SEDIMENTOLOGY AND ALLOSTRATIGRAPHY OF THE UPPER CRETACEOUS (CAMPANIAN) LEA PARK - BELLY RIVER TRANSITION IN CENTRAL ALBERTA, CANADA

By

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Once in his life a man ought to concentrate his mind upon the remembered earth.
He ought to give himself up to a particular landscape in his experience, to look
at it from as many angles as he can, to wonder upon it, to dwell upon it.
He ought to imagine that he touches it with his hands at every season
and listens to the sounds that are made upon it.

N. Scott Momaday

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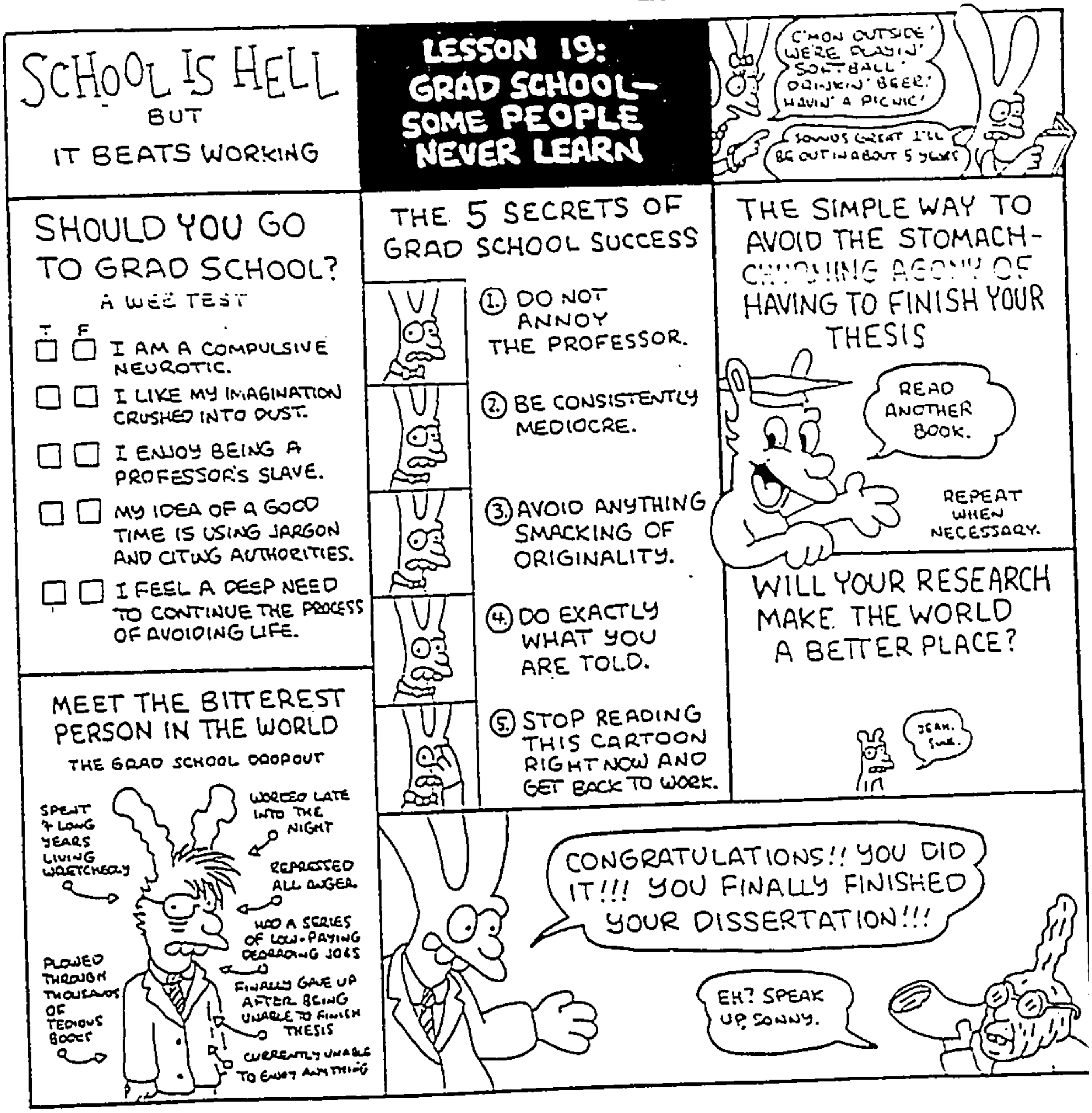
Life In Hell by Matt Groening



ABSTRACT

Sedimentological and allostratigraphic analysis of the Upper Cretaceous (Campanian) Lea Park - Belly River transition in central Alberta reveals that the sediments comprise a downlapping wedge of interbedded shallow marine to non-marine sediments.

Vertical and lateral facies relationships indicate that the coastal sediments of the Lea Park - Belly River transition were deposited in sandy, fluvially-dominated deltaic successions. The dominant feature of these deltaic successions is the general absence or scarcity of well-developed sedimentary structures. Instead, the successions are dominated by stacked beds of fine-grained, structureless or vaguely laminated sandstone which often comprise the basal portion of massive-to-laminated beds. The structureless sandstones are present within both the proximal and distal portions of the deltaic sandbodies, as well as in all locations alongshore. In some cycles, the massive-to-laminated beds are present tens of kilometres offshore of the last known position of the shoreline. Deltaic sand body geometries show the presence of elongate, shore-normal tongues of sandstone up to 70 km in length, which are surrounded by thinner, lobate sandstone sheets. These sand bodies are also characterized by an apparent absence of major fluvial distributaries cutting into the tops of the deltaic cycles. The dominant form of channel sediments deposited on the tops of the deltaic successions are thin successions, often do

not show demonstrably erosive bases, and contain poorly developed internal successions that are difficult to distinguish from upper shoreface successions associated with the delta.

This combination of facies relationships and coastal sand-body geometries can not be interpreted using existing deltaic facies models. A non-actualistic model is proposed as a possible solution. This model proposes that the coastal sediments of the Lea Park - Belly River transition were deposited in fine-grained, sandy "braid deltas", in which a maze of shallow channels would distribute sediment all along the shoreline. Rather than having one major fluvial distributary depositing the delta, the deltaic shoreline is interpreted to have been created by numerous smaller fluvial systems simultaneously depositing sediment at the shoreline. The massive-to-laminated beds are interpreted to have been deposited by turbidity currents which carried very large amounts of sand in suspension. These turbidity currents were most likely generated by slumping of pre-deposited delta mouth sediments, but may also have formed as hyperpycnal fluvial density currents.

The deltaic coastal systems are laterally equivalent to thick units of fluvial and associated non-marine sediments to the west. These fluvial systems were undoubtedly supplying the deltaic shorelines further to the east. Sediments indicative of both fine-grained, meandering fluvial systems and coarser-grained, higher energy fluvial systems are present within the study area.

There are eight separate shoreline successions, termed cycles, in the Lea Park - Belly River transition preserved within the study area. These cycles are stacked in a prograding, downlapping wedge that becomes younger to the northeast. This wedge is characterized by a regressive-transgressive cyclicity, wherein each cycle is separated from the overlying cycle by an interval of transgressive marine sediments. The shoreline succession in each cycle is very sharply-based, such that the transition between the mid/lower shoreface and the deeper water shelf sediments is either absent or very thin. The shoreline cycles are laterally equivalent to incised fluvial sediments, which in some locations sit directly on fine-grained marine shelf sediments. This combination of sharply-based shoreline cycles and incised fluvial sediments is interpreted to be indicative of forced regressions, where a relative drop in sea level causes the shoreline to move rapidly basinward.

This interpretation of rapid drops in relative sea level indicates that the regressive - transgressive cyclicity of the Lea Park - Belly River transition was allocyclically controlled. The overall time frame for deposition of all eight cycles within the study area is thought to be on the order of 1.0-1.25 million years, indicating that the Lea Park - Belly River cycles were deposited in response to fourth-order fluctuations in relative sea level. The underlying control of these fluctuations is speculative, but they may be due to fluctuations in subsidence rates due to active loading in the Cordillera acting in concert with a third-order drop in relative sea level.

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First and foremost, I would like to thank Dr. Roger Walker, who has supervised and supported both this project, and my Master of Science thesis four years ago. This project would not have been possible to undertake and complete without Roger's financial support and intellectual advice over the past four years. Many problems arose during this time, and Roger was always able to give constructive advice on how to find a solution. I have spent six years here at McMaster learning about sedimentology and stratigraphy from Roger, and his limitless energy and ability to pass on his knowledge is unmatched by any person I have encountered in the scientific community. I will always be grateful for all that Roger has taught me about the science of geology.

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All of the cores logged for this study are stored at the Energy Resources Conservation Board Core Research Centre in Calgary. The staff at this fine institution is thanked for their friendly and efficient service during the past three summers.

A number of other people also made significant contributions to this study. Jack Whorwood provided expert photographical services for this manuscript. His ability to turn out prints of artistic quality from my poorly-taken slides was remarkable. Janok Bhattachraya provided countless useful discussions over the course of the study on sequence stratigraphy and the meaning of life, the universe, and everything . Indraneel Raychaudhuri provided field assistance during an eventful first summer of research in 1987.

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TABLE OF CONTENTS

CHAPTER 11: ANALYSIS OF DELTAIC AND FLUVIAL SYSTEMS

LIST OF FIGURES

CHAPTER 5

CHAPTER 6

CHAPTER 7

CHAPTER 8

CHAPTER 9

LIST OF TABLES

CHAPTER 1: THE SCIENTIFIC PROBLEM

1.1: Introduction

This study of the sediments of the Lea Park - Belly River transition in central Alberta addresses the manner in which a large wedge of sediment progrades into a foreland basin during a relative drop in sea level. This problem involves an analysis of how coastal depositional systems prograde into the basin as the relative sea level drops and the sediment wedge fills in the basin, and the manner in which these depositional systems are affected by high frequency relative sea level fluctuations superimposed on the longer term relative drop in sea level.

The Upper Cretaceous (Campanian) Belly River Formation in Alberta is a thick wedge of dominantly non-marine sediment. Previous work has shown that the base of this wedge is diachronous, younging to the east, and that the basal sediments of the Belly River Formation and the uppermost sediments of the underlying Lea Park Formation in central Alberta are generally known to have been deposited in deltaic and associated coastal and nearshore environments. Prior to this study, there have been no public detailed stratigraphic and sedimentological analysis of these sediments, which are important hydrocarbon reservoirs in the subsurface of Alberta. An abundant data base of subsurface well logs and cores exists for the basal sediments of the Belly River Formation, making this interval an ideal subject for this study.

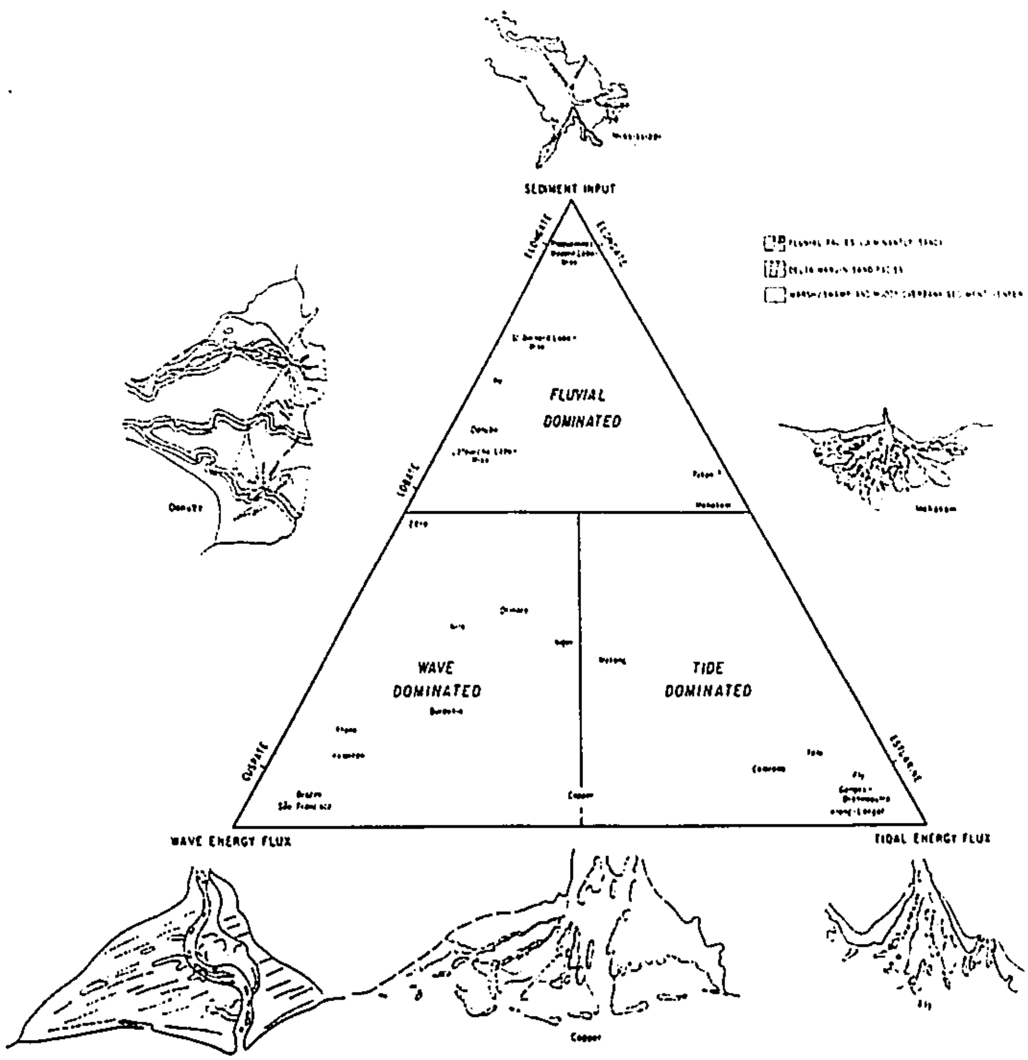
This analysis of the Lea Park - Belly River transition can be divided into two main areas. The first deals with applying techniques of facies modelling to determine the nature of the depositional environments of the sediments, especially with regards to how deltaic facies compare to existing facies models for deltas. Existing deltaic facies models are based on a small number of modern and ancient delta systems, with all of the modern systems being deposited during rising sea level. The Lea Park - Belly River transition presents a great opportunity to study the facies relationships and development of a series of ancient deltaic shorelines during a time of falling sea level.

The second problem deals with determining the three dimensional geometry and scale of the depositional systems, and how small-scale relative sea level fluctuations have affected these systems. There are very few existing studies of ancient sedimentary successions which present a three-dimensional analysis of the effects of high-frequency relative fluctuations in sea level on coastal depositional systems. Most well-doucumented examples of such studies that do exist focus on sedimentary successions at the depositional edge of a wedge, where the effects of relative fluctuations in sea level may be more dramatic. This study of the Lea Park - Belly River transition in central Alberta study will analyse cyclicity of shoreline systems within the context of falling sea level, rather than at the feather edge of a sediment wedge. This will be addressed by using the vertical facies relationships to set up an allostratigraphic subdivision for the sediments of the Lea Park - Belly River transition at the base of the Belly River wedge of sediment. An allostratigraphic subdivision of the sediments will allow for the interpretion of the succession with regards to the effects of fluctuations in relative sea level. In addition to the scientific questions on how high-frequency sea level fluctuations affect the establishment of coastal depositional systems at the base of a prograding wedge of sediment, there are possible industrial applications of this knowledge. Increased understanding of how relative sea level fluctuations affect the development and distribution of shoreline sediments can allow for better models of hydrocarbon reservoir geometries to be used for exploration and production. The following sections are a brief discussion of the background surrounding these questions, and how a study of the sediments of the Lea Park - Belly River transition may contribute to a better understanding of these problems.

1.2: Deltaic Facies Models

Facies models and the classification of deltaic systems are based on the relative importance of fluvial processes vs. basinal processes (Coleman and Wright, 1975). This has resulted in a classification scheme (Fig 1.1) where river-, wave-, and tide-dominated deltas form end members of a tripartite facies model based on the relative importance of these processes (Galloway, 1975). This classification scheme is based on the plan view morphology of the delta, especially with regard to the shape of the delta front. The implicit assumption

Figure 1.1: Triangular classification scheme for deltaic shorelines, with classification of major modern deltas plotted. Deltas are classified on the basis of their plan view morphologies, which are controlled by the relative influence of wave, tidal, and fluvial processes at the delta front. River-, wave-, and tide-dominated deltas form the end members of the triangle (from Galloway, 1975).



SEDIMENT INPUT
ELONGATE
ELONGATE
LOBATE
FLUVIAL DOMINATED
WAVE DOMINATED
TIDE DOMINATED
CUSPATE
ESTUARINE
WAVE ENERGY FLUX
TIDAL ENERGY FLUX
São Francisco
Copper
Fly

in this scheme is that the morphology of the delta will reflect the degree and nature of basinal reworking of the delta front, and is thus a reflection of depositional processes. The resultant sediment-body geometry should reflect the depositional environment, and thus the delta type.

The study of modern deltaic systems and those within the ancient record has been heavily biased toward river-dominated, lobate or bird-foot deltas such as the Holocene Mississippi Delta complex (Penland et al., 1988, 1987; Coleman and Wright, 1975; Frazier, 1967) and the Carboniferous deltas of the British Isles and the United States (Pulham, 1989; Elliott, 1975, 1976; Horne et al., 1978; Barrell, 1912). It is difficult, if not impossible to determine the three-dimensional geometry of ancient deltaic systems in outcrop, simply because of the limited data base. Subsurface studies, however, allow the possibility of determining the three-dimensional geometry of ancient deltaic systems. In the subsurface, river-dominated lobate deltas have been recognized in the Tertiary of the Gulf Coast of Texas in numerous studies (eg. Galloway, 1968). Due to the lack of core, detailed vertical and lateral facies relationships have not to date been included in most subsurface studies. Exceptions to this are the works of Bhattacharya (1989) and Bhattacharya and Walker (1991a, 1991b), which contain detailed facies relationships and sand-body geometries of fluvial-dominated deltas in the Cretaceous Dunvegan Formation of Alberta.

Other delta types are, by comparison, poorly studied. The modern Rhone and Sao Francisco deltas have been classified as wave-dominated deltas (Coleman and Wright, 1975; Oomkens, 1970). The morphology of the Rhone delta is an arcuate to lobate sheet of sand indicative of some degree of wave-reworking, but little is known about internal and three-dimensional facies relationships within the delta complex. Until recently, the same was true for the Sao Francisco. A recent study by Dominguez et al. (1987) has greatly elaborated on the internal facies relationships and stratigraphy of the Sao Francisco coastline, and calls into question its classification as a delta. it may be more properly classified as a strandplain. The best-known subsurface example of wave-dominated delta systems is the Upper Cretaceous San Miguel Formation in Texas (Weise, 1979). This study documented the three-dimensional sand-body geometries of San Miguel deltas, but contained little information on the facies relationships. Deltas within the Jurassic Brent Group under the North Sea have been interpreted as storm and wave-dominated on the basis of lithofacies successions, but geometries of the systems were not discussed (Brown and Richards, 1989). Bhattacharya (1989) has also recognized and delineated wave-dominated delta geometries within the Dunvegan Formation. This study also includes detailed facies relationships. Other ancient examples from outcrop studies include the Upper Cretaceous Rock Springs of Colorado (Kirschbaum, 1986) and the Upper Cretaceous Cody-Parkman delta (Hubert et al., 1972). These studies emphasized vertical facies relationships, but lack three-dimensional sand-body geometry control.

Tide-dominated deltas are even more poorly documented. The Ord River delta in Australia (Coleman and Wright, 1975) is commonly cited as an example of a tide-dominated modern delta, but may in fact be better classified as an estuary, rather than a delta. The best candidate for a modern tide-dominated delta is the Mahakam delta in Indonesia (Allen, 1985). Ancient examples of tide-dominated deltas are very rare. Ramos and Galloway (1990) recently proposed that the Eocene Queen City Formation in Texas contained examples of tide-dominated delta-embayments. This study details outcrop facies relationships and subsurface well-log geometries, but lacks three-dimensional facies relationships. Another possible example is the Jurassic Cloughton Formation in Yorkshire, England, which is interpreted as a small-scale wave- and tide-dominated delta (Livera and Leeder, 1981).

The work of Bhattacharya (1989) may be the most comprehensive recent study of ancient deltaic systems that contains both detailed facies relationships and three-dimensional sediment-body geometries. In order to improve our knowledge of deltaic systems, especially with regards to how these systems may have reacted to relative changes in sea level, other studies of ancient systems that combine three-dimensional geometries and detailed vertical and lateral facies relationships are needed. The sediments of the Lea Park - Belly River transition provide an opportunity to add to our knowledge in this area.

The lack of detailed examples showing facies relationships and three-dimensional geometries of ancient deltas may not be the only problem that needs further study. There is a need to address the problem of whether the present tripartite morphological classification scheme is a satisfactory method for classifying all deltas. Deltas are complex depositional systems, and cannot in all cases be adequately described by a scheme which focuses on one major characteristic, such as the coastal morphology of a system. Although sedimentary processes are, to some extent, implicity included in the present classification scheme, Martinsen (1990) feels that the river-mouth processes of deltas are not adequately documented using the tripartite method. He proposes that classification schemes should be developed that; (1) emphasize the nature of the river-mouth processes that actually transport and deposit the sediment, and (2) the resultant large-scale morphology. This is based on his interpretation of the Scar House Beds (Namurian) of England as the deposits of a fluvial-dominated deltaic system that is not adequately described and classified in the present scheme. Wright (1978) discussed the nature of processes that interact at the mouths of rivers, and further integration of process-oriented studies such as this into deltaic classification schemes may be a useful method of addressing this problem. Other features such as tectonic setting, effects of sea level state, climate, and sediment type have rarely been discussed in deltaic classifications systems. There are no well-documented examples of modern delta systems in foreland basins. The fact that all modern deltas are forming under conditions of eustatic sea level rise has not been addressed by the classifications schemes. Elliott (1989) summarizes the present state of delta research, and notes that there is increasing evidence that shows that many ancient delta systems are noticeably different from any modern delta, and that non-actualistic delta models may be needed to describe them. If sufficient data are available, models may be created that describe distinctive delta types that are dissimilar in many aspects from modern deltas.

The Lea Park - Belly River transition (Campanian) of central Alberta provides a possible vehicle for further study of these problems. These sediments were deposited at the base of a large prograding wedge of sediment within a foreland basin, and previous work (Wasser, 1988; Storey, 1982) indicates that the succession contains the deposits of deltaic systems. These sediments have been extensively drilled and logged in the search for hydrocarbons, and there is an abundant core data base as well. The excellent data base allows for the development of detailed facies relationships as well as three-dimensional sediment body geometries for deltas in a foreland basin setting.

1.3: Allostratigraphy

Stratigraphers have traditionally subdivided stratigraphic successions based on either lithostratigraphy or biostratigraphy. Recently, however, sedimentologists and stratigraphers have begun to attempt to subdivide stratigraphic successions in a manner that allows for a grouping of genetically associated depositional systems, without regard for aspects such as lithology.

One method for such stratigraphic subdivision is allostratigraphy. The North American Committee on Stratigraphic Nomenclature (1983) defines allostratigraphy as the correlation of rock strata units defined on the basis of;

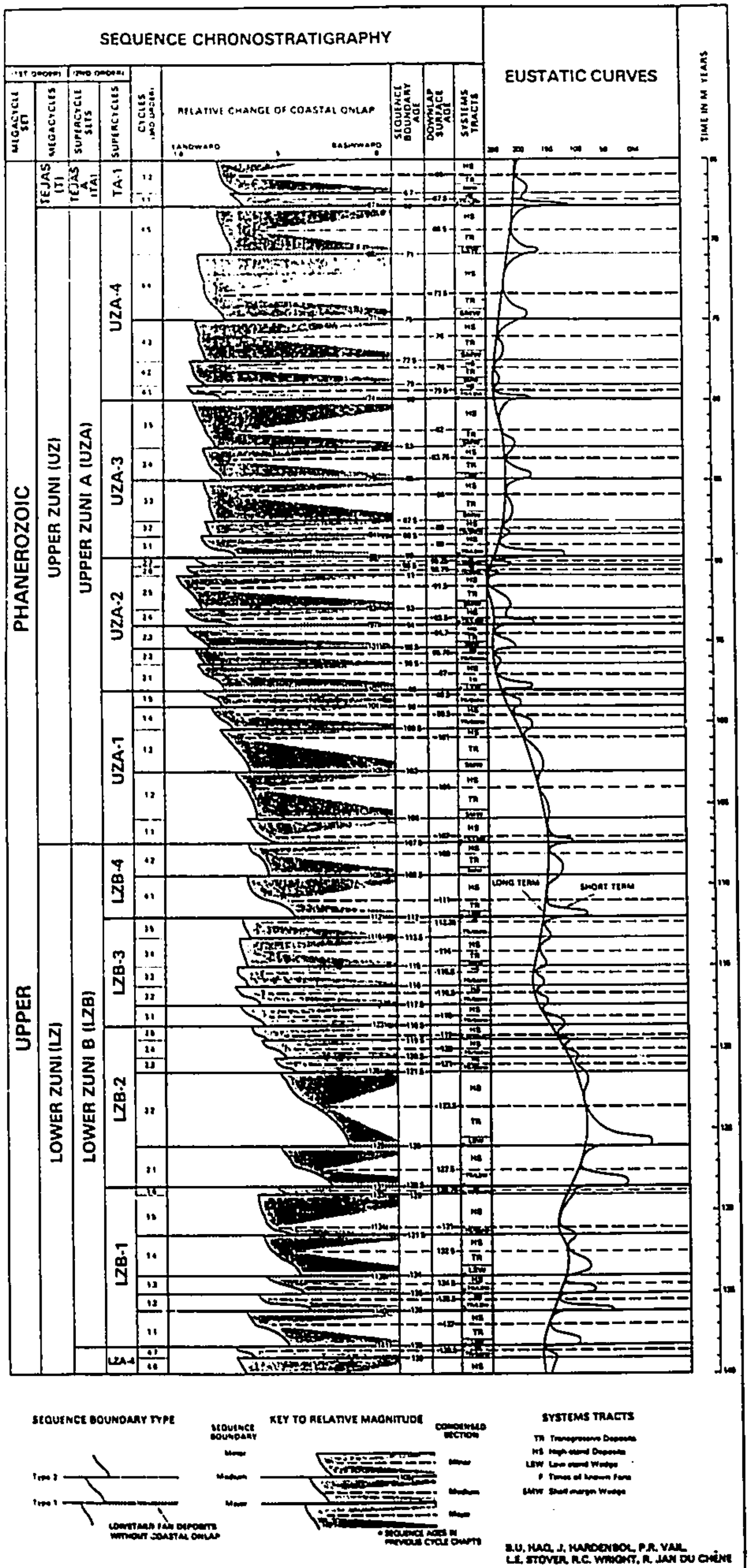
> *"superposed discontinuity bounded units ... internal characteristics may vary laterally and vertically, but the unit boundaries are laterally traceable discontinuities."*

By definition, allostratigraphy is simply a means of formalizing rock units in a non-interpretive manner. The importance of the bounding discontinuities can subsequently be analyzed for interpretive purposes. In most cases, these bounding discontinuities can be viewed as being geologically time-synchronous surfaces created as a result of rapidly changing conditions of sedimentation. The dominant mechanism for creation of large-scale bounding discontinuities in sedimentary successions is thought to be fluctuations of relative sea level.

In recent years, the concepts of how relative sea level fluctuations affect deposition has been the topic of much interest. In order to examine this problem, ancient sedimentary deposits must be examined and arranged in a stratigraphic framework which allows for the identification of relative sea level fluctuations by identifying the deposits characteristic of rising and/or falling sea level. Allostratigraphy allows the rocks to be correlated in such a way as to facilitate interpretation of the effects of relative fluctuations in sea level. Using regional correlations of bounding discontinuities one can ideally reconstruct the history of relative sea level fluctuation within the basin at a given time. There are, however, limitations to using allostratigraphy to subdivide a sedimentary succession. Bounding discontinuities often cannot easily be traced into successions of non-marine or deep marine sediments, and therefore any stratigraphic framework proposed for laterally equivalent coastal and "shallow" marine sediments cannot be correlated into the non-marine or deep marine sections.

Much of the recent interest in determining the history of sea level fluctuations in sedimentary basins can be related to the development of seismic stratigraphy by Peter Vail and his colleagues at the Exxon Production Research Co.. The basic concepts of seismic stratigraphy are reviewed in Vail et al. (1977). Seismic stratigraphy involves identifying distinct seismic reflection horizons which appear to represent erosional unconformities, and are interpreted to have been created during relative drops in sea level. These unconformities are the basis of correlation for sedimentary successions. The data used to develop the concepts of seismic stratigraphy consists mostly of seismic data collected from passive margin sedimentary basins around the world (Vail et al., 1977). These concepts have led to the development of the Exxon sea level fluctuation/coastal onlap charts, which record global fluctuations of sea level. The latest of these is shown in Haq et al. (1988) (Fig. 1.2). These charts are based on information (much of it unpublished) which suggests that, in a broad sense, the timing of many regional unconformities is

Figure 1.2: Exxon third order eustatic sea level curve and changes in relative coastal onlap for the Cretaceous (from Haq et al., 1988).


SEQUENCE CHRONOSTRATIGRAPHY
MEGACYCLE SET
MEGACYCLES
SUPERCYCLE SETS
SUPERCYCLES
CYCLES
RELATIVE CHANGE OF COASTAL ONLAP
LANDWARD
BASINWARD
SEQUENCE BOUNDARY AGE
DOWNLAP SURFACE AGE
SYSTEMS TRACTS
EUSTATIC CURVES
TIME IN M YEARS
PHANEROZOIC
UPPER
UPPER ZUNI (UZ)
LOWER ZUNI (LZ)
TEJAS (T)
TEJAS A (TA)
UPPER ZUNI A (UZA)
LOWER ZUNI B (LZB)
TA-1
UZA-4
UZA-3
UZA-2
UZA-1
LZB-4
LZB-3
LZB-2
LZB-1
LZA-4
LONG TERM
SHORT TERM
SEQUENCE BOUNDARY TYPE
Type 2
Type 1
LOWSTAND FAN DEPOSITS WITHOUT COASTAL ONLAP
SEQUENCE BOUNDARY
KEY TO RELATIVE MAGNITUDE
CONDENSED SECTION
Minor
Medium
Major
SYSTEMS TRACTS
TR Transgressive Deposits
HS High-stand Deposits
LSW Low-stand Wedge
F Times of known Fans
SMW Shelf-margin Wedge
B.U. HAQ, J. HARDENBOL, P.R. VAIL,
L.E. STOVER, R.C. WRIGHT, R. JAN DU CHÊNE

similar in sedimentary basins around the world, indicating that the controlling sea level fluctuations may be global in extent (Vail et al., 1977).

Further work by the Exxon Production Research Co. has led to the development of *sequence stratigraphy* (Posamentier et al., 1988; Van Wagoner et al., 1990). Sequence stratigraphy takes the principles of seismic stratigraphy and uses them to interpret the rock record in outcrop or in the subsurface. The concepts used in sequence stratigraphy are not new. They can be traced back to the sequence concepts of Sloss (1963) and the time-rock stratigraphic concepts of Wheeler (1958). These authors were in turn influenced by work on unconformities by Blackwelder (1909). These early works dealt with separating stratigraphic units on the basis of laterally continuous unconformities, and the importance of unconformities in constructing a time-rock stratigraphy of a succession of sediments. The unifying concept behind these ideas is that sediment packages can be grouped on the basis of the regional unconformities that separate them. These unconformities record major fluctuations in the spatial position of base level, with the underlying concept that sediments within a "*sequence*", or between two unconformities, are genetically related.

Applications of the concepts of sequence stratigraphy to intracontinental sedimentary basins have begun to appear in the last 5 years, especially within foreland basins. A number of recent studies have focused on the deposits of the Alberta foreland basin, in which sea level cyclicity is very rapid (cycles on a scale of 100,000-400,000 years), and may be more influenced by mechanisms other than eustatic sea level fluctuations, such as tectonic activity. Examples of such work include studies on the Cardium Formation of Alberta (Leggitt et al., 1990; Plint et al., 1986, 1987), the Dunvegan Formation (Bhattacharya, 1989; Bhattacharya and Walker, 1991a, 1991b), and the Viking Formation (Boreen and Walker, 1991; Power, 1988; Downing and Walker, 1988). Even including these studies, applications of sequence stratigraphic ideas to sediments within foreland basins is still at an early stage.

One of the areas of sequence stratigraphy that requires more in-depth research is the nature of high-frequency relative sea level fluctuations. Much of the Exxon work has focused on third-order sea level fluctuation cycles (their terminology) of several million years in duration (average duration of 1.8 my during the Cretaceous (R. Walker, pers. comm.)). Sediments deposited during longer term (third-order) falls or rises in relative sea level will, in many cases, contain evidence of sediments deposited as a result of short-term, higher frequency relative fluctuations in sea level. Due to increased rates of subsidence which allow for greater sediment aggradation, foreland basins may have a better chance of preserving the deposits of these higher-frequency fluctuations. An example of this is the Cretaceous Muskiki-Marshybank Formations of Alberta, which contain several high-frequency cycles of relative sea level fluctuation with durations of approximately 100,000 years (Plint, 1991).

The sediments of the Lea Park - Belly River transition provide a useful vehicle in which to apply the concepts of allostratigraphy to help determine the effects of high-frequency relative sea level fluctuations in a foreland basin. Higher-frequency fluctuations of sea level (fourth-order) superimposed on a longer-term drop in sea level will have a significant effect on how coastal systems prograde into the basin. Analysis on a fourth-order scale that will allow facies relationships to be integrated into allostratigraphic relationships, and help to understand the effect of sea level fluctuations on coastal depositional systems.

1.4: Primary Questions of this Study

(1) What are the nature of the facies relationships and depositional environments of the sediments in the Lea Park - Belly River transition?

(2) Are coastal systems within the Belly River Formation deltaic? If so, can these deltas be adequately described and classified by present schemes?

(3) What are the allostratigraphic relationships of sediments of the Lea Park - Belly River transition?

(4) What are the effects of small-scale relative sea level fluctuations on the nature of coastal systems?

- Are prograding shorefaces gradationally or sharply-based?
- To what degree do fluvio-distributary channels incise during small-scale drops in sea level?

- Can the effects of small-scale sea level fluctuations be recognized within the non-marine portions of a coastal plain? If so, how are they expressed.

(5) What factors may be controlling the large-scale and small-scale fluctuations in relative sea level which took place during the deposition of the sediments of the Lea Park - Belly River transition?

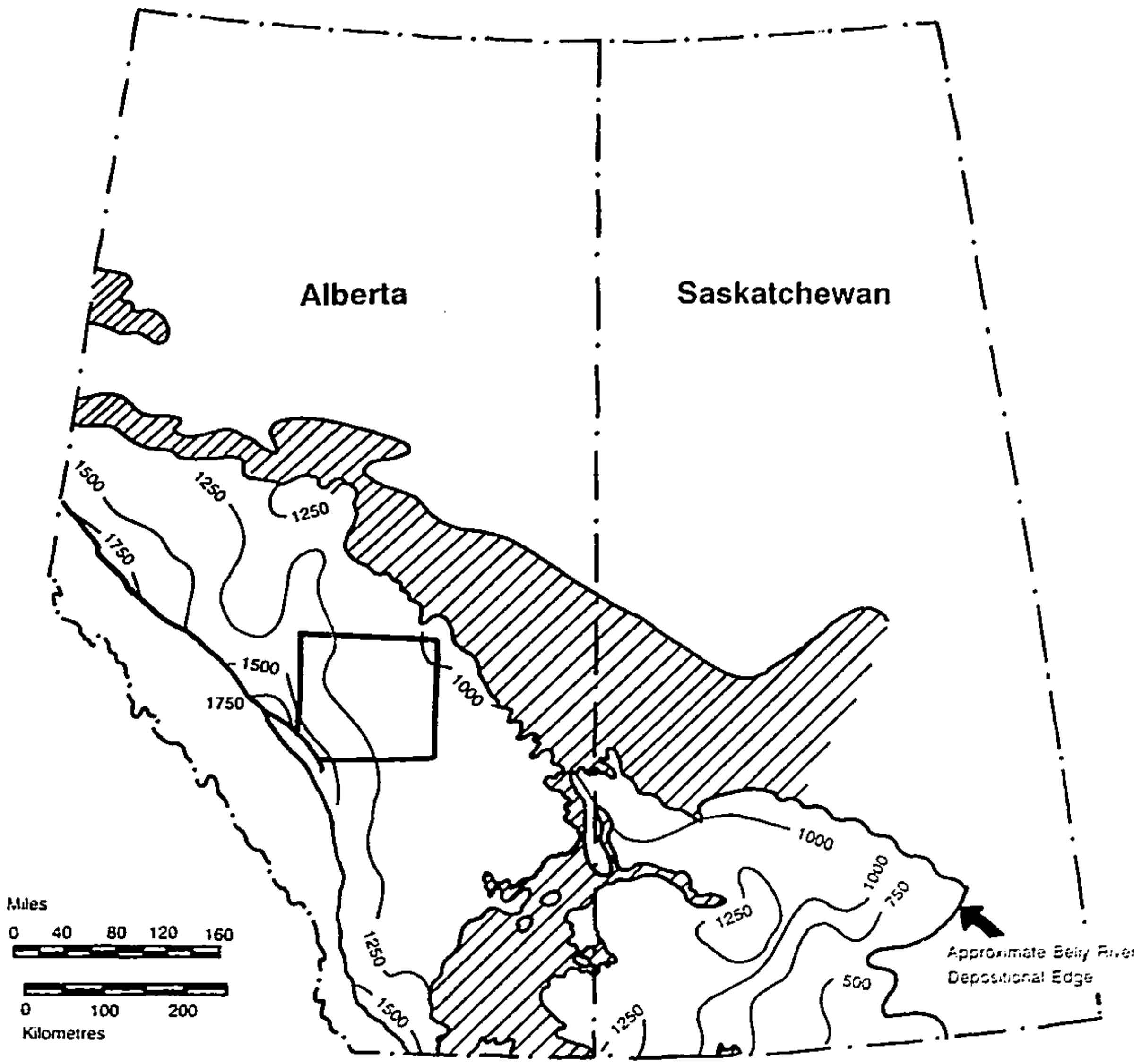
(6) Can Exxon sequence stratigraphy be applied to the allostratigraphic framework of the Lea Park - Belly River transition? Is sequence stratigraphy as it presently exists the best way to divide the sediments of the Lea Park - Belly River transition for sedimentological study?

CHAPTER 2: THE LEA PARK AND BELLY RIVER FORMATIONS

2.1: Definition, Distribution and Stratigraphy

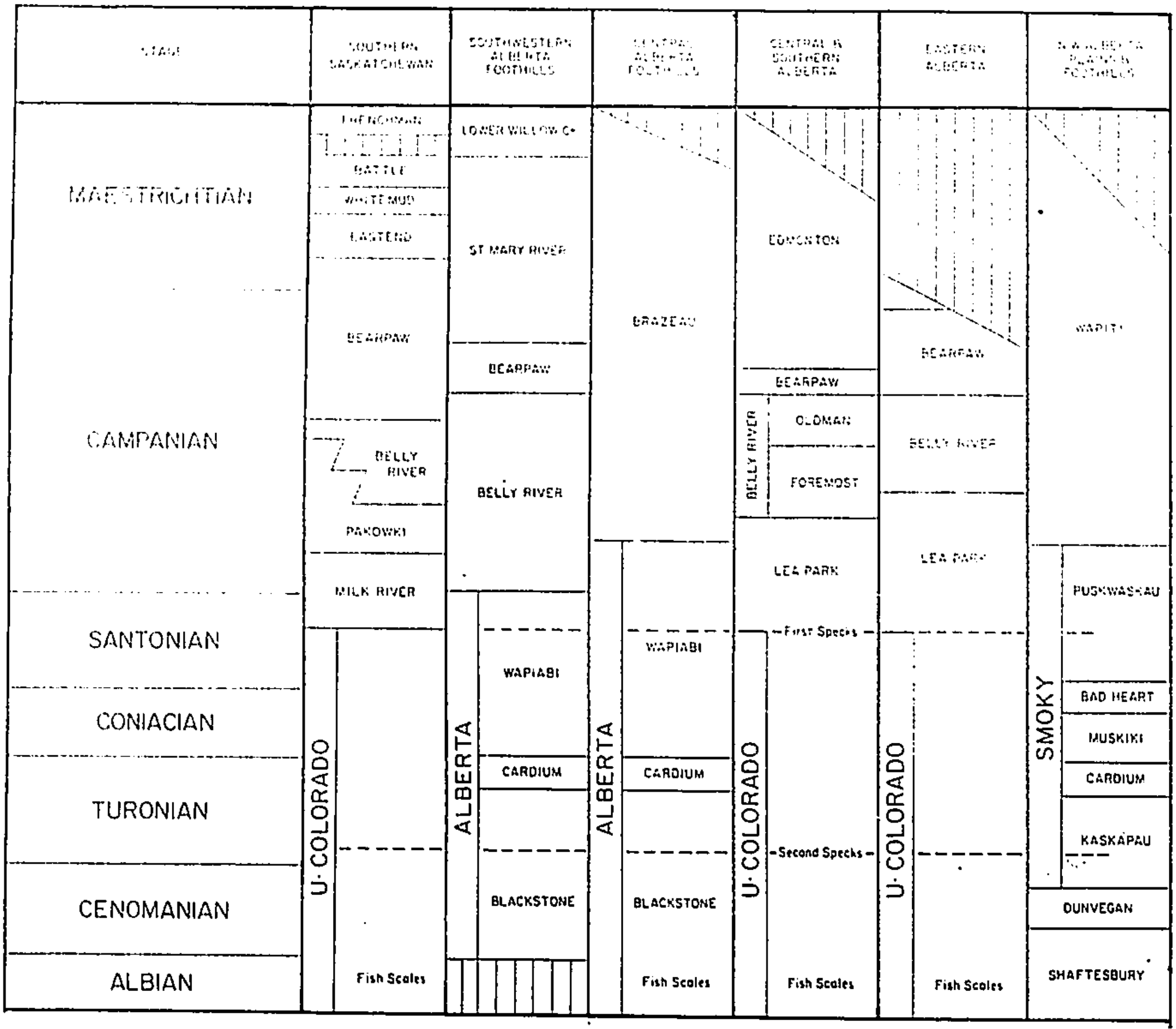
The transition between the Upper Cretaceous (Campanian) Lea Park Formation and the Belly River Formation consists of a series of interbedded marine and non-marine sandstones, siltstones, and mudstones, that were first described by Dawson (1883). The name Belly River was first used by Dawson (1883) to describe the rocks of the "Belly River series" exposed along the Belly River (now Oldman River) from Lethbridge, Alberta downstream. The Belly River Formation and its equivalents have since been determined to be widespread throughout western North America. Figure 2.1 shows a gross isopach map of the combined thickness of the Lea Park and Belly River Formations and their equivalents in western Canada. The Belly River Formation crops out in the Foothills of the Rocky Mountains, and within the badlands of southeastern Alberta and northern Montana. In the subsurface, it is present in central Alberta, northern Montana, and southwestern Saskatchewan, where it reaches its eastern depositional edge. These sediments are thickest in western Alberta, where sediments of the Belly River Formation or its equivalents can be just over 1 km thick. The sediment wedge thins in an easterly direction, eventually pinching out in southwestern Saskatchewan (Stott, 1984; McLean, 1971).

Figure 2.1: Gross isopach map of the combined thickness of the Lea Park and Belly River Formations in Alberta. Contours are in feet. Hatched areas denote Lea Park or Belly River outcrop. Location of this study area is outlined by thick black box (modified from Williams and Burk, 1964).


Alberta
Saskatchewan
1500
1250
1250
1750
1500
1750
1000
1000
1000
750
1250
500
1250
1250
1500
Approximate Belly River Depositional Edge
Miles
0 40 80 120 160
0 100 200
Kilometres

The lithostratigraphy of these sediments can be fairly confusing, as there is more than one set of names for these rocks. Figure 2.2 shows the lithostratigraphic correlation of Upper Cretaceous sediments in Western North America. The sediments underlying the Belly River Formation and overlying the Milk River Formation consist of marine mudstones. They are known as the Nomad Member of the Wapiabi Formation in outcrop of the Foothills of the Rocky Mountains in Alberta, and the Pakowki Formation in outcrop in southern and eastern Alberta. In northern Montana, these sediments comprise the Claggett Formation (McLean, 1971). In the subsurface of central Alberta, these sediments are known as the Lea Park Formation. Terminology of the wedge of dominantly non-marine sedimentary rocks overlying the Lea Park/Pakowki/ Nomad Formation is equally variable. In the Foothills outcrops of central and southern Alberta, the sediments are assigned to the Belly River Formation (Brazeau Group). In the northern Foothills, the Brazeau Group is downgraded to formation status within the Saunders Group. In the northern plains the Brazeau Group is known as the Wapiti Formation. In southeastern Alberta, western Saskatchewan, and northern Montana, the Belly River Group is known as the Judith River Formation. In southern Montana these sediments are assigned to the Two Medicine Formation. In the southern plains, the Belly River Group is often broken down into the Foremost (lower) and Oldman (upper) Formations. The Foremost Formation is also informally known as the Basal Belly River Formation, and represents the dominantly sandy sediments which

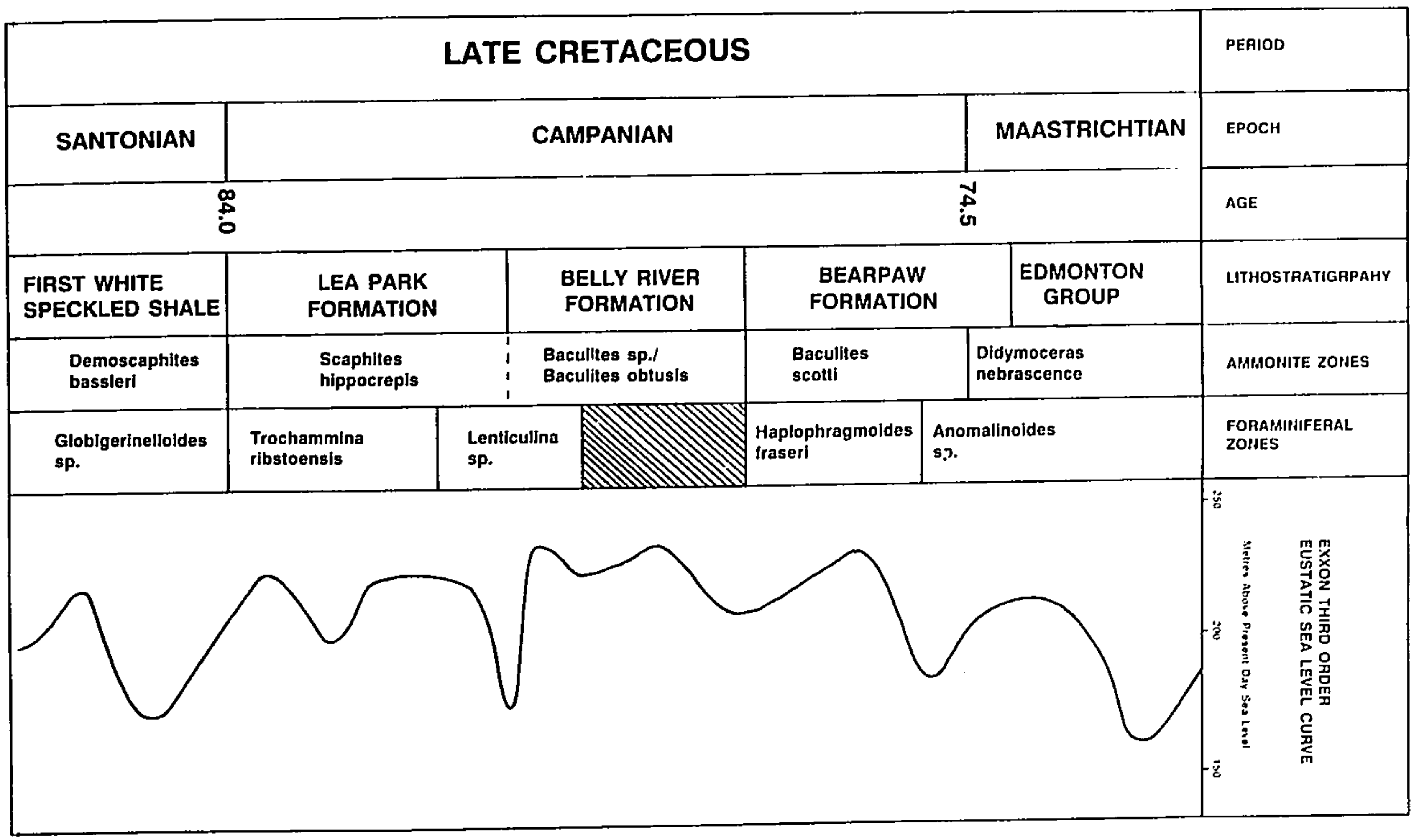
Figure 2.2: Cretaceous lithostratigraphy of the Canadian interior plains (modified from Williams and Burk, 1964).


STAGE
SOUTHERN SASKATCHEWAN
SOUTHWESTERN ALBERTA FOOTHILLS
CENTRAL ALBERTA FOOTHILLS
CENTRAL & SOUTHERN ALBERTA
EASTERN ALBERTA
MAESTRICHTIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
CENOMANIAN
ALBIAN
FRENCHMAN
BATTLE
WHITEMUD
EASTEND
BEARPAW
BELLY RIVER
PAKOWKI
MILK RIVER
U. COLORADO
Fish Scales
LOWER WILLOW C.
ST MARY RIVER
BEARPAW
BELLY RIVER
ALBERTA
WAPIABI
CARDIUM
BLACKSTONE
BRAZEAU
ALBERTA
WAPIABI
CARDIUM
BLACKSTONE
Fish Scales
EDMONTON
BEARPAW
BELLY RIVER
OLDMAN
FOREMOST
LEA PARK
First Specks
U. COLORADO
Second Specks
Fish Scales
BEARPAW
BELLY RIVER
LEA PARK
U. COLORADO
Fish Scales
WAPITI
SMOKY
PUSKWASKAU
BAD HEART
MUSKIKI
CARDIUM
KASKAPAU
DUNVEGAN
SHAFTESBURY

are transitional between the Lea park and the Belly River. In the subsurface of the central plains, the Belly River Group is again downgraded to formation status, and is referred to as either the Belly River or the Judith River Formation. Overlying the sediments of the Belly River Group in central and southern Alberta are marine mudstones and shales of the Bearpaw Formation (Williams and Burk, 1964). This study deals dominantly with the subsurface of central Alberta, and to avoid confusion, the sedimentary rocks in question will be referred to throughout the rest of the text as the Lea Park Formation and the Belly River Formation.

The biostratigraphy of the uppermost Lea Park and lowermost Belly River Formations is uncertain. Figure 2.3 shows the biostratigraphy of the Upper Cretaceous in central Alberta. Both the Lea Park and the lower portions of the Belly River Formation are thought to be early-mid Campanian in age. The base of the Campanian epoch in western North America is placed at the boundary between the ammonite zone *Demoscaphites bassleri* Reeside and the overlying *Scaphites hippocrepis* (DeKay) zones (Obradovich and Cobban, 1975). The Lea Park and Belly River Formations were deposited during the *S. hippocrepis* zones and the overlying *Baculites sp.* (weak flank ribs) and *Baculites obtusus* zones, but the precise time of transition between the Lea Park and Belly River is unknown, as information on the *S. hippocrepis* zones is sparse (Obradovich and Cobban, 1975). Jeletsky (1968) correlated the Oldman and Foremost Formations in southern Alberta with the *Baculites gregoryensis* Zone; however,

Figure 2.3: Chart showing correlation of chrono-, litho-, and biostratigraphy of the Upper Cretaceous in central Alberta. The Exxon third order eustatic sea level curve is also correlated to the sediments. Placement of ages is taken from Haq et al., 1988. The ammonite biostratigraphy is taken from Obradovitch and Cobban, 1975, and the foraminiferal biostratigraphy is taken from Caldwell et al., 1978.



PERIOD
EPOCH
AGE
LITHOSTRATIGRPAHY
AMMONITE ZONES
FORAMINIFERAL ZONES
EXXON THIRD ORDER EUSTATIC SEA LEVEL CURVE
Metres Above Present Day Sea Level
250
200
150
LATE CRETACEOUS
SANTONIAN
CAMPANIAN
MAASTRICHTIAN
84.0
74.5
FIRST WHITE SPECKLED SHALE
LEA PARK FORMATION
BELLY RIVER FORMATION
BEARPAW FORMATION
EDMONTON GROUP
Demoscaphites bassleri
Scaphites hippocrepis
Baculites sp./ Baculites obtusis
Baculites scotti
Didymoceras nebrascence
Globigerinelloides sp.
Trochammina ribstoensis
Lenticulina sp.
Haplophragmoides fraseri
Anomalinoides sp.

this zone most likely applies to sediments within the upper portions of the Belly River Formation. In the foraminiferal zonation scheme of Caldwell et al. (1978), the Lea Park Formation and the basal sediments of the Belly River Formation both belong to the *Lenticulina* sp. zone.

The age of the transition between the Lea Park and the Belly River is difficult to determine for a number of reasons. The transition between the two formations is diachronous throughout the Western Canada Basin, becoming gradually younger in an eastward direction. Also, as mentioned previously, the biostratigraphic relationship of the transition is poorly understood. However, some age relationships for early Campanian sediments of the Western Interior Seaway have been determined. Obradovich and Cobban (1975) date the Santonian/Campanian boundary at 82.5 my (+/- 1 my). The Exxon sea level curve dates the Santonian-Campanian boundary at 84.5 my (Haq et al., 1988). Russell (1970) reported an age of 87.4 my for a bentonite near the base of the Pakowki Formation, but this age is anomalously old and thought to be incorrect (McLean, 1971; Thomas et al., 1990). Thomas et al. (1990) have determined an age of 74-75 my for the uppermost sediments of the Judith River Formation (Belly River Formation) in eastern Alberta. This gives a time span of 7.5-9.5 my for the deposition of the Lea Park and the entire Belly River Formation. These ages were obtained using radiometric dating techniques of bentonite beds. Goodwin and Deino (1989) have reported an age of approximately 78 my for two bentonites within and just above the Taber Coal Zone in northern Montana.

According to McLean (1971), the Taber Coal Zone defines the boundary between the Oldman and Foremost Formations. Thus, there is a period of 4-6 my for the deposition of the Lea Park Formation and the basal sediments of the Belly River Formation. Assuming that most of this time involved slow deposition of marine mudstones of the Lea Park Formation, it is probable that, within the study area, deposition of the sediments representing the transition between the Lea Park and the Belly River Formations occurred over a time span of less than 2 my.

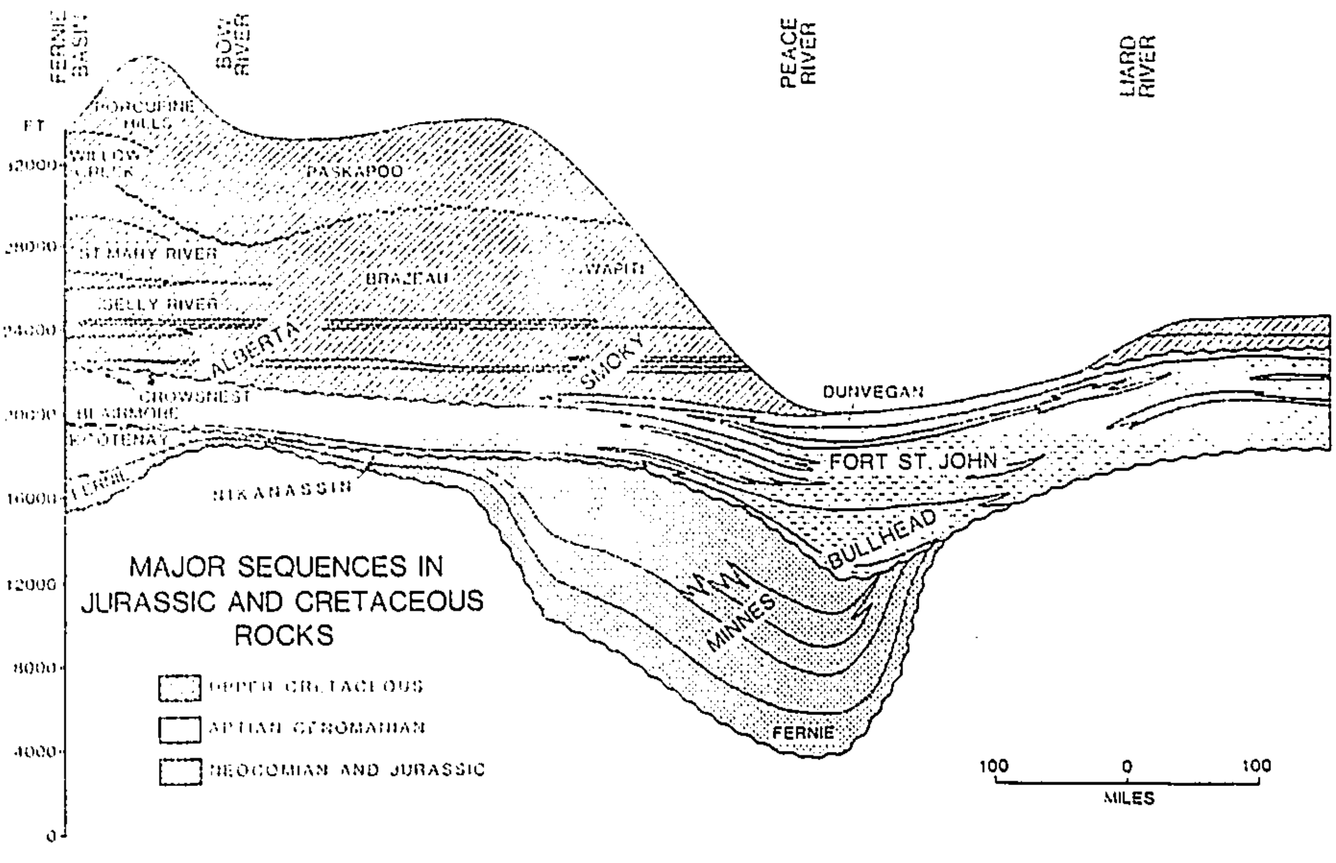
Correlation of the base of the Belly River Formation with global transgressive/regressive cycle charts or global sea level curves is somewhat uncertain. The Exxon sea level curves of Haq et al. (1988) reveal a relatively large third order drop in sea level at 80 my (Fig. 1.2, 2.3). The timing of initiation of Belly River deposition in central Alberta is constrained between 82.5 and 78 my, so this third-order drop of sea level may be coincident with the initiation of Belly River deposition (Fig. 2.3). If the 80 my sea level drop of Haq et al. (1988) does correspond to the initiation of Belly River deposition in Alberta, correlation with the transgressive-regressive scheme of Kauffman (1984) is uncertain, as 80 my falls between the R_7 or R_8 regressions according to this scheme.

2.2: Tectonic Setting and Regional Paleogeography

During the Cretaceous Period, a broad, inland sea covered much of what is now western North America from the Arctic Ocean to the Gulf of Mexico. This inland sea was flanked on the west by the tectonically active Cordillera, which provided large quantities of detrital sediment to the seaway, and to the east by the lower relief Precambrian shield, which provided little sediment input. The tectonically active Cordillera to the west was a result of the transformation of the western continental margin of North America during the late Jurassic from a passive margin to an active margin. The accretion of allochthonous terranes resulted in crustal shortening and loading, which produced a foreland basin to the east of the Cordillera. The sediments of the Lea Park and Belly River Formations were deposited in this foreland basin.

The two main periods of orogeny associated with the development of the Cordillera and the Alberta foreland basin were the Columbian Orogeny, which occurred in two stages during the Late Jurassic to early Late Cretaceous, and the Laramide Orogeny, which occurred during the Late Cretaceous to Early Tertiary (Stott, 1984). These major periods of orogeny are generally thought to be correlative with the deposition of major clastic wedges within the foreland basin. Stott (1984) recognizes three main clastic wedges within the Alberta basin (Fig. 2.4). The first clastic wedge consists of the Jurassic Fernie Formation and the Late Jurassic-earliest Cretaceous Kootenay and Minnes Groups, which correlates to the first stage of the Columbian Orogeny. The

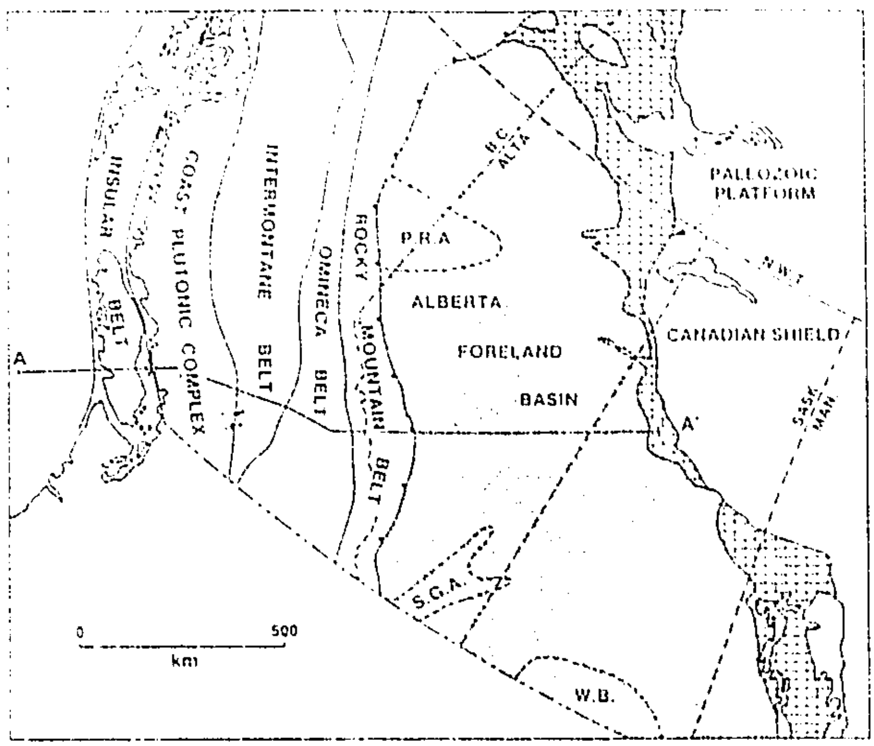
Figure 2.4: Schematic diagram illustrating the three main clastic wedges within Jurassic and Cretaceous rocks in the Rocky Mountains and Foothills in south-central Alberta (from Stott, 1984).


FERNIE BASIN
BOW RIVER
PEACE RIVER
LIARD RIVER
FT
PORCUPINE HILLS
WILLOW CREEK
PASKAPOO
ST MARY RIVER
BRAZEAU
WAPITI
BELLY RIVER
ALBERTA
SMOKY
CROWSNEST
BLAIRMORE
KOOTENAY
FERNIE
NIKANASSIN
DUNVEGAN
FORT ST. JOHN
BULLHEAD
MINNES
FERNIE
16000
12000
8000
4000
0
MAJOR SEQUENCES IN JURASSIC AND CRETACEOUS ROCKS
UPPER CRETACEOUS
APTIAN CENOMANIAN
NEOCOMIAN AND JURASSIC
100
0
100
MILES

second wedge comprises the Late Neocomian to Early Cenomanian Blairmore, Bullhead, and Fort St. John Groups and the Dunvegan Formation, which correlate to the second stage of the Columbian Orogeny. The final wedge consists of the Late Cenomanian to Early Tertiary Alberta and Smoky Groups, as well as a thick succession of dominantly continental clastic sediments, including the Lea Park and the Belly River Formations. This wedge is correlative with the Laramide Orogeny (Stott, 1984). Cant and Stockmal (1989) and Stockmal et al. (in press) have studied the relationship between terrane accretion and sediment deposition in the Foreland in more detail. Cant and Stockmal (1989) suggest that the accretion of the Insular superterrane (Fig. 2.5) during the Late Cretaceous directly initiated deposition of the Belly River Formation. Stockmal et al. (in press) slightly revise this correlation, stating that the accretion of the Insular superterrane was in fact over by the end of the Santonian stage, but that compressive tectonic stresses associated with this accretion persisted through to the Campanian. This caused a rapid period of overthrust loading which flexurally downwarped the basin, allowing for the accommodation of the Belly River Formation sediments.

Sediments of the Belly River Formation are therefore probably related either directly or indirectly to the accretion of the Insular superterrane during the late Cretaceous, and to the resultant development of the Laramide Orogeny in the Cordillera. Sediments deposited on the western flank of the basin were involved in uplift and erosion associated with the later stages of the Laramide

Figure 2.5: Regional map showing the Western Canada Basin and the Canadian Cordillera, showing tectonic subdivisions and features. Cant and Stockmal (1989) suggest that the accretion of the Insular Superterrane in the late Cretaceous may be correlative with the Laramide orogeny in the Cordillera and the deposition of the Belly River Formations sediments in the Alberta Basin (from Cant and Stockmal, 1989).


INSULAR BELT
COAST PLUTONIC COMPLEX
INTERMONTANE BELT
OMINECA BELT
ROCKY MOUNTAIN BELT
P.R.A
ALBERTA
FORELAND
BASIN
PALEOZOIC
PLATFORM
CANADIAN SHIELD
N.W.T
SASK
MAN
B.C.
ALTA
S.G.A.
W.B.
A
A'
0
500
km

Orogeny, and are not always preserved. Uplift and erosion associated with the Sweetgrass Arch in Montana has exposed and eroded the sediments of the Belly River Formation to the south.

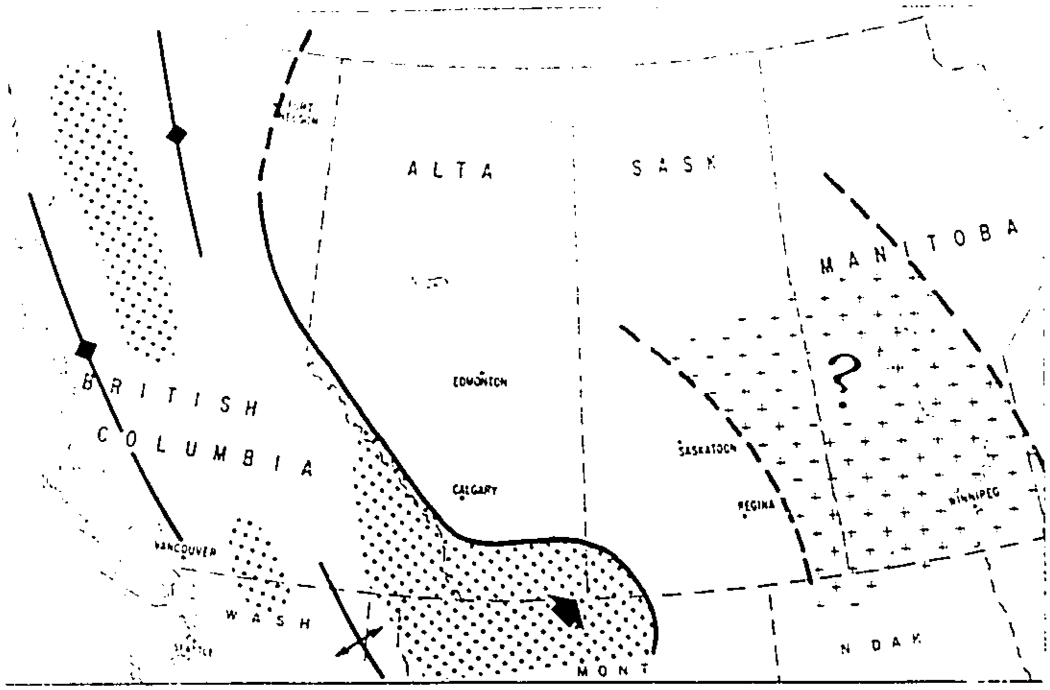
The paleogeography of the Cretaceous Western Interior Seaway (CWIS) during Lea Park and Belly River time has been studied by Williams and Burk (1964) and by Stott (1984). During Lea Park time, the seaway was open to the north and to south, and marine waters covered most of Alberta as far west as the present-day B.C. border (Fig. 2.6). During early to mid-Belly River time (Fig. 2.7) the seaway retreated to the southeast, so that most of present-day Alberta was subaerially exposed. The seaway continued to retreat throughout late Belly River time (Fig. 2.8), before transgression in the Upper Campanian re-established the seaway in southeastern Alberta, and allowed for deposition of the sediments of the Bearpaw Formation (Fig. 2.9).

2.3: Previous Work

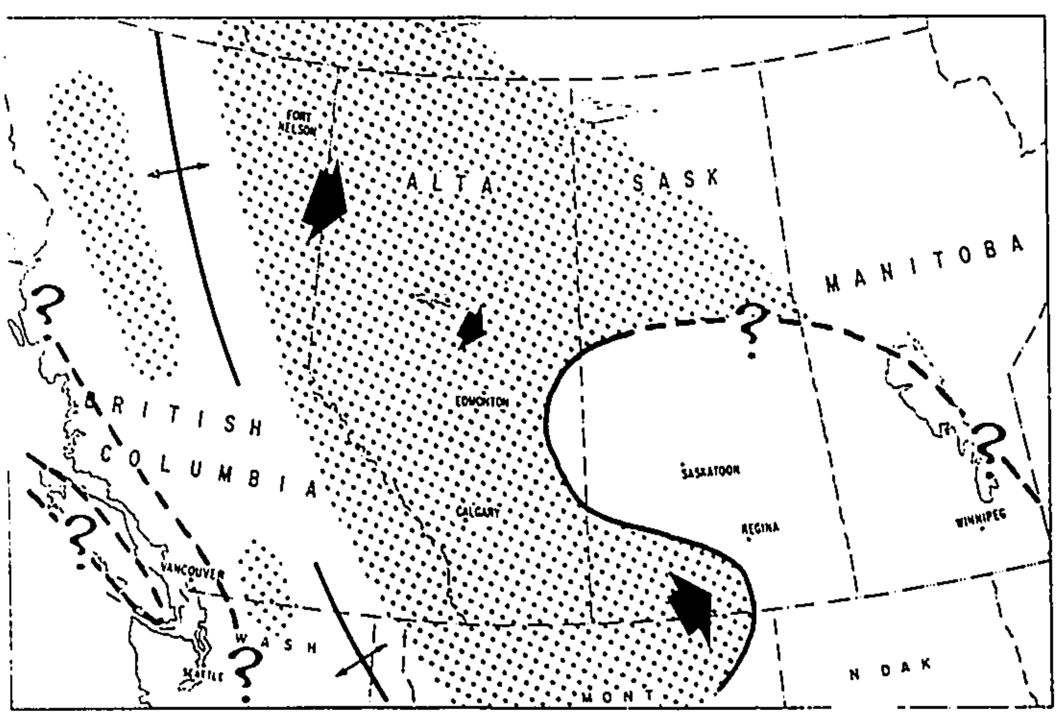
The sediments of the Belly River and the Lea Park Formations have been studied for over 100 years, since the initial work of Dawson (1883), who introduced the term "Belly River series". Since this time, these sediments have been studied by numerous workers, and a complete historical review of the early literature is given in McLean (1971). The reader is referred to this study for more information regarding these early studies. The following discussion

Figures 2.6A, 2.6B: Paleogeography of the Cretaceous Western Interior Seaway in Alberta during Lea Park and Belly River time. (A) Lea Park time. (B) Early Belly River time. The stippled areas denote regions of non-marine deposition, with the thick black line denoting the inferred position of the paleoshoreline. Directions of sediment input are shown by large arrows (from Williams and Burk, 1964).

(A) LEA PARK TIME

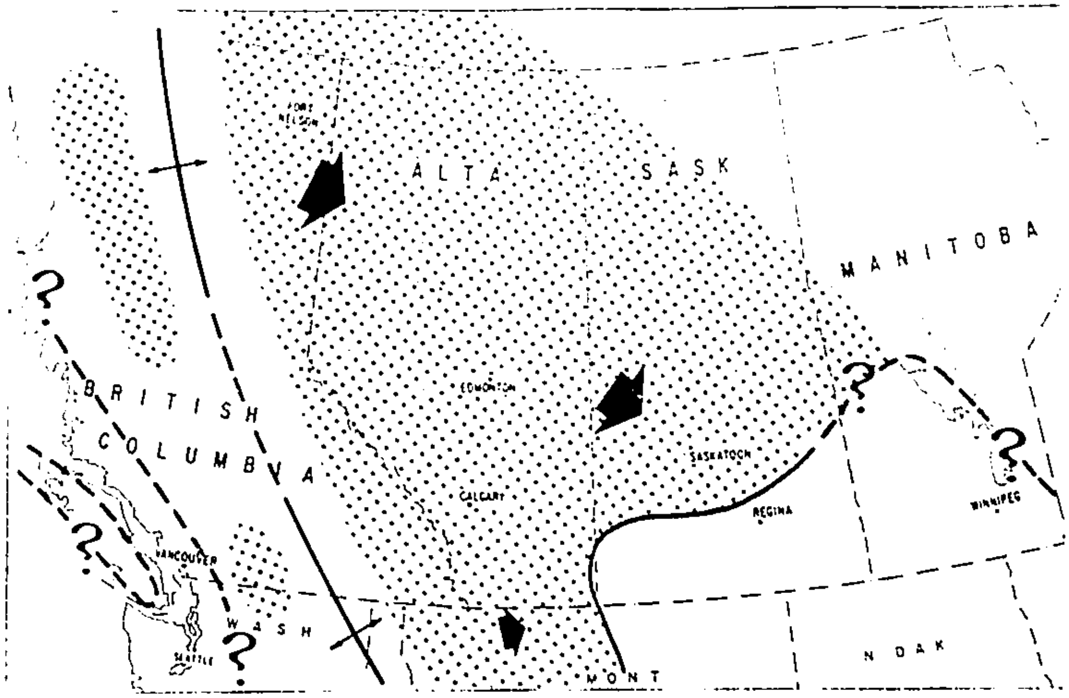


(B) EARLY BELLY RIVER TIME

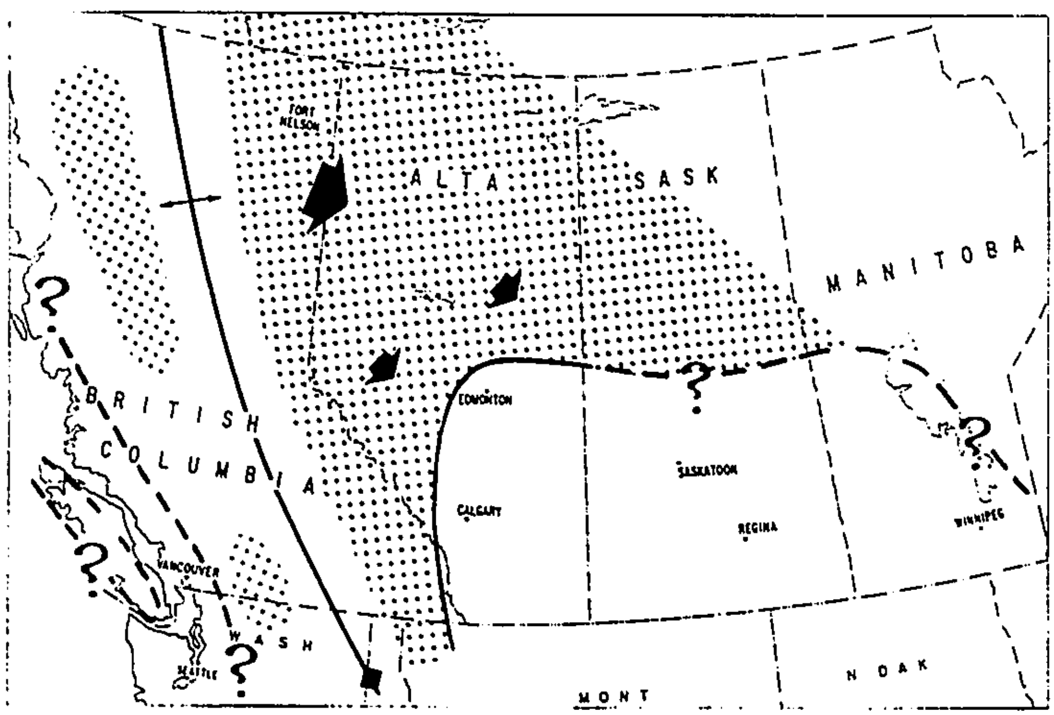


Figures 2.6C, 2.6D: Paleogeography of the Cretaceous Western Interior Seaway in Alberta during Belly River and Bearpaw time. (C) Late Belly River time. (D) Bearpaw time. Stippled areas denote regions of non-marine deposition, with the thick black line denoting the inferred position of the paleoshoreline. Directions of sediment input are shown by large black arrows (from Williams and Burk, 1964).

(C) LATE BELLY RIVER TIME



(D) BEARPAW TIME



will summarize the conclusions of subsequent studies, especially with regards to depositional environments and stratigraphy.

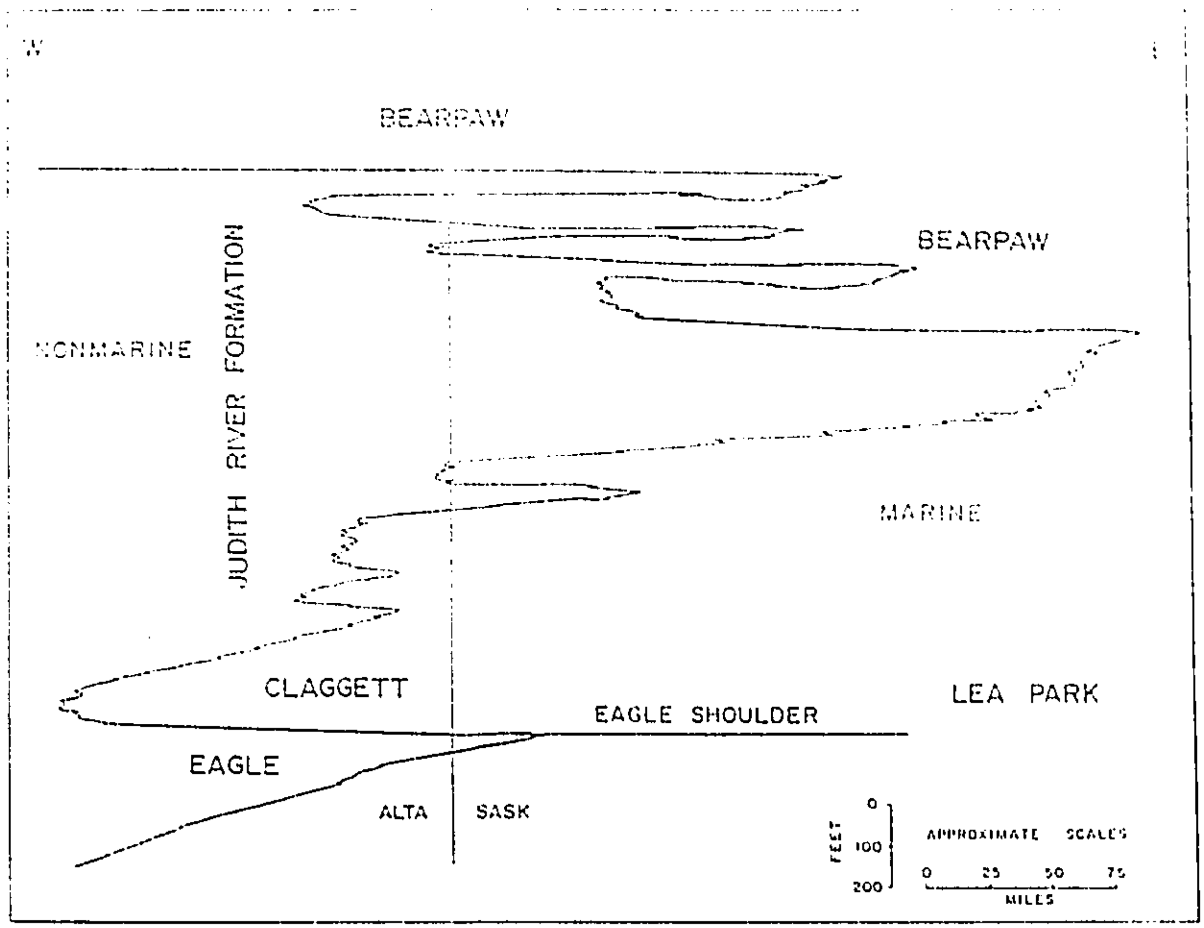
The work of McLean (1971) is the most extensive recent work on the Belly River Formation (Judith River Formation in his terminology). His study was restricted to southeastern Alberta and southwestern Saskatchewan. Most of this work was, however, related to aspects other than the sedimentology of the deposits. He did note, however, that the lowermost sediments of the Judith River Formation were shallow marine to transitional marine/non-marine in nature, and the contact with the underlying Pakowki Formation was diachronous and could be gradational in places (Fig. 2.7)(McLean, 1971).

Ogunyomi and Hills (1977) undertook a regional outcrop study in southeastern Alberta where they interpreted the Foremost Formation to consist of shallow marine sands, barrier island and beach sands, lagoons, salt and freshwater marshes, while the upper Oldman Formation consisted of non-marine fluvial sediments.

Shouldice (1979) examined subsurface reservoir properties of the Belly River Group, but did not deal with sedimentary facies or depositional environments.

Hunter (1980), Bullock (1981), and Haywick (1982) examined the Nomad - Belly River transition along the Highwood River, at Lundbreck Falls, and at Ghost Dam, respectively, in southwestern Alberta. They determined that the uppermost Nomad sediments consisted of a coarsening-upward, storm-

Figure 2.7: Schematic cross section of intertongued marine and non-marine sediments of the Upper Cretaceous in eastern Alberta and western Saskatchewan. Note the diachronous transition between the Claggett/Lea Park and the Judith River Formation (Belly River Formation) (from McLean, 1971).


W
BEARPAW
BEARPAW
NONMARINE
JUDITH RIVER FORMATION
MARINE
CLAGGETT
LEA PARK
EAGLE SHOULDER
EAGLE
ALTA
SASK
FEET
0
100
200
APPROXIMATE SCALES
0 25 50 75
MILES

dominated shelf to shoreface sequence. The basal Belly River sediments sharply overlie this, and were deposited in a fluvial and non-marine environment. Reich (1983) examined the Nomad - Belly River transition at Burnt Timber Creek, also in southwestern Alberta. The uppermost sediments of the Nomad were interpreted as storm-influenced shelf deposits. These are sharply overlain by thin coastal sandstones and mudstones of the Belly River Group, which in turn are overlain by fluvial sandstones and associated sediments. All of these four studies involve detailed facies analysis, but each is restricted to a single outcrop, and do not involve any lateral correlation.

Storey (1982) examined the sediments of the Belly River Formation in the subsurface in the Keystone-Pembina area, and concluded that they represent deposition in a shallow lobate deltaic system. However, the publication is very brief, did not distinguish between facies, and provided almost no data.

Wasser (1988) and Hartling and Wasser (1990) also concluded that the Basal Belly River sediments in the Keystone-Pembina and Ferrybank regions comprised a series of overlapping deltaic lobes. As with Storey (1982), however, these studies did not discuss facies in any detail, and provided very little data.

Iwuagwu and Lerbekmo (1984) related the petrology of Basal Belly River sediments to interpreted depositional environments. They interpreted the sediments to represent a wide variety of environments, including deltaic, prodeltaic, and fluvial floodplain environments. This study did not deal with lateral facies relationships or propose any stratigraphy for the sediments of the Basal Belly River Formation.

Eberth (1990), and Wood (1985, 1989) studied sediments of the Judith River Formation in Dinosaur Provincial Park, in eastern Alberta. Both of these studies interpreted the sediments to have been deposited by a variety of fluvial depositional mechanisms. These studies, however, deal with sediments within the upper portions of the Judith River Formation, and do not deal with sediments transitional with the Pakowki (Lea Park) Formation.

Gardiner et al. (1989) studied the lowermost sediments of the Belly River Formation with regards to depositional environments and reservoir properties in the Peco Field of west-central Alberta. They interpreted the sediments of the uppermost Lea Park Formation to have been deposited in a prograding marine shoreface environment, overlain by transgressive deeper marine sediments. The lowermost sediments of the Belly River Formation consist of braided fluvial sediments which unconformably overlie the Lea Park Formation.

Thomas et al. (1990) and Goodwin and Deino (1989) have studied the geochemistry of bentonites within the Judith River Formation. They have determined radiometric ages for sediments of the Judith River Formation, and these ages are used in this study.

Sabry (1990) studied sandstones of the Basal Belly River Formation in the subsurface of south-central Alberta. He studied these sediments from a lithostratigraphic perspective, and interpreted sandstones to have been deposited in a wide variety of environments, including deltaic, estuarine, tidal channels, tidal flats, tidal sand ridges, and fluvial environments. The area covered by this study is quite large (about 18,000 km^2), but the data base was small (less than 50 cores examined), and little data was presented.

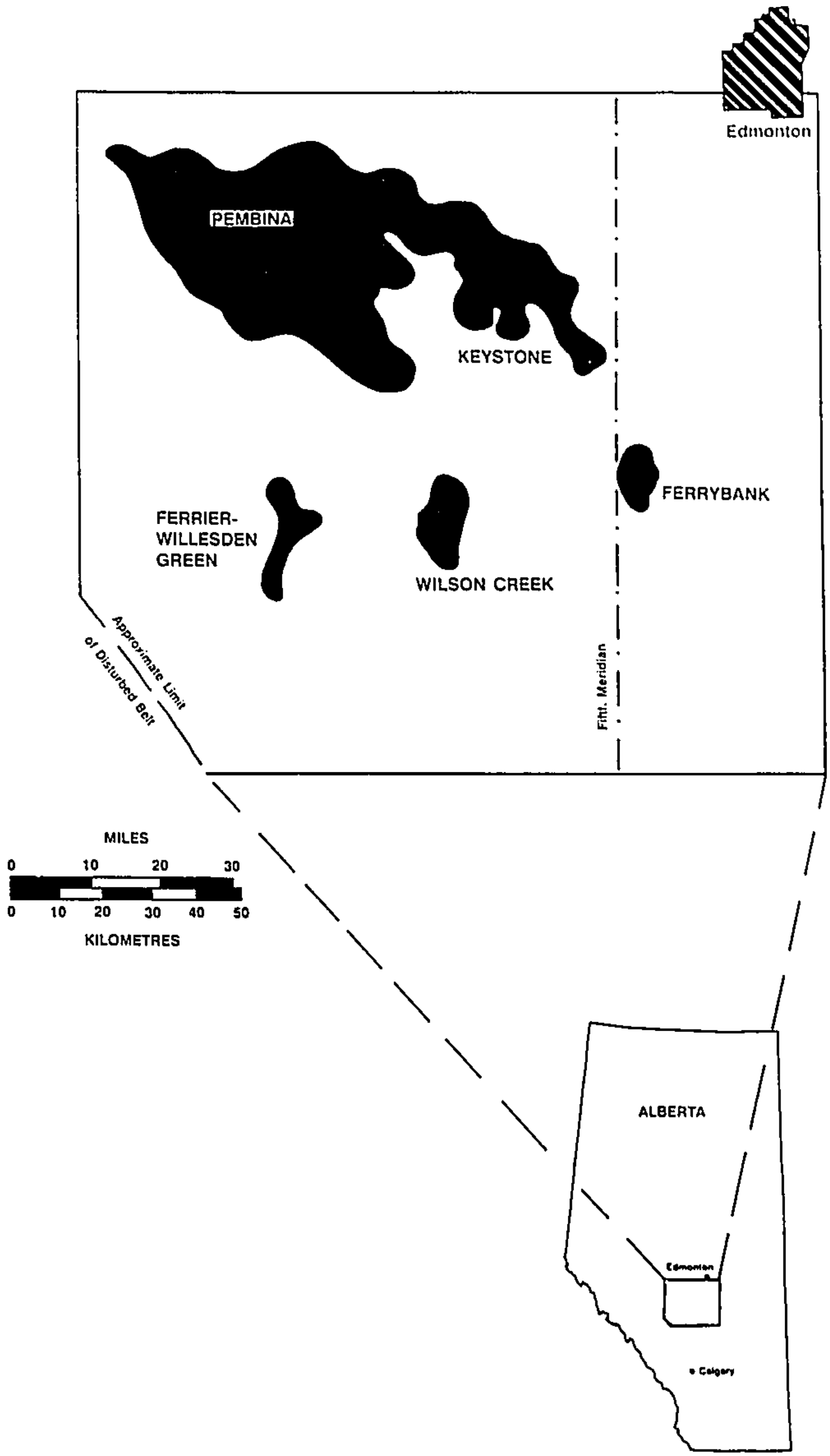
The preliminary conclusion to be drawn from the previous work is that the broad depositional environment of the sediments of the Lea Park - Belly River transition has been determined to be transitional from shallow marine to deltaic/non-marine. The coastal systems are dominated by sand, with lesser amounts of mud and silt. However, there have been no published studies with detailed examinations and three-dimensional correlations of the lowermost sediments of the Belly River Formation in the subsurface of Alberta. This study will present a more detailed discussion of facies and facies relationships within the Lea Park - Belly River transition in central Alberta and will attempt to integrate these facies relationships into a stratigraphic interpretation based on allostratigraphic rather than lithostratigraphic techniques.

2.4: Selection of the Study Area

One of the main purposes of this study is to construct an allostratigraphic subdivision of the sediments of the Lea Park - Belly River transition, so as to determine the history of relative sea level fluctuation. The most logical place to do this would be at the depositional edge of the Belly River Formation, as it is in this region that the effects of fluctuations in relative sea level would have their greatest effect. The depositional edge of the Belly River Formation is in southwestern Saskatchewan. However, this thesis is a subsurface study that requires abundant core data, and cores are present in abundance only where commercial hydrocarbon wells exist. There is not an abundant data base of cores available for the region of tne Belly River depositional edge. As a result, this area was not chosen for study.

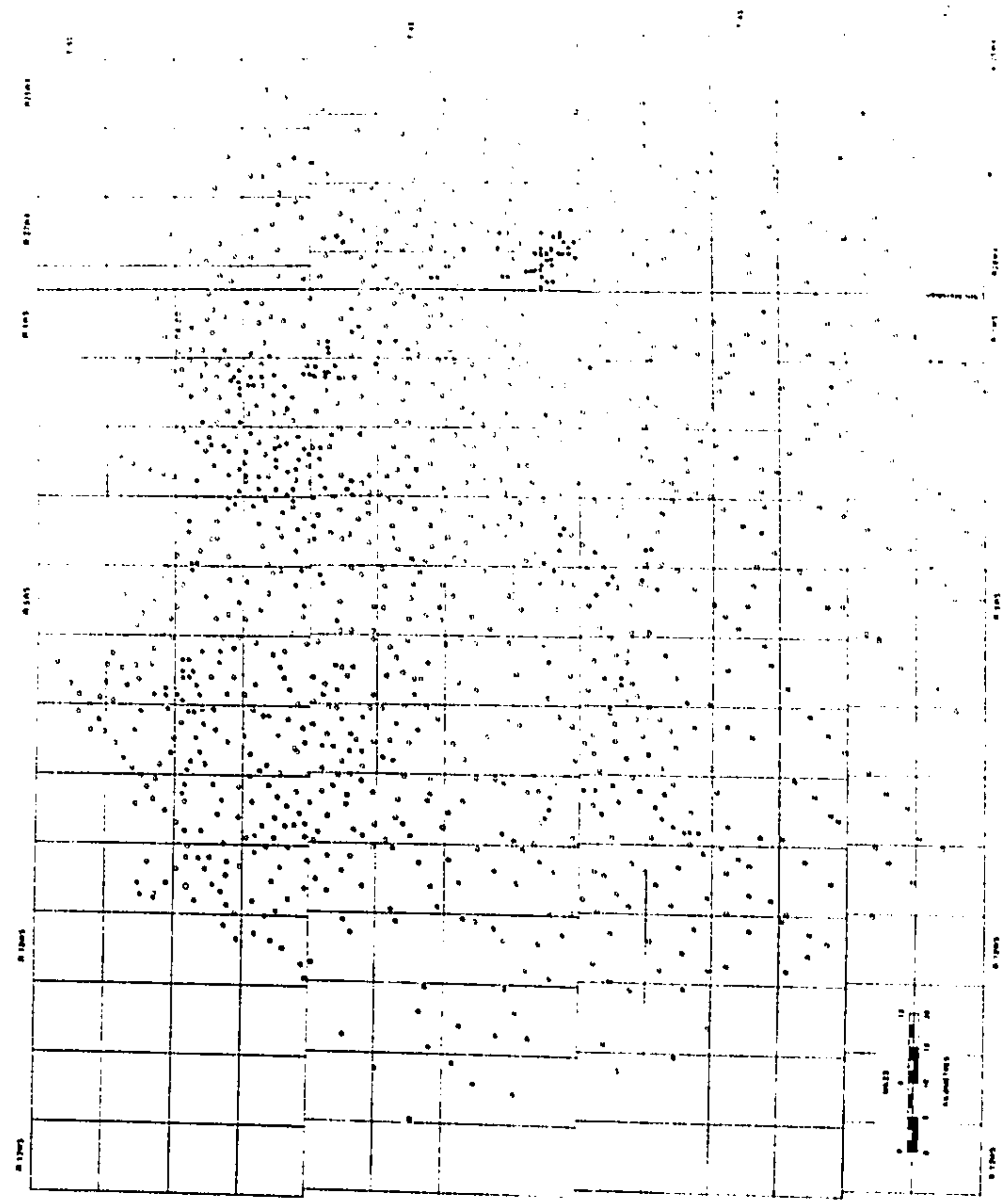
Abundant core data for the Lea Park - Belly River transition exists in central Alberta, where there are numerous oil and gas fields which produce from the sandstones of the Belly River Formation. The study area is shown in Figure 2.1. It extends from Township 37 to Township 50, and from Range 25W4 to Range 10W5, and covers approximately 20,000 km^2. The study area was chosen because it contains most of the major producing fields for the Belly River Formation. It is only in this area that there is enough core data to determine detailed three-dimensional facies relationships for the sediments of the Lea Park - Belly River transition. An enlargement of the study area showing the individual oil and gas fields is depicted in figure 2.8. This area is several hundred kilometres to the west of the depositional edge of the Belly River Formation in Saskatchewan. As a result, the study will focus on constructing an allostratigraphic framework for sediments at the base of a prograding wedge of sediment, rather than at the depositional edge. As mentioned previously, the nature of coastal systems deposited at the base of a prograding wedge may

Figure 2.8: Expanded view of the study area showing the major hydrocarbon fields which produce from the Belly River Formation.



Edmonton
PEMBINA
KEYSTONE
FERRYBANK
FERRIER-
WILLESDEN
GREEN
WILSON CREEK
Approximate Limit
of Disturbed Belt
Fifth Meridian
MILES
0 10 20 30
0 10 20 30 40 50
KILOMETRES
ALBERTA
Edmonton
Calgary

Figure 2.9: Map showing data base of the study area. Locations where core was logged are shown in solid circles, whereas locations where well logs only were used are shown in open circles. Each township is 36 square miles in area.



be different from those deposited at the feather edge because of the effects that falling sea level may have on progradation.

The data base for the study is shown in figure 2.9. It consists of over 1200 wireline logs (open circles) and 221 measured core sections in the Lea Park - Belly River transition (closed circles). Most of the data is concentrated within the major Belly River producing fields of Keystone, Pembina, Ferrybank, Ferrier - Willesden Green, and Wilson Creek.

CHAPTER 3: FACIES ASSOCIATIONS

Introduction

The sediments of the Lea Park - Belly River transition in central Alberta consist of a variety of marine and non-marine facies, which tend to occur in generally predictable facies associations. Rather than describing both the individual facies that exist and their associations, this chapter will simply present and describe the facies associations observed in the Lea Park - Belly River transition within the study area. Descriptions of the individual components of the facies associations will be brief. A brief interpretation of the depositional environment of each facies association is also given. More detailed interpretations of the depositional environment of the facies associations are given in chapter 11. Eight facies associations were observed, and these were grouped into six categories, with two of the categories containing two facies associations.

Facies Association 1a: Offshore Shelf Sediments

This facies association consists of (1) background sediments of cm-scale interbedded mudstones, siltstones, and very fine and fine-grained sandstones with (2) occasional thicker interbeds of sandstones. Figure 3.2 is a typical core section though this facies association, and figure 3.3 shows the core photographs of this section. The legend of symbols for all stratigraphic sections is shown in figure 3.1. The association is common and occurs

Figure 3.1: Legend of symbols used in stratigraphic sections in this thesis.

FACIES LEGEND

LITHOLOGY

- Sandstone
- Pebbly Sandstone
- Siltstone
- Mudstone
- Coal
- Paleosol
- Bentonite

STRUCTURES

- Wave Ripples
- Current Ripples
- Cross Bedding
- LAIS
- Flat Bedding
- Synaeresis Cracks
- Mud Clasts
- Soft Sediment Deformation
- Hummocky Cross Stratification

FOSSILS

- Roots
- Oysters
- Burrows
- Plant Debris
- P *Planolites*
- S *Skolithos*
- T *Teichichnus*
- H *Helminthopsis*
- Th *Thalassinoides*
- O *Ophiomorpha*
- Tr *Trichichnus*
- M *Macaronichnus*
- R *Rosselia*
- C *Chondrites*
- A *Asterosoma*

Figure 3.2: Stratigraphic section of Facies Association 1. Legend of symbols is given in figure 3.1.

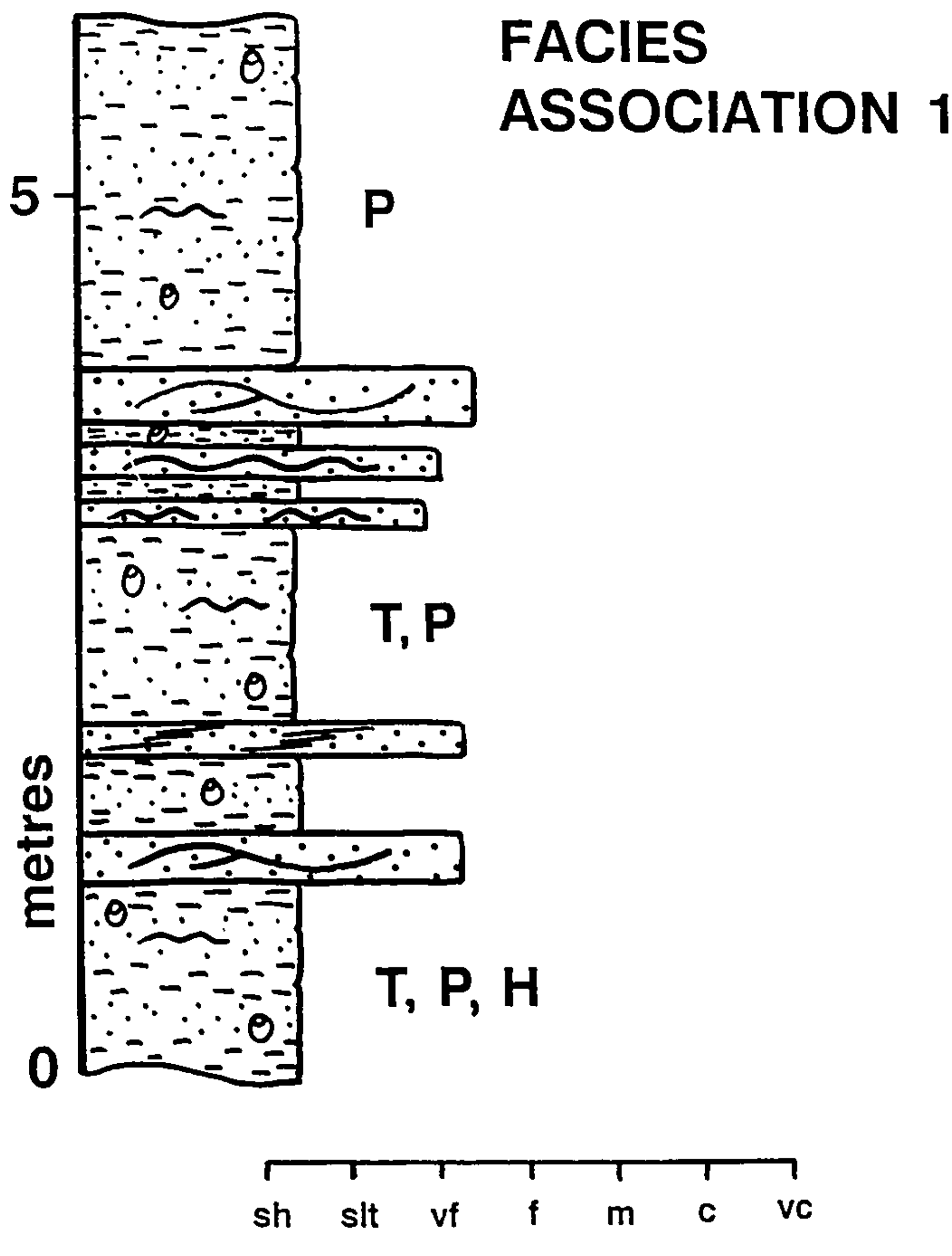
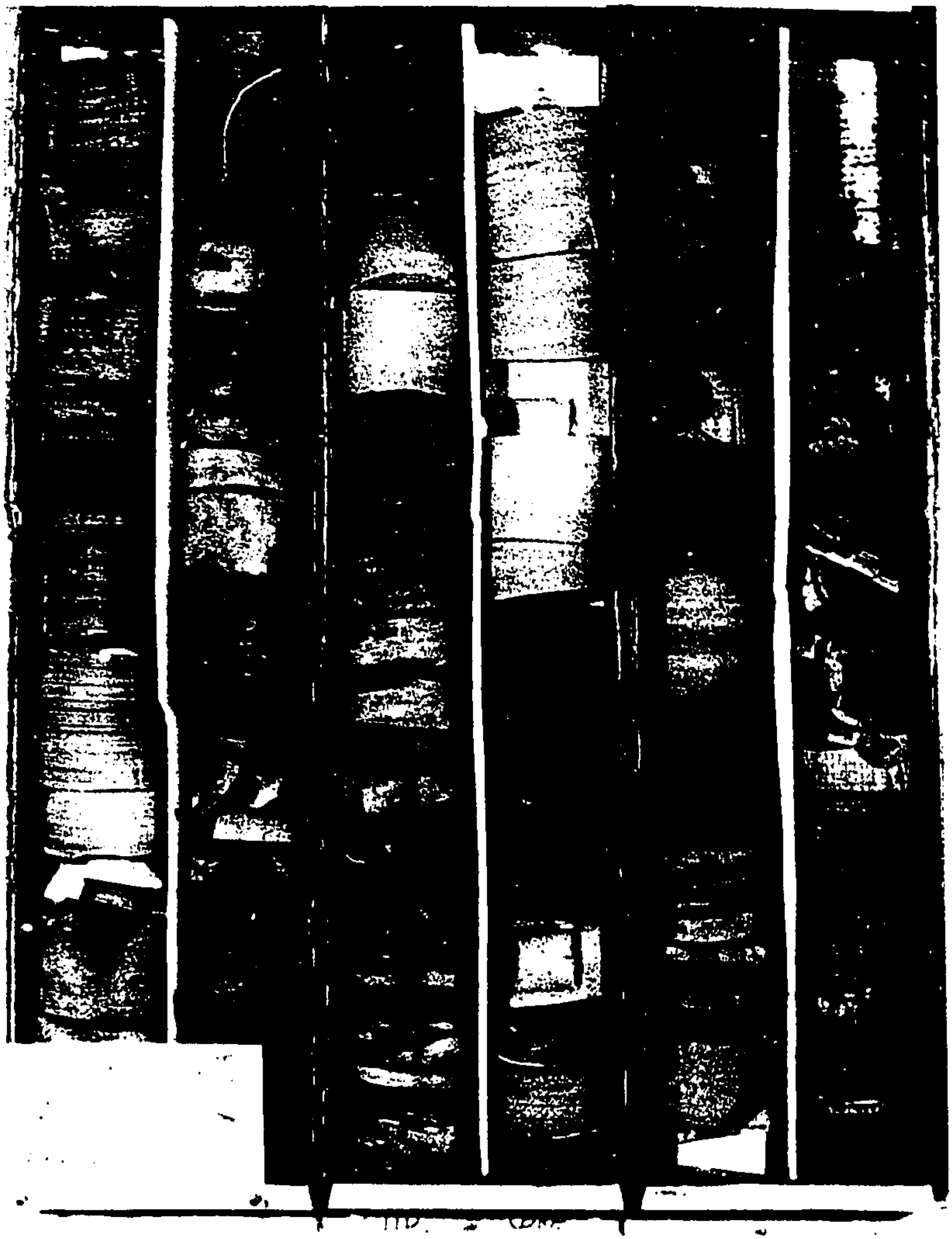

FACIES
ASSOCIATION 1
5
P
T, P
metres
T, P, H
0
sh
slt
vf
f
m
c
vc

Figure 3.3: Box photographs of core section of Facies Association 1. Well location is 6-16-46-1W5. The darker sediments are the background interbedded mudstones, siltstones, and sandstones. The lighter-coloured beds are the thicker interbeds of wave-rippled and HCS sandstone.



throughout the study area with thickness ranging from a few tens of cm to in excess of 15 m.

Interbedding in the background mudstones, siltstones, and sandstones occurs on a scale 0.5 to 15 cm with the average bed thickness being 1-3 cm (Fig. 3.4a). The percentage of mudstone ranges from 20-90% of the rock volume. The percentage of sandstone ranges from 10-50%, and siltstone from 20-60%.

The background interbeds of sandstone and siltstone are usually sharp-based, and often show colour grading (light to dark) and/or size grading (fining up) within the beds (Fig. 3.4b). Sandstone and siltstone beds contain small-scale wave ripples (wavelength less than the core diameter of 4.8 or 6.4 cm (Fig. 3.4c), small-scale current ripples (which are often expressed as climbing ripples)(Fig. 3.5a), flat lamination, and large-scale low angle inclined stratification (LAIS), where the dip of lamination is less than 10°. It is difficult to determine whether this lamination represents large-scale two-dimensional wave ripples or three-dimensional hummocky cross stratification because the wavelength of the original bedform is greater than the diameter of the core. Sandstone and siltstone beds are also rarely structureless. Organic matter is common within the interbedded mudstones, siltstones, and sandstones. It usually occurs as small, macerated plant fragments which are dispersed within the mudstones or form distinct laminations within the sandstones and siltstones.

Thicker beds of very fine to fine-grained sandstone are interspersed

Figure 3.4

(A) Typical background sediments of Facies Association 1. Note vertical *Teichichnus* burrow in the middle of the core. Scale in all photographs is 3 cm. Location: 4-23-47-3W5; 1057.2 m.

(B) Graded beds of siltstone/very fine-grained sandstone. Location 12-36-46-2W5; 991.8 m.

(C) Wave-rippled sandstone interbedded with mudstone. Location 14-5-49-6W5; 1074.9 m.

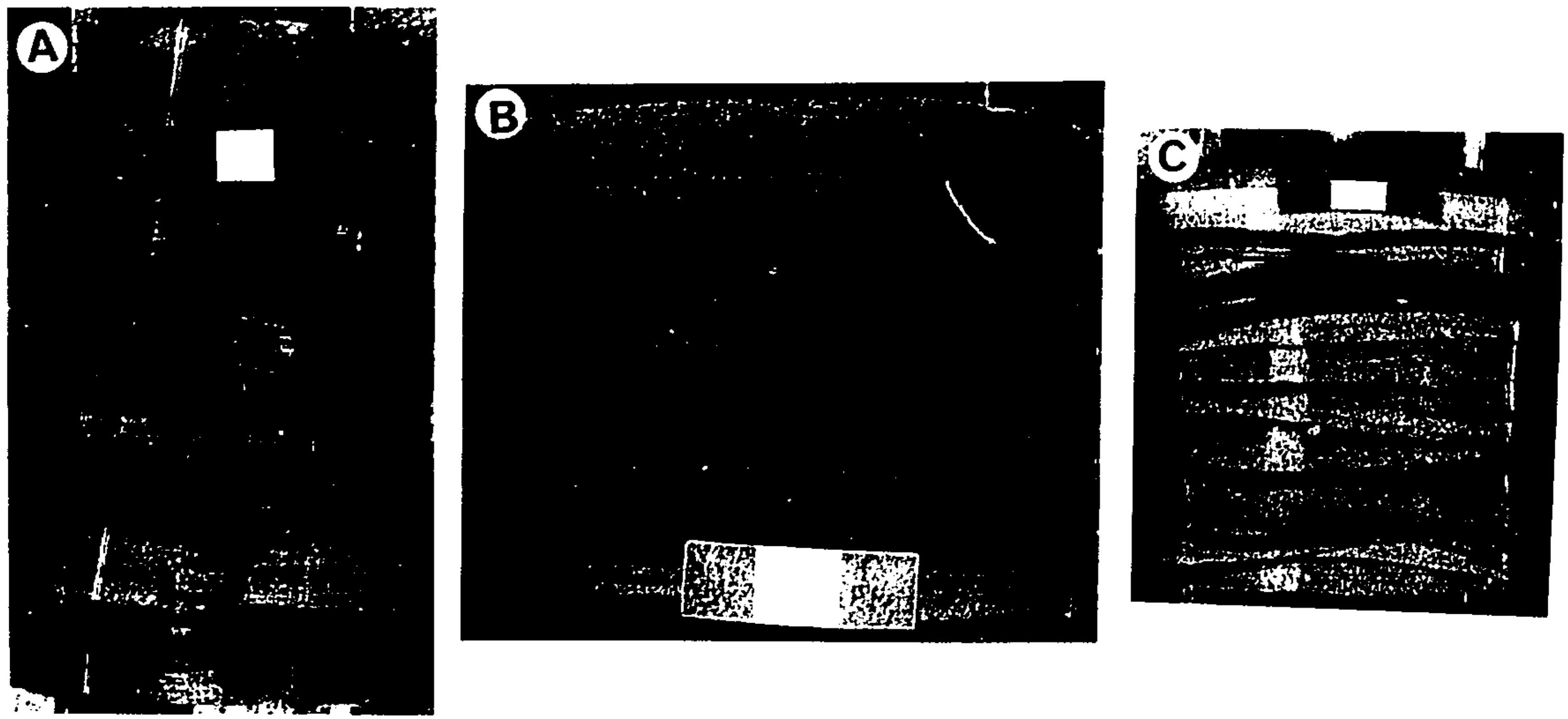
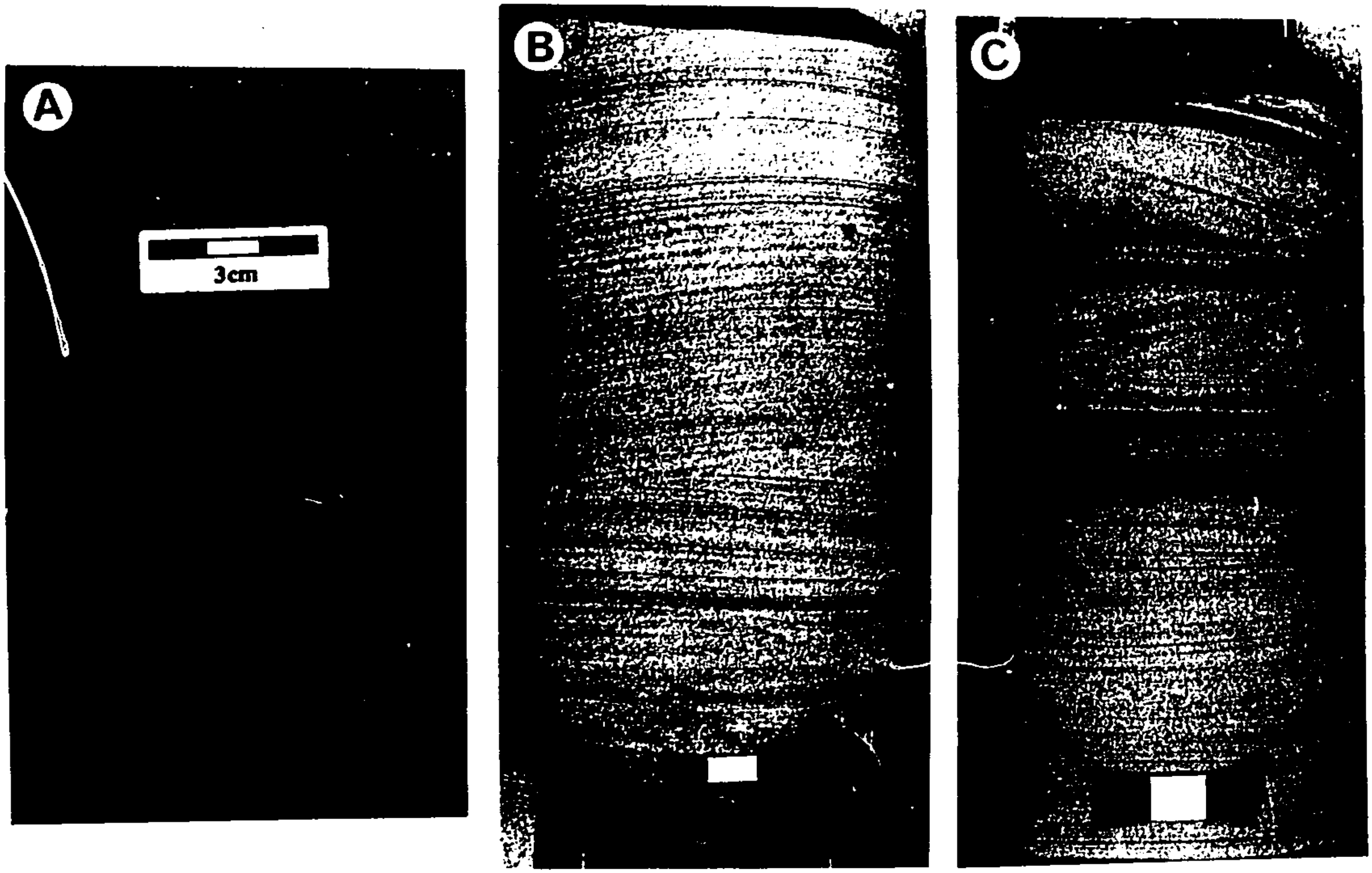

A
B
C

Figure 3.5

(A) Climbing-rippled sandstone bed within Facies Association 1. Location: 6-23-47-2W5; 951.5 m.

