STRUCTURAL GEOMETRY OF THE FRONTAL OUACHITAS-ARKOMA BASIN TRANSITION ZONE IN WESTERN ARKANSAS

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ABSTRACT

The Choctaw Fault and Ross Creek Fault are the leading-edge thrusts of the Ouachita fold-thrust belt and form the southern boundary of the Arkoma foreland basin in Oklahoma and Arkansas, respectively. Strain partitioning between the fold-thrust belt and the foreland basin is accommodated by the Wilburton Triangle Zone in the footwall of the Choctaw Fault of Oklahoma; however, few studies document the geometry of Ouachita-Arkoma transition zone in Arkansas. This study uses depth-converted 2-D seismic reflection profiles and well log data to clarify the subsurface structure and establish the presence or lack of triangle zone elements within the accommodation zone between the Choctaw Fault and Ross Creek Fault in north-central Scott County, western Arkansas. Structural interpretation of three depth-converted 2-D seismic profiles shows a triangle zone containing the surficial tip-out of the Choctaw Fault in the footwall of the Ross Creek Fault. This triangle zone is called the Waldron Triangle Zone after nearby Waldron, Arkansas, and is composed of three stacked wedges that share a roof thrust, the north-dipping lower Atokan Decollement, which meets a floor thrust, the south-dipping Stanley Decollement, at a tip line below the Poteau Syncline. Knowledge of accommodation zone structure predicts that the Waldron Triangle Zone dies out west of the seismic data into Oklahoma and coincides with the formation of the laterally equivalent Wilburton Triangle Zone.
DEDICATION

If I have seen further it is by standing on the shoulders of giants.

– Sir Isaac Newton

For my grandparents, John and Lillian Yezerski & Maurice and Jean Knight.
### LIST OF ABBREVIATIONS AND SYMBOLS

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td>2-D</td>
<td>Two dimensional</td>
</tr>
<tr>
<td>3-D</td>
<td>Three dimensional</td>
</tr>
<tr>
<td>AI</td>
<td>Acoustic impedance</td>
</tr>
<tr>
<td>API</td>
<td>American Petroleum Industry</td>
</tr>
<tr>
<td>CALI</td>
<td>Caliper</td>
</tr>
<tr>
<td>CF</td>
<td>Carbon Fault</td>
</tr>
<tr>
<td>CHF</td>
<td>Choctaw Fault</td>
</tr>
<tr>
<td>DT</td>
<td>Sonic</td>
</tr>
<tr>
<td>ft</td>
<td>Feet</td>
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<tr>
<td>ft/s</td>
<td>Feet per second</td>
</tr>
<tr>
<td>GR</td>
<td>Gamma ray</td>
</tr>
<tr>
<td>LAD</td>
<td>Lower Atokan Detachment/Decollement</td>
</tr>
<tr>
<td>m</td>
<td>Meter</td>
</tr>
<tr>
<td>m/s</td>
<td>Meters per second</td>
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Ma  Millions of years before present

MD  Measured depth

µs/ft  Microseconds per foot

RC  Reflection coefficient

RCF  Ross Creek Fault

RHOB  Density

s  Second

TD  Total depth

WTZ  Wilburton Triangle Zone
ACKNOWLEDGEMENTS

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CHAPTER 1
INTRODUCTION

The Late Paleozoic Ouachita Orogeny resulted from a continent-continent collision between the southern margin of North America and Gondwana (Poole et al., 2005). Thrust loading and sediment loading during the orogeny formed and ultimately incorporated synorogenic sediment into the northward-propagating Ouachita fold-thrust belt (Viele, 1979; Houseknecht and Kacena, 1983). Orogenic structures crop out for 250 miles (400 kilometers) along strike across eastern Oklahoma and western Arkansas and are unconformably overlain by Cretaceous units to the south (Figure 1). The Arkoma basin extends northward to the Ozark uplift. The Ouachita-Arkoma system may be continuous with the Marathon fold-thrust belt of Texas to the south (Poole et al., 2005), and may be continuous with the Appalachian fold-thrust belt to the east (Thomas, 1989); however, subsurface data show no interference structures between the Alleghanian fold-thrust belt and the Ouachita fold-thrust belt, indicating that the Alleghanian fold-thrust belt may continue toward the southwest (Robinson et al., 2012).

Strain partitioning between the frontal portion of fold-thrust belts and associated foreland basins in many regions is expressed by a tectonically thickened wedge known as a triangle zone, a term that first entered the literature as a description of the triangular zone of poor seismic imaging between the foreland- and hinterland-dipping reflectors in the foothills region of the eastern Cordillera of Alberta, Canada (Gordy et al., 1977; Jones, 1982; DeCelles and Giles, 1996). Subsequent study of the Canadian Cordillera and other orogenic belts containing triangle
Figure 1 – Map showing the structural provinces of the Ouachita-Arkoma system. Satellite imagery from Google Earth. Modified from Sutherland (1988).
zones demonstrated that this type of frontal structure exhibits similar geometries worldwide (Vann et al., 1986, Çemen et al., 2001a; 2001b).

In a triangle zone, foreland thrusting is commonly an intercutaneous wedge bound by a floor thrust and a roof thrust. Couzens and Wiltschko (1996) propose an end-member model for the internal geometry of the wedge (Figure 2). Type I triangle zones contain a wedge that remains relatively undeformed between opposing thrusts that emanate from a master detachment. Type II triangle zones contain internally duplexed wedges that possess two or more detachment levels. In the case of both end-member geometries, the roof and floor thrusts merge at a tip line or point of zero displacement, passively uplifting a frontal syncline as transport of the wedge forces the tip line to propagate into the foreland (Jones, 1982; 1996).

**Figure 2** – Schematic cross section of triangle zone end-members. Modified from Couzens and Wiltschko (1996).
Triangle zones yield information about the processes of deformation during the formation of fold-thrust belts, although the specific role that the triangle zone occupies in the spatial and temporal evolution of a thrust belt is not well understood and could vary among orogenic belts (Jones, 1996). Finite-element models demonstrate the viability of a propagating syncline or continuous generation model of triangle zone development if the roof strata are of sufficiently low friction (Jamison, 1996).

Çemen et al. (2001a; 2001b) interpret the frontal Ouachitas to Arkoma basin transition zone in eastern Oklahoma as a triangle zone in the footwall of both the Choctaw Fault (CHF), the leading-edge thrust of the Ouachita fold-thrust belt, and the Carbon Fault (CF), a north-dipping backthrust, and refer to this structure as the Wilburton Triangle Zone (WTZ) after nearby Wilburton, Oklahoma (Figure 3). In westernmost Arkansas, the termination of the trace of the CF coincides with the Ross Creek Fault (RCF), which continues as the leading-edge thrust of the eastern margin of the orogen (Figure 1) (Van Arsdale & Schweig III, 1990; Arbenz, 1989).

While the structural geometry of the transition zone is well documented in eastern Oklahoma (Çemen, et al. 2001a; Çemen, et al, 2001b), there has been little detailed analysis of the transition zone along strike in western Arkansas. Previous studies of frontal structures of the eastern margin of the Ouachita fold-thrust belt (Figure 3) are mostly limited to regional cross sections based on surface mapping and crustal-scale geophysical surveys (Arbenz, 2004; Blythe et al., 1988; Van Arsdale & Schweig III, 1990; Roberts, 2004).

The main purpose of this investigation is to describe the structural geometry of the frontal Ouachita fold-thrust belt in westernmost Arkansas based on available surface and subsurface data. Structural interpretations derived from depth-converted 2-D seismic reflection profiles are used to demonstrate the structural geometries of the area between the overlap of the major frontal
thrusts, the CHF and RCF (Figure 1). Describing the structures that accommodate strain partitioning between the Ouachita fold-thrust belt and the Arkoma basin will aid in understanding the structural evolution of Ouachita system, in addition to providing insight about the lateral continuity with the WTZ in Oklahoma.
Study Area

The study area is at the boundary between the southern Arkoma basin and the northern Ouachita fold-thrust belt in central Scott County, western Arkansas. This structural transition is thought to occur in the Waldron and Boles 7.5”-series quadrangles (Figure 4). This rural area extends approximately 16.5 miles (26.4 kilometers) north-south and approximately 6.5 miles (10.4 kilometers) east-west. 2-D seismic reflection profiles extend across undulating topography with over 1,700 (520 meters) of local relief between forested and linear ridgelines and cultivated floodplains, and are partially contained within the Ouachita National Forest. Although the 2-D seismic and well data used to construct depth-converted profiles extend beyond the footprint of the Waldron and Boles quadrangles, interpretations made during this investigation are focused in or near this area.
Figure 4 – Topographic map of the study area. Map position noted by red box within inset map at lower left. Map projection is Lambert Conformal Conic. Seismic lines (black lines) labeled A-H. Locations of wells used in this study indicated by white circles. Blue boxes outline the map footprint of the Waldron (north) and Boles (south) 7.5"-series quadrangles.
Methodology

Southwestern Energy Company provided a dataset consisting of 145 wells and ten 2-D time-migrated seismic reflection lines partially contained within Scott County, western Arkansas (Figure 4). The seismic data were acquired by CGG between 1982 and 1991 and processed by Kelman Technologies Incorporated in 2005 and 2006. Eight of the seismic reflection lines and three wells were selected for this study. Well and seismic reflection data were imported into the Petrel interpretation software for interpretation. Formation tops for selected wells were provided by Southwestern Energy Company. Well log data available at the Arkansas Geologic Survey also were used to identify formation tops and establish stratigraphy in adjacent wells. Well data for the Potoco LLC Milton #1-2 and the El Paso Natural Gas Company Cheesman #1 wells were tied to seismic data by construction and projection of synthetic seismograms to proximal seismic lines for identification of seismic reflectors using the well-seismic tie process within Petrel. Stratigraphic horizons were extrapolated across the seismic coverage from the synthetic seismograms to complete the time-domain interpretation.

