THE GEOMETRY OF THRUST SYSTEM IN THE WILBURTON GAS FIELD AND SURROUNDING, LATIMAR COUNTY, OKLAHOMA

BY

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CHAPTER I

INTRODUCTION

The Arkoma Basin is well-recognized as a foreland basin of the Late Paleozoic Ouachita Orogeny. The basin contains a thick sequence of sedimentary rocks over a Paleozoic rock succession. These rocks succession is represented by a Cambrian-Devonian sedimentary sequence. The initiation of the Arkoma basin is usually considered to coincide with the Pitkin and Fayetteville Formations, units of the Mississippian Delaware Creek Shale or Caney Shale (Johnson, 1989). The basin contains a thick sequence of Pennsylvanian Atokan stage rocks. The base of the Atokan is the Spiro Sandstone which is a major reservoir in the Arkoma Basin. The study area (Figure 1) includes twelve townships in the Arkoma Basin (T.3N.- 6N., R.17E.-19E.), which are the parts of the Latimar and Pittsburgh Counties. This study is undertaken as part of the OCAST project investigating Overthrusted Gas Reservoirs in the Arkoma basin. The Spiro Sandstone produces gas from the overpressured reservoirs in the Wilburton gas field.

The main objectives of this investigation are to determine:

1. The geometry of thrust system in the Wilburton gas field and surrounding areas.
2. The nature of deformation along the thrust surfaces (especially the Spiro Sandstone).
3. The position of regional detachment horizons in the Wilburton field and surrounding areas.

This investigation deals mainly with the subsurface structure of the Wilburton Gas Field and surrounding areas. The previous investigators had a long debate about the presence of a roof thrust over the duplex structure of the Spiro Sandstone, and the

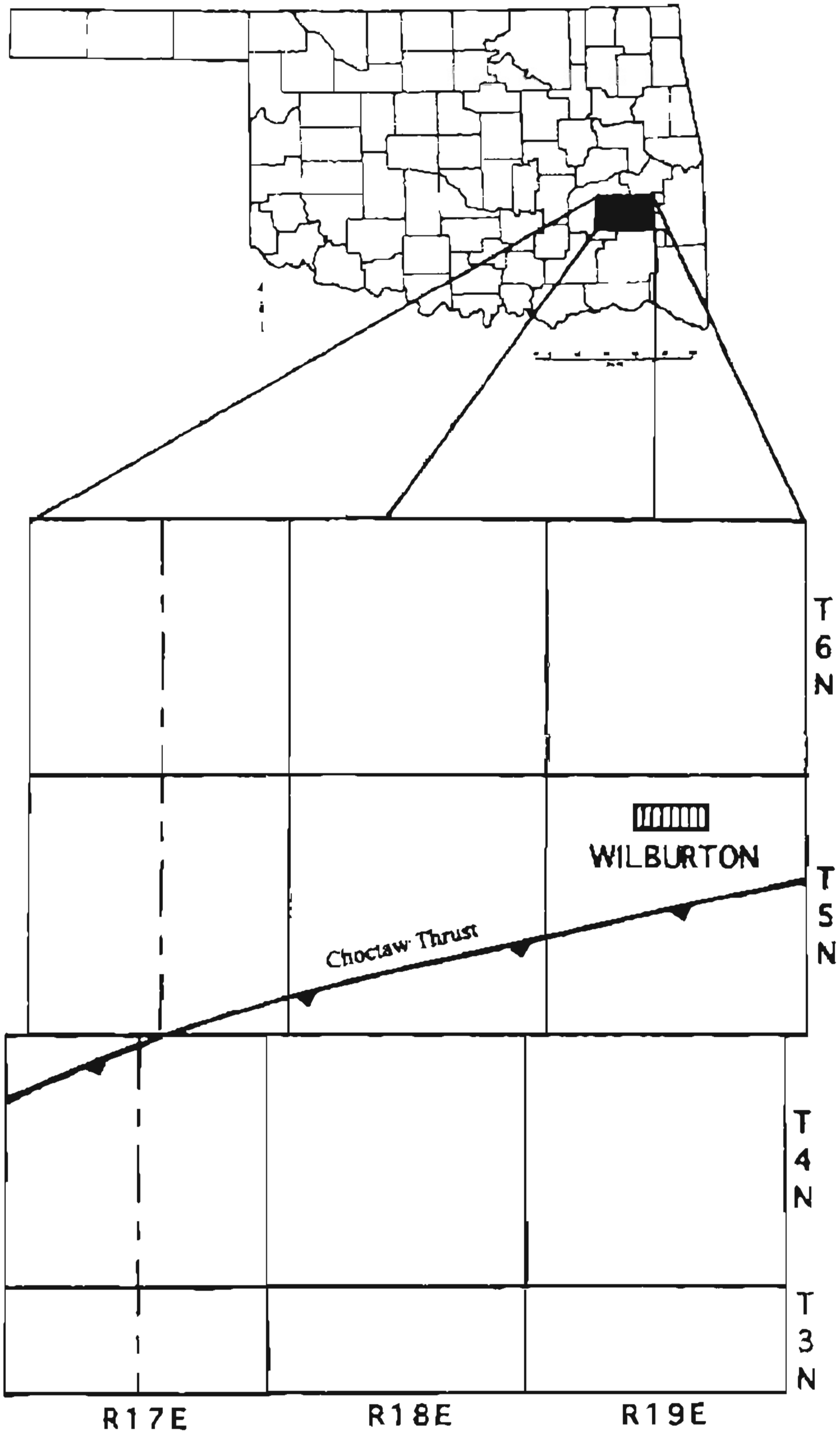


Figure 1.--Location map showing study area and location of the Choctaw fault.

geometry of the Choctaw and Carbon faults. The Carbon fault is thought to be a south dipping fault by some investigators, while others presented evidence that the Carbon fault is north-dipping fault.

METHODS OF INVESTIGATION

In order to achieve the objectives of this study the following tasks were performed:

1. A surface geological map was prepared (plate I) based on the information obtained from geological maps prepared by Neil H. Suneson and Charles A. Ferguson (1989), and by several visits to the study area.
2. Subsurface information was obtained from induction logs, gamma ray logs of approximately 150 well logs (Appendix I).
3. The subsurface information was interpreted from a seismic reflection profile provided by EXXON Corporation.
4. Eight balanced cross sections (plates II-IX) were constructed to determine the geometry of the Wilburton gas field and surrounding areas. The cross-sections were constructed using the classic method outlined by Clint Dahlstorm (1969).
5. The structural cross sections were restored to calculate the amount of shortening in the Wilburton field area.

The interpretation of the available subsurface data, including the reflection seismic profile, was an essential part of the balanced cross-section preparation. The balanced cross-sections were constructed parallel to the tectonic transport direction to determine the accurate geometry of the thrust system in the Wilburton area. The Spiro Sandstone interval and the adjacent rock units, as defined in the subsurface of the study area are illustrated with the aid of a "type" electric log (Figure 2).

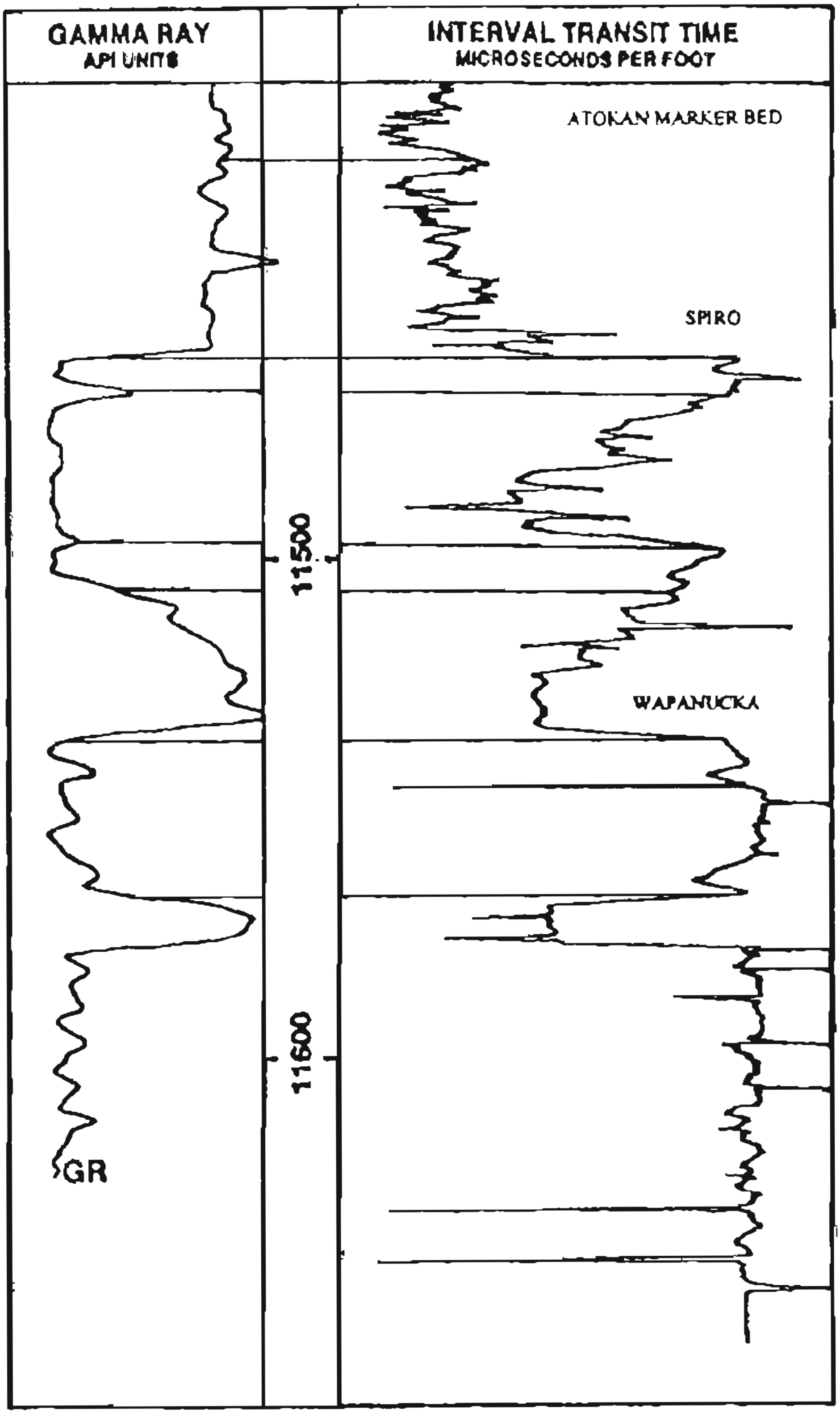


Figure 2.--Type log of the Spiro Sandstone interval and adjacent units.

PREVIOUS INVESTIGATIONS

Tectonics of Ouachita Mountains

The tectonic history of the Ouachita Mountains has been subject of much debate during the past two decades, numerous models have been proposed to explain the geological history of the Ouachita Orogenic belt and the associated foreland. Most of the tectonic interpretations have been focused on a scenario that involves consumption of oceanic crust and consequent collision between the southern margin of North America and a continental plate.

According to this scenario, the North American craton was sutured to a large landmass before the late Precambrian, the supper continent, "Proto-Pangea" (Walper, 1977). The major episode of rifting and the development of numerous rift basins along the southern margin of North America resulted in the opening of the Proto-Atlantic (Iapetus) ocean basin during the latest Cambrian or earliest Paleozoic (Figure 3A) (Thomas, 1977). As a consequent of this opening, the southern margin of the North America evolved into an Atlantic-type margin that remained through the middle Paleozoic (Figure 3B). The prism of sediments that accumulated along this passive margin includes Cambrian to lowermost Mississippian strata deposited on the shelf and off-the-shelf environments (Houseknecht, 1983). The shelf facies are present throughout the Arkoma Basin and the lower Paleozoic portion is believed to be extend southward beneath the highly deformed central uplift of the Ouachitas. The off-the-shelf facies are exposed only in allochthonous positions within the Ouachitas (Thomas, 1985). A recent study by Lowe (1985) suggested that the rifted Ouachita trough was "relatively" narrow during the Devonian system (Figure 4). During the Late Devonian, the Ouachita Geosyncline developed to the south of the passive continental margin, named Arkoma

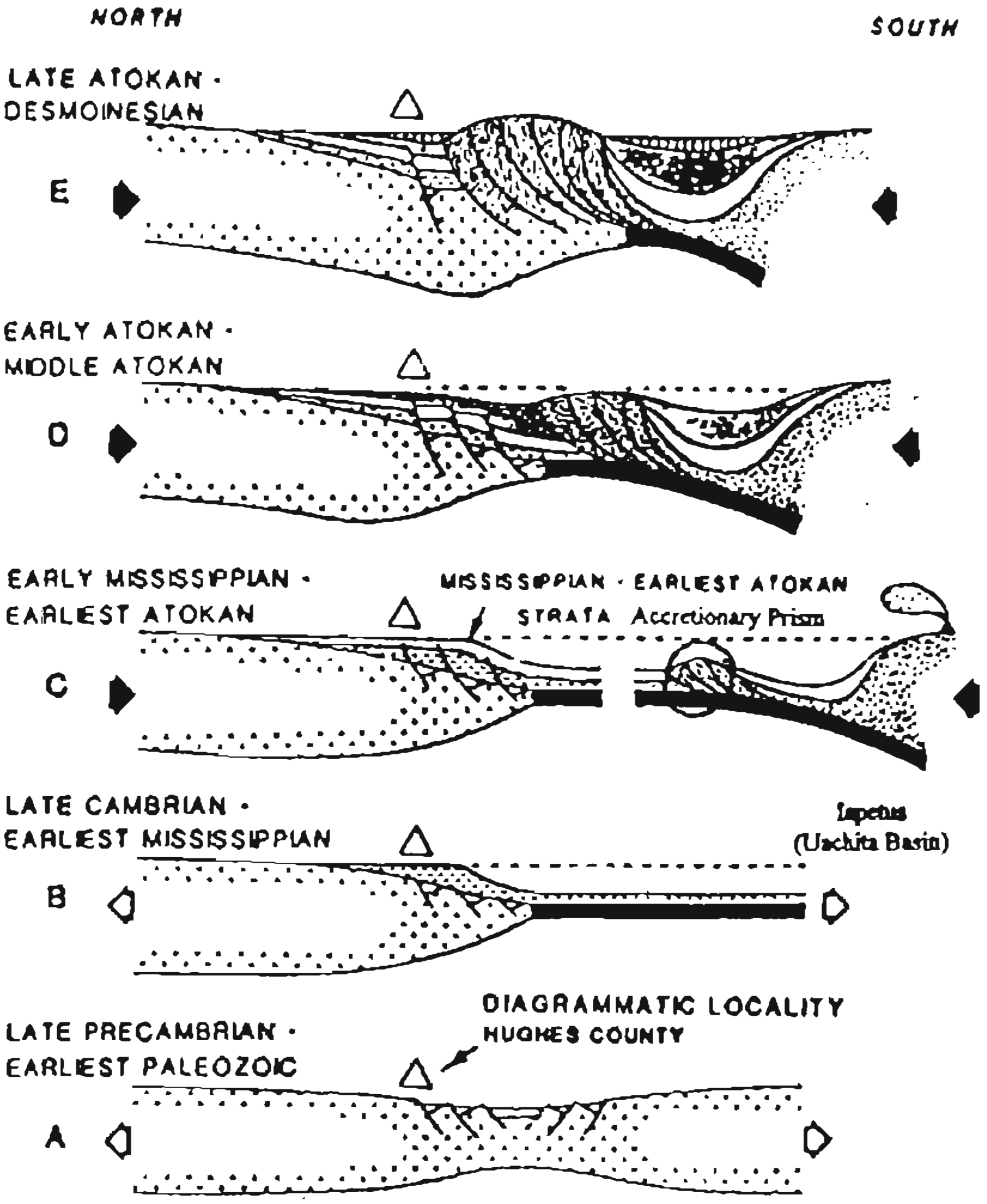


Figure 3.--Diagramatic evolution of Southern Margin of North America (modified from Houseknecht, 1983, p.15].

Shelf (Sutherland, 1988).

During the late Devonian or early Mississippian time (Figure 3C), the ocean basin began to close by southward subduction beneath a continental plate commonly called Llanoria (Houseknecht, 1983). Even though it is almost impossible to point out precisely when subduction began, it must have developed during the Mississippian time, as suggested by detritus indicative of an Orogenic provenance (Morris, 1974) and locally abundant volcanic detritus in the Stanley Formation of the Ouachitas. The Mississippian volcanic rocks that occur in the subsurface to the south of Ouachitas, along the flanks of the sabine uplift (Nicholas and Waddle, 1982) suggest the presence of a magmatic arc (Figure 3C) which developed along the northern margin of the Llanoria (Houseknecht, 1983). This convergent tectonic setting of the Ouachita Mountains also indicate that the Ouachita Orogenic belt began to form as an accretionary prism associated with the subduction zone (Figure 3C).

Throughout the interval from the late Mississippian and the early Atokan time, the remaining ocean basin had been consumed by subduction and the northward advancement of Llanoria was obducted on to the rifted continental margin of the North America (Figure 3D). The continued subduction resulted in the closure of the Ouachita ocean basin by Atokan time. The upper portion of the Wapanucka Limestone and shallow marine Foster and Spiro Sandstones were deposited on the southern Arkoma shelf during this time. The margin of the North American continent was partially drawn into southward-dipping subduction complex (Figure 3D) allowing vertical loading and "flexural bending " of previously rifted continental margin (Dickinson, 1974, and Houseknecht, 1983). Apparently, the resultant breakdown of the continental margin was initiated by a series of step-like normal faults in the foreland. The normal faults generally strike parallel to the Ouachita trend. They are mostly downthrown to the south, and offset both crystalline basement and overlying Cambrian to basal Atokan strata of the rifted margin prism (Houseknecht, 1983).

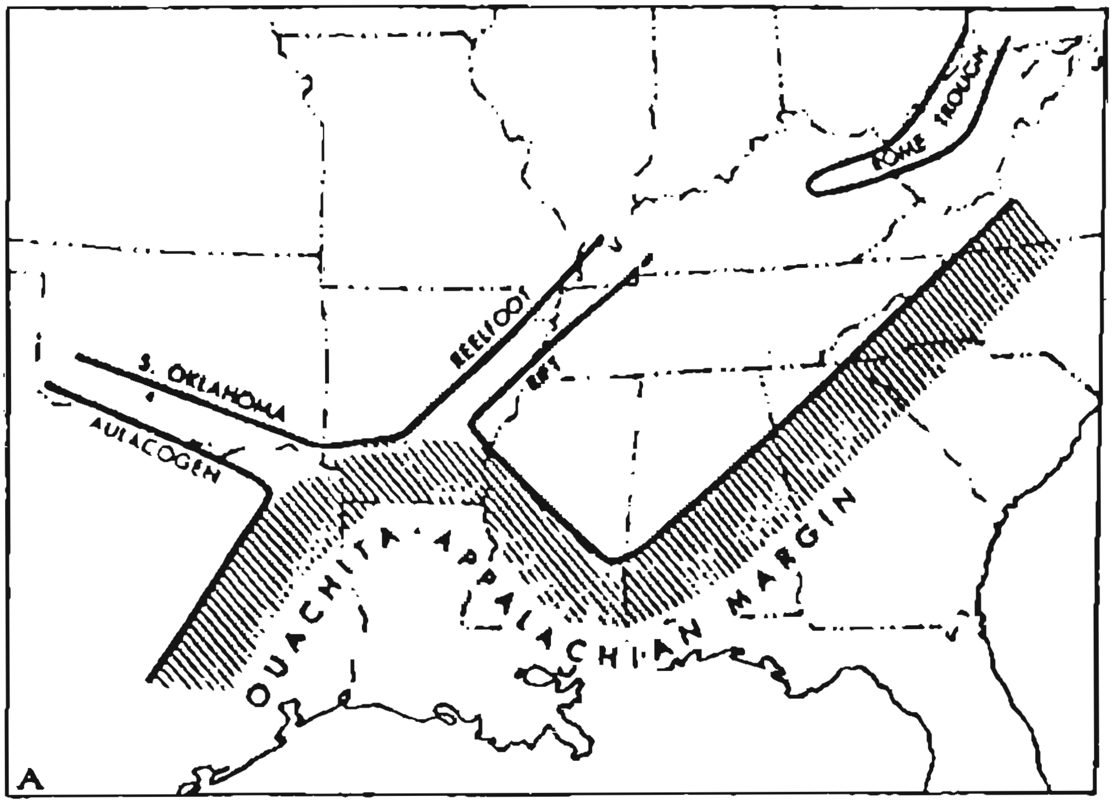


Figure 4.--Interpretation of rifted continental Margin (from Lowe, 1985,p.793).

By the late Atokan time (Figure 3E), the most intense collision events occurred, foreland-style thrusting became predominant and the subduction complex was uplifted as the Ouachita Mountains (Houseknecht, 1983). The resultant uplift along the frontal thrust belt of the Ouachitas completed the formation of a foreland basin (Dickson, 1974). Sutherland (1988) indicated that upper Atokan strata are not displaced by lower through middle Atokan syndepositional faults. Underwood and others (1988) reported that Desmoinesian strata were affected by compressive forces in a manner similar to compression of Atokan strata.

During the upper Atokan through the middle Desmoinesian time, a typical-coal bearing mollase sediments deposited from the Ouachita Orogenic Belt northward into the basin began, in addition, to the sediments derived from the Appalachian Mountains, Ozark Uplift, and Hunton Arc. The northern margin of the basin migrated northward through the Desmoinesian (Weirich, 1953). At Desmoinesian time the overall structural configuration of the Arkoma-Ouachita system was probably the same as at the present time.

Arkoma Basin

Cline (1960) mentioned that the term Arkoma Basin was defined by Hendricks in the late 1950's as the asymmetric structural depression between the southern flank of the Ozark Uplift (Boston Mountains) and the thrust front of the Ouachita Mountains.

Buchanan and Johnson (1968) were the first investigators to conclude that the Arkoma Basin is a foreland basin, where continental basement is overlain by an incomplete Cambrian through Morrowan platforms sequence of about 1,200 m of thickness. Houseknecht (1983) proposed that by the beginning of Desmoinesian time, the

Arkoma Basin had become a mild east-west topographic depression that was being filled by westward prograding deltaic system.

During the early Mississippian time the northern margin of the geosyncline become more active tectonically and a southward-dipping subduction complex evolved to the south. By the early Atokan time great strength of collision between the two plates occurred and this tectonic evolution led to the formation of Ouachita Mountains as the uplift and the foreland basin (Arkoma Basin) to the west of the Ouachita Mountains (Sutherland, 1988).

Arbenz (1989), while studying in the Arkoma Basin found Desmoinesian chert deposits whose tectonic southern boundary is well defined as the Choctaw fault. He also concluded that the Choctaw fault dies out in the westernmost Arkansas and the choice of the boundary between the Ouachita Thrust Belt and the Arkoma Basin becomes quite arbitrary. Arbenz (1989) considered that the Arkoma Basin is a mixed assemblage of structural styles. He described that in the southern two -thirds of the province, the surface geology and the shallow subsurface are dominated by a thin-skinned compressional fold belt of open mildly shortened synclines and anticlines detached from the underlying block -faulted lower Atokan and older rocks.

Wilburton Gas Field

The Wilburton gas field is located in the Latimar and Pittsburgh Counties, Oklahoma. It was discovered in 1929 and several wells drilled in the central part of Townships 5N., R. 18E., produced gas from an upper Atokan sandstone found at a depth of approximately 2,500 feet. In 1960, Ambassador Oil Corporation completed the #1 W. M. William's Unit, sec. 25, T. 5N., R. 18E., in Latimar County, as a discovery in the Spiro Sandstone.

Hendricks (1939) proposed that the Wilburton gas field underlies several prominent surface structural features. The gas field is on the east end of the Adamson anticline and the companion carbon fault, the Hartshorne Syncline, the Wilburton Anticline, and the southwest end of the North Wilburton Fold.

Lathan and Stuat (1963) stated that there is apparently a facies change between the Wilburton gas field and in wells drilled in sec. 13 and 17, T. 5N., R. 16E., in which the lowermost Atokan mainly consist of sandy limestone. They determined based on constructed cross-sections and available well-log data that the basal Atokan zone is not present west of Wilburton. By this explanation, they suggested that the depositional limit of the unit was somewhere west of Wilburton.

Berry and Trumbly (1967) drew the structural contour maps of the Wilburton field based on well control which show that the surface features continue downward in the subsurface as the middle Atokan. They also prepared some cross-sections and concluded that the Choctaw fault is a south-dipping thrust in the subsurface of the Wilburton gas field. Some unnamed thrusts within the subsurface were also recognized.

Tilford (1989) studied the Wilburton field using the well log and seismic data. He mentioned that the Wapanucka-to-basement rock units flooring the Arkoma Basin occur without depositional break from the shelf areas for tens of miles (at least) south of the Choctaw fault. Moreover, sometimes overturned thrust sheets of Spiro and Wapanucka-to-Cromwell age rocks, are bounded by Atokan or Morrowan shales. He suggested that this situation may be the result of axial plane thrusting of nearly isoclinal folds. He calculated that since 1960, Wilburton gas field has produced in excess of 1.1 TCF gas, primarily from Spiro Sandstone and secondarily from the Red Oak Formation.

