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REGIONAL CHARACTER OF THE LOWER TUSCALOOSA FORMATION DEPOSITIONAL SYSTEMS AND TRENDS IN RESERVOIR QUALITY

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by

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Dedication

To Julie, Trevor, and Lily

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Regional Character of the Lower Tuscaloosa Formation Depositional Systems and Trends in Reservoir Quality

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For decades the Upper Cretaceous Lower Tuscaloosa Formation of the U.S. Gulf Coast has been considered an onshore hydrocarbon play with no equivalent offshore deposits. A better understanding of the Lower Tuscaloosa sequence stratigraphic and paleogeographic framework, source-to-sink depositional environments, magnitude of fluvial systems, regional trends in reservoir quality, and structural influences on its deposition along with newly acquired data from offshore wells has changed this decades-long paradigm of the Lower Tuscaloosa as simply an onshore play.

The mid-Cenomanian unconformity, underlying the Lower Tuscaloosa, formed an extensive regional network of incised valleys. This incision and accompanying low accommodation allowed for sediment bypass and deposition of over 330 m thick gravity-driven sand-rich deposits over 400 km from their equivalent shelf edge. Subsequently a transgressive systems tract comprised of four fluvial sequences in the Lower Tuscaloosa Massive sand and an overlying estuarine sequence (Stringer sand) filled the incised valleys. Both wave- and tide-dominated deltaic facies of the Lower Tuscaloosa are located at the mouths of incised valleys proximal to the shelf edge. Deltaic and estuarine depositional environments were interpreted from impoverished trace fossil suites of the

Cruziana Ichnofacies and detailed sedimentological observations. The location and trend of valleys are controlled by basement structures.

Lower Tuscaloosa rivers were 3.8m – 7.8m deep and 145m – 721m wide comparable to the Siwalik Group outcrop and the modern Missouri River. These systems were capable of transporting large amounts of sediment indicating the Lower Tuscaloosa was capable of transporting large amounts of sediments to the shelf edge for resedimentation into the deep offshore.

Anomalously high porosity (>25%) and permeability (>1200md) in the Lower Tuscaloosa at stratigraphic depths below 20,000 ft. are influenced by chlorite coating the detrital grains. Chlorite coatings block quartz nucleation sites inhibiting quartz cementation. Chlorite coats in the Lower Tuscaloosa are controlled by the presence and abundance of volcanic rock fragments supplying the ions needed for the formation of chlorite. Chlorite decrease to the east in sediments derived from the Appalachian Mountains. An increase in chlorite in westward samples correlates with an increase of volcanic rock fragments derived from the Ouachita Mountains.

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Chapter 1: Introduction, Regional Setting, Summary of Data and Methods, and Objectives

INTRODUCTION

The Lower Tuscaloosa Formation (Mid-Cenomanian – Turonian) has been a prolific hydrocarbon producer in the central and eastern Gulf Coastal Plain of the southern United States since the 1940's (Hansley, 1996). Despite this long history of production it is thought to have significant remaining potential (Dowty and Moody, 1991; John et al., 1997; Meylan, 1997; Dubiel and Pitman, 2002; Mancini and Puckett, 2003; Pitman et al., 2007). Production to date has been entirely from the fluvial-deltaic to nearshore marine facies of the Lower Tuscaloosa, in the modern onshore. Until recently, these onshore deposits were thought to be the extent of the Lower Tuscaloosa Formation; however recent exploration of new plays in the Tuscaloosa-equivalent section of the offshore Gulf of Mexico has revealed Tuscaloosa-age gravity deposits in both slope and basin plain paleo-positions. These exploration efforts, proving the existence of Tuscaloosa-aged clastics in the deepwater Gulf of Mexico, greatly increases what was traditionally thought to be the depositional extent of the Lower Tuscaloosa Formation and shifts the paradigm of the Tuscaloosa as solely an onshore play. This paradigm shift necessitates a new look at the Lower Tuscaloosa Formation deposits and processes responsible for its development. As such, this dissertation aims to clarify the: (1) Sequence stratigraphic and paleogeographic framework of the Lower Tuscaloosa Formation and the influences on its sediment distribution and depositional environments; (2) Distribution of Lower Tuscaloosa facies and depositional systems from source-to-sink; (3) The size and nature of the Lower Tuscaloosa Fluvial systems that provided sediment for the deepwater Tuscaloosa interval; and (4) Regional distribution of chlorite in the Lower Tuscaloosa Formation with insights into the controls of chlorite occurrence and implications to reservoir quality.

In clarifying these major questions for the Lower Tuscaloosa Formation larger stratigraphic, sedimentologic, and petrographic questions were answered that can be applied to formations with similar characteristics. Additionally, this dissertation enables a better understanding of Late Cretaceous geology particularly in the Gulf of Mexico. This is the first highly detailed regional paleogeographic study of the Lower Tuscaloosa Formation documenting the regional distribution and trangressional nature of facies and the first to apply an incised valley/estuarine model to this formation. This model documents times of sediment bypass into the basin and can be applied to other incised valley systems for the recognition and better prediction of downdip facies. The detailed paleogeography and petrography documented in this study indicate that prior to the Tertiary much of the sediment deposited into the Gulf of Mexico was derived from the Appalachia Mountains.

This dissertation also shows that basement structures are critical to an understanding of depositional trends in the onshore and nearshore Tuscaloosa Formation. Basement structures influenced the trend on incised valleys, dammed sediments behind prominent highs, and interacted with coastal forces to influence wave and tide regimes controlling the distribution of facies. It follows, that these structures may have been important during the deposition of other formations in the Gulf of Mexico and may aid in a better understanding of their distribution of facies. This dissertation also highlights the utility of combining detailed sedimentology with ichnology to define facies and depositional environments.

Understanding the size and capacity of fluvial systems can help predict the presence of downdip deposits. Morphometric estimates can be used to determine outcrop and modern analogs to better understand the amount of sediment fluvial systems are capable of moving. Using measurements of both cross sets in core and channel deposits interpreted from wireline logs lead to similar estimates in channel size.

The formation of chlorite rims inhibits quartz cement growth by blocking quartz nucleation sites ultimately leading to better reservoir quality. It is now more apparent that the presence of chlorite coating grains in the Lower Tuscaloosa Formation is controlled by provenance and sediment derived from the relatively volcanic-rich Ouachita Mountains provides the necessary ions for good chlorite rim growth. It is likely that provenance plays an important part in many chlorite bearing formations. This understanding of provenance control can greatly aid in the prediction of areas with good reservoir quality.

REGIONAL SETTING

The Tuscaloosa Formation, located in the central to eastern Gulf Coast of the United States (see Figure 1.1 for map of study area), was deposited in the Late Cretaceous from the mid-Cenomanian through Turonian (95 Ma through 88.5 Ma) (Mancini et al., 1987; Sohl et al., 1991). At the time of the Tuscaloosa Formation deposition, the Appalachian Mountains were thought to be topographically high, as were the Ouachita Mountains of present-day Arkansas. The Western Interior Cretaceous Seaway opened to the northwestern Gulf of Mexico, forming an extensive carbonate platform across what is present-day central and south Texas (Figure 1.2) (Goldhammer and Johnson, 2001). Several major regional basement structures and related paleogeographic highs appear to have been active or high during deposition of the Tuscaloosa Formation (Jackson and Laubach, 1991; Stephens, 2009).

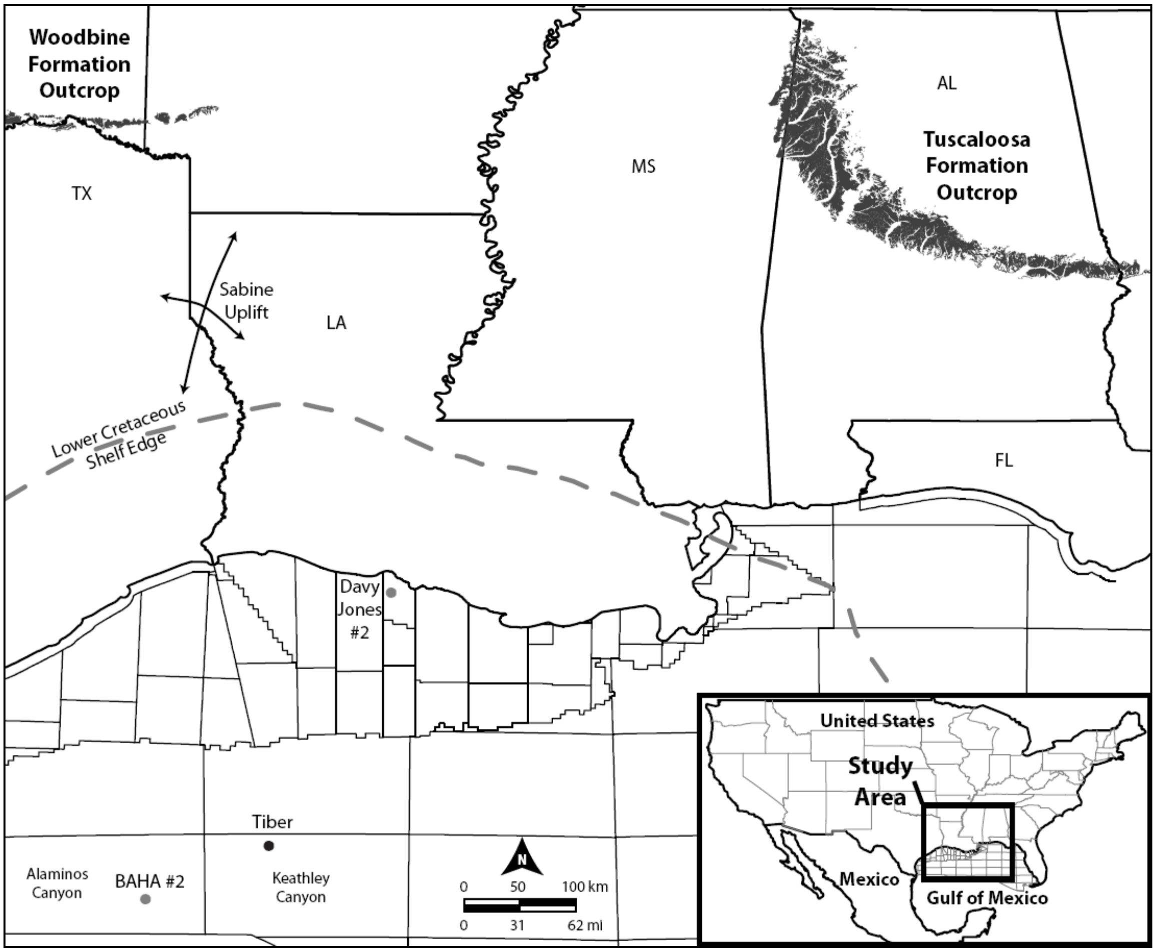

Figure 1.1. Base map of the study area of this dissertation showing the location of the Sabine Uplift which marks the western limit of the study area, outcrops of the Tuscaloosa and Woodbine formations, lower Cretaceous shelf edge, offshore blocks, and the locations of the three deep well penetrations (BAHA #2, Davy Jones #2, and Tiber) of Tuscaloosa age sediments discussed in the text.

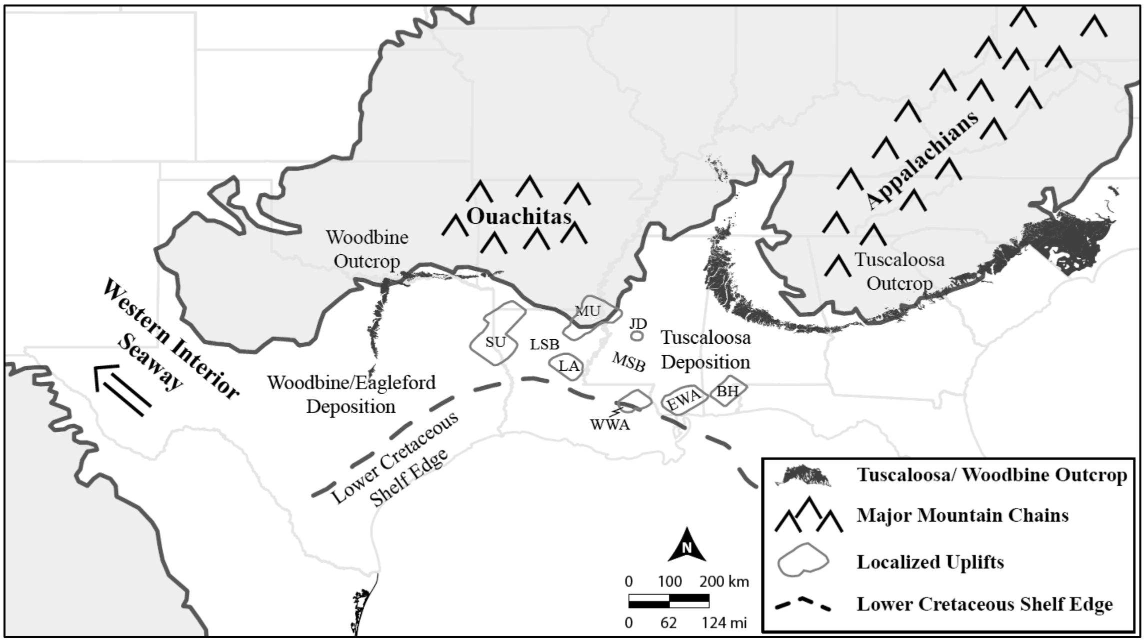

Figure 1.2. Map showing the paleogeography near the maximum flooding event during the Middle Tuscaloosa Formation. Paleo-shoreline is derived from Blakey's 85-Ma paleogeographic map (2011). Gray areas denote exposed land. Note the location of two main source areas of Tuscaloosa sediments; the Appalachian, and Ouachita Mountains, and the Gulf of Mexico's connection to Western Interior Seaway of North America. Modern day outcrops of the Tuscaloosa and Woodbine/Eagleford are located to the east and west of Sabine uplift, respectively. Structural highs important during deposition of the Tuscaloosa are shown and include the La Salle arch (LA), west Wiggins arch (WWA), east Wiggins arch (EWA), Baldwin high (BH) Jackson's dome (JD), Monroe uplift (MU), and Sabine uplift (SU). Basins hosting deposition of Lower Tuscaloosa sediment include the North Louisiana salt basin (LSB) and Mississippi salt basin (MSB) (Stephens, 2009).

Lithostratigraphy/Paleogeography

In Mississippi and Alabama the Tuscaloosa Formation has been divided into the Lower, Middle, and Upper Tuscaloosa (Mancini et al., 2008) (Figure 1.3). The Lower Tuscaloosa Formation in Mississippi and Alabama is divided into three informal members—the Massive, Stringer, and Pilot sands (Mancini et al., 1980), which are chronostratigraphically equivalent to the entire Tuscaloosa Formation in Louisiana. The Massive sand is the lowermost member overlain by the Stringer then Pilot sands (McGlothlin, 1944; Karges, 1962; Chasteen, 1983; Hansley, 1996; Mancini and Puckett, 2005). The Stringer sand of the Lower Tuscaloosa and the Upper Tuscaloosa pinch out toward the south and west and are not present in Louisiana. The Tuscaloosa Formation unconformably overlies the Washita/Fredricksburg Groups, the Dantzler Formation, and undifferentiated Paleozoic-aged sediments. The Tuscaloosa Formation is unconformably overlain by the Eutaw Formation in Alabama and Mississippi and conformably overlain by the Eagle Ford Shale in Louisiana (Figure 1.3). The Tuscaloosa Formation is thought to be temporally equivalent to the Woodbine Formation located west of the Sabine uplift in Texas (Harrison, 1980; Chasteen, 1983; Klicman et al., 1988).

The onshore Lower Tuscaloosa Formation was deposited in a fluvial-deltaic to nearshore marine system (Karges, 1962; Gruebel, 1985; Minter et al., 1992). The Massive sand of the Lower Tuscaloosa is a medium- to coarse-grained, fluvial-deltaic sandstone (Sohl et al., 1991), which contains abundant chert pebbles and small amounts of chlorite that coat the detrital grains (Thomson, 1979; Hearne and Lock, 1985; Hamlin and Cameron, 1987; Genuise, 1991; Hansley, 1996; Ryan and Reynolds, 1996; Weedman et al., 1996). The fluvial facies of the Massive sand is interpreted to contain elements of both braided- and meandering-stream systems, as evidenced by upward-fining cycles found in gamma logs and core. The lower part of the Massive sand, which is more

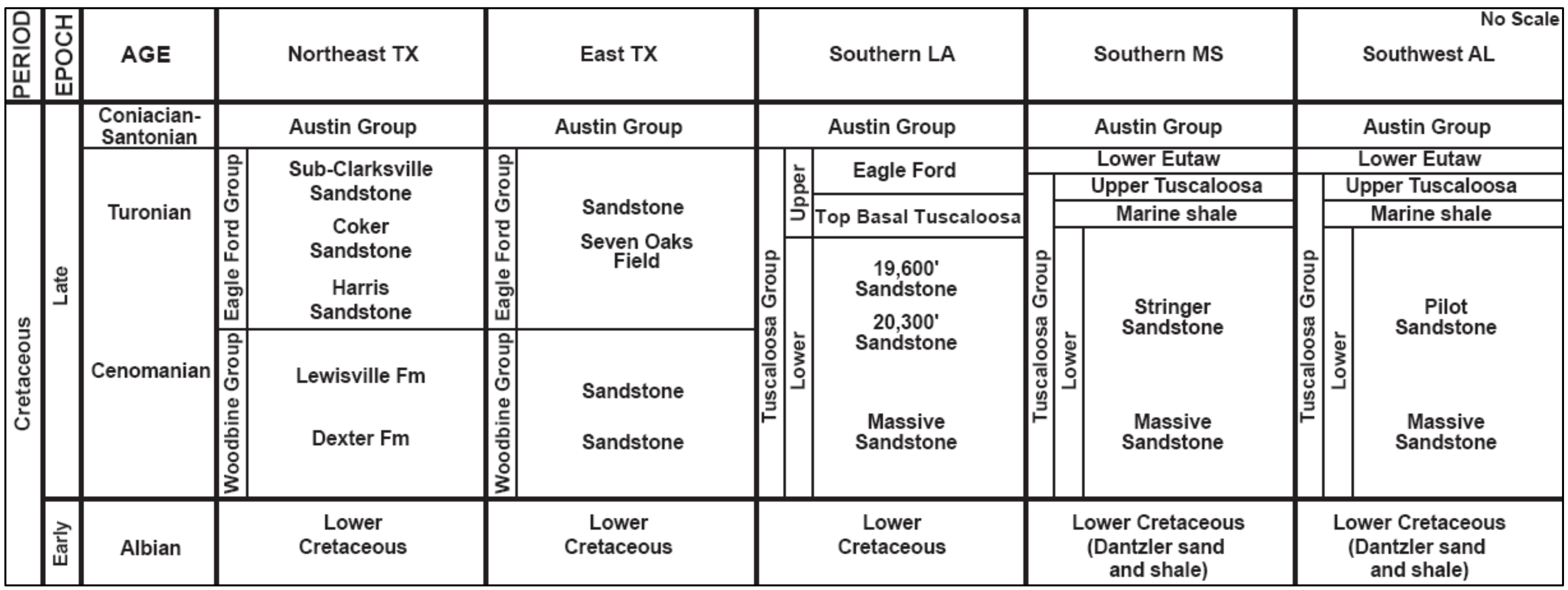

Figure 1.3. Stratigraphic column modified from Genuise (1991) showing relationship and variable lithostratigraphic nomenclature and correlation of Cenomanian to Turonian deposits from northeast Texas to southwest Alabama.

massive, with abundant large, chert pebbles and fewer shale breaks, has been interpreted as a braided stream system. The upper part of the Massive sand shows basal pebble lags, cross bedding, and mud drapes, indicating channel abandonment—all indicative of a more meandering, fluvial depositional style (Berg and Cook, 1968; Chasteen, 1983). In more distal locations as the system approaches the more balanced sediment supply to accommodation environment at the shelf edge it transitions to a shelf edge delta facies and several growth faults cause a dramatic thickening of the Lower Tuscaloosa (Barrell, 1997; Petty, 1997).

The Stringer sand is composed of gray, micaceous shales interbedded with thin sands and siltstones. The presence of oysters in the shales and intense bioturbation in the sands indicate deposition in a nearshore marine to brackish environment (Drennen, 1953; Chasteen, 1983). Drennen (1953) noted minimal marine influence in the outcrop, located in a belt across northern Mississippi to central Alabama, Georgia, South Carolina, and North Carolina (Figure 1.1), although marine influence increases downdip until it becomes almost entirely marine. The Pilot sand, which caps the top of the Lower Tuscaloosa, has been identified as a shallow shelf marine-bar complex by several workers (Berg and Cook, 1968; Mancini et al., 1980; Hamilton and Cameron, 1986).

Previous workers have interpreted the presence of slope-to-basin-floor deposits of Tuscaloosa age in deep stratigraphic wells and seismic sections in the Gulf of Mexico (Wu et al., 1990; Sawyer et al., 1991; Sohl et al., 1991; Dubiel and Pitman, 2002; Meyer et al., 2007). It was not until the BAHA #2 well (Meyer et al., 2007; Rains et al., 2007) penetrated more than 300 m of Tuscaloosa-age clastic gravity flow deposits that a significant deepwater facies was recognized. These sediments have since been tied to onshore, Woodbine-equivalent, shelf-edge deltas (Wornardt, 2010). The Davy Jones #2 well, located in a paleo-midslope position (McMoRan Exploration Co., 2011), encountered thick sections of pay in the Tuscaloosa interval. Logs from the well show what appear to be slope gravity flow sands. Newly released data from the Tiber well represents the most complete data set of deep offshore Tuscaloosa-aged deposits to date. This well indicates thick sequences of Tuscaloosa-age, sand-rich gravity deposits travelled over 400 km from their shelf edge source. However, the exact regional extent and nature of Tuscaloosa deepwater facies remains unknown.

Sequence Stratigraphy

The large-scale cyclicity of the Tuscaloosa Formation was well described by Mancini and Puckett (2005) in more shelf and shoreline deposits as an overall transgressive/regressive cycle. The transgressive/regressive model defines aggrading, backstepping, and infilling cycle intervals bounded at the top and base by surfaces of subaerial exposure (SA). Additional surfaces include a transgressive surface of ravinement (TS/RS) between the aggrading and backstepping intervals and a surface of maximum transgression (SMT) between the backstepping and infilling intervals (Figure 1.4).

The unconformity at the base of the Tuscaloosa Formation is known as the “mid-Cenomanian unconformity” and records widespread erosion throughout the Gulf of Mexico margin (Salvador, 1991).The magnitude of incision along this unconformity into Lower Cretaceous-, Jurassic-, Triassic-, and even Paleozoic-age strata suggests that the exposure and erosion associated with the base of the Tuscaloosa Formation are most likely due to the combined effects of both eustatic sea-level drop and regional tectonic uplift of the northern Gulf Coast (Salvador, 1991; Mancini and Puckett, 2005). This unconformity, known as the mid-Cenomanian unconformity or the mid-Cretaceous sequence boundary (Buffler et al., 1980; Buffler and Sawyer, 1985; Winker and Buffler,

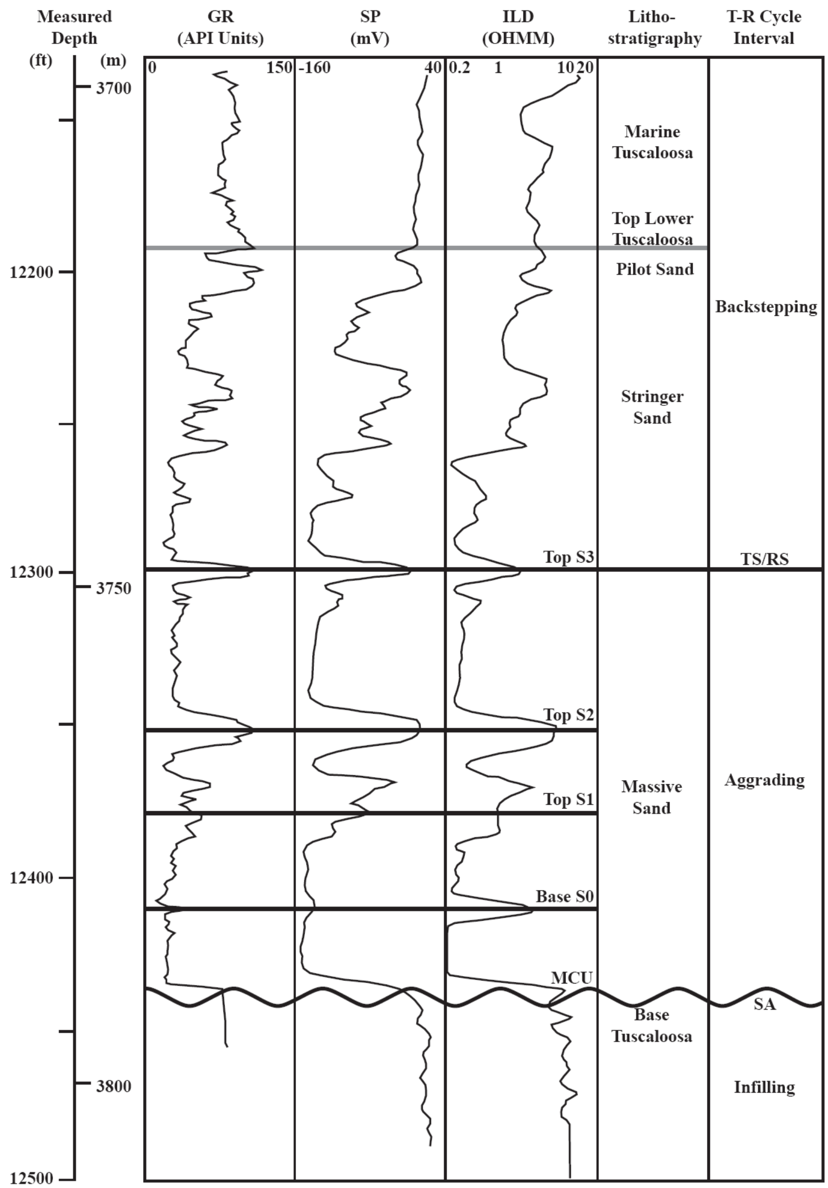

Figure 1.4. Type log of Lower Tuscaloosa interval from the Anderson Fred AIII #1 well in Liberty field, Amite County, Mississippi showing relationship of lithostratigraphic packages to sequence stratigraphic surfaces and intervals. Lower Tuscaloosa constitutes transgressive part of overall transgressive to regressive cycle; contains aggrading interval and a part of backstepping interval (Mancini and Puckett, 2005). Surface of subaerial exposure (SA) marks base of the Lower Tuscaloosa, and a transgressive surface/ravinement surface (TS/RS) marks boundary between Massive and Stringer sands and aggrading and backstepping intervals. Massive sand has been further divided into four higher frequency sequences (S0, S1, S2, S3) in Chapter 2.

1988), is an important event for at least two reasons. First, it is widely recognized and easily correlates across the Gulf of Mexico region. Second, it represents a major change in sedimentation style from a predominantly carbonate section below to a predominantly clastic section above (Salvador, 1991).

Following exposure and formation mid-Cenomanian unconformity at the base of the Tuscaloosa Formation a widespread marine transgression followed extending to the foothills of the paleo-Appalachian and Ouachita Mountains and northwest into the Western Interior Seaway (Figure 1.2) (Wu et al, 1990; Salvador, 1991; Sohl et al., 1991; Blakey, 2011). The initial rise in base level marked the deposition of the aggradational interval of the nearshore and onshore Lower Tuscaloosa (Sohl et al., 1991). It was during this time that the Massive sand was deposited (Figure 1.4). A continued and more rapid rise in base level created the transgressive surface/ravinement surface between the Massive and Stringer sands, followed by the backstepping interval, which includes the upper section of the Lower Tuscaloosa (Stringer and Pilot sands) and the lower part of the marine shale Middle Tuscaloosa (Figure 1.4) and is capped by a surface of maximum transgression. The infilling interval that followed was marked by a decrease in accommodation, and the upper part of the marine shale of the Middle Tuscaloosa and prograding sands of the Upper Tuscaloosa were deposited. The entire sequence is capped by another major subaerial exposure between the Upper Tuscaloosa and the Eutaw Formation (Salvador, 1991; Mancini and Puckett, 2005). This dissertation focuses on the Lower Tuscaloosa Formation.

Structure

The Gulf of Mexico formed as a result of the rifting of Pangea, which is widely thought to have started in the Triassic and continued into the Jurassic, with a northeast-trending system of rifts offset by northwest-striking transform faults (Thomas, 1988; Salvador, 1991). Several workers have confirmed the presence of northwest-striking transform faults in the basement of the Gulf of Mexico (Klitgord and Schouten, 1986; Simmons, 1992; Adams, 1993, 1997; Bradshaw and Watkins, 1995; MacRae and Watkins, 1996; Karlo and Shoup, 2000; Colling et al., 2001). However the diversity of data and methods in each study has led to a non-specific understanding of the exact trend, spacing and location of these transform faults. The maps of Stephens (2001, 2009) provide a good summary of several of these differing interpretations, supplemented by new interpretations by Stephens that are largely based on 2D and 3D seismic data.

Strong evidence suggests that uplifts and transform fault zones in the Gulf of Mexico were active throughout the Mesozoic and that they may have had a strong influence on depositional and petroleum systems in sediments as young as Tertiary (Hohlt 1977; Hogg 1988; Meylan 1991; Salvador, 1991; Sohl et al., 1991; Stephens, 2001, 2009). Basement highs important during the deposition of the Tuscaloosa Formation include the Wiggins arch, Jackson's dome, Monroe uplift, Louisiana arch, Sabine uplift, and Baldwin high (Figure 1.2). Additionally, work by Stephens (2009) suggests that the Wiggins arch may have been separated by the Pearl River transform fault into an eastern arch and a western arch. Data from this dissertation seems to support that interpretation and therefore, we refer to the Wiggins arch as being two components, the eastern Wiggins arch and the western Wiggins arch. These uplifts most likely formed topography with which shoreline and coastal-plain depositional systems interacted in the Late Cretaceous to influence environments of deposition, hinterland slopes, coastal processes, and accommodation regimes. The major basins that created accommodation and hosted sedimentation of the onshore Tuscaloosa Formation were the Mississippi and north Louisiana salt basins (Figure 1.2).

Summary of Data and Methods

The summary of data and methods in this section briefly describes the data and methods used for this entire dissertation. A more detailed explanation of the data and methods used in each chapter can be found in that chapter.

Several tasks were completed to achieve the objectives outlined for this study. An extensive number of publications and original work was compiled into a common ArcGIS database. This allowed for an easy comparison of the numerous studies that have been done on the trend. To date, this database contains over 200 maps (paleogeographic, isopach, and structure), 100 cross sections, and 70 point files containing information from wells detailing depositional environments, oil chemistry, and reservoir quality in the Tuscaloosa Formation.

668 well log suites were used to regionally map several chronostratigraphic sequences of the Lower Tuscaloosa. Individual channelized sequences in the Lower Tuscaloosa were identified based on gamma log patterns (considered a pseudo-sand/shale indicator). Although individual sands and shales may not correlate between wells and their thicknesses may alter dramatically between wells, the overall pattern of sandy, channelized aggradation and muddier periods of floodplain alluviation typifies cycles in the terrestrial-fluvial systems (see Miall, 2002; Takano and Waseda, 2003; Zhu et al., 2008 for discussion). Detailed regional-structure, isopach, and net-sand maps of major surfaces and intervals of the Lower Tuscaloosa were created to define spatial and temporal depositional trends.

Thirteen conventional cores were used in conjunction with information from three offshore wells that penetrate gravity deposits in the Tuscaloosa Formation equivalent section. Detailed core descriptions identifying lithologies, textures, grain size, sedimentary structures, fauna, and ichnogenera were used as the basis for identifying facies in these core and observations were integrated into interpretation of regional depositional system relationships. High resolution photographs were taken of the core and are presented in this dissertation. Facies were grouped into larger-scale facies associations which enabled, along with wireline log patterns, an interpretation of depositional environments. Ichnological observations, placed in the context of sedimentology, were considered important diagnostic features in the designation of depositional environments. Cross sections using both cored and uncored wells enabled core interpretations to be integrated with regional mapping that documents the paleogeomorphology and structural landscape of the Lower Tuscaloosa Formation.

Equations that estimate morphometrics of fluvial channels were used to estimate channel depth, channel width, and channel belt width from measurements of cross sets in 4 core and point bars interpreted from 136 wireline logs. Core observations were used to confirm the log motif of point bar deposits. Calculated dimension estimates were used to assess outcrop and modern day systems with similar dimensions and these systems were considered as analogs for the Lower Tuscaloosa fluvial systems.

From cored intervals in 13 different wells 115 thin sections impregnated with blue-dyed epoxy were made from rocks sampled regionally from traditionally hydrocarbon productive locations in Mississippi and Louisiana. These samples were taken over several different stratigraphic depths and depositional environments. These data enabled an assessment of the relationships between chlorite and both provenance and depositional environment. Thin sections were examined by optical microscopy for general observations of the diagenetic sequence. The diagenetic sequence was determined from observations of crosscutting relationships, dissolution of framework grains and cements, and growth of authigenic cements. Thin sections were point counted for a minimum of 300 points per slide. Thin section photomicrographs and porosity and permeability data from the offshore Tiber well were also used to make a preliminary assessment on the reservoir quality of the Tuscaloosa-aged deepwater deposits.

SUMMARY OF CHAPTERS

This summary of chapters serves to highlight the major goals and conclusions of each chapter.

1. Chapter 1 (this chapter) explains the importance of studying the Lower Tuscaloosa Formation. It also introduces the previous work that has been done on the Tuscaloosa Formation. This introduction discusses the regional setting, lithostratigraphy and paleogeography, large-scale sequence stratigraphy, and structural setting of the Tuscaloosa Formation. A summary of the data and methods used to accomplish this dissertation are then outlined. The principle research objectives and how the discussion of each of those objectives is organized in this dissertation are also outlined.
2. Chapter 2 builds on a previously interpreted large scale sequence stratigraphic framework for the Tuscaloosa Formation (Mancini and Puckett, 2005) to develop a higher-frequency sequence stratigraphic framework for the Lower Tuscaloosa Formation. This newly interpreted framework is used as a basis for much of the work in Chapter 2 and throughout the rest of the dissertation. The paleogeographic history of the Lower Tuscaloosa Formation is detailed within this framework and represents the first time the paleogeographic history of the Lower Tuscaloosa has been described in great detail. 668 well log suites are used to: (1) Assess the nature of sequences and to better understand the regional deposition, lithology, and sedimentology of the onshore Lower Tuscaloosa; (2) identify the paleofluvial and deltaic

depocenters that may have been possible feeder systems to the deeper Tuscaloosa fan system during Lower Tuscaloosa time; and (3) discuss the probability of deepwater Tuscaloosa-aged sediments in the context of this proximal evidence.

Five sequences (S0, S1, S2, S3, and Stringer) were mapped in the Lower Tuscaloosa. The Lower Tuscaloosa was derived dominantly from the ancient Appalachians to the northeast and from a large drainage trending within the ancestral Mississippi Valley. Several thick fairways of Lower Tuscaloosa sediments terminate southward at the highly extensional paleoshelf edge. Basement-rooted and long-stable structural uplifts influenced drainage directions and formed buttresses that resulted in ponded thicknesses of fluvial, estuarine, and coastal plain deposits. This chapter has been submitted to the AAPG Bulletin for publication.

3. Chapter 3 integrates the paleogeographic history from Chapter 2 with cored intervals from 13 wells and associated well-log motif data to detail the geographic distribution of Lower Tuscaloosa Formation facies and depositional environments from source-to-sink (onshore-to-offshore). Observations of key ichnofacies are integrated with sedimentological observations from the core to better define the depositional environments. Wireline well log patterns are also closely tied to core descriptions to improve interpretation of logs where only log motif exists. Importance is placed on the onshore study area in Mississippi and southeast Louisiana, and observations are made using limited information available in the offshore wells Davy Jones #2, BAHA #2, and Tiber.

This dissertation constitutes the first study to integrate the feeder systems of the Lower Tuscaloosa with the equivalent far-travelled gravity deposits, and the first study of the Tuscaloosa Formation to apply an incised valley/estuarine model to the paleogeography of the Tuscaloosa Formation to explain the regional variability in facies. Finally, this is the first study to integrate ichnology from the Tuscaloosa Formation with detailed core information to characterize Tuscaloosa facies and depositional environments.

4. Chapter 4 uses information on the location of Lower Tuscaloosa fluvial systems from Chapters 2 and 3 to better understand the size and capacity of channels from those systems. Morphometric estimates on the depth and width of Lower Tuscaloosa channels and on the width of their channel belts were made from cross set measurements in core (n=4 core) and from measurements of point bar heights from wireline logs (N=136 logs). Lower Tuscaloosa channel dimensions were then compared with documented dimensions in ancient and modern fluvial systems to gain a better understanding of the Lower Tuscaloosa fluvial systems.

   Outcrop analogs for the Lower Tuscaloosa Formation include the Kinderscoutian sediments in the Central Pennine basin, the Siwalik Group in Pakistan, the Escanilla Formation in Spain, the Chuckanut Formation in Washington, and the Canyon Creek Member of the Ericson Formation. Appropriate modern analogs for the Lower Tuscaloosa Formation are the Missouri, Ohio, and Alabama Rivers.

5. Chapter 5 discusses the petrographic nature of the Lower Tuscaloosa Formation sandstones and determines the regional distribution of chlorite in the formation to test Thomson's (1979) assertion of chlorite being controlled

by provenance. The depositional history determined in Chapter 2, which indicates the main source of sediments in the Lower Tuscaloosa Formation, is used to assess trends in reservoir quality that may be influenced by provenance. With the better understanding of facies, depositional environments, and the sequence stratigraphic framework, as defined in Chapters 2 and 3, a correlation between depositional systems and grain size with chlorite is also assessed. A paragenetic sequence for the Lower Tuscaloosa Formation is proposed. The reservoir quality of Tuscaloosa deepwater deposits in the Tiber well (discussed in Chapters 2 and 3) is also determined. Point counts of 115 thin sections of Lower Tuscaloosa sandstones from the core described in Chapter 3 were used to accomplish these goals.

This chapter concludes from observations in regional trends in chlorite, volcanic rock fragments, and quartz cement that reservoir quality is influenced by provenance. This is also the first study to discuss reservoir quality for proposed Tuscaloosa reservoirs in the deep offshore Gulf of Mexico.

6. Chapter 6 reviews the conclusions of chapters 2, 3, 4, and 5 and discusses the limitations of this dissertation as the basis for future work.

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Chapter 2: Sequence Stratigraphic and Paleogeographic Framework: Influences on Sediment Distribution and Depositional Environments

INTRODUCTION

Regional stratigraphy of the Lower Tuscaloosa Formation, its environments of deposition, provenance, and basinward extent, are all poorly understood. As trends in exploration and production of hydrocarbons move farther offshore and stratigraphically deeper, an increase in understanding of the Lower Tuscaloosa onshore and its implications for new, deeper opportunities in this interval push the need for additional study. A well-constrained understanding of locations where sediments are being fed into the paleo-offshore allow for a better understanding and prediction of the location and nature of deepwater slope and basin-floor-fan deposits. Downdip facies of the Tuscaloosa Formation have been alluded to by several authors (Wu et al., 1990; Sawyer et al., 1991; Sohl et al., 1991; Barrell, 1997; Dubiel and Pitman, 2002), although published research focusing on these facies is completely lacking. This lack is partly due to the difficulties of imaging and drilling the great stratigraphic depths and deepwater locations in which these deposits are presumed to be located. Recent exploration wells, although primarily targeted to the deep offshore Wilcox trend, have been drilled to test the offshore Tuscaloosa interval and results indicate the presence of clastic sediments in the deepwater offshore trend. Although the focus of these deep studies to date has been the Wilcox interval, the attention has recently shifted with the drilling of the Tiber well in the Keathley Canyon area (see Figure 2.1 for location) and the Davy Jones #2 well (McMoRan Exploration Co., 2011), which has proven potential for the existence of hydrocarbons in the offshore slope and abyssal plain facies Tuscaloosa play.

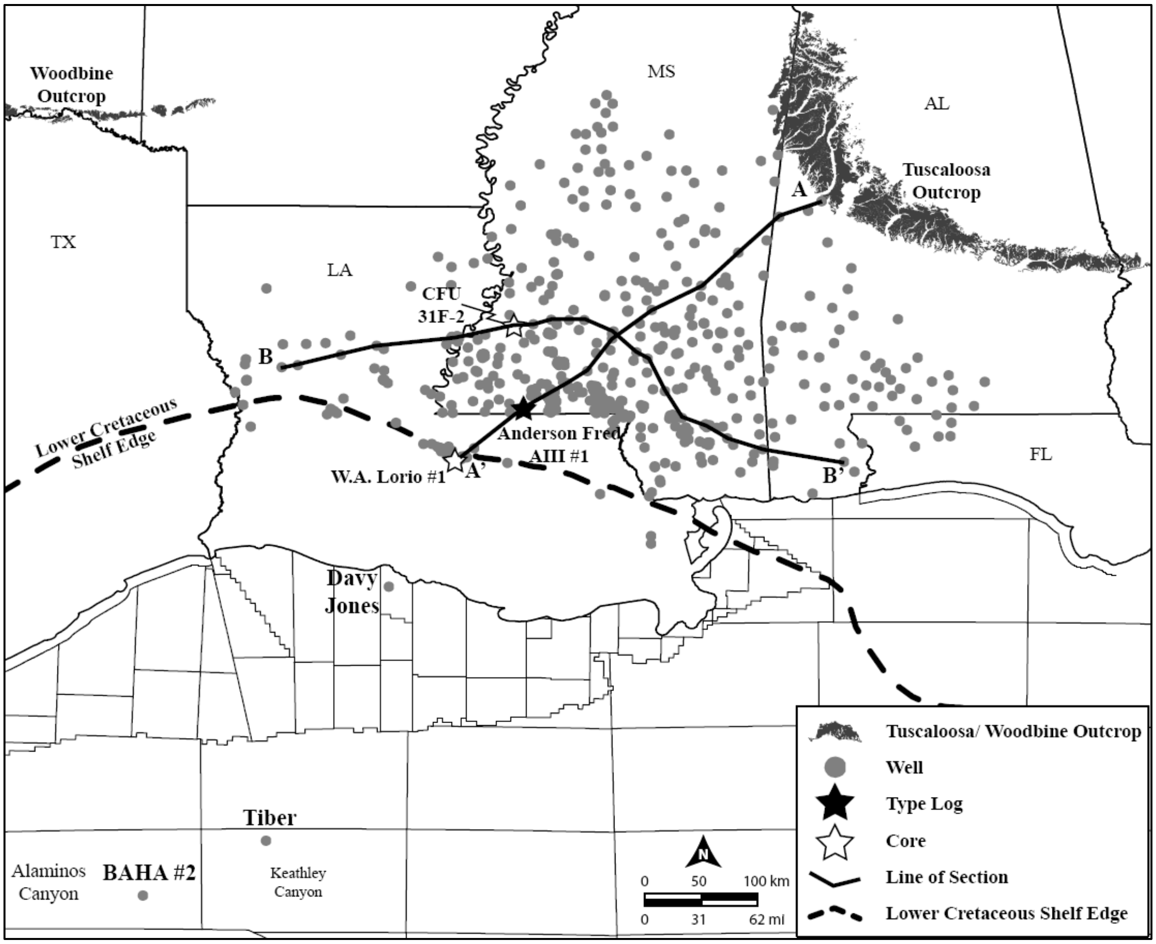

Figure 2.1. Regional map of the study area showing the locations of 668 wells used in this chapter, Tuscaloosa and Woodbine outcrops, and Lower Cretaceous shelf edge. Locations of key cross sections (A-A′ and B-B′), key cored wells (CFU 31F-2, Lorio #1), recent deepwater Tuscaloosa penetrations (BAHA #2, Davy Jones #2, and Tiber), and type log used for this chapter (Anderson Fred AIII #1 well) (Figure 1.4) are also noted.

Goals of this chapter are to (1) assess the nature of sequences and regional deposition, lithology, and sedimentology of the onshore fluvial and nearshore deltaic systems of the of Lower Tuscaloosa; (2) identify the paleofluvial and deltaic depocenters that may have been possible feeder systems to the deeper Tuscaloosa fan system during Lower Tuscaloosa time; and (3) discuss the probability of deepwater Tuscaloosa-age sediments in the deepwater Gulf of Mexico in the context of this proximal evidence.

DATA AND METHODS

The large-scale sequence stratigraphic framework of the Tuscaloosa Formation as a transgressive/regressive cycle (Mancini and Puckett, 2005; see Chapter 1 for discussion) was identified on 668 well log suites (see Figure 2.1 for locations). These same wireline well logs were used to regionally map several smaller scale chronostratigraphic sequences of the Lower Tuscaloosa. Individual channelized sequences in the Lower Tuscaloosa were identified on gamma logs (considered a pseudo-sand/shale indicator) by the sharp-based, upward-shaling or blocky log motifs. Deepening over the tops of these cycles was identified by the increasing gamma reading, or upward shaling, of the overlying interval or by high-gamma-reading shale capping the intervals. Although individual sands and shales may not correlate between wells or their thicknesses may alter dramatically between wells, the overall pattern of sandy, channelized aggradation and muddier periods of floodplain alluviation typifies cycles in the terrestrial-fluvial systems (Blum and Tornqvist 2000). Dramatic increases in muds associated with transgressive deepening within the Lower Tuscaloosa have been recognized by numerous authors as a regional phenomenon (Karges, 1962; Chasteen, 1983; Hansley, 1996; Mancini and Puckett, 2005).

Detailed regional-structure, isopach, and net-sand maps of major surfaces and intervals of the Lower Tuscaloosa were created to define spatial and temporal depositional trends. Trends were assessed using 2D seismic data, core, and well logs. Regional cross sections were produced from an integration of well log and seismic, and core data. Cored intervals from 13 wells in the study area were described and integrated into the regional framework in Chapter 3 to provide ground truth to log-motif interpretations and specific control points on environments of deposition.

OBSERVATIONS

Sequence framework, accumulation trends, and log character

The Massive sand of the Lower Tuscaloosa is composed of four sequences, which are, from oldest to youngest, the basal sequence (S0), sequence 1 (S1), sequence 2 (S2), and sequence 3 (S3) (Figures 1.4, 2.2, 2.3). Although these sequences can be mapped from well log motifs, they change in net sand and architecture across the study area according to their nature as either meandering-stream deposits (sharp based, upward fining), channel core deposits (blocky), prograding deltaic deposits (upward coarsening), prodelta muds and hyperpycnites (ratty), or nearshore marine estuarine (variable blocky to ratty). Core in these units aid in associating log motif with depositional environments (Chapter 3). Lower Tuscaloosa sequences shale upward in fluvial deposits and marine shales cap deltaic deposits. The general trend of the entire Tuscaloosa Formation (Lower, Middle, and Upper) shows an overall increase in thickness from west to east, with thickest accumulations (~213 m) occurring at the paleoshelf edge across the Mississippi panhandle and into Alabama. Additionally, the Lower Tuscaloosa thickens upsection from the basal sequence through each successive sequence (Figures 2.4–2.7).

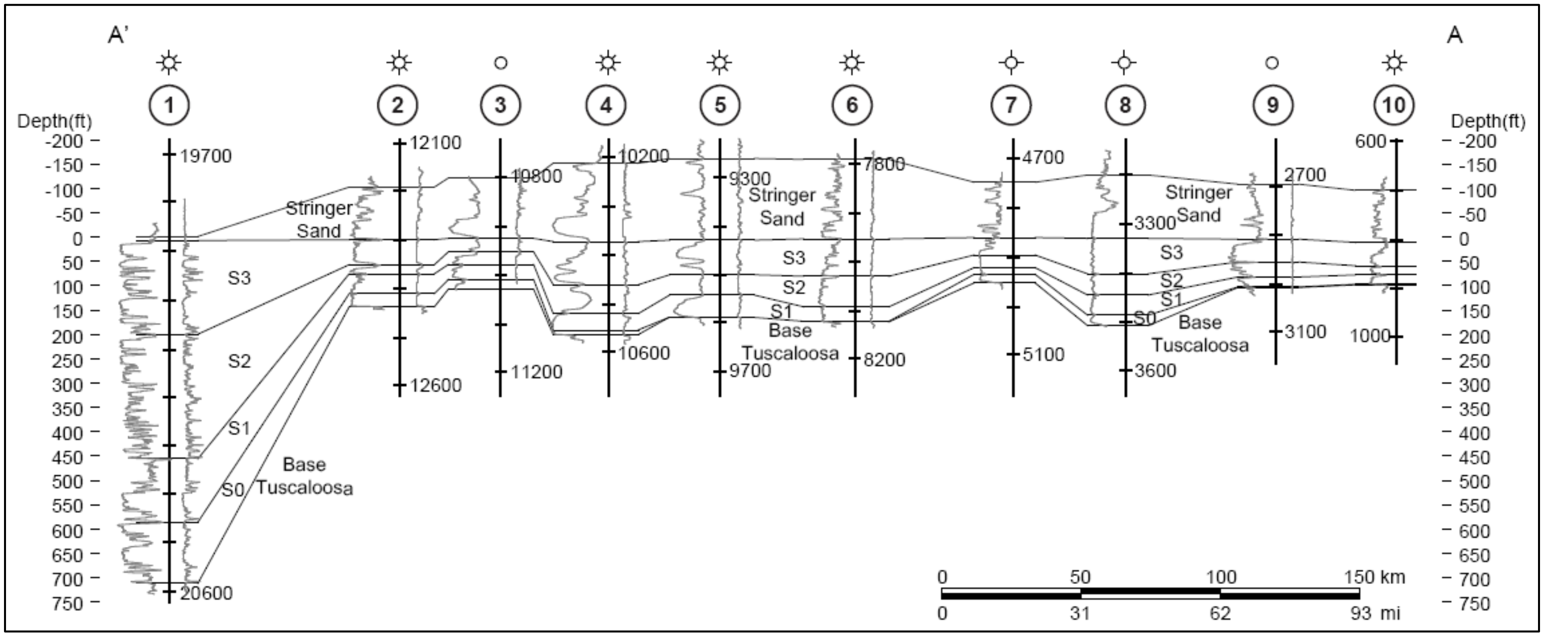

Figure 2.2. Dip cross section of Lower Tuscaloosa (see Figure 2.1 for location) flattened on top of sequence 3 (S3). Basal sequence (S0), sequence 1 (S1), sequence 2 (S2), and Stringer sand also picked on cross section. Note dramatic expansion from well 1 to well 2 as the system expands across the shelf edge. Detailed information on well name, operator, and location of each well is shown in Appendix IA.

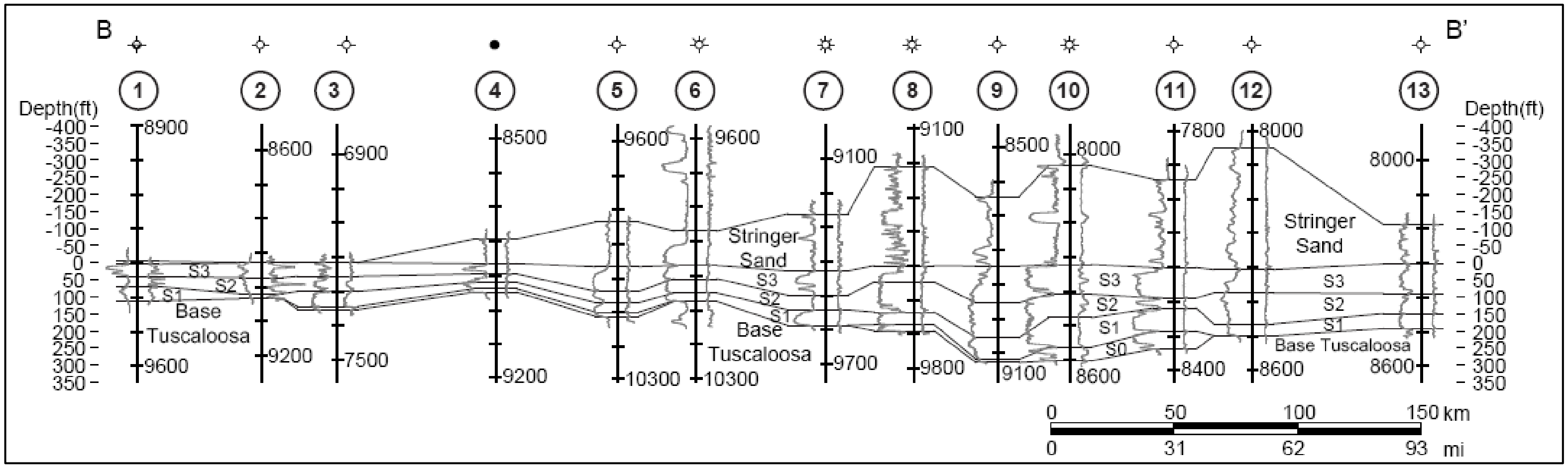

Figure 2.3. Strike cross section of Lower Tuscaloosa (see Figure 2.1 for location) flattened on top of sequence 3 (S3). Basal sequence (S0), sequence 1 (S1), sequence 2 (S2) and Stringer sand are also picked on cross section. Note how the Lower Tuscaloosa thickens to east. Detailed information on well name, operator, and location of each well is shown in Appendix IB.

The basal sequence of the Lower Tuscaloosa (S0) ranges in thickness from 0 m to about 33.5 m, with a maximum thickness at the paleoshelf edge (Figure 2.4). In several places it is absent to very thin, particularly over Jackson's dome, the Monroe uplift, the Louisiana arch, and the east and west Wiggin's arch. In contrast, sequences 1 through 3 of the Lower Tuscaloosa (S1, S2, S3) are thicker and more laterally continuous across the region than is the underlying S0 (Figures 2.5–2.7). S1 ranges from 12 m to 55 m in thickness in its main sediment fairways and reaches its maximum thickness at the paleoshelf edge (Figure 2.5). S2 is slightly thicker than S1, ranging from 12 m to 61 m and reaching its maximum thicknesses at the paleoshelf edge. The increase in thickness of these sequences is most apparent in the main sediment fairway across the Mississippi panhandle (Figure 2.6). S3 ranges in thickness from 15 m to 46 m. Although the maximum thickness of S3 is less than that of S2, the S3 does show some exceptional thicknesses in downdip locations, north of the east and west Wiggins arches (Figure 2.7).