Converting from the time domain to the depth domain, which is required for proper structural interpretation, necessitates the use of a velocity model to recognize and correct velocity-induced interpretation errors. Three seismic profiles were depth-converted through application of a velocity model derived from log-based velocities applied between seismic horizons and calibrated using formation tops within the Lone Star Producing Company C.J. Williams #1 well as structural control, using a velocity modeling function.
The existing geologic mapping at the 7.5”-series scale is poor and intermittent. A published geologic map exists for the Waldron quadrangle; however the Boles quadrangle, containing the trace of the Ross Creek Fault, was mapped as part of an incomplete 15”-series preliminary geologic map that lacks dense structural data (Reinemund and Danilchik, 1957). A composite geological map of the study area was created by digitizing the Waldron quadrangle from the 7.5”-series geologic map and combining it with the digitized copy of the Boles quadrangle from the 15”-series Waldron geologic worksheet. Strike and dip measurements were added to the geological map of the Boles quadrangle and the geologic map of the Waldron quadrangle was field checked during field work performed in April, 2012. Surface structural data helped control the interpretations along the trace of the depth-converted seismic reflection profiles in areas of poor seismic imaging.
Tectonic Setting of the Ouachita Orogen and the Arkoma Basin

The Arkoma basin is located south of the Ozark uplift and north of the stratigraphically continuous Ouachita fold-thrust belt (Figure 1). Sedimentation began in the Late Precambrian to Early Cambrian during breakup and rifting of the Rodinian supercontinent, which resulted in the development of the Laurentian platform (Figure 5) (Poole et al., 2005). Extension resulted in the formation of oceanic crust outboard of the southern continental margin, which was a passive margin through Early Mississippian time. The southward-thickening passive margin wedge consists of carbonate-dominated shallow shelf facies to the west with increasing clastic input to the east (Sutherland, 1988). South of the Laurentian platform, a sediment-starved deep water basin, known as the Ouachita embayment, hosted a succession of dark shale containing minor carbonate and chert that characterize the ‘Ouachita facies’ exposed in the thrust sheets of the fold-thrust belt (Houseknecht, 1986).

The onset of the Ouachita Orogeny is marked by deposition of a thick flysch sequence above deep water sediments beginning in Mississippian time due to the northward advance of a Gondwanan volcanic-plutonic arc that includes the Yucatan block, above south-directed subduction of Laurentian oceanic crust (Viele, 1979; Houseknecht, 1986). Continued convergence between southern Laurentia and the Yucatan block consumed the oceanic crust of the proto-Atlantic or Ouachita oceanic basin resulting in a growing accretionary wedge (Houseknecht, 1986). As the basin closed, the accretionary wedge was uplifted to near the sea surface and was a major
Figure 5 – Cross sections showing the tectonic development of the Ouachita-Arkoma system. A: Precambrian rifting B: Late Cambrian to Early Mississippian passive margin development C: Early Mississippian to Early Pennsylvanian south-directed subduction D: Late Mississippian to Early Pennsylvanian normal faulting E: Late Pennsylvanian thrusting. Cross sections modified from Houseknecht (1986).
source of sediment that began filling the Ouachita foreland basins, based on paleocurrent indicators (Houseknecht and Kacena, 1983).

Flexure and sediment loading of the continental platform resulted in major tectonic subsidence of the foreland basin, along with movement of south-dipping, east- to northeast-striking syndepositional normal faults that propagated northward into the basin. These extensional faults resulted from flexural extension of the lower plate during subduction (Van Arsdale, and Schweig III, 1990; Bradley and Kidd, 1991). Flexure-induced extension cut the passive margin section and accommodated extreme sediment thickening above the downthrown block, especially during deposition of the middle Atoka Formation at ~312 Ma (Sutherland, 1988). The greatest syndepositional movement occurred in the southwestern portion of the Arkoma basin in eastern Oklahoma (Houseknecht, 1986). Continuity of the base Pennsylvanian disconformity eastward into Arkansas limits the onset of major basement-involved down-to-south faulting to earlier than 318 Ma (Arbenz, 1989; Van Arsdale and Schweig, 1990; Thomas, 2010).

Folding of synorogenic sediments in the southern Arkoma basin indicate that compression began during Late Pennsylvanian time after the deposition of the Atoka Formation (Houseknecht and Kacena, 1983). Gentle folding of the post-Atokan section suggests minor compressional deformation continued through the end of the Pennsylvanian, controlling the deposition of shallow marine through alluvial systems along the axis of the gently contracting basin before complete tectonic quiescence beginning in Permian time (Houseknecht and Kacena, 1983).
CHAPTER 2

STRATIGRAPHY AND STRUCTURAL OVERVIEW OF THE
FRONTAL OUACHITAS-ARKOMA BASIN TRANSITION ZONE

Arbenz (2004) divided the Paleozoic stratigraphy of the Ouachita-Arkoma foreland basin by regional tectonic setting with a deepening north-to-south transition from the North American continental margin (Arkoma Basin and Foredeep sections) to the deep-water facies of the Ouachita Basin (Ouachita Basin section) (Figure 6). The lithology and extent of the stratigraphy involved in the frontal Ouachita fold-thrust belt and southern Arkoma basin, especially in western Arkansas, is poorly documented. Therefore, it is necessary to divide any discussion of transition zone stratigraphy into pre-orogenic and synorogenic stratigraphy.
**Figure 6** – Stratigraphic column and correlative facies of the Ouachita-Arkoma system. Adapted from Arbenz (2004).
Pre-Orogenic Stratigraphy

Pre-orogenic stratigraphy is Laurentian passive margin sediments that accumulated prior to tectonic subsidence during the early phases of the Ouachita Orogeny (Figure 5). The pre-orogenic Late Cambrian to Early Mississippian stratigraphy is not exposed in the frontal portion of the Ouachita fold-thrust belt and has not been penetrated by local wells; therefore, descriptions presented here are drawn from measured sections in the central Ouachita Mountains (Lowe, 1989).

The oldest sedimentary unit of the Ouachita Basin section is the Late Cambrian to Early Ordovician Collier Formation. Although its basal contact with underlying crystalline rocks is not exposed, more than 1000 feet (305 meters) of section is exposed that contains gray to black shale containing minor amounts of chert, sandstone, and conglomerate, as well as limestone beds up to 50 feet (15 meters) thick in the upper portion of the formation (McFarland, 2004). In Arkansas, the upper limestone beds contain a greater abundance of sand and are often lenticular (Lowe, 1989). Lying above Collier Formation is the 500- to 850-foot (150 to 260 meters) thick Early Ordovician Crystal Mountain Formation, which consists of massive, coarse-grained, and carbonate-cemented gray quartzose sandstone interbedded with thin shale (McFarland, 2004). Few large boulders of metaarkose up to 50 feet (15 meters) in diameter have been reported in conglomerate beds in eastern Arkansas (Lowe, 1989). The overlying Early Ordovician Marzan Formation ranges from 1000 feet to 2500 feet (305 to 760 meters) of gray to
black shale that contains minor siltstone, sandstone, conglomerate, and chert (McFarland, 2004). The overlying Middle Ordovician Blakely Formation contains up to 700 feet (215 meters) of black to green shale interbedded with few coarse-grained quartzose sandstone beds less than three feet (one meter) thick (Lowe, 1989). The overlying Middle Ordovician Womble Formation is a 500- to 1200-foot (150 to 365 meters) thick section of predominately black and green shale (McFarland, 2004). Quartzose sandstone beds less than one foot (30 centimeters) thick are present in the lower portion of the formation while thin limestone and chert beds are present in the upper portion (Lowe, 1989). The overlying Middle to Late Ordovician Bigfork Formation consists of 450 feet (135 meters) of highly fractured chert and is interbedded with thin shale and limestone (McFarland, 2004). The overlying Late Ordovician Polk Creek Formation consists of up to 225 feet (70 meters) of black carbonaceous shale containing interbedded black chert near the base of the formation and a characteristic conglomerate at the top (Lowe, 1989).

The Early Silurian period is recorded by the 1200-foot (365 meters) thick Baylock Formation, which consists of fine- to medium-grained quartzose gray and tan sandstone, interbedded with black shale (Lowe, 1989). The Late Silurian period is recorded by the Missouri Mountain Formation, a 300-foot (90 meters) thick dark shale succession interbedded with thin quartzose sands and containing a four-foot (one meter) thick conglomerate at its base and thin bedded novaculite near the top of the formation (McFarland, 2004).

The overlying Devonian through Early Mississippian section is recorded in the Arkansas Novaculite, a white and black chert unit that is up to 900 feet (275 meters) thick in the Ouachita Mountains but thins quickly north (McFarland, 2004). Throughout the formation, the novaculite is interbedded with thin shale and contains sparse conglomerate beds (McFarland, 2004).
The dominantly clastic pre-orogenic interval is thought to record slow deposition rates of pelagic and hemi-pelagic deep marine clays between periods of sand deposition by turbidity currents and sediment-gravity flows (Lowe, 1989). Paleocurrent indicators indicate that sediment dispersal was northeast-southwest with periodic quartzose influx from the south, which Lowe (1989) attributes to a southern continental source, possibly a rift-isolated segment of Laurentian crust. Alternatively, axial current redistribution of Laurentian-derived detritus along the base of the shelf-slope rise could produce similar indications of a southern provenance (Viele and Thomas, 1989).
Synorogenic Stratigraphy

The Late Mississippian Stanley Group overlies the Arkansas Novaculite as the first synorogenic unit and is 3,500 to 10,000 feet (1,065 to 3,050 meters) thick, thinning to the north (McFarland, 2004). Although mostly conformable with the underlying novaculite, basal chert conglomerate indicates some submarine erosion (McFarland, 2004). The Stanley Group consists dominantly of dark gray shale; however, beds of laterally discontinuous sandstone are common as well asolistostrome that contain exotic sandstone and limestone blocks (Morris, 1989). The occurrence of tuff and the paleocurrent indicators contained therein demonstrate dispersal from a southern volcanic arc (Figure 7) (Morris, 1989).