(B) Intersecting low-angle inclined stratified (LAIS) sandstone bed interpreted as hummocky cross stratification (HCS). Location 8-2-42-5W5; 1368.2 m.

(C) Flat-laminated sandstone bed grading up into wave-rippled sandstone. Location: 14-24-46-2W5; 1016.5 m.


A
3cm
B
C

within the background interbedded mudstones, siltstones and sandstones. These beds can range in thickness from a few tens of centimetres to just over a metre. The structures within the sandstone beds are similar to those described for the background interbeds. The most common structure is LAIS, and where intersecting laminae occur, it can be identified as hummocky cross stratification (Fig. 3.5b) These thicker beds of sandstone can grade upward from structureless, flat-laminated, or LAIS into wave-rippled sandstone (Fig. 3.5c). Plant fragments also commonly form laminae within structured sandstones. These laminae are often present as distinct couplets, or pairs, and are present within the LAIS, wave-rippled, and flat-bedded sandstones and siltstones (Figs. 3.5b, 3.5c).

Soft-sediment deformation occurs within this facies association, and is neither common nor rare (Fig. 3.6a, 3.6b). Individual deformation features such as balls or load casts are commonly quite large, reaching in excess of ten centimetres in diameter. Rare synaeresis cracks occur at the boundaries between sandstone and mudstone beds (Fig. 3.6c).

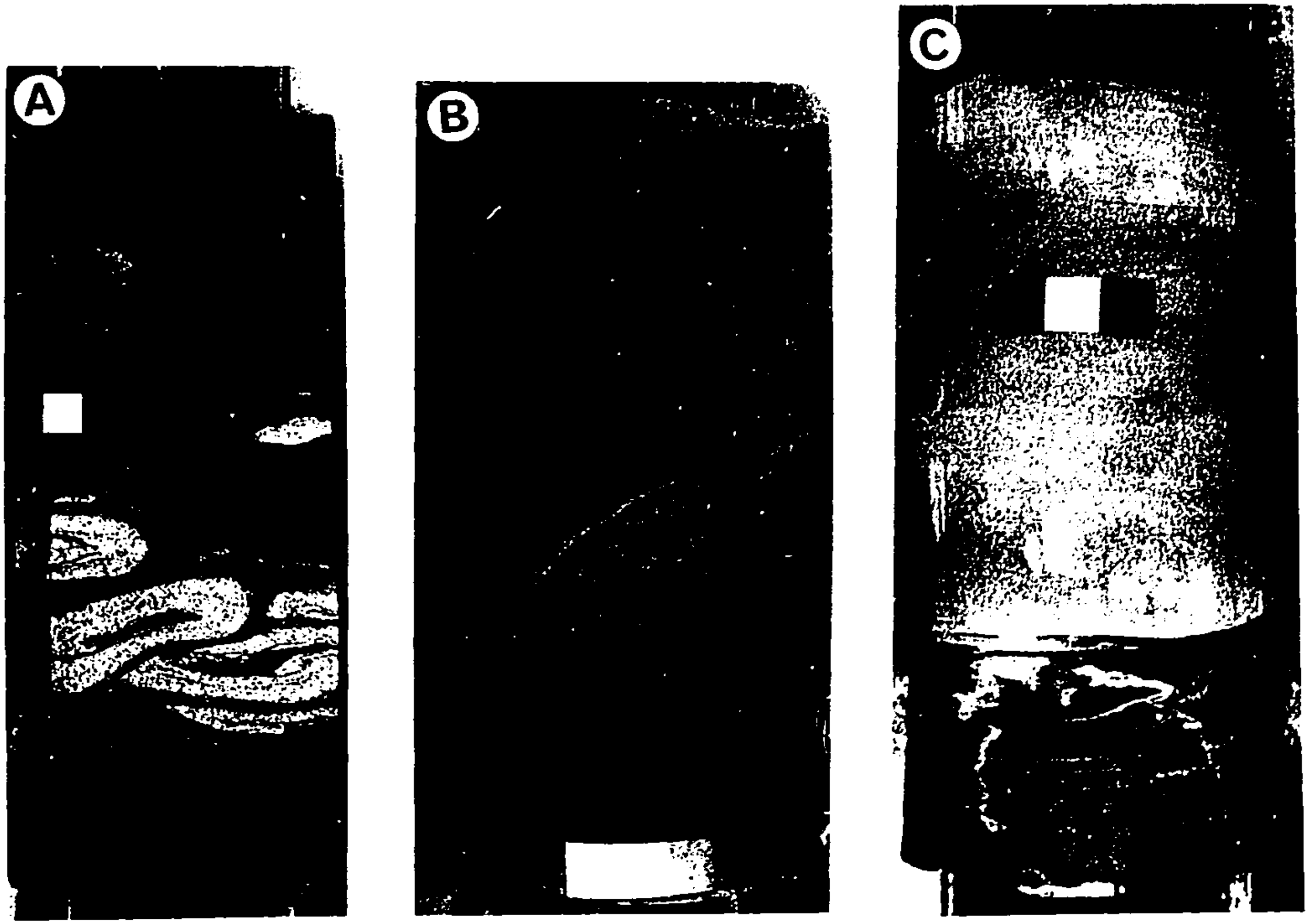
Macrofossils are absent from this facies association, but trace fossils are common. Traces usually occur as distinct burrows which do not completely disrupt lamination or bedding. As a result the term burrowed is preferred over bioturbated. Trace fossils present (in decreasing order of abundance) within the background interbedded sediments are *Planolites, Skolithos, Teichichnus, Helminthopsis, Rhizocorallium, and Chondrites.* Examples of some of these

Figure 3.6

(A) Soft-sediment deformation within Facies Association 1. Note the deformed, but relatively continuous sandstone bed. Location: 10-9-47-2W5; 1017.7 m.

(B) Soft-sediment deformation within siltstones and mudstones of Facies Association 1. Location: 6-3-48-2W5; 936.6 m.

(C) Synaeresis cracks filled by sandstone within Facies Association 1. Location: 16-29-47-2W5; 1016.3 m.


A
B
C

are shown in figures 3.4a and 3.7a-3.7b. Within the thicker sandstones beds, apparent trace fossils are uncommon. Mud-lined *Teichichnus* are present in a few of the thick sandstone beds.

Interpretation: The style of interbedding, sedimentary structures such as wave ripples and hummocky cross stratification, and the presence of marine trace fossils such as *Teichichnus* and *Helminthopsis* suggest that the sediments of Facies Association 1a were deposited in a shallow to moderately deep open marine shelf environment.

Facies Association 1b: *Helminthopsis*-Burrowed Shelf Mudstones

A sub-variety of Facies Association 1 exists, but is only present in geographically restricted areas. The geographical restriction of this subfacies will be discussed further in a later chapter of the thesis.

The background sediment of the facies association again consists of cm-scale interbedded mudstone, siltstone, and very fine to fine-grained sandstone. The composition of the mudstone is different from that of Facies Association 1a, consisting of a much higher proportion of swelling clays. The volume of sandstone rarely exceeds 20-25%, and is generally less than in Facies Association 1a. Fine-grained sandstone is less common than in Facies Association 1a, with most of the sandstone beds being very fine-grained.

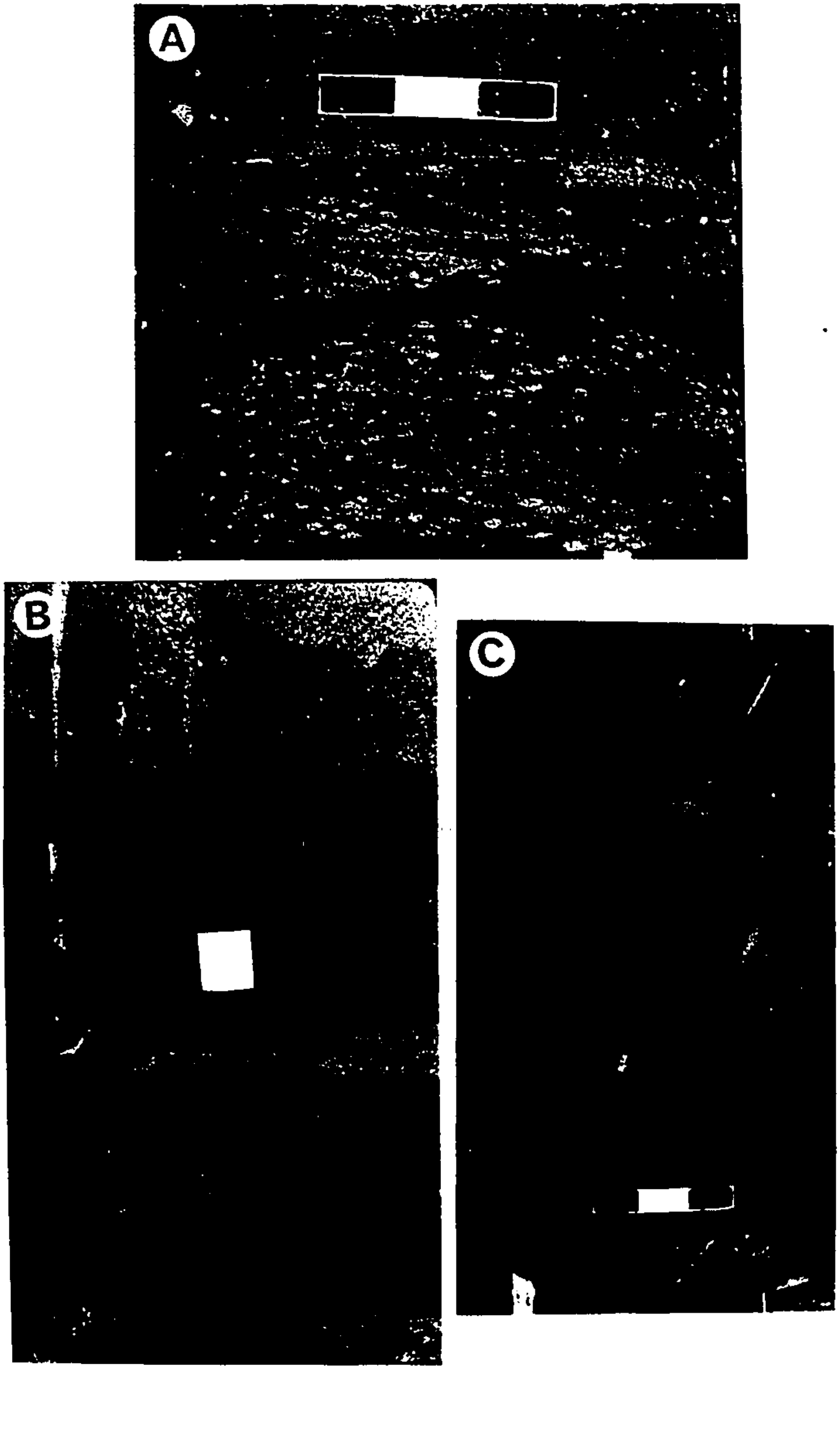
Sedimentary structures are not as common within the sandstone and siltstone beds, with most beds being either structureless or containing vague

Figure 3.7

(A) *Planolites* burrows in mudstone. Location: 10-3-47-27W4; 949 m.

(B) *Helminthopsis* burrows (black flecks of mud in basal half of the photo) within Facies Association 1a. Also note the vertical mud-filled burrow in the top of the photo. 14-28-48-6W5; 1080.2 m.

(C) *Helminthopsis* burrows in Facies Association 1b. Location: 14-6-43-27W4; 937.4 m.


A
B
C

LAIS or flat lamination. Wave ripples are present but are not common, and current ripples are rare.

The defining feature of this facies association is the abundance of *Helminthopsis* burrows (Fig. 3.7c). As in Facies Association 1a, burrowing is not usually intense enough to completely destroy bedding, but Facies Association 1b does have a more bioturbated appearance than 1a. Other traces are relatively scarce, and consists of *Planolites, Skolithos, Terebellina* and *Teichichnus*. Macrofossils are absent from this facies association.

Interpretation: The dominantly muddy and silty sediment and the presence of abundant *Helminthopsis* as well as more rare *Teichichnus* indicates that these sediments were deposited in a marine environment with low levels of wave or current energy. *Helminthopsis* is thought by some ichnologists to be a possible indicator of anoxic depositional environments (G. Pemberton, pers. comm.), which might occur in a quiet water, fairly protected marine environment that was not disturbed by strong currents or waves on a regular basis.

Facies Association 2a: Shoreface Sediments

This facies association consists of very fine to medium-grained marine sandstones which form coarsening upward sequences. Figure 3.8 is an example of this facies association, and the accompanying core photographs are shown in figure 3.9. Facies Association 2 is very common within the study

Figure 3.8: Stratigraphic section of Facies Association 2a. Based on core in well 6-16-46-1W5.

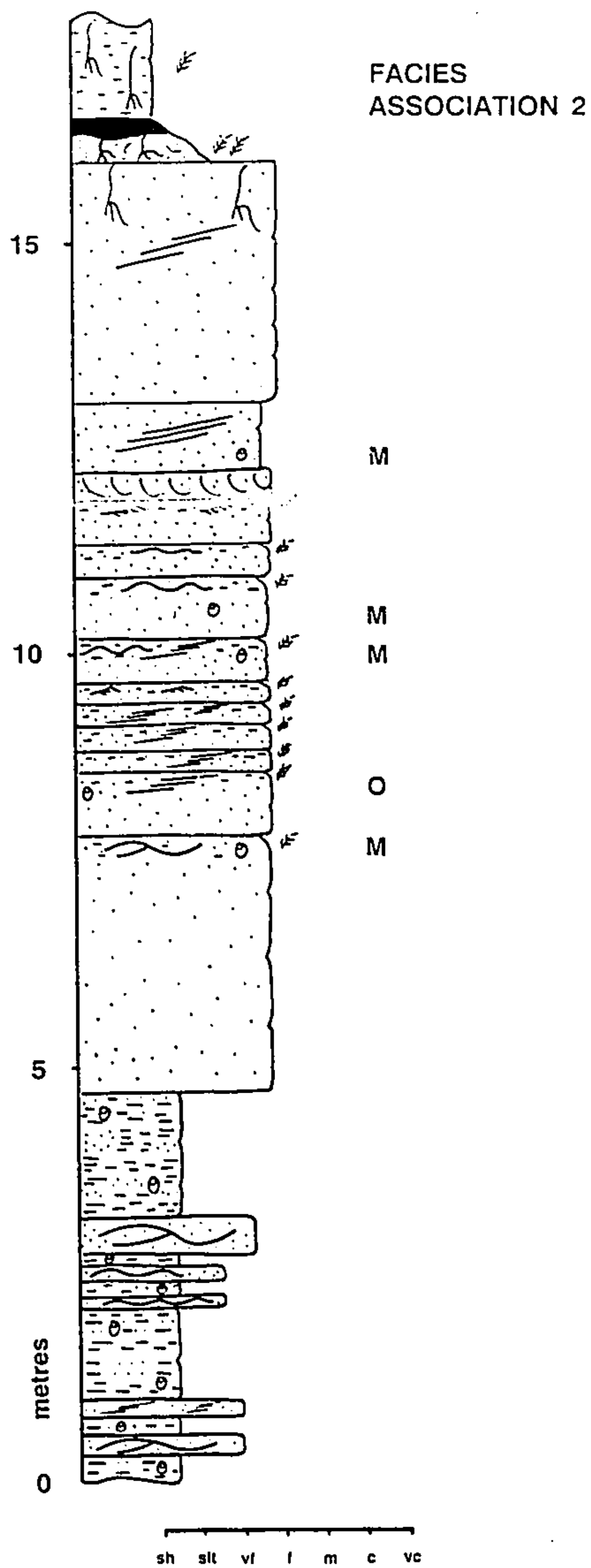
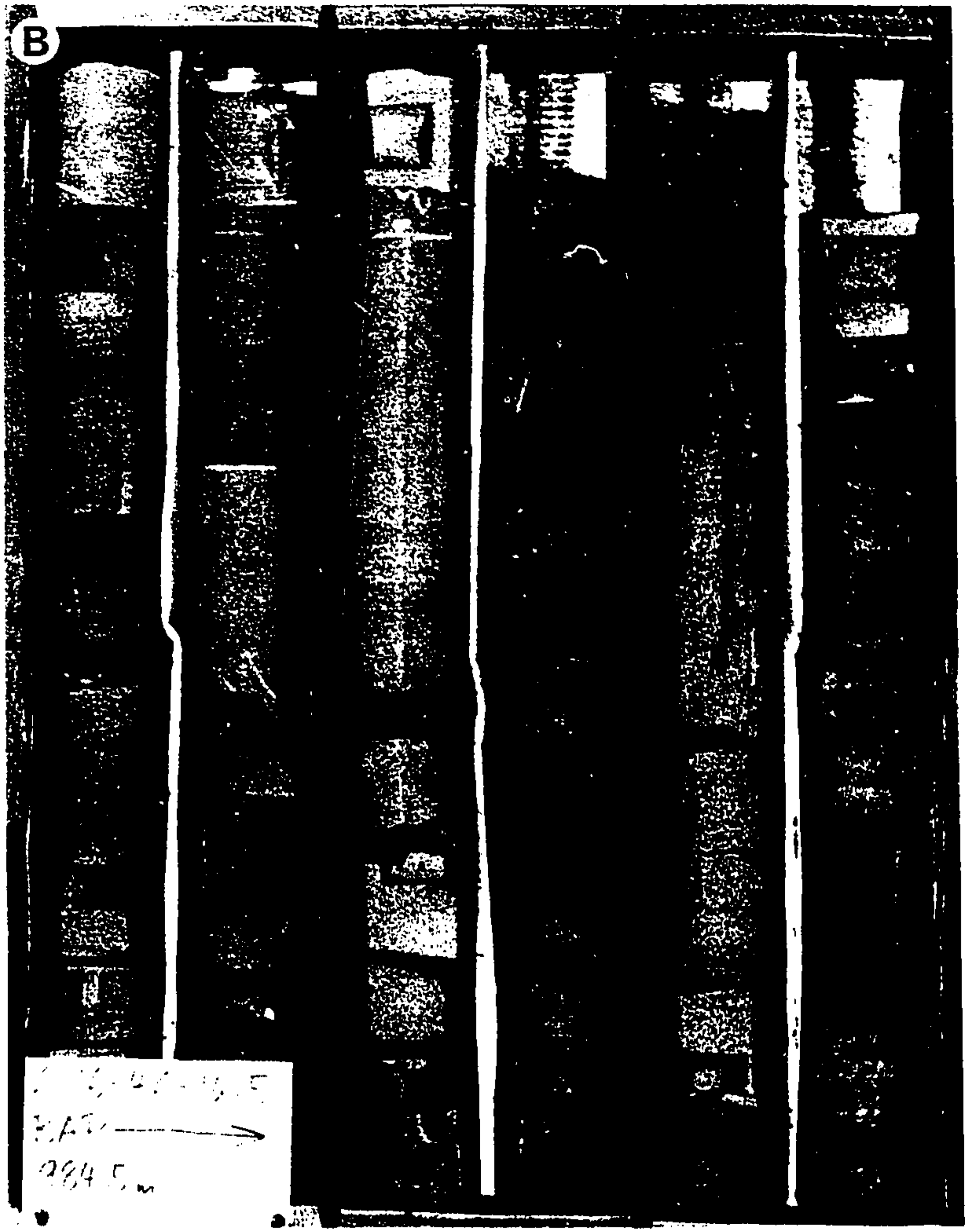
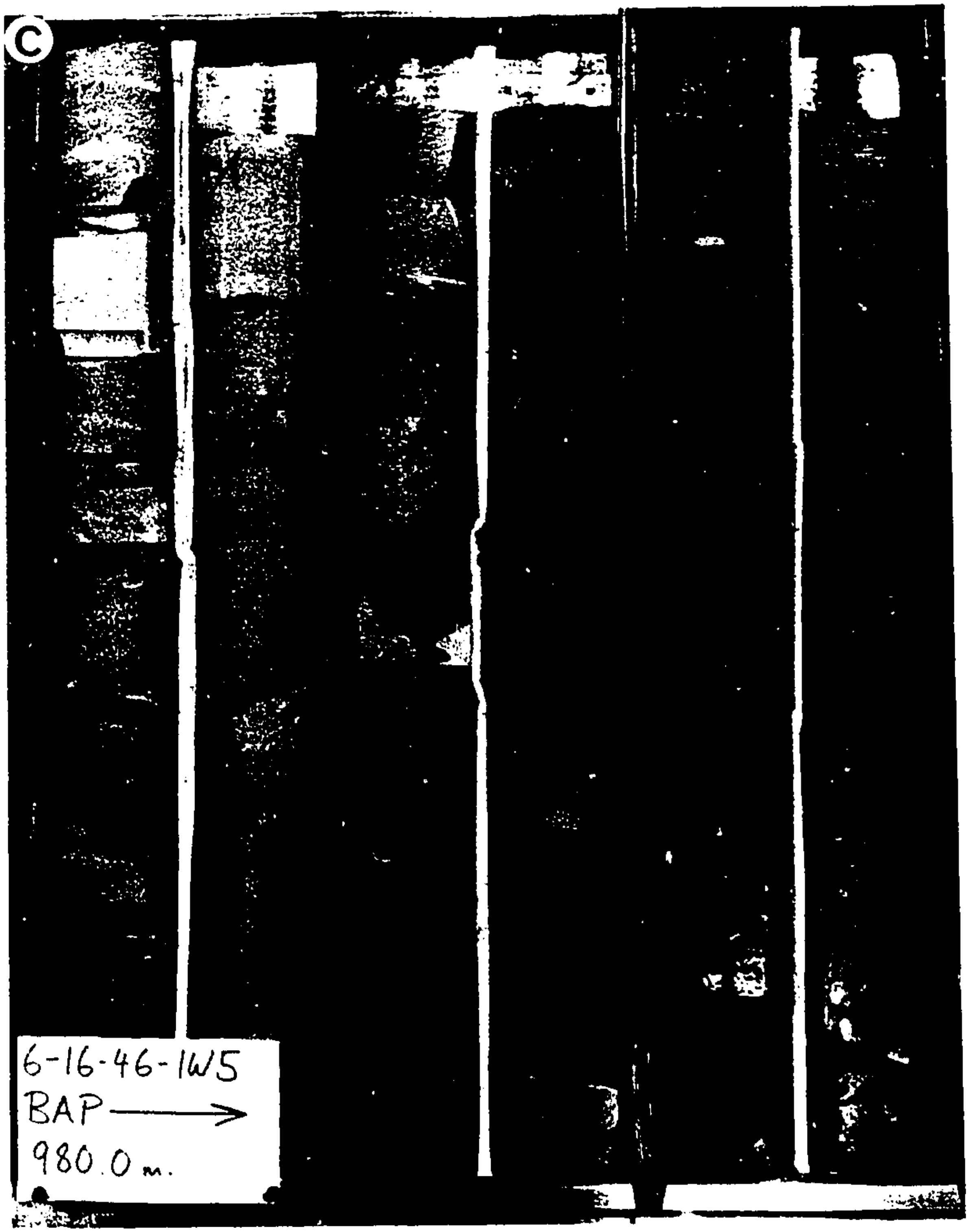

FACIES
ASSOCIATION 2
M
M
M
O
M
15
10
5
0
metres
sh
slt
vf
f
m
c
vc

Figure 3.9: Box photographs of succession in Facies Association 2a in well 6-16-46-1W5. The base of the succession in each plate is at the bottom left hand corner, and the top is at the top right hand corner. Each core tube is 75 cm in height. The base of the succession of Facies Association is at the base of Plate A. Note the dominance of massive, unbedded sandstone in Plate A. Plate B contains numerous stacked massive-to-laminated beds, each being several tens of centimetres thick. The sandstone in Plate C are dominantly LAIS/flat-laminated, and the top of the succession is denoted by the arrow in Plate D.


A
189 0 m


B


C
6-16-46-1W5
BAP
980.0 m.

area, especially within the Keystone, Ferrybank, and Pembina fields. The coarsening-upward sequence, when fully preserved, ranges in thickness from 6 to almost 20 m.

The base of the sequence is invariably sharp, and consists of very fine or fine-grained sandstones which sharply overlie the shelf sediments (usually mudstones) of Facies Association 1a. In some cases, there is visible erosion at this contact, represented by an angular contact between the two associations and/or mud chips from the underlying shelf sediments (Fig. 3.10a).

The sequence begins with a series of fine-grained sandstone beds ranging from 10 cm to 6 m in thickness in which structureless or very vaguely stratified sandstone is the dominant component (Fig. 3.10b). These beds usually consist of a succession in which structureless sandstone gradually grades upward into vaguely-defined LAIS or flat-laminated sandstone, in which the stratification becomes more clearly defined upwards (Fig. 3.11a), and which in turn may grade upwards into wave-rippled or current-rippled sandstone (Fig. 3.11b). Structureless sandstone can also grade directly into wave-rippled or current-rippled sandstone (Fig. 3.11c). Well-laminated sediments often comprise only the top few centimetres or tens of centimetres of these beds, with the bulk of the bed being structureless or vaguely lamianted. The tops of beds are sometimes, but not commonly, reworked into wave-rippled or hummocky cross-stratified sandstone (Figs. 3.12a, 3.12b). The base of each succession is usually sharp, and there is often a gradual and subtle internal

Figure 3.10

(A) Base of the shoreline succession in Facies Association 2a. Note the angular contact with the underlying mudstones. Location: 14-23-43-28W4; 1033.5 m.

(B) Structureless sandstone typical of the massive-to-laminated beds of Facies Association 2a. Location: 12-36-46-2W5; 987.9 m.

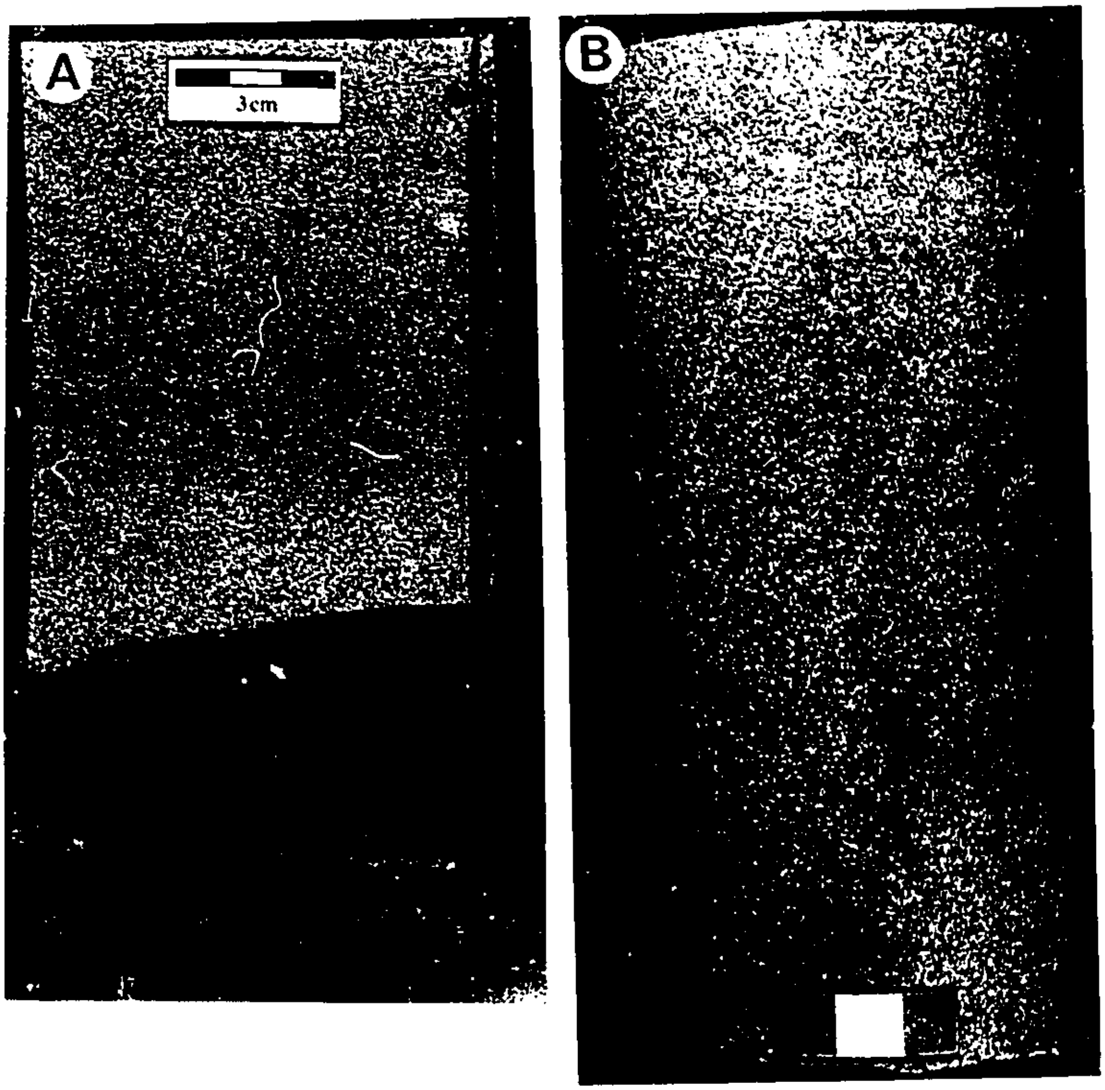

A
3cm
B

Figure 3.11

(A) Vaguely flat-laminated fine-grained sandstone grading up into well-defined flat-laminated sandstone. Plant matter helps define laminations in upper half of photo. Location: 8-4-48-5W5; 1063.2 m.

(B) Massive sandstone grading up into LAIS/wave-rippled sandstone. Deformation of lamination at the top is due to an escape burrow. Location: 6-29-46-1W5; 943.3 m.

(C) Structureless sandstone grading directly up into current-rippled sandstone. Location: 6-16-46-1W5; 1000.5 m.

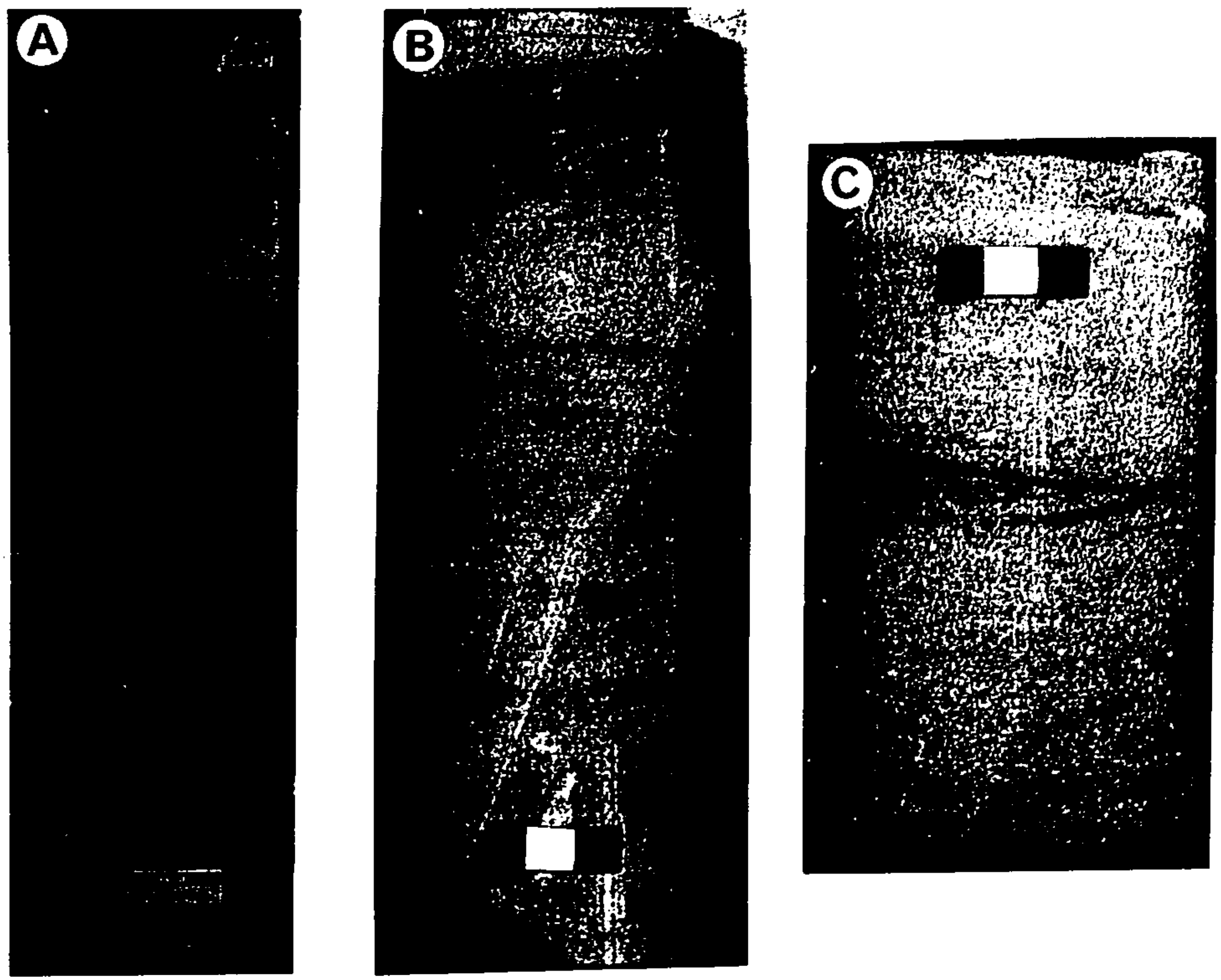
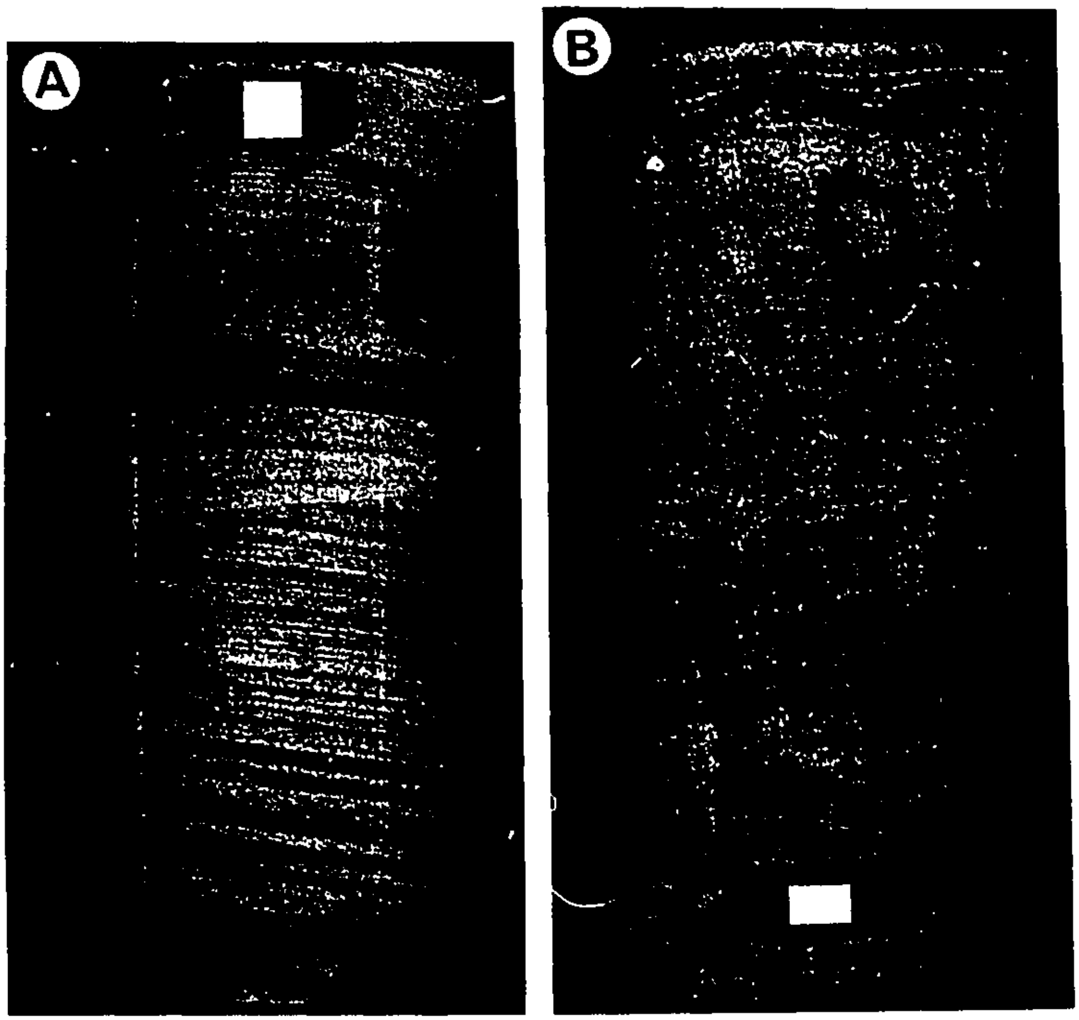

A
B
C

Figure 3.12

(A) HCS sandstone within Facies Association 2a. Location: 10-7-47-3W5; 1019.9 m.

(B) Wave-rippled sandstone at the top of a massive-to-laminated bed in Facies Association 2a. Note the vertical and horizontal *Ophiomorpha* burrows. Arrows point to pairs of organic laminations (see text for discussion of these features).


A
B

fining-upwards trend within the succession. Plant matter commonly forms distinct laminations within the stratified portion of the beds, and often is the only way in which lamination can be detected. These stacked successions of massive to laminated sandstones comprise the bulk of the shoreface sediments facies association.

In the ideal, or normal sequence of Facies Association 2a, the stacked massive to laminated successions become less abundant upwards, and pass into fine-grained or medium-grained trough cross-bedded and LAIS sandstones (Fig. 3.13a). This unit of cross-bedded sandstone ranges in thickness from 0-5 m, and averages 2-3 m. As with the lower sediments, the stratification within the sandstones is often vague. The transition from the massive to laminated successions up into the cross-bedded/LAIS sandstones is often, but not always, accompanied by an increase in grain size of as much as 2 phi divisions (eg. lower fine to lower medium). When the transition occurs with such a distinct change in grain size, the contact between the massive to laminated sandstones and the cross-bedded sandstones is sharp and commonly angular. If the cross-bedded sandstones are not present, the successions of massive-to-laminated sandstone beds continue upwards in its place.

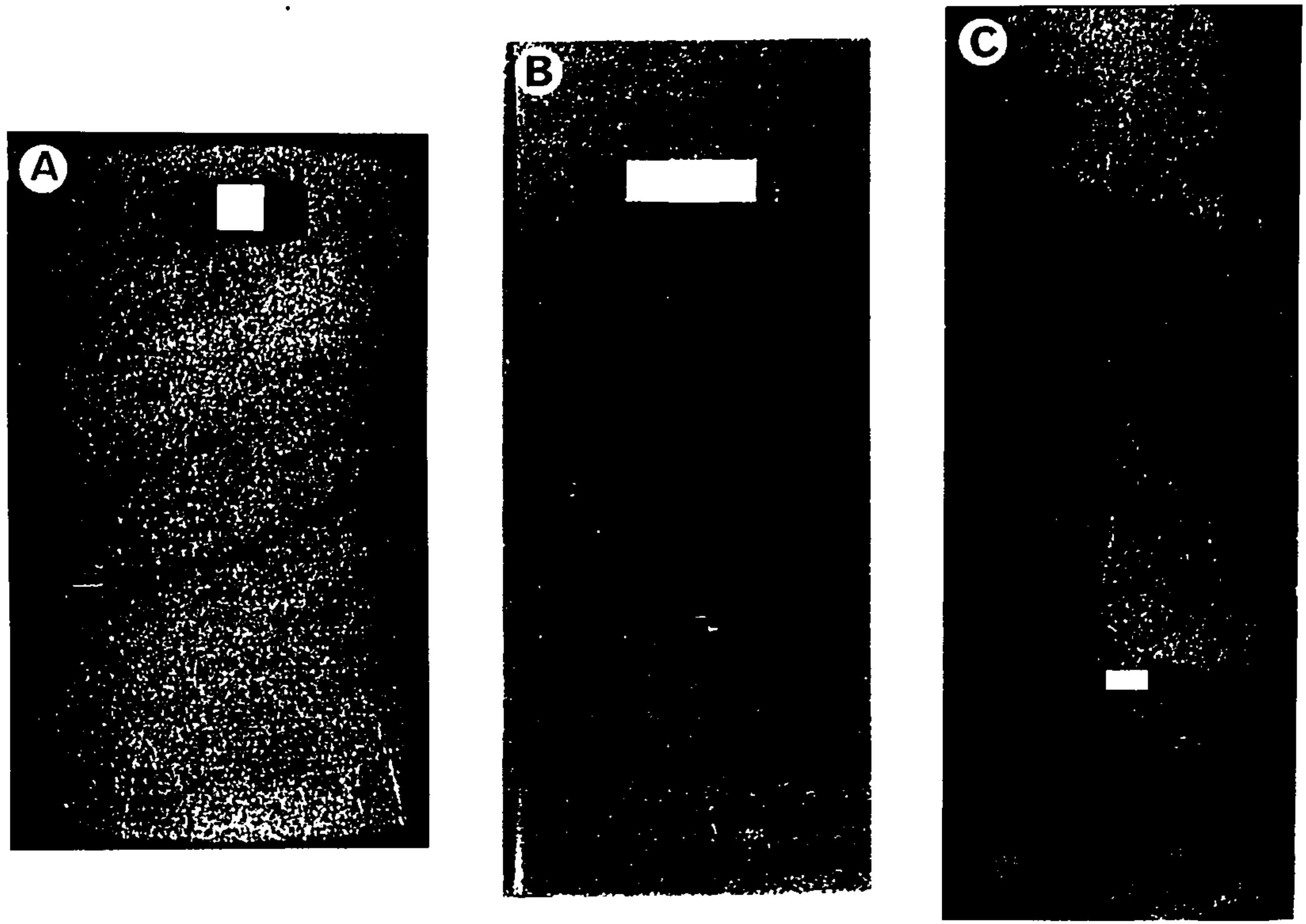
The cross-bedded sandstones sometimes form the top of the coarsening-upward sequence, but more commonly they are overlain by vaguely stratified flat-laminated or LAIS sandstone (angle of dip $<5°$) (Fig. 3.13b) which can be up to 4-5 m in thickness. The uppermost portions of this sand often

Figure 3.13

(A) Cross bedded sandstone in upper portion of Facies Association 2a. Location: 10-7-47-3W5; 1024.8 m

(B) Flat-laminated sandstone in the upper portion of Facies Association 2a. Location: 6-11-40-1W5; 1079.2 m.

(C) Root traces indicating subaerial exposure at the top of Facies Association 2a. Location: 8-22-49-7W5; 1040.8 m.


A
B
C

contain well-preserved root traces (Fig. 3.13c). As with the cross-bedded sandstone, but more rarely, this package of flat-laminated sandstones is sometimes absent.

Macrofossils are absent within this facies association, but several genera of trace fossils are present. Figures 3.12c, 3.14-3.16 show examples of some of the characteristic traces. *Ophiomorpha, Paleophycus,* and *Asterosoma* are common within the massive to laminated successions, usually in the lower half of the sequence. *Macaronichnus* is also very common within the massive to laminated successions, usually higher up within the sequence or within the cross-bedded or flat-laminated sandstones. It can occur as both large robust forms, or as very small, less distinct forms. Mud-lined *Skolithos, Teichichnus,* and *Rosselia* are also present within the shoreface sediments, but are not very common. *Trichichnus* burrows (Fig. 3.16a) are occasionally present near the top of the succession, and are sometimes associated with root traces. A single occurrence of *Teredolites* was also observed (Fig. 3.16b). In some cases, the tubes of vertical traces such as *Skolithos* or *Rosselia* are sharply truncated in what otherwise appears to be the homogeneous portion of the massive to laminated successions (3.16c), indicating that one very thick succession may actually be several amalgamated successions, with the laminated part of only the uppermost succession being preserved.

Interpretation: The coarsening upward sequence capped by roots, the presence of sedimentary structures such as wave ripples and hummocky cross

Figure 3.14

(A) Mud-lined horizontal burrow (*Ophiomorpha*?) in shoreline sandstones of Facies Association 2a. Location: 10-14-47-4W5; 1045.9 m.

(B) Large *Macaronichnus* burrows in massive sandstones. Location: 6-11-40-1W5; 1082.8 m.

(C) Small *Macaronichnus* burrows in vaguely flat-laminated sandstone. Location: 8-4-48-5W5; 1057.8 m.

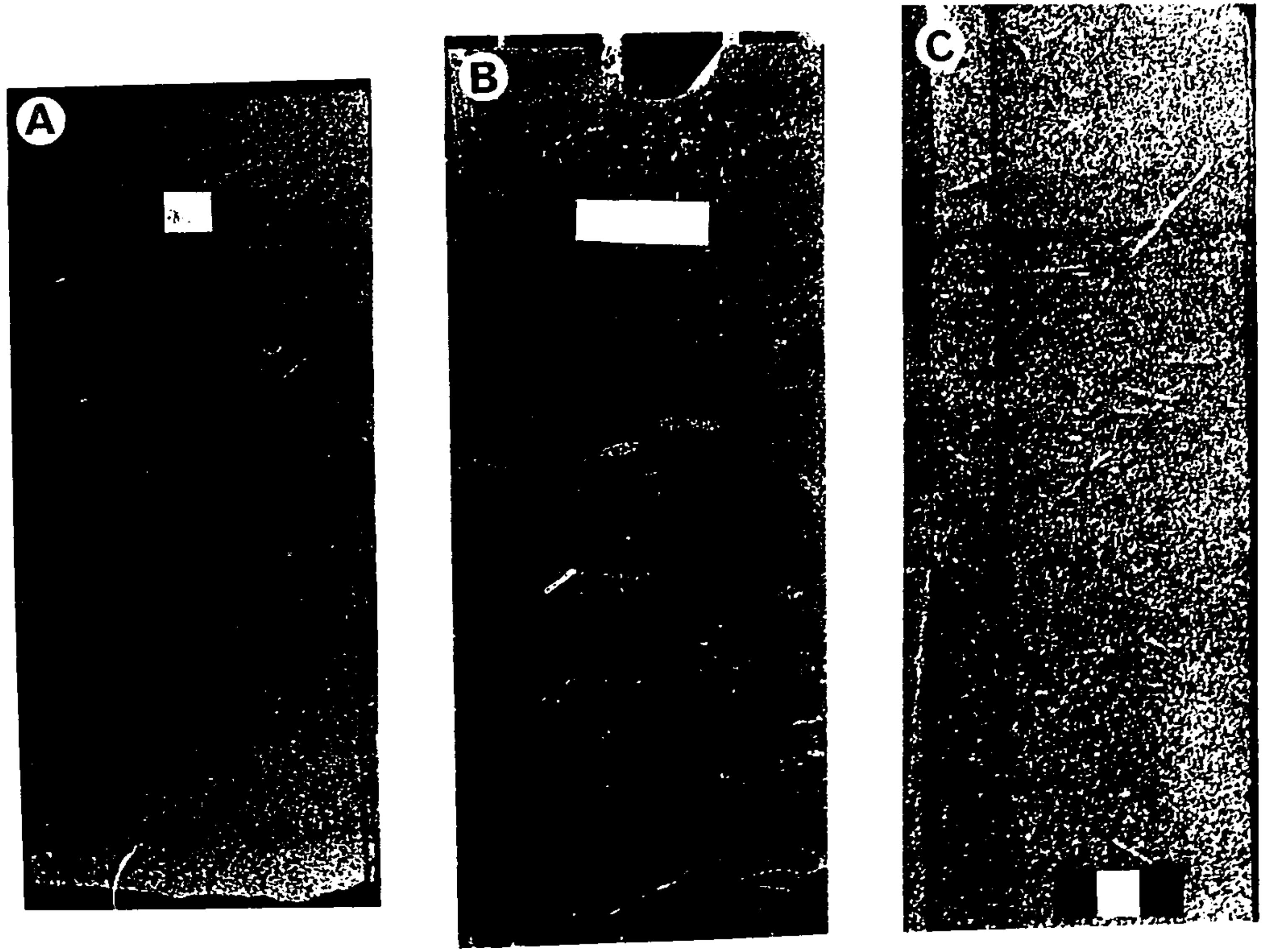

A
B
C

Figure 3.15

(A) *Rosselia* burrow (top of photo) within shoreline sediments. Location: 10-3-43-27W4; 946.7 m.

(B) *Teichichnus* burrow within shoreline sandstones. Location: 14-33-33-28W4; 1179.7 m.

(C) Mud-lined *Skolithos* burrows in shoreline sandstones. Location: 8-6-43-27W4; 935 m.

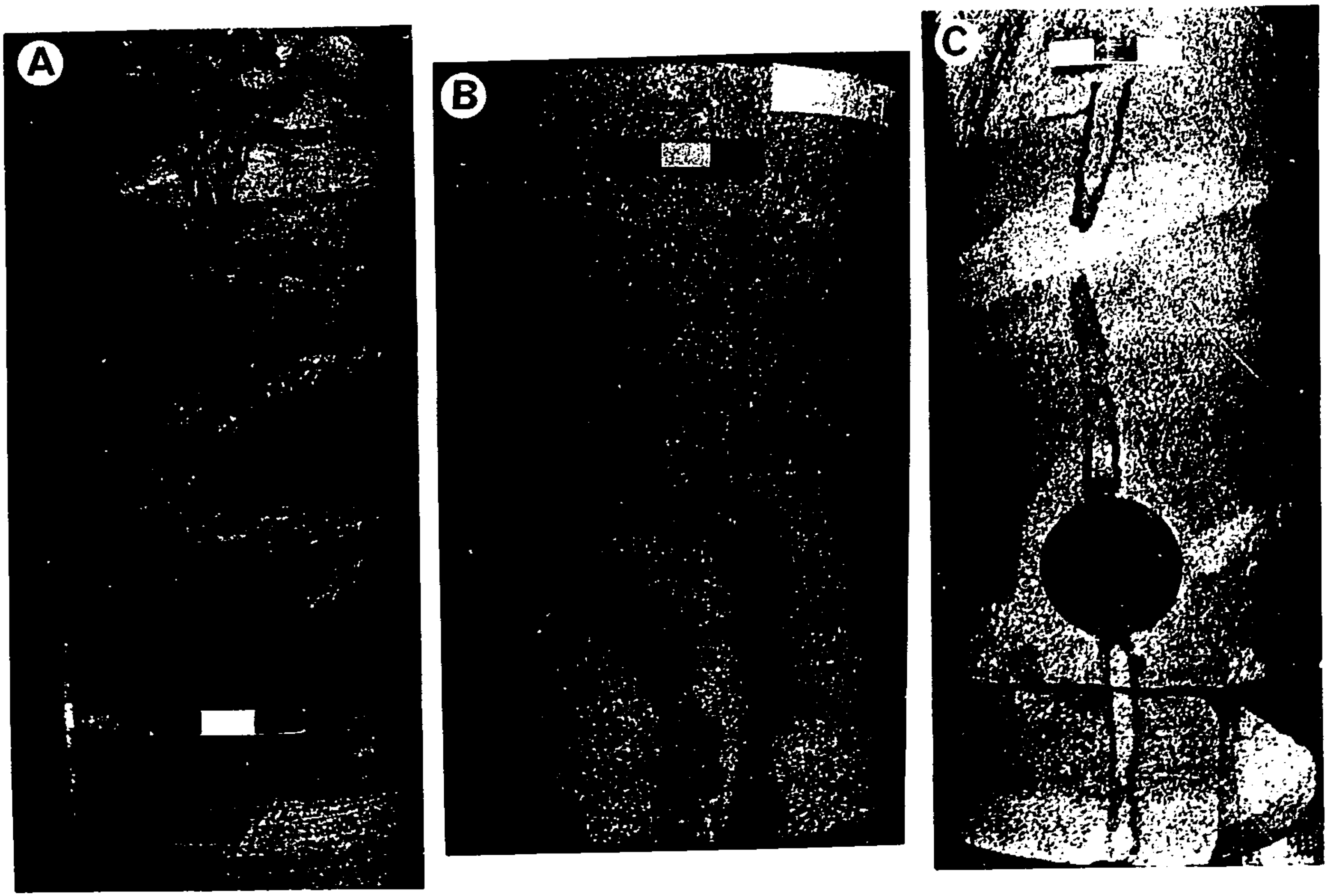
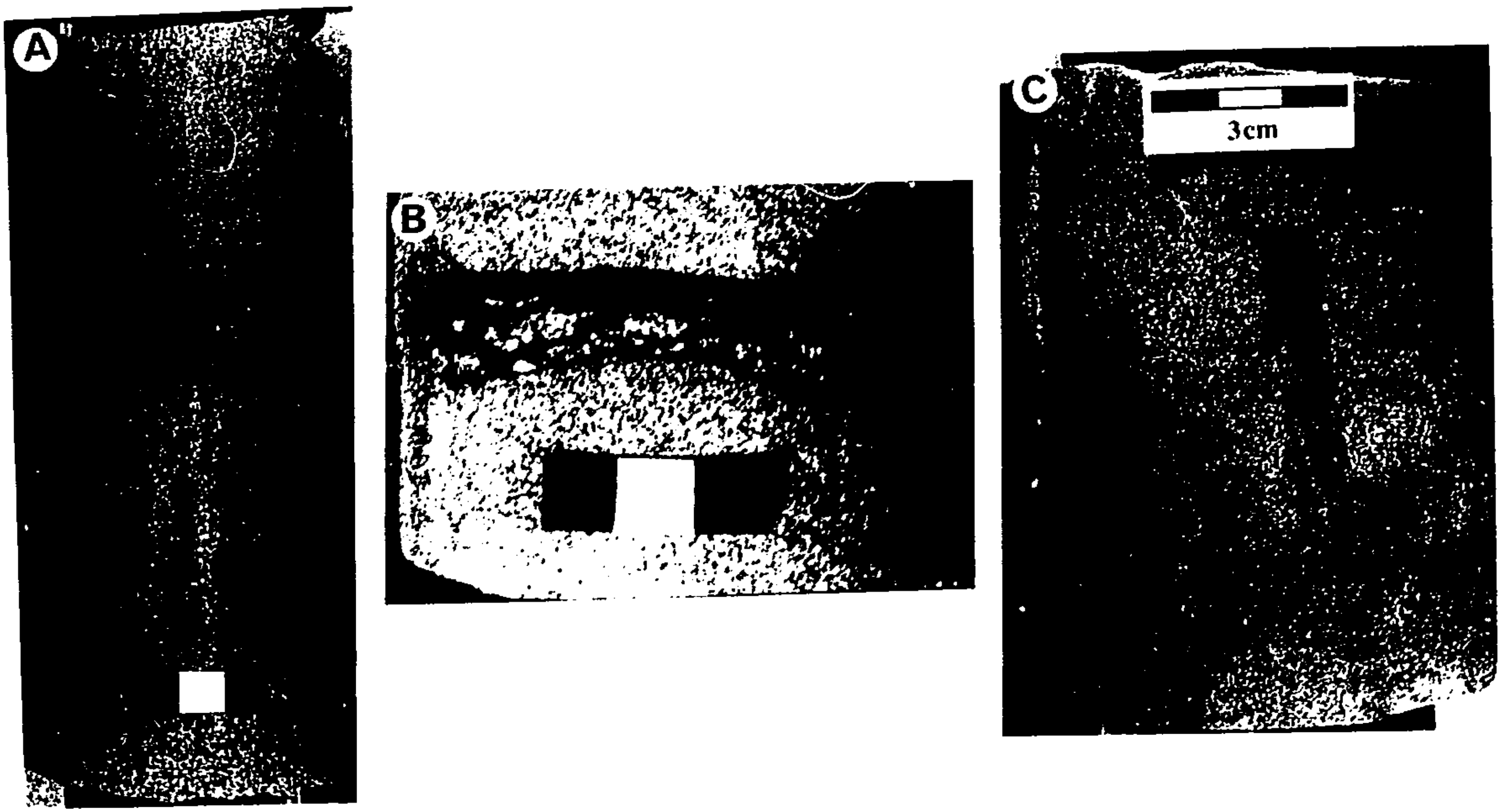

A
B
C

Figure 3.16

(A) *Trichichnus* burrows (small vertical tubes lined with plant matter) in the upper portion of Facies Association 2a. Location: 16-32-47-3W5; 980.1 m.

(B) *Teredolites* burrows in coal fragment within shoreline sandstones. Location: 6-35-48-7W5.

(C) Truncated *Skolithos* or *Rosselia* burrows (shown by arrows) indicating erosional amalgamation of sandstone beds within apparently continuous structureless sandstone. Location: 6-10-43-4W5; 1276 m.


A
B
C
3cm

stratification, and marine trace fossils such as *Macaronichnus, Teichichunus,* and *Rosselia* indicate that these sediments were deposited in a prograding shoreface environment. The abundance of massive sandstones and plant matter may be indicative of a nearby fluvial source.

Facies Association 2b: Cross-Bedded Shoreline Sandstones

This facies association consists of very fine to fine-grained marine sandstones which form a coarsening upward successions. Figure 3.17 shows a typical core log of this facies association, and the accompanying core photos for this log are shown in figure 3.18. The association is similar in some aspects to Facies Association 2a , but instead of being dominated by beds of massive sandstone which grade upward into straified sandstone, it is dominated by cross-bedded sandstone throughout the lower and middle portions of the succession. It is not common within the study area, and is restricted to a few occurrences in the western and northern areas of Pembina. When fully preserved, the succession ranges from 6-10 m in thickness.

The base of the succession is sharp, and as in Facies Association 2a, consists of very fine or fine-grained sandstones which sharply overlie the self sediments (usually mudstones) of Facies Association 1a. No evidence of visible erosion was observed at this contact. However, the basal contact is only penetrated in a few cores, providing a small data base.

The sandstones which overlie the basal contact are usually cross

Figure 3.17: Stratigraphic section of Facies Association 2b. Based on cored section in well 16-10-48-8W5.

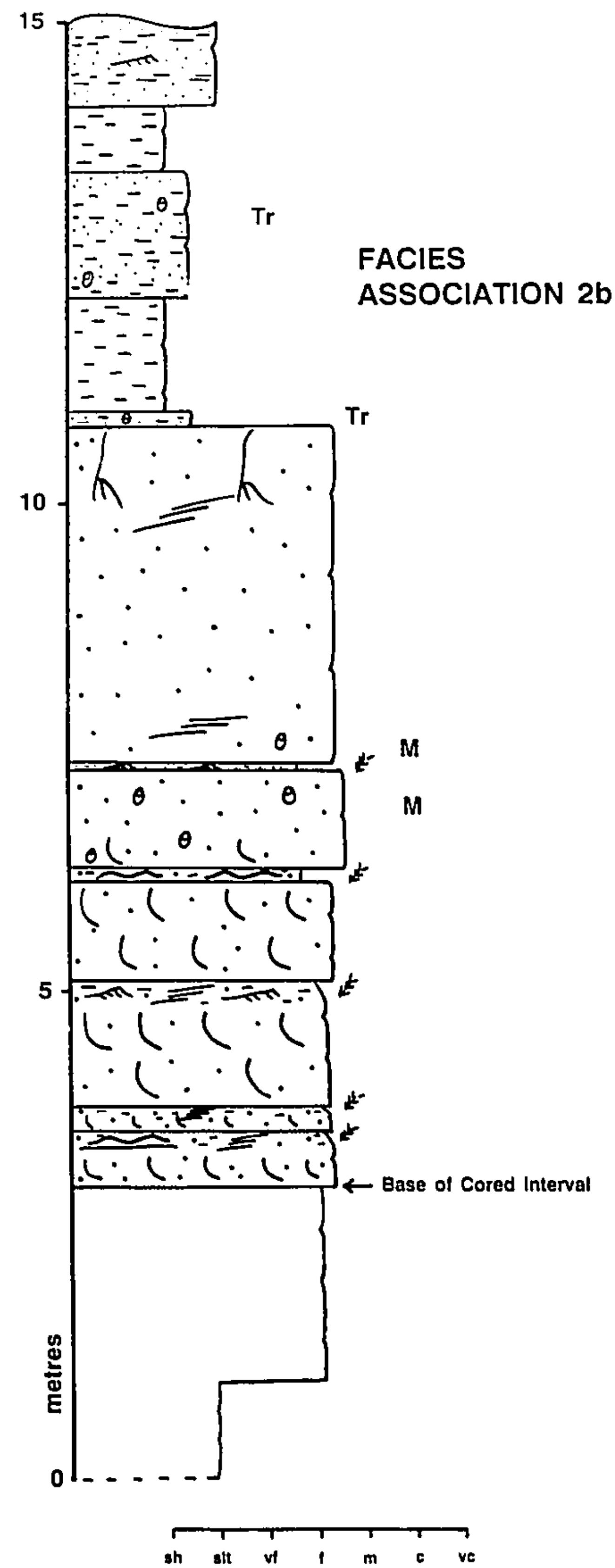
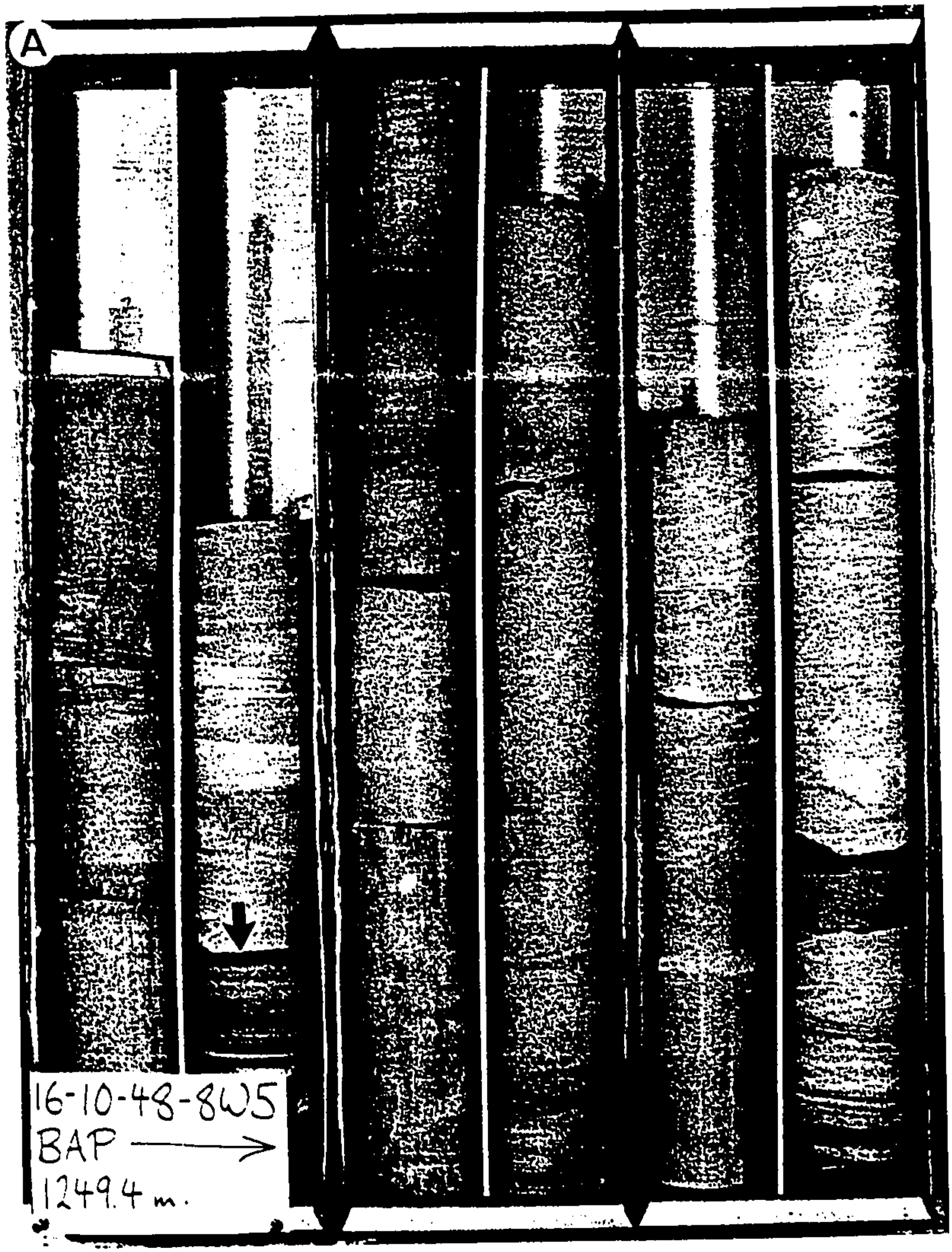

15
Tr
FACIES
ASSOCIATION 2b
Tr
10
M
M
5
Base of Cored Interval
metres
0
sh
slt
vf
f
m
c
vc

Figure 3.18: Box photographs of Facies Association 2b from well 16-10-48-8W5. The sandstone is almost completely cross-bedded above the arrow in Plate A. The top of the succession is rooted and is located at the top of Plate B.


A
16-10-48-8W5
BAP
1249.4 m.


B
16-10-48-8W5
BAP
1244.9 m.

C

bedded, as opposed to the massive sandstones of Facies Association 2a. Trough cross-beds which dip at angles ranging from 10-20 degrees are the dominant sedimentary structure, although come beds contain minor amounts of current rippled and wave-rippled sandstones. This cross-bedded zone ranges from 4-6 m in thickness, and may coarsen upwards. The succession then passes upward into a 3-5 m-thick zone of LAIS and/or flat-bedded fine-grained sandstone which commonly contains roots at the top. This upper portion of the succession is very similar to the upper portions of Facies Association 2a.

The massive to laminated sandstone beds which dominated Facies Association 2a are a minor component, if present at all.

Macrofossils are absent from this facies association. The trace fossils present are similar to those of Facies Association 2a. *Macaronichnus* is the most common trace fossil observed, and is usually located within the upper third of the succession. *Ophiomorpha* and *Planolites* have also been observed.

Interpretation: The coarsening-upward succession capped by rooted sandstones, and the presence of trace fossils such as *Macaronichnus* indicate that these sediments were deposited in a prograding shoreline environment. The abundance of cross-bedded sandstones within the shoreface succession indicates that unidirectional currents were the dominant process of sediment transport in the shoreface environment.

Facies Association 3: Fining-Upwards Channelized Sandstones

This facies association is very common within the study area, particularly in Keystone and Pembina. It ranges in thickness from 0.5 m to over 20 m, and consists of one or more stacked successions of fining-upwards sandstones. A core section of Facies Association 3 is shown in figure 3.19, with the accompanying core photographs in figure 3.20.

The base of each succession is always sharp and usually erosive. This is commonly indicated by the presence of mud clasts (commonly sideritic) up to several cm in diameter which immediately overlie the base (Fig. 3.21a). The succession consists of sandstones ranging in grain size from very fine to medium-grained, commonly forming a fining-upwards succession. Siltstone can also be present within the upper portions of a succession. An ideal succession consists of structureless or vaguely flat-bedded/LAIS sandstone which grades upward into trough cross-bedded sandstone or LAIS sandstone (dip less than 10°) and is capped by current-rippled sandstones or siltstones. Structureless sandstone is often absent from the succession, but trough cross-bedded/LAIS (Fig. 3.21b) and current-rippled sandstone (Fig. 3.21c) are almost always present, with cross-bedding being the dominant component. The definition of stratification in the cross-bedding and the current ripples is usually quite good, in contrast to the stratification in Facies Association 2. The association itself can consist of a single fining-upwards succession to as many as 10-15 stacked successions. The thickness of each succession ranges from less than 1 m to

Figure 3.19: Stratigraphic section of Facies Association 3. Based on core in well 10-30-47-3W5.

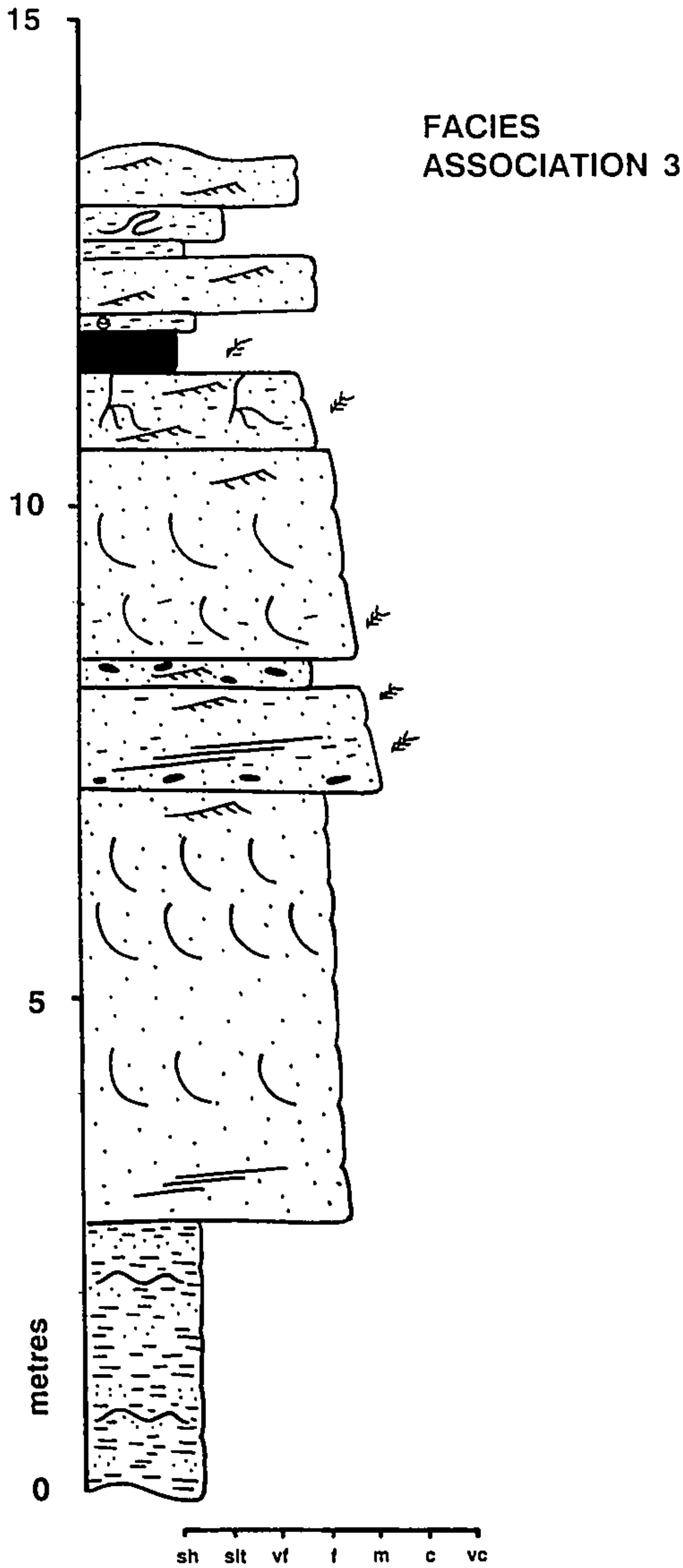
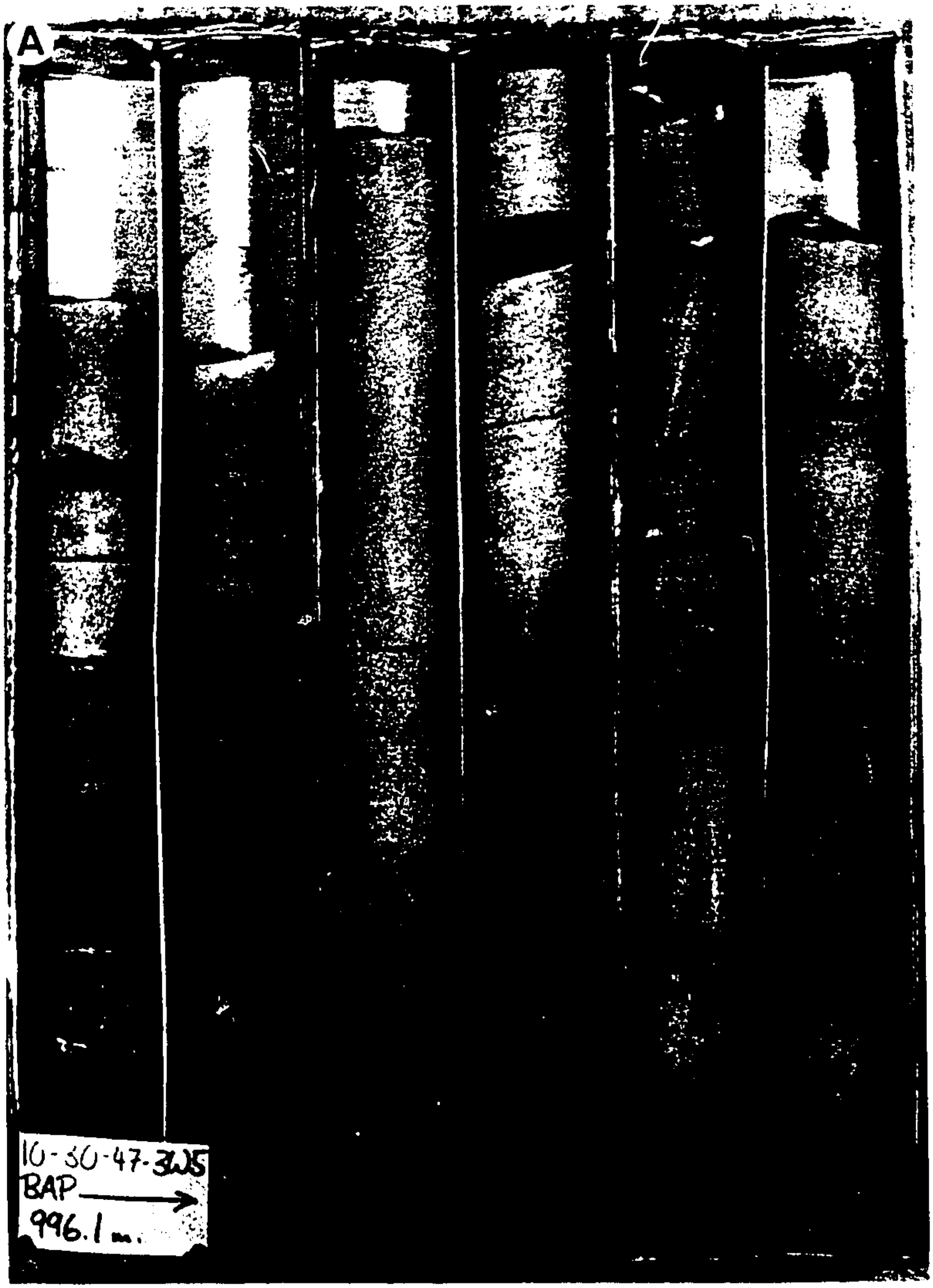

FACIES
ASSOCIATION 3
15
10
5
0
metres
sh
slt
vf
f
m
c
vc

Figure 3.20: Box photographs of Facies Association 3 in well 10-30-47-3W5. The base of the succession in each plate is at the bottom left hand corner and the top of the succession is at the top right hand corner. Each core tube is 75 cm in height. The base of the channel sandstone is shown by the arrow in Plate A. Several intervals of mud clasts can be seen in Plate B. Plate C shows the upper transition of Facies Association 3 from cross-bedded into finer-grainea current-rippled sandstone, and subsequently into the overbank sediments of Facies Association 5.


A
10-30-47-3W5
BAP
996.1 m.


B
10-30-47-3W5
BAP
991.6 m

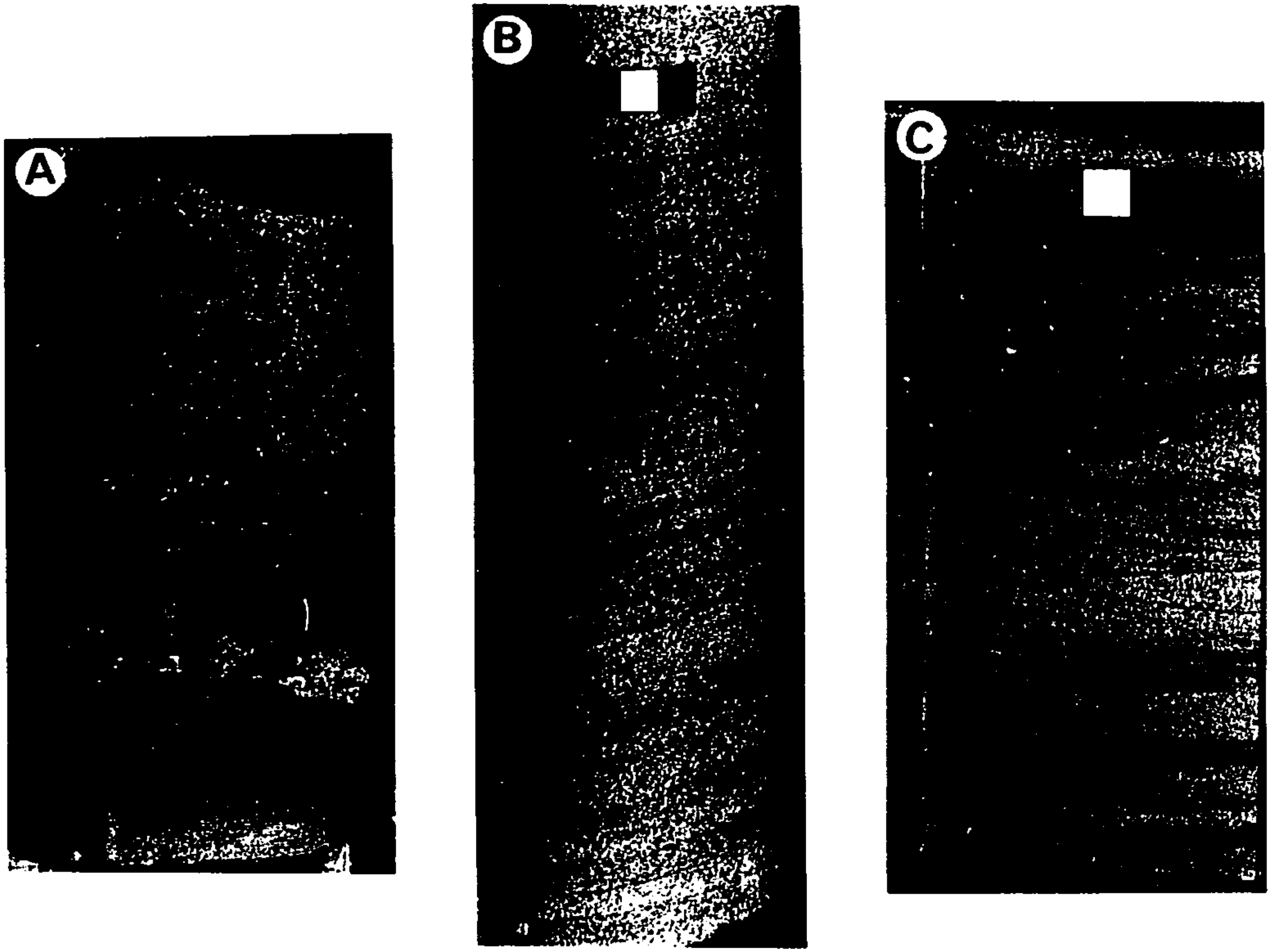
C

Figure 3.21

(A) Base of channel sandstone (shown by arrow) in Facies Association 3 characterized by rip-up clasts of mudstone. The core is 3.5 cm in width. Location: 6-36-47-3W5; 1018.9 m.

(B) Cross-bedded sandstone in Facies Association 3. Note the improved definition of lamination in comparison to the cross-bedding in Facies Association 2a. Location: 10-25-48-5W5; 975.8 m.

(C) Current-rippled sandstone. In this case the current-ripples are climbing ripples. Location: 10-947-2W5; 1010.4 m.


A
B
C

almost 10 m.

Plant matter is very common within the sandstones, and is usually concentrated into distinct laminations within the cross-bedded or current-rippled sandstones (Fig. 3.22a). These laminations sometimes occur as distinct couplets or pairs (Fig. 3.22b), similar to those of Facies Association 1a or 2. Mud clasts, both angular and rounded, are common features of this facies association. They can occur within a succession (Fig. 3.23a) as well as at the base, and can be as large as the diameter of the core. Root traces are often present, especially near the top of a succession.

Macrofossils are absent form this facies association. Bioturbation is generally absent , but thin, mud-lined *Trichichnus* or *Skolithos* burrows up to several cm long and a few mm wide are not uncommon (Fig. 3.23b). It is sometimes difficult to distinguish between these burrows and sand-filled root traces.

Interpretation: The fining-upwards successions with erosive bases and the general lack of marine indicators such wave ripples or distinctive trace fauna indicates that these sediments were most likely deposited within channels in a non-marine or coastal environment.

Facies Association 4: Pebbly Channelized Sandstones

Facies Association 4 is present only in the western portion of Pembina and Ferrier - Willesden Green, where it is the most common facies association.

Figure 3.22

(A) Laminations of macerated plant fragments within channel sandstones.

Location: 8-10-48-3W5; 961.4 m.

(B) Paired organic laminations in channel sandstones (see text for discussion). Location: 14-28-48-6W5; 1078.5 m.

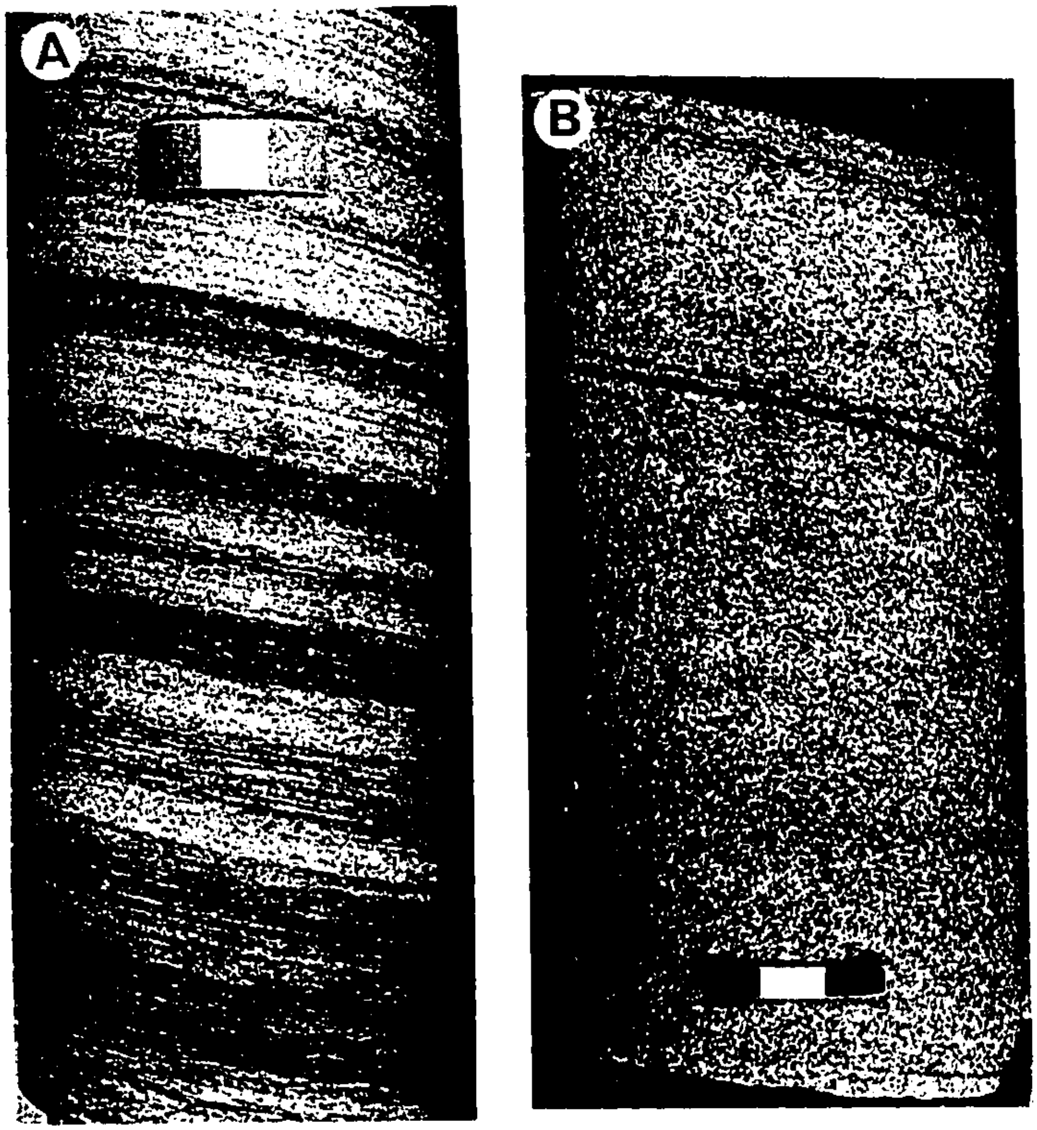
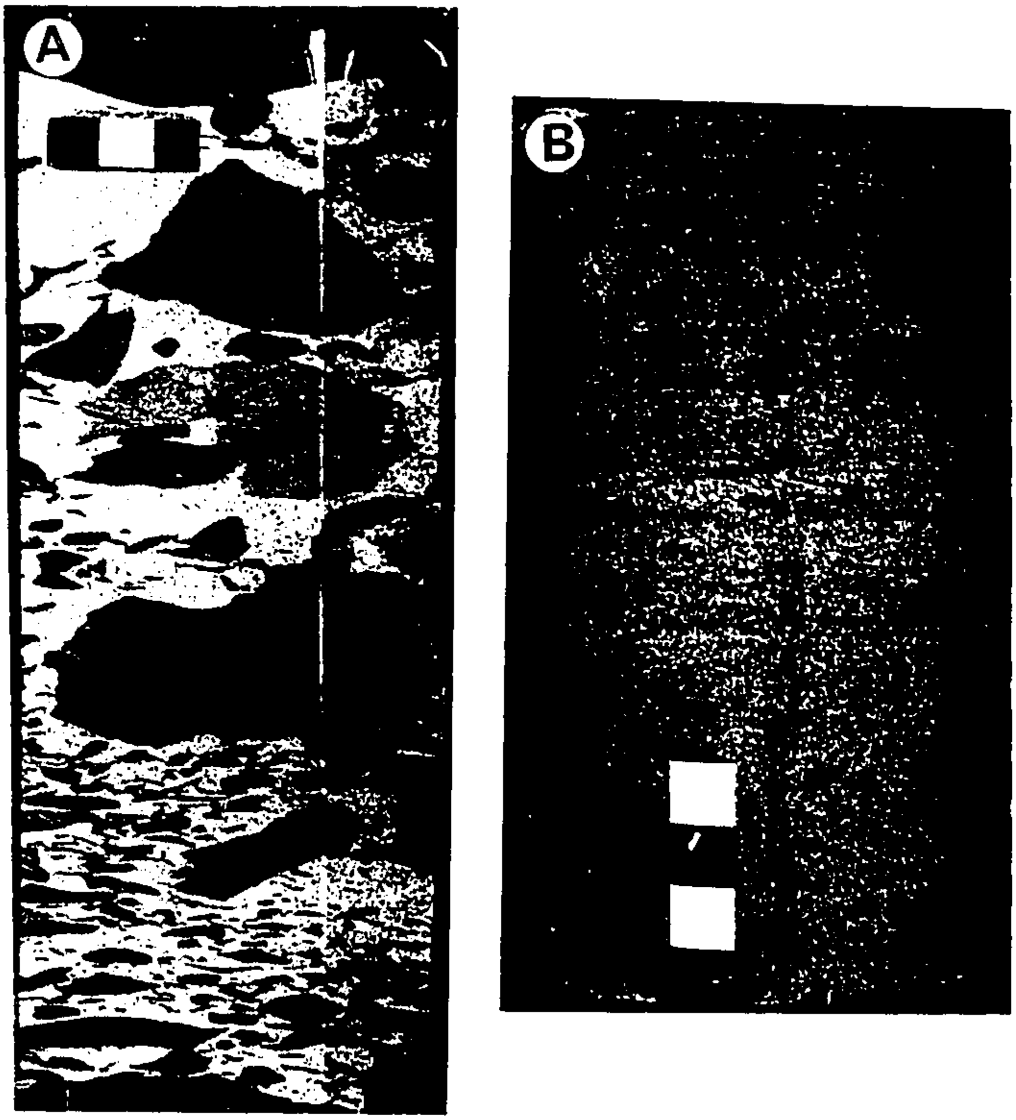

A
B

Figure 3.23

(A) Angular rip-up clasts of mudstone within channel succession.

Location: 4-36-47-4W5; 998 m.

(B) *Trichichnus* burrows within channel sandstones. Location: 8-8-48-3W5; 934.6 m.


A
B

It can range from 2->20 m in thickness. It consists of fine to medium grained pebbly sandstones which can occur in both fining-upwards and coarsening-upwards sequences, as well as units which contain no large-scale vertical trend in grain size. Figure 3.24 shows a core section of this facies association, with figure 3.25 showing the accompnaying core photographs.

The base of the association is usually sharp and erosive, with mud clasts being fairly common immediately overlying the basal contact. The association itself is dominated by stacked beds of pebbly sandstones in which the percentage of pebbles ranges from <1-50% of the volume, averaging between 5-20%. The pebbles are < 2 cm indiameter and are composed of chert. In rare instances, the percentage of pebbles within the sandstones exceeds 50% (Fig. 3.26a), and the rock becomes a sandy, matrix- or clast-supported pebble conglomerate. Fine- to medium-grained sandstones lacking any pebble component are also fairly common within this facies association. Structureless sandstones (Fig. 3.26b) are equally as common as stratified sandstones. Sedimentary structures present within the sandstones consist mainly of trough cross bedding or LAIS (dip <10°) (Figs. 3.26c, 3.27a, 3.27b). Other structures include flat lamination (Fig 3.27c) and current ripples, which are only present within the fine grained sandstones lacking any pebbles.

Sandstone beds are mostly in the range of 0.5-2 m in thickness. Internal successions can show both fining-upwards and coarsening-upwards trends. Fining-upwards successions are the most common, and usually consist of

Figure 3.24: Stratigraphic section of Facies Association 4. Based on cored section in well 10-26-42-8W5.

FACIES ASSOCIATION 4

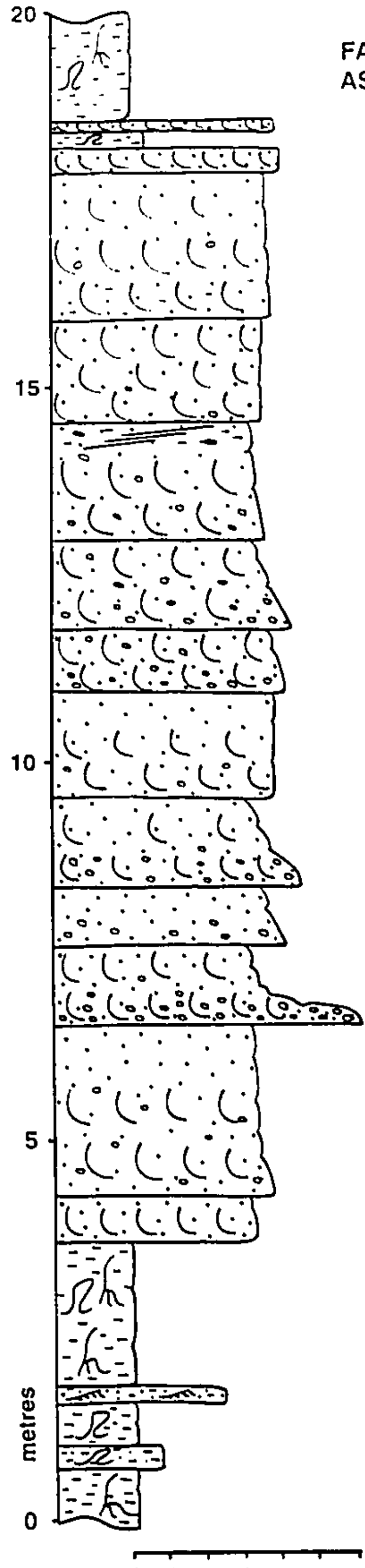
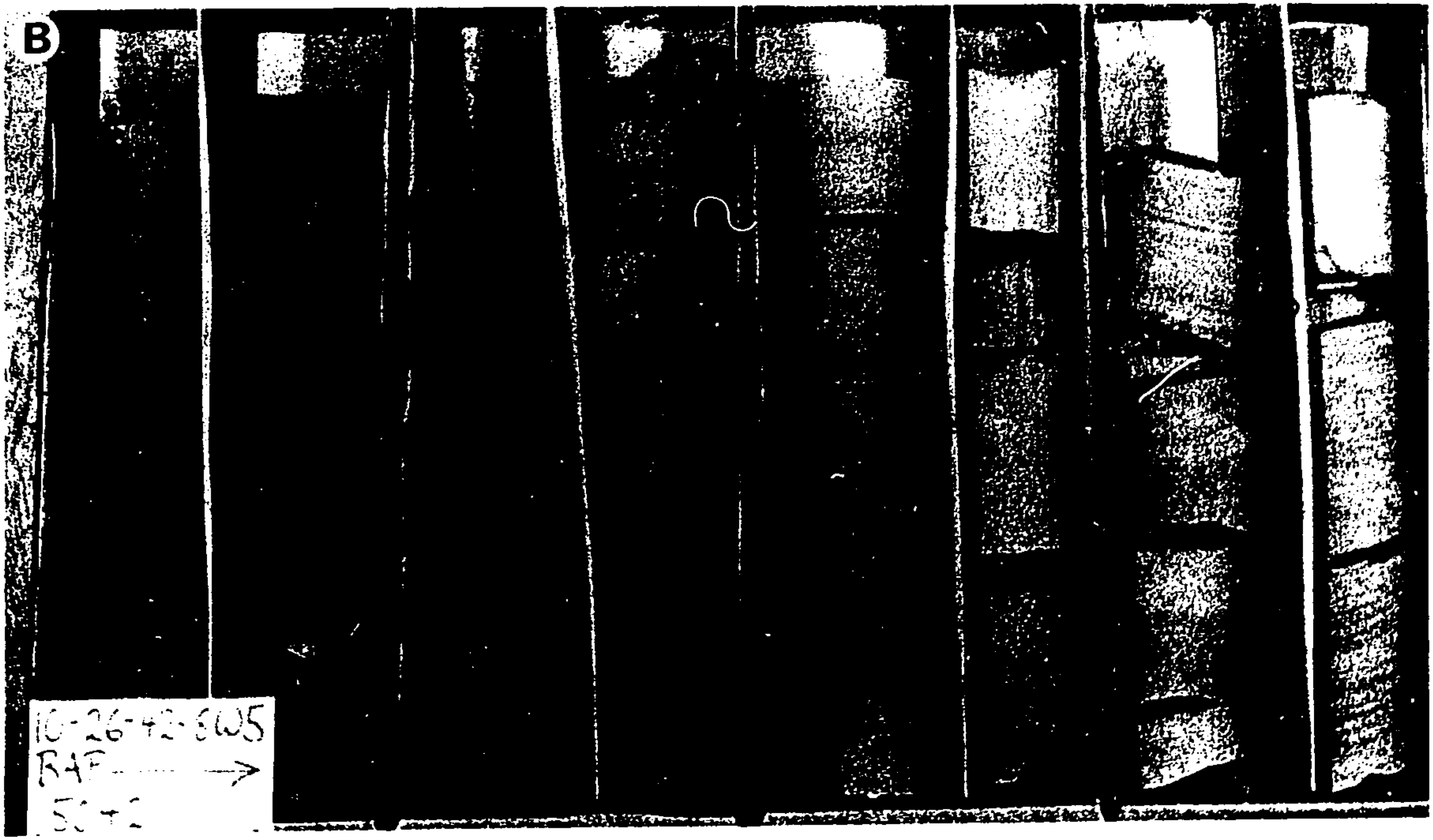
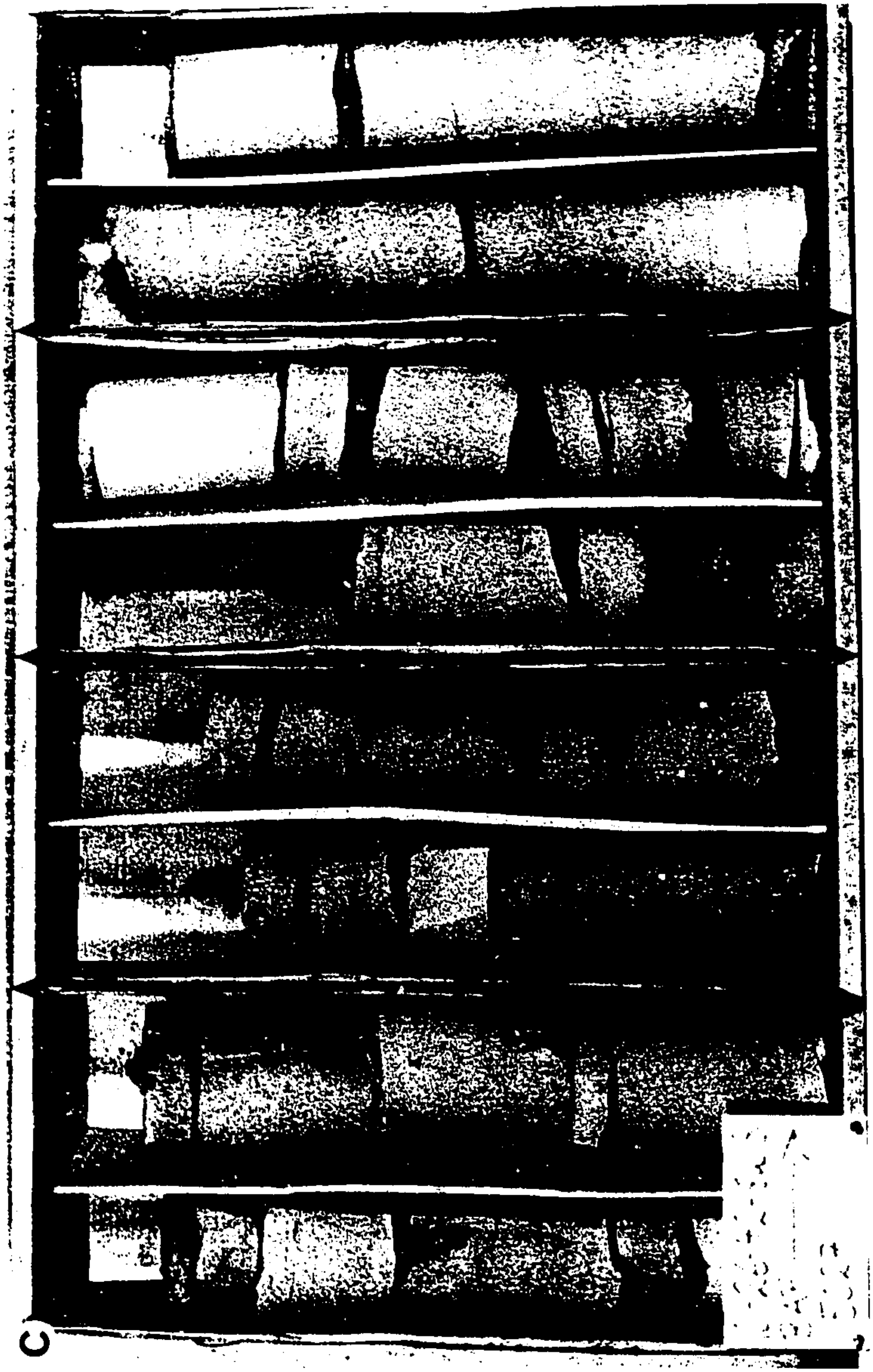


Figure 3.25: Box photographs of Facies Association 4 in well 10-26-42-8W5. The base of the succession in each plate is at the bottom left hand corner and the top is at the top right hand corner. Each tube of core is 60 cm in height. The base of the channel is shown by the arrow in Plate A. Several thin pebbly sandstone intervals can be seen in Plate B. The sandstone also becomes cross-bedded towards the top of Plate B. Plates C and D show cross-bedded pebbly sandstone.


A


B
10-26-42-6W5
RAF



C

D

Figure 3.26

(A) Massive pebble conglomerate in channelized succession of Facies Association 4. Location 16-18-45-6W5; 1314 m.

(B) Massive sandstone with minor amount of small pebbles. Location: 16-18-45-6W5; 1308 m.

(C) Cross-bedded medium-grained sandstone with minor amount of small pebbles. Location 7-28-32-28W4; 1206 m.

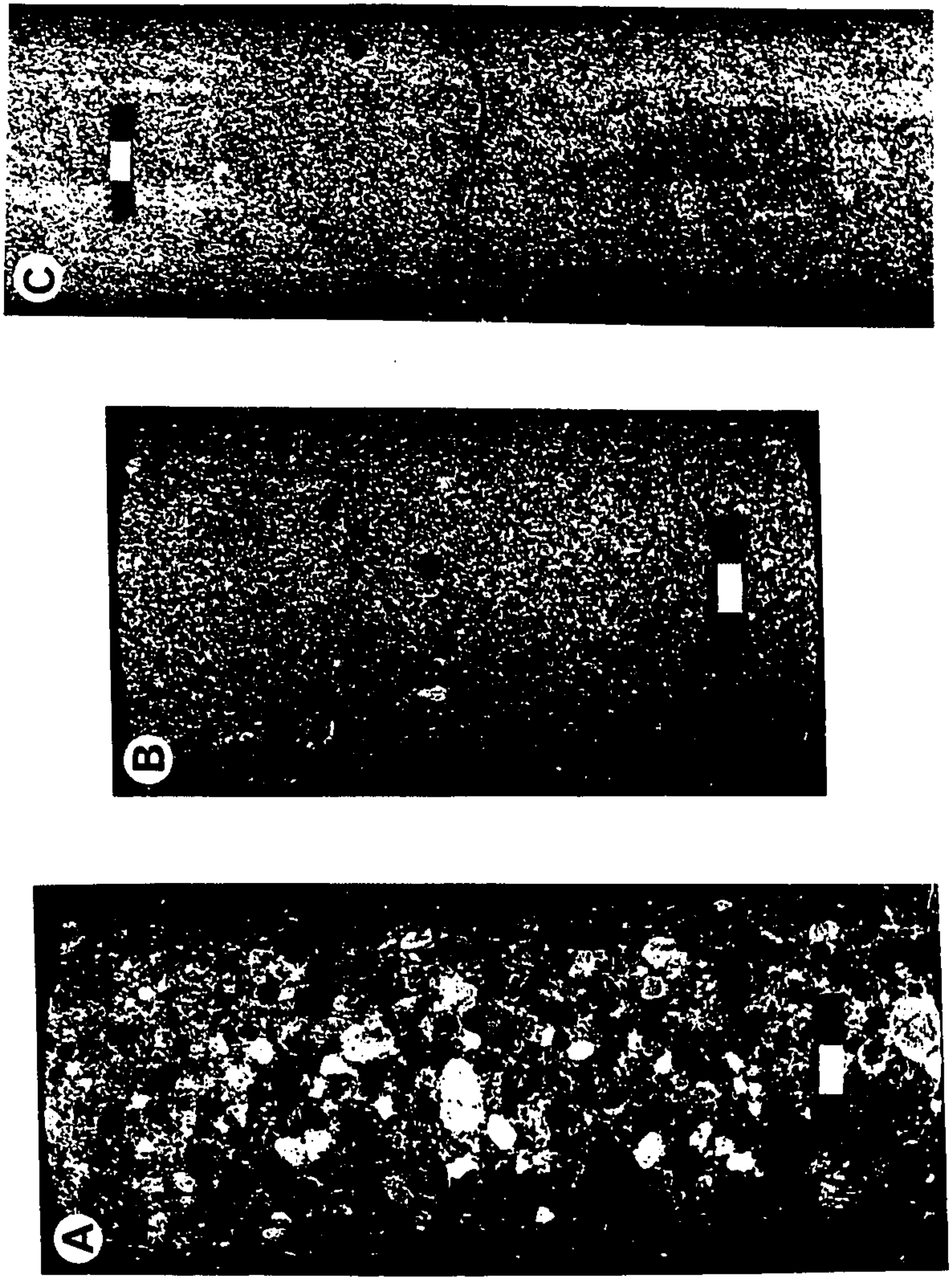
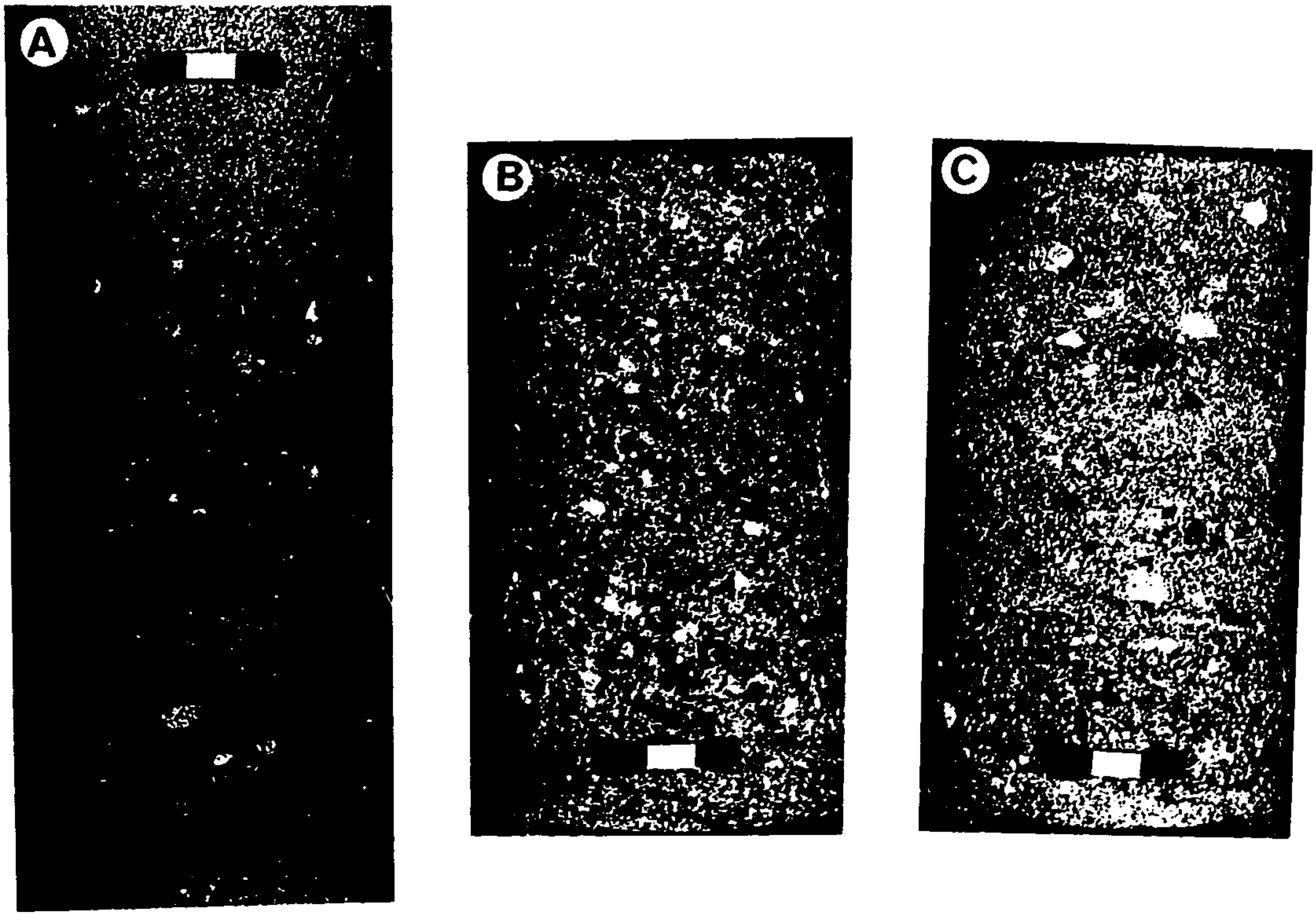

C
B
A

Figure 3.27

(A) Cross-bedded/LAIS pebbly sandstone with approximately 10% pebble content. Location: 4-17-41-8W5; 1714 m.

(B) Coarse-grained cross-bedded pebbly sandstone with 25-35% pebbles. Location 16-18-45-6W5; 1307 m.

(C) Flat-bedded coarse-grained pebbly sandstone with 15-20% pebbles. Location 16-18-45-6W5; 1308 m.


A
B
C

massive pebbly sandstone grading up into cross-bedded or LAIS sandstone with fewer pebbles. In many beds, the bulk of the pebbles occur within a few tens of cm from the base, and quickly decrease in abundance upwards. Coarsening-upwards successions are not very common, but when present occur as massive or cross-bedded pebbly sandstones in which the percentage of pebbles increases upwards. Beds with no vertical succession in grain size or sedimentary structures are also common. A typical example of this facies association will contain between 5 and 15 stacked beds. The contacts between beds are usually sharp and often characterized by sharp changes in grain size between the two beds.

Macrofossils are absent from this facies association, and no identifiable trace fossils or other evidence of faunal bioturbation were observed. Root traces are not common, and when present usually occur in finer-grained sediments near the top of the association.

Interpretation: The coarse grain size suggests that the sediments were deposited by high energy currents moving sand and gravel as bedload. The sharp, erosive base of the association, general lack of a large-scale fining-upwards or coarsening-upwards sequence, the numerous sharply-based beds of varying grain size, lack of marine bioturbators, and evidence of subaerial exposure (roots) all indicate that these sediments were deposited within non-marine channelized, high energy environments in which channel switching may have been rapid, and where channels were temporally unstable. This would allow for the stacking of numerous channel fill deposits. Alternatively, the numerous channel fill successions may represent successive fill events within the same channel. Without outcrop data, it is difficult to distinguish the true nature of these channels.

Facies Association 5: Coastal Plain/Floodplain Sediments

This association is very common in sediments of the Belly River Group, particularly in the Pembina-Keystone areas. The association can range in thickness from a few tens of cm to 15 m in observed cores. It consists of mudstones, sandstones, siltstones, coals, and paleosols which are interbedded on a variety of scales. This facies association can contain any or all of the following facies, and there is no predictable vertical facies succession. The vertical order of facies in any given example will vary. Figures 3.28 and 3.29 show examples of a core section and core photographs an example of this facies association.

This association is dominated by two types of background sediments. The first consists of interbedded mudstones, siltstones, and very fine to medium-grained sandstones which are interbedded or interlaminated on a scale of a few mm up to 10 cm (Fig. 3.30a). The interbedding is usually on a mm scale, which contrasts with the scale of interbedding in the background sediments of Facies Association 1a. This gives the rock a characteristic "pinstriped" appearance. Mudstone and siltstone are usually the dominant

Figure 3.28: Stratigraphic section of Facies Association 5. Based on cored section in well 10-1-49-7W5.