Accommodation and thickness distribution of the Massive sand, and it constituent sequences, contrasts dramatically with that of the Stringer sand. The Massive sand isopach ranges from 61 m to over 152 m; showing very thick accumulations along the paleoshelf edge (Figure 2.8) at both eastern, central, and western locations. A large thick accumulation trends east to west across central Mississippi and Alabama and may even extend to areas west in northern Louisiana. This trend separates linear northeast to southwest trending thick valley fills north of the trend with distinct deltaic depocenters south of this trend. The trend is herein interpreted as a paleoshoreline of the Lower Tuscaloosa. In contrast to the Massive sand unit, the Stringer sand isopach ranges from 49 m to 122 m thick, with its thickest accumulations in the sediment depocenters located proximal to the east Wiggins and west Wiggins arches (Figure 2.9). In fact, little to no sand seems to reach the shelf edge during deposition of the Stringer sand. A broad

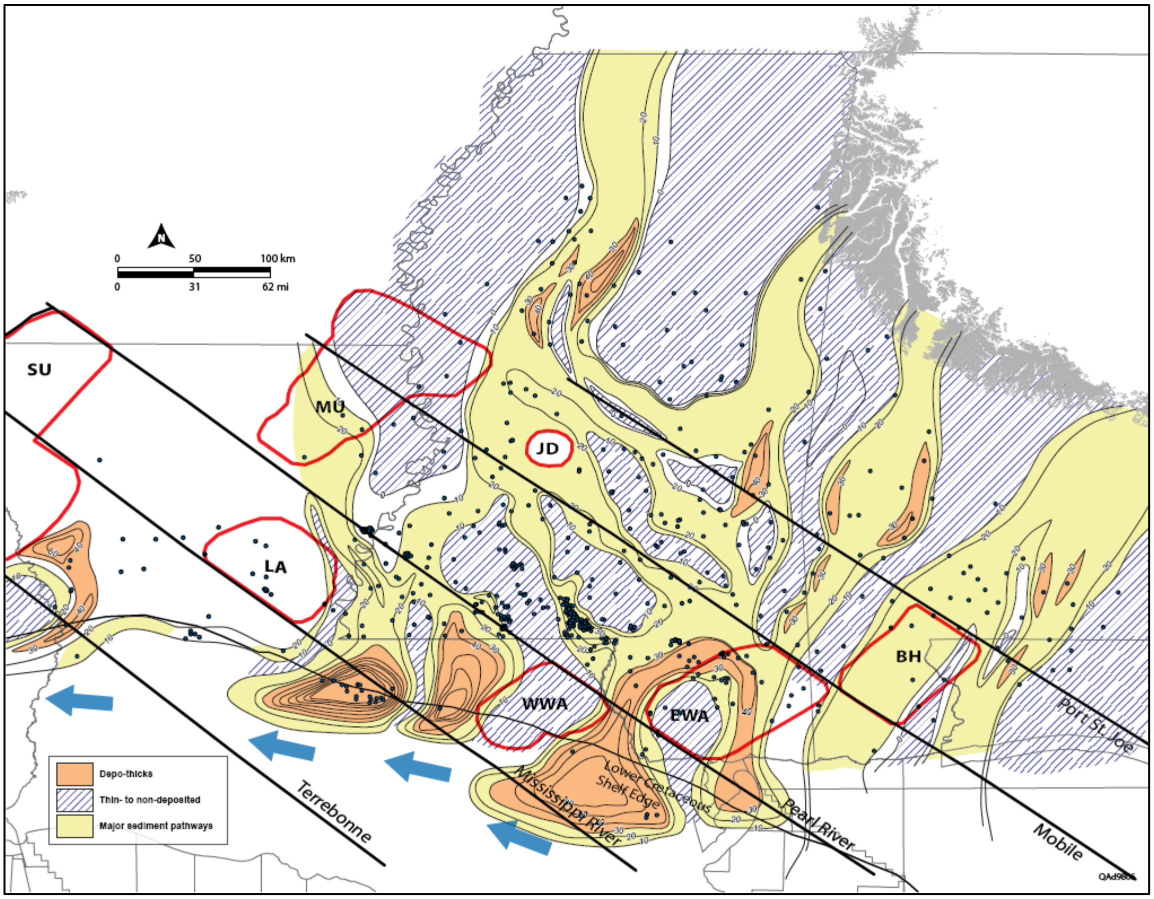

Figure 2.4. Isopach map of basal sequence of Lower Tuscaloosa (S0) ranges in thickness from 0 m to about 33.5 m, with maximum depth at the shelf edge. In several places absent to very thin. Main sediment fairway comes from northeast to southwest, with additional sources coming from north and northwest. Overlain on isopach are major basement structures, as interpreted by Stephens (2009), including key basement highs and transfer faults from rifting of Pangea during opening of the Gulf of Mexico. La Salle arch (LA), west Wiggins arch (WWA), east Wiggins arch (EWA), Baldwin high (BH) Jackson’s dome (JD), Monroe uplift (MU), and Sabine uplift (SU), Pearl River and Mobile transfer faults).

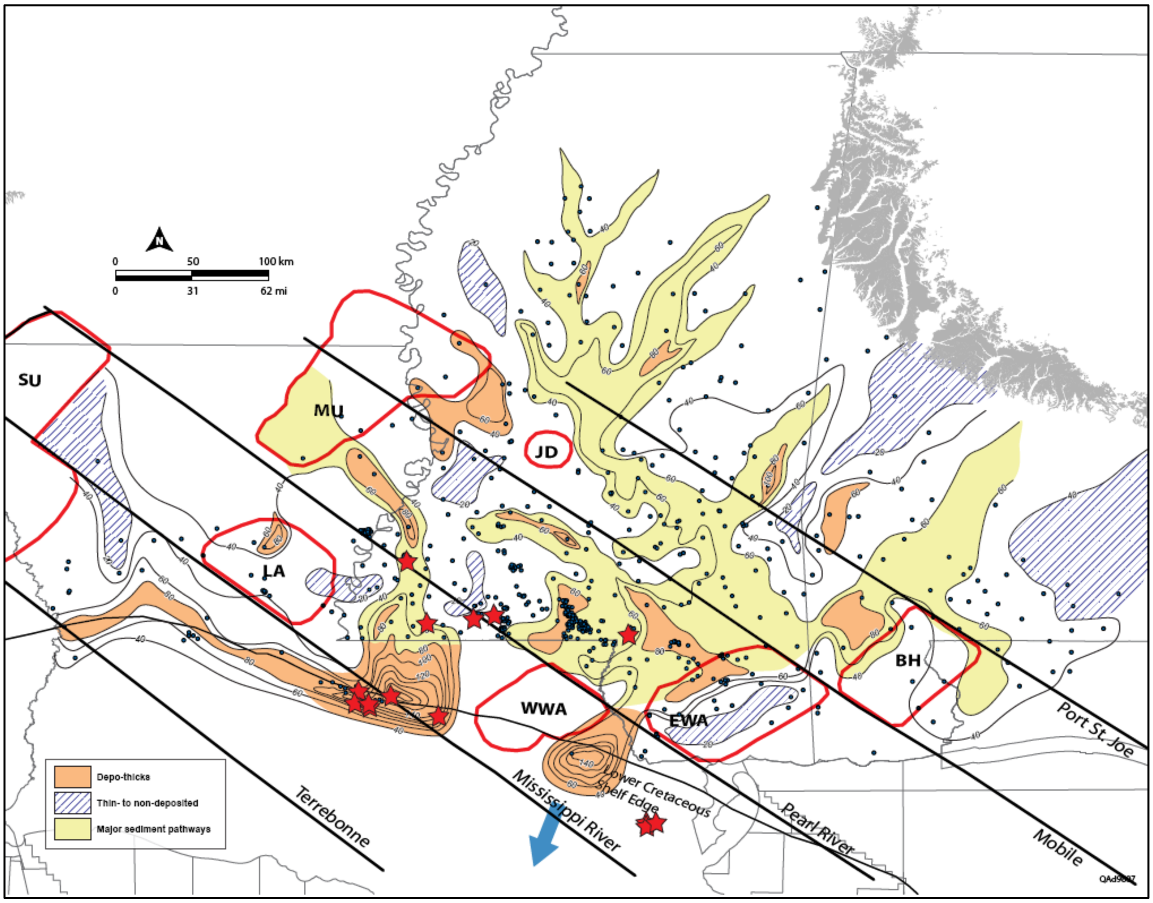

Figure 2.5. Isopach map of sequence 1 (S1) ranges from 12 m to 55 m in thickness in its main sediment fairways and reaches its maximum thickness at shelf edge. Main sediment fairway comes from northeast to southwest, with additional sources coming from north and northwest. Overlain on isopach are major basement structures, as interpreted by Stephens (2009), including key basement highs and transfer faults from rifting of Pangea during opening of the Gulf of Mexico. La Salle arch (LA), west Wiggins arch (WWA), east Wiggins arch (EWA), Baldwin high (BH) Jackson's dome (JD), Monroe uplift (MU), and Sabine uplift (SU), Pearl River and Mobile transfer faults).

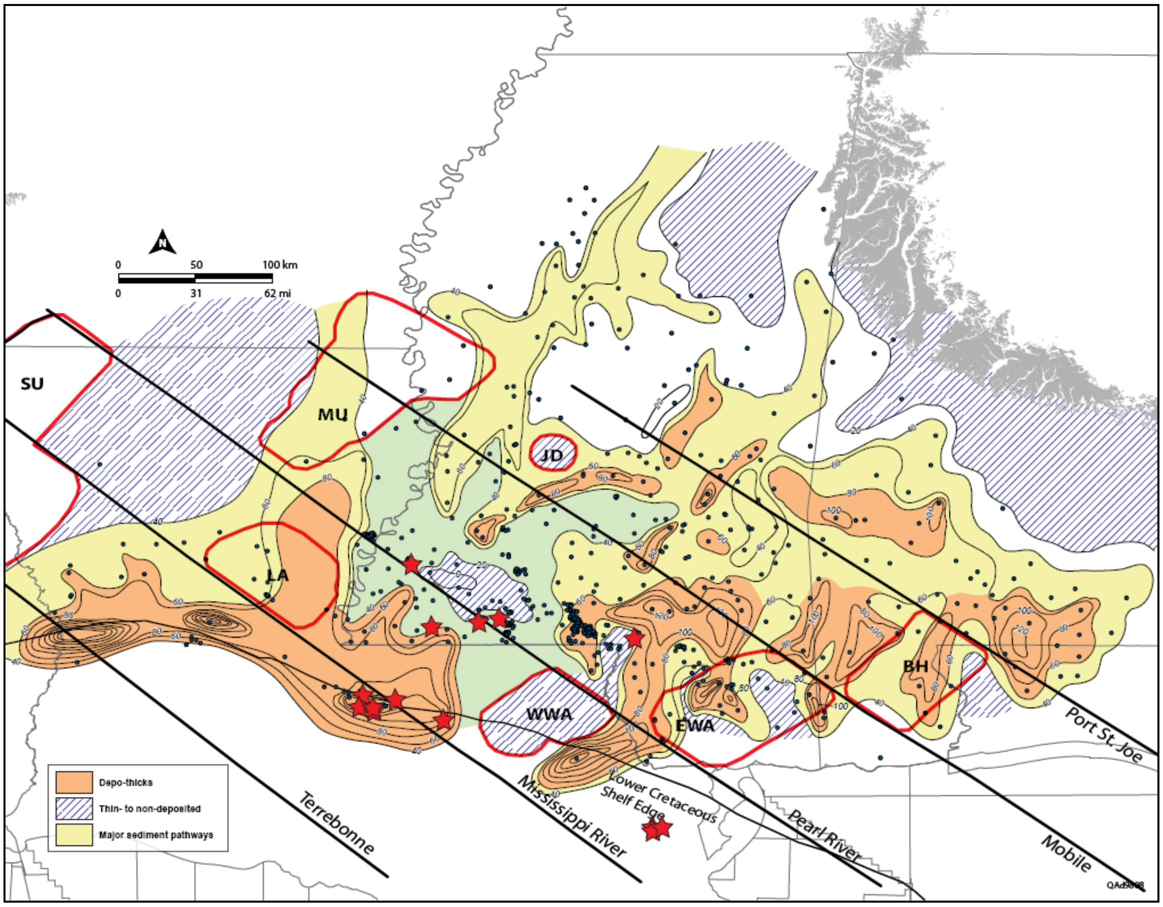

Figure 2.6. Isopach of sequence 2 (S2) shows overall increase in thickness from S1 and ranges from 12 m to 61 m, reaching its maximum thicknesses at shelf edge. This increase in thickness most apparent in main sediment fairway across Mississippi panhandle. Main sediment fairway comes from northeast to southwest, with additional sources coming from north and northwest. Overlain on isopach are major basement structures, as interpreted by Stephens (2009), including key basement highs and transfer faults from rifting of Pangea during opening of the Gulf of Mexico. La Salle arch (LA), west Wiggins arch (WWA), east Wiggins arch (EWA), Baldwin high (BH) Jackson's dome (JD), Monroe uplift (MU), and Sabine uplift (SU), Pearl River and Mobile transfer faults).

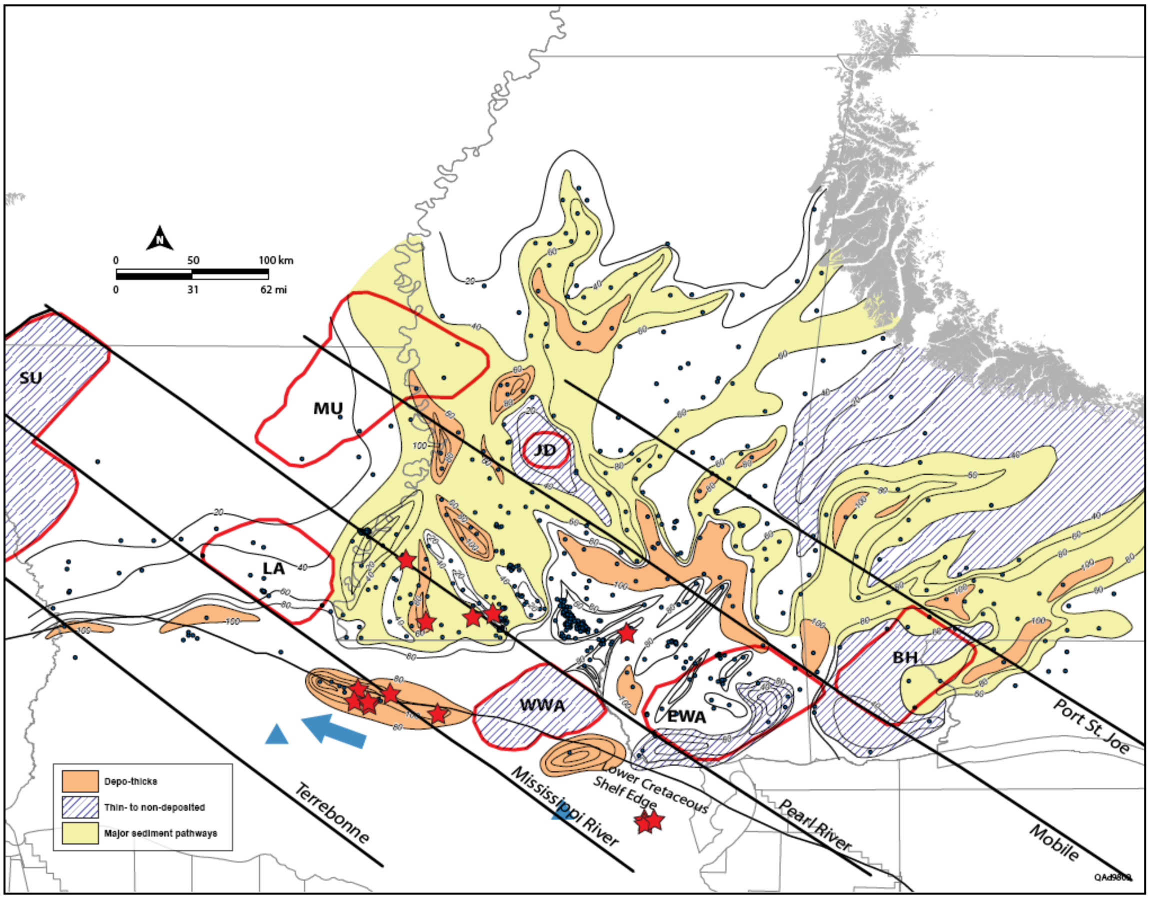

Figure 2.7. Isopach map of sequence 3 (S3) also showing increase in overall thickness in sediment fairways when compared with S2, ranging from 15 m to 46 m. Although maximum thickness of S3 less than S2, overall increase in thickness focused in proximal direction in main sediment fairways. Main sediment fairway comes from northeast to southwest, with additional sources coming from north and northwest. Overlain on isopach are major basement structures, as interpreted by Stephens (2009), including key basement highs and transfer faults from rifting of Pangea during opening of the Gulf of Mexico. La Salle arch (LA), west Wiggins arch (WWA), east Wiggins arch (EWA), Baldwin high (BH) Jackson's dome (JD), Monroe uplift (MU), and Sabine uplift (SU), Pearl River and Mobile transfer faults).

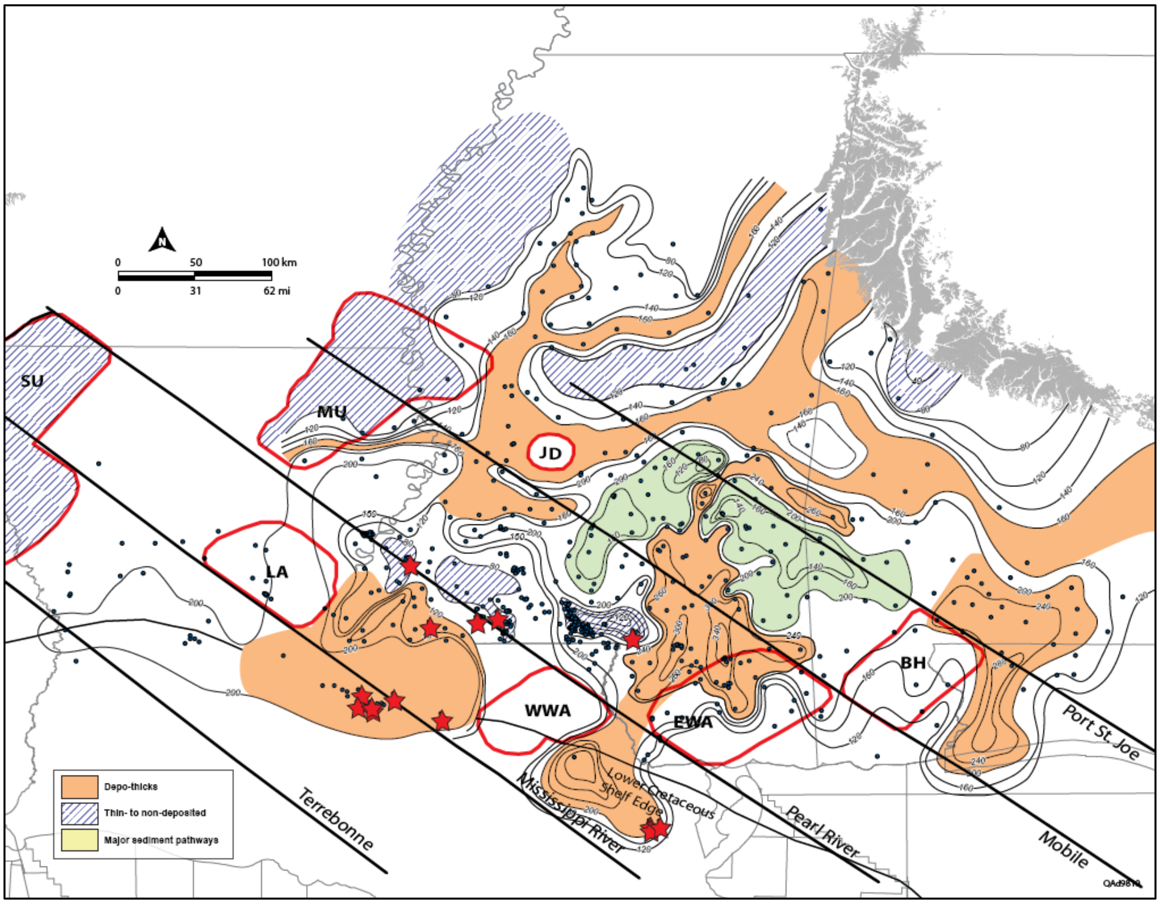

Figure 2.8. Isopach map of Massive sand. Massive sand ranges from 61 m to over 152 m, showing very thick accumulations at paleo-shelf edge as well as along the Massive sand corridor of central Mississippi and central Alabama . Overlain on isopach are major basement structures, as interpreted by Stephens (2009), including key basement highs and transfer faults from rifting of Pangea during opening of the Gulf of Mexico. La Salle arch (LA), west Wiggins arch (WWA), east Wiggins arch (EWA), Baldwin high (BH) Jackson's dome (JD), Monroe uplift (MU), and Sabine uplift (SU), Pearl River and Mobile transfer faults).

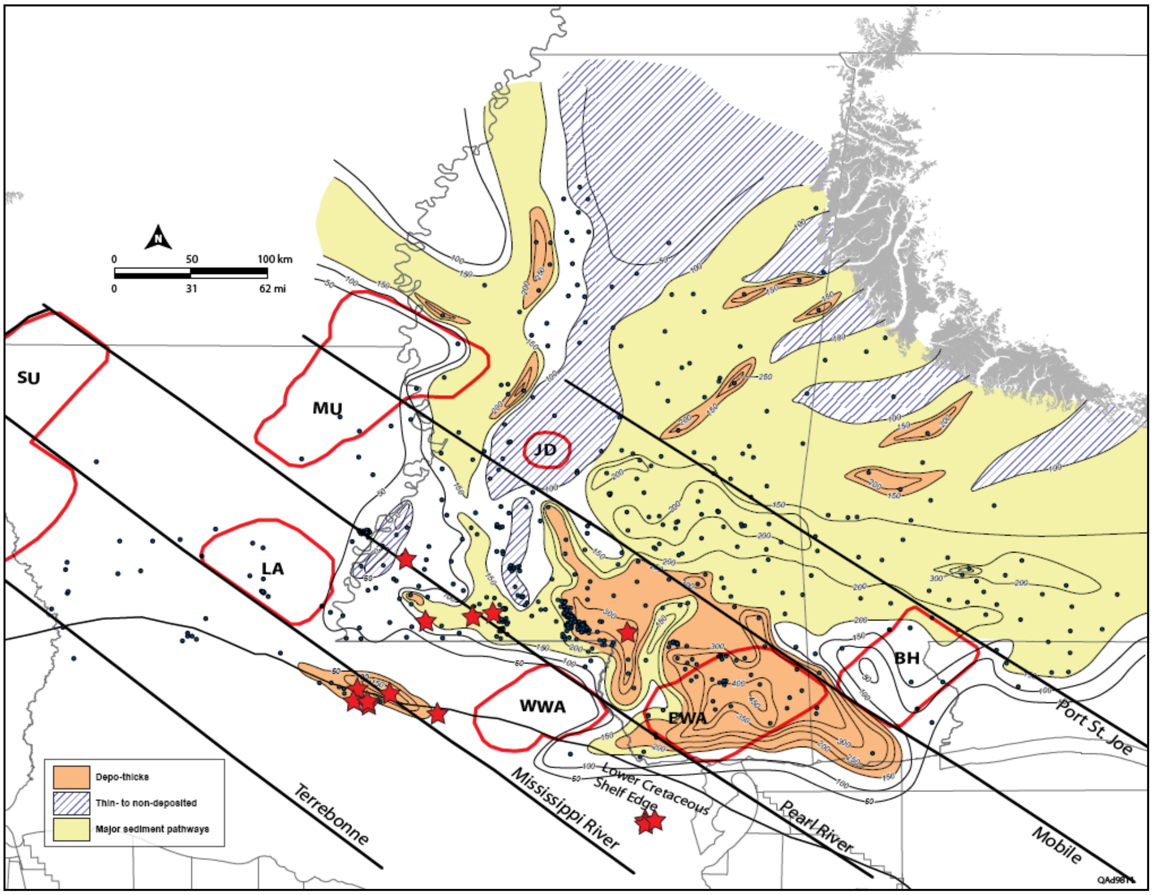

Figure 2.9. Isopach map of Stringer sand. Stringer sand isopach ranges from 49 m to 122 m, with thickest accumulations in sediment fairways in much more proximal direction from paleo-shelf edge. Overlain on isopach are major basement structures, as interpreted by Stephens (2009), including key basement highs and transfer faults from rifting of Pangea during opening of the Gulf of Mexico. La Salle arch (LA), west Wiggins arch (WWA), east Wiggins arch (EWA), Baldwin high (BH) Jackson's dome (JD), Monroe uplift (MU), and Sabine uplift (SU), Pearl River and Mobile transfer faults).

(~50 km) linear, east to west trending sand thick is present in the Stringer interval from the Jackson Dome (JD on Figure 2.9) to locations east of the study area. It is located slightly south of the similar trend in the Massive sand and is herein interpreted as the transgressive barrier shoreline of the Lower Tuscaloosa Stringer time.

Structure maps on the top of each sequence in the Lower Tuscaloosa show little change in the present-day structure from the base to the top of the Lower Tuscaloosa, suggesting there was little difference in regional dips and localized highs and lows during the deposition of each Lower Tuscaloosa sequence. Structure maps shown in Figure 2.10 exhibit similarity in structural character between the base and top of the Lower Tuscaloosa; likewise, structure maps created on the top of the other sequences also display this similarity.

Wireline logs over cored intervals in the Massive sand north of the paleoshelf edge show log motifs of sharp-based, slightly upward fining cycles (Figures 2.2 and 2.3), which are indicative of fluvial deposits (Miall, 1996; Bridge, 2003; Gibling, 2006). Core from the fluvial facies are closely tied to core descriptions in Chapter 3.

Log motifs for well penetrations at or south of the Lower Cretaceous paleoshelf edge change to inverted-funnel shapes with upward-coarsening cycles sharply capped by shale. Individual cycles are 46 m – 76 m thick. Core from the northeastern region of the Louisiana panhandle illustrates the character of the more shelf edge located deposits of the Lower Tuscaloosa interval which are tied closely to detailed descriptions of the deltaic facies in Chapter 3. Logs from wells to the south of the paleoshelf edge show the Lower Tuscaloosa section expanding to more than six times its thickness over very short distances, indicating that the Lower Tuscaloosa was highly affected by growth-fault expansion at its shelf edge (Figure 2.2). This observation is corroborated by observations on 2D seismic data (Figure 2.11).

Well log patterns in the Stringer sand show a “ratty” pattern, with smaller, hourglass-shaped curves indicating a more heterolithic deposit, with clays increasing upward. Sands are thin and become “dirtier” upward, with an overall increasingly serrated motif. These deposits reflect an increasing magnitude of transgression and more heterolithic, transgressive, coastal-plain environments. The very top of the Lower Tuscaloosa is sometimes capped by the Pilot sand, which has been interpreted to be a marine-bar complex (Berg and Cook, 1968; Mancini et al., 1980; Hamilton and Cameron, 1986). The intermittency of its occurrence is typical of offshore marine-bar complexes (Davies et al., 1971).

Offshore deposits interpreted to be basin-floor fans have been penetrated in the Tuscaloosa interval in BAHA #2 (Meyer, 2007) and Tiber wells and interslope basin deposits have been interpreted in data from the Davy Jones #2 well (McMoRan Exploration Co., 2011). The facies and log character of these wells will be discussed in detail in Chapter 3.

DISCUSSION

Isopach Trends and Paleogeography

The Massive sand ranges at its thickest from 61 m to152 m to less than 6.1 m and is absent to thin over paleohighs. Gross isopach maps, as well as individual sequence isopach maps show these sediments to be supplied dominantly by a group of fluvial valleys with drainage basins originating from north and northeast of the study area. The total isopach of the Massive sand shows four important things: (1) there is a clear east to west oriented thickness trend that demarcates northern versus southern architectural

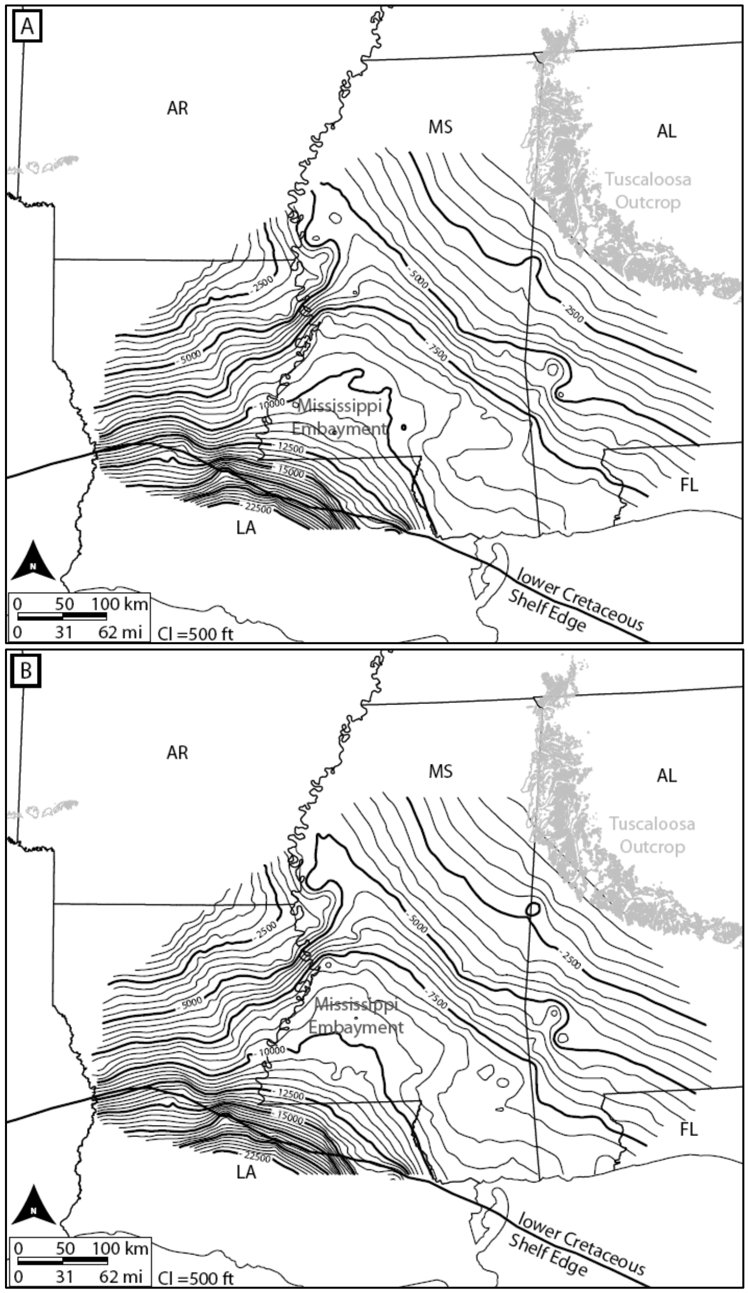

Figure 2.10. Structure maps on base of Lower Tuscaloosa Formation (A) and top of S3 (B). The Lower Tuscaloosa Formation ranges in depth from 0 ft. at outcrop in Alabama to more than -22,500 ft. beyond shelf edge in central Louisiana. Structure map also highlights location of the Mississippi embayment reentrant in southwest Mississippi. Note that the overall structure on these surfaces is similar indicating little change in structure during deposition of the Lower Tuscaloosa.

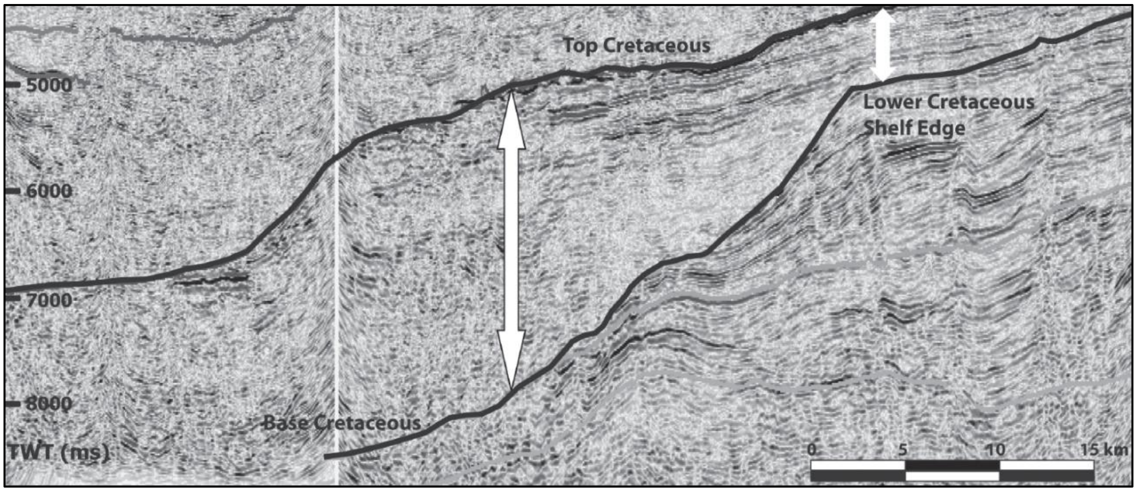

Figure 2.11. Amplitude 2D seismic line showing interpreted Cretaceous interval and Lower Cretaceous Shelf Edge. This image indicates about 4 times expansion of the Lower Tuscaloosa strata over the shelf edge in this area; however, the well log cross section in Figure 5 indicates other areas may have up to six times expansion. Data courtesy of ION Geophysical Inc.

regions in the depositional patterns of the Massive sand, and this feature is likely an old paleo shoreline trending from the southern edge of the Monroe uplift in northeast Louisiana through the Jackson dome across the east-west center line of Mississippi and Alabama, (2) the uplifts of the west and east Wiggins arch and the Baldwin high form an echelon series of highs that dam Massive-sand sediments north of these highs in southern Mississippi and Alabama, only allowing sediments to escape southward along narrow pathways between these uplifts, (3) Massive sand sediments thicken to the east away from the Louisiana border and (4) there appear to be three main depocenters for Massive sand sediments - a thinner basin that we term the southwest Mississippi basin and a much thicker depocenter that we term the Mississippi Panhandle basin and a still further eastward third depocenter that we term the Alabama-Florida Panhandle basin.

The five sequences of the Lower Tuscaloosa Formation; S0, S1, S2, S3, and the Stringer, youngest to oldest, respectively, individually and collectively map in depositional patterns that appear as fluvial to estuarine to deltaic in patterns (Coleman and Wright, 1975; Galloway, 1975). In addition, individual isopach maps show increasingly northward migrating depocenters over time, with more southward deposits from older sequences reworked into strike elongate sand bodies in the younger sequences. These later deposits form some of the best hydrocarbon fields in the Lower Tuscaloosa trend (Berg and Cook, 1968; Hamlin and Cameron, 1987; Wiygul and Young, 1987). Each sequence isopach map will be discussed in detail below, followed by a summary of the observations.

S0 Sequence

The S0 (Figure 2.4) represents the basal-most sand unit deposited directly over the top of the lowstand unconformity, known regionally as the mid-Cenomanian or

Middle Cretaceous unconformity and is the thinnest sequence of the Lower Tuscaloosa. This is likely due to the mixed bypass-aggradational nature of the sequence in keeping with its nature as a late lowstand to early transgressive fill of the Lower Tuscaloosa shoreline estuarine system. Four major fluvial axes that originate from the eastern ancestral Appalachia highlands and the central paleo-Mississippi embayment are identifiable in the S0 isopach map. A possible fifth system trends through northeastern Louisiana sourced from the paleo-Ouachita Mountains. Valleys in the Lower Tuscaloosa system range from 30 to 80 km wide in their fluvial/northern reaches widening in estuarine reaches in central and south Mississippi and Alabama to over 200 km. The northern reaches show distinct, thick sediment accumulations where nearly 15 m amalgamated channel sand thicknesses occur along valley margins and near the confluence of the feeder and the estuarine bay. Likewise, the estuarine bays that back up behind the long-lived west and east Wiggins arch and the Baldwin high are filled with through-going channels and bars, interspersed with thin interbar-complexes. The entire S0 complex terminates basinward in at least four major late lowstand shelf edge deltaic complexes with delta lobes amalgamating to thicknesses of up to 40 m. Lack of data preclude mapping the termination of the most eastward of the S0 valley systems. However, it is an extremely well-developed feeder system that likely hosts a well-developed shelf edge component south of the Alabama and Florida panhandles.

S1 Sequence

The S1 interval (Figure 2.5) marks a time of initial backstepping of the coastal system, initiating the reworking of previous S0 lowstand shelf edge deltas into transgressive barrier mouth bars that fill the distal openings of a thin transgressive estuary system couched northward of the west and east Wiggins arches, and the Baldwin high.

Distinct accumulations of thick sands occur immediately up depositional dip (north) from these highs. The S1 sequence, which represents a phase of transgressive fill in the previously defined S0 valley system, shows distribution patterns similar to its enclosing valleys with thick accumulations trending north-northeast from the shoreline, and a western system trending across the eastern Louisiana-western Mississippi region that links basinward to the most westward reworked shelf edge delta.

S2 Sequence

The S2 interval is much thicker than the underlying S0 and S1 sequences. Depocenters at shelf edge locations show thicknesses of up to 52 m. Depositional patterns of the S2 show a strong transgressive back-stepping character in the central Mississippi system while reflecting rejuvenation of the eastern Mississippi-Alabama system that continues to pond thick sediments northward of the east and west Wiggins arches and the Baldwin high. A large central mud basin characterizes the S2-paleo estuary of southeastern Mississippi (shown in green on Figure 2.6) and is reflected in the logs from this area's S2 interval. Sands near the shelf edge, located in front of this basin appear reworked into a wave-dominated estuarine mouth barrier. A western feeder system along the eastern Louisiana border, is adding sediments to the shoreline at this time, feeding a series of elongate sand bodies along the Louisiana portion of the paleo-Cretaceous shelf edge. In eastern portions of the study area, seven to eight separate, smaller, backstepping transgressive deltas are deposited along the Mississippi to Alabama to Florida Panhandle corridor, interspersed among the various highs that mark the region (Figure 2.6).

S3 Sequence

S3 time (Figure 2.7) shows a paucity of much sediment near the shelf edge or even basinward of the uplifts along the shelf and shelf break. A few thick pods of reworked shelf edge deltaic deposits persist near the shelf break, but the majority of sediment at S3 time is harbored in the paleo-estuary of southeastern Mississippi and located over 100 km landward of the shelf break in eastern locations. Valleys still fed these depocenters overlying the Mississippi central salt basin and the northern panhandles of Mississippi, Alabama, and Florida, and there is a valley fill associated with this S3 interval as far north as northern Mississippi. Elongate (north-south) isopach patterns within the paleoestuaries reflect sediments under increasing influence of transgressive tidal forces, being reworked into elongate bar/ bar complexes and channel/intrabar regions.

Stringer Sequence

The final transgressive sediment episode in the Lower Tuscaloosa; the Stringer sand, reflects the most transgressive appearing sequence of the all those mapped (Figure 2.9). By Stringer time, sedimentation in the southwestern Mississippi region has stepped northward and is almost entirely confined to the northern reaches of the central paleo-Mississippi feeder valley between the Monroe Uplift and the Jackson Dome. Sediments remaining southward of this system have been reworked into east-west elongate bodies within the distal reaches of a large flooded estuary. The eastern half of the study area is characterized by a thick, east-west oriented strandplain/barrier fed by a series of eastern-sourced river systems that stack over 122m of transgressive shoreline and shelf deltaic sediments along the northern sides of the east and west Wiggins arches.

Lower Tuscaloosa Formation Summary

In summary, the Lower Tuscaloosa was deposited during a period of transgression following a major mid-Cenomanian lowstand. Two major regions of deposition, a southeastern Mississippi-southwestern Alabama system and a southwestern Mississippi system are fed by a series of large 30-80 km wide incised valleys dominantly draining the central and eastern paleo-Appalachian highlands. Sediment accumulation, initially widespread during S0 time, shifts from western location during S1 and S2 time to eastern locations during S3 and Stringer time. Such a shift is possibly reflective of increased subsidence in the Central Mississippi salt basin, increased sediment supply from eastern sources, or decreased sediments from the central Mississippi drainage systems during S3 and Stringer time. Deltaic progradation of the Lower Tuscaloosa sequences occurs along distinct corridors between major uplifts in the region, with sediment accumulations sequestered behind uplifts and in shelf edge locations immediately basinward of these corridors. Sediment accumulations show increasingly elongate geometries as transgression proceeds, forming east-west oriented strandplains and reworked shelf deltas, and north-south oriented bars and channels within flooding estuaries.

Depositional History

The change in thicknesses and locations of depocenters for each sequence, S0 through Stringer, indicate a great deal about the depositional history of the Lower Tuscaloosa and support the interpretation of the Lower Tuscaloosa being a widespread transgressive fill. Additionally, a change upsection from braided to meandering fluvial systems to nearshore-marine and marine-bar-complex systems, as well as the increased backstepping of depocenters as indicated in isopach maps all suggest large-scale transgression during Lower Tuscaloosa time. The depocenters of each successively younger sequence shift landward. This northward (i.e., landward) temporal movement of depocenters is likely the result of larger scale transgression, which successively shifted depocenters northward, combined with initiation of salt mobilization in the older, Jurassic-age, Mississippi salt basin and generation of withdrawal and accommodation space for sediments to fill. These events combine to generate the thicknesses of the Lower Tuscaloosa sequences.

Actual spatial distribution of sediments within each Lower Tuscaloosa sequence appears to be influenced most by major topographic features; the shelf edge, the uplifts of the region, and the major salt basins. The underlying mid-Cenomanian unconformity, caused by a large drop in regional base level, created a high degree of differential topography including a major incised valley system. The initial rise in base level resulted in deposition of S0 within paleogeographic lows created by that unconformity (Figure 2.4). Therefore, the S0 sequence is commonly absent over highs and interfluves. Likewise, S1 time sediments appear to be mostly confined to the underlying incised lows of the S0 valley systems (Figure 2.5). However, by S2 time sediments are filling more proximal accommodation and depositing as transgressive shelf edge and shelf deltas (Figure 2.6). Although S0 and S1 time does show development of shelf edge deltas their thickness is limited, suggesting that during deposition of S0/S1, sediments were bypassing mainly into the deeper basin. Tuscaloosa-age clastic sediments observed in the offshore in BAHA #2 (Meyer, 2007), Tiber, and Davy Jones #2 (McMoRan Exploration Co., 2011) wells suggest that a time of early S0/S1 bypass of clastics into the basin indeed occurred. The amount of sediment located at the paleoshelf edge decreases most dramatically during S3 time. S3 thickening in both eastern Mississippi and western Mississippi in the proximal direction suggests a time of less bypass basinward and a transgressive landward shift in depocenters (Figure 2.7). An increase in subsidence within the Mississippi salt basin enhanced ongoing transgression in the western Mississippi region during Stringer time, resulting in a complete shift of sedimentation in those areas into more northern valley reaches, and shifting of more southern depocenters to southeastern Mississippi.

Structural influences on sediment distributions

Three major structural "features" influence the sequence stratigraphy of the Lower Tuscaloosa Formation. These are (1) The Cretaceous shelf edge, (2) The localized highs situated east to west including the Baldwin high, the east and west Wiggins, and the Louisiana arches, and (3) the Mississippi salt basin. The Mississippi salt basin works in tandem with the arches to create thick accumulation of estuarine/fluvial sediments in the Lower Tuscaloosa. Highs situated at or just north of the shelf edge create barriers to free movement of fluvial systems basinward, forming narrow outlets for sediments to move to shelf edge locations. Such funneling of deltas through narrow uplifts, cause changes in the cross sectional geometry of both the lobes and channels. The changes can cause the rivers and deltas to deposit sediments behind these highs that would be meant for more distal locations, and encourages the formation of significant depocenters landward of the shelf edge. Such "thieving" of sediment by more proximal depocenters is enhanced when accompanied by the accommodation provided by salt withdrawal in the region. The result of these tandem processes is significant thicknesses of fluvial/estuarine fill in the Lower Tuscaloosa situated in depocenters located in central and southeast Mississippi.

Paleogeography

The apparent structural stability of the south central and southeastern Gulf Coastal region is evidenced by the lack of temporal changes in structure maps (Figure 2.10). This stability means that basement uplifts remained high throughout deposition of the Lower Tuscaloosa, actively influencing geologic processes, both structurally and sedimentologically (Stephens 2001, 2009), at least through Lower Tuscaloosa time. This consistency of basement cored topography provided a structurally consistent paleogeographic setting that resulted in long-lived locations of valleys and sediment fairways. The separation of the Wiggins arch by Stephens (2009) into a western Wiggins arch and an eastern Wiggins arch (our terminology) separated by the Pearl River transform fault appears to be supported by the work done in this study. This break in the Wiggins arch acts as a conduit for sediments moving basinward through the southwestern tip of the Mississippi panhandle that is consistent through all of the Lower Tuscaloosa sequences. Only during Stringer time, does the eastern Wiggins arch appear to become a major site for sediment accumulations, suggesting a final burial by late stage transgressive deltas.

Deepwater Facies of the Lower Tuscaloosa

Sediments feeding potential eastern offshore Gulf of Mexico areas during Lower Tuscaloosa time were most likely hosted in the Mississippi panhandle region and at the shelf edge in Louisiana. The best potential time for sediment bypass in the Tuscaloosa Formation appears to have been early in Lower Tuscaloosa time (S0 and S1 times), coinciding with movement of sediments over the mid-Cenomanian (basal sand) unconformity. Core data and log-motif analysis, combined with regional fairway mapping, indicate that coarse-grained fluvial sediments were moved north-northeast to south-southwest, with minimal sequestering of shelf-edge or deltaic deposits in S0 and S1 times. S0 and S1 sediments would have had to traverse significant accommodation in the upper slope to move into Lower slope and abyssal locations. However, such accommodation sinks are not dissimilar to that observed in the Wilcox Formation system moving sediment to basinward locations nearly 300 km from its own shelf edge (Lewis et al., 2007). Distal shelf-edge core from the Lower Tuscaloosa suggests that these sediments fined rapidly from coarse-grained sediments of the braided and meandering Lower Tuscaloosa to medium- to fine-grained sediments at shelf-edge locations. Gravity flow deposits from these sediments could be far traveled. This is in the realm of possibility as even coarse-grained turbidites are known to move great distances down deepwater slopes, through complex topography (Chaderton and Wood, 2007).

The distal-most facies penetration of Tuscaloosa-equivalent sediments found in the Gulf of Mexico lie in the western Gulf, where the BAHA #2 well (Meyer, 2007) in the Perdido fold and thrust belt and the Tiber well in the Keathley Canyon area penetrated Tuscaloosa-age gravity flow deposits. Although the age and character of these sediments is not debated, the source area remains a question. Gravity flow deposits originating from Louisiana and eastern Gulf of Mexico systems would have had to travel a relatively long distance before settling in the western Gulf of Mexico. Alternatively, sediment may have passed through the East Texas basin, feeding the Perdido basins from Woodbine-named shelf edges in the vicinity of present-day Houston, Texas (Ambrose et al., 2009). Although the proximity of this source to the Perdido fold and thrust belt is attractive, the volume of sediments being transported through the East Texas basin is smaller than that of the systems documented in this paper occurring in the eastern Gulf of Mexico. This limited east Texas sediment volume may yield small possibilities of long-traveled gravity flows. A third, but as yet uninvestigated, alternative is a source through northern Mexico and into the Perdido foldbelt from the south. Additional work remains to be done, but recent wells have proven the existence of more distal clastic deposits associated with the Lower Tuscaloosa, which may lead to the next big deepwater play in the Gulf (Lewis et al., 2007; McMoRan Exploration Co., 2011).

Provenance

The provenance of the Lower Tuscaloosa Formation appears to be dominantly the ancient Appalachian highlands (Figure 2.2). There is a component that appears to be coming from an eastern Ouachita source, but the majority of valleys are trending from the east, and even the large, north-trending feeder valley of the central Mississippi paleo-estuary is likely receiving sediments from the combination of a northern and eastern sources. Such provenance issues may have some importance in understanding the reservoir nature of the Lower Tuscaloosa. Numerous authors have pointed out the importance of chlorite cement in preserving porosity, through inhibiting quartz cementation and, thus, prospectivity in the Lower Tuscaloosa (Thomson, 1979; Hearne and Lock, 1985; Hamlin and Cameron, 1987; Genuise, 1991; Hansley, 1996; Ryan and Reynolds, 1996; Weedman et al., 1996; Ambrose et al., 2009). However, the origin and extent of these chlorites are still up for debate. Several of these authors have suggested the origin of these chlorites to be volcanic rock fragments from northwestern sources. Although Upper Cretaceous volcanics are found in south Arkansas that can potentially supply such constituents to Lower Tuscaloosa depositional systems, mapping shows that the Lower Tuscaloosa rivers dominantly originated from the east and north. Whereas secondary input from the northwest (Ouachita region) dominantly appears in S3 and Stringer sand cycles, earlier cycles (S0, S1) appear to have a lesser influence from northwestern sources. Chapter 5 addresses the issue of the origin and controls on reservoir quality and chlorite distribution from detailed petrographic analysis of the Lower Tuscaloosa Formation.

CONCLUSIONS

The mid-Cenomanian in the Gulf of Mexico Basin marked a time of transition across much of North America from a carbonate-dominated depositional system to a clastic-dominated system. The Lower Tuscaloosa Formation, a thick, fluvial-deltaic to nearshore marine deposit in the north and east margins of the Gulf of Mexico marks this transition. Five sequences (S0, S1, S2, S3, and Stringer) were mapped in the Lower

Tuscaloosa using 668 well logs across Louisiana, Mississippi, Alabama, and Florida in the onshore and nearshore areas of the Gulf region indicating the Tuscaloosa Formation was sourced dominantly from the ancient Appalachians to the northeast and from a large drainage trending within the ancestral Mississippi valley that may have been similarly sourced from the ancient Appalachian plateau. Several thick fairways of Lower Tuscaloosa sediments terminate southward at the highly extensional paleoshelf edge. Basement-rooted and long-stable structural uplifts, including the east Wiggins and the west Wiggins arches, the Baldwin high, the Jackson dome, the Monroe uplift, and the Louisiana arch influenced drainage directions and formed buttresses that resulted in ponded thicknesses of fluvial, estuarine, and coastal plain deposits. This same Late Cretaceous topography influenced the distribution and magnitude of processes active along the transgressive-marine shoreline. Deepwater, low-density gravity flow deposits of the Lower Tuscaloosa were hosted from depocenters that formed in the Late Cretaceous in areas of the modern-day Mississippi and Alabama panhandles and areas of Louisiana. The sheer volume of sediments, coupled with recent drilling penetrations of Tuscaloosa-age slope facies in the modern Louisiana shelf, and basin floor fan facies in the Perdido foldbelt of the western Gulf, suggest the strong possibility of a much broader regional occurrence of deepwater facies Tuscaloosa Formation plays associated with the basal sequences (S0 and S1) of the Lower Tuscaloosa. Chapters 3, 4 and 5 use the provenance and sequence frameworks that have been developed here to examine sedimentology, facies, depositional systems, fluvial channel character, and petrography of the Lower Tuscaloosa reservoir systems including implications for regional reservoir quality.

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Chapter 3: Facies Analysis and Depositional Systems of the Lower Tuscaloosa Formation

INTRODUCTION

Recent exploration of new plays in the Tuscaloosa-equivalent section of the offshore Gulf of Mexico has revealed possible Tuscaloosa-age gravity deposits in both slope and basin plain paleo-positions. This recent exploration activity has increased interest in the character and regional distribution of depositional processes responsible for formation and preservation of the Lower Tuscaloosa facies and depositional environments. However, a detailed account of the distribution of facies and depositional environments along the Lower Tuscaloosa Formation paleoshoreline is essential to refine both local and regional models and improve the ability to predict the nature and presence of more distal deposits.

The purpose of this chapter is to describe and identify paleoenvironments in the Lower Tuscaloosa Formation and examine the paleogeographic distribution of the systems responsible for emplacement of these sediments. Emphasis is placed on an onshore study area in eastern Alabama, Mississippi, and southeast Louisiana. In addition, three offshore wells that penetrate the Tuscaloosa interval (Davy Jones #2 (South Marsh block, offshore Louisiana), BAHA #2 (Perdido fold and thrust belt), and Tiber (Keathley Canyon area)) (Figure 3.1) are incorporated to offer a complete source-to-sink perspective of the Lower Tuscaloosa. This is the first study to integrate ichnology with detailed core descriptions and wireline well logs to refine Lower Tuscaloosa Formation facies and depositional environments. Results indicate that the majority of source-to-sink sediments of the Lower Tuscaloosa Formation were deposited in three depositional systems (1) incised valleys; both fluvial and estuarine-filled, (2) shelf edge deltas; both wave and tide-dominated, and (3) deepwater slope and basin floor fans. A fluvial-estuarine incised valley-fill model of the Lower Tuscaloosa Formation is proposed to explain local and regional variability in brackish- to fresh-water facies. It is suggested that the fluvial systems within incised valleys and related sediments stored in deltas of the Lower Tuscaloosa are the source for what are believed to be temporally equivalent gravity-driven deepwater deposits discovered in wells of the offshore deep Gulf of Mexico. Facies and depositional environments are closely tied to well log responses.