The 3,500- to 6,000-feet (1,065 to 1,830 meters) thick Early Pennsylvanian Jackfork Group conformably overlies the Stanley Group and consists dominantly of brown to tan, thin- to thick-bedded, quartzose sandstone (Morris, 1989) that fines to the north and supports the elevated topography south of the transition zone in western Arkansas (Suneson, 2012). Interbedded gray to black shale increases in abundance to the north (McFarland, 2004). The Early Pennsylvanian Johns Valley Formation overlies the Jackfork Group and consists of dark shale with minor amounts of sandstone. Despite the high degree of internal deformation, the original thickness is estimated at 1,500 feet (455 meters) (McFarland, 2004). The Johns Valley Formation also contains a characteristic Cambrian to Pennsylvanian passive margin facies olistostromal deposit (Figure 7) (Morris, 1989). The blocks may have originated from a combination of slumping of the depositional shelf to the north and erosion from the footwall of
Figure 7 – Paleogeographic reconstructions of Mississippian and Pennsylvanian depositional systems. Reconstructions focus on present day eastern Oklahoma and western Arkansas. Modified from Sutherland (1988).
south-dipping syndepositional normal faults (Figure 8) (Sutherland and Manger, 1979). The overlying 60-foot (eighteen meter) thick Morrowan Cromwell Sandstone of the Union Valley Formation is a discontinuous calcareous sandstone interbedded with thin shale and overlain by less than twenty feet (six meters) of arenaceous limestone (Sutherland, 1988). Although local well tops indicate that the Cromwell Sandstone overlies the Johns Valley Formation in the transition zone, Morris (1989) interprets the Cromwell Sandstones equivalent to the Johns Valley Formation in Oklahoma, and Sutherland (1988) shows the Cromwell as equivalent to basin shale in the frontal Ouachita Mountains immediately to the south and presumably in the hanging wall of the Ross Creek Fault.

Figure 8 – Schematic cross section of the Arkoma basin. Note the intertonguing of shelfal and deep-water facies within the frontal Ouachitas. Modified from Sutherland and Manger (1979).
The overlying Early Pennsylvanian Bloyd Formation consists of interbedded dark shale, limestone, calcareous sandstone, and minor amounts of coal (McFarland, 2004). The Bloyd Formation is primarily restricted to the Ozark facies of northern Arkansas where it is up to 200 feet (60 meters) thick (McFarland, 2004); however, the shelfal Kessler Limestone member likely extends into the transition zone area as interpreted in local wells. The interpretation of deep-water facies at the Morrowan-Atokan boundary in the hanging wall of the Ross Creek Fault (RCF) suggests that the underlying Late Morrowan stratigraphy is carbonate-poor basinal equivalents (Sutherland, 1988).

Carbonate influence along the Early Pennsylvanian shelf ceased at the beginning of Atokan time with rapid deposition of the Atoka Formation (Figure 8), an asymmetric synorogenic clastic wedge with an estimated thickness greater than 20,000 feet (6,100 meters) at the southern margin of the Arkoma Basin where the lithology is dominantly shale (Zachry, 1983). The Atoka Formation is subdivided into three members, the lower, middle, and upper (Sutherland, 1988). The base of the lower member is recognized by the incision of the 25- to 200-foot (seven to 60 meters) thick Spiro Sandstone into shale of the Bloyd Formation (Zachry, 1983) and marks the end of northerly-sourced passive margin deposition before rapid subsidence accommodated deposition of a thick succession of turbidites preserved in the hanging wall of the RCF in Arkansas that grade towards neritic strata to the west in Oklahoma above the Choctaw Fault (CHF) (Figure 1) (Sutherland, 1988; Arbenz, 2004).

Syndepositional normal faulting during Middle Atokan time controlled deep seafloor bathymetry and resulted in axial channelization of easterly sand-rich sediment-gravity flows that rapidly thicken above the downthrown hanging walls that define the middle portion of the Atoka Formation in Oklahoma (Houseknecht, 1986). Accurate quantification of the extreme thicknesses
reached by the middle Atoka Formation is precluded by erosion of the upper intraformational contact, but the thickest measurable section is located within the study area between the overlap of the traces of the CHF and RCF (Figure 1) (Sutherland, 1988). Restored thickness estimates by Haley (1988) indicate that the formation thickens southward at 500 feet (150 meters) per mile. Due to the limited lateral extent of the submarine axial channels and fans, sandstone of the middle Atoka Formation are discontinuous and commonly change or lose character over distances of a quarter-mile (0.4 kilometers) and cannot be correlated over significant distances (Zachry and Sutherland, 1984). Cessation of down-to-south normal faulting marked the end of the deep marine facies of the middle Atoka Formation. The upper Atoka Formation is not cut by these faults and consists of fluvial and deltaic systems that prograded southward in Late Atokan time (Sutherland, 1988).

Conformably overlying the Atoka Formation is the Desmoinesian Krebs Group. Exposures are limited to the lower formations in the Poteau Syncline at the northern edge of the Waldron quadrangle. The basal unit of the Krebs Group is the tidal-deltaic Hartshorne Formation (Figure 7), a ten- to 300-foot (three to 90 meters) thick, brown to light gray, massive sandstone section (McFarland, 2004). Conformably overlying the Hartshorne Formation is the 500- to 2300-foot (150 to 700 meters) thick McAlister Formation, which consists dominantly of shale, isolated sandstone beds, and an increasing abundance of coal towards the top of the formation (McFarland, 2004). Conformably overlying the McAlister Formation is the 1600-foot (490 meter) thick Savanna Formation, which consists of dark gray shale, siltstone, fine-grained quartzose sandstone, and several coal beds (McFarland, 2004).
Structural Overview

Structural mapping of the southern Arkoma basin north of the Choctaw Fault (CHF) in eastern Oklahoma (Figure 1) shows a series of broad synclines separated by tight, north-vergent anticlines (Houseknecht, 1987). Availability of subsurface data increased in the late 1980s and 1990s with the discovery of natural gas in the Wilburton area of Oklahoma. Interpretations of well log and 2-D seismic data led Çemen et al. (2001a; 2001b) to propose the presence of a triangle zone, the Wilburton Triangle Zone (WTZ), in the footwall of the CHF to explain strain partitioning from the highly deformed frontal Ouachita fold-thrust belt to the south and the mildly deformed Arkoma basin to the north (Figure 9, line A).

Here, the intercutaneous wedge of the WTZ is composed of middle Atoka Formation that is detached above the lower Atokan Detachment (LAD), a horizontal to gentle (6°) south-dipping floor thrust that merges with the Carbon Fault (CF) at a tip line. The north-dipping CF, while mostly eroded, serves as the roof thrust that bounds the passively uplifted and transported foreland syncline (San Bois Syncline). The WTZ is underlain by a duplex that imbricates the basal Atokan Spiro Sandstone, which utilizes the LAD as a roof thrust and is detached above a floor thrust that ramps from the Late Devonian to Early Mississippian Woodford Shale through the Late Mississippian to Early Pennsylvanian Springer Shale.

Elements of the triangle zone extend eastward near Wister Lake in Leflore County, Oklahoma (Figure 9, line B; McPhail, 2001). Although the geometry is similar to the WTZ the triangle zone incorporates both lower and middle Atoka Formation in the intercutaneous wedge.
The LAD similarly merges with the CF at a tip line beneath the foreland syncline (Cavanal Syncline), but the CF dies out in the subsurface of the Heavener Anticline. The basal Spiro Sandstone-involved duplex is likewise present and does not change character from its lateral equivalent to the west. Geologic mapping by Suneson (2012) suggests that the Carbon Fault (CF) terminates in the Wister Lake area west of the Oklahoma-Arkansas state border while the CHF continues into western Arkansas before tipping out at the surface in central Scott County (Figure 3).
Figure 9 – Cross sections of the WTZ in Oklahoma. **Line A** modified from Çemen et al. (2001a); **Line B** modified from McPhail (2001). **LAD**: lower Atokan Detachment
**IPb/IPbg**: Boggy Formation **IPsv**: Savanna Formation **IPm**: McAlister Formation **IPh**: Hartshorne Formation **IPA**: Atoka Formation **IPro**: Red Oak Sandstone **IPC**: Cecil Sandstone **IPS**: Spiro Sandstone **IPjv**: Johns Valley Formation **IPjf**: Jackfork Group
CHAPTER 3

WELL-SEISMIC TIES AND DEPTH CONVERSION

The first step in subsurface structural interpretation is to define the lithology and stratigraphic units in wells using formation tops from scout tickets, a file summarizing the information available on a well, as well as analysis of well log data. Tying well data to a particular seismic reflection profile is accomplished by constructing a synthetic seismogram, where the seismic reflection response of the earth is modeled along the well borehole and then adjusted to achieve the best possible match with a seismic trace or series of traces that a particular well lies along or is projected (Ewing, 1997). Formation tops are then correlated to the reflections to define beds and delineate the structural geometry.
Formation Top Interpretations

Three well logs were utilized to identify the formations and provide known geologic markers that served as horizons during seismic interpretation (Figure 4). The Potoco LLC Milton #1-2 well is located northwest of the Waldron quadrangle within north central Scott County and reaches total depth (TD) at 12,209 feet (3,722 meters) measured depth (MD) (Figure 10). The Lone Star Producing Company C.J. Williams #1 well is located north of the Waldron quadrangle in north central Scott County and reaches TD at a depth of 12,397 feet (3,780 meters) MD (Figure 10). The El Paso Natural Gas Company Cheesman #1 well is located east of the Waldron quadrangle in central Scott County and reaches TD at a depth of 12,680 feet (3,866 meters) MD (Figure 10). Well logs for both the Milton #1-2 and Cheesman #1 wells were provided, while well logs for the C.J. Williams #1 well were digitized using Neuralog well log interpretation software from raster images available through the online database of the Arkansas Geological Survey and imported into Petrel. Formation tops were interpreted from operator picks for the C.J. Williams #1 well and used as a type log. Correlation of formation tops between the three wells are shown using the gamma ray (GR) and sonic (DT) logs, the only two log types contained in all three wells (Figure 10). The GR log is used as an indicator of formation shale content as clay-rich formations will exhibit a higher background level of radiation, and thus a higher GR count. Sandstone and carbonate units containing little to no clay will exhibit a lower GR count. The Cheesman #1 well also contains a spontaneous potential (SP) log, measured in millivolts, where increasing positive values correlate with greater shale content. The DT log is a measure of formation velocity,
Figure 10 – Formation top interpretations from well logs. GR and SP well logs show high shale content with few thin sandstone beds of the Atoka Formation above the Spiro Sandstone. Below the Spiro Sandstone and above the Jackfork Group, the low GR counts show low shale content and the high DT values indicate the presence of carbonate.
recorded as the one-way travel time of a sonic pulse between a source and a receiver (Ellis and Singer, 2008).