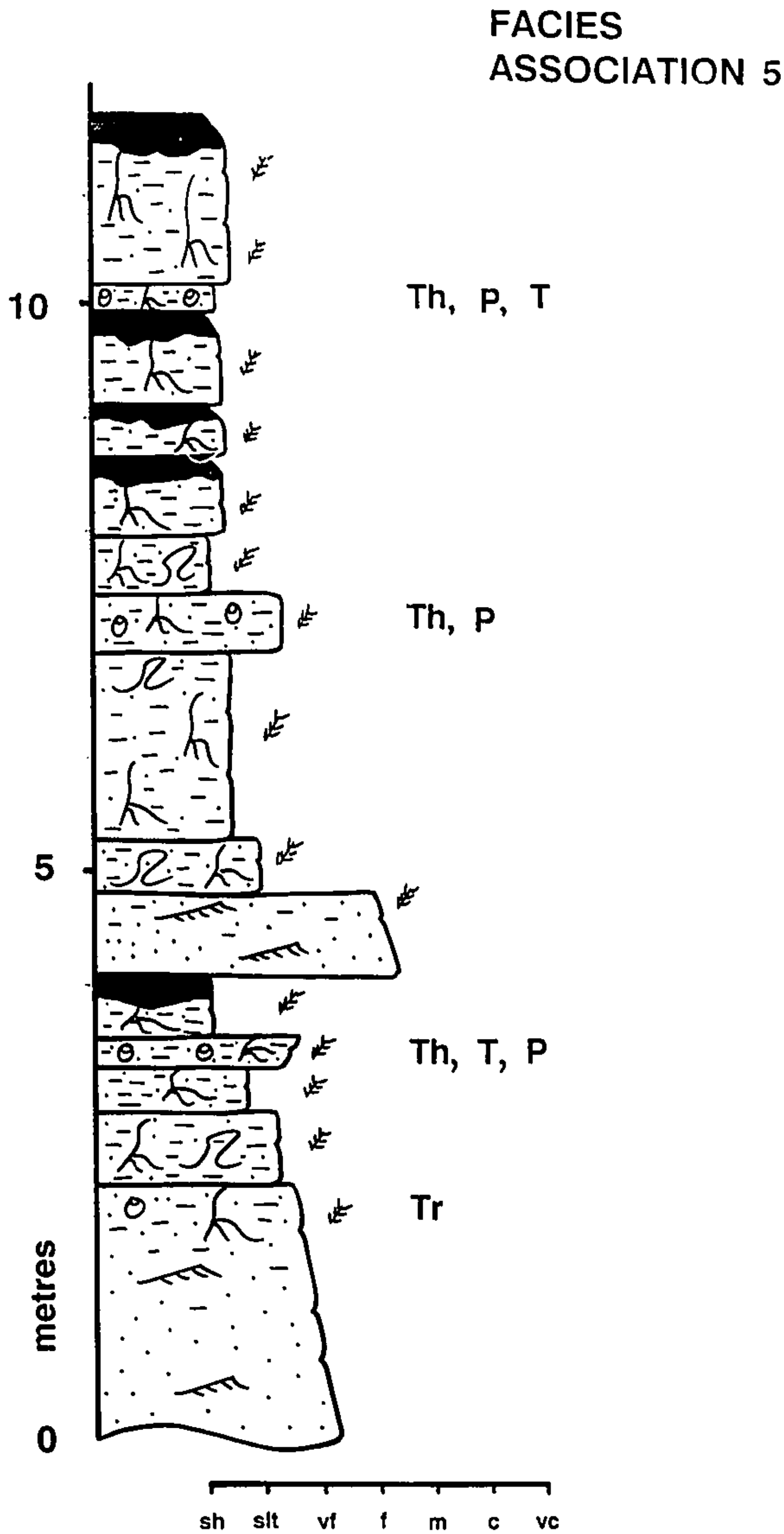
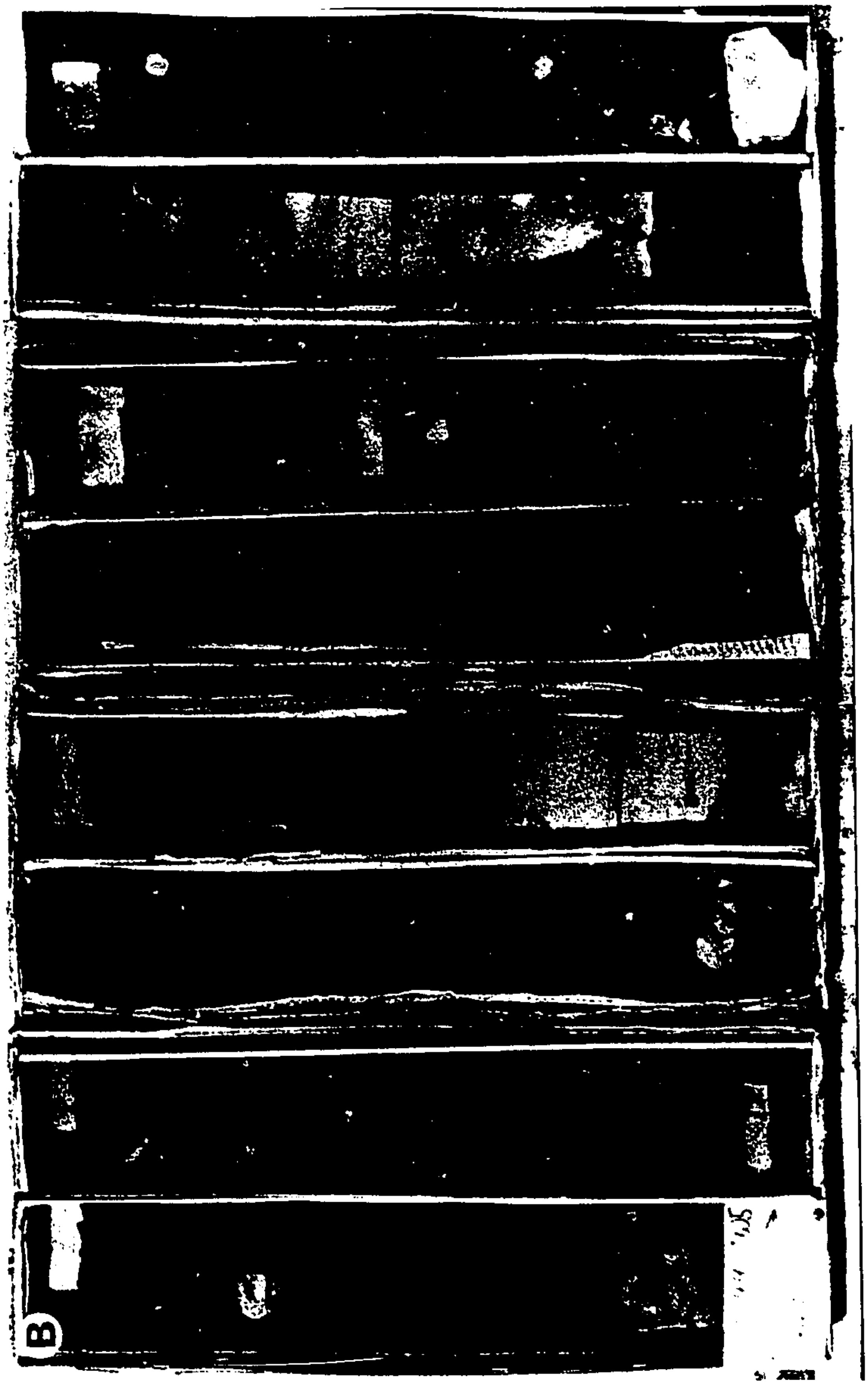

FACIES
ASSOCIATION 5
10
Th, P, T
Th, P
5
Th, T, P
Tr
metres
0
sh
slt
vf
f
m
c
vc

Figure 3.29: Box photographs of Facies Association 5 in well 10-1-49-7W5. The base of the succession in each plate is at the bottom left hand cornet and the top is at the top right hand corner. Each tube of core is 60 cm in height. Plate A begins showing the upward transition from a current-rippled channel sandstone into the muddy overbank sediments, which are capped by a coal bed (shown by arrow). Plates B and C are dominated by organic-rich fine-grained mudstones. The tops of three coal beds are shown by the arrows.


A
10-1-49-7W5
BAP
3486'


B

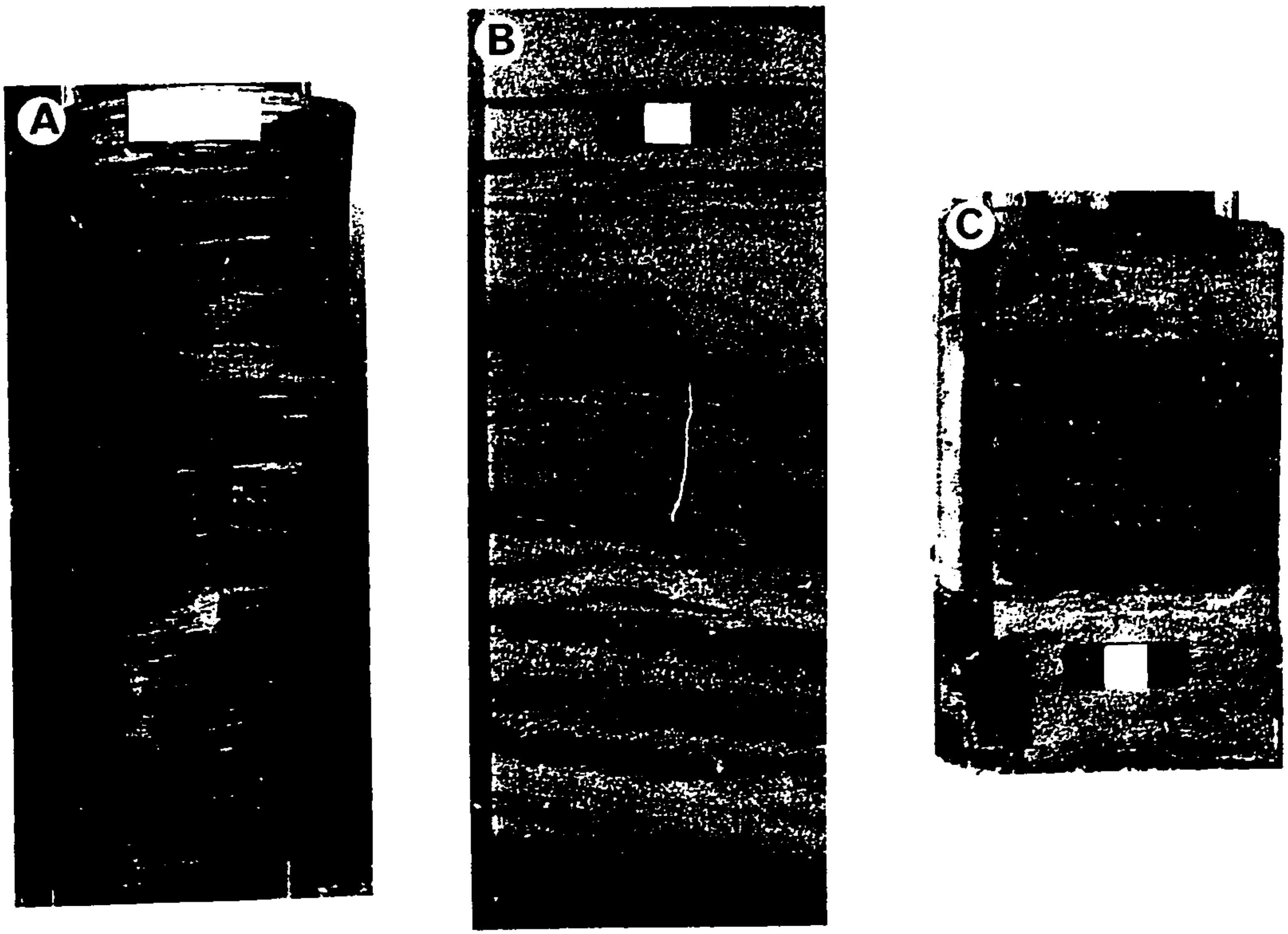
C
10-1-49-7W5
PAT
3454'

Figure 3.30

(A) Interlaminated mudstone and siltstone in Facies Association 5. Location: 8-11-48-4W5; 996.4 m.

(B) Current ripples in fine-grained sandstone. Location: 16-19-48-6W5; 1099.5 m.

(C) Structureless, organic-rich mudstone. Location: 7-15-47-2W5; 943.9 m.


A
B
C

component of this facies, with sandstone rarely exceeding 25%. On occasion, however, the volume of sandstone will exceed 50%. Sedimentary structures within the sandstones and siltstones are mainly current ripples of 1-2 cm in amplitude (Fig. 3.30b), sometimes expressed as climbing ripples. Flat lamination is also present, but is not as common. Units of this facies can range from ten cm to several m in thickness.

The second dominant background facies consists of massive, structureless silty mudstone (Fig. 3.30c), in which the percentage of silt ranges from <5-50%. This facies is usually a few tens of cm thick, but can be up to 2-3 m thick.

Macerated plant fragments and well-preserved root traces are very common within both of these facies (Fig. 3.31a). Siderite is also very common within both facies, being present as bands up to several cm thick, or as nodules. Faunal bioturbation (as opposed to root bioturbation) is rare within the background sediments of this facies association, but numerous small, thin *Trichichnus* or *Skolithos* burrows up to a few cm in length and a few mm in width are present within the mm-scale interbedded mudstones, siltstones, and sandstones (Fig. 3.31b). There are also a few places in which the interbedded and massive mudstones contain a suite of trace fossils which may include *Planolites, Skolithos, Teichichnus* and *Thalassinoides* (Figs 3.32a, 3.32b). The latter is the most common trace in these places, and is quite diagnostic of this sub-facies. Soft sediment deformation is very common within both of these

Figure 3.31:

(A) Rooted, silty, organic-rich mudstone. Roots shown by arrow

Location: 8-21-47-2W5;

971.5 m.

(B) Small vertical *Trichichnus* burrows in fine-grained

sandstone/mudstone. Location: 10-27-47-4W5; 975 m.

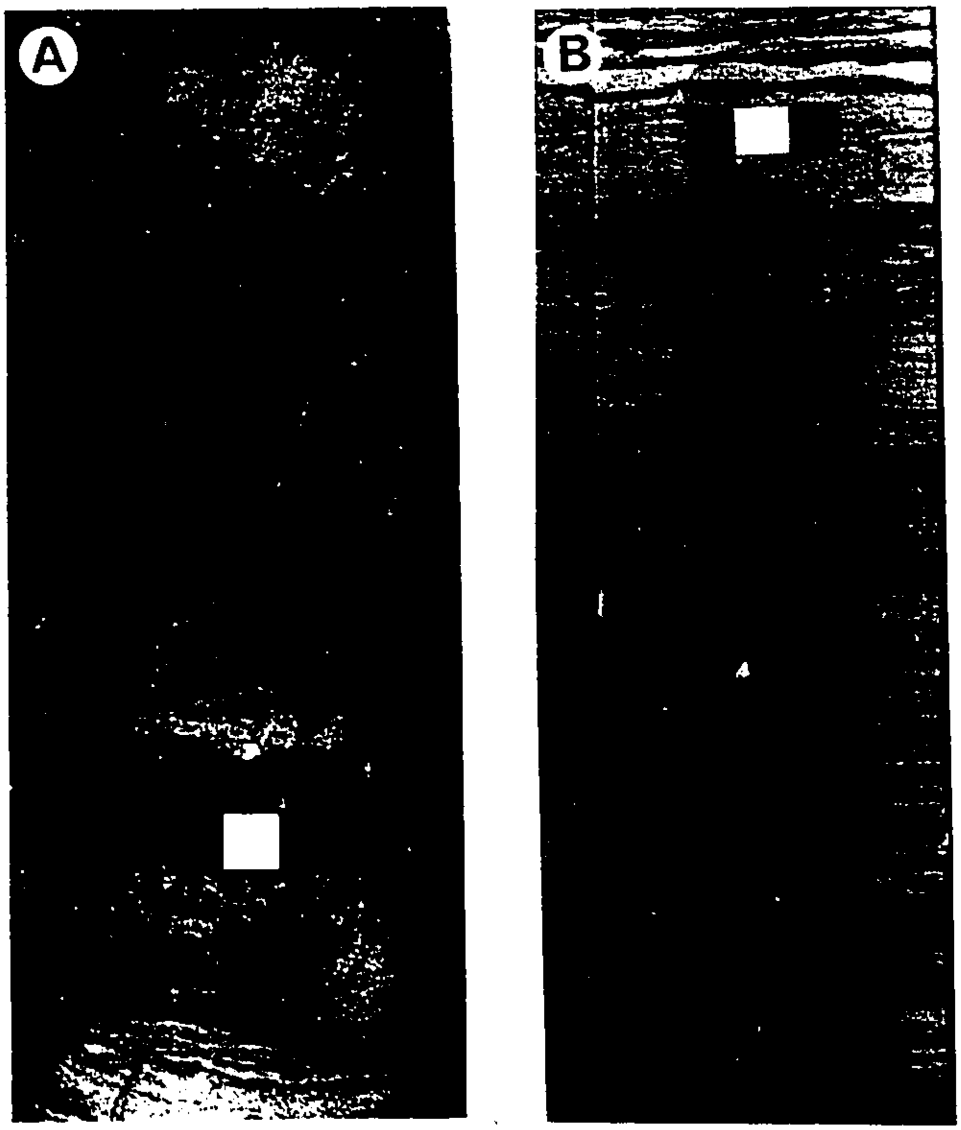
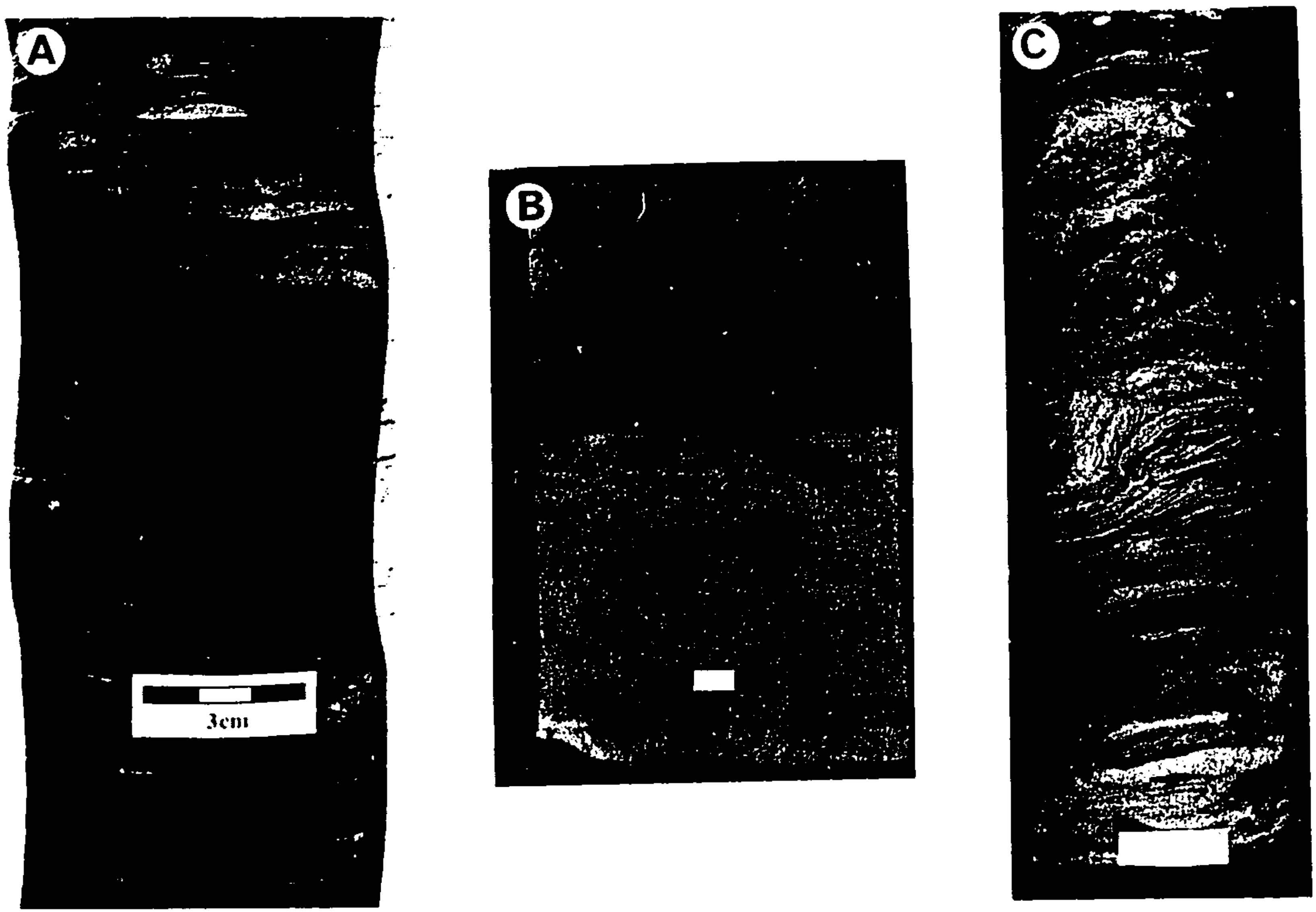

A
B

Figure 3.32

(A) *Teichichnus* burrows in coastal plain sediments. Location: 16-29-47-2W5; 1007 m.

(B) *Thalassinoides* burrows in coastal plain sediments. Location: 10-1-49-7W5; 1059.1 m.

(C) Soft-sediment deformation in coastal plain mudstones. Deformation is more pervasive than in marine mudstones. Location: 6-6-48-3W5; 991.9 m.


A
B
C
3cm

facies, and often appears to have completely churned up the sediment (Fig. 3.32c).

Numerous other facies are interbedded with the background sediments of this facies association. Cross-bedded and/or current rippled very fine to fine-grained sandstone beds are common. These beds range in thickness from 10 cm to a few tens of cm. If the thickness of one of these beds exceeds 50 cm, it is classified as being a thin example of Facies Association 3. The bases of these beds are usually very sharp and erosive, and they often contain mud clasts up to 1 cm in diameter. The beds commonly fine upwards, and cross-bedded sandstone will commonly grade upwards into current-rippled sandstone. Plant fragments and root traces are also common within these beds.

Coal is also a common component of this facies association (Fig. 3.33a). Coal beds are usually thin, ranging in thickness from a few cm to a few tens of cm. They usually are associated with rooted, organic-rich silty mudstones.

Paleosols (Fig. 3.33b) are present within the facies association, but are not common. They are usually only a few cm to a few tens of cm thick, and consist of green-gray, organic-rich mudstones which are very crumbly and often contain waxy slickenside-like surfaces. Root traces are common.

The final component of this facies association are black, massive carbonaceous shales, which often contain abundant preserved shells of bivalve oysters (Fig 3.33c, 3.33d). Preserved plant fragments are also common.

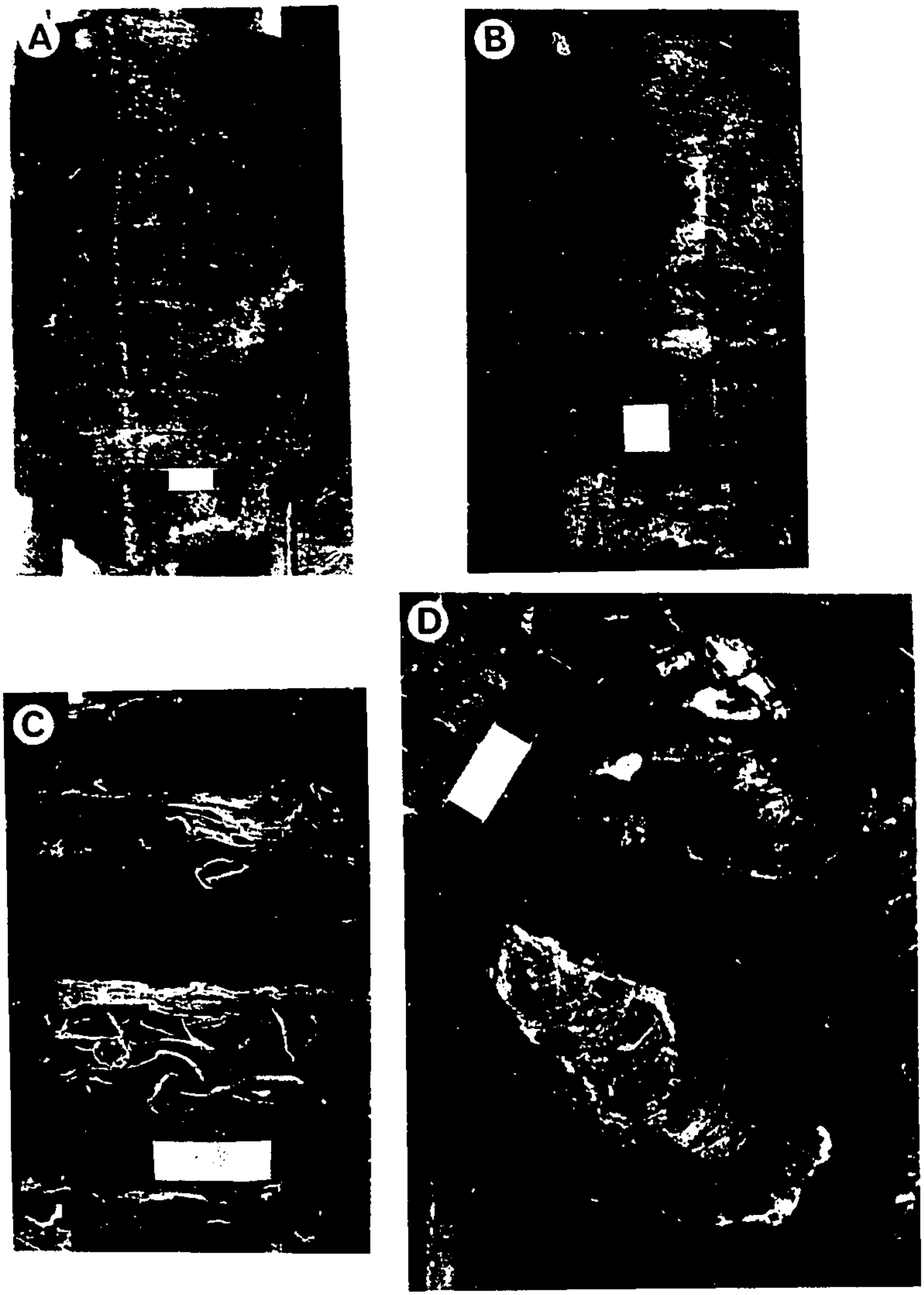
Figure 3.33

(A) Coal. Location: 15-17-49-9W5; 1307.6 m.

(B) Paleosol. Location: 11-19-48-2W5; 900.1 m.

(C) Black, organic-rich mudstones containing oyster shells. Location: 2-2-47-5W5; 975.5 m.

(D) Plan view of oyster shell found in black mudstones. Location: 16-16-36-1W5; 1187.8 m.


A
B
C
D

Visible bioturbation is absent. This facies is not common, and is present only in small, localized areas. It seldom exceeds a few tens of cm in thickness

Interpretation: This facies association is interpreted to represent deposition in a low energy non-marine to coastal plain environment. Coals, rooted mudstones, rooted interbedded mudstones, siltstones and sandstones are interpreted to be deposited in non-marine shallow marshy areas, such as delta plain or fluvial overbank environments. Massive, silty mudstones may be deposited in deeper ponded environments such as ponds or lakes. Periodic crevasse splays from nearby channels will bring in sharply-based cross-bedded and current-ripples sandstones. Periodic drying out of these environments and soil development allows for the formation of the paleosols. Occasional invasions of marine water due to minor transgression or storm flooding of coastal areas will allow for the deposition of black shales with oysters or for incursion of marine bioturbators such as *Thalassinoides* or *Teichichnus* into the non-marine environment. These marine incursion are short-lived, and non-marine conditions are quickly re-established.

Facies Association 6: Bioturbated Transgressive Sediments

This facies association is locally common in the eastern portions of Keystone and in Ferrybank. This association consists of a fining -upwards succession, and never exceeds 1-2 m in thickness. A core section of Facies Association 6 is shown in figure 3.34 and core photographs are shown in figure

Figure 3.34: Stratigraphic section through Facies Association 6. Based on cored section in well 14-24-46-2W5.

FACIES ASSOCIATION 6

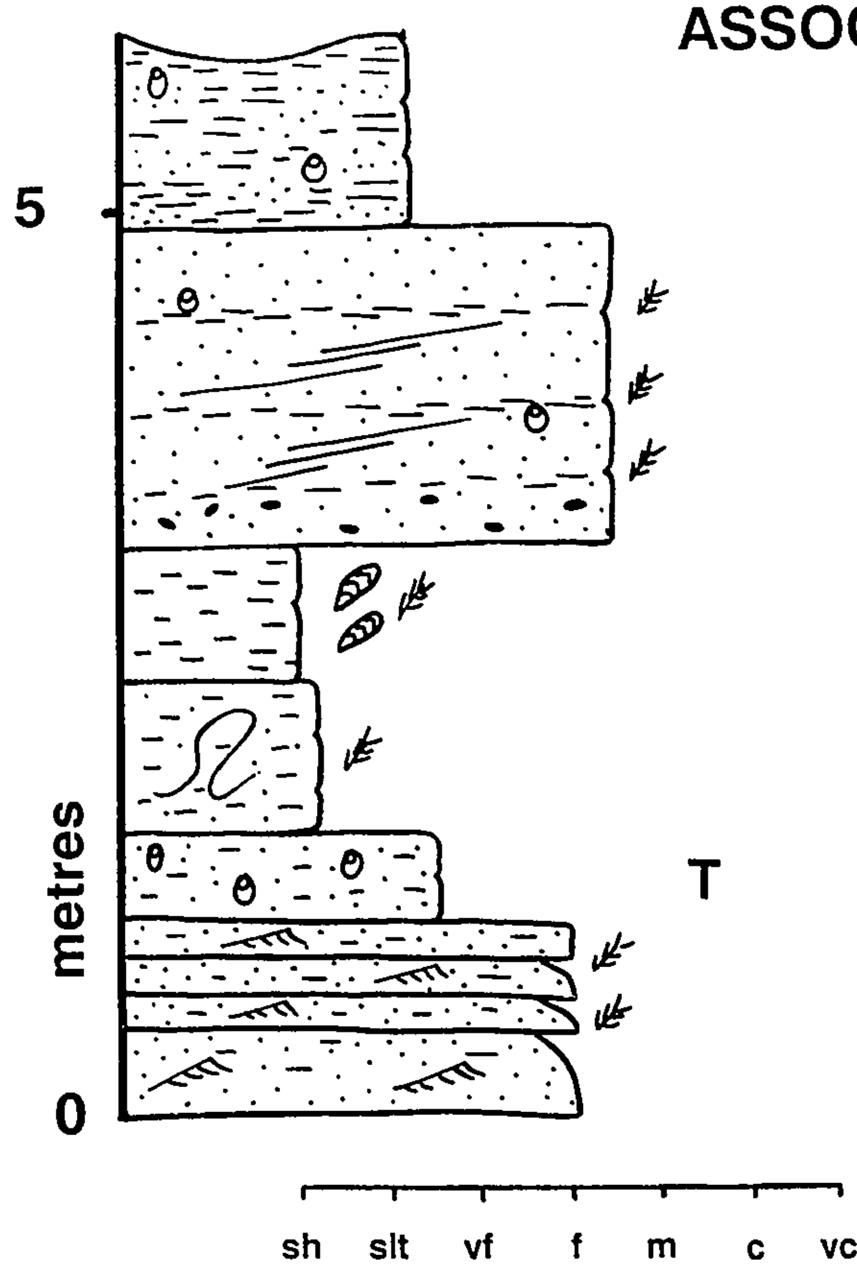
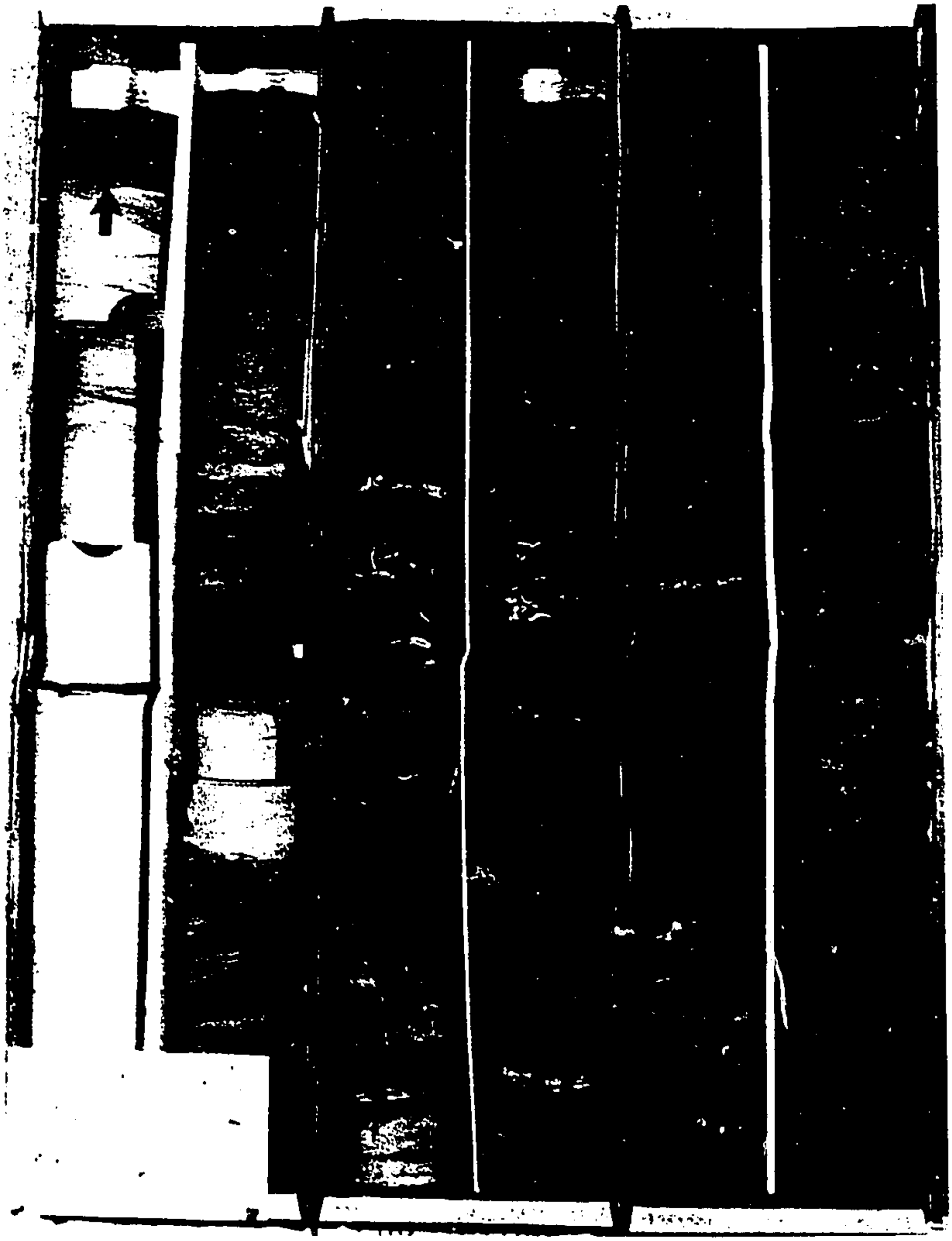


Figure 3.35: Box photograph of cored section through Facies Association 6 in well 6-16-46-1W5 (Note: not the same location as the stratigraphic section). Transgression begins at arrow marked, and gradually passes up through brackish water sediments (containing the oyster shells) into fully marine sediments in the top 75 cm of the succession.



3.35.

The association consists dominantly of a unit of moderately to heavily-bioturbated sandy mudstones or muddy sandstones. The percentage of mudstone ranges from 30-70%, with the remainder being very fine to medium-grained sand (Fig. 3.36a, 3.36b). Bioturbation is intense enough so that no beds or lamina are preserved, giving the rock a churned appearance (Fig. 3.36a). Distinct trace fossils are hard to find within this facies, perhaps because it has been more heavily bioturbated. The most common trace visible is *Planolites*. Also present, but much scarcer, are *Skolithos, Thalassinoides, Teichichnus,* and *Chondrites*. Organic matter is very common within this facies, being disseminated throughout by the bioturbation. This unit never exceeds 75 cm in thickness, and is usually less than 50 cm thick.

The bioturbated mudstones/sandstones may pass upwards into black, carbonaceous shales containing abundant oyster shells (Fig. 3.36c), however, this facies is not always present. This facies is identical to the shale with oyster shells of Facies Association 5. It is usually only a few tens of cm thick.

Sharp-based, massive or LAIS fine-grained sandstone beds with abundant organic debris are occasionally present within the black shale with oysters. These beds can contain evidence of bioturbation, but no distinct traces can be identified (Fig. 3.36d) These beds are always less than 2 m in thickness.

This facies association is always overlain by either Facies Association 1a

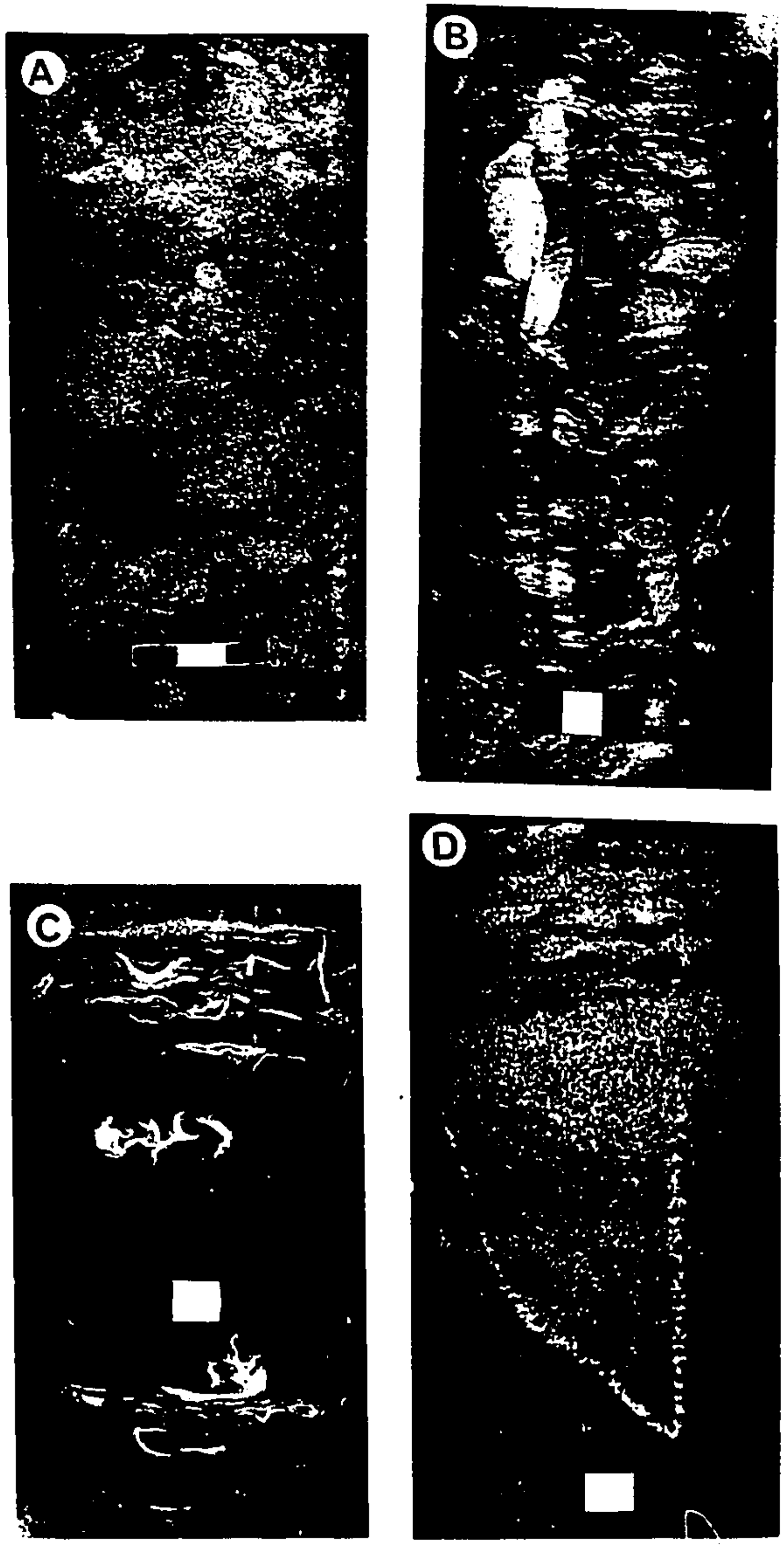
Figure 3.36

(A) Thoroughly bioturbated muddy sandstone overlying transgressive flooding surface. Location: 14-6-43-27W4; 941.7 m.

(B) Thoroughly bioturbated sandy mudstone overlying flooding surface. Location: 6-9-47-3W5; 975.8 m.

(C) Black mudstones with oyster shells. Location: 6-16-46-1W5; 994.0 m.

(D) Moderately bioturbated muddy sandstone from channel cutting into brackish sediments. Location: 6-23-46-2W5; 990.5 m.


A
B
C
D

or 1b.

Interpretation: This facies association is interpreted to represent sediments deposited during transgressive conditions. The unit of bioturbated mudstones/sandstones are interpreted to represent sediments from underlying deposits which were reworked by bioturbating organisms as the marine waters first began to transgress. In some places, this marine water allowed for the development of brackish bays or lagoons in which black shales containing oysters were deposited. Channels cut through the lagoon in places, depositing massive or LAIS sands.

CHAPTER 4

STRATIGRAPHY OF THE LEA PARK - BELLY RIVER TRANSITION IN CENTRAL ALBERTA

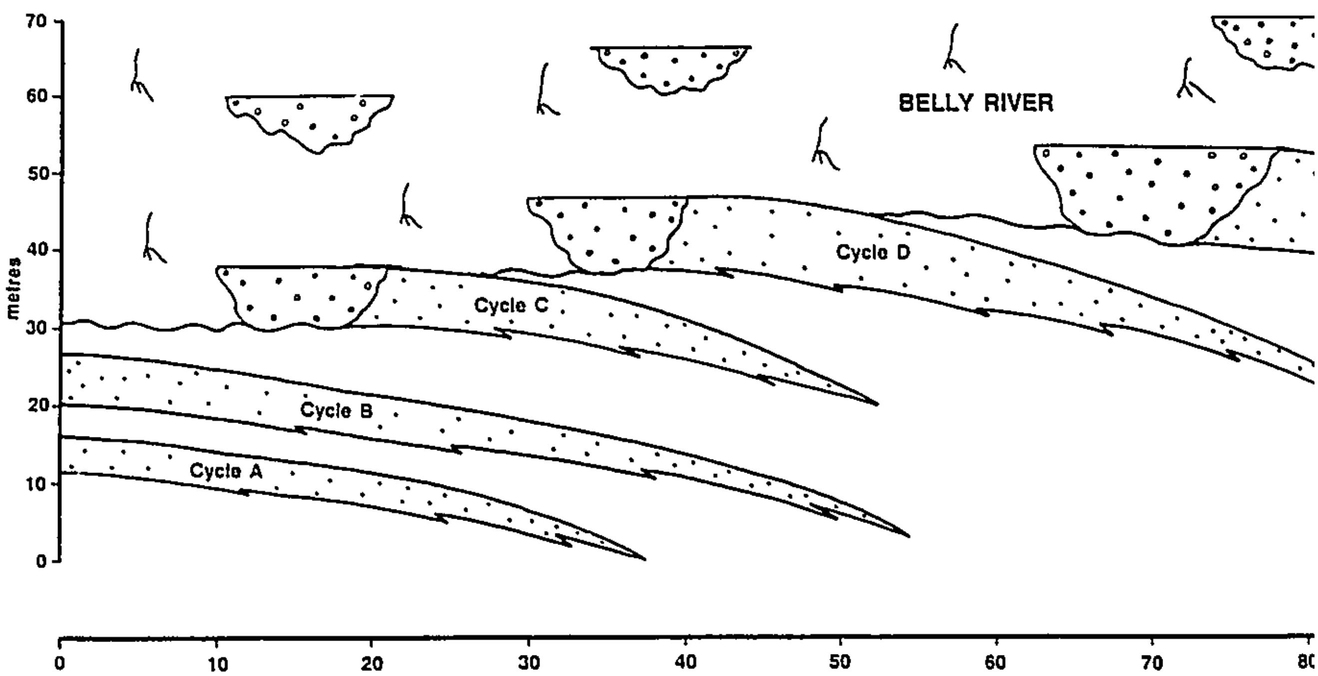
4.1: Introduction

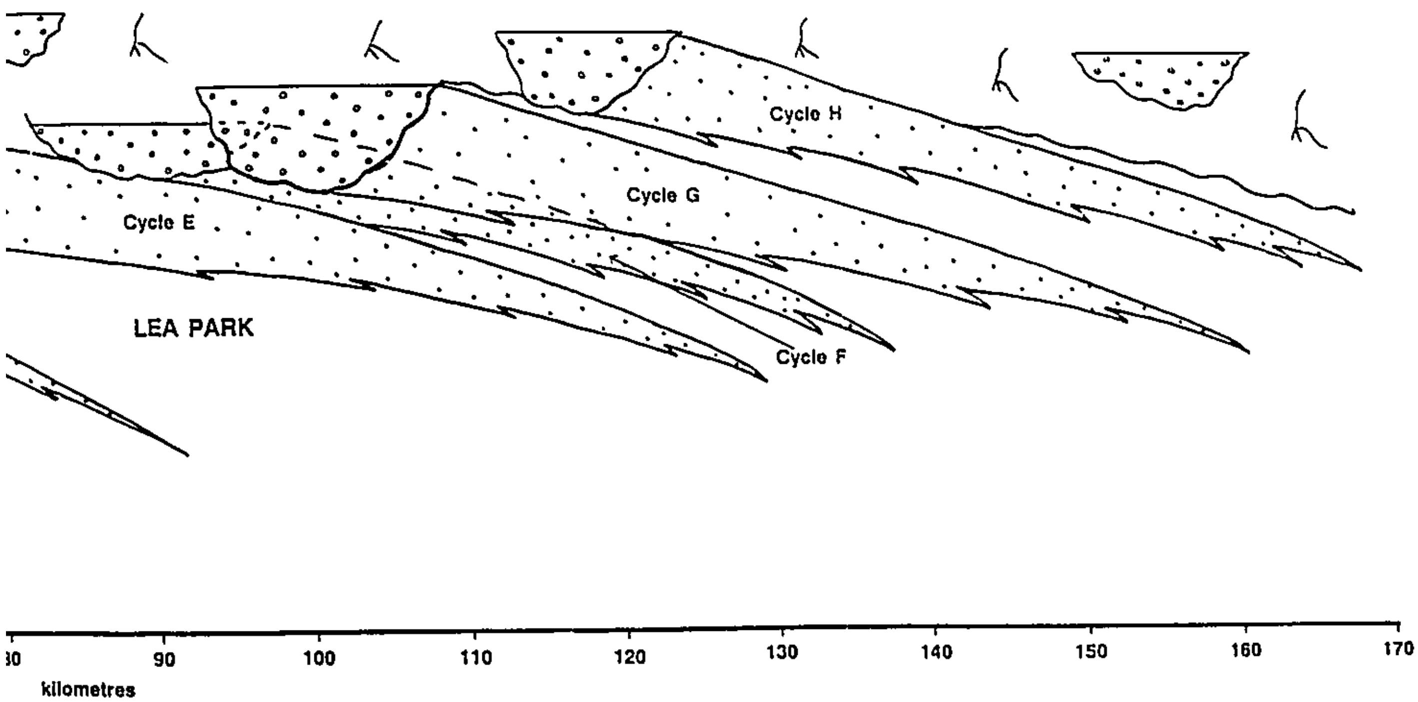
It has been known for some time, at least since the study of Mclean (1971), that the transition between the Lea Park and Belly River Formations in western Canada is diachronous, with tongues of Belly River sediment interfingering with the Lea Park Formation and becoming younger in an eastward direction. However, there have been no published studies of detailed stratigraphic subdivisions of the basal sediments of the Belly River Formation and the Lea Park Formation.

A detailed stratigraphic subdivision of the Lea Park - Belly River transition in the subsurface of central Alberta is proposed in this thesis. This study uses allostratigraphic principles and techniques to subdivide the Lea Park - Belly River transition into units defined by bounding discontinuities, but these units will not be presented as formal allostratigraphic members. The major allostratigraphic units interpreted in this study are instead referred to as "cycles". This term is not meant to imply any a priori definition that the "cycles" are repetitive in nature. It is simply a classification term for units. Both this and later chapters will, however, show that the stratigraphic arrangement of the cycles is repetitive within the study area. This chapter will show that, within the

Figure 4.1: Schematic dip-oriented cross section of the Lea Park - Belly River transition in the study area. The total thickness was estimated as the average of the thicknesses from the regional cross sections. Cycles F and G are located approximately along strike from each other, and thus occupy much of the same space in a dip section. Where Cycles F and G overlap, Cycle F is shown with a dashed line.

SCHEMATIC DIP-ORIENTED CROSS SECTION OF THE LEA PARK - BELLY RIVER TRANSITION





Cycle H
Cycle G
Cycle E
Cycle F
LEA PARK
30
90
100
110
120
130
140
150
160
170
kilometres

study area, the Lea Park - Belly River transition contains eight cycles (Fig. 4.1). Each of these cycles consists of the sediments of a single regressive phase of the Lea Park - Belly River transition. Each cycle is bounded at its top and bottom by regionally important bounding discontinuities, and is separated from other cycles by marine shelf sediments which were deposited between phases of regression. These sediments will be referred to throughout the thesis as "transgressive sediments" or "transgressive units". However, it is difficult, if not impossible to determine whether these marine shelf sediments were deposited during transgression of the underlying cycle, and/or during the initial stages of progradation of the overlying cycle. Because of this, the base of each cycle is placed at the base of the shoreface sandstone sediments of the cycle. This may seem to be a illogical place within the succession to place the base of the cycle, as some of the underlying sediments may be related to the progradation of the overlying shoreface. However, this contact at the base of the shoreface sediments in each cycle will later be shown to mark an important basinward shift in depositional environment. The internal complexities and details of each cycle vary, and will be discussed in chapters 5-10. Due to the fact that the bounding discontinuities can not be traced with any real confidence into non-marine sediments, this allostratigraphic subdivision can only be applied to the marine or transitional marine portions of the Lea Park - Belly River transition. Large portions of this study also focus on non-marine sediments of the Belly River Formation. The limitations of allostratigraphy are such that these sediments can not be incorporated into the stratigraphy proposed in this thesis, but the nature of their sedimentology and their possible relationships to the cycles will be discussed in a later chapter.

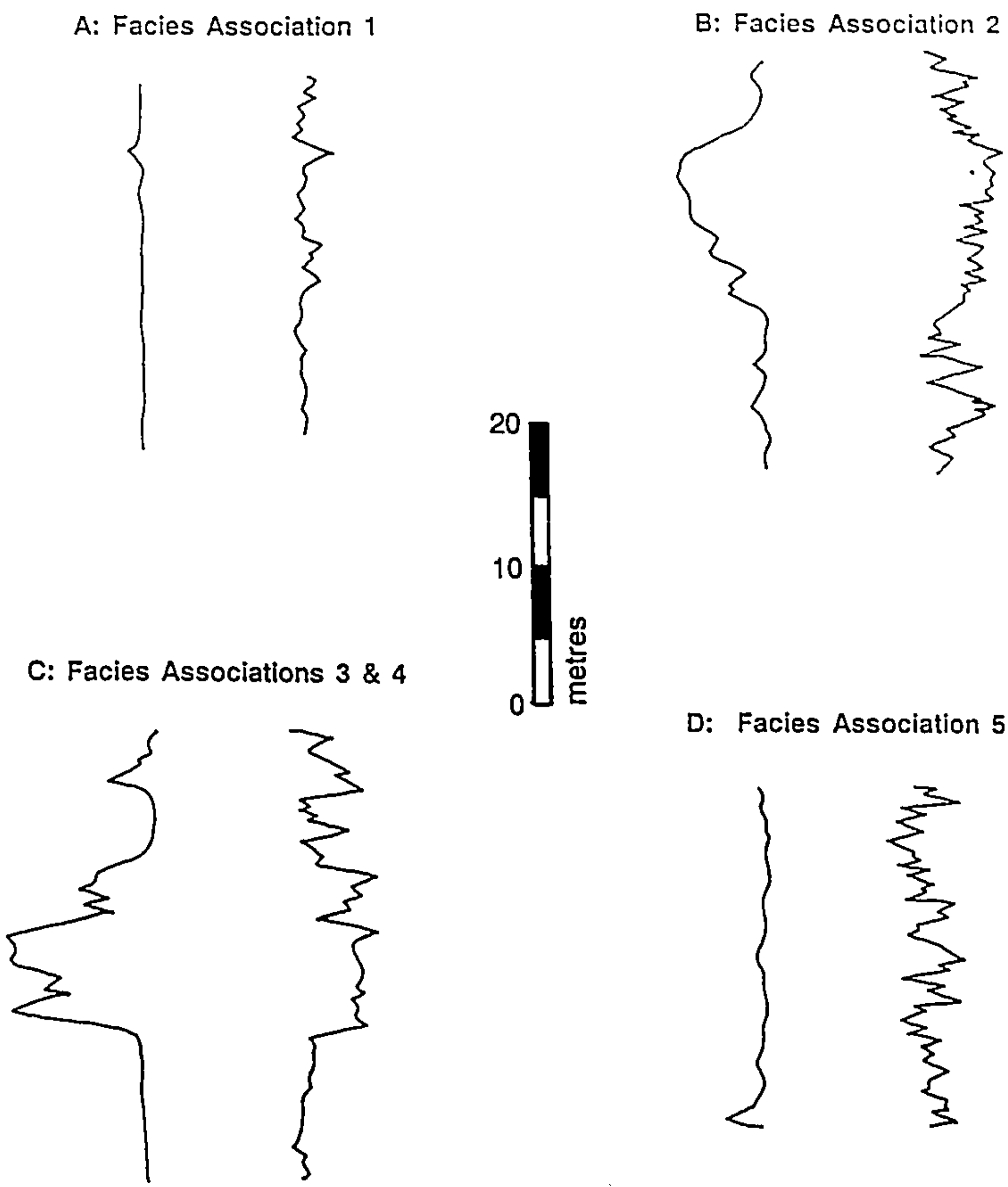
4.2: Interpretation of Facies Associations From Well Log Signatures

The study area of central Alberta is one of the most densely drilled portions of western Canada, and contains several hundred wells with cored intervals of the Lea Park - Belly River transition. Although the core control provides a large data base of information, wells containing core are only available for a relatively small percentage of wells in the study area. Correlations must in many cases be made using well log signatures alone. The success or failure of an allostratigraphic subdivision in the subsurface hinges upon how well facies associations and depositional environments can be interpreted from these well log signatures. The number of cores present within the study area makes this task somewhat easier. Where core exists, the facies associations present can be matched against their respective well log signatures. This allows for "typical" well log signatures for each facies association to be qualitatively determined. These "typical" well log signatures allow for interpretation and correlation in areas without core control. If a full suite of modern well logs were available for this study, these interpretations and correlations would have been based upon the signature of the gamma ray log, which measures the natural radioactivity of the rock, and is thus considered to be the best log for determining lithology. Unfortunately, gamma ray logs were not available for most of the well locations used in this study, as these wells were drilled several decades ago, before the gamma ray tool was developed. In most cases, interpretations and correlations in this study were based on the signature of the dual induction (resistivity) log. However, in some cases, both the resistivity and SP logs were used for interpretation.

Figure 4.2 shows the typical well log signatures for the different facies associations. Facies Association 1 consists of fine-grained marine mudstones and siltstones which typically show a flat, relatively featureless resistivity profile of low amplitude due to the low amount of pore fluids in the fine-grained shelf sediments (Fig. 4.2a) . Facies Association 2 consists of shoreface sandstones, and is generally typified by a "funnel-shaped" negative SP (leftward deflection) and positive resistivity (rightward deflection) logs, indicating higher permeability in the upper portions of the succession (Fig. 4.2b). Facies Associations 3 and 4 are channelized sandstones and are typically represented by sharp-based, internally variable deflections of both the SP and resistivity logs (Fig. 4.2c). Facies Association 5 is typified by an irregular, "sawtooth" resistivity signature (Fig. 4.2d) indicative of many minor fluctuations in pore fluid content due to grain-size fluctuations within the non-marine floodplain environment, as well as numerous heavily negative resistivity deflections indicative of coal beds with very low pore fluid content.

Figure 4.2: Well log signatures of the facies associations: (A) Facies Association 1 (shelf sediments); (B) Facies Association 2 (shoreface sediments); (C) Facies Associations 3 & 4 (fluvial channels); (D) Facies Association 5 (non-marine floodplain).

TYPICAL WELL LOG SIGNATURES



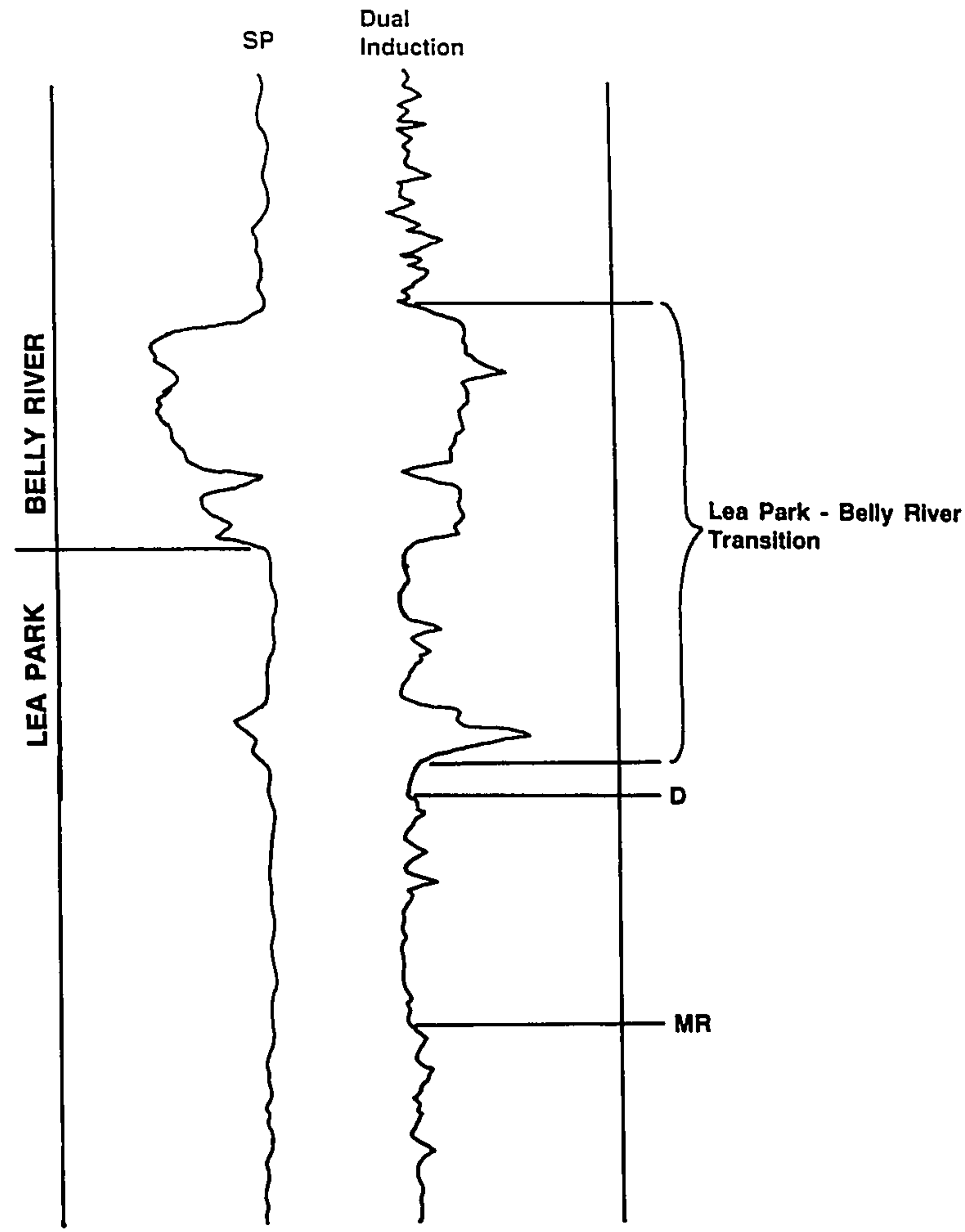
It should be pointed out that, in the absence of core, it is not possible to be certain about facies association interpretations and the resultant correlations. For example, the well log signatures for shoreface and channel sandstones can sometimes be very similar. In a coastal environment where channels and shoreline sediments are laterally and contemporaneously connected, it is possible to interpret a channel signature as representing shoreline sediments, or vice versa. At times, techniques other than profile shape of the well log signature can be used to make interpretations and correlations. Shoreline sandstones tend to be far more laterally continuous than channel sandstones, and this property was often used in deciding the facies association of well log signatures used in the correlations. Using a combination of core data, shape of well log profiles, and sediment unit continuity, the correlations used in this study can be developed with a high degree of confidence.

4.3: Well Log Datums

When correlating stratigraphic units, it is necessary to have stratigraphic datums so that individual stratigraphic sections and/or well logs can be placed in the proper topographic position and correlated to each other with confidence. Ideally, these datum horizons should be present throughout the study area, easily picked on well logs, and lie either above or below the stratigraphic unit of interest. It is also desirable that the datum represents a surface that is relatively flat with respect to the basin, although in reality it is unlikely that any surface is truly flat. Because the Lea Park - Belly river transition is overlain throughout the study area by a thick succession of non-marine sediments, no regionally persistent surfaces are available above the transition to use as datums for correlation. Locally extensive coal beds are present within the fluvial portion of the Belly River Formation, but they are not regionally extensive enough for use as datums. As a result, the well log datums used in this study are markers located below the Lea Park - Belly River Formation, within the marine mudstones of the Lea Park Formation. Figure 4.3 shows a typical well log signature of used in this study. The primary datum used in this study is the top of a distinctive positive deflection on the dual induction log noted as MR. This is the "Milk River shoulder", and is thought to be the basinal representation of the top of the Milk River Formation, a clastic wedge which underlies the Belly River Formation. This marker can be picked throughout the study area, and is therefore a useful datum. Unfortunately, other datums were required because many wells used in this study do not penetrate this log marker. As a result, other markers are used locally for datums if the Milk River shoulder is not penetrated. One of these is labelled "D" in Figure 4.3. These markers are the transgressive surfaces marking the top of the distal sediments of cycles within the Lea Park - Belly River transition. Locally these surfaces are roughly sub-parallel to the Milk River shoulder. They are not ideal datums because they lie within the unit of interest, and by using them to hang correlations, the resultant stratigraphic patterns are flattened out.

Figure 4.3: Typical SP and Resistivity well log signature for the Lea Park - Belly River transition, showing markers used as datums for correlation. "MR" is the Milk River shoulder, which is the regional datum used whenever available. Marker "D" represents the top of a distal lower cycle within the Lea Park - Belly River transition. These markers are used as datums if the MR marker is not available.

11-23-45-26W4


SP
Dual Induction
BELLY RIVER
LEA PARK
Lea Park - Belly River Transition
D
MR

However, they are the only surfaces available for datums in many cases, and they allow for confident correlation of cycles. In many of the log cross sections used in this thesis, there will be log signatures present which do not appear to penetrate deeply enough to reach any datum markers below the Lea Park - Belly River transition. In a few cases, this is true, and these logs are then correlated either using local internal markers such as coals or black, lagoonal shales which give distinctive log traces. In most cases, the lower datum marker is simply not shown in the expanded scale recording of the well log (which often does not contain a record of the complete sedimentary section, but rather only strata of interest to the oil company which drilled the well). Where this occurs, the stratigraphic position of the well log was determined using the condensed scale recording, which contains the entire section, and the expanded scale cross section was constructed using this correlation.

4.4: Variability of the Lea Park and Belly River Transition

The previous chapter detailed the internal characteristics of the vertical facies associations within the Lea Park - Belly River transition. The contact between the Lea Park and the Belly River Formations is a lithostratigraphic contact, and is usually placed at the base of the first "major" sandstone succession or where the sediments change from marine to non-marine in nature. Before lateral correlation of the facies successions could be accomplished, it was important to realize that the exact nature of the Lea Park -

Belly River transition varies considerably from place to place within the study area, and that superposition of facies associations can also vary considerably. Figure 4.4 shows the well log signatures of the variety of facies associations which can exist at the contact between the Lea Park and the Belly River.

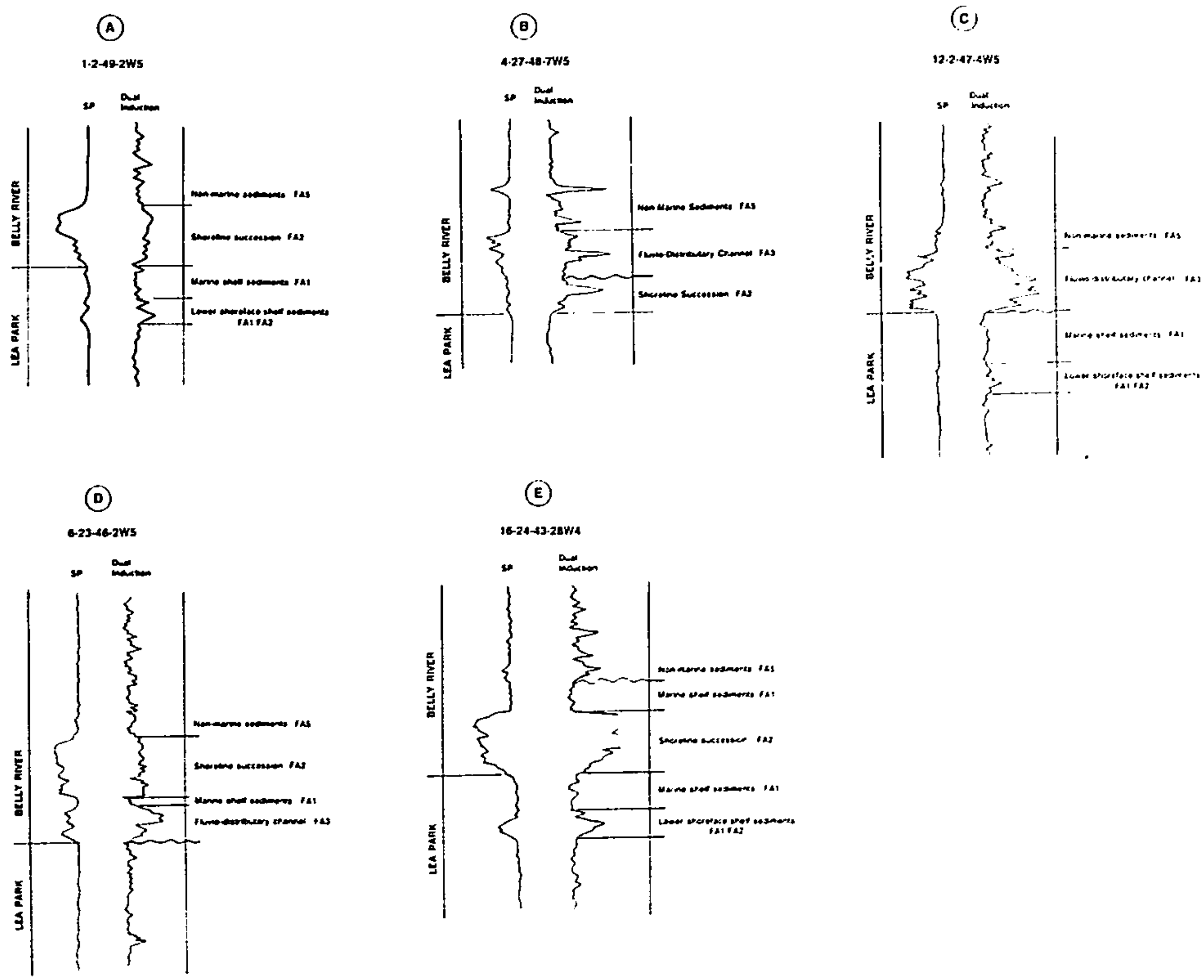
Figure 4.4a is probably the most common form of the transition observed in this study. The contact between the Lea Park and the Belly River is located at the base of a succession of shoreline sandstones, which represents the first "major" sandstone of the Belly River Formation. Overlying this sandstone is an uninterrupted succession of non-marine sediments of the Belly River Formation. Underlying the shoreline succession are marine shelf sediments which are in turn underlain by lower shoreface/shelf sandstones, which are likely the distal sands deposited by an earlier shoreline.

In figure 4.4b, the transition is similar to Figure 4.4a, except that the basal sandstone of the Belly River Formation is a composite sandstone. The uppermost sediments of the shoreline succession have been eroded and replaced by channelized sandstones.

In figure 4.4c, no shoreline sandstones are present, and the base of the Belly River Formation is at the base of the channel sandstone, which sits on marine shelf sediments of the Lea Park Formation.

In figure 4.4d, the contact between the Lea Park and the Belly River is also placed at the base of a channelized sandstone. The underlying Lea Park sediments show little evidence of distal shoreline deposits, which were present

Figure 4.4: Variability of the nature of the Lea Park - Belly River transition in central Alberta. (A) Distal shoreface succession overlain by subsequent fully developed marine shoreface succession. (B) Marine shoreface succession overlain by fluvial channel. (C) Fluvial channel. (D) Fluvial channel overlain by marine shoreface succession. (E) Marine shoreface succession overlain by shelf sediments, which is in turn overlain by non-marine sediments.


A
1-2-49-2W5
SP
Dual Induction
BELLY RIVER
LEA PARK
Non-marine sediments FA5
Shoreline succession FA2
Marine shelf sediments FA1
Lower shoreface shelf sediments FA1 FA2
B
4-27-48-7W5
SP
Dual Induction
BELLY RIVER
LEA PARK
Non-Marine Sediments FA5
Fluvio-Distributary Channel FA3
Shoreline Succession FA2
C
12-2-47-4W5
SP
Dual Induction
BELLY RIVER
LEA PARK
Non-marine sediments FA5
Fluvio-distributary Channel FA3
Marine shelf sediments FA1
Lower shoreface shelf sediments FA1 FA2
D
6-23-46-2W5
SP
Dual Induction
BELLY RIVER
LEA PARK
Non-marine sediments FA5
Shoreline succession FA2
Marine shelf sediments FA1
Fluvio-distributary channel FA3
E
16-24-43-28W4
SP
Dual Induction
BELLY RIVER
LEA PARK
Non-marine sediments FA5
Marine shelf sediments FA1
Shoreline succession FA2
Marine shelf sediments FA1
Lower shoreface shelf sediments FA1 FA2

in the previous three examples. The channel sandstone is overlain by marine shelf sediments, which are in turn overlain by shoreline sandstones, in a similar manner to figure 4.4a.

Figure 4.4e is similar to figure 4.4a. The only difference is that the sediments overlying the shoreline succession at the base of the Belly River Formation are marine shelf sediments, rather than non-marine floodplain sediments, indicating that two complete regressive-transgressive successions are preserved. These marine sediments are directly overlain by non-marine sediments, either channelized as in figure 4.4b, or floodplain deposits, as in figure 4.4e.

These figures show that the internal variability of the Lea Park - Belly River transition reflects a history of rapidly changing depositional environments, in which the position of shorelines and relative sea level has fluctuated in both a basinward and a landward direction. The stratigraphic scheme used to subdivide and describe these sediments should act as a basis for understanding the geologic events which occurred during the Lea Park - Belly River transition. Correlating these sediments using a lithostratigraphic scheme would not enable any significant understanding of how these various facies associations are spatially related and how they record the establishment of coastal plain depositional environments. The use of allostratigraphic principles is considered the best approach to this problem because it involves the recognition of genetically related packages of sediment.

4.5: Recognition of Bounding Discontinuities

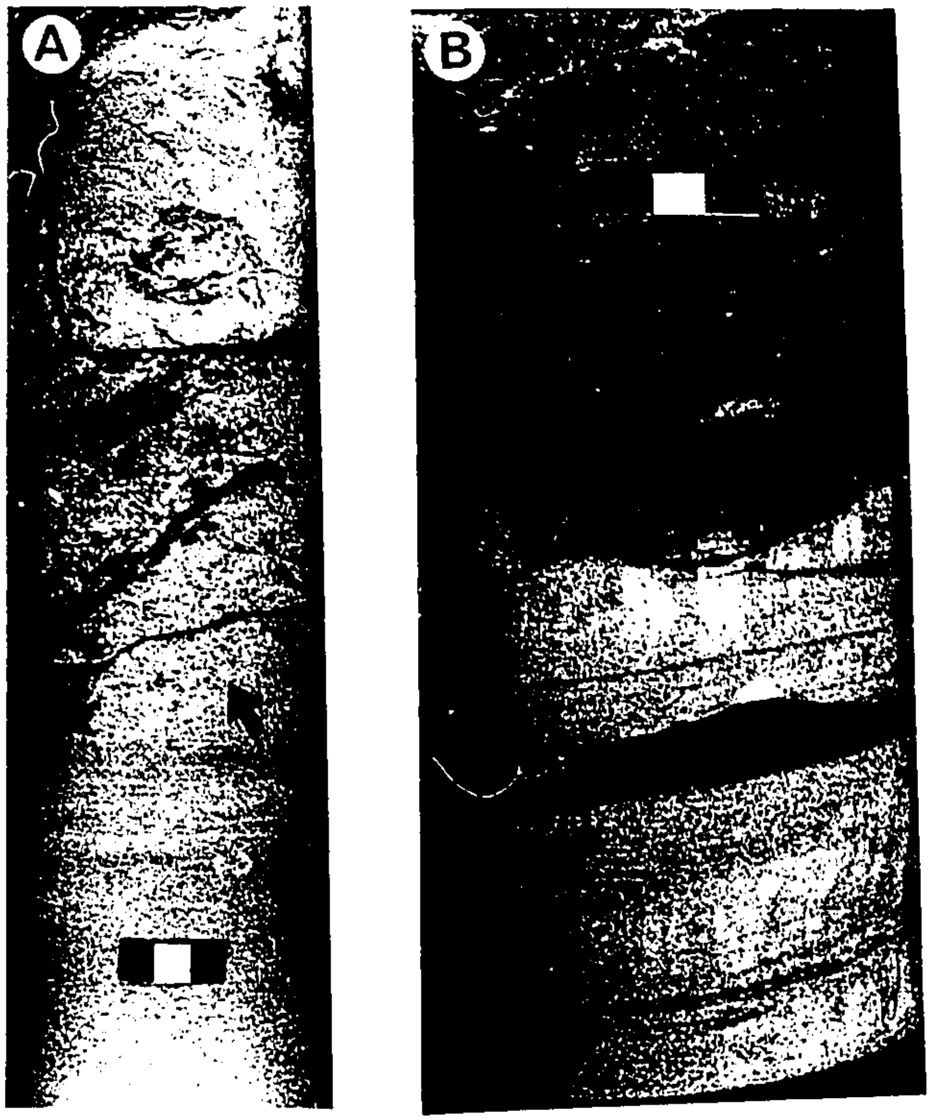
In order to subdivide a succession of rocks into an allostratigraphic or sequence stratigraphic scheme, it is important to be able to characterize the nature of the bounding discontinuities which will define the stratigraphic units. These are surfaces of erosion or non-deposition, which are interpreted to have been created by relative fluctuations in sea level. In many cases these surfaces also define the tops and/or bases of facies associations discussed in chapter 3. There are two main types of bounding discontinuity surfaces which can be recognized within the sediments of the Lea Park -Belly River transition. These are: (1) Regressive surfaces of erosion (both non-marine due to channel incision, and marine erosion surfaces associated with wave and/or current activity) and (2) Transgressive flooding surfaces.

Examples of these surfaces are shown in figures 4.6-4.8. In all cases, the surfaces are characterized by abrupt vertical changes in facies and facies associations. In some cases, the surfaces are sharp and easily defined. In other cases, most notably with the transgressive surfaces, the contact or discontinuity may be diffuse and spread over several centimetres or tens of centimetres.

Figure 4.5: Regressive surfaces of erosion due to non-marine erosion.

(A) Regressive surface of erosion (marked by arrow) due to fluvial channel incision.

(B) Regressive surface of erosion due to subaerial exposure of marine sediments (sandstones), which are then covered by non-marine floodplain mudstones.


A
B

4.5.1: Regressive Surfaces of Erosion Due to Non-Marine Erosion

Surfaces of erosion due to channel incision are easily recognized by their location at the base of channel sandstone successions (Fig. 4.5a). Another type of surface of non-marine or subaerial erosion exists where non-marine floodplain sediments lie erosively over marine shelf or lower shoreface sediments (Fig 4.5b). These surfaces are also easy to recognize beacuse of the sharp distinction betwee marine and non-marine facies.

It should be noted that the regional importance of any of these surfaces can not be determined from analysis of individual vertical facies successions. In order to understand the regional significance of a given surface, it must be mapped out in three-dimensional detail.

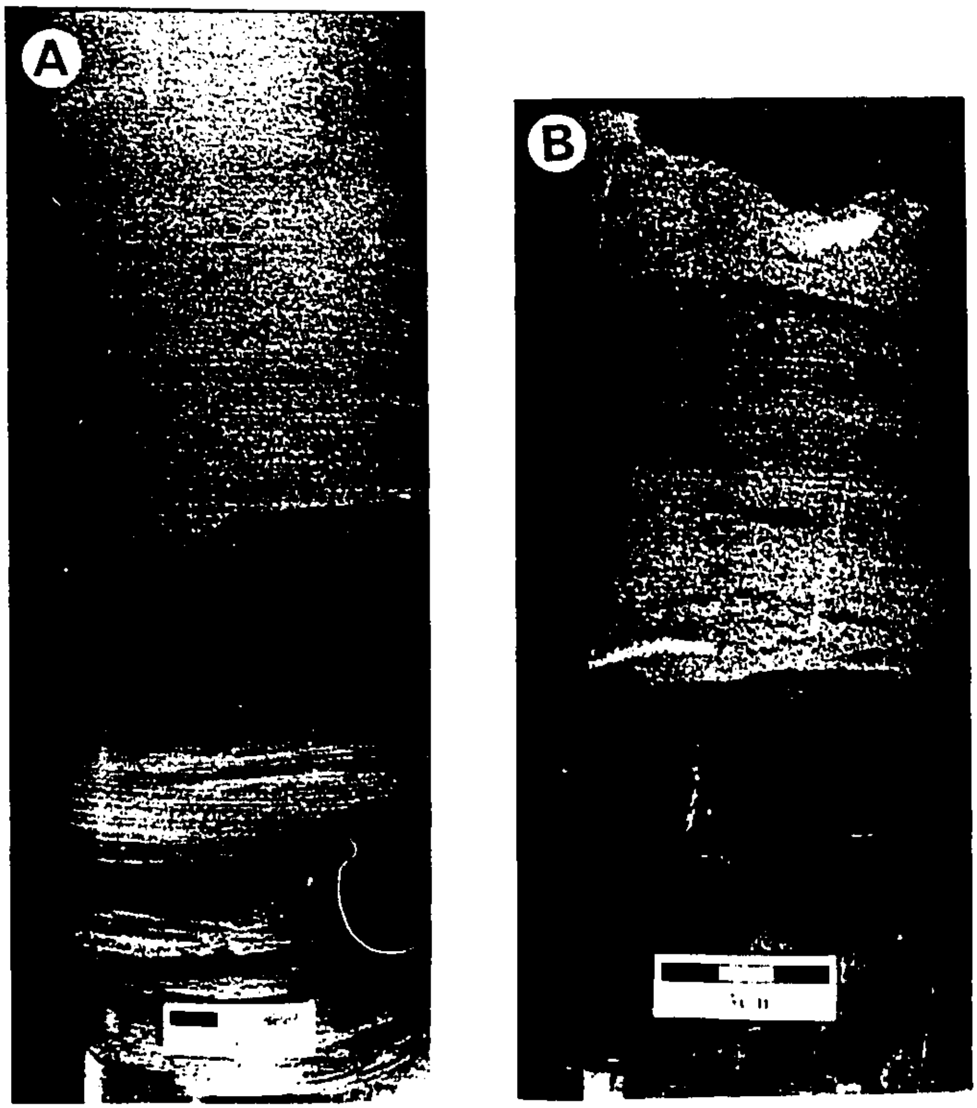
4.5.2: Regressive Surfaces of Erosion Due to Marine Erosion

Surfaces of marine erosion or their correlative conformities are surfaces which are either demonstrably erosional, or in the case of a correlative conformity, in which there is an abrupt change in facies, but no evidence of erosion. The facies associations above and below these surfaces are marine in nature, with the sediments above the surface being indicative of an abrupt decrease in water depth of the depositional environment (eg. shoreface sands overlying shelf mudstones). Two examples of such a surface are shown in figure 4.6.

Figure 4.6: Regressive surfaces of erosion due to marine erosion.

(A) Regressive surface of erosion due to marine erosion at the base of a shoreface succession. Note the angular contact.

(B) Regressive surface of erosion due to marine erosion at the base of a shoreface succession. Not the angular contact and mudstone rip-up clasts.


A
B

4.5.3: Transgressive Bounding Discontinuties

Transgressive surfaces are characterized by an upwards transition from coastal or non-marine sediments such as channel sandstones or shoreface sandstones into marine or brackish mudstones and siltstones (Fig 4.7). The transition is usually characterized by the sediments of Facies Association 6. These surfaces are indicative of deepening related to local or regional marine transgression (flooding surfaces). These surfaces are often the easiest to identify in the subsurface, because they mark sharp transitions from high to low resistivity. Another type of transgressive bounding discontinuity is the maximum flooding surface, which is the surface within the transgressive sediments denoting the time of maximum transgression. This surface is often represented in sedimentary sections by a "condensed section" due to low rates of deposition. Unfortunately, no maximum flooding surfaces could be identified within the sediments of the Lea Park - Belly River transition.

4.6: Regional Stratigraphy of the Lea Park - Belly River Transition

Thirty-three regional dip and strike cross sections, using data from over 1200 well locations, were constructed to determine the stratigraphic relationships of the Lea Park - Belly River transition in the study area. The location of these sections is shown in figure 4.8. The cycles correlated in any given section were traced back, using adjacent sections, to the starting location of the first section to insure that correlations matched throughout the study

Figure 4.7: Transgressive flooding surfaces

(A) Transgressive flooding surface (marked by arrow) showing bioturbated mudstones overlying coal.

(B) Transgressive flooding surface (marked by arrow) showing bioturbated muddy sandstones overlying channel sandstones.

(C) Transgressive flooding surface (marked by arrow) showing lower shoreface/shelf sandstones and mudstones overlying upper shoreface sandstones.

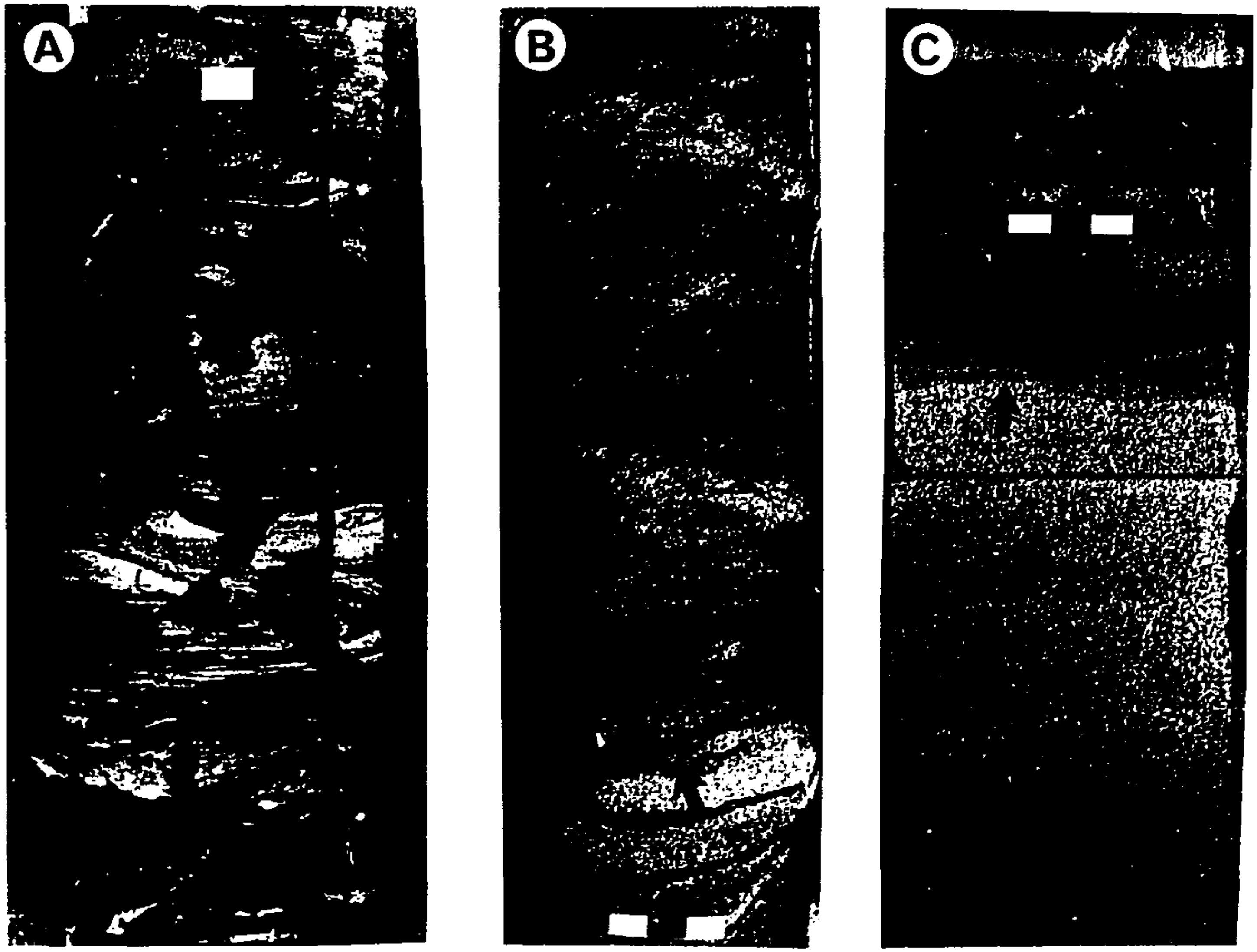
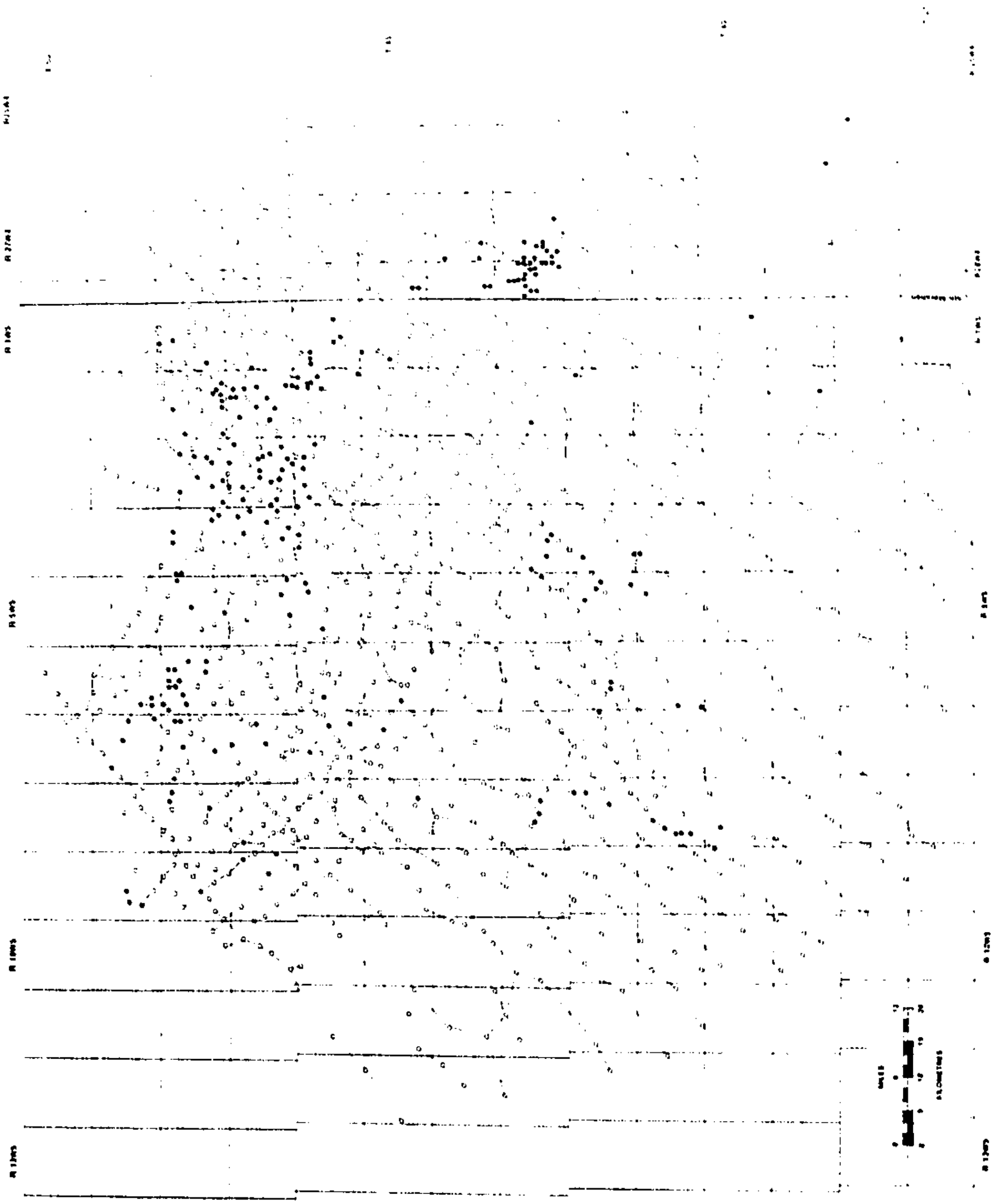

A
B
C

Figure 4.8: Grid of regional cross sections used to correlate and interpret the stratigraphy of the Lea Park - Belly River transition in this study. Correlations for any one sections are matched for correctness against all other sections.



area. The following section contains eight regional well log cross sections which together show the nature of the Lea Park - Belly River transition in the subsurface of central Alberta. The location of these sections is shown in figure 4.9. Each of the sections is a condensed version of one of the thirty-three working cross sections shown in figure 4.8. Cross sections A-A' through E-E' are dip-oriented sections and F-F', G-G', and H-H' are strike-oriented cross sections. Multiple cross sections are required to document the stratigraphy, as individual cycles are not present throughout the study area. It should be noted that none of these sections can be regarded as a true dip or strike section for all the cycles involved, as the precise direction of strike and dip changes somewhat between individual cycles. In general or regional terms, however, the strike of the depositional systems was determined to trend northwest-southeast, with rivers and shorelines prograding in a northeasterly direction.

Cross Section A-A'

Cross section A-A' (Fig. 4.10) is the most western of the dip cross sections. This section shows the basic stratigraphic pattern of relationships in the Lea Park - Belly River transition that is present throughout the study area. The transition itself is very diachronous, becoming younger to the northeast. Five regressive cycles can be identified and correlated within the section. These cycles are labelled A-E, with A being the youngest. Each cycle can be seen to downlap and pinch out towards the datum markers. Because of the

Figure 4.9: Location of regional cross sections discussed in this chapter.

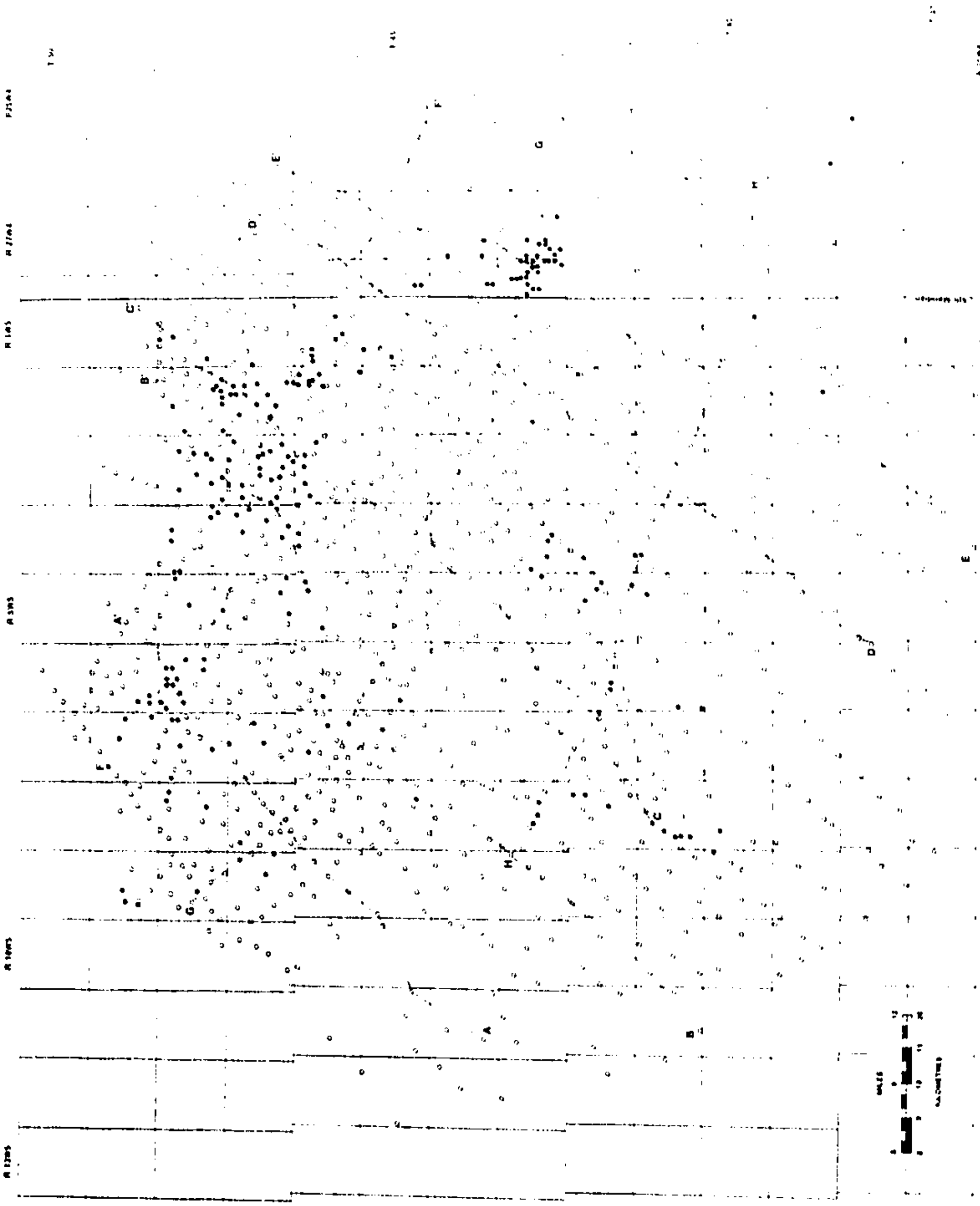


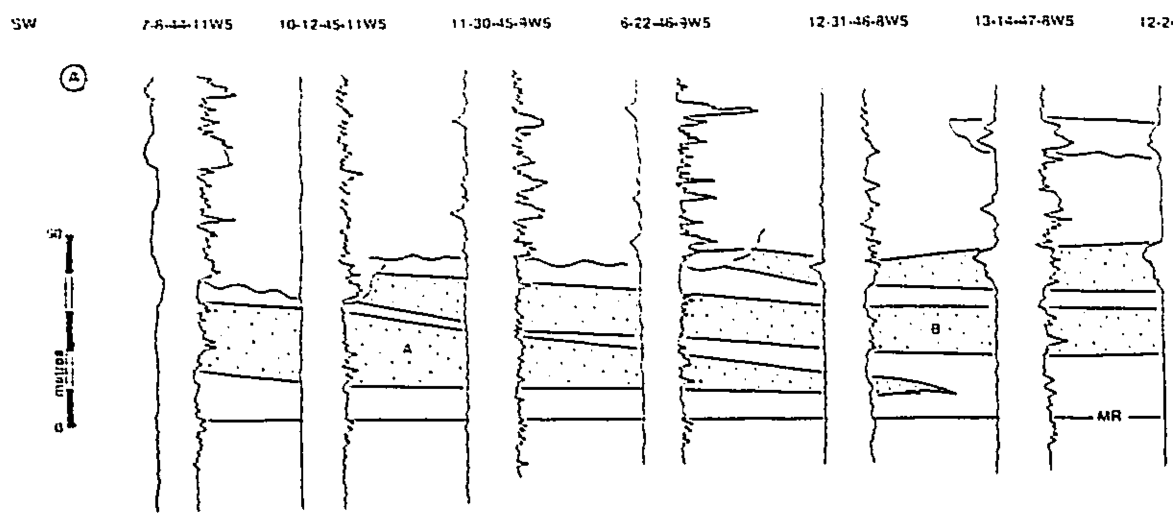
Figure 4.10: Cross section A-A'. This is a dip oriented section located near the western end of the study area. All locations show SP and Dual Induction log signatures. The section is approximately 70 km in length. Stippled pattern denotes marine sandstone of a given cycle. Undulating lines represent erosional contacts where non-marine sediments erosively sit on marine sediments. Vertical black strips indicate location and interval of examined core. These patterns and/or designations remain the same for all further sections in this study.

use of an interior datum in the northeastern portion of the section, evidence of the downlapping pattern is flattened out. The intercyclic marine sediments rise up in a landward direction. Both the regressive cycles and the transgressive sediments between each cycle abut in a landward direction against non-marine sediments, which may or may not be contemporaneous with deposition of the laterally equivalent marine cycle.

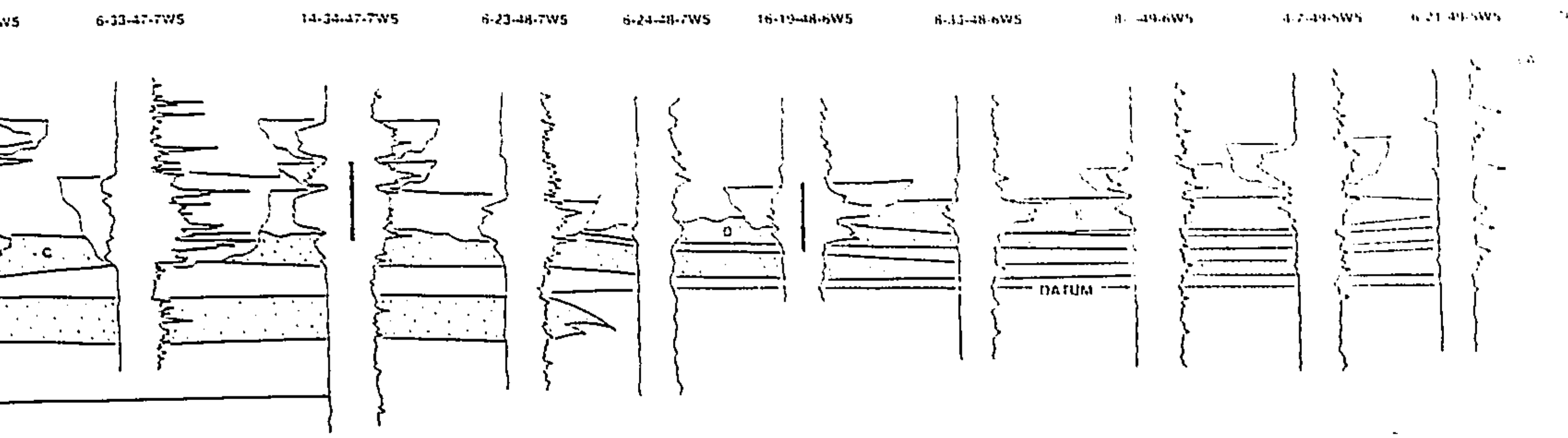
Cycle A is the oldest marine cycle identified within the study area. It is present only in section A-A'. Only the distal portions of the cycle are present within the study area. These exist as a 10-15 m-thick succession of muddy shelf sediments with minor sand content (according to log signature) at the southwestern edge of the study area which thins and pinches out approximately 30 km to the northeast, at 12-31-46-8W5.

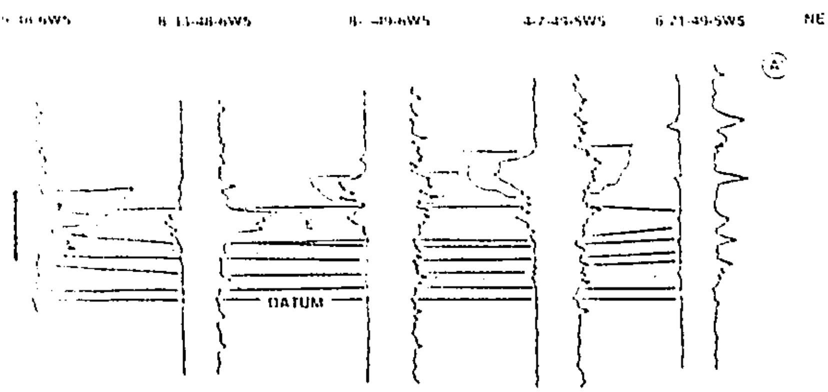
The transgressive sediments overlying Cycle A are approximately 6-8 metres thick at the basinward limit of the progradation of the underlying regressive unit. It rises and thins to the southwest, but is still present at the southwestern edge of the section.

Cycle B is similar to Cycle A in that only the distal sediments of the cycle are found within the study area. In section A-A', the cycle is present as a 10-12 m-thick interval of muddy shelf sediments with minor sand content very similar to the sediments of Cycle A. These sediments thin slightly to the northeast over 20 km, before gradually pinching out into marine shales. At the landward end of the section, Cycle B appears to have been eroded by later incision of non-



SW
7-6-44-11W5
10-12-45-11W5
11-30-45-9W5
6-22-46-9W5
12-31-46-8W5
13-14-47-8W5
12-2
A
0
metres
A
B
MR





6-21-49-5W5
NE
A'
DATUM

marine sediments at 10-12-45-11W5. These non-marine sediments appear to lie erosively on the transgressive mudstones between Cycles A and B.

The transgressive sediments overlying Cycle B range from 2-10 m thick, attaining maximum thickness at 14-34-47-7W5. These transgressive sediments stratigraphically rise slightly to the southwest before being erosively removed by the same non-marine sediments which eroded Cycle B at 10-12-45-11W5.

Cycle C is first observed as a thin (8-10 m) coarsening-upward or cleaning-upward (becomes sandier) sandstone unit at 12-31-46-8W5. To the southwest it is laterally equivalent to non-marine sediments which lie sharply on the transgressive sediments overlying Cycle B. The regressive sandstone succession thickens to the northeast to 12-24-47-8W5, where it appears as a well developed 15 m-thick coarsening-upward shoreface succession. To the northeast of this point, Cycle C is incompletely preserved, with the upper portions being removed by subsequent non-marine erosion. By 6-24-48-7W5, the regressive succession has thinned to a 6-8 m-thick unit of mudstone and sandstone shelf sediments, which continue to thin to the northeast. At the northeast end of the section, Cycle C has thinned almost to depositional edge.

The transgressive unit overlying Cycle C consists of a thin, 4-6 m thick unit of shelf mudstones that rises slightly to the southwest before being erosively removed by fluvial incision at 6-23-48-7W5.

Cycle D is present in section A-A' only as an incompletely preserved succession of lower shoreface sandstones at 6-24-48-7W5 which thin to the northeast into a 2-4 m thick unit of shelf mudstones/sandstones at A'.

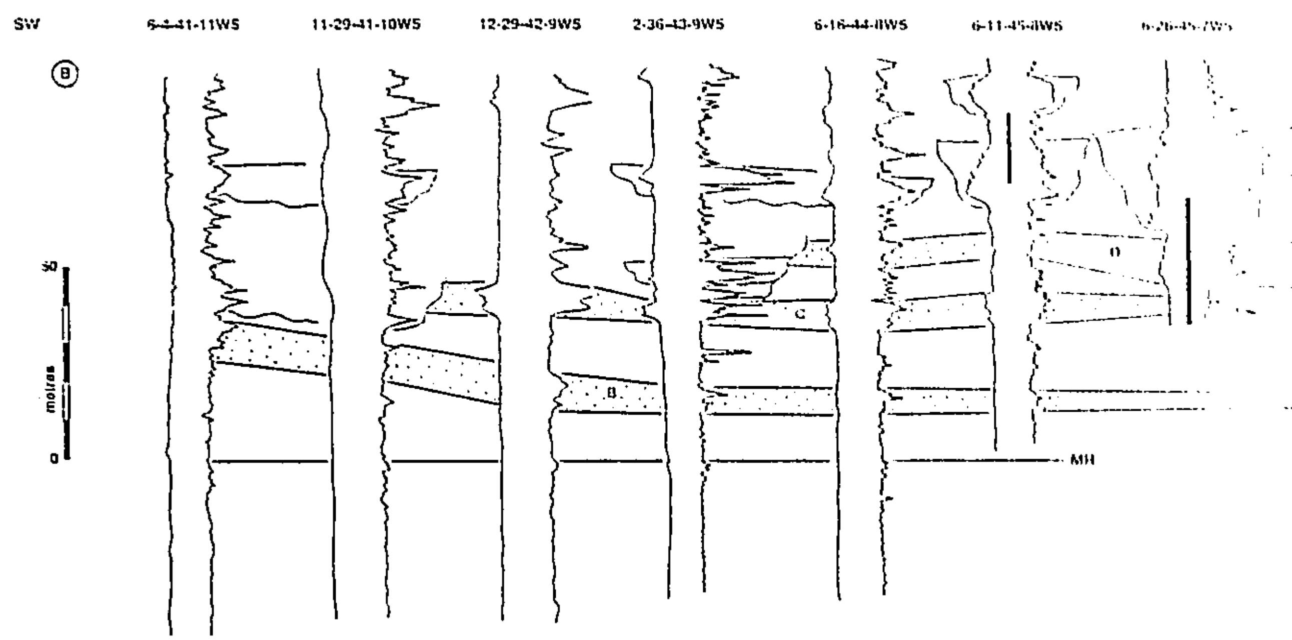
The transgressive sediments overlying Cycle D are present in this section only to the northeast of 8-33-48-6W5. To the southwest of this location, this unit has been removed by either fluvial incision or marine erosion due to the progradation of Cycle E sediments.

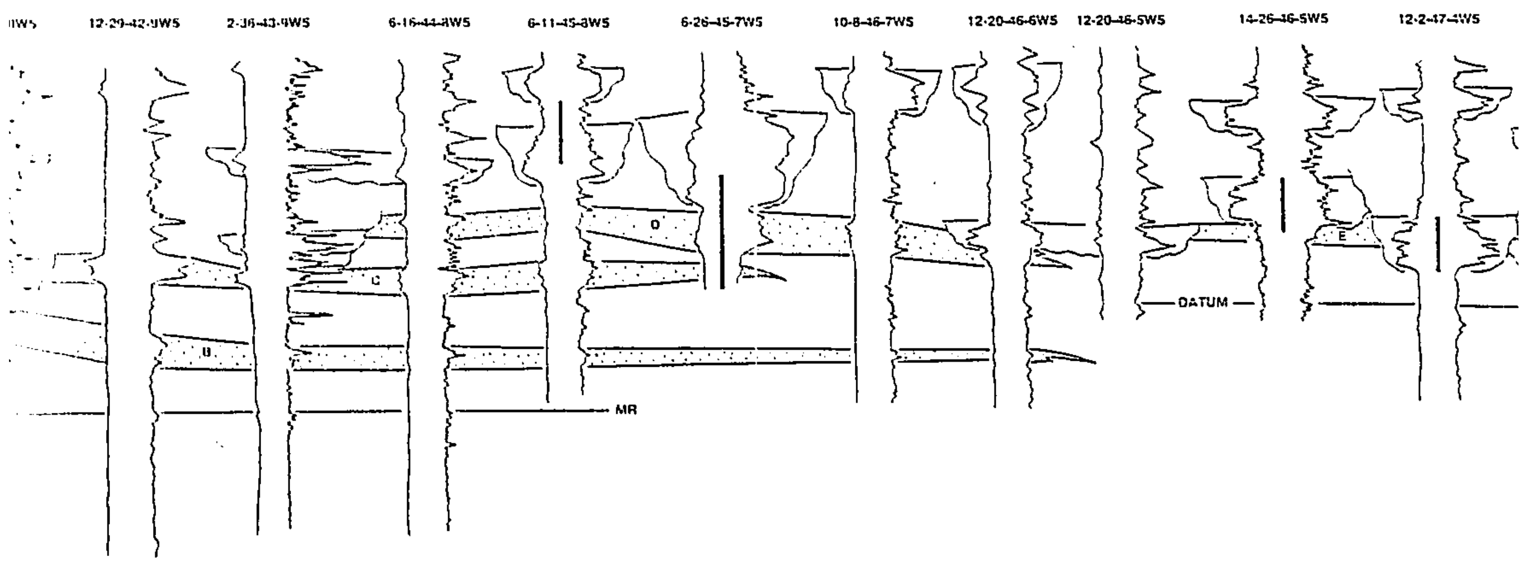
Cycle E is the youngest marine cycle preserved in section A-A'. It is first observed in 8-33-48-6W5 as a 10-12 m-thick shoreface succession. This thins to the northeast, and at A', exist only as a 4-5 m thick sand unit. The transgressive sediments overlying Cycle E is not present in this section.

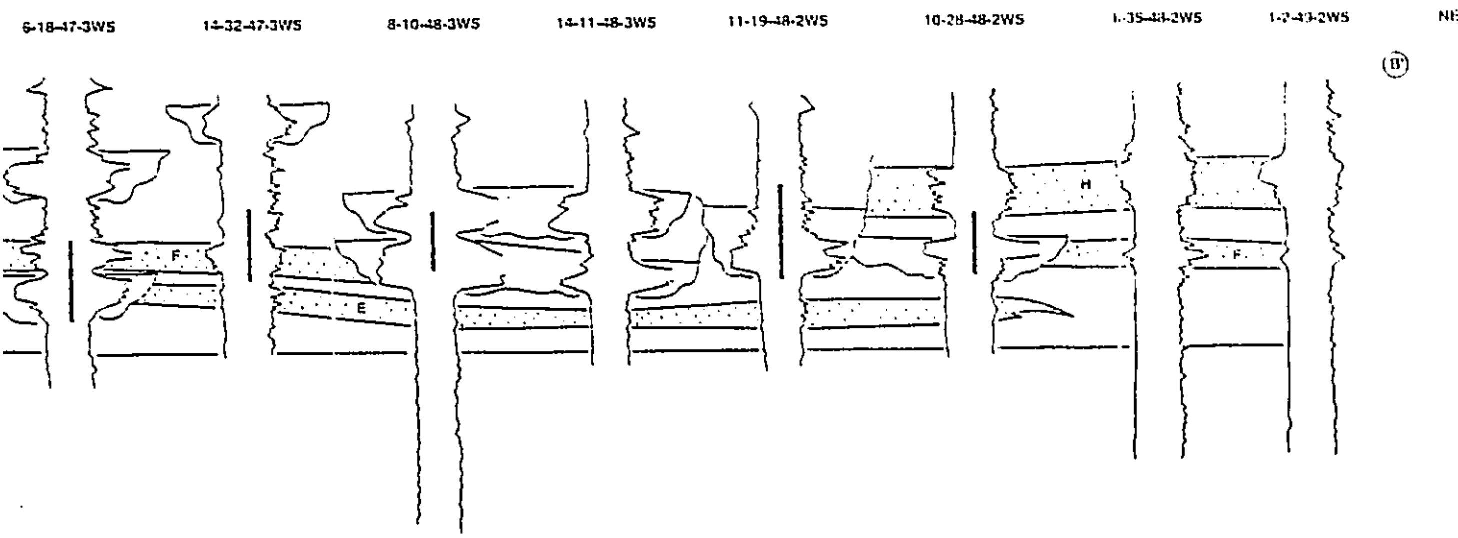
<u>Cross Section B-B'</u>

This dip section (Fig. 4.11) begins to the southwest of the Ferrier-Willesden Green field, and trends to the northeast through the western edge of the Keystone field, where much of the core data for this study is located (Fig. 4.9). The basic stratigraphic pattern observed in B-B' is the same as for A-A'. The transition consists of numerous downlapping regressive cycles separated by transgressive units which become younger to the northeast. Six cycles are present in section B-B'. In this section, large areas of the marine cycles have been removed by subsequent fluvial erosion, especially in the northeastern half of the section, in the Keystone field.