PREVIOUS WORK - DEPOSITIONAL ENVIRONMENTS

Based on core, cuttings, and wireline logs from both onshore and offshore wells in Alabama, Mississippi, and Louisiana workers have proposed numerous depositional environments for sediments of the Lower Tuscaloosa Formation (Chasteen, 1983; Mancini et al., 1987; Klicman et al., 1988; Miller and Groth, 1990; Corcoran et al., 1993; Petty, 1997). Although interpretations differ from field to field, likely driven by local and regional paleogeography, the consensus is that the overall succession is controlled by the transgressive nature of the Lower Tuscaloosa Formation. Both Chasteen (1983) and Klicman et al. (1988) provide an excellent summary of the Lower Tuscaloosa deposits in central Louisiana and southern Mississippi, classifying the base of the Massive sand as a braided fluvial system overlain by a meandering fluvial system. These fluvial systems are overlain by the transgressive Stringer sand, interpreted as shallow-marine to brackish-water deposits, which are conformably overlain by the fully-marine Middle Tuscaloosa shale (Chasteen,1983; Klicman et al., 1988). Petty (1997) identified several paleoenvironments from wireline log patterns of the Lower Tuscaloosa Formation in wells to the south and east of our study area in offshore Federal waters and state waters of Mississippi, Alabama, and Florida. He identified five sand-rich facies: meander belts, shoreline strandplains, stacked coastal barrier bars, distributary channels, and reworked

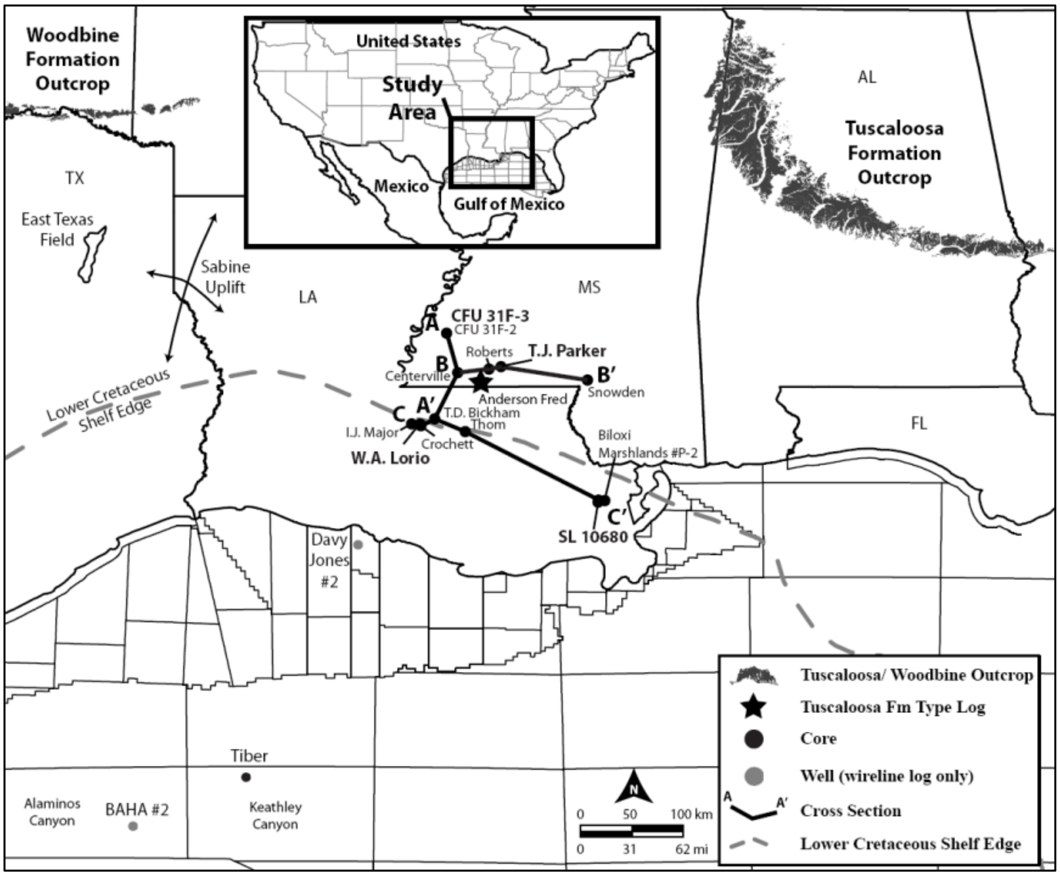

Figure 3.1**.** Regional map showing locations of the 13 core (Table 3.1), offshore wells (Davy Jones #2, BAHA #2, and Tiber), and cross sections presented in this paper. Also shown are the locations of the type log, Tuscaloosa and Woodbine Formation outcrops, Lower Cretaceous shelf edge, Sabine uplift, and East Texas field.

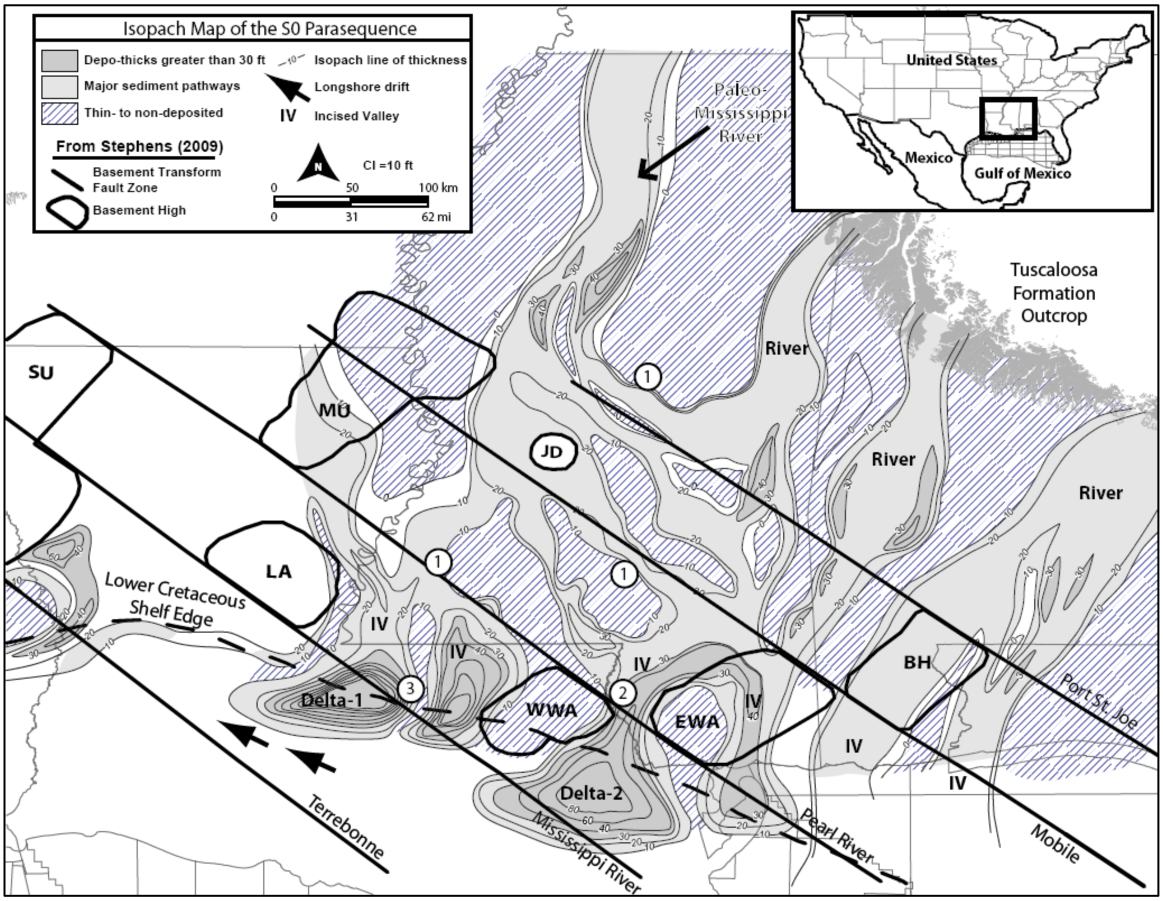

Figure 3.2. Isopach map of the basal sequence of Lower Tuscaloosa (S0) (Chapter 2). Main sediment fairways are shaded in gray. Incised-valleys (IV) are highlighted by main sediment fairways. Valleys generally trend from north - northeast to south – southwest. Overlain on isopach are major basement structures (diagonal black lines and bold polygons), including key basement highs and transfer faults as interpreted by Stephens (2009).

deltaic sands. Mancini et al. (1987) used core and wireline logs from Alabama to identify wave-dominated deltas in the Massive sand, and shelfal sands and shelfal clays in the Pilot sand (equivalent to the Stringer sand to the west) (Figure 1.3). Corcoran et al. (1993) and Barrell (1997) analyzed core, 2D and 3D seismic, well logs suites, and production information and identified a “deep Tuscaloosa” marine facies just to the south of the Lower Cretaceous shelf edge in Louisiana (the southern limit of our study area) as delta front and delta fan deposits. On the western side of the Sabine uplift in the East Texas field (Figure 3.1), Ambrose et al. (2009) used core and wireline logs to reinterpret the depositional environments of the Lower Tuscaloosa Formation-equivalent Woodbine Formation as lowstand valley-fills and fluvial-dominated deltas.

In Chapter 2 several incised valleys and valley-front deltas in isopach maps of the Lower Tuscaloosa Massive Sand (Figure 3.2). These incised valleys are thought to have formed during the drop in base level that also produced the mid-Cenomanian unconformity (Salvador, 1991). Valleys trend roughly north-northeast to south-southwest, are 30-80 km wide, and may be up to 100 m deep (Chapter 2). The valleys extend to the late Cretaceous lowstand shelf edge and appear to have developed between basement blocks that acted as structural paleotopographic highs (Figure 3.2, Chapter 2) (Stephens, 2001 and 2009). Two major shelf-edge delta systems, fed by the fluvial systems in the valleys, formed along the paleo shelf edge of the Lower Tuscaloosa Formation (Chapter 2, Figure 3.2). The northwestern delta, which is termed Delta-1 for this study, is located at the mouth of the valley that formed to the northwest of the west Wiggins arch in the Mississippi embayment. The southeastern delta, termed Delta-2, is located at the mouth of the valley that formed between the west Wiggins Arch and the east Wiggins arch. The recognition of these incised valleys and deltas in the subsurface, as well as mapping of the sequences that fill the valleys, along with the regional paleogeography outside of these valleys is the framework that our detailed sedimentologic and ichnologic study is built upon.

METHODS

Wireline logs (n=670) and conventional cores (n=12) from five onshore wells in southeastern Mississippi and seven onshore wells from the Louisiana panhandle were used in conjunction with wireline logs from two offshore wells (Davy Jones #2, and BAHA #2) and wireline logs and core photographs from one offshore well (Tiber) in the Tuscaloosa Formation equivalent section (see Figure 3.1 for locations; Table 3.1 for well data). Although original core were not available from the Tiber well, core photographs were interpreted and used for facies analyses. Core descriptions for the CFU 31F-2, T.D. Bickham, Roberts, Centerville, Thom, Crochett, I.J. Major, and Biloxi Marshlands #P-2 wells can be found in Appendix II. Where whole core was preserved, all available cored intervals were photographed using a Nikon D300s camera and a 16-85 mm Nikkor lens under daylight-corrected lighting (5000K). Photographs and core descriptions identifying lithologies, textures, grain sizes, sedimentary structures, trace fossils, and fauna were used as the basis for identifying facies, paleoenvironments, and depositional systems. Facies were grouped into facies associations to identify depositional environments. Trace fossils were identified using examples from literature (Nichol et al., 1997; McIlory, 2004; Bhattacharya, 2006; Boyd et al., 2006; MacEachern and Bann, 2008; Carmona et al., 2009) and were employed to refine paleoenvironmental interpretations. Wireline log patterns were used to cross-check interpreted depositional environments, measure thicknesses, and produce cross sections. Cross sections were created from core and well-logs, enabling this chapter to be integrated within the regional mapping of Chapter 2. Facies models of incised valleys and valley-fills (Dalrymple et al., 1992; Allen and

Table 3.1. Summary of core used in this study including: location, interval cored, interpreted depositional environment, and location of the core description in this paper. S1, S2, S3 refer to the individual sequences of the Massive sand discussed in Chapter 2.

Well Name	API	Location	Cored Interval	Depositional Environment	Core Description
CFU 31F-3	23-037-21486	Cranfield Field, Adams CO, MS	S2, S3, Base of Stringer	Incised Valley, Fluvial	Figure 3.3
CFU 31F-2	23-037-21485	Cranfield Field, Adams CO, MS	S2, S3, Base of Stringer	Incised Valley, Fluvial	Appendix IIa
T.D. Bickham	17-033-20040	Port Hudson Field, East Baton Rouge PA, LA	S1, S2, S3	Incised Valley, Fluvial	Appendix IIb
T.J. Parker	23-005-00191	McElveen Field, Adams CO, MS	Stringer Sand	Incised Valley, Estuarine	Figure 3.4
Centerville	23-005-20481	Thanksgiving Field, Adams CO, MS	Stringer Sand	Incised Valley, Estuarine	Appendix IIc
Roberts	23-005-00193	East Fork Field, Adams CO, MS	Stringer Sand	Incised Valley, Estuarine	Appendix IId
W.A. Lorio	17-077-20170	False River Field, Pointe Coupee PA, LA	S0, S1, S2, S3	Deltaic	Figure 3.8 Appendix IIi
SL 10680	17-087-20265	St. Bernard PA, LA	Unknown	Deltaic	Figure 3.9
Thom	17-063-20052	Lockhart Crossing Field, Livingston PA, LA	Unknown	Deltaic	Appendix IIe
I.J. Major	17-077-20250	Judge Digby Field, Pointe Coupee PA, LA	S1	Deltaic	Appendix IIf
Biloxi Marshlands P-2	17-087-20255	St. Bernard PA, LA	S0, S2, Middle Tuscaloosa	Deltaic	Appendix IIg
Crochett	17-077-20160	False River Field, Pointe Coupee PA, LA	S3, Stringer, Middle Tuscaloosa	Deltaic	Appendix IIh
Tiber	60-808-40015	Keathley Canyon	Lower Tuscaloosa	Deepwater, Gravity-driven	Figure 3.14

Posamentier, 1993; Nichol et al., 1997; Boyd et al., 2006), deltas (Galloway, 1975; Porebski and Steel, 2003; Bhattacharya, 2006), and gravity-driven deepwater deposits (Lowe, 1982; Shanmugam, 1996; Posamentier and Walker, 2006) were used as comparisons to produce depositional models. Depositional models were inserted into the paleogeomorphologic framework created in Chapter 2 to cross-check paleoenvironmental interpretations.

FACIES ANALYSIS

A total of thirteen cored intervals (twelve onshore wells across southern Mississippi and the Louisiana panhandle and the Tiber well in which core photographs were available) and wireline logs from three wells in the offshore Gulf of Mexico were used to define the facies and facies associations of the Lower Tuscaloosa Formation (see Figure 3.1 for locations; Table 3.1 for wells). These cores were selected based on the locations of incised valleys and shelf edge deltas identified in isopach maps (Chapter 2) while the offshore wells represent the only well penetrations of the deepwater facies of the Tuscaloosa Formation. Six of the cored wells were from two of the incised valleys identified in Chapter 2 and six were taken from outside the valleys in the unconfined, deltaic portion of the system (Table 3.1). Additionally, cores from both the Massive and Stringer sands were selected to determine any difference in facies between these two lithostratigraphic units. Sediments were separated into four unique depositional systems, two of which are confined within incised valleys (Fluvial Channels and Floodplains-Estuaries) one of which occurs only in the unconfined environment immediately basinward of valley mouths (Deltas), and the last which occurs only in deepwater settings (Gravity-Driven Deepwater Deposits).

Fluvial Channels and Floodplains

Three of the thirteen cored intervals were observed to occur in probable fluvial environments within incised valleys (Table 3.1). Figure 3.3 detailing the Cranfield CFU 31 F-3 core is included for reference. Descriptions of the two other cores exhibiting fluvial facies (CFU 31F-2 and T.D. Bickham) can be found in Appendix IIa and IIb. The key to all log descriptions including the abbreviations used for identified trace fossils in core descriptions and core photographs are found in Figure 3.4. Nine facies were identified and grouped into two facies associations (Figures 3.3, 3.4; Tables 3.2, 3.3; Appendix IIa, IIb).

Facies

FF-1 is a chert pebble conglomerate (Figure 3.5A) found at the base of fining upward successions. Chert pebbles in FF-1 are often imbricated. The basal contact of FF-1 is erosional and the upper contact is gradational. FF-1 is interpreted as a high energy, channel thalweg bedload deposit.

FF-2 through FF-6 typically overlie facies FF-1 and mostly occur within fining-upward successions. These facies display a variety of sedimentary structures that are likely dependent on a combination of their spatial and temporal location and the energy of the system. Facies include high- to low-angle, planar cross stratified sandstone (FF-2, Figure 3.5B), planar-laminated sandstone (FF-3, Figure 3.5C), current ripple-laminated sandstone (FF-4, Figure 3.5D), massive sandstone (FF-5), and bioturbated sandstone (FF-6, Figure 3.5E, F, G). The base of these sands may contain rare imbricated or floating chert pebbles and/or mud rip-up clasts. FF-4 contains rare climbing ripples (Figure. 3.5D). The basal contacts of each of these sandstones may be erosional, sharp, or gradational, while the upper contacts are gradational to sharp. No bioturbation was recorded in FF-2, 3, 4, or 5. In core from the T.D. Bickham well (Appendix IIb), located proximal to the paleo-shelf edge (see Figure 3.1 for location) a low-abundance trace fossil assemblage was found in FF-6 (Figure 3.5E, F, G). Ichnogenera in FF-6 include *Asterosoma*, *Chondrites*, *Cylindrichnus*, *Ophiomorpha*, *Palaeophycus*, *Planolites*, *Phycosiphon, Rosselia*, *Teichichnus*, and *Thalassinoides* (Figure 3.5E-G). FF-6 also contains rare mollusks. Fluvial facies FF-2 to FF-5 are interpreted as sandy channel-fills and sandy barforms of a channelized fluvial system. FF-6 is interpreted as a marine influenced, bioturbated, reworked sand originally deposited as a fluvial sand in an incised valley.

The siltstone facies (FF-7) is divided into two distinct subfacies, a mottled and bioturbated siltstone (FF-7a, Figure 3.5H) and a rippled siltstone (FF-7b). FF-7a occurs at the top of the succession in cores from the Cranfield area of western Mississippi (see Figure 3.1 for location) and contains a red/green mottled fabric along with rare *Planolites* and *Rhizocorallium*. The lower contact of FF-7a is gradational and the upper contact is not present in core. FF-7b is a rippled siltstone that lacks mottles or bioturbation and has a sharp basal contact. The upper contact may be truncated by the basal erosion surface of FF-5. FF-7b may also contain rare mollusk shells. FF-7a and FF-7b are interpreted as a thin bar-capping or overbank deposits. The upper surface of FF-7b is often eroded into by FF-5, suggesting that it was deposited in a channel setting or a near channel environment.

FF-8 is thinly-laminated shale (Figure 3.5I) with a sharp basal and upper contact and is interpreted as an overbank deposit or a muddy abandoned channel-fill. FF-9 is a coal containing abundant carbonized wood and plant fragments (Figure 3.5J) that is frequently interbedded with shale or mudstone (FF-8). Both the basal and upper contacts of FF-9 are sharp. FF-9 is typically found in close association with the bioturbated sandstone facies (FF-6) and is interpreted as a backswamp deposit.

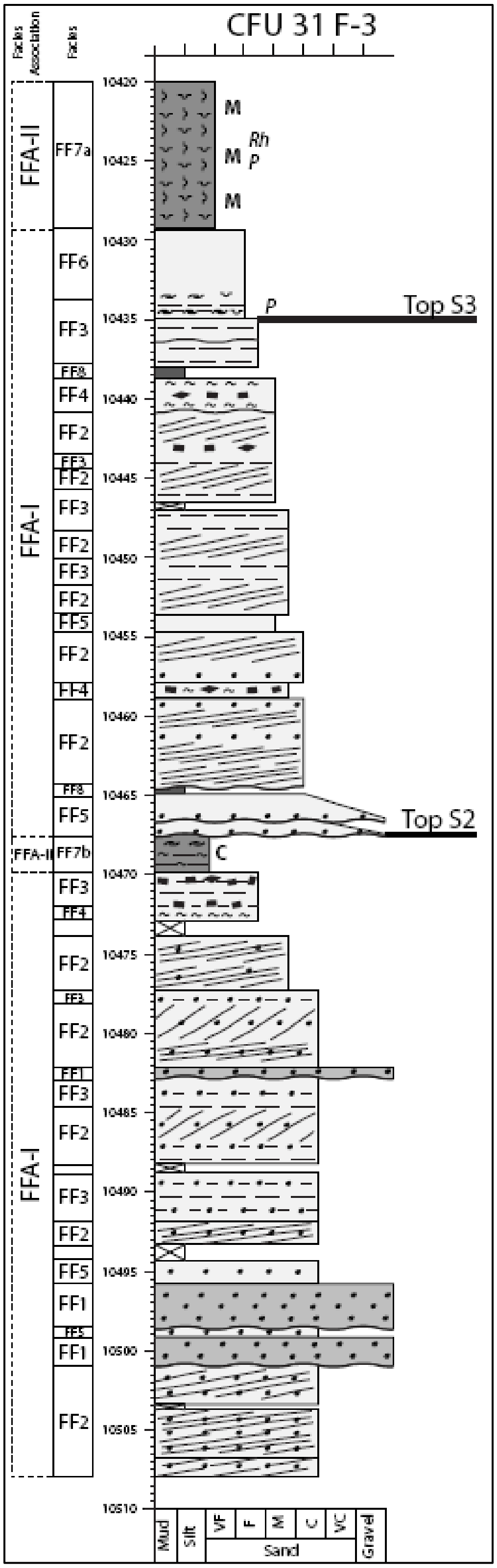

Figure 3.3. Core description of the Cranfield CFU 31F-3 well in southwest Mississippi (see Figure 3.1 for location). Figure contains interpreted facies and facies associations. Detailed descriptions of facies (FF) included in the Fluvial Channels and Floodplains Facies Association (FFA) are listed in Tables 3.2 and 3.3. Key for core description is found in Figure 3.4.

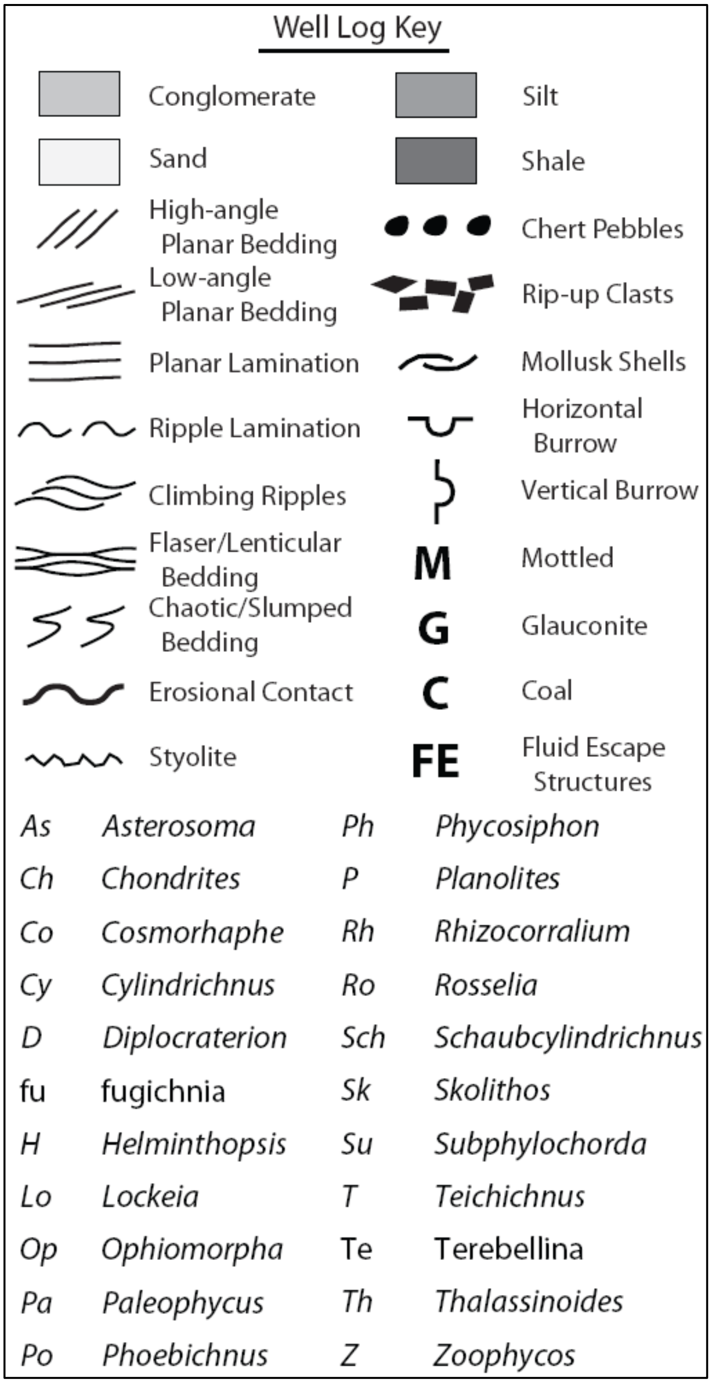

Figure 3.4. Key for core descriptions (Figures 3.3, 3.6, 3.8, 3.9, 3.11) and abbreviations for traces fossils used in Tables (Tables 3.2, 3.4, 3.6, 3.8) and core photographs (Figures 3.5, 3.7, 3.10).

Table 3.2. Description and interpretation of the nine fluvial facies identified in three core of the Lower Tuscaloosa Formation.

Facies		**Sedimentary Structures**	**Flora / Fauna**	**Ichnol ogy**	**EOD**
<u>FF-1</u> Pebble to Gravel Conglomerate		<u>Common</u> Imbricated chert pebbles	None	None	Channel thalweg bedload deposit
<u>FF-2</u> Fine- to coarse-grained high-to-low angle planar cross stratified sandstone		<u>Common</u> Imbricated chert pebble Organic rich laminations Mud rip-up clasts <u>Rare</u> Floating chert pebbles	None	None	Dune form channel deposits on a unit bar
<u>FF-3</u> Fine- to coarse-grained planar laminated sandstone		<u>Common</u> Imbricated chert pebble lags Mud rip-up clasts	None	None	Bed load sheet migration in the upper unit bar
<u>FF-4</u> Fine- to medium-grained ripple laminated sandstone		<u>Rare</u> Mud rip-up clasts Climbing ripples	None	None	Sand deposition near channel bank in lower energy environment
<u>FF-5</u> Fine- to medium-grained massive sandstone		<u>Common</u> Basal chert pebble lags <u>Rare</u> Mud rip-up clasts	None	None	Undifferentiat ed channel deposits

Table 3.2 continued

FF-6 Very fine- to fine-grained bioturbated sandstone		Common Ripple lamination	Common Bioturbation Rare Mollusk shells	*Op*, *T*, *Ro*, *P*, *Th*, *Ch*, *Ph*, *Cy*, *As*, *Pa*	Marine influenced reworked sands originally deposited as fluvial sands in the incised valleys.
FF-7 Siltstone	**FF-7a** Mottled, bioturbated Siltstone	Common Massively bedded Red/green mottling	Rare Bioturbation	*P*, *Rh*	Floodplain overbank/upper bar deposits
	FF-7b Rippled Siltstone	Common Rippled cross lamination Rare Mollusk shells	None	None	Floodplain overbank/upper bar deposits
FF-8 Shale		Common Thin laminations	None	None	Floodplain overbank or abandoned channel-fill
FF-9 Coal		Common Mud interbeds	Common Carbonized wood and plant fragments	None	Backswamp

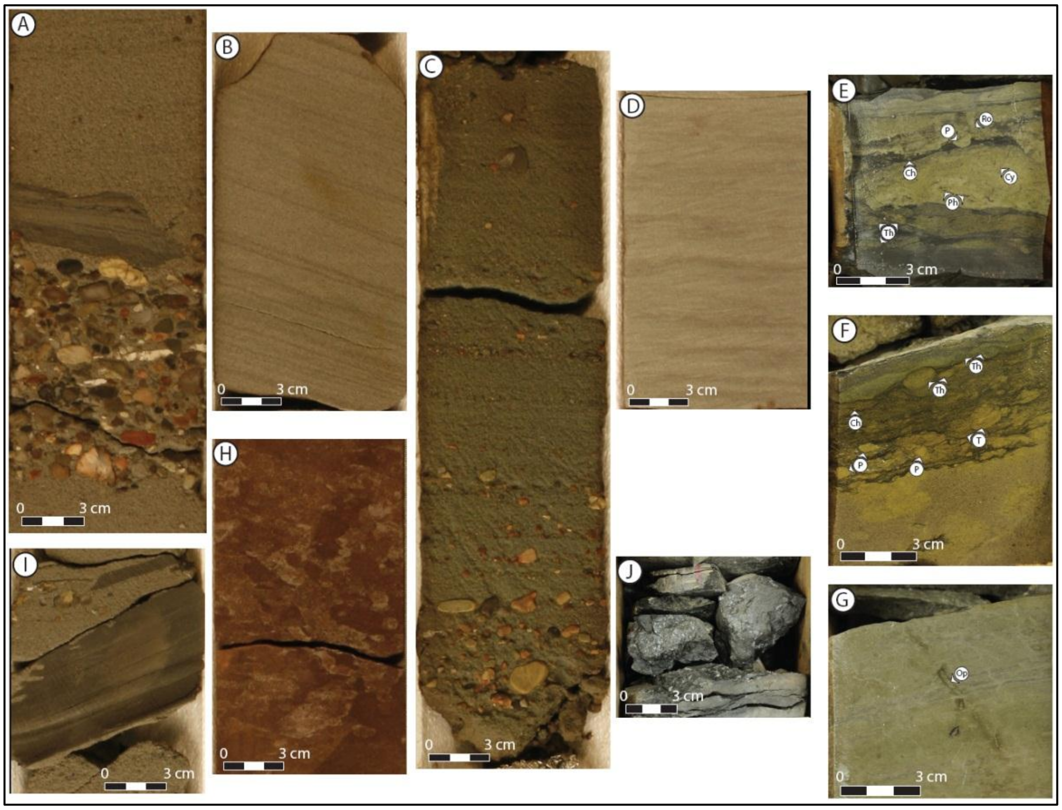

Figure 3.5. Core photographs of fluvial facies including identified trace fossils. (A) FF-1: chert pebble conglomerate; (B) FF-2: high- to low-angle planar cross stratified sandstone; (C) FF-3: planar-laminated sandstone; (D) FF-4: current ripple-laminated sandstone; (E, F, and G) FF-6: bioturbated sandstones interbedded with muddy beds; (H) FF-7a: mottled siltstone; (I) FF-8: thinly-laminated shale; and (J) FF-9: coal including carbonized wood and plant fragments interbedded with shale and mudstone. Key for trace fossil abbreviations is found in Figure 3.4.

Facies Associations

FF-1 through FF-10 are grouped into two facies associations (Table 3.3), a Channel-Fill Association (FFA-I) and a Floodplain-Abandoned Channel-Backswamp Association (FFA-II).

The Channel-Fill Association (FFA-I) includes FF-1, FF-2, FF-3, FF-4, FF-5, and FF-6. Diagnostic features of FFA-I include a basal, erosional contact with overlying imbricated chert pebble lags and/or mud-rip up clasts (FF-I). These basal facies are overlain by a fining upward succession that may be either erosionally-truncated by the base of another FFA-I succession or capped by shale or coal. FFA-I is similar to facies successions expected for fluvial channel-fill deposits (Miall, 1996; Bridge, 2003). These channel-fill successions are aggradational, dominated by sand and gravel, and lack evidence of lateral accretion surfaces or thick overbank successions suggesting that they are likely the deposits of braided channels (Williams and Rust, 1969; Miall, 1977, 1996; Bridge, 2003).

The Floodplain-Abandoned Channel-Backswamp Association (FFA-II) includes FF-7a, FF-7b, FF-8, and FF-9. FFA-II comprises siltstone, mudstone, and coal. Evidence for deposition of FFA-II predominantly on floodplains includes the fine grained nature of the deposits, small-scale ripples, and interbedding of coal and mudstone facies (Miall, 1977; Jones and Hartly, 1993). Mottling in F-7a suggests intermittent subaerial exposure in backswamp mudstones. The very limited occurrence of *Planolites* and *Rhizocorallium* in FF-7a capping the top of the fluvial facies of the Massive sand and underlying estuarine deposits of the Stringer sand indicate the transition from a fluvial to nearshore marine environment (MacEachern and Bann, 2008). Coals were likely deposited in backswamps (Jones and Hartly, 1993).

Table 3.3. Description and interpretation of the two fluvial facies associations identified in the three fluvial core of the Lower Tuscaloosa Formation. S1, S2, S3 refer to the individual sequences of the Massive sand discussed in Chapter 2.

Facies Association	**Paleo-environment**	**Inclusive Facies**	**Diagnostic Features**	**Occurrence**
FFA-I	Channel Fill Association	FF-1, FF-2, FF-3, FF-4, FF-5, FF-6, FF-8, FF-9	Fining upward cycles capped by shale, coal, paleosol, or erosional contact of overlying FFA-I cycle. Base of cycles often contain pebble lags and mud rip-up clasts	Port Hudson Field, East Baton Rouge Parish, Louisiana Cranfield Field, Adams County, Mississippi S1, S2, S3, Stringer
FFA-II	Floodplain/ Backswamp Association	FF-7a, FF-7b, FF-8, FF-9	Fine grained siltstone, mudstone, and coal. Caps FFA-I but is often missing due to erosion from the overlying FFA-I cycle. Some silts are heavily mottled.	Port Hudson Field, East Baton Rouge Parish, Louisiana Cranfield Field, Adams County, Mississippi S1, S2, S3, Stringer

Estuaries

Three of the thirteen cored intervals were identified as probable estuarine environments within incised valleys (Table 3.1). Figure 3.6 detailing the T.J Parker core is included for reference. Descriptions of the two other cores exhibiting estuarine facies (Centerville and Roberts) can be found in Appendix IIc and Id. Nine facies were identified and grouped into three facies associations (Figure 3.6; Tables 3.4, 3.5; Appendix IIc, IId).

Facies

Facies EF-1, EF-2, EF-3, and EF-4 are very fine- to medium-grained sandstones. Each facies displays a different dominant sedimentary structure that is likely dependent on a combination of their spatial and temporal location and the energy of the system. These facies include a planar cross stratified sandstone (EF-1, Figure 3.7A), a planar-laminated sandstone (EF-2, Figure 3.7B), a ripple cross laminated sandstone (EF-3, Figure 3.7C), and a massive-bedded sandstone (EF-4, Figure 3.7D). Basal contacts may be erosional, sharp, or gradational while upper contacts are sharp to gradational. EF-1, EF-3, and EF-4 all may contain mollusk shells and in one core (T.J. Parker) all three display abundant glauconite. EF-2 contains the trace fossils *Chondrites*, *Palaeophycus*, *Planolites*, and *Skolithos* (Figure 3.7C, D) while EF-3 rarely contains *Palaeophycus* and EF-4 rarely contains *Skolithos*. Estuarine facies EF-1 through EF-4 are interpreted as sandy barrier barforms in the seaward portion of an estuary.

EF-5 is the most heavily bioturbated of the very-fine sandstones and, unlike EF-1 through EF-4, bedding is either massive, disrupted, or absent. Both the basal and upper contacts of EF-5 range from sharp to gradational. Rarely, ripple cross laminations, mud rip-up clasts, and red to green mottling are observed. The low diversity trace fossil

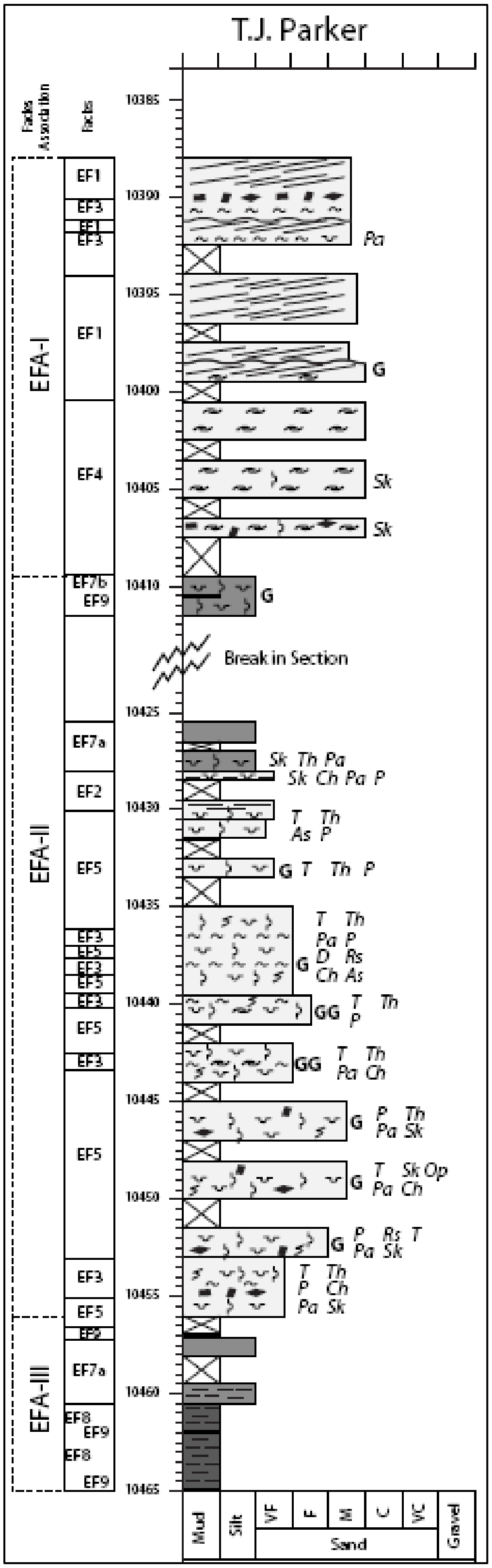

Figure 3.6. Core description of the T.J. Parker well in southwest Mississippi (see Figure 3.1 for location). Figure contains interpreted facies and facies associations. Detailed descriptions of facies (EF) included in the Estuarine Facies Association (EFA) are listed in Tables 3.4 and 3.5. Key for core description is found in Figure 3.4.

Table 3.4. Description and interpretation of the nine estuarine facies identified in three estuarine core of the Lower Tuscaloosa Formation.

Facies		**Physical Sedimentary Structures**	**Flora / Fauna**	**Ichnology**	**Depositional Environment**
EF-1 Very fine to medium-grained low-angle planar cross stratified sandstone		Common Low-angle planar lamination	Rare Isolated layers of abundant mollusk shells	None	Barrier bar, Hummocky shoreface (?)
EF-2 Very fine- to medium-grained planar laminated sandstone		Common Planar lamination	Rare Bioturbation	*P*, *Pa*, *Sk*, *Ch*	Barrier bar, Flood tidal delta, washover fans
EF-3 Very-fine to medium-grained ripple laminated sandstone		Common Ripple lamination Rare Climbing ripples Mud rip-up clasts	Rare Mollusk shells Bioturbation	*Pa*	Barrier bar, Flood tidal delta, washover fans

Table 3.4 continued.

EF-4 Very fine to medium-grained massive sandstone		Common Massively bedded Mud rip-up clasts Rare Red/green mottling	Common Mollusk shells Rare Bioturbation	*Sk*	Barrier bar, Flood tidal delta, washover fans
EF-5 Very-fine heavily bioturbated sandstone		Common Massive or disrupted bedding Rare Ripple lamination Mud rip-up clasts Red/green mottling	Common Heavy bioturbation	*Pa, T, P, Sk, Ro, Th, As, Ch, D, Op*	Central estuary, Bayhead delta plain with paleosol development
EF-6 Very fine to medium-grained convolute-bedded sandstone		Common Convoluted bedding	Rare Bioturbation	Unidentified	Flood tidal delta, washover fans

Table 3.4 continued.

<u>EF-7</u> Siltstone	**<u>EF-7a</u>** Laminated sparsely bioturbated Siltstone	<u>Common</u> Thin laminations <u>Rare</u> Soft sediment deformation	<u>Rare</u> Bioturbation	*P*, *Sk*, *Pa*, *Th*	Central estuary, Bayhead delta plain
	<u>EF-7b</u> Rippled Siltstone (more heavily bioturbated)	<u>Common</u> Lensoidal sandy silts Ripple laminations Red/green mottling Mud rip-up clasts	<u>Common</u> Heavy bioturbation	*Pa*, *P*, *Sk*, *T*, *Rh*, *Th*, *Ch*, *As*, *Lo*	Central estuary, Bayhead delta plain with paleosol development
<u>EF-8</u> Shale		<u>Common</u> Thin Laminations	None	None	Central estuary, Bayhead delta plain
<u>EF-9</u> Coal		<u>Common</u> Mud interbeds	<u>Common</u> Carbonized wood and plant fragments	None	Backswamps in bayhead delta plain

assemblage of EF-5 includes *Asterosoma*, *Chondrites*, *Diplocraterion*, *Ophiomorpha*, *Palaeophycus*, *Planolites*, *Rosselia*, *Skolithos*, *Teichichnus*, and *Thalassinoides* (Figure 3.7E). EF-5 is interpreted as sand episodically deposited in the central basin of the estuary by storm activity. This interpretation is based on more common mud rip-up clasts (Nichol et al., 1997) and the increase in bioturbation in EF-5 compared to the sandy barrier bar facies EF-1-EF-4. This bioturbation likely occurred as organisms recolonized the sands after emplacement by storm activity (Boyd et al., 2006). EF-6 is rare, very fine- to medium-grained sandstone dominated by convolute-bedding. Both the basal and upper contacts of EF-6 are gradational. Bioturbation is rare. Convolute bedding suggests that EF-6 was also rapidly deposited during episodic storm events; however EF-6 was not colonized by burrowing organisms to the same degree as EF-5.

Finer-grained facies include EF-7a, EF-7b, EF-8, and EF-9. EF-7 is divided into two subfacies including a laminated, sparsely-bioturbated siltstone (EF-7a) and a ripple cross laminated, heavily-bioturbated siltstone (EF-7b). Both the basal and upper contacts of EF-7a and EF-7b are sharp to gradational. EF-7a contains thin laminations, rare soft sediment deformation, and a rare and low diversity trace fossil assemblage including *Planolites*, *Skolithos*, *Palaeophycus and Thalassinoides* (Figure 3.7F). EF 7a is interpreted as either a central estuarine mudstone or a mudstone deposited in interdistributary bays of the bayhead delta in the proximal portion of the estuary (Allen and Posamentier, 1993; Nichol et al., 1997). EF-7b is commonly ripple cross laminated, contains interbedded lensoidal sands, and typically commonly contains a low diversity trace fossil assemblage including *Asterosoma*, *Chondrites*, *Lockeia*, *Palaeophycus*, *Planolites*, *Rhizocorallium*, *Skolithos*, *Teichichnus*, and *Thalassinoides* (Figure 3.7G). In the Centerville and Roberts cores (Appendix IIc and IId respectively), EF-7b has a red to

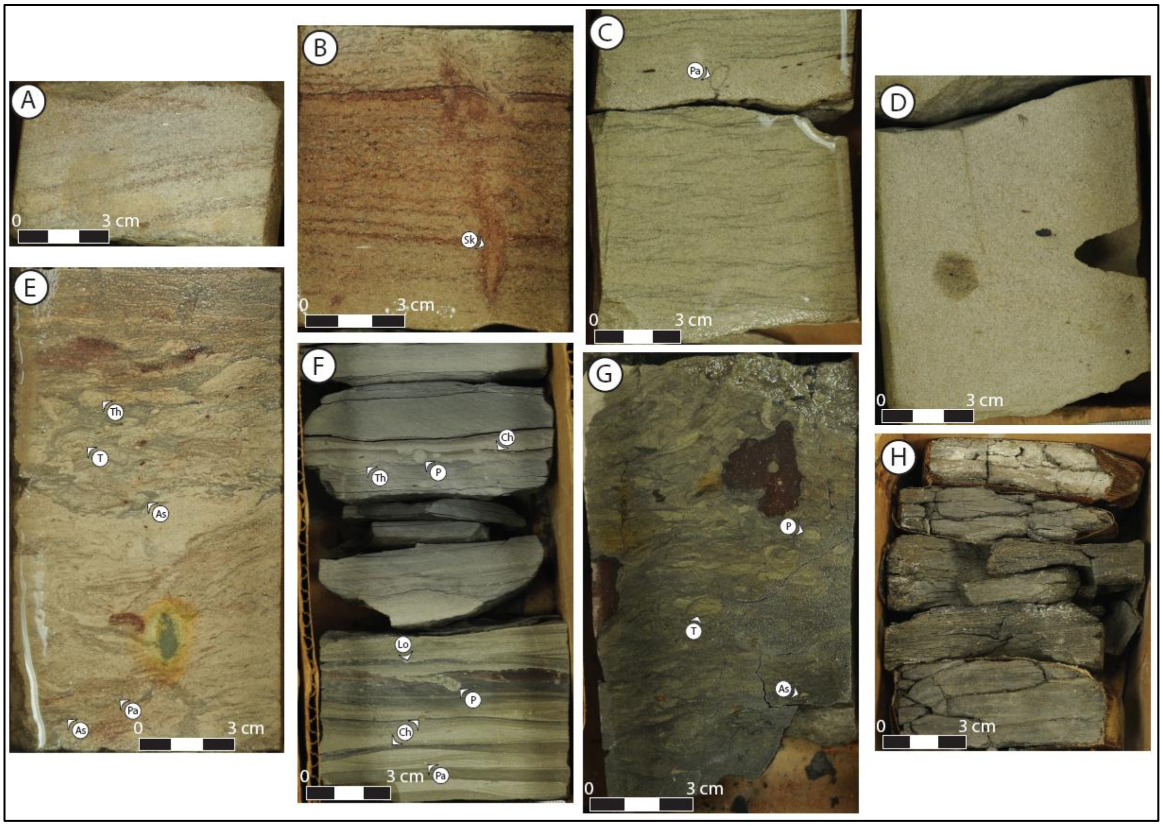

Figure 3.7. Core photographs of selected estuarine facies including identified trace fossils. (A) EF-1: planar cross stratified sandstone; (B) EF-2: planar-laminated sandstone; (C) EF-3: ripple-laminated sandstone; (D) EF-4: massive-bedded sandstone; (E) EF-5: heavily bioturbated very-fine sandstone; (F) EF-7a: laminated sparsely bioturbated siltstone interbedded with flaser-bedded sands; (G) EF-7b: ripple-laminated heavily bioturbated siltstone; and (H) EF-8: thinly-laminated shale interbedded with EF-9: coal. Key for trace fossil abbreviations is found in Figure 3.4.

green mottled fabric. A mottled fabric in estuarine facies has been noted in other studies (Boyd and Honig, 1992; Bryant et al., 1992). EF-8 is rare, thinly-laminated shale with a sharp basal contact and upper sharp to gradational contact that is commonly interbedded with coal (EF-9, Figure 3.7H). EF-8 is interpreted as a bayhead delta-plain mud and EF-9 is interpreted as a bayhead delta backswamp deposit.

Facies Associations

EF-1 through EF-9 are grouped into three facies associations (Table 3.5), a Shoreface-Barrier Association (EFA-I), a Central Basin Association (EFA-II), and a Bayhead Delta Association (EFA-III)

The Shoreface-Barrier Association (EFA-I) includes EF-1, EF-2, EF-3, EF-4, and EF-8. Characteristic features of EFA-I include medium- to coarse-grained sand rich facies accompanied by abundant high- to low-angled planar lamination, both indicative of the higher energies expected in the distal portions of a wave dominated estuary (Dalrymple et al., 1992; Allen and Posamentier, 1993; Boyd et al., 2006; MacEachern and Bann, 2008). Additionally, sandy EFA-1 facies overlie the central basin estuarine deposits (EFA-II) confirming the transgressive nature of the estuary (Allen and Posamentier, 1993; Boyd et al., 2006). In the T.J. Parker core the base of EFA-I is characterized by an approximately 10 foot interval of reworked mollusk shells interpreted to record the transgressive surface of the barrier bar (Allen and Posamentier, 1993). EFA-I is interpreted as deposits from the most distal parts of the estuary (Dalrymple et al., 1992; Allen and Posamentier, 1993; Boyd et al., 2006; MacEachern and Bann, 2008) in the wave-dominated zone characterized by the barrier bar, upper shore face, flood tidal deltas, and washover fans, similar to deposits described by Nichol et al. (1997) in the Holocene upper Hawkesbury River in New South Wales, Australia and by Allen and

Table 3.5. Description and interpretation of the three estuarine facies associations identified in the three estuarine core of the Lower Tuscaloosa Formation.

Facies Association	Paleo-environment	Inclusive Facies	Diagnostic Features	Occurrence
EFA-I	Outer Zone (Shoreface/Barrier) Wave-Dominated Estuary Association	EF-1, EF-2, EF-3, EF-4, EF-8	Very sandy wave dominated portion of the estuary. Planar cross bedded to massive sands. Transgressive lag full of mollusk shells at the base in some cases.	Found above EFA-II as the top of the transgressive estuary fill sequence. Amite County, Mississippi Stringer Sand
EFA-II	Central Zone (Lagoon) Wave-Dominated Estuary Association	EF-2, EF-3, EF-4, EF-5, EF-6, EF-7a, Ef-7b, EF-8	Interbedded sands silts and muds. Dominantly muddy but can be sandy at times. Interbeds often appear rhythmic suggesting tidal influence. Very heavy bioturbation and ripples are abundant.	Found above EFA-I and below EFA-III as the middle of the transgressive estuary fill sequence. Amite County, Mississippi Stringer Sand
EFA-III	Inner Zone (Bayhead Delta) Wave-Dominated Estuary Association	EF-4, EF-5, EF-7a, EF-7b, EF-8, EF-9	Muds, silts, and coals associated with the bayhead delta plain and marshes on the flanks of the estuary. Silts and very fine sands are heavily bioturbated and show moderate to heavy red and green mottling indicative of paleosols.	Found below EFA-II as the base of the transgressive estuary fill sequence. Amite County, Mississippi Stringer Sand

Posamentier (1993) found in the Holocene Gironde Estuary, France. The diversity and abundance of trace fossils in EFA-I are low, likely due to the combination of higher wave energies, sandier substrates, and salinity variations which inhibit many of the trace-making organisms found in both fully-marine environments and in the central basin of the estuary from occupying the substrate (Nichol et al., 1997; Boyd et al., 2006; MacEachern and Bann, 2008). MacEachern and Bann (2008) consider highly diverse trace fossil suites, such as those found in fully marine settings, to contain 15 – 24 ichnogenera. In contrast, impoverished trace fossil suites commonly found in more brackish–water environments contain 6 -10 ichnogenera with 3 – 7 recurring ichnogenera.

The Central Basin Association (EFA-II) includes EF-2, EF-3, EF-4, EF-5, EF-6, and EF-7a and b. EFA-II is interpreted as deposits from the mixed energy environment characteristic of the central estuary basin. Characteristic features of EFA-II include increased bioturbation, mud-draped ripple cross laminations in sands, and interbedded lensoidal sands and silts. These features are expected in the central basin of the estuary where there is an overall reduction in total energy and a mix of energy from tides, waves, and river currents (Dalrymple et al., 1992; Allen and Posamentier, 1993; Nichol et al., 1997; Boyd et al., 2006; MacEachern and Bann, 2008). Although bioturbation in EFA-II increases compared to EFA-I and EFA-III, the overall low abundance and low diversity of trace fossils indicate a stressed environment typical of brackish water conditions expected in the central estuary (Nichol et al., 1997; Boyd et al., 2006; MacEachern and Bann, 2008). Mud-draped ripples are common, suggesting strong tidal energies Allen and Posamentier, 1993; Boyd et al., 2006). Episodic storm and flood processes are evidenced by medium-grained sand beds containing mud rip-up clasts and soft-sediment deformation structures (Nichol et al., 1997; Boyd et al., 2006). Some of these beds also contain mollusk shells and all show bioturbated tops. These sandy beds likely record either storm-related, wash-over fans, (beds with mollusk shells) or periodic river-flood deposits (Nichol et al., 1997). The recolonization of episodic sand beds by burrowing organisms indicates hiatuses between storm activity (Nichol et al., 1997).

The Bayhead Delta Association (EFA-III) includes EF-4, EF-5, EF-7a, b, EF-8, and EF-9. EFA-III is interpreted as finer-grained bayhead delta-plain deposits from the most proximal portion of the estuary. Diagnostic features of the bayhead delta deposits include their fine-grained and organic rich nature and their position in the overall transgressive stacking pattern (Dalrymple et al., 1992; Allen and Posamentier, 1993; Nichol et al., 1997; Boyd et al., 2006). Lensoidal sands interbedded with EF-7b strongly suggest tidal influence in the estuary (Allen and Posamentier, 1993; Boyd et al., 2006). Red to green mottles in EF-5 suggests increased levels of bioturbation in the lower energy central basin, or intermittent subaerial exposure (Boyd and Honig, 1992; Bryant et al., 1992). Coals unique to EFA-III would have formed in backswamps of the floodplain (Allen and Posamentier, 1993; Nichol et al., 1997). Overall, core logs reflect the transgressive nature of this estuarine system with EFA-III (bayhead deltas) at the base of the succession overlain by EFA-II (central basin deposits) and capped by EFA-I (shoreface-barrier sands) (Figure 3.6, Appendix IIc, IId).

Deltas

Six of the thirteen cored intervals were identified as probable deltaic environments located in unconfined reaches distal of incised valleys (Table 3.1). Figures 3.8 and 3.9 detailing the W.A. Lorio and SL 10860 cores are included for reference. Descriptions of the four other cores exhibiting deltaic facies (Thom, I.J. Major, Biloxi

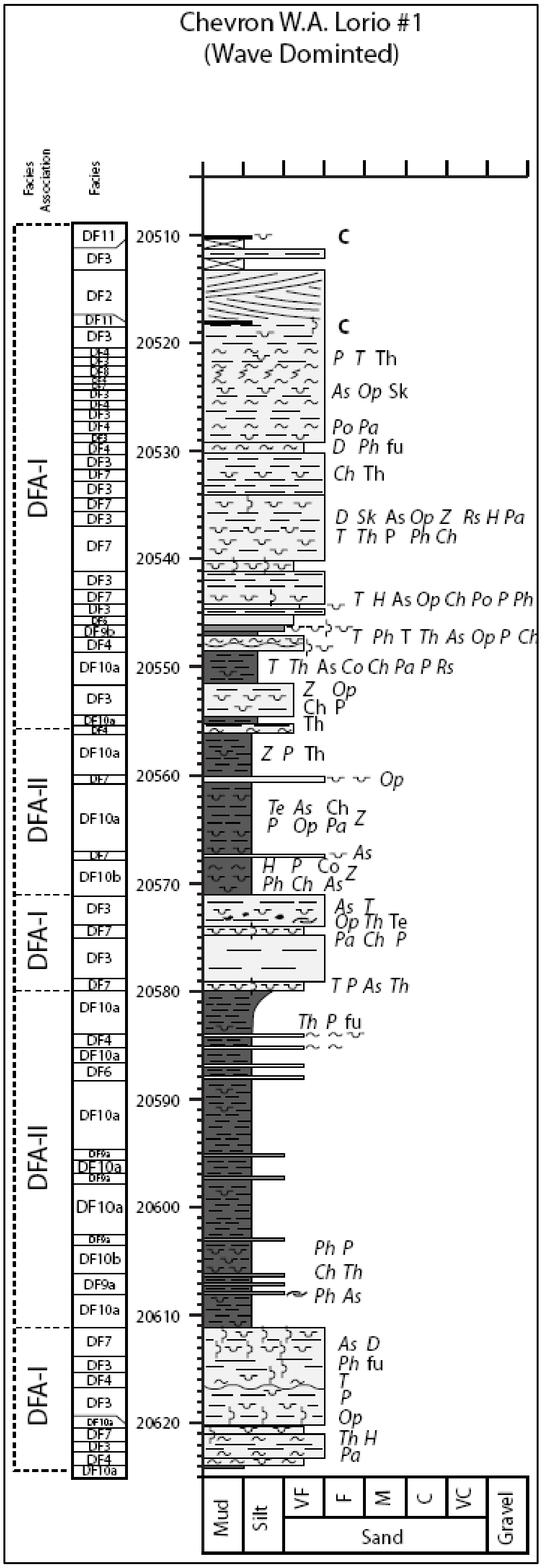

Figure 3.8. Core description of W.A. Lorio well from Delta-1 in central Louisiana (see Figure 3.1 for location). Figure contains interpreted facies and facies associations. Detailed descriptions of facies included in the Deltas Facies Association are listed in Tables 3.6 and 3.7. Key for core description is found in Figure 3.4.