CHAPTER II

GEOMETRY OF THRUST SYSTEM

Thrust systems are structural elements recognized as important components of most fold and thrust belts. Their geometries are very important in the kinematic balancing of geologic cross sections. They may provide clues to the kinematic and sequential development of geologic structures. Marshak and Mitra (1986) defined a thrust system as an exclusive collection of kinematically related faults that developed in sequence during a single regional deformation and are associated with deformation above a basal detachment. Generally there are two basic types of thrust systems (Figure 5):

1. An imbricate fan : and
2. A duplex.

Imbricate Fans

An imbricate fan is a type of thrust system in which faults cut up section from a basal detachment but do not rejoin at a higher stratigraphic level. There are three types of imbricate fans (Figure 5). First is the leading imbricate fan which has minor displacement on hanging wall of a major thrust (Figure 5 a). The second is the trailing imbricate fan which has its minor displacement on footwall (Figure 5 b). The third type of imbricate fan is termed as blind imbricate complex (5 c), in which the thrust tips lie in the subsurface. Its displacement is transferred upward into folds, either as a consequence

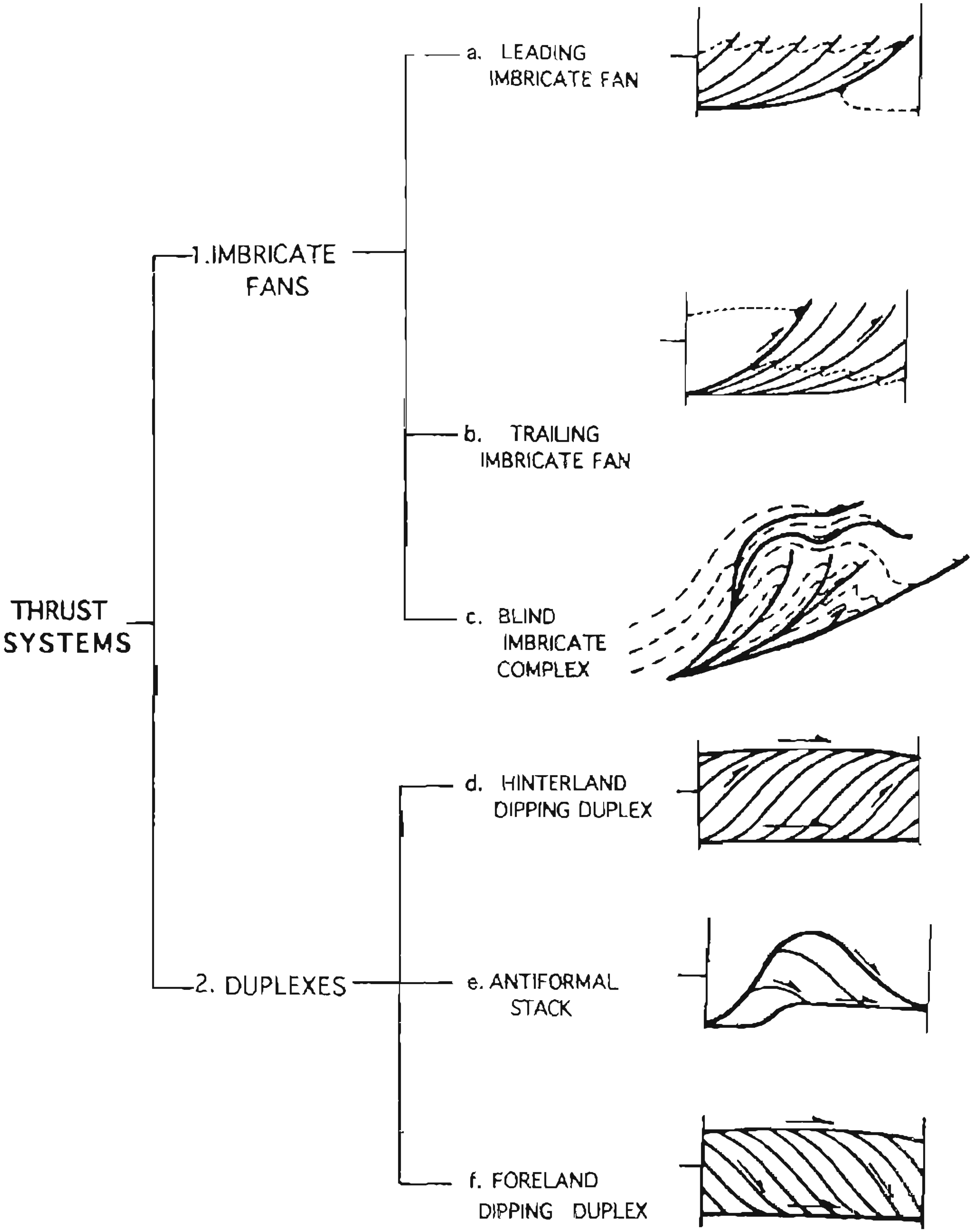


Figure 5.--Classification of different system of thrusts (after Boyer and Elliott, 1982).

of thrust propagation or by detachment. Blind thrust systems may contain large ramp folds and have geometrics similar to major thrust sheets, with the exception that the frontal thrust tips are not exposed.

Duplexes

A duplex is a thrust system in which the faults propagate upward from the basal detachment and rejoin to another detachment surface at a structurally higher level (Figure 5). A simplest duplex consists of a system of ramp-related anticlines with a floor thrust, a roof thrust, and a series of imbricate thrusts connecting the floor and roof thrusts. The faults cut up from the floor to the roof thrust surrounding by the bodies of rocks. These bodies, which are bounded on all sides by faults are called "horses". A duplex is also termed an imbricate family of horses, but it is possible for a horse to change along strike into an imbricate fan. This lateral change in the horses makes up a duplex. The change in shape of a horse is a result of oblique and lateral ramping giving the distinctive scoop shape of a roof, subsidiary, or floor thrusts. There is no limit to the size of a horse, it can vary from few meters to tens of kilometers. The size of horses plays an important role in determining the final geometry of duplex culminations. Generally, the duplexes exhibit a variety of forms depending on the amount of displacement of an individual horse.

Duplexes have characteristic internal features (Figure 5). The beds within a horse often trace out an elongate anticline-syncline fold pair, and the bedding near the central inflection point roughly parallels the subsidiary faults. Above and below the duplex the bedding may be relatively undisturbed, and for long distances a particular stratigraphic unit may compose the hanging wall of the roof or the footwall of the floor.

Sequential development of a duplex

There are two major geometric arguments for the successive development toward the foreland, or a forward progression of the subsidiary faults in duplexes. The oldest argument is best illustrated by Dahlstrom (1970), where a higher horse is folded over a lower one proving the forward development. This method is being used in this research. A second method of development is based on a series of simple graphic experiments developed by Boyer (1978). This method can be taken in account after constructing an idealized model based on typical dimensions and angles of observed duplexes assuming plane strains, constant bed lengths, and kink folding. In fact, the first major thrust with a slip (S_0) climbed upward in the section from a lower to an upper glide horizon making a steep footwall ramp in a more competent sequence. A second fracture with a slip (S_1) progrades from the base of the ramp for some distance in the lower glide horizon where it rejoins the major thrust. The first horse is formed. Afterward, the slip is transferred to next fracture (S_2) and the major thrust slips ($S_0 + S_1$) upward in front of the new horse which deactivate a portion of the major thrust that rides passively within the growing thrust sheet. Over the footwall ramp, the horse, which is an inactive portion of the major thrust and the rest of the overlying sheet are kink-like fold. And the process is repeated. The result is elongate folds, imbricate horses, undisturbed bedding above the roof thrust and same stratigraphic unit in the hanging wall along the roof thrust (Figure 6).

The third method proposed by Boyer and Elliot, 1982, in this method duplexes are mechanism for slip transfer from one glide horizon at depth to another at shallow levels. In the direction of movement, slip decreases along the floor and increases along the roof, and total slip at any point along floor or roof thrust is dependent on the number of horses which causes structural thickening, duplex growth, and an addition of mass to the moving thrust complex (Figure 7). Because the hanging wall rocks of the roof thrust are often flat-lying or only gently folded, one might suggest that the high-angle subsidiary faults

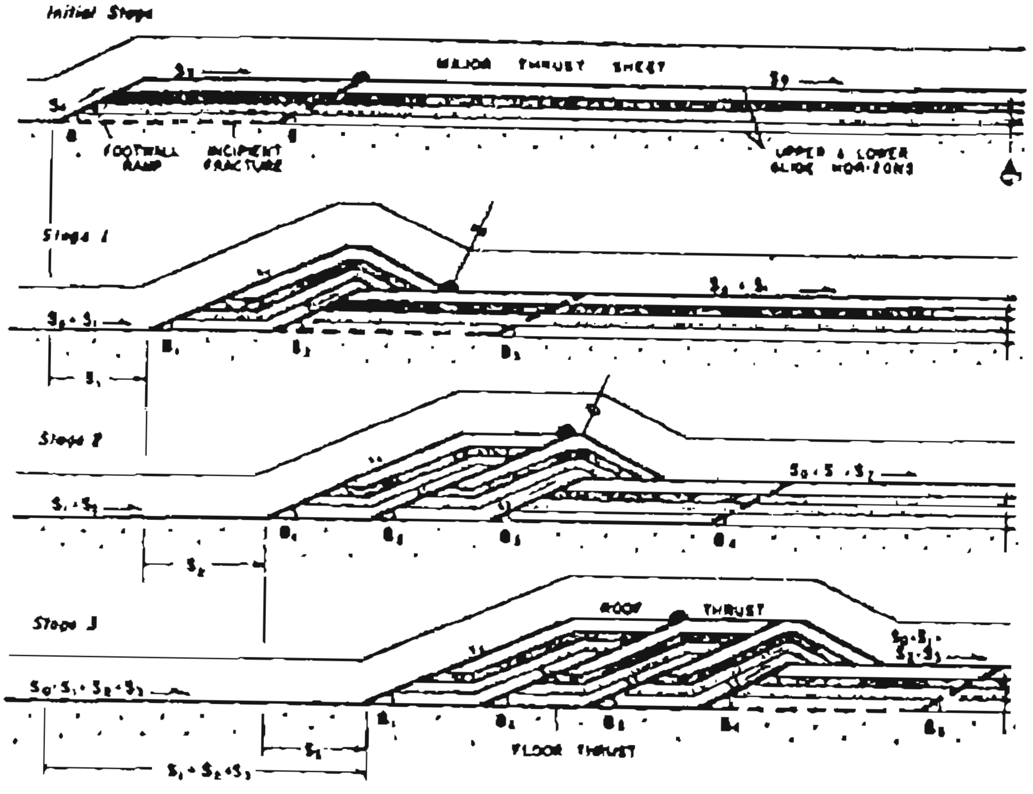


Figure 6.--Sequential development of duplex structures by measured graphic experiment (Boyer, 1978).

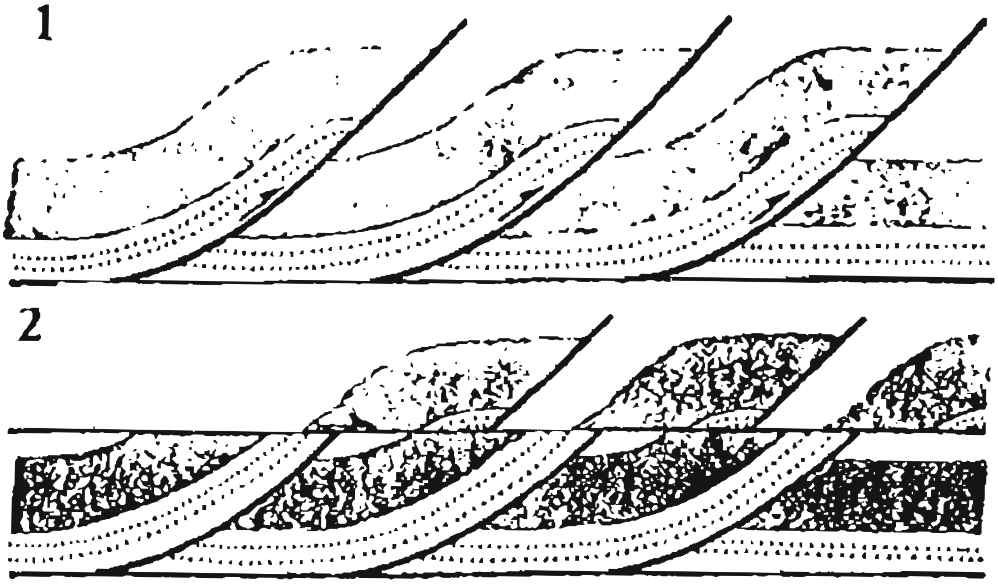


Figure 7.--Alternative method for developing duplex. Earlier formed imbricate faults in Stage 1, truncated by younger thrust in Stage 2 (Boyer & Elliott, 1982).

formed first by branching upward from the floor thrusts, then were truncated by the roof thrust (Figure 7).

A normal flat-roofed duplex will formed, if spacing of imbricates within a duplex and displacement are constant, and the single imbricate displacement is approximately half the pre-deformation spacing of imbricates (Woodward and others, 1985).

A hinterland dipping duplex is formed when the displacement of the horses are relatively small, they dip predominantly toward the hinterland and form a zone of roughly constant thickness between the roof and floor thrusts (Figure 5 d). In this kind of duplex, if the slip on the subsidiary faults is roughly equal to the length of the horse, the adjacent branch lines will bunch up and the horses will lie on the top of one another. This occurs from time to time in a normal duplex. On the other hand, if the bunching of the branch lines is widespread, the shingle-like imbricate pattern is destroyed (Boyer and Elliott, 1982). They are usually composed of upward-facing horses and always have a forward facing structure (Figure 5 d).

An antiformal stack is formed either when a duplex have a displacement equal to the spacing on the individual imbricates (Figure 5 e), or each higher horse is folded about the lower ones. This folding dies out downward and provides unambiguous evidence of sequence in which the horses are accumulated into the stack (Boyer and Elliott, 1982).

Foreland-dipping duplex is formed when displacement is twice the spacing and structurally dips towards the foreland (Figure 5 f). The dipping duplexes were first proposed by Woodward (1981). They usually consist of downward-facing horses (Figure 5 f), even when the beds are upside down and originated at a footwall syncline with a rolling hinge. They always have a forward structural facing like hinterland-dipping duplexes.

PROPOSED GEOMETRY OF THE WILBURTON GAS FIELD AND SURROUNDING

The geometry of the thrust system in the Wilburton gas field and surrounding areas has been the cause of debate among the geologists for many years. Most of the previous investigators came up with different results, especially the interpretation about the Carbon fault and the presence of a shallow triangle zone within the Wilburton gas field. Suneson (1995) summarized the proposed geometries of the thrust system in the frontal belt of the Ouachita Mountains. In essence, most of workers agreed that the thrust system in Wilburton field consists of a series of blind imbricate thrust faults, duplex structures, and associated folds. The series of thrusts is in the southern portion of the Arkoma Basin, south of the Latimar and Pittsburgh Counties.

Arbenz (1984) was the first who pointed out three main elements in the structural geology of the Arkoma Basin. He clearly recognized: 1. south-directed thrusts in the Arkoma Basin (presence of a shallow triangle zone); 2. blind imbricate thrusts extending basinward from the leading edge of the main part of the exposed fold and thrust belt; and, 3. a deep mostly bedding parallel decollement that served as the root of the imbricate thrusts and extended far into the basin (Figure 8).

Hardie (1988) introduced the concept of triangle zone in which Blanco thrust, located in T.3N., R.14E., is present as basinward roof. He showed in his cross-sections south-directed blind imbricate thrusts, backthrusts, and north-directed overturned folds beneath the Choctaw thrust and a basal detachment surface in Springer shale (Figure 9).

Milliken (1988) while studying the northern part of the Ouachita Mountains frontal belt, constructed a cross section showing a triangle zone floored by a north-directed imbricate fault and a basal detachment in Springer. During his investigation, he found deeper small displacement imbricate thrusts and backthrusts. He termed them "detached bi-vergent imbricates". The set of imbricates detach from the basal decolement

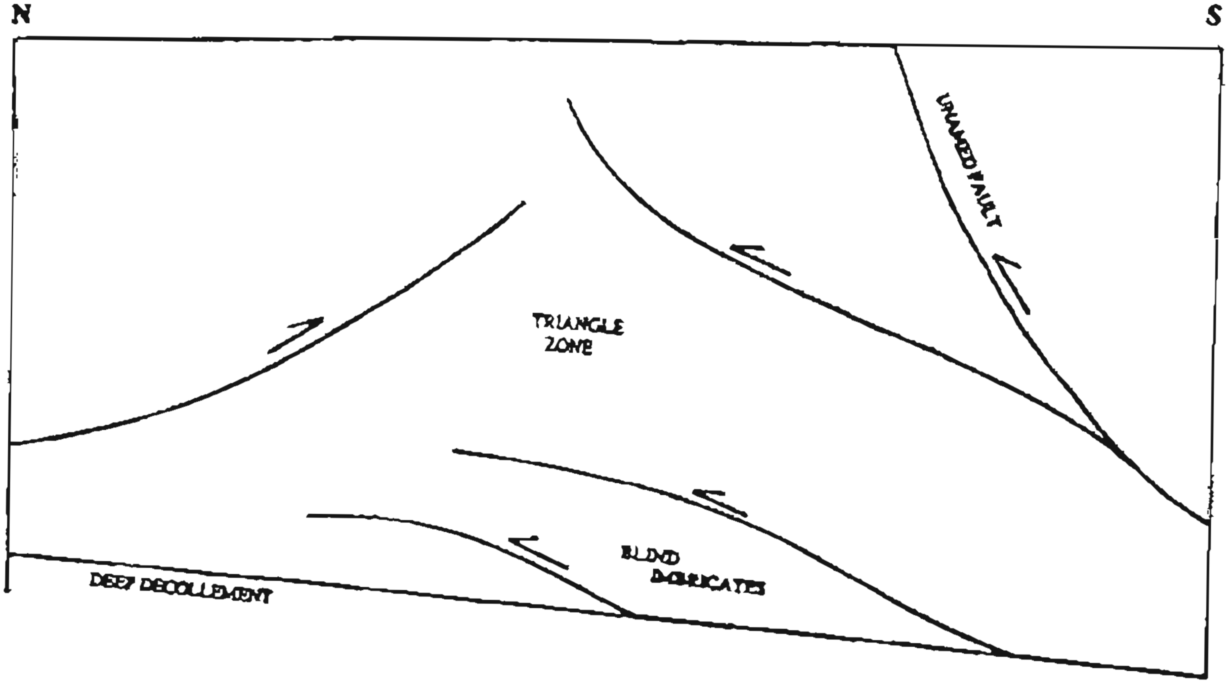


Figure 8.--Proposed thrust geometry of Ouachita Mountain frontal belt (Arbenz, 1984).

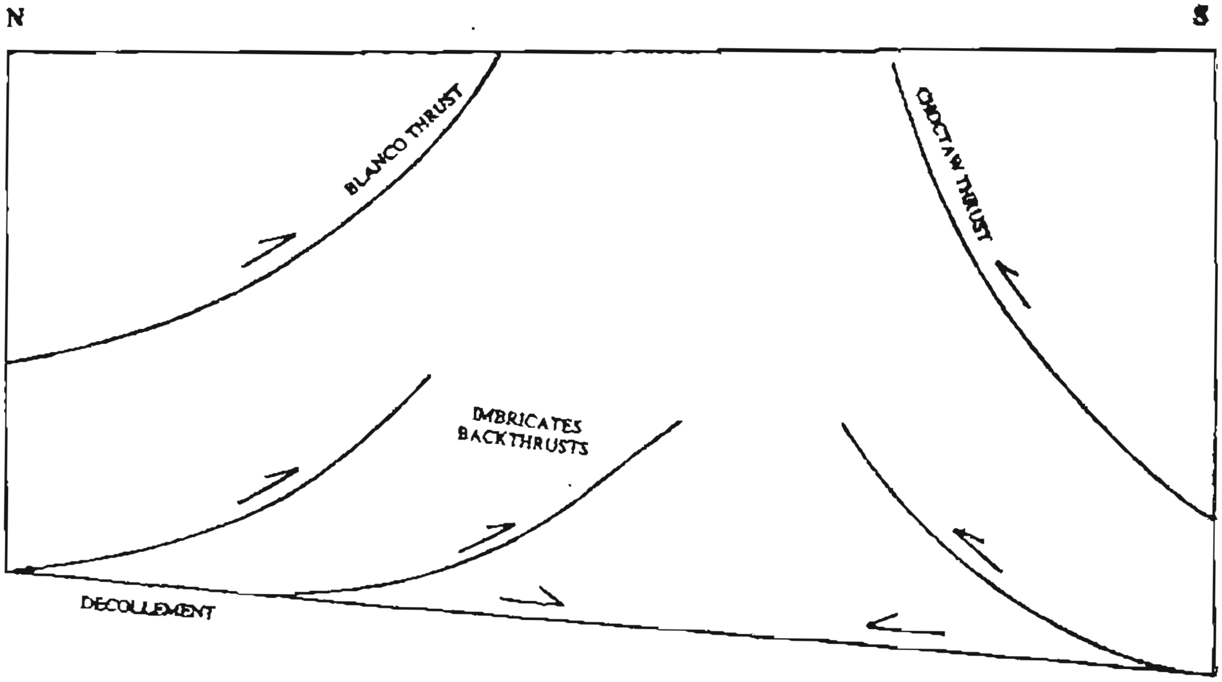


Figure 9.--Proposed geometry of Ouachitas showing decollement in th Springer Shale (Hardie, 1988).

and prograde northward as well as southward. These were interpreted to be beneath the major deep decollement, which extended southward beneath the imbricate thrust faults in the frontal belt. He also suggested that the detachment surface exposes on the surface as the Carbon fault instead of going basinward (Figure 10).

Camp and Ratliff (1989) constructed a computer generated balanced cross section based on surface mapping, seismic data, and well log information. They identified presence of a thick triangle zone floored by blind imbricate thrust faults, backthrusts and a deep north-directed detachment surface in Mississippian shale. Furthermore, they noticed imbricate backthrusts from the roof backthrust of the imbricate zone (Figure 11).

Reeves and others (1990) published two cross sections of northern frontal belt mostly based on seismic data. The first cross section showed complex structures similar to the "detached bi-vergent imbricates" of Milliken (1988). In contrast, the second cross section depicted a thin triangle zone floored by two north-directed duplex structures. Moreover, the detachment surface is relatively shallow (lower Atokan strata) and is separated by the duplex structures (Figure 12).

Perry and Suneson (1990) interpreted a north-south seismic profile across the frontal zone of the Ouachita thrust belt, and showed two triangle zones, in Hartshorne, Oklahoma (Figure 30). The deeper zone (passive roof duplex) underlying a shallow triangle zone, floored by foreland-directed imbricate thrust faults and south-directed backthrusts. They interpreted the Choctaw thrust as the frontal imbricate within the triangle zone. They placed a deep decollement in Woodford Shale. This interpretation was revised by Perry and others(1990); the roof thrust was interpreted as north-directed and the deep triangle zone as duplex structures consisting of north-directed imbricates and south-directed backthrusts (Figure 13).

Wilkerson and Wellman (1993) drew many cross sections across the Gale-Buckeye thrust system located to the southwest of the study area, based on closely spaced seismic lines and good well control. They identified a thin-triangle zone whose floor

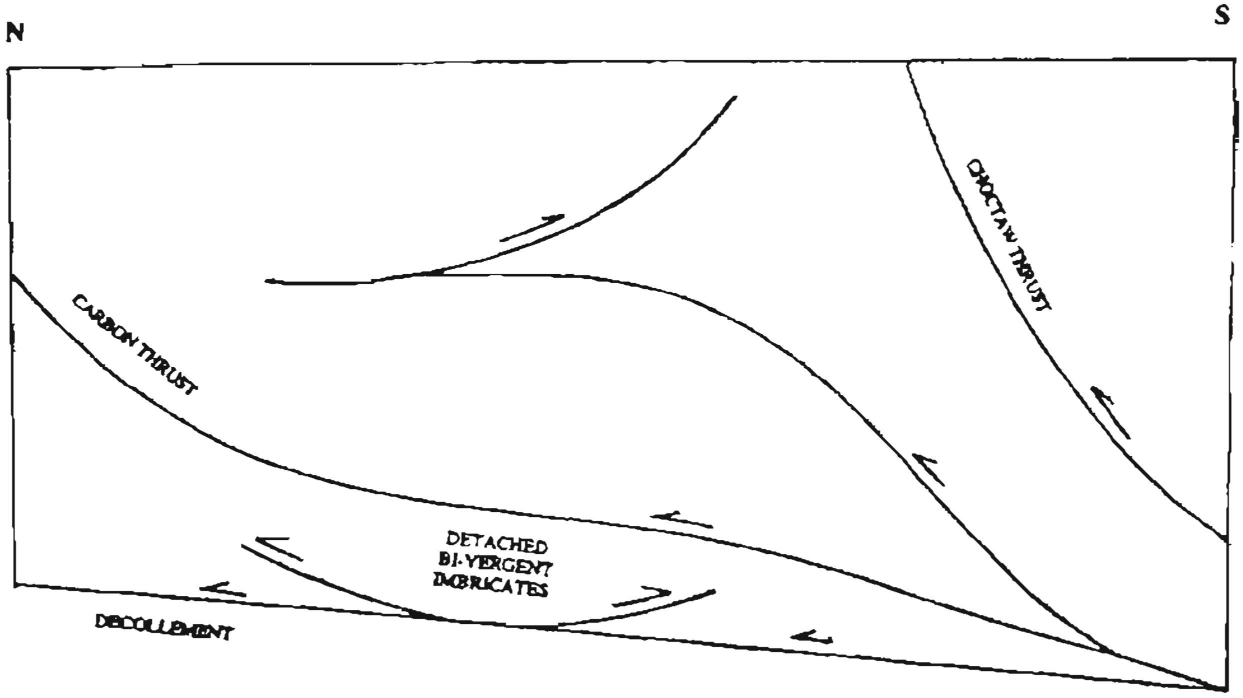


Figure 10.--Geometry showing detached bi-vergent imbricates and deep decolement (Milliken. 1988).