Figure 4.11: Cross section B-B'. This is a dip-oriented section trending through Ferrier and the eastern portion of Keystone. All locations show SP and Dual Induction log signature. The section is approximately 110-120 km in length.


SW
B
6-4-41-11W5
11-29-41-10W5
12-29-42-9W5
2-36-43-9W5
6-16-44-8W5
6-11-45-8W5
50
metres
0
MH


12-29-42-9W5
2-36-43-9W5
6-16-44-8W5
6-11-45-8W5
6-26-45-7W5
10-8-46-7W5
12-20-46-6W5
12-20-46-5W5
14-26-46-5W5
12-2-47-4W5
B
C
D
E
MR
DATUM



6-18-47-3WS
14-32-47-3WS
8-10-48-3WS
14-11-48-3WS
11-19-48-2WS
10-28-48-2WS
NE
B'
F
E
H
F

Cycle B is the oldest cycle present in this section. At the southwestern end of the section, the cycle is a 10 m-thick cleaning upwards shelf - lower shoreface sandstone succession. This thins basinward and reaches its depositional edge approximately 50 km to the northeast at 12-20-46-6W5.

The transgressive unit overlying Cycle B follows the pattern of those in section A-A'. It rises and thins to the southwest, and at the southwestern end of the section is just a thin (2-3 m) unit of shelf mudstones.

In section B-B', Cycle C is present as a laterally inextensive succession of shoreface and shelf sandstones. At 12-29-42-9W5, the regressive succession is a 5-12 m-thick cleaning-upwards shoreface sandstone, which thins slightly basinward, and then abruptly pinches out to the northeast of 6-26-45-7W5. To the southwest of 12-29-42-9W5, Cycle C is laterally equivalent to non-marine sediments which lie sharply on the transgressive sediments of Cycle B.

The transgressive sediments overlying Cycle C are a 10-12 m-thick unit of shelf mudstones, which are eroded southwest of 2-36-43-9W5 and replaced by non-marine sediments.

Cycle D begins as a thin (4-6 m) cleaning-upwards shoreface/shelf sandstone succession at 6-16-44-8W5. The cycle quickly thickens to the northeast into a 12-15 m-thick unit of shoreface/shelf sediments at 10-8-46-7W5. Basinward of this, it pinches out rapidly at 12-20-46-5W5. Transgressive sediments overlying Cycle D are nor preserved southwest of 12-20-46-5W5.

Cycle E is present in this section only as distal shoreface/shelf sediments 2-3 m thick. Much of the sediments of this cycle have been removed by fluvial erosion, most notably at 12-2-47-4W5, 6-18-47-3W5, and southwest of 14-26-46-5W5. Transgressive sediments overlying Cycle E are not present southwest of 12-2-47-4W5, due to post-depositional erosion by fluvial incision.

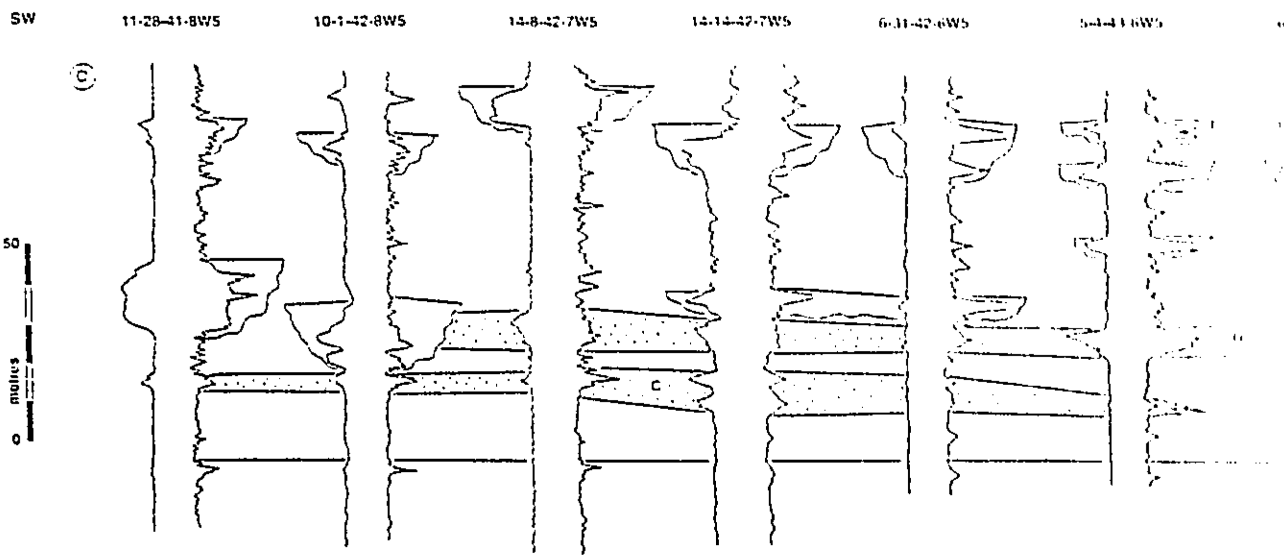
Cycle F is first observed at 6-18-47-3W5, where the regressive succession is a well-developed 12-15 m-thick shoreface succession, which sits sharply on the transgressive sediments overlying Cycle E. Southwest of this location, Cycle F has been removed by, or is laterally equivalent to, non-marine sediments. Basinward of 14-32-47-3W5, the regressive shoreface succession of Cycle F has been removed by fluvial erosion as far northeast as 8-35-48-2W5, where the succession returns as a 3-4 m-thick unit of shelf mudstones/sandstones. Fluvial incision in this area can be seen to have occurred in multiple stages. Fluvial sediments which eroded the shoreface sediments of Cycle F at 10-28-48-2W5 are overlain by the transgressive sediments overlying Cycle F, indicating that this channel was associated with the shoreline system of Cycle F. West of this location, both the earlier fluvial channel and the overlying transgressive sediments have been removed by fluvial erosion, leaving sand-filled channels lying erosively on the transgressive sediments between Cycles E and D.

The youngest cycle present in this section is Cycle H. Cycle G, which sits stratigraphically between F and H, is not present in this section. Cycle H begins at 10-28-48-2W5, where it is a well-developed 15-18 m-thick shoreface sandstone succession that sits sharply on the transgressive sediments overlying Cycle F (or possibly Cycle G). Southwest of this location, Cycle H is laterally equivalent to, or has been eroded by, non-marine sediments. At the northeastern end of the section, Cycle H is still a well-developed 10-11 m-thick shoreface succession. No transgressive sediments overlying Cycle H are present in this section.

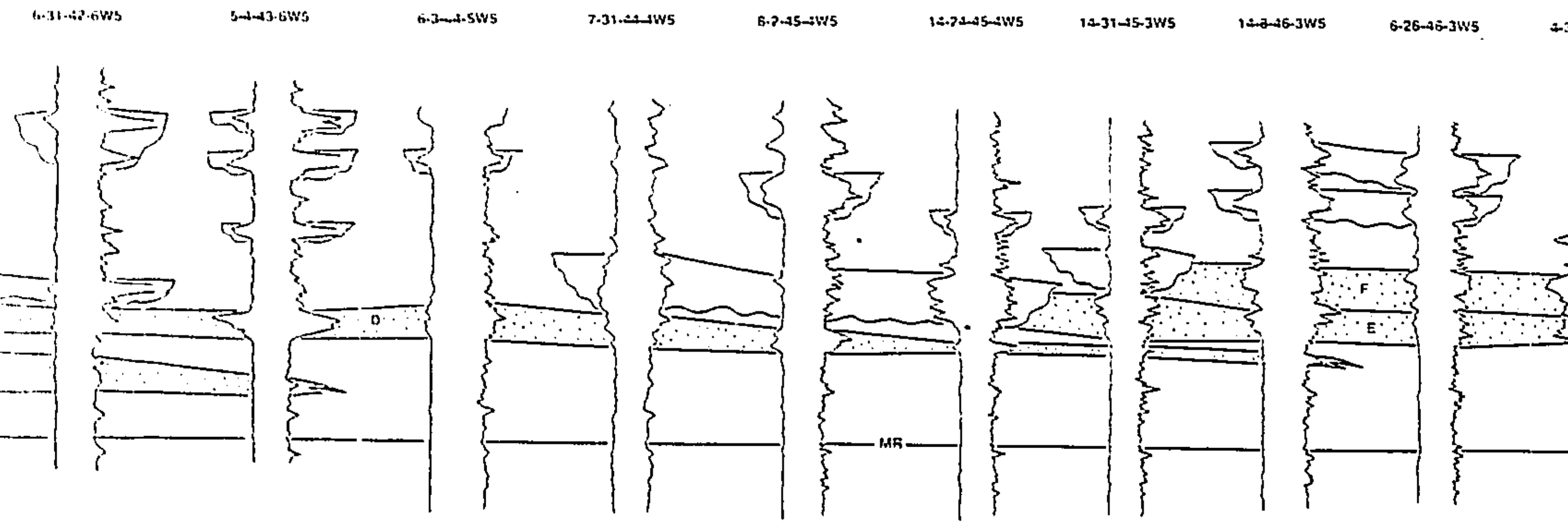
Cross Section C-C'

This dip section (Fig. 4.12) begins in the Ferrier - Willesden Green field and trends northeast through the middle of the Keystone field. The downlapping pattern of marine cycles is clearly evident in this section, and there is little post-depositional erosion of the marine cycles by fluvial processes. The thick fluvial sandstone present near the base of the Belly River Formation at the southwestern end of the section (11-28-41-8W5) is part of the succession of fluvial sediments which are the hydrocarbon reservoir in the Ferrier - Willesden Green area. The relationship of these fluvial sediments to any of the marine cycles is unknown. They may represent the deposits of rivers feeding one of the shorelines within the study area, or they may represent fluvial sediments unrelated to any of the cycles discussed in this thesis.

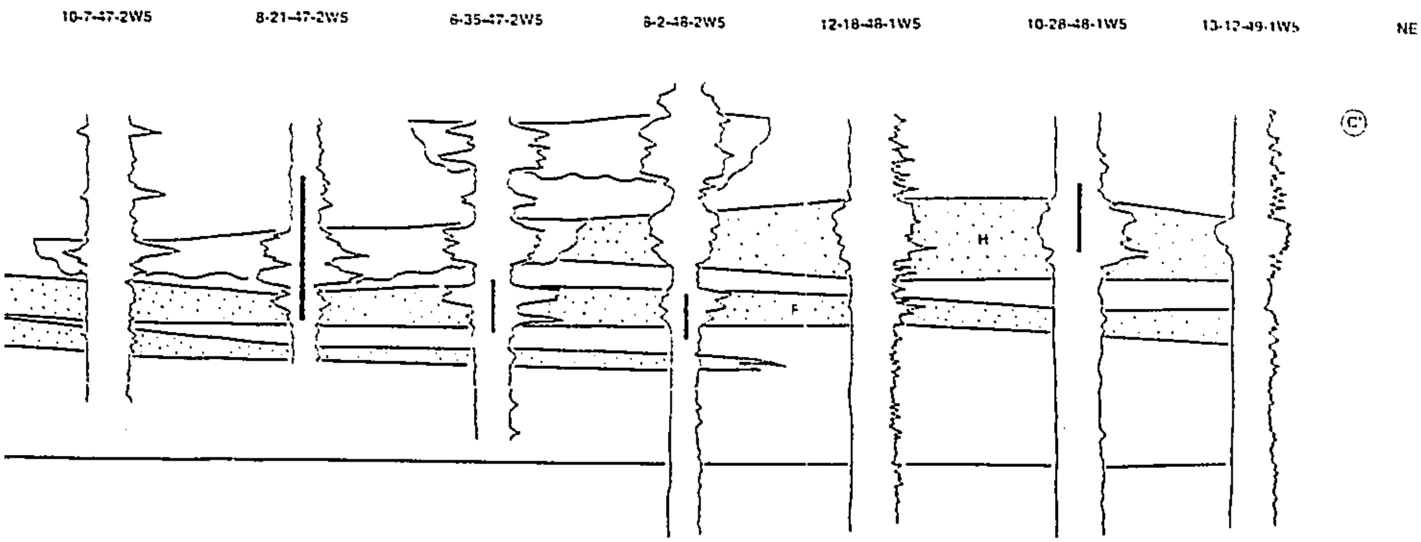
Figure 4.12: Cross Section C-C'. This is a dip-oriented section trending through the Keystone field. All locations show SP and Dual Induction log signatures. The section is approximately 90-100 km in length.



SW
11-28-41-8W5
10-1-42-8W5
14-8-42-7W5
14-14-42-7W5
(c)
50
metres
0



6-31-42-6W5
5-4-43-6W5
6-3-44-5W5
7-31-44-4W5
6-2-45-4W5
14-24-45-4W5
14-31-45-3W5
14-8-46-3W5
6-26-46-3W5
D
F
E
MR


10-7-47-2W5
8-21-47-2W5
6-35-47-2W5
6-2-48-2W5
12-18-48-1W5
10-28-48-1W5
13-12-49-1W5
NE
C'
H
F

Cycle C is the oldest cycle present in this portion of the study area. It exists only as a 3-7 m-thick succession of distal shelf mudstones and sandstones, and reaches its depositional edge by 6-3-44-5W5.

The transgressive sediments overlying Cycle C comprise a 5-10 m-thick unit of shelf mudstones that rises slightly to the southwest before being removed by later fluvial incision at 10-1-42-8W5.

Cycle D begins at 14-8-42-7W5 where the regressive succession is an 8-10 m-thick cleaning-upwards shoreface succession. Southwest of this point, Cycle D is equivalent to fluvial sediments. These sediments may be part of the Ferrier - Willesden Green trend, or they may represent fluvial sediments which are contemporaneous and associated with the Cycle D regression. The shoreface succession thins to the northeast over a distance of 40 km and reaches its depositional edge at 6-26-46-3W5. The transgressive mudstones between Cycles D and E are 3-5 m thick, rise to the southwest, and are eroded completely by fluvial incision at 7-31-44-4W5.

Sediments of Cycle E are first evident at 14-31-45-3W5, where the cycle is present as a well-developed 10-12 m-thick cleaning-upwards shoreface succession. Southwest of this point, Cycle E is laterally equivalent to incised fluvial channels 10-15 m thick, which lie erosively on the transgressive sediments between Cycles D and E. These fluvial sediments may be related to the regressive phase of Cycle E.

The transgressive unit overlying Cycle E is not preserved southwest of 10-7-47-2W5, as it has been eroded by the regressive phase of the following cycle.

The pattern of Cycle F is similar to that of Cycle E. The regressive succession is first present at 14-8-46-3W5 as a 12-15 m-thick well-developed shoreface succession, which sits sharply on Cycle E. Southwest of this are laterally equivalent fluvial channel sediments 10-15 m in thickness which may be related to the regressive phase of Cycle F. The cycle thins to northeast and at the northeastern end of the section is a 2-3 m-thick unit of shelf sandstones and mudstones.

The transgressive sediments overlying Cycle F also follow a similar pattern to those overlying the underlying cycles. They are present as far southwest as 8-21--47-2W5, after which they have been removed by fluvial erosion.

This section crosses Cycle H in an area where the regressive succession is noticeably thicker than in the previous section. Cycle H consists of a 15-20 m-thick shoreface succession which thins slightly to the northeast, but is still well-developed at the northeastern edge of the section. As with the previous cycles, erosionally-based fluvial channels 10-15 m thick are laterally equivalent to the southwest of the regressive phase of Cycle H. The transgressive sediments overlying Cycle H are not preserved in this cross section.

Cross Section D-D'

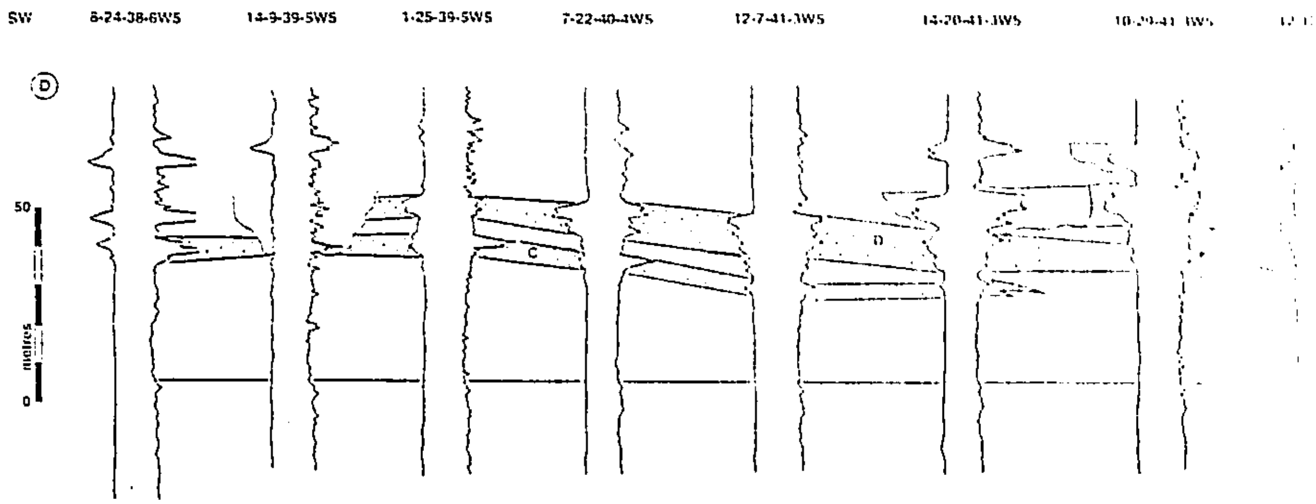
Cross section D-D' (Fig. 4.13) is the only one of the dip sections which is able to use the Milk River shoulder datum for the entire section. As such, it reveals the nature of the downlapping marine cycle geometry more distinctly than previous sections, which were somewhat flattened by the use of higher datums. This section is located between the Ferrybank and Keystone fields (Fig. 4.10).

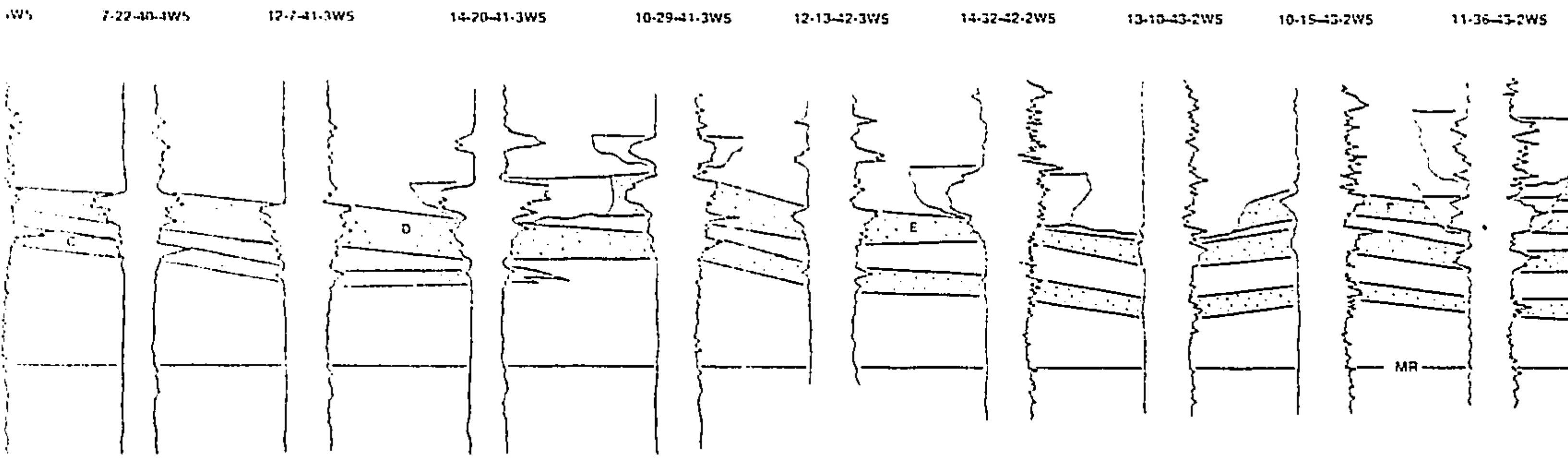
Cycle C is the oldest cycle present in this sections and occurs only as a thin unit of distal shoreface/shelf sandstones which pinches out by 14-20-41-3W5. The transgressive sediments overlying this cycle are eroded by fluvial sediments southwest of 14-9-39-5W5.

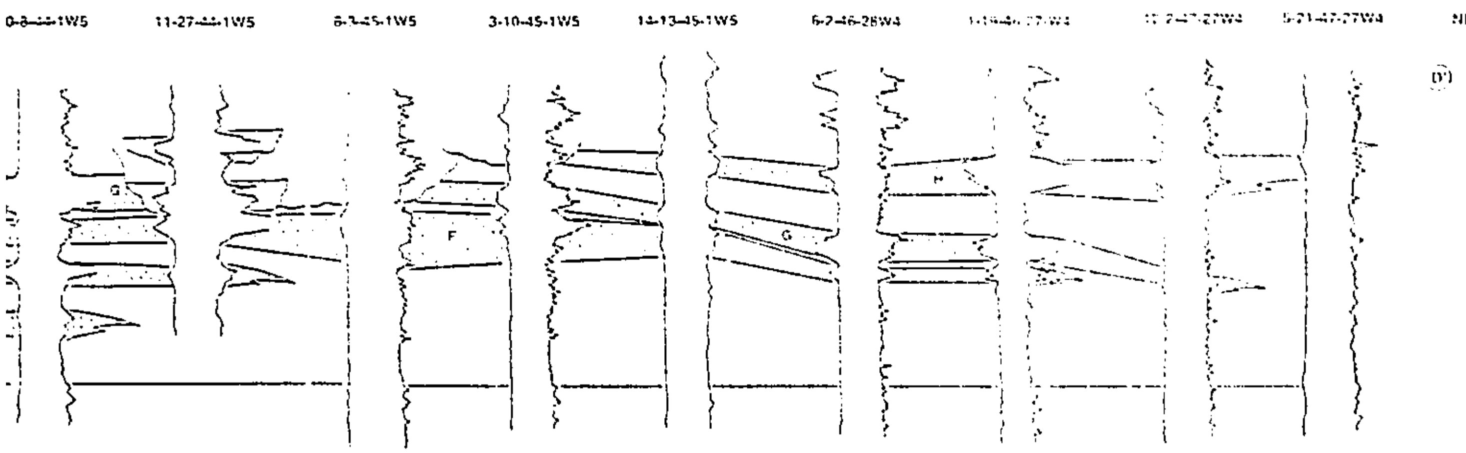
Cycle D is well preserved in this section, showing a well-developed cleaning-upwards shoreline succession at 12-7-41-3W5. It thins basinward over 35 km and pinches out at 10-8-44-1W5. The transgressive sediments overlying this cycle are eroded by fluvial sediments at 14-20-41-3W5 and southwest of this point.

Cycle E shows a similar pattern to that in section C-C'. A well-developed shoreline succession is evident at 10-29-41-3W5. Southwest of this point are laterally equivalent fluvial sediments which lie erosively on the transgressive sediments between Cycles D and E. The cycle thins basinward over 30 km and pinches out at 11-27-44-1W5. The transgressive sediments overlying the

Figure 4.13: Cross section D-D'. This is a dip-oriented section located between the Wilson Creek, Ferrybank, and Keystone fields. All locations show SP and Dual Induction log signatures. The section is approximately 110 km in length.


SW
6-24-38-6W5
14-9-39-5W5
1-25-39-5W5
7-22-40-4W5
12-7-41-3W5
14-20-41-3W5
D
C
D
50
metres
0




11-27-44-1W5
3-10-45-1W5
14-13-45-1W5
6-2-46-28W4
5-21-47-27W4
NE
G
F
G
H

cycle rise to the southwest until 14-32-42-2W5, where they are eroded by fluvial channelling.

The regressive succession of Cycle F does not appear to be as well-developed as in section C-C'. In this section it consists mainly of a thin unit of shoreface/shelf sands ranging from 5-12 m in thickness, which is present over about 25-30 km and pinches out at 1-19-46-27W4. These sediments are also bounded to the southwest by laterally equivalent fluvial sediments which sit erosively on the transgressive sediments overlying Cycle E.

This section is the first to contain sediments from Cycle G. A well-developed shoreline succession is present at 10-8-44-1W5. Immediately to the northeast, the shoreline succession is eroded by fluvial channelling, but reappears as a thinner shoreline/shelf sandstone unit at 3-10-45-1W5. This unit continues to thin basinward and abruptly pinches out at 12-2-47-27W4. Cycle G is also bounded to the southwest by fluvial and associated non-marine sediments, but the geometry of this relationship does not suggest that these sediments may be related to the regressive phase of Cycle G as clearly as for the fluvial channels to the southwest of Cycles E and F. The transgressive sediments overlying Cycle G comprise a 10-20 m-thick unit of shelf sediments which is present as far southwest as 8-3-45-1W5, where fluvial channelling has removed it.

The regressive succession of Cycle H is still present in this area, but is noticeably thinner than the previous section. It consists of a 8-12 m-thick shoreline/shelf succession of sandstone which seems to thin noticeably by the northeastern end of the section (5-21-47-27W4). As with the previous sections, no evidence of the transgressive sediments overlying Cycle H are present.

Cross Section E-E'

This section (Fig. 4.14) is the easternmost of the dip sections, and trends through the Ferrybank field and east of the Keystone field. The familiar downlapping pattern of marine cycles is again evident. Fewer cycles are present in this section than in others, as this section is located basinward of the depositional edge of several of the older cycles.

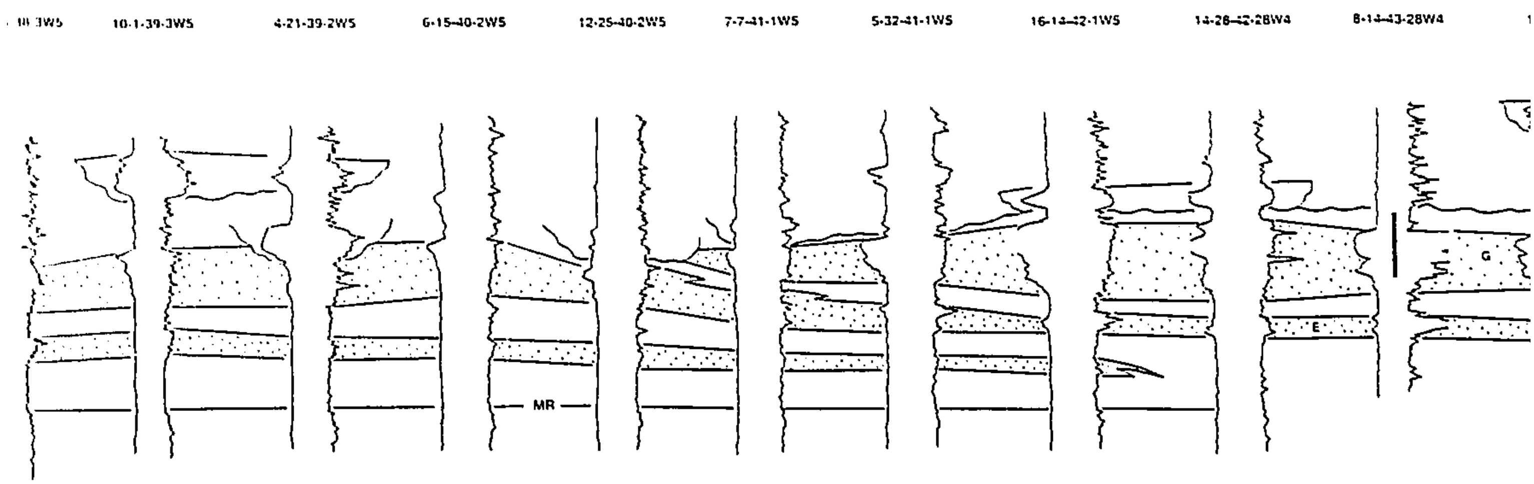
Cycle D is the oldest cycle present in this section, but exists here only as a 5-12 m-thick unit of lower shoreface/shelf sandstones and mudstones that pinches out by 16-14-42-1W5.

The transgressive sediments overlying Cycle D are absent southwest of 8-5-38-3W5, where they are eroded by the regressive phase of the following cycle.

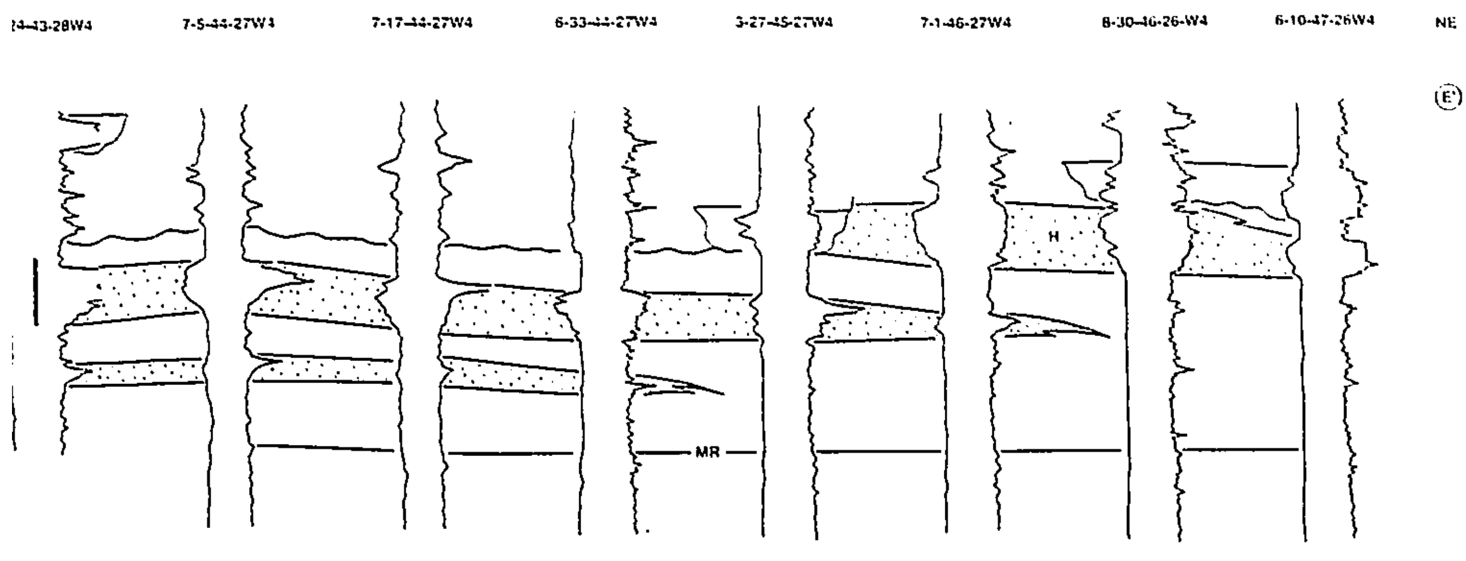
Cycle E is represented by a well-developed 10-15 m-thick shoreline sandstone succession. Immediately northeast of 2-5-37-4W5, this succession sits directly on the regressive succession of Cycle D. Further basinward, the underlying regressive sediments downlap stratigraphically, and the transgressive sediments between Cycles D and E are preserved between the two regressive phases. The regressive succession of Cycle E gradually thins in

Figure 4.14: Cross section E-E'. This is a dip-oriented section trending through Ferrybank and to the east of Keystone. All locations shows SP and Dual Induction log signatures. This section is approximately 100-110 km in length.

SW
2-5-37-4W5
16-13-37-4W5
8-5-38-3W5
10-23-38-3W5
10-1-39-3W5
4-21-39-2W5
E
50
metres
0
E
D
MH



18-3W5
10-1-39-3W5
4-21-39-2W5
6-15-40-2W5
12-25-40-2W5
7-7-41-1W5
5-32-41-1W5
16-14-42-1W5
14-28-42-28W4
8-14-43-28W4
MR
E
G



?4-43-28W4
7-5-44-27W4
7-17-44-27W4
6-33-44-27W4
3-27-45-27W4
7-1-46-27W4
8-30-46-26-W4
6-10-47-26W4
NE
E'
H
MR

a basinward direction over a distance of 80-90 km, pinching out at 6-33-44-27W4.

Transgressive sediments overlying Cycle E are eroded by fluvial incision at 12-25-40-2W5, approximately 40-45 km southwest of the depositional edge of the underlying cycle.

This section runs through the thickest occurrence of Cycle G, whose sediments constitute the reservoir of the Ferrybank field. The cycle first appears as a thin, 6-8 m-thick shoreface succession at 7-7-41-1W5, and quickly thickens to the northeast, reaching 22 m in thickness at 14-28-42-28W4, just southwest of the Ferrybank field. The succession quickly thins to the northeast of Ferrybank, pinching out at 7-1-46-26W4. The overlying transgressive sediments extend southwestward for a distance of over 50 km before they are eroded by fluvial incision at 7-7-41-1W5.

Non-marine sediments directly overlie the transgressive sediments overlying Cycle G for a distance of 40-45 km northeast of 7-7-41-1W5. Just before the northeastern end of the section, these fluvial and associated non-marine sediments are replaced laterally by Cycle H, which consists of a 15 m-thick shoreface succession. Cycle H is beginning to thin in a basinward direction at the northeastern end of the section.

The transgressive sediments overlying Cycle H are just visible in the last well of the section, and are eroded by fluvial sediments southwest of this location.

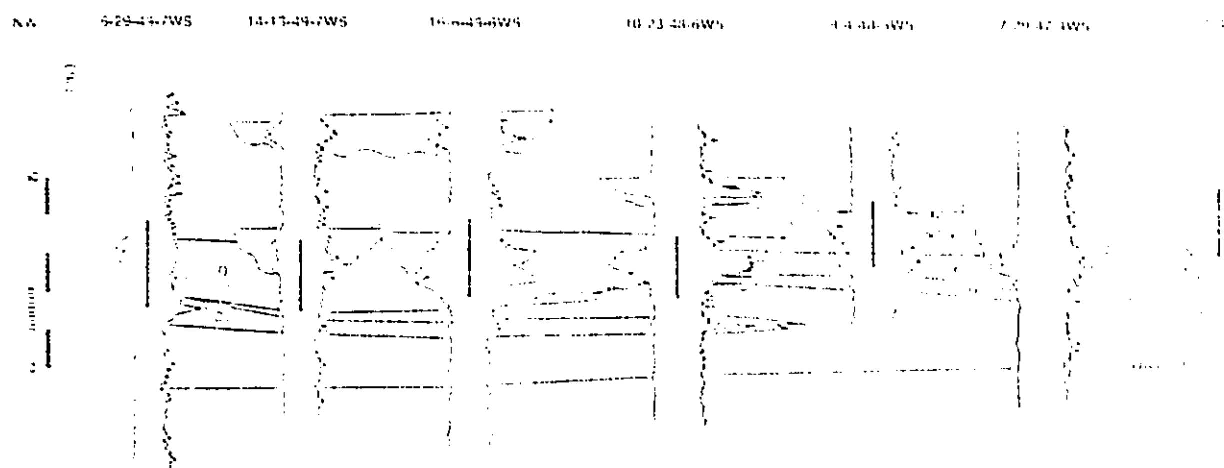
Cross Section F-F'

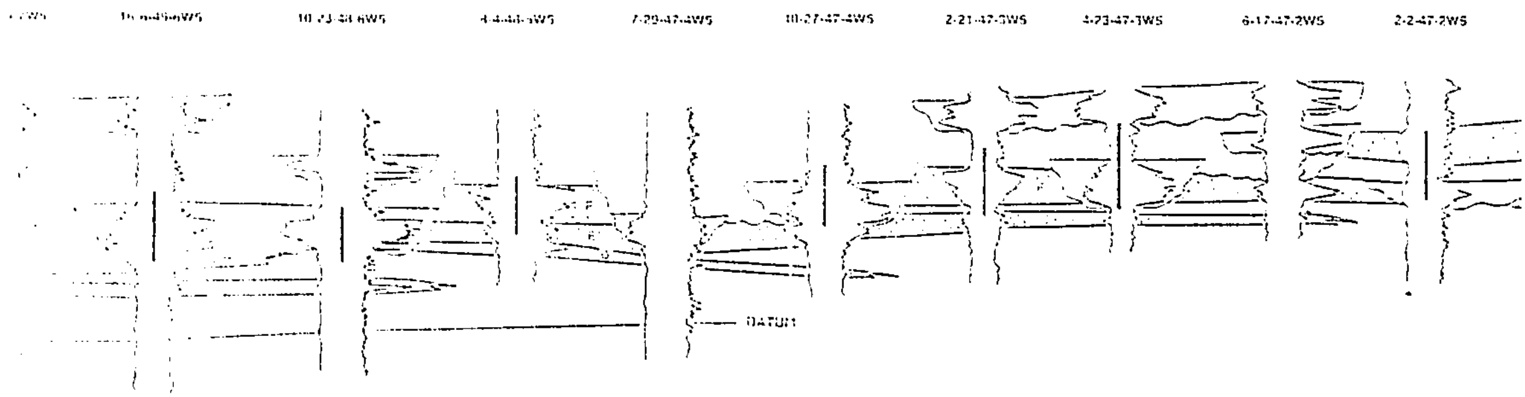
This section (Fig. 4.15) is the first of the "strike" sections to be discussed, and is the most northerly of the strike sections. It trends NW-SE, and runs through the Pembina and Keystone fields.

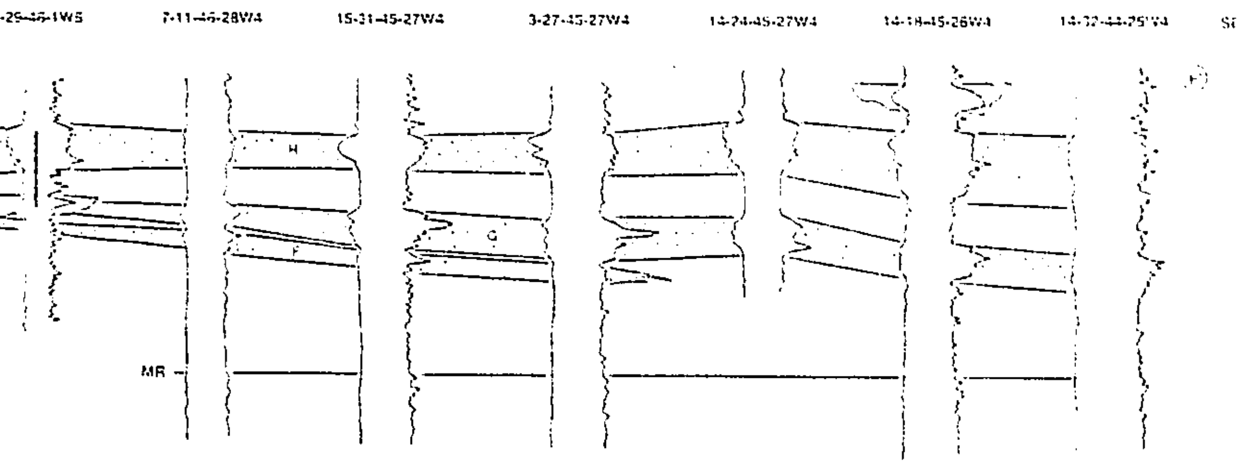
Although this section can be viewed as a regional strike section, it is obvious that it too shows the pattern of downlapping marine cycles that was evident in all of the dip sections. This is because the section actually trends oblique to several of the cycles, and therefore also shows them to be downlapping and pinching out to the southeast.

Cycles C-H are all present in this section. In the northwestern half of the section, the most obvious feature of this section is the erosion of marine shoreface sediments due to fluvial channelling. Channel sandstones up to 20 m thick have eroded into the shoreline sediments of Cycles D-H. This is the same region that showed abundant fluvial channelling in section B-B'. The southeastern half of the section reveals several things about the relationship between Cycles F and G. Most notably this section shows the "shingled" nature of these two cycles, with Cycle G shingled to the southeast over Cycle F. The fluvial channels which erode through Cycle F at 2-2-47-2W5 are laterally equivalent to the regressive shoreface succession of Cycle G, and may be related to the regressive portion of Cycle G. Cycle H is present over much of the southeastern half of this section. The regressive succession shows a thickening and thinning pattern which indicates that there may be a shingling of

Figure 4.15: Cross section F-F'. This is a strike-oriented section trending through Pembina and Keystone. All locations show SP and Dual Induction log signatures. This section is approximately 120 km in length.


14-13-49-7WS


10-23-48-6WS
7-29-47-4WS
10-27-47-4WS
2-21-47-3WS
4-23-47-3WS
6-17-47-2WS
2-2-47-2WS
DATUM



-29-46-1W5
7-11-46-28W4
15-31-45-27W4
3-27-45-27W4
14-24-45-27W4
14-18-45-26W4
14-32-44-25W4
H
G
F
MR

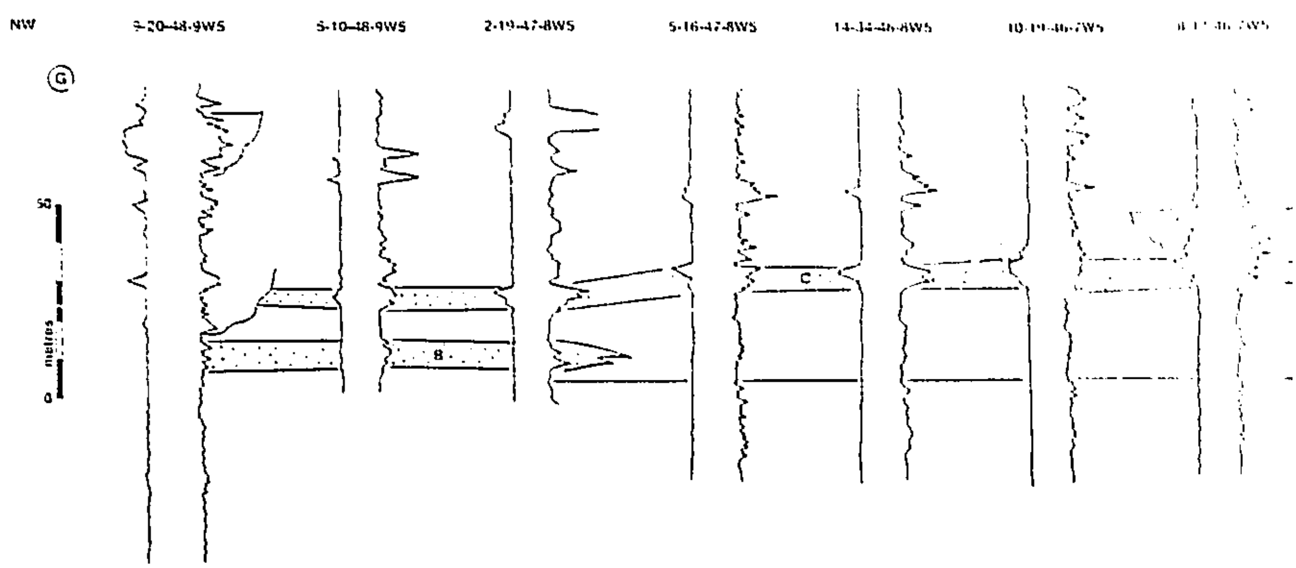
shoreline successions within the regressive portion of the cycle. This relationship will be discussed in further detail in the chapter dealing with Cycle H.

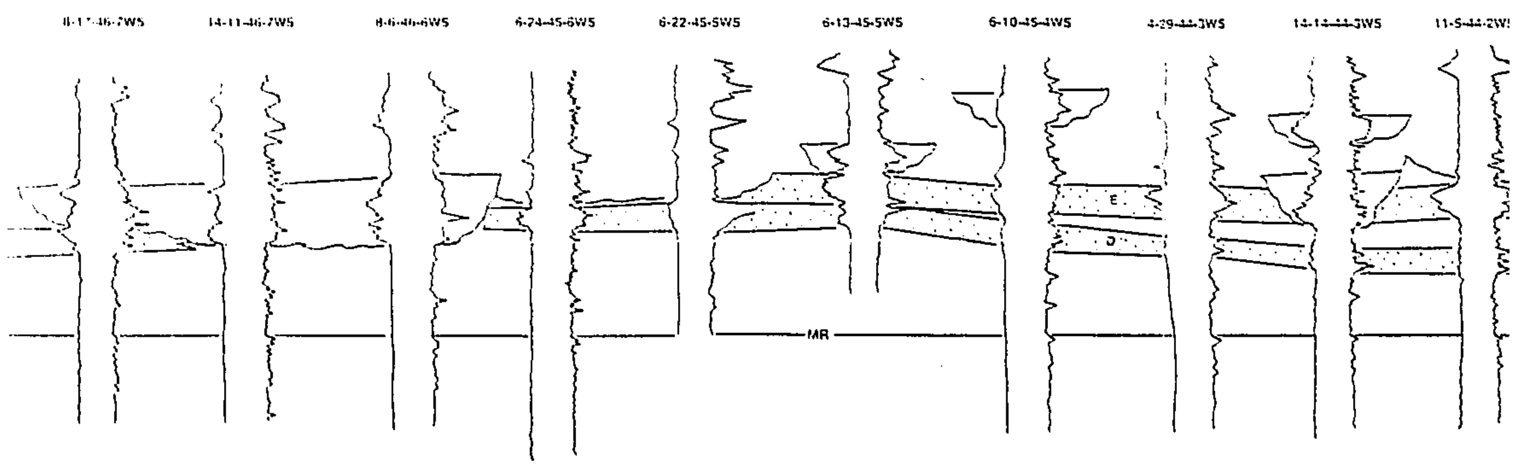
Cross Section G-G'

This section (Fig. 4.16) trends through the southern portion of the Pembina field , south of Keystone, and through the Ferrybank field. This section appears to be a more true strike orientation for most of the cycles present in this area, and the downlapping pattern of the dip sections is less evident.

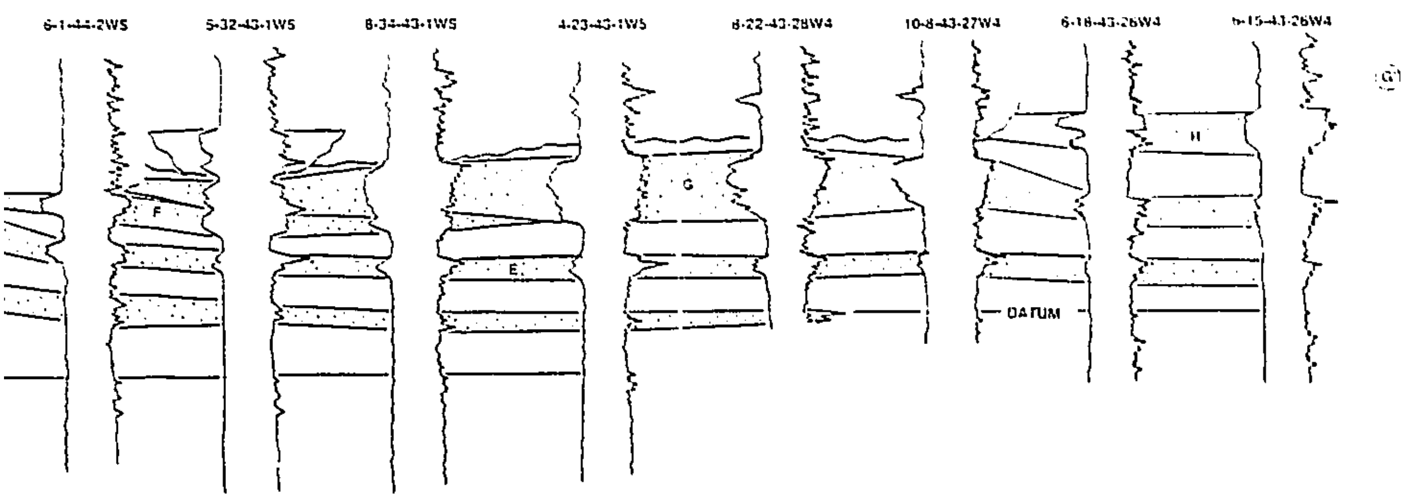
Cycles B-H are present in this section. Cycles B and C consist only of distal shoreface and shelf sediments, but the section contains fairly well-developed shoreline succession for Cycles D-H. Fluvial channelling and erosion of shoreline sediments is not as pervasive in this region as in the northern portion of the Pembina field. Perhaps the most obvious feature of this section is that the southeastern half of the section reveals a much greater vertical stacking of marine cycles in comparison to the northwestern half of the section. In some locations, most notably 8-34-43-1W5, evidence of four separate cycles can be found. This greater degree of aggradation of cycles may be indicative of changes in the relationship of relative sea level fluctuations to sediment supply and/or subsidence rates. This will be discussed later in the thesis. The nature of the shingled relationship between Cycles F and G can be seen in

Figure 4.16: Cross section G-G'. This is a strike-oriented section trending through the southern portions of Pembina, Wilson Creek, and Ferrybank fields. All locations show SP and Dual Induction log signatures. This section is approximately 120-130 km in length.


NW
9-20-48-9W5
5-10-48-9W5
2-19-47-8W5
5-16-47-8W5
14-34-46-8W5
G
50
0
B
C



6-13-45-5W5
6-10-45-4W5
E
D
MR



6-1-44-2W5
5-32-43-1W5
8-34-43-1W5
4-23-43-1W5
8-22-43-28W4
10-8-43-27W4
6-18-43-26W4
6-15-43-26W4
SE
G'
F
E
G
H
DATUM

further detail in this section. Wells 5-32-43-1W5 and 8-34-43-1W5 show that the regressive shoreline succession of Cycle G actually erodes and replaces Cycle F in the Ferrybank region. Cycle G is also very thick in this region as compared to the area just to the southeast of the Ferrybank field.

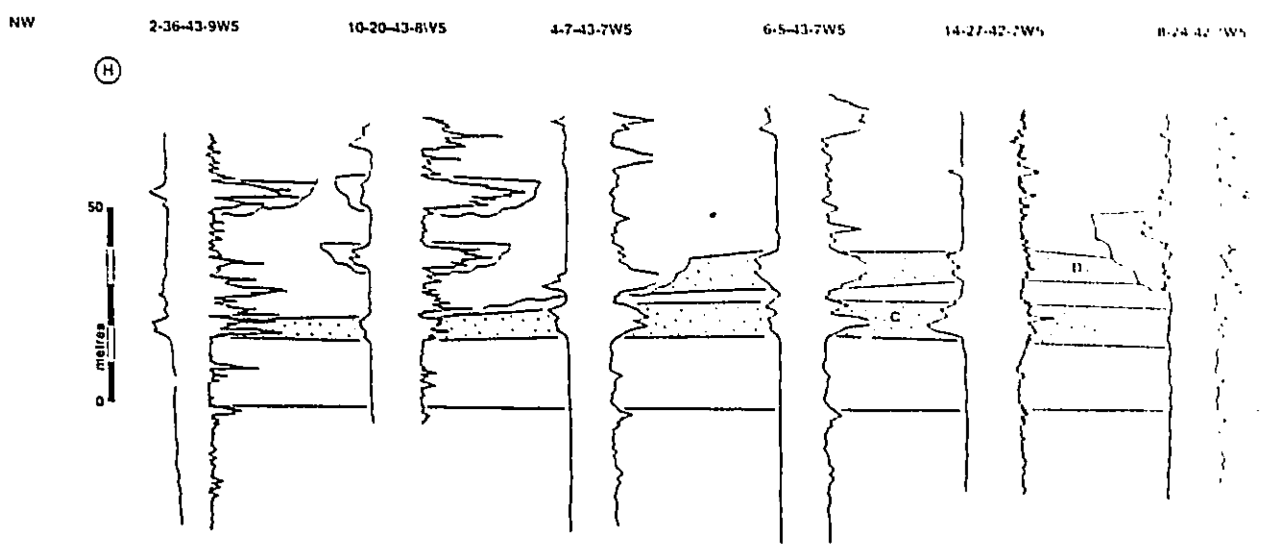
The regionally extensive unconformity between the transgressive sediments overlying Cycle G and the overlying non-marine sediments that was so evident in section E-E' is also visible in this section, extending over a distance of 20 km at the southeastern end of the section. As with section E-E', the unconformity also is laterally equivalent to the base of Cycle H, indicating that the fluvial incision may be related to this regressive episode.

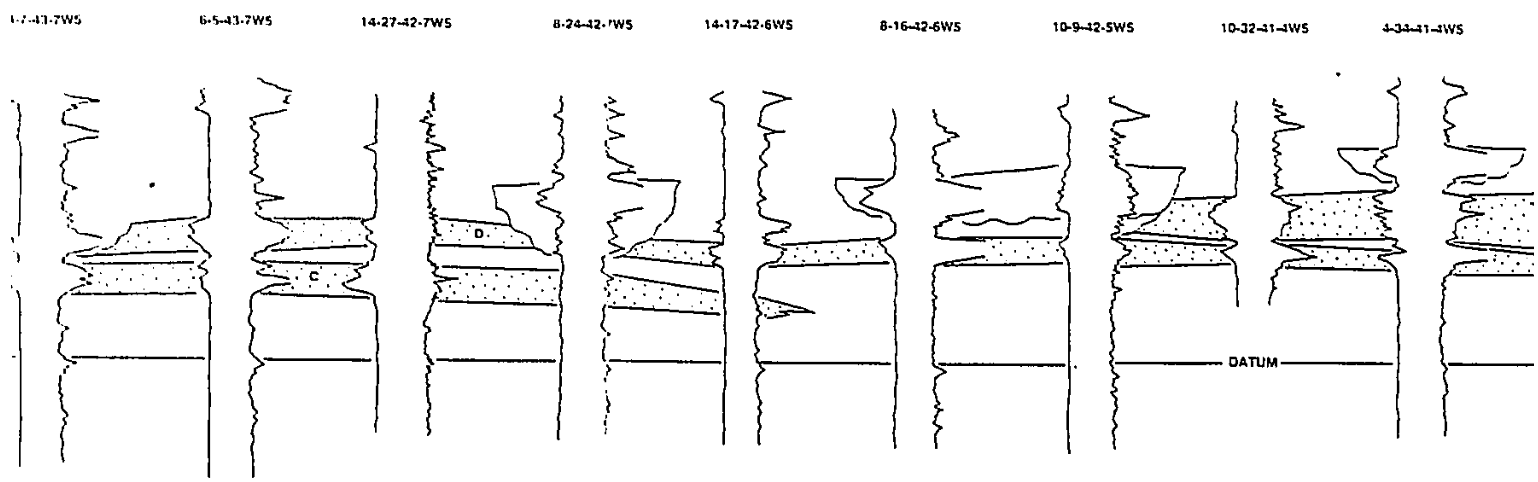
Cross Section H-H'

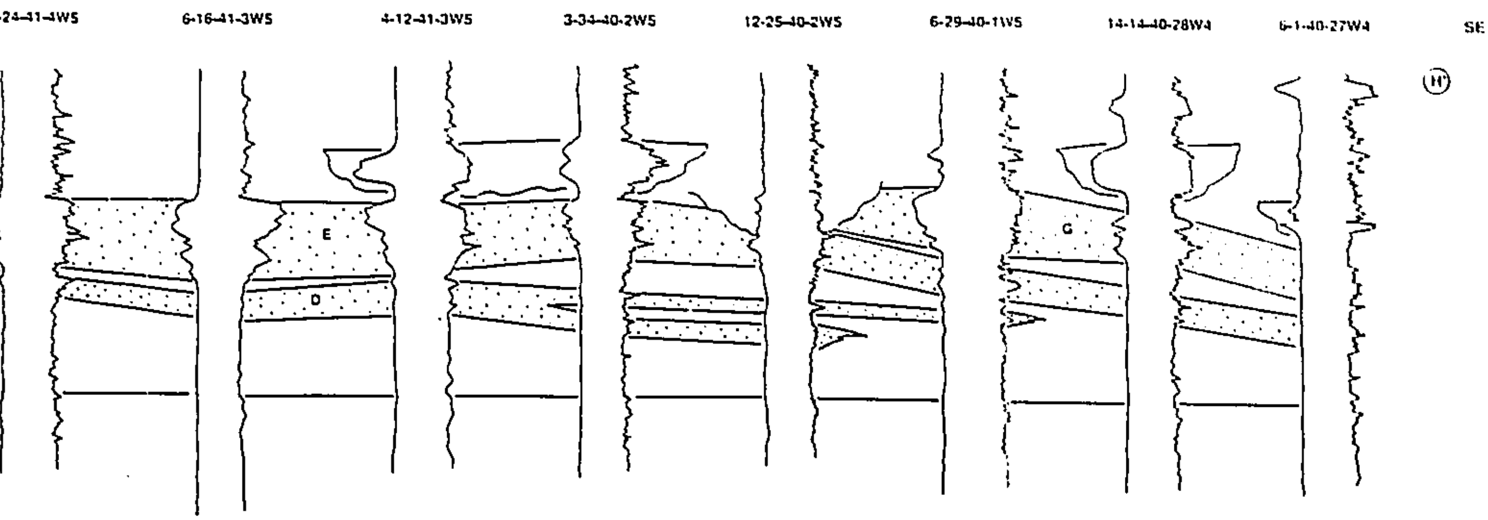
This section (Fig. 4.17) is the southernmost of the strike sections, beginning south of the Pembina field, running just to the north of the Ferrier - Willesden Green field, through the southern edge of the Wilson Creek field, and south of the Ferrybank field to the southeast.

The overall pattern of this section is very similar to section G-G'. Only four marine cycles are present in this section. There is relatively little removal of marine cycles by fluvial erosion. This section shows strike views of several of the older cycles, most notably Cycles C and D. Cycle C is present as a 5-10 m-thick succession of shoreface and shelf sandstones and mudstones. This section shows that Cycle D exhibits relatively little change in thickness along

Figure 4.17: Cross section H-H'. This is a strike-oriented section trending through the southern end of the study area. All locations show SP and Dual Induction log signatures. This section is approximately 90 km in length.


NW
2-36-43-9W5
10-20-43-8W5
4-7-43-7W5
6-5-43-7W5
H
50
metres
0
C
D


6-5-43-7W5
14-27-42-7W5
14-17-42-6W5
8-16-42-6W5
10-9-42-5W5
10-32-41-4W5
4-34-41-4W5
D
C
DATUM



-24-41-4W5
6-16-41-3W5
4-12-41-3W5
3-34-40-2W5
12-25-40-2W5
6-29-40-1W5
14-14-40-28W4
6-1-40-27W4
SE
B'
E
D
G

strike in this region. Cycle E is noticeably thicker in this region than in any of the other sections, with the regressive shoreline succession reaching 15-18 m just to the south of the Wilson Creek field at 10-24-41-4W5 and 6-16-41-3W5. Cycle F is not present in this region. It may be that it was not deposited in this area, or that it has been removed by Cycle G, in a similar manner as was detailed in section G-G'. Cycle G is still well-developed in this section, reaching 15 m in thickness at 6-29-40-1W5, but thins into shelf mudstones and sandstones rapidly to the southeast.

Summary of Regional Stratigraphy

The dip cross sections clearly show that the Lea Park - Belly River transition in central Alberta is characterized by a series of downlapping regressive cycles separated by transgressive sediments, in which each subsequent cycle is deposited in a farther basinward position than the previous cycle. The general direction of progradation of the Belly River wedge appears to have been from the southwest to the northeast, although the precise direction of strike and dip varies from cycle to cycle. The deposits of a cycle tend to have a basinward extent of 60-90 km. The transgressive sediment package overlying each cycle tends to be much thinner than the regressive succession of the underlying or overlying cycle. It also tends to be incompletely preserved, having been eroded at its southwestern end by fluvial and associated non-marine erosion or by marine erosion associated with the regressive phase of a subsequent cycle. In many cases it appears that the erosional unconformity associated with the fluvial incision into the transgressive sediments can be traced basinward and is laterally equivalent with the base of the next cycle (ie. the regressive surface of marine erosion).

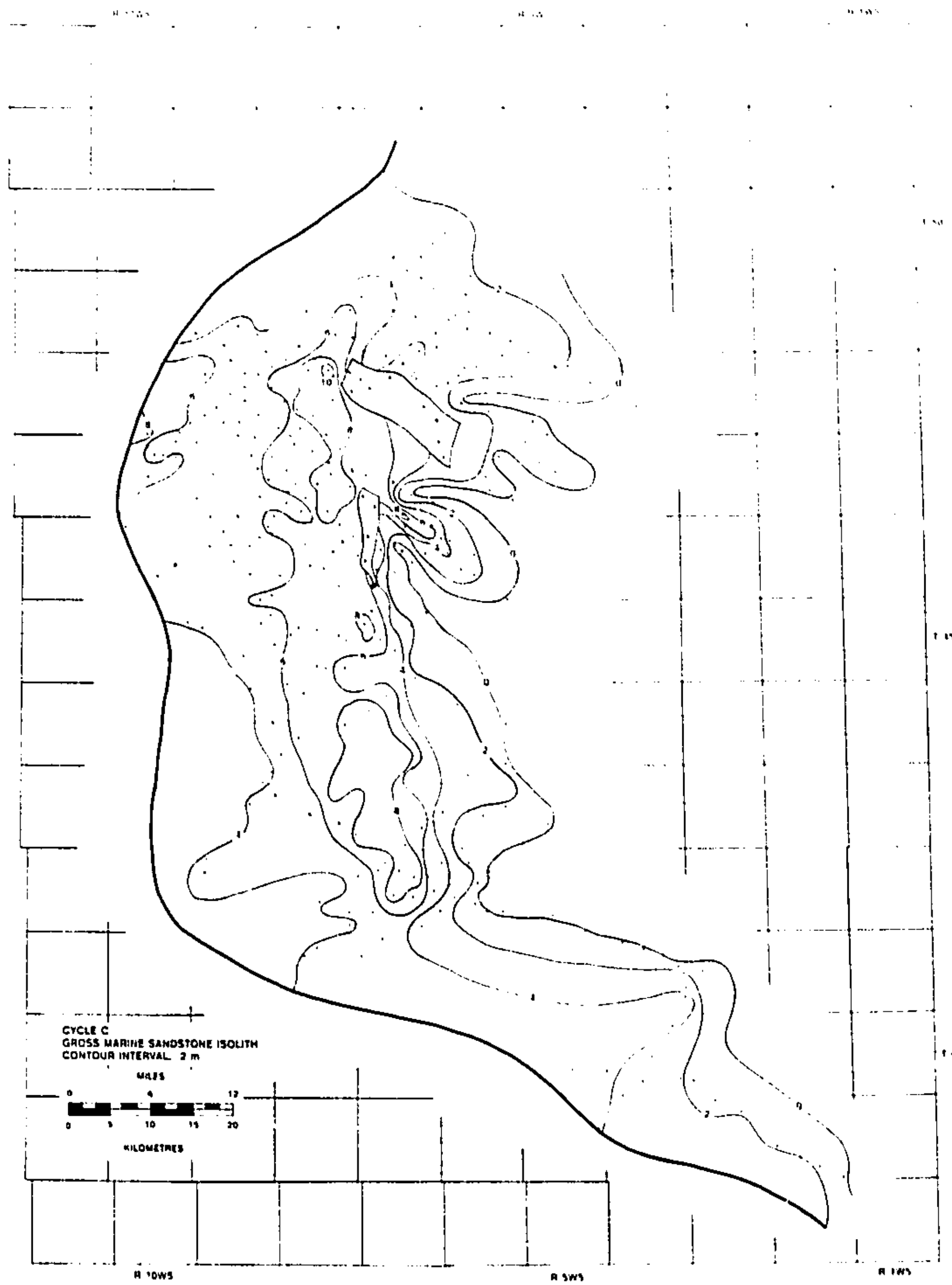
CHAPTER 5: CYCLE C

5.1: Introduction, Distribution, and Geometry

The following six chapters will discuss detailed observations of Cycles C-H. The previous chapter showed that the deposits of Cycles A and B are present within the study area only as distal shelf sandstones and mudstones. No cores of these two cycles were available for study, and therefore these deposits will not be discussed in further detail. Cycle C is the oldest of the cycles to be discussed in detail in this thesis.

The regional cross sections in the previous chapter showed that Cycle C is less than 10 m-thick, and concentrated in the western half of the study area. Figure 5.1 is a gross isolith map of the sandstone in Cycle C. The blocked-off regions denoted by "x" patterns are areas in which the deposits of Cycle C have been removed by post-depositional fluvial erosion, and were treated as null points when isopached. The map shows a relatively linear north-south sandstone body geometry, with isoliths trending parallel or sub-parallel to the paleoshoreline. The thickest deposits of sandstone (6-10 m) occur in a pod 60-70 km long and approximately 15 km wide centred over townships 42-7W5 to 49-8W5. The map shows that the sandstones thin relatively rapidly basinward of this thick area, and reach their depositional edge in 10-30 km, but thin very gradually to the west of the pod in a paleo-landward direction for approximately 30 km before reaching the landward limit of deposition of Cycle C. There are a

Figure 5.1: Gross isolith map of marine sandstones in Cycle C. Contours are in metres, and the contour interval is 2 m. The heavy black line represents the preserved landward edge of deposition of the marine sediments. Zero thickness data points are not plotted on map. All well locations within the study area, but outside the zero contour line contain no Cycle C sediments. This pattern is true for all of the isolith maps shown in this thesis.


CYCLE C
GROSS MARINE SANDSTONE ISOLITH
CONTOUR INTERVAL 2 m
MILES
0
4
12
0
5
10
15
20
KILOMETRES
R 10W5
R 5W5
R 1W5
T 40
T 45

few small protrusions of thicker sandstone trending perpendicular to the main sandstone body, most notably in township 46-6W5 and 7W5. However, this protrusion is mainly present in the 0-4 m isoliths. There is no real evidence of a larger-scale protrusion of the shoreline within the study area, and hence no evidence of a deltaic geometry of the shoreline in plan view.

5.2: Facies Associations

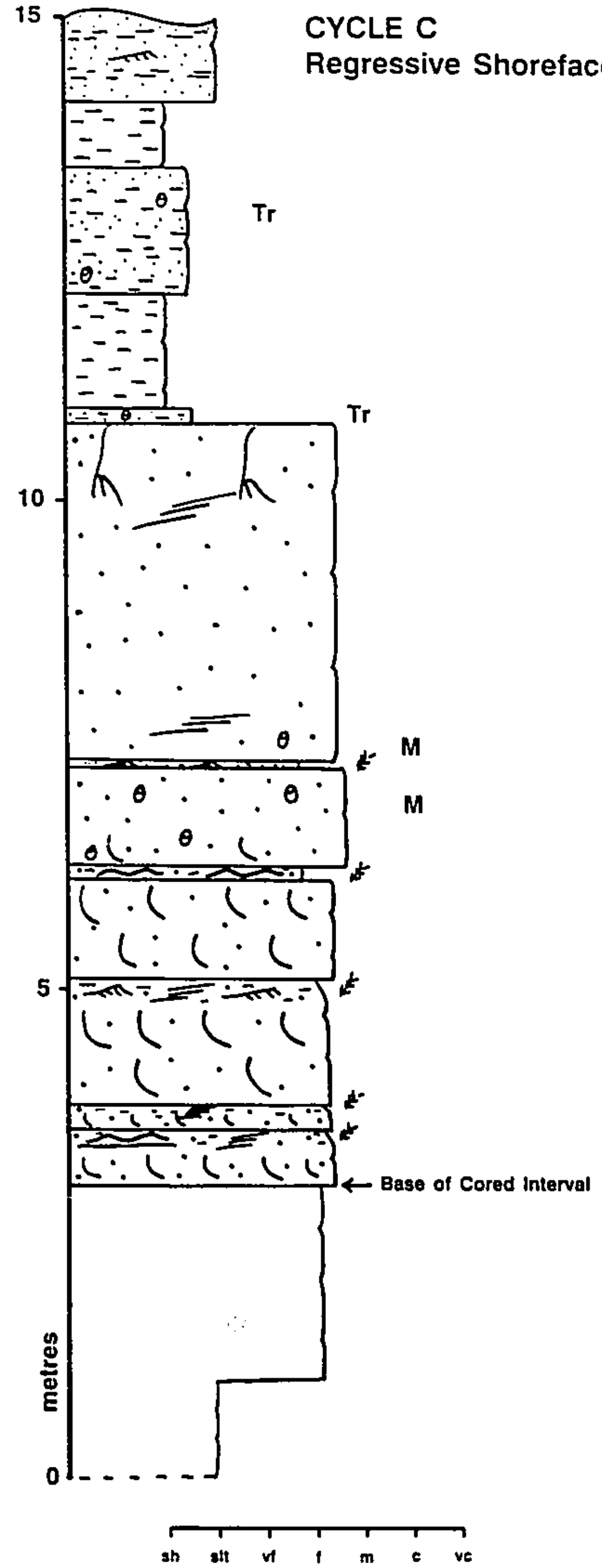
Information on the sedimentological nature of the shoreline sandstones of Cycle C is somewhat limited. There are no cores that completely penetrate the entire cycle except in the distal, basinward regions, and only four cores that contain portions of the cycle in the more landward sandy areas. One of these (16-10-48-8W5; Figs. 5.2) is located near the northern end of the thick, linear pod of Cycle C. The base of Cycle C is not present in the cored interval, but log traces indicate that the base of the cycle is 3 m below the base of the cored interval. This is the same well used to show the typical succession of Facies Association 2b, and the box photographs of the succession are shown in figure 3.17. The cycle would appear to be sharply-based, with little or no transition zone between the underlying transgressive sediments and the shoreface sandstones. The cycle is just under 10 m thick, and is characterized by a coarsening upward succession of marine shoreface sandstones with evidence of subaerial exposure at the top.

The succession preserved in Cycle C belongs to Facies Association 2b,

Figure 5.2: Stratigraphic section through Cycle C in well 16-10-48-8W5. Vertical scale is in metres. Legend of symbols is given in Figure 3.1. Horizontal grain size scale is as follows; sh - shale/mudstone, slt - siltstone, vf - very fine-grained sandstone, f - fine-grained sandstone, m - medium-grained sandstone, c - coarse-grained sandstone, vc - very coarse-grained sandstone.

16-10-48-8W5

CYCLE C
Regressive Shoreface Succession



and is dominated by cross-bedded sandstones. The cross-bedded sandstones contain abundant *Macaronichnus* burrows, and some beds show evidence of wave-reworking at the tops. The cross-bedded interval passes upward into a slightly coarser 4-5 m-thick unit of vaguely stratified LAIS sandstones, which are rooted at the top, and overlain by fine-grained non-marine sediments. Overall, the facies succession of Cycle C is characteristic of a prograding non-deltaic shoreline environment.

5.3: Cross Sections

Figure 5.3 shows the location of cross sections through Cycle C. Figure 5.4 is a log cross section which trends in a dip direction. The cross section is hung on the top of Cycle B as a datum and extends for a length of about 40 km. The internal stratigraphy of Cycle C appears to be fairly simple and consists of only one succession. It begins as a thin unit of sandstone which thickens to the east, reaching its maximum thickness of 10 m at 4-36-48-8W5. It then quickly thins again over a distance of 10 km, and by 6-24-48-7W5 is just a 3-4 m-thick unit of shelf sandstones and mudstones which continues basinward for approximately 20 km before reaching depositional edge.

Figure 5.5 is a core cross section along the same line as the dip log cross section. It is also hung on the top of Cycle B as a datum and trends from the middle of the thick, central pod for a distance of about 15 km in a basinward direction. The two westernmost cores in the section are within the

Figure 5.3: Simplified isolith map of Cycle C, showing the location of the cross sections I-I' and J-J'. The 2, 6, and 10-m isoliths are shown on this map, along with the preserved landward edge of deposition.

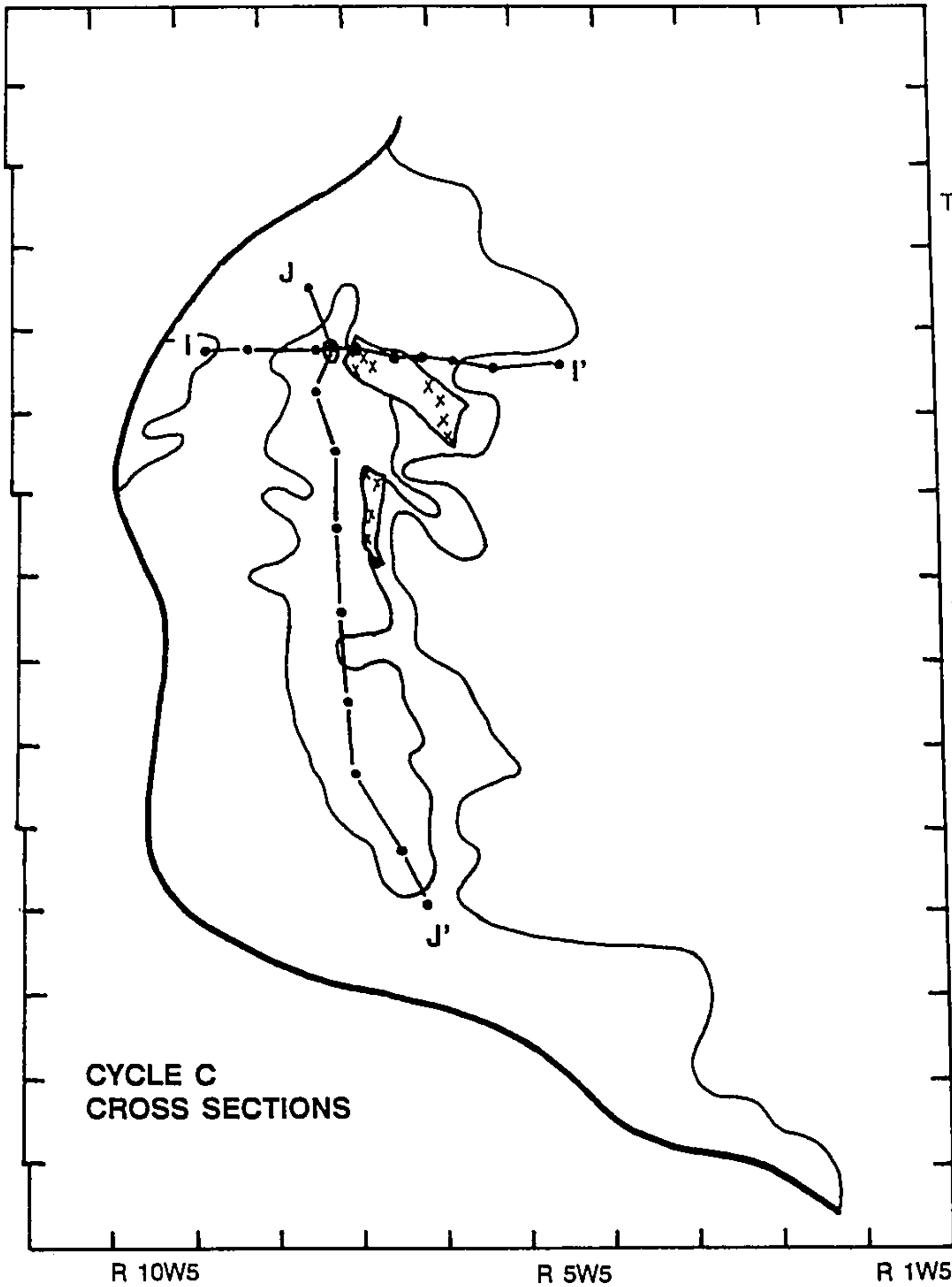

J
I
I'
J'
T 50
T 45
T 40
CYCLE C
CROSS SECTIONS
R 10W5
R 5W5
R 1W5

Figure 5.4: Log cross section I-I", trending parallel to dip. Stippled areas indicate marine sandstone, and each cycle is designated by letter. The cross section is hung using the top of Cycle B as a datum.

CYCLE C: LOG CROSS SECTION PARALLEL TO DIP

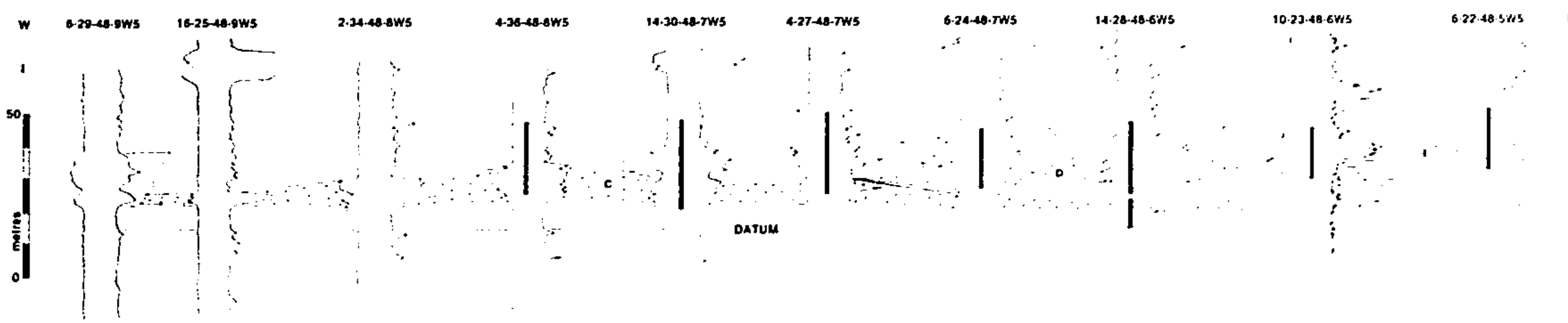
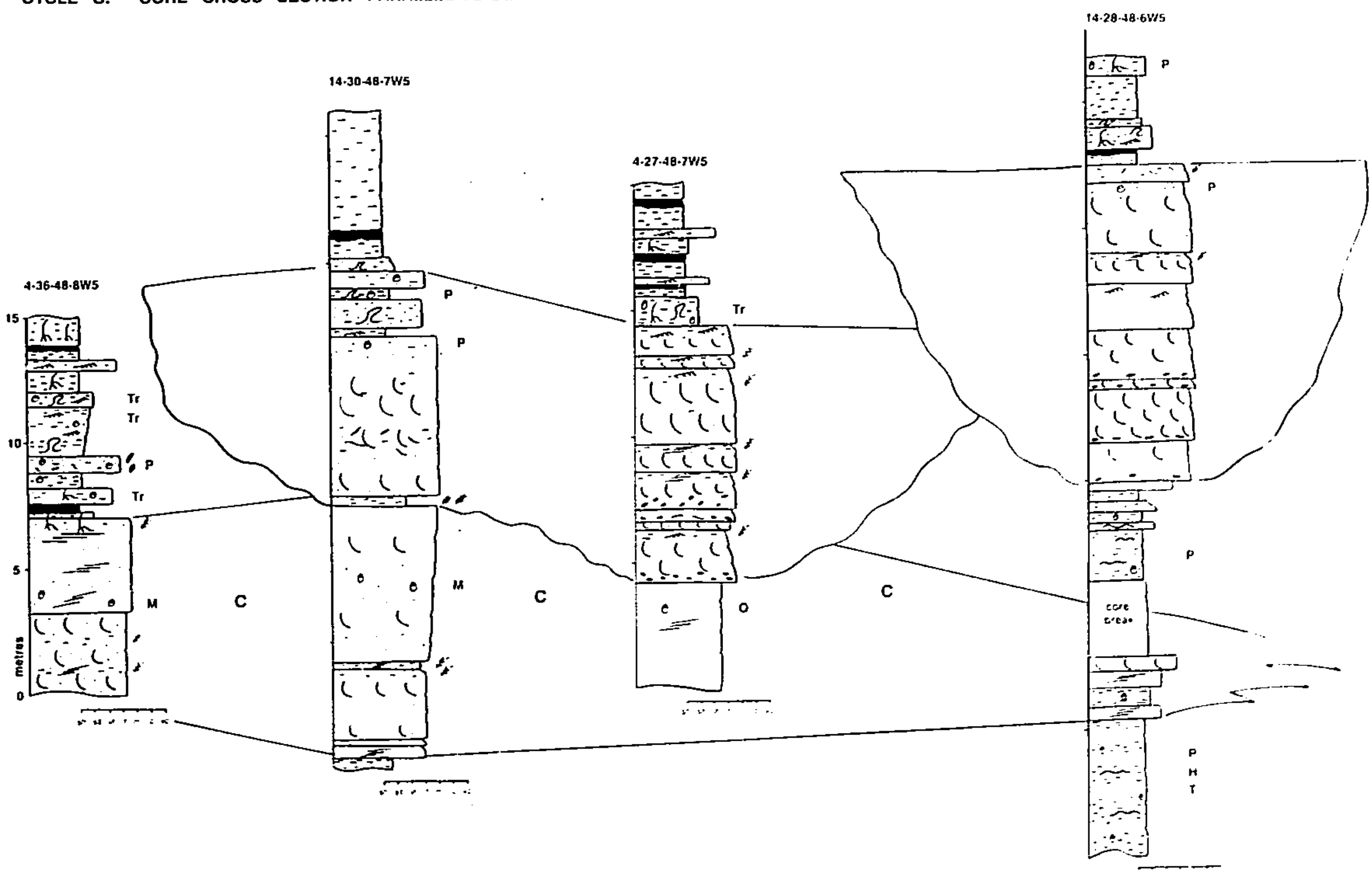


Figure 5.5: Core cross section parallel to dip. Section trends along the same line as log cross section I-I'. Wiggly lines indicate erosional surface due to fluvial incision. Note the removal of the upper parts of Cycle C by fluvial erosion at 14-30-48-7W5 and 4-27-48-7W5. The section is hung using the top of Cycle B as a datum.

CYCLE C: CORE CROSS SECTION PARALLEL TO DIP



thick central pod of sandstone and both show similar facies successions within the marine shoreface to figure 5.2. The base of Cycle C is just penetrated in 14-30-48-7W5, and shows that the shoreface is very sharply-based, with no transition zone from shelf mudstones to shoreface sandstones preserved. In 4-27-48-7W5 the cycle is composed of more massive, poorly stratified LAIS sandstones which thin over 10 km into 3-4 m of thin, interbedded LAIS sandstones and mudstones in 14-28-48-6W5.

The transgressive sediments which overlie Cycle C are shown in 14-28-48-6W5 in figure 5.5. They are typical of Facies Association 1, being composed of interbedded shelf mudstones, siltstones and sandstones with some burrowing and wave rippling of the sediment. The log cross section (Fig. 5.3) shows that these transgressive sediments are not preserved west of 4-27-48-7W5, having been removed by fluvial erosion.

Both the log and core cross sections show that immediately to the east of the thickest succession of Cycle C, fluvial channels erode into the cycle, removing the top portions of the succession. These are the same channels which eroded the transgressive sediments overlying Cycle C. These channels contain sediments typical of Facies Association 3. They are composed of stacked fining upward successions of cross-bedded to current-rippled sandstones. The exact relationship of these channels to shoreline sediments of Cycle C is unknown, but figure 5.3 shows that these channels are laterally equivalent to the shoreline sediments of Cycle D, and may be related to this cycle.

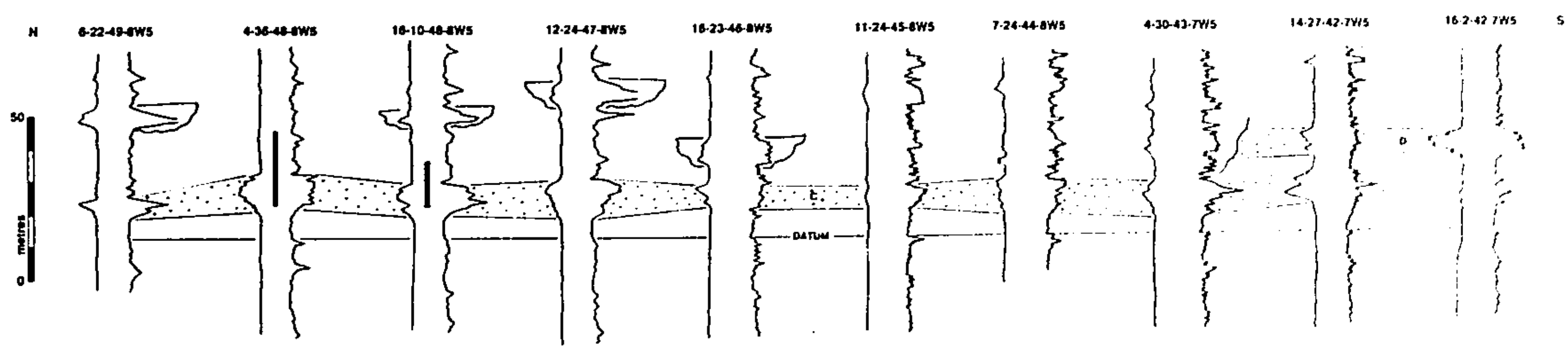
Figure 5.6 is a log cross section which is oriented in a strike direction, parallel to the paleoshoreline. It trends form north to south through the central thick sandstone over a distance of about 70 km. This section is also hung using the top of Cycle B as a datum. The section shows that Cycle C contains a single shoreface succession, which thickens and thins slighlty along strike, but is otherwise unvarying.

5.4: Interpretation

Cycle C is interpreted to represent the deposits of a prograding shoreface environment. The presence of *Macaronichnus* in the upper half of the succession indicates that the upper shoreface environment was a well-oxygenated high-energy environment, and the dominance of cross-bedded sandstones throughout the lower and middle portions of the succession indicates that unidirectional currents were capable of making dunes in fine grained sand. These currents may have been wave-induced longshore currents, tidal currents, or currents related to fluvial input at some location along the shoreline. Wave ripples at the tops of some sandstone beds indicate that the shoreline was at least influenced to some degree by wave activity, and wave processes may have generated the currents which were the dominant process in moving sediment in the shoreface environment of Cycle C. The study area contains no evidence of significant deltas within the shoreline

Figure 5.6: Log cross section J-J' trending parallel to strike through Cycle C. Section is hung using the top of Cycle B as a datum.

CYCLE C: LOG CROSS SECTION PARALLEL TO STRIKE



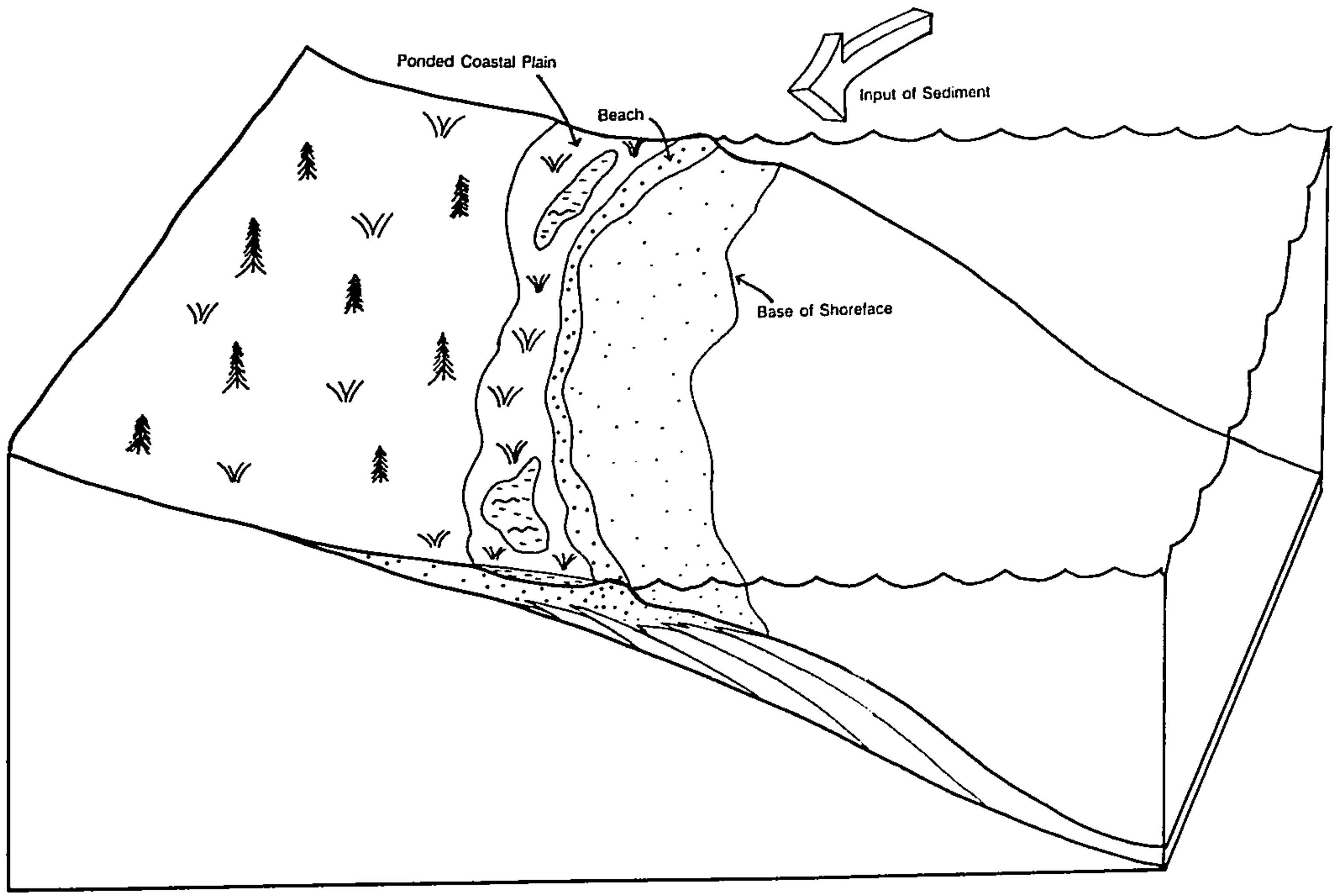
sediments of Cycle C, indicating that major point(s) of input of sediment into the system is(are) located outside the study area. The geometry of the system would therefore support the interpretation that longshore drift currents may have been the dominant process operating within the shoreface environment. The shoreline system of Cycle C is more analogous to a strandplain system, although the uppermost beach portions of the shoreline are not as well developed as in typical wave-dominated strandplain environments. The non-marine sediments which immediately overlie Cycle C in 4-36-48-8W5 contain a coal horizon and muddy sandstones containing oyster shells. This is indicative of a wet marshy environment which was at least partially saline at times, allowing for oysters to inhabit the region. This would be typical of a coastal plain environment landward of a shoreline, which progrades out over top of a shoreline as the shoreline advances basinward. Figure 5.7 shows a possible paleogeographic reconstruction of the depositional environment.

The base of the shoreline succession of Cycle C is very sharp, and there appears to be very little or no transition zone between the shelf sediments and the shoreface sediments. The lowermost sediments within the shoreface are also indicative of a relatively shallow, high energy environment (cross-bedded sandstone). This may indicate that the establishment of a shoreface environment in this area was rapid, and the water depth decreased from shelf to shallow shoreface depths over a short period of time.

The isopach map of Cycle C shows that the thickest sandstones are not

Figure 5.7: Interpreted depositional environment of Cycle C. The environment is interpreted to have been a relatively straight, non-deltaic sandy shoreline that was supplied by longshore drift.

DEPOSITIONAL ENVIRONMENT: CYCLE C



located at the landward limit of deposition, but are instead concentrated in a shore-parallel unit about 20 km east of the landward limit of deposition. The increased thickness in this location may indicate either the final, most basinward point of progradation of the shoreline system (and hence the most completely preserved succession), or that the shoreline was stable in this location for a relatively long time, and thus accumulated the thickest shoreline succession in this location. The former interpretation is somewhat favoured, simply because the isopach map also shows that Cycle C begins to thin rapidly immediately east of the linear thick trend into more distal shoreface and shelf sediments, and reaches depositional edge a short distance basinward. This indicates that the shoreline system did not prograde a great distance, if at all, past the location of the thick linear trend.

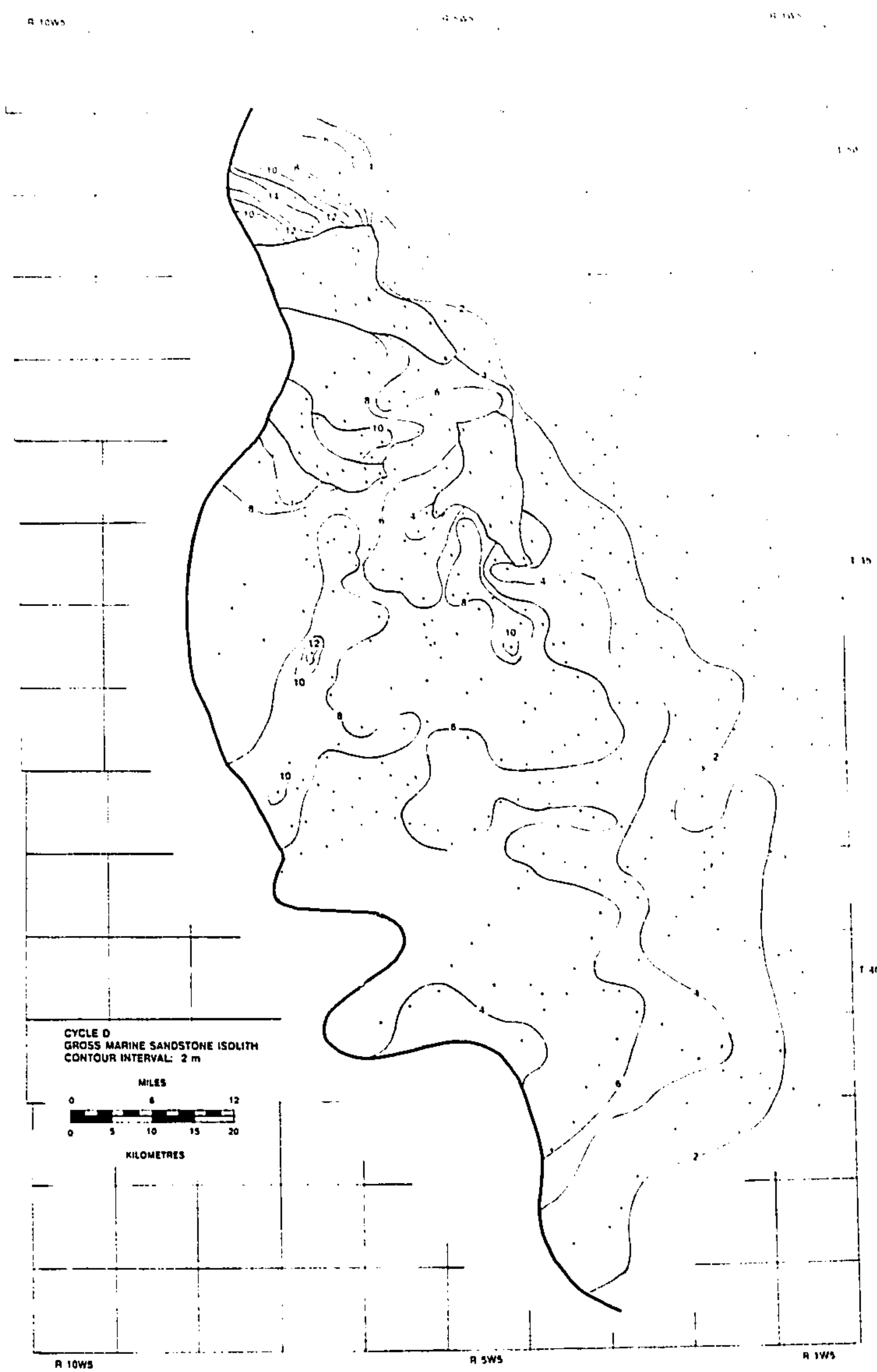
The contact with the overlying transgressive sediments is only observed in cores through the distal portion of Cycle C. As such, this bounding discontinuity is characterized by a deepening of facies from shelf sandstones and mudstones to deeper shelf mudstones. The transgression does not appear to have moved very far back towards the west, as the only evidence of marine waters in sediments overlying the thickest part of Cycle C is the presence of oysters in the coastal plain sediments above the cycle in 4-36-48-8W5. This may indicate a minor induction of saline water at the most landward edge of a transgression.

CHAPTER 6: CYCLE D

6.1: Introduction, Distribution, and Geometry

The regional cross sections in Chapter 4 show that Cycle D is present throughout much of the western and central parts of the study area, and reaches a maximum thickness of 14 m in the northern portion of the Pembina field (Fig. 6.1). The blocked-off regions denoted by "x" patterns in this figure are areas in which all or part of Cycle D has been removed by post-depositional fluvial erosion. The map shows that Cycle D strikes in a NNW-SSE orientation and thins to the northeast, except for the northern Pembina region, which shows a small, thick tongue of sandstone up to 15 m in thickness that thins from the northwest to the southeast. The general pattern of the isoliths shows that the cycle has a broad, lobate geometry in plan view, and extends basinward from its landward limit of deposition for approximately 70-80 km. Two distinct "protrusions" of thicker sandstones are visible within the study area. The first is the obvious thick tongue in the northern portion of Pembina. This tongue is also part of a broader lobate-shaped sandstone body outlined by the 8 m isolith. The second is an elongate tongue just north of the Ferrier - Willesden Green Field (Townships 42-44, Ranges 6-7W5) which thins to the northeast. Both of these tongues of sediment are laterally restricted along strike to local areas. The areas where fluvial erosion has removed Cycle D tend to begin and/or end abruptly, indicating that incision associated with a

Figure 6.1: Gross isolith map of marine sandstone of Cycle D. Contours are in metres. Contour interval is 2 m. Heavy black line represents the preserved landward edge of deposition of marine sandstones.


R 10W5
R 5W5
R 1W5
T 35
T 40
CYCLE D
GROSS MARINE SANDSTONE ISOLITH
CONTOUR INTERVAL: 2 m
MILES
0 6 12
0 5 10 15 20
KILOMETRES
R 10W5
R 5W5
R 1W5

particular episode of fluvial erosion does not necessarily downcut to a similar stratigraphic level throughout the region.

6.2: Facies Associations

Information on the sedimentological nature of Cycle D is more abundant then for Cycle C. There are 9 cores which contain all or part of the cycle. Most of these are located within or near the thick tongue of sandstone in northern Pembina. Figures 6.2 and 6.3 show an example of Cycle D, located at 8-22-49-7W5. The base of the cycle is not present in the core section, but logs and core sections from other locations indicate that the base of the cycle is sharp, with little or no transition zone between the cycle and the underlying sediments. In this location, the cycle is 15 m-thick, and is characterized by a coarsening-upward succession of marine shoreface sandstones with evidence of subaerial exposure at the top.

This succession is typical of Facies Association 2a. The lower half consists of thick beds of dominantly massive sandstone, which become laminated towards the tops of the beds. These beds can apparently reach 7-8 m in thickness. In figure 6.2, the thickest of these beds is just under 5 metres. The cross-bedded zone which overlies these dominantly massive beds is sharply-based, and the cross-bedded sandstones are noticeably coarser than the underlying sandstones (medium-grained vs. fine-grained). Where the succession is fully preserved, the cross-bedded zone is overlain by a 3-5 m-

Figure 6.2: Stratigraphic section through shoreline succession of Cycle D in well 8-22-49-7W5.

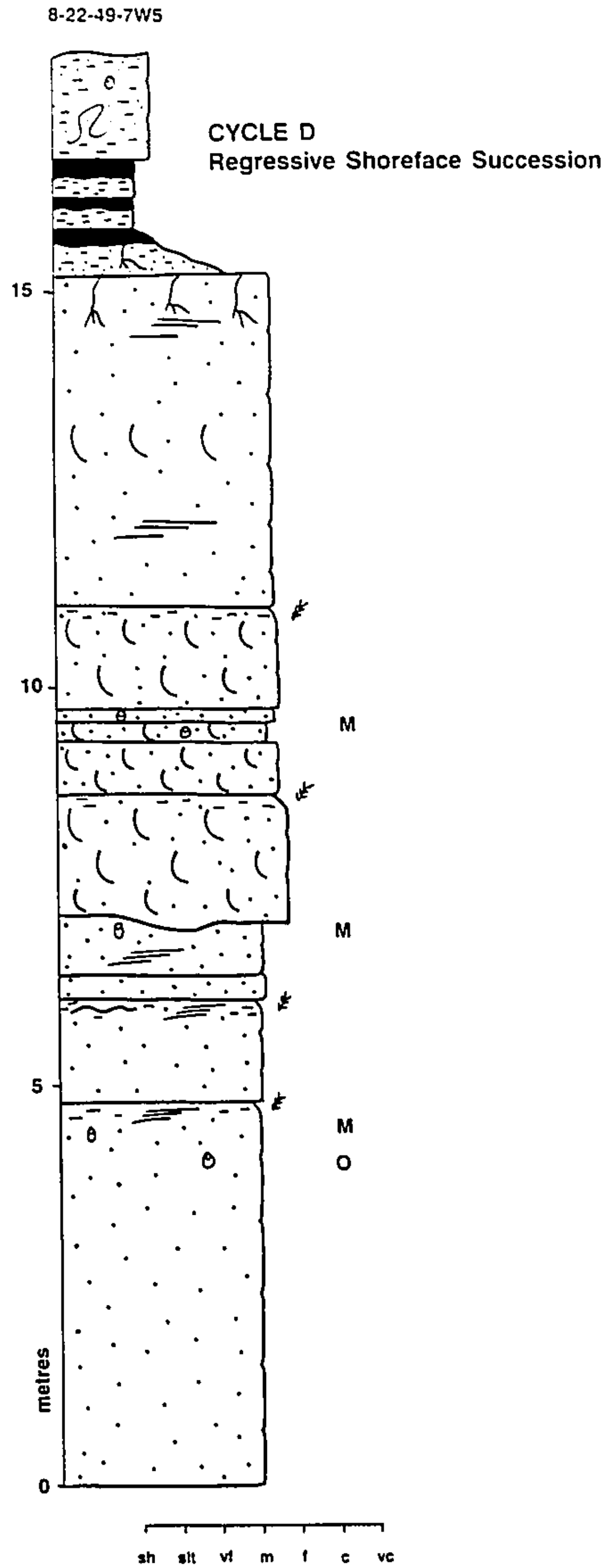
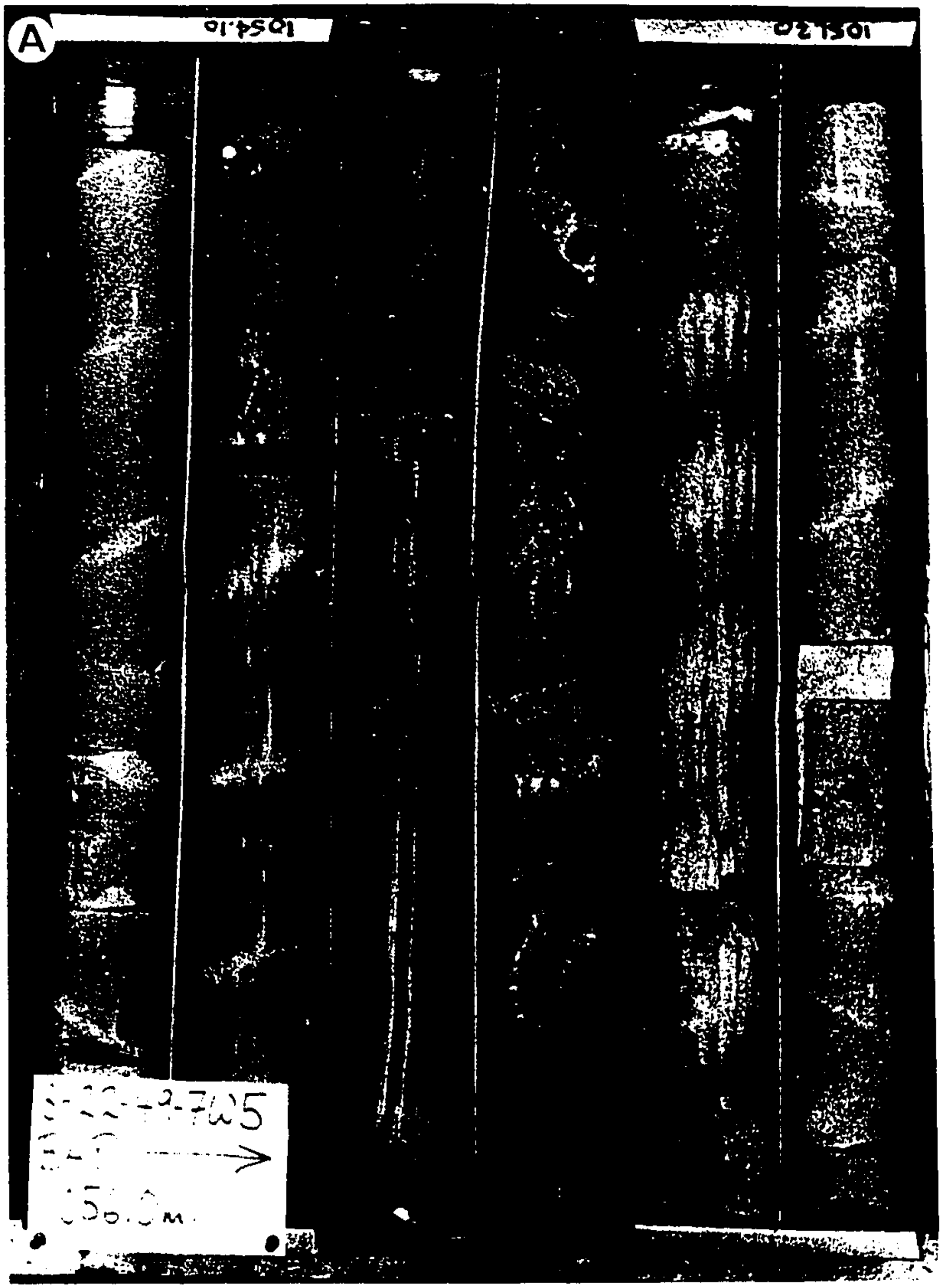
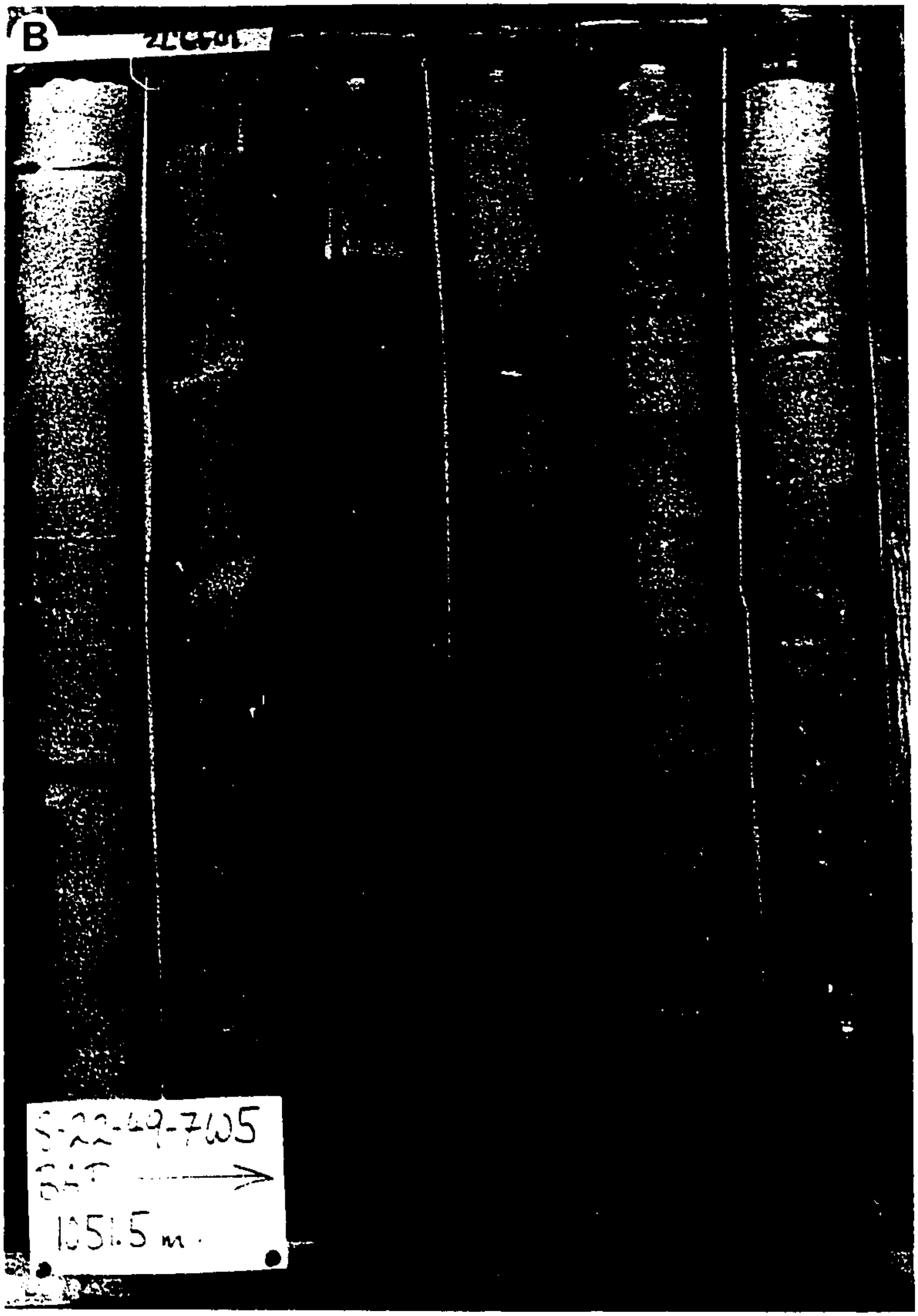
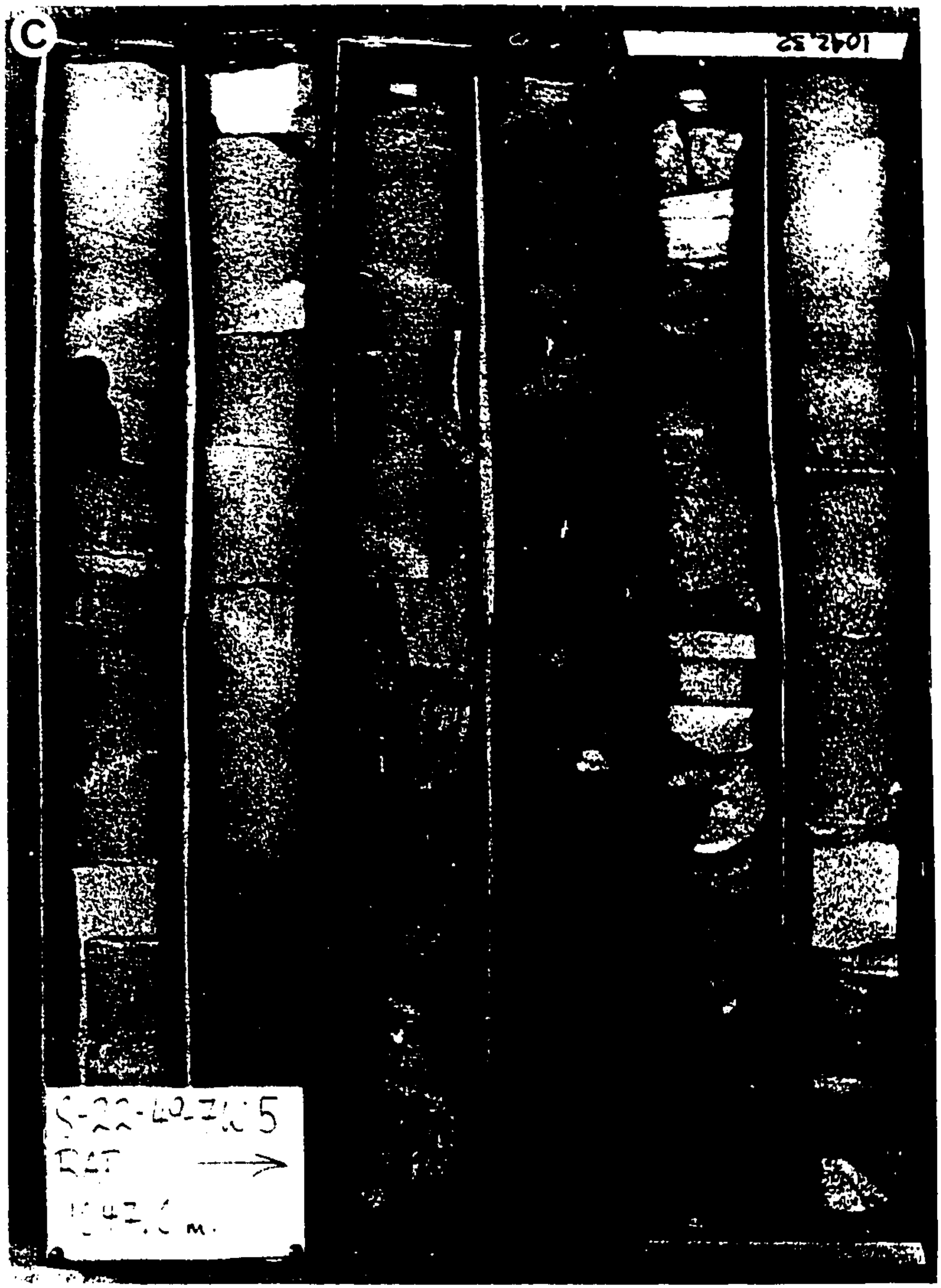
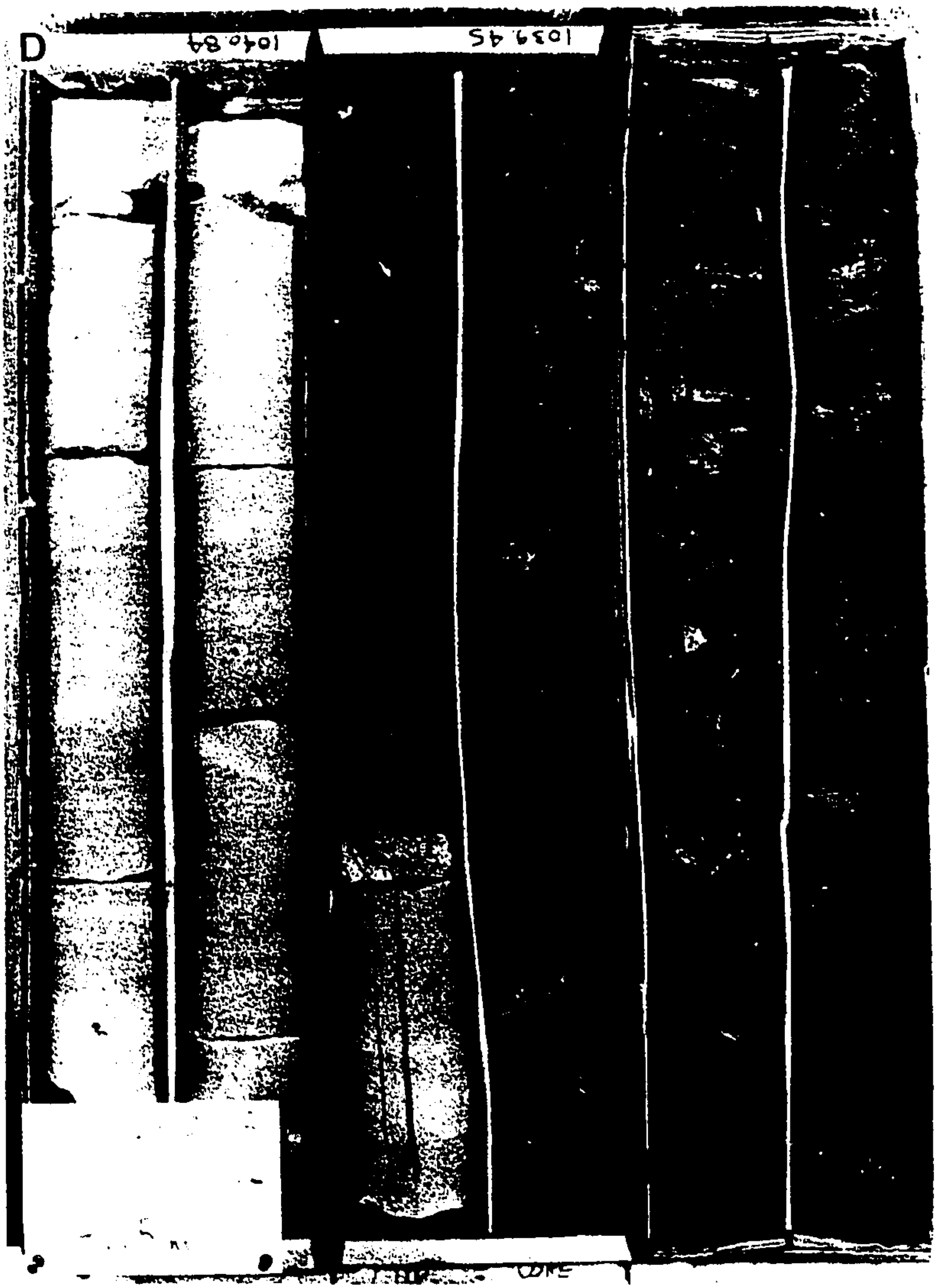

8-22-49-7W5
CYCLE D
Regressive Shoreface Succession
15
10
5
0
metres
M
M
M
O
sh
slt
vf
m
f
c
vc

Figure 6.3: Box photographs of shoreline succession of Cycle D in well 8-22-49-7W5. Core is 3.5 cm in diameter, and each tube of core is 75 cm in length. The base of the succession is at the bottom left corner of each plate, and the top is at the upper right corner. Plate A shows completely structureless sandstone of sub-unit D_1. The arrow in Plate B marks the contact between the massive sandstones of sub-unit D_1 and the cross-bedded sandstones of sub-unit D_2. the top of the shoreline succession in Plate D is flat-laminated and rooted, and is overlain by coal and non-marine sediments.