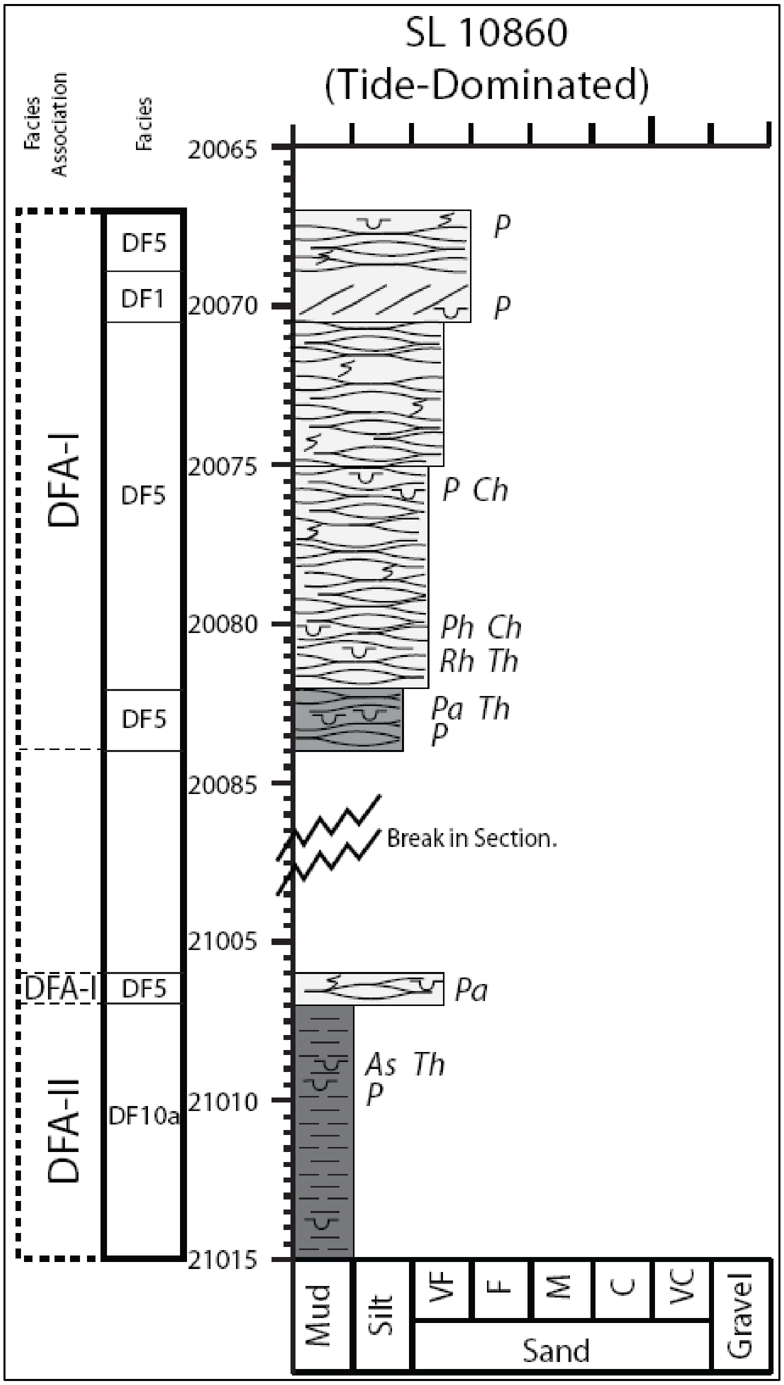

Figure 3.9. Core description of SL 10860 well from Delta-2 in eastern Louisiana Panhandle (see Figure 3.1 for location). Figure contains interpreted facies and facies associations. Detailed descriptions of facies included in the Deltas Facies Association are listed in Tables 3.6 and 3.7. Key for core description is found in Figure 3.4.

Marshlands #P2, and Crochett) can be found in Appendix IIe, IIf, IIg and IIh. Eleven facies were identified and grouped into three facies associations (Figures 3.8, 3.9; Tables 3.6, 3.7; Appendix IIe, IIf, IIg, IIh).

Facies

Eight of the facies associated with the deltaic systems of the Lower Tuscaloosa Formation are sandstones (DF-1, DF-2, DF-3, DF-4, DF-5, DF-6, DF-7, and DF-8). Each of these facies displays a different grain size, dominant sedimentary structure, and low diversity trace fossil assemblage that is likely related to its spatial and temporal location and the energy of the system. DF-1 (Figure 3.10A) is a high-angle, planar cross stratified very-fine to fine-grained sandstone with sharp-to-gradational basal contact and a sharp upper contact. DF-1 rarely contains the trace fossils *Palaeophycus, Planolites*, and *Skolithos*. DF-2 (Figure 3.10B) is a low-angle planar cross stratified very-fine to medium-grained sandstone with sharp-to-gradational basal and upper contacts. DF-2 contains rare stylolites, mollusk shells, and trace fossils including *Palaeophycus*, *Phycosiphon*, *Planolites*, *Subphylochorda*, *Teichichnus*, and *Thalassinoides*. DF-3 (Figure 3.10C) is a horizontal-bedded, planar-laminated very-fine to medium-grained sandstone with an erosional to gradational basal contact and a sharp-to-gradational upper contact containing common soft sediment deformation, rare floating chert pebbles, and mollusk shells. DF-3 trace fossils include *Asterosoma*, *Chondrites*, *Diplocraterion*, *Helminthopsis*, *Ophiomorpha*, *Palaeophycus*, *Phoebichnus*, *Phycosyphon*, *Planolites*, *Schaubcylindrichnus*, *Skolithos*, *Teichichnus*, Terebellina, *Thalassinoides*, and *Zoophycos*. DF-4 (Figure 10D) is a very-fine to fine-grained ripple cross laminated sandstone with an erosional to gradational basal contact and a sharp-to-gradational upper contact containing rare climbing ripples. DF-4 trace fossils include *Asterosoma*, *Chondrites*, *Cylindrichnus*, *Diplocraterion*, *Helminthopsis*, *Ophiomorpha*, *Palaeophycus*, *Phycosyphon*, *Planolites*, *Rosselia*, *Skolithos*, *Teichichnus*, *Thalassinoides*, and *Zoophycos* (Figure 10D). DF-5 (Figure 10E) is a very-fine to fine-grained, silt-rich, rhythmically-bedded sandstone containing common lenticular, flaser, and convolute bedding. Both the basal and upper contacts of DF-5 can be sharp to gradational. DF-5 trace fossils include *Chondrites*, *Helminthopsis*, *Palaeophycus*, *Phycosiphon*, *Planolites*, *Rhizocorallium*, *Teichichnus*, and *Thalassinoides* (Figure 10E). DF-6 (Figure 10F) is a very-fine to medium-grained massive sandstone with sharp-to-gradational basal and upper contacts. DF-6 contains the trace fossils *Asterosoma*, *Ophiomorpha*, *Palaeophycus*, *Planolites*, *Rosselia*, and *Thalassinoides* (Figure 10F). DF-7 is very-fine-grained, heavily-bioturbated sandstone with sharp-to-gradational basal and upper contacts that also contains rare, ripple cross laminations. DF-7 trace fossils include *Asterosoma*, *Chondrites*, *Cylindrichnus*, *Diplocraterion*, fugichnia, *Helminthopsis*, *Ophiomorpha*, *Palaeophycus*, *Phoebichnus*, *Phycosiphon*, *Planolites*, *Rhizocorallium*, *Rosselia*, *Skolithos*, *Teichichnus*, Terebellina, *Thalassinoides*, and *Zoophycos*, (Figure 10G). DF-8 (Figure 10H) is a very-fine to medium-grained sandstone with sharp-to-gradational basal and upper contacts containing abundant soft-sediment deformation. DF-8 contains the trace fossils *Asterosoma*, *Chondrites*, *Ophiomorpha*, *Palaeophycus*, *Planolites*, *Teichichnus*, and *Thalassinoides*. All of these sand-dominated facies are interpreted to record deposition in distributary mouth-bar and delta-front environments (Table 6).

Finer-grained facies include DF-9, DF-10, and DF-11. DF-9 is divided into two subfacies. DF-9a is a planar-laminated, sparsely-bioturbated siltstone with gradational to sharp basal and upper contacts that may contain rare mollusk shells. DF-9a contains the trace fossils *Asterosoma*, *Chondrites*, *Cylindrichnus*, *Diplocraterion*, *Helminthopsis*, *Palaeophycus*, *Phycosiphon*, *Planolites*, *Rosselia*, *Teichichnus*, *Thalassinoides*, and

Zoophycos. DF-9b (Figure 10I) is a ripple cross laminated, heavily-bioturbated siltstone with a gradational to sharp basal and upper contact that may also contain lensoidal sandy silts. DF-9b trace fossils include *Asterosoma*, *Chondrites*, *Diplocraterion*, *Helminthopsis*, *Palaeophycus*, *Phycosiphon*, *Planolites*, *Rhizocorallium*, *Rosselia*, *Schaubcylindrichnus*, *Skolithos*, *Teichichnus*, *Thalassinoides*, and *Zoophycos* (Figure 10I). DF-10 is also divided into two subfacies. DF-10a is thinly laminated shale with a sharp basal contacts and a sharp to gradational upper contact containing rare soft-sediment deformation, ripple cross laminations and mollusk shells. DF-10a contains the trace fossils *Asterosoma*, *Chondrites*, *Cosmorhaphe*, *Helminthopsis*, *Palaeophycus*, *Phycosiphon*, *Planolites*, *Rosselia*, *Teichichnus*, Terebellina, *Thalassinoides*, and *Zoophycos*. DF-10b is a massive, heavily-bioturbated mudstone with a sharp basal contact and a sharp to gradational upper contact. DF-10b contains the trace fossils *Arencolites*, *Asterosoma*, *Chondrites*, *Cosmorhaphe*, *Cylindrichnus*, *Helminthopsis*, *Palaeophycus*, *Planolites*, *Phycosiphon*, *Teichichnus*, *Thalassinoides*, and *Zoophycos* (Figure 10J). D-11 is a coal with sharp upper and lower contacts containing abundant plant and wood fragments and mudstone interbeds. The fine-grained nature of facies D9a, b and D10a, b combined with predominant sedimentary structures and trace fossils indicate they were likely deposited in the prodelta environment. Facies D-11 (coal) was likely deposited in backswamps of the delta plain.

Facies Associations

DF-1 through DF-11 are grouped into two facies associations (Table 3.7), a Delta Front Association (DFA-I) and a Prodelta Association (DFA-II).

The Delta Front Association (DFA-I) includes DF-1, DF-2, DF-3, DF-4, DF-5, DF-6, DF-7, DF-8, and rarely DF-10a and b, and DF-11. DFA-I is interpreted as deposits

Table 3.6. Description and interpretation of the eleven deltaic facies identified in the six Deltaic cores of the Lower Tuscaloosa Formation.

Facies		**Physical Sedimentary Structures**	**Flora / Fauna**	**Ichn-ology**	**Depositional Environment**
DF-1 Very fine to fine-grained high-angle planar cross stratified sandstone		Common High-angle planar lamination	Rare Bioturbation	*Pa, P, Sk*	Distributary mouth bar
DF-2 Very fine to medium-grained low-angle planar cross stratified sandstone		Common Low-angle planar lamination Rare Stylolites	Rare Bioturbation Mollusk shells	*Pa, Ph, P, Su, T, Th*	Distributary mouth bar
DF-3 Very fine- to medium-grained planar laminated sandstone		Common Soft sediment deformation Rare Floating chert pebbles Mud rip-up clasts	Rare Bioturbation Mollusk shells	*As, Ch, D, H, Op, Pa, Ph, P, Sch, Sk, T,* Te, *Th, Z*	Distributary mouth bar
DF-4 Very-fine to medium-grained ripple laminated sandstone		Common Ripple lamination Rare Climbing ripples	Rare Bioturbation	*As, Ch, Cy, D, H, Op, Pa, Ph, P, Ro, Sk, T, Th, Z*	Distributary mouth bar

Table 3.6 continued.

<u>DF-5</u> Very fine to fine-grained silty rhythmically bedded sandstone		<u>Common</u> Lenticular sands Convolute bedding	<u>Rare</u> Bioturbation	*Ch*, *H*, *Pa*, *Ph*, *P*, *Rh*, *T*, *Th*	Tidally influenced distributary mouth bar
<u>DF-6</u> Very fine to medium-grained massive sandstone		<u>Common</u> Massively bedded	<u>Rare</u> Bioturbation	*As*, *Op*, *Pa*, *P*, *Ra*, *Th*	Distributary mouth bar
<u>DF-7</u> Very-fine heavily bioturbated sandstone		<u>Rare</u> Ripple lamination	<u>Common</u> Heavy bioturbation	*As*, *Ch*, *Cy*, *D*, fu, *H*, *Op*, *Pa*, *Po*, *Ph*, *P*, *Rh*, *Ro*, *Sk*, *T*, Te, *Th*, *Z*	Distributary mouth bar
<u>DF-8</u> Very fine to medium-grained convolute-bedded sandstone		<u>Common</u> Convoluted bedding	<u>Rare</u> Bioturbation	*As*, *Ch*, *Op*, *Pa*, *P*, *T*, *Th*	Distributary mouth bar

Table 3.6 continued.

DF-9 Siltstone	**DF-9a** Laminated sparsely bioturbated Siltstone	Common Thin laminations	Rare Bioturbation Mollusk shells	*As, Ch, Cy, D, H, Pa, Ph, P, Ro, T, Th, Z*	Distal mouth bar, proximal prodelta
	DF-9b Rippled Siltstone (more heavily bioturbated)	Common Ripple laminations Rare Soft sediment deformation	Common Heavy bioturbation	*As, Ch, D, H, Pa, Ph, P, Rh, Ro, Sch, Sk, T, Th, Z*	Distal mouth bar, proximal prodelta
DF-10 Mudstone	**DF-10a** Shale	Common Thin Laminations Rare Soft sediment deformation	Rare Bioturbation Mollusk shells	*As, Ch, Co, H, Pa, Ph, P, Ro, T,* Te, *Th, Z*	Distal prodelta, interdistribut-ary bays, offshore
	DF-10b Bioturbated Mudstone	Massively bedded	Common Heavy Bioturbation	*As, Ch, Co, Cy, H, Pa, Ph, P, T, Th, Z*	Distal prodelta, interdistribut-ary bays, offshore
DF-11 Coal		Common Mud interbeds	Common Carbonized wood and plant fragments	None	Backswamps

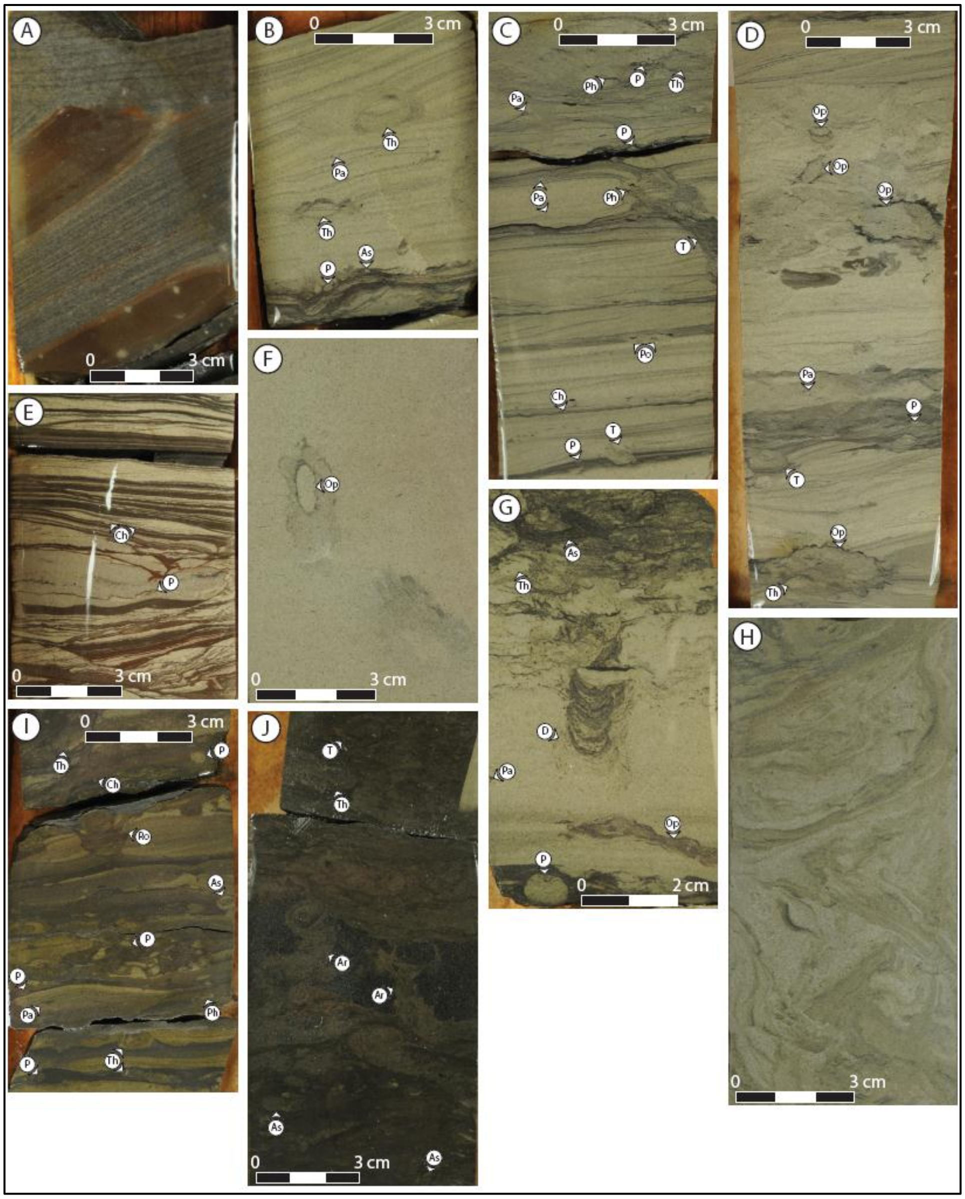

Figure 3.10. Core photographs of selected deltaic facies including identified trace fossils. (A) DF-1: high-angle planar cross stratified sandstone; (B) DF-2: low-angle planar cross stratified sandstone; (C) DF-3: planar-laminated sandstone; (D) DF-4: ripple-laminated sandstone; (E) DF-5: lenticular- to flaser- bedded sandstone; (F) DF-6: massive-bedded sandstone; (G) DF-7: very fine-grained bioturbated sandstone; (H) DF-8: very fine- to medium-grained sandstone containing soft-sediment deformation; (I) DF-9b: ripple-laminated heavily-bioturbated siltstone; and (J) DF-10b: massive heavily-bioturbated mudstone. Key for trace fossil abbreviations is found in Figure 4.

typically found in a delta-front environment (Mayall et al., 1992; Porebski and Steel, 2003; McIlroy, 2004; Bhattacharya, 2006; Carmona et al., 2009). Characteristic features of DFA-I include sand-rich facies, sedimentary structures indicative of high sedimentation rates (soft sediment deformation, slumping, and climbing ripples), and a low diversity trace fossil assemblage (Porebski and Steel, 2003; McIlroy, 2004; Bhattacharya, 2006; MacEachern and Bann, 2008; Carmona et al., 2009). Additionally, these deltaic facies are located proximal to the well documented Lower Tuscaloosa Formation paleo-shelf edge and exhibit repeated coarsening upward cycles, with the coarser-grained DFA-I located on top of the finer-grained prodelta facies of DFA-II. DFA-I exhibits varying degrees of wave and tidal influence. In cores from Delta-2 (SL 10860 and Biloxi Marshlands P-2) (Figure 3.9, Appendix IIg), DFA-I is dominated by lenticular- to flaser-bedded-, silty-sandstone that is typical of tidally-dominated deltas. Trace-fossil suites of DFA-I are low to very-low abundance, low diversity, and suggest a regionally restricted distribution of the *Cruziana* Ichnofacies. Highly impoverished suites of the *Cruziana* ichnofacies are typical of delta-front to prodelta environments (McIlroy, 2004; Bhattacharya, 2006; MacEachern and Bann, 2008; Carmona et al., 2009). Despite the similar impoverishment of traces in both tide and wave-influenced deltaic settings, bioturbation in the tide influenced environments is more highly impoverished when compared to the wave influenced environments (McIlroy, 2004; MacEachern and Bann, 2008; Carmona et al., 2009). In the Lower Tuscaloosa Formation cores from Delta-2 contain abundant lenticular and flaser-bedding and have a much more highly impoverished trace fossil assemblage compared to cores from Delta-1

The Prodelta Association (DFA-II) includes DF-9a and b, DF-10a and b, and rarely thin beds of DF-1, DF-2, DF-3, DF-4, DF-5, DF-6, DF-7, DF-8 (Table 3.7). DFA-II contains predominantly finer-grained facies including siltstones and mudstones that

Table 3.7. Description and interpretation of the two deltaic facies associations identified in six deltaic core of the Lower Tuscaloosa Formation. S0, S1, S2, S3 refer to the individual sequences of the Massive sand as determined in Chapter 2.

Facies Association	**Paleo-environment**	**Inclusive Facies**	**Diagnostic Features**	**Occurrence**
DFA-I	Delta Front Association	Common DF-1, DF-2, DF-3, DF-4, DF-5, DF- 6, DF-7, DF-8 Rare DF-10a, DF-10b, DF-11	Very fine to medium grained sands capping coarsening upwards cycles. Found on top of FA-II. Contain various sed. structures including; planar cross beds, planar lamination, ripples, and massive bedding. Sparse to heavily bioturbated.	In the proximity of the Lower Cretaceous shelf edge in Pointe Coupee, Livingston, and St. Bernard Parishes, Louisiana. False River, Judge Digby, Lockhart Crossing, and Biloxi Marshlands fields S0, S1, S2, S3, Stringer, Middle Tusc.
DFA-II	Prodelta Association	Common DF-9a, DF-9b, DF-10a, DF-10b Rare DF-1, DF-2, DF-3, DF-4, DF-5, DF- 6, DF-7, DF-8	Shales and siltstones found in between FA-I. Shales are laminated and sparsely to heavily bioturbated. Silts are rippled, contain lensoidal sands and are heavily bioturbated.	In the proximity of the Lower Cretaceous shelf edge in Pointe Coupee, Livingston, and St. Bernard Parishes, Louisiana. False River, Judge Digby, Lockhart Crossing, and Biloxi Marshlands fields S0, S1, S2, S3, Stringer, Middle Tusc.

likely settled out of suspension in a prodelta environment (Bhattacharya, 2006; Carmona et al., 2009). These fine-grained facies grade up into the coarser-grained facies of DFA-I preserving coarsening upward deltaic successions. Thin layers of low-angle planar-laminated sands (DF-1), ripple cross laminated sands (DF-4), and convolute-bedded sands (DF-8) interfinger with finer grained prodelta sediments. Soft-sediment deformation in the sand slump inclusions and finer grained facies is characteristic of the prodelta environment (Porebski and Steel, 2003; Bhattacharya, 2006). Low diversity and restricted distribution of ichnogenera of the *Cruziana* Ichnofacies is also typical of the prodelta environment (MacEachern and Bann, 2008; Carmona et al., 2009). The impoverished trace fossil assemblage reflects the mixing of fresh water from the fluvial system with marine waters and higher amounts of suspended sediment compared to fully marine environments (McIlroy, 2004; Bhattacharya, 2006; MacEachern and Bann, 2008). Similar to facies of the Delta Front Association (DFA-I), DFA-II facies from cores located within Delta-1 are more heavily bioturbated when compared to Delta-2 suggesting that Delta-1 may be wave-influenced while Delta-2 is tidally influenced (McIlroy, 2004; Bhattacharya, 2006; MacEachern and Bann, 2008; Carmona et al., 2009).

Gravity-Driven Deepwater Deposits

One of the thirteen cored intervals studied was identified as deposited in a probable deepwater environment. Physical core was not available; therefore interpretations for the Tiber well in the deepwater Gulf of Mexico (Table 3.1, Figure 3.1) are made from 72 m of continuous core photographs. Figure 3.11 details the Tiber well core photographs and is included for reference. Seven facies were identified and grouped into two facies associations (Tables 3.8, 3.9).

Facies

GF-1 through GF-5 are poorly sorted sandstones distinguished by differing sedimentary structures. GF-1 is a low-angle, planar cross stratified sandstone with sharp to gradational basal and upper contacts (Figure 3.12A). GF-2 is planar-laminated sandstone with an erosional, sharp, or gradational basal contact and a sharp or gradational upper contact (Figure 3.12B). GF-3 is ripple cross laminated sandstone with a sharp or gradational upper contact (Figure 3.12C). GF-4 is convolute- bedded sandstone with sharp basal and upper contacts. (Figures 3.12D-F). GF-5 is massive sandstone with an erosional, sharp, or gradational basal contact and a sharp or gradational upper contact (Figures 3.12G-H). Each of these sandstone facies displays varying amounts of floating pebbles and mud rip-up clasts as well as fluid escape features including dish structures and vertical pipes (Figures 3.12E,F). GF-1 through GF-5 are interpreted as lobate sheet sands associated with basin floor fans.

GF-6 contains interbedded silt and shale which also exhibits varying degrees of planar lamination, ripple cross lamination, soft sediment deformation, and bioturbation (Figure 3.12I) including *Planolites*. GF-7 is thinly bedded, thinly laminated shale that forms drapes between the thickly bedded GF-2, GF-3, and GF-5 sandstone facies (Figures 3.12G,H). The upper and lower contacts of GF-6 are sharp or rarely gradational. GF-6 and GF-7 are interpreted as deepwater muds of gravity-driven flows deposited during waning-flow stages on top of the sandier facies.

Facies Associations-

GF-1 through GF-7 are grouped into two facies associations (Table 3.9), a Sand-rich Association (GFA-I) and a Siltstone and Mudstone-rich Association (GFA-II).

The Sand-rich Association (GFA-I) includes facies GF-1, GF-2, GF-3, GF-4, and GF-5. Characteristic features of GFA-I are amalgamated, poorly sorted, sand-rich deposits with floating pebbles and shale clasts, few matrix muds, faint sedimentary structures such as planar and ripple lamination, and deformed and slumped beds. The combination of these features is characteristic of sandy debrites (Shanmugam, 1996; Amy et al., 2005; Shanmugam et al., 2009). The rare graded bedding, planar and ripple laminations, and massive bedding may indicate some deposition by turbidity currents (Amy et al., 2005; Shanmugam et al., 2009). There is an overall lack of trace fossils in the gravity-driven, deepwater facies likely due to frequent and high amounts of clastic input which does not allow time for organisms to burrow into the substrate.

The Siltstone and Mudstone-rich Association (GFA-II) includes facies GF-6 and GF-7. GFA-II is less abundant than GFA-I. Characteristic features of GFA-II include interbedded planar laminated or ripple cross laminated siltstones and shales that may contain soft sediment deformation which are in sharp contact with the sandier GFA-I facies. GFA-II is interpreted as being deposited during waning flow at the top of the sandier intervals of the gravity driven deposit (Talling et al., 2007; Haughton et al., 2009; Breien et al., 2010).

WIRELINE WELL LOGS

Fluvial Channels and Floodplains (FFA-I, FFA-II)

Wireline logs through FFA-I and FFA-II are found in the Massive sand of the Lower Tuscaloosa Formation from the outcrop belt to the shelf edge (Figure 3.13). Wireline patterns of the fluvial facies association display blocky (low gamma count) to inverted funnel (increasing gamma count) shapes indicating an upward increase in finer-grained facies (Miall, 1996; Bridge, 2003, 2006). FFA-I and FFA-II are dominated by thick, amalgamated sand packages correlated to low gamma ray readings, but contain intermittent high gamma ray readings interpreted as fine-grained abandoned channel-fills,

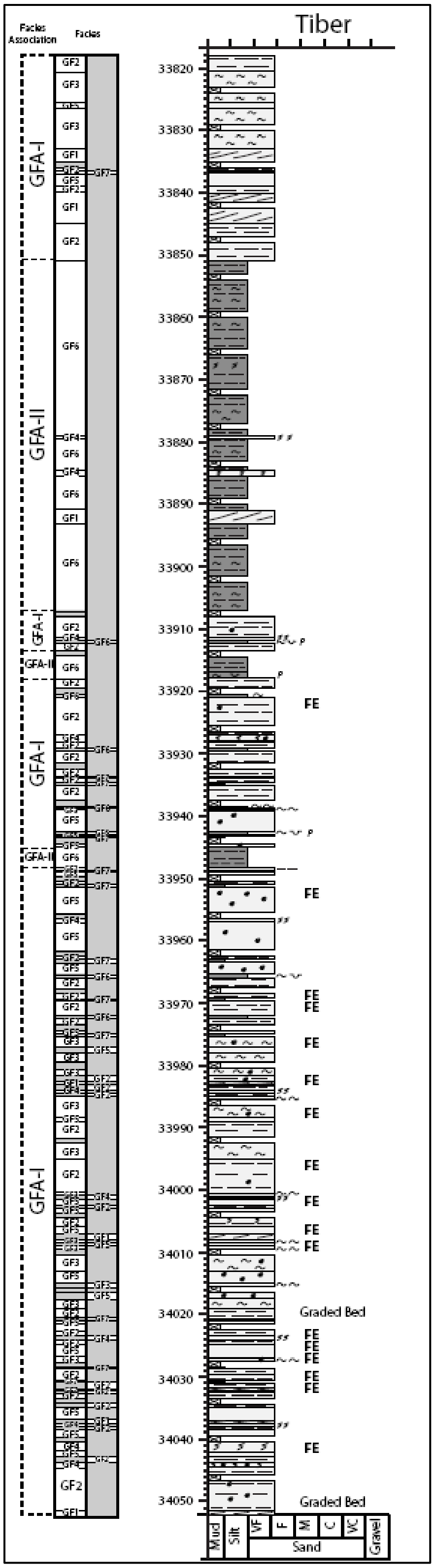

Figure 3.11. Core description (from core photographs) of Tiber well offshore Keathley Canyon area, Gulf of Mexico (see Figure 3.1 for location). Figure contains interpreted facies and facies associations. Detailed descriptions of facies included in the Deepwater Gravity Driven Facies Association are listed in Tables 3.8 and 3.9. Key for core description is found in Figure 3.4.

Table 3.8. Description and interpretation of the seven deepwater gravity driven facies identified in core photographs of the Lower Tuscaloosa Formation from the Tiber well (Figure. 3.1).

Facies	**Physical Sedimentary Structures**	**Depositional Environment**
GF-1 Low-angle planar cross stratified sandstone	Common Low-angle planar lamination Rare Fluid escape structures	Lobate sheet sands
GF-2 Planar laminated sandstone	Common Planar lamination Fluid escape structures Floating Pebbles Rare Graded bedding	Lobate sheet sands
GF-3 Ripple laminated sandstone	Common Ripple lamination Floating Pebbles Rare Fluid escape structures Thin mud drapes	Lobate sheet sands
GF-4 Convolute bedded sandstone	Common Soft sediment deformation Floating Pebbles Fluid escape structures	Lobate sheet sands
GF-5 Massive bedded sandstone	Common Fluid escape structures Floating Pebbles Rare Mud rip-up clasts	Lobate sheet sands
GF-6 Interbedded silt and shale	Common Interbedded ripple and planar lamination Bioturbation – *Planolites* burrows Soft sediment deformation	Waning gravity flow deposits
GF-7 Shale	Common Thin bedding Thin laminations	Waning gravity flow deposits

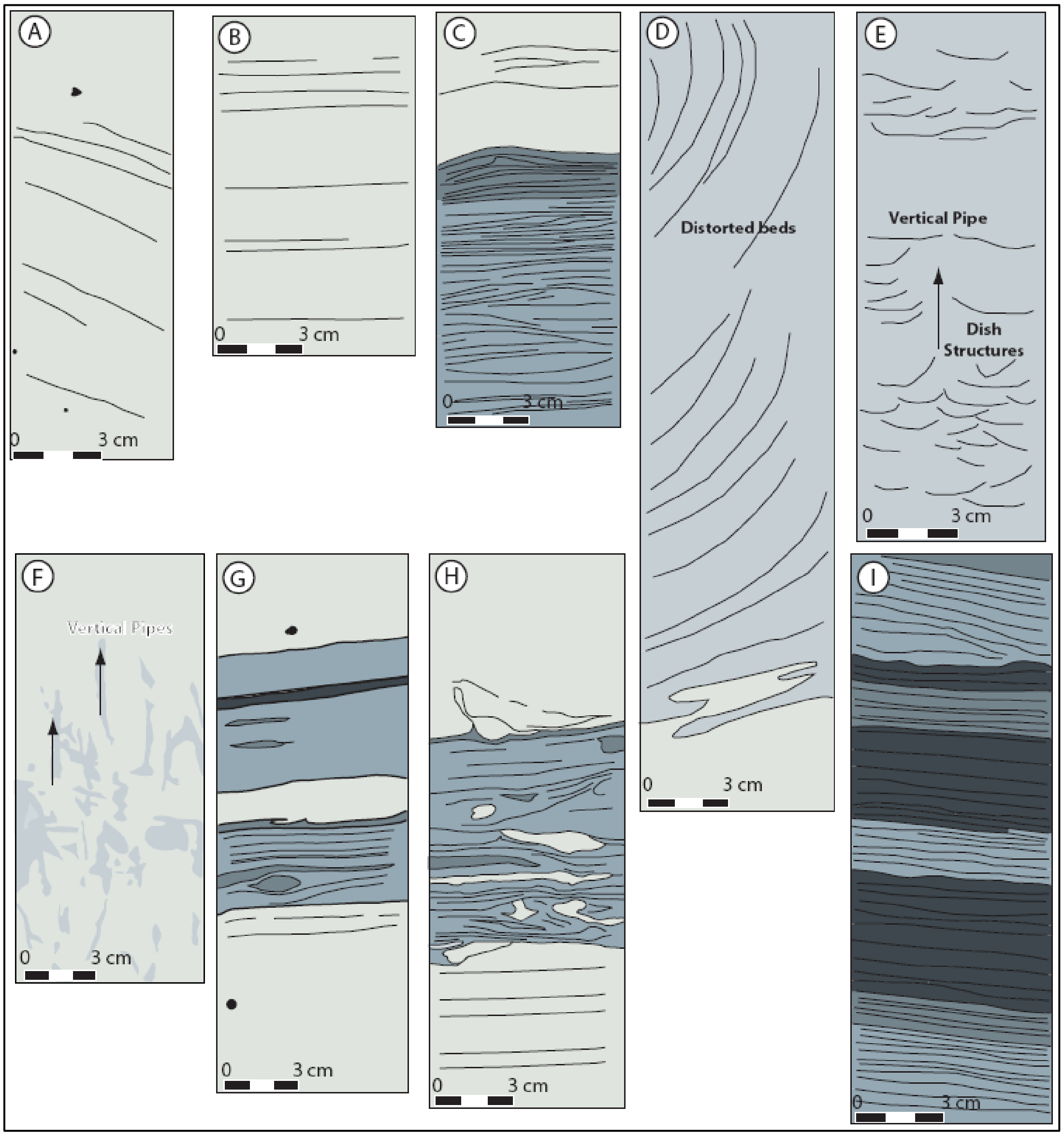

Figure 3.12. Interpreted line drawings from core photographs from Tiber well deep offshore Gulf of Mexico (see Figure 3.1 for location). Original photographs were redacted by BP. (A) GF-1: low-angle planar cross stratified sandstone; (B) GF-2: planar laminated sandstone; (C) GF-3: ripple cross laminated sandstone; (D-F) GF-4: convolute- bedded sandstones containing fluid escape features including dish structures and vertical pipes; (G and H) GF-5: massive sandstone sharply contacting finer grained facies GF-7; (I) GF-6: interbedded silt and shale exhibiting varying degrees of planar lamination, ripple cross lamination, and soft sediment deformation.

Table 3.9. Description and interpretation of the two deepwater gravity driven facies associations identified in core photographs of the Lower Tuscaloosa Formation in the Tiber well (Figure 3.1).

Facies Association	Paleo-environment	Inclusive Facies	Diagnostic Features	Occurrence
GFA-I	Sand-rich Association	GF-1, GF-2, GF-3, GF-4, GF-5	Thick bedded sandstones interbedded with thinner fine-grained siltstones and shales (GFA-II). Contain various sed. structures including; planar cross beds, planar lamination, ripples, convolute beds, and massive beds.	Deepwater Gulf of Mexico in the Keathley Canyon area
GFA-II	Siltstone and Mudstone-rich Association	GF-6 GF-7	Dominantly thin shales and siltstones found in between thick sandstone beds of GA-I. Shales are laminated. Silts are laminated and rippled, and sparsely bioturbated.	Deepwater Gulf of Mexico in the Keathley Canyon area

floodplain muds and silts, and backswamps. These amalgamated sand packages comprise repeated sharp-based fining upward successions.

Estuaries (EFA-I, EFA-II, EFA-III)

Wireline log patterns through EFA-I, EFA-II, and EFA-III are found in the Stringer sand interval capping the Massive sand of the Lower Tuscaloosa Formation (Figure 3.14). Wells containing an estuarine signature are predominantly located in southwest Mississippi and in the Louisiana panhandle north of the Lower Cretaceous shelf edge. Estuarine well log patterns display a tripartite signature that reflects the transgressive stacking pattern of the Bayhead Delta (EFA-III), Central Basin (EFA-II), and Shoreface-Barrier (EFA-I) facies associations. This stacking pattern is best represented in the T.J. Parker well (Figures 3.6, 3.14). This succession exhibits a fining upward log pattern from the coarser-grained bayhead delta, correlated to moderate gamma ray readings, transitioning to the central basin fine- grained facies, correlated to high gamma ray readings. This log motif is followed by a coarsening upward pattern from the central basin to the sand-rich shoreface/ barrier environment, correlated to very low gamma ray readings, capped sharply by very high gamma ray shales (Boyd et al., 2006). Although this succession is typified in the study area by the T.J. Parker well, estuarine wire line log patterns can display variability related to their range of facies (Boyd et al., 2006). Well log patterns commonly appear serrated because of the complex inter-fingering between the fine and coarse-grained facies (Figure 3.14).

Deltas (DFA-I, DFA-II)

Wireline-log patterns of DFA-I and DFA-II are found south of the well-defined Lower Cretaceous shelf edge but in close proximity to it (Figure 3.15). Well log patterns show repeated coarsening-upward successions capped sharply by shale, a common log

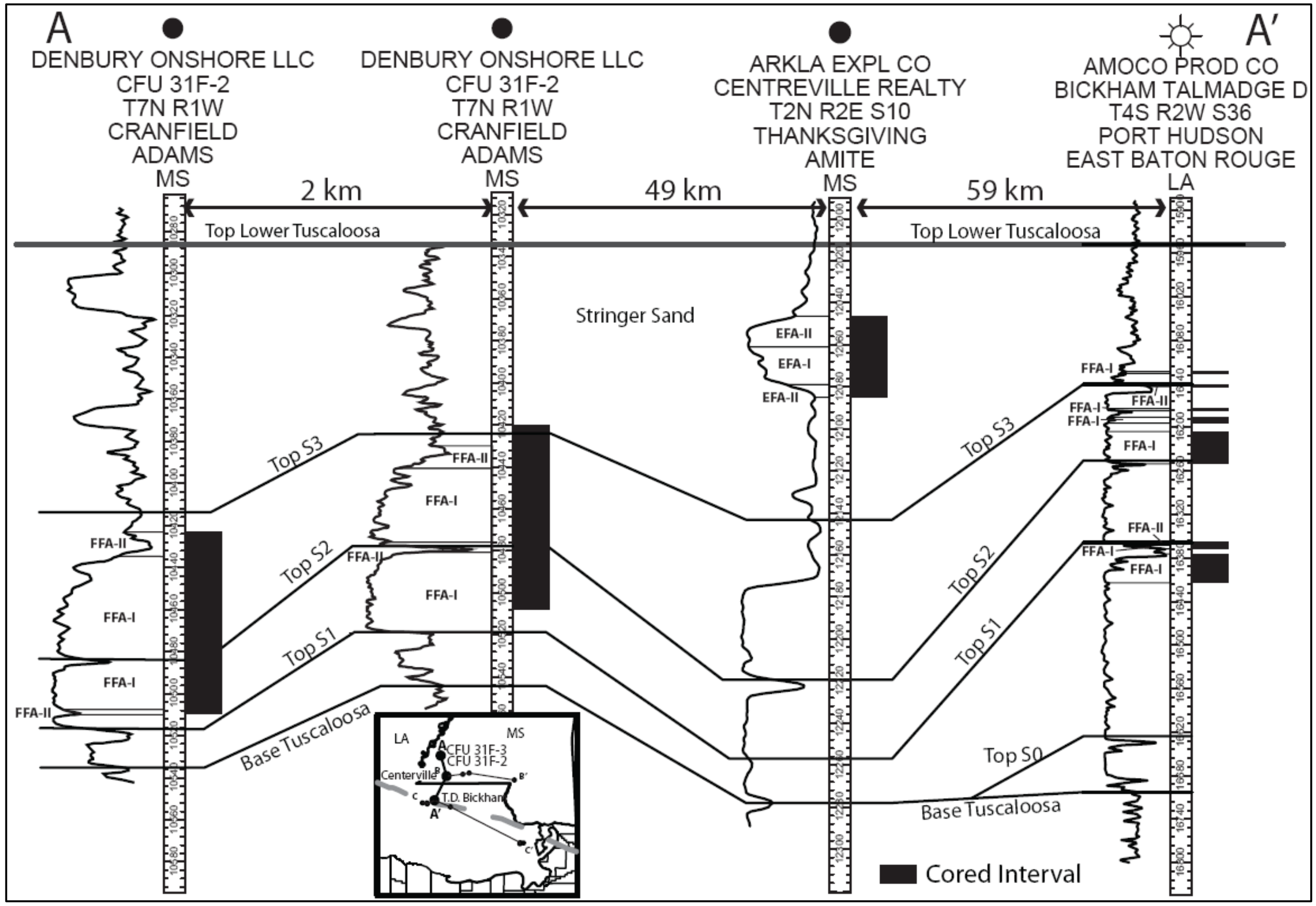

Figure 3.13. Cross section A-A’ flattened on the top of the Lower Tuscaloosa (see Figure 3.1 for location). The section passes through three core containing fluvial facies associations (FFA-I and FFA-II) and one core containing estuarine facies associations (EFA-I and EFA-II). Wireline logs are correlative to the cored intervals used in this study. Sequence tops are from the sequence stratigraphic framework determined in Chapter 2.

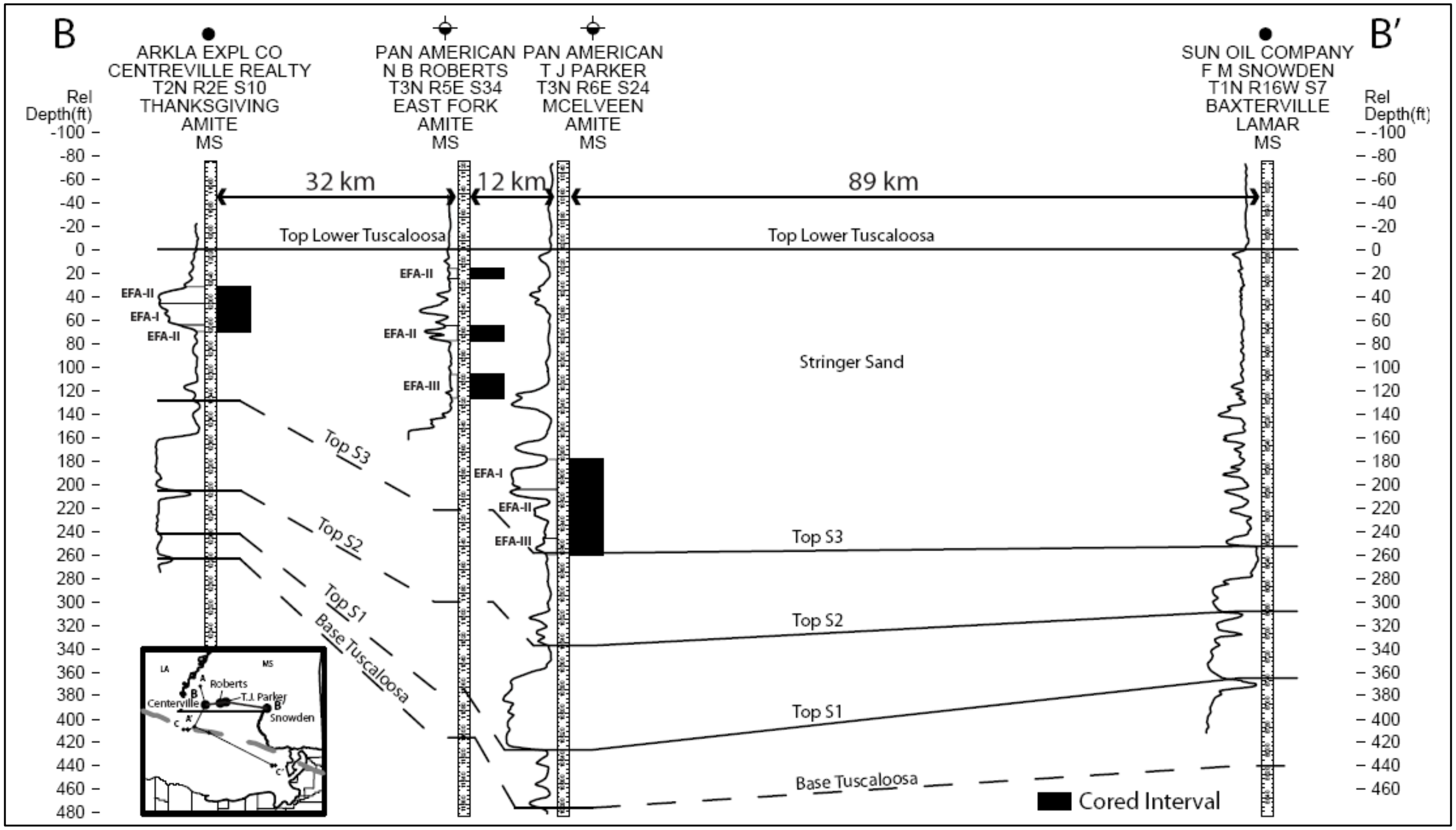

Figure 3.14. Cross section B-B’ flattened on top of the Lower Tuscaloosa (see Figure 3.1 for location). The section passes through three core containing estuarine facies associations (EFA-I, EFA-II, and EFA-III) and one additional wireline well log showing the Lower Tuscaloosa Formation character to the east where cored intervals were not studied. Wireline logs are correlative to the cored intervals used in this study. Sequence tops are from the sequence stratigraphic framework determined in Chapter 2.

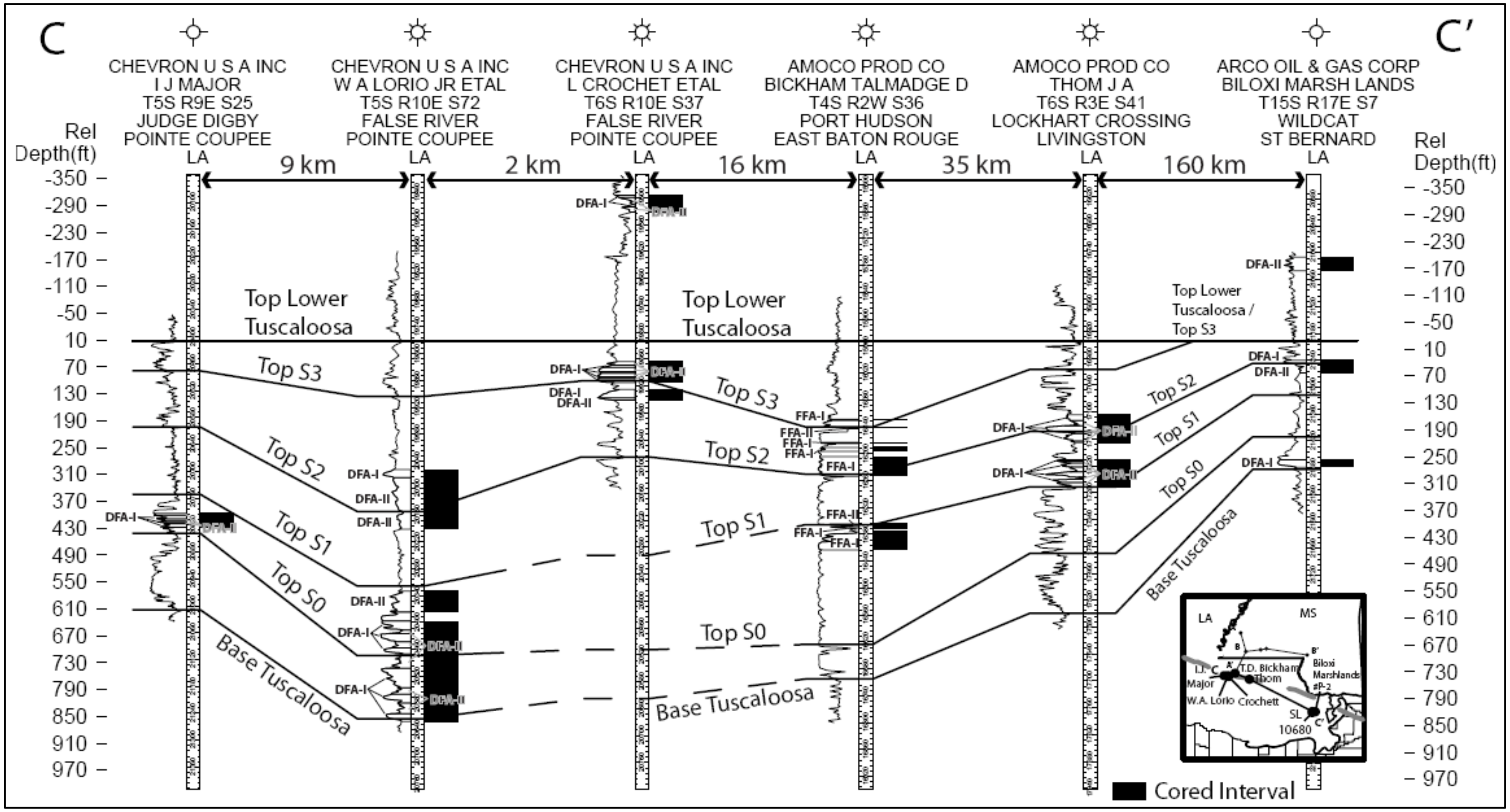

Figure 3.15. Cross section C-C' flattened on top of the Lower Tuscaloosa (see Figure 3.1 for location). The section is proximal to the Lower Cretaceous shelf edge and contains six cores with deltaic facies associations (DFA-I and DFA-II) and one with fluvial facies associations (FFA-I and FFA-II). Wireline logs are correlative to the cored intervals used in this study. Paleoenvironmental interpretations for the cored intervals are included on wireline logs. Sequence tops are from the sequence stratigraphic framework determined in Chapter 2.

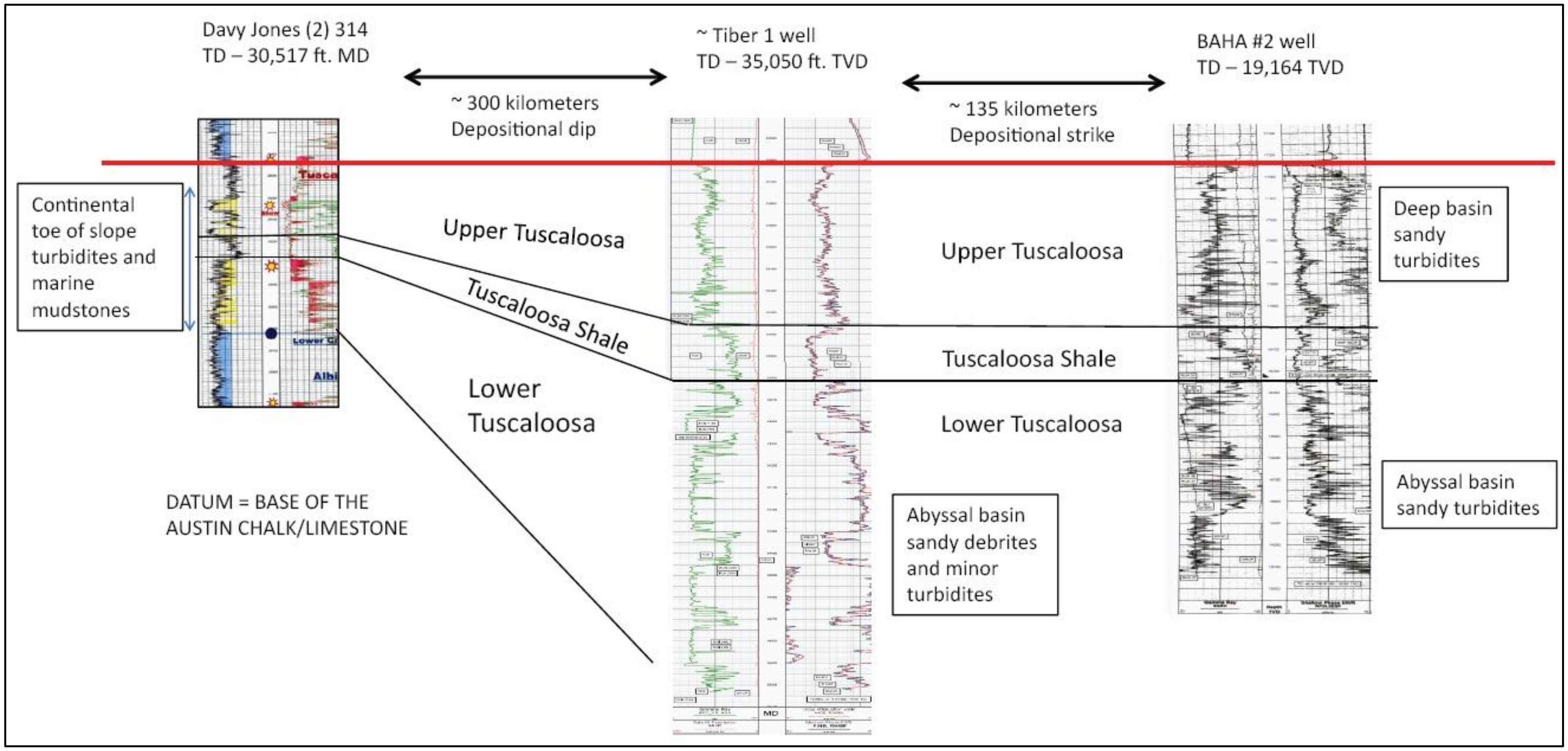

Figure 3.16. Cross section through three wells penetrating the Tuscaloosa interval in the offshore Gulf of Mexico (see Figure 3.1 for the locations of the three wells) and interpretation of the lithostratigraphic units of the Tuscaloosa Formation.

signature in deltaic deposits (Bhattacharya, 2006). The base of these coarsening upward cycles exhibit high gamma ray readings which correlate to the prodelta facies association (DFA-II). Capping these cycles are low gamma readings related to the sandy intervals of the delta front facies association (DFA-I).

