DISTRIBUTION AND DEPOSITIONAL SYSTEM OF SPIRO SANDSTONE IN ARKOMA BASIN OF EASTERN OKLAHOMA

(LeFlore, Sequoyah, Latimer, Haskel, Muskogee)

By

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DISTRIBUTION AND DEPOSITIONAL SYSTEM
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IN THE
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(LeFlore, Sequoyah, Latimer, Haskell, Muskogee)

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ABSTRACT

The Arkoma basin has become a significant dry-gas producing region in recent years. The lower Atokan Spiro sandstone (Pennsylvanian) is one of the main gas producing sandstone in the Arkoma Basin of Oklahoma. The recent explosion in gas exploration from the Spiro has been providing valuable new data for more comprehensive study of this formation.

The Spiro sandstone consists of vertically stacked sandstone bodies that deposited in a wave-dominated delta during the Earliest Atokan time. The mission of this study was to approach various aspects of sedimentation, stratigraphy, structure, and distribution of the Spiro sandstone that influence the distribution and production of dry-gas in the region.

Electric log signatures, isolith maps, isopach maps, and stratigraphic cross sections were used in the study to determine the depositional environment of the Spiro, to locate the location of the incised valleys, and picture the deltaic system which accumulated the Spiro sandstone. Also, based on the combination of the structural map of the top Wapanucka and the isolith map of Spiro sand stone, possible types of trapping mechanisms in the study area were prepared. Four channelized sandstone bodies, orienting NW-SE, were identified based on the isolith maps of both Spiro and sub-spiro shale. Spiro sand body is considered to be formed during three subsequent stages; (1) formation of the incised valleys and initiation of the Spiro Delta formation (regressive system), (2) Lowstand system tract, where wave-dominated Spiro Delta deposited most

of its strata, (3) rapid transgression and reworking of the Spiro delta were followed by drowning of the delta, depositing the overlying shale in a relatively high sea-level.

By enlarging the study area, the nature of Spiro sand stone may be understood more accurately. Location of subtle traps for hydrocarbon may be understood better if a regional depositional pattern of the Spiro sandstone is constructed throughout the Arkoma Basin.

INTRODUCTION

Purpose of Study

The lower Atokan Spiro formation is a well-known gas reservoir in the frontal Ouachite thrust belt and the Arkoma basin. The recent explosion in gas exploitation from spiro sands in the Arkoma basin is providing valuable new subsurface data for a comprehensive stratigraphic study of this formation. Few comprehensive studies of the Spiro formation in the study area exist. Therefore, the mission of this study is to provide a sedimentologic view of this interval through an investigation of the geometry and sedimentary characteristics of the sand units on a regional basis. In addition, a stratigraphic model is needed to predict reservoir quality facies in the subsurface. Also, it is a necessity to delineate the areal extent and trends of the sands within the Spiro sandstone, the primary gas reservoir.

Method of Investigation

Lower Atokan Spiro sandstone in the study area is completely in the subsurface. Spiro sandstone and Wapanucka limestone boundary can be reached at various depths ranging from about 3,000ft. in the northern portion, about 14,000 ft. in the southern portion of the study area.

Mechanical log surveys from the numerous gas wells were used as a source of data of this study. All of the well logs used in this study were contained in the Oklahoma City

Geological Society Log Library, OSU School of Geology Log Library. About 100 well logs were provided by Dr. Arthur Cleaves. Over 650 logs were selected from wells that penetrate the interval and that are free of faulting within the interval. Exploration in the northern and southern portion of the area has not proceeded as rapidly as other areas due to inconvenient depth of the interval, either too shallow or economically too deep. Any log that could be correlated in these areas was utilized. Wells selected for the study were assigned numbers and plotted on a well location map (Plate IX). Well name, location, and stratigraphic data are included in appendix A.

Electric and radioactive logs were used to correlate units across the study area. Numerous north-south and east-west stratigraphic cross-sections were prepared throughout the area to demonstrate the continuity of units across the basin and establish their stratigraphic relations. Seven stratigraphic cross sections are presented in this report. Six types of stratigraphic maps were constructed from the well logs. One structural and one mechanical maps were prepared. Thickness values for each units were plotted and countered to construct isopach maps. Isolith maps were constructed using footage of clean sand or lime for each unit. The gamma-ray logs curve is most useful in mapping matrix free sand and most isolith values were derived from gamma-ray curves. Where gamma-ray logs were not available the spontaneous potential curve was used for isolith values.

The various stratigraphic maps and cross-sections establish a basis for a lithostratigraphic framework, the direction of sediment transport and an interpretation of the depositional system active during the time of deposition.

Location of the Study Area

The Arkoma basin is an elongate tectonic province that extends about 250 miles across parts of eastern Oklahoma and central Arkansas (Figure 1). The Arkoma basin is bounded on the north by the Ozark uplift, it grades northwestward onto a northern shelf area (also called the Cherokee platform). To the south the Arkoma basin is bounded mainly by the frontal belt of the Ouachita mountains, and on the Southwest it abuts the Arbuckle Mountains. Arkoma basin is characterized by great thickness of sedimentary rocks in the Eastern Oklahoma; about 5,000ft.-20,000ft.(Johnson, Kenneth,1988). The basin is filled with a succession of Paleozoic rocks dominated by the Atoka formation of early Pennsylvanian age (Zachry, 1977).

The area of concern in this study lies in the northeastern part of Arkoma basin of Oklahoma, including LeFlore, Sequoyah, Latimer, Haskel counties, and the southeastern corner of the Muskogee. The area of investigation consists of thirty-five (35) townships (T.7N -11N., R.21E.-27E.) covering 1,185 square miles (Figure 2). The Study area covers the eastern part of Kinta, the northeastern part of Red Oak-Norris, Cameron, Cedars, Peno, S.Bokoshe, Poteau-Gilmore and several other small and several other small scale gas fields (Figure 3).

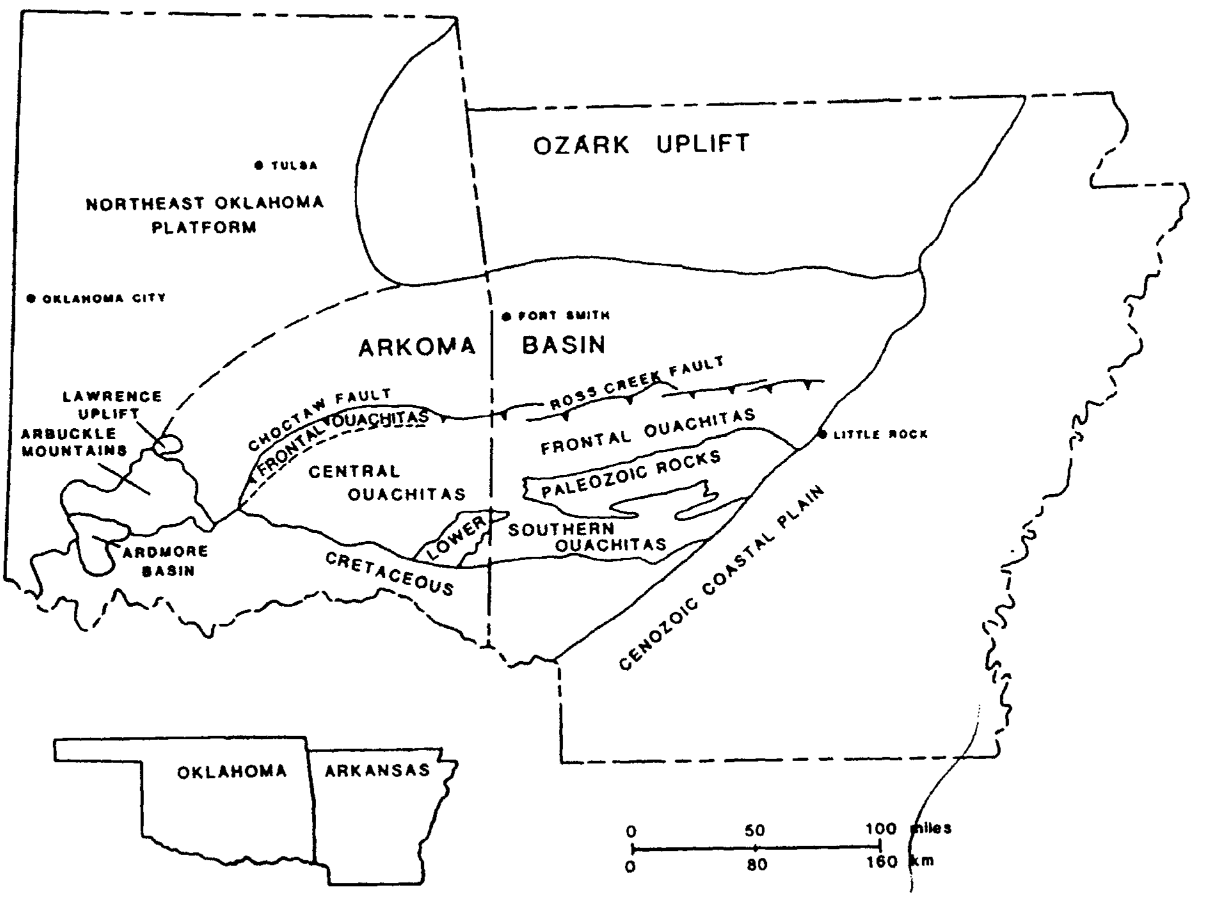


Figure 1. Arkoma Basin of Oklahoma and Arkansas, and surrounding geological provinces (From Zachry and Sutherland, 1984).

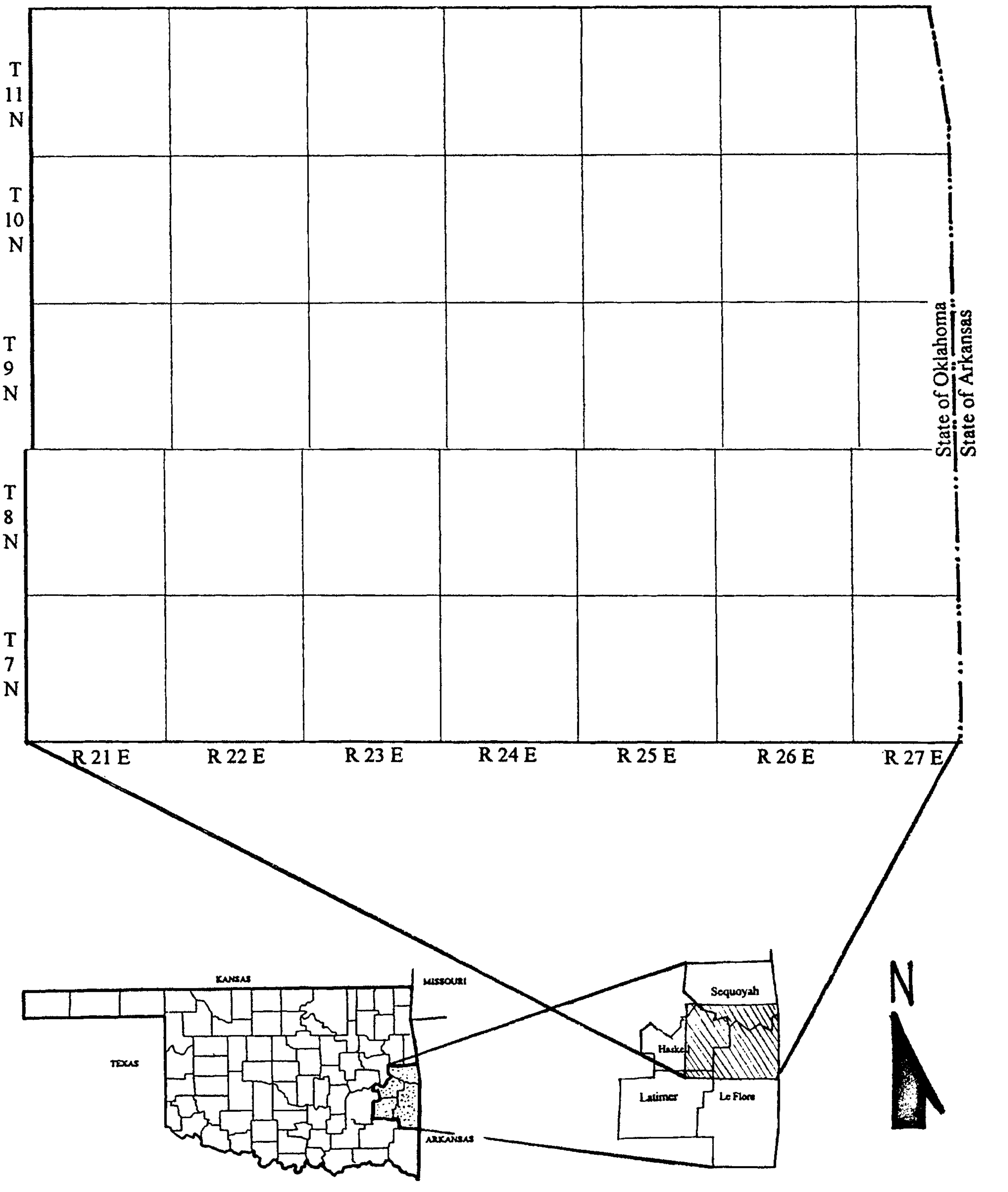


Figure 2. Geographic Location of the Study Area

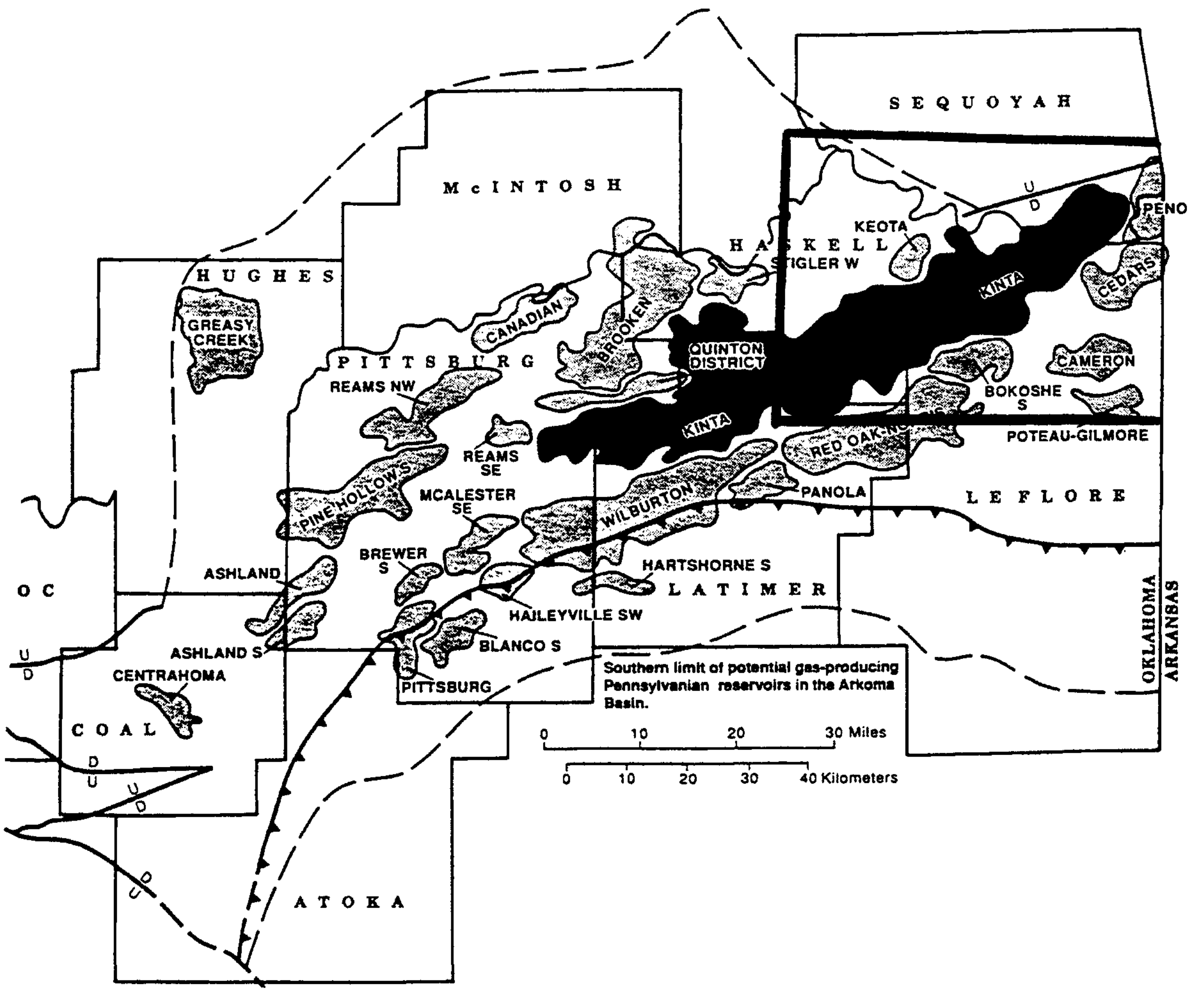


Figure 3. Oil and Gas Fields in parts of the Arkoma Basin and Ouachita Mountains of Eastern Oklahoma (From Johnson and Campbell, 1993)

Objectives

A better understanding of distribution, geometry, size, and trends of Spiro sandstone may lead to a more accurate approach for the hydrocarbon reserves in the Arkoma basin.

The main objectives of this study were to

- Determine the lateral and vertical distribution of depositional environments in the Spiro sandstone.
- Approach the nature of Wapanucka limestone in term of depositional environments utilizing the isolith and isopach maps.
- Determine erosional pattern and lateral distribution of the contact between the Spiro sandstone the underlying Wapanucka limestone.
- Determine the number of sandstone distributary channels and the character of the deltaic systems during the time of Spiro deposition.
- Link the results of this study and the ones done previously in order to come with a more regional interpretation in terms of deposition and distribution of Spiro sandstone.
- Determine the possible trapping mechanism and possible accumulation spots for hydrocarbons in the study area.

Previous Investigation

The initial regional stratigraphic framework and nomenclature for subsurface exposure of Bloyd and Atoka strata were provided by Owen (1858), Simond (1891), Taft and Adams (1900), Adama and Ulrich (1904), and Purdue (1907) for Arkansas, and Moore (1947) and Blythe (1959) for Oklahoma.

To define the evolution of the Arkoma basin, several tectonic models have been proposed by various scientists such as Keller and Cebull, 1973; Briggs and Roeder, 1975; Walper 1977). Houseknecht (1986) presented a model to explain the formation of the Arkoma basin as evolving from a passive margin at the time of Spiro sandstone deposition (Lower Atoka) to a foreland basin at the end of Atokan deposition (Figure 8). The Arkoma basin of Oklahoma and Ouachita mountains were divided into three tectonic provinces based on Ouachita deformation (Figure 1).

Moore (1947) investigated the Morrowan strata of Oklahoma, and Blythe (1959) interpreted the depositional environments and potential source areas for Atokan strata of Northeastern Oklahoma.

Maravich (1955) named the spiro sandstone after the town of Spiro, LeFlore, Oklahoma. An isopach map of the Spiro sandstone interval (as Foster sand trends and Spiro sheet-like sand) that identified fluvial channels trending northwest to southeast in Haskel, LeFlore, Seqouyah,and Haskel counties of Oklahoma was constructed by Lumsden and other (1971) (Figure 4). The sandstone bodies (trends) were named as Foster channels that were considered to be unconformable with underlying sub-Spiro shale and Morrowan Wapanucka limestone.

Sutherland (1988) interpreted that Spiro deposition occurred on a broad shelf from updip northerly fluvial system to downdip southerly shallow-marine environments (as barrier bar deposits). Houseknecht (1987) described the Spiro in Latimer and adjacent counties as fluvial and marine channels with the interchannel zones containing subtidal and tidal flat facies.

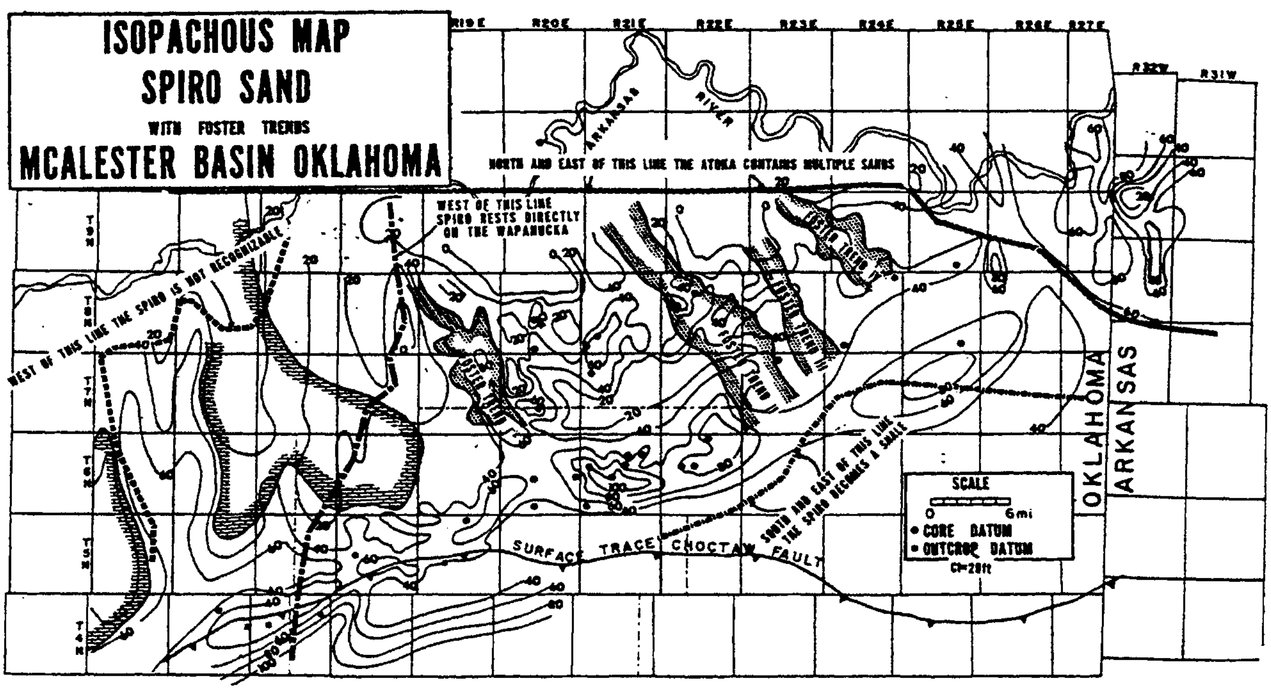


Figure 4. Isopach map of Spiro with four Foster sand trends outlined. Notice east-northeast trend of Spiro and northwest trend of Foster (Lumsden et al., 1971).

Also, he suggested three sediment transport direction during the Atokan. Hooker (1988) interpreted that Spiro was deposited in a marine environment based on the fossil and chemosite content in the sandstone facies. Hinde (1992) and Grayson and Hinde interpreted the Spiro as being Shallow marine bars from work they carried out on outcrop. Carlson (1989) defined the Spiro as being a shelf sandstone or offshore bar deposits. Gross et. al. (1995), utilizing cores and well logs, defined a clastic sequence in Spiro, a progradational and retrogradational facies (Figure 6). According to them, the sequence was formed in a barrier beach involved marine environment.

Several petrographic studies on Atokan Formation are available. Carlson (1989) stated that the Atokan sands were fine- to medium- grained moderately to very well sorted quartz-arenites, subarkoses, and sublitharenites. Based on the presence of the metamorphic rock fragments, he categorized the sand into two groups; northern sandstones and southern sandstones (Figure 5). The metamorphic fragments indicated from an orogenic source, possibly from the advancing Ouchita Fold Belt. The Northern Sandstones (quartz arenite) were drived from the Ozark Uplift.

Surface studies involving the regional stratigraphy of northeast Oklahoma and northwest Arkansas have been continued to date. Hendrics and Parks (1950) established stratigraphic boundaries and nomenclature for north-central Arkansas. Scull (1961) recognized depositional environments for the Atoka strata in the Arkoma basin and Branan (1966) investigated the stratigraphy and its relationship to gas production and emphasized taht the development of porosity zones stratigraphically controlled the Spiro reservoirs. Taff(1901) named the Wapanucka, Chichachoc chert, and Atoka formation as well as older units in Ouachitas. However, he interpreted the chickachoc as being Atokan in age and thus younger than the Wapanucka (Morrowan). Surface studies involving the regional stratigraphy of northeast Oklahoma and northwest Arkansas have been continued to date. Weirich (1953) devised guidelines for exploration geology on the northern shelf edge of the McAlester (Arkoma) basin in eastern Oklahoma.

Sutherland(1988) described the Mississippian and Pennsylvanian geology of eastern Oklahoma and western Arkansas. His article gives an excellent overwiev of the regional stratigraphic relations relating to tectonic events.

Zachry and Haley (1975) remapped the Bloyd-Atoka boundary, which was previously misidentified, and Zachry (1977) clarified this boundary relationship as it exists in northern Arkansas. The regional stratigraphic characteristics of the basal Atoka sandstones and corelations of Morrowan strata in northeastern Oklahoma were provided by Lumsden (1971) and Sutherland and Henry (1977), respectively.

A few studies concerning the local subsurface stratigraphy directly related to this study have been completed. Smith (1914) investigated the structure of the Forth Smith-Poteau gas field. Snider (1914) studied the stratigraphy of east-central Oklahoma, with special reference to the occurrence of hydrocarbons.

The most comprehensive study of Wapanucka/Spiro formation was a lithostratigraphic and biostratigraphic study by Grayson(1980). This study encompassed the entire outcrop belt of these formation. He, like Rowland, also refers to the Spiro as the upper sandstone-limestone member of Wapanucka formation.

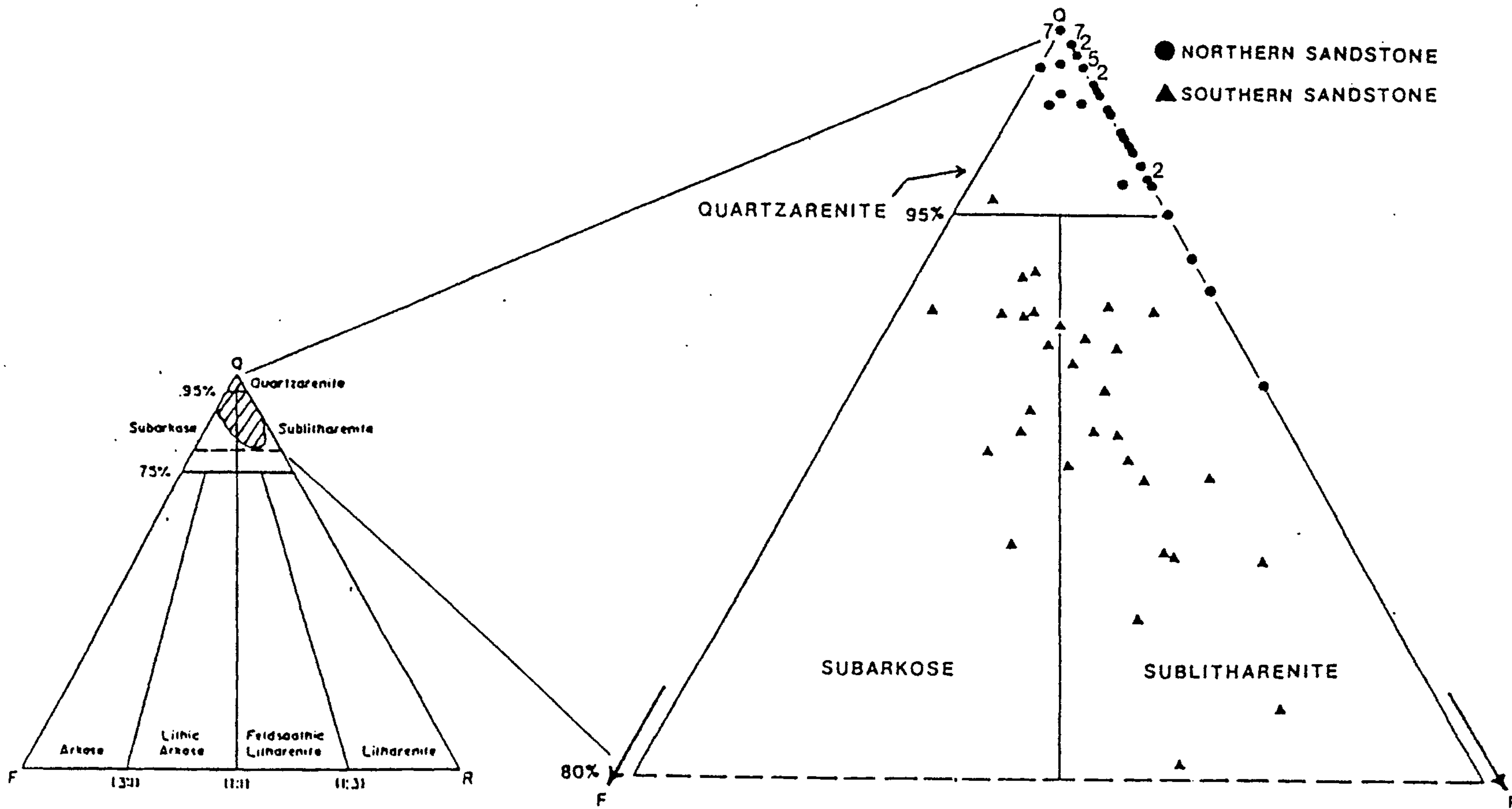


Figure 5: Ternary diagrams showing variations in QFR (Folk,1980) composition of Atoka Formation along western margin of Arkoma Basin (Carlson, 1989).

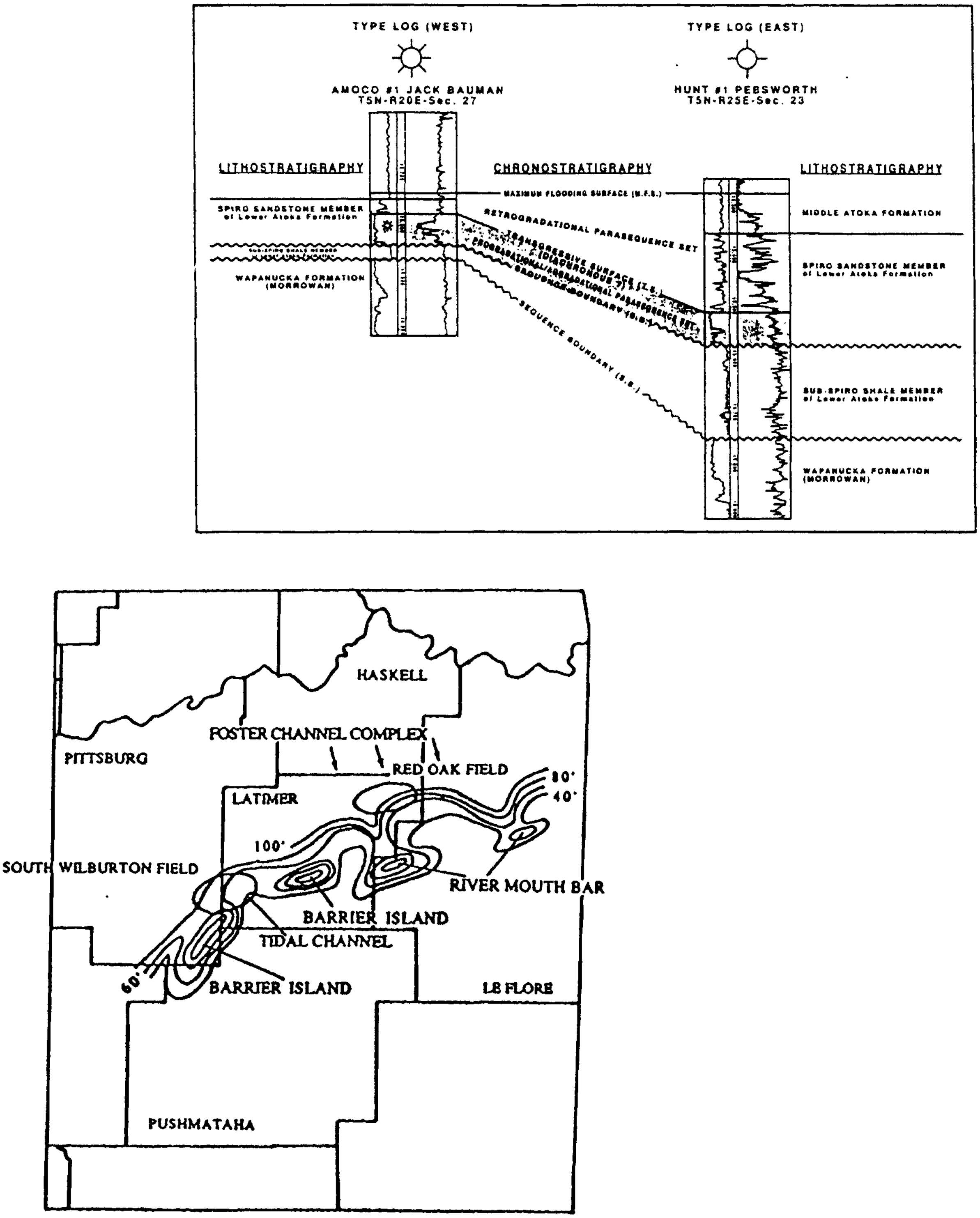


Figure 6: *Upper:* Type log display and stratigraphic column of formations in the Morrowan-Atokan section of the Arkoma Basin (Gross et al., 1995), *Lower:* Isopach map of the progradational/aggradational portion of tha Spiro (Gross et al.,1995).

