"In grad school there are no days, only deadlines."

"If you keep doing things like you've always done them,
what you'll get is what you've already got."
- Anon -

"Sacred cows make the best hamburger."
- Abbie Hoffman -

"Tell me, I'll forget. Show me, I may remember.
But involve me and I'll understand."
- Chinese Proverb -

"Education makes people easy to lead, but difficult to drive,
easy to govern, but impossible to enslave."
- Henry Peter Broughan -

"Never let your schooling interfere with your education."
- Mark Twain -

"The whole problem with the world is that fools and fanatics are always
so certain of themselves, and wiser people are so full of doubts."
- Bertrand Russell -

"The most exciting phrase to hear in science, the one that heralds new discoveries, is not,
"Eureka!" ("I found it!") but rather, "Hmm... that's funny..."
- Isaac Asimov -

"The great tragedy of science, the slaying of a beautiful hypothesis by an ugly fact."
- Thomas Henry Huxley -

"Was she told when she was young that pain would lead to pleasure,
Did she understand it when they said,
That a man must break his back to earn his day of leisure?
Will she still believe it when he's dead."
- The Beatles -

"Life is what happens to you while you're busy making other plans."
- John Lennon -

University of Alberta

Sedimentology and Ichnology of the Lower Cretaceous Kamik Formation in the Parsons Lake Gas Field, Mackenzie Delta, Northwest Territories

by

Michael David Hearn ©

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of Master of Science

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Your file Votre référence
ISBN:
Our file Notre référence
ISBN:

This thesis is dedicated to my grandparents,
Amy and Cecil Hearn and Karol and Josephine Ilnicki
If it weren't for the opportunities you provided myself and our family
I would not be where I am today.

Abstract

The Lower Cretaceous Kamik Formation of the Beaufort-Mackenzie Delta region is a sandstone-dominated succession interpreted to represent a moderately storm-influenced deltaic succession. Detailed ichnological and sedimentological analysis indicates that the Kamik Formation was deposited as a series of prograding delta lobes characterized by varying degrees of wave, storm and fluvial influence. Overall, wave- and storm-energy were the dominant processes influencing deposition, however some successions also display evidence of a strong riverine influence. Detailed sedimentological and ichnological analysis of fifteen cores identified fifteen facies (Facies A to O) which were subsequently grouped into five facies associations (FA1 to FA5) interpreted to represent a wave-dominated delta, meandering distributary channel system, braided distributary channel system, lower delta plain and a river-dominated delta. The Rhône delta is interpreted to represent a modern analogue for the majority of the wave-dominated Kamik deltas (FA1) and the Mississippi delta is interpreted to represent a modern analogue for the river-dominated deltas of FA5.

Acknowledgements

Well it's kind of hard to know where to start with my acknowledgements since pretty much everyone I know, including my family and friends, has had some type involvement with my thesis during the past 4 years! This thesis has been a huge part of my life and I can honestly say that I'm looking forward to having it done so I won't "have to go home to work on my thesis" anymore. The majority of this thesis was written between the hours of 10 pm to 7 am because I'm naturally a night owl and enjoy staying up late. But during the process of finishing this thesis, I truly developed into a night person. If I could only count up the number of hours of sleep I lost pulling all-nighters doing thesis work, it was definitely all worth it. Even as I type these acknowledgements I am sitting in the Ichnolab at 5:30 am after working through what I hope will be the last all-night thesis work session of my grad school career.

First and foremost, I would like to thank my supervisor, Dr. George Pemberton for his excellent supervision, guidance, endless support and patience throughout all of my years in graduate school. Throughout all the stages of my Master's thesis George has been an incredible mentor, a constant source of encouragement, scientific knowledge and an incredibly patient person when it came to my thesis progress, especially when it came to finally handing it in. George is an incredible teacher and has taught me how to be a true scientist and provided me with the opportunities to become fully encompassed in geological research. The atmosphere he provided with all his other graduate students in the Ichnolab and his amazing collection of books in his office gave me the scientific resources I needed to pursue my research interests to the furthest limits. In addition, his financial support throughout grad school provided me with many opportunities to continue my studies, meet and work with interesting people, see amazing parts of the world and to present my research at various geological conferences. George's skills and knowledge in ichnology and sedimentology are second to none, and in my opinion he truly represents the pinnacle of geological knowledge. When it comes to learning ichnology and sedimentology there's no one else I'd rather have had as a supervisor. The most memorable experience I have of grad school is during the summer of 2002 when I was logging my thesis core with George at the Geological Survey of Canada (GSC) core lab in Calgary. It was during this time that he taught me most of what I know about ichnology and sedimentology. Talking to George about ichnology and sedimentology is one thing, but when you have him on core it's amazing! I've never learned so much as when I had him helping me figure out my cores, it's definitely an experience I'll never forget. Some of the best advice he gave me was that the important stuff was in the details, he taught me that it wasn't about how many cores I could log, but the amount of information you could squeeze out of each one.

My first experience with George was in 3rd year stratigraphy, where as an undergraduate, I was introduced to his vast knowledge of Geology, and it was my first exposure to the world of ichnology. I really got to know George when I worked with him on my geology undergraduate

thesis on the Basal Belly River Formation during my 4th year. It was here that I learned how to log core, how to pronounce all the latinized trace fossil names and began spending long nights working on geology papers. The fruit of labour from my undergraduate thesis was an award for the "Best Geology Undergraduate Thesis", an award I am still highly honored to have received to this day because it got me my name engraved on a plaque in the halls of the Earth Sciences building. Finally I could show off my work to all my family and friends and prove that I wasn't wasting my time in University. If it weren't for George taking me on for an undergrad thesis back in 1999 I would've never have received this award. My years spent learning geology from George during undergrad and especially graduate school has made me the geologist I am today. George is incredibly wise both in life and in geology, the inspiration, guidance and advice he provided me throughout the years has got me where I am today. Without being a student of George's I don't think I could've ever called myself a geologist, and after all these years in grad school I now feel it is a title I rightly deserve. Thanks George, I know there's no way I will ever be able to pay you back for all the knowledge and wisdom you have imparted to me over the years and I'm definitely looking forward to collaborating with you in the future.

A very special thanks to Dr. Douglas Gardner for believing in me as a geologist when I was first starting out and most importantly for providing me with an amazing Master's project on the Kamik Formation. Doug was responsible for getting me started with a thesis project working on Lower Cretaceous rocks all the way up in the Mackenzie Delta. I first met Doug when I was a summer student at Petro-Canada back in 1999, when he was the Chief Geologist during my first summer there. Being my first job in the industry it was quite a learning curve when it came to being able to apply everything I had learned in school, but Doug offered his support and guidance and I ended up returning for two more summers at Petro-Canada. Doug's unwavering support for young geologists starting out is unparalleled by anyone I have ever met. He is an individual who truly appreciates the people he works with and has a keen eye for recognizing passionate geologists. Thank you Doug, you have no idea how much your help with getting me a thesis project changed my life.

There's a bit of a story behind how I got my Master's project working in the Mackenzie Delta region. It all started out my last summer at Petro-Canada in 2001, Doug had moved over from Petro-Canada to Gulf Canada (which soon became Conoco Canada) as the Vice President of Exploration. He was the man in charge of all the final decisions at Conoco in regards to the drilling of wells and the new projects. In the last couple of weeks of August 2001, I was wrapping up my project in the Foothills group at Petro-Canada and was desperately searching for a project to head back to grad school with. With only a few days left working downtown for the summer I gave him a call to meet up for lunch. In light of his busy work and vacation schedule we met up for a last minute meeting after work at the 11th Ave. Brewster's for hot wings and beer. During this conversation we discussed my future plans to start grad school in the following weeks with George and how I was looking for a really interesting Master's project to work on that could incorporate

ichnology. He mentioned a project that he had looking at core and interpreting the depositional environments from the Parsons Lake gas field, a discovery Gulf had made back in the early 1970's. It turned out that Conoco was re-evaluating this property and needed someone to look at all the core and figure out all the depositional environments. This was exactly the type of Master's project I was looking for, as it was a frontier area that had very little material published on the geology of the region and the type of study where I could be at the forefront of research in the area. Over the next couple of months Doug got me lined up with funding and the contacts at Conoco I would be working with over the next few years. Thanks Doug, you have no idea how much that short meeting after work opened up so many opportunities for me. Your assistance with getting me a thesis project in the Mackenzie Delta provided me with the opportunities to become the geologist I am today. I definitely look forward to working with you again in the future.

This project would not have been possible without the logistical and financial support from ConocoPhillips Canada. It was with their continued support that I was able to complete this project. I am very grateful for all the help, support and enthusiasm from Dr. Gary Prost, Dave Bywater and Guy Peasely throughout my thesis. Gary, Dave and Guy were the geologists and geophysicist at Conoco working on the Parsons Lake field. If it weren't for all the scientific interest, logistical and financial support you all provided me with over the years I wouldn't have been able to undertake and complete my thesis work. I really enjoyed working with all of you as a summer student and throughout my time in grad school. A special thank you goes to Gary Prost, for his support, enthusiasm, experience and incredible patience over the years, especially when it came to initially setting up this project and organizing the financial and logistical parts. I am especially grateful for Gary's assistance with the editing of my chapters and for always being available for discussions. No matter how long it took for me to finish he was always very supportive of my work. I would like to thank Dr. John Suter from ConocoPhillips for his help with the sedimentology and for the many discussions we had about deltas, especially when it came to identifying modern analogues for the Kamik. A special thanks also goes out to the great people I work with in the Foothills Team at ConocoPhillips for all their support over the last year while I was finishing writing. Also I would also like to acknowledge the financial support from the American Association of Petroleum Geologists (AAPG) Grant-In-Aid, Canadian Society of Petroleum Geologists (CSPG), the Province of Alberta Graduate Scholarship, Alberta Learning Graduate Student Scholarship and the Canadian Association of Petroleum Producers (CAPPS) Graduate Scholarship in Geology.

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Thank you to all the past and current members of the Ichnology Research Group (IRG) at the University of Alberta I had the pleasure to work with during my years at the U of A, you made the last few years a really enjoyable experience. I especially would like to thank Curtis Lettley, Dwayne Giggs, Michelle Spila, Scott Reid, Tom Saunders, Rozalia Pak, Rich McCrea, Jeff Reinprecht, Sean Miller, Kerrie Bann, Demian Robbins and Lynn Reich for all their help over the years. I really enjoyed being part of the IRG during all of our times at the Power Plant, the trips to AAPG and

CSPG conferences, trips to Salt Lake City, Grande Cache and Drumheller and all the other lab events. It was great being part of such a great group of people throughout the years. A special thanks goes out to Susan Fleming for organizing everything over the years, setting up my thesis defense and getting all the committee members together. I would like to thank my thesis defense committee members Dr. Charlie Stelck, Dr. Murray Gingras and Dr. Douglas Craig for making my thesis defense one the most memorable and enjoyable experiences of my grad school career. It was truly a once in a lifetime honor to have Dr. Stelck on my committee. I am in awe of his geological knowledge, experience and his unwavering passion for geology. I think I'm one of the only grad students who actually brought a fossil to my own thesis defense for him to identify!

Thanks to Dr. Jim Dixon from the Geological Survey of Canada (GSC) for being a source of knowledge and enthusiasm for working on the Kamik Formation. Dr. Dixon's knowledge of the geology of Beaufort-Mackenzie Delta region is unsurpassed. He was always very friendly and more than welcoming whenever I had any questions about the core or just wanted to stop by his office and discuss his work on the Kamik Formation. To this day he's still the man to talk to if I ever have questions about the Kamik.

Finally, I would like to give a very special thanks to my family, my Mom and Dad, Marlene and David Hearn and my brother Caley for their patience and encouragement over the last 4 years. My parents and my brother were always an unwavering source of love, support and encouragement throughout my years in University. I couldn't have made it this far without all of your help. I know it seemed at times that I would never finish, but here it is finally! Mom and Dad you can relax, I am finally done!

Table of Contents

Chapter One: Introduction

Chapter Two: Facies Descriptions and Interpretations

List of Tables

Chapter Two: Facies Descriptions and Interpretations

Chapter Three: Facies Associations, Depositional Environments and Modern Analogues

List of Figures

Chapter One: Introduction

Chapter Two: Facies Descriptions and Interpretations

Symbols and Abbreviations Legend for Core Logs

Lithology Symbols

- Fine-grained sandstone	- Siltstone	- Coal
- Medium-grained sandstone	- Muddy/shaley sandstone	- Massive coaly mudstone
- Coarse-grained sandstone	- Organic/Coaly mudstone	- Pyritized mudstone
- Silty sandstone	- Waxy to coaly mudstone	- Missing cored interval

Physical Sedimentary Structure Symbols

- Hummocky cross-stratification (HCS)	- Wavy bedding	- Mudstone rip-up clasts
- Swaley cross-stratification (SCS)	- Flaser bedding	- Pebbles/Granules
- Trough cross-bedding (TCS)	- Lenticular bedding	- Micro-fracture
- Low angle cross-bedding	- Large scale wave ripples	- Wood fragments
- Convolute bedding	- Ball & Pillow structures	- Synaeresis cracks
- Wave ripple lamination	- Stylolite	- Fracture/Slickenside
- Current ripple lamination	- Climbing ripples	Gc - Gutter Cast

Lithologic Accessories & Contact Symbols

Cal - Calcite	Rtm - Graded Rhythmites	- Sharp contact
Chrt - Chert Pebbles	Sid - Siderite	- Gradational contact
Mc - Micaceous material	Tidal - Tidal indicators	- Scoured contact
Palsl - Paleosol	Wd - Wood fragment	- Undulatory contact
Py - Pyrite	- Plant fragments	- Conglomeratic Lag
Qtz - Quartz Pebbles	- Shell Fragments	- Pebble Lag

Ichnofossil Symbols

Ar - *Arenicolites*	Glossi - *Glossifungites*	Ro - *Rosselia*	***Burrow Abundance***
As - *Asterosoma*	He - *Helminthopsis*	Sch - *Schaubcylindrichnus*	- Abundant
Bg - *Bergaueria*	Lo - *Lockeia*	Sc - *Scolicia*	- Moderate
Ch - *Chondrites*	Ma - *Macaronichnus*	Sk - *Skolithos*	- Common
Co - *Conichnus*	Oph - *Ophiomorpha*	Sp - *Siphonichnus*	- Rare
Cryptic - cryptic bioturbation	Pa - *Palaeophycus*	Te - *Teichichnus*	- Absent
Cy - *Cylindrichnus*	Ps - *Phycosiphon*	Th - *Thalassinoides*	
Dp - *Diplocraterion*	Pl - *Planolites*	Zo - *Zoophycos*	
Fg/Esc - Fugichnia (escape trace)	Rhz - *Rhizocorallium*	- Rootlets	

Chapter One: Introduction

1.1 Introduction

This study was undertaken to investigate the sedimentology, ichnology and depositional history of the Lower Cretaceous (late Valanginian to middle Hauterivian) Kamik Formation in the southeastern Mackenzie Delta region of the Northwest Territories. The Kamik Formation is a sandstone-dominated interval and is the uppermost unit in a collection of three formations, which together comprise the Lower Cretaceous Parsons Group.

Previous studies of the Kamik Formation focused on the stratigraphy, sedimentology, petrology and diagenetic history in both the subsurface of the Mackenzie Delta region and outcrop in the Richardson Mountains. The major portion of this research was completed over twenty years ago and since these initial studies, many significant developments have occurred within the field of sedimentary geology and ichnology. Recent developments within the field of sedimentology include improvements in the field of process sedimentology, development of refined facies and paleodepositional models, and the study of modern depositional systems. Within the field of ichnology, recent developments include the recognition of the environmental significance of trace fossils, applications for facies identification, paleoenvironmental reconstructions, applications in stratigraphy and the development of ichnofacies models. The aim of this study was to utilize these developments in sedimentary geology and ichnology to provide additional tools to improve the geological understanding of the Kamik Formation. This study focused on investigating the sedimentology, ichnology and depositional history of the Lower Cretaceous Kamik Formation in the Parsons Lake gas field to create a more detailed and refined paleodepositional and paleogeographic model.

1.2 Purpose of Study

This study was undertaken to investigate the sedimentological and ichnological characteristics to identify the lithofacies and ichnofacies present within the Kamik Formation in the Parsons Lake field. These interpreted facies were utilized to develop an integrated paleoenvironmental model in order to better to understand the paleodepositional environments and depositional processes responsible for the Kamik Formation. The paleoenvironmental model was utilized to identify the relationship between the interpreted paleodepositional environments and the preserved strata of the Kamik Formation present in the subsurface of the Beaufort-Mackenzie Delta region.

The overall goals for this study of the Kamik Formation were fourfold and included:

1. Examination and documentation of the sedimentological and ichnological characteristics.

2. Identification and interpretation of the lithofacies and ichnofacies.
3. Development of a paleodepositional model for the Kamik Formation.
4. Identification of modern depositional analogues for the Kamik Formation.

This thesis is presented in four chapters. The introductory chapter outlines the purpose of study, research methods, the study area, previous work, stratigraphic nomenclature, regional geology and stratigraphy, biostratigraphy, paleogeography, tectonic elements, structural setting and the geology of the Parsons Lake gas field. Chapter Two includes the facies descriptions and interpretations. Chapter Three includes the facies associations, a detailed discussion of deltaic environments, identifies modern analogues for the Kamik Formation and discusses the evolution of deltas through time and the influence of storms. The ideas presented in this thesis are summarized in Chapter Four. A study of the integrated sedimentological and ichnological aspects of the Kamik Formation will result in a more complete understanding of the paleoenvironmental conditions responsible for deposition of the Kamik Formation. The goal of this thesis is to enhance the geological understanding of the Kamik Formation to assist with future academic and industry related projects in the Beaufort-Mackenzie Delta region.

1.3 Research Methods

The Kamik Formation was analyzed in the subsurface of the Mackenzie Delta region by the examination of core material, well log and biostratigraphic data. Approximately 200 m of core from the Kamik Formation were logged in detail for sedimentary structures and ichnological characteristics. In addition to the Kamik core descriptions included in this thesis, cored intervals from the Martin Creek and McGuire formations were analyzed. Core material was examined at the Ichnology Research Group laboratory at the University of Alberta during January to April 2002 and at the Institute for Sedimentary and Petroleum Geology (ISPG) located at the Geological Survey of Canada in Calgary from May to August 2002. Sedimentological analysis of the Kamik strata concentrated on identifying lithology, grain size (measured with a CanStrat grain size chart), bed thickness, nature of bedding contacts, primary physical sedimentary structures and penecontemporaneous deformation structures. Ichnological observations included the identification of ichnogenera, their relative abundance, assemblage diversity, and the intensity of bioturbation. Each core was subsequently subdivided into lithofacies and ichnofacies on the basis of lithology, physical sedimentary structures, grain size, textures and ichnological characteristics. The ichnology and sedimentology were ultimately integrated in order to characterize the depositional lithofacies and ichnofacies present in the cores. The recognition of facies provided for the interpretation of the depositional environments responsible for the Kamik Formation.

All of the information gathered during core logging, including descriptions, hand drawn strip logs and photographs were organized and compiled. The hand-drawn lithologs were drafted into

AppleCORE© and then redrafted into Adobe Illustrator© software in an appropriate template for ease of presentation. The lithologies, sedimentary structures, ichnological symbols, facies, facies associations and interpreted paleodepositional environments were compiled and presented in separate columns. Facies were identified based on the observed lithologies, sedimentary and ichnological features. The core photographs provided a reference to verify the validity of the core descriptions. Once each core was subdivided into its respective facies, the paleodepositional environments were interpreted for each core and compared with the cores from other well locations.

The frontier nature of this study limited the amount of core material available for research compared to that available from more developed sedimentary basins (i.e. the Western Canadian Sedimentary Basin). Due to the limited amount of core material in the study area and the limited scope of the study, the author chose to focus on acquiring as much detailed information as possible from the limited data set. This required restricting the study area to the Parsons Lake gas field with the goal of presenting as high a resolution study as possible. There does exist additional core data from the Kamik Formation outside of the study area, but due to time restrictions it was not incorporated in this study. Studies by Myhr and Young (1975) and Dixon (1982a, 1982b, 1991) provided excellent sources of supplemental data and interpretations for the regional geology outside of the study area. The high-resolution nature of this study within the Parsons Lake area contrasts with the majority of core-based studies, which utilize appreciably larger study areas with lower scales of resolution due to the availability of larger data sets.

1.4 Study Area

The study area is located within the subsurface of the onshore, southeastern portion of the Beaufort-Mackenzie Delta region, Northwest Territories. Figure 1.1 illustrates the study area encompassing the onshore Parsons Lake gas field, located in the southwestern corner of the Tuktoyaktuk Peninsula. The detailed study area illustrated in Figure 1.2 encompasses the Parsons Lake gas field and includes a geographic region of approximately 840 km^2. There are 18 wells that penetrate the Kamik Formation within the Parsons Lake gas field (Fig. 1.2). Of this total, only 7 wells contain core (Table 2.1) with the majority of wells containing multiple cores from the Kamik Formation. The subsurface was the primary database for this study, and the reader is referred to Dixon (1991) for detailed descriptions and discussions of the sedimentology and stratigraphy of the Kamik Formation and the associated Parsons Group in outcrop from the northern Richardson Mountains and Northwest Territories.

1.5 Summary of Previous Work

The geology of the northern Yukon and Northwest Territories was first reported from outcrop by

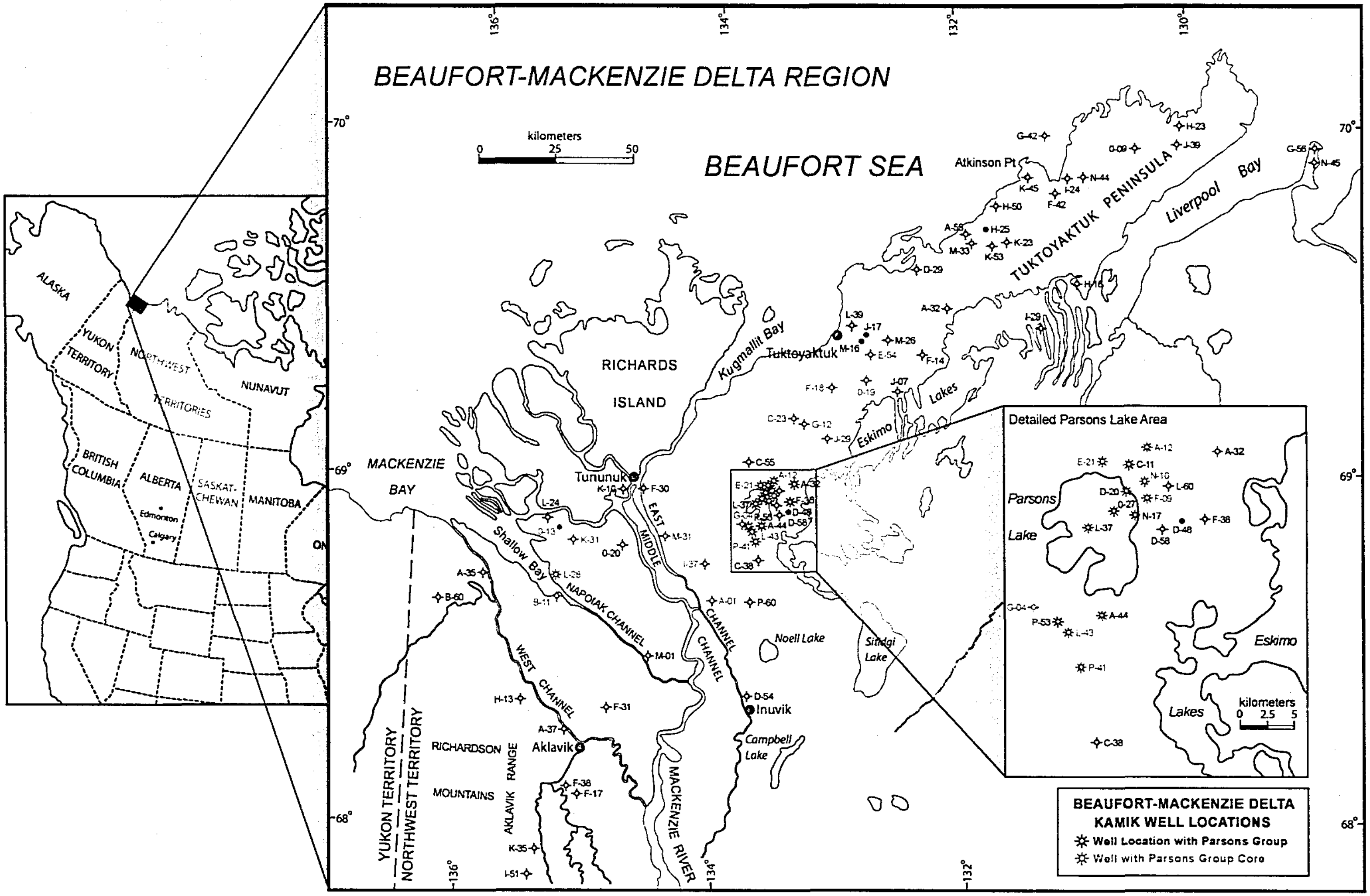


Figure 1.1 Location map of the study area within Beaufort-Mackenzie Delta region of the Northwest Territories. The map features the location of exploratory wells penetrating Lower Cretaceous strata in the region. The wells with core from the Kamik Formation are indicated in red (modified from Dixon, 1982b).

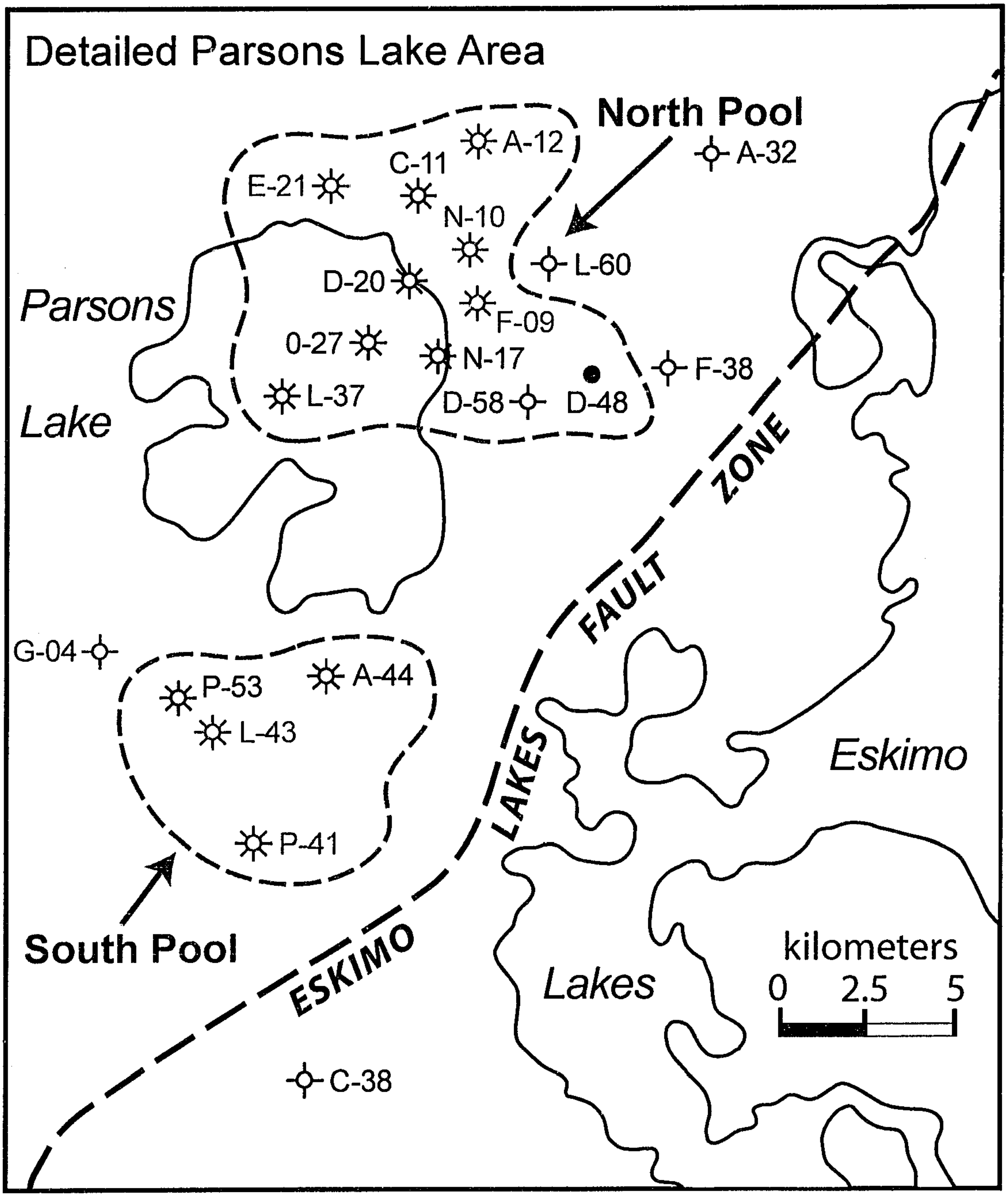


Figure 1.2 Location map of the study area in the Parsons Lake field located in the Beaufort-Mackenzie Delta region. The wells with core from the Kamik Formation are indicated in red. Note the geographic locations of the North Parsons and South Parsons gas fields and the approximate location of the Eskimo Lakes Fault Zone (modified from Dixon, 1982a).

R.G. McConnell (1891) during his traverse through the Aklavik Range. The outcrop region was subsequently briefly explored and traversed by numerous geologists employed by the Geological Survey of Canada including Camsell (1906), O'Neill (1924), Nauss (1944), Richards (1950) and Gabrielse (1957). The majority of the findings from this early exploration are summarized in papers by Hume and Link (1945) and Hume (1954). These geologists established the presence of Late Jurassic to Early Cretaceous age rocks on Mount Gifford and on the banks of the Donna River, and Early Cretaceous age rocks on Mount Goodenough (Jeletzky, 1958). These early studies did not include any detailed stratigraphic interpretations or precise age dating of these units. The first major geological publications of the regional geology, stratigraphy and depositional history of Mesozoic rocks from the northern Richardson Mountains were completed by Goodman (1954) and Martin (1959, 1961).

The first detailed investigations of Lower Cretaceous strata now attributed to the Kamik Formation and Parsons Group were completed by J.A. Jeletzky (1958, 1960, 1961a, 1961b, 1967, 1974, 1975, 1980) of the Geological Survey of Canada from the outcrop belt in the northern Richardson Mountains and the Aklavik Range. Jeletzky studied the stratigraphy, sedimentology and paleontology of Lower Cretaceous strata in the northern Richardson Mountains, where he informally subdivided the strata into formations. Jeletzky's work in the 1960's was part of the Geological Survey of Canada's large-scale geological mapping project termed Operation Porcupine, which was led by D.K. Norris. Jeletzky was mainly responsible for the stratigraphy and sedimentology of the Lower Cretaceous strata in the northern Richardson Mountains and Aklavik Ranges. Significant contributions to the knowledge of the surface and subsurface geology were also added by Young (1971, 1972, 1973a, 1973b, 1974). Since Jeletzky and Young's initial studies in the Richardson Mountains most of the subsequent work on the Parsons Group focused on its occurrence in the subsurface in the Beaufort-Mackenzie Delta region following the discovery of large natural gas reserves at Parsons Lake in 1972. Studies by Young (1972, 1973a, 1973b, 1974), however, did continue to focus on the Lower Cretaceous strata in the northern Richardson Mountains. The northern Yukon and northwestern Northwest Territories were mapped on a 1: 250,000 scale by Norris (1975, 1977, 1979). The regional stratigraphy and paleogeography of the Lower Cretaceous strata in the subsurface was reviewed by Lerand (1973) and Young (1973b) in their summaries of the geology of the Beaufort-Mackenzie and northern Yukon areas.

The first major geological studies of Lower Cretaceous strata in the subsurface were completed by Myhr and Gunther (1974), Coté *et al.* (1975), Myhr and Young (1975) and Brideaux and Myhr (1976). The tectonic history and structural geology of region was summarized in papers by Norris (1972, 1973, 1974), Yorath (1973) and by Yorath and Norris (1975). Encouraged by the major discovery of the Parsons Lake gas field in 1972, Coté *et al.* (1975) presented one of the first subsurface studies of the Lower Cretaceous stratigraphy and sedimentology and introduced the term "Parsons Sandstone". In their paper Coté *et al.* (1975) summarized the geology of the

Parsons Lake gas field and compared the subsurface stratigraphy with that done by Jeletzky on the eastern flank of the northern Richardson Mountains (Dixon, 1991). Further studies by Myhr and Young (1975) focused on the regional subsurface sedimentology of the Parsons Group strata in the Mackenzie Delta-Tuktoyaktuk Peninsula areas. Myhr (1974), Myhr and Gunther (1974), and Brideaux and Myhr (1976) provided additional subsurface data including biostratigraphy and the interpretation of depositional environments for cores from specific wells.

Young *et al.* (1976) presented one of the first syntheses of the regional geology of the surface and subsurface data from the Parsons Group of the Mackenzie Delta region and northern Yukon. Palynological studies of the Parsons strata have been published by Pocock (1976), Brideaux and Fisher (1976), McIntyre and Brideaux (1980) and Fowler (1986). Pocock (1976) introduced some informal formation names for the Lower Cretaceous strata in the northern Richardson Mountains, including Parsons Group equivalent rocks. The introduction of the informal terminology by Pocock was refuted by Brideaux *et al.* (1977) due to lack of formal definition within Pocock's paper. Pocock (1977) responded by accepting some of Brideaux *et al.*'s (1977) criticisms indicating he had not intended the names to be formal. Pocock (1976) also introduced the name Martin Creek for Parsons-equivalent strata, although it was not the same terminology subsequently used by Dixon, (1982a, 1991). A field guide of the Mackenzie Delta and northern Richardson Mountains was completed by Young (1978), which included descriptions of some Parsons-equivalent outcrops. A summary paper by Balkwill *et al.* (1983) referred to the Parsons Group strata with respect to regional geologic setting of the entire Arctic region.

Dixon (1982a) presented a detailed regional study of the subsurface stratigraphy and sedimentology of the Jurassic and Lower Cretaceous of the Mackenzie Delta and Tuktoyaktuk Peninsula. In the same paper Dixon (1982a) formally defined the name Parsons Group and introduced the Kamik Formation as a new formation of the group. Sykora (1984) focused on the investigation of the sedimentary petrology, diagenetic history, stratigraphy and sedimentology of the Parsons Group. This study also included a basic facies analysis based on sedimentological and ichnological analysis and provided paleoenvironmental interpretations. The regional geology of the Parsons Group and Kamik Formation were further summarized by Dixon (1991) from detailed studies of the stratigraphy and sedimentology from the outcrop belt in the northern Richardson Mountains. The formal terminology for the Parsons Group was finalized in a paper by Dixon and Jeletzky (1991) in which they introduced the Martin Creek and McGuire formations along with the previously defined Kamik Formation of Dixon (1982a) as comprising the Parsons Group. Subsequent work by Dixon *et al.* (1985) presented brief descriptions of the Parsons Group and Dixon (1986) reviewed the regional character of Parsons strata in a paper on the Cretaceous-Tertiary geology of the northern Yukon and adjacent Northwest Territories. The paleogeography of the Mackenzie Delta region and northern Yukon during the deposition of the Parsons Group was illustrated by Dixon (1987). A discussion of the Parsons Group and Kamik Formation was presented by Dixon (1992) in a review of Cretaceous and Tertiary stratigraphy in the northern Yukon and adjacent Northwest Territories. The hydrocarbon potential of the Parsons Group, including the Kamik Formation, has been discussed by Coté *et al.* (1975), Langhus (1980), Dixon (1983, 1996) and Dixon *et al.* (1985, 1988, 1994). A regional synthesis of the maps and publications from Operation Porcupine were summarized in a compilation edited by Norris (1997) on the geology, mineral and hydrocarbon potential of the northern Yukon and northwestern District of Mackenzie.

Preliminary work on the petrography of the Parsons Group rocks was completed on a few samples by Schmidt and MacDonald (1979) in a study of secondary porosity in sandstones. Dixon (1982a) completed limited petrographic work on the Parsons Group sandstones and shales. A study completed by Sykora (1984) provided the most detailed analysis of the petrography and diagenesis of the Parsons Group sandstones to date. Both Dixon (1982a) and Sykora (1984) recognized the highly quartz-rich and texturally supermature nature of these sandstones. For a further discussion of the sedimentary petrology and diagenetic history of the Kamik Formation the author is referred J. Sykora's (1984) master's thesis on the Parsons Group. Recent syntheses of the geology and tectonics of the Beaufort-Mackenzie Delta region are included in papers by Gabrielse and Yorath (1992), Stott and Aitken (1993) and Norris (1997). The most recent research in the Richardson Mountains and Beaufort-Mackenzie Delta region was completed by numerous geologists from the Geological Survey of Canada, who specifically focused on improving the understanding of the structural evolution, basin architecture, and deep crustal structure of the region. Regional and detailed studies of the geology of the Beaufort-Mackenzie Delta region are ongoing and are primarily driven by oil and gas exploration projects from the petroleum industry.

1.6 Stratigraphic Nomenclature

The Kamik Formation was originally part of the "Lower Sandstone Division" informally defined by Jeletzky (1958, 1960) for Lower Cretaceous strata from the Aklavik Range of the northern Richardson Mountains (Fig. 1.3). The "Lower Sandstone Division" was equivalent to part of the present day Parsons Group although the upper units were not yet fully recognized. Jeletzky (1958, 1960) initially subdivided the "Lower Sandstone Division" from the eastern flank of the northern Richardson Mountains into two internal units, the lower "Buff Sandstone Member" which is equivalent to the Martin Creek Formation, and the overlying "White Sandstone Member" which is equivalent to the lower part of the Kamik Formation (Dixon, 1991). In 1961 Jeletzky identified another unit above the "White Sandstone Member" that he termed the "Coal-bearing Division" which is also stratigraphically equivalent to the lower portion of the Kamik Formation. Jeletzky (1961a) also introduced the informal terms "Lower Sandstone Division", "Bluish-Grey Shale Division", "White Quartzite Division" and "Coaly Quartzite Division" for Lower Cretaceous strata on the west flank of the northern Richardson Mountains. The "Lower Sandstone Division" is equivalent to the Martin Creek Formation, the "Bluish-Grey Shale Division" equivalent to the

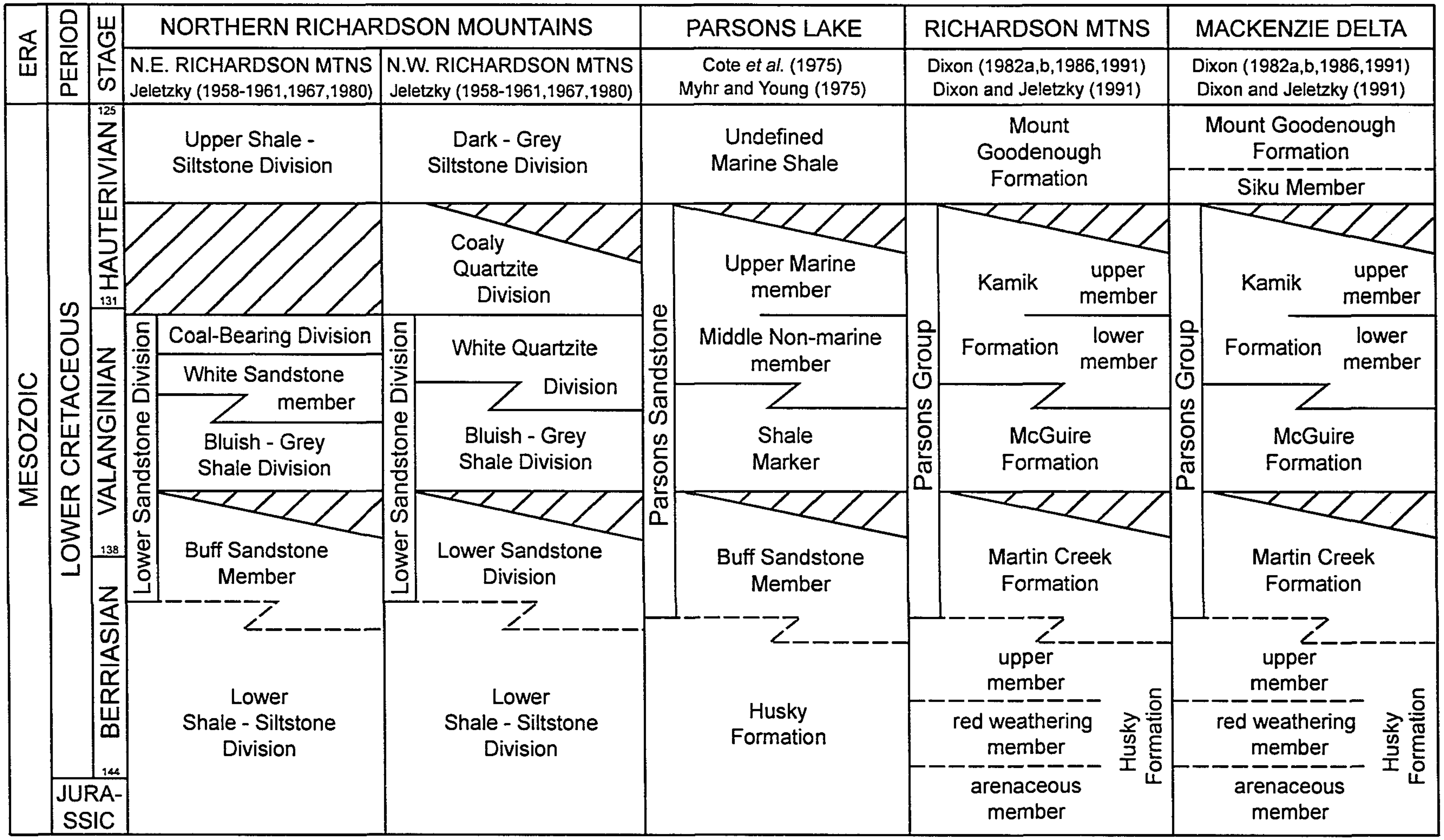


Figure 1.3 Regional stratigraphy of the Lower Cretaceous (Berriasian to Hauterivian) from the Beaufort-Mackenzie basin. The evolution of the Lower Cretaceous strata from informal to formally defined formation status is shown along with the approximate radiometric dates of the stages in millions of years (modified from Dixon, 1982a, 1982b, 1991).

McGuire Formation and the "White Quartzite" and "Coaly Quartzite Division" equivalent to the Kamik Formation. The "White Quartzite Division" is equivalent to the lower portion of the Kamik Formation and the "Coaly Quartzite Division" is equivalent to the upper portion of the Kamik Formation. Jeletzky did not recognize the "Bluish-Grey Shale division" during his initial studies in 1958 and 1960 and it was not identified until 1961 when fieldwork was completed on the west flank of the northern Richardson Mountains. Jeletzky's (1958, 1960, 1961a, 1961b, 1967, 1980) "White Quartzite Division" and "White Sandstone Member" are equivalent to the lower Kamik Formation and the "Coal-bearing Division" and "Coaly Quartzite Division" are equivalent to the upper Kamik Formation.

The term "Parsons Sandstone" was first informally introduced by Coté *et al.* (1975) for the sandstone-dominant basal Lower Cretaceous succession present in the subsurface of the Parsons Lake gas field. The usage of the term was subsequently expanded by Myhr and Young (1975) and applied to Lower Cretaceous strata present in the subsurface over the entire Mackenzie Delta region. The "Parsons Sandstone" is equivalent to the Parsons Group of Dixon (1982a) and includes what is now known as the Martin Creek, McGuire and Kamik formations. The Kamik Formation represented the upper sandstone-dominant portion of the "Parsons Sandstone" succession, which was underlain by Coté *et al.'s* (1975) "Shale Marker" and was overlain by an unnamed marine shale of Valanginian to Albian age. This upper unnamed marine shale was later named the Mount Goodenough Formation by Dixon and Jeletzky (1991). The upper portion of Coté *et al.*'s (1975) "Parsons Sandstone" was equivalent to the Kamik Formation and was subdivided into two members, a lower "Non-marine member" and the "Upper Marine member". Dixon (1982a) officially raised Coté *et al.'s* (1975) "Parsons Sandstone" to group status, naming it the Parsons Group. The Parsons Group was originally proposed by Dixon (1982a) to include the Martin Creek, McGuire and Kamik formations in the subsurface of the Mackenzie Delta. However, only the Kamik Formation was formally defined by Dixon (1982a), with the Martin Creek and McGuire formations subsequently formally defined by Dixon and Jeletzky (1991). The Parsons Group is dated as being late Berriasian to probably middle Hauterivian in age (Dixon, 1982a). The type locality for the Parsons Group as defined by Dixon (1982a) is in the Gulf Mobil Parsons F-09 well located at 68° 58' 34" N; 133° 31' 33" W between the log depths of 2698.1 and 3081.5 m. The F-09 well was the same used by Coté *et al.* (1975) for their original description of the "Parsons Sandstone" from the Parsons Lake area.

The Kamik Formation was first formally defined by Dixon (1982a) in the Mackenzie Delta region as the upper subsurface unit of the Parsons Group. The type section for the Kamik Formation was defined by Dixon (1982a) from the Gulf Mobil Kamik F-38 well located at 68° 57' 25" N; 133° 23' 54" W between the log depths of 3006.5 and 3236.4 m. The name Kamik Formation was extended by Dixon (1986) into the outcrop belts of the northern Yukon and adjacent Northwest Territories to replace the informal names "White Sandstone Member", "Coal-bearing Division", and "White Quartzite Division" and "Coaly Quartzite Division" previously used by Jeletzky (1958, 1960, 1961a, 1961b). Dixon (1991) noted that J.A. Jeletzky had intended to define the Martin Creek, McGuire, Fault Creek and Lower Canyon formations as the surface equivalents of Dixon's subsurface Parsons Group. Dixon (1982a) indicated that the proposed Fault Creek and Lower Canyon formations were to be equivalent to the lower and upper units of the subsurface Kamik Formation. In his study of the Parsons Group, Sykora (1984) subdivided Dixon's (1982a) Parsons Group in the subsurface into three informal members, a "lower non-marine member", a middle "transitional member" and an "upper marine member" based on regional mapping and stratigraphic work on the Parsons Group in the Mackenzie Delta region. Furthermore, Sykora (1984) also recognized Jeletzky's informal "Fault Creek Formation" and "Lower Canyon Formation" as surface equivalents of the subsurface Parsons Group from the northern Richardson Mountains.

The untimely death of J.A. Jeletzky in 1988 left the Fault Creek and Lower Canyon formations informally defined as surface equivalents of the Kamik Formation (Dixon, 1991). Although the Fault Creek and Lower Canyon formations were not formally defined, their usage by Norris (1981a-g, 1982, 1985b) and Dixon (1982b, 1986) had created a *de facto* state of recognition (Dixon, 1991). Following the summary of work by Dixon (1991) on the Parsons Group in outcrop belts in the northern Yukon and adjacent Northwest Territories, it was decided to discontinue the usage of Jeletzky's undefined Fault Creek and Lower Canyon formations because these names were never formally defined and these units represented local facies within the Kamik Formation restricted to the Aklavik Range of the northern Richardson Mountains (Dixon and Jeletzky, 1991). Further work by Dixon (1991) recognized a westward truncation of the upper Kamik strata by the overlying Mount Goodenough Formation that mostly preserved only lower Kamik strata in the Richardson Mountains. Dixon (1991) felt that it was unnecessary to create another set of names for Parsons Group equivalent strata and chose to extend the name Kamik Formation into the outcrop belt in the Richardson Mountains. Dixon and Jeletzky (1991) formally assigned the same formation names for surface and subsurface stratigraphic units of the Parsons Group to include the Martin Creek, McGuire and Kamik formations. Figure 1.3 summarizes the evolution of the Lower Cretaceous surface and subsurface lithostratigraphic terminology from the northern Richardson Mountains, Mackenzie Delta and Tuktoyaktuk Peninsula. The stratigraphic chart includes the informal and formal terminology from papers by Jeletzky (1958, 1960, 1961a, 1961b, 1967, 1980), Coté *et al.* (1975), Myhr and Young (1975), Dixon (1982a, 1986, 1991) and Dixon and Jeletzky (1991).

1.7 Regional Geology and Stratigraphy

The Kamik Formation is the uppermost sandstone-dominant unit of the Lower Cretaceous Parsons Group and is middle to late Valanginian to middle to late Hauterivian in age (Dixon 1982a, 1982b, 1991) (Fig. 1.3). The Kamik Formation is a sandstone and shale unit unconformably overlying the shale-rich McGuire Formation and is unconformably overlain by the shales of the Mount Goodenough Formation (Dixon, 1991; Dixon and Jeletzky, 1991). In the subsurface of the

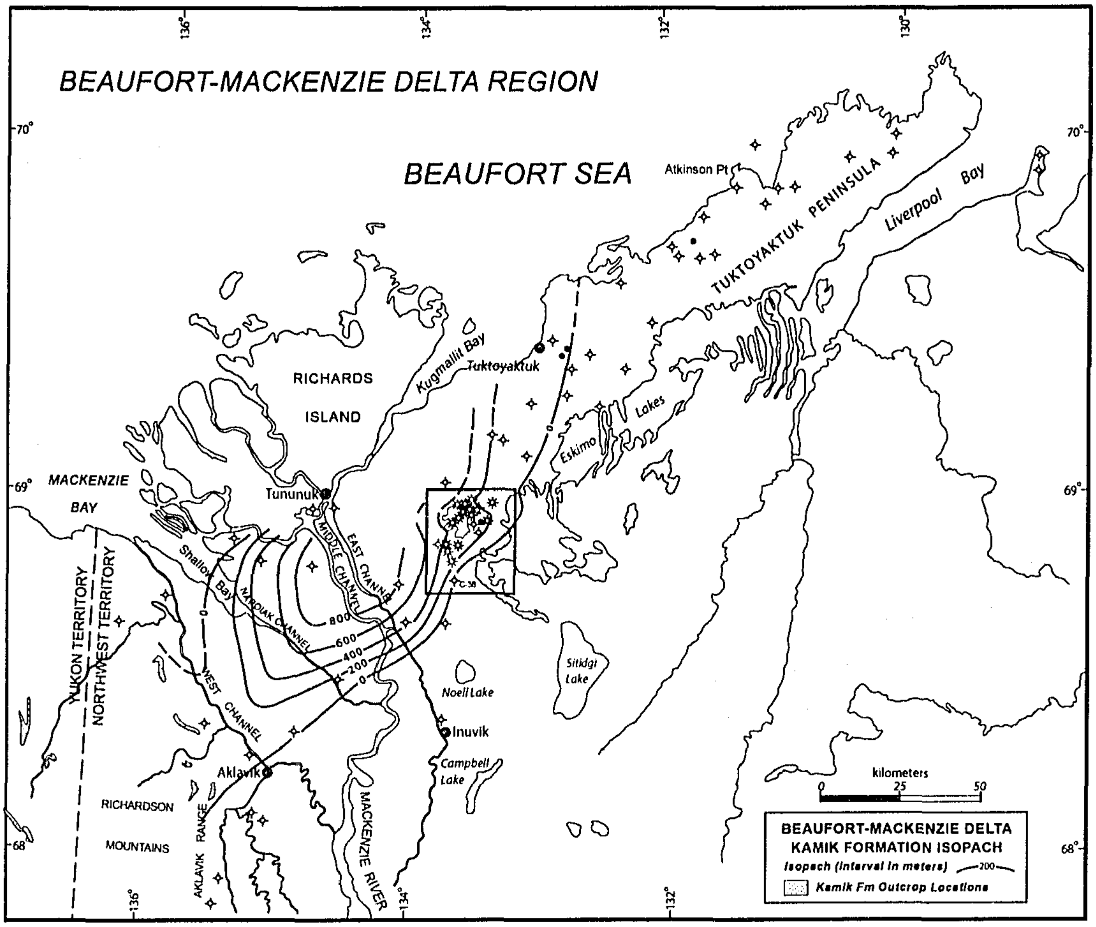


Figure 1.4 Isopach map of the Kamik Formation with outcrop locations indicated (modified from Dixon, 1982b)

Mackenzie Delta and Tuktoyaktuk Peninsula, the Kamik Formation is overlain by the Siku Member of the Mount Goodenough Formation (Dixon, 1991), whereas in the outcrop belt in the Richardson Mountains the Siku Member is undifferentiated from the Mount Goodenough Formation. The entire Parsons Group succession sits conformably on shales of the Upper Jurassic to Lower Cretaceous Husky Formation (Norris, 1973, 1974; Dixon, 1982a; Braman, 1985). The Parsons Group is biostratigraphically dated as late Berriasian to middle Hauterivian (Jeletzky, 1958, 1960, 1961a, 1961b, 1975; Dixon, 1982a, 1982b, 1991).

Regionally, the Kamik Formation extends west from the outcrop belt in the northern Yukon and adjacent Northwest Territories to the subsurface of the onshore Mackenzie Delta and the Tuktoyaktuk Peninsula. Kamik strata are found in outcrop in the northern Richardson Mountains, on the perimeter of the British and Barn mountains, western Keele Range and in the northern Ogilvie Mountains (Fig. 1.4) (Dixon, 1986, 1991). The Kamik Formation is speculated to extend north and northwest into the subsurface of the southern offshore region under the modern Mackenzie Delta. However, approximately 3,000 to 4,000 m of Upper Cretaceous and Tertiary sediments overlie the Lower Cretaceous strata and exploratory wells have been unable to identify Lower Cretaceous strata in the subsurface of the offshore region. In the subsurface of the Mackenzie Delta and Tuktoyaktuk Peninsula, the Kamik Formation is the thickest and most well preserved in the Kugmallit Trough, with thicknesses exceeding 800 m (Dixon, 1982a, 1991, 1992). Regionally, Kamik Formation strata thin in a westward direction towards the Richardson Mountains with the thickest preserved stratigraphic sections in the east and the thinnest successions preserved in the Richardson Mountains to the west. In addition, Kamik Formation strata thin in an easterly and southernly direction towards the Eskimo Lakes Arch, Cache Creek Uplift and Tununuk High due to depositional thinning and erosional truncation (Dixon, 1982a, 1991). Within the study area, Lower Cretaceous strata rapidly thicken northwest (basinward) of the Eskimo Lakes Arch and Fault Zone into the Kugmallit Trough (Young *et al.*, 1976; Dixon, 1982a).

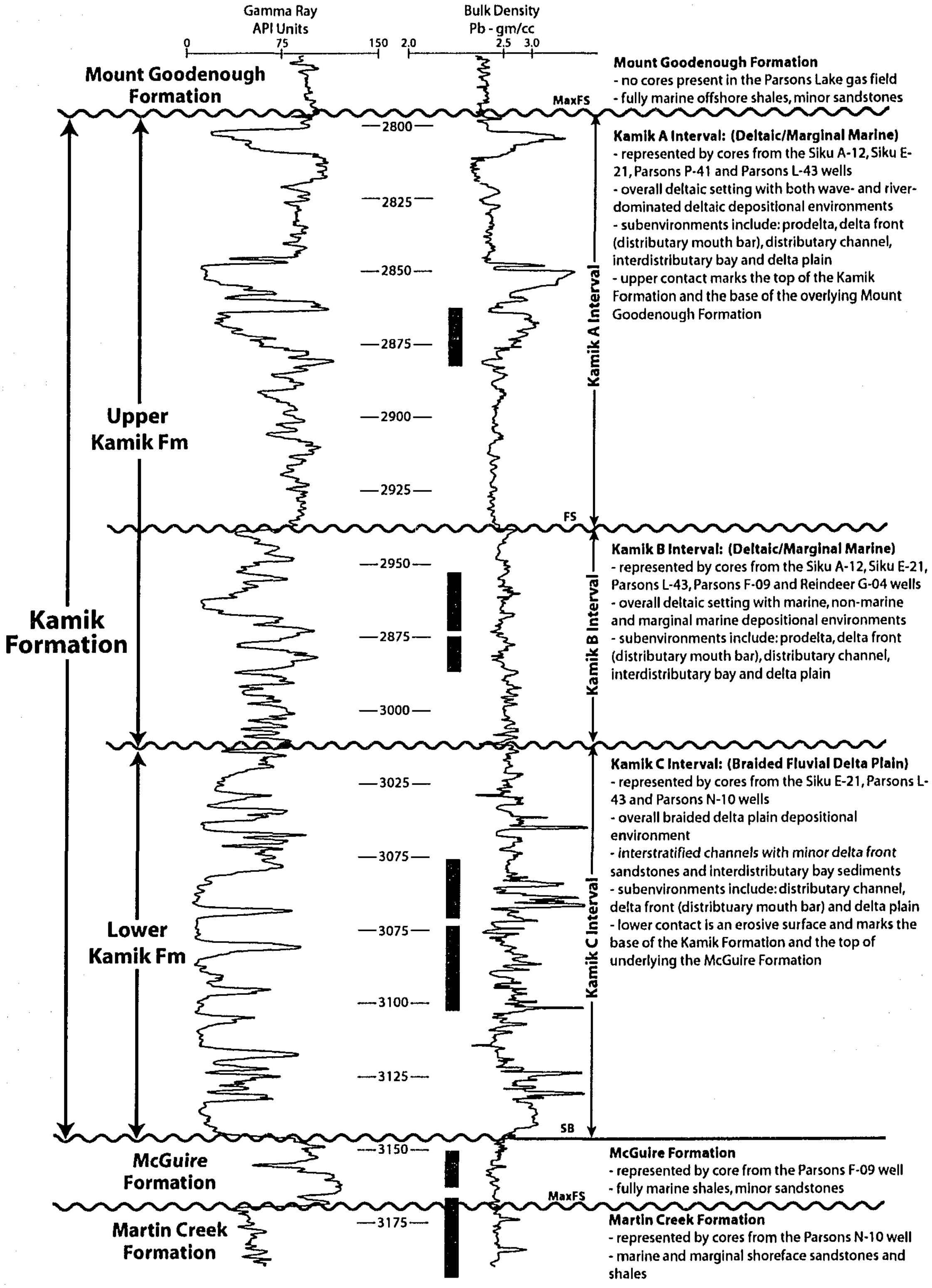
The Kamik Formation is informally subdivided into two members, a lower member and an upper member (Fig. 1.5). The stratigraphic and sedimentological character of each of these members is discussed further below. The thicknesses of the lower and upper members of the Kamik Formation are highly variable in the subsurface of the Mackenzie Delta. The lower member is approximately 100 to 200 m thick, while the upper member is up to 700 m thick, although it can be considerably thinner where removed by post-Kamik erosion (Dixon, 1991). Within the Parsons Lake study area, the Kamik Formation is approximately 300 to 350 m thick with the lower member ranging between 110 to 130 m thick and the upper member ranging between 190 to 220 m thick (D. Bywater pers. comm., 2004). In addition, many of the exploratory wells in the Mackenzie Delta region, especially those proximal to the Eskimo Lakes Arch and Eskimo Lakes Fault Zone, display truncated and partially preserved stratigraphic sections of Kamik strata due to erosion and fault displacement.

In the outcrop belt, Kamik strata range in thickness from 450 to 1800 m. Dixon (1992) noted that the average thickness of Kamik strata from outcrop ranges between 500 to 800 m, with many of the thicker measured sections located within the depocenters. Dixon (1991) noted that there are large variations in measured thicknesses over short distances in the outcrop belt and interpreted this to be due to fault repetition. Therefore, the exact thicknesses of Kamik Formation strata are difficult to determine in outcrop due to a lack of complete, well-exposed and accessible sections where reliable thicknesses can be measured (Dixon, 1991). The distribution of Kamik strata has been modified by late Hauterivian erosion that occurred post-Kamik time, but prior to deposition of the Mount Goodenough Formation. In many parts of the subsurface of the Tuktoyaktuk Peninsula (i.e. Eskimo Lakes Arch) the entire Kamik section has been erosionally removed. In some wells, remnants of Upper Jurassic Husky strata are directly overlain by the Barremian Mount Goodenough Formation. The preservation of Husky strata suggests that Kamik strata extended onto the structurally positive areas, such as the Eskimo Lakes Arch, but were subsequently removed by pre-Mount Goodenough erosion. Dixon (1991) also noted that the Kamik Formation was affected by local erosional events, especially during the late Aptian to early Albian.

The contact between the Kamik Formation and the underlying McGuire Formation is both erosional and gradational in nature (Fig. 1.5). Along the flanks of the Kugmallit Trough the contact is generally erosional, whereas in the deeper parts of the central trough the contact is gradational (Dixon, 1982a, 1982b). In the Richardson Mountains the contact between the McGuire and Kamik formations was found by Dixon (1991) to be either abrupt or transitional over a very short distance. Up to 10 m of transitional beds composed of interbedded mudstones and sandstones were found in outcrop locations where a gradational contact was recognized (Dixon, 1991). These transitional beds display a gradual upward increase in sandstone content and a decrease in shale content. These transitional beds were interpreted by Dixon (1991) as shoreface or delta front deposits of the lower Kamik Formation. The gradational contact between the lowermost beds of the Kamik Formation and the underlying McGuire Formation was interpreted by Dixon (1992) to represent the westward to northwestward progradation of a shoreline. Dixon (1991) interpreted the sharp to scoured nature of the lower contact at the base of the Kamik Formation to have resulted from channels resting erosionally on McGuire Formation strata. Within the Parsons Lake gas field the contact between the McGuire and Kamik Formation is also generally abrupt to erosional with a few wells displaying contacts that appear to be gradational on geophysical well logs. Unfortunately no core data is available from this contact, therefore the exact nature of the contact is speculative. However, in many of the wells, a 10 to 35 m thick blocky sandstone unit directly overlies McGuire Formation strata and is interpreted to represent a scoured contact composed of channel sands unconformably overlying fine-grained marine mudstones.

The contact between the Kamik Formation and the overlying Mount Goodenough Formation is generally abrupt and is recognized as regional unconformity (Dixon 1982a, 1991). Dixon (1982b) recognized a major erosional event that occurred in the late middle or late Hauterivian, during which older Mesozoic strata were removed from structurally positive areas. Young *et al.* (1976) also interpreted a period of late Hauterivian uplift and erosion of the Eskimo Lakes Arch, Cache Creek, Campbell and Rat Uplifts that removed large portions of Parsons Group strata from crestal areas of these structurally positive elements. Regionally, Kamik strata are generally thin or absent due to erosional removal from tectonic highs including the Eskimo Lakes Arch, Cache Creek Uplift, Eagle Arch, Campbell, Rat and Romanzof Uplifts and their flanks (Dixon, 1982b). Dixon (1982a) noted that within areas with closely spaced data, especially within the subsurface of the Tuktoyaktuk Peninsula, Kamik strata thin toward tectonic highs that are interpreted to have had significant influence on Kamik deposition. The magnitude of erosion along the top of the Kamik Formation was recognized by Dixon (1991) to have varied according to basin position and proximity to tectonic highs. As mentioned earlier, there is regional westward truncation of the upper Kamik strata by the overlying Mount Goodenough Formation that mostly preserved only lower Kamik strata in the Richardson Mountains. Tectonic depressions such as the Kugmallit Trough, Canoe Depression, Rapid Depression and Kandik Basin acted as the major depocenters during Kamik deposition. Erosion along the top of the Kamik Formation is minimal or undetectable in many of

Kamik Formation - Gulf Mobil Siku E-21



these tectonic depressions (Dixon, 1991).

As mentioned previously, the Kamik Formation is generally subdivided into two informal members, a lower member and an upper member (Dixon, 1991, 1992). This informal subdivision of the Kamik Formation was first introduced by Coté *et al.* (1975) and Myhr and Young (1975). Coté *et al.* (1975) subdivided the Kamik-equivalent of their "Parsons Sandstone" into a lower "Non-marine member" and an upper "Marine member" (Fig. 1.3). This subdivision was based mainly on lithological characteristics and interpreted depositional settings from the subsurface of the Parsons Lake gas field. This subdivision was continued by Dixon (1982a, 1982b, 1991) throughout its occurrence in the subsurface and in outcrop in the northern Richardson Mountains. Dixon (1991) informally designated the lower and upper members as part of the Kamik Formation and expanded upon the detailed stratigraphy and interpreted depositional environments. It is important to note that Dixon (1982a) recognized that the informal subdivision of Kamik-equivalent strata by Coté *et al.* (1975) and Myhr and Young (1975) into marine and non-marine lithogenic units was based in part on vertical succession and lateral equivalency.

Within the Parsons Lake area, the informal subdivision of the Kamik Formation into a lower non-marine member and an upper marine member is generally consistent with the vertical succession of facies recognized in core. In the study area, the lower portion of the Kamik Formation is generally composed of non-marine and marginal marine sediments and the upper portion composed of marine and marginal marine sediments. However, it is recognized that both the lower and upper members can include both marine and non-marine sediments throughout their occurrence in the subsurface and in outcrop. During deposition of the Kamik Formation, the study area was located in the most landward position in the basin; therefore, the lower non-marine and upper marine character is generally a localized occurrence as it varies according to basin position. Overall it is generally recognized that the nature of the Kamik Formation varies from non-marine to marine in a westerly and northwesterly (basinward) direction and the strict adherence to a non-marine interpretation for the lower member and a marine interpretation for the upper member is not always applicable as this basic subdivision is best applied in a local sense.

Dixon's (1982a, 1982b, 1991) basic two-fold subdivision for the Kamik Formation included: 1) a lower, sandstone-dominant interval and 2) an upper interval of interbedded sandstone and shale. Sykora (1984) further subdivided the Kamik Formation into three informal members, a lower non-marine member, a middle transitional member and an upper marine member. Sykora (1984) recognized a regional correlation between the grain size of the informal members with the non-marine member composed of the coarsest-grained sandstones and the transitional and marine members characterized by finer-grained sandstones. The Kamik Formation in the Parsons Lake gas field has also been informally subdivided by petroleum companies into three units, the A, B and C (Fig. 1.5). The "A" and "B" intervals correspond to the upper member and the "C' interval is roughly equivalent to the lower member of Dixon (1982a, 1991). For the purposes of this study, the informal subdivision of the Kamik Formation into a lower member composed predominantly of non-marine and marginal marine sediments and an upper member composed of marine to marginal marine is applied throughout the Parsons Lake study area.

The lower one-third of the Kamik Formation is a sandstone-dominant succession composed of thick sandstone units that generally display fining-upwards trends. This lower member was subdivided by Dixon (1982a) into two successive parts, a basal portion composed of thick sandstone units (average thicknesses between 10 to 35 m) and minor shale and thin coal interbeds and an upper portion composed of significantly thinner sandstone units interbedded with mudstone and minor coal beds. In outcrop the lower member is up to 200 m thick, although average thicknesses of 75 to 150 m are more common (Dixon, 1992). Lithologically, the lower Kamik Formation is composed of 70 to 80 percent sandstone with shale comprising less than 20 to 30 percent of the succession (Dixon, 1991, 1992). These sandstones are fine to coarse-grained, locally pebbly and generally become fine-grained upwards in the succession. The sandstones comprising the lower member are quartz-arenites with grain sizes ranging from very fine to very coarse with local occurrences of granular to pebble-sized conglomerates. The fining-upwards cycles typically have a basal scoured surface overlain by massive or cross-bedded, pebbly, medium to coarse-grained sandstone (Dixon, 1992).

The basal one-third of the Kamik Formation was interpreted by Shawa *et al.* (1974), Coté *et al.* (1975), Myhr and Young (1975) and Young *et al.* (1976) as a non-marine succession composed of delta plain deposits with prevalent distributary channels, overbank (levee, splay) and floodplain (lake, swamp, marsh) deposits. Coté *et al.* (1975) interpreted the upper part of their non-marine member as a transitional, lower coastal plain environment including floodplains, marshes and tidal flats distant from major river channels. Dixon (1982a, 1982b) interpreted the lower member from the subsurface to consist of delta plain deposits composed of fluvio-deltaic, fluvial-channel, crevasse splay, levee, floodplain and overbank deposits with some marginal marine sediments in the upper portions. Sykora (1984) agreed with the non-marine to coastal transition but specifically interpreted the channel deposits as representing a high-energy, braidplain environment in the east grading into a meandering river system in a basinward (westerly) direction. Sykora (1984) interpreted the depositional environments of the transitional member to range from delta front, meandering fluvial channel, floodplain (crevasse splay) to offshore. Dixon (1991) interpreted the depositional environment for the lower member from outcrop in the northern Richardson Mountains as a part of a large fluvio-deltaic setting, composed of delta front, distributary mouth bar, fluvial channel, overbank and interdistributary bay deposits. The largely non-marine lower succession of the Kamik Formation was also interpreted by Dixon (1991, 1992) as shoreface and delta front sediments with recognizable marine influences westward towards the Richardson Mountains. These delta front deposits in the Richardson Mountains were found by Dixon (1992) to be overlain by typical alluvial deposits in fining-upwards cycles that were interpreted to have been deposited on a prograding delta plain (Dixon, 1982b). In the Richardson Mountains, the lower Kamik Formation is comprised of lower shoreface to shallow shelf deposits, characterized by bioturbated strata and beds formed by storm-related processes (Dixon, 1992). The vertical change from delta plain to delta front sedimentation in the Richardson Mountains was interpreted by Dixon (1992) as reflecting a change from a progradation to transgression, during a period of overall high sedimentation rates. Dixon (1992) speculated that a decline in sedimentation rates could have induced transgression and abandonment of deltaic sedimentation. These changes in sedimentation rates coincided with continuous subsidence in the Brooks-Mackenzie Basin. However, it was noted that subsidence was not uniform, as evident from the growth of local structures, such as the faulted anticline of the Parsons Lake gas field (Coté *et al.*, 1975). The deltaic setting responsible for the deposition of the lower member was terminated by a marine transgression prior to deposition of the upper member (Dixon, 1991).

The upper two-thirds of the Kamik Formation consists of interbedded sandstones and mudstones, with minor coal beds. In outcrop the upper member ranges between 500 to 700 m thick, although average thicknesses of 200 to 300 m are most common (Dixon, 1992). The majority of the sandstone units form coarsening-upwards sequences with minor fining-upwards sequences. These coarsening-upwards sequences range in thickness from a few meters up to about 150 m (Dixon 1982a, 1982b). The upper member is composed of significantly less sandstone than the lower member, with the sandstone content ranging between 40 to 60 percent in the Parsons Lake area. Westward and southwestward the sandstone content decreases to between 20 to 30 percent (Dixon, 1982a). The sandstones in the upper member are also quartz arenites, but are generally finer-grained than those in the lower member. The upper member was interpreted by Coté *et al.* (1975) and Myhr and Young (1975) to consist of a complex interdigitation of marine and non-marine beds (Dixon, 1982a).

Myhr and Young (1975) interpreted the coarsening-upwards cycles in the upper Kamik Formation to consist of prodelta and delta front deposits interbedded with channel and delta plain deposits. Coté *et al.* (1975) interpreted the coarsening-upwards cycles present in the basal portion of the upper member in the Parsons Lake area to represent a transgressive barrier island-lagoon system. These sediments were interpreted by Coté *et al.* (1975) as being in a transitional to nearshore marine shelf deposited during the initial stages of a marine transgression, which flooded the coastal plain and terminated deposition of the Kamik Formation. Dixon (1982a) interpreted the upper member as consisting of a series of stacked, prograding barrier-island successions, including coarsening-upward cycles interpreted as barrier bar, tidal inlet and lagoonal deposits. Sykora (1984) also agreed with Dixon's (1982a, 1982b) interpretation of the upper member as a series of barrier island successions, composed of lagoonal, intertidal, nearshore and offshore deposits. Dixon (1982a) identified three coarsening-upward cycles within the Parsons Lake area, interpreted as barrier island deposits. The upper member is characterized by a series of small-scale transgressive-regressive events that resulted in the production of the coarsening-upwards cycles (Dixon, 1992). Towards the Richardson Mountains Dixon (1992) interpreted these coarsening-upward cycles as entirely marine and dominated by storm-generated sedimentary structures. A distinct east to west facies change is seen in the upper member of the Kamik Formation, with an increase in shale content and a decrease in the abundance, thickness and grain size of the sandstone beds in a westward direction. Dixon (1982a) was able to correlate these cycles up to 45 km between wells in the Parsons Lake area.

Dixon (1982a) recognized a consistent marker horizon within the subsurface of the Parsons Lake gas field that separated the basal one-third from the upper two-thirds of the Kamik Formation. This horizon was interpreted as significant, but Dixon (1982a) recognized that the entire succession of the Kamik Formation was related to a depositional event and considered it impractical to segregate units. The contacts between the informal lower and upper members of the Kamik Formation were interpreted by Coté *et al.* (1975) and Myhr and Young (1975) as major diachronous facies boundaries, however Dixon (1982a) suggested that only slight diachroneity existed within the subsurface. For a further discussion of the stratigraphic and sedimentological characteristics of the lower and upper members of the Kamik Formation in outcrop the reader is referred to Dixon (1991). The sedimentological and ichnological characteristics and paleoenvironmental interpretations of the Kamik Formation from this study are discussed in Chapters Two and Three.

1.8 Biostratigraphy

The Kamik Formation is Neocomian in age and is speculated to represent middle to late Valanginian to the middle to late Hauterivian stages (Fig. 1.6) (Jeletzky, 1958, 1960, 1961a, 1961b, 1967, 1975; Myhr and Gunther, 1974; Brideaux and Myhr, 1976; Dixon, 1982a, 1991, 1992). The exact age of the Kamik Formation is poorly constrained by biostratigraphic data due to the sparse presence of indigenous fossils. Dixon (1982a) noted that macrofossils and microfossils are not particularly abundant in the Kamik Formation and therefore it is difficult to accurately place biostratigraphic age limits on the strata. There are bivalve fossils found in the Kamik Formation, however most of these macrofossils tend to be non-age-diagnostic forms. The paucity of fossils from Kamik Formation strata led Jeletzky (1960, 1961a, 1961b) to rely mostly on the stratigraphic position of the more fossiliferous McGuire and Mount Goodenough strata to date the rocks. The youngest age of the Kamik Formation was derived from the age of the immediately overlying Mount Goodenough Formation, which is mostly likely Barremian, but was found by Jeletzky (1958, 1960, 1961a, 1961b) to possibly be as old as late Hauterivian (Dixon, 1991). The entire Parsons Group succession is biostratigraphically dated as Neocomian in age using macrofossils and represents the late Berriasian to probably middle Hauterivian stages (Jeletzky, 1958, 1960, 1961a, 1961b, 1975; Dixon, 1982a, 1982b, 1991).

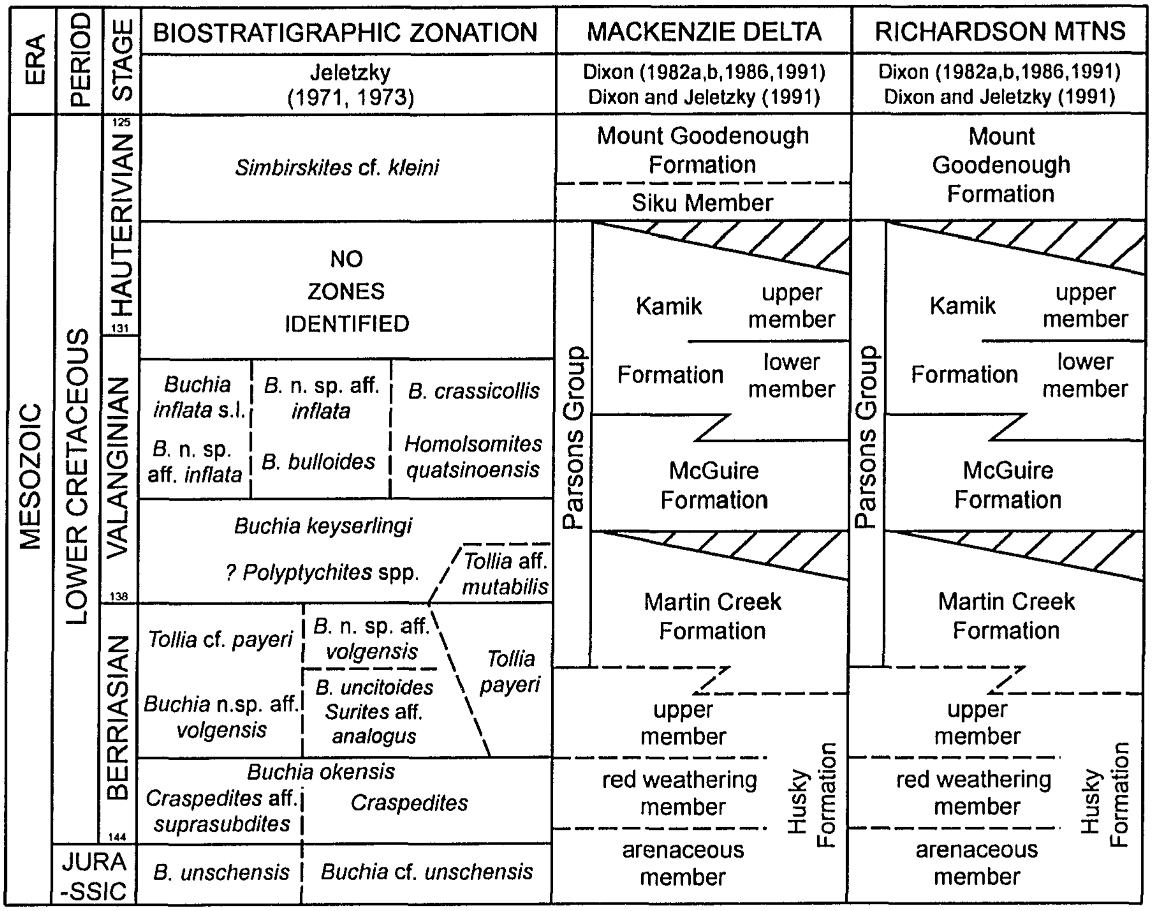


Figure 1.6 Biostratigraphic zonation of the Lower Cretaceous strata (Berriasian to Hauterivian) from the Beaufort-Mackenzie basin along with the approximate radiometric dates of the stages in millions of years (modified from Dixon, 1991).

The lowermost portion of the Kamik Formation contains bivalves from the *Buchia* n. sp. aff. *inflata* zone (Jeletzky, 1958, 1960, 1961a, 1961b, 1975) that were dated as late early to early late Valanginian (Dixon, 1991, 1992). Jeletzky (1973) also reported that bivalves of the *Buchia* ex gr. *inflata-sublaevis* Zone (Valanginian) in beds equivalent to the lowermost Kamik Formation. In addition, on the eastern slopes of the Richardson Mountains Jeletzky (1960) identified the bivalves *Buchia crassa* and *Buchia* n. sp. aff. *crassa* that were identified as middle or late Valanginian age (Dixon 1992). There are other bivalves present within the Kamik Formation but these species are undiagnostic, long-ranging types, such as *Astarte*. Jeletzky (1960) identified a few belemnites from the upper member of the Kamik Formation that could indicate a possible middle Hauterivian age for this interval. Foraminifera are locally abundant within the Kamik strata, especially in the shale-rich upper member, but these are also long ranging, non-age-diagnostic forms. The microfossils that have been recovered from the Kamik Formation indicated a Valanginian to Hauterivian age (Barnes *et al.*, 1974; Myhr and Gunter, 1974; Brideaux *et al.*, 1976; Brideaux and Myhr, 1976). Although microfossils are locally abundant within the Kamik Formation Dixon (1992) noted that recovery of these foraminifers and palynomorphs was usually poor and can only be used to support a general Neocomian age. Dixon (1991) noted that many of the foraminiferal species also occur in either the underlying McGuire Formation or the overlying Mount Goodenough Formation. Work by Brideaux and Myhr (1976) from the Parsons N-10 well included Kamik strata in their dinoflagellate zone IVe which they dated as "Valanginian?-Hauterivian". Dixon (1991) found that dinoflagellates and palynomorphs were of limited use in the outcrop belt because of high levels of thermal alteration. Recent biostratigraphic analysis contracted by ConocoPhillips Canada has identified a reworked Mississippian assemblage for the Kamik Formation. This Mississippian assemblage is interpreted as originating from the erosional reworking of Paleozoic rocks as sources for the sediments comprising the Kamik Formation.

1.9 Regional Paleogeography

The Kamik Formation was deposited as a thick, sandstone-dominant succession within the Mesozoic-aged Brooks-Mackenzie Basin during the Early Cretaceous. During the Mesozoic era, the Brooks-Mackenzie Basin was part of a shallow, rifted pericratonic marine basin in which thick successions of clastic sediments were deposited (Young *et al.,* 1976; Dixon 1982b). This marine basin was located on the northwestern corner of the North American craton (Fig. 1.7) and occupied a v-shaped region of the northern Yukon and adjacent Northwest Territories. The overall setting of the region was that of a broad cratonic shelf over much of the northern Yukon Territory with a deltaic complex located in the present location of the Mackenzie Delta (Dixon, 1993; Stott *et al.,* 1993). Throughout the Jurassic to Early Cretaceous, the region was influenced by extensional tectonics related to the rifting in the Brooks-Mackenzie Basin (Dixon, 1991). The basin was bounded to the southeast and east by the North American craton, Eskimo Lakes Arch and the Eskimo Lakes Fault Zone and to the north and northwest by the newly forming Arctic Ocean (Dixon, 1982b). Balkwill *et al.* (1983) identified an apparently extensive landmass to the north, which they interpreted to have bounded the newly forming Arctic Ocean and Brooks-Mackenzie Basin. However, it should be noted that some plate-tectonic reconstructions have interpreted that no cratonic landmasses were present immediately north or northwest of the Arctic North America during the Mesozoic. Due to the lack of a general consensus about the formation of the Arctic Ocean and the position of Alaska during the Early Cretaceous, definitive paleogeographic reconstructions of the Brooks-Mackenzie Basin are difficult (Balkwill *et al.,* 1983; Dixon, 1986).