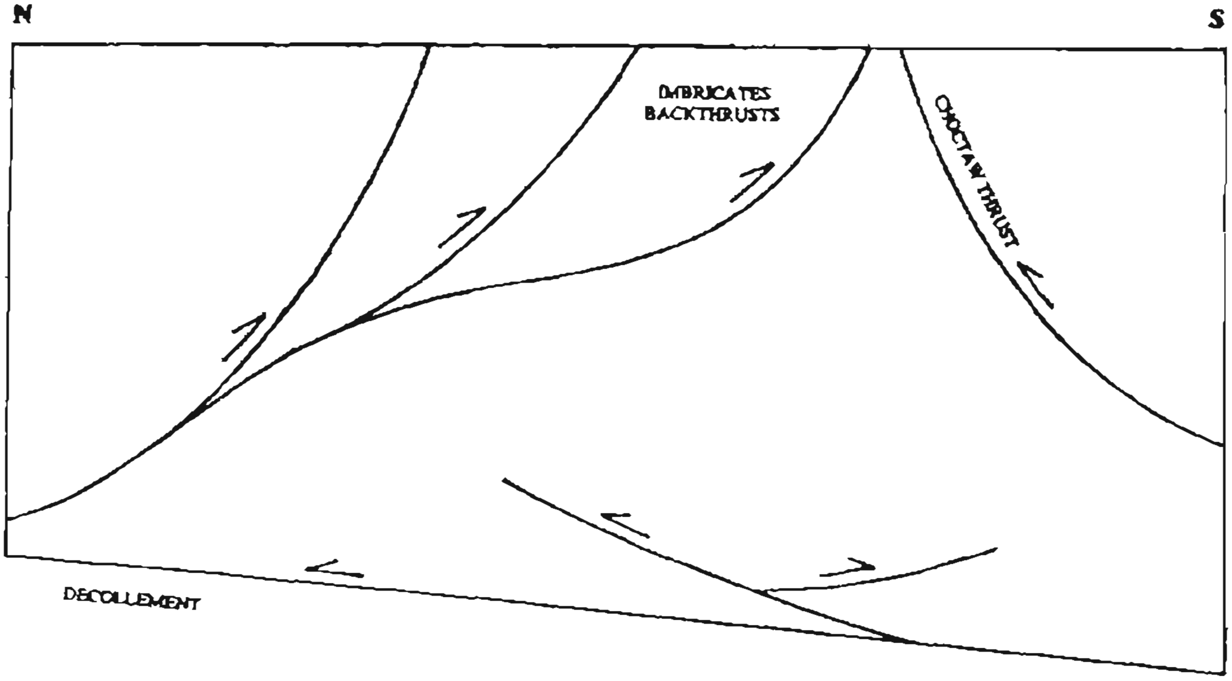


Figure 11.--Proposed geometry of the Ouachita Mountains transition zone near Wilburton (Camp and Ratliff, 1989).

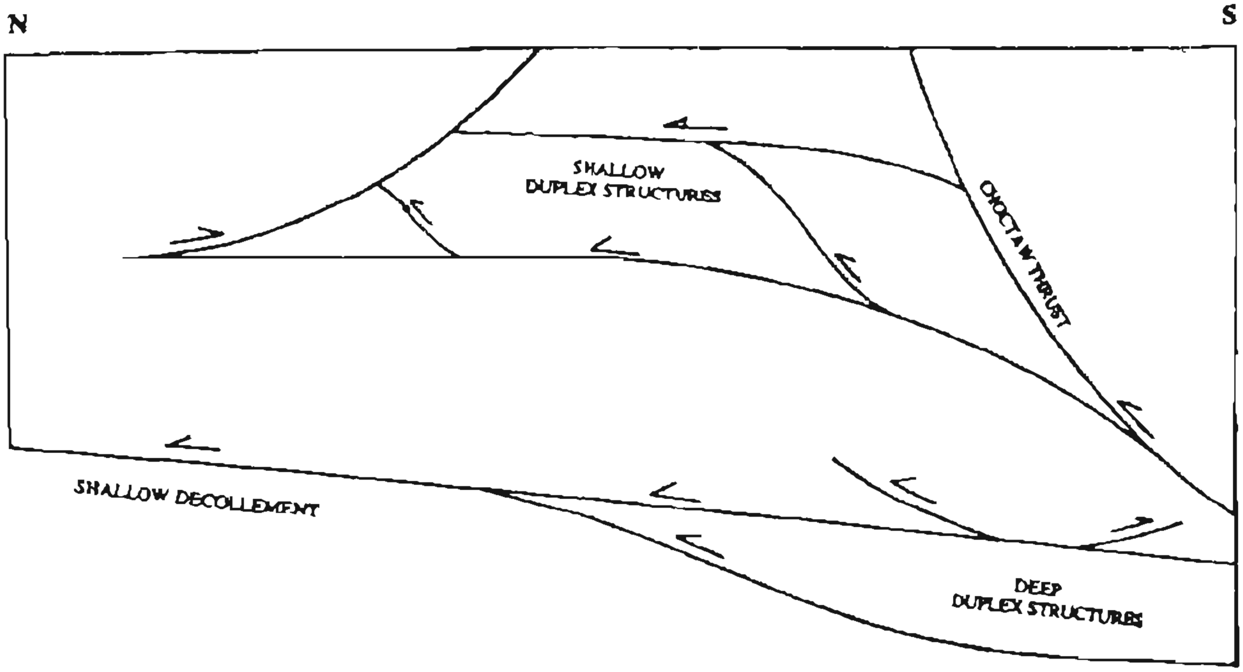


Figure 12.--Proposed geometry of northern frontal belt, showing complex imbricates and a shallow decollement (Reeves and others, 1990).

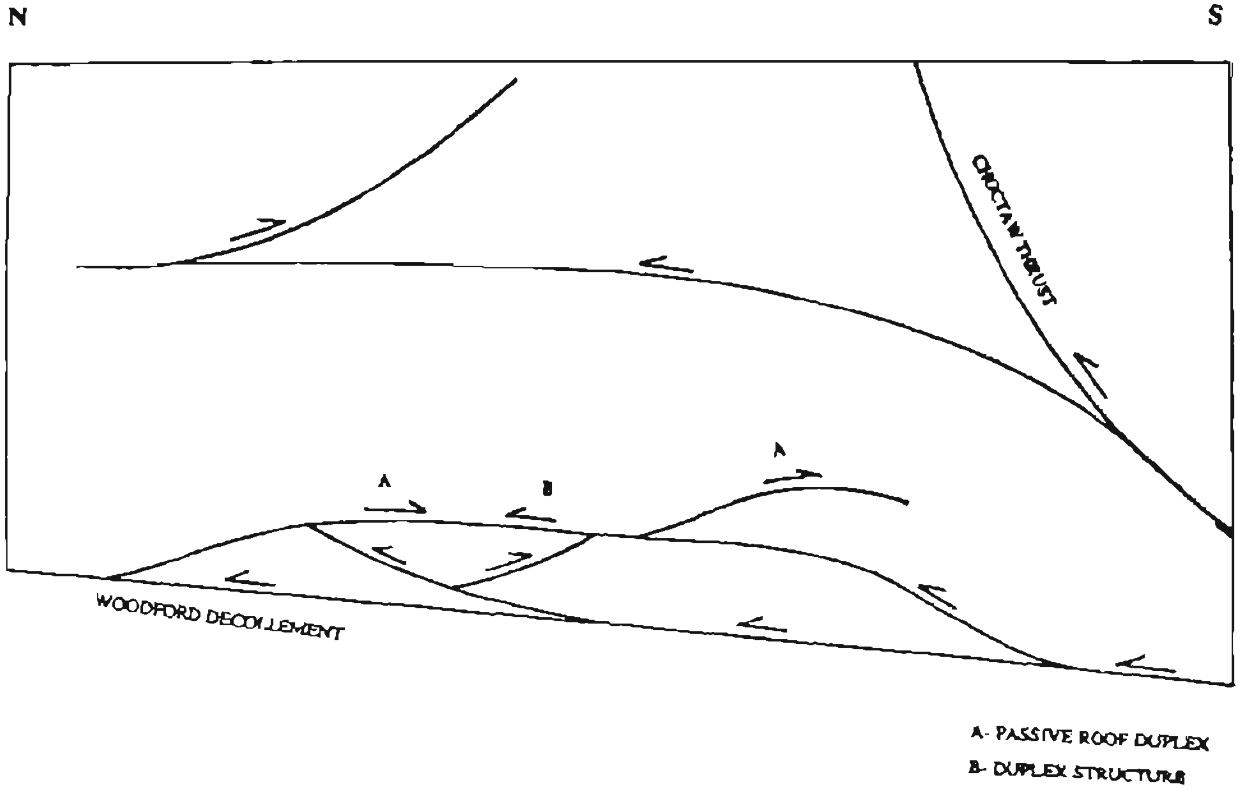


Figure 13.--Proposed geometry of Hartshorne, Oklahoma, depicting a shallow and deep triangle zone with deep decollement (Perry and Suneson, 1990).

thrust is the roof thrust of a duplex structure, a series of blind imbricates, and two detachment surfaces, Springer and Woodford that served as the root of imbricate thrusts (Figure 14).

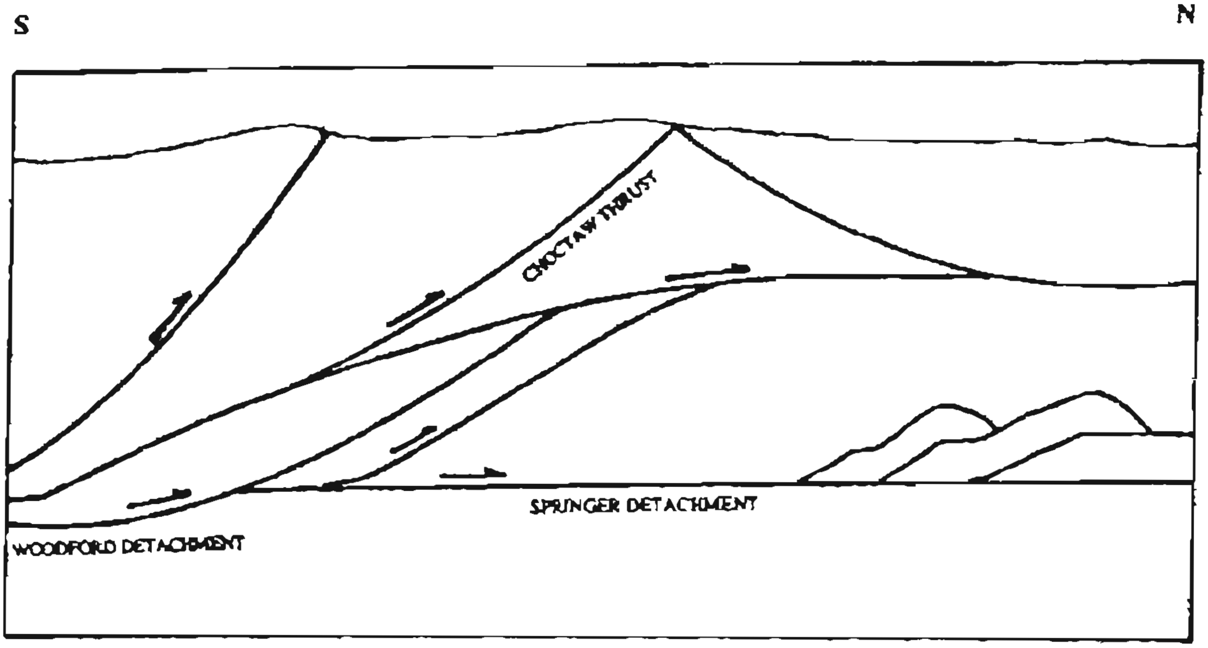


Figure 14.--Proposed geometry of west of Wilburton Field showing a shallow triangle zone and two decollements (Wilkerson and Wellman, 1993).

CHAPTER III

STRATIGRAPHIC FRAMEWORK

The Arkoma Basin contains a thick sequence of (oldest to youngest) pre-Morrowan (Cambrian to Mississippian), Pennsylvanian Morrowan, Atokan, and Desmoinesian rocks. The upper and basal Atokan sediments are the focus of this investigation, because these rocks units are well developed in the study area. As mentioned, this study deals mainly with the thrusted Spiro Sandstone. However, a brief overview of Cambrian to Mississippian rocks is necessary to understand the whole stratigraphy of the area (Figure 15).

Upper Cambrian - Lower Ordovician Systems

The Upper Cambrian Timbered Hills Group is underlain by Arbuckle Group, contain Reagan Sandstone and Honey Creek Formation. The Reagan Sandstone is a feldspathic and normally glauconitic sequence of Franconian age. The thickness of the Reagan Sandstone is ranges from 22.72-136.36 m (75-450 ft.). Above the Reagan, the Honey Creek Formation is present, consists of a thin, trilobite-rich pelmatazoan limestone which is 30.3 m (100 ft.) thick in the Arbuckle anticline (Ham, 1978).

The Arbuckle Group underlain by Timbered Hills Group, is the thickest sequence of lower Paleozoic strata in Oklahoma (Figure 15). It is Upper Cambrian to Lower Ordovician in age, and is divided into eight formations, six limestone and two dolomite are the dominated units. These formations in total are characterized by approximately

ERA	SYSTEM	SERIES	STAGE	GROUP	UNITS
PALEOZOIC	PENNSYVANIAN	Middle	Desmoinesian	Marmaton	Holdenville Sh. Wewoka Fm. Wetuma Sh. Calvin Ss.
				Cabaniss	Senora Fm. Stuart Sh. Thurman Ss.
				Krebs	Boggy Fm. Savana Fm. McAlester Fm. Hartshorne Fm.
		Lower	Atokan	"U. Dornick Hills"	Atokan Fm. "Spiro" Ss.
			Morrowan	"L. Dornick Hills"	Wapanucka Fm. Union Valley Fm. Springer Fm.
	MISSISSI	Upper	Chesterian		"Caney" Sh.
			Mermacian		
		Lower	Osagean		
			Kinderhookia		
	DEVONIAN	Upper	Chautauquan		Woodford Sh.
		Middle	Senecan	HuntonGp.	Frisco Fm. Quarry Mtn. Fm. Tenkiller Fm. Pettite Oolite
			Erian		
		Lower	Ulsterian		
	SILU-RIAN	Upper	Cayugan		
		Lower	Niagarian		
	ORDOVICIAN	Upper	Richmond	Sylvan Sh.	Sylvan Sh.
			Maysville Tenton	Viola Gp.	Wellming Fm. Viola Spring Fm.
		Middle	Black River Chazy White Rock	Simpson Gp.	Fite Fm. Tyner Fm. Burgen Fm.
		Lower	Canadian	Arbuckle Gp.	West Spr. Creek Fm. Kindblade Fm. Cool Creek Fm. Mckenzie Fm. Signal Mtn. Fm.
	CAMBRIAN	Upper	Franconian	Timbered Hills	Honey Creek Ls. Regean Ss.

Figure 15.--Stratigraphic column depicting Cambrian through Pennsylvanian rocks involved in the Wilburton thrust system (modified from Johnson, 1989).

1,000-7,000 feet. (303-2,121 m) thick of shallow marine carbonates (Johnson and Others. 1988). Ham (1959) used the term "Arbuckle Facies" for the sequence of thick carbonates and accompanying shales and sandstones, extending from the base of the Cambrian Reagan Sandstone to the Upper Devonian-Lower Mississippian Woodford Shale.

Middle-Upper Ordovician System

The Ordovician rocks include Simpson Group. Viola Limestone, and Sylvan Shale. Overlying the Arbuckle group is the middle Ordovician Simpson Group (Figure 15). which is mainly composed of alternating units of fossiliferous limestones, sandstones, shales. Above the Simpson Group the middle Ordovician Viola Limestone lies conformably (Figure 15). Its contact with the overlying Fernvale is conformably and probably represents a change in facies. The Viola nearly contains black, dense, white or light-tan crystalline limestone. Chert layers are also common, usually in the color of enclosing limestone. In general, the cherts of the Viola are tan to brown colored (Tulsa Geological Society Digest, 1961). The upper Ordovician Sylvan Shale is a dark. greenish-gray shale that is unconformable upon the Viola Limestone (Figure 15). The shale is also well-laminated and contains abundant graptolites and chitinozoans (Ham, 1978).

Silurian- Devonian Systems

The Lower Silurian through Lower Devonian rocks are represented by the Hunton Group which is predominantly limestone. The Hunton Group lies conformably both above the Sylvan Shale, and below the Woodford Shale. The thickness of the Hunton is the result of regional thinning from southwest to northeast, regional erosion during the post epeirogeny, and local differential erosion over pre-existing anticlines. Cline (1960), while studying in the Ouachita Mountains, concluded that the Hunton Limestones are completely different from the Ouachita units, although they occur in a comparable stratigraphic intervals. Ham (1959) suggested that the Missouri Mountains Shale of Silurian age is equivalent to the lower part of Hunton Group.

The Woodford Shale, Upper Devonian to Lower Mississippian in age, is lithologically and faunally similar to the upper Arkansas Novaculite. It lies conformably above Hunton Group and below the Caney Shale. It occurs throughout the Arkoma Basin and is known as a detachment surface in the Wilburton Field and surrounding areas.

Mississippian System

The Mississippian strata mostly recognized as the Caney Shale, which predominantly consists of sequences of shale that contain fauna of both Mississippian and Pennsylvanian affinites (Berry and Trumbly, 1967). The Caney Shale lies conformably above the Woodford Shale and unconformably beneath the Mississippian Springer Shale. Miser (1929) noticed the presence of thin slices of Caney Shale at the base of thrust sheets north of Ti Valley fault.

The Caney Shale is generally brown to black. Boyd (1938) noted that the top of the Caney is marked by a glauconite-rich stratum and it is unsuited for use as a structural horizon. On the wells Rutherford #1 sec.3, T.5N.,R.18E. and Ulysses #1 sec.35, T.4N.,R.18E., (see appendix) the Caney Shale is found around 15,670 ft. They show a calcareous nature and low resistivity wireline log response characteristics. This diagnostic log response was utilized in the delineation of the lithic boundary of the Caney Shale with that of overlying strata in the Arkoma Basin.

Pennsylvanian System

The lower and middle Pennsylvanian rocks in the Arkoma Basin are subdivided into Morrowan. Atokan, and Desmoinesian stages (Table 1). In the study area the upper Morrowan and lower Atokan rocks are mostly exposed on the surface and are penetrated by the wells drilled for oil or gas (see appendix). The upper Morrowan and lower Atokan rocks are differentiated on the basis of wireline log response. The upper Morrowan stage is mainly consist of calcareous to sandy shales, while the lower Atokan Spiro Sandstone is porous, clean, and highly resistive (Ulysses #1, sec.35, T.4N., R.18E.).

Morrowan Stage

The Morrowan Stage in the Arkoma Basin represents a continuation of the shelf-like sediments that are composed of shales, siltstones sandstones and shelf carbonates. This series consists of Springer Shale, and lower Dornick Hills Group; including Union Valley Formation (Cromwell sandstone is used as a subsurface term), and Wapanucka Limestone (Table 1).

The Morrowan Stage was described in northwestern Arkansas. the names Hale and Bloyd can be used only for a 25 km wide belt at the eastern margin of Oklahoma (Tulsa Geological Society Digest, 1961). The lithologic distinction is lost to the farther west, although an increase in the percentage of limestone has been observed westward and a corresponding decrease in the percentage of sandstones (Johnson & Others, 1989).

Springer Formation. The Springer Formation which was named by Goldstone (1922) for the town of Springer in Carter County, Oklahoma, is overlain by the Union Valley Formation (Cromwell Sandstone) and is underlain by Mississippian Caney Shale

TABLE 1

LOWER AND MIDDLE PENNSYLVANIAN STRATIGRAPHY

Desmoinesian Series	Marmaton Group	Holdenville Sh.
		Wewoka Fm.
		Wetumka Sh.
		Calvin Fm.
	Cabaniss Group	Senora Fm.
		Stuart Sh
		Thurman Ss.
	Krebs Group	Boggy Fm.
		Savanna Fm.
		McAlester Fm.
		Hartshorne Fm.
Atokan Series	"U. Dornick Hills Group"	Atoka Fm.
Morrowan Series	"L. Dornick Hills Group"	Wapanucka Fm.
		Union Valley Fm.
		Springer Fm.

in the Arkoma Basin (Figure 15). The Springer Formation is present in few wireline logs, those penetrating the Mississippian strata. In general, the Springer exhibits sandy/silty shale, a medium resistivity log response (Paschall #2, sec. 21 T.5N.,R.18E.) which contrast by being sandier and less limy than the underlying Caney Shale. This can be identified by the wireline log response characteristics (Kilpatrick #2, sec.16 T.5N.,R.18E.).

The sandstone of the Springer is very light to dark gray and very fine to fine-grained. It is commonly calcareous like the sandstone of the overlying Union Valley, but the smaller grain size and the absence of glauconite can be used to differentiate the sandstone units of Springer from the Union Valley Sandstone (Frezon, 1962).

Lower Dornick Hills Group

Union Valley Formation. The Union Valley Formation is the lowermost unit of the lower Dornick Hills Group; it contains two members: the upper Union Valley Limestone and underlying Union Valley Sandstone (Baker, 1951).

Cromwell Sandstone. The Cromwell Sandstone is used as the subsurface term for the Union Valley Sandstone. It extends throughout most of the Arkoma Basin. It occupies the stratigraphic interval between the Pennsylvanian Springer Shale and the Union Valley Limestone (Figure 15). The Union Valley Cromwell generally consists of a maximum of 265 feet of sandstone overlain by 5-15 feet of arenaceous limestone (Rutherford #1, see Appendix) that is broken by thin-shale patting. Locally the Cromwell Sandstone consists of multiple strata of sandstone, interbedded with dark-gray to black marine shales. The Cromwell Sandstone ranges from 0 to approximately 265 feet thick in the study area. (Rutherford #1, sec.5, T.3N.,R.18E., and Kilpatrick #2, sec.16, T.5N.,R.18E., respectively). Most of the wells do not penetrate the Cromwell Sandstone because the production is found in the stratigraphically higher unit (Spiro Sandstone). Therefore, the trend of thickening or thinning of the Cromwell Sandstone could not be determined in the study area. The Cromwell Sandstone is gray to brown, fine-medium grained, sub-angular to sub-rounded, calcareous, glauconitic, fossiliferous, and commonly bioturbated sandstone (Cline, 1968).

Union Valley Limestone. The Union Valley Limestone lies above the Union Valley Cromwell Sandstone and beneath the Wapanucka Limestone. Locally, the Union Valley Limestone is absent and the Cromwell Sandstone is separated by thin shales (Trust 1-5, sec.5, T.4N.,R.18E.), or it shows a gradational contact with the Cromwell. The sample and insoluble residue studies conducted by Jackson (1949) and Baker (1951) indicate an upward increasing carbonate content in the Cromwell, suggesting that a gradational contact exists with the limestone member. The Union Valley Limestone is gray to brown, fine to coarsely crystalline, fossiliferous, siliceous, and glauconitic limestone.

Wapanucka Formation. The upper Morrowan Wapanucka Formation described first by Taff (1901), is the upper formation of the lower Dornick Hills Group. The Wapanucka Formation conformably overlies the Union Valley Formation and is overlain by the Atokan Series. The Wapanucka Formation contains an unnamed lower shale and an upper limestone member. It is exposed repeatedly in a series of thrust-fault blocks in the frontal Ouachita Mountains of southeastern Oklahoma. In the frontal Ouachitas, Grayson (1979) subdivided the Wapanucka into three informal lithologic members: these are, in ascending order: lower limestone, middle shale, and upper sandstone-limestone. They show a continuous sedimentation and shallowing-upward sequences punctuated by reversals. The lower shale is commonly known to as "Wapanucka shale". The shale is gray to black and contains thin beds of limestone and sandstone. The upper limestone which holds the formation name is brown to gray and commonly sandy and/or oolitic. The Wapanucka in the Wilburton area is well exposed in natural outcrops and quarries, southwest of Latimar County, Oklahoma. It is dipping approximately 75° SE (Figure 16). The Wapanucka is proved to be an excellent seismic reflector, and a marker horizon on electric logs (Figure 2). The thickness of the Wapanucka Limestone ranges from approximately 100 feet in the northwest (Federal church #2, sec.36, T.6N.,R.18E.), to 282 feet in the southwest (Rutherford #1, sec.5, T.3N.,R.18E.).

