachock Chert separates the basal Atoka beds from underlying Bloyd Shale (Plate 2.2) (Stone and others, 1973). Farther west, the quartzose Spiro Sandstone separates the Atoka Formation from underlying strata (Plate 2.2). In the western part of the Arkoma basin where the Spiro Sandstone rests upon the Wapanucka Limestone, the contact between the two formation has been demonstrated to be unconformable (Shelton, 1996). No well has penetrated the base of the autochthonous Atoka Formation beneath the allochthonous eastern frontal belt of the Ouachita orogen in Arkansas. Although the age of the base of the Atoka Formation in the eastern part of the Arkoma basin cannot be definitely placed in the latest Morrowan, a regional unconformity in the Morrowan is established for the Ouachita foreland west of Arkansas (Plate 2.2).

**Southern Oklahoma aulacogen (Wichita and Arbuckle uplifts)**

Beds of the Atoka Formation and overlying Desmoinesian strata of Arkansas are laterally traceable as far west as the eastern flank of the Arbuckle uplift (Plate 2.1) (Ham and others, 1954). The Atoka Formation thins abruptly along the northeastern flank of the Arbuckle uplift where the Atoka Formation is progressively truncated beneath an unconformity at the base of the Desmoinesian succession of rocks (Ham, 1973). The Atoka Formation and younger rocks are absent from the crests of several basement uplifts that subdivide the broader Arbuckle uplift. Atoka Formation and overlying Desmoinesian through Missourian synorogenic strata crop out in several deep synclines (grabens) on the northeastern side of the Arbuckle uplift (Ham and others, 1954). Except for the basal part, the Desmoinesian and Missourian sequence of rocks that crop out on the northeastern flank of the Arbuckle uplift are locally conglomerates (Ham and others, 1954). The Missourian section includes the Seminole and Francis conglomerates which incorporate limestone clasts derived from the Arbuckle uplift (Ham and others, 1954).
The limestone content of the Desmoinesian stratigraphic section increases abruptly from north to south across the Arbuckle uplift and Ardmore basin (Plate 2.1) (Ham and others, 1954; Cooper, 1995). Whereas the Desmoinesian section is mostly clastic facies northeast of the Arbuckle uplift, increased proportions of limestone conglomerates and fossiliferous limestones crop out in the southeastern parts of the Arbuckle uplift (Ham and others, 1954). Across the southern Arbuckle uplift, the Desmoinesian section is called the Deese Group (Ham and others, 1954) (Plate 2.2). Farther south in the Criner Hills (Plate 2.1) and in the subsurface of the Ardmore basin, the Deese Group is interbedded limestone and red and gray shale (Cooper, 1995; Ham and others, 1954). South of the Arbuckle uplift, the Missourian stratigraphic section is broadly assigned to the Hoxbar Group (Plate 2.2) (Cooper, 1995).

Within the Southern Oklahoma aulacogen, predominantly west of the Arbuckle uplift, upper Pennsylvanian (Virgilian) clastic rocks are unconformable above Atoka Formation (and younger Pennsylvanian rocks). South-to-north across the Arbuckle uplift, Virgilian rocks grade from conglomerates to shales (Ham and others, 1954). The transition from Virgilian conglomerates (Collings Ranch, Ada, Vanoss) to Virgilian and Permian shales marks the onset of the post-“Arbuckle” orogenic phase of deposition in the Southern Oklahoma aulacogen (Plates 2.1 and 2.2). Although the Collings Ranch Conglomerate crops out within a tightly folded and faulted structure within the Arbuckle uplift, the overlying Ada and Vanoss conglomerates are generally undeformed and dip toward the northwest. Virgilian strata rest unconformably upon more tightly folded Cambrian to Desmoinesian rocks (Ham and others, 1954). The base of the post-“Arbuckle” orogenic phase is time-transgressive from north-to-south and varies from Virgilian (generally base of Ada Formation and Vanoss Group) across the northwestern Arbuckle uplift to late Wolfcampian (base of Wichita Group) in central and southwestern Texas (Plates 2.1 and 2.2).

**Ardmore basin and Muenster arch**

The Atoka Formation that crops out northeast of the Arbuckle uplift correlates with the upper part of the Dornick Hills Group which crops out in the cores of tight
synclines on the western flank of the Hunton arch and farther south on the northern flank of the Ardmore basin (Plates 2.1 and 2.2) (Ham and others, 1954). The Dornick Hills Group is subdivided into Morrowan “Wapanucka” strata and Atokan “Atoka” strata (Ham and others, 1954). Along the southern flank of the Arbuckle uplift and within the Ardmore basin, the Morrowan part of the Dornick Hills has a much greater proportion of sandstone beds which contrasts with the limestones of the Wapanucka (Ham and others, 1954; Cooper, 1995).

The Dornick Hills Group, consisting of dark-colored shales, sandstones, and thin limestones, is laterally discontinuous across the Arbuckle uplift, absent across structurally high areas and is thickest within the cores of synclines (Ham and others, 1954; Fay, 1986b). In general, the Dornick Hills Group thickens southward toward the Ardmore basin and has a maximum thickness of 1368 m (Ham and others, 1954; Fay, 1986b; Cooper, 1995). Along the southern flank of the Arbuckle uplift, an unconformity separates the base of the Dornick Hills Group from the top of the Springer Group sandstones and shales (Cooper, 1995; Ham, 1973). The unconformity at the base of the Dornick Hills Group is even more pronounced across the Criner Hills arch southwest of the Arbuckle uplift (Cooper, 1995) (Plates 2.1 and 2.2). Well data indicate that the unconformity at the base of the Dornick Hills Group extends from northeast to southwest across the Criner Hills uplift (Cooper, 1995). Because of lack of well data, it is uncertain how far the unconformity extends southeast toward the deeper parts of the central Ardmore basin (Flawn and others, 1961; Ham, 1973; Cooper, 1995). Farther south toward the crest of the Muenster arch, Dornick Hills Group rocks rest on progressively older rocks. Well data within the southeastern part of the Muenster arch show Morrowan through Atokan rocks (Dornick Hills-Atoka/“grey” Strawn) resting above Middle Ordovician calcareous sandstones and limestones (Plates 2.1 and 2.2) (Bradfield, 1957a-c).

Surface geology within the Criner Hills and well data for surrounding regions show that an unconformity separates the top of the Dornick Hills Group from a thick succession of Desmoinesian shales, sandstones, and limestones classified as the Deese Group (Plates 2.1 and 2.2) (Miser, 1954; Ham, 1973; Cooper, 1995). Southwest of the
Criner Hills uplift, toward the crest of the Muenster arch, Deese Group strata rest unconformably upon Mississippian and older formations (Plates 2.1 and 2.2) (Cooper, 1995). In comparison to the underlying Dornick Hills Group, the Deese Group contains a greater proportion of limestone (Cooper, 1995). The Deese Group thickens southeastward toward the Ouachita thrust front, and more abruptly northeastwards towards the Arbuckle uplift (Cooper, 1995). Across the Ardmore basin (Plate 2.1), the thickness of the Deese Group ranges from 608 m to 1520 m (Cooper, 1995). Maximum regional thickness of the Deese Group is estimated to be 2965 m (Fay, 1986b). South of the Ardmore basin, Deese Group rocks correlate with the reddish-colored upper part of the Strawn Group (“red” Strawn) (Plate 2.2).

Conformably above the Deese Group within the Ardmore basin is the Missourian Hoxbar Group (Plates 2.1 and 2.2) (Cooper, 1995). The Hoxbar Group is predominantly shale with lesser amounts of sandstone and limestone (Cooper, 1995). Conglomerates are more abundant in the Hoxbar Group than in the underlying Deese Group, and the upper part of the Hoxbar Group contains thin coal beds (Ham and others, 1954; Ham, 1973). The Hoxbar Group thickens southeastward toward the Ouachita thrust front, and northeastwards towards the Arbuckle uplift suggesting provenance from both locations (Cooper, 1995; Cleaves, 1997). Hoxbar Group strata also thin abruptly across subsurface basement uplifts along the Criner Hills uplift (Cooper, 1995). Across the northwestern Ardmore basin, thickness of the Hoxbar Group ranges from 304 m to 912 m (Cooper, 1995). The Hoxbar Group reaches a maximum thickness of 1216 m in the southeastern part of the Arbuckle uplift and Ardmore basin (Fay, 1986b). South of the Ardmore basin, the Hoxbar Group strata correlate with the Canyon Group strata which crop out south of the Muenster arch in central Texas (Plates 2.1 and 2.2).

Capping the synorogenic phase of deposition within the Ardmore basin is the Virgilian age Cisco Group (Plates 2.1 and 2.2). Within the Ardmore basin, a limestone marks the base of the Cisco Group (Cooper, 1995). The Cisco Group is predominantly gray shale and sandstone, with minor limestone beds (Ham, 1973). The Cisco Group is thickest within cores of basement synclines that flank the Criner Hills uplift,
Hills uplift (Plate 2.2) (Cooper, 1995). Cisco Group rocks are absent along the Criner Hills uplift and several other subsurface basement uplifts within the Ardmore basin (Cooper, 1995). The maximum thickness of the Cisco Group within the northwestern Ardmore basin is 608 meters (Cooper, 1995). An unconformity separates the Cisco Group from overlying Permian rocks (Cooper, 1995).

Lower Virgilian fine-grained shales and limestones common to the Cisco Group of the Ardmore basin are absent farther north (Plate 2.2). North of the Ardmore basin, across the Arbuckle uplift, Virgilian limestone conglomerates (Collings Ranch Conglomerate and lower Vanoss Group) rest unconformably above Desmoinesian and older rocks (Ham and others, 1954). Lower Vanoss Group limestone conglomerates grade westward into a limestone conglomerate at the base of the Permian Stratford Formation (Ham and others, 1954). The Upper Vanoss Group shales are locally arkosic and grade westward into the Permian Stratford Formation (Ham and others, 1954). The gradational change from Pennsylvanian (Virgilian) Vanoss to Permian (Wolfcampian) Stratford rocks suggests the unconformity at the base of the Vanoss in the Arbuckle uplift is the time-transgressive equivalent of the unconformity at the base of the Permian in the Ardmore basin (Plate 2.2) (Ham and others, 1954; Cooper, 1995).

**Llano uplift**

Northeast of the Llano uplift, in the Fort Worth basin (Plate 2.1), the Strawn Group is separated into lower (“grey”), and middle and upper (“red”) parts (Plate 2.2) (Bradfield, 1957a-c; Johnson and others, 1988). The lower “grey” Strawn (late Morrowan-Atokan) roughly correlates with, and is genetically related to, the Atoka “Group” of north-central Texas (Dornick Hills Group along the southeastern Muenster arch, Ardmore basin, and southern Arbuckle uplift) (Brown and others, 1973) (Plates 2.1 and 2.2). Few wells penetrate lower Strawn Group rocks in deeper parts of Fort Worth basin (Johnson and others, 1988). On the basis of distribution of Atokan aged rocks in north-central Texas, Yancey and Cleaves (1990) suggest that the lower Strawn rocks have northern provenance and are related to uplift along the Muenster arch.
The middle and upper “red” Strawn correlates with the Deese Group of southeastern Oklahoma, and is considered genetically related to Canyon Group of north-central Texas (Plates 2.1 and 2.2) (Brown and others, 1973). The term “red” applied to the middle and upper Strawn Group is derived from an observed increased proportion of red-colored arkosic clasts likely derived from eroded basement uplifts located farther north (Bradfield, 1957a-c; Yancey and Cleaves, 1990; Cleaves, 1996). However, cross sections constructed across the Ardmore basin and distribution of Desmoinesian-Virgilian rocks in north-central Texas suggest that subsidence of the Muenster arch began in the Desmoinesian (Yancey and Cleaves, 1990; Cooper, 1995; Cleaves, 1996). Yancey and Cleaves (1990) suggest that arkosic rock fragments within the middle and upper Strawn Group of the northern Fort Worth basin were derived from the Criner Hills arch (Plate 2.1) and possibly areas farther north. “Ouachita facies” rock fragments and chert clasts included within middle and upper Strawn Group rocks suggest uplift and erosion of the Ouachita orogen east of the Fort Worth basin during the Desmoinesian (Bradfield 1957 a,b; Yancey and Cleaves, 1990; Cleaves, 1996).

Within the Fort Worth basin, Strawn Group rocks are predominantly fluvial-deltaic and slope mudstone and sandstone with localized limestones (Figure 2.7) (Kier and others, 1979; Johnson and others, 1988). The boundary between the Strawn Group and overlying Canyon Group is gradational, but marked by an increase in proportion of limestone (Figure 2.7) (Cleaves, 1982). Along the crest of the Bend arch, and farther west in north-central Texas, the relatively thin latest Morrowan-Atokan Smithwick Shale separates Ordovician-Pennsylvanian (Morrowan) carbonates of the Caddo shelf carbonate system from Desmoinesian-Permian (latest Wolfcampian) cyclic clastic rocks and limestones (Figure 2.7, and Plates 2.1 and 2.2) (Kier and others, 1979; Johnson and others, 1988).

West of the Bend arch, the Smithwick Shale interfingers with carbonates of the Caddo shelf carbonate system (Figure 2.7 and Plate 2.1). The upper part of the Caddo shelf carbonate system correlates with the late Morrowan-Atokan Dornick Hills Group of the Ardmore basin and Arbuckle uplift regions of southeastern Oklahoma (Figure
The cross section through the Fort Worth basin (Figure 2.7) shows that the base of the Strawn Group rests unconformably upon the Smithwick Shale and upper Caddo shelf carbonate system in north-central Texas. Desmoinesian (Strawn Group) rocks that crop out adjacent to Llano uplift contain pebbles of Marble Falls Limestone (Chesterian-Morrowan) encapsulated by prodelta shale (presumably Morrowan-Atokan-aged Smithwick Shale) (Johnson and others, 1988). Inclusion of Morrowan-Atokan-aged rock fragments (Smithwick Shale) within the Desmoinesian Strawn Group suggest an erosion surface with minimal time gap.

West of the Bend arch, the Strawn Group maintains a relatively constant thickness of 304 m in north-central Texas (Figure 2.7) (Plate VIII in Cheney, 1929; Kier and others, 1979; Johnson and others, 1988). East of the Bend arch, the Strawn Group thickens abruptly to a maximum that exceeds 2432 m near the Ouachita thrust front (Plate VIII in Cheney, 1929; Kier and others, 1979; Johnson and others, 1988). The abrupt thickening of the Strawn Group adjacent to the Ouachita thrust belt indicates significant subsidence of the Fort Worth basin prior to overthrusting. West of the Bend arch the Desmoinesian Strawn Group strata dip to the west and are subparallel with underlying Ordovician to Atokan strata; however, east of the Bend arch, Desmoinesian (middle and upper Strawn Group) and younger rocks dip to the west, whereas Atokan (Smithwick Shale) and older rocks dip to the east (Figure 2.7) (Kier and others, 1979; Johnson and others, 1988).

Although limited well data make stratigraphic analysis of the deep Fort Worth basin imprecise, the lower Strawn Group apparently represents a transition between marginal- to deep-marine clastic deposition to shallow-water fluvial-deltaic deposition (Figure 2.7). A transition between south- to southwest-directed sediment transport and more westerly directed sediment transport is evident within the lower Strawn Group (Atokan) in the deep eastern part of the Fort Worth and Strawn basins (Figure 2.7 and Plate 2.1) (Cleaves, 1996). The combination of southwest-directed sediment transport in the eastern Fort Worth basin, which is parallel to strike of Ouachita orogen farther east, and inclusion of Ouachita-facies chert clasts in the overlying Desmoinesian part of the Strawn Group suggests that the lower Strawn Group (Atokan) was deposited in
a restricted and shrinking foreland basin sandwiched between the Bend arch to the west and an uplift located to the east within the Ouachita orogen (Figure 2.8 and Plate 2.1).

The Atokan-age rocks that crop out within the Ouachita Mountains, across the Arkoma basin, southern Ozark dome, and Arbuckle uplift are classified as Atoka Formation (as in Johnson and others, 1988) (Plates 2.1 and 2.2). In the subsurface of north-central Texas, Atokan-age rocks are classified as Atoka Group (both clastic-facies “Atoka” and shallow-water limestone facies “Caddo” type lithologies) (Figure 2.7) (Thompson, 1982; Cleaves, 1996). In deeper parts of the northeastern Fort Worth basin, the “Atoka Group” clastic rocks are also recognized as “grey” Strawn (Figure 2.7) (Bradfield, 1957a-c). For purpose of discussion, the clastic facies “Atoka Group” rocks of the northeastern Fort Worth basin are classified hereafter as “grey” Strawn; and the Atokan-aged clastic rocks of the Ouachita Mountains as Atoka Formation.

To the northeast in Arkansas, paleocurrent analyses of the marginal- to deep-marine Atoka Formation (latest Morrowan-Atokan) also suggest a sediment transport direction parallel to the Ouachita orogen (Figure 2.8) (Lowe, 1989). Paleocurrents in the lower grey Strawn strata show westward sediment transport, and southwestward sediment transport in middle and upper grey Strawn strata (Figure 2.8) (Sullivan, 1966). Paleocurrents in Atoka Formation strata in the Arkoma basin of Oklahoma suggest southwestward sediment transport similar to that of the lower Strawn Group of the eastern Fort Worth basin (Briggs and Cline, 1967). Although the inferred sediment transport direction of the Atoka Formation in the western Arkoma basin is similar to that of the lower Strawn Group in the eastern Fort Worth basin, unconformities along the Muenster arch, Criner Hills uplift, and Arbuckle uplift and different lithology of clasts suggest that the two basins were not connected (at least not completely) (Figure 2.8) (Ham, 1973; Yancey and Cleaves, 1990; Cooper, 1995). Control of stratigraphic variation of Paleozoic strata decreases abruptly beneath the Mesozoic Gulf Coastal Plain (Plate 2.1). On the basis of inferred paleocurrents, the easternmost part of the Fort Worth basin may be a southern extension of a sediment trough that during the late Morrowan through Atokan extended from north-central
Texas (eastern Fort Worth basin), through southeastern Oklahoma and central Arkansas (southern Arkoma basin and frontal Ouachita orogen), eastward towards the Appalachian orogen (Figure 2.8). Similar to the lower part of the Strawn Group in the eastern Fort Worth basin, the Atoka Formation of Arkansas is thickest in a trough sandwiched between a relative basement high to the north and an uplift located within the Ouachita orogen to the south (Benton uplift) (Figure 2.9 and Plate 2.1).

Conformably above the Strawn Group in north-central Texas rests a Pennsylvanian (Missourian) through Permian (Wolfcampian) cyclic succession of shallow-marine shelf and shelf-edge limestones, and fluvial-deltaic and slope mudstone and sandstone (Figure 2.7). The boundary between the upper Strawn Group and the lower Canyon Group is gradational (Figure 2.7). The Missourian to Wolfcampian strata dip westward and are broadly subdivided into the Canyon Group (Missourian), and the Cisco Group (Wolfcampian) (Figure 2.7). In north-central Texas, the Canyon through Cisco Group succession consists of westward-prograding shelf-carbonate banks and marine slope sediment fans and has a maximum cumulative thickness of 600 meters (Canyon Group is approximately 200 m thick) (Figure 2.7) (Kier and others, 1979; Johnson and others, 1988; Cleaves, 1996).

North of the Muenster arch in the southeastern parts of the Arbuckle uplift and Ardmore basin, the correlative of the Canyon Group (the Hoxbar Group) thickens abruptly to a maximum which exceeds 1200 m (Fay, 1986b). The overlying cyclic limestone/clastic rocks of the Virgilian part of the Cisco Group also thicken towards the center of the Ardmore basin. North of the Ardmore basin, Cisco Group strata grade into the shales, arkosic sandstones, and arkosic conglomerates of the Vanoss and Ada formations (Ham and others, 1954; Cooper, 1995). In the Ardmore basin, Permian (post-“Arbuckle” orogenic) rocks of the Pontotoc Group rest unconformably above the Virgilian Cisco Group (Plates 2.1 and 2.2) (Cooper, 1995). South of the Ardmore basin, Wolfcampian beds of the Pontotoc Group grade into Wolfcampian-age Cisco Group. Farther south in central Texas, west of the Llano uplift, the Wichita Group (late Wolfcampian to Guadaloupian) rests conformably above the Cisco Group.
F) Rapid deposition phase of the deep-marine facies syn-orogenic clastic wedge—Johns Valley Shale and the Atoka Formation of the Ouachita Mountains (late Morrowan-Atokan)

The Morrowan Johns Valley Shale rests stratigraphically between overlying Atoka Formation and underlying Jackfork Group rocks (Plate 2.3). Across the northwestern part of the Ouachita Mountains, the Johns Valley Shale contains clasts and boulders of predominantly shallow-water carbonates ranging in age from Late Cambrian to Pennsylvanian (Figure 2.10) (Shideler, 1970). The shallow-water carbonate clasts are primarily “Arbuckle facies,” with lesser amounts of “Ozark facies” clasts (Shideler, 1970). No direct evidence exists for source area (or areas) of the clasts incorporated in the Johns Valley Shale (see Shideler, 1970; and Viele and Thomas, 1989; for further discussion of the “Johns Valley boulder” problem).

The shallow-water carbonate clasts are likely derived from a source area that is now beneath the frontal zone of the Ouachita thrust belt (possibly a submarine canyon or fault scarp). Because most of the Johns Valley “boulders” are “Arbuckle facies” carbonates, an area east of the Arbuckle Mountains is a likely source for most of the Johns Valley boulders. A regional unconformity at the base of the late Morrowan and Atokan Dornick Hills Group and shallow-water facies Atoka Formation across the foreland adjacent to the Ouachita thrust-belt in Oklahoma suggests uplift of the Arbuckle uplift region during the Morrowan (Cooper, 1995).

Besides shallow-water carbonates, lesser quantities of “Ouachita facies” clasts are also found within the Johns Valley Shale (Shideler, 1970). Ouachita facies clasts include: Wapanucka-Chickachoc Chert (Morrowan), Arkansas Novaculite (Devonian-lower Mississippian), Pine Top Chert (Devonian), Polk Creek Shale (Ordovician), and Bigfork Chert (Ordovician) (Shideler, 1970). If the Ouachita facies clasts were deposited from subaerial or submarine sources prior to overthrusting, the Wapanucka-Chickachoc and Pine Top Chert clasts were likely derived from an autochthonous facies-transition zone located in the subsurface southeast of the frontal thrust fault of the Ouachita thrust belt, and the older Arkansas Novaculite and Bigfork Chert were derived from an autochthonous source area located still farther southeast. The
Arkansas Novaculite and Bigfork Chert clasts may also have been derived from subaerial or submarine allochthonous sources and transported northward. Ouachita facies clasts in Desmoinesian Strawn Group of the eastern Fort Worth basin in north-central Texas were likely derived from erosion of a similar (but younger) uplift within the Ouachita orogen (Yancey and Cleaves, 1990).

The deep-water facies Atoka Formation has the greatest maximum thickness and fastest estimated rate of deposition of all the Mississippian to Atokan synorogenic rocks that crop out within the Ouachita Mountains (Morris, 1974). The estimated rate of deposition of 912 m per m.y. for the Atoka Formation dwarfs the estimated rates for the underlying Stanley Group through Johns Valley Shale (range from 167 m per m.y. to 114 m per m.y.) (Morris, 1974).

Along the frontal zone of the Ouachita thrust belt, shallow- to marginal-marine Atoka Formation grades southward into deep-marine Atoka Formation. The deep-marine Atoka Formation consists of interbedded shale and sandstone turbidites (Morris, 1974; Lowe, 1989). In eastern parts of the frontal zone in Arkansas, the Atoka Formation thickens abruptly to a maximum of at least 8360 meters (Bush and others, 1977). The maximum thickness of deep-marine Atoka Formation in southern parts of the Ouachita Mountains exceeds 1520 m (Walthall, 1967). Seismic data suggests that the Atoka Formation thickens to nearly 3040 m in the subsurface south of the Ouachita Mountains (Nicholas and Waddell, 1989). The lower part of the deep-marine Atoka Formation strata correlate with the upper part of the Pottsville Formation of the Black Warrior basin (Plates 2.1, 2.2, and 2.3) (Thomas, 1977).

**G) Post “Ouachita orogenic” rocks of the Ouachita foreland (late Virgilian-Permian)**

For simplicity, discussion of post-orogenic stratigraphy of the Ouachita foreland is limited to part of the Permian of the north-central Texas and Oklahoma region. An attempt is made here to separate rock strata formed during imbrication and emplacement of the Ouachita orogen from rock strata deposited after emplacement. The age of the transition from the “Ouachita synorogenic” phase to the “post-Ouachita orogenic” phase of deposition is inexact, in many places gradational, and ranges from...
late Virgilian to Wolfcampian. The attempt to correlate and separate the stratigraphy into these two groups is further complicated because conglomeratic facies derived from basement uplifts within the Southern Oklahoma aulacogen (Plate 2.1) locally interfinger with the finer grained Virgilian and Wolfcampian strata which are derived from the Ouachita orogen.

In the Arbuckle Mountains of southern Oklahoma, onset of the “post- Ouachita orogenic phase is marked by deposition of Virgilian conglomerates and shales of the Vanoss Group that unconformably overlie all older rocks (Plates 2.1 and 2.2) (Ham and others, 1954). In northern parts of the Arbuckle Mountains, the slightly younger Ada Formation (Virgilian) may represent the earliest onset of the post-orogenic phase in the region. On the western flank of the Hunton arch, Vanoss Group rocks are not folded and dip westward; however, farther south, the Vanoss Group rocks are folded around the Arbuckle anticline (Figure 2.11). The conglomeratic facies of the Vanoss Group are part of a Morrowan to Wolcampian succession of limestone and arkosic conglomerates located on the surface and in the subsurface adjacent to the Arbuckle-Wichita-Amarillo uplifts (Plates 2.1 and 2.6) (Adler and others, 1971; Johnson and others, 1988; Yancey and Cleaves, 1990; Cleaves, 1996).

In the subsurface west of the Arbuckle Mountains and north of the Wichita Mountains, the Virgilian Vanoss Group conglomerates grade into limey shales; and farther west into limestones (Rascoe and Adler, 1983; Johnson and others, 1988). Although an unconformity in the Virgilian at the base of the Vanoss Group of the Arbuckle Mountains is clearly defined, no unconformity is clearly defined within the Virgilian strata of the Anadarko basin (Plate 2.6). A north-elongated area of “starved-basin” deep-water shales is located in the subsurface north of the Amarillo arch (Rascoe and Adler, 1983, Johnson and others, 1988). West of the eastern Texas-Oklahoma Panhandle, Virgilian “starved-basin” shales grade into limestones (Rascoe and Adler, 1983, Johnson and others, 1988). Still farther west, Virgilian limestones grade into clastic rocks with western provenance (Rascoe and Adler, 1983; Johnson and others, 1988).
West of the Arbuckle Mountains (Figure 2.11 and Plate 2.1), Virgilian Vanoss Group strata grade up-section into the Permian (Wolcampian) Stratford Formation (Ham and others, 1954). Wolcampian rocks located along the northern flank of the Wichita uplift and northwestern flank of the Arbuckle uplift are predominantly limestone and arkosic conglomerates (Figure 2.11 and Plate 2.6) (Ham and others, 1954). Farther northwest in central Oklahoma, the Wolcampian conglomerate facies grades into a red sandstone and shale facies (Stratford Formation-Pontotoc Group) (Ham and others, 1954; Johnson and others, 1988). Still farther west, towards the northwestern Oklahoma-Texas Panhandle region, the Wolfcampian red sandstones and shales grade into limestones (Plate 2.6). In this part of the Anadarko basin (Plate 2.1), the Wolfcampian is separated into three Groups (Admire, Council Grove, and Chase) (Johnson, 1978; Hills and Kottlowski, 1983; Johnson and others, 1988).

Upper Wolfcampian limestones cover much of central Kansas, central Oklahoma, and north-central Texas (Dutton and others, 1982; Rascoe and Adler, 1983; Johnson and others, 1988; Cleaves, 1996). These limestones grade into clastic rocks along the Wichita-Amarillo uplifts (Plate 2.6) and towards eastern Oklahoma (Rascoe and Adler, 1983; Johnson and others, 1988).

South of the Wichita-Amarillo uplift trend in the Palo Duro basin, a relatively thin layer of late Wolfcampian limestones separate underlying interbedded limestones and shales from overlying Leonardian evaporite rocks (Dutton and others, 1982; Johnson and others, 1988) (Plates 2.1, 2.2, and 2.6). The Pennsylvanian-Wolfcampian interbedded limestones and shales roughly correlate with the Strawn/Canyon and Cisco Groups of the Fort Worth basin (Figure 2.7 and Plate 2.2). In contrast to the Strawn/Canyon-Cisco Group strata of the Fort Worth basin that prograde westward, subsurface facies distributions of the Pennsylvanian-Wolfcampian limestones and clastic rocks in the Palo Duro basin suggest a southward sediment transport direction (away from the Amarillo-Wichita uplift) (Figure 2.7 and Plate 2.6) (Handford and Dutton, 1980; Johnson and others, 1988; Yancey and Cleaves, 1990; Cleaves, 1996). The maximum thickness of the Pennsylvanian-Wolfcampian (Strawn-Cisco correlatives) in the Palo Duro basin is approximately 1000 m (Handford and Dutton,
Farther north in the Anadarko basin, the Pennsylvanian to Wolcampian succession is nearly twice as thick (Plate 2.6).

Across northwest Texas, western and central Oklahoma, and Kansas, Leonardian evaporite rocks rest in stark contrast above late Wolcampian limestones (Plates 2.1 and 2.6). The abrupt change from limestone to evaporites suggests an increase in regional aridity and relative sea level fall (Johnson and others, 1988). North of the Amarillo-Wichita uplift trend in the Anadarko basin, the basal Leonardian anhydrite is called the Wellington Formation (Plate 2.6) (Johnson, 1978; Hills and Kottlowski, 1983; Johnson and others, 1988). Resting above the Wellington Formation is an interbedded succession of red-colored sandstone, shale, halite, and gypsum beds (Plate 2.6) (Johnson, 1978; Hills and Kottlowski, 1983; Johnson and others, 1988). In the Anadarko basin, Leonardian strata above the Wellington Formation are broadly assigned to the Henessy Group (Plates 2.1, 2.2, and 2.6). Conformable above the Henessy Group rest Guadalupian, and locally Ochoan, evaporite and red-colored clastic rocks (Plates 2.2, and 2.6) (Johnson, 1978; Hills and Kottlowski, 1983; Johnson and others, 1988). Farther east in the Arbuckle uplift, post-Hennesy Group strata are absent (Plates 2.1, and 2.2). The Leonardian to Guadalupian-Ochoan rocks in the Anadarko basin are thickest (≈1200 m) in the Texas-Oklahoma Panhandle on the northern flank of Amarillo uplift (Plate 2.6).

South of the Amarillo uplift in the Palo Duro basin, the correlative of the Wellington Formation anhydrite is the Wichita Group and Red Cave Formation (Plates 2.1, 2.2, and 2.6). Whereas the Wellington Formation is mostly anhydrite, the Wichita Group and Red Cave Formation are predominantly evaporitic dolomite with lesser anhydrite (Johnson and others, 1988). The Red Cave Formation is the basal formation of the Clear Fork Group (Nicholson, 1960; McGillis and Presley, 1981; Johnson and others, 1988). The Clear Fork Group is a Leonardian-early Guadalupian succession of interbedded salt, evaporitic dolomite and red-colored clastic rocks (Plate 2.6) (Nicholson, 1960; McGillis and Presley, 1981; Johnson and others, 1988). The Clear Fork Group of the Palo Duro basin roughly correlates with the Henessy Group of the Anadarko basin (Plates 2.1, 2.2, and 2.6).
Sequentially above the Clear Fork Group rests the San Andres Formation anhydrite (Nicholson, 1960; McGillis and Presley, 1981; Johnson and others, 1988). Northward toward the Amarillo uplift, the San Andres Formation thins and is called the Blaine Formation (Plates 2.2 and 2.6) (Nicholson, 1960; McGillis and Presley, 1981; Johnson and others, 1988). Guadalupian-Ochoan strata that rest conformably above the San Andres-Blaine anhydrite are broadly classified as the post-San Andres interval and post-Blaine red beds (Nicholson, 1960; McGillis and Presley, 1981; Johnson and others, 1988). Southward from the Amarillo uplift, the post-Blaine red beds (red-colored clastic rocks) thicken and interfinger with evaporite rocks of the post-San Andres interval (Plates 2.2 and 2.6).

North of the Amarillo-Wichita uplifts, Tertiary clastic rocks of the Ogallala Formation rest unconformably upon Guadalupian-Ochoan rocks (Plates 2.2 and 2.6). South of the Amarillo-Wichita uplifts, southward-thickening Triassic red-colored shales rest above Permian evaporite rocks (Plates 2.1 and 2.6). East of the Palo Duro basin, Mesozoic rocks are absent across the Bend arch (Figure 2.7 and Plate 2.1). Farther east in the Fort Worth basin, Cretaceous rocks thicken eastward towards the Ouachita thrust front (Figure 2.7). Mesozoic strata thicken eastward and southward towards Louisiana and the Texas Gulf Coast (Ewing, 1991).

**H) Post orogenic rocks southeast of the Ouachita thrust-front**

**Texarkana platform**

Several wells located south of the Ouachita Mountains in northeast Texas, southern Arkansas, and northern Louisiana have penetrated Desmoinesian through Permian rocks (Paine and Meyerhoff, 1970; Vernon, 1971; Meyerhoff, 1973; Woods and Addington, 1973; Nicholas and Waddell, 1989). The Desmoinesian strata are predominantly shallow-water “platform” limestones (Nicholas and Waddell, 1989). Therefore, the name Texarkana platform has been applied to the region of northeastern Texas, southern Arkansas, and northern Louisiana where Desmoinesian carbonates have been drilled (Nicholas and Waddel, 1989). Regional seismic data indicate that these Desmoinesian-Permian rocks dip gently southward and rest unconformably upon folded Atoka Formation (Atokan) (Nicholas and Waddel, 1989). In contrast to the
Arkoma basin where Desmoinesian strata are folded, the Desmoinesian strata of the Texarkana Platform are undeformed and broken by Triassic-age extensional faults (Nicholas and Waddell, 1989, Thomas and others, 1989).

On the basis of fusulinid biostratigraphy, a regional unconformity across the Texarkana platform is placed between the Desmoinesian (and locally Missourian) interbedded marine clastic rocks and limestones and overlying Wolfcampian through Leonardian (and locally Guadalupian) succession of marine clastic rocks and limestones (Nicholas and Waddell, 1989). The Desmoinesian succession are referred to by drillers as the lower and middle part of the “Morehouse,” whereas the Permian clastic rocks are referred to as the upper “Morehouse” (Plate 2.2) (Fay and others, 1986). Correlation with formations farther west and north is imprecise. Lower and middle “Morehouse” rocks chronostratigraphically correlate with Deese Group rocks of the Ardmore basin and Arbuckle uplift and Hartshorne-Savanna Formation rocks of the Arkoma basin (Plates 2.1 and 2.2). However, these lower and middle “Morehouse” limestones may be diachronous equivalents of “post-orogenic” Wolfcampian limestones found farther west in the subsurface of the Anadarko and Palo Duro basins (Plates 2.1, 2.2, and 2.6).

The upper “Morehouse” (Wolfcampian-Guadalupian) marine clastic rocks and limestones of the Texarkana platform are chronostratigraphic equivalent to Wolfcampian-Guadalupian limestones, evaporites and clastic rocks of the Anadarko and Palo Duro basins (Plates 2.1, 2.2, and 2.6). The “Morehouse” strata are evidently a deeper-water, marine-facies equivalents of Wolfcampian-Guadalupian strata found west of the Arbuckle Mountains and Fort Worth basin. The maximum drilled thickness of Desmoinesian-Permian rocks in the Texarkana Platform is 2195 m; however most wells have penetrated less than 300 m (Nicholas and Waddell, 1989).

**Sabine uplift**

Eight wells in the Sabine uplift area of eastern Texas and northwestern Louisiana have drilled Desmoinesian through Wolfcampian rocks. Desmoinesian limestones rest nonconformably above a 600 m-thick layer of rhyolite porphyry and tuff (Nicholas and Waddell, 1989). The rhyolite/tuff layer rests unconformably upon
deformed Mississippian-Pennsylvanian turbiditic sandstones and shales (flysch) (Nicholas and Waddell, 1989). Age and correlation of the rhyolite/tuff is unclear. Nicholas and Waddell (1989) correlate the rhyolite/tuff layer with Mississippian-aged lower Stanley Group tuffs of the Ouachita Mountains. The Rb-Sr isotopic age of 255 ± 15 Ma (Late Permian-Early Triassic) derived for the rhyolite is interpreted as a cooling age related to uplift and not emplacement (Nicholas and Rozendal 1975, Nicholas and Waddell, 1989).

The Desmoinesian carbonate and clastic succession of the Sabine uplift is approximately 200 m thick and is unconformably overlain by a much thicker Missourian through Wolfcampian succession of limestones, shales, and sandstones (Nicholas and Waddell, 1989). Missourian limestones grade upsection into Virgilian-Wolfcampian deeper-marine clastic rocks and limestones (Nicholas and Waddell, 1989). The entire Missourian-Wolfcampian section in the Sabine uplift area is approximately 1500 m thick (Nicholas and Waddell, 1989). The rhyolite/tuff layer and Desmoinesian-Wolfcampian strata dip eastwards, and were likely rotated during Mesozoic extension (Nicholas and Waddell, 1989).

Figure 2.12 compares stratigraphy and local structure of the Texarkana platform and Sabine uplift. The Desmoinesian limestones of the Sabine uplift correlate with the lower and middle “Morehouse” strata of the Texarkana platform (Figure 2.12 and Plate 2.2). The Wolfcampian strata of the Sabine uplift roughly correlate with the Wolfcampian-Guadalupian upper “Morehouse” strata of the Texarkana Platform (Figure 2.12 and Plate 2.2). Except for one well, Missourian-Virgilian strata are absent across the Texarkana platform (Nicholas and Waddell, 1989). Desmoinesian-Wolfcampian strata of the Sabine uplift also correlate with Strawn Group through Cisco Group strata of the Fort Worth basin and Bend arch regions of north-central Texas (Figures 2.7, 2.12, and Plates 2.1 and 2.2).

The Texarkana platform and Sabine uplift also have differently shaped Mesozoic unconformity surfaces. Across the Sabine uplift, Jurassic rocks drape over a structural high on the underlying Paleozoic section (Figure 2.12). In contrast, seismic data in part of the Texarkana Platform show Triassic through Cretaceous strata nearly
parallel with underlying Desmoinesian-Permian rocks (Figure 2.12). Although well control for the Paleozoic section is imprecise across eastern Texas and Louisiana, the different geometries of the Mesozoic unconformity illustrated in Figure 2.12 show an irregularly shaped pre-Mesozoic surface.
Figure 2.1: Comparison between Late Cambrian basal sedimentary strata of the Slick Hills, southwestern Oklahoma and the Llano uplift, central Texas.
Figure 2.2: Part A is a cross section of the Black Warrior basin showing generalized stratigraphy of the Cambrian to early Mississippian passive-margin carbonate shelf strata. Part B is a map of the Black Warrior basin and adjacent structures. Location of cross section is shown.

Figure 2.3: Pre-Hunton Group Cambrian-Late Ordovician strata of the Ouachita orogen foreland of northeast Texas, southeast Oklahoma, and western Arkansas.
Shown above is a stratigraphic cross section of the Cambrian-early Mississippian (Meramecian) sedimentary rocks of the region that extends from the Arbuckle Mountains (column 1), across the Ouachita Mountains (columns 2, 3, and 4'), to the eastern Arkoma basin (5). In the cross section, deep-water facies strata above the basal detachment (2', 3', and 4') are placed in approximate restored position above autochthonous strata. Selected timelines are shown to correlate shelf facies strata (1 and 6) with equivalent deep-water facies. Location of a suggested transitional facies is shown.

Inset map shows the location of cross section 1-2-3'-4'-5 with reference to several regional structures. Points 2', 3', and 4' are schematically restored.

Figure 2.4 (preceding page):
Stratigraphic cross section
of the Cambrian-early Mississippian (Meramecian) sedimentary rocks
of the region that extends from
the Arbuckle Mountains, across the Ouachita Mountains, to the eastern Arkoma basin.
Schematic cross section of the South Ozark arch region of northwestern Arkansas (see Plate 2.1 for location). Pre-upper Devonian strata beneath the unconformity at the base of the Meiser/Sylamore sandstone and Woodford/Chattanooga shale succession are parallel with the basement arch. The basal Mississippian (Kinderhookian) unconformity at the base of the Boone Chert truncates Woodford/Chattanooga shale east of the South Ozark arch. Chesterian and younger strata dip westward.

Figure 2.5 (preceding page):
Schematic cross section of the South Ozark arch region of northwestern Arkansas.
From Cheney (1929), Nicholas and Rozendal (1975), Nicholas (1989), and Cleaves (1996).

**MAP LEGEND**

- **Fort Worth basin**
- **Ouachita thrust front**
- **Low-grade metamorphism (serpentinites)**
- **Luling front**
- **Waco uplift**
- **Precambrian basement**
- **Unmetamorphosed allochthonous strata**
- **Intrusive Mesozoic igneous rocks**

**FORELAND FACIES**

- **Miop (turbidites)**
- **EM (shales, distal turbidites and limestones)**
- **PFm (limestones and dolostones)**
- **Po-C (basement)**

**MISSISSIPPIAN**

- **Atokan**
- **Morrovan**
- **Mississippian-Morrovan**
- **Cambrian-Ordovician**

**DEVONIAN-EARLY MISSISSIPPIAN**

- **Oxocene**

**PERMIAN**

- **Emsian**

**MISSISSIPPIAN-MORROVIAN**

- **Shoestring limestone**

**MISSISSIPPIAN-ATOKAN**

- **Cretaceous**

**DEVONIAN-LATE MISSISSIPPIAN**

- **Cretaceous**

**EXPLANATION**

**OUACHITA FACIES**

- **Mississippian Atokan (undivided)**
- **Devonian-early Mississippian**
- **Ordovician-Silurian (undivided)**

**Structure and stratigraphy**

- **Based on cross section F-G**
- **Based on cross section F-F**

Map references: Arbenz, 1989; Nicholas, 1989; Thomas and others, 1989; Cleaves, 1996.
Figure 2.6 (preceding page):
Cross sections F-F' and G-G' of part of the Ouachita orogen and adjacent foreland of northeastern Texas and southern Oklahoma. Ordovician to early Mississippian deep-water facies strata shown southeast of Fault A in cross section G-G' are conjectural. Local subcrops of Ordovician-early Mississippian deep-water facies strata southeast of Fault A to the north and south of G-G', shown in a regional geologic map of pre-Mesozoic strata (inset map), suggest that Ordovician-early Mississippian strata rest at variable depths along Fault A.

Southeastward truncation of Ordovician-early Mississippian (cherty) strata northwest of Fault A in cross section G-G' and the apparent lack of cherty strata south of the Luling Front (according to the Shell-Barret well in the center of the Waco uplift) are enigmas. Plate 2.5 examines further the lack of cherty strata southeast of the Luling front.
Figure 2.7: Stratigraphy and generalized structure of north-central Texas. A-A' is a generalized structural cross section across the Bend arch and Fort Worth basin of northern Texas. Generalized stratigraphy of the Paleozoic succession is illustrated. The inset map shows large-scale late Paleozoic structures of northeastern Texas, southern Oklahoma, and western Arkansas. Location of cross section A-A' is illustrated.

West of the Bend arch, all Cambrian to Pennsian strata dip towards the west. East of the Bend arch, middle Desmoinesian and younger strata dip west, whereas middle Desmoinesian and older strata dip east towards the center of the Fort Worth basin. West of the Bend arch, Cambrian to middle Atokan strata are mostly carbonates; whereas, east of the arch, Meramecian and middle Morrowan to middle Atokan strata are mostly clastic facies. In the Fort Worth basin, the base of the Desmoinesian marks an abrupt transition from south-directed sediment transport in the upper Morrowan-Atokan succession to west-directed transport in younger strata. West of the Fort Worth basin, westward-prograding carbonate banks comprise the Missourian to Wolfcampian succession.

References: Morris, 1974; Johnson and others, 1988; Arbenz, 1989e; Yancey and Cleaves, 1990; Cleaves, 1996.
Figure 2.8: Maps A and B compare paleogeography and distribution of facies of the middle Pennsylvanian (early Atokan) and late Pennsylvanian (Virgilian) of northeastern Texas, Oklahoma, and western Arkansas. Inferred paleocurrent directions are also shown.

In the early Atokan (map A), terrigenous sediment derived from northern source areas east of the Nemaha ridge is deflected southeastward by, and merges with sediment derived from an eastern source. This sediment continues to flow southwestward within a narrow, deep trough sandwiched between the North American craton to the northwest and an advancing Ouachita orogen to the southeast. Sediment derived from the Muenster arch flows southeast and fills the Fort Worth basin.

By Virgilian time (map B), the front edge of an emergent Ouachita orogen has over-thrust the early Atokan shelf edge. By the Desmoinesian, the Fort Worth basin is completely filled. Desmoinesian-Virgilian rocks cover the subsiding Muenster arch and prograde westward. By Virgilian time, predominant paleocurrent direction has switched from southward to westward.

References: Johnson and others, 1988; Nicholas and Waddell, 1989; Yancey and Cleaves, 1990; Cleaves, 1996; Visher, 1996.
Figure 2.9 (preceding page):

Comparison between stratigraphy and structure of the Bend arch-Fort Worth basin, northeast Texas, and the Benton uplift-Arkoma basin, Arkansas. Part A is a cross section of the Bend arch-Fort Worth basin of northeast Texas that shows generalized lithology of the Paleozoic succession.

Part B is a cross section of the southern part of the Arkoma basin and north-central part of the Ouachita Mountains that also shows generalized Paleozoic stratigraphy. Note that the post-Morrowan succession in the southern part of the Arkoma basin is mostly clastic facies; whereas, the post-Morrowan succession in the Bend arch-Fort Worth basin (Part A) contains several carbonate units.

Part C is a map showing large-scale structures of southern Oklahoma, western Arkansas, and northeastern Texas. Area covered by Mesozoic Gulf Coastal Plain shown in light gray; outcrop of Precambrian-early Mississippian (Meramecian) in dark gray; Meramecian-Permian in white. Locations of cross sections (A and B) are illustrated.
Figure 2.10: Outcrop map of the Johns Valley Shale olistostrome and distribution of exotic clasts. Part A is a map of the western part of the Ouachita Mountains which shows outcrop of Morrowan Johns Valley Shale. In some areas, the Johns Valley contains exotic clasts and is considered an olistostrome. In other areas, exotic clasts are absent. Also shown on the map are large-scale structural features. Note that all olistostrome beds are between the Ti Valley fault and the Octavia fault. Also note that exotic clasts are absent from the Johns Valley Shale outcrops southeast of the Potato Hills and north of the Octavia fault. From Viele and Thomas (1989).

Part B shows distribution of exotic clasts within the Johns Valley olistostrome beds sorted by geologic age of clasts. The upper right panel (Pennsylvanian boulders) labels regional structures for geographic reference. Clasts are mostly shallow-water carbonates, except for marginal- to deep-water cherts (Bigfork Chert-Pine Top Chert-
Map modified from Ham, and others, 1954
Figure 2.11 (preceding page):

Geologic map of the Arbuckle uplift of southern Oklahoma showing structural subdivisions, major faults and folds. Precambrian and Cambrian igneous rocks are exposed in the central core of the southern part of the uplift. Cambrian (Franconian) to Mississippian (Meramecian) shallow-water carbonate shelf strata crop out across the remainder of the uplift. Mississippian (Meramecian) to Pennsylvanian (Missourian) clastic rocks crop out within fault-bounded synclines that flank the Arbuckle uplift.

West of the uplift, Pennsylvanian (Virgilian) to Permian conglomerates, sandstones, and shales drape unconformably above older strata. In one location, north of the Arbuckle anticline along the Washita Valley Fault (A), Virgilian conglomerates are juxtaposed against pre-Meramecian carbonates. Virgilian-Permian strata are folded around the eastern flank of the Arbuckle anticline.

Inset map (upper right) shows location of Arbuckle uplift.
A. Cross section of part of southwestern Arkansas showing unconformities at the bases of the Cretaceous, Permian (Wolfcampian), and the Desmoinesian. Desmoinesian-Guadalupian strata collectively comprise the Texarkana platform. References: Fay, and others, 1986; Johnson and others, 1988; Nicholas and Waddell, 1989.

B: Cross section of part of the Sabine uplift showing discordance between Jurassic and Carboniferous-Wolfcampian strata. Desmoinesian to Wolfcampian carbonates, sandstones, and shales, are part of the Texarkana platform. From Nicholas and Waddell (1989).

C: Map of Texarkana platform showing locations of cross sections A and B. Paleozoic outcrop of the Ouachita Mountains shown in light and dark gray shading. White area is covered by Mesozoic Coastal Plain strata. Desmoinesian-Guadalupian strata comprise the Texarkana platform. BB = Broken Bow uplift. BU = Benton uplift. References: Arbenz, 1989e; Nicholas and Waddell, 1989.

Figure 2.12: Two cross sections of different parts of the Texarkana platform showing variation in discordance between Mesozoic and pre-Mesozoic strata. Part A, from southwestern Arkansas, shows a cross section of the northeastern margin of the Texarkana platform. Part B, from the Sabine uplift of eastern Texas and western Louisiana, shows a cross section of part of the southwestern Texarkana platform. Part C shows the location of cross sections illustrated in A and B.
Chapter Three
Regional structure

The following is a systematic description of large-scale regional structural and tectonic domains within the late Paleozoic Ouachita-Marathon orogen and adjacent foreland. The discussion begins at the intersection of the Ouachita-Marathon and Appalachian orogens in Mississippi and Alabama and terminates at the Llano uplift and adjacent Ouachita orogen in central Texas. For each area along the trend, a general discussion of surface data and subsurface control is given. More detailed descriptions of small-scale structural features for localities within the the project boundary (see Plate 3.1 for location) of this dissertation are given in Chapters Four and Five.

Black Warrior basin

Surface exposure

Except for a small area in part of northeast Mississippi and a large area to the east in Alabama, Paleozoic rocks in the Black Warrior basin are buried by rocks of the Mesozoic-Recent Gulf Coastal Plain (Thomas, 1972b). Exposure of Paleozoic rocks increases toward the east near the Appalachian orogen and to the north on the flanks of the Nashville dome (Figure 3.1).

Subsurface control

Subsurface control in the Black Warrior basin consists of wells and seismic reflection profiles. In east-central Mississippi, at the northeastern fringe of the Ouachita-Marathon orogen, northwest-southeast trending deep-water Paleozoic “Ouachita-facies” rocks encountered in wells are truncated and over-ridden farther to the east by east-west to southwest-northeast-trending shallow-water foreland-facies carbonates and synorogenic clastic rocks of the Appalachian orogen (Thomas, 1972b, 1988, 1989). These same “Ouachita-facies” grade northward into foreland-facies equivalents in the Black Warrior basin.

The Black Warrior basin is a foreland basin situated to the north of the intersection of the Ouachita-Marathon and Appalachian orogens (Figure 3-1). Subsurface control of Paleozoic strata of the Black Warrior basin includes core samples, cuttings, and geophysical logs for more than 1250 wells (Thomas, 1972b, 1988). Most of the wells penetrate only the Mississippian-Pennsylvanian synorogenic clastic wedge rocks of the upper part of the Paleozoic succession and into the top of the Mississippian shelf facies (Thomas, 1972b, 1988). Many wells penetrate beneath the clastic wedge rocks into Mississippian and older shallow-water (passive-margin) carbonates (Thomas, 1972b, 1988). A few wells drilled to Precambrian basement. Subsurface control for
deeper parts of the Paleozoic sedimentary rock sequence and underlying basement consist of interpreted seismic profiles and down-dip projection from rock units exposed on the surface or encountered in four basement-penetrating wells located in northern Mississippi (Thomas, 1972b, 1988).

Within the western part of the Black Warrior basin, Paleozoic strata dip homoclinally toward the southwest (Figure 3.1) (Thomas, 1972b, 1973, 1988). The southwest-directed dip increases from < 0.5° in the northeastern part of the basin (on the southern flank of the Nashville dome) to > 3° in the southwestern part adjacent to the subsurface Ouachita-Marathon orogen (Figure 3.1) (Thomas, 1972b, 1973, 1988). The homocline is cross cut by several northwest-southeast striking basement normal faults. Maximum vertical displacement across these “horst and graben” type faults is approximately 1.5 to >2.0 kilometers (Figure 3.1) (Thomas, 1972b, 1973, 1988). Displacement across these basement faults is generally down-to-the-south and is greater towards the Ouachita-Marathon orogen.

The deep part of the Black Warrior basin is bordered on the south by a thrust belt consisting of mostly lower Paleozoic rocks. The structural relief between this “Central Mississippi deformed belt” (CMDB) structural high and the adjacent basin to the north is at least 3650 m (Figure 3-1) (Thomas, 1972b, 1973). On the basis of seismic reflection data and one well that drilled through the frontal thrust in Mississippi, the northern boundary of the CMDB is a south-dipping, north-vergent thrust fault (Thomas, 1972b, 1988). Additional north-vergent thrust faults are located south of the CMDB boundary fault (Thomas, 1972b).

Wells have drilled Cambrian to Pennsylvanian age rocks across the CMDB (Figure 3-1). The Cambrian-Ordovician succession is predominantly shallow-water passive-margin carbonates (Knox Group-Chickamauga Group) (Plate 2.2) (Thomas, 1972b, 1988). The Silurian-Devonian succession is primarily shale and cherty carbonates (Thomas, 1972b, 1988). The thin Silurian-Devonian succession is capped by a much thicker Mississippian-Pennsylvanian succession of basal carbonates grading upward into clastic rocks (Thomas, 1972b, 1988). In contrast to the “Appalachian” facies carbonates and clastic rocks in the CMDB; farther west, low-grade slates with included quartz veins and cherts comprise the Western Mississippi slate belt (Thomas, 1972b, 1973, 1988). These low-grade slates and cherts are similar to the Ouachita facies slates and cherts exposed farther west in the Ouachita Mountains. The contrast in rock types between the CMDB and the Western Mississippi slate belt indicates that a shallow-water to deep-water facies transition zone is located at the western edge of the
CMDB (Thomas, 1972b, 1973, 1988). The difference between the east-west structural trend within the CMDB and the northwest-southeast trend of the Western Mississippi slate belt farther west indicates that the north-vergent thrust fault at the northern boundary of the CMDB continues west to where CMDB strata over-ride Ouachita facies rocks of the Western Mississippi slate belt (Figure 3.1). Because the CMDB strata are Appalachian style rocks, the CMDB is the southeastern terminous of the Appalachian orogen.

**Mississippi Valley graben/Reelfoot rift (Precambrian-early Cambrian)**--

**Mississippi embayment/Mississippi Valley syncline (Mesozoic-Recent)**

**Subsurface control**

For purpose of distinction, the term “Mississippi Valley syncline” applies to Mesozoic-Recent stratigraphy and structures within the Mississippi embayment of the Gulf Coastal Plain (Plate 3.2). The term “Mississippi Valley graben” applies to pre-Mesozoic stratigraphy and structures within the same region. The Mississippi Valley syncline consists of structures formed during the Mesozoic break-up of Pangaea and the opening of the Gulf of Mexico. The Mississippi Valley graben refers to a deep and roughly colinear late Precambrian to Middle Cambrian graben related to break-up of an older supercontinent (late Precambrian to Early Cambrian) (Plate 3.2) (Thomas, 1989, 1991).

The term “Reelfoot” has both structural and geographic connotations. The term Reelfoot rift coined by Ervin and McGinnis (1975) refers to a pre-Mesozoic graben structure that extends in a northeast-trending zone from northwestern Arkansas to western Kentucky and southern Illinois. Although the graben was renamed the “Mississippi Valley graben” by Kane and others (1979), the term “Reelfoot” persists in more recent publications (Kane and Hildebrand, 1981; Hildebrand, 1985; Marshak and Paulsen, 1996; Dart and Swolfs, 1998). To complicate matters, faults responsible for present-day seismic activity within northern part of the Mississippi embayment (western Kentucky, western Tennessee, eastern Missouri) are also considered “Reelfoot” structures.

Paleozoic rocks are completely buried beneath the Mesozoic-Recent formations of the Mississippi embayment. Paleozoic strata crop out both to the west and east of the Mississippi embayment (Plate 3.2). The margin of the Mississippi embayment coincides with the outer edge of a broad, south-plunging, syncline comprised of Mesozoic-Cenozoic strata. Beneath the basal Mesozoic unconformity, Paleozoic strata are broadly folded into south-plunging anticline-syncline pair (Thomas, 1989, 1991).
Well data show fault separation of the top of Precambrian basement along the boundary faults of the Mississippi Valley graben as on the order of 1800 m, whereas fault separation of Ordovician and younger rocks generally is less than 500 m (Thomas, 1989, 1991). The difference results from reactivation of earlier faults (Thomas, 1991). To the east, Paleozoic formations, mostly carbonates, are exposed on the Nashville dome (Precambrian in the subsurface) (Plate 3.2) (Thomas, 1991). To the west, on the Ozark dome (Plate 3.2), both Paleozoic sedimentary rocks and Precambrian igneous basement rocks crop out.