During deposition of the Kamik Formation two major depocenters are interpreted to have been present within the Mackenzie Delta region, the Canoe Depression and the Kugmallit Trough (Stott *et al.,* 1993). Sediments comprising the Kamik Formation prograded westward and northwestward across the active rifted margin of the North American craton into the Arctic Ocean to the north (Dixon, 1982a). Kamik Formation clastic sediments were shed from highlands to the east, south and southeast and prograded towards the west and northwest. Dixon (1982b) recognized a northwesterly shale-out trend into basinal deposits and interpreted a general southerly source area

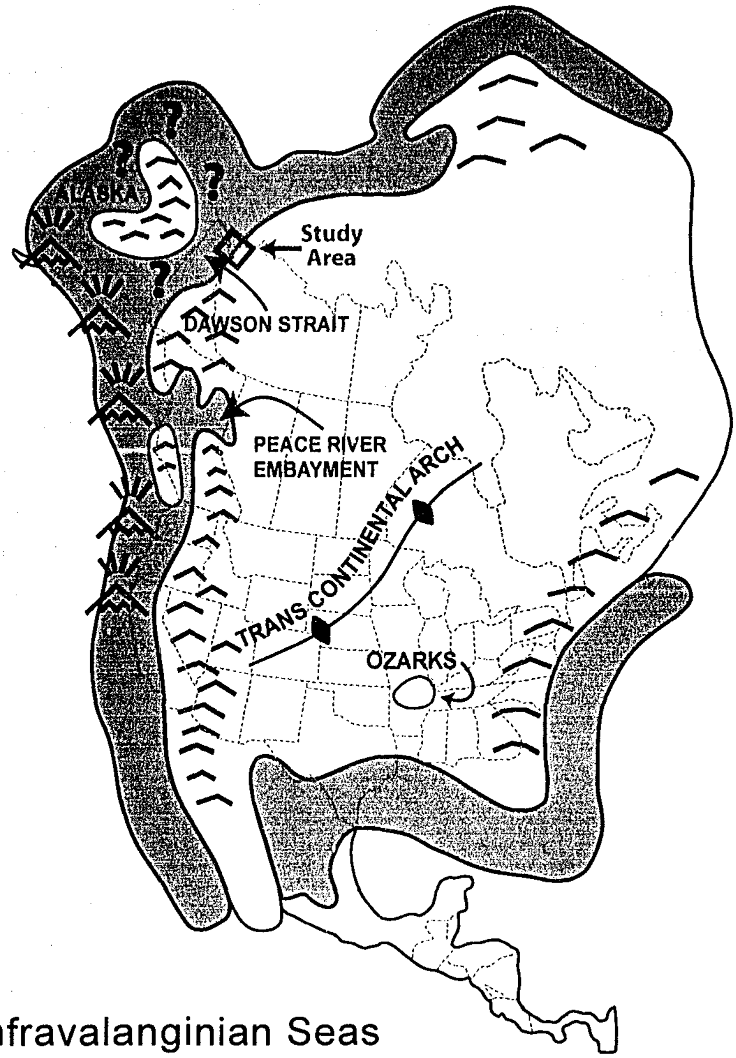


Figure 1.7 Paleogeographic map showing the extent of the earliest Cretaceous Infravalanginian seas. Note the location of the regional study area marked by the red box (modified from Williams and Stelck, 1975).

for the coarse clastic material comprising the Kamik Formation. The regional paleo-shoreline during Kamik deposition is interpreted to have trended southwest to northeast, approximately parallel to the trend of the Eskimo Lakes Arch (Dixon, 1982a, 1982b, 1986, 1991). The interpreted orientation of the paleoshoreline and inferred directions of sediment sources for the lower and upper members of the Kamik Formation is illustrated in Figures 1.8 and 1.9.

The large quantities of clastic material comprising the Kamik Formation are interpreted to have been derived from adjacent tectonic highs including the North American craton, Eskimo Lakes Arch, Cache Creek High and the Campbell Uplift. These tectonic elements influenced the deposition of the Kamik Formation and were interpreted to have formed due to oscillatory tectonic movements resulting in the emergence of short-lived and localized source areas (Jeletzky, 1975).

It was also noted by Jeletzky (1975) that these tectonic elements were merely 2nd and 3rd order features occurring within an overall regionally subsiding basin under the influence of regional extension. The strongest evidence for these tectonic highs acting as sediment sources includes partial to complete removal of pre-Cretaceous strata (i.e. Triassic and Jurassic) in some areas and non-deposition and regional stratigraphic thinning of Kamik strata over these structures.

The quartz-rich, sandstone-dominant nature of the Kamik Formation is attributed to the reworking and cannibalization of older sedimentary rocks from the uplifted margins of the Brooks-Mackenzie Basin. The most probable provenance sources for the large quantities of reworked quartz-rich, clastic rocks include Proterozoic, Paleozoic and early Mesozoic-aged (Triassic and Jurassic) sedimentary rocks uplifted and eroded from the adjacent highlands. The main source of sediment throughout the Early Cretaceous is attributed to a principally cratonic source which provided quartz-rich, relatively mature sediments (Dixon, 1991). This clastic material was likely sourced from adjacent tectonic highs including the emergent North American craton and especially the Eskimo Lakes Arch and to a lesser degree the Cache Creek Uplift. Dixon (1991) did not recognize any texturally and mineralogically immature sediments that would indicate direct influence of Cordilleran tectonics during the Neocomian. Sediment transport into the Parsons Lake area was mainly from the east and southeast and was controlled by the emergent Eskimo Lakes Arch. Young *et al.* (1976) interpreted the development of large deltaic complexes along the northwestern flank of the Eskimo Lakes Arch partly in response to tectonic uplift of the arch. In addition, they interpreted that the locus of deltaic sedimentation during the lower member shifted numerous times with the greatest accumulations of deltaic sediments deposited in the area proximal to the Ikhil I-37 well, north of Inuvik. During deposition of the upper member of the Kamik Formation Young *et al.* (1976) interpreted that deltaic sedimentation shifted away from the present area of the Tuktoyaktuk Peninsula toward the southwest in the area of Aklavik. Dixon (1982b) was unable to identify any single point source for the younger clastics and interpreted a broad alluvial plain that developed adjacent to a southern source area.

The Kamik coastline was likely positioned near the crest of the Eskimo Lakes Arch, as indicated by the large amounts of coarse-grained sandstone proximal to the arch. The uplifting of Eskimo Lakes Arch and the associated downward movement of normal faults along the Eskimo Lakes Fault Zone created a major depocenter proximal to a large source of mature clastic sediments to the west and northwest. The uplift of clastic rocks on the Eskimo Lakes Arch and the creation of accommodation space by extensional tectonics initiated a period of major west and northwesterly deposition of clastic sediment into the Brooks-Mackenzie Basin. During the late Berriasian to Valanginian the main depocenter was in the Kugmallit Trough north of Inuvik (Young *et al.*, 1976). In Hauterivian time, the main deltaic depocenter shifted to the northern Yukon and a thick, shallow-marine facies accumulated in the Kugmallit Trough (Young *et al.*, 1976). Across the Eskimo Lakes Arch much of the Paleozoic strata and older Mesozoic sedimentary rocks has been erosionally removed. Within

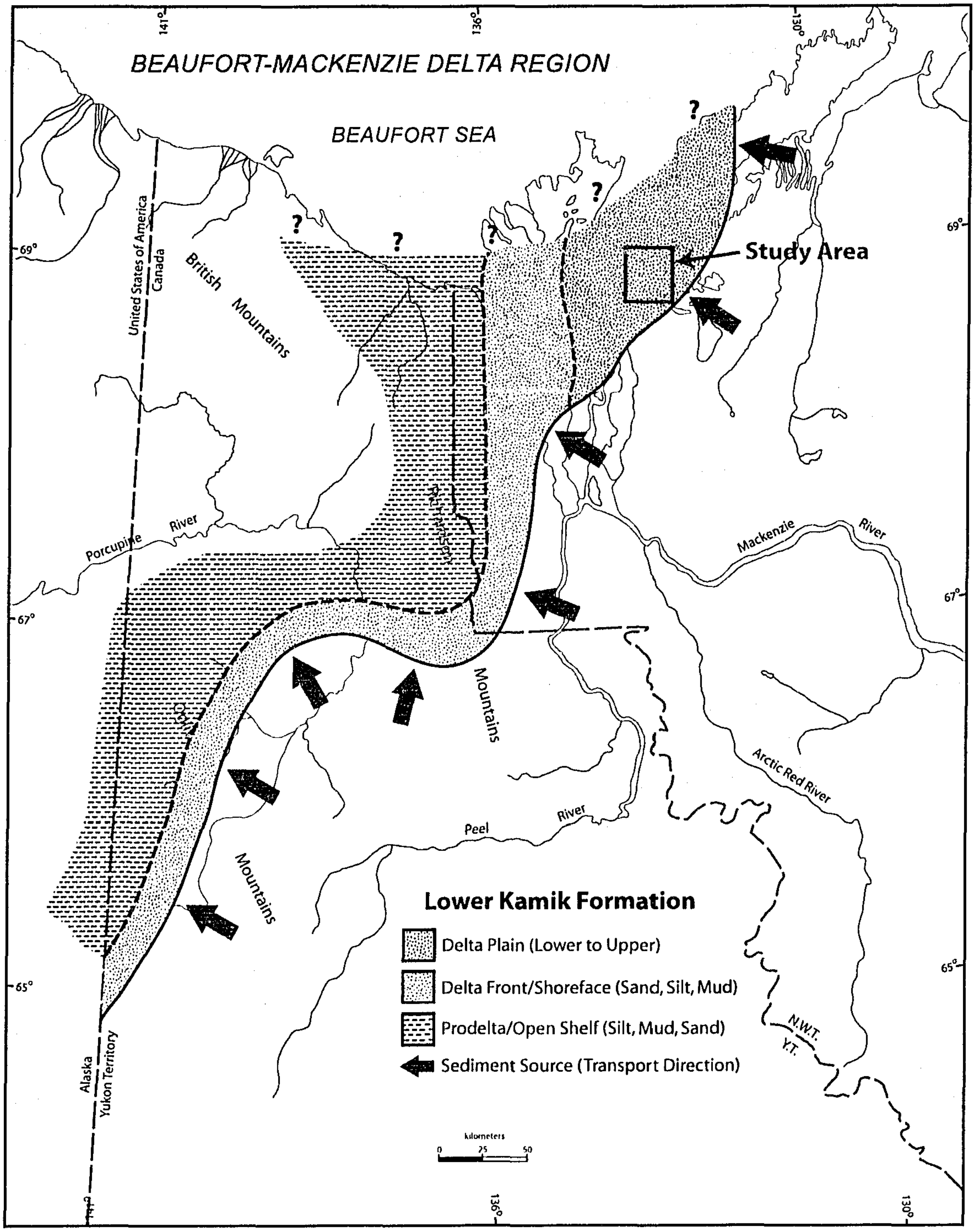


Figure 1.8 Paleogeographic map of late Valanginian time during deposition of the Lower Kamik Formation in the Beaufort-Mackenzie Delta region. The inferred directions of sediment transport is indicated by the red arrows. Note the geographic location of the study area relative to inferred paleoshoreline. No palinspastic reconstruction is attempted (modified from Dixon, 1991).

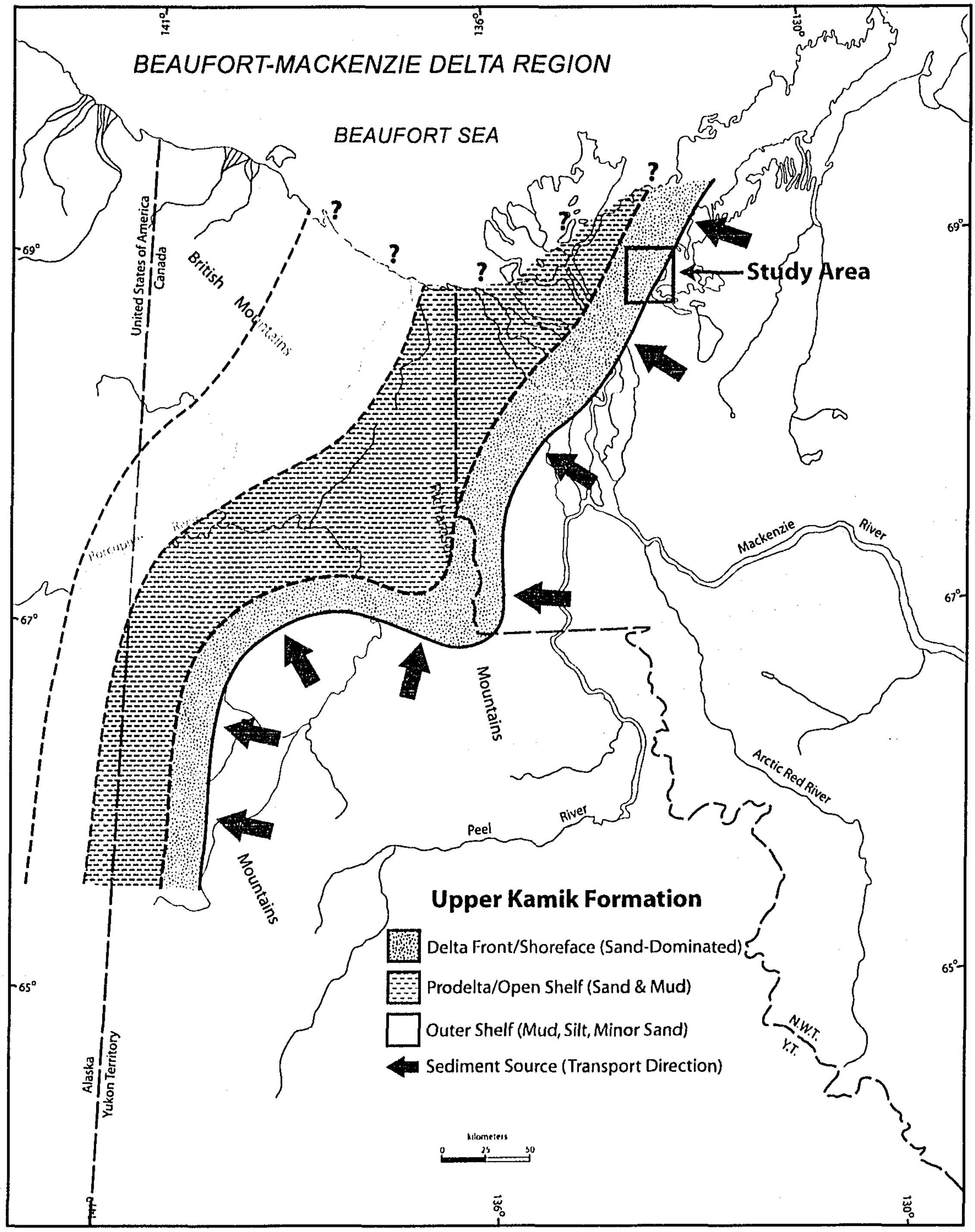


Figure 1.9 Paleogeographic map of middle Hauterivian time during deposition of the upper Kamik Formation in the Beaufort-Mackenzie Delta region. The inferred directions of sediment transport is indicated by the red arrows. Note the geographic location of the study area relative to inferred paleoshoreline. No palinspastic reconstruction is attempted (modified from Dixon, 1991).

the Parsons Lake area, thickening of Lower Cretaceous strata is attributed to syndepositional movement related to fault movement along the Eskimo Lakes Fault Zone. This suggests that there was growth faulting along the margins of the Eskimo Lakes Arch and that the arch was a major source of the coarse clastics comprising the Kamik Formation and that fault movement was long-lived and of significant magnitude (Dixon, 1982a). In addition, it was suggested by Dixon (1982a) that throughout Kamik deposition fault movement declined in frequency and magnitude, which resulted in reduced and intermittent sediment supply along with down-faulting and overall regional subsidence. The author believes that prior to and during Kamik deposition, active tectonic uplift and erosion of early Mesozoic strata from the Eskimo Lakes Arch initiated the erosion and deposition of large amounts sediment into the pericratonic Brooks-Mackenzie Basin.

During the Early Cretaceous, the Kamik Formation was deposited within the early to middle Mesozoic-aged Brooks-Mackenzie Basin. The term Brooks-Mackenzie Basin was first introduced into the literature by Balkwill *et al.* (1983) for a depositional basin in Arctic North America that was active between the Late Paleozoic to middle Early Cretaceous. This basin name was created as an encompassing term to illustrate the relationship and similarity of the stratigraphy, depositional history and tectonic evolution of the northern Richardson Mountains, northern Yukon and northern Alaska (Balkwill *et al.*, 1983; Dixon *et al.*, 1992). The term "Brooks Geosyncline" was generally used by previous authors for describing the geology of northern Alaska and northern Yukon, and the name "Beaufort-Mackenzie Basin" (Lerand, 1973; Yorath, 1973; Yorath and Norris, 1975; Young *et al.*, 1976; Hea *et al.*, 1980) in reference to the Mackenzie Delta, Tuktoyaktuk Peninsula, northern Richardson Mountains, Yukon Coastal Plain and adjacent offshore region beneath the Beaufort shelf region (Balkwill *et al.*, 1983).

The Brooks-Mackenzie Basin along with the Sverdrup Basin of the Canadian Arctic Archipelago were recognized as the two Late Paleozoic to middle Early Cretaceous depositional centers of the Arctic North America (Balkwill *et al.*, 1983). It was also recognized by Balkwill *et al.* (1983) that these basins represented separate depositional centers, but similarities in marine faunas indicated a connection between the Beaufort-Mackenzie and Sverdrup Basin since at least the Late Jurassic. For the purposes of this thesis, the author chose to retain the term Brooks-Mackenzie Basin in reference to the depositional basin active in the Mackenzie Delta region during deposition of the Kamik Formation. However, much of the previous research completed by Canadian workers (Lerand 1973; Yorath 1973; Yorath and Norris, 1975; Young *et al.*, 1976; Hea *et al.*, 1980) referred to the Lower Cretaceous strata in the Canadian portion of the Brooks-Mackenzie Basin as part of the Beaufort-Mackenzie Basin. The term Beaufort-Mackenzie Basin has now become relatively standard terminology used by authors (Dixon *et al.*, 1992) in reference to the Tertiary-aged depositional basin present in the Mackenzie Delta region.

The Brooks-Mackenzie Basin was interpreted to have formed as a structural depression resulting from extensional and strike-slip tectonics near the continental margin. Young *et al.* (1976)

interpreted that deposition in the Brooks-Mackenzie Basin began during the Late Triassic period and the basin became a major sedimentary depocenter in the Early Jurassic to Early Cretaceous. Lerand (1973) first subdivided the Phanerozoic stratigraphic succession of Arctic North America into three regional megasequences based on the identification of several major interregional unconformities, which represented tectono-stratigraphic boundaries. These tectono-stratigraphic boundaries were defined by Lerand (1973) in northwest Canada and adjacent Alaska to identify depositional basins based on their stratigraphic limits and tectonic history (Dixon *et al.*, 1992). The sequences identified include the Franklinian (Upper Cambrian to Devonian), Ellesmerian (Mississippian to Jurassic) and Brookian sequences (Cretaceous to Recent) (Lerand, 1973). These sequences were created to illustrate that the stratigraphy and tectonics within these two basins during specific time periods were part of a common tectono-stratigraphic assemblage. Balkwill *et al.* (1983) modified the stratigraphic limits of Lerand's (1973) tectono-stratigraphic assemblages and included the Mississippian to mid-early Cretaceous succession in the Ellesmerian regime and the mid-early Cretaceous to Tertiary in the Brookian regime. Balkwill *et al.* (1983) restricted strata above a major late Hauterivian unconformity (at the base of the Mount Goodenough Formation) to the Beaufort-Mackenzie Basin, and the Carboniferous to middle Hauterivian strata below to the Brooks-Mackenzie Basin (Dixon *et al.*, 1992). Dixon *et al.* (1992) noted that the late Hauterivian interregional unconformity has been used by several Canadian workers to separate two major tectono-stratigraphic assemblages; the Carboniferous to middle Hauterivian Ellesmerian sequence and the younger late Hauterivian to Holocene Brookian sequence (Lerand, 1973; Norris and Yorath, 1981; Balkwill *et al.*, 1983; Norris, 1985a).

This regional unconformity formed following deposition of the Kamik Formation but prior to deposition of the Mount Goodenough Formation and marks a period of regional uplift and erosion of Kamik Formation and Parsons Group sediments in the Mackenzie Delta region. The period of uplift is coincident with a change in regional tectonic regime that was likely related to the opening of the Canada Basin to the north. Lower Cretaceous sediments were included in the Ellesmerian Sequence by Balkwill *et al.* (1983) in order to distinguish the Mount Goodenough and younger Cretaceous strata from the underlying Kamik Formation and older Mesozoic strata (Balkwill pers. comm., 2004). According to Balkwill *et al.*'s (1983) definition the Kamik Formation is part of the Mississippian to Early Cretaceous-aged (late Hauterivian) Ellesmerian Sequence of the Brooks-Mackenzie Basin. The Ellesmerian sequence corresponds to a tectonic period dominated by extensional faulting and a basin fill consisting mainly of shallow water clastic deposits derived from the North American craton to the south and southeast (Lerand, 1973).

In the literature there are currently two opposing viewpoints on Jurassic to Early Cretaceous paleogeographic reconstructions. Early work by Jeletzky (1971, 1972, 1974, 1975) interpreted the presence of a north to south oriented depositional and bathymetric trough termed the Porcupine Plains-Richardson Mountains Trough (Dixon, 1991). This trough was interpreted to extend from the Rapid Depression (*sensu* Blow Trough) located between the present-day Richardson and Barn

Mountains, southward through the Eagle Plain. Jeletzky's Porcupine Plains-Richardson Mountains trough separated an eastern source terrain, called the "Peel Landmass" from a western source, known as the "Old Crow-Keele Landmass" (Dixon, 1991). According to Jeletzky's model, sediment was supplied to the trough from both landmasses. Lerand (1973) based his paleogeographic reconstructions of the Brooks-Mackenzie Delta region model and many conclusions on Jeletzky's work.

Recent work by Young (1975), Young *et al.* (1976), Poulton (1982, 1984) and Dixon (1982a, 1982b 1986, 1991) presented a different paleogeographic model from Jeletzky's, and this model is the basis for the paleogeographic interpretations presented within this thesis. This paleogeographic model identified a Jurassic and Lower Cretaceous shoreline that trended southwest to northeast and extended from the Tuktoyaktuk Peninsula through the northern Eagle Plain and then turned toward the south at the northern end of the Ogilvie Mountains (Figs. 1.8; 1.9). This paleogeographic reconstruction differs from Jeletzky's model in that it does not recognize a bathymetric trough through the Eagle Plain and did not clearly identify a western source terrain. The Old Crow-Keele Landmass was included in some of Young *et al.*'s (1976) paleogeographic reconstructions, however, Dixon (1991) noted that definitive stratigraphic and sedimentological evidence has not been found for the Old Crow-Keele Landmass. According to their paleogeographic interpretations, Young *et al.* (1976) implied that the Peel Landmass was the predominant source area. Poulton (1982) reviewed the arguments on the existence of a western source area during Early to Middle Jurassic time and concluded that only an eastern source terrain existed. Based on evidence from recent exploration in the Beaufort-Mackenzie Delta region the author agrees with the paleogeographic model presented by Young (1975), Young *et al.* (1976), Poulton (1982, 1984) and Dixon (1982a, 1982b, 1986, 1991) for a southwest to northeast trending shoreline that was sourced from an eastern source area.

Currently there exists controversy regarding the paleographic setting of Alaska during the early Mesozoic. Some authors, including Tailleur, (1973), Tailleur and Brosge (1970), Newman *et al.* (1977), Sweeny *et al.* (1978) and Grantz *et al.* (1990) have proposed that there is evidence for a counter-clockwise rotation of Alaska away from northern Canada beginning in the Early Jurassic. The rotational tectonic model is still actively debated and recent papers by Lane (1997, 1998) provide strong arguments against the rotational model. The present controversy regarding the rotational model is beyond the scope and focus of this study and the reader is referred to papers by the aforementioned authors for additional information.

1.10 Tectonic Elements and Structural Setting

The Kamik Formation was deposited on the structurally active southeastern, rifted-margin of the Brooks-Mackenzie Basin during the Early Cretaceous period. The Brooks-Mackenzie Basin began forming in the Mesozoic due to active rifting and regional extension associated with the opening of the Arctic Ocean to the north. A period of rifting initiated possibly as early as the Early Jurassic, and continued into the Early Cretaceous and was centered along an interpreted southwest to northeast spreading center (G. Prost pers. comm., 2004). Deposition of the Kamik Formation was initiated due to a major period of structural activity in the Brooks-Mackenzie Basin, which spanned the Late Jurassic to the earliest Late Cretaceous. This period of tectonic activity resulted in normal faulting and associated folding throughout the Beaufort-Mackenzie Delta region (Dixon, 1986). This tectonic activity was responsible for creating numerous uplifts, structural depressions and fault zones throughout the region including the Eskimo Lakes Arch, Eskimo Lakes Fault Zone, Cache Creek Uplift and Tununuk High (Fig. 1.10). In addition, there is evidence for strike-slip faulting along some faults in the Kaltag-Rapid Fault Array (Young *et al.,* 1976; Lane, 1988) in the Rapid Depression and Keele Block associated with the period of regional extension in the Brooks-Mackenzie Basin. The main tectonic elements interpreted to have influenced deposition of the Kamik Formation during Lower Cretaceous time include the Eskimo Lakes Arch, Eskimo Lakes Fault Zone, Cache Creek Uplift, Tununuk High, Rat Uplift, Napoiak High, Rapid Depression and tectonically stable North American craton (Fig. 1.11) (Dixon, 1982b, 1991). Most of these tectonic elements are interpreted as Mesozoic features that may have had much older precursor elements. Dixon (1992) noted that the present expression of these elements is mostly the result of Jurassic to Early Cretaceous tectonic activity with rejuvenation and modification in Tertiary time. The main structural elements affecting Kamik Formation deposition are illustrated in Figure 1.11 and are discussed and summarized below.

Structural activity in the southeastern portion of the Brooks-Mackenzie Basin was centered within a zone of large, down-to-the-basement normal faults referred to as the Eskimo Lakes Fault Zone (Coté *et al.,* 1975; Young *et al.,* 1976) or also known as the Tuktoyaktuk Fault Flexure Zone (Lerand, 1973). Structural activity in the Eskimo Lakes Fault Zone was dominated by vertical movement along normal faults, growth faults and minor folding (Lerand, 1973; Dixon, 1982a, 1982b). The northeastern part of the Aklavik Arch Complex (Eskimo Lakes Arch) became the faulted continental margin following the opening of the Canada Basin (Dixon, 1991). Depositional thinning of Lower Cretaceous strata toward the structural highs and local thinning and truncation within individual fault blocks was interpreted by Dixon (1991) to indicate that there was syndepositional movement along major faults zones, especially the Eskimo Lakes Fault Zone during deposition of the Kamik Formation. Coté *et al.* (1975) also identified growth faulting and minor folding associated with the Eskimo Lakes Fault Zone. This growth faulting was interpreted by Coté *et al.* (1975) to have resulted in the development of syndepositional folds adjacent to the Eskimo Lakes Fault Zone and the development of the structure at Parsons Lake. Lane (2002) noted that normal faults were active in the Brooks-Mackenzie Basin from at least the Late Jurassic to Middle Cretaceous time, which accommodated northwestward extension leading to the formation of the continental margin in post-Albian time. These normal faults are best developed beneath the Mackenzie Delta and Tuktoyaktuk Peninsula and trap significant oil and gas accumulations including the hydrocarbons

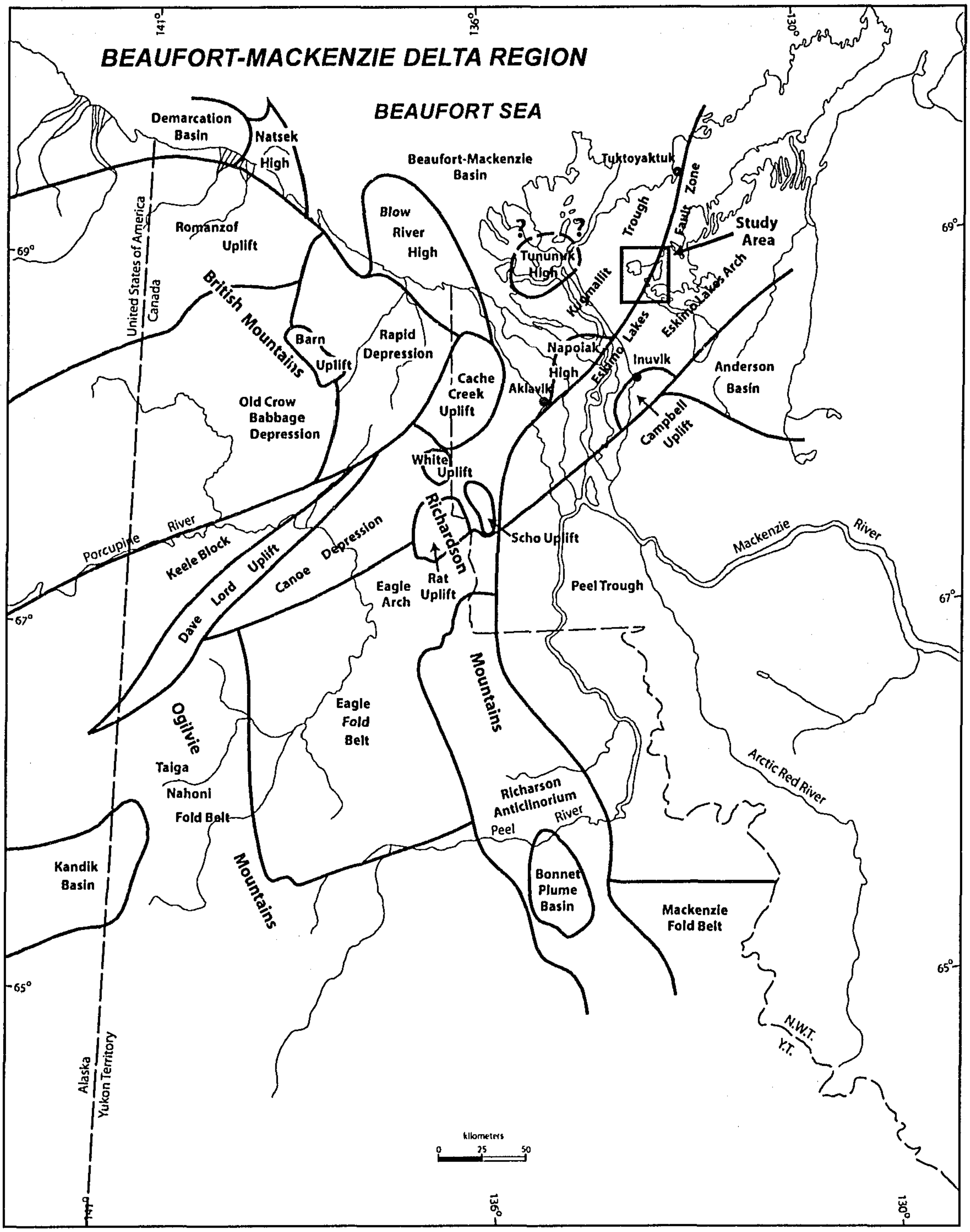


Figure 1.10 Regional map of the major tectonic elements of the northern Yukon and Northwest Territories. Note the location of the study area indicated by a black box (modified after Norris, 1983; Yorath and Cook, 1981; Dixon *et al.,* 1985 and Dixon, 1991).

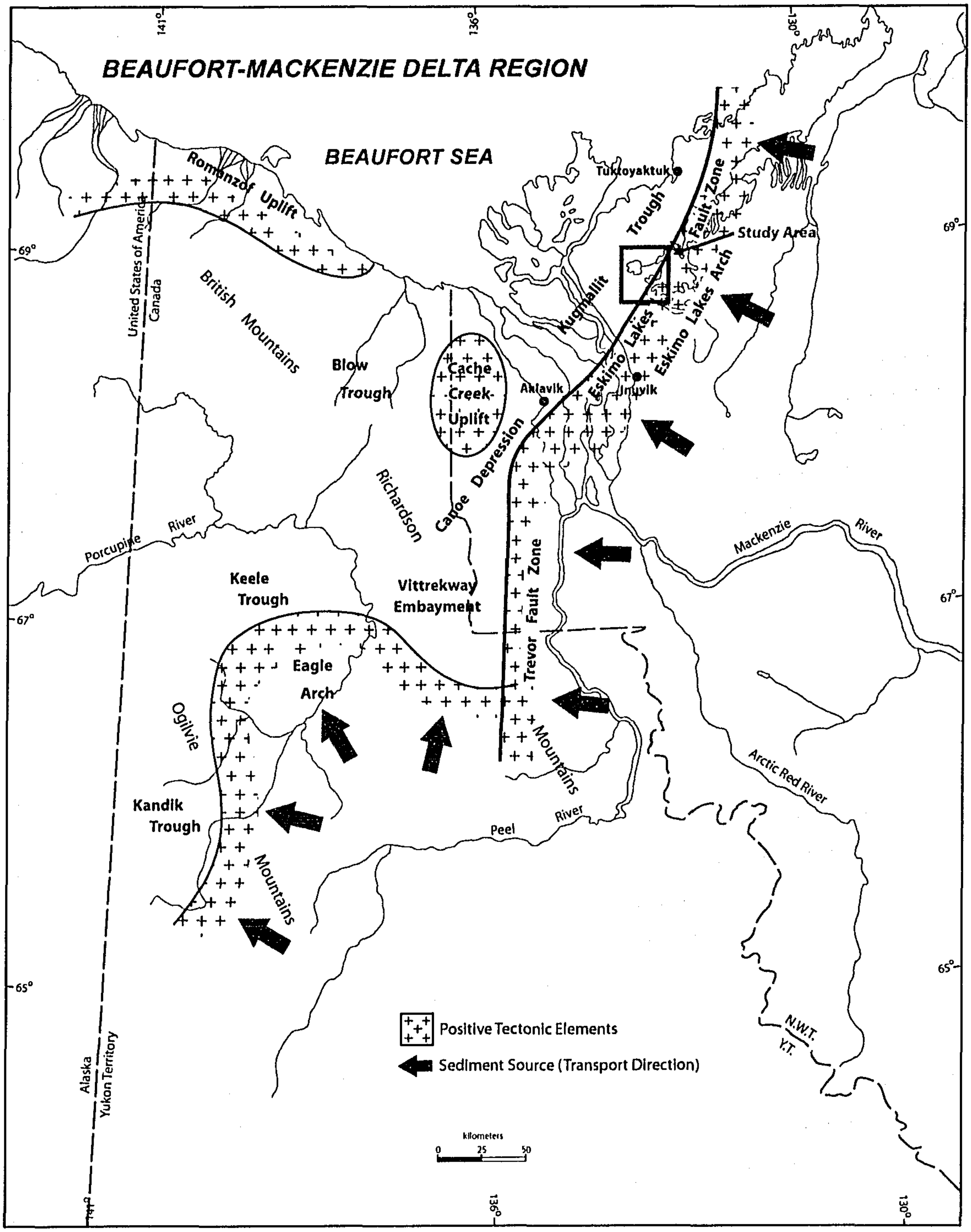


Figure 1.11 Lower Cretaceous (Neocomian) tectonic elements of the northern Yukon and Northwest Territories expressed as relative highs and lows based on depositional and erosional stratigraphic thickening and thinning trends. Note the location of the study area indicated in red (modified after Young *et al.*, 1976 and Dixon, 1991).

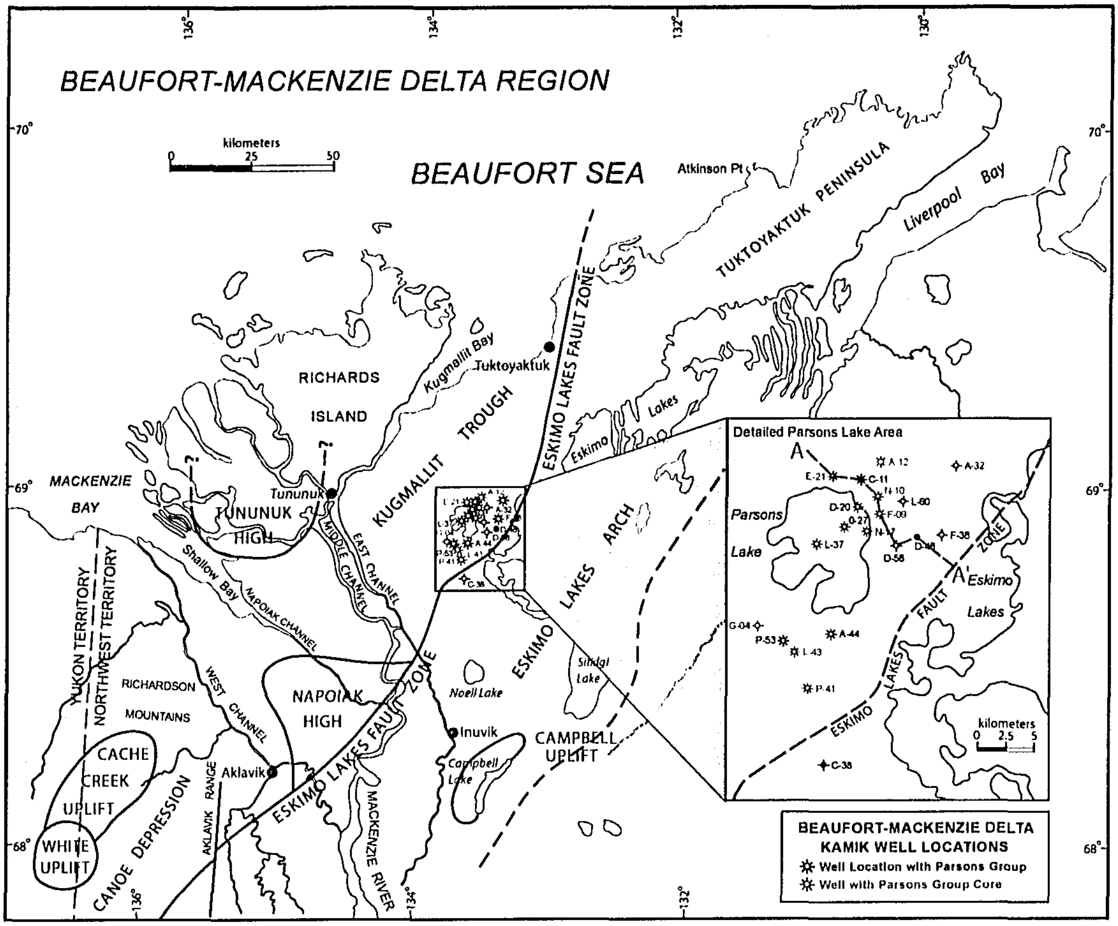


Figure 1.12 Structural map of the major tectonic elements influencing Kamik Formation deposition during the Lower Cretaceous within the Beaufort-Mackenzie Basin (modified from Dixon, 1982a).

at Parsons Lake gas field. The structural elements of the Brooks-Mackenzie Basin have been discussed by numerous authors including: Jeletzky (1961a, 1961b), Norris (1972), Lerand (1973), Tailleur (1973), Coté *et al.* (1975), Yorath and Norris (1975), Young *et al.* (1976), Lane (1988) and Lane (1997). For the purposes of this study, only the structural elements deemed pertinent to the Early Cretaceous are discussed below. For a complete summary of the tectonic history of the region the reader is directed to the references above.

Eskimo Lakes Arch

The Eskimo Lakes Arch is a relatively narrow, elongate, northeast-trending, antiformal structure that formed the southeastern margin of the Brooks-Mackenzie Basin and currently forms the southeast flank of the younger Beaufort-Mackenzie Basin (Figs. 1.11; 1.12). The Eskimo Lakes Arch structure was first recognized in the northern Richardson Mountains and named the Aklavik Arch by Jeletzky (1961b). The term Kugaluk Arch was also used by Yorath (1973) to describe a positive gravity trend along the Tuktoyaktuk Peninsula. The term Kugaluk Arch was later replaced by the more commonly used term Eskimo Lakes Arch by Young *et al.* (1976). The northwestern margin of the Eskimo Lakes Arch is bound by normal faults of a major fault zone referred to as the Tuktoyaktuk Fault Flexure Zone (Lerand, 1973) or the Eskimo Lakes Fault Zone (Coté *et al.*, 1975; Young *et al.*, 1976). During the Early Cretaceous, the uplifted Eskimo Lakes Arch is interpreted to have provided the main eastern and southeastern source of clastic material for the Kamik Formation. The Eskimo Lakes Arch was originally intended to have been a Columbian (Jurassic to Early Cretaceous) element and the Eskimo Lakes Fault Zone as its Laramide (Late Cretaceous to Early Cretaceous) expression. However, the common usage of these terms has not followed the original intentions of Young *et al.* (1976) as these terms are now used in reference to the present-day expressions of these structures (Dixon, 1991).

The Eskimo Lakes Arch was originally the northeasternmost element of the Aklavik Arch Complex of Norris (1974) and Yorath and Norris (1975). The Aklavik Arch Complex was defined as a composite, northeast-trending tectonic element extending 600 km in Canada from the Keele Range adjacent to the Alaskan border onto the Beaufort Shelf northeast of the Tuktoyaktuk Peninsula (Norris, 1974). The structurally positive components of the Aklavik Arch include the Dave Lord, Cache Creek, Rat and Campbell Uplifts (Yorath and Norris, 1975). Myhr and Young (1975) interpreted the Eskimo Lakes Arch, Cache Creek Uplift, and Rat Uplift as the main components of the Aklavik Arch Complex. The arch complex was interpreted by Lerand (1973) and Yorath and Norris (1975) as a complex structure that lacked a single, simple, well-defined crest and was characterized by culminations and depressions and by small *en echelon* normal faults along its axis. The arch complex is essentially a breached anticline on which Cretaceous strata unconformably overlie truncated sediments of Precambrian to Devonian age (Lerand, 1973). The Eskimo Lakes Arch plunges to the northeast, cuts obliquely across the Tuktoyaktuk Peninsula and disappears approximately near the tip of the Tuktoyaktuk Peninsula (Lerand, 1973).

The arch is composed of rocks ranging in age from Proterozoic to late Tertiary (Yorath and Norris, 1975; Young *et al.*, 1976). Under the Tuktoyaktuk Peninsula, the Eskimo Lakes Arch is cored by Proterozoic quartzites and dolomites and is flanked by Lower Paleozoic carbonates and clastics (Hea *et. al.*, 1980). The arch is composed of a thin cover of Mesozoic to Cenozoic rocks and a thick core of lower Proterozoic and Paleozoic rocks that outcrop within the Campbell Uplift. The Eskimo Lakes Arch initially showed a major expression following the Late Devonian, when it became a positive feature and remained above sea-level until the Late Cretaceous. The Eskimo Lakes Arch was a structurally active feature and experienced many vertical movements and uplifts along major faults extending from the Late Paleozoic until the Late Cretaceous (Lerand, 1973). Jeletzky (1975) interpreted the Aklavik Arch to have been uplifted and flexed in late Early to early Middle Jurassic and again in Valanginian to mid-Hauterivian time. During the Early Cretaceous, the Eskimo Lakes Arch existed as a positive structural element and represented a major sediment source for the Kamik Formation.

Eskimo Lakes Fault Zone

The Eskimo Lakes Fault Zone was originally referred to as the Tuktoyaktuk Fault Flexure Zone by Lerand (1973) as a strongly faulted hinge zone on the northwestern margin of the Aklavik Arch (*sensu* Eskimo Lakes Arch) (Figs. 1.11; 1.12). The aforementioned Kugaluk Arch of Yorath (1973) was also used to describe the Eskimo Lakes Arch and the assemblage of complex faults now known as the Eskimo Lakes Fault Zone (ELFZ). The term Eskimo Lakes Fault Zone was first used by Coté *et al.* (1975) and subsequently by Young *et al.* (1976) to replace Lerand's (1973) term Tuktoyaktuk Fault Flexure Zone. The Eskimo Lakes Fault Zone is composed of a series of down-to-the-basement normal faults and represents the southeastern basin bounding fault margin of the Brooks-Mackenzie Basin (Dixon 1982a). The faults comprising the ELFZ mark the transition from the Eskimo Lakes Arch to the Kugmallit Trough and Canoe Depression (Lerand 1973; Dixon 1982a). The fault movement of the Eskimo Lakes Fault Zone ranges in age from Paleozoic to Recent and is characterized by throws up to 1600 m (Dixon, 1982a). Cook *et al.* (1987a, 1987b) noted that deep-reflection seismic data across the Eskimo Lakes Fault Zone show that the faults are listric, soling-out at considerable depths within strata that are presumed to be Proterozoic in age. Across this fault zone, Mesozoic and Cenozoic strata rapidly thicken and reach thicknesses of approximately 10,000 to 12,000 m (Coté *et al.*, 1975).

During deposition of the Kamik Formation structural activity in the southeastern part of the Brooks-Mackenzie Basin was dominated by vertical fault movements and minor folding along the Eskimo Lakes Fault Zone (Dixon, 1982b). Throughout the Cretaceous and Tertiary, sediment deposition along the Eskimo Lakes Fault Zone was associated with syndepositional movement and active growth faulting (Lerand, 1973). The syndepositional growth faulting along the Eskimo Lakes Fault Zone initiated the development of folds adjacent to the fault zone and resulted in the development of the structure at Parsons Lake.

Interior Platform

The ancient Interior Platform was originally called the Peel Landmass and was first identified and described by Jeletzky (1961b) as an extensive, region of the northern Interior Platform exposed throughout much of Jurassic and Early Cretaceous time. Jeletzky (1975) interpreted the Peel Landmass as the eastern bounding margin of his interpreted Porcupine Plains-Richardson Mountains Trough. During the Early Cretaceous, this landmass was flanked on the west by a belt of arenaceous to argillaceous marine to nonmarine sediments that were primarily derived from the eastern exposed portion of the landmass (Jeletzky, 1975). Poulton *et al.* (1982) interpreted this feature as the unstable, northeasterly trending margin of the North American craton. The western margin of the Interior Platform comprises a series of uplifts and depressions including part of the Eskimo Lakes Arch, Eskimo Lakes Fault Zone, Cache Creek Uplift, Rat High and Campbell Uplift.

These structural features, in addition with the Kugmallit and Blow Troughs, constituted the Aklavik Arch Complex (Young *et al.*, 1976; Yorath and Norris, 1975).

Cache Creek Uplift

The Cache Creek Uplift is a linear, northeast to southwest trending structural high that straddles the boundary between the Yukon Territory and Northwest Territories on the west flank of the Mackenzie Delta (Figs. 1.11; 1.12) (Young *et al.*, 1976; Norris, 1997). The Cache Creek Uplift is cored by Ordovician and Silurian-aged rocks and is bounded on the northwest by the Eagle Fault (Norris, 1977) and on the southwest by the Cache Creek Fault. The Cache Creek Uplift was identified as a principal component of a right-hand *en echelon* array of uplifted and depressed segments comprising the Aklavik Arch complex (Yorath and Norris, 1975; Young *et al.*, 1976). Jeletzky (1975) interpreted the Cache Creek Uplift as a probable northward offshoot of the Aklavik Arch.

The Cache Creek Uplift is interpreted to have been tectonically active since at least the Paleozoic and was most influential on deposition of sediments during the Late Jurassic to Cretaceous. However, Jeletzky (1975) interpreted the Cache Creek Uplift and Canoe Depression to have been active only during the Valanginian to mid-Hauterivian. The uplift was responsible for separating the Brooks-Mackenzie Basin into two elements: the Blow Trough (Young *et al.*, 1976) or the Rapid Depression (Norris, 1972) to the west, and the Kugmallit Trough (Young *et al.*, 1976) and Canoe Depression (Norris, 1972) to the east. The Cache Creek Uplift is interpreted to have been an emergent positive tectonic feature during deposition of the Kamik Formation.

Tununuk High

The Tununuk High was originally introduced by Lerand (1973) for a positive gravity anomaly that coincided with a broad, domed structure occurring in what is now the southern part of the Richards Island in the modern Mackenzie Delta (Fig. 1.12). The Tununuk High was identified by Lerand (1973) as representing the northwest flank of the Kugmallit Trough. Although, the high was originally defined on the basis of its positive gravity expression, Dixon (1982a) identified that some Lower Cretaceous units thin or are truncated on the high.

Young *et al.* (1976) extended the Columbian Cache Creek High northeastward to incorporate the Tununuk High. However, isopach maps and gravity data did not support Young *et al's*. (1976) extension of the Cache Creek High due to the recognition of a depression with a more complete Mesozoic succession that separates the two structural elements (Dixon, 1982a). The Tununuk High was a minor, localized influence during deposition of the Kamik Formation.

Kugmallit Trough

The Kugmallit Trough is a large structural depression located on the northwest side of the Eskimo Lakes Arch. During the Jurassic to Early Cretaceous, this structure was manifested as a half-graben with its downthrown side to the northwest (Figs. 1.11; 1.12). The Kugmallit Trough mainly influenced deposition of the Jurassic to Early Cretaceous-aged rocks; however, it did continue to influence Tertiary deposition through differential subsidence (Dixon, 1992). The Kugmallit Trough trends northeast to southwest, approximately parallel the Eskimo Lake Fault Zone. In the subsurface of the Mackenzie Delta the northwest flank of the trough rises into the Tununuk High (Lerand, 1973). Mesozoic strata including the Kamik Formation are thickest and most well preserved within the Kugmallit Trough.

Norris (1972) originally named this feature the Canoe Depression, and Young *et al.* (1976) renamed it the Kugmallit Trough. Young *et al.* (1976) interpreted the Kugmallit Trough as a Jurassic Columbian tectonic element and expanded its margins to the northeast into the region previously occupied by the Canoe Depression of Norris (1972). Dixon (1982a) identified a small, relative positive trend named the Napoiak High, which partially separated this structural depression into the Canoe Depression to the southwest and the Kugmallit Trough in the northeast. The Napoiak High has not been recognized as strong influence on the deposition of the Kamik Formation. The Kugmallit Trough is interpreted to have limited influence during Kamik Formation deposition. Its primary expression was during the late Early Cretaceous (post-Kamik time), when it represented a major depocenter for Mesozoic strata. Sykora (1984) interpreted the western margin of the Kugmallit Trough as mostly uplifted during the Late Cretaceous to Early Tertiary Laramide orogeny in the Richardson Mountains.

1.11 Parsons Lake Gas Field

The Beaufort-Mackenzie Delta region of the Northwest Territories is a major petroleum province that contains a thick sedimentary column host to several major hydrocarbon deposits. The region is estimated to contain recoverable reserves of approximately 1.5 to 2.0 billion barrels of oil and 12 trillion cubic feet of natural gas in 53 significant discoveries (Dixon *et al.*, 1994). The Lower Cretaceous Kamik Formation is a major hydrocarbon-bearing clastic succession and is host to significant reserves of natural gas and lesser amounts of oil in the subsurface of the Beaufort-Mackenzie Delta region. The onshore, giant-class Parsons Lake gas field contains approximately 2.8 to 4.5 trillion cubic feet (TCF) of natural gas-in-place and represents the largest discovery to date from Lower Cretaceous strata in the region (Prost *et al.*, 2005). The Parsons Lake gas field is located on the southeastern margin of the modern Beaufort-Mackenzie Basin, within a zone of large down-to-the-basement normal faults that comprise the flank of the Eskimo Lakes Fault Zone (Fig. 1.13) (Coté *et al.*, 1975).

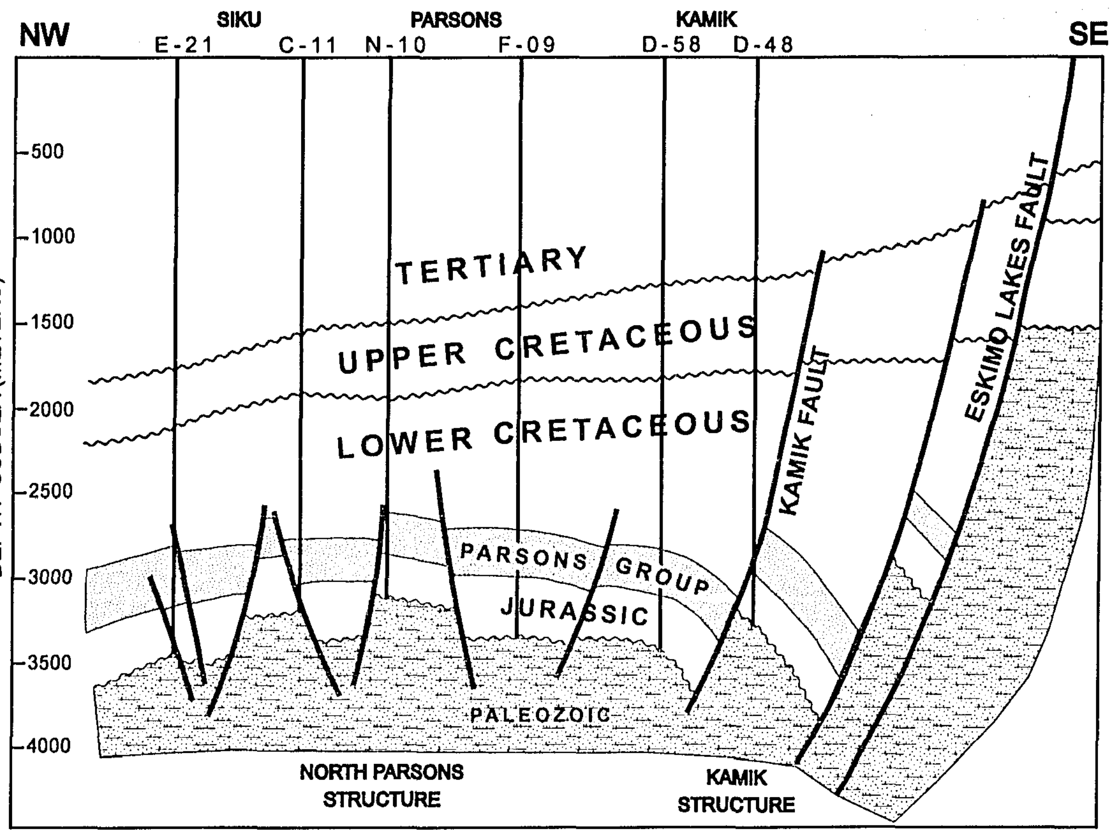


Figure 1.13 Schematic structural cross-section through the north pool of the Parsons Lake gas field. Note the location of exploratory wells, stratigraphic units, interpreted faults and geographic names. The line of section is shown in Figure 1.12 (modified from a Gulf Internal Report).

The Parsons Lake gas field was discovered by Gulf Canada Resources and Mobil Oil in the spring of 1972 by the drilling of the Gulf Mobil Parsons F-09 well on a seismic structure located beneath the Arctic Coastal Plain (Coté *et al.,* 1975). Following the discovery, an additional 17 exploratory wells were drilled in the Parsons Lake area resulting in the delineation of the extent and nature of hydrocarbons trapped in the Parsons Lake field. The Parsons Lake field consists of two gas pools, the north pool containing the majority of reserves and the smaller south pool containing the remaining reserves (Fig. 1.2). The gas reservoirs of the Kamik Formation at Parsons Lake are comprised of multiple, thick, delta front and distributary channel sandstones. The structure consists of a north and south dome, formed by growth faulting and folding related to syndepositional structural movement of the Eskimo Lakes Fault Zone. Hydrocarbons within the Parsons Lake gas field are structurally trapped within a rollover anticline on the downthrown side of the Eskimo Lakes Fault (Fig. 1.13). The field is highly compartmentalized by numerous small normal faults that trend both parallel and perpendicular to the main southwest to northeast trend of the Eskimo Lakes Fault Zone. The South Parsons pool contains gas and minor amounts of condensate, whereas the North Parsons pool contains gas, condensate and one oil well. The hydrocarbons of the Parsons Lake field were interpreted to have migrated from the deep, thermally mature Husky Formation, across a fault zone, into the adjacent Paleozoic carbonates under the Parsons structure and then upwards into the Kamik Formation reservoir (Langhus, 1980). The seal for the Parsons Lake gas field is the overlying thick marine shales of the Mount Goodenough Formation. For a further discussion of the petroleum geology of the Kamik Formation and Parsons Group in the Parsons Lake gas field and Beaufort-Mackenzie Delta region, the reader is referred to papers by Coté *et al.* (1975), Langhus (1980), Dixon (1983), Dixon *et al.* (1985, 1988) and Dixon (1996).

Chapter Two: Facies Descriptions and Interpretations

2.1 Introduction

Analysis of approximately 200 meters of core from fifteen cored intervals within the Kamik Formation (Table 2.1) resulted in the identification of fifteen facies (Facies A to Facies O) (Table 2.2). Usage of the term *facies* follows the definition given by Walker (1992, p. 2) as "a body of rock characterized by a particular combination of lithology, physical and biological structures that bestow an aspect different from the bodies of rock above, below or laterally adjacent". Within the Kamik Formation, facies are largely delineated on the basis of sedimentologic and ichnologic characteristics. Sedimentological analysis concentrated on the identification of lithology, grain size, sorting, bed thickness, nature of bedding contacts, primary physical sedimentary structures, penecontemporaneous deformation structures and accessory components. Ichnological criteria, which in many instances provided rationale for the subdivision of facies, included the identification of ichnogenera, along with their relative abundance, assemblage diversity and sizes, and the intensity of bioturbation. Twenty-three types of trace fossils were identified; their ethological classification, trophic strategy and possibly phylogeny are summarized in Table 2.3. A summary table of the grain size designations used within this study is presented in Table 2.4. Sedimentology and ichnology were ultimately integrated to characterize the paleoenvironmental conditions and depositional conditions present during deposition of the Kamik Formation. The facies identified from the Kamik Formation are presented in approximate interpreted depositional order within each facies association, starting with the distal facies and concluding with the proximal facies. The facies are arranged into two overall coarsening-upward successions and two fining-upward successions, followed by one irregular stacked facies succession exhibiting both coarsening- and fining-upward successions.

2.2 Facies Descriptions and Interpretations

Facies A - Thoroughly bioturbated mudstone, silty mudstone and very fine-grained silty sandstone

Lithology and Sedimentary Features:

Facies A consists of light to dark grey, organic mudstone and silty mudstone, interbedded with light grey to buff, upper very fine to lower fine-grained silty sandstone. The mudstone and silty mudstone beds are thoroughly bioturbated with primary sedimentary structures almost entirely obliterated (Figs. 2.1A-D). Wavy bedding is the dominant sedimentary structure within this facies with lesser amounts of horizontal to low angle parallel lamination. Mudstone is the dominant lithology, with beds ranging in thickness from less than 1 cm to approximately 15 cm. The intervening very fine-

Well Name	Cored Interval	Core Length
Gulf Mobil Siku A-12	2713.0m to 2732.2m (8901ft to 8964ft)	19.2 m
Gulf Mobil Siku A-12	2804.2m to 2823.1m (9200ft to 9262ft)	18.9 m
Gulf Mobil Siku E-21	2865.4m to 2867.6m (9401ft to 9408ft)	2.2 m
Gulf Mobil Siku E-21	2953.5m to 2971.8m (9690ft to 9750ft)	18.3 m
Gulf Mobil Siku E-21	3045.0m to 3055.9m (9990ft to 10026ft)	10.9 m
Gulf Mobil Parsons P-41	2961.0m to 2979.7m (9716ft to 9776ft)	18.7 m
Gulf Mobil Parsons F-09	2845.9m to 2864.2m (9337ft to 9397ft)	18.3 m
Gulf Mobil Parsons L-43	2772.5m to 2790.7m (9096ft to 9156ft)	18.2 m
Gulf Mobil Parsons L-43	2849.3m to 2858.4m (9348ft to 9378ft)	9.1 m
Gulf Mobil Parsons L-43	2932.8m to 2945.9m (9622ft to 9665ft)	13.1 m
Gulf Reindeer G-04	2920.0m to 2929.4m (9580ft to 9611ft)	9.4 m
Gulf Reindeer G-04	2930.7m to 2940.4m (9615ft to 9647ft)	9.7 m
Imperial Wagnark C-23	3780.7m to 3789.9m (12404ft to 12434ft)	9.2 m
Gulf Mobil Parsons N-10	2750.8m to 2765.1m (9025ft to 9072ft)	14.3 m
Gulf Mobil Parsons N-10	2795.5m to 2804.8m (9172ft to 9202ft)	9.3 m
	Total:	198.8 m

Table 2.1 List of cored intervals examined from the Kamik Formation in the study area.

Facies	Facies Descriptions	Physical Sedimentary Structures	Biogenic Sedimentary Structures	Depositional Environment
A	Thoroughly bioturbated mudstone, silty mudstone and very fine-grained silty sandstone	Faint horizontal to low angle (<5°) parallel lamination, wavy bedding, wave & combined flow ripples, distal HCS	He (a), Ps (a), Te (m), Ch (m-c), As (c), Sch (r-c), Dp (r), Pa (r), Pl (c)	Distal Prodelta (Lower Offshore)
B	Wavy to lenticular bedded, bioturbated, organic mudstone and fine-grained silty sandstone	Wavy to lenticular bedding, horizontal to low angle (<5°) parallel lamination, distal HCS, wave & combined flow ripples, synaeresis cracks	He (a), Ps (a), Te (a), Ch (a), Rhz (a-m), Pl (m), Pa (m), Dp (m), Sch (m), Ro (m), Sc (c), As (m-c), Co (r), Sk (r), Th (r), Zo (r), Ar (r), Fg (r)	Proximal Prodelta (Upper Offshore)
C	Interbedded, bioturbated, very fine to fine-grained, low angle parallel-laminated sandstone and bioturbated organic mudstone	Hummocky cross-stratification (HCS), swaley cross-stratification (SCS), wave & combined flow ripples, graded rhythmites, soft-sediment deformation, scour surfaces	He (a-m), Ps (a), Rhz (a-c), Ch (a-m), Pl (a-c), Pa (m), Oph (c), Ro (a-m), Te (m), Dp (c), Sch (c), Sp (r), Th (c-r), Ma (r), As (r), Sk (c), Sc (c-r), Cy (c), Co (r), Lo (r), Bg (r), Fg (r)	Distal Delta Front (Lower Shoreface)
D	Medium to high angle cross-stratified, sporadically burrowed, fine to medium-grained sandstone	Trough cross-bedding (TCS), planar tabular cross-bedding (~15° to 27°), wave ripples, massive bedding, graded rhythmites, ball & pillow structures, hyperpycnal muds, scours	Oph (a), Rhz (a-m), Co (a-m), Cy (m), Ma (m-c), Te (m-c), Fg (c)	Proximal Delta Front (Upper Shoreface)
E	Horizontal to low angle parallel-laminated fine to medium-grained sandstone	Horizontal to low angle (<5°) parallel lamination, wave & current ripple lamination	Ma (a), Oph (m-c) & Rootlets (c-r)	Proximal Delta Front (Foreshore)
F	Poorly-sorted, scour-based, matrix-supported conglomerate and medium to very coarse-grained sandstone	Poorly sorted, scoured bases, sandstone intraclasts, chert and quartz pebbles, mafic and metamorphic pebbles	No Trace Fossils were identified	Transgressive Lag
G	Medium to very coarse-grained, planar tabular and trough cross-bedded, salt and pepper textured sandstone and minor conglomerate	Planar tabular & trough cross-bedding (TCS), medium to high angle (~15° to 27°) cross-bedding, current ripples, pebble lags, mudstone rip-ups, coalified wood fragments	No Trace Fossils were identified	Distributary Channel (Meandering Fluvial Channel)
H	Poorly sorted, scour-based, medium to high angle cross-bedded, medium to very coarse-grained sandstone and conglomerate	Planar tabular & trough cross-bedding (TCS), medium to high angle (~15° to 32°) cross-bedding, current ripples, graded bedding, pebble lags, mudstone rip-ups, scour surfaces	No Trace Fossils were identified	Distributary Channel (Braided Fluvial Channel)
I	Wavy to lenticular bedded, bioturbated, organic mudstone, siltstone and fine-grained silty sandstone	Wavy & lenticular bedding, convolute bedding, low angle (<5°) parallel lamination, wave & current ripples, synaeresis cracks	Ar (a), Sk (m-c), Dp (m-r), Pl (c), Pa (c), Bg (r), Te (m), Lo (r), Fg (r) & Rt (c)	Interdistributary Bay (Bay Fill)

Table 2.2 Summary table of all facies identified in this study from the Kamik Formation, along with their names, basic descriptions, physical and biogenic sedimentary structures and interpreted environments of deposition.

Facies	Facies Descriptions	Physical Sedimentrary Structures	Biogenic Sedimentary Structures	Depositional Environment
J	Fine to medium-grained, low angle cross-bedded to flaser bedded sandstone and minor, dark grey siltstone and silty mudstone	Flaser bedding, low angle (<5°) parallel lamination, wave & current ripple lamination, climbing ripples, sigmoidal bedding, soft-sediment deformation, minor wavy bedding	Dp (m), Sk (m), Ar (c), Te (c-r), Pa (c-r), Pl (m), Th (c), Lo (r), Bg (r), Fg (r) & Rt (a)	Crevasse Splay (Interdistributary Bay Fill)
K	Well-burrowed, interbedded, fine-grained silty sandstone and silty mudstone with abundant siderite	Wavy bedding & flaser bedding, low angle (<5°) parallel lamination, burrow mottling, convolute & chaotic bedding, siderite	Te (a), Dp (a-m), Sk (m-c), Pl (c), Pa (c) & Rt (c-r)	Restricted Interdistributary Bay (Brackish Bay Fill)
L	Dark grey to black, coaly mudstone with rootlets	Faint to poorly laminated, massive to structureless beds, mottled bedding	Rt (a), *Glossifungites* (r)	Marsh (Lower Delta Plain)
M	Dark grey to black, waxy, carbonaceous mudstone	Waxy, chaotic bedding, massive to structureless, pyritized intervals	No Trace Fossils were identified	Paleosol (Lower Delta Plain)
N	Interbedded, unburrowed, low angle, parallel-laminated, soft-sediment deformed, fine-grained sandstone, siltstone and silty organic mudstone	Low angle (<5° to 10°) cross-bedding, soft-sediment deformation, convolute bedding, climbing ripples, wave ripples, ball & pillow structures, hypopycnal & hyperpycnal muds	No Trace Fossils were identified	Distal Delta Front (River-dominated Delta Front)
O	Medium-grained, unburrowed, low, medium to high angle cross-bedded sandstone	Planar tabular and trough cross-bedding (TCS), low, medium to high-angle (<5° to 32°) cross-bedding, wave & current ripple lamination, scour & fill structures, mudstone rip-ups	No Trace Fossils were identified	Proximal Delta Front (River-dominated Delta Front)

Table 2.2 (Continued) Summary table of facies identified in the Kamik Formation, along with their basic description, physical and biogenic sedimentary structures and interpreted environment of deposition.

Abbreviations: Ar - *Arenicolites*, **As** - *Asterosoma*, **Bg** - *Bergaueria*, **Ch** - *Chondrites*, **Co** - *Conichnus*, **Cy** - *Cylindrichnus*, **Dp** - *Diplocraterion*, **Fg** - Fugichnia (escape trace), **He** - *Helminthopsis*, **Lo** - *Lockeia*, **Ma** - *Macaronichnus*, **Oph** - *Ophiomorpha*, **Pa** - *Palaeophycus*, **Ps** - *Phycosiphon*, **Pl** - *Planolites*, **Rhz** - *Rhizocorallium*, **Ro** - *Rosselia*, **Rt** - Rootlet, **Sch** - *Schaubcylindrichnus*, **Sc** - *Scolicia*, **Sp** - *Siphonichnus*, **Sk** - *Skolithos*, **Te** - *Teichichnus*, **Th** - *Thalassinoides*, **Zo** - *Zoophycos*, **HCS** - Hummocky cross-stratification, **SCS** - Swaley cross-stratification, **TCS** - Trough cross-stratification, **a** - abundant, **m** - moderate, **c** - common, **r** - rare

Ichnofossil	Ethology	Trophic Group	Possible Tracemaker
Arenicolites	domichnia	suspension feeder	annelid or other worm-like phyla
Asterosoma	fodinichnia domichnia	deposit feeder grazing trace	annelid worm decapod crustacean
Bergaueria	cubichnia	passive carnivore	anemone
Chondrites	fodinichnia	deposit feeder	nematode; (?) sipunculid or polychaete or other worms
Conichnus	fodinichnia; cubichnia	passive carnivore	anemone
Cylindrichnus	domichnia	suspension feeder	polychaete or other worms
Diplocraterion	domichnia	suspension feeder	amphipod crustaceans; (?) polychaete worms
Escape Trace	fugichnia	locomotion trace	bivalves; worms
Helminthopsis	pascichnia	grazing trace	annelid, polychaete or other worm-like phyla
Lockeia	cubichnia	suspension feeder	bivalves
Macaronichnus	fodinichnia	deposit feeder	opheliid polychaetes (bloodworms)
Ophiomorpha	domichnia fodinichnia	suspension feeder	crustacean (thalassinid decapod)
Palaeophycus	domichnia	suspension feeder; carnivore (?)	annelid or polychaete worm
Phycosiphon	fodinichnia	deposit feeder	annelid or other worm-like phyla
Planolites	fodinichnia	deposit feeder	annelid or other worm-like phyla
Rhizocorallium	fodinichnia; domichnia	deposit feeder; suspension feeder	crustacean or annelid worm
Rosselia	fodinichnia	deposit feeder	annelid or other worm-like phyla
Schaubcylindrichnus	domichnia	suspension feeder	annelid, phoronid or other worm-like phyla
Scolicia	fodinichnia	deposit feeder	echinoids, gastropods
Siphonichnus	cubichnia	suspension feeder	bivalves
Skolithos	domichnia	suspension/deposit feeder; carnivore	annelid, phoronid or other worm-like phyla
Teichichnus	fodinichnia	deposit feeder	annelid, phoronid or other worm-like phyla
Thalassinoides	domichnia/ fodinichnia	deposit feeder	crustacean (thalassinid decapod)
Zoophycos	fodinichnia	grazing trace/ deposit feeder	annelid or other worm-like phyla

Table 2.3 Summary of trace fossils identified within the Kamik Formation, along with their names, ethological classification, trophic group and interpreted trace maker.

Can-Strat Terminology	Diameter (mm)	Phi (ø) size	
Cobble	64.0 - 256.0	-6.0 to -8.0	
Pebble			
Upper	16.0 - 64.0	-4.0 to -6.0	
Lower	4.00 - 16.0	-2.0 to -4.0	
Granule			Gravel
Upper	2.38 - 4.00	-1.5 to -2.0	
Lower	2.00 - 2.38	-1.0 to -1.5	
Very coarse-grained sand (vc)			
vcU	1.41 - 2.00	-0.5 to -1.0	
vcL	1.00 - 1.41	0.0 to -0.5	
Coarse-grained sand (c)			
cU	0.71 - 1.00	0.5 to 0.0	
cL	0.50 - 0.71	1.0 to 0.5	
Medium-grained sand (m)			
mU	0.35 - 0.50	1.5 to 1.0	Sand
mL	0.25 - 0.35	2.0 to 1.5	
Fine-grained sand (f)			
fU	0.177 - 0.250	2.5 to 2.0	
fL	0.125 - 0.177	3.0 to 2.5	
Very fine-grained sand (vf)			
vfU	0.088 - 0.125	3.5 to 3.0	
vfL	0.0625 - 0.088	4.0 to 3.5	
Silt			
coarse	0.0310 - 0.0625	5.0 to 4.0	
medium	0.0156 - 0.0310	6.0 to 5.0	
fine	0.0078 - 0.0156	7.0 to 6.0	Mud
very fine	0.0039 - 0.0078	8.0 to 7.0	
Clay	0.00006 - 0.0039	14.0 to 8.0	

U = Upper Size Limit L = Lower Size Limit

Table 2.4 Summary of grain size designations utilized in this study.

grained silty sandstone beds occur sporadically but are commonly isolated within the mudstone and silty mudstones. These silty sandstone beds are generally thin, ranging from less than 1 cm to approximately 2.5 cm. The thickness and abundance of individual sandstone beds generally increases upward toward the top of the interval. Some of the thicker beds display sharp, erosive bases with burrowed to diffuse tops, although the majority of the contacts are burrowed and appear gradational. Most of the thin sandstone beds are generally intensely bioturbated; thicker beds remain relatively unburrowed with remnant, undulatory, parallel laminae preserved. Where bedding has not been completely destroyed by bioturbation, the sandstone beds are characterized by horizontal to low angle (<5°) parallel lamination and symmetrical ripple lamination, although some beds appear structureless (Fig. 2.1D). Symmetrical and asymmetrical ripples occur isolated within the mudstone beds are common. Pyrite nodules averaging 1 cm in diameter are rare.

Facies A is stratigraphically poorly represented and is only preserved as a 36 cm thick interval in core #4 from the G-04 well. This facies is gradationally overlain by Facies B. The lower contact is not preserved in core however, it is assumed to be gradational with underlying finer-grained, distal facies. The contact between facies A and B is identified by a decrease in sediment homogenization and mudstone abundance, along with an increase in sandstone abundance and bed thickness. Facies A occurs at the base of a coarsening upward succession that is composed of Facies B, C, D and E. Due to the poor representation in core and limited penetration depths, the exact thicknesses cannot be determined and it is likely that thicker successions exist.

<u>Ichnological Characteristics:</u>

This facies displays a moderate to high abundance of burrows with an overall low to moderate diversity of trace fossils. There is a notable difference in the trace fossil assemblage within the mudstone and silty mudstone, versus the silty sandstone. The mudstone interbeds contain *Helminthopsis* (Figs. 2.1B,C), *Chondrites* (Fig. 2.1D), *Planolites*, and rare *Thalassinoides*. The sandstone intervals contain *Phycosiphon* (Figs. 2.1A-D), *Asterosoma* (Fig. 2.1A), *Schaubcylindrichnus* (Figs. 2.1A,C), *Diplocraterion* (Fig. 2.1C), *Teichichnus* (Figs. 2.1A,C,D) and *Palaeophycus*. Overall, *Helminthopsis* and *Phycosiphon* are the dominant ichnogenera in Facies A and comprise the majority of the burrow-mottled texture. *Phycosiphon* is especially abundant within sandstone beds, with most of the burrowing restricted to the uppermost portions of the beds. *Schaubcylindrichnus* is associated with the mudstone and silty mudstone beds and is commonly thin-walled (Figs. 2.1A,C). In addition, the majority of ichnogenera present in this facies are diminutive in size.

<u>Interpretation and Discussion:</u> Distal Prodelta (Lower Offshore)

Thoroughly bioturbated mudstone, silty mudstone and very fine-grained silty sandstone beds of

Figure 2.1

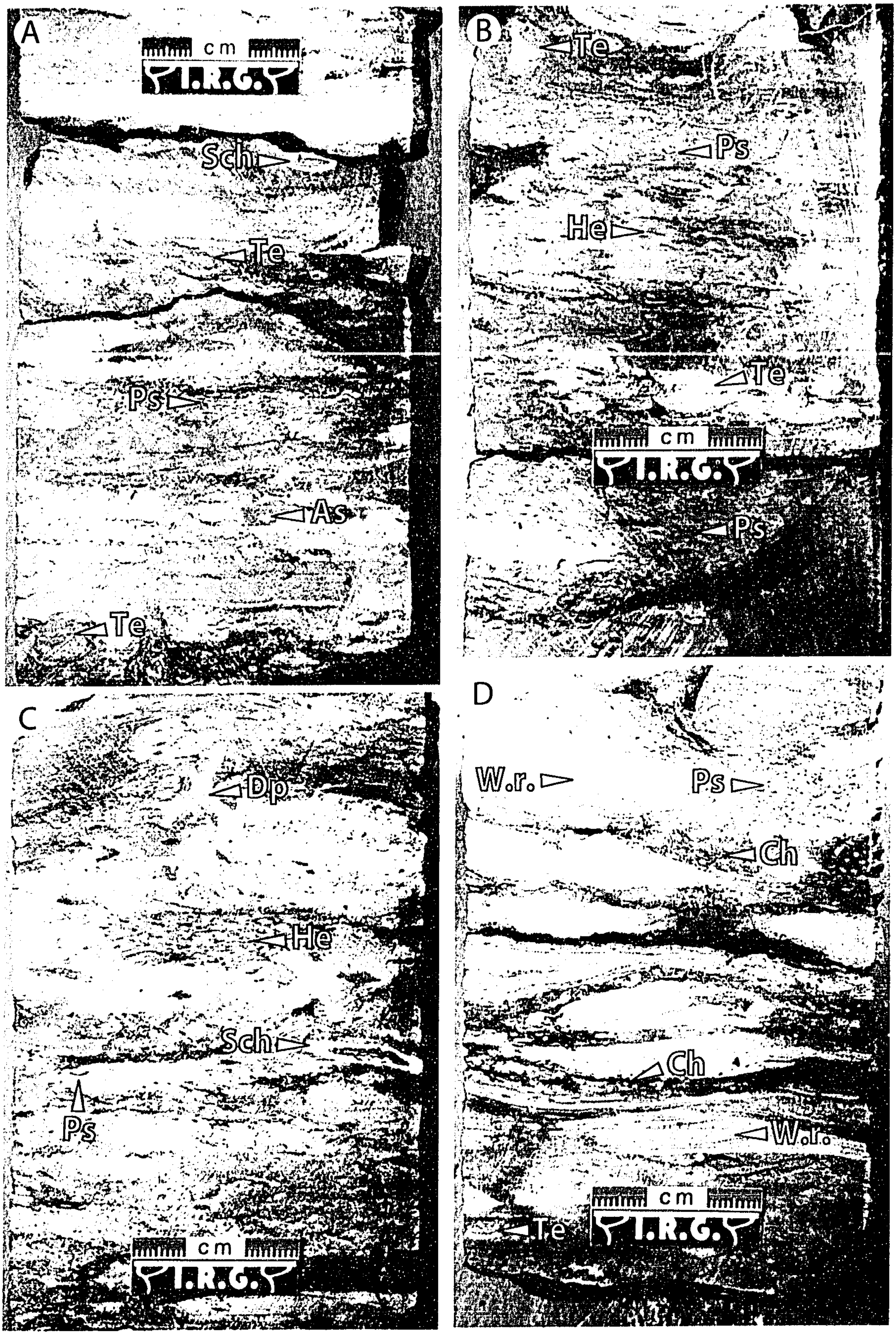
Facies A - Thoroughly bioturbated mudstone, silty mudstone and very fine-grained silty sandstone

A. Interbedded, thoroughly burrowed mudstone, silty mudstone and thin, very fine-grained silty sandstone with *Phycosiphon* (Ps), *Asterosoma* (As), *Teichichnus* (Te), and *Schaubcylindrichnus* (Sch). Note the sharp to burrowed lower contacts and the burrowed to gradational upper contacts between the sandstone and mudstone beds (Reindeer G-04, Core #4, depth ~2928.94 m).

B. Interbedded, thoroughly bioturbated mudstone, silty mudstone and thin, discontinuous, very fine-grained silty sandstone with *Phycosiphon* (Ps), *Helminthopsis* (He) and *Teichichnus* (Te) (Reindeer G-04, Core #4, depth ~2929.11 m).

C. Very fine-grained, intensely bioturbated sandstone, mudstone and silty mudstone with *Phycosiphon* (Ps), *Helminthopsis* (He), *Schuabcylindrichnus* (Sch) and *Diplocraterion* (Dp). Note the abundance of *Phycosiphon* near the upper portions of the sandstone beds (Reindeer G-04, Core #4, depth ~2929.02 m).

D. Thoroughly bioturbated mudstone and silty, very fine-grained sandstone with *Phycosiphon* (Ps), *Chondrites* (Ch) and *Teichichnus* (Te). Note the truncated, isolated oscillation ripples (W.r) (Reindeer G-04, Core #4, depth ~2929.13 m).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Facies A are interpreted to have been deposited in a moderately storm-dominated, distal prodelta (lower offshore) of a wave-dominated deltaic environment. The prodelta is the subaqueous, basal portion of the actively prograding delta front where fine-grained clays and silts are deposited mainly from suspension or from dilute turbidity current flows (Coleman and Prior, 1980). The prodelta or equivalent lower offshore zone is characteristically found well below fairweather (minimum) wave base but above storm (maximum) wave base (MacEachern and Pemberton, 1992). The prodelta gradationally merges seaward into the open shelf and landward into the delta front environment. The overall setting is that of a low energy, open marine environment dominated by quiescent, fairweather periods of mud deposition intermittently interrupted by high-energy storm events. Fine-grained sediments comprising the wavy mudstone and silty mudstone beds are interpreted as fairweather deposits characterized by suspension fallout with minor traction and saltation processes. The interstratified fine-grained silty sandstone beds are interpreted as distal tempestites that were deposited during periodic high-energy conditions associated with storm events. Locally preserved, remnant low angle, parallel lamination is interpreted to represent distal hummocky cross-stratification (HCS) (MacEachern and Pemberton, 1992). Conditions are optimal for the removal of sediment from the upper parts of the shoreface and transportation of this sediment into the relatively low energy conditions of the lower offshore zone. Alternatively, this sediment could have also been delivered to this setting during periods of heightened fluvial influx from distributary channels associated with storm events. Pemberton and Frey (1984) noted that in the lower offshore zone, renewed deposition following the cessation of scour by storm waves may incorporate storm layers within a sequence of otherwise low-energy deposits. These thin, unburrowed, distal storm beds generally have a high preservation potential because they are deposited well below fairweather wave base, where physical processes during non-storm periods are not competent enough to modify and rework these beds (Dott, 1983, 1988; Wheatcroft, 1990). The isolated asymmetrical and symmetrical ripples are interpreted as relict, storm wave and combined flow ripples. Harms *et al.* (1975) noted that combined flow ripples are formed by the superimposition of wave and current action by shoaling waves. These ripple structures are interpreted to have been deposited during major storm events when storm wave base interacts with the substrate. The silty mudstones are interpreted to have been deposited from suspension as buoyant plumes of sediment carried into the delta front during flooding events (Reading and Collinson, 1996).

The homogenization of sediments and virtual obliteration of bedding structures is interpreted to be the result of the burrowing activities of organisms in the lower offshore environment. The trace fossil assemblage of Facies A is characterized by a suite of burrows corresponding to the distal *Cruziana* ichnofacies (Seilacher, 1967) which is characteristic of quiet, open marine conditions within the lower offshore (Frey *et al.*, 1990; Pemberton *et al.*, 1992). The ichnological suite corresponding to the distal *Cruziana* ichnofacies is generally a low diversity assemblage composed of a high abundance and uniform distribution of both deposit feeding and to a lesser degree, grazing/ foraging structures (Frey *et al.*, 1990; MacEachern and Pemberton, 1992). Within Facies A, this assemblage can be described as a impoverished *Cruziana* assemblage comprised of *Helminthopsis, Phycosiphon, Chondrites, Planolites, Schaubcylindrichnus, Teichichnus* and rare *Asterosoma, Diplocraterion,* and *Palaeophycus. Helminthopsis* and *Chondrites* are associated with the mudstone interbeds and are interpreted to reflect a fairweather assemblage. *Schaubcylindrichnus* is common in lower energy parts of the offshore zone and is interpreted to represent suspension or deposit-feeding behaviour by a tubicolous organism (Frey and Pemberton, 1991; MacEachern and Pemberton, 1992). The thin-walled nature of this burrow was also described within lower offshore sediments from the Cretaceous Viking and Cardium formations of the Western Canadian Sedimentary Basin (MacEachern and Pemberton, 1992). The suspension-feeding structure *Diplocraterion* may reflect behaviour of organisms swept basinward during storms and survived to burrow the tempestite. The low abundance of this burrow suggests that organisms were unsuited to the fairweather conditions of this setting (Pemberton *et al.,* 1992; Pemberton and MacEachern, 1997) and likely represent "doomed pioneers" (Föllmi and Grimm, 1990).