Upper Dornick Hills Group

Atokan Stage

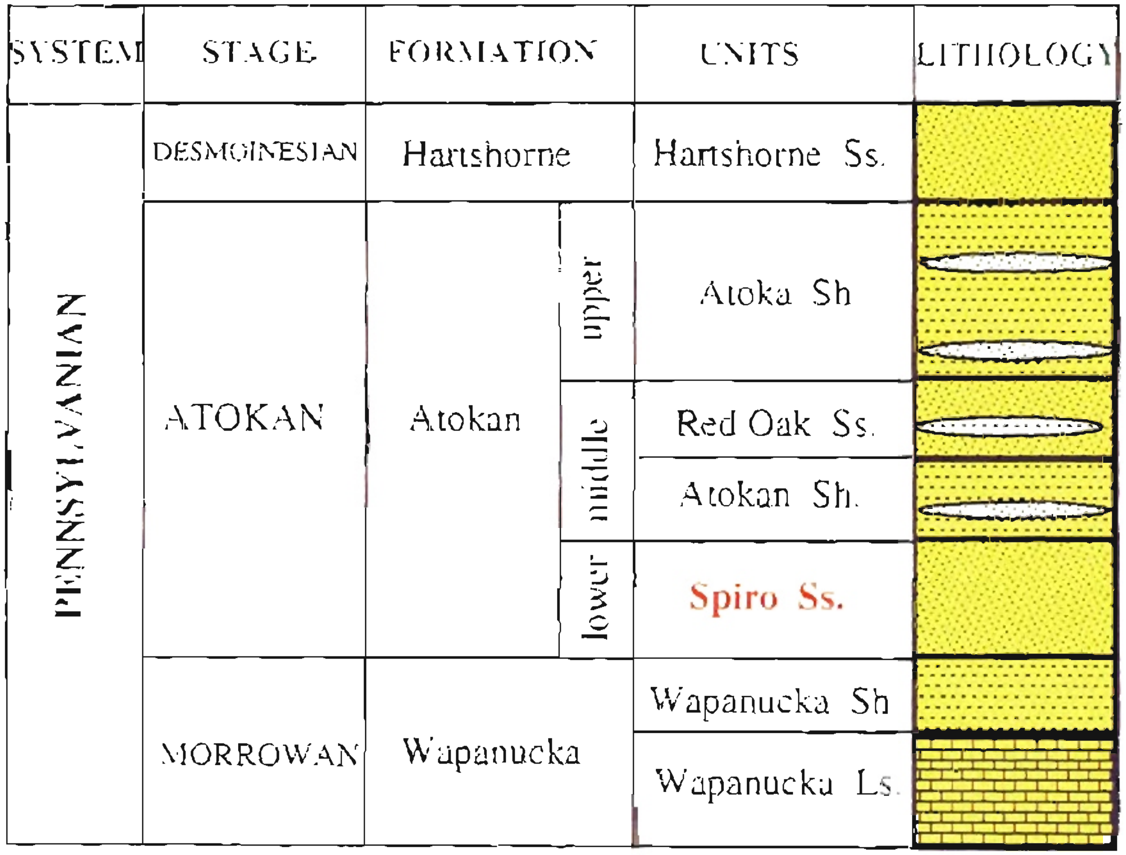
The Atokan Stage is mainly represented by the Atokan Formation which thickens southeastward in the Arkoma Basin. It consists of upper Dornick Hills Group which is made out of Spiro Sandstone of Atokan Formation. Table 2 illustrates the detailed stratigraphy of the Atokan rocks. The Atokan rocks are composed of alternating sandstones and shales deposited in non marine, shallow marine, and deep marine environment (Houseknecht and Kacena, 1983). A regional unconformity exits between the basal Atokan sandstone and the underlying Morrowan Wapanucka Limestone (Johnson and Others, 1989).

Basal Atokan Spiro Sandstone. The Spiro Sandstone is the basal unit of the Atokan Formation. It is the principal gas producing zone at the Wilburton gas field and surrounding areas and is underlain by the Morrowan Wapanucka Formation (Figure 2).

Figure 16.--Wapanucka Limestone showing amount of dip, estimated as 75° in the study area

TABLE-2

Generalized Atokan Stratigraphic Column



SYSTEM	STAGE	FORMATION		UNITS	LITHOLOGY
PENNSYLVANIAN	DESMOINESIAN	Hartshorne		Hartshorne Ss.	
	ATOKAN	Atokan	upper	Atoka Sh	
			middle	Red Oak Ss.	
				Atokan Sh.	
			lower	**Spiro Ss.**	
	MORROWAN	Wapanucka		Wapanucka Sh	
				Wapanucka Ls.	

Branan (1966) suggested that the Spiro was deposited prior to the significant subsidence of the Arkoma Basin and subsequent thickening of the Atoka section. The trend of the Spiro represents the deposition in the broad delta complex with distributory channels, tidal channels, and shallow-water, interfingering marine sandstone bars and carbonates (Houseknecht, 1983). Consequently, the lithology and the distribution pattern of the Spiro Sandstone in the Wilburton and surrounding areas suggest that the Spiro is a shallow marine deposit with the sand having been derived from the northeast (Cline, 1968). There seems to have continuous deposition from Morrowan through Atokan time as the rocks appear to be conformable in the Wilburton field (Cline, 1968).
The Spiro Sandstone is gray-black or tan, fine to medium grained, sub-rounded to sub-angular, quartzitic, gilsonitic, and calcareous. It becomes more calcareous along northern side of the town of Wilburton Berry and Trumbly, 1967). The surface data indicate the steeper dips of Spiro on the overthrust block of the Choctaw fault, approximately 75° SE (Figure 17). The Spiro Sandstone ranges from approximately 35 to 120 feet thick (Wildlife Cons.#1, sec.7, T.6N., R.19E., and Rutherford #1, sec.5, T.3N., R.18E. respectively) with an average thickness of 45 feet in the study area.

Middle - Lower Atokan Formation. The Atokan Formation was named by Taff and Adams (1900) for the town of Atokan in Atoka County, Oklahoma. The formation , which is the only unit of Atoka Group was initially described as a sequence of alternating sandstone and shale beds, 7,000 ft. thick, overlain by Hartshorne Sandstone and underlain by the Wapanucka Limestone. It is found almost in all the wireline logs in the Study area (see appendix I). The Atoka Formation is exposed at the surface and is present in the subsurface throughout the Arkoma Basin in Oklahoma and Arkansas. It is a complex terrigenous deposit that was accumulated prior to and during the development of a foreland basin associated with the Ouachita fold belt. The Atoka Formation within the

Figure 17.--Spiro Sandstone showing the amount of dip in the study area, located: sec 5, T.3N., R.18E.

basin is divided into lower, middle and upper intervals.

The lower Atoka is characterized by a widespread basal sand, the Spiro Sandstone and the overlying shale. The upper boundary of the lower Atoka interval occurs within a shale succession and is poorly defined. The middle Atoka interval is dominantly composed of shale with some major sandstone units. It was deposited during the development of syndepositional faults and is characterized by marked increase in thickness southward side of these faults. However, the Atoka Formation is predominantly consists of dark shales, usually gray-black or dark brown. Intercalated throughout the shales are micaceous siltstones and very fine-grained sandstones. The well log data show thick sequence of Atoka Formation in the T.4N. (southward) (Rutherford #1-24, sec.24, T.4N., R.17E.) than in the T.6N. (northward) (Quid B #1, sec.32, T.6N., R.19E.) along all the cross-sections. In addition, the well logs (see Appendix) used to construct the balanced cross-sections show the eastward increase in sandstone in the Atoka Formation. This increase suggests that the source of the sand was to the east or southeast of the Arkoma Basin.

Red Oak Sandstone. The term Red Oak Sandstone has been applied to a sandstone in the middle Atokan. The Red Oak Sandstone is a major sand unit in the middle Atoka interval: it is a significant producer of natural gas. It lies approximately 5,000 ft. above the base of the Atoka. It is interbedded with shale and exhibits moderate permeability and porosity and medium resistivity wireline log response. This sand unit is sub-angular, fine-grained, well-sorted, and quartzose. Individual sandstone beds are relatively thin and contain sole marks, load casts, horizontal bedding, small-scale cross-bedding, and ripple marks (Bowsher and Johnson, 1968). The sandstone is confined to the south side of the San Boise syncline. Vedros and Visher (1979) suggested from an analysis of sedimentary structures and sand unit geometry, that the Red Oak accumulated in a submarine fan environment supplied with sediments from submarine canyons cut into the scarp of an active growth fault to the north. The available data were not sufficient to determine the geographic limits of the Red Oak Sandstone.

Desmoinesian Stage.

The Desmoinesian Stage is divided into Krebs, Cabaniss, and Marmaton Groups. Only the Hartshorne, McAlester, Savana, and Boggy Formations of the Krebs Group are exposed on the surface in the study area. They are penetrated by the wells in the subsurface. Desmoinesian rocks in the Arkoma Basin are dark shales, siltstones, sandstones, thin coals, and sandy limestones in lenses (Branson, 1956).

The Hartshorne lies above the Atoka shale, apparently unconformable. It is a cyclothem sandstone consisting of micaceous, carbonaceous, sideritic siltstones and very fine-grained sandstones (Tulsa Geological Society Digest, 1961). The rest of the Desmoinesian rocks in the Arkoma Basin are dark shales, siltstones, thin coals and sandy limestones (Branson, 1956). These rock units were deposited in alluvial and shallow marine environments. This group show marked thickening southward into the subsiding foreland basin (Houseknecht, and Kacena, 1983).

CHAPTER IV

PETROLOGY, DIAGENESIS AND DEPOSITIONAL ENVIRONMENT OF THE SPIRO SANDSTONE

The petrologic and diagenetic studies have been done to determine the distribution, depositional environment, and diagenetic overprints of the Spiro Sandstone. Both field and lab investigations were needed to complete this work. The samples were collected from two different outcrops: the Gulf Course outcrop and near the Exxon Rutherford #1.sec.5. T.3N., R.18E. (well). Overall fourteen thin-sections from those samples were examined in this study in order to discuss the diagenesis and infer a depositional environment of the Spiro Sandstone.

Petrology and Diagenesis

Detrital Constituents

The detrital grains include quartz, little amount of feldspar, metamorphic rock fragments, glauconite and intraformational clasts of clays and siltstones. Some accessory minerals are also present such as, zircon, tourmaline, muscovite, and pyrite (Figure 18). Muscovite and pyrite occur as quantity. The major detrital constituent is quartz (75-95%). Most of the quartz grains are monocrystalline (90%). The grains exhibit straight to slightly undulose extinction (Figure 19). The large variation in grain size can be an

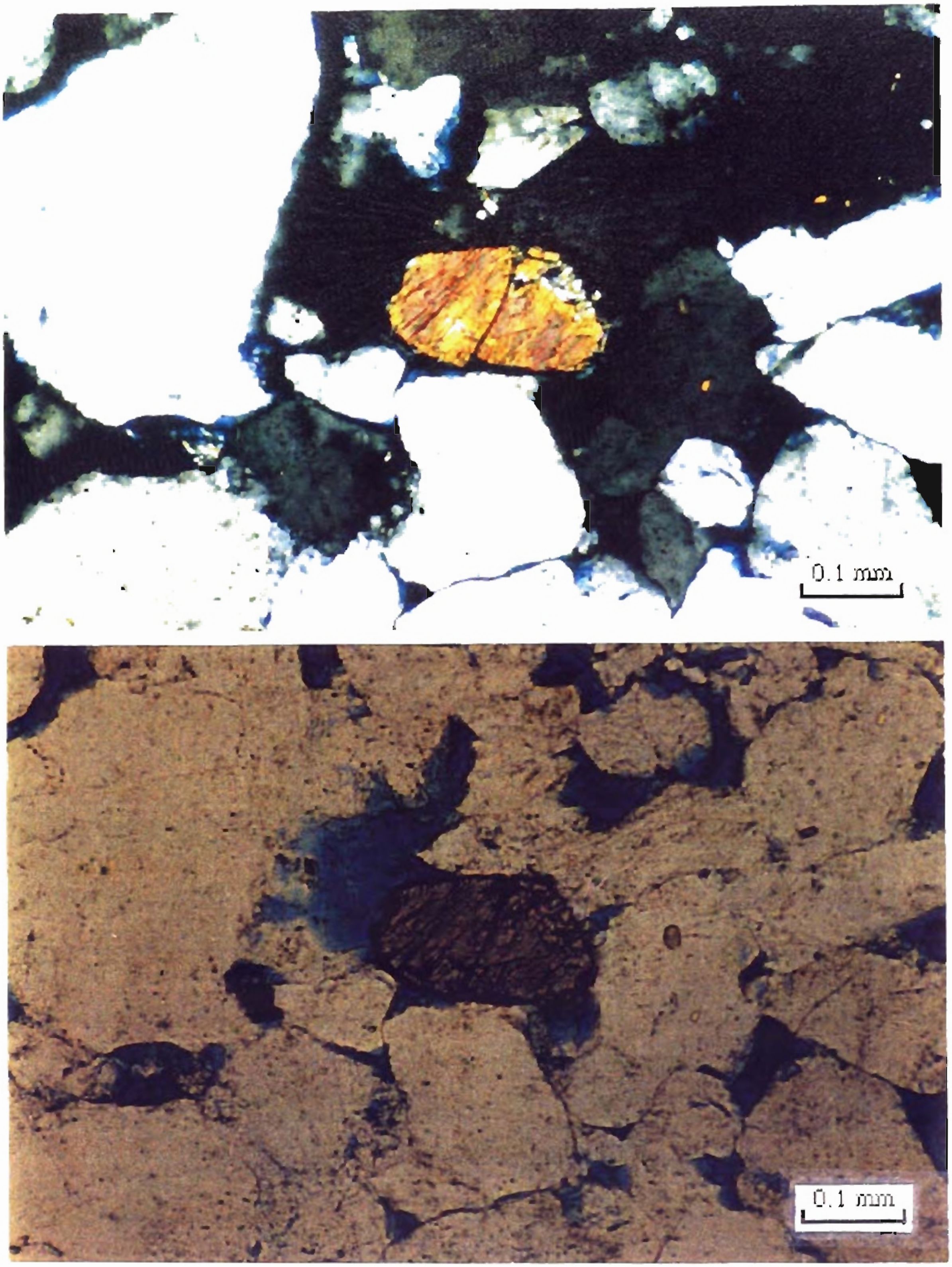


Figure 18.--Detrital tourmaline grain (accessory mineral) probably broken during the transportation (above, crossed) with some porosity and opaque grains in same section (below, parallel).

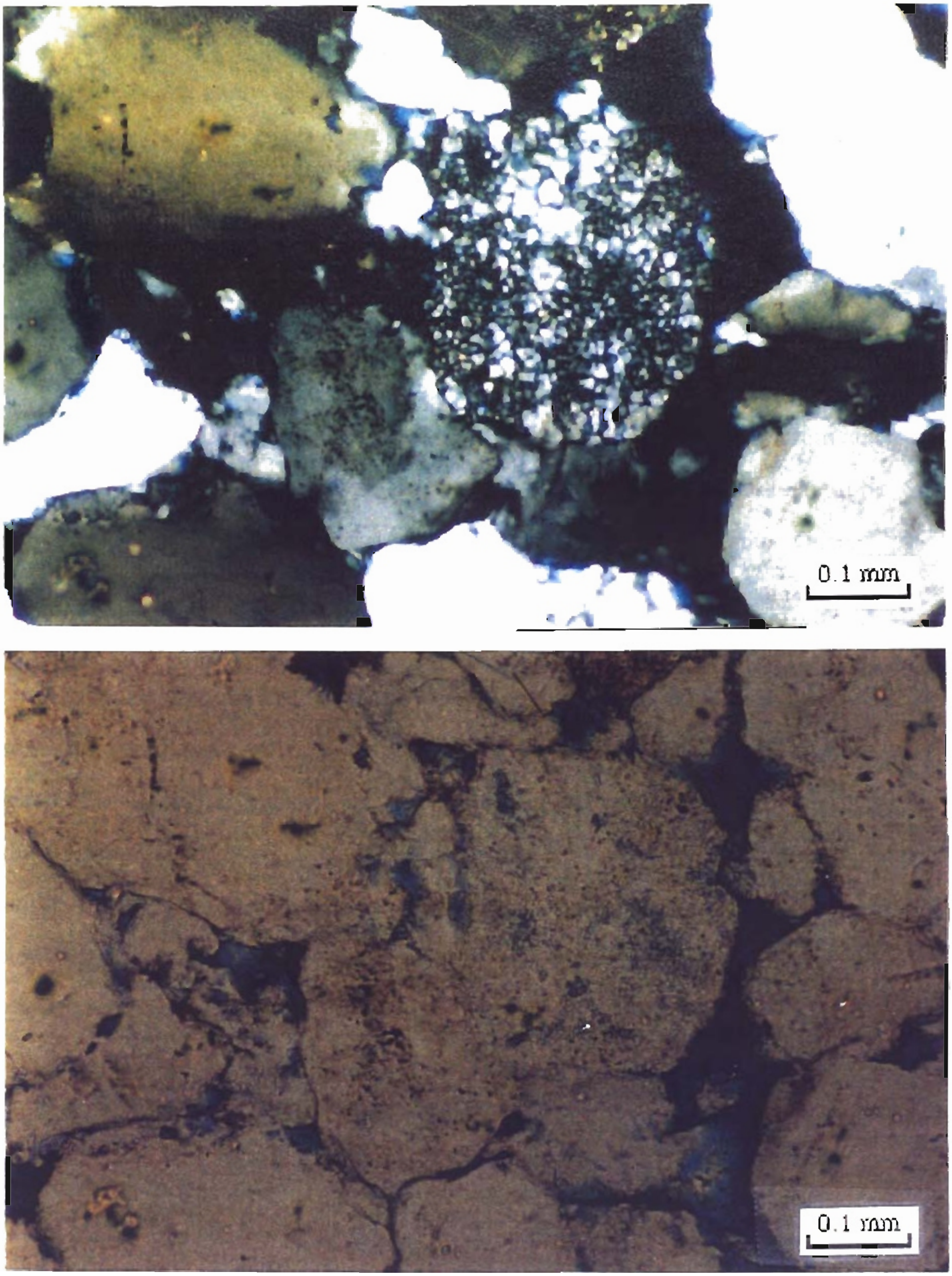


Figure 19.--Undulose extinction of quartz grain with a very visible chert rain in the middle, sampled near the Exxon Rutherford #1, located: sec. 5, T.3N., R. 18E.

evidence for fracturing during the compaction or/and tectonic activity (Figure 20). On the whole, the quartz grains are moderately well-sorted.

Detrital matrix consists of silt size metastable minerals and rock fragments which show a alteration to hematite or clays (illite and/or kaolinite).

Diagenetic Constituents

Kaolinite and Illite are the principal clays of the examined sandstone from outcrop (Figure 21). The illite could have been produced by the alteration of either feldspar or glauconite. In the presence of acid water conditions, illite alters to kaolinite. Quartz overgrowth is the primary cement. Dolomite is a diagenetic alteration of calcite. Phosphate and glauconite exist within the rock but only few grains were found.

Generally, the chlorite (chamosite facies) is almost absent in thin-sections. The presence of chamosite facies in the study area is extremely important since it is an indicator of hydrocarbon bearing rock. The footwall block of the Choctaw thrust is the producing zone in the study area and chamosite facies is the dominant diagenetic constituent in the Spiro Sandstone (Al-Shaieb, 1989). In fact, the hanging wall block and the southernmost portion of the footwall block do not have chamosite facies. Therefore, they are non-producing zones of the area.

Porosity. The Spiro Sandstone contains both primary and secondary porosity. The majority of the porosity found in the surface outcrop of Spiro Sandstone is enlarged (oversized) pore space classified as the secondary porosity (Figure 22). The secondary porosity is related to the weathering processes. The highest porosity occurs where the primary porosity is preserved in coarse-grained sandstone. The secondary porosity is mainly recognized by presence of oversized pore space, partial dissolution, and fractured grains.

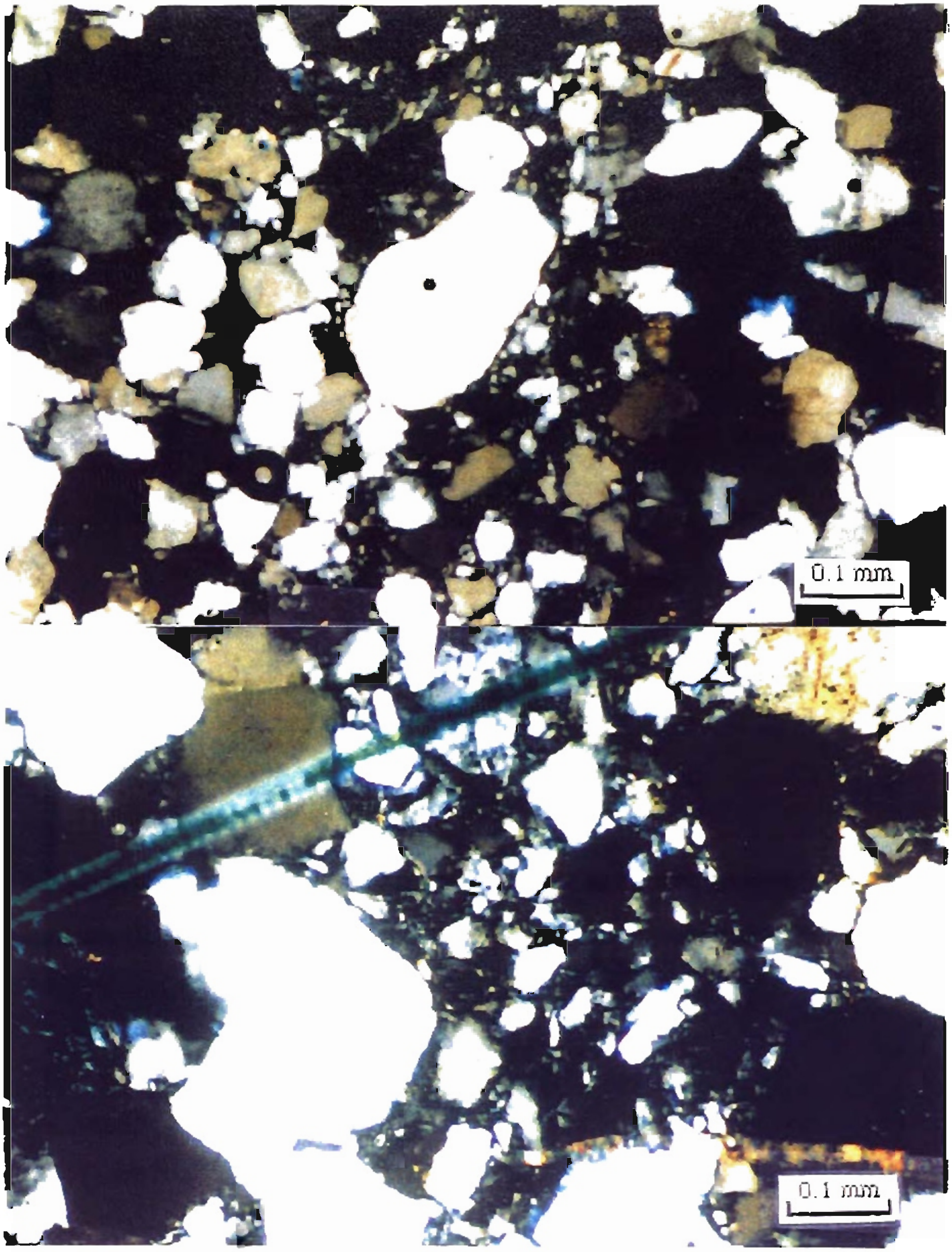


Figure 20.--Variation in the grain size due to the compaction indicate the presence of the tectonic activity in the area.

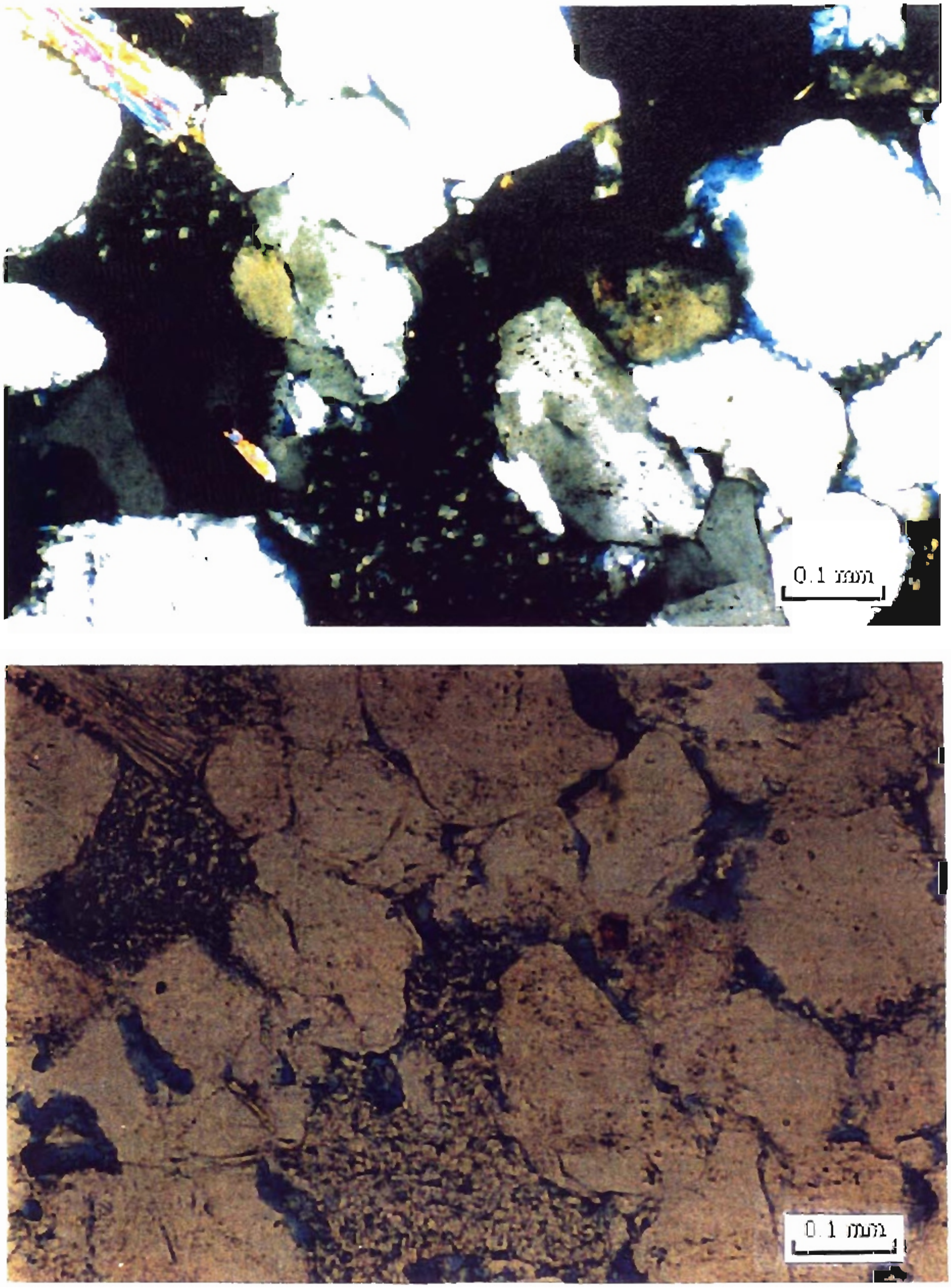


Figure 21.--Macro quartz grains with some mica flakes in the right corner (above, crossed). Kaolinite pore fillings with intergranular porosity (below, parallel).