The northwest-trending Paleozoic shallow-water-facies rocks of the subsurface Black Warrior basin are truncated by north- and northeast-trending basement faults that extend along the eastern margin of the Mississippi Valley graben (Plate 3.2). Farther west, a sub-parallel northeast-trending basement fault separates the Mississippi Valley graben from the Ozark dome (Plate 3.2). Well and seismic data show that deep-water Ouachita facies rocks of the Ouachita thrust belt over-ride the Paleozoic rocks of the southern Mississippi Valley graben (Thomas, 1989, 1991). Lithologic control for Paleozoic strata outside the edges of the Mississippi Valley graben is provided by scattered deep wells and down-dip projection of rocks exposed to the west on the crest of the Ozark dome and to the east on the crest of the Nashville dome (Figure 3-1 and Plate 3.1) (Thomas, 1989, 1991). The lower Paleozoic formations on the northwestern edge of the Mississippi Valley graben, closer to the flank of the Ozark dome, grade upward from basal arkosic sandstones to calcareous mudstones, and limestones (Thomas, 1991). In contrast, a more homogeneous succession of mudstone, siltstone, and sandstone comprises the lower Paleozoic strata of the northeastern part of the Mississippi Valley graben (Thomas, 1991).

Although well control for deep rocks within the buried Mississippi Valley graben is scattered, seismic and geophysical control for the entire Mississippi Valley region is extensive. Historical accounts of the series of New Madrid earthquakes of 1811-1812 (northeastern Arkansas/westernmost Kentucky/southeastern Missouri) sparked the interests of geologists in universities and the US federal government in matters concerning possible large future earthquakes produced by slip along regional basement faults (Kane and others, 1981). In 1974, the United States Geological Survey (USGS) began extensive seismic and aeromagnetic and gravity surveys of the region in search of potential seismogenic faults (Kane and others, 1981).

The fruit of the USGS survey is a series of magnetic and gravity anomaly maps for the Mississippi Valley region (Plate 3.3). Magnetic and gravity anomaly maps
clearly mark the boundaries of the Mississippi Valley basement fault system (Ervin and McGinnis, 1975; Kane and others, 1981; Hildebrand, 1985). Gravity and magnetic anomaly maps cannot be used to directly differentiate between rocks of different ages; however, they do show local variations which are a function of composite vertical thickness and density of rock types. A magnetic anomaly “high” indicates a greater composite thickness of iron-rich mafic rocks in that location (Plate 3.3). This may result from basement uplift or shallow-depth emplacement of a mafic igneous body. A gravity anomaly high commonly indicates that crystalline basement is relatively shallow and covered by a thinner sedimentary succession (Plate 3.3).

Interpretation of gravity anomaly maps and seismic refraction data within the Mississippi Valley graben indicates that crust beneath the Mesozoic strata of the Mississippi embayment is subdivided into zones of different density (seismic velocity) (Ervin and McGinnis, 1975; Keller and others, 1983). Both Ervin and McGinnis (1975) and Keller and others (1983) show an anomalously dense layer above the crust/mantle interface (Moho). The density of the anomalous lower crustal layer (calculated from seismic velocities) is intermediate between the densities of continental crust and mantle; therefore, it is suggested that the lower crustal layer is composed of transitional crust (Ervin and McGinnis, 1975). Bouguer gravity profiles show that the transitional crustal layer is thickest in the center of the Mississippi Valley graben and thins progressively towards the margins (Ervin and McGinnis, 1975; Keller and others, 1983). Above this lower transitional crustal layer, Ervin and McGinnis (1975) separate the overlying crust into three layers; whereas, Keller and others (1983) show two overlying crustal layers.

Interpretations based on the integration of magnetic and gravity anomaly data show shallow pods of high density rocks suspended within the crust (Ervin and McGinnis, 1975; Kane and others, 1981; Hildenbrand, 1985). The shallow depth, high density pods are interpreted from magnetic field data as mafic or ultramafic intrusive igneous bodies of uncertain age (Hildenbrand, 1985). The closest exposed mafic rocks are Cambrian basalts and gabbros in the Arbuckle and Wichita Mountains of the southern Oklahoma aulacogen (Plate 3.1).

Ozark dome, Arkoma basin, Ouachita Mountains, Athens Plateau

Surface exposure

The Ozark dome is a basement upwarp centered in southeast Missouri (Plate 3.4). The center of the Ozark dome consists of Precambrian basement rocks, including weakly metamorphosed igneous rocks dated at <1500 Ma and are considered part of the
“1500 Ma granite-rhyolite province” of North America (Van Schmus and Bickford, 1993).

To the south, along the rim of the Ozark dome, rocks dip southward, and toward the south, progressively younger strata are exposed. The narrow South Ozark arch extends westward from the southern Ozark dome and connects with the Northeast Oklahoma platform (Plate 3.4) (Suhm, 1997). Mostly Cambrian-Mississippian passive margin carbonates are exposed across the South Ozark arch and Northeast Oklahoma platform (Thomas, 1991; Suhm, 1997). Southward into the Arkoma basin, Cambrian-Mississippian passive-margin carbonates are overlain by Mississippian-Pennsylvanian clastic facies rocks. The older clastic rocks along in northern and western parts of the Arkoma basin are predominantly shallow-water facies and grade down-section into Mississippian carbonates (Thomas, 1989). Across southern and eastern parts of the Arkoma basin, shallow-water clastic rocks grade up-section into younger Mississippian-Pennsylvania marginal-marine and deltaic clastic facies that increases in thickness progressively towards the south (Morris, 1974; Thomas, 1975; Bush and others, 1977).

East-west-striking, north-vergent thrust faults mark the abrupt transition from the eastern Arkoma basin to the eastern Ouachita thrust belt in Arkansas (Plate 3.4). Farther west, in Oklahoma, the frontal fault of the Ouachita thrust belt bends abruptly from east-northeast-striking to a northeast strike (Plate 3.4). From west to east along strike, the Choctaw, Dutch Creek, and Ross Creek faults are the frontal thrust fault separating the Frontal Ouachita thrust belt from the Arkoma basin (Plate 3.4). In the easternmost part of the Frontal Ouachita thrust belt, structures bend sharply from northeast-striking to southeast-striking (Plate 3.4). An abrupt transition from predominantly marginal-marine and deltaic facies rocks to deep-marine turbidite facies within the Mississippian-Pennsylvania succession coincides with the northern and northwestern boundaries of the Ouachita thrust belt. However, exposure of Pennsylvanian deep-water turbidites in the cores of narrow anticlines within the eastern Arkoma basin indicates that the marginal-marine to deep-marine transition zone extends north of the Ouachita thrust front in the subsurface (Arbenz, 1989a, d; Thomas, 1989; Viele, 1989; Haley and others, 1993). From west to east along strike of the Frontal Ouachita thrust belt, the general structural fabric changes from a zone of numerous fault imbricates to upright, broad faulted folds (Arbenz, a,d; Viele, 1989; Haley and others, 1993). Although thrust faults in the western part of Frontal Ouachita thrust belt are mostly north and northwest-vergent, thrust faults in the eastern part are both north-vergent and south-vergent (Arbenz, 1989a,b; J.K. Arbenz, unpublished cross sections).
Pre-orogenic (pre-Meramecian) rocks are exposed in several isolated locations within the western Frontal Ouachita thrust belt. Four of these areas are located along the Pine Mountain fault, and a larger area is Black Knob Ridge located in the southwestern corner of the frontal part of the Ouachita thrust belt (Plates 3.1 and 3.4). Exposed within the isolated areas along the Pine Mountain fault are Devonian chert (transitional facies Pine Top Chert) and cherty Woodford Shale (Hardie, 1988; Arbenz, 1989b,d). The Pre-Meramecian succession exposed within Black Knob Ridge is Ordovician cherty and carbonaceous shales and slate, calcareous chert, chert, and Silurian-Devonian novaculite (Hendricks and others, 1937; Finney, 1988; Lowe, 1989). In contrast to the rest of the western frontal thrust belt where marginal- to deep-marine synorogenic rocks are imbricated, at Black Knob Ridge, deep-water pre-orogenic rocks are exposed near the frontal thrust fault with a small amount of fault imbrication and shortening to the west.

An area referred to in Plate 3.4 (Map B) as the Central Ouachita thrust belt spans an area south and southeast of the Frontal Ouachita thrust belt, northeast of the Broken Bow uplift, and north of the Cross Mountains-Trap Mountains trend. Included within the eastern part of the Central Ouachita thrust belt are the Benton uplift and the Eastern “nappes” area (Plate 3.4). Except for the Potato Hills (Plate 3.4), within the Central Ouachita thrust belt, the overall structural fabric changes from northwest- and north-vergent long-wavelengthfaulted folds in the west, to a tightly folded complex of predominantly south-vergent folds and folded thrust faults in the east (Arbenz, 1989a,b,d; Nielsen and others, 1989). The Potato Hills is a small northeast-trending elliptically shaped anticlinorium in the western part of Central Ouachita thrust belt where pre-Meramecian rocks are exposed (Plate 3.4). Tight folds and folded thrust faults similar in structural style to those exposed farther east along strike in the eastern Central Ouachita thrust belt are common in the Potato Hills (Plate 3.4). Although very complex structurally, the overall strike of thrust faults in the easternmost part of the Central Ouachita thrust belt bends sharply from east-west to a southeast (Plate 3.4).

Upper Mississippian (upper Stanley Group) to Pennsylvanian (Atoka Formation) synorogenic deep-water sandstone and shale turbidites form the broad folds of the western Central Ouachita thrust belt of the Ouachita Mountains (Morris, 1974; Arbenz, 1989a,b,d). The dominant ridge-forming unit in the western Central Ouachita thrust belt is the uppermost Chesterian-Morrowan Jackfork Sandstone (Morris, 1974). The Pre-Meramecian units exposed within the Potato Hills consist of a succession of Ordovician carbonaceous slates, cherty slates, cherts, and novaculites (Womble Shale-Arkansas Novaculite) (Allen, 1994). Exposed in the Benton uplift across the eastern part of the
Central Ouachita thrust belt is a more expansive succession of Cambrian to lower Mississippian (Kinderhookian-Osagean) pre-orogenic carbonaceous slate, distal turbidite, chert, and novaculite (Collier Shale-Arkansas Novaculite). Except for the area encompassed by the Mazarn basin (Plate 3.4), where Chesterian-Morrowan (middle and upper Stanley Group through Jackfork Sandstone) rocks are exposed, the lower part of the syn-orogenic Stanley Group (Meramecian) caps the pre-orogenic succession of the eastern Central Ouachita thrust belt (Arbenz, 1989a,b,d; Haley and others, 1993). A relatively wide expanse of tightly folded Stanley Group shale turbidites separates the western and eastern parts of the Central Ouachita thrust belt.

South of the Central Ouachita thrust belt is the Southern Ouachita thrust belt. As shown in Plate 3.4 (Map B), the Southern Ouachita thrust belt includes the Broken Bow uplift, and the area to the east that is south of the Cross Mountains-Trap Mountains trend, and north of the Mesozoic Gulf Coastal Plain. The westward along-strike change from broad, faulted folds in the east, to a complex zone of tight folds, folded thrust faults, and faulted folds in the west within the Southern Ouachita thrust belt is the reverse of the structural style previously described for the Central Ouachita thrust belt (Plate 3.4). Structural grain across the Broken Bow uplift bends sharply from a northeast-strike in the west, to an east-west-strike in the east. Similar to those of the Benton uplift to the northeast, the folds of the Broken Bow uplift are overturned to the south or southeast.

Very complex structures are located on the northern flank of the Broken Bow uplift. In general, from south to north perpendicular to strike, vergence of folds switches from south in the Broken Bow uplift, to north in the Cross Mountains (Figure 3.2). In several locations, vergence direction of folds changes abruptly across a single fault, or fault zone and folds fan-out (diverge) in cross-section (Part B in Figure 3.2). The east-west strike of folds and faults on the western edge of the Broken Bow uplift changes to an east-northeast strike farther west (Figure 3.2 and Plate 3.4).

The southern part of the Southern Ouachita thrust belt is generally a south-dipping, faulted homoclone, which to the south is covered by Mesozoic rocks of the Coastal Plain. This part of the Ouachita Mountains is also called the Athens Plateau (Plate 3.4). Mississippian-Pennsylvanian (Chesterian-Atokan) upper Stanley Group-Atoka Formation deep-water turbiditic sandstones are the resistant, ridge forming units of the Athens Plateau (Morris, 1974). Thrust faults within the Athens Plateau are predominantly north-vergent, and locally cross-cut by north- and northwest-striking transverse faults (Plate 3.4) (Walthal, 1967; Arbenz, 1989b; and Haley and others, 1993). A narrow zone of tightly folded Mississippian Stanley Group shale which separates the
northeastern part of the Athens Plateau from the Cross Mountains-Trap Mountains trend widens westward towards the Broken Bow uplift (Figure 3.2 and Plate 3.4). The core of the Broken Bow uplift contains a Cambrian-early Mississippian (Osagean) succession of pre-orogenic slate, distal turbidite, chert, and novaculite similar to that exposed to the northeast in the Benton uplift (Plate 2.3) (Lowe, 1989; Morris, 1989).

Subsurface control

Subsurface control for the region south of the Ozark dome is primarily well data, regional geophysical surveys and local seismic reflection and refraction profiles. The depth, location, number of wells, and therefore subsurface control, is a function mainly of commercial interests (such as oil and gas exploration). Many commercial oil and gas wells are located throughout the Arkoma basin, especially in the western part. The Wilburton field (Map B in Plate 3.4) in southeastern Oklahoma is an example of an area with exceptional well (and proprietary seismic) control (Arbenz, 1989d; Gatewood and Fay, 1991). For the most part, commercial oil and gas interests are concentrated in the autochthonous strata of the Arkoma basin and along the Frontal Ouachita thrust belt, mostly to the west and north of the Potato Hills (Figure 3.3) (Suneson and Campbell, 1989; Suneson and others, 1989; Gatewood and Fay, 1991).

There are few deep wells within the core areas of the Benton and Broken Bow uplifts. One such well, the Sohio 1-22 Weyerhaueser (figure 3.3) well drilled through more than 3560 m of deformed lower Paleozoic “Ouachita facies” rocks similar to those exposed at the surface across the Broken Bow uplift before it penetrated very low-grade metamorphosed and undeformed “Arbuckle facies” passive-margin carbonate shelf rocks similar to those exposed farther west in the Arbuckle Mountains (Leander and Legg, 1988; Denison, 1989). On the basis of lithologic comparison with rock formations encountered in nearby wells and exposed in the Arbuckle Mountains to the west (Figure 3.3 and Plate 3.4), the autochthonous sedimentary sequence beneath the basal detachment consists of Cambrian to Ordovician Arbuckle Group and lower Simpson Group (Denison, 1989). The overlying succession of the Ordovician-lower Mississippian shallow-water Arbuckle facies rocks which crops out in the Arbuckle Mountains to the west is absent from the Sohio 1-22 Weyerhaueser well.

The Hunt 1 Neeley well (Figure 3.3) in Lamar County of northeast Texas records a similar transition beneath the basal detachment of the allochthonous sequence. Here, the Paleozoic strata lie beneath the rocks of the Mesozoic Gulf Coastal Plain. At a depth of 4100 m, the deformed and slightly metamorphosed allochthonous strata are underlain by weakly metamorphosed “clean metaquartzite and dolomitic marble” which
are considered to be Arbuckle Group through lower Simpson Group (Denison, 1989). Similar to the Sohio 1-22 Weyerhaueser well, the Upper Ordovician-lower Mississippian part of the passive-margin shelf-carbonate succession is also absent in the Hunt 1 Neeley well. The Hunt-Neeley well penetrated 1800 m of shallow-water shelf carbonates before it bottomed in graphite-schist (presumably Precambrian basement) (Denison, 1989).

Well control south of the Athens Plateau (Plate 3.4) beneath the Gulf Coastal Plain is limited to sparse wells in south-central Arkansas, northeastern Texas and north-central Louisiana (Ark-La-Tex) and the Sabine uplift (Plate 3.1) of eastern Texas and western Louisiana. Approximately four dozen wells penetrated Paleozoic strata in the Ark-La-Tex region. In the northern and western part, wellsbottomed in Carboniferous deep-water turbidites (mostly Atoka Formation). However, in the southern part, wells bottomed in either upper Pennsylvanian (Desmoinesian) shallow-water carbonates, or Permian clastic rocks and associated thin carbonates (Nicholas and Waddell, 1989). Where both Desmoinesian and Permian rock units are encountered, they are separated by an unconformity. The entire Desmoinesian-Permian succession is classified as the “Morehouse” formation (Fay and others, 1986). To the southwest, on the Sabine uplift, seven deep wells encountered Desmoinesian-Permian shallow-water carbonates and clastic rocks beneath Mesozoic cover, one well bottomed in Mississippian-Pennsylvanian deep-water turbidites, and two others encountered volcanic rocks (Nicholas and Waddell, 1989).

**Seismic, gravity and magnetic anomaly data**

For most of the Central and Southern Ouachita thrust belts and areas farther southeast where Paleozoic and older strata are deeply buried beneath the Mesozoic-Recent Gulf Coastal Plain, subsurface control is limited to structural interpretations based upon gravity and magnetic anomaly maps and profiles, and seismic reflection and refraction profiles (for example, Kruger and Keller, 1986; Mickus and Keller, 1992). The following paragraphs summarize some of the various geophysical data for the Ouachita Mountains region, and list some of the interpretations derived from the data.

Long, deep seismic profiles across the Ouachita Mountains and extending south into Louisiana were produced by two consortia. In 1981, the Consortium for Continental Reflection Profiling (COCORP), a combination of personnel from geological surveys, academia, and industry funded by a grant from the National Science Foundation, acquired five deep seismic reflection profiles across the Ouachita Mountains (Lillie and others, 1983). The total length of profiles approximated 200 km, of which two north-
south profiles approximated 130 km, and three shorter, intersecting east-west sections comprised the rest of the survey (Lillie and others, 1983).

Lillie and others (1983) subdivide the interpreted cross section derived from the COCORP profile, from north to south, into five zones (Plate 3.5). The northernmost “frontal thrust zone,” is characterized by south-dipping coherent reflectors above a two-way travel time depth of 2 seconds (5 km at an average seismic velocity of 5 km/second) at the northern margin of the section, to 4 seconds (10 km at an average seismic velocity of 5 km/second) at the southern edge of the “frontal thrust zone.” The frontal thrust zone of Lillie and others (1983) is roughly equivalent to the Frontal Ouachita thrust belt (Plate 3.4), but extends into the southern part of the Arkoma basin. The frontal thrust zone has a width of 30 km, making the average dip of the basal coherent reflector approximately 18.5° to the south.

Halfway across the frontal thrust zone, the reflectors are interrupted by a south-dipping, north-vergent thrust fault. This fault is shown on regional geologic maps as the Ross Creek-Dutch Creek fault zone (Plate 3.1 and 3.4). Surface geology and wells correlate most of the reflectors with Mississippian-Pennsylvanian syn-orogenic shallow-to marginal-marine deltaic and deep-water turbidite rock units (Plate 3.5). On the basis of projection of formation tops from nearby wells, the basal reflector is determined to be the top of the Cambrian-early Mississippian pre-orogenic shallow-water shelf carbonates. Surface structures in the frontal thrust zone consist of broad (15 km wide), open synclines with narrow, fault-cored anticlines (Plate 3.5).

To the south of the frontal thrust zone is a location along the COCORP profile aptly referred to as the “Maumelle chaotic zone” (Plate 3.5). Except for the southernmost part, the Maumelle chaotic zone is included in the Frontal Ouachita thrust belt (Plate 3.4). In the Maumelle chaotic zone, seismic reflectors “wash-out” and appear as short discontinuous reflectors of various attitudes, suggesting that strata are tightly folded and steeply dipping (Plate 3.5). The surface formations in this area are indeed highly deformed Mississippian to Pennsylvanian syn-orogenic deep-water turbidites (Stanley Group, Jackfork Sandstone, and Johns Valley Shale) (Nielsen and others, 1989; Haley and others, 1993). Tight semi-chevron folds are common to the shaley beds (especially in the Stanley Group), and all formations are cross-cut by cleavages and locally by quartz veins (Nielsen and others, 1989). Many of the shaley beds, especially the Johns Valley Shale are heavily cleaved, including pencil cleavages, and locally have a swirly texture with included rock fragments of various types (“Johns Valley Boulders”) (Shideler, 1970; Nielsen and others, 1989).
A curious phenomenon of the Maumelle chaotic zone is the contrast between the south-dipping seismic reflectors at depth (Plate 3.5), and predominantly north-dipping strata at the surface. It is possible that the seismic waves are relecting off of relatively smooth, south-dipping faults within a sequence of highly deformed rocks.

To the south of the Maumelle chaotic zone, in the center of the north-south COCORP profile, is the Benton uplift (Plate 3.5). The Benton uplift is located in the eastern part of the Central Ouachita thrust belt (Map B in Plate 3.4), and according to the COCORP seismic reflection profile (Lillie and others, 1983), structures exposed on the surface within the Benton uplift are folded over the top of a broad subsurface arch (Plate 3.5). In general, seismic reflectors above 3.5 seconds (two-way travel time) in the center of the uplift, bend somewhat symmetrically to the north and south to 5.0 and 6.5 seconds respectively. At an estimated lithologic velocity of 5 km/second (used by Lillie and others, 1983), this corresponds to a basement arch with a minimum structural relief of 3.75 kilometers.

No well has penetrated beneath allochthonous strata within the Benton uplift; therefore, Lillie and others (1983) show several possible interpretations of autochthonous stratigraphy and geometry beneath the allochthonous terrain of the Benton uplift (Plate 3.5). Two interpretations illustrated in Plate 3.5 show the basal detachment of the allochthon resting on Cambrian-Mississippian “Arbuckle facies” type passive-margin shelf carbonates. The southward extension of a passive margin shelf to the southern boundary of the Benton uplift agrees with a larger scale lithospheric model of the region illustrated in Keller and others (1989) and Mickus and Keller (1992). The Keller and others (1989) and Mickus and Keller (1992) model shows that the southern edge of continental crust coincides with the southern edge of the Benton uplift (Plate 3.6). The sub-thrust presence of autochthonous Arbuckle facies carbonates in the Sohio 1-22 Weyerhaueser well located nearby to the southwest in the Broken Bow uplift also suggests the likely presence of autochthonous carbonates beneath the Benton uplift. All interpretations illustrated in Plate 3.5 show a large-displacement south-dipping north-directed thrust fault on the northern flank of the Benton uplift. However, geometry of this fault zone is poorly constrained because the Cambrian-Mississippian carbonate reflector is not clearly traceable beneath the Maumelle chaotic zone (as shown in uninterpreted seismic profile included in Lillie and others, 1983).

Predominant vergence of exposed surface structures changes direction from north to south across the Benton uplift. South-verge (southward-overturning) is common in the northern part of the Benton uplift, whereas north-verge (southward-overturning) is common in
the southern part (Plate 3.5). For the case where all layers are originally horizontal (no pre-existing folds), the axial planes of small-scale folds within a much larger arch tilt towards the axial plane of the larger arch, vergence directions change across the top of the basement arch, and no folds are overturned. However, the presence of south-overturned structures in the Maumelle chaotic zone (Frontal Ouachita thrust belt in Map B in Plate 3.4) suggests a large scale south-vergent thrust fault is located north of the Benton uplift.

Lillie and others (1983) show a zone called the “southern Ouachitas” south of the Benton uplift on their COCORP-based interpreted cross section (Plate 3.5). The “southern Ouachitas” zone illustrated in Plate 3.5 is located in the eastern part of the Southern Ouachita thrust belt (Map B in Plate 3.4). The “southern Ouachitas” zone where seismic reflectors are coherent and have southward dips correlates with the Athens Plateau homocline (Plate 3.4). Some signal loss is evident below 4.5 seconds (two-way travel time) (10 km) at the southern end of the Benton uplift on the COCORP profile in Lillie and others (1983); however, a discontinuous reflector can be traced farther south to a depth of 8.0 seconds (two-way travel time) (40 km). The nature of this surface is unknown. The shallow south-dipping seismic reflectors correlate with south-dipping Mississippian (Meramecian)-Pennsylvanian (Atokan) synorogenic deep-water turbidites (Stanley Group through Atoka Formation) exposed at the surface. The Paleozoic strata of the southern Ouachitas (Athens Plateau) continue beneath the Mesozoic-Recent Gulf-Coastal Plain farther to the south beyond the southern end of the Lillie and others (1983) COCORP profile.

Studies of the pre-Mesozoic subsurface structures south of the northern boundary of the Gulf-Coastal Plain in southern Arkansas, east Texas, and Louisiana rely heavily on seismic data collected under the guidance of the Program for Array Seismic Studies of the Continental Lithosphere (PASSCAL) and earlier studies (Mickus and Keller, 1992). A series of seismic profiles, including three long, deep transects are used to constrain models of lithospheric composition and structure across the region that extends from the southern part of the Ouachita Mountains in Arkansas southward into the Gulf of Mexico (Plate 3.6). Vertical variations in seismic velocity derived from seismic refraction data are used to estimate vertical changes in rock densities. The calculated rock densities for the seismic profile are compared with densities of known rock types to produce a cross section that illustrates large-scale, deep structural geometry (Plate 3.6). Regional gravity and magnetic anomaly data further constrain lithospheric models.
Regional gravity anomaly maps published by Kruger and Keller (1986) are constrained by more than 35,000 geophysical data collection points and 96 deep wells. Briefly stated, the model parameters include densities of 2.50-2.70 g/cm³, 3.0 g/cm³ and 3.3 g/cm³ for upper crust (including sedimentary cover), lower crust and mantle respectively. Kruger and Keller (1986) produced several types of gravity anomaly maps. One type called Bouger maps are designed to correct for local increases in gravity caused by topographic highs. The resulting map illustrates high and low gravity areas related to broad regional structures, such as uplifts of basement (gravity highs) or deep sedimentary basins (gravity lows). Another type called residual gravity maps remove any broad regional trends and accentuate local variations. Patterns shown in the residual and bouger gravity maps for the Ouachita region are very similar, although numerical values are different (Figure 3.4).

A residual gravity anomaly map with regional structures superimposed show many distinct “highs and lows” for the area encompassing the Ouachita Mountains and its surroundings (Figure 3.5). Distinct three northwest-southeast-trending gravity highs and lows separate two gravity lows that roughly coincide with the Fort Worth and Arkoma foreland basins (Figure 3.5). The northwest-southeast-trending three gravity anomalies separated by steep-gradients coincide with three fault-bounded basement uplifts and depressions that comprise the Southern Oklahoma aulacogen (Plate 3.7). Granites, rhyolites, gabbros, basalt dikes, and meta-igneous basement rocks are exposed within the Arbuckle and Wichita uplifts of the Southern Oklahoma aulacogen (Ham and others, 1954; Gilbert, 1982, 1983). Precambrian basement rocks are exposed in the eastern parts of the Arbuckle Mountains near the intersection with the Ouachita-Marathon orogen (Ham and others, 1954). Although the more easily weathered gabbros and basalts are found as scattered exposures, the gravity anomaly highs shown in Figure 3.5 suggest that a much greater amount is buried along the Arbuckle-Wichita uplift trend.

The eastern side of a roughly linear gravity high parallel with the margin of Ouachita thrust front shown as the “interior zone gravity maximum” in Figure 3.5 roughly marks the northwestern boundary of the Mesozoic basins located in east Texas and Louisiana. Three of semi-circular to elongate gravity anomaly highs and lows are found to the south and east of the “interior zone gravity maximum” (Figure 3.5). A roughly circular gravity high (positive gravity anomaly) coincides with the Sabine uplift of east Texas and western Louisiana (Figure 3.5 and Plate 3.6). Integration of deep seismic refraction data (Hales and others, 1970) and gravity anomaly data indicates that a
large semi-circular pod of upper-crustal-density rock (2.75 g/cm³) is located beneath the Sabine gravity high (Figure 3.5 and Plate 3.6) (Mickus and Keller, 1992). As shown in Plate 3.6, at its center, a crustal block beneath the Sabine gravity high has a crest at a depth of approximately 5 km beneath the Mesozoic cover, and it pinches out laterally at an approximate depth of 20 kilometers. The diameter of the Sabine block is nearly 800 kilometers (Plate 3.6).

General structural features of the Mickus and Keller (1992) model (Plate 3.6) are listed as follows. From the northern edge of the Mickus and Keller (1992) cross section, thickness of continental crust decreases abruptly from approximately 43 km in the Arkoma basin to 0 km beneath the southern part of the Ouachita Mountains. A 100 km wide zone of transitional crust separates the North American craton from the 800 km wide Sabine microcontinent/island arc terrane (Plate 3.6). Thin oceanic crust extends from the southern side of the Sabine terrane southward into the Gulf of Mexico (Plate 3.6). The Mesozoic Gulf Coastal Plain succession thickens southwards from zero at the southern edge of the Ouachita Mountains to more than 16 km in the Gulf of Mexico (Plate 3.6).

One of the most interesting features of the residual gravity anomaly map shown in Figure 3.5 is a -100 mgal gravity minimum (Arkoma basin gravity low) centered on the southeastern Oklahoma-Arkansas border. Because the long-axis of the depression is oblique to the trend of the Ouachita Mountains thrust belt, thrust loading and sediment deposition related to the Ouachita orogen alone cannot explain this feature. Cross sections of the western part of the Ouachita Mountains show an estimated depth of 11 km for the top of pre-Upper Cambrian basement in the area that coincides with the Arkoma basin gravity low (Arbenz, 1989a). West of the Ouachita thrust front, in the deep part of the Anadarko basin located along the northern side of the Wichita uplift, estimated depth to top of pre-Upper Cambrian basement is 12 km (Brewer and others, 1983; Ewing, 1991). The deep part of the Anadarko basin coincides with a residual gravity low of –20 mgal which is far less than the –100 mgal measured in the Arkoma basin gravity low; however, interpreted depth to the top of pre-Upper Cambrian “basement” is greater depth in the Anadarko basin.

Kruger and Keller (1986) suggest that the Arkoma basin gravity minimum is related to a late Precambrian-Early Cambrian (“Eocambrian”) cratonic sedimentary basin. A localized, thick succession of sedimentary rocks beneath the Cambrian-Pennsylvanian stratigraphy would accentuate the gravity low. A lithospheric model (Kruger and Keller, 1986) calculates that approximately 15 km of sedimentary rocks
(autochthonous and allochthonous) rests above an abruptly thin (≈ 9 km) upper (granitic) crustal layer in the deepest center of the Arkoma basin gravity minimum. The Kruger and Keller (1986) model also shows a slightly thinner underlying lower crustal layer. Because no well has penetrated basement in the area northwest of the Broken Bow uplift and southeast of the Frontal Ouachita thrust belt, the exact nature of basement is unknown.

The depths to basement in the Anadarko basin are determined from deep well data, and a COCORP seismic profile constructed across the Wichita Mountains (Brewer and others, 1983). The exposure of several types of mafic igneous rocks in the Wichita Mountains (Gilbert, 1982) and the steep gravity and magnetic gradients on the northern and southern sides of the Wichita are consistent with a thick, deeply rooted mafic igneous body beneath relatively thin granitic continental (upper) crust within the uplift (Hildenbrand and others, 1983; Kruger and Keller, 1986). The gravity low to the north of the Wichita uplift suggests that the adjacent Anadarko basin sits upon much thicker continental crust (Kruger and Keller, 1986). Another lithospheric model (Kruger and Keller, 1986) shows nearly 19 km-thick granitic upper crustal layer beneath approximately 1.5 km of sedimentary rocks in the shallow, southeastern part of the Anadarko basin. Analogous to the Arkoma basin gravity minimum, the granitic upper crustal layer within the Anadarko basin is likely thinner in the deeper parts of the basin. The smaller negative values within the Anadarko basin gravity anomaly, in comparison to the greater negative values within the Arkoma basin gravity minimum, are consistent with a thinner sedimentary succession (≈ 13 km, maximum, Ewing, 1991) for the deep Anadarko basin in contrast to 15 km for the center of the Arkoma basin gravity minimum. The stark contrast between the minimum values (-20 mgal, Anadarko basin, in contrast to –100 mgal, Arkoma basin gravity minimum) also suggests a thinner granitic upper crustal layer and/or thicker, more dense, lower crustal layer in the Anadarko basin as compared to the Arkoma basin gravity minimum.

**Southern Oklahoma aulacogen (Arbuckle, Wichita, and Amarillo uplifts) and adjacent northern and western Texas (overview)**

Continuing in a counterclockwise direction along the foreland of the Ouachita orogen, the northeast-southwest trending structures of the western Arkoma basin and Ouachita Mountains intersect the northwest-southeast trending structures of the Southern Oklahoma aulacogen (Plate 3.7). Although for some, the term aulacogen includes the failed-arm of a Burke-Dewey (1973) type triple junction, in its strictest sense, the term as initially defined by Schatski in 1946 is descriptive with no genetic
implications. An aulocogen is simply a linear system of continental basement faults that intersect a thrust belt at a large angle. The structures of the southern Oklahoma region fit Schatski’s description regardless of genetic model.

The Southern Oklahoma aulacogen is one of the few locations along the Ouachita foreland where pre-Mississippian, pre-Ouachita orogenic strata are exposed. In the strictest sense, the Southern Oklahoma aulacogen is defined by outcrops across two broad igneous-cored basement uplifts (Arbuckle, Wichita) and adjacent deep sedimentary basins; however, the subsurface Amarillo uplift in the Texas Panhandle is another basement uplift that marks the northeastern terminus of structures of the Southern Oklahoma aulacogen (Plate 3.7). The Southern Oklahoma aulacogen also includes many smaller-scale structural subdivisions. In the Wichita uplift, resistant Cambrian-age granitic igneous rocks produce mountains which rise abruptly more than 304 m above the adjacent plains (Gilbert, 1982). Lesser quantities of more easily weathered mafic igneous rocks are locally exposed along the northern flanks of the Wichita Mountain granites. Variation of amount of cover vegetation visible on air photography and satellite images of the region studied by this author, and cursory field study in the region indicate that the Wichita Mountain granites, in general, are a resistant sill sitting above mafic rocks which are deeply weathered. Mafic igneous rocks are locally exposed within the Wichita Mountains, and many wells have drilled mafic rocks (Gilbert, 1982). Although covered by Pennsylvanian and Permian strata, regional seismic data show that the steep north-dipping, northwest-southeast-striking Burch fault separates the Wichita Mountains uplift block from the Hollis-Hardeman and Palo Duro basins located to the southwest (Plate 3.7) (Brewer, 1982; Brewer and others, 1983).

**Subsurface south of the Amarillo and Wichita uplifts (overview)**

COCORP seismic data from several transects that cross the Wichita Mountains, in addition to scattered well data, show 10 km of layered rocks resting beneath a thin veneer of Cambrian-Mississippian sedimentary rocks, and above basement along the southwestern flank of the Wichita uplift (Plates 3.1, 3.7 and 3.8) (Brewer, 1982; Brewer and others, 1983). The thin layer of Cambrian-Mississippian rocks shown in the COCORP-derived cross section shown in Plate 3.8 is interpreted as the basal layer of Upper Cambrian-Ordovician Arbuckle Group carbonates and underlying Upper Cambrian Timbered Hills Group clastic rocks and carbonates. Pennsylvanian and Permian sedimentary rocks rest unconformably above the Cambrian-Mississippian rocks shown in the COCORP cross section (Plate 3.8). A thin layer of unconformable Mississippian carbonates such as those that rest directly on basement south of the
Amarillo uplift (Johnson and others, 1988) may also be included in the Cambrian-Mississippian succession of rocks in the southern part of the COCORP cross section (Plate 3.8).

The southern end of the COCORP study area is located in the eastern part of the Hollis basin in southwestern Oklahoma (Plate 3.8). Regional cross sections and structure contour/isopach maps of the Cambrian-Ordovician Arbuckle Group show that the Arbuckle Group is less than 150 m thick in the eastern part of the Hollis basin (Bartram and others, 1950; Ewing, 1991). The Arbuckle/Ellenburger Group carbonates are also very thin, and locally absent, along the crests of the Matador-Red River and Muenster arches (Bartram and others, 1950; Ewing, 1991). Pre-Mississippian strata are absent from the Hardeman basin and much of the Texas Panhandle region south of the Amarillo uplift (Bartram and others, 1950; Johnson and others, 1988; Ewing, 1991).

The exact age and nature of the rocks represented by the pre-Late Cambrian (pre-Timbered Hills Group) succession of distinct southwest-dipping seismic reflector horizons visible in the southern part of the COCORP-derived cross section (Plate 3.8) are uncertain. The most accepted interpretation, based upon on-lap and down-lap patterns visible on the seismic profiles, suggests that these reflectors represent 10 km of layered clastic sedimentary rocks (Brewer, 1982; Brewer and others, 1983). Another interpretation suggests that that the layered reflectors represent layered igneous rocks (possibly rhyolites) (Lynn, 1980). The upper part of the layered reflector package is no younger than Franconian (Late Cambrian) which is the age of the basal part of the overlying Timbered Hills Group (Reagan Sandstone). A well located just south of the COCORP study area penetrated a “1265 ± 40 Ma micrographic microgranite porphyry” within the layered sequence (Denison, unpublished data referenced by Brewer, 1982). Farther north toward the Wichita Mountains, wells in southwestern Oklahoma encountered a succession of metamorphosed sedimentary rocks classified as the Tillman Metasedimentary Group (Brewer, 1982; Sides and Miller, 1982) which rest upon basement. Brewer (1982) places the approximate age of the pre-Late Cambrian layered sequence at 1200 to 1400 Ma. The pre-Franconian layered seismic reflectors visible in the southern part of the COCORP cross section are absent north of the Wichita uplift (Plate 3.8).

**Wichita Mountains and Slick Hills (surface exposure)**

To the northeast, the northwest-southeast-striking Meers fault separates the Wichita Mountains uplift from another locally famous outcrop area called the Slick Hills. At the surface, the Meers fault appears vertical, however, COCORP seismic data
suggests that the fault is south-dipping at depth (Plate 3.8). Exposed across the Slick Hills is the Cambrian-age Carlton Rhyolite overlain by Cambrian-Ordovician Timbered Hills Group clastic rocks and carbonates and Arbuckle Group carbonates (Donovan, 1986). Small localized exposures of gabbros are also found in this region. The Carlton Rhyolite is the cogenetic extrusive equivalent of the Wichita Mountain granites and both are considered to have been emplaced in a shallow-crustal, hypabyssal setting (Gilbert, 1982, 1983). Field relationships show that the clastic rocks that form the lower part of the Timbered Hills Group, thin abruptly along the flanks of several exposures of Carlton Rhyolite (Donovan, 1986). This indicates that the Cambrian igneous terrain of the Slick Hills and surrounding igneous terrains were initially exposed and subsequently overlapped by carbonates during the Franconian transgression that affected the entire region (Donovan, 1986).

Splaying off the Meers fault to the north are more northerly striking faults that swing abruptly parallel with the Meers fault at some distance. Stratigraphic throw across these faults increases westward toward the abrupt bends in fault trend. The strata exposed in the Slick Hills are tightly folded, and where the Arbuckle Group carbonates are steeply dipping, one recognizes a “tombstone” topography littered with broken slabs of dolomitized limestones. These “tombstone” layers are separated from each other by weathered intervals formed above more shaley interbeds. In some locations, the Arbuckle Group carbonates are nearly isoclinally folded. The general map pattern of folds in the region shows a series of en-échélon double-plunging folds that splay off a dominant northwest trend parallel with the Meers fault (Figure 3.6). The south-dipping, large-displacement Mountain View fault zone separates the Wichita-Slick Hills uplift blocks from the deep Anadarko basin (Plate 3.7). Maximum down-to-north offset of Arbuckle Group carbonates across the Mountain View fault exceeds 9 km (Johnson, 1991).

The isolated Wichita Mountain and Slick Hills terrains are surrounded and unconformably overlain by Permian-age conglomerates and younger deposits. Wolcampian rocks rest unconformably above older rocks around the entire Southern Oklahoma aulacogen. Surface exposures and wells show that conglomeratic facies of Permian rocks thicken towards the basement uplift regions of the Southern Oklahoma aulacogen (Johnson and others, 1988). The conglomeratic facies of the Permian interfingers with red-bed shales and evaporites in the Anadarko basin and basins to the south of the Wichita uplift (Johnson and others, 1988).
**Arbuckle uplift (surface and subsurface)**

Southeast of the Wichita Mountains uplift and Slick Hills, separated by a width of more than 100 km of Permian to recent cover rocks, pre-Mississippian strata crop out in the Arbuckle uplift of southeastern Oklahoma (Plates 3.7 and 3.9). The Arbuckle uplift is separated into several fault-bounded uplift blocks. One of these, the Arbuckle anticline, is located in the Western Arbuckle Mountains (Plate 3.9). The other uplift blocks, the Tishomingo and Belton anticlines, Hunton arch, Lawrence uplift, and Clarita anticline are located in the Eastern Arbuckle Mountains (Plate 3.9).

In the Arbuckle anticline of the Western Arbuckle Mountains (Plate 3.9), two cores of Cambrian Colbert Rhyolite (equivalent to Carlton Rhyolite exposed in the Slick Hills farther west) are surrounded by late Cambrian-Mississippian pre-orogenic passive-margin carbonates and shallow-water clastic rocks (Timbered Hills Group through Sycamore Limestone). On the southwestern flank of the Arbuckle anticline, strata dip south. Along the southernmost exposed units of the Arbuckle anticline, dip increases southeastward from 30° to nearly 90° at the Washita Valley fault zone (Ham and others, 1954).

The northwest-southeast-striking Washita Valley fault zone separates the Arbuckle anticline on the southwest, from the Tishomingo anticline on the northeast (Plate 3.9). The Washita Valley fault zone is exhibits variations of stratigraphic throw and direction along strike (Plate 3.9). Surface relationships along the northeastern flank of the Arbuckle anticline in the Western Arbuckle Mountains indicate north-directed offset along a steeply south-dipping fault. Farther southeast along strike, along the southwestern flank of the Tishomingo anticline in the Eastern Arbuckle Mountains, the Washita Valley fault zone is an anastomosing series of near-vertical faults with little stratigraphic throw. Farther to the southeast along strike in the subsurface, well and seismic data show the Washita Valley fault as a south-directed reverse-fault with a steep north dip (Huffman and others, 1978, 1987; and proprietary seismic reflection profiles).

Because it is impossible for the attitude and displacement of a single fault surface to vary in such a manner, the Washita Valley fault zone must represent a series of faults that intersect in a narrow zone. A careful examination of the Ham and others (1954) geologic map of the Arbuckle Mountains shows that the north-vergent reverse fault that marks the northern boundary of the Arbuckle anticline terminates along strike to the southeast (Plate 3.9). Conversely, the map also shows that the south-vergent reverse-fault that marks the southern boundary of the Tishomingo anticline terminates along strike to the northwest (Plate 3.9). The Washita Valley fault zone is roughly
parallel with the Meers fault farther to the west. Both these faults parallel the general
trend of the Southern Oklahoma aulacogen (Plate 3.9).

Several northwest-southeast-striking fault zones separate the Eastern Arbuckle
Mountains into a series of uplifts (anticlines) and fault-bounded synclines (Plate 3.9).
Differing amount of vertical motion across the fault zones beneath the Permian
(Wolfcampian) unconformity is evidenced by exposure of different parts of the
Cambrian-Pennsylvanian succession in different locations. Precambrian-Cambrian
igneous basement and Cambrian to Ordovician pre-orogenic Arbuckle Group carbonates
are found along the crests of anticlines, whereas Late Ordovician Simpson Group to
Mississipian-lower Pennsylvanian syn-orogenic foreland facies Caney through
Goddard-Springer formations are found in the troughs of synclines (Plate 3.9).

In the western part of the eastern Arbuckle Mountains and western Arbuckle
Mountains, the general northwest-plunge of folds northeast of the Washita Valley fault
zone contrasts with the overall southeast plunge of folds on the southwestern side of the
fault zone. The trends of smaller folds are also rotated clockwise approximately 20°
from the overall strike of the Washita Valley fault zone (Plate 3.9). Numerous faults that
splay off the Washita Valley fault zone also make similar angles (Plate 3.9). The
contrast of fold plunges and apparent rotation of fold axes away from strike of adjacent
faults suggests a transpressional tectonic setting. Historic arguments have been made
suggesting that some of the structures in the region represent strike-slip “flower-type”
structures (Harding, 1974; Harding and Lowell, 1979). That idea is not argued here;
however, evidence for oblique slip of some magnitude is compelling.

The near-vertical Sulfur fault zone separates the Hunton arch to the north, from
the Belton anticline to the south (Plate 3.9). The geologic map of Ham and others
(1954) shows that rocks are more tightly folded in the area between the Sulfur and
Washita Valley fault zones than to the north in the Hunton arch (Plate 3.9). The
northwestern end of the Belton anticline separates the Sulfur syncline on the north from
the Mill Creek syncline on the south (Plate 3.9). The steep, down-to-south Mill Creek
fault separates the Belton anticline from the Mill Creek syncline (Plate 3.9). The Mill
Creek “syncline” is a fault-dissected asymmetric northwest-plunging syncline-anticline
pair. Southeast along strike, the Mill Creek fault merges with Blue River fault zone
(Plate 3.9). The north-vergent Reagan reverse fault marks the southern boundary of the
Mill Creek syncline, and the fault merges southeastward along strike with the Mill
Creek-Blue River fault zone (Plate 3.9). Ham and others (1954) suggest that the Blue
River fault loses displacement to the southeast near the margin of the Mesozoic Gulf
Coastal Plain (Plate 3.9). However, subsurface well and seismic data indicate that the Blue River fault zone bends slightly southward and continues beneath Mesozoic cover (Huffman and others, 1978, 1987; Gatewood and Fay, 1991).

Multiple northwest-plunging tight folds are located along the western part of the Eastern Arbuckle Mountains south of the Sulfur fault, within the Sulfur and Mill Creek synclines, and within an area to the southwest between the Reagan and Washita Valley fault zones (Plate 3.9). These tight folds of relatively thin-bedded Upper Ordovician to Pennsylvanian (Desmoinesian) limestones and clastic rocks grade downsection southeastward along along fold trend into the broad Tishomingo and Belton anticlines where thick-bedded Cambrian-Ordovician Timbered Hills and Arbuckle Group carbonates (and lesser clastic units) surround a core of Precambrian and Cambrian igneous rocks located farther southwest (Plate 3.9).

In the eastern part of the Arbuckle uplift, the southeastern ends of the Tishomingo and Belton anticlines appear to merge and plunge southeast beneath the Ouachita thrust front (Plates 3.1, 3.7, and 3.9). The Belton anticline is not clearly defined southeast of the northern margin of the Mesozoic Gulf Coastal Plain (Plate 3.9). Well data indicate that shallow-water Cambrian-Pennsylvanian rock formations flank the igneous core of the southeastern end of the Tishomingo anticline beneath Mesozoic cover (Huffman and others, 1978, 1987, Oklahoma Geological Survey, formation-top data). Three fault-bounded depressions known as the “Coleman half-grabens” are located on the southeastern end of the Tishomingo anticline (Plates 3.7 and 3.9) (Ham and others, 1954; Huffman and others, 1978, 1987; Gatewood and Fay, 1991; proprietary seismic reflection profiles). The westernmost of the “Coleman half-grabens” is shown on Plate 3.9 as the Coleman syncline.

North of the Tishomingo anticline and north of the Sulfur fault zone, Ordovician upper Arbuckle Group carbonates are exposed on the crest of the fault-dissected Hunton arch (Plates 3.7 and 3.9). Upper Ordovician Simpson Group rocks are exposed in several fault-bounded synclines located along the western margin of the Hunton arch (Plates 3.7 and 3.9). Here, Ordovician strata extend unconformably beneath roughly northwest-striking, gently folded Permian conglomerates and shales (Plate 3.9). The central part of the Hunton arch extends eastwards as the east-plunging Clarita anticline (Plate 3.9). The eastern tip of the Clarita anticline bends to the north and is aligned with the basement fault trends of the western Arkoma basin (Plates 3.1, 3.4, and 3.9). Ordovician to Pennsylvanian (Morrowan) rocks are exposed within the Clarita anticline, and the northeast-striking, south-dipping, down-to-south Bromide-Olney fault zone
separates the southern margin of the Clarita anticline from the Wapanucka syncline (graben) to the south (Plate 3.9). The Bromide-Olney fault zone is a composite of many steep south-dipping, southwest-northeast striking normal faults. The southeast-striking, north-dipping, down-to-north Franks and Clarita fault zones separate the northern margin of the Clarita anticline from the Franks syncline (graben) to the north (Plate 3.9).

The Bromide-Olney fault zone marks the southern boundary of the Clarita anticline, and the southwestern end of the fault zone intersects the Sulfur fault (Plate 3.9). Stratigraphic throw of the Bromide-Olney fault is more than 210 m down-to-south (Ham and others, 1954). Stratigraphic throw of the Sulfur fault zone increases southeastward along strike. In the center of the Hunton arch, where the Sulfur fault is nearly vertical, stratigraphic throw is less than 152 m down-to-north (Ham and others, 1954). South of the Wapanucka syncline, where the Sulfur fault is mapped as a north-vergent reverse fault, cumulative estimated stratigraphic throw across the fault zone is more than 3000 m down-to-north (Plate 3.9) (Ham and others, 1954). Here, a narrow faulted zone of steep to vertical, north-dipping Ordovician to Mississippian rocks separates Precambrian igneous rocks on the south from gently folded Pennsylvanian rocks exposed at the surface on the north in the Wapanucka syncline (Ham and others, 1954). The Wapanucka syncline plunges to the east toward the western Arkoma basin.

Subsurface well and seismic data southwest of the northern edge of the Mesozoic Gulf Coastal Plain show that the Sulfur fault is a large-displacement fault beneath the Ouachita thrust belt at the southeastern terminus of the Arbuckle uplift (northeastern side of the southeast-plunging nose of the Tishomingo anticline) (Plates 3.7 and 3.9) (Tarr and others, 1965; Huffman and others, 1978, 1987; Gatewood and Fay, 1991; Oklahoma Geological Survey, formation-top data; proprietary seismic reflection profiles). Figure 3.7 includes a generalized cross section which is perpendicular to the trend of the Tishomingo anticline and shows the basal detachment of the Ouachita allochthon folded sharply over the Sulfur fault zone. The nature of the root zone of the Sulfur fault at depth is uncertain; however, west of the Mesozoic Coastal Plain in the center of the Arbuckle uplift, the Sulfur fault separates the tightly-folded rocks on the south from the relatively flat Hunton arch on the north (Plates 3.4 and 3.9).

North of the Clarita anticline, the western margin of the Franks syncline (graben) is delineated by a number of northwest-striking, northeast-dipping faults (Clarita and Franks fault zones) that are intersected and offset by numerous east- and northeast-striking faults (Ham and others, 1954) (Plate 3.9). Several longer and larger-offset northeast-striking faults, the Coal Creek fault zone, extends from the eastern Hunton arch
into the western Franks syncline (graben) (Plate 3.9). Roughly parallel to, and north of the Coal Creek fault zone is the Stonewall fault which separates the northern part of the Franks syncline located on the south side of the fault from the Lawrence arch on the north side (Plate 3.9). The Franks fault is nearly vertical with a down-to-north throw, whereas the Stonewall fault has a steep southward dip and has a down-to-south throw (Ham and others, 1954). Estimated maximum displacement of the Franks-Clarita and Stonewall fault zones exceeds 610 meters in the western part of the Franks syncline (Ham and others, 1954). Displacement across the southeast-striking Franks-Clarita fault zone (Plate 3.9) decreases southeastward along strike (Ham and others, 1954). The Clarita fault is not mapped east of the Clarita anticline where the fault offsets only the lowest part of the Atoka Formation (Ham and others, 1954). In contrast, the nearly east-west striking Coal Creek and Stonewall faults lose displacement in the younger Desmoinesian and Missourian clastic rocks which rest unconformably above the Atoka Formation (Ham and others, 1954). Abrupt thickening of, and prevalence of conglomerates in, the Desmoinesian-Missourian succession in the Franks syncline suggest that the Hunton arch and Lawrence arch began uplift prior to the Desmoinesian (Ham and others, 1954). Same as the Wapanucka syncline, the Franks syncline plunges east towards the western Arkoma basin.

North of the Stonewall fault, the Lawrence arch is the northernmost fault-bounded uplift of the Eastern Arbuckle Mountains (Plates 3.7 and 3.9). Upper Ordovician Simpson Group through Pennsylvanian (Morrowan) Wapanucka Formation rocks dip gently towards the east-northeast (Ham and others, 1954). From east to west along the northern Lawrence arch, the base of the Desmoinesian unconformity covers progressively older rocks. In the east, Desmoinesian rocks rest unconformably upon Morrowan Wapanucka Formation limestones (Ham and others, 1954). In the west, Desmoinesian rocks rest unconformably upon Late Ordovician Viola Group limestones and Sylvan Shale (Ham and others, 1954). On the western side of the Lawrence uplift, Desmoinesian and Missourian rocks extend under northwest-dipping Virgilian and Permian rocks which rest unconformably upon all older stratigraphy along the margins of basement uplifts within the Southern Oklahoma aulacogen (Ham and others, 1954).

**Ardmore and Marietta basins (synclines), Sherman fault block and Muenster arch (subsurface)**

In the subsurface beneath the Mesozoic Gulf Coastal Plain, the southeastern end of the Washita Valley fault zone separates the Tishomingo anticline on the northeast from three several long, northwest-trending fault-bounded uplifts located on the
southwest. Flanking the southwestern margin of the Tishomingo anticline, the Ravia-Cumberland-Sand Canyon nappe fault block (Plate 3.1 and Figure 3.7) contains a southwest-overturned anticline of Cambrian-Late Ordovician carbonates (Arbuckle Group-Simpson Group) (Huffman and others, 1978). The Cumberland syncline separates the Ravia-Cumberland-Sand Canyon nappe on the northeast, from a northwest-trending zone of basement uplifts known as the Mannsville-Madyll-Aylesworth flexure on the southwest (Figure 3.7) (Huffman and others, 1987). Located still farther southwest is the near vertical, down-to south Bryan fault which separates basement uplifts to its north from the deep Ardmore basin to its south (Figure 3.7) (Huffman and others, 1987).

The Ardmore basin (syncline) axis plunges southeast toward the Ouachita thrust belt (Plate 3.7). The depth of the Ardmore basin increases towards both the Anadarko basin to the northwest and the buried Ouachita thrust belt to the southeast. The depth to igneous basement in the central Ardmore basin is approximately 10 km (Cooper, 1995; Ewing, 1991). Estimates of maximum depth to igneous basement within the deepest parts of the Ardmore and Anadarko basins northwest of the Ouachita thrust front range from just more than 11 km (Davis and Northcutt, 1989) to nearly 14 km (Ewing, 1991). Depths to igneous basement are based upon structure contour maps of top of Arbuckle Group illustrated in Ewing (1991) and Davis and Northcutt (1989). An additional 1976 m (maximum measured thickness of Timberred Hills-Arbuckle Group listed in Ham, 1973) are added to the top-of-Arbuckle depths to give the estimated top of basement depths. A proprietary seismic reflection profile indicates that the depth to the igneous basement nears 12 km in the southeastern part of the Ardmore basin beneath the Ouachita thrust belt.

The narrow Criner Hills uplift separates the southwestern edge of the deep Ardmore basin on the north from the shallow Marietta basin to the south (Plate 3.7). The Criner Hills uplift fault zone extends southeastward from the southeastern part of the Wichita uplift toward the Ouachita thrust belt (Plate 3.7). Along this fault zone southwest of the Arbuckle Mountains, surrounded by Pennsylvanian and Permian rocks, are the isolated Ordovician-Mississippian rocks of the Criner Hills (Plate 3.7). The southeastern end of the Criner Hills fault zone extends beneath the Ouachita thrust belt where it merges with the northeastern edge of the Sherman fault block (Plate 3.7) (Bradfield, 1957a-c; Ewing, 1991; proprietary seismic reflection profiles). The Marietta basin narrows southeastward and wedges out where the Criner Hills fault zone merges with the Sherman fault block.
The Sherman fault block is a northwest-southeast trending fault-bounded segment of the eastern end of the Muenster arch (Plates 3.1 and 3.7). The southeastern end of the Sherman fault block is wider than the northwestern end where it connects with Muenster arch (Ewing, 1991). A narrow down-dropped fault block separates the southwestern side of the Sherman fault block from the northeastern side of the southeast-plunging nose of the Muenster arch (Bradfield, 1957a-c; Ewing, 1991; seismic reflection profiles). The southeastern end of the Muenster arch dips eastward toward the Ouachita thrust front. The Muenster arch and the Criner Hills uplift fault zone on the northeast merges to the northwest into the southeastern Wichita uplift (Plate 3.7).

Fort Worth / Strawn basin

The northwest-southeast-striking Muenster-Waurika fault zone separates the southern part of the Muenster arch from the northwestern part of the Fort-Worth/Strawn basin (Plate 3.7). The Muenster-Waurika fault zone extends primarily in the subsurface from the southeastern interior of the Wichita uplift southeastwards to the Ouachita thrust front. Brewer (1982) maps the northwestern part of the fault zone as a southwest-vergent reverse fault, whereas Ewing (1991) maps the southeastern part as a southdipping normal fault.

