

PALEONTOLOGY, PALEOECOLOGY AND DEPOSITIONAL ENVIRONMENT
OF THE LOWER EAGLE FORD GROUP IN NORTH CENTRAL TEXAS

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DEDICATION

To my husband Howard and my daughter Joanna
for their unconditional patience and support.

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OF THE LOWER EAGLE FORD GROUP IN NORTH CENTRAL TEXAS

by

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THESIS

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Palaeontology begins and finishes in the field

September 2004

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OF THE LOWER EAGLE FORD GROUP IN NORTH CENTRAL TEXAS

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The University of Texas at Dallas, 2004

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The Eagle Ford Group in North Central Texas has been the subject of geological research for nearly a century. Nevertheless, little has been said about the paleontology, paleoecology and the depositional environment of the Britton Formation and the underlying Tarrant beds, the lowest member of the Eagle Ford Group. The sole assignment of this lowest member to the Eagle Ford Group or to the underlying Woodbine Group is to this date controversial. The focus of this research is to evaluate the impact of the Cenomanian-Turonian Oceanic Anoxic Event (OAE) on the paleontology and paleoecology of the vertebrate and invertebrate assemblage, as well as the depositional environment of the lower Eagle Ford Group of North Central Texas during this global event. The outcrops chosen for this research are almost all newly recognized and are geographically located in the Dallas-Fort Worth Metroplex as follows: Mansfield (Tarrant County, Ellis County), Dallas-Fort Worth International Airport (Tarrant County), Grand Prairie (Dallas County, Ellis County), and Alvarado (Johnson County).

The outcrops yielded an abundant and diverse ichthyofauna (*Cretoxyrhina*, *Squalicorax*, *Cretolamna*, *Enchodus*, *Ptychodus*, *Protosphyraena*, *Xiphactinus*), as well as the oldest record of saurodontid fish in North America. Reptile remains have also been found

(plesiosaurs, turtles and the rare and enigmatic *Coniasaurus*). Of special interest is the rich vertebrate ichnofauna. Vertebrate coprolites in an outstanding state of preservation are found in almost all the outcrops of the research area. Most of these coprolites contain inclusions. They are attributed to sharks and/or other large fishes on the basis of their morphology, size, inclusion content, and due to the fact that after coprolites shark and other fish teeth, vertebrae and other remains, are the most abundant body fossils in these lower Eagle Ford new localities. Coprolite horizons possess potential application not only in paleoecological reconstruction but in the application of biostratigraphic correlation, which is also one of the main focuses of this thesis.

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CHAPTER 1

INTRODUCTION

In the late 1800s R.T. Hill established the stratigraphic framework of the Cretaceous in Texas. Cretaceous age strata are the stratigraphic backbone of the state, forming a broad band at places more than 320 km wide that trends across southwest, central, and north central Texas and includes most of the big cities in the state except Houston (Fig.1.1). In north Texas, the Cretaceous sequence is more than 1,220 m thick and is divided into two parts: Lower Cretaceous strata on the northwest and Upper Cretaceous strata on the southeast. These rocks have contributed to the Texas economy since early days. The rich soil that developed along the eastern outcrop of these strata has provided bumper crops of cotton, corn, and wheat since the late 1890s. Significant quantities of oil and gas have been extracted from Upper Cretaceous rocks in the large East Texas oil field as well as from the south Texas fields at Giddings and Pearsall. Cretaceous rocks in some areas are also a major source of artesian water, brick clay, building stone, and lime for cement (Reaser 2002). The rocks exposed in the Dallas-Fort Worth Metroplex are all sedimentary rocks. No metamorphic or igneous rocks crop out in the area. Bedrock strata were deposited during the Cretaceous Period, which extended from 140 to 64 million years before the present. The Cretaceous was characterized by warm climatic conditions and extensive flooding of the North American continent by a shallow epicontinental sea. Texas was located on the southeastern margin of this seaway. During the 75 million year Cretaceous interval about 1,280 m of sedimentary rock were deposited in the Metroplex vicinity. This sequence

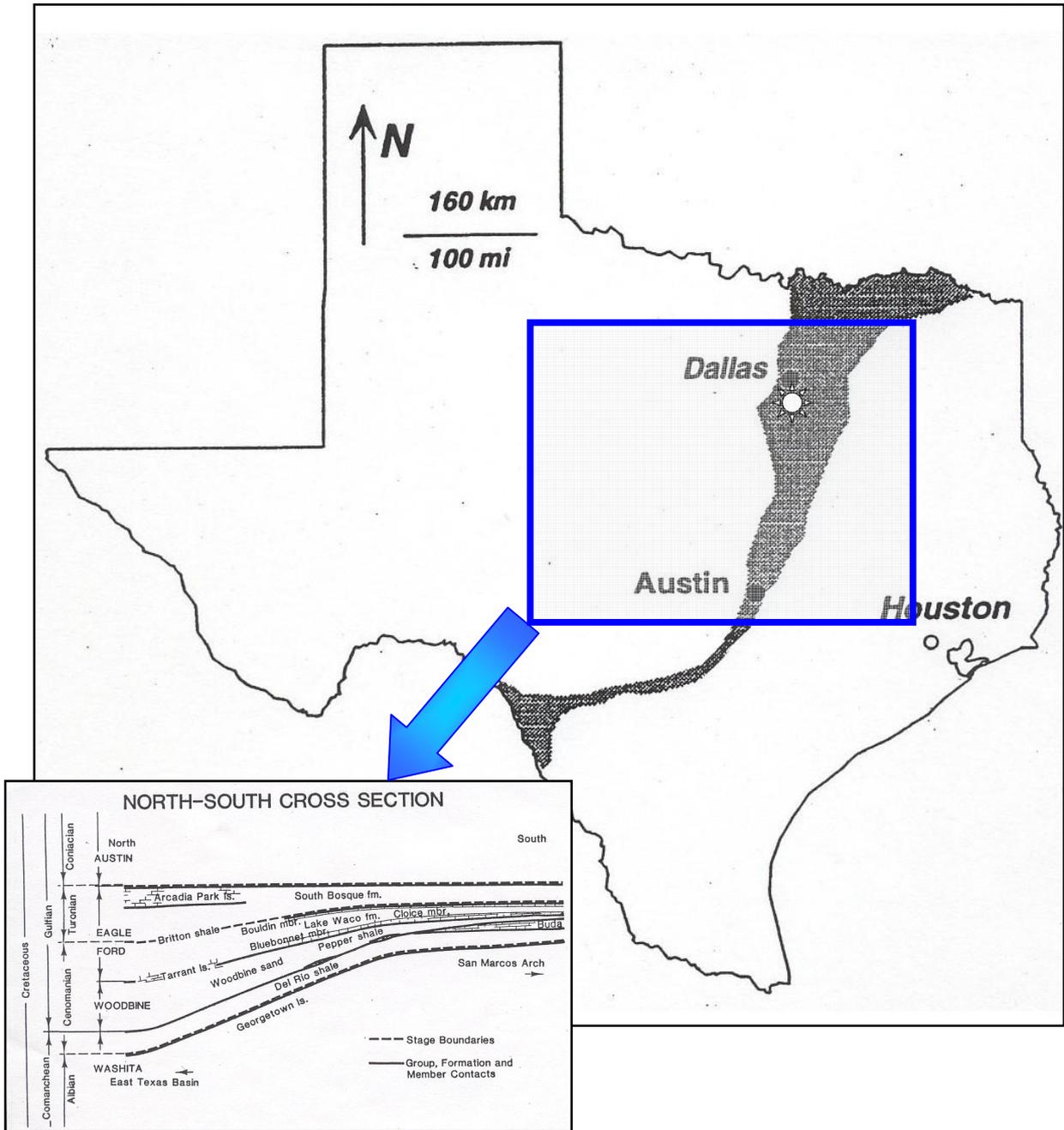


Figure 1.1. Map showing Cretaceous outcrops in Texas (Mod. after Dawson, 1997).
 Cross section (after Adkins and Lozo, 1951).

consists of about 60 percent limestone and marl, 30 percent conglomerate and sandstone, and 10 percent argillaceous rocks. Lower Cretaceous strata, deposited during the Berriasian to Albian European stages, crop out in the western part of the Metroplex. Upper Cretaceous rocks, deposited during the Cenomanian to Maastrichtian European stages, are exposed in the eastern part of the Metroplex. The local Cretaceous stratigraphic sequence includes both quartzose clastic and bioclastic or chemical rocks as well as a variety of mixed clay-carbonate rocks. The quartzose rocks include conglomerate, sandstone, shale and mudstone. The bioclastic/chemical rocks include different varieties of limestone. These limestones are dominant in the Cretaceous stratigraphic sequence in north Texas and reflect deposition in a semi-tropical to tropical region; warm shallow water, agitated by wave action that drives off carbon dioxide and promotes production of carbonate material (Reaser, 2002).

The most prominent topographic feature in the Metroplex is the White Rock Escarpment, a well-defined cuesta formed by the indurated chalky limestone of the Austin Group. This west facing bluff extends almost north-south through the western part of Dallas County, reaching its maximum elevation about 259 m above sea level near Cedar Hill. The cuesta has an average relief of about 91.4 m. The valley to the west is formed in the less resistant Britton and Arcadia Park Formations of the Eagle Ford Group (Smith, 1981). The scarp is a product of differential erosion at the Austin-Eagle Ford contact, and it's an extensive sinuous feature severely dissected by stream action (Meier, 1964).

The Cretaceous rocks of the Metroplex unconformably overlie the Pennsylvanian Mineral Wells Formation of the Strawn Group throughout the western part of the area; on the east, older Paleozoic rocks are present below the Cretaceous sequence. Possible Jurassic strata (Cotton Valley) have been reported in the subsurface at some places in the extreme

southeast part of the area. Lower Cretaceous rocks in the Metroplex were deposited on a broad low-relief surface designated as the Wichita paleoplain by Hill (1901). This feature was a sloping plateau developed on planed-off Paleozoic rocks. Extensive erosion degraded mountains down to low rolling hills or plains during a 100-million year interval. Marine Triassic and Jurassic rocks, if deposited in the area, were removed during this immense interval of time (Reaser, 2002). Cenozoic rocks cover more than 25 percent of the Metroplex and consist both of Pleistocene terrace deposits and Holocene alluvium. The combined thickness of the terrace deposits is generally less than 30 m at most places. Individual terrace levels range up to 55 m above the present thalweg of the Trinity River (Reaser, 2002).

According to Adkins (1933) ancient seas advanced northward from the south and east across the low-lying Wichita surface. In early Jurassic time the Gulf sea extended northward across northern Mexico to the Rio Grande. During Early and middle Cretaceous times the sea spread over the remainder of Texas and in Late Cretaceous time it transgressed further north to join the southern extent of the Arctic sea. Thus, the area became a part of the extensive Western Interior Sea that connected the Arctic Ocean and the Gulf of Mexico (Fig.1.2). The Cretaceous rocks in Texas are divided into two series, the older Comanche Series and the younger Gulf Series. There is an older Cretaceous Series (Coahuila Series) but this is found only in the subsurface in Texas and outcrops in Mexico. The age of the Comanche-Gulf boundary is mid-Cenomanian (Smith et al., 2000). In the Metroplex the Upper Cretaceous rock column consists of eight units that range from Cenomanian to Maastrichtian. This sequence, approximately 610 m thick, accumulated during a 32-35

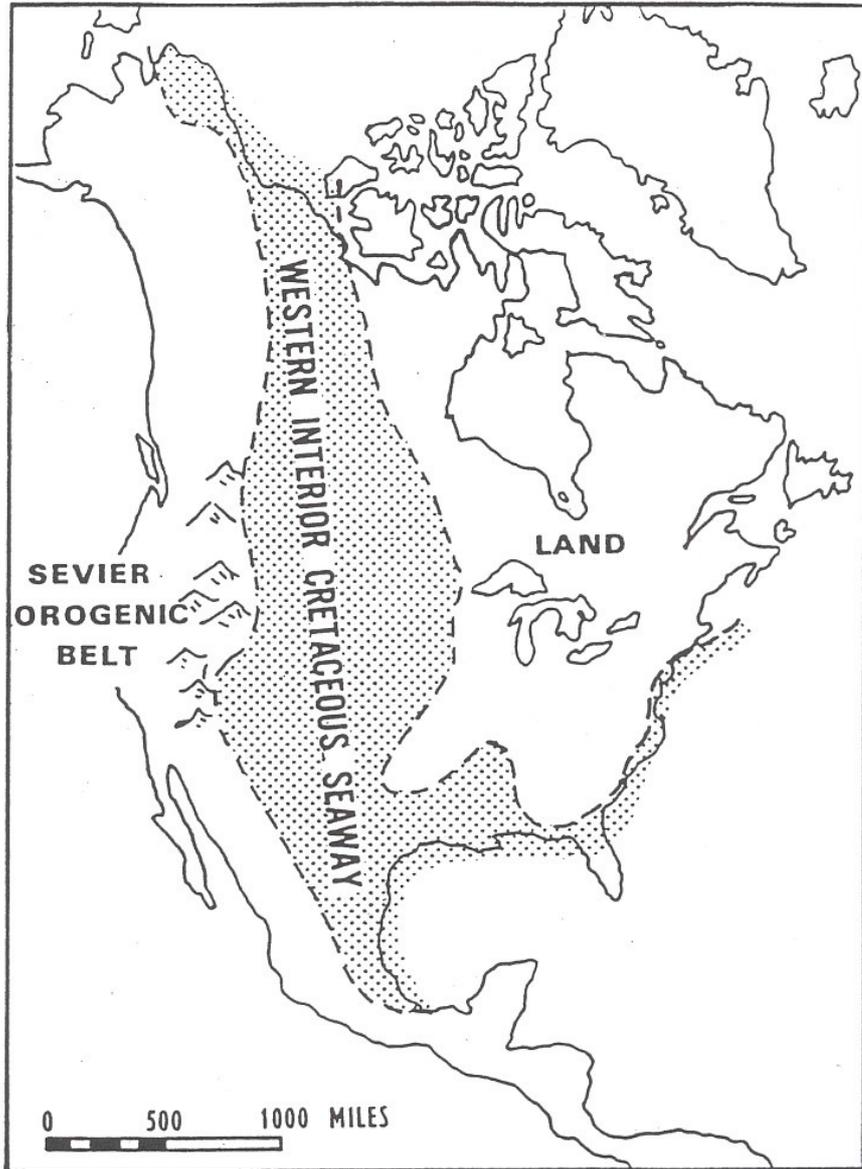


Figure 1.2. Generalized map of the Western Interior Seaway showing its extension at maximum transgression (Mod. after Kauffman, 1977a).

million-year interval. The Cretaceous-Tertiary boundary is not exposed in the area. The nearest exposure of the contact is a short distance east of Terrell in Kaufman County, TX. (Reaser, 2002).

CHAPTER 2

GEOLOGIC SETTING

The Eagle Ford Group is very extensive in Texas, forming an outcrop across the entire state from the Red River to the Rio Grande. The thickness varies at different localities, reaching a maximum of 152 to 183 m in North Texas between Dallas and Sherman, while in Austin the thickness is only about 15 m. West of Austin the thickness increases to 60 to 90 m in the Rio Grande region. The decreasing thickness southwards from the Red River is probably due to a gradual disappearance in that direction of the lower beds until, at Austin, only the upper part of the group is represented. The earliest mention of the term “Eagle Ford” in geologic literature was by Ferdinand Roemer. In 1847, he included “black Eagle Ford shales with fish remains” in the New Braunfels area. The name “Eagle Ford shales” was formally applied to these strata by R.T. Hill in 1887. He derived the name from the village of Eagle Ford along the Texas and Pacific Railway about 6 miles west of Dallas where the upper part of the strata are typically exposed (150 m). The lithologic character of the Eagle Ford is unlike at different levels. Calcareous sandstone, siltstone, calcareous shale, and flaggy limestone overlying the Woodbine Group and conformably or disconformably overlain by the Austin Chalk are recognized as the Eagle Ford Group (Smith, 1981). The Eagle Ford shales are an important source of petroleum.

In 1927, Moreman divided the Eagle Ford into three formations. In ascending order these units are the Tarrant, Britton and Arcadia Park Formations. In 1951, Adkins and Lozo presented a north-south diagrammatic section of Eagle Ford rocks showing lithologic units

and zonal stages as shown in Fig.2.1.

Dawson (2000) divided the Eagle Ford Group into two regional depositional units: an older transgressive and “partly condensed” sequence and a younger highstand sequence. These two depositional units have distinctive sedimentological and geochemical characteristics.

Another division of the Eagle Ford was made by Turner (1951) who subdivided the formation on the basis of calcium carbonate content into two informal units -a lower clayey marl and an upper clay. The first one contains more than 25% calcium carbonate by weight and includes a 7 m section of white or yellowish laminated marl at the top and an underlying medium gray compact shale. Turner was unable to establish a lower limit to the unit. This clayey marl grades upward and laterally into the clay unit which is described as brownish gray to black with a calcium carbonate content under 25% (Meier, 1964).

The Eagle Ford Group is also subdivided into lower Eagle Ford (Tarrant and Britton Formations) and upper Eagle Ford (Arcadia Park Formation). The dip of the Eagle Ford formations varies slightly, but is for the most part approximately 2 degrees southeast (Winton and Adkins, 1919). The Eaglefordian sequence has been interpreted to range from late Cenomanian (94 Myr), to mid-Turonian (89.5 Myr). Stratigraphically in outcrop, the Eaglefordian sequence extends from the unconformity, at the base of the Eagle Ford Group (Tarrant Formation), to the unconformity at the base of the Arcadia Park Formation in the Dallas County area (Phillips, 1987).

The Eagle Ford shale is a very unstable material due to its geo-technical properties (BGS, 1978). Injudicious residential and commercial construction on the Eagle Ford shale will almost certainly result in failures as has been seen in countless cases in the Metroplex.

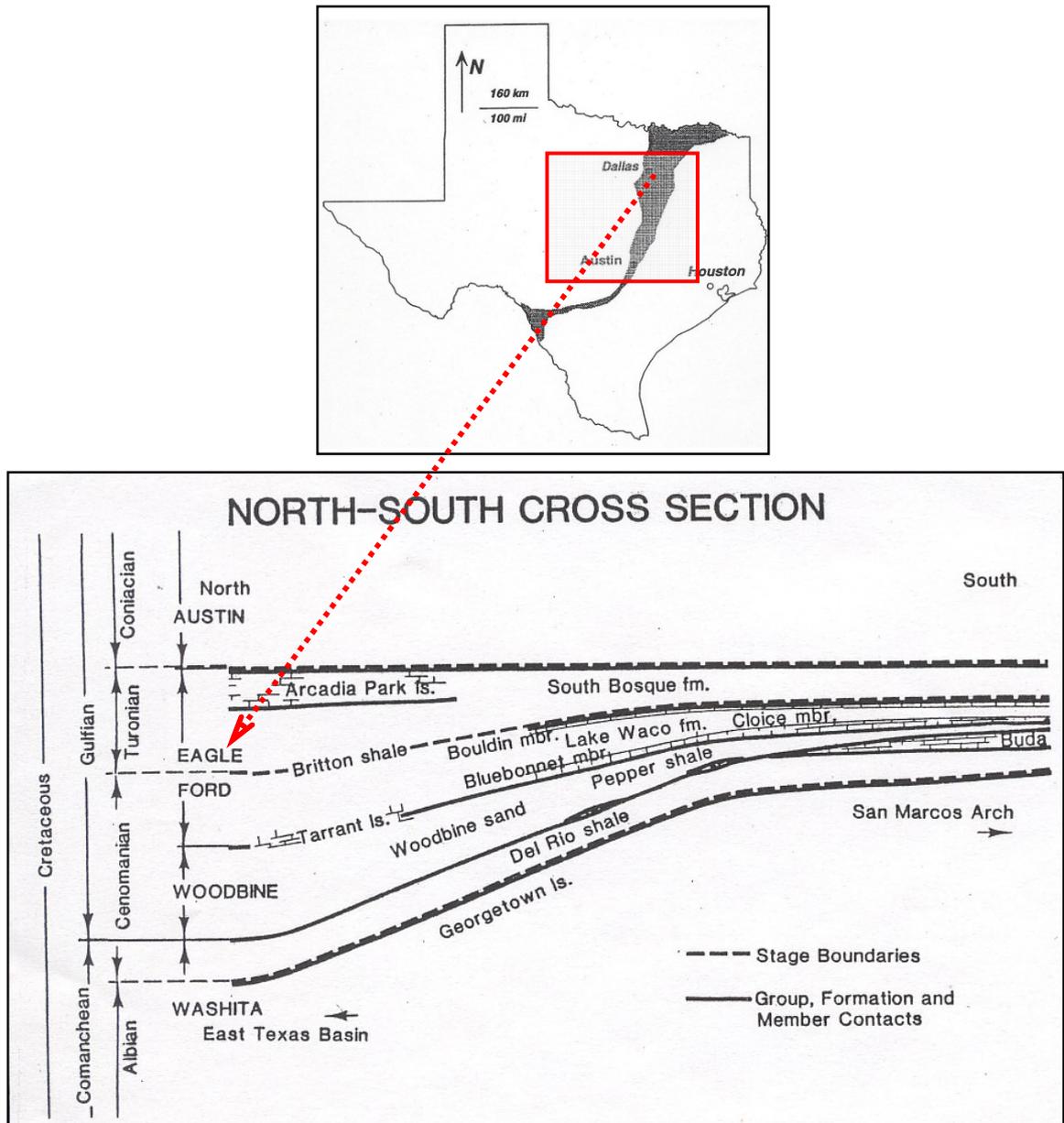


Figure 2.1 North-south diagrammatic section of Eagle Ford rocks, showing lithologic units and zonal stages (after Adkins and Lozo, 1951). Map showing Cretaceous outcrops in Texas (after Dawson, 1997).

TARRANT FORMATION

Beds between typical sandstones of the Woodbine and typical shales of the Britton Formation of the Eagle Ford Group in Tarrant County were described by Moreman (1933) as the “Tarrant sandy clay and limestone” and as the “Tarrant Formation” by the same author in 1942, when he regarded them as the basal unit of the Eagle Ford Group. Norton (1965) and Powell (1968) also classified it in the Eagle Ford Group. Stephenson (1952) believed the Tarrant to be part of the Lewisville Member of the Woodbine.

The idea that the basal Eagle Ford may at places be the time equivalent of the Upper Woodbine is not new. In northeastern Tarrant County as well as other places, the basal Eagle Ford Group is represented by the Tarrant beds. These beds are unlike the basal Eagle Ford as developed elsewhere and are of limited thickness (aprox. 5 m) and geographic extent. Jarvis (1948) attempted to show by means of surface exposure that the Tarrant beds are merely a special development of the uppermost Woodbine. The Tarrant beds are described as being 5 m of sandy clays with intermittent limestone strata and having at the base a thin bed of conglomeratic material containing phosphatic pebbles and fish remains. The lower-most unit of the Eagle Ford, the Tarrant sandy clay, consists of 5 to 7 m of gray to brownish-gray calcareous sandstone interbedded with brown siltstone, brownish limestone and shale (Surles, 1987).