The low to moderate diversity trace fossil assemblage is interpreted to reflect stressful environmental conditions, i.e., conditions were not conducive for the development of a diverse assemblage of organisms and only highly resilient forms were able to survive and successfully exploit the sediment (MacEachern and Pemberton, 1992; Pemberton *et al.,* 2001). The diminutive size of many of the ichnogenera is attributed to biological stresses including fluctuating sedimentation rates, salinity variations, diminished food supply, water turbidity, low oxygen levels, and light and temperature variations (MacEachern *et al.*, 2005). The diminutive size of the burrows suggests that juveniles were mainly present and that the aforementioned stresses inhibited growth and resulted in high mortality rates (Pemberton and MacEachern, 1997). *Phycosiphon,* which is abundant along the tops of the distal tempestite beds, is commonly described an opportunistic colonizer of storm-deposited sands (Goldring *et al.,* 1991; Pemberton *et al.,* 1992). These burrows decrease upward into the overlying mudstone beds, suggesting that fairweather conditions were less suitable for the colonization by the *Phycosiphon* trace maker. The overall high degree of bioturbation and near complete destruction of bedding is attributable to fluctuating sedimentation rates that allowed the burrowing organisms plenty of time to rework the mudstone. The presence of relatively unburrowed silty sandstone beds indicates higher sedimentation rates than for the mudstone, inhibiting biogenic reworking. The predominance of deposit-feeding and grazing structures, relatively low diversity of trace fossils, intense nature of burrowing and the sedimentary structures, all indicate that these sediments were deposited within an overall low energy, nutrient-rich, distal prodelta or equivalent lower offshore environment.

Facies B - Wavy to lenticular bedded bioturbated, organic mudstone and fine-grained, silty sandstone

Lithology and Sedimentary Features:

Facies B consists of a bioturbated, dark grey to black, organic mudstone interlaminated to interbedded with light grey to buff, lower very fine to lower fine-grained, silty sandstone (Figs. 2.2A-D). The silty mudstone beds are moderately to highly bioturbated, whereas the sandstone beds are weakly to moderately bioturbated (Figs. 2.2A,C). The mudstone and silty sandstone interbeds are thinner than in Facies A, ranging in thickness from 1 cm to approximately 10 cm. The mudstone beds range in thickness from 1 to 5 cm with some beds up to 10 cm thick. The sandstone beds range in thickness from 1 to 10 cm with some beds reaching 20 cm thick; bed thickness generally increase towards the top of the facies. Most of these sandstone beds have sharp to scoured bases and burrowed to gradational tops (Fig. 2.2D).

Sandstones are dominated by horizontal to low angle (<5°), parallel lamination that is commonly accentuated by finely comminuted, organic detritus (Fig. 2.3A). The sandstone beds typically have scoured bases but vary from biogenically homogenized to those with negligible bioturbation. Symmetrical and asymmetrical ripples commonly cap the tops of these sandstone beds. Other sandstone beds occur as unburrowed, isolated lenses within the mudstones. These isolated sandstone beds are generally low angle (<5°), parallel-laminated, possessing remnant ripple cross-lamination, appearing as load casted ripples. Interlaminated to interbedded mudstone and silty mudstone intervals display wavy to lenticular bedding. Synaeresis cracks are rare to common with lengths ranging from 1 cm up to 3 cm (Figs. 2.3C,D). There are also rare centimeter-scale chert grains and pyrite nodules.

Facies B is identified within five core intervals including core #2 from A-12, core #2 from E-21, core #1 from P-41, core #1 from F-09 and cores #4 and #5 from the G-04 well. The lower contact is gradational with the underlying Facies A. This contact is marked by a change in the ichnogenera as well as an increase in bed thickness and overall sandstone content. Facies B is gradationally to sharply overlain by Facies C. As well, it typically occurs within a coarsening upward succession composed of Facies A, C, D and E. Facies B varies in thickness from 56 cm up to 4.85 m, although due to limited core penetration depths, it is likely that thicker successions exist.

Ichnological Characteristics:

The interbedded mudstone and very fine-grained, silty sandstone beds of Facies B are characterized by an overall intensely to sporadically bioturbated texture. These deposits contain a uniformly distributed, highly diverse and abundant trace fossil suite. The mudstone interbeds generally

Figure 2.2

Facies B - Wavy to lenticular bedded bioturbated, organic mudstone and fine-grained, silty sandstone beds

A. Interbedded, well-burrowed, organic mudstone and thin, fine-grained sandstone with *Helminthopsis* (He), *Scolicia* (Sc), *Teichichnus* (Te), *Phycosiphon* (Ps) and *Planolites* (Pl) and pyrite nodules (Py). Note the moderate to high degree of bioturbation of the middle sandstone bed by *Helminthopsis* (He) and *Phycosiphon* (Ps) and the isolated (starved) wave-rippled (W.r.) sandstone beds near the top (Siku E-21, Core #2, depth ~2958.64 m).

B. Interbedded, well-burrowed, organic mudstone and thin, very fine-grained sandstone with *Planolites* (Pl) and *Teichichnus* (Te). The sandstone beds are wavy, parallel-bedded with low angle, parallel lamination, interpreted to reflect HCS (Parsons P-41, Core #1, depth ~2979.23 m).

C. Interbedded, well-burrowed, organic mudstone and thin, fine-grained sandstone with *Zoophycos* (Zo) and *Teichichnus* (Te). Note the isolated, wave-rippled sandstone beds within the mudstone dominated interval (Parsons P-41, Core #1, depth ~2970.92 m).

D. Interbedded, well-burrowed, organic mudstone and thin, very fine-grained sandstone with *Phycosiphon* (Ps), *Chondrites* (Ch), *Asterosoma* (As), *Teichichnus* (Te) and *Planolites* (Pl). The lower mudstone bed is highly bioturbated by a monospecific assemblage of *Chondrites* and to a lesser degree, *Planolites* (Pl) (Siku E-21, Core #2, depth ~2956.56 m).

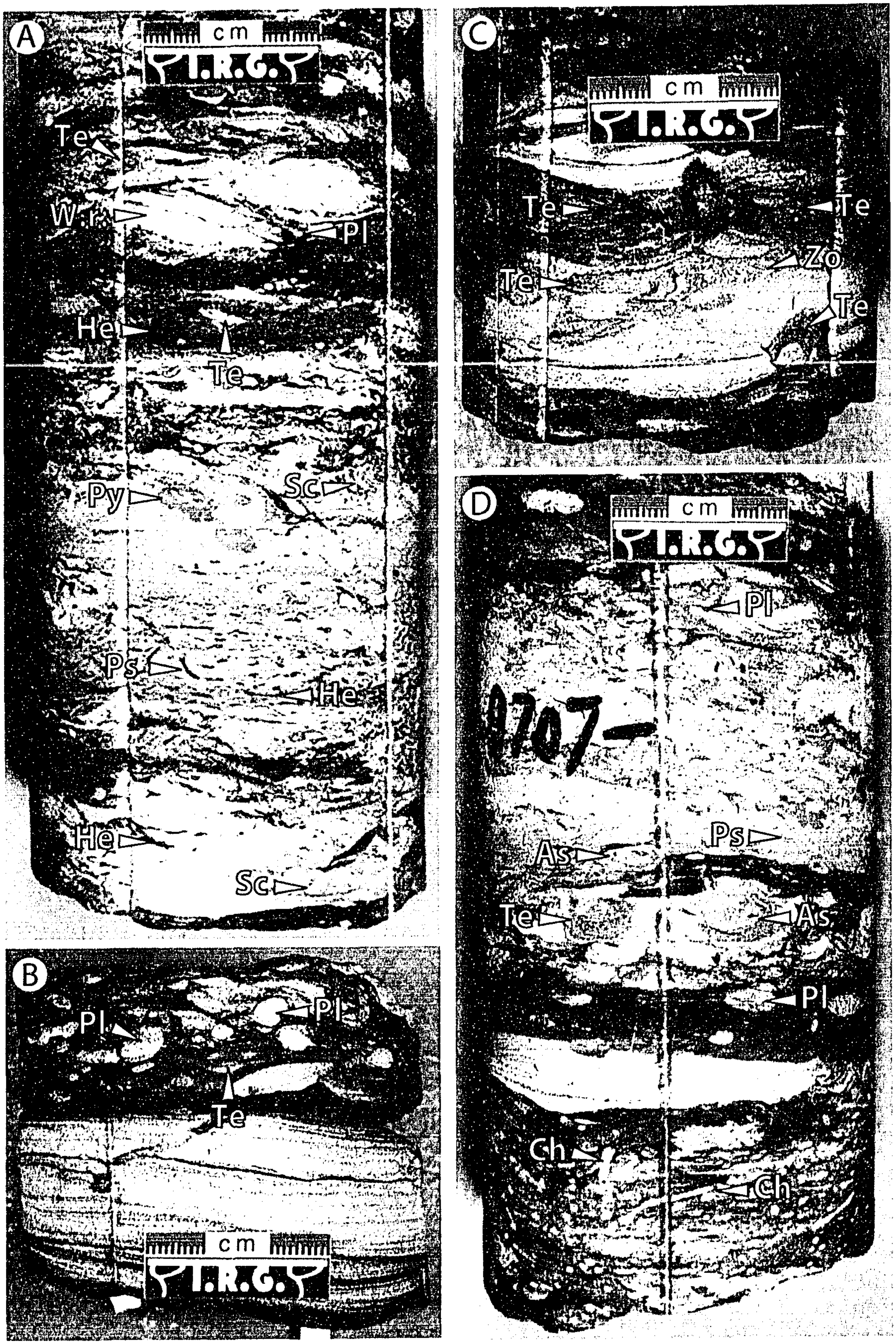
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Figure 2.3

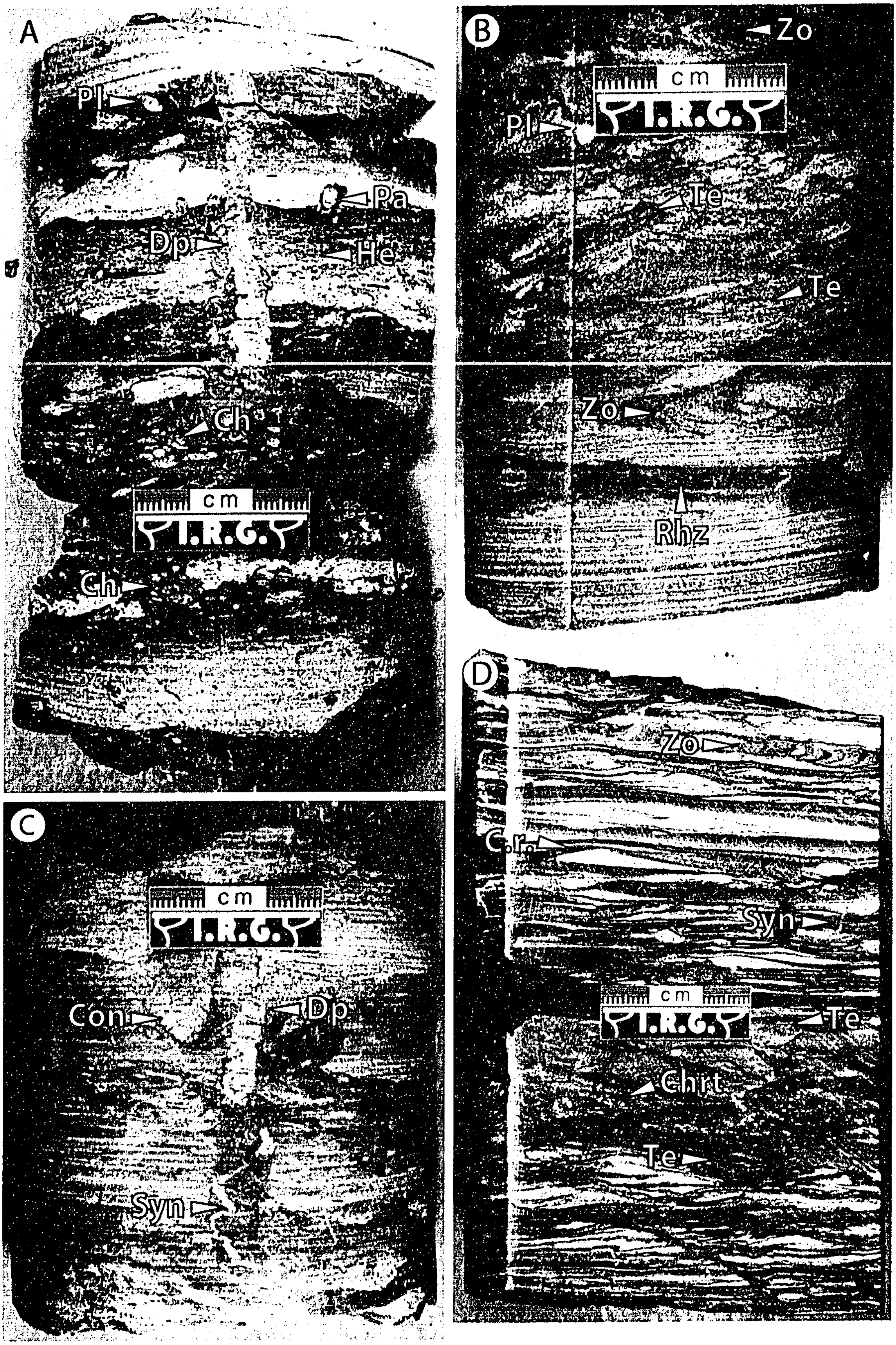
Facies B - Wavy to lenticular bedded bioturbated, organic mudstone and fine-grained, silty sandstone beds

A. Interbedded, well-burrowed, organic mudstone and thin, very fine-grained sandstone with *Helminthopsis* (He), *Diplocraterion* (Dp), *Planolites* (Pl) and *Chondrites* (Ch), *Palaeophycus* (Pa). The sandstone beds have scoured to sharp bases and gradational tops, are low angle (<5°), parallel-laminated, and are interpreted to reflect distal HCS. The *Diplocraterion* (Dp) burrow is infilled with sandstone that is coarser-grained than the surrounding sandstone beds, marking a concealed bed junction. The burrow fill is the only preserved evidence of a coarser-grained bed that has been subsequently removed by erosion (Reindeer G-04, Core #5, depth ~2934.76 m).

B. Well-burrowed, organic mudstone overlying a low angle, parallel-laminated very fine-grained silty sandstone. Bedding and laminae within the sandstone bed is marked by abundant, finely comminuted, organic detritus and phytodetrital material. The mudstone bed is well-burrowed by *Teichichnus* (Te), *Zoophycos* (Zo) and *Planolites* (Pl). The sandstone bed contains an excellent example of *Rhizocorallium* (Rhz) with both tubes and spreiten preserved in cross-section (Parsons P-41 Core #1, depth ~2969.19 m).

C. Interbedded, organic mudstone and thin, very fine-grained silty sandstone with *Diplocraterion* (Dp) and *Conichnus* (Co). These sandstone beds are low angle (<5°), parallel-laminated and are interpreted to reflect distal HCS. Note the large synaeresis crack (Syn) within the base of the *Diplocraterion* (Dp) burrow (Reindeer G-04, Core #5, depth ~2933.97 m).

D. Wavy to lenticularly bedded organic mudstone with thin, very fine-grained silty sandstone exhibiting with minor current and possible combined flow ripples. Identified ichnofossils include *Zoophycos* (Zo) and *Teichichnus* (Te). Note the small current ripple (C.r.), chert grains (Chrt) and the synaeresis crack (Syn) that cross-cuts the laminae (Parsons F-09, Core #1, depth ~2860.34 m).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contain a higher diversity of ichnogenera and degree of bioturbation than the sandstone interbeds. The ichnofossil assemblage contains a moderate abundance of ichnogenera including *Helminthopsis* (Fig. 2.2A), *Phycosiphon* (Figs. 2.2A,D), *Chondrites* (Figs. 2.2D; 2.3D), *Planolites* (Figs. 2.2B,D; 2.3A,B), *Thalassinoides*, *Arenicolites*, *Zoophycos* (Fig. 2.3D), *Rosselia*, *Schaubcylindrichnus*, *Asterosoma* (Fig. 2.2D), *Scolicia* (Fig. 2.2A), *Rhizocorallium* (Fig. 2.2A), *Diplocraterion* (Figs. 2.3A,C), *Teichichnus* (Figs. 2.2A-D; 2.3B,D), *Skolithos*, *Conichnus* (Fig. 2.3C), *Palaeophycus tubularis* (Fig. 2.3A), *Palaeophycus heberti* and rare fugichnia. There is a higher degree of biogenic reworking of sandstone beds by deeper-penetrating, robust forms than was evident in the underlying deposits of Facies A. Similar to Facies A, bioturbation by *Phycosiphon* is abundant, especially in the upper portions of sandstone beds (Figs. 2.2A,D). Both *Phycosiphon* and *Helminthopsis* occur in discrete horizons, typically within silt or fine-grained sandstone beds. Sandstone beds contain rare, vertical burrows that are sporadically distributed, including *Skolithos*, *Arenicolites* and *Diplocraterion*. Some of the organic mudstone interbeds are intensely bioturbated and contain a monospecific suite dominated by *Chondrites* (Fig. 2.2D). It is important to note that not all ichnogenera were identified within every cored interval, but multiple cored intervals contained the complete ichnological assemblage.

<u>Interpretation and Discussion:</u> Proximal Prodelta (Upper Offshore)

Interbedded to interlaminated bioturbated mudstone and very fine to fine-grained silty sandstone beds of Facies B are interpreted to have been deposited in a moderately storm-influenced, proximal prodelta, or upper offshore equivalent, of a wave-dominated, deltaic environment. The proximal prodelta or upper offshore setting is characteristically situated above storm-weather wave base (maximum) but below fairweather wave base (minimum) (Howard, 1971, 1972; Howard and Reineck, 1981; Howard and Frey, 1984; Walker and Plint, 1992; MacEachern and Pemberton, 1992). The overall depositional setting was a low energy, marine environment dominated by fairweather deposition and only periodically exposed to higher-energy storm events. The interbedded silty mudstone beds are interpreted to reflect fairweather deposition. The interstratified, sharp-based sandstone beds are interpreted to represent distal tempestite deposits.

The locally preserved, low angle, parallel lamination in the tempestites is interpreted to represent distal hummocky cross-stratification (HCS) formed by high-energy, storm-generated, oscillatory currents. HCS is recognized to form below fairweather wave base during intense storms (Harms *et al.*, 1975; Dott and Bourgeois, 1982; Hunter and Clifton, 1982). The thinner beds with wavy, parallel, ripple cross lamination are interpreted as oscillation to combined flow ripples, and are also thought to represent distal storm deposits. These ripples are interpreted as combined flow structures, as they display an overall symmetrical profile, but with small-scale foresets dipping in one direction. Combined flow ripples are formed by superimposed wave and current action, and by the actions of shoaling waves (Harms *et al.*, 1975). The relative abundance of current-generated structures provides evidence that considerable volumes of sediment originated from the distributary channels and was carried basinward in buoyant plumes, or by sediment gravity flows (Reading and Collinson, 1996). Synaeresis cracks are interpreted to form in settings with rapidly fluctuating salinities, that causes shrinkage of clays, and can form either both at the sediment-water interface or substratally (Burst, 1965; Plummer and Gostin, 1981). Synaeresis cracks are most common in the finely interlaminated mudstone and sandstone intervals, which suggests that they were formed during salinity variations caused by the influx of density underflows of sediment-laden freshets (MacEachern *et al.*, 2005) into the prodelta environment (Burst, 1965; Plummer and Gostin, 1981). Fresh water is introduced into this setting with the influx of sediment from distributary channels during increased fluvial discharge that typically accompany storms. These density underflows of sediment-laden fresh water occur due to high sediment concentrations, which may have facilitated inertia-dominated hyperpycnal flows (Coates, 2001). The sediment laden hyperpycnal flows travel down the delta front and stay near the bed in the prodelta and delta front. Synaeresis cracks and associated pyrite reflect fluctuating salinities in marginal marine settings (Wightman *et al.,* 1987), which further support the interpretation that Facies B was deposited in a wave-dominated, proximal prodeltaic environment.

The sharp-based, very fine-grained sandstone beds with scoured, sharp bases and burrowed to gradational tops interstratified with the mudstone reflect a "laminated-to-scrambled" or "lam-scram" bedding, as described by Howard (1972). These "lam-scram" beds record a transition from rapid sedimentation to quiescent fairweather conditions. Figure 2.8 illustrates an idealized model for "lam-scram" texture for a storm-deposited tempestite comprised of hummocky cross-stratified (laminated) sandstones and burrowed (scrambled) shales. The sandstone beds are commonly demarcated by an erosive scour surface and represent tempestite beds, which range from intensely burrowed to rarely unburrowed. The lack of burrowing is interpreted to reflect rapid sedimentation rates that inhibited infaunal colonization. The intensely bioturbated, interstratified mudstone reflects slow, continuous deposition during fairweather conditions. The prevalence of sandstone beds, deposited during major storm events suggests that the depositional environment was affected by persistent large storms. Normal grading, oscillation ripple-laminated tops, and suspension fall-out of silts reflect deposition during the waning stages of storms (MacEachern and Pemberton, 1992). Upper offshore deposits are considerably more variable than lower offshore deposits, due to the greater degree of influence during deposition from storm and wave interaction at the sediment-water interface (MacEachern and Pemberton, 1992). Some intervals within this offshore zone display decreased levels of bioturbation and more regular interbedding of mudstone and sandstone. These features were recognized by MacEachern and Pemberton (1992) to reflect greater storm domination and/or frequency, as well as higher sedimentation rates. Such deposits are commonly referred to as reflecting deposition in the "offshore transition". The increased sandstone proportion, thickness of beds, higher degree of preserved HCS in tempestite beds, and wave-generated ripples suggests deposition in a slightly higher energy setting and thus, potentially shallower water depths than Facies A.

The trace fossil assemblage of Facies B is characterized by burrows corresponding to the archetypal *Cruziana* ichnofacies (Seilacher, 1967). The trace fossil assemblage is dominated by deposit and grazing feeding behaviours, characteristic of open marine conditions. The archetypal *Cruziana* ichnofacies assemblage suggests deposition within a low energy environment, characteristically below fairweather wave base (minimum) but above storm wave base (maximum) (Frey *et al.*, 1990; Pemberton *et al.*, 1992). Burrowing within the upper offshore zone is typically intense with a high diversity and uniform distribution of ichnogenera reflecting a k-selected or equilibrium behaviour (MacEachern and Pemberton, 1992). The archetypal *Cruziana* assemblage is generally dominated by deposit-feeding behaviours with a comparatively low abundance of structures produced by grazing and foraging organisms (Frey *et al.*, 1990; Pemberton *et al.*, 1992). Suspension-feeding structures, such as *Diplocraterion, Arenicolites* and *Skolithos*, associated with both the fairweather mudstones and the storm-deposited tempestites reflect elements of the *Skolithos* ichnofacies. These suspension-feeding structures are interpreted to share a similar origin as discussed previously in Facies B. These structures are interpreted to reflect the behaviour of organisms that were swept basinward during storms and survived to colonize the tempestites. The lack of persistence of these structures throughout this facies suggests that they were unsuited to the fairweather conditions of this setting (Pemberton *et al.*, 1992, Pemberton and MacEachern, 1997). These organisms are carried basinward during emplacement of storm deposits and are analogous to the "doomed pioneers" of Föllmi and Grimm (1990). The ichnological assemblage of Facies B is a diverse suite of 18 ichnogenera. The dominant deposit-feeding structures include *Chondrites, Planolites, Teichichnus, Schaubcylindrichnus, Asterosoma, Scolicia, Rhizocorallium, Palaeophycus* and *Thalassinoides*. The grazing and foraging structures include *Helminthopsis* and *Phycosiphon*, which record colonization by opportunistic infauna following the storm event. Rarer suspension feeding and dwelling structures represent opportunistic colonization (or r-selected behaviour) following storm abatement (Pemberton and Frey, 1984; Pemberton *et al.*, 1992; Pemberton and MacEachern, 1997). MacEachern and Pemberton (1992) suggested that a paucity of suspension-feeding structures may reflect rapid burial of the sandstone beds by organic silty sediments, thus the colonization by organisms that favour sandy substrates was precluded. MacEachern and Pemberton (1992) noted that distal tempestites in the upper offshore typically display a higher degree of biogenic reworking compared to the thinner storm beds of the lower offshore. This is attributed to a greater degree of vertical penetration of the substrate by more robust, benthic organisms found in the upper offshore compared to those of the lower offshore (Dörjes and Hertweck, 1975). The trace fossil *Zoophycos* has numerous ethological interpretations, and its origins are actively debated. *Zoophycos* has been interpreted as a grazing/foraging structure, a deposit-feeding structure (Kotake, 1989, 1991), and a deep tier, deposit-feeding structure (Bromley, 1991). *Zoophycos* is rare within this facies.

Pervasive *Chondrites* within organic-rich mudstones are interpreted to reflect the opportunistic colonization by an organism that employed a deep burrowing strategy in order to feed on the nutrient rich, bacterial blooms, degrading from storm buried organic material (Bromley and Ekdale, 1984; Vossler and Pemberton, 1988). Such conditions are present in chemically reduced, oxygen-depleted conditions below the redox boundary (Bromley and Ekdale, 1984). The *Chondrites* trace maker was able to maintain communication with the oxygen-rich waters of the seafloor while burrowing deeply into the sediments. Vossler and Pemberton (1988) recognized that post-storm muds are buried before surface recolonization and bioturbation can take place. The monospecific assemblage of *Chondrites* within these organic rich mudstone layers suggests that oxygen was a limiting factor on benthic organisms, at least during deposition of some of these beds. The *Chondrites* trace maker employs an opportunistic behaviour to take advantage of an unexplored niche that other species are not able to. This type of episodic deposition could occur in storm-influenced or delta influenced shoreface settings where sediment is distributed in pulses to the marine environment via adjacent rivers.

Facies C - Interbedded, bioturbated, very fine to fine-grained, low angle parallel-laminated sandstone and bioturbated organic mudstone

<u>**Lithology and Sedimentary Features:**</u>

Facies C consists of moderately bioturbated, light grey to buff, upper very fine to lower medium-grained, low angle, parallel-laminated sandstone, interbedded with bioturbated, dark grey to black, organic mudstone and silty mudstones (Figs. 2.4A-D; 2.5 A-D). The sandstone beds are characterized by low to medium angle (<5° to 10° average up to 15° maximum), parallel lamination, with some beds appearing structureless (Figs. 2.4A-D; 2.5A,C). In addition, some of the siltier and muddier sandstone intervals display wavy to lenticular bedding. Bedding and lamination display a gentle curvature, including both concave-upward and concave-downward geometries with many internal truncations (Fig. 2.4C). The majority of sandstone beds are stacked, display laminasets that thicken and thin, and show diverging and converging geometries (Figs. 2.4A,D). These sandstone beds commonly display scoured to sharp bases and burrowed to gradational tops (Figs. 2.4B,D; 2.5B,D; 2.7B). Symmetrical and asymmetrical ripples are commonly preserved along the tops of some low angle, parallel-laminated sandstone beds (Fig. 2.4B). Many of the low angle, parallel-laminated sandstone intervals are characterized by soft-sediment deformation structures including decimeter-scale intervals of convolute bedding (Fig. 2.5A). Thickness of sandstone beds ranges from centimeter to decimeter scale with most beds ranging from 5 to 20 cm, with amalgamated intervals up to 50 cm. Bedding is on average thicker than in Facies B. The intervening mudstone beds are thinner, ranging from less than 1 cm to 10 cm in thickness.

The mudstones and silty mudstones beds that overlie low angle, parallel-laminated sandstone are intensely burrowed and display a mottled to scrambled texture. They are present in low abundance and are absent within amalgamated intervals, where they have been erosionally removed. These mudstone beds are organic rich (carbonaceous) and weakly fissile. The lower contacts are burrowed to gradational and the upper contacts are scoured to sharp with the overlying sandstone interbeds. Thickness of the mudstone interbeds averages between 1 to 10 cm. Overall, there is a decrease in the thickness and abundance of mudstone interbeds towards the top of the succession, becoming almost completely absent in the upper portion of the facies. Organic material in the lower portion of this facies is relatively abundant (15% to 20%). It occurs in the form of phytodetrital material, disseminated macerated organics and micaceous material (biotite and muscovite), which marks the laminae and bedding structures. Some of the sandstone beds display centimeter-scale, fining-upward, graded beds capped by the finely disseminated organic and micaceous material (Fig. 2.7A). Coalified pieces of wood are commonly preserved on bedding planes. The majority of these wood fragments are 5 to 10 cm in length and 6 to 7 cm in width. The preserved size of these fragments is limited to the diameter of the core. Chert and quartz granules and pebbles are commonly found floating within mudstone and silty mudstone interbeds. Large siderite nodules (up to 5.5 cm in diameter) are commonly found within mudstone and sandstone beds, with some beds displaying complete sideritization (Fig. 2.6D). Pyrite nodules and synaeresis cracks are rare.

As previously mentioned, upper portions of this facies display amalgamation of sandstone beds with abundant internal scour and truncation surfaces and absence of mudstone interbeds. These stacked, amalgamated intervals are characterized by abundant, low angle, scour surfaces, which are commonly marked by the truncation of bedding and/or shell hashes composed mainly of disarticulated bivalve shells (Fig. 2.5C). The majority of these shells are oriented concave-downward and commonly pyritized. Some scour surfaces are also marked by thin pebble lags of chert granules and mudstone clasts (less than 1 cm diameter) with associated iron staining. The sandstone beds are generally sporadically burrowed, with many beds completely unburrowed. Examples of Facies C displaying this erosional amalgamation occur in core #2 from the A-12 well, core #1 from the L-43 well and core #5 from the G-04 well.

Facies C is present in eight core intervals including core #2 from A-12, core #2 from E-21, core #1 from L-43, core #1 from P-41, core #1 from F-09, and cores #4 and #5 from the G-04 well. It is commonly underlain by Facies B, G or L. Facies C normally overlies Facies B, where the contact is gradational and marked by a change in the trace fossil assemblage, a decrease in mudstone content and an increase in bed thickness. The upper contact of Facies C is generally gradational with Facies D, although Facies B and G can scour into it. The upper contact with Facies D is marked by an increase in the overall proportion of sandstone and grain size, as well as a decrease in the amount of interstratified mudstone and bioturbation. Facies C typically occurs within coarsening-upward successions composed of Facies A, B, D and E. Intervals of Facies C ranges in thickness from 1.93 m up to 7.67 m, although it is likely that both thinner and thicker successions exist.

Ichnological Characteristics:

The sandstone intervals of Facies C display a low to moderate degree of bioturbation with a moderate to high diversity of trace fossils, whereas the mudstone interbeds are generally thoroughly bioturbated, displaying a moderate to high diversity of trace fossils. Bioturbation within the sandstones is overall sporadic, but commonly restricted to the lower and upper portions of the beds. Facies C is dominated by burrows corresponding to a mixed occurrence of the *Cruziana* and *Skolithos* ichnofacies and displays the "lam-scram" texture as described by Howard (1972) and Pemberton *et al.* (1992), and is similar to that seen in Facies B albeit on a larger scale. The overall ichnological suite is of a moderate diversity proximal *Cruziana* assemblage with lesser contributions from elements of the *Skolithos* ichnofacies.

The ichnological suite of this facies contains 22 ichnogenera including *Helminthopsis* (Figs. 2.4D; 2.5B; 2.7B), *Phycosiphon* (Figs. 2.5D; 2.6D; 2.7B,D), *Rhizocorallium* (Figs. 2.4C; 2.8A,B), *Teichichnus* (Fig. 2.8C), *Ophiomorpha* (Figs. 2.5D; 2.8A), *Rosselia* (Figs. 2.6A,B; 2.7B; 2.7D), *Palaeophycus* (Figs. 2.5D; 2.6A,B), *Planolites* (Figs. 2.4A; 2.6A,C,D; 2.7D), *Skolithos* (Fig. 2.6B), *Diplocraterion* (Figs. 2.7C; 2.8D), *Chondrites* (Figs. 2.4A,D; 2.5B), *Siphonichnus* (Fig. 2.6C), *Thalassinoides* (Figs. 2.5D; 2.6B), *Asterosoma, Macaronichnus* (Fig. 2.7A), *Scolicia, Schaubcylindrichnus* (Fig. 2.7B), *Cylindrichnus* (Fig. 2.5B), *Bergaueria, Conichnus, Lockeia* and fugichnia. *Helminthopsis, Chondrites, Scolicia,* and *Planolites* are typically restricted to the mudstone interbeds. Facies C in both core #2 from the A-12 and core #1 from the F-09 well contain abundant *Rhizocorallium* imparting a biogenic fabric texture best described as a 'Psycho Rhizo' texture (Fig. 2.8A). These occurrences of *Rhizocorallium* display tubes exhibiting vertical migration and have *Ophiomorpha* and *Teichichnus* characteristics. *Rosselia* within this facies are large and robust, and display an inclined nature. A single example of a zone with large, thickly walled, *Macaronichnus* occurs within the upper portion of Facies C in the P-41 well (Fig. 2.7A). The *Macaronichnus* display an unsorted, infill with a sorted lining and appear to have characteristics of *Palaeophycus*, *Planolites* and *Macaronichnus*. The unsorted nature of the burrow fills and sorted lining deviates from the diagnostic characteristics of *Macaronichnus* where burrow fills are biogenically sorted and the walls are composed of unsorted micaceous material. These burrows occur in tightly packed clusters that tend to not crosscut each other. Many intervals are dominated by cryptic bioturbation, which imparts a fuzzy nature to the internal laminae of the sandstone.

Interpretation and Discussion: Distal Delta Front (Lower Shoreface)

Interstratified, low angle, parallel-laminated sandstone, bioturbated, mudstone of Facies C is interpreted to have been deposited in a low to moderately storm-influenced, distal delta front or upper offshore to lower shoreface equivalent, of a wave-dominated, deltaic environment. The distal delta front is defined as beginning at the lower limit of fairweather (minimum) wave base,

Figure 2.4

Facies C - Interbedded, bioturbated, very fine to fine-grained, low angle parallel-laminated sandstone and bioturbated organic mudstone

A. Fine-grained, low to medium-angle (<5° up to 10°), parallel laminated sandstone overlain by wavy bedded sandstone and mudstone. This core slab displays a portion of an interpreted hummock, which is larger than the diameter of the core. The lower sandstone displays thickening and thinning of laminasets with truncation of the underlying beds interpreted to represent HCS. Note the abundant *Chondrites* (Ch) and *Planolites* (Pl) within the organic mudstone beds. There is also abundant phytodetrital and finely comminuted organic detritus highlighting the laminae (Siku A-12, Core #2, depth ~ 2804.71 m).

B. Low angle (<5° up to 10°), parallel-laminated, well-sorted, fine-grained sandstone interpreted to represent swaley cross-stratification (SCS). The sandstone beds thicken and thin and are capped by wave-ripple (W.r.) lamination. Note the abundance of organic detritus comprised of phytodetrital and micaeous material (Mc) marking the laminae and bedding (Siku A-12, Core #2, depth ~2806.92 m).

C. Amalgamated, low angle (<5° up to 10°), parallel-laminated, fine-grained sandstone displaying erosional truncation of the lower beds. Note the thickening and thinning of the laminae displaying diverging and converging geometries, typical of hummocky cross-stratification (HCS). The only burrow identified is a *Rhizocorallium* (Rhz) (Parsons G-04, Core #4, depth ~2920.67 m).

D. Fine-grained, low angle (<5°), parallel-laminated sandstone with laminasets that thicken and thin, as well as display diverging and converging geometries interpreted to represent HCS. Identified trace fossils include *Helminthopsis* (He) and *Chondrites* (Ch), which are mainly restricted to the intervening mudstone beds. The sandstone contains a moderate amount of finely comminuted organics and phytodetrital material which mark the laminae and bedding (Siku A-12, Core #, depth ~2933.70 m).

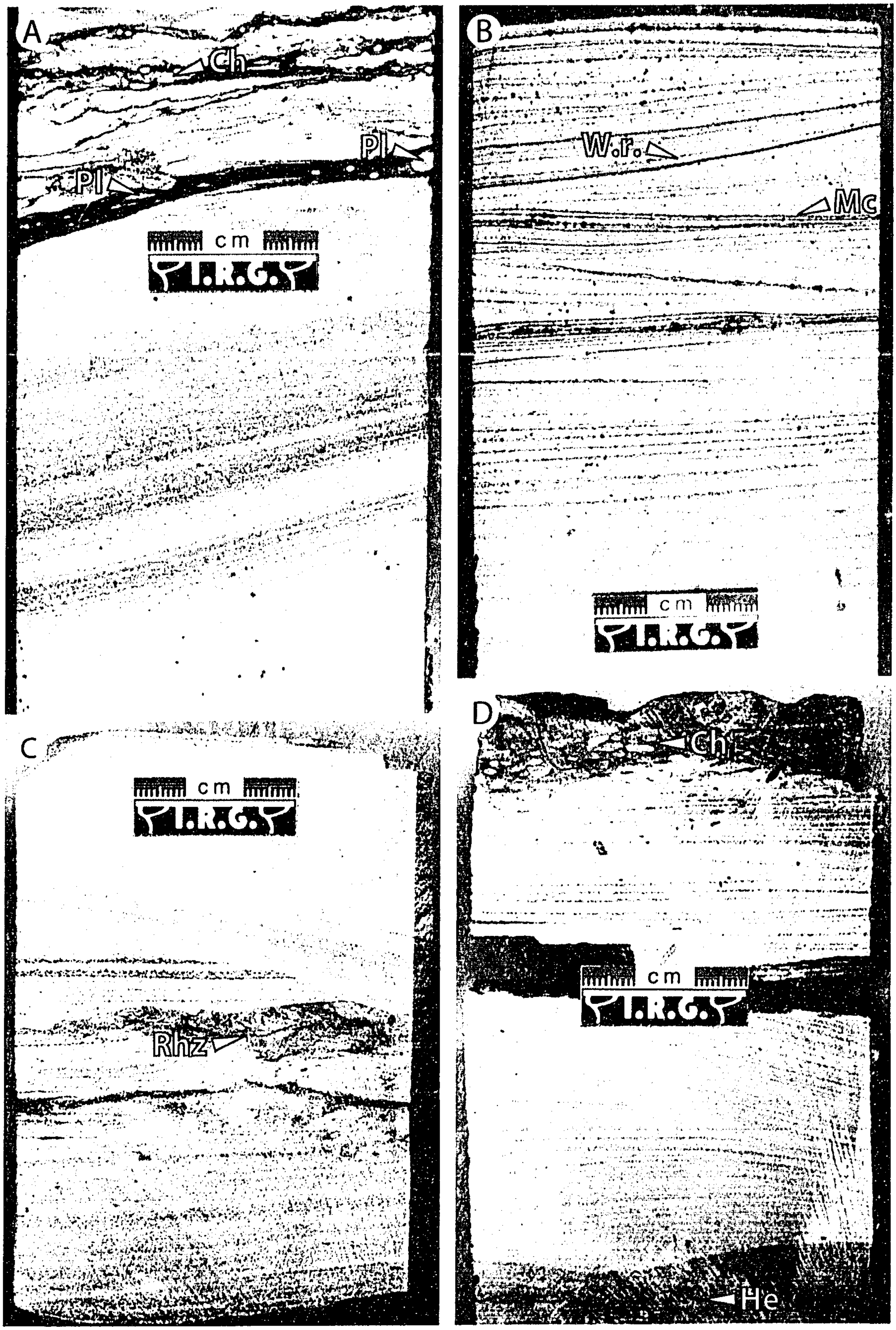
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Figure 2.5

Facies C - Interbedded, bioturbated, very fine to fine-grained, low angle parallel-laminated sandstone and bioturbated organic mudstone

A. Fine-grained, convoluted sandstone bed overlain and underlain by low angle (<5°), parallel-laminated sandstone. Note the sharp contacts between the low angle, parallel-laminated sandstone beds and the intervening, convolute bedded interval. The sandstone beds are marked by thin micaceous (Mc) laminae (Siku A-12, Core #2, depth ~2809.23 m).

B. Interbedded, well-burrowed, silty mudstone and horizontal to low angle (<5° up to 10°), parallel-laminated fine-grained sandstone. The mudstone beds are highly burrowed by *Helminthopsis* (He), *Chondrites* (Ch) and *Cylindrichnus* (Cy). The sandstone bed contains *Skolithos* (Sk) and *Rhizocorallium* (Rhz). Note the scoured base of the sandstone bed (Siku A-12, Core #2, depth ~2806.85 m).

C. Amalgamated, low angle (<5° up to 10°), parallel-laminated fine-grained, well-sorted sandstone. Note the subtle scour surface between the two amalgamated sandstone beds marked by a thin shell hash composed of disarticulated pelecypod (Plcpd) shell fragments (Reindeer G-04, Core #5, depth ~2937.57 m).

D. Interbedded, low angle (<5° up to 10°), parallel-laminated, fine-grained sandstone and well-burrowed, organic mudstone. The sandstone beds contain *Ophiomorpha* (Oph) and *Palaeophycus* (Pa), whereas the mudstone is burrowed by *Phycosiphon* (Ps), *Palaeophycus* (Pa) and *Thalassinoides* (Th). Note the sharp to burrowed nature of the contact between the sandstone and mudstone beds (Siku E-21, Core #2, depth ~2957.61 m).

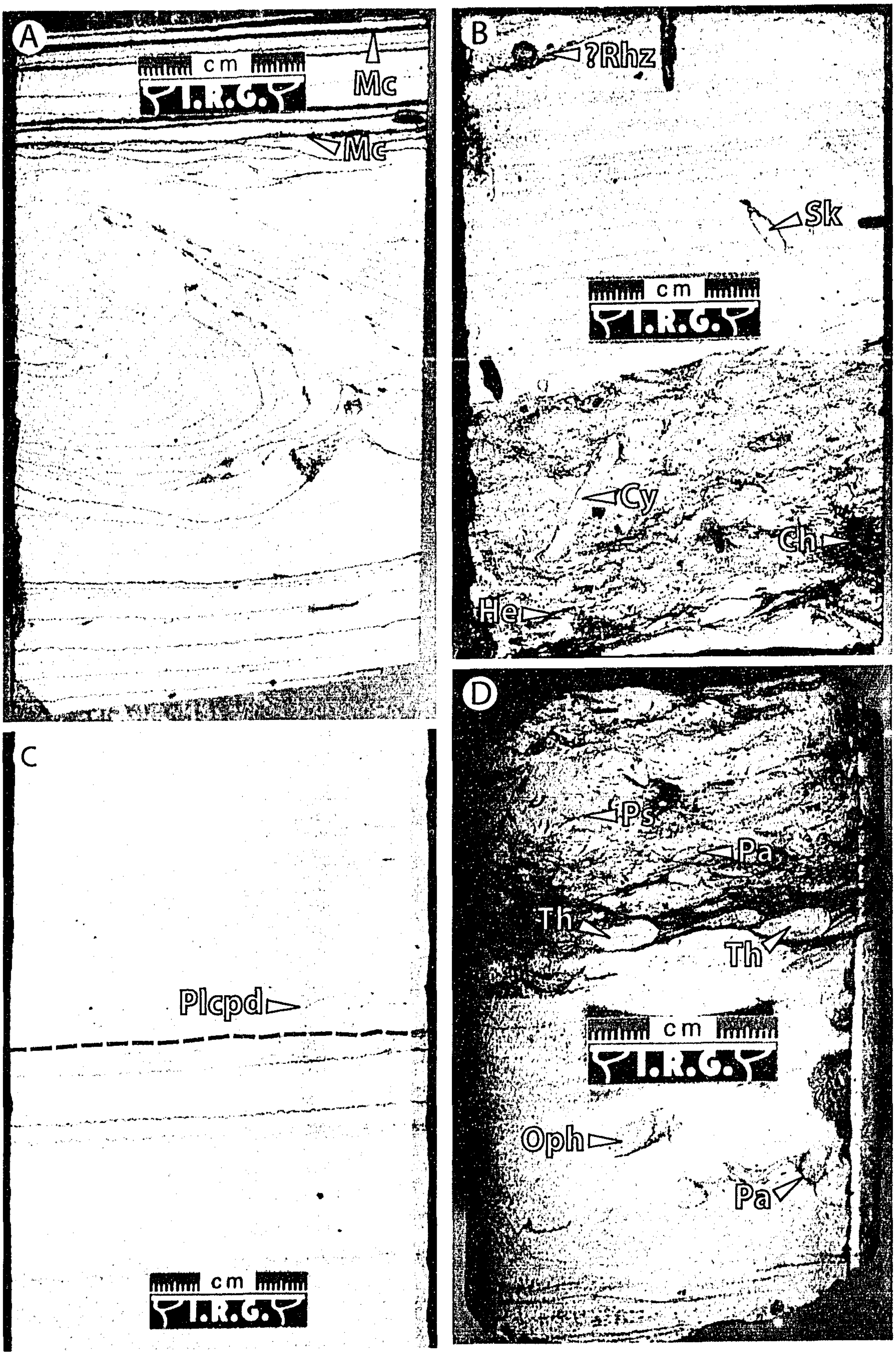
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Figure 2.6

Facies C - Interbedded, bioturbated, very fine to fine-grained, low angle parallel-laminated sandstone and bioturbated organic mudstone

A. Well-burrowed, low angle, parallel-laminated, fine-grained sandstone with large *Rosselia* (Ro), *Rhizocorallium* (Rhz), *Planolites* (Pl) and *Palaeophycus* (Pa). Note the large *Rosselia* mud 'bulb' with concentric mud layering and a sand-filled central dwelling tube (Siku E-21, Core #2, depth ~2957.71 m).

B. Well-burrowed, fine-grained sandstone with *Skolithos* (Sk), *Rosselia* (Ro), *Thalassinoides* (Th) and *Palaeophycus* (Pa). Note the multiple, mud-lined burrows that could possibly represent truncated stalks of *Rosselia*, with the mud 'bulbs' erosionally removed (Siku E-21, Core #2, depth ~2956.00 m).

C. Well-burrowed, low angle, parallel-laminated, fine-grained sandstone with large *Siphonichnus* (Sp) and smaller *Planolites* (Pl). Note the thin, millimeter thick fecal pellets (Fp) associated with *Siphonichnus* (Sp). The internal structure of *Siphonichnus* is likely formed by a bivalve, and there appears to be evidence of re-adjustment of the trace making organism with an overall upward vertical migration of the trace (Parsons G-04, Core #4 depth ~2924.67 m).

D. Low angle (<5°), parallel-laminated, fine-grained sandstone with sub-rounded to sub-angular siderite (Sid) and chert pebbles (Chrt). *Phycosiphon* (Ps), *Planolites* (Pl) and *Palaeophycus* (Pa) are the only identified burrows. Note the abundant organic detritus consisting of finely comminuted organics and phytodetrital material marking laminae and bedding structures (Reindeer G-04, Core #4, depth ~2925.71 m).

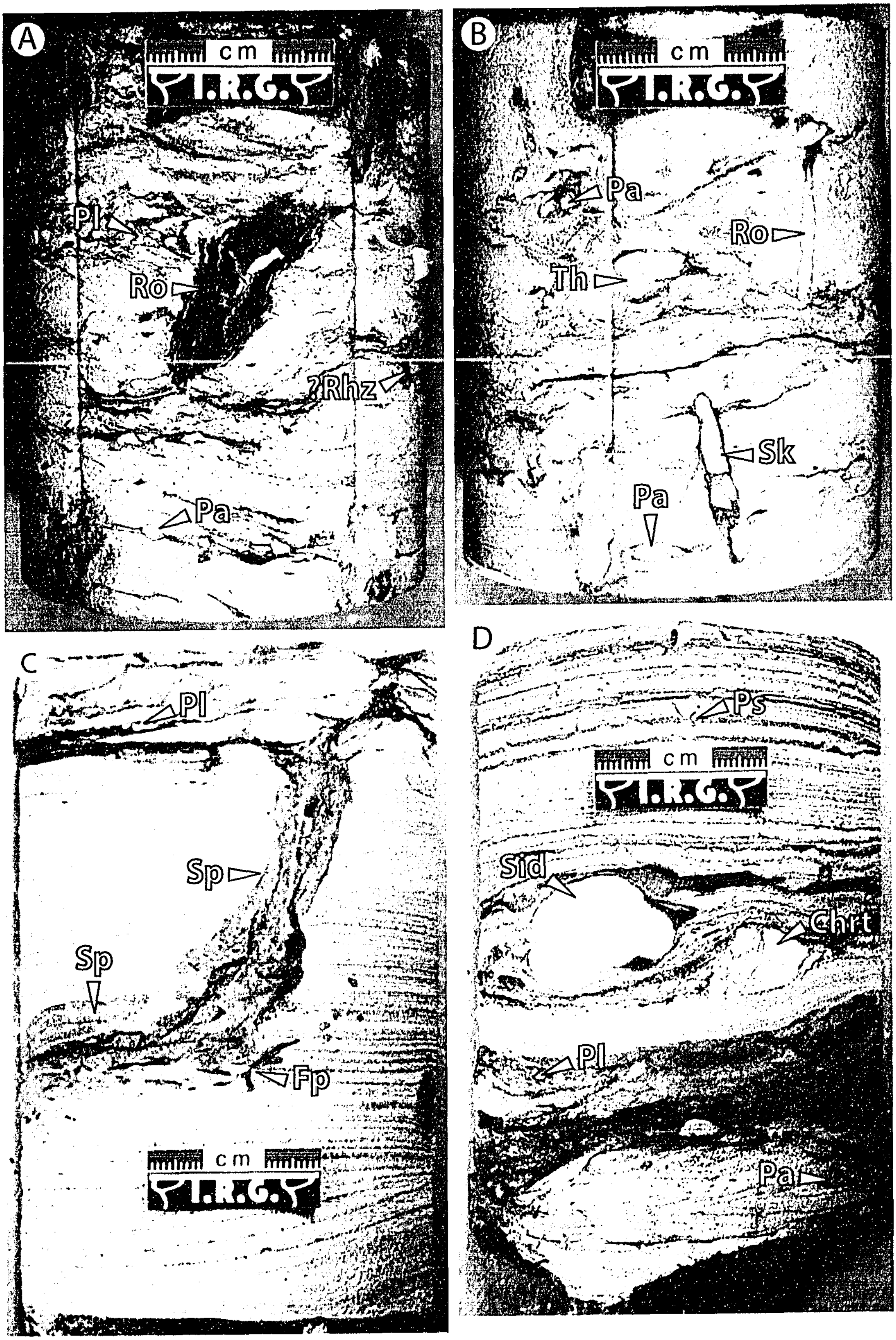
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Figure 2.7

Facies C - Interbedded, bioturbated, very fine to fine-grained, low angle parallel-laminated sandstone and bioturbated organic mudstone

A. Horizontal to low angle (<5° up to 10°), parallel-laminated, fine-grained sandstone with large *Macaronichnus* (Ma). These burrows display an unsorted fill with a sorted lining, a characteristic which deviates from the "normal" appearance of *Macaronichnus,* where burrow fills are biogenically sorted and the linings are composed of unsorted material. The sandstone displays centimeter-scale graded rhythmites (Rtm), consisting of small-scale, fining upward trends that are generally capped by finely comminuted organics and phytodetrital material (Parsons P-41, Core #1, depth ~ 2976.79 m).

B. Interbedded, fine-grained, low angle (<5°), parallel-laminated, silty sandstone and well-burrowed organic mudstone. The lower sandstone bed is capped by wave-ripple lamination, which hosts traces extending from the overlying mudstone beds. *Phycosiphon* (Ps), *Helminthopsis* (He), *Schaubcylindrichnus* (Sch), *Rosselia* (Ro), *Planolites* (Pl) and fugichnia (Fg) are typical trace fossils. The upper sandstone bed contains *Phycosiphon* (Ps), which display hook-shaped cross-sections, versus *Helminthopsis,* which is dominantly a horizontal trace. Note the sharp to erosional lower contact and the burrowed upper contact of the sandstone beds, with escape traces (Fg) within the lower sandstone bed (Reindeer G-04, Core #4, depth ~2926.66 m).

C. Wavy bedded, fine-grained sandstone and organic mudstone. Ichnofossils include *Diplocraterion* (Dp) and *Teichichnus* (Te). The burrows in this core slab mark a concealed bed junction where the burrows infilled with coarse-grained sands and granules originating from an overlying bed that was subsequently erosionally removed (Siku A-12, Core #2, depth ~2814.66 m).

D. Moderately-burrowed, fine-grained sandstone with *Rosselia* (Ro), *Phycosiphon* (Ps) and smaller *Planolites* (Pl). Note that the sand filled central dwelling tube of the large *Rosselia* is thickly mud mantled. The underlying organic mudstone bed is thoroughly bioturbated by *Planolites* (Pl) (Siku E-21, Core #2, depth ~2957.05 m).

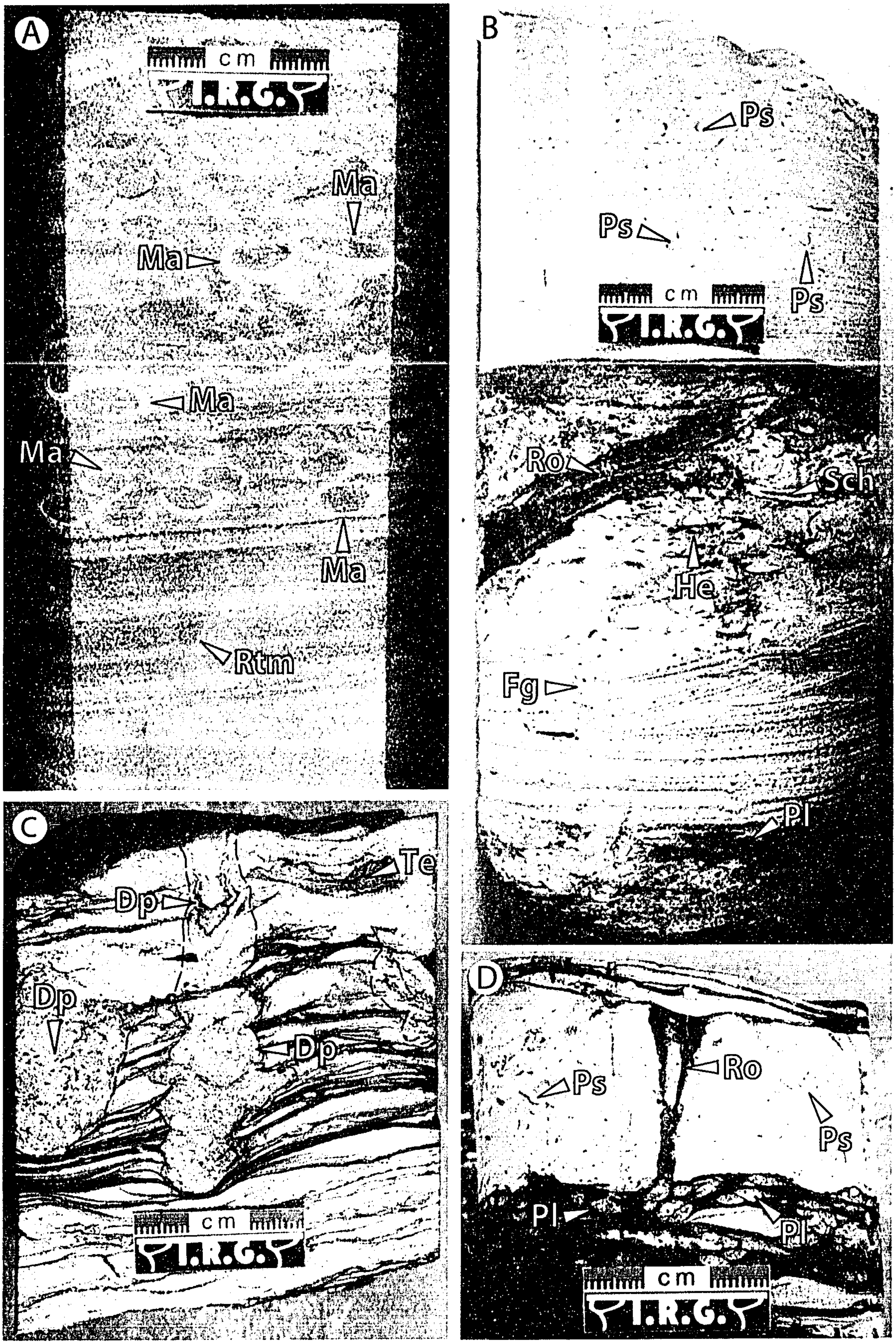
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Figure 2.8

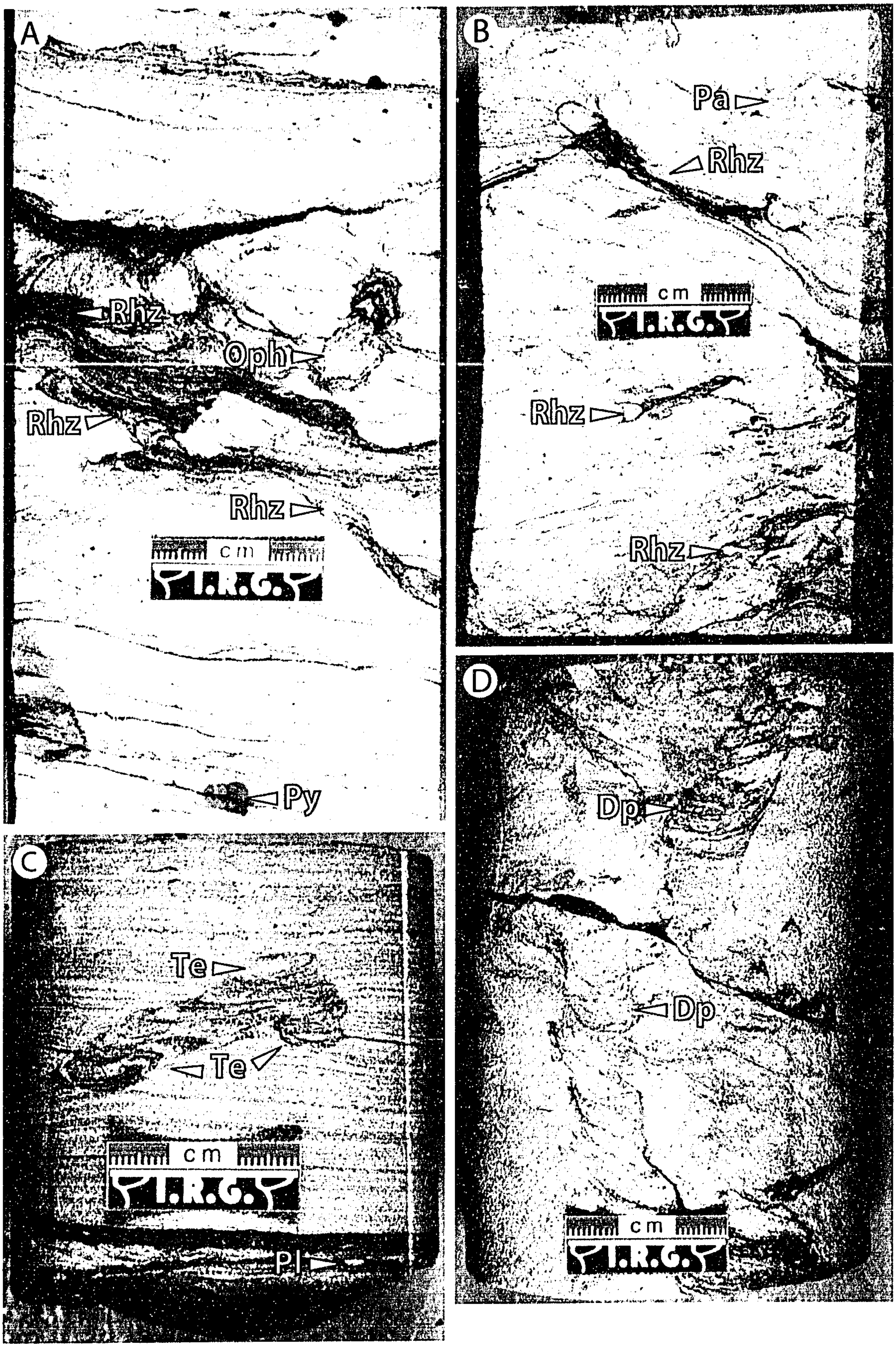
Facies C - Interbedded, bioturbated, very fine to fine-grained, low angle parallel-laminated sandstone and bioturbated organic mudstone

A. Low angle, parallel-laminated, fine-grained sandstone with large *Rhizocorallium* (Rhz) and *Ophiomorpha* (Oph). Note the vertical migration of *Rhizocorallium* tubes and the interconnecting spreiten (Parsons F-09, Core #2, depth ~2816.88 m).

B. Low angle, parallel-laminated, fine-grained sandstone with an excellent example of a large *Rhizocorallium* (Rhz) with the two burrow tubes displayed (Parsons F-09, Core #1, depth ~2860.65 m).

C. Low angle, parallel-laminated, fine-grained sandstone with large *Teichichnus* (Te) displaying a zig-zag migration of the burrow structure. This morphology is commonly recognized in shoreface sandstone and given the informal name of *Teichichnus zig-zagus* (S.G. Pemberton pers. comm., 2003) (Parsons P-41, Core #1, depth ~2969.97 m).

D. Fine-grained sandstone with large, upward-migrating *Diplocraterion* (Dp) that displays retrusive spreite (Parsons F-09, Core #2, depth ~2856.76 m).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but where offshore processes continue to operate (Reinson, 1984; Walker and Plint, 1992). The distal delta front merges seaward into the prodelta and landward into the proximal delta front and lower delta plain Storms are the dominant influence on sediment deposition in the lower shoreface, with the majority of the depositional record composed of storm deposits (MacEachern and Pemberton, 1992). The lower portion of this facies is characterized by interbedded, low to medium angle (<5° up to 15°), parallel-laminated sandstone beds displaying concave-upward and concave-downward stratification interpreted to reflect hummocky cross-stratification (HCS) (Dott and Bourgeois, 1982). Additional distinguishing features, such as low angle laminations, diverging and converging internal geometries, scoured basal surfaces, symmetrical wave ripples are also characteristically associated with HCS. The upper portion of this facies is generally composed of erosionally amalgamated, low angle (<5° up to 15°), parallel-laminated sandstones with concave-upward stratification interpreted as swaley cross-stratification (SCS) (Leckie and Walker, 1982). Distinguishing features of SCS include low angle parallel laminations, preferential preservation of concave-upward geometries, internal truncations and rare wave ripples. Both HCS and SCS are currently thought to reflect a storm-generated origin (Harms *et al.,* 1975; Duke, 1985; Duke *et al.*, 1991; Dott and Bourgeois, 1982; Leckie and Walker, 1982).

Hummocky cross-stratification was originally identified by Campbell (1966) as "large-scale truncated wave ripple laminae". Harms *et al.* (1975) formally introduced the term "hummocky cross-stratification" for these same sedimentary structures. Distinguishing characteristics of HCS identified by Harms *et al.* (1975) included: 1) bedsets with erosional lower bounding surfaces with dips of less than 10° and some up to 15°, 2) laminae that are parallel or approximately parallel with the lower erosional boundaries, 3) laminae that thicken and thin laterally within a bedset with similar decreasing dips from the crests 4) variable dip directions of the erosional set boundaries and overlying laminae. In terms of scale, the hummocks are between 10 to 50 cm and spaced 1 m to a few meters apart (Harms *et al.,* 1975). HCS beds commonly have sharp bases and exhibit drag or prod marks with wave and/or combined flow ripples capping the tops.

There is an ongoing controversy regarding the origins and hydrodynamic processes responsible for hummocky cross-stratification. Harms *et al.* (1975, 1982), Dott and Bourgeois (1982), Swift *et al.* (1983), Allen (1985), Duke (1985), Nottvedt and Kreisa (1987), Hunter and Clifton (1982), Leckie and Krystinik (1989), Southard *et al.* (1990) and Duke *et al.* (1991) discussed the nature, possible origins and environments of HCS deposition. The exact process responsible for the formation of HCS is still actively debated, however most workers attribute the bedform to storm activity (Harms *et al.,* 1975; Dott and Bourgeois, 1982; Swift *et al.,* 1983; Duke, 1985), which typically affects sediment deposition below fairweather wave base. There are currently two working models for the formation of hummocky cross-stratification; 1) a purely oscillatory flow model and 2) a combined oscillatory and unidirectional current flow model. The purely oscillatory flow model for the formation of HCS is favoured by authors such as Harms *et al.* (1975, 1982) and Dott and

Bourgeois (1982). The combined oscillatory and unidirectional current flow model is favoured by authors including Swift *et al.* (1983), Allen (1985) and Nottvedt and Kreisa (1987). A variation of the combined oscillatory and unidirectional current flow model proposed by Hunter and Clifton (1982), Leckie and Krystinik (1989) and Duke *et al.* (1991) suggests the main process is a very strong oscillatory dominant flow, with a only a weak, shore-normal, combined flow component.

Harms *et al.* (1975, 1982) stated that the process responsible for the formation of HCS is similar to that of wave ripples, but stronger wave action with surges of greater displacement and velocity than those required to form ripples is necessary. In this purely oscillatory model stronger wave energy is interpreted to have created elevated wave orbitals that reworked the hummocks and swales (Harms *et al.*, 1975; Dott and Bourgeois, 1982; Duke, 1985). The main lines of evidence used by Harms *et al.* (1975, 1982) for the purely oscillatory origin for HCS includes the close association of wave ripples, the lack of well organized directional trends, the mantling relationships of laminae to lower set boundaries, and the sharp basal contact with some drag marks. Recent flume experiments by Southard *et al.* (1990) have reproduced hummocky cross-stratification in fine-grained sands and further support the purely oscillatory flow model as suggested by Harms *et al.* (1975, 1982) and Dott and Bourgeois (1982).

The combined oscillatory and unidirectional current flow model for favoured by Swift *et al.* (1983), Allen (1985) and Nottvedt and Kreisa (1987) suggests that geostrophic currents are required for the formation of hummocky cross-stratification. On modern coasts intense wave energy is typically associated with geostrophic currents, which typically flow shore-parallel or shore-oblique due to the Coriolis Effect (Davis 1977; Leckie and Krystinik 1989). This modern interaction of wave energy and shore-parallel geostrophic currents was interpreted by Swift *et al.* (1983), Allen (1985) and Nottvedt and Kreisa (1987) as the process responsible for the combined oscillatory and unidirectional current flow regime required to form HCS. However, paleocurrent studies collected from numerous HCS deposits were unable to provide conclusive evidence that HCS was formed in response to geostrophic currents flowing parallel to the shoreline. In fact, most paleocurrent data collected from HCS deposits supported a combined flow model with a dominant oscillatory flow and a weaker, shore-normal combined flow component (Leckie and Krystinik, 1989). Therefore, its is important to note that no conclusive data has been presented thus far to support that geostrophic currents, similar to those which operate in modern day shorelines were important for the formation and deposition of hummocky cross-stratification or any other storm-related deposit.

Regardless of the exact nature of the depositional processes responsible for the formation of HCS, there is consensus among sedimentologists that HCS forms below fairweather wave base. However, bathymetric conditions for the formation of hummocky cross-stratification are variable. Harms *et al.* (1975) estimated that HCS formed in water depths ranging from 5 to 30 m, whereas Hunter and Clifton (1982) indicated that hummocky cross-stratification could form in depth as

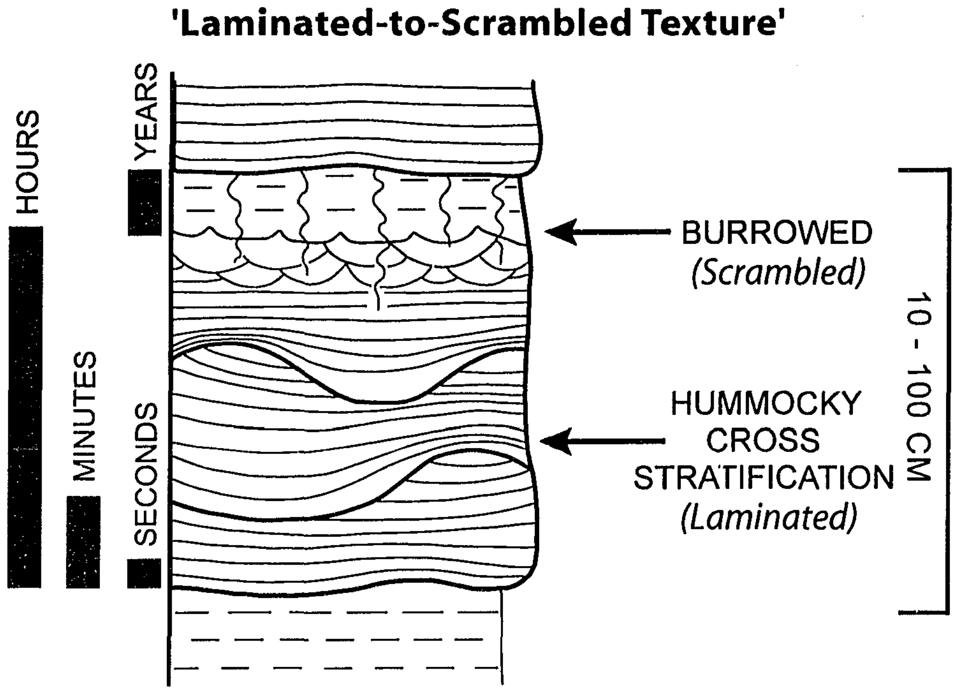


Figure 2.9 Schematic model of a storm deposited tempestite displaying the interstratified, scour-based hummocky cross-stratified sandstone beds and the gradationally based burrowed silty shales. Note the temporal significance of the storm deposited HCS beds and the interstratified fairweather burrowed shales. The storm deposited sandstone beds are deposited during seconds to hours and the interstratified mudstones represent periods of months to years (modified after Dott and Bougeois, 1982; Dott, 1983).

shallow as 2 m. Two main types of hummocky cross-stratification were identified by Duke (1985) including: 1) interbedded lithologies, where HCS sandstone beds are interbedded with bioturbated mudstone, siltstone, and 2) amalgamated sandstone, which consists of many HCS beds in erosional contact (Figs. 2.9; 2.10). Both types of hummocky cross-stratification are identified from the Kamik Formation especially within Facies C. The variation in the type and abundance of hummocky cross-stratification identified within Facies C is attributed to varying degrees of storm frequency and dominance that affected the delta front/shoreface successions.

The first type of HCS, as described by Duke (1985), is the most common type of hummocky cross-stratification observed in Facies C, especially in the lower portions of this facies. It essentially represents the "laminated-to-scrambled" or "lam-scram" texture described by Howard (1972) in Facies B and illustrated in Figure 2.9. Lam-scram texture records a transition from quiescent low energy, fairweather conditions of mudstone deposition to abrupt and rapid deposition of sandstone tempestite beds marked by a basal scoured surface. These tempestite deposits are deposited during storm events when storm wave base interacts with the substrate and reflect periods of higher energy storm conditions interrupting fairweather quiescent periods of mud deposition. The sequence of tempestite deposition producing "lam-scram" texture, as outlined by Dott and Bourgeois (1982) and Dott (1983), is initiated with an erosional surface that may or may not display sole marks or a lag. During emplacement of a storm-generated tempestite deposition and erosion occur in the same process. Hummocky cross-stratified sandstone represents the peak storm energy and reworking of the sediment when storm wave base interacts with the substrate during large persistent storm events. The hummocky cross-stratified interval grades upward and is overlain by symmetrical and asymmetrical ripple laminae interpreted to reflect wave and combined-flow ripples (Harms *et al.,* 1975). These ripples are formed by the reworking of the tops of the storm beds during the waning stages of the storm surge following storm abatement (Dott and Bourgeois, 1982; Dott, 1983; MacEachern and Pemberton, 1992). Typically the rippled interval is overlain by a burrowed mudstone or shale, which represents the final waning stages of the storm event and a return to fairweather conditions. Oscillation ripples associated with fairweather conditions may also be present, but they are relatively uncommon features (MacEachern and Pemberton, 1992). This sequence is an idealized model and there are bound to be variations due to the following factors: relative sand supply, water depth, tidal range, frequency, duration and intensity of storms, and relative productivity of burrowing organisms (Dott and Bourgeois, 1982). The lack of reworking of these types of HCS beds by fairweather processes and the preservation of the interstratified fairweather mudstone beds is attributable to deposition above storm wave base but below fairweather wave base (minimum), where only large storms rework the sediment (Dott and Bourgeois, 1982; Walker and Plint, 1992).

The second type of HCS composed of amalgamated, wavy, low angle, parallel-laminated beds in erosional contact, is typically found in the middle to upper portions of Facies C. The origin of this type of HCS is similar to that of the first type however, this type varies from the idealized model of Dott and Bourgeois (1982) (Fig. 2.9) as the interstratified well-burrowed mudstone beds are absent due to removal by erosion during deposition of the overlying HCS sandstone bed (Fig. 2.10). This creates an erosionally amalgamated succession of very fine to fine-grained sandstone beds scoured with both scoured bases and tops (Fig. 2.11). These tempestite beds can also be mantled by wave and combined flow ripples, which form due to the reworking of these storm beds by the waning currents associated with storm currents or rare fairweather waves (Dott and Bourgeois, 1982; Duke, 1985). These amalgamated HCS beds are interpreted to have been deposited above fairweather wave base and their lack of reworking is attributable to the relatively high frequency of storm activity which prevents the deposit from being modified under fairweather processes. This facies displays a general upward increase in fine sand, the amalgamation of the hummocky cross-stratified beds and a decrease in the mudstone interbeds into the amalgamated swaley cross-stratified sandstones, which commonly characterize the upper portion of this facies. The observed increase in bed thickness and abundance in the upper portion of this facies is interpreted to reflect increasingly shallow water conditions in the lower shoreface. MacEachern and Pemberton (1992) recognized that this general trend corresponds to a combination of enhanced depth of scouring due to water shoaling depth and storm frequency. Amalgamated HCS, and the features discussed herein,

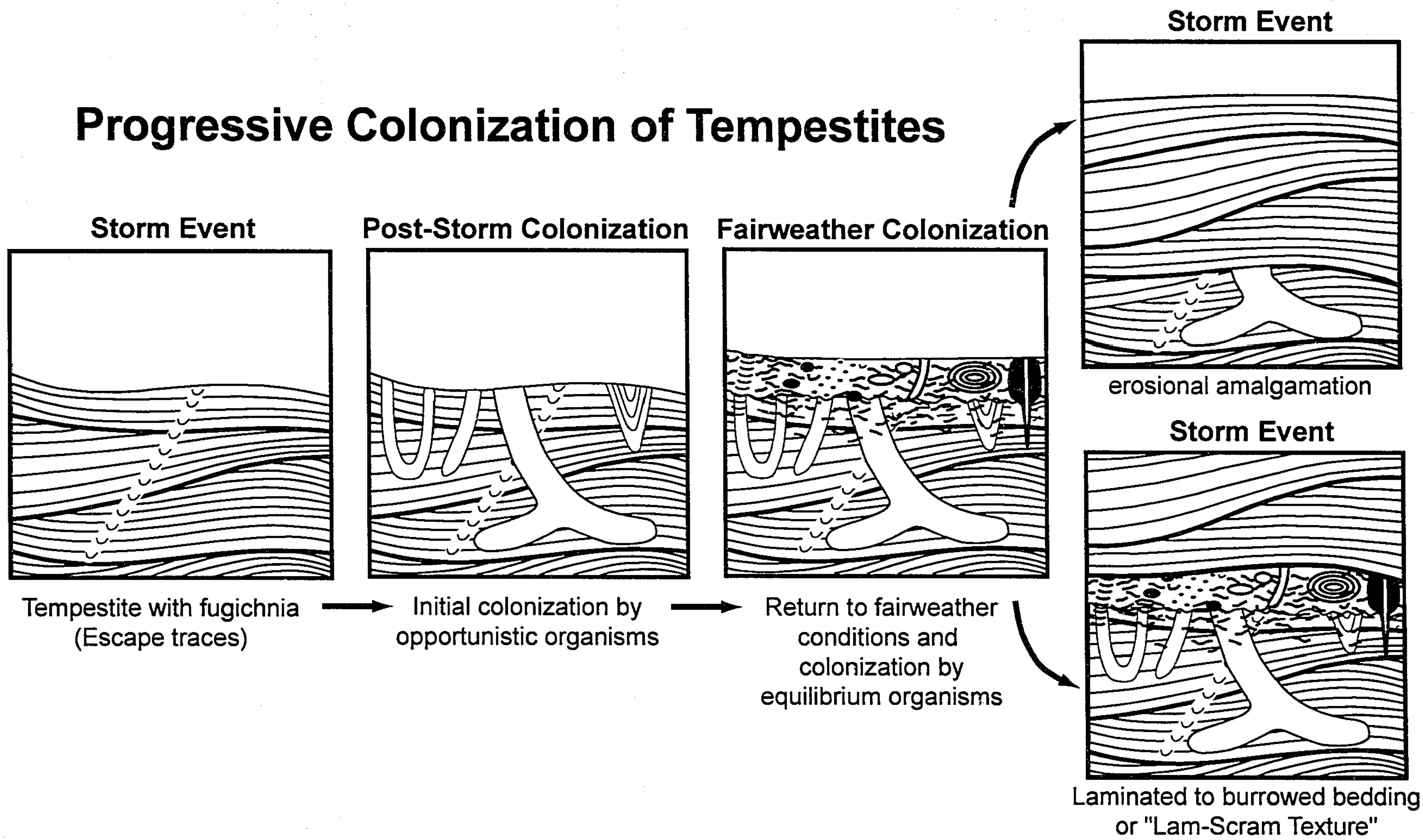


Figure 2.10 Sedimentological and ichnological model of a storm deposited tempestite. Note the interstratified scour-based hummocky cross-stratified sandstone beds and the gradationally based burrowed silty shales. Colonization of a tempestite and preservation potential of the assemblages depends upon the degree of erosional amalgamation of successive storm beds (modified after Dott, 1983 and Pemberton and MacEachern, 1997).

are interpreted to reflect a highly storm-dominated setting. Thus, hummocky cross-stratification present in Facies C developed both below and above fairweather wave base. The interbedded HCS and well-burrowed mudstone characterizing the lower portion of this facies is interpreted to have been deposited in the upper offshore to lower shoreface below fairweather wave base. The amalgamated HCS that characterizes the middle and upper portion of this facies is interpreted to reflect deposition strictly in the lower shoreface above fairweather wave base.

The upper portion of this succession is generally composed of erosionally amalgamated, low angle parallel-laminated sandstone interpreted as swaley cross-stratification (SCS). Intervals characterized by SCS were only identified within three cores (core #2 from the A-12 well, core #1 from the L-43 well and core #5 from the G-04 well). SCS was first introduced by Duke (1985) to describe a type of hummocky cross-stratification where the concave-upward swales are preferentially preserved due to down cutting by the overlying bed. Leckie and Walker (1982) formally defined SCS as a series of superimposed, concave-upward, shallow scours, essentially the same as the swales of HCS. SCS is distinguishable from trough cross-stratification (TCS) as the former generally has narrower and deeper troughs, and is associated with stratification that approximates angle-of-repose cross-stratification. An additional distinguishing feature of swaley cross-stratification is the tendency for dip of the internal laminae to progressively decrease upward toward the top of a bedset. In addition to occurring interstratified with each other, the trend from HCS to SCS in upper portions of Facies C suggests that the two bedforms are genetically linked. Both HCS and SCS are associated with large storms such as hurricanes and intense winter storms (Duke, 1985), which are further discussed in Chapter 3. The genetic a relationship between HCS and SCS has been recognized by previous authors, including Leckie and Walker (1982) who suggested that SCS originates in shallower water depths, thus is probably deposited above fairweather wave base. This rationale originated from the observation that in prograding sequences SCS typically overlies HCS (Leckie and Walker, 1982). In addition, the absence of interstratified mudstone beds within these amalgamated SCS beds also suggests deposition in a generally shallower, persistently wave agitated setting compared to that of the HCS dominated zones (Aigner and Reineck 1982; MacEachern *et al.*, 1992; Walker and Plint, 1992; Pemberton and MacEachern, 1997). These interpretations are consistent with the observations from the Kamik Formation where SCS dominated intervals are commonly found gradationally overlying HCS dominated intervals, suggesting a shallowing-upward succession. The inverse relationship where HCS is found overlying SCS, is also common in many cores from the Kamik Formation, indicating a deepening upward succession.

Duke (1985) suggested that a combination of shallower water depths with more energetic and frequent storms would allow for the amalgamation of storm deposits and the preservation of swaley cross-stratification. MacEachern and Pemberton (1992) noted that in shoreface settings under progressive storm-domination, the facies of lower shoreface are deposited basinward of fairweather wave base due to effective wave base being significantly lowered. A strongly storm-

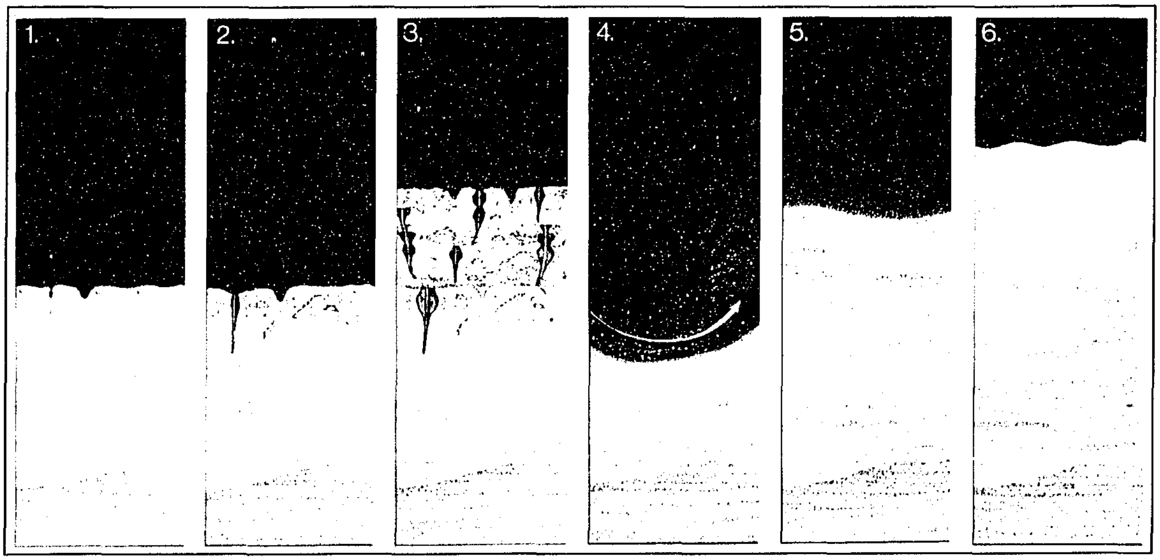


Figure 2.11 Schematic diagram illustrating the idealized formation of amalgamated low angle parallel laminated storm-bed successions formed due to frequent winter storms. The interstratifed fair-weather deposits, displayed in stage 2 and 3, are erosionally removed by subsequent storm scouring (stage 4). A combination of shallow water depth, frequency of storms, intensity of storms contribute to the amalgamation of storm beds. This sequence of events is speculated for the development of thick, amalgamated storm deposits of Facies C from the Kamik Formation (illustration courtesy of T.D.A Saunders).

dominated setting would act as an important influence in the deposition and preservation of swaley cross-stratification in a setting above fairweather wave base. In addition, persistent storm action can result in the near continuous reworking and removal of the record of fairweather deposition. Also, continuous deposition would favour the formation of swaley cross-stratification, over the interbedded type of hummocky cross-stratification, which forms under episodic sedimentation. Thus, Facies C was deposited in an environment dominated by storms. The presence of bivalve shell hashes and pebble lags marking the internal scour surfaces within SCS beds is due to removal and transport of this material from shallower settings such as the upper shoreface and foreshore, into the lower shoreface and offshore zones during high energy storm conditions. The soft-sediment deformation and convolute bedding may reflect dewatering structures, which occur in response to wave-induced liquefaction (Greenwood and Davidson-Arnott, 1979; Greenwood and Mittler, 1985; Pemberton and MacEachern, 1997). Convolute bedding could also reflect the rapid deposition of sediment typically associated with a deltaic environment (MacEachern *et al.*, 2005). The massive to structureless nature of some storm beds is attributable to rapid deposition and "freezing" of suspended sediment clouds from storm-generated currents (Hunter and Clifton, 1982). The preservation of little to no mud in these amalgamated intervals also suggests a generally shallower, persistently wave agitated environment than the underlying laminated to burrowed sandstones (MacEachern *et al.*, 1992; Walker and Plint, 1992; Pemberton and MacEachern, 1997).