The Fort Worth/Strawn basin is a composite foreland and cratonic basin located south of the Muenster arch and north of the Llano uplift (Plate 3.7). The deepest part of the Fort Worth/Strawn basin trends northeast-southwest and is parallel with the Ouachita thrust front south of the Muenster arch (Plate 3.7). Sedimentary rocks of the deep Fort Worth foreland basin extend northwestwards into the Strawn cratonic basin which trends northwest-southeast and separates the southern part of the Muenster arch from the northern part of the Llano uplift (Plate 3.7). In the deepest part of the basin, east of the Bend arch (Plate 3.7), a thick sequence of Morrowan-Desmoinesian clastic rocks, which rest upon a thinner succession of Cambrian-Ordovician shelf carbonates and shales, dips eastwards towards the Ouachita thrust front (Kier and others, 1979). On the crest of the Bend arch and farther west, Desmosnesian-Permian clastic rocks and carbonates, which rest unconformably upon Cambrian-Atokan shelf carbonates, dip westward towards the Midland basin (Kier and others, 1979) (Plate 3.7).
Matador-Red River arch, and Bend arch:

In northern Texas, the northwest-southeast-trending Muenster arch intersects the east-west trending Matador-Red River basement arch (Plates 3.1 and 3.7). On the southwestern flank of the Wichita-Amarillo uplift trend, the Palo Duro and Hollis-Hardeman basins are north of the Matador-Red River uplift trend (Plates 3.1 and 3.7). The eastern end of the Palo Duro and Hollis-Hardeman basins wedge-out between the Wichita-Muenster and Matador-Red River uplift trends. The north-plunging, northern end of the Bend arch is located just south of the intersection of the Muenster and Matador-Red River basement arches and the southern end of the Bend arch merges with the Llano uplift where pre-Pennsylvanian rocks are exposed in the core (Plate 3.7). The Bend arch separates the Fort Worth/Strawn foreland basin on the east from the Midland cratonic basin on the west (Plate 3.7).

Intersection of thick-skinned and thin-skinned structures along the western Ouachita salient (Tishomingo small-scale recess, Bryan small-scale salient, and northern Texas recess)

Seismic and well data show that in the southeastern part of the Ardmore basin (syncline), imbricated “Ouachita facies” rocks are thrust over foreland “Arbuckle facies” rocks (Plate 3.7) (Huffman and others, 1978). Seismic and well data also show that the Bryan fault and Mannsville-Madyll-Aylesworth structures continue southeastward along strike beneath the imbricated Ouachita thrust belt (Huffman and others, 1978, 1987). The southeastern part of the Bryan fault coincides with the northern part of the Bryan small-scale salient (BSSS) (Huffman and others, 1978 and Huffman and different others, 1987) (Plate 3.7). The BSSS is a small westward bend in the Ouachita thrust front south of the southeastern end of the Tishomingo anticline (Plate 3.7). Seismic and well data indicate that north of the BSSS, the Ouachita thrust front bends sharply towards the east around the southeastern nose of the Tishomingo anticline (Plate 3.7) (Huffman and others, 1978, 1987; proprietary seismic reflection profiles). The eastward bend in the Ouachita thrust front across the southeastern end of the Tishomingo anticline is show in in Plate 3.7 as the Tishomingo small-scale recess (TSSR) (Plate 3.7).

The roughly northeast-southwest-striking, convex-to-northwest Kingston thrust fault marks the western boundary of the BSSS (Huffman and others, 1987; Hardie, 1990) (Plate 3.7). This part of the BSSS, crossing the Texas-Oklahoma border, has numerous commercial petroleum wells and is known as the Isom Springs oil well field.
Huffman and others, 1987). Hydrocarbons are trapped between thrust-fault imbricates of Ordovician-early Mississippian Bigfork Chert-Arkansas Novaculite (Huffman and others, 1987). Allochthonous strata in the western part of the BSSS are steeply dipping to vertical and, in some places, are overturned toward the center of the salient (Huffman and others, 1987; Hardie, 1990). Well data from the Isom Springs field show the Pine Top Chert beneath the Arkansas Novaculite (Huffman and others, 1987). Pine Top Chert is a Devonian transitional facies chert that is also found in isolated fault blocks on the southern side of the Pine Mountain thrust fault in the western part of the Ouachita Frontal Ouachita thrust belt (Hardie, 1988; Arbenz, 1989b). The presence of Pine Top Chert suggests that the Ouachita facies rocks of the western BSSS may have initially been located near the transition zone between shallow-water and deep-water facies.

The southern end of the Kingston fault zone bends sharply to the southeast and continues as the southern boundary of the BSSS which extends southeastward along the northern flank of the Muenster arch (Plate 3.7). The fault marking the southern boundary of the Bryan salient bends sharply to the southwest at the southeastern end of the Muenster arch (Plate 3.7). The frontal fault of the Ouachita thrust belt continues with a southwest-northeast strike from the Muenster arch to just southeast of the Llano uplift in central Texas (Plate 3.7). This segment of the Ouachita thrust front is the northern part of the Texas recess (Plate 3.1).

**Waco uplift (basement uplift beneath Ouachita allochthon)**

East of the Fort Worth basin in the subsurface beneath both the allochthonous strata of the Ouachita orogen and the Mesozoic Gulf Coastal Plain is the Waco basement uplift. Regional seismic reflection profiles, gravity anomaly maps, and well data indicate that the Waco uplift is a northeast-trending, double-plunging basement arch that is parallel with the Ouachita thrust front (Flawn and others, 1961; Nicholas and Rozendal, 1975; Kruger and Keller, 1986; Nicholas and Waddell, 1989). From northwest to southeast across the Waco uplift, thickness of the Mesozoic cover strata increases from less than 150 m (west of the Ouachita thrust front) to more than 4700 m at the western edge East Texas basin (Nicholas, 1989).

Cross section F-F’ in Plate 3.10 shows a modified version of part of the Nicholas (1989) cross section across the Waco uplift. The Arco (Atlantic) No. 1 Chapin well, located 5 km east of the frontal thrust fault, penetrated nearly 4000 m of deformed allochthonous Ordovician through Devonian slates and cherts (Womble-Arkansas Novaculite type) before penetrating a thin sequence of foreland facies sandstones and shales. The well Bottomed in the upper part of the Ordovician Ellenburger shallow-water
carbonates. West of the Ouachita frontal thrust fault towards the Bend arch, west-dipping Atokan “gray” Strawn Group clastic rocks rest unconformably upon east-dipping Cambrian-Ordovician Ellenburger Group carbonates (Cheney, 1929; Kier, 1988). Near the Ouachita thrust front (and likely farther east beneath the Ouachita allochthon), Atokan “gray” Strawn Group clastic rocks rest upon a thin succession of Mississippian-Atokan shales and limestones (Chappel-Barnett-Marble Falls-Smithwick) (Cheney, 1929; Kier, 1988). Depending on location and proximity to the Ouachita thrust front, the thin Mississippian-Atokan strata rest unconformably either upon very thin Upper Ordovician to Devonian strata, or directly upon Cambrian-Early Ordovician Ellenburger Group shelf carbonates (Cheney, 1929; Kier, 1988). Progressively more of the Upper Ordovician-Atokan succession is preserved eastward towards the Ouachita thrust front (Cheney, 1929; Kier, 1988).

East of the Ouachita frontal thrust fault, the basal décollement dips eastwards towards the Luling metamorphic front (Plates 3.7 and 3.10). The Luling front separates the Waco uplift on the east from the frontal zone of the Ouachita thrust belt on the west. The basal décollement ramps westward over several listric normal fault blocks with east-dipping, down-to-east throws (3.10). Cumulative displacement of autochthonous terrain across these normal faults beneath the basal décollement is nearly 6 km (3.10). Nicholas (cross section F-F’, 1989) shows a deformed layer of Ordovician-Devonian chert beds above the basal décollement extending from the frontal thrust fault eastward toward the Luling front (Plate 3.10). Above the chert beds is a contorted succession of Mississippian to Pennsylvanian synorogenic deep-water sandstone and shale turbidites. These strata roughly correlate with the Stanley Group-Jackfork Sandstone-Johns Valley Shale (and possibly deep-water Atoka) formations of the Ouachita Mountains.

Cross section F-F’ illustrated in Plate 3.10 shows that several wells the penetrate Paleozoic rocks beneath Mesozoic cover strata. One of these, the Shell No. 1 Barrett well located on the crest of the Waco uplift arch penetrated 2954 m of highly deformed and weakly metamorphosed “phyllites and quartzites with calcite and quartz veins” beneath the basal Mesozoic unconformity (Nicholas and Rozendal, 1975). The seismic reflection profile through the Waco uplift show these strata as “transparent” or without clear reflectors, suggesting tight folds (Nicholas and Rozendal, 1975). Below the deformed clastic rocks, the Barrett well penetrated 1847 m of “intensely deformed carbonates” (Nicholas and Rozendal, 1975). The top of the carbonates is a distinct seismic reflector. Although devoid of useful fossils, comparison of lithology with carbonates encountered in wells and in Paleozoic outcrop belts indicates that these
carbonate rocks are Cambrian-Devonian (or lower Mississippian) shallow-water carbonates (Nicholas and Rozendal, 1975). Beneath the carbonate strata, the Barret well penetrated a thin 38 m unit of “quartz-biotite schist” and a “basal quartzite,” before it bottomed in 162 m of “quartz-diorite” (interpreted as Precambrian) (Nicholas and Rozendal, 1975).

**Llano uplift**

The Llano uplift is on the southeastern end of the Bend arch (Plates 3.7 and 3.10). The central part of the Llano uplift and much of the crest of the Bend arch are not covered by the Mesozoic Gulf Coastal Plain; however, the eastern part of the Llano uplift, and adjacent Ouachita thrust belt are covered by Mesozoic strata (Plates 3.7 and 3.10). The southern part of the Fort Worth basin narrows along the southeastern side of the Llano uplift. Southwest of the Llano uplift along the Marathon orogenic belt are the Val-Verde and Kerr basins and Devils River and Marathon uplifts of southwestern Texas (Plate 3.7). The Marathon orogenic belt is the continuation of the Ouachita thrust belt in the area south of the Llano uplift (Texas recess), and the sharp westward bend in the thrust front southwest of the Llano uplift is the Marathon salient (Plates 3.1 and 3.7). The strike of the frontal thrust fault and interior structures of the Ouachita orogen bend 90° from a northeast-strike east of the Llano uplift, to a northwest-strike on the southwestern side of the Llano uplift (Marathon orogen) (Plates 3.1 and 3.7). At the western end of the Val-Verde basin, the strike of the Marathon orogen bends 90° from a northwest-strike to a northeast-strike (Plates 3.1 and 3.7). The Marathon orogen continues with a northeast-strike through the Marathon uplift of southwest Texas, and farther south into Mexico (Plates 3.1 and 3.7).

The core of the Llano uplift contains Precambrian meta-igneous and metasedimentary rocks (Mosher, 1993) is surrounded successively by Cambrian to Ordovician Ellenburger Group carbonates through Atokan shales and sandstones. East of the Llano uplift, in the autochthonous terrain beneath the Ouachita thrust front, west-dipping Desmoinesian clastic rocks rest upon Cambrian-Atokan strata (adapted from Cheney, 1929; Kier and others, 1979; Nicholas, 1989). Southwest of the Llano uplift, a much thicker north-dipping Desmoinesian-Permian succession of clastic rocks on-laps upon south-dipping Cambrian-Atokan shallow-water carbonate strata (Nicholas, 1989). The combination of the west-dips of the Atokan “grey” Strawn strata located in the foreland to the west of the Waco uplift, the north-dips of the Desmoinesian-Permian strata located in the foreland to the southwest of the Llano uplift, and north of the Marathon thrust front, suggests that the Ouachita and Marathon orogens were being
uplifted during the Atokan. Map patterns of the Llano uplift and surrounding areas combined with cross sections of the Bend arch show that Desmoinesian-Permian strata prograde westward toward the Midland basin (Cleaves, 1996). The westward progradation of the Desmoinesian-Permian succession suggests that the Midland basin was subsiding, whereas the frontal zone of the Ouachita-Marathon orogen farther east was being uplifted (Cleaves, 1996). However, Desmoinesian-Permian strata that unconformably overly strata of the Ouachita orogen southeast of the thrust front of the Ouachita salient indicate the southeastern interior of the Ouachita orogen also began subsiding in the Desmoinesian (Plate 3.1) (Nicholas and Waddel, 1989).

The Llano uplift of central Texas differs in several respects from the Arbuckle uplift to the north in the southern Oklahoma. The Llano uplift is broad dome, whereas the Arbuckle uplift is a composite of northwest-southeast-elongated fault blocks and folds (Plates 3.1, 3.7 and 3.9). Faults have a predominant northeast strike in the Llano uplift; whereas, major faults in the Arbuckle uplift have a predominant northwest strike. Also, the frontal fault of the Ouachita-Marathon orogen does not sharply wrap around the southeastern terminus of the Llano uplift in the same manner as it bends sharply around the southeastern end of Arbuckle uplift (Tishomingo-Belton small scale recess) (Plate 3.7).

Cross section G-G’ in Plate 3.10 illustrates the general structure of the southeastern Llano uplift and adjacent Ouachita orogen. The cross section is constrained by wells and COCORP seismic reflection data (Cullotta and others, 1992). Thickness of Mesozoic cover strata increases from zero on the southeastern end of the Llano uplift to 2500 m at a distance of 120 km southeast (cross section G-G’ in Plate 3.10). Stratigraphy and structure of the frontal zone of the Ouachita allochthon and underlying autochthon are constrained by the Shell No. 1 Johnson well which penetrated nearly 1500 m of Late Ordovician allochthonous slates and chert beds beneath Mesozoic cover rocks (Plate 3.10). Beneath the chert beds, the Shell No. 1 Johnson well penetrated approximately 1200 m of Pennsylvanian (possibly Morrowan-Atokan Strawn Group) before it bottomed in Precambrian basement (cross section G-G’ in Plate 3.10).

As shown in cross section G-G’ in Plate 3.10, Ordovician-lower Mississippian chert-beds (dark gray color on map) pinch out 15 km southeast of the frontal thrust fault. Southeast of the pinchout, Mississippian-Pennsylvanian clastic rocks rest upon the basal décollément (cross section G-G’ in Plate 3.10). A simplified restoration of cross section G-G’ with a pre-Strawn datum shown in Plate 3.10 shows one possible explanation for the pinchout of Ordovician-lower Mississippian chert-beds southeast of the thrust front.
in cross section G-G’. The basal décollement crossed a down-to-southeast normal fault located at the restored location of the pinch-out. Another possible explanation, the Ordovician-Devonian chert-beds are truncated by an erosional unconformity on a pre-thrust structural high located southwest of the restored location of the pinch-out. Ordovician-Devonian chert beds also pinch-out southwards along the frontal thrust fault and are absent along the frontal thrust fault southwest of the Llano uplift (Plates 3.1, 3.7 and 3.10). Still another possible solution requires the basal detachment coincide with the base of Ordovician-lower Mississippian cherty strata southeast of fault A (inset map in Plate 3.10).

The southeastern half of cross section G-G’ shown in Plate 3.10 is not well constrained beneath the base of the Mesozoic unconformity. Seismic reflection data and data from the Shell No. 1 Forgy well, show a southeast-dipping, northwest-vergent thrust fault (fault A) splaying off of the basal décollement (Plate 3.10). In cross section G-G’ (Plate 10), Fault A separates a relatively broad northwest-tilted syncline on the northwest, from three smaller north-vergent folds which have short northwest-dipping limbs and much longer southeast-dipping limbs. Although only the uppermost beds are penetrated by wells, comparison with the cross section F-F’ (Plate 3.10) indicates that the strata between Fault A and the Luling front farther southeast is predominantly Mississippian-Pennsylvanian clastic rocks (Plates 3.1 and 3.10). Ordovician-Devonian cherty formations may, or may not, rest above the basal décollement.

The Luling front separates unmetamorphosed “Ouachita-facies” shales and sandstones on the northwest from weakly metamorphosed (at most lowest greenschist) sandstones, slates, and phyllites on the southeast (Plate 3.10). Cross section G-G’ in Plate 3.10 shows deformed strata above a northwest-vergent thrust fault that emerges at the Luling front. These contorted beds dip generally to the southeast (cross section G-G’ in Plate 3.10). Farther southeast, a thick succession of undeformed strata dip more steeply southeast (cross section G-G’, Plate 3.10). The discordance between the contorted and undeformed strata is similar to the discordance between the Mississippian-Pennsylvanian rocks and overlying Desmoinesian-Permian rocks determined from wells located in the Sabine uplift (Plate 3.1) (Nicholas and Waddell, 1989).
Figure 3.1: Pre-Mesozoic geologic map of the Black Warrior basin showing large-scale tectonic features.

References: Thomas, 1972b; Morris, 1974; Hatcher and others, 1989; Thomas, 1989; Thomas and others, 1989; Thomas, 1991.
Figure 3.2: Geologic map and cross section of the northern flank of the Broken Bow uplift, Linson Creek synclinorium, and Cross Mountains anticlinorium, southeastern Oklahoma.
Figure 3.3: Map of the western Ouachita Mountains and southwestern Arkoma basin showing distribution of deep wells. Map illustrates several major thrust faults of the Ouachita thrust belt and basement faults of the Arbuckle uplift. The thick dashed line shows the southernmost extent where deep wells penetrate the entire Simpson Group (Middle Ordovician, Whiterockian-Mohawkian shallow-water carbonates and clastic rocks) (Suhm, 1996). Two wells in the Broken Bow uplift (*) penetrated interpreted lower Simpson Group (McLish Fm) and underlying inferred Arbuckle Group dolostones (Late Cambrian-Early Ordovician) beneath a thick succession of allochthonous “Ouachita facies” deep-water strata (Leander and Legg, 1988; Denison, 1989).
Figure 3.4: Comparison between Bouguer gravity and residual gravity for part of the southern mid-continent region of North America. Part A is a Bouguer gravity map which is corrected to remove the gravity related to topography. Part B is a residual gravity map in which the regional trend (as represented by the second-order polynomial surface) was subtracted from the Bouguer gravity surface. Both maps are from Kruger and Keller (1986). Consult Kruger and Keller (1986) for further discussion of theory and model parameters.

In general, trends of the gravity anomalies in both maps are similar; however, anomalies are accentuated in the residual gravity map. Key features of both maps include a large negative gravity anomaly centered on southeastern Oklahoma, and several northwest-trending positive and negative anomalies along the Oklahoma-Texas border.
Figure 3.5: Large-scale tectonic features and faults of the Ouachita orogen and adjacent foreland to the northwest are superimposed upon a residual gravity map of part of the southern mid-continent region. The Arbuckle and Wichita uplifts, Ardmore basin, and Muenster arch together comprise the Southern Oklahoma "aulacogen" (foldbelt). The large gravity "low" centered on "A" is possibly an area where a thick succession of Precambrian sedimentary rocks underlie the Paleozoic succession (Eocambrian basin of Kruger and Keller, 1986). This large area of low gravity is oblique to strike of surface faults; therefore, this feature cannot be purely a foreland basin. Gravity low "B" may be another area where Precambrian sedimentary rocks are anomalously thick. Paleozoic strata thicken west of the Bend arch towards the Midland basin (off the map) (Johnson and others, 1988). Gravity lows "C and D" are likely areas where Mesozoic salt formations are thick, and, or areas where other Mesozoic sedimentary rocks or Paleozoic rocks are thicker than adjacent areas.
Figure 3.6: Geologic map of part of the Slick Hills of southwestern Oklahoma. The Cambrian-Ordovician strata of the Slick Hills are unconformably overlapped and surrounded by Permian strata. The Blue Creek Canyon fault separates the more tightly folded Western Slick Hills terrain, to the west, from the less contorted Eastern Slick Hills terrain. The Western Slick Hills terrain consists of a larger scale anticline-syncline pair that plunges to the northwest ("A" and "B"). These folds contain numerous smaller scale folds and are cross-cut by many reverse faults. The most laterally continuous of these is the Stumbling Bear fault. The Eastern Slick Hills terrain contains two large-scale folds that plunge towards the southeast ("C" and "D").

Note that many of the small-scale folds, and some of the small-scale reverse faults, within the Western Slick Hills terrain are oblique to the trend of the Meers fault. In some locations, such as "E," fold axes appear to be en échelon. The clockwise-rotated angle between many compressive structures within the Slick Hills suggests left-lateral transpressive strain. For this reason, the Meers fault is by some considered a left-lateral, high-angle, wrench fault (such as Donovan, 1986; Donovan and others, 1989); however, others suggest a less steeply dipping reverse fault geometry (Brewer, 1982; McConnell, 1988).
Figure 3.7: Generalized pre-Mesozoic structural geology of the southeastern Arbuckle uplift, southern Oklahoma. Part A is a generalized pre-Mesozoic geologic map of the eastern Arbuckle uplift. Major fault zones, structures, and intersection of Arbuckle terrain with Ouachita thrust belt are shown.

Part B is a generalized cross section of the southeastern end of the Arbuckle uplift that extends northeast from the Bryan small-scale salient (BSSS) to the western part of the Ouachita Mountains.

Map and cross section references shown above.
Chapter Four

Structural Geology of the Ouachita Mountains and Arkoma basin

Introduction

The Ouachita Mountains straddle the border between Oklahoma and Arkansas (Figure 4.1). Deformed and thrust-faulted rocks exposed in the Ouachita Mountains belong to the late Paleozoic Ouachita orogen. The Ouachita orogen extends south-eastward in the subsurface beneath the Gulf Coastal Plain into central Mississippi where it is overridden and truncated by rocks of the Appalachian orogen (Thomas, 1973, 1989). The Ouachita orogen extends south-eastward beneath the Gulf Coastal Plain toward central and western Texas (Figure 4.1). The orogen bends sharply north-westward south of the Llano uplift of central Texas, continues towards the southern end of the Delaware basin where the orogen bends sharply to the south-west and continues through the Marathon Mountains into Mexico. The Ouachita and Marathon Mountains regions constitute the only large outcrops of rocks belonging to the Ouachita orogen.

The study of the complex structural geology of the Ouachita Mountains of Oklahoma and Arkansas has often caused researchers to engage in heated discussions and debates with respect to regional tectonic models. Description of regional structural geology of the Ouachita Mountains and establishment of a tectonic model consistent with the structural geology depends upon analysis of several key parameters which are to variable degrees either known, inferred, or speculated (Arbenz, 1984).

Surface geology is the best known and described parameter that constrains interpretation of regional structural geology. Geologic maps based upon field mapping show distinct along-strike variation in fold wavelength and fault spacing within the Ouachita Mountains (Arbenz, 1989b; Haley and others, 1993). The Ouachita Mountains can be subdivided into three thrust belt zones: The Frontal Ouachita thrust belt, the Central Ouachita thrust belt, and the Southern Ouachita thrust belt (Figure 4.2). The Frontal Ouachita thrust belt changes eastward along strike from a zone of numerous, north-vergent thrust-fault imbricates to a zone of broad synclines with narrow, fault-cored anticlines (Figure 4.2) (Arbenz, 1989a, b, d). In the northern
part of the eastern Frontal Ouachita thrust belt, folds are upright to north-vergent; however, in the southern part, folds are predominantly south-vergent (Figure 4.2) (Viele, 1966a, 1974; Lillie and others, 1983; Blythe and others, 1988; J.K. Arbenz, unpublished cross sections, 1996). The Central Ouachita thrust belt changes eastward along strike from a zone of broad, upright and north-vergent folds of Carboniferous turbiditic sandstones and shales to a zone of south-vergent tight folds and folded thrust faults encompassed within the Benton uplift of Upper Cambrian-lower Mississippian carbonaceous deep-water shales, turbiditic sandstones, cherts and novaculites (Figure 4.2) (Morris, 1974; Arbenz, 1989a-d; Nielsen and others, 1989). The structural style of the Southern Ouachita thrust belt changes westward along strike from broad, north-vergent folds of Carboniferous turbidites to a zone of south-vergent, tightly folded Upper Cambrian-lower Mississippian carbonaceous clastic rocks and cherts in the Broken Bow uplift (Figure 4.2). From north to south across the central part of the Central and Southern Ouachita thrust belts, a wide area of Mississippian Stanley Group shales separates long wavelength folds in overlying strata from short wavelength folds in underlying strata (Figure 4.2).

Another important parameter that constrains interpretation of regional structural geology is subsurface depth control. Although structures of the allochthonous terrain are exposed within the Ouachita Mountains, the geometry of structures at depth is far less constrained. A COCORP seismic reflection profile, and a larger scale PASSCAL seismic reflection-refraction profile across the Ouachita Mountains, and regional gravity and magnetic anomaly patterns constrain geometry of the autochthon of the Ouachita Mountains region on a large scale (Lillie and others, 1983; Kruger and Keller, 1986; Keller and others, 1989; Mickus and Keller, 1992). The magnetic and gravity anomaly patterns are heavily overprinted by Mesozoic rift structures (Kane and Hildebrand, 1981; Keller and others, 1983; Hildenbrand, 1985; Kruger and Keller, 1986). Subsurface geology in the Frontal Ouachita thrust belt is constrained by numerous wells, seismic data, and subsurface projections of surface structures (Blythe and others, 1988; Arbenz, 1989a, d; Mazengarb, 1995; Valderamma and others, 1996). Except for the well located in the Broken Bow uplift region of southern Oklahoma.
(Sohio 1-20 Weyerhausser), and another located nearby in northeastern Texas (Hunt 1 Neeley), no well has penetrated autochthonous strata southeast of the Frontal Ouachita thrust belt within the Ouachita Mountains (Figure 4-2) (Leander and Legg, 1988; Denison, 1989; Gatewood and Fay, 1991).

The third major controlling parameter that appears to affect the structural complexity of the Ouachita thrust belt is variation in mechanical stratigraphy. Mechanical stratigraphy, or the variation in relative thickness of alternate stiff-layer rocks and weak-layer rocks within a stratigraphic succession affect fold and fault geometry.

Cross sections constructed across the southern Appalachian orogen, in which passive-margin shelf carbonates are imbricated, generally show décollements (flats) in ductile weak-layer shaley units or along abrupt changes in rock type (mechanical properties) (Thomas, 1987; Hatcher, Jr., and others, 1989). The transition from carbonate platform to off-shelf shales marks a significant change in mechanical properties (décollement C in Figure 3, Thomas, 1987). The shelf edge where these pre-orogenic facies changes are located is the most likely place for the basal décollement to be initiated. Increased fluid pressure caused by the weight of overlying syn-orogenic strata (which lowers effective stress) helps initiate motion within the basal off-shelf shale. It is likely that the frontal fault of the basal décollement parallels the shelf edge facies transition zone throughout the entire continental margin. The shape of this continental margin is reflected in the shape of the frontal fault of the allochthon, and to a lesser extent in the shape of thrust faults in the allochthon interior.

In contrast to the southern Appalachian thrust belt, where the ductile layer within the shelf facies is relatively thin, the Ouachita thrust-belt has a relatively thick composite ductile layer succession of strata (shales, turbidites, and cherts) which is overlain by a thicker stiff layer (turbiditic sandstones) (Figure 4.3) (Arbenz, 1989d; Titus and Legg, 1995). The structural geology of the Ouachita Mountains is further complicated because the lower weak layer is subdivided into a succession of alternate stiff layers (distal turbiditic sandstones, cherts, and novaculites) and weak layers (shales) (Figure 4.3). Folding of these alternate stiff and weak layers has resulted in folds of different wavelengths and shortening (Arbenz, 1989d; Titus and Legg, 1995).
The end result is stratigraphy that is disharmonically folded, and wavelengths of folds vary in proportion to thickness of stiff layers (Arbenz, 1989b; Titus and Legg, 1995). The weak-layer shales behave as wavelength transition zones which compensate for the different wavelengths of adjacent stiff layers.

The along-strike change of structural styles within the Ouachita Mountains combined with limited well control of deeper subsurface structures produces a variety of plausible cross sections constructed across various parts of the Ouachita Mountains (Blythe and others, 1988; Arbenz, 1989a; Viele, 1989). Taken by themselves, palinspastic restorations of these cross sections produce differing tectonic models. This chapter examines the along-strike changes in structural style and highlights some of the structural complexity of the region. In general, this chapter will examine the tectonic significance of variation in mechanical stratigraphy across the Ouachita Mountains and adjacent foreland.

**Mechanical stratigraphy: variations across Ouachita Mountains**

The stratigraphy of the Ouachita Mountains allochthonous terrain is separated into alternate formations that behave as stiff (rigid) and weak (ductile) layers. The lower part of the stratigraphic succession, the pre-orogenic Collier Formation (Late Cambrian) through Arkansas Novaculite (early Mississippian), contains several stiff layers (distal turbiditic sandstones and cherts) separated by weak-layer shales (Morris, 1974; Lowe, 1979; Titus and Legg, 1995). The stiff-layer formations, which are also more resistant to erosion, are more easily mappable in the field, and they produce the sinuous outcrop patterns shown on geologic maps and aerial photographs of the core regions of the Benton and Broken Bow uplifts (Melton, 1975; Arbenz, 1989b; Haley and others, 1993).

Titus and Legg (1995) describe the differences in observed wavelengths of the various stiff-layer folds for the Broken Bow uplift part of the Ouachita Mountains (Figures 4.2 and 4.3). Folds of the Lower Ordovician Crystal Mountain Sandstone, which is approximately 304 m thick, have the longest observed wavelength of the pre-orogenic formations, estimated at 9100 m (Figure 4.3). The stiff-layer Crystal Mountain Sandstone overlies the weak-layer Collier Formation (mostly Upper
Cambrian to Lower Ordovician shales). The approximate thickness of the Collier Formation is 304 meters. More than 760 m of predominantly weak-layer Ordovician shales of the Mazarn through Womble formations overlies the Crystal Mountain Sandstone (Figure 4.3). Locally, in the Broken Bow and Benton uplifts, the Mazarn and Womble shales are separated by the thin stiff-layer Blakely Sandstone (Figures 4.2 and 4.3). Folds of the Blakely Sandstone have a nearly 920 m wavelength which is superimposed upon broader folds similar in scale to the fold wavelength in the Crystal Mountain Sandstone. In the Broken Bow uplift, the weak-layer shales have observed wavelengths of less than 30 m (Titus and Legg, 1995) and most likely fill the volumes between adjacent disharmonic stiff-layer folds via a combination of flexural-slip and flexural-flow mechanisms.

The succession from the Late Cambrian-Ordovician Collier Formation through lower Womble Shale can be considered as the lower mechanical stratigraphic unit of the Ouachita Mountains allochthonous terrain (Figure 4.3). The large-scale fold geometries of this succession conform to the wavelength of the folds of the Crystal Mountain Sandstone (the dominant stiff-layer). A disharmonic boundary in the Womble Shale separates the lower mechanical stratigraphic unit from a more complexly interbedded middle unit. The middle unit consists of part of the Ordovician Womble Shale, Silurian through early Mississippian (Osagean) cherts, novaculites, shales, and distal turbiditic sandstones (Bigfork Chert-Arkansas Novaculite) and the lower Mississippian (Meramecian) syn-orogenic shales and tuffs of the lower Stanley Group. The estimated total thickness of these formations in the Broken Bow region is 729 m (Figure 4.3).

The middle mechanical stratigraphic unit could indeed be separated into several smaller sub-units each with a thin stiff-layer (Figure 4.3). However, the fold geometries of this middle unit appear to conform to the wavelength of folds of the Ordovician Bigfork Chert (Figure 4.3). Folds of the Bigfork Chert have an observed wavelength of 3040 m (Figure 4.3). The shorter 608 m fold wavelength of folds of the overlying Devonian-lower Mississippian Arkansas Novaculite appear to be parasitic folds that conform to the broader folds of the Bigfork Chert (Figure 4.3). A dishar-
monic boundary coinciding with the Silurian Missouri Mountain Shale and locally discontinuous Blaylock Sandstone separates the Arkansas Novaculite folds from the Bigfork Chert folds. Another disharmonic boundary separates the Arkansas Novaculite folds from the 304 to 456 m wavelength folds of the stiff-layer tuffs in the lower part of the early Mississippian Stanley Group (Figure 4.3).

Across most of the Ouachita Mountains, the upper mechanical stratigraphic unit consists of a succession of synorogenic clastic-wedge deep-water sandstone and shale turbidites that ranges in stratigraphic thickness from two to four times greater than the middle and lower mechanical stratigraphic units combined (Arbenz, 1989a, Lillie and others, 1983). Figure 4.3 shows combined tectonic thickness of the lower and middle mechanical stratigraphic units in the Broken Bow uplift. A major regional disharmonic boundary separates the upper mechanical stratigraphic unit from the underlying middle unit. Approximately 122 m of tightly folded lower Stanley Group rocks are included with the middle mechanical stratigraphic unit in the Broken Bow uplift region (Figure 4.3). The rest of the Mississippian Stanley Group and the lower Pennsylvanian Jackfork Sandstone, Johns Valley Shale, and interbedded sandstones and shales of the Atoka Formation belong to the upper mechanical stratigraphic unit. The total estimated thickness of this upper unit is approximately 7290 m around the Broken Bow uplift (Figure 4.3).

Adjacent to the Broken Bow and Benton uplifts of the southeastern part, and across much of the western part of the Ouachita Mountains, the upper mechanical stratigraphic unit can be subdivided into a basal weak-layer and an upper stiff-layer. The basal 2130 m of the upper mechanical stratigraphic unit, lower and middle Stanley Group, behaves as a weak-layer with observed fold wavelengths of less than 30 m (Figure 4.3). A disharmonic boundary separates the top of the weak-layer Stanley Group from the overlying stiff-layer rocks. The overlying 5170 m of upper Stanley Group (Chesterian) through Atoka Formation (Atokan) behaves as the dominant stiff layer in the parts of the Ouachita Mountains surrounding the core regions of the Benton and Broken Bow uplifts (and presumably above them prior to uplift and erosion) (Figures 4.2 and 4.3). These dominant stiff-layer formations are deformed
into broad folds with wavelengths greater than 9120 m, and locally are disrupted by thrust faults (Figure 4.3).

**Structural geology of the Arkoma basin (overview)**

The Arkoma basin of eastern Oklahoma and Arkansas is cross cut by both thin-skinned thrust faults which root in a shallow-level basal décollement and thick-skinned basement faults which root in Precambrian basement. The southern and western parts of the Arkoma basin contain several isolated thin-skinned north- and northwest-vergent thrust faults and south-vergent backthrusts (Figure 4.2 and Plate 4.1). Cross sections of the Arkoma basin show that, depending on location and proximity to the Ouachita thrust front, the thin-skinned faults root either in a lower Atokan or Morrowan shale detachment (Lillie and others, 1983; Blythe and others, 1988; Arbenz, 1989a, d; Van Arsdale and Schweig, 1990; Valderamma and others, 1996). Basement faults cross the entire Arkoma basin from the Ouachita orogen north to the Ozark dome, and west to the Arbuckle uplift (Figures 4.1, 4.2 and Plate 4.1) (Arbenz, 1989b-d; Gatewood and Fay, 1991). Predominant strike direction of basement faults changes eastward from southwest-northeast in Oklahoma, to east-west in western Arkansas, to northwest-southeast, and to southwest-northeast near the Mississippi embayment (overlap of Mesozoic rocks) (Figure 4.2 and Plate 4.1) (Arbenz, 1989b-d).

Many smaller displacement basement faults of the Arkoma basin are pre-Pennsylvanian structures that clearly offset Cambrian-Ordovician formations but do not penetrate a basal Morrowan unconformity (Figure 4.4) (Arbenz, 1989a-d; Van Arsdale and Schweig, 1990; Gatewood and Fay, 1991). Other more prominent large-displacement down-to-south (or southeast) normal faults are Atokan-age (or latest Morrowan-age) structures marked by an abruptly thicker lower Atoka Formation on the southern side of these faults (Figure 4.4) (Arbenz, 1989a, b, d). The coincidence of abrupt increase in thickness with abrupt change from shallow-water to deep-water facies southward across these large-displacement normal faults (growth faults) suggests either late Morrowan initiation of fault motion (submarine fault scarp), or rapid early Atokan offset which exceeded sediment influx. Along the northern edge of
the Arkoma basin near the Ozark dome, and at the southwestern end near the Arbuckle uplift, basement faults offset Atokan, Desmoinesian, and Missourian strata (Figures 4.1 and 4.2) (Ham and others, 1954; Arbenz, 1989b).

**Along strike changes in structures of the Arkoma basin**

**Western Arkoma basin (north of Choctaw thrust fault)**

The Choctaw fault separates the western part of the Frontal Ouachita thrust belt from the western Arkoma basin (Figure 4.2 and Plate 4.1). The westernmost part of the Arkoma basin terminates as two northeast-plunging, fault-offset down-dropped synclines (Wapanucka and Wardsville) on the northeastern flank of the Arbuckle uplift (Plate 4.1). Northwest-southeast oriented cross sections constructed across the western part of the Arkoma basin and frontal Ouachita Mountains show a series of northeast-trending, large-offset basement normal faults beneath a lower Atokan detachment (Arbenz, 1989b-d). A cross section of the southwestern Ouachita Mountains, constructed across Black Knob Ridge, shows the geometry of one of these large-offset, pre-Atokan normal faults (Figure 4.5). This basement fault beneath the Ouachita allochthon, referred to as the “Bengalia fault” in Gatewood and Fay (1991), has a southeast dip, and is located in the subsurface beneath the Frontal Ouachita thrust belt (Figure 4.5). A triangle zone in the subsurface west of the Choctaw fault is positioned above the upthrown side of the Bengalia fault beneath Black Knob Ridge (Figure 4.5). The cross section of the western Ouachita Mountains, based upon well and seismic data, shows that the triangle zone roots in a basal décollément located in the lower Atoka Formation (Arbenz, 1989d; unpublished cross sections of J.K. Arbenz, 1996). East of the triangle zone, the basal décollément ramps along an erosional unconformity on the upthrown side of the Bengalia fault where Atoka Formation clastic rocks rest upon pre-Mississippian Arbuckle facies shallow-water carbonate rocks (Figure 4.5). A simplified restoration of the western part of the cross section illustrated in Figure 4.5 shows an abrupt thickening of Mississippian-Morrowan shallow-water facies clastic rocks (Caney-Goddard/Springer-Wapanucka) on the southeastern side of the Bengalia basement fault. Therefore, motion along this normal fault began at latest in the Mississippian.
Gatewood and Fay (1991), show the “Bengalia fault” as a near-concentric fault that is roughly parallel with the western part of the Ouachita thrust front. The locations of the Bengalia fault and other sub-allochthonous faults are constrained by sparse deep well data and proprietary seismic reflection profiles constructed across the Ouachita Mountains (Gatewood and Fay, 1991; R. O. Fay, personal communication, 2000). The arcuate-shaped Bengalia fault zone also coincides with a residual gravity anomaly high determined from proprietary modeling (Gatewood and Fay, 1991; R. O. Fay, personal communication, 2000). Regional structural cross sections and basement fault trends in the Arkoma basin suggest that the Bengalia fault comprises interconnected basement faults of differing trends (Lillie and others, 1983; Blythe and others, 1988; Arbenz, 1989a, d; Viele, 1989). From west to east, the strikes of segments of the Bengalia fault bend abruptly, and the fault segments are parallel with basement faults of the Arkoma basin (Gatewood and Fay, 1991).

Although Gatewood and Fay (1991) show the Bengalia fault as part of a structural high that is parallel and southeast of Ouachita thrust front (Choctaw-Dutch Creek-Ross Creek faults), the western segment of the Bengalia fault is parallel with, and similar in geometry to the large-offset, south-dipping Sans Bois normal fault in the subsurface beneath the Sans Bois syncline in the Arkoma basin (Figures 4.5 and 4.6). A cross section of part of the western Arkoma basin shows abrupt thickening of the Atoka Formation along the south-dipping Mulberry fault (Figure 4.5). On the northern upthrown block of the Sans Bois fault, 608 m to 1216 m of shallow-water middle and upper Atoka Formation rests unconformably upon Mississippian shelf-carbonate rocks (Arbenz, 1988a,d; Visher, 1996; J. K. Arbenz, unpublished cross sections, 1996). On the southern, down-dropped side of the fault, an additional 1824 m of deeper-water lower and middle Atoka Formation underlies shallow-water middle and upper Atoka Formation (Figure 4.5). The total thickness of deep-water facies Atoka thickens southward from 1824 m on the southeastern side of the Sans Bois fault, to more than 3647 m south of the Choctaw fault (Figure 4.5).

The erosional unconformity on the upthrown side of the Sans Bois fault is structurally similar to the erosional unconformity along the “Bengalia fault” beneath
the Choctaw fault west of Black Knob Ridge (Figures 4.5 and 4.6). The major difference is in timing of fault motion. Whereas abrupt thickening of Mississippian-Morrowan rocks on the southeastern downthrown block of the Bengalia fault at Black Knob Ridge shows that movement on the Bengalia fault began in the Mississippian (at latest), abrupt thickening of the lower and middle Atokan rocks on the southeastern downthrown block of the Sans Bois fault shows the fault movement started in Atokan (or latest Morrowan) (Figures 4.5 and 4.6). The thickening of Atoka Formation on the southeastern down-thrown block of the Sans Bois fault suggests an Atokan-age growth fault (Arbenz, 1989d); however, the abrupt appearance of deep-water facies Atoka Formation indicates that the Atoka beds were deposited in deep water along a pre-existing fault scarp.

Central and eastern Arkoma basin (north of Dutch Creek–Ross Creek–Cadron thrust faults)

The east-northeast-striking Dutch Creek-Ross Creek-Cadron thrust faults separate the central and eastern parts of the Arkoma basin on the north from the eastern Frontal Ouachita thrust belt on the south (Figure 4.2 and Plate 4.1). North-south oriented cross sections illustrate predominant down-to-southeast (and local down-to-northwest) offset of the northeast-trending basement faults of the eastern Arkoma basin (Viele, 1989; Van Arsdale and Schweig, 1990; J. K. Arbenz, unpublished cross sections, 1996). A cross section of the easternmost Arkoma basin shows that the early Mississippian Boone Formation is offset by three south-dipping normal faults (Figure 4.4) (Van Arsdale and Schweig, 1990). Individual fault offsets are approximately 1000 m down-to-south (Figure 4.4). In contrast to the Mulberry and Sans Bois faults to the west (Figure 4.4), no abrupt thickening of Atoka Formation is visible on the down-dropped sides of the normal faults in the eastern Arkoma basin. Instead, the Atoka Formation and underlying Morrowan clastic facies thicken progressively towards the Ouachita thrust front (Van Arsdale and Schweig, 1990). The Van Arsdale and Schweig (1990) cross section of the eastern Arkoma basin shows a décollement surface that coincides with a regional unconformity at the base of the Morrowan (Figure 4.4). This décollement separates the thin-skinned structures of the
eastern Arkoma basin from underlying thick-skinned basement faults. In some locations along the up-thrown sides of the normal faults, the base of the Morrowan succession rests on pre-Mississippian strata (possibly Chattanooga Shale) (Van Arsdale and Schweig, 1990). Progressive southward truncation of Mississippian strata within down-thrown fault blocks indicates late Mississippian-early Morrowan down-to-south listric fault motion followed by erosion prior to deposition of Morrowan strata across the southeastern Arkoma basin.

The lack of abrupt thickening of the Atoka Formation and underlying Morrowan clastic rocks south of the normal faults in the eastern Arkoma basin suggests that the basement flexed southwards under the weight of clastic sediment which began to be deposited in the Morrowan, and became rapidly deposited in the Atokan. However, cross sections of the eastern Frontal Ouachita thrust belt show abrupt thickening of deep-water facies lower (and middle) Atoka Formation southeast of the Ross Creek-Cadron thrust faults (cross section C-C’ in Plate 4.2, and cross sections ARK 1 to ARK 4 in Plate 4.3) (Viele, 1989; J.K Arbenz, unpublished cross sections, 1996). Restoration of these cross sections suggests that large-offset, down-to south, normal faults are located in the subsurface beneath the allochthonous strata of the eastern Frontal Ouachita thrust belt.

In summary, northwest-southeast and north-south oriented cross sections of the western and eastern parts of the Arkoma basin and autochthon beneath the Frontal Ouachita thrust belt suggest variation in timing of basement fault motion throughout the region. A set of large-offset down-to-south (and southeast) basement growth faults in the Arkoma basin outline the northern boundary of an Atokan (possibly latest Morrowan) small-scale embayment marked by an abrupt southward facies transition in the lower Atoka Formation (Figure 4.6). The abrupt change from shallow-water to deep-water facies in lower Atoka Formation beds on the down-thrown sides of the basement faults suggests deposition along pre-existing submarine fault scarps.

The western edge of the Atokan (latest Morrowan) small-scale embayment extends northeastward from the subsurface west of Black Knob Ridge towards the central Arkoma basin (Figure 4.6). The Mulberry fault marks the northwestern limit
of the shallow, northern part of the small-scale embayment (Figures 4.4 and 4.6). Lower Atokan rocks are thin or absent north of the Mulberry fault (Figure 4.6). The northern margin of the embayment stretches eastward from the Mulberry fault along three nearly continuous basement faults (Figure 4.6). The eastern end of the embayment bends abruptly to a northeastward trend in the subsurface near the Mississippi embayment (Figure 4.6). South of the Mulberry fault, the Sans Bois fault marks the northwestern limit of thick lower Atoka Formation.

Sparse well, seismic, and gravity anomaly data indicate that several large-offset basement faults beneath the Frontal Ouachita thrust belt (Arbenz, 1989a; Gatewood and Fay, 1991; J. K. Arbenz, unpublished cross sections, 1996; proprietary seismic data). One of these segmented fault zones is referred to as the “Bengalia” fault (Figures 4.5 and 4.6). East of Black Knob Ridge, the Bengalia fault is located in the subsurface beneath the allochthonous strata (Figures 4.5 and Figure 4.6). Although the Bengalia fault is southeast of the Sans Bois fault, cross sections of the Black Knob Ridge part of the Ouachita Mountains show abrupt thickening of deep-water-facies lower Atoka Formation east of the Bengalia fault (Figures 4.5 and 4.6). However, in contrast to the Sans Bois fault, abrupt thickening of Mississippian-Morrowan strata is apparent east of the Bengalia fault (Figures 4.5 and 4.6). This indicates that displacement of the Bengalia fault began in the Mississippian or Morrowan (at the latest).

**Structural subdivisions of the Ouachita Mountains**

**Frontal thrust belt (overview)**

Although the detailed geometry of the Ouachita Mountains is complex, the region can be subdivided into several structural zones (Figure 4.2 and Plate 4.1). The Frontal Ouachita thrust belt, a zone of variable map width dominated by surface exposures of Mississippian to lower Pennsylvanian (Atokan) rocks separates the interior thrust belts of the Ouachita Mountains to the south from the Arkoma basin to the north (Figure 4.2 and Plate 4.1). The Choctaw and Dutch Creek-Ross Creek thrust faults mark the Ouachita thrust front. However, folds, blind thrust faults, and locally exposed small-displacement thrust faults which parallel the Ouachita thrust front are
found north of the Choctaw-Dutch Creek-Ross Creek faults in the Arkoma basin (Arbenz, 1989a, b, d; Mazengarb, 1995; Valderrama and others, 1996).

The frontal faults, Choctaw fault and Dutch Creek-Ross Creek faults, are not a single continuous fault rimming the Ouachita thrust belt. The Choctaw fault loses displacement and ends eastward along the spine of the Honest anticline (Figure 4.2 and Plate 4.1). The Dutch Creek fault passes south of the Choctaw fault and extends into the Oklahoma part of the Ouachita frontal belt, where it either ends westward or merges with another thrust fault (Figure 4.2 and Plate 4.1). The western terminus of the Dutch Creek fault is not clearly defined on generalized regional structural geologic maps (Arbenz, 1989b; Viele and Thomas, 1989). The Dutch Creek fault appears to terminate north of the Black Fork syncline; however, it may continue westward as a fault splay in the western Frontal Ouachita thrust belt (Figure 4.2 and Plate 4.1).

The southern boundary of the Frontal Ouachita thrust belt is difficult to locate precisely. Traditionally, the southern boundary of the frontal belt has been placed along the Windingstair fault in Oklahoma, coinciding with the northernmost exposure of Mississippian Stanley Group, and north of the Benton uplift in Arkansas at the northernmost exposure of Devonian-Mississippian Bigfork Chert-Arkansas Novaculite (Plate 4.1). The Windingstair fault is an intuitively obvious boundary from the northern end of Black Knob Ridge to the Potato Hills area of Oklahoma, where Stanley Group shales are in fault contact with Jackfork-Atoka sandstone formations. Southwestward along Black Knob Ridge, the Windingstair fault loses displacement, and the Stanley Group forms both hanging wall and footwall (Plate 4.1).

The structural relationship between the Windingstair fault east of the Potato Hills and the southern boundary of the Frontal Ouachita thrust belt in Arkansas is very complex. The Windingstair fault is a north-vergent thrust fault in Oklahoma; however, folds along the same structural trend farther east in Arkansas are south-vergent. Many small-displacement faults with older rocks on the southern side parallel the south-vergent folds (Haley and others, 1993). Although many of these faults are interpreted to be south-overturned thrust faults, dips of faults in many areas south of the Frontal Ouachita thrust belt are not discernable in the field (Blythe and others,
1988), and without fault-dip information, several alternative fault geometries can fit the field outcrop geometry (Blythe and others, 1988). If older rocks are to the north, and fault dip is to the north, then the fault can either be a backthrust or a north-vergent hangingwall folded over an arched footwall (Figure 4.7). If older rocks are on the south, and if the dips of these faults are to the north, then they can be described as thrust faults overturned towards the south (Figure 4.7) (Zimmerman and others, 1982; Viele, 1989; J. K. Arbenz, unpublished cross sections, 1996). However, axial-planar south-vergent reverse faults (such as Fault A in Figure 4.7) which have been reactivated as normal faults would also have older rocks on the south side of the fault if late-stage normal offset exceeded early-stage reverse offset.

The apparently complex relationship between the eastern end of the Windingstair fault and the western end of the south-vergent southern boundary of the eastern Frontal Ouachita thrust belt is somewhat resolved if the part of the Windingstair fault that is north of the Potato Hills is interpreted as a fault splay which intersects other fault splays northwest of the Potato Hills (Plate 4.1). It is suggested here that, northwest of the Potato Hills, the Windingstair fault splits and the northernmost fault belonging to this set of fault splays is recognized east of the Potato Hills and south of the Black Fork syncline as the Briery fault (Plate 4.1). The Briery fault, which marks the northernmost exposure of Stanley Group in the Oklahoma-Arkansas border area, is more structurally analogous to the part of the Windingstair fault west of the Potato Hills, than the eastern end of Windingstair fault is similar to the western end.

An eastern terminus of the Briery fault is not clearly defined on regional structural geologic maps (Plate 4.1) (Arbenz, 1989b; Viele and Thomas, 1989; Haley and others, 1993). The Briery fault may continue eastward as a fault in the western end of the eastern Frontal Ouachita thrust belt (Plate 4.1). However, the Briery fault is a clearly defined, north-vergent structure in contrast to the dominantly south-vergent fabric of most of the eastern Frontal Ouachita thrust belt. Like the Choctaw fault, the Briery fault may lose displacement and end eastwards. According to regional geologic maps, the Honess fault located between the Briery and Windingstair faults also
appears to lose displacement eastward (Plate 4.1) (Arbenz, 1988b; Viele and Thomas, 1989; Haley and others, 1993). The Briery, Honess, and Windingstair faults also appear to structurally overlap with, but are not continuous with, the south-vergent structures farther east along strike in the eastern part of the Frontal Ouachita thrust belt (Plate 4.1) (Blythe and others, 1988; Arbenz, 1989b; Haley and others, 1993; Viele and Thomas, 1989; J. K. Arbenz, unpublished cross sections, 1996).

The easternmost exposed part of the southern boundary of the Frontal Ouachita thrust belt has the most complicated geometry. Between the Paron and Jessieville nappes of the eastern Ouachita Mountains, the Mississippian Stanley Group is folded complexly around Ordovician to Mississippian pre-orogenic strata (Mazarn Shale through Arkansas Novaculite) (Figure 4.2 and Plate 4.1). At the eastern end of the southern boundary of the exposed Frontal Ouachita thrust belt, along most of the northern margin of the Paron nappe, tightly folded, south-vergent Morrowan-age Jackfork Sandstone strata are in fault contact with tightly folded, south-vergent Ordovician to Mississippian strata (Bigfork Chert through Stanley Group) (Figure 4.2 and Plate 4.1).

Central and Southern Ouachita thrust belts (overview)

The Central Ouachita thrust belt is bounded on the north and west by the Windingstair fault zone (in the west) and roughly by the northernmost exposure of pre-Mississippian deep-water facies strata along the northern flank of the Benton uplift, Jessieville nappe, and Paron nappe (in the east) (Figure 4.2 and Plate 4.1). The western part of the southern boundary is roughly located along the southern limb of the Bethel syncline on the northwestern flank of the Broken Bow uplift. At the northern end of the Broken Bow uplift, the southern boundary of the Central Ouachita thrust belt continues along an east trend along the southern part of the Cross Mountains and extends farther east toward the Trap Mountains (Figure 4.2 and Plate 4.1).

From west to east along strike, the predominant strike of structures in the Central Ouachita thrust belt varies from northeast-striking, to east-west-striking, to
northwest-striking (Figure 4.2 and Plate 4.1). The Central Ouachita thrust belt is widest in the far western part along the margin of the Mesozoic Gulf Coastal Plain (Figure 4.2 and Plate 4.1). North of the Broken Bow uplift, the width of the Central Ouachita thrust belt decreases abruptly, but is relatively constant farther east (Figure 4.2 and Plate 4.1). In general, broad folds of upper Mississippian-Atokan turbidites comprise the western part of the Central Ouachita thrust belt; whereas, tightly folded pre-Mississippian cherts, sandstones, and low-grade phyllites (Benton uplift) comprise the eastern part of the belt.

The Southern Ouachita thrust belt covers the area of the Ouachita Mountains that contains the Broken Bow uplift and the area farther east and to the south of the Cross Mountains-Cossatot Mountains-Trap Mountains trend (Figure 4.2 and Plate 4.1). In contrast to the Central belt, tightly folded pre-Mississippian strata comprise the western part (Broken Bow uplift) of the Southern Ouachita thrust belt; whereas, long wavelength folds of upper Mississippian-Atokan formations comprise the eastern part. The eastern part of the Southern Ouachita thrust belt consists of predominantly south-dipping Jackfork and Atoka Formations which collectively are recognized as the Athens Plateau (Figure 4.2 and Plate 4.1). The Paleozoic strata of the Athens Plateau are unconformably covered on the south by the Mesozoic Gulf Coastal Plain (Figure 4.2).

**Along strike changes in structures of the Ouachita Mountains**

**Western Frontal Ouachita thrust belt (overview)**

The western part of the Frontal Ouachita thrust belt, north of Black Knob Ridge, is dissected by numerous thrust faults which encompass numerous thin fault blocks (Figure 4.2 and Plate 4.1). For this reason, the western part of the Frontal Ouachita thrust belt can be referred to as an imbricated fault zone. Although there are numerous faults throughout this zone, two larger displacement through-going faults, the Pine Mountain and Ti Valley faults, are located southeast of the Choctaw fault (Plate 4.1).

In the westernmost part of the Frontal Ouachita thrust belt, along Black Knob Ridge (Figure 4.2 and Plate 4.1), the Ti Valley fault is a northwest-vergent thrust fault
which places Ordovician to Mississippian deep-water shales and cherts (Womble through Stanley Group) in contact with shallow-water lower Pennsylvanian sandstones and shales (Atoka Formation) on the west. Surface maps and cross sections through Black Knob Ridge show little imbrication of the rocks west of the Ti Valley fault (Figure 4.2, and Plates 4.1, 4.2).

North of Black Knob Ridge, the Ti Valley fault splits in map view (Plate 4.1) (Hendricks and others, 1937; Hardie, 1988, 1990). The western fault splay deflects westward and intersects the Choctaw fault. The eastern fault branch continues northward roughly parallel with the Choctaw fault. Approximately 15 km north of the northern end of Black Knob Ridge, the Ti Valley fault intersects the Pine Mountain fault (Plate 4.1). The Pine Mountain fault bends westward and roughly parallels the Choctaw fault. The Pine Mountain fault intersects the Choctaw fault northwest of the Black Fork syncline (Plate 4.1).

North of the intersection with the Pine Mountain fault, the Ti Valley fault continues north and curves eastward (Plate 4.1). The Ti Valley fault, sensu stricto, ends north of the Simmons Mountain syncline in the eastern part of the western Frontal Ouachita thrust belt (Plate 4.1). According to regional geologic maps, it is not clear if the structural equivalent of the Ti Valley extends east of the Simmons Mountain syncline (Plate 4.1) (Arbenz, 1988b; Viele and Thomas, 1989). The Ti Valley fault may continue eastward to become the Briery fault, or it may continue as a fault within, or along the northern boundary of, the Black Fork syncline (Plate 4.1).

The variation in geometry and lithologic composition of strata within the imbricated fault zone of the western Frontal Ouachita thrust belt (Plate 4.1) illustrate along-strike changes in thickness and stratigraphic level of detachments. The hanging wall of the Choctaw fault carries thrust-fault imbricated repeats of the Morrowan-age shallow-water Springer Group (and locally Mississippian-age Caney-Goddard) shales and overlying Morrowan (and lower Atokan?) Wapanucka Limestone, and Atokan-age Atoka Formation. Detachments are within the shales at bases of stiff layers within the succession (Caney, Goddard, Springer) (Hardie, 1988; Arbenz, 1989a, J. K. Arbenz, unpublished cross sections, 1996). One of the detachments is within the
Spinger Group below the base of the Wapanucka-Atoka succession (Arbenz, 1989d; Valderrama and others, 1996).