The precise boundary between the Woodbine and Eagle Ford in the Metroplex vicinity has been questionable for at least 70 years. The focus of the controversy is the stratigraphic position and lithic associations of these Tarrant beds, a 5 m sequence that consists of gray sandy clay some thin brown limestone, and calcareous concretions that yielded the ammonites *Acanthoceras tarrantense* (Adkins) and *Metengoceras dumbli* (Cragin). The

type section was in Tarrant County about 1.6 km east Tarrant Station at the crossing of the St. Louis, San Francisco and Texas railroad over Bear Creek (Adkins, 1933).

Kauffman and others (1977) designated the Tarrant as a member of the Eagle Ford Formation in north central Texas and indicated that it was partly equivalent to Woodbine. Reaser (2002) included the Tarrant strata in the upper Woodbine. Others (i.e. Hensleigh, 1983) have also placed the Tarrant in this stratigraphic position. Jarvis (1948) and other geologists concluded that the 5 m Tarrant sequence was a “special development in the uppermost Woodbine.”

In any event, the lithic relationships between the Woodbine and Eagle Ford in the Metroplex involve a complex interplay of nearshore continental and marine environments. The exact relationships are yet to be solved (Reaser, 2002).

The Tarrant is a valid, mappable rock-stratigraphic unit in northeastern Tarrant County, Texas. It represents sediments deposited in a near-shore marine environment and is part of the overall early Gulfian marine transgression in North Texas. The member is wholly contained within the range of *Acanthoceras wintoni* Adkins. The Tarrant Formation is a sequence of beds that represent the marine phase of the transition from non-marine and paralic Woodbine environments to younger offshore marine Eagle Ford environments.

This transition is marked by lithologic variation, both vertically and along strike, spanning a stratal thickness of from 0.45 to 2.13 m. Petrographic studies indicate that this member, though mapped with the Eagle Ford Formation, is genetically more closely related to the Woodbine Formation and represents a complex tongue of the Woodbine limestone (Powell, 1968). The same author regarded the boundary between the Woodbine and the Eagle Ford

as a definite contact and that it was nearly always selected at the top of the youngest ferruginous sandstone with *Crassostrea soleniscus* (Powell, 1968).

In cores from both McLennan and Dallas counties (Socony Mobil Field Research Laboratory coreholes numbers 16 and 18), the Woodbine-Eagle Ford contact is also reported to be unconformable (Brown and Pierce, 1962; Pessagno, 1969; Smith 1981; Phillips, 1987). It is represented in the cores by a reworked zone of lignitic mudstone pebbles, borings, glauconite and black phosphate pebbles (Brown and Pierce, 1962). These distinctive characteristics suggest that where it is present, the base of Tarrant Formation marks the Woodbian-Eaglefordian sequence boundary (Phillips, 1987). No planktonic foraminifera were recovered from the Tarrant Formation in the Dallas County core to document its age. However, palynomorph assemblages (Brown and Pierce, 1962) and megafossil (ammonite) evidence indicate that the Tarrant is in fact late Cenomanian (early Eaglefordian) in age (Pessagno, 1969).

A locality in Bear Creek is stated to show the contact between the Woodbine and the basal Eagle Ford shales, but this contact has always been controversial (Norton, 1965).

BRITTON FORMATION

The Britton is the thickest formation within the Eagle Ford Group. It was named for exposures in the vicinity of Britton, a small community in northwest Ellis County. The formation crops out along north-south line from Grapevine in Denton County to Mansfield in southern Tarrant County. It underlies most of DFW International Airport and surrounds the west side of Joe Pool Lake. The type section is along Newton Branch, a tributary of Mountain Creek approximately 6 km south of Britton (Reaser 2002). Bentonites are concentrated in the Britton, where at least 34 bentonite seams have been reported within

21 m of strata (Brown and Pierce 1962).

In 1970, Powell informally divided the Britton in the Dallas area into a lower calcareous (chalky) unit and an upper noncalcareous unit separated by a thin (0.3 - 0.6m) “transition zone”. Later Powell and Reaser (2001) formally subdivided the 112 m formation into two members: The lower Turner Park equivalent to the calcareous unit and the upper Camp Wisdom equivalent to the noncalcareous unit (Reaser, 2002).

TURNER PARK MEMBER

The Turner Park consists mostly of 36.6 m of dark gray to light-yellowish brown calcareous shale with numerous bentonite seams that range in thickness from a few millimeters to more than 5 cm. The bentonites were formed by a succession of volcanic eruptions during early Britton deposition. According to Charvat (1985) the sources of these ash deposits include “Arkansas, west Texas, and the western interior area”. The lower member is named after rocks exposed at places near Turner Park in Grand Prairie, Texas. The type section is designated as “a north-facing bluff along an abandoned meander of the West Fork of the Trinity River, about 305 m southeast of the intersection of Interstate Highway 30 and Belt Line Road” (Powell and Reaser 2001). Approximately 12 m of shale and thin bentonite seams are exposed. The age of the Turner Park is late Cenomanian (Reaser, 2002).

CAMP WISDOM MEMBER

The Camp Wisdom is composed mostly of about 76 m of olive gray shale which weathers to olive brown to glistening “blue-gray flakes on bare exposures” (Powell and Reaser, 2001). The member is named after a “Boy Scouts of America camp” of that name near the headwaters of Artesian Creek and on top of the White Rock escarpment” (Powell and

Reaser, 2001). The type section of the member is designated as an exposure along the north side of Mansfield Road just south of Lake Joe Pool (locality 25 of Powell and Reaser, 2001). According to these authors the member is distinguished in outcrop by the “ocherish, calcareous clay-ironstone nodules and gray septarian concretions in dark olive gray shale”. The age of the Camp Wisdom ranges from late Cenomanian to earliest Turonian (Reaser, 2002). The Dallas Geological Society (Norton, 1965) informally divided the Britton sequence into nine informal units, as follows:

LOWER BRITTON (Bentonitic Sub-Member). The bentonite-bearing beds of the Britton are divided into two identifiable units, Unit 1 and Unit 2.

UNIT 1

The main and lowermost bentonitic portion of the Eagle Ford in Dallas County consists of about 31 m of clay shale with interbedded very thin limestone beds and bentonite seams. A prominent bentonite nearly 30 cm in thickness is found. This bed is well exposed at Turner Park and at other areas in and near Grand Prairie. The bentonitic clays weather normally to a light tan or cream color. The unit is characterized by numerous bentonite seams—estimated to be 69 or more by the Dallas Geological Society (1965). The lack of fossils matter in the bentonites themselves is an interesting and unresolved problem.

UNIT 2

This 6 m unit consists of light brown clay interbedded with white-weathering limestone and thin bentonite seams. The limestone yields the small clam *Inoceramus labiatus*. The unit thins markedly from southern Dallas County into northern Ellis County. The clays normally weather to the light tan or cream color characteristic of Unit 1.

Units 1 and 2 combined make up the lithologic unit here recognized as the lower (bentonitic)

sub-member of the Britton member of the Eagle Ford.

UPPER BRITTON SUB-MEMBER

UNIT 3

The interval next above the lower (bentonitic) sub-member of the Britton marks a radical change in sedimentation, inaugurating the concretion-bearing darker clays of the upper Britton. The unit consists of 12 to 17 m dark clay shale with brown to red clay-ironstone nodules. The unit contains the ammonites *Metoicoceras irwini* and *Sciponoceras (Baculites) gracilis*.

UNIT 4

The succeeding lithic unit, easily recognized in the section, is one of thin sandstone beds and streaks of some 3 m in thickness. The sandstone is fine grained, partly micaceous, and very finely cross bedded. It is pitted on a weathered surface. Prominent beds mark the top and bottom of the unit. The member has yielded the coiled ammonites, *Mentengoceras irwini acutum* and *Proplacentoceras* sp.

UNIT 5

Above the sandstone beds of unit 4 more shales occur. Unit 5 is composed of 8 m of shale interbedded with ferruginous and calcareous concretions. The unit is capped by rough-weathering, golden yellow limestone concretions. Shale beneath these concretions contains small concretions with possible dasycladacean algal material. The sequence contains the following mollusks: *Allocrioceras pariense*, *Inoceramus* sp., *Metoicoeras whitei*, *Proplacentoceras* sp. and *Sciponiceras (Baculites) gracilis*. Decapod crustaceans: *Notopocotystes dichrous*, etc.

UNIT 6

Unit 6 consists of from 2 m to 2.5m of clay with a distinctive pock-marked light gray to white limestone concretion bed that serves as a marker bed. At some places a thin sandstone occurs beneath the concretion bed. The concretion bed has yielded *Proplacenticerias stantoni* bolli, *Metoicoeras whitei* and *Inoceramus* sp.

UNIT 7

Unit 7 is composed of from 15-23 m of clay shale interbedded with both calcareous and ferruginous fossiliferous concretions. The massive-yellowish weathering septarian limestone concretions are characteristic of the unit and have yielded well preserved ammonites including some *Metoicoceras whitei* and *Baculites* sp. with a distinctive nacreous luster. The top of the unit was arbitrarily placed at the highest occurrence of “red-brown clay ironstone concretions”

UNIT 8

Unit 8 consists of from 8 to 12 m of dark shale with large septarian limestone concretions. Some of these massive bodies are 1 m in diameter. Fossils are not plentiful in the large concretions, but there are some very large ammonites represented as well as *Inoceramus* sp. and occasional shark teeth in the surrounding shales.

UNIT 9

Unit 9 is the highest unit assigned to the Britton Member of the Ragle Ford, is largely a shale interval between the uppermost large septarian concretion horizon of unit 8 and the base of the Kamp Ranch limestone. This unit consists mostly of dark shale with small, flat calcareous concretions, peculiar log-like concretions, and thin fossil-bearing limestone

lenses. This unit is well exposed along the old Mansfield Road adjacent to Joe Pool Lake in Cedar Hill State Park along FR1382. According to the Dallas Geological Society (1965) a large accumulation of secondary selenite occurs in the unit from 4.5 to 6 m below the Kamp Ranch beds in the southern part of Denton County. The unit contains *Inoceramus labiatus* and a few unidentified ammonite fragments. Numerous worm tubes *Serpula* sp. are attached to *Inoceramus* prisms at some places. A variety of crustaceans is known from the Britton Formation: *Cenomanocarcinus vanstraeleni*, *Enoploclyia* sp. *Homerus brittonestrus*, *Linuparus grimmeri*, and *Necrocarcinus ovalis*. There is some evidence suggesting an unconformity between the top of the Britton Formation and the Arcadia Park Formation.

ARCADIA PARK FORMATION

The Arcadia Park is the upper formation of the Eagle Ford Group. It was named for exposures in the Arcadia Park area in west Dallas. The formation consists of approximately 37 m of dark olive-gray calcareous shales with some beds of flaggy limestone in the upper and lower parts (Kamp Ranch Member). Calcareous septarian concretions occur about 12 m and 27 m above the base. The minerals gypsum, jarosite and pyrite occur within the shale sequence. This shale yielded in the past a wealth of ammonites from quarries in west Dallas. The age of the Arcadia Park is Turonian (Powell et al., 1970).

KAMP RANCH MEMBER

Limestone flags near the base of the Arcadia Park have been designated as "Arcadia Park flagstone" or "hard calcarenites" by several workers (Powell et al., 1970). These beds were named Kamp Ranch after the Kamp Ranch in Dallas County by Meier (1964). The Kamp Ranch is a yellow, sandy, bioclastic limestone that forms an irregular but very distinctive and easily identifiable marker bed in the upper Eagle Ford. The sequence consists of

approximately 1 m of flaggy conquinoid limestone interbedded with thinner flags and calcareous shale. The bed is almost 28 m below the Arcadia Park -Austin Chalk in Dallas County and only about two meters above the underlying Britton Formation. The prominent light brown limestone yields abundant oyster fragments (*Ostrea* sp.) and small coiled ammonite impressions (*Prionotropis* sp.). Reid (1952) described the unit at Dallas in detail, reporting that the beds consist mostly of bivalve shell remains (probably *Inoceramus* prisms) with small amounts of biotite, colophene, quartz, pyrite and lignite. The bed contains numerous shark and other fish teeth and inoceramid prisms. Dawson (2000) referred to the Kamp Ranch as stratigraphically superjacent regressive limestones, consisting of arenaceous inoceramid grainstones which contain abundant megafossils (bivalves and ammonites) and trace fossils (*Rhizocorallium*). Bishop et al.,(1992) showed an unconformity separating the Britton and Arcadia Park Formations. This surface is placed below the Kamp Ranch limestone sequence. The Eagle Ford-Austin Chalk contact is a disconformity marked by a concentration of glauconite/phosphate grains and vertebrate remains. Some geologists have interpreted this boundary as a condensed zone (Reaser, 2002).

CHAPTER 3

THE CENOMANIAN IN NORTH CENTRAL TEXAS

The Cenomanian Stage (97- 90 Ma) of the Upper Cretaceous was first named in 1847 by D'Orbigny for sediments around LeMans in the Paris Basin, northern France. Cenomanum was the Latin name for this Cretaceous Stage. The 1999 Geological Society of America time scale places the Albian-Cenomanian boundary at 99 Ma and the Cenomanian-Turonian boundary at 93.5 Ma. Thus, the total duration of the Cenomanian is 5.5 Ma. Gale (1995) presented a composite cyclostratigraphy for the Cenomanian Stage in Western Europe, indicating a duration of the stage of 4.45 Ma. Corroboration of this cyclochronology comes from radiometric dating (4.0-4.5 Myr) (Gale, 1989). The discovery of a distinctive Cenomanian magnetostratigraphy, including several reversals was worked out by Vanderberg and Wonders (1980). The continuing recognition of Milankovitch cyclicity: the periods of orbital precession (~21ky), obliquity (~41ky) and eccentricity (~100ky and ~400ky) in the geological record permitted the development of an astrochronological time scale giving a notable refinement to our knowledge of the duration and dating of events within the geological time scale (Gale, 1989; Sageman et al.,1997). In the Late Cretaceous in particular, recognition of Milankovitch cycles, coupled with high resolution radiometric dates (Obradovich, 1993) provides very precise control for the chronostratigraphic calibration of the biozonations for this time period (Gradstein et al., 1994, 1995). The orbitally-tuned timescale developed by Gale (1989, 1995) and Gale et al., (1999) allows a calibration down to time slices of approximately 20 ky.

Biostratigraphy studies by Hancock (1960), Kennedy (1969), Wright et al., (1984) and Gale (1995) showed that ammonites provide the basic framework for biostratigraphy of the Cenomanian Stage. Where faunas are well and continuously preserved, a total of 13 successive assemblages can be recognized through the Cenomanian.

The major positive $d^{13}C$ excursion present in the latest Cenomanian and earliest Turonian has been recognized as a useful stratigraphic marker by many workers, and is associated with the widespread deposition of organic-rich shales on the shelves and ocean basins at that time (Jenkyns, 1980). The similarity of fine detail of this curve between Europe and Pueblo, Colorado in the Western Interior basin of the United States (Gale et al., 1993) suggests that $d^{13}C$ excursion can be a powerful tool in high resolution correlation.

Gale (1995) stated that integration of ammonite, inoceramid and planktonic foraminiferal biostratigraphy with the $d^{13}C$ curve and sequence stratigraphy and cyclostratigraphy provides a high- resolution framework into which most marine Cenomanian sections in Europe can be fitted.

Nevertheless, Sliter (1989) argued that planktic foraminifera, although widely used in Tethyan pelagic successions, do not offer a very fine subdivision of the Cenomanian Stage. The exact correspondence between the base of the Cenomanian as recognized using ammonites and that based on the entry of *Rotalipora brotzeni* (Sigal) is not established. The zone of *R. brotzeni* corresponds roughly to the Lower Cenomanian of ammonite workers; the zonation in the Middle and Upper Cenomanian includes successively the zones of *R. reicheli* (Monrod), *R. cushmani* (Morrow) and *W. archeocretacea* (Pessagno). The extinction of *R. cushmani* has been widely used as a

datum, but may well be significantly diachronous from region to region according to Gale et al., (1993).

The Eaglefordian sequence is interpreted to extend from late Cenomanian (94 Myr), to mid- Turonian (aprox. 89.5 Myr). Stratigraphically in outcrop, the Eaglefordian sequence extends from the unconformity, at the base of the Eagle Ford Group (Tarrant Formation), to the unconformity at the base of the Arcadia Park Formation in the Dallas County area (Phillips, 1987).

In North Central Texas the Cenomanian Stage is represented by the Grayson Formation of the Washita Group which is early Cenomanian in age (Ford, 1982). The middle Cenomanian comprises the Woodbine Group and the late Cenomanian the lower Eagle Ford Group.

Even though the Cenomanian/Turonian boundary lies within the Britton Formation of the Eagle Ford Group, its exact position has yet to be determined.

The Cenomanian ICS Stage ranges from 99.6 +/- 0.9 to 93.5 +/- 0.8 Ma. Start is defined by the lowest occurrence of planktic foraminifer *Rotalipora globotruncanoides*. End is defined by the lowest occurrence of ammonite *Watinoceras devonense* (Gradstein et al., 2004).

CHAPTER 4

AGE OF THE EAGLE FORD GROUP IN NORTH CENTRAL TEXAS

Paleontologic and biostratigraphic studies of the Eagle Ford Group have been carried out by several workers (Scott, 1926; Moreman 1942; Adkins, 1933, 1949; Adkins and Lozo, 1951; Stephenson 1955; Brown and Pierce 1962; Pessagno, 1967, 1969; Kauffman et al., 1977; Christopher, 1982) (Fig.4.1).

Some of those studies were based on studies conducted on material provided by the three complete cores of the Eagle Ford Group located in Dallas, McLennan and Wood counties, TX. drilled by Socony Mobil Field Research Laboratory. The core drilled in Dallas County, designated Core Hole 16 (drilled near Loop12, east of Mountain Creek Lake, lat. 32° 41'44''N, long. 96°54'16''W) contains the longest section of the Eagle Ford Group (Fig. 4.2). This core penetrated the basal beds of the Austin Chalk, the entire Eagle Ford Group, and the upper part of the Woodbine Formation (Christopher, 1982). In general, the above mentioned workers agree that: 1) the Cenomanian-Turonian boundary occurs within the Eagle Ford Group, and 2) only the upper Cenomanian is represented in the unit.

Disagreement exists however, over 1) the exact position of the Cenomanian-Turonian boundary within the Eagle Ford Group, and 2) whether or not the lower Turonian is represented (Christopher, 1982).

Whereas Adkins and Lozo (1951) and Brown and Pierce (1962) placed the Cenomanian-Turonian boundary in the lower half of the Britton Formation, Pessagno (1969) and Kauffman, Hattin and Powell (1977) placed the boundary within a few meters of the top

MICROFOSSIL ZONATION (from Pessagno, 1969)				MEGAFOSSIL ZONATION (from Kauffman et al., 1977)			PALYNOLOGIC ZONATION (from Christopher, 1977, 1979)		
				EUROPEAN STAGE	LOCAL ZONE (NORTHWESTERN OKLAHOMA)	REGIONAL ZONE (WESTERN INTERIOR)	EUROPEAN STAGE	ZONE	
				EUROPEAN STAGE	ZONE	SUBZONE	EUROPEAN STAGE	ZONE	
AUSTIN GROUP (in part)		SANTONIAN	<i>Gibbotruncana bulloides</i> Assemblage-Zone (in part)	<i>Marginotruncana concavata</i> Subzone	NO DATA			CONIACIAN	
ARCADIA PARK FORMATION		UPPER TURONIAN	<i>Marginotruncana helvetica</i> Assemblage-Zone	<i>Whiteinella archaeocretacea</i> Subzone	NO DATA			UPPER TURONIAN	unzoned (pre <i>Complexiopollis exigua</i> - <i>Santalacites minor</i> Zone)
BRITTON FORMATION		UPPER CENOMANIAN	<i>Rotalipora s</i> Assemblage-Zone (in part)	<i>Rotalipora cushmani-greenhornensis</i> Subzone	LOWER TURONIAN	<i>Mytiloides mytiloides</i> <i>Mytiloides opalensis</i>	<i>Mytiloides "labiatus"</i>	LOWER TURONIAN	
TARRANT MEMBER					UPPER CENOMANIAN	<i>Sciponoceras gracile</i>	<i>Sciponoceras gracile</i>	UPPER CENOMANIAN	<i>Complexiopollis-Atlantopollis</i> Assemblage-Zone (= zone IV)
WOODBINE FORMATION (in part)					UPPER CENOMANIAN	unzoned <i>Phleoptera quadrate</i> sp. aff. <i>P. minuta</i> <i>Phleoptera minuta</i> <i>Inoceramus frivus pictoides</i> unzoned <i>Inoceramus prefragilis stephensoni</i> <i>Ostrea beloiti</i> <i>Inoceramus rutherfordi</i>	<i>Dunveganoceras albertense</i> <i>Dunveganoceras conditum</i> <i>Dunveganoceras pondi</i> zones 1-7	UPPER CENOMANIAN	unzoned (post zone III)
WOODBINE FORMATION (in part)					UPPER CENOMANIAN	unzoned	<i>Inoceramus bellvuensis-l. dunveganensis</i>	UPPER CENOMANIAN	

Figure 4.1. Comparison of microfaunal, megafaunal, and palynological zonations of the Eagle Ford Group and adjacent strata of Texas (Christopher, 1982).

of the Britton Formation. In addition, Adkins and Lozo (1951) found no evidence of lower Turonian strata in Waco, TX. Pessagno (1969) also considered the lower Turonian to be absent in the Eagle Ford Group to the north, in the Dallas area, but he placed the Cenomanian-Turonian boundary higher in the Britton Formation than did Adkins and Lozo (1951). Kauffman, Hattin and Powell (1977), concurred with Pessagno (1969) in placing the Cenomanian-Turonian boundary near the top of the Britton Formation. Also, extensive field work led by Powell concluded that a nearly complete but condensed section of lower Turonian rocks is represented by the uppermost 4.5- 6 m of the Britton Formation and the basal 3- 9.5 m of the Arcadia Park Formation (Christopher, 1982). The presence of lower Turonian strata is indicated primarily by the occurrence of *Mytiloides mytiloides* (Mantell) Kauffman, Hattin and Powell (1977), *Inoceramus labiatus* Schlotheim, *Watinoceras* sp. and the foraminifers *Marginotruncana difformis* (Gandolfi) and *M. sigali* (Reichel). Middle and upper Turonian strata are represented by the upper part of the Arcadia Park Formation as evidence of the occurrence of *Collignoceras woollgari* (Mantell) and late forms of *Mytiloides labiatus* and *M. latus* (middle Turonian) and *Inoceramus cuvieri* Sowerby (late Turonian). No data are provided that permit an age determination for the remaining upper part of the Arcadia Park Formation.