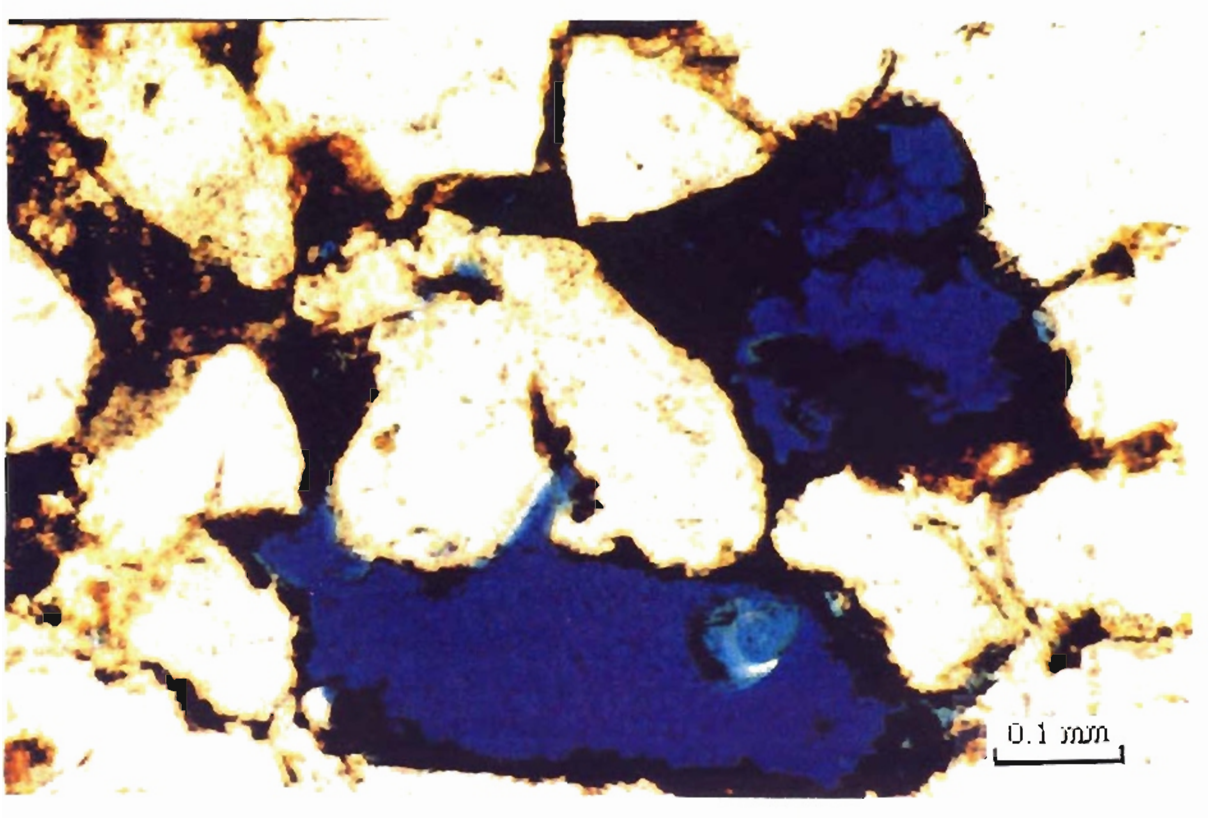


Figure 22.--Oversized pore space (secondary porosity) and the hematite and pyrite associated with mega quartz is present

Depositional Environment

The Spiro Sandstone in the study area is interpreted as having been deposited in shallow-marine environment. This interpretation was derived from outcrop studies of sedimentary structures, and examination of thin-sections.

Specific authigenic constituents, notably kaolinite and glauconite found in the thin-sections, are indicative of shallow marine deposits. The Spiro Sandstone unit was, therefore, deposited in a fairly shallow -water environment, prior to the major basin down warping and subsequent thickening of the Atoka section. The major down warping of the Arkoma Basin occurred in middle-early Atokan time after the deposition of the Spiro unit (Sutherland, 1988).

Many authors came up with the same results. Nevertheless, some authors related the deposition of the Spiro Sandstone to the fluvial deposition coupled with the shallow marine environments.

Medium to coarse-grained non-fossiliferous sandstone in the Spiro are also considered as marine deposits. Sharp lateral and lower boundaries of major sandstone trends and the associated coarser grain size are featured. That could suggest a fluvial origin. Lumsden et al. (1971) referred to these channelized deposits as the "Foster sand" and suggested they were deposited by stream channels. Houseknecht (1987) also discussed the presence of a fluvial dominated facies in the Spiro. However, because chamosite pellets (present in the subsurface) in sandstone are developed only in shallow marine environments (Porrenga, 1967) and are unstable as allochthonous elements, it is thought that these thick channelized sandstones were also deposited in a shallow marine environment.

In contrast to most marine-bar deposits, Spiro bar facies are commonly associated with carbonate facies. Some of the carbonates also represent marine bars, whereas other carbonates record interbar environments. Spiro limestone lithologies include spicule, biomicrite, crionoidal biosparite, and rarely oosparite (Grayson and Hinde, 1993).

Sutherland (1988) considered that the Arkoma Basin was depositionally part of a stable shelf along a passive continental margin during much of its history (Cambrian to Early Pennsylvanian time). Further, he suggested that the Atokan Spiro Sandstone in Oklahoma possibly came in part from Nemaha ridge in Kansas, which was higher topographically than the ridge in the Oklahoma.

Buchanan and Johnson (1968) studied the basal Atokan in Oklahoma. It includes the Spiro Sandstone and underlying shale (Sub-Spiro shale). They determined that sedimentation in this interval was initiated by the development of fluvial systems and small deltas on the eroded surface of the underlying Wapanucka Limestone with a source from the north west (Figure 23).

Lumsden and others (1971) confirmed the Buchanan and Johnson's (1968) studies and added that this all was followed by a rapid northward transgression of a coastal sand, named Spiro, that form a complex sand unit. In addition, they found that the Spiro Sandstone is typically composed of moderately to very well sorted quartz grains which is made up over 95% of the detrital grains, mostly quartz arenite.

The observable local unconformity at the base of thick Spiro Sandstone may have formed during a second period of emergence (Houseknecht, 1986). However, the transitional sections between the Spiro and Sub-Spiro shale suggest a prominent unconformity, as well as subsequent channels infilling, may have formed entirely in a marine environment (Hooker, 1988).

A rapid northward transgression of coastal sand, basically named Spiro forms a blanket-like sand unit in some areas proceeded channeling, and this geometry best developed after channel infilling (Sutherland, 1988).

The Atoka formation embodies three members: the lower, middle, and upper Atoka distinguishable on the base of the depositional history. During the Atokan time, as far as the structural history is concerned, the Lower Atoka was deposited prior to the

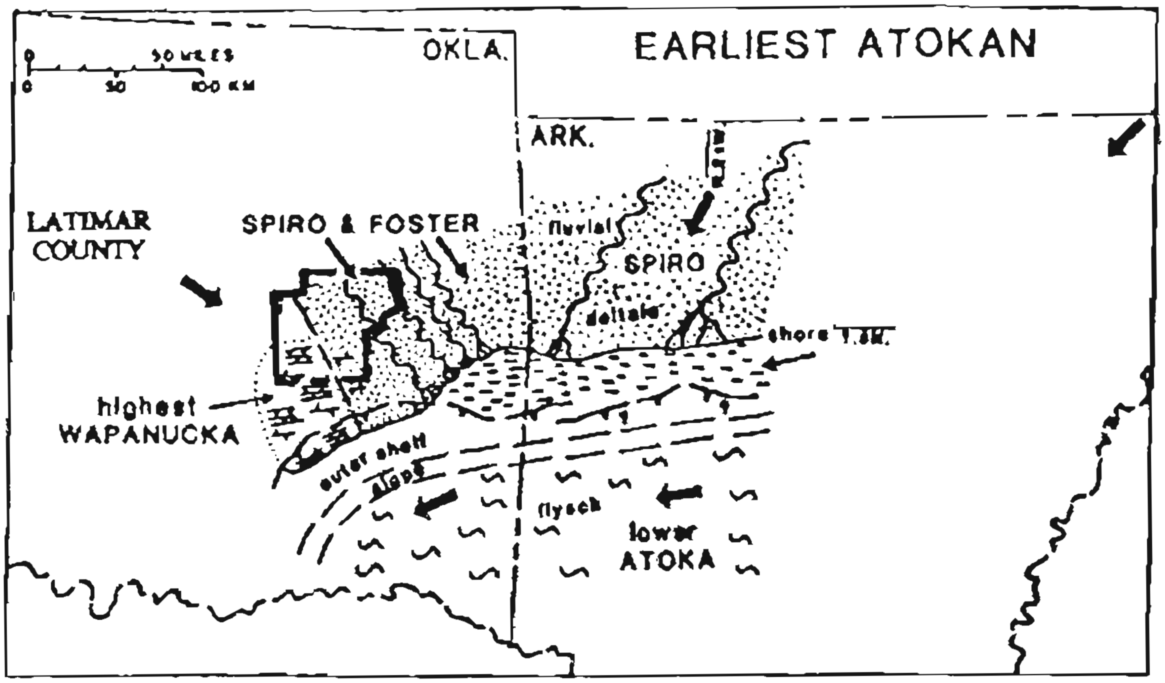


Figure 23.--Depositional model showing Atokan paleo-sedimentation with a source from north west (modified from Sutherland. 1988).

thrusting. The Middle Atoka was deposited during the development of thrusting and was characterized by increase in thickness on the downthrown side of the faults. The Upper Atoka is undisturbed by the faulting. The compressional conditions which resulted in the formation of the Ouachita Mountains and the Arkoma Basin ceased by the Desmoinesian time.

CHAPTER V

STRUCTURAL GEOLOGY

The Arkoma Basin is a foreland Basin situated in Southern Oklahoma and west-central Arkansas. It formed during the Ouachita Orogeny. The Ouachita Orogenic belt extends from east-central Mississippi through Arkansas, Oklahoma, Texas, and into northern Mexico. The Ouachita tectonic province is usually divided into three belts. From north to south these are: the Frontal belt, consisting of closely spaced thrust faults and relatively tight folds; the Central belt, consisting of widely spaced thrust faults and mostly broad, open synclines; and the Broken Bow Uplifts, consisting of Ordovician through lower Mississippian strata deformed into isoclinal recumbent folds. The Windingstair fault forms the boundary between the Frontal and Central belts (Wickham, 1978).

The Arkoma Basin is located within the Ouachita Frontal belt. It is bordered by the Ozark Uplift to the north, the Central Oklahoma Platform to the northwest, the Arbuckle Uplift to the southwest, and by the Choctaw fault to the south (Figure 24). As it is pointed by Arbenz (1984) that the Arkoma Basin is a mixed assemblage of structural styles, characterized by extensional and compressional features. Extensional features consist primarily of basement-involved down-to-the-south syndepositional faults, and are restricted to the northern part of the basin. In the frontal belt of Ouachitas, the extensional features are overridden by the compressional structural features which are detached from the underlying block faulted Atokan and older rocks. The compressional anticlines with associated thrust faults along their axes commonly are tightly folded and asymmetric southward toward the Ouachita Orogenic belt.

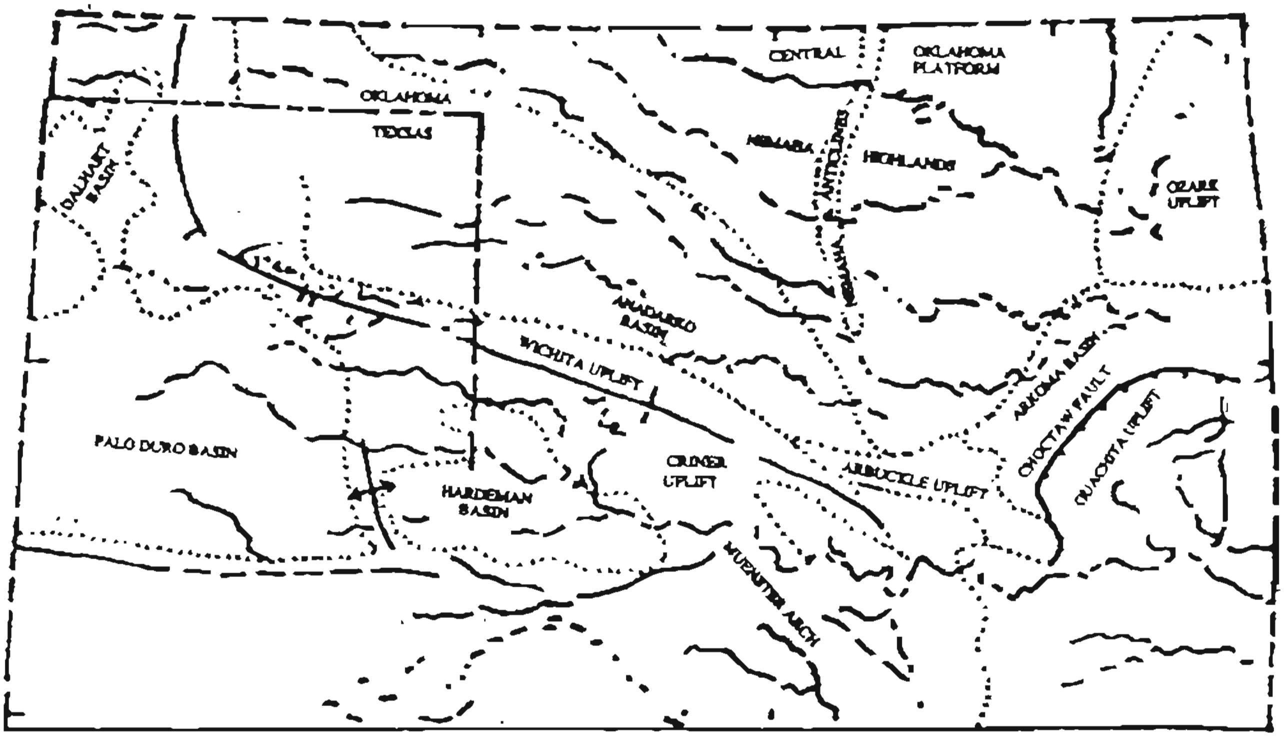


Figure 24.--Regional Geologic setting of Arkoma basin and Ouachita foldbelt (from Frezon and Dixon 1975, p.179).

The surface and subsurface data were used to construct the structural cross-sections in the Wilburton gas field and surrounding area.. The study area is within the frontal belt of the Ouachita Mountains. Although the surface geology (plate I) is provided by Niel H. Suneson and Charles A. Ferguson (1989), the structural features were examined on the surface during several field trips. The cross-sections constructed depict the subsurface geometry of the folds and faults in the area based on available well-log data, and the interpretation of several seismic profiles donated by EXXON Corporation.

The structural setting of the study area contains compressional and extensional structural features (Figure 25, plate II). Extensional features are dominantly down-to-the-south normal faults which usually trend east to east-northeast and are subparallel to the overall basin trend (plate III). Extensional features were probably forming the growth faults during the subsidence of the basin. Koinm and Dickey (1967) and McQuillan (1977) while studying in the Arkoma Basin. found an abrupt increase in the thickness of middle and lower Atoka along the normal faults, accompanied by the turbidite facies to the south. Therefore, they concluded that the faults were growth faults and were active during the Atokan sedimentation. The well-log and seismic data of the area do not show any indication of growth faulting during the upper Atokan deposition. Hartshorne Sandstone had been found unfaulted in many of the wireline well-logs (USA J.W.Mc. Tierman #1, sec.6, T.5N., R.19E.).

The anticlines present in the Wilburton gas field area are asymmetrical to the north, and where outcrop quality permits, show thrust-imbricated cores (plates II-IX). This suggests that they are probably thrust faulted at depth. These thrust faults are foothill-like imbricates of the Choctaw fault to the south.

The structural configuration of the study area was evaluated on the basis of eight balanced cross-sections and their restorations, and the interpretation of several seismic profiles. A simplified geological map (Figure 26) of the study area shows the lines of

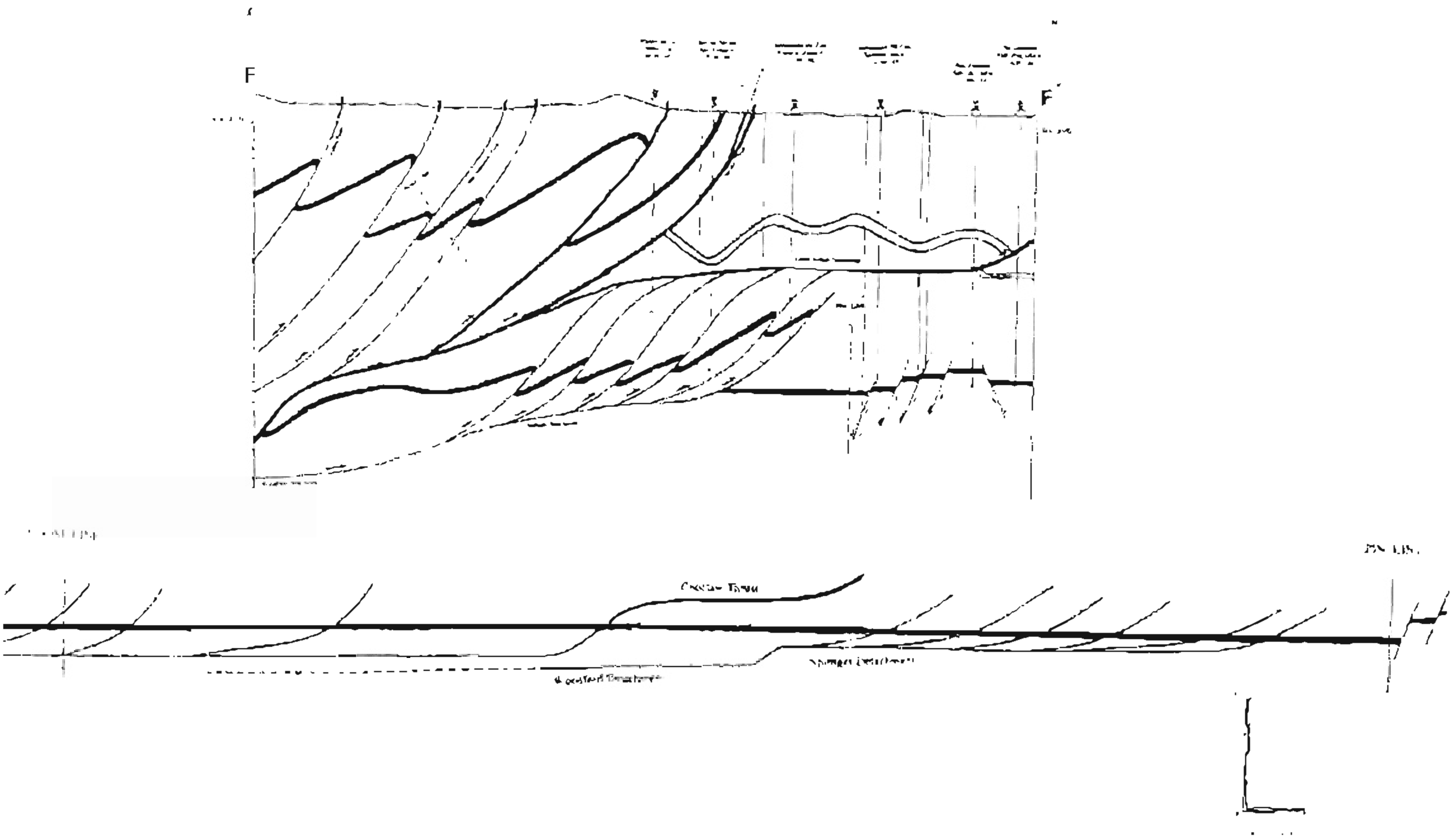
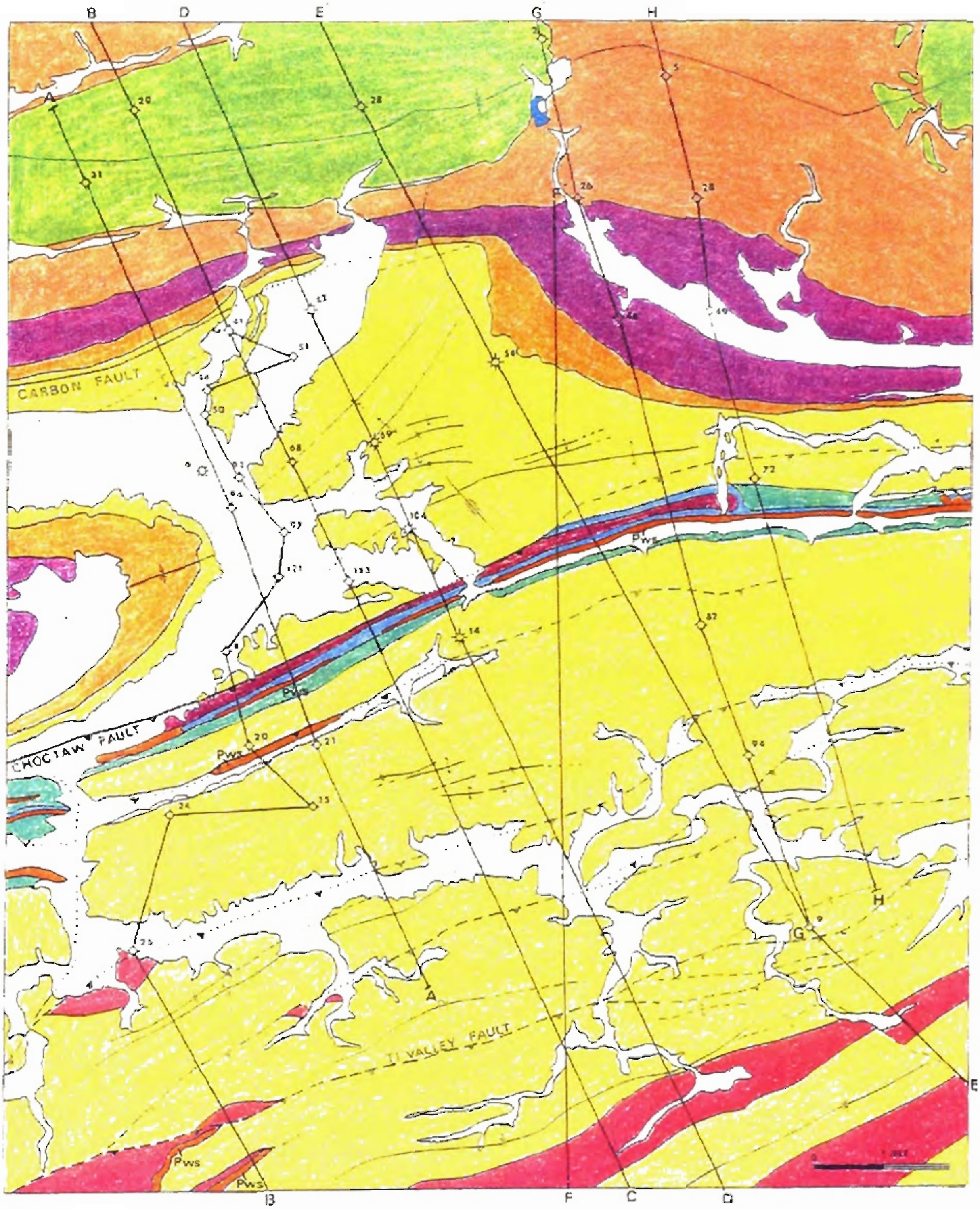


Figure 25. --Regional balanced and restored cross-sections through the Wilburton Field, illustrating the compressional and extensional features.



SIMPLIFIED GEOLOGICAL MAP OF THE WILBURTON GAS FIELD AREA
(From Suneson and Ferguson 1989; Hemish 1992, Suneson and Ferguson 1990)

Figure 26 -Simplified Geologic map showing location of cross sections in the study area (Plates II-IX).