The deepest stratigraphy recorded by the three well logs occurs within the Chessman #1 well, which reaches TD below the top of the Jackfork Group. Jackfork Group sandstone exhibits a serrate GR signature with an abrupt upper contact at the base of the Johns Valley Formation. The high shale content of the Johns Valley Formation is indicated by a smooth, elevated baseline near 150 API. The Cromwell Sandstone of the Union Valley Formation exhibits a smooth, clean shape and slightly gradational contacts with the overlying and underlying shale-rich units. The Spiro Sandstone contains a characteristic smooth and relatively clean GR signature with abrupt upper and lower contacts. Atokan sandstones above the Spiro Sandstone all exhibit irregular and serrate GR signatures contained between abrupt upper and lower contacts.

Due to lack of production and the confusing informal nomenclature of Atokan sandstones that are often field specific, subsurface terminology used here is applied from the adjacent Mansfield gas field in southeastern Sebastian and northernmost Scott Counties (Figure 3). The Spiro Sandstone represents the base of the Atoka Formation, and the base of the Hartshorne Sandstone marks the top of the formation (Houseknecht, 1986). Dividing the Atoka Formation into lower, middle, and upper members; however, is complicated by the limited lateral extent of key beds. The lower Carpenter Sandstone is accepted as the top of the lower Atoka Formation in the data accompanying the original formation tops for the C.J. Williams #1 well. The top of middle Atoka Formation in the southernmost portion of the productive trend of western Arkansas commonly utilizes the Mansfield Sandstone (Boland, G., pers. comm., 2013). The position of Atokan sandstones delineates structural geometries. Figure 11 illustrates the recognized
sandstones and their association in the lower, middle, or upper members of the Atoka Formation in eastern Oklahoma and western Arkansas.

**Figure 11** – Stratigraphic chart showing the correlative named sandstone units between eastern Oklahoma and western Arkansas. It is important to note that this correlation refers only to approximate time-equivalent deposition and stratigraphic position and that lateral continuity is unlikely due to the isolated nature of submarine channel and debris flows (Sutherland and Manger, 1979).
Construction of Synthetic Seismograms

A DT log is typically calibrated by a vertical seismic profile (VSP). A VSP generates a time-depth relationship by measuring the travel time to receivers placed within the borehole. The time-depth pairs generated constitute a checkshot survey that is used to calibrate the DT log. The caliper log, a measurement of borehole diameter, can be used to qualitatively evaluate borehole quality and therefore the reliability of the DT log values (Cordier, 1985). The DT values are multiplied by the bulk density (RHOB) values to generate an acoustic impedance (AI) log. Modeling the subsurface as a layered series of interfaces, each containing its own AI, it is possible to predict the amplitude of a seismic response by first calculating the reflection coefficient at a boundary between two layers through the equation:

\[ R = \frac{(\text{AI}_{n+1} - \text{AI}_n)}{(\text{AI}_{n+1} + \text{AI}_n)} \]

where \( \text{AI}_n \) is the AI of the upper layer and \( \text{AI}_{n+1} \) is the AI for the lower layer and where \( R \) is the reflection coefficient (Bacon et al., 2003). The derived reflectivity sequence is then convolved with a wavelet, either idealized or extracted from the seismic data being matched, to generate a synthetic. Seismic data are initially collected using a minimum phase wavelet where the first amplitude zero crossing is aligned with the acoustic impedance boundary. Because seismic interpreters are concerned with describing the profile in terms of events, the wavelet is commonly converted to zero-phase, which forces the pulse to align symmetric around the impedance boundary responsible for the particular amplitude (Bacon et al., 2003).
Convolution is a mathematical operation that describes the change in waveform shape during passage through a filter. The convolution operation is described by the function:

\[ y(t) = g(t) * f(t) \]

where \( g(t) \) is the input seismic impulse, \( f(t) \) is the impulse or Earth response, and \( y(t) \) is the filtered output or synthetic (Kearey et al., 2002).

By convolving a seismic pulse with an AI series in a layered Earth model, it is possible to predict the seismic response of the subsurface in one dimension along the trace of a borehole logged by DT and RHOB. A synthetic allows for correlation in time of geologic events observed along the borehole with seismic events along a particular seismic trace by matching the seismic signature. Due to borehole washouts, logging tool malfunctions, and missing log data, ideal correlation between a synthetic and a seismic trace is rarely possible and requires interpretation (Box et al., 2004).
Synthetic Seismograms

The synthetic for the Milton #1-2 well was projected 3.4 miles (5.4 kilometers) at 082.6° to the northern edge of seismic line B; the synthetic for the Cheesman #1 well was projected 3.0 miles (4.8 kilometers) at 257.4° to the south central portion of seismic line H (Figure 4). Neither well contains a RHOB log and thus a log was created with a constant value of 1000 kg/m³ after the methodology of Bailey (2007). Because checkshot surveys were not available for the two wells, an approximate time-depth relationship was assigned to each well by integrating the respective DT logs and calculating a total travel time to each formation top. This relationship was not manually adjusted because of the absence of independent time-depth control.

For the Milton #1-2 and Cheesman #1 wells, a wavelet was extracted from the seismic data for 500 and 100 traces symmetric about the projected trace, respectively (Figure 12). Convolving each wavelet with the respective modeled reflectivity series produced the raw synthetics. The seismic expression of the formation tops prior to synthetic calculation was unknown as no seismic profiles imaging the upper subsurface are publically available in western Arkansas, and shallower images along strike in eastern Oklahoma are rendered unreliable due to the dramatic shift in correlative facies; however, the initial qualitative match between the Milton #1-2 synthetic and its projected trace in the interval below the Mansfield Sandstone and above the lower Alma Sandstone corresponds well with the expected time position of the formations from the integrated time-depth relationship. Similarly, high amplitude responses in the Cheesman #1 synthetic at the Johns Valley Formation and Jackfork Group tops match the
Figure 12 – Parameters of extracted wavelets used in synthetic construction. Graph A shows the power in decibels (dB) of the signal across a frequency spectrum in Hertz. Graph B displays the wavelet extracted from the seismic data, showing amplitude of the function versus time in milliseconds. Graph C shows the wavelet phase, or initial angle of the function, across a frequency spectrum in Hertz.
seismic data well at the expected time position. After pinning the synthetic at these high confidence tie points, minor stretching and squeezing of the modeled trace with the correlative trace produced the calibrated synthetics used for formation top identification (Figure 13).
Figure 13 – Synthetic seismograms and well-to-seismic ties. The Milton #1-2 and Cheesman #1 synthetics shown with well log data and interpreted formation tops.
Process of Depth Conversion

Proper structural interpretation of seismic reflection profiles requires conversion of seismic data from two-way travel time to depth, which is accomplished through the process of depth migration or depth conversion. Depth migration is a processing technique where geophysicists focus and adjust the lateral position of reflectors during processing using derived velocity functions (Etris et al., 2001). Depth conversion is accomplished by multiplying the time difference between horizons by an interval velocity to fit seismic time to depth, often calibrated to formation tops if well data are available. Modeling the velocity structure of a layer is accomplished through the Adlinvel equation:

\[ V = V_0 + K \cdot (Z - Z_0) \]

where \( V \) equals the instantaneous velocity, \( V_0 \) equals the initial velocity of the layer, \( K \) equals the velocity gradient or factor by which velocity changes vertically, and the quantity \( (Z - Z_0) \) equals the distance between a depth and the top of the interval (Snodgrass, K., pers. comm., 2013).

Beginning with a complete horizon interpretation in the time domain, an interpreter must decide how many and which horizons to use to define the velocity model based on analysis of interval velocities. Interval velocities and their rate of change with depth are typically calculated from local well data. Changing the number, top, initial velocity, or velocity factor of an interval will affect the resulting depth conversion.
Velocity Model for Depth Conversion

Depth conversion was accomplished using the velocity modeling function within Petrel using local well-based velocities. A time interpretation of seismic line F (Figure 4) served as the foundation for the resulting velocity model (Figure 14). Seismic line F strikes approximately perpendicular to the local structural trend across the Waldron and northern Boles quadrangles and provides the highest quality images of the structures. Furthermore, the projection of high-confidence formation tops available for the C.J. Williams #1 well 0.4 miles (0.6 kilometers) along strike to the well-imaged strata underlying Poteau Mountain provides subsurface control as little structural deviation over this distance is expected.