The abundant, disseminated organic material found within these sediments is recognized as phytodetrital material deposited as a "phytodetrital pulses" (Rice *et al.*, 1986). The large amount of carbonaceous detritus is interpreted to reflect an abundant influx of terrestrial plant material into the water column, which is commonly associated with proximity to fluvial input (Reading and Collinson, 1996). The centimeter-scale, fining-upwards, graded beds interpreted to represent graded rhythmites. Graded rhythmites are interpreted to correspond to 'pulses' of sedimentation in deltaic environments generated by decaying sediment plumes entering the delta front from distributary channels. Deposition of phytodetrital material is associated with periods of short-lived, increased rates of fluvial discharge through rivers and distributaries, due to higher amounts of precipitation that typically accompany storms (Leithold, 1989; MacEachern, 1994). This phytodetrital material was transported into the delta front by storm-generated currents where it is rapidly deposited in the form of suspended fines from buoyant plumes or density underflows generated at the river mouth during times of high discharge (Coleman and Wright, 1975; Wright *et al.*, 1988). Leithold (1989) showed that short-lived, river flood events play an important role for the accumulation of fine sediments in the offshore, even within high-energy, storm-dominated coasts. Moslow and Pemberton (1988) recognized that the rapid deposition of fine material in the Harmon Member from Alberta was consistent with deposition in prodelta to distal delta front environments, representing accumulation in a storm/wave-dominated prodelta, rather than in the offshore of a strandline system.

The coalified pieces of wood material preserved within the mudstone interbeds are interpreted as rafted wood fragments that were transported to the lower shoreface/distal delta front during periods of heightened fluvial discharge. During storms, flood-induced transport carries woody material from adjacent highlands into fluvial channels, then into the delta front. Following storm abatement, the coalified pieces of wood sink and are incorporated into fairweather mudstone. The chert and quartz granules and pebbles found as stringers within mudstone intervals were likely emplaced by high-energy pulses associated with storm conditions. The presence of large siderite concretions within Facies C is reflective of incipient diagenesis catalyzed by the *in situ* breakdown of the abundant organic material present within this facies (Berner, 1980). Teodorovich (1961) also noted that siderite is common in littoral sediments that contain an abundance of organic material. MacEachern *et al.* (2005) recognized that abundant, early-formed siderite nodules are a common feature of deltaic settings and likely indicate the dilution of seawater, due to the mixing of fresh and marine waters, with a concomitant reduction in sulphate activity and low eH conditions reflecting low oxygenation of the substrate. Synaeresis cracks are interpreted to reflect fluctuations in salinity (Plummer and Gostin, 1981). These structures form in response to fresh water entering the delta front from distributary channels, especially during storm periods.

The ichnological assemblage of Facies C is characterized by a relatively high diversity suite of deposit feeding and grazing structures corresponding to a proximal expression of the *Cruziana* ichnofacies. Suspension feeding structures reflecting elements of the *Skolithos* ichnofacies. The assemblage is best described as a somewhat "stressed" *Cruziana* ichnofacies consistent with deposition in a distal delta front or equivalent lower shoreface equivalent setting, with considerable influence from storms. Where present, the fine-grained mudstone and silty mudstone interbeds contain a fairweather trace fossil suite, composed of deposit-feeding and grazing structures, which corresponds to a proximal expression of the *Cruziana* ichnofacies. In contrast, the storm-deposited sandstones contain a low diversity, opportunistic trace fossil assemblage that corresponds to the *Skolithos* ichnofacies. The degree of bioturbation and preservation of burrows is dependent on the amount of erosion of the interstratified, fairweather deposits and the degree of storm bed amalgamation.

Within mudstone intervals of Facies C, the *Cruziana* ichnofacies is represented by structures such as *Rhizocorallium*, *Rosselia*, *Teichichnus*, *Asterosoma*, *Cylindrichnus*, *Planolites*, *Thalassinoides*, *Chondrites*, *Schaubcylindrichnus*, and *Macaronichnus*. This assemblage commonly overlies the hummocky cross-stratified beds and represents a normal, fairweather, equilibrium behaviour. Examples of grazing/forging structures include *Helminthopsis*, *Scolicia* and *Phycosiphon*, which are generally found within discrete horizons. *Helminthopsis* is largely confined to the fairweather mudstone intervals, but is found within some of the storm-generated, tempestite beds. The sandstone-dominant tempestite beds contain an assemblage composed of an opportunistic suite of suspension-feeding structures including *Skolithos*, *Ophiomorpha*, *Conichnus*, *Diplocraterion*, *Siphonichnus*, *Arenicolites*, *Bergaueria* and fugichnia, the passive carnivore burrow *Palaeophycus* and the resting trace *Lockeia*. This suite is generally restricted to the upper portions of HCS beds. Phycosiphon is particularly abundant along the top of the tempestite beds, reflecting opportunistic (r-selected) behaviour immediately following storms (Goldring *et al.*, 1991). These burrows decrease in abundance upward into the overlying mudstone beds, suggesting that fairweather conditions were unsuitable for colonization. Fugichnia (escape traces) are also common at the top of tempestite beds and likely reflect rapid sedimentation rates. Escape traces record the attempt of organisms, entrained within or buried by the flow, to burrow up through the storm-generated tempestite in order to reach the new sediment-water interface (MacEachern and Pemberton 1992). The predominance of deposit-feeding structures, the low abundance of suspension-feeding ichnogenera and the diminutive size of ichnogenera is characteristic of a deltaic environments and has been attributed to increased water turbidity, reduced water salinity, fluctuating sedimentation rates, reduced oxygen levels (Moslow and Pemberton 1988; MacEachern and Pemberton, 1992; Gingras *et al.*, 1998; Coates and MacEachern, 1999; Coates, 2001; Bann and Fielding, 2004; MacEachern *et al.*, 2005). High water turbidity with suspended sediment is associated with fluvial discharge from distributary channels. An increased volume of suspended sediment within the water column is thought to interfere with the efficiency of organism's suspension feeding apparatus (Gingras *et al.*, 1998). Increase in water turbidity does not seem to affect the deposit-feeding and passive carnivores, as reflected by the dominance of these structures within this facies.

The large, thick-walled, tubular burrows found in the upper part of Facies C in the P-41 well are interpreted as a variant of the ichnogenus *Macaronichnus*. *Macaronichnus* generally reflects the behaviour of a specialized, grain-selective deposit-feeding organism and is normally associated with upper shoreface to foreshore environments or 'toe-of-the-beach' position within high-energy shorefaces (Saunders and Pemberton, 1986; Saunders *et al.*, 1994, Pemberton *et al.*, 2001). Most *Macaronichnus* burrows have fills that are biogenically sorted and linings that are composed of unsorted micaceous material. The variety of *Macaronichnus* found in Facies C have unsorted fills and sorted linings. Similar types of *Macaronichnus* were recognized by Bann and Fielding (2004) from the Permian of Australia, who interpreted these burrows as a composite burrow system in which some elements where lined and remained open to be passively infilled whereas other parts were actively filled through back-filling or as waste-stuffed chambers. It was also noted that the composite morphology of this burrow structure suggests that a time lag existed between formation of the burrow wall and the infill suggesting reprobing and possible collapse of branches (Bann and Fielding, 2004). The organism producing this burrow appears to be employing a similar depositing feeding behaviour as exemplified by *Macaronichnus,* however, it appears to inhabit a large, composite burrow structure reflecting a more complex feeding strategy. The cryptic bioturbated sandstones are interpreted to reflect the action of meiofauna that live between the sand grains, disrupt grain arrangement and obscure or obliterate physical structures (Saunders *et al.,* 1994; Pemberton *et al.,* 2001).

Rosselia and *Asterosoma* burrows are relatively abundant in Facies C. Similar abundances have also been recognized in storm-dominated open marine settings such as the lower shoreface and equivalent delta front environments (MacEachern and Pemberton, 1992; Bann *et al.*, 2004; Bann and Fielding, 2004). *Rosselia* and *Asterosoma* reflect the ability of the trace making organism to exploit its own deposits, rather than relying on deposited food in the substrate and are excellent indicators of storm-dominated, shallow marine settings, such as lower shoreface and comparable delta front environments (Howard and Frey 1984; Pemberton and Frey 1984; Saunders and Pemberton 1986; MacEachern and Pemberton, 1992; Pemberton *et al.*, 1992; Saunders *et al.*, 1994). *Rosselia* burrows are characterized by thickly lined tubes and their mud bulbs are commonly truncated by storm beds where they are commonly found detached nearby. *Rosselia* will commonly migrate upward, indicating the organism was able to readjust its structure in response to shifting sediment-water interfaces. This was interpreted to reflect the persistent shoaling of waves above fairweather wave base displacing suspended mud offshore while keeping food particles just above the shifting sand at the sediment-water interface (MacEachern and Pemberton, 1992).

The *Cruziana* and *Skolithos* assemblages found within Facies C are interpreted to represent pre- and post-storm assemblages. The pre-storm or fairweather assemblage is composed of mainly deposit-feeding structures characteristic of the *Cruziana* ichnofacies. The moderate to high degree of bioturbation is due to low rates of deposition and is mainly dependent upon the amount of time available for the fairweather assemblage to bioturbate. The post-storm assemblage is composed primarily of vertical burrows of the *Skolithos* ichnofacies reflecting suspension-feeding behaviours. This ichnological suite reflects an opportunistic assemblage formed within the storm deposited sandstones following storm abatement. The paucity in burrowing, especially within the sandstone beds, is interpreted to be a preservational bias reflecting high frequency and high intensity of erosionally amalgamated storm beds. Storms events tend to uproot, destroy and/or bury resident (fairweather) benthic communities (Pemberton and Frey, 1984; Frey, 1990; Pemberton *et al.*, 1992). The presence of unburrowed SCS-dominated intervals is attributable to high intensity and frequency storm events that do not allow organisms sufficient time to colonize the substrate. If these structures did exist, they were removed by subsequent erosion. For a further discussion of the ichnology of storm deposits, including r-selected and k-selected behaviours, the reader is referred to Pemberton *et al.* (1992).

Facies D - Medium to high angle, cross-stratified, sporadically burrowed, fine to medium-grained sandstone

Lithology and Sedimentary Features:

Facies D consists of a sporadically burrowed, light grey to buff, upper fine to lower medium-grained, well-sorted sandstone, with minor mudstone interbeds (Figs. 2.12A-D; 2.13A,D). Sedimentary structures include low angle (<5°) parallel lamination, medium to high angle (ranging from <10° to 15° up to 27°) cross-bedding and planar tabular cross-bedding, with some intervals massively bedded. Symmetrical and asymmetrical ripple laminations are present, and ripple structures commonly mantle the tops of individual bedsets (Figs. 2.12B,C). Soft-sediment deformation structures, including ball and pillow structures (Fig. 2.13B) and convolute bedding are also present. Similar to Facies C, the sandstone beds of this facies also display centimeter-scale, fining-upward, graded beds capped by the finely disseminated organic and micaceous material (Fig. 2.13A). Beds and bedsets with Facies D range in thickness from 5 to 25 cm and commonly erosionally amalgamated with numerous internal scour surfaces (Fig. 2.13D).

Some of these scour surfaces are marked by mudstone rip-up clasts and pebble stringers (Fig. 2.12A). Certain intervals exhibit centimeter-scale, fining-upward, graded beds generally capped by finely comminuted organic detritus (Fig. 2.13A). This organic detritus is termed 'coffee ground' material consisting of both phytodetrital material (ground up terrestrial plant material) and micaceous material (biotite and muscovite) and commonly highlights the laminae and bedding within many of the intervals. Interstratified within this facies are relatively common, centimeter to decimeter-scale, unburrowed, sharp-based and sharp-topped, massively bedded to structureless mudstone beds (Fig. 2.13C). These mudstone beds average 2.5 to 10 cm, and the upper and lower contacts are commonly pyritized. The sandstones contain rare coalified wood fragments, rare siderite concretions and disseminated pyrite nodules. A few of the sandstones beds contain relatively high concentrations of the heavy mineral garnet. Grain size, percentage of sand and bed thickness increases upward within this facies.

Facies D is identified in four cored intervals including core #2 from E-21, core #1 and #2 from L-43, core #1 from the P-41 well, and core #1 from the F-09 well. The lower contact is generally sharp to undulatory with Facies C and is marked by an increase in sandstone content, grain size and bed thickness as well as a decrease in mudstone content and interbed thickness. In addition, the lower contact is highlighted by a change in the dominant ichnogenera. Facies D typically occurs at the top of coarsening upward successions composed of Facies A, B, C and E. This facies is commonly stratigraphically overlain by Facies F and I and rarely by Facies G. The upper contact is generally gradational, however it can be sharp to undulatory with Facies I and is commonly scoured into by Facies G. Facies D varies in thickness from 3.39 m up to 6.05 m.

Ichnological Characteristics:

Trace fossils within Facies D are sporadically distributed but locally common and rarely abundant. The diversity is characteristically low. The main ichnological elements represent the activity of suspension-feeding organisms corresponding to the *Skolithos* ichnofacies with contributions of deposit-feeding structures from the *Macaronichnus* assemblage. Identified trace fossils include *Ophiomorpha* (Figs. 2.14A; 2.15C,D), *Conichnus* (Fig. 2.14B), *Cylindrichnus* (Fig. 2.14D), *Macaronichnus* (Figs. 2.15A,D), *Rhizocorallium* (Fig. 2.15B), *Teichichnus*, fugichnia (escape structures) (Fig. 2.14C) and cryptic bioturbation (Fig. 2.13A). The majority of biogenic structures are vertical or deeply penetrating burrows and heavily lined such as *Ophiomorpha nodosa* and *Conichnus conichnus* (Figs. 2.14A,B). *Macaronichnus segregatis* is particularly abundant in some intervals, however where it is relatively widely spaced (Figs. 2.15A,D).

Interpretation and Discussion: Proximal Delta Front (Upper Shoreface)

Medium to high angle cross-stratified, sporadically burrowed, fine to medium-grained sandstone of Facies D is interpreted as the moderately storm-dominated proximal delta front of a wave-dominated delta. The proximal delta front occurs in the equivalent position of the upper shoreface zone and is situated in the high-energy build-up and surf zone, and lies landward of the breaker zone (Clifton *et al.*, 1971; Davidson-Arnott and Greenwood, 1976; Hunter *et al.*, 1979; Greenwood and Mittler, 1985). The delta front is a transitional portion of a delta between the fluvial environments of the distributary channels and the marine environment of the prodelta (Coleman and Prior, 1980; Suter, 1994). The delta front merges seaward into the prodelta and distal delta front and landward into the lower delta plain. This zone overlies the lower shoreface and is situated well above fairweather wave base with its landward margin corresponding to the low tide mark. The upper shoreface zone

Figure 2.12

Facies D - Medium to high angle cross-stratified, sporadically burrowed, fine to medium-grained sandstone

A. Medium-grained, well-sorted sandstone displaying low to high angle (~5° up to 27°) cross-bedding interpreted as trough cross-bedding (TCS). The distinguishing feature of trough cross-bedding is the increasing foreset angle toward the top of the bed. Note the sharp contact marked by shale rip-up clasts between the cross-bedded sandstone and the overlying horizontal bedded interval (Parsons P-41, Core #1, depth ~2973.27 m).

B. Well-sorted, medium-grained sandstone displaying an oblique cut through a wave ripple. The laminae are marked by phytodetrital material (Parsons P-41, Core #1, depth ~2963.41 m).

C. Low angle, cross-bedded, well-sorted, fine-grained sandstone displaying an excellent example of a symmetrical wave ripple (Parsons P-41, Core #1, depth ~2964.12 m).

D. Medium-grained, well-sorted sandstone displaying medium to high angle (~10° up to 27°) cross-bedding interpreted as trough cross-bedding (TCS). Note the sharp contact between the cross-bedded sandstones below and the horizontal bedded interval above marked by shale rip-up clasts (Parsons P-41, Core #1, depth ~2973.27 m).

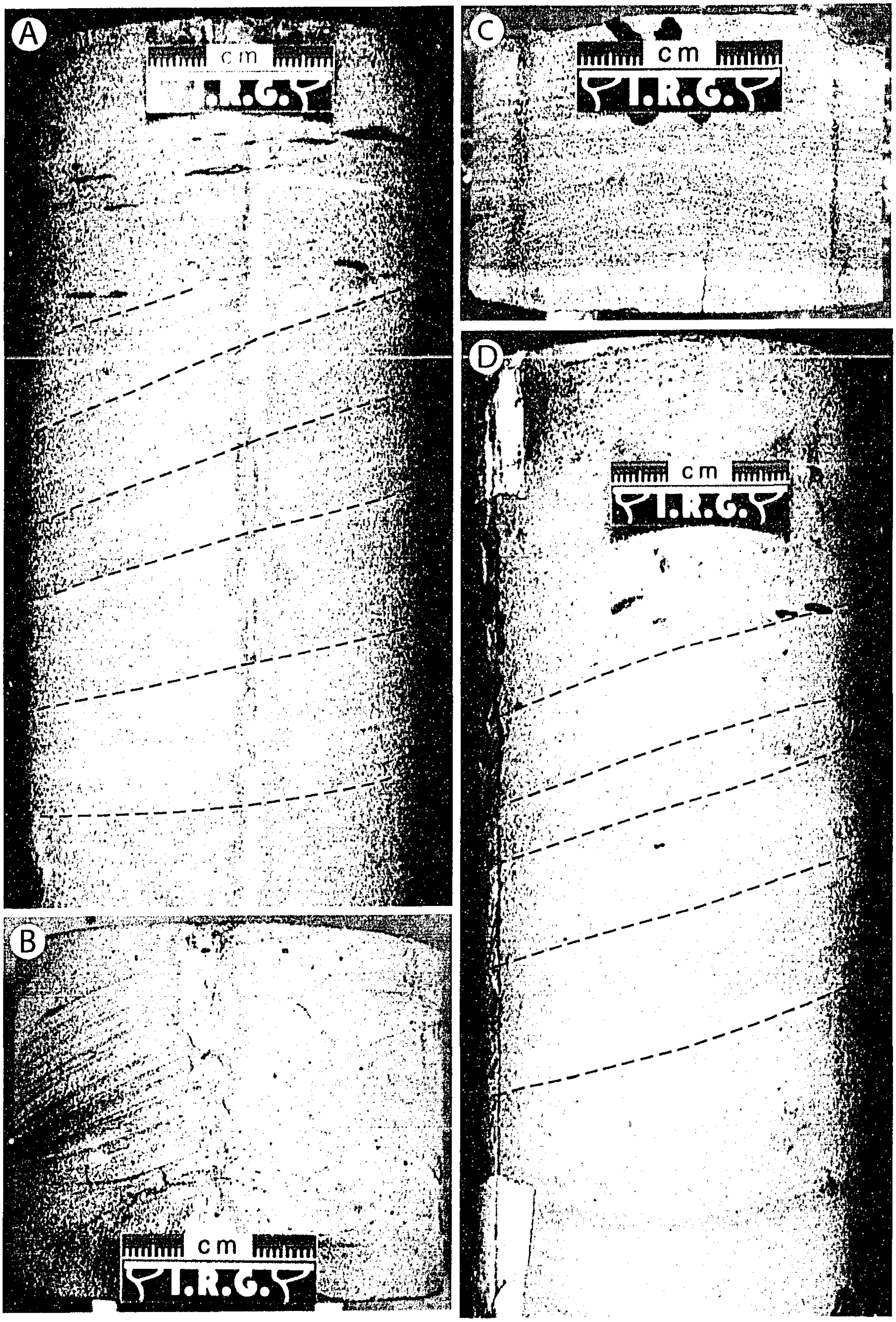

A
B
C
D
cm
cm
cm
cm

Figure 2.13

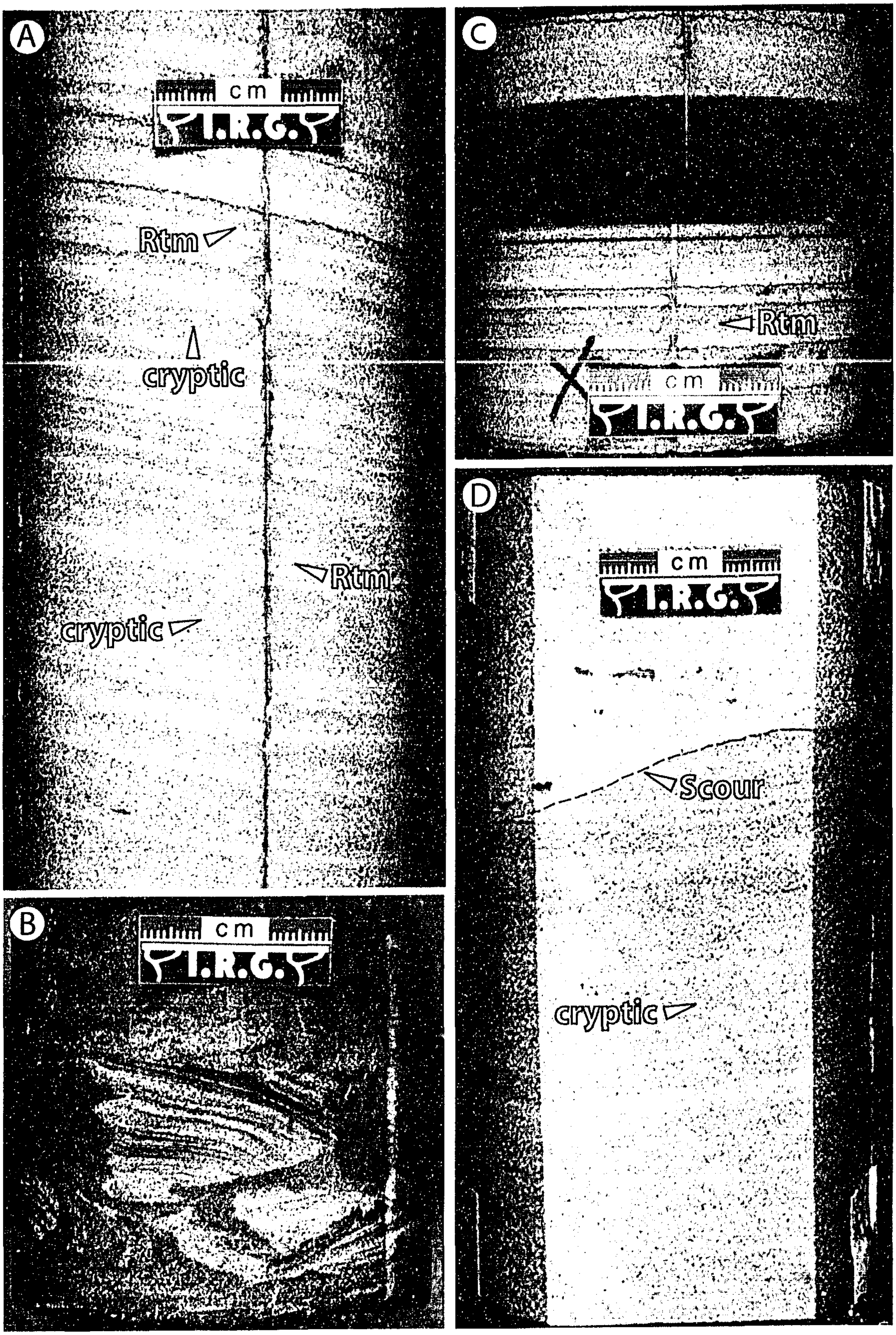
Facies D - Medium to high angle cross-stratified, sporadically burrowed, fine to medium-grained sandstone

A. Fine-grained, well-sorted sandstone displaying low angle (~5° to 15°) cross-bedding with multiple centimeter-scale normally graded beds interpreted as graded rhythmites (Rtm). Note organic rich laminae capping the tops of these sandstone beds. These sands are pervasively cryptic bioturbated (cryptic) by meiofauna imparting a fuzzy nature to all the sandstone beds and laminae (Parsons L-43, Core #1, depth ~2965.24 m).

B. Mudstone displaying ball and pillow structures. The contorted sandstone beds appear to be loaded cast ripples. Note the organic detritus within the sandstone beds (Parsons L-43, Core #1, depth ~2977.13 m).

C. Hyperpycnal mudstone bed within a fine-grained sandstone displaying near horizontal bedding with multiple centimeter-scale graded rhythmites (Rtm). Note the sharp bottom and top to these mudstone beds, characteristic of hyperpycnal mud deposition in the delta front. The graded rhythmites are capped by thin laminae of organic detritus (phytodetrital material) (Parsons L-43, Core #1, depth ~2965.66 m).

D. Fine-grained, well-sorted sandstone displaying a storm-generated scour surface marked by the truncation of underlying low angle (~5° to 10°) cross-bedding. Note small shale rip-up clasts in the overlying sandstone associated with the scour surface. These sands are pervasively cryptic bioturbated (cryptic) by meiofauna imparting a fuzzy nature to all the sandstone beds and laminae (Parsons L-43, Core #1, depth ~2966.58 m).



A
cm
T.R.G.
Rtm
cryptic
Rtm
cryptic
B
cm
T.R.G.
C
Rtm
cm
T.R.G.
D
cm
T.R.G.
Scour
cryptic

Figure 2.14

Facies D - Medium to high angle cross-stratified, sporadically burrowed, fine to medium-grained sandstone

A. Medium-grained, well-sorted sandstone displaying horizontal to low angle (~5°) cross-bedding with a large *Ophiomorpha* (Oph). Note that the upper wall of the burrow lining is thicker than the basal wall (Parsons L-43, Core #1, depth ~2781.98 m).

B. Well-sorted, medium-grained sandstone with a *Conichnus* (Con) (Parsons P-41, Core #1, depth ~2972.45 m).

C. Well-sorted, medium-grained, low angle cross-bedded sandstone with a large fugichnia (Fg) representing the escape trace of an infaunal organism. The trace making organism was probably a bivalve (Parsons P-41, Core #1, depth ~2973.02 m).

D. Well-sorted, medium-grained sandstone with *Cylindrichnus* (Cy). Note the inclined nature and multiple concentric mud linings (Parsons P-41, Core #1, depth ~2973.11 m).

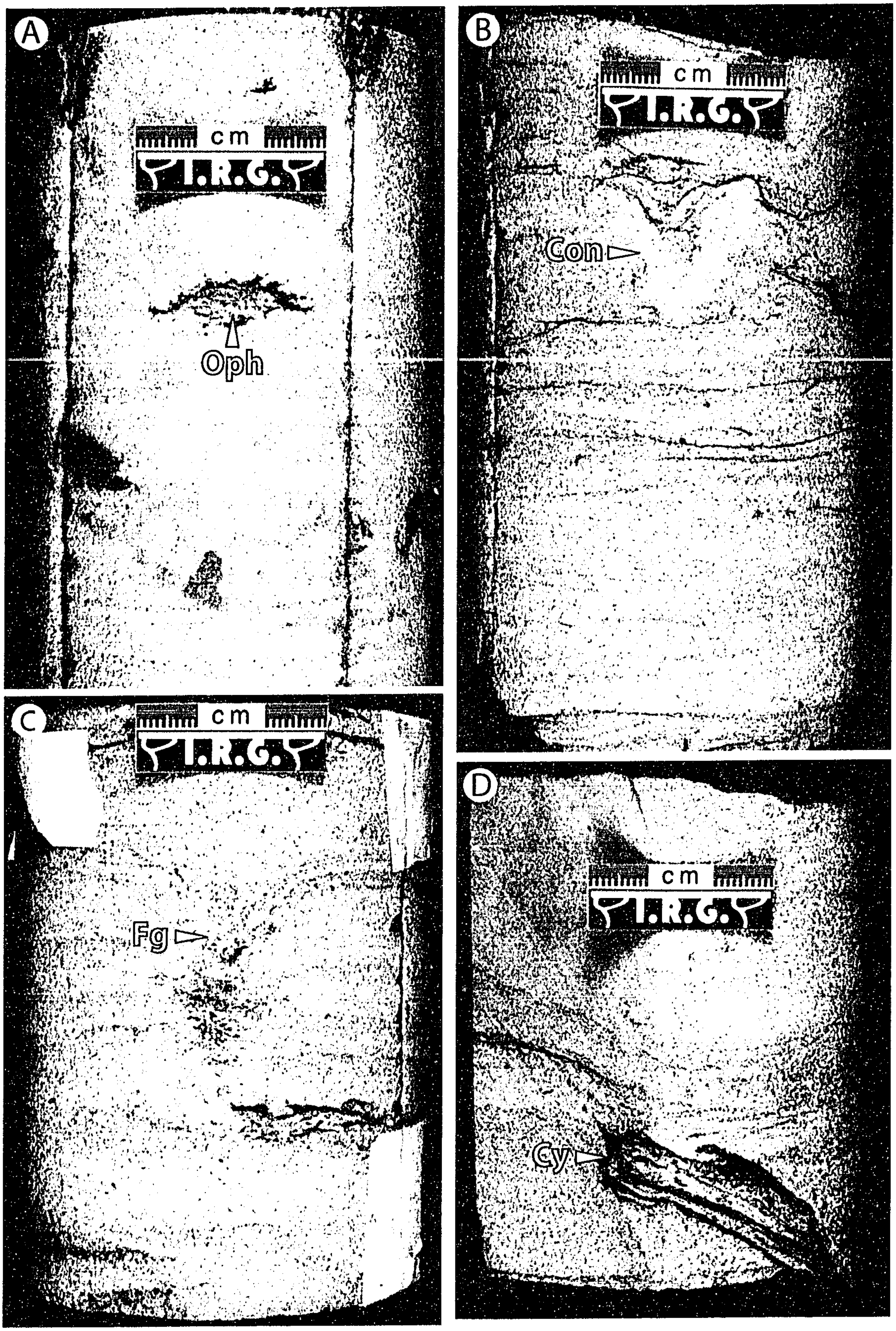

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Figure 2.15

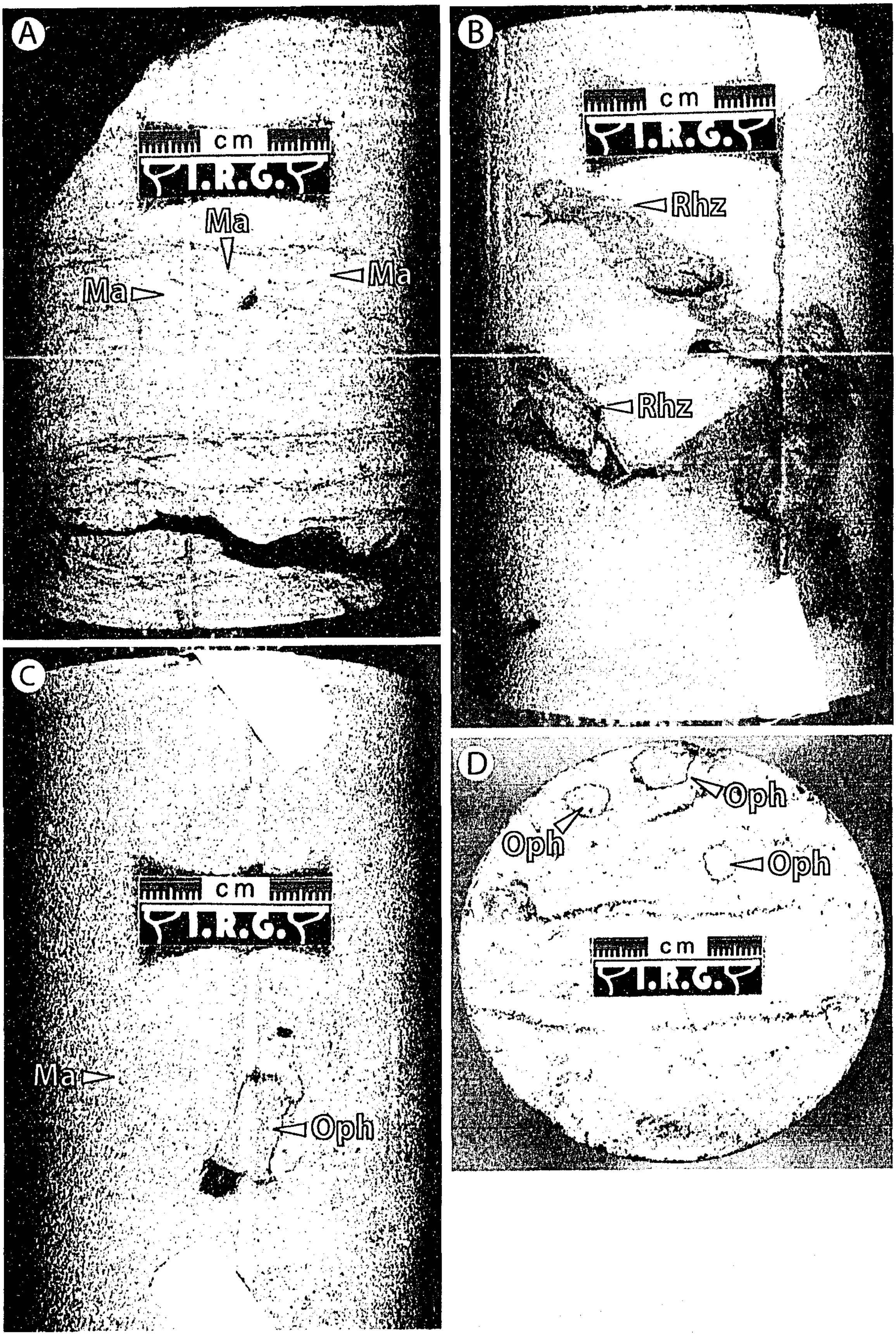
Facies D - Medium to high angle cross-stratified, sporadically burrowed, fine to medium-grained sandstone

A. Medium-grained, well-sorted sandstone displaying the trace fossil *Macaronichnus* (Ma). Note the clean quartz rich fills with the exterior of the burrows enriched in micaceous material (Parsons L-43, Core #1, depth ~2962.27 m).

B. Low angle, parallel-laminated, fine-grained sandstone with large *Rhizocorallium* (Rhz). Note the vertical migration of the *Rhizocorallium* tubes (Parsons L-43, Core #1, depth ~2782.79 m).

C. Apparently structureless, medium-grained, well-sorted sandstone displaying a large *Macaronichnus* (Ma) with a thick lining. This burrow displays an unsorted fill with a sorted lining, a characteristic which deviates from the "normal" appearance of *Macaronichnus* where burrow fills are biogenically sorted and the linings are composed of unsorted material. Note the *Ophiomorpha* (Oph) with a think wall lining (Parsons L-43, Core #1, depth ~2780.78 m).

D. Bedding plane view of a medium-grained, well-sorted sandstone displaying a transverse section through an *Ophiomorpha* (Oph). Note the distinct agglutinated pelletoidal lining of the burrow margins (Parsons L-43, Core #1, depth ~2780.62 m).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is affected by wave- and storm-driven currents paralleling the shoreline (i.e. longshore drift), which interact with shore-normal currents generated by translatory flow associated with plunging waves (MacEachern and Pemberton, 1992). This interaction produces multidirectional, sinuous-crested subaqueous dunes.

Sedimentary structures within this facies suggest deposition in a high-energy setting characterized by scouring, suspension and traction processes, gravity-induced processes of deposition and oscillatory flow produced by waves. The medium to high angle cross-bedding displaying the steepening-upward angles is interpreted to represent vertical sections cut through trough cross-bedded bedforms. Trough cross-bedding (TCS) is interpreted to have formed by the migration of three-dimensional (3D) subaqueous dunes or "megaripples" (Harms *et al.,* 1975) in a high-energy shoreface setting dominated by oscillatory flow. TCS consists of elongate erosional scours filled with curved laminae, with the long axes of the scours parallel to the local flow directions (Harms *et al.,* 1975). Depending on the cut through the trough (perpendicular or parallel to flow direction) the erosional surfaces and laminae appear as a series of upward-truncated arcs. Observed dips from core range from 15° to 27°, well within the 25° to 30° range noted by Harms *et al.* (1975). Characteristically, trough cross-beds are a few decimeters thick, with widths ranging from less than a meter to more than 4 m (Harms *et al.,* 1975). Davidson-Arnott and Greenwood (1976), Roy *et al.* (1980), Reinson (1984) and Thom *et al.* (1986) noted that upper shoreface deposits are typically dominated by 15 to 45 cm thick, multidirectional trough cross-stratified beds intercalated with low angle, bi-directional, planar cross-bedded sets. The minimum water depths required to form trough cross-bedding is approximately twice the thickness of the individual sets (Harms *et al.,* 1975, 1982). However, this bedform can form under deeper conditions and therefore cannot be used to estimate water depths. The symmetrical ripples found capping the individual bedsets are interpreted as wave and to a lesser extent, combined-flow ripples produced by oscillatory flow from wave-generated processes and minor current processes (Harms *et al.,* 1975).

The presence of wave and current ripples, along with soft-sediment deformation structures, suggests the depositional environment was characterized by a dynamic interplay of wave processes and current action, attributable to close proximity to distributary channels (Reading and Collinson, 1996). The sediments comprising the proximal delta front or upper shoreface equivalent are continuously reworked by wave processes resulting in re-ordering, improved sorting, and concomitant tighter packing of the sediment grains (Coleman and Prior, 1982). The multiple scour surfaces are due to erosional amalgamation of beds via storm activity. In addition, these sediments contain little to no preserved mud in these sediments. This suggests a generally shallower, more persistently wave agitated environment than the underlying laminated to burrowed sandstones of Facies C which is consistent with deposition in a relatively high-energy proximal delta front or equivalent upper shoreface environment (MacEachern *et al.*, 1992; Walker and Plint, 1992; Pemberton and MacEachern, 1997). The low proportion of penecontemporaneous deformation structures (as compared to Facies N and O) is due to an increase in overall wave-energy, which was effective at winnowing these sediments, and increased distance and limited influence from fluvial point sources. Soft-sediment deformation structures such as ball and pillow structures are indicative of the rapid deposition. Ball and pillow structures form when sand deposited onto unstable, soupy muddy sediments, sinks into the underlying mud bed (Reineck and Singh, 1980). Such soft and unstable substrates cannot disperse fluids quickly enough, thus allowing deformation during early compaction (Coleman and Prior, 1982; Bhattacharya and Walker, 1991; Reading, 1996). The rapid deposition of sediment is a characteristic feature of deltaic settings where active deposition takes place proximal to the fluvial point source (i.e. a distributary channel) with high fluvial discharge (Coleman and Prior, 1982; Wright, 1985).

The coalified wood fragments and phytodetrital material contained within these sandstones are interpreted to represent rafted wood fragments eroded in the fluvial system and carried in suspension, likely during periods of high fluvial discharge, especially during storm periods. The centimeter scale, fining-upward beds capped by micaceous and macerated organic material is interpreted to represent graded rhythmites deposited in the delta front. Graded rhythmites correspond to "pulses" of sedimentation in deltaic environments generated by decaying sediment plumes entering the delta front from distributary channels. The graded nature indicates that traction and suspension processes were present during deposition, a feature typical of deltaic environments. Moslow and Pemberton (1988) recognized similar rhythmically bedded, laminated siltstones in distal delta front facies, interpreting these deposits to indicate cyclicity in sedimentation patterns reflecting seasonal fluctuations in sediment supply or fluvial discharge. The organic-rich laminae and beds cap the graded rhythmites reflect deposition from "phytodetrital pulses". These phytodetrital pulses originate from either the settling of material carried out into the delta front in suspension as buoyant plumes or from density underflows generated at the river mouth during periods of high fluvial discharge (Coleman, 1981; Wright *et al.,* 1988). The presence of graded rhythmites and phytodetrital material suggests that sediment plumes from distributary channels significantly influence deposition in the delta front.

The interstratified, centimeter to decimeter thick, unburrowed, massive mudstone beds displaying sharp tops and sharp bases, represent hyperpycnal or fluid muds. Hyperpycnal mudstone beds, deposited as sediment-laden gravity flows (density currents), migrate along the substrate-water interface in the delta front environment (Bates, 1953; Bhattacharya and Walker, 1992). Hyperpycnal deposits are common in deltaic settings where sediment-laden river water of a higher density flows beneath basin water; generating a vertically oriented, plane jet flow (Bates, 1953). These types of flows are generated at river mouths in marine basins during times of high fluvial discharge, typically during or closely following storm events, where the dense, sediment-laden river outflow enters the delta front and flows down along the bottom of the delta front slope as density underflows (Wright *et al.*, 1988; Mulder and Syvitsky, 1995). The absence of internally graded bedding indicates that deposition was not from suspension fallout, thus differentiating them from hypopycnal deposits. Hyperpycnal muds are characteristic of riverine influenced delta fronts and are commonly identified in ancient deltaic settings (S.G. Pemberton pers. comm., 2004). The dark color of these mudstones is attributed to the high organic content. These interspersed mudstone beds are organic-rich and pyrite nodules commonly found at the contacts reflect redox reactions between the reduced, organic rich mudstone and oxygenated sandstone. The lack of biogenic structures may be due to periods of dysoxic to anoxic conditions, thereby limiting or even precluding the establishment of a benthic community (MacEachern *et al.*, 2005).

The trace fossil assemblage of Facies D is dominated by elements corresponding to *Skolithos* ichnofacies with minor contributions from the *Macaronichnus* assemblage. Biogenic features are somewhat similar to that found in Facies C, however there is a marked decrease in the presence of *Cruziana* elements coupled with the total domination of elements from the *Skolithos* ichnofacies. Elements from the *Skolithos* ichnofacies include *Ophiomorpha* and *Conichnus*. The high-energy nature and overall harsh environmental conditions associated with the upper shoreface environment is responsible for the overall scarcity of animals inhabiting and creating domiciles in this setting (MacEachern and Pemberton, 1992). Conditions in the upper shoreface, including continually shifting, sandy substrates and migrating bedforms, present a major problem to endobenthic organisms and inhibit the construction of permanent domiciles (Howard, 1972; Saunders, 1989). Any biogenic structures that are established with this setting are subject to a low preservation potential and in general, only deeply penetrating or heavily lined structures are preserved (Howard and Frey, 1984). The relatively rare abundance of the deposit-feeding *Teichichnus* within these sediments reflects the opportunistic behaviour of an organism that can tolerate a wide range of environmental conditions.

The *Macaronichnus* present within this facies represent a unique biogenic element with significant environmental implications. *Macaronichnus* is typically associated with the very high-energy conditions of the upper shoreface environment. Saunders (1989) and Saunders and Pemberton (1990) noted that *Macaronichnus* is more common near the upper shoreface-foreshore contact, however, under more reflective shoreface states, which are characterized by steeper depositional profiles and coarser grain sizes, the *Macaronichnus* zone may occur lower down in the upper shoreface. *Macaronichnus* elements present within this facies are widely spaced and lack the density associated with "toe-of-the-beach" assemblage (Saunders *et al.*, 1994). A further discussion of the behaviour of *Macaronichnus* is presented in Facies E. Cryptic bioturbation is interpreted reflects the action of meiofauna that live between the sand grains; the organisms disrupt grain arrangement and obscure or obliterate physical structures (Saunders *et al.*, 1994; Pemberton *et al.*, 2001). The presence of multiple unburrowed intervals within this facies is attributable to intense and frequent storm events allowing insufficient time for organisms to colonize the substrate. If these structures did exist, they were removed by subsequent erosion.

Facies E - Horizontal to low angle, parallel-laminated, fine to medium-grained sandstone

Lithology and Sedimentary Features:

Facies E is composed of light to dark grey, upper fine to lower medium-grained sandstone that is generally quartz-rich, well-sorted and very well winnowed (Figs. 2.16A-D). Thin, centimeter-scale intervals of coarse-grained sandstone are common. White quartz grains and black mica/lithic grains which lend the sandstone a "salt and pepper" texture. Sedimentary structures include planar to low angle (2° to 5°) parallel lamination, asymmetrical and symmetrical ripple cross-lamination, and occasional massive bedding (Figs. 2.16B,C). Low angle parallel lamination occurs in bedsets ranging in thickness from 1 to 15 cm. Some intervals contain vertical coalified rootlets. This sandstone is also interbedded with and overlain by centimeter to decimeter thick coal beds.

Facies E is identified in a single cored interval from the C-23 well. Due to limited recovery and the poor condition of the core, the exact nature of the bounding contacts and relationships with other facies is unclear. However, this facies is assumed to be underlain by sediments similar to Facies E with the lower contact likely gradational. Both Facies D and E share a similar ichnological signature, which is used to interpret a genetic stratigraphic relationship between the two facies. The upper contact is sharp with overlying coaly mudstone and coal of Facies L. Facies E typically occurs at the top of coarsening upward successions composed of facies A, B, C and D and varies in thickness from 1.69 m up to 1.75 m.

Ichnological Characteristics:

Trace fossils within Facies E are sporadically distributed but relatively high in abundance. The diversity is characteristically low, however many intervals display a high degree of bioturbation. The main ichnological elements are deposit-feeding organisms of the *Macaronichnus* assemblage and rarer suspension-feeding organisms representing to the *Skolithos* ichnofacies. Only two ichnogenera were identified, *Macaronichnus segregatis* and vertical *Ophiomorpha*. There are several discrete zones containing *Macaronichnus segregatis*, which are generally 1 to 2 mm in diameter and dominantly horizontal in nature, although some burrows display a significant vertical component (Figs. 2.16A,B,D). The individual burrows generally do not cross-cut one another and where associated with primary sedimentary structures, they do not appear to disrupt the stratification. This relationship suggests that the burrows are in equilibrium with the bedforms. *Macaronichnus* within Facies E are found in dense concentrations and are more closely spaced than in Facies D. The *Ophiomorpha* within these sandstones are heavily lined and typically display inclined to vertical orientations (Fig. 2.16D). In addition there are 1 to 5 cm long, vertical, coalified rootlets penetrating downward into the *Macaronichnus*-burrowed zones (Fig. 2.16B). The root traces are wispy and sinuous in nature and commonly thin downward with very thin bifurcations

Figure 2.16

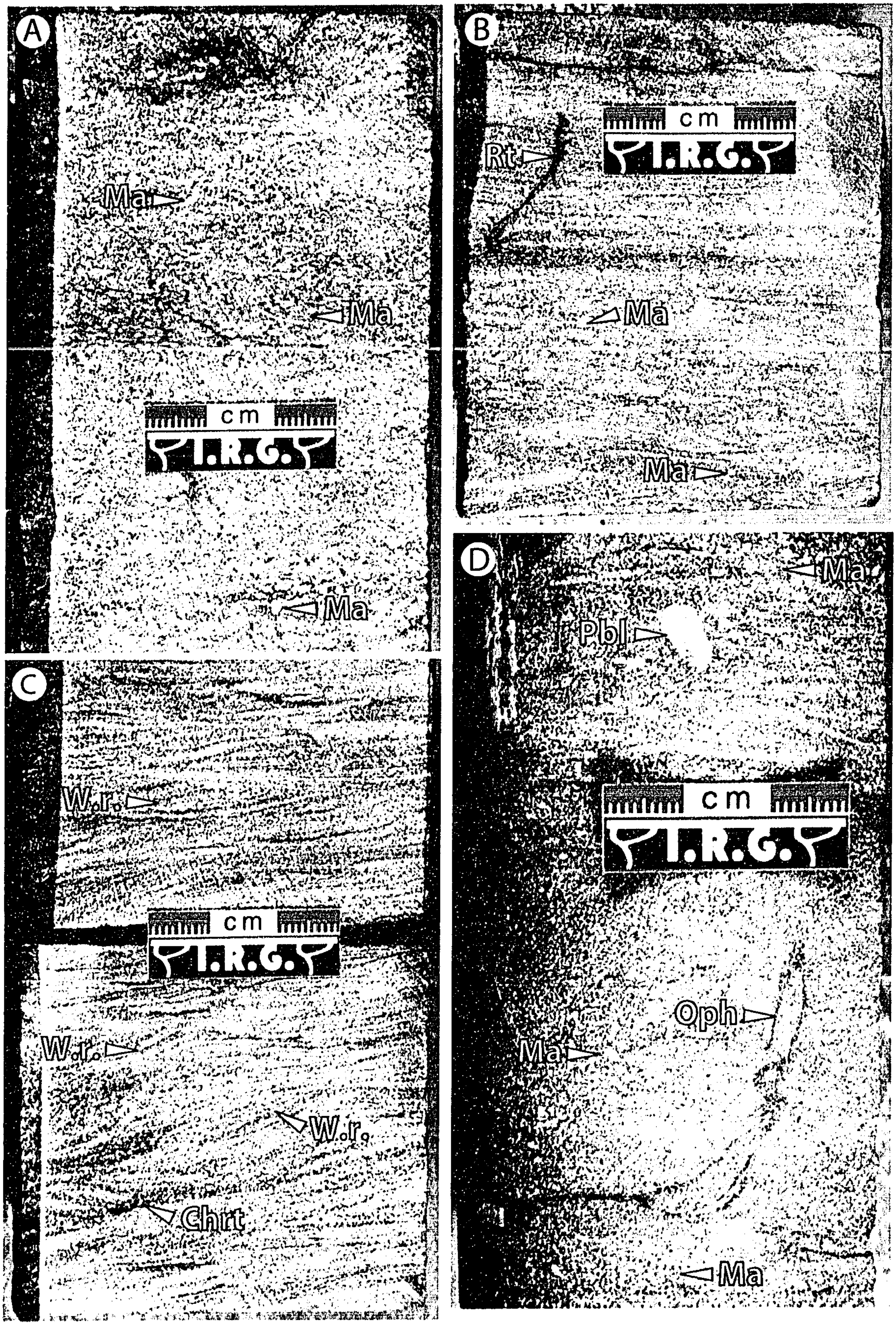
Facies E - Horizontal to low angle, parallel laminated, fine to medium-grained sandstone

A. Planar bedded, fine-grained sandstone thoroughly bioturbated by a monospecific assemblage of *Macaronichnus segregatis* (Ma). Note that even with the thoroughly burrowed nature of the sandstone by *Macaronichnus*, the sedimentary structures are still preserved. Note the subtle contact between the less burrowed sandstone below and the well-burrowed interval above (Wagnark C-23, Core #1, depth ~3783.46 m).

B. Fine-grained, horizontal to low angle (<5°), parallel laminated sandstone with a well-burrowed interval of *Macaronichnus segregatis* (Ma). Note the subtle contact between the *Macaronichnus*-burrowed, low angle cross-bedded sandstone below and the relatively unburrowed, horizontal to low angle (<5°), parallel-laminated sandstone interval above with a root (Rt) trace extending downwards (Wagnark C-23, Core #1, depth ~3784.62 m).

C. Wave-ripple laminated, fine-grained sandstone. These structure were identified as wave ripples (W.r) by the presence of internal dipping foresets and undulatory bases. In addition, some of the minor bedding planes are marked by thin granule lags composed of chert (Chrt) grains (Wagnark C-23, Core #1, depth ~3788.25 m).

D. Fine-grained, sandstone with a high abundance of *Macaronichnus segregatis* (Ma) and a single, inclined to vertical *Ophiomorpha* (Oph). The vertical character of *Ophiomorpha* is indicative of high-energy settings such as upper shoreface and foreshore environments. The lower zone pervasively bioturbated by *Macaronichnus* is overlain by an interval with horizontal to low angle (<5°) parallel laminated sandstone with a lower abundance of *Macaronichnus* and a subangular quartz pebble (Pbl). Note the subtle contact between the less burrowed sandstone below and the well-burrowed interval above (Wagnark C-23, Core #1, depth ~3782.95 m).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Chrt
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Oph
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Ma

at their terminations.

<u>Interpretation and Discussion:</u> Proximal Delta Front (Foreshore)

The light to dark grey, fine to medium-grained, low angle, parallel-laminated to ripple laminated sandstone of Facies E is interpreted as beach deposition in the foreshore environment. The foreshore is defined as the intertidal zone landward of the upper shoreface between low and high tide (Reineck and Singh 1980; MacEachern and Pemberton, 1992). Both wave swash and surf zone processes operate in the foreshore, with wave swash the dominant physical process (Clifton *et al.,* 1971). The foreshore characteristically has a highly mobile, non-cohesive substrate, which is subjected to very high-energy conditions (MacEachern and Pemberton, 1992). Distinct planar to low angle (2° to 5°) parallel lamination is a characteristic feature of foreshore and beach environments and is formed due to swash and backwash processes (Clifton, 1969; MacEachern and Pemberton, 1992). Stratification in the foreshore commonly referred to as swash-zone stratification, occurs as wedge-shaped sets with internal laminae parallel to subparallel to lower set contacts (MacEachern and Pemberton, 1992). This stratification reflects deposition from suspension clouds (Reineck and Singh 1980).

The limited nature of core makes identifying the characteristic, large, seaward-dipping foreshore wedge sets difficult due to the large scale of these bedforms. Therefore, the interpretation of these sedimentary features is based on the nature of stratification, grain size, overall depositional context and ichnology. The low amplitude asymmetrical and symmetrical ripple cross-lamination in some intervals represents current and wave ripples, respectively (Harms *et al.,* 1975). Wave and current ripples are commonly modified during the last phase of emergence by flowing water (Reineck and Singh 1980). Low angle wave ripples were noted by Saunders and Pemberton (1986) to be associated with 10 to 15 cm thick swash zone bed sets from foreshore deposits of the Appaloosa Sandstone, Bearpaw-Horseshoe Canyon Formation transition in Drumheller, Alberta. The effect of storm events on the upper shoreface and foreshore results in virtually no storm-induced deposition and mainly pronounced erosion where short-period waves cut ridge and runnel systems and plane off the accretionary profile of the beach-face (Kumar and Sanders 1976; Fox and Davis, 1978; Niedoroda *et al.,* 1984; Swift *et al.,* 1985). During storms sediment is eroded from the upper shoreface and foreshore and transported basinward into the lower to middle shoreface. Under prolonged fairweather conditions, much of the sediment removed during storms is returned to the beach in the form of longshore bars, which migrate landward and weld to beach in response to long-period waves (MacEachern and Pemberton, 1992). Some sediment is lost permanently when transported to the lower and middle shoreface as storm-induced tempestites.

Trace fossils in the foreshore display a low diversity and abundance, attributable to the predominant of harsh environmental conditions, including continuously shifting substrates and overall low preservation potential (Howard and Frey, 1984). Elements of the *Skolithos* ichnofacies, such as vertical *Ophiomorpha nodusa,* are deeply penetrating domiciles of organisms reflecting a suspension-feeding behaviour. The presence of discrete zones of abundant *Macaronichnus segregatis* is important to the interpretation of these sediments as foreshore deposits. The trace fossil *Macaronichnus* is characteristically associated with high-energy beach settings (Saunders and Pemberton, 1986; Saunders, 1989). *Macaronichnus segregatis* was originally described and defined by Clifton and Thompson (1978) and was attributed to the deposit-feeding behaviour of the opheliid polychaete *Ophelia limicina.* MacEachern and Pemberton (1992) also recognized that the modern polychaete *Euzonus mucronata,* also from the family *Ophelidae,* formed modern *Macaronichnus* burrows on Long Beach, Vancouver Island, Canada. The *Macaronichnus* tracemaker ingests sand grains in order to feed on the epigranular food source of bacteria and other grain attached organics, and after feeding on these food sources excrete the "cleansed" sand to form the fill of the burrow structure (MacEachern and Pemberton, 1992; Pemberton *et al.,* 2001). The polychaetes that make *Macaronichnus* preferentially ingest quartz and feldspar sand and avoid mica flakes and mafic grains. This biogenic preference for quartz and feldspar grains forms an unlined burrow of which the fill tends to be lighter in colour than the surrounding host sediment (MacEachern and Pemberton, 1992; Pemberton *et al.,* 2001).

The presence of *Macaronichnus* suggests a high-energy setting where oxygenated waters are circulated several meters into the sand, well below the reach of surface wave disturbance. Such settings are found within intertidal and innermost surf zones of the upper shoreface and foreshore of wave-exposed beaches, where high oxygen windows occur as permanent, hydrodynamically sustained features (MacEachern and Pemberton 1992; Pemberton *et al.,* 2001). The action of high energy waves pounding the beach filters tremendous volumes of dissolved and finely particulate organic matter through the sand, where it passes through the porous system of the beach and is mineralized at depth (MacLachlan *et al.,* 1985). These processes create a deep sand habitat in the upper shoreface and foreshore, which is very stable, predictable, and possesses a potentially abundant and constantly replenishing food supply (MacEachern and Pemberton, 1992; Pemberton *et al.,* 2001). The congregation of *Macaronichnus* in a narrow zone at the 'toe-of-the-beach' (Saunders *et al.,* 1994) is interpreted to reflect the exploitation of a zone of nutrient convergence (MacEachern and Pemberton, 1992). This zone of convergence occurs within a specific zone where nutrients are delivered from both the landward side, via low-tide drainage through the intertidal prism, and from the seaward side, through intense "pumping" in the innermost surf (MacEachern and Pemberton, 1992; Pemberton *et al.,* 2001).

Macaronichnus has also been identified in Lower Cretaceous boreal beaches from western Canada by Saunders (1989) and Saunders *et al.* (1994). These ancient examples of *Macaronichnus* were interpreted to reflect colonization by a macrofaunal deposit feeder, such as an opheliid or opheliid-like polychaete similar to those found in modern beaches (Clifton and Thompson, 1978; Saunders, 1989). The creation of a well-oxygenated zone in the foreshore and upper shoreface facilitates the exploitation of this unique, deep sand habitat by the *Macaronichnus* trace making organism. The swash activity on the intertidal part of the beach, causes large quantities of water to move through the beach sand (Riedl *et al.*, 1972), in which the redox-system of the sediments can be strongly influenced up to a depth of 4 m (Reineck and Singh, 1980). Saunders (1989) recognized that within the Appaloosa Sandstone from the Bearpaw-Horseshoe Canyon transition from Drumheller, Alberta, *Macaronichnus segregatis* is found preferentially concentrated within a narrow zone situated in the upper foreshore and persists downward in lower densities into the foreshore-upper shoreface transition, but never below this zone. However, MacEachern and Pemberton (1992) noted that in virtually all other shoreline cycles from the boreal Lower Cretaceous in Alberta, *Macaronichnus* display the opposite zonal relationship. In these sediments, the burrows generally occur in a toe-of-the-beach position with rare examples extending into the overlying swash-laminated foreshore sandstones. Overall, the presence of *Macaronichnus* is characteristic of higher energy shoreface environments where domiciles cannot be established because of shifting substrates and filter feeding is precluded because of intense wave action (MacEachern and Pemberton, 1992).

Facies F – Poorly sorted, scour-based, matrix supported conglomerate and medium to very coarse sandstone

Lithology and Sedimentary Features:

Facies F is comprised of a poorly-sorted, matrix-supported conglomerate with lower medium to upper very coarse-grained, quartzite sandstone (Figs. 2.17A-D). This conglomerate lag deposit is composed of laminated sandstone intraclasts (up to 6.0 cm in diameter) along with pebbles and cobbles composed of chert, quartz and other metamorphic and mafic material (Figs. 2.17B,D). This facies has a scoured basal contact with Facies D and is sharply overlain by the well-bioturbated mudstone of Facies B (Fig. 2.17A). In core from the P-41 well, this facies occurs at the top of a coarsening-upward succession and separates two coarsening-upward sequences. Facies F is only identified from core #1 of the P-41 well and is 6.5 cm thick.

Ichnological Characteristics:

No trace fossils were identified within this facies.

Interpretation and Discussion: **Transgressive Lag Deposit**

The poorly sorted, matrix-supported conglomerate of Facies F is interpreted as a lag deposit formed during transgressive erosion of the delta front or shoreface. Similar lag deposits were interpreted by Pemberton and MacEachern (1995) as the proximal facies of transgressive erosion, or ravinement

Figure 2.17

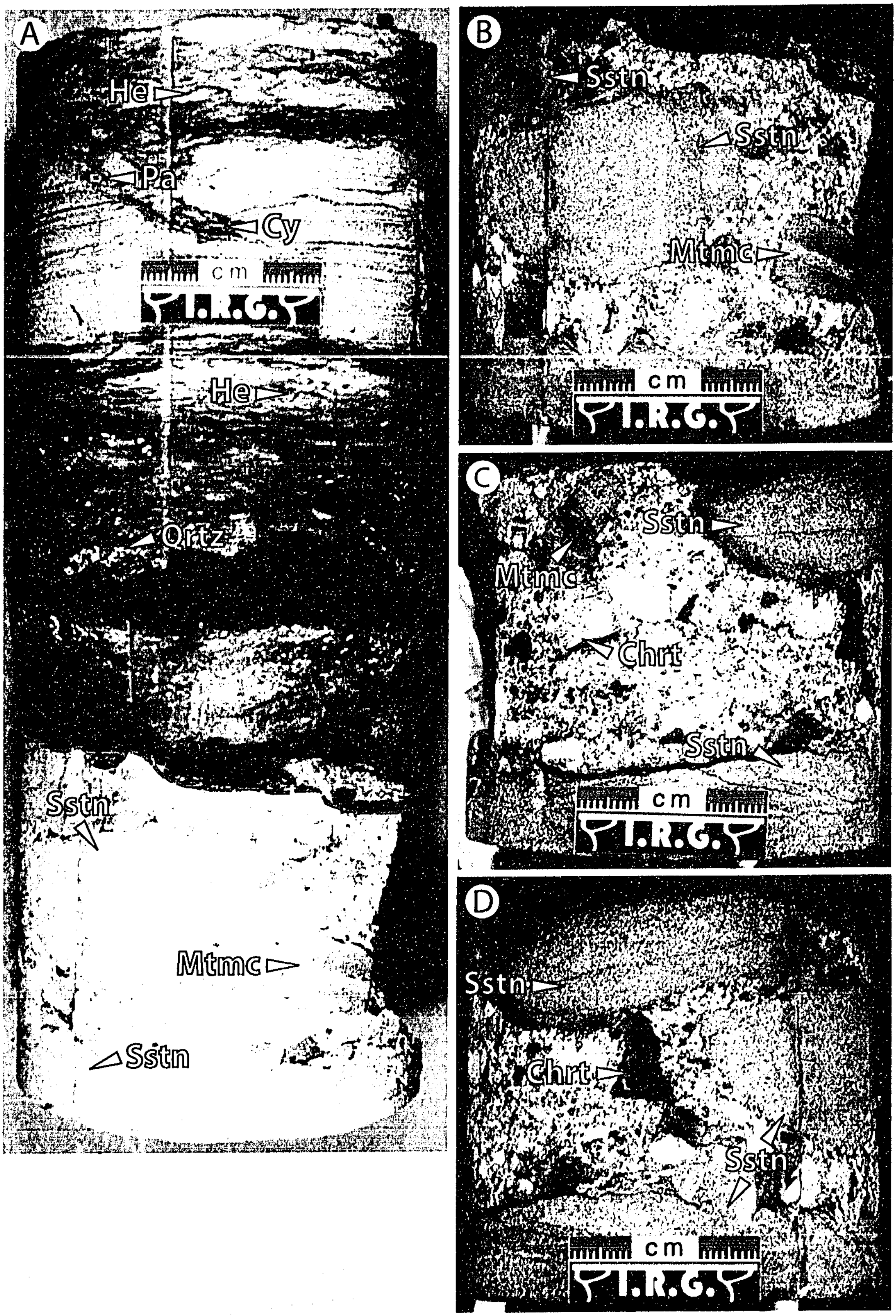
Facies F - Poorly sorted, scour-based, matrix supported conglomerate and medium to very coarse sandstone

A. Undulatory contact between a matrix-supported conglomeratic lag of Facies F and an interbedded, very fine-grained sandstone and a well-burrowed organic mudstone of Facies B. The mudstone and sandstone beds are burrowed by *Helminthopsis* (He), *Cylindrichnus* (Cy) and *Palaeophycus* (Pa) and contain numerous quartz (Qrtz) granules. The underlying matrix-supported conglomerate contains large sandstone (Sstn) and metamorphic (Mtmc) clasts (Parsons P-41, Core #1, depth ~2971.30 m).

B. Matrix-supported conglomerate with large sandstone (Sstn) intraclasts, metamorphic (Mtmc) clasts and chert (Chrt) pebbles. The matrix for the conglomerate is a medium to very coarse-grained sanstone (Parsons P-41, Core #1, depth ~2971.30 m).

C. Matrix-supported conglomerate with large sandstone (Sstn) intraclasts, metamorphic (Mtmc) clasts and chert (Chrt) pebbles. The matrix for the conglomerate is a medium to very coarse-grained sanstone (Parsons P-41, Core #1, depth ~2971.30 m).

D. Matrix-supported conglomerate with large sandstone (Sstn) intraclasts and chert (Chrt) pebbles. The matrix for the conglomerate is a medium to very coarse-grained sanstone (Parsons P-41, Core #1, depth ~2971.30 m).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processes, corresponding to the erosional remnants of a landward movement of the shoreface during erosional shoreface retreat. Overall, the main constituents comprising this facies are intra-formationally derived sandstone rip-up clasts, chert and pebble clasts with mafic and metamorphic clasts. These poorly-sorted sediments support a genetic link between erosion of the substrate and emplacement of the lag deposit (MacEachern *et al.*, 1992). The erosion surface on which this lag sits is interpreted to represent a transgressive surface of erosion (TSE) (Bhattacharya and Walker, 1991; MacEachern *et al.*, 1992), which is equivalent to the ravinement and marine erosion surfaces of Nummedal and Swift (1987) and the high-energy flooding surface of Pemberton and MacEachern (1995). Transgressive surfaces are characteristically low relief, erosional surfaces that are cut by wave action and current processes typically affecting the upper shoreface during transgression and associated erosional shoreface retreat (Reading, 1996; Pemberton and MacEachern, 1995). These types of surfaces are discontinuous and limited spatially (Pemberton and MacEachern, 1995).

In the P-41 well, the conglomeratic deposits of Facies F separate two stacked delta front successions, and represent either an autocyclic or allocyclic surface. However, due to its single occurrence in the study area, its stratigraphic significance has not been further explored. It is likely that other examples of this facies exist in other wells; unfortunately no other cored intervals contain this facies.

Facies G - Medium to very coarse-grained, planar tabular and trough cross-bedded, salt and pepper textured sandstone and minor conglomerate

<u>**Lithology and Sedimentary Features:**</u>

Facies G consists of light grey to buff, salt and pepper textured, upper medium to upper very coarse-grained sandstone interbedded with minor, multi-coloured conglomerate (Figs. 2.18A-D; 2.19A-D). Some sandstone intervals contain granule and pebble-sized grains, but most are generally quartz-rich and moderately to well-sorted. The interstratified conglomerate beds are generally poorly to moderately sorted, and contain significant amounts of subangular to subrounded lithic fragments, chert and quartz pebbles and lesser amounts of mudstone rip-up clasts (Figs. 2.18A; 2.19A,B,D). The chert and quartz pebbles and mudstone rip-up clasts are generally concentrated at the basal, scour surfaces of individual units, although some pebbles are scattered within the units themselves. In some of the basal intervals, matrix-supported chert pebble conglomerates 10 to 15 cm thick or thinner clast-supported conglomerates are observed. This facies typically has a scoured base and displays an overall fining-upward grain size profile, ranging from very coarse to pebble-sized grains at the base with medium-grained sandstone toward the top of the succession. Sedimentary structures include low, medium to high-angle, planar tabular and trough cross-stratification (Figs. 2.18B,C). Planar tabular cross-stratification is characterized by bedding angles greater than 10°, but mostly averaging between 15° to 25°. Some intervals also contain trough cross-stratification that display bedding angles that progressively increase upward from 10° to 27°. Many of the coarser-grained intervals within the sandstone are massively bedded and normally graded with no recognizable sedimentary structures. Within the finer-grained, upper portions of the sandstone, minor asymmetrical ripple lamination is observed with foresets marked by organic detritus. Overall these ripple structures are relatively rare within this facies.

Coalified wood fragments and their impressions are especially abundant within this facies. They are generally preserved on the bedding planes and range in size from centimeters to decimeters in size with many pieces larger than the core diameter (Figs. 2.19A,B,D). Some of these wood fragments occur as thin coal beds up to 2 cm thick. Finely comminuted organic detritus, occurring as finely ground up wood fragments and/or phytodetrital material, is also particularly abundant within this facies. The abundant organic detritus and coalified wood fragments are generally found marking bedding planes enhancing the recognition of sedimentary structures. Both nodular and disseminated pyrite is common throughout sandstone intervals and is generally associated with coalified wood fragments. Overall, Facies G consists of stacked depositional units, exhibiting fining-upward trends.

Facies G is identified in five cores, including core #2 from A-12, core #2 and #3 from E-21, and core #2 and #3 from the L-43 well. The lower contact of Facies G is a scoured surface commonly incised into underlying Facies C, D, L and I. The upper contact is sharp to gradational and is generally stratigraphically overlain by Facies C. However, in many cores the upper contact is not preserved and the nature of the upper contact is speculative, but assumed to be gradational with overlying fine-grained deposits. In core #2 from the A-12 well Facies G incises into Facies C and has likely erosionally removed Facies D and E. Sections of Facies G in core varies in thickness from 2.87 m up to 7.04 m thick and are typically composed of stacked depositional sequences. The average thickness of individual sequences ranges between 1.79 to 4.50 m.

__**Ichnological Characteristics:**__

No trace fossils were identified within this facies.

__**Interpretation and Discussion:**__ **Distributary Channel (Meandering Fluvial Channels)**

Medium to very coarse-grained, quartz-rich sandstone and conglomerate intervals of Facies G are interpreted to represent distributary channel fill deposits of a meandering fluvial distributary channel system within a deltaic setting. The erosional base and overall fining upward grain size profile of this facies is characteristic of a channel fill sequence from a meandering fluvial system (Walker and Cant, 1984). The moderately to well-sorted sorted nature of the sandstone is characteristic of a moderate to high-energy hydraulic regime dominated by traction and saltation processes with

Figure 2.18

Facies G - Medium to very coarse-grained, planar tabular and trough cross-bedded, salt and pepper textured sandstone and minor conglomerate

A. Interstratified, matrix-supported conglomerate and medium to coarse-grained, moderately-sorted sandstone with faint low angle (<5°) cross-bedding. The conglomerate is composed of rounded chert pebbles (Chrt) and sandstone intraclasts (Sstn) (Siku A-12, Core #2, depth ~2810.75 m).

B. Medium to coarse-grained, moderately-sorted sandstone with faint low angle (<5°) cross-bedding. Note the coalified wood fragments (Wd) and minor scattered chert granules and pebbles (Siku A-12, Core #2, depth ~2812.25 m).

C. Medium angle (5° to 10°) cross-bedded, medium to coarse-grained, quartz-rich sandstone. Note the centimeter-scale graded bedding (Siku E-21, Core #3, depth ~3051.67 m).

D. Medium to coarse-grained sandstone with abundant coalified wood (Wd) fragments (Siku A-12, Core #2, depth ~2811.73 m).

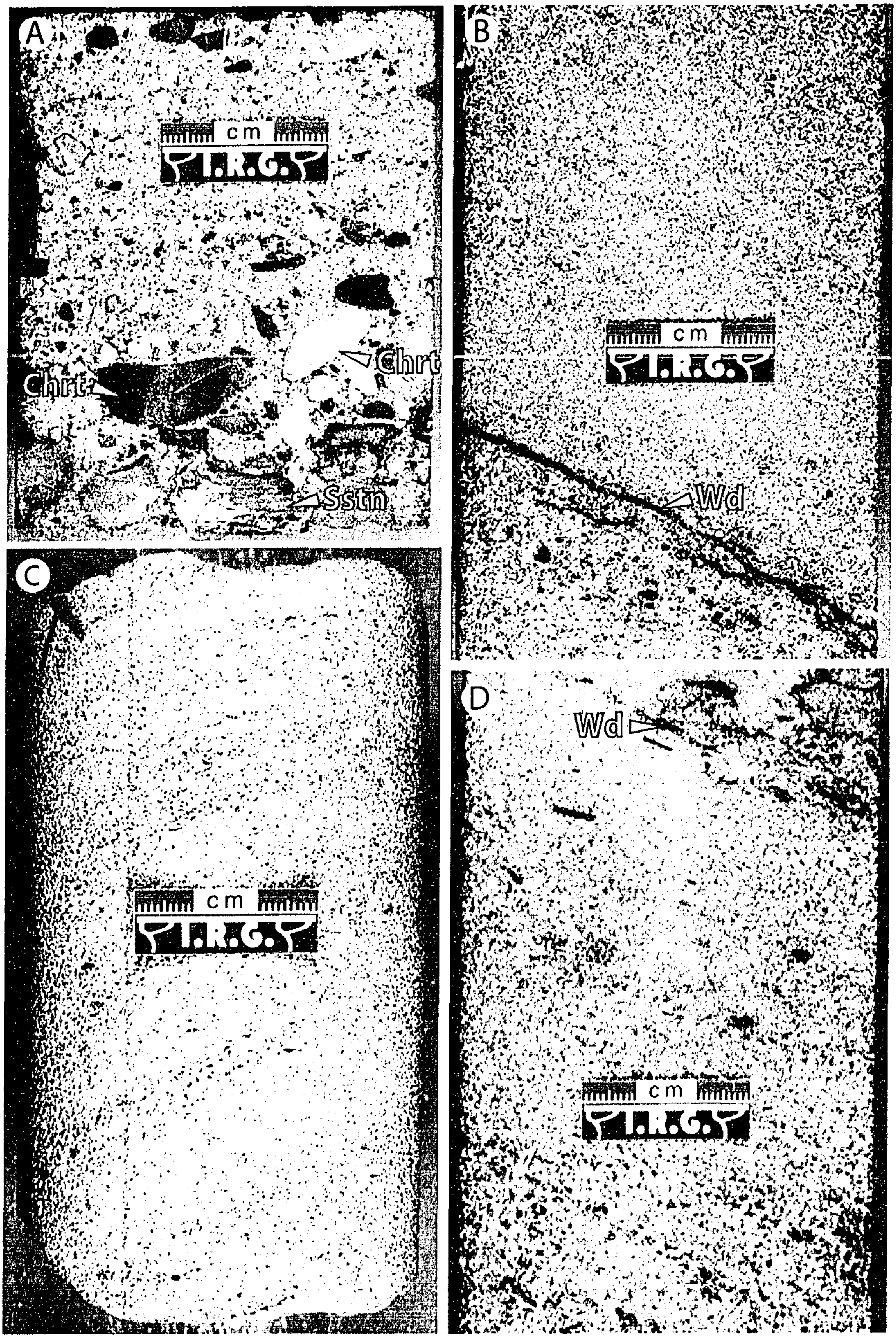

A
cm
T.R.G.
Chrt
Chrt
Sstn
B
cm
T.R.G.
Wd
C
cm
T.R.G.
D
Wd
cm
T.R.G.

Figure 2.19

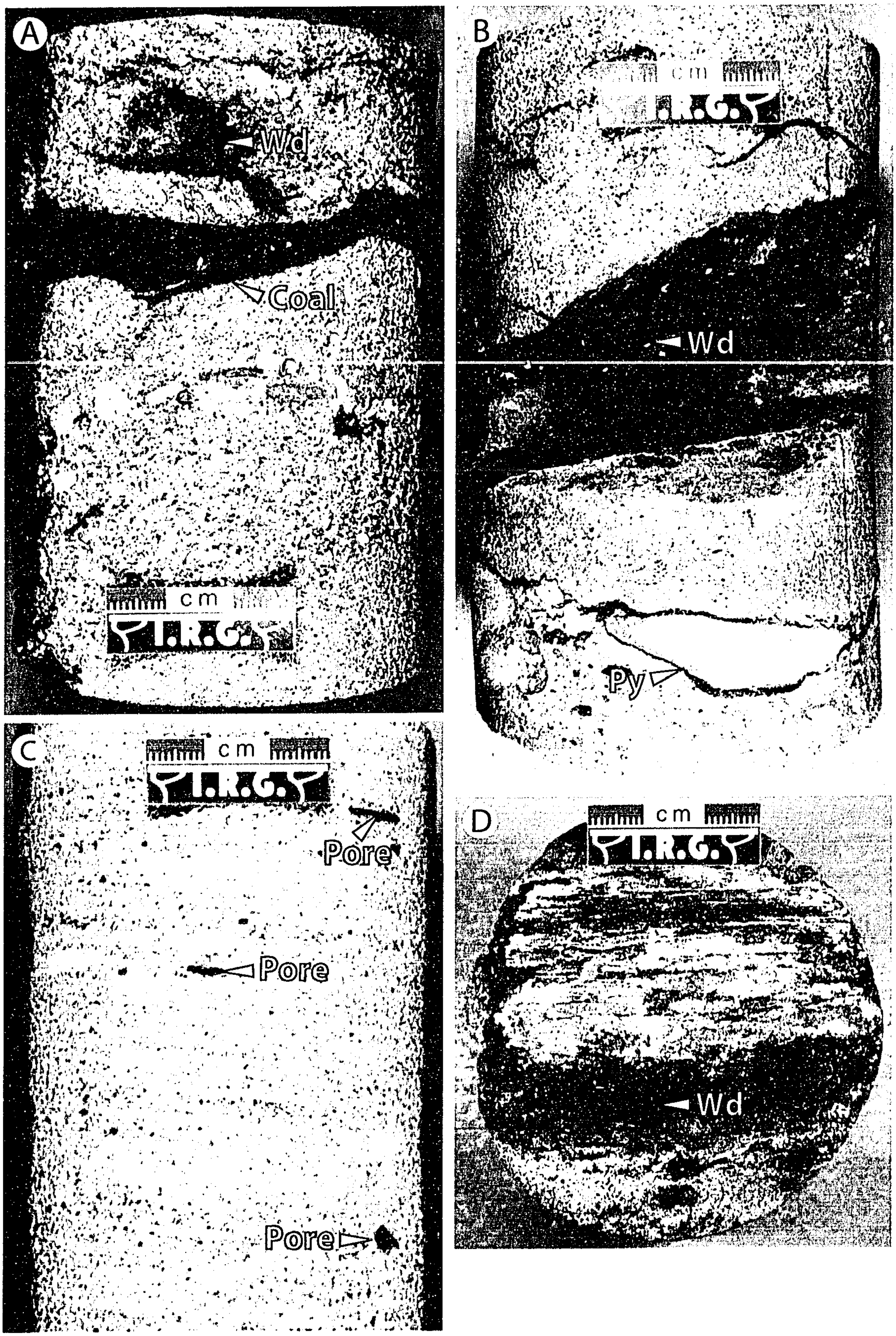
Facies G - Medium to very coarse-grained, planar tabular and trough cross-bedded, salt and pepper textured sandstone and minor conglomerate

A. Massive to low angle cross-bedded, poorly to moderately-sorted, medium to coarse-grained sandstone with scattered chert and quartz pebbles. The upper portions contains a 1 cm thick coal bed (Coal) and is overlain by coalified wood fragments (Wd). The coalifed wood fragments mark the bedding planes, especially near the top of the core slab (Siku E-21, Core #3, depth ~3046.28 m).

B. Low angle cross-bedded, medium-grained sandstone with a thick zone of coalified wood fragments. The middle portion contains a 3 to 5 cm thick coal bed comprised of coalified wood fragments (Wd). Note the pebble lag comprised of pyritized wood fragments (Py) and chert and quartz pebbles and the base of the core slab (Parson L-43, Core #2, depth ~2853.36 m).

C. Low angle cross-bedded, medium to coarse-grained sandstone with centimeter-scale graded bedding. Note the numerous open pore spaces (Pore) marked by dissolved mudstone rip-up clasts and small pebbles (Siku E-21, Core #3, depth ~3046.89 m).

D. Bedding plane view of a large coalified wood fragment (Wd) preserved within a medium-grained sandstone (Siku A-12, Core #2, depth ~2941.39 m).



A
Wd
Coal
cm
T.R.G.
B
cm
T.R.G.
Wd
Py
C
cm
T.R.G.
Pore
Pore
Pore
D
cm
T.R.G.
Wd

varying energy levels and current flows associated with a fluvial channel.

The low, medium to high angle cross-bedding present within this sandstone is interpreted to represent both planar tabular and trough cross-stratification (TCS). TCS intervals exhibit cross-bedding angles greater than 10°, but mostly range between 15° to 25° and progressively steepen-upward. Most sets are a few decimeters in thickness, with widths ranging from less than a meter to more than 4 m (Harms *et al.,* 1982). Within the coarser-grained and pebbly sandstone intervals, TCS is generally faint and poorly defined because of the coarse-grained size. Planar tabular cross-stratified intervals are generally parallel and consistent with measured angles that also range between 10°, but mostly between 15° to 25°, with some with steeper angles up to 32°. Bounding surfaces are planar and parallel, and internal laminations do not have tangential relationships to the basal surfaces. The average dip angle for foresets can average around 30° or more, which approximates the angle of repose for sand (Harms *et al.,* 1982; Reineck and Singh, 1980). These bedforms range in thickness from a few decimeters to a meter or more with some up to 10 m thick (Harms *et al.,* 1982). Planar tabular cross-bedding is developed in both sandy and gravely sediments as well as in both meandering and braided fluvial systems (Reineck and Singh 1980). It should be noted that the limited diameter of the core material makes positive identification of both trough cross-stratification and planar tabular cross-bedding difficult. In reality, a three dimensional view of the bedform is required for proper identification.

Both of these types of cross-stratification are interpreted to represent the migration of numerous dunes and small bars along the base of a fluvial channel. Planar tabular cross-stratification and current rippled laminated sandstone reflect the migration of sand waves. The shift from planar tabular to current rippled sandstones at the top of bedsets reflects a decrease in velocity of flow (Harms *et al.,* 1975). These structures are evidence of deposition during waning current flow on a point bar. Planar tabular, medium to high angle cross-bedded sandstone is interpreted as channel point bar deposits. The asymmetrical ripples found capping the individual bedsets are interpreted as current-ripples formed in a depositional environment dominated by unidirectional current flow (Harms *et al.,* 1975). The massively bedded nature of some portions of this facies is indicative of rapid sedimentation rates within a high-energy hydraulic regime dominated by traction and saltation processes. The relatively poorly-sorted, matrix-supported conglomeratic beds commonly found at the base of this facies are interpreted to represent a channel lag deposits formed at the base of a distributary channel. Gravels are generally concentrated at the base of channels as lags because the finer-grained components are moved as bed load carried away in suspension (Reineck and Singh, 1980). Where present, the channel lags typically occur as thin, discontinuous layers, which occupy the lowest part of a channel or point bar sequence (Reineck and Singh, 1980). Lags composed primarily of mudstone rip-ups clasts likely represent collapse sediments due to erosion of the cut bank.