**Western Frontal Ouachita thrust belt (between Choctaw and Pine Mountain faults)**

Fault duplex splays and imbricate fans dissect the strata between the Choctaw fault and Pine Mountain fault (Arbenz, 1989a, b, d). These fault splays duplicate mostly Morrowan-Atokan strata (Springer Group-Wapanucka-Atoka Formation) (Hardie, 1988; Arbenz, 1989a, b), and in several locations, repeated Springer-Wapanucka beds fan out along several fault splays (Plate 4.1). Many of the faults lose displacement eastward along-strike in the lower Atoka Formation, and fault terminations may coincide with a detachment surface in the lower Atoka Formation. Cross sections of the western Arkoma basin suggest that a detachment surface is located in the subsurface within the lower Atoka Formation north and northwest of the Choctaw fault (Mazengarb, 1995; Valderrama and others, 1996) (Figure 4.8). The northern ends of the imbricated fans of Springer-Wapanucka beds located between the Choctaw and Pine Mountain faults may mark the position where the lower Atoka Formation delamination surface intersects present ground level. The Choctaw fault is a larger displacement, thrust fault which cross cuts (and therefore postdates) the lower Atoka Formation delamination surface (Figure 4.8).

In the area near the Carbon fault, the Choctaw fault curves from a northeast strike to a nearly east-west strike (Plate 4.1). Most of the fault splays between the Choctaw and Pine Mountain faults south of the bend in the Choctaw fault lose displacement along strike toward the north or east at the end of the imbricate fans (Plate 4.1). On the basis of map patterns (Plate 4.1), it is unclear whether the Pine Mountain fault truncates fault slices located between the Choctaw and Pine Mountain faults, or coincides with the top of a duplex. In some locations, splay faults west of the Pine Mountain fault lose displacement eastwards and do not intersect the Pine Mountain fault (Plate 4.1). At least one of these splay faults intersects the Pine Mountain fault at a large angle (Plate 4.1). The large intersection angle suggests that the Pine Mountain fault truncates (beheads) the underlying fault slices. However, a down-plunge view
cross section constructed of the western end of the Frontal Ouachita thrust belt suggests that the southern end of the Pine Mountain fault coincides with top of a fault duplex (Arbenz, 1989b). It is possible that the Pine Mountain fault cuts up-section toward the north (above the lower Atoka detachment surface).

Along the east-west striking part of the Choctaw fault, the amplitude and number of Springer-Wapanucka-lower Atoka thrust-fault repetitions decreases. At the eastern ends, each of these faults intersects the Choctaw fault within the lower Atoka Formation. The western ends of the faults appear to lose displacement within (and likely root in) the Springer Group (locally Caney-Goddard) (Plate 4.1). It is unclear from map view whether the Springer-Wapanucka-lower Atoka duplex splays continue farther east in the subsurface or whether the Springer detachment level loses displacement farther east. Farther east in Arkansas in the subsurface, the Wapanucka Limestone thins and grades into a Morrowan shale (Bloyd or Johns Valley) (Gatewood and Fay, 1991; Shelton, 1996; C.G. Stone, personal communication, 2000).

On the basis of field studies, Mazengarb (1995) proposed that the east-west striking part of the Choctaw fault (north of the Potato Hills) is at the apex of a triangle zone. His interpreted cross section suggests that the part of imbricated zone of the western Frontal Ouachita thrust belt north of the Pine Mountain fault represents the eroded core of a duplex that once was beneath a passively uplifted Atoka Formation. This model suggests that a regional detachment level is located in the lower Atoka Formation. The many repetitions of Springer-Wapanucka-lower Atoka strata visible on the surface and interpreted in the subsurface along the northwestern side of the Choctaw fault (Valderrama and others, 1996) suggest the validity of a lower-Atoka detachment-level passive-uplift model for the front edge of the western Frontal Ouachita thrust belt (Figure 4.8). The original duplexed strata beneath the passive-uplift backthrust has been offset and intersected by a set of younger larger displacement thrust-faults (Choctaw, Pine Mountain, and perhaps Ti Valley north of Black Knob Ridge).
The similarity in along-strike stratigraphy within imbricated thrust slices bounded by the Choctaw and Pine Mountain faults indicates that all the included thrust faults root at the same stratigraphic detachment level. For purposes of discussion of detachment level, the part of the Ti Valley fault that extends between the northern end of Black Knob Ridge and the split with the Pine Mountain fault may be considered a continuation of the Pine Mountain fault (Plate 4.1). All of the thrust faults between the Choctaw and Pine Mountain fault root beneath the Morrowan Wapanucka Limestone in either the Mississippian-Morrowan Springer Group (including Goddard) or Mississippian Caney Formation. The Caney and lower part of the Springer Group are predominantly shale and these shale layers likely provide detachment horizons and glide planes (Figure 4.9).

The lower Pennsylvanian basal Spiro Sandstone bed and the rest of the overlying Atoka Formation cover the Morrowan Wapanucka Formation throughout the western Frontal Ouachita thrust belt between the Choctaw and Pine Mountain faults. However, structural cross sections constructed across the western Frontal Ouachita thrust belt show a progressive increase in deep-water facies Atoka Formation northward along strike (Arbenz, 1988a, d; J. K. Arbenz, unpublished cross sections, 1996). At Black Knob Ridge, the exposed Atoka Formation is entirely shallow-water facies. Farther north along the Choctaw fault, southeast of the nearby Carbon fault, deep-water facies Atoka Formation rests upon the Wapanucka Limestone. Farther east along-strike, the thickness of the deep-water facies Atoka Formation increases abruptly eastward.

Western Frontal Ouachita thrust belt (between Pine Mountain and Ti Valley faults)

The ages of strata imbricated between the Pine Mountain and Ti Valley faults range from Devonian to Early Pennsylvanian (Atokan). Strata between the Pine Mountain and Ti Valley faults are predominantly clastic facies. Interpreted facies vary from shallow-water marine to marginal- or deep-water marine (Arbenz, 1989c, d). The predominant thrust-repeated strata are middle Mississippian (Meramecian) to lower Pennsylvanian (Atokan). The overlying lower Pennsylvanian (Atokan) Atoka
Formation is entirely deep-water marine turbidites where exposed between the Pine Mountain and Ti Valley faults (Plate 4.1).

Along the Pine Mountain fault, isolated outcrops of Devonian to Mississippian cherts and shales are found in several locations (Plate 4.1). These isolated blocks may be detached olistolithic blocks incorporated in deep-water facies Atoka Formation (Gatewood and Fay, 1991), or small tectonically emplaced fault blocks. The formations included in these olistolith/fault blocks are Devonian Pinetop Chert (an intermediate facies between the upper part of the Hunton Group cherty limestone and Arkansas Novaculite), Devonian to Mississippian Woodford Formation (shale and chert), and transitional-depth Caney Formation-Springer Group shales and sandstones (Figure 4.9 and Plate 4.1). In contrast to strata northwest of the Pine Mountain fault, the Wapanucka Limestone is absent from the olistolith/fault blocks and is not exposed between the Pine Mountain and Ti Valley faults (Figure 4.9 and Plates 4.1 and 4.2). Here, lower Atoka Formation is unconformable above Springer Group (Figure 4.9 and Plates 4.1 and 4.2).

If the isolated blocks of Pine Top-Woodford-Caney-Springer rocks are olistoliths, the source area must be near the restored position of the Pine Mountain fault. A possible source area could be a fault scarp to the northwest of the restored location of the Pine Mountain fault. The presence of a cherty facies also suggests that the “olistoliths” are from a facies transition zone. Another possible explanation for the isolated blocks of Devonian to Morrowan strata is that the Pine Mountain fault restores palinspastically southeast of an irregular late Morrowan/early Atokan unconformity surface. The isolated blocks may have been scooped-up where the relief of the unconformity surface was relatively high and transported northwest along a detachment fault. Late Morrowan unconformity surfaces are recognized throughout the foreland area of the Ouachita thrust belt northeast of the Muenster uplift (Johnson and others, 1988; Van Arsdale and Schweig, 1990; Cooper, 1995; Shelton, 1996).
Western Frontal Ouachita thrust belt (between Ti Valley and Windingstair faults)

Exposed strata between the Ti Valley and Windingstair faults in the imbricated fault zone of the western Frontal Ouachita thrust belt are Pennsylvanian (Morrowan-Atokan) deep-water clastic facies (Jackfork Sandstone-Johns Valley Shale-Atoka Formation). Regional cross sections constructed across this zone show Mississippian Stanley Group rocks in the subsurface beneath the Jackfork Sandstone (Arbenz, 1989a,d; J. K. Arbenz, unpublished cross sections, 1996). The thrust faults between the Ti Valley and Windingstair faults root in a detachment glide plane in the Stanley Group shales.

Cross sections through the western part of the Frontal Ouachita thrust belt show several southeast-dipping thrust-fault slices (Plate 4.2). Relatively thin Morrowan Jackfork Sandstone-Johns Valley Shale are beneath much thicker deep-water facies of the Atoka Formation (Plate 4.2). Relatively thin Stanley Group rests beneath the Jackfork Sandstone in the deeper, southeastern root zones of the thrust sheets (Plate 4.2).

Northward along strike, a transition in structural style is recognizable between the Ti Valley and Windingstair faults. West of the Potato Hills, the Ti Valley and Windingstair faults bound a tight anticline-syncline pair that is overturned toward the northwest (Plates 4.1 and 4.2). The tight anticline-syncline pair is cross cut by several steep, north-vergent thrust faults (Plates 4.1 and 4.2). Many of these faults appear to parallel the axial planes of the fold pair (Plates 4.1 and 4.2). Several of these axial planar faults intersect with the Windingstair fault west of the Potato Hills (Plates 4.1 and 4.2).

The northwest-vergent crest of a tight anticline of Jackfork Sandstone-Johns Valley Shale is separated from relatively horizontal Stanley Group on the southeast by the Windingstair fault (Plates 4.1 and 4.2). North of the Potato Hills, the Windingstair fault places relatively horizontal Stanley Group (and in some locations Jackfork Sandstone) in fault contact with south-dipping Atoka Formation (Plates 4.1 and 4.2).
This Atoka Formation is located on the southern limb of a tight, north-verging anticline that extends westward to an area west of the Potato Hills in the footwall of the Windingstair fault (Plate 4.2). Northeast of the Potato Hills, between the Ti Valley fault and Windingstair fault (as shown in Plates 4.1 and 4.2), a pair of tight, fault dissected, north-vergent folds, changes abruptly to the broad, upright Simmons Mountain syncline.

**Transition zone between western and eastern parts of the Frontal Ouachita thrust belt (overview)**

The area containing the Simmons Mountain, Rich Mountain, and Black Fork synclines, and the area west of the Fourché/Perryville syncline, marks a transition between structural styles of the western and eastern parts of the Frontal Ouachita thrust belt (Plates 4.1 and 4.2). Along strike to the west of the Simmons Mountain syncline, Mississippian-Morrowan strata are thin, locally deformed into northwest-vergent tight folds, and imbricated by numerous faults (Plates 4.1 and 4.2). Along strike to the east of the Black Fork and Rich Mountain synclines, Mississippian-Morrowan strata are much thicker and deformed into broad folds that vary from north-vergent to upright folds on the north to predominantly south-vergent tighter folds farther south (Plates 4.1 and 4.2). The along-strike changes in fold style and fault spacing within the Frontal Ouachita thrust belt correlates with an along-strike increase in thickness of the Mississippian-Morrowan succession (Caney-Wapanucka/Stanley-Johns Valley).

In addition to an eastward increase in thickness of the Mississippian-Morrowan succession, map patterns of the transitional zone of the Frontal Ouachita thrust belt show several thrust-fault displacement transfers both north and south of the Black Fork syncline (Figure 4.10). North of the Black Fork syncline, the north-vergent Choctaw and Dutch Creek faults overlap along strike in map view (Figure 4.10). Along strike to the east, displacement along the Choctaw fault decreases, whereas displacement along the Dutch Creek-Ross Creek fault zone increases (Figure 4.10). South of the Black Fork syncline, the Briery, Honess, and Windingstair faults overlap with north- and south-vergent structures extending from the east (Figure 4.10).
**North of Ti Valley-Briery faults**

The surface geology north of the Ti Valley-Briery faults in the transition zone between the western and eastern parts of the Frontal Ouachita thrust belt consists entirely of deep-water Pennsylvanian (Atokan) Atoka Formation (Figure 4.2 and Plate 4.1). The northernmost exposure of pre-Atoka rocks in the transition zone is south of the Ti Valley-Briery faults. The Briery fault marks the northernmost exposure of Mississippian Stanley Group in the transition zone of the Frontal Ouachita thrust belt. Therefore, the Briery fault occupies the same structural position as the Windingstair fault southwest of the Potato Hills.

A cross section constructed across the transition zone along the Oklahoma-Arkansas border, based on the COCORP seismic reflection profile of Lillie and others (1983), shows that the thrust faults between the Choctaw and Ti Valley-Briery faults root in a detachment above the Cambrian-Mississippian shallow-water carbonate succession (Plate 4.2). The detachment may coincide with a regional unconformity surface at the base of the Morrowan which extends across the eastern Arkoma basin (Van Arsdale and Schweig, 1990). The imbricates of Springer Group-lower Atoka Formation shown in geologic maps and illustrated in cross sections constructed across the western Frontal Ouachita thrust belt are missing from cross sections across the transitional zone of the Frontal Ouachita thrust belt along the Oklahoma-Arkansas border (Plates 4.1 and 4.2). The COCORP seismic reflection profile (Lillie and others, 1983) clearly shows the entire Morrowan-Atokan-Desmoinesian succession north of the Ti Valley-Briery fault is displaced northward (Plate 4.4). However, careful examination of the COCORP profile shows a detachment surface above the base of the lower Atoka Formation (Plate 4.4). This detachment surface is likely the eastern extension of the lower Atoka detachment illustrated in cross sections of the western Arkoma basin (Figure 4.8) (Arbenz, 1989d; Mazengarb, 1995; Valderrama and others, 1996).
**Abrupt increase in thickness of Atoka Formation**

The Atoka Formation thickens abruptly eastward in the transitional zone of the frontal belt south of the Dutch Creek-Ross Creek faults (Plates 4.1 and 4.2). The exposed Atoka Formation south of the Dutch Creek-Ross Creek faults is entirely deep-water turbidite facies and contains a variety of submarine sedimentary slump structures (Visher, 1996; and C.G. Stone, personal communication, 2000). The combination of sedimentary slump structures, great thickness of Atoka Formation (estimated maximum thickness of more than 8662 m in the eastern part of the transitional zone of the Frontal Ouachita thrust belt); (Branon, 1961; Bush and others, 1977; Lillie and others, 1983; Fay and others, 1986; J.K. Arbenz, unpublished cross sections, 1996), and cross section restoration suggests that a large-offset, down-to-south normal fault may be buried beneath the southern part of the eastern Frontal Ouachita thrust belt.

**Age of the lower Atoka Formation?**

The age of the lower part of the Atoka Formation, especially in the subsurface, is uncertain. The base of the Atoka Formation is commonly an unconformity. In the western Frontal Ouachita thrust belt, the Morrowan Wapanucka Formation limestone and sandstone beds are easily recognizable marker beds. Farther southwest in the Ardmore basin (Figure 4.1 and Plate 4.1), the limestone facies of the Wapanucka Formation is missing, and the Morrowan-age clastic facies part of the Dornick Hills Group rests unconformably on shales of the Springer Group (Cooper, 1995). Atokan-age clastic beds of the Dornick Hills Group are conformable above the Morrowan beds of the Dornick Hills (Cooper, 1995). The Dornick Hills Group beds are unconformably beneath a thick series of Desmoinesian and Missourian clastic rocks in the Ardmore basin (Cooper, 1995). South of the Ardmore basin, along the crest of the Muenster arch (Figure 4.1 and Plate 4.1), a succession of Desmoinesian through Permian clastic rocks rests unconformably on Cambrian-Ordovician carbonate rocks and, in some locations, directly on Precambrian basement (Ewing, 1991).
Southeast of the Pine Mountain fault, and along-strike towards the eastern part of the Frontal Ouachita thrust belt, the limestone facies of the Morrowan Wapanucka Formation thins, is progressively chert-enriched, and is classified as Chickachock Chert (Cline and Shelburne, 1959). Within the eastern Frontal Ouachita thrust belt, the Chickachock spiculitic chert has been found above the base of a thick succession of Atoka-type sandstones. It is unclear whether the basal Atoka-type sandstone below the Chickachock Chert represents a sandy, turbiditic facies of the Wapanucka Formation, or represents the base of the Atoka Formation. To complicate the correlation, it is difficult to locally distinguish between upper Jackfork, Johns Valley and lower Atoka formations in parts of the eastern Frontal Ouachita thrust belt where these strata are greatly deformed (C.G. Stone, personal communication, 2000). Also, deeply buried beds of the lower Atoka Formation beneath the eastern Frontal Ouachita thrust belt have not been penetrated by wells.

Regardless of whether the Atoka-type sandstones beneath the Chickachock Chert are included in the Wapanucka Formation or assigned to the base of the Atoka Formation, these basal turbidites mark the onset of a phase of rapid clastic deposition along the eastern Frontal Ouachita thrust belt in Arkansas that began in the late Morrowan. Deposition continued to be rapid through the Atokan, before slowing in the Desmoinesian. In contrast, farther west in the Ardmore basin of Oklahoma, the Morrowan/Atokan Dornick Hills Group is relatively thin (Figure 4.1). Here, a much thicker succession of Desmoinesian through Missourian shallow-water clastic rocks unconformably overlies the Dornick Hills Group (Ham, 1973; Cooper, 1995).

The observed southwestward age progression in the onset of the phase of rapid clastic deposition in the Pennsylvanian succession of the Ouachita Mountains, Arkoma basin, and Ardmore basin is consistent with the overall southwestward decrease in age of the onset of rapid deposition along the foreland of the Ouachita orogen. On the east, in the Black Warrior basin (Figure 4.1), abrupt thickening of the Late Mississippian-early Morrowan Parkwood and Pottsville Formations signals the onset of rapid clastic facies deposition. Along the eastern Frontal Ouachita thrust belt, the phase of rapid deposition is represented by slight thickening of Morrowan strata and a
much more noticeable increase in thickness of the Atoka Formation. Along the western Frontal Ouachita thrust belt the Atokan section thins, and the Desmoinesian-Missourian section thickens. Farther southwest along the Ouachita thrust front in the Marathon Mountains (Figure 4.1), the phase of rapid clastic deposition is represented by the upper Desmoinesian (Pennsylvanian) through Wolfcampian (Permian) Gaptank Formation (Ethington and others, 1989; Muehlberger and Tauvers, 1989).

**South of Ti Valley-Briery faults**

The Ti Valley-Briery faults mark the northernmost exposure of pre-Atoka strata in the transitional zone of the Frontal Ouachita thrust belt (Figure 4.2 and Plate 4.1). South of the Black Fork syncline, the Briery fault is a north-vergent thrust fault separating Atoka Formation on the north from a narrow syncline containing Mississippian Stanley Group through Morrowan Jackfork Sandstone and Johns Valley Shale in the hanging wall (Plate 4.1). The eastern end of the Briery fault is not clearly defined but appears to intersect (or overlap) structures of the eastern Frontal Ouachita thrust belt (Plate 4.1).

The western end of the Briery fault appears to splay into several faults (Plate 4.1). The northernmost splay continues westward as the Ti Valley fault (or as a fault that connects with the Ti Valley fault). In contrast to the Briery fault farther east, a much smaller apparent stratigraphic offset is observed along the Ti Valley fault in the western part of the transitional Frontal Ouachita thrust belt. Except for the area north of the western end of the Rich Mountain syncline, where a narrow zone of Jackfork/Johns Valley is thrust northward over the Atoka Formation, the surface trace of the fault that connects the Ti Valley fault with the Briery fault remains within the Atoka Formation (Plate 4.1). However, the Atoka Formation is thick in this location, and regional geologic maps lack the precision necessary to accurately compare along-strike changes in offset of the Ti Valley-Briery fault zone (Plate 4.1). Apparent stratigraphic offset of the Ti Valley fault increases at the western end of the transitional zone of the Frontal Ouachita thrust belt where the Ti Valley fault extends into the northeastern end of the imbricated fault zone of the western Frontal Ouachita thrust belt (Plate 4.1). Here, the north-vergent Ti Valley fault separates relatively thin
Morrowan Jackfork Sandstone-Johns Valley Shale on the south from the Atoka Formation (Plate 4.1).

Map patterns suggest that fault displacement increases south of the Ti Valley fault in the western part of the transitional zone of the Frontal Ouachita thrust belt (Plate 4.1). In contrast to the northern fault splay which connects the Briery fault with the Ti Valley fault, two other fault splays connect farther east and extend as a single fault along the northern margin of the Simmons Mountain syncline (Plate 4.1). This southern fault splay continues west of the Simmons Mountain syncline and extends into the imbricate fault zone of the western Frontal Ouachita thrust belt (Plate 4.1). The Simmons Mountain syncline is a broad, east-southeast-elongate basin (Plate 4.1). The fault along the northern margin of the Simmons Mountain syncline is north-vergent thrust fault that separates Morrowan Jackfork Sandstone (on the south) from Atoka Formation (on the north). Atoka Formation surrounded by a rim of thin Morrowan Johns Valley Shale forms the center of the Simmons Mountain syncline.

A narrow, southeast-trending fold pair cuts obliquely across the western end of the Simmons Mountain syncline. In this location, thrust-faults along both the northern and southern margins of the Simmons Mountain syncline are deflected around a southeast-trending anticline (Plate 4.1). This narrow fold at the western end of the Simmons Mountain syncline may result from along-strike shortening associated with the sharp bend in strike of the thrust belt in this location. The Windingstair fault marks the southern boundary of the Simmons Mountain syncline, and separates it from the Central Ouachita thrust belt.

A narrow hyperbolically shaped outcrop belt of Mississippian Stanley Group separates the eastern up-plunge end of the Simmons Mountain syncline from the western up-plunge end of the Rich Mountain syncline (Plate 4.1). The narrow belt of Stanley Group coincides with the crest of a narrow northeast-trending anticline (Plate 4.1). The Rich Mountain syncline is another broad, east-southeast-elongate basin with a fold axis along the same trend as the Simmons Mountain syncline. Morrowan Jackfork Sandstone-Johns Valley Shale comprises the core of the Rich Mountain syncline.
The north-vergent Honess thrust fault (marked by steep, south-dipping Mississippian Stanley Group at the base of the hangingwall) separates the Rich Mountain from another more narrow syncline on the north (Plates 4.1 and 4.2). The surface trace of the Honess fault dissapears to the west within the Simmons Mountain syncline where the fault loses displacement within Jackfork strata (Plate 4.1). The eastern terminus of the Honess fault is not intuitively obvious. Map patterns suggest that the Honess fault either loses displacement within, or overlaps with structures of the eastern Frontal Ouachita thrust belt (Plate 4.1). The eastern continuation of the Windingstair fault marks both the coincident southern boundaries of the Rich Mountain syncline and the transitional zone of the Frontal Ouachita thrust belt.

Definitive cross sections through the transitional zone of the Frontal Ouachita thrust belt are not published. However, cross sections constructed across the region (cross sections ARK-4 and OK-1 in Plate 4.3) suggest that the Briery, Honess, and Windingstair faults root within a relatively thin succession of pre-Mississippian (pre-orogenic) deep-water cherts and shales (Womble Shale through Arkansas Novaculite). This suggestion fits well with the observed map patterns roughly along strike in the Potato Hills (Central Ouachita thrust belt) and at Black Knob Ridge (westernmost exposed part of the Frontal Ouachita thrust belt) (Plate 4.1). In both of these locations relatively thin Ordovician-Mississippian (Womble-Arkansas Novaculite) sections are beneath much thicker Mississippian-Atokan successions.

**Eastern Frontal Ouachita thrust belt**

*Between Dutch Creek-Ross Creek-Cadron faults and Fourché-Perryville faults*

The north-vergent Dutch Creek-Ross Creek thrust faults strike east-northeast and separate the eastern part of the Frontal Ouachita thrust belt from the eastern Arkoma basin (Plate 4.1). The eastern end of the Ross Creek fault loses displacement abruptly and is a blind thrust fault south of the Cadron anticline. The small-displacement Cadron thrust fault marks the northern limit of the eastern Frontal Ouachita thrust belt. Farther east, Paleozoic formations of the Ouachita Mountains and Arkoma basin are buried by Mesozoic Gulf Coastal Plain deposits. However, sub-
crop maps of Paleozoic rocks indicate a southeastward bend in strike of the Ouachita thrust front beneath the Gulf Coastal Plain (Thomas and others, 1989).

Exposed rocks between the Dutch Creek-Ross Creek-Cadron and the Fourche-Perryville fault zones in the eastern Frontal Ouachita thrust belt consist mostly of Atokan-age Atoka Formation (Plate 4.1). The surface traces of the Dutch Creek and western part of the Ross Creek thrust faults separate shallow-water middle and upper Atoka Formation on the north from abruptly thicker lower Atoka Formation deep-water turbidites. Farther east along strike, the thickness of the middle and upper Atoka Formation is relatively constant; however, the thickness of the deep-water lower Atoka Formation decreases abruptly eastward along the Cadron fault.

The western end of the Fourché-Perryville fault extends westward into the transitional zone of the Frontal Ouachita thrust belt where it intersects the Y-City fault (Plate 4.1). It is unclear whether the Fourché-Perryville continues as a fault in the western part of the Frontal Ouachita thrust belt; however, inference based upon geologic maps suggests that the structural equivalent of the Fourché-Perryville fault west of the Black Fork syncline is the Ti Valley fault (Plate 4.1).

The surface trace of the Fourché-Perryville fault exhibits small displacement and is entirely within the Atoka Formation (Plate 4.1) (Haley and others, 1993). For most of the length of the fault, the upper part of the lower Atoka Formation in the hanging wall on the south is separated from middle Atoka Formation in the footwall (Plate 4.1) (Haley and others, 1993). Stratigraphic throw increases at the eastern end of the Fourché-Perryville fault, where the syncline to the north of the Fourché-Perryville fault is tighter (Plate 4.1, cross section C-C’ in Plate 4.2, and cross section ARK 1 in Plate 4.3) (Haley and others, 1993). The eastern end of the Fourché-Perryville fault bends sharply to the south (at the Arkansas River Valley) and continues as a fault on the southern side of the Cato syncline (Plate 4.1) (Haley and others, 1993).

A small, circular basin with basal Desmoinesian rocks in the core separates the eastern end of the Fourche-Perryville syncline on the west from the Bayou Meto anticline and two flanking synclines farther east (Cato syncline on the south, unnamed
A north-vergent thrust fault (eastern extension or structural equivalent of the Fourché-Perryville fault) separates the shallow- to marginal-depth upper and middle part of the Atoka Formation (on the north) exposed along the southern limb of the Cato syncline from deep-water lower Atoka Formation exposed on the south within the hanging wall (Plates 4.1 and 4.3). The abrupt eastward loss of displacement along the Ross Creek and Cadron faults combined with tightening of folds toward the south around the Bayou Meto anticline suggest a displacement transfer zone across the eastern end of the Frontal Ouachita thrust belt where north-directed displacement is transferred from the Ross Creek fault to the Fourché-Perryville fault (Plate 4.1 and cross sections ARK-1 and ARK-2 in Plate 4.3). In contrast to areas farther west where displacement along the Fourché-Perryville fault is minimal or absent, and stratigraphic separation small, displacement along the fault is much greater east of the Arkansas River valley where the surface trace of the Fourché-Perryville fault places Morrowan strata in the hanging wall on the south in contact with middle and upper Atokan strata in the footwall on the north (Plate 4.1).

Within the easternmost part of the Frontal Ouachita thrust belt, where north-directed displacement and shortening along the Ross Creek and Cadron thrust faults decrease abruptly, deformation is confined to structures farther south (Plates 4.1, 4.2, and 4.3). Cross sections of the eastern end of the Frontal Ouachita thrust belt suggest that, in this area, shortening is accommodated by fault imbrication of Morrowan strata
and folding of the thicker Atokan strata (Plate 4.3). Imbrication and underplating beneath the Morrowan (Johns Valley) detachment terminates south of the Cadron anticline where the overlying Atoka Formation thins abruptly along an east-west-striking basement fault (Plate 4.3). The Morrowan-age Johns Valley Shale in the core of the Bayou Meto anticline is along the crest of an imbricated triangle zone (Plate 4.3). Because the Morrowan-age Johns Valley Shale is in stratigraphic contact with the lower Atoka Formation around the nose of the narrow, west-plunging Bayou Meto anticline (Haley and others, 1993), the Johns Valley Shale in the core of the anticline is above a detachment.

**Between Fourché-Perryville fault and southern boundary**

The eastern part of the Frontal Ouachita thrust belt south of the Fourche-Perryville fault consists mostly of Mississippian to Morrowan sandstone and shale turbidites. In one location, Atokan sandstones are exposed in the core of a narrow syncline (Plate 4.1, and cross section ARK2 in Plate 4.3). The northernmost exposure of Arkansas Novaculite and overlying basal units of the Stanley Formation mark the southern boundary of the eastern Frontal Ouachita thrust belt.

The western part of the eastern Frontal Ouachita thrust belt strikes east-northeast and parallels the Dutch Creek and Ross Creek thrust faults (Figure 4.2 and Plate 4.1). Resistant Morrowan Jackfork Sandstone is separated from lower Atoka Formation shales by the Y-City fault (Plate 4.1). In the area east of the transitional zone of the Frontal Ouachita thrust belt and west of the Bayou Meto anticline (Figure 4.2), the Morrowan Johns Valley Shale crops out along the Y-City fault. South of the Bayou Meto anticline, a north-vergent thrust fault (eastern continuation of Fourche-Perryville fault) separates Jackfork Sandstone within the hanging wall on the south in fault contact with, from east to west, lower to middle Atoka Formation strata (Plate 4.1).

For most of its length, the Y-City fault separates upright or north-vergent folds (north of fault) from predominantly south-vergent folds (south of fault) (cross section C-C’ in Plate 4.2 and cross sections ARK1 and ARK 2 in Plate 4.3). The western end of the Y-City fault bends northward and intersects the Fourche-Perryville fault in the
transitional zone of the Frontal Ouachita thrust belt (Plate 4.1). West of the western end of the Fourche-Perryville syncline, folds are also predominantly north-vergent south of the Fourche-Perryville fault. The degree of south vergence increases both southward towards the Benton uplift and eastward along strike within the Frontal Ouachita thrust belt (Plates 4.2 and 4.3).

The Morrowan Jackfork Sandstone-Johns Valley Shale are much thicker than correlative formations penetrated by wells farther north in the Arkoma basin. Morrowan strata north of the Dutch Creek, Ross Creek, and Cadron faults range from less than 122 m thick (west) to more than 608 m thick (east) (Plate 4.3) (Van Arsdale and Schweig, 1990). The Morrowan section south of the Y-City fault has a maximum thickness of approximately 2128 m (Plate 4.3).

The large difference in thickness of the Morrowan strata north and south of the Y-City fault suggests northward transport of the Morrowan Jackfork Sandstone-Johns Valley Shale south of the Y-City fault. The Jackfork Sandstone-Johns Valley Shale are stratigraphically above Cambrian-Mississippian off-shelf cherts and distal turbidites that crop out on the south along the crest of the Benton uplift (Figure 4.3 and cross section D-D’ in Plate 4.2). According to the COCORP seismic profile shown in Lillie and others (1983), and the lithospheric model of Keller and others (1983) (also Mickus and Keller, 1992), the Cambrian-Mississippian shelf-edge is south of the Benton uplift. Therefore, the Jackfork Sandstone-Johns Valley Shale strata restore palinspastically south of the Benton uplift.

The combination of implied north-transport of the allochthonous strata south of the Y-City fault with predominant south-vergent folds and faults is one of many puzzles of Ouachita geology. Many published cross sections of the eastern Frontal Ouachita thrust belt model the Y-City fault as a steep south-dipping north-vergent thrust fault in the western part (Haley and others, 1977; Blythe and others, 1988) or a south-overturned thrust fault in the eastern part (Viele, 1989). However, the fault plane is not exposed, and formations dip to the north on both sides of the Y-City fault east of the transitional zone of the Frontal Ouachita thrust belt.
The combination of sub-parallel dips north and south of the Y-City fault and confinement of the fault to one stratigraphic unit suggest that the Y-City fault is a bed- 
ding-parallel back-thrust detachment rather than an overturned thrust fault. This Y-
City fault may be a surface expression of the southward continuation of the subsurface 
Morrowan detachment imaged on the COCORP seismic profile (Plate 4.4). The 
Morrowan (Johns Valley Shale) detachment is offset by the cross-cutting north-
vergence Dutch Creek-Ross Creek faults and the Fourche-Perryville fault (Plates 4.2 
and 4.3). Current geometry of the surface geology in the eastern Frontal Ouachita 
thrust belt is a function of the along-strike variation in throw of the Dutch Creek-Ross 
Creek faults and the Fourche-Perryville fault. East of the transitional zone of the 
Frontal Ouachita thrust belt and west of the Bayou-Meto anticline, where offset along 
the Fourche-Perryville fault is small, the Y-City backthrust detachment dips to the 
north and intersects the Fourché-Perryville fault in the subsurface (Plate 4.3). East of 
the Arkansas River valley, where offset along the Fourche-Perryville fault is much 
greater, the Y-City backthrust detachment is absent in the hangingwall where it has 
been displaced above present erosional level (Plates 4.1 and 4.3). In this location, the 
surface trace of the Fourche-Perryville fault separates Morrowan strata on the south in 
the hangingwall from middle Atoka Formation on the north (Plates 4.1 and 4.3). 

Whereas the northern boundary of the eastern Frontal Ouachita thrust belt is 
clearly defined by the Dutch Creek, Ross Creek, and Cadron faults, the southern 
boundary is much less precisely located. The southern boundary of the eastern Frontal 
Ouachita thrust belt is located roughly along the northernmost exposure of the 
Ordovician-Mississippian pre-orogenic cherts and low-grade slates that crop out along 
the northern flanks of the Benton uplift and eastern nappes (Figure 4.2 and Plate 4.1). 

Several laterally discontinuous south-vergent back-thrust faults parallel the 
northern flank of the Benton uplift (Plate 4.1). These faults range from at the top to 
within the Mississippian Stanley Formation. Observed wavelengths of surface folds 
vary dramatically north and south of the Stanley Formation outcrop belt along the 
northern flank of the Benton uplift. To the north, longer wavelength folds 
predominate in the thick stiff-layer Chesterian-Morrowan strata that include the upper
part of the Stanley Group and the Jackfork Sandstone. To the south, much shorter and variable wavelengths are found in the Cambrian-Mississippian (below upper Stanley Formation) formations in the core of the Benton uplift (Plates 4.1 and 4.3). The geometries within the core of the Benton uplift result from folding of numerous interlayered thin stiff-layer (sandstone turbidites and cherts) and thicker ductile-layer (low-grade slates/phyllites) strata.

The shale-dominated middle part of the Stanley Formation can be considered a wavelength/volume transition zone. The internal geometry within the middle layers of the Stanley Formation laterally varies to conform to the geometries of the surrounding stiff-layer formations. The Stanley Formation in the wavelength/volume transition zone is likely deformed by a combination of flexural slip and flexural flow processes.

Evidence for flexural flow seems more apparent along the southern boundary of the easternmost Frontal Ouachita thrust belt (Figure 4.2 and Plate 4.1). West of the surface trace of the basal Mesozoic unconformity which marks the western boundary of the Mississippi embayment in Arkansas, the Panther Creek fault marks the southern boundary of the easternmost Frontal Ouachita thrust belt (Plate 4.1). The Panther Creek fault separates tightly folded Mississippian-Morrowan turbidites on the north from complexely folded Cambrian-Mississippian cherts and distal turbidites of the Paron nappe on the south (easternmost part of the Central Ouachita thrust belt) (Figure 4.2 and Plate 4.1). The outcrop of Stanley Formation is very narrow along the Panther Creek fault and widens abruptly west of the Paron nappe. The Stanley Formation also wraps around a complex narrow, southeast-trending fold along the northeastern flank of the Alum Fork nappe (Figure 4.2 and Plate 4.1). A complexely folded thrust fault separates the Mississippian Stanley Formation in the footwall from Cambrian-Mississippian strata on the south in the hanging wall of the Alum Fork nappe. The abrupt variation in outcrop width combined with the complex folding in the region suggests flow of the Stanley Formation shales west of the Paron nappe (Plates 4.1 and 4.3).
Central Ouachita thrust belt (overview)

The lack of imbricated shallow-water facies rocks at the surface indicates that the basal décollement is in deep-water facies beneath the entire allochthonous terrain of the Central Ouachita thrust belt (Plates 4.1 and 4.3). The primary along-strike variation in the Central Ouachita thrust-belt is transition from long-wavelength folds in the Chesterian (upper part of Stanley Group) through Atokan sandstone-dominated turbidites on the west to short-wavelength folds in Cambrian-Chesterian (below upper part of Stanley Group) cherts and distal turbidites on the east in the Benton uplift. Long-wavelength folds are predominant in the western part of the Central Ouachita thrust belt where most of the outcrop belt consists of thick, stiff-layer sandstones (Figure 4.2 and Plate 4.1). In contrast, short-wavelength folds are predominant in eastern parts of the Central Ouachita thrust belt, where thin, stiff-layer cherts and laterally discontinuous sandstones are sandwiched between much thicker ductile-layer shales (Figure 4.2 and Plate 4.1).

The westward transition from exposed Cambrian-Ordovician rocks in the core of the eastern part of the Central Ouachita thrust belt to Mississippian and Pennsylvanian rocks exposed in the western part indicates that the basal décollement of the allochthon dips to the west and cuts up-section along strike. COCORP seismic data show that the allochthonous strata of the eastern part of the Central Ouachita thrust belt (Benton uplift) are draped over a broad basement anticline (Plate 3.5) (Lillie and others, 1983). Interpreted cross sections of Lillie and others (1983) show the northern flank of the buried basement anticline bordered by a north-vergent basement reverse fault. The presence of a thrust fault in the subsurface beneath the northern flank of the Benton uplift is consistent with the observed north-vergent surface structures within the northern part of the eastern Frontal Ouachita thrust belt (Plate 3.5 and cross section X-X’ in Plate 4.2). However, basement reflectors visible on the COCORP line are lost beneath the thickest part of the frontal thrust belt. Therefore, illustrations of basement geometry beneath the thick, central and southern
parts of the eastern Frontal Ouachita thrust belt and adjacent eastern part of the Central Ouachita thrust belt are poorly constrained.

The general west-directed plunge of surface folds across the western part of the Benton uplift show that the underlying basement anticline beneath the allochthonous surface strata also plunges to the west. In contrast to the eastern part of the Central Ouachita thrust belt where allochthonous Cambrian-Ordovician strata drape over a basement anticline, Mississippian and Pennsylvanian rocks exposed in the western part rest in the core of a basement syncline (or basin) (Plate 4.1).

Cross sections of the western part of the Central Ouachita thrust belt suggest that the southeastern margin of a northeast-elongate basement syncline (basin) beneath the allochthon terminates along a northeast-striking basement fault which extends in the subsurface beneath the western margin of the Broken Bow uplift (cross section D-D’ in Plate 4.2 and cross section OK2 in Plate 4.3). Allochthonous strata of the Broken Bow uplift are draped over the underlying basement anticline. The western limb of the basement anticline dips steeply to the northwest and is cross-cut by a northwest-vergent thrust fault (cross section D-D’ in Plate 4.2 and cross section OK2 in Plate 4.3). The northwest and northern sides of the Broken Bow uplift mark the southern boundary of the western Central Ouachita thrust belt. Except for the Potato Hills (Figure 4.2 and Plate 4.1), pre-Mississippian formations are not exposed northwest of the Broken Bow uplift within the western part of the Central Ouachita thrust belt.

The pre-Mississippian succession is much thinner within the Potato Hills than within the Broken Bow uplift (cross section D-D’ in Plate 4.2 and cross section OK2 in Plate 4.3). Surface geology and local wells indicate that only Ordovician-Mississippian (Womble-Arkansas Novaculite) rocks are in fault contact with underlying younger rocks in the Potato Hills (Allen, 1994). Thickness of the Potato Hills stratigraphic succession is closer to that of Black Knob Ridge in the westernmost part of the Frontal Ouachita thrust belt (Figure 4.2, Plate 4.1, cross sections D-D’ and E-E’ in Plate 4.2 and cross sections OK2 and OK 4 in Plate 4.3) (Hendricks and others, 1937; Morris, 1974; Allen, 1994).
Farther east along strike, a much thicker section of allochthonous pre-Mississippian rocks crops out in the Benton uplift (Figure 4.2, Plate 4.1, cross sections X-X’ and C-C’ in Plate 4.2 and cross sections ARK2, ARK3, and ARK4 in Plate 4.3). An east-northeast elongate, tightly folded rim of Ordovician-Mississippian cherts and novaculites surrounds the Benton uplift core. Complex folds and folded faults within Cambrian-Ordovician shale-dominated section comprise the central core of the Benton uplift. The eastern end of the Benton uplift intersects a southeast-trending core (Alum Fork nappe) which is also bounded on its flanks by Ordovician-Mississippian cherts and novaculites (Figure 4.2 and Plate 4.1). According to sparse well data and seismic profiles, this easternmost core region of the central Ouachita thrust belt extends southeastward beneath Mesozoic formations of the Mississippi embayment (Thomas and others, 1989).

No well has penetrated the autochthonous rocks within the eastern core of the Central Ouachita thrust belt. The nearest wells that penetrate autochthonous rocks are located to the southwest in the Broken Bow uplift (Leander and Legg, 1988; Denison, 1989). On the basis of data from these wells and from regional seismic and geophysical surveys, the allochthonous terrain of the eastern Central Ouachita thrust belt is inferred to overlie a buried carbonate shelf (Lillie and others, 1983; Kruger and Keller, 1986; Leander and Legg, 1988; Denison, 1989; Mickus and Keller, 1992). However, restorations of cross sections across the Benton uplift suggest that a layer of Mississippian-Morrowan transitional-to-deep-water clastic facies rocks cover the Cambrian-Mississippian carbonate platform east of the Broken Bow uplift (Plates 4.2 and 4.3) (Blythe and others, 1988).

A Bouger gravity anomaly map produced by Kruger and Keller (1986) shows a northeast-trending zone of steep gradient that parallels the western flank of the Broken Bow uplift and extends to the northwestern end of the Benton uplift (Figure 3.4). On the basis of this steep gravity gradient, geologic maps and cross sections suggest that a continuous fault is located along the northern flank of the Benton uplift and western flank of the Broken Bow uplift (Arbenz, 1989d; J.K. Arbenz, unpublished cross sections, 1996).
Alternatively, the steep Bouger gravity gradient and associated gravity low to the northwest may represent a localized depression of the Moho that may be unrelated to the uplifts of the Ouachita core regions. The contours of the gravity anomaly low and steep gravity gradient are oblique to the predominant trend of compressive structures of the central and eastern parts of the Ouachita Mountains (Figure 3.4 and Plate 4.1). This suggests that the stress field that caused the formation of the basement depression that coincides with the gravity anomaly low is not parallel to that which produced the surface folds exposed across the central and eastern Ouachita Mountains. Kruger and Keller (1986) suggest that an “Eocambrian” sedimentary basin may be located in the autochthon beneath the western part of the Central Ouachita thrust belt.

Another bouger gravity anomaly map published in Gatewood and Fay (1991) shows that the gravity low west of the Broken Bow uplift is separated into a set of semi-isolated elongated gravity anomaly lows (Figure 4.11). The Gatewood and Fay (1991) map is proprietary and model parameters are not listed for comparison with the Kruger and Keller (1986) map. The Gatewood and Fay (1991) map shows southeast-trending bends in the gravity anomaly pattern between the Broken Bow and Benton uplifts. This suggests that a southeast-striking basement fault separates the basement anticline beneath the Broken Bow uplift from the one beneath the Benton uplift (Figure 4.11).

Surface geology also suggests that the Broken Bow and Benton uplifts overly separate basement anticlines (Figure 4.2 and Plate 4.1). Tight folds that extend across the western nose of the Benton uplift plunge west-northwest. East-west trending tight anticlines cored by pre-Mississippian rocks become progressively narrower and shorter with distance north of the Broken Bow uplift (Plate 4.1). The northward younging of allochthonous strata north of the Broken Bow uplift shows that the northern nose of the basement anticline beneath the Broken Bow uplift plunges northeast (Plate 4.1). The allochthonous strata between the Broken Bow and Benton uplifts appear to be tightly folded along the axis of a northwest-plunging syncline. Farther east, strata exposed within east-west trending folds along the southern flank of the Benton uplift are progressively younger with distance southward from the core of the
Benton uplift. This southward younging of allochthonous strata, combined with consistent south-dip of beds south of the Benton uplift, indicate that the southern limb of the underlying basement anticline also dips southward beneath the eastern part of Southern Ouachita thrust belt (Plate 4.1, cross section X-X’ and C-C’ in Plate 4.2 and cross sections ARK1, ARK2, and ARK3 in Plate 4.3).

**Western Central Ouachita thrust belt**

**North of Octavia fault**

The Octavia fault is a laterally continuous thrust fault that marks the southern boundary of the Lynn Mountain syncline (Plate 4.1). From southwest to northeast, the Octavia fault varies from south-overturned thrust fault, where overlapped by the Gulf Coastal Plain, to a north-vergent thrust fault in eastern part of western Central Ouachita thrust belt. Both the Lynn Mountain syncline and Octavia fault extend farther east into the transitional Central Ouachita thrust belt (Figure 4.2 and Plate 4.1).

The Lynn Mountain syncline is a doubly plunging fold that is widest towards the center of the fold. The wide central part of the fold is located south of the Potato Hills (Plate 4.1). In this central part, Atoka Formation deep-water turbidites fill the core of the Lynn Mountain syncline. In one location in the center of the Lynn Mountain syncline, Morrowan deep-water turbidites crop out in the relatively narrow Hardy Creek anticline (Plate 4.1). The base of the Atoka Formation surrounding the Hardy Creek anticline outlines a tight anticline-syncline pair (Plate 4.1). The Morrowan formations exposed within the Hardy Creek anticline are cut by northwest-vergent thrust faults and southeast-vergent back-thrusts (Plate 4.1).

The location of a set of tight folds in close proximity to the Potato Hills anticline suggests a genetic relationship. Cross sections constructed perpendicular to strike through the Potato Hills and the central part of the Lynn Mountain syncline show the deep-water facies rocks of the Central Ouachita thrust belt deformed above a fault ramp located in the subsurface beneath the Potato Hills (cross section D-D’ in Plate 4.2 and cross section OK2 in Plate 4.3). The footwall northwest of the fault ramp is deformed and fault imbricated. Arching of the Hardy Creek anticline in the central part of the Lynn Mountain syncline, south of the Potato Hills, may be a
response to increased strain caused by folding above and behind the subsurface fault ramp.

Westward along strike from the central part of the fold, the width of the Lynn Mountain syncline decreases abruptly from a maximum of 20 km to approximately 5 km at the edge of the Gulf Coastal Plain (Plate 4.1). The westernmost exposed part of the Lynn Mountain syncline has a northeast strike and contains a core of tightly folded Morrowan Jackfork Sandstone with Mississippian Stanley Group along the northwestern margin. The tightly folded strata within the western part of the Lynn Mountain syncline are bounded by northwest-vergent thrust faults and sandwiched between tightly folded, northeast-striking anticlines of Mississippian Stanley Group (Plate 4.1, cross section D-D’ in Plate 4.2, and cross sections OK2 and OK3 in Plate 4.3). Near the margin of the overlapping Mesozoic Gulf Coastal Plain, the Octavia fault, which marks the southeastern boundary of the Lynn Mountains syncline, dips steeply to the north and is overturned towards the southeast (Plate 4.1, cross section E-E’ in Plate 4.2, and cross section OK4 in Plate 4.3). The tight folding and southeast-overturning of the Octavia fault indicate increased shortening and folding of the westernmost part of the Central Ouachita thrust belt. The distance between the Broken Bow anticline and the southeastern plunge of the Tishomingo-Belton anticline (Arbuckle uplift) decreases abruptly south of the Mesozoic Gulf Coastal Plain overlap (Plate 4.1). Thin-skinned structures of the western part of the Central Ouachita thrust belt are refolded within a basement depression located northwest of the Broken Bow uplift and northeast of the Tishomingo and Belton anticlines (Arbuckle uplift) (Plate 4.1). The center of the basement depression roughly coincides with the widest part of the Lynn Mountain syncline (Plate 4.1).

**Northwest of Lynn Mountain syncline**

Width and wavelengths of exposed folds within the western part of the Central Ouachita thrust belt differ greatly along strike northwest of the Lynn Mountain syncline (Plate 4.1, cross section D-D’ in Plate 4.2, and cross sections OK2, OK3, and OK4 in Plate 4.3). The map width of the Central Ouachita thrust belt northwest of the Lynn Mountain syncline narrows from more than 35 km in the western part, where
overlapped by Mesozoic strata, to less than 15 km east of the Potato Hills (Plate 4.1). In general, the westernmost part of the Central Ouachita thrust belt, northwest of the Lynn Mountain syncline, consists of several broad synclines and intervening narrow, fault-cored anticlines (Plate 4.1). A thick succession of Morrowan and Atokan deep-water turbidites crops out in the cores of the synclines (Plate 4.1, cross section E-E’ in Plate 4.2, and cross sections OK3, and OK4 in Plate 4.3). The Morrowan-Atokan succession consists of alternate sandstone and shale beds that comprise part of a thick composite stiff layer. In contrast, farther northeast along strike the long-wavelength folds of Morrowan-Atokan strata change into a composite antiform of Upper Ordovician to lower Mississippian strata at the Potato Hills (Plate 4.1, cross section D-D’ in Plate 4.2, and cross sections OK2 in Plate 4.3). Exposed within this composite antiform is a complexly and tightly folded and faulted succession of cherts, novaculites, and shales (Womble Shale-Bigfork Chert-Arkansas Novaculite) (Plate 4.1, cross section D-D’ in Plate 4.2, and cross sections OK2 in Plate 4.3). Except for Black Knob Ridge, the Potato Hills is the only outcrop area of pre-Meramecian (pre-Stanley Group) deep-water facies strata in the Ouachita Mountains northwest of the Broken Bow uplift (Plate 4.1, cross section E-E’ in Plate 4.2, and cross sections OK4 in Plate 4.3). Several structural characteristics shown in a cross section of the Potato Hills (cross section E-E’ in Plate 4.2, and cross sections OK4 in Plate 4.3) illustrate the kinematic and structural complexity of the western Central Ouachita thrust belt (and indeed that of the entire allochthon). One characteristic is the difference between wavelengths of folds in the Upper Ordovician-lower Mississippian (Kinderhookian-Osagean?) cherts and wavelength of folds in the upper part of the overlying Mississippian Stanley Group and Morrowan-Atokan strata exposed along strike southwest and northeast of the Potato Hills. The cherty formations are relatively thin and tightly folded, whereas the beds of the upper part of the Stanley Group (and overlying Morrowan-Atokan beds) comprise a much thicker mechanical stratigraphic unit with a much greater fold wavelength (cross section E-E’ in Plate 4.2, and cross sections OK4 in Plate 4.3). Differences in fold wavelengths indicate differential
shortening between the pre-Meramecian strata and overlying Mississippian-Atokan strata. In addition, several thrust faults that displace the pre-Mississippian strata appear to lose displacement in the lower part of the overlying Stanley Group (cross section D-D’ in Plate 4.2). The structural characteristics suggest that a bedding parallel detachment surface, and or disharmonic boundary, is located in the Stanley Group across the Potato Hills antiform and adjacent parts of the Central Ouachita thrust belt (and possibly the entire Ouachita Mountains).

Several authors disagree regarding kinematics of several faults in the northwestern part the Potato Hills antiform. Allen (1994) suggests that several north-dipping faults are antithetic back-thrusts with a south vergence that contrasts with the general north vergence interpreted from most of the surrounding structures. Other authors suggest that these north-dipping faults are north-vergent thrust faults that are part of a folded thrust fault that has corresponding southeast-dipping faults on the southwestern side of the Potato Hills antiform (cross section D-D’ in Arbenz, 1989a; and Miller and Smart, 2001). In either interpretation, these faults of debated kinetic origin are below the detachment/disharmonic boundary located in the overlying Stanley Group.

Cross sections of the western part of the Central Ouachita thrust belt show that the Potato Hills antiform and the broad folds located along strike to the west are within the footwall of the Lynn Mountain syncline thrust slice (cross sections D-D’ and E-E’ in Plate 4.2 and cross sections OK2, OK3, and OK4). Wells in the Potato Hills show that along the Potato Hills thrust fault (cross section D-D’ in Plate 4.2), Upper Ordovician-lower Mississippian Womble Shale-Arkansas Novaculite (pre-orogenic deep-water “Ouachita” facies) strata ramp over, and are fault-imbricated above, Mississippian-Pennsylvanian Stanley Group-Jackfork-Johns Valley strata (syn-orogenic deep-water clastic wedge) (Allen, 1994, Miller and Smart, 2001). Imbricated syn-orogenic clastic rocks (Stanley Group-Atoka Formation) emerge to the northwest in the imbricated zone of the western part of the Frontal Ouachita thrust belt (Figure 4.2 and Plate 4.1).
Northwest of the Lynn Mountain syncline, the along-strike change in stratigraphy and structural style of the Central Ouachita thrust belt on the northeastern and southwestern sides of the Potato Hills indicate an abrupt increase in depth of the Potato Hills thrust fault (detachment) on both sides of the Potato Hills antiform (Plate 4.1). For example, exposure of progressively younger strata with increased distance on both sides of the Potato Hills suggests an increase in depth of the Potato Hills detachment away from the core of the Potato Hills antiform. The increase in depth is greater toward the southwest where Morrowan and Atokan strata are exposed. This increase in depth can be attributed to various subsurface changes. For example, the westward decrease in number and volume of thrust-fault imbricates in the western Frontal Ouachita thrust belt suggests that westward decrease in volume allotted to the surface and subsurface part of the Frontal Ouachita thrust belt is counter-balanced by thickening and expansion of the Central Ouachita thrust belt (Figure 4.2 and Plate 4.1).

**South of Octavia fault**

The western Central Ouachita thrust belt southeast of the Octavia fault is broadly subdivided into an area of long-wavelength synclines and narrow, fault-cored anticlines flanking the Broken Bow uplift (Figure 4.2 and Plate 4.1). Map patterns and cross sections across the western Ouachita Mountains show progressive tightening of folds toward the westernmost exposed part of the Central Ouachita thrust belt (Figure 4.2, Plate 4.1, cross sections D-D’ and E-E’ in Plate 4.2, and cross sections OK3 and OK4 in Plate 4.3).

Map patterns illustrated in Plate 4.1 show at least two fault displacement-transfer zones between the Lynn Mountain syncline and the Broken Bow uplift. One displacement-transfer zone is located along the southern flank of the Boktukola syncline (Plate 4.1). Here displacement of the eastern end of the surface trace of the Big One fault terminates abruptly eastward in Morrowan Jackfork Sandstone. Nearby on the south, the western end of the surface trace of the Boktukola fault loses displacement westward in the upper part of the Morrowan Jackfork Sandstone or Johns Valley Shale (Plate 4.1). Another displacement-transfer zone
surrounds the Pickens anticline (Plate 4.1). Map patterns and regional cross sections show that both the Cloudy fault and connecting Big Waterhole Creek fault are folded over the southwest-plunging nose of the Pickens anticline (Arbenz, 1989d) (cross sections E-E’ in Plate 4.2, and cross sections OK4 in Plate 4.3). Here the Big Waterhole Creek/Cloudy fault appears to terminate abruptly eastward along the southern flank of the Pickens anticline. It is unclear whether displacement of the eastern end of the Big Waterhole Creek/Cloudy fault transfers to the Boktukola fault, transfers to another adjacent fault, or whether it merely loses displacement towards the east.

**Transitional Central Ouachita thrust belt**

**North of Octavia fault**

Map width between the Octavia and Windingstair faults decreases abruptly in the transitional part of the Central Ouachita thrust belt (Figure 4.2 and Plate 4.1). Approximate distance between the two faults decreases from 30 km east of the Potato Hills to 15 km across the transitional Central Ouachita thrust belt. To maintain constant volume, this abrupt decrease in map width is likely compensated by increase in fold amplitude (depth) and fault imbrication.

The west-plunging, eastern nose of the Lynn Mountain syncline is in close proximity to the west-plunging, western end of the Benton uplift anticlinorium (Plate 4.1). The tight folds of Upper Ordovician-lower Mississippian Bigfork Chert-Arkansas Novaculite strata, which delineate the shape of the western margins of the Benton uplift and Cossatot Mountains anticlinoria, appear to be out-of-phase with the Mississippian-Atokan strata within the Lynn Mountain syncline, which projects above the western Benton uplift (Figure 4.12 and Plate 4.1). An alternative interpretation suggests that the Lynn Mountain syncline tightens eastward north of the Cossatot Mountains anticlinorium (Figure 4.12).

Contrast between Meramecian-Chesterian Stanley Group on the southeast side of the Octavia fault and the uppermost Chesterian-Morrowan Jackfork Sandstone on the northwest clearly marks the position of the Octavia fault through most of the Central Ouachita thrust belt (Plate 4.1). However, the eastern end of Octavia fault is
not clearly defined (Plate 4.1). The geologic map of Arbenz (1989b) suggests that the
eastern end of the Octavia fault loses displacement within the northwestern part of the
Cossatot Mountains (as shown in Plate 4.1). It is also possible that the Octavia fault
continues farther east along the northern part of the Cossatot Mountains and extends
into the central part of the Mazarn syncline (Plate 4.1).