Early velocity models were assigned interval velocities by converting available DT log values into velocities after the methodology of Bailey (2007). The Milton #1-2 well contains a continuous DT log from above the upper Carpenter Sandstone within the upper Atoka Formation to the top of the Spiro Sandstone near the base of the lower Atoka Formation and provided velocities for this stratigraphic interval. The Cheesman #1 well is spudded in the middle Atoka Formation and reaches TD in the upper Jackfork Group. In this well, an incomplete DT log provides initial velocities for the lowermost lower Atoka Formation through the upper Morrowan section. Despite the missing section of DT log within the lower Atoka Formation in the Chessman #1 well (Figure 10), initial velocities were calculated over the entire stratigraphic interval when the Milton #1-2 and Cheesman #1 wells were considered together.
Analysis of the log-based velocities demonstrates that the Atoka Formation does not exhibit dramatic shifts in character, but rather increases steadily from 12,500 ft/s to 14,200 ft/s (3,810 m/s to 4,328 m/s) with increasing depth due to its gross lithologic homogeneity. The presence of high velocity Morrowan carbonate units below the Spiro Sandstone results in slightly elevated velocities. No well in the region penetrates the section below the upper Jackfork Group; therefore, little is known about the velocity structure below these units. To honor the structural control provided by the C.J. Williams #1 well, the interval velocities were varied between selected seismic horizons to match the corresponding well depth. To keep consistency with mapped units, the Atoka Formation was subdivided into three intervals:

1) the base of the Spiro Sandstone to the top of the lower Carpenter Sandstone to define the lower Atoka Formation at an average velocity of 13,800 ft/s (4,207 m/s),

2) the top of the lower Carpenter Sandstone to the top of the Mansfield Sandstone to define the middle Atoka Formation at an average velocity of 13,200 ft/s (4,024 m/s), and

3) the top of the Mansfield Sandstone to the top of the upper Carpenter Sandstone to define the upper Atoka Formation at an average velocity of 11,400 ft/s (3,476 m/s).

Although base of the Hartshorne Sandstone marks the top of the upper Atoka Formation, the upper Carpenter Sandstone is used as a working top in the velocity model because it is the highest logged marker. The Desmoinesian section was not subdivided and was assigned a constant velocity of 11,500 ft/s (3,506 m/s), a velocity that forced the base of the interval, the upper Carpenter Sandstone, to match its corresponding depth in the C.J. Williams #1 well. The interval between the topographic surface and the above processing datum was assigned the original replacement velocity of 10,000 ft/s (3,049 m/s).
As the C.J. Williams #1 well does not penetrate stratigraphy below the Spiro Sandstone, no logged depth is available for calibration of interval velocities for the underlying section. Therefore, the sub-Spiro Sandstone stratigraphy is grouped into two intervals in the velocity model based on lithology:

1) the carbonaceous units are assigned to a single interval between the top of the Jackfork Group and the base of the Spiro Sandstone at a constant velocity of 14,736 ft/s (4,493 m/s), calculated from the Chessman #1 DT log.

2) the units below the top of the Jackfork Group are assigned an initial velocity of 13,500 ft/s (4,116 m/s) at the top of the zone after the simple single-layer velocity model of Roberts (2004) and a velocity gradient of -0.194 was applied. This velocity gradient lies within the range of commonly observed values (Snodgrass, K., pers. comm., 2013) and is calculated from the Carboniferous flysch interval of the PASSCAL Ouachita Experiment velocity model used to interpret the upper crustal structure of southwestern Arkansas (Lutter et al., 1990).

This six-layer velocity model honoring the C.J. Williams #1 well-based structural control for seismic line F (Figure 4) is applied to the same intervals imaged on subsequent seismic lines G and H (Figure 4) without modification.
CHAPTER 4
STRUCTURAL GEOLOGY

Three 2-D seismic reflection profiles that trend approximately perpendicular to the structures of interest within the Waldron and Boles quadrangles and provide the highest quality images of the subsurface were selected from the dataset and depth-converted based on the velocity model developed from seismic line F (Figure 14). The three depth-converted 2-D seismic reflection profiles are shown overlain on the geologic map of the Waldron and Boles quadrangles in Figure 15. These profiles trend approximately north-south; seismic profile 1 is cropped from the northern portion of original line G; seismic profile 2 is converted from the entire length of original line F; and seismic profile 3 is cropped from the northern portion of original line H (Figure 4). Both uninterpreted and interpreted profiles are shown in the depth domain from the topographic surface to -30,000 feet (-9,150 meters) with no vertical exaggeration (H=V). Formation tops interpreted within local wells (Figure 10) and correlated from geologic maps are shown as seismic horizons to illustrate structural geometry.
Figure 15 – Geologic map of the Waldron and Boles quadrangles. Numbered black lines show the location of depth-converted seismic profiles.
Seismic Reflection Profile 1

The cross section plane for seismic reflection profile 1 (Figure 15) strikes 345° for 6.8 miles (10.9 kilometers) within the northwestern Waldron quadrangle and lies approximately perpendicular to the major structural trend. The structural geometry (Figure 16) indicates an in-sequence thrust system merging at a tip line with a backthrust in typical triangle zone geometry (Figure 2). Due to an ambiguous package of reflectors in the north-dipping limb of the Poteau Syncline, two possible interpretations are presented.

The first interpretation (Figure 16) suggests the presence of a high-angle (60°) south-dipping normal fault (N1) in the footwall of a south-dipping thrust that ramps from pre-Jackfork Group through the Morrowan units and decreases in dip angle from 45° to a bedding-parallel flat within the middle portion of the shale-rich lower Atoka Formation. This is the Stanley Decollement (SD) detached within shale units at the base of the Stanley Group south of the seismic profile. The Choctaw Fault (CHF) is present in the subsurface of the Hon Anticline as a steep (60°) splay of the SD hinterland of the primary thrust surface, consistent with geologic mapping (Figure 15) that shows the CHF increasing in displacement to the west. The CHF offsets the Stanley Group and lower Jackfork Group but does not displace the upper Jackfork Group. Along the SD, offset of the top Jackfork Group horizon indicates dip separation of 2,100 feet (640 meters) that decreases up-dip before merging at a point of zero-displacement with a bedding-parallel backthrust within the lower Atoka Formation, the lower Atokan Decollement (LAD).
Figure 16 – Seismic profile 1 showing interpretation 1. **HT**: Hartshorne Sandstone **UC**: upper Carpenter Sandstone **UA**: upper Alma Sandstone **MAN**: Mansfield Sandstone **MA**: middle Alma Sandstone **LA**: lower Alma Sandstone **LC**: lower Carpenter Sandstone **SS**: Spiro Sandstone **CS**: Cromwell Sandstone **JV**: Johns Valley Formation **JKFK**: Jackfork Group
The LAD is the eastern continuation of the LAD mapped in the western Arkoma Basin in the Wilburton gas field area by Çemen et al. (2001a; 2001b). In seismic profile 1, the LAD continues to the south but is folded in the hanging wall anticline of the SD. Backthrusting along the LAD is created by the relative motion of the northward transported footwall passively uplifting the hanging wall, forming a wedge that delaminates the lower Atoka Formation as the tip line propagates into the foreland. The amount or existence of bedding-parallel motion thrusting along the LAD north of the tip line could not be determined by the data available.

The second possible interpretation (Figure 17) retains the geometry of the SD, CHF, and N1 faults in the footwall of the LAD, but differs where a splay of the LAD is shown to offset the lower Carpenter Sandstone (top lower Atoka Formation) and merges with a north-dipping bedding-parallel backthrust at a tip line within the middle Atoka Formation. The in-sequence thrust splaying from the LAD exhibits a listric shape down-dip from its termination in the structurally higher tip line and is attributed to later folding during development of the Hon Anticline. This geometry is similar to the listric thrusts observed by Banks and Warburton (1986) in the structurally highest and foreland most structures of the Kirthar and Sulaiman Ranges of Pakistan.

The structurally and stratigraphically higher backthrust is the Carbon Fault (CF) because it occupies a similar structural position to the CF in eastern Oklahoma (Çemen et al., 2001a; 2001b); however, the CF is not mapped in western Arkansas (Figure 15) and lateral continuity of the CF in eastern Oklahoma with the interpreted CF in western Arkansas is not clear because of the lack of study between these areas. The CF bounds the north-dipping limb of the Poteau Syncline as an inclined bedding-parallel flat within the middle Atoka Formation approximately 900 feet (275 meters) below the top of the lower Alma Sand. At its up-dip limit,
Figure 17 – Seismic profile 1 showing interpretation 2. HT: Hartshorne Sandstone UC: upper Carpenter Sandstone UA: upper Alma Sandstone MAN: Mansfield Sandstone MA: middle Alma Sandstone LA: lower Alma Sandstone LC: lower Carpenter Sandstone SS: Spiro Sandstone CS: Cromwell Sandstone JV: Johns Valley Formation JKFK: Jackfork Group
the CF cuts up section, increasing dip angle to 20° creating 450 feet (135 meters) of dip separation observed in the offset of the lower Alma Sand before emerging at the topographic surface within the crest of the Hon Anticline (Figure 17).

The need for two interpretations centers on a package of bright reflectors near the change in dip of the LAD from moderately north-dipping to horizontal between -5,000 and -10,000 feet (-1,525 and -3,050 meters) (Figure 18). The first interpretation suggests the lower Carpenter Sandstone is folded but continuous from a bright reflector to the north to a less bright reflector to the south. It implies that the vertical change in reflector strength is independent of the position of the lower Carpenter Sandstone. The second interpretation suggests that the brightness of the reflectors does not change vertically and that the apparent offset is due to a north-dipping thrust that soles into the folded LAD to the south. Both interpretations are valid based on available data.
Figure 18 – Comparison of interpretations 1 and 2 of seismic profile 1. UC: upper Carpenter Sandstone UA: upper Alma Sandstone MAN: Mansfield Sandstone MA: middle Alma Sandstone LA: lower Alma Sandstone LC: lower Carpenter Sandstone SS: Spiro Sandstone CS: Cromwell Sandstone JV: Johns Valley Formation JKFK: Jackfork Group
Seismic Reflection Profile 2

The cross section plane for seismic reflection profile 2 strikes 002° for 16.2 miles (25.9 kilometers) within the Waldron quadrangle and crosses the Ross Creek Fault (RCF) in the northern Boles quadrangle imaging the hanging wall south to Buffalo Mountain (Figure 15). Due to the plunge of the folds in the Waldron quadrangle and the strike of the RCF, minor apparent dip is introduced along the central portion of the line while true structural dip is preserved along the northern and southern portions of the profile. The structural geometry suggests a dominantly forward-propagating series of thrusts with two detachment levels toward the foreland that form a triangle zone that is composed of three wedges (Figure 19).