The term spiro was first used by Wilson(1935) to describe the first sand unit at the base of the Atoka formation in the subsurface. Maravich (1955) named Spiro sand stone after town of Spiro, LeFlore County, Oklahoma. Bowsher and Johnson(1968) were the first to use the term Spiro formation for Hendrics and others (1947) and Rowlands (1974) upper sandstone-limestone member of the Wapanucka formation. Lumsden and other(1990) describe the sedimentology and petrology of the Spiro in the subsurface Arkoma basin.

Kenneth and Campbell (1995) described the characteristics of the Kinta Gas Field, 45% production of which is derived from Spiro sand stone. They constructed a detailed structural map of top Spiro to locate stable traps in the study area.

More recently Hess(1995) descibed Spiro as channel sandstone and sheet-like sand sandstone deposits of a wave-dominated delta (Figure 7).

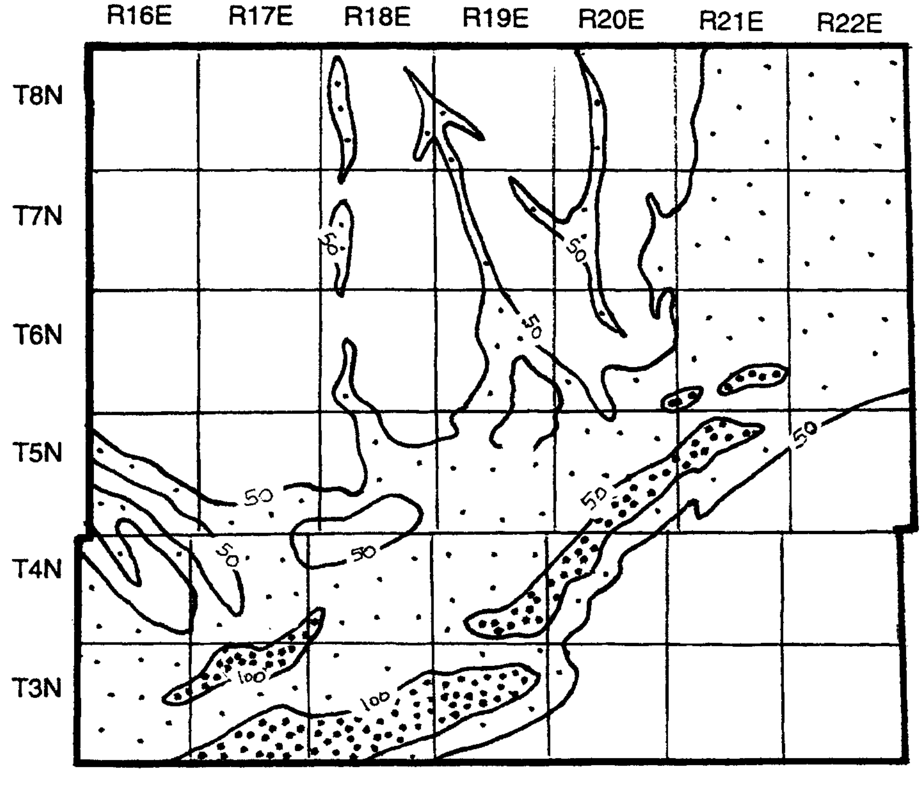


Figure 7. Isolith map of the Spiro sandstone (After Hess, 1995)

STRUCTURAL FRAMEWORK

Overview

The Arkoma basin is a peripheral foreland basin associated with the Ouachita Fold Belt of southern Arkansas and southwestern Oklahoma. The basin, as a structural feature, extends from the Gulf Coastal Plain of central Arkansas westward to its end against the Arbuckle Mountains of south-central Oklahoma (Figure 1). The Arkoma basin is bounded to north by the Ozark Uplift of northern Arkansas and the eastern Shelf of north eastern Oklahoma. The frontal zone of the Ouachita mountains bounds the basin to the south.

Formation of Arkoma Basin and Ouachita Fold Belt

The evolution of the Arkoma basin and the Ouachita orogenic belt reflects opening and subsequent closing of a Paleozoic ocean basin (Houseknecht and Kaneca, 1983). Since the late 1970s various tectonic models have been proposed to explain the geological history of the Ouachita orogenic belt and Arkoma basin, an associated foreland basin. One of the widely accepted tectonic models is a collision model with south-dipping subduction.

Arkoma basin was depositional part of a broad, stable shelf along a passive continental margin during much of its history (Sutherland et. al. 1989). There had been a mainly deposition of carbonate strata until early Atokan time. The complete Atokan strata of the Arkoma basin and the Ouachita fold belt display the history of gradual collapse and the development of this foreland basin as a result of continental collision (Sutherland, 1988).

Houseknecht and Matthews (1985) proposed a model showing five (5) hypothetical cross-sections depicting tectonic evolution of souther margin of North America Continent (Figure 8). According to the model, a major episode of rifting resulted in the opening of an ocean basin during the latest Precambrian or earliest Paleozoic (Figure 8A). As a consequence of this rift opening, the southern margin of North America evolved into an Atlantic-type margin that persisted through the middle Paleozoic (Figure 8B). In Cambrian to early Mississippian time a thick sequence of shallow-water carbonates deposited on the shelf, and thinner, and deep water black shales and cherts accumulated in the basin (Sutherland et. al. 1989). During the early Mississippian, the ocean basin began to close beneath continental plate (South America Continent) (Figure 8C). Even though it is impossible to determine accurately when the subduction began, it was clearly under its way during the Mississippian, as suggested by detritus indicative of an orogenic provenance (Morris, 1974) and locally abundant volcanic detritus in the Stanley formation of Ouachitas (Houseknecht, 1989). Within this convergent tectonic setting, the incipient Ouchita Orogenic belt began to form as an accretionary prism associated with the subduction zone (Houseknecht, 1986) (Figure 8C). Throughout the Mississippian and into the earliest Atokan (Spiro sandstone), the shelf along the southern margin of North America remained a site of relatively slow sedimentation in shallow marine through non-marine environment (Figure 8C). However, the deep remnant ocean basin (Dickinson, 1974) become the site of rapid

Hypothetical cross sections depicting tectonic evolution of southern margin of North America (From Housecknect, 1986).

- (A) Late Pre-Cambrien-earliest Cambrien,
- (B) Late Cambrien-earliest Mississippian,
- (C) Early Mississippian-earliest Atokan,
- (D) Early-middle Atokan,
- (E) Late-Atokan-Desmoinesian.

**** **Keys to patterns**:

* Crosses=continental crust,
* Black=oceanic crust,
* Heavy dots=basal Paleozoic strata,
* Horizantal hachures = upper Cambrien-basal Mississippian strata,
* White = Mississippian-basal Atokan strata,
* Sand stippling = lower-middle Atokan strata,
* Vertical lines = upper Atokan-Desmoinesian strata,
* Mottled = Ouachita accretionary prism,
* Wavy lines = Ouachita foreland thrust belt,
* Black triangle = magmatic arc volcanoes.

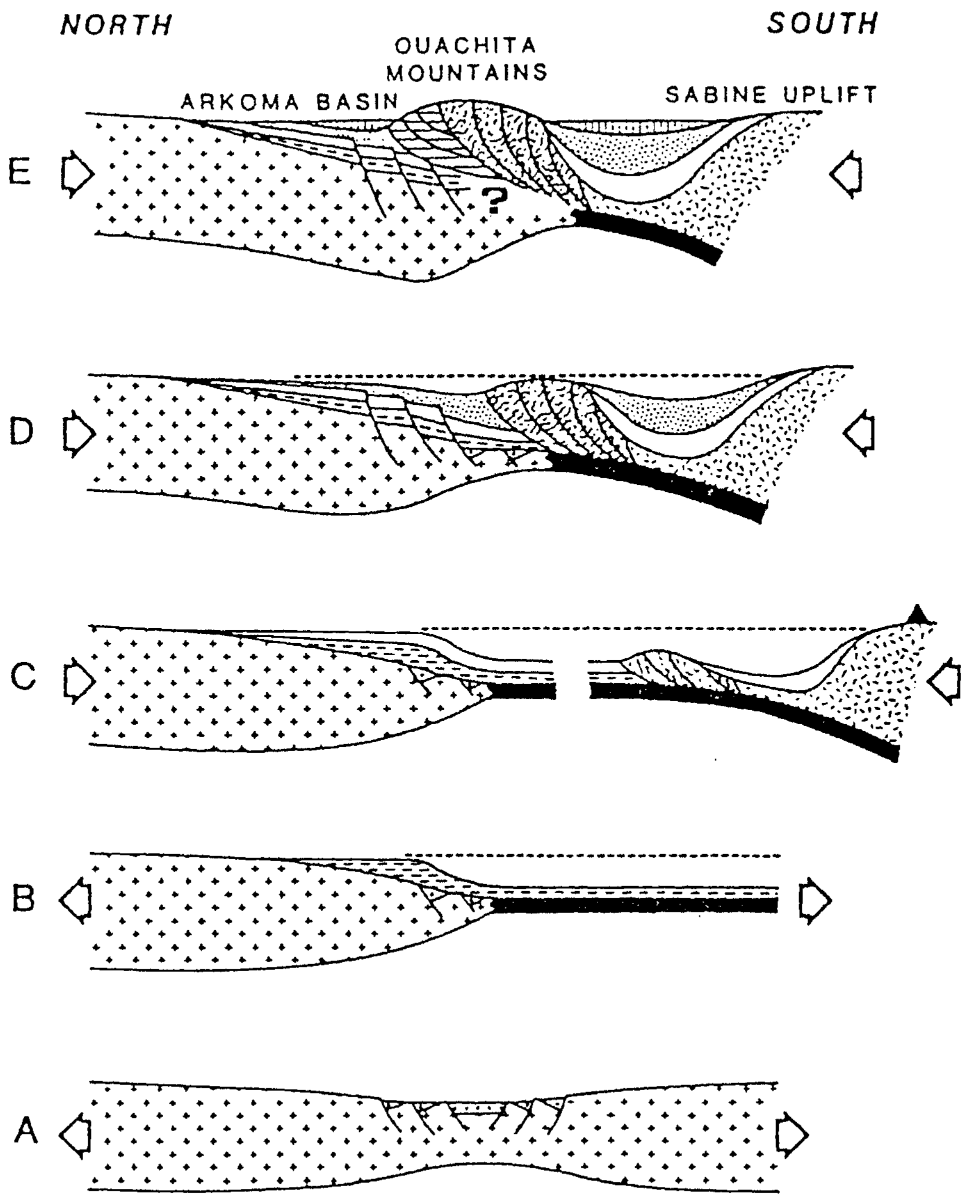


Figure 8: Hypothetical cross sections depicting tectonic evolution of southern margin of North America (From Houseknect, 1986).

deposition of flysch. As a result, more than 5 km. of flysch was deposited during Mississippian and Morrowan time (Stanley, Jackfork and Johns Valley formations) (Figure 10).

By early Atokan time, the remnant ocean basin had been consumed by subduction and northward advancing subduction complex was being abducted onto rifted continental margin of North America (Figure 8D). Partly as a result of attenuated continental crust being drawn into the subduction zone and partly because of vertical loading by the overriding accretionary prism (Dickinson, 1974, 1976), the southern margin of the North American continental crust was subjected to flexural bending. Apparently, as a result of this flexural bending, widespread normal faulting occurred in the foreland (Houseknecht, 1986).

The normal faults generally strike parallel to the Ouachita trend, are mostly downthrown to the south and offset both crystalline basement and overlying Cambrian to basal Atokan strata of rifted margin prism (Houseknecht, 1986). Normal faults broke the shelf-slope-rise geometry that had prevailed along the passive continental margin since the early Paleozoic. Also, these normal faults greatly increased the rate of sedimentation and subsidence (Figure 11). Fault movement was contemporaneous with deposition of lower to middle Atokan shales and sandstones, resulting in abrupt thickness across faults (Figure 8D). These Atokan strata represent a critical transition between passive margin sedimentation and foreland basin sedimentation (Houseknecht, 1986). By late Atokan time (Figure 8E), foreland-style thrusting became predominant as the subduction complex pushed northwards against strata deposited in settings illustrated by Figure 8 C,D.

Resultant uplift along the frontal thrust belt of the Ouachitas completed the formation of the peripheral foreland basin (Dickinson, 1974) in which shallow marine, deltaic, and fluvial sedimentation occurred, with sediment dispersal patterns following those established during the Atokan. Upper Atokan and Desmoinnesian strata of the Arkoma basin contains a typical coal-bearing molasse (Houseknecht, 1986).

Local Features:

Discussion of structures within the area of investigation is divided into two categories; extensional (Normal growth faults) and compressional (folds). Figure 9 illustrates the both types of structural features in the study area.

a. Syndepositional Normal Faults:

As a result of flexural bendind during the early to middle Atokan, widespread normal faulting occured in the southern margin of the North American continental crust. These normal faults are mostly downthrown to the south, and offset both cristalline basement and overlying Cambrien through basal Atokan (Houseknecht, 1987).

From a depositional edge along the northern margin of the basin, Atokan strata thicken to a maximum of more than 18,000 feet along the northern edge of the Ouachita frontal thrust belt (Figure 10). Most of the increase in thickness is accommodated by abrupt

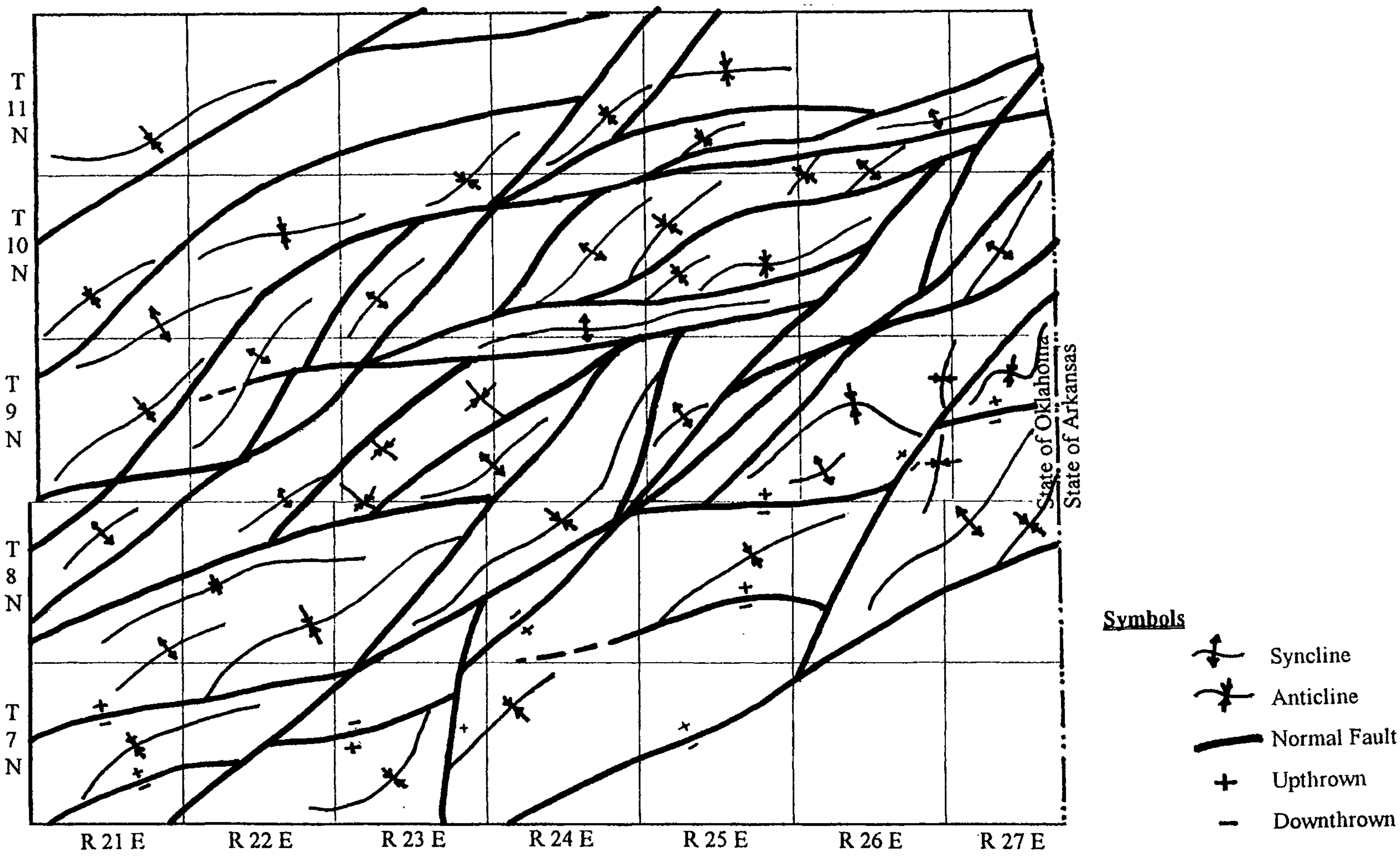


Figure 9: Location of Normal Faults and Folds in the Area of Study

expansion of the lower through middle part of the Atoka formation across syndepositional normal faults, the simplefied location of which are shown in Plate B..

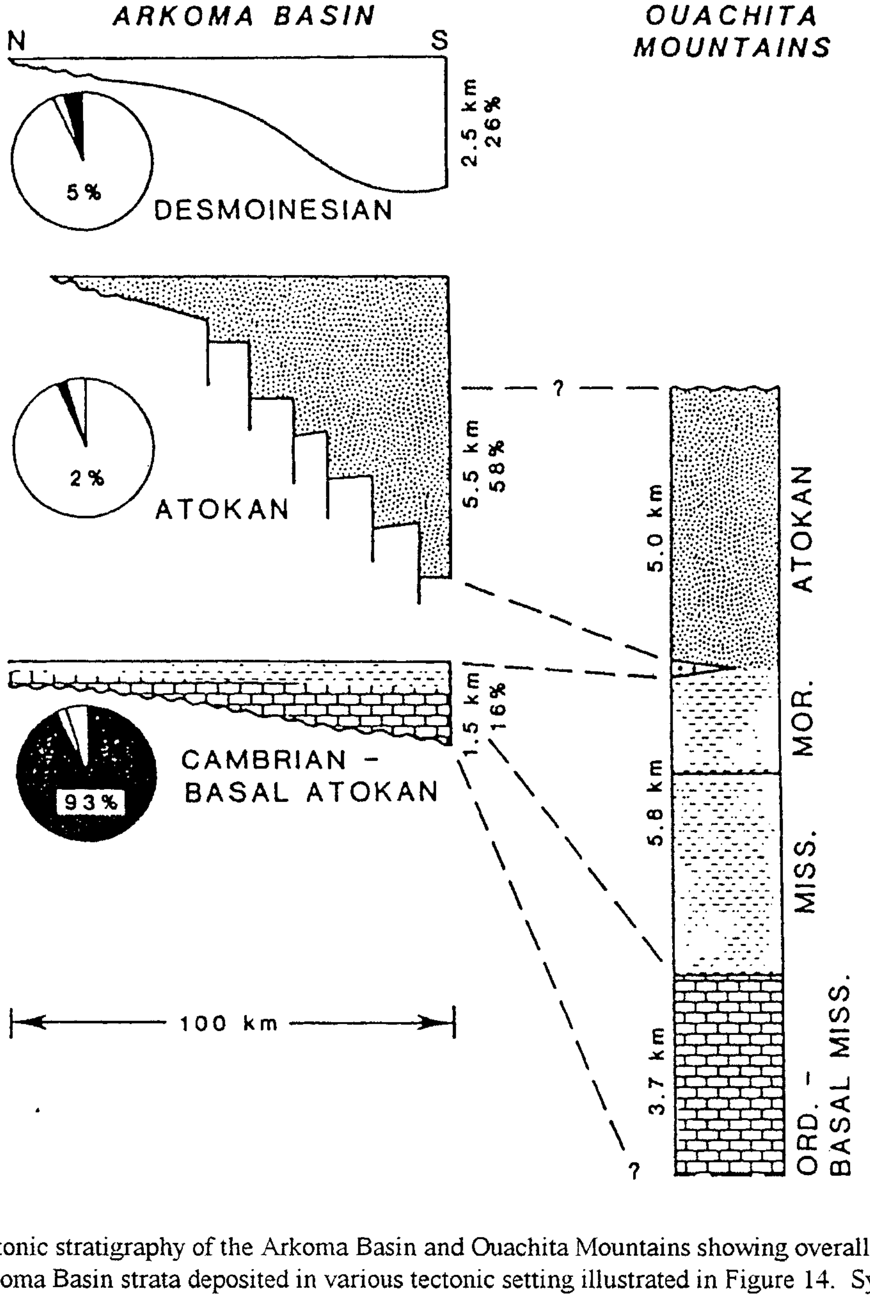


Figure 10: Tectonic stratigraphy of the Arkoma Basin and Ouachita Mountains showing overall geometry of Arkoma Basin strata deposited in various tectonic setting illustrated in Figure 14. Symbols denote genetically related strata and do not imply lithologies. Pie diagrams illustrate the amount of time represented by each genetic package of strata, expressed as a percentage of total Paleozoic time during which sedimentation occured.(Houseknect, 1987).

The first major tectonic event in the study area occurred after deposition of the Spiro sandstone, possibly during early Atokan time and resulted in extensional down to-the-south block faulting. During Middle Atokan time, extensional faulting continued and was accompanied by a rapid influx of clastic sediments (Koinmand and Dickey, 1967) resulting in a zone of growth fault with southwest-northeast extension covering the complete area of the study.

A structural map of the study area contoured on the top of the Wapanucka limestone indicates these extensive normal faults (Figure 9). The fact that the thickness of Atokan strata is significantly greater on the southern block of these faults indicates that movement of the faults is contemporaneous with deposition (Figure 11).

The San Bois fault zone, along the southern edge of the field, has down to-the-basin faults with as much as 6,000 feet of vertical separation. The Atoka formation almost doubles in thickness at the south of this significant structural boundary (Northcut and Brown, 1993).

Folds:

In the Study area, folds are oriented southwest-northwest. The axis of folds almost parallel those of faults in the area. Folds in the study area are considered to be produced by two different mechanisms;

- Extensional forces (syndepositional forces):

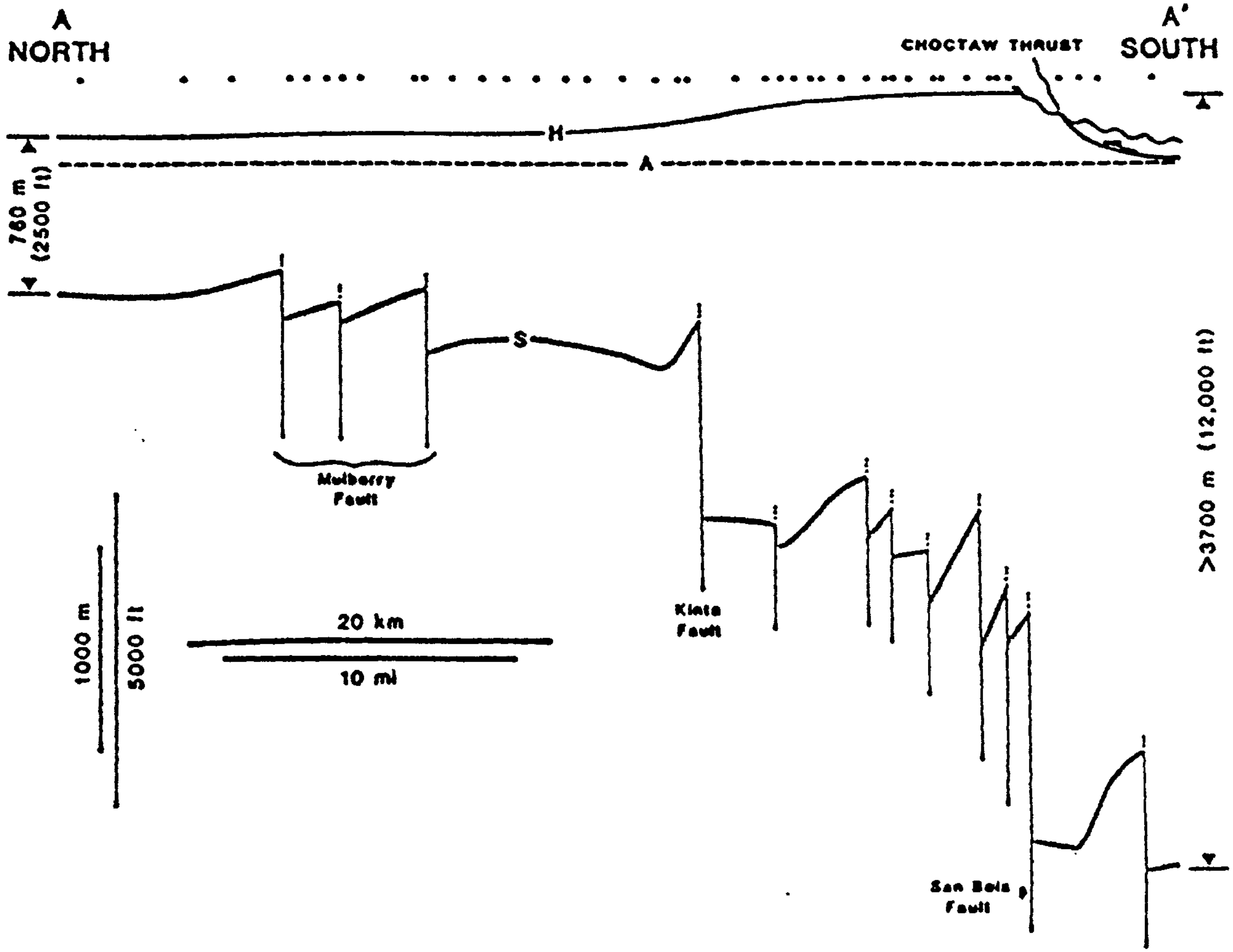


Figure 11: Thickening of the Middle Atokan as a Result of Syndepositional Faulting. The Spiro Sandstone (S) and Upper Atokan bed (A) mark the boundaries of the Middle Atokan (From Houseknecht, 1983).

During the depositional of Atokan formation, rollover anticlines were produced along the southern side of normal growth faults. These folds are considered to be syndepositional features formed during Early Atokan-Desmoisenian time.

- Compressional forces (post-depositional):

By late Atokan time (Figure 8E, D), foreland-style thrusting became significant as the subduction complex pushed northwards against strata deposited in setting illustrated by Figure 8(C,D). At that time, the gross structural configuration of the Arkoma-Ouachita system was essentially the same as at present, although relatively minor folding and thrusting continued after Desmoinesian (Houseknecht, 1986). As a result of this compressional forces, southwest-northeast oriented folds, which host the natural gas in the area, were formed.

STRATIGRPAHY OF ARKOMA BASIN

Figure 12 schematically represents the Paleozoic stratigraphy of the Arkoma basin and Ouachita Mountains. However, stratigraphic numeculature for the basin is completely different in northeastern Oklahoma than those in northwestern Arkansas. Also, it differs from surface to subsurface (McGee, 1971; Figure 13). The Arkoma basin of Oklahoma contains a great thickness of sedimentary rocks whose ages range from upper Cambrien time to upper Pennsylvanian time. Although throughout the Mississippian and into earliest Atokan the shelf along the southern margin of the North America remained a site of relatively slow sedimentation in shallow marine through non-marine environments, during the Atokan time (while the deep trench and remnant ocean basin (Dickinson, 1974) became the site of rapid deposition of flysh) a great thickness of clastic sediments deposited in the Arkoma Basin of Oklahoma (Housekneht, 1987) (Figure 14).

In the Arkoma basin of Oklahoma, the upper Mississippian strata are represented by the Chesterian Series. The lower and middle Pennsylvanian rocks in the Arkoma basin are subdivided into the Morrowan, Atoka, and Desmoines Series (Table 15). This study is of upper Morrowan and basal Atokan age sediments but a brief overview of Pennsylvanian age rocks will be given.

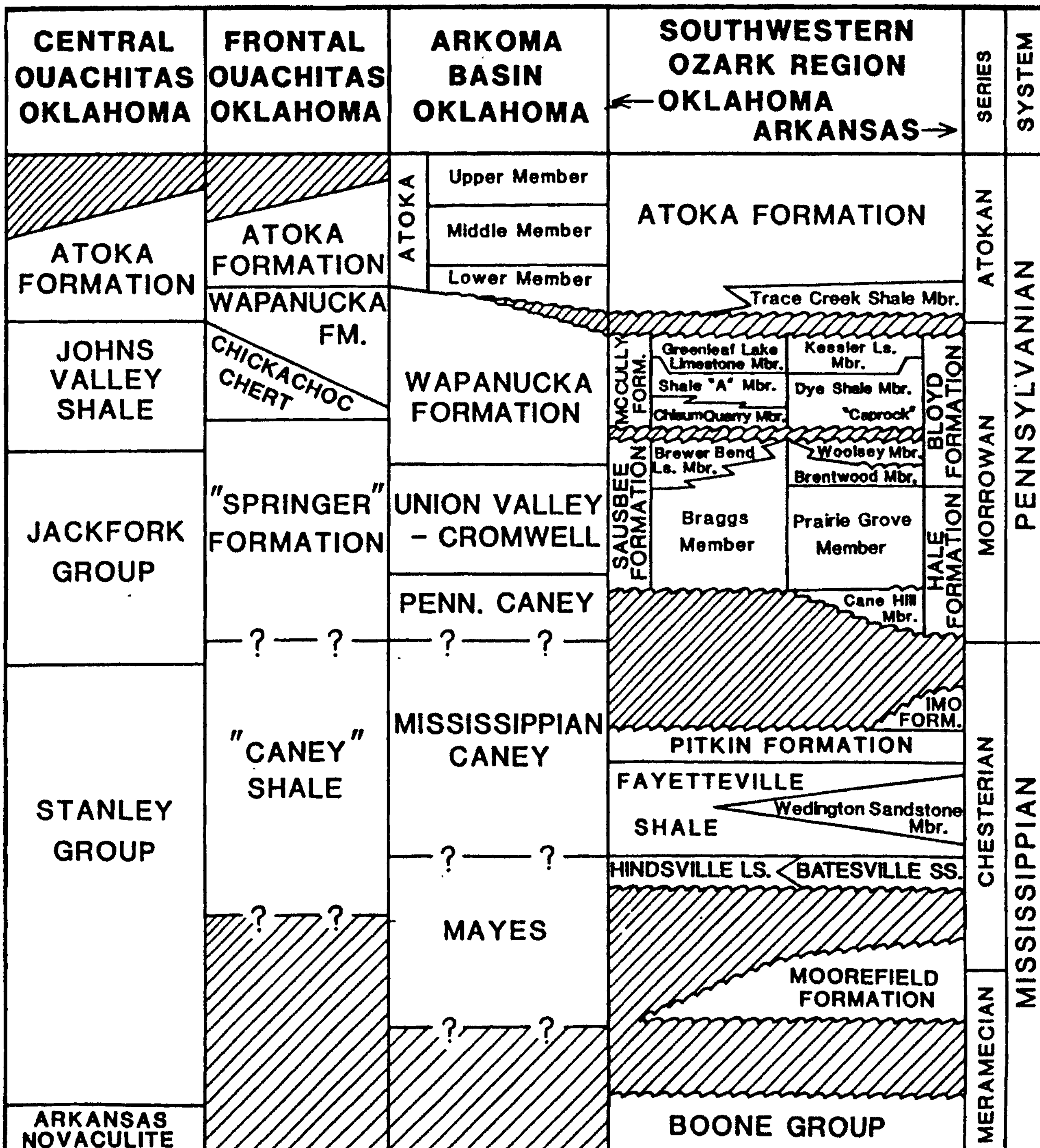


Figure 12: Stratigraphic Column of The Arkoma Basin (From Sutherland, 1988).