Gravity-Driven Deepwater Deposits (GFA-I, GFA-II)

Wireline-log patterns of GFA-I and GFA-II are only found in distal Tuscaloosa equivalent penetrations in the marine offshore of the Gulf of Mexico in the Davy Jones #2, BAHA #2, and Tiber well (Figure 3.16); however core photographs of the Tiber core allow for correlation of the facies with the wireline log patterns. The Tiber well in the Keathley Canyon area (see Figure 3.1 for location) was drilled to 35,030 ft. total depth and provides the most complete data set of all the Cretaceous-age deepwater penetrations. The top of the Tuscaloosa Formation is interpreted to be at 33,730 ft. total depth and data over the interval show over 305 m of thick blocky sands broken into numerous packages each ranging from ~15 to 45 m thick (GFA-I) (Figure 3.16). Each is separated by maximum hundred foot intervals of shales (GFA-II) interbedded with sharp-based, sharp-topped 3-4.5 m-thick sandy deposits. These deposits are interpreted to represent deepwater deposition of coarse-clastics fed from the Tuscaloosa-age shelf edge canyons located roughly along modern day southern Louisiana (Chapter 2).

DISCUSSION

Valley Incision

Incised valleys often result from erosion by fluvial systems driven by a variety of mechanisms. Incision at the base of incised valleys commonly occurs during lowstands driven by eustatic sea level fall and/or local tectonics; however, incision can also occur due to an increase in discharge or decrease in sediment supply (Blum and Tornqvist, 2000; Ardies et al., 2002; Boyd et al., 2006). Valley incision typically correlates with a regionally extensive unconformity, the location and extent of which is often partially controlled by paleotopographic features and structural trends (Ardies et al., 2002; Wadsworth et al., 2003; Boyd et al., 2006). Although sediments are intermittently deposited in valleys throughout the entire depositional-cycle, it is only during base level rise(s) following the valley-cutting lowstand that the valley typically fills with sediments (Boyd et al., 2006). Incised valleys commonly contain a compound fill in which several cut and fill cycles incise into and subsequently fill the valley (Boyd et al., 2006).

In the Lower Tuscaloosa Formation, we surmise that valleys were initially cut during the large scale drop in base level associated with the regionally-extensive mid-Cenomanian unconformity (Figure 3.17). Each of the subsequent sequences (Chapter 2) represents an individual fill cycle of fluvial channel or estuarine deposits bounded by erosional surfaces controlled by fourth-order fluctuations in sea level. It appears as if little to no deposition occurred in the valleys during erosional events and sediments were instead bypassed to the shelf and basin floor.

Deltas

Deltas and their depositional processes have been well studied (Galloway, 1975; Bhattacharya, 2006; Porebski and Steel, 2006; Steel et al., 2008). Deltas can be progradational, aggradational, or retrogradational in nature and form coarsening-upward successions as delta-front sands and sandy, mouth bars are deposited on top of muddy, prodelta deposits (Bhattacharya, 2006). Deltas can be further classified according to the dominant physical processes (waves, tides, and stream flow) acting on the delta during deposition. Wave-, tide-, and river-dominated deltas can be distinguished based on their differing morphologies, facies, and facies architectures (Galloway, 1975; Steel et al., 2008). Wave-dominated deltas commonly have an arcuate shape in map view and sedimentary structures formed by wave currents (i.e. hummocky cross stratification) (Galloway, 1975; Bhattacharya, 2006). Tide-dominated deltas are characterized by sedimentary structures formed by tidal processes (i.e. wavy, flaser, and lenticular bedding, and mud draped structures) (Bhattacharya, 2006; Carmona et al., 2009). Similar to estuaries, deltas can also display impoverished trace fossil suites of the *Cruziana* Ichnofacies (McIlroy, 2004; MacEachern and Bann, 2008; Carmona et al., 2009). Ichnology can further help distinguish between different types of deltas (McIlroy, 2004; MacEachern and Bann, 2008; Carmona et al., 2009). River-dominated deltas generally exhibit the most stressed conditions for trace-makers due to higher suspended sediment concentrations and a large, constant influx of fresh water, and therefore contain the most impoverished trace fossil suites. In contrast, wave-dominated deltas generally contain the least impoverished trace fossil suites with a comparatively lesser amount of suspended sediment and less salinity variations. Tide-dominated deltas typically fall between river and wave-dominated deltas with intermediate suspended sediment concentrations and moderate salinity variations (McIlroy, 2004; MacEachern and Bann, 2008; Carmona et al., 2009).

In the Lower Tuscaloosa Formation deltas were formed at the mouth of incised valleys by deposition of sediments delivered by the fluvial systems during times of lowstand and when sediment supply was sufficient to deliver sediment to the shelf edge despite rising sea level (Figures 3.17). Delta-2 is most likely a tide-influenced delta because of the combination of abundant lenticular and flaser bedding and an extremely limited trace fossil suite. Delta-1 was likely a wave-influenced delta because of its less impoverished trace fossil suite and asymmetrical, arcuate shape in isopach maps indicating it was reworked by waves. Wave currents created longshore drift which redistributed sediments along strike to the shoreline in a northwesterly direction towards the opening of the Western Interior Seaway (Figures 3.2) (Chapter 2). This asymmetry in the deltaic morphology is common to wave influenced deltas where waves obliquely meet the shoreline (Bhattacharya, 2006).

Deepwater Gravity-driven Deposits

Over the past two decades there has been much debate surrounding sand-rich deepwater gravity-driven deposits focusing on turbid vs. laminar flow (Lowe, 1982; Shanmugam, 1996; Mulder and Alexander, 2001; Baas, 2004; Plink-Bjorklund and Steel, 2004; Amy et al., 2005; Talling et al., 2007; Haughton et al., 2009; Shanmugam et al., 2009; Breien et al., 2010). Adding to the complexity of the debate, a combination of stratified flow types can occur where laminar flow is found at the base of the flow and turbid flow dominates the uppermost flow (Talling et al., 2007; Breien et al., 2010). Combined flow types may also occur together when flow rheology differs along the transport path due to changes in slope, flow speed, and flow composition (Shanmugam, 1996; Haughton et al., 2009; Breien et al., 2010). This complexity has led many workers to group all deepwater gravity-driven deposits into “turbidites” (for discussion see Lowe, 1982; Shanmugam, 1996; Mulder and Alexander, 2001; Baas, 2004; Breien et al., 2010). However, Lowe (1982) used the term “high density turbidites” to explain thick, sand-rich, essentially “structureless” deepwater deposits. More recently, Shanmugam (1996) suggested Lowe’s (1982) “high density turbidites” be renamed “sandy debrites” because they (1) lack the expected stratification of the traditional Bouma Sequence (Bouma, 1962) and (2) were likely deposited by laminar flow supported by grain-to-grain contact. Some sand-rich gravity flows have been attributed to hyperpycnal flows which deposit sediment as turbid flow but have almost the same appearance as sandy debrites (Plink-

Bjorklund and Steel, 2004). However, in contrast to sandy debrites, hyperpycnal flows travel only a few km from their source, lack floating clasts, are generally better sorted, and can be organic-rich (Plink-Bjorklund and Steel, 2004).

Deepwater deposits of the Tuscaloosa Formation, located 400-500 km from their shelf source are not likely to be hyperpycnal flows because of the 100's of km separating them from their source. Rare graded bedding in the Tuscaloosa Formation deepwater deposits does indicate that some of these flows exhibit turbidite characteristics (Bouma, 1962; Amy et al., 2005; Shanmugam et al., 2009). However, facies analysis from core photographs in the Tiber well indicate the majority of these deposits resemble sandy debrites as they are dominated by amalgamated sandstones with a low percentage of mud, contain structureless bedding and a general a lack of graded bedding, soft sediment deformation, moderate to poor sorting, floating clasts, fluid escape structures, and sharp contacts (Lowe, 1982; Shanmugam, 1996; Amy et al., 2005; Shanmugam et al., 2009). It is not unusual for gravity-driven deposits to be found several hundreds of kilometers from the shelf edge (Wynn et al., 2002; Posamentier and Walker, 2006; Anka et al., 2009). Thick sequences of Tuscaloosa Formation clastics in the Davy Jones #2, BAHA #2, and Tiber wells prove that this system transported sediment 100's of km from the shelf edge changing our understanding of the Tuscaloosa Formation as simply a present day onshore deposit. Far-travelled sand-rich gravity deposits have also been identified in deepwater Gulf of Mexico wells containing the Wilcox Group which is found stratigraphically above the Tuscaloosa Formation (Zarra, 2006; Meyer et al., 2007; Rains et al., 2007; Wornardt, 2010).

The Davy Jones #2 well (see Figure 3.1 for location) was drilled to a total depth of 30,546 ft. and contains Tuscaloosa-age strata equivalent to the wells penetrating Cretaceous reservoirs in the onshore False River field of Louisiana (McMoRan

Exploration, 2011). Davy Jones #2 wireline logs record 230 m of Tuscaloosa Formation clastics penetrated at 29,000 ft. These deposits rest unconformably on Albian limestones (Figure 3.16) (McMoRan Exploration, 2011). Predominantly low gamma ray readings across this interval are interpreted as sandy intervals that can be broken down into three distinct packages. The lowermost package is approximately 120 m thick, composed of interbedded sands and shales, and shows a ratty log motif with a gradational base and a slight overall fining-upward character before being sharply overlain by shales. The shale (likely a flooding surface) is overlain by a 34 m thick coarsening up package, which is in turn overlain by a smaller, 12 m thick coarsening up package. This coarsening-up package is capped by a sharp-based, blocky sand package overlain by marine shales and carbonates. On the basis of log motif and overall paleogeographic location in the Late Cretaceous continental slope, these deposits are interpreted to be slope fan gravity flow deposits overlain by a series of prograding clastic wedges (Mitchum et al., 1993). These more proximal base-of-slope fans and prograding wedges located in the Davy Jones #2 well are evidence of significant ponding of coarse-grained gravity flow deposits in the slope and toe of slope positions.

The BAHA #2 well in the Perdido fold and thrust belt is the offshore well located most distal from the paleo-shelf edge in the subsurface of southern Louisiana (see Figure 3.1 for location). The well was drilled to 19,164 ft. true vertical depth and penetrated over 305 m of Tuscaloosa-aged clastics. The log of the BAHA #2 shows six coarsening- and fining-up sequences grouped into two major cycles, each containing three sequences (Figure 3.16). The entire well-motif shows an overall ratty appearance indicating shale and sand interbedding. Sand percentage decreases upward. These cycles are overlain by over 300 m of deepwater chalks and limestone. Sands in this interval are reported to be turbiditic in origin (Zarra, 2006; Meyer et al., 2007; Rains et al., 2007; Wornardt, 2010) and microforaminifera indicate that sediments are equivalent to deposits of the Late Cretaceous Tuscaloosa/Woodbine shelf edge located roughly near present day Houston, Texas (Wornardt, 2010).

It is likely that during the time of valley incision, when base level was lowest and sediment supply was high, bypass allowed large volumes of sediment to travel far out into the basin as deepwater gravity deposits (Figure 3.17). The source for these deepwater gravity flows are most likely deltas near the shelf edge in southeast Louisiana.

Fluvial and Estuarine Incised Valley-fills

During base level rise(s) following the valley-cutting lowstand, incised valleys typically fill with sediments (Dalrymple et al., 1992; Boyd et al., 2006). These sediments may be fluvial, estuarine, brackish-water, and/or marine deposits (Dalrymple et al., 1992; Boyd et al., 2006). During the transgression that followed the mid-Cenomanian drop in base level, fluvial systems characterized by fining upward cycles aggraded in the Lower Tuscaloosa valleys (Figure 3.17). Our data show that amalgamated braided stream deposits 5 - 40 m thick per sequence initially filled the valley. This fill is well-known regionally as the “base of the Massive sand”. It has been suggested that the fluvial systems of the Lower Tuscaloosa originated as braided systems and then transitioned to meandering systems during base level rise (Chasteen, 1983; Klicman et al., 1988) although data in this chapter does not capture this transition. Trace fossils including *Asterosoma*, *Chondrites*, *Cylindrichnus*, *Ophiomorpha*, *Palaeophycus*, *Planolites*, *Phycosiphon*, *Rhizocorallium*, *Rosselia*, *Teichichnus*, and *Thalassinoides* found in some channels within incised valleys suggest rare marine water incursions into valleys during deposition of the Massive sand. These bioturbated channels are found at the distal end of the northwest valley north of the shelf edge and Delta-1 in the T.D. Bickham well (Figure

3.1). A transgressive surface is commonly found in incised valley-fills between the fluvial and estuarine fill (Boyd et al., 2006). Channel-fills in the most proximal Lower Tuscaloosa Formation cores of the Cranfield field are capped by fine-grained mottled silts containing an impoverished suite of marine trace fossils including *Rhizocorallium* and *Planolites* suggesting a stressed, nearshore marine environment. This change in grain size, facies, and ichnology likely marks an increase in base level and the transition of the valley from fluvial-dominated to an estuary. This siltstone also marks the boundary between the Massive and Stringer sands. As base level continued to raise estuarine facies (Stringer sands) filled the valley above previously deposits fluvial sands (Figure 3.18).

Estuaries have been classified as either wave- or tide-dominated (Dalrymple et al., 1992; Allen and Posamentier, 1993; Nichol et al., 1997; Boyd et al., 2006). Wave-dominated estuaries exhibit a complex tripartite environmental zonation with fluvial processes dominating proximal environments, waves and/or tidal currents dominating distal environments, and a relatively low energy central basin occurring where current and wave-driven forces meet to cancel out landward directed and basinward directed energies (Dalrymple et al., 1992; Allen and Posamentier, 1993; Nichol et al., 1997; Boyd et al., 2006). Conversely, in tide-dominated estuaries the tripartite stacking pattern is not as pronounced as stronger tidal energies from the seaward portion of the estuary commonly infiltrate well into the low-energy zone (Boyd et al., 2006). This influx of tidal energy also results in sedimentary structures common to tidally influenced systems such as flaser bedding and mud draped structures (Dalrymple et al., 1992; Boyd et al., 2006). The transgressive nature of all estuarine deposits drives the preservation of a retrogradational stacking pattern of the tripartite zonation. In the Lower Tuscaloosa Formation above the transgressive surface at the top of the Massive sand, wave dominated estuarine facies of the Stringer sand were emplaced in a transgressive sequence consisting from base-to-top of bayhead delta deposits, estuarine central basin deposits, and barrier bar deposits (Figure 3.18). At the base of the barrier bar facies a reworked interval marks the transgression of the barrier bar. Trace fossil suites in the estuarine cores display a low abundance and diversity as would be expected in a stressed estuarine environment (Boyd et al., 2006; MacEachern and Bann, 2008). The highest diversity and degree of bioturbation occur in the lower energy central basin deposits. Overall, core and wireline logs record a fining upward pattern from the bayhead delta facies to the fine-grained central basin facies and a coarsening upward pattern from the central basin to the sandy barrier/shoreface facies similar to that expected for a wave-dominated estuarine succession (Figure 3.14).

PALEOTOPOGRAPHICAL/STRUCTURAL INFLUENCE ON FACIES

The regional slope during Lower Tuscaloosa time was from north-northeast to south-southwest, away from the uplift of the Appalachian Plateau (Chapter 2). Regional structural highs and lows had a strong influence on the overall development of valley systems in the Lower Tuscaloosa Formation (Chapter 2). Valleys trended north-northeast to south-southwest, forming between basement highs and were also redirected to the northwest-southeast by structural lineaments along major transfer faults formed during the opening of the Gulf of Mexico (Figure 3.2). Structures proximal to the paleo shelf edge helped direct currents along the shelf and at the shoreline which in turn influenced shoreline, estuary, and delta geometries (Chapter 2). In our study, the incised valley associated with Delta-2 (Figures 3.2, 3.17) and the basement uplifts associated with it may have served to focus tidal energy resulting in the observed increased tidal signatures in deposits of Delta-2.

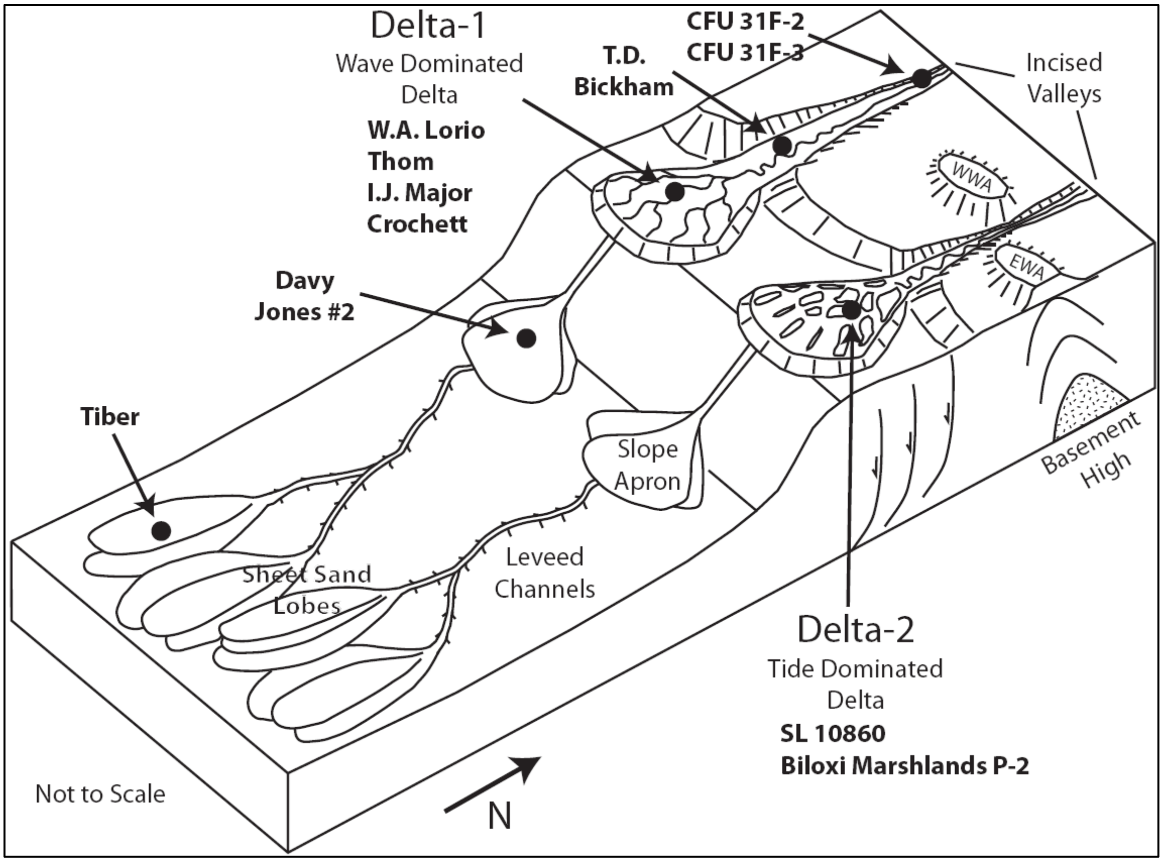

Figure 3.17. Paleoenvironmental reconstruction of the depositional systems of the early Lower Tuscaloosa Formation. Depositional systems are dominated by incised-valleys containing fluvial systems that feed wave and tide dominated deltas at the shelf edge. Bypass during times of relative lowstand both at the base of the valleys (MCU) and during the compound fill of the valleys allowed for far travelled gravity deposits into the basin plain. Reconstruction shows the incised valley between the west (WWA) and east (EWA) Wiggins arch and associated tidally-influenced valley-front delta. In contrast, the valley to the west was influenced by waves.

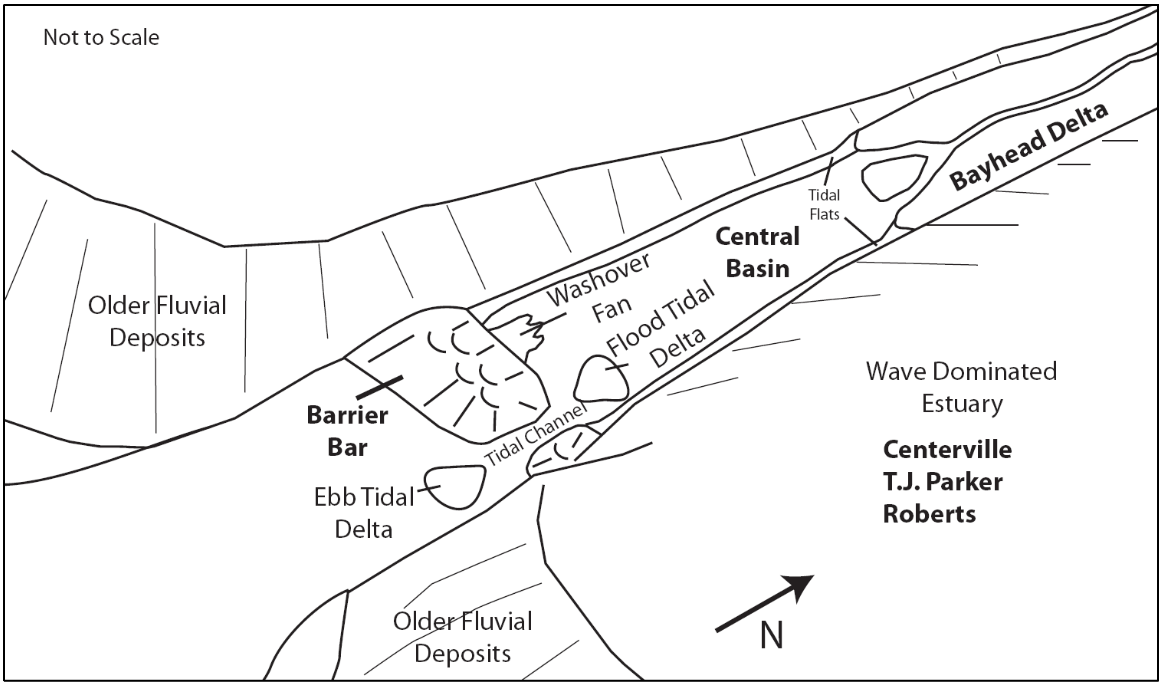

Figure 3.18. Paleoenvironmental reconstruction of the depositional systems of the late Lower Tuscaloosa Formation. Depositional systems are dominated by estuarine systems (barrier bar, central basin, and bayhead delta) within the incised valleys as overall transgression progresses.

Conversely, other valleys in western Mississippi and the Louisiana panhandle trend northwest-southeast, nearly parallel to their shoreline creating a broad estuary during transgression (Figure 3.2). The combination of a wider valley located within a sheltered embayment may have created more favorable conditions for a wave-dominated delta (Delta-1) in this part of the system.

CONCLUSIONS

The succession of depositional environments in the Lower Tuscaloosa Formation from source-to-sink includes fluvial and estuarine incised valley fills, deltas, and deepwater gravity-driven deposits. Incised valleys formed during an initial large-scale lowstand that produced the mid-Cenomanian unconformity. The trends, locations, and sizes of these incised valleys were influenced by basement structures. During this lowstand and subsequent cut-and-fill cycles within the valleys, sediment bypass allowed sediments to reach the shelf-edge. Some of the sediment reaching the shelf-edge was deposited as shelf-edge deltas while some of the sediment continued into deepwater environments. As sea level rose during transgression fluvial sediments aggraded in the incised valleys. High sedimentation rates continued to supply sediment to tide and wave-influenced deltas and deepwater gravity deposits continued to aggrade. As transgression progressed wave-dominated estuaries formed within the incised valleys. The deposition of the Middle Tuscaloosa Marine Shale marks the end of the complex cycle of deposition in the Lower Tuscaloosa Formation.

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Chapter 4: Fluvial channel morphometrics of the Lower Tuscaloosa Sandstone

INTRODUCTION

The Lower Tuscaloosa Formation is currently of great interest to companies involved in hydrocarbon exploration in the offshore Gulf of Mexico due to the possibility of deep oil and gas reservoirs in what may be a large un-explored deepwater clastic fan system that rivals the Wilcox fans in extent. Although, little is known about the extent and amount of Tuscaloosa-aged sediment that exists in this offshore frontier area due to scarcity of well penetrations and poor quality seismic data, the extensive data in the more proximal feeder systems to this deepwater play offer opportunity to examine the capacity of the source systems and thus estimate the potential for sediments to overcome proximal accommodation sinks and spread basinward.

Currently, the BAHA #2, the Davy Jones #2, and the Tiber wells are the three wells known to have penetrated the late Cretaceous Tuscaloosa Formation in paleo-slope and basin floor positions in the offshore Gulf of Mexico. These offshore deposits, identified as sandy debrite basin floor fans are fed by shelf-edge deltaic and fluvial systems confined to a regional network of incised valleys located in modern day onshore Mississippi and Louisiana (Chapters 2 and 3). Onshore strata in the Lower Tuscaloosa are much better documented and accessible than their offshore equivalents. Tens of thousands of wireline well logs and dozens of cored intervals exist through Lower Tuscaloosa proximal deposits, and offer the foundation for assessing knowledge on the size and capacity of the fluvial systems that may have fed these deepwater deposits. In addition, improved understanding of the nature and capacity of the Lower Tuscaloosa fluvial systems, allow for a selection of feasible analogs from ancient outcrop and modern fluvial systems for additional insights into architecture of onshore reservoirs currently under development, and to examine the capacity of these analogs to produce their own deepwater fan deposits. Such insights are the focus of this chapter.

Measurements of thickness and grain size were collected from 4 cored intervals of the Lower Tuscaloosa Formation and measurements of thickness were calculated from 136 wireline logs of fluvial channels in Lower Tuscaloosa Formation. This process provided data on 384 separate point bar deposits and was used as a basis for this study. The cored intervals are located in the CFU 29-12, CFU 31-2, and CFU 31-3 wells in the Cranfield field of southwest Mississippi and the T.D. Bickham well in Louisiana (Figure 4.1). The wireline well logs are located in southwest Mississippi (Figure 4.1). These measurements were used to calculate channel depths, channel widths, and channel belt widths (Bridge, 2003) for paleochannels in the Lower Tuscaloosa Formation. Results were compared to interpreted fluvial outcrop and modern day fluvial systems to identify potential ancient and modern day analogs for the Lower Tuscaloosa Formation (Gibling, 2006). Such characterization of the Lower Tuscaloosa channel systems has never been done and can aid in understanding the three-dimensional nature of the Lower Tuscaloosa depositional systems, the carrying capacity of the Lower Tuscaloosa fluvial systems, and ultimately lead to insights on the volume of sediments that have passed through these late Cretaceous valleys to residence at shelf edges and in canyons feeding the deepwater environments during late Cretaceous time.

PREVIOUS WORK - QUANTIFYING FLUVIAL CHANNELS

Determining the thickness and width of subsurface channel and channel belt deposits is essential to many aspects of the exploration and development in these deposits. Scientists typically need such information for calculating reservoir volumes, deciding placement and density of development wells, locating new and unpenetrated

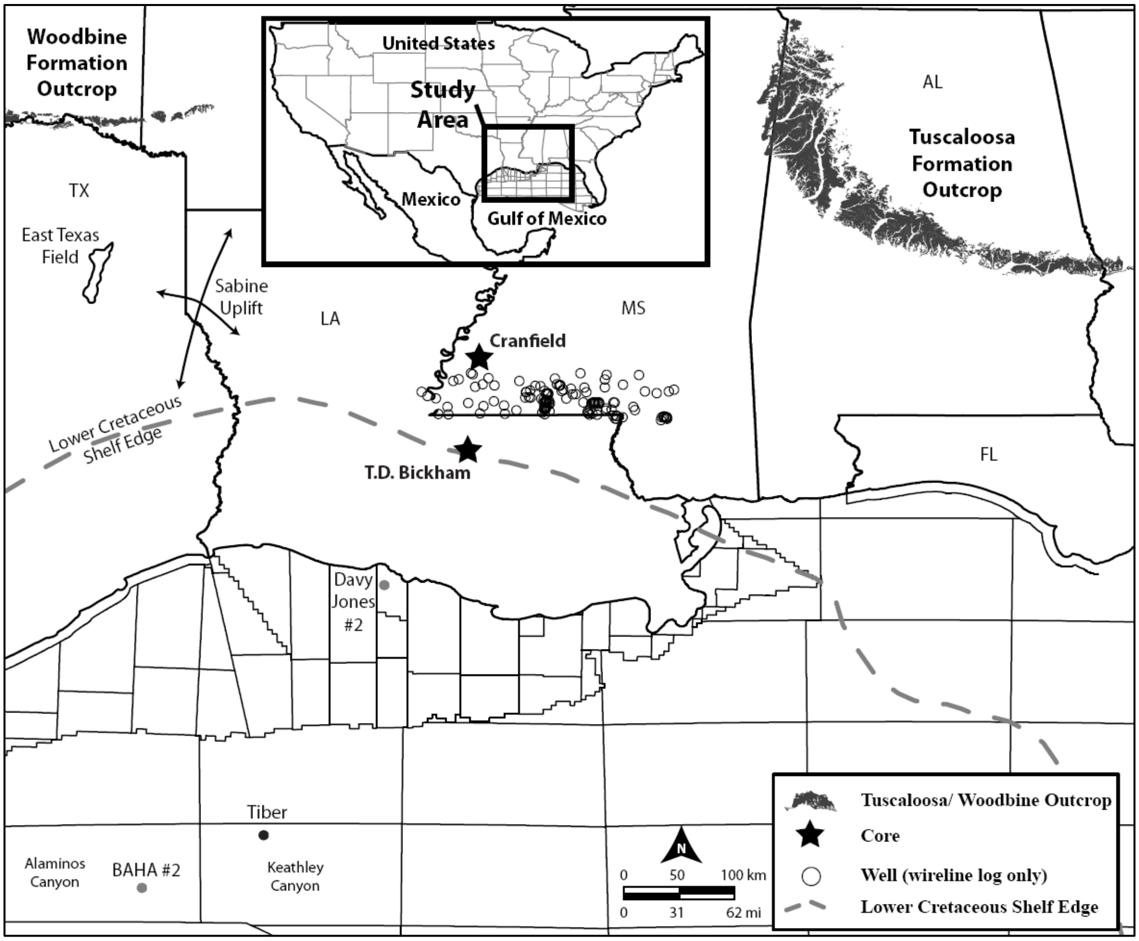

Figure 4.1. Base map showing the study area and location of core and wireline well logs used in Chapter 4. The locations of deep offshore wells that have penetrated Tuscaloosa-aged basin floor facies are shown for reference.

compartments of the fluvial reservoir system, mapping lateral connectivity, defining producibility, and 3D reservoir modeling. Bridge (2003) discussed traditional methods for determining channel and channel belt widths such as correlating log signatures between wells, outcrop analogs, empirical equations derived from studies of modern rivers, and estimation from amplitude analysis of 3D time slices. He noted that all of these methods present significant drawbacks in accurately determining channel dimensions. Bridge (2003) went on to develop an independent means for calculating channel flow depth using the relationship between distributions of dune height and cross set thickness and the known relationship of dune height to water depth. Subsequently, channel width and channel belt width can be estimated from flow depth. Bridge (2003) also describes how channel depth can be estimated from wireline well logs. The method for making such estimations is discussed in greater detail in the methods section below.

Gibling (2006) compiled a large amount of morphometric data from literature focusing on fluvial systems. The results of this work showed broad similarities in channel widths and depths for several different river types (e.g. Meandering, braided, valley fill, etc…) including maximum, minimum, and "common range" width and depth values for each river type. This compilation of data was presented in graphs for each fluvial type and allows for an easy and reliable way to classify any fluvial system being studied in the subsurface and identify outcrop analogs for that system (Gibling, 2006). The quantitative work done herein is integrated with the observations of Gibling (2006) to identify possible outcrop analogs for the Lower Tuscaloosa Formation. Data of channel depths and widths of modern rivers derived from the USGS National Hydrography Dataset (NHD) (Kiel, personal communication) were compared with the calculated channel depths and widths of the Lower Tuscaloosa to determine viable modern analogs for these ancient systems.

METHODS

The method used in this study to estimate paleochannel depth and width and channel belt width of the Lower Tuscaloosa fluvial systems from core and wireline log data is outlined in detail in Bridge (2003) and is briefly discussed herein. To determine paleochannel depth from cored intervals two relationships were utilized; (1) the relationship between distributions of dune height and cross set thickness , and (2) the known relationship of dune height to water depth. Additionally, the relationship between and channel depth to channel belt width was used. To use the Bridge (2003) method thicknesses were measured of as many cross sets as possible from each of the four selected core, and the mean (s_m) and standard deviation (s_d) of cross set thickness was determined. An initial test that $s_d/s_m \approx 0.88$ (±0.3) allows for this method to be used with confidence (Bridge, 2003). From these measurements of s_m and s_d the mean dune height (H_m) was estimated. An estimation of H_m (Leclair and Bridge, 2001; see Bridge, 2003 for complete derivation of the formula) for the Lower Tuscaloosa Formation was made by the following equation:

$$H_m = 5.3(s_m /1.8) \quad \text{(eq. 1.1)}$$

Formative flow depth (d) is related to H_m such that d/H_m averages between 6 and 10 for all types of river dunes (Bridge, 2003). From this a maximum, minimum, and mean bankfull flow depth was estimated using the following equations:

$$d_{max} = 10H_m \quad \text{(eq. 1.2)}$$

$$d_{min} = 6H_m \quad \text{(eq. 1.3)}$$

$$d_m = 0.5d_{max} \quad \text{(eq. 1.4)}$$

Mean bankfull flow depth (d_m) is related to the channel width (w) (Leeder 1973) and the maximum and minimum channel belt width (cbw_{max} and cbw_{min}) (Bridge and

Mackey, 1993; Bridge, 2003). An estimation of these values was made for the Lower Tuscaloosa Formation from the following equations:

$$w = 6.8d_{max}^{1.54} \quad \text{(eq. 1.5)}$$

$$cbw_{max} = 192d_m^{1.37} \quad \text{(eq. 1.6)}$$

$$cbw_{min} = 59.9d_m^{1.8} \quad \text{(eq. 1.7)}$$

In situations where core was not available an estimate of maximum bankfull flow depth was made from the full height of point bar deposit interpreted from well logs. Point bars were identified in wireline logs as sharp-based, fining-upward to blocky cycles. Care was taken to avoid erroneous estimations of point bar height by measuring complete, fining-upward, point-bar cycles. While the bases of the point bars were often easy to identify, identifying the tops of the point bar deposits often proved more problematic. The upper portion of the bar often grades into a muddy upper bar which can be mistaken for a floodplain deposit leading to an underestimation of point bar height (Bridge, 2003). In addition, point bars in the Lower Tuscaloosa often appeared to be truncated by an overlying point bar, which can lead to amalgamated channel deposits causing an overestimation of point bar height. These risks regarding the accuracy of measurement are somewhat mitigated statistically by the sheer number of point bar measurements taken. Figure 4.2 represents a typical point bar measured from the Lower Tuscaloosa Formation for this study. Once these point bars were measured equations 1.4, 1.5, 1.6, and 1.7 were used to estimate the mean bankfull flow depth, channel width, and channel belt maximum and minimum.

Estimates of channel depth and width for the Lower Tuscaloosa Formation were graphed alongside data provided by Gibling (2006). These graphs helped to classify the type of fluvial systems in the Lower Tuscaloosa based on channel geometry, and allowed

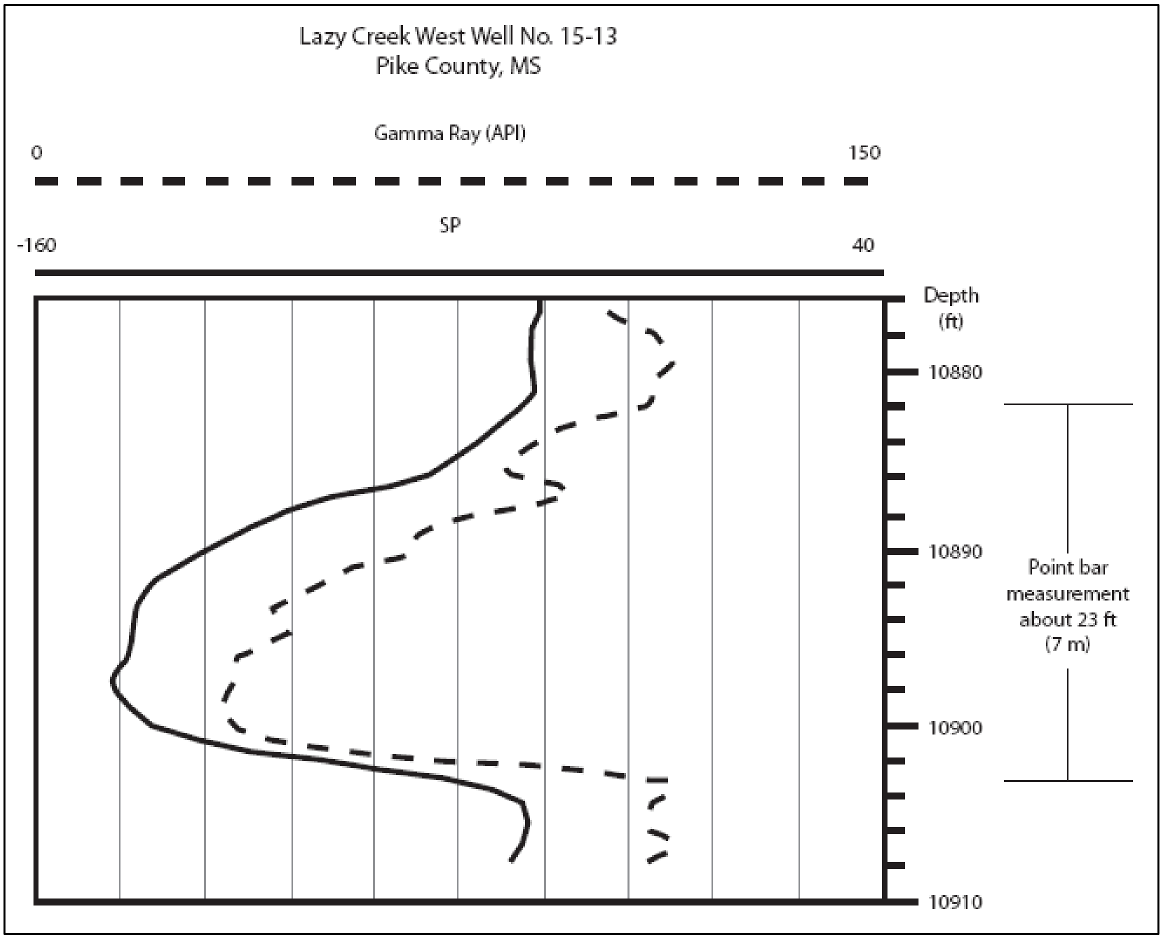

Figure 4.2. Example of a typical Lower Tuscaloosa Formation point bar interpreted from wireline well logs.

for comparison of the Lower Tuscaloosa channel morphologies with a large number of potential outcrop analogs. Likewise, modern analogs were determined by comparing Lower Tuscaloosa channel morphometrics with channel depths and widths of modern U.S. Rivers derived from the USGS National Hydrography Dataset (NHD) (Kiel, personal communication).

OBSERVATIONS

The study area was divided into three separate areas: Area 1 representing an incised valley located to the west of the study area, Area 3 representing an incised valley located to the east, and Area 2 the interfluve area between these two incised valleys (Figure 4.3). This geographic division of the wells and resultant distribution of analyses was based upon the mapped geomorphology of the Lower Tuscaloosa and the desire to accurately assess any trends related to processes unique to individual regions versus those that might be more regional in nature. Cores from the Cranfield field appear to be on the edge of Areas 1 and 2 but are on an isopach thin similar to Area 2. The T.D. Bickham core is located just 5 km to the north of the shelf edge distal to Area 1 (Figure 4.3).

Area 3 contains the thickest deposits of the Massive sand (Figure 4.3) with thicknesses ranging from 35 to over 90 m. Calculations show that Area 3 also contains the largest channels and channel belt widths. Average channel depth for Area 3 is 7.77 m and average channel width is 340 m. Channel belt width ranged from 2748 to 3346 m (Table 4.1).

Area 1 is the second thickest area on the Massive sand isopach map (Figure 4.3) with sand thicknesses ranging from 35 to 60 m. Channel dimensions in Area 1 are also the second largest with channel depth and width averaging 5.62 m and 306 m respectively. Channel belt width ranged from 1533 to 2148 m (Table 4.1). Area 1 offered

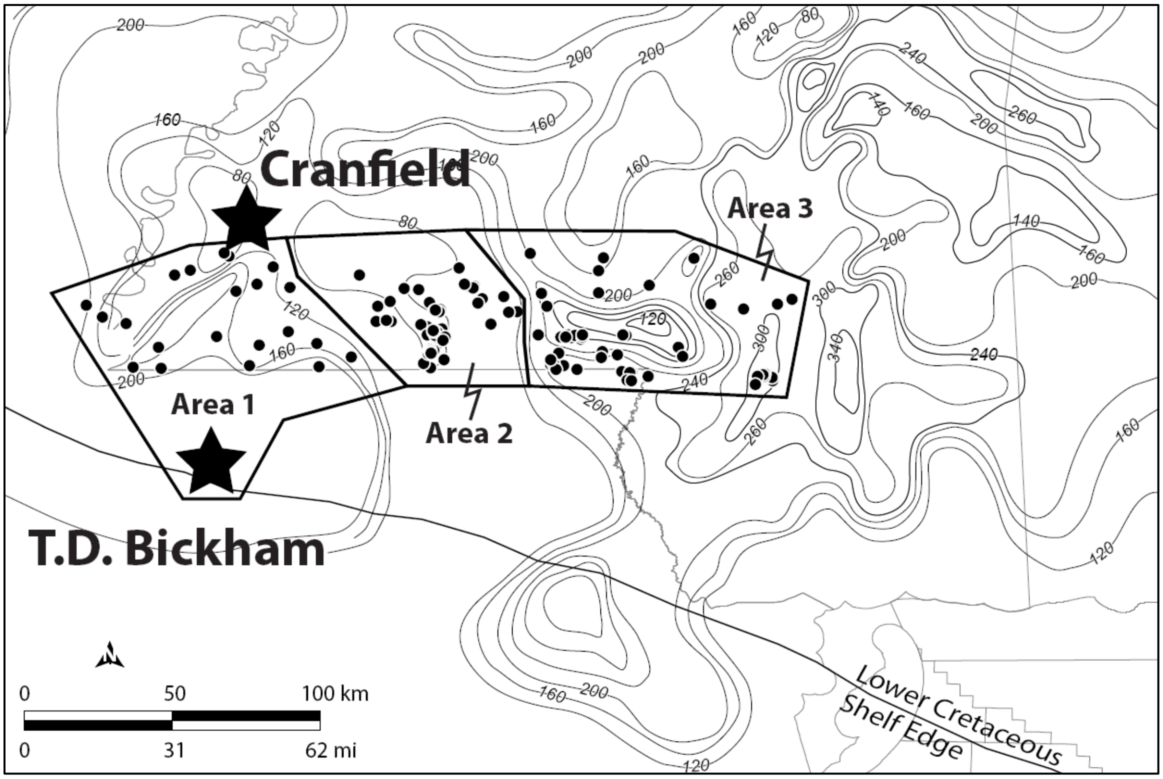

Figure 4.3. Map delineating the division of the three areas used in this study. Locations of wireline well logs are overlain on the isopach map for the Massive sand (Chapter 2). Note the changes in isopach thickness between the three areas. Also, note the similarity in isopach thickness of the T.D. Bickham core with Area 1 and the Cranfield cores to Area 2.

an opportunity to compare log-based morphometric calculations to those of core derived cross set data in a core from the T.D. Bickham well. Although the core was located only 5 km to the north of the Lower Cretaceous shelf edge and over 40 km south of the wireline logs of Area 1 these data occupy the same incised valley-shelf edge delta system with wireline well logs located in more proximal regions of the valley and core (T.D. Bickham) located in more distal portions of the valley (Figure 4.3). Morphometric estimates from measuring cross sets in the T.D. Bickham core did not meet the initial test of the Bridge (2003) method (i.e. $s_d/s_m \approx 0.88\ (\pm 0.3)$). This was likely due to the limited number of fluvial dune cross sets identified in the core (n-12). Despite this limitation, the morphometric estimates from those cross sets that were available compare closely to those morphometrics derived from the wireline log data collected in more proximal areas of the valley. Although fluvial channels diminish in number as one moved basinward they appear to maintain their size and flow capacity. Cross set measurements from the T.D. Bickham core estimate an average channel depth of 5.92 m and an average channel width of 306 m. Estimates from wireline logs and from cross set measurements in the T.D. Bickham provide independent corroboration of the two methodologies, with each approach showing similarity in calculated channel dimensions. The estimated channel belt widths from the T.D. Bickham core range from 2043 m to 5674 m which is approximately twice the width estimated from the Area 1 wireline logs to the north. Such widening of the channel belt is expected as the incised valley widens as it approaches the shelf edge and the channel belt becomes less confined.

Although Area 2 is located in an overbank or interfluve area it does contain fluvial deposits from smaller channels typical of those that occupied overbank areas and floodplains between the major valley systems (e.g. O'Byrne and Flint, 1996; McCarthy and Plint, 1998). This area, like Area 1, offers both core and wireline log data for comparative analysis. The sands in this area, still considered part of the Massive sand interval, range from less than 24 m to 35 m (Figure 4.3). Channel morphometrics estimated from wireline logs in Area 2 indicate an average channel depth of 3.79 m and an average channel width of 167 m. Channel belt widths ranged from 759 m to 1253 m (Table 4.1). Some core from the Cranfield area is also located in Area 2. Channel morphometrics estimated from measuring cross sets indicate an average channel depth of 4.45 m and average channel width of 197 m. Channel belt widths ranged from 1219 to 3832 m (Table 4.1). Comparable to Area 1, there is a similarity of channel dimension estimates between the estimates made from wireline logs and cross set measurements confirming the validity of the two methods.

DISCUSSION

Outcrop Analogs

The data on channel width and depth derived from analysis of the Lower Tuscaloosa Massive sand was plotted against similar data derived by Gibling (2006) (Figures 4.4-4.7). These plots showed that the dimensions of the Lower Tuscaloosa Massive sand channels compared to other “braided and low sinuosity”, “meandering”, and “valley fills within alluvial and marine strata” systems as defined by Gibling (2006) (Figure 4.4) . The overlap of these three different types of fluvial environments (Gibling, 2006) in the Lower Tuscaloosa lends support to the interpretation of these systems as incised valleys filled with transitioning braided to meandering systems (Chapters 2 and 3). These valleys contain a variety of channel types as the systems transition temporally from lowstand braided to late lowstand meandering (Chasteen 1983; Klicman et al. 1988). Alternatively, the classification of these channels as either braided or meandering end members may be over simplified. Miall (1996) cites many examples of fluvial types

Table 4.1. Morphometric estimates from wireline logs and cross sets measured from core for the Lower Tuscaloosa Formation. Average channel depth (d_m), width (w), and maximum and minimum channel belt width (cbw_{max} and cbw_{min}) are shown.

	Wireline Well Logs				Cranfield Core	TD Bickham Core
	Area 1	Area 2	Area 3	All Areas		
d_m (eq. 1.4)	5.62 m	3.79 m	7.77 m	5.87 m	4.45 m	5.92 m
w (eq. 1.5)	306 m	167 m	503 m	340 m	197 m	306
cbw_{max} (eq. 1.6)	2148 m	1253 m	3346 m	2416 m	3832 m	5674
cbw_{min} (eq. 1.7)	1533 m	759 m	2748 m	1853 m	1219 m	2043

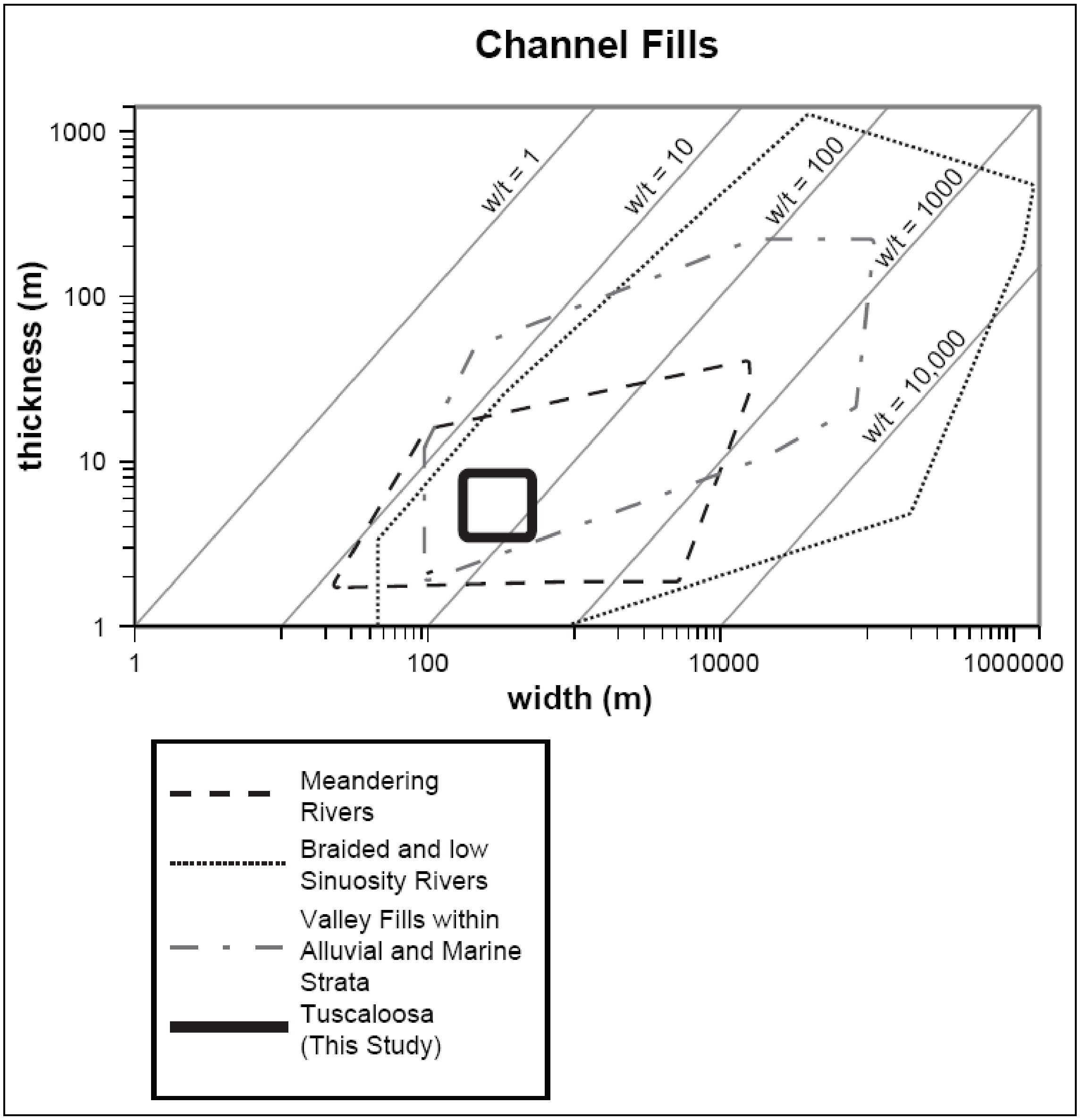

Figure 4.4. Channel fill graph modified after Gibling (2006). Graph contains fields that summarize channel depth and width estimates for “meandering”, “braided and low sinuosity”, and “valley fills within alluvial and marine strata” and how the Lower Tuscaloosa Formation compares to those fields.

that fall between the strictly meandering and braided end members. Core-based sedimentologic analyses show that the Lower Tuscaloosa Massive sand deposits are often rich in gravel and very coarse sands (Chapter 3), however, the assumption of braided patterns may have been too confining. These channels may have had a meandering pattern similar to the "gravel wandering" or "gravel meandering" fluvial style of Miall (1996). Although Lower Tuscaloosa fluvial facies channel deposits are sand-rich and often contain large components of gravel and pebbles (often imbricated and found in lags at the base of sand bodies), they similarly contain fining upward sand bodies bounded by erosional surfaces and sand bodies often transition up to a red mottled siltstone (Chapter 3), suggestive of a more meandering nature.