At the southern edge of the profile, the south-dipping RCF intersects the topographic surface as a high angle (80°) bedding parallel thrust immediately north of High Point Mountain (Figure 15). The RCF exhibits the listric characteristic of decreasing dip angle with depth to a bedding-parallel flat detached within the upper Jackfork Group at -15,000 feet (-4,575 meters). The hanging wall shows moderately south-dipping reflectors of the lower Atoka Formation near the surface that similarly decrease in dip with depth. Morrowan stratigraphy below the basal Atokan Spiro Sandstone and above the upper Jackfork Group detachment surface is interpreted at the base of the hanging wall block and extends to the shallow subsurface. The RCF splays into two segments near the topographic surface and ramps up-section, offsetting the south-dipping limb of a minor hanging wall anticline of sub-Spiro Sandstone section.
The upper wedge interpreted along profile 2 is a triangular region of chaotic reflectors in the immediate footwall of the RCF and a north-dipping backthrust within the lower Atoka Formation (Figure 20). This backthrust continues north where it occupies the same structural position as the LAD in profile 1; therefore, it is interpreted as the southern continuation of the LAD where it cuts shallowly up section before emerging south of Ross Mountain (Figure 15). This poorly-imaged region is floored by a bedding-parallel thrust within the Union Valley Formation above the Cromwell Sandstone from where it enters the profile from the south to a ramp that merges with the LAD at the northern edge of the chaotic zone and is probably composed entirely of ductily deformed shale of the lower Atoka Formation. The hanging wall of the LAD consists of north-dipping middle and upper Atoka Formation that is passively uplifted in an in-sequence series of thrusting. The unconstrained depositional thickening of the lower Atoka Formation and poor imaging prohibit reasonable seismic interpretation and estimation of the magnitude of shortening accommodated within the upper wedge. Therefore, the interpreted position of the Spiro Sandstone near the base of the structure is schematic and represents a minimum (undeformed) bed length.

The middle wedge is located below the Waldron Syncline and consists of a thrust sheet within the lower Atoka Formation and upper Union Valley Formation, and contains the basal Spiro Sandstone (Figure 21). The thrust sheet is bound by sigmoidal south-dipping thrusts D1 and D2 that splay from the up-dip limit of the decollement within the Union Valley Formation (floor thrust) and sole into the LAD (roof thrust) (Figure 21). Thrust D1 separates the upper wedge from the Spiro Sandstone-involved thrust sheet of the middle wedge. Thrust D2 is the basal thrust of the middle wedge and soles into the LAD and its up-dip limit. Within the thrust sheet, a minor south-dipping
Figure 20 – Detailed image of the upper wedge of seismic profile 2. **MA**: middle Alma Sandstone **LA**: lower Alma Sandstone **LC**: lower Carpenter Sandstone **SS**: Spiro Sandstone **CS**: Cromwell Sandstone **JV**: Johns Valley Formation **JKFK**: Jackfork Group
Figure 21 – Detailed image of the middle wedge of seismic profile 2. LC: lower Carpenter Sandstone SS: Spiro Sandstone CS: Cromwell Sandstone JV: Johns Valley Formation JKFK: Jackfork Group
thrust offsets the Spiro Sandstone before displacement dies out into the overlying section.

The lower wedge consists of the Stanley Group through the lower Atoka Formation folded in a hanging wall anticline detached above the SD, which passively transports the overlying upper and middle wedges northward in-sequence. The frontal edge of the lower wedge (Figure 22) is the lateral continuation of the structure interpreted in profile 1 as the SD ramps above the normal fault, N1. The SD cores the fault propagation fold before merging with the LAD at the tip line underlying the Poteau Syncline. The interpretation does not show a stratigraphically and structurally higher tip line within the middle Atoka Formation as seen in the second interpretation of profile 1 (Figure 17), requiring the CF and associated north-dipping listric thrust to die out west of profile 2. North of normal fault N1, a second south-dipping normal fault N2 is interpreted to offset the pre-Morrowan stratigraphy, but does not offset the Jackfork Group (Figure 19).
Figure 22 – Detailed image of the lower wedge of seismic profile 2. UC: upper Carpenter Sandstone UA: upper Alma Sandstone MAN: Mansfield Sandstone MA: middle Alma Sandstone LA: lower Alma Sandstone LC: lower Carpenter Sandstone SS: Spiro Sandstone CS: Cromwell Sandstone JV: Johns Valley Formation JKFK: Jackfork Group
Seismic Reflection Profile 3

The cross section plane for seismic reflection profile 3 strikes 341° for 10.0 miles (16 kilometers) through the eastern Waldron quadrangle and crosses the trace of the RCF near the northeastern corner of the Boles quadrangle (Figure 15). Apparent dip on the order of 4° is expected along the northern portion of the profile due to the curvature of the traces of the major fold axial surfaces; however, the southern portion of the profile images true structural dip as it is near perpendicular to the structural trend. The structural geometry indicates two detachment levels forming an in-sequence series of thrusts described along seismic profile 2 with a few key lateral variations (Figure 23).

At the southern edge of the profile, the RCF is interpreted as a subvertical, bedding-parallel, south-dipping listric thrust detached within the upper Jackfork Group that decreases in dip to horizontal near -13,500 feet (-4,115 meters). At the surface, the hanging wall exposes south-dipping strata of the lower Atoka Formation; seismic interpretation shows the same Morrowan stratigraphy at the base of the hanging wall, forming a faulted anticline near, but not exposed at, the surface underlying Dutch Creek Mountain (Figure 15). The upper wedge described in profile 2 is present and exhibits a triangular region of uninterpretable and chaotic reflectors bound to the south by the RCF and to the north by the LAD (Figure 24). The footwalls of the RCF and LAD are composed of ductily deformed shale of the lower Atoka Formation. Within the upper wedge, conformable reflectors are apparent above the Spiro Sandstone, suggesting that the floor thrust bounding the base of the ductile duplex is detached within the
Figure 23 – Seismic profile 3. **MC**: McAlister Formation **HT**: Hartshorne Sandstone **UC**: upper Carpenter Sandstone **UA**: upper Alma Sandstone **MAN**: Mansfield Sandstone **MA**: middle Alma Sandstone **LA**: lower Alma Sandstone **LC**: lower Carpenter Sandstone **SS**: Spiro Sandstone **CS**: Cromwell Sandstone **JV**: Johns Valley Formation **JKFK**: Jackfork Group
Figure 24 – Detailed image of the upper wedge of seismic profile 3. MAN: Mansfield Sandstone MA: middle Alma Sandstone LA: lower Alma Sandstone LC: lower Carpenter Sandstone SS: Spiro Sandstone CS: Cromwell Sandstone JV: Johns Valley Formation JKFK: Jackfork Group
lower Atoka Formation.

The most significant change in structural geometry between profile 2 and 3 is that the middle wedge consisting of the Spiro-involved thrust sheet is not present (Figure 25). The structural level continuous with the middle wedge in profile 2 to the west is well-imaged and shows undeformed and continuous reflectors, leaving no ambiguity concerning the absence of the middle wedge.

The lower wedge is composed of Stanley Group through lower Atoka Formation detached above the SD, similar to the geometry interpreted along profile 2. The northward termination of the lower wedge is the lateral continuation of the same tip line structure interpreted in profiles 1 and 2 (Figure 26). The triangle zone exhibits a box-like geometry due to a north-dipping blind thrust creating a drape fold in the strata above the Jackfork Group. The CHF offsets the Spiro Sandstone before displacement dies out toward the frontal limb of the Hon Anticline (Figure 23). The SD tips out below the Poteau Syncline north of the profile edge and slip is transferred to the LAD as a backthrust.
Figure 25 – Detailed image of the lack of a middle wedge in seismic profile 3. **LC:** lower Carpenter Sandstone **SS:** Spiro Sandstone **CS:** Cromwell Sandstone **JV:** Johns Valley Formation **JKFK:** Jackfork Group
Figure 26 – Detailed image of the lower wedge of seismic profile 3. LC: lower Carpenter Sandstone SS: Spiro Sandstone CS: Cromwell Sandstone JV: Johns Valley Formation JKFK: Jackfork Group
CHAPTER 5

DISCUSSION

Accommodation Zone Structure

The trace of the Choctaw Fault (CHF) in western Arkansas strikes east-west and tips out in the north-dipping limb of the Hon Anticline within the Waldron quadrangle (Figure 15), where it overlaps the east-west to southwest-northeast striking trace of the Ross Creek Fault (RCF) approximately six miles (ten kilometers) to the south within the Boles quadrangle (Figure 15). This en echelon arrangement of the leading-edge thrusts that define the transition from the Ouachita fold-thrust belt to the Arkoma foreland basin is consistent with the surface expression of an accommodation zone, where slip is transferred between separate faults as a decrease in displacement along one fault segment coincides with an increase in displacement along a parallel fault segment sharing a common decollement (Figure 27) (Dahlstrom, 1969). Accommodation zones maintain displacement along a continuous belt of deformed rock despite the limited lateral extent of a single fault (Dahlstrom, 1969).