A relatively narrow zone of tightly folded Stanley Group shales separates the
tightly folded pre-Meramecian strata of the western Benton uplift from the broadly
folded Chesterian-Morrowan interbedded sandstones and shales (upper Stanley
Group-Jackfork Sandstone) of the eastern part of the Lynn Mountain syncline (Plate
4.1). An abrupt transition separates the long wavelength fold of the upper Stanley
Group-Jackfork strata and the underlying middle and lower part of the Stanley Group.
This transition has been interpreted either as a disharmonic boundary (Arbenz, 1989d)
or as a folded unconformity/detachment surface (Babei and Viele, 1992). Recent field
work south of the Octavia fault on the eastern flank of the Boktukola syncline also
suggests that an low-angle detachment fault (1 m thick brecciated zone) marks the
abrupt transition (Williamson and Nielsen, 2001).

Another less abrupt transition boundary separates the tight folds of the lower
and middle Stanley Group from the tight folds of basal Stanley Group and Arkansas
Novaculite of the western Benton uplift. In general, the area of tightly folded lower
and middle Stanley Group that is sandwiched between the two fold-wavelength transi-
tion boundaries can be considered a wavelength/volume transition zone. Here, the
lower and middle part of the Stanley Group serves as a ductile layer and deforms to
compensate for, or fills the gap between, the folds of the overlying stiff layer (upper
Stanley-Jackfork-Atoka) and the underlying stiff layer (Arkansas Novaculite-basal
Stanley Group) (Plate 4.1).

Between Octavia and Boktukola faults

South of the Octavia fault, the area of the lower and middle Stanley Group
wavelength/volume transition zone is much greater than the corresponding area to the
north of the Octavia fault (Figure 4.2). An abrupt fold-wavelength transition along the
eastern margin of the west-plunging Boktukola syncline marks the western boundary
of the Stanley Group wavelength/volume transition zone (Figure 4.2 and Plate 4.1). A contorted fold-wavelength transition boundary along the western end of the Cossatot Mountains marks the eastern margin of the Stanley Group wavelength/volume transition zone (Figure 4.2 and Plate 4.1). Tight folds of basal Stanley Group (Meramecian) and underlying Arkansas Novaculite (Devonian-Kinderhookian-Osagean?) on the western flank of the Cossatot Mountains change from northwest-plunging in the north, to west-plunging in the south (Figure 4.2 and Plate 4.1). The tight folds of the western Cossatot Mountains can be interpreted as either out-of-phase or in-phase with the broad folds to the west (Plate 4.1). The Stanley Group wavelength/volume transition zone is cross-cut by three east-striking, south-dipping and north-vergent thrust faults (Figure 4.2, Plate 4.1, and cross section ARK1 in Plate 4.3).

South of Cossatot Mountains-Trap Mountains trend

South of the Boktukola fault, the east-west trending Cross Mountains are located in the center of the southernmost part of the transitional zone of the Central Ouachita thrust belt (Figure 4.2 and Plate 4.1). The Cross Mountains is a doubly plunging anticlinorium with fold noses that plunge both east and west. Upper Ordovician Bigfork Chert is exposed in the cores of the larger anticlines, surrounded by Silurian strata and Devonian-lower Mississippian Arkansas Novaculite and basal Stanley Group shale. A simplified cross section of the central part of the Cross Mountains anticlinorium shows that the folds are nearly upright and possibly slightly north-vergent (Figure 3.2) (Nielsen and others, 1989). This slight north-vergence is in stark contrast to the clear south-vergence of the same strata within most of the Benton uplift on the northeast, and within the western and northern parts of the Broken Bow uplift on the southwest (cross sections X-X’ and C-C’ in Plate 4.2 and cross sections ARK2 and ARK3 in Plate 4.3). However, folds and faults within the Stanley Group wavelength/volume compensation zone north of the Boktukola fault, and folds within the westernmost parts of the Cossatot Mountains and Benton uplift are generally near-vertical or north-vergent (cross section X-X’ in Plate 4.2 and cross sections ARK3 and ARK4 in Plate 4.3) (Zimmerman and others, 1982).
The eastern end of the Cross Mountain anticlinorium nearly merges farther east with the southern flank of the Cossatot Mountains anticlinorium (Plate 4.1). The southern edge of the Cross Mountains anticlinorium and eastern part of the Cossatot Mountains anticlinorium mark the southern margin of the transitional zone of the Central Ouachita thrust belt (Figure 4.2). Farther east, the west-plunging Trap Mountains anticlinorium marks the southern boundary of the eastern part of the Central Ouachita thrust belt (Figure 4.2). South of the Cross Mountains-Cossatot Mountains-Trap Mountains trend is the Southern Ouachita thrust belt (Figure 4.2).

Eastern Central Ouachita thrust belt

North of Cossatot-Trap Mountain trend (overview)

The eastern part of the Central Ouachita thrust belt is subdivided into two broad structures; the Benton uplift and the Mazarn syncline (Figure 4.2 and Plate 4.1). The eastern part of the Benton uplift is subdivided into the Jessievile nappe, Paron nappe, and Alum Fork nappe (Figure 4.2 and Plate 4.1). The eastern Central Ouachita thrust belt is bounded on the north by late Mississippian-Morrowan turbidites of the eastern Frontal Ouachita thrust belt (Figure 4.2 and Plate 4.1). Tightly folded Ordovician-Mississippian cherts, novaculites, and distal turbidite strata of the Cossatot Mountains (northwest-southeast-trending folds) and Trap Mountains (east-west-trending folds) mark the southern boundary of the eastern Central Ouachita thrust belt (Figure 4.2 and Plate 4.1).

Map patterns and cross sections of the eastern Ouachita Mountains show progressive eastward tightening of folds across the Benton uplift and farther east within the eastern nappes of the Central Ouachita thrust belt (Plate 4.1, cross section C-C’ in Plate 4.2, and cross section ARK1 in Plate 4.3). The eastward increase in shortening is best evidenced by the eastward change from the single folded Benton uplift anticline north of the Cossatot Mountains trend to the more complexly folded Benton uplift-Mazarn syncline fold-pair along strike farther east (Plate 4.1 and cross sections ARK2 and ARK3 in Plate 4.3). In addition to eastward tightening, folds and folded thrust faults within the Benton uplift exhibit progressively greater southward overturning towards the east (Plate 4.1 and cross sections ARK2 and ARK3 in Plate 4.3). Cross
sections and map patterns of the Benton uplift show several folded thrust fault imbricates within the eastern part of the core area which are absent from the western part (Plate 4.1, and cross sections ARK2, ARK3, and ARK4 in Plate 4.3).

The Mazarn syncline separates the broader core area of the Benton uplift, where Cambrian-early Mississippian pre-orogenic cherts, novaculites, and distal turbidites are exposed, from the much narrower Trap Mountains anticlinorium, where Late Ordovician-early Mississippian cherts and novaculites protrude through surrounding Mississippian (Meramecian-Chesterian) synorogenic turbidites (Figure 4.2 and Plate 4.1). The Mazarn syncline trends east-northeast and is narrower and deeper toward the west. The central and western part of the central axis of the Mazarn syncline is cut by a steep, south-dipping, north-vergent thrust fault (Figure 4.2 and Plate 4.1). Here, tightly folded Devonian-Mississippian (Arkansas Novaculite and Stanley Formation) strata on the south are juxtaposed against Mississippian-Morrowan (Stanley Formation-Jackfork Sandstone) strata on the north side of the fault. A central core of Morrowan Jackfork Sandstone is surrounded by Mississippian Stanley Formation and Ordovician-Mississippian cherts and shales (Bigfork Chert-Arkansas Novaculite) (Plate 4.1 and cross section ARK2 in Plate 4.2). The abrupt change from predominantly south-vergent structures within the Mazarn syncline (and the Benton uplift to the north) to north-vergent structures on the south side of the Trap Mountains fold trend marks the boundary between the eastern parts of the Central Ouachita and Southern Ouachita thrust belt (Figure 4.2, and Plates 4.1 and 4.2).

**Eastern nappes region (overview)**

Strata within the core area of the eastern part of the Benton uplift and eastern nappes are more complexly folded and faulted than strata along strike to the west in the western part of the Benton uplift (Figure 4.2, Plate 4.1, cross sections X-X’ and C-C’ in Plate 4.2, and cross sections ARK2, ARK3, and ARK4 in Plate 4.3). A brief summary of the structural style of each of the eastern nappes is given in the following paragraphs.
**Jessieville nappe (window)**

The smallest of the eastern nappes, the Jessieville nappe, is located near where the strike of the northeastern part of the Central Ouachita thrust belt changes abruptly from east-northeast-striking to southeast-striking (Figure 4.2 and Plate 4.1). The Jessieville nappe is a northeast-elongated outcrop belt of Upper Cambrian to Ordovician rocks surrounded by Ordovician and younger rocks that are bounded by an irregularly shaped combination of faults; therefore, a more appropriate name for the outcrop belt of the Jessieville area is the Jessieville “window” (Plate 4.5). The northeast-striking part of the Jessieville fault marks the northwestern boundary of the Jessieville nappe (window) (Plate 4.5). Strata on both sides of this part of the Jessieville fault dip approximately 45° towards the northwest (Haley and Stone, 1994). From the southwestern end of the Jessieville fault (sensu stricto) where Ordovician Mazarn Shale crops out on both sides of the fault, apparent stratigraphic displacement along the fault increases towards the northeast where Upper Cambrian Collier Formation on the southeast is separated from Ordovician Mazarn Shale on the northwest (Plate 4.5). On the basis of conodont and trilobite biostratigraphy, Upper Cambrian Collier Formation beds which crop out along the northern boundary of the Jessieville nappe (window) in this location are the oldest exposed strata of the entire Ouachita Mountains (Plate 4.5) (Hart and others, 1987; Ethington and others, 1989; Nielsen and others, 1989).

Strata within the Jessieville nappe (window) (Plate 4.5) consist of a tightly folded, southwest-plunging anticline/syncline pair with south-vergent axial surfaces (Nielsen and others, 1989; Haley and Stone, 1994). A relatively large area of Upper Cambrian to Ordovician Collier Formation crops out across the northeastern part of the Jessieville nappe (window); whereas, overlying Ordovician Crystal Mountain Sandstone and Mazarn Shale outcrop along the southeastern part (Plate 4.5). In general, folds within the Jessieville nappe (window) are chevron to isoclinal; cleavage is parallel with the axial planes of the folds (Nielsen and others, 1989). North-dipping small-scale faults are pervasive and either offset strata within the cores of synclines as south-vergent reverse faults, or offset limbs of tight isoclinal folds as extensional
faults (Nielsen and others, 1989). In the western and central parts of the Jessieville nappe, beds, cleavage, and axial planes all strike northeast-southwest and are parallel with the predominant structural trend of the Benton uplift (Nielsen and others, 1989). In the eastern part of the Jessieville nappe (window), the strike of all planar surfaces bends abruptly to a northwest-southeast strike on the western side of the Alum Fork small-scale recess (Plate 4.5) (Nielsen and others, 1989).

Whereas the western part of the Jessieville fault (Jessieville fault *sensu stricto*) exhibits a relatively straight fault trace, the shape of the eastern part of the fault varies according to author (Arbenz, 1989; Nielsen and others, 1989; Haley and Stone, 1994). A map of the Jessieville area illustrated in Nielsen and others (1989) does not precisely delineate the eastern part of the Jessieville fault; however, the fault is shown to extend with a northeast strike and bend abruptly to a southeast strike on the eastern side of the Jessieville nappe (window). The other faults bounding the Jessieville nappe (window) are not illustrated on the map of Nielsen and others (1989). The Arbenz (1989b) map shows a sharp northward bend in the fault on the northern side of the Jessieville outcrop belt where the fault connects with or continues into other faults farther north (Plate 4.5). The Arbenz (1989b) map shows an irregularly shaped fault along the eastern boundary of the Jessieville nappe (window). The unpublished CoGeo geologic maps of the Jessieville and adjacent Goosepond Mountain 7.5 minute quadrangles of Haley and Stone (1994) provides much more precise interpretation of surface geology and shows the eastern part of the Jessieville nappe (window) bounded by an irregularly shaped extension of the trace of the Jessieville fault (Plate 4.5). The irregular trace suggests that the fault surface dips less steeply along the eastern margin of the Jessieville window than it does farther west.

The southern and western boundary of the Jessieville nappe (window) is a composite boundary consisting of a set of intersecting faults. One of these sets of faults is bedding parallel within the Ordovician Mazarn Shale near the base of the Blakely Sandstone and can be named for simplicity the Mazarn/Blakely detachment (Plate 4.5). The Mazarn/Blakely detachment surface drapes over tightly folded Mazarn Shale and underlying Crystal Mountain Sandstone strata and is delineated
along the southern and western margins of the Jessieville nappe (window) (Plate 4.5). An irregularly shaped, arcuate and gently south-dipping fault along the southeastern margin of the Jessieville nappe (window) places north-dipping Ordovician Blakely Sandstone on the south (little Blakely slab) against Upper Cambrian-Ordovician Collier Formation on the north (Plate 4.5) (Haley and Stone, 1994). The fault at the base of the little Blakely slab truncates the Mazarn/Blakely detachment to the west and is truncated to the south by a northeast- to east-striking thrust fault (Fault A) with Blakely Sandstone on the south (Plate 4.5). Map patterns and cross sections indicate that the fault at the base of the little Blakely Slab is a displaced fault slice that may have been a splay of Fault A and not an extension of the Mazarn/Blakely detachment (Plate 4.5).

The unpublished Jessieville and Goosepond Mountain, and selected adjacent geologic quadrangle maps of Haley and Stone (1990, 1991, 1993, and 1994) suggest that faults of three genetic types cross-cut the vicinity of the Jessieville nappe (window). One type of fault (Type 1) consists of detachment faults that are generally restricted to one stratigraphic unit and are locally bedding parallel and tightly folded (Plate 4.5). Another type of fault (Type 2) consists of laterally continuous faults that cross cut the Jessieville and surrounding areas and appear to truncate, and therefore postdate, the detachment faults (Plate 4.5). Another type of fault (Type 3) mapped in the Jessieville area consists of north- and northeast-vergent thrust faults and south- and southwest-vergent backthrusts that intersect with, or splay off, the more laterally continuous cross-cutting faults (Plate 4.5).

One example of a Type 1 detachment faults is within the Ordovician Mazarn Shale and is described as the Mazarn/Blakely detachment. On the western side of the Jessieville window, Haley and Stone (1994) show bedding-parallel faults that locally are in the upper part of the Mazarn Shale, parallel with the overlying Blakely Sandstone, and a fault in the lower part of the Mazarn Shale that is parallel with the underlying Crystal Mountain Sandstone (Mazarn/Crystal Mountain detachment) (Plate 4.5). Another of the Type 1 detachment faults is found east of the Jessieville nappe.
(window) and separates Ordovician Womble Shale and/or Bigfork Chert from underlying older Ordovician Womble Shale, Blakely Sandstone, or Mazarn Shale.

Fault A (Plate 4.5) is an example of a laterally continuous Type 2 fault. Fault A cuts up-section from the west, where it contains Ordovician Mazarn Shale and Blakely Sandstone on the southern side of the fault in the hangingwall, to the east, where it contains Ordovician Womble Shale through Mississippian Stanley Group strata (Plate 4.5). In the area enclosed by the Goosepond Mountain quadrangle, the eastern part of Fault A cross-cuts the Womble detachment (Plate 4.5). In this location, both Fault A and the Womble detachment, and associated strata within the hanging walls, are deformed into tight folds with northwest-southeast trends (Plate 4.5). Because the predominant dip of beds is toward the north (including south-overturned beds) in the Jessieville area and much of the surrounding parts of the Benton uplift and eastern nappes areas, the Type 2 faults are interpreted as predominantly north-dipping, south-overturned thrust faults, and locally as backthrusts (Plate 4.5).

On the basis of map patterns, it is not possible to assign the Jessieville fault precisely to either the Type 1 detachment fault or Type 2 laterally continuous cross-cutting fault category. If modeled as a south-overturned thrust fault, the southwestern part of the fault cuts upsection to the southeast from Upper Cambrian Collier Formation to Ordovician Mazarn Shale and Blakely Sandstone where it intersects the Mazarn/Blakely detachment. If the Jessieville fault is viewed as a north-dipping fault that marks the northern side of a tectonic window, where younger Ordovician Mazarn Shale and overlying strata slide over older Ordovician and Upper Cambrian strata, then the Jessieville fault should be classified as a detachment fault, and possibly the eastward extension of either the Mazarn/Blakely or Mazarn/Crystal Mountain detachment (Plate 4.5).

From west to east across the Jessieville and Goosepond Mountain quadrangles, strikes of Type 3 thrust faults and backthrust faults change from east-west or northeast-southwest strikes across the western side of the Jessieville nappe (window) to northwest-southeast strikes (east) near the western margin of the Alum Fork small-
scale recess (Plate 4.5). Map patterns and truncation of these type 3 thrusts and backthrusts at adjacent laterally continuous (type 2) faults suggests that the type 3 faults were formed in response to folding of the previous faulted terrain.

**Alum Fork and Paron nappes (overview)**

East of the Jessieville window, the Alum Fork small-scale recess and Little Rock small-scale salient separate the Jessieville window and the rest of the eastern part of the Benton uplift from the Alum Fork and Paron nappes (Figure 4.2, Plates 4.1 and 4.5). A narrow, southeast-trending exposure of Mississippian Stanley Group turbidites is located on the northern side of several intersecting faults which outline the center of the Alum Fork recess. In a broader sense, the Alum Fork recess is a location where a boundary fault, which separates pre-Upper Ordovician (pre-Bigfork Chert) strata on the south from Upper Ordovician Bigfork Chert through Mississippian Stanley Formation on the north, bends sharply from a northeast- or east-strike (in the Little Rock salient) to a southeast-strike (Plate 4.5). The predominant strike of the pre-Stanley Formation strata and faults northwest of the boundary fault is northwest-southeast (Plate 4.5). Outcrop patterns of the Jessieville nappe (window) part of the eastern Benton uplift indicate that pre-Bigfork Chert strata dip northwest beneath, or are folded underneath, overlying Bigfork Chert through Mississippian Stanley Formation strata (Plate 4.5).

Because geometry of formation boundaries and fault traces vary according to interpretation, it is not possible to precisely connect, or correlate, detachments and faults that are mapped on both sides of the Alum Fork recess (Nielsen and others, 1989; Haley and others, 1993; Haley and Stone, 1990, 1991, 1993, and 1994). The Lonsdale fault marks the northwestern boundary of the Alum Fork recess, whereas the southeast-striking part of the Alum Fork thrust fault marks the eastern boundary (Plate 4.5). The Lonsdale fault is a southwest-vergent, overturned thrust fault where Ordovician-Mississippian (Bigfork Chert-Stanley Formation) strata on the southwest are juxtaposed against Stanley Formation on the northeast (Plate 4.5). The southeast-striking segment of the Alum Fork thrust fault separates southeast-vergent Ordovician Mazarn Shale strata (Haley and others, 1993; and Haley and Stone, 1990, 1991, 1993,
and 1994), and possibly some laterally discontinuous Blakely Sandstone beds (Nielsen and others, 1989; Haley and others, 1993), on the southeastern side of the Alum Fork thrust fault from southeast-, to east-striking Mississippian Stanley Formation and underlying older strata on the northwestern side (Plate 4.5). Strata within the hanging wall of the Alum Fork thrust fault are folded over Mississippian Stanley Formation and older strata in the complexly folded and faulted footwall (Plate 4.5).

Both the small-scale Alum Fork recess and Little Rock salient mark the northern margins of the Alum Fork and Paron nappes (Plate 4.5). Tightly folded, resistant Upper Ordovician-lower Mississippian chert and novaculite strata (Bigfork Chert-Arkansas Novaculite) exposed within the Zig Zag Mountains separate western part of the Alum Fork nappe from the eastern part of the Mazarn syncline (Figure 4.2, Plate 4.1). The east- to southeast-striking Paron fault separates the northeastern part of the Alum Fork nappe on the south, from the narrow Paron nappe on the north. In contrast to parts of the central core of the Benton uplift and Jessieville nappe (window) to the southwest where rocks as old as the Upper Cambrian-Ordovician Collier Shale are exposed, rocks older than the Ordovician Mazarn Shale are absent within the Alum Fork and Paron nappes (Nielsen and others, 1989; Haley and others, 1993; Haley and Stone, 1990, 1991, 1993, and 1994).

**Paron fault and detachment fault rooted in the Womble Shale**

The location of the Paron fault, which separates the Paron nappe on the north from the larger Alum Fork nappe on the south is somewhat arbitrarily located (Figure 4.2, Plates 4.1 and 4.5) (Nielsen and others, 1989). A number of pod-shaped serpentinite, meta-gabbro, and metamorphosed ultra-mafic rock outcrops (interpreted ophiolites) roughly delineate the trace of the Paron fault (Arbenz, 1989d; Nielsen and others, 1989). In general, strata on both sides of the Paron fault dip to the northeast with slightly older rocks on the north side of the fault.

Mapped stratigraphy on both the north and south sides of the Paron fault varies according to several authors. Both Nielsen and others (1989) and Haley and others (1993) show that on the north side of the fault, exposed strata range from Ordovician Womble Shale on the southeast to slightly older Blakely Sandstone on the northwest.
Nielsen and others (1989) and Haley and others (1993) also show that Womble Shale is exposed along the entire length of the south side of the Paron fault. According to either of the above geologic maps, the Paron fault may be interpreted as either a backthrust, or the erosional trace of a northeast-dipping detachment fault that drapes over an anticlinal bulge in the allochthon south of the Paron fault.

A geologic map derived from unpublished geologic quadrangle maps of Haley and Stone (1990, 1991, 1993, and 1994) of the Paron and Alum Fork nappes show a different interpretation (Plate 4.5). In this interpretation, an outcrop belt of Mazarn Shale, cross-cut by east- to southeast-striking backthrusts, is located along the entire length of the north side of the Paron fault (Plate 4.5). Blakely Sandstone, shown as laterally discontinuous in the southern part of the Paron nappe and the northeastern part of the Alum Fork nappe in Haley and others (1993) is not shown in the Haley and Stone (1990, 1991, 1993, and 1994) interpretation (Plate 4.5). A narrow, fault-bounded belt of Womble Shale separates the Mazarn Shale of the southern part of the Paron nappe from the tightly folded and faulted, east- to southeast-striking outcrop belt of Ordovician Womble Shale to Mississippian Stanley Formation located farther north (Plate 4.5).

Another significant change in the interpretation illustrated in Haley and Stone (1990, 1991, 1993, and 1994) is the outcrop of a large area of Mazarn Shale south of the Paron fault, within the Alum Fork nappe, which in the Nielsen and others (1989) and Haley and Stone (1993) maps is shown as the younger Womble Shale (Plate 4.5). South of the Paron fault, an east-trending outcrop belt of faulted and folded Womble Shale and Bigfork Chert located near the Alum Fork recess is separated by a larger, east- to southeast-trending outcrop belt of the same strata by an outcrop belt of faulted, east-striking Mazarn Shale (Plate 4.5). In the Haley and Stone (1990, 1991, 1993, and 1994) interpretation, the roughly east-west-striking Paron “fault” is a composite of several intersecting faults (Plate 4.5). Along the Paron “fault” the northern sides of the detachment faults rooted in the Womble Shale intersect and are nearly parallel with an east- to southeast-striking backthrust which offsets Mazarn Shale (Plate 4.5). The younger Womble Shale and Bigfork Chert strata above the detachment are tightly
folded and are structurally above the older tightly folded Mazarn Shale. This geometry suggests a younger over older detachment fault or displacement which operated along a weak shale layer during deformation of the entire allochthon.

Because of structural complexity, it is difficult to trace faults from the Alum Fork and Paron nappes westward; however, detachment faults shown at the base of the Womble and BigFork Chert formations farther west in the Jessieville area (Plate 4.5) may be westward continuation of detachment faults at the base of the Womble Shale in the southern part of the Paron nappe and northern part of the Alum Fork nappe (Plate 4.5).

**Significance of the geology of the northern Paron nappe**

North and northeast of the Paron fault, a relatively narrow outcrop belt of Ordovician Womble Shale and Blakely Sandstone (Nielsen and others, 1989; Haley and others, 1993), or Mazarn Shale (Haley and Stone, 1990, 1991, 1993, and 1994), which comprises the southern part of the Paron nappe, is in fault contact with a wider outcrop belt comprised of tightly folded Ordovician Womble Shale through Mississippian Stanley Formation (Plate 4.5). Folds on both sides of the fault are overturned toward the south, and the fault may be interpreted as a south-overturned thrust fault or a backthrust formed during folding of the allochthon. In the Haley and others (1993) interpretation, stratigraphic throw of the fault increases from the southeast, where Womble is in contact with Womble, to the northwest, where Bigfork Chert on the north is juxtaposed against Blakely Sandstone on the south (Plate 4.5). In the Haley and Stone (1990, 1991, 1993, and 1994) interpretation, stratigraphic throw along the fault is essentially constant along the fault where Womble Shale on the north is juxtaposed against Mazarn Shale on the south (Plate 4.5).

The northern boundary of the Paron nappe (or thrust wedge) is the north-dipping Panther Creek fault which separates tightly folded, thin-bedded, Ordovician to Mississippian strata on the south from broadly folded, thick-bedded, Mississippian to lower Pennsylvanian ( Morrowan) Stanley Formation to Jackfork Sandstone turbidites (Plate 4.5). Haley and Stone (1990, 1991, 1993, and 1994) show a small, east-west trending outcrop of Jackfork Sandstone in the core of a narrow syncline in the central
part of the Paron nappe (Plate 4.5). An interpreted thrust fault is shown at the base of the Jackfork Sandstone (Plate 4.5). The predominant south-vergence of folds to the south of the Panther Creek fault formed during north-directed underplating of early Mississippian and older allochthonous strata beneath the Panther Creek backthrust (cross section C-C’ in Plate 4.2 and cross section ARK 1 in Plate 4.3).

A key feature of the Paron nappe is the sedimentology of the Devonian to early Mississippian (Kinderhookian-Osagean?) Arkansas Novaculite. In contrast to the Arkansas Novaculite which crops out farther west along the northern side of the Benton uplift and Jessievile window area which are classified as a “northern facies,” the Arkansas Novaculite strata of the Paron nappe are more similar to the “southern facies” Arkansas Novaculite strata which crop out along the south side of the Benton uplift and southwest of the Alum Fork nappe (Nielsen and others, 1989). It is also key that the strata within the Paron nappe are above an interpreted detachment fault rooted in the Womble Shale (Plate 4.5) (based upon Haley and Stone, 1990, 1991, 1993, and 1994). Map patterns suggest that the Womble Shale through Stanley Formation strata exposed within the northern part of the Paron nappe are above the northern continuation of the Womble detachment (Plate 4.5). The final observation one needs to make is to notice that strata at the the northern end of the Paron nappe are imbricated south of a thick succession of Mississippian to Atokan turbidites on the northern side of the Panther Creek fault (Plate 4.5). The placement of a “southern facies” of Arkansas Novaculite at the northern end of the eastern Central Ouachita thrust belt is consistent with a break-back fault rooted in the Womble Shale (Womble detachment). Younger Womble Shale-Arkansas Novaculite (and lower Stanley Formation) are displaced northward over older Mazarn Shale and underlying allochthonous strata. Allochthonous strata above and below the Womble detachment which are imbricated beneath (and south of) the thick Mississippian-Atokan turbidites on the north side of the Panther Creek fault (Figure 4.14).

**Alum Fork nappe (key points)**

The predominant surface geology of the Alum Fork nappe, south of the Paron fault, varies according to interpretation. According to Viele (1966) and Haley and
Ordovician Womble Shale crops out across most of the Alum Fork nappe. Haley and others (1993) show an outcrop belt of slightly older Ordovician Mazarn Shale and Blakely Sandstone along the northwestern margin of the Alum Fork nappe (Plate 4.5). In the north-central, and northeastern part of the Alum Fork nappe, Upper Ordovician Bigfork Chert (and possibly Polk Creek Shale) are exposed in the cores of a series of tightly folded synclines (Plate 4.5).

According to Haley and Stone (1994), Middle Ordovician Mazarn Shale crops out across most of the Alum Fork nappe (Plate 4.5). Unlike in older maps, outcrop of Blakely Sandstone is not shown in the northwestern part of the nappe, and Upper Ordovician Womble Shale crops out along the southwestern margin of the nappe (Plate 4.5). Tightly folded, southwest-plunging synclines of Upper Ordovician Bigfork Chert/Polk Creek Shale through Devonian-lower Mississippian Arkansas Novaculite, exposed within the Zig Zag Mountains, mark the southwestern boundary of the Alum Fork nappe and the northeastern margin of the Mazarn syncline (Plate 4.5).

Womble Shale and overlying Bigfork Chert which crop out within the north-central and northeastern parts of the nappe are above a folded detachment fault rooted in the Womble Shale (Plate 4.5). In general, a north-vergent thrust fault (Fault M) separates the north-central, and part of the northeastern, Alum Fork nappe where Womble Shale and Bigfork Chert are exposed, from the rest of the Alum Fork nappe on the south, where Mazarn Shale is the predominant exposed formation (Plate 4.5). South-vergent faults, which may be backthrusts or overturned thrust faults, cross-cut the southern part of the nappe (south of Fault M as described above) and most appear to be truncated southwestward by a northwest-striking detachment fault at the base of the Womble Shale (Plate 4.5). It is unclear whether a detachment at Womble level extends farther southwest into the subsurface of the Mazarn syncline. The Mazarn Shale northeast of the Womble detachment apparently is exposed in the core of an arch where the overlying Womble and younger strata are eroded (Plate 4.5).

sections of Viele (1966) which suggest that overturned and upside-down Womble Shale comprises much of the surface exposure of the southeastern part of the nappe. Viele (1966) proposed that these overturned Womble beds rest in the core of a large recumbent nappe fold with a fold nose that verges toward the north. Although this model cannot be refuted at this stage, map patterns suggest another interpretation.

Even if the Viele (1966) interpretation is correct (and Womble Shale crops out across most of the southern part of the Alum Fork nappe), geometry of faults suggests northwest-directed transport of the Ouachita allochthon. In the southeastern part of the Alum Fork nappe, surface traces of thrust faults bend abruptly from northeast-striking, in the west, to southeast-striking, in the east (Plate 4.5) (Arbenz, 1989b, Haley and others, 1993). Therefore, northeast trending cross sections, such as in Viele (1966, and 1989), which are perpendicular to strike in the northeastern part of the Alum Fork nappe are parallel to strike farther south where the interpreted Womble beds are upside down. Therefore, fold axes in this location are likely parallel with the plane of the cross section, which in turn indicates the larger scale fold nose points either to the northwest or southeast, and not to the northeast as suggested in Viele (1966 and 1989). Small-scale folds of the Mazarn Shale and Blakely Sandstone east of the Alum Fork thrust fault trend northeast, which indicates northwest-directed translation of the Alum Fork nappe strata within the central axis of the nappe (Plate 4.5). Finally, the folds which delineate the shape of the Zig Zag Mountains (those which mark the southwestern boundary of the Alum Fork nappe) trend northeast and are overturned toward the southeast (Plate 4.5).

The upside-down Womble beds in Viele (1966) which are said to crop out in the southeastern part of the Alum Fork nappe may belong to numerous southeast-overturned (recumbent) folds which are parallel with the faults that cross cut the area (Arbenz, 1989b; Haley and others, 1993). In areas farther to the west, northwest, and northeast, folds are more upright, and in general are overturned to the southwest or southeast. Although overturning is toward the southeast or southwest, fold trends that are perpendicular to the Alum Fork thrust fault on the eastern side of the Alum Fork
small-scale recess indicate northwest-directed transport of Alum Fork strata (Plate 4.5).

If the Haley and Stone (1990, 1991, 1993, and 1994) interpretation is correct, and Mazarn Shale crops out across most of the southern part of the Alum Fork nappe, structural features still suggest northwest-directed translation of the allochthon in this part of the Central Ouachita thrust belt (Figure 4.2, and Plates 4.1 and 4.5). The upside-down beds in the southeastern part of the Alum Fork nappe previously classified as Upper Ordovician Womble Shale (Viele, 1966; Nielsen and others, 1989; Haley and others, 1993) would now be classified as Middle Ordovician Mazarn Shale. According to strike and dip information derived from unpublished geologic quadrangle maps of Haley and Stone (1990, 1991, 1993, and 1994), which span the southeastern part of the Alum Fork nappe, upside-down beds and those with low dip angles (<15°) are not common. The upside-down beds of Viele (1966) may be local features that belong to small scale folds. According to Haley and Stone (1994) the predominant dip of Mazarn Shale strata of the southern part of the Alum Fork nappe change from northwest towards the south and southwest, to north-east farther east. The Womble Shale and younger strata located in the northern part and southwestern part of the Alum Fork nappe overlie a detachment surface that is folded above underlying deformed Mazarn Shale (Plate 4.5).

**Southern Ouachita thrust belt**

The Southern Ouachita thrust belt is the smallest structural subdivision of the Ouachita Mountains (Figure 4.2 and Plate 4.1). The southeastern limb of the Bethel syncline marks the western boundary (Figure 4.2 and Plate 4.1). The southern flank of the Cross Mountains anticlinorium, the southeastern end of the Cossotot Mountains, and the Trap Mountains mark the northern boundary (Figure 4.2 and Plate 4.1). The surface trace of the Mesozoic unconformity marks the southern and eastern boundary (Figure 4.2 and Plate 4.1).

**Broken Bow uplift**

The largest structural feature of the western part of the Southern Ouachita thrust belt is the Broken Bow uplift (Figure 4.2 and Plate 4.1). Tightly folded and
faulted Upper Ordovician Bigfork Chert through lower Mississippian Arkansas Novaculite strata outline the shape of the Broken Bow uplift (Figure 3.2 and Plate 4.1). The predominant trend of folds changes abruptly across the Broken Bow uplift. Along the northwestern flank of the Broken Bow uplift, folds are mostly northeast-trending and overturned towards the southeast (Figure 4.2, Plate 4.1, and cross section ARK4 in Plate 4.3). In the northern part, folds are very tight, east-trending, and near-vertical toward the north; south-overturning increases progressively toward the south (cross section D-D’ in Plate 4.2 and cross section ARK4 in Plate 4.3). Along the eastern flank of the uplift, folds are east-trending and change from south-overturned on the north, to upright and north-tilted farther south (cross section D-D’ in Plate 4.2 and cross section ARK4 in Plate 4.3).

In general, the Broken Bow uplift is a northeast-plunging arch of pre-Meramecian (pre-Stanley Group) strata subdivided into two structural domains by cross-cutting faults (Figure 3.2). The large southwestern part of the uplift, the Hochatown dome, contains exposures of Cambrian-Ordovician (Collier-Womble) formations flanked by tight folds of Upper Ordovician-lower Mississippian (Bigfork Chert-Arkansas Novaculite) formations (Figure 3.2). North of the Hochatown dome, the Crater Mountain anticlinorium is an area of tightly folded Upper Ordovician-lower Mississippian (Womble-Arkansas Novaculite) formations cross-cut by numerous east-striking faults (Figure 3.2). A generalized cross section of the Carter Mountain anticlinorium shows progressive south-overturning of Arkansas Novaculite from north to south across the anticlinorium (Figure 3.2).

The Linson Creek synclinorium separates the Carter Mountain anticlinorium on the south from the Cross Mountains anticlinorium to the north (Figure 3.2). The central axis of the Linson Creek synclinorium straddles the boundary between the transitional zone of the Central Ouachita thrust belt and the western part of the Southern Ouachita thrust belt (Figure 3.2). Tight, east-trending folds of Upper Ordovician Bigfork Chert through Devonian-lower Mississippian Arkansas Novaculite crop out in the core of the Linson Creek synclinorium.
**Athens Plateau**

East of the Broken Bow uplift within the Southern Ouachita thrust belt is an area generally classified as the Athens Plateau (Figure 4.2 and Plate 4.1). The Athens Plateau is an area of east-northeast-striking Morrowan-Atokan deep-water turbidites that crop out along the southern boundary of the Ouachita Mountains (Plate 4.1). In general, the Athens Plateau is a homocline where all formations dip southward. Regional cross sections show that surface dips steepen towards the west and north within the Athens Plateau (cross section X-X’ in Plate 4.2 and cross sections ARK2 and ARK3 in Plate 4.3) (Blythe and others, 1988). Cross sections also show that the formations exposed in the Athens Plateau continue and dip less steeply beneath the Mesozoic Gulf Coastal Plain (Blythe and others 1988; Viele, 1989) (cross section C-C’ in Plate 4.2). The northward steepening of dips coincides with the subsurface location of a basal detachment thrust ramp located at the Paleozoic shelf-slope transition (Lillie and others, 1983; Mickus and Keller, 1992).

The stiff-layer formations of the Athens Plateau (upper Stanley Formation-Atoka Formation) are cross-cut by a series of near vertical transverse faults. Most of these are north-northeast-striking faults that cut the Jackfork Sandstone south of the Cowhide fault (Plate 4.1). These faults appear to be strike-slip tear-faults or oblique-normal faults (Walthal, 1967). Regardless of type, cumulative displacement along these faults is small. These short transverse faults have the same strike as a number of small transverse faults that cross cut Jackfork Sandstone of the northern limb of the Lynn Mountain syncline to the south of the Potato Hills (Plate 4.1).

Another larger transverse fault visible in the Athens Plateau is the northwest-striking Amity fault (Plate 4.1) (Walthal, 1967). The Amity fault appears to have much greater displacement than the smaller transverse faults located farther west. The northern part of the Amity fault separates tightly folded Mississippian Stanley Formation on the east from Morrowan-Atokan formations on the west (Plate 4.1). The southern end of the fault is not clearly defined and either intersects with (or cross cut with minimal displacement) or is over thrust by the Cowhide fault (Plate 4.1). The northern part of the Amity fault appears to be an oblique fault (tear-fault?) with down-
to-west offset (Walthal, 1967) (Plate 4.1). The Amity fault is parallel with northwest-striking transverse faults that offset Jackfork Sandstone strata in parts of the western Ouachita Mountains southwest of the Potato Hills (Plate 4.1).
Figure 4.1: Generalized geologic map of the Ouachita salient and surrounding areas showing location of major tectonic features. Black lines with bars denote thrust faults which delineate the general strike of the Ouachita thrust belt. Black lines with ticks are normal faults with ticks on down-thrown side.

References: Morris, 1974; Johnson and others, 1989; Nicholas, 1989; Thomas and others, 1989.
Figure 4.2: Geologic and structure map of the Ouachita Mountains and adjacent areas of Oklahoma and Arkansas. The Ouachita Mountains are subdivided into three thrust belts: the Frontal Ouachita (FTB); the Central Ouachita (CTB), and the Southern Ouachita (STB). Because of along-strike variations of structure within these thrust belts, each thrust belt is further subdivided into an eastern (e) and a western (w) part separated by a transitional zone (t). A wide area of Mississippian (Stanley Group) separates folds of vastly different wavelengths on opposite sides of a wavelength transition zone (see above). See Figure 4.7 for cross sections E-E', D-D', X-X', and C-C'.
Figure 4.3: Summary of the mechanical stratigraphy of the deep-water facies rocks of the Ouachita Mountains. Shown schematically are variations in observed fold wavelengths for stiff-layer units. Major detachment horizons (or at least disharmonic boundaries) are located within the Womble Shale and lower Stanley Group (shown as thick dashed lines). These are horizons where fold wavelengths change abruptly as a result of fault motion (detachment) or fold dynamics without large displacement (disharmonic boundary) (Arbenz, 1989d). Other detachment/disharmonic boundaries are shown as thinner dashed lines.

The sandstone-dominated upper Stanley Group-lower Atoka Formation strata form the dominant stiff layer. The Collier Formation-lower Stanley Group form a composite weak layer. Deformation of the middle Stanley Group compensates for difference in fold wavelengths between the upper stiff layer and the lower weak layer.

Figure 4.4: Cross sections of the eastern and central part of the Arkoma basin and Ouachita thrust front. The Mississippian-Morrowan succession north of the Ouachita thrust front is thicker farther east (cross section A-A'). A laterally continuous unconformity surface at the base of the Morrowan, which evidently forms a detachment, is another characteristic of cross section A-A'. As illustrated in the cross sections, along strike to the west of cross section A-A', in the central part of the Arkoma basin, the lower Atoka is much thicker than farther east; however, the middle and upper Atoka are much thinner. The thick succession of Atoka and older strata within the hanging-wall of the Ross Creek/Cadron fault suggests that a large-offset growth fault is located in the subsurface south of the Ross Creek/Cadron fault.

See Figure 4.2 for locations of cross sections. References listed above.
Figure 4.5: Two cross sections of the western Arkoma basin and Ouachita thrust front of Oklahoma and Arkansas. Cross sections compare the timing of offsets of the basement fault beneath the Sans Bois syncline (Sans Bois fault) of the central part of the Arkoma basin, to that of the Bengalia fault located beneath the westernmost part of the Frontal Ouachita thrust belt near Black Knob Ridge.

In cross section D-D', deep-water facies Atoka Forination thickens abruptly south of the Sans Bois basement fault. The combination of increased thickness and deep-water facies suggests that the Sans Bois fault initiated fault motion prior to rapid deposition of lower Atoka Formation. In contrast, in cross section E-E'', deep-water facies lower Atoka Formation is absent above the Bengalia fault. Here restoration of the thrust fault below "A" indicates abrupt thickening of Mississippian-Morrowan strata to the south. This suggests fault motion began before the Atokan.

See Figure 4.2 and Plate 4.1 for location of the Sans Bois syncline and Black Knob Ridge. The Bengalia fault is located in the subsurface beneath the surface trace of the Windingstar fault. See Figure 4.2 for location of end points of each cross section.
Figure 4.6: Map showing correlation between sharp bend in Ouachita thrust belt and location of thick lower Atokan strata. Note the abrupt thickening of lower Atokan moderate- to deep-water facies strata south of the Sans Bois fault. The strike of the frontal thrust fault of the Ouachita thrust belt (Chf.-DCF.-RCF.) and thrust faults farther south bend sharply south of the area of thickest Atokan deposition. Lower Atokan strata thin abruptly along strike of the Sans Bois fault towards the Bengalia fault (near Black Knob Ridge).

Figure 4.7: Schematic diagrams showing three possible fault geometries for the northern flank of the Benton uplift. All models explain the observed south-vergent folds common to the northern Benton uplift. Model A requires the exposed rocks of the northern Benton uplift (at position X) be above the north-dipping limb of a folded thrust fault. South-overturned folds form either in response to backthrust to the north, or in response to uplift of basement. Model B requires strata at position X to be above a backfolded, south-overturned thrust fault. Back-folding is in response to a backthrust farther north. Model C requires the strata at position X be on either side of a south-vergent backthrust.

Because fault dips are uncertain throughout most of the northern Benton uplift, study of stratigraphy adjacent to the fault zone is necessary to decide which of the above models best fits surface geology. Because older strata tend to be on the northern side of most faults along the northern flank of the Benton uplift, Models A and C (or a combination) are more appropriate than model B. Overturned thrust faults, such as in Model B, may be a smaller scale feature of local importance.

Figure 4.8: Cross sections of part of the western Frontal Ouachita thrust belt and Arkoma basin showing major detachments. Cross sections show stratigraphic levels of four detachments. Two of these, the late Devonian-Mississippian detachment (Woodford/Caney shales), and the Morrowan detachment (Springer/basal Wapanucka), are northwest-vergent detachments. Depending on interpretation, in the vicinity of the Kiowa syncline, the Lower Atokan detachment is everywhere a south-vergent delamination surface (A above), or north-vergent southeast of the Kiowa syncline and south-vergent northwest of the syncline (B above).

See Figure 4.2 for location of cross sections.
Figure 4.9: Regional stratigraphy of the western Ouachita Mountains and western Arkoma basin, southeastern Oklahoma. Horizontal line pattern shows strata absent because of erosion or non-deposition. Area above diagonal line pattern represent strata in the hanging wall of selected thrust faults. Diagram identifies several regional detachment levels and shows a general northwestward stratigraphic up-cutting in detachment level.

Figure 4.10: Geologic/structure map of part of the central Ouachita Mountains and adjacent Arkoma basin showing fault-displacement transfer zone north of the Black Fork syncline, Oklahoma-Arkansas border. North of the Black Fork syncline, the northeasterly displacement along the Choctaw fault decreases eastward toward "A" where the fault becomes a blind thrust. South of the Choctaw fault, north-directed displacement increases along the Dutch Creek-Ross Creek fault. The Dutch Creek-Ross Creek fault is not clearly defined west of "B," but may merge with another fault farther west in the western part of the Frontal Ouachita thrust belt.

The map is an expanded view of part of Figure 4.2. See Figure 4.2 for a regional perspective. References: Walthal, 1967; Blythe and others, 1988; Arbenz, 1989b,d; Viele and Thomas, 1989; Haley and others, 1993.
Figure 4.11: Bouguer gravity map of the western Ouachita Mountains. Superimposed are some residual gravity high areas in the core of the Ouachita Mountains and along the Ouachita thrust front (dark gray shading). Several large scale regional structures are shown (see Figures 4.1 and 4.2). Thick dashed lines denote several northwest-trending lineations (lineaments). Modified from Gatewood and Fay (1991).
Figure 4.12: Example of fold disharmony in the central Ouachita Mountains between the eastern end of the Lynn Mountain syncline and the northwestern end of the Cossatot Mountains anticlinorium (CMA), Oklahoma-Arkansas border. Location of the area of fold disharmony is shown on the geologic/structure map of part of the central Ouachita Mountains and adjacent Arkoma basin shown at the top of the page (reduced size version of Figure 4.10).

Map A shows the apparent mismatch of fold wavelengths between the northwestern end of the Cossatot Mountains anticlinorium (CMA) and the stratigraphically higher Lynn Mountain syncline along strike to the west. Map B shows an alternative interpretation where the folds on the northern flank of the CMA continue westward as a tight anticline on the southern flank of the Lynn Mountain syncline. A synform to the north of the CMA continues westward as the Lynn Mountain syncline.

Figure 4.13: Geologic/structure map of the eastern part of the Benton uplift, Arkansas, showing location of the Jessieville (JN), Alum Fork, and Paron nappes (eastern nappes). Also shown are the locations of two sharp small scale bends in regional strike on the northwestern side of the Alum Fork nappe identified as the Alum Fork recess and the Little Rock salient. The sharply curved Alum Fork thrust fault separates Lower Ordovician Mazam shale on the east from Upper Ordovician-lower Mississippian (Bigfork Chert–Stanley Formation) on the west.

The map shown above is an expanded view of part of Figure 4.2. See Figure 4.2 for a regional perspective.

A) "Break forward" imbricate thrust belt

B) Example of a "break-back" (out-of-sequence) thrust fault

Figure 4.14: Schematic diagram of an out of sequence break-back fault. Part A compares the geometry of a set of break-forward imbricate thrust faults to a break-back thrust fault shown in B. The break-back fault is younger than the imbricate faults underneath. Evidence for a break-back (out-of-sequence) fault such as shown in part B is deeper-water facies strata for a given formation at position X rather than at position Y.

See Figure 4.13 and Plate 4.5 for location of Panther Creek fault and Plate 4.5 for location of Womble detachment.
Chapter Five

Sub-Mesozoic structural cross sections showing intersection of the autochthonous structures of the Arbuckle uplift, Ardmore basin, and Muenster arch with the allochthonous structures of the Ouachita thrust belt in southeastern Oklahoma and northern Texas

Location of study area

The study area for which seismic reflection profiles have been studied for this dissertation is located in southeastern Oklahoma and northeastern Texas. The study area includes in the autochthon (all buried beneath Mesozoic strata): the southeastern part of the Arbuckle uplift (Tishomingo and Belton anticlines, Coleman half-grabens, Cumberland-Ravia-Sand Canyon nappes), the southeastern part of the Ardmore basin, the southeastern projection of the Criner Hills uplift, and the southeastern end of the Muenster arch (Figure 4.1 and Plate 4.1). The study area in the allochthon includes the exposed southwestern edge of the Ouachita Mountains, and extends in the subsurface farther south and includes the Bryan salient and areas southeast of the Muenster arch (Figure 4.1 and Plate 4.1)

Autochthon

The southeastern part of the Arbuckle uplift is subdivided into several structural terrains (Tishomingo and Belton anticlines, Coleman half-grabens, Cumberland-Ravia-Sand Canyon nappes) (Plates 3.9 and 4.1). Combining the present outcrop and subsurface extent, the southeastern part of the Arbuckle uplift consists of a core of Precambrian and Cambrian igneous rocks surrounded by a thick succession of Cambrian to Ordovician carbonates (with a basal transgressive sandstone at the base and increased proportion of sandstone and shale in the Middle Ordovician part). These thick carbonates, sandstones and shales are flanked by relatively thin Late Ordovician to Mississippian carbonates, shales, sandstones, and cherty shales and limestones.
Most of the southeastern Arbuckle uplift is a broad arch bounded on the north by the Reagan-Sulfur faults and on the south by the Washita Valley (Plates 3.9 and 4.1). This broad arch includes the Tishomingo and Belton anticlines and the three Coleman half-grabens (Plate 4.1). The western part of the near vertical Sulfur fault separates the Belton anticline on the south, from the Hunton arch on the north (Plate 3.9). The southeastern part of the Sulfur fault separates the Belton anticline on the south from the Wapanucka syncline on the north, which the westernmost extent of the Arkoma foreland basin (Plates 3.9 and 4.1). The steep, northeast-dipping, southwest-vergent Washita Valley fault separates the Tishomingo anticline on the northeast from a set of relatively narrow fault blocks collectively known as the Cumberland, Ravia, and Sand Canyon nappes (Plate 4.1). South of the Cumberland, Ravia, and Sand Canyon nappes is the deep Ardmore basin and overlying allochthonous strata in the Bryan small-scale salient (BSSS) (Plates 3.9 and 4.1). Finally, the southeastern end of the Tishomingo and Belton anticlines plunge beneath the Ouachita thrust belt.

North of the Tishomingo and Belton anticlines, the Hunton arch is a relatively flat dome capped by Ordovician Arbuckle Group carbonates; younger formations are concentrated in narrow, graben blocks along the eastern and western flanks. The eastern flank of the Hunton arch drops abruptly along normal faults into the western end of the Arkoma foreland basin. Two such normal fault zones are the Clarita and Olney fault zones (Plate 3.9).

The Ardmore basin separates the structurally high Tishomingo and Belton anticlines of southeastern Oklahoma from the Muenster arch of northeast Texas. Within the study area, the Criner Hills uplift loses amplitude southeastward along strike and merges with the Sherman fault block (Plates 3.4 and 4.1) (Hardie, 1990; Ewing, 1991; proprietary seismic reflection profiles). The Sherman fault loses displacement to the northeast where the Sherman block merges with the much larger Muenster arch (Plate 4.1) (Ewing, 1991, Bradfield 1957a-c). The Muenster arch plunges southeastward beneath the Ouachita thrust belt (Plate 4.1).
Allochthon

Average strike of the frontal fault of the Ouachita orogen within the primary study area is northeast. The frontal fault deflects sharply eastward around the Tishomingo-Belton anticline (Plate 4.1). In the autochthon, the eastern part of the Tishomingo-Belton anticline includes three structures known as the Coleman half-grabens (numbered 1, 2, and 3 in Plate 4.1). Note that Coleman half-graben #1 (Plate 4.1) is the same structure that is shown as the Coleman syncline (name given in Ham and others, 1954) in Plate 3.9. The easternmost of the Coleman half-grabens extends beneath the Ouachita allochthon (Plate 4.1). The Ouachita thrust front is bent sharply on a smaller scale by several basement faults on the southeastern margin of the Tishomingo-Belton anticline and includes a westward bend into Coleman half-graben #3 (Plate 4.1) (Huffman and others, 1987).

South of the Tishomingo-Belton anticline on the northern end of the Bryan small-scale salient (BSSS), the Ouachita thrust front bends abruptly westward and is parallel with regional basement faults which extend northwest of the Ouachita allochthon (Plate 4.1). The N-15°E-striking Kingston fault that marks the northwestern end of the Bryan small-scale salient is nearly vertical or overturned towards the east (Plate 4.1) (Huffman and others, 1978; Hardie, 1990). The Kingston fault separates imbricate thrust slices of Ordovician-lower Mississippian deep-water shales and cherts (Womble Shale-Bigfork Chert-Arkansas Novaculite-Stanley Group) on the east from a triangle zone of uppermost Mississippian- coarser grained shallow-water facies clastic rocks (Springer-Dornick Hills-Deese Groups) (Huffman and others, 1978; Hardie, 1990).

In the subsurface beneath Mesozoic cover, allochthonous strata in the northwestern end of the Bryan small-scale salient are tightly folded above underling autochthonous strata within the axis of the Ardmore basin between the northwest-striking Criner Hills and Bryan basement fault zones (Plate 4.1). The Bryan fault on the north separates the abruptly deepening central part of the Ardmore basin from a series of narrow, northwest-trending fault-bounded folds known as the Madyll-
Aylesworth flexure (Plate 4.1) (Huffman, 1978). The Criner Hills fault zone marks the southern boundary of the narrow, northwestern end of the Bryan small-scale salient (Plate 4.1). The frontal fault of the Ouachita allochthon, marking the southwestern margin of the Bryan small-scale salient, bends sharply toward the south around several narrow structures in the underlying autochthon. On a larger scale, the frontal fault bends sharply southward and southwestward around the southeast-plunging nose of the Muenster arch (Plate 4.1).

**Discussion of cross sections**

In the following paragraphs, the cross sections are described in numerical order according to the number scheme shown on Plate 4.1. The cross sections are grouped into two grids. The northern grid (containing cross sections 1 through 5) is along the northeastern flank of the Tishomingo-Belton anticline (Plate 4.1). The seismic reflection profiles used as the template for these five cross sections were shot in 1993 and were contributed by Richardson Seismic. The datum level for cross sections 1 through 5 is 600 ft (182.4 m) above sea level. The southern grid (containing cross sections 6 through 9) is located southwest of the Tishomingo-Belton anticline (Plate 4.1). Cross sections 6 through 8 are located in the Bryan small-scale salient, and cross section 9 extends southeast from the southeastern end of the Muenster arch (Plate 4.1). The seismic reflection profiles used as the template for cross sections 6 through 8 are older than the lines in the northern grid and were contributed by Digicon. Seismic data for cross section 9 were contributed by Delta Exploration. Summaries of interval seismic velocities used to calculate formation thickness for all the structural cross sections discussed in this chapter are listed in Table 5.1.

Except for the eastern half of cross section 1 and the northern ends of cross sections 4 and 5 (which intersect cross section 1), the area encompassed by all nine cross sections is covered by Mesozoic strata of the Gulf Coastal Plain. For areas where Paleozoic strata are exposed, geologic maps of the region were used to located seismic reflectors that correlate to mapped formation contacts (Miser, 1954; Ham and others, 1954; Arbenz, 1989b; Hardie, 1990). Published regional cross sections (Ham and others, 1954; Arbenz, 1989a; Hardie, 1990), and unpublished structural cross
sections provided by J. Kaspar Arbenz were also used as further constraints. In areas where Paleozoic strata are completely buried by Mesozoic strata, constraint of structural interpretation consists of formation-top data for selected deep wells supplied by the Oklahoma Geological Survey (Table 5.2), formation-top data for the Sohio Natural Resources No. 1 Taylor well supplied by Neil Suneson of the Oklahoma Geological Survey, structure contour maps of the top of the Arbuckle/Ellenburger Group beneath frontal parts of the western Ouachita thrust belt and adjacent foreland (Bartram and others, 1950; Ewing, 1991; Gatewood and Fay, 1991), and pre-Mesozoic subcrop maps (Ham and others, 1964; Huffman and others, 1978; Huffman and others, 1987; and Hardie, 1990). Another important source of information was Flawn and others (1961), which includes well data and pre-Mesozoic subcrop maps. Well data and structural cross sections of Huffman and others (1978; 1987) helped greatly to constrain cross sections 6 to 8 (Plate 4.1). Bradfield (1957a-c) was used as the primary constraint on Paleozoic stratigraphy for the Muenster arch (cross section 9), specifically that part surrounding the Sherman fault block (see Plate 4.1). The Bradfield (1957a,b) publications also supplied interval seismic velocities for several Paleozoic formations (Middle Ordovician McLish and Tulip Creek Formations, Morrowan/Atokan Dornick Hills Group, and Atokan/Desmoinesian Strawn Group). These velocities were used as initial estimates for these strata during construction of cross sections. Cross section 9 is also constrained by cross sections and a detailed description of the Mesozoic succession by Wood and Guevara (1981). Cross section 9 contains the thickest Mesozoic succession of all the cross sections constructed for this dissertation, and formation contacts within the Mesozoic succession are clearly imaged.

The following discussion of cross sections 1 through 9 includes key features of each cross section and identifies key seismic reflectors. For each cross section, rocks of the autochthon beneath the basal detachment fault of the overthrust allochthon are discussed first, followed by a discussion of the allochthon. Regional stratigraphy and structure of the Ouachita salient and adjacent foreland are summarized in chapters 2 and 3, and the structural complexity of the Ouachita Mountains is discussed in
chapter 4. This chapter examines the interaction of the Ouachita thrust belt and the
Arbuckle and Muenster uplifts (Plate 4.1).