A
1054.10


B
5-22-49-7W5
1051.5 m.


C

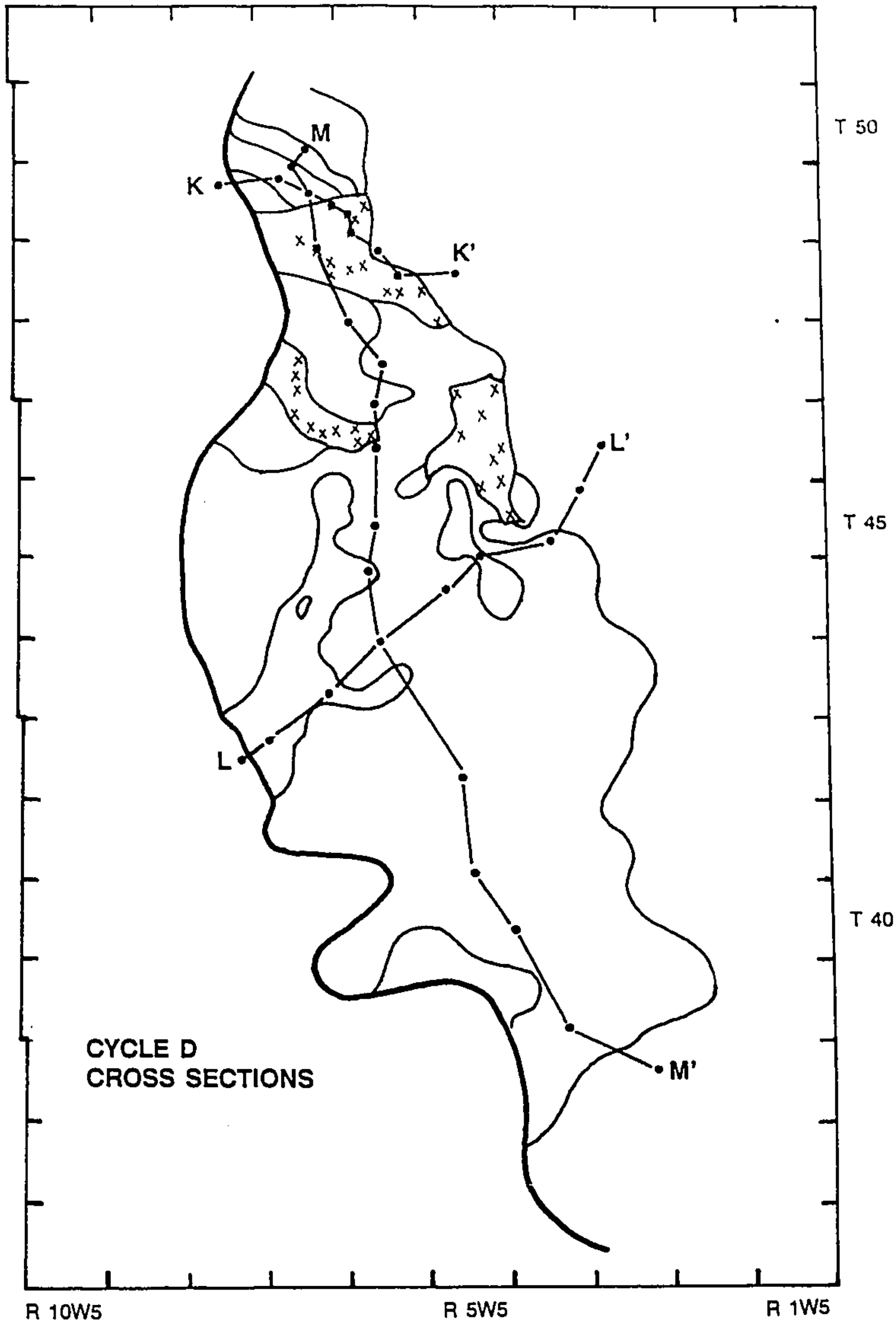

D
1040.84
1039.45

thick unit of flat-laminated/LAIS sandstone, which commonly contains root traces at the top. *Macaronichnus* is the most common trace fossil within the succession, and is usually concentrated within the cross-bedded sandstones or in the higher beds of the underlying massive-to-laminated sandstones. Overall, the facies succession in Cycle D is typical of a prograding shoreline system. The abundance of massive sandstone and lack of well defined beach lamination may indicate that the sediments were not extensively reworked by marine processes such as longshore currents, indicating that the sediments are more indicative of a lobate deltaic shoreline setting rather than a straight shoreline.

6.3: Cross Sections

Figure 6.4 shows the location of detailed cross sections through Cycle D. Figure 6.5 is a log cross section (K-K') oriented parallel to dip and trending through the northern portion of Pembina, where the thickest tongue of Cycle D sandstone occurs. The section is approximately 30 km in length, and is hung using the top of Cycle B as a datum. Figure 6.6 is a core cross section along the same line, and is also hung on the same datum. Just basinward of the preserved landward edge of its deposition (6-29-49-7W5, 8-22-49-7W5), Cycle D is up to 15 m-thick and sits directly on Cycle C, with no transgressive sediments separating the two cycles. The contact between Cycles C and D is unfortunately not contained in either core section, so it is unknown whether or

Figure 6.4: Simplified isolith map of Cycle D showing location of cross sections K-K', L-L', and M-M'. The 4, 8, and 12-m isoliths are shown.



T 50
T 45
T 40
M
K
K'
L'
L
M'
CYCLE D
CROSS SECTIONS
R 10W5
R 5W5
R 1W5

Figure 6.5: Log cross section K-K' oriented parallel to dip and trending through the northern Pembina region. Note the presence of the two sub-units D_1 and D_2 on the log signatures. The boundary between these two units is shown with a dashed line. The section is hung using the top of Cycle B as a datum.

CYCLE D: LOG CROSS SECTION PARALLEL TO DIP - NORTHERN PEMBINA REGION

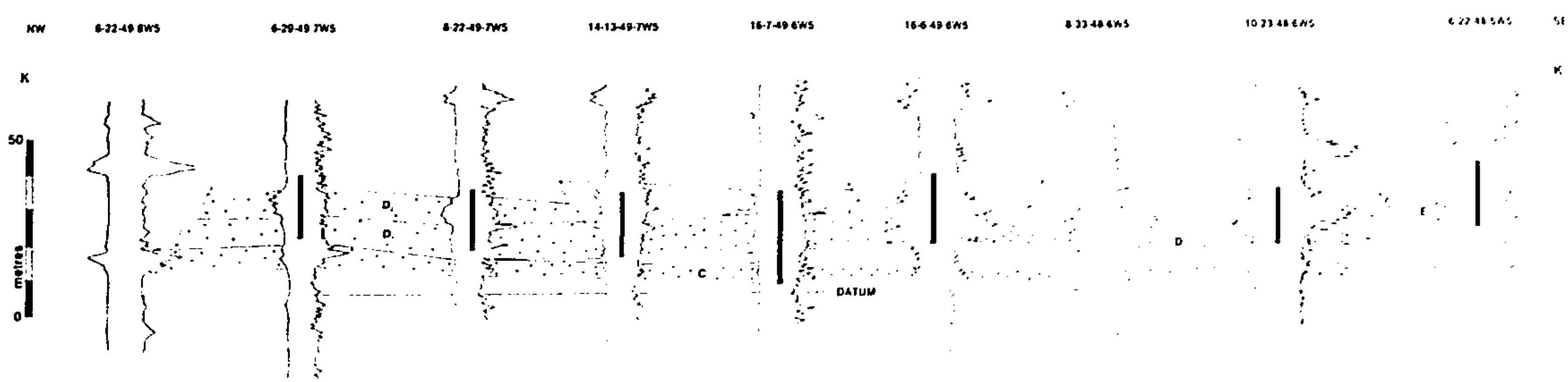
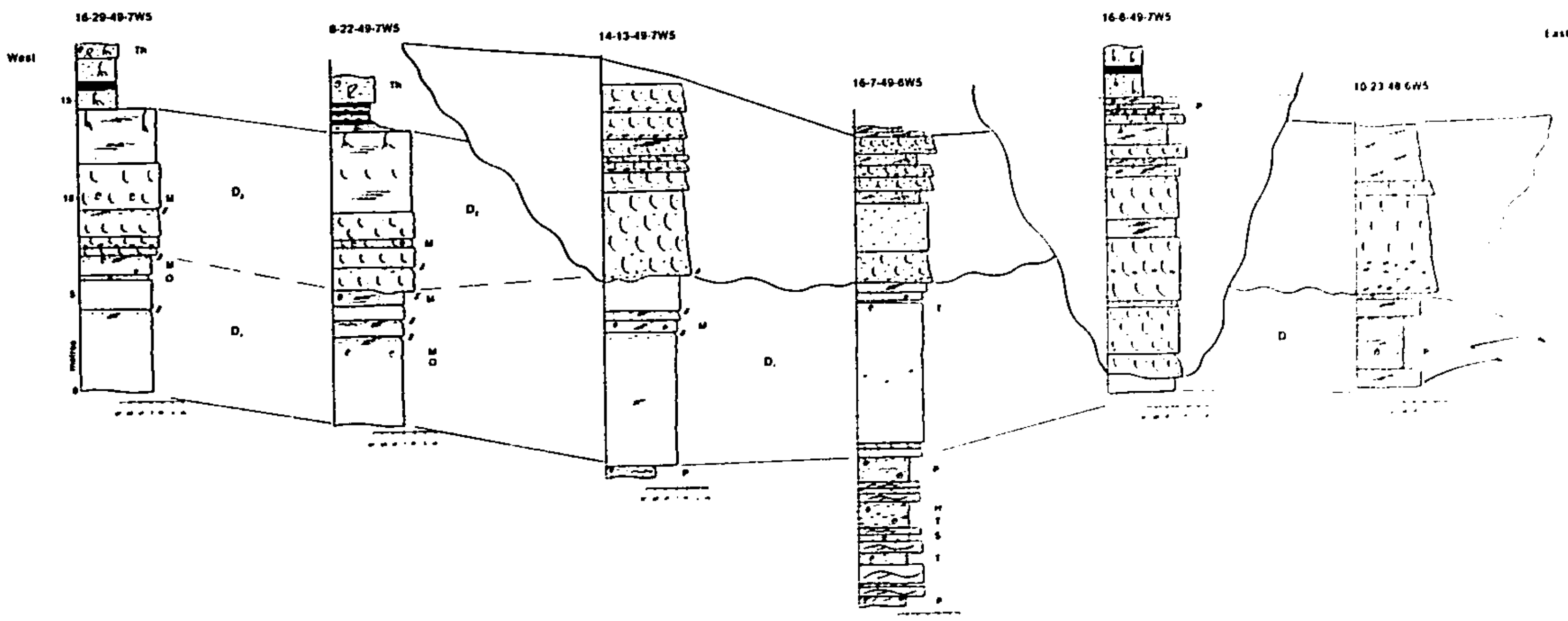


Figure 6.6: Core cross section oriented parallel to dip. The section follows the same line as log section K-K'. Sub-unit D_1 is characterized by the massive-to-laminated beds, whereas sub-unit D_2 is characterized by cross-bedding and flat lamination. Note the removal of sub-unit D_2 in the eastern portion of the section. The HCS and wave-rippled beds beneath Cycle D in well 16-7-49-6W5 are distal storm sediments of Cycle C. The section is hung using the top of Cycle B as a datum.

CYCLE D: CORE CROSS SECTION PARALLEL TO DIP - NORTHERN PEMBINA REGION



not this contact is erosive. Figure 6.5 shows that the log response of Cycle D is composed of two distinct sub-units in this area. These two units are labelled D_1 (lower) and D_2 (upper) respectively. D_1 is characterized by a low amplitude response on both the SP and Dual Induction logs. The overlying D_2 unit shows a much more positive SP response, indicating greater permeability within this unit, although the Dual Induction response is still of low amplitude, indicating that the fluids within the sandstone are mostly water. The contact between the two units corresponds to the contact within the succession between the lower massive sandstones beds and the cross bedded sandstones. Immediately eastward of 8-22-49-7W5, the D_2 unit is removed by fluvial erosion, and stacked fluvial channel sandstones typical of Facies Association 3 sit erosively on the D_1 sandstones in 14-13-49-7W5 and 16-7-49-6W5. In both of these locations, the massive sandstone beds are very thick, reaching almost 8 m in thickness. Cycle D is entirely removed by fluvial erosion at 16-6-49-6W5. Southeast of this point, figure 6.5 indicates that cycle D is thinning and beginning to pinch out. At 10-23-48-6W5, fluvial channel sediments are sitting on distal shoreface or shelf sandstones of Cycle D. A short distance basinward of this location, Cycle D reaches its basinward depositional edge. The time relationship of the channels which cut into Cycle D in this area to the shoreface succession of Cycle D is not known.

The transgressive sediments which overlie Cycle D are not preserved at any location along this section. However, Figure 6.6 shows that in 16-29-49-

7W5 and 8-22-49-7W5 there is a thin unit of sandy mudstone containing *Thalassinoides* burrows situated 2-3 m above the top of Cycle D, with the coastal plain sediments of Facies Association 5. *Thalassinoides* is generally considered to be a marine trace fossil, and its presence in this location may indicate that this unit represents a minor flooding of the coastal plain with saline marine waters.

Figure 6.7 is another log cross section (L-L') oriented parallel to dip, located south of the Pembina field. Cycle D in this section is simpler internally than section K-K', and consists of a single coarsening-upward succession up to 8-10 m thick which is present for 40-50 km basinward of its landward edge of deposition. The transgressive sediments which overlie Cycle D are removed by fluvial erosion southwest of 2-20-44-5W5, leaving a similar stratigraphic pattern as in the regional cross sections of Chapter 4. Figure 6.8 is log cross section M-M', which is oriented in a strike-parallel direction (Fig. 6.4). The composite nature of Cycle D can be seen more clearly in this section. The presence of the two sub-units of Cycle D in the northern half of the section is clearly indicated on both the SP and Dual Induction logs. At 16-29-44-6W5 D_1 and D_2 have begun to separate stratigraphically, and a short distance south of this location, D_1 pinches out into shelf sediments. Sub-unit D_2 is present in all wells of the section except at the far southern end where it has also pinched out into shelf sediments. The transgressive sediments overlying Cycle D are only present in this section at the far southern end where Cycle D has thinned and begun to

Figure 6.7: Log cross section L-L' oriented parallel to dip and trending through the southern Pembina region. The section is hung using the top of Cycle B as a datum.

CYCLE D: LOG CROSS SECTION PARALLEL TO DIP - SOUTHERN PEMBINA REGION

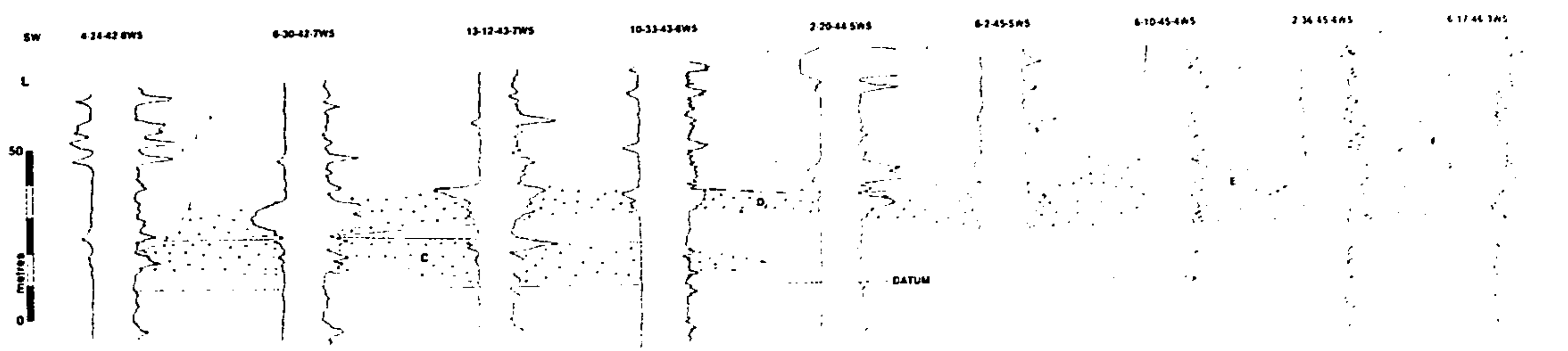
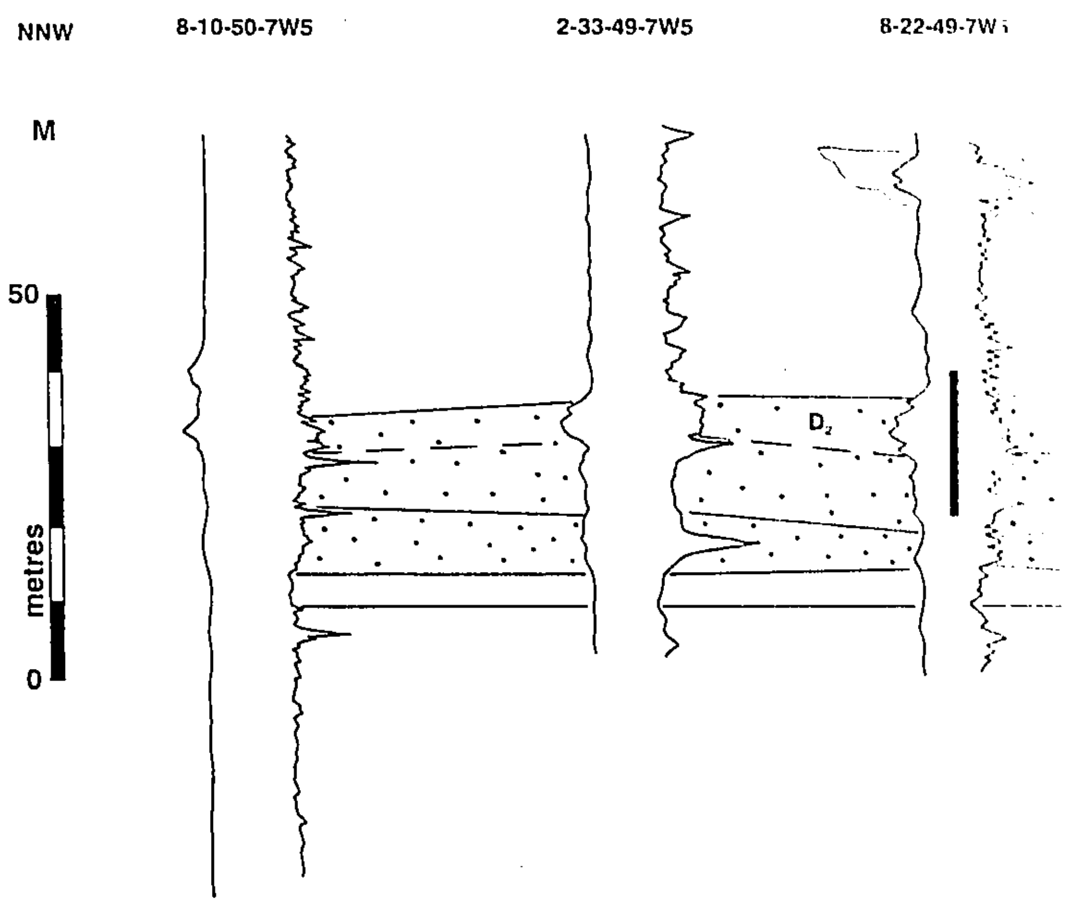
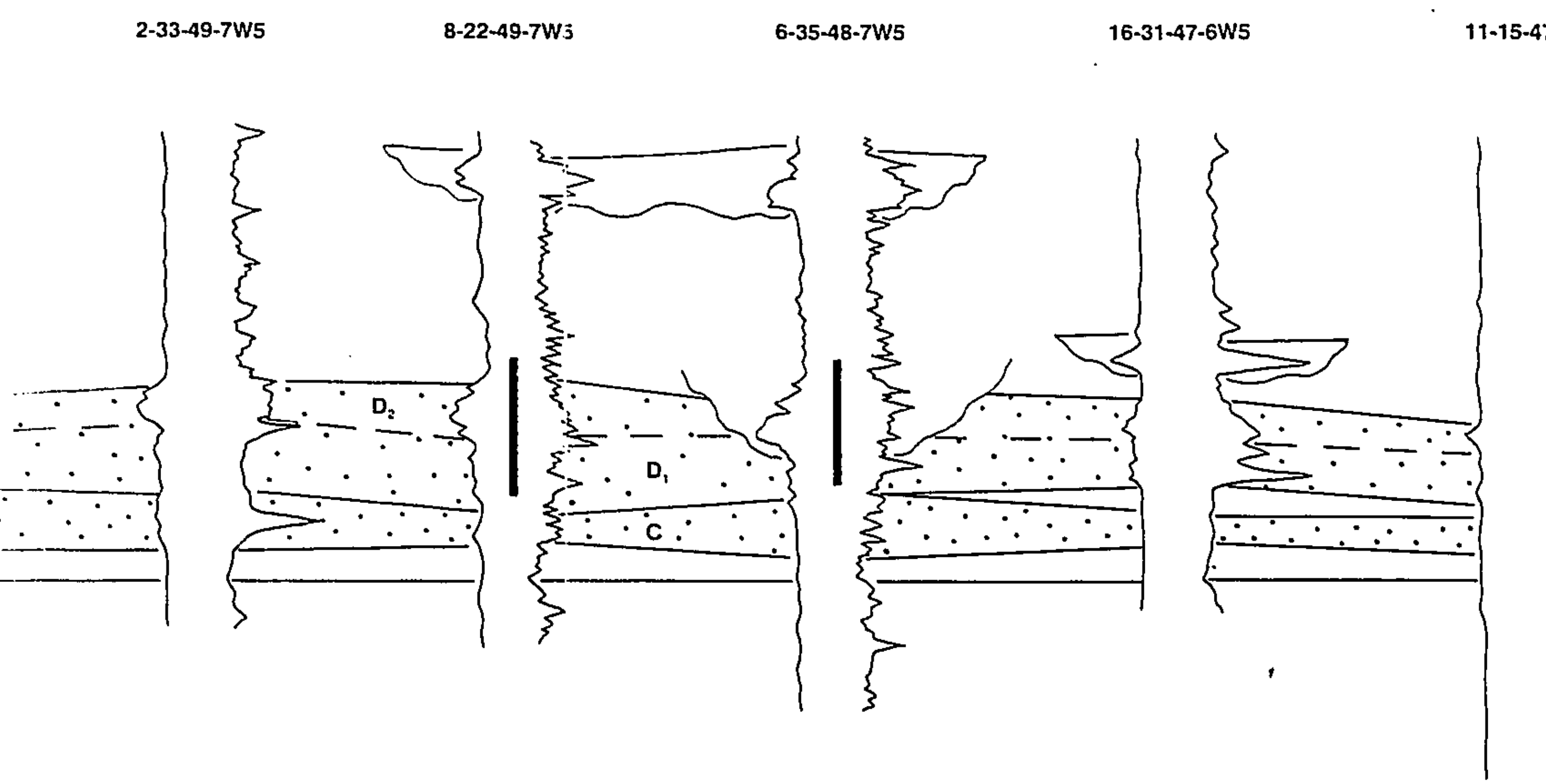
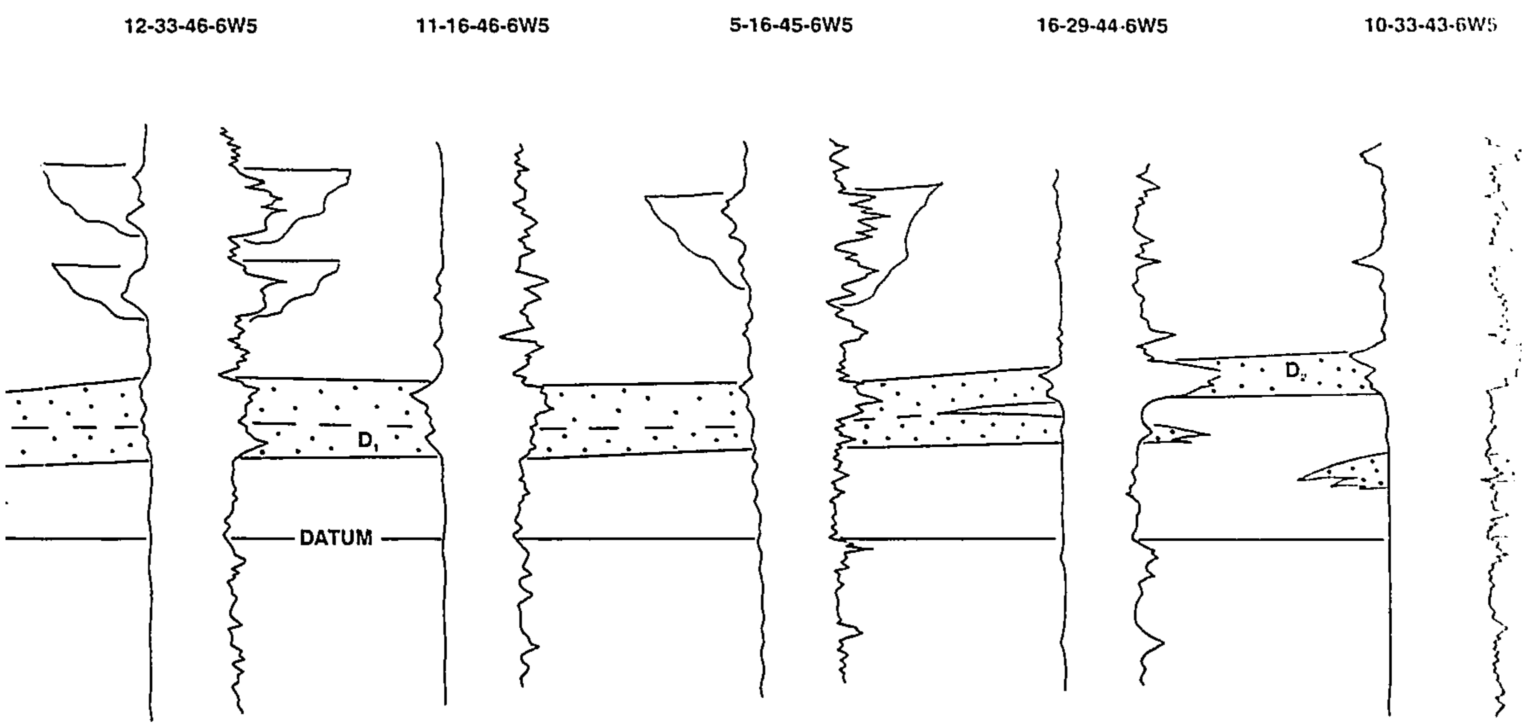


Figure 6.8: Log cross section M-M' oriented parallel to strike of Cycle D. Note the presence of the two sub-units in the northern two-thirds of the section. The section is hung using the top of Cycle B as a datum.

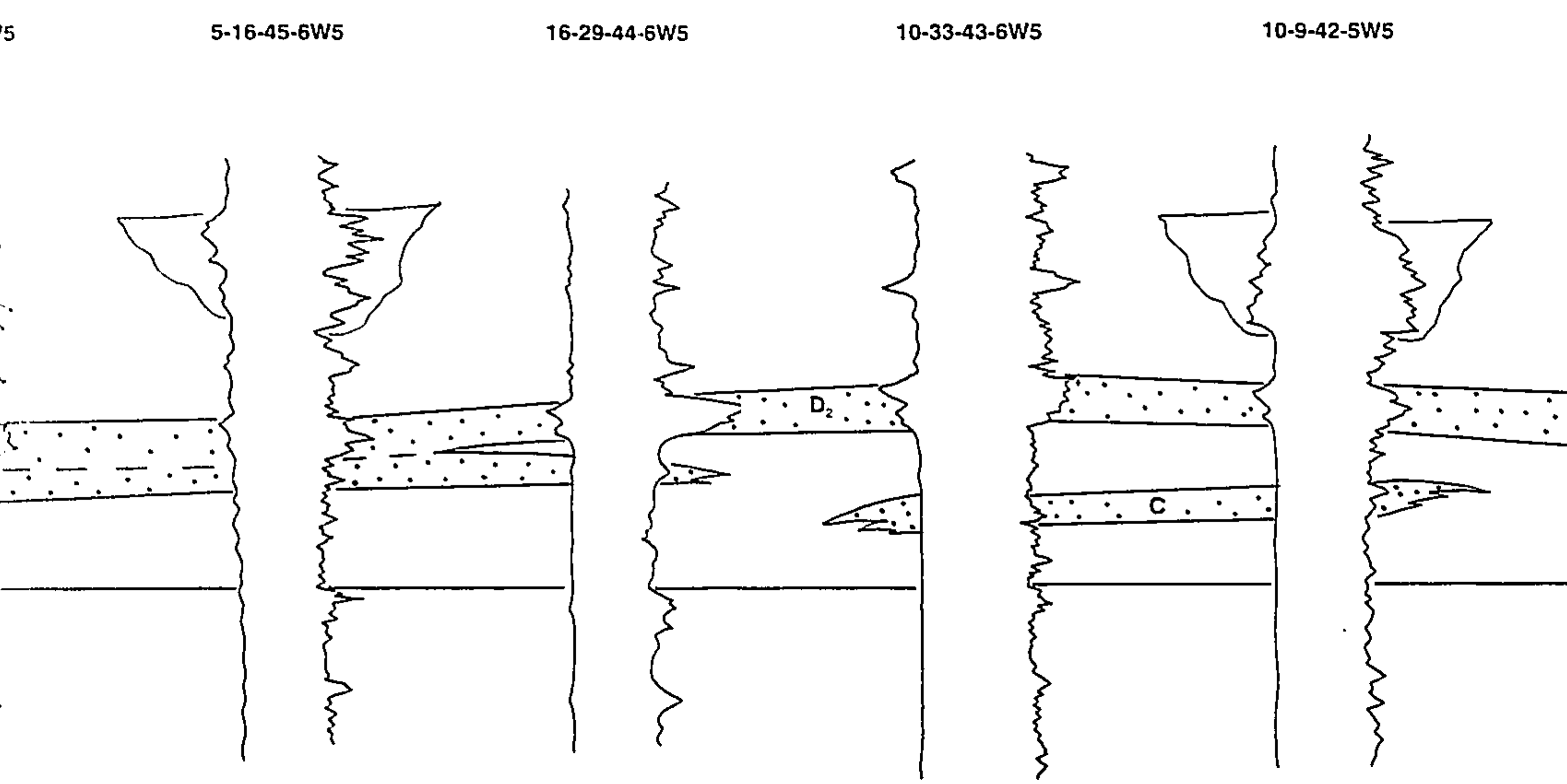

NNW
8-10-50-7W5
2-33-49-7W5
8-22-49-7W5
M
50
0
metres
D2

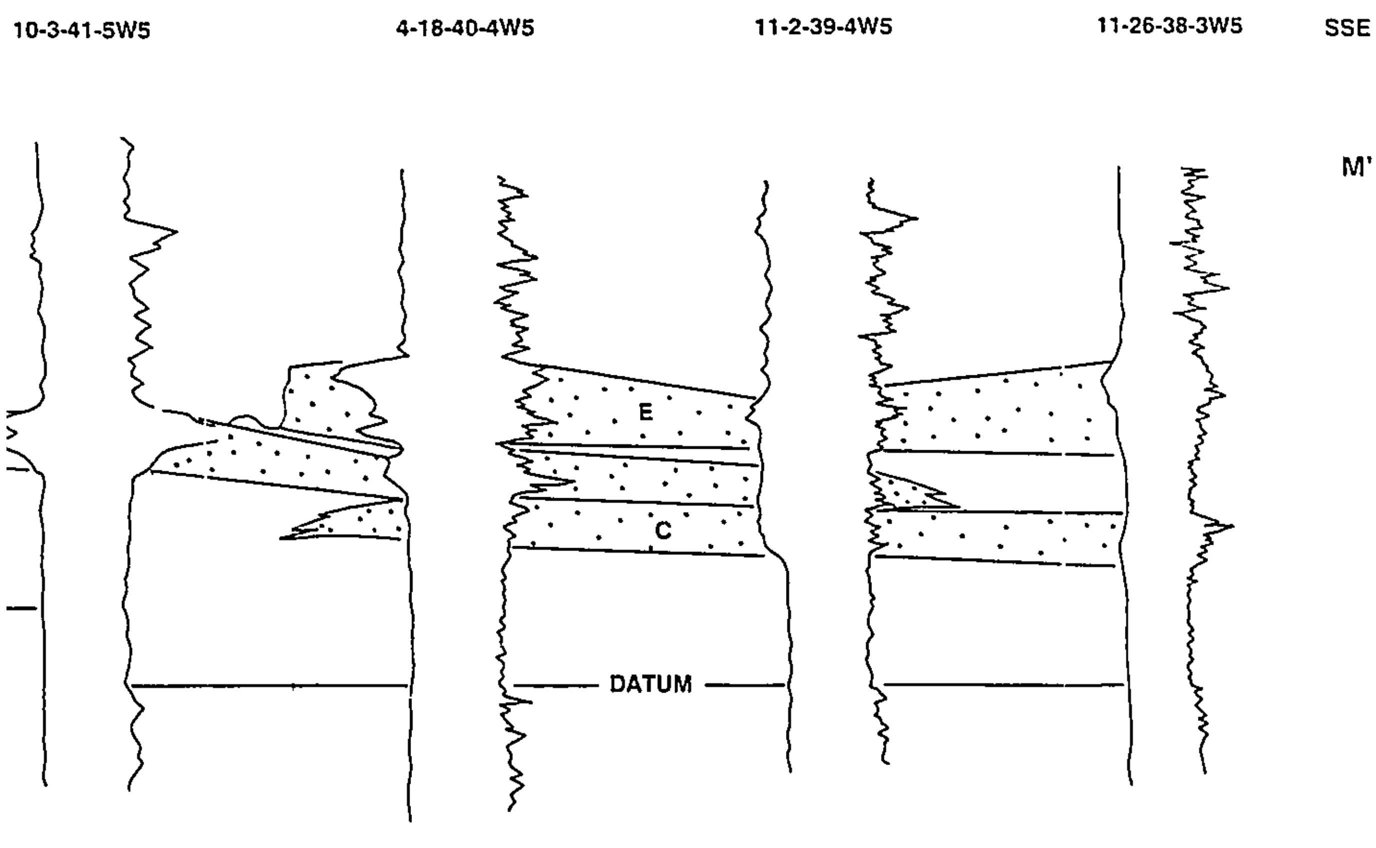

2-33-49-7W5
8-22-49-7W5
6-35-48-7W5
16-31-47-6W5
11-15-47
D2
D1
C

CYCLE D: LOG CROSS SECTION PARALLEL TO STRIKE



ION PARALLEL TO STRIKE




10-3-41-5W5
4-18-40-4W5
11-2-39-4W5
11-26-38-3W5
SSE
M'
E
C
DATUM

pinch out. Further north of 4-18-40-4W5 these sediments are eroded or were never deposited.

6.4: Interpretation

Cycle D is interpreted to contain the deposits of a prograding, deltaic shoreline environment. The plan view geometry of the cycle indicates that the system was characterized by localized "protrusions" of thick sandstone. These protrusions are interpreted to represent locations where the shoreline underwent active deltaic progradation due to localized input from fluvial sources. The elongate "tongue" shape or lobate shape of these protrusions indicates that the proximal deltaic sediments were not widely reworked by basinal processes, and as such the deltaic system might be classified as fluvially-dominated. The widespread lobate areas of sandstone surrounding the proximal lobes indicate, however, that some reworking and redistribution of sediment by basinal processes may have occurred. However, the abundance of massive-to-laminated sandstone beds within the lower portions of Cycle D is not indicative of significant marine reworking, and is somewhat problematic to interpret in a deltaic environment. These problems, along with other aspects of the sedimentological nature of the delta sediments will be discussed in greater depth in Chapter 11.

Cycle D is internally composed of two sub-units, which are interpreted to have been deposited as separate sand bodies due to at least two and possibly three different episodes of delta progradation. The lowermost D_1 unit is present only in the northern half of the study area. The location of its source of input is not precisely known, but may have been associated with the > 8 m thick lobate-shaped sandstone body in northern Pembina.

The younger D_2 sub-unit is present wherever Cycle D is present. The sediments of this unit may have been deposited at two deltaic depocenters within the study area. One may be associated with the SW-NE trending tongue of sediment near Ferrier - Willesden Green, and the other with the NW-SE trending tongue in northern Pembina. The latter is re-occupying the depocenter of sub-unit D_1. Therefore it appears that the thickest sandstone deposits of Cycle D may be due to the stacking of the deposits of two different deltas lobes, which occupied the same depocenter at different times. The 8 m isolith outlines the deposits of the older lobe, while the 10-14 m-thick tongue reflects the deposits of the younger lobe. The more southerly tongue of sediment may have been deposited between the two times of delta building at the northern depocenter.

The sediments of the D_2 lobe are noticeably coarser than the deposits of the D_1 lobe, and are more characterized by cross-bedding as opposed to massive-to-laminated beds. This may reflect a closer proximity to the source of sediment. The coarser grain size is more likely to be transported as bedload, and thus form bedforms such as dunes which will be preserved as cross-bedded sandstone. The shore-normal plan view geometry of these deposits does not suggest that these cross-beds could have been created by basinal processes such as those interpreted to have created the cross-bedded shoreface sediments of Cycle C, which were oriented sub-parallel to the paleoshoreline. These sediments may also represent the deposits of channels which cut into the massive-to-laminated beds as the shoreline prograded basinward. It may be rather difficult, given the appropriate depositional conditions, to distinguish between sediments deposited in distal portions of distributaries, and between sediments deposited immediately seaward of the distributary mouth. The cross-bedded sandstones in this succession are therefore interpreted to be the preserved record of dunes which were formed by currents with either fluvial origin in an open marine environment proximal to the delta mouth. In a sense they are similar to delta mouth-bar deposits, although the size and geometry of the thick tongue of sandstone would indicate that these mouth bars are present over tens of kilometres. They may be more generally and perhaps more properly referred to simply as shallow delta front sediments. The sedimentolgoical nature of all deltas in the Lea Park - Belly River transition will be discussed in further detail in Chapter 11.

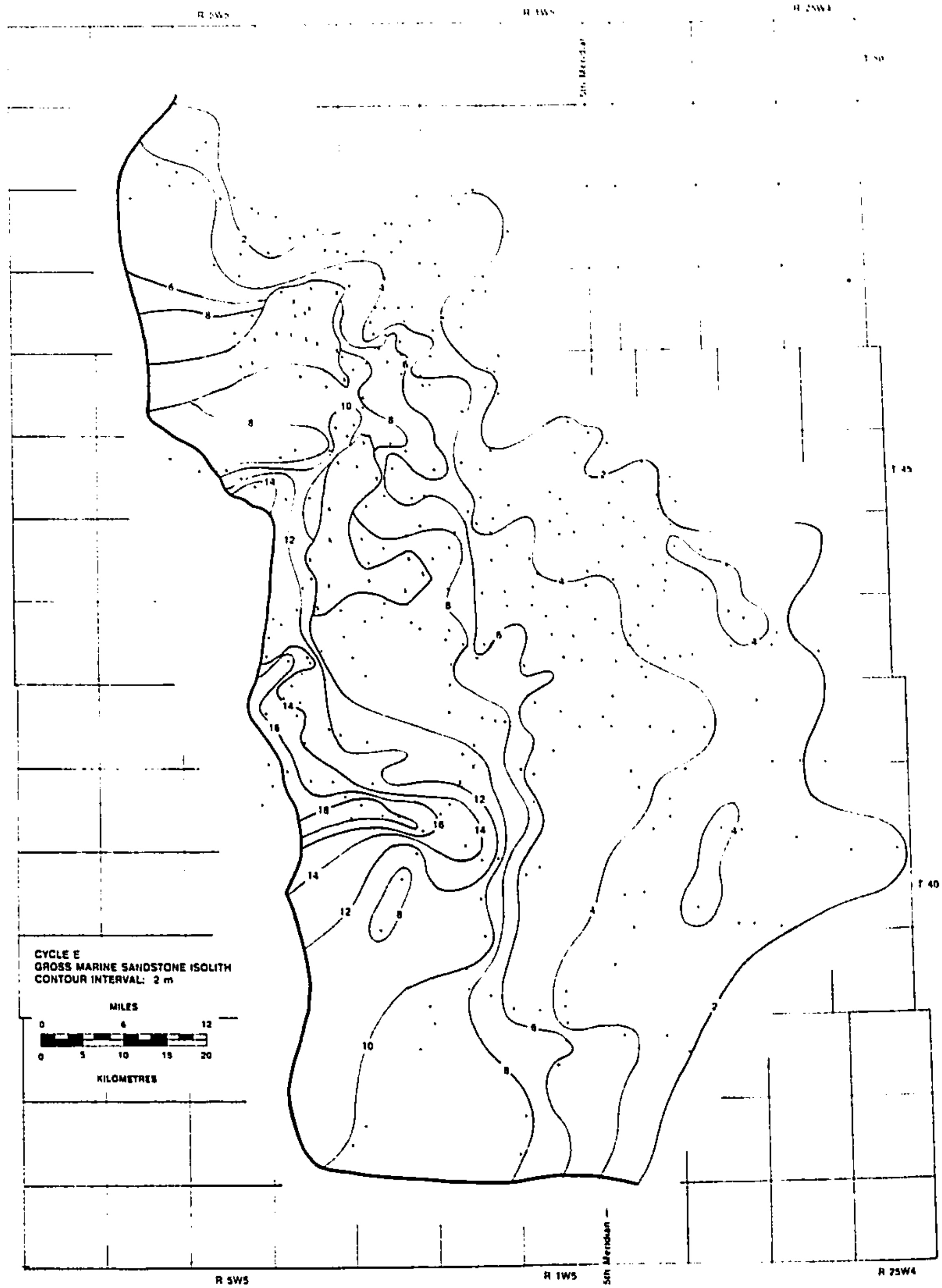
The transgressive sediments which overlie Cycle D are rarely preserved, except in regions of distal Cycle D deposits. The coastal-plain equivalent of the transgressive unit which overlies Cycle D may be preserved as the thin *Thalassinoides*-burrowed mudstone unit which sits on coal beds overlying the top of Cycle D. This unit represents a minor flooding of the coastal plain with marine waters, and sits at approximately the same stratigraphic level as the transgressive unit overlying Cycle D. It is likely that the *Thalassinoides*-burrowed beds are the most landward deposits of the transgressive unit. Their position immediately overlying coal beds may also support this interpretation. McCabe (1984) states that peat deposits (which upon burial will become coal) are difficult to erode, and therefore basinal processes associated with transgression may not erode the peat deposits, but instead will deposit transgressive sediments on top of them.

CHAPTER 7: CYCLE E

7.1: Introduction, Distribution, and Geometry

Cycle E can be found throughout much of the study area (Fig. 7.1). It is present in the eastern portions of the Pembina field, throughout the Keystone field to south of Wilson Creek, and east of the fifth meridian. The general plan view geometry is similar to that of Cycle D. The large-scale geometry is broadly lobate, striking NW-SE and dipping to the northeast. The basinward extent of deposition ranges from 30-70 km along strike. A distinct protrusion of thick sandstone is present within and to the south of the Wilson Creek Field (Townships 42 and 43, Ranges 2-4W5). The protrusion consists of one dominant tongue elongated in an east-west direction joined to a smaller tongue 10 km to the north oriented SW-NE. Cycle E is thickest within the larger tongue, reaching a maximum of 19 m. A third, very small protrusion is evident in the vicinity of the southwest corner of township 45-4W5, immediately basinward of the preserved landward edge of deposition. Cycle E reaches 15 m in thickness at this location. The three protrusions together form a thin elongate/lobate area 50 km long in a strike direction which is greater than 12 m thick in all locations. The 2-10 m isoliths broadly surround this thick central area. Large areas of fluvial incision have removed Cycle E in eastern Pembina and immediately south of Keystone. The abrupt boundaries of these areas of fluvial erosion are similar to those which removed portions of Cycle D.

Figure 7.1: Gross isolith map of marine sandstone in Cycle E. Contours are in metres, and the contour interval is 2 m. Heavy black line represents the preserved landward edge of marine deposition.



CYCLE E
GROSS MARINE SANDSTONE ISOLITH
CONTOUR INTERVAL: 2 m
MILES
0 6 12
0 5 10 15 20
KILOMETRES
T 45
T 40
R 5W5
R 1W5
5th Meridian
R 25W4

7.2: Facies Associations

Although Cycle E is present over much of the study area, there are relatively few cores which contain complete successions through this interval. Some are present in the western and southern portions of Keystone, and several more are located within the Wilson Creek Field. One of the latter is located at 6-10-43-4W5 (Figs. 7.2, 7.3). Cycle E consists of a 15 m-thick coarsening-upward succession of marine sandstones that is very similar to the succession contained in Cycle D.

The shoreface succession is underlain by a thin transition zone less than 2 m-thick which contains beds of very fine-grained HCS or LAIS sandstone up to 60 cm thick. The shoreface sediments sharply overlie this transition zone. The lower half of the shoreface succession consists of beds of dominantly massive sandstone up to 2-3 m-thick. These beds grade up into LAIS or cross-bedded sandstone towards their tops. This lower zone of dominantly massive sandstone is overlain by 3-4 m of dominantly cross-bedded sandstone. The transition is accompanied by an upwards coarsening from lower fine-grained sandstone to upper fine-grained sandstone. In the Keystone region, this cross-bedded zone is often absent. The cross-bedded sandstone is overlain by 1-2 m of massive bioturbated sandstone which in turn is overlain by 2-3 m of poorly stratified flat-bedded sandstone. *Macaronichnus* is the most common trace fossil. It is present in the upper beds of the massive-to-laminated beds and

Figure 7.2: Stratigraphic section through the shoreline succession of Cycle E in well 6-10-43-4W5.

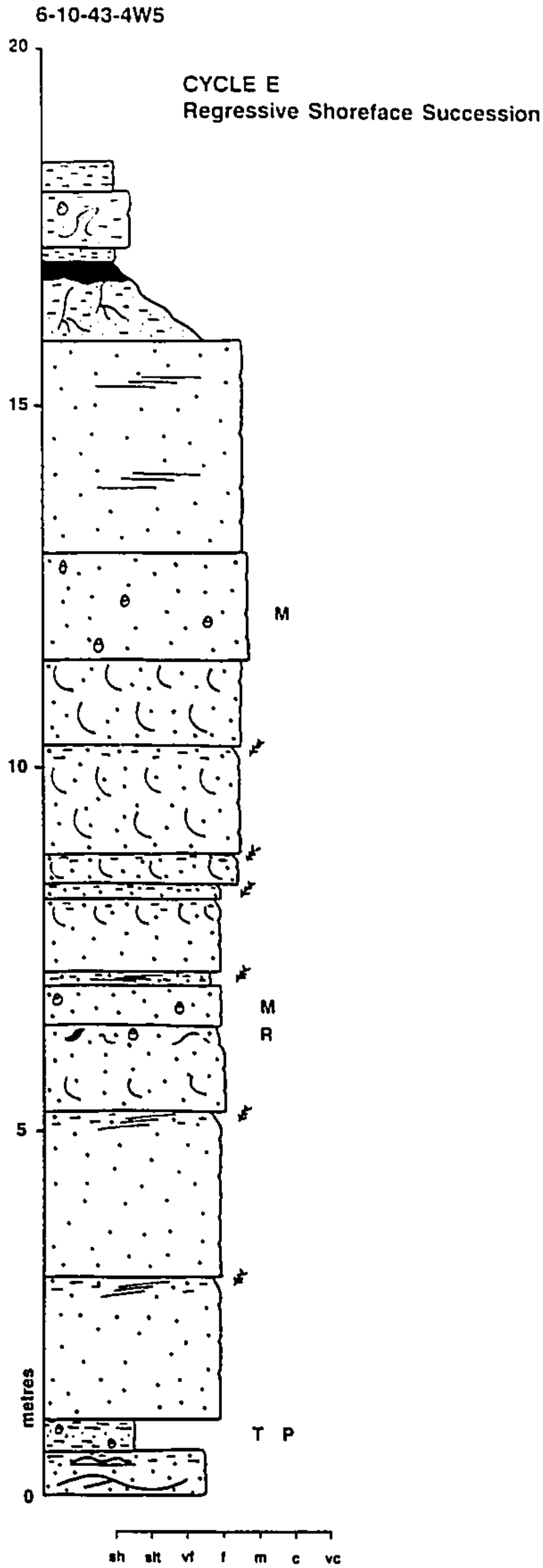
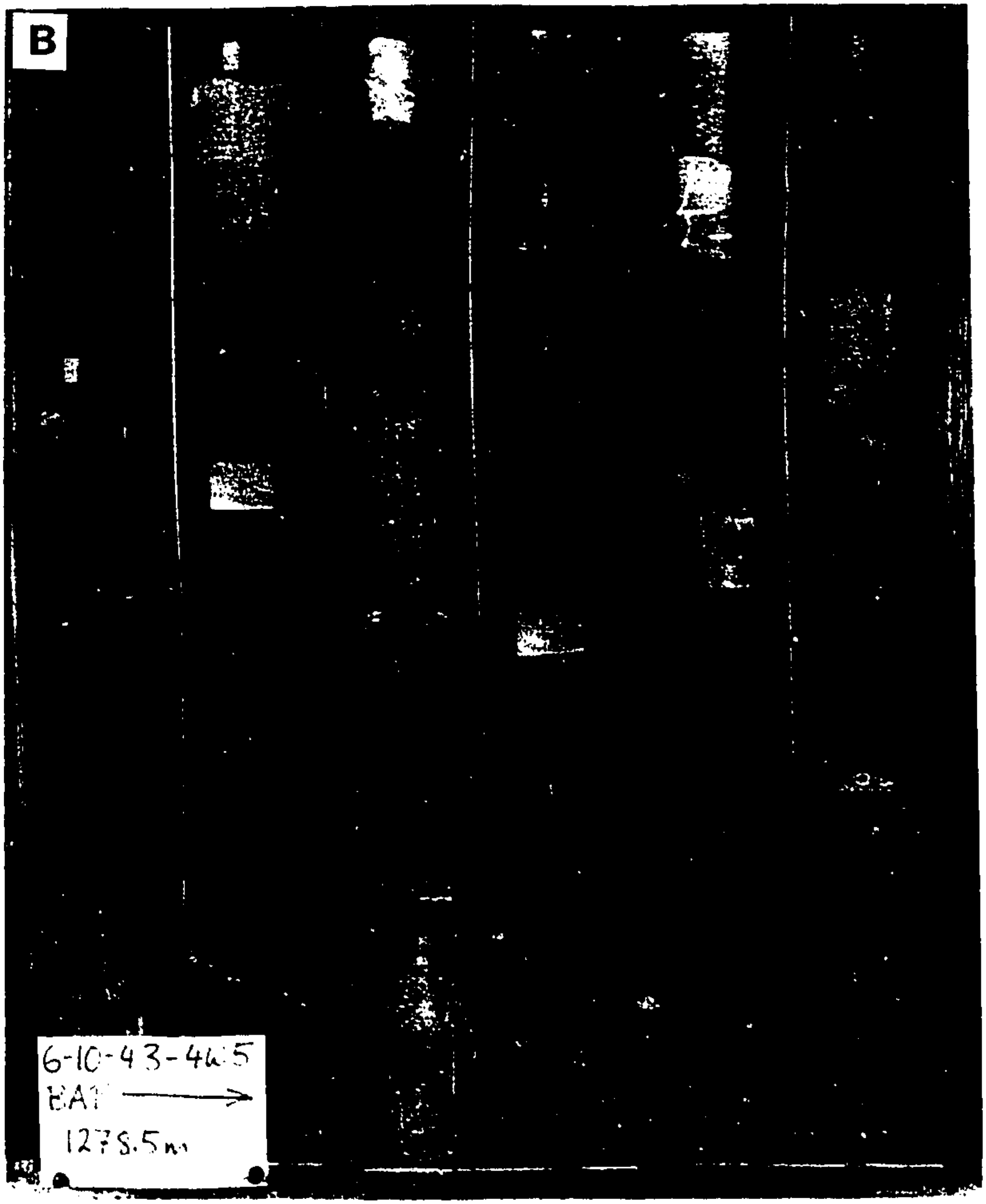
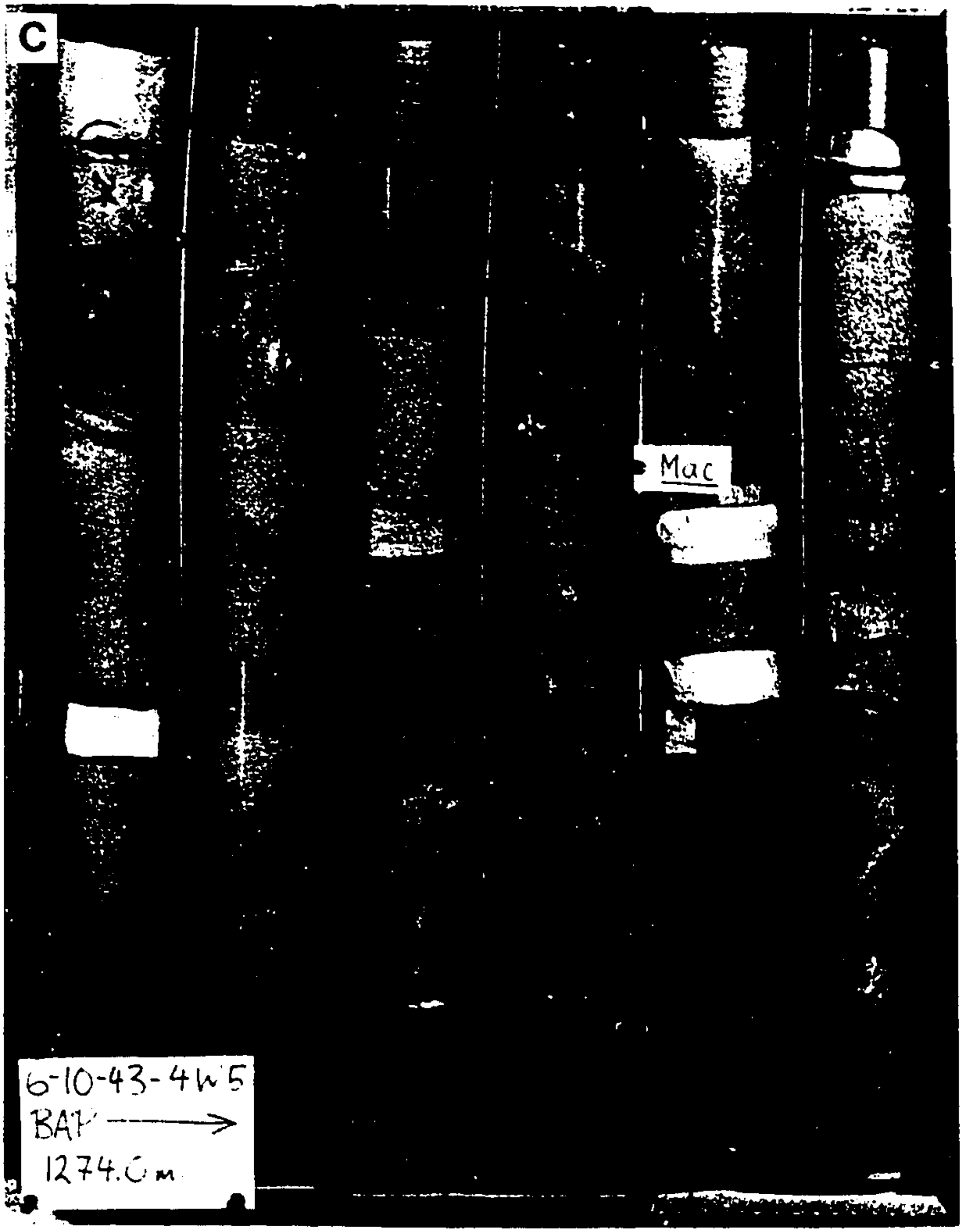
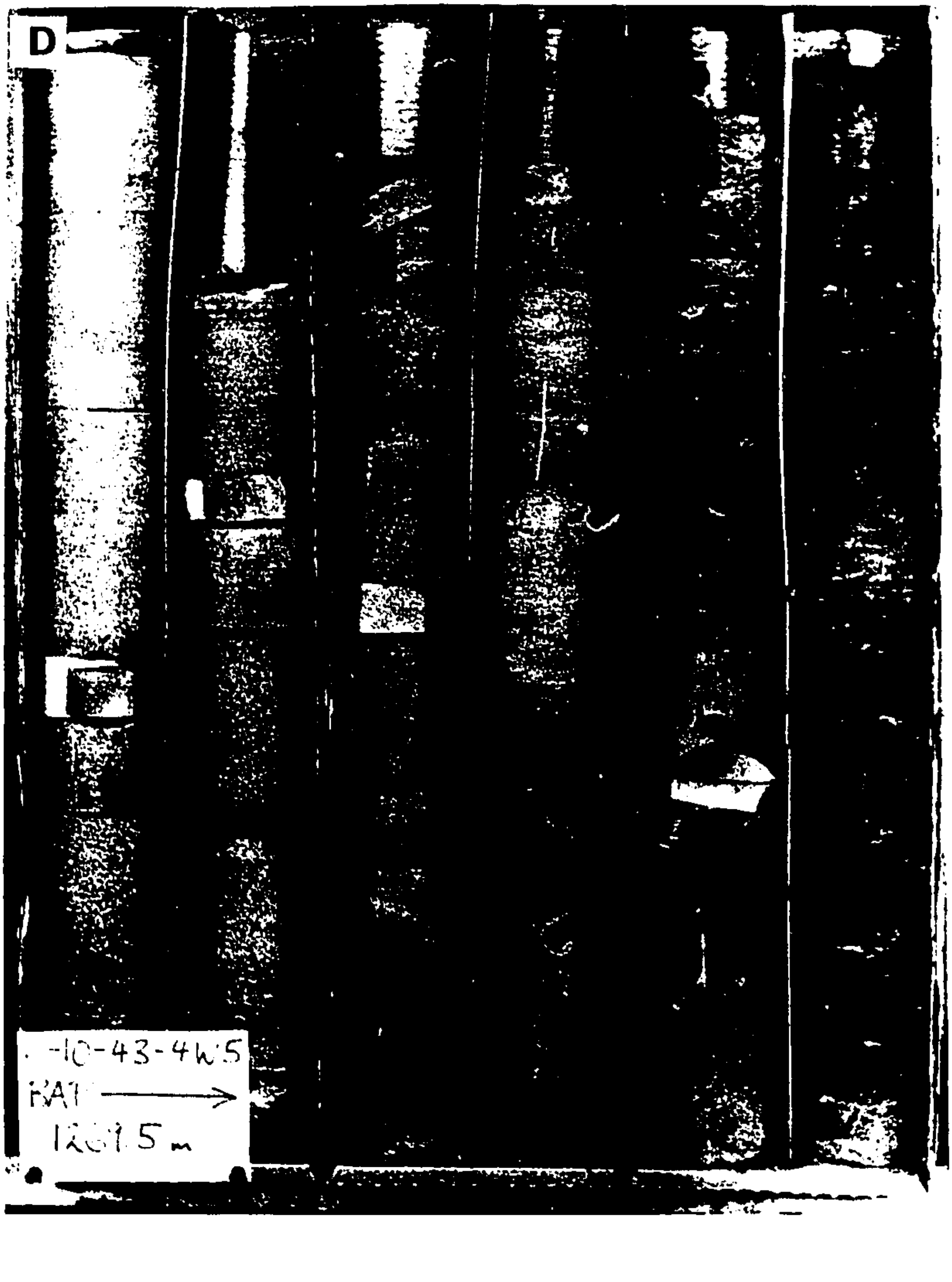

6-10-43-4W5
CYCLE E
Regressive Shoreface Succession
20
15
10
5
0
metres
M
M
R
T P
sh slt vf f m c vc

Figure 7.3: Box photographs of the shoreline succession of Cycle E in well 6-10-43-4W5. Core is 3.5 cm in diameter and each tube of core is 75 cm in length. The base of the succession is at the bottom left corner of each plate, and the top is at the to right corner. The first arrow in Plate A marks the base of the cycle. The second arrow in Plate A marks the top of the first massive-to-laminated bed, which is about 2 m thick. Plate B shows several massive-to-laminated beds. Plate C contains cross-bedded sandstone near the base, which grade upwards into LAIS sandstones (Mac tag denotes Macaronichnus burrows). Plate D shows the vaguely flat-laminated top of the succession, which grades into non-marine coastal plain sediments at the top.


A
6-10


B
6-10-43-4W5
BAT
1278.5m


C
Mac
6-10-43-4W5
BAP
1274.0m

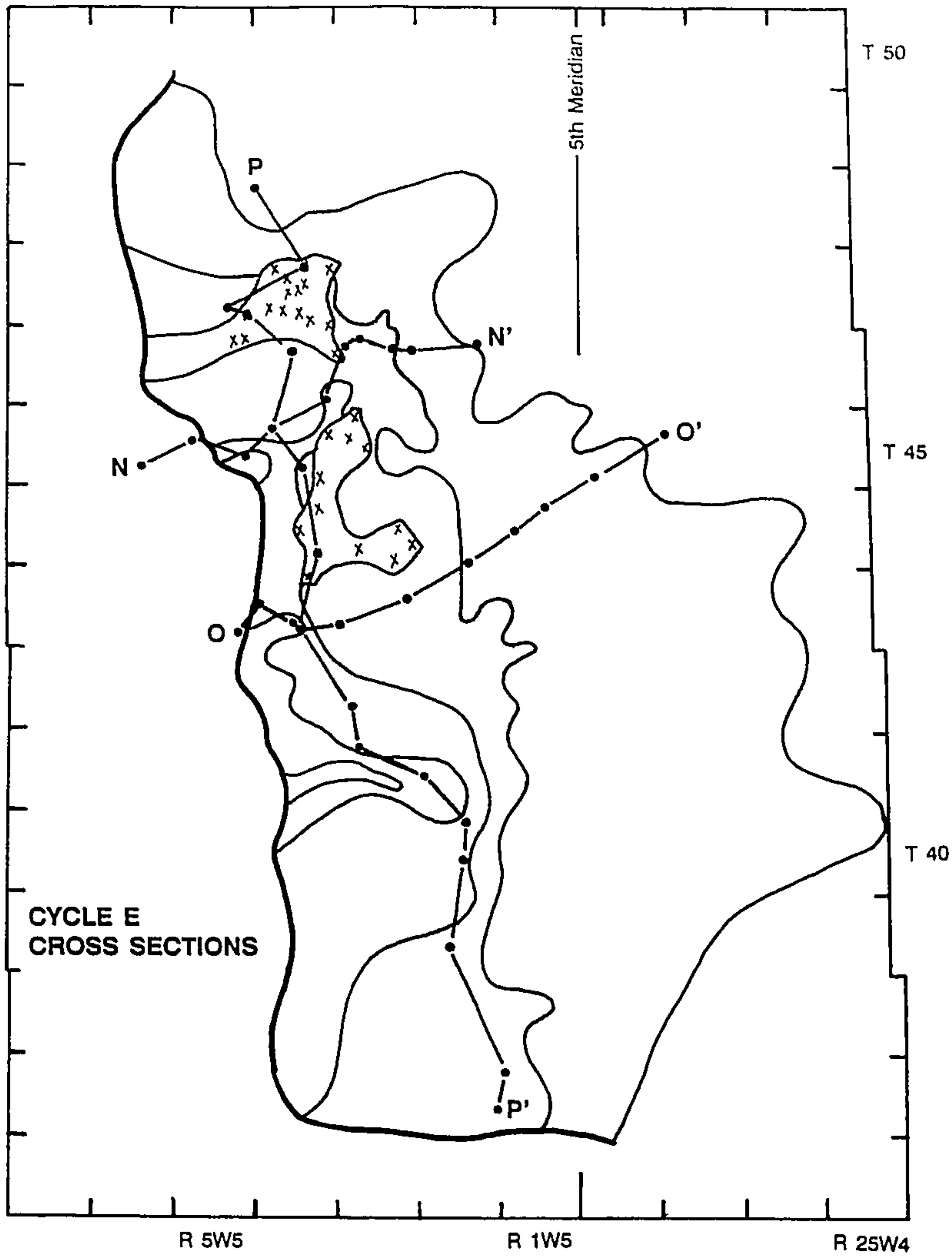

D
10-43-4W5

within the upper bed of structureless sandstone above the cross-bedded sandstone. The shoreface succession in Cycle E is typical of Facies Association 2a, and indicative of a prograding shoreline environment which may be deltaic (non-straight), or in close proximity to a nearby fluvial source.

7.3: Cross Sections

Figure 7.4 shows the location of detailed cross sections through Cycle E. Cross section N-N' (Fig. 7.5) trends parallel to dip through the small northernmost thick protrusion of sandstone basinward into the Keystone region. The section is approximately 40 km in length. Figure 7.6 is a core cross section which trends along the same line, with a proximal core section added on at the beginning. Cycle E begins immediately basinward of the preserved landward depositional edge as a 10-14 m-thick coarsening-upward succession, which appears to be sharply-based and resting directly on the distal deposits of Cycle D. The succession at 6-11-47-5W5 (Fig. 7.6) is fully developed with evidence that the shoreline became subaereally exposed at this location. The succession thins gradually to the northeast. At 10-30-46-3W5, the upper section of Cycle E appears to have been removed and the base of the overlying Cycle F sits directly on Cycle E. Two kilometres further basinward, a thin unit of transgressive deposits separates the two cycles. Cycle E is still 8-9 m thick at this location, but does not contain any evidence of subaerial exposure. The uppermost sediments would appear to be subaqueous upper

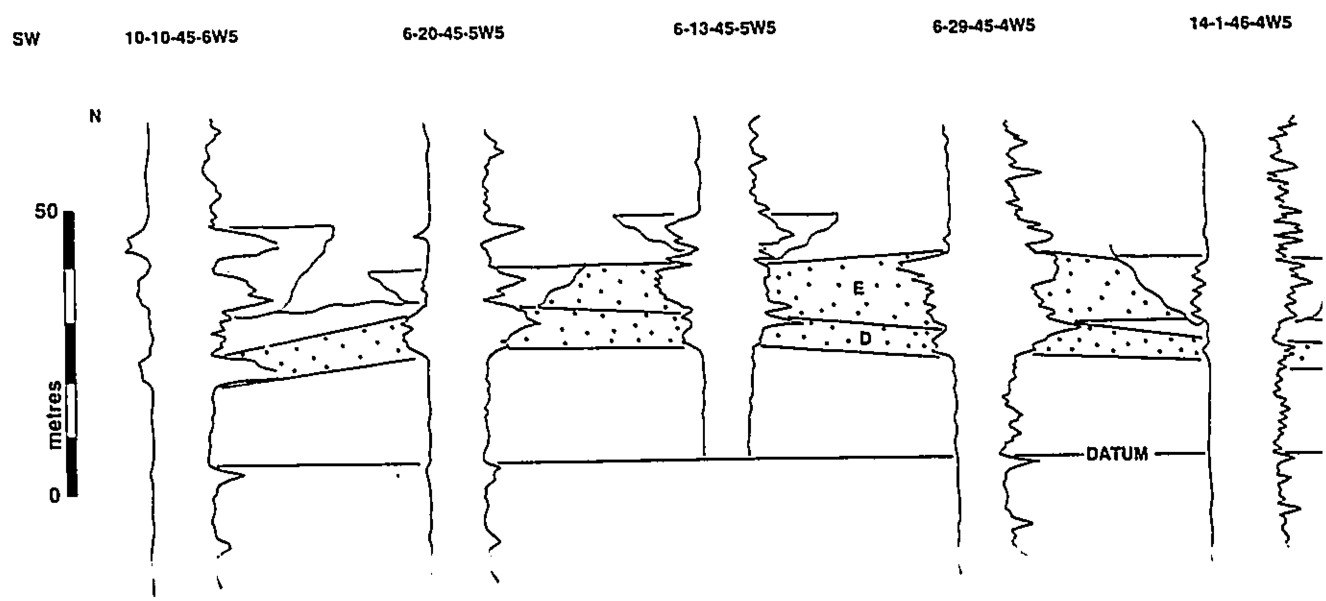
Figure 7.4: Simplified isolith map of Cycle E showing the location of cross sections N-N', O-O', and P-P'. The 2, 6, 10, 14, and 18-m contours are shown.



5th Meridian
T 50
T 45
T 40
P
N'
N
O'
O
P'
CYCLE E
CROSS SECTIONS
R 5W5
R 1W5
R 25W4

Figure 7.5: Log cross section N-N' oriented parallel to dip and trending through the Keystone region.

CYCLE E: LOG CROSS SEC



ːTION PARALLEL TO DIP - KEYSTONE REGION

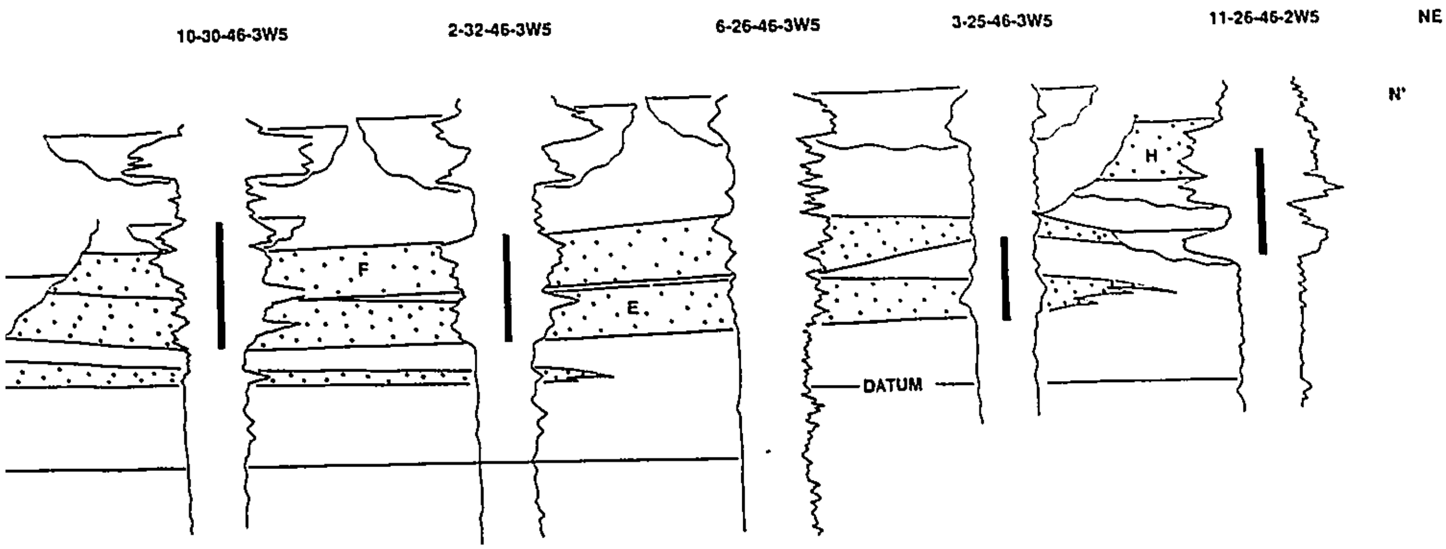
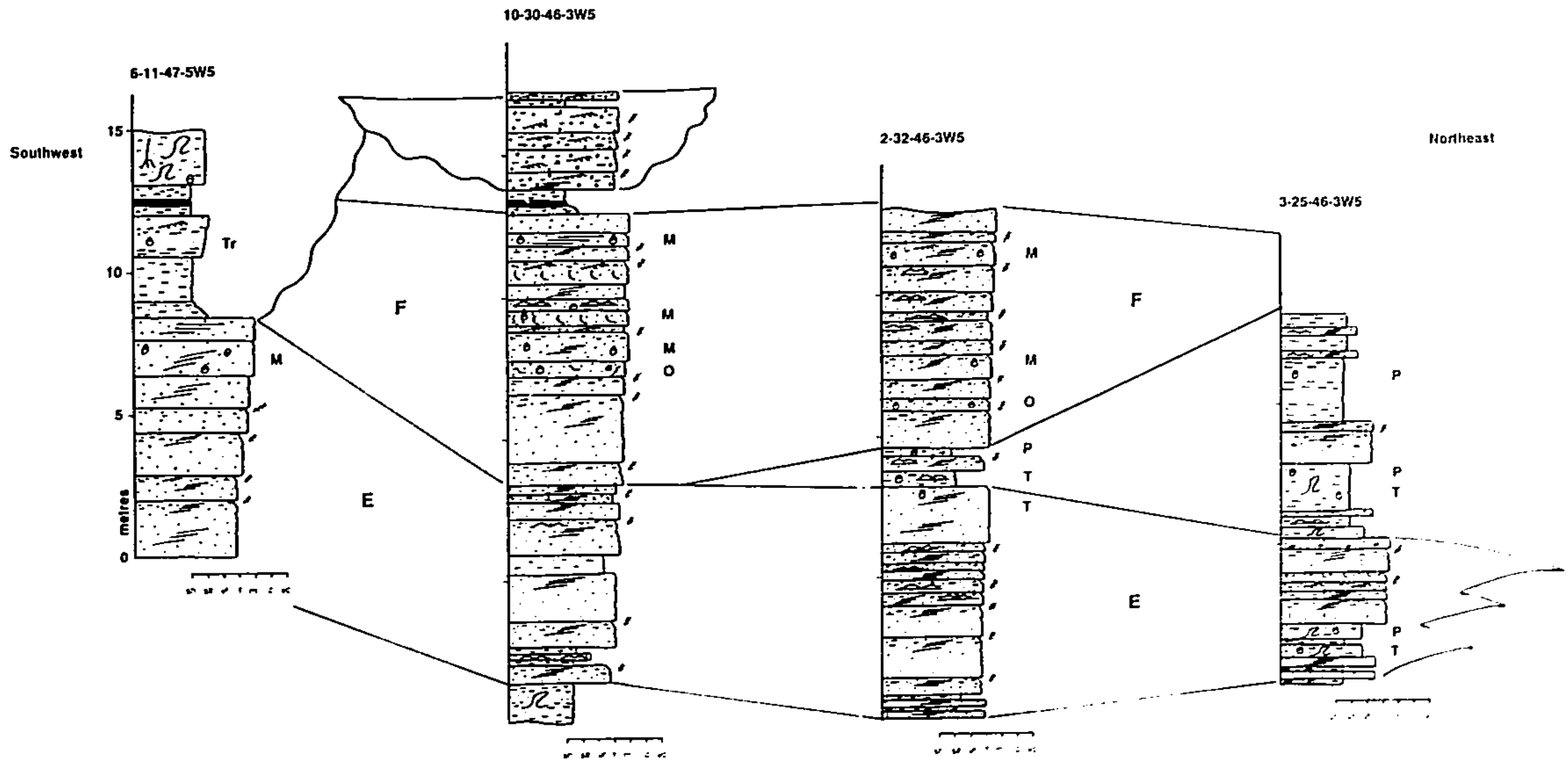


Figure 7.6: Core cross section oriented parallel to dip and trending through the Keystone region. The section follows the same line as log section N-N', except for well 6-11-47-5W5, which is added on at the western end of the core section. Note the stacking of shoreline successions in well 10-30-46-3W5, with the overlying Cycle F sitting directly on Cycle E.

CYCLE E: CORE CROSS SECTION PARALLEL TO DIP - KEYSTONE REGION

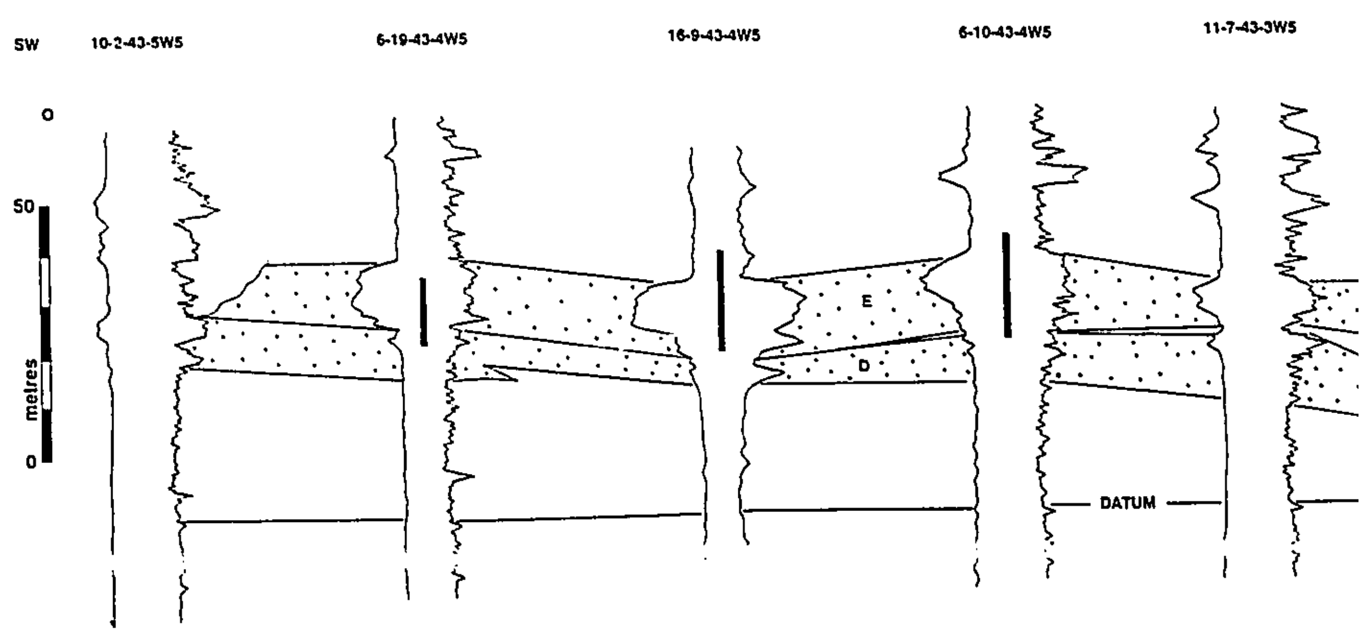


shoreface sediments. The cycle continues to thin basinward and at 3-25-46-3W5 consists of 5 m of interbedded distal shoreface sandstones and shelf mudstones. The log section (Fig. 7.5) indicates that the cycle pinches out a short distance basinward of this location. At all locations in the core section, the cycle is almost completely dominated by stacked beds of structureless (or vaguely stratified) beds of sandstone which become LAIS towards their tops. The cross-bedded zone of sandstone is absent throughout this cross section. In numerous beds, the tops are reworked into small-scale wave-rippled sandstone. The massive-to-laminated beds are still the dominant component in the distal shoreface sandstones at 3-25-46-3W5, and small mudstone rip-up clasts can be present near the bases of these beds.

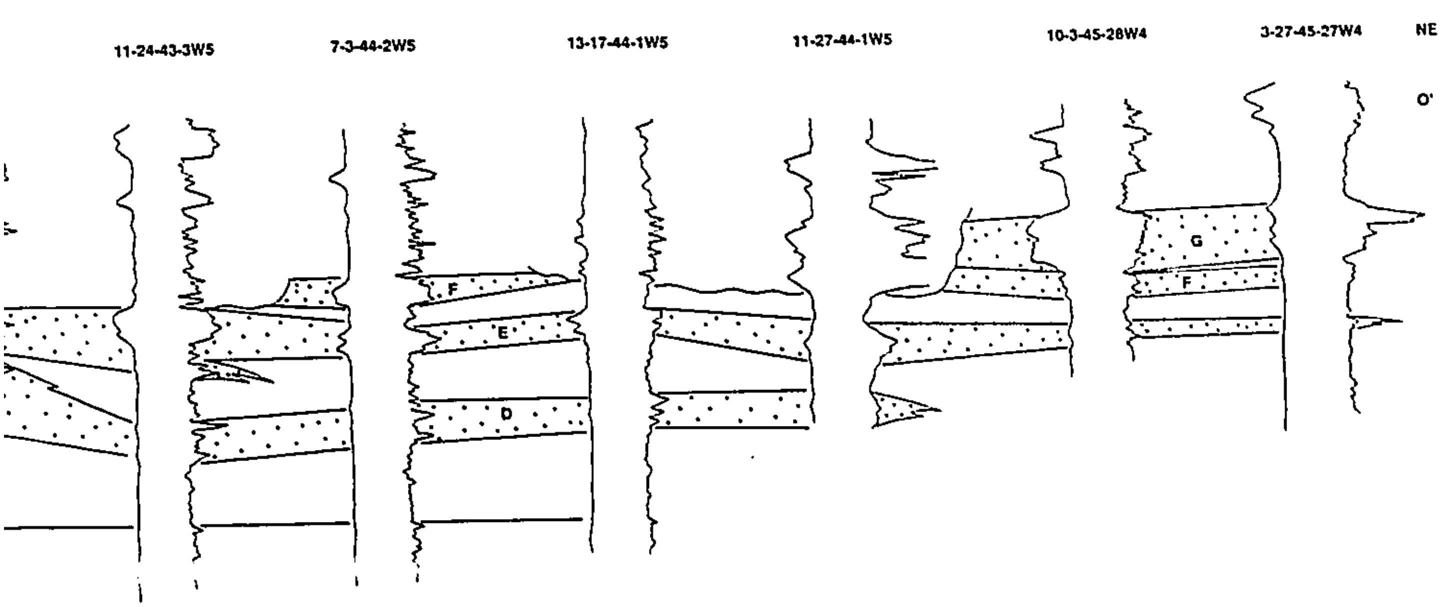
Cross section O-O' (Fig. 7.7) is a log section oriented in a dip direction and trends through the middle of the three tongues of thick sandstone where the Wilson Creek field is located. While core control is present in the Wilson Creek field, it is all concentrated in a small area, and all the locations contain successions very similar to that shown in figure 7.2 and 7.3. Therefore no core cross section was constructed. At the southwestern edge (6-19-43-4W5) the cycle consists of a sharply-based 15 m-thick shoreface succession which sits directly on Cycle D. The succession is thick and well developed for 8-10 km in a basinward direction, but then rapidly thins to about 8 m at 11-7-43-3W5. The cycle then gradually thins over 30-40 km to its depositional edge. The transgressive unit overlying Cycle E is 5-6 m thick at the northeastern edge of

Figure 7.7: Log cross section O-O' oriented parallel to dip and trending through the Wilson Creek region. Note the dramatic thinning of the cycle just east of the proximal elongate tongue of sediment at 11-24-43-3W5.

CYCLE E: LOG CROSS SECTION PARA



'ARALLEL TO DIP - WILSON CREEK REGION



the section and gradually thins back to the southeast to 11-24-43-3W5 where it is removed by fluvial erosion.

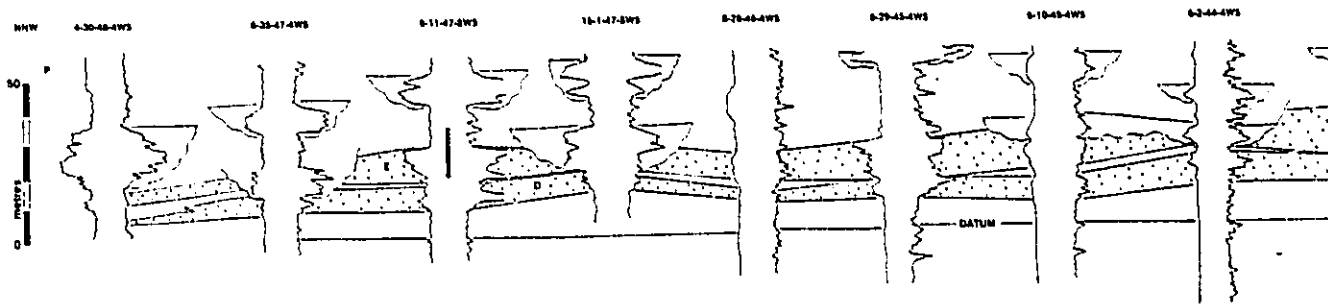
Section P-P' is a strike section (Fig. 7.8) that trends NW-SE for approximately 90-100 km sub-parallel to the paleoshoreline, and is offset a distance of 5-20 km basinward from the preserved landward edge of deposition. The section shows that Cycle E rises and falls stratigraphically along strike, that the underlying transgressive unit between Cycles D and E is removed in certain locations, and that Cycle E sits directly on Cycle D. As this section is located a short distance basinward of the preserved landward edge of deposition, the coarsening-upward succession that comprises Cycle E is thick and well-developed throughout most of the section. Successions within the three major thick sandstone tongues or protrusions can be seen in wells 6-29-45-4W5, 6-10-43-4W5, and 4-20-41-2W5 respectively. It is difficult to determine how many internal units or overlapping lobes are present within the cycle. The sections shows that Cycle E rises stratigraphically to the south, and it is conceivable that the two thicker successions in the south are stratigraphically younger than the northern succession documented in section N-N' (Fig. 7.5). The transgressive sediments which overlie Cycle E are not present at any location along this section.

7.4: Interpretation

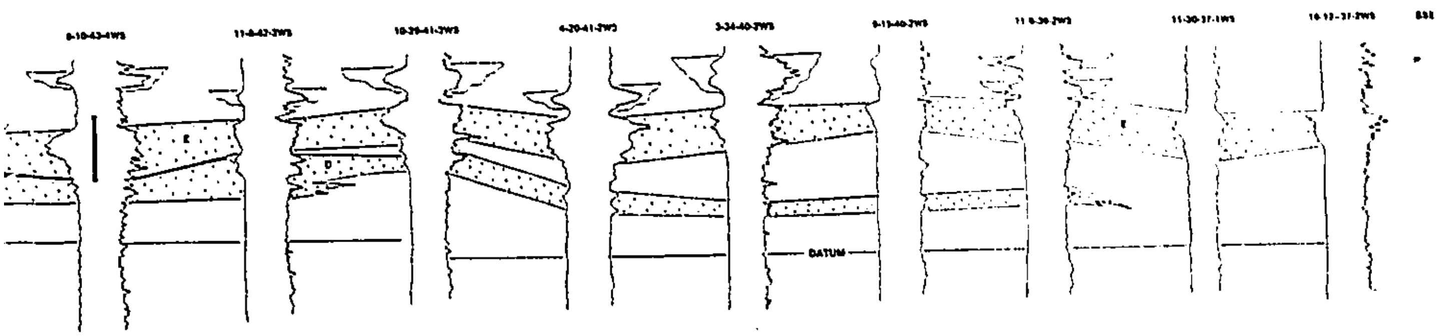
Cycle E is interpreted to represent the deposits of a prograding deltaic

Figure 7.8: Log cross section P-P' oriented parallel to strike through Cycle E. Note the three thick regions at 6-29-45-4W5, 6-10-43-4W5, and 4-20-41-2W5 corresponding to the three thick tongues of sandstone present in the study area.

CYCLE E: LOG CR



IG CROSS SECTION PARALLEL TO STRIKE



shoreline. The nature of the deltas is thought to be similar to those identified in Cycle D. The elongate tongue shape of the thick sandstone protrusions indicates that the proximal lobes were probably fluvial-dominated. As with Cycle D, the broad lobe of sandstone surrounding the elongate tongues is difficult to explain , seeing as how it is dominantly composed of massive sandstones. Some of this sand may have been redistibuted by waves. The core cross section (Fig. 7.6) shows that in areas 10-20 km basinward of the preserved landward edge of deposition, wave influence was sufficient to rework the tops of many beds into wave-rippled sandstone. However, this wave energy was apparently not sufficient to rework the sand into sedimentary structures more typical of wave-dominated deltas and shorelines, these being abundant HCS and SCS, and higher within the succession, cross-bedding and beach lamination. The cross-bedded zone higher in the succession within the Wilson Creek field is interpreted to represent the deposits of a shallow delta front environment/mouth bar environment, similar to the cross-bedded sandstones of Cycle D. All of these observations will be discussed in further detail in Chapter 11.

Internal subdivisions of Cycle E are not as apparent as in Cycle D. It does appear that the shoreline had at least three points of fluvial input, each of which built a thick, locally restricted elongate lobe into the basin. Temporal relationships between these three lobes are unclear. The two southern lobes may be younger than the northern lobe, but this is uncertain. Two or all of these lobes may have been active at the same time, or over a geologically short period of time. Sediments which were transported seaward of the proximal lobes together formed a coalesced lobate sheet of prodeltaic shoreface sediments. The dominant processes of deposition of these sediments are still interpreted to be related to their initial discharge at the delta mouth by fluvial processes.

The contact with the overlying transgressive sediments is not preserved (or was never present) in locations where Cycle E is fully developed and was subaereally exposed. Therefore no transgressive lag is observed. There is also no observed evidence of coastal plain flooding such as was observed overlying Cycle D. Further seaward, where the transgressive sediments overlie incompletely developed shoreface successions (eg. 2-32-46-3W5, Fig. 7.6), the transition between Cycle E and the overlying transgressive sediments is simply characterized by a basinward shift in facies of shoreface sandstones into shelf mudstones. There appears to be no significant erosion associated with the transgression, and the transgressive surface is simply a flooding surface.

CHAPTER 8: CYCLE F

8.1: Introduction, Distribution, and Geometry

Cycle F is present in the northeastern portion of the study area, mainly within and immediately to the southeast of the Keystone field (Fig. 8.1). The plan view geometry of the sandstone shows it to be a lobate body of sediment. Like the underlying cycles, Cycle F strikes NW-SE, and the sandstone thins from the southwest to the northeast. The lateral extent in a dip direction ranges from 20-40 km, and the cycle is present along strike for approximately 80 km within the study area. Large areas of Cycle F have been removed in various places due to both erosion by fluvial incision and by erosion and replacement by overlying marine cycles. The isolith pattern shows that the sandstone body is characterized by a lobate protrusion of thick sandstone in the northwest that is outlined by the 8 and 10 m isoliths. The thickest successions of Cycle F are 10-11 m thick, and occur within this area. This is the only major protrusion of thick sandstone within the study area. The 2-6 m isoliths surround this thicker area, resulting in a large, broadly lobate sandstone body of Cycle F.

8.2: Facies Associations

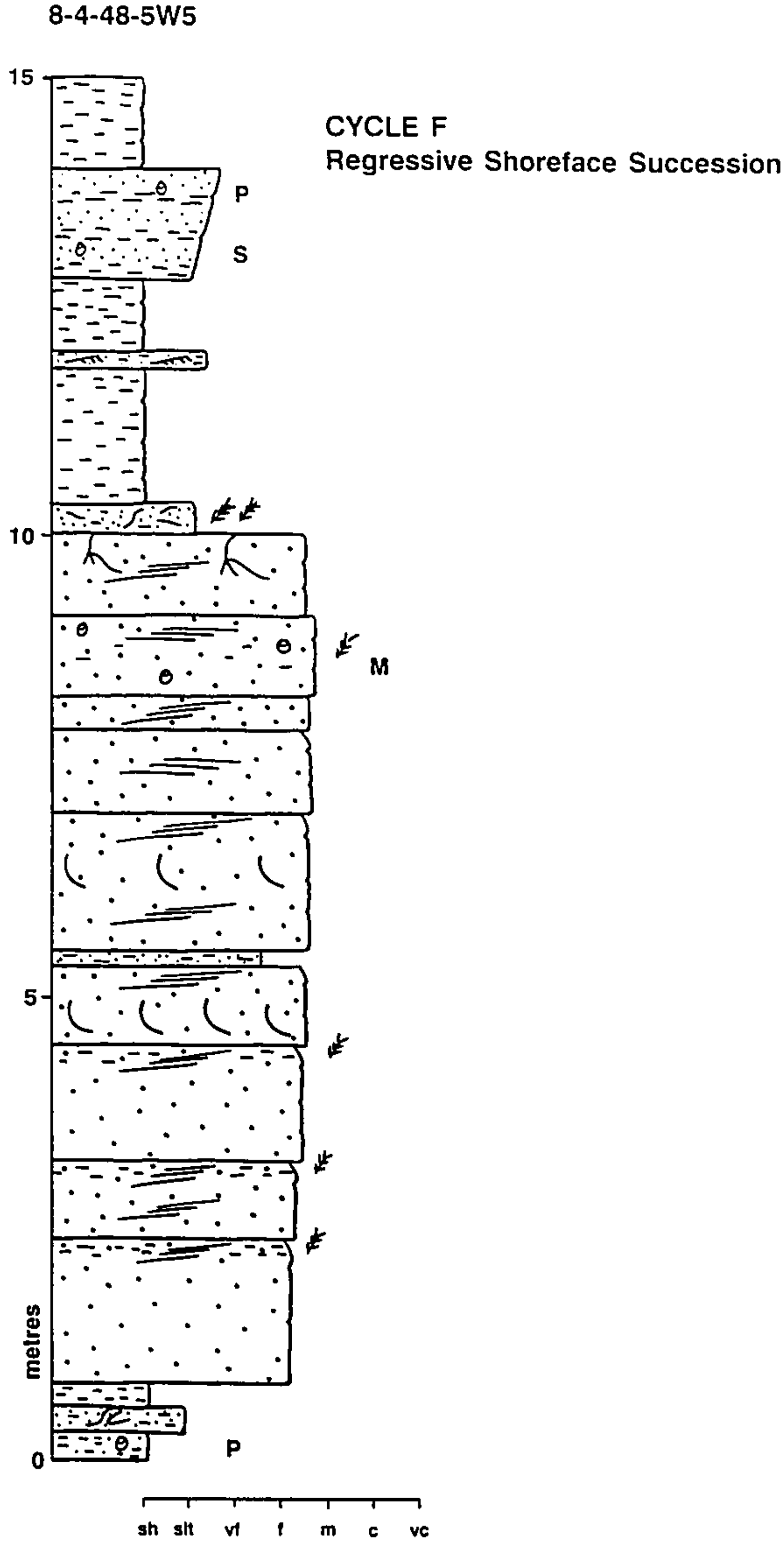
Although Cycle F is smaller in lateral extent then the underlying cycle, there are numerous cores through this interval, all of which are concentrated within the Keystone field. There are also many cores of the fluvial sandstones

Figure 8.1: Gross isolith map of marine sandstone in Cycle F. Contours are in metres, and the contour interval is two metres. The heavy black line represents the preserved landward edge of deposition.


R 5W5
R 1W5
R 26W4
5th Meridian
T 50
T 45
T 40
CYCLE F
GROSS MARINE SANDSTONE ISOLITH
CONTOUR INTERVAL: 2 m
MILES
0
6
12
0
5
10
15
20
KILOMETRES
R 5W5
R 1W5
R 28W4
R 26W4

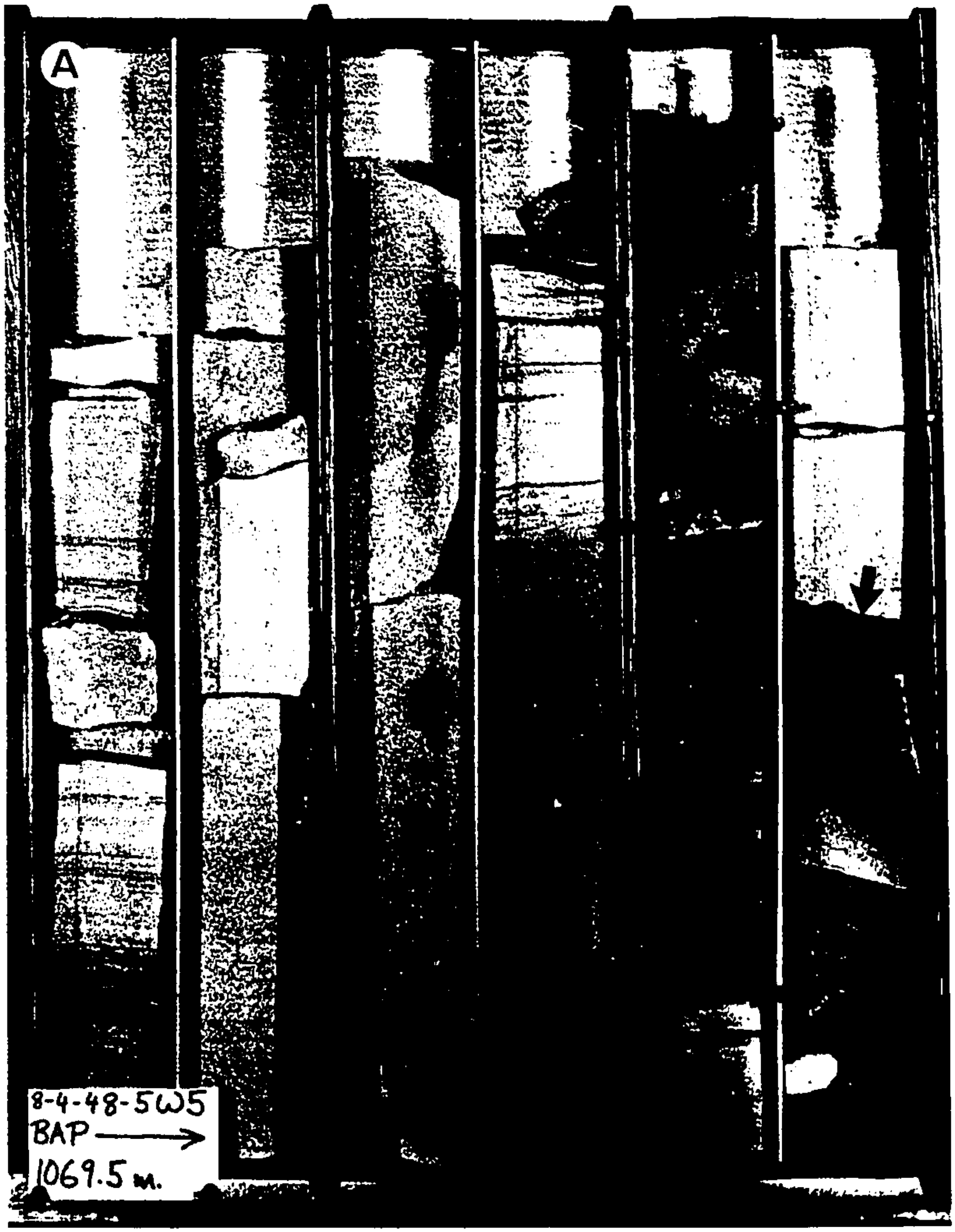
which incised and eroded Cycle F in this area. Figures 8.2 and 8.3 show an example of the shoreface succession of Cycle F which is located within the thick region of sandstone in Keystone (8-4-48-5W5). The shoreface succession typical of Cycle F is thinner than in underlying cycles, with fully developed successions (those showing evidence of subaerial exposure) being 8-11 m in thickness. The base of Cycle F successions, like those in underlying cycles, is very sharp, with shoreface sandstones sharply overlying shelf mudstones and sandstones typical of Facies Association 1. The succession itself is very similar to those discussed in Cycles D and E. It is dominated by beds of structureless or vaguely LAIS fine-grained sandstone which commonly become better stratified near the tops of the beds. The tops of beds in some wells also show wave-ripples, although theses were not observed in 8-4-48-4W5. Within the upper half of the succession, cross-bedded sandstone can be interbedded with the LAIS sandstone, as shown in figures 8.2 and 8.3. The uppermost sandstone beds of the succession are flat-bedded or LAIS sandstones which commonly contain root traces at their tops. *Macaronichnus* is the most common trace fossil present within the succession, and is usually located in the upper half of the succession. Other trace fossils present include *Ophiomorpha* and *Skolithos*. As with cycles D and E, the shoreface succession of Cycle F is characteristic of Facies Association 2a, and is interpreted to have been deposited in a prograding deltaic shoreline environment.

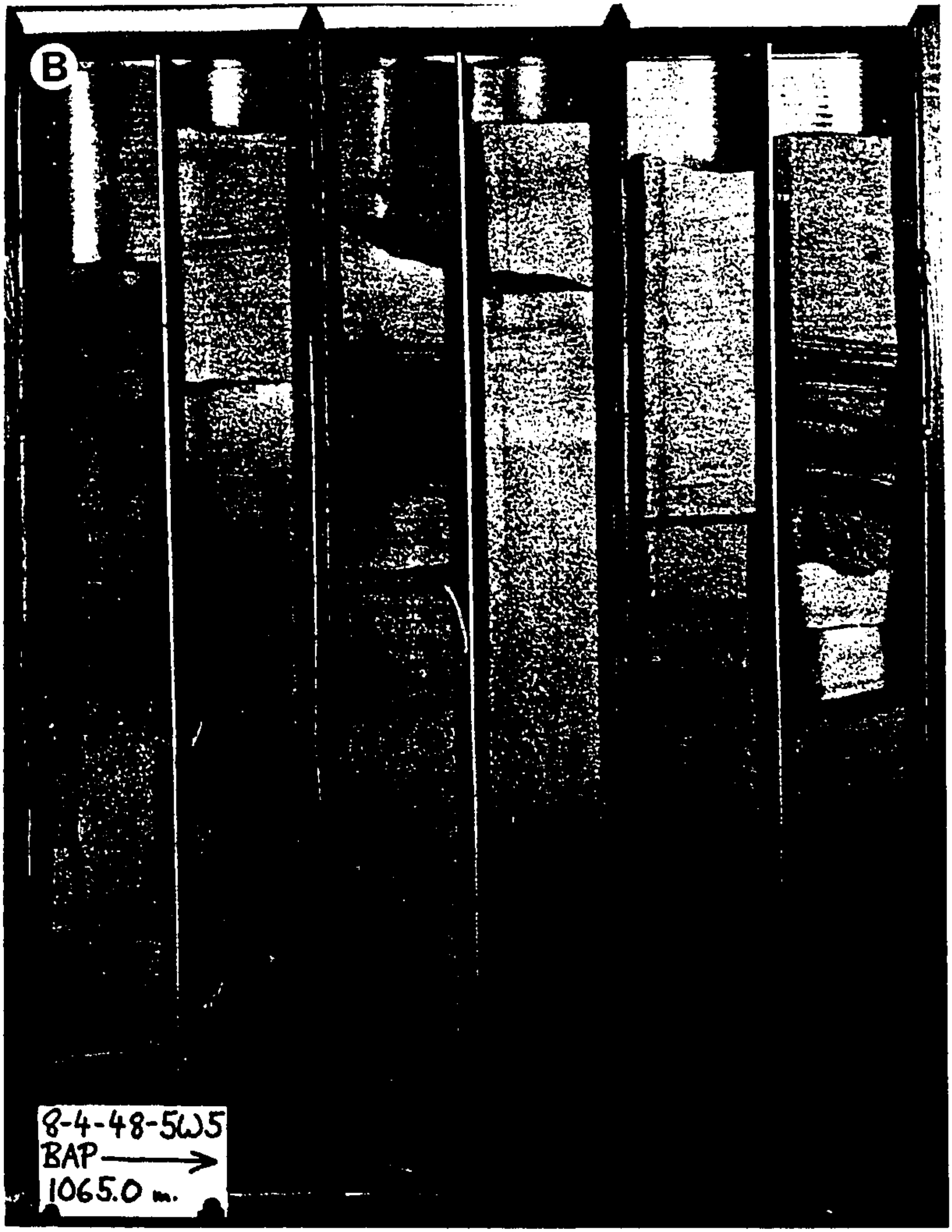
Figure 8.2: Stratigraphic section of the shoreline succession of Cycle F in well 8-4-48-5W5.



8-4-48-5W5
CYCLE F
Regressive Shoreface Succession
15
10
5
0
metres
P
S
M
P
sh
slt
vf
f
m
c
vc

Figure 8.3: Box photographs of Cycle F in well 8-4-48-5W5. The base of the succession in each plate is at the bottom left hand corner and the top is at the upper right hand corner. Each tube of core is 75 cm in height. The base of the Cycle F shoreline succession is shown by the arrow in Plate A. The sandstone unit below this is part of Cycle E. An excellent example of one of the massive-to-laminated beds is located between the two arrows on Plate B. Note the sharp base of the bed and the gradual transition into better defined lamination at the top of the bed. Plate C contains dominantly LAIS sandstones, with some *Macaronichnus* burrows. The top of the succession is shown near the base of Plate D.


A
8-4-48-5W5
BAP
1069.5 m.


B
8-4-48-5W5
BAP
1065.0 m.