MATERIAL

Forty sediment samples, each with a dry weight of approximately 500 g were collected during fieldwork (Summers of 2001, 2002, and 2003) from the 10 sites under study of the lower Eagle Ford in North Central Texas designated henceforth as: LEF-1, LEF-2, LEF-3, LEF-4, LEF-5, LEF-6, LEF-7, LEF-8, LEF-9 and LEF-10 to be examined for microfossils.

In addition 7 samples were collected (sites LEF-1, LEF-4, LEF-5, LEF-9) for calcareous nannofossil biostratigraphy as backup data (Watkins, written comm. 2001 and 2003).

MICROPALAEONTOLOGICAL TECHNIQUES

There are a host of techniques available depending on the nature of preservation and the matrix containing the microfossils. Different fossil groups require specific analytical procedures. For forams, the sediment samples are crushed first, when necessary, with a geological hammer. Unconsolidated sediments may be crumbled by hand and soft rock might be dissolved overnight in a bucket of water. Since the samples under study are mudstones and shales they are relatively soft and therefore disaggregation is straight forward. Each sample is then placed in an industrial blender to further pulverize the sediment for about 2 minutes at high speed depending on the sample. Maximum care should be exercised during this step in order not to contaminate one sample of sediment with the subsequent one. Therefore, the blender should be washed thoroughly between samples using a brush and detergent. After this step, the sample is placed in a glass flask or jar with a disaggregating agent like Calgon (sodium hexametaphosphate) and placed in a ultrasonic tank for about half an hour. After that, the sediment is washed through a stainless steel fine-mesh 0.063 mm sieve and filtered in a funnel using Watman No. 2 filter paper. At this point, it is advisable to take a small sample of the residue and place it under a microscope to ensure that the recovered forams are matrix-free. If not, the sample can be treated with a hydrogen peroxide solution (10%) and/or treated again with more calgon and the ultrasonic bath to ensure a high quality sample. The sample is washed again through the stainless steel sieve and filtered in a funnel using filter paper. After properly labeling this filter paper, the sample is placed in a microbiological oven for about 8h at 60 °C. The

63 micron residue is then evenly distributed on a black picking tray. Each of the samples is qualitatively and quantitatively analyzed using incident light stereomicroscopy in search of biostratigraphic key taxa. At least 300 forams were counted for each sediment residue. Microfossils were later mounted in gridded slides (32 cells) for reference as well as for future research. The classification of the foraminifera was mainly carried out according to the descriptions and figures of Robaszynski and Caron (1979) and Caron (1985). The data for each sample were compiled to produce an occurrence chart (Table 1).

AGE DETERMINATION

Based on foraminiferal data and following the biozonation of Pessagno (1969), the age of the sediments of localities: LEF-1, LEF-2, LEF-3, LEF-4, LEF-5, LEF-6, LEF-7 and LEF-8 is late Cenomanian corresponding to the *Rotalipora cushmani-greenhornensis* Subzone of Pessagno (1969) or *Rotalipora cushmani* zone of Caron (1985).

Additional data: Calcareous nannofossil biostratigraphy showed that the age of outcrops LEF-1, “is late, but not latest, Cenomanian in age. It lacks a peculiar and distinctive couple of fossils (*Corollithion kennedyi*, *Gartnerago theta*) that go extinct in the lower part of the *Sciponoceras gracile* zone, setting a floor on the age of mid-late Cenomanian. It contains two distinctive markers (*Rhagodiscus asper* and *Axopodorhabdus albianus*) that disappear at (or very near) the Cenomanian-Turonian boundary, placing a ceiling on the age at the boundary, In addition, it contains the older part of the lineage (*Eprolithius floralis*), but not the younger representative (*Eprolithius octopetalus*) that appears near the top of the Cenomanian (top *Neocardioceras juddi* zone). The nannos indicate an age of late *S. gracile* Zone to early *N. juddi* zone(= mid-late Cenomanian)” (Watkins, written comm. 2001).

TABLE 1. PLANKTIC FORAMINIFERA OCCURRENCE CHART

TAXA	LOCALITIES									
	LEF-1	LEF-2	LEF-3	LEF-4	LEF-5	LEF-6	LEF-7	LEF-8	LEF-9	LEF-10
<i>Heterohelix moremani</i>	A	A	A	A	A	A	A	A	NP	NP
<i>Heterohelix</i> sp.	C	C	A	C	A	A	C	A	NP	NP
<i>Hedbergella amabilis</i>	A	A	C	C	C	C	C	C	NP	NP
<i>Hedbergella brittonensis</i>	A	A	A	A	A	A	A	A	NP	NP
<i>Hedbergella delribensis</i>	A	A	A	A	A	A	A	A	NP	NP
<i>Hedbergella</i> sp.	C	C	C	C	A	C	A	R	NP	NP
<i>Guembelitra harrisi</i>	R	R	NP	R	R	NP	NP	NP	NP	NP
<i>Schakoina cenomana</i>	R	NP	NP	NP	R	NP	NP	NP	NP	NP
* <i>Rotalipora cushmani</i>	C	C	R	R	C	R	R	R	NP	NP
* <i>Rotalipora greenhornensis</i>	R	R	R	R	R	NP	NP	R	NP	NP

LEGEND
 A = Abundant (15 or more specimens)
 C = Common (5 - 14 specimens)
 R = Rare (1 - 4 specimens)
 NP = Not Present

* Index fossils (following planktic zonation of Pessagno, 1969).

Lithostratigraphic Units:

- LEF-1 to LEF-8 Lower Eagle Ford Group, lower Britton Fm., Turner Park Member
- LEF-9 Lower Eagle Ford Group, upper Britton Fm., Camp Wisdom Member
- LEF-10 Lower Eagle Ford Group, Tarrant Formation

For LEF-4 calcareous nannoplankton showed that “these samples are similar to LEF-5, but differ in one respect. Two species (*Lithraphidites acutus* and *Helenea chiastia*) are absent, suggesting that these samples are close to the C/T boundary and, perhaps, stratigraphically higher than LEF-5” (Watkins, written comm. 2003).

It is inferred therefore that the C/T boundary is contained within this locality. In addition, rare ammonite impressions of *Watinoceras reesidei* were found in the black shale horizon. This small ammonite is indicative of the lower Turonian worldwide (Kirkland, 1996). This confirms Pessagno’s (1969) assignment of the Cenomanian-Turonian boundary to a position high in the Britton Formation.

LEF-5 is an outcrop located across the Type section of the Turner Park Member of the Britton Fm. (as designated by Reaser, 2002) and is reported as being late Cenomanian in age (Reaser, 2002). Watkins (written comm. 2003) wrote: “These two samples have very similar assemblages containing several taxa indicating a late, but not latest Cenomanian age. These taxa include *Corolithion kennedyi*, *Rhagodiscus asper*, *Lithraphidites acutus*, *Helenea chiastia* and *Axopodorhabdus albianus*. All of these species have extinctions just below the C/T boundary. The lack of *Quadrum intermedium* indicates that this is not just beneath the boundary (but it could be within a meter or two). The preservation is good in sample #2 (black shale) and excellent in #1 (tan sediments)”.

In addition, the age assigned to locality LEF-8 is late Cenomanian based on the presence of megafossils (*Sciponoceras gracile*) (Kauffman et al., 1977; Kennedy et al., 2003).

The presence of the ammonite *Metoicoceras geslinianum* reported from the upper and middle Cenomanian (Kennedy and Cobban, 1990) and upper Cenomanian (Kennedy et al., 2003). *Metoicoceras* cf. *geslinianum* was recorded in outcrops LEF-1, LEF-2,

LEF-4, LEF-5, LEF-7, LEF-8 and LEF-9. Only one complete specimen was found (LEF-1).

In all other sites just fragments are present.

All sediment samples collected at locality LEF-9 (upper Britton Fm. Camp Wisdom Member) proved to be devoid of calcareous nannoplankton (Watkins, written comm. 2003) as well as of foraminifera. The age of this unit was assigned on the presence of molluscan megafauna assemblage and on published data since this locality is the type section for the Camp Wisdom Member of the Britton Fm. (Reaser, 2002). The age of the Camp Wisdom ranges from late Cenomanian to earliest Turonian (Reaser, 2002).

Locality LEF-10 is of late Cenomanian age based on the presence of several specimens of ammonite *Conlinoceras tarrantense in situ* (Figs. 4.3 and 4.4). No forams were recovered from this site.

Christopher (1982) reported palynological assemblages of the Tarrant Formation and underlying Woodbine as quite similar, supporting the findings of Brown and Pierce (1962). The microfaunal assemblage of the sections (LEF-1 to 8) shows that the diversity of the planktic foraminifera of the lower Britton (Turner Park Member) is low, with abundant and widely distributed, shallow-dwelling species of *Hedbergella* and *Heterohelix*. It is important to note also the absence of benthic forams. The assemblage is entirely planktic in all the outcrops under study. Low diversity foraminiferal populations consisting of high planktonic/benthic ratios indicate stressed conditions. This strongly suggests that the bottom sediments were anoxic creating an environment unsuitable for the benthic forams to inhabit. The bottom water conditions at the time of deposition of these deposits were anoxic and



Figure 4.3 Locality LEF-10 Tarrant Formation at Johnson Creek.



Figure 4.4 Locality LEF-10 *Conlinoceras tarrantense* in situ

close to the C/T boundary. Last occurrences of the planktic foraminifera *Rotalipora cushmani* and *R. greenhornensis* are events recognized worldwide and are commonly used for correlation. The extinction of species of *Rotalipora* probably reflects changes in paleoceanographic conditions related to the C/T Anoxic Event (Lamolda and Peryt, 1995; Lamolda et al., 1997).

LITHOSTRATIGRAPHIC DETERMINATION

Localities LEF-1, LEF-2, LEF-3, LEF-4, LEF-5, LEF-6, LEF-7, and LEF-8 are assigned to lithostratigraphic unit known as Turner Park Member of the Britton Formation of the lower Eagle Ford Group (Reaser, 2002) based on micro- and macrofaunal assemblages and lithological characteristics. It is also worth mentioning here that local professional as well as avocational paleontologists of the area refer to locality LEF-4 as the “Tarrant beds”. This stratigraphic assignment is refuted here by the writer on the basis of extensive fieldwork and the data gathered herein. The lithostratigraphic unit of this locality (LEF-4) is the lower Britton Fm. (Turner Park Member) and not the “Tarrant beds” as it has been erroneously assumed for many years. Locality LEF-9 is the type locality of the Camp Wisdom Member of the Britton Formation (Reaser, 2002). Locality LEF-10 (Fig. 4.3) is identified as the Tarrant Formation, basal lithostratigraphic unit of the Eagle Ford Group, based on ammonite *C. tarrantense in situ* (Fig. 4.4). No forams were recovered from this unit. Pessagno (1969) reported “no planktonic foraminifera were recovered from the Tarrant Formation”. However, megafossil (ammonite) evidence indicates that the Tarrant Formation is of late Cenomanian age (Pessagno, 1969).

CHAPTER 5

DEPOSITIONAL ENVIRONMENT

During the Late Cretaceous the greatest inundation of the continents by the sea since the Ordovician occurred. Eighteen percent of the Earth's surface was land, compared to today's twenty eight percent (Matsumoto, 1977). The Western Interior Seaway (WIS) extended approximately 6000 km meridionally over the central North American continent throughout most of the Cretaceous (Kaufman and Caldwell, 1993). This epeiric sea provided a conduit between the sub-mediterranean Boreal Sea and subtropical water masses from the proto-Gulf of Mexico and Atlantic Ocean (Tethys). The WIS resulted from the development of an elongated foreland basin east of the emerging North American Cordillera in conjunction with the highest global sea levels of Mesozoic time.

Although the Cretaceous saw the waxing and waning of several trans-continental seaways in different climatic settings such as the trans-African- (Reyment, 1980), the trans-Asian-, and the trans-Australian epeiric seas (Veevers, 1984), all providing relatively shallow conduits between marine basins, the paleoceanography of these settings is relatively unique. Compared to its contemporaneous equivalents, the WIS was by far the most extensive especially during the Cenomanian-Turonian second order cycle known as the Greenhorn Cyclothem (Kauffman, 1977a). There are no modern analogs to the WIS, although it has been suggested that oceanographic conditions of this seaway are somewhat similar to large estuarine systems (Slingerland et al.,1996).

The Upper Cretaceous marine biota of the Western Interior Basin indicate restricted marine

conditions, uncharacteristic of open marine basins. Two factors indicate abnormal marine conditions. The first factor is water column stratification throughout a considerable part of the basin history (Kauffman and Sageman, 1990). This gave way to a dysoxic to anoxic bottom-water environment supporting relatively abundant opportunists and low oxygen-adapted taxa. The second factor inferred from the biota is the predominance of brackish-water conditions for surface waters with influx of cooler, less saline waters from the northern circumpolar ocean (Hay et al., 1993).

The Eagle Ford sea was part of the mid-Cretaceous Greenhorn Sea, which in turn was part of the WIS. The Eagle Ford Group was deposited at the south-southeastern end of the Western Interior Basin. The Eagle Ford shales were deposited in an epicontinental sea which had an anoxic water column or bottom environment. Three models of Eagle Ford deposition have been defined:

- 1) anoxic silled basins
- 2) anoxic layers caused by upwelling
- 3) anoxic open oceans.

In the first model (silled basin), a physical barrier restricts the vertical mixing of water and enhances water stratification. This, in turn, causes anoxia, as in the Black Sea of today. According to Heckel (1972), the Eagle Ford depositional environment, with its thin laminations and horizontal bedding fits into this stagnant marine environment.

In the second model: the low sea level bottom waters stay anoxic, meanwhile high sea Level (transgression) spreads, establishing a permanent thermocline over most of the epicontinental sea. Upwelling into the sea from the open ocean causes a planktonic boom, organic matter from the bloom settles into the oxygen-minimum zone where its

decay further depletes oxygen and causes the deposition of black shales (Heckel, 1977). In this upwelling model, anoxia occurs when oxygen-poor but nutrient-rich waters are welled upward into the euphotic zone and mix with oxygenated water to cause high biologic production (bloom) within the upwelling area (Demaison and Moore, 1980). The anoxic open ocean model (c) is best explained on continental margins during major transgressions (Schlanger and Jenkyns, 1976; Arthur and Schlanger, 1919; Demaison and Moore, 1980.). In this model, an expanded oxygen-minimum layer in the open ocean on the western side of continents at low latitudes is caused by deep oceanic circulation and the Coriolis effect (Demaison and Moore, 1980). However, it has been suggested that these anoxic events are a series of local anoxic event (Waples, 1983) similar to those caused by upwelling.

Eagle Ford anoxic sediments appear to have been deposited under somewhat peculiar circumstances, most closely related to the upwelling model, but the other two models anoxic open ocean and anoxic silled basins could also be related (Charvat, 1985).

BENTONITES

The Eagle Ford Group contains many seams of bentonite interbedded with shales, lime mudstones, and limestones. In all but a few of the outcrops the individual bentonite seams are conspicuous and continuous. They range from 3 to 30 cm in thickness. These are particularly conspicuous in lower Eagle Ford rocks (Figs. 5.1 and 5.4). The upper and lower contacts of the bentonite with the overlying and underlying beds of calcareous mudstones or shales are always sharp. They are conspicuous in weathered outcrops where their color ranges from creamy white



Figure 5.1 Bentonite at locality LEF-3.

to buff or tan to bright orange.

The term bentonite was originally suggested by Knight (1898) for a peculiar clay-like material with soapy properties from its occurrence in the Fort Benton unit of Cretaceous age in Wyoming. Ross and Shannon (1926) presented the following definition, which has been widely quoted: “Bentonite is a rock composed essentially of a crystalline clay-like mineral formed by devitrification and the accompanying chemical alteration of a glassy igneous material, usually a tuff or volcanic ash in a marine environment. The characteristic clay mineral has a micaceous habit and facile cleavage, high birefringence, and a texture inherited from the volcanic tuff or ash, and it is usually the mineral montmorillonite, but less often beidellite”.

Bentonite is a clay mineral formed by the weathering of volcanic ash and is composed of particles of colloidal or near colloidal size. Bentonite itself is not a distinct mineral, but is composed dominantly of a mixture of smectite clay minerals. The physical properties of bentonites from various localities and stratigraphic horizons are quite variable and this variation is to some extent due to varying proportions of the different clay minerals present. The physical properties of bentonite determine its value for different industrial uses, and the most important property is the varying abilities of different bentonites to absorb water and thus increase in volume (Spivey, 1940).

Bentonites are believed to have been developed principally from finely divided volcanic ash that settled at various times during the geologic past in shallow interior seas. Whether the water in these seas was any different from that of the present oceans cannot be determined. It is probable, however, that the ash of the various falls may have varied both in size of the particles and in composition. Further, the conditions which caused

devitrification and other changes in the ash may have varied. Consequently, there are at least two main types of bentonite and many intermediate varieties. The higher grades of bentonite, have the property of absorbing large quantities of water and swelling enormously. They will remain indefinitely in suspension in thin water dispersions. The lower grade bentonites are like the ordinary plastic clays and do not have the property of swelling, nor will they remain in suspension long (Wing, 1940).

Neoformation of smectite in a sedimentary environment is certainly a possible mode of origin, for some occurrences, where there is no evidence of any related volcanic activity (Grim and Guven, 1978).

The Eagle Ford bentonites are composed principally of smectite (calcium montmorillonite) with minor amounts of sodium montmorillonite, illite, quartz, feldspars, and calcite (Charvat, 1985). There are several possible source areas for the volcanic ash, which was altered to bentonite in the Eagle Ford. The source of the volcanic ash would be any volcanic activity that was contemporaneous with deposition of the Eagle Ford shales. It is possible that the volcanoes were located in southwestern Arkansas (Ross et al., 1928; Hunter and Davis, 1978) or in the Western Interior (Reeside, 1957; Kauffman 1977a). Hunter and Davis (1979) believed that the center of Woodbine-Eagle Ford volcanism was located in southwest Arkansas. Eaton (1963) and Slaughter and Hamil (1970) showed that volcanic ash can be transported great distances (>320 kilometers) by the wind and still form substantial accumulations.

Charvat (1985) in his studies on the Eagle Ford Group bentonites concluded that the layered bentonites of the Eagle Ford are derived from windblown volcanic ash that was altered in a shallow marine environment, *in situ*. This author indicated also that the dark colored, thinly

laminated rocks of the Eagle Ford Group are not bioturbated and lack most benthic organisms indicating an euxinic (anoxic) depositional environment, possibly the product of upwelling of deep waters and that the environmental conditions necessary for bentonite formation also indicate an anoxic environment. Therefore, the shales, lime mudstones and limestones which are interbedded with the bentonites, were deposited in an environment that was similar to the environment during bentonite formation, thus placing Eagle Ford deposition into the euxinic (anoxic) environment (Charvat, 1985).

TOTAL ORGANIC CARBON (TOC%)

The Cenomanian/Turonian boundary coincides with an oceanic anoxic event (OAE) which is expressed globally as a positive excursion of the stable carbon isotopic composition of carbonates and organic carbon (Simons and Kenig 2001). Oceanic anoxic events (OAEs) were short-lived episodes of organic carbon burial that are distinguished by their widespread distribution as discrete beds of black shale and/or pronounced carbon isotopic excursions OAE1a in the early Albian (~120.5 Ma) and OAE2 at C/T boundary (~93.5 Ma) were global in their distribution and associated with heightened marine productivity (Leckie et al., 2001). The Cenomanian-Turonian Stage boundary interval (90-93 m.y.) is recognized as the time of maximum eustatic rise in the Cretaceous. This sea-level rise was accompanied by a global OAE (Schlanger and Jenkyns 1976), commonly marked by an easily recognized dark organic-rich shale layer, the “black band” of England and the “Bonarelli Layer” of Italy. Biogenic fractionation and the increased burial of ^{12}C -enriched organic matter at this time probably resulted in the late Cenomanian global positive ^{13}C excursion (Schlanger et al., 1987), which was initiated several hundred thousand years before the paleontological C/T

boundary, as defined, by first appearance of Turonian ammonites (*Watinoceras devonense* fauna) and Turonian species of the inoceramid bivalve *Mytiloides* (Orth et al, 1993).

The widespread deposition of organic-rich sediments, which can be correlated between different basins and oceans were described as OAEs by Schlanger et al., (1986). According to these authors the mid-Cretaceous includes two main OAEs: Aptian-Albian (OAE1) and the Cenomanian-Turonian transition (OAE2) or Bonarelli Event of Arthur and Premori-Silva (1982).

The deposition of organic carbon-rich sediments in marine successions implies special geochemical conditions for short time intervals during the mid-Cretaceous. The extraordinarily organic carbon-rich sediments deposited during this time are mostly characterized now by black shales alternating with silty-sandy mudstones. These marine sediments have been recorded in different basinal settings such as mid-ocean plateaus in the Pacific basin, the interior seaways of the North American craton, European shelf and the interior seaways, circum-African embayments and seaways, Tethyan margins and the Caribbean region (Scholle, 1974; Schlanger et al.,1986). They have also been recovered from Deep Sea Drilling Project (DSDP) sites in the Atlantic and Pacific Oceans and more recently from a number of Ocean Drilling Program (ODP) Sites (Arthur, 1979; Jenkyns, 1980).

In the light of many previous studies (Schlanger et al.,1986; Arthur et al.,1986, 1990; Jenkyns, 1980; Gale et al.,1993; Orth et al, 1993) deposits of the Cenomanian/Turonian OAE have been defined to have: 1) global occurrence, 2) time constancy, 3) extraordinarily high values of organic matter composed mainly of type II marine kerogen, 4) carbon isotope anomalies with positive $d^{13}C$ excursions, 5) an abundance of some trace metals, such as Cr,

Au, Cd, Mo, Ir, Pt, Sb, V, Zn, 6) finely laminated sediments, 7) low sedimentation rates as a result of little or no terrestrial input, 8) rare or none bioturbated layers, 9) rare or none macro- and microbenthos, 10) extinction/radiation events and important evolutionary changes in marine plankton (foraminifera and radiolarians) within them.

Many depositional models have been proposed to explain the mechanisms of organic carbon-rich sedimentation. These are: 1) a stagnant ocean (De Graciansky et al., 1984), 2) an expanding ocean minimum zone (Schlanger et al., 1984) and 3) slow circulation affecting the development of anoxic deeper oceanic waters (Sarmiento et al., 1988; Southern et al., 1982) and others (Yurtsever et al., 2003).