The abundance of coalified wood fragments and phytodetrital material within this sandstone represents the remnants of terrestrial organic debris eroded from the surrounding floodplains and cut banks of fluvial channels during storm periods. In deltaic settings, storms are generally associated with high amounts of precipitation and flooding in the lower and upper delta plain environments. The high amounts of precipitation associated with these storm events elevate the rate of fluvial discharge resulting in flooding of the distributary channels. During flood stage these distributary channels fill with sediment-rich waters in the form of both bedload and suspended sediment. Flooding and increased fluvial discharge associated with these high levels of precipitation erode and transport organic detritus from the surrounding floodplains and riverbanks into the fluvial channels where it is carried in suspension to the distributary channels as rafted material. As this rafted organic material is transported downstream, it is either incorporated into the fluvial sediments and preserved on bedding planes of migrating dunes and ripple bedforms or transported further into the delta front setting. Reineck and Singh (1980) noted that in the lower reaches of rivers where coarse-grained materials are unavailable drifted wood fragments and mud pebbles can become concentrated as lag deposits. The finer-grained organic detritus or "coffee ground" material within these sandstones is interpreted to have a similar origin as the larger wood fragments, however this material likely reflects the fine-grained component.

The lack of trace fossils within Facies G further suggests a non-marine origin for these deposits. Distributary channels are characterized by harsh ecological conditions with numerous biological stresses that do not favour the colonization of these sediments by infaunal organisms. Such biological stresses include overall high-energy conditions, fluctuations in salinity, high sedimentation rates, turbid water conditions and continuously migrating bedforms.

Facies H - Poorly sorted, scour-based, medium to high angle cross-bedded, medium to very coarse-grained sandstone and conglomerate

<u>Lithology and Sedimentary Features:</u>

Facies H consists of interstratified, light grey to buff, upper medium to upper very coarse-grained sandstone and conglomerate (Figs. 2.20A-D; 2.21A-D). Sandstone intervals are dominantly quartz-rich and contain granule and pebble-sized grains resulting in poor to moderate sorting. The interstratified conglomerate intervals are matrix-supported, composed of granule to pebble-sized grains, generally poorly to moderately-sorted. The dominant lithology of Facies H is coarse to very coarse-grained sandstone; the ratio of coarse-grained sandstone to conglomerate is approximately 85 to 15. Overall, this facies is composed of multiple, stacked, commonly truncated, decimeter-scale, fining-upward depositional units with scoured bases marked by conglomeratic lags composed of mudstone rip-up clasts, chert and quartz pebbles, sandstone intraclasts and coalified wood fragments. The main distinguishing feature between the sediments of Facies G and H included thin, multiple, stacked, fining-upward sequences and the overall coarser-grained nature and poorer sorting of the sediments comprising Facies H compared to finer-grained and moderately to well-sorted sediments of Facies G.

Sedimentary structures identified within Facies G include interstratified, medium to high-angle, planar tabular and trough cross-stratification. Planar tabular cross-bedding is the dominant structure recognized with the majority of measured bedding angles greater than 10° and up to 32°, but averaging 15° to 25° (Figs. 2.20A,C; 2.21A). Trough cross-stratification is less common within this facies. It is recognized from beds displaying the progressively upward-steepening of bedding angles from 10° to 25° up to a maximum of 32° (Fig. 2.21D). The foresets of both the planar tabular and trough cross-stratified intervals are composed of alternating graded beds (1 to 3 cm thick) of medium to very coarse-grained sandstone (Figs. 2.20A; 2,21B). The majority of coarse to very coarse-grained sandstone is faintly horizontal to low angle cross-bedded. However, several intervals are massively bedded or normally graded exhibiting no recognizable sedimentary structures. Asymmetrical ripple lamination is particularly abundant within the finer-grained upper portions of this facies (Fig. 2.21C). These ripple structures are commonly found interstratified or overlying planar tabular cross-stratified beds.

Although the sediments are generally quartz-rich, many intervals contain thick beds of lithic fragments with significant amounts of chert and quartz pebbles, sandstone intraclasts and mudstone rip-ups clasts. The chert and quartz pebbles are subangular to subrounded and range in diameter from less than 1 cm up to 4 cm (Fig. 2.20A). Mudstone rip-up clasts ranging in size from 1 to 3.5 cm are especially abundant within many intervals, however they are not as abundant as in Facies G (Fig. 2.20D). Coalified wood fragments are common within sandstone beds and are also found associated with conglomerate and mudstone rip-ups. These wood fragments are centimeter to decimeter scaled with most pieces ranging in size from 2 to 5 cm; some pieces are larger than the core diameter. The brecciated to conglomeratic intervals composed of lithic fragments, chert and quartz pebbles, mudstones rip-up clasts and coalified wood fragments commonly occur as lags marking the basal scour surfaces of multiple, fining-upward sequences that average 30 cm to over 1 m in thickness (Fig. 2.20A). Disseminated and nodular pyrite is also common throughout Facies H. Vertical fractures are also common features in some sandstone beds. These fractures are both open and sealed with pyrite and cross-cut sedimentary bedding, indicating that they are likely post-depositional features.

Facies H comprises approximately 85% to 90% of the lower portion of the Kamik Formation. It is identified in three cores from the study area including core #3 from L-43 and core #1 and 2 from the N-10 well. The lower contact of Facies H is scoured and generally overlies the fine-grained deposits of Facies I and L. The upper contact is only present in core #3 from the L-43 well where sediments of Facies G are overlain by the fine-grained deposits of Facies K. The upper contact is

Figure 2.20

Facies H - Poorly sorted, scour-based, medium to high angle cross-bedded, medium to very coarse-grained sandstone and conglomerate

A. Medium to coarse-grained, well-sorted, planar tabular cross-bedded sandstone, with bedding angles between 15° and 20°, overlain by a poorly sorted, medium-grained sandstone. The contact is marked by a scour and is overlain by rounded sandstone intraclasts (Sstn), chert pebbles (Chrt) and coalified wood fragments (Wd) (Parsons N-10, Core #1, depth ~2753.59 m).

B. Very coarse-grained, moderately sorted sandstone with faint, low angle (5° to 10°) cross-bedding. Note the coalified wood fragments (Wd) and numerous chert and quartz granules. This interval is extremely porous with measured porosity of 19.9% and permeability of 1088 millidarcies (Parsons N-10, Core #2, depth ~2761.05 m).

C. Medium-grained, planar tabular cross-bedded sandstone, with bedding angles between 15° and 20°, overlain by a massively bedded, medium-grained sandstone. Note the centimeter-scale graded bedding alternating from medium to coarse-grained sandstone (Parsons L-43, Core #3, depth ~2940.71 m).

D. Medium to very coarse-grained sandstone with large (centimeter-scale) angular mudstone rip-up (Mdstn) clasts, coalified wood fragments (Wd) and chert pebbles (Chrt) (Parsons L-43, Core #3, depth ~2943.78 m).

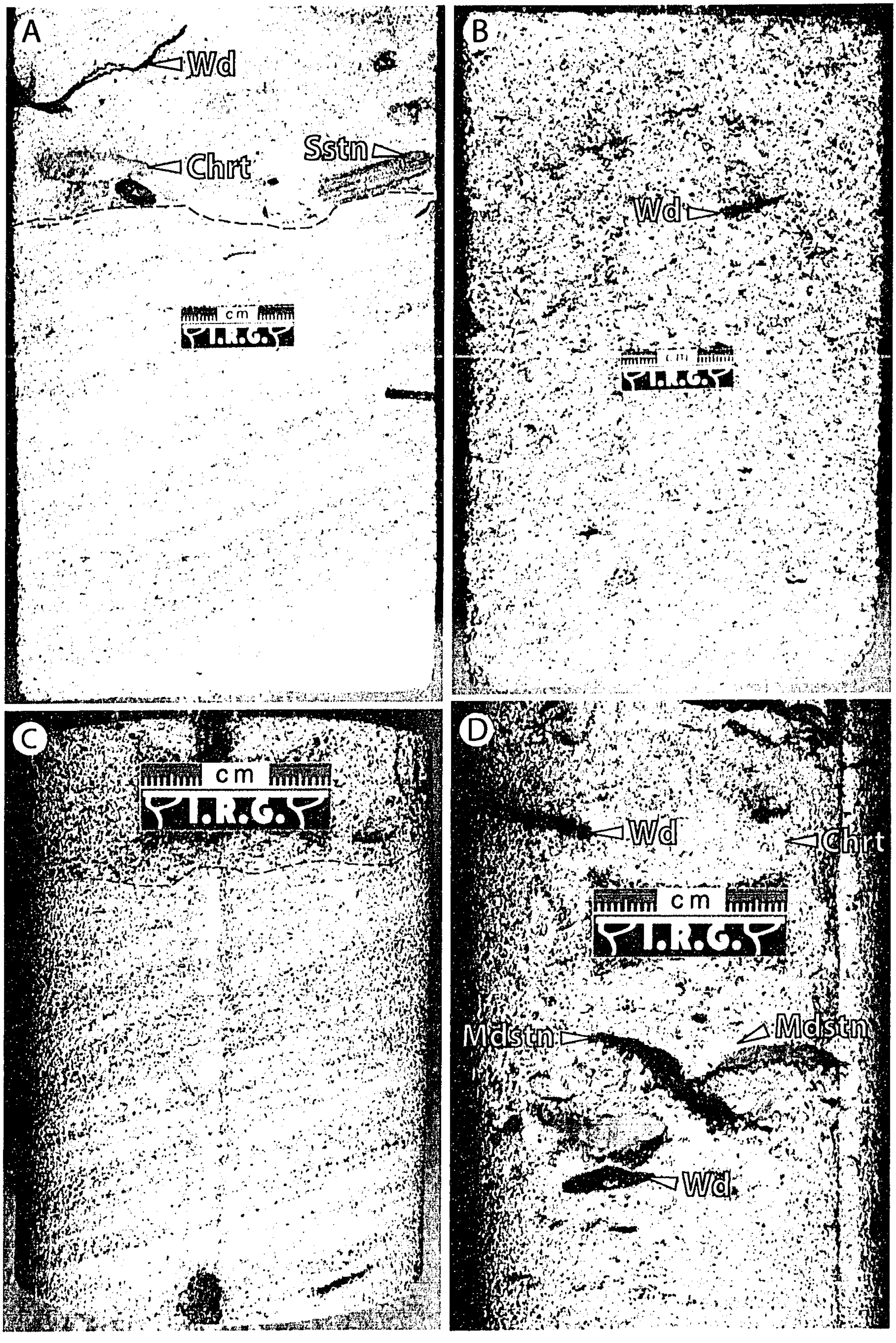

A
Wd
Chrt
Sstn
cm
T.R.G.
B
Wd
cm
T.R.G.
C
cm
T.R.G.
D
Wd
Chrt
cm
T.R.G.
Mdstn
Mdstn
Wd

Figure 2.21

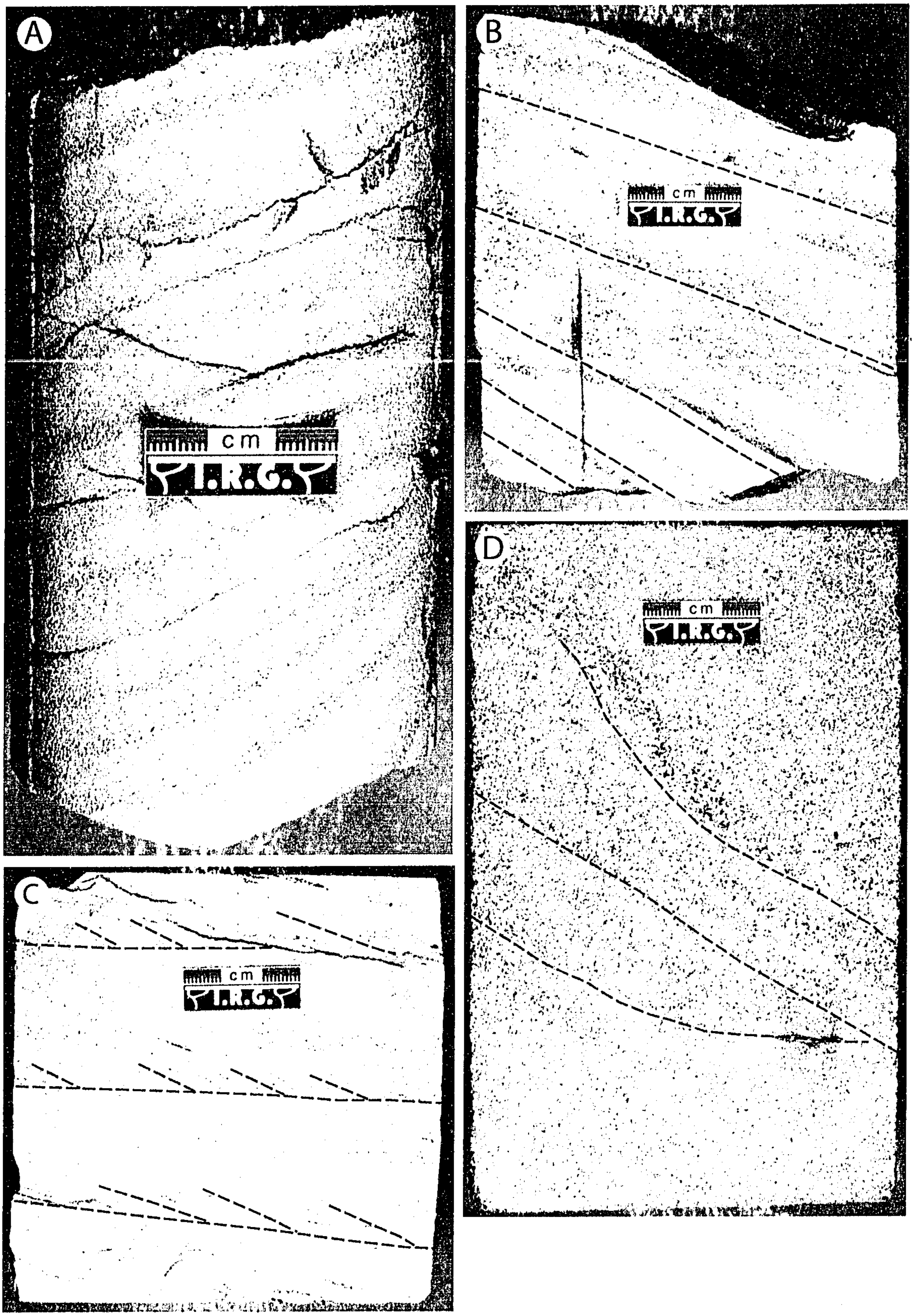
Facies H - Poorly sorted, scour-based, medium to high angle cross-bedded, medium to very coarse-grained sandstone and conglomerate

A. Medium-grained, well-sorted, medium to high angle (15° to 20°) planar tabular cross-bedded sandstone (Parsons L-43, Core #3, depth ~2745.80 m).

B. Medium to coarse-grained, medium to high angle (22° to 32°) cross-bedded sandstone. Note the steep dip of the cross-beds in the lower portion interpreted to represent the toesets of a braid bar (Parsons N-10, Core #2, depth ~2797.99 m).

C. Current ripple cross-lamination marked by dipping asymmetrical foresets. Note the stacking of successive ripple cross-laminated intervals (Parsons N-10, Core #2, depth ~2795.63 m).

D. Medium to coarse-grained sandstone displaying trough cross-bedding (TCS). Note the progressive upward increase in the bedding angles from base to top (15° to 32°) with bedding in the upper portions oversteepened above the angle of repose for sand (32°) (Parsons N-10, Core #1, depth ~2758.25 m).


A
B
C
D
cm
I.R.G.

marked by a decrease in sandstone content and an increase in mudstone content, interbed thickness and amount of bioturbation. In core, Facies H varies in thickness from 3.07 m to 6.11 m up to 9.85 m however; many of these cored intervals are partial sections. In well logs, Facies H appears to occur in thicker successions averaging 25 to 30 m.

Ichnological Characteristics:

No trace fossils were identified within this facies.

Interpretation and Discussion: Distributary Channel (Braided Fluvial Channel)

Medium to very coarse-grained, quartz-rich sandstone and interstratified conglomerate of Facies H are interpreted to represent the channel fill deposits of a sandy, braided distributary channel system from the upper delta plain. The sandstone-rich nature of these deposits, along with the multiple, stacked, commonly truncated, scoured-based, fining-upward depositional sequences are characteristic of a braided channel fill sequence. The poor to moderate sorting of the coarse-grained sandstone and interstratified conglomerate suggests deposition in a braided system dominated by bedload transport of sediment and constantly fluctuating discharge rates. In addition, the sheet-like nature, overall thickness, wide lateral extent, stratigraphic continuity and thin or absent interstratified floodplain deposits also indicate a braided fluvial system (Campbell, 1976; Walker and Cant, 1984).

The sedimentary structures comprising this facies are indicative of deposition in a setting dominated by unidirectional sediment transport by currents. Medium to high angle cross-bedding within sandstone intervals represents planar tabular and trough cross-stratification. Planar tabular cross-stratification represents the migration of two-dimensional (2D) dunes (Harms *et al.*, 1975). It commonly develops in both coarse sand and gravely sediments and this type of cross-stratification has been recognized as the dominant bedding structure in sediments from braided rivers (Reineck and Singh, 1980). The intervals displaying cross-bedding that progressively steepen-upward from angles of 10° to 25° or greater is interpreted to reflect trough cross-stratification. Trough cross-stratification is interpreted to originate from the migration of three-dimensional (3D) tabular bars or dunes (Harms *et al.*, 1975). Both the planar tabular and trough cross-stratified sandstones represent the migration of active bars and dune bedforms along the base of a braided fluvial channel. The asymmetrical ripples found capping the individual bedsets are interpreted to represent current ripples, formed in an environment dominated by traction and saltation processes. The formation of these structures is attributed to waning flow conditions within a braided channel fill sequence dominated by unidirectional current flow (Reading and Collinson, 1996). As previously discussed in the interpretation for Facies G, the identification of planar tabular and trough cross-stratification in core is difficult. The proper identification and differentiation between these bedforms requires a three-dimensional perspective, which is generally unattainable from the limited diameter of most subsurface cores. The alternating graded beds of medium, coarse to very coarse-grained sandstone comprising the foresets of planar tabular and trough cross-bedded intervals form by the sorting action of small bedforms on the bar surface or from the migration of bars composed of sandstone with variable grain sizes (Reineck and Singh, 1980). These graded beds form due to the avalanching of medium, coarse to very coarse-grained sized sediments down the slipfaces of bar bedforms. The size of sediment comprising the beds within individual dune bedforms is controlled by the flow rate and the available sediment sources within the channel. This is consistent with deposition in a braided river system as the sediments within these settings are typically poorly to moderately-sorted with variable grain sizes and flow or discharge rates are variable and sporadic. The massively bedded nature of some portions of this facies are indicative of high sedimentation rates within a relatively high-energy hydraulic regime dominated by bedload transport of sediment.

The abundance of coarse-grained sandstone and the lack of mud in the channel fill deposits indicates a high bedload transport, more characteristic of a braided system rather than a meandering one (Hamblin and Walker, 1979). The absence of thick, fining-upward channel fill sequences is due to the rapid and continuous shifting of channel positions, a characteristic feature of braided river systems (Miall, 1978). The mudstone rip-up clasts are interpreted to have originated from the active erosion of numerous mid-channel alluvial islands and braid bars (Miall, 1978; Reineck and Singh, 1980). The lower abundance of mudstone rip-up clasts compared to Facies G is attributed to the channel banks of braided fluvial systems, being composed of fine-grained floodplain mudstones and siltstones, which are not as thick or extensive as those found in meandering river systems (Cant, 1982). The coalified wood fragments preserved within these sandstones are interpreted to have a similar origin as those in Facies G. This organic material represents remnants of vegetative material eroded from the margins and emergent, inactive bars of the braided channel system during periods of high discharge. It was likely that this organic material was transported downstream and incorporated into the braided channel fill deposits during storms when high amounts of precipitation and increased fluvial discharge facilitated erosion and sediment transport within the braided channel system. This organic material would have been carried downstream both in suspension and as bedload and incorporated within the sandstones and bedforms comprising these deposits.

Reineck and Singh (1980) noted that the channel fills of braided river systems are characterized by a scoured base and are comprised of large-scale cross-bedding with individual sets up to 1 m thick (Coleman and Prior, 1982). This is consistent with the thickness (ranging from 30 cm up to 1 m) of individual, fining-upwards, channel fill cycles observed in Facies H. These cycles are interpreted to reflect ephemeral bars that were commonly truncated due to the relatively shallow depths and lateral migration of individual braid channels (Reineck and Singh, 1980). However, it should be noted that the braid channel fills of Facies H in core occur as much thicker successions (ranging from 6.11 to 9.85 m). These thicker successions are interpreted to represent, multiple, stacked channel fills of a braided river system. This contrasts to meandering fluvial systems that are characterized by thicker point bars deposits, which display thicker individual channel fills. In addition, the sediments in braided river systems tend to aggrade vertically by deposition, whereas in meandering systems tend to deposit sediments by lateral accretion and to a lesser extent vertical accretion within the channels and surrounding flood plains (Reineck and Singh, 1980; Cant, 1982). The concentration of the coarse-grained material including, conglomeratic and breccia intervals, composed of chert and quartz pebbles, mudstone rip-up clasts, coalified wood fragments and sandstone intraclasts within certain intervals is interpreted to represent the base of individual braid channels. This material is concentrated at the base of individual braid channels due to its inability to be transported by the fluctuating discharges characteristic of braided fluvial systems.

The absence of trace fossils within Facies H also suggests a non-marine origin for these deposits. The lack of bioturbation is attributed to similar reasons as interpreted for Facies G. The harsh environmental conditions within a braided distributary channel during deposition including, overall high-energy conditions, fluctuating sedimentation rates, salinity variations, turbid water conditions, continuously migrating bedforms and numerous other stresses, created an environment inhospitable for infaunal organisms to colonize the sediments.

Facies I - Wavy to lenticular bedded, bioturbated, organic mudstone, siltstone and fine-grained silty sandstone

Lithology and Sedimentary Features:

Facies I is composed of wavy to lenticular bedded, bioturbated, dark grey to black, organic mudstone, siltstone and light grey, upper very fine to upper fine-grained silty sandstone (Figs. 2.22A-D). The mudstone, siltstone and sandstone beds are interlaminated to interbedded on a millimeter to decimeter scale and exhibit sporadic burrowing. The mudstones are highly organic, massive, but generally fissile and unburrowed. Bedding thickness of the sandstone beds averages 1 to 3 cm in the lower portion and 5 to 15 cm toward the top of Facies I intervals. The amount and thickness of mudstone interbeds decreases upward from 1 cm or less in the lower portion to 10 to 20 cm in upper portions. This facies corresponds to the distal equivalent of the sandstone-dominant sediments of Facies J.

Sedimentary structures include wavy and lenticular bedding, low angle (<5°) parallel lamination, asymmetrical and symmetrical ripple lamination (Figs. 2.22A,B). Flaser bedding is rare to common in some of the thicker sandstone beds (Fig. 2.22B). Asymmetrical ripple lamination is more common than symmetrical ripple lamination; however, symmetrical ripples are locally relatively abundant within some intervals. Soft-sediment deformation structures are abundant in this facies and include decimeter-thick intervals of convolute and overturned bedding (Fig. 2.22D). These structures are particularly common in core #3 from the L-43 well. Mudstone beds commonly contain synaeresis cracks, which are infilled with very fine to fine-grained sandstone (Fig. 2.22A). In cross-section, the cracks are subvertical to vertical, generally crenulated, extend downward from overlying sandstone beds and varying in length from 1 cm to 6 cm. Carbonaceous detritus in the form of coalified wood and plant fragments is especially abundant within mudstone and siltstone intervals and sparse to moderate in sandstone beds. Large pyrite and rare siderite nodules are also present, most with diameters ranging from 1 to 3 cm.

Facies I is identified in seven cored intervals including core #1 and #3 from E-21, core #1, #2 and #3 from L-43, and core #1 and #2 from the N-10 well. Facies I is commonly interstratified with Facies G, H, J and L. The lower contact is sharp to gradational with Facies H, J and L and undulatory to gradational with Facies D. The upper contact is generally sharp, gradational to undulatory with overlying Facies H, J and L and scoured with Facies G and H. The upper contact is marked by an increase in sandstone content and bed thickness, along with a decrease in mudstone content. Facies I varies in thickness from 1.46 m to 2.43 m.

Ichnological Characteristics:

Trace fossils within Facies I are sporadically distributed, exhibit a low diversity and impart a low degree of bioturbation. Most ichnofossils are restricted to the arenaceous sediments. The main ichnological elements are suspension-feeding burrows corresponding to the *Skolithos* ichnofacies, with a lesser degree of deposit-feeding organisms, and include *Arenicolites* (Fig. 2.22A), *Skolithos*, *Teichichnus*, *Diplocraterion*, *Planolites*, *Palaeophycus*, *Thalassinoides*, *Cylindrichnus* and rare *Lockeia*, *Bergaueria* (Fig. 2.22B) and fugichnia. Some sandstone beds contain vertical escape structures. Carbonaceous rootlets 1 to 3.5 cm in length are also present within mudstone beds, particularly in core #1 and #3 from E-21 and core #2 from the N-10 well.

Interpretation and Discussion: Interdistributary Bay (Bay Fill)

Wavy to lenticular bedded, bioturbated, fine-grained sandstone, siltstone and organic mudstone of Facies I represent an interdistributary bay in the lower delta plain environment. The lower delta plain extends landward from the shoreline to the upper limit of tidal influence (intertidal) and is subject to both marine and fluvial processes (Coleman and Prior, 1982). The mostly fine-grained nature of these sediments is indicative of deposition within a relatively low to moderate-energy bay environment with deposition dominated by slow suspension and minor traction processes. Lenticular to wavy bedding is characteristic of low-energy settings and suggests that both sand and mud are available for deposition and that periods of current activity alternated with periods of quiescence (Reineck and Singh, 1980). Wavy and lenticular bedding develops under

Figure 2.22

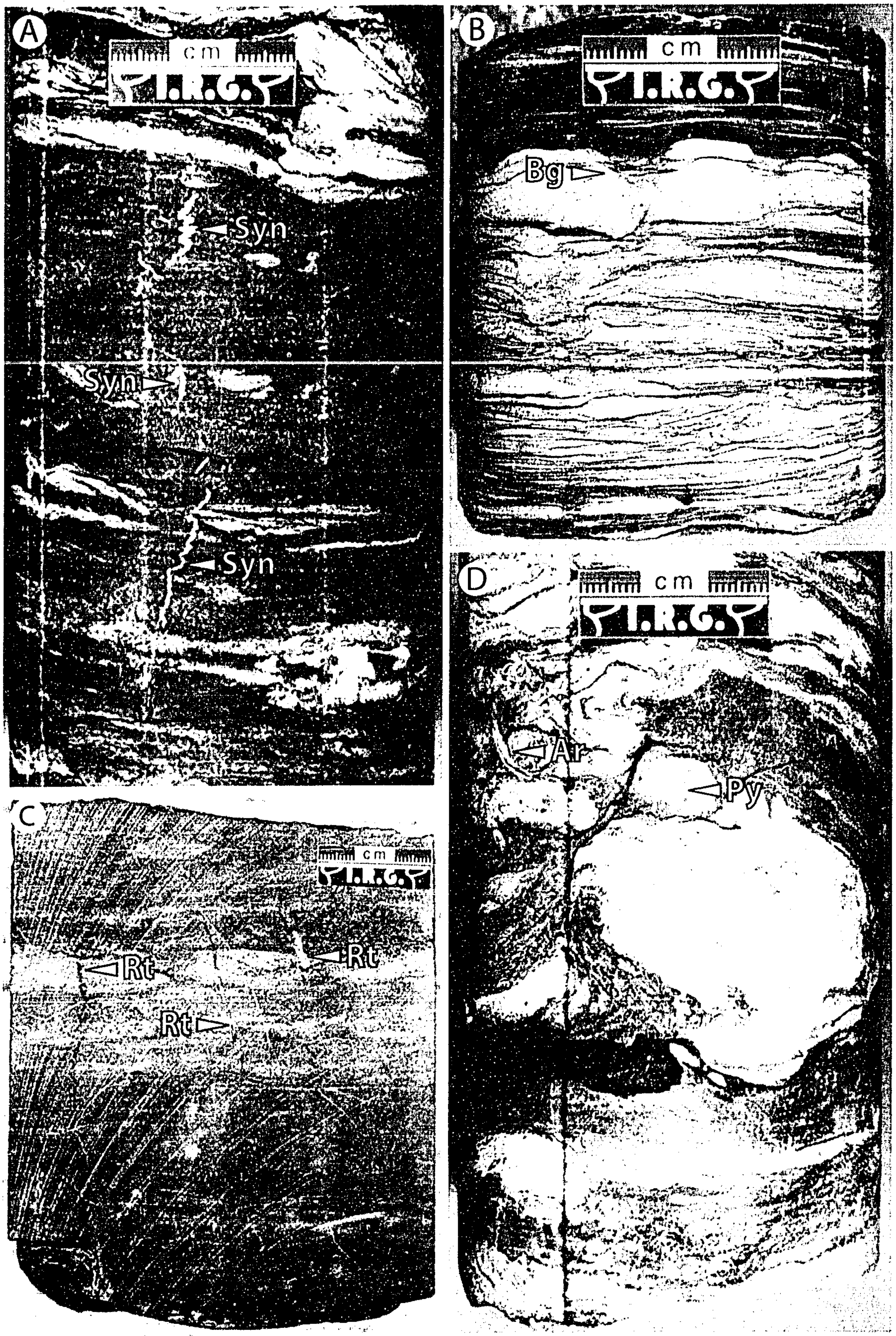
Facies I - Wavy to lenticular bedded, bioturbated, silty fine-grained sandstone, siltstone and organic mudstone

A. Wavy to lenticular bedded, highly organic mudstone interbedded with a fine to medium-grained sandstone. Note the long (1.5 to 5.5 cm) crenulated synaeresis cracks (Syn) infilled with fine-grained silty sandstone (Siku E-21, Core #3, depth ~3050.62 m).

B. Wavy bedded mudstone and very fine-grained, flaser-bedded sandstone with *Bergaueria* (Bg) extending downward into the flaser bedded sandstone from the overlying mudstone (Parsons L-43, Core #1, depth ~2774.42 m).

C. Faint, low angle, parallel-laminated to massively bedded, silty, highly organic, coaly mudstone with minor coalified rootlets (Rt) (Parsons N-10, Core #1, depth ~2801.65 m).

D. Convolute bedded, organic silty mudstone and fine to medium-grained sandstone. Note the single *Arenicolites* (Ar) in the upper portion and the pyrite nodules (Py) (Parsons L-43, Core #3, depth ~2936.75 m).


A
cm
I.R.G.
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Syn
Syn
B
cm
I.R.G.
Bg
C
cm
I.R.G.
Rt
Rt
Rt
D
cm
I.R.G.
Ar
Py

conditions characterized by slack water and turbulent flow, generally within subtidal and intertidal environments. These structures are not purely restricted to tidal environments; they have been described from marine delta fronts by Coleman and Gagliano (1964).

Deposition of fine-grained, rippled sandstone is due to punctuated, sporadic, higher energy events, such as crevasse splaying and overbank flooding, depositing sediment into the quiescent bay environment. Thin, rippled silty sandstone reflects transport by traction currents and represent periods of weak wave and current activity. The asymmetrical ripples are interpreted as current ripples, which are consistent with deposition in an environment dominated by unidirectional current flow (Harms *et al.*, 1975), such as crevassing or overbank flooding into an interdistributary bay. The symmetrical ripples represent wave ripples, which form via wave processes reworking the thin sand beds. Wave ripples are differentiated from current ripples by their symmetrical slopes with rounded troughs and the absence of foresets. The shrinkage cracks are interpreted as synaeresis cracks rather than desiccation cracks due to the irregular form and the absence of any evidence of subaerial exposure. Synaeresis cracks form in settings with rapidly fluctuating salinities that causes the shrinkage of clays, and can either form at the sediment-water interface or substratally (Burst, 1965; Plummer and Gostin, 1981). The crenulated nature of these cracks indicates that they formed prior to compaction. Synaeresis cracks are often found extending downward from fine-grained sandstone into underlying mudstone and siltstone. This relationship suggests that the deposition of the sandstone beds was associated with the influx of fresh water into the marine waters of the bay environment. The co-existence of pyrite with the synaeresis cracks is also indicative of fluctuating salinities within marginal marine depositional settings (Burst, 1965; Plummer and Gostin, 1981; Wightman *et al.*, 1987) and further supports that Facies I was deposited in a open, interdistributary bay with periodic influxes of fresh water. The introduction of fresh water-laden sediments into the marine waters of the bay environment is consistent with the transport and rapid deposition by crevasse splays and overbank flooding of distributary channels. Soft-sediment deformation structures also originate from rapid deposition of sediments and are consistent with the deposition of crevasse splays into the bay environment.

The trace fossil assemblage is mixed, containing dwelling traces such as *Arenicolites*, *Skolithos*, *Diplocraterion*, *Thalassinoides*, *Palaeophycus*, and *Cylindrichnus*, feeding structures including *Planolites*, *Teichichnus*, resting traces such as *Bergaueria* and *Lockeia*. This trace fossil assemblage corresponds mainly to the *Skolithos* ichnofacies and is indicative of shallow, subtidal, low-energy marine environments. The low to moderate diversity suggests that the depositional environment was open marine, however the intervals with rare to absent burrowing suggest deposition in proximity to fresh water sources and/or periods of restriction from marine waters. Occasional brackish water within the bay environment could create conditions that would be responsible for the lower diversity and diminutive nature of some of the forms (Pemberton and Wightman, 1992). MacEachern and Pemberton (1992) noted that *Arenicolites* and *Diplocraterion* burrows are generally associated with arenaceous substrates in low energy sandy tidal flats. *Cylindrichnus* is commonly recognized from brackish water settings such as brackish-water bays, sandy lagoonal settings, and some sandy tidal flats (MacEachern, 2001). *Bergaueria* is generally indicative of normal marine conditions however, diminutive forms are commonly found within brackish-water environments. Fugichnia reflect the escape behaviour of benthic organisms rapidly covered by sediment. Following burial, the organism must escape upward to the newly formed sediment-water interface in order to respire and ingest food (MacEachern, 2001). This type of trace fossil is most common in settings subjected to very rapid or episodic sedimentation and is consistent with deposition in a bay environment characterized by crevasse splaying and overbank flooding. The trace fossil assemblage of Facies I is interpreted to represent a relatively "unstressed" environment compared to that of Facies K.

Facies J - Fine to medium-grained, low angle cross-stratified to flaser bedded sandstone and minor, dark grey siltstone and silty mudstone

Lithology and Sedimentary Features:

Facies J is composed of light grey to buff, lower fine to upper medium-grained silty sandstone with minor, dark grey to black, siltstone and organic mudstone (Figs. 2.23A-D; 2.24A-D; 2.5A-D; 2.6A-D). Mudstone becomes almost completely absent within upper portions of Facies J intervals. Sandstone beds average 5 to 30 cm thickness. The abundance of sandstone and thickness of beds increases toward the top of the facies. The mudstone beds are characteristically thin and range from less than 1 cm up to 5 cm thick. This facies characteristically displays a coarsening-upward profile. A typical succession is composed of low angle, parallel-laminated, flaser bedded and ripple laminated sandstone capped by rooted, fine to medium-grained sandstone and siltstone.

Sandstone intervals are characterized by low angle (<5°) cross-bedding, flaser and wavy bedding, ripple lamination, and soft-sediment deformation structures, with some intervals appearing massive to structureless (Figs. 2.23A-D; 2.24A-D; 2.25A). The flaser bedded intervals are marked by wisps of carbonaceous detritus, phytodetrital material and clay drapes (Fig. 2.23A). Asymmetrical and symmetrical ripples are particularly abundant and commonly display excellent examples of climbing ripple lamination (Figs. 2.23C,D; 2.24A,B,D). Some excellent examples of asymmetrical ripple foresets are highlighted by abundant organic detritus and phytodetrital material (Fig. 2.23D). The climbing ripple laminated intervals display a high degree of aggradation and are marked by well-preserved upward and laterally migrating ripple crests and troughs (Figs. 2.23C; 2.24D). In addition, a portion of this facies displays sigmoidal bedding and double mud drapes (Fig. 2.23B). Siltstone and mudstone intervals are massive to parallel-laminated with minor siderite beds and nodules. The upper parts of this facies are characterized by massively bedded to soft-sediment deformed sandstones with scattered horizontal laminae and ripple cross-stratification (Fig. 2.24C).

Organic detritus and finely comminuted "coffee ground" and micaceous material is abundant in Facies J (Figs. 2.23C; 2.4A,B,D). This organic detritus is found concentrated in the troughs of ripples and accentuate the laminae and bedding. Within core #1 from the A-12 well, carbonized plant fragments including leaf and stem imprints are observed on a bedding plane within a mudstone bed (Fig. 2.25E). Some of the sandstones contain coal fragments in addition to the finely comminuted, macerated, organic material. Extensively rooted zones and thin coal beds characterize the tops of many coarsening-upward successions.

Facies J is identified in six cores from the study area, including core #1 from the A-12, core #3 from E-21, cores #1 and #3 from the L-43 and cores #1 and #2 from the N-10 well. The lower contact is sharp to gradational with Facies H, I and L and undulatory with underlying Facies I and O. The upper contact is generally sharp to gradational with Facies I, K and L and marked by a decrease in sandstone content and bed thickness. Facies J is commonly interstratified with Facies H, I, K and L. Facies J typically occurs as a coarsening-upward successions and varies in thickness from 0.45 to 2.42 m.

Ichnological Characteristics:

This facies displays a low to moderate degree of bioturbation and diversity. Trace fossils are more abundant in the lower portions whereas root structures are more common in the upper portions of coarsening-upward sequences. The main ichnological elements are dwelling, suspension- and deposit-feeding burrows corresponding mainly to the *Skolithos* ichnofacies. Identified trace fossils include: *Diplocraterion* (Figs. 2.24A; 2.25A,D), *Skolithos*, *Arenicolites* (Figs. 2.25B,C), *Teichichnus*, *Palaeophycus*, *Planolites*, *Bergaueria* and fugichnia. Many of the *Arenicolites*, *Skolithos* (Fig. 2.26A) and *Diplocraterion* burrows range between 2 to 5 cm in length and truncate underlying sandstone and mudstone laminae. It should be noted that the burrows characterizing this facies were not identified in each occurrence of Facies J, and in many cores only 2 to 3 ichnogenera were identified with many intervals containing no trace fossils. The abundance and diversity of trace fossils varies between all occurrences of Facies J within the Kamik Formation. Root structures dominating upper parts of this facies are characteristically carbonaceous and identified by their vertical, wispy, downward-tapering and branching nature (Figs. 2.26A-D). They are sinuous in nature, crosscut the laminae and bedding structures, commonly thin and thicken along their lengths, and average 2.5 to 16 cm in length and 1 to 3 mm in width. Rootlet branches are distinct and have daughter roots and root hairs, which are smaller than the primary root and also taper downward. In core #3 from the E-21 well a large carbonaceous root approximately 6.5 cm long and 1 cm wide is preserved at the top of a sandstone bed (Fig. 2.26A). In core #2 from the N-10 well rootlet traces are associated with *Palaeophycus*.

Figure 2.23

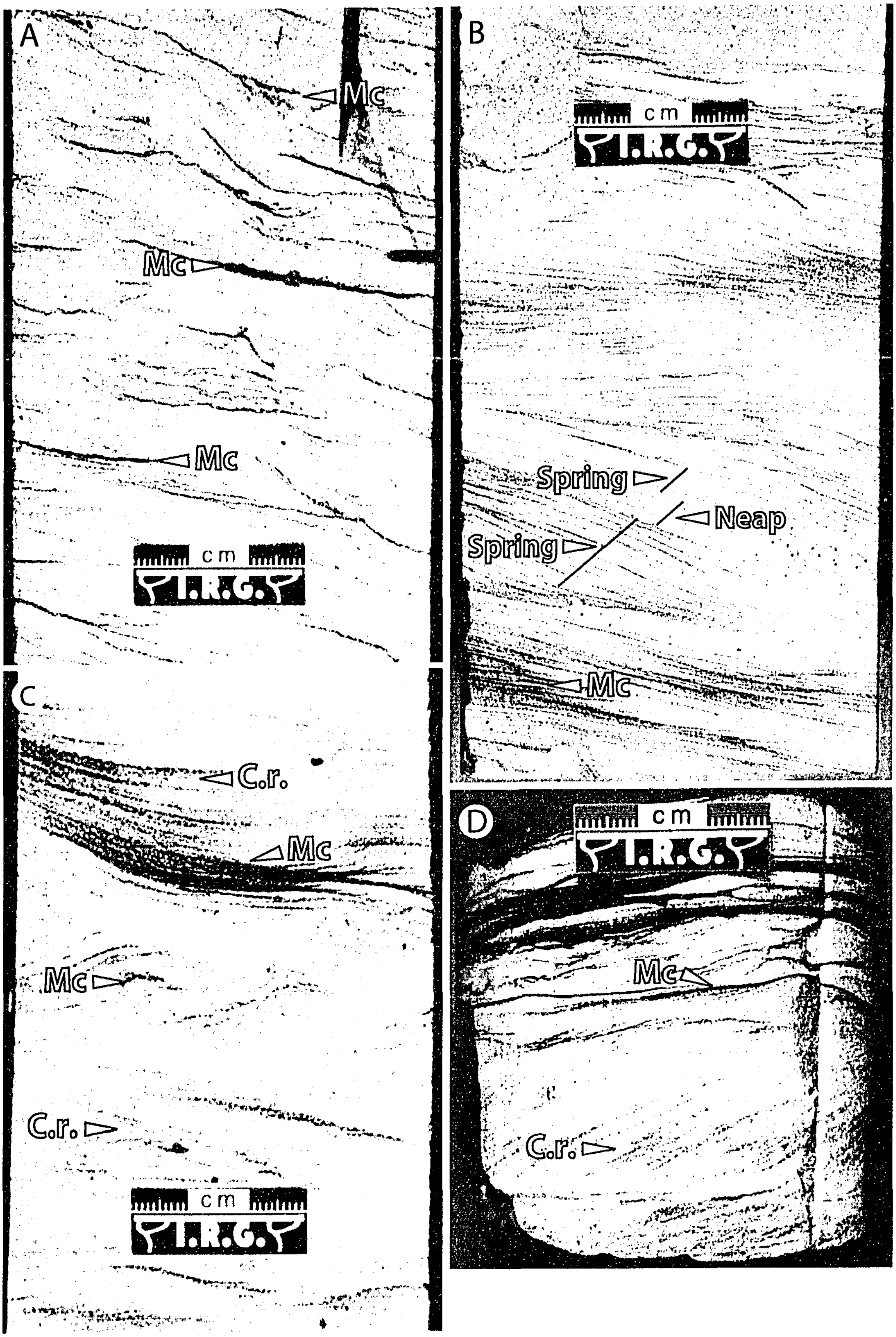
Facies J - Fine to medium-grained, low angle cross-stratified to flaser bedded sandstone and minor, dark grey siltstone and silty mudstone

A. Flaser bedded, current-rippled, fine-grained sandstone with wispy laminae of finely comminuted organic detritus (Mc) and phytodetrital material (Siku A-12, Core #1, depth ~2718.17 m).

B. Very fine-grained silty sandstone exhibiting sigmoidal bedding with preserved bottomsets and foresets. Tidal bundles, representing spring and neap, diurnal to semi-diurnal deposition are highlighted by organic material (Mc) (Siku A-12, Core #1, depth ~2719.43 m).

C. Fine-grained, silty sandstone with climbing ripple and current ripple (C.r.) laminae. Note the enrichment of organic material (Mc) in the troughs of the ripples (Siku A-12, Core #1, depth ~2719.08 m).

D. Fine to medium-grained sandstone with current ripple (C.r.) lamination marked by macerated organic detritus (Mc) (Siku E-21, Core #3, depth ~3053.25 m).


A
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D
cm
T.R.G.
Mc
C.r.
cm
T.R.G.

Figure 2.24

Facies J - Fine to medium-grained, low angle cross-stratified to flaser bedded sandstone and minor, dark grey siltstone and silty mudstone

A. Fine-grained, silty sandstone with wavy bedding and climbing current ripples. Micaceous material is preserved in the trough of the ripple bedforms. These climbing ripples display a relatively high degree of lateral migration and vertical aggradation, reflecting high sedimentation rates. Note the vertical *Diplocraterion* (Dp) and *Skolithos* (Sk) burrows cross-cutting the laminae and bedding (Parsons L-43, Core #1, depth ~2773.71 m).

B. Fine-grained sandstone with climbing wave-ripples (W.r.) where laminae are marked by finely comminuted organic detritus and phytodetrital material (Parsons L-43, Core #1, depth ~2776.05 m).

C. Soft-sediment deformed, medium-grained sandstone with coalified wood fragment (Wd) (Parsons L-43, Core #1, depth ~2774.66 m).

D. Fine-grained, silty sandstone exhibiting climbing current ripples (C.r.). Micaceous material is preserved in the trough of the ripples. These climbing ripples display a relatively high degree of lateral migration and vertical aggradation, reflecting high sedimentation rates (Parsons L-43, Core #1, depth ~2773.71 m).

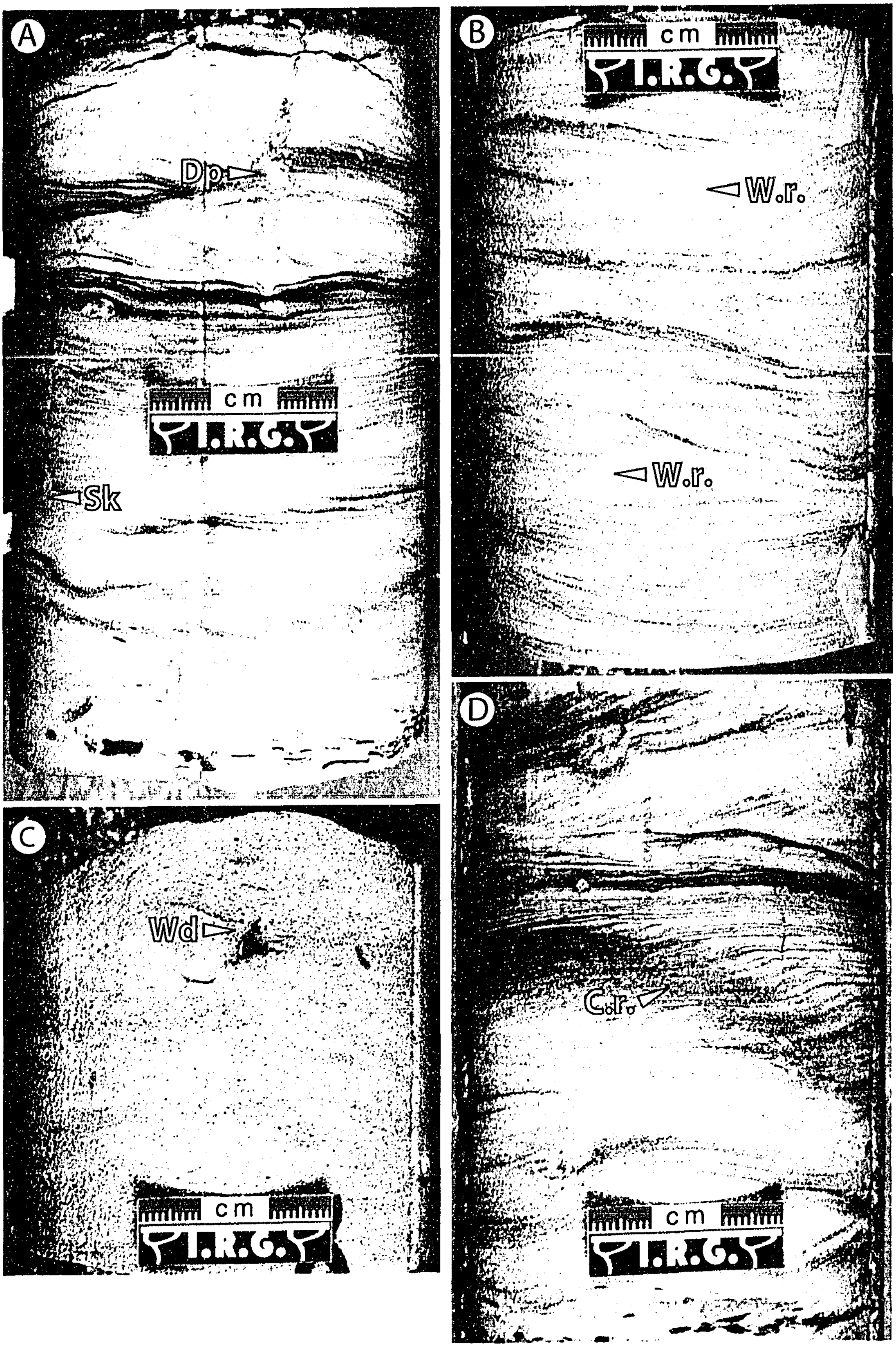

A
Dp
cm
Sk
B
cm
W.r.
W.r.
C
Wd
cm
D
C.r.
cm

Figure 2.25

Facies J - Fine to medium-grained, low angle cross-stratified to flaser bedded sandstone and minor, dark grey siltstone and silty mudstone

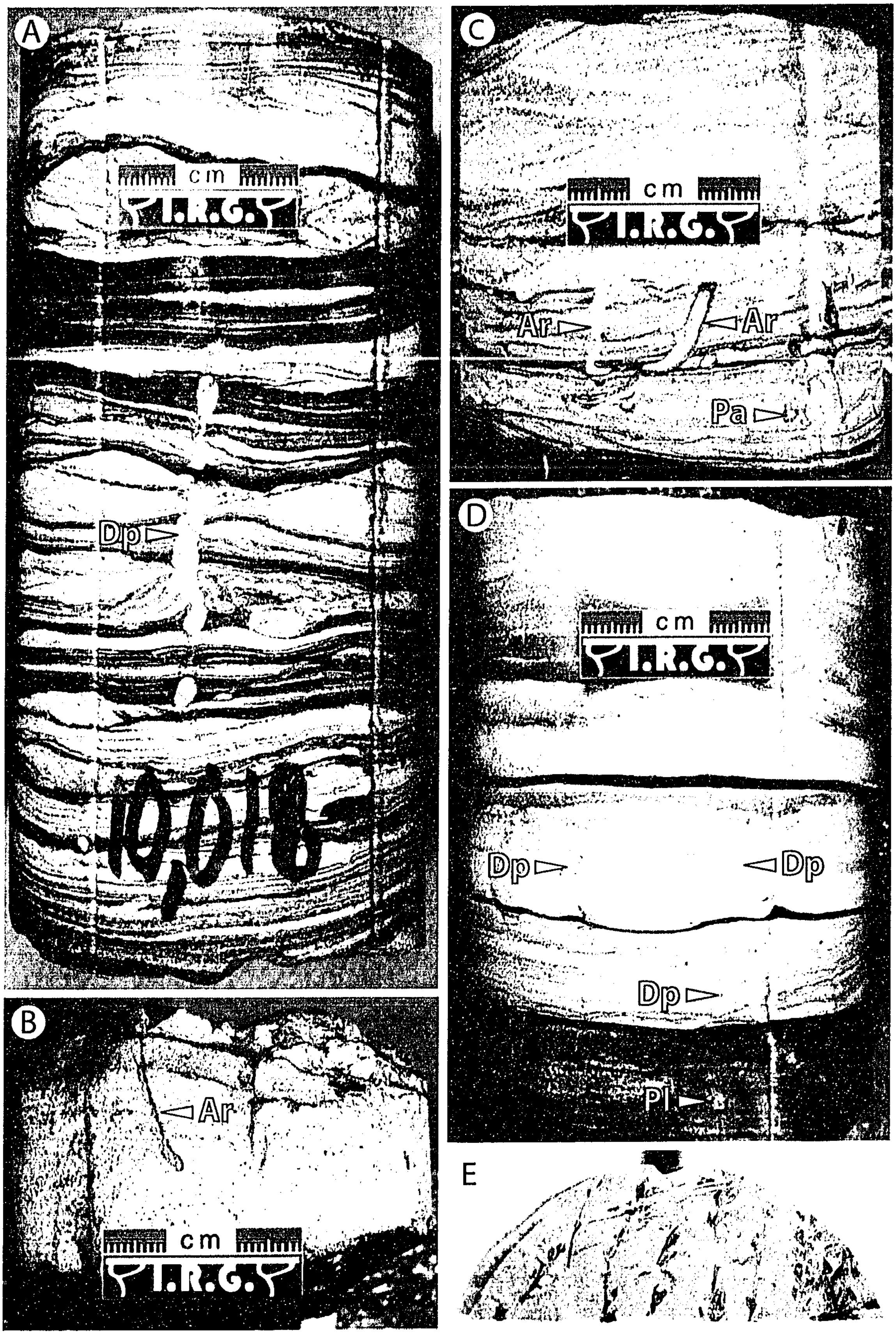
A. Wavy bedded, fine-grained sandstone and organic mudstone with large *Diplocraterion* (Dp) cross-cutting the laminae (Siku E-21, Core #3, depth ~3053.51 m).

B. Medium-grained sandstone with a partial cut through *Arenicolites* (Ar) (Parsons L-43, Core #3, depth ~2939.65 m).

C. Medium-grained, current rippled sandstone with minor wavy bedding. Note the numerous *Arenicolites* (Ar) cross-cutting the laminae and a single *Palaeophycus* (Pa) (Siku E-21, Core #3, depth ~3054.23 m).

D. Fine to medium-grained, silty sandstone with numerous *Diplocraterion* (Dp) overlying a highly organic, black mudstone with *Planolites* (Pl) (Siku E-21, Core #3, depth ~3048.34 m).

E. Bedding plane view of carbonaceous fern (?) leaf imprints within a coalified mudstone (Siku A-12, Core #1, depth ~2717.73 m).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T.R.G.
Dp
Dp
Dp
Pl
E

Figure 2.26

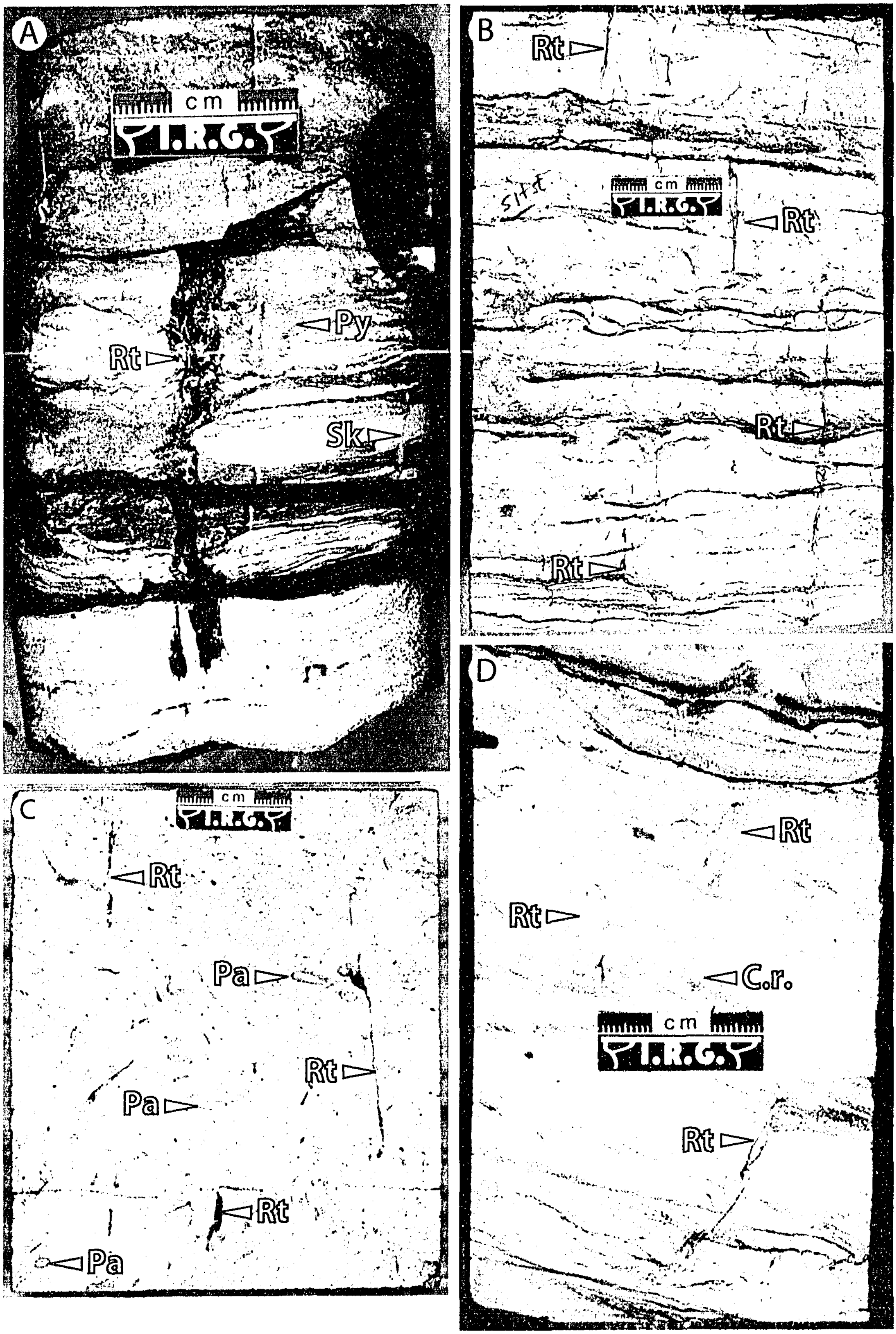
Facies J - Fine to medium-grained, low angle cross-stratified to flaser bedded sandstone and minor, dark grey siltstone and silty mudstone

A. Wavy to flaser bedded, fine to medium-grained, silty sandstone with a large coalified rootlet (Rt) and *Skolithos* (Sk) cross-cutting the bedding. Note the large pyrite (Py) nodule associated with the coalified rootlet (Siku E-21, Core #3, depth ~3052.65 m).

B. Flaser bedded, fine-grained sandstone cross-cut by numerous coalified rootlets (Rt). Note the wispy, downward tapering nature of the rootlets (Parsons N-10, Core #2, depth ~2800.66 m).

C. Fine-grained sandstone with coalified rootlets (Rt) and *Palaeophycus* (Pa) (Parsons N-10, Core #2, depth ~2800.96 m).

D. Current rippled, fine-grained, silty sandstone with bedding cross-cut by coalified rootlets (Rt). Note the abundant current ripple foresets (C.r.) marked by finely comminuted organic detritus (Parsons N-10, Core #2, depth ~2717.35 m).


A
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<u>Interpretation and Discussion:</u> Crevasse Splay (Interdistributary Bay Fill)

Fine to medium-grained, low angle cross-stratified to flaser bedded sandstone, minor dark grey siltstone and mudstone of Facies J represent crevasse splay deposits from a lower delta plain environment. The sandstone-dominant lithology, coarsening-upward nature, abundance of carbonaceous detritus, episodic nature of sedimentation, and thickness (1 to 3 m) represent high rates of episodic sedimentation characteristic of crevasse splay deposits. These sediments were rapidly deposited in a low to moderate-energy setting where sand, silt and mud were readily available. Wavy and flaser bedding styles reflect an environment where mud and sand is deposited under suspension and traction and saltation processes (Reineck and Singh, 1980). Flaser bedding is characterized by ripple bedding in which finely comminuted organic detritus (coffee grounds) is preserved completely in the troughs and partly on the crests as flasers. This type of bedding generally forms during periods of current activity alternating with periods of quiescence. During periods of current activity, the sand is transported and deposited as ripples, while the organic material and mud is held in suspension (Reineck and Singh, 1980). Once the current passes, the finely comminuted organic detritus held in suspension flocculates and is deposited mainly in the troughs or can completely cover the ripples. During the next cycle, the ripple crests are eroded away and new sand is deposited as ripples, burying and preserving the ripple beds with the mud and organic flasers in the troughs (Reineck and Singh, 1980). Flaser bedding is commonly produced in environments that have an abundant source of sand and where the preservation of sand is more favourable than for mud or organic material.

The ripple cross-stratified sandstone beds are deposited in a low to moderate-energy shallow water setting dominated by traction currents with minor suspension processes. Asymmetrical ripples are interpreted as current-ripples formed under unidirectional current flow (Harms *et al.*, 1975). The symmetrical ripples with foresets that display bipolar directions are interpreted as wave ripples. These wave-formed structures are interpreted to form in an interdistributary bay under the occasional influence of oscillatory current processes from the open marine realm, which are able to rework the sands deposited by crevasse splaying. The climbing ripple lamination represents aggrading current ripples, which form in a setting with high sedimentation rates, typical of a crevasse splay deposit. Crevasse splay deposits are typically characterized by waning flow near the end of sheet flood conditions resulting in current ripple lamination, which is commonly aggradational due to high sediment content (MacEachern, 2001). The abundant soft-sediment deformation structures also result from rapid deposition of sandy sediments onto soupy, unconsolidated, muddy substrates. The sigmoidal beds reflect deposition in a tidally influenced portion of the interdistributary bay. Shanley *et al.* (1992) interpreted sigmoidal beds as tidal bundles that were deposited and modified in response to neap-spring-neap tide fluctuations. The finely comminuted organic detritus or "coffee grounds" that highlight the sigmoidal beds are attributed to deposition during relatively slack water periods. Individual tidal bundles probably formed in response to migration during a single tidal cycle at or near spring tide. Tidal bundles have been identified from the more open, sand-dominated portions of estuaries and bays (Nio and Yang, 1991).

The trace fossil assemblage of Facies J is a mixed assemblage, containing both suspension-feeding and dwelling traces such as *Skolithos, Arenicolites, Diplocraterion, Palaeophycus* and feeding structures including *Planolites, Teichichnus* and resting traces such as *Bergaueria.* This assemblage corresponds mainly to the *Skolithos* ichnofacies and is indicative of deposition in a low energy, shallow, subtidal conditions. The relatively low to moderate diversity of morphologically simple trace fossils suggests that the environment was restricted (Ekdale *et al.*, 1984; Wightman *et al.*, 1987; Beynon *et al.*, 1988). The low to moderate diversity of traces suggests that the depositional environment was characterized by numerous biological stresses including high sedimentation rates, salinity variations and turbid water. MacEachern and Pemberton (1992) noted that *Arenicolites* and *Diplocraterion* burrows are generally associated with arenaceous substrates in low energy sandy tidal flats. *Bergaueria* is generally indicative of normal marine conditions, however it can also found associated with the brackish-water environments with smaller morphologies than occur in the fully marine realm. Some of the sandstones contain vertical escape structures (fugichnia), which reflect the behaviour of benthic organisms rapidly covered by sediment. Following burial the organisms must escape upwards to the newly formed sediment-water interfaces in order to respire and ingest food (MacEachern, 2001). Escape traces are most common in settings subjected to very rapid (episodic sedimentation) and is consistent with deposition in an environment characterized by rapid deposition by crevassing. The trace fossils present within this facies are consistent with deposition in a proximal crevasse splay infilling a brackish to fully marine interdistributary bay environment. This assemblage is similar to Facies I, however it displays a slightly lower diversity and abundance of trace fossils attributed to the rapid sedimentation rates and higher energy conditions associated with crevasse splaying.

The rooted, interbedded, silty sandstone comprising the upper portion of this facies indicates a shoaling upward trend, and reflects infilling of the bay and the initial establishment of a marsh environment on previously deposited sediments crevasse splay sediments. Rhizoliths in sandstones are generally indicative of subaqueous colonization by plants, although marginal marine or terrestrial rooting is also common (MacEachern, 2001). Subaerial exposure of substrates is commonly associated with rapid podsolisation resulting in low preservation potentials of rootlets. The presence of abundant rhizoliths reflects deposition in a shallow, subaqueous environment where sediment deposition was intermittent and vegetation was able to establish itself. This vegetation was likely tolerant to saline or at least brackish conditions and able to withstand prolonged inundation (Suter, 1994). In addition, the shallow water overlying the sediment was brackish to marine due to the association of rootlets and *Palaeophycus*. The overprinting of rootlets in the *Palaeophycus*-burrowed sandstone is likely due to fluctuations in water levels during deposition, resulting in conditions shallow enough to allow the establishment of a rooted horizon. The top of a crevasse splay deposit represents an ideal low energy setting allowing for the establishment of vegetation and their preservation as rooted sandstones. Finally, the crevasse splay sediments of Facies J are commonly interstratified with the interdistributary bay deposits of Facies I and the marsh deposits of Facies L.

Facies K - Well-burrowed, interbedded, fine-grained silty sandstone and silty mudstone with abundant siderite

Lithology and Sedimentary Features:

Facies K consists of a burrowed, interlaminated/interbedded, light to dark grey mudstone, silty mudstone and light grey to buff, upper very fine-grained silty sandstone (Figs. 2.27A-D; 2.28A-D). Silty, very fine-grained, laminated sandstone is dominated by flaser and wavy parallel bedding with minor lenticular bedding and are well burrowed (Fig. 2.27C). The mudstones are silty, organic rich, finely-laminated, and also well-burrowed. Sedimentary structures also include rare synaeresis cracks, convolute bedding, asymmetrical and symmetrical ripples.

This facies is characterized by abundant siderite beds up to 6 cm thick and rounded siderite nodules 2 to 5 cm in diameter (Figs. 2.27B,D; 2.28A,D). The thicker beds of siderite may represent large, flattened concretions (unknown due to the limited diameter of core). The siderite beds are also associated with abundant centimeter-scale pyrite nodules. Carbonaceous organic material commonly occurs in the form of disseminated macerated coal and wood fragments, plant debris, "coffee ground" material and micaceous material. There are also coalified rootlets that display cross-cutting relationships with laminae in sandstone intervals. This facies displays an overall coarsening-upward succession with the proportion of mudstone and very-grained sandstone increasing toward the top of each succession.

Facies K is only identified in core #1 from the A-12 well. The lower contact is generally sharp with underlying Facies J and is marked by an overall decrease in sandstone content, grain size and ichnogenera, along with an increase in the proportion of mudstone. The upper contact of this facies is not preserved in core, thus the nature of this contact with overlying sediments is unknown. Facies K is found capping a coarsening-upward succession composed of Facies N, O and J and is 3.41 m thick.

Ichnological Characteristics:

This facies has a markedly reduced burrowing intensity, abundance and diversity compared to that of Facies I. Trace fossils are sporadically distributed and most intervals display a low to degree of bioturbation, however some intervals are dominated by *Teichichnus* resulting in the complete destruction of bedding (Figs. 2.27A,C). The main ichnological elements are deposit-feeding structures, as well as rare suspension-feeding and dwelling traces. Identified trace fossils mainly include large, robust *Teichichnus* (Figs. 2.27A; 2.28B) with *Diplocraterion* (Figs. 2.28A,C), *Planolites* (Fig. 2.27C), *Palaeophycus, Skolithos* (Figs. 2.28B,C) and rare fugichnia. The majority of the burrows display diminutive, simple forms, with the exception of *Teichichnus*. In addition to the trace fossils, root traces are also particularly common within this facies (Fig. 2.28B). These rootlets are thin, wispy, taper downward, and cross-cut the laminae and bedding structures. They are also sinuous in nature, commonly thicken and thin along their lengths and range in length from 3 to 16 cm and 1 to 2 mm in diameter.

<u>Interpretation and Discussion:</u> Restricted/Brackish Interdistributary Bay

Interbedded, burrowed, siderite-rich silty mudstone, mudstone and fine-grained, silty sandstone of Facies K represent the infilling of a restricted/brackish interdistributary bay environment in the lower deltaic plain. Distinguishing characteristics that suggest a brackish bay environment include the prevalence of mudstone and low diversity ichnofossil assemblage. The deposits of Facies I and K have similar physical sedimentary structures; however, the main difference between the two facies includes the diversity and abundance of the ichnological suite. Trace fossils can often provide the least equivocal, and most readily accessible data for to assist with the recognition of brackish water deposits (Ekdale *et al.*, 1984; Wightman *et al.*, 1987; Beynon *et al.*, 1988; Pemberton and Wightman, 1992).

Flaser to finely, wavy laminated, very fine-grained sandstone and mudstone of Facies K is indicative of a low energy depositional environment dominated by slow suspension and rare traction processes. Wavy and flaser bedding is characteristic of low energy settings, suggests that both sand and mud are available for deposition, and that periods of current activity alternated with periods of quiescence (Reineck and Singh, 1980). Deposition of the fine-grained, flaser to wavy bedded sandstone with associated ripples is due to punctuated, higher energy events in a quiescent, brackish bay environment. Processes such as crevasse splaying and overbank deposition during periods of high fluvial discharge can distribute sediment rapidly into the interdistributary bay environment (Elliott, 1974; Coleman and Prior, 1980, 1982). Facies J is similar to Facies I and likely represents the distal portion of splay deposits. The presence of rooted intervals suggests that the setting was a stable, low energy setting where vegetation was able to establish itself. The association of rootlets with *Teichichnus* suggests that plant systems were established subaqueously in an environment with marine or at least brackish waters.

The abundance of siderite within Facies K as both concretions and decimeter-scale thick beds has important implications for understanding the conditions present during deposition. Siderite is indicative of littoral, shallow-water strata that contain an abundance of organic material

Figure 2.27

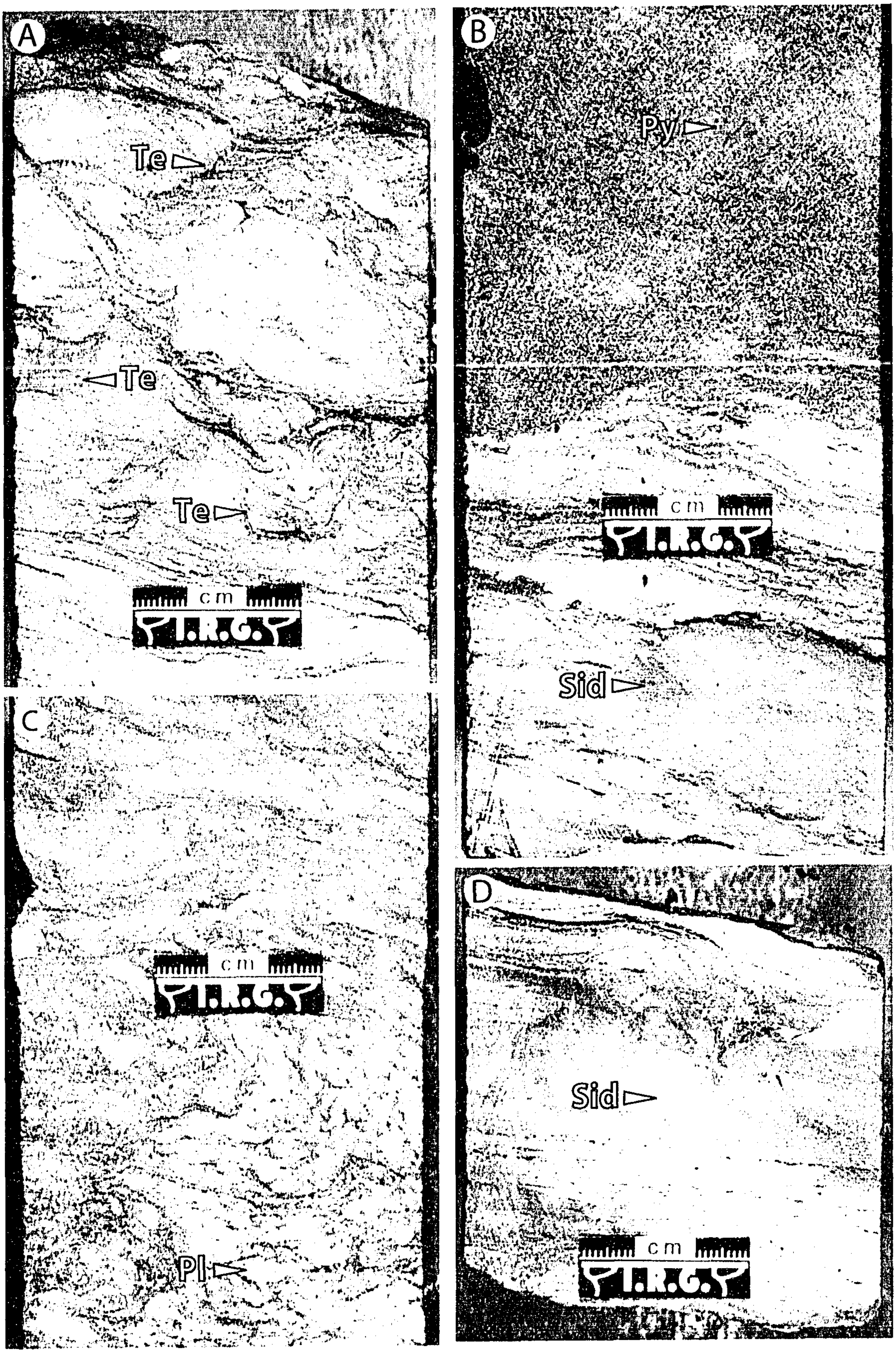
Facies K - Well-burrowed, interbedded, fine-grained silty sandstone and silty mudstone with abundant siderite

A. Flaser to wavy bedded, very fine-grained sandstone and minor organic mudstone with large *Teichichnus* (Te) cross-cutting the laminae (Siku A-12, Core #1, depth ~2712.92 m).

B. Wavy bedded, very fine-grained sandstone overlain by a silty mudstone. Note the large siderite (Sid) and pyrite (Py) nodule in the lower, wavy bedded, very fine-grained sandstone (Siku A-12, Core #1, depth ~2713.83 m).

C. Very fine-grained sandstone with a biogenically mottled texture that resembles soft-sediment deformation structures (Siku A-12, Core #1, depth ~2712.66 m).

D. Sideritized, low angle cross-bedded, fine-grained silty sandstone (Siku A-12, Core #1, depth ~2712.66 m).


A
Te
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cm
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Sid
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Pl
D
Sid
cm
I.R.G.

Figure 2.28

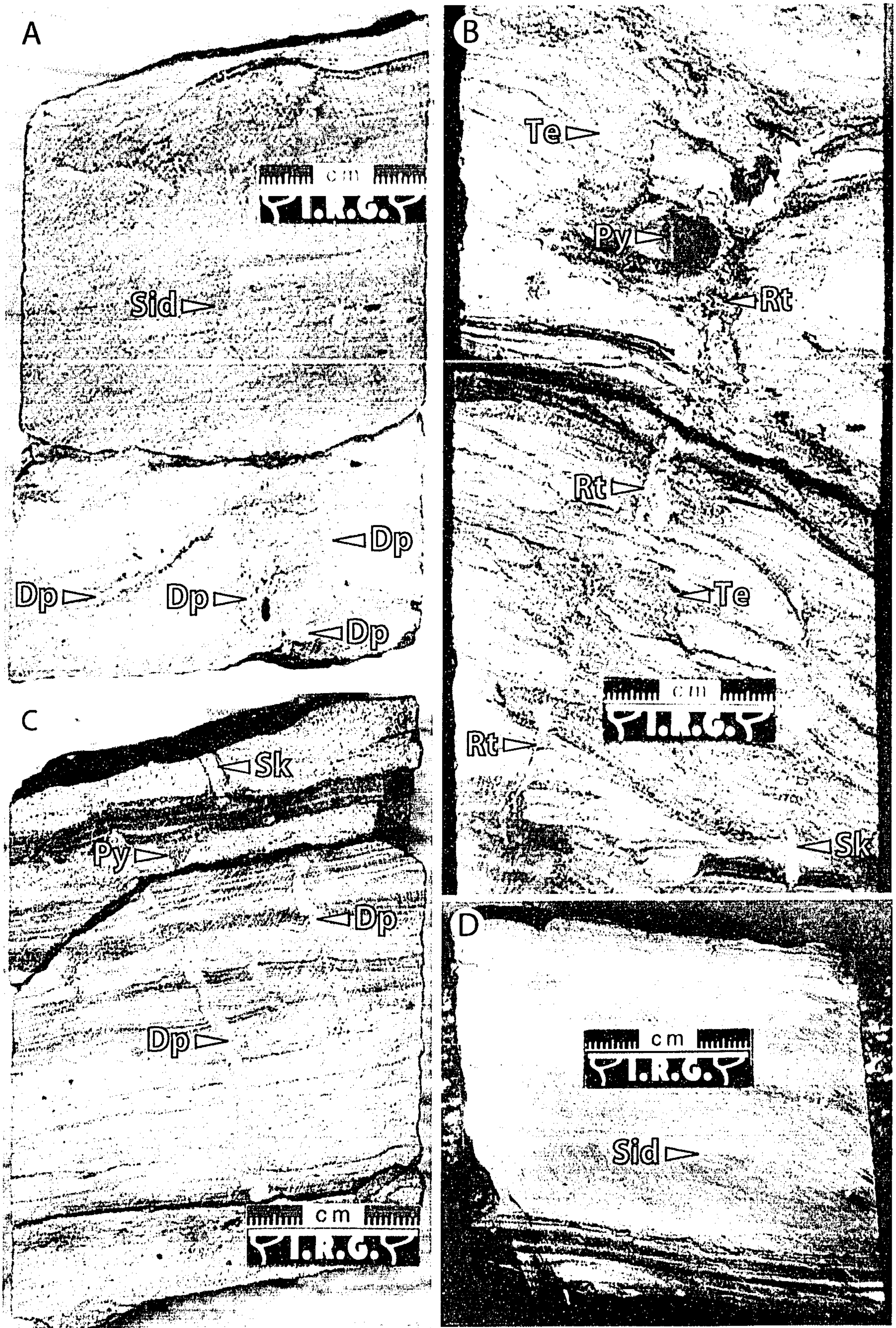
Facies K - Well-burrowed, interbedded, fine-grained silty sandstone and silty mudstone with abundant siderite

A. Sideritized (Sid), low angle cross-bedded, fine-grained sandstone with minor organic mudstone exhibiting with multiple large *Diplocraterion* (Dp) cross-cutting the laminae (Siku A-12, Core #1, depth ~2714.30 m).

B. Flaser to wavy bedded, silty, very fine-grained sandstone containing *Teichichnus* (Te), *Skolithos* (Sk) and rootlets (Rt). Note the numerous pyrite nodules (Py) (Siku A-12, Core #1, depth ~2714.23 m).

C. Wavy bedded, very fine-grained sandstone with multiple *Diplocraterion* (Dp) and *Skolithos* (Sk) (Siku A-12, Core #1, depth ~2714.47 m).

D. Sideritized (Sid), very fine-grained, silty sandstone with minor wavy bedding near the base (Siku A-12, Core #1, depth ~2715.03 m).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Sid

(Teodorovich, 1961). Baird *et al.* (1986) suggested that the presence of siderite requires three key conditions. These include: 1) availability of iron, b) rapid burial of organic material, and c) low to non-existent supply of seawater sourced sulphate, required for methanogenesis. These conditions are commonly found in both fresh water and brackish water environments (Baird *et al.*, 1986). In marine waters with abundant sulphate, iron sulphides are produced at the expense of iron carbonates. However, in fresh water environments that are depleted in sulphate, the formation of iron sulphide is precluded, and the precipitation of iron carbonates such as siderite is favoured. Likewise, the formation of siderite requires a low Eh, which can be produced by anaerobic bacterial decomposition of organic matter (Berner, 1971). It is believed that the decomposition of organic material forms carbon dioxide, due to mixing of marine waters with fresh water, combines with ferrous iron to form siderite ($FeCO_3$). This process continues until all organic matter has been destroyed, or all the sulphate has been used up. Macerated organic detritus, coal fragments and coalified rhizoliths within Facies K are possible sources of the organic material required for siderite formation. Coleman and Gagliano (1965) identified that siderite concretions were common in the lower delta plain deposits of the Mississippi Delta and were developed around plant and root fragments. For a further discussion of the formation of siderite, the reader is referred to Teodorovich (1961), Baird *et al.* (1986), Berner (1971) and Maynard (1984). The co-existence of siderite and pyrite is thought to be indicative of fluctuating salinities within marginal marine settings (Wightman *et al.*, 1987; Hubbard *et al.*, 1999) and further supports the interpretation of this facies as a brackish, interdistributary bay deposit. Siderite has been recognized to be common in modern salt water marsh environments (Pye, 1981; Postma, 1981) and sideritic sediments have been recognized in cores from the Atchafalaya Bay in the modern Mississippi delta plain (Ho and Coleman, 1969).

The trace fossil assemblage of Facies K, comprised of large, robust *Teichichnus* with rare *Diplocraterion*, *Planolites*, *Palaeophycus*, and *Skolithos*, is interpreted to reflect the general traits of a brackish water assemblage as described by Pemberton and Wightman (1992) and MacEachern and Pemberton (1992). Brackish water trace fossil suites reflect inherently fluctuating environmental parameters, are characterized by low diversities and are generally composed of an impoverished marine assemblage consisting of size-reduced, diminutive forms relative to those found in fully marine environments (Howard and Frey, 1983; Wightman *et al.*, 1987; Pemberton and Wightman, 1992; Hubbard *et al.*, 1999; MacEachern and Pemberton, 1992; Pemberton *et al.*, 2001). In addition, they are characterized by simple vertical and horizontal structures constructed by trophic generalists, common to both the *Skolithos* and *Cruziana* ichnofacies (Pemberton and Wightman, 1992; MacEachern and Pemberton, 1992).

Brackish water environments such as interdistributary bays represent ecologically stressed habitats that are characterized by a low diversity suite of morphologically simple trace fossils, typically dominated by high densities of one or two ichnogenera (MacEachern and Pemberton, 1992;

Pemberton *et al.*, 2001). The dominance and almost monospecific nature of *Teichichnus* within Facies K is consistent with the observations of Pemberton and Wightman (1992), Wightman *et al.* (1987) and MacEachern and Pemberton (1992) for brackish water deposits such as interdistributary bays. Deposit-feeding burrows such as *Teichichnus* and *Planolites* are produced by trophic generalists and combine to form biogenic fabrics exhibiting overall low diversity with high individual densities. The overall lower diversity and diminutive nature of burrow forms in this facies indicates a setting more biologically stressed than that of the open interdistributary bay deposits of Facies I. The combination of the sedimentological and ichnological characteristics of Facies K strongly suggests deposition in a restricted, brackish, interdistributary bay setting.