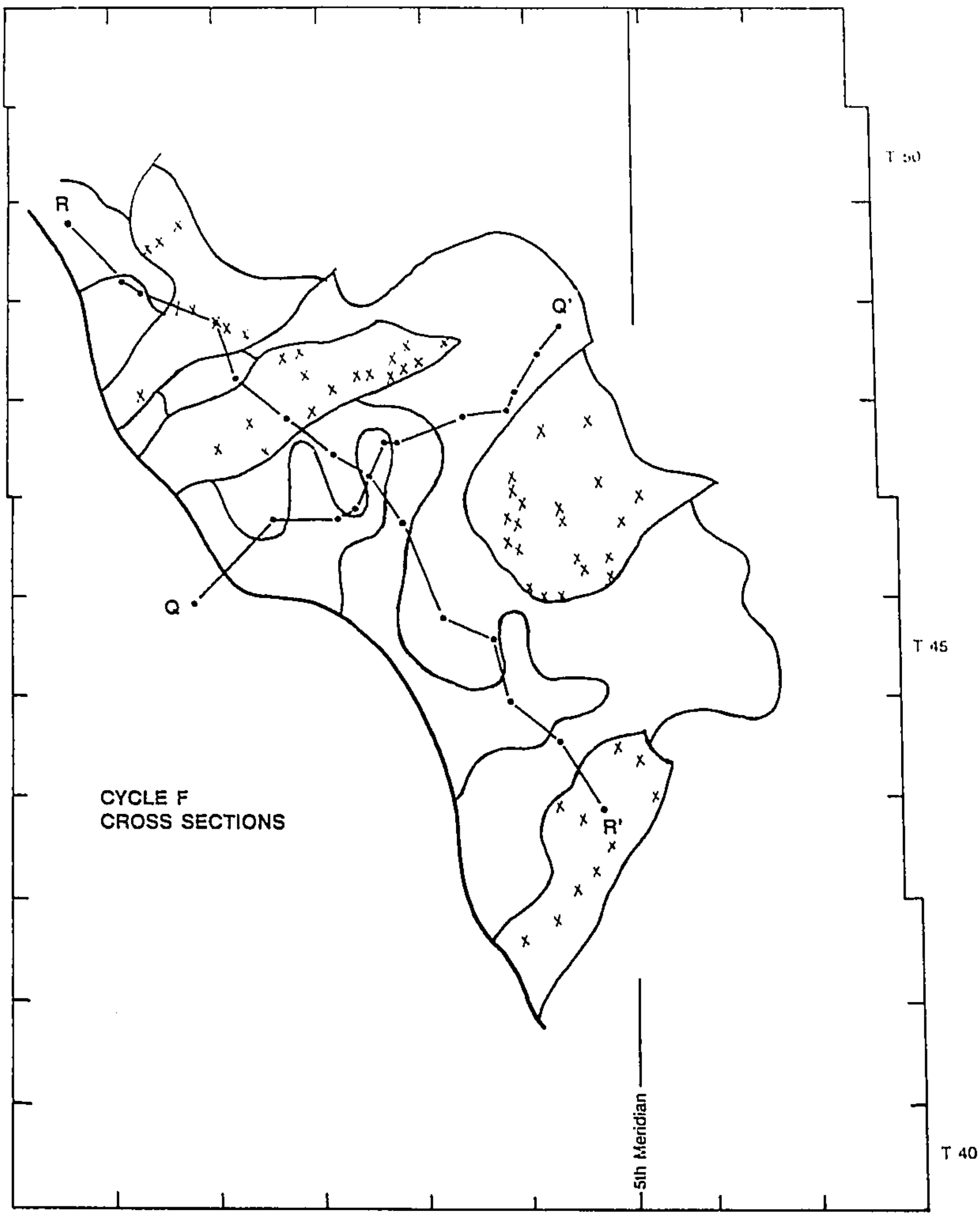

C
Mac.
8-4-48-5W5
BAP
1060.5m


D
8-4-48-5W5
BAP
1056.0m

8.3: Cross Sections

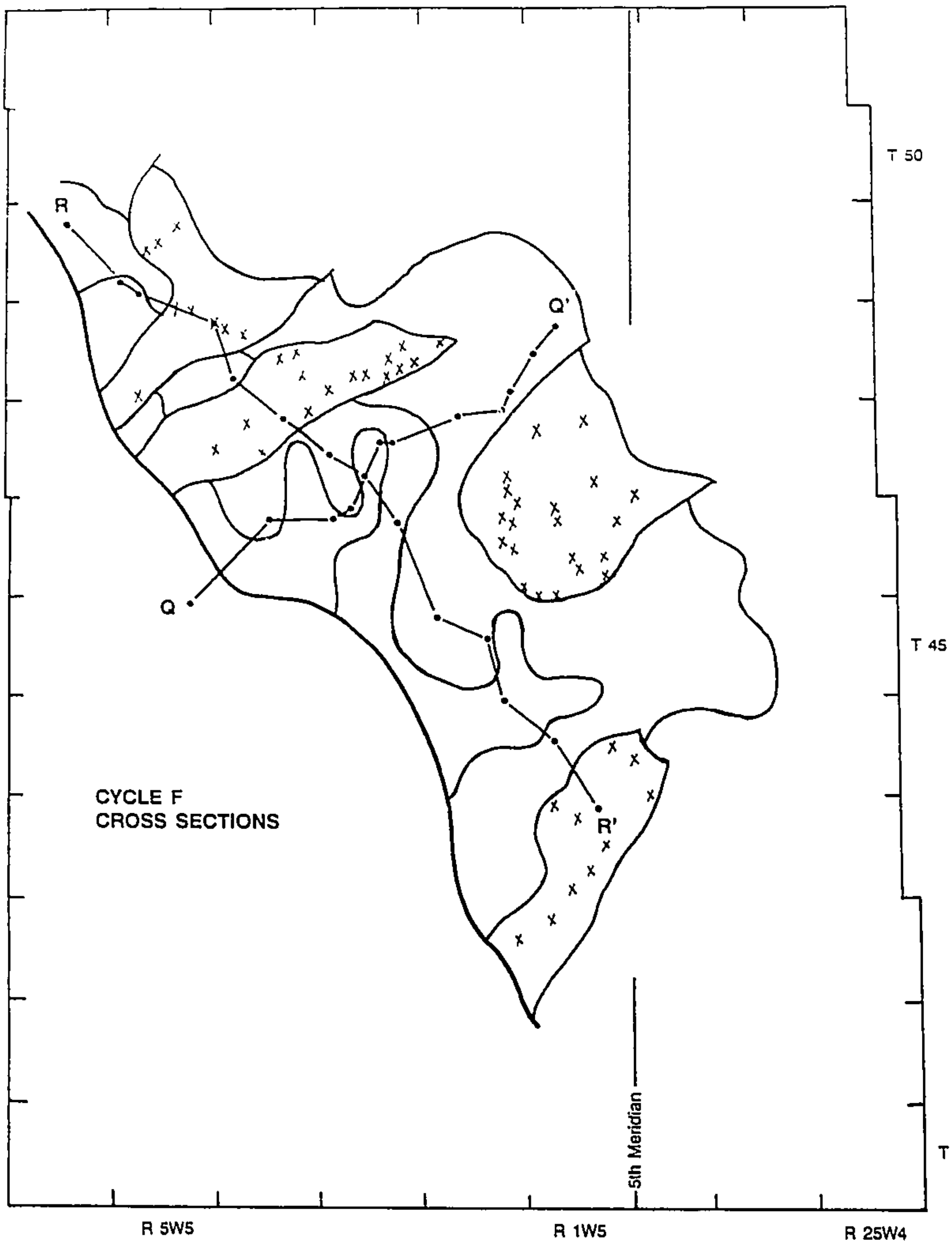
Figure 8.4 shows the location of detailed cross sections through Cycle F. Section Q-Q' (Fig. 8.5) is a log cross section trending parallel to dip through the thick lobe of sandstone in Keystone and further basinward to the northwest. The section is approximately 40 km in length. Figure 8.6 is a core cross section along the same line. The log section is hung for the most part using the top of Cycle E as a datum, although in the southwest, where logs penetrated deeper, the top of Cycle C was used as a datum. The core section is hung entirely using the top of Cycle E as a datum. Cycle F begins in the southwest as a relatively thick 8-11 m-thick succession which sits directly on the deposits of Cycle E with no intercyclic transgressive sediments present. In 10-30-46-3W5, the base of Cycle F may have eroded into the top of Cycle E, although this is uncertain. It may simply sit non-erosively on the top of Cycle E. Immediately basinward (10-32-46-3W5) a thin unit of transgressive marine mudstone and sandstone is present between Cycle E and F. The sharp base of Cycle F is readily apparent in this core. Cycle F remains approximately 10 m thick for about 10 km to the northeast. Wells 6-9-47-3W5 and 4-22-47-3W5 show a more distinct coarsening-upward profile than the succession immediately to the southwest, and the dominance of stacked beds of massive-to-laminated sandstone is still evident. The base of the succession in 4-22-47-3W5 is characterized by rip-up clasts of the underlying shelf mudstones. The coastal plain sediments immediately overlying the shoreface succession of

Figure 8.4: Simplified isolith map of Cycle F showing the location of the detailed cross sections used in this chapter.

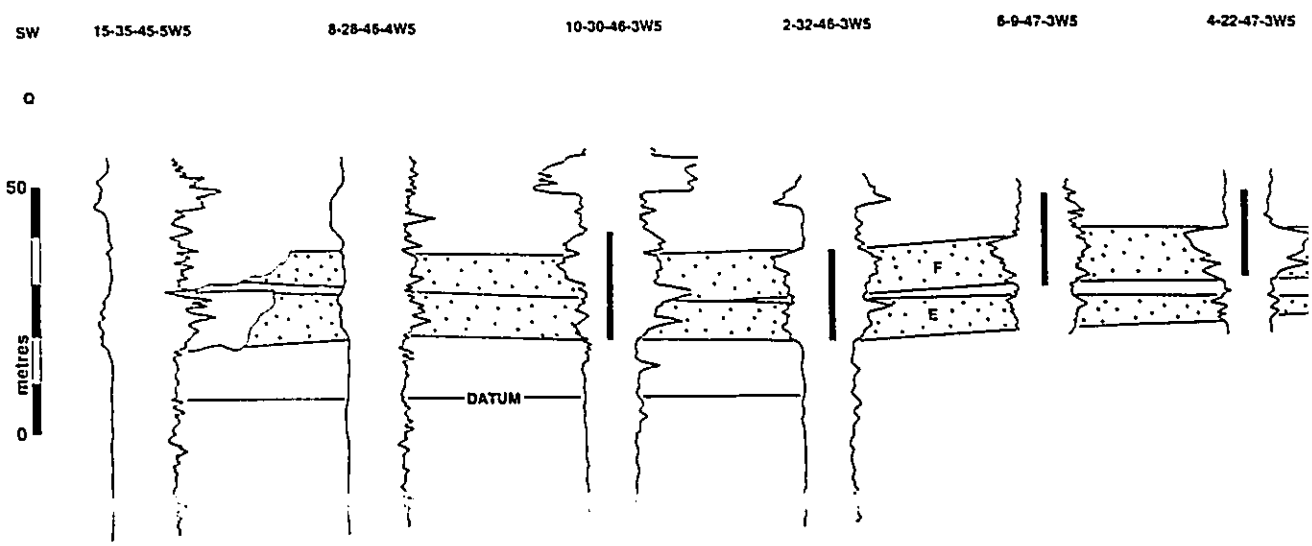


T 50
T 45
T 40
R
Q'
Q
R'
CYCLE F
CROSS SECTIONS
5th Meridian
R 5W5
R 1W5
R 25W4

Figure 8.5: Log cross section Q-Q', which is oriented parallel to dip and trends through the Keystone region. The section is approximately 40 km long. Note the stacking of Cycles F and E at 10-30-46-3W5 and the gradual splitting of these two cycles further to the northeast.


T 50
T 45
T 40
R
R'
Q
Q'
CYCLE F
CROSS SECTIONS
5th Meridian
R 5W5
R 1W5
R 25W4

CYCLE F: LOG CROSS



SECTION PARALLEL TO DIP

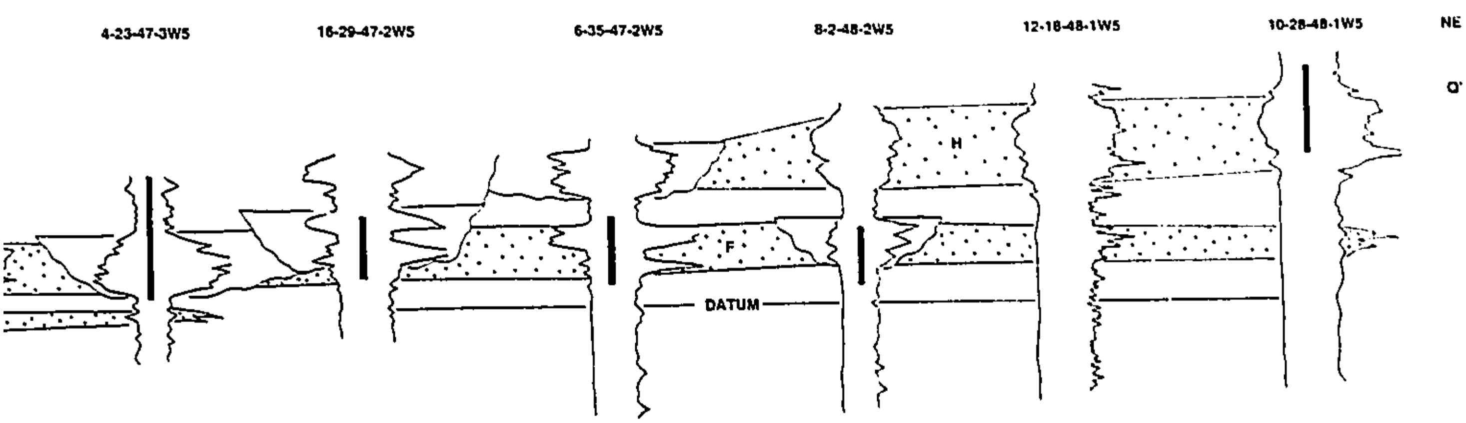
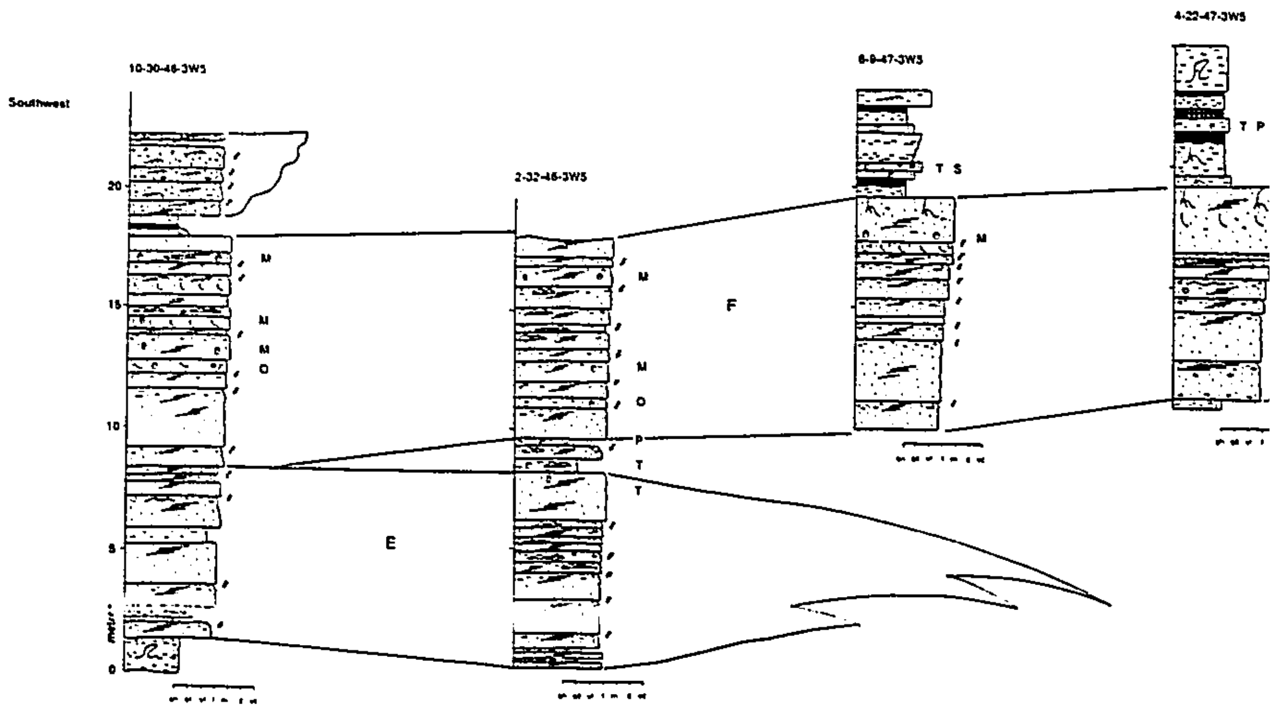
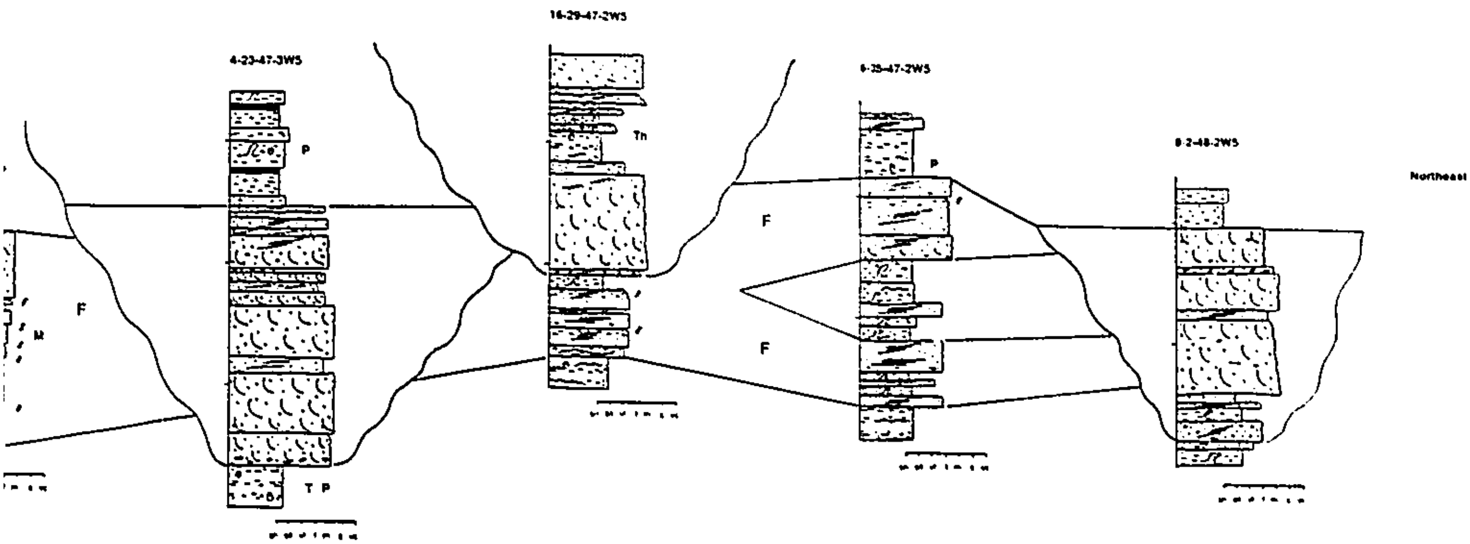


Figure 8.6: Core cross section oriented parallel to dip and following the same line as log cross section Q-Q'. Note the stacking of Cycles E and F in 10-30-46-3W5 and the gradual splitting of the tow cycles in 2-32-46-3W5. Also note the lack of a well-developed cross-bedded interval in any of the fully developed shoreline successions of Cycle F.

CYCLE F: CORE CROSS SECTION PARALLEL TO DIP




4-23-47-3W5
16-29-47-2W5
Northeast
P
Th
F
M
T P

Cycle F in these two wells contain coal beds which are overlain by thin beds of silty, coastal plain mudstones containing marine trace fossils such as *Teichichnus*. At 4-23-47-3W5, Cycle F has been eroded by fluvial incision, and is replaced by a 10 m-thick succession of cross-bedded fluvial sandstones typical of Facies Association 3. The upper portions of Cycle F have also been removed by fluvial incision at 16-29-47-2W5. This core contains the lower portions of Cycle C, and indicates that the succession is starting to thin in a basinward direction into finer-grained, thinner beds of LAIS sandstone, with occasional interbeds of marine mudstones. Well 6-35-47-2W5 shows more evidence of the basinward thinning. In this location, Cycle F is a succession of interbedded lower shoreface sandstones and shelf mudstones. The sandstone beds are still dominated by LAIS structures. The interbedded mudstones often show soft sediment deformation, and the bases of the sandstone beds can contain mudstone rip-up clasts. Northeast of this point, Cycle F is again removed by fluvial incision (8-2-48-2W5), and reaches its depositional edge 5-10 km further basinward.

The transgressive sediments overlying Cycle F are 10-15 m-thick at the depositional edge of the cycle. These thin back to the southwest to about 5-10 m at 6-35-47-3W5, but further southwest of this location, they have been removed by fluvial erosion. The core at 6-35-47-3W5 indicates that these transgressive sediments are typical of Facies Association 1, being composed of interbedded shelf mudstones and wave rippled sandstones.

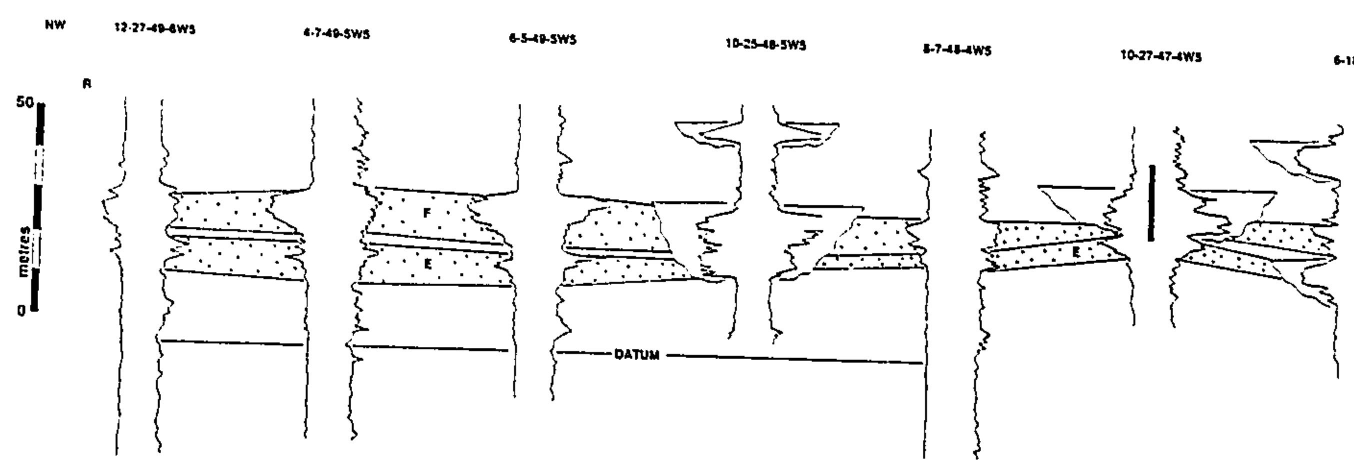
Figure 8.7 is log cross section R-R', which trends parallel to strike through Cycle F. The section shows that Cycle F is composed of only one coarsening-upward succession within the study area. At the northwestern end of the section, Cycle F is a thick, well developed succession approximately 10 m in thickness. The cycle retains this thickness within the northwestern lobate area (up to 6-9-47-3W5). Southeast of this location, Cycle F begins to thin, and by 13-17-44-1W5 is just a 3-5 m thick poorly-developed cleaning-upward signature on the dual induction log. Further to the southeast (8-34-43-1W5), Cycle G is suddenly present, and sits directly upon the distal deposits of Cycle F. Cycle F either pinches out south of this location or is removed by downcutting associated with the establishment of Cycle G. The latter is thought to be the case in some areas, most notably the southerly "x" swath where Cycle F has been removed.

8.4: Interpretation

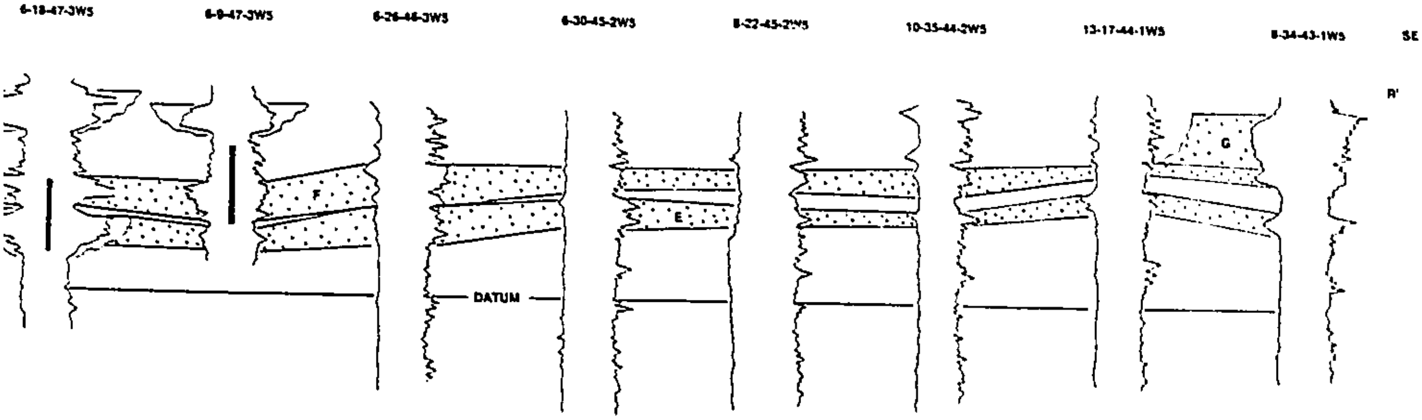
The sediments of Cycle F are interpreted to have been deposited in a prograding deltaic shoreline environment. Facies successions indicate that the delta was similar in nature to those interpreted for Cycles D and E. The abundance of massive-to-laminated and very vaguely LAIS stratified fine-grained sandstone beds indicates that these sediments were deposited very rapidly, with much of the sediment settling out of suspension rather than being deposited as bedload. The thick lobate section of sandstone centred in the

Figure 8.7: Log cross section R-R', oriented parallel to strike through Cycle F. Note the sudden appearance of Cycle G at he southern end of the section and how it sits directly on Cycle F.

CYCLE F: LOG CROSS S



SS SECTION PARALLEL TO STRIKE



Keystone field indicates that the source of fluvial input was probably most proximal in this area. Cycle F does not contain the elongate shore-normal tongues of delta front or mouth-bar sediments that are present in Cycles D and E. This is also evidenced in the paucity of cross-bedded sandstones within the upper portions of the shoreline succession. In both Cycles D and E, the shore-normal tongues of sandstone contained intervals of coarser-grained (up to medium-grained) cross-bedded sandstones up to 5-6 m in thickness, indicating a closer proximity to the source of sediment than the underlying massive-to-laminated beds. The succession in Cycle F is almost totally dominated by the massive-to-laminated beds, and is thinner than those of Cycle E (11 m vs. 18 m). The delta system of Cycle F may have been somewhat smaller than those in the underlying cycles, and did not deposit a large mouth-bar tongue of sandstone. Alternatively, the most proximal areas of marine deposition in Cycle F may not be preserved, and may lie southwest of the preserved landward edge of deposition. The former is more likely, simply because numerous cores in the Keystone field indicate that the deltaic succession was completely developed, and became subaerially exposed. This necessitates that the source of sediment was very proximal, but was probably somewhat smaller than those of Cycles D and E, and thus also deposited a thinner succession. Only one major depocenter of the delta system is present. If others exist, they are outside the study area.

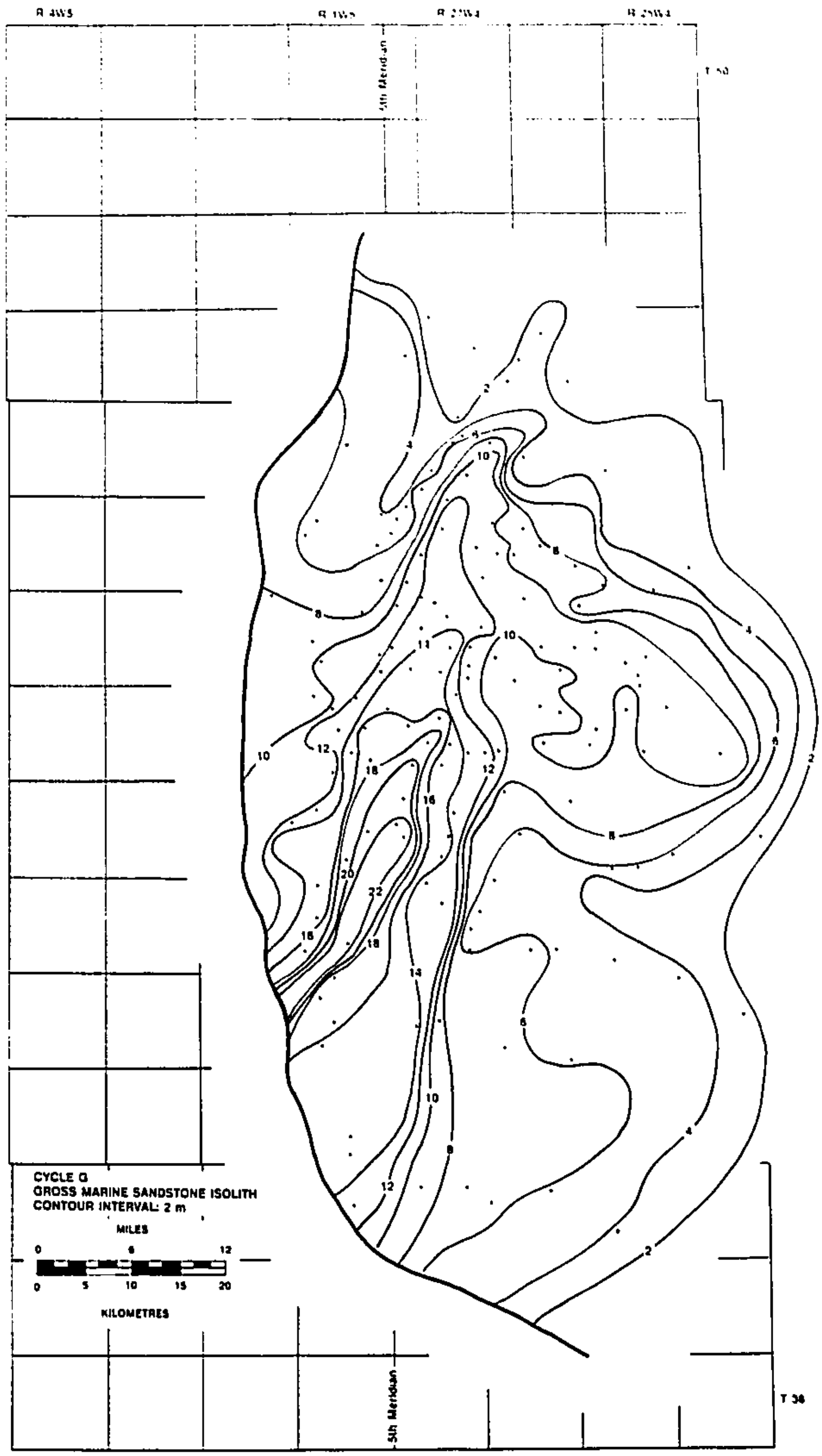
The contact with the overlying transgressive sediments is only preserved in distal successions of Cycle F (eg. 6-35-47-2W5). The contact is characterized by a shift in facies from lower shoreface sandstones to shelf mudstones. No erosion is evident at this contact, and is thus interpreted as a flooding surface rather than a ravinement surface. The transgressive sediments themselves are typical of Facies Association 1, and are interpreted to have been deposited in a mid to shallow shelf environment, where waves and storms could still move sediment. The landward edge of the transgressive deposits may be represented by the thin unit of *Teichichnus*-burrowed mudstone and siltstone immediately overlying the coal beds at 6-9-47-3W5 and 4-22-47-3W5 (Fig. 8.6). The presence of these trace fossils necessitates that the waters in which these sediments were deposited were at least partially saline. These units may be the landward edge of the transgressive sediments, in a similar manner to the flooded coastal plain sediments overlying Cycle D.

CHAPTER 9: CYCLE G

Introduction, Distribution, and Geometry

Cycle G is present only in the eastern portion of study area, with most of the cycle located east of the fifth meridian (Fig. 9.1). Shoreline sediments from this cycle form the hydrocarbon reservoir in the Ferrybank field. The regional cross sections of chapter 4 showed that Cycle G can attain thicknesses in excess of 20 m, and is the thickest of all the cycles in the study area. Like the underlying cycles, the sandstone body strikes NW-SE and thins to the north-northeast. The cycle is present along strike for a distance of 100-110 km and in a dip direction for 30-50 km. Unlike the underlying cycles, there are very few locations where Cycle G has been removed by fluvial erosion. The dominant morphological feature of the cycle is the single thick elongate, tongue of sandstone, outlined by the 12 m isolith, which trends to the northeast. The tongue is approximately 6-30 km wide and 50 km in length, and attains a maximum thickness of 23 m at the centre. The apparent trend of all the underlying paleoshorelines has been NW-SE. If this is also true for Cycle G and the preserved landward edge of deposition is approximately parallel to the paleoshoreline, this large tongue of sediment is shore-normal to shore-oblique in orientation. Isoliths are very closely space in the central portions of this tongue, especially in a "strike" direction. This protrusive tongue of sandstone is similar to those identified in underlying cycles, but noticeably larger in areal

Figure 9.1: Gross isolith map of marine sandstone in Cycle G. The contours are in metres and the contour interval is 2 m. The heavy black line represents the preserved landward edge of deposition. Note the presence of the single thick tongue of sandstone oriented SW-NE, with no evidence of fluvial incision into this tongue.



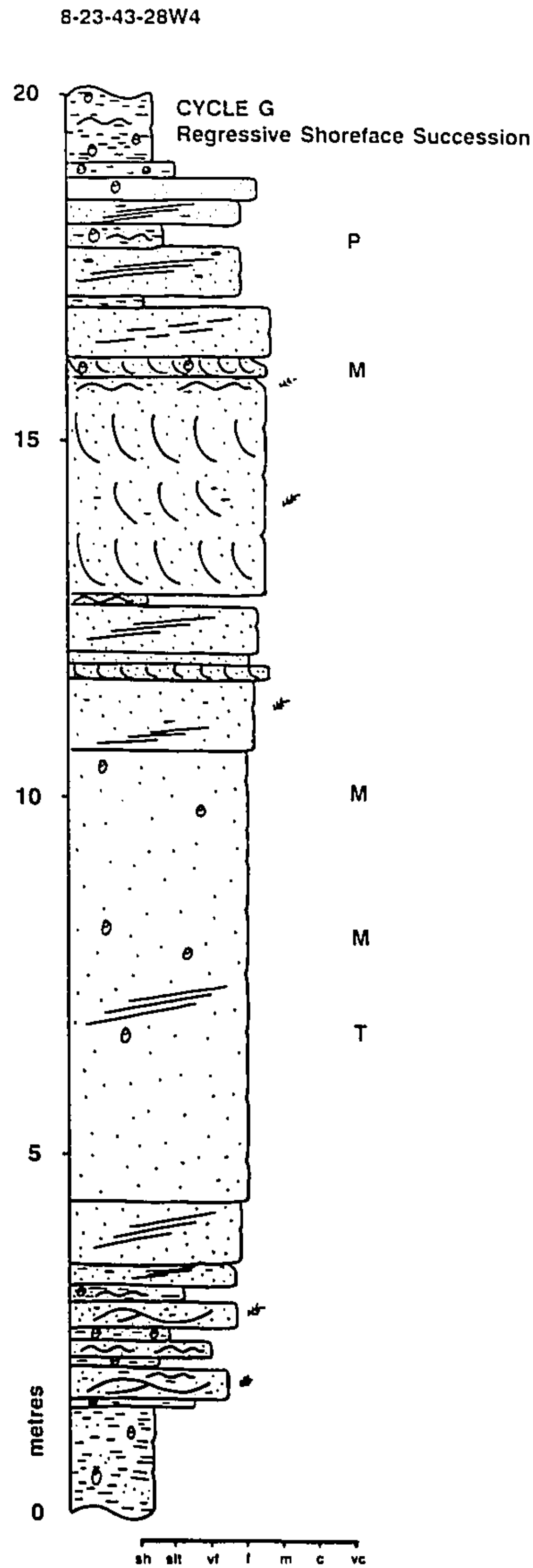
R 4W5
R 1W5
R 28W4
R 25W4
T 50
5th Meridian
T 38
CYCLE G
GROSS MARINE SANDSTONE ISOLITH
CONTOUR INTERVAL: 2 m
MILES
0
6
12
0
5
10
15
20
KILOMETRES

extent. It is also oriented somewhat more parallel or oblique to the preserved landward edge of deposition tha tongues in underlying cycles. The 2-10 m isoliths outline the thick tongue in some areas, but also reveal a more broad lobate geometry to the east of the tongue, which appears to strike in a almost true N-S direction and dip directly to the east. Thickness changes in this area are less dramatic than within the thick tongue to the west.

9.2: Facies Associations

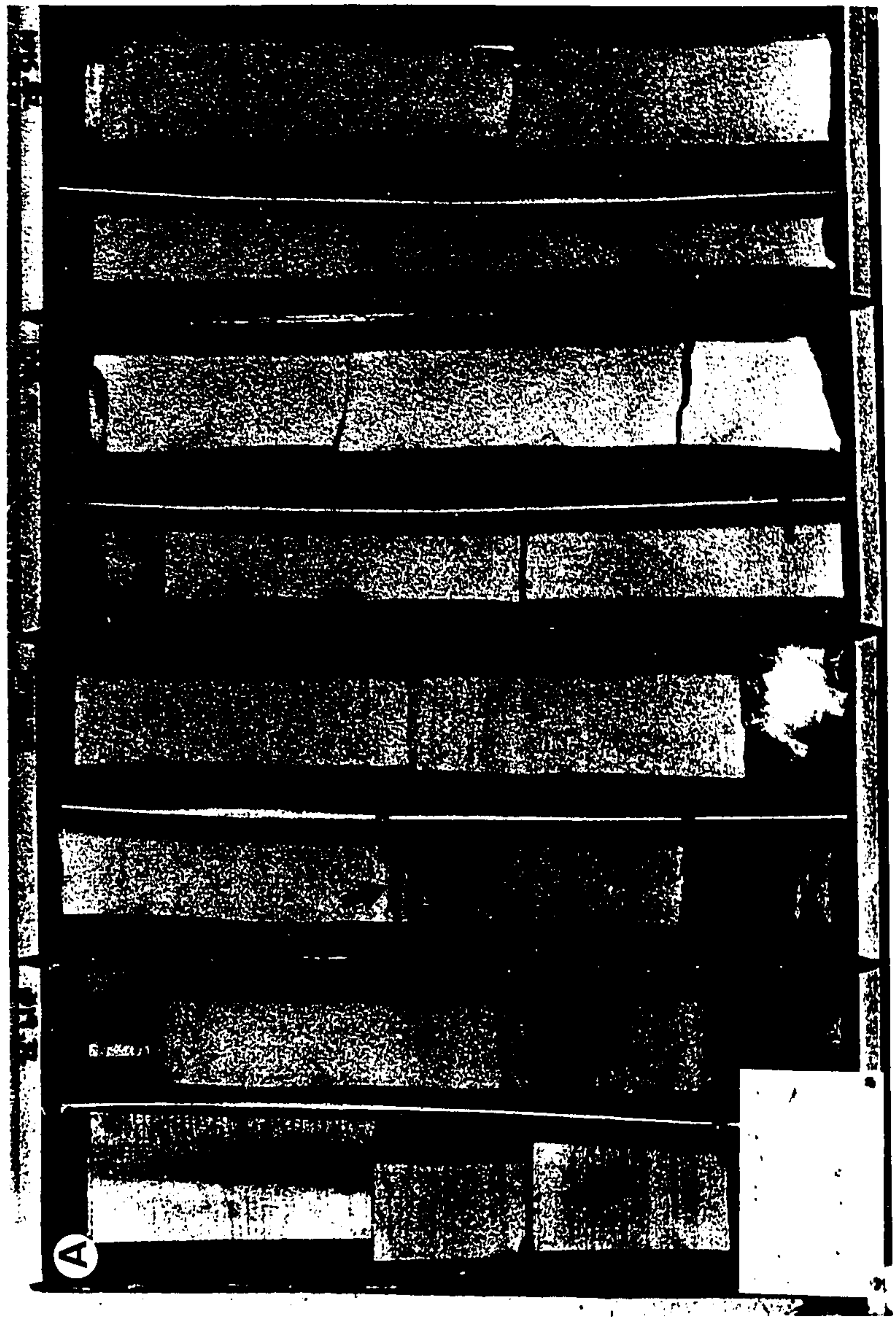
A large data base of core through Cycle G is present, mainly because it is the reservoir for the Ferrybank field. All of the core for this cycle is located in or near to Ferrybank. A typical example of the succession in Cycle G can be seen in well 8-23-43-28W4 (Figs. 9.2, 9.3). These figures show the base of the cycle can be characterized by a thin, 1-2 m thick interval of interbedded LAIS or wave-rippled sandstone and mudstone which is transitional between the shelf mudstones of Facies Association 1 and the overlying shoreface sediments. This transition zone is not present in all locations. Some cores show the familiar sharp base in which the shoreface sediments of the cycle sit directly on the shelf mudstones. The succession is classified as being typical of Facies Association 2a, and is very similar in most ways to those detailed in the underlying Cycles D-F, being dominantly composed of massive-to-laminated beds of fine-grained sandstone in the lower half to two-thirds. Beds of this structureless or very vaguely stratified sandstone can be greater than 6 m in

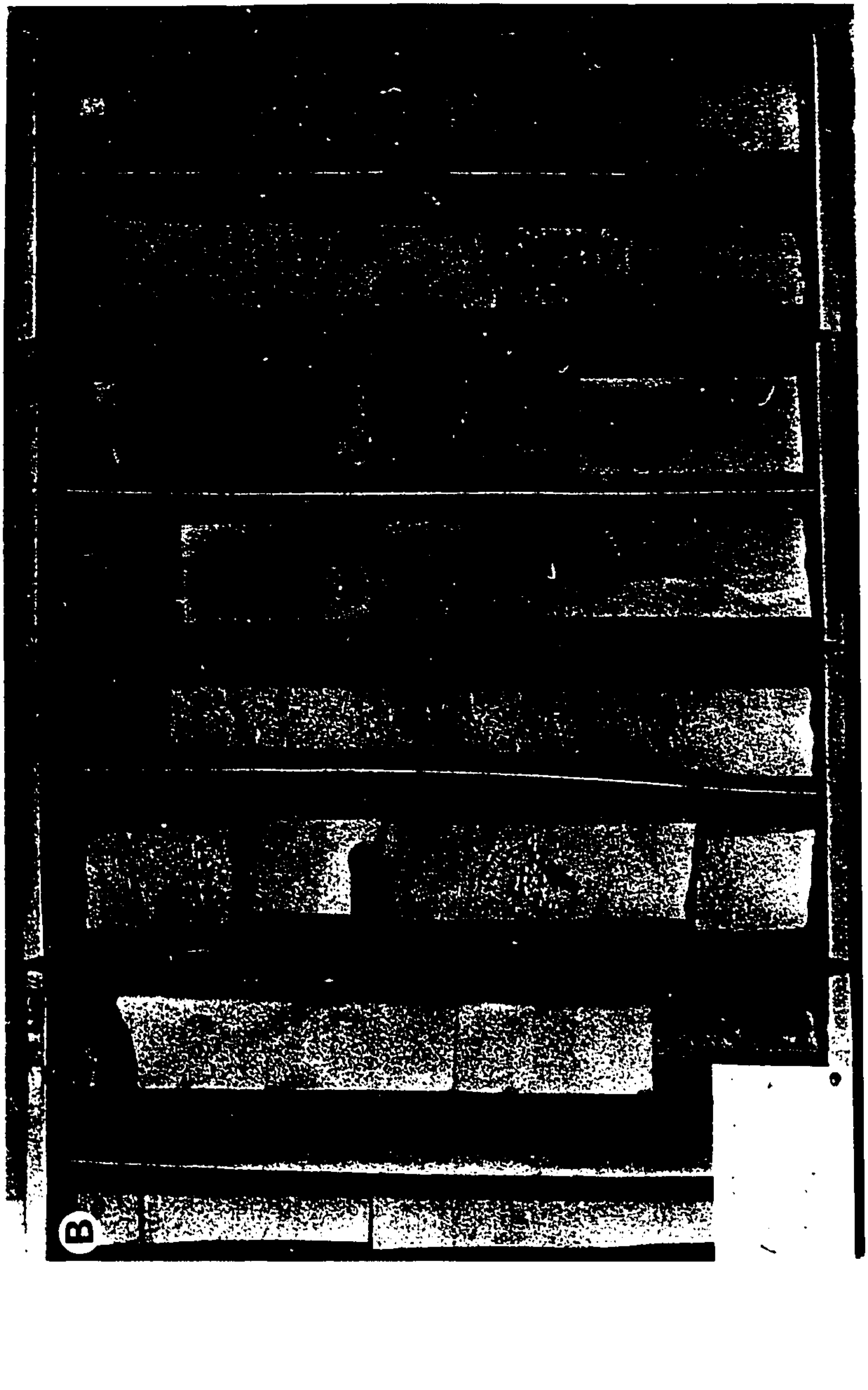
Figure 9.2: Stratigraphic section of the shoreline succession of Cycle G in well 8-23-43-28W4. Note that the succession does not show evidence of subaerial exposure at the top.

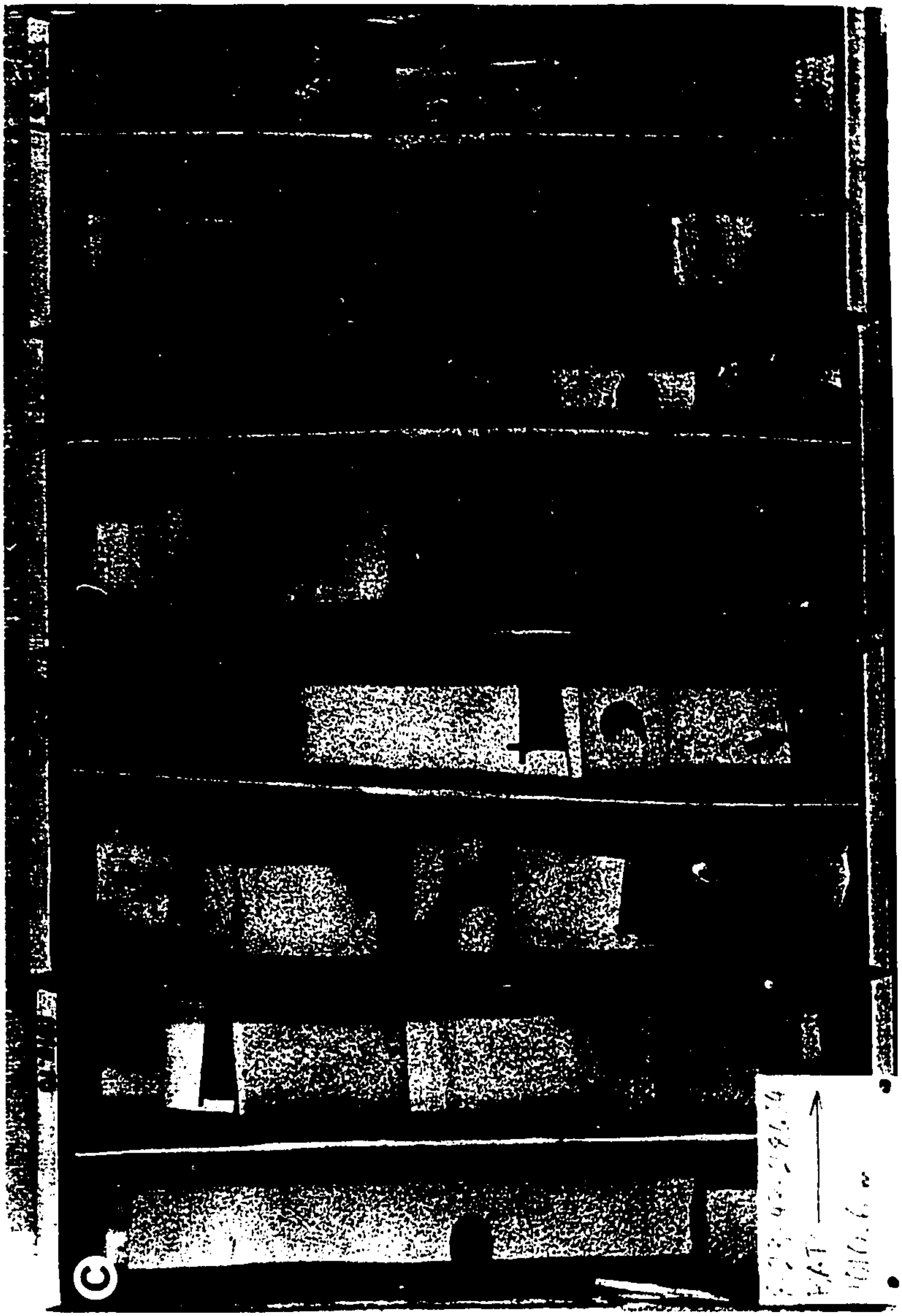


8-23-43-28W4
CYCLE G
Regressive Shoreface Succession
20
15
10
5
0
metres
P
M
M
M
T
sh slt vf f m c vc

Figure 9.3: Box photographs of Cycle G in well 8-23-43-28W4. The base of the succession in each plate is at the bottom left hand corner and the top is at the top right hand corner. Each tube of core is 60 cm in height. The basal metre in Plate A shows a transitional interval between the shelf and shoreface sediments of storm sandstones interbedded with mudstones. A good example of the massive-to-laminated beds is shown between the two arrows in Plate A. The arrows in Plate B point to zones of *Macaronichnus* burrows. The arrow in Plate C marks the transition from dominantly massive-to-laminated beds to cross-bedded sandstone. The first arrow in Plate D marks the surface of initial transgression, and the second arrow marks the surface of final transgression.


A


B


C

D
1002.00
1003.18
%-23-43-28W4
BAP
1005.8m

thickness. The main difference between the succession in Cycle G and others is the thickness of the cross-bedded interval overlying the massive-to-laminated beds. In wells within the thick tongue of sandstone, the massive sandstone beds are overlain by up to 4-5 m of cross-bedded sandstone. The cross-bedding is usually moderatley to well-developed, in contrast to the vague stratification of other cycles. The cross-bedded sandstones are the uppermost sediments within the succession. *Macaronichnus* is common within the succession, usually within the massive-to-laminated beds, but it is also present in the cross-bedded sandstones. Other traces present include *Rosselia*, and *Teichichnus*. These are restricted to the lower massive sandstones.

The shoreface succession of Cycle G is incompletely developed or preserved within the study area. Another major difference between Cycle G and other cycles is that there is no evidence of subaereal exposure in any core. The cross-bedded sandstones appear to represent the maximum preserved extent of the shoreline progradation. The cross-bedded sandstones are commonly overlain by 1-3 m of finer-grained LAIS sandstone, which is interbedded with thin beds of marine mudstone. This interval is in turn always overlain by *Helminthopsis*-burrowed mudstones of Facies Association 1b.

The abundance of the massive-to-laminated beds indicates that Cycle G was likely deposited in a similar environment to the underlying cycles. Therefore, Cycle G is interpreted to have been deposited by the progradation of a deltaic shoreline. The cross-bedded interval, again, may be indicative of deposition in a shallow marine environment immediately seaward of distributary channels, or possibley within the channels themselves. The transition from cross-bedded sandstones to *Helminthopsis*-burrowed mudstones at the top of the cycle represents a progressive deepening of the depositional environment associated with the ensuing transgression.

9.3: Cross Sections

The location of the detailed cross sections through Cycle G are shown in figure 9.4. Log cross section S-S' (Fig. 9.5) is a dip-oriented section which trends through the thick central tongue of sandstone. Figure 9.6 is the accompanying core cross section. The log cross section shows that at the preserved landward edge of deposition, Cycle G sits sharply on a thin transgressive unit of marine mudstones which separate it from the underlying Cycle E. Cycle F is not present in this area. The cycle appears to be incompletely preserved at its southwestern end, as the upper portions are removed by fluvial erosion in 6-35-40-2W5 and 7-7-41-1W5. A short distance to the northeast, at 6-10-41-1W5, the cycle is over 20 m thick. This is located within the thickest portion of the sandstone tongue, just to the southwest of the Ferrybank field. The log signature on both the SP and the dual induction logs in 6-10 and 16-4-42-1W5 is somewhat muted and of low amplitude. It has been interpreted as representing a shoreline succession because it is more similar to log signatures in Ferrybank, which definitely represent shoreline sediments,

Figure 9.4: Simplified isolith map showing the location of the detailed cross sections used in this chapter.

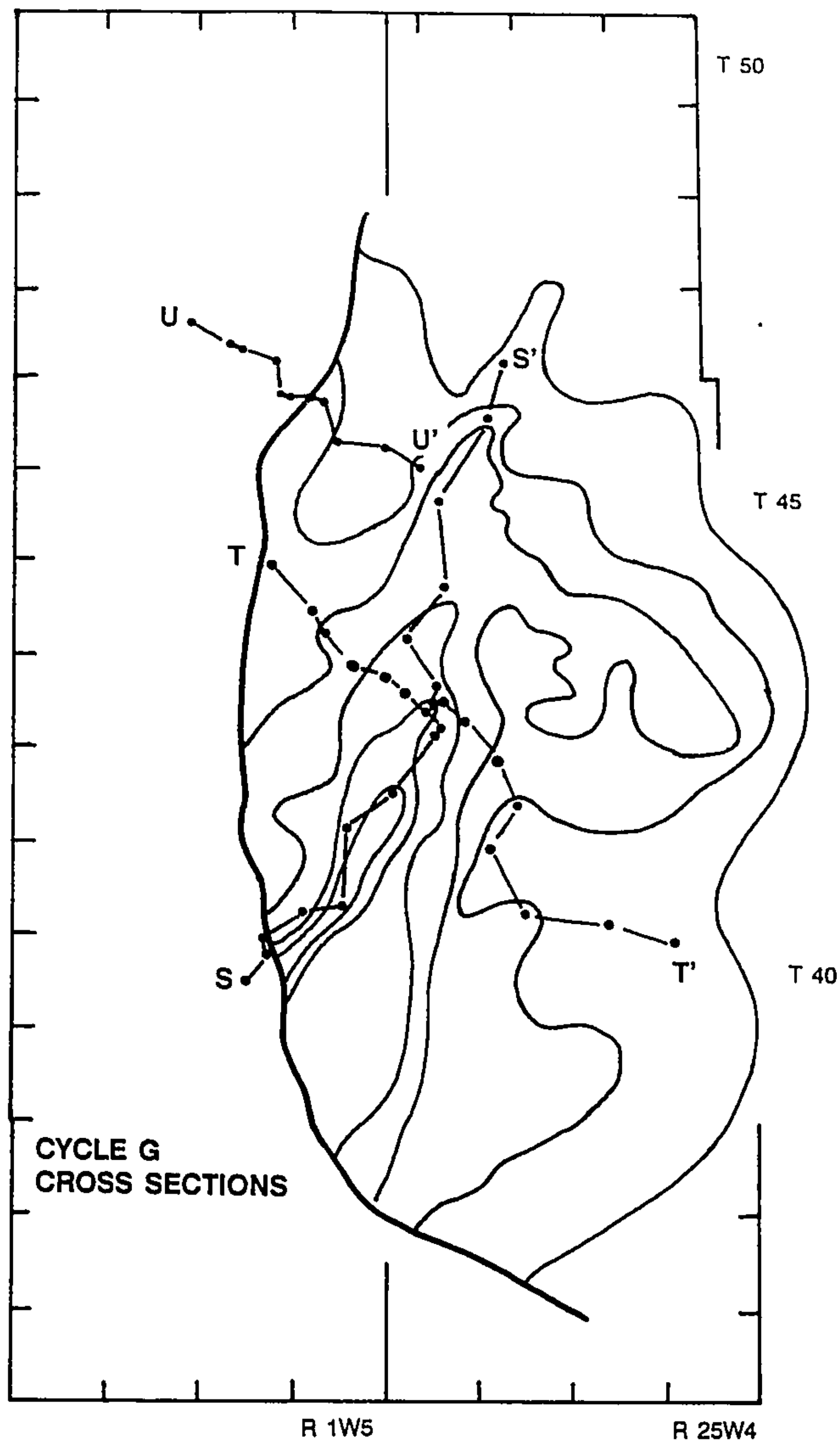
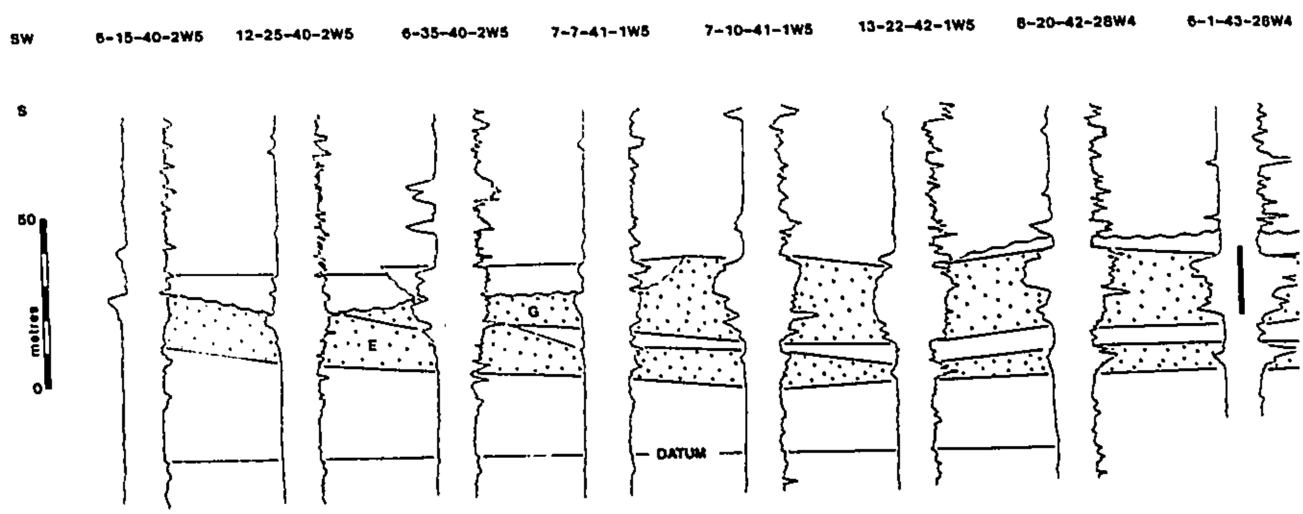

T 50
T 45
T 40
U
U'
S'
T
S
T'
R 1W5
R 25W4
CYCLE G
CROSS SECTIONS

Figure 9.5: Log cross section S-S', oriented parallel to dip. Note the large thickness of the sandstone within the southern portion of the thick central tongue, and the lack of evidence of fluvial incision into the cycle.

CYCLE G: L



SW
6-15-40-2W5
12-25-40-2W5
6-35-40-2W5
7-7-41-1W5
7-10-41-1W5
13-22-42-1W5
6-20-42-28W4
6-1-43-28W4
S
50
metres
0
G
E
DATUM

.OG CROSS SECTION PARALLEL TO DIP

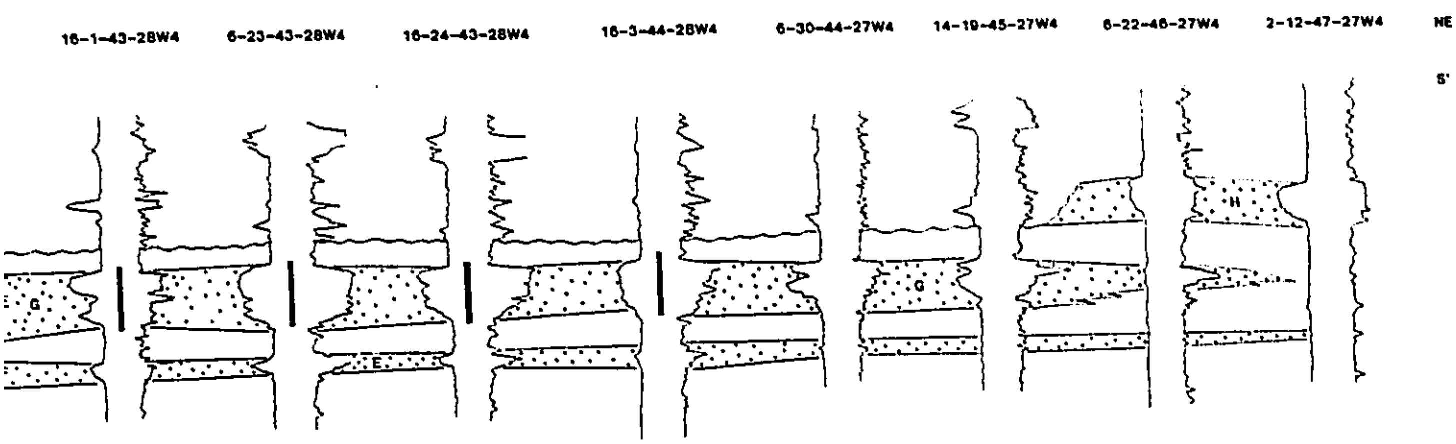
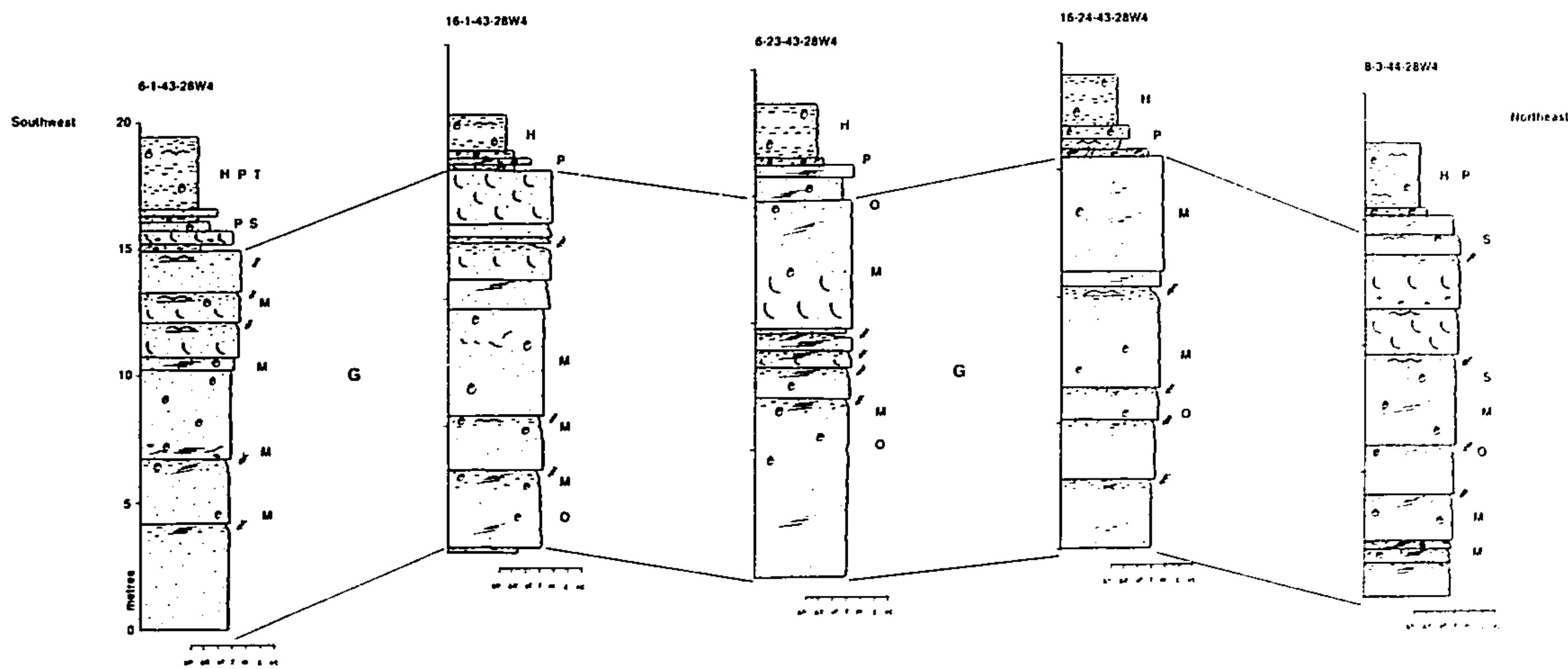


Figure 9.6: Core cross section oriented parallel to dip and following the same line as section S-S'. Note the well-developed interval of cross-bedding in the upper portion of the succession.

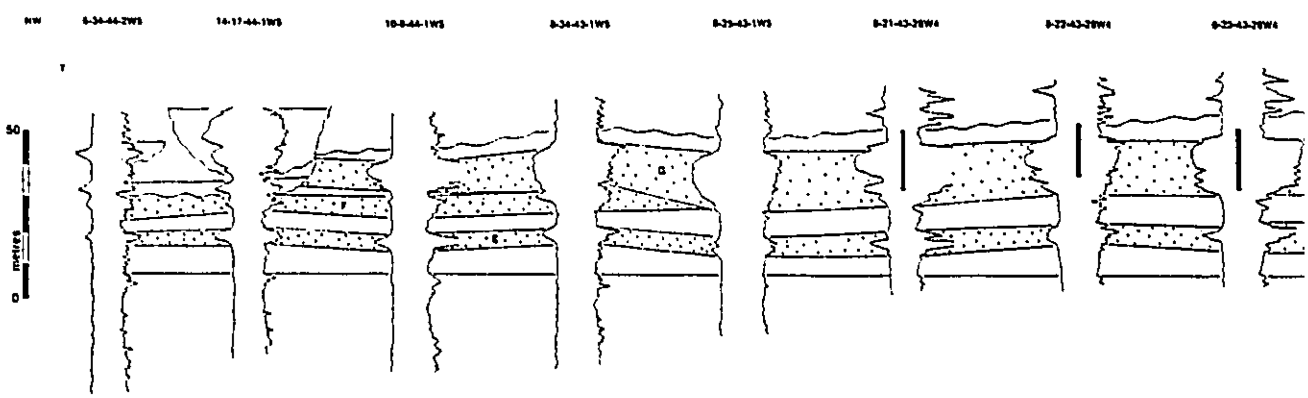
CYCLE G: CORE CROSS SECTION PARALLEL TO DIP



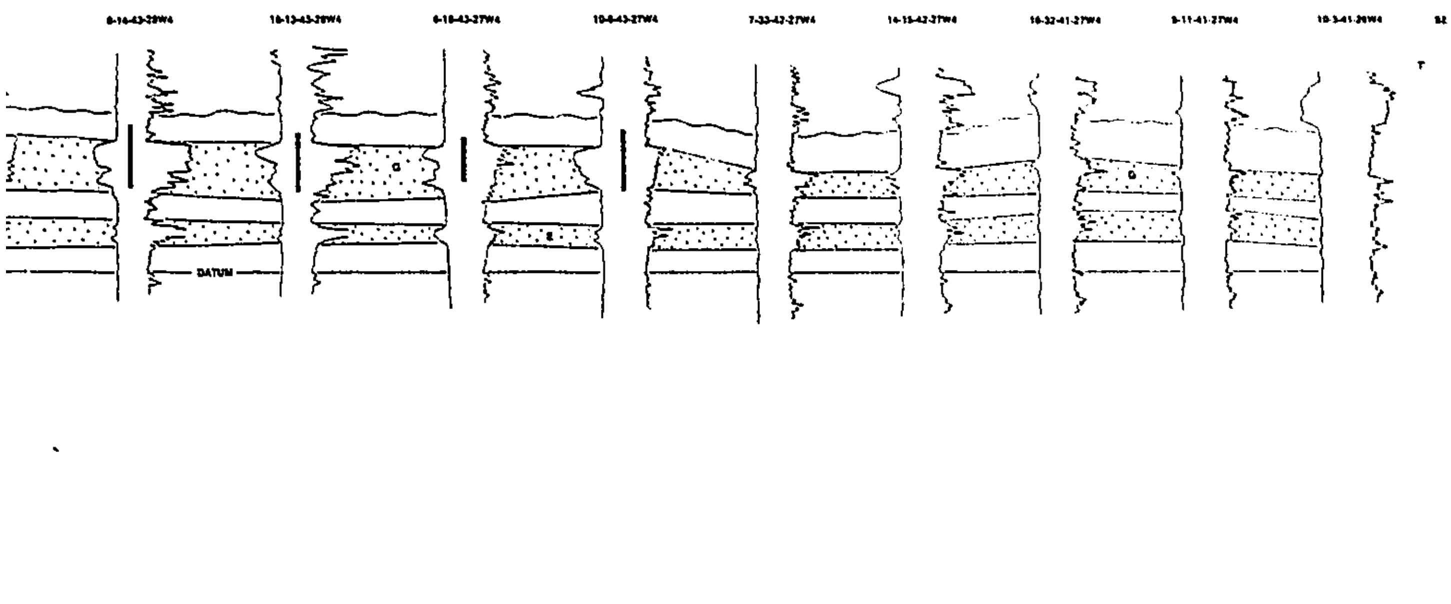
than to definite fluvial channel signatures to the southwest. However, the upper portions of the sandstone in this area may be fluvial rather than shoreline in nature. If true, the tongue of sandstone would not be quite as thick as shown in figure 9.1 and on the cross sections, but the overall geometry of the system would still be the same. The fluvial channels which are incised into the top of Cycle G at 6-35 and 7-7 may be have been feeding the delta in these locations of thicker shoreline successions. Cycle G retains its thickness in excess of 20 m for a further 10 km to the northeast, and just at the southern edge of Ferrybank, it thins to approximately 15-18 m. The log and core cross sections show that within the Ferrybank field, Cycle G is approximately 15 m thick, and the successions in wells 6-1-43-28W4 to 16-3-44-28W4 are all very similar. The log signatures of many wells in the Ferrybank field show the succession to be apparently composed of two distinct units (eg. 6-1-43-28W4, 16-1-43-28W4). The upper unit appears to have much higher permeability and hydrocarbon content than the lower unit. The core cross section (Fig. 9.6) indicates that the succession is often characterized by an increase in grain size (usually lower fine grained sandstone to upper fine-grained sandstone) at the contact between the massive-to-laminated sandstones and the cross-bedded sandstones. A similar feature was observed in Cycle D, where the two units were interpreted as representing separate lobes of the delta system. This may be the case again in Cycle G, but the relationship is not as clear as in Cycle D. Comparison of the core and logs for numerous wells indicates that the contact between the massive-to-laminated and cross-bedded sandstones does not occur at the contact between the two "units" on the well log, but is usually 2-3 metres above this inflection on the well log. The contact between the intervals of high and low hydrocarbon content is lower in the succession, within the upper beds of the massive-to-laminated sandstone. The well log signature seems to be responding to properties other the original physical properties of the sediment. In this case it is most likely a diagenetic factor related to clay or carbonate cements within the sandstones. The succession preserved in Cycle G may be interpreted as a single rather than a composite shoreline succession. There is no clear separation of two "units" within the succession as there is in Cycle D. The upper boundary of Cycle G is similar in most cores in the cross section to the model succession of figure 9.2. There is a 1-3 m transition interval of deeper shoreface sandstones sitting between the cross-bedded upper shoreface/delta front sandstones and the overlying transgressive mudstones.

Log cross section T-T' (Fig. 9.7) is oriented parallel or sub-parallel to strike through Cycle G, and trends across the thick sandstone tongue. Figure 9.8 is the accompanying core cross section. At the northwestern edge of preserved deposition (10-8-44-1W5) Cycle G is only about 10 m thick, and sits sharply on a transgressive unit of mudstone which is between it and the underlying distal sediments of Cycle F. AT 8-34-43-1W5, Cycle G has become thicker and now sits directly on the distal deposits of Cycle F, apparently having eroded away the intercyclic mudstones. This feature was observed in the

Figure 9.7: Log cross section T-T', oriented parallel to strike through Cycle G and trending across the thick central tongue of sandstone. Note the sudden thickness changes on either side of the central tongue.


NW
14-17-44-1W5
10-8-44-1W5
50
metres
0

YCLE G: LOG CROSS SECTION PARALLEL TO STRIKE



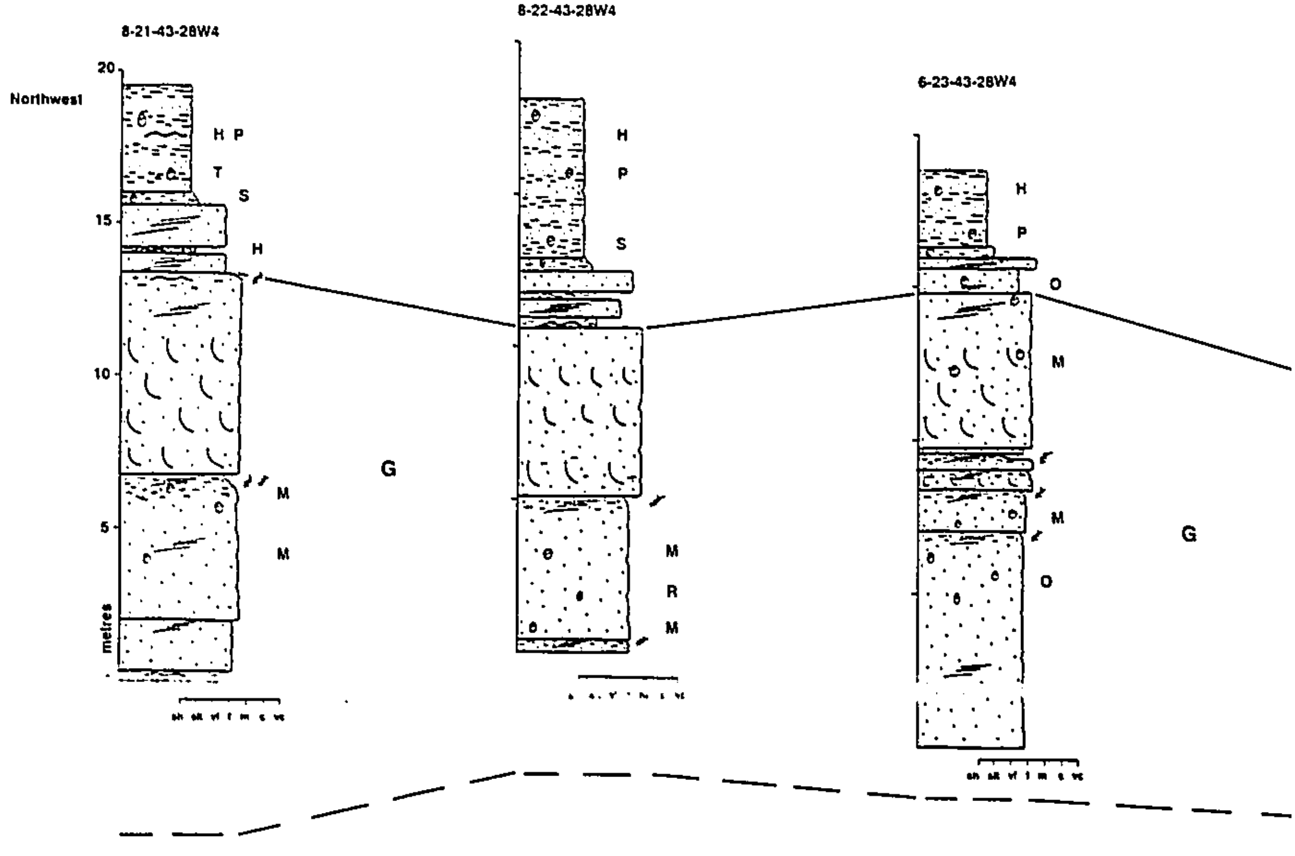
previous chapter on Cycle F. At 8-25-43-1W5, Cycle G is almost 20 m thick, has completely removed Cycle F, and sits directly on the post-Cycle E transgressive sediments. The cycle remains approximately 15-20 m thick across the cored area in Ferrybank field. The core cross section (Fig. 9.8) shows that the succession of Cycle G is very similar throughout this region. The two cores at the southeastern end of the section (6-18 and 10-8-43-27W4) contain noticeably thinner intervals of cross-bedded sandstone. Both of these cores are located on the edge of the thick central tongue of sandstone, whereas the other cores are located in more cental portions of the tongue. This association of thick cross-bedded intervals in central portions of the tongues was also observed in Cycles D and E. Cycle G thins rapidly to the southeast of the central tongue of sandstone to under 10 metres, and then gradually thins into shelf sediments over 10-15 km.

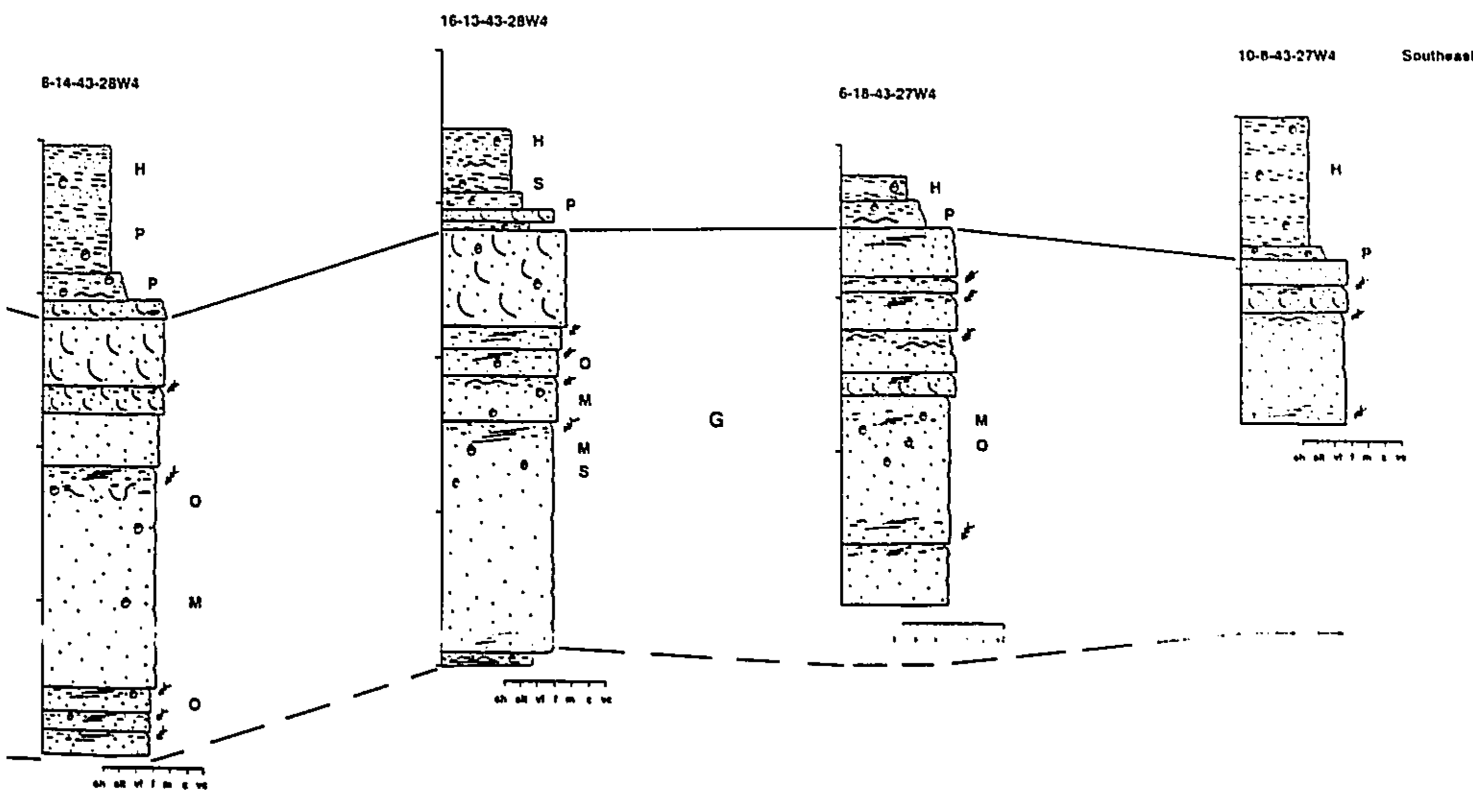
One obvious feature of these cross sections is the lack of fluvial channels in all locations more than a few kilometres basinward of the preserved landward edge of deposition. The sand body is almost entirely marine shoreface in all locations. This separates it from all the other cycles, which show evidence of fluvial incision into the shoreline sediments.

All of the cores in figure 9.8 also show the transition zone from Cycle G into the overlying transgressive mudstones. The interval is thickest and contains the thickest sandstone beds in the northwest, and becomes thinner

Figure 9.8. Core cross section oriented parallel to strike and following the same line as section T-T'. Note how the well-developed cross-bedded interval is not present east of the central tongue (6-18-43-27W4).

CYCLE G: CORE CROSS SECTION PARALLEL TO STRIKE





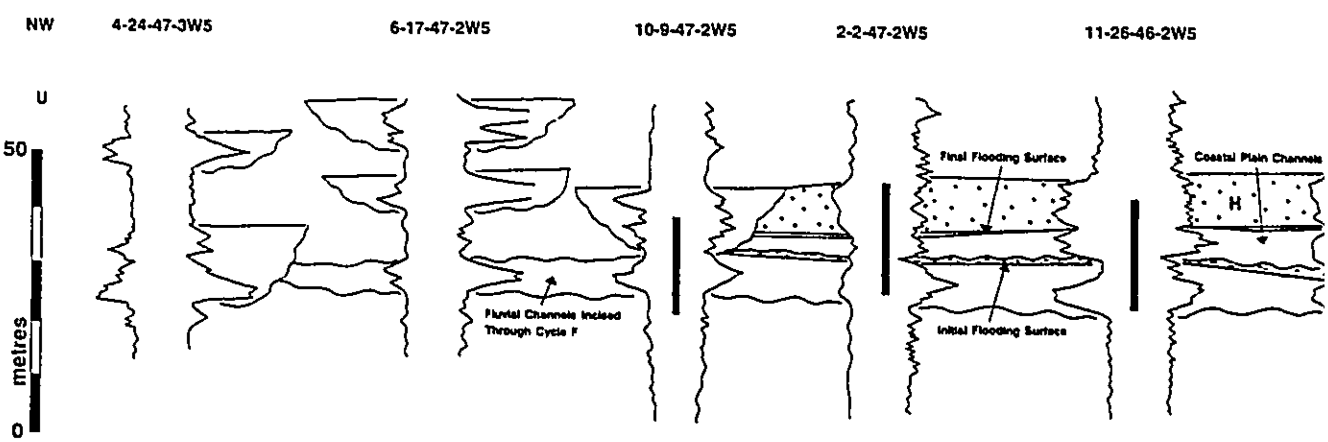
6-14-43-28W4
16-13-43-28W4
6-18-43-27W4
10-8-43-27W4
Southeast
H
P
S
O
M
G
ch slt vf f m c vc

and muddier to the southeast. The transgressive mudstones in all locations are the *Helminthopsis*-burrowed mudstones of Facies Association 1b.

Cycle G is different from all of the other Cycles in this study in that the subsequent transgression seems to have been more extensive than the other transgressive episodes, and its record is preserved in areas which are to the west of the preserved landward edge of Cycle G deposition. Figures 9.9 and 9.10 are log and core cross sections respectively which detail the nature of the post-Cycle G transgression in the location of the Keystone field. Section U-U' is the log cross section (Fig. 9.9). It is oriented WNW-ESE, and is approximately 30 km in length. The basal contact between the Lea Park and the Belly River over much of this section is characterized by fluvial channels which sit directly on shelf mudstones. These channels have eroded into and replaced the deposits of Cycle F in this region. The channels sit at approximately the same stratigraphic level as the deposits of Cycle G (note relationship between 6-16-46-1W5 and 15-12-46-1W5), and may very likely be contemporaneous coastal plain channels which fed the shoreline of Cycle G. The channels were definitely subaereally exposed at times as evidenced by the root traces at their tops in wells 11-26-46-2W5 and 8-26-46-2W5 (Fig. 9.10). These fluvial channels are overlain throughout the section by black mudstones containing abundant oyster shells. The presence of the oysters in this sediment is indicative of brackish or partially saline waters at the time of deposition. In some locations the oyster-rich mudstone is incised into by cross-bedded or LAIS fine grained, bioturbated

Figure 9.9. Log cross section U-U', showing the nature of the post-Cycle G transgressive sediments. The unit denoted by dash marks represents brackish mudstones. See text for discussion.

POST-CYCLE G TRANSGRESSIVI



'E SEDIMENTS: LOG CROSS SECTION

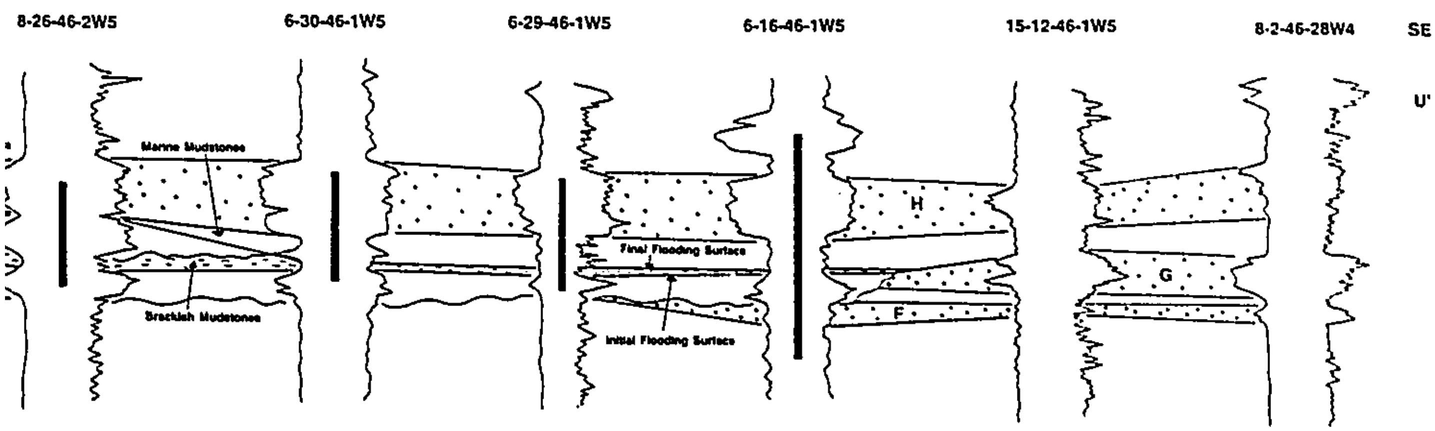
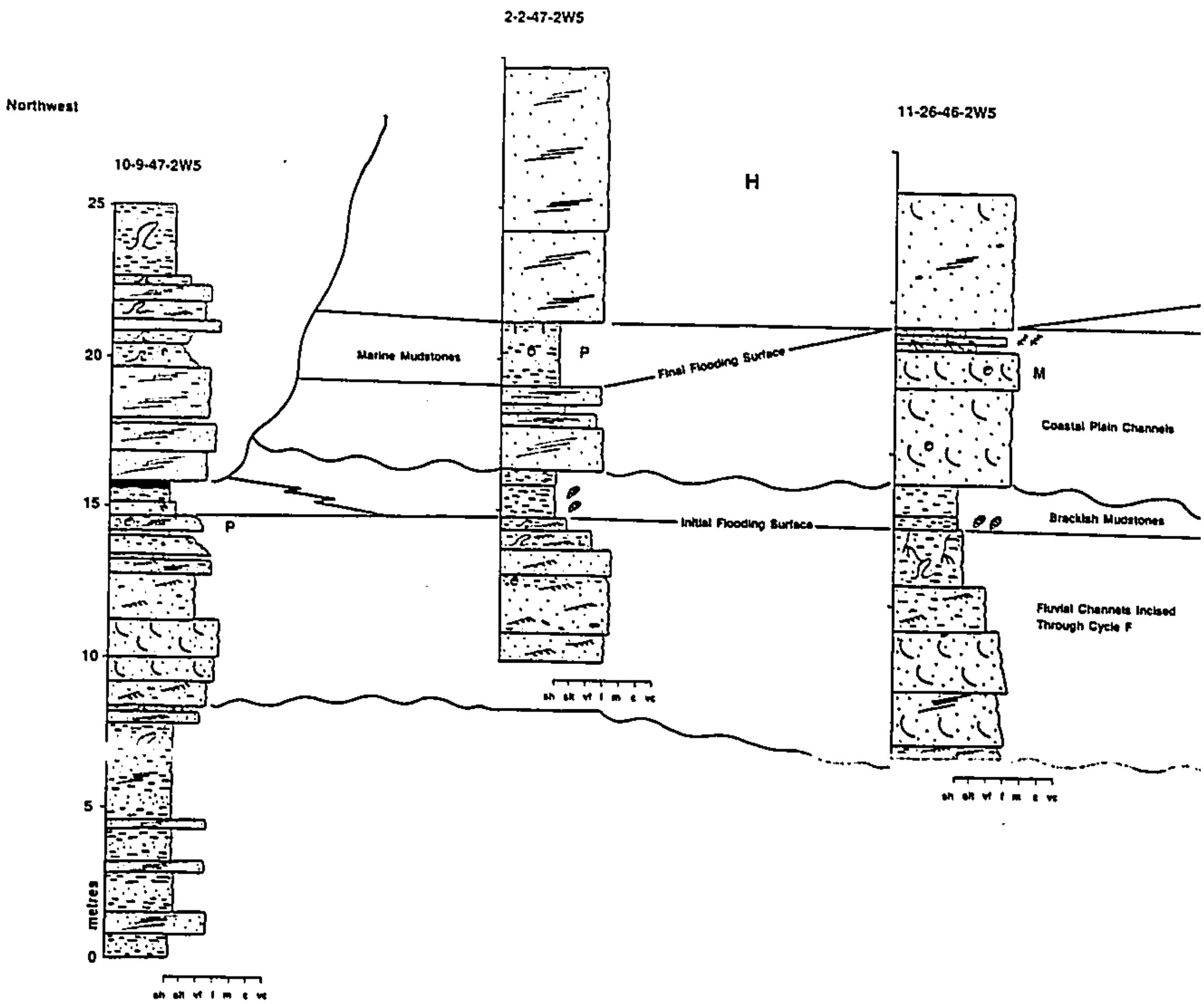
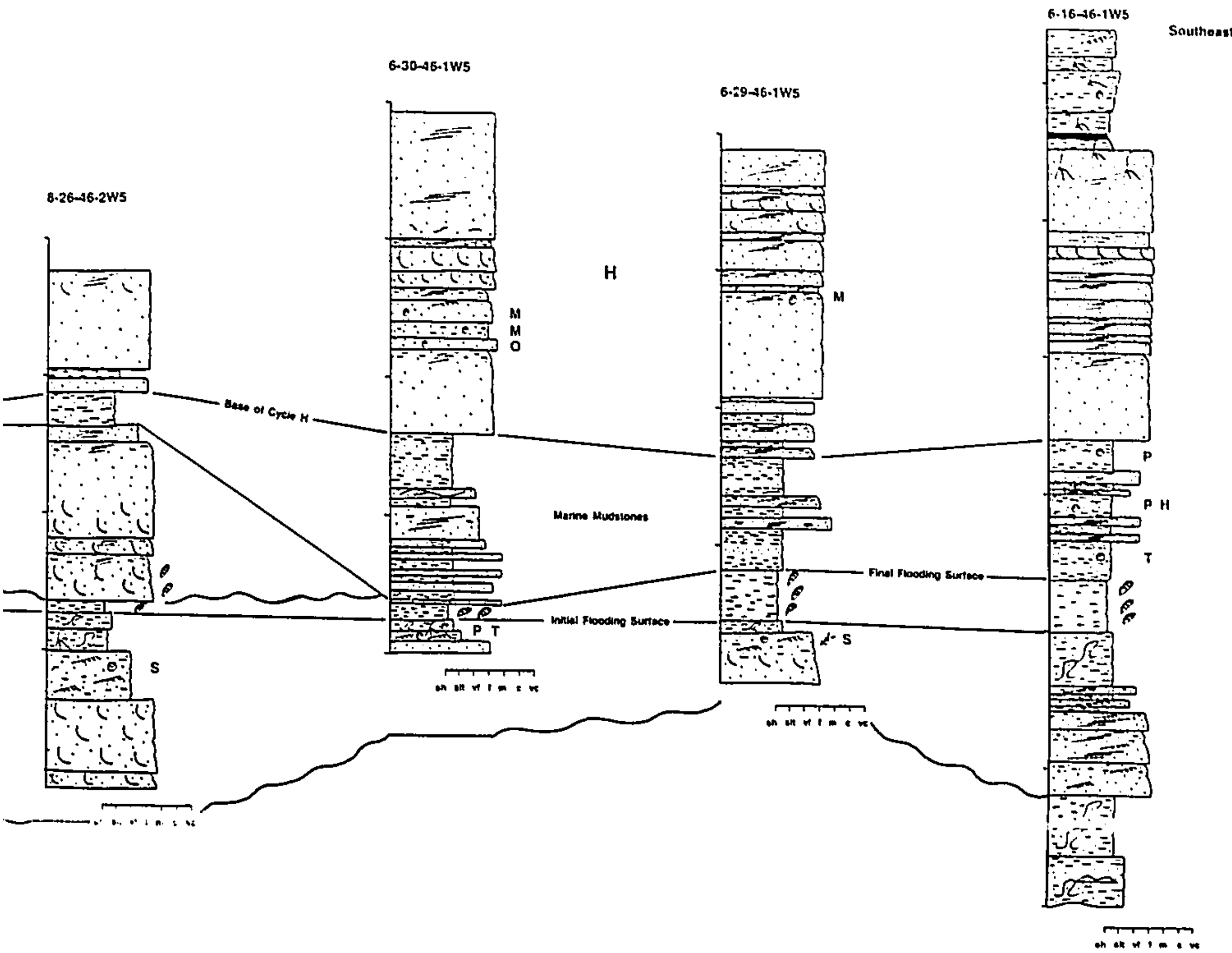


Figure 9.10: Core cross section showing the nature of the post-Cycle G transgressive sediments and following the same line as section U-U'. Note the nature of the two types of flooding surfaces, and the presence of coastal plain channels cutting into the brackish water sediments.

POST-CYCLE G TRANSGRESSIVE SEDIMENTS: CORE CROSS SECTION





8-26-46-2W5
6-30-46-1W5
6-29-46-1W5
6-16-46-1W5
Southeast
H
M
M
O
M
Base of Cycle H
Marine Mudstones
P
P H
T
Final Flooding Surface
Initial Flooding Surface
P T
S
S
sh slt vf f m c vc

channel sandstones (11-26, 8-26-46-2W5). In locations further to the east, the oyster-rich mudstones are sharply overlain by interbedded shelf mudstones, siltstones, and fine-grained sandstones of Facies Association 1a. The lower portion of the shelf sediments interval contains numerous LAIS sandstone beds, which may be of storm origin, although there is no clearly defined HCS within these beds. These beds are less common in the upper portions of the transgressive unit. These deposits also overlie the channel sandstones incised into the brackish mudstones. At the western end of the section, the entire transgressive interval has been eroded by subsequent fluvial incision. This cross section appears to record the deposits of a gradual transgression of a coastal plain environment, which went from non-marine (fluvial channels) to brackish (oyster-rich mudstones with incised channels) to fully marine (shelf mudstones and sandstones).

9.4: Interpretation

Cycle G contains the deposits of a prograding deltaic shoreline, but with some noticeable differences from the deltaic shorelines of the underlying cycles. The dominance of massive-to-laminated sandstones in the lower half of the succession indicates that similar processes were responsible for the deposition of these beds as for the massive-to-laminated beds of other successions. However, interpretation of Cycle G as being an elongate delta lobe similar to those of underlying cyels creates a problem. The elongate sandstone tongue of Cycle G is at least 60-70 km in length, and appears to be almost completely marine in nature, with no channels cutting into it. It is difficult to envisage how this tongue could prograde 60-70 km into the basin without channels following the progradation as it moved basinward. As a result, the origin of the large tongue of sandstone remains problematic. Some speculations on its origin can be made, and these will be discussed in detail in Chapter 11. One possible interpretation is that , unlike the other cycles, Cycle G is not fully preserved, and therefore the geometry of Cycle G is not indicative of its original depositional environment.

The succession preserved in Cycle G shows no definite evidence of subaerial exposure, such as root traces. The cross-bedded shoreface/channel sediments represent the shallowest preserved environment of deposition. Fully-developed or preserved shoreline successions may be present southwest of the Ferrybank field, where the thickest deposits of Cycle G are located. Unfortuantely, no core is available in this area to test this specualtion.

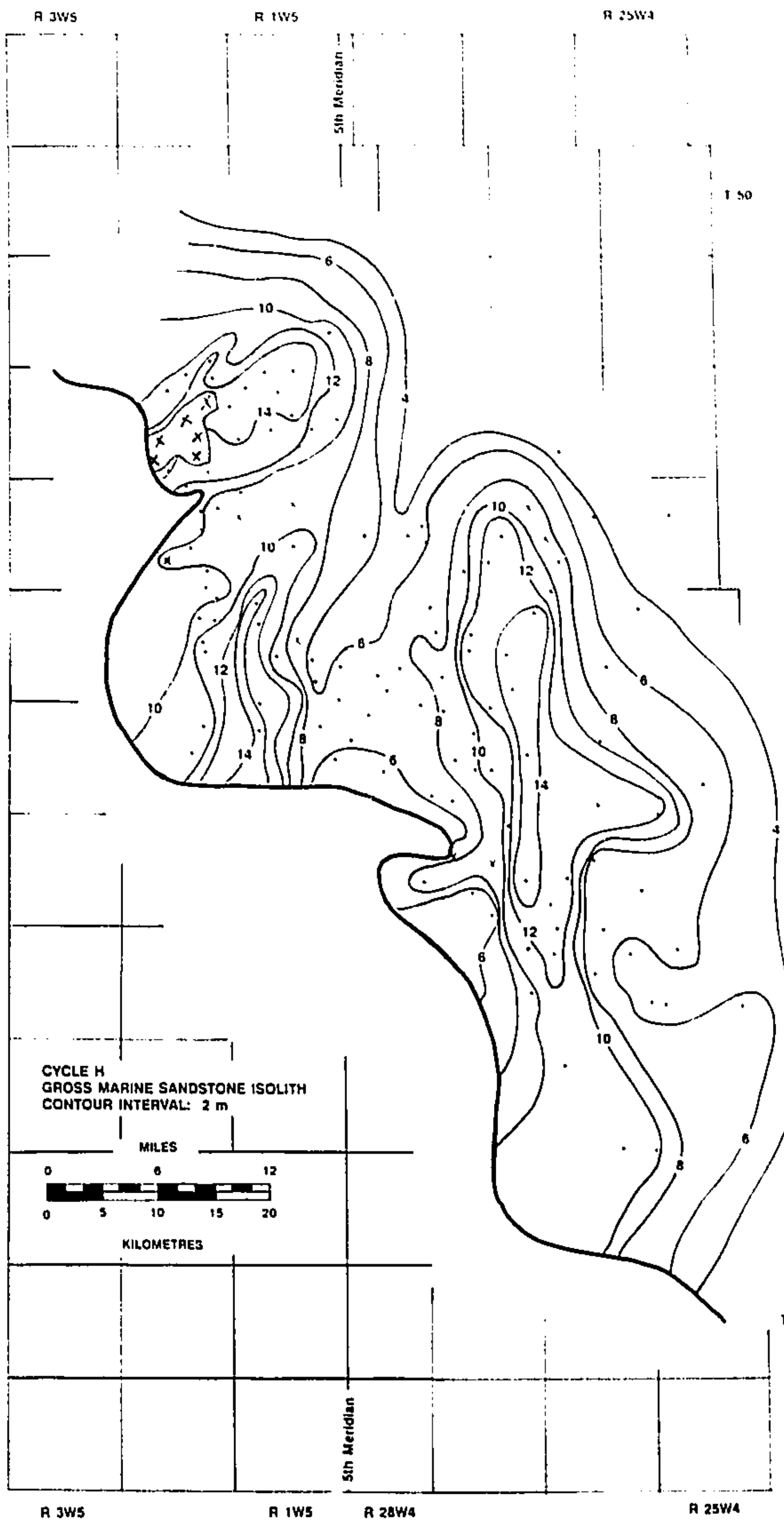
The transgressive deposits which overlie Cycle G show evidence of a more gradual transgression than in the underlying cycles. The marine shoreface successions show a thin fining-upward interval at the top of Cycle G, where the succession changes from upper shoreface cross-bedded sandstones to lower shoreface/shelf LAIS sandstones and finally into shelf mudstones. Sections through the interpreted coastal plain of Cycle G also show a gradual transgression. The initial flooding surface is located at the contact between the fluvial sandstones and the brackish oyster-rich mudstones. This records the initial flooding of the coastal plair In places, this brackish, pond-like environment contained channels which may have been tidal or fluvial in nature, but probably contained a mixture of fresh and saline waters, as evidenced by a more bioturbated nature than the fluvial channels which underlie the brackish mudstones. As the transgression continued, this brackish coastal environment was flooded completely by marine waters, and was replaced by a shallow shelf environment. The contact between the brackish and the open marine sediments is referred to as the final flooding surface. There is no preserved record of the transgressive shoreface sediments.

CHAPTER 10: CYCLE H

10.1: Introduction, Distribution, and Geometry

Cycle H is the youngest of the regressive marine cycles of the Lea Park - Belly River transition within the study area. The regional cross sections showed that it is present in the northeastern portion of the study area, in the vicinity of the Keystone and Ferrybank fields. As with the older cycles, Cycle H strikes in a NW-SE direction, and the sandstone thins to the northeast (Fig. 10.1). It is present along strike for a distance of about 90-100 km and in a dip direction for about 30 km. The plan view morphology of the cycle is characterized by three elongate, tongue shaped lobes of thick sandstone which are oriented normal to the preserved landward edge of deposition. These thick tongues of sandstone are similar in geometry to those identified in the underlying Cycles D, E, and G. The two northern tongues within the keystone region are each about 15-20 km long and about 10 km wide. The northern tongue also contains a central area in which fluvial channels have eroded the cycle. The southern tongue, located in and east of the Ferrybank region is 50-60 km in length and about 10-15 km wide. The sandstone reaches a maximum thickness of about 15 m at the centre of each of the three tongues, and thins to 8-10 m in the regions between the tongues. The 4-6 m contours broadly surround all three tongues, resulting in a sheet-like geometry approximately 30 km wide in all places along the paleoshoreline.

Figure 10.1: Isolith Map of the gross marine sandstone thickness in Cycle H. The contours are in metres and the contour interval is 2 m. The heavy black line represents the preserved landward edge of deposition.

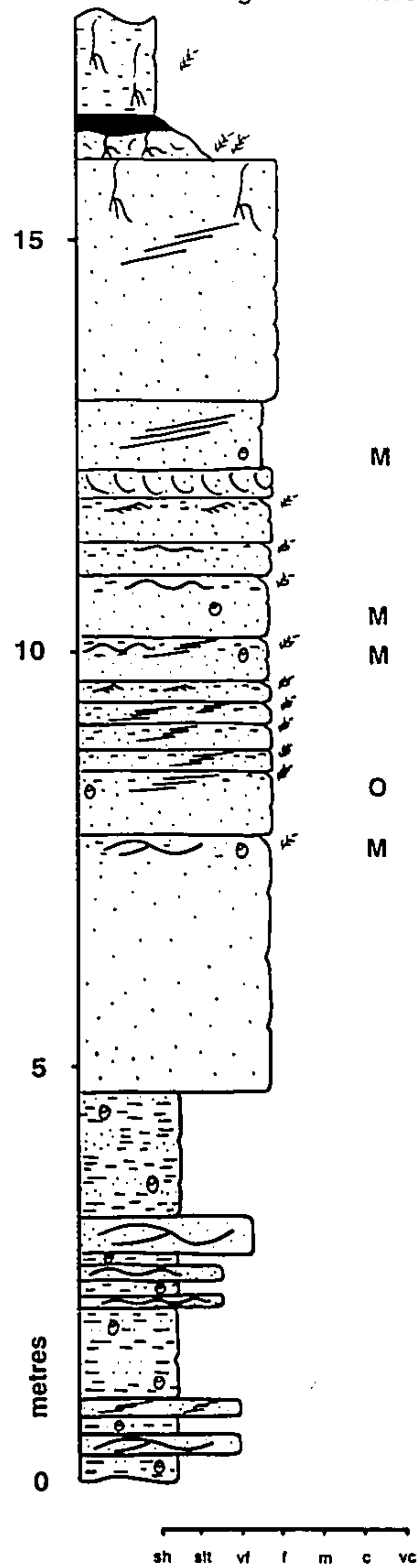

R 3W5
R 1W5
R 25W4
5th Meridian
T 50
T 40
CYCLE H
GROSS MARINE SANDSTONE ISOLITH
CONTOUR INTERVAL: 2 m
MILES
0 6 12
0 5 10 15 20
KILOMETRES
R 3W5
R 1W5
R 28W4
R 25W4

10.2: Facies Associations

Numerous cores through Cycle H are available in the Keystone area. Most of these are located in the region of the middle of the three tongues of sandstone, although a few cores are located within the northern thick tongue. No cores through Cycle H exist for the southern tongue in the Ferrybank region. A typical succession through Cycle H (6-16-46-1W5; Fig. 10.2) shows that the base of the cycle is typically very sharp, with the shoreface sandstones sitting directly on shelf mudstones. This is the same well used to depict the typical succession of Facies Association 2a, and the box photographs of this succession are shown in figure 3.9. Some successions contain a thin, 1-2 m thick transition interval of interbedded LAIS very fine-grained sandstones and shelf mudstones, but most locations show the familiar sharp base. The succession is typical of Facies Association 2a, and is very similar to deltaic successions of the underlying cycles. It is dominated throughout by the massive-to-laminated beds. The cross-bedded interval which typically overlies the massive-to-laminated beds is reduced or absent in Cycle H, being typically 1-2 m thick. The uppermost flat-laminated or LAIS sandstones often contain root traces. Fully developed successions range from 9-15 metres in thickness. *Macaronichnus* is the most common trace fossil present, and is most commonly located within the middle portions of the succession. In all cored examples, Cycle H is overlain by non-marine floodplain deposits of Facies

Figure 10.2: Stratigraphic section through the shoreline succession of Cycle H in well 6-16-46-1W5.

CYCLE H
Regressive Shoreface Succession



Association 5. The transgressive deposits which overlie Cycle H are present only in wells at the very eastern edge of the study area.

10.3: Cross Sections

Three dip-oriented log cross sections were constructed through Cycle H (Fig. 10.3). Cross section V-V' (Fig. 10.4) trends in a dip direction through the northernmost thick tongue of sandstone in the northern Keystone region. The first four wells show fluvial channels at the stratigraphic level of Cycle H. The first is located west of the landward edge of deposition, while the next three are within the swath where Cycle H has been removed. At the southwestern edge, Cycle H is a well-developed, sharply-based 12-14 m-thick succession. It maintains this thickness for about 5-6 km to the northeast, becoming nearly 15 m thick at 14-33-48-1W5. The cycle then abruptly thins to 6-8 m at 10-12-49-1W5, and continues to thin to about 4-5 m at 14-24-49-28W4. The cycle reaches its depositional edge a short distance to the northeast of the end of the section.

Cross section W-W' (Fig. 10.5) trends through the middle of the thick sandstone tongues, where there is the most core data, in the southern Keystone region. Figure 10.6 is the accompanying core cross section. As with the northern tongue, Cycle H in this section is a well-developed 15 m- thick succession immediately northeast of the preserved landward edge of deposition (8-19-45-1W5). The base of Cycle H in this location shows a sharp,

Figure 10.3: Simplified isolith map showing the location of the detailed cross sections used in this chapter.

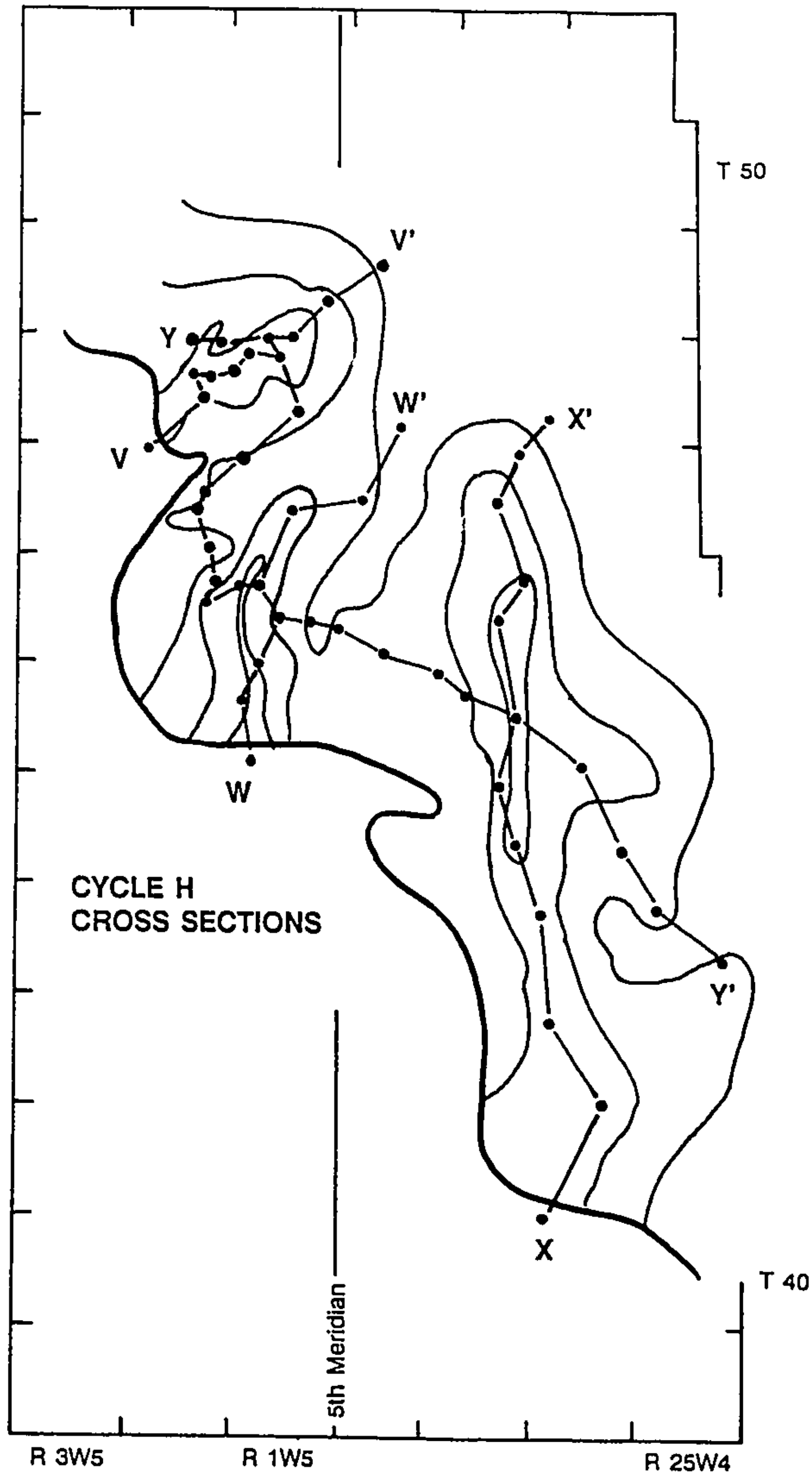
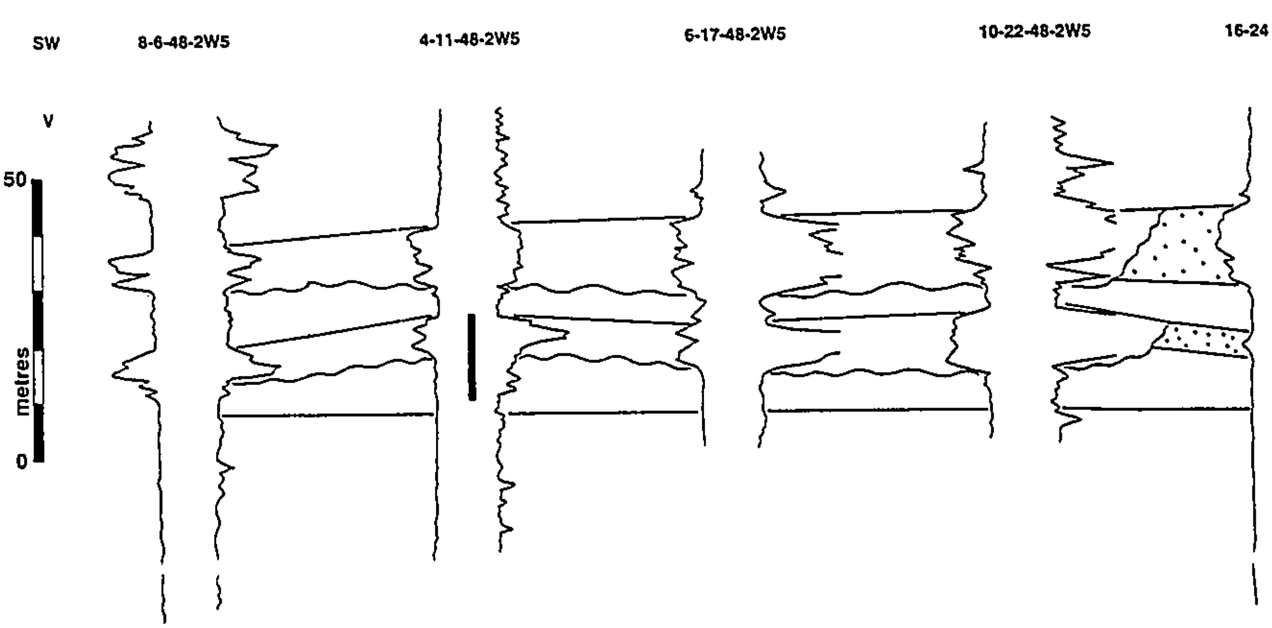

T 50
V'
Y
W'
X'
V
W
CYCLE H
CROSS SECTIONS
Y'
X
T 40
5th Meridian
R 3W5
R 1W5
R 25W4

Figure 10.4: Log cross section V-V', oriented parallel to dip and trending through the thick tongue of sandstone in the northern Keystone region.