Black shales are very rich in organic matter, the precursor of petroleum. The quantity or richness of organic matter content of a sediment is normally determined by the total of organic carbon (TOC or C org). The total organic carbon content is obtained from a carbon analyzer, which is the common method to determine source rock potential. Source rocks are sedimentary rocks that may become or have been able to generate petroleum.

The minimum values of organic carbon for potential source rocks are 0.3% for carbonates and 0.5% for shales. As a general rule, shales containing less than 0.5% TOC are considered as a negligible source, having poor potential to be a petroleum source rock. Shales with 0.5%-1.0% TOC are considered as slight source that have fair potential. Shale with 1.0% to 2.0% TOC are considered modest source, indicates a good source potential. Shales containing more than 2.0% TOC are considered as good to excellent sources indicating a very good and excellent source potential. TOC as well as whole rock mineralogy were performed in the area under study. The results of these analyses are reported in Appendix II. Only two localities were chosen due to financial constraints. LEF-4 is a section within the

lower Britton and LEF-9 is a section within the upper Britton. LEF-4 is a locality particularly significant because in this site was observed a clear interfingering of the tan color mudstones and the black shales. The TOC% results of these two samples show 0.20% for the tan mudstones and 4.65% for the black shales.

Even though no TOC % analyses were done on localities LEF-1, LEF-2, LEF- 3, LEF-5, LEF- 6, LEF-7 and LEF-8 it is reasonable to infer that the TOC% would be similar to LEF-4 since they all present the same lithological characteristics (tan color mudstones). Locality LEF-9 is a section of the upper Britton (Camp Wisdom Member). This is a typical black shale that shows a TOC of 1.21%.

These results are consistent with previous studies (Surles, 1987; Liro et al.,1994; Robison, 1997) in that the Eagle Ford is known to display significant variability in organic enrichment. Differences can be observed between two outcrops and within each of the localities (Surles, 1987). Furthermore, these findings suggest that the lower Eagle Ford has good potential as a source rock (i.e. rocks containing above-average quantities of organic carbon, TOC >1.0 wt.%) (Bissada, 1982).

The results of the whole rock mineralogy of these same localities LEF-4 (tan color horizon), LEF-4 (black shales horizon) and LEF-9 are also presented in Appendix II. The X-ray diffraction analyses show that the tan color mudstones (LEF-4) present a high calcite content (57%) compared to the interbedded black shales in this same locality (37%) and to the LEF-9 (black shale, also) (5.3%). Sediment color is controlled by variations in total organic carbon and total carbonate content. Thus, the light color mudstones have a lower TOC% and a higher carbonate content than the black color shales.

From the whole rock mineralogy is also observed the high percentage of total phyllosilicates (illite/smectite) in the samples: LEF-4 (tan sediments) 38%, black shales same locality, 42% and LEF-9 (black shales also), 74%. The quartz content in LEF-4 (tan sediments) is 4.5%, for the black shales same locality (15%) and LEF-9 (dark shales also), 19%. There is also a difference in pyrite content among the localities: LEF-4 (tan sediments) 0%, black shales (same locality) 6.0% and LEF-9, 2.2%.

These results confirmed the X-ray diffraction analyses of Frost et al., (1982) which indicated that their Eagle Ford samples consisted predominately of smectite (70 to 95%), with accessory amounts of kaolinite, quartz, glauconite, calcite, apatite and biotite.

From these data it is reasonable to conclude that the Eagle Ford displays significant variability in organic enrichment and that those differences are not only observed between the localities, but also within the same locality. Even though color is a characteristic of any given sediment and that it should be interpreted with caution, there appears to be enough differences in mineral composition and TOC% between the tan mudstones and the black shales that they are interpreted to have been deposited in different depositional environments. One would tend to believe that the tan color sediments (Figs. 5.2 and 5.3) so prevalent in the lower Britton (Turner Park Member) were deposited under less anoxic conditions than the black shales (Fig. 5.4) of the Eagle Ford. However this does not seem to be the case, since there are very few indications of benthonic life in these tan color sediments as well as the typical dark shales of the rest of the Eagle Ford Group. Another possible explanation for the difference in rock composition and TOC% between the interfingering tan mudstones and dark shales of the lower Eagle Ford is the pulsatory



Figure 5.2 Buff color mudstones at locality LEF-2.



Figure 5.3 Buff color mudstones at locality LEF-1.



Figure 5.4 Black shales at locality LEF-4. Note bentonite above hammer.

nature of the Cenomanian transgression (Wilmsen, 2003).

Liro et al., (1994) stated that the Eagle Ford Group records mixed siliciclastic/carbonate deposition during the Late Cretaceous (Cenomanian-Turonian) transgression on the Texas craton. His detailed analysis of Eagle Ford outcrops in Central Texas revealed considerable lithologic variability at several different scales. The lower section is interpreted to represent transgressive, in part “condensed” deposits, whereas the overlying section is interpreted as a highstand deposit. Detailed organic geochemical analyses clearly differentiated the upper and lower Eagle Ford. The shales in the lower transgressive unit are organically enriched, exhibit higher generation potential, and are more oil-prone than the shales in the overlying regressive interval. Even within the oil-prone intervals there is considerable variation in geochemical attributes (Liro et al.,1994).

Previous work by Surles (1987) clearly documented that at least limited portions of the Eagle Ford Group are organically enriched, having organic carbon values greater than 9%. However his work, did not examine either the actual generation potential of the Eagle Ford shales or the nature of the organic matter contained within them. Such information is critical to any meaningful source rock study because of the numerous examples in which organic-rich rocks are incapable of hydrocarbon generation because of the hydrogen-depleted nature of their kerogen (Deroo et al.,1979).

Such analyses include total organic carbon, total sulfur, Rock-Eval pyrolysis, elemental analysis of isolated kerogen, and gas chromatography of the saturated hydrocarbon fraction. It is beyond the scope of the present thesis to undertake a complete source rock study, so only total organic carbon and whole rock mineralogy were determined. The results of both are presented in Appendix II. It is also important to notice that while every attempt was

made to collect fresh samples, the measured total organic carbon (TOC) contents may have been reduced by exposure and weathering as Clayton and Swetland (1978) pointed out. Therefore, reported values should be considered minimums.

Much of the Eagle Ford Group was deposited during a relative global transgression and highstand of eustatic sea level (Haq et al., 1988; Jiang, 1989). The transgressive episode containing the Eagle Ford Group occurred following the regional Middle Cretaceous Unconformity in the Gulf of Mexico Basin (Winker and Buffler, 1988). Within the Eagle Ford Group is a major “condensed section” (sediment starvation and/or omission during transgressive and early highstand systems tracts) developed during cycle 2.5 of the UZA-2 supercycle (Haq et al., 1988). Loutit et al., (1988) showed evidence that sediments associated with this transgressive event have characteristic isotopic properties that are observed in time-equivalent sections in many parts of the world. This time interval is also coincident with deposition of numerous organic-rich source rocks throughout the world (Arthur and Schlanger, 1979).

In addition, the Eagle Ford Group contains many elements (i.e glauconite, enrichment in organic matter, and bentonites) observed in marine condensed sections (Loutit et al., 1988). Regional geologic factors suggest that Eagle Ford-age seas attained their maximum extent during the Britton Formation deposition. Deltas prograding from the north and northwest delivered siliciclastic detritus into the shallow Eagle Ford-age marine basin (Surles, 1987). Consequently, shales, siltstones, and fine-grained sandstones are interstratified with marine limestones and bentonites. The predominance of marine fossils attests to a largely marine origin for the Eagle Ford Group (Pessagno, 1969). However, the siltstones and sandstones

contain a mixture of marine fossils and lignitic, terrestrial, plant debris which is indicative of a near-shore depositional setting (i.e. interdeltic or lagoonal). The well-preserved thin bentonites and finely laminated nature of the shales and siltstone are indicative of low-energy (below storm wave base) depositional conditions. The fine sandstone and siltstone beds may record periodic higher energy events (i.e. storms) (Liro et al, 1994).

In 2000 Dawson reported that the Eagle Ford Group records a second order transgression and contains a major (third-order) condensed interval. Eagle Ford strata are overlain disconformably by the Austin Chalk. The Eagle Ford/Austin contact represents the Turonian/Coniacian (89 Ma) boundary. The Cenomanian/Turonian boundary (92 Ma) occurs within the Eagle Ford.

Regionally, the Eagle Ford Group consists of two major depositional units: a lower transgressive and partly condensed unit and an upper highstand unit. These two depositional units have distinctive sedimentological and geochemical characteristics. Eagle Ford strata exhibit considerable lithologic variability. Six major argillaceous microfacies compose Eagle Ford strata: 1) fissile shale 2) phosphatic shale 3) bentonitic shale 4) fossiliferous (foraminiferal) shales 5) silty (quartzose) carbonaceous shale and 6) sapropelic shales and claystones. Eagle Ford shales are composed predominantly of smectite. Microfacies 1, 2 and 3 comprise the condensed part of the Eagle Ford interval. Microfacies 4 and 5 represent highstand Eagle Ford strata, and Microfacies 1 and 6 comprise the transgressive part of the Eagle Ford. These six shale microfacies are thinly interstratified with bentonites, carbonaceous siltstones and limestones (foraminiferal, inoceramid and peloidal grainstones) (Dawson, 2000).

Condensed Eagle Ford strata represent poorly-oxygenated, low-energy, marine

paleoenvironments of deposition. In contrast, overlying regressive Eagle Ford lithofacies accumulated in relatively high-energy, oxygenated, shallow marine environments (Dawson, 1997). Charvat and Grayson (1981) concluded that Eagle Ford deposition occurred in water depths ranging from 18 to 30 m and under anoxic conditions.

Frost et al.,(1982) reported that the Eagle Ford shale is the source rock for much of the hydrocarbons found in the Austin Chalk and Buda limestone. The Eagle Ford shale is a predominantly fractured, brittle, often micaceous, and fossiliferous shale with some siltstone and occasionally recrystallized dolomitic lime streaks that exhibit a highly oil-saturated matrix. In some areas, these carbonate stringers range in thickness from 3.0 to 9.1m.

Depending on the location within the Cretaceous trend, Eagle Ford shale may vary from a typically dark, organic-rich shale that exhibits high potassium (K), high thorium (Th) and excessively high uranium (U) to indicating those calcareous, brittle, fractured, and often productive Eagle Ford intervals that exhibit low potassium, low thorium and excessively high uranium. Uranium concentrations ranging as high as 7 to 15 ppm are frequently observed (Frost et al.,1982). Although the Eagle Ford Group may be considered to represent a single, regional condensed interval according to sequence stratigraphy studies, there are significant variations in source rock attributes within the unit. These differences occur both laterally and stratigraphically. Lateral variations appear to be the result of differences in proximity to the coast. Stratigraphic variations appear to be at least in part the result of relative changes in sea level (Liro et al.,1994).

HARDGROUNDS

Hardgrounds are sites of primary submarine lithification of the sea floor, usually stable,

high energy, well oxygenated environments with a good supply of food for the suspension feeders that constitute the great proportion of their faunas (Wilson and Palmer, 1992). Hardgrounds are syndimentarily lithified carbonate seafloors that became hardened *in situ* by the precipitation of a carbonate cement in the primary pore spaces. Ancient hardground faunas are attractive to paleontologists for two reasons: First, fossils are preserved *in situ* on hardgrounds, so their distribution on the surface and proximity to neighbors reflects their original spacing in life. This makes hardgrounds appropriate for various types of paleoecological analysis (Wilson and Palmer, 1992).

There is evidence of hardgrounds in some of the localities under study. Several slabs were found float at localities (LEF-2 and LEF-4) and *in situ* at localities (LEF-1 and LEF-8). The slabs from locality LEF-4 are dark gray in color and under stereoscope present countless disassociated shiny inoceramid prisms pyritized, tiny shark teeth as well as many microcoprolites. The LEF-2 slabs are dark yellow also very well lithified and contain an array of shark and other fishes teeth small and big, many microcoprolites and inoceramid prisms. The LEF-8 slab is light gray and contains large shark teeth and rare isolated reptile vertebrae identified as *Coniasaurus* sp. The slabs recovered from LEF-1 are rather sandy poorly-indurated and are interpreted as a storm deposit. According to Marsaglia and Klein (1983), fine sandstone and siltstone beds record periodic higher energy events (i.e. storms). According to these authors during the Late Cretaceous this area had the optimum paleogeographic position for seasonal (winter) storms.

SUMMARY

The Eagle Ford, a highly calcareous, thinly laminated shale, consists of undisturbed laminates suggesting periodic stagnation and anoxic bottom conditions alternating with richly fossiliferous limestone beds which suggest oxygenated conditions. In addition, numerous thin bentonite seams punctuate the mudstones, each of which represents a single ash fall. A shallow marine shelf, platform or lagoonal setting has been proposed as the depositional environment for the Eagle Ford Group. In addition, mud-dominated deltas supplied by sediment from the north have been suggested to explain some facies within the Eagle Ford.

The Tarrant Formation (lower-most unit of the Eagle Ford Group) in particular represents a transition from the deltaic Woodbine strata to more typical shales of the rest of the Eagle Ford. This unit is mainly gray to brownish gray calcareous sandstone, indicative of lower destructive energy (Surles, 1987). The rest of the lower Eagle Ford Group (lower and upper Britton) is interpreted as have been deposited under low energy (below wave base) or quiet, nearshore, shallow to moderately deep water marine environment, where a variety of fish, sharks, turtles, coniasaurs and plesiosaurs inhabited and accumulated on the anoxic ocean floor devoid of bottom-dwelling scavengers, where they were slowly covered by fine sediments.

CHAPTER 6

SYSTEMATIC PALEONTOLOGY

The biostratigraphic subdivision of Cretaceous time rests upon an invertebrate faunal zonation (Cobban, 1984; Kennedy 1988; Kennedy and Cobban 1990; Kauffman, 1977b, among others). The Cretaceous marine record of vertebrate life in the Western Interior is stratigraphically and geographically more complete than the terrestrial record. The lower Eagle Ford outcrops of North Central Texas being marine and part of the Western Interior Seaway are no exception to this scenario and are classified as Niobraran “age” (Russell, 1993). This author presented the following North American Marine Vertebrate “Ages” (acronym NAMVA) based on marine vertebrate fossils. Their conceptual base is taken from North American Land Mammal “Ages” (acronym NALMA), which were not intended to fit within a formalized category for the purpose of biostratigraphic correlation, and have flexible limits not defined by a lithostratigraphic referent:

Trinitian “Age”.- This “age” is proposed as a provincial time term, based on the vertebrate assemblage from the Paluxy Formation of the Trinity Group in north-central Texas.

Assemblages from the Travis Peak Formation to the Buda Limestone and equivalents (of approximately Aptian to early Cenomanian age) are included. “Trinitian Age” fauna includes: diverse primitive durophagous fishes and ichthyosaurs.

Woodbinian “Age”.-This “age” is based on the marine vertebrate assemblage of the Woodbine Formation in North Central Texas (of approximately middle Cenomanian age).

This “Age” includes: less diverse durophagous fishes, no ichthyosaurs.

Niobrara “Age”.- This “age” is proposed as a provincial time term , based on the vertebrate assemblage from the Niobrara Chalk of western Kansas. Assemblages from the Eagle Ford Formation to the Ozan Member of the Taylor Formation and equivalents (of approximately late Cenomanian to early Campanian age). The “Niobran Age” fauna includes abundant large teleosts, primitive mosasaurs and hesperornithiformes.

Navesinkan “Age”.- This “age” is proposed as a provincial time term , based on the vertebrate assemblage from the Navesink Formation of New Jersey. Included here are assemblages from the Pecan Gap Member of the Taylor Formation to the end of the Cretaceous Period and equivalents (of approximately late Campanian to terminal Maastrichtian age). Faunal elements include: large teleosts and less abundant advanced mosasaurs (Russell, 1993).

It is worth mentioning here that Texas has an excellent fossil record of Cretaceous sharks, rays and other fishes, including the oldest record of saurodontid fishes in North America, found by the author in one of the localities under study (LEF-1) (Friedman, 2001; Stewart and Friedman, 2001; Friedman 2002).

Capetta (1987) listed approximately 96 genera of elasmobranchs that he considered to be valid taxa from the Cretaceous worldwide. One fourth of these genera occur in Texas and numerous undescribed forms are awaiting study (Welton and Farish, 1993).

Fish first appeared during the early Paleozoic Era more than 450 million years ago.

They were the first true vertebrates to appear and give rise to all higher forms of life.

The estimated 20,000 species have adapted themselves to almost every variation of their environment. Approximately 14,000 of the living fishes are marine, the remaining 6,000

species live in fresh water. The living fishes comprise three vertebrate classes: Agnatha, the jawless fishes, including the lamprey; the Chondrichthyes, possessing true jaws and skeleton of cartilage, as in the sharks; the Osteichthyes (or Bony Fish) such as the trout, possessing jaws and a bony skeleton.

The lower Eagle Ford localities under study contain a very rich vertebrate assemblage, that includes a great abundance of articulated and disarticulated shark and other fish remains, as well as pleisosaurs, turtles, crocodyliforms, marine lizards and countless vertebrate coprolites. Most of the fossils collected were found by surface collecting and sometimes by digging during field work (Summers 2001, 2002 and 2003).

A detailed systematics of the vertebrate faunal assemblage from the lower Eagle Ford is beyond the scope of the present study. The taxa present at each locality is presented in Table 2 (page 56).

Identification of the fossil specimens found in the study area was made by reference to published taxonomic studies and by discussions with acknowledged experts on the specific taxonomic groups under consideration (Welton and Farish, 1993; Case 1967, 2001; Williston, 1900; Woodward 1912; Frickinger 1987; Russell 1993; Hill, 1976; Siverson, 1992; Shimada, 1994a, 1994b, 1996, 1997a, 1997b, 2000, 2001; McNulty and Kienzlen 1970; Capetta, 1973; Willimon, 1973; Welles and Slaughter 1963; Kennedy and Cobban 1990; Kirkland 1996; Bardet et al., 2000; Hattin and Cobban, 1977; Kauffman, 1977b; Lazo and Cichowolski 2003; Kennedy and Juignet 1994; White, 1940; Welles 1949; McClung 1926; Applegate 1965; Stewart, 1899; Shuler, 1950; Young and Powell, 1978; Bardack, 1965; Fishman et al, 1995; Cavin, 1999; Raab and Chalifa, 1987; Hoch,

1992; Maisey 1996; Hirayama 1997; Finsley, 1989; Bell, 1986; Schumacher and Shimada, 2002; Hamlett, 1999; Cumbaa and Tokaryk 1993; Everhart and Everhart, 1993; Casier, 1953; Budker, 1971; Akers and Akers 2002; Antunes and Capetta 2002).

However, poorly preserved or fragmentary material or material found that is not sufficiently complete to allow a more precise systematic identification was given the designation *gen. et sp. indet.* Also, some of the specimens collected during fieldwork may be genera and species new to science and are currently under study by various researchers with the results pending or will be the subject of future research. The localities under study are important sections that further define the stratigraphy and most importantly the paleontological significance of the basal Eagle Ford Group of North Central Texas. The fossil megafaunal assemblage collected at the study area is presented in Table 2.

Table 2. Megafossil assemblage at the lower Eagle Ford (Continued)

ORDER TESTUDINES Batsch, 1786										
Gen. et sp. indet.	8 pleastron fragments 1 humerus	12 pleastron fragments 2 vertebrae	4 pleastron fragments	1 pleastron fragment						
ORDER SQUAMATA Oppel, 1811										
Family Dolichosauridae Gervais, 1852										
<i>Coniasaurus</i> sp.	3 limb bones 10 vertebrae	52 vertebrae 3 jaws	3 vertebrae	1 vertebra	1 vertebra					
<i>Coniasaurus</i> cf. <i>crassidens</i>		2 jaws								
INVERTEBRATA										
CLASS BIVALVIA Linnaeus, 1758										
Family Anomidae Rafinesque, 1815										
<i>Anomia</i> sp.		12								
Family Inoceramidae Glebel, 1952										
<i>Inoceramus</i> sp.	420 shell fragments	500 shell fragments	120 shell fragments	12 shell fragments	20 shell fragments	50 shell fragments	120 shell fragments	12 shell fragments	40 shell fragments	
<i>Inoceramus pictus pictus</i> Sowerby, 1829		10								
<i>Inoceramus tenuimbonatus</i> Warren, 1930		8								
<i>Inoceramus arvensis</i> Stephenson, 1952		3								
<i>Inoceramus flavus pictoides</i> Somay		7								
Family Ostreidae Rafinesque, 1815										
<i>Pseudoperna congesta</i> Conrad, 1843		80	3	12						
CLASS CEPHALOPODA										
Family Acanthoceratidae De Grossouvre, 1894										
<i>Coninoceras tarrantense</i> Adkins 1928										12
<i>Meliticoceras</i> cf. <i>gesslinium</i> d'Orbigny, 1950	1		2 fragments			3 fragments			2 fragments	2 fragments
<i>Watinoceras reesidei</i> Warren, 1930			4							
Family Baculitidae Gill, 1871	3 fragments									
<i>Sciponoceras gracile</i> Shumard, 1860.									1	
TRACE FOSSILS										
Vertebrate coprolites	1500	1134	1800	1283	226	120				
Chondrites			one horizon							
<i>Thalassinoides</i>										8 slabs
PLANTAE										
Petrified wood										
Gen. et spec. indet.		4 pieces								
Carbonized plant impressions										
Gen. et spec. indet.	9	8	2	10						

Localities:

- LEF-1 to LEF-8 Lower Eagle Ford Group, lower Britton Fm., Turner Park Member
 LEF-9 Lower Eagle Ford Group, upper Britton Fm., Camp Wisdom Member
 LEF-10 Lower Eagle Ford Group, Tarrant Formation

CHAPTER 7

THE VERTEBRATE ICHNOFAUNA: COPROLITES

The lower Eagle Ford Group, specifically the lower Britton calcareous unit (Turner Park Member) has been found to contain a very rich ichno-assemblage composed of vertebrate coprolites in an outstanding state of preservation (Friedman, 2001; Friedman, 2002).

The abundance and occurrence of such vertebrate trace fossils is so rich throughout the entire study area and beyond (27 fossil localities in total in the DFW area) that it warrants a separate chapter in the present thesis and is presently the subject of ongoing research (Friedman and Hunt, 2004 in prep.)

INTRODUCTION

Coprolites (from the Greek: kopros, excrement, and lithos, stone) are ichnofossils or trace fossils. As such, they reflect the activities and behavior of long extinct organisms, as do tracks, eggs, burrows, trails, nests. The term coprolite was originally proposed by Buckland (1829), who subsequently demonstrated that they were of fecal origin (Buckland, 1835).