Facies L - Dark grey to black, coaly mudstone with rootlets

Lithology and Sedimentary Features:

Facies L is a dark grey to black, blocky, highly carbonaceous mudstone with minor, interbedded coal beds (Figs. 2.29A-D; 2.31A,C). Thin, discontinuous, 1 to 4 cm thick "stringers" of siltstone and very fine to lower fine-grained sandstone are commonly interstratified with the carbonaceous mudstones. No significant sedimentary structures are identifiable, although there is faint parallel lamination within the mudstone. This mudstone is characterized by abundant organic detritus including in-situ rootlets, centimeter-scale coal fragments, and plant fragment imprints (Figs. 2.29A-D; 2.31A,C). Some intervals contain abundant nodular and disseminated pyrite with occasional siderite nodules.

Facies L is identified in five cored intervals including cores #1 and #3 from E-21, core #3 from L-43, core #1 from F-09 and core #2 from the N-10 well. It is found interstratified with Facies H, I, J, M and rarely C. Both upper and lower contacts with Facies H, I, J and M are generally gradational to sharp, but in core #1 from the F-09 well, the upper contact is scoured to undulatory and marked by a 3 cm thick pebble lag. Facies L commonly overlies the coarsening-upward successions composed of Facies J and I and varies in thickness from 0.47 to 1.23 m.

Ichnological Characteristics:

The only trace fossils identified within Facies L are rhizoliths (Figs. 2.29A,B,C; 2.31C). These root traces share similar characteristics to those identified in Facies J; however, they occur in a carbonaceous mudstone with abundant coalified wood fragments. These rootlets are carbonaceous, vertical, wispy, thin, taper downward and display a branching nature. In addition, they are sinuous in nature, commonly thin and thicken along their lengths and cross-cut the laminae and bedding structures. A few rootlets display distinct daughter roots and root hairs, which are smaller than the primary root but also taper downwards. On average, the roots range in length from 3 to 6.5 cm with

Figure 2.29

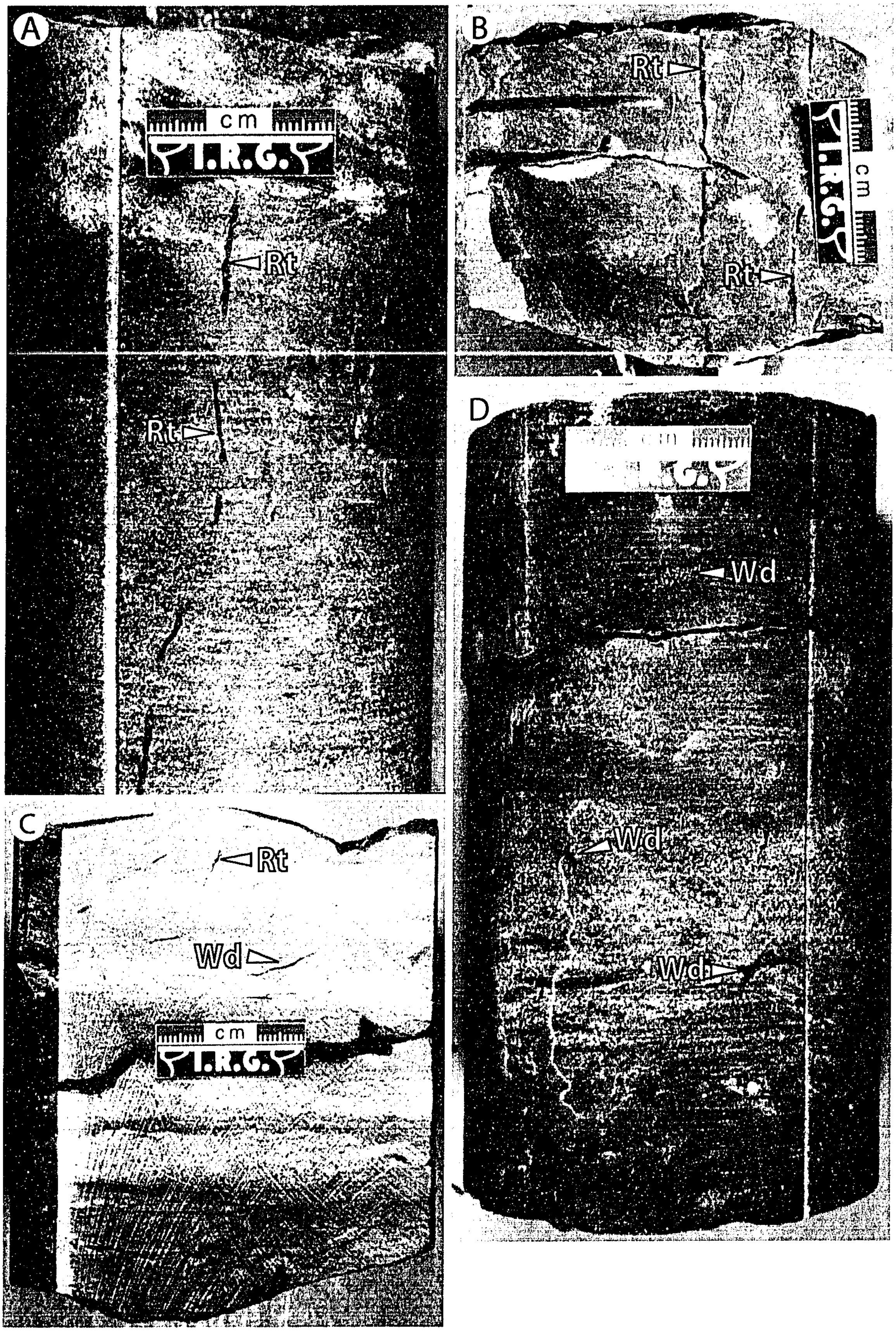
Facies L - Dark grey to black, coaly mudstone with rootlets

A. Massively bedded, silty, highly organic mudstone cross-cut by a long (~15.5 cm) coalified rootlet (Rt) (Parsons L-43, Core #1, depth ~2935.63 m).

B. Massively bedded, silty, organic mudstone cross-cut by two long (~5.5 cm) coalified rootlets (Rt) (Parsons L-43, Core #1, depth ~2934.86 m).

C. Faint, low angle to massively bedded silty, highly organic mudstone with minor coalified rootlets (Rt) and coalified wood fragments (Wd) (Parsons F-09, Core #1, depth ~2863.90 m).

D. Massively bedded, silty, highly organic mudstone with coalified wood fragments (Wd) (Parsons L-43, Core #1, depth ~3052.45 m).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Wd
Wd

widths averaging 1 to 3 mm. They occur within the middle and upper portions of this facies. In core #3 from the L-43 well, a large carbonaceous root approximately 16.5 cm long and 3 mm wide is preserved within a dark grey carbonaceous mudstone (Fig. 2.29A). In core #3 from the E-21 well, the lower contact between Facies G and L is marked by 0.5 to 1 cm diameter *Thalassinoides* (Fig. 2.31A). The burrows are infilled with medium-grained sandstone of Facies G and extend downward from the base of Facies G into underlying coaly mudstone of Facies L.

<u>Interpretation and Discussion:</u> Marsh

Dark grey to black coaly mudstone interbedded with thin siltstone laminae, very fine-grained sandstone and minor coal beds of Facies L represent a marsh environment in the lower delta plain. Abundant in-situ carbonaceous rootlets, organic detritus and plant fragment imprints are indicative of very low energy conditions, transport distances and sedimentation rates. The formation and preservation of rootlets implies that these deposits were formed subaqueously in a shallow water setting that was favourable for the establishment and colonization by plants. The thin coal and coaly mudstone beds are also indicative of deposition in a dominantly non-marine to paralic environment. It has been recognized that sediments that are subaerially exposed are subjected to rapid podsolisation with generally low preservation potentials of rootlets (S.G. Pemberton pers. comm., 2004). The thin siltstone and fine-grained sandstone laminae and beds are transported from either the marine or non-marine realm. Although root traces indicate that plants colonized a surface, they do not indicate that genetic soil horizons were developed, or that the surface was subaerially exposed for prolonged periods of time.

The lack of burrowing and abundance of pyrite suggests harsh, reducing environmental conditions. However, it should be noted that pyrite might have also formed post-depositionally due to later diagenetic processes. The presence of *Thalassinoides*, representing the *Glossifungites* ichnofacies at the base of a channel sequence in core #3 from the E-21 well suggests that following the deposition of Facies L, and prior to the deposition of the channel fill deposits of Facies G, there was an incursion of marine waters that allowed the establishment of these burrows within the mudstone. This indicates that there was a depositional hiatus between deposition of the coaly mudstone of Facies L and the channel sandstone of Facies G. The upward gradation of Facies L into subaerially exposed rocks of Facies M also suggests deposition in a non-marine setting. The stratigraphic position of this facies relative to Facies J and I, the sedimentary and ichnological features strongly supports the interpretation of Facies L as a fresh water marsh that was rarely inundated by marine waters.

The *Glossifungites* trace fossil assemblage in this core occurs at a sharp contact between the black coaly mudstones of Facies L and the overlying unburrowed, massive channel sandstones of Facies G. The medium-grained channel sandstones of Facies G infill *Thalassinoides* within the coaly

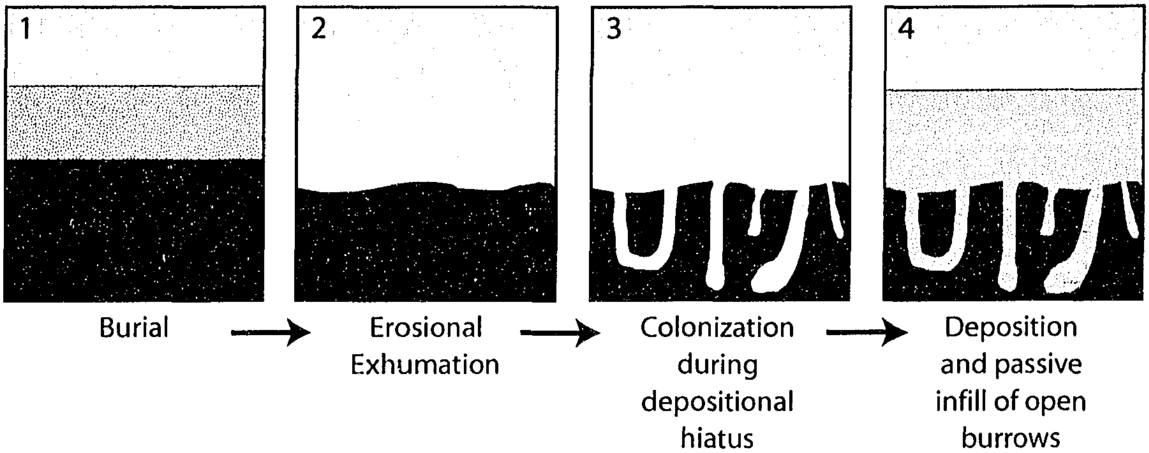


Figure 2.30 Schematic diagram illustrating the development of a *Glossifungites* demarcated erosional discontinuity. 1) The muddy substrate is initially buried, compacted and dewatered, resulting in a compacted, stiff character. 2) The mudstone bed is erosionally exhumed, exposing a firm substrate. 3) Colonization of the discontinuity surface of trace makers of the *Glossifungites* ichnofacies proceeds under marine conditions during a depositional hiatus. 4) The open burrow structures are passively infilled during a succeeding depositional episode (modified after Pemberton and Frey, 1984 and MacEachern *et al.*, 1992).

mudstones of Facies L (Fig. 2.31A). The sandstones infilling the *Thalassinoides* were piped down approximately 40 cm into the underlying black organic mudstones of Facies L through the passive infilling of large *Thalassinoides* from above. The burrows comprising the *Glossifungites* ichnofacies were formed by burrowing organisms during a period of non-deposition prior to deposition of the overlying channel sandstones and therefore marks a hiatal surface. Burrow traces mark a hiatal surface and develop on erosionally exhumed semi-lithified substrates including dewatered muds of (bays, lagoons and marshes) and other fine-grained deposits, which are ultimately colonized by organisms in the marine and marginal marine environments. The development of a *Glossifungites* demarcated erosional discontinuity is illustrated in Figure 2.30.

The substrate in which the burrows of the *Glossifungites* assemblage occur are made firm through the process of dewatering, which can form either due to subaerial exposure and desiccation, or by burial, compaction, and exhumation due to erosion (Pemberton and Frey, 1985; Gingras *et al.*, 2001; Pemberton *et al.*, 2001). Surfaces demarcated by the *Glossifungites* ichnofacies can be useful for genetic-stratigraphic interpretations due to their associations with allocyclic and autocyclic base level changes in marginal marine environments (Pemberton and Frey, 1985; MacEachern *et al.*, 1992; Pemberton *et al.*, 1992; Pemberton and MacEachern, 1995; Gingras *et al.*, 2000). The *Glossifungites* ichnofacies can represent significant stratigraphic surfaces, as it allows for the recognition of depositional hiatuses, and records changes in the depositional history of the intervals. These surfaces occur on both local or regional scales and can potentially represent minor scour surfaces, transgressive surfaces of erosion (TSE) or sequence boundaries. However mapping is required for the identification of these assemblages as important stratigraphic surfaces.

The *Glossifungites* ichnofacies identified in core #3 from the Siku E-21 well occurs at the base of a channel and thus, likely only represents a locally developed scour surface.

The development of the *Glossifungites* ichnofacies in this core indicates that the following deposition of the marsh mudstones of Facies L marine waters inundated the system prior to deposition of the distributary channel. This indicates that there was a hiatus between the deposition of the marsh mudstones of Facies L and the channel sandstones of Facies G. The surface marked by the *Glossifungites* ichnofacies is interpreted to represent a localized scour surface developed due to incision by the channel into the underlying marsh mudstones. The localized development of the *Glossifungites* ichnofacies in a single core suggests the surface it occurs on does not represent a significant surface with any a regional stratigraphic implications as noted by Pemberton and Frey (1985) and MacEachern *et al.* (1992).

Facies M - Dark grey to black, waxy, carbonaceous mudstone

<u>Lithology and Sedimentary Features:</u>

Facies M consists of dark grey to black, crumbly to slightly friable, highly carbonaceous mudstone exhibiting a waxy texture (Figs. 2.31B,D). Carbonaceous detritus is particularly abundant within this facies and includes coalified wood fragments and thin coal beds. Additional features include variable colour, glossy slickensides, abundant plant fragments and thick pyrite beds (Figs. 2.31B,D). This facies is very commonly found as pieces of rubble that easily falls apart with even minimal handling. Coal beds interstratified within these sediments are closely associated with rooted intervals of Facies L.

Facies M is only identified from two cores including core #1 from E-21 and core #3 from L-43 well. This facies directly overlies and is commonly interstratified with Facies L. The contact is gradational to sharp and is transitional into waxy shales, thick pyritized beds and coal beds. Facies M varies in thickness from 0.47 to 1.23 m.

<u>Ichnological Characteristics:</u>

No trace fossils were identified within this facies.

<u>Interpretation and Discussion:</u> Paleosol/Subaerially Exposed Marsh (Lower Delta Plain)

Dark grey to black, waxy, carbonaceous mudstone of Facies M is interpreted as a paleosol deposit formed in a subaerially exposed portion of a marsh from the lower delta plain environment. Paleosols generally develop during periods of prolonged subaerial exposure. Podsolisation occurs

Figure 2.31

Facies L - Dark grey to black, coaly mudstone with rootlets

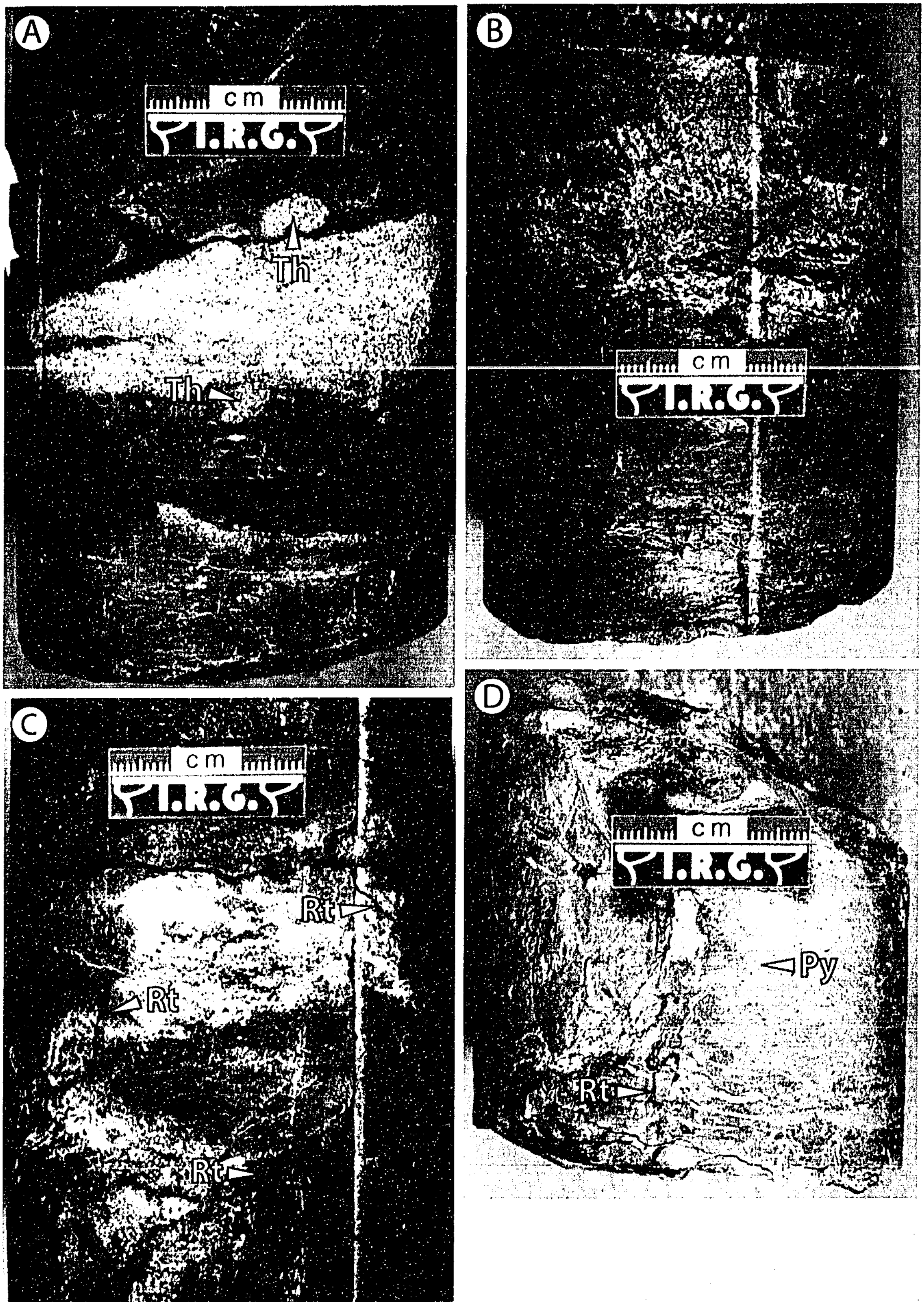
A. Medium-grained sandstone with *Thalassinoides* (Th) infilled with a medium-grained sandstone within a black, highly organic mudstone. The sand-filled burrows are characteristic of the *Glossifungites* ichnofacies. The sands infilling the *Thalassinoides* (Th) were piped down approximately 40 cm into the underlying mudstone from an overlying channel sandstone. The *Thalassinoides* (Th) are interpreted to have been created by organisms during a period of non-deposition before the overlying channel sandstones were deposited. This surface therefore marks a hiatal surface and likely represents a locally developed scour surface (Siku E-21, Core #3, depth ~3047.03 m).

C. Unburrowed, highly organic, black mudstones with interspersed medium-grained sandstone with large carbonaceous rootlets (Rt) (Siku E-21, Core #3, depth ~3047.24 m).

Facies M - Dark grey to black, waxy, carbonaceous mudstone

B. Massively bedded, highly organic, waxy mudstone (Parsons L-43, Core #1, depth ~2866.02 m).

D. Massively bedded, pyritized (Py), coaly mudstone with a possible rootlet (Rt) (Parsons L-43, Core #1, depth ~2865.96 m).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due to varying degrees of exposure, wetting and drying, drainage patterns and oxidation (Wright, 1986). The waxy texture, variable colour, rubbly appearance and presence of slickensides are characteristic features of pedogenic alteration. Abundant, thick, pyrite beds and nodules suggest that these deposits formed in a setting characterized by reducing conditions. Early diagenetic products such as pyrite and siderite are found in abundance within marshes in the modern Mississippi delta (Coleman and Prior, 1980). The absence of trace fossils and indications of reducing conditions suggests overall harsh environmental conditions that precluded the colonization of these deposits by infaunal organisms. In addition, Facies M and L are genetically related, suggesting that these deposits likely formed due to autocyclic processes and simply represent the non-marine deposits of a prograding delta.

Facies N - Interbedded, unburrowed, low angle, parallel-laminated, soft-sediment deformed, fine-grained sandstone, siltstone and silty organic mudstone

<u>Lithology and Sedimentary Features:</u>

Facies N consists of light grey to buff, fine-grained sandstone interbedded with siltstone and organic mudstone (Figs. 2.32A-D; 2.33A-D; 2.34A-C). The sands are characterized by low angle (<5° to 15°) parallel lamination, ripple cross-lamination and abundant soft-sediment deformation structures. The lower portion of this facies is dominated by decimeter thick intervals of convolute, slumped, oversteepened and curvilinear bedding, (Figs. 2.32A-D; 2.34A,B) as well as ball and pillow structures and load casts (Fig. 2.34A). Angular sandstone rip-up clasts and small, cm-scale fractures are also identified (Figs. 2.32A,B). Low angle, parallel-laminated sandstone is typically found interbedded with deformed intervals. Large-scale, aggrading, symmetrical and asymmetrical ripple lamination occurs in decimeter-scale (averaging 6 to 15 cm) bedsets of anticlinal or synclinal bedforms that have wavelengths larger than the core diameter (Figs. 2.32D; 2.33C). These dunes display stacked troughs and crests with minimal crest migration. Micaceous material is concentrated in the troughs, and less commonly on the crests of the structures. Interstratified with these large-scale dune structures are also excellent examples of symmetrical ripples, commonly displaying ripple-drift and climbing ripple cross-lamination (Figs. 2.33A,B).

Organic detritus in the form of finely comminuted phytodetrital and micaceous material is moderately abundant (5% to 10%) within Facies N. This organic material is also commonly found concentrated in the troughs of the symmetrical ripples (Figs. 2.33A,B). Numerous centimeter to decimeter thick, unburrowed, massively bedded to structureless, sharp-based and sharp-topped, highly organic mudstone beds, 5 to 23 cm thick, are commonly interspersed throughout this facies (Fig. 2.34C). Commonly sideritized, mudstone rip-up clasts with lengths averaging 0.5 to 2 cm are abundant in the upper portion of Facies N. Disseminated and nodular pyrite is particularly abundant at upper and lower contacts between the mudstone and sandstone beds.

Facies N is only found in core #1 from the A-12 well. The lower contact is not preserved however, from geophysical well logs it is interpreted to be gradational with distal deposits. The contact between Facies N and overlying Facies O is scoured and marked by an increase in grain size and sandstone content, along with a decrease in mudstone content and soft-sediment deformation structures. In core, Facies N is 7.09 m thick, but this thickness is limited by the maximum depth penetration of the core; it is likely that a thicker succession exists.

Ichnological Characteristics:

No trace fossils were identified within this facies.

Interpretation and Discussion: Distal Delta Front (River-Dominated Delta Front)

Unburrowed, interbedded, low angle, parallel-laminated, soft-sediment deformed, fine-grained sandstone, siltstone and mudstones of Facies N are interpreted as distal delta front deposits of a river-dominated delta. The distal delta front begins at the lower limit of fairweather (minimum) wave base where wave energy dominates producing structures reflecting storm deposition (MacEachern and Pemberton, 1992). In an interdeltaic shoreface setting, these deposits would be equivalent to the lower shoreface.

Sedimentary structures within the distal delta front suggest deposition in a low to moderate-energy setting characterized by rhythmic sedimentation events (i.e. seasonal floods) with a dominance of suspension, traction, and gravity induced processes in combination with marine and wave processes (Moslow and Pemberton, 1988). Low angle (<5° to 15°), parallel-laminated sandstone reflects hummocky cross-stratification (HCS) (Duke, 1985; Duke *et al.*, 1991) and/or swaley cross-stratification (SCS) (Leckie and Walker, 1982). Symmetrical ripples represent wave ripples formed in a setting dominated by oscillatory flow and negligible current drift (Harms *et al.*, 1975). Coarse material accumulates on the crest of wave ripples while organics and fines accumulated in the troughs. These wave ripples were differentiated from current ripples by their symmetrical slopes with rounded troughs and pointed crests. They probably represent weaker storm deposits or waning storm energies. The climbing ripples displaying ripple-drift cross-lamination reflect aggradational bedforms in which sediment was deposited at higher rates than the migration of the ripple crest (Walker, 1963). Ripple-drift cross-lamination is produced when each ripple climbs onto the stoss slope of the ripple immediately underlying it. Harms *et al.* (1975) noted that aggradational ripples form in settings with abundant, temporarily suspended sand supplies, consistent with deposition in a deltaic setting proximal to a fluvial point source. Coleman and Gagliano (1965) recognized climbing ripples were relatively abundant in delta front deposits from the Mississippi Delta. The large-scale aggrading/climbing symmetrical and asymmetrical dune structures characterizing the lower and middle portion of this facies are interpreted as large-scale wave and combined-flow

Figure 2.32

Facies N - Interbedded, low angle, parallel-laminated, soft-sediment deformed, fine-grained sandstone, siltstone and silty organic mudstone

A. Soft-sediment deformed (overturned), low angle, parallel-laminated, fine-grained sandstone with fractures (Frac) cross-cutting laminae. Note the finely comminuted organic detritus marking some of the internal laminae (Siku A-12, Core #1, depth ~2731.01 m).

B. Soft-sediment deformed, fine-grained sandstone with an angular sandstone rip-up clast. Note the disseminated, organic detritus (Siku A-12, Core #1, depth ~2731.24 m).

C. Soft-sediment deformed, low angle, parallel-laminated, fine-grained sandstone with convolute bedding and minor siderite cement in the lower portion (Siku A-12, Core #1, depth ~2726.84 m).

D. Fine-grained, silty sandstone with wave ripple lamination, convolute bedding, and large-scale, climbing megaripples (W.r.) in the upper portion of this core slab. Note the finely comminuted organic detritus and phytodetrital material marking the laminae (Siku A-12, Core #1, depth ~2727.71 m).

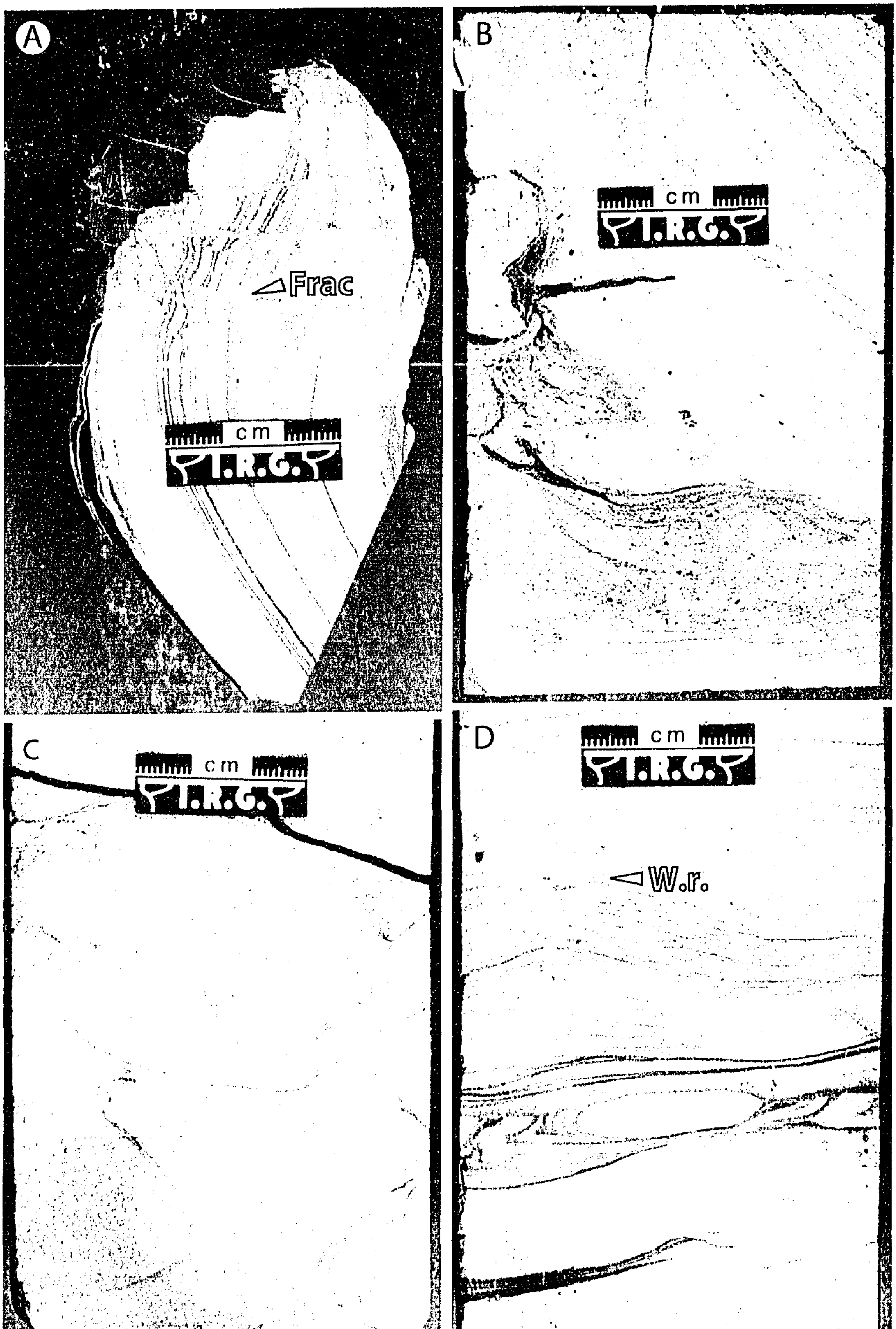
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Figure 2.33

Facies N - Interbedded, low angle, parallel-laminated, soft-sediment deformed, fine-grained sandstone, siltstone and silty organic mudstone

A. Fine-grained sandstone with climbing wave ripples (W.r.) displaying high lateral and vertical migration of the ripples. These ripples are interpreted to represent wave or combined-flow ripples. Note the enrichment of organic material (Mc) in the troughs of the ripples and the minor pyrite nodules (Py) (Siku A-12, Core #1, depth ~2730.36 m).

B. Low angle, parallel-laminated, fine-grained sandstone with wave ripples (W.r.). Note the upward transition from low angle parallel lamination into overlying wave ripples (W.r.). Finely comminuted organic detritus and phytodetrital material mark the internal laminae (Siku A-12, Core #1, depth ~2727.91 m).

C. Fine-grained sandstone with large-scale, climbing wave ripples (W.r.). Note the disseminated, organic detritus (Mc) at the crests of the bedforms (Siku A-12, Core #1, depth ~2729.70 m).

D. Fine-grained, low angle (<5° to 10°), parallel-laminated sandstone with laminae marked by finely comminuted organic detritus and phytodetrital material (Siku A-12, Core #1, depth ~2729.18 m).

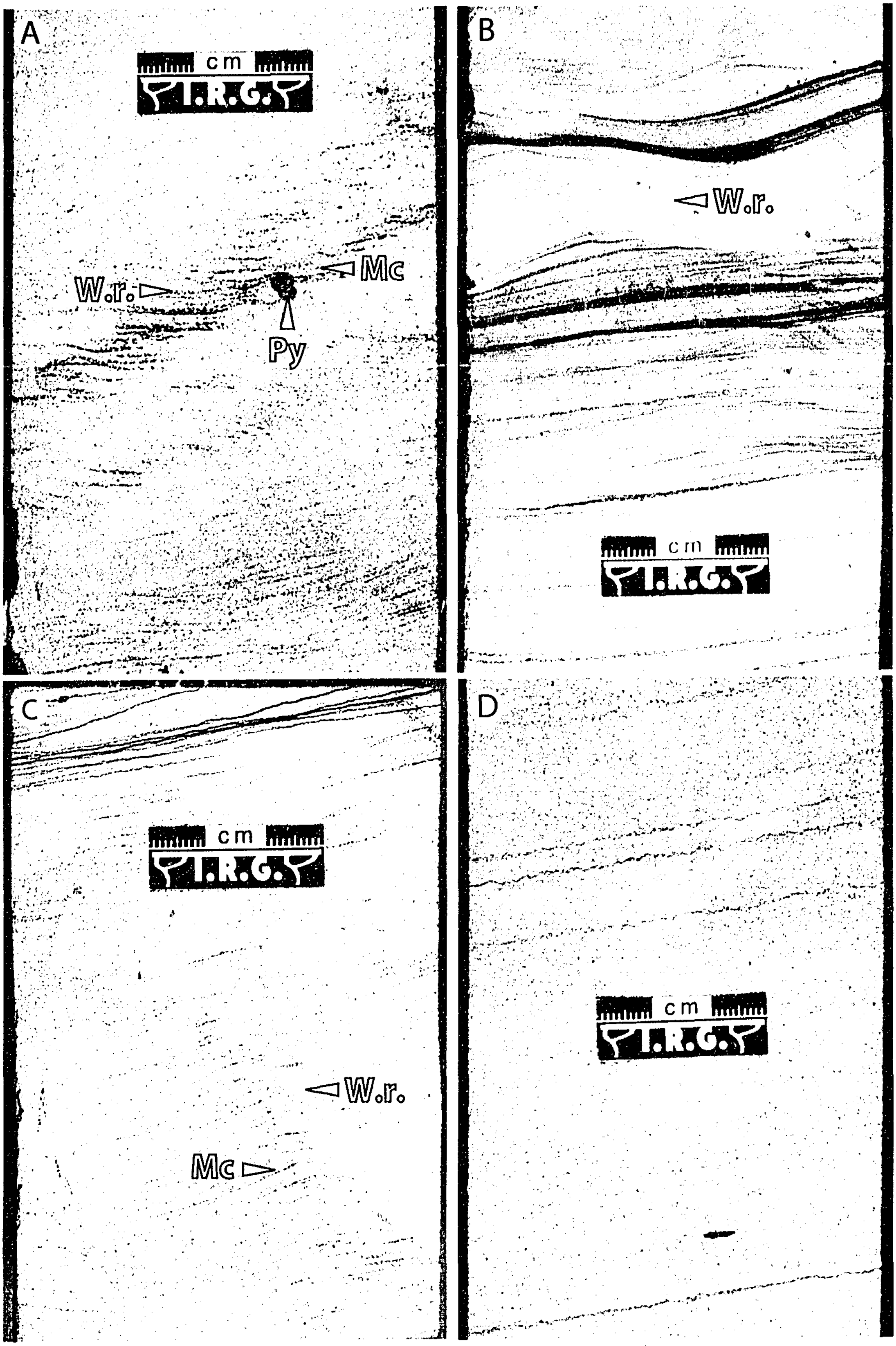
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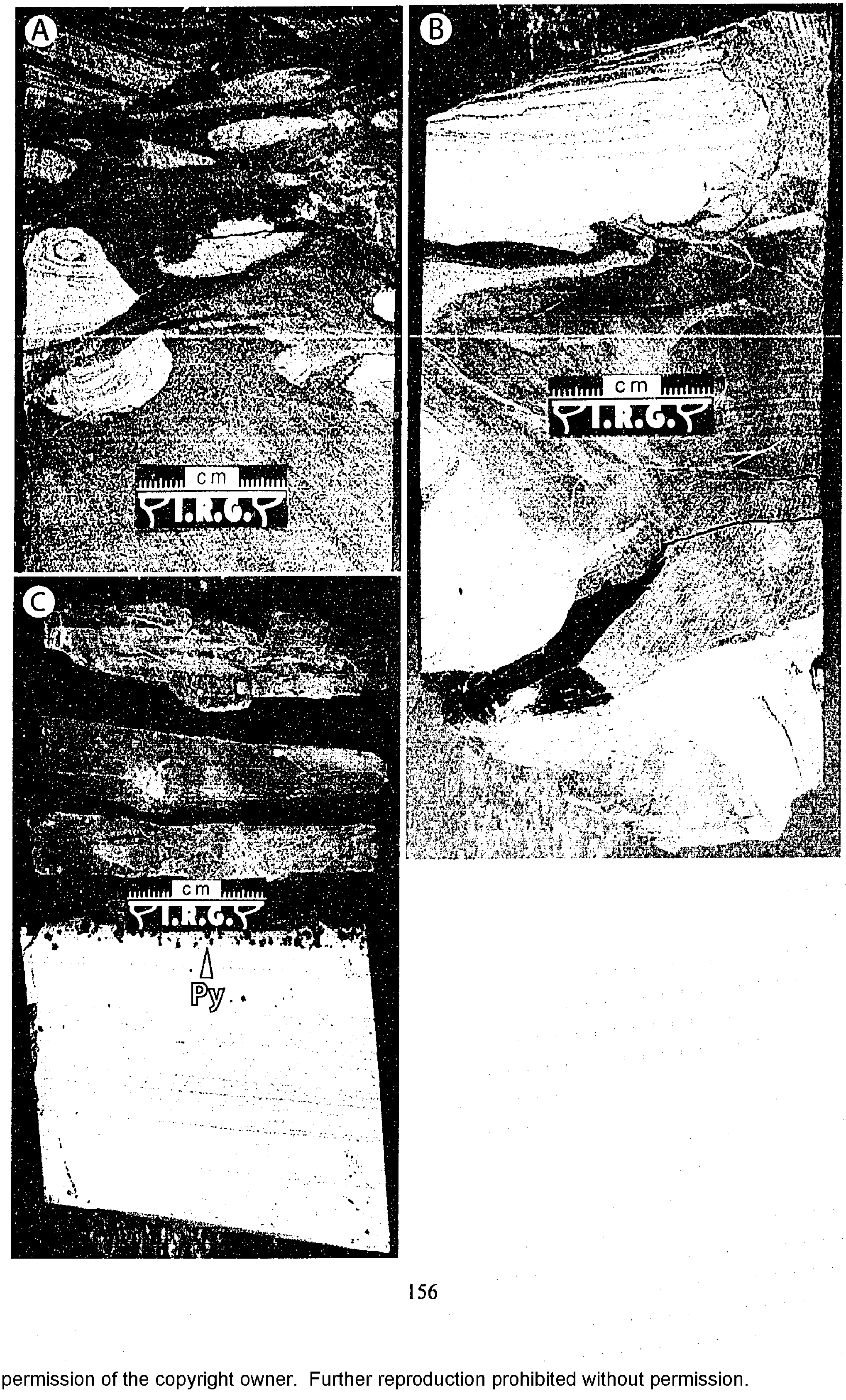
Figure 2.34

Facies N - Interbedded, low angle, parallel-laminated, soft-sediment deformed, fine-grained sandstone, siltstone and silty organic mudstone

A. Highly organic mudstone with ball and pillow structures. The ball and pillow structures likely represent ripples which have sunken into the underlying soupy muds penecontemporaneously with deposition (Siku A-12, Core #1, depth ~2728.48 m).

B. Soft-sediment deformed, very fine to fine-grained, silty sandstone interbedded with mudstone and siltstone. Note the convolute, oversteepened and curvilinear bedding with mud injected into siltstones (Siku A-12, Core #1, depth ~2731.01 m).

C. Massively bedded, sharp based, highly organic mudstone, interpreted to reflect a hyperpycnal mud, overlying a low angle (<5° to 10°), parallel-laminated, fine-grained, well-sorted sandstone. Note the numerous disseminated pyrite (Py) nodules at the contact between the sandstone and mudstone (Siku A-12, Core #1, depth ~2728.66 m).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Py

ripples. These structures reflect high sedimentation rates and large amounts of sediment carried in suspension (Harms *et al.,* 1975). The minimal migration of the large-scale ripple crests suggests that current drift was negligible, and aggradation and reworking by waves was the dominant depositional process. Wright (1977) noted that in the delta front, the breaking of waves enhances mixing and momentum exchange between the effluent and ambient waters, causing very rapid deceleration and loss of sediment transportability within short distances from the distributary outlet. This process would cause the rapid deposition of large of amounts of suspended sediment, which if subjected to minimal wave reworking, would form conditions suitable to form large-scale dune bedforms.

The abundance of interstratified soft-sediment deformation structures including slumped, oversteepened, and curvilinear bedding, in addition to ball and pillow structures and load casts, are indicative of high sedimentation rates and rapid sediment deposition onto unstable, soupy substrates. Convolute bedding develops due to sediment loading increasing pore pressures and subsequent dewatering after burial. Ball and pillow structures form when rapidly deposited sand sinks into underlying soupy muds (Reineck and Singh, 1980). The formation of soft-sediment deformation structures suggests rapid deposition where the substrate becomes soft and unstable, as it cannot disperse fluids quickly enough, allows deformation during early compaction (Coleman and Prior 1982; Bhattacharya and Walker 1991; Reading 1996). The rare fractures within the convoluted intervals were likely generated by loading, increased pore pressures, and subsequent dewatering after burial. High sedimentation rates and rapid deposition of sediments onto unstable, soupy substrates are characteristic features of river-dominated deltas, where active deposition takes place in lobes proximal to the fluvial point source (Coleman and Prior, 1982; Wright, 1985). Coleman and Gagliano (1965) noted that deformation structures were very common in delta front deposits from the Mississippi Delta. They attributed to these structures to gravity-induced slumps, differential overloading and slow mass movement of saturated sediments on an inclined slope. These structures are more common in river-dominated settings as opposed to wave-dominated settings as the majority of sediments in wave-dominated settings are subsequently reworked by wave action following deposition.

The abundant finely comminuted organic and micaceous material marking laminae and bedding is interpreted as "phytodetrital pulses" (Rice *et al.,* 1986). The origin, nature and environmental implications of this phytodetrital material is discussed further in Facies O where it occurs in greater abundance. The pyrite nodules commonly found at mudstone contacts reflect redox reactions between reduced, organic rich mudstone and oxygenated sandstone. Some interstratified, centimeter to decimeter thick, unburrowed, massive mudstone beds displaying sharp tops and sharp bases reflect hyperpycnal or fluid muds similar to those described in Facies D. Most of these unburrowed mudstone beds however, display faint internal laminae. These types of flows are generated at river mouths in marine basins during times of high fluvial discharge, typically during or closely following storm events, where the dense, sediment-laden river outflow enters the delta front and flows down along the bottom of the delta front slope as density underflows (Wright *et al.*, 1988; Mulder and Syvitsky, 1995). The presence of internal graded bedding indicates deposition from suspension fallout, thus differentiating them from the aforementioned hyperpycnal or fluid mud deposits. These mudstone beds are interpreted to represent hypopycnal muds that form due to the raining out of suspended sediments from the delta front. Hypopycnal flow forms where the river outflow is less dense than the receiving basin waters and the river waters flow outward on top of the basin water as a horizontally oriented plane jet (Bates, 1953; Bhattacharya and Walker, 1992). Fine sediment is carried in suspension some distance outward from the river mouth where subsequent mixing of the fresh and marine waters causes suspended clays to flocculate and be deposited rapidly from suspension. Within this facies, it is likely that both hyperpycnal and hypopycnal conditions were prevalent; the type of deposition was dependent on the characteristics of the sediment influx into the delta front from the point sources.

The absence of trace fossils suggests highly stressed, harsh environmental conditions, which precluded the colonization of the sediments by infaunal organisms. Ecological stresses in the distal delta front of a river-dominated delta include high sedimentation rates, high water turbidity, unstable, soupy substrates, fluctuating salinities, as well as oxygen and temperature variations. This facies was likely deposited near a distributary outlet thereby receiving large amounts of rapidly deposited sediments, abundant carbonaceous detritus, in addition to mud and silt from suspended sediment plumes. High sedimentation rates and increased water turbidity relating river discharge would make the substrate unstable and difficult to colonize by tracemakers that produce permanent dwelling structures (Gingras *et al.*, 1998; Coates 2001; MacEachern *et al.*, 2005). Increased amounts of suspended sediment and water turbidity in the delta front also preclude suspension-feeding organisms, which are susceptible to the clogging of their filter feeding apparatus by fine-grained sediments in the water column (Gingras *et al.*, 1998; Coates, 2001; MacEachern *et al.*, 2005). The abundance of suspended sediment also decreases the percentage of food per unit volume of material ingested by deposit feeders (Moslow and Pemberton 1988; Gingras *et al.*, 1998; Coates and MacEachern, 1999; MacEachern *et al.*, 2005).

Facies O - Medium-grained, unburrowed, low, medium to high angle cross-bedded sandstone

<u>Lithology and Sedimentary Features:</u>

Facies O consists of light grey to buff, medium-grained sandstone interbedded with silty organic mudstone (Figs. 2.35A-D; 2.36A-D). The sands are much coarser than Facies N and are characterized by low angle cross-bedding, ripple cross-lamination, trough and tabular cross-bedding (Figs. 2.35A; 2.36A,D). The sands are dominated by horizontal to low, medium and high angle (<5° to 25°) cross-bedding with some beds appearing massive to structureless. Low angle cross-bedding dominates, with minor flaser bedded intervals. Intervals from upper portion of this facies are characterized by asymmetrical ripple lamination, which displays well-preserved foresets marked by abundant, finely comminuted, organic detritus and phytodetrital material (Figs. 2.35B,D). Some intervals display excellent examples of trough cross-bedding (TCS) where bedding angles progressively increase from <5° up to 28° (Fig. 2.35A).

Sandstone intervals contain a relatively high abundance (15% to 20%) of finely comminuted phytodetrital material and micaceous material, which commonly accentuates the laminae and bedding structures (Figs. 2.35A,B,D; 2.36A,C,D). It commonly concentrates in the troughs of bedding structures (i.e. asymmetrical ripples and cross-bedding). Some laminated beds (up to 10 cm thick) are composed almost entirely of finely comminuted/macerated organic detritus and micaceous material (Fig. 2.36B). Angular mudstone rip-up clasts 1 to 3 cm in diameter that are commonly pyritized are found throughout Facies O (Fig. 2.35C). Grain size and phytodetrital material content generally increases toward the upper portion of the facies, displaying an overall coarsening-upward sequence.

Facies O is only found in core #1 from the A-12 well. It directly overlies Facies N via a scoured contact marked by an increase in sandstone grain size and a decrease in the proportion of the muddy sediments and soft-sediment deformation structures (Fig. 2.34A). The upper contact with fine-grained sediments of Facies J is undulatory and marked by an overall decrease in grain size, an increase in mudstone and siltstone content and the presence of rootlet traces. The single occurrence of Facies O is approximately 4.23 m thick.

Ichnological Characteristics:

No trace fossils were identified within this facies.

Interpretation and Discussion: Proximal Delta Front (River-Dominated Delta Front)

Facies O is interpreted as weakly storm-influenced, proximal delta front deposits of a river-dominated delta. It occurs equivalent position to the upper shoreface and foreshore of non-deltaic shoreface environments. The proximal delta front zone (upper shoreface) is situated well above fairweather wave base in the high-energy build-up and surf zones landward of the breaker zones (Clifton *et al.*, 1971; Davidson-Arnott and Greenwood, 1976; Hunter *et al.*, 1979; Greenwood and Mittler, 1985). This zone is characterized by the interaction of both marine and fluvial processes.

The low angle cross-bedding, ripple cross-lamination, trough and tabular cross-bedding structures are indicative of an environment characterized by moderate to high-energy conditions induced by wave and storm reworking. The symmetrical ripple structures are interpreted as wave or combined-

Figure 2.35

Facies O - Medium-grained, unburrowed, low, medium to high angle cross-bedded sandstone

A. Angular contact marked by a scour and fill structure (Scour). The surface is marked by a grain size change from upper fine to a lower medium-grained sandstone displaying low angle (~5°) cross-bedding. Note the finely comminuted organic detritus and phytodetrital material marking the bedding. Some of these organic laminae are stylolitized (Siku A-12, Core #1, depth ~2721.45 m).

B. Fine-grained, low angle (~5°) parallel laminated sandstone with current ripple (C.r.) cross-lamination composed almost entirely of finely comminuted organic detritus (Mc) and micaceous material interpreted to reflect phytodetrital material (Siku A-12, Core #1, depth ~2721.45 m).

C. Medium-grained sandstone with angular mudstone rip-up clasts and abundant finely comminuted organic detritus (Siku A-12, Core #1, depth ~2725.48 m).

D. Fine-grained sandstone with current ripple (C.r.) cross-lamination and abundant finely comminuted organic detritus (Mc), some of which are stylolitized (Siku A-12, Core #1, depth ~2721.45 m).

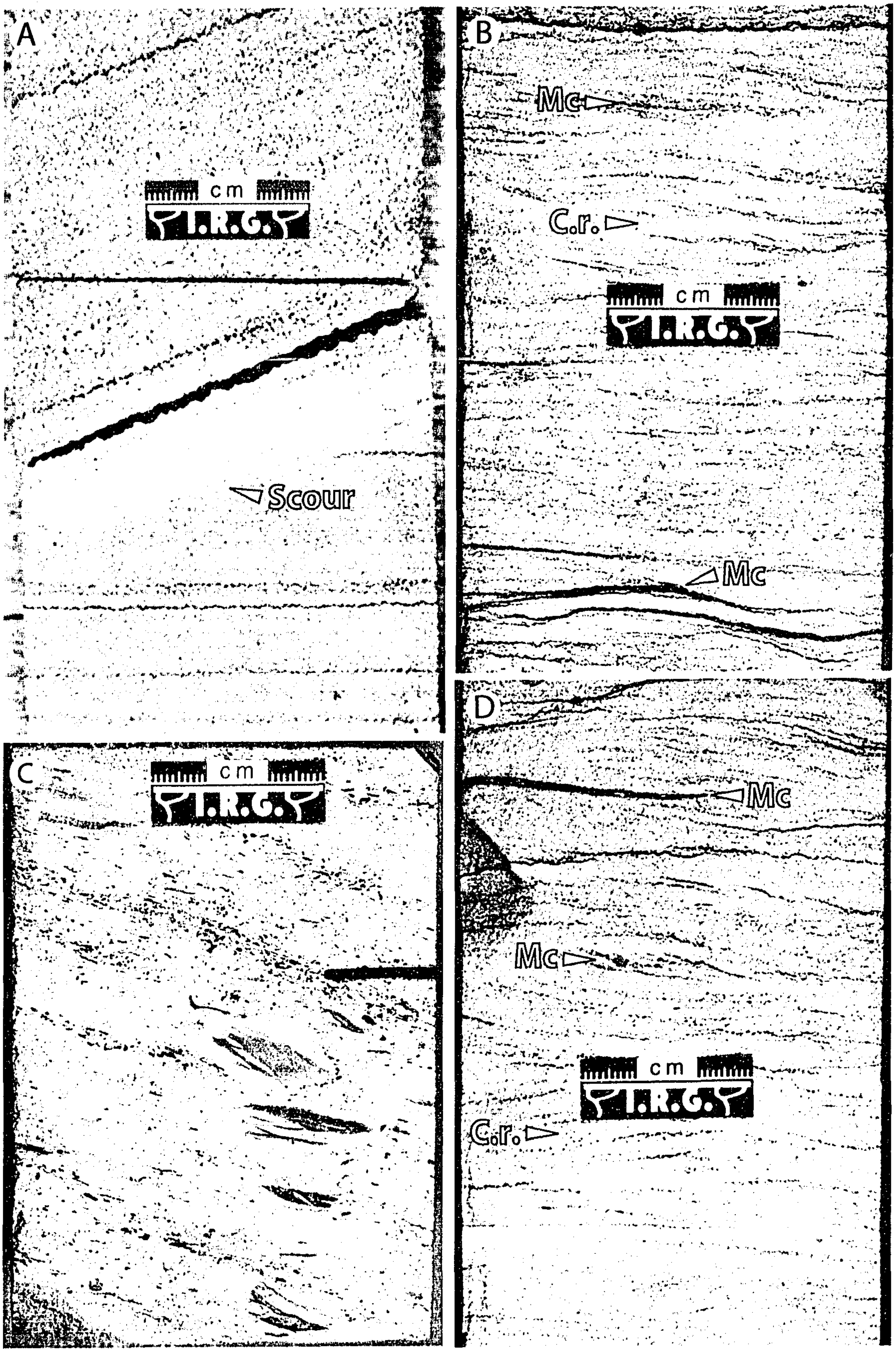

A
cm
T.R.G.
Scour
B
Mc
C.r.
cm
T.R.G.
Mc
C
cm
T.R.G.
D
Mc
Mc
cm
T.R.G.
C.r.

Figure 2.36

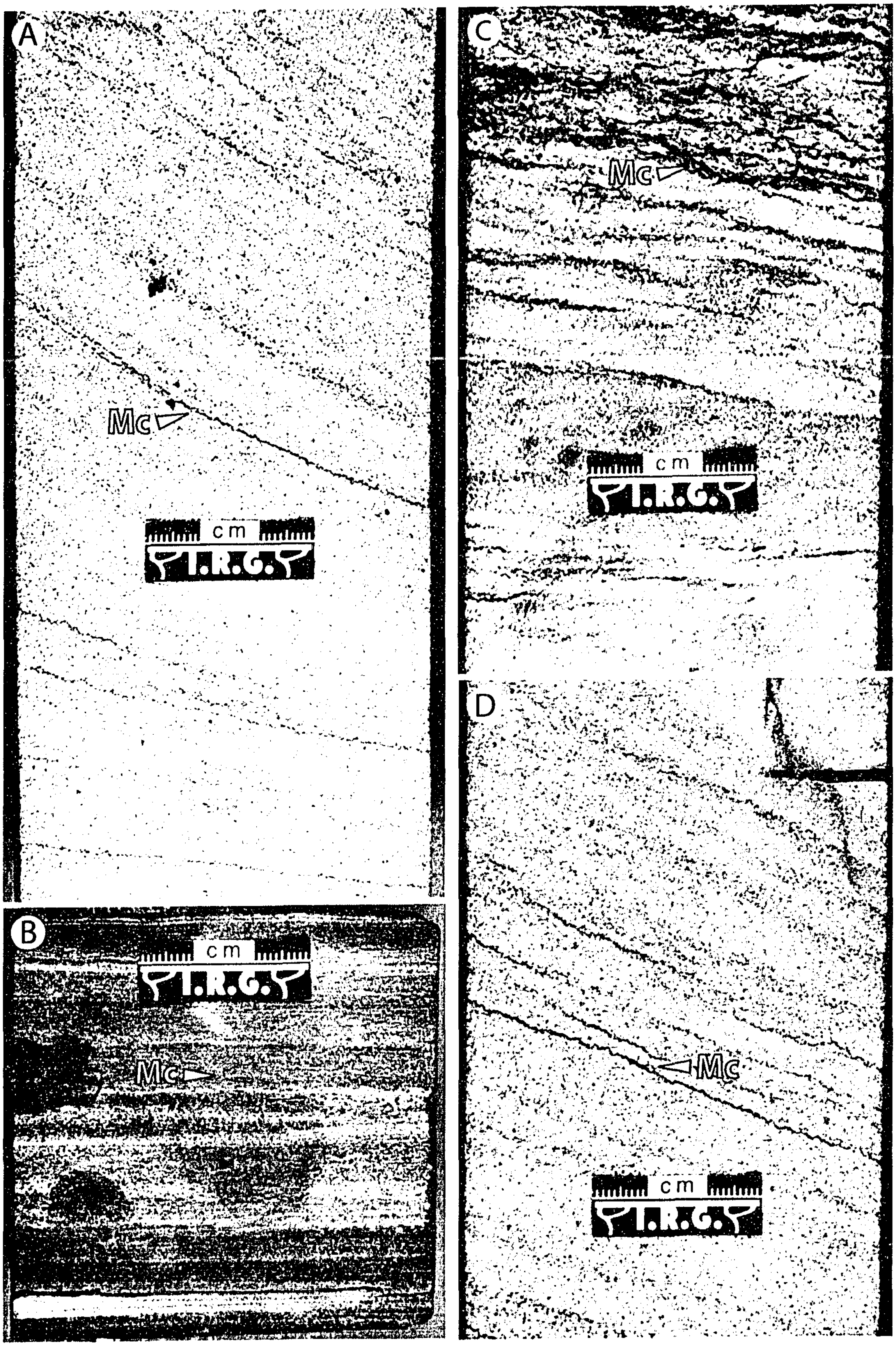
Facies O - Medium-grained, unburrowed, low, medium to high angle cross-bedded sandstone

A. Trough cross-bedded, medium-grained sandstone with progressively increasing bedding angles (~5° up to 32°). The distinguishing feature of trough cross-bedding is the increasing bedding angles towards the top part of the bed. The upper portion of this piece of core displays oversteepened bedding above the angle of repose for sand (~32°). Note the finely comminuted organic detritus (Mc) and phytodetrital material marking the laminae. Some of these organic laminae are stylolitized (Siku A-12, Core #1, depth ~2721.45 m).

B. Fine-grained, parallel-laminated sandstone composed almost entirely of finely comminuted organic detritus (Mc) and micaceous material interpreted to reflect phytodetrital material (Siku A-12, Core #1, depth ~2721.56 m).

C. Medium-grained sandstone with current ripple lamination in which the foresets are marked by abundant finely comminuted organic detritus (Mc) and phytodetrital material. Some of the laminae comprised of organic detritus are 1 to 2 mm thick (Siku A-12, Core #1, depth ~2725.48 m).

D. Planar tabular cross-bedded, medium-grained sandstone with abundant finely comminuted organic and micaceous detritus (Mc). Some of the organic laminae are stylolitized (Siku A-12, Core #1, depth ~2721.45 m).


A
Mc
cm
T.R.G.
B
cm
T.R.G.
Mc
C
Mc
cm
T.R.G.
D
Mc
cm
T.R.G.

flow ripples formed in a marine environment characterized by the interaction of oscillatory and current regimes (Harms *et al.,* 1975). The asymmetrical ripples with the well-preserved foresets are interpreted as current ripples indicating unidirectional flow (Harms *et al.,* 1975, 1982) and may reflect flooding discharge from closely positioned distributary channels. These flooding events are likely related to periods of high precipitation generally associated with storm events and the flooding of distributary channels. River-dominated delta fronts are characterized by a confluence of oscillatory currents from wave approach and unidirectional currents from the distributary channels entering the delta front. Trough cross-stratification has been recognized from other upper shoreface deposits by Davidson-Arnott and Greenwood (1976), Roy *et al.* (1980), Reinson (1984), Thom *et al.* (1986). TCS becomes more dominant in the upper portion of this facies, reflecting higher energy and shallowing-upward conditions associated with the progradation of the delta front.

The sediments comprising the proximal delta front or upper shoreface equivalent are continuously reworked by wave processes, which results in re-ordering, improved sorting, and concomitant tighter packing of the sediment grains (Coleman and Prior 1982). Sedimentary structures in the proximal delta front facies suggest deposition characterized by scour, differential loading of sediment, wave and current reworking, and erosion. Wave reworking reduces variability in substrate consistency and limits the development of deformation structures (Coates, 2001). The multiple scour surfaces form during storm activity and are consistent with deposition in a relatively high-energy proximal delta front environment. The coarse-grained nature of sediments comprising this facies represents deposition as bedload from distributary channels at the drop point in the delta front. When riverine distributary channel waters enter the delta front, current velocity and competency decreases and bedload sediments are deposited (Coleman and Prior, 1980). The mudstone rip-up clasts either represent erosional remnants from underlying hyperpycnal and hypopycnal mudstone beds or reworking of diapiric mud lumps in the delta front.

This facies is similar to Facies N in that it contains the abundant finely comminuted organic detritus or "coffee ground" material interpreted to reflect "phytodetrital pulses" (Rice *et al.,* 1986). This phytodetrital material marks the laminae and bedding within the sandstones of Facies O and is composed of ground up terrestrial plant, woody and micaceous material. Reading and Collinson (1996) noted that large amounts of carbonaceous detritus reflect an abundant influx of terrestrial plant material into the water column, which is commonly associated with discharge from a proximal fluvial point source such as a distributary channel. This phytodetrital material is transported to the delta front by distributary channels during periods of high precipitation that typically accompany storm events (Leithold, 1989; Raychaudhuri and Pemberton, 1992; Coates and MacEachern, 1999; MacEachern *et al.*, 2005). During storms, short-lived increases in river discharge related to storm precipitation carry the phytodetrital material in buoyant plumes into the delta front as suspended load where it settles out from suspension and is incorporated as organic rich laminae within the delta front sandstones. Deposition of phytodetrital material is associated with fluctuations in river sediment volumes and discharge and is associated with proximity to a fluvial point source with abundant organic matter held in suspension, such as a distributary channel (Reading and Collinson, 1996). Coates (2001) noted that these "pulses" are typically more common in river-dominated deltaic settings, compared to wave-dominated settings. This is attributed to the suppression of the buoyant plumes at the river mouth by wave action and the subsequent reworking the sediments prior to deposition. Wright (1977) recognized that the "breaking of the wave" enhances the mixing and momentum exchange between the effluent and ambient water, causing very rapid deceleration and loss of sediment transporting ability within short distances from the distributary outlet. Coleman and Prior (1980) also recognized that in modern deltaic environments large accumulations of river-transported organic detritus such as ground up plant material and water-saturated logs were common near the top of distributary mouth bars. These authors noted that the organic material is transported into the delta front due to high fluvial discharge during flooding events when this organic detritus is transported down the rivers and discharged into the nearshore zone. In the high-energy, nearshore zone, wave action grinds the plant fragments and coarser wood particles into large concentrations of ground-up organic detritus referred to as "coffee grounds". In the delta front this phytodetrital material becomes incorporated into the nearshore sediments and accumulates in the lows of large bedforms. In addition, Coleman and Prior (1980) noted that mica content is also generally extremely high in deltaic environments and in many cases mica forms a significant number of individual laminae within delta front sandstones.

The absence of ichnofauna within these sediments is attributable to numerous biological stresses on the ecological community. Stresses include overall high sedimentation rates, rapid sedimentation (slumps), water turbidity (sediment in suspension), salinity fluctuations related to a fluvial influx into marine waters, unstable substrate consistencies, temperature variations and reduced oxygenation (Moslow and Pemberton, 1988; MacEachern and Pemberton, 1992; Saunders *et al.*, 1994; Gingras *et al.*, 1998). Delta front settings are characterized by high water turbidity due to plumes of mud and sediment entering the system from distributary channels (MacEachern *et al.*, 2005). The absence of suspension feeders, including representatives of the *Skolithos* ichnofacies, is attributable to high water turbidity and increased volume of suspended sediment within the water column that interferes with the ability and efficiency of the organisms filter feeding apparatus (Gingras *et al.*, 1998). Many burrowing organisms would be unable to keep pace with the high sedimentation rates associated with delta front environments and would become buried within the sediments. Salinity fluctuations typically results in a general decrease in the number and diversity of ichnogenera compared to shoreface environments. The influence of frequent, high-energy storms on the sediments of the proximal delta front is associated with erosional amalgamation allowing insufficient time for infaunal communities to colonize the substrate. The paucity of bioturbation within these sediments is not attributable to a single factor, but rather a combination of biological stresses working simultaneously to create an inhospitable environment for infaunal organisms.

The overall coarsening-upward sequence of Facies N and O represents the transition from distal to proximal delta front deposits and is consistent with deposition by a prograding river-dominated delta front. In addition, the transition from proximal delta front deposits of Facies O into overlying interdistributary bay fill and crevasse splay deposits of Facies J and K, reflects continued shoaling and progradation into the lower delta plain. Crevasse splay and interdistributary bay fill deposits are more commonly associated with river-dominated or lobate deltas versus shore parallel, wave-dominated delta or strandplain settings (Bhattacharya and Walker, 1991).

2.3 Summary

Within the study area, fifteen facies (Facies A to O) are identified from the Kamik Formation. These facies are grouped into five facies associations representing a two overall coarsening-upward (or shallowing-upward) successions, two erosionally-based fining-upwards successions and one irregular succession comprised of both fining- and coarsening-upward successions. Chapter Three provides further interpretation and discussion of the facies associations and their relationship to the paleodepositional environments represented in the Kamik Formation.

Chapter Three: Facies Associations, Depositional Environments and Modern Analogues

3.1 Introduction

This chapter groups the fifteen facies, previously identified and interpreted from the Kamik Formation in Chapter Two, into five distinct sedimentary packages termed *facies associations*. The sedimentology, ichnology and the associated environmental interpretations of the fifteen facies will be utilized to develop a larger scale depositional model to determine the paleoenvironmental conditions prevalent during deposition of the Kamik Formation. The usage of the term *facies association* in this study follows the definition of Reading (1996, p. 107) as "...groups of facies that occur together and are considered to be genetically or environmentally related". The grouping of these facies into associations allows for the recognition of recurring vertical successions of genetically related depositional environments and the interpretation of all the facies in context of a genetically related stratigraphic succession. Facies associations provide a more intuitive framework for recognizing the relations between facies and the interpretation of paleodepositional environments. Facies associations are useful for characterizing a depositional setting, however, facies successions are susceptible to the perturbations of a facies due to local variability and heterogeneities of a particular depositional environment. The placement of genetically related facies within facies associations allows for the recognition of depositional cyclicity and for Walther's Law to be more reliably applied within a facies succession to assist with the interpretation of a particular depositional setting. The resultant facies association can be compared with established facies models to assist with the interpretation of the depositional environments. The following five facies associations identified in this study were compared with well-documented facies models and modern analogues to aid with the interpretation of the depositional environments, the details of which are discussed below.

3.2 Facies Associations

Five distinct facies associations (FA1 to FA5) were identified in the Kamik Formation (Table 3.1). These facies associations follow three basic stratigraphic patterns: 1) coarsening-upwards (or sandier-upwards) successions, 2) erosionally based fining-upwards successions and 3) an irregular succession composed of both fining- and coarsening-upwards successions. The majority of the facies associations display gradational to sharp contacts between successive facies and are commonly bounded by stratigraphic discontinuities. Facies Association 1 (FA1) is composed of Facies A, B, C, D, E and F and is interpreted to represent a prograding, moderately storm-influenced, wave-dominated delta. Facies Association 2 (FA2) is composed of Facies G and is interpreted to represent a meandering distributary channel system. Facies Association 3 (FA3) is composed of Facies H and is interpreted to represent a braided distributary channel system. Facies Association 4 (FA4) is composed of Facies I, J, K, L and M and is interpreted to represent the interdistributary

Facies Association	Facies Association Descriptions	Depositional Environment
Facies Association One (FA1)	- comprised of Facies A, B, C, D, E and F - overall coarsening-upwards and shallowing-upward successions - identified in core #2 from A-12, core #2 from E-21, core #1 & #2 from L-43, core #1 from P-41, core #1 & #2 from G-04, core #1 from F-09 well	**Moderately Storm-Influenced Wave-Dominated Delta (Upper Kamik Formation)**
Facies Association Two (FA2)	- comprised of Facies G - fining-upward channel fill successions - medium to very coarse-grained sandstone - identifed in core #2 from A-12, core #2 & #3 from E-21, core #2 & #3 from L-43 well	**Meandering Distributary Channel (Upper Kamik Formation)**
Facies Association Three (FA3)	- comprised of Facies H - multiple, stacked, erosionally based, fining-upward channel fill successions - coarse-grained sandstone & conglomerates - identified in core #3 from L-43, core #1 and #2 from N-10 well	**Braided Distributary Channel (Lower Kamik Formation)**
Facies Association Four (FA4)	- comprised of Facies I, J, K, L and M - includes both fining-upwards and coarsening-upwards successions with interstratifed very fine-grained sandstones mudstones & siltstones - identified in core #1 from A-12, core #1 & #3 from E-21, core #1 & #3 from L-43, core #1 from F-09, core #1 and #2 from N-10 well	**Lower Delta Plain (Interdistributary Bay, Brackish Bay, Crevasse Splay, Marsh, Paleosol (Upper & Lower Kamik Formation)**
Facies Association Five (FA5)	- comprised of Facies N and O - overall coarsening-upwards & shallowing-upwards succession - prograding river-dominated deltaic system - identified in core #1 from the A-12 well	**Moderately Storm-Influenced River-Dominated Delta (Upper Kamik Formation)**

Table 3.1 Facies associations identified from the Kamik Formation including descriptions and interpreted depositional environments.

bay, bay fill (crevasse splay), restricted interdistributary bay, marsh and paleosol subenvironments of a lower delta plain. Facies Association 5 (FA5) is composed of Facies N and O and is interpreted to represent a prograding, moderately storm-influenced, river-dominated delta.

3.2.1 Facies Association 1 (FA1) - Moderately Storm-Influenced, Wave-Dominated Delta

Description:

Facies Association 1 (FA1) forms a gradual coarsening and shallowing-upward succession comprised of Facies A, B, C, D, E and F. FA1 typically displays an overall upward increase in grain size, bed thickness, and proportion of sandstone versus mudstone. The deposits of FA1 are the most common facies association identified in this study, with portions of this facies association identified in most of the cores from study area. FA1 has been identified from core #1 from A-12, core #2 from E-21, cores #1 and #2 from L-43, core #1 from P-41, core #1 from F-09 and cores #4 and #5 from G-04 (Figs. 3.3; 3.4; 3.5). However, it should be noted that most cored intervals only contain partial sections of this facies association and a complete succession is not preserved in any one core presented in this study. FA1 is interstratified with FA2 and FA4 and is restricted to the upper member of the Kamik Formation in the study area. The lower bounding surface of FA1 is generally gradational to sharp in nature. The upper bounding surface is commonly gradational to slightly undulatory with FA4 or scoured when directly overlain by FA2. Deposits of FA1 range from 9.13 to 13.2 m in thickness; however, most of the cores only contain partial sequences of the entire facies association and it is likely that thicker, more complete successions exist.

Interpretation:

The strata comprising FA1 are interpreted to represent the progradation of a moderately storm-influenced, wave-dominated deltaic system with attached strandplain shorelines (Fig. 3.6). This succession includes distal prodelta, proximal prodelta, distal delta front and proximal delta front environments. In core #1 from the P-41 well this succession is overlain by a thin, poorly sorted conglomerate of Facies F that is interpreted to represent a transgressive lag deposit (see Chapter 2 for a further discussion). The equivalent shoreface terminology is provided in Figs. 3.1 and 3.2 along with the deltaic terminology due to the similarity in the sedimentological and ichnological features of these facies with those of a non-deltaic shoreface. The basal portion of FA1 is composed of distal and proximal prodelta and distal delta front sediments of Facies A, B and C. Facies A and B are composed of interlaminated to interbedded, thoroughly bioturbated organic-rich mudstones, silty mudstones and very fine-grained silty sandstones interpreted to reflect deposition within a prodelta or equivalent offshore environment. The prodelta can be subdivided into distal and proximal zones; this subdivision is highly arbitrary and is based on sand content, sedimentary structures and ichnological assemblages (Pemberton and MacEachern, 1995; Pemberton *et al.*, 2001). In wave-dominated deltaic settings, such as the one interpreted for FA1, the prodelta deposits closely resemble the offshore to lower shoreface deposits of wave-dominated shoreface deposits. Distinguishing between these two environments requires recognition of subtle ichnological and sedimentological characteristics.

The predominance of clays, fine-grained mudstones and silty mudstones comprising the basal portion of FA1 (Facies A and B), indicates that these sediments were deposited in a low energy, prodelta environment dominated by fairweather suspension sedimentation. The occurrence of interstratified hummocky cross-stratified (HCS) distal tempestite deposits with these fine-grained fairweather offshore deposits indicates that this setting was influenced by episodic, storm events. The rare isolated wave and combined flow ripples found interstratified within the mudstones are interpreted to have been deposited during major storm events when storm wave base interacts with the substrate. The presence of rare current ripple structures is attributed to the influx of

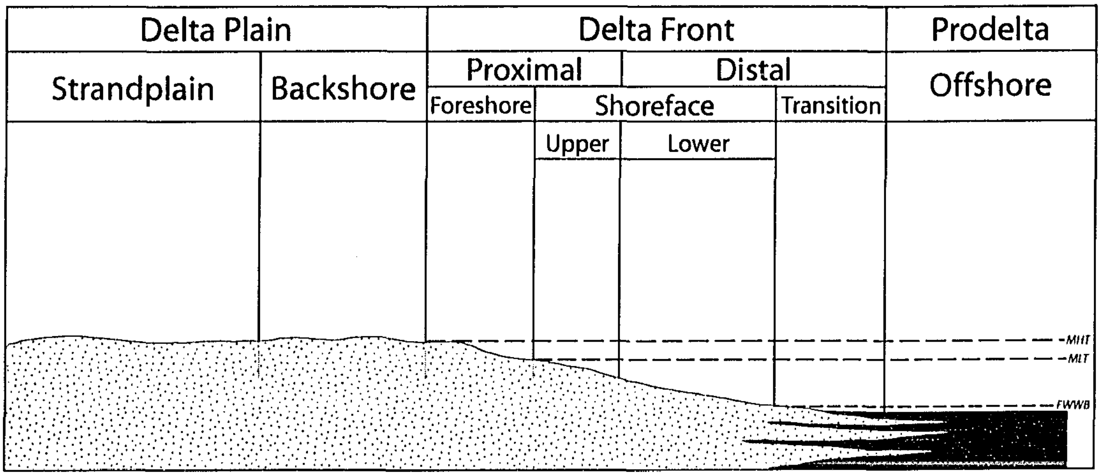


Figure 3.1. Schematic diagram comparing the lateral distribution of depositional environments along the seaward margin of a wave-dominated deltaic depositional system (above) and a strandplain or barrier island depositional system (below). This diagram illustrates the differences in terminology between deltaic and the non-deltaic strandplain and shoreface depositional environments. The sedimentary facies recognized in the Kamik Formation cored sequences are a product of deposition in one or more of these environments (modified after Moslow and Pemberton, 1988).

sediment from distributary channels during increased fluvial discharge that typically accompanies storms. Synaeresis cracks present within this facies are interpreted to reflect salinity variations (Burst, 1965; Plummer and Gostin, 1981) attributed to the influx of freshwater laden sediments from distributary channels during storm events (MacEachern *et al.*, 2005). Overall these deposits are characterized by a low to moderate abundance of storm-generated deposits suggesting that the prodelta and delta front was under moderate storm influence (MacEachern and Pemberton, 1992). The thoroughly to well-burrowed nature of the sediments of facies A and B by elements of the *Cruziana* ichnofacies is consistent with deposition within the offshore complex. The distal and proximal prodelta (lower offshore and upper offshore) typically contains a moderate to high diversity ichnological suite corresponding to the distal and archetypal *Cruziana* ichnofacies (Seilacher, 1967; Frey *et al.,* 1990). The ichnological suite identified in Facies A and B is best described as a slightly 'stressed' mixed *Cruziana* and *Skolithos* suite composed of deposit-feeding, dwelling and grazing structures including *Helminthopsis*, *Phycosiphon*, *Teichichnus*, *Asterosoma*, *Chondrites*, *Rosselia*, *Thalassinoides*, *Zoophycos*, *Scolicia*, *Rhizocorallium*, *Palaeophycus*, *Planolites*, *Zoophycos* and suspension-feeding traces including *Skolithos*, *Conichnus*, *Arenicolites*, *Diplocraterion* and *Schaubcylindrichnus*.

The interbedded, well-burrowed, very fine to fine-grained hummocky (HCS) and swaley (SCS) cross-stratified sandstones and organic mudstones of Facies C are interpreted as storm-generated tempestites and interbedded fairweather mudstones deposited within the distal delta front or equivalent lower shoreface. The distal delta front is characterized by an abundance of episodic storm-generated tempestite deposits including HCS and SCS sandstones. HCS is generally

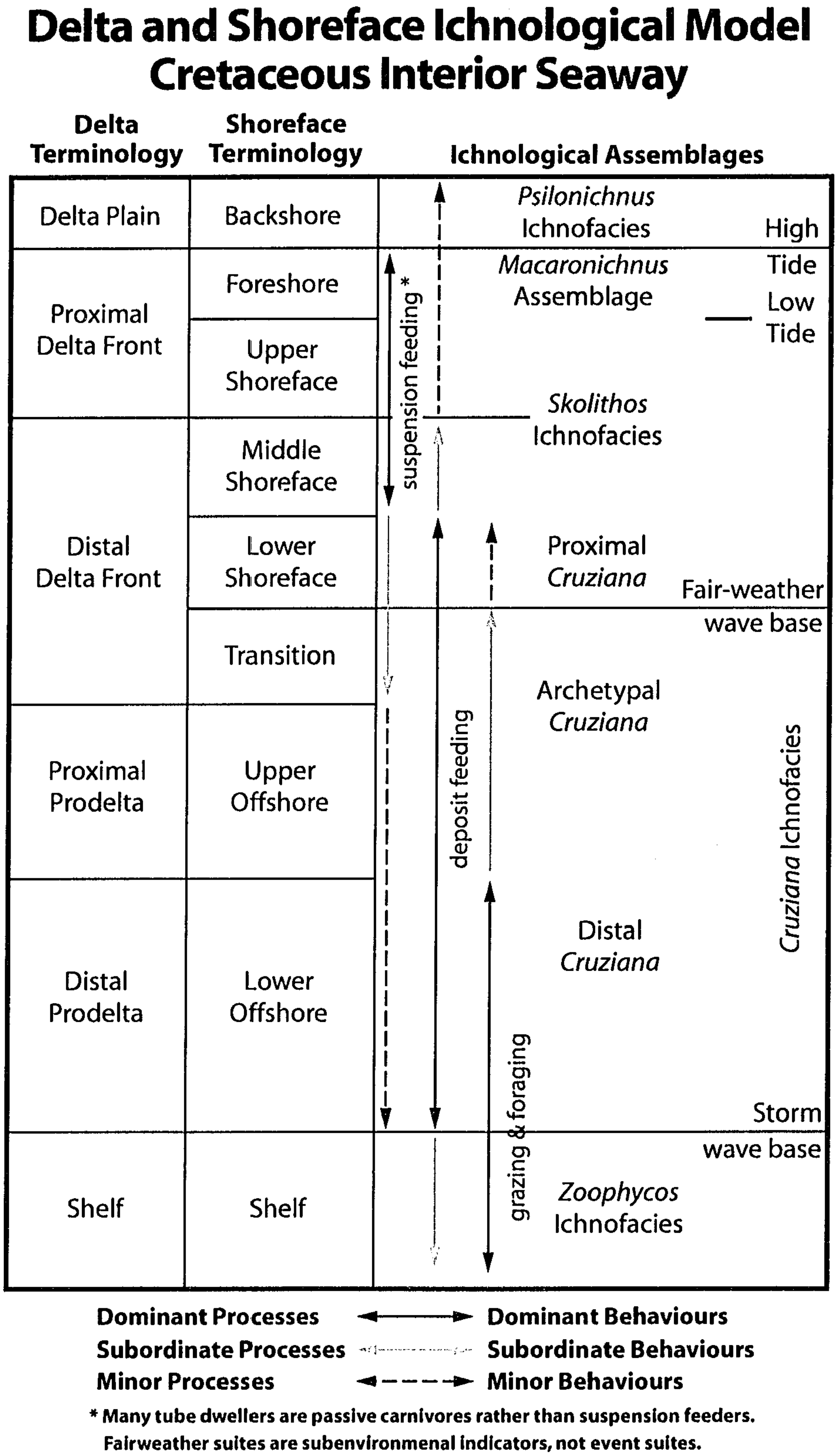


Figure 3.2 Idealized ichnological model based on Cretaceous strata from the Western Interior Seaway of North America. The subdivisions of the delta and shoreface are identified, along with the corresponding typical ichnofacies (modified from Pemberton and MacEachern, 1995).

interpreted to form below fairweather wave base during intense storms (Harms *et al.,* 1975; Duke *et al.,* 1991) and has been identified in deltaic deposits (Dott and Bourgeois, 1982). SCS is also interpreted to represent storm deposition and generally forms above fairweather wave base. The distal delta front or equivalent lower shoreface is situated above fairweather wave base (minimum) in a wave-dominated delta where offshore processes continue to operate (Reinson, 1984). Walker and Plint (1992) noted that fairweather wave base generally occurs in water depths ranging from 5 to 15 m, however it has been recognized that under progressive storm-domination by high frequency storms fairweather wave base can be significantly lowered with facies of the lower shoreface deposited below fairweather wave base (MacEachern and Pemberton, 1992).

Compared to Facies A and B, Facies C is characterized by an increase in the proportion of sandstone and increase in the abundance of tempestites which suggests deposition in a slightly shallower and higher energy setting dominated by wave and storm currents. The thickness of the storm beds and degree of erosional amalgamation increases upward within this facies and reflects a gradation from HCS to SCS. This facies also displays a few examples of soft-sediment deformation structures, including decimeter-scale intervals of convolute bedding. These structures are attributed to the rapid deposition of sediments, a feature commonly associated with the deposits of wave-dominated deltas (Coleman and Wright, 1975; Heward, 1981). The strata of Facies C are typical of the subaqueous distal delta front of a wave-dominated delta, but are virtually indistinguishable from the lower and middle shoreface of a storm-dominated shoreface (MacEachern and Pemberton, 1992). The abundance of storm-generated HCS and SCS bedforms, well-sorted sandstones winnowed of fine-grained particles by wave activity, the presence of lam-scram bedding and planar-tabular to low angle cross-bedded sandstone in the coarser grained facies and minor soft-sediment deformation structures are typical features of the deposits of wave-dominated deltas. These features indicate that these wave-dominated deltas were moderately influenced by storm events and that storm energy was a constant influence on sediment deposition (J.R. Suter pers. comm., 2004).

The subtly stressed trace fossil assemblage and the presence of storm-generated sedimentary structures within FA1 are strong indicators of deposition in a wave-dominated deltaic setting. The frequency and thickness of the storm beds increases upward within the distal delta front deposits and reflects a shoaling-upward trend. The distal delta front facies contain a moderate to high diversity *Cruziana* trace fossil assemblage (Seilacher, 1967; Frey *et al.,* 1990). Although the assemblage is highly diverse it is best described as slightly 'stressed' composed of *Helminthopsis*, *Phycosiphon*, *Asterosoma*, *Chondrites*, *Rosselia*, *Thalassinoides*, *Ophiomorpha*, *Arenicolites*, *Zoophycos*, *Schaubcylindrichnus*, *Scolicia*, *Rhizocorallium*, *Diplocraterion*, *Teichichnus*, *Skolithos*, *Palaeophycus*, *Planolites*, *Cylindrichnus*, *Conichnus*, *Macaronichnus*, *Siphonichnus*, *Lockeia*, cryptic bioturbation and fugichnia.