Qa,Qt Quaternary

🟩 Pb Boggy Formation (Pennsylvanian)

🟧 Psv Savana Formation (Pennsylvanian)

🟪 Pm Mc.Alester Formation (Pennsylvanian)

🟧 Ph Hartshorne Formation (Pennsylvanian)

🟨 Pa Atoka Formation (Pennsylvanian)

🟦 Pal Lower Atoka Shale (Pennsylvanian)

🟦 Pw Wapanucka Formation (Pennsylvanian)

🟧 Pws Spiro Sandstone member (Informal) of Wapanucka Formation (Pennsylvanian)

🟥 Pjv Johns Valley Shale of Wapanucka Formation (Pennsylvanian)

🟥 Psp Springer Formation (Pennsylvanian) cross-sections. Figure 27 shows an interpreted seismic profile which is along the cross-section F-F'. The cross-sections show the changes in the geometry of structural features from surface to the subsurface and indicate that the study area comprises two main blocks named as upper block (hanging wall block of the Choctaw fault) and lower block (footwall block of the Choctaw fault). They also depict the changes taking place from west to east within the study area.

FAULTS

There are many thrust faults exposed on the surface in the study area. and some other thrusts that are delineated in the cross-sections based on subsurface data. In this thesis only the major faults will be discussed.

Ti Valley Fault

The Ti Valley fault was first named by Powers (1928). It is located in T.3N.-T.4N., R.17E., through 19E. in the study area and is generally believed to be the one of the largest thrust faults in the Ouachita Mountains frontal belt (Figure 28) in southern Oklahoma (Suneson. 1988). The significance of the fault as a boundary between different geologic provinces was first recognized by Ulrich (1927). He suggested that a great overthrust fault in the valley, north of the Windingstair Mountains. formed the northern edge of the Ouachita depositional basin represented by John's Valley through Jackfork strata. The fault possibly extends far to the southwest and east in the subsurface. The Atokan Formation is the only unit which is common on the both sides of the Ti Valley fault.

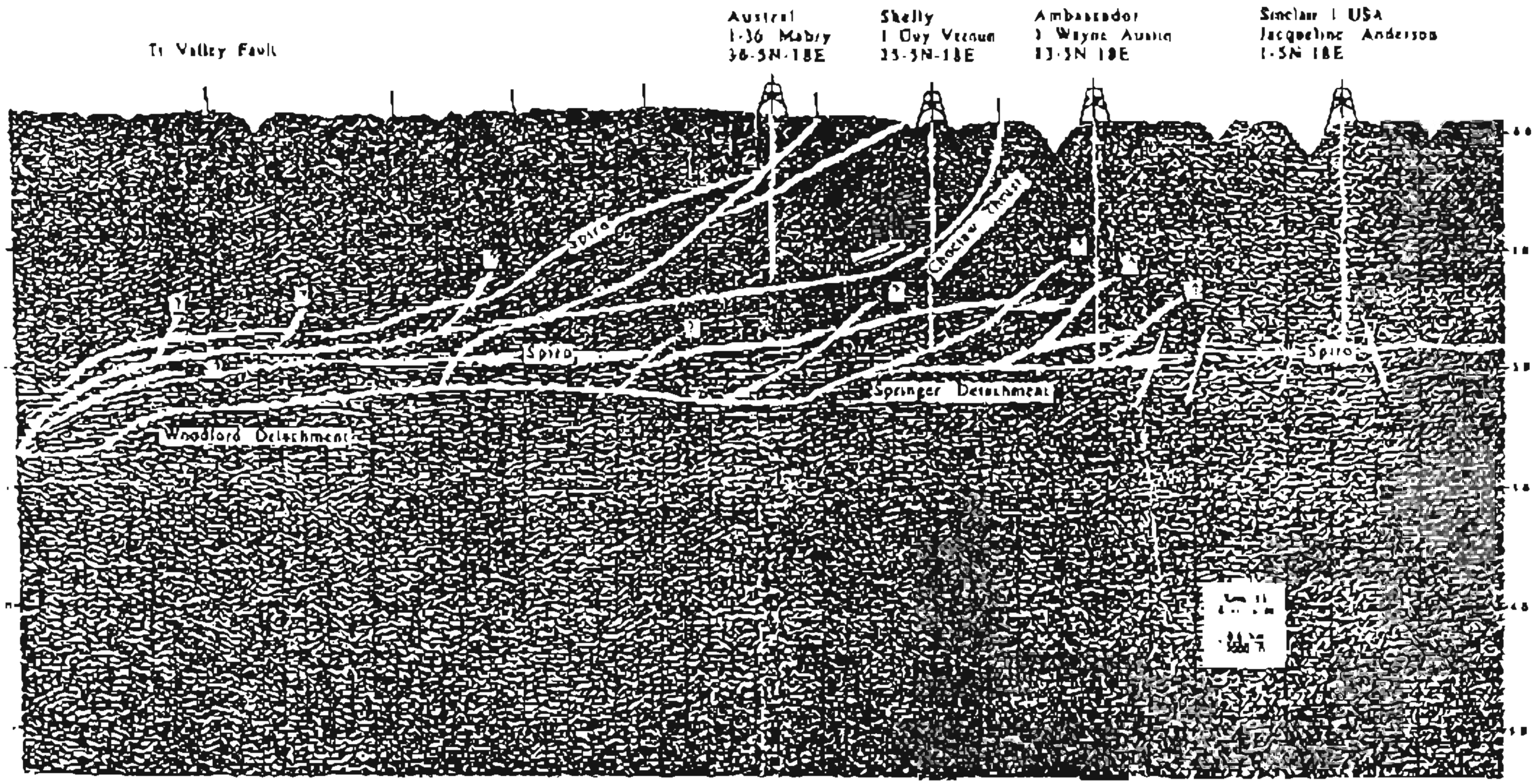


Figure 27. --Expression of interpreted seismic profile showing the geometry of the study area.

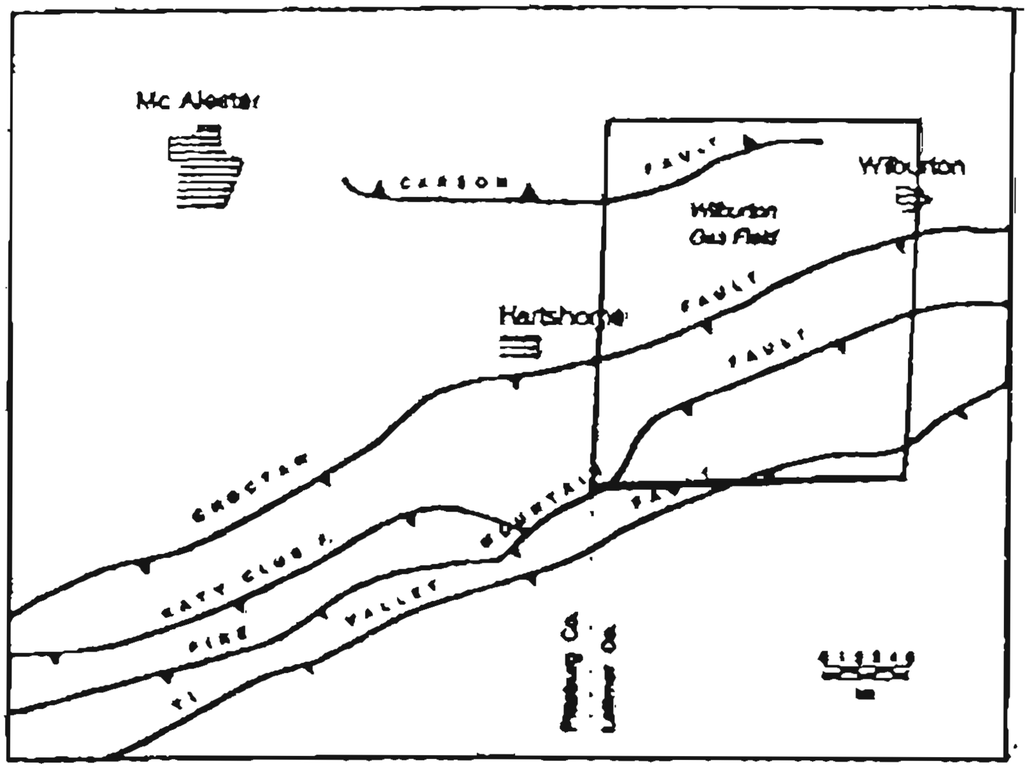


Figure 28.--The surface trace of main thrusts in the study area (after Valderrama and others, 1994).

Several geological aspects of the Ti Valley fault suggest that it may not be a fault with the largest displacement in the frontal belt of the Ouachita Mountains in Oklahoma. The position of the fault in the southwest Ti Valley has been the subject of controversy because Hendricks and others (1947) mapped a large klippe there, clearly indicating that the fault has a very low dip. In the southern part of Ti Valley, Hendrick (1947) observed the facies change of the entire Paleozoic section from the upper Ordovician Womble shale to the Morrowan Johns Valley shale with olistromes of foreland facies. In addition, he recognized that the slices between the Choctaw and Ti Valley faults show a gradual southward facies change in the Morrowan sequence, especially the Wapanucka Limestone changes into a massive spiculite. The presence of thick flysch sequences of black shale and turbidites also shows a facies change in the Atoka Formation.

Cline (1960) while studying in the Ouachita Mountains, found some sudden stratigraphic changes. A thick sequence (17,000 ft.) of Stanley-Jackfork turbidites, wedge their way into the stratigraphic succession above the Arkansas Navaculite (Devonian - Mississippian) and below thge Wapanucka equivalent. This thiock flysh is absent to the north of the Ti Valley fault. In addition, the stratigraphic throw indicates that it is the largest fault in the entire Ouachitas.

The Ti Valley fault is present in the southern portion of the study area (plate I). The surface attitudes of the Ti Valley are generally quite steep (55-65°) in the study area, as mapped by Suneson and Ferguson (1989). The fault runs west-east, parallel to the bedding. The cross-sections (BB' to HH') show the displacement along the Ti Valley decreasing from west to east. The displacement along the Ti Valley on the westernmost cross-section (B-B') is estimated about 3,600 ft. (0.72 mile) and on the eastern side (F-F') about 1,100 ft. (0.22 mile). The geometry of the Spiro Sandstone in the hanging wall of the Ti Valley fault is associated with imbricate thrust faults. In the extreme western portion of the study area, the Spiro Sandstone appears on the surface at T.3N.- 4N., R.18E. (plate I). The Spiro Sandstone on both sides of the Ti Valley fault shows a gentle dipping pattern with some blind imbricate thrust faults propagating from the Choctaw thrust (plate II-V).

Pine Mountain Fault

The Pine Mountain Fault is located in T. 4N. - T.5N., R. 17E. through 19E. in the Southern part of the study area. The surface trace, as mapped by Suneson and Ferguson (1989) is not continuous throughout the area (plate I). The surface geometry indicates that it is a high angle dipping fault, approximately 65-70 degree southeastward. But the balanced cross-sections constructed in the study area depict that the Pine Mountain fault becomes gently dipping in the subsurface around the depth of 10,000 feet. The Pine Mountain is present between two major faults, the hanging wall of the Choctaw thrust makes the footwall of the Pine Mountain fault and the footwall of the Ti Valley constitutes the hanging wall of the Pine Mountain fault in the study area. It is present as an imbricate from the Choctaw Fault in most of the cross-sections. The fault is running west to east, almost parallel to the bedding and the Ti Valley fault. The displacement along the Pine Mountain decreases from West- East. The Pine Mountain fault on the westernmost cross section (B-B') shows about 0.75 miles of displacement of Spiro Sandstone. The cross section (F-F') at the eastern part of the study area shows the minimum displacement of about 1,000 ft (0.10 miles). The Spiro Sandstone in the footwall block of Pine-Mountain fault is associated with some unnamed blind imbricate thrusts (plates II - VI).

Choctaw Fault and Associated Structures

Choctaw Fault

The Choctaw fault was first mapped by Hendricks and others (1947), and the surface trace is recognized to be the boundary between the Arkoma basin and the Ouachita Mountains. It is located in T.3N.-4N., R.17E. through 19E., and is one of the major fault in the study area (Figure 28). Most of the geological maps show that the fault extends about 120 miles in Oklahoma. Berry and Trumbly (1967) estimated a vertical stratigraphic throw of 11,000 feet in wells located, sec.33, T.5N.,R.18E., along the Choctaw thrust in the southeastern Oklahoma. They stated that the minimum horizontal displacement of the Choctaw fault must be in excess of 15,000 feet judging from overriding of minor slices of the Choctaw fault as demonstrated by the well in sec.4, T.4N., R.18E. Briggs (1962), estimated the stratigraphic throw of six to eight miles based on paleocurrent measurements. Suneson and Perry (1990) interpreted the Choctaw thrust as a simple frontal imbricate of a complex, partially eroded, upper triangle which represents approximately six miles of shortening.

In the study area, the cross-sections (AA'-HH') display the displacement along the Choctaw fault decreasing from west to east. The cross-sections (BB'-DD') in the western part of the study area portray the maximum displacement of the Spiro Sandstone along the Choctaw fault around 9-10 miles (plates II-IV). The cross-section F-F' in the eastern part, is the one which exhibits the least displacement of the Spiro Sandstone along the Choctaw fault, estimated around five and a quarter of miles. The configuration of the surface trace of Choctaw fault from most of the cross-sections indicates that it is eroded at least a quarter mile. This estimation is based on the extrapolating the Choctaw thrust and the Spiro Sandstone in the air until they join, and measure the eroded Spiro Sandstone according to the scale (showed by dashed lines in cross-sections).

Some unnamed faults have been found on the hanging wall of the Choctaw thrust. They join the Choctaw fault in the subsurface. Those faults are mainly identified as the blind imbricate thrusts recognized on the basis of well and seismic data. Only, few of them are mapped by Suneson and Ferguson (1989) on the surface. The surface configuration of these faults shows that they are dipping high angle close to the surface, but in the subsurface they become nearly flat as they join the Choctaw thrust.

The Spiro Sandstone in the hanging wall block of the Choctaw fault is gently dipping, and is associated with numerous imbricate thrust faults. All the cross-sections (AA'-HH') show the thrusted Spiro Sandstone on surface, dipping almost parallel to the Choctaw thrust. The thrust faulting is the only cause that brought the Spiro Sandstone at the surface. The leading imbricate thrust faults which propagade upward from the base of the Choctaw fault brought the Spiro Sandstone at the structurally higher level in the hanging wall block of the Choctaw fault. As a result, the gently dipping Spiro Sandstone in the hanging wall block becomes shallower and steeper southward and finally appears on the surface at T.4N., R.19E. (plates II and IV).

The geometry of Spiro Sandstone in the footwall block is a series of duplexes; steep-gentle dipping Spiro is enclosed by a probable roof thrust and a basal detachment surface. An overturned Spiro Sandstone is also recognized in one cross-section (plate V) and is interpreted as an overturned limb of an anticline. The Spiro Sandstone in the footwall block is also associated with few back-thrusts present only on the western cross-sections of the study area (plates II-IV).

In the Wilburton field, along nearly the entire length of the Choctaw thrust, its footwall on the north and northwest displays north-dipping Atoka shales and Desmoinesian beds. The hanging wall of the dipping Morrowan beds are overlying massive Atokan turbidite sequences (Hopkins, 1968).

Suneson and Ferguson (1989) constructed many cross-sections across the Wilburton gas field area. They concluded that the Choctaw fault is a high angle 70-80° southeast dipping fault. All the cross-sections (AA'-HH') constructed in the area of investigation during this study are based on geologic maps prepared by Suneson and Ferguson (1989). The cross-sections run southward and show high angle 70-80° southward dipping Choctaw fault. The subsurface information indicates that the fault flattens considerably at depth around 12,000 ft. and the structure is basically a series of imbricate overriding thrusts. It serves as the root for the imbricate thrusts of the hanging wall block. The flatness of the Choctaw fault is confirmed by the cross-sections (plates II-IX) and seismic interpretation (Figure 27).

Duplex Structures

Duplex structures are found in the study area in the lower block (footwall block) of the Choctaw fault zone. They are mainly composed of hinterland dipping duplexes. These duplexes probably formed horses within the imbricate slices due to the upward propagation of fault from the basal decollement. They show approximately the same displacement among the different imbricates. The passive roof thrust in this investigation is named as "Lower Atokan Detachment" (LAD). The position of the LAD has been recognized below a sand unit named as the Atokan marker bed which is mostly present at approximately 6,000-7,000 feet in the subsurface. Therefore, the probable roof thrust has been identified around 7,000 feet below the sea level in all cross-sections. The duplexes are floored by Woodford and/or Springer detachment surfaces.

The nature and geometry of duplex structures in the Wilburton gas field and surrounding areas were the most important topic among the structural geologists during the 1980's, working in the frontal Ouachita belt. Arbenz (1984) was the first worker who pointed out the presence of the duplex structures in the Wilburton gas field area. He proposed that the thrust system in the Wilburton area is composed of a series of duplexes and a shallow triangle zone. The credit also goes to Hardie (1988), Reeves (1990), Perry and Suneson (1990), Wilkerson and Wellman (1993), and Cemen and others (1994) who proposed the geometry of the duplex structures in the footwall of the Choctaw fault.

During this study the duplex structures were found in the footwall of the Choctaw fault (plates II-IX) which is mainly composed of hinterland dipping duplexes. The duplex structures consist of south-dipping imbricates (plates II-IX) and north-dipping back thrusts (plates II-V). The cross-sections AA' to GG' exhibit a wide range of duplexes in the area. The geometry of duplex structures changes progressively from west to east (AA'-GG'). The western side of the area (plates II-IV) shows well-developed series of steep dipping pattern of duplexes. Generally, the duplexes are prograding northward forming horses within the imbricate slices due to the upward propagation of faults from the basal decollement. The cross-sections constructed on the west side (plates II-IV) do not show the amount of displacement of Spiro along the Choctaw fault, for the corresponding Spiro on the footwall block is absent. This is the reason why a loose line was placed in the footwall block of the Choctaw fault to restore the cross-sections. The roof thrust and the two basal detachment surfaces are well developed in the western part of the area. Some north-dipping back thrusts are also observed only on the western side of the area (plates II-IV).

The number of horses decreases (11 to 4) eastward in the study area (Figure 29) They become larger and dip gently (plates VIII and IX). The hanging wall Spiro has its corresponding footwall Spiro along the Choctaw fault, and a piercing point can be established along the cross-sections in the eastern part of the study area. Consequently, there is no need to place a loose line on these cross-sections (plates VII-IX). The roof thrust and the Woodford detachment are still observable, but the Springer detachment is not very well developed. In addition, the back-thrusts are not observed in the eastern part of the area

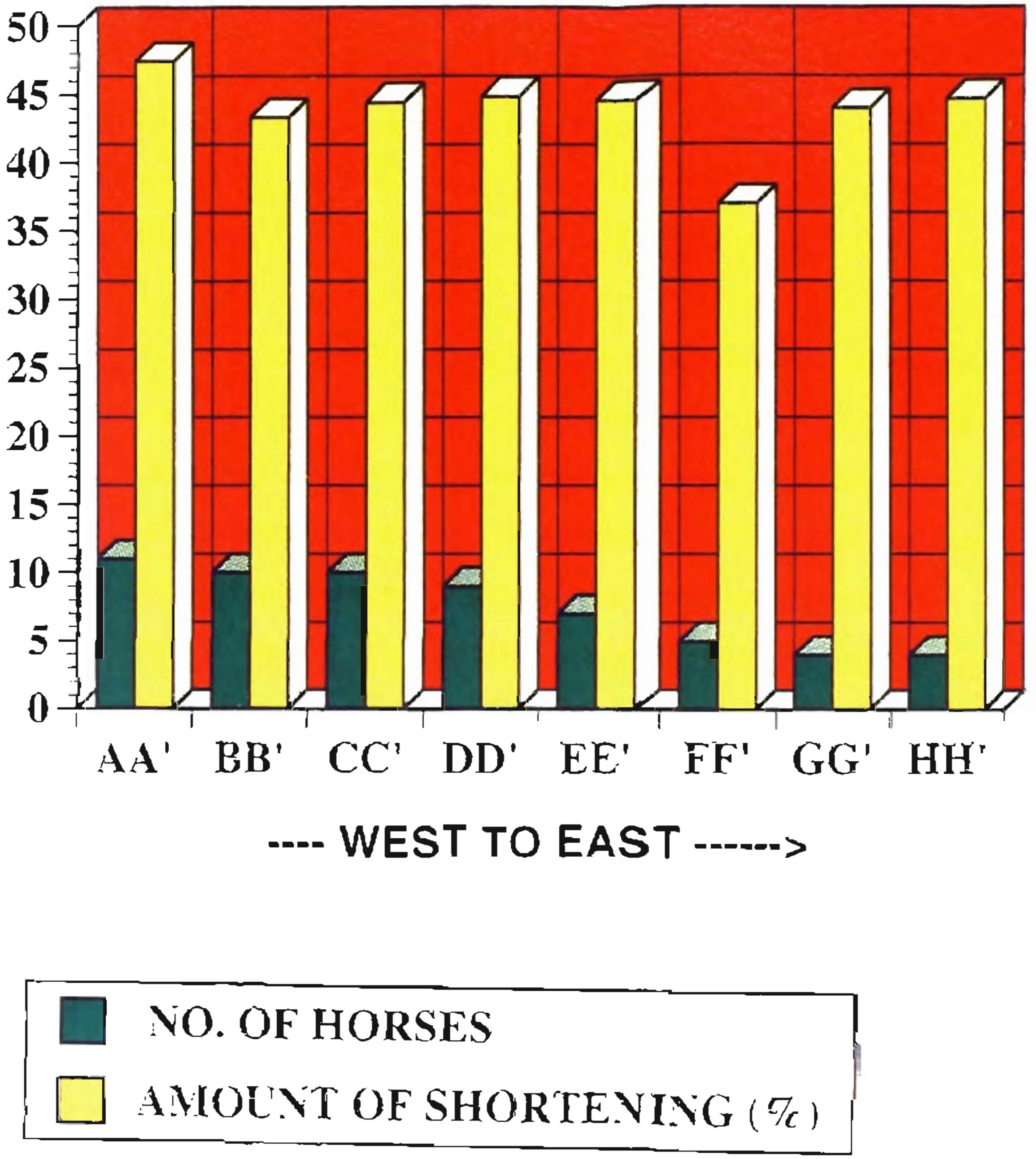


Figure 29 --Histogram showing change in number of horses and the amount of shortening from west to east in the study area.

Carbon Fault

The surface geometry of the Carbon fault is described on the basis of the Suneson and Ferguson maps (1989) (plate I). The Carbon fault is located in T.5N., R.17E., through 19E. It is underlain by the Wilburton and Adamson anticlines and overlain by San Bois syncline in the study area. It runs from west to east and dips almost parallel to the overlying beds at about 35 degree northward. The surface trace of this fault is very prominent in the western side (T.5N., R.18E.) and it becomes discontinuous toward east (T.6N., R.18E.)

The geometry of the Carbon fault has been a subject of discussion for more than two decades. Many authors proposed different interpretations about the geometry of this fault. According to Berry and Trumbly (1967), the detachment below the Choctaw fault had been identified as the Carbon fault splaying upward to the north from the duplex zone (Figure 30). Arbenz (1984) interpreted the Carbon fault as a north dipping low angle fault which is probably detached from the basal detachment (Figure 31). Suneson and Perry (1990) interpreted a seismic profile across the Ouachita frontal zone. They showed the Carbon fault as a north -dipping fault formed by a back-thrust movement of the roof thrust (Figure 32). On the other hand, Valderrama and others (1994), interpreted the Carbon fault as a basement-involved reverse fault below the Choctaw fault (Figure 33). Figure 34 represents the proposed thrust geometry in the Wilburton gas field and surrounding areas along the cross-section line D-D'.

The cross-sections (A-A' to E-E') constructed during this investigation show that the characteristics of the Carbon fault in the subsurface are concordant with the surface geometry of the fault (plate I). The well log data in the eastern part of the study area show that the Carbon fault dips very gently northward and flattens at depth around 3,000 ft. The western cross-sections (plate II-V) display the north-dipping Carbon fault

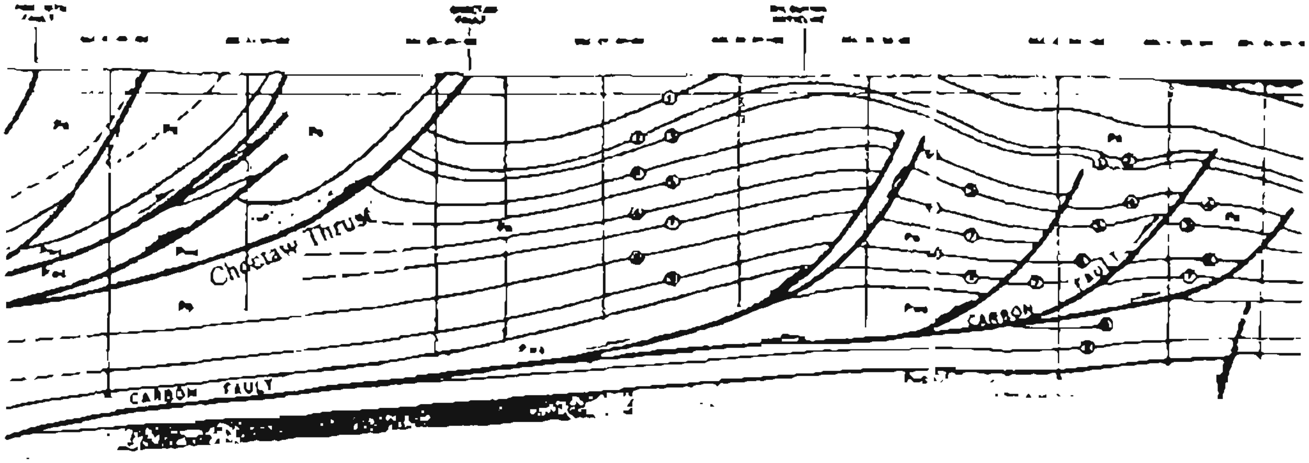


Figure 30.--Interpretation of thrust system in the Wilburton gas field (Berry and Trumbly, 1967).