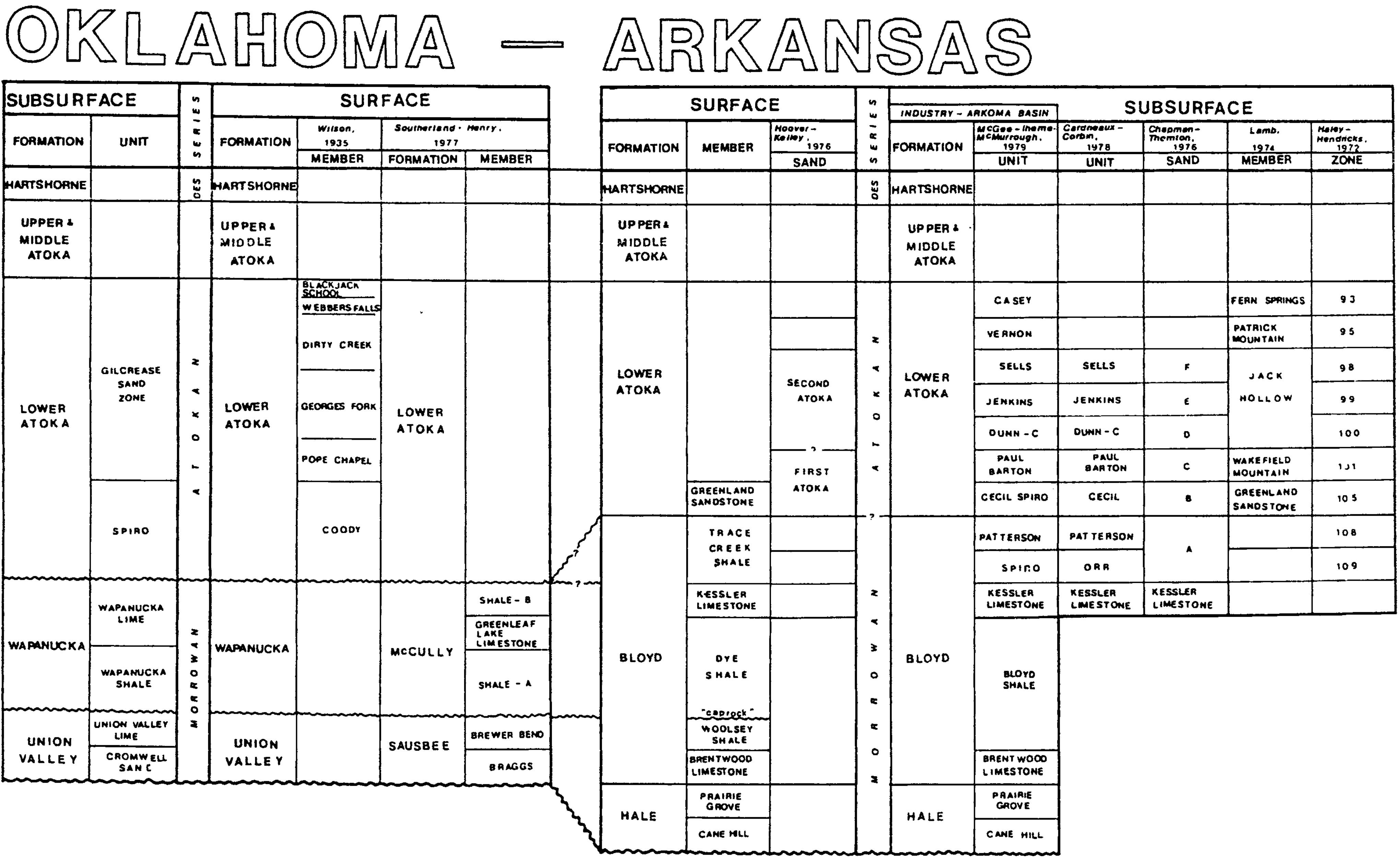


Figure 13. Surface and subsurface stratigraphic nomenclature chart for northeastern Oklahoma and northwestern Arkansas (From McGee, 1971).

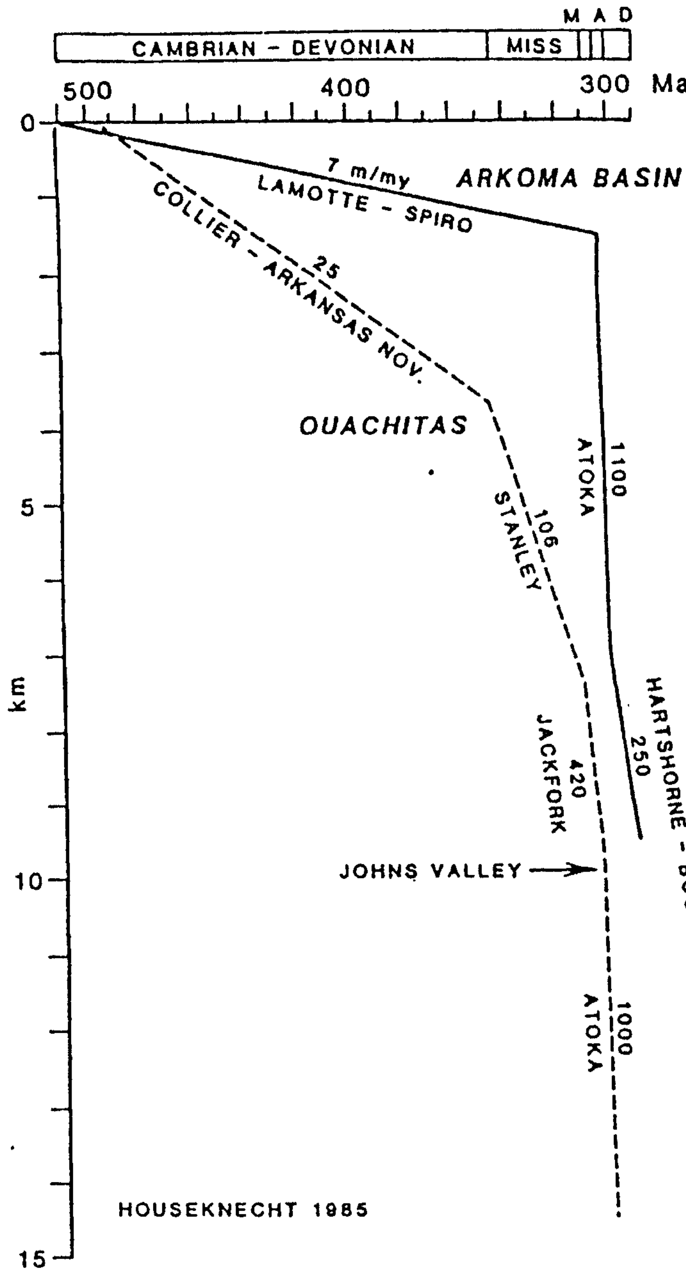


Figure 14: Sediment accumulation curves for Arkoma Basin and Ouachita strata (Not corrected for compaction). Major break in slope on each curve marks change in tectonic setting and nature of sediment. Older strata contain slightly feldispathic, quartzose sandstones and oil prone organic material; younger strata contain lithic sandstones and gas prone organic material (Houseknecht, 1985).

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.

Figure 15

Morrowan Series

The Morrowan series (Figure..) consists of the Wapanucka and Union Valley Formations and Springer formation (Figure 13 and 15). The Wapanucka Formation consists of (1) upper sandstone/limestone unit, (2) middle shale unit, (3) Lower limestone or Chickachoc chert unit. The union Valley Formation is composed of upper Union Valley lime member and lower Cromwell sandstone. Wapanucka formation is included in this study. The thickness of Morrowan Succession ranges to several hundred feet and is essentially constant throughout the basin.

Morrowan strata in the Arkoma basin of Oklahoma are composed of limestone, shale, and sandstone units that accumulated in various shallow-shelf environments, adjacent to the deep Ouachita trough immediately to the south (Zachry and Sutherland, 1984). The absence of Morrowan rocks from northeastern Oklahoma is partially the result of post-Morrowan, pre-Atokan erosion. The apparent conformable boundary relationship between Morrowan and Atokan strata in the central part of the Arkoma basin becomes unconformable towards the northern basin margin; there Atokan and Desmonian rocks unconformable overlie the Wapanucka limestone, the underlying Wapanucka shale, and the basal Union Valley-Cromwell sandstone. To the north, the Morrowan strata progressively disappear shelfward by truncation beneath the overlying Atokan and Desmoinesian rocks (Sutherland, 1986). In the early Morrowan, the sea transgressed north from the ouachita trough onto the Arkoma basin, across the truncated Chesterian surface (Sutherland, 1988; Figure 16). Depositional patterns were complex on the Arkoma shelf in the middle Morrowan time (Figure 17). There was a gradual regression, causing emergence and nonmarine environmental conditions in northwestern Arkansas during the middle Morrowan time (Sutherland, 1988; Figure 17).

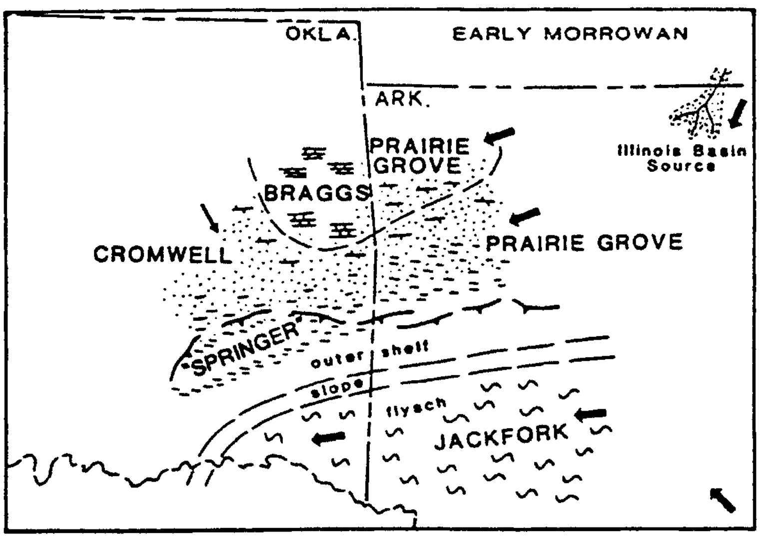


Figure 16. Early Morrowan paleogeographic map (From Sutherland, 1988).

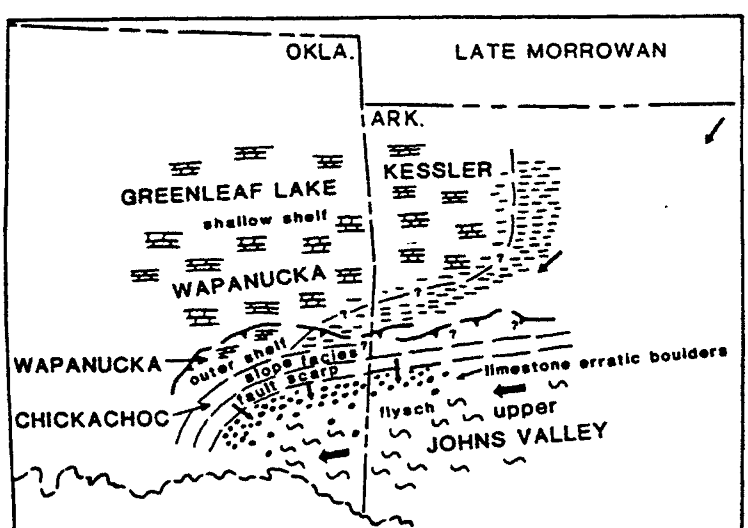


Figure 17. Middle Morrowan paloegeographic map (From Sutherland, 1988)

Late Morrowan:

Late Morrowan of Arkoma basin in Oklahoma is composed of Wapanucka limestone and the overlying Trace Creek shale member. The Trace Creek Shale is only present in the Ozark Region. It is not present in the study area.

The wapanucka limestone Formation represents the youngest massive limestone in the Paleozoic strata of the Arkoma basin of Oklahoma. The isopach (Plate IV) map indicates that the Wapanucka formation generally thickens to the south, with thickness ranging from maximum of 317 to minimum of 65 feet in the Study area. The total thickness of Wapanucka limestone in the study area, ranging from 200 to 700 feet throughout the Basin, gradually increases southward. The ratio of Limestone/shale decrease southward that probable indicate a transition from shelf to basinward. There is not any published studyof Wapanucka Formation in the study area.

The formation is a subsurface unit in the study area. The Wapanucka formation is exposed in three geographical separate area in Oklahoma. These areas are the frontal Ouachita Mountains, the northeastern flank of the Arbuckle Mountains, and the Mill Creek syncline in the central Arbuckle Mountains. However, there are several outcrops exposed in region of the Ouachita Mountains (Grayson, 1984). Grayson(1979) described the Wapanucka Formation in detail throughout the frontal belt. He distinguished four informal members (Figure 18): a lower limestone member, a middle shale member, an upper sandstone-limestone member, and the Chickachoc chert member. The Chickachoc Chert member is the basinward (or an outer-shelf) facies time equivalent of the lower limestone member.

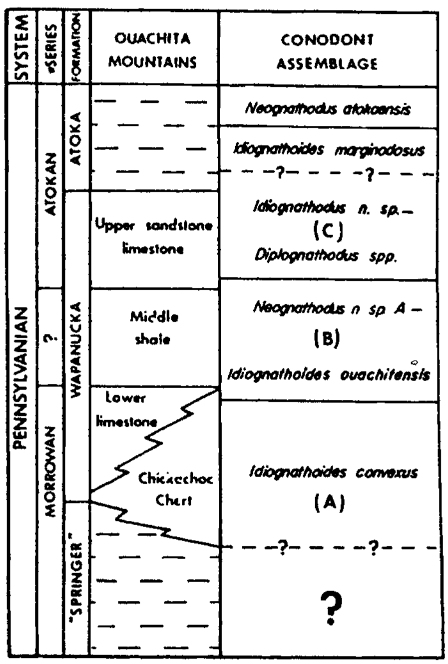


Figure 18: Lithostratigraphic and biostratigraphic subdivision of Wapanucka Formation, Frontal Ouachita Mountains, Oklahoma (Grayson, 1984).

The Wapanucka Formation consists of a diversity of lithofacies that are commonly repetitive both laterally and vertically (Grayson, 1984). Ferguson (1988) emphasized that basinward facies changes of the Wapanucka are found in fault blocks progressively to the south. In other words, the Wapanucka Formation represents shelf to slope facies (the Chickachoc Chert), basinward (Southward) (Figure 19).

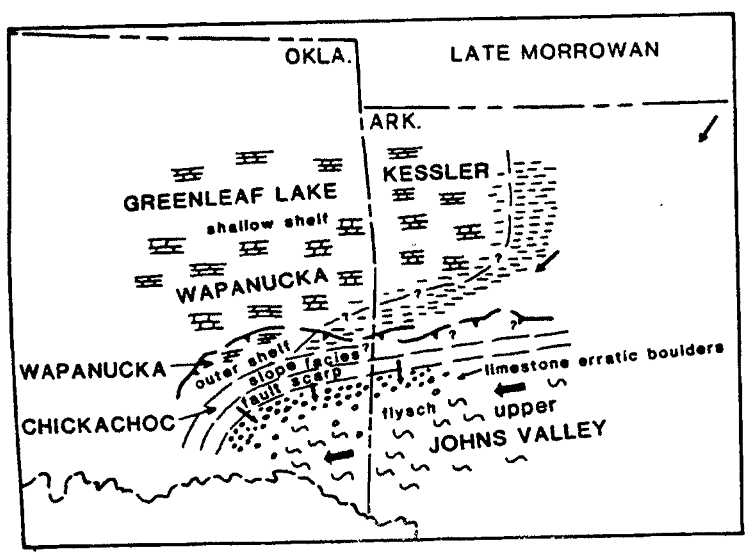


Figure 19. Late Morrowan paloegeographic map (From Sutherland, 1988)

The most recognizable basinward changes are a dramatic increase in thickness of the middle shale member, and a transition from spicular limestone to spiculite in the lower two members (Ferguson, 1988). The Wapanucka Formation represent mostly nearer shore and shallower water deposits (Grayson, 1984).

Atoka Strata

The name Atoka (Figure 21) was first proposed by Taff and Adams (1900) as"beds of sandstones in alternating strata occurring in regular succession below the Hartshorne Sandstone Formation (Desmonian in age) and above the Wapanucka limestone Formation (Morrowan in age), to a thickness of nearly 7,000 feet near Atoka, Oklahoma. The Atoka formation is a complex succession of sandstone, shales, and siltstones with sporatic occurrences of discontinuous limestones and thin-bedded coal in its upper half that accumulated prior to and during the development of a peripheral basin associated with the Ouachita Fold Belt.

The distribution and character of the various detrital rocks were controlled by the volume of sediments supplied to the region, the geographic position where sediment entered the area, and the tectonic activity that accompanied sedimentation (Zachry, and Sutherland, 1984).

The number of individual sandstone units increases toward the east in the Arkoma basin (Branan, 1966). The abundant sandstone units to the north and east become thin, fine-grained and less abundant toward the south and southwest, where the predominance of marine shales constitute the greatest volume of Atoka strata that accumulated in the subsiding Arkoma basin during early and middle Pennsylvanian time. Shelton (1974) describes the southward-thickening of Atoka succession to being deposited contemporaneously with active faulting within a prograding delta-marine environment, where the repetitious interbedded sand and shale sequence is indicative of a distal delta-fringe environment. The cyclicity of Atoka strata is a reflection of reoccurring transgressive, and more predominant, regressive environments (Glick, 1975). Shelton (1974) suggested a north cratonic source for Atokan strata on the shelf and in the basin. The Atoka strata increases in thickness from a featheredge to a maximum of 10,000 feet in eastern Oklahoma (Figure 20), and a maximum of 20,000 feet in western Arkansas (Branan, 1960). Suneson and Hemish (1984) devided the Atokan Formation of the Arkoma basin of Oklahoma into lower, middle, and upper intervals (Figure 21).

The lower unit ranges in thickness from 900 feet adjacent to the northern margin of the basin, to approximately 2,000 feet near the southern margin (Sutherland, 1984). In Arkansas and extreme eastern Oklahoma, the interval is composed of multiple sandstone units separated by shale units. Individual sand stone units range from 20 to 200 feet in thickness, and are continuous throughout the northern and central part of the basin (Zachry and Sutherland, 1984). A general increase in unit thickness occurs to the east in Arkansas, and shale intervals are thinner. Sandstone of the east in the southern part of the basin and except for basal unit, pass into shale. The lower Atoka of Oklahoma is characterized by a widespread basal sand, the spiro Sandstone, overlain by shale.

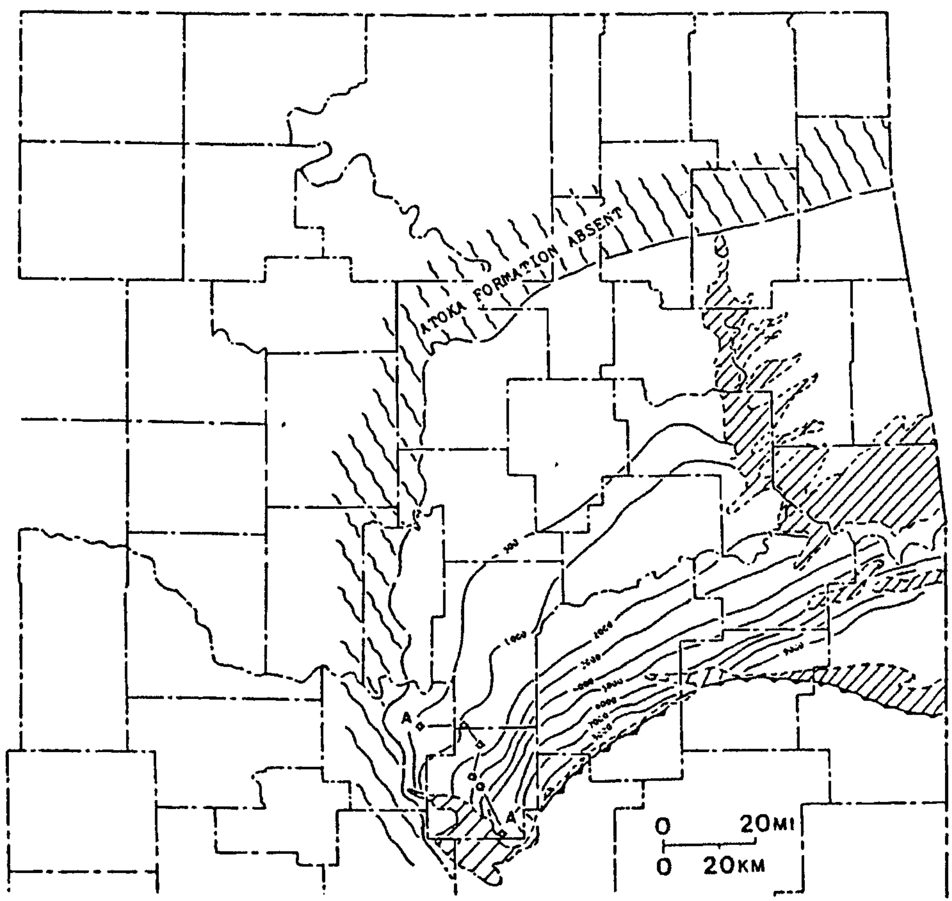


Figure 20. Thickness map of Atoka Formation in Eastern Oklahoma. Thickness contours (in feet)from Frezon and Dixon (1976). Atoka outcrops indicated by diagonal lines.

PENNSYLVANIAN	MIDDLE	DESMOINESIAN	McAlester Formation		(Keota Sandstone) (Tamaha Sandstone) (Cameron Sandstone) Booch (Warner) Sandstone
			Hartshorne Formation		Hartshorne Sandstone
		ATOKAN	Atoka Formation	upper	(Webbers Falls Sandstone) Gilcrease Sandstone Fanshawe Sandstone
				middle	Red Oak Sandstone Panola Sandstone Diamond Sandstone Brazil Sandstone Bullard Sandstone Cecil Sandstone Shay Sandstone
				lower	Spiro Sandstone Foster Sandstone
	LOWER	MORROWAN			Wapanucka Limestone

Figure 21. Stratigraphy of the Atokan and Desmoinesian in the Arkoma basin of Oklahoma (From Suneson and Hemish, eds., 1984).

The multiple sand succession of lower Atoka, above Spiro, of Arkansas is absent in Oklahoma. Pre-Spiro channel systems (Foster sandstone channels) are cut into the underlying Wapanucka limestone and contains elongate sandstone bodies that trend southeastward across the northern and central parts of the basin (Lumsden and others, 1971) (Figure 4). These elongate sand bodies are vertically continuous with the overlying Spiro blanket sand. In the absence of correlateable sand units above the Spiro in Oklahoma, the upper boundary of the lower Atoka interval occurs within a shale succession and is poorly defined in Oklahoma (Zachry and Sutherland, 1984).

The middle Atoka unit of Arkansas and Oklahoma is confined to the southern part of the basin. The middle Atoka interval in Oklahoma is composed dominantly of shale, with few major sandstone units. It consists of laterally discontinuous sandstone units, ranging to 100 feet in thickness, that are separated by thick intervals of shale in western Arkansas and extreme eastern Oklahoma (Zachry and Sutherland, 1984). The unit thickens from approximately 1,200 feet in T.7N. to more than 10,000 feet in T.4N. (Buchanan and Johnson, 1968). Increment of thickening occur successively on the downthrown sides of large, east- and northeast- trending normal faults, in a stepwise manner from north to south (Zachry and Sutherland, 1984; Figure 23). The relationship of the faults to southward thickening within the terrigenous clastic deposit indicates that faulting was contemporaneous with sedimentation. Buchanan and Jonhson (1968) suggested that faulting began in the south, focusing the site of earliest Middle Atoka deposition. Later development of east-trending faults successively to the north controlled the accumulation of younger sediments within the middle Atoka wedge. Middle Atoka strata are very thin to the north of T.7 N. in Arkansas (Zachry and Sutherland, 1984). A major sand unit within this interval, the Red Oak sandstone, is a significant producer of natural gas. The red Oak is confined to the south side of the San Bois Fault. Vedros and Fisher (1979) suggested, from an analysis of sedimentary structures and sand unit geometry, that the Red Oak accumulated in a submarine fan environment supplied with sediment from submarine canyons cut into the scarp of an active growth fault to north. the Fanshawe sandstone, a unit that occur higher in the middle Atoka interval, was also interpreted a submarine-fan deposits.

The upper Atoka strata in the Akoma basin of Oklahoma are composed of shale units with thin, discontinuous sands. The unit displays a pattern of gradual thickening to the south, and is present from the northern margin of the basin southward (Zachry and Sutherland, 1984). Large normal faults that cut middle Atoka interval are buried by the upper Atoka rocks, indicating a cessation of growth fault activity (Buchanan and Johnson, 1968).

Depositional history

Lower Atoka strata in the Arkoma basin accumulated in a shallow-shelf setting not unlike the setting of earlier Morrowan sedimentation. Deposition was continuous from Morrowan to Atokan time on the southern part of the shelf in Oklahoma and probably in Arkansas (Zachry and Sutherland, 1984). Widespread Atoka sedimentation was initiated northward transgression of Atokan seas across an erosion surface cut on Morrowan strata

(Zachry and Sutherland, 1984). According to Sutherland (1989), the basal Spiro Sandstone (including the Foster sands) accumulated in pre-transgression channel systems and in widespread coastal sand complexes to form a blanket-sand unit at the base of the formation (Figure 22).

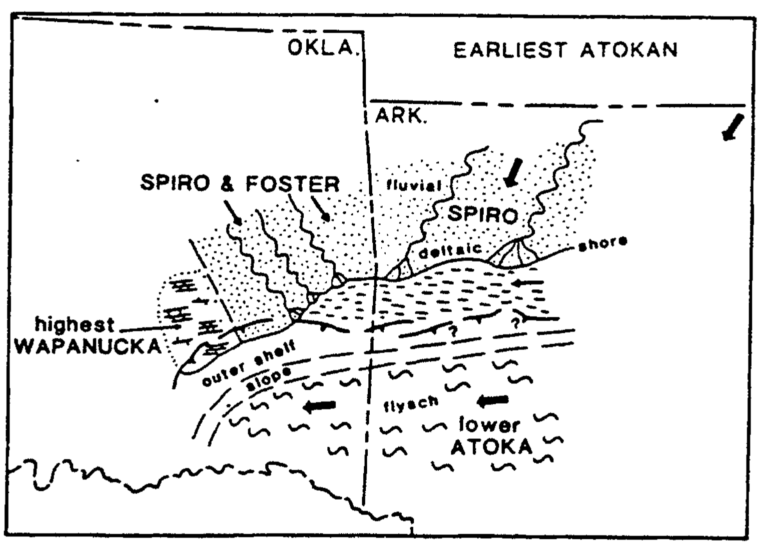


Figure 22. Earliest Atokan paleogeographic map (From Sutherland, 1988).

Multiple sand units above the basal sand thicken eastward in Arkansas and are absent in Oklahoma, suggesting that the dominant sediment supply entered the basin from the northeast and north in Arkansas (Sutherland, 1988). High, destructive cratonic deltas that prograded southward across a shelf characterized by little subsidence formed widespread but thin sand units in delta and delta-related environments. Progradation was interrupted by periodic northward transgression, bringing open-shelf environments to the Arkoma Basin area and forming shale units (Zahry and Sutherland, 1984). Distal to the major sediment source, the shelf in Oklahoma was a site of shale accumulation. Thin, calcareous sand units were deposited in shoreface environments in the western part of the area adjacent to the Arbucle Mountains (Zachry and Sutherland, 1984).

During the deposition of the middle Atoka formation (Figure 24), the lower Atoka was subjected to tentional stress, and large, east-and northeast- trending normal faults developed in a stepwise fashion from south to north. Large volumes of sediment by-passed the northern shelf areas and accumulated on the downthrown sides of active faults, forming the middle Atokan clastic wedge (Zachry and Sutherland, 1984).
In Arkansas, proximal to northeastern source, submarine-fan systems delivered sand to deep-water environments. Distal to the source, in Oklahoma the rate of sedimentation were lower, and submarine scarps developed along the faults. Submarine canyons transferred sand, brought westward by littoral drift, to base of the fault scarps, constructing submarine fans (Zachry and Sutherland, 1984).

Upper Atoka strata were deposited after the cessation of normal faulting that produced the great structural relief and thick sediment fill characteristic of the southern

Arkoma basin. Sandstone units within the upper Atoka Formation of Arkansas are related to the progradation of deltaic systems southward across a muddy shelf (Figure 25). Periodic regional transgressions interrupted deltation and spread open-shelf conditions across the basin, forming the shale units. Sediment supply remained high, but subsidence slowed, allowing deltaic sands to be moved laterally into strand-plain and coastal sand environments. This redistribution of sand allowed for the development of sandstone units with great lateral continuity during progradation.

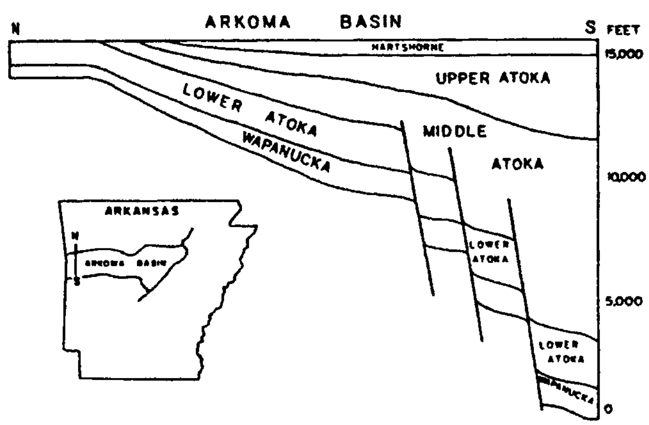


Figure 23. North-south section across Arkoma basin depicting syndepositional faults that controlled sedimentation during deposition of middle Atoka Formation (From Zachry and Sutherland, 1984).

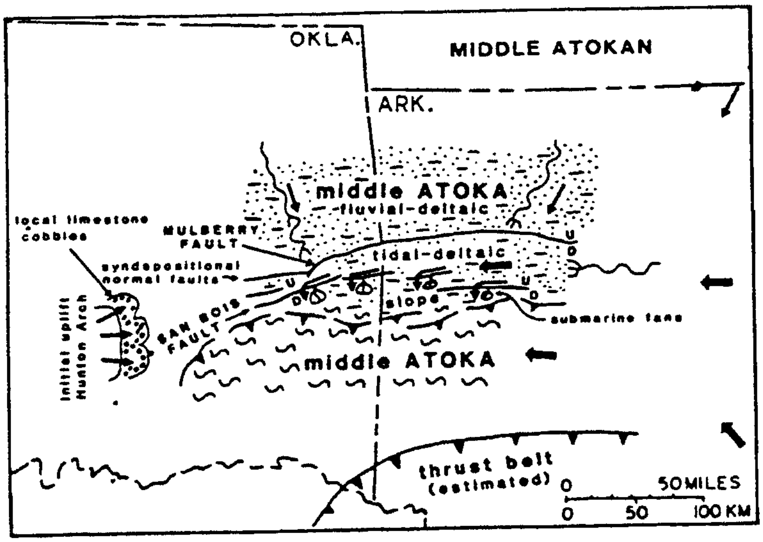


Figure 24. Middle Atokan paleogeographic map (Sutherland, 1988).