CYCLE H



.E H: LOG CROSS SECTION PARALLEL TO DIP - NORTHERN KEYSTONE REGION

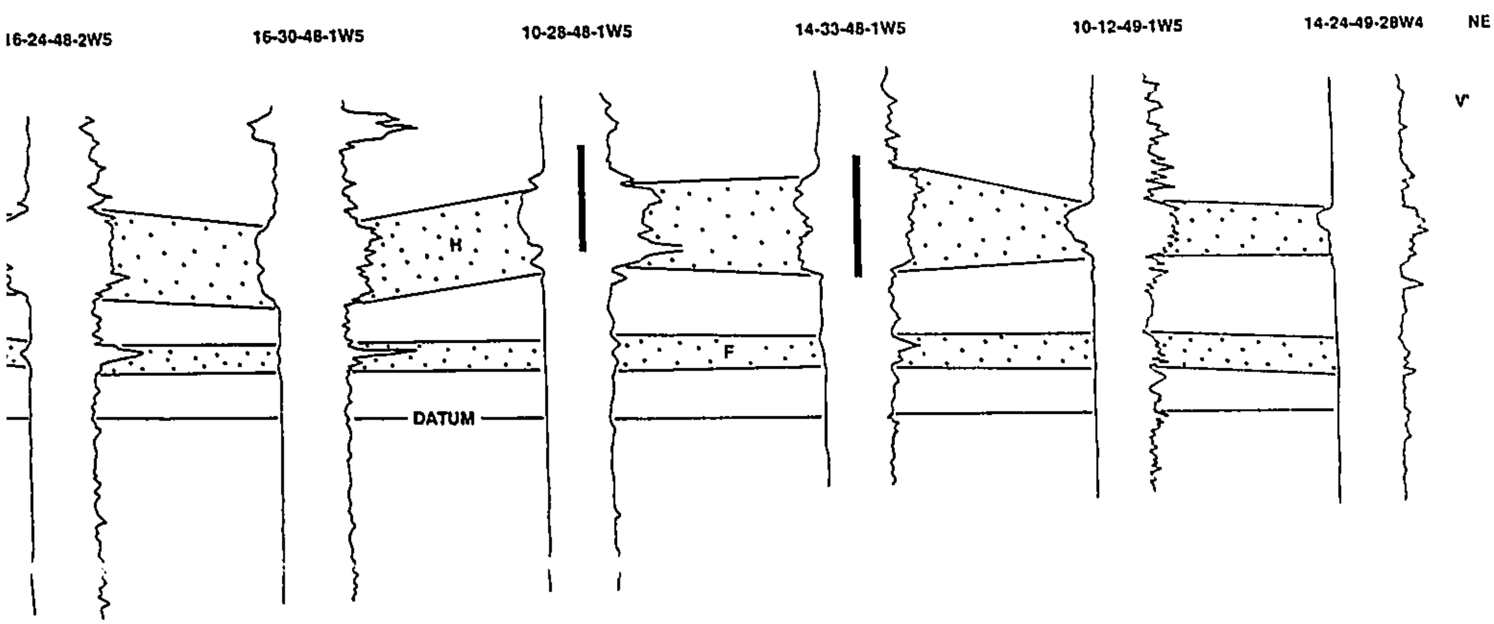
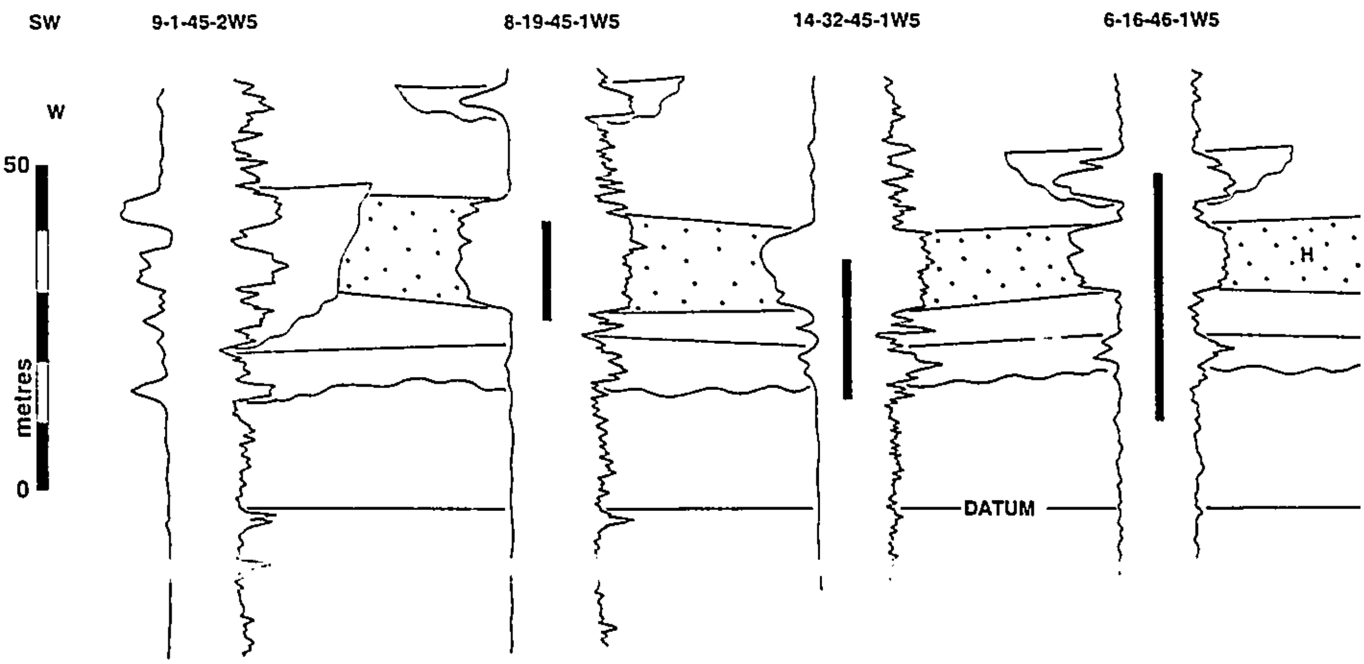


Figure 10.5: Log cross section W-W', oriented parallel to dip and trending through the thick tongue of sandstone in the southern Keystone region.

CYCLE H: LOG CROSS SECTION PAR



ARALLEL TO DIP - SOUTHERN KEYSTONE REGION

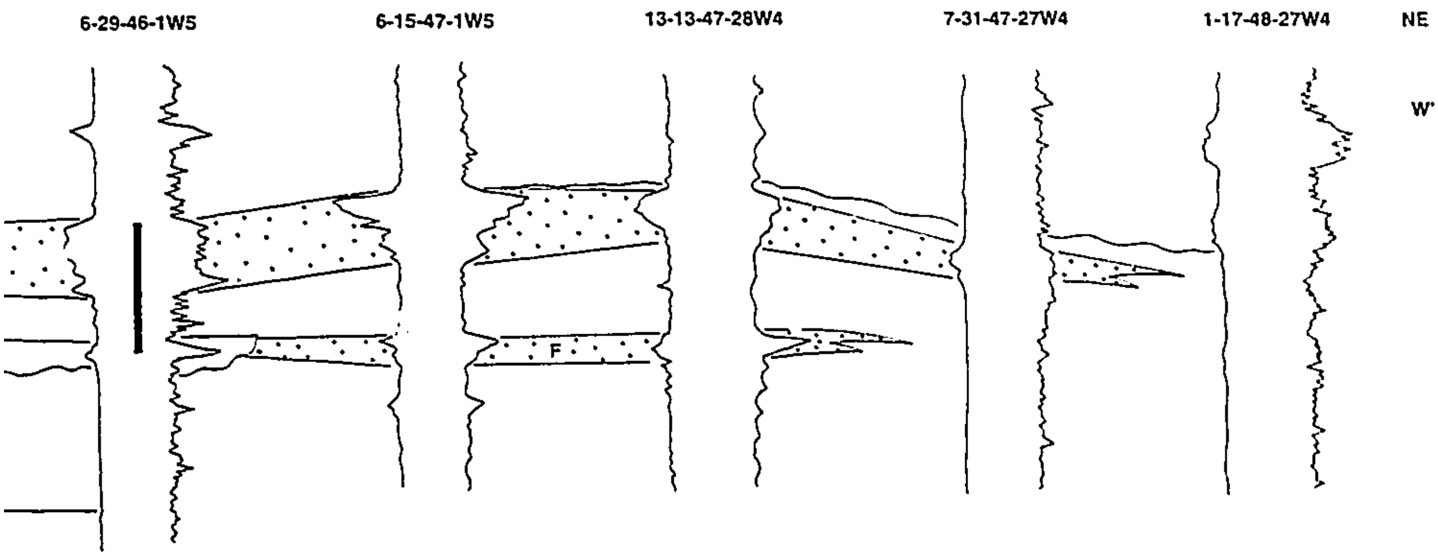
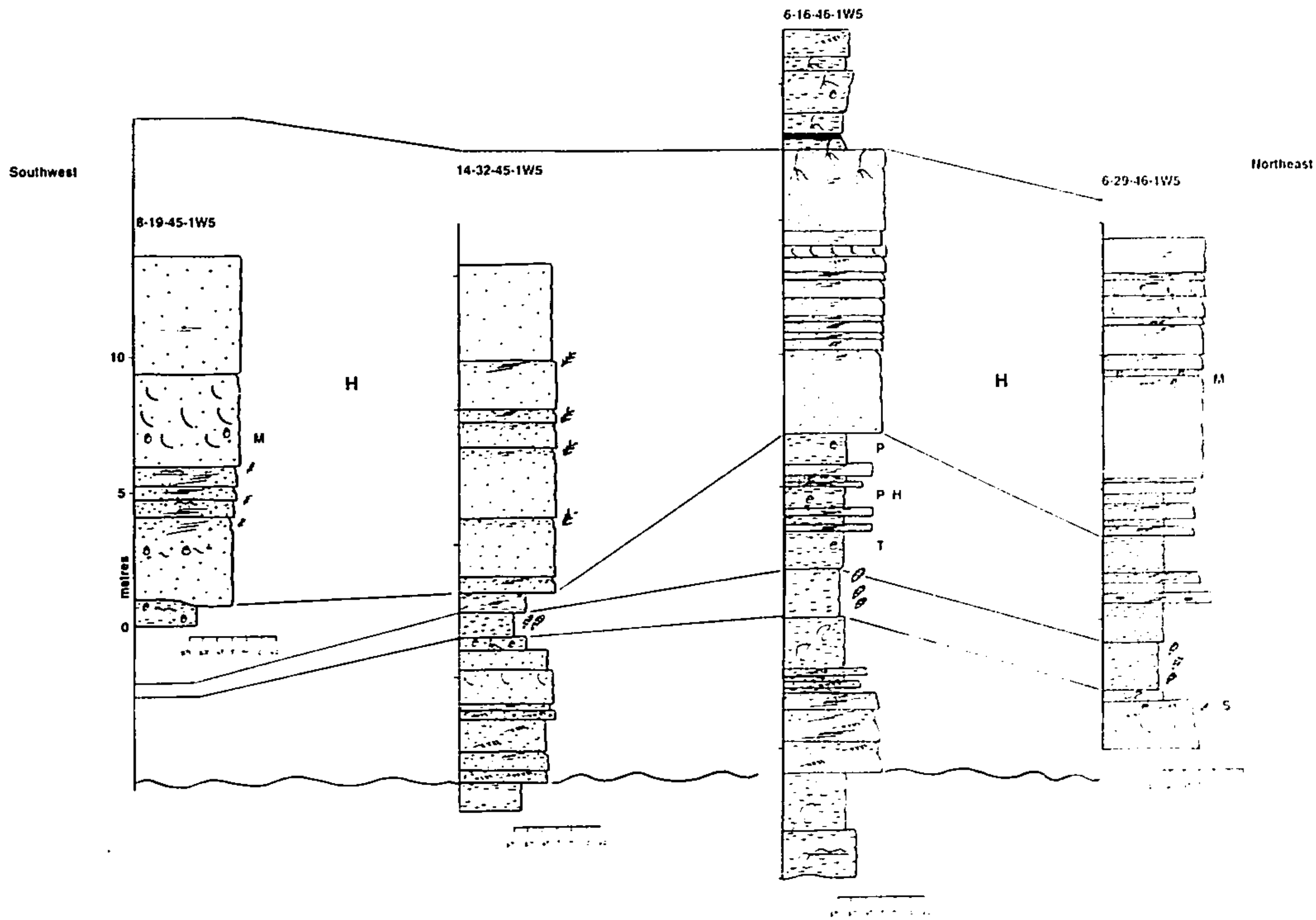


Figure 10.6: Core cross section oriented parallel to dip through the thick tongue of sandstone in the southern Keystone region. The section follows the same line as log cross section W-W'. The units below Cycle H are the post-Cycle G transgressive sediments.

CYCLE H: CORE CROSS SECTION PARALLEL TO DIP - SOUTHERN KEYSTONE REGION

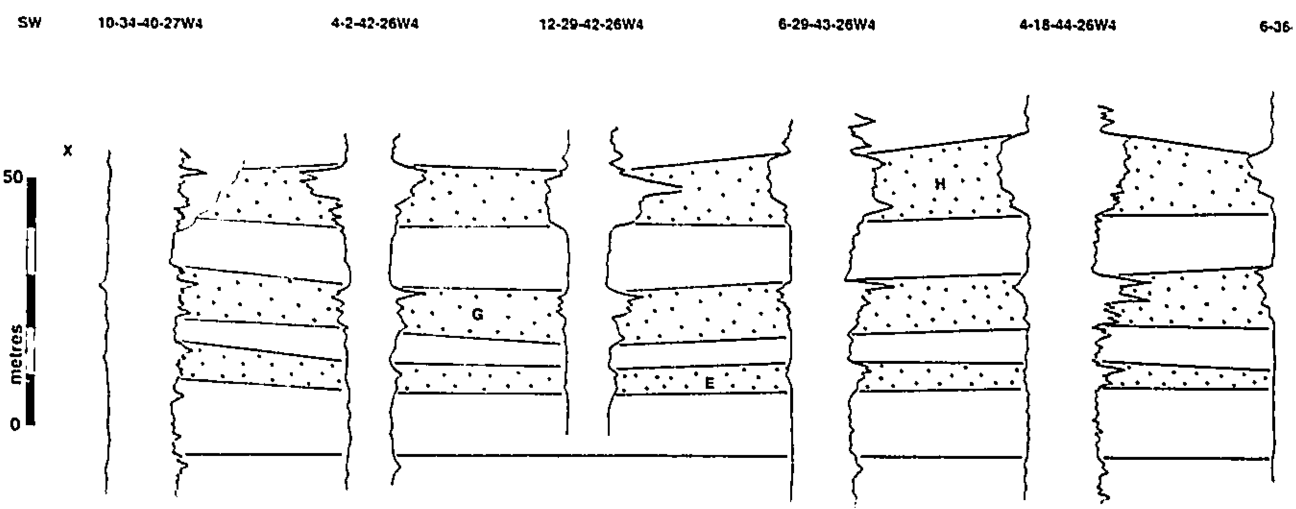


angular contact with the underlying marine mudstones. The cycle gradually thins to about 12 m at 14-32-45-1W5 and 6-16-46-1W5, and to about 11 m at 16-29-46-1W5. All of these wells show typical Facies Association 2a successions, and the bases are very sharp. The underlying transgressive deposits show evidence of the gradual brackish to fully marine transgression of Cycle G discussed in the previous chapter. Cycle H continues to thin to the northeast, and at 13-13-47-28W4 is 8-10 m thick. At 7-31-47-27W4, it is reduced to 5-6 m, and reaches depositional edge a short distance to the northeast of this. A thin unit of transgressive deposits up to 2-3 m thick overlies Cycle H at the northeastern end of the section, and is present as far southwest as 13-13-47-28W4, after which it is removed by fluvial incision.

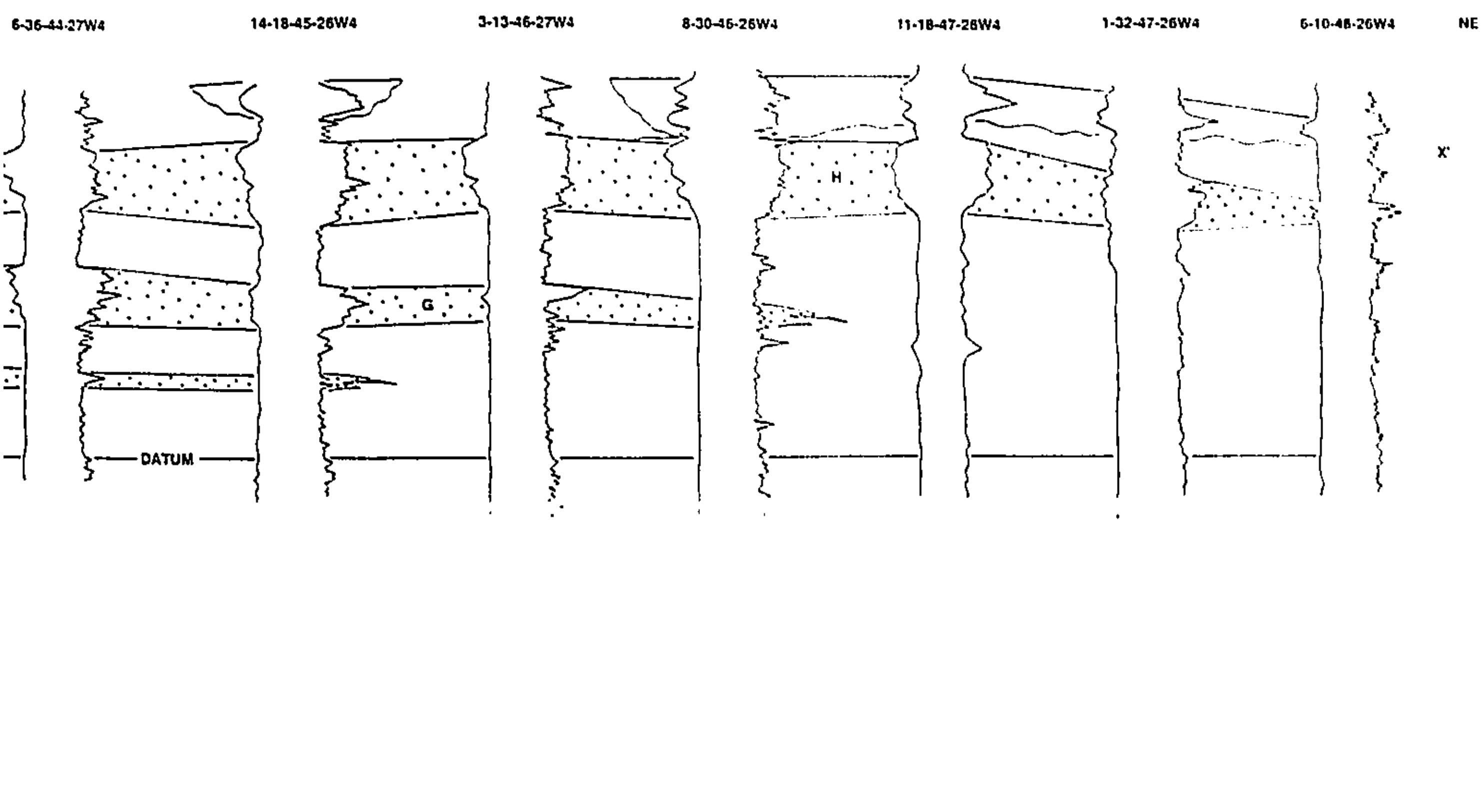
Cross section X-X' (Fig. 10.7) trends through the southernmost thick tongue of sandstone, in an almost true N-S direction. The cycle begins at 4-2-42-26W4 as a well-developed 10-12 m thick succession. It thickens gradually to the north over a distance of 25-30 km, and reaches 15 m at 14-18-45-26W4. It then thins gradually back to 10-12 m over a distance of 10 km, before abruptly thinning to 8 m at 1-32-47-26W4 and to 3 m at 6-10-48-26W4. It reaches depositional edge a short distance to the north of the end of the section. This lobe is significantly larger than the northern lobes, and like the thick tongue in Cycle G, contains no evidence of channel sandstones. The data base for this lobe is admittedly poor, with no cores and very few well locations at the southern end, where channels may be present. The transgressive deposits

Figure 10.7: Log cross section X-X', oriented parallel to dip and trending through the thick sandstone tongue east of Ferrybank.

CYCLE H: LOG CROSS SE



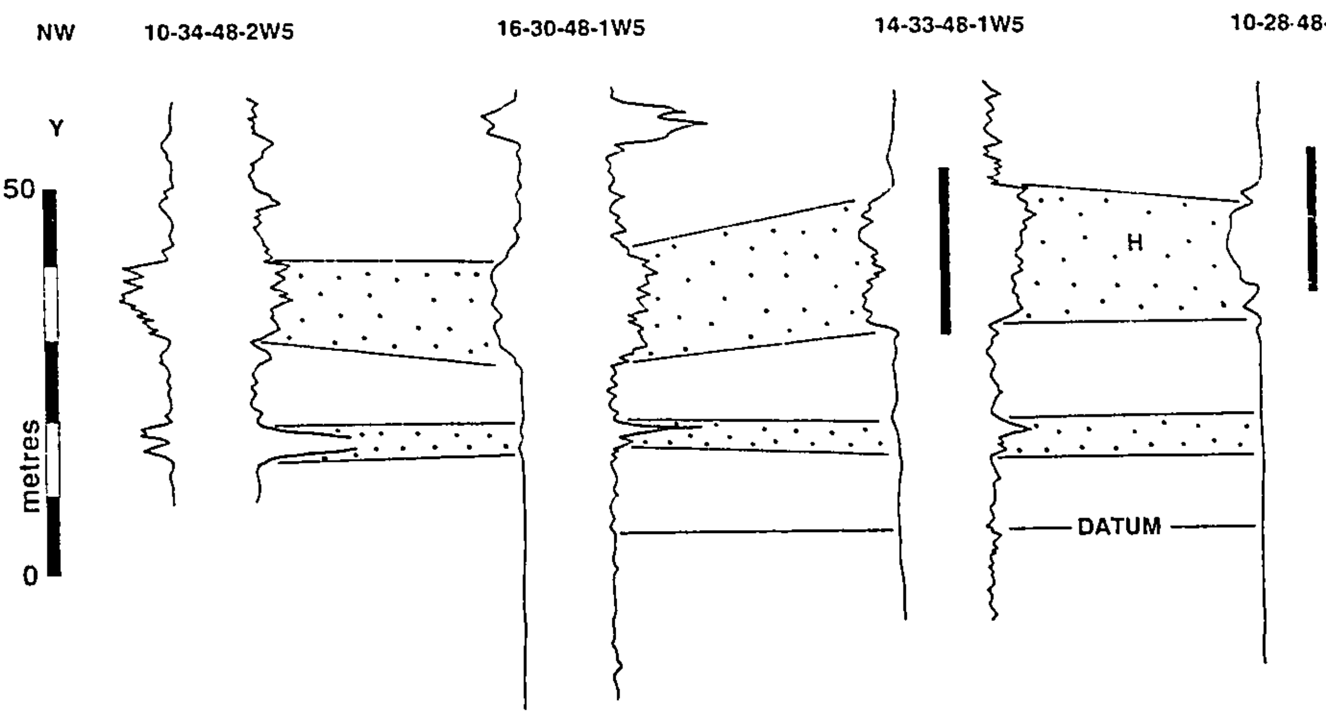
; SECTION PARALLEL TO DIP - FERRYBANK REGION

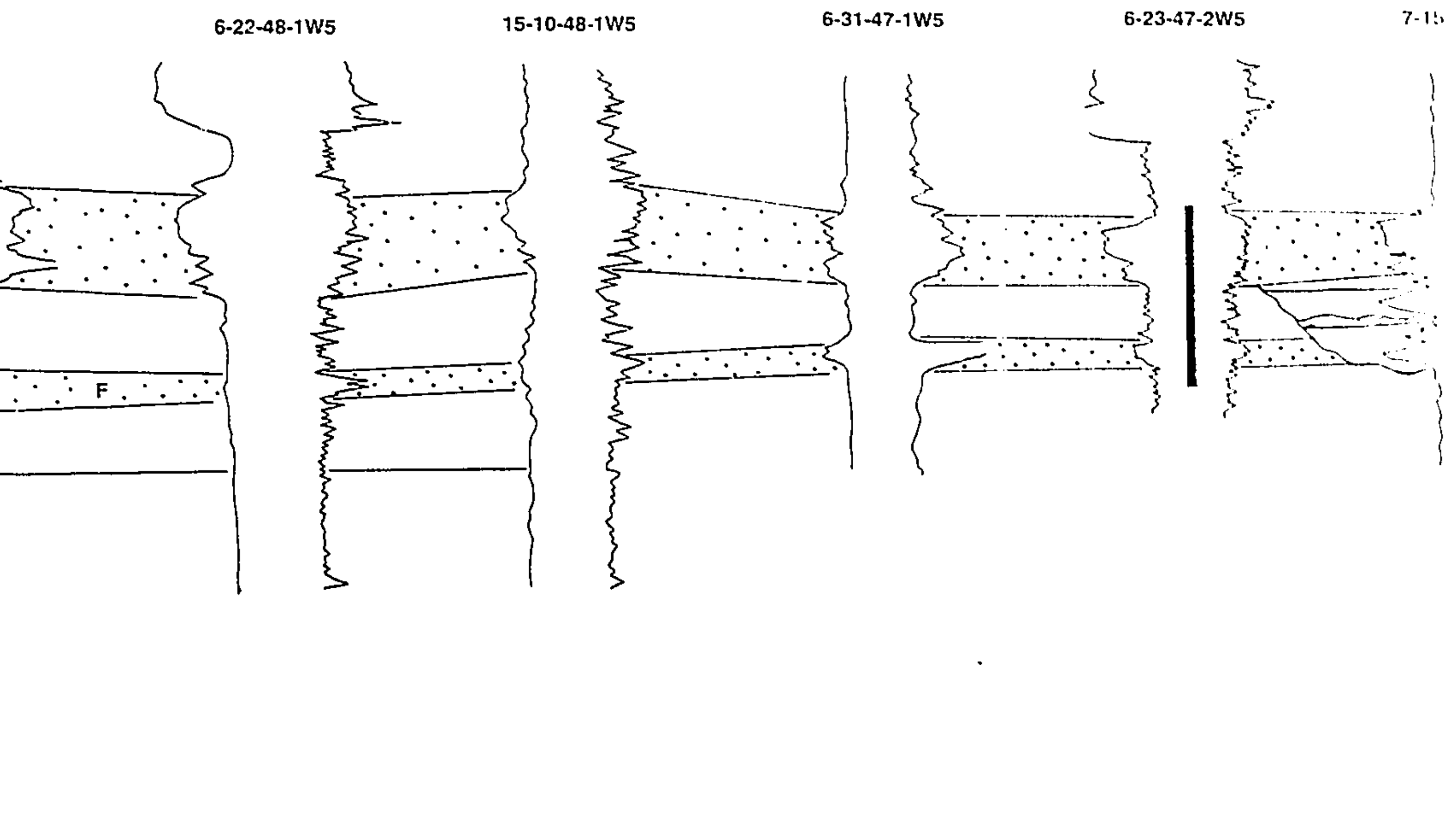


overlying Cycle H are up to 10 m thick at the northern end of the section, and thin to the south to about 3-4 m at 11-18-47-26W4. South of this location, the transgressive unit is removed by fluvial erosion.

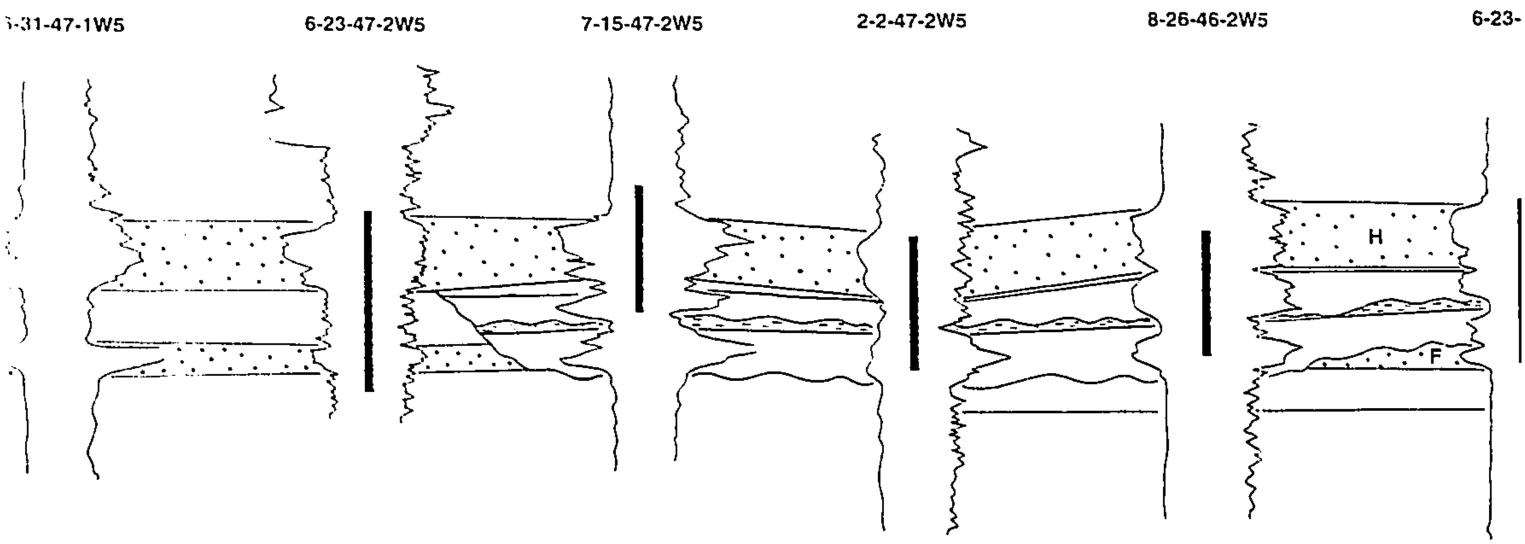
Cross section Y-Y' (Fig. 10.8) is approximately 90-100 km long, oriented in a strike direction, and passes through all three of the thick sandstone tongues. Figure 10.9 is the accompanying core cross section. The section begins just to the northwest of the thick tongue in northern Keystone, where Cycle H is 10-11 m thick. It thickens across the central portion of the northern Keystone tongue to 15 m (14-32 and 10-28-48-1W5). The cycle then thins to a minimum of 8 m in the area between the two tongues in Keystone (6-23-47-2W5). Between 6-23 and 7-15-47-2W5, the transgressive sediments underlying Cycle H and the distal sediments of Cycle F are removed by fluvial erosion, as was documented in the previous chapter. The surface of incision is not evident in 6-23, and is interpreted to have been removed by the later progradation of Cycle H in this location. This implies that there was at least 10-12 metres of incision associated with this downcutting event which removed Cycle F in this region. At 7-15, the transgressive sediments underlying Cycle H are the brackish coastal plain deposits, as evidenced by the coastal plain channel at the base of the core (Fig. 10.9). Cycle H itself is still relatively thin (9 m), but fully developed in this location. For the next 10 km, the section crosses the middle of the thick tongues of sandstone, and is essentially the same section as shown in section U-Ú' (Figs. 9.9, 9.10) of the previous chapter. Cycle H

Figure 10.8: Log cross section Y-Y', oriented parallel to strike through Cycle H.

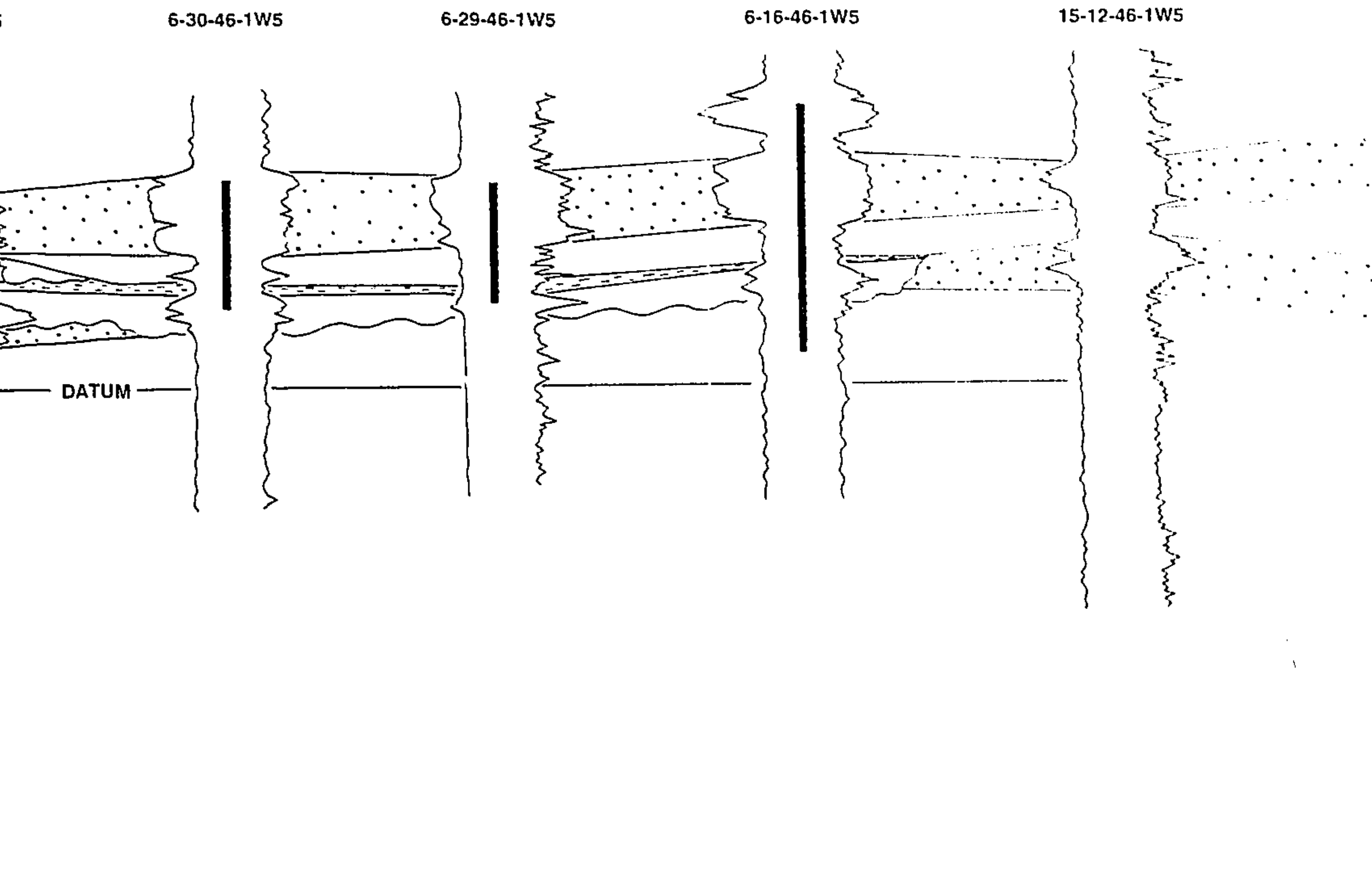

NW
10-34-48-2W5
16-30-48-1W5
14-33-48-1W5
10-28-48·
Y
50
0
metres
H
DATUM

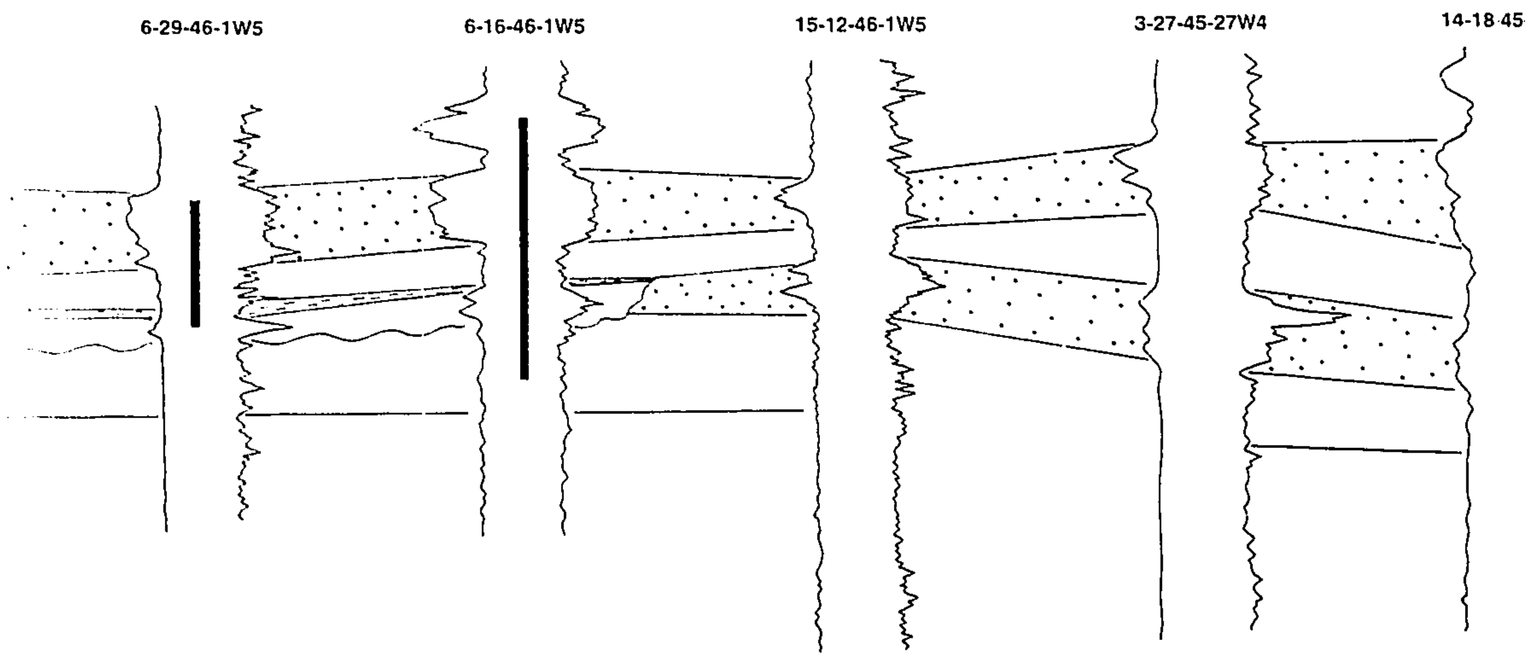

6-22-48-1W5
15-10-48-1W5
6-31-47-1W5
6-23-47-2W5
7-15
F

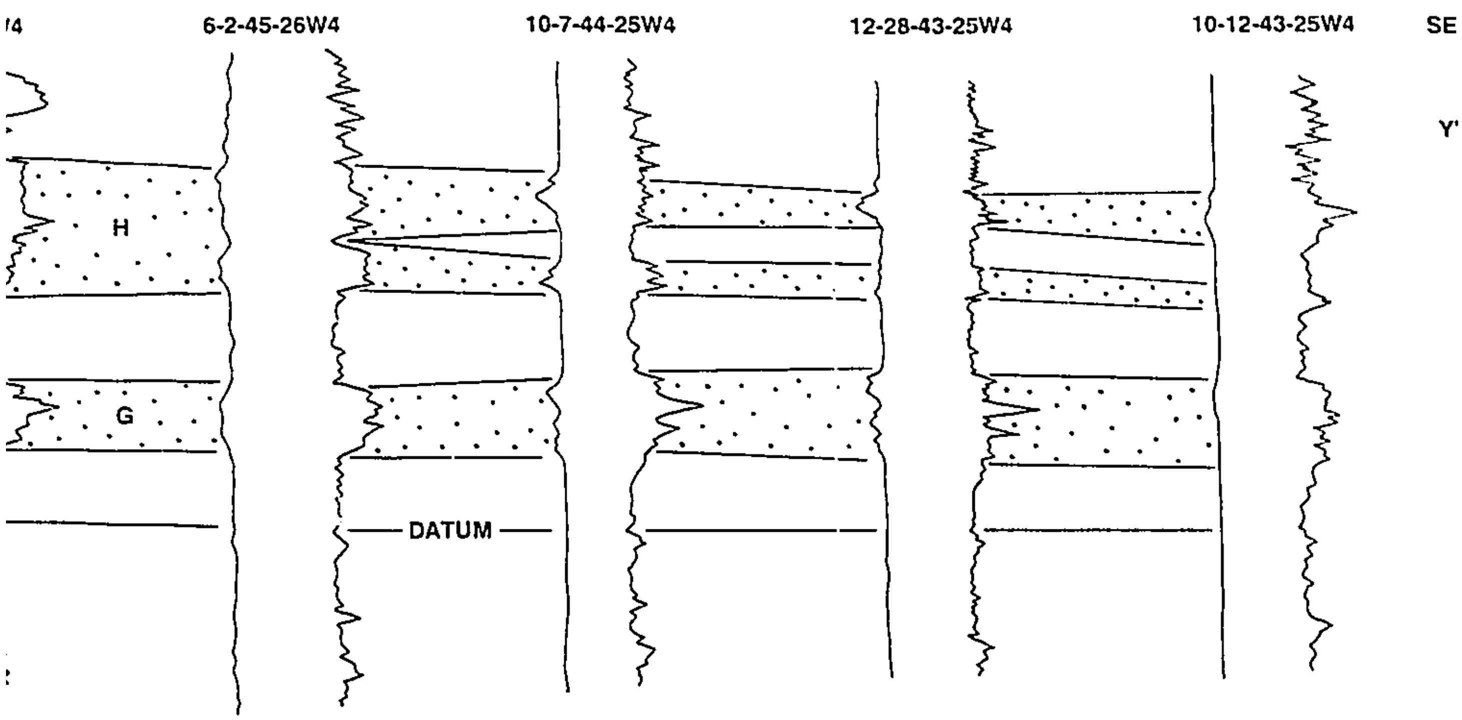
CYCLE H: LOG CROSS SECTION PARALL



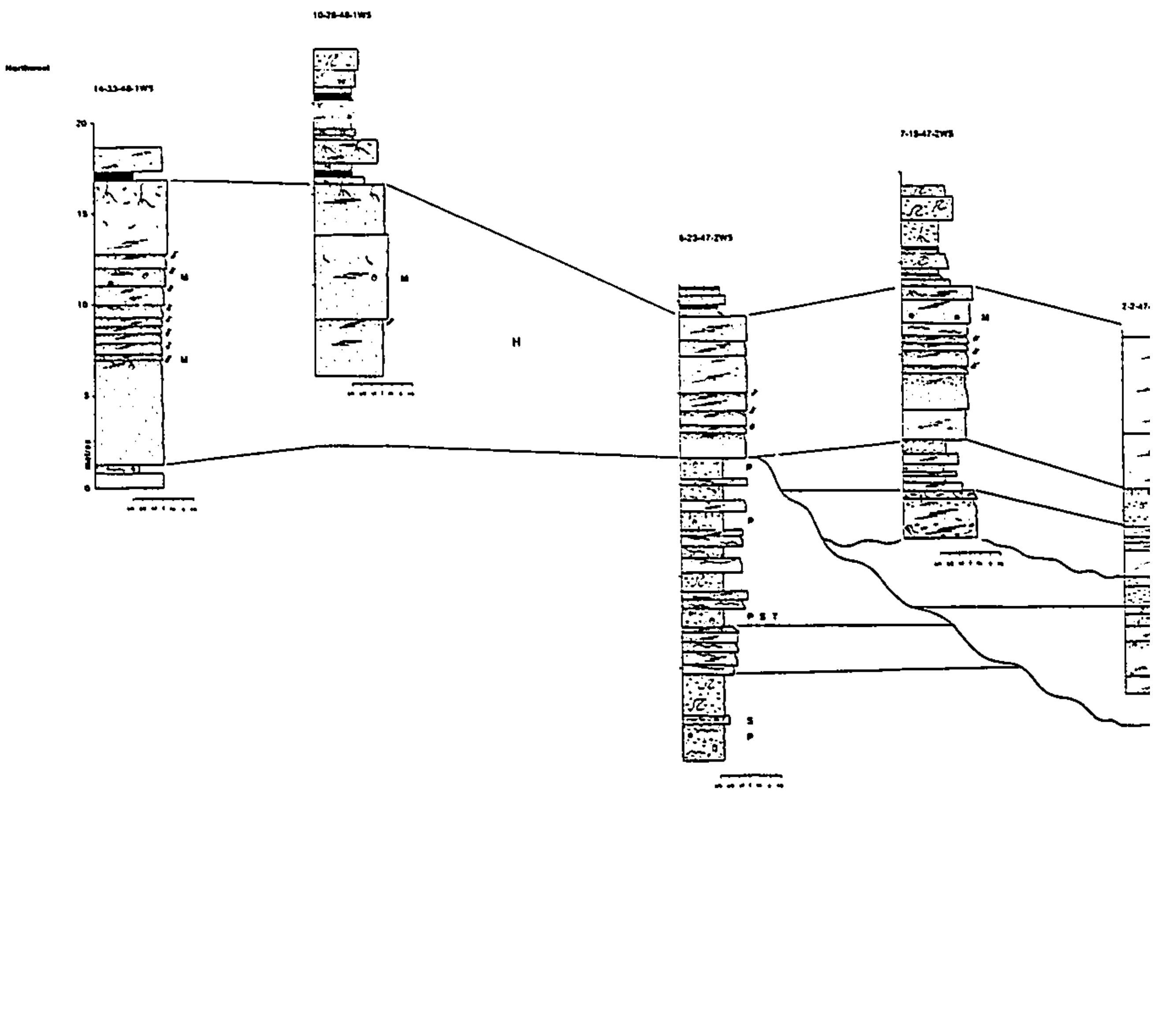
O STRIKE

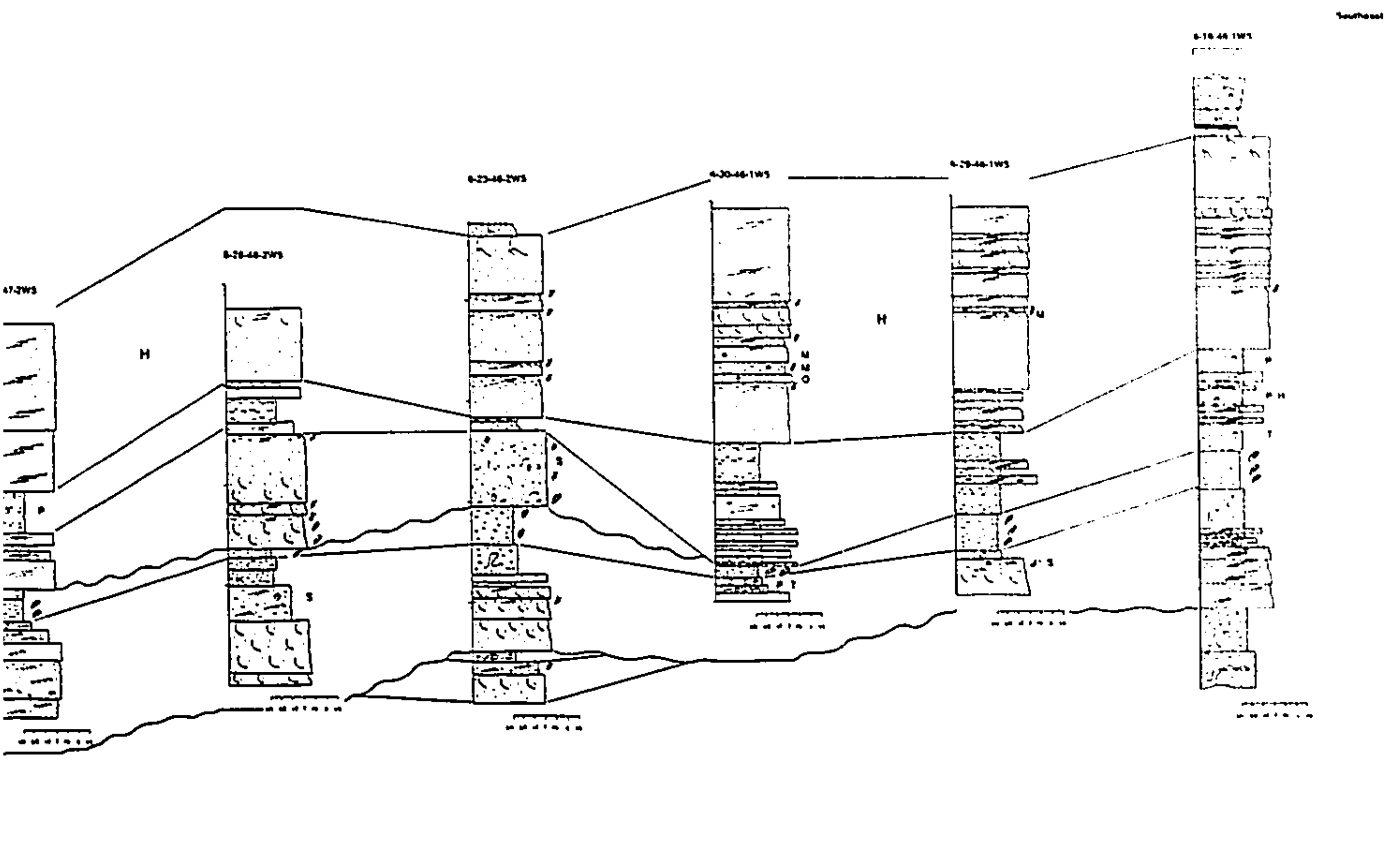



6-29-46-1W5
6-16-46-1W5
15-12-46-1W5
3-27-45-27W4
14-18-45


/4
6-2-45-26W4
10-7-44-25W4
12-28-43-25W4
10-12-43-25W4
SE
Y'
H
G
DATUM

CYCLE H: CORE CROSS SECTION PARALLEL TO STRIKE





Southeast
H
H
S

thickens to about 13 m at 6-30-46-1W5, and is still 12 m thick at 6-16-46-1W5. All the wells in this region contain the brackish to fully marine transgressive sediments underlying Cycle H. South of the southern Keystone thick tongue, Cycle H thins to a minimum of 6-7 m at 15-12--46-1W5, before thickening again to 14-15 m at 14-18-45-26W4 across the central portion of the southern tongue in the Ferrybank region. To the southeast of this, Cycle H begins to split into two sub-units separated by up to 5-6 m of muddier shelf sediments. The lower unit may be a continuation of the Cycle H sediments located to the northwest of the southerly thick tongue, while the upper unit would seem to be restricted to this region. The log signature within the southerly tongue (14-18-45-26W4, 6-2-45-26W4) also shows two distinct units, indicating that the thick tongue in this region may be related to the progradation of the upper unit, stacked on top of the more distal deposits of the lower unit. No transgressive sediments overlying Cycle H are present along this section.

10.4: Interpretation

Cycle H is interpreted to contain the deposits of a deltaic shoreline system similar to those described in the underlying cycles. The cycle rests sharply on the transgressive deposits overlying Cycle F in the northern Keystone region, and on the transgressive deposits overlying Cycle G in southern Keystone and areas southeast of this. The succession is dominated by the massive-to-laminated beds throughout, with a very thin cross-bedded upper shoreface interval. The plan view geometry of the system indicates that three distinct deltaic lobes were constructed within the study area. All three lobes are elongate and shore-normal. This indicates that there was little reworking of the proximal sediments by basinal processes. The temporal relationships between the three lobes is unclear, especially for the two lobes in the Keystone region. The southernmost lobe may be the youngest of the three, as indicated by the strike cross section Y-Y' (Fig. 10.9). The channels which incise into the northernmost lobe are likly related to the progradation of the lobe itself, and are therefore interpreted to be coeval to the lobe. The rapid thinning of the lobe 5-6 km northeast of 14-33-48-1W5 in section V-V' indicates that these locations were probably seaward of the most basinward position of shoreline progradation, and do not represent fully developed shoreline successions. The southern lobe, which contains no definite evidence of channels cutting into the lobe is, like Cycle G, problematic to interpret. No cores exist to determine the facies associations present in the this tongue. This is critical to understanding its origin, and core data from this region would be required to answer this question to any degree of satisfaction.

The transgressive sediments overlying Cycle H are not present throughout most of the study area, and are never cored. The coastal plain sediments overlying Cycle H in the Keystone region contain no examples of partially saline ponded sediments which may the landward edge correlative deposits of the transgression, as were seen in some of the underlying cycles.

CHAPTER 11: ANALYSIS OF DELTAIC AND FLUVIAL SYSTEMS

11.1: Introduction

The shoreline facies associations discussed in chapter 3, and the lateral facies relationships and geometries of the shoreline cycles discussed in chapters 5-10 led to the interpretation of most of the shoreline cycles as deltaic. The variety of fluvial channelized sandstone facies associations was also discussed, and brief interpretations were made of the fluvial environment of deposition. This chapter will attempt a more detailed discussion of the sedimentary mechanisms responsible for deposition within both the deltaic and the fluvial systems.

11.2: Lea Park - Belly River Deltas

11.2.1: Introduction

The deltaic interpretation is based on the observed sedimentary successions and the sand body geometries. The sedimentary succession which comprises the deltaic cycles (Facies Association 2, Chapter 3) is best interpreted as a prograding shoreline environment. At the base, shoreface sandstones lie sharply on marine mudstones, and at the top the succession shows evidence of subaerial exposure. The sand body geometries of Cycles D-H (Chapters 6-10) do not indicate straight shorelines, but are regionally restricted, lobate features with discrete, shore-normal or shore-oblique "tongues" of thick sandstone. The sand bodies are all approximately 100 km in a shore-parallel direction, and 25-50 km in a shore-normal direction. The thick shore-normal to shore-oblique sandstone tongues tend to be approximately 10-30 km long, and 5-15 km wide. These lobate shoreline cycles are laterally equivalent to channelized fluvial sediments to the west. Each shoreline cycle is therefore interpreted to be indicative of a discrete protrusion of the shoreline due to fluvial input of sediment, and therefore fits the definition of a delta (Elliott, 1986).

However, there are numerous aspects of the facies associations and geometries of Lea Park - Belly River deltas which make these systems distinctly different from any other delta system, modern or ancient, which has been described in the literature.

11.2.1.a: Sedimentary Successions

The sedimentary successions are unlike those of any other previously described deltaic succession. They are dominated by fine-grained sandstone, but there is a general absence of sedimentary structures. The shoreline successions are dominated by thick beds of structureless sandstone which grade upward into poorly-defined LAIS sandstones. These beds appear to be up to 7 m in thickness, although at least some of these thick beds are amalgamations of more than one bed. These massive-to-laminated beds comprise the lower 50%-80% of the shoreline successions in the cycles.

Trough cross-bedding is generally only present within the upper 30%-50% of the succession, where it is vaguely to moderately-well defined. It is never as well defined as the cross bedding within definite fluvial channels. Well-developed parallel lamination, interpreted as forming in a beach environment, is generally absent at the top of the successions.

11.2.1b: Sand Body Geometries

The sand-body geometries are also unusual. The combination of a broadly lobate large-scale geometry with the presence of very elongate, shore-normal thick "tongues" of sandstone has not been previously observed in any deltaic system. There is also the problem of the lack of definite channel successions cutting into the marine portion of the deltaic succession. This is most notable in Cycle G (Fig. 9.1), where the single elongate lobe is approximately 70 km long, most of which is apparently marine in nature.

11.2.1c: Problems of Interpretation

The combination of structureless sandstones and lobate shoreline geometries in the Lea Park - Belly River deltas presents several problems with regards to interpretation and formulation of an acceptable deltaic depositional model. The first problem relates to the abundance of structureless sandstones. Structureless sandstones can be difficult to interpret in any environment. If this structureless nature is a primary feature, the abundance of structureless sandstones within a shoreline sedimentary succession is even more difficult to explain.

The interpretation that the structureless nature of the sandstones is a primary depositional feature compounds several other problems, and makes the interpretation of the deltaic systems even more difficult. These include:

(1) How are the currents necessary to deposit these sediments generated at or near a shoreline? The depositional processes operating in shoreline environments are not generally thought to be capable forming and preserving large amounts of structureless sand.

(2) The geometry of the deltaic cycles indicates that massive-to-laminated sandstones are present up to tens of kilometres offshore from the last known shoreline position. Without an abundance of sedimentary structures, it is difficult to interpret the sand as having been moved tens of kilometres by normal, day-to-day marine processes.

(3) It is also difficult to interpret the structureless sands as having been deposited directly from fluvial currents debouching at the mouth(s) of distributaries. This would require that such currents flow tens of kilometres offshore before depositing thick beds of structureless sandstone. If this is the case, It raises problems concerning the lateral growth of the delta lobes. If they resulted from deposition from single large fluvial distributaries, it is difficult to understand how sediment could

be deposited over such a large area alongshore and offshore without any evidence of longshore drift or wave reworking.

(4) The **apparent** lack of definite fluvial channels within the deltaic lobes is also a problem. These deltaic lobes are interpreted to be progradational, and should contain numerous fluvial channels which have fed the delta.

It is the compounding of these problems which really makes the deltaic systems difficult to interpret. Any interpretation of the depositional processes responsible for the structureless sandstones makes interpretation of sand body geometries very difficult, and vice versa. Neither modern analogues, nor comparable ancient systems have been documented. A non-actualistic model must therefore be developed. Such a model is proposed in this chapter, but it is not completely satisfactory in explaining all of the problems listed above. This proposed depositional system does, however, attempt to explain the combination of characteristics observed within the Lea Park - Belly River deltas.

11.2.2: Depositional Mechanisms

The main problem concerns the depositional mechanisms of the thick, structureless or vaguely-stratified sandstone beds. X-radiographs of several examples of structureless or vaguely stratified sandstone do not indicate the presence of internal laminations (Fig. 11.1), indicating that the sandstone truly is structureless. There are two possible explanations for this structureless

Figure 11.1: X-radiograph showing structureless sandstone from Cycle G in 16-13-43-28W4. Small rectangular dark patch is unknown contaminant. Core is 4 inches (10.16 cm) wide.

[redacted]

nature; (1) the sand was deposited in bedforms, but post-depositional bioturbation or diagenesis has since destroyed the sedimentary structures; (2) the structureless sandstone beds were deposited very rapidly from sediment-laden suspension currents, with relatively little transport of the sediment as bedload, and relatively little time for the formation of bedforms such as dunes or ripples.

11.2.2a: Bioturbated Structureless Sandstone

Post-depositional bioturbation has been proposed as one mechanism to explain the presence of massive sandstones in shoreline sediments (eg. Bhattacharya and Walker, 1991b). Sandstones within the Lea Park - Belly River shoreline successions do contain *Macaronichnus* burrows and other trace fossils indicating some biogenic reworking of the sediment, so this explanation must be considered a possibility. Organisms similar to those which created the large visible *Macaronichnus* burrows, but much smaller, might have thoroughly churned the sand, leaving no evidence their presence. Some intervals of sandstone have a mottled appearance, which may be due to intense bioturbation. However, no definite evidence for extensive bioturbation could be found macroscopically, and when examined in thin section, no evidence of sorting of heavy minerals into discrete areas, as is typical of *Macaronichnus*, was observed. None of the marine sediments, mudstones or sandstones, observed in this study are intensely burrowed. Burrowing is usually confined to isolated burrows, rather than a pervasively churned mass of sediment. The existence of vague stratification in many of the "structureless" beds also argues against bioturbation. When present, this stratification, however vague or poorly defined, is continuous across the width of the core, with no disruptions apparent. If the structureless or vaguely stratified sandstones were the result of bioturbation, one would not expect stratification to be continuous. The presence and preservation of the apparently waning-flow beds also argues against a bioturbated origin. It seems unlikely that bioturbation would occur in such a manner as to give the appearance of structureless sandstone grading gradually upwards into poorly stratified flat bedded sandstones and then into well stratified LAIS sandstone. All of these features favour an interpretation in which the lack of structure or the presence of only very vague stratification is a primary characteristic of the sediment, and reflects the mechanism which deposited it. In conclusion, it seems that bioturbation was not responsible for the formation of the large amount of structureless sandstone in the shoreline successions.

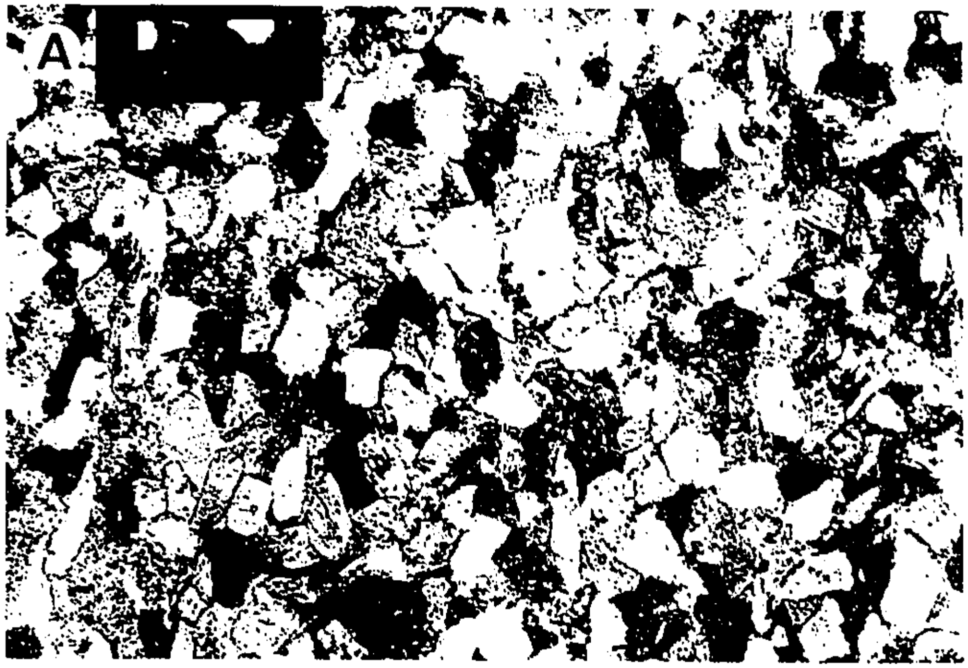
11.2.2b: Diagenetically-Created Structureless Sandstone

Another possible explanation for the abundance of structureless sandstone is that sub-surface diagenesis following deposition and burial has resulted in the realignment of sand grains in a manner that destroyed most sedimentary structures.

Essentially this process would require extensive chemical compaction and partial dissolution of the sandstone grains so that as pressure increased during burial, a repacking of the sediment would result in a destruction of sedimentary structures. This is a process that is not thought to have been common in unmetamorphosed sedimentary rocks, as even sandstones that have undergone extensive chemical compaction and dissolution usually retain some sedimentary structures.

The structureless sandstones within the Lea Park - Belly River transition appear to have a fairly complex diagenetic history, which will not be dealt with in detail. Calcite cement is a common precipitate within the deltaic successions, and is present in distinct intervals of heavily-cemented sandstones. Structureless sandstones are very common in both the calcite-cemented and non-calcite-cemented intervals, and both types of sandstone may also contain sedimentary structures. Thin sections from both calcite-cemented and non-calcite-cemented structureless sandstone are shown in figures 11.2 and 11.3. These thin sections are from the structureless shoreline sediments of Cycle H. Both samples show the grains to be of approximately similar size and shape. In the non-calcitic sample (Fig. 11.2), much of the pore space is filled with diagenetic clays, and there is little calcite cement. The calcite-cemented sample (Fig. 11.3), also contains interstitial clays, which are overgrowing the calcite cement, indicating that the cementation was a fairly early diagenetic event. Both samples have different diagenetic histories, and both are

Figure 11.2: Thin section photographs of non-calcified structureless sandstone from deltaic succession in Cycle H at 6-29-46-1W5. (A) Plane polarized light. Blue staining is porosity impregnation. (B) Cross-polarized light. Note abundant pore-filling clays. Magnification in both photographs is 50X.


A

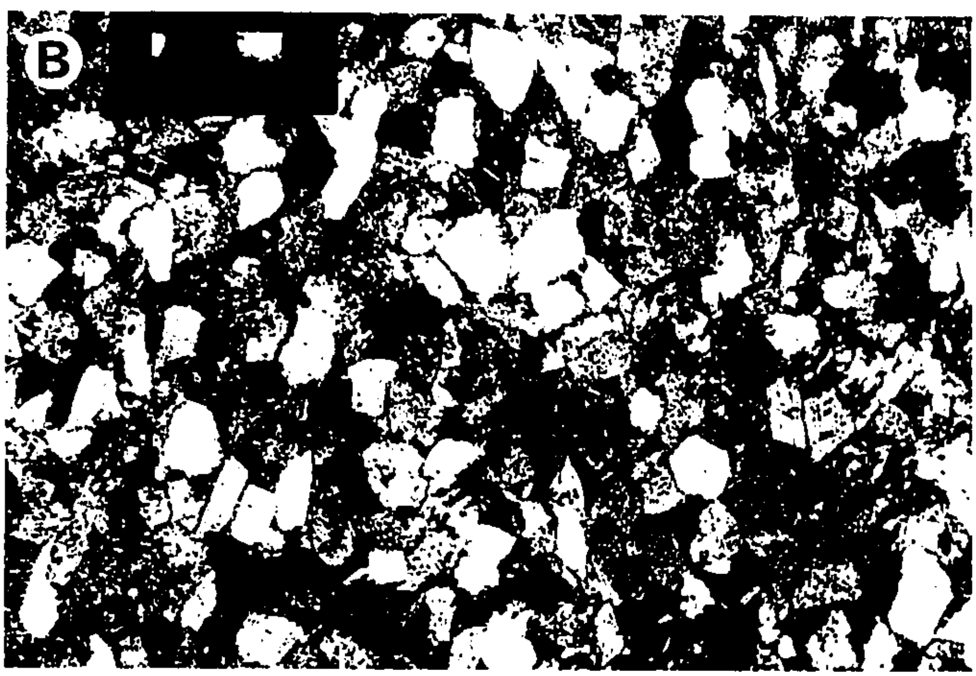
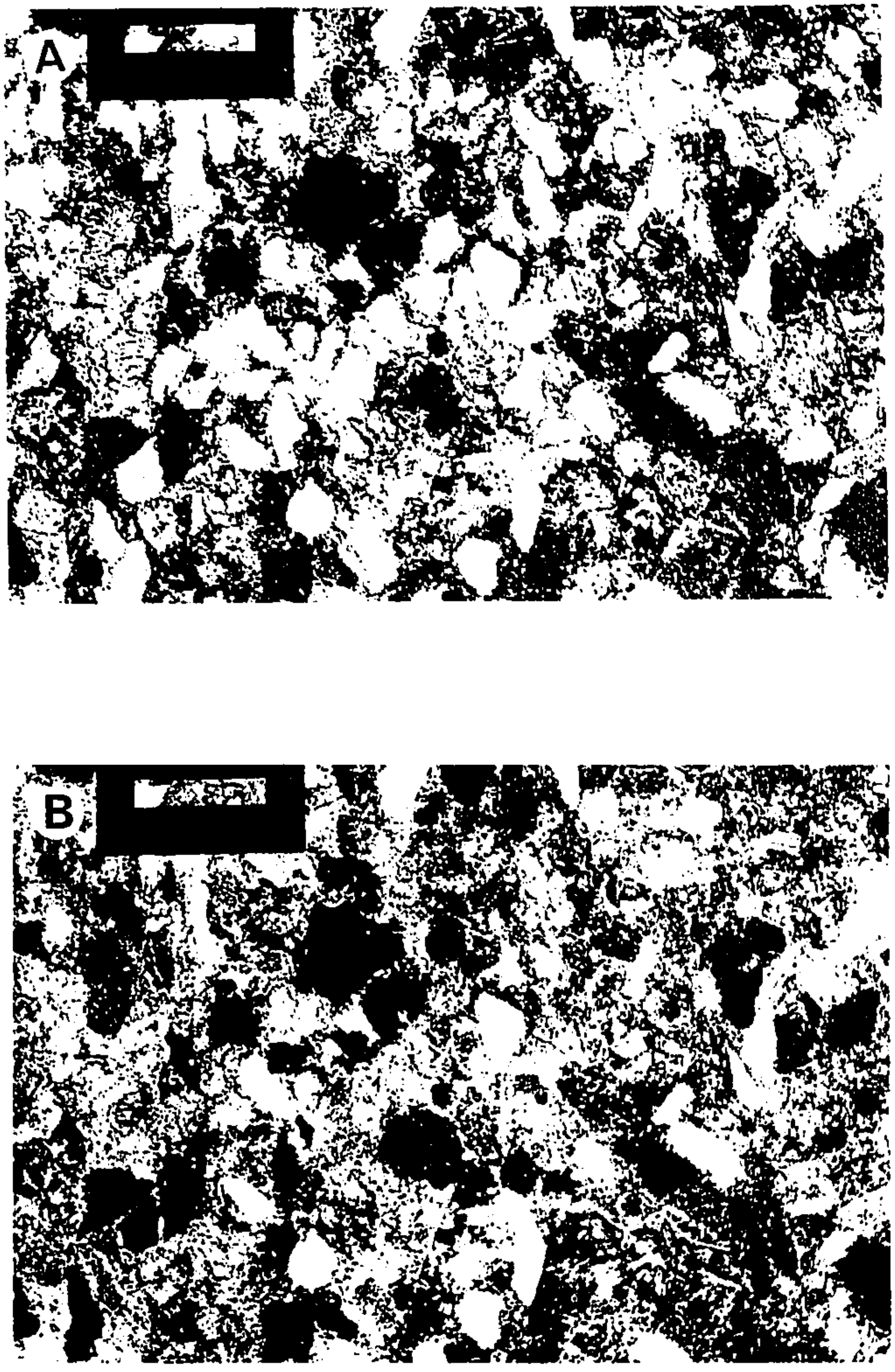

B

Figure 11.3: Thin section photographs of calcified structureless sandstone from deltaic succession in Cycle H at 6-29-46-1W5. (A) Plan polarized light. (B) Cross-polarized light. Magnification is 50X.


A
B

structureless, indicating that diagenesis was likely not the cause of their structureless nature.

11.2.2c: Direct Deposition of Structureless Sandstone

Sedimentary structures reflect the sorting of the sediment as it is transported as bedload and subsequently deposited in bedforms. If all of the sand being transported were of a similar size and composition, it is possible that poorly defined sedimentary structures would result. Much of the sand in the shoreline successions is very well-sorted, usually being fine sand. However, it is not unimodal in its composition. An abundance of both chert and quartz grains exist, as well as heavy minerals and other rock fragments. This should allow for compositional sorting of the sediment if it were being transported as bedload. The structureless nature of the sandstones observed in this study is therefore interpreted to be indicative of the processes which deposited it. This implies that the sands were deposited very rapidly, and that much of the sediment was travelling in suspension rather than as bedload. The most likely interpretation is that the massive-to-laminated beds are the deposits of shallow marine turbidity currents (turbidites). Critical evidence is the presence of a waning-flow succession within some of the beds, suggesting that they were deposited rapidly from a decelerating flow. The currents may have been similar to high-density turbidity currents discussed by Lowe (1982). The presence of large amounts of suspended sediment in these flows will inhibit the formation of equilibrium bedforms (Lowe, 1988). Sediment moving as bedload would be more likely to form poorly developed dunes, with vague internal stratification due to the constant rainout of sand from suspension. As the flow finally decelerates, more sediment would begin to move as bedload, creating better-defined stratification. Particles with lower density than sand, such as plant matter, would tend to settle out in the upper portions of the bed, helping to define the stratification. The concentration of organic matter in the upper stratified portions of the beds is a characteristic feature of the massive-to-laminated beds in the Lea Park - Belly River shoreline sediments.

Direct deposition from suspension of fine- to very fine-grained sand is thought to be rare (Lowe, 1988). In such situations, the sand is usually moved some distance as bedload prior to deposition. This may explain the abundance of very vague stratification, which is perhaps indicative of poorly-defined bedforms where sand is rapidly settling out of suspension.

There are other sedimentary structures present within the Lea Park - Belly River transition that are also indicative of rapid deposition from sediment-laden flows. Climbing ripples are present in distal shoreline deposits (Fig. 3.5a), and are also occasionally present at the top of structureless sandstone beds. Climbing ripples indicate high sedimentation rates from flows with relatively high sediment concentrations.

Within a shoreline environment there are a limited number of ways that turbidity currents with such high sediment loads could originate. The first possibility is that the flows may have been created by slumping, or similar mass movement of previously deposited sediment. Second, the flows were the result of rivers in flood debouching into the basin. Third, the flows may have been created by storm activity within the basin, and were unrelated to fluvial discharge.

The first possibility is that the massive-to-laminated beds were deposited by turbidity currents generated by mass movement such as slumping within the upper delta front. Slumping within deltaic systems is common, and has been most notably documented from the Mississippi Delta (Coleman and Prior, 1982; Coleman et al., 1974). It is most frequent in the outer regions of the delta front or prodelta, where depositional rates are high and slopes are relatively steep and unstable, and could conceivably generate density currents. However, shallower deposits are also prone to slumping. Approximately half of the mouth-bar deposits of the Mississippi delta are later slumped into deeper water (Lindsay et al, 1984). Abundant slumping generally characterizes deltas with high rates of sediment accumulation. The sediments have a high pore water content and are therefore prone to instability (Coleman and Prior, 1982).

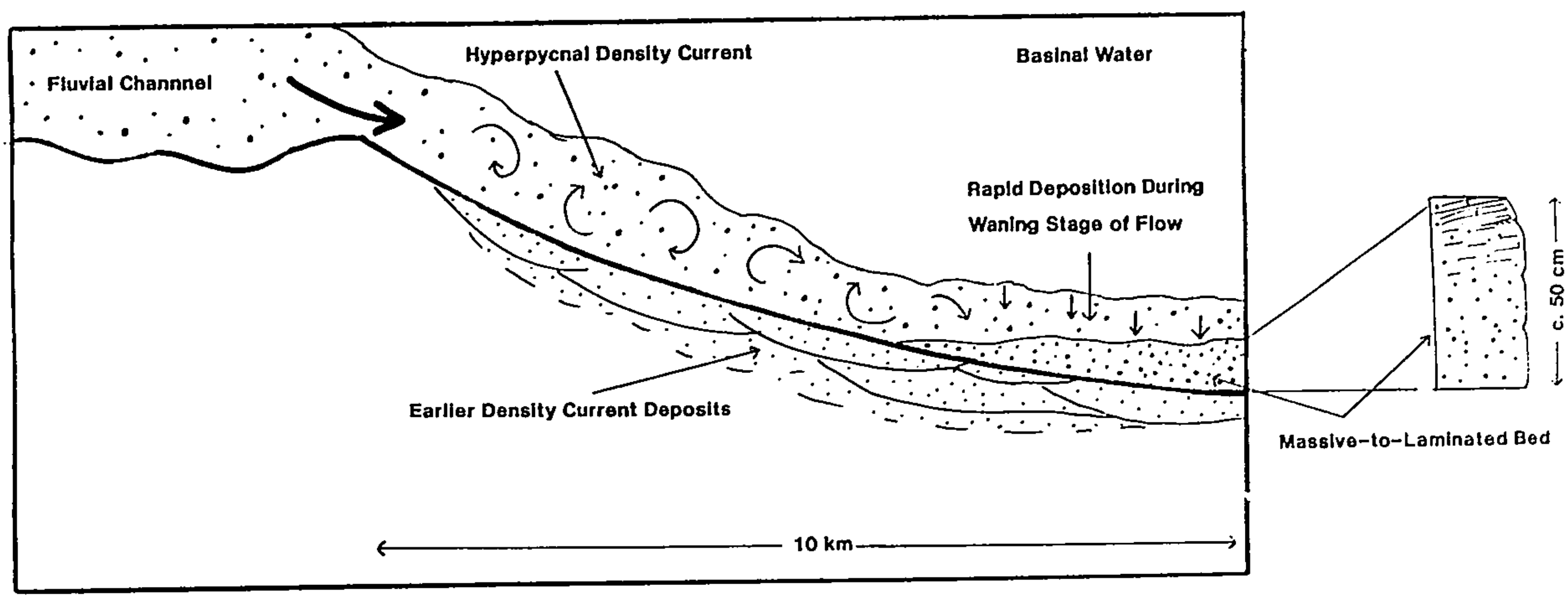
This interpretation is attractive in certain respects for the Lea Park - Belly River deltas. Soft sediment deformation, possibly indicative of slumping, is common within the interbedded mudstones and sandstones at the base of the massive sandstones. The triggering of the turbidity currents could be related to the instability of thick blankets of sand deposited in the delta mouth environment by river floods. These sands would be unstable due to a high pore-water content, and numerous mechanisms, such as sudden loading by a subsequent deposit, could trigger slumping of the sands. A slumped origin for the massive-to-laminated beds would also remove the requirement that the density flows travelled long distances, as the flows could travel incrementally over the required tens of kilometres. Middle Pliocence shelf edge deltas of the Mississippi Canyon region in the Gulf of Mexico contain thick, structureless and waning-flow sandstone beds (up to 4 m thick) which are interpreted to have formed by mass movement of previously deposited mouth bar sands (Mayall et al., 1992).

There are, however, several characteristics of the Lea Park - Belly River successions which raise questions regarding a slumped origin of the massive-to-laminated beds. The first is that slumping would have occurred so pervasively over the whole deltaic system such that almost all the sediment in both proximal and distal areas was redeposited. Second, if such a large amount of slumping had occurred, one would expect to find abundant evidence of soft sediment deformation (SSD) within the shoreline deposits. As previously mentioned, some deformation is observed in the lower shoreface/shelf sediments, but SSD is rarely observed within the sandy shoreface successions themselves. However, the known abundance of slumping within deltaic sediments and the fact that this interpretation does not require the proposal of non-actualistic processes makes it the most likely cause of the turbidity currents

The second possible cause of the flows is that they were generated by fluvial flood currents debouching into the basin. Bates (1953) classified deltaic effluents as being hypopycnal (less dense), homopycnal (equally dense), or hyperpycnal (more dense) than the basinal waters. The turbidity currents which deposited the massive-to-laminated beds would are interpreted to have carried very high amounts of sand in suspension and were more dense than the basinal waters. Using this interpretation, the massive-to-laminated sandstone beds were deposited by flows that debouched from the mouths of rivers as hyperpycnal density underflows, in which inertial processes dominated the flow. After the currents travelled some distance into the basin, they would begin to lose the capacity to transport sediment, and the structureless or vaguely laminated sandstones would be rapidly deposited from suspension (Fig. 11.4). The uppermost laminated portion of the bed would be deposited during the waning stages of flow. In this interpretation, the hyperpycnal flows would be required to travel up to tens of kilometres offshore from the mouth of the river before depositing the sand. However, there is no record of such processes depositing beds of sandstone more than a few cm thick per event in modern deltas, and no other studies of ancient deltas have proposed hyperpycnal currents capable of travelling such long distances. This is the main problem with this interpretation, but it is still viewed as being a definite possible cause of the turbidity currents which deposited the massive-to-laminated beds.

Figure 11.4: Diagrammatic representation of deposition of massive-to-laminated sandstone beds by hyperpycnal density current flow from river mouths.

DEPOSITION OF STRUCTURELESS SAND BEDS BY HYPERPYCNAL DENSITY CURRENTS



The third option is that the structureless and vaguely stratified sandstones were either deposited or reworked by storm currents and/or waves. However, our present understanding suggests that in shallow, nearshore environments, storm currents are not likely to generate sediment laden currents (Elliott, 1986). The observed facies successions are very different from those normally associated with wave or storm-dominated shorelines. The preserved succession requires large amounts of sediment to be deposited out of suspension without forming sedimentary structures, and without subsequent reworking of the sand by storm waves to form HCS or ripples. A storm interpretation would require the generation of currents with a large amount of sand in suspension at the shoreline. In this interpretation, the currents would then have to move as density underflows, depositing the sand as massive-to-laminated beds as the flow waned. The deposits would have undergone little subsequent reworking by waves associated with storms or everyday shoreline processes.

11.2.2d: The Upper Portion of the Deltaic Successions

In some of the cycles (most notably Cycle F, but also Cycle H), the massive-to-laminated beds dominate the succession. In the other deltaic cycles, however, the upper portions of the succession commonly contain cross-bedding, LAIS, and flat lamination, suggesting that the sediment was transported as bedload. The presence of these structures, along with roots and other evidence of subaerial exposure at the tops of the successions, suggests that these sediments were deposited very close to the fluvial source, either in a delta-front/mouth-bar environment or possibly within deltaic channels a short distance landward of the actual mouth. The sediments represent deposition during "normal" river flow between the floods which were responsible for the massive-to-laminated beds. It is unclear whether these cross-bedded sandstones were deposited within the open marine basin, or within distributary channels. It is very likely that both environments are represented by these deposits, and are difficult to distinguish from one another.

The stratification in the upper portions of the deltaic successions could also represent reworking by basinal processes, such as wave- or tide-induced currents in the upper shoreface. However, there is little supporting evidence of wave and tide processes within the shoreline systems, and the inferred geometry of the shoreline systems makes it unlikely that basinal processes were responsible for reworking the shallow water sediments. The poorly defined nature of the stratification also makes it unlikely that it was formed by basinal currents reworking previously deposited sediment.

11.2.3: Deltaic Geometries

One of the major problems in formulating a model for the Lea Park - Belly River deltas focuses on the geometry of the delta systems, especially with respect to how these geometries can be related to the observed facies associations.

There are several features of the internal geometries that are difficult to explain. The isopach maps of the cycles show the presence of narrow, elongate shore-normal or shore-oblique tongues of sandstone which are thicker than the surrounding more lobate sandstones. These tongues are situated at the landward edges of each cycle (Figs. 6.1, 7.1, 8.1, 10.1). The elongate tongues are up to several kilometres in width and 10-30 km in length. The surrounding lobate areas of sandstone are present for 20-30 km in an offshore direction and for up to 100 km in a shore-parallel direction. In most of the cycles, both the proximal elongate lobes and the surrounding sandstone body appear to be dominantly marine in nature. Broadly lobate sandy deltas similar in morphology to the Lea Park - Belly River deltas are usually present in wave-dominated environments, where wave and storm processes redistribute sediment into a broadly lobate morphology. There is only minor evidence for such processes within the Lea Park - Belly River deltaic cycles, especially in the shallow shoreface/foreshore deposits. The distal lobe deposits show more evidence of wave structures, but it is still not interpreted to be the dominant depositional process. It is therefore apparent that wave and storm processes are not dominantly responsible for the preserved morphology of the cycles. Instead, it appears that the lobate geometry is the reflects the distribution of turbidite deposition. The narrow, proximal tongues of thick sandstone may represent locations along the shoreline where there was greater sediment input from rivers. The increased sedimentation rate would cause an increase in the local subsidence rate as a result of compaction of the underlying mudstones. These thick tongues may also represent areas where increased amounts of sediment accumulated in pre-existing topographic lows. No evidence exists, however, to prove this. The locations of the thick tongues are apparently **not** controlled by focusing sedimentation between underlying topographic highs created by the thick tongues of the underlying cycle(s). For the most part, each successive cycle is located several kilometres basinward of the thick tongues of the underlying cycle.

There is also little evidence of large, major deltaic distributaries eroding into the marine sediments. This raises the problem of explaining how all of this sand was deposited in a marine environment with relatively little apparent evidence of major fluvial channels through earlier deposits as the shorelines prograded. The lack of channels may be more apparent than real, due to the limitations of core and log control. It is possible that the channels exist within significant gaps in the data base of the shoreline cycles. It is also possible, as previously mentioned, that the cross-bedded or LAIS sandstones in the upper portions of the shoreline successions may represent the deposits of distributaries. Unless this is the case, and if the channels are being fed by large, single distributaries, one is left with the conclusion that these thin tongues of sandstone must have prograded up to 20-30 km seaward without significant preservation of channels during delta progradation.

11.2.4: Proposal of a Non-Actualistic Deltaic Model

The preceding discussion indicates that all questions regarding the depositional nature of the Lea Park - Belly River cycles are far from answered. Certain things about the deltas can be stated with a relatively high degree of confidence.

(1) The deltas are very sandy systems, and appear to dominated by deposition from fluvial processes and shallow turbidity currents resulting in thick deposits of structureless sandstones. Turbidite deposits are present throughout the shoreline system in locations along the shoreline and tens of km offshore. Wave reworking appears to have been of lesser importance in the distribution of sediment.

(2) No existing deltaic facies model adequately describes these systems. Most delta models have only one or two major distributaries active at any one time. This is true for both river dominated deltas such as the Mississippi delta (Coleman and Wright, 1975), and for wave-dominated deltas such as the Rhone delta (Oomkens, 1970; Van Andel and Curray, 1960). This concept does not appear to apply to the Lea Park - Belly River deltas for a number of reasons:

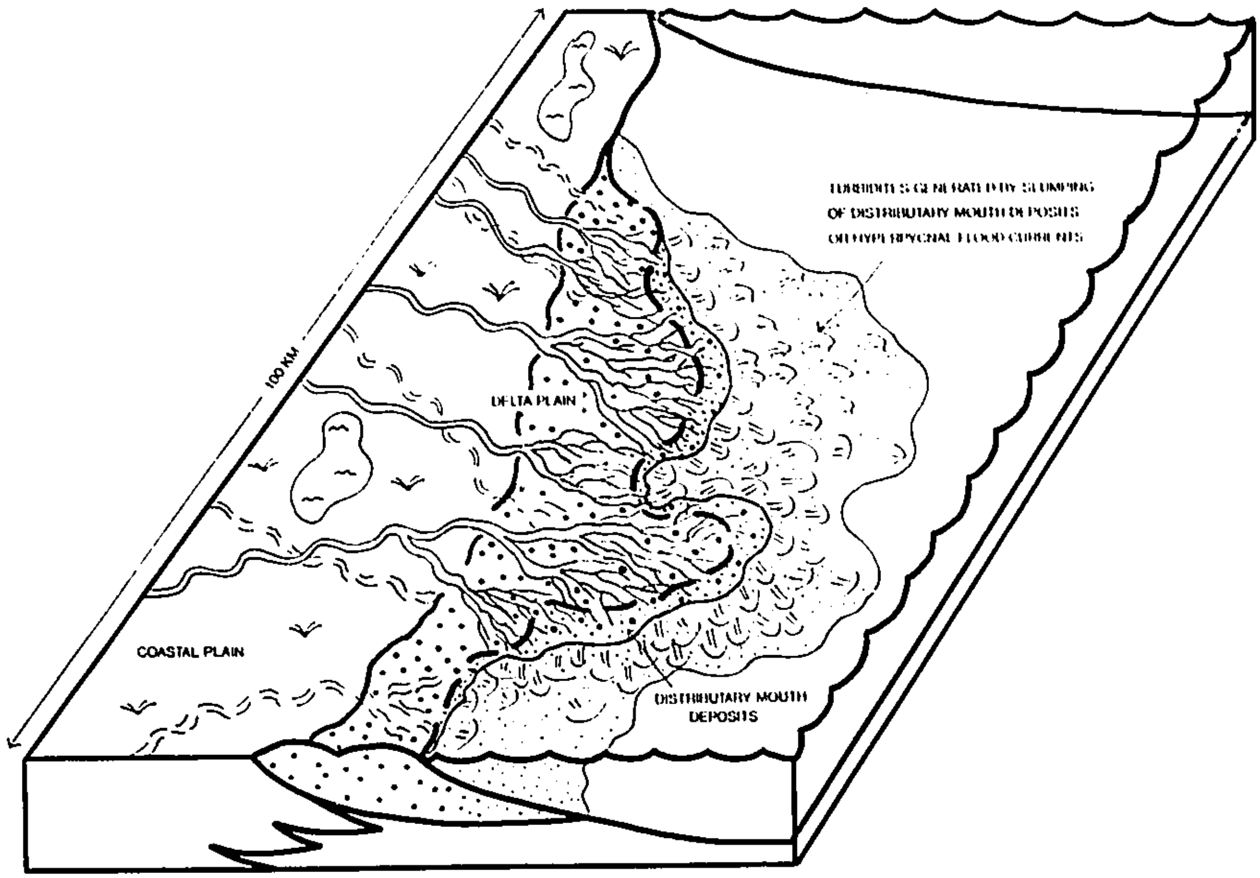
(A) There is little evidence of major contemporaneous distributary feeder channels on top of the marine deltaic sediments.

(B) The deltaic cycles have an alongshore extent of up to 100 km. All along the delta front the sediments are dominated by structureless sandstones. Therefore there is little or no evidence of longshore transport of these sediments.

(C) One or two major distributaries cannot account for the abundance and distribution of the massive sandstones.

The only way to reconcile all of these observations is to propose a deltaic model based upon deposition from many smaller distributaries, rather than one or two major distributaries. Thus, I propose that the Lea Park - Belly River deltas were characterized by a laterally extensive network of narrow, shallow channels on the delta top (Fig. 11.5). Sedimentation within the marine portion of the delta was dominated by deposition of shallow, sandy turbidites. These turbidites were likely created by the slumping of pre-existing, unstable delta mouth sands, although they may also have resulted from hyperpycnal density underflow from the river mouths. The lower delta plain of this system may have been morphologically similar to fine-grained versions of braid deltas proposed by Nemec and Steel (1988) and McPherson et al. (1987), although it would differ in many important characteristics from previously described braid deltas. The deltaic shorelines are interpreted to have been fed by numerous smaller

Figure 11.5: Possible depositional model of Lea Park - Belly River deltas in Cycles D, E, F, and H. The delta is created by numerous small rivers simultaneously feeding the shoreline over a 50-100 km wide area. Sedimentation within the marine protion of the delta is dominated by turbidite deposition, either due to slumping of distributary mouth sands or from hyperpycnal density flows from the river mouths.



SCHEMATIC DEPOSITIONAL MODEL OF
LEA PARK – BELLY RIVER DELTAIC SHORELINES

distributary systems which were simultaneously active, rather than one major fluvial system.

One of the central aspects of this proposed delta system is the nature of the channels which feed the shoreline, both in the delta plain environment, and further upstream in the alluvial plain environment. The delta plain itself would have been characterized by a maze of small, shallow channels as opposed to a single major distributary creating a distributary mouth bar at the delta front. These channels would switch positions rapidly, and would never become large and temporally stable. The alluvial plain is also interpreted to have been characterized by numerous smaller rivers actively feeding the shoreline, rather than a drainage system in which tributary rivers flow into a major distributary which then feeds the shoreline. In such a deltaic system, numerous active channels would provide many points of input into the basin, accounting for the problem of the wide lateral distribution of structureless sandstones deposited by turbidity currents. Such a deltaic model is non-actualistic for two reasons. First, there are no modern examples of sandy braid deltas with similar morphologies, and secondly, there are no known examples of deltas containing such abundant shallow sandy turbidite deposits.

Evidence to support the existence of this maze of shallow channels is found mostly in the stratified, upper portion of the shoreline succession (Facies Association 2a). These intervals are cross-bedded, are often characterized by subaerial exposure at the top (root traces, coal beds), and can show sharp grain size increases at the base of the interval (Fig. 11.6). The latter is especially true for Cycles D and G. These characteristics are consistent with a channel interpretation. The stratified intervals of Facies Association 2 are distinctly different from the definite fluvial channels of Facies Association 3, mainly in the sense that fluvial channels of Facies Association 3 contain sediment that is more poorly-sorted, cross-bedding that is much better developed, abundant current ripple lamination, abundant rip-up clasts, and well-developed fining-upward successions. This indicates that the interpreted channels of the lower delta plain were fundamentally different in nature from the meandering fluvial channels further inland. The lack of well developed internal successions is consistent with the interpretation that the delta plain channels were shallow and temporally unstable.

Figures 11.7-11.10 are versions of the dip-oriented log and core cross sections through Cycles D and G (Sections K-K' and S-S'). These sections are identical to the ones shown in figures 6.5, 6.6 and 9.5, 9.6, except that the upper cross-bedded portions of the successions are interpreted to represent distributary channel deposits in some locations, rather than marine shoreface sediments. This interpretation does not change the position of the last known exposed location of the shoreline within any cycle, so the requirement that the hyperpycnal flows must travel several kilometres offshore is still necessary. The consequences of this interpretation will be discussed in more detail in Section 11.2.5.

Figure 11.6: Sharp, angular grain size change from upper fine-grained sandstone to lower medium-grained sandstone (marked by arrow). This grain size change marks the boundary between the dominantly structureless or vaguely stratified shoreface sandstones and the cross-bedded interval of Facies Association 2 of Cycle D. The upper cross-bedded sandstones may represent distributary channel sandstones which prograded out over top of the shoreface/mouth bar environment. Scale bar is 3 cm. Location: 8-22-49-7W5; Depth 1048.9 m.

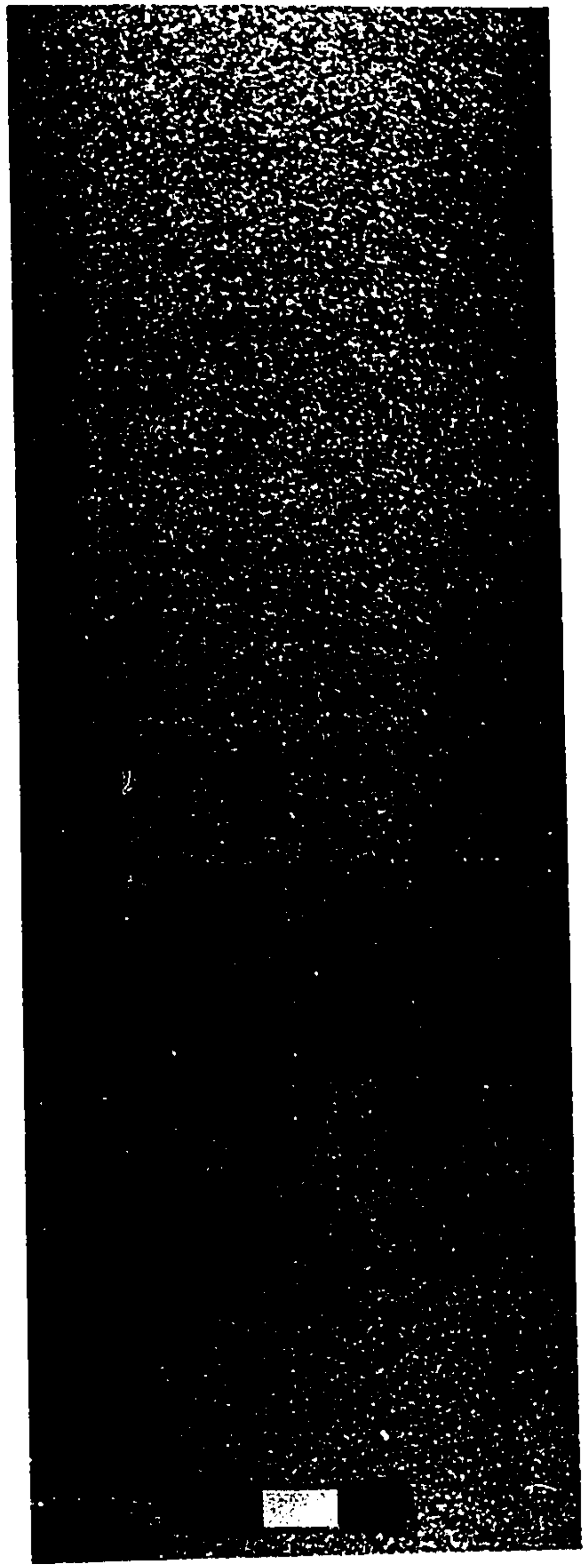
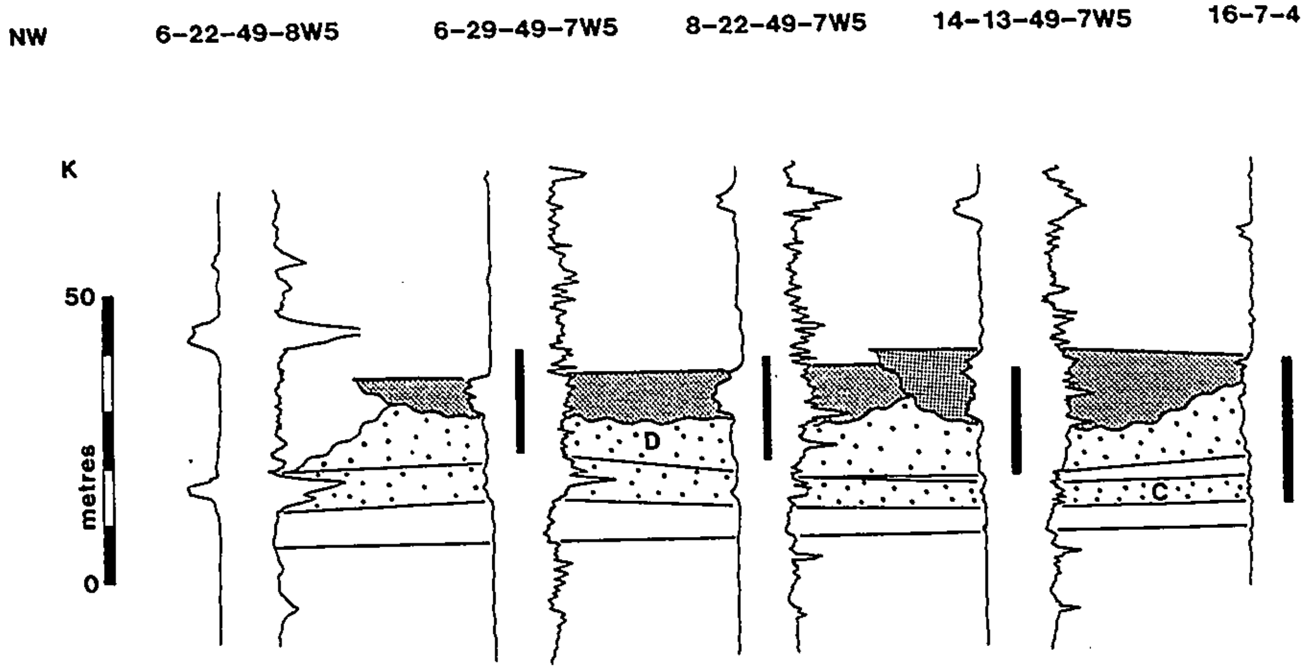


Figure 11.7: Log cross section K-K'. This is a dip-oriented section through Cycle D, and is identical to Figure 6.5 except that the upper, cross-bedded interval of the shoreline succession is interpreted to represent distributary channel deposits (shaded grey), rather than upper shoreface/mouth bar deposits. Definite fluvial channel sediments are not shaded, and are not thought to be related to Cycle D deposition.

CYCLE D: LOG CROSS SECTION PARAL

VER



.LEL TO DIP – NORTHERN PEMBINA REGION

ƗSION 2

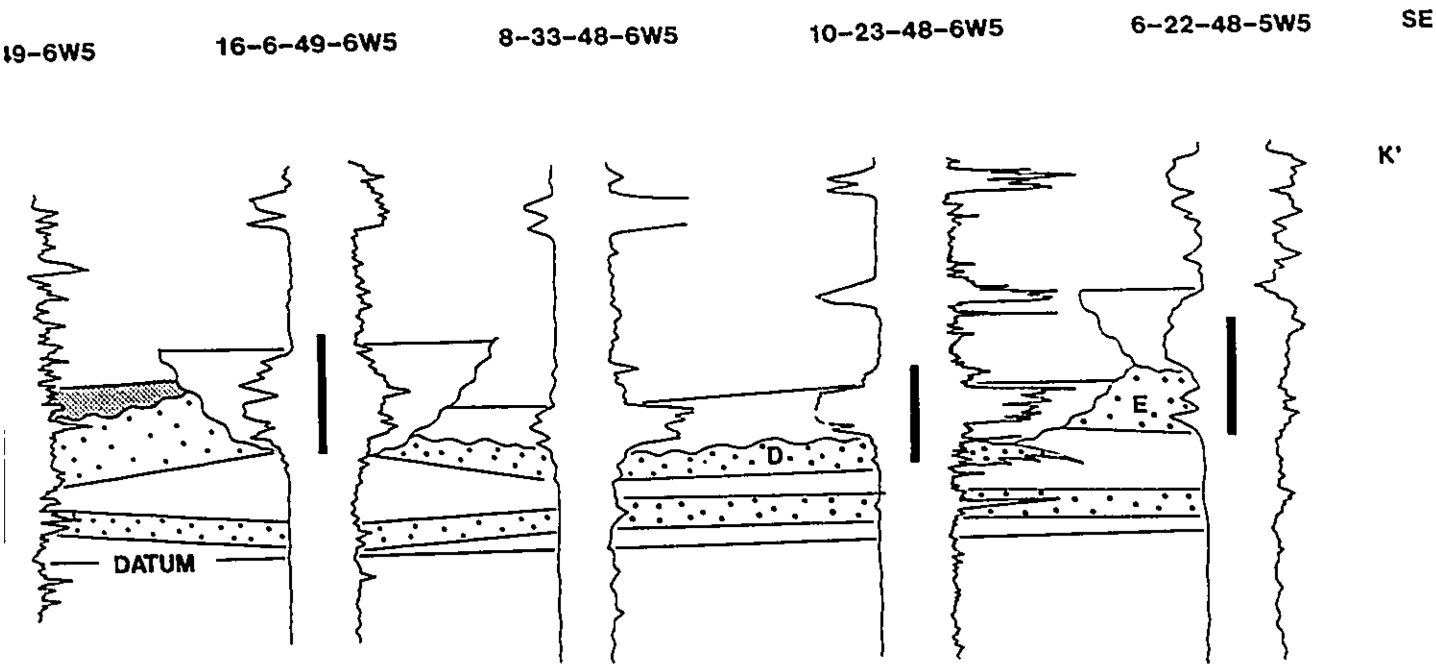


Figure 11.8: Core cross section for Figure 11.6, showing cross-bedded intervals in the shoreline succession of Cycle D interpreted as deltaic distributaries (speckled area).

CYCLE D: CORE CROSS SECTION PARALLEL TO DIP - NORTHERN PEMBINA REGION

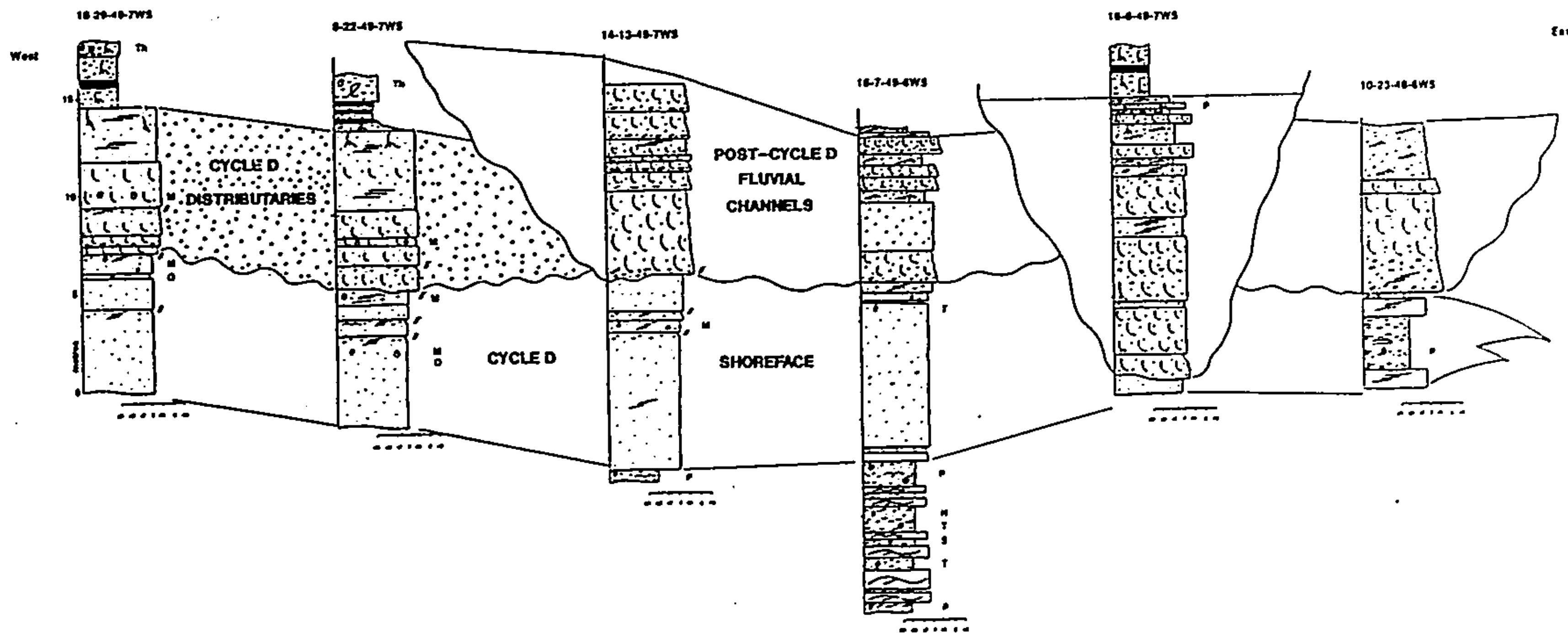
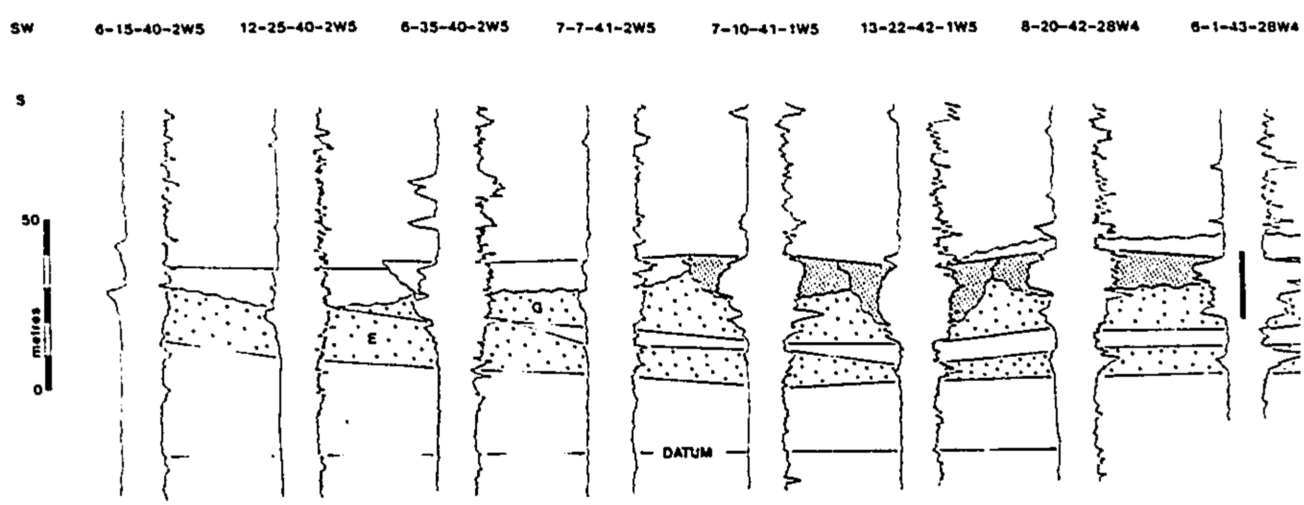


Figure 11.9: Log cross section S-S'. This is a dip-oriented section through Cycle G, and is identical to Figure 9.5, except that the upper, stratified interval of the shoreline succession in several cored wells and the corresponding log interval in several uncored wells is interpreted to represent deltaic distributary channel sediments (grey, speckled areas) rather than upper shoreface sediments.

CYCLE G: L



SW
6-15-40-2W5
12-25-40-2W5
6-35-40-2W5
7-7-41-2W5
7-10-41-1W5
13-22-42-1W5
8-20-42-28W4
6-1-43-28W4
S
50
metres
0
E
G
DATUM

LOG CROSS SECTION PARALLEL TO DIP

VERSION 2

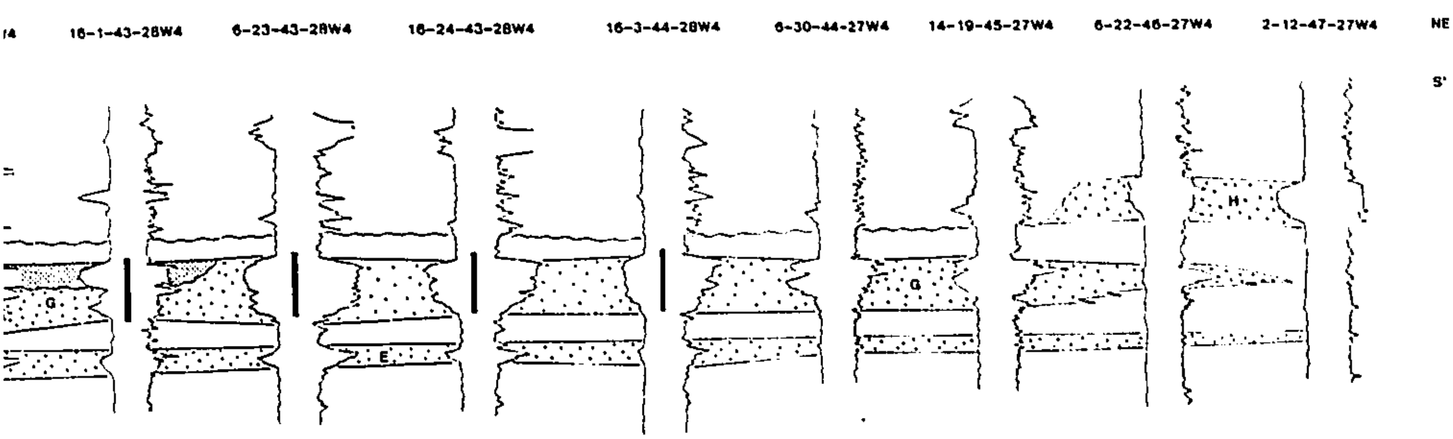
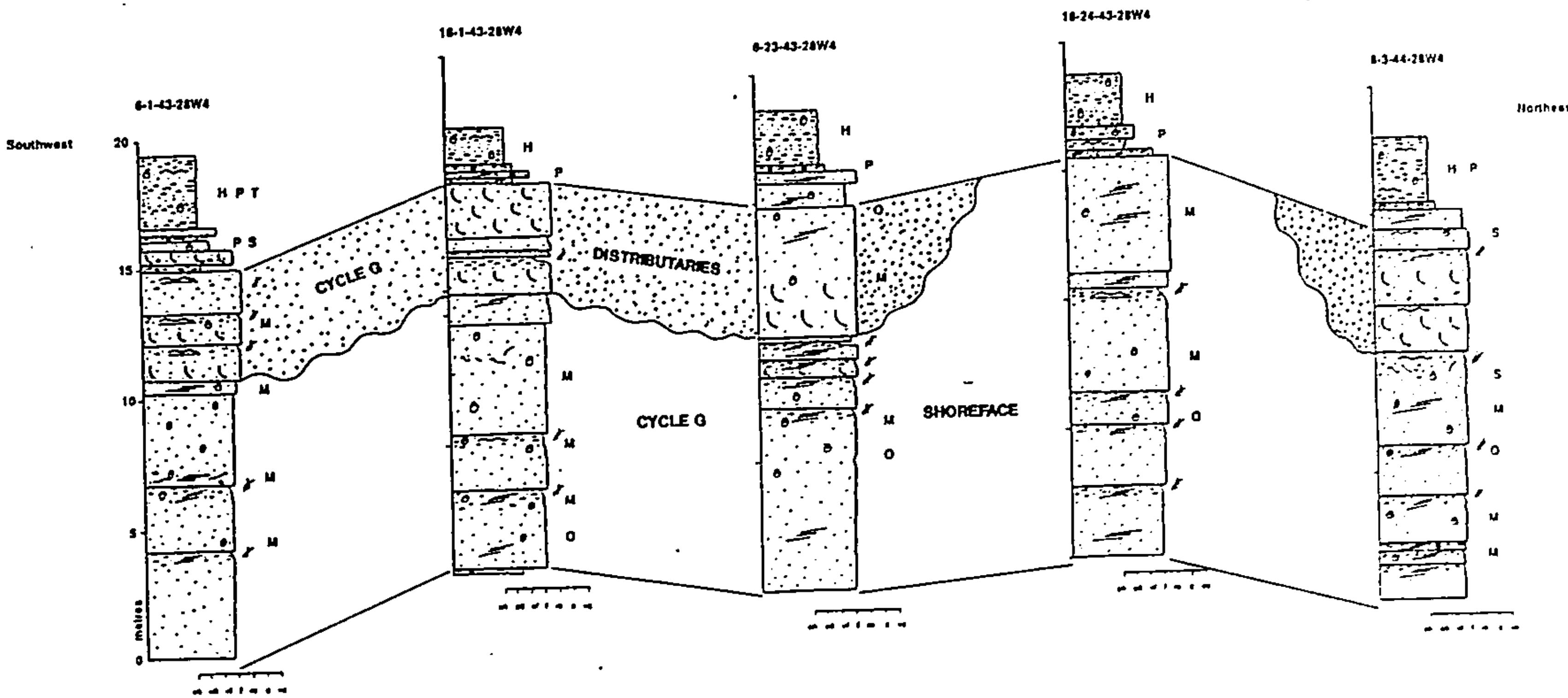


Figure 11.10: Corresponding core cross section from Figure 11.8 showing shoreline successions form Cycle G with the upper, stratified interval interpreted as distributary channel deposits.

CYCLE G: CORE CROSS SECTION PARALLEL TO DIP



The cross-bedded interval of Facies Succession 2 is **not** uniformly interpreted to represent the deposits of distributary channels. Many of the cross-bedded intervals contain features such as bioturbation or wave-rippling, which favour an open marine interpretation. The model proposes that the cross-bedded/LAIS sediments could represent the deposits of either shallow distributary channels, or the delta front environment immediately seaward of the channels. In such a case, the delta front would not be a single, discernable shoreline where the marine meets the non-marine, but instead would be a relatively indistinct transition between a very shallow marine delta front and a delta plain consisting of a maze of shallow channels. The maze of channels on the braid delta would pass landwards into larger, more definable distributaries or fluvial channels that are apparent on logs and in core. This would be morphologically similar to the conceptual Type I or Type J braid delta models of Nemec and Steel (1988), although the Lea Park - Belly River deltas are much finer-grained than these deltas are assumed to be. The nature of the shallow channels near the shoreline would make their deposits difficult to distinguish from the deposits of the area immediately seaward of the channel mouths. Deposition of sand in this environment during non-flood times would have formed the upper, stratified portions of the successions. The depositional environment of these sediments is therefore rather ambiguous, but is interpreted to be this shallow channel/marine delta mouth setting.