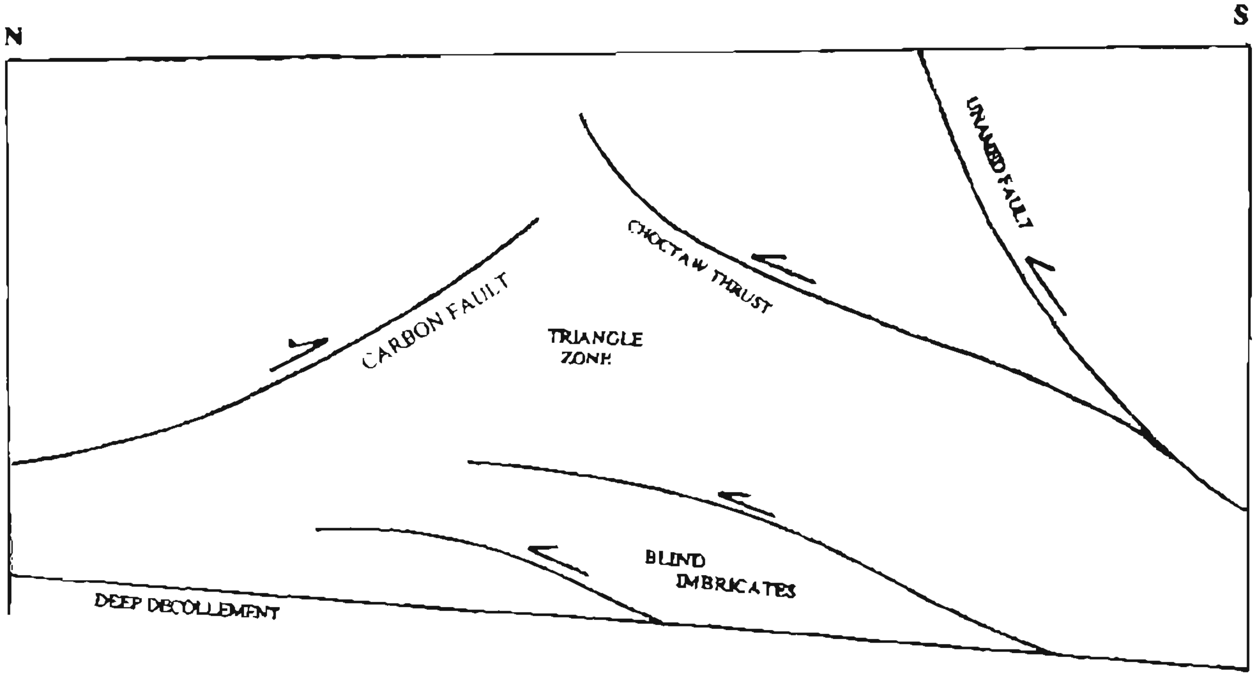


Figure 31.--Geometry of thrust system in the Ouachita Mountain frontal belt (Arbenz, 1984).

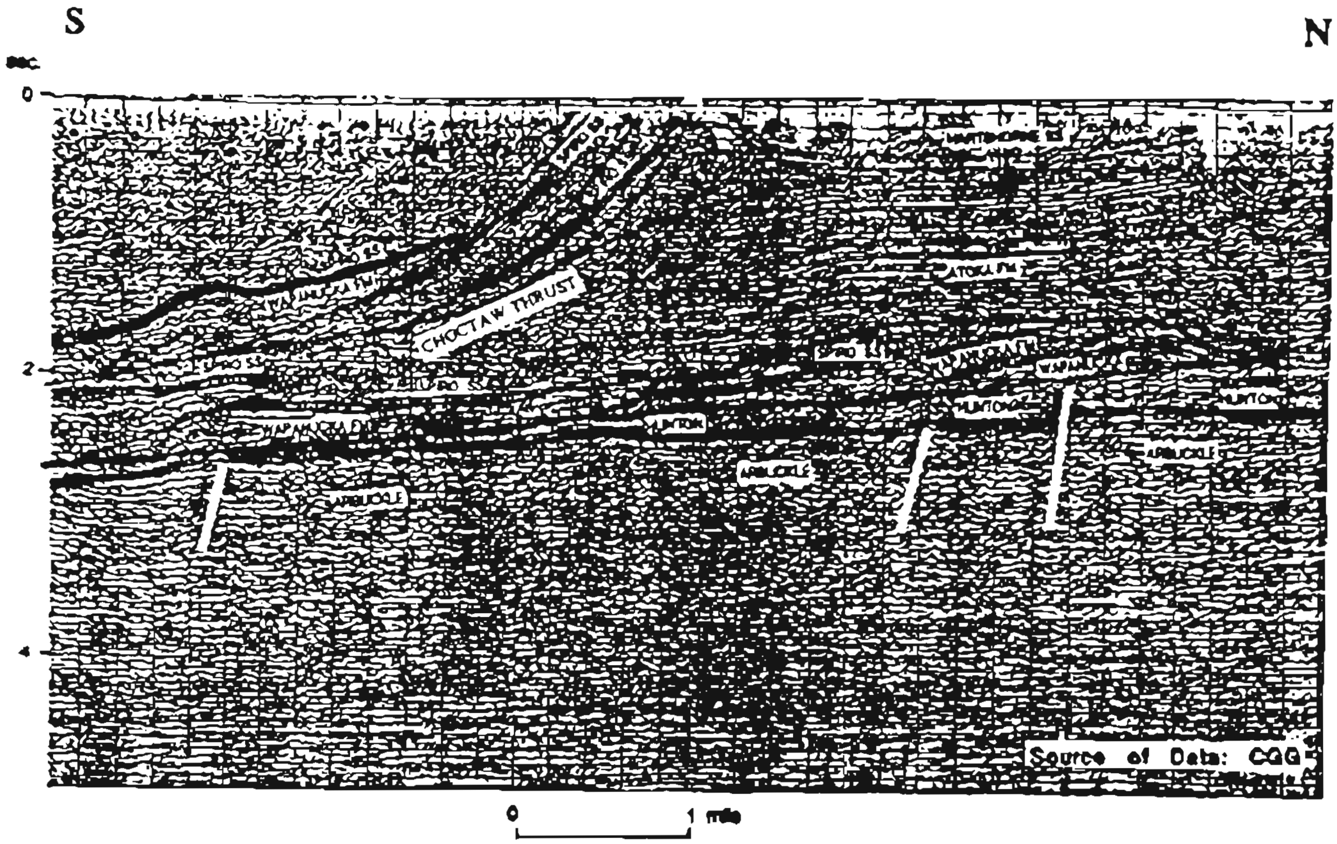


Figure 32 --Interpreted seismic profile across the Ouachita frontal belt (Perry & Suneson, 1990).

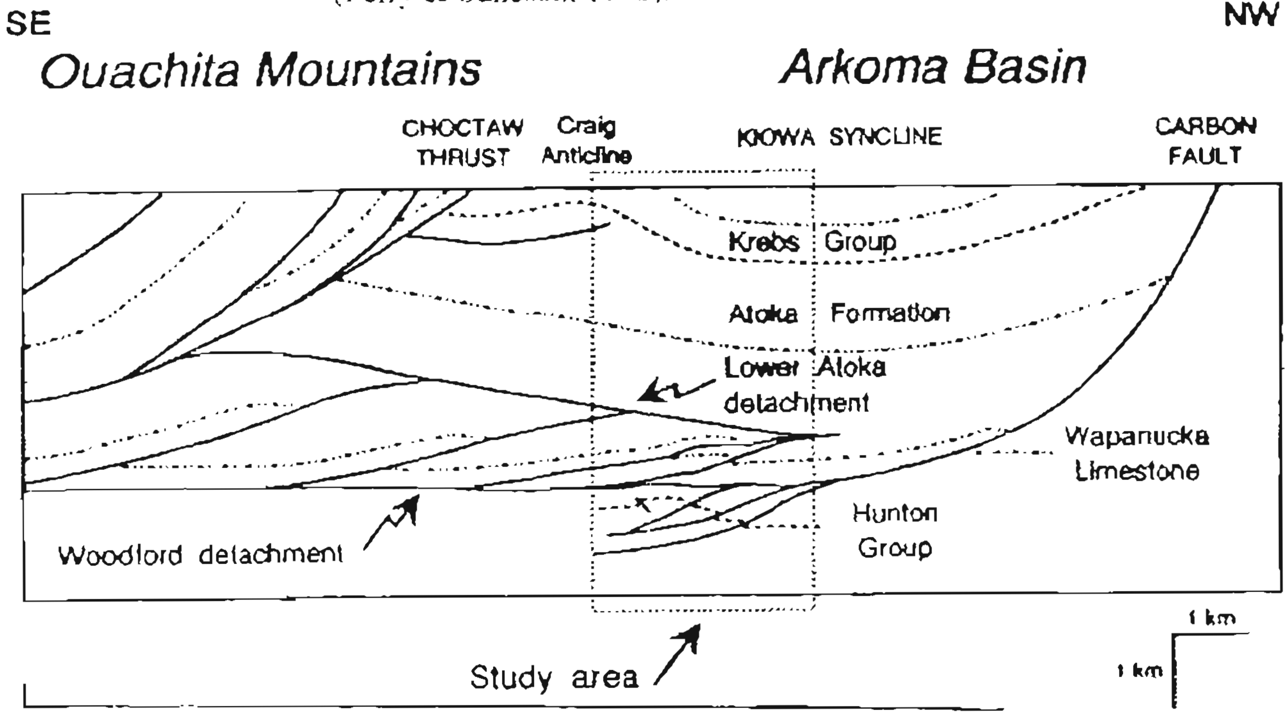


Figure 33.--Schematic cross-section in Ouachita Mountains (Valderrama and others, 1994).

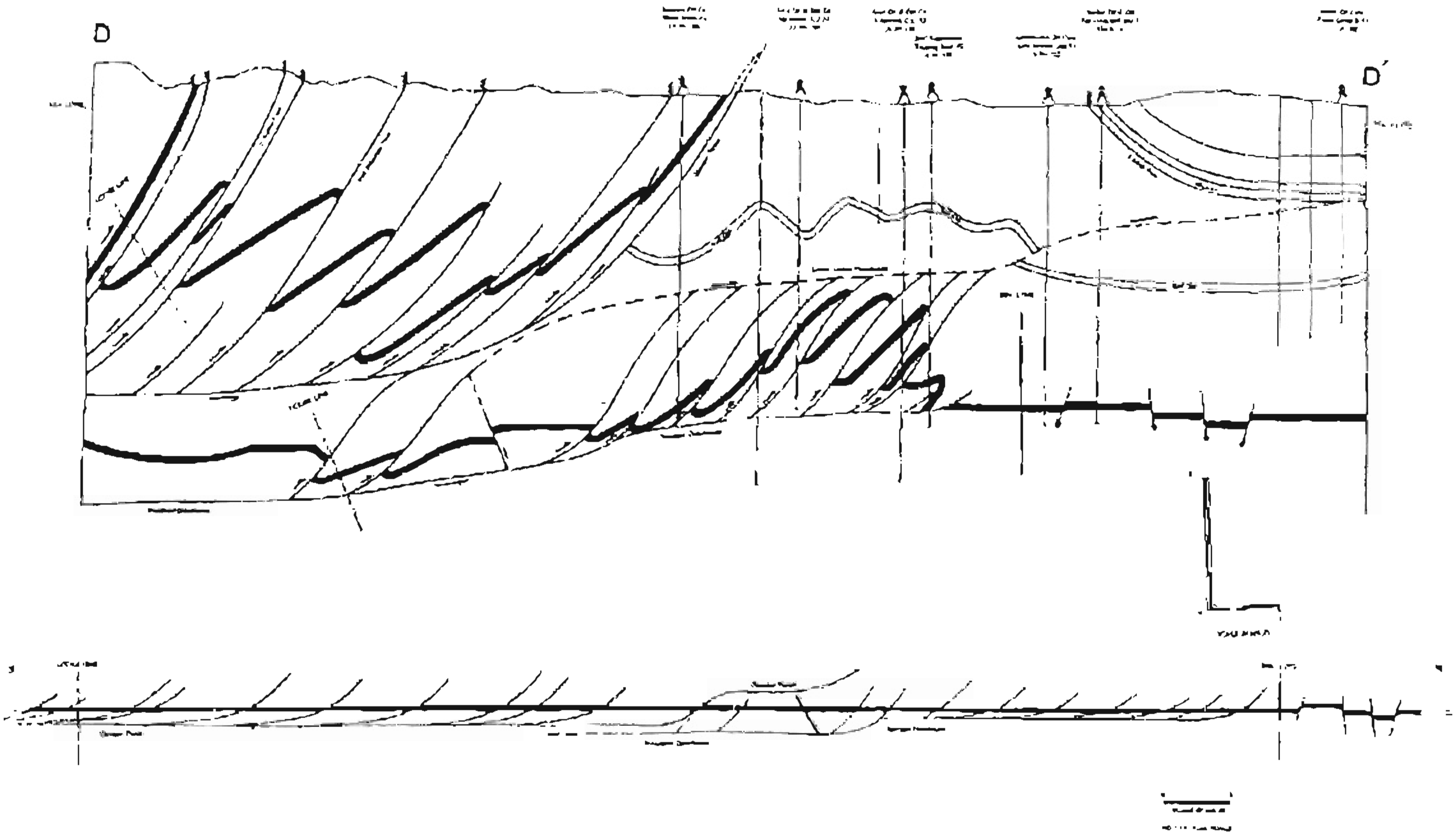


Figure 34.--Cross-section along the D-D' showing the geometry of the study area (see plate 1 for location).

clearly, but those on the eastern side (plates VI-IX) do not show a fault. This research inferred the presence of a probable roof thrust above the duplex structures. With the theoretical assumption that the foreland-verging thrust movement should be accommodated by the back-thrusts (Jones, 1983). The formation of the Carbon fault is then interpreted as the back-thrust of an upward progradation of the probable roof thrust. The upward progradation of the probable roof thrust continuing within the Atoka shale northward when this prograding roof thrust hit the coarse-grained Hartshorne Sandstone, it formed a back thrust within the Atokan Shale because it is difficult to penetrate in the coarse-grained sandstone than the shale. This north-dipping back thrust named as the carbon fault on the surface (plate II-IV). The Desmoinesian rocks which are lying above the carbon fault present unfaulted evident that upward prograding roof thrust is making a back thrust in presence of zero point displacement.

Lower Atokan Detachment (LAD)

The cross-sections constructed during this study depict a probable south-north prograding roof thrust named Lower Atokan Detachment (LAD). The imbricate thrusts from the basal detachment join to the south-north propagating roof thrust (LAD). This upper detachment is probably present in the study area, even though it does not have a clear seismic expression, probably due to the fact that the LAD is within the shales of Atokan Formation and do not provide a well-developed velocity contrast. The wireline well-log data gives a hint about the presence of the probable roof thrust, in terms of change in log signatures at depth where the roof thrust is present. Therefore, the LAD is the zone where the duplexes from the basal detachment merge under the overlying gently folded syncline. The depth of the upper detachment approximates 7,000 ft. in most of the cross-sections. The LAD probably joins the Carbon fault as it propagates northward (plate II-IV). The most convincing evidence for the forward thrust propagation also involves interaction between thrust sheets due to the tectonic transport direction. Milliken (1988), Reeves and others (1990), Perry and Suneson (1990), Wilkerson and Wellman (1993) noticed the presence of a roof thrust for the duplexes whereas Arbenz (1984), Hardie (1988), and Camp and Ratillif (1989) did not recognize the roof thrust (chapter II). By definition, duplex structures are bounded by a floor thrust and a roof thrust.

Basal Detachment

Two basal detachment horizons exist within the study area. One at the base of the Springer Shale, named "Springer" detachment (Berry and Trumbly, 1968 : Hardie, 1988) and the other one at the base the Woodford Shale, referred as the "Woodford" detachment (Hardie, 1988). In fact, there is only one detachment surface, but it is termed in respect to the host formation (plates II-IX). These detachment surfaces are interpreted on the basis of wireline well-log and seismic data. The Springer detachment (Figure 27) exhibits sandy/silty shale, a medium resistivity log response (Exxon, Rutherford #1), and in seismic profiles, it lies below the Union Valley formation (Cromwell Sandstone). The Woodford detachment (Figure 27) shows tight sandy shale, a low resistivity log response (Rutherford #1), and a seismic expression above the Hunton group. The depth of the Springer detachment ranges between 12,000-13,000 ft., while the Woodford detachment is identified around 16,000 ft. below the sea level in the study area.

Triangle Zone

The term triangle zone was first used in petroleum exploration at the edge of the Alberta foothills in the 1950's. It has been applied to similar structures in many other deformed belts. "Passive roof duplex" (Bank and Warburton, 1986) and "Buried thrust front (Morley, 1986) are both more descriptive, but the older term "triangle zone" is still most common. According to Jones (1982), the ideal triangle zone is termed by emplacement of one or more foreland-vergent blind step thrust which, rising from a basal bedding plane detachment , flattens with an upper bedding plane detachment in the subsurface. Butler (1982) used the term "triangle zone" for a structure bounded by opposed thrusts within a hanging wall of a major thrust fault.

The triangle zone is usually applied to that part of a deformed belt where erosion has breached the fold belt to expose the underlying thrust belt depending on the triangle zone's taper and the depth of erosion through it. it may constitute the part of a deformed belt which involves the sedimentary rocks (Jones, 1982). He further described that the formation of a simple triangle zone in which he utilized the concept of a ramping monocline (Figure 35): a structure that contains a duplex and that is responsible for the termination of the eastern directed thrusting (back-thrust).

The present investigation shows a shallow triangle zone floored by a probable upper detachment (plates II-V). The geometry of the triangle zone is represented by the Lower Atokan Detachment LAD in the Lower Atoka shale. The flanks of the triangle zone are delimited by the south-dipping Choctaw thrust to the south and the north-dipping Carbon fault to the north (plate II-V). The cross-sections (plates II-IX) show that the Arkoma Basin seems to have an incipient triangle zone which becomes better developed toward the west. The eastern part of the study area (plates VII-IX) does not

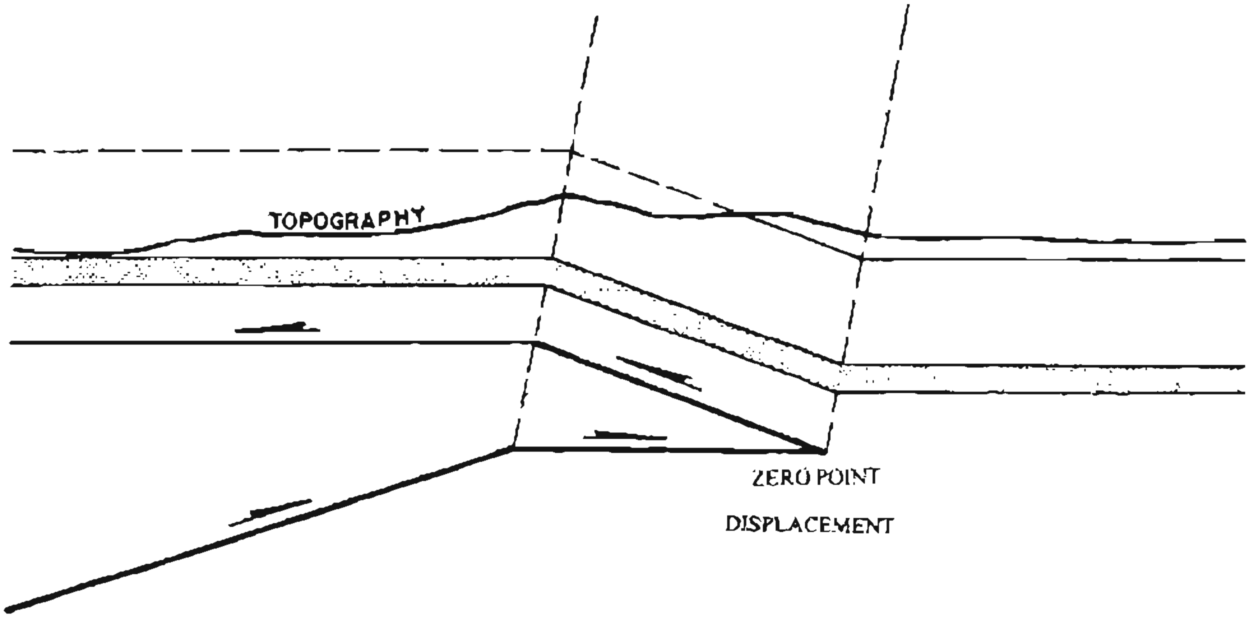


Figure 35.--Cross section shows the formation of back-thrust resulted from the upper detachment (Jones, 1982).

show a triangle zone because the limited available data show that the north-dipping Carbon fault does not exist. The cross-sections also indicate that the triangle zone becomes smaller as the Choctaw fault and the Carbon fault get closer towards the east (compare plates II-VI) and finally disappears with the absence of the Carbon fault in the east (T.6N., R.18E.).

The presence of a shallow triangle zone within the Wilburton gas field can be summarized as :

Firstly, the detachment below the Choctaw is interpreted as a Woodford detachment. Secondly, the Carbon fault is flattened in the northern most part of the area, as it propagates down-to-the north. Thirdly, the Woodford detachment extends upward as a lower Springer detachment in Springer shale. These two decollements and a probable roof thrust, locally named as "lower Atokan Detachment" formed a duplex type of structure in the area of investigation. Later, this upper detachment joins the Carbon fault as it propagates northward. All these features end up with the formation of a triangle zone.

Amount of Shortening in the Wilburton Gas Field and Surrounding Area

Eight restored cross-sections have been constructed from the balanced cross-sections to determine the amount of shortening along the Spiro Sandstone within the study area (plates II-IX). The method used to construct the restored cross-sections is called 'sinuous-bed method, that was introduced first by Dahlstrom (1969). The estimated average shortening from the restored cross-sections is approximately 45 % in the study area (Figure 29). Hardie (1988) drew many cross-sections in Pittsburgh quadrangle, T.3N., R.14E. to the south-west of the present study area. He estimated the shortening about 40%.

The calculation for the amount shortening in the area is given below:

Shortening along the for Cross-section A-A'

l_o = 32.7 x 4000 / 5280 = 24.75 miles.
l_f = 17.2 x 4000 / 5280 = 13.02 miles.
Δl = $(l_o - l_f)$ = (24.75 - 13.02) = 11.72 miles.
e = $- \Delta l / l_o$ - 11.72 / 24.75 x 100 = 47.39 % of shortening.

Shortening along the Cross-section B-B'

l_o = 43.83 x 4000 / 5280 = 33.15 miles.
l_f = 24.8 x 4000 / 5280 = 18.77 miles.
Δl = $(l_o - l_f)$ = (33.15 - 18.77) = 14.37 miles.
e = $- \Delta l / l_o$ - 14.37 / 33.15 x 100 43.36 % of shortening.

Shortening along the Cross-section C-C'

l_o = 44.3 x 4000 / 5280 = 33.53 miles.
l_f = 24.6 x 4000 / 5280 = 18.62 miles
Δl = $(l_o - l_f)$ = (33.53 - 18.62) = 14.91 miles.
e $- \Delta l / l_o$ - 14.91 / 33.53 x 100 = 44.46 % of shortening.

Shortening along the Cross-section D-D'

l_o = 23.6 x 4000 / 5280 = 17.86 miles.

l_f = 13.0 x 4000 / 5280 = 9.84 miles

Δl = $(l_o - l_f)$ = (17.86 - 9.84) = 8.01 miles.

e = $- \Delta l / l_o$ - 8.01 / 17.86 x 100 = 44.89 % of shortening.

Shortening along the Cross-section E-E'

l_o = 24.0 x 4000 / 5280 = 18.16 miles.

l_f = 13.3 x 4000 / 5280 = 10.06 miles

Δl = $(l_o - l_f)$ = (18.16 - 10.06) = 8.10 miles.

e = $- \Delta l / l_o$ - 8.10 / 18.16 x 100 = 44.60 % of shortening.

Shortening along the Cross-section F-F'

l_o = 31.2 x 4000 / 5280 = 23.61 miles.

l_f = 15.3 x 4000 / 5280 = 15.29 miles

Δl = $(l_o - l_f)$ = (23.61 - 15.29) = 8.31 miles.

e = $- \Delta l / l_o$ - 8.31 / 23.61 x 100 = 37 % of shortening.

Shortening along the Cross-section G-G'

l_o = 29.4 x 4000 / 5280 = 21.91 miles.

l_f = 16.2 x 4000 / 5280 = 12.26 miles

Δl = $(l_o - l_f)$ = (21.91 - 12.26) = 9.64 miles.

e = $- \Delta l / l_o$ - 9.64 / 21.91 x 100 = 44 % of shortening.

Shortening along the Cross-section H-H'

$$l_0 = 26.8 \times 4000 / 5280 = 20.28 \text{ miles.}$$

$$l_f = 14.8 \times 4000 / 5280 = 11.20 \text{ miles}$$

$$\Delta l = (l_0 - l_f) = (20.28 - 11.20) = 9.08 \text{ miles.}$$

$$e = -\Delta l / l_0 = -9.08 / 20.28 \times 100 = 44.7\ \% \text{ of shortening.}$$

Normal Faults

Normal faults in northern part of the cross-sections (plates II-IX) in the study area were studied to understand their relationship with the thrust system. The normal faults are earlier event and developed during the middle Atokan time. They show a northeast to north-northeast trend (Cline, 1967). These faults have throws as much as 2.4 km.