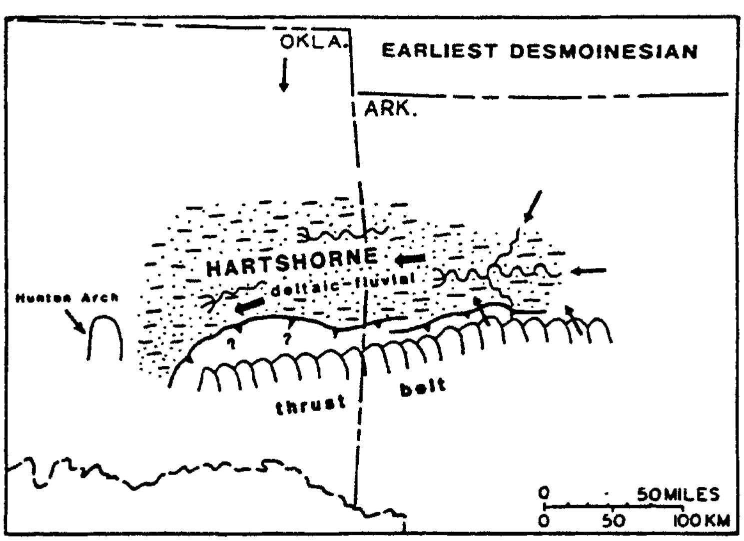


Figure 25. Earliest Desmoinesian paleogeographic map (From Sutherland, 1988).

DELTAIC SYSTEMS

A. Introduction

Since ancient times, river deltas have been of fundamental importance of civilization. Historically, deltas have received great attention as significant agricultural lands. More recently, subsurface deltaic facies have played a significant role in accommodating the world's energy needs; ancient deltaic sediments have provided source beds and reservoirs for a large percentage of the known petroleum reserves.
Deltaic environmental settings normally exhibit many of the characteristics required of a potential hydrocarbon-producing basin-multiple types of reservoir rocks, immediate or nearby source beds, a wide variety of structural and stratigraphic traps, and a buildup of considerable thickness of clastic rocks that are rapidly varied (Coleman et al., 1980). Deltaic depositional facies result from interacting dynamic physical processes (wave energy, tidal action, climate, etc.) which modify and disperse riverborne clastics.

B. Deltaic Setting

The term "delta" was first applied by the Greek historian Heredotus in approximately 450 B.C. to the triangular alluvial deposits at the mount of the Nile River. In broader terms, Deltas are discrete shoreline protuberances formed where rivers enter oceans, semi-enclosed seas, lakes or barrier-sheltered lagoons and supply sediment more rapidly than marine processes can remove it (Elliot, 1978). Deltas have fan-shaped sedimentary bodies in plan, and lenticular in cross-section. The various physical, chemical, and biological processes that control delta development vary appreciably on a global scale, and hence the landforms in deltaic region span nearly the entire spectrum of coastal features and include distributary channels, rivermouth bars, interdistributary bays, tidal flats, tidal ridges, beach dune complexes, swamps, marshes, and evaporate flats (Coleman et al., 1980) (Figure 27). A river system generally consists of four primary components: drainage basin, alluvial valley, delta plain, and receiving basin (Figure 26).

The *drainage basin* basically is the source of the water and sediments, and processes within this component determine the sediment-water supply and the initial size and composition of the sedimentary load. Eventually the distributaries within the drainage basin merge into one or more major channels, and *the alluvial valley* is formed. The alluvial valley is essentially a conduit in which the river flows over and through its own deposits. As a result of migration of alluvial channels, there is often continuos sorting of the sediments suite entering the valley from the drainage basin. At some point along its length the river ceases to function primarily as a transporting agent and becomes a dispersal system. Sediments accumulate and begin the formation of the *delta plain*. This component of a river system results primarily from interaction between riverine and marine processes.

The morphology and geometry of the deltaic deposits reflect the hydraulic regime, sediment load, geologic structure and tectonic stability, climate, tides, winds, waves, water density contrast, coast currents, and the innumerable interactions of all these factors. The characteristics of the receiving basin are among the most important

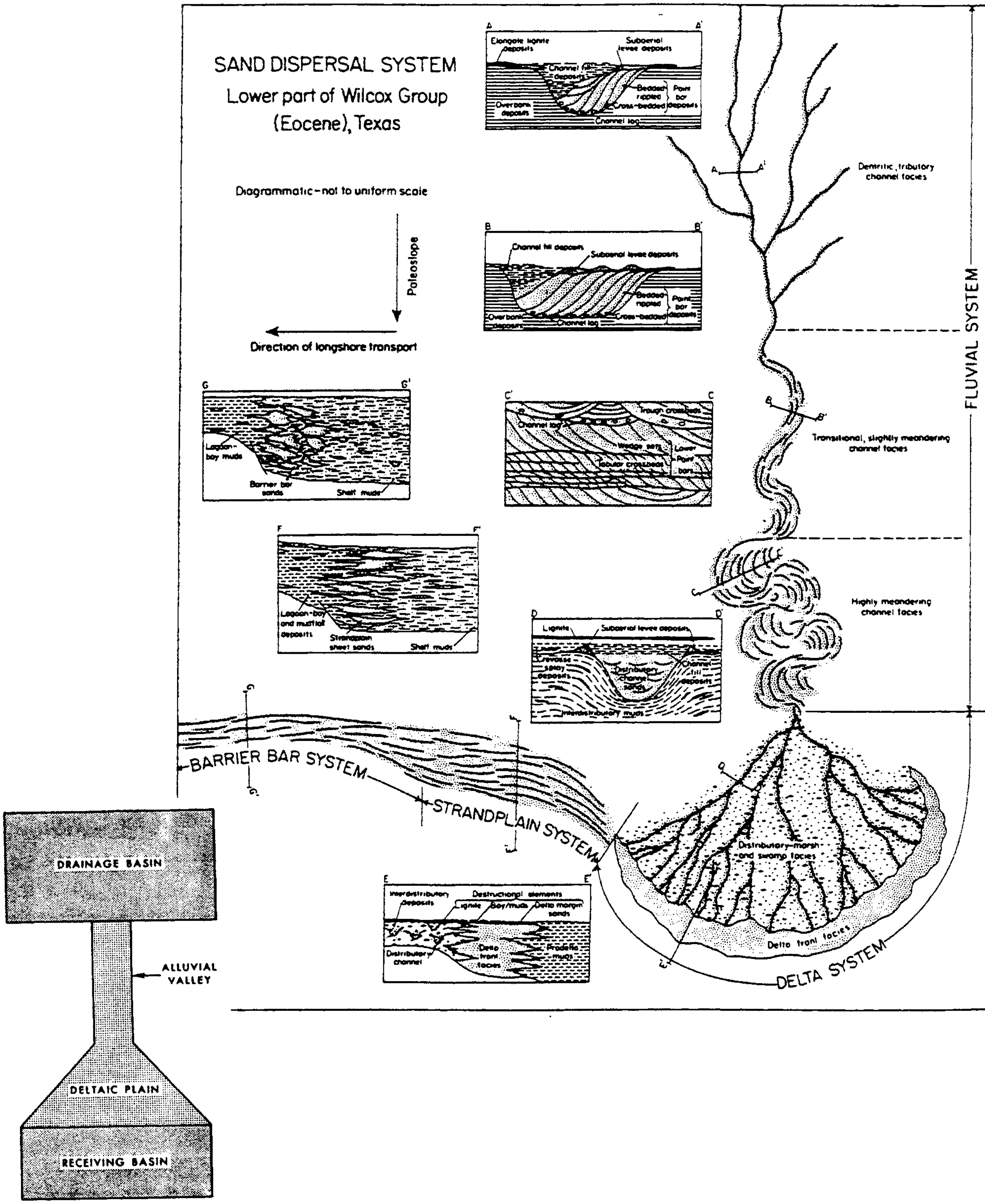


Figure 26. Sand dispersal systems, Lower Wilcox high-constructive delta and associated depositional systems. (From Fisher and McGowen, 1967, 1969).

process controls in the development of a delta, and the various oceanographic nearshore processes on a delta coast are crucial to the shaping and remolding riverborne sediments (Coleman et al, 1980).

DELTAIC	UPPER DELTAIC PLAIN	MEANDER BELTS		CHANNELS
				NATURAL LEVEES
				POINT BARS
		FLOODBASINS		STREAMS, LAKES & SWAMPS
	LOWER DELTAIC PLAIN	DISTRIBUTARY CHANNELS		CHANNELS
				NATURAL LEVEES
		INTER-DISTRIBUTARY AREAS		MARSH, LAKES, TIDAL CHANNELS & TIDAL FLATS
	FRINGE	DELTA FRONT	INNER	RIVER-MOUTH BARS BEACHES & BEACH RIDGES TIDAL FLATS
			OUTER	
	DISTAL			

Figure 27. Transitional deltaic environments. (From Leblanc, 1972).

C. Delta-Forming Processes

A delta forms when a river of sediment-laden fresh water enters a standing body of water, loses its competence to carry sediment, and deposits it. The general form of the deltaic deposit depends upon (Buhattachaeya et al., 1992):

* the density contrast between the river outflow and the standing body of water
* the extend to which the deposits are reworked by marine (Wave and Tide) processes.

The type of jet flow established by a river entering a body of relatively still water depends on the presence or absence of a density contrast between the two water bodies. Three mixing types occur between sediment-laden water and the waters of the receiving basin at the distributary channel mouth (Fisher et al, 1969). Coleman (1980) contrasted stuations in which the river waters were equally dense, more dense and less dense than the basin water (homopycnal, hiperpycnal, and hypopycnal flow; Figure 28). Contrasted stuations in which the river waters were equally dense, more dense and less dense than the basin water (homopycnal, hiperpycnal, and hypopycnal flow; Figure 28).

- Where the flow are ***hyperpycnal(Inertia-dominated)*** and the sediment load is relatively coarse-grained, small, steeply deeding elongate deltas tend to form (Giblet-type mouth bar). Finer-grained material may be deposited further offshore as density underflows. This kind of deposition is common where sediment-laden streams enter fresh waterlakes.

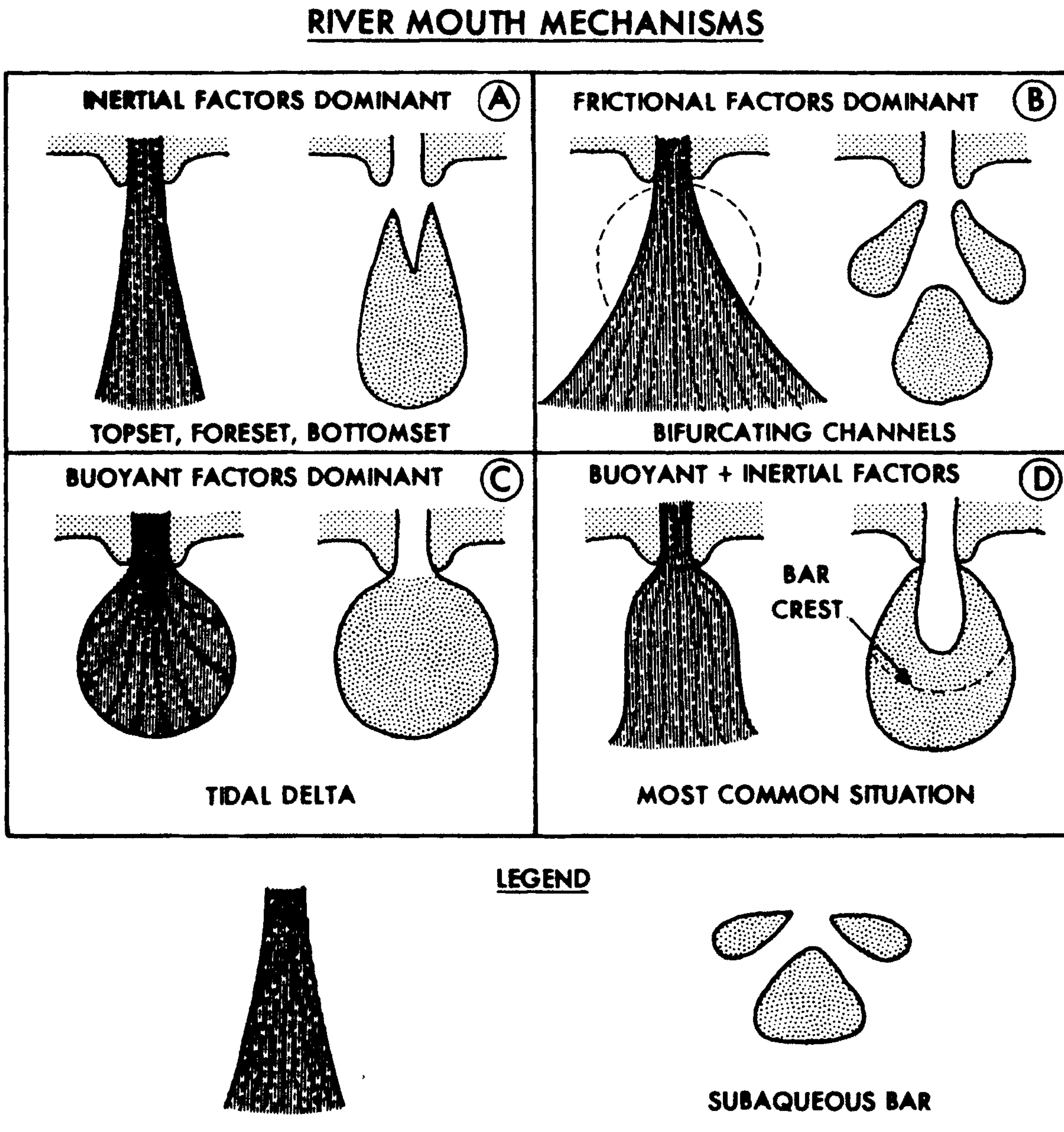


Figure 28. River-mouth mechanism. (From Coleman, 1980).

- In ***homopycnal(friction-dominated)*** settings there may be a greater degree of mixing between the river and the standing body of water. This stuations are common in freshwater deltas, especially where the suspended sediment load is high. In shallow water, friction at the bed causes rapid deceleration and development of a middle-ground bar that tends to cause the associated distributary to bifurcate (Wright, 1977). So-called friction-dominated mount bars also tend to be more fan-shaped and may be dominated by traction current features such as ripples and cross bedding. These types of deltas are commonly characterized by steeply deeping foreset beds, separating less steeply dipping topset and bottomset beds.
- In most cases where a river enters salt water, the density of the river plus sediment load is less than that of the sea water (hypopycnal flow). Under these conditions, ***buoyancy-dominated*** river moths form. The suspended sediment is relatively fine-grained, and tends to be carried farther out into the receiving basin as a buoyancy plume, resulting in lower depositional slopes and more extensive deltaic deposits.

In any given river mount or delta, inertial, frictional, and buoyant forces may be operative in varying portions. Deposition at river-mouth bar usually involves a mix of inertial, frictional, and bouyancy processes. Discharge fluctuations of the rivers are particularly important in this respect. Interacting physical, biological, and chemical processes during the time of deposition exert significant control over the distribution, orientation, and internal geometry of deltaic sand bodies (Coleman et al., 1980). Figure 29 illustrates schematically some of these major process variations that exist within the drainage basin, alluvial valley, delta plain and receiving basin. Factors such as climate, vegetation, and soil and the geology and tectonic history of the receiving basin play an significant role in controlling the size and composition of the sediments that are debounced into the alluvial valley. Factors such as wave energy, tidal processes, and tectonic of the receiving basin play a major role in the development of the sand bodies (Coleman et al., 1980).

Various interacting dynamic processes acting on sediment produce the wide variety of deltaic sand body geometries and distributions in modern deltas. Coleman et al (1980) summarized the major control exerted on deltaic facies by each process.

A. climate:

- Controls sediment-water yield
- Controls in situ delta deposits; tropical: large, thick accumulation of peat; temperature: thin, high, laterally continuous peat layers; arid: complex interfingering supra tidal and evaporite deposits.

B. Water Discharge Regime:

- Erratic discharge regimes produce braided channels displaying wide laterally continuity
- Nonerratic discharge regimes produce more stable meandering channels (shoestring sands)
- Nonnerratic discharge regimes commonly result in numerous complex interfingering finning-upward sequences displaying highly variable porosity-permeability relationships

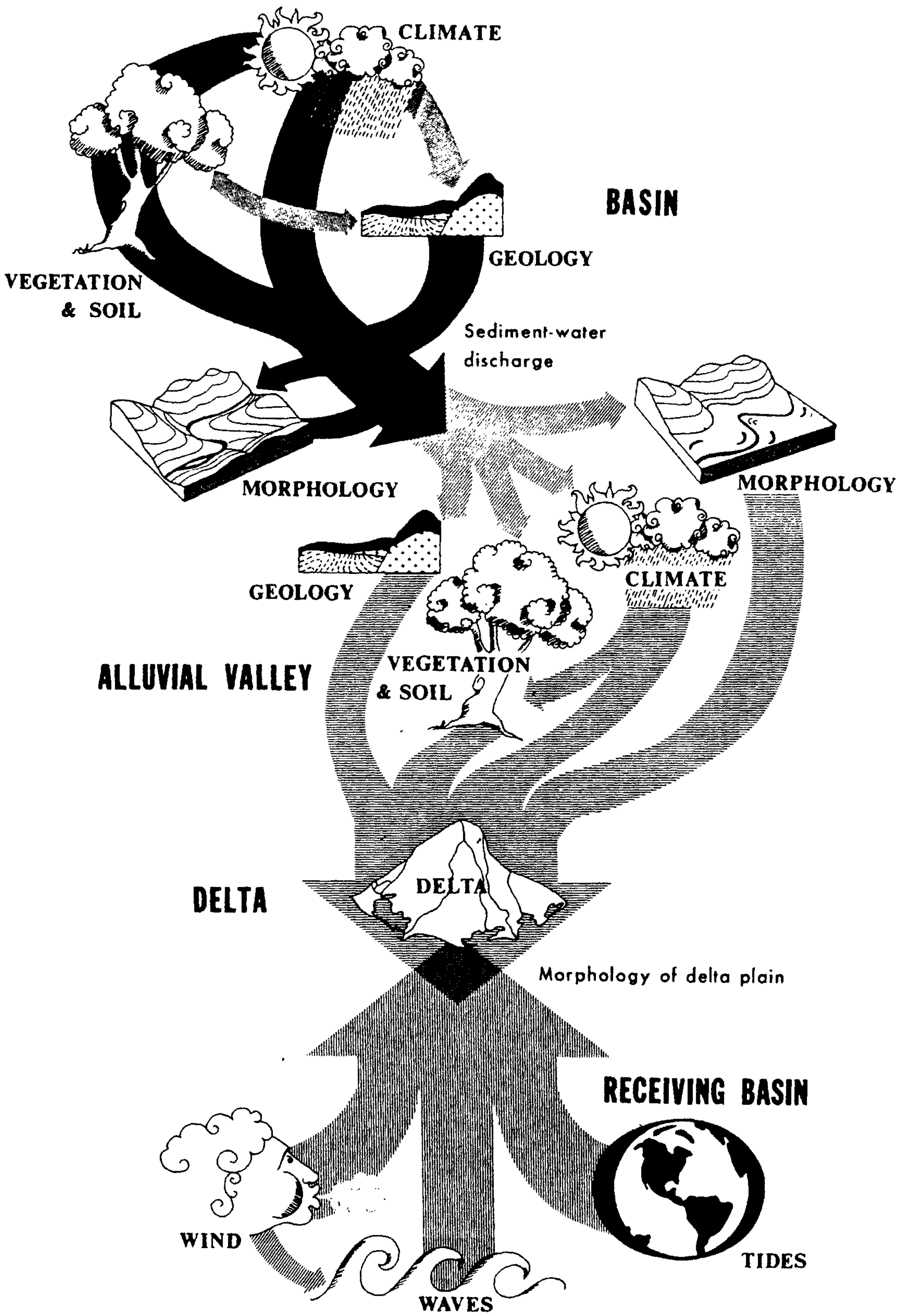


Figure 29. Mojor process controls on river systems. (From Coleman, 1980)

C. Sediment Yield

- River with high fine-grained suspended load build extensive unstable platforms characterized by complex compectional and deformational features.

D. Rivermount Processes

- Distributary-mount-bar sand-body geometry and distribution controlled by three major forces: internal (narrow, linear sand bodies); buoyant (thin, widespread, coalescing sand bodies); and frictional (bifurcated channels with middle ground sand bars capped by subaqueous natural-levee deposits)
- All distributary-mount-bar deposits display coarsening-upward sequences.

E. Wave Power

- High, persistent wave power produces straight delta shorelines; sand bodies display marine characteristics and are oriented parallel to depositional strike or form clean, well-sorted sheet; sand bodies display high quartzose content independent of parent material.
- Low wave power results in irregular indented delta shorelines and sand bodies oriented at high angles to depositional strike; sand bodies often clay bound and poorly sorted.

F. Tidal Processes

- High-tide-range deltas have sand-filled channels; numerous sandy overbank crevasse splays; large complex meander belt sand bodies in upperdelta plain

- Macrotidal ranges in narrow seaways, straight, etc., result in formation of major linear offshore tidal ridges and bars that often attain lengths of tens of kilometers and thickness that approach 20-30 m.

G. Wind Processes

- Directional variability in coastal wind systems often produces multidirectional cross bedding to coastal sand dunes.
- In microtidal regions wind stress on nearshore water masses usually controls currents that shape and orient offshore sand bodies.

H. Nearshore Currents

- Sediment transporting currents in delta regions are driven by several forces: permanent deep oceanic currents impinging on the shelf, tidal progradation, wind- and wave-driven and complex density currents
- Currents are responsible for orienting offshore sand bodies parallel or subparallel to depositional strike; sand bodies are in many cases located considerable distances offshore or downcurrent from active delta lob; sand bodies commonly display strong marine bedding characteristics

I. Shelf Slope

- Subsequeous slope controls offshore wave dissipation, which strongly influences sorting and hence porosity of nearshore sand bodies
- Low offshore slopes are commonly associated with multi-channel distributary patterns, which can form complex sand body relationships

- Shelf Topography, especially presence of submarine canyons, plays major role in topping of delta sediments and formation of deep marine sand bodies

J. Tectonic of Receiving Basin

- Rapidly subsiding basins result in overthickening and localization of deltaic sand bodies, whereas relatively stable basins display widespread and laterally continuous delta sand facies
- Localized differential weighting and dewatering of sediments in receiving basin results in formation of large-scale penecontemporaneuos structures, and sediments often display sediment, and complex slumping

K. Receiving Basin Geometry (Figure 30)

- Large-scale geometry of receiving basin exerts significant control on delta configuration and delta switching patterns
- The large-scale geometry of most receiving basins results in formation of similar delta relationships for large distances in an alongshore.

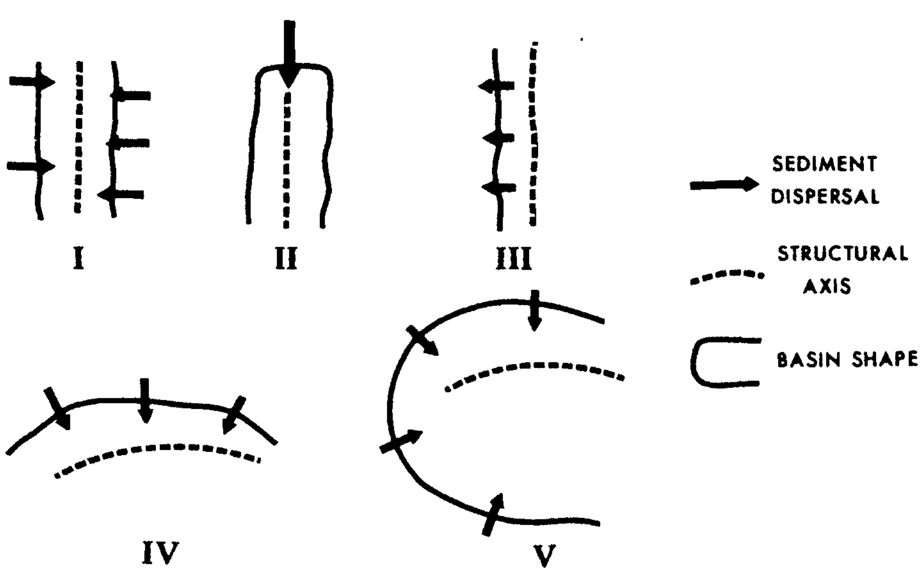


Figure 30. Major configuration of receiving basin. (Coleman, 1980)

L. Relative /sea Level Changes

- Relative sea level rise will first lead reworking of and then abondonment of delta
- Relative sea level fall will lead a lobate delta progradation (seaward).

D. Classification of Deltas

Fisher, Brown et al. (1969) distinguished high-constructive deltas dominated by fluvial processes from high-destructive deltas dominated by basinal processes. Overall, deltas are characterized by coarsening-upward stratigraphic sections (Figure 41).

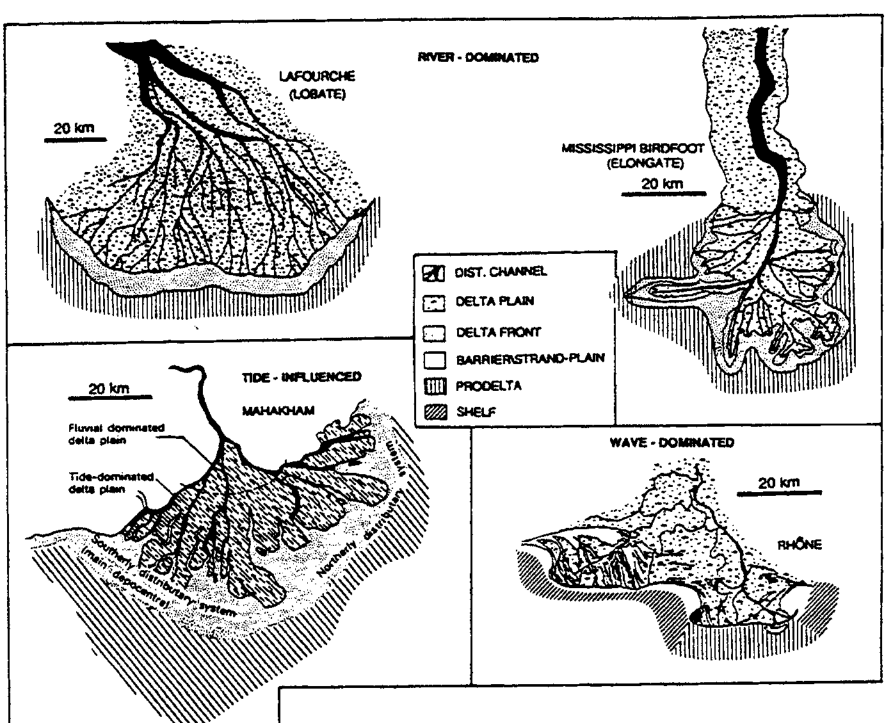


Figure 31. Representative modern example of river-dominated, wave dominated and tide-dominated deltas (Coleman after Allen et al., 1979).

High-Constructive Delta Systems	High-Destructive Delta Systems (wave dominated)
Holocene examples: Mississippi System	Holocene examples: Rhone, Apalachicola, Tabasco coast, Surinam coast, Po.
Gulf Coast examples: Lower Wilcox (Eocene), Yegua (Eocene), Jackson (Eocene), Woodbine (Cretaceous), Cotton Valley (Jurassic)	Gulf Coast examples: Upper Wilcox (Eocene), Vicksburg (Oligocene), Frio (Miocene)
1. Source: distant, inland, continental interior; on order of 2,000 miles; fluvial system and facies concentrated along edge of depositional basin; streams heading delta system meandering	1. Source: local, basin margin; order of 200 miles; fluvial system and facies relatively uniformly developed along edge of depositional basin; streams heading delta system meandering and braiding
2. Sediment input: high volume and more or less continuous rate of input; distinct phases of construction and destruction; pronounced coastal progradation	2. Sediment input: moderate volume and sporadic rate of input; constructional and destructional phases interrelated and not vertically distinct; slight to moderate coastal progradation
3. Constructional facies: well-developed and extensive constructional sequences (progradational and aggradational); thick delta plain deposits; numerous lignites, peats, or coals; distributary channels chiefly straight and bounded by well-developed levees	3. Constructional facies: constructional sequences local and areally restricted; in situ carbonaceous deposits not well developed; distributary channels generally of meander belt type
4. Destructional facies: marine destruction commonly restricted to distal part of constructed delta mass; landward destruction marked by extensive peat or marsh deposits; temporally and vertically distinct from constructional facies; relatively low proportion of destructional sediments in delta system	4. Destructional facies: extensive, developed contemporaneously with construction; marine transgressions commonly far inland with extensive shallow embayments; commonly associated with nondeltaic (strandplain, coastal barrier) systems; high proportion of marine reworked sediments in delta systems
5. Associated systems: commonly large scale laterally associated (down longshore drift) and principally including strandplain, barrier bar, and lagoonal systems	5. Associated systems: generally cannot support large-scale laterally associated depositional systems; local barrier bar and strandplain facies included in delta system
6. Delta flank systems: well developed as independent systems	6. Delta flank systems: poorly developed as independent systems
7. Prodelta: thick, commonly the thickest facies of delta system	7. Prodelta: relatively thin, generally no thicker than other delta facies and commonly indistinguishable from nondeltaic shelf muds
8. Shape: lobate to elongate; main axes perpendicular to regional depositional strike; numerous lobes	8. Shape: chevron to cuspate with axes subparallel as well as perpendicular to regional depositional strike; few lobes
9. Sand/mud ratio: commonly low, results in common differential compaction features; mud diapirs; stabilization of distributaries with prominent levee development	9. Sand/mud ratio: relatively high; distributaries commonly not well stabilized; muddy aggradational deposits not well developed as part of delta plain facies

Figure 31a. Salient features of high-constructive and high-destructive delta systems (From Fisher et al., 1969)

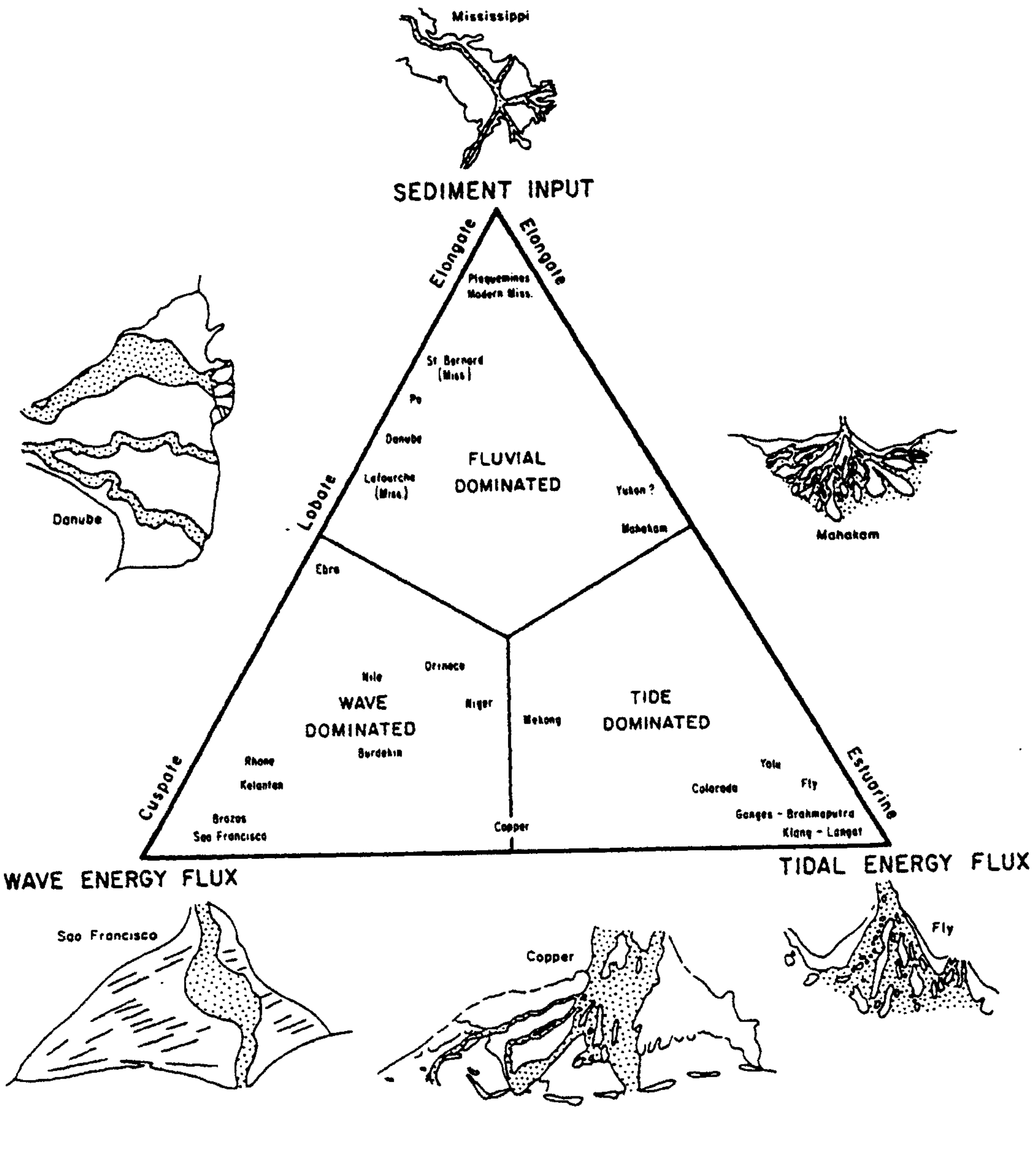


Figure 31 b. Morphologic and stratigraphic classification of delta systems based on relative intensity of fluvial and marine processes (From Galloway, 1975).