Valley Fill Analogs

Outcrop analogs for the Lower Tuscaloosa Formation identified from the work of Gibling (2006) for "valley fills associated with underlying alluvial and marine strata" appear to include the Dinosaur Park and Horseshoe Canyon Formations (both upper Cretaceous) which are located in Alberta Canada (Eberth, 1996; Gibling, 2006) (Figure 4.5). However, the Dinosaur Park and Horseshoe Canyon Formations are potential analogs to the Lower Tuscaloosa Massive sand in size only as they are dominated by muddy, heterolithic fills interpreted to be dominantly estuarine in nature (Eberth, 1996). Depositionally and sedimentologically these deposits appear a more suitable analog for the estuarine Stringer sand of the Lower Tuscaloosa Formation (Chapter 3). McCabe (1977) identified sand-rich fluvial facies inside of paleovalleys cut into the Kinderscoutian delta in the Central Pennine Basin in England. The majority of these channels ranged from 20 m to 40 m deep and 500 m to 1500 m wide and are characterized by medium to very coarse grained sands with scattered pebbles and pebbles

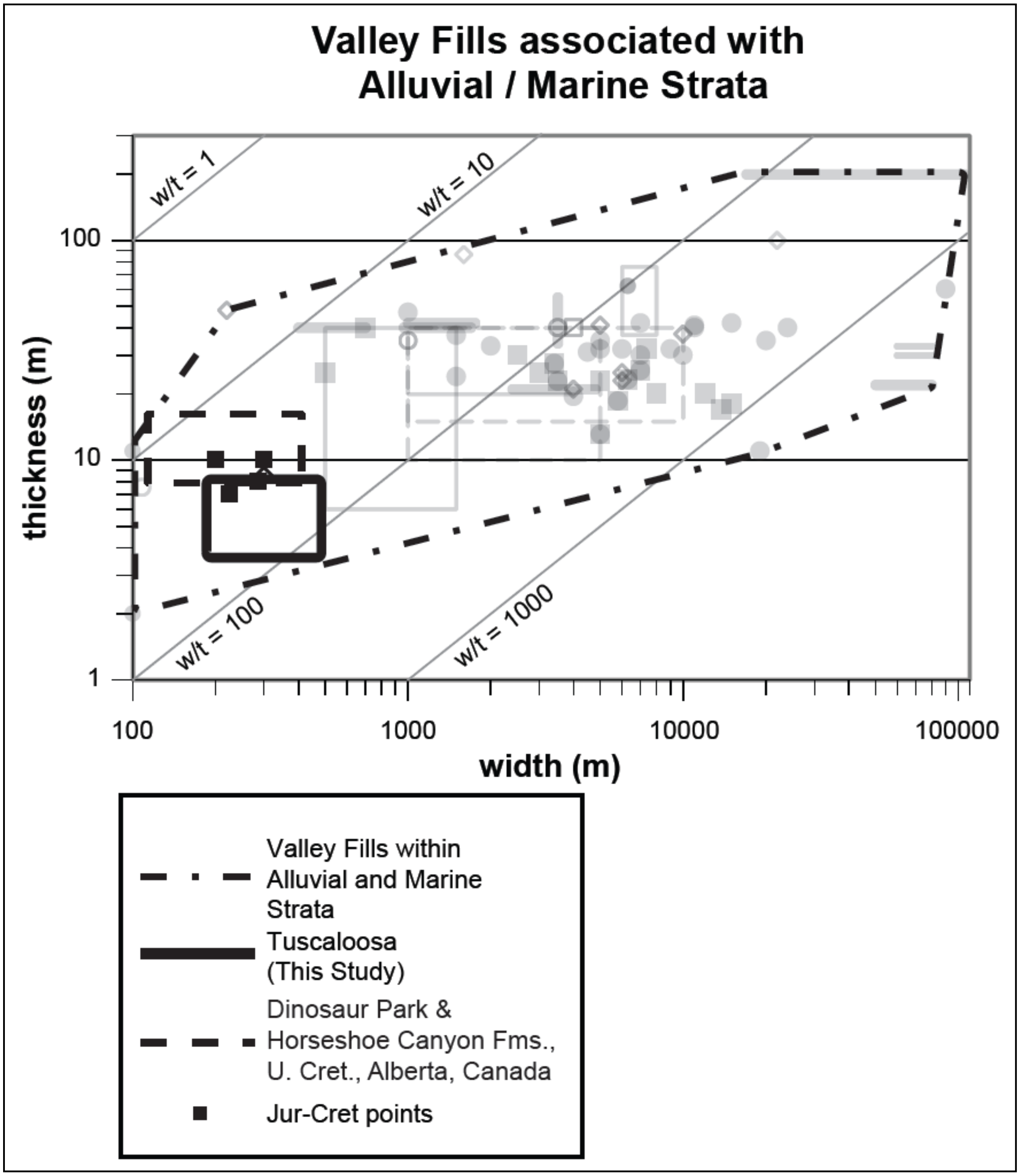

Figure 4.5. Valley fills within alluvial and marine strata graph, modified after Gibling (2006), used compare the Lower Tuscaloosa channel dimensions with ancient analogs. Light gray data points are from the original Gibling (2006) graph but do not compare to the Lower Tuscaloosa Formation.

at the erosional bases of cross sets. These observations led McCabe (1977) to conclude that a “major river”, slightly smaller than the modern Mississippi, fed the Kinderscoutian delta. The lower end of the estimates by McCabe (1977) of the Kinderscoutian channels are similar to the upper end of the Lower Tuscaloosa channel dimensions from this study indicating that the Lower Tuscaloosa was slightly smaller than the Kinderscoutian. McCabe (1977) also identified several smaller channels that averaged 6 m deep and 85 m wide which match closely to the Lower Tuscaloosa channel dimensions.

Braided and low sinuosity analogs

“Braided and low sinuosity fluvial” type outcrop analogs for the Lower Tuscaloosa systems include the Siwalik Group in Pakistan (Willis 1993a,b; Friend et al. 2001), the Escanilla Formation of Spain (Bentham et al. 1993), and the Chuckanut Formation of Washington (Johnson, 1984; Gibling, 2006) (Figure 4.6). The Siwalik Group in Pakistan appears to be an excellent analog for the Lower Tuscaloosa Formation. The channels of the Siwalik Group are underlain by an erosional surface, composed of coarse- to very fine-grained sandstones with a gravel lag at the base, and capped by a paleosol (Willis, 1993a). The morphometric estimates indicate channel depths of 4 to 13 m, a channel widths of 80 to 200 m, and channel belt widths of 1 to 2 km (Willis, 1993a). Willis (1993a) also estimated discharges for the Siwalik Group of 400 to 800 m^3/s. The sedimentology and morphometrics are nearly identical to those in the Lower Tuscaloosa Formation (Figure 4.6) (Chapter 3). Similarly, the Escanilla Formation appears to closely resemble the Lower Tuscaloosa Formation in both morphometrics and sedimentology. Bentham et al. (1993) described the individual sandstone bodies of the Escanilla Formation as erosionally based followed by gravels and pebbly sandstones at the base fining upward into progressively finer sandstones and capped by mottled overbank

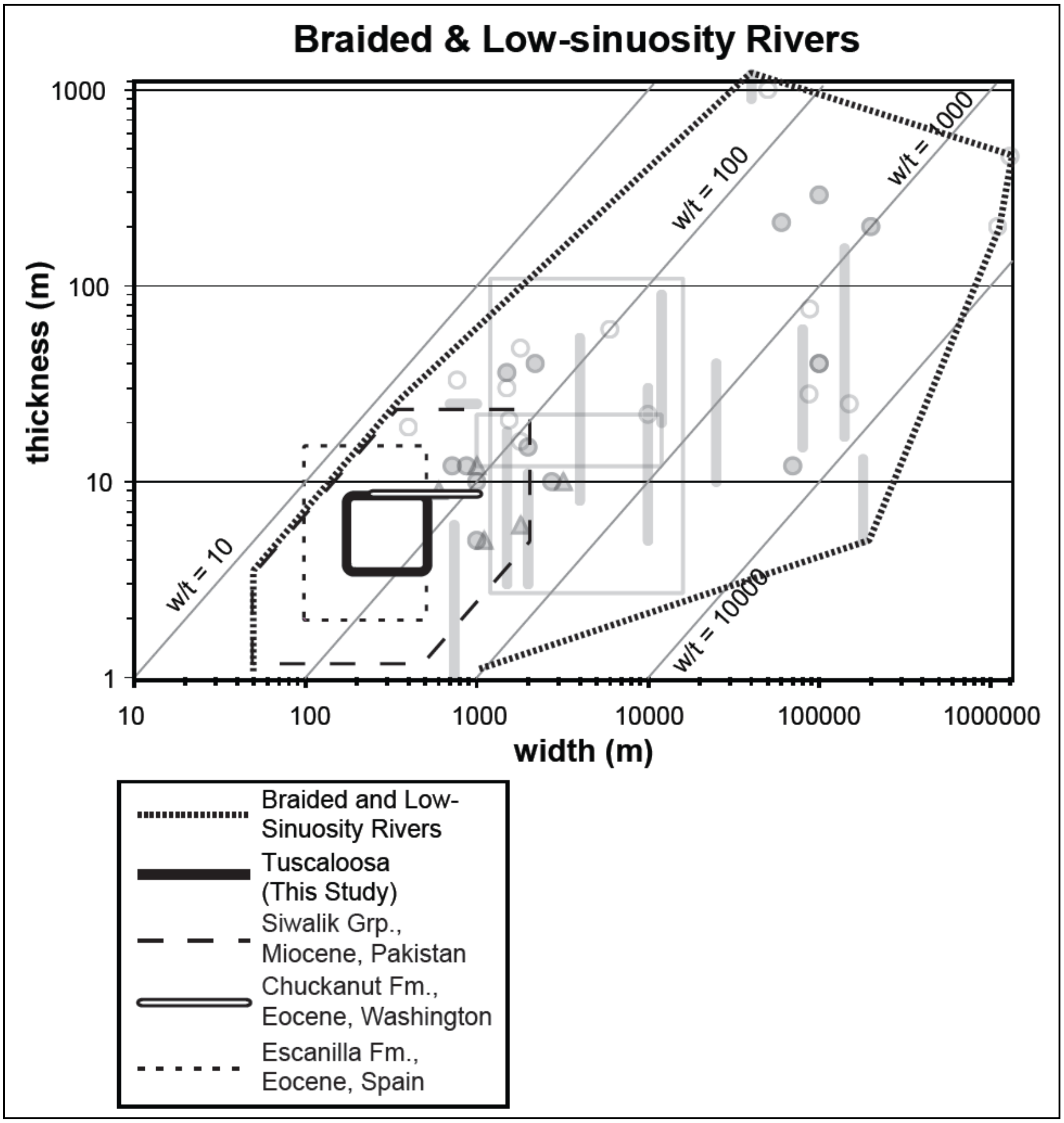

Figure 4.6. Braided and low sinuosity graph, modified after Gibling (2006), used compare the Lower Tuscaloosa channel dimensions with ancient analogs. Light gray data points are from the original Gibling (2006) graph but do not compare to the Lower Tuscaloosa Formation.

siltstones and mudstones. Bentham et al. (1993) also proposed a depositional model which confines the channel belt within fine grained overbank deposits of the flood plain which approximates the Lower Tuscaloosa channels having been deposited within an incised valley cut into alluvium and marine strata. The range of channel depths and widths of the Escanilla Formation closely matches the Lower Tuscaloosa Formation estimates from this study (Figure 4.6). The interpreted low sinuosity Chuckanut Formation (Johnson, 1984) also shows fining upward cycles of minor conglomerates and coarse grained sandstones that alternate with very-fine grained sandstones, siltstones, mudstones, and minor coal. The base of these fining upward cycles is erosional. Morphometric estimates of channel depth and width in the Chuckanut Formation closely resemble the Lower Tuscaloosa, Siwalik, and Escanilla Formations (Figure 4.6) (Gibling, 2006).

Meandering Analog

A good outcrop analog for meandering architectures in the Lower Tuscaloosa is the Canyon Creek Member of the Ericson Formation (Martinsen et al., 1999; Gibling, 2006) (Figure 4.7). The Canyon Creek Member has been interpreted as a sand-rich meandering channel which contains several fining upward cycles with erosional surfaces at their base showing pebbly to gravely lags. These units amalgamate to form several multistory channel bodies (Martinsen et al. 1999). The Canyon Creek Member sedimentologically also closely resemble the Lower Tuscaloosa Formation (Martinsen et al. 1999). The Canyon Creek illustrates a fluvial system that possibly lies between the exact definitions of the meandering and braided fluvial system end members (Miall, 1996). The Canyon Creek is interpreted as being meandering in nature, yet is sedimentologically sand-rich (Martinsen et al. 1999). Additionally, Martinsen et al.

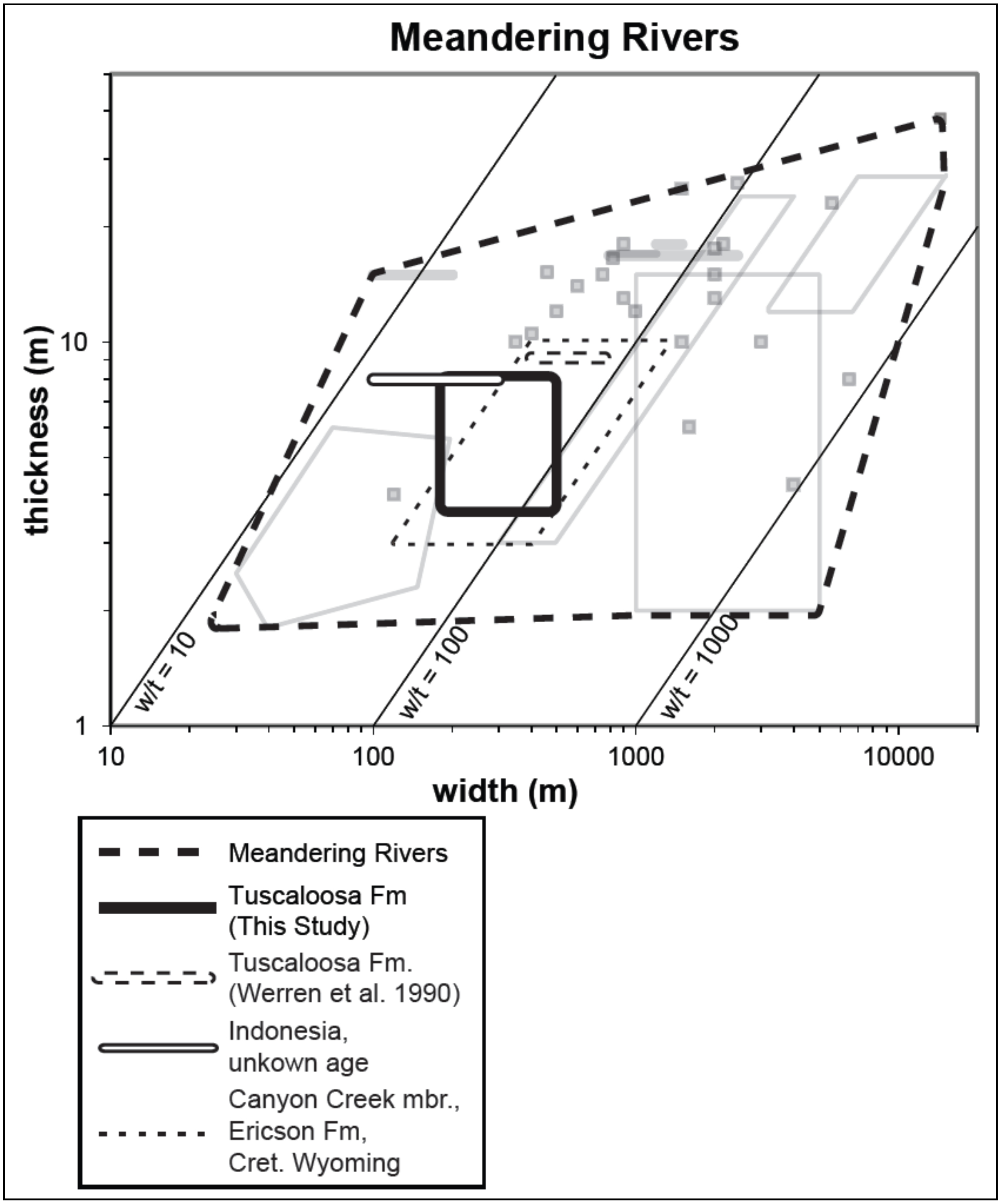

Figure 4.7. Meandering river graph, modified after Gibling (2006), used compare the Lower Tuscaloosa channel dimensions with ancient analogs. Light gray data points are from the original Gibling (2006) graph but do not compare to the Lower Tuscaloosa Formation.

(1999) interpreted the Canyon Creek to have a basal nested or amalgamated channel facies association which transitions to a meandering channel facies association. This transition appears analogous to the braided to meandering transition in the Lower Tuscaloosa Formation noted by this author and several others (Chasteen, 1983; Klicman et al., 1988). Morphometric estimates of channel depths and widths in the Canyon Creek member also fall within the range of morphometrics measures for the Lower Tuscaloosa channel systems (Figure 4.7). Of interesting note in Figure 4.7 another study of the Tuscaloosa Formation (Werren et al., 1990) plots just outside the estimates of this study with slightly larger channel depths and widths.

Modern Analogs

Modern rivers with similar channel depths and widths to the Lower Tuscaloosa Formation include the Missouri, Ohio, and Alabama Rivers (Table 4.2). The Ohio, Alabama, and the downdip portion of the Missouri Rivers are meandering systems, while the upper portion of the Missouri River is a braided system that displays a meandering pattern.

The Missouri River

The Missouri river has an average channel depth of 6.2 m, an average channel width of 351 m and a mean annual discharge of 1253 m^3/s (Figure, 4.8; Table 4.2) (Kiel, personal communication). Prior to major human influences on the Missouri River (e.g. dams, channelization, and irrigation) the Missouri-Mississippi River system, of which the Missouri River is the main sediment contributor, transported nearly 400 million metric tons per year of sediment to the Gulf of Mexico (Meade and Moody, 2010). Much of the sediment drained by the Missouri river is suspended load which has led to the nickname “Big Muddy” (Meade, 1995; USGS.gov). The Missouri river still contributes nearly half

Table 4.2. Comparison of average channel depth and width of the Lower Tuscaloosa Formation with potential modern day rivers.

	Average Channel Depth (m) (d_m)	Average Channel Width (m) (w)
Lower Tuscaloosa Fm.	3.8 – 7.8	145 - 721
Missouri River	6.2	351
Ohio River	7.2	556
Alabama River	5.0	191

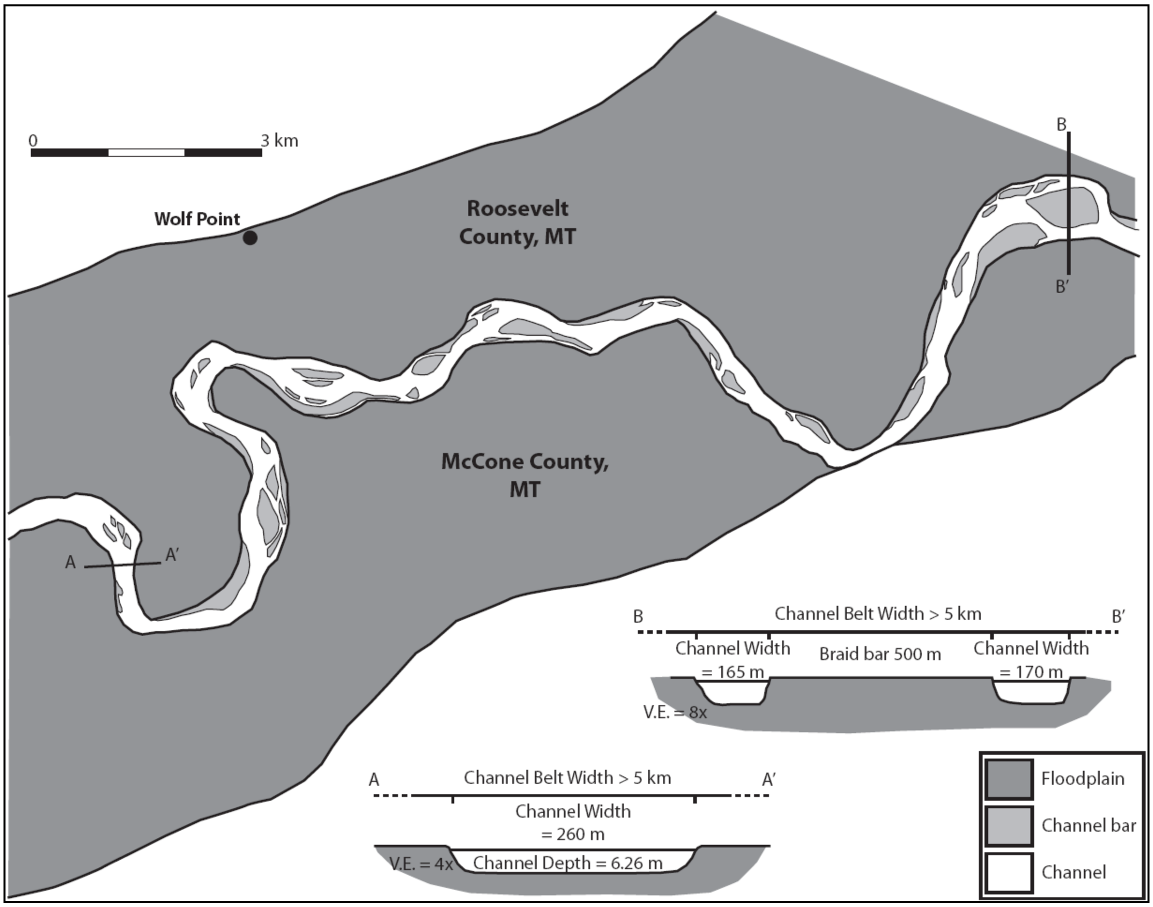

Figure 4.8. Outline and channel cross sections of the Missouri River in Roosevelt and McCone Counties, Montana. Note the meandering nature of the channel coupled with mid-channel bars.

of the sediment delivered to the Gulf of Mexico and drains approximately 1/6 of the United States (Meade, 1995) indicating that a river this size is capable of delivering a significant amount of sediment to its depositional basin. The USGS indicates that before major human interaction with the river the large amount of sediment transported by the Missouri allowed for braided channels to form in the meandering river similar to what is proposed for the Lower Tuscaloosa by this author (Chapter 3). Subsequent modifications to the lower Missouri River have eliminated braid bars to open the river for navigation (USGS.gov); however the upper Missouri in Montana still contains braided channels (Figure 4.8).

The Ohio River

Like the Missouri River the Ohio is one of the major tributaries of the Mississippi River and one of the major drainages of the United States, but there are several differences in the character of the two rivers. The Ohio River has an average channel depth of 7.2 m and average channel width of 556 m (Kiel, personal communication) which is similar to the Missouri and to the Lower Tuscaloosa estimates from this study (Figure 4.9; Table 4.2). In contrast to the Missouri River the Ohio discharges nearly three times the amount of water at an average of 3206 m^3/s and contributes nearly half of the water to the Mississippi River (Meade and Moody, 2010). Despite this increase in discharge the Ohio River contributes relatively minor amounts of sediment to the Mississippi River drainage into the Gulf of Mexico (Meade and Moody, 2010). The Ohio River is entirely a meandering fluvial system.

The Alabama River

The Alabama River runs just north of Montgomery, Alabama to its confluence with the Tombigbee River just north of Mobile Bay. The Alabama River appears to be the most sinuous of the three modern analogs discussed in this study and has well developed pointbars (Figure 4.10). Average channel depth in the Alabama River is 5.0 m and average channel width in 191 m well within the range of the Lower Tuscaloosa channel dimensions (Figure 4.10; Table 4.2). Mean annual discharge for the Alabama has been measured at 950 m^3/s.

Modern Analogs Summary

The ability of these modern rivers to transport large amounts of sediment to the Gulf of Mexico combined with the architectural elements of both a meandering and braided river systems indicate the Lower Tuscaloosa fluvial systems may have had similar geometries and abilities to transport sediment to the basin. The Mississippi River, although bigger than the Lower Tuscaloosa fluvial systems, deposits a well-developed fan into the Gulf of Mexico (Bouma et al., 1983). This fan is the result of strong influences from the Missouri and Ohio Rivers, both possible modern analogs for the Lower Tuscaloosa. These strong influences are high sediment input from the Missouri River and high water input from the Ohio River. The resulting Mississippi fan has a low sand:clay ratio, however, Bouma et al. (1983) noted that much of the sand in the system may be transported to deeper water. The Lower Tuscaloosa Formation was coarser-grained than the modern Mississippi fan perhaps because the Lower Tuscaloosa sediment was less mature being closer to its main source of sediment. Nonetheless, the example of the Mississippi fan indicates the Lower Tuscaloosa fluvial systems could transport large amounts of sediment to the basin. The Lower Tuscaloosa's proximity to its source is more analogous to the Missouri River in Montana. This indicates the Missouri River (i.e. gravel meandering or gravel-sand meandering after Miall (1996)) is perhaps the best architectural analog for the Lower Tuscaloosa (Figure 4.8). Although, elements of the

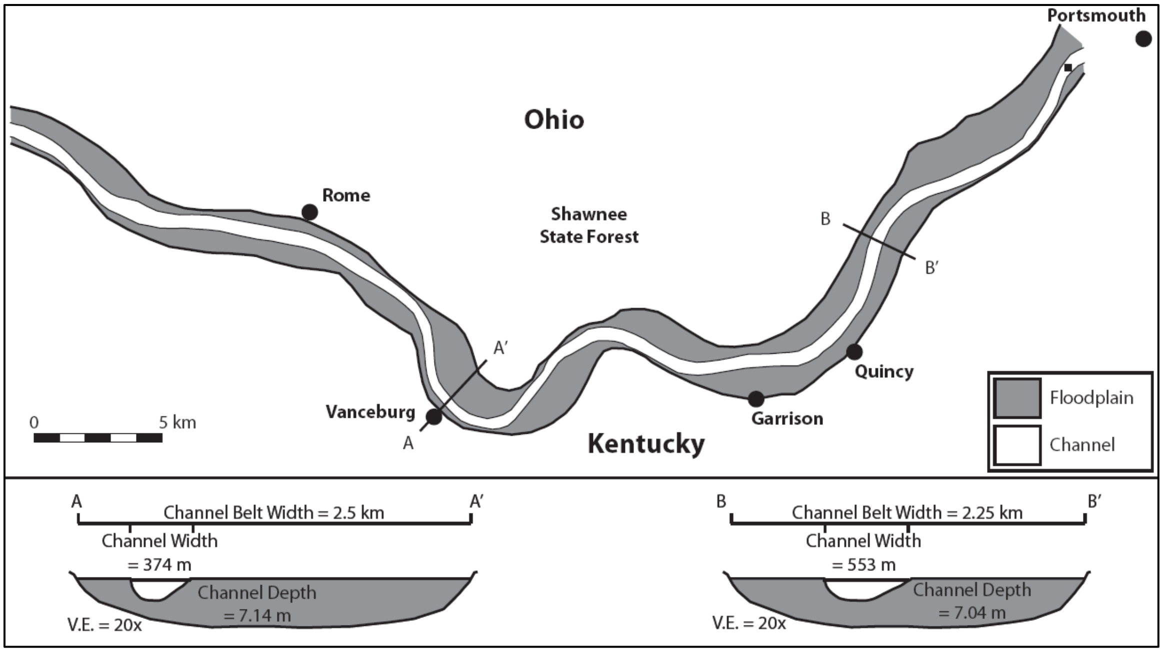

Figure 4.9. Outline and channel cross sections of the Ohio River on the border between Ohio and Kentucky just East of Portsmouth, Ohio. Note the meandering nature of the channel.

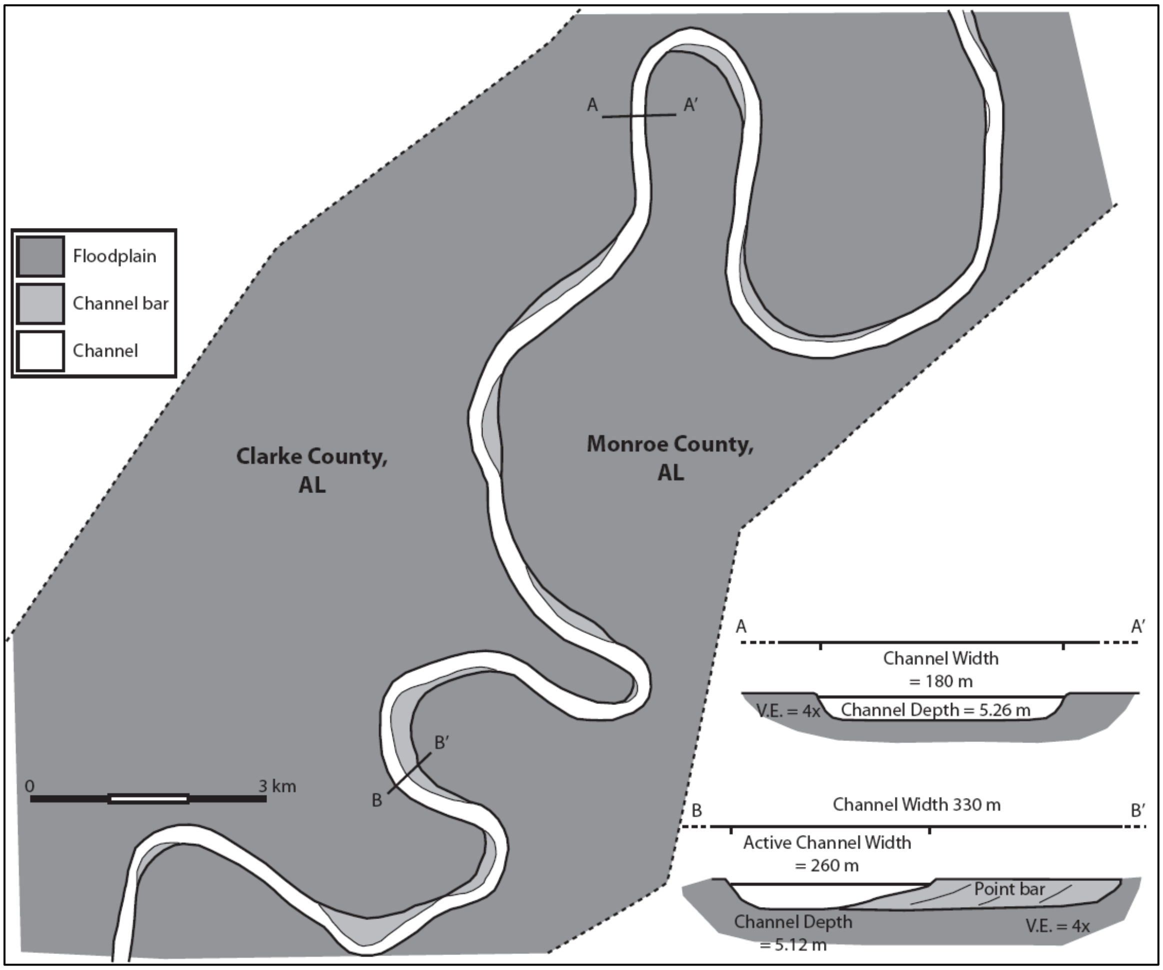

Figure 4.10. Outline and channel cross sections of the Alabama River in Clarke and Monroe Counties, Alabama. Note the meandering nature of the channel and point bar development.

Lower Tuscaloosa fluvial systems do not rule out a meandering system (e.g. Ohio and Alabama Rivers) (Figures 4.9, 4.10).

Source for Basin Floor Deposits

Morphometric calculations and comparison to both ancient and modern day analogs supports the conclusion that the Lower Tuscaloosa Formation fluvial systems were capable of transporting large amounts of sediment to the late Cretaceous shelf edge, for subsequent resedimentation to the deep offshore basin floor. Recent exploration activity in the Gulf of Mexico has discovered thick sequences of Tuscaloosa-aged clastic deposits over 400 km sourced from the time-equivalent shelf edge. The Davy Jones #2, Tiber, and BAHA #2 wells have each penetrated thick Tuscaloosa-aged, sand-rich deepwater, gravity deposits (Chapters 2 and 3). The few well penetrations into the offshore Tuscaloosa interval are significant in thickness and in the distance sediments must have travelled from the shelf edge. These three wells, although the only three to currently have penetrated this interval, provide an initial sampling of the true extent of these deposits. The offshore Tuscaloosa has the potential to be as extensive as the overlying Wilcox Formation which blankets the Gulf Basin. A substantial amount of work and continued exploration effort needs to be completed before a true understanding of the extent and nature of these deposits is reached.

CONCLUSIONS

Morphometric estimates of the size and capacity of the Lower Tuscaloosa fluvial systems has led to a better understanding of the nature and sediment generation capacity of onshore fluvial systems, which have been discovered recently to have been the feeder systems for a major late Cretaceous-age, deepwater clastic system in the Gulf of Mexico.

Independent estimates of channel depth, channel width, and channel belt width made from wireline logs and from dune architecture cross set measurements from core (Bridge, 2003) show complimentary results confirming the validity of using either method to estimate paleo-channel dimensions.

Channel depth and width estimates in the Lower Tuscaloosa reveal the nature of the Lower Tuscaloosa Formation fluvial systems as incised valleys cut into alluvium and marine strata and filled with braided to low sinuosity and meandering rivers. This variety of fluvial systems in the Lower Tuscaloosa highlights its dynamic nature supporting the interpretation of incised valleys and the transitional nature of the Massive sand from braided to meandering channels within transitioning lowstand to transgressive valleys.

Outcrop analogs for the Lower Tuscaloosa Formation include the braided and low sinuosity Siwalik Group in Pakistan, the Escanilla Formation of Spain, and the Chuckanut Formation of Washington. Likewise, the Canyon Creek Member of the Ericson Formation appears to be a good outcrop analog for the meandering channelized components of the Lower Tuscaloosa. Data suggest appropriate modern analogs for the Lower Tuscaloosa to be the Missouri, Ohio, and Alabama Rivers.

The Lower Tuscaloosa Formation fluvial systems were capable of transporting large amounts of sediment to the shelf edge which were later transported into the deep offshore basin floor. These late Cretaceous feeder systems supplied sediment over 400 km from the paleo shelf edge to the basin floor as thick, sandy basin floor fans which have been recently penetrated by the BAHA#2, Tiber and Davy Jones #2 wells.

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Chapter 5: Regional Distribution of Chlorite in the Lower Tuscaloosa Formation: Insights into the Controls of Chlorite Occurrence and Implications to Reservoir Quality.

INTRODUCTION

For over three decades the presence of chlorite in the Lower Tuscaloosa Formation has been a major focus of study. Much of this focus stems from the reported anomalously high porosity and permeability at depths below 20,000 ft. in the downdip producing trend of the Lower Tuscaloosa Formation (Thompson, 1979). This anomalously high porosity and permeability, thought to be greatly influenced by the presence of thin chlorite coats on detrital grains, has resulted in favorable reservoir quality in highly productive gas reservoirs at these great depths. Despite the interest in the presence of chlorite in these reservoirs there still exist many points of disagreement and confusion regarding the source of chlorite, mechanisms of formation, and role that chlorite plays in the preservation of porosity and permeability (See Dowey et al., 2012 for review).

One of the major questions that still exists in the Lower Tuscaloosa Formation is the degree to which provenance controls the abundance of chlorite. The recent discovery of Tuscaloosa-aged deepwater basin floor deposits in the deep offshore Gulf of Mexico has served to enhance the urgency to develop understanding of the origin of chlorite and how its distribution is affected by both chemical and mechanical processes of source-to-sink sediment movement. It has been noted by other authors that many believe the chlorite in the Lower Tuscaloosa to be provenance controlled (Dowey et al., 2012; Bloch, 2002; Ehrenberg, 1993; Thomson, 1979; Geniuse, 1991). Despite claims of the Lower Tuscaloosa being derived from the relatively volcanic-rich Ouachita Mountains (Thompson, 1979), regional mapping of Lower Tuscaloosa sequences (Chapter 2) shows that much of the sediment was sourced from the Appalachian Mountains to the east. The Appalachians have relatively fewer volcanic rocks (Mann and Thomas, 1968; Birdwell and Hill, 1997) to provide volcanic rock fragments (VRFs) thought to be the source of chlorite development. A west to east decrease of VRFs in the Lower Tuscaloosa Formation has been speculated on by other scientists (Hearne and Lock, 1985; Hansley, 1996), however such a hypothesis has not been presented through examination of the data.

Alternative to a provenance argument, the presence of chlorite may be controlled by the depositional environment or detrital grain size. It was decided to utilize the detailed study outlining the facies and depositional environments from Lower Tuscaloosa core data (Chapter 3) to assess any correlation of depositional environment or detrital grain size to chlorite abundance.

The goal of this chapter is to determine the distribution of chlorite in Lower Tuscaloosa sandstones across the region to test Thomson's (1979) assertion of chlorite being controlled by provenance. Additionally, with the framework of facies and depositional environments established in Chapter 3 correlations between depositional systems and grain size and chlorite are assessed. A paragenetic sequence for the Lower Tuscaloosa formation is proposed. The reservoir quality of Tuscaloosa deepwater deposits is also discussed. Data from 13 core (See Figure 5.1 for locations), 115 thin sections of sandstones from those core, publically available data of the deep offshore Tiber well, and previously published data were used to accomplish these goals.

PREVIOUS WORK

Thomson (1979) first proposed an explanation for the origin of anomalously high porosity and permeability in reservoirs of the Lower Tuscaloosa Formation at

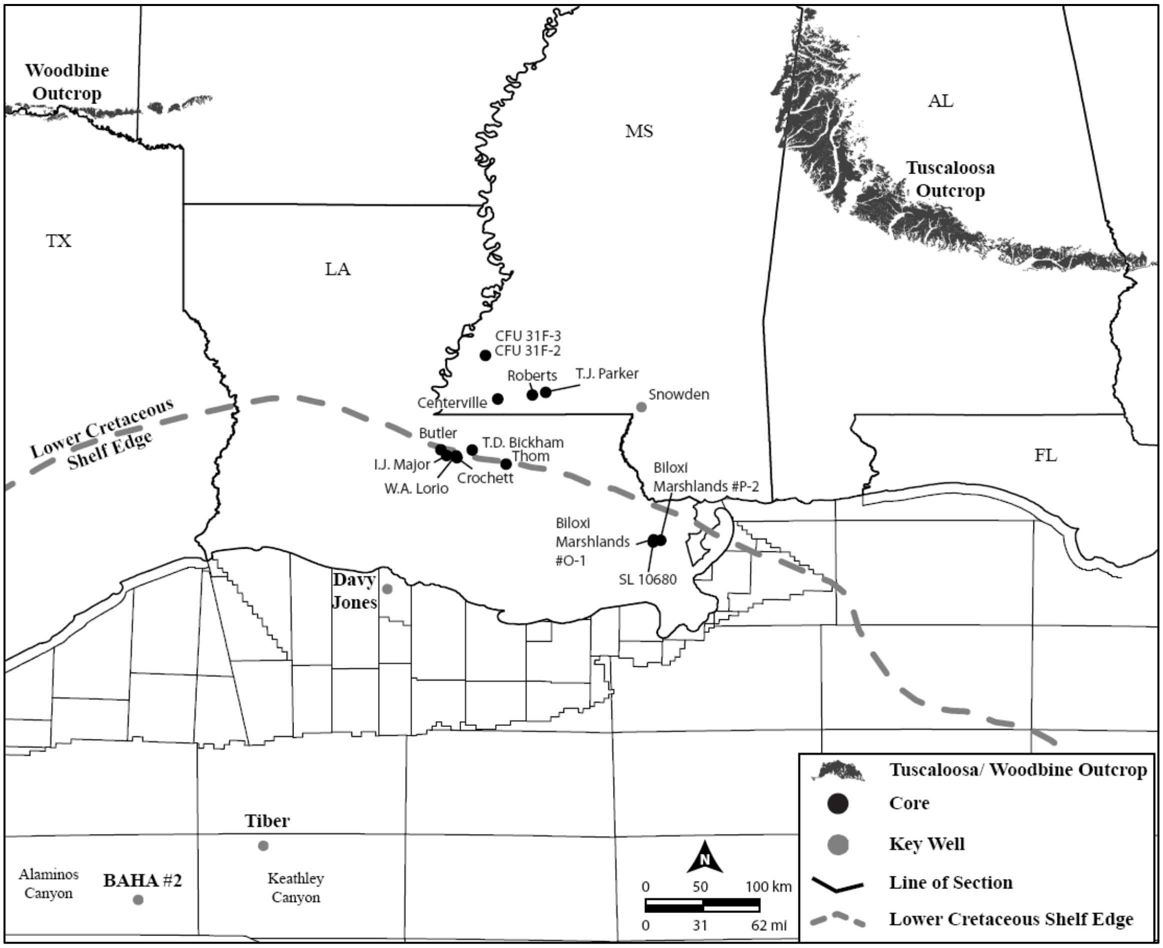

Figure 5.1. Regional map of the study area showing the locations of the 13 core and key offshore wells (Davy Jones #2, BAHA #2, and Tiber) discussed in this paper. Also shown are the locations of the Tuscaloosa and Woodbine outcrops and the Lower Cretaceous shelf edge.

stratigraphic depths below 20,000 ft. These observations included; (1) chlorite riming the detrital grains in the sandstone help to inhibit quartz cementation by blocking nucleation sites for quart cement thus preserving primary porosity; (2) the suggestion of an origin for the iron (Fe) and other ions needed for the formation of chlorite coats being from the dissolution of volcanic rock fragments (VRFs); and (3) the author further suggested that the VRFs in the Lower Tuscaloosa Formation were derived from the Ouachita Mountain belt.

Since the publication of Thomson (1979) his observations and conclusions have become the standard reference for much of the research focused on the reservoir quality of Lower Tuscaloosa Formation across the region (Smith, 1981; Thomson, 1982; Dahl, 1984; Larese et al., 1984; Hearne and Lock, 1985; Stancliffe and Adams, 1986; Wiygul and Young, 1987; Minter et al., 1992; Hansley, 1996; Ryan and Reynolds, 1996, 1997). Subsequent authors although in agreement regarding the source of iron for chlorite nucleation, have inferred that VRFs in the Lower Tuscaloosa are less abundant in reservoirs in the more eastern portion of the Lower Tuscaloosa trend. This observation supports a dual origin for Lower Tuscaloosa sands with some being derived from the volcanic-poor Appalachian Mountains (Hansley, 1996; Hearne, 1985). Since the publication of Thompson (1979) similar observations regarding the ability of chlorite coats to inhibit quartz cementation have been observed in other reservoirs around the world (Dowey et al., 2012; Billault, 2003; Bloch, 2002; Ehrenberg, 1993; Geniuse, 1991).

With broader observation, other researchers generally agree that the source of chlorite coats in Lower Tuscaloosa reservoirs stems from the dissolution of VRFs, however, the formation of chlorite is interpreted by these authors as a dominantly late diagenetic process (Watkins, 1985; Hamilton and Cameron, 1986; Hamlin and Cameron, 1987; Klicman et al., 1988; Geniuse, 1991; Cameron et al., 1992; Weedman et al., 1992, 1996; Corcoran et al., 1993). These researchers attribute the higher porosity values to early calcite cementation when primary porosity was high followed by dissolution of that cement after deep burial. Finally, the question remains regarding the source of Fe needed for the formation of Fe-rich chlorite in the Lower Tuscaloosa Formation. This Fe may have originated from Fe-rich fluids from the dewatering of surrounding shales during compaction and/or other Fe-rich constituents which are commonly observed in the Lower Tuscaloosa Formation such as siderite, ankerite, and Fe-Ti oxide cements (Hearne and Lock, 1985; Hansley, 1996).

METHODS

Data from over 1200 ft. of cored intervals from 13 wells (See Figure 5.1 for locations) and 115 thin sections from these cores were used as the basis for this study. Thin sections were impregnated with blue-dyed epoxy and sampled regionally from traditionally hydrocarbon productive locations in Mississippi and Louisiana over several different depths, depositional environments, and grain sizes allowing for the assessment of how these factors affect the abundance of chlorite. All 115 thin sections were examined by optical microscopy for identification of the diagenetic sequence and point counted a minimum of 300 points per slide. The diagenetic sequence was determined from observations of crosscutting relationships, dissolution of framework grains and cements, and growth of authigenic cements. Based on these point counts each of the samples was classified on the basis of the relative percentage of their framework grains (Folk, 1980) and on the basis of provenance governed by plate tectonics (Dickinson and Suczek, 1979). Ternary plots for these classification schemes were generated using Excel spreadsheets and graphs (Zahid and Barbeau, 2011). Trends in the whole rock percentages of porosity, chlorite coats, quartz cement, volcanic rock fragments, and porosity were assessed to determine the control of provenance on these elements important to reservoir quality. Porosity values for the onshore wells cited in this study were derived from point counting. The samples were arranged by well from west to east and the key elements were averaged for each well with an error of 1 standard deviation. Grain sizes were determined by measuring a minimum of 100 random grains along the long axis of the grain and then taking the median grain size as the grain size of the sample. Intergranular volume (IGV), the sum of intergranular porosity, cement, and detrital matrix, for each of the onshore samples was used as a quantitative measure to determine the degree of compaction and timing of cementation.

Photomicrographs of samples from the Tiber well in the offshore Gulf of Mexico (see Figure 5.1 for location) were qualitatively observed for petrographic-scale crosscutting relationships, dissolution of framework grains and cements, and growth of authigenic cements. Quantitative measurements of grain size were also taken as discussed above. Measurements of porosity and permeability were assessed to initially determine the reservoir quality of this deep offshore play area.

SANDSTONE COMPOSITION

The sandstones studied in this paper are dominantly sublitharenite (n=100) with minor occurrences of quartz arenite (n=3), litharenite (n=6), and subarkose (n=6) (Figure 5.2A). Sublitharenites were more specifically classified as sedarenite (n=46), phyllarenite (n=31), and volcanic arenites (n=23) (Figure 5.2B) (Folk, 1980). Additionally, 18 of the sublitharenites contained between 10% and 50% muddy detrital matrix and are more accurately classified as muddy sandstones (Folk, 1980). Samples range from coarse (n=1) to very fine (N=15) sandstones but are dominated by medium (n=59) to fine grained

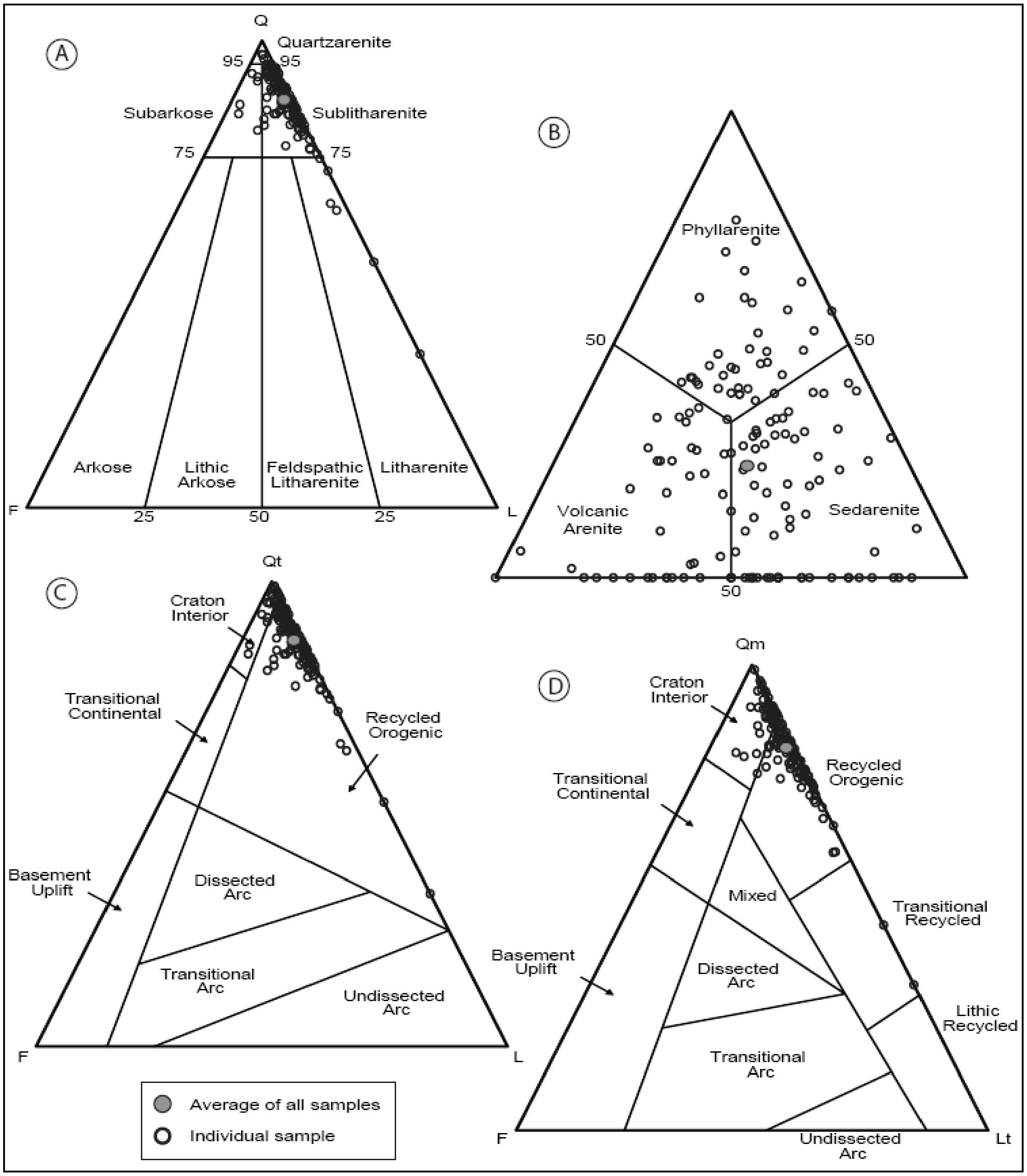

Figure 5.2. Ternary diagrams classifying sandstones in the Lower Tuscaloosa Formation. (A and B) Samples classified on the basis of the relative percentage of their framework grains (Folk, 1980) and (C and D) on the basis of provenance governed by plate tectonics (Dickinson and Suczek, 1979). Open symbols represent an individual sample. Q = quartz, Qm = monocrystalline quartz, Qt = total quartz, F = feldspar, L = lithic fragments, Lt = total lithic fragments.

(n=41) sandstones. Tectonic provenance plots (Dickinson and Suczek, 1979) indicate Lower Tuscaloosa sediment was dominantly derived from a recycled orogenic source with a component having been derived from the craton interior (Figure 5.2C, D) as would be expected of Lower Tuscaloosa sediments originating from the Ouachita and Appalachian Mountains. Intergranular volume of the samples studied ranged from 6% to 57% (avg. 32% ± 8%). Sandstone compositions determined from point counting and averaged by well and by rock type in Tables 5.1 and 5.2 respectively. Complete whole rock percentages for each sample are available in Appendix III.

Detrital Grains

Quartz

Quartz displaying normal to slightly undulatory extinction is the dominant grain type in the Lower Tuscaloosa sandstones (Figure 5.3). Small percentages of polycrystalline quartz with varying degrees of undulatory extinction also exist. In many of the samples a few quartz grains are fractured due to compaction. Pressure solution between quartz grains is observed as a rare occurrence. Some of the quartz grains have an obvious rim of quartz overgrowth that appear to have been somewhat rounded. Such rounding of these overgrowths indicates a reworked sedimentary source.

Feldspar

Feldspar grains compose a small percentage of the rock volume in the Lower Tuscaloosa sandstones (Tables 5.1 and 5.2). Both plagioclase and potassium feldspar grains are present and have undergone varying degrees of alteration and dissolution (Figure 5.3A, B). Plagioclase grains display both simple and polysynthetic twinning and are often preferentially altered along twin planes. Many plagioclase grains are replaced by illite and/or carbonate cement. Potassium feldspar is most often untwinned and generally heavily replaced by clays. Microcline is observed in some of the samples and identified by its tartan twinning. In some cases feldspar grains have been totally dissolved leaving behind moldic pores rimmed in chlorite that exhibit an elongate crystal shape similar to other undissolved feldspar grains found in the sample. A few of the eastern most samples have the highest percentage of feldspar with these grains making up to 8 percent of the total rock volume.

Lithic Rock Fragments

Metamorphic (MRFs), volcanic, and sedimentary rock fragments make up an important percentage of the rock volume (up to 20%). MRFs are dominated by polycrystalline quartz grains rich in foliated muscovite (Figure 5.3C). These grains also display undulatory extinction and slight deformation of the muscovite crystals. Small degrees of slate may be present although difficult to distinguish from shale clasts. MRFs do not display alteration from diagenetic affects.

Volcanic rock fragments (VRFs) are commonly heavily altered to completely dissolved which can make their identification difficult (Figure 5.3D, E). Completely dissolved grains are inferred from remnant round chlorite coats similar to what is observed in the dissolution of feldspar (Figure 5.3B, E, L). VRFs are often dark brown to dark green in plane-polarized light and contain small, elongate, plagioclase laths within them. In samples that show a greater degree of compaction VRFs can be compacted around grains forming a pseudomatrix. There is an overall decrease in VRFs from west to east as Lower Tuscaloosa streams increasingly originated from a more Appalachian provenance (Figure 5.4A).

Several different types of sedimentary rock fragments (SRFs) were observed. Clastic SRFs include siltstone and shale fragments. Both occur as dark brown easily

Table 5.1. Whole rock percentages (averaged for each well) of the major components of the sandstones in this study.

Sample	Framework Grains			Detrital Matrix	Cement						Other	Porosity	
	Q	F	L	Clay	Qtz	Chl-orite	Carb-onate	Kao-linite	Side-rite	Opa-que		Prim.	Sec.
I.J. Major	53.3	0.3	10.4	8.8	7.0	6.8	9.5	0.1	0	2.0	1.3	0.7	0
Butler	64.6	0	8.6	0	1.7	18.9	0	0	0	0	0.3	4.8	1.0
W.A. Lorio	56.6	0.7	7.3	0.2	10.7	9.1	9.3	0	0	3.1	0.8	2.0	0.4
Crochet	51.8	1.0	9.5	9.9	10.2	9.5	0.8	0.6	0	1.8	2.1	2.7	0.1
T.D. Bickham	56.5	0.3	15.4	2.5	3.1	12.3	2.2	0	0	0.4	0.2	7.0	0
Cranfield	64.9	0	10.4	0	1.0	19.8	0	0	0	0	0	2.2	1.6
Centerville	49.1	0.6	9.5	9.7	5.7	4.1	15.3	0.4	0	3.3	1.1	0.9	0.3
Thom	52.1	0.7	6.1	9.7	8.6	3.1	16.2	0.1	0	2.1	0.9	0.2	0.2
Roberts	44.0	1.1	6.4	21.9	9.4	2.1	10.7	0.2	0	2.7	0.9	0.8	0
T.J. Parker	49.7	1.2	4.3	4.5	11.5	4.2	4.6	0.7	5.8	1.2	6.3	4.7	1.1
Snowden	62.5	3.5	6.5	0.1	8.6	1.4	0.7	0.9	0	0.5	3.0	12.4	0
BM #P2	57.1	2.9	2.5	4.5	9.8	3.9	6.3	3.5	0.1	0.6	3.4	5.1	0
SL 10860	46.3	3.0	4.1	21.8	6.1	0	14.5	0.3	0	2.9	0.9	0	0

Table 5.2. Whole rock percentages (averaged for each rock type) of the major components of the sandstones in this study.