Northern seismic cross section grid

Cross section 1

Overview/location

Cross section 1 (plate 4.1) is an east-west oriented interpreted structural cross
section that extends from Atoka County to western Pushmataha County in
southeastern Oklahoma. Although not illustrated because the base of the Mesozoic is
above seismic datum level, most of the western half is covered by a thin veneer of
Mesozoic Gulf Coastal Plain strata. Cross section 1 is oblique to trend of the Ouachita
thrust front and the foreland structures and cannot be strictly kinematically balanced.
The eastern end of the cross section nearly intersects the published cross section E-E’
of Arbenz (1989a) just east of the Jumbo anticline (Figure 4.2 and Plate 4.1). Because
cross section 1 has the thinnest Mesozoic cover, the datum-level geology is the best
constrained by surface geology of all the cross sections discussed in this chapter.

Autochthon

The interpretation shown for the Ouachita thrust front in the western part of
cross section 1 is based upon the clearest seismic reflectors visible on the entire
seismic reflection profile. Clearly defined gentle east-dipping autochthonous strata are
truncated abruptly at a more steeply east-dipping fault surface. The most clearly
identified reflective surfaces in the western part of cross section 1 are the Wapanucka-
Spiro formations (limestone and calcareous sandstone) at the base of the Atoka
Formation, the Bromide-Viola Formations (dense limestones), and the top of the
Arbuckle Group (dense dolostone) (Plate 5.1). The north-vergent, south-dipping
Choctaw thrust fault separates contorted reflectors containing uncertain formations in
the hanging wall above the gently east-dipping layered reflectors of the autochthonous
Atoka Formation. This frontal thrust fault is the subsurface projection beneath the
Coastal Plain of the Choctaw fault which is exposed farther to the north (Plate 4.1) and
is the frontal fault (emergent base) of the Ouachita allochthon. East of the Choctaw
fault, the basal décollement of the allochthon abruptly deepens and over-thrusts
progressively older strata and intersects another thrust fault. This steeper fault is along strike from the Ti Valley fault that is exposed farther northeast (Plate 4.1). In this location, all seismic reflectors corresponding to formations above the Arbuckle Group to lower Atoka Formation (lower Ordovician to lower Pennsylvanian) terminate abruptly eastward against the basal décollement.

Footwall geometry east of the Ti Valley fault beneath the basal décollement of the Ouachita thrust belt is not clearly defined. The bottom of the Arbuckle Group appears to be in normal offset at a possible fault zone that corresponds geographically with part of the Bengalia fault of Gatewood and Fay (1991); therefore, the fault is labeled Bengalia on cross section 1 (Plate 5.1). However, no offset is visible for the upper levels. The upper Cambrian-Lower Ordovician Arbuckle Group reflectors so clearly visible on the western end of cross section 1 terminate abruptly beneath the basal décollement at the root of the Ti Valley fault (Plate 5.1). Faint east-dipping layered seismic reflectors are discernible for a distance of 10,000 ft (3040 m) east of the root of the Ti Valley fault.

East of the Bengalia fault, the only clear set of seismic reflectors within the autochthon is a roughly horizontal, 200 millisecond thick (2-way travel time) set of reflectors at a two-way travel time depth of 3 seconds. Westward projection of this set of reflectors places the base at the base of the Bromide Formation (upper Ordovician upper Simpson Group) and the top at the top of the Sycamore Formation (lower Mississippian). Because of the clarity of the set of reflectors, it was assigned a seismic velocity appropriate to dense carbonate rock (22,500 ft/sec). Dense chert also has a high seismic velocity (approximately 20,000 ft/sec). Although the set of reflectors is clearly defined at the top and bottom, no clear horizons are recognized within the 200 millisecond thick succession, suggesting that either the entire succession has the same density and a consistent lithology, or there is a loss of seismic resolution. The lack of clear internal reflectors within the 200 millisecond thick succession east of the Bengalia fault contrasts with many clearly defined reflectors within correlative Bromide-Woodford-Sycamore succession west of the Bengalia fault. Because this
difference, the 200 millisecond set of reflectors east of the Bengalia fault is labeled “Bromide-Sycamore” in the cross section (Plate 5.1).

East of the Bengalia fault, the only reflector beneath the “Bromide- Sycamore” reflector is a much less clearly defined reflector 400 milliseconds (2-way travel time) deeper. Westward projection of this reflector places it at the top of Cambrian-Ordovician Arbuckle Group carbonate succession. For this reason, this reflector is considered to be the eastward continuation of the top of Arbuckle Group. The Simpson Group strata between the top of the Arbuckle Group and base of the upper Simpson Bromide Formation thicken eastward in cross section 1. The composite thickness of the Joins-Oil Creek-McLish Formations increases from approximately 100 ft (30.4 m) thick at the western end of cross section 1 to 2500 ft (760 m) in the central and eastern parts. A seismic velocity of 12,500 ft/sec, and thickness of 2500 ft for the pre-“Bromide- Sycamore” strata east of the Bengalia fault combined with estimated seismic velocities and thicknesses for all overlying strata (Table 5.1) produce a depth to top of Arbuckle Group consistent with published structure contour maps (Gatewood and Fay, 1991). The seismic velocity of 12,500 ft/sec is slightly greater than the 12,300 ft/sec used for the Oil Creek Shale-Tulip Creek-McLish part, but less than the 14,500 ft/sec used for the Joins-basal Oil Creek sandstone west of the Bengalia fault (Table 5.1, and Plate 5.1).

The maximum estimated thickness of the entire Simpson Group in cross section 1 increases eastward from greater than 2000 ft (608 m) (Suhm, 1997, Ham, 1973) west of the Ouachita thrust front to approximately 2600 ft (790 m) east of the Bengalia fault. Maximum thickness estimates for the Simpson Group. The Simpson Group is a Middle to Upper Ordovician succession subdivided into several formations which are in ascending order Joins Formation limestone-basal Oil Creek sandstone, Oil Creek, McLish, Tulip Creek, and Bromide limestone. The Joins is a limestone-clast conglomerate at its base and grades upward and laterally into the basal Oil Creek Sandstone (Suhm, 1997). The Oil Creek, McLish, and Tulip Creek, vary laterally and vertically between quartz sandstones, shales, and limestones, and the Bromide is primarily a limestone (Suhm, 1997). The Simpson Group thickens abruptly towards
the Ardmore basin (Plates 2.4 and 4.1) and has proportionally greater shale content (Suhm, 1997). The thickening of the Joins-Oil Creek-McLish Formations of the Simpson Group shown in cross section 1 may also coincide with a similar increase in shale content. Increased shale content and absence of thick interbeds of limestone and sandstone may partly explain the eastward change in seismic expression in the pre-Bromide Simpson strata of the autochthonous terrain (east of the Bengalia fault) in cross section 1. Increased depth and seismic signal attenuation caused by overlying strata may also cause a loss of clear reflectors or improper processing of seismic data.

**Possible facies transition in upper Simpson-Woodford-Sycamore succession**

The lithology of the “Bromide-Sycamore” reflector interval is unknown east of the Bengalia fault in cross section 1. Traditionally, large-scale regional cross sections lump all the upper Cambrian to lower Mississippian sedimentary succession in the deeply buried footwall east of cross section 1 and west of the Broken Bow uplift (Figure 4.2 and Plate 4.1) into “shallow-water passive-margin shelf carbonate” facies (for example, Arbenz, 1989a-d). However, more detailed stratigraphic examinations indicate laterally variable lithologies (and possible facies) above the Cambrian-lower Ordovician Arbuckle Group (Denison, 1997; Finney, 1997; Suhm, 1997).

The limestones in the Simpson Group range from shallow-water limestones in the Hunton arch to moderately deep-water lime mudstones in the Ardmore basin (Figure 4.2 and Plate 4.1). Similarly to the Simpson Group limestones, Viola Formation limestones exhibit laterally and vertically variable facies and grade from shallow-water skeletal limestones on the Hunton arch, to moderate- to deep-water lime mudstones in the Ardmore basin (Figure 4.2 and Plate 4.1). These lime-mudstones are dark grey, are rich in organic matter, and contain graptolites (Finney, 1997).

The distribution of deep-water facies limestone indicate that a Late Ordovician transgression of moderate- to deep-water facies rocks rimmed the southeastern part of the Arbuckle uplift and extended into the Ardmore basin which was an embayment (or relative topographic low) on the Late Cambrian-Early Ordovician passive-margin
shelf. This passive margin shelf extended in arcuate shape from the Black Warrior basin, in the east, westward to the Arbuckle uplift, southwestward to the east of the Llano uplift, and through northwest southwest jogs to the Marathon uplift of southwest Texas (Figure 4.1). It is not suggested that the shelf edge for the entire Cambrian-Ordovician carbonate succession is located close to the Arbuckle uplift, but rather that the older Arbuckle-Group and lower Simpson Group rocks represent more rapid subsidence in the Ardmore basin than in areas on the north and south (Denison, 1997; Finney, 1997; Suhm, 1997). Strata within the Ardmore basin and the Anadarko basin farther northwest (Figure 4.1) suggest deepening during the Mohawkian (Late Ordovician); whereas, strata within the foreland to the northeast and southwest indicate a eustatic low stand during the Mohawkian (Finney, 1997).

Several lines of evidence indicate that a facies transition in the Late Ordovician-Mississippian upper Simpson Group through Sycamore Formations is located east of the Bengalia fault in the autochthon and rims the southeastern nose of the Tishomingo-Belton anticline (Plate 4.1). First, the interpreted cross section 1 indicates a dramatic eastward thickening of the Simpson Group (Plate 5.1), suggesting subsidence and deepening east of the Arbuckle uplift during the Middle Ordovician. Second, isopach maps and upsection changes in lithology within the Simpson Group from coarser-grained limestone-clast conglomerates, quartz sandstones, and shales to shales and lime-mudstones suggest relative sea-level rise during the Middle Ordovician (Suhm, 1997). Third, relative deepening of the nearby Ardmore basin (Figure 4.2 and Plate 4.1) in the Late Ordovician is indicated by the graptolitic laminated lime-mudstones of the basal Viola Formation which overlie the Simpson Group (Finney, 1997). Fourth, the Sylvan shale above the Viola is a green-colored transitional facies shale (Ham, 1973; Morris, 1974). Fifth, the overlying Hunton Group-Woodford strata contain many cherty limestone beds (Ham, 1973). The “foreland-facies” stratigraphic succession indicates an overall upsection increase in chert and subsidence of the Paleozoic shelf including the Arbuckle uplift and surrounding regions (Figure 4.2).
The abrupt change in seismic character of the “Bromide- Sycamore” interval east of the Bengalia fault also suggests a facies transition. This interval which extends eastward beneath the basal décollement of the allochthon is similar in appearance (internally transparent, or no clear reflectors within the interval) and thickness to that of the deep-water facies chronostratigraphic equivalent Bigfork Chert-Arkansas Novaculite (Plate 2.2) strata in the allochthon. The “Bromide- Sycamore” interval in the autochthon and the Bigfork Chert-Arkansas Novaculite interval within the allochthon are much more “internally transparent” in appearance than the same interval west of the Bengalia fault. Because interval seismic velocities are similar (both more than 20,000 ft/sec), it may not be possible to discern from seismic data the difference between a reflector intervals composed of limestone, cherty limestone, or chert (also loss of seismic resolution with depth).

If the “Bromide- Sycamore” interval is indeed a cherty carbonate or chert, then a facies transition in the Upper Ordovician-lower Mississippian is located in the autochthon beneath the northwestern flank of the Round Prairie syncline (Plate 4.1). If this interpretation is correct, it places the facies transition beneath the western Ouachita thrust belt far west of the currently accepted location east of the Broken Bow uplift (Figure 4.2 and Plate 4.1) (for example, Arbenz, 1989a).

The “Bromide- Sycamore” interval and the underlying top of Arbuckle reflector end abruptly eastward in the eastern part of cross section 1, and the geometry of the autochthon is unconstrained farther east. A reason for this abrupt end in these reflectors is uncertain. This may be the location of a large-offset normal fault (the “detachment” fault shown in Gatewood and Fay, 1991) which drops the “Bromide-Sycamore” and other older autochthonous strata to great depths.

The abrupt end of the “Bromide- Sycamore” interval and underlying reflectors within the autochthon in cross section 1 (Plate 5.1) is near the southeastern end of the surface trace of the Octavia fault in the allochthon to the north (Plate 4.1). The position of the Octavia fault (Plate 4.1) marks the location where the footwall cutoff of the Ti Valley fault restores as estimated from regional cross sections constructed across the westernmost Ouachita Mountains (cross section E-E’ in Plate 4.2)
(modified from cross section E-E’ in Arbenz, 1989a). The abrupt end of the “Bromide-Sycamore” interval may represent the southwestward subsurface continuation of the Ti Valley fault footwall cutoff.

**Allochthon**

The allochthonous Ouachita thrust belt covers most of the autochthonous terrain in cross section 1. Structural geometry of the allochthon is constrained by formation-top data for 3 wells supplied by the Oklahoma Geological Survey (A4, A15, and PU8, Table 5.2), and formation tops for one well (Sohio Natural Resources No. 1 Taylor) provided by Neil Suneson of the Oklahoma Geological Survey. Because of the close proximity of well A4 and the Sohio Natural Resources No. 1 Taylor well, and the difference in information supplied for each well, a single composite interpretation is illustrated on the cross section. Surface structural geology from published maps (Arbenz, 1989b; Hardie, 1990) was also used as a constraint. The upper parts of the allochthon exhibit clearly visible layered seismic reflectors. The primary velocity contrasts in the layered strata are located at the base of the Pennsylvanian (Atokan) Atoka Formation and at the base of the Pennsylvanian (Morrowan) Jackfork Sandstone (see Table 5.1 for velocities).

The upper part of the Mississippian Stanley Group is shown by layered reflectors; however, the geometry of the lower part is less well defined. On the basis of thicknesses of Stanley Group derived from wells, the Stanley Group has a lower seismic velocity than the overlying Jackfork Sandstone. The determined velocity of 13,000 ft/sec is nearly the same as the 13,500 ft/sec determined for the shallow-water facies equivalent Caney-Springer formations (as calculated from thickness in well A4/Sohio Natural Resources No. 1 Taylor). A sharp velocity contrast marks the top of the Upper Ordovician-lower Mississippian Bigfork Chert-Arkansas Novaculite. A composite velocity of 18,560 ft/sec for the Bigfork Chert-Arkansas Novaculite yields a thickness equal to that from well data. The Bigfork Chert-Arkansas Novaculite interval is clearly expressed throughout most of the seismic profile.

An interesting structural characteristic of the allochthon shown in cross section 1 consists of triangular shaped zones at depth between overlying surface synclines.
(consisting of exposed interbedded sandstone and shale turbidites). The best example is the triangular shaped zone of contorted and discontinuous seismic reflectors is located in the footwall west of the Jumbo fault and east of an interpreted east-vergent backthrust (Plates 4.1 and 5.1). A tight anticline is formed by the triangular shaped zone. The triangle beneath the Jumbo fault is geometrically linked to the broad folds in the upper part of the allochthon. The seismic expression of structures in the upper part of the allochthon suggests folding by flexural slip, consistent with evenly interbedded sandstone and shale. To maintain constant thickness, the cores of anticlines are tighter with depth as shown beneath the Jumbo anticline. The tight Jumbo anticline is the “cusp” of a cuspate-lobate fold pair (tight anticline-broad syncline) (Arbenz, 1989b).

Another interesting feature is the difference between the wavelengths of the upper Stanley Group-Atoka Formation reflectors and those of the Bigfork Chert-Arkansas Novaculite. Although nearly in-phase throughout most of the cross section, the Bigfork Chert-Arkansas Novaculite is clearly out-of-phase with the overlying strata in the tight core of the Jumbo anticline. This mismatch of wavelength, amplitude, and phase is likely accommodated by a transition from flexural slip in the upper Stanley through Atoka strata to flexural flow in the more shale-rich lower parts of the Stanley Group.

Another interesting aspect of cross section 1 is the great variation in thickness of the Ordovician Womble Shale beneath the Bigfork Chert-Arkansas Novaculite interval. Thickness of the Womble Shale above the basal décollement and beneath the Bigfork Chert-Arkansas Novaculite ranges in cross section 1 from about 200 ft to more than 10,000 ft (which far exceeds estimates of maximum stratigraphic thickness of 3000 ft (for example, Gatewood and Fay, 1991). An overall eastward increase in thickness of the Womble Shale is explained by placing the basal décollement at progressively lower stratigraphic levels with distance east of the Ouachita frontal fault; however, variation within individual thrust slices is increased by flexural flow or fault imbrication of the Womble Shale, as suggested by lateral disruptions of seismic reflectors within the Womble interval. Map patterns of the Womble Shale through
Arkansas Novaculite (narrow fold limbs and thickened fold noses) in the westernmost Ouachita Mountains at Black Knob Ridge (Figure 4.2 and Plate 4.1) suggest flexural flow in the Womble Shale (Hendricks and others, 1937; Arbenz, 1989d; Hardie, 1990). The cause for the abrupt thickening of the Womble Shale in the western part of the cross section beneath the anticline east of the Round Prairie syncline (Plates 4.1 and 5.1) is uncertain. It may result from increased imbrication or flow in this location.

**Cross section 2**

**Overview/Location**

Cross section 2 is oriented east-west, extends from northeast Bryan County to western Choctaw County in southeastern Oklahoma (see Plate 4.1 for location). Cross section 2 is oblique to trends of autochthonsous and allochthonsous structures. For this reason, this cross section cannot be strictly kinematically balanced. The Mesozoic strata of the Gulf Coastal Plain cover the entire cross section. Cross section 2 is not constrained by directly available well data; however, is constrained by seismic profiles used for cross sections 4 and 5, which both intersect cross section 2, and by the seismic profile used to create cross section (which also intersects cross sections 4 and 5). Seismic reflectors not as clear as in cross section 1.

**Autochthon**

At the west end of cross section 2, Mesozoic strata of the Gulf Coastal Plain directly overlie Precambrian (or Early Cambrian) basement (Ham and others, 1964; Huffman and others, 1978; Thomas and others, 2000). The Sulfur fault separates the basement rocks on the west from Upper Cambrian-Mississippian strata on the east side of the fault (Ham and others, 1954; Ham and others, 1964; Huffman and others, 1978). The Sulfur fault is a steep basement fault which extends farther west into the Arbuckle uplift (Plate 4.1). The Sulfur fault is shown on cross section 2 as a steep, east-dipping fault; however, the fault is mapped as a northeast-vergent reverse fault for part of its length north of the Tishomingo-Belton anticline (Plate 4.1) (Ham and others, 1954). For some distance east of the Sulfur fault and west of the Bengalia fault no reflectors image strata younger than Arbuckle Group. East of the Sulfur fault and west of the
Bengalia fault, Simpson Group-Woodford-Sycamore intervals evidently are truncated beneath the basal décollement of the allochthon.

Absence of any reported imbricates of shallow-water facies Simpson Group through Woodford strata within the southwestern part of the Frontal Ouachita thrust belt (Figure 4.2) leaves two possible explanations for the absence of Simpson Group-Woodford strata in part of cross section 2. First, the missing Simpson Group-Woodford strata were transported in the hanging wall west of the present day Ouachita frontal fault (Choctaw fault) (Plate 4.1) and eroded. McBee (1995) suggests the frontal fault (Choctaw fault) may once have extended as much as 25 miles farther northwest than its present location (Plate 4.1); however, no stratigraphic evidence is explicit. The second possible solution requires that the basal décollement coincide with a pre-thrust angular unconformity surface. Pennsylvanian (Morrowan) unconformities are common within foreland facies stratigraphic successions northwest of the Ouachita thrust belt in Arkansas, Oklahoma, and north Texas (Johnson and others, 1988; Van Arsdale and Schweig, 1990; Cooper, 1995). Morrowan uplift of the Tishomingo-Belton anticline (Plate 4.1) may have caused part to emerge above sea level resulting in erosion of the uppermost Arbuckle-Woodford (and younger pre-Morrowan) part of the stratigraphic succession. In cross section 2, the basal décollement/unconformity surface (Plate 5.1) places allochthonous rocks above upper Arbuckle Group (West Spring Creek Formation). According to Shideler (1970), the West Spring Creek is most abundant of clasts in the Morrowan Johns Valley Formation within the Ouachita Mountains. Cross section 2 may show evidence of a source of West Spring Creek clasts close to the Ouachita thrust front.

In the same manner as in cross section 1, the “Bromide-Sycamore” interval ends abruptly at the east end of cross section 2 (Plate 5.1). The cause is uncertain, and may be evidence of a large-offset normal fault (Gatewood and Fay, 1991), or the possible footwall cutoff for subsurface trace of the Ti Valley fault (see argument in discussion of cross section 1).
**Allochthon:**

In general, there is an eastward deepening of the basal décollement and corresponding eastward thickening of the allochthon. A narrow fault slice with uncertain stratigraphy is sandwiched between the subsurface Choctaw fault and Ti Valley faults. Stratigraphy within the fault slice between the Choctaw fault and Ti Valley fault (in the allochthon) is uncertain (Plate 5.1). The fault slice may contain upper Arbuckle and younger strata; however, the restored length of foreland-facies strata in that fault block is far less than that required to cover the length in cross section 2 where Simpson Group-Woodford strata are absent (Plate 5.1). Alternatively, the fault slice may also contain only younger foreland facies strata, such as the Atoka Formation.

Two distinct faults east of the Ti Valley fault, and are labeled as Faults B and C (Plates 4.1 and 5.1). Fault B is interpreted as a small-offset, west- or southwest vergent thrust fault, and may be the subsurface continuation of the Windingstair fault (Plates 4.1 and 5.1). A broad syncline composed of layered seismic reflectors consistent with upper Stanley Group-Jackfork Sandstone is located between Faults B and C, and another syncline is east of Fault C. A triangularly shaped zone of contored and discontinuous seismic reflectors is located west in the footwall of Fault C and east of an interpreted east- or northeast-vergent back-thrust (Plate 5.1). A tight anticline is formed by the triangular shaped zone. The tight anticline and adjacent broad synclines shown in the allochthon in the central and eastern part of cross section 2 is another example of a “cusp” of a cuspate-lobate fold. Cuspate-lobate folds are common to the Stanley-Atoka succession of the Ouachita Mountains (Arbenz, 1989a, d). As in cross section 1, a fold wavelength mismatch (disharmony) is apparent in cross section 2 between the broad folds of upper Stanley-Jackfork strata (Mississippian-Pennsylvanian) toward the top of the allochthon, and the interpreted tighter, fault-imbricated, folds of the Womble-Arkansas Novaculite- strata (Ordovician-middle Mississippian) (Plate 5.1). As in cross section 1, the lower and middle parts of the
Stanley Group appear to be distorted to compensate for the differences in fold wavelength and amplitude of the underlying and overlying strata (Plate 5.1).

**Cross section 3**

**Overview/location**

Cross section 3 is oriented east-west and located in southeastern Bryan County and southwestern Choctaw County, Oklahoma (near border with Texas) (Plate 4.1). The cross section is highly oblique to the trend of Tishomingo-Belton anticline (Plate 4.1), and is less oblique to the trend of Ouachita thrust belt. Like cross sections 1 and 2, cross section 3 cannot be strictly kinematically balanced. The line traversed by the cross section is completely covered by the Mesozoic Gulf Coastal Plain. Geometry of and stratigraphy of both the autochthon and allochthon are constrained by formation-top data for three wells (B3, C5, and C6) supplied by the Oklahoma Geological Survey (Table 5.2). Cross section 3 is also constrained by seismic profile used to create cross section 5 (which intersects cross section 3).

**Autochthon**

The western end of cross section 3 shows a narrow horst flanked by normal faults. This narrow horst is the southeastern extension of a narrow fault block that separates two of the Coleman half-grabens (Plate 4.1) (Ham and others, 1964; Huffman and others, 1978). Northwest of the western end of cross section 3, Cretaceous strata rest unconformably upon Precambrian Tishomingo Granite (Ham and others, 1964; Huffman and others, 1978). However, an irregular (distorted) appearance of the seismic reflector at the top of the autochthon in the narrow horst suggests a weathered, possibly karstic surface. For this reason, it is suggested that the base of the allochthon above the narrow horst nearly coincides with an unconformity within the upper part of the Cambrian-Ordovician Arbuckle Group carbonates; however, the basal décollement may also be at the top of Tishomingo Granite.

The down-dropped fault block shown in the western end of cross section 3 is a segment of the southernmost part of Coleman half-graben #3 (Plate 4.1) (Huffman and others, 1978). The most clear seismic reflector in the graben is at the top of the autochthon (base of the allochthon). With correlation to well B-1 (Table 5.2), the next
deeper most distinct seismic reflector correlates with the base of the Upper Ordovician Bromide Formation.

East of the narrow horst, the only distinct reflector is located at the top of the autochthon (base of the allochthon). This reflector corresponds to the top of the “Bromide-Sycamore” interval previously described in cross sections 1 and 2. As in cross sections 1 and 2, the eastern end of cross section 3 is also marked by an abrupt termination of the seismic reflectors in the autochthon.

If the abrupt termination of the “Bromide-Sycamore” interval near the eastern end of cross section 3 represents the footwall cutoff of the Ti Valley fault, as suggested for cross sections 1 and 2, restoration of the hanging wall cutoff of the Ti Valley fault (unknown and eroded) to this location requires that a large-offset, down-to-east normal fault coincide with the footwall cutoff. A large difference in thickness distinguishes the deep-water facies Mississippian to Pennsylvanian strata (Stanley Group through Atoka Formations) in the allochthon from the shallow-water equivalent strata (Caney-Goddard-Springer-Atoka) in the autochthon shown in cross sections 1, 2, and 3, indicating a necessary large offset fault to accommodate the change in thickness. On the basis of minimum thickness of the deep-water facies Stanley Group-Atoka Formation measured in the western part of cross sections 1, vertical offset of more than 10,000 ft would be necessary along this fault. The abrupt eastward increase in thickness of the Mississippian Pennsylvanian strata produced by this hypothetical restoration would require this large-offset fault to be a growth fault which began in the Mississippian. This offset would place the top of Arbuckle at a depth of more than 35,000 ft (12.2 km). This required depth places the top of Arbuckle below the depth of the bottom of the seismic profile. A depth of 35,000 ft is similar to that in the deepest part of the Anadarko basin (Figure 4.1) (Bartram and others, 1950; Brewer, 1982; Visher, 1989; Ewing, 1991).

**Allochthon:**

Only the upper part of the allochthon exhibits distinct seismic reflectors. On the bases of intersection with stratigraphic interpretations of cross section 5 which intersects cross section 3, and wells C5 and C6, these layered reflectors correlate with
the upper parts of the Stanley Group through Jackfork Sandstone in the eastern end of
the cross section (Plates 4.1 and 5.2). The lower part of the allochthon exhibits more
diffuse seismic reflectors. The Bigfork Chert-Arkansas Novaculite interval is
discontinuous and not clearly defined, and appears to be offset by several small-offset
faults. These offsets could also be steeply dipping limbs of folds which don’t image
well on seismic reflection profiles.

In the western part of cross section 3, the sole fault of the allochthon bends
sharply upward over the eastern side of the narrow basement horst (Plate 5.1).
However, the basal décollement does not bend down over the western side of the
narrow horst and is nearly flat across the Coleman half-graben (Plates 4.1 and 5.1).
This geometry of the basal décollement shows that the Coleman half-grabens were
formed prior to overthrusting. East of the narrow horst, the basal detachment dips
gently eastwards, but is unconstrained in the eastern end of cross section 3 (Plate 5.1).

In cross section 3, the allochthonous strata are imbricated by two small-
displacement west-or northwest-vergent thrust faults (Faults B and C) (Plates 4.1 and
5.1). Disharmonic folds similar to those shown in cross sections 1 and 2 are shown in
cross section 3; however, degree of disharmony is much less in cross section 3. The
broad folds of the upper part of the Stanley and Jackfork contrast with those of the
shorter-wavelength faulted folds of the underlying Womble-Bigfork Chert-Arkansas
Novaculite strata. A small fault-bounded triangularly shaped zone of deformed
Womble-Arkansas Novaculite-middle Stanley Group strata is located west of Fault C
and east of an adjacent interpreted backthrust (Plate 5.1).

Cross section 4

Overview/location

Cross section 4 is oriented northeast-southwest and extends from northeast
Bryan County to Atoka County in southeastern Oklahoma. Cross section 4 is
perpendicular to strike of the structures in the autochthon and to structures in the
southwestern part of the allochthon (Plates 4.1 and 5.1). The northeastern end of the
cross section is parallel to fold axis trend of Prairie Mountain syncline; therefore, the
northwestern end of cross section 4 is parallel with strike of allochthonous structures.
Cross section 4 is constrained by surface geology derived from published geologic maps (Arbenz, 1989b; Hardie, 1990) and formation-top data for well A1 (Table 5.2 and Plates 4.1 and 5.1) supplied by the Oklahoma Geological Survey. Cross section 4 is also constrained by intersecting seismic reflection profiles used to create cross sections 1 and 3.

**Autochthon**

In cross section 4, the autochthon is subdivided into two main parts. The Sulfur fault separates the northeastern side Tishomingo-Belton anticline (which includes Coleman half-graben #2) on the southwest from the westernmost fringe of the Arkoma basin on the northeast (Figure 4.1 and Plates 4.1 and 5.1).

West of cross section 4, the vertical offset of the Sulfur fault decreases abruptly northwestward along strike. West of the intersection of the Sulfur and Olney faults (Plates 3.9 and 4.1), in the core of the Arbuckle uplift (Figure 4.1 and Plate 3.9), minimal vertical displacement along the Sulfur fault zone is suggested where the Upper Ordovician beds of the upper part of the Arbuckle Group on the northeast are juxtaposed against the lower part of the overlying Simpson Group on the southwest (Ham and others, 1954). East of the intersection of the Sulfur and Olney faults (Plate 4.1), stratigraphic separation increases abruptly where Precambrian Tishomingo Granite on the southwest of the Sulfur fault is juxtaposed against steep, northeast-dipping Middle Ordovician Simpson Group and overlying Upper Ordovician to Pennsylvanian strata (Ham and others, 1954). Southeast of the Olney fault, and northwest of the frontal fault of the Ouachita allochthon, the Sulfur fault marks the southwestern boundary of the Wapanucka syncline (Plate 4.1). The Wapanucka syncline is the southwestern terminus of the Arkoma basin (Figure 4.1).

In cross section 4 southwest of the Sulfur fault, the most distinct reflector beneath the base of the Mesozoic Gulf Coastal Plain rocks is the base of the allochthon. Less distinct imaged reflectors beneath the base of the allochthon include the top of the Viola Group limestone (Upper Ordovician at top), the base of the upper Simpson Group Bromide Formation limestone (Middle Ordovician), and the top of the Upper Cambrian-Lower Ordovician Arbuckle Group carbonates.
Northeast of the Sulfur fault, reflectors dip moderately northward, and flatten farther northeast beneath the allochthon. Two northeast-dipping basement normal faults offset the strata in the autochthon northeast of the Sulfur fault. The Clear Boggy fault extends northwest of the front edge of the allochthon along the northeastern flank of the Tishomingo-Belton anticline (Huffman and others, 1978; Hardie, 1990; Gaines and Yates, 1986). Vertical offset of the top of Arbuckle Group at the Clear Boggy fault in cross section 4 is approximately 500 feet. Another normal fault, labeled the Bengalia fault (Plate 5.1), marks the northeastward limit of clear seismic reflectors characteristic of the shallow-water facies Simpson Group and younger strata exhibited in the southwestern end of cross section 4 (and common west of the Bengalia fault in cross sections 1, 2, and 3). Northeast of the Bengalia fault, the “Bromide-Sycamore” interval exhibits the same “internally transparent” character common to that interval in cross sections 1, 2, and 3 (Plate 5.1). Faint layered seismic reflectors beneath the base of the “Bromide-Sycamore” interval dip northeastward and suggest a northeastward thickening of underlying Simpson Group strata, the same as described for cross section 1. The estimated vertical displacement of the top of the Arbuckle Group at the Bengalia fault is 3000 feet. On the basis of calculated, and/or estimated, interval seismic velocities for allochthonous and autochthonous strata (Table 5.1), the estimated depth of top of Arbuckle in the deepest part of cross section 4 is approximately 22,000 ft. This depth is consistent with a published estimate shown in Gatewood and Fay (1991).

**Allochthon**

The seismic reflectors in the allochthon are more distinct and more laterally continuous than those exhibited in cross sections 1, 2, and 3. Four key reflectors are expressed in the allochthon of cross section 4. The shallowest is at the base of the Atoka Formation. The base of the deep-water facies Atoka formation is less than 2000 ft below datum level in the northeastern end of the cross section. The next deepest key reflector is at the top of the Stanley Group (approximately the top of Mississippian). In Oklahoma, a slightly more dense and higher velocity cherty unit, the Chickasaw Creek Member of the upper Stanley, separates the sandstone and shale turbidites of the
Jackfork Sandstone from similar strata of the upper Stanley Group (Bush and others, 1977; Morris, 1989; Gatewood and Fay, 1991). This Chickasaw Creek Member accentuates the velocity contrast between the base of the Jackfork Sandstone and the top of the lower velocity Stanley Group. The lower seismic reflectors are the top of the Arkansas Novaculite (lower Mississippian at top) and base of the Bigfork Chert (Upper Ordovician).

The basal décollement of the allochthon is bent abruptly upward above foreland facies formations southwest of the Clear Boggy fault. This upward bend in the base of the allochthon could have formed either from post-overthrust or pre-overthrust relative uplift of the southwestern block of the Clear Boggy fault. The frontal thrust fault strikes northwest-southeast at cross section 4, and is probably the subsurface continuation of the Choctaw fault (Plates 4.1 and 5.1). A narrow imbricated fault slice of uncertain stratigraphy is sandwiched between the Choctaw fault and another adjacent thrust fault. Because this northeast-dipping fault has deep-water facies Ordovician to Mississippian strata (Womble through Stanley Group) in the hangingwall, it is interpreted as the southward subsurface continuation of the Ti Valley fault (Plates 4.1 and 5.1).

Northeast of the Ti Valley fault, the structure of the allochthon is separated laterally into two distinct styles. The southwestern part of the allochthon, along the thrust front, consists of two, southwest-vergent faulted folds. The seismically distinct Bigfork Chert-Arkansas Novaculite interval displays the amplitude of these structures. In contrast, the northeastern part of the cross section has the geometry of the southeastern limb of a broad syncline.

The abrupt northeastward change from imbricated tight folds at the southwestern front edge of the allochthon to the broad syncline in the northeastern part in cross section 4 evidently resulted from a combination of two main factors. One of these factors is structural and is caused by the upward bend of the basal décollement of the allochthon above the Clear Boggy fault (Plates 4.1 and 5.1), and the other is geometrically caused by the orientation of cross section 4. The upward bend of the basal décollement in the southeastern part of the cross section, either caused by pre-
thrust or post-thrust vertical offset of the Clear Boggy fault, resulted in a progressive southwestward increase in folds and fault imbricates within the allochthon (Plate 5.1). Because the strike of the Ouachita thrust front changes abruptly from northwest-striking in the southwestern part of cross section 4, to northeast-striking in the northeastern part, cross section 4 changes from a strike-perpendicular cross section (southwest) to a strike-parallel cross section (northeast) and coincides with the central axis of the Prairie Mountain syncline (Plates 4.1 and 5.1).

In the northeastern part of the cross section, where the upper part of the allochthon is not truncated by the Mesozoic unconformity, the marked wavelength mismatch between the upper Stanley-Atoka Formation strata and the Bigfork Chert-Arkansas Novaculite strata, common in cross sections 1, 2, and 3, is not evident in cross section 4 (Plate 5.1). In the northeastern part of the cross section, the Bigfork Chert-Arkansas Novaculite interval is nearly parallel to the overlying reflectors. The contrast in fold wavelengths is not evident because the northeastern part of the cross section is parallel to strike in the central axis of the Prairie Mountain syncline (Plates 4.1 and 5.1).

**Cross section 5**

**Overview/location**

Cross section 5 extends north to south from Atoka County to the southeastern tip of Bryan County, Oklahoma. Cross section 5 intersects cross section 1, 2, and 3 and is constrained by all three. The northern end of cross section 5 is constrained further by surface geology as shown on geologic maps (Arbenz, 1989c, Hardie, 1990). The southern part of cross section 5 is further constrained by formation-top data from well C3 supplied by the Oklahoma Geological Survey (Table 5.2).

**Autochthon**

The only reflector clearly visible beneath the allochthon in cross section 5 is the “Bromide-Sycamore” reflector. This reflector is offset by numerous small-displacement faults. In several locations adjacent to faults, the upper part of the “Bromide-Sycamore” interval is inclined to the base of the allochthon and parts of the interval are truncated. This truncation could be evidence that the base of the
allochthon propogated along an post-early Mississippian (possibly Morrowan) erosional unconformity. The amplitude of the Tishomingo-Belton anticline is far less in cross section 5 than in nearby cross section 1.

**Allochthon**

The upper part of cross section 5 is based on clearly defined layered reflectors that correlate with Mississippian to Pennsylvanian turbidites. Two broad synclines, the Farris syncline at the north end and one labeled Syncline A in the center, are shown on the cross section. A triangularly shaped zone of uncertain internal geometry separates the two synclines. On the basis of faint seismic reflectors correlated with the Bigfork Chert-Arkansas Novaculite interval at the intersection of cross sections 3 and 5, discontinuous and faulted Bigfork Chert-Arkansas Novaculite and younger formations are interpreted as arched above tectonically thickened Womble Shale (Plate 5.1). This triangularly shaped zone is interpreted as the tight core of the broader anticline which separates the two synclines at a higher structural level. The two faults that mark the boundary of this zone, labelled Fault A and Backthrust X, are possibly two parts of the same fault that is bent sharply out of the plane of the cross section (Plates 4.1 and 5.1). In the southern part of the cross section, two faults of opposite vergence are labelled as Fault B and are interpreted with more conviction as two parts of the same fault bent out of the plane of the cross section (Plates 4.1 and 5.1).

Thickness of the Womble varies greatly across cross section 5. Thickness ranges from less than 1000 ft beneath the centers of the two synclines to a maximum of nearly 10,000 ft in the southern part of cross section 5. This far exceeds regional estimates of maximum stratigraphic thickness of 3500 ft (Thomas, 1977); therefore, significant tectonic thickening is evident in the southern part of the cross section and between Syncline A and the Farris syncline (Plates 4.1 and 5.1). The extreme variation of thickness of Womble in cross section 5 is also a function of the geometrical variation of strike of structures in the allochthon with respect to orientation of the cross section. For example, south of the northern limb of Syncline A (Plates 4.1 and 5.1), the line of section is nearly parallel with the Ouachita thrust front and is an oblique cross section. Because a significant component of the dip of strata in cross
section 5 is into the plane of the section, apparent variation in thickness of strata is exaggerated.

**Southern seismic cross section grid**

The “Bryan salient” is the traditional geographical name given for the sharp, northwestward bend in the Ouachita thrust front located south of the Arbuckle uplift (Tishomingo-Belton anticline) and north of the Muenster arch (Figure 4.1) (Huffman and others, 1978; Hardie, 1990). To avoid any possible confusion of scale, the feature traditionally named the “Bryan salient” is referred to in this dissertation, and on Plate 4.1, as the Bryan small-scale salient (BSSS). The axis of the Bryan small-scale salient coincides with the central axis of the Ardmore basin. The Bryan salient is a small-scale bend in the much larger Ouachita salient (Figure 4.1). Using similar nomenclature, the small southeastward bend in the Ouachita thrust front around the nose of the Tishomingo-Belton anticline to the north could be called the Tishomingo-Belton small-scale recess. Folded deep-water facies allochthonous rocks of the Bryan small-scale salient overlie shallow-water facies autochthonous rocks of the Ardmore basin. Because the entire Bryan small-scale salient is covered by Mesozoic Gulf Coastal Plain formations, precise geometry of the Paleozoic structures is not known.

**Cross section 6**

**Overview/location**

Cross section 6 is based on a northeast-trending seismic reflection profile across southwestern Bryan County, Oklahoma. The interpreted cross section is perpendicular to strike of structures in the autochthon. Cross section 6 is also constrained by interpreted structure of intersecting cross section 8 (also based on a seismic profile). Formation-top data for 3 wells (Fergusen 1 Childers, Sunray 1 Beal, and Atlantic Refining Company No. 1 Brown) provided additional constraint. Data for the Fergusen 1 Childers and Sunray 1 Beal wells were calculated from published cross sections of Huffman and others (1978). Data for the Atlantic Refining Company No. 1 Brown well from Flawn and others (1961). On the basis of subcrop maps showing formations below the Mesozoic Gulf Coastal Plain, the frontal fault of the Ouachita allochthon is oblique to strike of the autochthonous structures (Plate 4.1)
(Huffman and others, 1978; Hardie, 1990). Structures within the allochthonous Bryan small-scale salient and underlying “autochthonous” Ardmore basin and adjacent Cumberland, Ravia, and Sand Canyon nappes are very complex (Plate 4.1). Although in the strictest sense, strata within nappes are “allochthonous,” the foreland facies strata within the Cumberland, Ravia, and Sand Canyon nappes are included in the discussion of the autochthonous strata beneath the Ouachita allochthon. Furthermore the boundary faults of the Cumberland, Ravia, and Sand Canyon nappes are perpendicular to, and tectonically unrelated to the structures of the Ouachita allochthon.

**Autochthon**

In cross section 6, a steeply dipping, south-vergent reverse fault is visible at the northeastern end of the seismic profile. This fault is the Washita Valley fault, a major regional fault which marks the southern boundary of the Tishomingo-Belton anticline (Plate 4.1). In cross section 6, uplifted Precambrian igneous and meta-igneous rocks are located on the northern side of the fault. A shallow-depth basement nappe extends southward in the footwall of the Washita Valley fault. The fault at the base of this nappe structure is the Cumberland-Ravia fault, and the nappe is called the Ravia nappe in the vicinity of cross section 6 (Plate 4.1).

In the seismic profile, the base of Ravia nappe is a clear seismic reflector. The 22,500-feet-per-second interval seismic velocity of the Cambrian-Ordovician Arbuckle Group carbonates above the Cumberland-Ravia fault contrasts greatly with the 11,555-feet-per-second interval seismic velocity of the underlying Mississippian Goddard shales. The seismic velocities are calculated on the basis of thicknesses of strata at two wells, Fergusen 1 Childers and Sunray 1 Beal (Plate 5.1), shown in published cross sections by Huffman and others (1978). Other clear reflectors in the autochthon beneath the Ravia nappe are the top of the Viola Formation and the base of the Bromide Formation (upper Simpson Group), and the top of the Arbuckle Group carbonates beneath the Ravia nappe.

The Ravia nappe overlies a basement anticline. The Ravia nappe is one of three nappes along the southwestern side of the Washita Valley fault along the
southwestern flank of the Tishomingo-Belton anticline. From southeast to northwest, these are the Cumberland, Ravia, and Sand Canyon nappes (Plate 4.1). Strata within these nappes are primarily steep, south-dipping to south-overturned Upper Cambrian-Lower Ordovician Arbuckle Group, Middle to Upper Ordovician Simpson Group, and in some areas Upper Ordovician Viola through Mississippian Woodford Formation (Tarr and others, 1965; Huffman and others, 1978).

Two south-dipping faults (Fault X and Fault Y) offset basement beneath the Ravia nappe. On the basis of seismic data, both faults exhibit apparent variable dip separation along the length of each fault which indicates that Fault X and Fault Y are transpressive, oblique-slip faults. At Fault X, the base of the Arbuckle Group appears to be down-dropped toward the south; whereas, structurally higher, the fault exhibits apparent reverse-separation where Cambrian-Ordovician Arbuckle Group strata on the south are juxtaposed against Ordovician-Mississippian stata on the north (Simpson-Caney-Springer). There is also an great difference in apparent thickness of the Arbuckle Group across Fault X; however, reflectors on the northern side of the fault are not as clear as those on the southern side.

To the south, the amount of dip-slip offset on the southern side of Fault Y decreases progressively upwards. The base of the Arbuckle Group carbonate succession on the south side of Fault Y is offset 8000 feet in a reverse sense; whereas, the top of the Arbuckle Group is offset in reverse sense less than 2000 feet (Plate 5.1). No dip separation is apparent at the northern, upper tip of Fault Y where it appears to flatten into bedding and terminate within the Mississippian Caney Shale. Oblique slip of FaultX and Fault Y is consistent with transpressive tectonic models used to explain fold and fault orientations in the Ardmore basin (Harding, 1974; Harding and Lowell, 1979).

South of Fault Y, a near vertical, south-dipping normal fault separates the tightly folded and faulted basement rocks beneath the Ravia nappe from a deep and broad basement syncline to the south (Plate 4.1). This fault is the southeastward, sub-allochthon extension of the Bryan fault, a clearly defined basement fault that parallels the Washita Valley fault (Plate 4.1) (Huffman and others, 1978). At cross section 6
the Bryan fault has a normal offset of 2000 feet down to the south. The Bryan fault marks the northeastern boundary of the Ardmore basin (Plate 5.1). Seismic reflectors are not as clearly defined south of the Bryan fault. One reflector that can be traced reliably correlates with the base of the low-velocity Goddard Formation-Springer Group (upper Mississippian). The other most reliable reflector is the top of Arbuckle Group carbonates (lower Ordovician).

The southern nose of the Ravia nappe is thrust against interpreted steeply dipping, southward-overturned Mississippian-Pennsylvanian Goddard Formation-Springer Group Formation shales and sandstones (Plate 5.1). Pennsylvanian Dornick Hills sandstones and shales sit above the Goddard/Springer Formations in the core of the south-overturned syncline (Syncline D) (Plate 5.1). Because the lower part of the Goddard-Springer strata beneath the Ravia nappe is horizontal, whereas the upper part and overlying Dornick Hills strata are overturned toward the south, a detachment is apparent within the Springer Group. Evidently the south-vergent nose of the Ravia nappe delaminated and overturned strata above the Springer detachment (Plate 5.1). Cross sections across the Ardmore basin to the northwest also show evidence of a significant detachment level in the Springer (Cooper, 1995).

**Allochthon**

Dornick Hills strata within the core of the south-overturned Syncline D separate the south-verging Ravia nappe from the allochthonous Ouachita strata of the Bryan small-scale salient (Plates 4.1 and 5.1). The northeastern edge of the basal décollement of the allochthon has a steep southward dip. Farther south, the dip of the décollement decreases with depth and parallels the bedding in underlying Goddard-Springer strata.

The internal geometry within the allochthon is complex. Except for the southwestern end of cross section 6, where there are no clear seismic reflectors and the stratigraphy within the allochthon is uncertain, the key reflectors within the allochthon are the top of the Arkansas Novaculite (lower Mississippian at top), and the base of the Upper Ordovician Bigfork Chert. The chert-dominated Bigfork Chert-Arkansas Novaculite interval contrasts sharply in velocity with shales of the overlying Stanley
Group and underlying Womble Shale (Plate 5.1). It is interpreted that the Upper Ordovician Bigfork Chert through Mississippian Arkansas Novaculite are contained within three folded imbricate thrust slices (1, 2, and 3) (Plate 5.1). Because the southeastern half of cross section 6 is nearly parallel to strike of the frontal fault of the Bryan small-scale salient (The Kingston fault) (Plate 4.1), the vergence directions of thrust slices 1, 2, and 3 are northwest (into the plane of the cross section).

The Bigfork Chert-Arkansas Novaculite interval is locally offset by internal reverse faults. These reverse faults diverge upwards from décollements at the bases of repeats of Womble shale (Plate 5.1). These divergent faults are examples of out-of-syncline escape structures (Hardie, 1990). The chert-rich, Bigfork Chert-Arkansas Novaculite interval acts as a stiff-layer, and faulted “pop-ups” are surrounded by ductile shales of the Stanley and Womble (Plate 5.1). The rocks in the allochthon in cross section 6 are also located in the upper core of the autochthonous Ardmore basin. There are two non mutually exclusive possible scenarios for the formation of the “pop-up” structures. In one scenario, the “pop-up” structures evidently resulted from the increased lateral compressive stress in the upper parts of the pre-thrust syncline, as the allochthon advanced into a northward-narrowing space. In another scenario, the “pop-ups” resulted from the same increased lateral compressive stress, but within the upper part of a post-thrust syncline. In either scenario, the basal décollement acted as a strain partitioning surface, in effect separating the autochthon from the allochthon. For this reason, the tightly folded strata in the allochthon directly overlie a broad basement syncline.

The geometry of the allochthon at the southern end of cross section 6 is not certain. Approximately 8 miles farther south on the flank of the Criner Hills uplift (adjacent to Criner Hills fault zone, CHFZ in Plate 4.1), the frontal fault of the allochthon is truncated at the unconformity at the base of the Mesozoic Gulf Coastal Plain (Hardie, 1990). Between the southern end of cross section 6 and the Criner Hills uplift, published subcrop maps and cross sections show tight folding of the allochthon (Hardie, 1990).
Cross section 7

Overview/location

The northeast-southwest oriented cross section 7 is located southeast of, and parallel with cross section 6 (Plate 4.1). Cross section 7 extends from southern Bryan County, Oklahoma, through northwestern Fannin and southeastern Grayson Counties, Texas (Plate 4.1). Cross section 7 extends across the southeastern part of the Bryan small-scale salient (Plate 4.1). The Bryan small-scale salient is twice as wide at cross section 7 as at cross section 6 (Plate 4.1). A narrow horst on the northeastern flank of the Muenster uplift, known as the Sherman block marks the southeastern margin of the Bryan small-scale salient. South of the Sherman block, the strike of the frontal fault of the Ouachita allochthon changes abruptly from southeast-striking, to south- and southwest-striking (Plate 4.1).

The primary constraint for cross section 7 is a seismic reflection profile contributed by Digicon. Other constraints are interpreted structure from the intersecting cross section 8 (also based on a seismic profile) and formation-top data for four wells (Pan American Production Company No. 1 J. Umphress, Elkay Oil and Gas Company No. 1 Wilson Lane, No. 1 Ellet, and Mobil No. 1 Col Clazier). Data for the first three listed wells are from Flawn and others (1961), and the fourth is from Robert O. Fay of the Oklahoma Geological Survey. The No. 1 Ellet well is the same as well B4 for which formation-top data was also supplied by the Oklahoma Geological Survey. Subcrop maps and cross sections of Hardie (1990) constrain interpretations of the northern half of the cross section.

Autochthon

In the seismic profile used as the template for cross section 7, the autochthon is completely covered by the allochthonous Ouachita thrust belt. The northern end of cross section 7 is constrained by the nearby Mobil Col Clazier well. In correlation to the Mobil Col Clazier well, the primary reflector in the autochthon at the northern end of cross section 7 is the top of the dense Upper Cambrian-Lower Ordovician Arbuckle Group carbonates. Another clear reflector is the top of the Viola Group limestones. The tops of both the Arbuckle Group and Viola Group mark abrupt velocity contrasts
between denser carbonate rock below, and more shale- and sandstone-rich rock above. On the basis of thicknesses of formations from the Mobil Col Clazier well and two-way travel time between the top of Arbuckle and top of Viola reflectors, an average seismic velocity of 17,000 ft/sec was calculated for the Middle Ordovician Simpson Group through Upper Ordovician Viola Group. An average seismic velocity of 13,800 feet per second was calculated in the same fashion for the overlying Upper Ordovician Sylvan Shale through Mississippian Caney Shale. A lesser velocity contrast separates the Caney Shale from the overlying Mississippian Goddard Shale. See Table 5.1 for a summary list of seismic velocities used for cross section 7.

The northern end of cross section 7 is south of the southeastern part of the Cumberland-Ravia nappe (Plate 4.1). Strata in the autochthon dip southward towards the center of the Ardmore basin (Plate 4.1). Two small-offset south-dipping normal faults offset the autochthonous strata in the northern part of cross section 7. The southernmost of the two faults is likely the along-strike southeastern continuation of the Bryan fault (Plates 4.1 and 5.1). In cross section 7, the deepest part of the Ardmore basin is located approximately 15,000 feet south of the Bryan fault, where the top of Arbuckle Group has a maximum depth of 28,000 feet. The tight folds and faults in the autochthon northeast of the Bryan fault in cross section 6 are absent northeast of the Bryan fault in cross section 7 (Plate 5.1). The autochthon is likely more tightly folded and faulted north of cross section 7 in closer proximity to the Cumberland nappe (Plate 5.1).

The southern end of cross section 7 is constrained by a deep well that penetrated into Precambrian crystalline basement. The Pan American Production Company No. 1 J Umphress well drilled to a total depth of more than 20,000 feet. The seismic reflection profile indicates that the autochthonous strata in the southern part of cross section 7 are within a down-dropped basement graben (Plate 5.1). The northern edge of this graben is separated from the narrow, uplifted Sherman fault block by a south-dipping normal fault (Sherman fault) (Plate 5.1). A kink in the autochthonous strata, coinciding with a small-offset fault, marks the northern side of the Sherman fault block. Down-to-north displacement increases progressively toward the
northwest along the fault that defines the northern side of the Sherman fault block (Plate 4.1). The Sherman fault block and the graben to the southwest are located in the northeastern part of the Muenster arch (Plate 4.1).

Formation-top data from the Umphress well and other wells near the Sherman fault block (Bradfield, 1957a-c) indicate that two regional unconformity surfaces separate stratigraphic successions above the Arbuckle Group carbonates. These unconformity surfaces are clearly defined on the seismic profile even though velocities of adjacent formations are nearly identical.

The uppermost of the two unconformities separates a thick Desmoinesian succession (called “red” Strawn in Texas and Deese Group in Ardmore basin and Arbuckle uplift of southeastern Oklahoma) (see Plate 2.1 for a chronostratigraphic correlation and Figure 4.1 for locations of Arbuckle uplift and Ardmore basin) from the underlying Morrowan-Atokan succession (called Atoka-“grey” Strawn in Texas and Dornick Hills Group in the Ardmore basin and southwestern part of the Arbuckle uplift in Oklahoma) (Plates 2.1 and 5.1). The length of time represented by the unconformity at the base of the Desmoinesian strata is small; however, the unconformity marks the onset of rapid deposition of synorogenic formations in this part of the Ouachita orogenic foreland. Cleaves (1996) also gives evidence that the base of the Desmoinesian also marks an abrupt change from southwest-directed transport of sediment in the Morrowan and Atokan to northwest-directed transport in the foreland strata of the northern part of the Fort Worth basin, located south of the Muenster arch (Figure 4.1 and Plate 4.1).

A much greater hiatus of geologic time is marked by the unconformity that separates the base of the Morrowan succession (Atoka-grey Strawn/Dornick Hills Group) from the top of the Middle Ordovician Simpson Group (Joints-Oil Creek-McLish). In the southern part of cross section 7, the Middle Ordovician, upper part of the Simpson Group (Bromide Formation) through Mississippian (Goddard-Springer strata) are stratigraphically absent. This sub-Morrowan unconformity is a regional unconformity recognized in the foreland stratigraphy of much of northeastern Texas, Oklahoma, and Arkansas (Johnson and others, 1988; Van Arsdale and Schweig, 1990;
Well and seismic data show that this Morrowan unconformity cuts progressively down section from north to south from the southwestern margin of the Arbuckle uplift, across the Ardmore basin, and across the Muenster Arch (Bradfield 1957a-c; Cooper, 1995).

South of the Sherman fault, the autochthonous strata are horizontal within the plane of cross section 7. The strike of the formations in this location is parallel with the cross section and true dip is to the southeast. Two small-displacement normal faults offset the strata below the Morrowan unconformity surface. One of these faults offsets both the base of the Arbuckle Group and the base of the Simpson Group, and the other only the base of the Simpson Group. Because the upper part of the Simpson Group strata are not displaced, evidently these two faults moved during the Middle Ordovician prior to deposition of Middle Ordovician McLish Formation.

North of the Sherman fault block, formations within the autochthon dip northward towards the center of the Ardmore basin in the plane of cross section 7. The Ardmore basin is asymmetric with a longer, and less steeply dipping southern limb. Two complex fault zones (the along strike southeastward sub-allochthon extension of the Criner Hills fault zone, and Fault Zone Z) disrupt the autochthonous strata in the longer southern limb of the Ardmore basin (Plate 5.2).

A vertical offset of 4000 feet separates the Arbuckle and Simpson groups on opposite sides of the Sherman fault (Plate 5.2). To the north, the Arbuckle and Simpson Group reflectors are clearly defined between the Sherman fault and of the Criner Hills fault zone. Within the Criner Hills fault zone, the seismic profile indicates a narrow down-dropped fault block that decreases abruptly in width upward (Plate 5.2). The stratigraphy within this narrow fault block is uncertain; however, if thicknesses of Arbuckle Group and Simpson-Viola strata are similar to thicknesses on opposite sides of the fault-block, Dornick Hills strata would comprise the upper part of the fault-block. A set of upward diverging faults marks the northern end of the Criner Hills fault zone (Plate 5.2).

In cross section 7, south of the deepest part of the Ardmore basin, thickness of the Simpson Group-Viola Group strata decreases progressively southward toward the
northern margin of the Sherman fault block (Plate 5.2). This progressive southward thinning of Simpson Group-Viola Group is evidently caused by erosional truncation beneath the unconformity at the base of the grey Strawn/Dornick Hills strata. Evidence for a similar southward truncation of progressively older strata beneath the sub-Morrowan unconformity in the northwestern part of the Ardmore basin (northwest of cross section 6) is shown in Cooper (1995).