In the field, coprolites are most of the times overlooked, or if they are found, they are confused with concretions most of the time. However, coprolites can be distinguished by a combination of some or all of the following features: 1) extrusive external morphology e.g. spiral 2) internal structure is ordered, e.g. spiral or with longitudinal canals 3) longitudinal or spiral external striations 4) similarity of morphology to animal guts 5) morphology reflecting ranges of viscosity in modern fecal matter 6) flattening of ventral side 7) inclusions of organic matter 8) presence of evidence of gas bubbles or gas-escape

structures 9) composition of calcium phosphate 10) very fine-grained matrix (Amstutz, 1958). Coprolites along with all trace fossils are excellent environmental indicators often giving clues where no body fossils have been found or are scarce. Under certain sedimentological conditions (i.e. anoxic environments) the fecal remains of organisms can become fossilized. In some outcrops, these trace fossils are more abundant than other vertebrate body fossils. Coprolites become then an important tool as paleoenvironmental indicators, as well as stratigraphic markers. Due to the initial consistency (hard, formed, loose, or watery) of animal feces most of them will not become fossilized. Specifically, in a marine environment, feces must have a certain degree of viscosity and be rather formed or hard to be preserved as coprolites. Coprolites display a wide variation of forms and this can vary even within one single individual. However, as this study shows, there are certain guidelines as to the identity of the producer.

Regarding color, coprolites display a wide variety from black, brown, reddish, orange, gray, to some even with a lavender color. For the purpose of standardization a color rock chart was used (Goddard et al., 1980). Coprolites also vary in degree of petrification. Some are light in weight and show an undetermined degree of porosity, while others are rather indurated, stony. Some coprolites have a sideritic or siliceous mineralogical composition while most are either calcitic, or more commonly phosphatic in nature. All of this is affected by diagenetic variability, depending on the sedimentological as well as paleoenvironmental factors where the feces were first deposited. The size of the coprolites is an important factor, since it can provide information about possible animal producers. There are microcoprolites (i.e. fecal pellets) of just a few millimeters in diameter, small coprolites of only few centimeters in length, to presumed dinosaurian coprolites of several decimeters.

Regardless of the animal producer of the coprolite, during the digestive process, most food ingested will be thoroughly processed and this will become part of the ground mass of the coprolite. However, sometimes, depending on the producer, part of the food (i.e. bony parts, teeth, scales) will remain rather intact and will be excreted in the feces. The identification of such remains will provide clues as to trophic interaction between the organism producer and its paleoecosystem. The locally abundant well-preserved coprolites provide paleoecological information and have a definitive application in biostratigraphic correlation (Appendix I).

ANALYSIS

Vertebrate coprolites are abundant in 27 new outcrops of the Turner Park Member, Britton Formation of the lower Eagle Ford Group in North Central Texas (Friedman 2001; Friedman 2002; Friedman and Hunt, 2004 in prep.). The age of the sediments has been assigned to late Cenomanian based on planktic foraminiferal data as described elsewhere in this work.

More than 4000 vertebrate coprolites were collected. Many are complete and show great detail, others are just fragments. All the coprolites present a “greasy” and “powdery” texture. They weather out easily from the surrounding calcareous matrix and are most of the times excellently preserved (Figs.7.1 and 7.2). As a result, the external morphology of most coprolites is clearly visible and shows exquisite detail. Studies on coprolites are based mainly on external features, but thin sections were prepared as well in order to study their inclusions. The samples of coprolites chosen for thin sectioning were those with bone inclusions clearly visible. Broken or dented coprolites were chosen in order not to sacrifice the best specimens. The coprolites were chosen randomly from several of the localities



Figure 7.1. Vertebrate coprolite *in situ* (locality LEF-8).



Figure 7.2. Vertebrate coprolite *in situ* (locality LEF-6).

under study. The thin sections were examined under a petrographic microscope. They all present a fine apatitic groundmass and an assortment of undigested fragments of fish bones, as shown in Fig. 7.3. Most of them are not identifiable, but in some cases it is possible to recognize very small undigested fish vertebrae (Fig. 7.4). Fig. 7.5 shows in addition to bone fragments, calcite prisms probably belonging to an inoceramid bivalve. It is not clear whether the coprolite maker preyed on inoceramids or it merely represents debris swallowed accidentally. All the thin sections show undigested bone fragments that are rather intact and unetched.

The fine-grained groundmass of the coprolites does not react with diluted HCL (10%) as the surrounding matrix readily does. The presence of bone inclusions suggested also a phosphatic chemical composition. Based on thin section petrography as well as X-ray diffraction analyses, the mineralogical composition of the coprolites was determined to be primarily apatite with minor mineralogical constituents such as quartz and calcite, rare goethite, illite and mica. (Appendix II).

The specific gravity of 50 randomly chosen coprolites ranges from 2.6 to 2.9 g/ml. Since the specific gravity of apatite ranges from 3.1- 3.2 g/ml, the discrepancy between the apatite density and the density of the coprolites is most likely due to the porosity that many coprolites present.

Coprolite color was measured dry with a Rock Color Chart (Goddard et al., 1980). They range in size from 1.1 to 4.8 cm in length and are assigned to 7 morphotypes: discoidal, ovoid, elongated, oblong and spherical, folded, and amorphous (Fig. 7.6 and Table 3) (Friedman and Hunt, 2004 in prep.).

The coprolites are recognized as such on the basis of the presence of semiarticulated

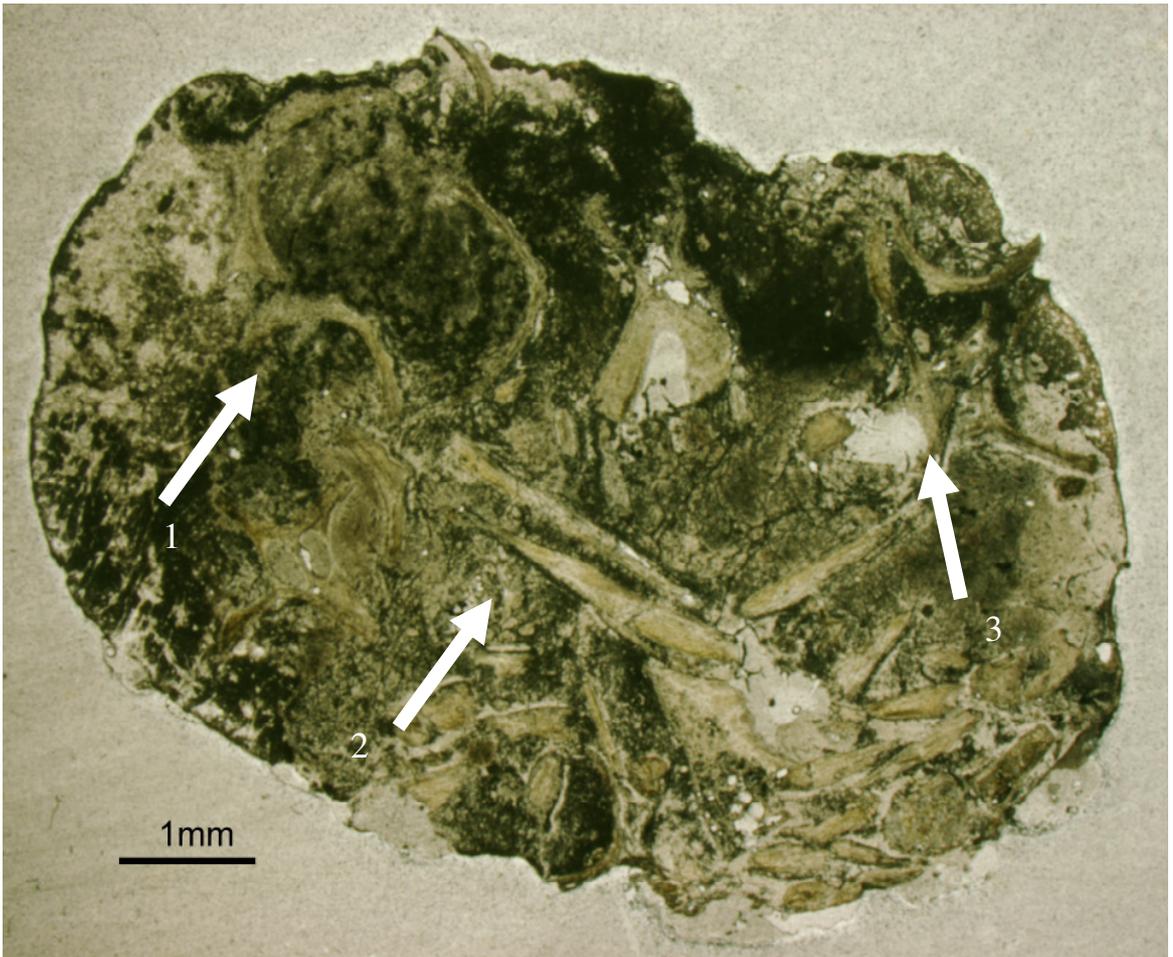


FIGURE 7.3. Coprolite thin section showing numerous fish bone fragments (1, 2, and 3).

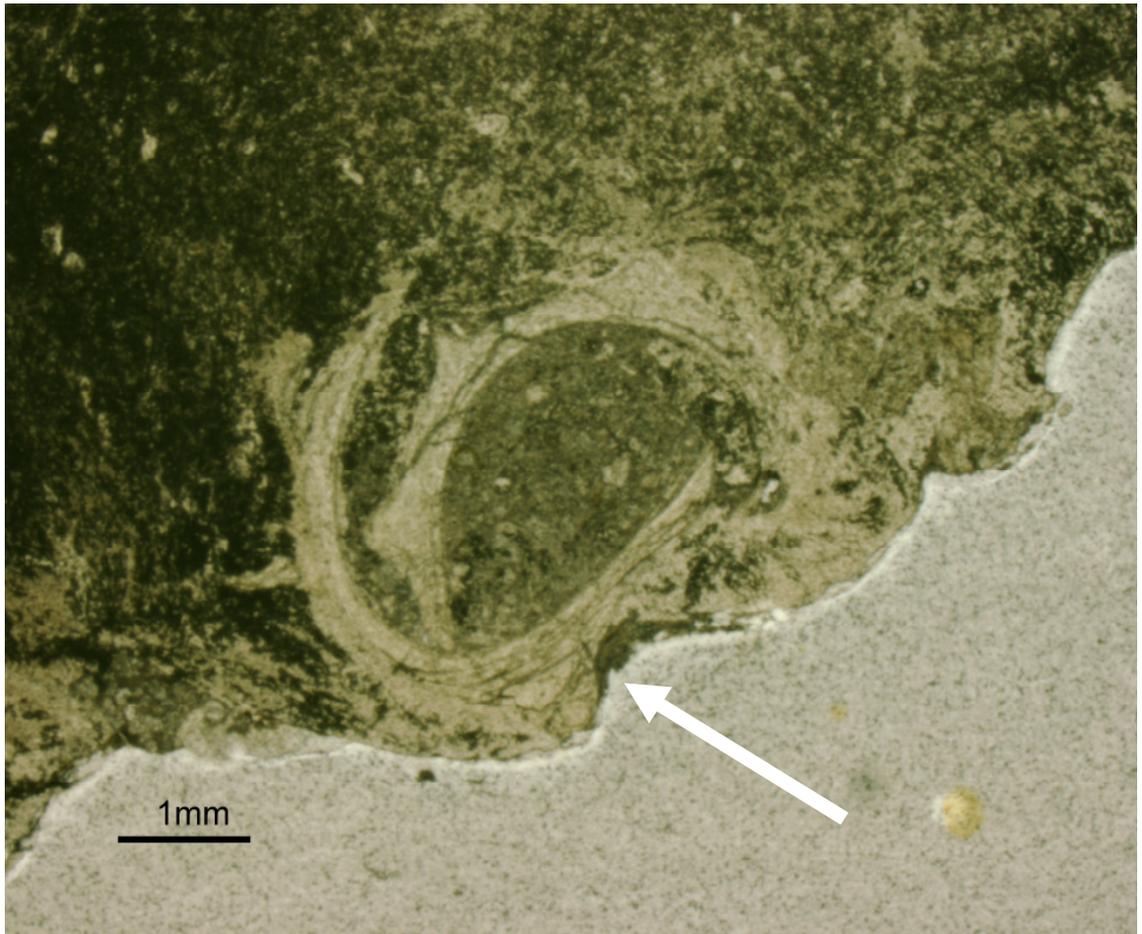


FIGURE 7.4. Coprolite thin section showing unidentifiable fish vertebra.

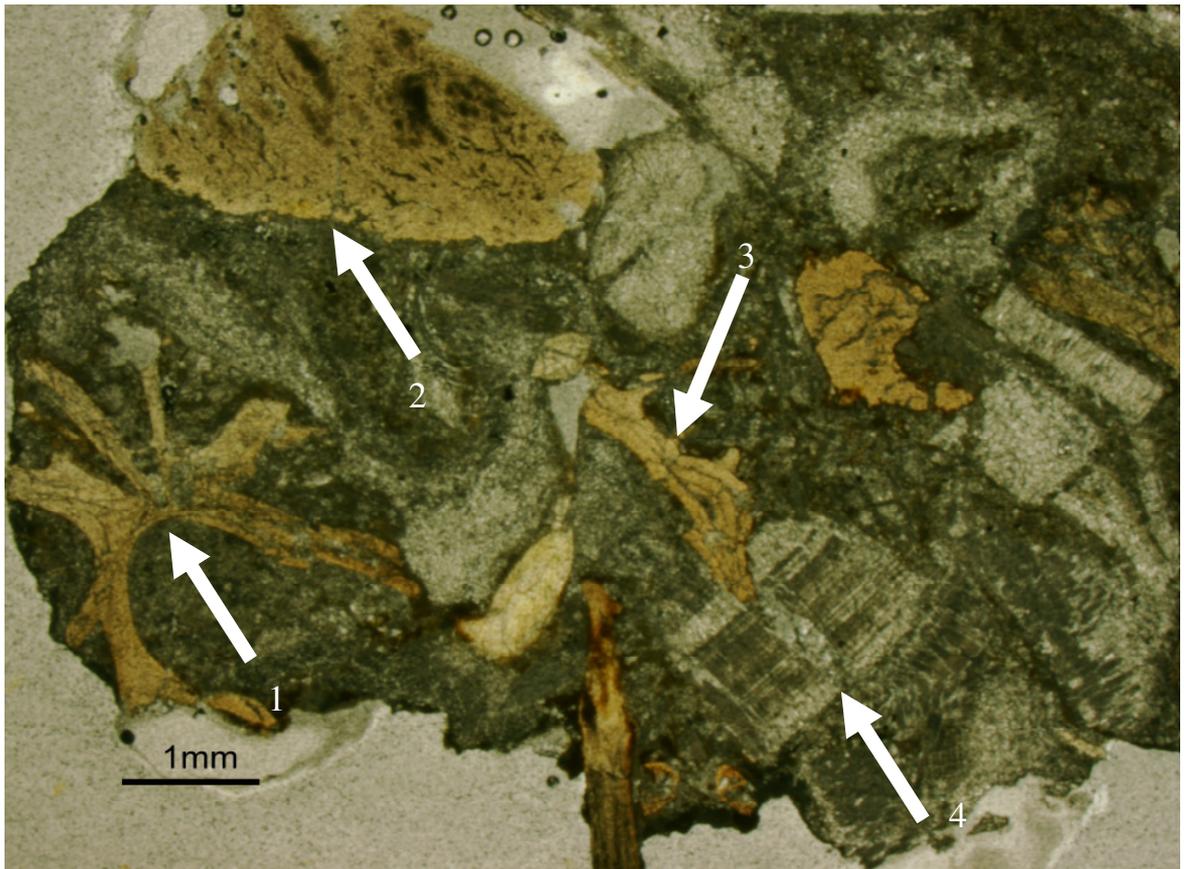


FIGURE 7.5. Coprolite thin section showing numerous bone fragments (1,2,3) and inoceramid calcite prisms (4).

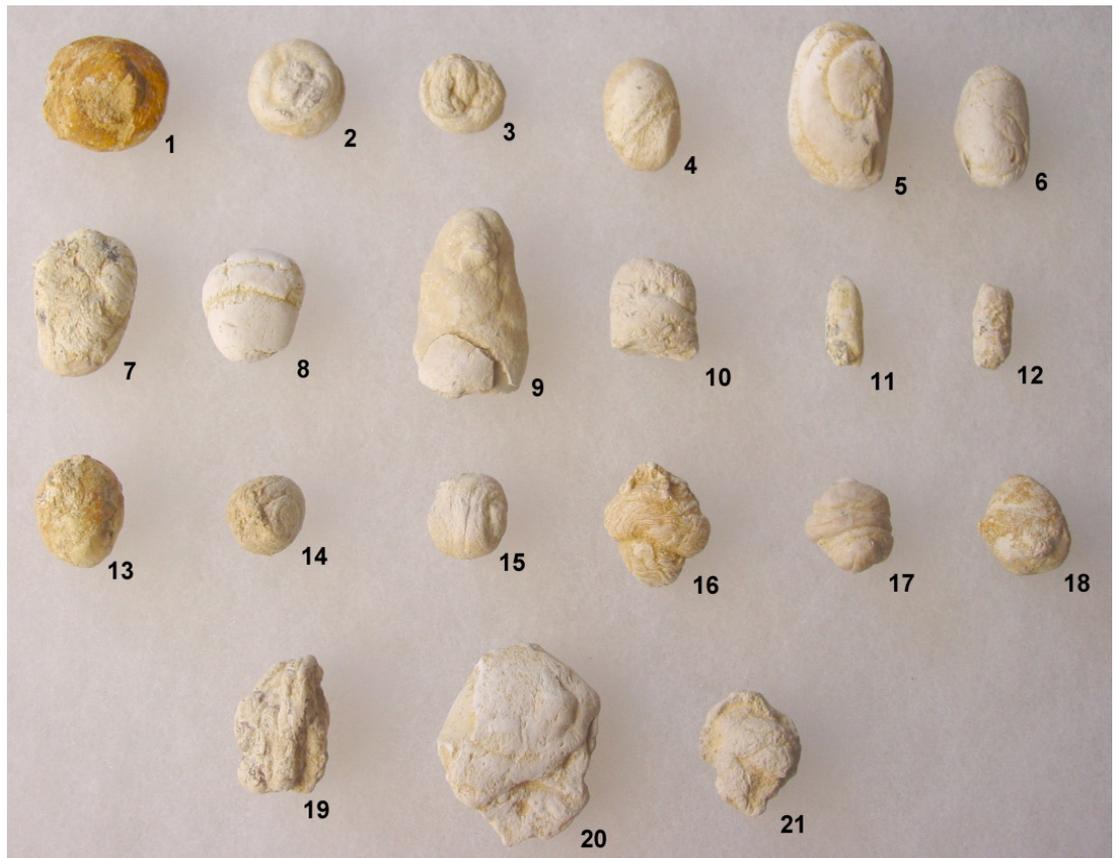


FIGURE 7.6. Coprolite morphotypes (1, 2, 3, discoidal; 4, 5, 6, ovoid; 7, 8, 9, oblong; 10, 11, 12, elongated; 13, 14, 15, spherical; 16, 17, 18, folded; 19, 20, 21, amorphous).

LENGTH	WIDTH	WEIGHT	COLOR	MORPHOTYPE	INCLUSIONS	REMARKS	FIGURE
31.7	29.3	11.5	Light brown 5YR 5/6	Discoidal	Rare	Spirally coiled fine striations	7.3.1
24.1	24.4	5.4	Very pale orange 10YR 8/2	Discoidal	Occasional	Spirally coiled	7.3.2
22.0	19.3	2.6	Very pale orange 10YR 8/2	Discoidal	Rare	Spirally coiled many fine striations	7.3.3
30.0	19.4	6.1	Grayish orange 10YR 7/4	Ovoid	Not Visible	Faint spiral striations	7.3.4
42.9	26.7	17.6	Very pale orange 10YR 8/2	Ovoid	Not Visible	Scroll-shaped	7.3.5
33.0	18.9	6.0	Very pale orange 10YR 8/2	Ovoid	Rare	Scroll-shaped	7.3.6
36.4	23.9	8.2	Grayish orange 10YR 7/4	Oblong	Abundant	Many fine striations	7.3.7
27.9	24.6	5.8	White N9	Oblong	Occasional	Spiral thick folds	7.3.8
45.7	27.4	15.1	Grayish orange 10YR 7/4	Oblong	Rare	Vaguely scroll-shaped	7.3.9
25.7	21.2	6.5	Very pale orange 10YR 8/2	Elongated	Rare	Many fine striations, thick spiral folds	7.3.10
24.6	10.3	1.9	Very pale orange 10YR 8/2	Elongated	Rare	No striations or folds, very smooth surface	7.3.11
21.8	10.0	1.6	Very pale orange 10YR 8/2	Elongated	Occasional	No striations or folds	7.3.12
25.4	20.6	7.4	Light brown 5YR 5/6	Spherical	Abundant	Occasional fine striations, thick spiral folds	7.3.13
19.3	18.1	3.4	Grayish orange 10YR 7/4	Spherical	Not Visible	Faint spiral striations	7.3.14
20.2	18.7	3.8	Very pale orange 10YR 8/2	Spherical	Abundant	Faint spiral folds	7.3.15
29.1	25.1	5.8	Grayish orange 10YR 7/4	Folded	Rare	Thick folds, many fine striations, rolled-up	7.3.16
24.1	20.7	4.7	Pale yellowish brown 10YR 6/2	Folded	Not Visible	Thick folds, rolled-up	7.3.17
23.2	21.8	6.3	Grayish orange 10YR 7/4	Folded	Not Visible	Vague thick folds, rolled-up	7.3.18
30.3	20.2	7.3	Very pale orange 10YR 8/2	Amorphous	Abundant	Many fine striations, twisted folds	7.3.19

TABLE 3 Coprolites Description (All measurements for length and width are in mm, for weight in grams.) Color was measured using Rock Color Chart.

TABLE 3 Coprolites Description (Continued).