Displacement along the northern edge of the Ouachita fold-thrust belt is accommodated by the CHF in Oklahoma and the RCF in Arkansas (Figure 1). Structural interpretation of depth-converted 2-D seismic reflection profiles demonstrate that the RCF is the dominant leading-edge thrust of the frontal Ouachitas in north-central Scott County, western Arkansas (Figure 19, 23). The CHF is interpreted as a blind thrust in the footwall of the RCF that merges with the south-
Figure 27 – Schematic cross sections demonstrating the three dimensional geometry of an accommodation zone. From the upper right to lower left, slip along thrust fault A decreases to zero while slip along thrust fault B reaches a maximum in the center of the accommodation zone and thrust fault C increases from zero. Modified from Dahlstrom (1969).
dipping Stanley Decollement (SD) (Figure 16, 17, 19, 23). At the southern end of seismic profile 2 and 3, the RCF is interpreted as a thrust flat within lower Atokan shale; however, the geometry of an accommodation zone requires that the RCF sole into a basal decollement within the Stanley Group south of the seismic data.

The RCF, LAD, and SD thrusts bound a triangular region internally composed of three stacked wedges. The CHF is interpreted within the lower wedge (Figure 17, 17, 19, 23); however, as displacement increases to the west, the Wilburton Triangle Zone (WTZ) develops within the footwall (Figure 9). The geometry of an accommodation zone predicts that displacement along the RCF decreases to the west with increasing displacement along the CHF, although the nature of the westward termination of the RCF is not well known (Arbenz, 2004). The triangle zone in the footwall of the RCF also may die out to the west with the increasing structural dominance of the CHF and WTZ in Oklahoma, although further study is required to document the structure to the west along strike.
Implications of Mechanical Stratigraphy

Structural interpretation of depth-converted 2-D seismic reflection profiles in western Arkansas demonstrates that the footwall of the Ross Creek Fault (RCF) consists of hinterland-dipping thrust sheets above blind thrusts that sole into a foreland-dipping backthrust at a tip line that merges with the relatively undeformed strata of the Arkoma basin (Figures 16 through 26). These structural elements are consistent with the general framework of the triangle zone concept (Figure 2); the internal structure of the triangle zone interpreted in western Arkansas contains multiple detachment levels forming three stacked wedges that best fit the description of a Type II triangle zone based on the classification of Couzens and Wiltschko (1996).

Strain partitioning between the frontal Ouachitas and southern Arkoma Basin among upper, middle, and lower wedges is a product of the interaction of Early Pennsylvanian to Early Permian compression with Late Mississippian to Early Pennsylvanian pre-compressional structures and is controlled by the heterogeneous mechanical properties of the Late Mississippian to Middle Atokan stratigraphy. Based on the lithology and observed structural behavior of the stratigraphy elsewhere in the orogen (Arbenz, 2004), the mechanical behavior can be predicted from a composite stratigraphic column (Figure 28) and is used to explain the development and location of particular structures. Thick sandstone and carbonate units are competent, while the shale-rich units are more likely incompetent (Çemen et al., 2001a; 2001b; Arbenz, 2004). Shale of the Stanley Group and Atoka Formation are incompetent as these units host detachments or are internally deformed, while Jackfork Group sandstone and the Spiro Sandstone are considered competent and form rigid thrust sheets (Figure 28).
Figure 28 – Stratigraphic column showing mechanical behavior and location of detachments. Modified from Arbenz (2004).
The upper wedge consists of a poorly imaged region in the footwall of the RCF and lower Atokan Decollement (LAD), and is composed of tectonically thickened lower Atoka Formation in a ductile duplex (Figures 20 and 24). The lower Atokan section is composed of incompetent shale suggested by GR values from the Cheesman #1 well that increase from 90 API to 240 API upwards through the formation indicating clean, thick shale beds (Figure 10). A ductile duplex in the immediate RCF footwall is further supported by the zones of poor seismic imaging. These zones are also recognized in the frontal portion of the southern Appalachians in Alabama and Georgia (Figure 29) and are known informally as MUSHWAD (Malleable Unctuous SHAle Weak-layer Accretion in a Ductile duplex) (Thomas, 2001; Pashin et al., 2012). Along profile 2, the middle wedge consists of a thrust sheet of lower Atoka Formation, including the Spiro Sandstone. Thrusting of the Spiro Sandstone is absent along profile 3. This interpretation requires that displacement along the middle wedge dies out laterally eastward along the 3.5 miles (5.6 kilometers) separating profiles 2 and 3. The 5,000 feet (1,524 meters) of displacement of the lower wedge along the SD remains consistent in profiles 1, 2, and 3; however, the blind Choctaw Fault (CHF) increases in displacement in profile 3, offsetting the Spiro Sandstone and folding the LAD before dying out in the core of the Hon Anticline. Due to the lack of piercing points along the CHF in profiles 1 and 2, the increase in displacement in profile 3 is not quantifiable. Below the SD, normal faults due to flexural bending of the uppermost crust (Van Arsdale, 1990; Bradley and Kidd, 1991) are present. Two normal faults, N1 and N2, are interpreted at the northern end of profile 2; N1 is interpreted on profiles 1 and 3. Fault N2 extends to but does not offset the top of the Jackfork Group, which requires displacement prior to Late Morrowan deposition. The footwall of the SD is well imaged to the base of the profiles to the north; however, imaging quality decreases below the top Jackfork Group toward the south. Despite the
Figure 29 – MUSHWAD structure of the southern Appalachians. Modified from Pashin, et al. (2012).
imaging quality, N1 is interpreted in the SD footwall below the top Jackfork Group.

The presence of deep-water facies of the Ouachita Basin section (Stanley Group-Jackfork Group-Johns Valley Formation) in the N1 hanging wall represents a major facies change from the shallow water carbonate of the Morrowan shelf to the north (Sutherland, 1988). The facies change coupled with N1 suggests that the transition zone lies near the location of the Morrowan shelf and fits the syndepositional model proposed by Sutherland and Manger (1979) for the Stanley Group-Johns Valley Formation section above a major down-to-south normal fault (Figure 8). The presence of N1 is further supported by regional COCORP seismic profiles of the frontal thrust belt showing a major down-to-south normal fault occupying the same structural position along strike in eastern Scott County (Lillie et al., 1983).

The Stanley Group and Johns Valley Formation contain olistostromes of exotic carbonate blocks to the south (Figure 30). These carbonate blocks suggest exposure or gravitational instability of shallow water strata of the North American continental platform section in the footwall of a large active normal fault during deposition of the Stanley Group and Johns Valley Formation. Reinemund and Danilchik (1957) report the mining of large blocks of Ordovician dolomite within the Johns Valley Formation near Boles, Arkansas, within the Boles quadrangle and footwall of the RCF (Figure 15), requiring a proximal source still further to the north, consistent with the position of N1.

N1 affected the structural development of the transition zone during Late Pennsylvanian thrusting. The N1 footwall presented a significant barrier to thrusting within the Stanley Group and acted as a buttress, forcing the SD to ramp through the competent sandstone of the Jackfork Group before flattening into lower Atokan strata where it meets the LAD.
Figure 30 – Field photographs showing olistostromes within the Ouachita flysch. Carbonate boulder within the Johns Valley Formation (left) and carbonate cobble within the Stanley Group (right).
This style of ramp-flat thrusting controlled by pre-existing basement normal faults is also present within the Black Warrior Basin of Alabama (Figure 31) (Thomas and Bayona, 2005; Groshong et al., 2010).

**Figure 31** – Schematic cross section showing the response of growth faulted sections to thrust ramp localization. Modified from Thomas (1982).
Kinematic Evolution

The kinematic evolution of the triangle zone in western Arkansas is mostly unconstrained due to a lack of structural relationships that impose a specific deformation history outside of in-sequence thrusting along the Ross Creek Fault (RCF), Choctaw Fault (CHF), and Stanley Decollement (SD) that transfers slip to the lower Atokan Detachment (LAD) at a tip line that propagates into the foreland. The kinematic evolution presented here is one possible evolution consistent with the data presented and is not a unique solution. The structural configuration of the southern Arkoma basin prior to Late Pennsylvanian compression consists of the undeformed and southward-thickening Atoka Formation above the growth-faulted Stanley Group, Jackfork Group, and Johns Valley Formation lying above and separated from shallow water carbonate in the N1 footwall (Figure 32, panel A). Initiation of thrusting at the location of the present transition zone commenced with the establishment of the RCF to the south (Figure 32, panel B) after deposition of the lower formations of the Krebs Group above the Hartshorne Sandstone, indicating thrusting began after 313.4 Ma. Erosion of the hanging wall cutoff prevents accurate estimation of the magnitude of displacement (Arbenz, 2004). Continued compression resulted in the formation of a ductile duplex within the lower Atoka Formation, tectonically thickening the formation to approximately twice original depositional thickness above a detachment within the Union Valley Formation and offset of the Spiro Sandstone along thrust D1 that joined the LAD at a tip line, forming an incipient triangle zone that constitutes the upper wedge of the present triangle zone geometry (Figure 32, panel C).
Figure 32 – Schematic reconstruction of the kinematic evolution of the triangle zone in western Arkansas. Reconstruction based on geometry interpreted along profile 2. RCF: Ross Creek Fault LAD: Lower Atokan Detachment UVD: Decollement with the Union Valley Formation D1: Hinterland thrust of the middle wedge duplex D2: Foreland thrust of the middle wedge duplex SD: Stanley Detachment N1: southern pre-Atokan normal fault N2: northern pre-Atokan normal fault CHF: Choctaw Fault HT: Hartshorne Sandstone (top of upper Atoka Formation) MAN: Mansfield Sandstone (top of middle Atoka Formation) LC: lower Carpenter Sandstone (top of lower Atoka Formation) SS: Spiro Sandstone (base of lower Atoka Formation) JV: Johns Valley Formation (top of growth-faulted section).
Continued compression resulted in the propagation of the decollement within the Union Valley Formation and establishment of thrust D2 north of D1, forcing the LAD and tip line into the foreland, further uplifting the passive roof above the LAD, and completing the development of the thrust sheet belonging to the modern middle wedge (Figure 32, panel D). Slip was then transferred from the upper and middle wedges to the basal SD, which ramps from the base of the Stanley Group into the lower Atoka Formation due to the buttress effect of the pre-orogenic carbonates in the footwall of the south-dipping N1 (Figure 32, panel E). Ramping of the SD into shale of the lower Atoka Formation synchronously establishes a new tip line with the LAD north of the middle wedge. The final stage of Late Pennsylvanian thrusting forming the current triangle zone geometry in western Arkansas is the out-of-sequence blind thrusting of the CHF, which splays from the SD south of and in response to the buttressing effect of the N1 footwall (Figure 32, panel F). Formation of the CHF in the subsurface increases the structural elevation of the hanging wall anticline composing the lower wedge, during which the LAD is passively folded.