**Allochthon**

From the northern end of cross section 7, the basal décollement of the Ouachita thrust belt cuts progressively down-section southward toward the deep center of the Bryan small-scale salient. At the northern end of cross section 7, interpreted Ordovician Womble Shale above the basal décollement is above autochthonous Mississippian Goddard Shale. In the deepest part of the Bryan small-scale salient, Ordovician Womble Shale above the basal décollement is above Ordovician Simpson Group strata. Southward from the deepest part of the allochthon, Ordovician Womble Shale above the basal décollement is in fault contact with progressively younger Ordovician to Pennsylvanian rocks. At the southern end of the cross section, autochthonous strata (red Strawn/Deese Group) are below the basal décollement.

The central and northern part of cross section 7 contains layered seismic reflectors in the allochthon beneath the Mesozoic Gulf Coastal Plain. According to formation-top data for two wells shown on the cross section (from Flawn and others, 1961), these reflectors correlate with upper parts of the Mississippian Stanley Group and the lower Pennsylvanian Jackfork Sandstone and define two broad synclines. A triangular, fault bounded zone (inbetween adjacent synclines) is located in the deep, center of the Bryan small-scale salient (Plate 5.2). The stratigraphy of the uppermost part of this anticline within the fault bounded zone is uncertain, and published subcrop maps suggest either Jackfork or lower Atoka formations (Flawn and others, 1961; Hardie, 1990). The fault on both sides of this anticline is interpreted as parts of the same fault (Fault D) that bends out of the plane of cross section 7 around the anticline in the footwall on the west (Plates 4.1 and 5.2).
Because the entire length of cross section 7 is oriented approximately perpendicular to translation direction of the allochton, internal structures (including the anticline inbetween Fault D (Plates 4.1 and 5.2) may be pre-thrust or post-thrust structures. Internal structures in the allochton could be syn-thrust structures formed in the center of a pre-thrust embayment, post-thrust structures formed by later folding, or structures formed by a combination of both mechanisms.

Except for the shallow, southern end of cross section 7, the geometry of the lower part of the allochthon is unclear. A faint and discontinuous Bigfork Chert-Arkansas Novaculite interval can be traced across the cross section. The upper and lower boundaries of this interval are constrained by wells at the northern and southern ends of cross section 7.

In the northern end of the seismic profile, the Bigfork Chert-Arkansas Novaculite interval appear to outline a series of tight folds overturned to the south (Plate 5.2). Some of these folds are cut by faults. A bedding parallel fault zone appears to separate the clearly defined reflectors of the upper Stanley and Jackfork strata from underlying contorted Womble Shale through lower Stanley strata. On the basis of the subcrop map of Hardie (1990), this bedding parallel fault zone is the northern continuation of Fault D (Plates 4.1 and 5.2). Within the triangular zone sandwiched between the folded Fault D (center of Bryan small-scale salient), the Bigfork Chert-Arkansas Novaculite interval is disrupted by several north-vergent faults. The faults that disrupt the Bigfork Chert-Arkansas Novaculite interval flatten downward at the basal décollement, and upward in the lower part of the Stanley Group (Plate 5.2). South of the triangular zone, the Bigfork Chert-Arkansas Novaculite is roughly parallel with the basal décollement, and is locally offset by south-vergent and north-vergent faults (Plate 5.2). At the southern end of cross section 7, Mesozoic Gulf Coastal Plain strata rest directly on Ordovician Bigfork Chert.

**Cross section 8**

**Overview/location**

Cross section 8 is located in southern Bryan County, Oklahoma. The seismic reflection profile used as the template for cross section 8 is oriented northwest-
southeast, and is highly oblique to strike of structures in both the autochthon and the allochthon and gives an interesting view in comparison to cross sections 6 and 7. Cross section 8 intersects both cross sections 6 and 7, and is constrained by both. Depths to formations in cross section 8 are constrained by well data derived from published cross sections by Huffman and others (1978) and Hardie (1990). Formation-top data for well B3 (the Enox-Justice Unit well in Huffman and others, 1978) was provided by the Oklahoma Geological Survey.

**Autochthon**

Paleozoic rocks along the entire cross section are unconformably covered by Cretaceous Gulf Coastal Plain strata. Except for the northwestern end of cross section 8, the entire autochthon in the cross section is also covered the Ouachita allochthon (Plate 5.2). The Enox-Justice Unit well (well B3, Table 5.2) (Flawn and others, 1961) located in the northwestern part of cross section 8 bottomed in the lower part of the Middle Ordovician Simpson Group, just above the top of the Arbuckle Group carbonates. For this reason, stratigraphy within the autochthon in the northwestern part of cross section 8 is well constrained.

In the northwestern part, cross section 8 cuts highly obliquely across the basal décollement of the allochthon (Plate 5.2). Also, in the same area, cross section 8 cuts highly obliquely across several basement faults in the autochthon. The seismic profile indicates oppositely dipping faults that bound a fault block in the area marked by three question marks on cross section 8 (Plate 5.2). This structure is an artifact of orientation of the seismic profile. The seismic line cuts back and forth across the same normal fault at two places in this location. For map-scale purposes, and to honor proprietary conditions for use of the seismic profile, precise location of cross section 8 is not shown (Plate 4.1).

Southeast of the Enox-Justice Unit well, the primary seismic reflectors are the base of the Mississippian-Pennsylvanian Goddard-Springer succession and the top of the Cambrian-Ordovician Arbuckle Group carbonates. As calculated from formation thicknesses reported for the Enox-Justice Unit well, an average seismic velocity of 11,555 feet per second for the Goddard-Springer succession contrasts with an average
seismic velocity of 16,430 feet per second for the Upper Ordovician to Mississippian Simpson Group-Caney Shale succession. An estimated velocity of 22,500 feet per second for the dense Arbuckle Group carbonates contrasts sharply with the overlying Simpson Group carbonates and clastic rocks. See Table 5.1 for summary list of seismic velocities.

Where cross section 8 intersects cross section 6, a perpendicular view of some of the structures shown in cross section 6 is shown in cross section 8 (Plate 5.2). A clearly defined fault (Fault Y) offsets the Arbuckle Group. The same Fault Y is shown in cross section 6; however, contrary to cross section 6, offset above the top of the Arbuckle Group is not evident in the seismic profile used to create cross section 8. This indicates that Fault Y loses vertical separation along both strike and dip directions of the fault.

Southeast of Fault Y, cross section 8 shows two other faults with obvious vertical displacement. One of these is a steep-to-southeast-dipping (in plane of cross section) normal fault, which is interpreted as the same Bryan fault that is shown in cross section 6 (Plate 5.2). Seismic reflectors are not as clearly defined southeast of the Bryan fault; however, the top of Arbuckle Group appears to be offset by approximately 7000 feet. The Arbuckle Group also appears to thicken abruptly southeast of the Bryan fault (Plate 5.2). Farther southeast, the Arbuckle Group is offset by a steep, northwest-dipping (in plane of cross section) normal fault. Here, the base of the Arbuckle Group is offset by approximately 2000 ft.

Southeast of the Bryan fault, northwestward toward the shallower frontal part of the Bryan small-scale salient, progressively younger strata in the autochthon are truncated beneath the basal décollement of the allochthon (Plate 5.2). At the Bryan fault, interpreted deep-water facies Middle-Upper Ordovician Womble Shale in the allochthon above the décollement is above Mississippian-Pennsylvanian Goddard-Springer strata (Plate 5.2). Farther southeast, where the basal décollement has the greatest depth, allochthonous Middle-Upper Ordovician Womble Shale is above autochthonous Lower Ordovician strata at the top of the Arbuckle Group carbonate succession (Plate 5.2). There is no evidence of imbricated shallow-water facies strata
within the allochthon in the Bryan small-scale salient. One possible explanation for
the stratigraphically absent autochthonous strata beneath the deep, southeastern central
part of the Bryan small scale salient requires that the basal décollement propagated
across a pre-thrust sub-Morrowan unconformity. This solution requires that the center
of the Ardmore basin (beneath the present day Bryan small-scale salient) was a pre-
Morrowan structural high compared to adjacent areas to the northeast and southwest.
Another possible explanation is that the basal décollement of the allochthon within the
Bryan small-scale salient propagated along a facies transition zone within the Middle
Ordovician-Pennsylvanian succession, where the basal décollement coincides with an
abrupt change in facies (mechanical properties).

**Allochthon**

The total thickness of allochthonous formations in cross section 8 increases
southeastward from less than 3000 feet to approximately 20,000 feet (Plate 5.2). Near
the northwestern end of cross section 8, the basal décollement of the allochthon
separates Upper Ordovician Bigfork Chert from Mississippian-Pennsylvanian
Goddard-Springer strata (Plate 5.2). A narrow zone of uncertain stratigraphy
interpreted to be bounded by oppositely dipping faults separates the thin,
northwesternmost slice of the allochthon from the remainder of the allochthon (Plate
5.2). On the basis of a subcrop map of the Bryan small scale salient by Hardie (1990),
the narrow zone is shown as containing Mississippian Stanley Group strata.

Farther southeast, where cross sections 8 and 6 intersect, four thin imbricate
thrust slices are clearly visible (Plate 5.2). A steep, southeast-dipping fault separates
the southeastern end of the stack of thin thrust slices from a much thicker and less
imbricated allochthon (Plate 5.2). The southeastern part of cross section 8 shows two
broad synclines in the upper part of the allochthon. Layered reflectors correlating to
the upper part of the Mississippian Stanley Group through the Morrowan Jackfork
Sandstone are clearly defined in the seismic profile. Two oppositely verging faults
merge at a near vertical fault that bisects the intervening tight anticline. Beneath the
two oppositely diverging faults is a highly deformed triangular zone of interpreted
Middle Ordovician Womble Shale through Middle Mississippian Stanley Group strata.
The internal geometry within the triangular zone is uncertain; however, seismic reflectors correlated with the Bigfork Chert-Arkansas Novaculite interval are contorted and locally faulted. A relatively flat-lying segment of the Bigfork Chert-Arkansas Novaculite interval, bounded by two oppositely diverging faults, appears to be a “pop-up” structure.

**Cross section 9**

*Overview/location*

Cross section 9 is located at the southern end of the part of the study area where contributed proprietary seismic reflection profiles were studied as part of this dissertation. The cross section is divided into two parts (Northwest and Southeast) and is based upon two connecting seismic reflection profiles which extend from northeastern Collin to southeastern Hunt County, Texas. Constraints on interpretation include well data for four wells (Deep Rock Oil Corporation No. 1 W. M. Sherley, Humble Oil and Refining Co. No. 1, Humble Oil and Refining Co. E.M. Anderson, Humble Oil and Refining Co. No. 1 Norman, Humble Oil and Refining Co. No. 1 Rutherford) in Flawn and others (1961) and well data for one well (No. 1 Hawkins Gas Unit) in Wood and Guevara (1981), and a pre-Mesozoic subcrop map in Flawn and others (1961).

Of all the cross sections discussed in this chapter, cross section 9 is buried most deeply beneath the Mesozoic Gulf Coastal Plain. The Mesozoic succession is clearly defined by seismic reflectors in the cross section, and cross section 9 is the only cross section which shows Jurassic age strata. The Mesozoic formations in the southeastern end of the cross section are offset by faults of the Mexia-Talco fault zone which mark the northwestern boundary the Mesozoic age East Texas structural and sedimentary basin.

*Autochthon*

In the middle of the northwest part of cross section 9, strata in the autochthon are cut by a steep, southeast-dipping normal fault (labeled Fault 1) (Plate 5.2). Autochthonous strata are clearly defined by seismic reflectors northwest of Fault 1. Northwest of Fault 1 in cross section 9, seismic reflectors are correlated on the basis of
the data from the Deep Rock Oil Corporation No. 1 W. M. Sherley well (Flawn and others, 1961), and interpreted stratigraphy from the southwestern end of cross section 7 which is located nearby to the north (Plates 4.1 and 5.2). The two most distinct seismic reflectors northwest of Fault 1 are the base of the Desmoinesian (red Strawn/Deese) succession and the top of the Upper Cambrian-Lower Ordovician Arbuckle Group (Plate 5.1). At the northwestern end of cross section 9, two northwest-vergent basement reverse faults displace the Arbuckle Group through lower part of the red Strawn/Deese strata.

Southeast of Fault 1 in cross section 9, stratigraphy in the autochthon is poorly constrained. Only several faint reflectors near the base of the overtrust Ouachita allochthon show southeast dips. Cross section 9 shows a plausible (albeit hypothetical) stratigraphy for the autochthon in the fault block southeast of Fault 1. On the basis of southeast-overturned folds of Bigfork Chert-Arkansas in the overlying allochthon which are interpreted as formed by south-east directed slumping along the basal décollement, Fault 1 is interpreted as a post-thrust normal fault. For this reason, in cross section 9, thicknesses of strata are maintained across Fault 1. However, Fault 1 could also be a growth fault marked by abrupt thickening of Morrowan-Desmoinesian strata on the southeast. Southwest of cross section 9 (and southwest of the Muenster arch) beneath the Ouachita thrust front, Morrowan-Atokan (post-Marble Falls Limestone) down-to-southeast growth faults offset strata within the autochthon (Plate 3.10).

In the northwest part of cross section 9, one northwest-dipping, and two southeast-dipping normal faults are shown southeast of Fault 1 (Plate 5.2). One of the normal faults (Fault 2) is located at the southeastern end of the northwest part of cross section 9 (Plate 5.2). For a distance of approximately 7.6 miles (≈ 12.2 km) southeast of fault 2, stratigraphy and structure within the autochthon are completely unconstrained.

Farther southeast, the autochthon shown in southeast part of cross section 9 is cut by one large-displacement, northwest-vergent basement reverse fault (Fault 3). Basement Fault 3 displaces both autochthonous and overlying allochthonous strata.
Within the allochthon, the location where Fault 3 intersects the Mesozoic unconformity is the Luling front (Plates 2.1, 3.1, and 4.1), which is a fault zone that is approximately parallel with strike of the Ouachita thrust front and separates slightly higher grade metamorphosed strata on the southeast from lower grade and unmetamorphosed strata on the northwest (Arbenz, 1989e; Thomas and others, 1989).

Stratigraphy within the autochthon southeast of Fault 3 is poorly constrained; however, two deep wells located along strike to the northeast in the Broken Bow uplift, and one along strike to the southwest in the Waco uplift penetrated pre-Mississippian shallow-water carbonates. On the basis of lithologies of rocks in the two wells in the Broken Bow uplift, the basal décollement of the allochthon in that area is interpreted to rest upon Middle Ordovician Simpson Group strata (Leander and Legg, 1988; Denison, 1989). In the Waco uplift, autochthonous strata possibly as young as Devonian are interpreted to rest below the basal décollement of the allochthon (Nicholas and Rozendal, 1975). In cross section 9, the autochthon is contorted into several broad folds, with a distinct syncline beneath the Mesozoic structures of the Mexia-Talco fault zone (Plate 5.2). At this syncline, one reflector (shown as Reflector A) on cross section 9 separates a contorted package of discontinuous reflectors above, from underlying undulatory continuous reflectors. It is uncertain whether Reflector A marks the base of the allochthon, or the abrupt transition to deformed shaley autochthonous strata. On the basis of well data from the Broken Bow and Waco uplifts, Reflector A is shown on cross section 9 as the top of Upper Cambrian-Middle Ordovician shallow-water carbonate strata (Plate 5.2).

**Allochthon**

At the northwestern end of cross section 9, the basal décollement of the allochthon is flat lying and separates Middle Ordovician Womble Shale above, from Pennsylvanian (Desmoinesian) red Strawn/Deese strata below the décollement (Plate 5.2). The allochthon is very thin in the northwestern part of cross section 9, where Mesozoic Gulf Coastal Plain strata rest upon Upper Ordovician Bigfork Chert (Plate 5.2). Thickness of the allochthon increases abruptly at Fault 1 where the basal décollement also increases abruptly in depth from a depth of approximately 6000 ft on
the northwest side of the fault, to more than 15,000 ft on the southeast side (Plate 5.1). Southeast of Fault 1, the dip of basal décollement is more gentle and is nearly horizontal northwest of Fault 2 (Plate 5.2). At Fault 2, the basal décollement bends downward and can be traced on seismic farther southeast for a distance of approximately 4.7 miles (7.6 km). Still farther southeast, and northwest of Fault 3, location of the basal décollement is poorly defined.

Between basement Faults 1 and 3 in cross section 9, two broad synclines and an intervening fault-cored anticline (Fault D) are clearly defined by seismic reflectors that correlate to the upper part of the Mississippian Stanley Group through Pennsylvanian Jackfork Sandstone. On the basis of published subcrop maps (Flawn and others, 1961; Hardie, 1990) and cross sections in Hardie (1990), the base of the Jackfork Sandstone is imprecisely located as shown in cross section 9, and Fault D is interpreted to be the southerly along-strike continuation of Fault D shown in cross section 8 (Plates 4.1 and 5.2). Location of the base of the Jackfork is also constrained by published estimates of regional maximum stratigraphic thickness of the underlying Stanley Group (for example, Thomas, 1977).

Near the center of cross section 9, the northwest-vergent Fault D roots at the basal décollement at basement fault 2 (Plate 5.2). Between Fault D and basement Fault 2, within the allochthon, a faint contorted set of reflectors, interpreted as Upper Ordovician-Mississippian Bigfork Chert-Arkansas Novaculite (on the basis of stratigraphic position), are roughly parallel with the basal décollement of the allochthon. Strata below the base of the Bigfork Chert-Arkansas Novaculite interval is interpreted as Middle Ordovician Womble Shale. A triangular zone of tightly folded Womble through lower Stanley Group strata is apparently defined by seismic reflectors northwest of Fault D (Plate 5.2). This triangular zone is tightly folded between the northwest-vergent Fault D and a southeast-vergent backthrust fault (Plate 5.2). Between Fault D and the large-displacement basement Fault 3 which uplifts the allochthon on the southeast side, geometry of strata beneath the upper part of the Stanley Group as shown in cross section 9 is poorly defined on the seismic reflection profile.
Southeast of basement Fault 3 (southeast of the Luling front), the allochthon is much thinner and much shallower than northwest of the fault. As shown on cross section 9, three wells that bottomed in Mississippian Stanley Group strata (Flawn and others, 1961; Wood and Guevara, 1981) constrain interpretation (Plate 5.2). For most of the part of cross section 9 that is southeast of the Luling front, relatively thin Mississippian Stanley Group shales of variable thickness rest between overlying Jurassic Smackover limestones and underlying cherty Upper Ordovician-Mississippian Bigfork Chert-Arkansas Novaculite strata (Plate 5.2). On the seismic profile, the relatively low velocity Stanley Group shales contrast with the higher velocity of the overlying Smackover and the underlying Bigfork Chert-Arkansas Novaculite. In two locations of particular interest southeast of the Luling front on cross section 9, the Smackover interval is shown to pinch out laterally on the flanks of Bigfork Chert-Arkansas Novaculite strata which evidently were pre-Jurassic topographic highs (Plate 5.2).

Southeast of the Luling front in cross section 9, stratigraphy and structure beneath the clearly defined Bigfork Chert-Arkansas Novaculite interval is poorly constrained. The only distinct reflector beneath the Bigfork Chert-Arkansas Novaculite interval is the top of the autochthonous Upper Cambrian-Middle Ordovician shallow-water carbonates (top of Arbuckle-Simpson)(Plate 5.2). An especially difficult-to-interpret assemblage of contorted seismic reflectors is located in the allochthon beneath the Mesozoic structures of the Mexia-Talco fault zone (Plate 5.2). Because interpretation is very difficult, two possible locations of the basal décollement (1 and 2) are shown on the southeastern end of cross section 9 (Plate 5.2). Because the strata between these two possible levels of basal décollement lacks clearly defined reflectors, it is interpreted to be either non-cherty allochthonous strata or non-calcareous autochthonous strata.

Summary

Geometry within the allochthon

The concentration of tectonically thickened Womble Shale and Stanley Group in steep anticlines between broad synclines is a common feature of cross sections 1-9,
and of regional cross section constructed across the Ouachita thrust belt (Arbenz, 1989a). This geometry of broad synclines surrounding a narrow fault-cored anticline (or cuspate-lobate fold) is found throughout the Ouachita thrust belt where the stratigraphy is predominantly thick layered sandstones and shales (upper part of the Mississippian Stanley Group through Pennsylvanian Atoka Formation). Extreme variation in thickness of the Womble Shale in cross section 1-5 (Plate 5.1) may be exaggerated by abrupt changes in thrust-belt strike which causes apparent thickness of units within the plane of the cross section change abruptly, and/or substantial deformation of the shale.

Possible Middle Ordovician to Lower Mississippian autochthonous transitional/deep water facies between Arbuckle uplift and Broken Bow uplift

The interpretation that the basal décollement rests on Middle Ordovician Simpson Group strata in the northern part of cross section 7 and on Lower Ordovician strata at the top of Arbuckle Group in the southeastern part of cross section 8, may be a key to understanding regional structural geology. Both cross sections cross part of the Bryan small scale salient, a northwest-directed bend in the much larger Ouachita thrust belt, located between the Arbuckle uplift to the north and Muenster uplift to the south (Figure 4.1 and Plate 4.1). In both cross sections, toward the center of the Bryan small-scale salient, where the basal décollement is deepest, it rests upon progressively older autochthonous strata, and the allochthonous strata in the Bryan small scale salient are folded above the autochthonous strata of the Ardmore basin (Figure 4.1 and Plate 5.1). There is no indication of imbricated shallow-water facies strata in cross sections 6, 7, or 8 which cross the Bryan small-scale salient, and virtually none in any of the other cross sections discussed in this chapter. Furthermore, published cross sections in Hardie (1990) and subcrops maps of the Bryan small-scale salient and adjacent foreland show no evidence of imbricated foreland-facies strata older than Mississippian-Pennsylvanian Springer Group (Huffman and others, 1978; Hardie, 1990). Neither is there evidence of significant imbricated pre Mississippian shallow-water facies strata northeast of the Arbuckle uplift along the Ouachita thrust front (Hardie, 1988; Arbenz, 1989a, b). Furthermore, two deep wells drilled in the Broken
Bow uplift region of southeastern Oklahoma and northeastern Texas (Figure 4.1) penetrated autochthonous low-grade quartzites and carbonates interpreted as Middle Ordovician Simpson Group and underlying Upper Cambrian-Lower Ordovician Arbuckle Group beneath the basal décollement of the allochthon (Leander and Legg, 1988; Denison, 1989).

Because the décollement cuts stratigraphically upward toward the west, progressively older autochthonous strata remain beneath the footwall cutoff toward the interior of the Ouachita allochthon. The lack of an extensive area of corresponding hanging-wall cutoffs or imbricated footwall horses poses a puzzle. One possible solution is that the missing hanging-wall stratigraphy has been transported westward beyond the presently preserved frontal fault of the allochthon, and ensuing erosion has removed the translated foreland strata. This scenario seems unlikely because a large mass of rock would have been translated at the base of relatively weak rocks of the lower part of the allochthon. There is also a lack of evidence of clasts that might have been derived from erosion of such imbricates in formations west of the Bryan structural salient.

Another possible solution for the missing Ordovician to Mississippian strata requires that the basal décollement of the Ouachita allochthon coincide with an unconformity surface at depth across the southwestern part of the Ouachita Mountains (west of the Broken Bow uplift, Figure 4.1) and beneath the Ouachita allochthon as far south as the Muenster arch. In this scenario, the shallow-water facies carbonates and clastic rocks were removed by erosion prior to deposition of the Morrowan and later cover. Northwestward translation of the allochthon imbricated Morrowan and younger strata above the unconformity/basal décollement along the Ouachita thrust front and emplaced progressively older and deeper-water facies strata (derived from an area east of the Broken Bow uplift) westward toward the interior of the Ouachita allochthon. Across much of the foreland of the Ouachita orogen north of the Muenster arch (Plate 3.7) and west of the Mississippi embayment (Plate 3.7) an unconformity either marks the base of Morrowan, or marks the base of uppermost Morrowan (or Atokan) through Desmoinesian succession (Van Arsdale and Schweig, 1990; Cooper,
The upper Morrowan Johns Valley Shale which is imbricated within part of the Ouachita allochthon in the Ouachita thrust belt also contains numerous exotic clasts, blocks, and boulders of Cambrian to Pennsylvanian foreland-facies lithologies (Shideler, 1970), further documenting erosion of rocks adjacent to the Ouachita depositional sites.

Still another possible solution to the missing Ordovician-Mississippian strata requires that the basal décollement coincide with an abrupt facies transition zone in the Ordovician at depth across the western Ouachita Mountains and beneath the Ouachita allochthon as far south as the Muenster arch. On the basis of data from two deep wells within the Broken Bow uplift which drilled interpreted middle Simpson Group strata (McLish Formation) beneath the basal décollement (Leander and Legg, 1988; and Denison, 1989), the facies transition can be no older than Middle Ordovician which correlates with deep-water facies Womble Shale. Cross sections of the Potato Hills and Black Knob Ridge areas of the western Ouachita Mountains (Plate 4.1), which show no imbricated deep-water facies older than Womble Shale, are consistent with the Ordovician facies change scenario.

**Conclusion**

Footwall geometry illustrated in the newly interpreted structural cross sections, included as part of this dissertation, provide a possible solution to restoration problems encountered in published cross sections across the western part of the Ouachita Mountains. Evidence is given to suggest that a facies transition from shallow-water carbonates to deep-water cherts and shales is located in the upper Ordovician rocks of the upper Simpson Group through Woodford/Sycamore Formations far to the west of the Broken Bow uplift along the southeastern flank of the Arbuckle uplift. This facies boundary bends westward up the axis of the Ardmore basin. Finally, the basal décollement appears to coincide with an unconformity (possibly Morrowan) in the southeastern end of the Tishomingo-Belton anticline, also suggesting a source for Arbuckle facies clasts that are common in the Morrowan Johns Valley Shale in the western part of the Ouachita Mountains (Shideler, 1970).
Table 5.1: Summary chart showing interval seismic velocities for formations shown in cross sections 1 through 9 in Plates 5.1 and 5.2. Each chart is an interval seismic velocity stratigraphic column for a location specified above each chart. Interval seismic velocity determined from: \( w = \) interval thickness in well, \( p = \) published interval velocities, and or constrained by other published data, \( e = \) estimation based on regional calculations of velocity for a specific lithology, \( a = \) average.

### CROSS SECTION 1 (west of frontal fault of Ouachita thrust belt)

<table>
<thead>
<tr>
<th>TERRAIN</th>
<th>AGE</th>
<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>autochthonous</td>
<td>Pennsylvanian</td>
<td>Atoka</td>
<td>10,300w</td>
<td>shallow-marine clastic</td>
</tr>
<tr>
<td>footwall</td>
<td>(Atokan)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Penn. (Atokan)</td>
<td>basal Atoka member (Spiro Sandstone)</td>
<td>17,733w</td>
<td>shallow-marine clastic</td>
</tr>
<tr>
<td></td>
<td>Penn. (Morrowan)</td>
<td>Wapanucka</td>
<td>17,733w</td>
<td>shallow-marine limestone and sandstone</td>
</tr>
<tr>
<td></td>
<td>Mississippian-Penn.</td>
<td>Caney Formation-Springer Group</td>
<td>13,500w</td>
<td>shallow-marine shales and sandstone</td>
</tr>
<tr>
<td></td>
<td>(Morrowan)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Silurian-Mississippian</td>
<td>Sylvan-Woodford-(Sycamore?)</td>
<td>13,500w</td>
<td>shallow-marine carbonates, shales, and cherts</td>
</tr>
<tr>
<td></td>
<td>Late Ordovician</td>
<td>Simpson Group-(Bromide)-Viola</td>
<td>20,200w</td>
<td>shallow- to deep-marine carbonates (carbonaceous)</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson Gp. (Oil Creek-Tulip Creek)</td>
<td>12,300w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson Gp. (Joins and basal Oil Creek)</td>
<td>14,500w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
<td>22,500e</td>
<td>shallow-marine carbonates</td>
</tr>
</tbody>
</table>

### CROSS SECTION 1 (east of frontal fault of Ouachita thrust belt)

<table>
<thead>
<tr>
<th>TERRAIN</th>
<th>AGE</th>
<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>allochthonous</td>
<td>Pennsylvanian</td>
<td>Atoka</td>
<td>10,500w</td>
<td>deep-marine turbidites</td>
</tr>
<tr>
<td>Ouachita</td>
<td>(Atokan)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>thrust belt</td>
<td>Penn. (Morrowan)</td>
<td>Johns Valley Shale</td>
<td>14,433w</td>
<td>deep-marine turbidites</td>
</tr>
<tr>
<td></td>
<td>latest Miss. (Chesterian) to Penn. (Morrowan)</td>
<td>Jackfork Sandstone</td>
<td>14,433w</td>
<td>deep-marine turbidites</td>
</tr>
<tr>
<td></td>
<td>Mississippian</td>
<td>Stanley Group</td>
<td>13,000w</td>
<td>deep-marine turbidites and tuffs</td>
</tr>
<tr>
<td></td>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert-Arkansas Novaculite</td>
<td>18,560w</td>
<td>deep-marine cherts and clastics</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Womble Shale</td>
<td>15,500e</td>
<td>deep-marine shale</td>
</tr>
<tr>
<td>autochthonous</td>
<td>Mississippian</td>
<td>Caney-Goddard or Stanley Group</td>
<td>15,500e</td>
<td>shallow- to transitional-depth clastic</td>
</tr>
<tr>
<td>footwall</td>
<td>Late Ordovician to early Mississippian</td>
<td>&quot;Bromide-Sycamore&quot; interval</td>
<td>22,500e</td>
<td>cherty limestones or limey cherts (transition facies?)</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson Group-(Joins-Tulip Creek)</td>
<td>12,300p</td>
<td>shallow-marine to transitional-depth carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
<td>22,500e</td>
<td>shallow-marine carbonate</td>
</tr>
</tbody>
</table>
### Table 5.1: (continued)

**CROSS SECTION 2 (east of Sulfur fault and west of frontal fault of Ouachita thrust belt)**

<table>
<thead>
<tr>
<th>TERRAIN</th>
<th>AGE</th>
<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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</thead>
<tbody>
<tr>
<td>Gulf Coastal Plain</td>
<td>Cretaceous</td>
<td></td>
<td>5,283w</td>
<td>shallow-marine limestone and sandstone</td>
</tr>
<tr>
<td>autochthonous footwall</td>
<td>Pennsylvanian (Morrowan)</td>
<td>Springer Group</td>
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<td>shallow-marine shales and sandstones</td>
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<td>Mississippian</td>
<td></td>
<td>Caney/Goddard-lower Springer Group</td>
<td>13,500w</td>
<td>shallow-marine shales and sandstones</td>
</tr>
<tr>
<td>Silurian</td>
<td></td>
<td>Sylvan Shale</td>
<td>13,500w</td>
<td>shallow- to deep-marine shales</td>
</tr>
<tr>
<td>Late Ordovician</td>
<td></td>
<td>Simpson Group</td>
<td>20,200w</td>
<td>shallow- to deep marine carbonates (carbonaceous)</td>
</tr>
<tr>
<td>Middle Ordovician</td>
<td></td>
<td>Simpson Group (McLish-Tulip Creek)</td>
<td>12,300w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td>Middle Ordovician</td>
<td></td>
<td>Simpson Group (Oil Creek)</td>
<td>14,500w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td>Cambrian-Ordovician</td>
<td></td>
<td>Arbuckle Group</td>
<td>22,500e</td>
<td>shallow-marine carbonates</td>
</tr>
<tr>
<td>basement</td>
<td>Precambrian-Cambrian (pre-Franconian)</td>
<td>layered reflectors</td>
<td></td>
<td>Rhyolite flows or gneissic granite gneiss or Amphibolite</td>
</tr>
</tbody>
</table>

**CROSS SECTION 2 (just west of Clear Boggy fault)**

<table>
<thead>
<tr>
<th>TERRAIN</th>
<th>AGE</th>
<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gulf Coastal Plain</td>
<td>Cretaceous</td>
<td></td>
<td>5,283w</td>
<td>shallow-marine limestones, sandstones, and shales.</td>
</tr>
<tr>
<td>allochthonous Ouachita thrust belt</td>
<td>Ordovician</td>
<td>Womble Shale (no discernible reflectors)</td>
<td>11,393w (low velocity)</td>
<td>deep-marine shales</td>
</tr>
<tr>
<td>autochthonous footwall</td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
<td>22,500e</td>
<td>shallow-marine carbonates</td>
</tr>
</tbody>
</table>

**CROSS SECTION 2 (eastern part of cross section)**

<table>
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<tr>
<th>TERRAIN</th>
<th>AGE</th>
<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gulf Coastal Plain</td>
<td>Cretaceous</td>
<td></td>
<td>5,283w</td>
<td>shallow-marine facies limestone and sandstone</td>
</tr>
<tr>
<td>allochthonous Ouachita thrust belt</td>
<td>latest Miss. (Chesterian) to Penn. (Morrowan)</td>
<td>Jackfork Sandstone</td>
<td>14,433w</td>
<td>deep-marine turbidites</td>
</tr>
<tr>
<td>Mississippian</td>
<td></td>
<td>Stanley Group</td>
<td>11,393w</td>
<td>deep-marine turbidites and tuffs</td>
</tr>
<tr>
<td>Late Ordovician</td>
<td></td>
<td>Bigfork Chert-Arkansas Novaculite</td>
<td>18,560w</td>
<td>deep-marine cherts and clastics</td>
</tr>
<tr>
<td>triangle-shaped deformed zone</td>
<td>Ordovician to Mississippian</td>
<td>(Womble Shale?)-Stanley Group</td>
<td>12,156p</td>
<td>deep-marine shales</td>
</tr>
</tbody>
</table>

286
<table>
<thead>
<tr>
<th>TERRAIN</th>
<th>AGE</th>
<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>autochthonous footwall</td>
<td>Mississippian</td>
<td>Caney-Springer or Stanley Group</td>
<td>12,156</td>
<td>shallow-marine to deep-marine shales and sandstones</td>
</tr>
<tr>
<td></td>
<td>Late Ordovician to early Mississippian</td>
<td>Bromide-Sycamore interval</td>
<td>22,500e</td>
<td>cherty limestones or limey cherts (transitional facies?)</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson Group (Joins-Tulip Creek)</td>
<td>12,300p</td>
<td>shallow-marine to transitional-depth carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
<td>22,500e</td>
<td>shallow-marine carbonates</td>
</tr>
</tbody>
</table>

### CROSS SECTION 3 (Coleman half-graben #3, western end of cross section, Plate 4.1)

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<thead>
<tr>
<th>TERRAIN</th>
<th>AGE</th>
<th>FORMATIONS</th>
<th>SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gulf Coastal Plain</td>
<td>Cretaceous (top)</td>
<td></td>
<td>4,255w</td>
<td>shallow-marine (sandstone)</td>
</tr>
<tr>
<td></td>
<td>Cretaceous (bottom)</td>
<td></td>
<td>7,000w</td>
<td>shallow-marine (limestone)</td>
</tr>
<tr>
<td>allochthonous Ouachita thrust belt</td>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert-Arkansas Novaculite</td>
<td>18,560w</td>
<td>deep-marine cherts and shales</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Womble Shale</td>
<td>15,500w</td>
<td>deep-marine shale</td>
</tr>
<tr>
<td>autochthonous footwall</td>
<td>Pennsylvanian (Morrow)</td>
<td>Springer Group</td>
<td>13,500w</td>
<td>shallow-marine shales and sandstones</td>
</tr>
<tr>
<td></td>
<td>Mississippian</td>
<td>Caney/Goddard-Springer Group</td>
<td>13,500w</td>
<td>shallow-marine shales and sandstones</td>
</tr>
<tr>
<td></td>
<td>Silurian</td>
<td>Sylvan Shale</td>
<td>13,500w</td>
<td>shallow to deep marine shales</td>
</tr>
<tr>
<td></td>
<td>Late Ordovician</td>
<td>Simpson Group (Bromide)-Viola</td>
<td>20,200w</td>
<td>shallow-marine to deep-marine carbonates (carbonaceous)</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson Group (Joins-Tulip Creek)</td>
<td>14,250w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td>Ordovician-Cambrian</td>
<td>Arbuckle Group</td>
<td>22,500e</td>
<td>shallow-marine carbonates</td>
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<tr>
<td>basement</td>
<td>Precambrian-Cambrian (pre-Franconian)</td>
<td>inclined layered reflectors</td>
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<td>Rhyolite flows or granitic gneiss or Amphibolite</td>
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### CROSS SECTION 3 (narrow horst east of Coleman half-graben #3, Plate 4.1)

<table>
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<th>AGE</th>
<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<tr>
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<td>4,255w</td>
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<td>Cretaceous (bottom)</td>
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<td>shallow-marine (limestone)</td>
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<tr>
<td>allochthonous Ouachita thrust belt</td>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert-Arkansas Novaculite</td>
<td>18,560w</td>
<td>deep-marine cherts and clastics</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Womble Shale</td>
<td>15,500w</td>
<td>deep-marine shale</td>
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<tr>
<td>autochthonous footwall</td>
<td>Ordovician-Cambrian</td>
<td>Arbuckle Group (possible karst)</td>
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<tr>
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<td>Rhyolite flows or granitic gneiss or Amphibolite</td>
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Table 5.1: (continued)

CROSS SECTION 3 (central part of cross section)

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<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<td>4,255w</td>
<td>shallow-marine sandstone</td>
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<tr>
<td></td>
<td>Cretaceous (bottom)</td>
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<td>7,000w</td>
<td>shallow-marine limestone</td>
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<td></td>
<td>allochthonous Ouachita thrust belt</td>
<td>Mississippian</td>
<td>13,625w</td>
<td>deep-marine turbidites and tuffs</td>
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<td></td>
<td></td>
<td>Stanley Group</td>
<td>18,560w</td>
<td>deep-marine cherts and clastics</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Bigfork Chert-Arkansas Novaculite</td>
<td>15,500w</td>
<td>deep-marine shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(Womble Shale?)-Stanley Group</td>
<td>14,250w</td>
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<td>Mississippian-Caney-Springer Gp. or Stanley Group</td>
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<td>shallow-marine to deep-marine sandstones and shales</td>
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<td></td>
<td>&quot;Bromide-Sycamore&quot; interval</td>
<td>22,500w</td>
<td>cherty limestones or limey cherts (transitional facies?)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Simpson Group (Joins-Tulip Creek)</td>
<td>12,300p</td>
<td>shallow-marine to transitional-depth carbonates, sandstones, and limey shales</td>
</tr>
<tr>
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<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
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CROSS SECTION 4 (in Coleman half graben #2, southwestern end of cross section, Plate 4.1)

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<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<tr>
<td></td>
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<td>Pennsylvanian (Atokan) Atoka</td>
<td>10,500w</td>
<td>shallow-water shales and sandstones</td>
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<tr>
<td></td>
<td></td>
<td>basal Atoka member (Spiro Sandstone)</td>
<td>17,733w</td>
<td>shallow-marine sandstone</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Wapanucka</td>
<td>17,733w</td>
<td>shallow-marine limestones and sandstones</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Springer Group</td>
<td>13,500w</td>
<td>shallow-marine shales and sandstones</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Caney/Goddard-Springer Group</td>
<td>13,500w</td>
<td>shallow-marine shales and sandstones</td>
</tr>
<tr>
<td></td>
<td>Silurian</td>
<td>Sylvan Shale</td>
<td>13,500w</td>
<td>shallow to deep marine shale</td>
</tr>
<tr>
<td></td>
<td>Late Ordovician</td>
<td>Simpson Group (Bromide)-Viola</td>
<td>20,200w</td>
<td>shallow-marine to deep-marine carbonates (carbonaceous)</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson Group (McLish-Tulip Creek)</td>
<td>12,300w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson (Oil Creek)</td>
<td>14,500w</td>
<td>shallow-marine carbonates and limey shales</td>
</tr>
<tr>
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<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
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Table 5.1: (continued)

### CROSS SECTION 4 (northeastern end)

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<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<tr>
<td>allochthonous Ouachita thrust belt</td>
<td>Pennsylvanian (Atokan)</td>
<td>Atoka</td>
<td>10,500w</td>
<td>deep-water turbidites</td>
</tr>
<tr>
<td></td>
<td>Penn. (Morrowan)</td>
<td>Johns Valley Shale</td>
<td>14,433w</td>
<td>deep-water turbidites</td>
</tr>
<tr>
<td></td>
<td>latest Miss. (Chesterian to Penn. (Morrowan)</td>
<td>Jackfork Sandstone</td>
<td>14,433w</td>
<td>deep-water turbidites</td>
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<tr>
<td>Mississippian</td>
<td>Stanley Group</td>
<td>13,000w</td>
<td>deep-water turbidites</td>
<td></td>
</tr>
<tr>
<td>Late Ordovician to early Mississippian</td>
<td>Arkansas Novaculite</td>
<td>18,560w</td>
<td>deep-marine cherts and clastics</td>
<td></td>
</tr>
<tr>
<td>autochthonous footwall</td>
<td>Mississippian</td>
<td>Caney-Springer or Stanley Group</td>
<td>15,500e</td>
<td>shallow-marine to deep-marine sandstones and shales</td>
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<tr>
<td></td>
<td>Late Ordovician to early Mississippian</td>
<td>&quot;Bromide-Sycamore&quot; interval</td>
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<td>cherty limestones or limey cherts (transitional facies?)</td>
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<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson Group</td>
<td>12,300w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(McLish-Tulip Creek)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson Group</td>
<td>14,500w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(Joins-Oil Creek)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ordovician-Cambrian</td>
<td>Arbuckle Group</td>
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### CROSS SECTION 5

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<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<td>top</td>
<td>4255w</td>
<td>shallow-marine sandstone</td>
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<td></td>
<td></td>
<td>bottom</td>
<td>7000w</td>
<td>shallow-marine limestone</td>
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<tr>
<td>allochthonous Ouachita thrust belt</td>
<td>Mississippian</td>
<td>Stanley Group</td>
<td>13,000w</td>
<td>deep-marine turbidites</td>
</tr>
<tr>
<td></td>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert-</td>
<td>18,560w</td>
<td>deep-marine cherts and clastics</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Arkansas Novaculite</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Womble Shale</td>
<td>15,500w</td>
<td>deep-marine shale</td>
</tr>
<tr>
<td>triangle-shaped deformed zone</td>
<td>Ordovician to Mississippian</td>
<td>Womble Shale-</td>
<td>14,250 a</td>
<td>deep-marine shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Stanley Group</td>
<td></td>
<td></td>
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<tr>
<td>autochthonous footwall</td>
<td>Mississippian</td>
<td>Caney-Goddard or Stanley Group</td>
<td>14,250a</td>
<td>shallow-marine to deep-marine sandstones and shales</td>
</tr>
<tr>
<td></td>
<td>Late Ordovician to early Mississippian</td>
<td>&quot;Bromide-Sycamore&quot; interval</td>
<td>22,500a</td>
<td>cherty limestones or limey cherts (transitional facies?)</td>
</tr>
<tr>
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<td>Ordovician</td>
<td>Simpson Group</td>
<td>12,300w</td>
<td>shallow-marine carbonates, sandstones, and limey shales</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(McLish-Tulip Creek)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td>Simpson (Joins-Oil Creek)</td>
<td>14,500p</td>
<td>shallow marine carbonate + clastic</td>
</tr>
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<td></td>
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</tr>
<tr>
<td></td>
<td>Ordovician-Cambrian</td>
<td>Arbuckle Group</td>
<td>22,500e</td>
<td>shallow marine carbonate</td>
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</table>
CROSS SECTION 6 (in thickest part of Ravia nappe southwest of Washita valley fault, northeast end of cross section)

<table>
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<th>AGE</th>
<th>FORMATIONS</th>
<th>INTERVAL SEISMIC VELOCITY (FT/SEC)</th>
<th>FACIES and/or DOMINANT LITHOLOGIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gulf Coastal Plain</td>
<td>Cretaceous</td>
<td>7,220w shallow-water limestone and sandstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td>basement nappe (hanging wall)</td>
<td>Middle Ordovician</td>
<td>Simpson Group (Oil Creek-McLish)</td>
<td>13,800w,p shallow-marine carbonates and sandstones</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Middle Ordovician</td>
<td>Simpson Group (Joins-Oil Creek)</td>
<td>17,800w,p shallow-marine carbonates and limey shales</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
<td>22,500e shallow-marine carbonates</td>
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</tbody>
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<table>
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<th>TERRAIN</th>
<th>AGE</th>
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<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<tbody>
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<tr>
<td>basement nappe</td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
<td>22,500e shallow-marine carbonates</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Pennsylvanian (Morrowan)</td>
<td>Springer Group</td>
<td>11,555w,p shallow-marine to transitional-depth sandstones and shales</td>
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<tr>
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<td>Mississippian (Chesterian)</td>
<td>Goddard</td>
<td>11,555w,p shallow-marine to transitional-depth sandstones and shales</td>
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<tr>
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<td>Mississippian (Meramac)</td>
<td>Caney Shale</td>
<td>13,800w,p shallow-to transitional-depth shale</td>
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<tr>
<td></td>
<td>Silurian to early Mississippian</td>
<td>Sylvan Shale - Woodford</td>
<td>13,800w,p shallow-marine to transitional-depth carbonates, cherts, sandstones, and shales</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Late Ordovician</td>
<td>Simpson Group (Bromide-Viola)</td>
<td>20,200w,p shallow-marine to deep-marine carbonates (carbonaceous)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Middle Ordovician</td>
<td>Simpson Group (Oil Creek-McLish)</td>
<td>13,800w,p shallow-marine carbonates and sandstone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Middle Ordovician</td>
<td>Simpson Group (Joins-Oil Creek)</td>
<td>17,800w,p shallow-marine carbonates and limey-shales</td>
<td></td>
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<tr>
<td></td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
<td>22,500e shallow-marine carbonates</td>
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CROSS SECTION 6 (in thin part of Ravia nappe southwest of Washita valley fault, northeast end of cross section)

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<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<td>7,220w shallow-marine carbonates</td>
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<td></td>
</tr>
<tr>
<td>basement nappe</td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
<td>22,500e shallow-marine carbonates</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Pennsylvanian (Morrowan)</td>
<td>Springer Group</td>
<td>11,555w,p shallow-marine to transitional-depth sandstones and shales</td>
<td></td>
</tr>
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<td></td>
<td>Mississippian (Chesterian)</td>
<td>Goddard</td>
<td>11,555w,p shallow-marine to transitional-depth sandstones and shales</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Mississippian (Meramac)</td>
<td>Caney Shale</td>
<td>13,800w,p shallow-marine to transitional-depth shales</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Silurian to early Mississippian</td>
<td>Sylvan Shale - Woodford</td>
<td>13,800w,p shallow-marine to transitional-depth carbonates, cherts, sandstones, and shales</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Late Ordovician</td>
<td>Simpson Group (Bromide-Viola)</td>
<td>20,200w,p shallow-marine to deep-marine carbonates (carbonaceous)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Middle Ordovician</td>
<td>Simpson Group (Oil Creek-McLish)</td>
<td>13,800w,p shallow-marine carbonates and sandstones</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Middle Ordovician</td>
<td>Simpson Group (Joins-Oil Creek)</td>
<td>17,800w,p shallow-marine carbonates and limey-shales</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
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Table 5.1: (continued)

CROSS SECTION 6 (inbetween frontal fault of Bryan small-scale salient and southern edge of Ravia nappe, central part of cross section)

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<th>FORMATIONS</th>
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<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<tr>
<td>Gulf Coastal Plain</td>
<td>Cretaceous</td>
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<td>7,220w</td>
<td>shallow-marine limestone and sandstone</td>
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<tr>
<td>autochthonous</td>
<td>Pennsylvanian (Atokan)</td>
<td>Dornick Hills Group</td>
<td>11,613w,p</td>
<td>shallow-water sandstones and shales</td>
</tr>
<tr>
<td></td>
<td>Pennsylvanian (Morrowan)</td>
<td>Dornick Hills Group</td>
<td>11,613w,p</td>
<td>shallow-water sandstones and shales</td>
</tr>
<tr>
<td></td>
<td>Pennsylvanian (Morrowan)</td>
<td>Springer Group</td>
<td>11,555w,p</td>
<td>shallow-marine to transitional-depth sandstones and shales</td>
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<tr>
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<td>Mississippian (Meramecian)</td>
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<td>13,800w,p</td>
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<td>Silurian to early Mississippian</td>
<td>Sylvan Shale-Woodford</td>
<td>13,80w,p</td>
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<td>Simpson Group (Bromide)-Viola</td>
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<td>Simpson Group (Joins-Oil Creek)</td>
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CROSS SECTION 6 (deepest central part of Bryan small-scale salient)

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<td>Mississippian</td>
<td>Stanley Group</td>
<td>13,000w</td>
<td>deep-marine turbidites and tuffs</td>
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<td>Ouachita thrust belt (top slice)</td>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert-Arkansas Novaculite</td>
<td>19,800w</td>
<td>deep-marine cherts and clastics</td>
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<td>Womble Shale</td>
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<td>Womble Shale</td>
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<td>deep-marine shale</td>
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<td>(third slice)</td>
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<td>Bigfork Chert-Arkansas Novaculite</td>
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<td>(bottom slice)</td>
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Table 5.1: (continued)

CROSS SECTION 6 (deepest central part of Bryan small-scale salient)

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<td>Mississippian (Meramecian)</td>
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<td>13,800 w,p</td>
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<td></td>
<td>Middle Ordovician-Mississippian</td>
<td>Simpson Group-Woodford</td>
<td>16,430 w,p</td>
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<td>Middle Ordovician</td>
<td>Simpson Group (basal Oil Creek)</td>
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<td>Cambrian-Ordovician</td>
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CROSS SECTION 6 (southwest end)

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<td>Ordovician? to Mississippian</td>
<td>Womble Shale?- Stanley Group (uncertain geometry)</td>
<td>14,250 w</td>
<td>deep-marine turbidite, shales and tuffs</td>
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<td>autochthonous footwall</td>
<td>Mississippian (Chesterian)</td>
<td>Goddard</td>
<td>11,555 w,p</td>
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<tr>
<td></td>
<td>Mississippian (Meramecian)</td>
<td>Caney Shale</td>
<td>13,800 w,p</td>
<td>shallow-marine to transitional-depth shales</td>
</tr>
<tr>
<td></td>
<td>Ordovician-Mississippian</td>
<td>Simpson Group-Woodford</td>
<td>16,430 w,p</td>
<td>shallow-marine/transitional carbonates, cherts and clastics</td>
</tr>
<tr>
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<td>Ordovician</td>
<td>Simpson Group (basal Oil Creek)</td>
<td>17,800 w,p</td>
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<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
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<td>shallow-marine carbonates</td>
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CROSS SECTION 7 (north end)

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<td>Stanley Group</td>
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<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert- Arkansas Novaculite</td>
<td>19,000 w</td>
<td>deep-marine cherts and clastics</td>
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<tr>
<td></td>
<td>Ordovician</td>
<td>Womble Shale</td>
<td>14,152 w</td>
<td>deep-marine shale</td>
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<td>Mississippian (Meramecian)</td>
<td>Caney Shale</td>
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<tr>
<td></td>
<td>Silurian to early Mississippian</td>
<td>Sylvan Shale- Woodford</td>
<td>13,800 w</td>
<td>shallow-marine/transitional carbonates, cherts and clastics</td>
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<td></td>
<td>Ordovician</td>
<td>Simpson Group-Viola</td>
<td>17,000 w</td>
<td>shallow-marine ls., ds., ss., sh.</td>
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<td>Arbuckle Group</td>
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Table 5.1: (continued)

CROSS SECTION 7 (deepest part of allochthon, at Ellet well)

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<td>Ouachita thrust belt</td>
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<td>Pennsylvanian (Morrowan)</td>
<td>Johns Valley Shale</td>
<td>13,566w</td>
<td>deep-water turbidites</td>
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<td>latest Miss. (Chesterian) to Penn.</td>
<td>Jackfork Sandstone</td>
<td>13,566w</td>
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<td>Mississippian</td>
<td>Stanley Group</td>
<td>12,133w</td>
<td>deep-water turbidites and tuffs</td>
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<tr>
<td>triangle-shaped deformed zone</td>
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<tr>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert-</td>
<td>19,000w</td>
<td>deep-marine cherts and clastics</td>
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<tr>
<td>autochthonous</td>
<td>Simpson Group</td>
<td>17,000p</td>
<td>shallow-marine/transitional</td>
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<td>Ordovician</td>
<td>Caney Shale</td>
<td>11,555p</td>
<td>shallow-marine to transitional-depth shales</td>
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<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
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CROSS SECTION 7 (south of Ellet well in center of "triangle-shaped deformed zone")

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<td>Ouachita thrust belt</td>
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<td>Pennsylvanian (Atokan)</td>
<td>Atoka ?</td>
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<td>Johns Valley Shale</td>
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<td>deep-marine turbidites</td>
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<td>latest Miss. (Chesterian) to Penn.</td>
<td>Jackfork Sandstone</td>
<td>13,566w</td>
<td>deep-marine turbidites</td>
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<td>Mississippian</td>
<td>Stanley Group</td>
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<td>deep-water turbidites and tuffs</td>
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<td>triangle-shaped deformed zone</td>
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<tr>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert-</td>
<td>19,000w</td>
<td>deep-marine cherts and clastics</td>
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<td>Simpson Group</td>
<td>17,000p</td>
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<td>Ordovician</td>
<td>Caney Shale</td>
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<td>22,500e</td>
<td>shallow-marine carbonates</td>
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Table 5.1: (continued)

**CROSS SECTION 7 (on horst block north of Sherman fault)**

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<td>Ouachita thrust</td>
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<td>Jackfork Sandstone</td>
<td>13,566w</td>
<td>deep-marine turbidites</td>
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<td>Mississippian</td>
<td>Stanley Group</td>
<td>12,133w</td>
<td>deep-water turbidites and tuffs</td>
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<td></td>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert: Arkansas Novaculite</td>
<td>19,800w</td>
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<td>footwall</td>
<td>Pennsylvania (Morrown-Atokan)</td>
<td>Dornick Hills Group</td>
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<td>Middle Ordovician</td>
<td>Simpson Group (Oil Creek)</td>
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<td>Cambrian-Ordovician</td>
<td>Arbuckle Group</td>
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**CROSS SECTION 7 (south end)**

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<td>Bigfork Chert</td>
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<td>deep-marine cherts and clastics</td>
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<td>Womble Shale</td>
<td>14,152w</td>
<td>deep-marine shale</td>
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<tr>
<td>footwall</td>
<td>Pennsylvania (Desmoinesian)</td>
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<td>Simpson Group (Oil Creek)</td>
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<td>Arbuckle Group</td>
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**CROSS SECTION 8 (southeast end)**

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<td>latest Miss. (Chesterian) to Penn. (Morrown)</td>
<td>Jackfork Sandstone</td>
<td>13,566w</td>
<td>deep-marine turbidites</td>
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<td></td>
<td>Mississippian</td>
<td>Stanley Group</td>
<td>12,133w</td>
<td>deep-water turbidites and tuffs</td>
</tr>
<tr>
<td></td>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert: Arkansas Novaculite</td>
<td>19,800w</td>
<td>deep-marine cherts and clastics</td>
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<td></td>
</tr>
<tr>
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<td>Orдовician</td>
<td>Womble Shale</td>
<td>14,152w</td>
<td>deep-marine shale</td>
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## Table 5.1: (continued)

### CROSS SECTION 8 (southeast end)(continued)

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<th>FACIES and/or DOMINANT LITHOLOGIES</th>
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<td>Mississippian (Chesterian)</td>
<td>Goddard</td>
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<td>shallow-marine to transitional-depth sandstones and shales</td>
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<td>Middle Ordovician to Miss. (Meramecian)</td>
<td>Simpson Group</td>
<td>16,430w,p</td>
<td>Shallow-marine to transitional-depth carbonates and clastics</td>
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<td>Arbuckle Group</td>
<td>22,500e</td>
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### CROSS SECTION 8 (through "triangle-shaped deformed zone")

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<td>Jackfork Sandstone</td>
<td>13,566w,p</td>
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</tr>
<tr>
<td></td>
<td>Mississippian</td>
<td>Stanley Group</td>
<td>12,133w,p</td>
<td>deep-water turbidites and tuffs</td>
</tr>
<tr>
<td>triangle-shaped deformed zone</td>
<td>Mississippian</td>
<td>Stanley Group</td>
<td>13,383w,p</td>
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<tr>
<td></td>
<td>Late Ordovician to early Mississippian</td>
<td>Bigfork Chert - Arkansas Novaculite</td>
<td>19,800w,p</td>
<td>deep-marine cherts and clastics</td>
</tr>
<tr>
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<td>Ordovician</td>
<td>Womble Shale</td>
<td>14,152w</td>
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<td>Simpson Group</td>
<td>17,800w,p</td>
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### CROSS SECTION 8 (on southeast side of Bryan fault)

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<td>12,133w,p</td>
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<td>Arkansas Novaculite</td>
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<td>Ordovician</td>
<td>Womble Shale</td>
<td>14,152w,p</td>
<td>deep-marine shale</td>
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<td>Miss. (Chesterian) to Penn. (Morrowan)</td>
<td>Goddard: Springer Group</td>
<td>11,555w,p</td>
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<td>Middle Ordovician to early Mississippian</td>
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<td>Simpson Group (Joins-Oil Creek)</td>
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Table 5.1: (continued)

### CROSS SECTION 8 (through "triangle-shaped deformed zone" northwest of Bryan fault)

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<td>Arkansas Novaculite</td>
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<td>Ordovician</td>
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<td>19,800w,p</td>
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<td>early Mississippian</td>
<td>Arkansas Novaculite</td>
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<td>Ordovician</td>
<td>Womble Shale</td>
<td>14,152w,p</td>
<td>deep-marine shale</td>
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<td>Goddard-</td>
<td>11,555w,p</td>
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<td>Springer Group</td>
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<td>Mississippian</td>
<td>Caney Shale</td>
<td>16,430w,p</td>
<td>shallow-marine to transitional-</td>
</tr>
<tr>
<td></td>
<td>Silurian to</td>
<td>Sylvan Shale-</td>
<td>13,800w,p</td>
<td>carbonates, cherts and clastics</td>
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<td>Woodford-Sycamore</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Late Ordovician</td>
<td>Simpson Gp</td>
<td>17,800w,p</td>
<td>shallow-marine carbonates and</td>
</tr>
<tr>
<td></td>
<td>(Joins-Oil Creek)</td>
<td>(Bromide)-Viola</td>
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### CROSS SECTION 8 (northwest end)

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<td>Miss. (Penn. (Morrowan))</td>
<td>Springer Group</td>
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<tr>
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<td>Silurian to</td>
<td>Sylvan Shale-</td>
<td>13,800w,p</td>
<td>carbonates, cherts and clastics</td>
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<td>Woodford-Sycamore</td>
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<td></td>
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<td>Simpson Gp</td>
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<td>(Joins-McLish)</td>
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<td>limey shales</td>
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## Cross Section 9 (Northwest End)

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<td></td>
<td>Fredericksburg Group</td>
<td>11,000w,p</td>
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<td></td>
<td></td>
<td>Paluxy Sand</td>
<td>7,273w,p</td>
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<td>Trinity Group</td>
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### Cross Section 9 (Northwest End) (continued)

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<td>19,800w,p</td>
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<td>(Join-Atoka)</td>
<td>(Join-McLish)</td>
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<td>15,789w,p shallow-marine limestones</td>
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<td>22,000w,p shallow-marine limestones</td>
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<td>Stanley Group</td>
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<td>Ouachita thrust belt</td>
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<td>Late Ordovician to early Mississippian</td>
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<td>Ordovician</td>
<td>Simpson Group (Joins-McLish) (uncertain geometry)</td>
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<td>Stanley Group</td>
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<td>Arbuckle Group</td>
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Table 5.2: Formation-top data for wells in Atoka, Bryan, Choctaw, Marshall, and Pushmataha Counties, Oklahoma.