LENGTH	WIDTH	WEIGHT	COLOR	MORPHOTYPE	INCLUSIONS	REMARKS	FIGURE
43.4	37.0	12.7	White N9	Amorphous	Not Visible	No folds or striations, very smooth surface	7.3.20
28.2	23.3	5.6	Very pale orange 10YR 8/2	Amorphous	Occasional	No folds or striations	7.3.21
28.5	26.3	4.73	Grayish orange 10YR 7/4	Discoidal	Occasional	occasional fine striations	N/A
27.4	26.3	4.75	Dark yellowish orange 10YR 6/6	Discoidal	Rare	Many fine striations	N/A
17.7	17.5	2.03	Grayish orange 10YR 7/4	Discoidal	Not visible	Thick spiral concentric folds	N/A
21.2	18.9	2.26	Grayish orange 10YR 7/4	Discoidal	Not visible	Fine striations, thick folds	N/A
16.7	15.4	1.22	Grayish orange 10YR 7/4	Discoidal	Rare	Many fine striations	N/A
18.0	17.0	1.76	Very pale orange 10YR 8/2	Discoidal	Rare	Faint fine striations	N/A
18.2	16.8	1.53	Very pale orange 10YR 8/2	Discoidal	Rare	Faint fine striations	N/A
30.6	17.5	5.02	Very pale orange 10YR 8/2	Ovoid	Rare	Faint striations	N/A
33.3	17.6	6.58	Grayish orange 10YR 7/4	Ovoid	Abundant	No striations	N/A
27.5	18.8	6.16	Dark yellowish orange 10YR 6/6	Ovoid	Occasional	No striations	N/A
32.1	16.8	5.09	Grayish orange 10YR 7/4	Ovoid	Occasional	Fine striations	N/A
32.1	21.2	5.85	Grayish orange 10YR 7/4	Oblong	Occasional	Rare striations	N/A
21.3	18.2	1.97	Very pale orange 10YR 8/2	Oblong	Occasional	Faint striations	N/A

TABLE 3 Coprolites Description (Continued).

LENGTH	WIDTH	WEIGHT	COLOR	MORPHOTYPE	INCLUSIONS	REMARKS	FIGURE
30.6	22.6	4.86	Grayish orange	Oblong	Not visible	Many striations, thick folds	N/A
26.8	19.6	5.50.	1 OYR 7/4 Pale yellowish brown	Oblong	Rare	Many striations	N/A
21.6	21.2	4.24	10YR 6/2 Grayish orange	Spherical	Not visible	Many striations	N/A
18.7	16.8	3.02	10YR 7/4 Pale yellowish brown	Spherical	Rare	Few striations	N/A
16.2	15.7	2.19	10YR 6/2 Grayish orange	Spherical	Rare	Faint striations	N/A
24.5	15.7	5.10	10YR 7/4 Very pale orange	Elongated	Rare	No striations	N/A
25.2	11.6	1.66	10YR 8/2 Grayish orange	Elongated	Not visible	Rare striations	N/A
28.6	11.7	2.02	10YR 7/4 Grayish orange	Elongated	Not visible	Rare striations	N/A
19.8	6.6	0.65	10YR 7/4 Very pale orange	Elongated	Not visible	Occasional longitudinal striations	N/A
18.9	18.6	2.62	10YR 8/2 Moderate yellowish brown	Folded	Occasional	Many striations	N/A
16.9	16.8	1.91	10YR 5/4 Dark yellowish orange	Folded	Rare	Many convoluted striations	N/A
23.2	18.7	2.50	10YR 6/6 Very pale orange	Folded	Not visible	Many transverse striations	N/A

and disarticulated fish remains within a very fine-grained groundmass. Their content is dominated by undigested fish remains belonging to different taxa. The inclusions are small vertebrae and assorted bones of unidentified fishes, but also *Enchodus* palatine teeth and pachyrhizodontids fin fragments have been identified within the coprolites. No coprolite was found to contain shark, ray or in general invertebrate material. All coprolites contain marine faunal remains, no non-marine remains were found.

Many coprolites present thin spiral striations and folds. Primitive fishes including sharks, gars and lungfishes possess spiral intestinal valves (Gilmore, 1992), hence the coiled shape of most of the coprolites. The spiral valve is absent in teleost fishes (Romer and Parson, 1986). Under favorable conditions the form of the feces retains outlines which are characteristic. It is also clear that, the greater the number of chambers in the intestinal valve, the greater the number of coils in the feces (Price, 1927).

One of the most distinguishing features of shark coprolites is the unique way in which they are twisted around themselves. This pattern is produced by the internal structure of the shark intestine. In sharks, a sheet of tissue, called a valve, increases the surface area of the intestine for more efficient absorption of food. The exact form of the valve is different in the various major groups of sharks. The more primitive form (e.g., in many lamniforms) is called a spiral valve because the valve coils like an auger. More advanced sharks (e.g., many carcharhiniforms) have a valve that is rolled like a scroll. As food passes through the intestine unabsorbed material is molded by the valve. Coprolites from a spiral valve have a closely spaced spiral groove, while those from a scroll valve have a rolled appearance. So, the other sharks have loosely coiled valves that produce coprolites of intermediate form (i.e. with a very widely spaced spiral groove) (Kent, 1994). Most of the coprolites (70 %)

collected in the lower Britton formation show a more-or-less spiral or coiled morphology. These bone-bearing coprolites are attributed to sharks and/or other fishes, due to their morphology (spiral or scroll-folded), inclusion content and due to the fact that after coprolites, sharks and other fish remains (teeth, vertebrae, fins, etc.) are the most abundant body fossil found in the faunal assemblage of the Britton Fm. of the lower Eagle Ford Group. The coprolites (30 %) that do not present spiral folding or striations have undergone severe weathering and therefore the outer layers were removed or they were produced by teleost fishes, that lack spiral valves in their intestines.

The ichthyofaunal assemblage of the locality is multitaxic and includes the genera:

Cretoxyrhina, *Cretodus*, *Squalicorax*, *Cretolamna*, *Carcharias*, *Enchodus*, *Ptychodus*, *Protosphyraena*, *Xiphactinus*, *Pachyrhizodus*, as well as the oldest saurodontid fish record in North America (Stewart and Friedman, 2001). Reptile remains have also been found belonging to plesiosaurs, turtles, as well as the primitive and rare, *Coniasaurus*.

Coprolite horizons possess a potential application not only in paleoecological reconstructions but also in biostratigraphic correlation (Price, 1927; Johnson, 1934; Hunt et al., 1993). This work shows this potential application of coprolite horizons in biostratigraphic correlation of 8 new outcrop sections of the lower Britton Formation (Turner Park Mbr.) within an area of 64 kilometers in North Central Texas (Appendix I).

CONCLUSIONS

1. The bone-bearing coprolites of the lower Eagle Ford Group are attributed primarily to sharks and/or other primitive fishes, due to their basic spiral or scroll-shape morphology. The coprolites that do not present this morphology, have suffered some degree of decay

(i.e., weathering) or are attributable to teleost fishes, which do not possess spiral valves in their intestinal tract.

2. Coprolites are valuable indicators of the paleodiet of extinct organisms and most important they provide additional information in reconstructing ecosystem relationships of fossil flora and fauna of a given place in space and time.

3. The abundance of coprolites in some environments is indicative of specific sedimentological and environmental conditions. Poorly-oxygenated conditions prevailed during the entire Eagle Ford Group deposition. The excellent preservation of the vertebrate coprolites is attributed to their initial viscosity and semi-solid form as they were deposited, lack of coprophagic organisms, and to their deposition in a shallow-marine, low-energy, thinly laminated dysoxic/anoxic environment, where practically no bacterial decay took place. This scenario is consistent with the C/T global anoxic event worldwide.

4. The recognition of a the coprolite-rich horizon in the lower Eagle Ford Group provides a potential correlation proxy for the lower Eagle Ford Group in North Central Texas. The coprolite-rich horizon is easy to recognize, since the coprolites are very abundant and are a distinctive sedimentological feature that appears in every single locality of the lower Britton (Turner Park Mbr.) under study. Although the origin of the coprolite-rich interval is unclear, possible factors include anomalous water chemistry (i.e., anoxic bottom), or a deposit formed simply by winnowing currents is not discarded.

CHAPTER 8

PALEOECOLOGY

A large epeiric sea existed on the western North American craton, with chalks forming in the central and southern parts of this seaway. This great interior sea divided North America into an eastern and western continent during the Late Cretaceous. The Austin Chalk and Eagle Ford shales formed in the southern end of this epeiric sea, on a gently sloping ramp that deepened into the Gulf of Mexico. The seas during Eagle Ford deposition were comparatively shallow and were filled with abundant invertebrate and vertebrate life forms. These ancient Texas environments produced one of the most intensely fossiliferous marine deposits found anywhere in the world.

HABITAT: EAGLE FORD TIME AND SEA

In their struggle for life all living entities past and present are linked to one another and to their surroundings in cycles and complex networks which constitute what we call ecosystems. Paleoenvironmental reconstructions must examine a wide variety of data sets, analyze them and synthesize them into a consistent and coherent picture of an ancient world. Sedimentary structures found within the lithofacies shed light on past environments. The fossil content - macrofossil, microfossil and ichnofossil - of the units must be taken into account. No single piece of information can stand by itself, unsupported by others (Thompson et al.,1991).

Several considerations have to be taken into account, such parameters as salinity, oxygen

content, temperature, pressure, nutrient supply, tidal fluctuations, turbulence, biotic character and abundance can be observed directly by an ecologist, but can only be indirectly studied or inferred with the older materials available to the paleontologist.

SALINITY

Distribution of aquatic organisms is closely controlled by salinity and variations in salinity. Fossils are a powerful tool for distinguishing marine from nonmarine environment in the stratigraphic record. Salinity tolerances - particularly of stenohaline organisms do not seem to change readily through time, an association of fossils whose modern descendants and/or relatives are marine indicates that the ancient environment in which they live was marine (Heckel, 1972).

The number of species is greatest at normal marine salinity and declines at either higher or lower salinities. Nevertheless, salinity tolerance of marine organisms differs from taxon to taxon (Heckel 1972).

The faunal fossil assemblage found in the study area indicates exclusively a marine environment and as such should contain representatives of a normal salinity (3.5%) marine fauna (i.e., corals, gastropods, sponges, brachiopods, bryozoans, cephalopods, echinoderms). However, the normal salinity of this ancient environment is questionable. It is inferred that it might not have been normal due to the fact that known stenohaline organisms (including echinoderms, corals or sponges) are not present in any of the localities under study.

There are essentially no benthic invertebrate organisms in the study area (except for inoceramids and rare oysters). This may be due primarily to poorly oxygenated waters rather than to abnormal salinity. Apparently the greatest biodiversity thrives in an environment that is stable, one in which all factors, such as temperature, salinity, oxygen, sedimentation, etc.

vary only within a narrow range over a long period of time. Most organisms cannot tolerate a rapid and extreme change of physical environmental factors (Heckel, 1972).

The fossil assemblage at the localities under study are all marine fauna, except for occasional taxonomically unidentifiable woody plant remains. However, Heckel (1972) pointed out that whereas nonmarine fossils may readily be carried into a marine environment, marine organisms are rarely transported into nonmarine environments. Thus a minor amount of exclusively nonmarine fossils in a marine assemblage does not invalidate a fully marine interpretation of the environment.

TEMPERATURE

The middle Cretaceous (~105-89 Ma) encompasses one of the warmest climate intervals of the Phanerozoic. Polar ice caps and related climate contrasts were not present in the Cretaceous. The time was characterized by some of the highest atmospheric CO₂ concentrations, warmest climate, and the highest sea level in Earth's history (Haq et al., 1987; Clarke and Jenkyns, 1999). The mid-Cretaceous was a period of unusually active tectonism that drove enhanced volcanic outgassing and high seafloor spreading rates. This intense tectonic activity is coincident with dramatic events in the marine environment, oceanic anoxic events, high biological turnover rates, and a thermal maximum (Poulsen et al., 2001). The interval spanning the Barremian to Turonian (~125-88 Ma) is known as the mid-Cretaceous "green-house" world (Barron, 1983). Both marine and terrestrial proxies indicate that the mid-Cretaceous climate was significantly warmer with a smaller equator-to-pole temperature gradient than the modern climate. It has been suggested that this resulted from an anomalous amount of oceanic volcanism, leading to 3 to 12 times higher atmospheric levels of carbon dioxide during the Cretaceous than the present (Berner,

1992). The transition from the Cenomanian to the Turonian marks a major turning point, with globally dispersed oxygen isotope data indicating that the mid-Cretaceous warming trend ceased at the C/T boundary and was followed by a deterioration at least until the Maastrichtian (Jenkyns et al., 1994).

Cretaceous climate and sea level history are also important to the production of hydrocarbon source rock. The warmer middle Cretaceous climate and higher sea level coincide with an increase in accumulation rates of organic carbon. Approximately 60% of all known oil reserves are from the Cretaceous (Irving et al., 1976; Mysak and Wang, 2000).

Since the mid-Cretaceous is generally thought to have been warmer than the present (6 to 12° C warmer than today's) (Mysak and Wang, 2000) and with higher concentrations of atmospheric greenhouse gases such as carbon dioxide, it has been suggested that this time period could be used by modelers as an analogue for future climate change. However, new ocean paleotemperature records for Cenomanian sites in the Atlantic and Pacific oceans, together with a re-evaluation of published data, cast doubt on the idea that the Cretaceous period was generally warmer. The findings suggest that the Cretaceous cannot be used as an analogue for future climate change (Sellwood et al., 1994).

DEPTH

Sea level was higher during most of the Cretaceous than at any other time in Earth history, with sea levels as much as 250 m higher than present with fluctuations of 100 m over 10 Myr (Haq et al., 1987). The high Cretaceous sea level is thought to have been primarily the result of water in the ocean basins being displaced by the enlargement of the mid-oceanic ridges. As a result of higher sea levels during the Late Cretaceous, marine waters inundated the continents, creating relatively shallow epicontinental seas in North America, South America,

Europe, Russia, Africa and Australia.

Epicontinental or epeiric seas lie upon the continents and spread for perhaps 1600 to about 3200 km in breadth over the craton. As they probably only rarely reached depths of 200 m and generally may have attained only 30 m. Epicontinental seas are rare today, but they provided much, perhaps a majority in some continents of the preserved and accessible stratigraphic record (Heckel, 1972).

Climate, land-sea distribution, sea level, and fauna were different during the Cretaceous than they are today. Barron et al., (1980) suggested that during the middle Cretaceous, the paleogeographic distribution was different and continental patterns were more symmetrical than today and that the sea level was 100-200 m higher than present and about 20% of the continental area was flooded.

Kauffman (1969) reported that the thick dark calcareous and non-calcareous shales of the upper Cretaceous in the Western Interior were deposited in water depths of 90 to 150 m, and that the basinward calcareous shales, calcarenites, flaggy impure limestones, and massive fine-grained limestones of the Greenhorn and Niobrara cycles represent deposition within the photic zone. In order to explain relatively shallow water in the central part of the Western Interior seaway, Kauffman (1969) suggested that the basin center did not subside as much as the marginal areas, and that it even rose from depths of 90 to 150 m at mid-transgression to depths of 30 to 60 m during maximum transgression. This process would have reversed as the sea regressed. In Kauffman's view, a similar sedimentary model is applicable throughout the Texas-Western Interior Seaway as far south as Mexico.

DISOLVED OXYGEN

The Oceanic Anoxic Event (OAE2) of the middle Cretaceous is characterized in North Central Texas, as well as elsewhere in the world, mainly by dark sediments enriched in organic matter due to their enhanced preservation during global anoxic conditions within the water column. Oxygen deficiency in bottom waters occurs when the oxygen demand created by the decay of metabolizable organic matter exceeds the rate of oxygen supply. Oxygen consumption rates exceed oxygen supply with excess organic matter, resulting in decreasing dissolved oxygen concentrations. The pH of sea water becomes acidic (5.5-7) in contact with black mud.

Poorly oxygenated waters have a pronounced effect on benthic biota. If anoxia is prolonged, conditions at the sediment-water interface, and in the water column above, may permit survival of sulfate reducing bacteria, normally found in deeper sediments. These bacteria produce H_2S as a metabolic by-product, which is toxic to aerobic life. The formation of hydrogen sulfide causes the almost total extermination of benthic faunas. Although H_2S is toxic to many organisms, a world-wide sulfide biotope exists below the aerobic surface layer of many sediments. Permanent residents of this interstitial habitat include the prokaryotic phyla: Cyanophyta, Eubacteria, Spirochaeta, Myxobacteriae, and Actinomycota. Many eucaryotic Protists are represented as well as fungi. Two metazoan phyla, platyhelminthes and aschelminthes are also present (Fenchel and Riedl, 1970). These two authors proposed that the members of the sulfide biome (>12 phyla) represent the oldest biosystem on earth. In 1971 Rhoads and Morse discussed the evolutionary and ecological significance of oxygen- deficient marine basins, and described three biofacies for the distribution of benthonic invertebrates in the low oxygen regions of the Black Sea, Gulf of California

and continental borderland basins of California. These biofacies are as follows:

1. Anaerobic zone is where the concentration of oxygen is less than 0.1 ml O₂/l H₂O.

This zone lacks metazoans- the azoic zone.

2. Dysaerobic zone is where the oxygen concentration is between 0.3 to 1.0 ml O₂/l H₂O.

This zone supports a low diversity assemblage of small soft bodied infauna.

3. Aerobic zone is where the oxygen concentration is greater than 1.0 ml O₂/l H₂O. Here diversity increases, especially for calcareous species.

The saturation of oxygen in sea water depends upon two factors salinity and temperature (Rhodes and Morse, 1971).

Anoxic conditions prevailed during Eagle Ford deposition. Environmental indicators include the generally black color of the Eagle Ford shale, millimeter laminations, a general absence of infauna, authigenic pyrite, and the high ratio of pelagic to benthic fossils. (Charvat and Grayson, 1981).

Even though stressed conditions prevailed during the deposition of these organic-rich shales, these conditions did not prevent a benthic colonization of epifaunal inoceramid bivalves (localities LEF-1 to LEF-8) and oysters *Pseudoperma congesta* (localities LEF-2, LEF-3, LEF-4) as well as the presence of a rare horizon of bioturbated sediments (*Chondrites*) in LEF-5. The duration of benthic colonization of the abundant inoceramids and their occasional epizoan oysters, nevertheless in the case of the *Chondrites* horizon (LEF-5) given the thickness of the horizon (15 cm) the duration of the colonization event may have been only few months. Seasonal fluctuations may have caused short-term variations in salinity and also in oxygen availability within the fauna.

The Eagle Ford with its finely laminated shales and lime mudstones in general lacking

bioturbation is typical of deposits of the anaerobic and the lower part of the dysaerobic zone. According to Charvat (1985) the Eagle Ford Group is representative of deposits of the stagnant marine environment in epicontinental seas. The late Cenomanian expanding oxygen-minimum zone sequentially affected the biota in an order dependent on their particular niche on the sea floor or in the water column. The oxygen-minimum zone did not reach surface waters (Leary et al., 1989). This is evident by the rich pelagic ichthyofauna in the study area, as well as by the presence of large marine reptiles (plesiosaurs) and marine lizards (coniasaurs).

Heckel (1972) reported the existence of unfossiliferous black shales (as is the case of locality LEF-9) and stated that the origin of unfossiliferous black shales has long been a subject of controversy. Black color in these shales is caused by a relatively high content of either unoxidized organic matter or finely divided iron sulphide. Both substances are preserved or formed in large quantities only under reducing conditions which develop in "anoxic" environments where the normal processes of oxygenation have ceased to operate on a scale sufficient for continual oxidation of organic material normally accumulating on the bottom. As soon as oxygen is depleted in the environment, anaerobic bacterial decay of organic matter begins on a larger scale, producing hydrogen sulphide which, along with the lack of oxygen, fouls the substrate and bottom waters and makes them inhospitable for benthonic and nektobenthonic life. One model for development of such an environment is loss of circulation that normally brings oxygenated waters to the bottoms of most modern bodies of water including the oceanic depths. Loss of bottom circulation takes place where water bodies, or portions thereof, are barred or silled, and density stratification occurs as water of slightly lower salinity flows in over the denser, normal marine water which is

prevented from flowing out of the basin along the bottom by the barrier. This density stratification is stable, and as vertical circulation ceases, bottom stagnation eventually results. The barrier may be a submarine topographic high, or sill. The general end result is loss of bottom life and deposition of black shale (Heckel, 1972).

Another model for attaining an oxygen-depleted environment in shallow water involves offshore upwelling of nutrient-rich and oxygen-poor deeper water in response to driving of surface waters seaward by constant strong winds from the adjacent land. Upwelling brings nutrients, especially phosphorous and nitrogen, into the photic zone where they are utilized by phytoplankton which blooms in enormous numbers and produce toxins that can cause mass mortality of the neritic fauna, particularly fish. The great amount of decaying organic material both from the blooms and the mass mortalities further depletes the already oxygen-poor water brought up from the oxygen-minimum zone. As a result of either periodic or continual upwelling the benthonic fauna is eliminated, and much organic matter accumulates in the sediment, in this case on account of circulation of anoxic water. Although upwelling is known today mainly in pericontinental seas along the west coast of continents, this model has been extended to ancient epicontinental seas (Heckel, 1972).

MICROFAUNA

The Cenomanian/Turonian boundary corresponds to the second largest extinction event in the Cretaceous (Raupe and Seposki, 1986). The area under study contains a stressed microfauna. The lower Eagle Ford microfaunal assemblages found at the localities under study (LEF-1 to LEF-8) shows a marked dominance of planktic foraminifers, mostly (*Hedbergella* and *Heterohelix*) and no benthic microfossils (Table 1, p.28).

These epipelagic foraminifers, *Heterohelix* and *Hedbergella*, have been described as shallow-marine opportunistic organisms able to survive in stressed environmental conditions (Nederbragt et al., 1998; Gasinski, 1997; Rodriguez Lazaro, 1998).

An increase in abundance of *Hedbergella delrioensis* provides an indirect argument for the presence of low-oxygen levels. It is a form that can be assumed to be tolerant of low-oxygen conditions since they were able to survive the anoxic conditions during the Cenomanian/ Turonian Oceanic Anoxic Event (Nederbragt et al., 2001).

Heterohelix moremani, on the other hand, was the only biserial planktonic foraminiferal species from its first appearance in the late Albian up to the Cenomanian/Turonian boundary. Within that time, it increased gradually in abundance relative to other planktonic foraminifera. It is generally rare in upper Albian sediments, common in most of the Cenomanian and very abundant in sediments representing the latest Cenomanian Oceanic Anoxic Event (Nederbragt, 1991, Nederbragt et al., 1998). *H. moremani* was found abundantly in practically all sections under study (Table 1, p.28). During the anoxic event, *Heterohelix* started to diversify. Heterohelicids in general, remained an abundant part of the planktonic foraminiferal faunas until the end of the Cretaceous (Nederbragt, 1991; Nederbragt et al., 1998). Maximum rain rates of *H. moremani* during the Cenomanian/ Turonian OAE2 show that this species was an opportunist that did well in extreme conditions (Table 1, p.28). Stable oxygen and carbon isotope measurements on foraminiferal species indicate that *H. moremani* was a surface water dweller at least in part of its geographic range (Nederbragt et al., 1998).

Leckie (1987) concluded also that *H. moremani* is an organism indicative of epicontinental sea deposits. Eicher and Worstell (1970) also suggested that *Heterohelix* is a form which

frequented shallower levels of the water column. They determined that *Heterohelix* was first among planktonics to enter a newly transgressed region and last to be forced out during a regression.