The faults were active during the deposition of lower Atokan. Koinm and Dickey (1967) documented an abrupt increase in the thickness during the Atokan time, especially of the middle and lower members of Atokan, along these normal faults. The thickness range is accompanied by the appearance of turbidite facies to the south, and shows convincingly that the faults were growth faults and active during the Atokan sedimentation. Most of the faults do not cut the upper Atokan, therefore, it can be concluded that the fault movement had been ceased before the upper Atokan sedimentation.

Growth folds are broad and relatively gentle. Where faults present, they are usually normal and downthrown to the south. They have probably been affected to some extent by compression, but they own their origin to differential movement of the basement during basin subsidence (Berry and Trumbly, 1967).

Folds

The rocks exposed in the study area are folded into a series of anticlines and synclines, trending generally east northeast-west southwest, sub-parallel to the overall basin trend (plate I). Faults usually strike parallel to the fold axes (Figure 36), and most of the folds are open (Cline, 1968). The folds in the study area can be traced for tens of miles in an east-west direction. The folds found in Oklahoma portion of the Ouachitas are shorter than the Arkansas portion and display both frequent relaying and changes of strike from east-west to northeast-southwest (Arbenz, 1989).

The Sans Boise syncline is a broad syncline and is located above the Carbon fault in the northern part of the study area within the Desmoinesian strata. In the Wilburton area, anticlines are asymmetric to the north, and they are thrust-faulted at depth. These thrust faults are foothill-like imbricates of the Choctaw fault to the south.

The pattern of folding in the Oklahoma may be the result of a two-phase compression, first from the north and then from the northwest, based on structural maps drawn by previous investigators. But it is more likely that the fold pattern is strongly influenced by the pre-existing Atokan growth fault pattern (Nielson and Leonhardt, 1983). The individual folds have the tendency to be asymmetrical to the north where outcrop quality permits anticline to show the thrust-imbricated cores (Haley and Hendricks, 1968).

Folds are mostly found in the southern part of the Arkoma basin near the Ouachita front. They exhibit considerable structural disturbance, and are frequently associated with thrust faults.

Figure 36 --Overturned syncline in Atokan sandstone showing the overturned faults.

CHAPTER VI

SUMMARY AND CONCLUSIONS

The major accomplishments of this study can be summarized as follow :

1. The study area comprises both compressional and extentional features; two blocks have been identified: the thrust system in the upper block classified as leading imbricate fan, and the lower block classified as hinterland dipping duplexes.
2. The leading edge of the thrust belts exhibits many structural features: triangle zone, duplex structures, imbricate thrusts, back thrusts, upper detachment, and basal decollement.
3. A shallow triangle zone delimited by Choctaw thrust to the south and Carbon fault to the north is floored by the lower detachment surface (LAD). The rock units forming the floor of the triangle zone are more often gently deformed, and primarily consist of south dipping, block-faulted structures dominated by high-angle, deep seated reverse faults.
4. Duplex structures present in the footwall block of the Choctaw fault form horses.
5. Choctaw thrust is recognized as the major thrust fault, dipping approximately 70-80° southward. Moreover many blind and emergent thrusts are associated with the Choctaw thrust.
6. Two basal decollement surfaces have been recognized within the study area, named on the basis of the host formation: Springer and Woodford detachments.
7. The average amount of shortening calculated from the restored sections of the eight balanced cross sections approximates 45%.

8. In the study area the Spiro is mainly composed of rarely fossiliferous sandstone. The presence of fossils, limestones, burrows and bioturbations, and chamosite pellets (in the subsurface only) suggests a shallow-marine environment.
9. Medium to coarse grained non-fossiliferous sandstones contain the highest porosity in the area of investigation.
10. Hydrocarbon production in the Wilburton gas field area mainly comes from the Spiro Sandstones, most prolific reservoir in the study area.

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APPENDIX

WELL LOGS USED TO CONSTRUCT VARIOUS CROSS SECTIONS

S.NO.	OPERATOR	WELL NAME	LOCATION	DEPTH SPIRO (feet)
1	EXXON	RUTHERFORD #1	05-3N-18E	5,900/14,900
2	SHELL	RUTHERFORD #1-24	24-4N-17E	6661/12826
3	SINCLAIR	MOSE C WATTS #1	03-4N-18E	N.D.E.
4	SHELL	WILLIAMS-MABRY #1-4	04-4N-18E	544
5	TENNECO	MABRY TRUST 1-5	05-4N-18E	11,372
6	SHELL	EVERY 1-5A	05-4N-18E	N.D.E.
7	SHELL	MABRY 1-5	05-4N-18E	N.D.E.
8	AUSTRAL	MABRY 1-7	07-4N-18E	3,182
9	SHELL	MARBY 1-9	09-4N-18E	716/4973/12826
10	ARCO	DOLLINS # 1-13	13-4N-18E	13,374
11	BTA	JVP-AMASON # 1	24-4N-18E	12,629
12	UNKNOWN	UNKNOWN	25-4N-18E	14,307
13	ARCO	ULYSSES # 1	35-4N-18E	14,486
14	MOBIL	E M.LAWLESS # 1	01-4N-19E	13,446
15	HELMRICH	GARY NO.1-5	05-4N-19E	12,187
16	ARCO	HART #1	06-4N-19E	12,148
17	HELMRICH & PAYNE	BURGER TRUST 1-6	06-4N-19E	12.260
18	EXXON	YOURMAN # 1	09-4N-19E	13,785
19	ARCO	HOLSTEN #1	11-4N-19E	4591/14,077
20	WITEHEAD	SEPTER 1-A	11-4N-19E	N.D.E
21	WITEHEAD	SEPTER 1-B	11-4N-19E	N.D.E
22	SHELL	DIPPLE # 1-17	12-4N-19E	10,301
23	ARCO	JAMES # 1-17	17-4N-19E	11,487
24	ARKOMA	KENNEDY B-1	22-4N-19E	N.D.E
25	H & H STAR	COLONY, 1-23	23-4N-19E	2503/15757
26	SINCLAIR	USA ANDERSON # 1	01-5N-18E	10,725
27	AMBASSADOR	DAVIS UNIT #1	02-5N-18E	10,977
28	AMBASSADOR	KINNIKIN PATE # 1	03-5N-18E	10,966
29	AMBASSADOR	LELA SAWYER # 1	05-5N-18E	10,981
30	AMBASSADOR	CHAUDION UNIT # 1	06-5N-18E	11,276
31	BROCK	BODDY 1-6	06-5N-18E	11280
32	PE PETROLEUM	HUNT # 1	07-5N-18E	N.D.E.
33	PETROLEUM	FERGUSON UNIT # 1	07-5N-18E	N.D.E.
34	SINCLAIR	GARDEN UNIT # 1	07-5N-18E	11,358
35	AMBASSADOR	RAUNIKER UNIT # 1	08-5N-18E	10,985

36	JONES & PILLOW	GEORGE McLAIN # 1	08-5N-18E	10,772
37	AMBASSADOR	TOPPING STATE # 1	09-5N-18E	11,028
38	JMC	TOPPING STATE # 2	09-5N-18E	10,153
39	ARCO	McALESTER # 2A	10-5N-18E	9716
40	AMBASSADOR	McALESTER #1A	10-5N-18E	11072
41	AMBASSADOR	DAVIS A # 1	11-5N-18E	11,212
42	ARCO	DAVIS A # 2	11-5N-18E	11,402
43	AMBASSADOR	ROBINSON UNIT # 1	12-5N-18E	11,235
44	SAMSON	JUNIOR # 1	12-5N-18E	11,591
45	AMBASSADOR	AUSTIN uNIT # 1	13-5N-18E	8,376
46	SAMSON	COSTILOW # 4	14-5N-18E	7,855
47	AMBASSADOR	COSTILOW # 1	14-5N-18E	7,875
48	SAMSON	COSTILOW # 5	14-5N-183	8,110
49	LIMESTONE	McCURRDY # 1	15-5N-18E	N.D.E.
50	ARCO	YOURMAN # 2MA	15-5N-18E	8,274
51	ARCO	YOURMAN # 3	15-5N-18E	7,813
52	ARCO	KILPATRIK # 3	16-5N-18E	8,031/9,567
53	ARCO	KILPATRIK # 2	16-5N-18E	7,936
54	AMBASSADOR	FAZAKAS UNIT # 1	17-5N-18E	8,126
55	ARCO	STEVE FAZAKAS # 2	17-5N-18E	7,892/9,615
56	SINCLAIR	BENNET STATE # 1	19-5N-18E	8,344
57	ARCO	BUD HAMPTON # 1	18-5N-18E	8,384
58	ARCO	SMITH UNIT MA # 2	20-5N-18E	7364/8444/10019
59	AMBASSADOR	PASCHAL UNIT # 1	21-5N-18E	7,618
60	ARCO	PASCHAL UNIT # 2	21-5N-18E	8,244/9,755
61	ARCO	R.F. McALESTER # 2	22-5N-18E	10,933
62	ARCO	R.F.McALESTER # 3	22-5N-18E	7945/8247/10791
63	AMBASSADOR	WILLIAMS A # 3	23-5N-18E	8,310/10,812
64	AMBASSADOR	JAMES UNIT "A" # 1	24-5N-18E	8,320
65	AMBASSADOR	JAMES UNIT # 1	24-5N-18E	8,410
66	SKELLY	GUY VARNUM # 1	25-5N-18E	9,658
67	FERGUSON	WAGGONER # 1	26-5N-18E	95
68	ARCO	E.V.ENIS # 2	27-5N-18E	9,194
69	AMBASSADOR	DOBBS STATE # 1	29-5N-18E	8,066
70	ARCO	STATE C # 2	28-5N-18E	8,397/10,843
71	ARKOMA	HUNTER TUCKER # 3	31-5N-18E	9,163/9,959
72	ARKOMA	KENNEDY B-2	32-5N-18E	10,257
73	SINCLAIR	McWATTS # 1	33-5N-18E	10,608
74	ARCO	DOBBS UNIT MA # 2	29-5N-18E	8,065
75	COQUINA	WATTS # 1	34-5N-18E	10,760
76	SAMSON	MOSE WATS # 1	35-5N-18E	10,679
77	SAMSON	WATTS # 1	35-5N-18E	2,356
78	AUSTRAL	MERBY 1-36	36-5N-18E	2,990
79	PANAM	KIER UNIT # 1	02-5N-19E	15,148
80	PANAM	USA CHOC. 1-3	04-5N-19E	0,674
81	PANAM	REUSCH # 1	03-5N-19E	10,802

82	PANAM	CHOC. T 4 # 1	05-5N-19E	11,363
83	PANAM	CHOC. T-4 # 2	05-5N-19E	10,546
84	PANAM	USA McTIERMAN # 1	06-5N-19E	10.661
85	PANAM	QUIAD # 1	07-5N-19E	11,066
86	SINCLAIR	D.J.BISHOP UNIT # 1	08-5N-19E	11,756
87	HADSON	EAST OK STATE 1-8	08-5N-19E	10,918
88	WILLFORD	BULLARD NO. 1	09-5N-19E	12,103
89	AMAREX	WILLBURTON TOWNSITE 1	09-5N-19E	11,819
90	WILLFORD	BURGER 1-16	16-5N-19E	12,355
91	DONALD SLAWSON	PACE NO. 1-16	16-5N-19E	N.D.E.
92	PITCO	POTEET # 1-17	17-5N-19E	11,519
93	HUMBLE	J.D.HUMPHREY # 1	17-5N-19E	12,525
94	AKERS	JUGRAY # 1	17-5N-19E	N D E
95	SAMSON	DRESSEN UNIT # 1	19-5N-19E	N.D.E
96	FERGUSON	McKEOWN # 1	20-5N-19E	1,842
97	HUMBLE	J.A.RAY UNIT # 1	20-5N-19E	11,567
98	FERGUSON	V.F.W. # 1	21-5N-19E	7,960
99	TEXAS	GIVENS # 1	21-5N-19E	6,959
100	HUMBLE	ENVIR JEWELL # 1	21-5N-19E	12,506
101	SKELLY	M.C.JOHNSON # 1	22-5N-19E	782
102	SHELL	WILLIAMS 32-27	27-5N-19E	2268 12829/14339
103	DANIEL PRICE	CHURCH LAKE # 1	29-5N-19E	12,165
104	CHAPARREL ENERGY	V.F.W. # 1-29	29-5N-19E	11,022
105	SUPERIOR	BABB UNIT # 2	29-5N-19E	N.D.E.
106	AUSTRAL	DIAMOND UNIT 1-30	30-5N-19E	10,174
107	SUN	DIAMOND NO 2	30-5N-19E	10,433
108	AMOCO	A.J.MABRY# 1	31-5N-19E	3,716/11,427
109	AMOCO	VIRGINIA WALKER #1	32-5N-19E	11,657
110	AMOCO	ERLAINE WHEELER # 25-1	36-5N-19E	4493/6824 13816
111	AMAX	PARKER # 1	24-6N-17E	6,984
112	ARCO	BROWN # 1	28-6N-17E	8,645
113	GLAXY	STATE A # 1	13-6N-18E	9,432
114	SNEE & EBERLY	SCRUGGS # 1-14	14-6N-18E	9,321
115	MOBIL	FRANK GLENN "B" # 1	18-6N-18E	N.D.E
116	MOBIL	FRANK GLENN "B" # 2	18-6N-18E	N.D.E.
117	SAMSON	JAN KOWSKY 1-19	19-6N-18E	N.D.E.
118	CLEARY PETRO	CARVER # 2-23	23-6N-18E	10,985
119	PANAM	DUREMUS UNIT # 1	26-6N-18E	11,213
120	ATLANTIC	CONWAY UNIT # 1	30-6N-18E	N.D.E
121	MONSANTO	WILDLIFE # 1	24-6N-18E	10,669
122	PANAM	USA AND. PRITCHARD 1	25-6N-18E	10,927
123	SINCLAIR	MITCHELL UNIT # 1	32-6N-18E	10,915
124	SINCLAIR	R.H.LOWRY # 1	33-6N-18E	10,853
125	BROCK	HECKMAN # 2-34	34-6N-18E	10,590
126	SAMSON	SUNFLOWER # 1-35	35-6N-18E	10,531
127	PANAN	CLAUDE WILSON # 1	35-6N-18E	10,751

128	SUN/TEXAS PACIFIC	FEDERAL CHURCH # 1	36-6N-18E	10,723
129	SAMSON	CHOCTAW # 1	01-6N-19E	8.765
130	ARCO	WILDLIFE # 1	07-6N-19E	9,076
131	MOBIL	FRED SLAWSON UNIT #1	09-6N-19E	8.907
132	MOBIL	TRACY JOHNSON # 1	14-6N-19E	9.527
133	PITCO	SHAW # 1-14	14-6N-19E	9,699
134	MOBIL	HOMER JOHNSON # 2	15-6N-19E	9.366
135	SUNSET	WEAVER # 1	16-6N-19E	9,034
136	OXLEY	WEAVER # 2	16-6N-19E	9,295
137	MOBIL	DOVIE WEAVER D #1	17-6N-19E	9,310
138	LOQUINA	ROBBER CAVE # 1	18-6N-19E	10,442
139	MONSANTO	CAVE # 1	19-6N-19E	10,546
140	BHP	CAVE # 2	19-6N-19E	10,067
141	MOBIL	DOVIE WEAVER B #1	20-6N-19E	10.096
142	MOBIL	DOVIE WEAVER E #1	21-6N-19E	9.946
143	TENNECO	ARK.KRAFT. # 1-25	25-6N-19E	12.590
144	TENNECO	ARK. KRAFT # 1-26	26-6N-19E	12.725
145	SAMSON	YOUNG RANCH # 2	27-6N-19E	10,083
146	AUSTRAL	YOUNG RANCH # 1-28	28-6N-19E	10,212
147	SAMSON	KILPATRICK # 2-29	29-6N-19E	10,309
148	MOBIL	HACKEY # 1	30-6N-19E	10,550
149	PANAM	E. OK AM # 1	31-6N-19E	10,540
150	PANAM	QUAID UNIT "B" # 1	32-6N-19E	10.308
151	AMOCO	ADAMS C # 2	33-6N-19E	10.276
152	PANAM	ADAMS C # 1	33-6N-19E	10,375
153	WILLIFORD	WILSHIRE YOUNG #1	34-6N-19E	10,894
154	TENNECO	ARK. KRAFT #1-35	35-6N-19E	12,918
155	TENNECO	SHAW # 1-36	36-6N-19E	12.944

VITA

Saleem Akhtar

Candidate for the Degree of

Master of Science

Thesis : THE GEOMETRY OF THRUST SYSTEM IN THE WILBURTON GAS FIELD AND SURROUNDING, LATIMAR COUNTY, OKLAHOMA

Major Field : Geology

Biographical :

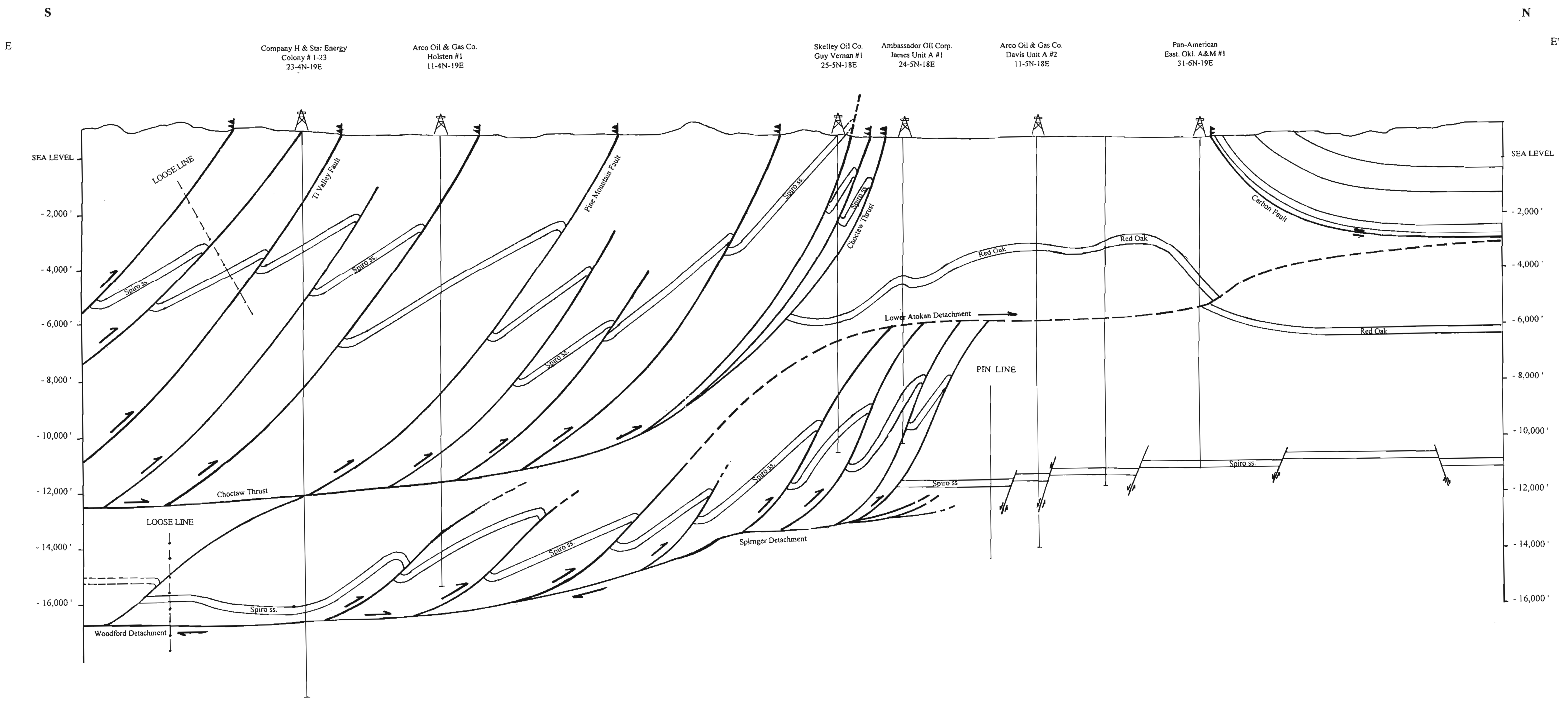
Personal Data : Born in Karachi, Pakistan, April 6, 1965, the son of Mr and Mrs. B.A.Khan.

Education : Received Bachelor of Science in Geology from University of Karachi, Pakistan in 1989; received Master of Science degree in Geology from University of karachi, Pakistan in 1990. Completed the requirements for the Master of Science degree at Oklahoma State University in May 1995, with a major in Geology.

Professional Experience :Research Assistant, School of Geology, Oklahoma State University, August 1992 to present; Assistant Observer & Seismologist, Seismograph Services Ltd. (SSL), February 1991 to November 1991.

Professional Memberships : Member American Association of Petroleum Geologists, Member Houston Geological Society.

Plates 1, 2, 3, 4, 5, 7, 9, are missing. Plates 6 and 8 are present.

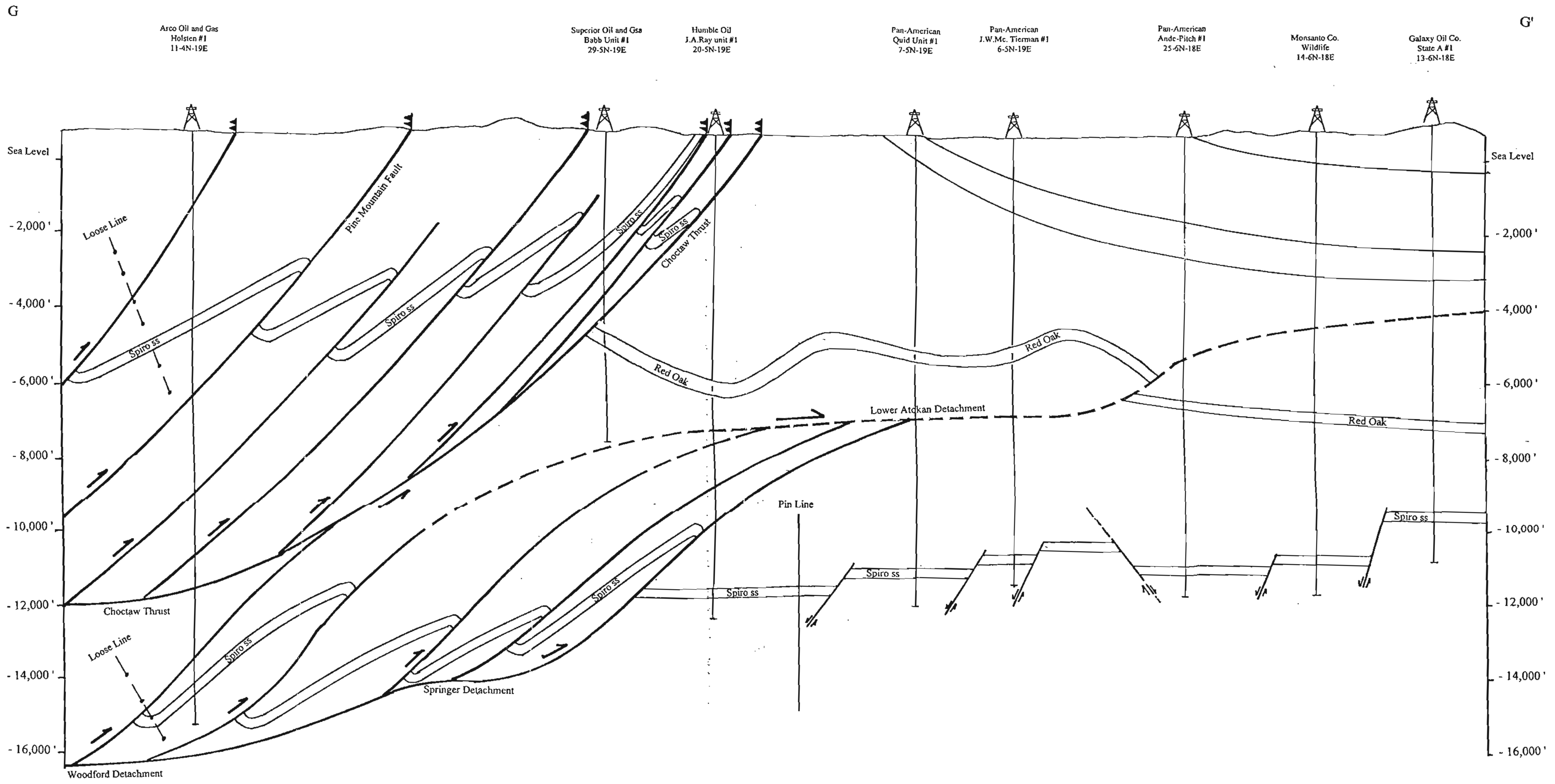


SOUTH - NORTH
Balanced and Restored Cross Sections
E-E'

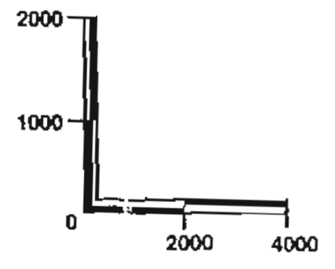
2000
1000
0
2000
4000
Scale in ft.

Arkoma Basin
Plate VI

Saleem Akhtar
M.S. 1995



SOUTH - NORTH
Balanced and Restored Cross Sections
G G'



Scale in ft.

Arkoma Basin Plate VIII

Saleem Akhtar M.S. 1995