Lobate and birdfoot (Elongate) types were recognized in the high-constructive class, and wave-dominated (cuspate and strandplain) and tide-dominated (radial) in the high-destructive class (Figure 31). Each type has a characteristics morphology and facies pattern, described in terms of vertical sequences (Figure 41), areal facies distribution and sand body geometry (Figure 31a, and 31b).

Scott, in Fisher et al., 1969 illustrates three basic delta morphologies. This is a model of the process-response mechanisms operating along deltaic coasts that gives a picture of how riverine and marine processes interact (Figure 32).

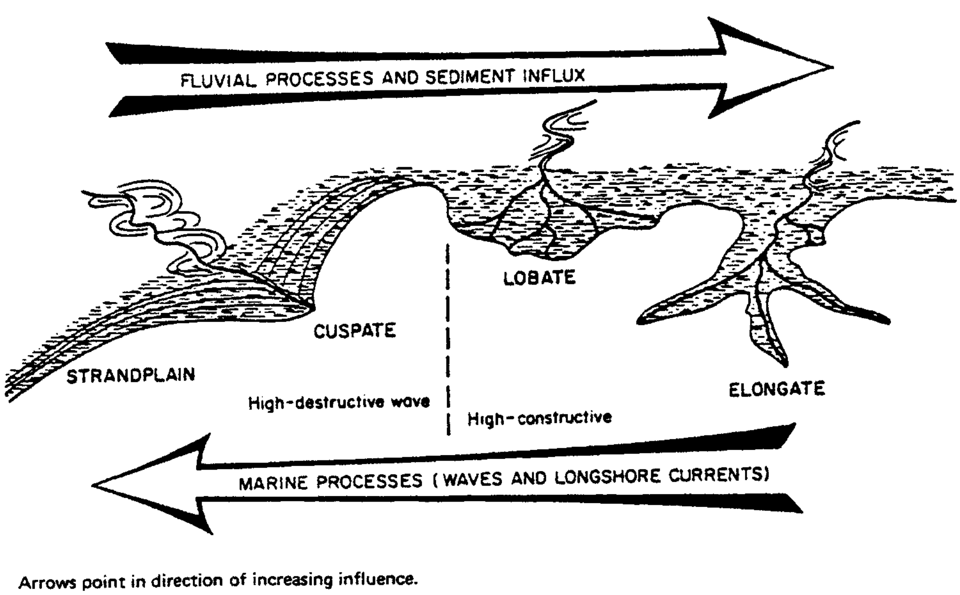


Figure 32. Relative influence of fluvial and marine processes on general configuration of deltas (From Scott, 1969, in Fisher et al., 1969).

D. 1. High-Constructive Systems

High-constructive systems are river-dominated deltas formed by progradation under low bed load/suspended load ratio. Progradational facies (prodelta, delta-front, channel mouth-bar facies) and aggradational facies (distributary channel, levee, crevasse-splay, coal, interdistributary-bay facies) constitute the bulk of the system. Contemporaneous destructional processes are minor and destructional facies are relatively minor in such systems (Brown et al.,1973).

Progradational sand facies are built basinward over relatively thick prodelda muds; after delta abondonment, the delta subsides, allowing thin destructive facies (and eventually shelf facies) to transgress the foundering delta front (Brown et al., 1973). Source areas are relatively distant, and fluvial systems at the head of the delta are meanderbelt. In modern deltas sediment input is relatively high and uniform. Modern deltas exhibit pronounced constructional and destructional phases. Constructional facies are well developed and include both pogradational nad aggradational facies. Destructional facies commonly restricted to distal parts of the system and consist of small barriers, marsh, and reworked strand plain sands. The low sand/mud ratio is the factor that causes soft sediment deformation in high-constructional delta systems (Brown et al., 1973).

D. 1. a. River-dominated delta

The best-known is the recent Mississippi delta. This tpye of deltas are characterised by a skeleton of radiating, shore normal barfinger sands that formed by the progradation of distributary mouth bars (Fisk,1961). The very straight nature of the bar fingers is largely due to lack of reworking of sands in the marine environment and the muddy nature of the sediments; wave action is slight and tidal range is very low. Overall, the facies are dominated by muds with well-developed coarsening-upward mouth bar succession.

D. 2. High-destructive deltas

High-desutructive deltas are dominate by the environments and processes of the receiving basin. In high-destructive deltas, sediment input is relatively low compared to high-constructive ones. In these deltas, fluvially intriduced sediments are comtemporaneously reworked by marine processes; wave, tide , or combination of both. In wave dominated, high-constructive deltas, principal accumulation is as a series of coastal barriers flanking the river mouth, giving a cuspate to arcuate trend of the main sand units (Fisher et al.,1969). In tide-dominated high-destructive deltas, fluvially introduced sediments are reworked by tidal currents into a series of digitatesand units radiating from the front of the river mouth (Fisher et al., 1969).

D. 2. a. Wave-Dominated Deltas

The general structure of wave-dominated deltas is marked by coarcening-upward progradation: first muds of the shelf and prodelta, then an irregullar alternation of sands, silts, and bioturbated muds of the distal delta front, overlain by sand-bars with sedimentary structures resulting from waves and currents, and finally well-sorted plane-bedded beach sands (Figure 41).

One of the best studied ancient wave-dominated delta occurs in the upper Cretaceous San Miguel Formation, Texas (Weise, 1980). The isopach maps of ten stacked lobes in the formation constructed by Wiese (1980) showed a wide variety of sand body morphologies (Figure 33). Weise (1980) related the differences in sand body geometry to differences in wave and fluvial energy (Figure 33).

D. 2. b. Wave-Influenced Delta

Wave-influenced deltaic systems show facies that are intermediate between the river- and wave-dominated end members (Bhattacharya and Walker, 1991). The Upper Cretaceous Dunvegan Formation in Alberta (Bhattacharya and Walker, 1991) show well-developed coarsening-upward facies succession.

D. 2. c. Tide-Influenced Delta

Tide-dominated deltas contain an abundance of linear features oriented about perpendicular to the overall trend of the shoreline and parallel to tidal flow (digitate or radial) (Davis, 1983). Klang-Langat delta of Malaysia is one of the best studied tide-dominated deltas (Coleman, 1974). This delta is characterized by low sediment discharge and low wave action. However, tidal action are the physical proccesses that form the delta. the delta plain is consists of tidal flats, mangrove swamps, and two-way flowing tidal channels.

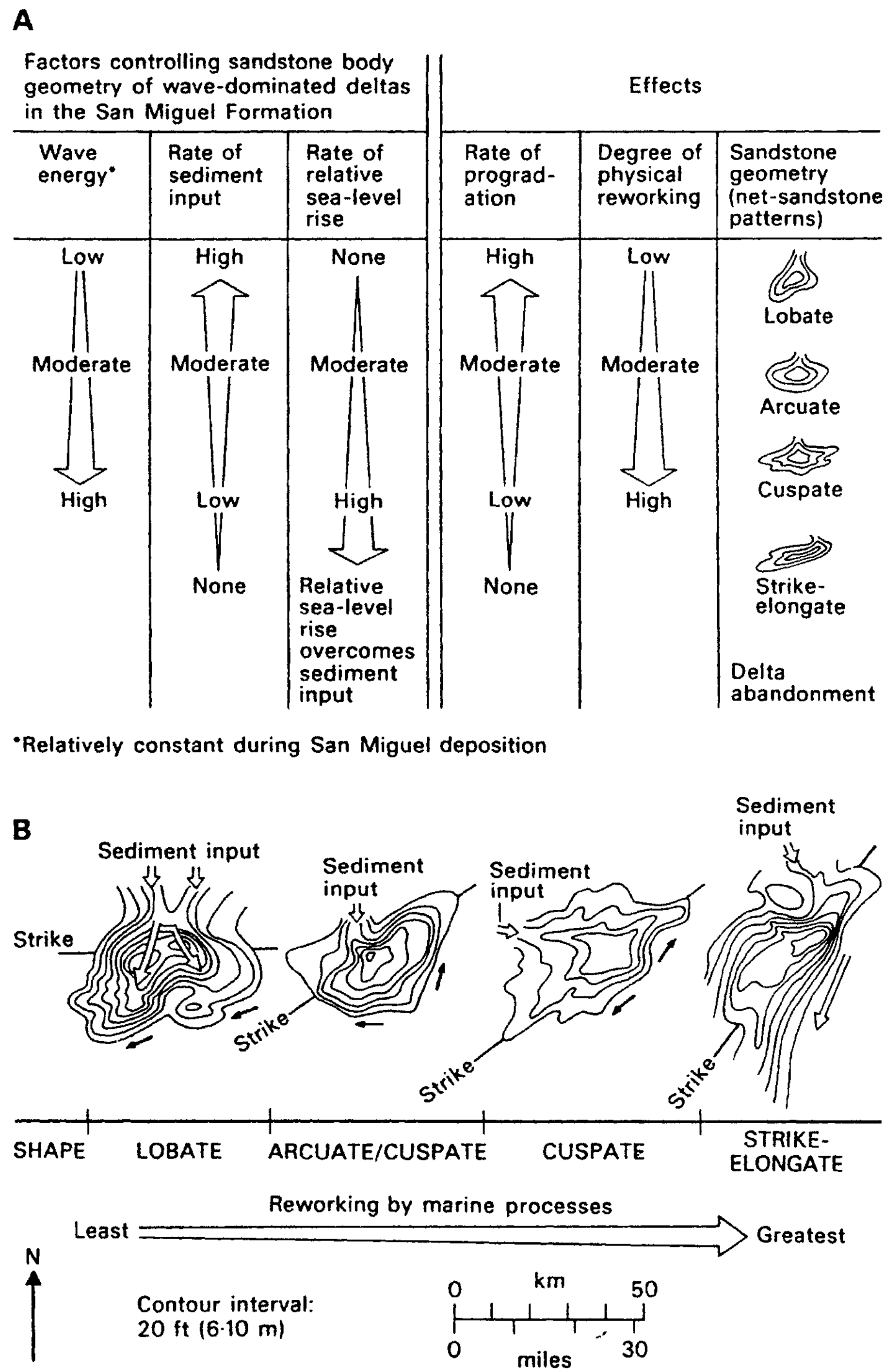


Figure 33. Controls and responses of the Cretaceous San Miguel delta system, Texas (From Weise, 1980)

The vertical sequence of this delta consists of basal muds of the prodelta zone, tidal sand bodies of the delta front zone, and a delta plain of tidal mudflats cut by two-way tidal channel sands overlain by tidal flats and mangrove swamps.

Niger Delta is also one of the best-studied tide-dominated deltas.

E. Deltaic sand distribution models and basic deltaic morphology

Deltas exhibit a wide variety of areal configurations, depending on the interrelationships between sediment supply and the major processes in this dynamic environment (Davis,1993). Primary effects that must be considered in the development of delta morphology are the riverine processes and their related sediment load, wave energy flux, logshore currents, and tidal energy flux (Figure 31b).

Coleman and Prior (1980) classified deltaic sand bodies into six types of sand distribution pattern based on the various process settings in which modern deltas are accumulating (Figure 34).

- Type one(I) consists of a wide spread body of sand composed of primarly of distributary mouth-bar deposits. Within the sand body are definite fingerlike thickenings of sands associated with individual distributaries. This type sand distribution pattern is formed by strong fluvial energy, exetremely low wave energy, low tide range, low offshore slope. low lotoral drift, and a high fine grained suspended-sediment load. Elongate depositional pods are normally oriented at a high angle to the shore line or depositional strike.

- Type two(2) consists of finger-like protrusions of channel sands and numerous isolated sand bodies lying seaward of the shoreline. The sand fingers represand sand-filled channels and nearly always display a scored base. Linaer sand bodies offshore represent deposition by tidal processes. Environmental setting conductive to the formation of this type of pattern include low wave energy; high tide range; narrow, resitricted depositional basins; and normally low littoral drift.
- Type three(3) is formed and sand bodies begin to aling with the depositional strike as wave energy is increased. Sand-filled channels still persist and form sand bodies trending at high angles to the shore line, but increased wave energy is responsible for the formation of beach-dune-ridge complexes that parallel the shoreline and intersect channel sands at high angle.
- Type four(4) consists of distributary-mouth-bar deposits in the more landward setting, and barrier-beach deposits in seaward of the delta sedimentation. Distributary-mouth-bar deposits form in some instances long finger-like sand bodies and in distributary-mouth-bar deposits merge a widespread sheet sand with only localized thickening. Intermediate wave energy, extremely low shore slope, and low sediment input tend to be associated with this pattern.

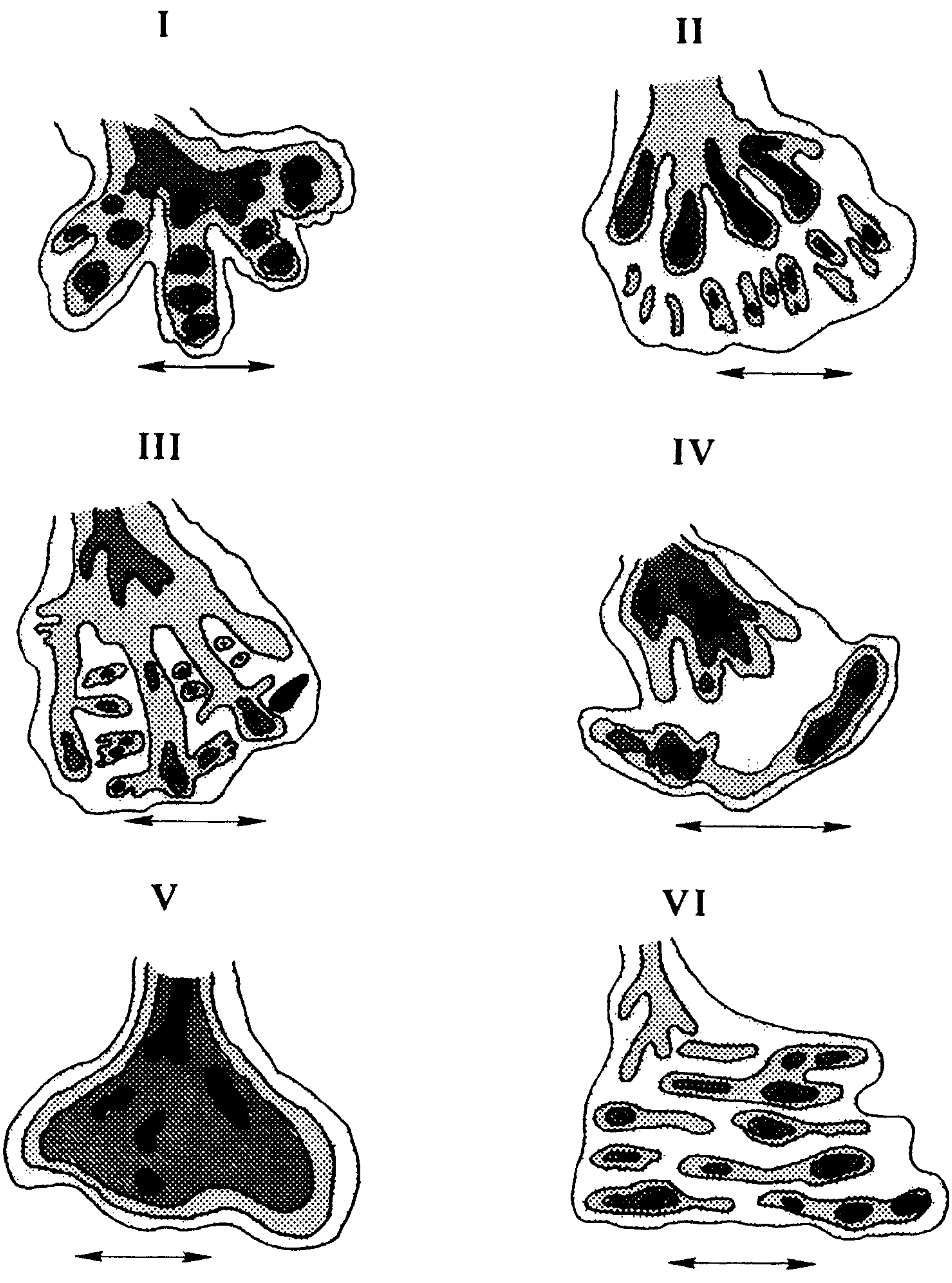


Figure 34. Net sand distribution pattern in modern deltas (From Coleman, 1980).

- Type five(5) is characterized by high lateral continuity of sands and a sheetlike geometry. Clean, well-sorted sands that display quite uniform thickness over large areas are common, and local thicknesses within the sand sheet are due primarily to scoured channels, which may or may not be sand filled. High persistent wave energy, low litoral drifft, and steep offshore slope characterize the environmental setting.
- Type six(6) is characterized by more than one sand ridges that parallel the shoreline. Extremely high wave energy, Low sediment input rate, low tide, and extremely low offshore slope are the characteristics of this kind of delta pattern. Senegal delta of Africa is the best example for this kind deltas.

E. Deltaic depositional environments

Deltas commonly exhibit numerous sedimentary depositional environments within the depositional systems. Each of these is characterized by its own assemblage of sediments and sedimentary structures as shown in Figure 41.The term delta includes all the delta plain, delta front, and prodelta deposits of a particular river (Figure 35). Figure 42 illustrates generalized stratigraphic cross-section for unique types of deltas as delta fronf prograde seaward.

E. 1. Delta plain

Delta plains are extensive lowland areas which comprise active and abandoned distributary channels separated by shallow-water environments and emergent or near-emergent areas. Usually, deltas have more than one distributary channels spread across the delta plain. Between the channels is a varied assemblage of bays, floodplains, lakes, tidal flats, marshes, swamps and salinas which are extremely sensitive to climate (Elliott, 1992). For example, in tropical settings, luxuriant vegetation prevails over large areas of the delta plain. In contrast, delta plains in arid and semi-arid areas tend to be devoid of vegetation and are characterized by calcretes (Ebro Delta) or salinas with gypsum and halite (the Nile Delta). Alternatively, arid delta plains are dominated by eolian dune fields, particularly in sandy, wave-influenced deltas where sand is eroded from active and abandoned beach ridges (Elliott,1992). Most delta plains are affected by fluvial or tidal processes but only rarely by major waves as wave-influenced deltas are characterized by beach-barrier shorelines which enclose and protect the delta plain.

Fluvial distributary channels are characterized by unidirectional flow with periodic stage fluctuations, and are therefore, similar to channels in strictly alluvial system. High sinuosity patterns are common.

Facies and sequences of distributary channels resemble those of alluvial channels to a large extent. The overall finning-upward results either from lateral migration of the channel, or more commonly, from channel abandonment, with the upper fine member representing infilling of the channel by diminishing flow and perhaps later by overbank flooding from an adjacent active channel. Large -scale bank slumping can be an important feature of distributary channels as they often have fine-grained, cohesive bank materials (Coleman, 1981).

The switching or avulsive behavior of fluvial distributary channels causes them to be short-lived channels relative to upstream alluvial equivalents. Sand bodies of the

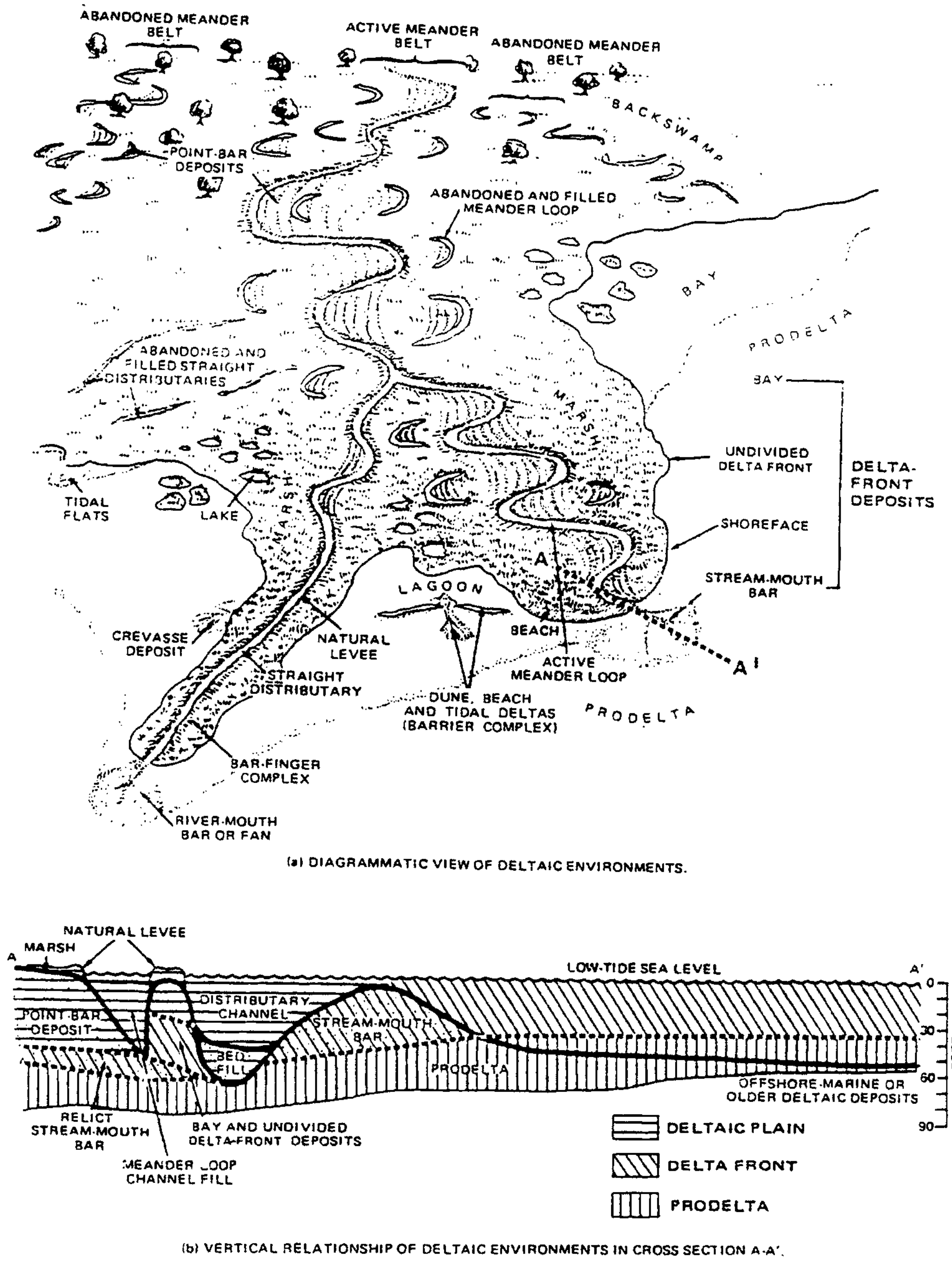


Figure 35. Principal shallow-water deltaic environments and facies with their relationship to one another and to sea level (From

distributaries often have a lower width to depth ratio as a result.

Interdistributary areas of fluvial - dominated delta plains are generally enclosed, shallow water environments. Flood-generated processes are the principal means of sediment supply to the interdistributary areas and features which results from these processes include levees, various types of crevasse channels. These features collectively fill large areas of the shallow bays and provide platform for vegetation growth, gypsum and halite precipitation, or calcrete development, depending on prevailing development (Elliott, 1974b).

Overbank flooding involves sheet-flow of sediment-laden water over the channel banks. Fine-grained, laminated sediment is deposited over the entire area, although frequently the laminations are destroyed by subsequent bioturbation. Coarser sediments are confined to the channel margins and contributes to the growth of levees. As a result, levees facies comprise repeated alternations of thin, erosive-based sand beds deposited from sediment-laden flood and silt-mud bed deposited from suspension.

One of the major facies associated with many deltas is the large areal extent of *the bay fills or crevasses* that break off main distributaries and infill the numerous interdistributary bays in the lower delta plain. Most of the bays are usually elongate shallow water bodies surrounded by march and ditributary channels.

The spatial distribution of processes operating in a fluvial-dominated interdisributary area is determined by the distance from active distributary channels. Near-channel facies will be dominated by levee sequences and numerous crevasse splay lobes, whilst distal or central facies may comprise fine-grained bay floor sediments or crevasse channel/minor mouth bar couplets (Elliott, 1992).

Of crevasse splay, the sediment may be deposited in many small, anastomosing streams in which case the deposit comprises numerous small channel lenses (Arndorfer, 1973).

Stage variations are particularly important in the crevasse channels as the channels may be temporarily abandoned at low river stage, resulting in complete cut-off of sediment supply until next river flood (Elliott, 1992). Crevasse channel facies may therefore comprise sands with undirectional, current-produced structures, together with numerous reactivation surfaces and fine-grained drapes (Elliott, 1992).

b. Tide-dominated delta plains

In areas of moderate to high tidal range, tidal currents enter the distributary channels during tidal flood stage, spill over the channel banks and inundate the adjacent interdistributary areas. The tidal waters are stored temporarily and subsequently released during the ebb stage. Tidal currents are therefore predominant in the lower distributary courses and the interdistributary areas assume the characteristics of intertidal flat (Elliott, 1992) (Figure 36).

Tidally influenced distributary ***channels*** have a low sinuosity, flared and sometimes funnel-shaped form with a high width to depth ratio which contrasts with the

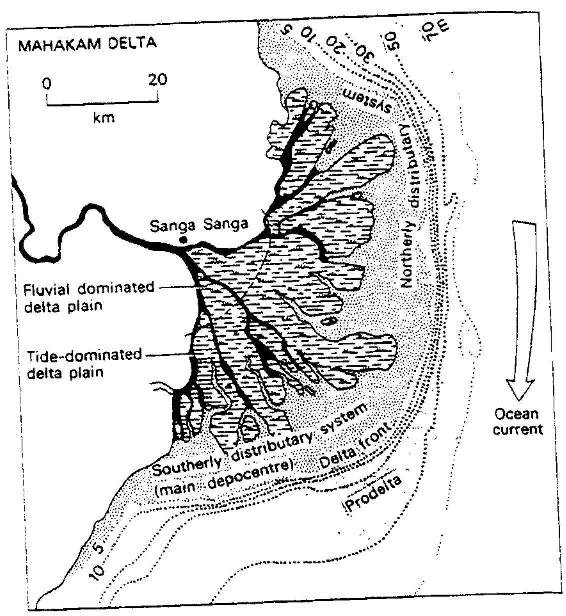


Figure 36. Mahakam delta, Indonesia; a fine-grained, tide-dominated delta with an extensive area of tidal flats, estuarine channels, tidal channels and creek dominating the delta plain (From Elliott, 1991, after Allen, Laurier and Thouvenin, 1979).

almost parallel-sided nature of fluvial distributary channels in areas of low tidal range. (Wright, Coleman and Thom, 1973). Tidal distributary channel are less prone to switching and abandonment, but can migrate laterally. Sand body shape and dimensions are therefore a function of the size and form of the channels and the degree of lateral migration (Elliott, 1992).

Interdistributary of tide-dominated delta plains include lagoons, minor tidal creeks and intertidal-supratidal flats which are sensitive to the climate. The entire delta plain probably comprises a sheet-like complexes of small-scale, erosive-based sequences which pass upwards from point bar sands-silts into the mangrove swamp facies, with localized clay plags representing infilled channels.

E. 2. Delta Front

The seaward portion of the delta (Figure 35) in the shallow subtidal zone is generally called the delta front (Allen, 1970b). This zone is typically no more than a few kilometers wide and extends to a depth of about 10m or less. It is characterized by sand-size sediment with grain size decreasing seaward where some mud is mixed with the sand. This zone is dominated by marine processes. The delta front is primarly a sheet sand accumulation with local sandbars (Wright, 1985).

Rive mouth processes and sediment load are the most important characters that determine the delta front type. The river mouth is the point at which the seaward-flowing water leaves the confines of the channel banks and spreads and mixes with ambient waters of the receiving basin. The resulting geometry and distribution of river-mouth bar sand bodies is determined by riverine flow conditions, density contrast between issuing and ambient waters, bottom slope seaward of the mouth, tidal range, tidal currents with the lower river channel, and the ability of waves and other forces to obstruct the outlet. It is the dynamic dissemination point for sedimentation which contribute to continuing delta progradation and is responsible for forming one of the major sand bodies associated with deltaic sequences, the distributary-mouth bar (Coleman, 1980). This is the area in which sediment-laden fluvial currents enter the basin and are dispersed while interacting with basinal processes (Figure 28). Delta front successions are usally characterized by coarsening-upward sand bodies (Figure 41)

E. 2. a. Fluvial-dominated delta fronts

These types of deltas are characterized by elongate shorelines. The uniform prograding sequence of the delta front is well known in the modern Mississippi. In the Mississippi delta, sedimentation is dominated by fluvial processes with minimal interference from basinal processes. Two type of mouth bars are recognized (Figure 39): *Upper:* on the eastern side of the delta, distributaries empty into relatively shallow waters and characterized by bifurcating channels and middle ground bars which result from a predominance of frictional processes at the mouth and, *Lower:* on the southern side of

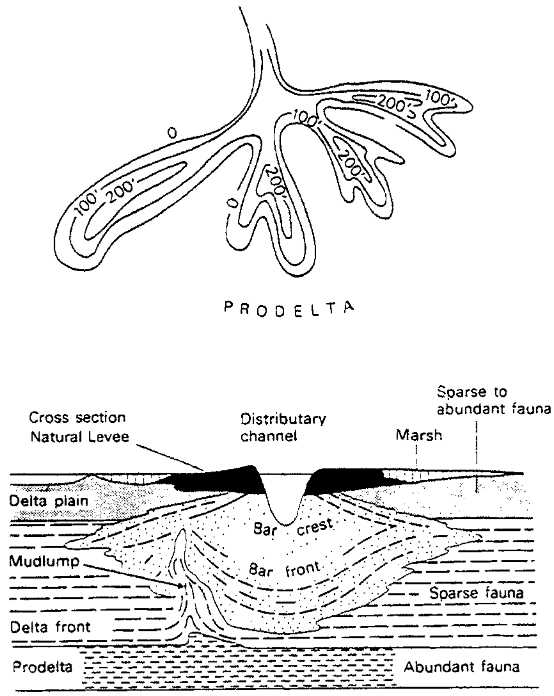


Figure 38. Distinguishing geometric and sedimentary charactertistics of bar-finger deposits.

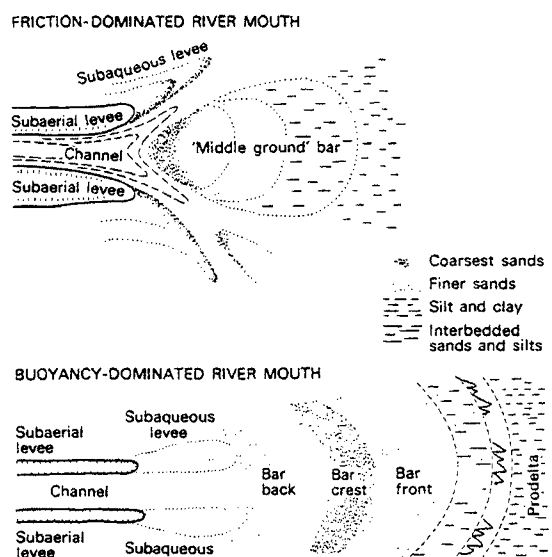


Figure 39. Friction-dominated and buoyancy-dominated river mouth bars which respectively develop in shallow-water and deep-water areas of fluvial-dominated deltas (From Elliott, 1991, after Wright, 1977).

the delta, distributaries enter deeper water which favors the operation of buoyancy processes during periods of low to intermediate river discharge (Elliott, 1992). Progradation of these mouth bars produces large-scale (60-150 m) coarsening-upwards sequences which record a transition from prodelta clays upwards into sands of the upper bar front and bar crest (Fisk, 1961; Coleman and Wright, 1975). As the delta prograde seaward and directly overlie prodelta muds, prodradation of distributary channel-mouth systems produce a series of radiating " bar finger sands which form the birdfoot framework of the delta. In Mississippi delta, these finger sand bodies are bi-convex, elongate bodies up to 30 km long, 5-8 km wide with thickness of 70 m (Fisk, 1961; Figure 38).