Hedbergellids and heterohelicids were opportunist foraminifera that were found abundantly in the study area as shown in Table 1 (p.28). Opportunists are eurytopic taxa employing r-selection in reproduction, and thus capable of exploiting open ecospace in stressed environments through prolific generations of offspring. Eurytopes (ecological generalists) are those taxa which have a broad adaptive range, and which are little affected by the perturbations associated with mass extinctions (Harries and Kauffman, 1990). Opportunistic taxa (r-strategists) have the greatest advantage in surviving changed conditions. These conditions may be related to such factors as drastic sea-level changes, severe flooding episodes, extremely high temperature and/or enhanced hydrological and nutrient cycles, i.e. conditions in which new niches are created for successful colonization by 'visitors'. The mid-Cretaceous was a period that was rich in such extreme global events with global warming, relatively fast opening of the South Atlantic, numerous upwelling phenomena, transgression and regression episodes, and related dysoxic-anoxic conditions. These changing conditions created possibilities for 'visitors' (both Boreal and Tethyan) to successfully colonise newly created niches. Rapid changes of biotope conditions produced a wide variety of new niches, which were immediately colonised by opportunistic r-strategists (Gasinski, 1997; Ekdale, 1985). In almost all assemblages, including those with *Rotalipora*, non-keeled *Hedbergella* became the dominant component of all 'typical' Tethyan microassemblages (Gasinski, 1988, 1992). The localities under study confirm this Scenario for North Central Texas during the mid-Cretaceous.

In 1972 Sliter defined a classical scheme in which *Hedbergella/Heterohelix* are classified as epipelagic and *Rotalipora/Praeglobotruncana* as bathypelagic forms.

The microassemblage at the study area shows rare rotaliporids (*Rotalipora cushmani* and *R. greenhornensis*) in a *Heterohelix/Hedbergella*-dominated assemblages. According to Hueber et al., (1999) the rotaliporid extinction occurs at the level where middle bathyal temperatures are estimated to have increased from 15 to 19 °C, which is warmer than any other time during the Cretaceous and Cenozoic. This deep water warming may have caused a breakdown in the vertical structure of the water column, and could explain the extinction of deeper dwelling planktic species, including *Rotalipora*. On the other hand sea surface temperature estimates remained steady throughout the Cenomanian/Turonian transition varying between 23 to 26 °C (Hueber et al., 1999). Rotaliporids are among the taxa to have been environmentally most sensitive of all the mid-Cretaceous planktonic foraminifers (Leckie, 1987). Rotaliporids were also rarely found in the microfaunal assemblage of the area under study (Table 1, p.28). Micronutrient availability was also somehow restricted during the upper Cenomanian, as indicated by the low diversity of the microfaunal assemblage. Low diversity foraminiferal populations consisting of a high planktic/benthic ratio indicate stressed conditions during this event.

MICROFLORA

Certain planktonic protistid groups (i.e. coccoliths) can also be used as paleoecological indicators (i.e., depth) since they live only in near-surface waters because they require sunlight for photosynthesis. The euphotic zone is the area in which sufficient sunlight can penetrate to support the growth of phytoplankton, which is approximately 100 meters in the marine environment. Many of the localities under study (LEF-1 to LEF-8) contain

excellently preserved calcareous nannofossils and were also used in determining the age of the sediments of some of the localities under study (Watkins written comm., 2001 and 2003). This autotroph microfauna is part of the phytoplankton that constitutes the base of the food chain in the oceans past and present.

INVERTEBRATE MACROFAUNA

CLASS BIVALVIA

Even though anoxic/dysoxic conditions prevailed during the deposition of the lower Eagle Ford, these conditions could have been interrupted by short oxygenated periods indicated by benthic colonization of *Inoceramus* (localities LEF-1 to 8) and its epizoan the oyster *Pseudoperna congesta* (localities LEF-2, LEF-3 and LEF-4). *P. congesta* is a small irregular oyster attached on other shells (i.e. inoceramids) in colonies. Large attachment areas are a major feature of this oyster. Also a horizon of bioturbated sediments (*Chondrites*) was found at locality LEF-5. The paucity of infaunal and epibenthic invertebrates, other than inoceramids and occasional oysters, reflects an inhospitable sediment-water interface, possibly as a result of low oxygen levels (Harries and Kauffman, 1990).

Inoceramid bivalves are the most common benthic macroinvertebrate in the lower Eagle Ford localities under study. Kauffman (1990) reported that Inoceramidae favored basinal fine-grained basins with dysaerobic oxygen levels. Sageman (1989) and Kauffman and Sageman (1990) suggested that certain taxa, chiefly Inoceramidae, evolved highly opportunistic life strategies to exploit basinal paleoenvironments. It has also been suggested that inoceramids may have depended on a relationship with photosynthetic or chemotrophic symbionts, and anomalous stable isotope signatures of their shell carbonate has been cited

as evidence of such a relationship (MacLeod and Hoppe, 1992). Taxa capable of bacterial-chemosymbioses might have been able to fulfill their physiological requirements in otherwise inhospitable environments created by factors forcing mass extinction (Harries and Kauffman, 1990).

Traditionally, it has been thought that these extinct bivalves most likely rested on the seafloor on their left valves. Intriguingly, though, these unusual bivalves show common encrustation by the oyster *Pseudoperma congesta*, sometimes on both valve exteriors (Hattin, 1986). If the left valve had been resting on the soft substrate, it would not be expected that encrusters such as these oysters would have colonized it. This has led to the hypothesis that the inoceramids potentially had a pseudoplanktonic life mode, attached to floating logs or algae, or that the valves were flipped over by large predators such as mosasaurs in search of food, which would have allowed colonization of both valves by encrusters (Hattin, 1982 and 1986). This same author finds however, that the preponderance of evidence supports the original hypothesis that these enormous bivalves reclined on the seafloor during life.

CLASS CEPHALOPODA

Ammonites are not abundant in the localities under study. It is not clear if this is due to lack of fossilization or that they actually were rare inhabitants in the lower Eagle Ford sea. Ammonites were all nektonic organisms in adult stage. If the upper part of the water column could support a diverse fauna it is probable that they were also present but for unknown reasons few have been preserved. Nevertheless the few found at the study area provided an additional source for biozonation (Chapter 4, p. 32).

TRACE FOSSILS

Trace fossils are useful indicators of depositional environments. They are exclusively autochthonous. Unlike body fossils, trace fossils cannot be transported or reworked so they provide highly dependable indicators of final conditions experienced before ultimate burial of sediments and as “way-up” indicators.

In the localities under study there are practically no signs of bioturbation, suggesting oxygen stress below the sediment-water interface. One single bioturbated horizon, 18 cm in thickness was observed at locality LEF-5. It was readily identified as *Chondrites*.

Bromley and Ekdale (1984) indicated that *Chondrites* classically is a trace fossil that may indicate poorly-oxygenated settings. *Chondrites* is a regularly branched burrow system constructed by endobenthic deposit-feeding animals of unknown taxonomic affinity. It is a distinctive trace fossil, characterized by a root-like structure of branching shafts and tunnels in which the branching angle is a relatively constant 30 degrees to 40 degrees and the shaft or tunnel diameter, which may be 0.1 to 10 mm, is uniform throughout any single system. *Chondrites* is common in a wide variety of sedimentary rocks deposited over the past half billion years. The burrows are still being produced today, but all known modern occurrences are in deep-sea deposits. Although no living or dead animal actually has been discovered inside a *Chondrites* burrow, let alone in the process of constructing a *Chondrites*, there is some circumstantial evidence reported suggesting the burrower to be an infaunal abyssal nematode (Bromley and Ekdale 1984).

Low diversity assemblages of *Planolites*, *Trichichnus* and/or *Chondrites* are commonly associated with mudrocks representing oxygen-deficient paleoenvironments conditions in offshore lithofacies (i.e. fine-grained mudstones, laminated shales, and marls reflecting

quiet-water sedimentation are typical of offshore depositional environments). Mieras et al., (1993) reported that distal offshore environments, poorly-oxygenated facies, which include predominantly non-burrowed, well-laminated shales, were deposited under low bottom water oxygen levels (probably dysoxic to anoxic) and that they represent soft to soupy substrate conditions. These facies commonly contain epifaunal macro-invertebrate associations of low-oxygen tolerant “flat-clams” and their epibionts. This suggests that the redox boundary may have maintained a position at or near the surface water interface, possibly aided by the presence of bacterial mats on the sediment surface. Although bottom waters were dysoxic, anoxic conditions and hydrogen sulfide in the soft substrate inhibited colonization by trace-making taxa.

Episodic fluctuation of this boundary below the surface-water interface may have allowed brief burrowing of the substrate by the most low-oxygen tolerant foraminifera (e.g. *Planolites* and *Chondrites*) (Mieras et al., 1993).

Another trace fossil, *Thalassinoides* was found at locality LEF-10 (Figs. 4.3 and 8.1). This locality represents an important section of the lowermost Eagle Ford Group, the Tarrant Formation, which was deposited in a transitional environment from paralic Woodbine to classical deeper marine Eagle Ford shales.

Thalassinoides is an ichnofossil that belongs to *Cruziana* ichnofacies and is indicative of nearshore or coastal environments. *Thalassinoides* are relatively large burrows systems consisting of smooth-walled cylindrical components. Branches are Y- to T-shaped and are enlarged at points of bifurcation. These are believed to be the dwelling/feeding burrows of deposit-feeding decapods (crustacea), *Callinassa* sp. These trace fossils are associated



Figure 8.1. *Thalassinoides* at locality LEF-10.

with the *Cruziana* ichnofacies in lower shoreface to offshore environments, but are also found in low diversity, brackish-water suites (Pemberton et al., 2001).

COPROLITES

These trace fossils are so abundant in the study area that they were discussed in the previous chapter and are the subject of on going research (Friedman and Hunt, 2004 in prep).

SELACHIAN MACROFAUNA

Sharks are one of nature's great success stories. Since their first appearance about 400 million years ago, they have changed very little, merely improving their ability to hunt and feed through evolving more resilient tooth structures and more streamlined body shapes. Sharks tend to be environmental generalists. There is always a direct relationship between tooth morphology and trophic ecology. Paleoecologic data are provided for each taxonomic group when possible, as well as chronologic ranges based on published literature.

CLASS CHONDRICHTYES

ORDER LAMNIFORMES

The sharks of the order Lamniformes are all carnivorous, feeding mostly on fish which are swallowed nearly whole and are characterized by long slender (grasping) and triangular (tearing) teeth (Capetta, 1987). An exception is *Squalicorax*, whose serrated teeth are adapted to cutting large prey (Kirkland, 1990). Selachian tooth morphologies closely reflect diets and not only allow for identification but also give some indication of trophic adaptations. Thus selachian fossils have a great potential for biostratigraphy and paleoenvironmental interpretation (Williamson et al., 1993). Fossilized remains of these sharks were found at localities: LEF-1, LEF-2, LEF-3, LEF-4, LEF-5, LEF-6, LEF-7

and LEF-8 (Table 2, p. 56).

FAMILY ANACORACIDAE

Squalicorax were mainly scavenger sharks (Shimada, 1997a). They were also trophic generalists that fed on a variety of animals. Hank and Shimada (2002) reported the first shark-bitten bird bone: a tibiotarsus of a hesperornithiform. It is uncertain whether *Squalicorax* attacked or scavenged this Cretaceous marine bird. In the study area *Squalicorax* teeth were found in localities LEF-1 to LEF-8.

Squalicorax falcatus.- The recurved, serrated teeth of this distinctive Cretaceous version of the tiger shark are very common in the fossil record. Its distribution indicates that this shark ranges from brackish water lagoons into pelagic marine environments (Kirkland, 1990). The total length of *Squalicorax* did not exceed 3 m (Shimada and Cicimurri, 2001). The teeth of this taxon show a progressive increase in size from the Cenomanian to Santonian (Welton and Farish, 1993).

Squalicorax teeth are common in Cretaceous marine deposits worldwide (Shimada and Cicimurri, 2001). Chronologic range: Cenomanian - Santonian

Squalicorax curvatus.- Chronologic range: Cenomanian

Squalicorax sp.- Chronologic range: Albian

Microcorax crassus.- Chronologic range: Cenomanian

FAMILY CRETOXYRHINIDAE

Based on their large body size, tearing or prehensile dentition, inferred patterns of heterodonty, paleogeographic distribution and the depositional environments in which they are found, sharks of the family Cretoxyrhinidae clearly were the largest and most voracious of all fish predators in Cretaceous seas. The family Cretoxyrhinidae is almost

exclusively Cretaceous, ranging from late Albian through Paleocene. As their living counterparts they were coastal as well as oceanic fishes, and found worldwide in tropical, warm temperate and temperate seas (Welton and Farish, 1993). Shark teeth of this family were found in localities LEF-1 to LEF-8.

Cretolamna appendiculata.- These were 2-3 m in length pelagic sharks. The distribution of this taxon appears to indicate a distinct preference for shallow water environments. Through time the teeth of *Cretolamna appendiculata* appear to increase in size.

Chronologic Range: Albian to Maastrichtian.

Cretoxyrhina mantelli.- This taxon probably occupied the apex of the food chain in the Late Cretaceous seas (Figs. 8.2 from LEF- 8 and Fig 8.3 from LEF-2)). This top predator however, was probably scavenged frequently by anacoracid sharks (Shimada, 1997a).

Cretoxyrhina is one of the best guides to recognizing pelagic shark faunas (Kirkland, 1990; Welton and Farish, 1993; Shimada, 1997b). Meyer (1974) reported that in the Upper Cretaceous rocks of the Texas Gulf Coast, *Cretoxyrhina* is rare in nearshore deposits, but abundant in deep water. Similarly, *C. mantelli* is represented in the deep-water assemblage of Cenomanian -Turonian selachians in Arizona (Williamson et al., 1993). Complete skeletons of *Cretoxyrhina mantelli* exceed 6 m in length (Welton and Farish, 1993). Together with the large mosasaurs, *Cretoxyrhina mantelli* was undoubtedly one of the top predators in the food chain of the shallow sea. Its very large size and fearsome dentition must have allowed *C. mantelli* to feed on prey like the huge teleost fish *Xiphactinus*, sea turtles, elasmosaurs, small mosasaurs and other sharks (Siverson 1992; Shimada et al.,1992). Shimada (1997a) reported bite marks and a tooth of this shark found on mosasaur vertebrae. Shimada et al.,(2002) reported ichthyodectid fish



Figure 8.2. *Cretoxyrhina mantelli* lateral tooth in lingual view at locality LEF-8.



Figure 8.3. *Cretoxyrhina mantelli* teeth from locality LEF-2.

(*Xiphactinus*) and protostegid turtle bitten by *C. mantelli*. The selachians *Cretoxyrhina*, *Ptychodus* and *Squalicorax* probably occupied the highest positions in the food chain during the maximum invasion of the Niobrara Sea (Shimada 1996).

Chronologic range: Cenomanian -Coniacian. Its distribution was worldwide.

Creodus semiplicatus.- The teeth of this species are the largest Cenomanian fossil teeth found in Texas (Welton and Farish, 1993) (Fig. 8.4 from LEF-2). According to Kirkland (1990), teeth of this shark have been found in nearshore environments of Cenomanian age (Kirkland, 1990).

Chronologic Range: Cenomanian.

FAMILY MITSUKURINIDAE

Scapanorhincus raphidon.- Sharks of this type were abundant in shallow water environments throughout the late Cretaceous (Kirkland, 1990).

Chronologic range: Turonian to Coniacian. This species may also occur in the Cenomanian of Texas (Welton and Farish, 1993). Teeth of this shark are not very abundant. They were found at localities LEF-1 and LEF-2 only.

FAMILY ODONTASPIDIDAE

Carcharias tenuiplicatus.- Very small teeth, rarely exceeding 5 mm.

Chronologic range : Cenomanian (Welton and Farish, 1993).

Carcharias sp. A .- Teeth very common in the open marine facies of the Eagle Ford Group. These teeth are almost always associated with *C. mantelli* and *S. falcatus* (Welton and Farish, 1993). Shark teeth of this family were not very abundant either. They were found only at localities LEF-1, LEF-2 and LEF-4. This may be due to their minute size, that makes them easy to be overlooked, rather than their lack of preservation



Figure 8.4. *Cretodus semiplicatus* lateral tooth in labial view (Locality LEF-2).

at the localities under study.

ORDER HYBODONTIFORMES

FAMILY PTYCHODONTIDAE

The selachian genus *Ptychodus* are ray-like benthic sharks. There is some evidence that they were durophagous (shell-crushing) predators of the Late Cretaceous bivalves *Inoceramus* or *Mytiloides*. Kauffman (1972b) reported an inoceramid bivalve believed to have been bitten by *Ptychodus*. No reported ptychodontid specimens have shown direct evidence of diet, but many partial dentitions contain deposits of finely divided carbonate particle between the tooth bases. These masses are apparently made of crushed bivalves and often contain oral denticles of *Ptychodus*. Specimens from shale facies also demonstrate that these masses were present with the teeth prior to deposition. Some carbonate fragments belong to osteid bivalves and others to inoceramids. Some may be from cirripeds. It remains uncertain if *Ptychodus* targeted the inoceramids, or their epizoa, or both. Extirpation of *Ptychodus* from the North American Western Interior apparently antedated the decline of abundant inoceramids and epizoan bivalves (Stewart, 1988). Johnson et al., (2002a) reported that *Ptychodus* is a cosmopolitan Albian to Maastrichtian selachian genus. In North America, its principal occurrences are in strata deposited in the Cretaceous Western Interior Seaway, with some rare records along the Atlantic and Pacific margins. Williamson et al., (1993) found in Arizona that *Ptychodus* are most abundant in deep-water deposits and are rare or absent in nearshore deposits. They also found that deep-water assemblages tend to be lower in diversity and are identified by the association of *Ptychodus* sp. and *Cretoxyrhina mantelli*. This assemblage is identical to the *Ptychodus-Cretoxyrhina* association of Meyer

(1974) from the Texas Gulf Coast. The selachian assemblage found in the study area (North Central Texas) confirms these findings. In addition, many of the selachian fauna collected from the upper Cenomanian localities under study are known from other Cenomanian localities in other parts of North America as well as Europe and Africa. The pelagic sharks *Squalicorax*, *Scapanorhynchus*, *Cretodus semiplicatus*, *Cretoxyrhina mantelli* and *Cretolamna appendiculata* have a nearly worldwide distribution. Benthic genera such as *Ptychodus* have also a large geographic distribution (Williamson et al., 1993).

Ptychodus anonymus.-As with other species of *Ptychodus* this species is interpreted to have fed on inoceramid bivalves as the distribution of these sharks follows that of the inoceramids very closely. The only substantiated report of predation of an inoceramid by *Ptychodus* is that of Kauffman (1972b). The lack of documentation may be a result of the efficiency with which *Ptychodus* preys on inoceramids. Hattin (1975) suggested that the source of inoceramid prisms making the bulk of many calcarenites are inoceramids processed by *Ptychodus*. Therefore it is possible to infer that the lensoid calcarenites loaded with pyritized inoceramid prisms found at locality LEF-4 could have been at least partially processed by *Ptychodus* sharks.

Chronologic range: Cenomanian - Turonian.

Ptychodus whipplei. -Chronologic range: Cenomanian -Coniacian (Johnson et al., 2002b).

Ptychodus decurrens.- Chronologic range: Late Albian-Cenomanian.

P. anonymus was collected at localities LEF-1 to LEF-8, *P. whipplei* only at locality LEF-8, and *P. decurrens* at localities LEF-1 and LEF-8.

ORDER ORECTOLOBIFORMES

FAMILY RHINCODONTIDAE

The teeth of this species are very small, generally less than 2 mm, superficially resemble teeth of the extant whale shark *Rhincodon*. They were only found at locality LEF-1.

Chronologic range: Cenomanian - Turonian.

NON-SELACHIAN MACROFAUNA. CLASS OSTEICHTHYES

An abundant teleost ichthyofauna was found at the localities under study. Many remain undescribed and undoubtedly will be the subject of future work.

ORDER PYCNODONTIFORMES

The pebble-like teeth of these extinct fish indicate that they may have fed on shelled invertebrates as part of their durophagous diet (Kirkland, 1990). Teeth of this order were found at localities LEF-1 and LEF-5.

ORDER PACHYCORMIFORMES

FAMILY PACHYCORMIDAE

Protosphyraena sp. This is one of the largest Upper Cretaceous fish reaching 2.5 m in length. There are at least four species recognized for this genus. All presented a long shallow skull extending into a strong very elongate rostrum. They were powerful predators. The significance of the long rostrum is uncertain (Frickhinger, 1995).

Chronologic range: Cenomanian - Turonian.

The remains of this large fish were found at localities LEF-2 (1 complete rostrum, 2 additional rostrum fragments, pectoral girdle, and 2 hypural bones), LEF-1 (3 rostrum fragments) and at locality LEF-4 another complete pointy rostrum 21 cm in length was found lying at a bedding plane.

ORDER ICHTHYODECTIFORMES

FAMILY ICHTHYODECTIDAE

Ichthyodectes sp. Large predatory fishes with slender elongated bodies. Caudal fin very large, deeply forked. Remains of this large fish were found at localities LEF-1, LEF-2 and LEF-3.

Chronologic range: Lower to Upper Cretaceous (Frickhinger, 1995).

Xiphactinus audax.- This is the largest upper Cretaceous fish reaching at least 5 m in length. *Xiphactinus* was perhaps the largest bony fish that ever lived (Figs. 8.5 and Fig 8.6). This very large carnivorous fish is a typical element of pelagic faunas during the upper Cretaceous (Kirkland, 1990). Remains of this gigantic fish were found at localities LEF-1 and LEF-2. At locality LEF-3, a partial skeleton was found (at least 27 vertebrae measuring 6 cm in diameter were collected). The rest of the skeleton could not be excavated due to lack of man power and overburden of at least 3 m. The locality has now been developed for residential construction. At locality LEF-4 remains of this gigantic fish (8 vertebrae measuring 5.5 cm in diameter, skull bones and operculum) were found all scattered around due to heavy construction at the site. Salvage of the bones was done over several days of scouting around. Figs. 8.5 and 8.6 show some of the remains rescued from this locality.

Chronologic range: Cenomanian -Turonian.

FAMILY SAURODONTIDAE

Middle -sized fishes with shallow elongate heads. The snout was elongate, with the lower jaw projecting and a pointed edentulous prementary. They were formidable predators.



Figure 8.5. *Xiphactinus audax* (vertebrae, operculum and some cranial elements) LEF-4.