Facies C is generally gradationally overlain by the medium to high angle cross-stratified sandstones

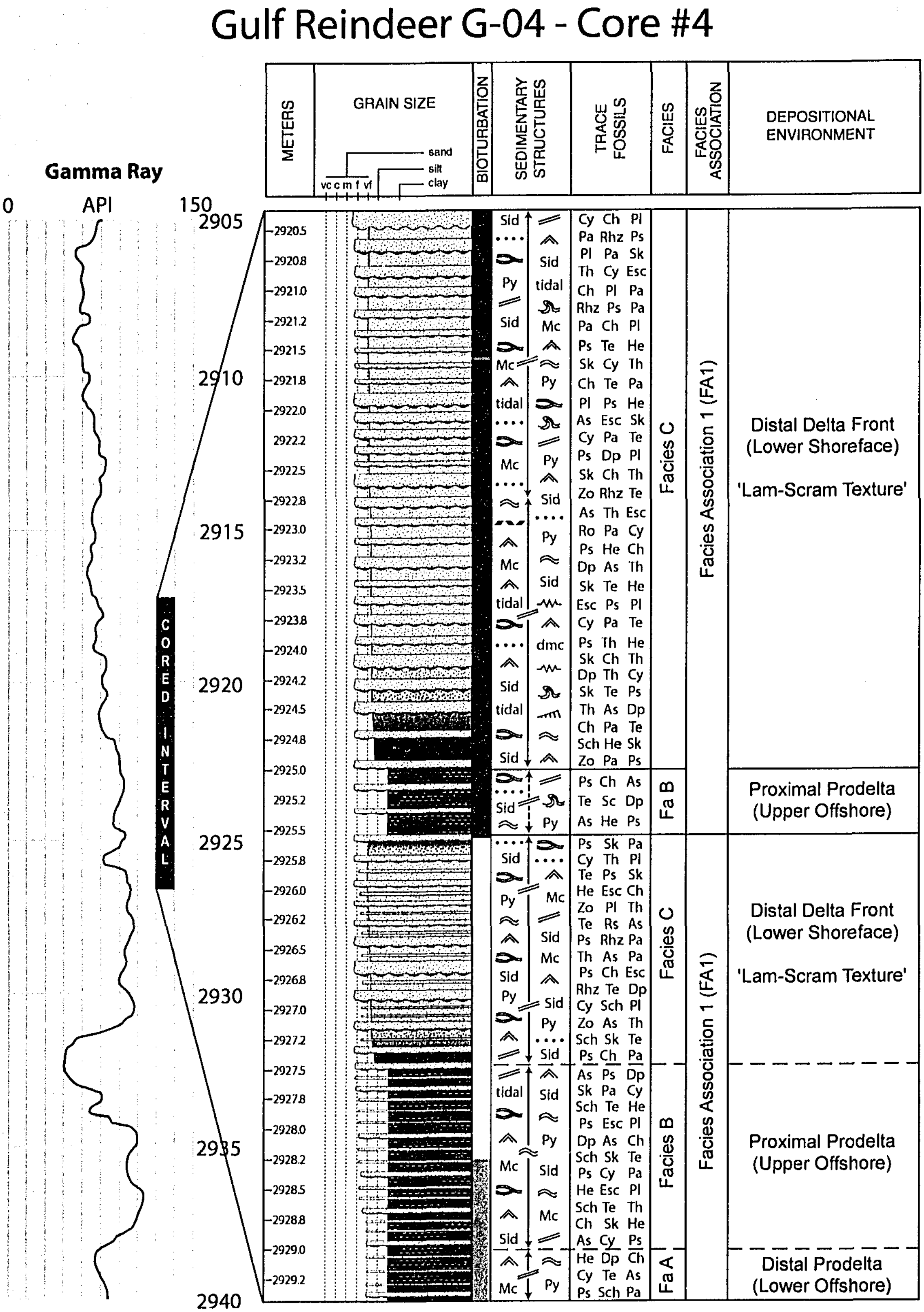


Figure 3.3 Core log of core #4 from the Reindeer G-04 well from the Kamik Formation displaying grain size, lithology, sedimentary structures, ichnological structures, facies, facies associations and interpreted depositional environments.

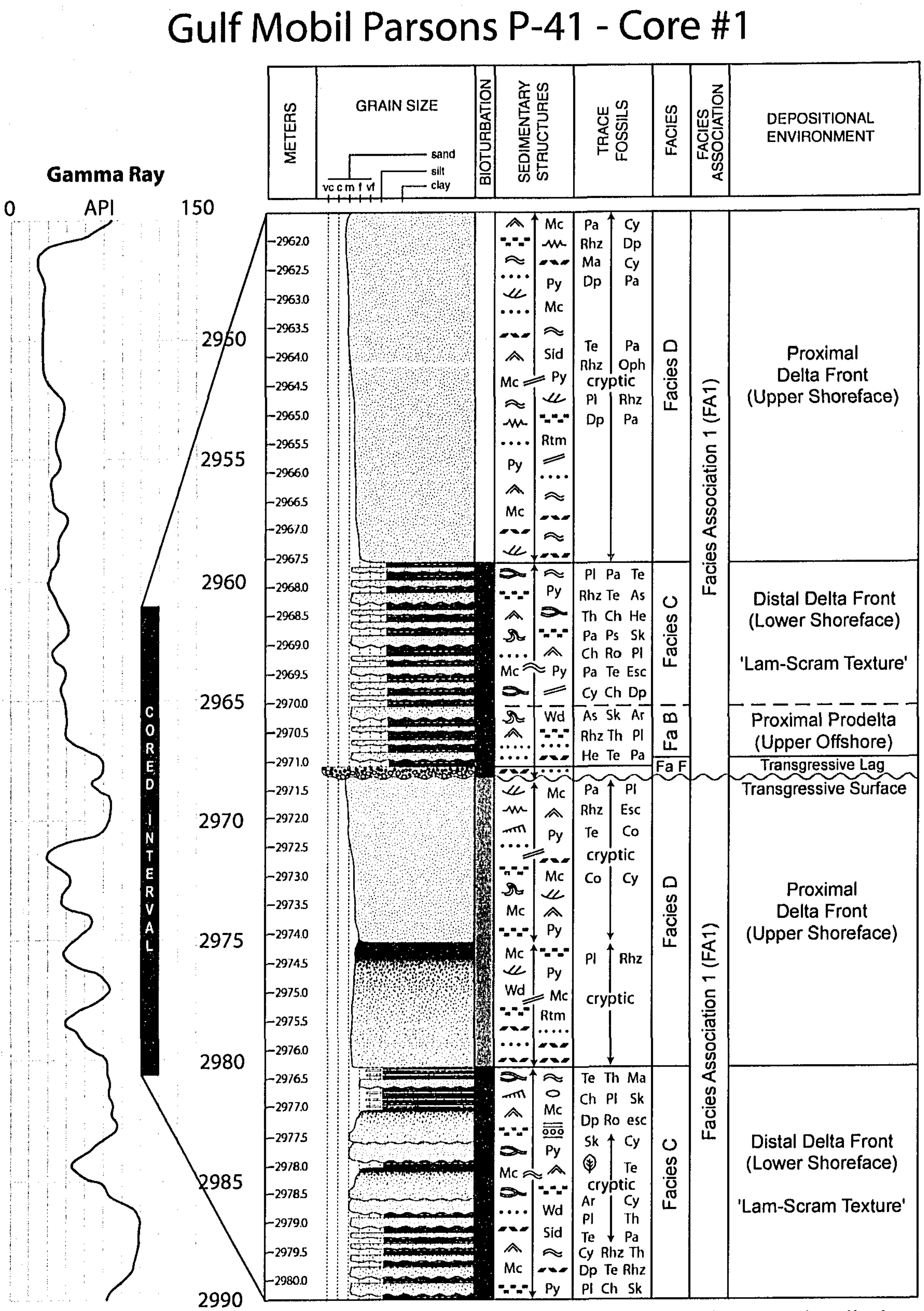


Figure 3.4 Core log of core #1 from the Parsons P-41 well from the Kamik Formation displaying grain size, lithology, sedimentary structures, ichnological structures, facies, facies associations and interpreted depositional environments.

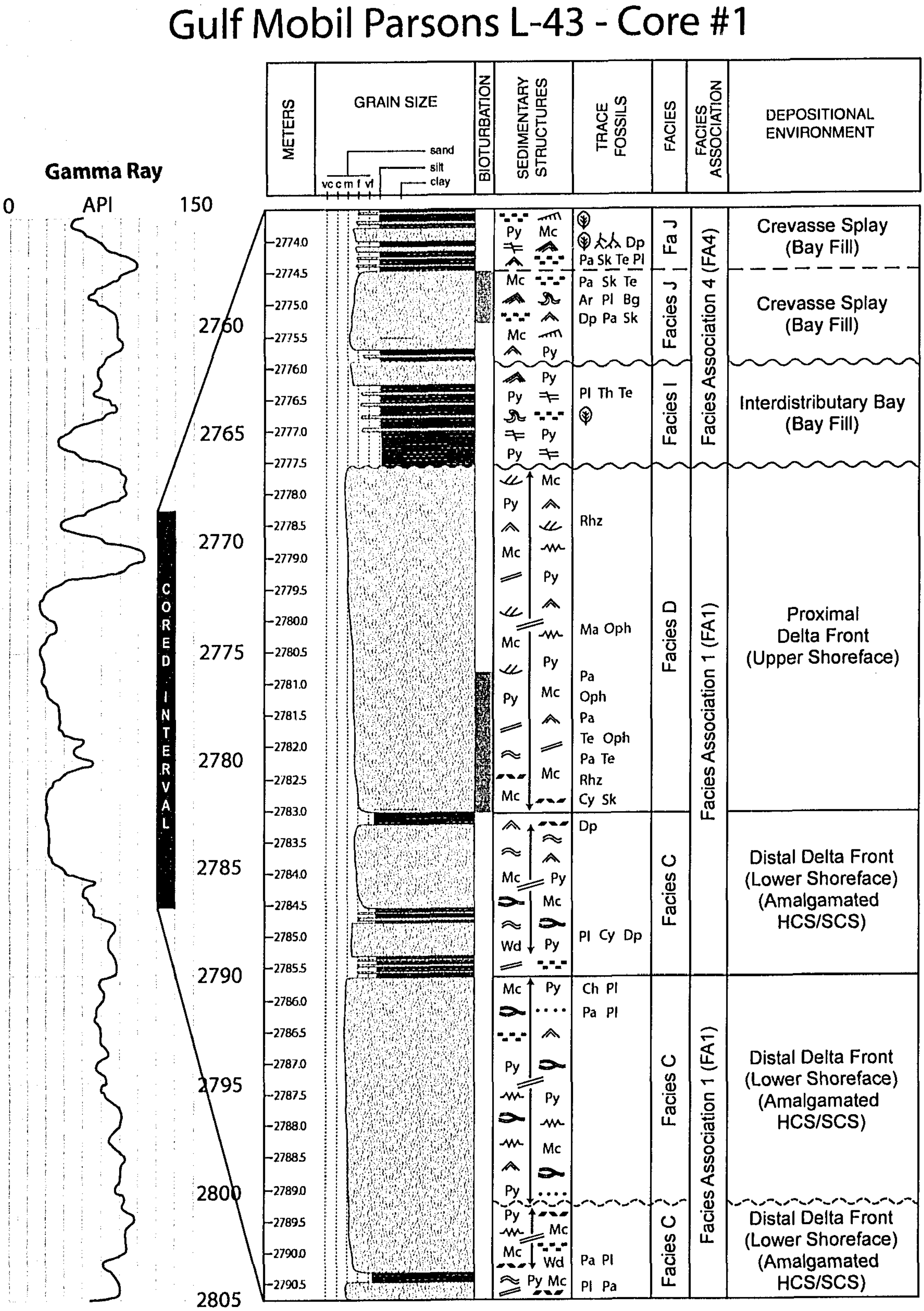


Figure 3.5 Core log of core #1 from the Parsons L-43 well from the Kamik Formation displaying grain size, lithology, sedimentary structures, ichnological structures, facies, facies associations, and interpreted depositional environments.

of Facies D and E, which represent the most proximal portions of FA1. Overall, these facies are part of the coarsening-upwards succession of FA1 displaying a landward progression from the distal delta front to proximal environments of the proximal delta front (upper shoreface and foreshore equivalents) associated with progradation of a wave-dominated delta. The proximal delta front deposits of Facies D and E are comprised of well-sorted, fine to medium-grained sandstones displaying trough and planar cross-stratification, low relief truncations and symmetrical ripple laminae reflecting deposition in a higher-energy, wave-dominated proximal delta front. These sedimentary structures suggests a dominance of suspension, traction and gravity induced processes of deposition (Moslow and Pemberton, 1988). Preserved symmetrical ripples indicate that wave-generated oscillatory flow processes were dominant (Harms *et al.,* 1982). Graded rhythmites and the abundance of phytodetrital laminae or "coffee ground" material correspond to 'pulses' of sedimentation in deltaic environments generated by decaying sediment plumes entering the delta front from distributary channels (Rice *et al.,* 1986). Deposition of these graded rhythmites requires that both traction and suspension processes were operating during deposition, a process common in delta front settings (S.G. Pemberton pers. comm., 2004). Similar structures have been recognized from other ancient deltaic deposits and are attributed to cyclic sedimentation patterns reflecting seasonal fluctuations in sediment supply and fluvial discharge (Moslow and Pemberton, 1988). The sharp-topped and based hyperpycnal mudstone beds are also common feature of delta front settings. The hyperpycnal mud beds identified within this facies are interpreted to have formed due to the introduction of dense, sediment-laden hyperpycnal mud flows into the delta front during flood stages accompanying storm events (Mulder and Syvitsky, 1995; Leithold, 1989). In addition, the occurrence of ball and pillow structures reflects the rapid deposition of sandy sediments onto unstable, soupy substrates, a feature typical of deltaic environments (Coleman and Prior, 1982; Wright, 1985). These deposits are capped by the delta plain sediments mudstones, siltstones, silty sandstones and coals of FA4.

The proximal delta front facies are characterized by a low to moderate diversity *Skolithos* ichnofacies assemblage composed of *Ophiomorpha*, *Rhizocorallium*, *Conichnus*, *Cylindrichnus*, *Macaronichnus*, *Teichichnus*, cryptic bioturbation and rare fugichnia. The presence of mainly suspension-feeding burrows indicates that Facies D and E were deposited in high-energy settings, characterized by well-oxygenated conditions with suitable amounts of suspended food available for suspension-feeding behaviours (MacEachern and Pemberton, 1992). This ichnological assemblage is best described as an impoverished assemblage of suspension-feeding burrows due to the lower diversities and abundances of these organisms compared to those in non-deltaic shoreface environments. The *Macaronichnus* trace fossil is typically found in high-energy shorefaces in the upper shoreface and foreshore zones (Saunders and Pemberton, 1986; MacEachern and Pemberton, 1992) consistent with those that can occur in a wave-dominated delta. The rootlet traces that are found at the top of FA1 reflect a shoaling-upward transition into the lower delta plain and indicate that stable substrates were available for colonization by plants.

The ichnological assemblage of FA1 displays an overall gradation from mainly deposit-feeding and grazing structures of the *Cruziana* ichnofacies in the prodelta and distal delta front facies (Facies A, B and C) to the mainly suspension-feeding structures of the *Skolithos* ichnofacies associated with the higher energy conditions of the proximal delta front facies (Facies D and E). The ichnological assemblage is described as 'stressed' and is characterized by relatively low to moderate diversities and abundance, diminutive morphologies and sporadic distributions of ichnofossils compared to those of non-deltaic shorelines. Deltaic environments represent one of the most biologically stressed marine and marginal marine environments for burrowing organisms. The stressed nature of ichnological assemblages is attributed to numerous biological stresses present in deltaic environments including high and/or fluctuating sedimentation rates, hypopycnal-induced water turbidity, salinity fluctuations, substrate instability, distributary flood discharges with accompanying phytodetrital (comminuted plant debris) pulses, hyperpycnal-induced sediment gravity flows and fluid mud deposition, low oxygen levels, light and temperature variations, all of which contribute to impede infaunal colonization (Moslow and Pemberton, 1988; Gingras *et al.*, 1998; MacEachern *et al.*, 2005). In the wave-dominated deltas responsible for FA1, increased levels of water turbidity related to fluvial discharge from distributary channels periodically impeded filter-feeding behaviour by suspension-feeding organisms. Suspension-feeding organisms require clean water so that clay does not clog their filter feeding apparatus (Gingras *et al.*, 1998). High amounts of suspended sediment within the water column results in a reduced percentage of food per unit volume of material ingested by the organisms (Moslow and Pemberton, 1988; Gingras *et al.*, 1998; Coates and MacEachern, 1999). In wave-dominated deltaic settings, such as those interpreted for FA1, these ecological stresses are not as profound as in river-dominated settings due to being mediated by wave-energy.

Wave-dominated deltaic settings are characterized by reduced sedimentation rates, decreased turbidity and greater sediment stability in comparison to river-dominated deltaic settings. Wave energy generally buffers the effects of fluvial discharge, by dispersing suspended sediment offshore and encouraging the thorough mixing of waters of contrasting salinities and densities (MacEachern *et al.*, 2005). Strong longshore drift can also operate to extend river-derived stresses considerable distances down-drift from the mouths of distributary channels. This also results in the mediation of the river-derived stress and limits their effects on infaunal communities. However, it should be noted that recent work on asymmetric deltas has recognized that markedly different ichnological expressions are expressed on either side of distributary channel mouths. The updrift settings retain classic shoreface assemblages and the down-drift environments commonly acquire markedly stressed suites (MacEachern *et al.*, 2005). The main ecological stresses interpreted to have been present during the deposition of FA1 included fluctuating energy conditions associated with storms activity, high sedimentation rates, substrate instability and hypopycnal-induced water turbidity all of which resulted in limited development of infaunal communities. For a more detailed discussion of the ichnology of deltas the reader is referred to MacEachern *et al.* (2005).

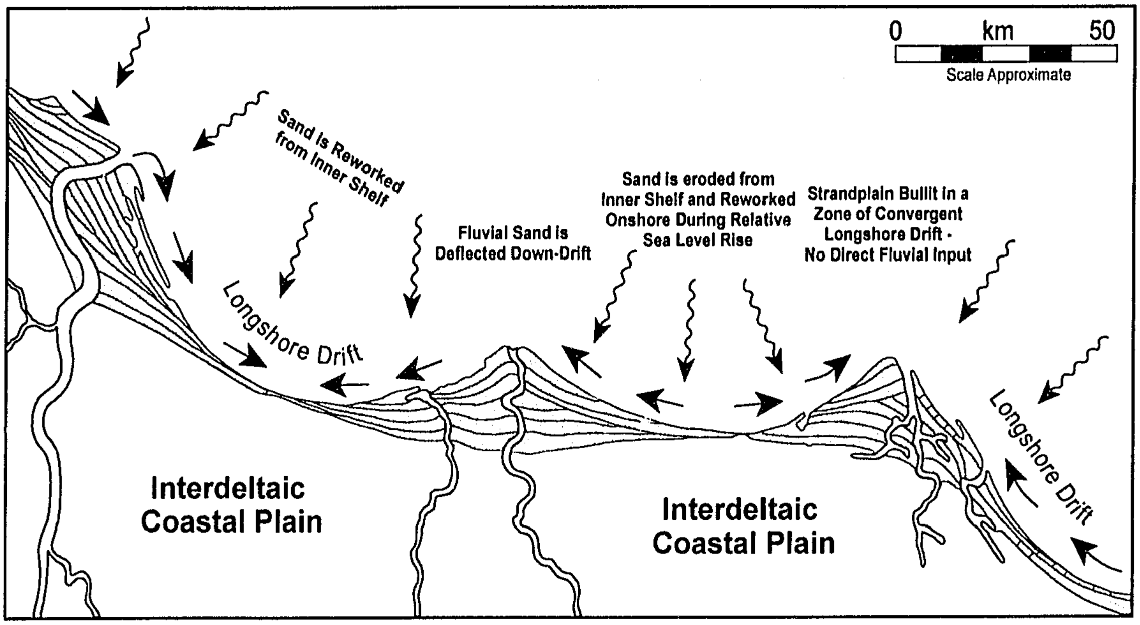


Figure 3.6 Wave-dominated deltas and associated strandplain systems (modified after Walker and Plint, 1992).

Overall, there are many characteristics of FA1 that support its interpretation as representing a moderately storm-influenced, wave-dominated delta (Fig. 3.6). This is due to a number of key characteristics including: 1) the presence of wave-generated structures such wave ripples and trough cross-bedding, minor soft-sediment deformation structures such as ball and pillow structures, convolute bedding, synaeresis cracks, graded rhythmites, hyperpycnal and hypopycnal mudstone beds and abundant phytodetrital laminae are all features common in wave-dominated deltas; 2) abundant storm-generated HCS and SCS bedforms and the well-winnowed, sand-dominated nature of the deposits reflect the dominance of wave and storm processes during the deposition of FA1; 3) the stressed ichnological assemblages attributed to the numerous biological stresses prevalent in deltaic environments. The stratigraphic relationship between the deltaic deposits of FA1, with the distributary channel deposits of FA2, and the lower delta plain deposits of FA4 supports the interpretation that these sediments were deposited as part of a wave-dominated delta system.

The strata of FA1 are remarkably similar to those described from normal marine, shoreface environments of the Cretaceous Western Interior Seaway (MacEachern and Pemberton, 1992) (Fig. 3.2). Discrimination between wave-dominated delta fronts and non-deltaic, strandplain shoreface environments is difficult. However, the recognition of subtle ichnological and sedimentological aspects of FA1 supports a deltaic interpretation rather than that of a strandplain. Modern examples of wave-dominated deltas include the modern Rhône Delta of Southern France (van Straaten, 1959, 1960a, 1960b; Oomkens, 1967, 1970; Fisher *et al.*, 1969), the Saõ Francisco Delta of southern Brazil (Coleman, 1976; Dominguez, 1996), the Brazos Delta of east Texas (Bernard *et al.*, 1970; Rodriguez *et al.*, 2000) and the Senegal Delta of West Africa (Wright and Coleman, 1972, 1973).

Studies of the sedimentology and ichnology of ancient wave-dominated deltas include the Upper Cretaceous Dunvegan Formation in Alberta (Bhattacharya, 1989; Bhattacharya and Walker, 1991; Gingras *et al.*, 1998) and the Upper Cretaceous Basal Belly River Formation of Alberta (Coates and MacEachern, 1999, 2000; Coates, 2001). The details of modern and ancient wave-dominated deltas are expanded upon in section 3.4.

3.2.2 Facies Association 2 (FA2) - Meandering Distributary Channel Fill (Lower Delta Plain)

Description:

Facies Association 2 (FA2) consists of a fining-upwards, sandstone-dominant succession composed of moderately to well-sorted, medium to very coarse-grained planar tabular and trough cross-bedded, salt and pepper sandstone and minor conglomerates of Facies G. FA2 was identified in five cores from the study area including, core #2 from the A-12 well, cores #2 and #3 from the E-21 well, and cores #2 and #3 from the L-43 well (Fig. 3.7). FA2 is commonly found interstratified with the deposits of FA1 and FA4. The lower bounding surface of FA2 is generally a scoured, erosive surface and the upper bounding surface varies from sharp to gradational. In the study area, this facies association is generally restricted to the upper member of the Kamik Formation. Deposits of FA2 range in thickness from 2.87 to 7.04 m.

Interpretation:

The sediments comprising FA2 are interpreted to represent channel fill deposits of a meandering distributary channel system from a lower delta plain (Fig. 3.8). These meandering channel fill deposits are interpreted to have been part of a larger prograding delta plain complex that was feeding sediments to the time-equivalent deltaic shorelines of FA1 and FA5 within the study area. The stratigraphic relationship between the channel fill deposits of FA2 and the deltaic shoreline deposits of FA1 and FA5 strengthens the interpretation that these meandering distributary channels were acting as sediment point sources for contemporaneous prograding wave- and river-dominated deltas.

Meandering rivers are moderate-energy, sinuous, single channeled, systems with moderate width-to-depth ratios and depositional gradients (Nanson and Gibling, 2004). These river systems tend to develop in settings with low to moderate downslope gradients, cohesive vegetative banks, mixed loads of sand, gravel and mud, and relatively regular levels of fluvial discharge (Reineck and Singh, 1980; Cant, 1982). This type of setting is interpreted to have been present during deposition of the upper Kamik Formation and resulted in the meandering channel fill deposits of FA2. Most river systems generally adopt a meandering profile in the down-valley, lower gradient reaches of a river system (Coleman and Prior, 1982). The river channels are relatively narrow and deep with an asymmetric, v-shaped profile and are generally confined within a single major channel. Meandering river systems are generally more laterally stable because they are surrounded by thick, heavily vegetated, cohesive floodplain deposits and abandoned mud-filled meander loops that are difficult to erode (Cant, 1982). These river systems form by the continual shifting of their positions across a valley bottom and by the deposition of sediments on the inside of meander bends and the simultaneous erosion on the outer banks (Walker and Cant, 1984). Flow in meandering channels is helicoidal with a component of surface flow towards the outer bank and bottom flow towards the inner bank (Fisk, 1947). Within the channel, the axis of maximum flow velocity, the thalweg, alternates from side to side of the channel as helicoidal flow changes its sense of rotation in successive meander loops. Sediment is eroded from the outer, concave bank (cut-bank) of the meander bend by the thalweg and is transported obliquely across the channel and deposited on the adjacent, inner convex bend as a laterally accreting point bar. Point bars display an overall fining-upwards profile which forms due to the progressive waning of flow-energy toward the bar top which results in a decrease in bedform and grain size toward the top of the point bar surface (Allen, 1970). Some of the channel fill deposits of FA2 do not display the characteristic fining-upwards profile of point bars. An absence of these fining-upward cycles within these channel fill successions is attributed to incision and removal of the fine-grained deposits from the upper portions of the point bar sequence by development of a large chute bars.

Point bars are a characteristic feature of meandering river systems and represent the dominant sites for sediment deposition within these river systems (Allen, 1965, 1970; Walker and Cant, 1984). Point-bar surfaces are generally shallow sloping with low dips angle (~4°) that locally exceed 12° and widths about two-thirds that of the channel. The thicknesses of point bars range from 10 to 30 m and widths of individual bars may extend up to 15 km (Coleman and Prior, 1982). The width of the meander-belt deposit is dependent on the discharge and channel depth and in the lower Mississippi River the channel depth averages 50 m with the preserved point bar sand bodies within these channels displaying similar thicknesses (Coleman and Prior, 1980, 1982). A point bar facies model was developed by Miall (1978, 1987), which predicted the pattern of distribution of grain size and bedforms comprising the vertical succession of facies produced by lateral accretion. The vertical sequence of a point bar begins with a scoured base with a channel lag, composed of coarse-sandstone and conglomerate with coalified wood fragments and mudstone rip-up clasts. These basal lag deposits are overlain by a massive, thick sandstone unit composed of plane beds, large-scale trough cross-stratified medium to coarse-grained sandstone that is subsequently overlain by an interval dominated by planar tabular cross-stratified sandstone. The large-scale dunes responsible for the trough and planar tabular cross-bedding migrate primarily during flood periods. Overlying the large-scale cross-bedding is a lower interval of fine-grained, well-sorted sand sequence composed of repeated cyclic sedimentation units of ripple cross-stratified sandstones, climbing ripples, convolute bedding and parallel well-sorted sand laminations and interbedded mudstone layers (Allen, 1970; Harms *et al.*, 1982, Walker and Cant, 1984). These cyclic sedimentation units

Gulf Mobil Parsons L-43 - Core #2

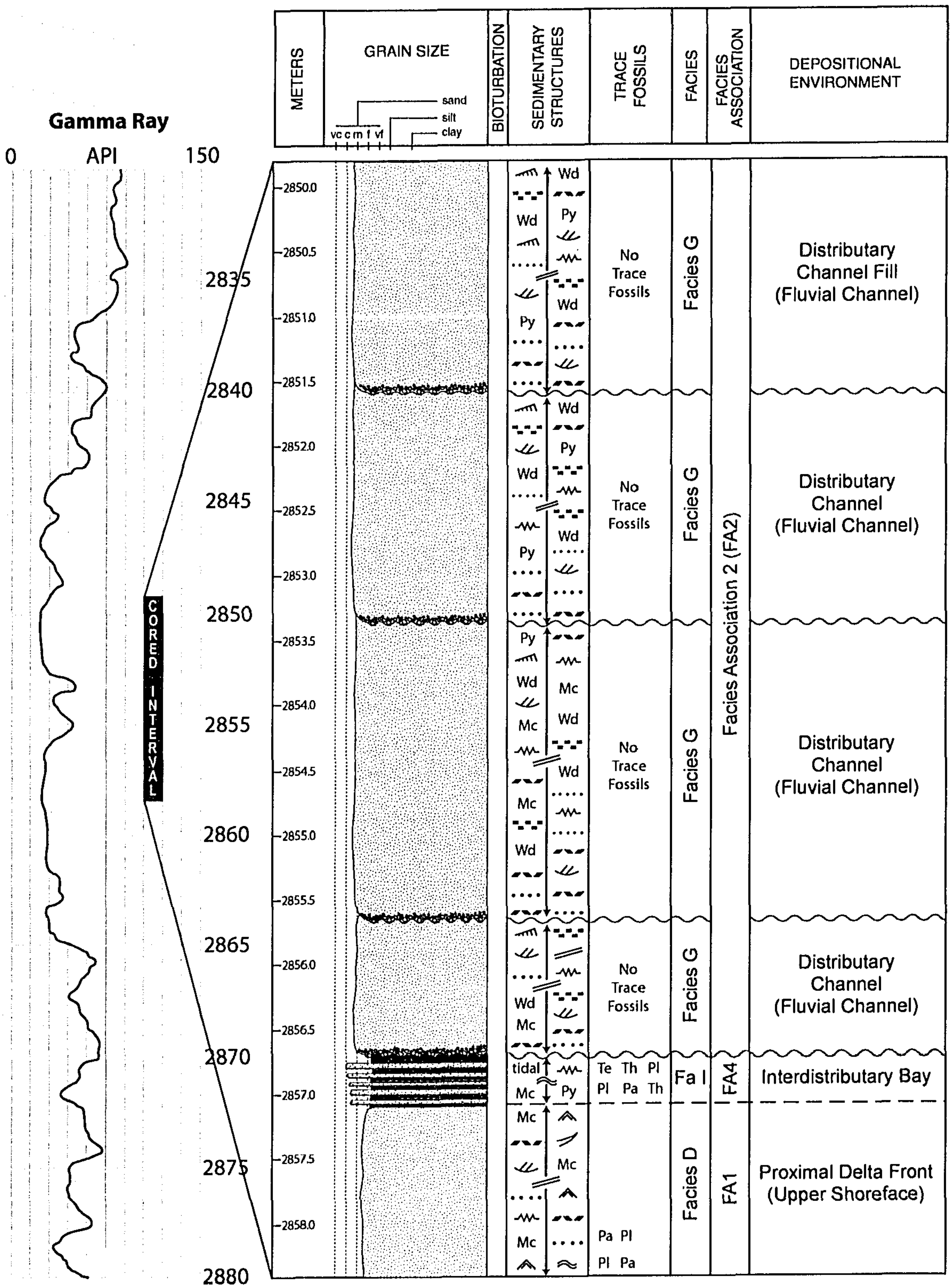


Figure 3.7 Core log of core #2 from the Parsons L-43 well from the Kamik Formation, displaying grain size, lithology, sedimentary structures, ichnological structures, facies, facies associations and interpreted depositional environments.

range in thickness from 0.5 to 1.5 m and represent deposition during a single flood and a change from upper to lower flow regime (Coleman and Prior, 1982). The channel fill successions of FA2 display similar characteristics as point bar deposits described by Allen (1965, 1970) and Walker and Cant, 1984). An example of a meandering channel fill succession of FA2 is illustrated in core #2 from the L-43 well (Figure 3.7). This core is comprised of four, stacked fining-upwards channel fill successions of FA2 that are approximately 1.0 to 2.5 m thick. These successions display many similar features as discussed above, including, a scoured base with a lag, abundant coalified wood fragments and mudstone rip-up clasts and vertical sequence of sedimentary structures including trough and planar tabular cross-bedding and current ripples. The uppermost portion of the fining-upwards point bar sequence is overlain by thick, fine-grained floodplain deposits consisting of clays, silts and ripple laminated fine-grained sandstones which commonly contain roots and caliches depending on the climate (Cant, 1982; Coleman and Prior, 1982).

A distributary channel is a natural flume that accommodates and directs a portion of the discharge and transported sediment from the trunk river system (Coleman and Gagliano, 1965). Observations of the meandering lower Mississippi River shows that distributary channels can be a variety of sizes and shapes with widths ranging from a few meters up to 10 km. These channels can generate thick channel fill deposits, ranging from 1 to 2 m for smaller channels and up to 30 m for the larger channels (Coleman and Prior, 1982). Many of these distributary channel sandstone bodies can also stack both vertically and laterally, forming thick (up to 30 m or greater) multi-storey complexes. Distributary channels systems with high bedload sediment supplies typically contain well-sorted, coarse-grained channel fill sequences with lesser amounts of interbedded silts and shales. Distributary channels from the lowermost portion of lower delta plain systems are rather stable and display minimal degrees of lateral migration (Coleman and Prior, 1980). The stability of the channels is due to incision of distributary channels into underlying prodelta and delta front marine muds. During incision, the channels commonly scour down into, and remove their distributary mouth bar deposits and eventually incise down into the underlying prodelta and delta front clays and/or interdistributary bay muds become entrenched and are unable to meander laterally (Coleman and Prior, 1980, 1982). The resulting lack of lateral channel migration limits the formation of point bars and meander belts in the active delta region forming relatively straight isolated sand bodies. Channel sands and point bars in this setting are deposited by accretion on the narrow inner banks of broad meanders or by the meandering of the thalweg within the confines of an essentially straight channel. However, in sand-rich, high-energy deltas such as those interpreted for FA2, the distributary channels will tend to migrate laterally and form point bar sands (Coleman and Prior, 1982).

The channel fill deposits of FA2 are interpreted to reflect point bar deposits of a meandering river system that was discharging into a deltaic system (Fig. 3.8). These deposits were interpreted as meandering channel fill deposits due to the poor to moderate sorting of the sediments, presence

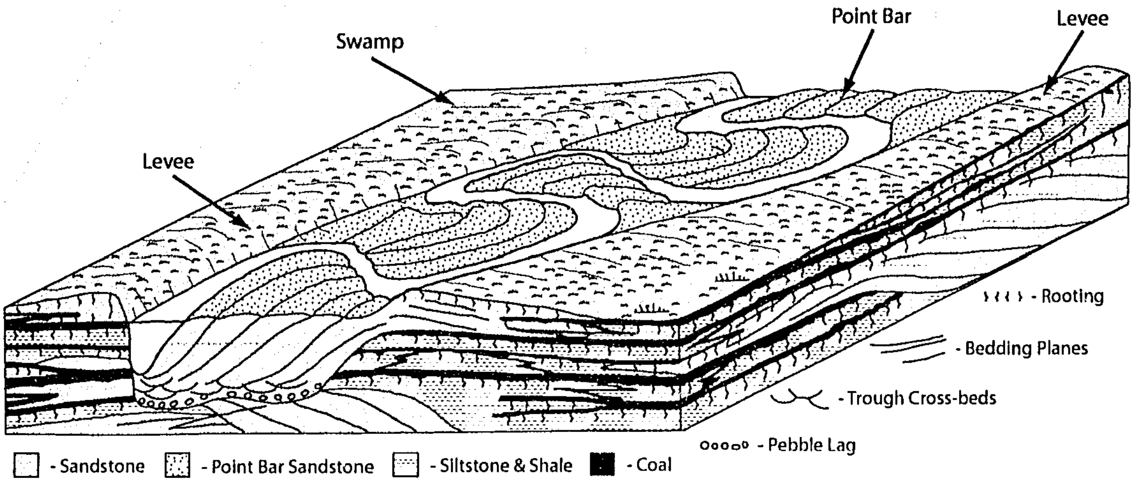


Figure 3.8 Schematic depositional facies model for a meandering distributary channel system (modified from Horne *et al.*, 1978)

of a scoured basal surface with associated lags, and the fining-upwards grain size profiles. The overall fining-upwards succession with a basal lag, overlain by trough and planar tabular cross-bedding capped current ripples is attributed to waning flow conditions characteristic of a point bar deposited in a distributary channel (Reading and Collinson, 1996). The prevalence of trough and planar tabular cross-bedding and current ripples originates from traction and saltation processes and reflects deposition in a moderate to high-energy setting dominated by unidirectional current flow, consistent with deposition in a fluvial channel (Harms *et al.*, 1975, 1982). The sub-cycles within the overall fining-upwards succession suggests channel infilling by a series of waning flows, likely associated with the flood stages of a river. The abundance of coalified wood fragments and mudstone rip-up clasts in these deposits is interpreted as intraformational material from the surrounding vegetated floodplains along the cut banks of the channel. The absence of trace fossils within FA2 is attributed to the harsh ecological conditions associated with deposition in a distributary channel. Numerous biological stresses including overall high-energy conditions, very low salinities, high sedimentation rates, turbid water conditions and constant migration of bedforms all contribute to impede the colonization of these sediments by infaunal organisms (Pemberton and Wightman, 1992). Deposition of these channel fill deposits is interpreted to have originated from increased sediment supplied by the uplift of the adjacent Eskimo Lake Arch located to the east of the study area. The predominance of meandering channel deposits of FA2 in the upper Kamik Formation, compared to the older braided distributary channels of FA3 from the lower Kamik is attributed to a decrease in the downslope gradients and sediment supply associated with a decrease in tectonic activity and the initiation of a period of regional tectonic quiescence in the Brooks-Mackenzie Basin.

3.2.3 Facies Association 3 (FA3) - Braided Distributary Channel Fill (Lower Delta Plain)

Description:

Facies Association 3 (FA3) consists of multiple, stacked fining-upwards successions comprised of poorly-sorted, scour-based, interstratified medium to very coarse-grained, salt and pepper textured sandstones and conglomerates of Facies G. FA3 was identified in three cores from the study area including core #3 from the L-43 well and cores #1 and #2 from the N-10 well (Fig. 3.9). This facies association was only identified from the strata of the lower member of the Kamik Formation and is intercalated with deposits of FA4. The lower bounding surface of FA3 is a scoured surface and the upper bounding surface is sharp to gradational. Strata of FA3 are widespread throughout the study area and display a sheet-like geometry with relatively good stratigraphic continuity. FA3 is interstratified with, and commonly incises into deposits of FA4 and ranges in thickness from 6.11 up to 9.85 m.

Interpretation:

The sediments comprising FA3 are interpreted to represent channel fill deposits of a braided distributary channel system from a lower delta plain (Fig. 3.10). These braided channel fill deposits are interpreted to have been part of a larger prograding delta plain complex that was feeding sediments to a time-equivalent deltaic shoreline that was situated to the west of the study area. In addition, the stratigraphic relationship of FA3 with the lower delta plain deposits of FA4 supports the interpretation that these channels were part of a large prograding delta plain. This braided distributary system may have extended from the delta plain all the way to the delta front as a braided system, but due to limited core data from the lower Kamik to the west of the study area, this idea is speculative. It was noted by Coleman and Prior (1982) that modern deltas from tectonically active settings commonly display braided distributary channel systems that extend all the way to the delta front.

Braided rivers are relatively high-energy systems with large width-to-depth ratios and multiple branching channels that divide and rejoin around alluvial bars (Coleman and Prior, 1982; Nanson and Gibling, 2004). The development of braided river systems is favoured in tectonically active settings with high downstream gradients, rapid, large fluctuations in river discharge, large quantities of bedload-transported coarse sediment and easily erodible, non-cohesive banks (Cant, 1982; Walker and Cant, 1984). These river systems tend to be shallow, relatively wide and composed of unstabilized mid-channel bars of sand or gravel which are emergent during periods of low to moderate discharge (Cant, 1982). Braided river systems are characterized by rapid and continuous shifting of sediment and channel positions and are marked by successive divisions and rejoining of the individual channels around numerous mid-channel alluvial islands and braid bars (Miall, 1978;

Reineck and Singh, 1980; Coleman and Prior, 1982). The channels tend to have depths and widths which are highly variable and do not conform to the simple pattern shown by meandering systems. Braided rivers tend to aggrade vertically by deposition of sediment in one part of the system with diversion of flow to other parts of the system (Cant, 1982). Sediments are typically dominated by gravel and sand, but silt-dominated systems are known (Nanson and Gibling, 2004). The coarse-grained sands and conglomerates are transported as bedload, whereas the finer-grained materials (clays and silts) tend to be transported through the system without accumulating (Walker and Cant, 1984). Most deposition within braided channels occurs during the waning stages of floods when sediments rapidly aggrade and form vertically accreted channel deposits. Braiding is typically a feature of the upstream reaches of a river, with meandering becoming more common downstream as the slope and coarseness of load decreases (Cant, 1982). Braided rivers are prominent near mountain and glacial fronts, range from humid to arid regions, and commonly change downstream into fine-grained, single-channel systems, although some feed directly into oceans as braid deltas.

The deposits of FA3 are identified as braided channel fill deposits due to their poorly-sorted nature consisting of medium to very coarse-grained sandstone and conglomerate. The channel fill successions of FA3 display similar characteristics as described from modern and ancient braided river systems (Cant, 1982; Walker and Cant, 1984). An example of a braided channel fill succession of FA3 is illustrated in core #1 from the N-10 well (Figure 3.9). The braided channel fill succession shown in this core is sand-dominated and is comprised of multiple, stacked, erosionally-based, fining-upward cycles of sandstone and conglomerate that are approximately 0.25 to 1.5 m thick. Each fining-upward cycle is comprised of a basal lag deposit overlying a scoured base, which in turn is overlain by an interval dominated by trough and planar tabular cross-bedded sets capped by current ripple lamination. The trough and planar tabular cross-bedding is formed in the deeper channels by the migration of sinuous-crested dunes along the channel floor (Walker and Cant, 1984). Wright (1985) observed that planar tabular and trough cross-stratification is characteristic of river systems such as braided systems which transport a high percentage of bedload and are generally composed of pronounced sand-rich channel fills.

It has been recognized by Hamblin and Walker (1979) that an abundance of coarse-grained sediments and a lack of mud in the distributary channel deposits indicates a high bedload transport system, more typical of a confined, braided system rather than a meandering one. The absence of mud in the channel fills of FA3 further supports the interpretation of a braided distributary system. During waning stages of floods, the channel beds aggrade and flood stage sedimentary structures are preserved in the upper portions of these cycles, including current ripple lamination, scour and fill structures, small-scale convolute bedding, organic detritus and thin mudstone layers (Reading and Collinson, 1996). The preserved thickness of the stacked, fining-upward channel fill cycles is also dependent on variations in the scouring depths of successive overlying channels. The absence of the upper, fine-grained portion of the channel fill cycles in FA3 is attributed to erosional truncation

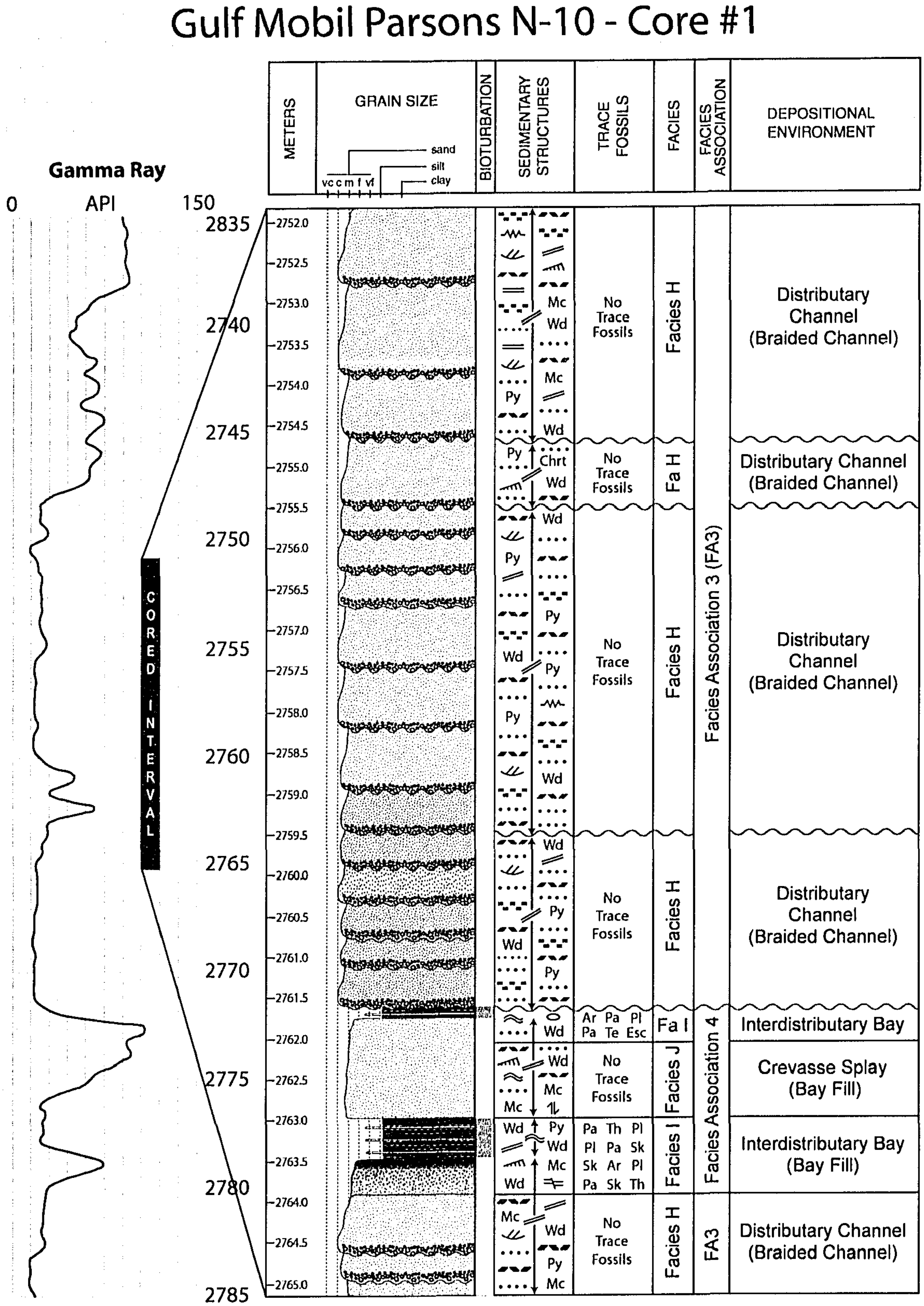


Figure 3.9 Core log of core #1 from the Parsons N-10 well from the Kamik Formation displaying grain size, lithology, sedimentary structures, ichnological structures, facies, facies associations and interpreted depositional environments.

due to scouring from the lateral migration of successive overlying channels. These deposits are generally complex due to being formed by the coalescence and accretion of smaller bedforms, and the repeated periods of erosion and deposition resulting from many different flood events. Due to the high lateral migration rates and shallow scour depths, most braided channel fill deposits display a sheet-like geometry and good lateral stratigraphic continuity as recognized for FA3. These deposits are typically thin, with most deposits averaging 15 to 25 m thick with maximum thicknesses up to 30 m (Coleman and Prior, 1982). These braided channel fill deposits of FA3 are distinguished from the meandering channel fills of FA2 by their internal stacking patterns consisting of decimeter-scale, stacked, fining-upwards cycles. Channel fills from meandering fluvial systems typically display thicker, single fining-upward successions that do not display numerous erosional basal contacts and internal truncations (Walker and Cant, 1984).

Braided river systems are most commonly found in tectonically active settings with arid to arctic climates (Coleman and Prior, 1982). During deposition of the lower Kamik Formation there was significant tectonic activity proximal to the study area, including fault movements in the adjacent Eskimo Lakes Fault Zone and uplift of the Eskimo Lakes Arch (Dixon, 1982a, 1991). This tectonic uplift likely resulted in high downstream gradients and provided large quantities of coarse clastic sediments proximal to the study area and encouraged the development of the braided distributary systems responsible for FA3. The predominance of braided channel fill deposits of FA3 in the lower Kamik Formation compared to the meandering nature of the channels of FA2 in the upper Kamik is attributed to the aforementioned uplift of the Eskimo Lakes Arch and tectonic activity in the Eskimo Lakes Fault Zone. It is speculated that during deposition of the younger distributary channel fills of FA2 there was a decrease in the downstream gradients and sediment supply associated with a decrease in tectonic activity in the Eskimo Lakes Fault Zone and the initiation of a period of regional tectonic quiescence.

During the Lower Cretaceous, the Mackenzie Delta region was positioned in a northern setting and the climate was likely arid to humid and was much warmer than the present Arctic climate (Frakes, 1979). Climates in these northern settings are strongly seasonal and channels developed are generally characterized by sporadic discharge and a high proportion of bedload and the poor development of vegetative material required to stabilize channel bars. The poor development of vegetated floodplains and cohesive channel banks would have encouraged large, erratic seasonal fluctuations in fluvial discharge rates, rapid and continuous lateral migration of bars and channel position across wide areas, resulting in the development of a braided channel system. However, it should be noted that the climate of the Brooks-Mackenzie Basin did not significantly change between the deposition of FA2 and FA3, therefore tectonics are interpreted to been the most significant factor in the development of the braided versus meandering channel systems. Similar to the deposits of FA2, there is an absence of trace fossils within FA3 that is attributed to the harsh ecological conditions associated with deposition in a distributary channel. The absence of trace

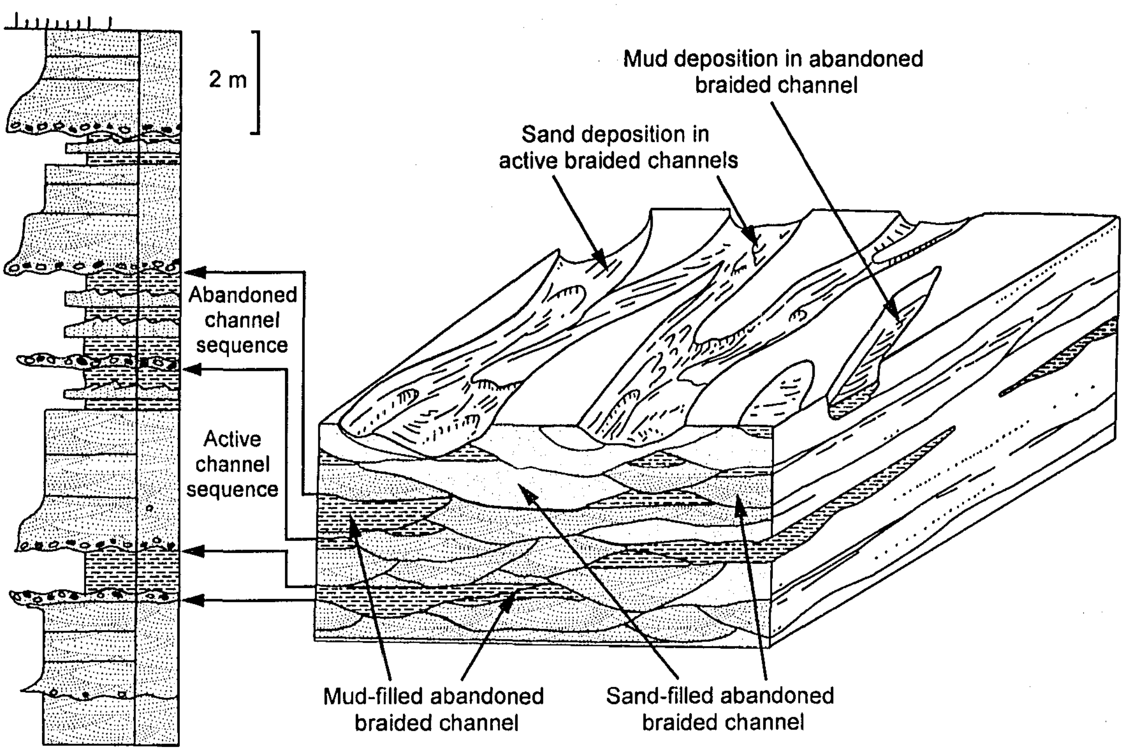


Figure 3.10 Depositional facies model for a braided channel system (modified from Selley (1976, 1982)

fossils within these deposits is attributed to numerous biological stresses including overall high-energy conditions, very low salinities and/or fluctuating water chemistry, high sedimentation rates, turbid water conditions and continuously migrating bedforms. Any of these individual stresses or a combination of them would have impeded the colonization of these sediments by infaunal organisms (Pemberton and Wightman, 1992).

A modern analogue for the braided fluvial system responsible for FA3, and the majority of the channels comprising the lower member of the Kamik Formation is the sandy, braided South Saskatchewan River (Miall, 1977; Cant and Walker, 1978). The braided channels comprising the South Saskatchewan River are 3 to 5 m deep and are up to 200 m wide and are characterized by channel bars, sand flats, vegetated islands and floodplains. The channels of the South Saskatchewan River are characterized by long, straight-crested, channel bars and dunes which form trough and planar tabular cross-bedding with smaller tabular sets with parallel lamination, trough cross-bedding and ripple cross-lamination (Cant and Walker, 1978). A typical channel fill sequence is similar to those described for FA3 and consists of a basal lag, overlain by trough cross-beds which decrease in scale upwards commonly interstratified with planar tabular cross-bedding finally overlain by an upper portion composed of ripple cross-laminated sandstones and mudstone layers (Cant and Walker, 1978). These sequences are generally greater than 5 m thick and are overlain by 0.5 to 1 m of silty and muddy vertical accretion deposits (Fig. 3.10). The sandy braided distributary

Figure 3.11 Air photo of a wave-dominated delta and associated emerged strandplains from Andersrag Beach, north of Cape Storm from the southwestern coast of Ellesmere Island of the Canadian Arctic Islands. This strandplain beach system is fed by a coarse-grained braided distributary channel system which extends all the way to the delta front. This system is formed due to post-glacial rebound of the land surface following Laurentide Glaciation. The modern environment is a excellent analog for the lower member of the Kamik Formation. Note: the air photo has been reversed to approximate the orientation of the shoreline and distributary channel system during deposition of the lower Kamik Formation (from Blake, Jr., 1975).

channel deposits of FA3 are interpreted to represent the fluvial point sources of a deltaic system that was responsible for the deposition of the lower member of the Kamik Formation. Figure 3.11 illustrates a wave-dominated delta from Andersrag Beach located north of Cape Storm from southwestern coast of Ellesmere Island from the Canadian Arctic Islands. This wave-dominated delta and associated emergent strandplain is interpreted as an excellent modern analogue for the lower member of the Kamik Formation. The important feature of this delta is the coarse-grained braided distributary channel system, which extends to the delta front. The distributary channel is similar to that interpreted for FA3 and illustrates that in some deltaic settings braided channels can extend to the delta front. This modern analogue is not a true wave-dominated delta and was formed by post-glacial rebound of the land surface following Laurentide Glaciation (Blake Jr., 1975).

3.2.4 Facies Association 4 (FA4) - Interdistributary Bay/Brackish Bay/Crevasse Splay/Marsh/ Paleosol (Lower Delta Plain)

Description:

Facies Association 4 (FA4) consists of an interstratified succession composed both of coarsening- and fining-upwards successions of sandstones, siltstones, mudstones and coals of Facies I, J, K, L and M. FA4 has been identified in from study area core #1 from the A-12 well, core #1 and #3 from the E-21, core #1, #2 and #3 from the L-43 well, core #1 from the F-09 well and core #1 and #2 from the N-10 well (Fig. 3.12). FA4 is interstratified with FA1, FA2, FA3 and FA5 and was identified both within the lower and upper members of the Kamik Formation. The lower bounding surface of FA4 can be sharp, gradational to slightly undulatory with the upper bounding surface generally scoured, undulatory or sharp. Deposits of FA4 range in thickness from 4.00 to 11.24 m with most cores containing only partial successions.

Interpretation:

The strata comprising FA4 are interpreted to represent the deposits of a prograding lower delta plain environment (Fig. 3.13). FA4 is comprised of interdistributary bay, brackish bay, crevasse splay, marsh and paleosol deposits of Facies I, J, K, L and M. The lower delta plain lies within the realm of river and marine interaction and extends landward from the low tide mark to the landward limit of tidal influence (Coleman and Prior, 1980). Delta plains are complex depositional settings that are composed of many subenvironments including interdistributary bays, marshes, swamps, tidal flats, distributary channels, beaches, beach ridges, dunes and bay fill deposits (Coleman and Prior, 1982). The lower delta plain is typically a brackish to marginal marine environment and along with the delta front, is a site of active deposition, mostly by crevassing or overbank flooding of the distributary channels (Suter, 1994). These lower delta plain deposits are interpreted to have been part of a larger prograding delta plain that was deposited contemporaneously with the wave-dominated deltas of FA1, the meandering and braided distributary channels of FA2 and FA3 and the river-dominated deltas of FA5. During the deposition of each of the other facies associations, the environments responsible for FA4 existed; however, it is important to recognize that they were all not time equivalent with each other.

Interdistributary bays were defined by Coleman and Gagliano (1964) as the areas between deltaic distributaries, irrespective of whether the bays are open to the sea, partially open or entirely closed. Bays are typically infilled by muddy sediments and are also infilled by overbank spilling of fine-grained material from the river during flood stages (Bhattacharya and Walker, 1992). Interdistributary bay and crevasse splay sediments represent one of the most extensive facies and depositional environments associated with many deltas (Coleman and Prior, 1982; Wright, 1985). These sediments form the major land areas in the lower delta plain and can contain freshwater, brackish or marine waters. Bays are normally elongate, shallow, open bodies of water ranging in size from a few hundred meters to approximately 15 to 20 km (Coleman and Prior, 1982). Most bays are shallow water bodies, rarely exceeding 7 to 8 m depths with average depths of approximately 4 m (Elliott, 1974; Coleman and Prior, 1980). Interdistributary bays sediments are composed of levees, crevasse splay deposits and minor channels and are generally completely surrounded by distributary channels and extensively vegetated marshes (Elliot, 1974; Coleman and Prior, 1980). Bays are often infilled by fine-grained sediments such as clays and silts that are brought into the bays during flood stages or abnormally high tides associated with storms. The settings are normally bypassed by active coarse clastic sedimentation. Interdistributary bays are more common in fluvial-dominated deltas than in wave-dominated deltas (Elliott, 1996). In wave-dominated deltas, interdistributary bays are often completely closed off by barrier island and beach complexes (i.e. the Nile and Saõ Francisco deltas), which results in the development of extensive backbarrier lagoons (Bhattacharya and Walker, 1992).

Bays are influenced by a complex interaction of shallow marine depositional processes, riverine influence and brackish-water faunal activity that impart lithologic and textural variability and produce a wide range of sedimentary structures (Wright, 1985). Open bays can be significantly affected by wave influence and diurnal tides can play important roles in water circulation and the distribution of fine, suspended sediment (Arndorfer, 1973). Under normal conditions bay waters are quiet and commonly ponded, however this placid regime is frequently interrupted during floods when excess discharge from distributaries is diverted into the bay. Elliot (1974) noted that these flood-generated incursions dominate interdistributary bay sedimentation and combine to produce an association of sedimentary sequences, each of which represent the infilling of the shallow bay. The main processes responsible for the infilling of shallow interdistributary bays include overbank flooding and crevasse splaying during flood periods and high fluvial discharge (Elliott, 1974). Overbank flooding operates during a single flood event and involves the transport of sediment-laden waters spilling over the channel banks as sheet flow, with no breaches or crevasse channels

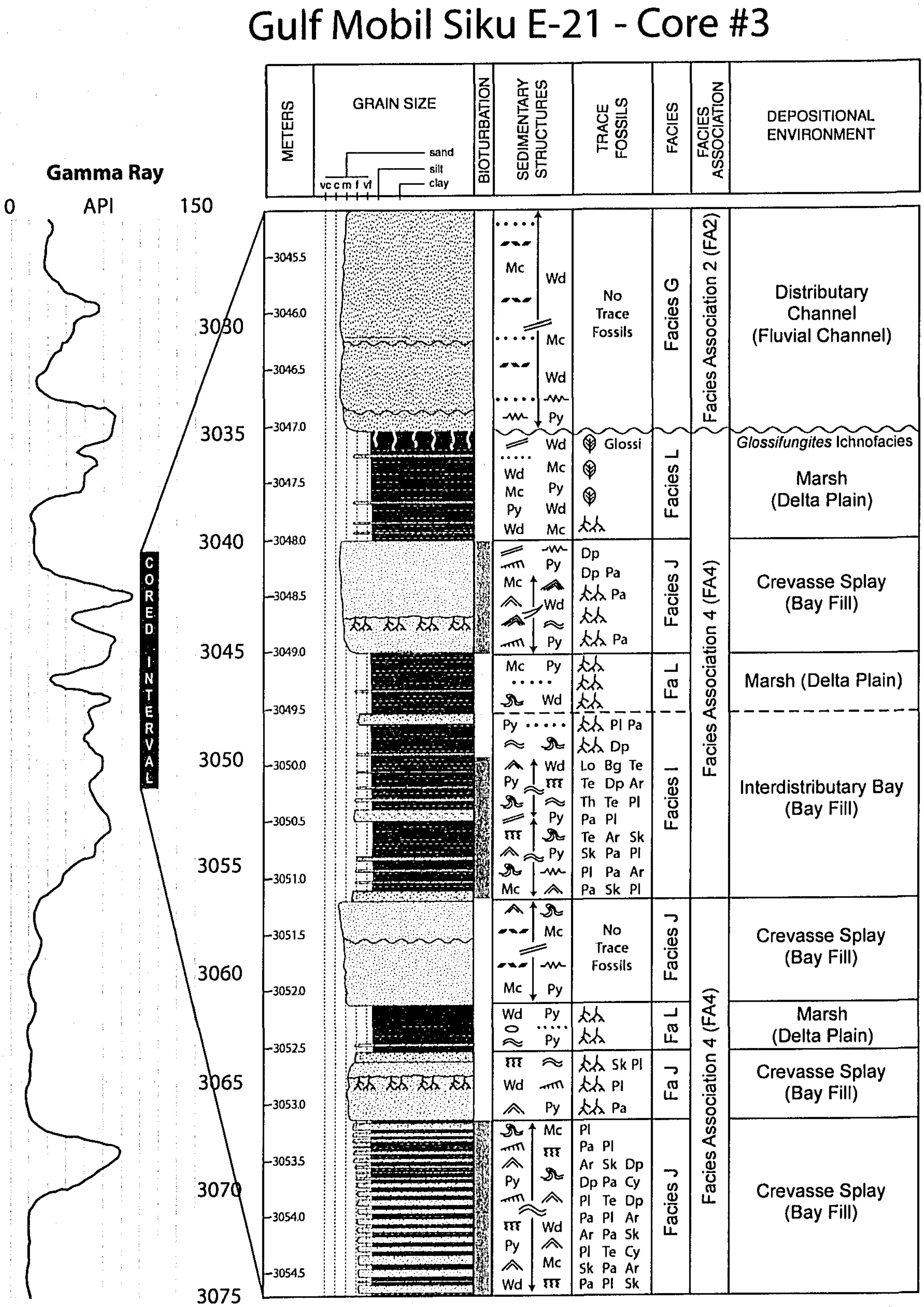


Figure 3.12 Core log of core #3 from the Siku E-21 well from the Kamik Formation, displaying grain size, lithology, sedimentary structures, ichnological structures, facies, facies associations and interpreted depositional environments.

(Coleman, 1969). During this flooding event fine-grained, suspended sediment is deposited over the entire bay and thin muds and silts with coarse sediment confined to the bay margins which contributes to the development of the channel levees. Crevasse splays are deposited into the shallow bays as sudden, discrete incursions of sediment-laden density currents and sheet sands generally during floods when the distributary channel margins are breached and sediment-laden floodwaters are rapidly deposited in the bay via crevasse channels (Elliott, 1974).

The wavy to lenticularly bedded, bioturbated, fine-grained silty sandstones, siltstones and organic mudstones of Facies I and J are interpreted to reflect deposition in an interdistributary bay environment. The predominance of fine-grained sediments including silty sandstones, siltstones and organic mudstones comprising Facies I and J indicate that the sediments were deposited in a low energy environment dominated by suspension sedimentation. The presence of current ripples is consistent with deposition in an environment dominated by unidirectional current flow (Harms *et al.*, 1975), such as crevasse splaying or overbank flooding into an interdistributary bay. The wave ripples are a result of waves from the open marine realm reworking the thin sands in the bay. Synaeresis cracks are indicative of a setting with rapidly fluctuating salinities reflecting intermittent influx fresh water into the marine waters of the bay (Burst, 1965; Plummer and Gostin, 1981). The soft-sediment deformation structures are attributed to the rapid deposition of sediment associated with crevasse splaying and overbank flooding of distributary channels. In addition, the presence of siderite is indicative of deposition in a littoral, shallow-water setting that contained an abundance of organic material (Teodorovich, 1961). The ichnological suite is mixed, comprised mainly of dwelling, suspension and deposit-feeding burrows including *Arenicolites*, *Skolithos*, *Diplocraterion*, *Thalassinoides*, *Palaeophycus*, and *Cylindrichnus*, feeding structures including *Planolites*, *Teichichnus*, resting traces such as *Bergaueria* and *Lockeia*. Some intervals are dominated by *Teichichnus* and are interpreted to reflect stressed brackish water interdistributary bays (Wightman *et al.*, 1987; Pemberton and Wightman, 1992). The overall low diversity assemblage is consistent with deposition in a low energy, open to brackish marine interdistributary bay environment (Pemberton and Wightman, 1992).

Crevasse splays and bay fill sediments represent one the most significant facies and accumulations of sediment in the lower delta plain of many deltaic environments (Coleman and Gagliano, 1964; Coleman and Prior, 1982; Wright, 1985). Crevassing is more frequent in the lower delta plain because natural levees are usually more poorly developed in this setting than they are in the upper delta plain (Suter, 1994). The sediments comprising crevasse splays generally build into shallow bays between or adjacent to major interdistributary channels and extend themselves seaward through a system of radial bifurcating channels (Fig. 3.13) (Coleman and Prior, 1982). Deposition of splay deposits within interdistributary bays generally originates during flood stage periods with the initial breaching of distributary channel margins and the debouching of sediment-laden floodwaters into the surrounding bay via small crevasse channels in response to a local surface gradient advantage.

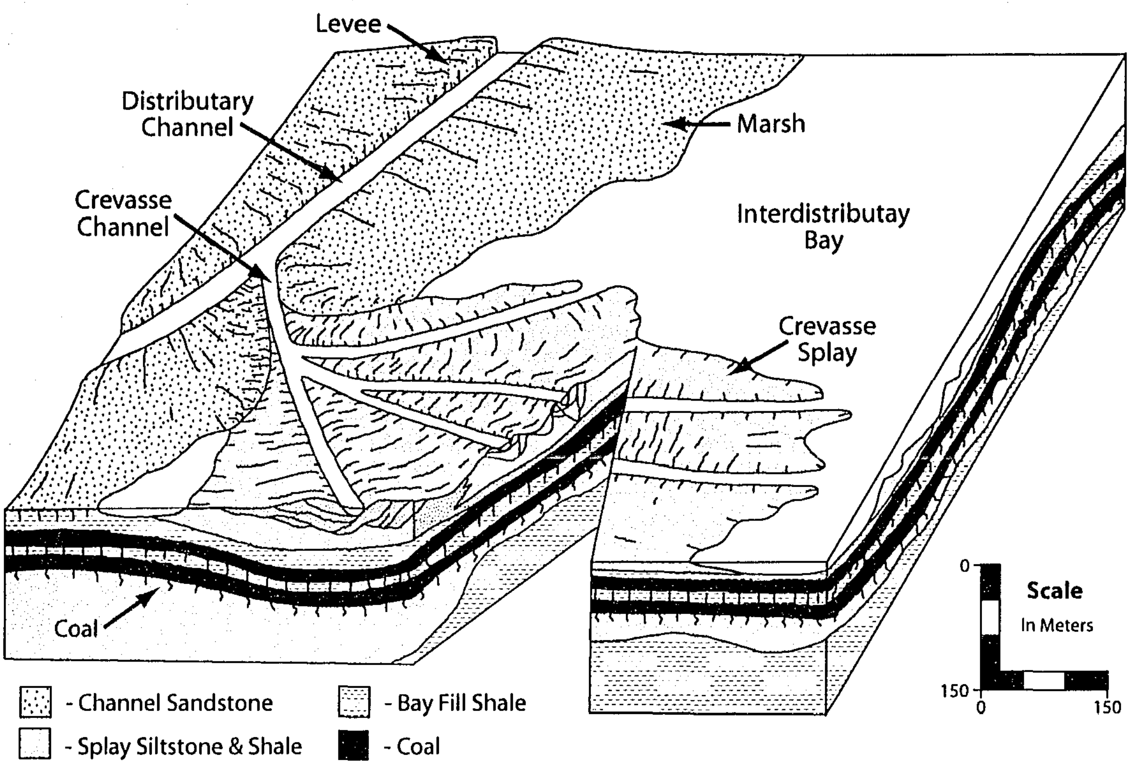


Figure 3.13 Schematic diagram showing the shape and depositional associations of crevasse splay and crevasse channel sand bodies, interdistributary bay and marsh deposits from outcrop exposures in Carboniferous of eastern Kentucky, USA (modified from Bangz *et al.,* 1975).

The breaching of the channel is immediately followed by a rapid influx of sediment building of the crevasse splay as a fan-shaped deposit with a geometry similar to a Gilbert type delta. The splaying process is responsible for the rapid deposition of coarse-grained crevasse splay sediments into the interdistributary bays and floodplains surrounding the distributary channels (Elliott, 1974). Once the crevasse channel is established, flow through the distributary channel margins gradually increases through successive floods until it reaches a peak of deposition, wanes, and becomes inactive (Coleman and Prior, 1982). Due to the lack of confinement of the flow, crevasse splay deposits generally have a fan-shaped morphology (MacEachern, 2001).

The process of bay infilling by crevasse splaying generally forms a vertical sequence of relatively thin, stacked coarsening-upwards and fining-upwards facies successions. The splay succession typically grades into rooted coaly mudstones or coals representing a variety of swamp and marsh environments (Bhattacharya and Walker, 1992). An example of this vertical sequence is illustrated in core #3 from the E-21 well where three crevasse splay successions grade upward into marsh and interdistributary bay deposits (Figure 3.7). The sediments of the crevasse splay are typically coarser nearer the main channel and fine upward toward the open interdistributary bay. The lower portion is generally composed of shallow brackish water clays and organic detritus and the upper portion is composed of well-sorted sandstones. The upper sandstone-dominant portion of the crevasse splay is essentially the distributary mouth bar deposits associated with a prograding distributary channel and delta front (Coleman and Prior, 1980, 1982). The process of crevassing typically creates lobe-shaped sand bodies that vary in thickness from 3 to 15 m but are generally less than 10 m thick (Elliott, 1974; Coleman and Prior, 1980, 1982). Modern crevasse splay deposits in the Balize complex of the Mississippi delta cover an average area of approximately 2 square kilometres (Arndorfer, 1973) and generally form over a period of approximately 100 to 150 years (Coleman and Prior, 1982). Most bay fills are relatively thin, however continuing subsidence and stacking of bay fill sediments builds up a thick sequence of lower delta plain deposits.

Pemberton and Wightman (1994) identified a typical series of events for the deposition of the coarsening-upwards crevasse splay deposits. Initial sediment deposition originates with a detrital flux of fine-grained sediments into the bay at relatively low rates, as indicated by the moderate to high degree of bioturbation of the mudstones. Following the breach of the distributary channel margin and the development of a crevasse splay, coarse-grained sediments and plant debris are deposited into the bay. As deposition of the crevasse splay continues, the overall sedimentation rate becomes greatly increased but deposition is also highly variable with flood events in the channel corresponding to event bed deposition in the splay. Shallowing occurs as the bay is progressively infilled by the splay deposits and once the bay becomes shallow enough the bay fill sediments provide a platform for the establishment of plant growth. Rooting along the tops of the splay deposits is a similar feature in FA4 and indicates that the top of the splay became stable enough for vegetation to become established. The final filling of the bay by the splay deposits causes an abandonment of the crevasse splay and the establishment of a marsh overlying the splay sediments. Once the crevasse splay is abandoned and becomes inactive, the deposits subside and the crevasse system is inundated by marine waters and then reverts back to an open bay environment (Coleman and Prior, 1982). The process continues in a cyclical succession of crevasse splay deposition, subsidence and subsequent abandonment. The crevasse splay deposits of the Kamik Formation are very similar to the lower delta plain bay fill sequences described from the modern Mississippi delta (Coleman, 1976; Coleman and Prior, 1982) and from the Mannville Group of Alberta (Pemberton and Wightman, 1994).

The fine to medium-grained, cross-bedded to flaser bedded sandstones of Facies K are interpreted to reflect crevasse splay deposits infilling an interdistributary bay similar to those described by Coleman (1976), Coleman and Prior (1982) and Pemberton and Wightman (1994) (Fig. 3.13). The overall coarsening-upward nature, sandstone-dominant lithology, abundance of organic detritus and micaceous material, episodic nature of sedimentation and relatively thin nature of these deposits (ranging from 0.5 to 3.0 m) is characteristic feature of splay sandstones (Coleman and Prior, 1980). The relatively well-sorted nature of these sandstones with low angle cross-bedding, climbing ripple laminae, current and wave ripples indicate are also consistent with deposition of a crevasse splay. The presence of abundant climbing ripples and soft-sediment deformation structures reflects high sedimentation rates associated with rapid deposition of sediments by crevassing. The predominance of current ripple structures throughout these deposits reflects that unidirectional currents strongly influenced sediment deposition (Harms *et al.*, 1975). Coleman and Gagliano (1965) identified similar cross-bedded sandstones, climbing ripples, wave-ripple lamination and trough cross-bedding from crevasse splays from the lower delta plain of the modern Mississippi delta. The abundant organic and micaeous material within these sediments originates from the sediment-laden waters of the crevasse channel. Crevasse channels are relatively shallow compared to major distributary channels and these channels only tap the upper portion of the water column from the main channel and mainly divert the suspended load (MacEachern, 2001). As a result of this process, splay deposits are relatively fine-grained and sediments such as organic detritus and mica carried in suspension are deposited along with the crevasse splay sands. The ichnological suite identified in the crevasse splay deposits includes mainly suspension-feeding burrows such as *Arenicolites*, *Skolithos*, *Diplocraterion*, *Bergaueria* and *Lockeia* and lesser amounts of deposit-feeding and dwelling burrows such as *Teichichnus, Planolites*, *Palaeophycus* and *Thalassinoides.* This assemblage corresponds mainly to the *Skolithos* ichnofacies and is indicative of deposition in a low to moderate energy, sandy, shallow marine environment consistent with the infilling of bay. The abundance of coalified root traces at the top of many of these crevasse splay successions reflects colonization of these substrates by vegetation due to overall shallowing-upwards conditions associated with infilling of the bays by crevasse splays and the establishment of a marsh.

The dark grey to black coaly mudstones and coals of Facies L are interpreted to reflect deposition in a marsh environment in the lower delta plain. The deposits of Facies L and M display many characteristics that are similar to the marsh deposits described by Coleman and Prior (1980) from the modern Mississippi delta. The high amount of preserved organic material and in-situ coalified rootlets reflects deposition in a very low energy environment with minor suspension sedimentation. Marshes from the Mississippi delta are characterized by extreme organic activity that results in the deposition of abundant organic detritus and evidence of active plant rooting. The abundance of organic material in FA4 including coalified rootlets and wood fragments reflects the abundant of plant life in this setting and supports the interpretation that these sediments were deposited in a marsh. The presence of preserved rootlets implies that these deposits were formed subaqueously in a shallow water setting that was favourable for colonization by plants and was not subaerially exposed for any significant amount of time. The dark grey to black, waxy, carbonaceous and pyritized mudstones of Facies M are interpreted to reflect a paleosol, developed in the subaerially exposed portion of the marsh. The absence of burrowing within these sediments is consistent with deposition in a marsh environment that was inhospitable for infaunal organisms. Overall, the strata of FA4 display many characteristics of that support the interpretation that these sediments were deposited in a lower delta plain.

3.2.5 Facies Association 5 (FA5) - Moderately Storm-Influenced, River-dominated Delta

Description:

Facies Association 5 (FA5) forms gradual coarsening and shallowing-upward succession comprised of Facies N and Facies O. FA5 displays a similar overall upward increase in the overall grain size, bed thickness, and proportion of sandstone as recognized for FA1. FA5 was only identified from a single core (core #1 from the Siku A-12 well) from the study area and this facies association was only recognized from the upper member of the Kamik Formation. FA5 is overlain by FA4 with an upper bounding surface that is undulatory to gradational in nature. The lower bounding surface is not preserved in core and its exact nature is unknown, however it is assumed to be gradational with underlying distal deposits. Deposits of FA5 from the A-12 well are 11.23 m thick, but this core only represents a partial sequence of FA5 and it is likely that complete successions are greater in thickness.

Interpretation:

The strata of FA5 are interpreted to represent deposition from the progradation of a moderately storm-influenced, river-dominated deltaic system. FA5 displays an overall coarsening-upwards succession displaying the landward progression from the distal delta front deposits of Facies N to the proximal delta front deposits of Facies O (Fig. 3.14). Prodelta deposits were not identified as part of FA5 due to limited core data, although it is likely that these facies do exist in the study area. The distal delta front succession of FA5 is represented by the cross-bedded to soft-sediment deformed, fine-grained, silty sandstones and mudstones of Facies N. The proximal delta front succession consists of medium-grained, cross-bedded sandstones with abundant phytodetrital material. FA5 displays a similar coarsening-upwards profile as the wave-dominated deltaic sediments of FA1, however its distinguished by it is more irregular profile and the greater proportion of interbedded mudstone and soft-sediment deformation structures.

The distal delta front sandstones and mudstones of Facies N are characterized by ubiquitous soft-sediment deformation structures intercalated with low-angle parallel lamination, interpreted to represent HCS and SCS. The soft-sediment deformation structures are interpreted to result from high sedimentation rates and rapid deposition of sediments onto unstable, soupy substrates that undergo minimal degrees of wave reworking (Moslow and Pemberton, 1988; Bhattacharya and Walker, 1992). Soft-sediment structures are especially common in modern river-dominated deltas and numerous authors have recognized an abundance of similar convoluted intervals from the delta front of the modern Mississippi delta (Coleman and Gagliano, 1965; Coleman and Garrison, 1977; Prior and Coleman, 1978, 1980; Coleman and Prior, 1980, 1982; Coleman *et al.*, 1983; Wright *et al.*, 1988). In addition to the soft-sediment deformation structures are the abundant, large-scale aggradational wave ripples that are attributed to high sedimentation rates, rapid burial and minor reworking by waves. The predominance of soft-sediment deformation structures and aggradational structures in Facies N suggests rapid deposition and unstable substrates (Moslow and Pemberton, 1988). The structures are interpreted to have originated from episodes of high precipitation which resulted in high flood discharge through rivers and deltaic networks that deposited sediments faster than they could be reworked and stabilized by basinal processes (Pemberton *et al.*, 2001). These structures can be attributed to gravity-induced slumps, differential overloading and slow mass movement of saturated sediments on an inclined slope, processes that are common in river-dominated delta fronts.

The interstratified, 10 to 15 cm thick, low angle parallel laminated sandstone beds are interpreted to represent storm-generated HCS and SCS (Harms *et al.*, 1975; Leckie and Walker, 1982; Duke *et al.*, 1991). The presence of these storm beds, although in lower abundance to those within FA1, indicates that storm events constantly influenced sediment deposition in these river-dominated deltas and suggests that the storm climate of the basin was similar during the deposition of FA1 and FA5. Overlying the storm beds are climbing ripple wave- and current ripples indicating that although fluvial processes dominated during deposition, oscillatory currents from waves were still active. The preservation of the current-generated structures is attributed to the high sedimentation rates and rapid burial preventing wave reworking. The convolute silty sandstones and mudstones and low parallel laminated sandstones comprising the distal delta front are typically devoid of bioturbation. The absence of bioturbation is attributed the heightened rates of sediment deposition which makes it difficult for infaunal organisms to colonize the substrate and construct permanent domiciles (Saunders, 1989; MacEachern *et al.*, 2005).

The presence of abundant hyperpycnal and hypopycnal muds within the delta front deposits of FA5 also support a river-dominated delta interpretation. Hyperpycnal and hypopycnal mud beds form due to the density contrasts between the sediment-laden river waters and the waters of the receiving basin. These sediments are deposited either by the settling of the suspended material carried out into suspension by a buoyant hypopycnal plume or by hyperpycnal density underflows generated during periods of high fluvial discharge (Wright *et al.*, 1988). The interstratified, centimeter to decimeter thick, unburrowed, faintly stratified mudstone displaying sharp tops and sharp bases are interpreted to have originated due to hypopycnal conditions. Hypopycnal flows are formed in settings where river outflow and sediment load is less dense than the saline waters in the receiving basin (Bates, 1953; Fisher *et al.*, 1969). These types of flows are most common in deltaic environments in marine basins where the buoyant freshwaters from the rivers flow on top of the denser marine seawater of the basin. Hypopycnal flows typically generate buoyant plumes of suspended fine-grained sediments, which are widely distributed and carried further into the delta front and prodelta environments. Turbulent mixing of the river and marine water results in deposition of the sediment load, with the coarser material being deposited near the river mouth.