E. 2.b. Fluvial-wave interaction delta fronts

The ratio of fluvial processes/wave processes determines the morhology of a particular delta. The Cretaceous San Miguel delta system of Texas represents a spectrum of net sandstone pattern (Wiese, 1980) resulting in degree of marine reworking by waves and longshore drifft (Figure 33). Present-day examples occur in the Danube, Ebro, Nile and Rohane deltas, all of which are located in enclosed seas with moderate wave action but minimal tidal processes. In, general, this type of delta front is characterized by a smooth, cuspate or arcuate, beach shoreline.

Sand distribution pattern in the Rhone delta consists of a laterally extensive beach-barrier sand, cut locally by distributary channel sands (Oomkens, 1967). Progradation is by beach ridges accretion and mouth bar progradation, and is most pronounced in the vicinity of the main distributary of the Grand Rhone. Oomkens (1967) described delta front coarsening-upwards sequences, having bioturbated offshore clays that pass upwards into finely laminated silt and sand in the intermediate part of the sequence.

E. 2. c. Wave-dominated delta fronts

In this type, wave processes are capable of redistributing most of the sediment supplied to the delta front. Therefore, it is characterized by a regular beach shoreline with only a slight deflection at the distributary mouth and a relatively steep delta front slope. Mouth bars do not form and bathymetric contours parallel to the shoreline. Progradation involves the entire delta front, rather than particular points and is generally slow by comparison with other types. Abandoned beach-ridges occur behind the active shoreline and the delta plain is often dominated by aeolian dunes and shallow, elongate lagoons between beach ridges. Changes in the direction of longshore drift or the position of distributary channels affect the shoreline configuration (Psuty, 1967).

Delta front of the Sao Francisco display a coarsening-upward sequence (Coleman and Wright, 1975). Bioturbated, fossiliferous muds at the base pass upwards into alternating mud, silt and sand beds with wave induced scouring, grading and cross-lamination and finally into a well-sorted sand with parallel and low-angle lamination representing a high energy beach face. As progradation involves the entire delta front the resultant sand body is a sheet-like unit which parallels the shoreline (Elliott, 1992).

E. 2. d. Tide-dominated delta front

In tide-dominated delta fronts the shoreline and distributary mouth areas are often an ill-defined maze of tidal current ridges, channels and islands which may extend a considerable distance offshore before giving way to the delta front slope (Coleman, 1969). The main features of this type of delta front are the tidal current ridges which radiate from the distributary mouths. In the Ord River delta the ridges are on average 2 km long, 300m wide and range in height from 10 to 22 m (Elliott, 1992). Channels between the ridges contain shoals and bars covered by flood- and ebb-oriented bedforms (Coleman and Wright, 1975). In an idealized vertical succession from this delta, the tidal current ridge sands at the top of delta front coarsening-upwards sequence are composed of bi-directional trough cross-beds with occasional clay drapes and numerous minor channels (Coleman and Wright, 1975; Figure 40). In terms of sand body characteristics, this type of delta front will probably produce relatively thick, elongate bodies aligned normal to the shoreline trend.

D. Deltaic System Tract

System tracts (linkages of related contemporaneous depositional systems) form an important conceptual framework for understanding the larger scale relationship between depositional systems (Bhattacharya and Walker, 1991).They also help in the interpretation of relative sea level changes and allow sedimentary rocks to be studied in the context of sequence stratigraphy (Brown and Fisher, 1977; Van Wagoner et al., 1990).

Deltas deposited during times of relatively high sea level (highstand systems tracts) are usually confined to the shelf and deposited in shallow water. These shallow water deltas are characterized by rapid lobe switching. During time of sea level fall, elongate deltas may result form from progradation into depeer water on th mid to lower shelf, especially when sediments are relatively fine-grained. If sedimentation rate just keeps up with relative sea level fall, lobe and channel switching will be supressed, enhancing the elongate geometry (Bhattacharya and Walker, 1991). Deltas deposited after a fall of relative sea level (low stand or shelf margin system tracts) commonly overlie an incised topographic surface caused by the preceding fall.

During times of active sea level rise, (transgressive system tracts) deltaic deposition will commonly be suppressed because river-borne sediment accomulates on the aggrading flood-plain. Where active deltaic deposition does accure, the fluvial influence is minimal and tide- and wave-dominated deltas will tend to dominate. Perhaps more importantlly, flooding of river valleys during transgression forms eustaries (Bhattacharya and Walker, 1991).

Active development of deep water submarine fans is commonly related to the development of shelf edge deltas in lowstand systems tracts Development of fans may also correlate with the incision of alluvial systems in a landward direction. Active submarine fan development commonly ends with a rise of relative sea level, and may coincide with development of shoal water deltas in a transgressive or high systems tracts (Bhattacharya and Walker, 1991).

F. Autocyclicity and allocyclic controls of delta development

The paralic setting of deltaic depositional systems means that they tend to be very sensitive indicators of relative sea level change. the repeated development of similiar facies successions ("cyclicity") due to switching delta lobes is avery common feature in many deltaic depositional settings (Elliott, 1990).

Autocyclic processes are intrabasinal in origin, and are related to the sedimentological behaviour of the depositional system and can include eustacy, tectonics in the source area and receiving basin, climate and other factors. allocyclic tend to produce more wide spread effects (Bhattacharya and walker, 1991).

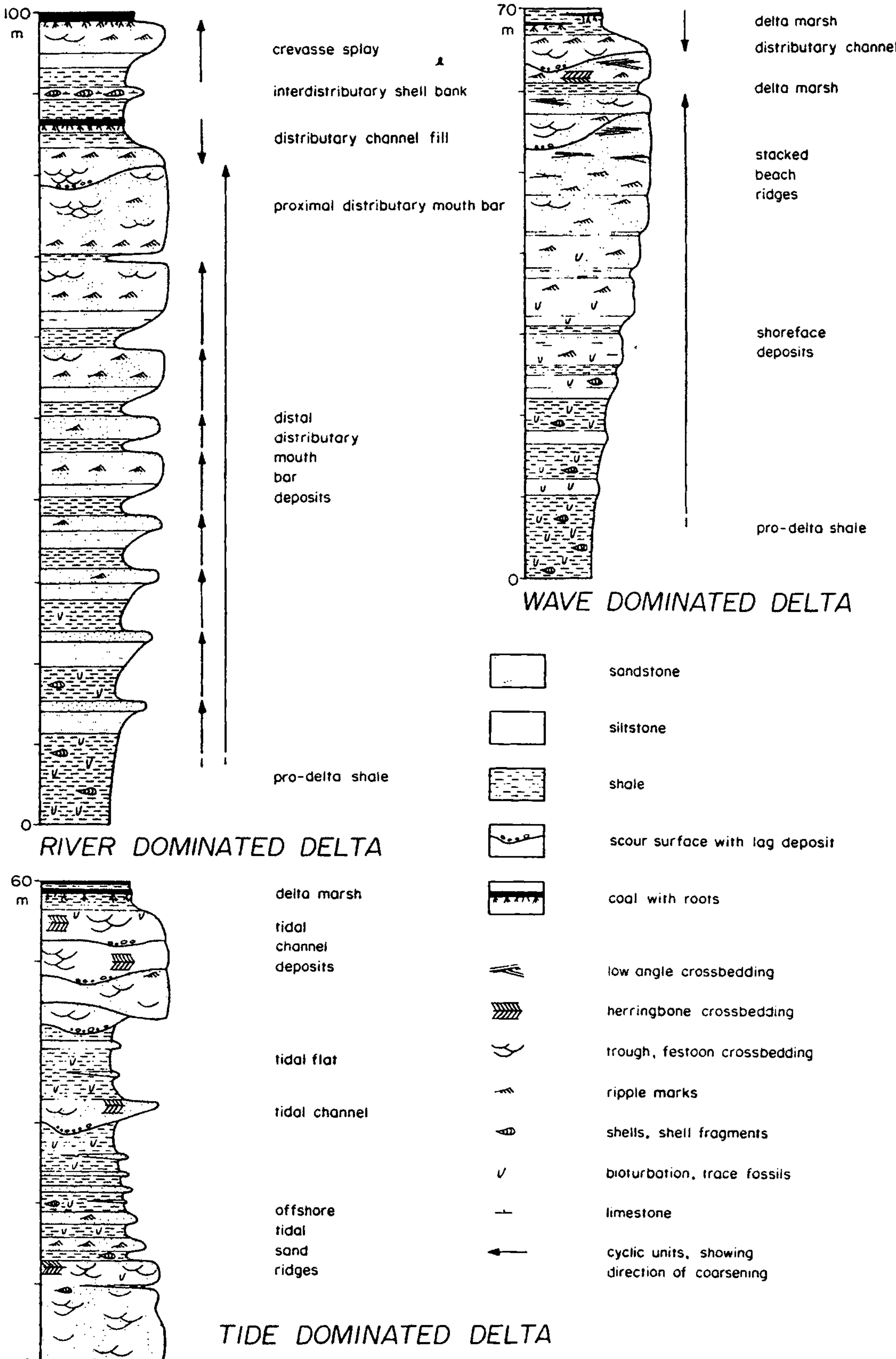


Figure 41. Vertical sequences through idelized river-dominated, wave-dominated, and tide-dominated deltas (From Miall 1976)

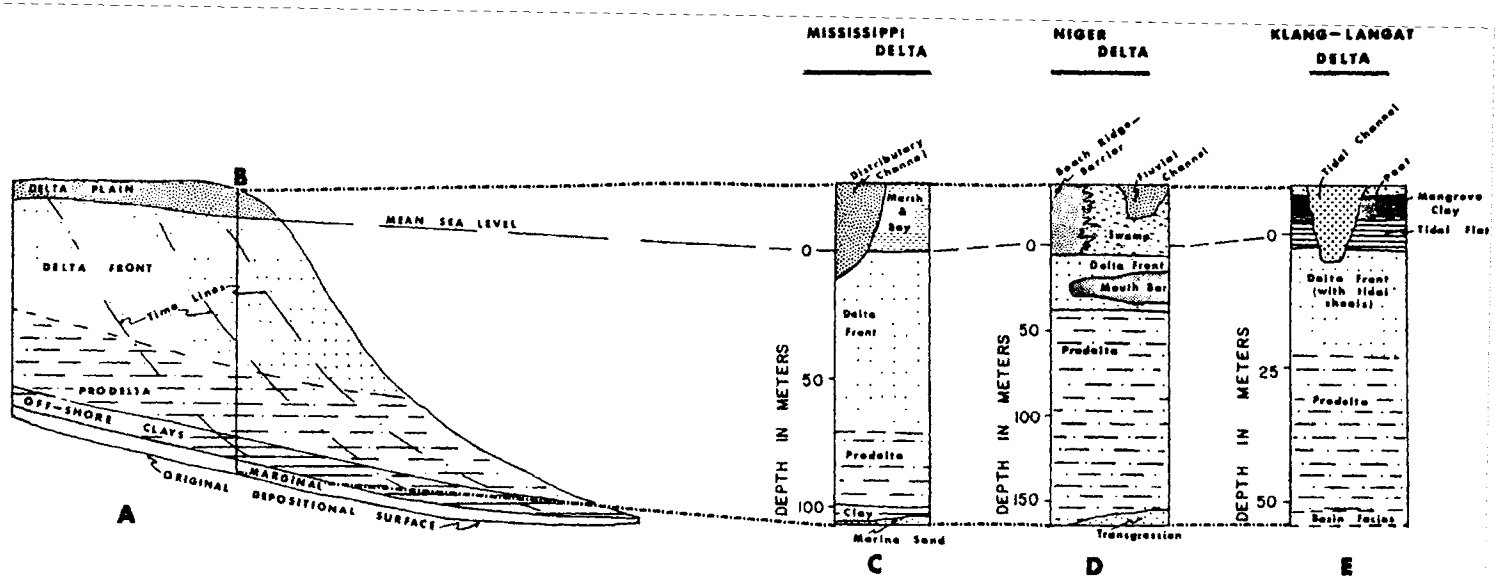


Figure 42. Vertical and lateral distribution of sediments on idelized delta based on vertical succession of deltaic subenvironments (from Klein, 1974)

STRATIGRAPHIC PROCEDURES

Plate III provides a representation of the mechanical well-logs used in this study. Several of these logs are usually recorded after reaching a casing point or at the total depth of the drilled well. These mechanical well logs graphically illustrate response of a mechanical recording tool to its position in the borehole. They are continuous recordings of the acoustical, electrical and radioactive properties of the penetrated strata.

These measured responses are recognized as a well-log signatures, and were used stratigraphically locate, delineate, and correlate reservoir units within the subsurface. Thickness data for specific units and inferences about their lithic compositions were also obtained from the logs.

WIRELINE LOGS

Wireline logs measure the electrical, radioactive, and acoustic properties of rocks and these properties are related to lithology, grain size, density, porosity and the pore fluids. Since they run continuously up a bore hole they are particularly valuable as indicators of sequences on a scale of meters to hundreds meters and can be used in sequential environmental analysis provided the logs reflect the sedimentological parameters and not fluid properties or other secondary features (Reading, 1986).

The *induction-electric* and *induction-gamma ray* logs from the numerous gas wells within the study area serve as a data base for this report. Mechanical well logs are graphic illustrations derived from continuous recordings of the acoustical, electrical, and radioactive properties of the formations penetrated by the well bore. These responses are observed as well-log signatures, and were used to delineate and correlate the sandstone units of concern in this study. Once correlation was achieved, the various stratigraphic maps and cross-sections were constructed. Unit tops and bottoms provided thickness values which allowed construction of isopach maps. Comparison of the magnitude of deflections for specific log curves allowed sandstone isolith values to be derived and maps constructed. By using well-logs, various stratigraphic cross-sections were constructed to show the characteristics of the study interval, to recognize and delineate depositional environments, and to determine the stratigraphic relationships in the area.

The spontaneous potential (SP) curve of an electric log or induction-electric log is a record of the naturally occurring potentials in the well bore as a function of depth. The spontaneous potential curve is useful to detect permeable beds, to locate bed-boundaries for correlation purposes, to provide qualitative indications of bed shaliness, and to calculate formation water salinity and water resistivity parameters used in calculating petroleum and water saturations. Due to the changes in permeability, changes in the spontaneous potential can sometimes be used to infer changes in gross gain-size distribution. The spontaneous potential curve is recorded in Track-1(left l side) of the mechanical well log, and is commonly run simultaneously with resistivity or other logging devices. Recording the spontaneous potential requires electrical continuity between the tool and the penetrated strata, thus restricting its use to fluid-filled holes. The potential cannot be recorded in the holes that are air-filled, cased or in holes filled with nonconductive (oil-base) fluids. Maximum intensity of the spontaneous potential occurs opposite permeable beds. Branan (1968) states that Atoka sands in the Arkoma basin of Arkansas have low interstitial water saturation, a factor which increases the resistivity of the rock and subdues the spontaneous potential. North of the Mulberry fault zone, the sandstones contain sufficient connate water to permit good spontaneous potential development, however, south of this fault zone the sandstones are very tight. the consequence is a poorly developed (flat) response (Schlumberger, pers. comm. 1978). The potential response is further subdued by the presence of resistive hydrocarbons (Goetz, 1977), intergranular cement(Selley, 1976), and where the sands are shaley resulting in very shallow invasion of the drilling fluids (Tixier, 1962). Gas recovery has been achieved in sandstones unit with less than two-millivolt deflection of the spontaneous potential curve from the shale-base line. In cases where the spontaneous potential cannot be recorded, as in air-drilled wells of the Arkoma basin, or where it is ill-defined, the gamma ray tool is effectively substituted.

Resistivity-logs record the resistance of rock formations to electric current flow. Recorded resistivity profiles are useful for correlation and for inferring lithology. In general, shales and salt-water saturated and porous rocks have low resistivities; tight and hydrocarbon formations and coals have high resistivities (Readings, 1986). In conventional resistivity logs, currents are passed through the strata via an array of electrodes and recorded voltages provide the resistivity curves (Schlumberger, 1972). The various electrode arrangements provide resistivity measurements in close proximity to the borehole, the invaded zone, and the distal noninvaded rock strata. These curves are always recorded in Tracks 1 and 2 (right side) of the well log, with increasing resistivities shown by deflections to the right.

The resistivity of the rock particles and hydrocarbons is considered infinite, however, the ionized interstitial water reacts as an electrolite and can be considered the only conductor of current in sediments. Rocks with low resistivity such as shale are generally water saturated, whereas highly resistive rocks include the “dry”, dense carbonates, evaporties, quartzites and coals. Variations in the resistivity occur with changes in porosity, permeability, relative fluid saturations (gas, oil and water) in the pore space, salinity, distribution of formation waters and reservoir temperatures.

The short normal curve was employed in this study because the short normal reads the invaded-flushed zone, and it provides subtle changes in rock porosity and permeability. Also, the long normal tends to “average-out” the bed boundaries and the lateral device presented asymmetrical curves.

Resistivity measurements by the induction method eliminates the effects of the borehole. Unless the mud filtrate undergoes deep invasion or is salty, the long-spaced-electrode induction log reads near true resistivity. The deep-focusing induction device was a resolving power approximately equivalent to that of its electrode spacing (about 40 inches). For rocks with low resistivity and where the borehole effects are minimal, it provides a more accurate definition of a bed boundaries, bed thickness, indications of lithology and data for quantitative evaluation for possible productive zones. Where the beds are greater than five feet in thickness, the focusing induction devices are very efficient in reducing the effect of the surrounding beds.

The gamma ray log provides a measurement of natural gamma radiation which spontaneously emitted by some naturally occurring radioactive elements such as potassium, thorium and uranium that are commonly associated with shales and are usually less abundant in sandstones. The gamma ray curve is always presented in Track-1 (left side) of the log. They are commonly taken as indicator of grain size because a high reading normally indicates clays. With increasing radioactivity, the curve is deflected to the right, where maximum deflection, at least in the Arkoma basin, generally represents the shale base line.

Patterns of sequences are apparent in all types of logs, in particular spontaneous potential logs, and gamma ray logs (Figure 45). They can be used to discern facies relationship and evolving environments.

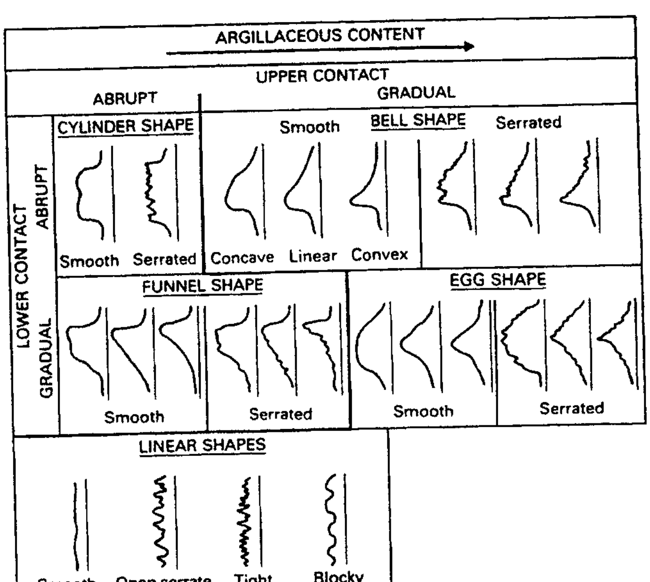


Figure 45. Some shapes of GR/SP log profiles, based on argillageous content. Notice the nature of the upper and lower contacts (Serra and Sulpice, 1975).

If related argillaceous content or grain size, cylinder-shapes on gamma ray or spontenous potential logs indicate either thick relatively homogenous sediments bounded by shaley sediments or channel-fills with sharp tops. Bell-shaped profiles indicate upward fining, possibly due to channel-fills. Funnel-shaped profiles indicate upward coarsening probably produced by prograding systems such as deltas, submarine-fan lobes, regressive shallow marine bars, barrier islands or carbonate forereefs prograding over basin mudstone. Egg-shaped profiles might suggest fining-upwards channel fills with basal shale clast, conglomerates or breccias, progradational-regressional sequences or submarine fan channel-lob systems. Linear profiles can indicate thick mudstone sequences, possibly with interbedded sandstones or siltstones, interfluvial deposits, marsh coals or shales (Readings, 1986).

Induction-gamma ray logs are available for most of the wells drilled in the study area and are one of the most beneficial log-types used in this study. Because of the excellent definition of sand-shale contacts, the gamma ray and induction resistivity curves improved the accuracy of cross-sections, correlations and the interpretations of the depositional processes of the sands.

UNIT BOUNDARIES

The sedimentary succession considered in this report is basically composed of alternating sandstone and shale units. The boundaries between Spiro sandstone units and adjacent shale units are both abrupt and gradational. Where Spiro sand unit directly overlies the Wapanucka limestone the is a sharp contact at the boundary. In order to prepare isopach of the sandstone units, uniform procedures must be applied to establish unit boundaries from mechanical logs so that accurate thickness data can be obtained.

The gamma ray log does not require permeability contrasts in the rock, and is an excellent tool for inferring lithology for correlation purposes.

In a sand-shale sequence, lithology and unit thickness can be inferred from a well-developed spontaneous potential curve. Opposite shale units, the spontaneous potential curve tends to flat, and defines the shale base line, and in a sand-shale sequence, these anomalies indicate sandstone. A line drawn through the points of maximum negative deflection of the spontaneous potential curve to the left of, and parallel to the shale base line. For consistent results, the sand shale boundaries were selected at a point on the spontaneous potential curve located about one-tenth the distance from the shale base line to the sand line.

Unit boundaries defined in this report evolve contrasts between sandstone and shale except in the central part of the area where the Spiro sand rests directly on the Wapanucka Limestone Member. Well-developed spontaneous potential curves usually indicate a well-defined contrasts between the commonly permeable Spiro sand and the underlying dense Wapanucka limestone member. Even though the Spiro sand is permeable, a featureless spontaneous potential curve can result because of the content and sparse interstitial water in the sand (Branan 1969; Schlumberger, 1978) from the presence of gas,or from too much resistivty contrast proximal to the bore hole between the drilling fluid and the formation water.

The induction resistivity curves often indicate that the Spiro sand like the Wapanucka Limestone Member, is very resistive. However, a gamma ray profile most commonly serves to differentiate the sandstone from the less radioactive limestone unit. In addition, the formation-density log provide more definite placement of the Spiro sandstone-Wapanucka boundary. This device emits and counts returning gamma ray energy which pass through the strata. The device will operate in air and fluid-filled holes. The gamma radiation responds to total electron density, which is related to bulk density (interstitial fluid density and grain density of the rock). Porosity can be calculated when the bulk density is measured and the grain and fluid densities are known. An increase in the counting rate at the detector indicates a decrease in bulk density.

Where the Spiro sandstone is calcareous, the spontaneous potential curve is subdued and the impermeable sand is very resistive, making it difficult to distinguish from the underlying limestone. However, in most cases, the density contrast is easily recognizable on the formation-density profile.

MECHANICAL LOG SIGNATURES

The response a logging device to a particular interval of rock as represented by curves recorded during logging procedures is the mechanical log signature for that interval. Log signatures do not provide direct information about the lithic properties of the interval. They do record parameters that allow the inference of textural and compositional variations that characterize the unit.

Many depositional systems produce intervals characterized by specific variances in texture and composition as recorded in vertical successions. The construction of lithic profiles reflecting these variations has been widely used in reconstructing depositional systems where the intervals are exposed at the surface (Fisher, 1968 and 1969). Gamma ray curve forms log signatures that reflect qualitatively the clay content of an interval. The gamma ray logs signature records gradational changes in composition from shale to interbedded sandstone and shale to sandstone, or abrupt changes from shale to sandstone at either the upper or lower boundary of a sandstone unit. It thus provides a lithic profile of compositional changes within a unit although in a much more qualitative fashion than direct observation in surface sections.

The spontaneous potential curve reflects changes in permeability throughout a unit in a qualitative fashion. Maximum permeability maybe expected in shale-free sandstone units with coarse-grain size. Minimum permeability may be expected in shale successions with gradations existing between the two that reflect either gradual or abrupt changes. The log signatures generated by a spontaneous potential curve thus suggests textural changes in a unit related to grain size. Tightly cemented coarse sandstone intervals or intervals that are dry or contain hydrocarbons respond as impermeable rocks and diminish the effectiveness of this curve and the signature it produces.

Mechanical log signatures can be simply described as being either lens-, bell-, box-, or funnel-shaped or any combinations of these basic shapes (Figure 45). The lens-, bell-, and funnel-shapes are transitional patterns that reflect gradual changes in the grain-size from one depositional environment to another. Abrupt boundaries are indicative of sudden changes in the particle size and the depositional environment. Visher (1969) indicated that a tentative distinction can be made between marine and fluvial sand deposits, depending on whether the log signature is coarsening-upward (Funnel shape) or if it is fining-upward (bell-shape). Smooth spontaneous potential and gamma ray log signatures suggest homogeneity within a lithic interval and may represent periods of uniform depositional conditions. Serrated spontaneous potential and gamma ray log signatures suggest the disruptive influence of shaleness within a sand unit possibly due to changes in the water level, current type and sediment carrying power.

With the all of supporting evidence, mechanical log signatures can be useful in making predictions in sandstone sequences. It must be emphasized that log signatures are characteristics rather than diagnostic of environments and that additional information should always be used (Selley, 1965). The environmental interpretation from a spontaneous potential curve can be regarded as tentative until confirmed by detailed study of continuous core or observing the same strata in outcrop (Carring, 1970). Selley (1976) states that while some modern clastic sedimentary environments tend to generate characteristic vertical sequences of grain-size and sedimentary structures, no environment possesses a unique sequence, and that similar patterns can be generated by different environment.

Stratigraphic correlations, cross-sections, isopach, and log-signature maps in this report were compared to interpret the responsible depositional environment for the interval.

An **electrofacies** is a combination of log responses that characterize a particular rock and allows it to be separeted from other sediments (Serra and Abbott, 1982). Electrofacies correlate well log signitures with facies recognized in core when facies are defined on parameters measureable by logging tools, such as lithology or grain size patterns- assumming variations in porosity, minerallogy, and rock fluids are minimal (Cant, 1984). In this study, electro facies is used in large part as a synonym for log-curve shape (Figure 46).

Much of the literature on interpreting depositional environments from log-curve shapes is extremelly simplistic, attempting to equate particular shapes with depositional environments (Cant, 1984). No log-curve shapes is unique to a particular depositional environment. Electrofacies are best interpreted by considering them with context that includes sand-body geometry and orientation, and the character and distribution of the associated electrofacies (Bulling and Breyer, 1989; Figure 46). Besides scale is also an important factor in interpreting electrofacies. Bulling and Breyer (1989) summarized electrofacies in Reklaw 1 interval (Eocene) of South Texas which can be applied to other interval in any geological time period (Figure 46).

Various electrical well-log signitures that are to characterize specific depositional environments of ancient and/or modern delta system have been studied by many geologists. The deltaic environment represents a complex set of subcomponent depositional settings which experience a constructional and destructional history. Each of these depositional settings is characterized by its own SP and gamma ray log characteristics which are recognizable in the subsurface. these characteristics permit easy interpretation of the history of a specific delta lobe, and enable the petroleum geologists to predict new step-out wells and new locations for drilling (Sanena 1976; in Klein, 1980). Coleman (1980) illustrated various typical electrical log response in deltaic facies (Figure 48). Galloway (1968) illustrated electric log profiles of deltaic sandstone bodies base on the Tertiary fluvial-dominated Holly Springs delta system in the Gulf Coast, USA (Figure 47).

Klein (1980) illustrated several electric-log motifs and sequences to have diagnostic characteristics for identification of facies in deltaic systems. These are;

(A) electrofacies characteristics for river-dominated delta with mounh bar system (Figure 49),

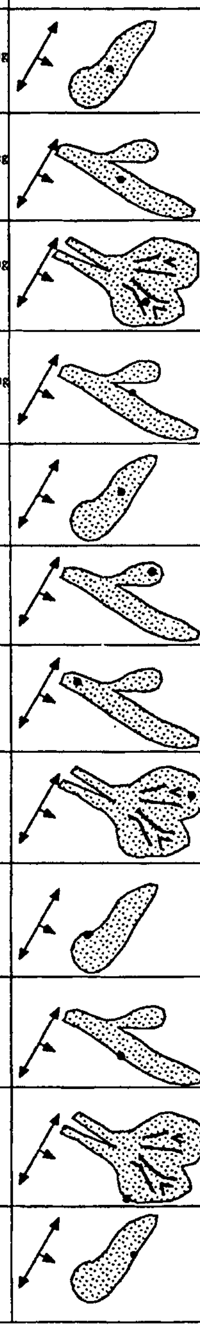
(B) series of electric-log motifs showing1,beach-barrier; 2, delta desutructive systems; 3, distributary overlying by crevasse splay fan; 4, distributary overlying by point bar; and 5, a series of crevasse splays in bay sequence (Figure 54),

(C) electric-log motif through delta destructive system (Figure 51) and idelized electric-log pattern of a delta destructive system (Figure 50),

(D) electric-log motifs of Figure 52 (upper) shows Sp log pattern that developes when a well penetrates a sequence of sediments of delta distributary abondonment followed by a transgressive destructive phase and later deposition of a transgressive barrier bar analogous to the Chandeleur Islands. The lower panel of Figure 52 shows the log pattern of delta progradation that was followed by seaward extention of the meandering alluvial upper delta plain over the distributary mount bar, (E) electric-logs of

Figure 53 characterize both channel progradation over barrier island system (lef…

channel progradation over a distributary mouth bar.



Name	Log Pattern	Lithology	Sand-Body Geometry	Associated Facies	Example	
BLOCKY	a.	•sand, 40-60' thick •abrupt contact with overlying and underlying shales		•tapered spike electrofacies (l) along basinward edge •simple spike electrofacies (i) along landward edge	Atkinson Field	BA BA BA
	b.	•sand, 40-60' thick •abrupt contact with overlying and underlying shales		•pinches out into shales along depositional strike	Hysaw Field	AX PR BA
	c.	•sand, 40-60' thick •abrupt contact with overlying and underlying shales		•surrounded by funnel-shaped electrofacies (h)	Hondo Creek Area	BAR THR FRO
SERRATE BLOCKY	d.	•interbedded sand and shale •abrupt contact with overlying and underlying shales		•on the margins of isopach thicks defined by blocky (b,c) electrofacies	Hysaw Field	MA BA SA
BELL SHAPED	e.	•fining-upward sequence, >35' thick •abrupt contact with underlying shales		•cuts isopach thick defined by blocky electrofacies (a)	Atkinson Field	TII
FUNNEL SHAPED	f.	•coarsening-upward sequence, 1-10' thick •gradational contact with underlying shales		•thin lobes extending from main isopach thick •passes into simple spike and tapered spike electrofacies as it thins and pinches out	Hysaw Field	CR
	g.	•coarsening-upward sequence, 15-40' thick •gradational contact with underlying shales		•lateral to blocky (b) and serrate-blocky (d) electrofacies • may be associated with simple spike electrofacies (j)	Flax Field	BA
	h.	•coarsening-upward sequence, 5-40' thick •gradational contact with underlying shales		•surrounds isopach thicks defined by blocky (c) and serrate-blocky (d) electrofacies •simple spike electrofacies (k) at seawardmost extent	Hondo Creek Area	DE SII
SIMPLE SPIKE	i.	•thin sand between shales •abrupt contact with overlying and underlying shales		•landward of blocky electrofacies (a)	Atkinson Field	WA
	j.	•thin sand between shales •abrupt contact with overlying and underlying shales		•lateral to blocky electrofacies (b)	Hysaw Field	PIN BA
	k.	•thin sand between shales •abrupt contact with overlying and underlying shales		•seaward margin of funnel-shaped electrofacies (h)	Hondo Creek Area	DIS DE SII
TAPERED SPIKE	l.	•thin sand •subtle coarsening upward from underlying shales		•basinward of blocky electrofacies (a)	Atkinson Field	SEA OF

*SP and resistivity logs from actual wells situated as indicated by dots. Double-headed arrows indicate depositional strike; single-headed a… paleoslope.