Data provided by Oklahoma Geological Survey.

Only selected key wells are shown in Plates 4.1, 5.1 and 5.2.

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9744 2971 07S-12E-02 1980 12/13

9919 1835 07S-07E-11 1979 11/20

3819 05S-08E-15 1988/11/13

12525 3819 05S-08E-15 1988/11/13

7712 2351 -7033 -2144

12525 3819 05S-08E-15 1988/11/13
### Table 5.2: (continued)

Bryan County wells (continued)

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<th>Wells</th>
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<th>Hunton Gp. (cherty limestone)</th>
<th>Sylvan (green shale)</th>
<th>Viola Gp. (limestone)</th>
<th>Simpson Group</th>
<th>-Bromide (dense limestone)</th>
<th>-McClish (shale and limestone)</th>
<th>-Oil Creek (shale)</th>
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**Notes:**

- All depths are in feet.
- The depth values are approximate and may vary slightly due to rounding.
- The dates provided are the latest known analysis dates for each well.
- The table includes additional columns for well identifiers, formation names, and other relevant information.

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<th>SYCAMORE (CHERTY LIMESTONE)</th>
<th>WOODFORD (SHALE AND CHERT)</th>
<th>HUNTON G/ (CHERTY LIMESTONE)</th>
<th>SYLVAN (GREEN SHALE)</th>
<th>VIOLA G/ (LIMESTONE)</th>
<th>SIMPSON GROUP</th>
<th>-BROMIDE (DENSE LIMESTONE)</th>
<th>-MCCILSH (SHALE AND LIMESTONE)</th>
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**Notes:**

- The table includes additional columns for well identifiers, formation names, and other relevant information.
- The dates provided are the latest known analysis dates for each well.
- The table includes additional columns for well identifiers, formation names, and other relevant information.
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### Choctaw County wells

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### Marshall County wells

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Table 5.2: (continued)

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<td>McLish Lower</td>
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<td>Oil Creek Upper</td>
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305
### Table 5.2: (continued)

**Marshall County wells (continued)**

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| Ma1  | Sampson Group |  | Joins (limestone) | 1778 | 541 | -898 | -274 |
|      | Arbuckle Group (carbonates) |  | 1998 | 609 | -1120 | -341 |
|      | Timbered Hills Group |  | 3718 | 1134 | -2840 | -866 |
|      | -- Honey Creek? ["Brown Zone"] |  | 5420 | 1652 | 04S-04E-35 | 1986 |

| MA2  | 003561 | 3 | Cretaceous—Coastal Plain        | 945  | 288 | -225 | -69  |
|      | (--) --unconformity --> |  | 1890 | 576 | -1170 | -357 |
|      | Springfield Group (shale)—foreland-facies synorogenic |  | 6984 | 2129 | -6264 | -1910 |
|      | Caney (shale)—foreland-facies synorogenic |  | 7115 | 2169 | -6395 | -1950 |
|      | Woodford (shale) |  | 7334 | 2242 | -6634 | -2023 |
|      | (--) --unconformity --> |  | 7933 | 2425 | -7233 | -2205 |
|      | Viola Gp. (limestone) |  | 8565 | 2611 | -7845 | -2392 |
|      | --Bromide (dense limestone) |  | 8848 | 2698 | -8128 | -2478 |
|      | --Bromide First Sand (sandstone) |  | 8882 | 2708 | -8162 | -2488 |
|      | --Bromide Second Sand (sandstone) |  | 9030 | 2747 | -8280 | -2527 |
|      | --Bromide Basal (sandstone) |  | 9135 | 2783 | -8415 | -2566 |
|      | --McLish (shale and sandstone) |  | 9220 | 2811 | -8500 | -2591 |
|      | --McLish Sand (sandstone) |  | 9618 | 2932 | -8308 | -2713 |
|      | --Oil Creek (shale) |  | 9730 | 2966 | -8010 | -2747 |
|      | --Oil Creek Basal (sandstone) |  | 10444 | 3184 | -9724 | -2965 |

| MA3  | 002134 | 09520380 | 395 | Atoka—foreland-facies-synorogenic | 8180 | 2494 | -7390 | -2253 |
|      | Davis (lower Atoka) |  | 9830 | 2986 | -9010 | -2747 |

| MA4  | 002120 | 1 | Atoka—foreland-facies-synorogenic | 1525 | 465 | -781 | -238 |
|      | (--) --unconformity --> |  | 5065 | 1544 | -4321 | -1317 |
|      | Davis Zone (lower Atoka) |  | 5985 | 1819 | -5221 | -1592 |

| MA5  | 002906 | 3-A | Goddard (shale) ["False Caney"] | 2372 | 723 | -1766 | -538 |
|      | Caney (shale)—foreland-facies synorogenic |  | 2582 | 787 | -1976 | -602 |
|      | Sycamore (limestone) |  | 2876 | 877 | -2270 | -692 |
|      | Woodford (shale an chert) |  | 3988 | 945 | -2492 | -760 |
|      | Hunton Group (cherty limestones) |  | 3345 | 1020 | -2739 | -835 |
|      | Chimney Hill subgroup (cherty limestones) |  | 3668 | 1100 | -3002 | -915 |
|      | Sylvan (green shale) |  | 3642 | 1110 | -3036 | -926 |
|      | Viola Gp. (limestone) |  | 3939 | 1192 | -3304 | -1007 |
|      | Sampson Group |  | 4480 | 1566 | -3874 | -1181 |
|      | --Bromide (dense limestone) |  | 4735 | 1450 | -4150 | -1265 |
|      | --Bromide First Sand (sandstone) |  | 4865 | 1483 | -4259 | -1298 |
|      | --Bromide Second Sand (sandstone) |  | 5012 | 1528 | -4406 | -1343 |
|      | --McLish (sandstone) |  | 5530 | 1686 | -4924 | -1504 |
|      | --Oil Creek (limestone and shale) |  | 5624 | 1715 | -5018 | -1530 |
|      | --Oil Creek Upper (sandstone) |  | 6082 | 1854 | -5476 | -1670 |
|      | --Oil Creek Intermediate (shale) |  | 6196 | 1889 | -5590 | -1704 |
|      | --Oil Creek Basal (sandstone) |  | 6256 | 1907 | -5660 | -1723 |
|      | --Joins (limestone) |  | 6480 | 1975 | -5854 | -1755 |
|      | Arbuckle Group (carbonates) |  | 6605 | 2014 | -5990 | -1829 |
|      | --Cool Creek |  | 9050 | 2759 | -8444 | -2574 |

| MA6  | 003061 | 09500037 | 1 | Cretaceous—Coastal Plain          | 0    | 0    | 805   | 245  |

<p>|      | Hoxbar (clastic)—foreland-facies synorogenic/post-orogenic |  | 470 | 143 | 335 | 102 |
|      | Deese (clastic)—foreland-facies synorogenic |  | 2960 | 902 | -2155 | -657 |</p>
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### Table 5.2: (continued)

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Chapter Six

Conclusions and suggestions for future research

Questions and problems in Ouachita geology

This dissertation examines a number of important questions pertaining to Ouachita geology. The primary question concerns the relative timing of the autochthonous structures of the southeastern part of the Arbuckle uplift with respect to the allochthonous structures of the Ouachita thrust belt where they intersect in southeastern Oklahoma (Figure 4.1 and Plate 4.1). Did the Tishomingo-Belton anticline form first (i.e., pre-existing buttress) and the Ouachita allochthon thrust over and around, or did the curvature of the Ouachita thrust front form as a result of post-thrust uplift?

Another important question concerns the along-strike changes in structural style within the Ouachita Mountains and adjacent Arkoma basin (Figure 4.1). More specifically, what implications for the kinematics of the Ouachita salient (Figure 1.1) can be inferred from the along-strike differences in structural style within the center of the salient (Ouachita Mountains)? One particularly important along-strike difference of structural style is the eastward change from imbricate thrust belt in the western Frontal Ouachita thrust belt, to broad folds in the eastern Frontal Ouachita thrust belt (Figure 4.1). Another important along-strike difference is the eastward change from north-vergent structures in the western Frontal and Central Ouachita thrust belts, to predominantly south-vergent structures in the southern part of the eastern Frontal Ouachita, and most of the eastern Central Ouachita thrust belts (Figure 4.1).

Yet another important question concerns the lack of extensive imbricated Middle Ordovician-Morrowan shallow-water facies strata in the frontal part of the Ouachita allochthon west of the Ti Valley fault at Black Knob Ridge in the southwesternmost Ouachita Mountains (Figure 4.1 and Plate 4.1) and around the southeastern ends of the Tishomingo-Belton anticline and Muenster arch, and in the Bryan small-scale salient (Plate 4.1). Interpretation of two deep wells in the Broken Bow uplift, to the east of the Tishomingo-Belton anticline, show the basal décollement of the Ouachita allochthon directly above Middle Ordovician Simpson Group.
(Leander and Legg, 1998; Denison, 1989). What happened to the missing Middle Ordovician-Morrowan shallow-water facies strata?

Several methods were used to examine the perplexing questions of Ouachita geology. On a large scale, the stratigraphy and structures of the Ouachita allochthon and the adjacent foreland along the Ouachita salient (Figure 1.1 and Plate 2.1) were studied to gain a regional perspective. On a smaller scale, structures within the Ouachita Mountains and Arkoma basin were studied to determine the along-strike changes in structural geometry and kinematics of the center of the Ouachita salient. On the smallest scale, nine new interpreted cross sections (cross sections 1 through 9, Plates 5.1 and 5.2), based on seismic reflection profiles, were created to show structures along the southeastern part of the Tishomingo-Belton anticline, part of the Ardmore basin, and the southeastern Muenster arch (Plate 4.1). Interpreted structures shown in cross sections 1 through 9 are constrained further by well data, published surface and subsurface geologic maps, cross sections, top-of-Arbuckle structure contour maps, and conclusions derived from the larger scale studies of the Ouachita salient.

**Observations/conclusions**

Conclusions drawn in this dissertation fall into three categories: conclusions derived from examination of the entire Ouachita salient, conclusions derived from detailed study of the structural geology of the Ouachita Mountains and adjacent Arkoma basin, and conclusions derived from study of the intersection of the Ouachita thrust front with the southeastern ends of the Arbuckle uplift and Muenster arch (based on newly interpreted cross sections).

**Regional stratigraphy**

On a large scale, the examination of the Paleozoic stratigraphy of deep-water facies of the Ouachita orogen and the shallow-water facies of the foreland on the north and west indicate that two types of unconformities interrupt the strata across the region. The first type of unconformity is laterally continuous and is nearly parallel with underlying strata. In the chronostratigraphic correlation chart shown in Plate 2.2, examples of this type of unconformity are shown clearly as “horizontal” gaps of missing
strata which pinch out laterally. For the most part, these “horizontal” unconformities were formed by a process that affected a large area, such as a eustatic change in relative sea level. The second type of unconformity is characterized by clearly defined truncation of underlying strata which obliquely intersect the unconformity. Examples of this type of unconformity are illustrated in the chronostratigraphic correlation chart shown Plate 2.2 as areas where the margins of time gaps are steep and usually represent a large gap in geologic time (such as in the center of the Llano uplift or Muenster arch). These unconformities, characterized by localized absence of stratigraphy that represent a large amount of unrecorded geologic time, were formed by tectonic uplift.

**Mechanical stratigraphy of the Ouachita Mountains and Arkoma basin**

On a smaller scale, the effect of along-strike changes in mechanical stratigraphy on structural style was derived through study of the structural geology of the Ouachita Mountains and the adjacent Arkoma basin. The Paleozoic succession that crops out in the Ouachita Mountains is subdivided into lower, middle, and upper mechanical stratigraphic units. The lower unit is a composite weak (ductile) layer comprised of an Upper Cambrian-lower Mississippian succession of carbonaceous very low-grade to non-metamorphic shales, fine-grained sandstones, cherts, and novaculites (Collier Formation through basal Stanley Group (Stanley Formation in Arkansas) shale). The middle unit is the lower and middle part of the Mississippian Stanley Group (mostly shale), and the upper unit is comprised of upper Mississippian-Pennsylvanian (Atokan) interbedded sandstone and shale turbidites which form a composite stiff layer (upper Stanley Group-deep-water facies Atoka Formation).

Tight short wavelength folds cut by numerous small-displacement faults are common where strata of the lower mechanical stratigraphic unit crops out (such as Broken Bow and Benton uplifts). In contrast, broad synclines and intervening narrow, fault-cored, anticlines (cuspate-lobate folds) are common where the upper mechanical stratigraphic unit crops out (such as in much of the western part of the Central Ouachita thrust belt and in the eastern Southern Ouachita thrust belt). In between, the Stanley Group shale of the middle mechanical stratigraphic unit evidently serves as a
wavelength transition (and/or volume compensation) zone, and folds contort to fill the gaps between the base of the upper unit and top of the lower unit.

Several detachment levels are evident within the Paleozoic succession of the Ouachita Mountains and Arkoma basin. South of the Ouachita thrust front (Dutch Creek-Ross Creek fault) in the eastern part of the Frontal Ouachita thrust belt, the Y-City fault evidently marks the location of a small-displacement south-vergent delamination surface which coincides with the top of the Morrowan Johns Valley Shale. The predominance of south-vergent and south-overturned folds and faults south of the Y-City fault are underplated beneath the Y-City (John Valley detachment). The very thick, broad folds of Atoka Formation (and overlying strata) in thrust slices to the north of the Y-City fault (>9 km maximum stratigraphic thickness) is evidence that the Johns Valley detachment formed in response to a very thick succession of overlying Atokan strata.

In the subsurface of the southeastern Arkoma basin, north of the Ross Creek fault, the most evident detachments in the COCORP seismic reflection profile (Plate 4.4) are one in the lower Atoka Formation and one in the middle Atoka Formation. Cross sections of the Arkoma basin near the Ouachita thrust front (Figure 4.8) show that a detachment in the lower part of the lower Atoka Formation is common in the subsurface of the southwestern and south-central parts of the Arkoma basin and Ouachita thrust front. Other detachments in the southwestern Arkoma basin are in the Springer Group (Morrowan) beneath the thin Wapanucka Limestone and the basal Spiro Sandstone member of the Atoka Formation, and at the base of the Woodford Formation (coincides with an upper Devonian unconformity surface).

Outcrop of many thin thrust-fault imbricates of Morrowan Springer Group through lower Atoka Formation strata east of the frontal thrust fault (Choctaw fault) in the western part of the Frontal Ouachita thrust belt is evidence that the lower Atoka detachment originally extended farther east. The eastward extension of the detachment was displaced by the Choctaw fault and other faults within the western Frontal Ouachita thrust belt. Evidently, the eastward along-strike change from imbricate thrust belt in the western part to broad folds in the eastern part of the Frontal
Ouachita thrust belt (south of the Dutch Creek-Ross Creek faults) resulted from an abrupt eastward increase in thickness of the Atoka Formation.

In the Central and Southern Ouachita thrust belts, the structural geology is more complex. In some areas, according to various interpretation (Arbenz, 1989d, in contrast to Titus and Legg, 1995), several detachments may also be considered disharmonic boundaries. In either classification, folds in strata on either side of detachment/disharmonic boundary differ greatly in wavelength. The most laterally continuous and most readily identifiable of the detachments (disharmonic boundaries) is in the upper part of the Stanley Group and separates broad folds above from underlying much tighter folds. This “upper Stanley detachment” is recognized on the margins of broad synclines that are cored with Morrowan and Atokan strata. Another detachment in the lower part of the Stanley Group (“lower Stanley detachment”) encircles the exposed pre-Mississippian strata in the Ouachita Mountains. Other detachments recognized in the core areas of the Broken Bow and Benton uplifts include: Mazarn Shale 1 (near top of Crystal Mountain Sandstone), Mazarn Shale 2 (near base of Blakely Sandstone), and in the Womble Shale (above Blakely Sandstone). These detachments are laterally discontinuous and are cut by numerous faults.

The Mazarn Shale and Womble Shale “detachments” are clearly defined on geologic maps of the eastern Benton uplift region of the eastern Central Ouachita thrust belt (Plate 4.5) (Haley and Stone, 1990, 1991, 1993, 1994). In the Jessieville nappe of the Benton uplift, the Mazarn and Womble detachments are tightly folded and may be disharmonic boundaries. Map patterns in Haley and Stone (1990, 1991, 1993, 1994) show that farther east across the Alum Fork nappe and southern part of the Paron nappe, Mazarn and Womble detachments have larger displacement. The Mazarn Shale detachment coincides with the Alum Fork thrust fault that juxtaposes Lower–Middle Ordovician Mazarn Shale against underlying Upper Ordovician Bigfork Chert through Mississippian Stanley Group. On the basis of Haley and Stone (1990, 1991, 1993, 1994) the north-dipping Paron fault that separates the Paron nappe on the north, from the Alum Fork nappe on the south coincides with a detachment in
the Womble Shale (near the top of the Blakely Sandstone) (Plate 4.5). Outcrop of “southern facies” Arkansas Novaculite (similar to the type that crops out southeast of the Alum Fork nappe) north of the Paron fault, is evidence that the Paron fault is a north-dipping, north-vergent décollement that drapes over the uplifted core of the central Alum Fork nappe.

**Intersection of allochthonous structures of the Ouachita thrust belt and autochthonous structures of the Arbuckle uplift**

On a more detailed scale, new interpreted cross sections (based largely on seismic reflection profiles) constructed across the intersection of the southeastern Southern Oklahoma aulacogen (Arbuckle uplift) and the Ouachita thrust belt show evidence that transitional- or deep-water facies Upper Ordovician-lower Mississippian strata are located beneath the Ouachita allochthon between the Arbuckle uplift and the Broken Bow uplift. In general, seismic reflectors characteristic of foreland facies are traceable for a short distance east of the Ouachita thrust front beneath the basal décollement. East of an interpreted down-to-the-east normal fault (Bengalia fault), the interpreted Upper Ordovician-lower Mississippian “Bromide-Sycamore” seismic interval is more similar in appearance to the deep-water facies Upper Ordovician-lower Mississippian Bigfork Chert-Arkansas Novaculite interval in the overlying Ouachita allochthon, than to the shallow-water Bromide-Woodford/Sycamore interval west of the Bengalia fault. This similarity of seismic reflector characteristics suggests a lithologic similarity. Thickening of the Simpson Group (Suhm, 1997), and increased carbon content and graptolites in the overlying lower part of the Viola Group (Finney, 1986), indicate deepening along the southeastern margin of the Arbuckle uplift and farther southwest within the Ardmore basin.

In the autochthon, the reflectors of the “Bromide-Sycamore” interval are traceable for a short distance east of the Bengalia fault and end abruptly at an interpreted north-south trending large displacement down-to-east normal fault (possibly a growth fault) (Plate 5.1). In part of one cross section (cross section 3 in Plate 5.1), the “Bromide-Sycamore” reflector interval is absent and is evidently truncated beneath the basal décollement of the Ouachita allochthon. Because there is
no evidence of imbrication of strata equivalent to the “Bromide-Sycamore” interval farther west along the Ouachita thrust front; the basal décollement evidently coincides with a pre-thrust unconformity. The abundance of exotic clasts of mostly foreland facies rocks incorporated in the Morrowan Johns Valley Shale to the northeast in the western Ouachita Mountains suggests that the unconformity is late Morrowan in age.

Cross sections of the Bryan small-scale salient (Plate 4.1), located southwest of the Arbuckle uplift and north of the southeastern end of the Muenster arch, also show evidence that the basal décollement of the Ouachita allochthon coincides with a pre-thrust unconformity. Interpreted seismic reflection profiles show that the base of the Ouachita allochthon overlies progressively older autochthonous strata towards the deep central part of the Bryan small-scale salient (cross sections 7 and 8 in Plate 5.2). In the deepest part of the salient, the allochthonous strata are interpreted to rest upon upper Simpson Group (cross section 7 in Plate 5.2) or upper Arbuckle Group (cross section 8 in Plate 5.2).

Because pre-Morrowan shallow-water facies strata are not imbricated within the Bryan small-scale salient, the absence of pre-Mississippian strata in the autochthon beneath the base of the Ouachita allochthon in the deep central part of the salient is also consistent with a facies transition zone coincident with the basal décollement. Lack of Upper Ordovician-lower Mississippian autochthonous strata in two deep wells in the Broken Bow uplift (Leander and Legg, 1988; Denison, 1989) is evidence that the basal décollement the southwestern part of the Ouachita Mountains either coincides with a pre-thrust unconformity or with an abrupt increase in shale content (facies change) in the Middle Ordovician. Lack of pre-Middle Ordovician (pre-Womble Shale) strata in outcrop (or in the subsurface) in the Black Knob Ridge and Potato Hills regions west of the Broken Bow uplift favors the abrupt facies change interpretation. Furthermore, a palinspastically restored regional cross section places the Womble Shale and overlying strata of the Potato Hills above the Broken Bow uplift (on the basis of cross section D-D’ in Arbenz, 1989a).

Finally, the new interpreted cross sections (Plates 5.1 and 5.2) constructed as part of this dissertation show that the strata of the Ouachita allochthon are greatly
deformed internally on the flanks of the Arbuckle uplift and within the Ardmore basin. The basal décollement is bent sharply over several basement faults above the southeast-plunging nose of the Tishomingo-Belton anticline (Plates 4.1 and 5.1). The décollement is bent most sharply over the Sulfur fault. Farther southwest, in the northwestern part of the Bryan small-scale salient, several stacks of northwest-vergent Womble Shale-Stanley Shale (Middle Ordovician-lower Mississippian) are folded and cut by divergent reverse-offset faults (cross section 6 in Plate 5.2). Hardie (1990) suggests that these structures are “out-of-syncline” pop-up structures formed as a result of post-thrust folding within the core of the underlying autochthonous Ardmore basin (Plate 4.1). However, the pop-up structures could also have formed as a result of lateral constriction between two adjacent pre-thrust structural highs (Arbuckle uplift and Criner Hills uplift and Muenster arch) (Plates 3.4 and 4.1). An interpreted Morrowan-age pre-thrust unconformity coincident with the basal décollement on the southeastern flank of the Arbuckle uplift (especially evident in cross section 2, Plate 5.1) indicates a pre-existing structural high north of the Ardmore basin. South of the Ardmore basin, Morrowan- and Desmoinesian-age unconformities in the southeastern Muenster arch (cross sections 8 and 9, Plate 5.2) and sediment-dispersal patterns within the Atokan succession in the Marietta basin and northern Fort Worth basin (Cleaves, 1996) show that the Muenster arch and Criner Hills uplift were structural highs from the Morrowan until the early Desmoinesian. However, the thick Desmoinesian succession in the southeastern Muenster arch (as shown in the footwall southwest of the Ouachita thrust front in cross section 8, Plate 5.2), and the Desmoinesian strata in the northwestern part of the Ardmore basin (west of the Bryan small-scale salient) indicate that the Muenster arch and Ardmore basin began to subside in the Desmoinesian. Stratigraphy of the Morrowan-Desmoinesian succession of the Muenster arch-Ardmore basin-Arbuckle uplift and adjacent areas favor the assertion that the pop-up structures within the allochthonous strata of the northwestern Bryan small-scale salient formed primarily as a result of lateral constriction between two adjacent pre-thrust structural highs. However, an unconformity within the Virgilian succession (base of Collings Ranch Conglomerate and Vanoss Group, Ham
and others, 1954) across the western Arbuckle uplift (Plate 3.9), indicates that the Arbuckle uplift continued to be uplifted after cessation of the northwestward translation of the Ouachita allochthon; therefore, some amount of post-thrusting structural movement accentuated the curvature of the Bryan small-scale salient and Tishomingo small-scale recess (Plates 3.4 and 4.1).

**Suggestions for future research**

Ouachita geology is very complex and full of interesting structures. Future research that would greatly benefit the understanding of Ouachita geology is briefly stated in the next few sentences. Before large-scale regional kinematically balanceable cross sections can be constructed, numerous small scale cross sections need to be constructed across the Frontal Ouachita thrust belt to precisely and accurately examine the transition between the imbricate zone on the west and the broad folds farther east. Also, more study is needed to unravel the complicated structures of the Broken Bow and Benton uplifts. Large-scale cross sections through the Broken Bow and Benton uplifts lack the precision necessary to produce regional kinematic models. Finally, a few deep wells between the Arbuckle and Broken Bow uplifts would determine if autochthonous deep-water facies strata is located beneath the base of the Ouachita allochthon. An integrated synthesis of a variety of detailed research is necessary to put to rest the “sacred cows” (Arbenz, 1984) of Ouachita geology.
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Sources of unpublished data:
Formation-top data for selected wells in Atoka, Bryan, Choctaw, Marshall, and Pushmataha Counties, southeast Oklahoma, provided by the Oklahoma Geological Survey. Code numbers, well names and locations are listed in Table 5.2.

Formation-top data (copy of scout ticket) for Sohio Natural Resources No. 1 Taylor well supplied by Neil Suneson of the Oklahoma Geological Survey.

Well data for Mobil Oil Corp. No. 1 Col Clazier well supplied by Robert O. Fay of the Oklahoma Geological Survey (personal communication).

Unpublished cross sections provided by J. Kaspar Arbenz.

Proprietary seismic reflection profiles (for cross sections 1-9 in Plates 5.1 and 5.2).
Vita

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List of publications and abstracts:


Plate 2.1: Generalized pre-Mesozoic geologic map of the south-central and southeastern part of the USA showing large-scale and selected small-scale tectonic features of the Ouachita orogen, southern Appalachian orogen, adjacent foreland to the north and west, and the Texarkana platform.

References: Ham and others, 1954; Morris, 1974; Nicholas and Rozendaal, 1975; Brewer, 1982; Gilbert, 1982; Johnson and others, 1988; Arbent 1989a,b; Hatcher, and others, 1989; Nicholas, 1989; Thomas, 1989; Thomas and others, 1989; Cleaves, 1996; Cooper, 1995.

ROCK TYPES
- Precambrian to early Mississippian (Meramecian) -- includes: Precambrian igneous, meta-igneous, meta-sedimentary basement; Cambrian igneous rocks; Late Precambrian to early Mississippian shallow-water shelf carbonates, sandstones, and shales.
- Cambrian to early Mississippian/Meramecian -- includes: Cambrian/Late Ordovician deep-marine carbonaceous shales, distal turbidites; and Late Ordovician to early Mississippian cherts, novaculites, sandstones, and shales.
- Mississippian (Meramecian) to Permian (Ochoan) -- includes: Mississippian to Virgilian proximal turbidites (syn-orogenic clastic wedges) adjacent to Ouachita and Appalachian orogens; Desmoinesian to Wolfcampian clastic rocks and carbonates; and Wolfcampian carbonates, evaporites, and red beds in west Texas and western Oklahoma; and Desmoinesian to Wolfcampian carbonates, sandstones, and shales within the Texarkana platform.

TIME LINES
- Mississippian (Meramecian) -- includes: Cambrian-Late Ordovician deep-marine carbonaceous shales, distal turbidites; and Late Ordovician to early Mississippian cherts, novaculites, sandstones, and shales.
- Mississippian (Meramecian) to Permian (Ochoan) -- includes: Mississippian to Virgilian proximal turbidites (syn-orogenic clastic wedges) adjacent to Ouachita and Appalachian orogens; Desmoinesian to Wolfcampian clastic rocks and carbonates; and Wolfcampian carbonates, evaporites, and red beds in west Texas and western Oklahoma; and Desmoinesian to Wolfcampian carbonates, sandstones, and shales within the Texarkana platform.

ABBREVIATIONS
- A = Arbuckle anticline
- BB = Broken Bow uplift
- BFZ = Frontal fault zone
- H = Hunton arch
- SB = Sherman fault block
- T-B = Tishomingo-Belton anticline

FAULTS
- Late Paleozoic structures produced during orogenesis:
  - normal fault (ticks on down-thrown side)
  - normal fault (ticks on down-thrown side)
  - high-angle reverse fault (barbs on hanging wall side)
- Cratonic basement structures (northwest of Ouachita and Appalachian thrust fronts):
  - normal fault (ticks on down-thrown side)
  - overturned thrust fault (barbs on hanging wall side)
Plate 2.2: Chronostratigraphic correlation charts for the Paleozoic shallow-water facies strata of the foreland adjacent to the Ouachita orogen.


Plate 2.3: Chronostratigraphic correlation chart of Paleozoic formations for the Ouachita orogen and adjacent foreland showing correlation between shallow-water "foreland facies" strata and deep-water "Ouachita facies" strata.

The chart above shows strata in approximately restored locations and facies transitions between shallow-water and deep-water facies strata. The foreland facies strata is broadly subdivided into Late Cambrian-Late Ordovician platform carbonates and lesser sandstones and shales (Arbuckle-Simpson-Viola-Sylvan); latest Ordovician-Early Silurian platform carbonates / sandstones and shales ( ecosystem); and Mississippian-Atokan syn-orogenic clastic wedge turbidites (Stanley Group/Formation-Atoka Formation).


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Cooper (1995)
Visher (1996)
Finney (1997)
Suhm (1997)

abbreviations
M = Mississippi
S = Shale
F = Limestone
B = Sandstone
unconformities
MESOZOIC/ERODED
DEPOSITIONAL HESSIAN
MISSORIAN
"FALSEVILLE"
"SOUTHERN"
"HOODORF-DUNNAGE"
"FRISCO"
"PENTERS"
"STABLING-HARAGAN-BOIS D'ARC"
"COCHRANE"
"VIOLA"
"MOLISH"
"SAUK-TIPPECANOE-SIMPSON"
Note: Unconformities (in quotes) are informally named for formation or age of sedimentary sequence above unconformity.
Plate 2.4: Spatial and temporal variation in dominant facies of the Middle Ordovician Simpson Group of Oklahoma and Arkansas. Maps A-G show facies distributions and formation correlations for successive, discrete time intervals that begin in early Whiterockian (Armourian) and end in early Mohawkian (Caradocian). Map H is an isopach map of the middle part of the Simpson Group. Large scale regional structures and locations of proprietary seismic profiles are shown. In maps A-G, inferred paleocurrents are illustrated, and all maps show the position of the frontal fault of the late Paleozoic Ouachita orogen for geographic reference. All maps also show the approximate location of the Late Cambrian-Middle Ordovician (at earliest) shelf edge as determined from seismic and gravity anomaly data (Lilie and others, 1989; Kruger and Keller, 1990; Keller and others, 1999; Mickus and Keller, 1992). Thomas (1991) incorporated other data and shows a shelf edge map where a north-east trending, rift-parallel margin intersects a northwest-trending, transform-parallel margin in southeastern Oklahoma. In general, the “sandstone” bodies of the Simpson Group illustrated in maps A through E (Calico Rock Ss. through lower St. Peter Ss./McLish Fm.) are elongated northeast-southwest with paleocurrents that suggest a northeastern source area. The sandstone bodies appear to be deltaic-type fans dispersed laterally to the northwest and southeast along the northern sides of the Ardmore-Anadarko basins. The sandstones of this part of the Simpson Group are quartz-rich and have a possible source area as far north as Michigan (from Suhm, 1997). The sandstone bodies of the Tulip Creek (and upper St. Peter Ss.) and Bromide Formations (maps F and G) are also northeast-elongated and appear to be sediment fans that flow towards the Ardmore-Anadarko basins. The Tulip Creek-Bromide sandstones are more calcareous than the St. Peter Ss. and the underlying older Simpson Group sandstones. The up-section increase in limestone content of the Tulip Creek-Bromide strata indicates an abrupt relative sea level rise (and, or subsidence) of the Oklahoma-Arkansas shelf (especially the Ardmore-Anadarko trough).- Hypothetical transgression of deep-water facies Womble Shale over the older platform strata is shown in maps E-G. Abrupt increase in organic carbon content in the lower part of the Viola Group which overlies the top of the Simpson Group suggests rapid relative subsidence of the Ardmore-Anadarko trough and the adjacent shelf. References: Leander and Legg (1988), Denison (1989), Finney (1997), Suhm (1997).
Part A is a regional geologic map of part of northern Texas and southern Oklahoma showing location of large-scale structures. Also shown is location of Waco uplift and location of cross section illustrated in Part B. Map from Ham and others (1954), Arbenz (1989e), Thomas and others (1989), and Cleaves (1996).

Part B is a simplified cross section of the Waco uplift based on seismic data and data from the Shell No. 1 Barrett well (illustrated in Part C). From Nicholas and Rozendal (1975).

Part C shows the generalized Shell No. 1 Barrett well lithologic column. Also shown are parts of the column where radiometric ages were determined and where core samples were examined. To the left of the column, 4 possible stratigraphic and structural interpretations for the lithologic column are listed (1,2,3, and 4). From Ham (1973), Nicholas and Rozendal (1975), Fay (1985e,i), Denison (1989), Ethlington and others (1989), and Suhm (1996).

Model 1
Interpreted stratigraphy and structure

Model 2
Interpreted stratigraphy and structure

Model 3
Interpreted stratigraphy and structure

Model 4
Interpreted stratigraphy and structure

Model 1 (left) shows an approximate restored Waco uplift cross section based on stratigraphic and structural interpretation number “1” shown in Part C (above). Model 1 suggests that the shallow-water carbonate strata penetrated in the Shell No. 1 Barrett well are the same as those penetrated by two deep wells farther north along the same trend as the Waco uplift (Hunt-Neely and Sohio-Weyerhaeuser, see Leander and Legg, 1988; and Denison, 1989). Model 1 suggests an abrupt upsection transition from Cambrian-Ordovician carbonates to Ordovician-Mississippian chert and shale strata within the Waco fault block.

Model 2 (left) shows an approximate restored Waco uplift cross section based on stratigraphic and structural interpretation number “2” shown in Part C. Model 2 suggests that a basal detachment fault operates along an unconformity at the top of the Cambrian-Devonian carbonate strata.

Interpreted cross sections for Models 3 and 4 (far left) based on stratigraphic and structural interpretations numbers “3” and “4” shown in Part C (above). An approximate restored Waco uplift cross section is also shown (left). Models 3 and 4 suggest that the sandstones and shales above the carbonate succession are autochthonous. Both models require that the detachment fault northwest of Fault A was above the level of the current base of the Mesozoic and eroded prior to deposition of Mesozoic formations.
Plate 2.6: Paleozoic stratigraphy of west Texas and western Oklahoma. Part A shows a stratigraphic cross section of the western part of the Anadarko basin (north of the Amarillo uplift) (Johnson and others, 1988). Part B shows a stratigraphic cross section of the Palo Duro basin (south of the Amarillo uplift) (Johnson and others, 1988). Dominant lithologies, stratigraphic correlations, and time lines are illustrated. Part C is a generalized geologic map of the region showing major tectonic features and locations of cross sections (Arbenz, 1989e; Thomas and others, 1989).
Plate 3.1: Generalized pre-Mesozoic geologic map of the south-central and southeastern part of the USA showing selected large-scale tectonic features of the Ouachita orogen, southern Appalachian orogen, adjacent foreland to the north and west, and the Texarkana platform. Also shown is boundary of detailed study area and location of proprietary seismic profiles.

References: Ham and others, 1954; Morris, 1974; Nicholas and Rozendal, 1975; Brewer, 1982; Gilbert, 1982; Johnson and others, 1988; Arbenz, 1989a, b; Hatcher and others, 1989; Nicholas, 1989; Thomas, 1989; Thomas and others, 1989; Cooper, 1995; Cleaves, 1996.
Plate 3.2: Pre-Mesozoic paleogeologic map and cross sections of the Mississippi Valley graben and adjacent areas showing large-scale tectonic features. Time lines shown in the map and cross sections above are approximately located and shown to illustrate structures within the several parts of the graben. The northeast-striking faults that mark the boundaries of grabens A and B and the northeast-trending folds within, appear to be en échelon. This apparent en échelon offset suggests late Paleozoic, left-lateral oblique offset of boundary faults of the Mississippi Valley graben.

Plate 3.3: Comparison of aeromagnetic and Bouguer gravity maps of part of the Mississippi embayment and adjacent regions. Also shown are locations of several large-scale structures, Paleozoic faults and folds of the Mississippi Valley graben, and thickness of Mesozoic strata within the Mississippi embayment. See above for further description.

A: Aeromagnetic total field intensity map of part of the Mississippi embayment region. Contour interval (100 γ) from Kane and others (1981). Faults and fold axes shown illustrate offset and deformation of pre-Mesozoic strata (from Thomas, 1991). Grabens A, B, and C are fault-bounded subdivisions of the Mississippi Valley graben (see Plate 3.2). Paleozoic strata are thickest within the graben A part of the Mississippi Valley graben. The location of seven interpreted mafic-igneous plutons, coincident with aeromagnetic positive (and low-negative) “high” regions from Hildenbrand (1985). Many lineaments that cross-cut the map are mostly oriented with northeast trends parallel with the Mississippi Valley graben, or northwest trends that are nearly perpendicular with the graben. Exposure of Mesozoic igneous rocks along the western margin of the Mississippi embayment southwest of the Newport pluton (off the map) suggests that the plutons beneath the Mesozoic cover within the Mississippi embayment are Mesozoic age intrusive bodies. Relative magnetic “high” regions (either positive or low-negative values), especially those located northwest of the Mississippi embayment and southeast of the Mississippi Valley graben, may also correlate to Cambrian or Precambrian mafic plutons.

B: Regional Bouger gravity (corrected for low-density sedimentary strata of post-Paleozoic age). Solid contours represent regional gravity values. Dashed contours show thickness of Mesozoic-Recent strata of Mississippi embayment (in meters). From Kane and others, 1981. Faults and fold axes shown on the map are the same as those shown on the aeromagnetic map (part A, above). See part A for description. Regions with positive gravity values (gravity highs) represent areas where high-density rocks (such as intrusive mafic igneous) are relatively close to the Earth’s surface; and, or, where low-density, granitic continental type crust is relatively thin. In contrast, regions with negative gravity values (gravity lows) represent areas where high-density mafic rocks are more deeply buried; and, or, where continental crust is relatively thick.

PRE-MESOZOIC STRUCTURES (from Thomas, 1991):

- Near vertical basement fault:
  - Ticks on down-thrown side.
- Plunge direction:
- Anticline axis
- Syncline axis
Plate 3.4: Large-scale structures of the Ouachita structural salient and adjacent foreland. Both maps show regional faults, outcrop of pre-early Mississippian (Meramecian), and subcrop of pre-Meramecian strata beneath Mesozoic cover. Both maps also show approximate locations of proprietary seismic profiles. Part A shows a regional view which extends northeastward from the Muenster arch in northeast Texas to the Ozark dome in eastern Missouri.

Part B shows an expanded view of the center of the Ouachita salient. Map illustrates selected faults within the Ouachita thrust belt and adjacent foreland. The Ouachita Mountains can be subdivided into three "thrust belts: Frontal Ouachita, Central Ouachita, and Southern Ouachita (as shown). The thin, vertical line pattern in the western part of the Frontal Ouachita thrust belt represents an area where numerous imbricate thrust faults dissect that part of the thrust belt. This area is generally named the frontal imbricate zone.

References: Ham and others, 1954; Morris, 1974; Arbenz 1989a,b, e; Thomas, 1989; Thomas and others, 1989; Ewing, 1991.
Plate 3.5 Structural cross sections of the Benton uplift and southern Arkoma basin of Arkansas.

Cross section A illustrates the attitudes of major seismic reflectors interpreted from COCORP seismic reflection profiles (Lillie and others, 1983). Cross sections B through D show several interpretations of structure based on seismic data, well data, and regional geology (Lillie and others, 1983; Arbenz, 1989b,d). Cross section E is the interpretation favored by Lillie and others (1983).

Cross section E shows another interpretation which is based upon the COCORP seismic profile included in Lillie and others (1983) and several published sources (Blythe and others, 1988; Arbenz, 1989a,b,d; Ellington and others, 1989; Van Assdell and Schwab, 1990; Haley and others, 1993). Finney, 1996; Suhm, 1996. Structures within the alliduction above the Benton arch are very complex and fault geometries are approximated and represented schematically. The surface structures are subdivided into four structural domains: Arkoma basin, frontal Ouachita thrust belt, central Ouachita thrust belt, and southern Ouachita thrust belt. Surface structures across the Benton uplift change from south-vergent in the northern part, to north-vergent in the southern part. Tight folds of the Cambrian to early Mississippian strata within the Benton uplift (dark gray), schematically represented as folds in the upper contact with the overlying Mississippian Stanley Group, contrast with broad folds within the upper part of the Stanley Group through Pennsylvanian strata to the north and south of the Benton uplift. Interpretation cross section E shows that north-vergent thrust faults are folded over south-vergent folds and offset by several south-vergent reverse faults.

The geologic map to the left shows the location of the COCORP seismic profile and the approximate location of every 100th vibrator point, and the locations of two wells used to constrain the cross sections. The map shows the locations of faults and folds shown in cross sections A through E. The geologic map uses the same gray scale and pattern scheme as do the cross sections. Map references: Lillie and others (1983), Arbenz (1989b), and Haley and others (1993).
Plate 3.6: Lithospheric model of part of the southern mid-continent and Gulf Coast region of North America.

Part A shows a gravity profile (both measured and interpreted) constructed along a north-south transect that stretches from southwestern Missouri to the center of the Gulf of Mexico. Below the gravity profile (above left) is an interpreted lithospheric cross section based on the gravity profile and constrained further by regional gravity measurements and seismic surveys. Consult Mickus and Keller (1992) for discussion of model parameters and theory.

Part B is a generalized geologic map of the southern mid-continent and Gulf Coast region of North America. Map shows location of gravity profile/lithospheric transect and other regional gravity and seismic surveys which were used as constraining data in Mickus and Keller (1989). COCORP = Consortium for Continental Reflection Profiling. PASSCAL = Program for Array Seismic Studies of the Continental Lithosphere. UTIG = University of Texas Institute of Geophysics. An interpreted cross section of the Ouachita Mountains of Arkansas based on COCORP data is presented in Lillie and others (1983). Structural interpretation based on PASSCAL data is presented in Keller and others (1989).
Plate 3.7: Generalized geology, major faults, and tectonic features of the Southern Oklahoma aulacogen and adjacent areas of part of the southern mid-continent of North America.

Generalized pre-Mesozoic geologic map of part of the southern mid-continent region of North America showing major faults and tectonic features. Location of the Southern Oklahoma aulacogen (inset map B) is shown.

EXPLANATION

ROCK TYPES (both maps)

Cambrian to early Mississippian (Meramecian)—includes: Precambrian igneous, meta-igneous, meta-sedimentary basement; Cambrian igneous rocks; Latte Cambrian to early Mississippian shallow-water shelf carbonates, sandstones, and shales.

Mississippian (Meramecian) to Permian (Ochoan)—includes: Meramecian to Virgilian proximal turbidites (syn-orogenic clastic wedge) adjacent to the Ouachita orogen; Desmoinesian to Wolfcampian clastic rocks and carbonates (westward-prograding banks) in central Texas; and Wolfcampian to Ochoan carbonates, evaporites, and red beds in west Texas and western Oklahoma.

Faults (map A)

- Normal fault (ticks on down-thrown side)
- Thrust fault or high angle reverse fault (Arbuckle and Wichita uplifts)
- Ouachita frontal thrust fault (barbs on hanging wall)

Faults (map B)

- Normal fault (ticks on down-thrown side)
- Mesozoic faults

ABBREVIATIONS:

<table>
<thead>
<tr>
<th>Code</th>
<th>Name</th>
</tr>
</thead>
<tbody>
<tr>
<td>BSSS</td>
<td>Bryan small-scale salient</td>
</tr>
<tr>
<td>TSSR</td>
<td>Tahornings small-scale recess</td>
</tr>
<tr>
<td>SF</td>
<td>Sulfur fault zone</td>
</tr>
<tr>
<td>WVF</td>
<td>Washita Valley fault zone</td>
</tr>
<tr>
<td>WVF</td>
<td>Wichita Valley fault zone</td>
</tr>
<tr>
<td>WAP</td>
<td>Wapanucka graben</td>
</tr>
<tr>
<td>SC</td>
<td>Sand Canyon nappe</td>
</tr>
<tr>
<td>KL</td>
<td>Kingston fault</td>
</tr>
<tr>
<td>CHG</td>
<td>Coleman half-grabens</td>
</tr>
<tr>
<td>OFB</td>
<td>Oklahoma Fold Belt</td>
</tr>
<tr>
<td>SAKI</td>
<td>South Arkansas Klip Fault</td>
</tr>
<tr>
<td>TSSR</td>
<td>Tishomingo small-scale recess</td>
</tr>
<tr>
<td>BSSS</td>
<td>Bryan small-scale salient</td>
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<td>TSSR</td>
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<td>SAKI</td>
<td>South Arkansas Klip Fault</td>
</tr>
</tbody>
</table>

Plate 3.7: Generalized geology, major faults, and tectonic features of the Southern Oklahoma aulacogen with respect to the Llano uplift, and other tectonic features of central and northern Texas, and tectonic features of the western part of the Ouachita orogen (including part of the Marathon salient). Map B is an expanded view of the Southern Oklahoma aulacogen showing location of major faults and tectonic features.

Plate 3.8 Generalized structure of the Wichita uplift of southern Oklahoma. Map A shows examples of seismic reflection profiles which are used (along with well and other data) to determine large-scale structural geometries of the Wichita uplift and adjacent Hollis-Jackson/Hardenbasin to the south, and the Anadarko basin to the north. Reflector horizons 'b' (seismic line 1) may represent Precambrian sedimentary strata or layered igneous rocks. Reflector horizon 'c' may be an offset continuation of one of the 'b' reflector horizons. Reflectors at large arrows 1 and 2 (cross section Y-Y', seismic lines 2/2A) may represent distorted sedimentary strata; however, they are not as clearly defined as the sedimentary succession farther north in line 2A. See Brewer (1982) for further discussion. Part B is a generalized cross section of the Wichita uplift and adjacent basins (from Brewer, 1982). Part C is a fault map of the Wichita uplift and adjacent area of southwestern Oklahoma and northern Texas. Location of seismic lines illustrated in part A are shown. Map references: Tarr and others (1965), Brewer (1982), Ewing (1991).
Plate 3.9: Simplified geologic map of the Arbuckle Mountains of south-central Oklahoma showing location of major faults and large-scale folds (gray color).

Plate 3.9: Simplified geologic map of the Arbuckle Mountains of south-central Oklahoma showing location of major faults and large-scale folds (gray color). Predominant fault trace trends are northwest-southeast (~295°-305°), near north-south (~348°), and northeast-southwest (~032°-070°). Trends of many of the large scale fold axes are oblique to the trend of the surface traces of the major faults that cross-cut the Arbuckle Mountains. South of the Reagan fault, many of the large fold axes form acute, clockwise-rotated angles with respect to Reagan and Washita Valley faults. Several fold axes north of the Reagan fault (Hickory and Wapanucka synclines and Clarita anticline) form acute, counter-clockwise rotated angles with respect to the trend of the surface trace of the Reagan fault. The apparent sense of rotation of the large fold axes, and smaller scale field measurements, suggest left-lateral "transpressive" (or generally compressive) strain south of the Reagan fault, and right-lateral "transpressive" (generally compressive) strain north of the Reagan fault. The eastern ends of the Clarita anticline and Wapanucka syncline bend sharply to a northeast-southwest trend that conforms with the overall trend of the western part of the Arkoma foreland basin.

Several fault-displacement transfer zones are located within the Arbuckle geology. Note that the northwestern end of the Washita Valley fault is a north-vergent reverse fault, whereas, in contrast, the southeastern end is a south-vergent reverse fault. The Washita Valley "fault" (or fault zone) is actually a set of several faults with differing displacements that are closely spaced where they intersect the Earth's surface. Between points A and B, the north-vergent northwestern part of the Washita Valley fault zone loses displacement towards the southeast, and ceases near point B. southeast of point B, south-vergent displacement increases progressively along the southeastern end of Washita Valley fault zone. The switch to south-vergent displacement at point B appears to correspond structurally with an apparent transfer of north-vergent displacement from the eastern end of the Reagan fault at point C, to the western end of the southeastern part of the Sulfur fault zone at point D.

References: Ham and others (1954) [revised by Johnson, 1990], Ham (1973), Hardie (1980), and McCoss (1986).
Plate 3.10: Major tectonic features of central Texas. Cross sections F-F' and G-G' illustrate major structures of part of the Ouachita orogen and adjacent foreland east of the Llano uplift. Also shown are approximated restorations of cross sections F-F' and G-G'. The restored cross sections show a possible late Morrowan geometry that may have existed prior to deposition of thick Atokan (possibly latest Morrowan-Desmoinesian) deep-water facies strata and nowhere exceed maximum published thickness for equivalent strata in the Ouachita Mountains of Oklahoma and Arkansas (>5 km) (Thomas, 1977; Fay and others, 1986). Superimposed on the restored cross sections is the approximate location of the basal detachment. The restorations of cross sections F-F' and G-G' show a mechanism that would transport Ordovician-Mississippian cherts toward the Ouachita thrust front; however, the same strata are absent above the basal detachment across the Waco uplift (F-F') and in several locations along the basal detachment of the Ouachita allochthon east of the Llano uplift (G-G'). In both restored cross sections, the basal detachment abruptly changes to a higher (post-chert) stratigraphic level east of the Waco uplift in the vicinity of a hypothetical pre-thrust normal fault. In restored G-G', the basal detachment crosses three pre-thrust normal fault zones. Position markers A, B, C, and D show corresponding locations on both the current and restored cross sections. Shown to the right is a generalized Paleozoic geologic map of the Arkansas-Louisiana-Texas-Oklahoma region showing major tectonic features of the Ouachita orogen and adjacent foreland. Locations of cross sections F-F' and G-G' are shown. Ordovician to early Mississippian deep-water facies strata shown southeast of Fault A in cross section G-G' are suggested by the local subsrops of Ordovician-Mississippian strata southeast of Fault A to the north and south of G-G' which are illustrated in the map shown to the right. Map references: Huffman and others (1978), Arbent (1989b), Nicholas (1989), Thomas and others (1989), and Cleaves (1996), proprietary seismic reflection profiles.
Plate 4.1 Pre-Mesozoic surface/subcrop geologic map of the Ouachita Mountains and adjacent foreland between the Muenster arch and Mississippi embayment.

Map shows location of selected major structures. Formation contacts beneath Mesozoic based on proprietary seismic profiles (section lines 1-9), wells, published cross sections (C-C', D-D', E-E', and COCORP), and other published and unpublished maps and cross sections.

C-C' (from Viele, 1989), D-D' (from Arbenz, 1989), E-E' (from Arbenz, 1989), COCORP (from Liddle and others, 1983).

Other references: Bradfield (1957a.c), Flawn and others (1965), Ham (1973), Hoffman and others (1978, 1987), Arbenz (1988a,b,d).

Plate 4.2: Cross sections of the Ouachita Mountains of Oklahoma and Arkansas showing along-strike variations in structure.
Plate 4.3 (Part 1): Structural cross sections of the Ouachita Mountains. Included on Part 1 of Plate 4.3 are cross sections OK4, OK3, and OK2, lithologic/mechanical stratigraphic chart of units included in all cross sections (Part 1 and Part 2), schematic restorations of cross sections OK4 and ARK3, and a map showing location of all cross sections.

All cross sections (Parts 1 and 2) illustrate the along-strike change in structural style within the Ouachita Mountains and adjacent Arkoma basin. In general, upper Atoka strata behave as a thick stiff layer and form long wavelength folds. In contrast, lower Stanley and older strata behave as a composite ductile layer and form complicated short wavelength faults.

Two disharmonic boundary/detachment horizons are shown within the Stanley strata. In some locations, contorted Stanley strata occupy wavelength transition/volume compensation zones between the overlying stiff layer and underlying ductile layer.


NOTE: Items shown in Plate 4.3 are modified from a poster that was presented at the 1999 Geological Society of America meeting in Denver, Colorado.

Plate 4.3 (Part 2): Structural cross sections of the Ouachita Mountains. Included on Part 1 of Plate 4.3 are cross sections OK1, ARK4, ARK3, ARK2, and ARK1. Lithologic/mechanical stratigraphic chart of units included in all cross sections (Part 1 and Part 2), schematic restorations of cross sections OK4 and ARK3, and a map showing location of all cross sections are shown in Part 1.

Cross sections in Plate 4.2 (Part 2) show an eastwards along-strike (from OK1 towards OK4) increase in thickness of the Atoka formation north of the Choctaw and Dutch Creek - Ross Creek fault zones. All cross sections (Parts 1 and 2) illustrate the along-strike change in structural style within the Ouachita Mountains and adjacent Arkoma basin. In general, upper Stanley-Atoka strata behave as a thick fault. The Y-City fault is interpreted as a delamination backthrust that operates along the base of the thick Atoka Formation within the Johns Valley Shale. Evidently, strata south of the Y-City fault are underplated beneath, and overturned towards the south, south of the Y-City fault.

Plate 4.4: Two interpreted cross sections of part of the Frontal Ouachita thrust belt and adjacent Arkoma basin based on the COCORP seismic reflection profile in Lillie and others (1983). Cross section A shows the Lillie and others (1983) interpretation of the northern end of the COCORP profile. Cross section B shows an alternative interpretation superimposed on the COCORP profile. In cross section B, the basal detachment is at the base of the lower Atoka in the northern half of the cross section. Also shown are two south-vergent detachments, one in the lower Atoka, and the other in the middle Atoka. Location of cross sections coincide with northern part of cross section X-X’ in Plate 4.2.
Plate 4.5: Geologic-structure map of the eastern nappes part of the eastern Benton uplift showing major faults and detachments.

MODIFIED FROM
Haley and Stone, 1990
--Lonsdale, AR
Haley and Stone, 1991
--Fourche SW, AR
--Ferndale, AR
--Paron, AR
Haley and Stone, 1993
--Benton, AR
Haley and Stone, 1994
--Congo, AR
--Goosepond Mountain, AR
--Haskell, AR
--Jessieville, AR
--Lake Norrell, AR
--Lonsdale NE, AR

EXPLANATION

FAULTS

detachment fault (barbs on hanging wall)
overturned detachment fault (barbs on hanging wall)
thrust fault (barbs on hanging wall)
overturned thrust fault (barbs on hanging wall)
thrust fault/reverse fault reactivated as normal fault (barbs and ticks on hanging wall)
PLATE 5.1
CROSS SECTIONS SHOWING INTERSECTION OF OUACHITA THRUST BELT AND ARBUCKLE UPLIFT STRUCTURES BENEATH MESOZOIC GULF COASTAL PLAIN
(CROSS SECTIONS 1, 2, 3, 4, AND 5)

(Cross sections 1, 2, 3, 4, and 5)
PLATE 5.2
CROSS SECTIONS OF PART OF THE BRYAN SMALL SCALE SALIENT, SOUTHEASTERN MUENSTERTH, AND BASEMENT UPLIFT SOUTHEAST OF THE MUENSTERTH ARCH
(CROSS SECTIONS 6, 7, 8, and 9)
(SEE PLATE 4.1 FOR LOCATIONS OF CROSS SECTIONS)

CROSS SECTION 6

CROSS SECTION 7

CROSS SECTION 8

CROSS SECTION 9