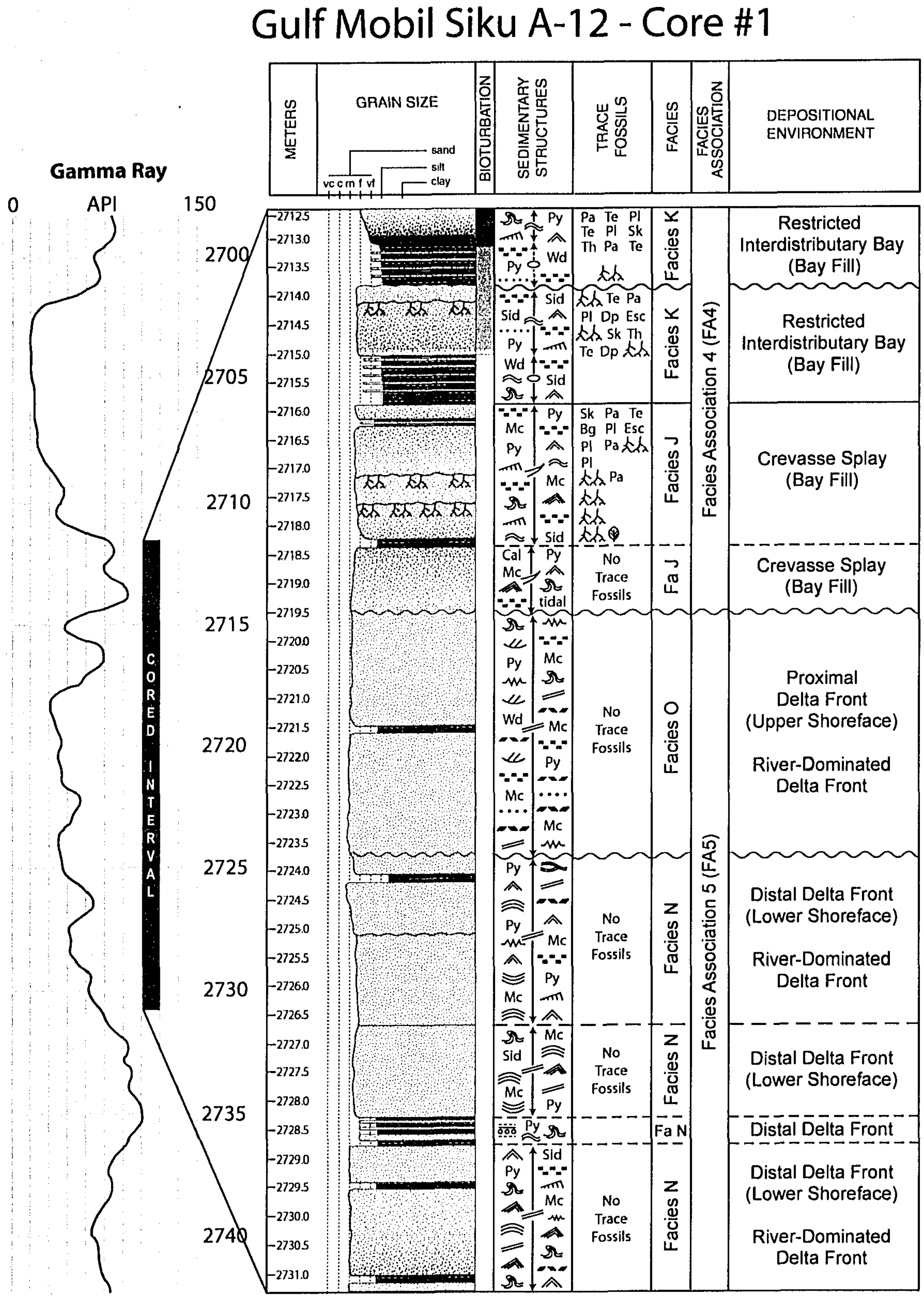


Figure 3.14 Core log of core #1 from the Siku A-12 well from the Kamik Formation, displaying grain size, lithology, sedimentary structures, ichnological structures, facies, facies associations and interpreted depositional environments.

The fine-grained sediments (silts, clays and fine-grained sands) are transported further offshore and deposited from suspension as the plume disperses (Suter, 1994). River-dominated deltas carry more finer-grained sediment in suspension, a major portion of which is deposited in the prodelta and offshore (Elliott, 1986). Within FA5 are similar interstratified, centimeter to decimeter thick, unburrowed, mudstone displaying sharp tops and sharp bases. However, these mudstone beds are differentiated from the hypopycnal mudstone beds discussed above. These mudstone beds are unique in that they do not display any internal stratification; therefore they are interpreted to originate from hyperpycnal processes in the delta front. Hyperpycnal conditions are developed where river outflow is denser than the waters in the receiving basin (Bates, 1953; Fisher *et al.*, 1969). These type of flows are generated at the river mouths in marine basins during times of high fluvial discharge, typically during or closely following storm events, where sediment-laden river outflows enter the delta front and flow down along the bottom of the delta front slope as density underflows (Wright *et al.*, 1988). These density underflows are responsible for depositing the unburrowed, highly organic sharp-topped and based hyperpycnal mudstone beds which typically mantle the underlying sandstone beds. Hyperpycnal flow is often associated with coarser-grained delta systems characterized by flash floods and sediment gravity flows (Orton and Reading, 1993) conditions, which can be enhanced in fine-grained delta systems during large storm events.

The distal delta front deposits of Facies N are abruptly overlain by the proximal delta front deposits of Facies O. Similar to the upper shoreface, the proximal delta front is a high-energy zone that is situated well above fairweather (minimum) wave base and is affected by wave- and storm-driven currents paralleling the shoreline (i.e. longshore drift). The proximal delta front sandstones of Facies O are generally well-sorted, medium-grained sandstones displaying trough and planar cross-bedding, current ripples and abundant disseminated organic material. These sedimentary structures indicate a higher-energy setting characterized by suspension, traction and gravity induced processes of deposition (Moslow and Pemberton, 1988). There is a general coarsening-upwards grain size trend with trough cross-stratification becoming dominant towards the upper portion of these successions. This facies association shares similar characteristics identified in the proximal delta front of FA1; however, there are several subtle features that differ from equivalent deposits in FA1. The presence of current ripple laminae within these sediments indicates that unidirectional current flow was dominant (Harms *et al.*, 1982). The origin of these unidirectional structures is attributed to proximity to fluvial input from nearby distributary channels. The preservation of the current ripple structures and the absence of wave-generated structures within these sediments is attributed to rapid sediment deposition within the river-dominated delta front, where sediment input from the distributary channels is able to overpower the wave-processes at the delta front. In the Siku A-12 core, the river-dominated delta front deposits of FA5 are overlain by the fine-grained, crevasse splay and interdistributary bay deposits of FA4 that are interpreted to represent delta lobe abandonment and subsequent sediment infilling of an interdistributary bay environment. Delta lobe switching is more common in river-dominated deltas than their wave-dominated counterparts due to the existence of a single, favoured distributary channel path that favours delta lobe switching (Suter, 1994).

The sandstones of Facies O contain a relatively high abundance of finely comminuted phytodetrital and micaeous material, which commonly accentuates the bedding structures. This organic material is commonly found concentrated in the troughs of bedding structures with a few beds comprised entirely of organic detritus. Coleman and Prior (1980) recognized a similar abundance of organic material in the proximal portions of river-dominated delta fronts due to the accumulations of large amounts of river transported organic debris. This organic material or "coffee grounds" originates from phytodetrital pulses, water-saturated logs and other organic debris being transported down the distributary channels during periods of storm-induced flooding. This material is eventually discharged from the distributary channels into the delta front as dense plumes of suspended sediment where wave action grinds the coarser wood particles into large concentrations of organic debris (Coleman and Prior, 1980). In river-dominated deltas the abundance of this macerated organic detritus and mica content can be extremely high and in many cases can form a significant number of the laminations within these deposits (Coleman and Prior, 1980).

The paucity of bioturbation in this river-dominated deltaic succession is attributed to the multitude of biological stresses associated with deltaic environments (Moslow and Pemberton, 1988; Gingras *et al.*, 1998; MacEachern *et al.*, 2005). As previously mentioned in the discussion of FA1, deltaic environments, especially river-dominated deltas represent one of the most biologically stressed and inhospitable environments for burrowing organisms (S.G. Pemberton pers. comm., 2004). The fluvial-derived stresses in river-dominated deltas are more profound than in wave-dominated deltas and can result in extremely impoverished ichnological assemblages characterized by very low diversities to a complete absence of bioturbation. The stresses are similar to those identified for FA1 but are much more abundant. Specific stresses include: heightened fluvial discharge, persistent high sedimentations rates, salinity fluctuations, substrate instability, hypopycnal-induced water turbidity, distributary flood and storm discharges with accompanying phytodetrital (comminuted plant debris) pulses, hyperpycnal-induced sediment gravity flows, fluid mud deposition, rapid influxes of sediment and freshwater near the bed, low oxygen levels, light and temperature variations, all of which result in environments that are inhospitable for burrowing organisms (Moslow and Pemberton, 1988; Gingras *et al.*, 1998; MacEachern *et al.*, 2005). The high amounts of river discharge associated with river-dominated deltas is a major stress for infaunal organisms in that it increases water turbidity making suspension-feeding difficult or impossible. Another stress associated with river-dominated deltas is increased discharge of large amounts of organic detritus associated with high rates of precipitation during that accompany storm events. These storm events transport large amounts of organic clay and organic detritus into the basin, which results in the rapid oxidation and concomitant oxygen at the sea floor (Leithold, 1989). This diminishes the ability of infauna to colonize the bed (Leithold, 1989; Raychaudhuri and Pemberton, 1992; Coates and MacEachern, 1999). Each of these individual ecological stresses can result in a stressed ichnological assemblage, however the combination of a number of these stresses in a deltaic environment can result in very low diversities or a total absence of burrows (MacEachern and Pemberton, 1992; Pemberton *et al.*, 2002). For a more detailed discussion of the ichnology of river-dominated deltas the reader is referred to MacEachern *et al.* (2005).

FA5 displays many features that support the interpretation of a moderately storm-dominated, river-dominated delta. In contrast to the deltaic deposits of FA1 there is abundant evidence of high rates of fluvial input and little evidence of strong wave reworking and winnowing by wave- and storm-energy. The presence of abundant soft-sediment deformation structures and climbing current ripple lamination reflect high sedimentation rates. Abundant evidence of unidirectional current processes, hypopycnal and hyperpycnal mudstone beds and large quantities of organic detritus further supports this interpretation. The paucity of trace fossils within these sediments indicates that during deposition of FA5 conditions were highly stressful and therefore impeded infaunal burrowing. This is attributed to the profound fluvial-derived stresses associated with river-dominated deltaic environments. FA5 is the least common facies association identified from the Kamik Formation and indicates that although the majority of deltas were wave-dominated some of the Kamik deltas were also river-dominated. This succession is similar to the ancient examples of river-dominated deltas including the Carboniferous of Western Ireland (Pulham, 1989), the Ferron Sandstone of Utah (Cotter, 1975) and Upper Cretaceous Dunvegan Formation (Bhattacharya and Walker, 1992) and the modern examples such as the Balize and Atchafalaya lobes of the Mississippi delta (Coleman and Prior, 1982; Coleman, 1988; Elliott, 1986; Coleman *et al.*, 1998). The details of modern and ancient river-dominated deltas are discussed further in section 3.4.3 and 3.4.4.

3.3 Summary of Facies Associations

The Kamik Formation was grouped into five facies associations (FA1 to FA5) based upon physical and biological characteristics observed in core. These five facies associations identified in this study include: FA1: moderately storm-influenced, wave-dominated delta; FA2: meandering distributary channel; FA3: braided distributary channel; FA4: lower delta plain; and FA5: moderately storm-influenced, river-dominated delta. The majority of the deltaic sediments comprising the Kamik Formation are interpreted to represent deposition in a moderately storm-influenced, wave-dominated deltaic environment (FA1); however, the deposits of a river-dominated deltaic environment (FA5) were also identified. Discrimination between the two types of delta systems identified within this study was based upon the detailed sedimentological and ichnological criteria (see section 3.2). The predominance of the wave-dominated deltaic deposits of FA1, in comparison to the river-dominated deltaic deposits of FA5 indicates that deposition of the upper Kamik Formation the Brooks-Mackenzie Basin was dominated by wave and storm processes. The interpreted morphologies of the deltas of FA1 and FA5 are plotted on the tripartite deltaic

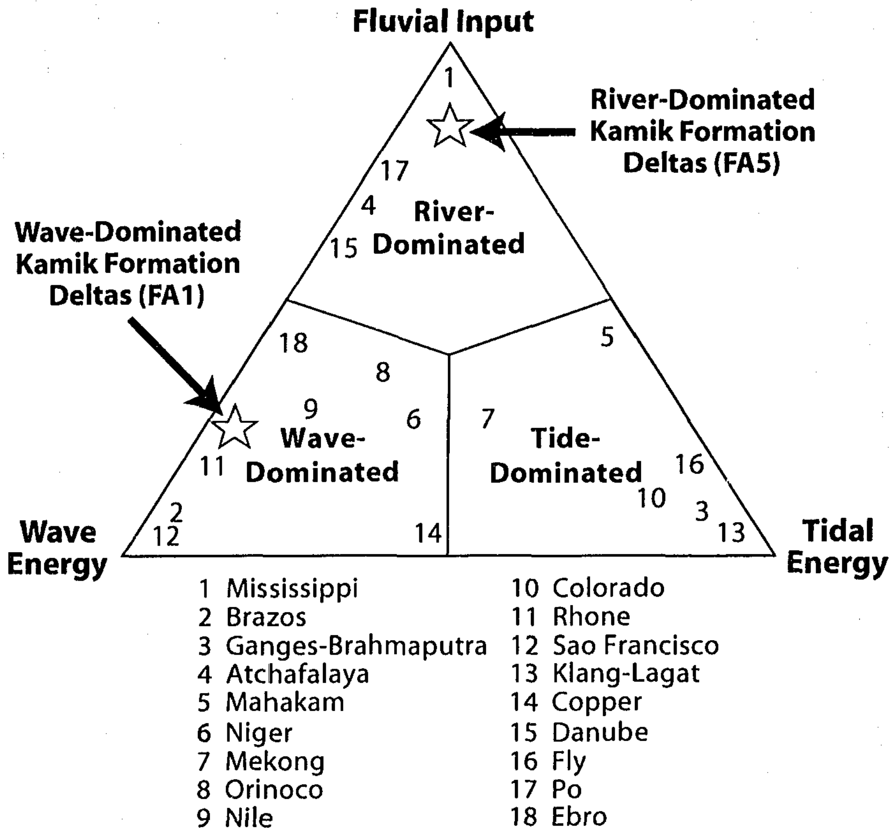


Figure 3.15 Process-based, tripartite classification of deltas into wave-, river- and tide-dominated. The positions of specific deltas are plotted on the basis of general delta-front morphology, which reflects the dominant processes acting upon the system. Note the interpreted position of Kamik Formation deltas on the classification scheme (modified from Galloway, 1975).

classification scheme of Galloway (1975) (Figure 3.15). The position of wave-dominated deltas of FA1 and the river-dominated deltas of FA5 is based on the inferred delta front morphology, which reflects the dominant processes acting upon the system. The presence of the deposits of river- and wave-dominated deltas within the Kamik Formation indicates conditions existed that favoured the development of deltas with both morphologies. The origins of these two types of delta morphologies are discussed in section 3.6.

3.4 Discussion of Wave- and River-Dominated Deltas

The definition of a delta as used in this study is "a discrete shoreline protuberance formed where rivers enter oceans, semi-closed seas, lakes or lagoons and supply sediment more rapidly than it can be redistributed by basinal processes" (Elliott, 1986, p. 113). Deltas form where sediment-laden distributary channels suddenly expand and decelerate on entering a depositional basin. As a result, the sediment load is dispersed and deposited, with coarse-grained bedload accumulating near the river mouth, whilst finer-grained, suspended sediment is transported offshore and deposited in

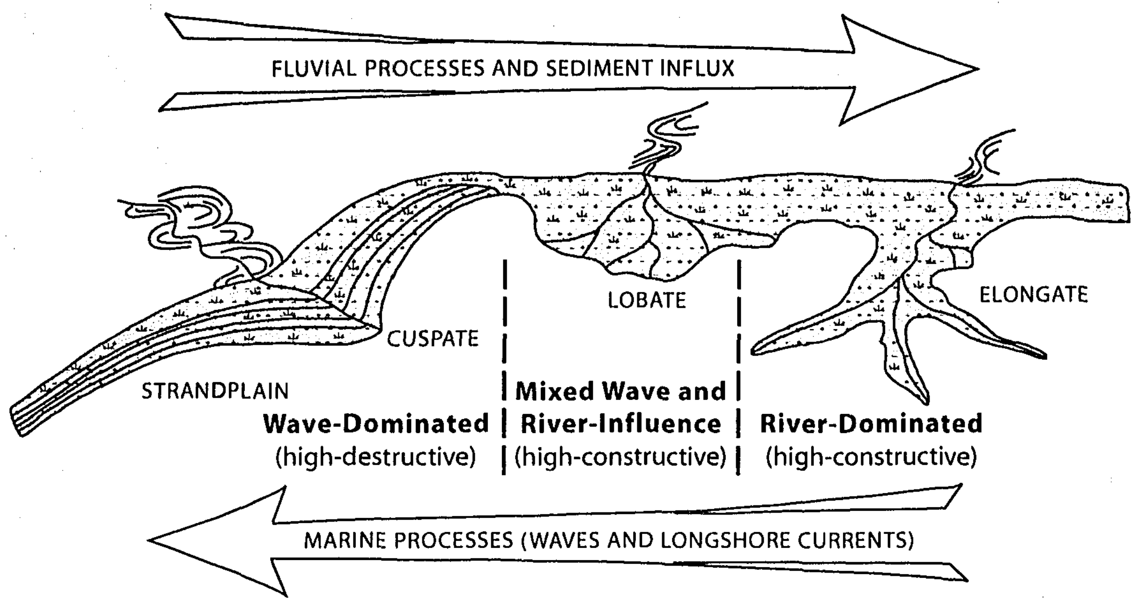


Figure 3.16 Comparative shoreline configurations of wave-dominated and river-dominated deltaic systems (modified from Fisher *et al.,* 1969).

deeper water (Elliott, 1996). The physical morphologies and distribution of sediment in deltas is largely controlled by the interaction of dynamic physical processes in the depositional basin. The relative dominance of fluvial processes, sediment input and marine processes, such as wave-energy, tidal range, oceanic currents all work in combination to modify and disperse the riverborne clastics (Fig. 3.16) (Wright and Coleman, 1973; Galloway, 1975; Coleman and Prior, 1980). Additional factors affect the form and type deltas including nature of the fluvial system, relative density of the river inflow with respect to the waters in the receiving basin, sediment sizes and compositions, shelf gradient, nature of the drainage basin and climate influence the nature and distribution of facies in deltaic environments (Galloway, 1975; Coleman and Wright, 1975; Wright, 1985; Orton and Reading, 1993). On a larger scale, the deltaic shape is influenced by the size and morphology of the receiving basin, tectonic setting, rates of relative sea level rise and subsidence rates (Bhattacharya and Walker, 1992; Suter, 1994). Evolutionary factors such as autocyclic delta switching or eustatic fluctuations are also important influences on the expressions of these delta systems (Penland *et al.*, 1988).

Marine processes affecting delta systems are strongly influenced by basin configuration and tectonics. Rivers entering open ocean settings are subject to the full effects of basinal processes, including waves, tides, storms, and semi-permanent ocean currents. Wave-dominated deltas are commonly developed in open-ocean basins, which are generally deeper and have greater wave fetch, therefore increasing the potential for greater wave-energy (Suter, 1994). River-dominated deltas, such as the modern Mississippi delta are commonly developed in tectonically inactive basins with limited marine energy (Suter, 1994). The majority of deltas are not dominated solely by wave, river or tidal input, rather their morphologies are a result of a continuous spectrum of basinal processes (i.e. river sediment input, wave-energy and tidal flux), which rework and redistribute the deltaic sediments. Most deltaic coasts receive different rates of sediment supply and it is not unusual for one part of the deltaic shoreline to rapidly prograde seaward, while other parts are being transgressed and reworked by marine processes (Coleman and Prior, 1982). The evolution of deltas is controlled by a balance between accumulation of fluvially supplied sediment, reworking, erosion and transport of deltaic sediments by marine processes (Wright, 1985). Therefore, the overall dynamic and complex depositional nature of modern and ancient deltas requires a detailed analysis of the sedimentology, ichnological and stratigraphic characteristics throughout its evolution to understand the dominant processes influencing deposition. These factors have been discussed by numerous authors including Galloway (1975), Coleman and Wright (1975), Coleman and Prior (1980, 1982), Bhattacharya and Walker (1992), Orton and Reading (1993) and Suter (1994).

3.4.1 Discussion of Wave-Dominated Deltas

Wave-dominated deltas are developed along coastlines with a strong wave climate, comparatively weak fluvial output and tidal energy and appreciable amounts of clastic sediment input (Heward, 1981; Davis and Hayes, 1984). These types of deltas were originally termed by Fisher *et al.* (1969) as "highly-destructive" (Fig. 3.16). The deposits are generally sand-dominated and display abundant evidence of reworking by basinal wave processes and limited evidence for discrete sediment point sources of sediment input (Pemberton *et al.*, 2001). These deltas are characterized by broad, sand-rich lobes with arcuate to cuspate geometries and exhibit straight, smooth coastlines (Bhattacharya and Walker, 1992; Bhattacharya and Giosan, 2003). Deposits of wave- and storm-dominated deltas are characterized by relatively thick, continuous coarsening-upwards delta front successions comprised of large quantities of sand. The large amounts of sediment comprising wave-dominated deltas originate from the influx of sands, silts and clays from one or more distributary channels. Strong wave- and storm-energy in the basin results in the rapid diffusion and deceleration of the river outflow and deposition of sediments at the delta front as distributary mouth bars, which are reworked by wave action and redistributed along the deltaic coastline by strong longshore currents (Coleman and Prior, 1982). The redistribution and reworking of sediment results in the deposition of the delta front sand bodies as a series of parallel shorelines with well-developed delta front beaches, shore-parallel beach-ridge complexes, barrier bars and sand spits centered around the mouths of distributary channels (Wright and Coleman, 1973; Coleman and Wright, 1975; Davis and Hayes, 1984; Bhattacharya and Walker, 1992). The overall morphology of wave-dominated deltas reflects a balance between sediment supply from distributary channels and the rate of wave reworking and redistribution of sediments. Due to the constant reworking of sediment by waves and storms, wave-dominated deltas display sedimentological and ichnological characteristics which make them difficult to distinguish from prograding, wave-dominated strandplain system.

The persistence of high wave-energy flux in these systems results in delta front deposits that are typically well-sorted due to continuous reworking by wave action and longshore drift (Coleman and Wright, 1975). Wave action winnows and carries the finer-grained sediments seaward in suspension, where it is deposited in the prodelta and distal delta front or accumulates in the interdistributary bays and lagoons behind the shore parallel beach and beach ridges (Coleman and Prior, 1980). Ancient wave-dominated deltas are characterized by abundant wave-formed structures and pervasive storm-generated HCS and SCS bedforms. The abundance of storm deposits depends upon the degree of storm-influence on the depositional system; whether it is low, moderate to highly storm-dominated (MacEachern and Pemberton, 1992). The presence of abundant wave-generated structures, minor soft-sediment deformation structures, indicates rapid sediment deposition, a feature characteristic of wave-dominated deltas (Curray *et al.,* 1969; Coleman and Wright, 1975; Heward, 1981). These deltaic systems are characterized by stressed ichnological assemblages with lower diversities and abundance of ichnofossils compared to non-deltaic wave-dominated shorefaces (Moslow and Pemberton, 1988; Gingras *et al.*, 1998; MacEachern *et al.*, 2005). The majority of the deltaic sediments of the Kamik Formation are well-sorted and are characterized by an abundance of wave-formed sedimentary structures and minor soft-sediment deformation structures. These features combined with the thick coarsening-upwards and/or sandier-upwards nature of FA1 successions support the interpretation of a wave-dominated deltaic environment.

The Kamik Formation was deposited in the actively rifting Brooks-Mackenzie Basin. This extensional basin was bound along its southeast margin by the Eskimo Lakes Fault Zone and the uplifted margin of the Eskimo Lakes Arch (see Chapter 1, section 1.10). The Kamik paleo-shoreline is interpreted to have been oriented approximately southwest to northeast, parallel to the trend of the Eskimo Lakes Fault Zone and the uplifted margin of the Eskimo Lakes Arch (Dixon, 1982a, 1991). It has been recognized that wave-dominated deltas are commonly occupy structural lows such as the down-thrown sides of growth faults (Heward, 1981). During deposition of the Kamik Formation extension in the Brooks-Mackenzie Basin resulted in the movement of normal faults in the Eskimo Lakes Fault Zone and uplift of the Eskimo Lakes Arch. This structural activity created a structural low on the down-thrown side of the Eskimo Lakes Fault Zone and resulted in the development of topographic highlands to the southeast. The tectonic uplift proximal to the study area resulted in the supply of large quantities of quartz-rich sediments to the actively prograding wave- and river-dominated delta shorelines of FA1 and FA5. The basin configuration resulted in the development of a strongly storm-influenced deltaic shoreline that was exposed to the north and northwest to the open ocean basin. The morphology of the basin did not provide shelter from storms or from the development of large waves and facilitated the deposition of storm-generated bedforms (i.e. HCS and SCS) and the development of wave-dominated deltas.

Modern examples of wave-dominated deltas and strandplain systems include the Rhône delta of southern France (Oomkens, 1967, 1970), the Saõ Francisco river delta of Brazil (Coleman and Wright, 1975; Coleman, 1981), the Brazos Delta of east Texas (Bernard *et al.*, 1970; Rodriguez *et* *al.*, 2000), the Nile delta of Egypt (Coleman and Wright, 1975; Coleman, 1981) and the Costa de Nayarit from the west coast of Mexico (Curray and Moore, 1964). These wave-dominated deltas and strandplain systems can extend hundreds of kilometers alongshore, and are appreciably larger when fed by multiple river systems. The Saõ Francisco delta and Coast of Nayarit strandplain are characterized by long, linear beach and beach-ridge complexes of continuous progradational shoreface sands (Curray and Moore, 1964; Coleman and Wright, 1975; Coleman, 1981). These environments are characterized by abundant sediment, which can lead to rapidly prograding sheet sands, some of which have been documented to prograde one to three kilometers in 100 years (van Straatan, 1960). This is illustrated by the Coast of Nayarit, which has prograded 10 to 15 km over a distance of 225 km since the stabilization of sea level after the Holocene transgression (Curray *et al.*, 1969). There is debate as to the origin of many wave-dominated deltas. Many of these 'deltas' are actually composite features formed by longshore drift of sediments from other sources and should be considered as prograding strandplains (Curray *et al.*, 1969; Dominguez and Wanless, 1991; Boyd *et al.*, 1992). The distinguishing feature between deltas and strandplains is that a true delta has a fluvial point source and a coeval delta plain, whereas, strandplains are strike-fed depositional systems which are welded onto a pre-existing coastal plain (Boyd *et al.*, 1992; Suter, 1994). For a more thorough discussion of wave-dominated deltas see papers by Coleman and Wright (1973), Galloway (1975), Coleman and Prior (1980, 1982), Wright (1985), Elliott (1986), Bhattacharya and Walker (1992) and Suter (1994).

3.4.2 Rhône Delta - Modern Analogue

The Rhône delta of southern France is interpreted to represent a modern analogue for the wave-dominated deltas of FA1. The Rhône delta is classified as a "highly-destructive" wave-dominated delta system that displays a lobate morphology (Fig. 3.17) (Fisher *et al.*, 1969; Galloway, 1975). The lobate morphology of the this delta is the result of the mixed influence of river input from the Rhône River and moderate wave-energy, but minimal tidal processes in the Mediterranean Sea (Kruit, 1955; Lagaaij and Kopstein, 1964; van Straaten, 1959; Oomkens, 1967, 1970). The tides in the northwestern part of the Mediterranean Sea are weak and the average tidal range in the Rhône delta is approximately 0.20 m (Oomkens, 1970). The tidal currents do not significantly influence sediment deposition on the Rhône deltaic shorelines (van Straaten, 1959; Oomkens, 1970). Wave action reworks the delta front sediments and longshore drift transports sediments westward along the coast where it is subsequently redeposited in broad, sandy beach ridges and interdeltaic strandplains. The Rhône delta is subject to intense marine processes, and therefore displays fewer distributary channels and river-built features such as bays, marshes and crevasse splays.

Geographically, the Rhône delta occurs in the northwest portion of the Mediterranean Sea and covers a large part of the shore of the Gulf of Lions and drains an area of approximately 95,830 km^2 of which 2590 km^2 are subaerial delta plain deposits (Oomkens, 1967, 1970). The Rhône

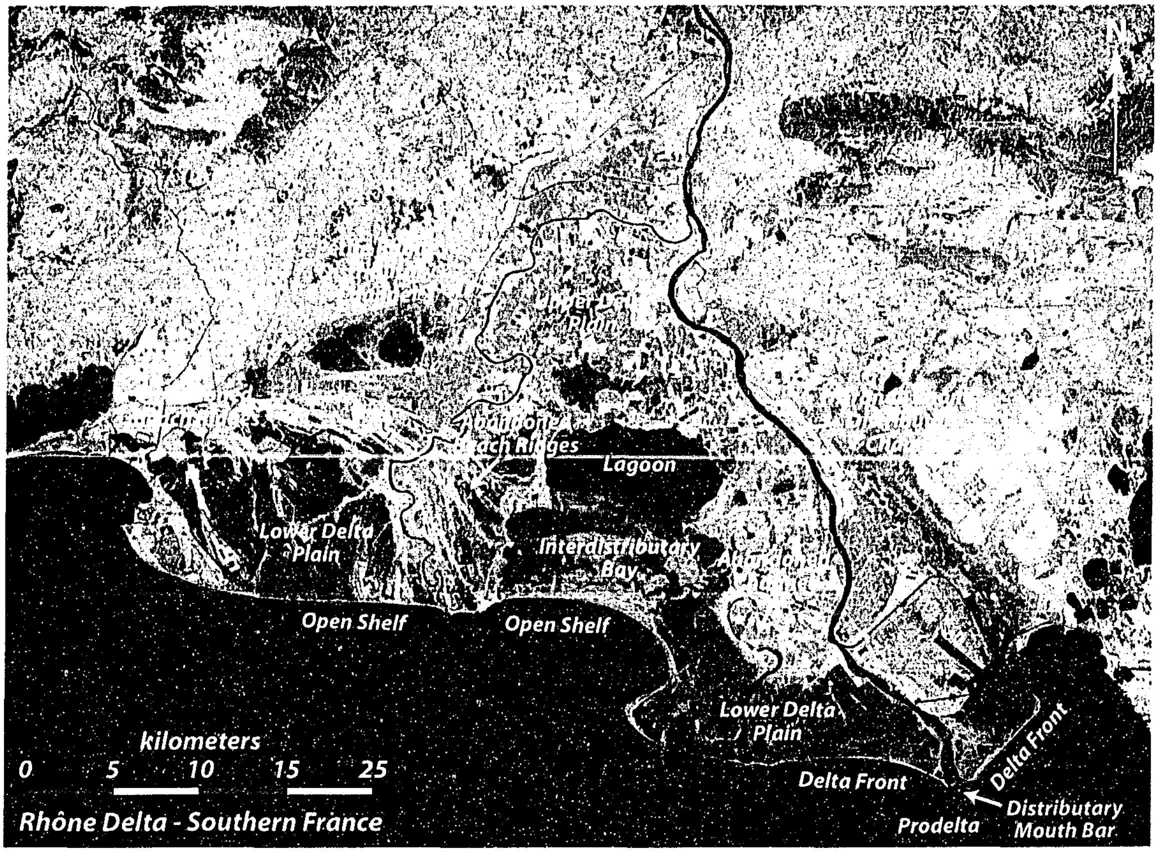


Figure 3.17 Modern wave-dominated Rhône Delta of southern France. This delta is interpreted to represent a modern analogue for the upper Kamik Formation, especially the deposits of FA1. The delta contains the two main delta lobes fed by the main distributary channels. The eastern Grande Rhône is the most active, prograding delta lobe and the western Petit Rhône is currently under transgression. Note the contrast between the constructional (active) and the transgressive destructional (non-deltaic) stages (modified from NASA's https://zulu.ssc.nasa.gov/mrsid/mrsid.pl website, 2005).

delta is composed of two main distributary channels, which form two delta lobes, the western Petit Rhône, which is currently undergoing transgression, and the Grande Rhône that is actively prograding across the continental shelf. The Grande Rhône river is the main distributary of the Rhône delta and provides the main supply of sediment to the actively prograding delta front (van Straaten, 1959). The supply of sediment provided by the Grande Rhône River has resulted in the development of a small distributary mouth bar in the immediate vicinity of the river mouth (Fig. 3.17). Progradation of the delta front is most pronounced in the vicinity of the Grande Rhône distributary channel where the distributary mouth bar deposits are accreted onto the shoreline as a series of beach ridges. In the inactive parts of the delta, the abandoned delta lobes are reworked by wave action and are actively retreating landwards (van Straaten, 1960b).

The Rhône delta is subdivided into two main geomorphologic elements, 1) the northern upper delta plain and 2) the southern lower delta plain, delta front and prodelta. The upper delta plain is composed mainly of active and abandoned fluvial channels, point bars, natural levees, crevasses, dunes and coastal plain basins (Gensous *et al.*, 1993). The distributary channels of the Rhône river system meander significantly and form meander belts throughout the delta plain. The southern portion of the delta includes the lower delta plain and delta front and is comprised of active and abandoned distributary channels, point bars, natural levees, crevasses, dunes, interdistributary bays, lagoons, lakes, marshes and an extensive pattern of both active and abandoned beach ridges. The Rhône delta is typified by interdistributary bays, saltwater to brackish-water swamps, fresh and saltwater marshes and lakes, crevasse splays and fluvial channels (van Straaten, 1960b). The shoreline is smooth with common wave-dominated beaches, coastal barriers, spits and beach ridges, particularly in proximity to the abandoned river mouths. Riverine-dominance is reflected by the pronounced lobate form with interdistributary bays and marshes behind the sandy beaches ridge complexes of the delta shoreline.

The delta front at the mouth of the Grande Rhône River is comprised of laterally extensive sandy beach ridges that form thick, coarsening-upwards successions (Oomkens, 1967; Gensous *et al.*, 1993). These delta front sediments consist of well-sorted sands that display a gradual seaward decrease in sand content and grain size. These deposits grade seaward into interbedded sands, silts and clays of the prodelta. The majority of the delta front deposits were deposited during storms when river flooding and currents are much stronger than during normal conditions. Sedimentation rates in the delta front in front of the main river mouth are about 35 cm per year, with the average annual advance of the deltaic coastline at the mouth of the Grande Rhône approximately 30 m per year (van Straaten, 1959; Oomkens, 1970). Strong wave action and storm energy on the active Rhône delta lobe resulted in the longshore transport of sand and the development of sand spits and shoreline-parallel beach ridges adjacent to the river mouth (van Straaten, 1959). Sediment deposition and beach accretion along the coastline of the Rhône delta is mainly restricted to the immediate vicinity of river mouths. The entire coastline of the Rhône delta is characterized by subaqueous sandy longshore bars that are approximately parallel, or at a slight angle, to the shore. Some of the delta front sediments of the Rhône delta are derived by longshore drift from adjacent strandplains; however, the majority of the delta front sediments are still derived from distributary channels (Oomkens, 1970). Sediments comprising the abandoned lobes of the Rhône delta are reworked by waves and the eroded sand is moved toward the shore and transported along shore by longshore drift. The morphology of the delta is largely controlled by successive changes in courses of the Rhône River and by westward-directed longshore transport of sediments. The shifting and lateral migration of the main distributary channels has resulted in the accumulation of deltaic deposits along some 50 km of the shoreline (Oomkens, 1970). In the interdeltaic areas, where the distributaries have been abandoned and sediment is not supplied, the shoreline is retreating as a result of transgression and erosion. The Rhône delta has been actively prograding southward over the last 5000 years and overlaps numerous older submerged abandoned lobes and forms a thick sediment lens of imbricated deltaic deposits (Oomkens, 1970).

Studies of the distribution and abundance of infaunal and microfauna assemblages in the Rhône delta identified a decrease in the abundance and diversity of organisms proximal to the delta front. This is similar to the overall decrease in diversity and abundance of the stressed ichnological assemblages of FA1. A detailed analysis of the distribution of mollusc shell assemblages in the delta front by van Straatan (1960b) identified a number of controlling factors including variations in salinity, bottom texture, water depth, sedimentation rates and turbidity. It was found that the diversity and abundance of the mollusc assemblages decreased with increasing proximity to the distributary channels mouths. Oomkens (1970) presented a similar study where the microfauna content was used to distinguish between deltaic sequences produced by direct fluvial input of sediment near the active distributary mouth, and marine sequences in which the sediment was supplied by longshore drift from the distributary mouth. It was also found that the sediments deposited proximal to the mouth of the Grande Rhône distributary channel were characterized by a paucity or absence of benthonic fauna, both in the number of specimens and species diversity compared to contemporaneous marine sequences deposited away from the distributary channel mouths (Lagaaij and Kopstein, 1964; Lagaaij and Gauthier, 1965). The paucity or absence of organisms in these environments was attributed to the adverse effects resulting from the salinity variations of the shallow waters near the river mouth, the high rates of sediment deposition, sediment instability and high water turbidity (i.e. suspended clays and silts) which resulted in an increase in bottom darkness and made it all but impossible for the hardiest forms to survive (van Straaten, 1960b; Oomkens, 1970). The major biological stress identified in the Rhône delta was the high sedimentation rates at the mouth of the Grande Rhône distributary channel (van Straatan, 1960b, Lagaaij and Kopstein, 1964). The origin of these stressed biological assemblages in both settings is attributed to same the fluvial-derived stresses present in deltaic environments (i.e. salinity variations, high sedimentation rates and water turbidity) and further supports a deltaic interpretation for FA1.

The morphology of the modern wave-dominated Rhône delta is illustrated in Figure 3.17. The geometry of the delta, the dominance of wave and storm-energy and minimal tidal influence on the delta front, strong littoral drift and geomorphological characteristics are all similar features inferred for the wave-dominated deltas of FA1. The dominance of wave energy on the delta front of the Rhône delta resulted in the reworking of sediments into well-sorted, coarsening-upwards successions. The delta front sediments of the Rhône also contain a similar abundance of organic detritus as recognized in the deposits of FA1 (Oomkens, 1970). The lower diversities and abundances of trace fossils found in FA1 are interpreted to have resulted from the same adverse effects (i.e. salinity variations, rapid sediment deposition and water turbidity) that acted upon the microfauna in the Rhône delta front. The presence of delta plain deposits (FA4) including distributary channels, interdistributary bay, crevasse splay and marsh deposits that are interstratified with FA1 are features that are also present in the modern wave-dominated Rhône delta. It has been recognized that the more wave-dominated the delta, the greater the proportion of lobe sediment with limited amounts of interdistributary bay and distributary channel facies (Bhattacharya and Walker, 1991). Interdistributary bay deposits are poorly developed in the modern, strongly wave-dominated Saõ Francisco delta. This delta is characterized by strong littoral drift, limited development of interdistributary bays and lack of bifurcating distributary channels (Coleman, 1976; Dominguez, 1996). In comparison, the moderately wave-dominated Rhône delta is characterized by the development of multiple distributary channels, interdistributary bays and crevasse splay deposits. Therefore, the presence of abundant interdistributary bay, marsh and crevasse splay deposits of FA4 interstratified with the deltaic deposits of FA1 supports the interpretation that the deltas responsible for FA1 more closely resembled the modern Rhône delta than the more wave-dominated Saõ Francisco delta.

There are some significant differences between the depositional setting of the modern Rhône delta and the Kamik deltas responsible for FA1: 1) the basin configuration interpreted for the Kamik Formation was that of an pericratonic basin exposed to the wave energy of the open ocean. In comparison, the Rhône delta is located within the intercontinental Mediterranean basin, a setting with much lower wave energy and storm frequency than what was present during deposition of the Kamik Formation, 2) the Rhône delta is located in a humid, subtropical climate; whereas, the Kamik Formation was deposited in a northern region with an arid to temperate northern climate. This difference in climatic setting would have resulted in the variations in the development of vegetative material and fluvial system and would have influenced the geomorphology of the delta plain. Aside from these differences, it is interpreted that the modern wave-dominated Rhône delta represents an appropriate analogue for the deltas of FA1.

3.4.3 Discussion of River-Dominated Deltas

River-dominated deltas form in tectonically inactive basins where fluvial processes are able to overwhelm the reworking capabilities of basinal processes, such as wave and tidal-energy (Galloway, 1975; Coleman and Prior, 1980, 1982; Suter, 1994). Fluvial-dominated or "highly-constructive" deltas generally develop as a series of prograding, branching distributary channels that form elongate "birdsfoot" or lobate morphologies (Fisher *et al.*, 1969; Coleman and Prior, 1982; Wright, 1985) (Fig. 3.16). River-dominated deltas are characterized by highly irregular and protruding shorelines, a sparsity of wave formed features, and a low lateral continuity of sands (Wright and Coleman, 1973). The variability between the birdsfoot and lobate morphologies in fluvial-dominated deltas is due to a combination of water depth and sand content. Sandier, lobate deltas tend to form in shallow water settings and elongate, birdsfoot deltas are common in deeper water settings (Pemberton *et al.*, 2001). River-dominated delta fronts are characterized by deposition of fine-grained, mixed-load deposits deposited within the active lobes immediate to the distributary channel. These active lobes are comprised of a series of distributary mouth bars, which form at the seaward terminus of a distributary channel (Coleman and Prior, 1980). The limited basinal reworking of these fluvial supplied sediments in the marine environment results in the development of very straight distributary mouth bars. Sand is deposited in the delta front in bar fingers, channel mouth bars and within the distributary channels with silts and clays deposited in the prodelta, interdistributary bays and delta plain. The distributary mouth bar represents the portion of the delta environment with the highest sediment accumulation rates. For example, depositional rates of coarse sediments at the mouth of the modern Mississippi delta are as high as 4 to 5 m per year (Coleman and Prior, 1980).

Deposits of river-dominated delta lobes are characterized by a coarsening-upwards and shallowing-upwards succession comprised of thin sand laminations, which become thicker upward and generally have a limited lateral continuity (Coleman, 1981). The sediments of the distributary mouth bar are commonly clean, well-sorted sandstones due to the reworking by river currents, waves and tides generated in the open marine realm. These successions commonly display sedimentary structures that reflect deposition dominated by fluvial-generated unidirectional currents, abundant evidence of rapid, unstable deposition and low diversity ichnological assemblages (Pemberton *et al.*, 2001). The most common sedimentary structures include current ripples, trough and planar tabular cross-bedding and climbing ripples. High sedimentation rates and the subdued nature of wave action and tidal processes in the delta front leads to mass movement processes such as submarine landslides and localized slumps, resulting in abundant soft-sediment deformation structures including convolute bedding, loading structures, aggrading sedimentary bedforms and large scale slumps (Prior and Coleman, 1978; Coleman and Prior, 1980, 1982; Lindsay *et al.*, 1984). In addition, these sediments also contain significant amounts of carbonaceous detritus and mica associated with discharge from distributary channels (Reading and Collinson, 1996). The best known modern example of a river-dominated delta is the modern Balize lobe of the Mississippi delta (Coleman and Prior, 1982). For a more thorough discussion of river-dominated deltas see papers by Galloway (1975), Coleman and Prior (1980, 1982), Wright (1985), Bhattacharya and Walker (1992) and Suter (1994).

3.4.4 Mississippi Delta - Modern Analogue

The Mississippi River delta is interpreted to represent a modern analogue (albeit not entirely representative) for the deltas of FA5. The modern Balize lobe of the Mississippi River delta is classified as a "highly-constructive" river-dominated delta system (Fisher *et al.*, 1969). This river-dominated delta displays a digitate 'birdsfoot' morphology comprised of a series of radiating distributary channels and a highly irregular coastline (Galloway, 1975) (Fig. 3.18). The delta is dominated by fluvial processes and characterized by a series of prograding distributary channel mouths which form a skeleton of radiating narrow, elongate shore-normal finger sands (distributary

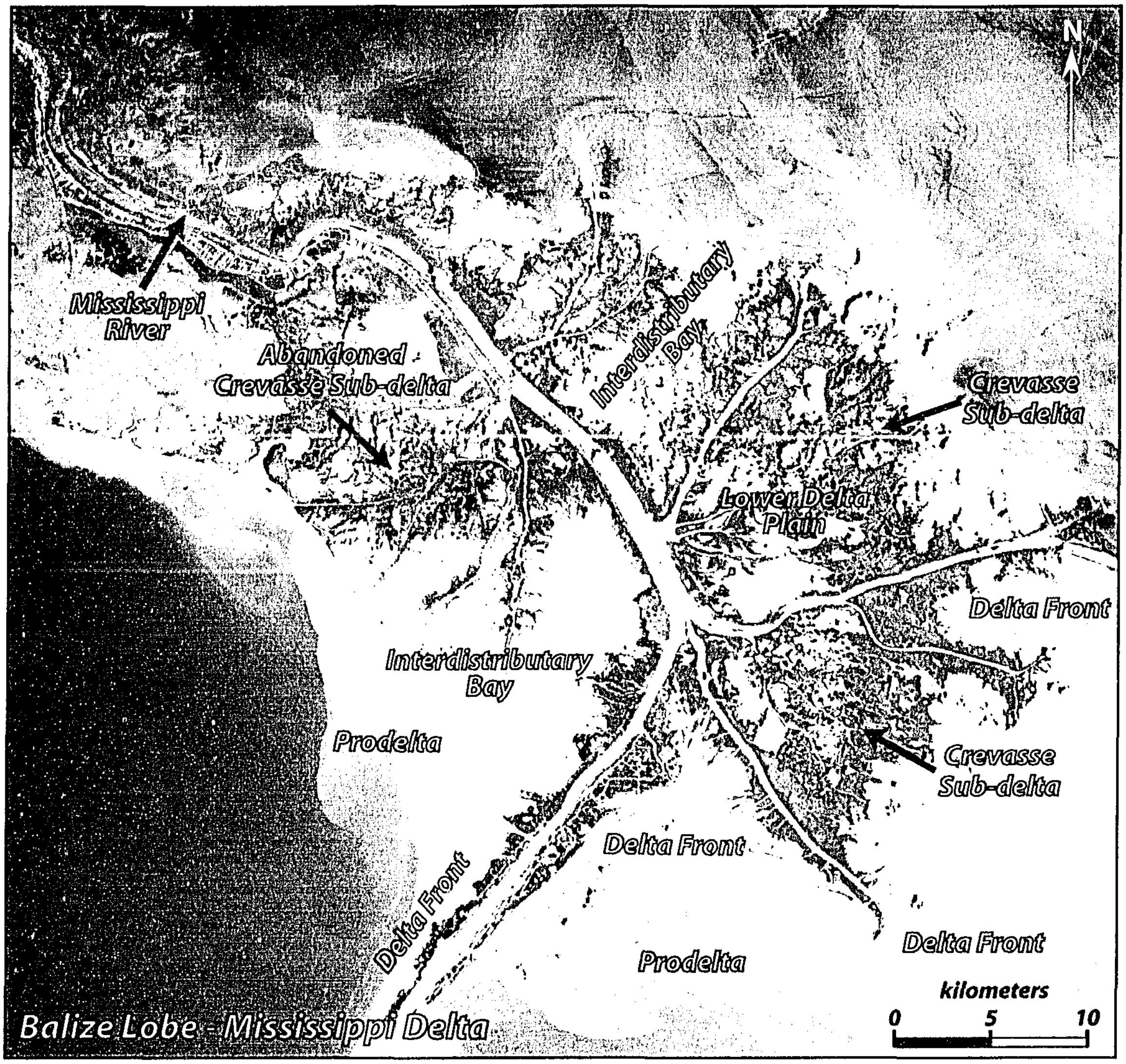


Figure 3.18 Modern river-dominated Mississippi Delta of southeast Louisiana as it enters the Gulf of Mexico. This delta is interpreted to represent a modern analogue for the deposits of FA5. Note the classic 'Birdsfoot' morphology and the active distributary channels. The image covers an area of approximately 54 kilometers by 57 kilometers (image fron NASA's Earth Observatory website http://earthobservatory.nasa.gov, 2005).

mouth bars) separated by shallow interdistributary bays (Fisk, 1961). The Mississippi delta is situated in the storm-dominated northern Gulf of Mexico where it experiences relatively low energy levels resulting from wind and wave processes, except during the occurrence of hurricanes and tropical storms (Boyd and Penland, 1988). The birdsfoot morphology and straight nature of the distributary mouth bars is due to the inability of the low wave-energy and very low tidal range in the Gulf of Mexico to significantly rework the large quantities of fine-grained sediments supplied by the Mississippi River. The birdsfoot morphology displayed by modern Balize lobe owes it unusual morphology to its position in relatively deep waters, close to the edge of the continental shelf (Boyd and Penland, 1988). This contrasts with the older lobes of the Mississippi delta which prograded across the shallow shelf as a series of 'shoal-water' lobes with lobate geometries (Frazier, 1967). These shallow water deltas were characterized by a greater degree to lobe and river-mouth switching which tends to form thinner and more lobate deltas (Bhattacharya and Walker, 1992). Since the early 1900's the Mississippi delta has been heavily modified by the actions of the United States Army Corps of Engineers. This anthropological modification has resulted in the development of a modern delta whose morphology is not truly representative of the older river-dominated deltas developed in the region. In light of this anthropological modification, the Mississippi delta still represents the best known example of modern river-dominated delta with its morphology considered to have most closely resembled the deltas responsible for FA5.

The Mississippi River delta is located in southeast Louisiana and occupies approximately 500 km of the coastline on the northern Gulf of Mexico. The Mississippi is the largest river system in North America and drains an area of approximately 3,344,560 km^2 (41% of continental United States). The delta plain has a total area of 28,568 km^2 of which 23,900 km^2 are subaerial deposits (Coleman and Prior, 1980; Wright, 1985). The annual sediment discharge of the Mississippi River at the delta front is extreme and is estimated at 6.21×10^{11} kg consisting of a high suspended sediment load of mostly clays, silts and fine-grained sands (Elliott, 1996). The overall shoreline wave power is extremely low, averaging only 0.03×10^7 ergs/second. When these sediments leave the confinement of the river mouth and discharge into the delta front they spread out for long distances. At the present time, sands are deposited near the river mouths of the two active delta complexes, the Balize and Atchafalaya. Finer-grained sediments are transported and settle farther offshore from suspension or as the result of downslope mass movement processes. Plumes of highly turbid freshwater effluent extend out considerable distances into the offshore where the fine-grained suspended sediments are deposited in the prodelta. Sedimentation rates in the delta front are extremely high, averaging 1 m per year resulting in progradation of the distributary mouths ranging from 50 to 100 m per year (Coleman and Prior, 1980). During periods of high floods, accumulations of 3 to 4 m of sediment over 2 to 4 months have been documented (Coleman and Prior, 1982). Regional subsidence rates range from 30 to 100 cm per century; however, in the delta front, where the distributary mouth bar sands prograde over the prodelta clays local subsidence rates can be as high as 200 cm per year (Coleman and Prior, 1980). High discharge rates and rapid deposition of sediments results in a variety of soft-sediment deformation and subaqueous slump features (Lindsay *et al.*, 1984; Suter, 1994). The rapid progradation of the thick delta front sandstones and subaqueous delta front leads to diapiric clay intrusions with mass wasting, slumping and abundant faulting across the shelf. These structures are developed due to the high sedimentation rates and substrate instability associated with the rapid progradation and sediment loading of distributary mouth bars over weak prodelta clays (Coleman and Prior, 1980). The presence of similar soft-sediment deformation structures in the delta front deposits of FA5 supports the interpretation that these deltas were river-dominated.

The morphology of the modern river-dominated Mississippi delta is illustrated in Figure 3.18. The geometry of the delta, the dominance of fluvial input, high sedimentation rates, limited effects of wave and storm-energy on the delta front and the pervasiveness of syndepositional deformation structures are all similar features inferred for the river-dominated deltas of FA5. The dominance of fluvial input results in the deposition of the deltaic deposits, which display ubiquitous soft-sediment deformation structures and an absence of bioturbation. Similar to the deltas of FA1, the presence of delta plain deposits (FA4) including interdistributary bay, crevasse splay and marsh deposits interstratified with FA5 are features that are characteristic of the modern river-dominated deltas. The absence of bioturbation within FA5 is attributed to the numerous river-derived, biological stresses present in fluvial-dominated deltas. Stresses such as high sedimentation rates, salinity variations, substrate instability, water turbidity, low oxygen levels, light and temperature variations are all enhanced in river-dominated deltas and therefore their effects are more profound on burrowing organisms. It is important to recognize that even though the Mississippi is regarded as an analogue for the deposits of FA5, there are a few differences between the depositional settings between these two deltas: Firstly, the Kamik deltas were deposited on the rifted margin of the pericratonic Brooks-Mackenzie Basin; whereas, the Mississippi delta is located on the passive continental margin of North America. The position of the Mississippi delta in the Gulf of Mexico is affected by much lower wave energy and storm frequency than the storms that were influencing deposition of the Kamik Formation (see section 3.8). Secondly, the Mississippi delta is located in a humid climate, whereas, the Kamik Formation was deposited in a northern region with an arid to temperate northern climate.

3.5 Comparison of the Rhône and Mississippi Deltas

Overall, it is important to recognize that there are major geomorphologic differences between the modern Rhône and Mississippi deltas that should be taken into consideration when comparing these two delta systems. Table 3.2 displays the contrasting characteristics of the Rhône delta and the Mississippi delta. The main difference between these two delta systems lies in the relative dominance of basinal processes such as river sediment input, wave-energy and tidal range in the depositional basin, all of which act to rework the riverborne sediments. The lobate Rhône delta is wave-dominated due to a strong wave-climate in the Mediterranean basin over the comparatively weaker fluvial and tidal energies. In contrast, the river-dominated, birdsfoot morphology of the Mississippi delta is a result of the large fluvial discharge supplied by the Mississippi River and the extremely weak wave power and tidal energy in the Gulf of Mexico. The shoreline of the Rhône delta is considerably more regular than that of the Mississippi, with common wave-deposited beaches, coastal barriers, spits and beach ridges. Riverine-influence in the Rhône delta is reflected by the pronounced lobate form with interdistributary bays and marshes behind the sandy beaches ridge complexes of the delta shoreline. It has been recognized that the more wave-dominated the delta, the greater the proportion of lobe sediment with limited amounts of interdistributary

Characteristics	Rhone Delta[1]	Mississippi Delta[2]
Drainage Area	95,830 km^2	3,344,560 km^2
Subaerial Area	2,590 km^2	23,900 km^2
Classification (Galloway (1975)	Wave-dominated	River-dominated
Delta Morphology & Geometry	Lobate	Birdsfoot
Annual Sediment Discharge	4.08 x10^{10} kg	6.21 x10^{11} kg
Content of suspended matter at surface, directly seaward of bar	5 to 15 mg/L	25 mg/L
Current velocity at surface, directly seaward of the bar	1.0 meter/sec	0.6 meter/sec
Content of suspended matter at surface, directly seaward of bar	5 to 15 mg/L	25 mg/L
Percentage >62u of suspended matter at surface, directly seaward of bar	5 to 25%	2%
Visibility (3.3 km from river mouth)	3 meters	0.6 meters
Visibility (5.0 km from river mouth)	5 meters	0.3 meters
Maximum slope of the delta front	2 degrees	1 degree

1 - Based on average river discharge data from van Straaten (1959) and Fisher *et al.* (1969)
2 - Based on low river discharge data from Scruton (1959) and Elliott (1996)

Table 3.2 Table displaying the contrasting characteristics of the Rhône delta and the Mississippi delta (from van Straaten (1959) with data from Scruton (1956).

bay and distributary channel facies (Bhattacharya and Walker, 1991). Interdistributary bay deposits are poorly developed in the modern wave-dominated Saõ Francisco delta; a delta which is characterized by strong littoral drift, limited development of interdistributary bays and lack of bifurcating distributary channels (Wright and Coleman, 1973; Coleman, 1981). In comparison, the modern Rhône delta, which is not as strongly influenced by wave-energy as the Saõ Francisco delta, is characterized by the development of multiple distributary channels, interdistributary bays and crevasse splay deposits. The strongly river-dominated Mississippi delta system is characterized by numerous distributary channels, interdistributary bays, marshes, crevasse sub-deltas and bay fill successions, which results in a highly irregular coastline.

Significant differences between the two deltas also include the size of the drainage areas, the discharge rates and the size of the subaerial delta plains (Table 3.2). In terms of scale, the Rhône delta is significantly smaller than the Mississippi delta, the latter with a drainage area 35 times larger and a subaerial area approximately 9 times larger. The annual sediment discharge from Rhône River at the delta front is approximately 4.08 x 10^{10} kg, consisting of fine-grained sands and minor silts and clays. In comparison, the annual sediment discharge at the delta front of the Mississippi River is extreme and is estimated at 6.21 x 10^{11} kg consisting of a fine-grained sediment load comprised clays, silts and fine sand (Elliott, 1996). Aside from the differences in scale, it was recognized by Andel and Curray (1960) that one of the dominant reasons for the differences in structure and lithology between the Rhône and Mississippi deltas was the higher proportion of sand in the Rhône. The significance and implications of sediment size and lithology for the morphology and configuration of the deltas and the distribution of deltaic sediments has been discussed in detail by Orton and Reading (1993).

3.6 Discussion of Depositional Setting

The presence of the deposits of river- and wave-dominated deltas within the upper Kamik Formation indicates conditions existed that favoured the development of deltas with both morphologies. As discussed in section 3.3, the wave-dominated delta deposits of FA1 are much more common than the river-dominated deltaic deposits of FA5. Therefore, it is interpreted that the majority of the deltas responsible for the deposition of the Kamik Formation were dominated by wave and storm processes and resembled the modern Rhône delta. However, the existence of river-dominated deltas of FA5 suggest that conditions also existed that were favourable for the development of river-dominated deltas with morphologies similar to the modern Mississippi delta.

Wright and Coleman (1973) observed that in a general sense, the greater the fluvial discharge of a delta, the greater the wave energy required to rework the sediments into wave-dominated configuration. These authors noted that during periods of high discharge (i.e. floods accompanying storm events) coincided with times of greatest wave power when the increased fluvial sediment supplies may be balanced more closely by the waves ability to redistribute the material. During periods when fluvial discharge and wave-power are out of phase, river outflow may dominate for part of the year, followed by a period of intense reworking by waves (Wright and Coleman, 1973). Periods when sediment supply and wave energy are relatively in phase favour the more regular and smooth progradation of a delta shoreline and the development of a series of successive spits, barrier beaches, or beach ridges along the coast flanking the river mouths (Wright and Coleman, 1973). During periods when discharge is low, wave-generated longshore drift would remove the river-derived sediments and waves would reach the shoreline undiminished and favour the development of wave-dominated deltas with beach ridges and barrier islands (Coleman and Wright, 1971). In comparison, periods when wave energy was low (i.e. during the summer months), the fluvial processes may become totally dominant and favour the development of an irregular shoreline and river-dominated deltas. If the discharge of a river is extremely high and waves are unable to remove sediments through littoral drift, the resultant delta front may attenuate wave power to the extent that the wave energy that reaches the shoreline would be almost totally ineffective, resulting in the development of a delta with a river-dominated morphology.

This formation of deltas with river-dominated morphologies during periods of enhanced fluvial discharge is one possible explanation for the development of both the wave- and river-dominated deltas of FA1 and FA5. It is recognized that during deposition of the majority of the Kamik deltas the basin was strongly influenced by wave and storm-energy and conditions favoured the development of wave-dominated deltas. It is likely that during the deposition of the river-dominated deltas of FA5 the nature of the basin itself did not change, but rather localized processes were responsible for the variations in delta morphologies. The development of fluvial-dominated deltas of FA5 may have resulted from relatively short-lived periods of enhanced fluvial deposition accompanying storms causing riverine processes to become dominant. During these periods of enhanced fluvial deposition, the normally wave-dominated Kamik deltas of FA1 would have become flooded and temporarily adopted a river-dominated morphology. A modern observation of this process has been recognized from recent studies of the Brazos delta, where the dominantly wave-dominated delta adopts a fluvial-dominated morphology during floods (Rodriguez *et al.*, 2000). The preservation of these river-dominated deposits would required favourable local conditions (i.e. tectonically enhanced subsidence rates) for the rapid burial and preservation of these river-dominated deposits, thereby limiting the effects of wave reworking. Active growth faulting in the Eskimo Lakes Fault Zone during deposition of FA5 could have lead to tectonically enhanced subsidence and the creation of accommodation space required to rapidly bury and preserve these fluvial-dominated deltaic deposits.

A second explanation for the variation in delta types is attributed to the development of the Brooks-Mackenzie Basin by active rifting and variations in sediment supplies and shoreline configuration during Kamik deposition. As mentioned earlier, the Kamik Formation was deposited on the rifted margin of the pericratonic Brooks-Mackenzie Basin along the Eskimo Lakes Fault Zone (Dixon, 1982a, 1991). At this time, the Eskimo Lakes Fault Zone was likely a step-like terrace composed of numerous fault blocks, grabens and half-grabens that resulted in the development of a highly irregular shoreline. The nature of this shoreline is interpreted to have resulted in the development of smaller depositional basins that may have been protected from the wave and storm-energy of the open ocean basin that were affecting the wave-dominated deltas of FA1. The sheltering of these deltas from the full effects of waves and storms in the open Brooks-Mackenzie Basin may have facilitated the development of the river-dominated deltas of FA5.

Finally, it has been recognized by numerous authors that when delta lobes become inactive or abandoned and destructional processes dominate with basinal processes the preserved record of deltaic successions can change markedly (Boyd and Penland, 1988; Penland *et al.*, 1988). An example of this process is the modern birdsfoot lobe of the Mississippi delta, which is currently undergoing transgression as a result of tectonic subsidence in the Gulf of Mexico. The delta front of the abandoned delta lobes is modified by wave action resulting in the formation of barrier bars and lagoon systems (Penland *et al.*, 1988). Although wave energy is weak in the Gulf of Mexico, the absence of active fluvial input results in waves being able to effectively redistribute fluvial derived sediments. The reworking of the river-dominated deltaic sediments by wave processes following abandonment can result in the preservation of a deltaic deposits which appear to have been deposited in a wave-dominated delta. An example of this process is occurring in the modern Mississippi delta, where river-dominated delta lobes are actively being reworked by waves following delta abandonment. If the preserved wave-reworked deposits were the sole basis for interpreting the delta morphology, this might lead to the erroneous interpretation that this delta was originally wave-dominated rather than river-dominated. Therefore, it should be recognized that the interpretation that the majority of the deltas responsible for the Kamik Formation were wave-dominated is susceptible to bias based the preservation of the deltaic deposits and the availability of core data from the Parsons Lake area. It is possible that additional core data may show that the majority of deltas responsible for the Kamik Formation were river-dominated, rather than the current interpretation of being predominantly wave-dominated.

3.7 Evolution of Delta Systems Through Time

Deltas are complex depositional systems, which constantly evolve throughout their existence. Deposition within these systems reflect processes that existed at a given time and may not be representative of the entire deltaic cycle (Suter, 1994). Delta deposition is cyclic in nature and consists of two phases; 1) a constructional phase during which the delta progrades, and 2) a destructional or abandonment phase initiated by a reduction in the amount of sediment supplied to the delta (Fisher *et al.*, 1969; Elliott, 1996). The evolution of deltaic systems is influenced by both autocyclic and allocyclic processes. Allocyclic processes include factors such as eustatic sea level fluctuations and climate changes which act independently of deltaic deposition. Autocyclic processes include delta lobe switching due to distributary channel avulsion when an alluvial system or distributary is abandoned or avulsed for a more hydraulically favoured route (Suter, 1994). Autocyclic processes are more dominant in deltaic systems and are a part of the system itself. Channel avulsion and delta lobe switching processes are more common processes in river- and tide-dominated deltas than in wave-dominated deltas (Elliott, 1996). Autocyclic processes occur in virtually all deltas, but the best studied examples are displayed in the Mississippi delta (Suter, 1994).

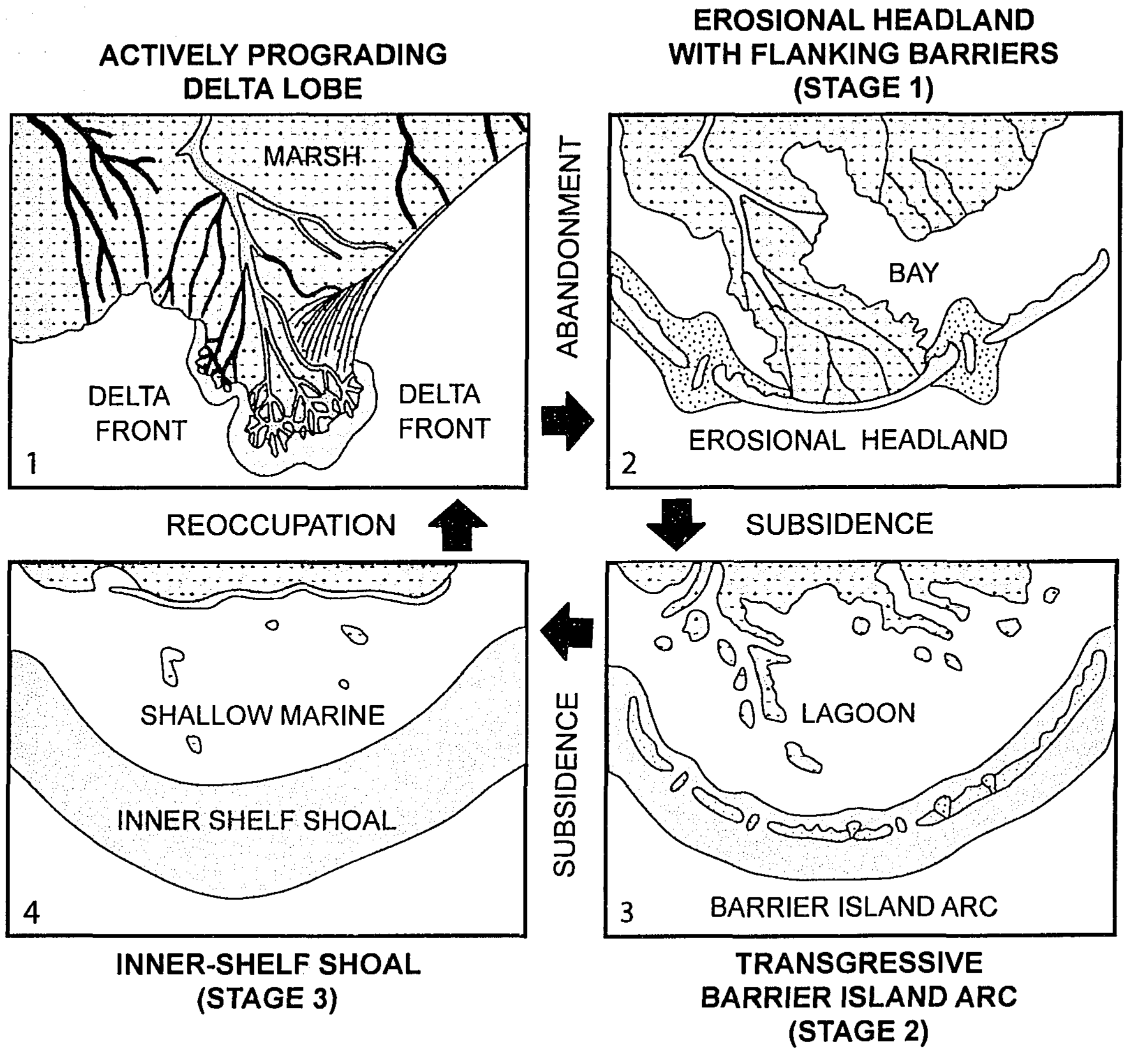


Figure 3.19 Stages in the evolution of a delta, based on examples from the ancient abandoned Mississippi Delta lobes. Note the contrast between the constructional (active) phase 1 and the transgressive destructional (non-deltaic) stages in panels 2 to 4. Reworking during transgression forms barrier islands arcs (panels 2 to 4) which when submerged form inner shelf shoals (panel 4). Simplified from Boyd and Penland (1988) and Penland *et al.* (1988).

Delta switching in the Mississippi delta occurs on two scales, delta lobe switching from the abandonment of a series of delta lobes by the over-extension of distributary channels, and avulsion of major trunk distributary channels. The abandonment of a delta lobe due to lobe switching initiates the transgressive or destructional phases of the delta cycle (Scruton, 1960); once abandoned, the older delta lobes of the Mississippi delta undergo rapid subsidence and coastal processes rework the seaward margin (Penland *et al.*, 1988). In the Rhône delta, lobe switching occurs due to channel extension, in which several distributaries originate from a common point, switching back and forth as the active conduit (Oomkens, 1970). This results in the formation of a series of beach ridges oriented about the various river mouths.

Penland *et al.* (1988) established a three-stage evolutionary model for the development of transgressive delta, based on examples from the abandoned Mississippi delta lobes (Fig. 3.19). This model consists of a constructional (active) or regressive phase of delta progradation (Fig. 3.19A) followed by a destructional (non-deltaic) or transgressive phase of deposition (Figs. 3.19B-D). The transgressive or destructional phase is initiated with the abandonment and subsequent subsidence of an actively prograding delta due to autocyclic process (i.e. delta lobe switching or channel avulsion). Abandonment of an active delta complex results in the formation of transgressive depositional systems, which follows a process of transgressive submergence in which the horizontal component of reworking occurs during shoreface retreat. This combined with a vertical component of submergence acts to preserve the depositional sequence.

The depositional history of each delta complex within the Mississippi delta starts with the progradational phase in which the Mississippi River deposits its sediment load into the Gulf of Mexico (Boyd and Penland, 1988). Progradation of the delta lobe leads to the over-extension of the distributary network and a decrease in hydraulic efficiency, followed by distributary avulsion. The channel switches it course to a shorter, more hydraulically efficient course with a steeper gradient. The abandonment of the older delta lobe results in the transgressive phase of the delta cycle (Boyd and Penland, 1988). The transgressive or destructional phase of delta evolution as defined by Penland *et al.* (1988) is summarized in a three stage sequence. The sequence is initiated with stage 1, which encompasses the reworking of the regressive sandstone deposits within the abandoned deltaic headlands. These deltaic sands are reworked by the eroding shoreface and dispersed by longshore currents into contiguous flanking barriers enclosing restricted interdistributary bays resulting in the formation of an erosional headland with flanking barrier beaches (Fig. 3.19B). During storms, sand is washed over the beach ridges onto the abandoned delta top, and the beach and beach ridge complex is reworked during transgression and moves landward and keeps pace with the transgressing sea. Normally, this process forms a distinct winnowed deposit or a transgressive lag; however, in the rapidly subsiding Mississippi delta lobes, the subsidence rates are high and the delta top becomes submerged as a shallow bay before the passage of the beach-ridge complex and is stranded out to sea as a barrier island. This results in stage 2, in which a transgressive barrier island arc is formed by the submergence of the delta plain during relative sea-level rise and results in the generation of an interdeltaic lagoon separating the stage 1 sand body from the shoreline (Fig. 3.19C). An example of this is the abandoned St. Bernard lobe of the Mississippi delta plain that is now separated from the Chandeleur barrier islands by Chandeleur Sound (Bhattacharya and Walker, 1992). The barrier islands protect Chandeleur Sound from most of the wave action and the deposits of the St. Bernard lobe are slowly transgressive with minimal wave reworking and winnowing at the lobe edge. Due to the attenuation of the transgressive wave energy by the barrier islands the mudstones and siltstones of the delta plain are overlain by brackish-water and marine mudstones of the delta margin bay. The last stage in the sequence is characterized by the inability of the landward-migrating barrier island arc to keep pace with relative sea-level rise and the retreating mainland shoreline, resulting in the submergence of the barrier island arc and the formation of the inner-shelf shoal (Fig. 3.19D) (Boyd and Penland, 1988).

The transgressive component of deltaic deposition accounts for the majority of the surface area in the lower Mississippi River delta plain and up to 50% of the total sequence thickness in shallow-water deltas (Penland *et al.,* 1988). The long term evolution of a delta system is a function of the rate of riverine sediment input and the rate and pattern of sediment reworking, transport, and deposition by marine processes after the initial deposition and subsequent abandonment. In the Mississippi delta, a lobe or complex evolves from fluvial dominance during its deposition, to wave dominance during its transgressive phases (Russell, 1936; Fisk, 1944; Scrunton, 1960).

Allocyclic processes represent a significant influence on the deposition of deltaic sediments. Eustatic fluctuations have large effects on the style of deltaic deposition (Suter and Berryhill, 1985; Elliott, 1989; Suter, 1994). Rising sea level creates greater amounts of accommodation space and favours aggradation or retrogradation over progradation and the formation of shelf-phase or shoal-water deltas (Suter, 1994). In contrast, during sea level lowstands falling sea level forces the deltaic system to prograde rapidly across the emerging continental shelf, which can result in entrenchment of a particular distributive network of incised fluvial systems and formation of deep water or shelf edge deltas (Suter and Berryhill, 1985). Climatic fluctuations can alter fluvial discharge and/or sediment supply by determining the rate of physical and chemical weathering, governing the amount and calibre of the sediment load supply (Galloway, 1975). Overall, deltas represent composite systems that record the evolution of depositional systems through a variety of phases of basin subsidence, relative sea level changes, nature of the receiving basin, and the rate of sediment influx.

The concept of delta abandonment was developed based on the Mississippi delta and its universality may not be directly applicable to all deltas, especially considering the variability of modern deltas. The initiation and abandonment of delta lobes or complexes is related to the frequency of channel avulsion. Rapid progradation in river-dominated deltas produce significant shoreline protuberances and numerous areas of gradient advantages, resulting in frequent channel avulsion and delta lobe abandonment (Elliott, 1996). In wave-dominated deltas the delta front advances slower over a broader front resulting in the creation of fewer gradient advantages and less frequent channel avulsions. In addition, after abandonment in wave-dominated deltas a greater proportion of the former delta is likely to be reworked by wave action. Therefore, distributary channel switching and the development of delta lobes are a preferential feature of fluvial-dominated deltas and a less common feature in wave-dominated deltas (Elliott, 1996). This process is illustrated by the wave-dominated Rhône delta, where only three lobes have formed during the same 6000 to 7000 year period in which the river-dominated Mississippi River delta built at least six major delta complexes (Frazier, 1967).

3.8 Intense Winter Storms

The predominance of storm deposits (i.e. tempestites) within the Kamik Formation indicates that storms were a strong influence on the deposition of these sediments. Storms represent one of the most important influences on sediment deposition on shorefaces and delta front environments as they are able to accomplish significant amounts of erosion and deposition in a few hours that normally takes fairweather processes years to accomplish (Morton, 1988). The sedimentological and ichnological affect of intense storms, such as hurricanes and winter storms on shorefaces has been discussed by numerous authors (see Hayes, 1967; Marsaglia and Klein, 1983; Pemberton and Frey, 1984; Duke, 1985; Morton, 1988; Barron, 1989; Pemberton *et al.,* 1992; Saunders *et al.,* 1994; MacEachern, 1994). Storms events are responsible for the offshore transport of sand, creation of significant erosion and suspension of sediment, modification of coastlines and river meandering paths and the deposition of storm layers and lag deposits, including HCS and SCS (Kreisa, 1981; Dott and Bourgeois, 1982; Duke, 1985; Barron, 1989). Recognizing the significance of storms on the deposition on the Kamik Formation provides insight into the paleogeography and paleodepositional setting of the Brooks-Mackenzie Basin during the Lower Cretaceous.

During the Cretaceous, global climatic conditions were much warmer than present conditions and it is generally agreed that the entire Mesozoic was a non-glacial time and polar regions were likely free of major ice caps (Frakes, 1979; Haq *et al.*, 1987). The ice-free nature of the polar oceans during the Early Cretaceous would have provided a greater opportunity for storm systems to develop and influence the deposition of the Kamik Formation. It has been recognized that the only storms capable of profoundly affecting shallow-marine depositional environments are hurricanes and mid-latitude intense winter storms (Duke, 1985). In the Early Cretaceous, it is interpreted that both hurricanes and intense winter storms occupied essentially identical latitudinal belts to those occupied by their modern counterparts. Hurricanes are relatively small, intense storms that originate in equatorial regions and travel from east to west due to the general flow of winds of the tropics (Morton, 2003). These storms generally form during the summer or early fall when warm seawater and atmospheric circulation favours storm generation (Simpson and Riehl, 1981). Due to the globally warmer climatic conditions during the Mesozoic, it was suggested by Duke (1985) that the modern hurricane distribution belt was broadened and hurricanes occurred more frequently at higher latitudes than today. Even though the distribution of hurricane belts was broadened, their distribution was still limited to the equatorial regions.

High latitude settings, such as those in which the Kamik Formation was deposited, favour the development of higher frequency storms, termed intense winter storms (Barron, 1989). Winter storms are large, mid- to high-latitude storms that are driven by cold, Arctic air masses that travel from west to east (Morton, 2003). These storms develop along fronts between cold and warm air masses (Strahler and Strahler, 1979) and may originate over continents or oceans (Duke, 1985).

Figure 3.20 Paleogeographic map illustrating the extent of storms in the earliest Cretaceous Infravalanginian seas. Winter storms are interpreted to have to dominated in the northern Pacific Ocean and in the Brooks-Mackenzie Basin, whereas hurricanes were dominant in the southern equatorial region. Note the location of the study area marked by the red box (Diagram based on the work of T.D.A Saunders; predicted after Williams and Stelck, 1975).

Winter storms are larger than hurricanes and inflict a greater influence on coasts because they occur more frequently (several each year) and may last for several days, whereas hurricanes occur less frequently and only last for a few hours as they cross the coast. Intense winter storms develop in the same manner as hurricanes; however, they can develop over continents or oceans and form along the fronts between cold and warm air masses. During the early Cretaceous, these high latitude settings were characterized by a strong seasonality and cyclicity between winter and summer storm seasons. The strong seasonality of these climates was conducive to the formation of cold polar air masses and these intense winter storms (Owens, 1977). The winter season is dominated by frequent and high magnitude intense storms, whereas the summer season is typically marked by less frequent and lower intensity storms (Owens, 1977). The higher frequency and magnitude of winter storms causes storm deposits to accumulate more rapidly during the winter than during the summer. It has been suggested that the lack of ice covering on the polar oceans would have extended the length of the winter season and provided a greater opportunity for these storms to influence the depositional basin (Duke, 1985; Morton, 1988). In addition, Barron (1989) noted that during the mid-Cretaceous severe winter storms were most prominent in the North Pacific and extended into the northwestern portion of North America. It follows that during deposition of the Kamik Formation in the early Cretaceous, there was similar prominence of severe winter storms in the adjacent Brooks-Mackenzie Basin.

During deposition of the Kamik Formation, the Brooks-Mackenzie Basin was situated in a northern region that was exposed to the north and northwest to the open ocean. Figure 3.20 is a paleogeographic map, which illustrates the extent of storms in the earliest Cretaceous seas. Intense winter storms are interpreted to have dominated in the northern Pacific Ocean and Brooks-Mackenzie Basin, whereas hurricanes were dominant in the southern equatorial regions. The basin at this time was an unsheltered, open ocean basin that was large enough to develop large, intense winter storms. Open-ocean basins, such as the Brooks-Mackenzie Basin, are known to facilitate greater wave fetch and therefore the potential for greater wave- and storm-energy and the development of large storms (Suter, 1994). The resulting basin configuration created a setting where the shorelines were strongly and continuously reworked by intense winter storm waves. The influence of these storms is represented by the pervasiveness of storm-generated HCS and SCS tempestite deposits throughout the deltaic deposits of the Kamik Formation. The overall degree of storm-influence is interpreted to have been moderate and the abundance of storm deposits indicates that intense, frequent winter storms were occurring on a regular basis. The paleogeography of the basin at the time of Kamik deposition favoured the development of intense winter storms and facilitated the development of wave-dominated deltas. Deltaic environments are extremely vulnerable to storms because the subaerial surfaces are flat and only slightly above the local mean seal level; a slight rise in sea level can extend the zone subject to storm surges and waves farther inland (Suter, 1994). This interpretation is similar to the interpretations of Barnett (2003) for the underlying Jurassic Aklavik Formation; however, the degree of storm amalgamation is not as high in the Kamik Formation as in the Aklavik Formation. The variation in the prevalence of storm deposits between the Kamik and Aklavik formations is attributed to differences in the geographic location of the study area and the overall differences in the configuration of the Brooks-Mackenzie Basin and local tectonics for the late Jurassic compared to the Early Cretaceous.

3.9 Summary

The fifteen facies (Facies A to Facies O) that were identified within the Kamik Formation were grouped into five distinct facies associations (FA1 to FA5). These five facies associations were interpreted to represent moderately storm-influenced, wave- and river-dominated deltas, meandering and braided distributary channel systems and lower delta plain sediments. During deposition of the Kamik Formation the depositional basin was strongly influenced by storms, specifically intense winter storms that created abundant storm-generated tempestite beds, which characterized the prodelta and delta front sediments of FA1 and FA5. The majority of the deltaic sediments comprising the Kamik Formation are interpreted to represent deposition in a moderately storm-influenced, wave-dominated delta. The modern Rhône delta is interpreted to represent an appropriate depositional analogue for the deltaic systems responsible of FA1. The modern Mississippi delta is interpreted to represent an appropriate depositional analogue for the deltaic systems responsible for FA5.

Chapter Four: Conclusions

4.1 Conclusions

1. Detailed ichnological and sedimentological analysis of approximately 200 meters of core from 14 cored intervals from the Lower Cretaceous Kamik Formation in the Parsons Lake gas field resulted in the identification of 15 facies (Facies A to O) which were subsequently organized into five facies associations (FA1 to FA5).

2. The five facies associations (FA1 to FA5) identified within the Kamik Formation are interpreted to represent a moderately storm-influenced, wave-dominated deltaic succession (FA1), a meandering distributary channel (FA2), a braided distributary channel (FA3), a lower delta plain (FA4) and a moderately storm-influenced, river-dominated deltaic succession (FA5).

3. The Kamik Formation is interpreted to be deltaic in origin and was deposited as a series of prograding delta lobes characterized by varying degrees of wave, storm and fluvial influence. Overall, wave- and storm-energy were the dominant processes influencing deposition, resulting in the predominance of wave-dominated deltas. Some of the successions also display evidence of a strong riverine influence and are interpreted to have been deposited in a moderately storm-dominated, river-dominated delta depositional system.

4. Facies Association 1 (FA1) corresponds most closely to a wave-dominated deltaic succession. Facies Association 5 (FA5) corresponds to a river-dominated deltaic succession. The discrimination of FA1 and FA5 as wave- and river-dominated deltas is based primarily on the basis of physical sedimentology and ichnology.

5. The modern Rhône delta of southern France is interpreted to represent an appropriate analog for the wave-dominated deltas of FA1 and the modern Mississippi delta southeastern United States is interpreted to represent an appropriate analog for the river-dominated deltas of FA5.

6. The development of both the wave- and river-dominated deltas of FA1 and FA5 is interpreted to be a result variations in the fluvial discharge at the delta front. During periods of enhanced fluvial deposition the normally wave-dominated Kamik deltas of FA1 would have become flooded and temporarily adopted a river-dominated morphology. The river-dominated deltas of FA5 is interpreted to have resulted from relatively short-lived periods of enhanced fluvial deposition during floods when riverine processes became dominant over wave processes. A modern example of this process has been recognized from the Brazos delta, where the normally wave-dominated delta adopts a fluvial-dominated morphology during large floods.

7. During deposition of the Kamik Formation, the Brooks-Mackenzie basin was an unsheltered, open ocean basin that was characterized by frequent intense winter storms. These intense winter storm waves continuously reworked the deltaic shorelines and facilitated the development of the pervasive storm-generated deposits and favoured the development of wave-dominated deltas.

8. The abundance of storm-related tempestite deposits (HCS and SCS bedforms) throughout the deltaic deposits of the Kamik Formation implies that storms were a constant influence on the deposition and that the storm climate of the basin was 'moderately storm-dominated'. The depositional basin was large enough to facilitate a large wave fetch and therefore a greater potential for wave- and storm-energy and the development of large storms.

9. Detailed ichnological and sedimentological analysis is key to understanding the dynamic interplay between wave and riverine processes in deltaic systems. The integration of this data provides an effective tool in differentiating between wave- and river-dominated deltaic successions from the Kamik Formation and the prediction of the geometries of other ancient deltaic successions.

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