Figure 46. Electrofacies based on log-curve shape and sandstone body geometry and orientati… Reklaw 1 interval (Eocene), Texas (Bulling and Breyer, 1989)

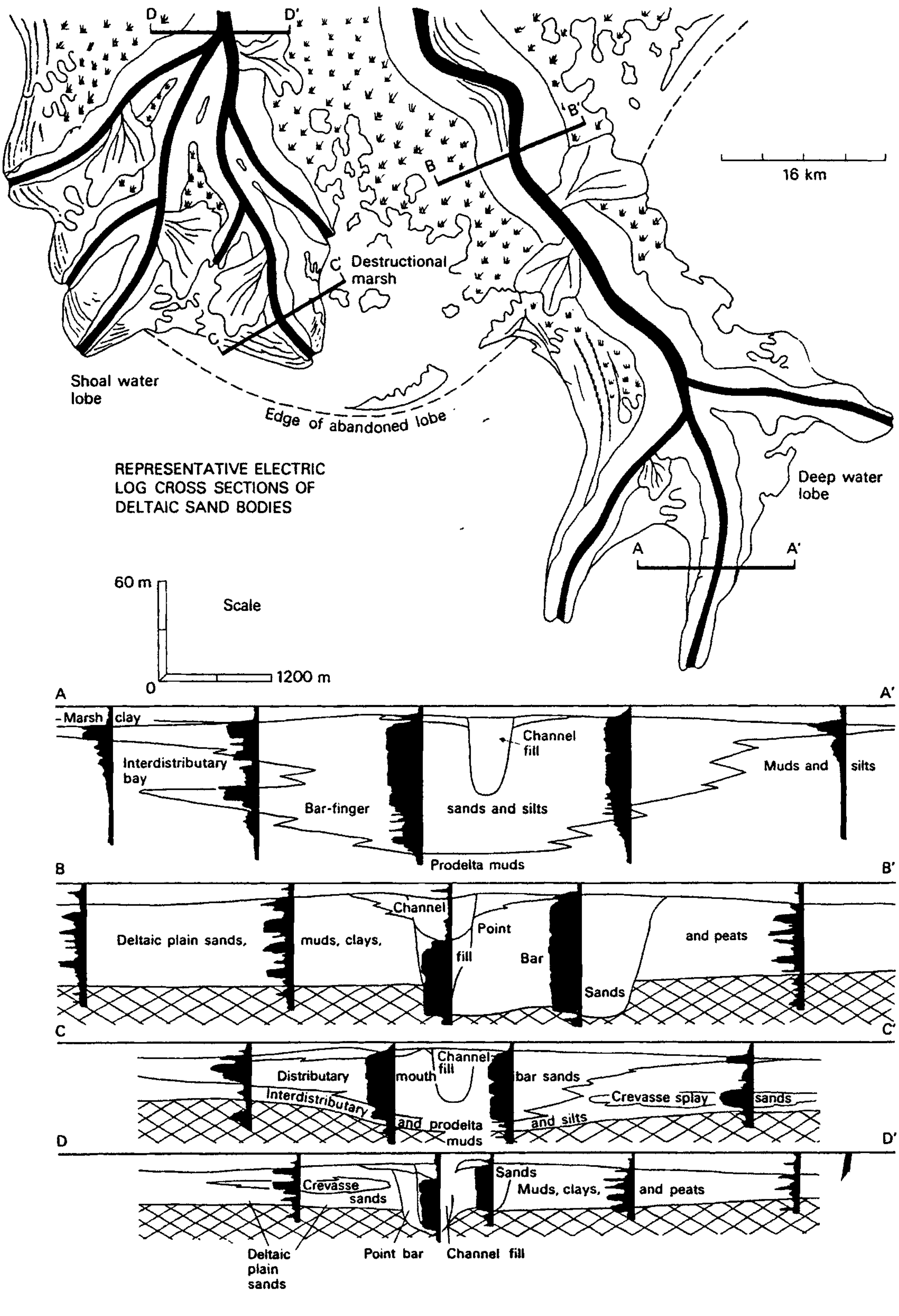


Figure 47. Electric log profiles of deltaic sandstone bodies based on the Tertiary fluvial-dominated Holly Springs delta systems in the Gulf Coast, USA (Galloway, 1968)

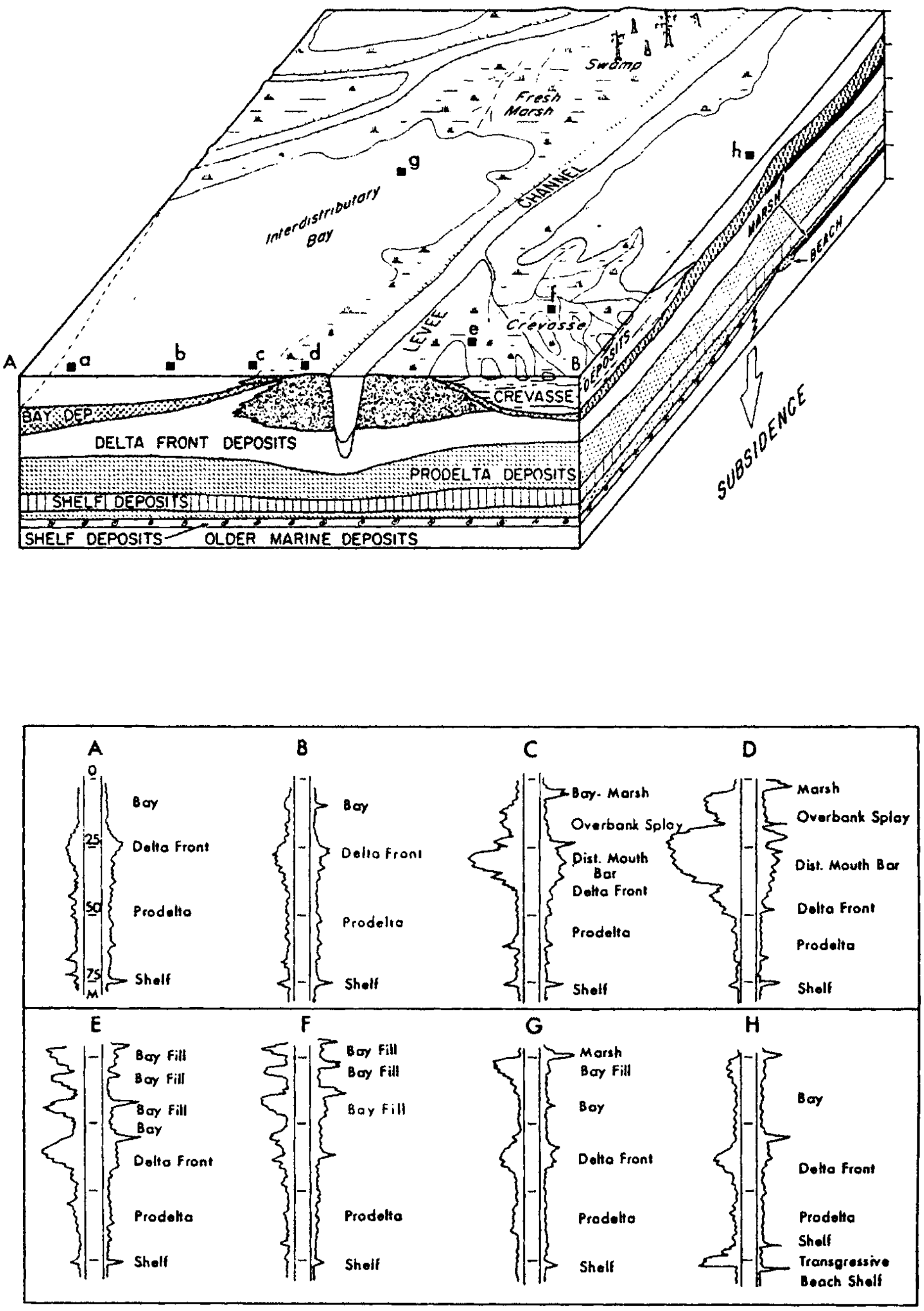


Figure 48. Schematic electrical log response in deltaic facies, and trhe block diagram showing vertical and lateral facies relationship of deltaic deposits (From Coleman, 1980).

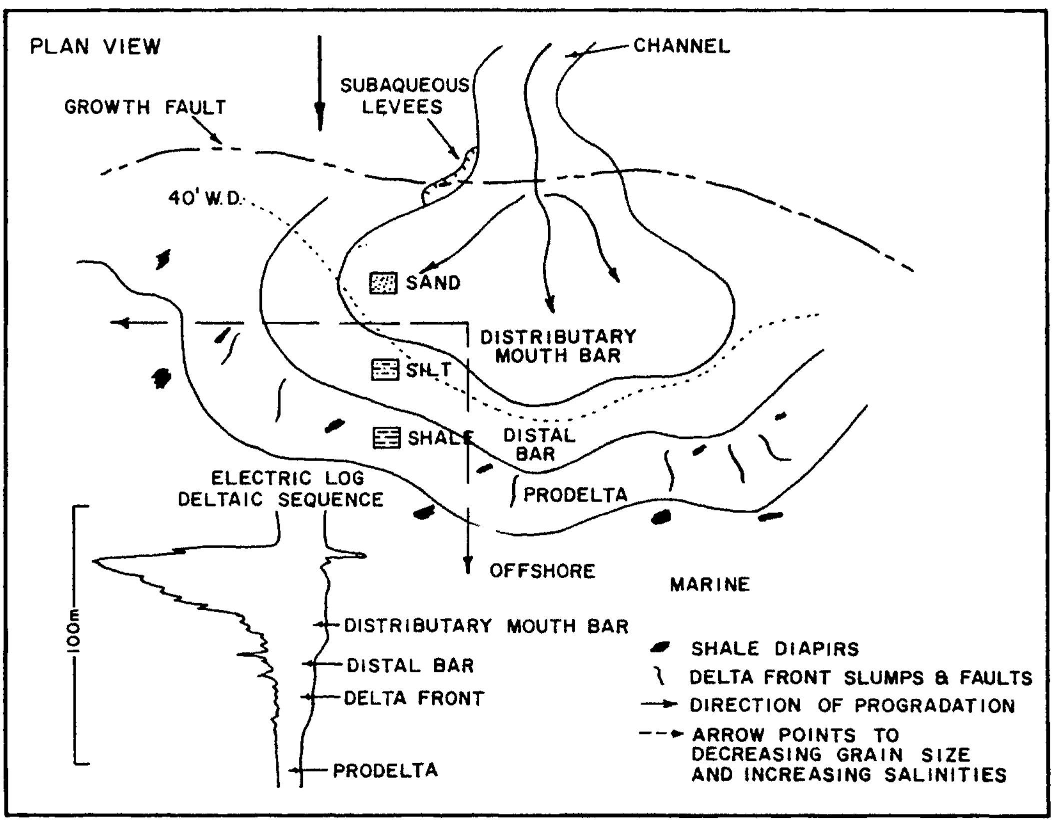


Figure 49. Plan view and diagnostic electric-log motif for river-dominated delta with distributary mouth bar system (From Redrawn after Saxena, 1976a,b).

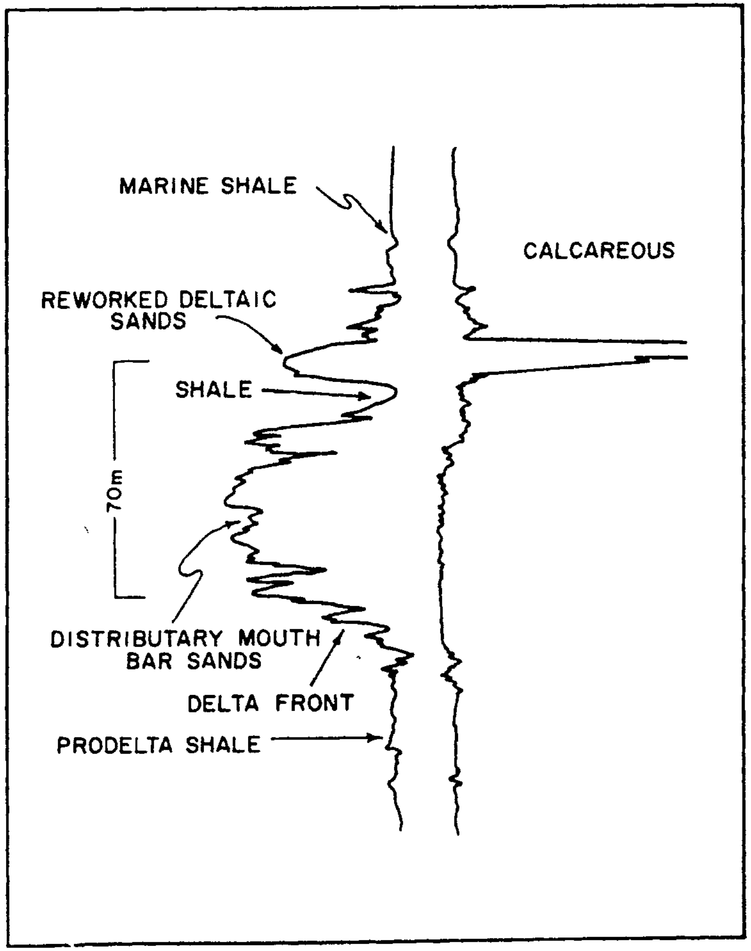


Figure 50. Idealized electric-log pattern of a delta destructive system (Redrawn after Saxena, 1976c).

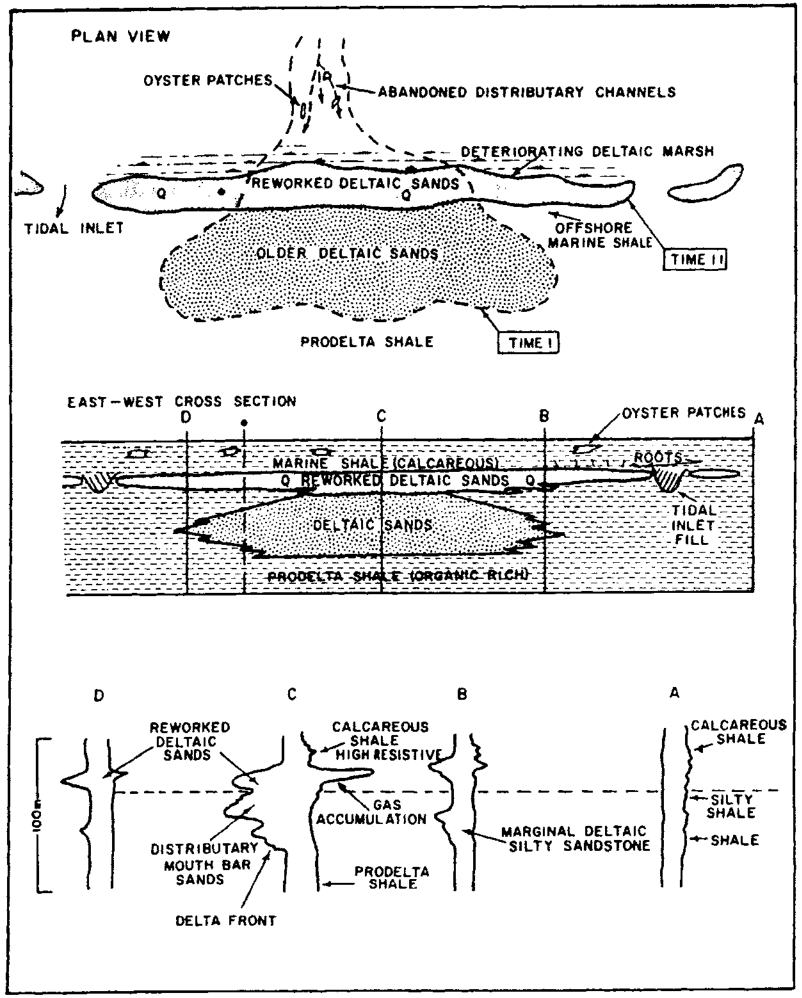


Figure 51. Map wiev, cross-section, and electric-log motif through delta destructive system. in plan view, *Time I* represents maximum constructional stage of distributary bar progradation, and *Time II* represents subsequent destructive stage with wave-dominated barrier transgressive sand overlying distributary. Cross-section shows both constructional and destructional stages, with lower panel showing electric-log features at specific places marked in cross-section (Redrawn after Saxena, 1976a, b, c).

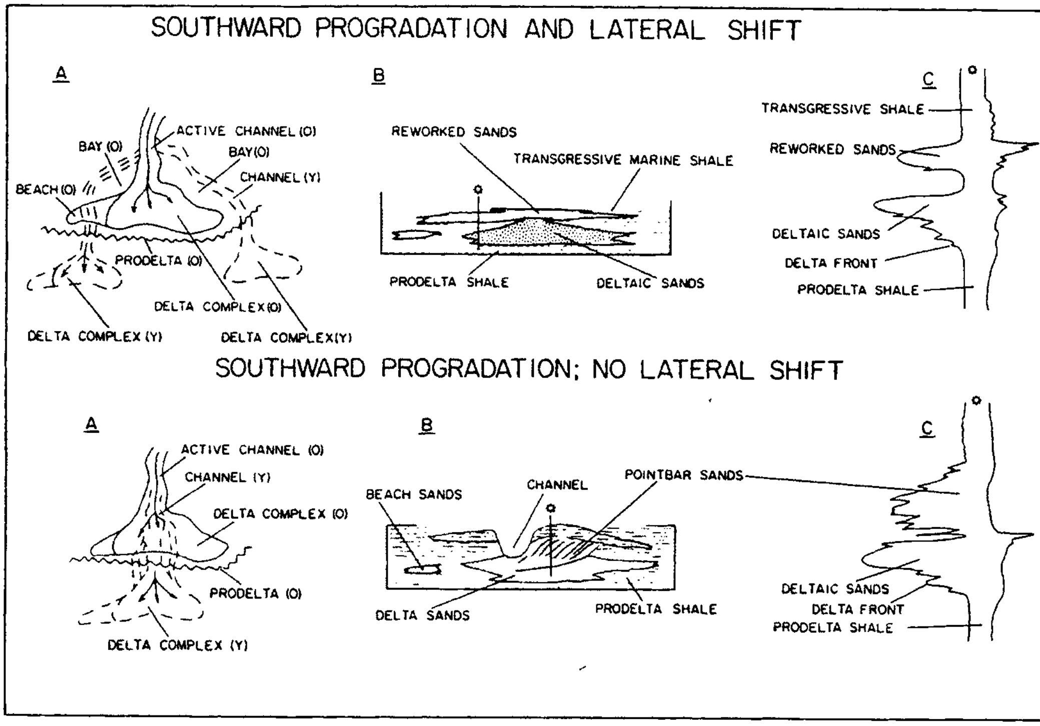


Figure 52. *Upper:* Suggested electric-log sequence representing southward progradation and lateral shifting leading to development of delta-destructive system. *A*, Plan, *B*, section, and *C*, electric-log. *Lower:* Suggested electric-log sequence showing history of southward progradation of delta system, without lateral shift. *A*, plan, *B*, section, and *C*, electric-log (Redrawn after Saxena, 1976c).

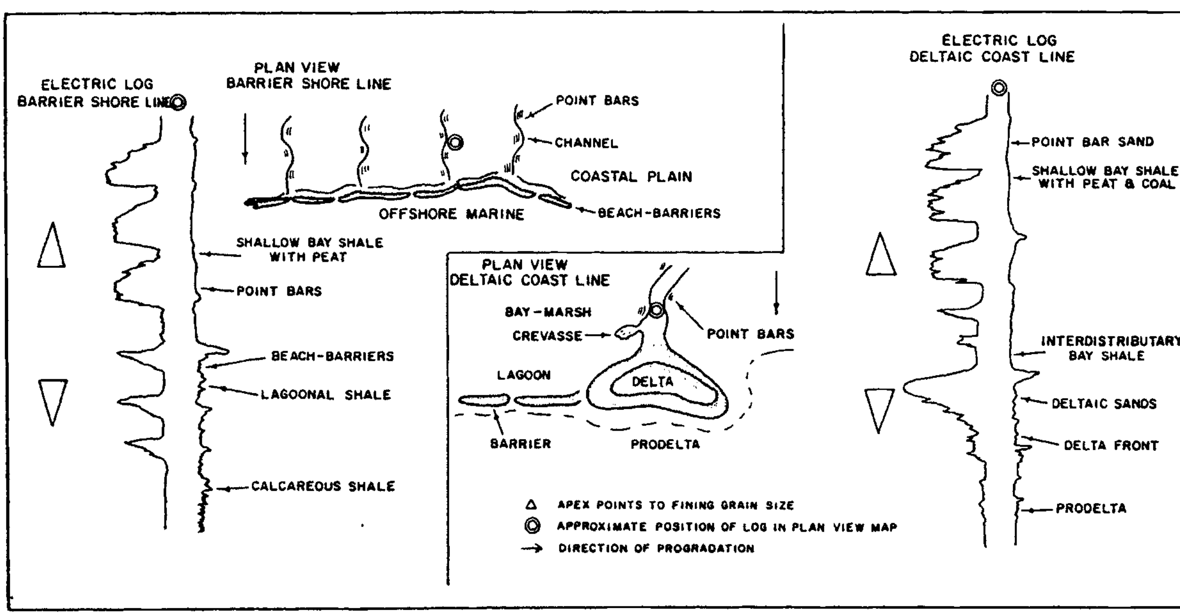


Figure 53. *Left:* Electric-log from well shown in map with double circle, indicating history of prograding barrier systems overlain by series of meandering channels. *Right:* electric-log from well shown in map with double circle, indicating a history of a distributary mouth bar being channelized and overlain by meandering point bar sands (Redrawn after Saxena, 1976c).

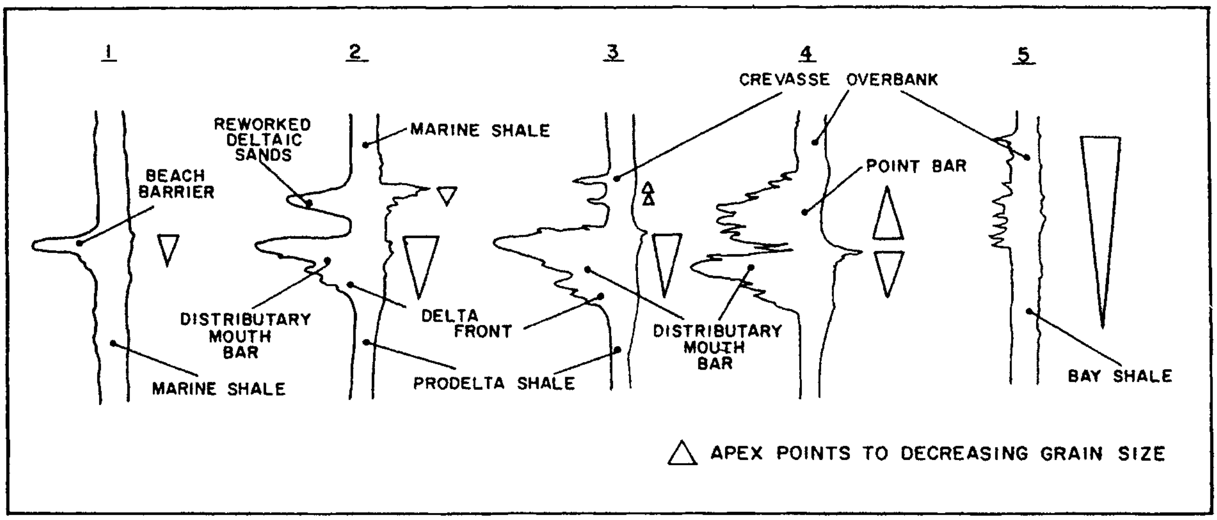


Figure 54. Series of electric-log motifs showing 1, beach-barrier; 2;Delta destructive systems; 3, distributary overlain by crevasse splay fan; 4, distributary overlain by point bar; and 5, a series of crevasse splay fans in bay sequnce (Klein, 1980).

THE SPIRO SAND

Introduction

The spiro sand occurs between the underlying Wapanucka limestone and the overlying Lower Atoka shale (Figure 12). In this study, Spiro sandstone is considered as the whole interval, combination of both Foster Channel sands and the overlying sheet-like sandstone of Lumsden (1971). The Spiro sand is stratigraphically equivalent to; the Orr sand (Cardneaux and Corbin, 1978), Zone 108 (Haley and Hendrix, 1972), and the Foster sand (Lumsden, et al., 1971). The Spiro sand is a well-developed sandstone complex within the subsurface of the Arkoma basin.

The stratigraphic cross-sections (Plate II) indicate that a complex relationship exists between the Spiro sand and the underlying Wapanucka limestone. Except the place underneath the sand distributary channel, there is a shale unit that separates the Spiro sandstones and the underlying Wapanucka limestone. The sub-Spiro shale is often absent along the Foster sand trends of the study area (Figure 55), and differentiating the units on the basis of mechanical well-log signatures is difficult. This is true where the Spiro is highly resistant or calcareous and the Wapanucka limestone is sandy.

In northeastern Oklahoma, Lumsden (1971) observed that the Foster sand (Spiro channel sandstone equivalent) partially truncates or completely replaces the underlying Wapanucka limestone, as it prograded toward the southeast, primarily along four parallel thickening (Foster channels). South and southeasterly directed delivery patterns of the

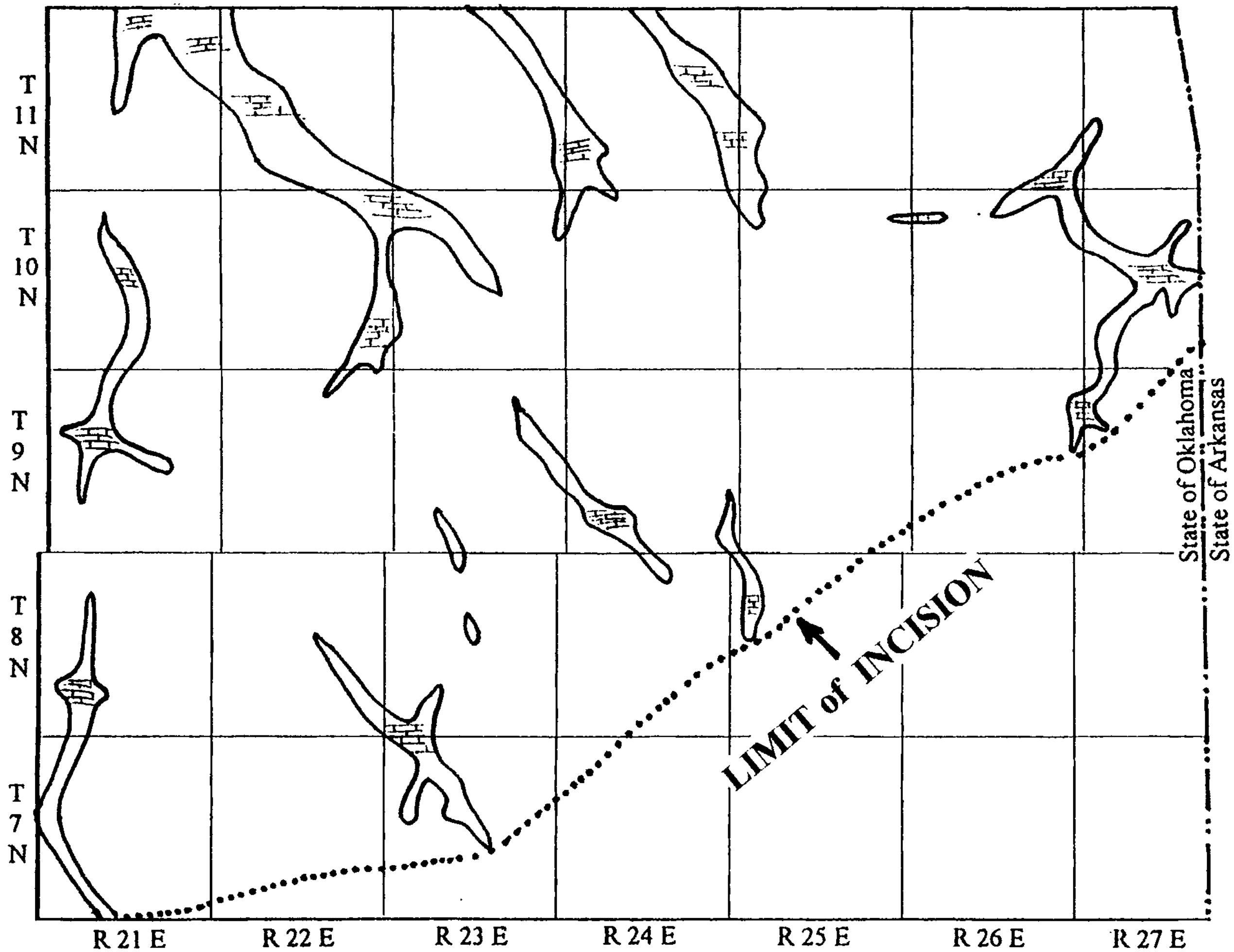


Figure 55. Generalized Sub-Spiro Shale Isolith Map. Solid contours indicate zero "0" thickness of sub-Spiro shale. Areas inside the closures (zero sub-Spiro contours) are localized places where Spiro directly overlies Wapanucka Limestone.

Spiro have been documented in the study area. Within the subsurface of this study area, the spiro sand is characterized by an overall blanket geometry which incorporates a series of four subparallel southerly and southeasterly trending thickening pattern (Figure 56). The primary basis for reconstruction of depositional environments are the isolith map pattern of the study area and the vertical sequences of depositional units indicated by the electric logs (Plate IV). Over much of the study area, control density was approximately one well per section (one square mile). this provided detail in mapping.