Sample	Framework Grains			Detrital Matrix	Cement						Other	Porosity	
	Q	F	L	Clay	Qtz	Chl-orite	Carb-onate	Kao-linite	Side-rite	Opa-que		Prim.	Sec.
Sublitharenit	52.1	0.8	6.9	11.4	8.0	6.4	6.8	0.4	0.4	1.8	0.9	3.0	0.3
Quartz	64.0	1.3	1.0	0	9.4	0.5	7.6	1.7	0	0.2	0.3	9.1	0
Subarkose	54.7	5.4	2.4	0.6	12.4	5.5	8.6	1.3	2.0	1.0	1.2	4.8	0
Litharenite	44.4	0.7	29.7	0.3	8.4	6.4	4.5	0.2	0	1.8	0.2	2.2	0

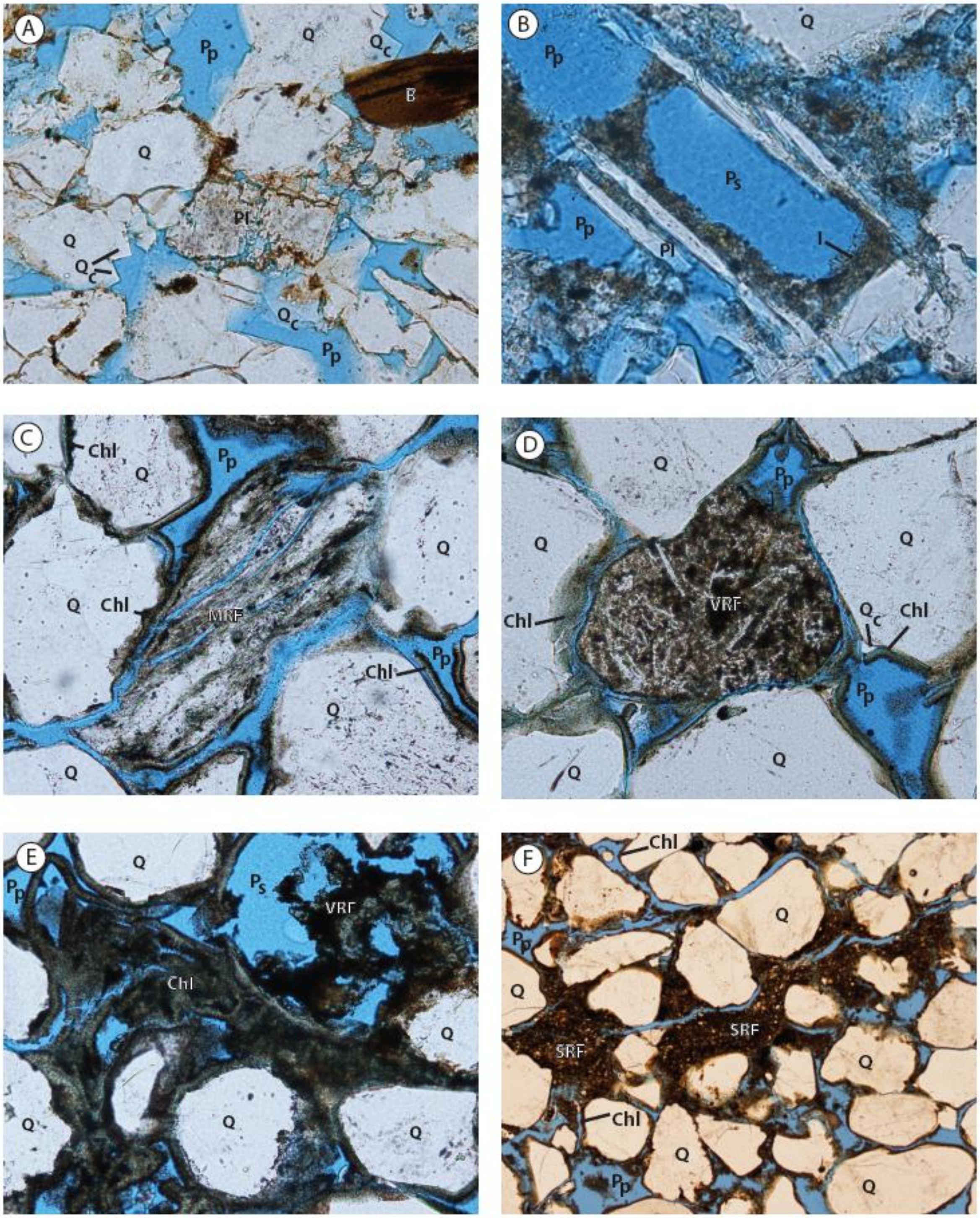

Figure 5.3. See following page for caption.

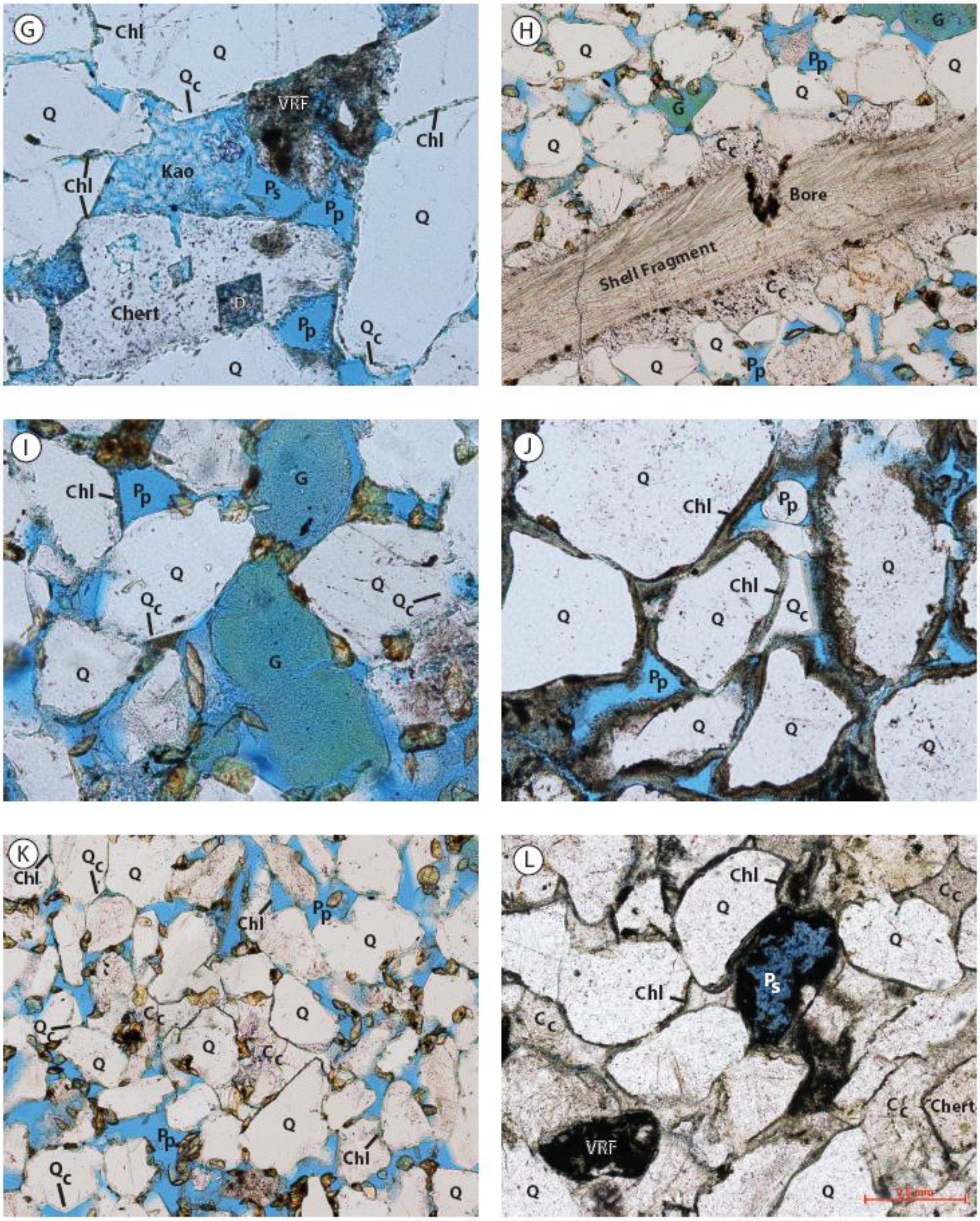

Figure 5.3. See following page for caption.

Figure 5.3. Photomicrographs showing the textural relationships between different components of the Lower Tuscaloosa Formation. (A) Snowden well (8671ft). Sample is absent of chlorite but contains incomplete drusy smectite coats around detrital grains including normal quartz. Relatively abundant quartz overgrowths take up some of the intergranular porosity. Also displayed, a partially dissolved plagioclase grain displaying polysynthetic twinning in cross polarized light and detrital biotite. (B) T.J. Parker well (10396 ft.). Partially replaced plagioclase by illite. Much of the grain has been completely dissolved creating a large intragranular pore. Intergranular porosity also exists. (C) T.D. Bickham well (16216 ft.). Metamorphic rock fragment composed of polycrystalline quartz and foliated muscovite partially altered to clays. Chlorite coats the detrital grains including quartz. (D) T.D. Bickham well (16400 ft.). Relatively well preserved volcanic rock fragment containing plagioclase laths. Detrital grains, including quartz, are rimmed in chlorite. Some radial chlorite also fills the intergranular porosity. A small amount of quartz cement predated the chlorite coats. (E) T.D. Bickham well (16216 ft.). Detrital quartz grains and dissolved volcanic rock fragments rimmed completely by chlorite. Note the chlorite mass of compacted remnant chlorite coats after dissolution of the detrital grains. (F) T.D. Bickham well (16400 ft.). Detrital quartz grains rimmed in chlorite. Sedimentary rock fragments are compacted forming a pseudomatrix and eliminating some of the intergranular porosity. (G) T.J. Parker well (10396 ft.). Dissolved dolomite rhombohedra within chert grain. Quartz overgrowths occur in locations where chlorite coats are incomplete. Heavily dissolved volcanic rock fragment produces some intragranular porosity. Intergranular porosity can be filled with vermicular kaolinite cement. (H) T.J. Parker well (10400 ft.). Large shell fragment containing a bore mark. Glauconite and quartz are the detrital grains. “Wheat seed” siderite is the small light brown to yellow rhombohedral crystals filling intergranular porosity. Calcium carbonate cement nucleated on the shell fragment and grew out from there. (I) T.J. Parker well (10400 ft.). Detrital glauconite grains. “Wheat seed” siderite is the small light brown to yellow rhombohedral crystals filling intergranular porosity. Sample also contains incomplete chlorite coats and quartz overgrowths. (J) T.D. Bickham well (16216 ft.). Detrital quartz grains with well-developed chlorite coats. Quartz overgrowth nucleates from a broken chlorite coat. (K) T.J. Parker (10400 ft.). Poikilotopic carbonate cement. “Wheat seed” siderite is the small light brown to yellow rhombohedral crystals filling intergranular porosity. Sample also contains incomplete chlorite coats and quartz overgrowths. (L) Butler well (18383ft). Intergranular porosity completely cemented with calcite. Intragranular porosity where a detrital rock fragments (likely VRFs) has dissolved. Chlorite coats the detrital quartz grains clearly predating carbonate cement. B = Biotite, Cc = carbonate cement, Chl = chlorite coats, D = Dolomite rhombohedron, G = Glauconite, (I) = illite, Kao = kaolinite, MRF = metamorphic rock fragment, Pl = plagioclase, Pp = intergranular porosity, Ps = intragranular porosity, Q = quartz, Qc = quartz cement, SRF = sedimentary rock fragment, VRF = volcanic rock fragment.

deformed clasts with the siltstone being a little coarser-grained (Figure 5.3F). Compaction can form a pseudomatrix with SRFs. Carbonate SRFs were also identified. Small rounded clasts of calcite containing dolomite rhombohedra were rare occurrences. Chert grains make up a substantial portion of the rock volume in some of the samples.

These grains were identified by their microcrystalline texture and occasional replacement by dolomite rhombohedra (Figure 5.3G). Chert grains are obviously altered and as a result are slightly "dirty" in appearance in plane-polarized light. In some of the samples mollusk shells were also identified. Some of these shells were marked by bores on the outer surface (Figure 5.3H).

Other accessory detrital grains which constitute less than 1 % of the total rock volume include muscovite, glauconite (Figure 5.3H, I), zircon, and organic material.

Cements

Diagenetic processes in the Lower Tuscaloosa, including cementation, can vary widely across the region. Of course, it is common for these variations to occur on a centimeter to meter scale within the same wellbore as well as across a region. Despite this, the Lower Tuscaloosa displays many regionally consistent diagenetic features that have been observed in this study and numerous other studies.

Chlorite Coats

Chlorite is one of the most common cements found in the Lower Tuscaloosa and can comprise up to 21 % of the rock volume averaging 8% (± 6 %) in samples that contain chlorite. 25 of the 115 samples studied were absent of chlorite (Figure 5.3A) with a large amount of these samples located to the east in the Snowden and Biloxi Marshlands wells (see Figure 5.1 for location). In the samples absent of chlorite VRFs were also absent or composed a very small percentage of the rock. Generally, chlorite is present as coatings of individual chlorite crystals which form perpendicular to the detrital grain surface (Figure 5.3). These coatings are commonly observed between detrital grains and in some cases are absent between the contacts of detrital grains. It is common for the detrital grains (often feldspar and volcanic rock fragments) to have completely dissolved leaving behind a chlorite coat and moldic pore (Figure 5.3E and L). In samples that continue to undergo compaction after dissolution these chlorite coats can be compacted together into a thick mass of chlorite destroying the intragranular porosity created by dissolution (Figure 5.3E). Small amounts of radially precipitated chlorite are also found within both the intergranular and intragranular pore space (Figure 5.3D, F, I). This later phase of chlorite has a spikier, less-organized appearance than the earlier more complete grain coats. Similar to volcanic rock fragments, there is an overall decrease in chlorite from west to east towards more Appalachian-derived sediments (Figure 5.4B).

Quartz Cement

Quartz cementation is a ubiquitous process in all of the samples studied. The maximum amount of quartz cement quantified was about 31 % and averaged 8.5 % (± 6 %). Quartz cement occurs as euhedral overgrowths which nucleate from detrital quartz grains and grow into the intergranular pore space. Most commonly these nucleation sites are on samples that contain incomplete or broken chlorite coats (Figure 5.3J). Quartz overgrowths are generally larger and more frequent in samples with incomplete and less frequent chlorite grain coats. A very small amount of quartz overgrowths predate chlorite and are rimmed with chlorite similar to the detrital grains (Figure 5.3D). There is a rough correlation in the increase of quartz cement with a decrease in chlorite coats (Figure 5.4D).

Carbonate Cement

Carbonate cements are common, although very sporadic, and when present often completely fill intergranular porosity (Figure 5.3L). These carbonate cements occur as nodules, streaks, and patches and are often poikilotopic encasing clusters of detrital grains, chlorite and carbonate cements, and “wheat seed” siderite (Figure 5.3K). Some intragranular moldic porosity is seen in areas completely cemented by carbonate except where a few of the detrital grains have been leached (Figure 5.3L). Minter et al. (1992) described in detail these carbonate cements from southwestern Mississippi in the estuarine facies of the Stringer sand and concluded the cement to be composed of ankerite. Most of the samples from the estuarine facies used in this study contained large amounts of carbonate cement, however less frequent and wider spaced carbonate cemented zones were observed in all areas and facies. In samples with mollusk shell fragments carbonate cement nucleated on these shells and grew into the surrounding pore space (Figure 5.3H). “Wheat seed” siderite (Figure 5.3H, I, K) was also observed in a six of the samples from the T.J. Parker well and one sample from the Biloxi Marshlands #P-2 well (see Figure 5.1 for location). This siderite occurs as small yellow to light brown rhombohedra and can be up to 28 % of the rock volume when present.

Accessory Cements

Minor to moderate cementation of kaolinite, smectite, and opaque minerals also occurs. Kaolinite fills pores as vermicular strands and small hexagonal plates and can represent up to 6% of the total rock volume (Figure 5.3G). Smectite occurs in samples were chlorite is absent as grain coats similar to chlorite (Figure 5.3A), however it is brown, generally thinner than chlorite coats, and displays a more chaotic arrangement of individual crystals. Opaque minerals are dominated by pyrite and can represent relatively large poikilotopic concretions.

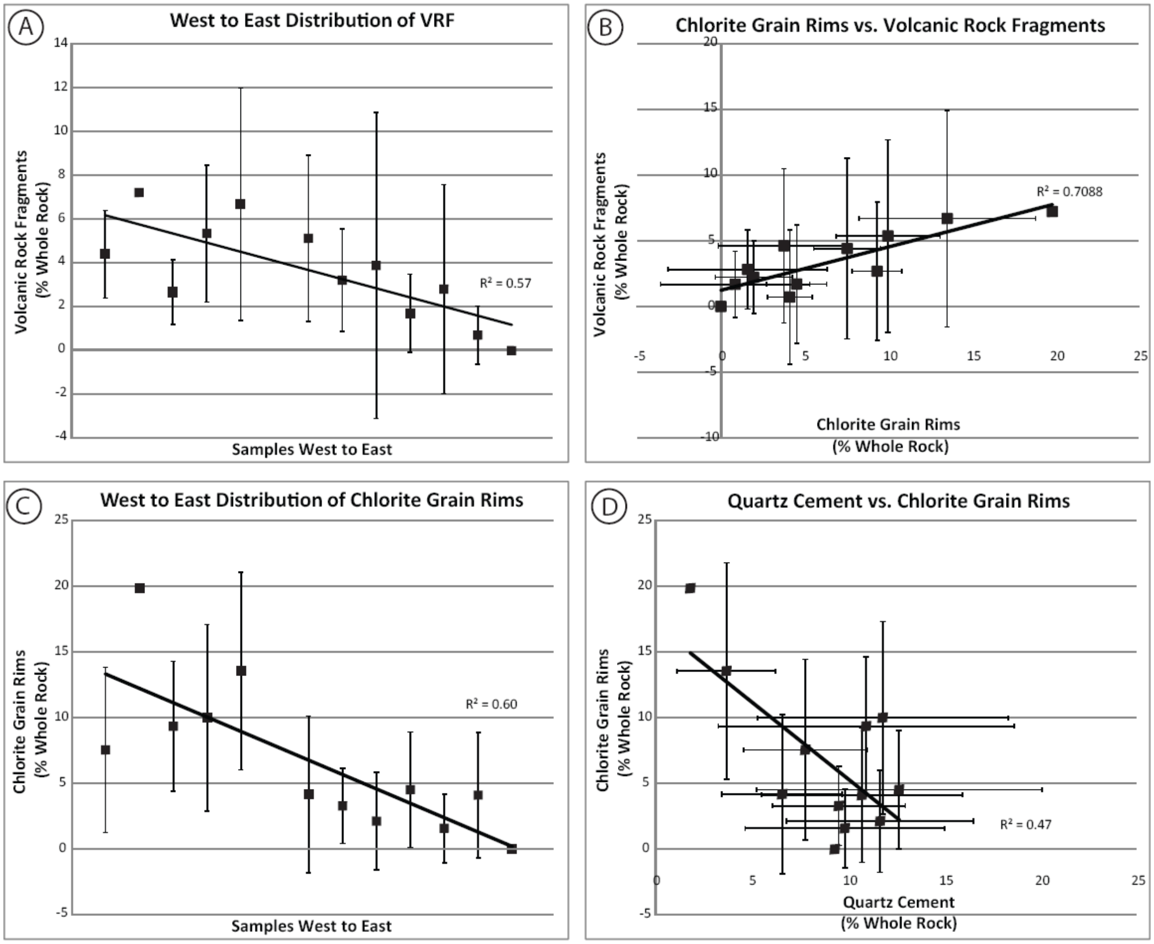

Figure 5.4. Graphs showing the west to east decline in (A) volcanic rock fragments (VRFs) and (C) chlorite coats and the relationship between (B) chlorite coats and VRFs and (D) chlorite coats and quartz cement.

Porosity

Most of the porosity observed in this study is intergranular porosity (Figure 5.3) although varying amounts of intragranular porosity are also present (Figure 5.3B, C, E, L). Intergranular porosity values quantified in this study ranged from 0 % to 15 %. Samples with 0 % intergranular porosity are either completely cemented with carbonate cement or contain a relatively large percentage of detrital matrix. Smaller amounts of intragranular porosity (maximum 2 % of the rock volume) created by the partial or total dissolution of framework grains was also observed. These moldic intragranular pores are easily identified within carbonate cemented zones and in grains where remnant chlorite coats are left after dissolution of the framework grain (Figure 5.3B, C, E, L).

Deepwater Deposits

Although thin sections and actual core from any deepwater wells through the Tuscaloosa interval were not available to the author, publically available photomicrographs and porosity and permeability measurements of the Tiber well (see Figure 5.1 for location) allow for some insight into petrographic nature of these deepwater deposits. Photomicrographs, porosity and permeability measurements, and X-ray diffraction analysis (XRD) were available from subsea depths 33,818 to 34,052 ft. in the Tiber well (drilled in 4,132 ft. of water). Previously these deposits have been reported as sand-rich gravity flow deposits fed from the onshore Lower Tuscaloosa deposits located at the shelf edge (Chapters 2 and 3). Photomicrographs show a poorly-sorted rounded to sub-angular sandstone. These sandstones are dominated by quartz and have minor amounts of potassium feldspar and plagioclase (Figure 5.5). Feldspar grains commonly show dissolution and replacement by illite and calcite. Rock fragments appear to be less abundant in the offshore samples compared to the onshore samples and when present appear to have undergone heavy dissolution. Similar to deposits in the onshore

Figure 5.5. Photomicrographs of the Tiber well in the deep offshore Gulf of Mexico (see Figure 5.1 for location) show the poorly sorted and sub-rounded to sub-angular nature of the deposits. Note the lack of well-developed grain coats. (A) Dominantly cemented by calcite (stained red). Dissolution and replacement (by kaolinite) of rock fragments and feldspars is also noted. (B) Abundant quartz cementation and dissolution of rock fragments and feldspars.

Lower Tuscaloosa there appears to be zones completely cemented by carbonate cement (Figure 5.5a). Smaller patches of poikilotopic carbonate cement are also fairly common. XRD analysis indicates that calcite can make up to 21 % of the rock volume. Unlike deposits in the onshore Lower Tuscaloosa there appears to be a complete lack of chlorite grain coats. A few samples show drusy clay coatings. X-ray diffraction analyses show about 1 % chlorite and up to 6 % illite make up the total rock volume. Additionally, many of the samples appear to have a much larger percent of quartz cement compared to their onshore equivalent (Figure 5.5b). Measured porosity ranges from 2.5 – 11 % and average 8 % (±2%), and measured permeability averages less than 0.1 md showing a greatly reduced reservoir quality when compared to the onshore deposits.

PARAGENETIC SEQUENCE – ONSHORE LOWER TUSCALOOSA

The paragenetic sequence of the onshore Lower Tuscaloosa appears to be dominated by approximately six different events: (1) compaction, (2) dissolution, (3) chlorite rim formation, (4) quartz cementation, (5) carbonate cementation, and (6) accessory mineral formation. This sequence is obviously complex and several of the events may have had long durations, overlap, and/or be repeated multiple times. Additionally, this study has sampled from core across the region at many different depths which has some advantages and disadvantages. The wide variability of diagenetic processes that can occur within the same well on a small scale can possibly be even more pronounced on a large regional scale. Despite this possibility, it appears that the Lower Tuscaloosa Formation is generally similar in the diagenetic processes that have acted on the rocks after deposition. The advantage to sampling from a wide range of locations, depths, and depositional environments is you can capture different windows of the paragenetic sequence. This advantage will allow a researcher to gain a better insight into how this sequence changes with burial, spatially across the region, and between depositional environments. This complexity should be taken into account when considering the proposed paragenetic sequence in this study.

Compaction

Mechanical compaction occurred shortly after deposition of the Lower Tuscaloosa and appears to vary between samples to some degree. This variation depends on the amount of ductile grains deposited in the sample and the degree of early cementation. Intergranular volume (IGV) was used as a quantitative measure of the degree mechanical compaction reduced intergranular porosities. In a rigid sandstone, like most of the samples in this study, mechanical compaction can reduce initial intergranular porosities of about 40 % (shortly after deposition) to a lower threshold of about 26 % (Paxton et al., 2002). Although the majority samples in this study were rigid sandstones there were some sandstones which contained relatively high amounts of detrital matrix that were considered non-rigid.

IGV calculations for the Lower Tuscaloosa sandstones ranges from 16 % to 57 % and averages 35 % (± 9 %). IGV values above 40% are samples with a higher content of detrital matrix in which initial IGV values are expected to be over 40%. There are also some occurrences of IGV values above 40% in samples lacking matrix. These samples had large amounts of carbonate cement that filled intergranular porosity and most likely replaced a certain percentage of unstable detrital grains, such as feldspars and rock fragments, leading to higher IGV values than expected.

In the Lower Tuscaloosa additional compaction is commonly observed in the compaction of volcanic and sedimentary rock fragments which create small amounts of pseudomatrix. Additionally, further compaction occurs where detrital grains have completely dissolved and the remnant chlorite coats have compacted together into masses of chlorite. Small amounts of chemical compaction due to pressure solution were also observed.

Dissolution

Dissolution and replacement of feldspar grains and rock fragments, especially volcanic rock fragments (VRFs) and chert, was likely an early and fairly consistent process. In many cases these grains are so heavily altered it is difficult to identify them. Feldspar is most commonly altered to illite and in some cases is partially replaced by carbonate cement. Volcanic rock fragments are often replaced by small euhedral crystals of what is thought to be Fe-Ti oxides. There is evidence that both feldspars and VRFs have completely dissolved leaving behind moldic pores. Chert grains (possibly argillaceous) are “dirty” in appearance having been altered and often contain dissolved dolomite rhombohedra within the grain.

Chlorite Formation

The formation of chlorite coats occurs as an early diagenetic process after minor dissolution of feldspar and VRFs but before major dissolution of these elements. In some samples very minor quartz and opaque cementation at the detrital grain surface predates chlorite coats. Samples containing carbonate cement can enclose chlorite that is coating the detrital grains clearly indicating that the carbonate cement clearly postdates the chlorite coat formation. In other samples where carbonate cement is a major component chlorite is absent. It appears in these samples that carbonate cement has formed prior to chlorite formation and appears to have used any space available for the formation of chlorite coats. In chlorite-rich samples in the deeper onshore reservoirs pore-filling radial chlorite can be relatively common. This radial chlorite appears to be one of the latest diagenetic events.

Quartz Cementation

The majority of quartz cementation occurs after chlorite in all samples studied. Quartz cement is more abundant in samples with poorly formed, incomplete, and/or broken chlorite coats. In such samples relatively large euhedral quartz overgrowths can diminish a good percentage of the intergranular porosity. In samples with well-developed chlorite coats only small overgrowths are present nucleating from detrital grains where breaks in the chlorite coats occur. This apparent correlation in chlorite rim completeness with the type and amount of quartz cementation is strong evidence for the previously reported control chlorite has on the cementation of quartz. Indeed, it does appear that chlorite coats inhibit quartz cementation. Quartz cement exhibits a similar relationship to carbonate cement as the chlorite coats in that it can both pre- and postdate carbonate cementation. In samples completely cemented by carbonate there does not appear to be a large component of quartz cement as the carbonate cement appears to have used the nucleation surface and room needed to initiate and sustain quartz cementation.

Carbonate Cementation

Carbonate cementation can volumetrically constitute a large percentage of the total rock volume. Samples completely cemented with carbonate cement were observed in several cases. Often smaller poikilotopic carbonate cements enclosed detrital grains, chlorite cement, quartz cement, and intragranular moldic pores. This indicates carbonate cement was a relatively later diagenetic process occurring after several other events and ceasing before the latest occurrence of dissolution. Although most of the carbonate cement appears to be a later diagenetic event, it may have played an important role in preserving IGV in some of the samples indicating that cementation may have started before mechanical compaction was complete. In cases where samples are completely filled with carbonate cement the cement was observed commonly replacing feldspar grains and possibly unstable rock fragments. Wheat seed siderite appears to occur prior to the poikilotopic carbonate cementation as it is often enclosed in that cement.

Accessory Mineral Formation

Additional cements that can make up fairly large percentages of the total rock volume include Kaolinite and Fe-Ti oxide. These are interpreted to occur at the latest stages of diagenesis as pore filling cements.

DISCUSSION

Regional Trends in Chlorite and Reservoir Quality

In the Lower Tuscaloosa the amount of volcanic rock fragments (VRFs) decreases towards the east (Figure 5.4A). It appears that the relatively volcanic-rich Ouachita Mountains did provide additional VRFs to the depositional systems in the western portions of the basin. Accompanying this decrease in VRFs to the east, chlorite also decreases toward the east (Figure 5.4B). The decrease of both VRFs and chlorite suggests that the additional Fe provided to the system by the dissolution of VRFs is an important process to the formation of chlorite as described by Thomson (1979). 25 samples, the majority of which are located in more eastern positions (see Figure 5.1 for location), contained no chlorite and contained very little to no VRFs. Although it appears that VRFs have a large control on the presence of chlorite, it is apparent that they are not the only control. For example, there are 10 samples with less than 2 % VRFs that have over 5% chlorite coats. Some of these samples with chlorite had no VRFs at all. In these cases, Fe may have come from other sources such as the dewatering of surrounding shales. There are many other Fe-rich elements in many of the samples including ankerite, siderite, and Fe-Ti oxides indicating a large amount of Fe was available throughout the system.

It appears that the chlorite coats do inhibit quartz cementation supporting the numerous studies that have previously observed this occurrence (Thomson, 1979; Ehrenberg, 1993; Bloch et al., 2002; Billault et al., 2003; Dowey et al., 2012). As chlorite coats on detrital grains increase the amount of quartz cement decreases (Figure 5.4D). Additionally, small quartz overgrowths nucleate exclusively from locations on the detrital grains where chlorite coats are incomplete, have broken, or are absent. In samples where there is little to no chlorite coats, quartz cement is relatively more abundant compared to samples with well-formed chlorite coats.

Anomalously High Porosity

A variable range of porosity values have been reported for the Lower Tuscaloosa Formation which are generally considered anomalously high for the age and depth of burial of the reservoir (Thomson, 1979; Hearne and Lock, 1985; Hamlin and Cameron, 1987; Geniuse, 1991; Weedman et al., 1992, 1996; Hansley, 1996). Some of the variable range in porosity can be attributed to the different methods used in obtaining these values along with natural variation in sampling. Porosity in the Lower Tuscaloosa Formation has been measured through derivation of log properties (neutron and density logs) (Thomson, 1979), core plug measurements, and point counting (Hearne and Lock, 1985; Hamlin and Cameron, 1987; Geniuse, 1991; Weedman et al., 1992, 1996; Hansley, 1996). It has been shown in the Lower Tuscaloosa Formation that up to 50% of the total porosity is microporosity which comes chiefly from the chlorite and kaolinite present in the rocks (Stancliffe, 1986; Hogg, 1988; Hansley, 1996). When comparing porosity values for the Lower Tuscaloosa it is apparent that point-counting methods, which do not take into account microporosity, can measure lower total porosity values than log derived porosity measurements which do account for microporosity. Porosity values reported in this study are from point count analysis.

Most of the porosity observed in this study is intergranular porosity (Figure 5.3) although varying amounts of intragranular porosity are also present (Figure 5.3B, C, E, L). Intergranular porosity values determined from point counting in this study ranged from 0 % to 15 %. The porosity values estimated in this study compare favorably to other petrographic studies which measured porosity with point counting methods in the Lower Tuscaloosa (Hearne and Lock, 1985; Hamlin and Cameron, 1987; Geniuse, 1991; Weedman et al., 1992, 1996; Hansley, 1996). The high end of the porosity range from this study (15%) also compares to the log derived porosity values reported by Thomson (1979) (over 25 %) after microporosity is taken into consideration.

A certain degree of anomalously high porosity and permeability may be the result of overpressure in some Lower Tuscaloosa reservoirs (Weedman et al., 1992; Weedman, 1996). Preservation of porosity due to fluid overpressure has been documented in reservoirs younger than the Lower Tuscaloosa Formation (Bloch, 2002). Bloch (2002) suggests that invoking an overpressure mechanism for preserving porosity in pre-Tertiary deposits (e.g. the Lower Tuscaloosa) is difficult because the complex history of overpressure and diagenesis in these reservoirs. This complex history can lead to diagenetic affects, such as quartz cementation, filling the porosity preserved by overpressure. Portions of the deep trend of the stratigraphically deep Lower Tuscaloosa have been identified as overpressured “tongues” which interfinger along strike with a series of normally pressured zones (McCulloh and Purcell, 1983; Weedman et al., 1992; Weedman, 1996). McCulloh and Purcell (1983) noted no correlation between the occurrence of hydrocarbon reservoirs and the occurrence of overpressured or normally pressured stratigraphic zones in the Lower Tuscaloosa Formation. In addition, these authors observed that the reservoirs with the highest porosity and permeability values exist in the normally pressured zones (“tongues” of McCulloh and Purcell, 1983). This indicates that diagenetic events in the Lower Tuscaloosa may play a more important role in the reservoir quality than does overpressure. However, Weedman et al. (1992) and Weedman (1996) noted a higher degree of compaction in normally pressured sandstones in the Lower Tuscaloosa supporting the view that overpressure does have some control in reservoir quality. The contrasting views documented in the discussion above serve to highlight the complex nature of porosity preservation in the deep Lower Tuscaloosa Formation.

Environment of Deposition and Grain Size Control on Chlorite

There was no obvious correlation between the abundance of chlorite and depositional environment, rock type, or grain size in the Lower Tuscaloosa Formation. However, Dowey et al. (2012) note some influence of depositional environment in rocks containing chlorite. Chlorite in the Lower Tuscaloosa can occur in relatively large percentages or be equally as absent in fluvial, deltaic, and estuarine depositional environments and among different rock types and grain sizes (Table 5.2). The lack of control of depositional environment was most pronounced in the fluvial depositional environment. The Lower Tuscaloosa in the Cranfield and T.D. Bickham wells (see Figure 5.1 for location) are dominated by fluvial facies and are located in the western portion of the study area where there is a greater influence of the Ouachita Mountains providing sediments to the system. Each of the samples studied from these cores contains over 20% chlorite of the total rock volume and abundant volcanic rock fragments (VRFs). In contrast, most of the samples from the fluvial facies of the Snowden well located further to the east of the Cranfield area (see Figure 5.1 for location) contain none to very trace amounts of VRFs and chlorite. Similar variations exist, although not as pronounced, within cored intervals interpreted as deltaic and estuarine Lower Tuscaloosa facies. Similarly, the abundance of quartz cement ranges widely from 0% to up to 20% and averages 8 % to 12.4% in all of the rock types and grain sizes studied (Table 5.2).

Future Exploration Efforts

The expense of wells and the often hostile drilling conditions in stratigraphically deep offshore Tuscaloosa plays highlight the need for a better understanding of the petrophysics, clay mineralogy of cements, and influence of petrography on pressure and the diagenetic history of this interval. Because sediments offshore are fed by point sources onshore, it is possible that these two deposits are compositionally similar in both the presence of VRFs and chlorite coats. If this were the case, it would be feasible that mechanisms preserving porosity and permeability in the deep reservoirs onshore may be similar to those acting on the deep reservoirs offshore. This possibility will be discussed below knowing that there is extremely limited data available and there undoubtedly exists a large variability in reservoir quality of these offshore deposits.

The only data available on the petrography and reservoir quality of the deepwater offshore Tuscaloosa deposits is from limited intervals in the Tiber well, located in Keathley Canyon (see Figure 5.1 for location). From this data it does not appear as if chlorite is an important element of these deposits. In contrast to what is seen onshore, no evidence of abundant VRFs or chlorite coats was observed in images of these deepwater Tuscaloosa deposits. The lack of VRFs in these deposits may have limited chlorite coat formation and allowed for the more pervasive quartz cementation noted in these samples. Calcite cementation, similar to what is seen onshore, is also common in the Tiber samples. These factors lead to low porosity (< 11 %) and especially low permeability (< 0.1 md) values that do not bode well for hydrocarbon production from these intervals. However, these intervals are saturated with hydrocarbons proving a working petroleum system. It is possible that the sediments from these reservoirs were sourced from the eastern volcanic-poor Appalachian provenance or another volcanic-poor source. It is also possible that VRFs were selectively sorted out as sediments became more mature over the greater than 400 km transport distance from the shelf edge to the basin floor. Wright and Anderson (1982) showed that sediment gravity flows can affectively sort sediments over a distance as little as 10 km. The presumed higher densities of these VRFs would have allowed them to be selectively deposited ahead of the other less dense detrital grains. Whatever the cause, sorting and/or provenance control, fewer VRFs would have resulted in less Fe ions for the formation of chlorite.

Of the other two offshore wells penetrating Tuscaloosa deepwater deposits, the Davy Jones #2 has been reported a discovery while the BAHA #2 was reported a dry hole. No data on the reservoir quality or petrography of these wells was available to the author other than what can be derived from wireline logs. BAHA #2 is the western most well in the deep offshore leading to the possibility of having more VRFs derived from the volcanic–rich Ouachita provenance. However, BAHA #2 is also located the farthest transport distance from the Ouachita provenance allowing for the possibility of VRFs having been selectively deposited well before they reached the deep offshore. Davy Jones #2 is the closest to the point sources documented in this study (Chapter 2). The proximity of the Davy Jones #2 to the source in the Ouachita provenance makes it a likely candidate to have VRFs and chlorite coats. To date, no production has come from any of these wells from the Tuscaloosa interval.

One possibility for future exploration efforts would be in sediments relatively rich in VRFs because of their connection to favorable reservoir quality. These areas may be located to the west and proximal to the volcanic-rich Ouachita provenance. Work on these deepwater deposits is in its infancy, and we are only beginning to understand some of the preliminary observations. Much more data obtained from future exploration efforts is needed before the exact nature of these deepwater deposits and the link to their onshore source can be determined. Although much work has been done on understanding the nature of chlorite in individual samples and even individual wells, and many causes for and impacts of its occurrence have been speculated on, in fact much more work needs to be done to assess the validity of all the hypotheses of chlorite origin and distribution. This work is the first to examine for any regional trends in chlorite relative to source areas and relative to depositional environments in the Lower Tuscaloosa. Its results stand in support of several speculations by previous authors regarding the relationship between porosity and permeability and chlorite, between chlorite presence and volcanic rock fragments, of chlorite, calcite and quartz development to burial of the host rock and between hydrocarbon production and chlorite presence. The question of whether chlorite plays similar rolls in nature and productivity of deepwater facies in the Tuscaloosa remains to be answered. However, if such a link does exist it will be imperative to understand the sink to source links that may have enhanced movement of VRFs from the shelf into the ultra-deepwater deposits.

CONCLUSIONS

This study helps to further the understanding of the regional trends in reservoir quality from onshore to offshore Tuscaloosa deposits. This study concludes that in the Lower Tuscaloosa sandstones; (1) The presence and abundance of chlorite coats is related to the presence and abundance of volcanic rock fragments which are in turn controlled by provenance; (2) The presence and abundance of chlorite coats is not affected by depositional environment, grain size, or rock type; (3) The formation of chlorite coats is a dominantly early diagenetic process which inhibits quartz cementation preserving intergranular porosity; (4) To date, from extremely limited data, chlorite does not appear to be an important component of the Tuscaloosa deepwater deposits and reported reservoir qualities in this interval are poor.

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Chapter 6: Conclusions

INTRODUCTION

The goal of this study was to further the knowledge of the Lower Tuscaloosa Formation in four areas: (1) Sequence stratigraphic and paleogeographic framework of the Lower Tuscaloosa Formation and the influences on sediment distribution and depositional environments; (2) Distribution of facies and depositional systems; (3) Size and capacity of the fluvial systems; and (4) Regional distribution of chlorite with insights into the controls of chlorite occurrence and implications to reservoir quality. Several conclusions as a result of this dissertation (listed below) are evidence of having accomplished these goals and have helped to further the knowledge of the Lower Tuscaloosa Formation. Additionally, larger stratigraphic, sedimentologic, and petrographic questions were answered that can be applied to formations with similar characteristics. This dissertation also enables a better understanding of Late Cretaceous geology particularly in the Gulf of Mexico.

1. Five sequences (S0, S1, S2, S3, and Stringer) are present in the overall regressive cycle of the onshore Lower Tuscaloosa.
2. The basal sequences (S0 and S1), with the base of S0 representing a large scale subaerial unconformity, marked times of sediment bypass in the Lower Tuscaloosa. This sediment bypass supplied a large amount of sediment to the toe-of-slope and abyssal plain during Tuscaloosa time.
3. The younger sequences (S2, S3, and Stringer) are increasingly dominated by transgressive systems tracts from S2 to Stringer time. This is marked by an increase of marine influence during S2 and S3, followed with deposition of nearshore estuaries during the Stringer, finally capped by the fully marine Middle Tuscaloosa Shale.

4. This is the first highly detailed regional paleogeographic study of the Lower Tuscaloosa Formation and the first to apply an incised valley/estuarine model to this formation. Incised valleys are initially filled with stacked fluvial channels followed by backstepping estuarine deposits. This model can be applied to other incised valley systems for the recognition and better prediction of downdip facies.
5. Several thick fairways of the Lower Tuscaloosa originating from the Appalachian Plateau to the northeast and from a large drainage trending within the ancestral Mississippi valley feed these depocenters and terminate southward at the highly extensional paleoshelf edge. The detailed paleogeography and petrography documented in this study indicate that prior to the Tertiary much of the sediment deposited into the Gulf of Mexico was derived from the Appalachia Mountains.
6. Deepwater facies made up of gravity flow deposits were hosted from depocenters that formed in the areas of the modern-day Mississippi and Alabama panhandles, as well as along areas of Louisiana. These sites provided significant sediment into the toe of slope fans and leveed channel systems (Davy Jones #2) and the basin floor fans (BAHA#2 and Tiber) associated with deepwater environments of the Lower Tuscaloosa. This is the first study describing the details of these deposits.
7. Both wave- and tide-dominated deltas are recognized as occurring in depocenters near the Lower Tuscaloosa-time shelf edge.
8. Integrating ichnofacies with detailed observations of sedimentology enables better differentiation of facies, with distinct and recognizable ichnogenera in both the deltaic and estuarine facies. Sporadic distribution and reduced

diversities of ichnogenera relating to the *Cruziana* ichnofacies represent the brackish conditions of these environments. This is the first study to describe and integrate ichnology from the Lower Tuscaloosa Formation with detailed core descriptions.

9. Recent drilling penetrations of Tuscaloosa-age slope and abyssal plain facies prove the presence of a deepwater depositional system associated with the Lower Tuscaloosa. The sheer volume of sediments located at the time equivalent shelf edge suggests the strong possibility of a much broader regional occurrence of deepwater facies than is presently known.
10. Channel dimensions determined by channel deposit cross set measurements from Lower Tuscaloosa core and point bar thicknesses from wireline log measurements (Bridge et al., 2003) of the Lower Tuscaloosa logs matched closely. This confirms the consistency of both methods and lends validity to the results.
11. Outcrop analogs for the Lower Tuscaloosa Formation include the Siwalik Group in Pakistan, the Escanilla Formation of Spain, the Chuckanut Formation of Washington, and the Canyon Creek Member of the Ericson Formation
12. Modern analogs for the Lower Tuscaloosa appear to be the Missouri, Ohio, and Alabama Rivers.
13. Basement-rooted and long-stable structural uplifts influenced drainage directions and formed buttresses that resulted in ponded thicknesses of fluvial, estuarine, and coastal plain deposits in locations well landward of shelf edge locations. These same structures focused drainages to very specific deltaic depo-sites at the shelf edge. These structures may have been important during

the deposition of other formations in the Gulf of Mexico and may aid in a better understanding of their distribution of facies and depositional fairways.

14. The Late Cretaceous topography influenced the distribution and magnitude of processes active along the transgressive-marine shoreline of the Lower Tuscaloosa. Narrowing of valleys between uplifts enhanced tidal processes within flooded valleys leading to lowstand tide-dominated deltas and transgressive tide-dominated estuaries.
15. The formation of chlorite coats is a dominantly early diagenetic process as determined by thin section analysis of textural and crosscutting relationships.
16. Chlorite grain coats in the Lower Tuscaloosa help to inhibit quartz cementation preserving intergranular porosity resulting in better than expected reservoir qualities even with deep burial.
17. The presence and abundance of chlorite coats is related to the presence and abundance of volcanic rock fragments (VRFs) As VRFs decrease in abundance so do chlorite coats.
18. Both VRFs and chlorite coats decrease eastward in the study area toward the volcanic-poor Appalachian Plateau. This suggests validity to Thomson's (1979) assertion that chlorite is provenance controlled.
19. The presence and abundance of chlorite coats in the Lower Tuscaloosa Formation is not affected by depositional environment, grain size, or rock type as no correlative trends between these variables were observed.
20. To date, from extremely limited data, chlorite does not appear to be an important component of deepwater deposits. Additionally, porosity and permeability values reported from the Tiber well indicate a poor reservoir quality in these intervals.

LIMITATIONS AND FUTURE WORK

The conclusions of this study emphasize the importance of this dissertation to the understanding of the Lower Tuscaloosa Formation. However, there are limitations to this study that should be noted.

The author recognizes the difficulty in correlating individual fluvial sequences on such a regional scale and especially in the fluvial facies of the Lower Tuscaloosa. It is possible that the individual sands and shales may not be connected and/or their thicknesses may alter dramatically between wells. As a result, the bounding surfaces of the five sequences should be regarded as a representation of an overall and regionally extensive change in the system and not a direct correlation of sand and shale bodies. A pattern of sandy amalgamated channels and heterolithic (muddier) periods of floodplain alluviation typifies cycles in the terrestrial-fluvial systems. Blum and Tornqvist (2000) suggest the effects of sea level change can be felt up to 400 km up the incised valley. Dramatic increases in muds associated with transgressive deepening within the Lower Tuscaloosa have been recognized by numerous authors as a regional phenomenon (Karges, 1962; Chasteen, 1983; Hansley, 1996; Mancini and Puckett, 2005). Numerous authors have discussed the documented effect of such changes in sea level, as well as climate and tectonics on system architecture and net-to-gross (see discussion chapter 4),

The locations of major basement structures used in this study are, at times, inexact and speculative. This is especially true with the transform faults under the Sigsbee Salt sheet which are difficult to image with seismic data and difficult to interpret with accuracy. The interpreter, at times, must infer the location of these faults, which are in fact zones of faulting. As a result the interpretation of these transform faults can appear too evenly spaced and straight to represent their presumed complexity. The exact planform and location of these fault zones are slightly flexible. In contrast, the locations of basement structures used in this study have been well accepted. Determining the exact location of these basement structures was beyond the scope of this study, however it is believed that the structures used in this study (taken from Stephens, 2001 and 2009) are a good representation of the location of these structures. Stephen's (2009) interpretation incorporates the interpretations of basement structures from a wide variety of sources with additional observations to get the best available interpretation. Additionally, regional isopach maps were compared to basement structures after mapping was completed. The resulting positive correlation between depositional fairways and basement structures is supporting data that these structures did exist syndepositionally with the Lower Tuscaloosa and do affect the location of sediment fairways and the interaction of currents and waves with shoreline depositional systems.

When discussing the deepwater deposits of the Tuscaloosa Formation it is important to recognize the scarcity of data that is currently available in these deposits. The most useful data providing permeability and porosity values as well as XRD analyses was available in only one well (Tiber) and from only limited intervals. As a result, the exact nature of these deposits, their reservoir quality, and their ability to produce hydrocarbons is not fully understood. Despite the sparse amount of data, these few data points have opened a plethora of new questions for future study. One such question, that also has bearing on Tertiary deepwater prospectivity, is what is the mechanism for how these sand rich gravity flows travel from their source at the shelf edge to over 400 km on the abyssal plain?

The dramatic increase in interest in shale plays throughout the world opens up the Middle Tuscaloosa marine shale for numerous future studies. A very limited amount of work has been done on this formation but the productive interval of the Middle Tuscaloosa appears to be the lower interval or the transgressive systems tract (retrogradational wedge) of the Tuscaloosa sequence. Therefore, the paleogeomorphology of the underlying estuaries and back bay regions of the Lower Tuscaloosa Stringer sand may have some important bearing on the distribution and nature of the lower Middle Tuscaloosa deposits. In addition, VRF concentrations in the Stringer intervals of the Lower Tuscaloosa may have bearing on VRF concentrations in reworked backstepping deposits of the Middle Tuscaloosa.

With the recognition of these limitations several points of future work would help to increase our understanding of the Tuscaloosa Formation and the Gulf of Mexico in general.

1. Collection and release of additional information on the deepwater facies of the Lower Tuscaloosa.
2. A better understanding of basement structures and how they affect sedimentation in the Gulf of Mexico.
3. Additional mapping of the Tuscaloosa Formation into Arkansas and northern Louisiana to the west and into eastern Alabama and Georgia to the east may reveal other point sources for offshore Tuscaloosa sediments and will further our understanding of the onshore sedimentology and stratigraphy.
4. Studies involving the flow mechanisms of far-travelled gravity flows may help better explain how these sediments travel these great distances.
5. Mapping of the lower Middle Tuscaloosa and analysis of its relationship to the petrography and thickness of the Lower Tuscaloosa sequences.
6. Modeling of ocean currents of the upper Lower Tuscaloosa time shelf and how those currents interacted with bathymetry of the flooding shelf and shelf edge to influence distribution of grains sizes and lithologies in the Middle Tuscaloosa.

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Appendices

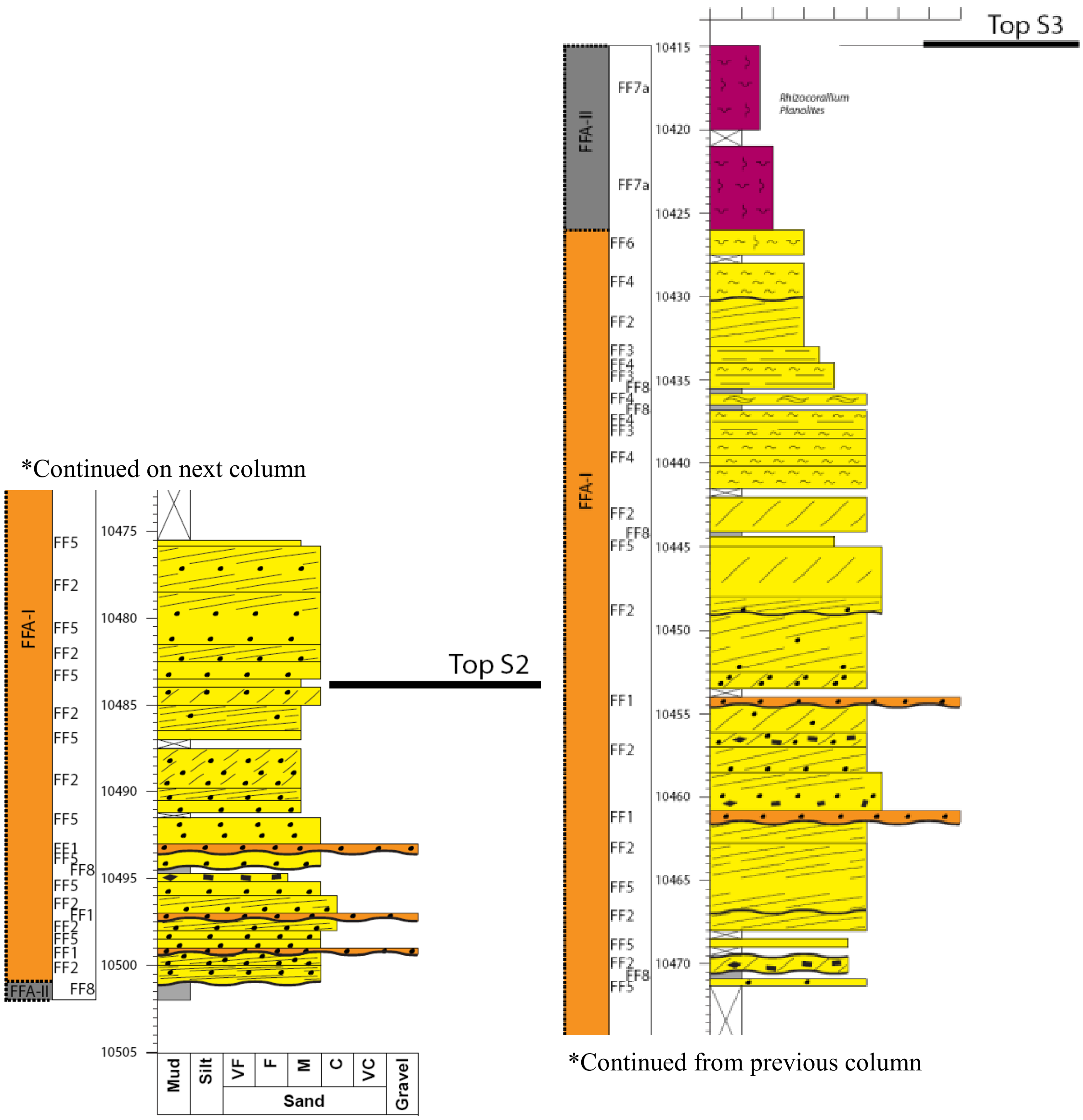
Appendix Ia: Well information for wells used in Figure 2.2

Well Number from Figure 2.2	Operator	Well Name	Well Number	County/ Parish	State	Field	Township and Range
1	BP America Production Company	Rougon J V Etal	3	Pointe Coupee	LA	Profit Island	T5S R11E S48
2	Stacey Bryan C	Anderson Fred AIII	1	Amite	LA	Liberty	T1N R4E S15
3	Denbury Onshore LLC	McComb Field Unit 1	2	Pike	MS	McComb	T3N R7E S14
4	Mobil Exploration Company Inc	Tynes	1	Lawrence	MS	Topeka	T5N R10E S11
5	Marathon Oil Company	Berry 21-12	1	Jefferson Davis	MS	Gwinville	T9N R19W S21
6	Tellus OPR Group LLC	Russel B R	1	Smith	MS	Ely Creek	T2N R7E S13
7	Wyatt L A	O G Horne	1	Newton	MS	Wildcat	T5N R12E S27
8	Amoco Prod CO	W H Lucky UN 6-16	1	Lauderdale	MS	Wildcat	T8N R16E S6
9	Range Production CO	Sumter Farm & Stock	1	Sumter	AL	Wildcat	T22N R3W S4
10	Meridian Oil Inc	Gulf States Paper C	1	Greene	AL	Wildcat	T24N R2E S16

APPENDIX IB: WELL INFORMATION FOR WELLS USED IN FIGURE 2.3

Well Number from Figure 2.3	Operator	Well Name	Well Number	County/ Parish	State	Field	Township and Range
1	Pruet Chelsey Drlg	USA	1	Natchitoches	LA	Wildcat	T6N R7W S31
2	Kadane Oil Co	Miller W B	1	Grant	LA	Wildcat	T7N R2W S23
3	Justiss Oil Co Inc	Pipes Mary Etal	1	La Salle	LA	Trout Creek	T8N R3E S19
4	Carter Oil Company	Frank Testa	1	Concordia	LA	Lake St. John	T9N R10E S30
5	Amoco Prod Co	Unit 14-7	1	Claiborne	LA	Wildcat	T10N R3E S14
6	Inexco Oil Co	P H Carraway	1	Copiah	LA	Wildcat	T1N R3W S21
7	Marathon Oil Company	Berry 21-12	1	Jefferson Davis	MS	Gwinville	T9N R19W S21
8	Eog Resources Inc	Green 26-16	1	Covington	MS	Williamsburg S	T7N R16W S26
9	Elf Aquitaine O&G	Board of Supervisor	1	Lamar	MS	Wildcat	T3N R14W S16
10	Humble Oil Refg Co	Maxie Unit SW-61	1	Forrest	MS	Maxie	T1S R13W S13
11	Jett Drilling	Dantzler LBR Co	1	George	MS	Wildcat	T3S R8W S20
12	Conoco Incorporated	Middleton W K	1	Jackson	MS	Wildcat	T4S R5W S35
13	National Energy Grp	McDaniel E L 9-10	1	Baldwin	MS	Wildcat	T6S R4E S9

APPENDIX IIA: CFU 31F-2 CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.2 AND 3.3 FOR DESCRIPTIONS OF FLUVIAL FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY. CONTINUOUS DESCRIPTION IS IN 2 SECTIONS IN ORDER TO FIT THE PAGE.