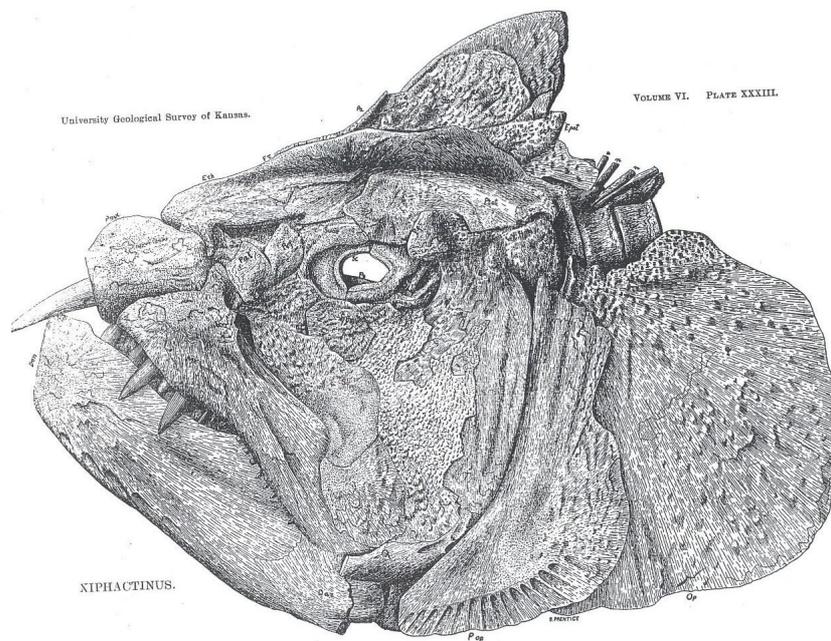


Figure 8.6. *Xiphactinus audax* (Williston, 1900).

The oldest record of saurodontids in North America was found by the writer and her daughter Joanna on June 6th 2000 at locality LEF-1 of the study area (Fig. 8.7) and was subsequently described by Friedman (2001) and Stewart and Friedman (2001) (Figs. 8.8 and Fig. 8.9). Practically all skull bones were found, jaw with teeth, 34 vertebrae and pectoral girdle. The estimated length of his fish in life was about 2 m. Isolated remains of *Saurodon* sp. were also collected at localities LEF-2, LEF-4 and LEF-5.

Chronologic range: Upper Cretaceous (Frickhinger, 1995).

ORDER CLUPEIFORMES

FAMILY PLETHODIDAE

Thryptodus sp. The plethodids are predatory fishes that possess round and rather massive snouts. Several of these rostra have been found at the study area: LEF-1 (one rostrum), LEF-2 (four rostra) and LEF- 4 (1 rostrum, Fig. 8.10). The rostrum of this last locality seems to be from a new species of plethodid fish and will be the subject of future collaborative research and comparison with Canadian specimens and with other North American specimens (Shimada and Cumbaa pers. comm. 2002).

Chronologic range: Upper Cretaceous (Frickhinger, 1995).

ORDER ELOPIIFORMES

FAMILY PACHYRHIZODONTIDAE

Pachyrhizodus sp. Middle-sized to large predators. Remains of these fishes were collected from localities LEF-1, LEF-2 and LEF-4 of the study area.

Chronologic range: Upper Cretaceous (Frickhinger, 1995).



Figure 8.7. Locality LEF-1 (Mansfield, TX) where the oldest record of saurodontid fish in North America was found on June 6th, 2000.



Figure 8.8. *Saurodon* sp. from locality LEF-1.

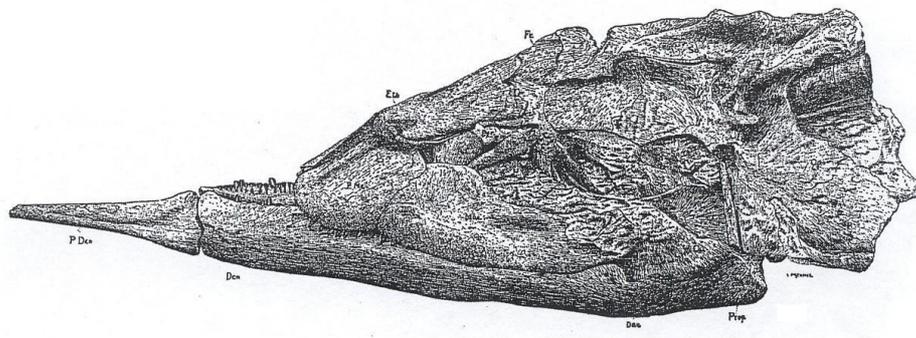


Figure 8.9. *Saurodon* sp. (Stewart, 1900).



Figure 8.10. Dorsal view of *Thryptodus* sp. rostrum from locality LEF-4.

ORDER AULOPIFORMES

FAMILY ENCHODONTIDAE

Enchodus sp. The fishes of this family were 30-90 cm in length and were very efficient moderate to deep water predators (Applegate, 1970). They possessed widely spaced conspicuously large teeth. The durability of enchodontid material in the fossil record is notorious. The variety of lithologies containing enchodontid remains, coupled with their worldwide latitudinal distribution reflects a 'warm' open sea habitat, which included the relatively shallow inland Cretaceous seas (Willimon, 1973). In the study area countless enchodontid teeth were found at localities (LEF-1 to LEF-8). Their distinctive sharp palatine teeth were found disarticulated practically in all the study area (Fig 8.11). Some of these teeth were found as major skeletal components of the hardgrounds at locality LEF-2. Although they were found most of the time disarticulated, several articulated jaws collected consist of dentary bones joined at the symphysis (LEF-1, LEF-2, LEF-3). At locality LEF-9 a partial enchodontid skeleton was collected enveloped in an ellipsoid-shaped reddish calcareous concretion.

Applegate (1970) gave the following depth ranges (fathoms) for the following taxa:

	0-50 F (shallow)	50-100 F (moderate)	100+ F (600 feet) (deep)
<i>Squalicorax</i>	X	X	X
<i>Ptychodus</i>	X	X	X
<i>Scapanorhyncus</i>	X	X	X
<i>Cretolamna</i>	?	X	X
<i>Saurodon</i>	X	X	X
<i>Enchodus</i>		X	X
<i>Protosphyraena</i>	?	X	X
<i>Cretoxyrhina</i>	?	X	X
<i>Pachyrhizodus</i>		X	X
<i>Ichthyodectes</i>	X	X	X
<i>Xiphactinus</i>	X	X	X



Figure 8.11. Partial jaw and palatine teeth of *Enchodus* sp. from locality LEF-1.

INDETERMINANT TELEOST REMAINS

Isolated, articulated and semiarticulated cranial elements, fins, vertebrae, teeth, scales of a wide range of fishes and other sharks are present at the localities under study and awaits study. The presence of this wealth of indeterminate remains indicates that a healthy fauna existed in the upper part of a stratified water column throughout the lower Eagle Ford Group (Britton Fm.).

CLASS REPTILIA

ORDER PLESIOSAURIA

Plesiosaurs were formidable marine tetrapods. Recent discoveries in northern Mexico, infer that some of them reached lengths of 15 m (Buchy et al, 2003). There are very rare freshwater remains found in Australia (Bartholomai, 1966). In this same continent, some opalized skeletons of this sauropterygian have been found (Etheridge, 1904). Plesiosaurs had large conical teeth, somewhat recurved at the tips with longitudinal striae. Their long sharp teeth indicate a fish-eating habit, and this has been confirmed by the fossilized stomach contents. The fish were evidently gulped down whole and ground up in a gizzard-like organ which contained gastroliths (stomach stones) (Welles, 1949). Plesiosaur remains are rather common in the Eagle Ford Group of North Central Texas and some of them have been rather spectacular, like the plesiosaur found during construction of the then Branniff terminal at DFW International airport (Hill, 1976) and the legendary *Elasmosaurus morgani* find of Cedar Hill, TX. (Shuler, 1950).

Several plesiosaur teeth and some disarticulated vertebrae were found in the study area at localities LEF-1, LEF-2 and LEF-4. A plesiosaur paddle bone was found at LEF-3 and a partial dentary at LEF-4.

Chronologic range: Middle Triassic- Maastrichtian. (Callaway and Nichols, 1997).

ORDER CHELONIA

Turtle remains were found at localities LEF-1, LEF-2, LEF-4 and LEF-5. However, they are too fragmented and remain unidentified. It is likely that they belong to sea turtles since no terrestrial animal remains were found at any of the localities in the study area.

ORDER SQUAMATA

The family Dolichosauridae represents the most conservative mosasauroids (Bell,1993).

These marine lizards had slender, almost snake-like bodies, small heads and paddle-shaped limbs. They were approximately 1 m total length and were probably inhabitants of inshore waters. *Coniasaurus crassidens* shares characteristics commonly found in a number of families of both marine and terrestrial lizards (Bell et al.,1982). Their remains are rare worldwide and they remain largely understudied. In the study area, surprisingly, remains were found in most of the localities under study (LEF-1, LEF-2, LEF-3, LEF-4, LEF-5 and LEF-6) and therefore these findings warrant future research as some of them may be new species (Caldwell, written comm., 2004) (Figs. 8.12 and 8.13).

Chronologic Range: Cenomanian - Turonian.

PLANTAE

Rare petrified wood fragments up to 12 cm in length were found at locality LEF-2.

Nevertheless carbonized plant remains are found occasionally in almost all localities.

They are found often on the bedding planes of the buff color mudstones.

Carbonization occurs when solution and other types of chemical action underwater transform the composition of plant and animal tissues to a thin film of carbon, the organic

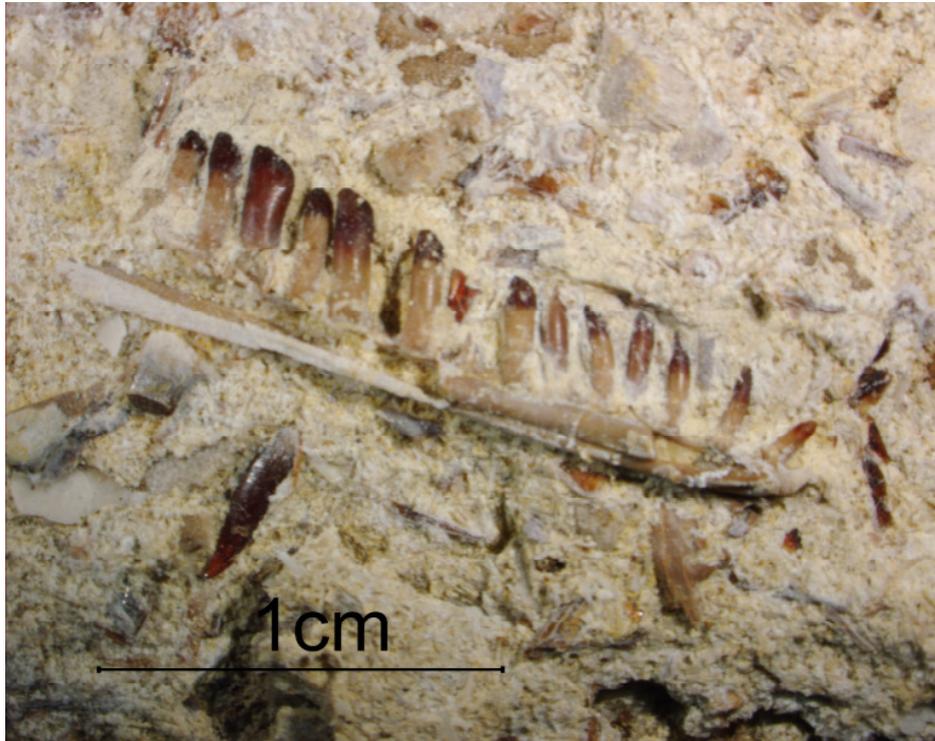


Figure 8.12. *Coniasaurus* sp. partial jaw in matrix (locality LEF-1).

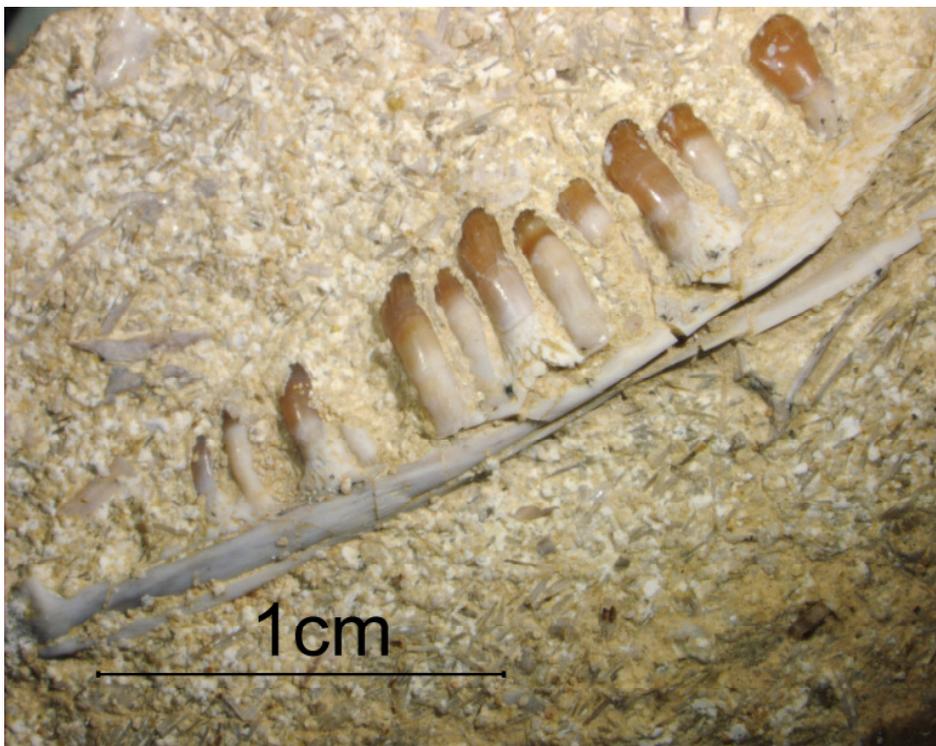


Figure 8.13. *Coniasaurus* sp. partial jaw in matrix (locality LEF-2).

remains are carbonized. During carbonization, volatile substances of the original organism- including hydrogen, nitrogen, and oxygen- are driven off leaving a thin film of carbonaceous material to show the outline of the original organism. Plants (as well as other organisms) are commonly preserved this way.

Were these plants growing in nearshore environments or were they transported by the currents? Is this a distal or proximal offshore environment? Or was this definitively a nearshore environment? Even though these questions cannot be assessed with complete certainty, one possibility is the drifting of the coastline through time because of fluctuations of the sea level due to the pulsatory nature of the Cenomanian (Wilmsen, 2003). According to Jacobs (1993), prior to the lower Eagle Ford deposition, during the uppermost Woodbine, the shoreline ran through what is now the Dallas-Fort Worth International Airport (locality LEF-7 of the study area) (Fig. 8.14).

Fossil transport is common enough to be seriously considered when making paleoenvironmental interpretation. However, whereas non-marine fossils may readily be carried into a marine environment, marine organisms are rarely transported into nonmarine environments. Thus a minor amount of exclusively nonmarine fossils in a marine assemblage does not invalidate a fully marine interpretation of the environment (Heckel, 1972).

THE FOOD WEB IN THE TIMES OF THE LOWER EAGLE FORD SEA

The lower Eagle Ford contains a rich ichthyofauna fossil consisting of a mixed assemblage of undoubted pelagic open sea forms along with some possible nearshore and perhaps shallow water fishes, but showing a dominance of deep-water fishes. All the evidence suggests oxygen-poor bottom conditions that resulted in a very restricted bottom fauna



Figure 8.14. Locality LEF-7 (Dallas-Forth Worth International Airport).

consisting only of inoceramids and their rare epizoan oysters. Certainly as far as the present evidence is considered, the lower Eagle Ford fauna lived in shallow to moderate depth water (30-70 m). At the same time, the carbonized plant impressions and rare petrified wood fragments that have been found may suggest a nearby shore. The possibility of a migrating paleoshore line through time due to fluctuations in sea level can not be discarded and therefore a proximal offshore environment for the lower Eagle Ford is a more likely setting.

If one is willing to assume that at least most of the common genera of vertebrates and invertebrates of the lower Eagle Ford have been preserved in the fossil record, then it is possible to make some inferences concerning the possible paleoecological relationships among these forms as depicted in Fig. 8.15. These trophic levels are based primarily on the relative abundance of finds of isolated or associated teeth. The hard dental components possess a high fossilization potential and large numbers of them are produced by any individual during its lifespan. Most importantly, due to their ease of identification and the direct relationship between tooth morphology and trophic ecology.

The following TROPHIC LEVELS are proposed for the paleoecological environment of the lower Eagle Ford Group in North Central Texas. (Fig. 8.15)

LEVEL 1. The vertebrate fauna represents the top of the food chain and in some cases interacts directly with the invertebrate fauna. At the top of the pyramid were the largest sharks *Cretoxyrhina*, *Cretodus* and *Cretolamna* which were formidable predators. Based on their fossil record members of the family Cretoxyrhinidae they were very common and swam rampant in practically all the study area (LEF-1 to LEF-8).

LEVEL 2. Marine reptiles: Plesiosaurs. These were formidable predators preying on

smaller and larger fish. Their scattered teeth, vertebrae and rare paddle bones were found at localities LEF-1, LEF-2, LEF-3, LEF-4).

LEVEL 3. Largest carnivorous fishes : *Xiphactinus*, *Pachyrhizodus*, *Saurodon*, *Protosphyraena*.

These fish have also a rich fossil record in localities LEF-1, LEF-2, LEF-3, LEF-4, LEF-5, LEF-6, LEF-7, LEF-8.

LEVEL 4. Middle -sized predatory fishes: abundant Enchodontids found in great abundance in practically all localities (LEF-1 to LEF-9).

LEVEL 5. Scavenger sharks: *Squalicorax*. Teeth of these sharks were found readily at LEF-1 to LEF-8.

LEVEL 6. Ray-like sharks *Ptychodus*. Present at localities LEF-1 to LEF-8. Based on their dentition morphological features they had a durophagous diet feeding mostly on invertebrates: *Inoceramus* ?

LEVEL 7. Small marine lizards (*Coniasaurus*) fed on seaweeds and small fishes? Their fragile remains are present at localities LEF-1 to LEF-6.

LEVEL 8. Invertebrate macrofauna: Abundant filter feeder bivalves (inoceramids and their epizoans oysters (*Pseudoperma congesta*). Inoceramids were present in most of the study area (localities LEF-1 to LEF-8). Although they were filter feeders organisms, they may likely have depended on a relationship with photosynthetic or chemotrophic symbionts.

LEVEL 9. Zooplankton (very abundant globular planktic foraminifers). Fossilized tests collected from LEF-1 to LEF-8.

LEVEL 10. Phytoplankton (represented in the sediment (LEF-1, LEF-4) by calcareous

nannoplankton, i.e. coccolithophores and other algae). This is the base of the food web then as it is today.

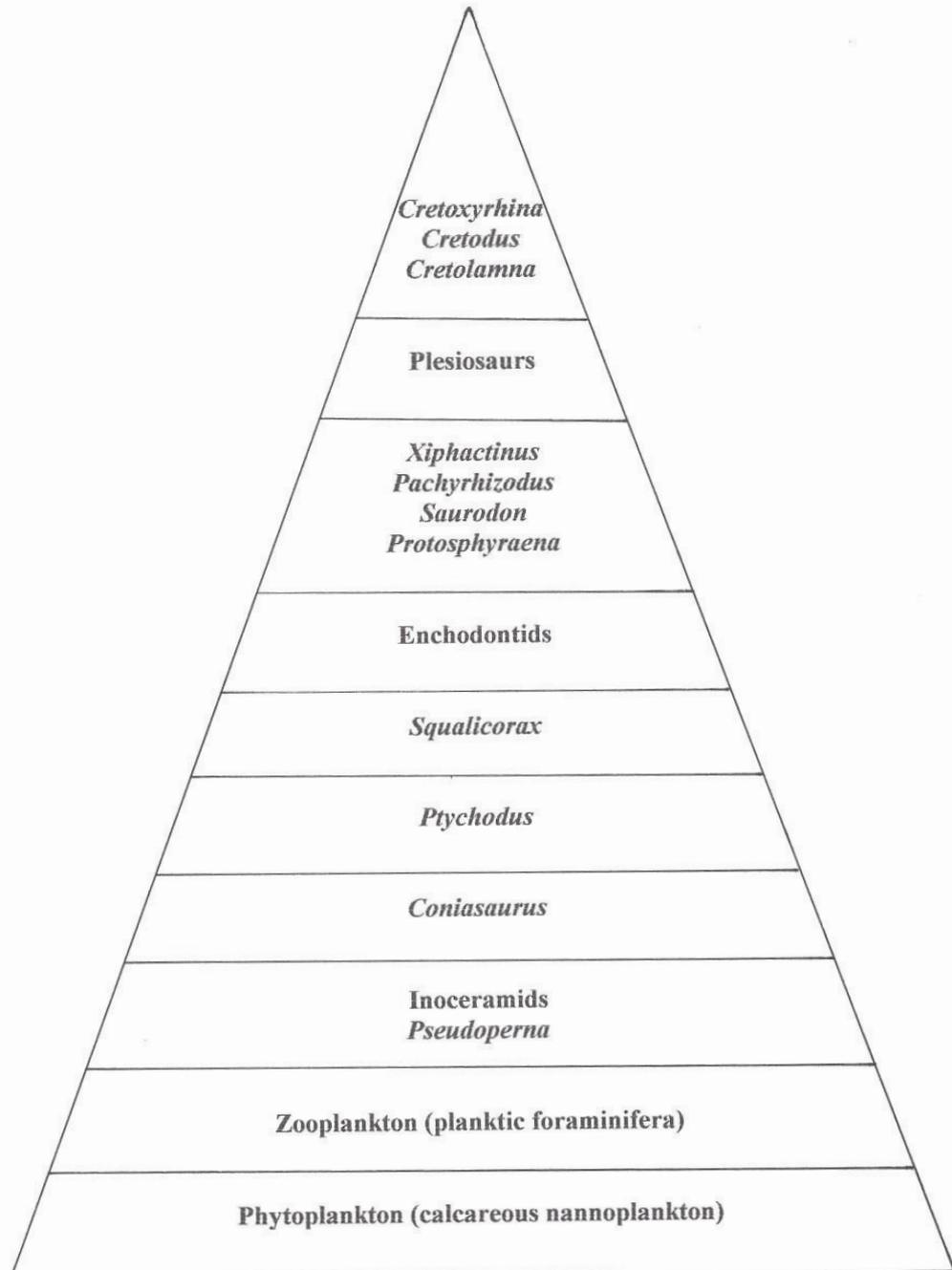


Figure 8.15 Food Web during lower Eagle Ford time in North Central Texas.