Geometry of the SPIRO sandstone

An isolith map of the Spiro was used to delineate the trends, geometry and lateral distribution of the sandstone (Plate II). The isolith values for the Spiro interval include both the Foster channels of Lumsden and other (1971) and the overlying blanket-like sandstone (Figure 57 and 58). Thickness was determined as the Spiro interval minus shale. Deflection of less than 80 A.P.I units, which is 20 A. P. I units less then shale base line value, were considered to show sandstone. This measurement was determined from avaraging the maximum and minumum A.P.I units of 50 gamma ray logs within the area of study. Also, seven stratigraphic cross-sections were constructed to show the lateral and vertical relationships with the Spiro sandstone, sub-Spiro Shale, and Wapanucka limestone (Plate VII and VIII). The datum for cross-sections is the lower boundary of the Wapanucka limestone that is present throughout the study area. A sub-Spiro shale isopach was constructed in order to locate erosional features in the shale and determine the locations of Spiro and Wapanucka contact (Figure55 or Plate II). Also, an isopach

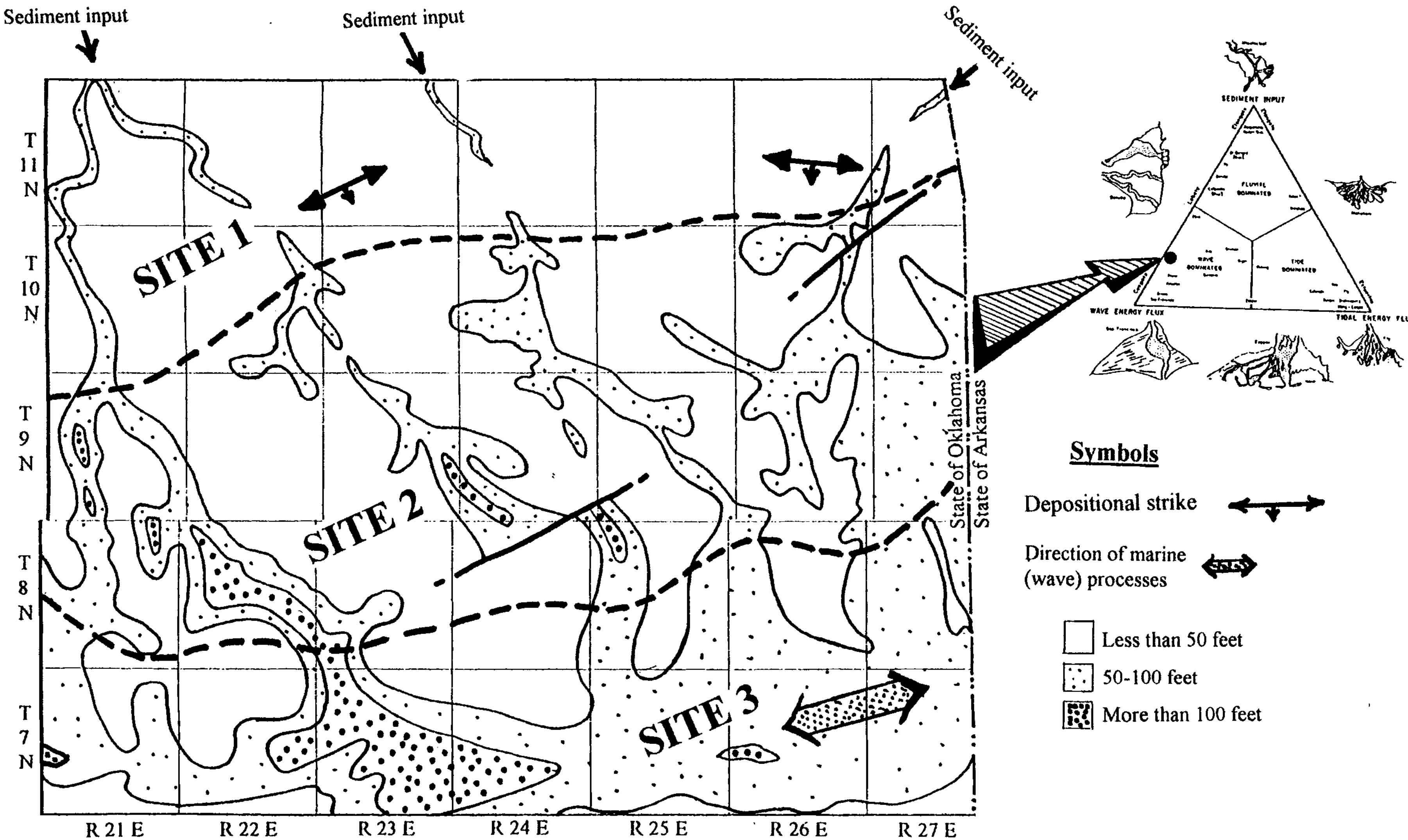


Figure 56. Generalized Isolith Map of the Spiro Sandstone and Its Depositional Sites

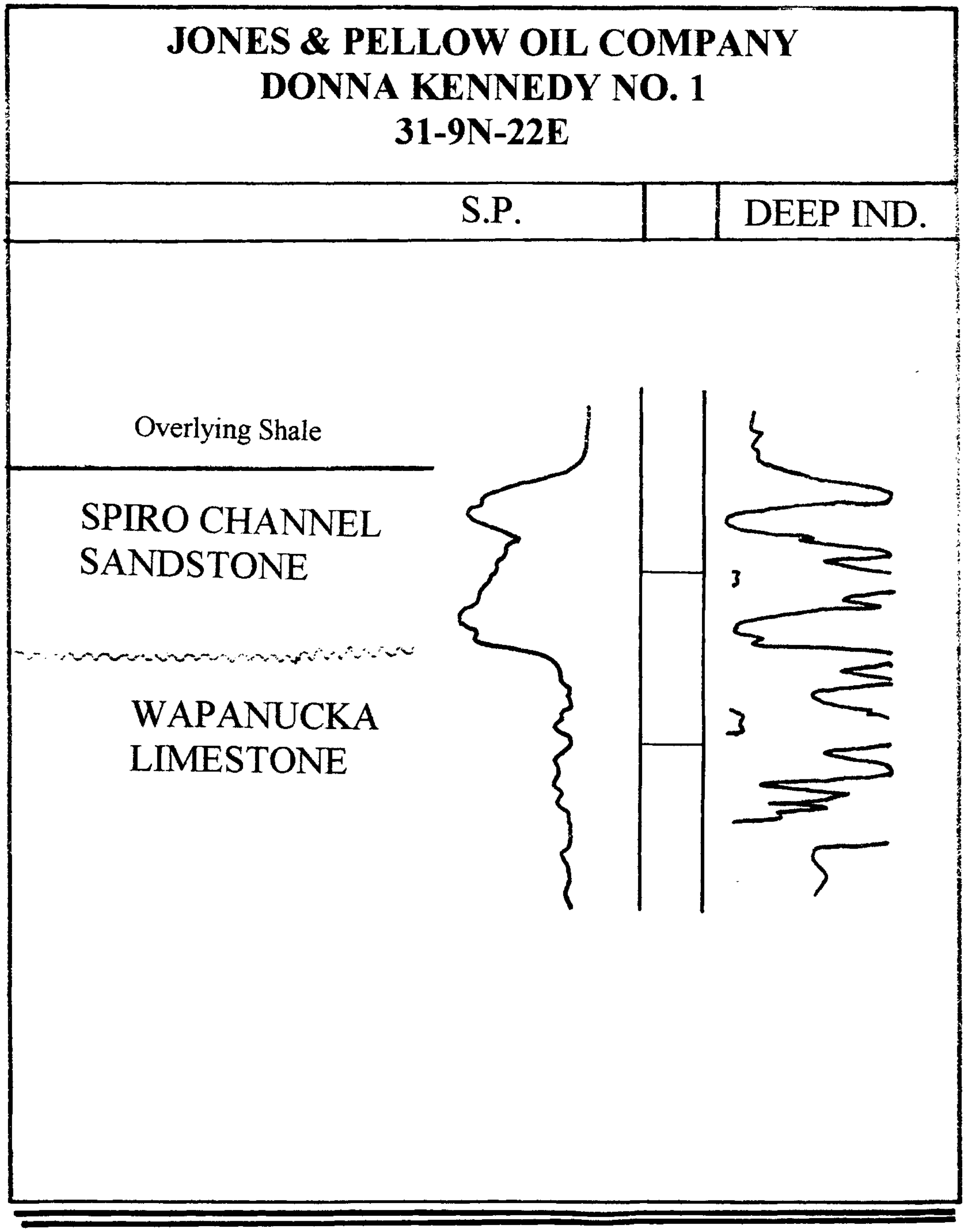


Figure 57. Electric Well-Log signature of Spiro Channel Sand deposits

map of top Spiro-base Wapanucka was constructed to show the overall geometry of the interval of interest (Plate VI).

Cross Section A-A

The cross section A-A' extends northwest-southeast across the western part of the study area (59). The cross section begins in section 19,11N,21E and continues to section 33,7N,23E (Plate VII). Spiro sandstone is approximately 10 feet thick in Sunray Mid-Cont. Oil #1Cozart Estate well, and has "0" feet thickness in Natol Petroleum #1 Chastain well. The major change in the Spiro thickness occurs IN Pan American Petroleum #1 Krisher Unit, Barrett Resources #4-21 Redd, and Mobil Oil #4 Lowery wells. the thickenings in these wells mark the Foster Channels.

The sub-Spiro shale unit is very thin in the southeast partion of the cross-section and thickens northwestward. In the Terra Resources No. 2-16 Blaylock, the sub-Spiro shale thickens up to 56 feet and thins southward, reaching "0" feet thickness in the Pan American Petroleum #1 Krisher Unit. In the Pan American Petroleum #1 Krisher Unit, Spiro directly overlies Wapanucka limestone with an erosional base that suggest the incision of the Foster channels. An overall thickening of Sub-Spiro shale is due to distance from main distributaries and indicate the lack of physical processes during deposition.

Cross section C-C'

Cross section C-C extends north to south across the eastern portion of the study area (Figure 59). The cross section begins with section 12,11N,26E and continues into section 23,7N,26E.

The Spiro sandstone is representedwith an almost constant thickness along the cross section, with a maximum of 65 feet (in Texas Oil & Gas #1 Bromly), and a minimum of 20 feet (in D-Pex Operating #1 Brown) (Plate VIII). The Spiro of this cross section cuts through the strike of the easternmost distributary channel of the Spiro delta.

Sub-Spiro shale is represented with a constant thickness of between 50-60 feet except in Texas Oil & Gas #1 Roberts "o" well where Spiro Overlies directly on Wapanucka limestone with an erosional base.

The Wapanucka thickens gradually from notrh to south. It is represented with a minimum thickness of 160 feet in Oxley Petroleum #1 Broocks well and a maximum of 235 feet in Midwest Oil #1 Robbs unit.

Cross section G-G'

Cross section G-G' extends west to east across the northern portion of the study area (Figure 59). The cross section begins in section 20,10N,21E and continues into 3,10N,27E.

The thickness of Spiro sandstone varies along the cross section (PlateVIII). The Spiro sandstone reaches a maximum thickness of 78 feet in Snee & Eberly #1-18 Reading well. this represents the westmost Spiro distributary channel . Spiro gradually thickens from Natol Petroleum #1 Chastain well with a thickness of "0" feet to Samson resources #1 Moore well with a thickness of 30 feet. there is a constant thickness (about 40 feet) of Spiro sandstone from Whitmar Exploration 313-1 Kelly to D-pPex Operating #1 Brown.

The thickness of sub-Spiro changes dramatically along the cross section. Sub-Spiro shale is characterized with "0" zero shale in four of the cross section, including Snee and Eberly #1-18 Reading; Samson Resources #1 Perry unit; Samson Resources #1 Moore; Stephens Production #2 Arkansas Valley Farms. Sub-Spiro shale reaches a maximum thickness of 55 feet in Natol Petroleum #1 Chastain well.

The thickness of Wapanucka limestone gradually thickens from west to east with a minimum of 75 feet in Snee and Eberly #1-18 Reading and a maximum of 220 feet in Stephens Production #2 Arkansas Valley Farms.

Sub-Spiro shale isopach

The thickness of sub-spiro shale varies throughout the entire area of study, with maximum of 142 feet (in 14, T8N, R26E) and minimum of 0 feet in several areas where Wapanucka had been subjected to the pre-Spiro erosion (Plate II and Figure 55). Localized, elongate southeast-northwest thining of the sub-Spiro shale geometries were mapped. Those thining of the sub-Spiro shale coincide (Figure 55) roughly with the channels mapped in the Spiro interval (Figure 56 and Plate I). The isolith map of the sub-Spiro shale presents some localized, west-east oriented thickening of the sub-Spiro shale in the southern portion of the study area (in 14, T8N, R26E; 23, T7N, R24E; 29, T8N, R24E.....). This may indicate lagoonal facies or mud injection into delta front sand.

Throughout the northern portion of the study area, the thickness of sub-Spiro shale tends to be constant except where Spiro channels reside.

Spiro net sand isolith

The net sand isolith of the Spiro is used to show the distribution and geometry of the Spiro sandstone reservoir rock (Plate I and Figure 56). Four prominent thickness trend were identified. Channel systems were mapped in the northern part of the study area. These channels coincide with the location of the Foster channels mapped by Lumsden and others (1971). The width and thickness of channels increase nouthwestward. Besides, an elongate sandstone body, mapped in the southern portion of the area, with northeast-southwest strike lies at the southmost ends of these four sand trends (Foster channels). The distribution of Spiro sandstone show a cuspate delta morphology in the study area (Plate I and Figure 56).

Although the Spiro is present throughout the area several distinct trends do exist, including four northwest-southwest trends. These trends were extracted from the both Spiro isolith and Sub-spiro shale isolith maps (Figure 55 and 55). Southwardly, all four trends extends into several branches and merge into a southwest-northeast oriented elongate sand body.

- *Trend 1* : This extends from T10N, R21E into T7N, R21E; R22E; R23E.
- *Trend 2:* This trend extends from T10N, R22E into T8N, R25E.

- *Trend 3:* This trend extends from T11N, R27E into T8N, R27E
- *Trend 4:* Trend four extends from T11N, R27E into T8N, R27E

Trends 1 and 2 merge northward into the township of T11N, R21E. The range of the sandstone trends in width is from 0.5 mile to 4 mile; length exceeds 30 miles.

Thickness

Although the spiro sandstone body display a generally continuous geometry over the whole area, the thickness of the spiro sandstone is not uniform throughout the study area (Figure 56). The thickness of Spiro is controlled by facies distribution. The Spiro thickens southward. At the axes and/or near areas around the distribution channels there are thick elongate, northwest-southeast oriented sand bodies, represented by a thick blocky wel-log signature. In Southern part of the area (south of T8N), the sandstone thickness tends to be relatively uniform-shaped. Thickness of Spiro sandstone (net sand) varies from 0 feet (in 23,T10N,R21E) to 162 feet (in 15,T7N,R23E).

The thickness of the sub-spiro shale is controlled by the spiro sandstone distribution; where the Spiro channels resides no or very thin layer of the sub-spiro shale is present (Figure 56 and Plate Ii). From the stratigraphic Cross-sections (Plate I, II), the thickness variation throughout the area can clearly be seen.

Boundaries

The lowermost Atokan in age spiro sandstone is overlain by the Atoka Formation and underlain by the Morrowan Wapanucka Formation or sub-Spiro shale (Figure 12).

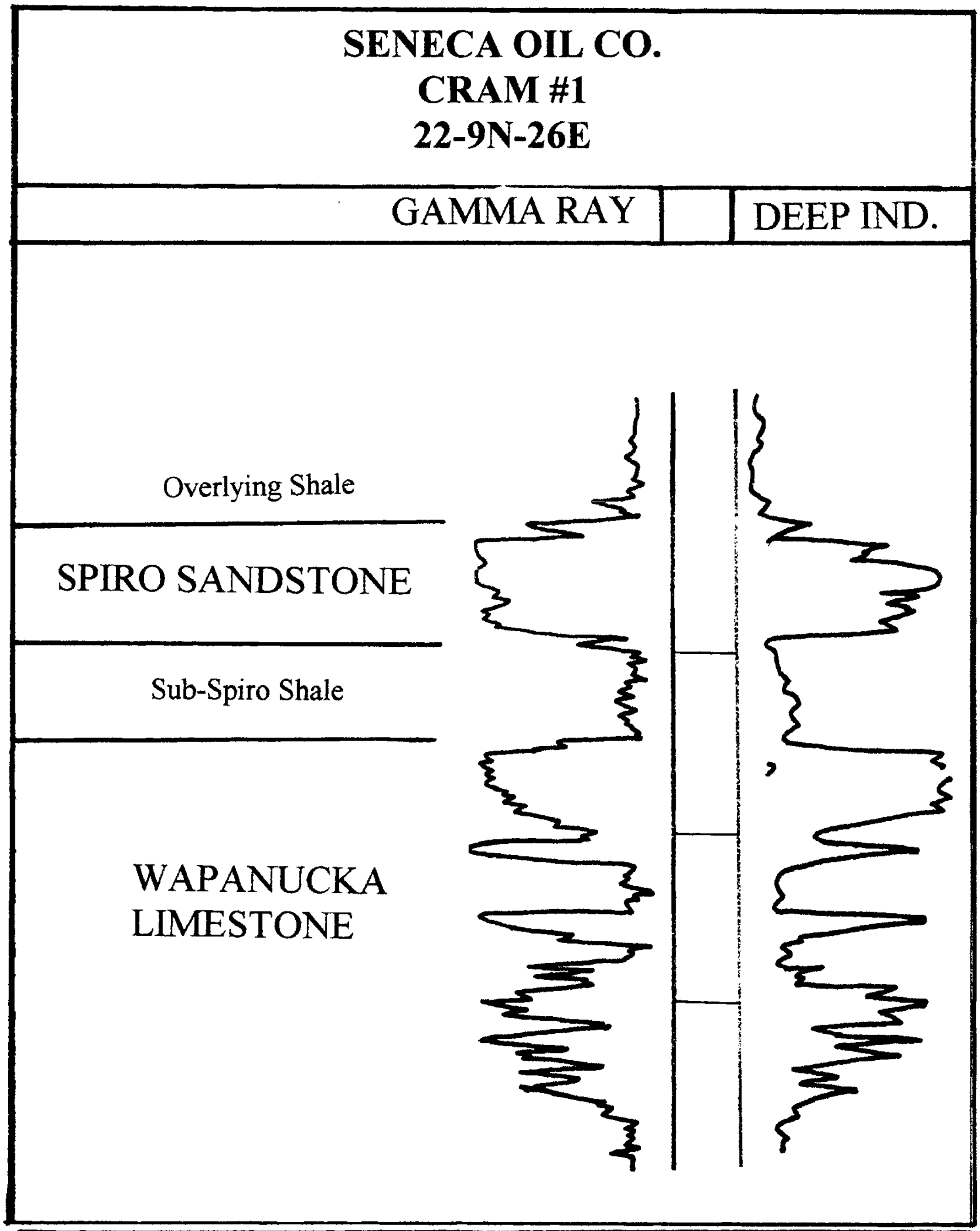


Figure 58. Type-log of the Spiro Sandstone for the Area of the Study.

Spiro has both a conformable and unconformable contact with the underlying Wapanucka. Grayson (1980) includes the Spiro as part of the upper sandstone member of Rowland (1974)'s Wapanucka Formation.

Throughout most of the study area sandstone of spiro is seperated from the Wapanucka by a shale unit,called sub-spiro shale (Plate VII and VIII). However, along the sandstone trends (Foster trends of Lumsden, 1971) the Spiro sanstone directly overlies the Wapanucka limestone with a scoured base where the sub-Spiro shale does not exist. Also, along the axes where the Spiro directly overlies the wapanucka, there is a significant thining in thickness of Wapanucka (Figure Plate VII and VIII). This is evidence, showing the incision of the Spiro channels into Wapanucka limestone in the study area, developed during a relative sea level fall.

It is believed that there is an unconformity, seperating the Spiro and the underlying wapanucka limestone in the Study area. Evidence of an unconformity is most clearly observed in along the axes of the Spiro Channel bases (Figure 57).

Depositional System of Spiro Sandstone

Introduction

Several workers have studied to describe the depositional systems responsible for the accumulation of Spiro sandstone. Lumsden et al. (1971) mapped four Foster channel trends and thought a high-destructive delta system was responsible for deposition of the Spiro sandstone. They also emphasized a north/northeast sediment source area which fed the Spiro delta system. Hess (1995) constructed an isolith map of Spiro of Latimer, southern Haskell, western LeFlore, and eastern Pittsburg counties that is adjacent to the area of this investigation (Figure 7). He indicated existance of an elongate sand trend (interpreted as deposits of a wave dominated delta system) perpendicular to the Foster channel and parallel to the paleo-shoreline (Figure 56). Gross et al. (1995) described Spiro in Pittsburg and Latimer counties as deposits of barrier islands based on the core studies and distribution of the Spiro sandstone (Figure 6). The result of previous studies and of this study were combined and a high-destructive cuspate delta morphology was extracted as Spiro net sand isoliths of each individual studies were compared. The electrofacies distribution of the study area also supports the existance of a high-constructive cuspate delta (Figure 61).

Mechanical log signatures indicate that the Spiro sandstone contains numerous thin, discontinuous and inconsistently-dispersed (Figure 61) shale intervals. These shales are locally developed. These shale intervals separate individual sand bodies. As shown in Figure (61j) thin shale unit seperates barrier bar sand from mouth bars.

The Spiro sands are characterized by a variety of mechanical log signatures. A predominance of either massive or multi-stacked serrated box-shaped well-log signatures characterize the Spiro sand that accumulated within thickening trends, and suggest that both fluvial and delta-marine distributary channels were active in these areas. The interbedded shales may indicate sporadic sediment supply due to channel abandonment followed by short-lived marine transgressions. A few intermitted serrated funnel-shaped well-log signatures also are recorded from the Spiro sand. These shaped are typical of distal delta-fringe sands, barrier bar sands or distributary mount bar sand, however, their isolation and infrequent occurrence suggest other alternatives. Brown et al., (1973) associated this pattern with delta distributary channels. Cardneaux (1978) concluded that the coarsening-upward profile of the Spiro sand represented reworked interdeltaic sand within a destructive delta system. In the northern part of the study area, occasional serrate bell-shaped well-log signatures (Plate III and Figure61) suggest a fluvially-deposited sand.

Depositionel Environments of the Spiro Sandstone

The isolith pattern of the Spiro sandstone displays a cuspate morphologic pattern (Figure 56 and Plate I). The isolith map were compared with those of previous studies and a depositional model pattern was constructed (Figure 56). A high-constructive cuspate delta model, called "The Spiro Delta", was interferred from the results.
Three distinct depositional environments within the Spiro Delta were distinguished in the study area based on the isolith map pattern (Plate I) and electrofacies (Plate III and Figure 56) of Spiro sandstone.

- *Site 1*: Alluvial Plain
- *Site 2:* Fluvial-dominated Delta plain
- *Site 3:* Wave influenced (dominated) Delta front

Each individuall depositional site is characterized by unique types of both (1) sandstone isolith pattern, and (2) electrofacies. However, the limit which seperates each depositional site is not certainly located. The locations are subjet to change.

SITE - 1 (Alluvial Plain):

This depositional site occupies the northern portion of the study area, including southeastern corner of Muscogee and southern portion of Sequoyan counties (Figure 2). Isolith map (Figure 56) of this Site includes four slightly sinisoidal sandstone trends, orienting south-north . These sands extend 11 to 3 miles along strike and reach width of less than .05 mile and maximum thickness of 78 feet (Plate I). The sand pinches out into shale along strike, updip, and downdip. Six electrofacies occur (Plate II and Figure 61) in the Site - 1, including blocky, simple spike, tapered spike, smooth linear, thight linear, blocky, and bell shaped. Most of log in this area have either bell shaped or simple spike log patterns. Bell shaped and blocky log patterns reach thickness of 70 feet. Whereas, simple spike only reaches a maximum thickness of 20 feet.

It is thought that the narrow sinusoidal sand strings were accumulated in a meandearing or anastomosed river of an alluvial plain. These sand strings are represented by both blocky and bell-shaped electrofacies. The shales alternating with thin sand intervals (simple and tapered spike, and tight linear electrofacies) along depositional strike probably represent flood plain or interchannel deposits.

SITE - 2 (Fluvial-dominated Delta plain)

This depositional site occur in the middle of the study area, covering most of the study area located in Leflore, Haskell, Seqouyah, and Latimer counties (Figure 2 and 56). The isopach map shows narrow ribbons of sands oriented northwest-southeast, perpendicular to depositional strike (Figure 56 and Plate I). These sand ribbons of Site - 2 are contiunity of sand strings from Site - 1, but are much wider and thicker. There are several small lobes branching from main distributaries. These are considered as "crevassse splays" of bay fill deposits (or interdistributary). Sand thickness ranges between 75-150 feet along axes of the sandstone ribbons, and reaches a maximum of 70 feet in the center of smaal lobeb (crevasse splays).

Five electrofacies are present in the producing Site - 2, including serrate blocky, smooth blocky , serrate funnel, bell-shaped, and simple spike. Logs adjoining the isopach thick usually have serrate-blocky pattern. Funnel-shaped electrofacies are found in centers lobes (Crevasse splays) joining the main isolith thick. Logs farther from the isopach thick of lobes usually display simple or teppared spike electrofacies (subtle coarsening-upward), having les than 20 feet thick, probably represents distal edge of a prograding crevasse splay. Bell-shaped electrofacies is often present in Site-2 and probably represents the channel abondonment, filled with mud. The presence of mud in these channel fills suggest absence of marine processes during channel filling. Serrate blocky electrofacies represents interfingering between distributary channels and laterally equavalent facies.

The sand body geometry and electrofacies distribution suggest the Site - 2 is a fluvial- dominated delta plain environment. The perfect isolith pattern of ribbon sands suggest lack of marine processes during deposition of these ribbons. Northern limit of the Site - 2 is not clear. However, it is considered to be transitional with the facies of the Site - 1.

The lower boundary of the Site - 2 coincides roughly with the limit of pre-Spiro deposition incision (Figure 55). the lower limit of Site - 2 has transitional relationship with its southern neighbour. Channel sand trends become wider as moves southward from Site - 1 to Site - 2 and isolith thick gains greater contour values.

SITE - 3 (Wave-dominated Delta Front):

This depositional site occurs in the south portion of the study area, including Leflore, latimer, and Haskel counties. The sand, in northern portion of this Site, forms several northwest-southeast oriented wide, channelized sand isolith thicks (Figure 56 and Plate I). These wide sand ribbons occur perpendicular to depositional strike. Farther southward (basinward) these sand thicks merge into an east-west oriented, thick, sheet-like lobate sand body. Along the wide ribbons, sand thickness reaches up to 162 feet. The sand thickness varies from 50 feet to more than 100 feet throughout the lowermost lobate sand body, oriented parallel to depositional strike. The sand ribbons of Site - 3 is the continuity of sand ribbons of Site - 2, but those of Site -3 are much wider and thicker and merge into one (Figure ..). Also, the channel sandstones of the Site - 2 thickens primarly at the expense of the underlying strata (sub-Spiro Shale), whereas the west-east oriented elongate sand body (barrier sand) thickens at the expense of overlying strata. There is an almost constant thickness of sub-spiro shale present in the Site - 3.

F our electrofacies occur in this Site: blocky, funnel, tight linear, and simple/tappered spike. The isopach thick ribbons are characterized by mostly funnel-shaped and sometimes blocky electrofacies. The lowermost west-east oriented, lobate sand isolith thick consists of mostly blocky and sometimes funnel-shaped electrofacies. Landward behind elongate sand body, it is charecterized by tight linear electrofacies which is characteristics of muddy lagoonal facies. The basinward edge of the isolith thick of elongate sand is represented by funnel shape electrofacies and distal edge of elongate sand body pinches out into pro delta mud as simple spike electrofacies. The southmost end of northwest-souteast oriented sand ribbons are usually represented by funnel-shaped electrofacies, indicating presence of shoreline progration.

Based on the isolith map pattern and the electrofacies distribution, the sand ribbons are considered ro be "bar-finger or mounth-bar " deposits. The east-west oriented, elongate sand body is thougth to be "barrier-bar or barrier-beach" deposits of a wave dominated delta front. The funnel-shaped and blocky electrofacies of Site - 3 represents delta front sheet-sand (barrier bar) and bar-finger sands, respectively. The updip and downdip shales are interpreted as lagoonal and shelf muds, respectivelly. Basinward taperred funnel and simple spike electrofacies represent transition zone from barrier-bar to pro-delta mud flat. Tigth linear electrofacies of landward distal part of the barrier bar represent lagoonal facies.

Depositional history of the Spiro sandstone

The stratigraphy of Spiro sandstone from mechanical well-log indicated that three major episodes occured from the initiation of to completion of the Spiro deposition

(Figure 60): (1) Incised valley formation, (2) formation of the Spiro delta , and (3) consequent drowning of the Spiro Delta.

The Wapanucka limestone deposited on a stable shelf during a rise in relative sea level in Late Morrowan time (Figure 60-1). The deposition of Wapanucka limestone was followed by a southward regression (Late Morrowan or Earliest Atokan) (Figure 60-2). Sutherland and Henry (1977) linked this regression with either tilting of the Arkoma shelf or the the development of a forebulge on the Arkoma shelf, establishing fluvio-deltaic environment. The lowering in relative sea-level restricted the limestone deposition of the Wapanucka formation and lead fine-grained clastic sediment influx to deposit the sub-spiro shale (Figure 60-2). Further fall in relative sea-level resulted in aerial exposure of the wapanucka and sub-Spiro formations and in developing incised valleys (Figure 60-2). Probably the magnitude and duration of the relative sea level-fall was not large enough to allow the incision to proceed further into the shelf (Figure 55). The Spiro Delta started developing near the end of lowstand (Figure 60-2 and -3).

Wave action was the primary agent in the formation of the Spiro delta. However, the wave processes were probably not effective enough to redistribute the delta mouth bar sands all over the delta plain which was dominated by fluvial-processes, depositing channel sandstones and interchannel mudy clastics. The delta prograded southward toward the basin center untill the end of lowstand. Probably the lowstand stage was not long enough to develope a complete regressional strata. The wave- dominated delta was followed by a rapid rise in relative sea-level which reworked the sediment of Spiro delta northward, forming a sheet-like sand body (Figure 60-4). The lateral extent of reworked delta sand was limited to the north. Because sea-level rise (transgressive system tract) was quite fast and short-lived which did not give enough time to marine processes to rework the sediments, previously deposited in the Spiro Delta, and consequently the Spiro Delta was drowned and the overlying shale deposited over the top of the deltaic sediments (Figure 60-5).

PETROLEUM GEOLOGY

The Arkoma basin is essentially a dry-gas province with about 25 gas-producing zone (Figure), ranging in age from early Desmoinesian (Penyslnvanian) to Simpson (Ordovician). Most of the gas production in the Arkoma basin is from lenticular , fine-grained sandstones within the Atoka Formation. The major productive zone in the basin is the Spiro sandstone of the Atoka Formation. Although traping in some instances is structural, previous studies indicate that trapping is primarly stratigraphic in the Arkoma Basin.

Hydrocarbon exploration in the basin first started with coal seams. The surface exploration for coal seams in the Arkoma basin provided maps which located the closed anticlines and encouraged wildcatters to search and drill for oil and gas (Branan, 1966). Bartlett (1966) stated that the first known test well in the Arkoma basin was drilled to 1,400 feet in 1887 by Choctaw Oil & Gas Co. near downtown Fort Smith, Arkansas. Natural gas first was discovered on a large surface anticline (the Hartford anticline) in the Arkoma basin at Mansfield, Sebastian County, Arkansas in March 1902.

The study area consists of several large and small gas fields. Notrh half of the Red Oak-Norris, Northeast half of the Kinta, Pteau-Gilmore, Cameron, Bokoshe, Cedars, Peno, and Keto gas fiels lie in the Study area. Among those Kinta and Red Oak-Norris gas field are major natural gas producers of the Arkoma basin.

Kinta gas field covers most of the study area. The field was discovered in 1951 by the Superior Oil Company. Production is from both Spiro and Cromwell sanstones. Although anticlines are dominant structural features of the field, the accumulation appears to be controlled stratigraphically by porosity of both the Spiro and Cromwell sands (Woncik, 1968).

Red Oak-Norris is situated directly under the surface expression of the Brazil anticline, located south of the study area. The field was first discovered discovered in 1912 by Gladys Belle Oil Co. by drilling a well in section 10, T6N, R21E. Desmoinesian Hartshorne sandstone is the main producer. The Spiro sandstone is a secondary producer in the Red Oak-Norris field.

Trapping Mechanism

Nearly all of the early wildcat discovery wells in the Arkoma basinwere drilled on surface anticlines. However, much of the gas production is from stratigraphic traps which have little or no relation to surface structural features, especially in Spiro sandstone (Brannan, 1968). Diagenesis has strongly influenced porosity development in the basal Atokan Spiro sandstone (Lumsden and others, 1971). Gross et. al., (1995) stated that depositional environment and termal maturity of Spiro sandstone are the key regional controls on production.

Possible trapping mechanism of the study are is considered to be a combination of both stratigraphic and structural features. Vertical and Lateral facies changes, diagenetic porosity development may play important role in the formation of stratigraphic entrapment. Folding and faulting are key mechanisms of structural entrapment. Figure 62 represents possible trapping mechanism of the Spiro sandstone reservoir in the study area.

Future Locations

Combined facies distribution, porosity distribution, and regional and local structural styles of the Spiro sanstone may lead new discoveries of natural gas in the study area. Wapanucka limestone may produce a good reservoir rock where pre-Spiro incision developed karstic features and where dolomitization was developed along normal faults and axes of folds.

CONCLUSIONS

A reconstruction of depostional system has been attempted for Spiro sandstone in the subsurface of the northeastern Oklahoma. The reconstruction is based on evidence presented by means of stratigraphic cross-sections for corelation purposes, isopachs maps, electrofacies distribution to emphasize the geometry and distribution of thickness trends, the mechanical well-log signature map to illustrate the vertical andlateral variations within the lithic successsion oh Spiro sandstone.

Spiro is belived to have accumulated in a high-destructive wave-dominated delta system. Fluvial systems of the northern Oklahoma and Arkansas structural platform supplied the delta system that actively prograded into the Arkoma basin to the south. The shale interval that separate Spiro from younger sandstone (Patterson sansstone) marks the termination of Spiro delta growth and the beginnig of wide spread marine transgression to the north.

The Spiro is thickest in the southern portion of the study area. Four thickness trends were map which coincides with the Foster channels of Lumsden et. al. (1971). A east-west oriented sand body was mapped in the southern portion of the study area. This elongate sand body is interpreted as wave-dominated delta front sediment (barrier bar, barrier islands). A combination of both electrofacies studies and sand isolith pattern suggets that Spiro deposition occurred in three different depositional regions;

(1). Alluvial plain

(2) Fluvial-dominated delta Plain

(3) Wave-dominated delta front.

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