APPENDIX IIB: T.D. BICKHAM CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.2 AND 3.3 FOR DESCRIPTIONS OF FLUVIAL FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY. CONTINUOUS DESCRIPTION IS IN 2 SECTIONS IN ORDER TO FIT THE PAGE.

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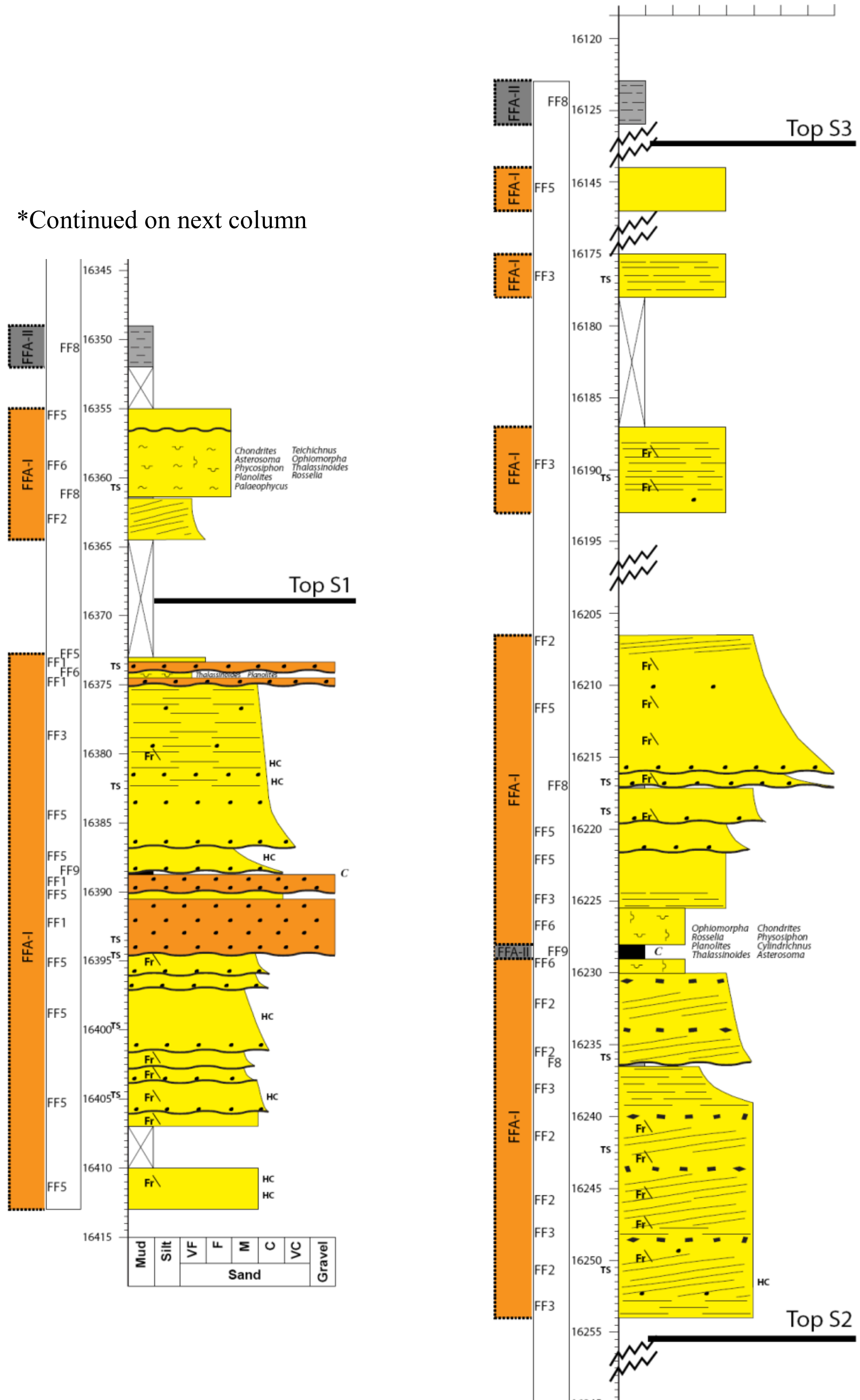
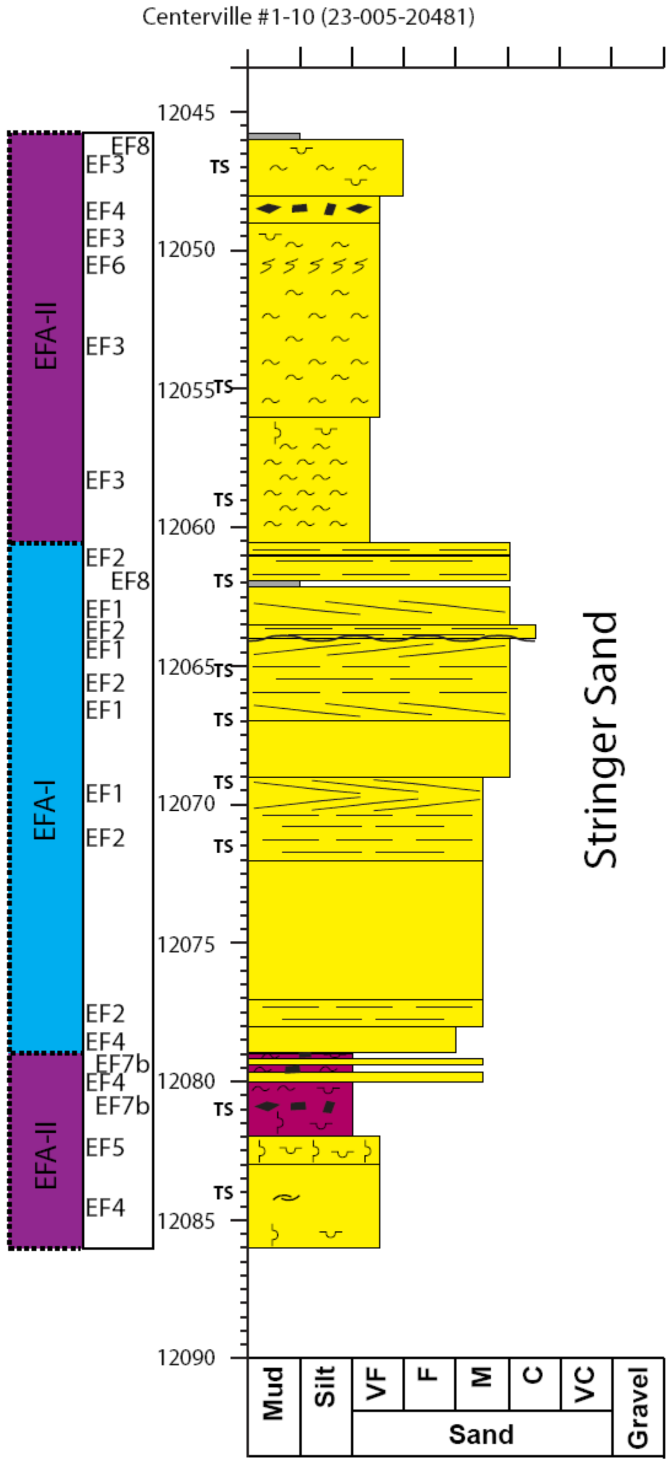
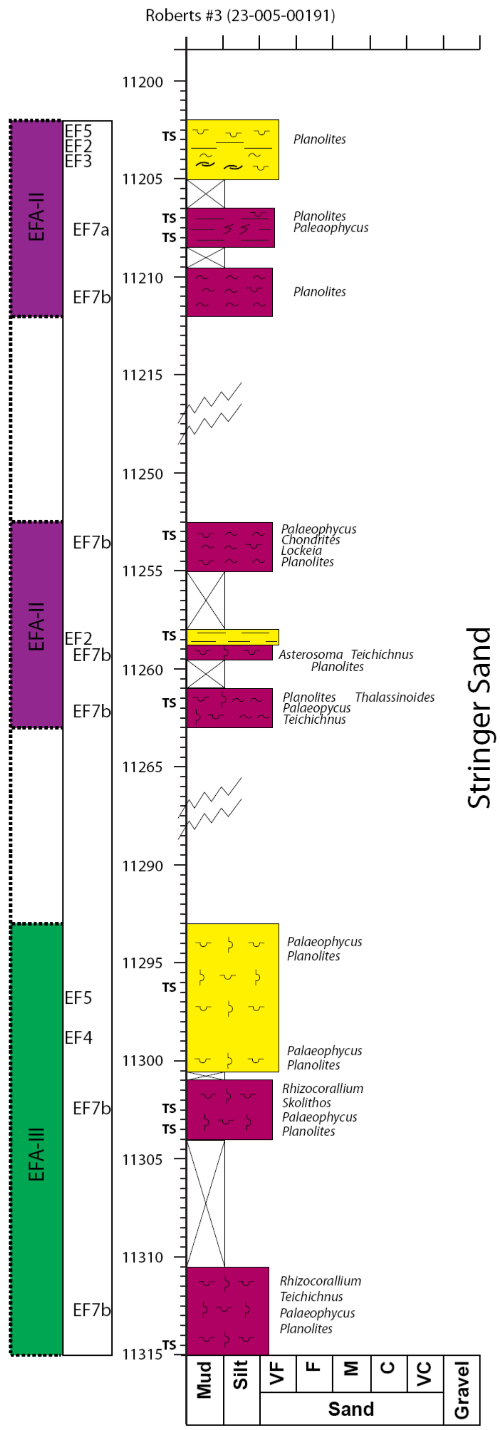


*Continued from previous column

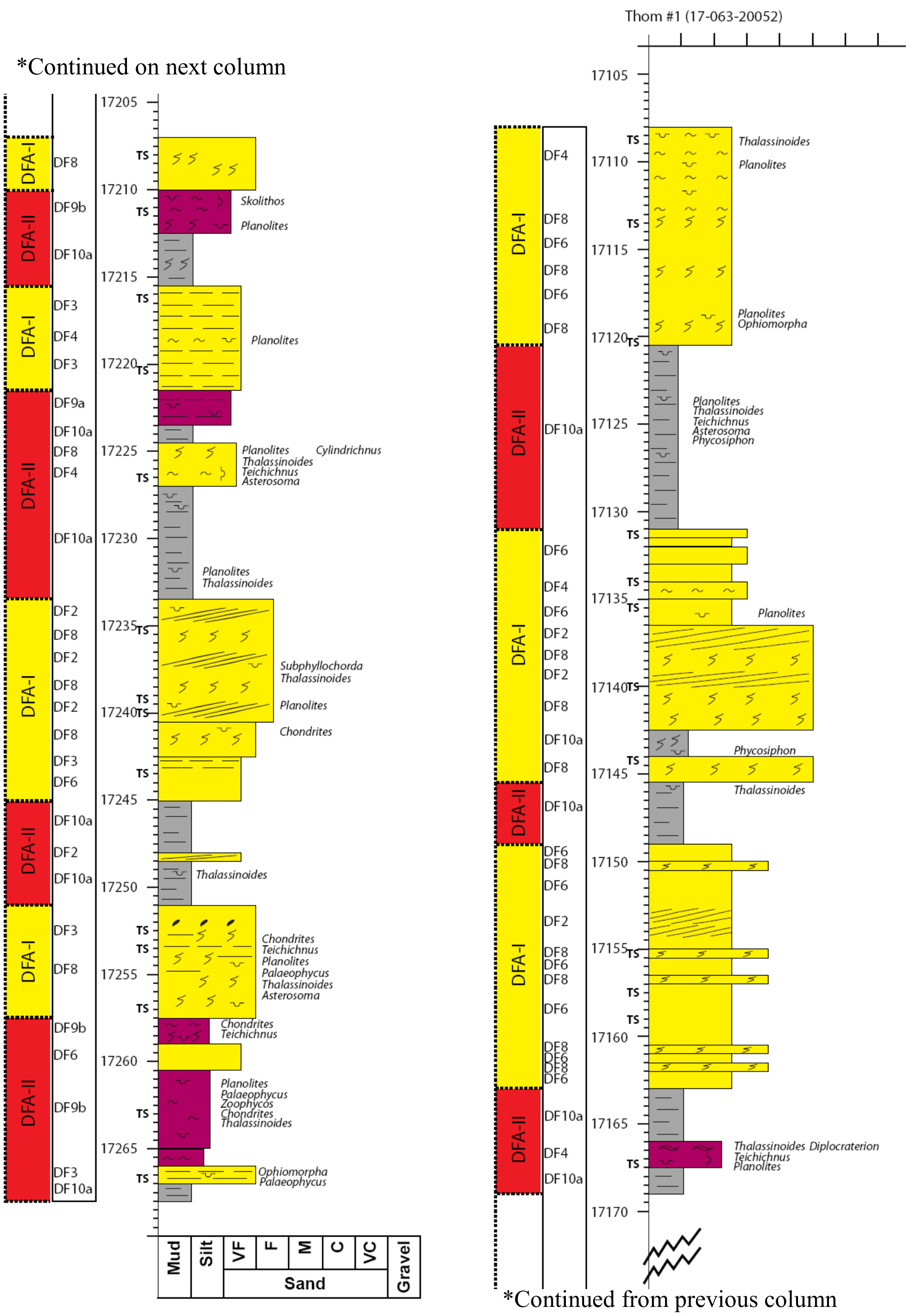
APPENDIX IIC: T.D. CENTERVILLE CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.4 AND 3.5 FOR DESCRIPTIONS OF ESTUARINE FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY.



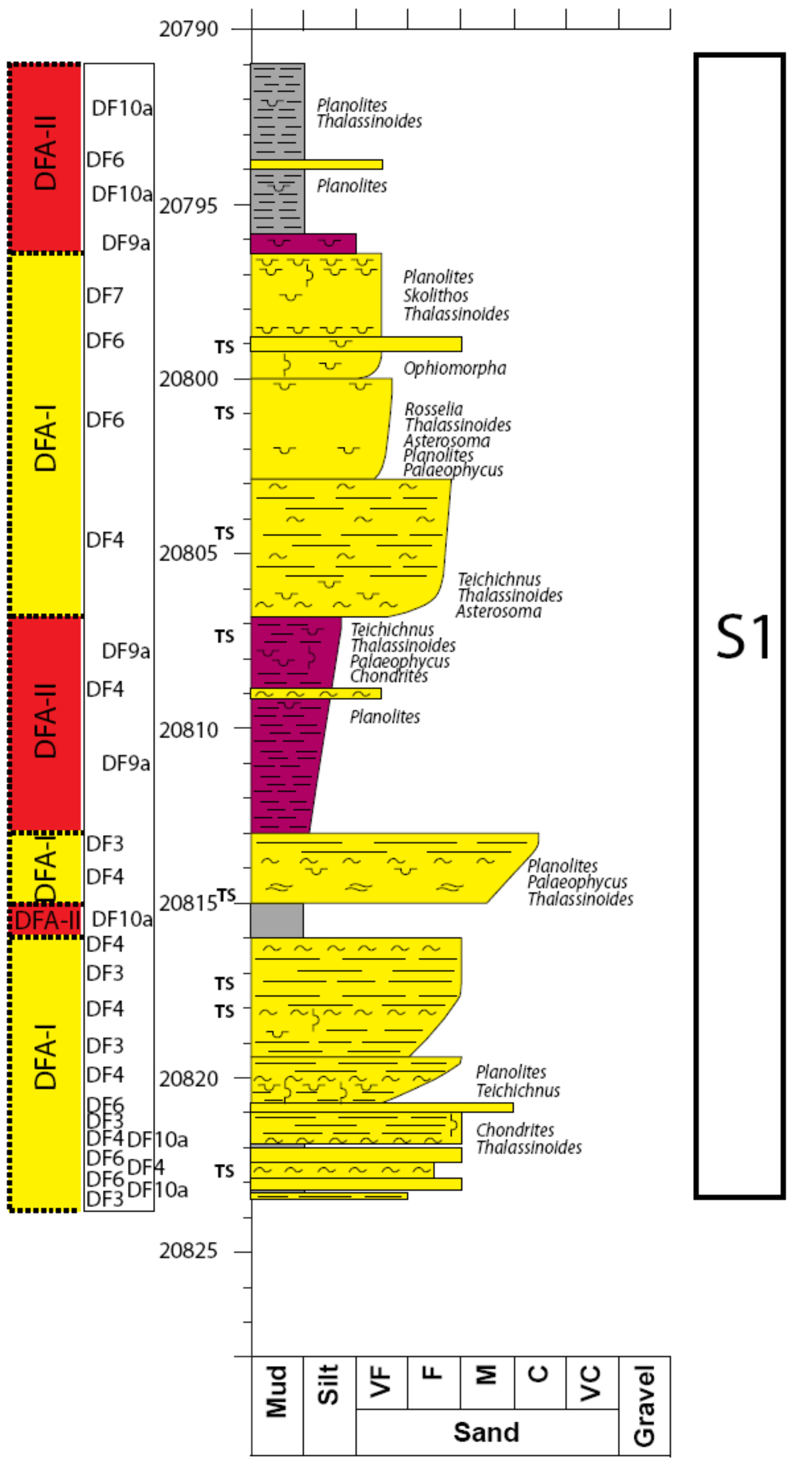
APPENDIX IID: ROBERTS CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.4 AND 3.5 FOR DESCRIPTIONS OF ESTUARINE FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY.



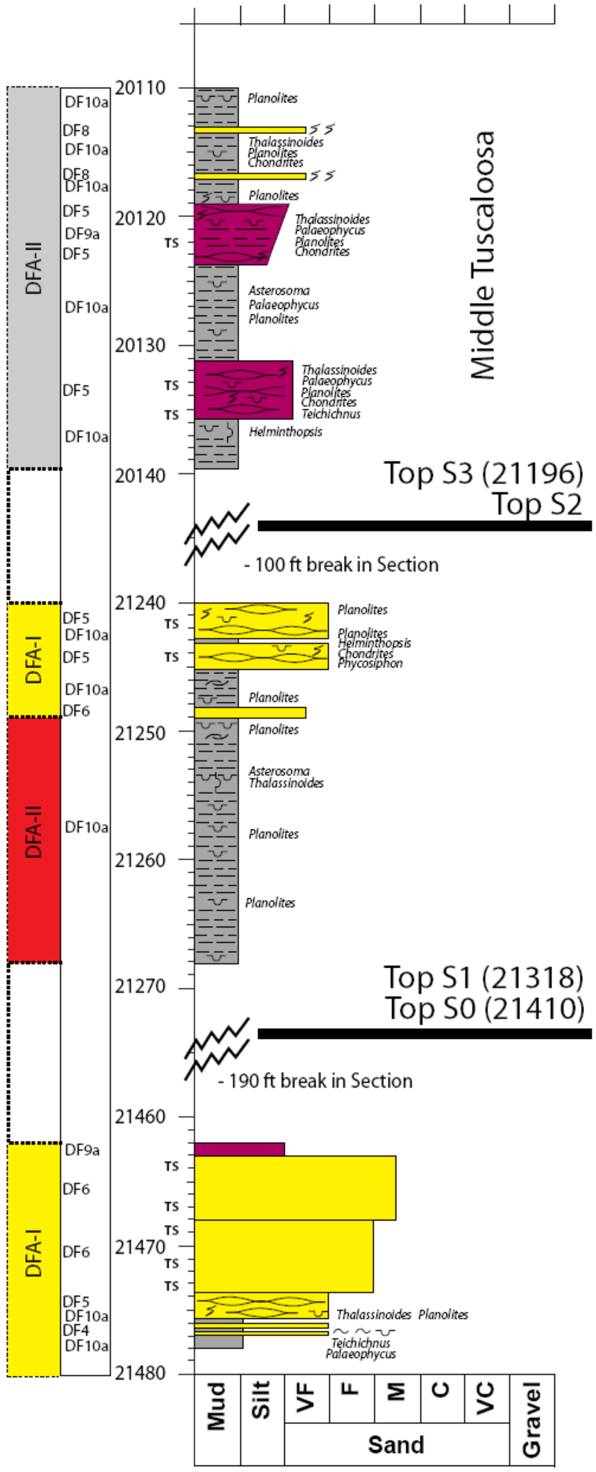
APPENDIX IIE: THOM CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.6 AND 3.7 FOR DESCRIPTIONS OF DELTAIC FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY. CONTINUOUS DESCRIPTION IS IN 2 SECTIONS IN ORDER TO FIT THE PAGE.



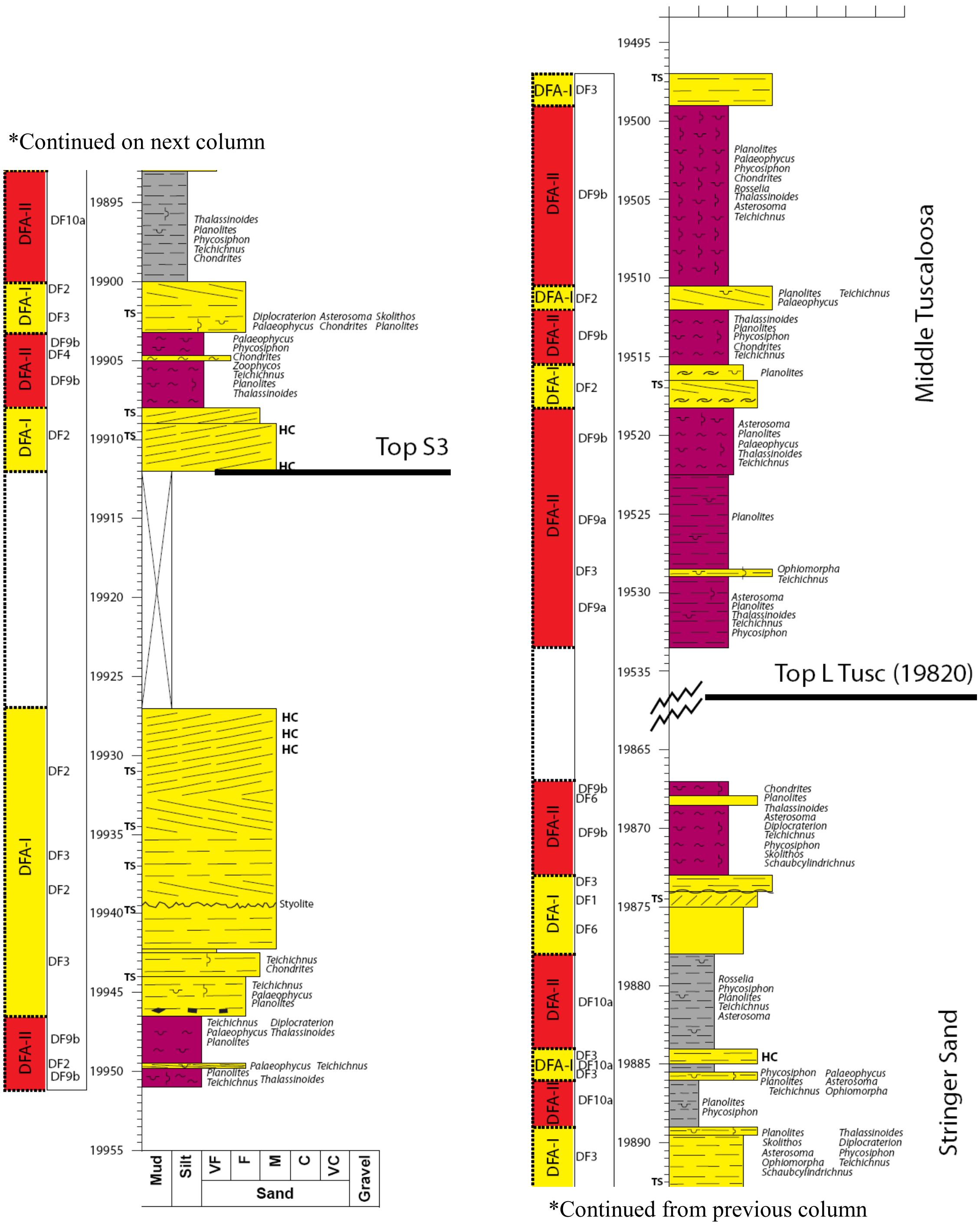
APPENDIX IIF: I.J. MAJOR CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.6 AND 3.7 FOR DESCRIPTIONS OF DELTAIC FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY.



APPENDIX IIG: BILOXI MARSHLANDS #P-2 CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.6 AND 3.7 FOR DESCRIPTIONS OF DELTAIC FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY.



APPENDIX IIH: CROCHETT CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.6 AND 3.7 FOR DESCRIPTIONS OF DELTAIC FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY. CONTINUOUS DESCRIPTION IS IN 2 SECTIONS IN ORDER TO FIT THE PAGE.



APPENDIX III: W.A. LORIO CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.6 AND 3.7 FOR DESCRIPTIONS OF DELTAIC FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY CONTINUOUS DESCRIPTION IS IN 4 SECTIONS IN ORDER TO FIT THE FOLLOWING PAGES.

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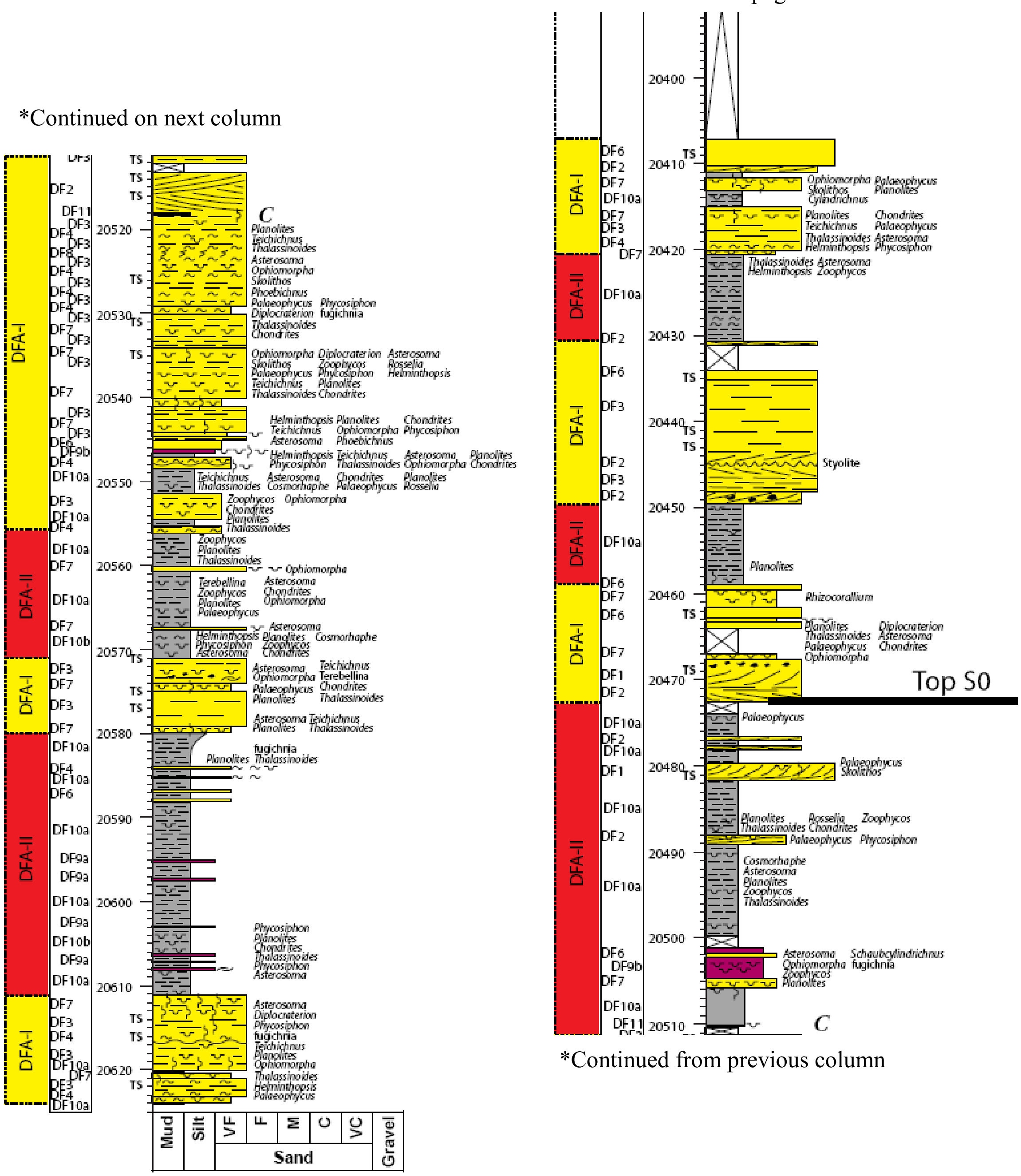

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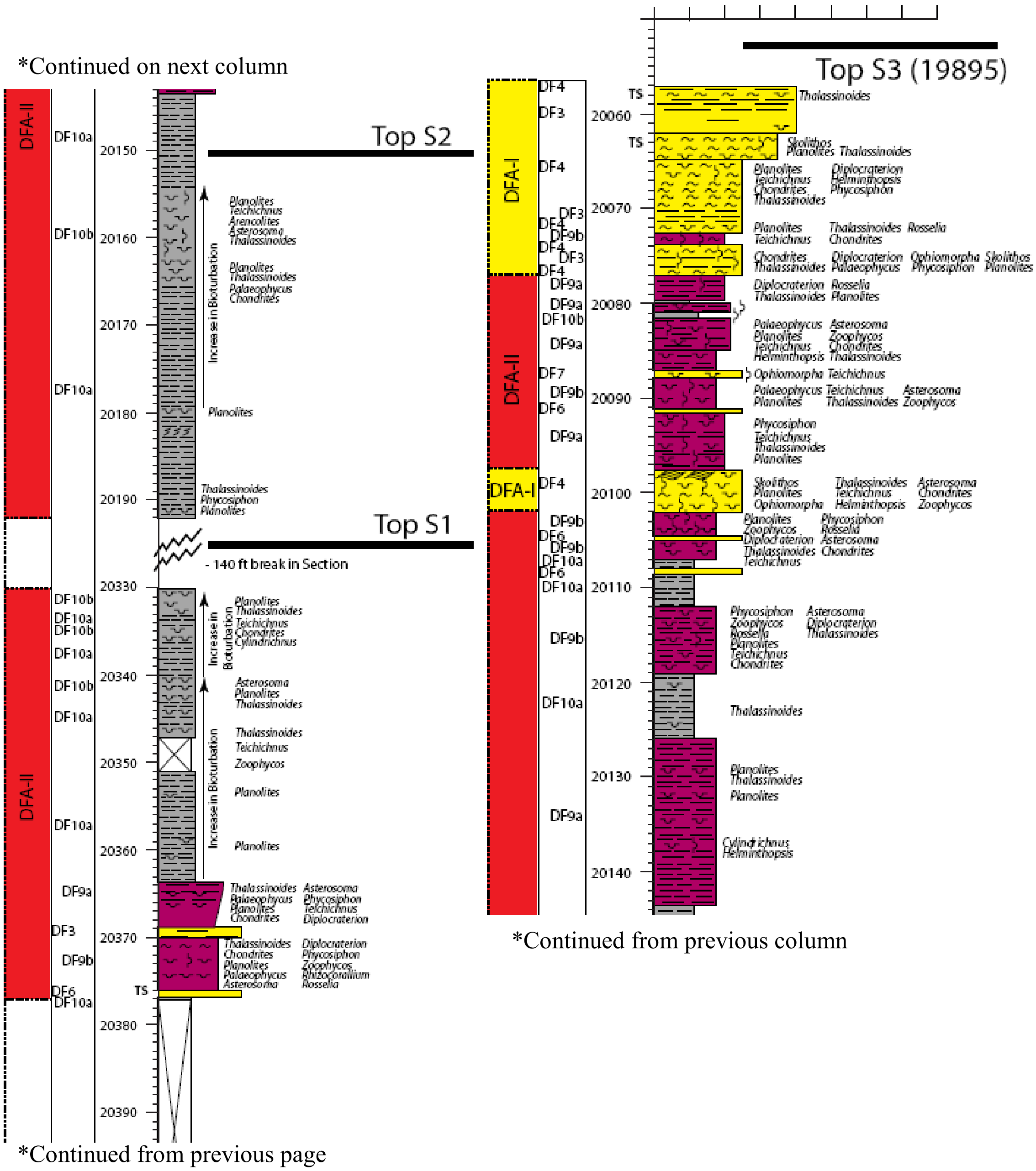
APPENDIX IIi CONTINUED: W.A. LORIO CORE DESCRIPTION. SEE FIGURE 3.4 FOR KEY TO SYMBOLS AND TABLES 3.6 AND 3.7 FOR DESCRIPTIONS OF DELTAIC FACIES AND FACIES ASSOCIATIONS, RESPECTIVELY. CONTINUOUS DESCRIPTION IS IN 4 SECTIONS CONTINUED FROM THE PREVIOUS PAGE.



APPENDIX III: WHOLE ROCK PERCENTAGES FOR EACH THIN SECTION POINT COUNTED FOR CHAPTER 5.

Sample	Framework Grains			Detrital Matrix	Cement						Other	Porosity	
	Q	F	L	Clay	Qtz	Chl-orite	Carb-onate	Kao-linite	Side-rite	Opa-que		Prim	Sec.
BM 20070	50.9	2.9	4.1	6.5	2.4	0.0	27.6	0.0	0.0	5.0	0.6	0.0	0.0
BM 20074.5	41.7	3.1	4.1	37.0	9.8	0.0	1.3	0.5	0.0	0.8	1.3	0.0	0.0
BMP2 20133	53.1	1.3	2.4	0.0	4.3	0.0	37.5	0.0	0.0	0.5	0.8	0.0	0.0
BMP2 20136	43.9	1.6	3.8	38.9	3.4	0.0	3.8	0.0	1.3	0.0	3.4	0.0	0.0
BMP2 21241	54.7	8.3	1.8	0.3	19.2	6.0	1.0	2.8	0.0	1.6	1.3	3.1	0.0
BMP2 21243	53.0	7.1	1.3	1.6	16.3	8.4	2.6	4.5	0.0	1.8	1.8	1.6	0.0
BMP2 21463.5	61.9	1.8	3.9	0.0	6.5	9.2	1.2	4.5	0.0	0.3	0.6	9.8	0.3
BMP2 21466.5	61.2	0.9	0.9	0.0	10.2	0.0	0.9	4.0	0.0	0.3	9.3	12.1	0.0
BMP2 21469.5	61.8	1.8	3.3	0.0	7.0	11.2	1.8	6.1	0.0	0.0	0.0	7.0	0.0
BMP2 21471	63.1	1.8	1.3	0.0	11.9	0.0	7.0	3.9	0.0	0.3	3.4	7.5	0.0
BMP2 21473	61.3	1.1	3.4	0.0	9.3	0.0	1.1	5.8	0.0	0.5	10.1	5.0	0.0
Butler 18227	64.6	0.0	8.6	0.0	1.7	18.9	0.0	0.0	0.0	0.0	0.3	4.8	1.0
Cent 12047	45.0	0.8	12.7	5.9	2.4	0.0	25.3	0.0	0.0	1.1	5.1	1.6	0.0
Cent 12055	47.5	2.1	23.3	0.0	10.1	8.3	0.9	0.0	0.0	4.3	1.5	1.8	0.0
Cent 12062	51.6	1.3	8.4	0.0	3.9	17.7	11.6	1.0	0.0	2.3	0.0	1.9	0.3
Cent 12065.2	52.4	0.0	9.1	0.0	5.0	0.9	28.5	0.3	0.0	1.2	0.9	1.2	0.6
Cent 12066.7	40.6	0.3	5.6	0.0	5.9	2.0	43.7	0.3	0.0	0.8	0.0	0.0	0.8
Cent 12069	55.1	0.9	12.5	0.0	3.7	1.4	22.7	1.1	0.0	1.4	0.0	0.0	1.1
Cent 12071	50.9	0.0	9.9	0.0	10.2	6.8	0.5	0.8	0.0	18.3	1.0	1.6	0.0
Cent 12081	52.3	0.3	0.6	41.8	3.7	0.0	0.0	0.0	0.0	0.0	1.2	0.0	0.0

APPENDIX III CONTINUED: WHOLE ROCK PERCENTAGES FOR EACH THIN SECTION POINT COUNTED FOR CHAPTER 5.

Sample	Framework Grains			Detrital Matrix	Cement						Other	Porosity	
	Q	F	L	Clay	Qtz	Chl-orite	Carb-onate	Kao-linite	Side-rite	Opa-que		Prim	Sec.
Cent 12084	46.2	0.0	3.5	39.2	6.4	0.0	4.7	0.0	0.0	0.0	0.0	0.0	0.0
CR 19498	52.2	0.9	11.3	0.0	24.8	0.0	0.3	0.0	0.0	1.6	8.5	0.0	0.3
CR 19517	53.8	1.7	7.0	14.8	12.9	4.5	0.0	0.0	0.0	2.5	2.8	0.0	0.0
CR 19874.5	38.8	0.9	2.4	48.8	5.3	0.0	0.0	0.0	0.0	1.2	2.6	0.0	0.0
CR 19892.5	33.8	1.9	4.4	45.8	7.6	0.0	2.7	0.0	0.0	1.1	2.7	0.0	0.0
CR 19902	45.5	1.0	9.6	9.0	18.3	6.1	0.6	0.0	0.0	5.8	3.5	0.6	0.0
CR 19908	53.1	0.9	12.5	0.0	9.6	16.7	0.3	1.8	0.0	2.1	0.6	2.4	0.0
CR 19910	65.4	1.3	7.4	0.0	3.2	9.7	0.0	2.9	0.0	0.0	1.3	8.4	0.3
CR 19931.5	63.2	0.0	7.0	0.0	4.9	15.1	0.3	0.0	0.0	0.5	0.0	8.9	0.0
CR 19934.5	60.8	0.3	9.9	0.0	7.5	15.5	0.0	0.0	0.0	1.7	0.0	4.4	0.0
CR 19937	60.3	2.0	13.6	0.0	4.1	13.9	0.0	0.9	0.0	1.2	1.7	2.3	0.0
CR 19940	48.7	0.3	16.2	0.0	14.4	13.4	0.3	0.5	0.0	2.5	1.3	2.5	0.0
CR 19944	45.7	0.8	13.0	0.0	10.2	19.1	5.5	1.1	0.0	1.1	0.6	2.5	0.3
Cran 10469	64.9	0.0	10.4	0.0	1.0	19.8	0.0	0.0	0.0	0.0	0.2	2.2	1.6
IJM 20799	51.5	0.3	10.0	7.8	4.7	1.7	23.1	0.0	0.0	0.6	0.3	0.0	0.0
IJM 20801	55.5	0.3	5.2	7.2	3.7	0.3	23.9	0.3	0.0	1.4	2.3	0.0	0.0
IJM 20804	61.3	0.6	9.8	3.0	5.4	3.9	11.6	0.0	0.0	2.7	1.5	0.0	0.0
IJM 20807	52.5	0.3	12.0	8.2	7.1	2.5	13.4	0.0	0.0	3.0	1.1	0.0	0.0
IJM 20815	50.9	0.3	10.7	12.2	9.8	10.4	0.3	0.0	0.0	3.7	1.8	0.0	0.0
IJM 20817	48.4	0.3	15.1	0.9	6.5	19.6	1.2	0.0	0.0	1.8	1.2	5.0	0.0

APPENDIX III CONTINUED: WHOLE ROCK PERCENTAGES FOR EACH THIN SECTION POINT COUNTED FOR CHAPTER 5.

Sample	Framework Grains			Detrital Matrix	Cement						Other	Porosity	
	Q	F	L	Clay	Qtz	Chl-orite	Carb-onate	Kao-linite	Side-rite	Opa-que		Prim	Sec.
IJM 20818	51.7	0.3	6.0	22.7	6.0	8.8	1.9	0.0	0.0	1.6	0.6	0.3	0.0
IJM 20822.5	54.4	0.3	14.3	8.2	12.5	7.0	0.9	0.3	0.0	0.9	1.2	0.0	0.0
RO 11206	45.3	2.4	1.5	2.1	8.5	0.6	36.8	0.0	0.0	1.8	0.6	0.6	0.0
RO 11258	40.5	1.5	21.6	1.8	18.0	7.6	0.0	0.6	0.0	3.0	2.7	2.4	0.0
RO 11296	47.5	0.0	0.6	39.5	5.6	0.0	6.2	0.0	0.0	0.6	0.0	0.0	0.0
RO 11315	42.8	0.6	1.8	44.0	5.3	0.0	0.0	0.0	0.0	5.3	0.3	0.0	0.0
SN 8608	49.8	1.2	9.2	0.0	16.6	6.8	0.6	0.9	0.0	2.2	0.0	12.6	0.0
SN 8656	55.2	4.4	11.6	0.0	11.9	3.0	0.8	0.3	0.0	0.6	0.8	11.3	0.0
SN 8671	62.5	8.1	6.7	0.0	5.2	0.0	1.5	0.6	0.0	0.0	1.2	14.2	0.0
SN 8694	66.1	2.4	5.2	0.0	9.5	0.0	0.3	1.2	0.0	0.9	5.2	9.2	0.0
SN 8709	63.9	6.1	6.7	0.9	7.6	0.0	1.5	0.6	0.0	0.0	0.6	11.9	0.0
SN 8728	69.7	0.9	1.3	0.0	5.6	0.0	0.0	0.9	0.0	0.0	6.3	15.3	0.0
SN 8754	70.0	1.3	4.7	0.0	3.4	0.0	0.0	1.9	0.0	0.0	6.6	12.2	0.0
TDB 16177	66.6	0.0	7.4	0.0	1.5	15.7	0.0	0.0	0.0	0.0	0.0	8.9	0.0
TDB 16190.5	61.8	0.0	12.8	0.0	2.4	13.4	0.6	0.0	0.0	0.0	0.3	8.7	0.0
TDB 16216.5	55.9	0.8	10.7	0.0	2.5	20.3	0.3	0.0	0.0	0.6	0.0	8.8	0.0
TDB 16218.5	56.9	0.3	8.6	0.0	2.9	19.0	0.0	0.0	0.0	0.6	0.0	11.8	0.0
TDB 16235	54.1	0.6	10.9	0.0	3.8	16.5	3.2	0.0	0.0	2.1	0.0	8.8	0.0
TDB 16242	56.8	0.3	11.6	0.0	2.0	15.9	0.0	0.0	0.0	0.9	0.0	12.5	0.0
TDB 16250	61.6	0.8	5.8	0.0	3.4	20.5	0.0	0.0	0.0	0.0	0.0	7.9	0.0

APPENDIX III CONTINUED: WHOLE ROCK PERCENTAGES FOR EACH THIN SECTION POINT COUNTED FOR CHAPTER 5.

Sample	Framework Grains			Detrital Matrix	Cement						Other	Porosity	
	Q	F	L	Clay	Qtz	Chl-orite	Carb-onate	Kao-linite	Side-rite	Opa-que		Prim	Sec.
TDB 16360	54.0	0.0	3.3	34.7	7.2	0.3	0.0	0.0	0.0	0.0	0.0	0.0	0.0
TDB 16374	27.5	0.0	56.0	0.0	0.0	1.1	14.6	0.0	0.0	0.3	0.0	0.5	0.0
TDB 16383	60.1	0.0	23.3	0.0	3.9	7.5	0.0	0.0	0.0	0.0	0.0	5.3	0.0
TDB 16393.5	42.2	0.0	38.1	0.0	3.8	0.6	11.5	0.0	0.0	0.6	2.7	0.6	0.0
TDB 16395	63.4	0.0	9.1	0.0	5.0	10.5	0.0	0.0	0.0	0.0	0.3	11.8	0.0
TDB 16400	62.7	1.0	7.2	0.0	1.6	20.9	0.0	0.0	0.0	0.7	0.0	5.9	0.0
TDB 16405.5	66.9	0.0	11.2	0.0	4.0	10.7	0.0	0.0	0.0	0.3	0.0	6.9	0.0
Thom 17134	48.3	0.8	4.5	0.0	4.3	0.0	41.6	0.0	0.0	0.3	0.3	0.0	0.0
Thom 17135.5	50.3	0.6	8.8	0.0	4.1	3.5	22.0	0.0	0.0	1.9	5.3	0.9	2.5
Thom 17140	59.3	0.3	3.7	0.0	6.6	4.5	23.9	0.3	0.0	1.3	0.0	0.0	0.0
Thom 17144	62.6	0.3	11.7	0.3	12.3	9.4	0.0	0.0	0.0	2.6	0.6	0.3	0.0
Thom 17154	42.0	0.2	6.1	22.7	9.8	3.5	12.9	0.0	0.0	1.8	0.8	0.0	0.0
Thom 17208	51.8	0.8	9.0	9.8	6.5	2.8	17.0	0.0	0.0	2.0	0.3	0.3	0.0
Thom 17211.5	40.6	0.0	4.8	43.3	3.9	0.0	6.3	0.0	0.0	1.2	0.0	0.0	0.0
Thom 17235	63.9	1.8	5.4	0.0	13.6	5.4	3.3	0.9	0.0	3.6	1.8	0.3	0.0
Thom 17240	60.7	1.1	5.0	6.1	10.6	2.7	8.0	0.5	0.0	2.1	2.4	0.8	0.0
Thom 17240.5	61.2	2.0	0.9	0.0	12.2	1.5	21.9	0.0	0.0	0.3	0.0	0.0	0.0
Thom 17252	51.7	0.0	11.4	10.0	12.0	8.3	2.9	0.0	0.0	3.1	0.3	0.3	0.0
Thom 17253.5	45.2	1.1	5.9	0.0	9.1	0.5	32.5	0.0	0.0	5.4	0.3	0.0	0.0
Thom 17257	40.9	0.6	5.5	5.5	7.0	2.9	32.5	0.0	0.0	3.5	0.3	0.6	0.0

APPENDIX III CONTINUED: WHOLE ROCK PERCENTAGES FOR EACH THIN SECTION POINT COUNTED FOR CHAPTER 5.

Sample	Framework Grains			Detrital Matrix	Cement						Other	Porosity	
	Q	F	L	Clay	Qtz	Chl-orite	Carb-onate	Kao-linite	Side-rite	Opa-que		Prim	Sec.
Thom 17263	44.1	0.3	2.8	47.5	3.7	0.0	0.6	0.0	0.0	0.8	0.3	0.0	0.0
Thom 17266	58.6	0.3	6.0	0.0	13.4	1.9	18.1	0.0	0.0	1.4	0.3	0.0	0.0
TJP 10388	56.6	0.6	8.5	0.0	12.9	3.5	0.3	1.5	0.0	0.6	0.6	11.7	3.2
TJP 10391	60.4	2.7	8.0	0.0	11.8	4.9	0.8	3.3	0.0	0.5	1.1	5.5	0.8
TJP 10392	63.9	1.5	8.0	0.0	10.4	3.1	0.0	1.2	0.0	0.6	0.6	9.5	1.2
TJP 10396	58.6	1.6	9.4	1.0	6.1	9.1	0.6	0.0	0.0	0.3	0.6	10.0	1.0
TJP 10398	60.7	1.4	4.8	0.0	17.8	4.0	0.0	1.1	0.0	0.6	0.3	8.2	1.1
TJP 10400.5	59.2	3.6	0.9	0.0	3.6	8.7	2.4	0.0	11.7	0.0	0.9	9.0	0.0
TJP 10401.5	35.2	0.0	1.1	0.0	4.5	0.8	41.5	0.3	2.4	0.0	7.7	5.0	1.6
TJP 10404	43.8	1.2	2.1	0.0	10.8	2.7	0.6	0.0	28.2	0.0	4.8	4.5	1.2
TJP 10406	42.3	0.8	1.5	14.1	10.3	3.3	3.8	0.0	9.6	1.8	6.8	3.8	1.5
TJP 10428	48.0	0.3	2.3	0.0	31.3	16.8	0.3	0.0	0.0	0.0	1.1	0.0	0.0
TJP 10439	23.8	0.0	1.0	0.0	2.8	0.0	12.3	0.0	26.1	4.5	29.3	0.0	0.3
TJP 10443	36.2	0.3	1.8	0.0	8.7	1.3	2.9	0.0	8.7	1.8	37.3	0.0	1.0
TJP 10448	61.8	1.4	5.6	0.0	17.5	1.7	3.6	1.4	0.0	2.2	1.4	1.7	1.7
TJP 10453	54.5	2.0	7.3	3.5	18.4	3.8	0.3	1.2	0.0	4.4	0.6	1.5	2.3
TJP 10455.5	40.7	0.3	3.0	48.2	5.7	0.0	0.0	0.0	0.0	0.8	1.3	0.0	0.0
WAL 20058	57.6	0.0	9.8	0.0	22.5	5.9	0.3	0.0	0.0	2.8	0.0	0.0	1.1
WAL 20063	55.8	0.3	5.9	1.3	7.5	1.8	22.7	0.0	0.0	3.6	0.5	0.0	0.5
WAL 20376	53.7	3.1	2.0	0.0	21.9	9.4	7.1	0.0	0.0	1.1	1.7	0.0	0.0

APPENDIX III CONTINUED: WHOLE ROCK PERCENTAGES FOR EACH THIN SECTION POINT COUNTED FOR CHAPTER 5.

Sample	Framework Grains			Detrital Matrix	Cement						Other	Porosity	
	Q	F	L	Clay	Qtz	Chl-orite	Carb-onate	Kao-linite	Side-rite	Opa-que		Prim	Sec.
WAL 20409	63.4	0.6	5.1	0.0	7.9	12.4	0.0	0.3	0.0	1.8	0.6	7.6	0.3
WAL 20435	59.8	1.5	8.0	0.0	6.5	14.6	0.6	0.0	0.0	1.2	0.0	7.7	0.0
WAL 20441	55.9	0.8	10.5	0.0	3.5	4.3	23.6	0.0	0.0	1.3	0.0	0.0	0.3
WAL 20443	55.4	1.0	6.8	0.0	4.5	3.9	27.3	0.0	0.0	0.8	0.3	0.0	0.0
WAL 20462.5	55.6	0.7	6.7	0.0	30.1	4.2	0.2	0.0	0.0	1.0	1.5	0.0	0.0
WAL 20469	57.6	1.3	7.2	0.0	21.8	6.4	0.8	0.0	0.0	2.1	0.5	0.5	1.9
WAL 20481	40.3	0.9	4.0	0.3	9.8	0.9	30.3	0.0	0.0	6.1	7.5	0.0	0.0
WAL 20512	61.4	0.0	3.8	0.0	3.3	6.6	22.7	0.0	0.0	2.2	0.0	0.0	0.0
WAL 20514	52.4	0.8	6.1	0.0	6.1	7.8	21.1	0.0	0.0	4.8	0.0	0.0	0.8
WAL 20516	56.3	0.2	9.0	0.0	4.4	5.8	19.7	0.0	0.0	3.6	0.0	0.0	1.0
WAL 20526	54.9	0.5	7.6	0.0	10.8	16.6	0.5	0.0	0.0	5.8	1.3	1.8	0.3
WAL 20531	59.6	0.0	8.9	0.0	10.0	14.4	1.1	0.0	0.0	3.9	0.8	1.1	0.3
WAL 20535	61.7	0.3	10.1	1.7	12.1	5.8	0.3	0.0	0.0	7.2	0.9	0.0	0.0
WAL 20571	61.2	0.3	9.6	0.0	4.1	12.0	1.1	0.0	0.0	1.1	0.8	9.0	0.8
WAL 20575	58.2	0.0	9.7	0.0	5.9	15.1	0.5	0.0	0.0	2.7	0.3	6.7	0.8
WAL 20577	56.6	2.1	10.3	0.0	5.6	14.7	0.0	0.0	0.0	1.5	0.6	8.6	0.0
WAL 20614	57.0	0.3	7.8	0.0	18.2	11.5	0.0	0.0	0.0	5.1	0.3	0.0	0.0
WAL 20616	56.1	0.0	4.3	0.0	3.2	8.4	23.7	0.0	0.0	3.5	0.3	0.0	0.5
WAL 20622	55.1	0.3	6.3	0.0	15.9	17.0	0.3	0.0	0.0	5.1	0.0	0.0	0.0

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