Although the maximum age of thrusting within the modern transition zone is constrained by the presence of deformed Desmoinesian stratigraphy, it is not possible to quantify the duration of compression and the age of each stage with the available data. Regional study by Houseknecht and Kacena (1983) suggest that compression ceased by the end of Pennsylvanian time at 298.9 Ma. This age allows for a maximum duration of deformation of 14.5 million years.
Structural Analogues

Triangle zones are structures commonly recognized at the boundary between fold-thrust belts and associated foreland basins (DeCelles and Giles, 1996); however, triangle zones are not recognized in several orogenic systems (Figure 33). The factors controlling the development of triangle zones are not fully understood but may be related to the mechanical stratigraphy of the deforming section and the pre-existing structure prior to compression (Couzens and Wiltschko, 1996). Empirical evidence compiled from known triangle zones, or transition zones containing a backthrust sense of motion, are more likely to form when the deforming section is composed of contrasting competent and incompetent units, allowing a wedge of competent strata to detach above (floor thrust) and below (roof thrust) an incompetent horizon (Couzens and Wiltschko, 1996). It is unknown whether triangle zone formation is a continuous process at work during orogenesis or is restricted to late stage evolution of a fold-thrust belt (Jones, 1996); however, incompetent shale-rich horizons provided by burial of late synorogenic deposits host backthrusts in numerous triangle zones worldwide (Couzens and Wiltschko, 1996; Couzens-Schultz and Wiltschko, 2000). Finite-element and sandbox models also suggest that the presence of a structural backstop promotes triangle zone formation, as the energy required to form a backthrust is reduced (Jamison, 1996). The geometry of the Waldron Triangle Zone is consistent with these requirements, having developed within thick synorogenic deposits of contrasting mechanical behavior (Figure 28) above a large south-dipping normal fault that exposes a competent footwall block as a barrier to northward thrusting. Although transition zones lacking a triangle zone (Figure 33) may be the result of bias due to erosion of all or part of the structure, a deforming
A section composed dominantly of competent units without layered incompetent horizons inhibits backthrusting and are instead expressed by forethrusts with decreasing displacement toward the foreland (Couzens and Wiltschko, 1996).

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<tr>
<th>Orogenic Belt</th>
<th>Frontal Thrust</th>
<th>Geometry</th>
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<tr>
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<tr>
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<td>Jura Mountains</td>
<td>Forethrust</td>
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**Figure 33** – Chart showing the location of orogenic belts with dominant backthrust or forethrust sense of motion at the foreland limit of thrusting. Those orogenic belts with a backthrust sense of motion contain triangle zones while orogenic belts lacking a significant backthrust do not contain a triangle zone. Modified from Couzens and Wiltschko (1996).
Comparison with the Wilburton Triangle Zone

The triangle zone of western Arkansas exhibits geometries consistent with frontal structures of both the North American Cordillera and the southern Appalachians; however, the most significant and relevant comparison is with the Wilburton Triangle Zone (WTZ) of eastern Oklahoma (Çemen et al., 2001a; 2001b), 40 miles (64 kilometers) to west along strike and occupying the same structural position at the southern boundary of the Arkoma basin (Figure 1) (Arbenz, 2004).

The primary difference in structural geometry between the two triangle zones is that the upper wedge of the triangle zone presented here (Figure 20, 24) is composed of a ductile duplex of lower Atoka Formation, while the intercutaneous wedge of the WTZ involves the younger middle Atoka Formation, which is broadly folded (Figure 9). This example illustrates the transient appearance of ductile duplexes. Additionally, the LAD in western Arkansas shifts to a north-dipping backthrust from a south-dipping in-sequence thrust within the WTZ.

The geometry of the Spiro Sandstone-involved thrust sheet of the middle wedge presented here is similar in major respects to the duplex at the base of the WTZ (Figure 9) (Çemen et al., 2001a; 2001b) as each imbricate the Spiro Sandstone between a floor thrust and roof thrust in age-equivalent formations. The single thrust sheet in the middle wedge interpreted along profile 2 coupled with its absence in profile 3 shows that the structure dies out within western Arkansas.
Within the WTZ, the Woodford and Springer detachments mark the base of thrusting (Figure 9). In western Arkansas, the base of thrusting also lies within Mississippian shale of the Stanley Group. Although no wells penetrate Early Mississippian strata in the frontal Ouachitas or southernmost Arkoma basin, this deepest detachment is inferred at the base of the Stanley Group (Caney Shale equivalent), based on well imaged stratigraphy down to the presumed top of the Arkansas Novaculite at -30,000 feet (-9,150 meters) at the southern end of profile 2, increasing in structural elevation to -10,000 feet (-3,050 meters) where it tips out into the LAD. The presence of a major decollement in the Stanley Group has been previously proposed by Arbenz (2004); its continuation into the transition zone area is inferred from outcrop relationships by Suneson et al. (2005) who concluded that the SD is the eastern continuation of the Woodford Detachment of the WTZ (Figure 9). Previous work in the Wilburton area of Oklahoma also suggests that south-dipping normal faults analogous to N1 and N2 underlie and likely control the placement of ramps in the Woodford and Springer Detachments (Figure 9) (Çemen et al, 2001a; 2001b; Ronck, 1997; Medhi, 1997; McPhail, 2001).

Because of the similarity in structural position and geometry, the triangle zone of western Arkansas is the eastern continuation of the WTZ of eastern Oklahoma (Çemen, et al., 2001a; 2001b); however, due the changing frontal thrusts and facies changes to deep-water strata, the triangle zone in western Arkansas is the Waldron Triangle Zone, named after the nearby town of Waldron, Arkansas. Further study is required to document the geometry of the Ouachita fold-thrust belt to Arkoma basin transition zone between the WTZ in the Wister Lake area, Oklahoma and the Waldron Triangle Zone area of Arkansas (Figure 3).
Interpretations of depth-converted 2-D seismic reflection profiles in north-central Scott County, Arkansas show an accommodation zone between the leading-edge thrusts of the Ouachita fold-thrust belt. Slip along the Choctaw Fault dies out in the subsurface north of the Ross Creek Fault, which continues as the leading-edge fault of the fold-thrust belt to the east. The accommodation zone between the Choctaw Fault and Ross Creek Fault serves as the boundary between the highly deformed frontal Ouachitas and the mildly deformed southern Arkoma basin. Strain partitioning between the fold-thrust belt and foreland basin is accommodated by a triangle zone consisting of three wedges in the footwall of the Ross Creek Fault. The present structural geometry evolved through both in-sequence and out-of-sequence thrusting during the Ouachita Orogeny in Late Pennsylvanian time. This triangle zone is the lateral equivalent to the Wilburton Triangle Zone of eastern Oklahoma; however, due to the changing associated frontal thrusts and involved stratigraphy, the triangle zone of western Arkansas is called the Waldron Triangle Zone.

The upper wedge of the Waldron Triangle Zone consists of a region of ductily deformed shale of the lower Atokan Formation bound by two oppositely dipping thrust faults, the Ross Creek Fault and the lower Atokan Decollement, similar to the ductile duplex or MUSHWAD recognized in the foreland basins of the southern Appalachians. The middle wedge consists of a
thrust sheet of lower Atoka Formation that contains the regionally extensive Spiro Sandstone; however, this thrust sheet is only present in profile 2. The thrust sheet dies out eastward through the southern Waldron quadrangle. The north-dipping lower Atokan Decollement serves as the roof thrust of both the ductile duplex and the Spiro Sandstone-involved thrust sheet, which soles into and accommodates slip transferred from the south-dipping Stanley Decollement, or floor thrust of the lower wedge, via a tip line below the Poteau Syncline. The location and mechanics of the Stanley Decollement ramp down-dip of the tip line is controlled by strain localization above a major south-dipping pre-Atokan normal fault, N1.

The geometry and subsurface formations present indicate that normal fault N1 is likely the source of the large exotic blocks of early Paleozoic shelf strata observed within the Stanley Group and Johns Valley Formation to the south. The Waldron Triangle Zone is likely the site of the Late Mississippian and Early Pennsylvanian shelf, consistent with the fault-controlled depositional model for the flysch units (Stanley Group, Jackfork Group, and Johns Valley Formation) of the Ouachita Basin section.

The presence of the Waldron Triangle Zone in the footwall of the Ross Creek Fault is consistent with previous studies that find deforming strata of variable mechanical strength at the foreland limit of thrusting in an orogenic belt typically host triangle zones. Furthermore, the presence of the Waldron Triangle Zone in the Ross Creek Fault footwall demonstrates that the triangle zone structure at the transition from the Ouachita fold-thrust belt to the Arkoma basin is not an ephemeral feature of the orogenic belt and persists across the accommodation zone eastward into central Arkansas. The surface tip-out and decreasing subsurface displacement along the Wilburton Triangle Zone and Choctaw Fault into western Arkansas also shows that the footwall structures follow the same accommodation zone pattern as the leading-edge thrusts.
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