The lack of wave action in the shallow shoreline environment would result in poorly developed beaches. Farther offshore, the delta front was dominated by the massive-to-laminated deposits of the turbidity currents. Successive currents would be erosive as well as depositional, allowing the amalgamation of turbidites into thick, apparently structureless beds several metres in thickness. These turbidity currents were most likely generated by slumping of unstable distributary mouth deposits. Occasional waves (most likely associated with storms) reworked the tops of some of these massive-to-laminated beds. More distal portions of the delta front and the prodelta were characterized by a combination of the distal deposits of turbidites and sediment reworked by storms and waves.

11.2.5: Discussion

The preceding sections discussed the reasons why the Lea Park - Belly River deltas have been interpreted in such a manner. The following section is a discussion of the two main non-actualistic characteristics of the proposed model, and focuses on previous studies which may contain some similarities to the Lea Park - Belly River system, as well as the problems that the non-actualistic interpretation creates.

<u>11.2.5.1: Shallow Turbidity Currents in Deltas</u>

Turbidity currents due to slumping are well-documented in deltaic systems. The problem in using this interpretation for the Lea Park - Belly River sediments is mainly concerned with the lack of SSD in the sandy shoreface sediments. Therefore, this discussion focuses mainly on other examples of shallow water turbidite deposits. Middle Pliocene shelf-edge deltas in the Mississippi Canyon region of the Gulf of Mexico contain structureless and vaguely-laminated sandstone beds which are very similar to those observed in the Lea Park - Belly River transition (Mayall et al., 1992). These beds form amalgamated units 2-4 m thick within the deltaic successions, which is very comparable in scale to the Lea Park - Belly River successions. These deposits, however, are attributed to mass movement of previously deposited mouth-bar sands, not to hyperpycnal fluvial underflows. The steep slopes (2-4°) of shelf-edge deltas are thought to cause large-scale instability of the sediment within the shelf-edge delta (Mayall et al., 1992). The slope of the Lea Park - Belly River deltas is unknown, but if it were also steep, the possibility exists that slumping could have been pervasive enough to dominate deposition to the required extent. The lack of evidence of soft-sediment deformation within the sandstones remains a potential problem.

Similar slumping of delta-mouth sediments is also very common in fjord deltas. Thick beds of structureless sandstone within fjord-delta successions are commonly interpreted to be deposited as a result of slumping (Syvitski and

Farrow, 1989). The tectonic and paleogeographic setting of fjords is very different from that of the Lea Park - Belly River shorelines, but the comparison of processes may still be valid. The nature of slumping and mass movement of sediments will be determined by the balance between suspended sediment load and bedload of the deltaic effluent. Semi-continuous failure of sandy mouth bar sediments as is interpreted for the Lea Park - Belly River deltas is thought to be favoured by situations where both bedload and suspended load are significant (Syvitski and Farrow, 1989). Bioturbation within such fjord deltas tends to be very low within both the proximal delta mouth and the areas affected by turbidity current deposition. This is thought to be due to a combination of high sedimentation rates and the fact that the substrate is often being destroyed be slumping and turbidity current flow (Syvitski and Farrow, 1989; Farrow et al, 1983).

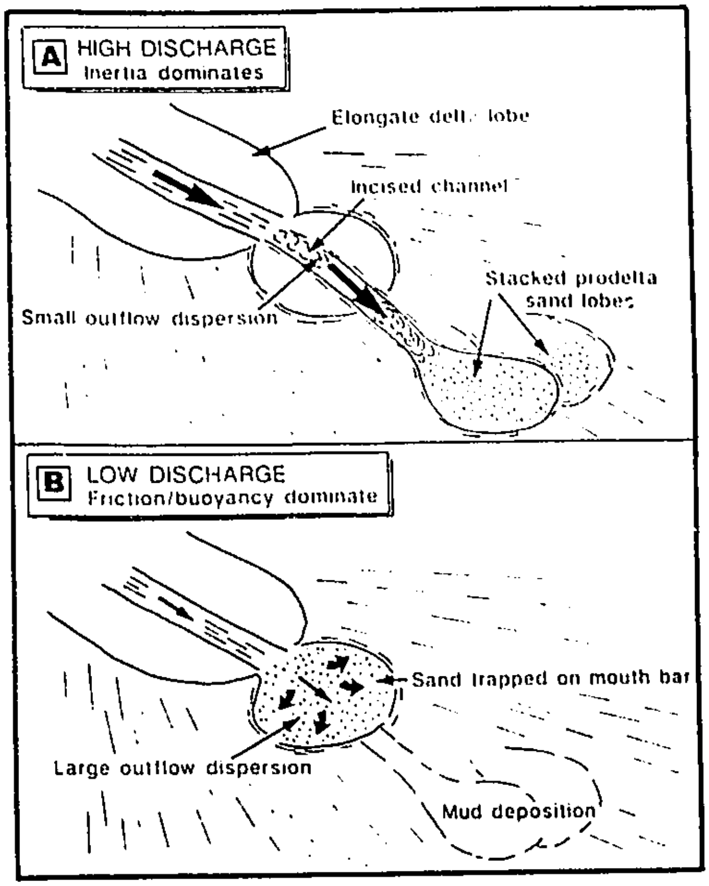
There does not appear to be any documentation of abundant shallow sandy turbidite deposition within "shoal water" deltas deposited on ramp margins or within foreland basins. Many ancient sandy delta deposits are interpreted as being wave-dominated systems. In such systems, turbidite deposition may have occurred, but the sediments would have been later reworked by wave/storm processes. Reworking of deltaic sediments by mass wasting is certainly very common in the modern Mississippi delta (Lindsay et al., 1984). As previously mentioned, it is thought that between 50-90% of all distributary mouth sediments are later subjected to mass movement such as slumping, and are redeposited in deeper water (Lindsay et al., 1984). It is interesting to note that the underlying cause of mass wasting is attributed to thick blankets of unstable sediment that are deposited during times of river flood (Lindsay et al, 1984). This situation is also interpreted to be the underlying cause of turbidity current generation within the Lea Park - Belly River deltas. Thick accumulations of sandy turbidites are not common in the Mississippi delta, however, because of the dominantly muddy load of the system.

11.5.2.2: Hyperpycnal Fluvial Currents in Deltas

The only documented modern delta to be characterized by hyperpycnal fluvial flows is the Huanghe (Yellow River) Delta in China (Wright et al., 1986, 1988). This river is characterized by extremely high densities of loess and silt in suspension, and as a result, the fluvial discharge is much denser than the oceanic waters (reverse of normal situation, where fresh water is less dense than oceanic water). This has resulted in widespread, dilute hyperpycnal plumes of sediment being deposited during the flood season (Wright et al., 1988). Unfortunately, this delta is depositing only fine-grained sediment, rather than sand, so it is perhaps not very useful as a modern analogue for the proposed Belly River deltas.

The only ancient deltaic system to contain sandstone beds interpreted to be deposited in a similar manner is the Namurian Scar House Beds of Yorkshire, England (Martinsen, 1990). These deltaic successions contain thick, sharply-based beds of vaguely stratified sandstone. Martinsen (1990) interprets these sediments to have been deposited from hyperpycnal underflows generated at flood times in which inertial processes were dominant. This process implies that the massive-to-laminated beds should not be deposited in the areas closest to the river mouth, but rather in locations somewhat more distal, as the hyperpycnal flow would bypass these most proximal areas (Fig. 11.11). The sedimentary succession of the Lea Park - Belly River deltas might support this interpretation. The massive-to-laminated beds usually comprise the lower half to two-thirds of the succession, indicative of more distal areas. As stated earlier, the distance that these hyperpycnal flows might travel is unknown, but the presence of the massive-to-laminated beds is not uncommon in locations tens of kilometres from the last known position of subaerial exposure of a given shoreline cycle within the Lea Park - Belly River transition. This is the major obstacle to the viability of an interpretation of fluvial origin of the currents. In the Yellow River Delta, hyperpycnal sediment-laden flows of are known to travel up to 20 km from the river mouth (Wright et al., 1986). This sediment is loess, however, and the distance sandy flows might travel is unknown. Elliott (1989) touches on this problem when, in a general discussion paper, he notes that portions of delta-front successions in several examples of ancient fluvially-dominated deltaic sediments contain "erosive-based beds of sandstone deposited by basinward-waning currents ... considered to have issued directly from the river mouth," and that no evidence for such deposits is

Figure 11.11: Theorized depositional mechanism for hyperpycnal flow deposits in the Namurian Scar House beds of England. The hyperpycnal flows bypass the proximal mouth bar and deposit their sediment in more distal regions (from Martinsen, 1990).


A
HIGH DISCHARGE
Inertia dominates
Elongate delta lobe
Incised channel
Small outflow dispersion
Stacked prodelta
sand lobes
B
LOW DISCHARGE
Friction/buoyancy dominate
Sand trapped on mouth bar
Large outflow dispersion
Mud deposition

found in modern systems (Elliott, 1989, p. 4). Unfortunately, there is little supportive evidence from environments which might be favourable to hyperpycnal density current deposition. For example, one might expect such density currents to be common in glacially-fed deltas, where temperature-related density differences due to the cold, sediment-laden water might cause underflows. However, there is no recorded evidence of glacially fed, hyperpycnal underflows which travel tens of km offshore.

The cause of the floods which would generate these flows is also speculative. The plumes of hyperpycnal sediment in the Yellow River Delta occur during regular, seasonal flood periods (Wright et al., 1988). This is possible in the case of the Lea Park - Belly River deltas. Another possibility is that the floods occurred as a result of the deltaic watershed response to large storms. Drake et al. (1972) describe the deposition of a large, significant sand delta which formed off the mouth of the Santa Clara River in California as a result of two intense rainstorms in January and February of 1969. From these two storms alone, a delta was created that had a subaerial length of 0.7 km, and extended further offshore for about 2 km. The bulk of this sediment is thought to have been transported in suspension, rather than as bedload (Drake at al., 1972). The presence of HCS and wave-rippled beds of sandstone within the shelf sediments below the deltaic shoreface successions of Lea Park - Belly River deltas indicate the existence of storms in the basin at this time. It is possible that severe rainstorms in the coastal deltaic watershed were the cause of the flood events.

The thickest massive-to-laminated beds within the Lea Park - Belly River transition are 7-8 m thick. However, the storm-generated fluvial floods would not have to be capable of depositing beds this thick. It is likely that all of these beds thicker than several tens of centimetres are amalgamated, and represent the deposits of numerous flood events. Each successive flow would be partially erosive as well as depositional. Only the basal structureless portion of each flood deposit is preserved, giving the appearance of an extremely thick, structureless bed of sandstone. Truncated burrows within otherwise structureless sandstone (Fig. 3.15c) support this interpretation of erosionally-amalgamated sandstone beds. The maximum thickness of the structureless portion of a single hyperpycnal flow is unknown. Laminated tops of beds which appear to record continuous deposition from one event can reach a thickness of 50-75 cm. However, most of this interval was deposited as bedload, as evidenced by the preserved structures. The structureless portion of this deposit may have been significantly thinner or thicker.

11.2.5.3: Sandy Braid Deltas

The deltaic model constructed shows some morphological similarities to the Type I or Type J braid deltas of Nemec and Steel (1988). These models are only hypothetical, and are thought to typify much coarser-grained systems than the Lea Park - Belly River deltas. Most braid deltas would be associated with fan-delta systems. There do not appear to be any examples of modern, sandy braid-type deltas similar to the Lea Park - Belly River deltas. However, some comparisons can be made with previous studies of sandy braid and fan-deltas.

The best-documented example of a modern, sandy braid delta is the Copper River delta of Alaska (Galloway, 1976). The Holocene Copper River delta system is approximately 80-100 km long in a shore parallel direction and 30-40 km in a shore-normal direction. The Copper River delta is therefore approximately the same size as the delta Cycles of the Lea Park - Belly River transition. However, there are many differences between the Copper River delta and the Lea Park - Belly River deltas. The Copper River delta is a glacially-fed, wave-dominated delta with significant tidal influence. Much of the shore-parallel extent of the delta system is due to sediment that has been reworked alongshore by waves and wave-induced currents, with the bulk of the delta system located to the west of the Copper River. The river is a single stream within a valley 20-30 km in width. The delta plain is braided only within the valley system. Therefore, the Copper River delta is being created by a single, major fluvial distributary, which is not the case for the model envisioned for the Lea Park - Belly River deltas. There is no evidence for such major valley systems in the Belly River fluvial plains. Instead, the Belly River fluvial plain appears to have been characterized by channel systems less than 1 km in width. The braid delta model envisioned for the Lea Park - Belly River deltas is the product of many of these rivers actively feeding the shoreline. Comparing the channel deposits of the Copper River delta to the interpreted distributary channel deposits of the Lea Park - Belly River deltas is difficult, as there is very little information available on the internal stratigraphy of the Copper River deltaic distributaries.

Another example of a prograding sandy shoreline fed by braided streams is the Skeidarársandur coast of southern Iceland (Hine and Boothroyd, 1978; Ward et al., 1976). This system is probably better classified as a strandplain than a delta, as the shoreline is straight. As with the Copper River delta, the Skeidarársandur coast is fed by glacial outwash. The fluvial system is dominated by sedimentation during meltwater flood stages, and as such might contain some similarities to the proposed Lea Park - Belly River delta systems. However, like the Copper River system, the Skeidarársandur coast is a poor analogy to the Lea Park - Belly River systems. The Skeidarársandur coastline is a wave-dominated system, and is characterized by extensive barrier-spit development. Sedimentary structures within the shoreline system are dominated by typical foreshore cross-stratification and low angle or flat lamination. There is no evidence of the presence of structureless sediments deposited by hyperpycnal fluvial outflows. The channels feeding the shoreline are dominated by trough cross-bedding. The channel systems themselves are poor analogies to the Belly River system. Like the Copper River, the rivers feeding the Skeidarásandur coast are major braided stream systems and are extremely wide (up to 9 km) (Hine and Boothroyd, 1978). The preserved deposits of this river system would be a wide sheet of sandstone. There is no evidence of this in the Belly River fluvial sediments.

There are also very few ancient examples of sandy braid deltas in the literature. One is the Pleistocene Athabasca braid delta of northeastern Alberta (Rhine and Smith, 1988). This was a paraglacial delta system deposited by the Athabasca River flowing into Glacial Lake McConnell following the retreat of the Laurentide ice sheet in the Late Pleistocene. The preserved deposits comprise a sandplain approximately 4000 Km^2, which is slightly larger, but comparable in size to the Cycles of the Lea Park - Belly River deltas. This braid delta was deposited over a very short period of time (1700 years). One similarity this delta system has with those of this study is that both are interpreted to have been deposited without any immediate proximal association with tectonism (ie. fault scarps). The deposits of the Athabasca braid delta also contain some similarities to the Lea Park - Belly River deltas. The Athabasca delta also contains density current deposits in the shoreface. These density current deposits, however, are very thin (1-5 cm), and are much more easily explained than the thicker deposits of the Lea Park - Belly River system. Unlike the Lea Park - Belly River deltas, the Athabasca delta contains abundant deposits of obvious pebbly sandstone channel sediments erosively overlying the marine deposits. As with the modern Copper River delta and the Skeidarársandur coast, the Athabasca delta is the result of deposition from a single major fluvial system, unlike the Lea Park - Belly River deltas. The braided nature of the Athabasca delta is interpreted to be caused by the influx of large amounts of paraglacial sand by eolian processes. This caused excessive bedload of the fluvial system, resulting in the braid-plain nature of the delta plain (Rhine and Smith, 1988).

The presence of flood-generated density current deposits has also been observed in Middle Ordovician fan and braid deltas in Wales (Orton, 1988). Graded sand beds up to 20 cm thick are present in distal mouth-bar deposits in both the Gwern Gof sandy fan delta system and the Capel Curig braidplain delta system. Other than these beds, the Welsh delta systems bear little resemblance to the Lea Park - Belly River deltas. The Gwern Gof and Capel Curig delta systems were deposited in a tectonically active back-arc basin, in direct association with active fault scarps. The fluvial systems supplying the deltas deposited obvious cross-bedded, coarse-grained pebbly sandstone-filled channel sediments.

There are many other examples of coarse-grained modern and ancient fan delta systems in the literature. Two recent volumes have focused on the sedimentology of these systems (Colella and Prior, 1990; Nemec and Steel, 1988). However, aside from those studies discussed above, the delta systems discussed in these volumes are of little use as analogies to the Lea Park - Belly River deltas. Their coarse-grained nature, direct relationship to either active fault scarps or glaciers, and the nature of their alluvial plains combine to make comparisons meaningless.

One notable problem raised by the proposed model focuses on the factor(s) that might have caused the fluvial systems to change from meandering channel systems to multiple channel braided or bifurcating systems at the shoreline.

Within the continuum between high-sinuosity single channel meandering rivers and low-sinuosity, multiple channel braided rivers, multiple channel and braided systems are generally favoured by high and variable discharges, steep depositional gradients (slopes), and abundant sediment supply (Allen and Allen, 1990).

It is very unlikely that an increase in depositional gradient is the cause of the change in channel pattern. This would be the reverse of what would normally occur. Depositional gradients invariably decrease towards the shoreline, and fluvial systems generally change from high-energy braided channel patterns to lower energy meandering patterns. Even rapid base-level drop associated with relative sea level fall should not cause rivers to become braided at the shoreline. This change in base level would be transmitted up the fluvial system to the headlands by downcutting and erosion of the fluvial valley, not by a change in channel pattern.

The braided channel pattern of the Pleistocene Athabasca delta is interpreted to have been caused by an excess of eolian-derived sand in the system, which caused the capacity of the river to be exceeded (Rhine and Smith, 1988). No evidence of abundant eolian sand supply exists in the Lea Park - Belly River system, but excess sediment supply might still theoretically be a cause of the interpreted change in channel pattern for the Lea Park - Belly River deltas. However, the origin of the necessary large amount of sand at or near the shoreline is unknown, and there is little evidence in the preserved Lea Park - Belly River stratigraphy of extensive sand deposits which could supply the fluvial system. One possible source of excess sand near the shoreline would have been the underlying shoreline deposits of the previous depositional cycle. However, there appears to have been relatively little "cannibalization" of underlying shoreline deposits in the Lea Park - Belly River system, which therefore does not favour this interpretation.

Perhaps the most likely cause of the proposed change in channel pattern would be periods of very high fluvial discharge (ie. storm-generated floods). These flood periods were also interpreted to be the cause of the hyperpycnal density currents which deposited the massive-to-laminated beds in the shallow and middle shoreface environment. These same storm floods would bring large amounts of sediment from the alluvial plain to the coastal plain. This sediment could be stored in the coastal plain, and during non-flood times, this abundant sediment supply in the lower delta plain would exceed the capacity of a meandering channel system, causing the channel system to switch to a multiple-channel, braided system.

Another problem with the multiple channel delta plain model is that braid deltas tend to form in relatively dry areas with little interchannel vegetation. The Lea Park - Belly River alluvial plain clearly was not dry, as evidenced by the abundant ponded overbank sediments and coal deposits. If vegetated interchannel regions were present on the delta plain itself, these would have acted to stabilize channel banks, making the channels more temporally stable.

One answer to this problem may be that the lower delta plain and shoreline was different from the alluvial plain in that the coastal areas were far sandier in nature, with little fine-grained mud component. The shoreline cycles show no evidence of interdistributary bays, but rather appear to be more continuous, coalesced sand lobes. Channels incised into sand would have less stable banks than those incised into mud, facilitating rapid channel switching.

In summary, there appears to be no definitive depositional model that is satisfactory for the Lea Park - Belly River delta systems. The proposed depositional model is non-actualistic. Documentation of abundant hyperpycnal sand-laden currents from deltas does not exist. Use of the term braid delta may even be misleading. The proposed model bears little resemblance to the glacial or fault-scarp braid deltas described in the literature. The cause of an evident change in fluvial style from single channel meandering systems in the upstream alluvial plain to multiple-channel "braided" or bifurcating systems in the coastal plain is presently unknown, and can only be speculated on. The largest problem raised by this model may relate to the absence of wave-reworked sand in the broad, distal lobes, and the consequent interpretation of hyperpycnal currents that are able to travel long distances offshore. Despite these problems, it is hoped that this proposed model will at least generate further discussion on the nature of structureless sandstones in deltaic and shoreline systems.

11.3: Fine-Grained Fluvial Systems

This thesis has concentrated on the marine or coastal sediments within the Lea Park - Belly River transition, mainly because it is these sediments that can be used to construct an allostratigraphic framework. Bounding discontinuities are very difficult to trace into the non-marine portion of the Belly River Formation. Without these bounding discontinuities, the non-marine sediments are, unfortunately, relegated to a minor role in the stratigraphic framework. However, over half of the cores through the Lea Park - Belly River transition examined in this study comprise definite non-marine sediments. Most of these are fluvial channel sediments, and these sandstones form important hydrocarbon reservoirs. Therefore, it is important that they be examined sedimentologically and some attempt is made to relate them to the marine cycles.

Fine-grained fluvial channels characteristic of Facies Association 3 are the most common facies association observed within the study area. The brief interpretation given in chapter 3 for this facies association simply stated that these sediments were deposited in non-marine, channelized environments. This facies association consists almost entirely of stacked fining-upwards successions of very fine- to medium-grained sandstones. Trough cross-bedding is the most abundant sedimentary structure and usually grades upwards into current ripples. This facies association is readily classified as being typical of sandy, meandering fluvial systems, indicative of deposition on meandering river point bars (Walker and Cant, 1984; Allen, 1964; 1965).

Most of the cored intervals through the fine-grained fluvial channels indicate the presence of multiple channel fills, suggesting that the channels migrated laterally somewhat, and each channel location contains the record of several periods of point bar deposition. On a larger scale, the cross sections in chapters 4-10 show that these channel sandstones are not usually part of laterally extensive sheet sandstones, but tend to be laterally-restricted, and encased in finer-grained floodplain sediments. This lateral restriction of channel sediments is typical of meandering fluvial systems.

One important feature of the Belly River alluvial sediments is that they are **not** characterized by major valley systems (ie. major distributaries). Instead, the alluvial stratigraphy is characterized by numerous channels deposits, most of which appear to be of equal dimensions. This is consistent with the proposed model, which implies that the shorelines were being fed by numerous, smaller active fluvial systems, rather than one major valley system. The reason for the presence of many smaller channel systems may be related to the nature of the alluvial drainage basin. It was a foreland basin, and rivers originated from the mountains to the west. The distance travelled by the rivers from the mountains to the shoreline was likely less than 100 km. There may not have been sufficient space for the fivers to form a drainage basin characterized by one major axial river being fed by smaller tributaries. If this had formed, the rivers draining the mountains would have been tributaries of this larger fluvial system, which would have flowed parallel to the mountain chain. Instead, the rivers appear to have flowed perpendicular to the mountain chain, directly across the proximal foreland basin. This would have resulted in many smaller streams actively feeding the shoreline.

Rahmani (1988) interpreted some of the channels in the basal portion of the Belly River Formation to be estuarine, based on the presence of features such as paired organic laminations (Fig. 3.21). In the present study, little evidence was found to support an estuarine interpretation. None of the Lea Park - Belly River channels contains any real biological evidence (bioturbation) of the increased salinity that one would expect in a tidally-influenced estuary. In addition, the paired organic laminations cannot be shown to occur with any lunar cyclicity. Even if some of the organic couplets of laminations were tidally-produced, it does not require that the channel be estuarine. Deltaic distributaries are also connected to the open ocean, and it is perfectly reasonable to expect that tides may have a minor influence on sedimentation within these channels. This does not make the distributaries estuarine, as they are not drowned river valleys. It therefore seems far more likely that these channels were fluvio-deltaic than estuarine in nature.

11.4: Coarse-Grained Fluvial Systems

The fine-grained sandy fluvial systems discussed in section 11.3 represent the dominant style of fluvial deposition within the study area, especially in the Pembina and Keystone fields. Coarser-grained fluvial systems, typified by Facies Association 4, are far less common. They are present only in the Ferrier - Willesden Green area, and in some portions of western Pembina.

This facies association is composed mainly of cross-bedded or structureless pebbly sandstones, with the sandstone being dominantly fine-grained or medium-grained. Trends within internal successions, both in terms of grain size and changes in sedimentary structures, tend to be less well-developed than in the finer-grained fluvial systems. The channel sandstones are composed of numerous internal successions, and sharp changes in grain size between successions are common.

In Chapter 3, this facies association was interpreted to represent deposition in a high-energy fluvial environment where channels were relatively unconfined. The coarse grain size is indicative of relatively high-energy stream conditions, and the lack of well-developed internal successions, as well as the large number of fill events within a fluvial sandstone in any given location may be indicative of rapid changes in channel conditions, and the presence of multiple, shifting channels. This facies association is similar to that of braided stream environments (Rust and Koster, 1984; Miall, 1978). However, due to the limitations of core data, it is very difficult to definitively interpret these channel sediments as being deposited in a braided stream environment. They apparent evidence of abrupt channel switching and poor internal succession development may also have been produced in a meandering stream environment in which channels migrated and reoccupied the same area many times.

It is difficult to determine the relationship of these fluvial systems to the marine cycles farther to the east. They are not directly landward of any particular marine cycle. They most likely represent upper alluvial plain sedimentation related either to one of the later marine cycles within the study area, or to some marine cycle further to the east of the study area. Without some way of extrapolating bounding discontinuities back into the non-marine section, it is impossible to determine the exact relationship between these sediments and the marine cycles.

11.5: Summary

The sedimentology of the deltaic and fluvial systems within the Lea Park - Belly River transition in central Alberta is a study in contrasts. The fine-grained fluvial systems are relatively easily interpreted as meandering stream deposits. The coarse-grained fluvial systems are more problematic. They are indicative of deposition in an high-energy environment where channel switching appears to have been fairly common, but this may have been either a braided or a meandering stream system. These models have somewhat limited stratigraphic applications because they do not emphasize detailed lateral and temporal relationships between the fluvial sediments. It is presently very difficult to analyze fluvial sediments in a chronostratigraphic manner due to the lack of correlatable bounding discontinuities on which to base an allostratigraphy.

The deltaic sediments observed in this study cannot be adequately described and classified according to existing deltaic facies models. They contain a combination of sedimentary characteristics and three-dimensional geometries which have not been previously observed in either modern or ancient deltaic systems. The following points show why these deltas are unique, and why it is difficult to formulate a model to explain their deposition.

(1) The abundance of structureless sandstone and the scarcity of sedimentary structures indicative of extensive basinal reworking of sediment leads to an interpretation that shallow water turbidity currents carrying large amounts of sand in suspension were the dominant process responsible for deposition of sand in the system. The most feasible explanation for the origin of these turbidity currents is that they were created by slumping of unstable distributary mouth deposits. An

alternate possibility is that hyperpycnal density flows were issued directly from the river mouths. However, the distribution of the apparent turbidites would require that the currents were able to travel tens of km basinward from the river mouth before depositing the sediment. Hyperpycnal flows such as these have not been found in any modern delta system, nor have these types of flows been interpreted to have deposited such thick amounts of structureless sandstone in any ancient deltaic succession.

(2) The geometries of the deltas indicate that the most proximal mouth bar areas were elongate and shore-normal, but that extensive, lobate areas of sand tens of kilometres wide in both shore-parallel and shore-normal directions surround these proximal lobes. These geometries have not been previously observed in deltaic systems.

(3) There is apparently little evidence of large deltaic distributaries feeding the deltaic cycles as they prograde into the basin. Instead, the delta top was apparently characterized by numerous, small, shallow braided or bifurcating distributaries.

(4) The problems in proposing a depositional model centre on trying to reconcile the geometry of the shoreline systems with the facies successions dominated by the massive-to-laminated sandstone beds. The geometry of the cycles would normally indicate that at least moderate reworking of the sand is probably required to form the

extensive lobate areas of sandstone surrounding the proximal lobes. The elongate proximal lobes, on the other hand, seem to be indicative of very little basinal reworking of delta-mouth sediments. The facies successions indicate that the lobate geometry is largely the result of turbidity currents deposition, and the proximal elongate tongues of thick sandstone likely reflect the locations of increased sediment input from rivers.

The resulting model proposed to explain these deltas therefore contains some elements which may seem contradictory. It is admittedly speculative, and there are questions as to whether or not the processes proposed to have deposited the sandstone are realistic on such a large scale. However, no existing deltaic model can reasonably explain these deposits, requiring that the non-actualistic model proposed at least be considered.

CHAPTER 12: EVOLUTION OF THE LEA PARK - BELLY RIVER TRANSITION

12.1: Introduction

The regional cross sections in chapter 4 showed that the Lea Park - Belly River transition in the study area is characterized by eight regressive marine shoreline cycles. Each cycle is separated from the overlying cycle by a unit of muddy shelf sediments, which is referred to as a "transgressive unit", although much of this sediment may have been deposited during progradation of the overlying cycle, rather than transgression of the underlying cycle. Chapters 5-10 discussed the nature of the youngest six cycles, and showed that most of the cycles contain the deposits of sandy, deltaic shoreline systems.

The purpose of this chapter is to analyze how the Lea Park - Belly River transition evolved within the study area, mostly with respect to how the cyclic regressive-transgressive nature of the transition developed.

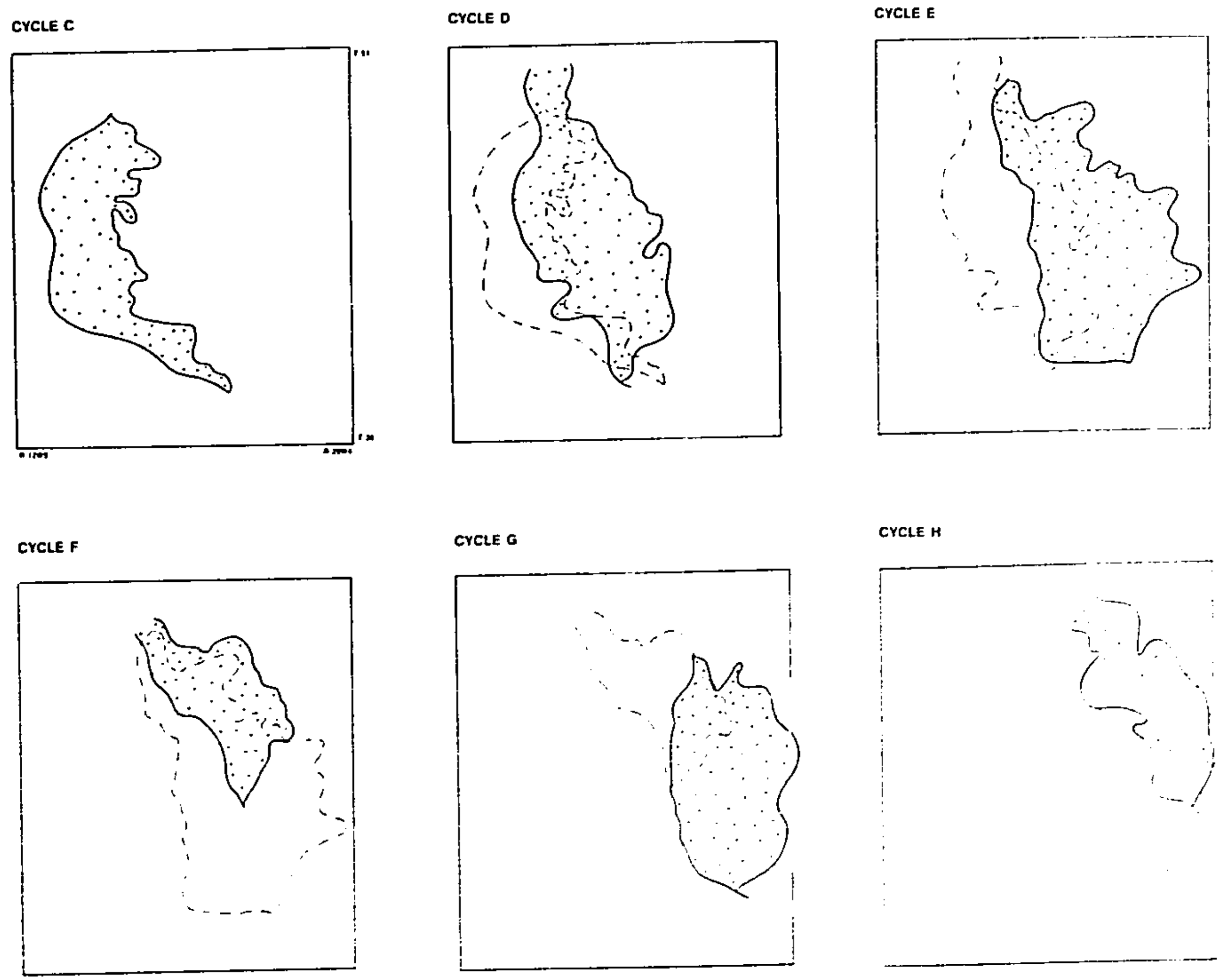
12.2: Spatial Relationship of Lea Park - Belly River Cycles

Figure 4.1 is the schematic cross section showing the distribution pattern of the eight cycles of the Lea Park - Belly River transition within the study area. The thickness of the cross section (75 m) is the average thickness of the transition in the regional cross sections of chapter 4. The thickness of each cycle is consistent with its true thickness. The lateral position of each cycle within the study area was taken from the isolith maps of the previous chapters and plotted along the horizontal axis of the diagram. This diagram clearly shows the regressive-transgressive nature of the Lea Park - Belly River transition, and the fact that each cycle is deposited in a progressively basinward position from the previous cycle. The figure also shows that the deposits of the regressive cycles are either preferentially preserved, or comprise the bulk of the original deposition, indicating that deposition during the intercyclic transgressive phase was relatively minor. The basinward progradation of successive cycles is also clearly visible in figure 12.1, which is a diagram showing the plan view distribution of Cycles C-H within the study area. The distribution of each cycle is shown relative to the previous cycle, which is shown in dashed lines. From the onset of Cycle C deposition to the onset of Cycle H deposition, the shoreline has migrated in a stepwise fashion over 100 km into the basin. Each basinward "step" is separated by a period of time during which the intercyclic transgressive shelf sediments were deposited.

12.3: Control of Regressions and Transgressions

The underlying control of the alternating regressions and transgressions can be interpreted in two ways. First, the system could be responding to internal sedimentological controls, such as deltaic lobe avulsion and abandonment. Second, the regressive-transgressive pattern could be reflective

Figure 12.1: Spatial relationship of Cycles C-H in plan view. In each cell, the previous cycle is marked by a dashed line, to show the amount of progradation into the basin associated with the onset of each cycle.



SPATIAL EVOLUTION OF SHORELINE CYCLES

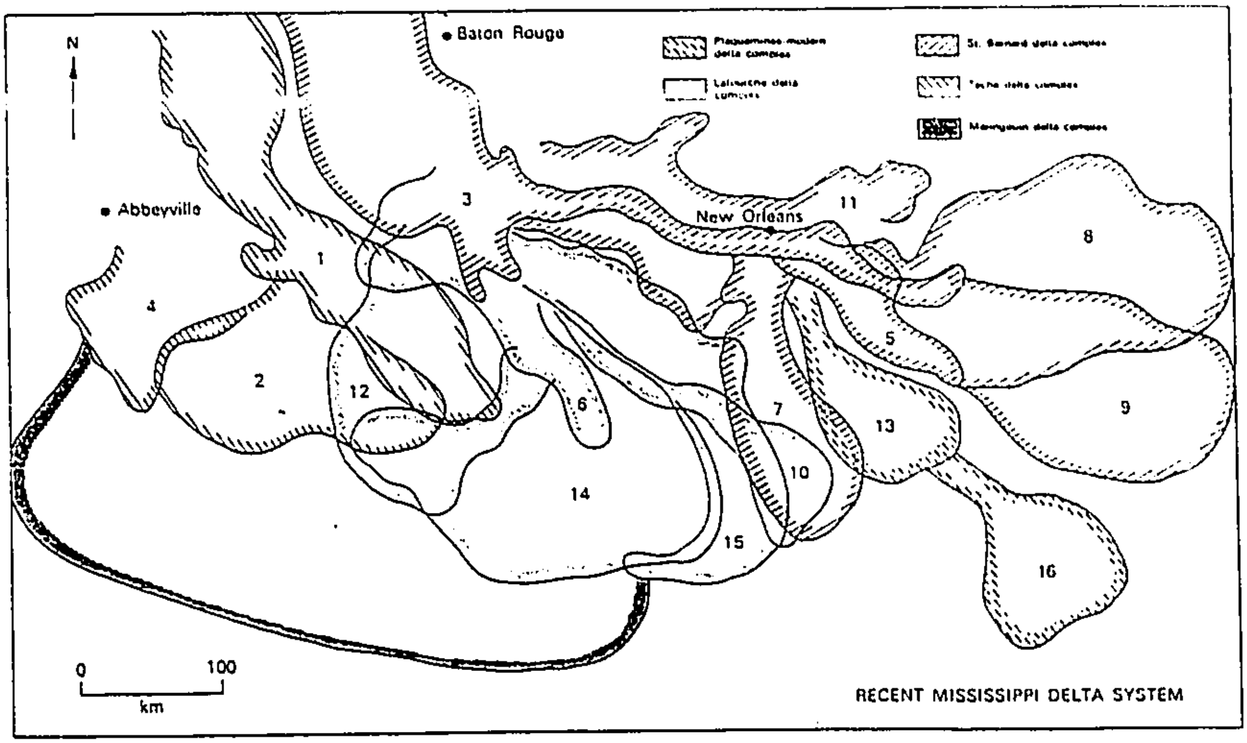
of allocyclic controls, which forced relative fluctuations in sea level upon the system.

12.3.1: Autocyclic Control of Regressions and Transgressions

Most of the cycles identified within the study area are the deposits of deltaic shoreline systems. Autocyclic switching of shoreline position due to deltaic avulsion is therefore a possible explanation for the Lea Park - Belly River sediments. Fluvial-dominated deltas periodically abandon the active delta lobe due to upstream fluvial avulsion and begin building a new lobe at a position along the shoreline (the "deltaic cycle" of Scruton (1960)). Relative sea level is assumed to remain constant throughout.

This process has been interpreted to have been one of the underlying controls on the development and deposition of deltaic lobes within the Mississippi Delta system (Penland et al., 1988; 1987; Frazier, 1967; Coleman and Gagliano, 1964; Scruton, 1960). According to Frazier (1967) 5 deltaic complexes consisting of 16 individual delta lobes have been deposited since the end of the Holocene transgression (Figure 12.2). However, Penland et al. (1987) have since shown that not all of the Holocene Mississippi lobes were deposited simply by autocyclic switching during sea level highstand. There is a major rise of sea level represented by the development of the Teche shoreline system which separates the deposits of delta systems 1-3 from 4 and 5 within the recent Mississippi Delta system. However, most of the lobes are still

Figure 12.2: The 16 lobes of the Holocene Mississippi Delta, as determined by Frazier (1967) (from Elliott, 1986).


Baton Rouge
N
Abbeyville
New Orleans
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
0
100
km
RECENT MISSISSIPPI DELTA SYSTEM

interpreted to be deposited due to autocyclic switching (Penland et al., 1987). Many of these lobes are much larger than the lobes found in this study, but the most recent lobe (number 16) is approximately the same size as several of the Lea Park - Belly River cycles.

Delta lobe switching was the most probable control on the development of deltaic sediments in numerous ancient examples. Elliott (1976, 1975, 1974) interpreted six deltaic successions within the Carboniferous Abbotsham Formation (Bideford Group) in Devon, England as having been constructed and abandoned by this process. Numerous deltaic systems of the Eocene Wilcox Group of Texas are also interpreted to have developed autocyclically (Fisher and Brown, 1967). Within the Cretaceous sediments of the Western Canada Basin, deposition of deltaic "shingles" (individual coarsening-upward delta successions) within the Dunvegan Formation of Alberta are interpreted to be autocyclically controlled (Bhattacharya and Walker, 1991a; Bhattacharya, 1989). The Dunvegan shingles are components of larger-scale allomembers which are interpreted to reflect larger-scale allocyclic fluctuations in relative sea level. The shingles are roughly comparable in both thickness and lateral extent to the sandy cycles of the Lea Park - Belly River transition detailed in this study.

If autocyclic switching of shoreline position by deltaic avulsion were responsible for development of shoreline systems within the Lea Park - Belly River transition, each cycle would represent the deposits of a single episode of deltaic construction. The "transgressive" deposits between each cycle would represent the abandonment sediments, which were deposited after a deltaic lobe was abandoned and had subsided into shallow shelf depths. Such an interpretation creates a number of problems. One would be the positioning of the Lea Park - Belly River cycles with respect to each other. Autocyclic switching of deltaic position generally involves lateral movement along a shoreline position which, over time, does not move a great deal either basinward or landward (eg. the position of lobes of the Mississippi Delta relative to each other in figure 12.2). In contrast to this, shoreline cycles of the Lea Park - Belly River transition are offset up to several tens of kilometres in a basinward direction from the previous cycle, rather than migrating laterally along strike. An exception to this is the relationship between Cycles F and G (Fig. 12.1), which are located along strike from each other. Another problem concerns the relationship of the transgressive units to the cycles. These deposits appear to lie stratigraphically between each regressive cycle. If autocyclic switching were the controlling process, one might expect to find these deposits as lateral and coeval with the following regressive cycle, rather than always underlying it.

12.3.2: Allocyclic Fluctuations of Relative Sea Level

The alternative interpretation to autocyclic switching of shoreline position is that the regressive-transgressive nature of the Lea Park - Belly River transition is controlled by allocyclically-induced changes in relative sea level.

Tectonically or eustatically-controlled fluctuations in sea level would affect the whole basin, and cause the shoreline position to move according to the direction of sea level migration. Chapter 1 briefly discussed several studies of sediments within the Cretaceous of Alberta in which the authors concluded that relative sea level fluctuation was the dominant control on the spatial deposition of shoreline sediments and the resultant stratigraphy of these sediments. These include studies of the Cardium Formation (Leggitt et al., 1990; Plint, 1988; Plint et al., 1986, 1987), the Viking Formation (Pattison, 1991; Boreen and Walker, 1991; Downing and Walker, 1988; Power, 1988), the Muskiki and Marshybank Formations (Plint, 1990; Plint, 1991), and the Dunvegan Formation (Bhattacharya and Walker, 1991a, 1991b; Bhattacharya, 1989).

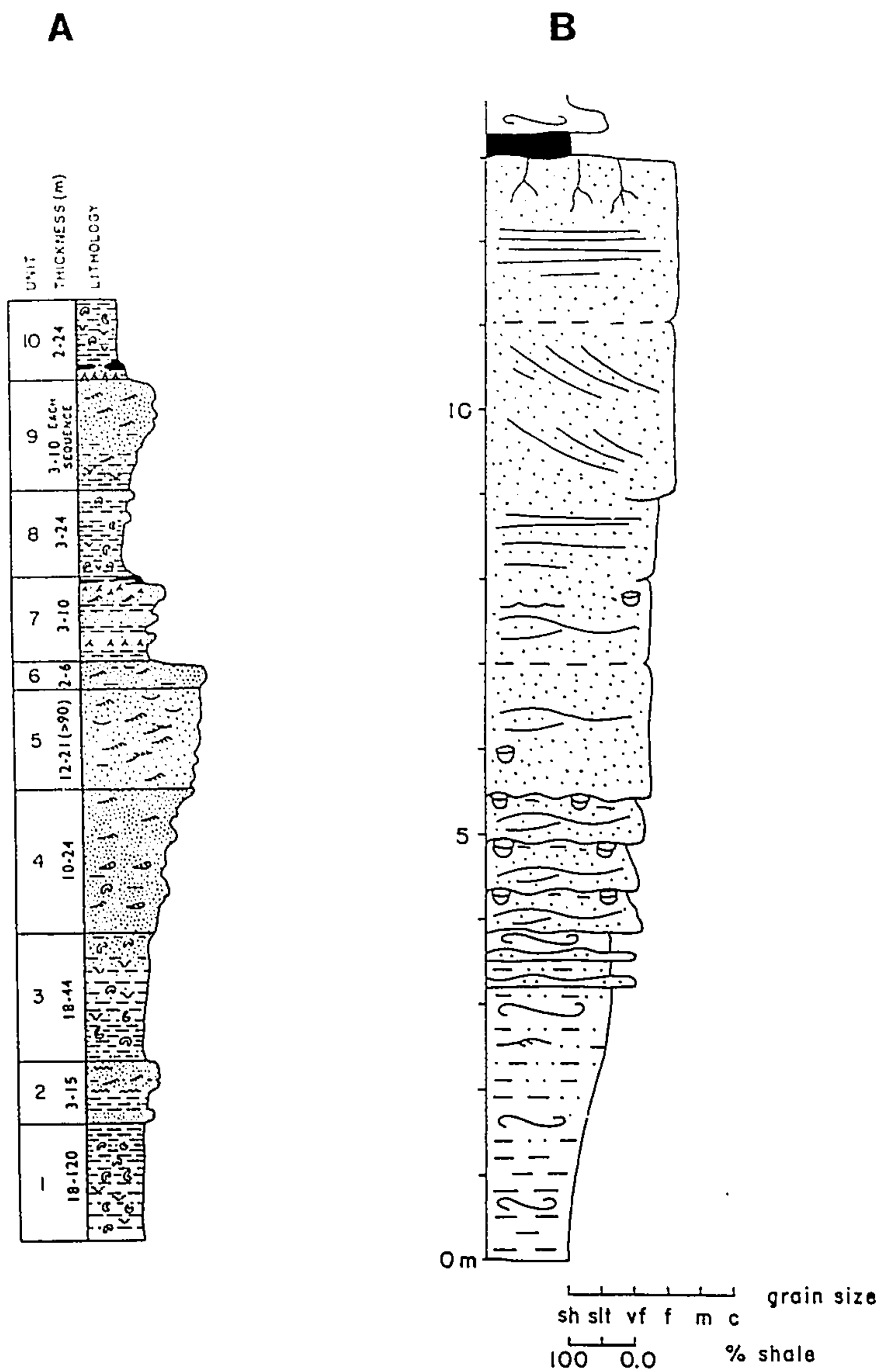
12.3.2a: Evidence For Allocyclicity

All of the above studies have shown that relative sea level fluctuations are recorded by flooding surfaces and/or erosional unconformities (or their correlative conformities) which are the bases and tops of allostratigraphic or sequence stratigraphic units, and which serve to separate sedimentary successions which are not genetically related. If fluctuations in relative sea level were the controlling factor on regressions and transgressions in the Lea Park - Belly River transition, these surfaces should also be present at the bases and tops of the cycles defined in this study. The cycles defined in this study are, in fact, bounded at their bases and tops by such surfaces. The following sections discuss the evidence for allocyclic fluctuations in relative sea level.

(1) Rapid Basinward Shift of Marine Cycles

A common feature of all the cycles described in this study is that the shoreface sediments of any given cycle sit very sharply on the underlying transgressive shelf sediments. This can be seen in almost any of the stratigraphic sections, core cross sections, or core photographs of shoreline successions in chapters 5-10 (eg. Figs. 10.2, 10.3). There is little or no transitional interval of interbedded shelf mudstones and sandstones, which would indicate a gradual shoaling from the shelf into the shoreface. The shelf sediments also show little or no evidence of shoaling upwards below the shoreface sediments. The ratio of mudstone, siltstone, and sandstone within the background sediments is usually approximately the same immediately below the shoreface as it is several metres lower in the succession. The contact itself is angular in numerous places, and small clasts of the underlying mudstones are commonly present immediately above the contact. In contrast, the progradational successions produced by autocyclic development of deltaic shorelines in the modern Mississippi Delta show a very gradual coarsening-upward succession, with a large transition zone between the shelf and the shoreface/distributary mouth bar (Fig. 12.3a) (Coleman and Wright, 1975). Deltaic successions of Cretaceous Dunvegan shingles (Fig. 12.3b), which are

Figure 12.3: Stratigraphic successions of the modern Mississippi Delta (A), and the Cretaceous Dunvegan Formation (B), showing gradual coarsening upwards successions due to autocyclic progradation of deltas (from Coleman and Wright, 1975; and Bhattacharya, 1989).

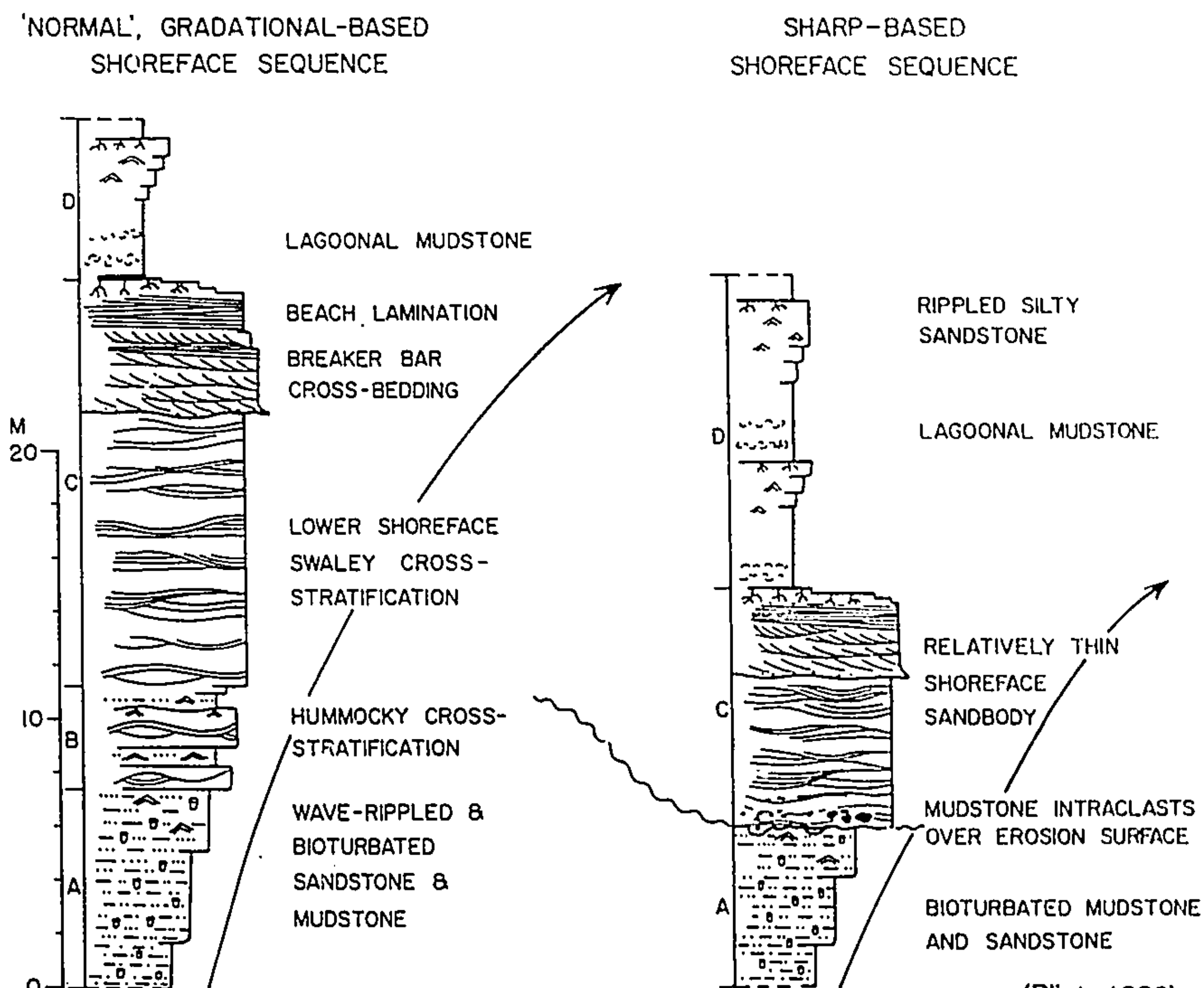

A
B
UNIT
THICKNESS (m)
LITHOLOGY
10
2-24
9
3-10 EACH SEQUENCE
8
3-24
7
3-10
6
2-6
5
12-21 (>90)
4
10-24
3
18-44
2
3-15
1
18-120
10
5
0 m
grain size
sh slt vf f m c
100 0.0
% shale

interpreted to be autocyclically produced, also show a gradual coarsening upward, indicative of slow progradation of the shoreline out across the shelf (Bhattacharya, 1989).

The sharp-based shoreface successions so characteristic of Lea Park - Belly River cycles would appear to indicate that the shoreface environment did not gradually prograde out over the shelf, but rather indicate that the depositional environment shallowed rapidly from shelf to shoreface depths. Sharp-based shoreface successions have been observed in other Cretaceous sediments within Alberta, most notably the Muskiki-Marshybank Formations and the Cardium Formation (Fig. 12.4) (Plint, 1988; Plint, 1991), where they are interpreted to have been produced by *forced regressions*. These result from a relative drop in sea level, which causes the shoreline to move rapidly basinward. While the sea level is dropping, the shelf will be eroded in front of the shoreface by waves and currents (Plint, 1988). This erosion surface is the contact between the shelf sediments and the rapidly prograded shoreface sediments. When relative sea level stabilizes at the new lower position, the shoreface will prograde out over the erosion surface. This process is shown diagrammatically in figure 12.5. Plint (1991) cites the sharp-based Ingleside sand of Texas as a Pleistocene model of sharp-based shorefaces. This sandbody prograded several tens of kilometres onto the Texas shelf during a mid-Wisconsinan sea level fall (Wilkinson et al., 1975). The Caravelas

Figure 12.4: Comparison between normal coarsening upward successions and sharp-based shoreface successions in the Cretaceous Cardium Formation (modified from Plint, 1988).

CARDIUM FORMATION

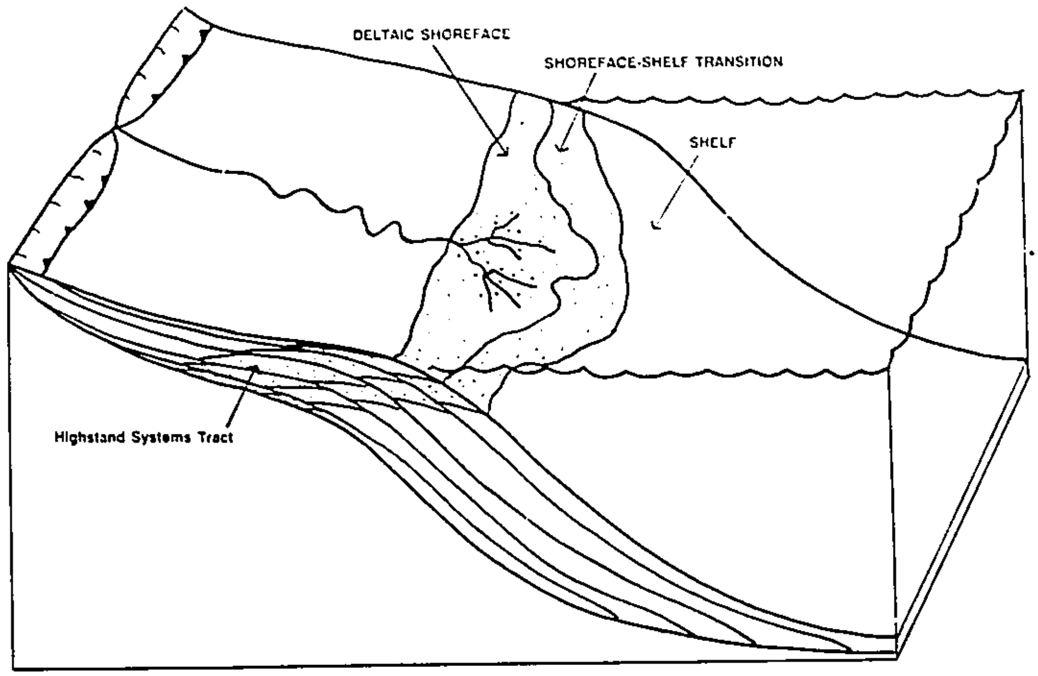


(Plint, 1988)

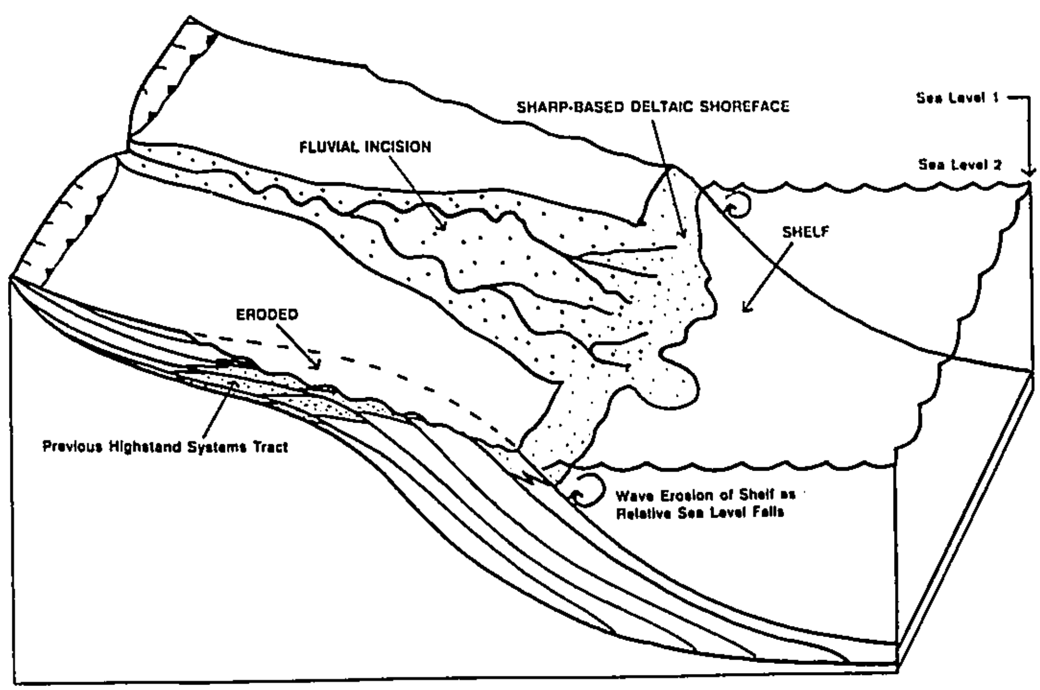
Figure 12.5: Interpreted mechanism for creation of sharp-based shoreface successions due to forced regressions caused by relative drop in sea level. (A) Prior to sea level fall. (B) After sea level fall. See text for discussion.

DEVELOPMENT OF SHARP-BASED SHOREFACE SUCCESSIONS

STAGE 1: RELATIVE SEA LEVEL HIGHSTAND



STAGE 2: FOLLOWING RAPID FALL IN RELATIVE SEA LEVEL



strandplain in southeast Brazil may be another example of a sharp-based shoreface formed due to forced regression (Dominguez et al., 1987).

The basal surface of each Lea Park - Belly River cycle is a regressive surface of erosion (RSE), but the preserved stratigraphy requires one minor modification of Plint's (1988) model. In the original model, the progradation and aggradation of the shoreface and the creation of the shelf unconformity are all occurring simultaneously. Deposition of a shoreline succession occurs over the entire area that the shoreline migrates across as sea level drops. The deltaic shoreline cycles of the Lea Park - Belly River are offset from each other, requiring that the process be slightly modified to include a period of erosional bypass and incision where the shoreline "jumps" to a new basinward position before it begins progradation. This explains the spatial relationship of the deltaic cycles, which are offset from one another by up to tens of kilometres. The Cardium shorelines of Plint (1988) were strandplains rather than deltas. One might expect that the nature of the shelf-shoreface transition might therefore be different for deltaic shorelines such as those in the Lea Park - Belly River transition due to increased local sediment supply. The only visible difference is that there is less visible dramatic erosion associated with the forced regressions of the Lea Park - Belly River transition than for the forced regressions of the Cardium Formation. This may be due to increased wave erosional energy during Cardium time. The deltaic shorelines observed in this study are interpreted to have been much less wave-dominated. Otherwise, the increased local sediment supply associated with a deltaic shoreline does not appear to have made the transition from shelf to shoreface significantly more gradual than for the strandplain shorelines of the Cardium Formation.

(2) Evidence of Subaerial Exposure and Incision

The basal contacts of the shoreline cycles are not the only regressive surfaces of erosion in the Lea Park - Belly River transition. There are many locations where fluvial and associated non-marine sediments sit directly on marine shelf sediments. The regressive surfaces of erosion where non-marine sediments sit directly on shelf sediments clearly represent subaerial erosion and/or exposure surfaces, and record times when relative sea level dropped, causing exposure of the shelf and incision of rivers into underlying sediments.

In many locations the fluvial channels or associated non-marine sediments sit directly on marine shelf sediments. The erosional surface at the contact between the marine and non-marine sediments is at the same approximate stratigraphic level as the regressive surface of erosion at the base of cycle. Regional cross section C-C' (Fig. 4.12) of chapter 4 shows fluvial channel sediments erosively sitting at approximately the same stratigraphic level as the base of the cycles in Cycles D, E, F, and H. Cross section E-E' (Fig. 4.14) shows that the regressive surface of erosion at the base of Cycle H, if traced back southwest of the preserved landward limit of deposition, sits at the same stratigraphic level as an extensive surface of exposure/erosion where non-marine sediments overlie the transgressive shelf sediments between Cycles G and H for a distance of 20-30 km. This apparent relationship between the subaerial surfaces of erosion and the regressive surfaces of erosion at the base of the cycles is very logical within the context of the forced regression model proposed to explain the sharp-based marine cycles. The forced regression model requires that there be a relative drop in sea level, which causes the shoreline to rapidly "jump" basinwards. The landward expression of this drop will be that the rivers will incise behind the migrating shoreline, creating subaerial surfaces of erosion which are coeval with the marine regressive surface of erosion at the base of the shoreline cycle. The two different types of surfaces are lateral expressions of the same drop in relative sea level.

There are other subaerial surfaces of erosion at the bases of fluvial channels and/or non-marine sediments which cannot be correlated so confidently with a specific cycle of shoreline sediments. In some cases, the process responsible for their development is envisaged to be the same as for those mentioned above. They are simply the landward representation of a relative drop in sea level which deposited a shoreline further out into the basin. In other cases, especially at the base of channels, these surfaces of erosion may be local in extent, and simply related to autocyclic migration of fluvial or deltaic channels over their floodplain. It would seem logical that these channels would not incise as deeply as those related to a relative drop in sea level, but this is unclear. It is difficult to distinguish between the two processes unless there is the immediate visible stratigraphic relationship with a deltaic shoreline cycle.

(3) Marine Flooding Surfaces

The two preceding sections interpret each shoreline cycle to have been initiated by a relative drop in sea level. Each cycle is also terminated by a relative rise is sea level. These rises are recorded in the flooding surfaces at the tops of each cycle. The transgressive surfaces generally do **not** show any evidence of erosion, such as winnowed lags at the surface, and for the most part appear to record non-erosional transgressions. The surfaces are simply referred to as flooding surfaces. They are characterized by an abrupt shift in facies from shallow shoreface/delta front sandstones to deeper shelf mudstones, siltstones, and sandstones. How much of the "transgressive sediments" are truly transgressive in nature is unclear. It was not possible to identify maximum flooding surfaces, and some of the muddy shelf sediments may be related to the progradation of the following cycle. However, it is clear that some relative rise of sea level did occur in the intercyclic times; this rise would be necessary to create the required accommodation space for the following regressive shoreline succession. It is likely, however, that these intercyclic relative sea level rises were relatively minor, and most of the permanent accommodation space was created by subsidence.

The cross sections suggest that the transgressions were generally not long-lived events, in which the shoreline gradually migrated back away from the basin, depositing a transgressive shoreline succession as it migrated. The only evidence of a gradual transgression that following Cycle G This records the gradual flooding of the coastal plain from brackish to fully marine, and further basinward, a shift from upper shoreface/fluvial to lower shoreface and finally to shelf sediments. If any transgressive shoreline sediments were deposited, they were either removed by transgressive erosion, or by erosion associated with the following drop in sea level which deposited the overlying cycle. This implies that the transgression migrated farther landward than its preserved record indicates. This is supported by the relationships indicated in all cross sections. When transgressive sediments are traced back in a landward direction, they do not just gradually "pinch out". Instead they are always eroded either by the base of the overlying cycle or by fluvial erosion landward of the overlying cycle.

The near total absence of transgressive shoreline successions may indicate that they were very poorly developed, which would be expected if the transgressive phases of relative sea level rise were of a short duration. During transgression, the shoreline and shallow shelf may have been "starved" of sediment because the coastal plain environment was acting as a depositional sink, and trapping sediment within fluvial channels and on floodplains as base level rose. Much of the sand filling the channels that were incised during the regression was presumably deposited during this infilling phase. This trapping of sediment in the channels and coastal plain would cause the end of active progradation of the shoreline.

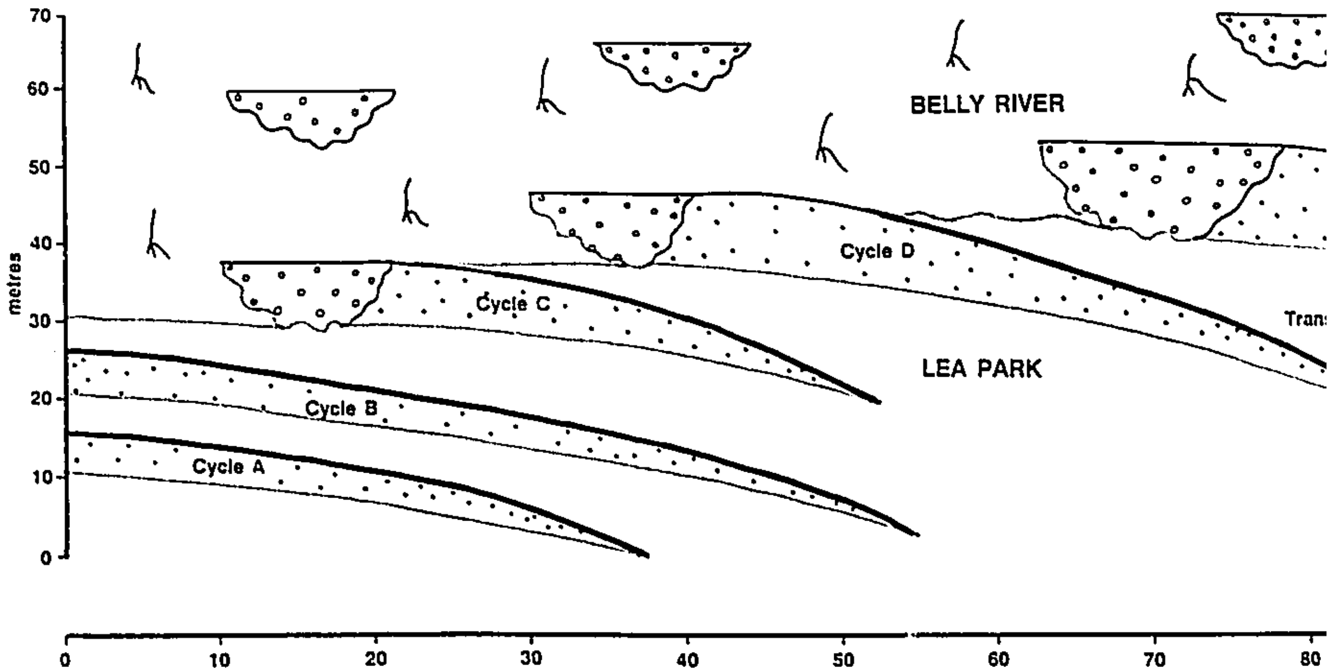
Holocene and Tertiary studies indicate that rapid rises in sea level affect sediment supply to the basin by trapping sediment in ephemeral coastal sinks (Kidwell, 1988), and can produce deepening-upward successions in coastal areas. An example of this may be the transgressive sediments which overlie the coastal plain deposits of Cycle G, and pass upwards from non-marine fluvial channels to brackish ponds, to open marine shelf sediments.

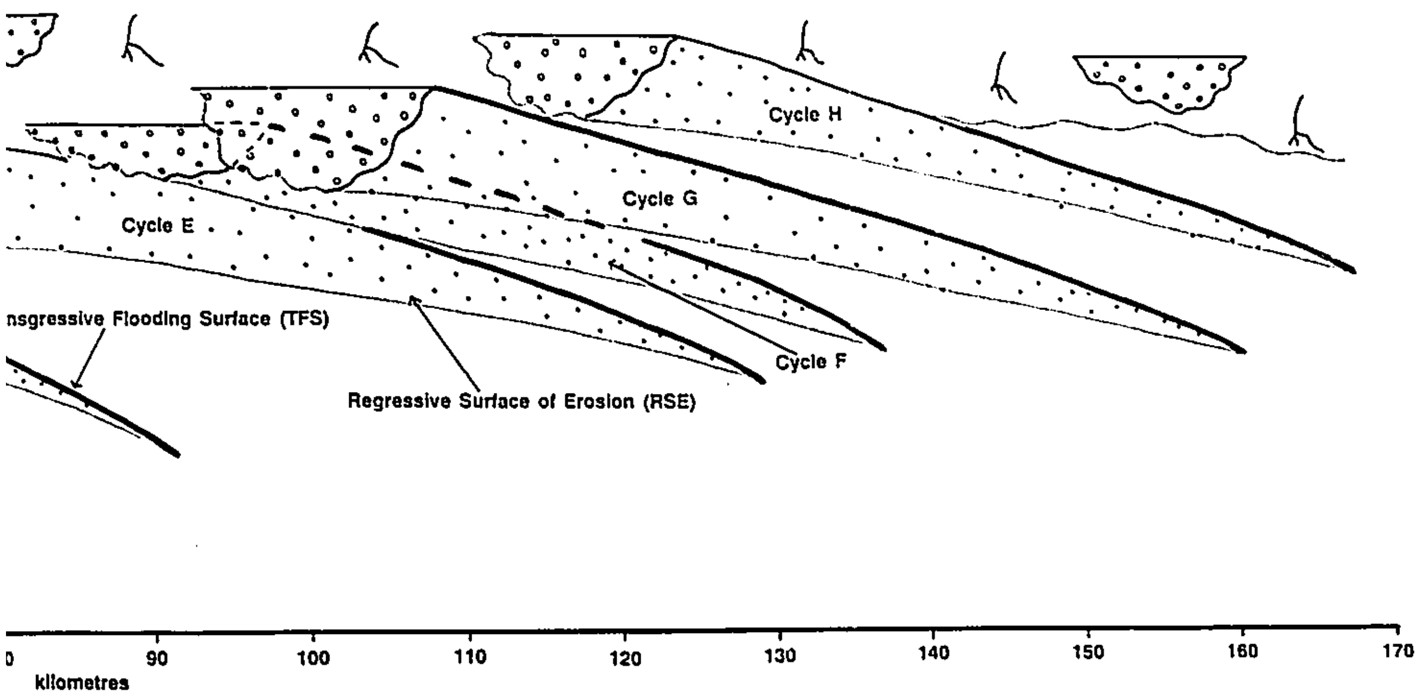
12.3.3: Proposed Allostratigraphy

The interpretation of the Lea Park - Belly River transition in the study area is that the sediments were deposited as a result of a series of relative falls and minor rises in sea level. These fluctuations in relative sea level created bounding discontinuities which allow the sediments to be divided into an informal allostratigraphy. Figure 12.6 is a modified version of the schematic cross section (Fig. 4.1) which shows the nature of the allostratigraphy, and the bounding surfaces upon which it is based. Each cycle can be thought of as an informal allomember. Its base is defined be a regressive surface of marine erosion, at the contact between the shoreface and the shelf sediments, and its top is defined by the flooding surface which was created by the transgression of the shoreline system. The regressive surface of erosion which defines the base of each cycle can be inferred (but not proven) to continue back in a landward

Figure 12.6: Modified version of figure 4.1, showing the informal allostratigraphy of the Lea Park - Belly River transition in the study area. The red lines indicate erosion surfaces due to relative fall in sea level and the heavy black lines denote transgressive flooding surfaces. Each cycle is bounded by these two types of discontinuities, defining each cycle as an allomember.

SCHEMATIC ALLOSTRATIGRAPHY OF THE LEA PARK - BELLY RIVER TRANSITION





Cycle H
Cycle G
Cycle E
nsgressive Flooding Surface (TFS)
Cycle F
Regressive Surface of Erosion (RSE)
0
90
100
110
120
130
140
150
160
170
kilometres

direction from the cycle, and is represented by subaerial erosional surfaces where non-marine sediments overlie shelf sediments. Following the drop of sea level at the base of each cycle, sediment began to aggrade and deltaic shoreline systems developed for most of the cycles. The fact that the shorelines were deltaic rather than strandplain shorelines does not appear to altered the nature of the effects of the forced regression model significantly from those observed by Plint (1988) for Cardium Formation strandplains. The lack of preserved transgressive shoreline deposits and the relatively non-erosive nature of the transgressive flooding surfaces indicates that relative sea level rises and their associated transgressions were less extensive, and far less important the geometry of the preserved stratigraphic successions than the relative drops in sea level. In essence, the Lea Park - Belly River transition in central Alberta is indicative of a period staggered relative sea level fall.

This interpretation does not imply that autocyclic processes of river avulsion and deltaic lobe switching were not occurring during deposition of Lea Park - Belly River sediments. Several of the cycles, such as Cycles D,E, and G, contain the evidence of more than one deltaic depocenter which were probably not active at the same time. This indicates that autocyclic delta switching was occurring within some of the cycles. The similarity in scale between the shingles of Dunvegan Formation (Bhattacharya, 1989), which are interpreted to be autocyclic deltaic lobes, and the cycles of the Lea Park - Belly River transition, which are interpreted as allocyclic units, would appear to be coincidental. It may imply simply that the rivers feeding Lea Park - Belly River shorelines were smaller and deposited smaller deltas than those interpreted for the Dunvegan Formation.

12.4: Determining the Amount of Relative Sea Level Fluctuation

A rough estimate of the amount of sea level drop associated with the base of each cycle can be determined by calculating the distance of the "jump" between the most landward position of a new cycle and the most seaward position of the underlying transgressive sediments. This can then be applied to a general figure for the slope of a shelf (determined from the literature) to determine the amount of vertical drop of sea level. Unfortunately, due to the amount of erosion of transgressive sediments at their landward edge within the Lea Park - Belly River transition, this becomes a difficult calculation to make because the data base is flawed, and the value of any resultant figure would be questionable. A very rough estimate of the amount of sea level drop associated with each cycle was, however, obtained by using the distance between the preserved landward edges of deposition of successive cycles instead of the distance between maximum transgression and cycle initiation. This assumption is also flawed, as the shoreline of each cycle prograded farther into the basin than just its landward edge of deposition, and the resultant calculation thus overestimates the distance between the position of the last shoreline of the underlying cycle and the first shoreline of the overlying cycle.

However, the assumption is that this overestimation is balanced out by the fact that the transgression between the two cycles may have flooded the underlying cycle back as far, if not farther, than its original shoreline position. Accordingly, these calculations should be "taken with a grain of salt", and treated as rough approximations.

There are numerous different values which can be used as a typical shelf gradient. Swift and Field (1981) give a gradient of 1/2000 for the Atlantic shelf off the northeastern United States. Reineck and Singh (1972) give a value of 1/1333 for the southern North Sea shelf off the coast of Germany. Johnson and Baldwin (1986) give a much steeper value of 1/576 as an average for shallow seas, but do not state how this value was obtained. The shelf gradient of the Cretaceous Seaway in Western Canada is thought to have been relatively low, perhaps more akin to the value for the U.S. Atlantic shelf of Swift and Field (1981). However, because the value of Johnson and Baldwin (1986) is said to be an average, calculations of drops in sea level will be compared using shelf gradients of both 1/2000 and 1/576.

The average distance between landward margins of deposition of successive cycles was estimated from figure 12.1. Because the landward margins of Cycles A and B lie to the west of the study area, this value could not be determined for Cycles A, B, or C. The values for Cycles D-H, along with the calculated values of sea level drop, are shown in Table 12.1.

Table 12.1: Estimation of amount of sea level drop associated with the initiation of each cycle in Cycles D-H. See text for discussion.

ESTIMATION OF AMOUNT OF RELATIVE SEA LEVEL DROP

Cycle	Distance of Shoreline Jump	Relative Sea Level Drop	
		1/2000	1/576
D	20 km	10 m	35 m
E	30 km	15 m	52 m
F	10-15 km	5-7.5 m	17-26 m
G	<5 km	<2.5 m	<9 m
H	20-30 km	10-15 m	35-52 m

This table shows that the amount of relative sea level drop at the base of a cycle with a shelf gradient of 1/576 can be quite large, and is probably an overestimate of the drops which occurred at the base of each cycle. A shelf gradient of 1/2000 gives results which are perhaps more appropriate. This shows that the sea level fluctuations involved are probably quite small, and average 10-15 metres. Regardless of which shelf gradient is used, it is clear that the process creating these relative drops in sea level should be capable of creating fluctuations on the scale of tens of metres.

12.5: Time Scale of Deposition of the Lea Park - Belly River Transition

The preceding sections discussed the nature of the regressive-transgressive character of the Lea Park - Belly River transition, and interpreted it as being the result of allocyclic fluctuations in relative sea level. The cause of these fluctuations in sea level has yet to be discussed. Before controls on sea level fluctuation can be discussed, the temporal scale of the sea level fluctuations must first be discerned.

Determining the frequency of cycle deposition within the Lea Park - Belly River transition is difficult for the same reason it is difficult to determine the age of these deposits. There is no direct radiometric dating of the Lea Park - Belly River transition in central Alberta available, and the biostratigraphy of this interval is poorly understood. As a result, there are no ages available for the

Lea Park - Belly River transition across the study area, and thus no direct way of determining the frequency of interpreted fluctuations in relative sea level.

An indirect, rough estimate of the amount of time taken up during the deposition of the Lea Park - Belly River transition within the study area can be determined by using published sedimentation rates for the Cretaceous of the Alberta Basin. Two recent studies by Chamberlain et al. (1989) and Lerbekmo (1989) give average sedimentation rates for the Campanian section of Alberta. Lerbekmo determined the average sedimentation rate during the Campanian to be about 60 m per million years, and rates determined by Chamberlain et al. (1989) vary between 77 and 90 m per million years (average of 83.5 m per million years). By applying these figures to the thickness of sediment accumulated in the study area, a very rough estimate of the time required to deposit the Lea Park - Belly River transition can be determined. As with the calculations of the amount of relative drop in sea level, these calculations should be treated only as approximations.

Figure 4.1 shows that approximately 75 m of sediment accumulated at the western edge of the study area from the time of initiation of Belly River deposition at this location to the time of Belly River deposition at the eastern edge of the study area. Using this figure, the time required to deposit the Lea Park - Belly River transition within the study area works out to 1.25 million years using the sedimentation rates of Lerbekmo (1989) or 898,000 years using the rates of Chamberlain et al. (1989). There are eight cycles present within the study area. Assuming that the time required to deposit each cycle is approximately equal, the time required to deposit one cycle (and the overlying transgressive sediments) is 112,250 years using the rates of Chamberlain et al. (1989) or 156,250 years using the rates of Lerbekmo (1989). The periodicity of relative falls in sea level therefore appears to be in the 100,000 to 150,000 year range.

12.6: Possible Mechanisms of Relative Sea Level Fluctuations

Much of the recent literature concerning sequence stratigraphy (or allostratigraphy), is concerned with determining the causative mechanism of relative sea level fluctuations. The relative sea level history of any basin depends upon three factors; (1) subsidence, (2) sedimentation rate, and (3) eustasy. Much of the work of Exxon seismic and sequence stratigraphers has focused on the eustatic nature of sea level fluctuations. In fact, they have often been forced into "an unspoken reliance" (Sloss, 1988, p. 1655) on eustasy to explain high-frequency sea level fluctuations for which no other suitable mechanism can be found. Other workers believe that the importance of subsidence and sediment supply, especially in tectonically active basins has been underestimated by Exxon researchers (Embry and Podrudski, 1988). Determining the relative importance of each factor can be very difficult, and has often proved impossible. Burton et al. (1987) state that with our present understanding of how these factors interact, and the sedimentary record they leave behind, it is beyond our ability to determine which factor is controlling relative sea level fluctuations

A detailed discussion of the various mechanisms which control relative sea level fluctuation is beyond the scope of this study. However, it is important to consider all of the possible mechanisms which may have controlled sea level during deposition of the Lea Park - Belly River transition. The following sections briefly discuss the known mechanisms of sea level fluctuation, and whether or not they can be considered as possible controls on the relative sea level fluctuations which were responsible for deposition of the coastal cycles observed in this study.

The most important factor concerning a given mechanism's appropriateness as a possible control on relative sea level fluctuation for any system is the time scale on which the mechanism affects sea level. In the case of the Lea Park - Belly River transition, the maximum time between each fall (or each rise) in relative sea level is thought to be between 100,000 and 150,000 years. This approximates the time span of fourth-order cycles of Exxon sequence stratigraphy (Vail et al., 1977). The controlling mechanism must be capable of causing relative sea level fluctuations of up to a few tens of metres over this time span. A secondary concern should focus on whether or not a given mechanism is likely to affect sea level with repetitive cyclicity. Eight shoreline cycles were deposited within the study area. Unless a mechanism is likely to cause cyclic rises and falls in sea level, it is probably not a possible control of relative sea level fluctuation during Lea Park - Belly River time.

The known mechanisms which may be responsible for sea level fluctuation can be broadly grouped into three categories:

(A) Tectonically-controlled fluctuations, both eustatic and local

(B) Glacio-eustatic fluctuations

(C) Non-glacial fluctuations induced by climatic change

12.6.1: Tectonically-Controlled Fluctuations in Sea Level

(A) Tectono-Eustatic Fluctuations

Tectono-eustatic fluctuations in sea level are generally considered to be caused by changes in mid-ocean ridge volume as a result of variations in sea-floor spreading rates (Hallam, 1984; Donovan and Jones, 1979; Pitman) 1978 or by variations in the volume of ridge volcanic material (Schlanger et al., 1981). Both of these processes serve to change the volume of the ocean basin itself, and thus affect the level of the sea surface. Hallam (1984, p. 239) regards these tectono-eustatic fluctuations in sea level as having been "generally more important than glaciation and deglaciation in controlling sea level variations prior to the Quaternary". However, evidence favouring this opinion is far from conclusive, especially with regards to relatively high-frequency fluctuations in sea level. The rate at which these processes might affect sea level is very slow, with the rate of sea level change being approximately 1 cm/1000 years (Gornitz and Lebedeff, 1987; Pitman, 1978). These processes are thought to be responsible for sea level fluctuations with a periodicity of 10->50 million years (Pitman and Golovchenko, 1983; Donovan and Jones, 1979), and thus are thought to be the main control on 1st and 2nd order Exxon cycles. These periodicities are several orders of magnitude too slow to be responsible for the Lea Park - Belly River fluctuations in sea level.

(B) Fluctuations Due to Regional Tectonics

Local tectonic mechanisms, which affect sea level on a geographically restricted scale, have also been proposed to explain sea level fluctuations, especially within the deposits of tectonically-active basins such as foreland basins. Relative sea level changes in foreland basins may be due to tectonically-induced changes in the subsidence rate of the basin due to increased loading or unloading of the lithosphere.

The Western Interior foreland basin was created by deformation associated with the accretion of allochthonous terranes onto the western edge of the North American Plate during the Late Jurassic and throughout the Cretaceous (Stott, 1984). Cant and Stockmal (1989) have proposed that there is a correlation between the docking of allocthonous terranes, and the deposition of numerous major clastic wedges in the Alberta basin, including the onset of Belly River deposition (caused by docking of the Insular superterrane). These clastic wedges are often bounded by unconformities caused by relative changes in sea level. The docking of allochthonous terranes causes thrust faulting and crustal shortening in the mountain belt adjacent to the foreland basin, which in turn cause loading of the crust, downwarping of the lithosphere and increases the subsidence rate within the basin (Beaumont, 1981; Jordan, 1981). The increased rate of subsidence would cause a relative rise in sea level. Progradation would occur during the following period of tectonic quiescence, when sediment shed from the mountains (due to the renewed uplift) would begin to fill in the new accommodation space.

The rates of the tectonic processes would appear to be rapid enough for them to be a possible cause of the Lea Park - Belly River fluctuations. Jordan et al. (1988) state that thrusts can move at rates of up to 30 m/1000 years, with an average of 6 m/1000 years. Lundberg and Dorsey (1988) state that changes in subsidence rates during collision events in the Western Taiwan foreland basin could be as high as 3 m/1000 years. Subsidence occurring at this rate could cause relative rises in sea level over a period of tens or hundreds of thousands of years. Relative sea level fluctuations in the Turonian Cardium and Cenomanian Dunvegan Formations of Alberta have been interpreted as most likely having been caused by local tectonic activity in the Cordillera (Bhattacharya, 1989; Bergman and Walker, 1988). King (1990) also attributes relative sea level fluctuations in the Campanian Arcola member of Alabama to be the result of foreland tectonic processes in the western United States.

There are, however, some additional problems associated with this interpretation if applied to the Lea Park - Belly River transition. The first concerns the cyclic nature of the transition. The Lea Park - Belly River transition in the study area would require eight separate thrusting events with a periodicity of 100,000-150,000 years to produce the eight shoreline cycles. The periodicity of these events precludes them being directly caused by the docking of eight different terranes. One possible solution to this problem is the application of the concept that subsidence related to tectonic events in foreland basins is not continuous (Jameison and Beaumont, 1988; Jordan et al., 1988, Lundberg and Dorsey, 1988; Paola, 1988). Lundberg and Dorsey (1988) show that subsidence rates during collision in the Western Taiwan foreland basin varied between 0.15-3.0 m/1000 years. The response of the mountain belt and foreland basin to a collision event may therefore have been thrusting, uplift in the mountains and foothills, and basinal subsidence which occurred as pulses of activity separated by periods of quiescence. These periods of thrusting and subsidence may have been able to occur on the time scale of Lea Park - Belly River fluctuations in relative sea level.

Staggered periods of subsidence related to foreland thrusting may be able to explain the relative rises in sea level which end each cycle in the Lea Park - Belly River transition, but do not really explain the problem of how the relative drops in sea level at the base of each cycle were caused. The incised fluvial channels and sharp bases of the shoreface cycles would require that, in addition to the rapid subsidence associated with loading in the Cordillera, there was rapid uplift in the basin (not adjacent to the basin). If the load on the lithosphere caused by thrusting is removed, the basin would isostatically rebound and uplift, causing a relative drop in sea level. The crust is capable of rebounding on a time scale of 1000-10,000 years (Turcotte and Schubert, 1982). The question is whether or not the load can be removed quickly enough. Under normal circumstances, it would not appear that the increased load in the Cordillera caused by the thrusting could be removed quickly enough to cause isostatic rebound of the lithosphere within the required range of periodicity. Average rates of continental erosion are only 0.05 m/1000 years (Pinet and Souriau, 1988), which are clearly too slow to account for the relative drops in sea level required. Rates are undoubtedly higher in mountainous regions, but are probably still too slow. Jameison and Beaumont (1988) state that the load may be removed by extensional unloading of the thrust blocks. In other words, once the compression causing the thrust has ceased, the thrust blocks may slide back down the planes of movement, causing unloading of the lithosphere. This theory is presently rather speculative, and no estimate of the rates at which unloading might occur have been made. The only driving force on unloading would be gravity, but this is unlikely to be able to overcome inertia and cause movement at the same rate as the thrusting, which was aided by the compressive force of accretion. It therefore seems unlikely that uplift associated with unloading of the basin crust could explain the relative falls in sea level within the Lea Park - Belly River transition. Another cause for the drops in sea level must be responsible.

One solution to this problem may be to superimpose subsidence related to foreland thrusting on a basin which is undergoing a sea level drop caused by an outside factor. In chapter 2, it was noted that the Lea Park - Belly River transition in central Alberta coincides approximately with the 80 ma third order eustatic drop in sea level of the Exxon curves of Haq et al. (1988). If sea level were dropping over a time period of 1-2 million years throughout deposition of the Lea Park - Belly River transition, superimposition of fluctuating subsidence rates with a periodicity of hundreds of thousands of years may be able to explain the fourth order fluctuations observed in this study. This may be accomplished by varying the rates of subsidence as described above while eustatic sea level is falling at a constant rate, or by varying the rate of eustatic fall while subsidence is occurring at a constant rate, or a combination of variations in both processes. As with foreland subsidence, there is no a priori reason to expect eustatic sea level fluctuations to occur a constant rate. The Alberta basin was undergoing increased subsidence during the Campanian as a result of the Laramide orogeny (Chamberlain et al., 1989). If the 80 ma eustatic drop did occur at varying rates, the net effect on Alberta basin sedimentation may have been to create the fourth order cycles observed in the Lea Park - Belly River transition. At times of higher rates of eustatic sea level drop or lower rates of subsidence due to relative tectonic quiescence, the effect would be to cause a rapid relative drop in sea level, resulting in fluvial incision and creating the sharp based shoreface successions. At times of greater subsidence due to active thrusting or lower rates of eustatic drop, the net result would be a relative rise in sea level, which would have caused the intercyclic transgressions observed. As was mentioned earlier, relative sea level fluctuations which occurred during deposition of the Campanian Arcola member of Alabama have recently been interpreted as having been caused by foreland tectonics related to the Laramide orogeny in the United States (King, 1990). The Arcola member is thought to be approximately 80 million years old, and is therefore time equivalent to the estimated age of the Lea Park - Belly River transition within the study area. The periodicity of the cycles in these sediments averages 103,000 years, which is very similar to the assumed periodicity of the sea level fluctuations in this study. King (1990) does not convincingly explain why he interprets the cyclicity in the Arcola as being tectonically-induced, but nonetheless, the similarities with the Lea Park - Belly River transition are striking.

The plausibility of this hypothesis hinges on what the effect of the 80 ma third order eustatic drop was in central Alberta, and whether it really was occurring during deposition of the Lea Park - Belly River transition within the study area. As mentioned in chapter 2, the absolute age of the transition in the study area is unknown. Van Wagoner et al. (1990) state that the effects of the 80 ma drop in the Powder River basin of the United States was to cause the shoreline to retreat as far east as South Dakota, and that "it is probable that the sea retreated from a large part of the North American craton at or about 80 Ma." (Van Wagoner et al, 1990, p. 34). This must have occurred during deposition of the Lea Park - Belly River transition somewhere in the Alberta basin. Onset of Belly River deposition at the western edge of the basin likely began prior to 80 Ma, and the seaway retreated beyond southeastern Alberta by 78 Ma (Goodwin and Deino, 1989). It is therefore at least possible, and perhaps probable, that the Lea Park - Belly River transition would have been deposited within the study area at or about 80 Ma, during the period of eustatic drop.

Another local tectonic mechanism that has been proposed as a cause of sea level fluctuation is that of intraplate stresses (Cloetingh, 1988a; 1988b; 1986). These studies suggest that changes in intraplate stress fields and their interaction with basin subsidence and uplift can cause relative fluctuations in sea level with a magnitude of up to a hundred metres and a rate of 0.01-0.1 m/1000 years. This translates to periodicities of sea level cycles in excess of 1 million years, which is appropriate for third-order cycles. Several authors have attributed sea level fluctuations to similar intraplate stresses. Embry and Podruski (1988, p. 73) attribute third-order fluctuations in sea level observed in the deposits in the Mesozoic Sverdrup Basin in Arctic Canada to the "Cloetingh Tectonic Model". Arnott (1988) has interpreted that changes in intrabasinal subsidence related to the Sweetgrass Arch (an intrabasinal feature of the

Western Interior basin) were responsible for creating relative sea level fluctuations with a periodicity of approximately 400,000 years during deposition of the Albian/Cenomanian Bootlegger Member in Montana. This may be too fast for the intraplate mechanism proposed by Cloetingh (1988a, 1988b), although it may be a possible contributing factor in Sweetgrass Arch movement. The rate at which this mechanism operates would seem to be an order of magnitude too slow to explain the Lea Park - Belly River fluctuations. It thus seems an unlikely cause for the fourth-order fluctuations, although it may have been a contributing factor in the third-order drop associated with progradation of the Belly River wedge.

12.6.2: Glacio-Eustatic Fluctuations in Sea Level

Changes in the volume of continental ice sheets, especially throughout the Pleistocene, have been shown to dramatically affect sea level on a global scale. Pitman (1978) states that the complete melting of all the present land ice would raise sea level by 40-50 metres. During the Pleistocene glaciation, sea level was at least 100 metres lower than at present, causing much of the continental shelf to be subaerially exposed (Donovan and Jones, 1979). This process operates at a relatively rapid rate (10 m/1000 years - Gornitz and Lebedeff, 1987), and is therefore more than capable of effecting large sea level fluctuations on a scale of as little as a few thousands of years.

The cause of the fluctuations in land ice mass is thought to be related to changes in solar insolation due to predictable and cyclic orbital variations of the Earth known as Milankovitch cycles (Hays et al. 1976). These orbital variations are caused by changes in the gravitational field experienced by the earth (Berger, 1984). The differences in incoming solar insolation are amplified by climatic factors, which causes growth or decay of the ice sheets (Barron et al., 1985). Calculations for the recent and Pleistocene show that Milankovitch periodicities occur at 400,000 and 100,000 years due to fluctuations in the eccentricity of the earth's orbit around the sun; at 41,000 years due to fluctuations in the obliquity of the earth's axis; and at 19,000 and 23,000 years (average 21,000) due to variations in precession of the equinoxes (related to the perihelion) (Berger, 1984). These periodicities correspond extremely well to cyclic variations in sea level during the Pleistocene (Hays et al., 1976). Variation in the orbital behaviour of the earth over time may mean that the periodicities of the Milankovitch cycles vary somewhat for pre-Quaternary sediments, but the overall magnitude is very similar (Collier et al, 1990). The cyclicity of sequences in deposits of ancient carbonate sediments over a wide geological time span have been attributed to the theory, including Triassic carbonates of northern Italy (Goldhammer et al., 1987) and Austria (Fischer, 1964), Cambrian cycles in the Appalachians of Virginia (Koerschner and Read, 1989), and Pre-Cambrian carbonate cycles in northern Canada (Grotzinger, 1986). Examples of cyclicity in siliciclastic sediments which are unequivocally due to Milankovitch forcing are few. A recent study by Mitchum and Van Wagoner (1991) attributes fourth- and fifth- order sea level fluctuations in the Eocene Wilcox and Queen City Formations of Texas to Milankovitch cycles. The main evidence for this interpretation is the periodicity of the sedimentary cycles, which are 100,000-200,000 years (fourth-order) and 10,000-20,000 years (fifth-order).

The major problem faced in attempting to interpret any Cretaceous sea level fluctuations as being due to Milankovitch forcing is that the Cretaceous is widely thought to have been an ice-free period, with a relatively equable (low seasonality) climate. This conclusion is based on the oxygen isotope record of these sediments, the presence of floras and faunas indicative of warm temperatures in areas of high paleolatitudes, and the lack of any physical sedimentary evidence for extensive land glaciation (Matthews, 1984; Barron et al, 1981). Without continental glaciers to act as a nucleus for changes in the volume of oceanic water, it is generally thought that Milankovitch forcing could not cause significant fluctuations in sea level. This has prevented the interpretation of several units in the Cretaceous of Alberta, which contain sedimentary sequences of fourth-order cyclicity, as having been produced as a result of glacio-eustatic fluctuations in sea level. These include the Cardium, Viking, and Dunvegan Formations.

Recently, some researchers have begun to question whether the Cretaceous truly was an ice-free period. Sloan and Barron (1990) state that computer models based on Cretaceous paleogeography and oceanic circulation patterns indicate that, even at its warmest times, continental interiors at high latitudes would have experienced below freezing temperatures during the winter months. Frakes and Francis (1988) interpret the presence of lonestones in the Albian Bulldog Shale of central Australia to indicate the existence of sea ice in Australia during the Cretaceous, when the continent was at high latitudes. Pirrie and Marshall (1990) concluded that stable isotope records from Santonian-Maastrichtian macrofossils in Antarctica indicate the presence of cool polar regions at this time. Rich et al. (1988) also state that isotopic and paleontological data indicate that mean annual temperatures at this time in Australia were less than 5 degrees Celsius, with some seasonality. Plint (1991) has cited some of these studies as possible evidence for Cretaceous glaciation during the Coniacian/Santonian, and suggests that a 100,000 year cyclicity of sea level fluctuations observed in the Muskiki and Marshybank Formations of Alberta may therefore be indicative of Milankovitch forcing of glacial cycles. The correspondence of the inferred Lea Park - Belly River cycle periodicity to that of the eccentricity periodicity (approx. 100,000 years) of the Milankovitch cycles makes this mechanism very attractive as a choice for the cause of the observed cyclicity.

None of this evidence, however, indicates the existence of large-scale, permanent continental glaciation during the Cretaceous. The presence of ephemeral ice in continental interiors may not be sufficient to cause major fluctuations in sea level. Therefore, in spite of the recent data, any interpretation involving Cretaceous glaciation is still questionable. Nonetheless, it remains a distinct possible forcing mechanism of both the third-order eustatic drop at 80 Ma associated with the initiation of Belly River sedimentation, and the fourth-order cyclicity observed within the Lea Park - Belly River transition.

12.6.3: Non-Glacial Climate-Induced Fluctuations in Sea Level

Changes in climate can be caused by a number of factors, among them changes in oceanic and wind circulation patterns caused by changes in plate orientations throughout geologic time. As mentioned earlier, the rate of tectonic processes such as plate movement is too slow to have caused the fourth-order sea level fluctuations observed in this study.

The glacial cycles caused by Milankovitch periodicities in the earth's orbit are another obvious pattern of climate change. Variations in the amount of glacial ice may not be the only effect of Milankovitch periodicities. Barron et al. (1985) have used computer simulations to show that orbital variations may also have affected storm and precipitation patterns, and thus the sedimentation rate, with a similar periodicity as the glacial cycles. Basically the theory holds that the amount of precipitation is a function of the land-sea thermal contrast, and that differences in the solar insolation cause much greater changes in land temperature than oceanic temperature due to the difference in the thermal inertia of land and water. Thus the orbital parameters which cause glacial cycles could also have caused periods of intensified storms, especially if the paleogeography was such that it amplified land-sea thermal contrast. Barron et al. (1985) hold that Cretaceous paleogeography did exert a substantial control on precipitation patterns, and that Milankovitch cyclicity may have controlled sedimentation patterns at this time. The theory has been used to explain Milankovitch periodicities observed in Cretaceous rhythmic bedded pelagic sediments in Colorado, Italy, and France (Barron et al., 1985). Periods of high precipitation/high sediment yield will be represented by organic mud deposits, while periods of low precipitation/low sediment yield are characterized by carbonate deposition. Cyclicity in the Cretaceous Arcola Limestone Member of Alabama has also been interpreted to represent climatic forcing (Bottjer et al., 1986), although, as previously mentioned, this has recently been reinterpreted by King (1990) as being tectonically-induced cyclicity. Barron (1989) also holds that the apparent abundance of storm-dominated shoreline deposits in the Cretaceous may be attributed to an increased intensity of storms at this time.

Interpretations of climatic-cyclicity in shallow-water siliciclastic sediments are far less common. The only known study to propose this interpretation is that of Wright et al. (1989), which states that climate-induced variations in sediment yield are the most likely explanation of fourth- and fifth-order sea level fluctuations in shoreline cycles of the Cretaceous Point Lookout Sandstone of the San Juan Basin in New Mexico and Utah. This interpretation is based upon the apparent correspondence with of the shoreline cycles with Milankovitch periodicities. Without good evidence for Cretaceous glaciation, Wright et al. (1989) feel that the climate-control model is the most likely mechanism of sea level fluctuation.

The principles of variations in precipitation patterns proposed by Barron et al. (1985) seem logical, but it remains to be proven whether these changes could produce the required changes in sediment yields to cause basin-wide fluctuations in sea level on the order of tens of metres, incision and erosion creating unconformities, and the rapid movements of shorelines required to explain the fluctuations in the Lea Park - Belly River transition. Further testing of this theory is required before it can really be seriously regarded as a possible control of Lea Park - Belly River relative sea level fluctuations.

Changes in climate can affect sea level in another manner. Changes in air temperature will cause the ocean to change temperature as well. With the change in temperature, the ocean will expand and contract, causing sea level to rise and fall (Gornitz and Lebedeff, 1987; Barnett, 1983; Gornitz et al., 1982). As with all other climate changes, this process could also be forced by Milankovitch periodicities. Gornitz et al. (1982) measured a rate of sea level rise of 12 cm/100 years over the past century, and attribute much of this to thermal expansion of the upper ocean as the air temperature has warmed. Etkins and Epstein (1982) calculated that between the years 1890-1940, the sea level rose 2.4 cm due to thermal expansion. This rate is fast enough to cause fourth-order cyclicity if continued for a period of tens of thousands of years.

12.6.4: Summary of Possible Controls on Sea Level Fluctuations

The record left by the relative sea level fluctuations during Lea Park - Belly River time is a combination of the effects of subsidence, sediment supply and possible eustatic variations. Discerning the relative importance of these factors is a difficult, and sometimes impossible task. However, the preceding discussion has allowed most of the mechanisms which cause sea level fluctuation to be ruled out in the case of the Lea Park - Belly River transition. Of all of the mechanisms described above which could possibly control the fourth-order fluctuations observed within the Lea Park - Belly River transition, as well as the 80 Ma eustatic third-order drop upon which the fourth-order fluctuations are superimposed, only two real possibilities exist. The first is that the fourth-order cyclicity is due to interactions between basin subsidence due to loading and the third-order drop, and the second is that all of the cyclicity and sea level fluctuations are due to Milankovitch-induced glacio-eustasy. Both explanations have their pros and cons. In the tectonic model, much hinges on the assumption that subsidence due to foreland thrusting was episodic, and that the basin was undergoing a third-order drop in sea level at the time. The discussion of this concept shows that this was indeed very likely. Accepting this process as the controlling mechanism on the fourth-order cyclicity still requires that an explanation regarding the third-order sea level drop be found. If this drop is eustatic, one may have to fall back on the words of Sloss (1988, p. 1655), and into an "unspoken reliance on glacio-eustatics". If one must rely on glacio-eustasy for the third-order drop, it therefore automatically should become the preferred choice for explaining the fourth-order cyclicity as well. The absence of evidence for large-scale Cretaceous glaciation may perhaps invalidate the possibility of glacio-eustasy as a control. It is apparent that these questions cannot be presently answered, and there is much we do not understand about sea level fluctuations throughout geologic history, especially during apparently ice-free times such as the Cretaceous.

12.7: Application of the Lea Park - Belly River Allostratigraphy to Published Concepts of Sequence Stratigraphy

Allostratigraphy is a broad term, and because any type of discontinuity can be an allostratigraphic bounding surface, all of the proposed methods of sequence stratigraphy can be classified as specific types of allostratigraphy. The most widely-used sequence stratigraphic scheme is, of course, Exxon sequence stratigraphy. Recently, however, as sequence stratigraphy has been applied by non-Exxon researchers, some workers have found that Exxon sequence stratigraphic principles do not adequately describe and classify the sediments that they have observed. The only other method presently published, however, is that of Galloway (1989), who has proposed a method called genetic sequence stratigraphy. This section briefly examines these two methods of sequence stratigraphy, and discusses their applications as a classification system and interpretive scheme for the allostratigraphy developed in this study for the Lea Park - Belly River transition.

12.7.1: Exxon Sequence Stratigraphy

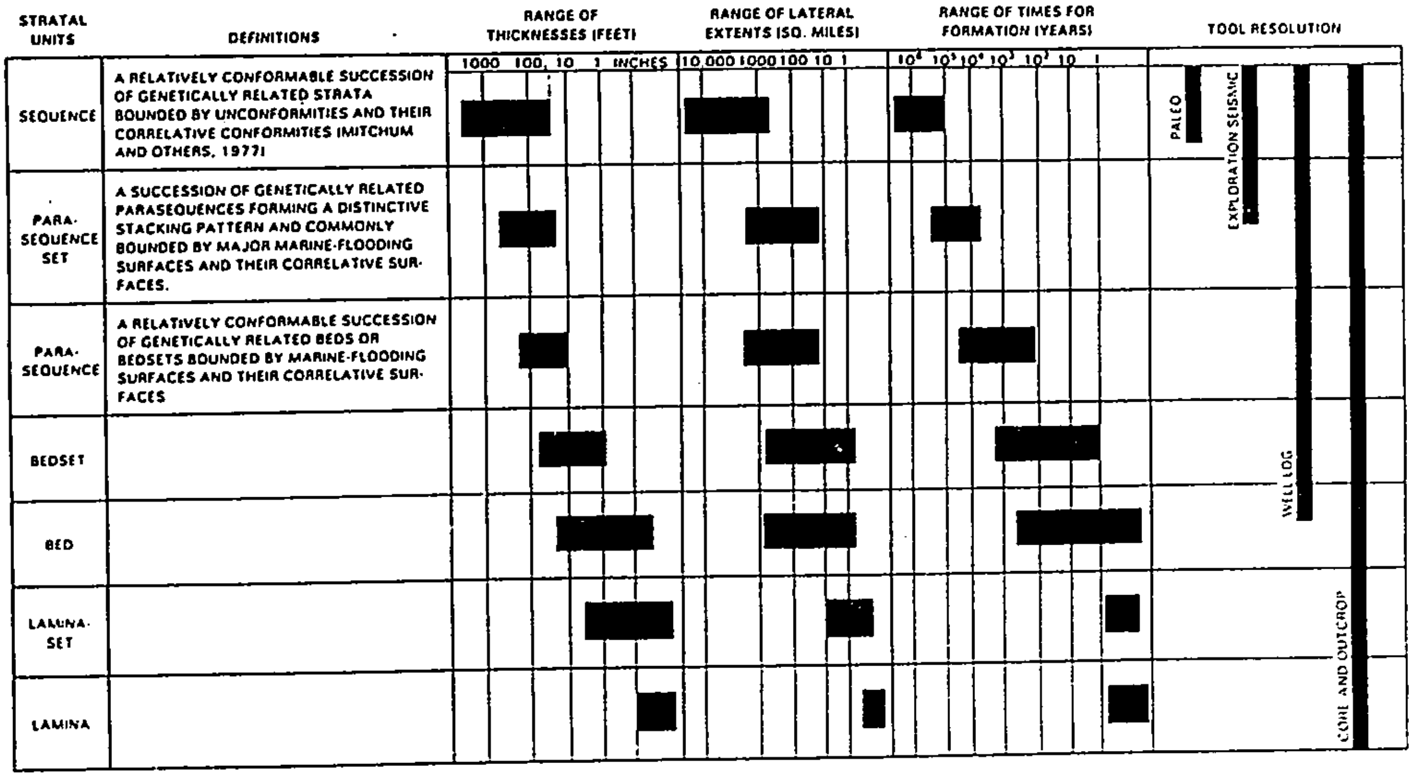
A brief history of the development of the concepts of sequence stratigraphy is given in Chapter 1. Since its original inception in the 1970's, seismic and sequence stratigraphy have undergone internal evolution within the Exxon corporation to include new concepts and terminologies as more information has become available, and new terms are continuing to be proposed presently. The basic terminology of stratal units used in sequence stratigraphy and their definitions presently in use is given in Table 12.2 (Van Wagoner et al., 1990). It should be noted here that the scales applied to each term in this table are not absolute, and some practitioners of Exxon sequence stratigraphy believe that the principles should apply on all scales, both temporal and spatial (H. Posamentier, pers. comm.).

The basic unit of sequence stratigraphy is the *sequence*, which is defined by Mitchum (1977) as:

> *"a relatively conformable succession of genetically related strata bounded by unconformities and their correlative conformities."*

Sequence boundaries are unconformities which form as a result of only a relative drop in sea level. This is a very restricted use of the term unconformity.

Table 12.2: Definitions of Exxon sequence stratigraphic terms (modified from Van Wagoner et al., 1990).


STRATAL UNITS
DEFINITIONS
RANGE OF THICKNESSES (FEET)
1000 100 10 1 INCHES
RANGE OF LATERAL EXTENTS (SQ. MILES)
10,000 1000 100 10 1
RANGE OF TIMES FOR FORMATION (YEARS)
10^6 10^5 10^4 10^3 10^2 10 1
TOOL RESOLUTION
SEQUENCE
A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED STRATA BOUNDED BY UNCONFORMITIES AND THEIR CORRELATIVE CONFORMITIES (MITCHUM AND OTHERS, 1977)
PARA-SEQUENCE SET
A SUCCESSION OF GENETICALLY RELATED PARASEQUENCES FORMING A DISTINCTIVE STACKING PATTERN AND COMMONLY BOUNDED BY MAJOR MARINE-FLOODING SURFACES AND THEIR CORRELATIVE SURFACES.
PARA-SEQUENCE
A RELATIVELY CONFORMABLE SUCCESSION OF GENETICALLY RELATED BEDS OR BEDSETS BOUNDED BY MARINE-FLOODING SURFACES AND THEIR CORRELATIVE SURFACES
BEDSET
BED
LAMINA-SET
LAMINA
PALEO
EXPLORATION SEISMIC
WELL LOG
CORE AND OUTCROP

A sequence is therefore the sediments deposited between two periods of relative sea level fall.

The building blocks of sequences are termed *parasequences.* A oparasequence is defined by Van Wagoner et al. (1990) as:

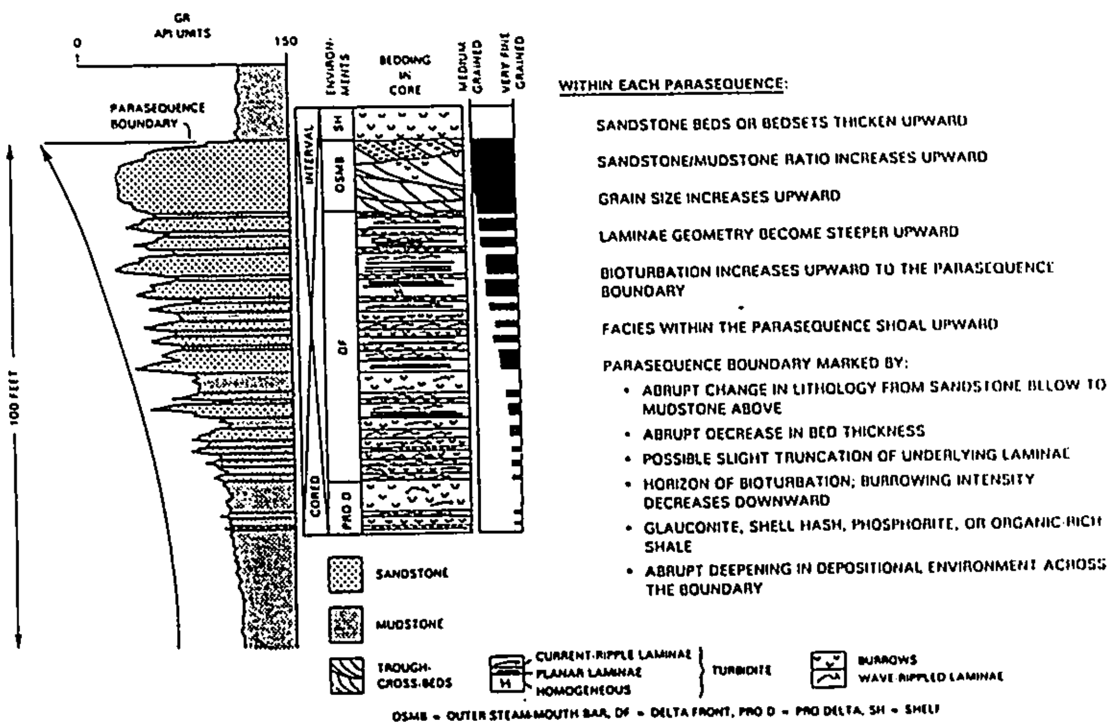
> *" a relatively conformable succession of genetically related beds or bedsets bounded by marine-flooding surfaces and their correlative surfaces."*

Parasequences tend to be composed of a single coarsening-upward succession, which are interpreted to record a gradual shallowing of water depth (Van Wagoner et al., 1990). An example of a theoretical parasequence due to progradation of a deltaic environment is shown in figure 12.7.

The allostratigraphic cycles defined by this study in the Lea Park - Belly River transition contain many of the characteristics of parasequences. The are composed of a single sedimentary succession reflecting a shallowing of the water depth, and are bounded at their top by a marine flooding surface, which is similar to a parasequence. The overall progradational pattern of the eight cycles is very typical of the progradational pattern of parasequences in what Exxon terms highstand systems tracts (Fig. 12.8). The scale of the individual cycles, both in terms of thickness and lateral extent is most similar to that of parasequences.

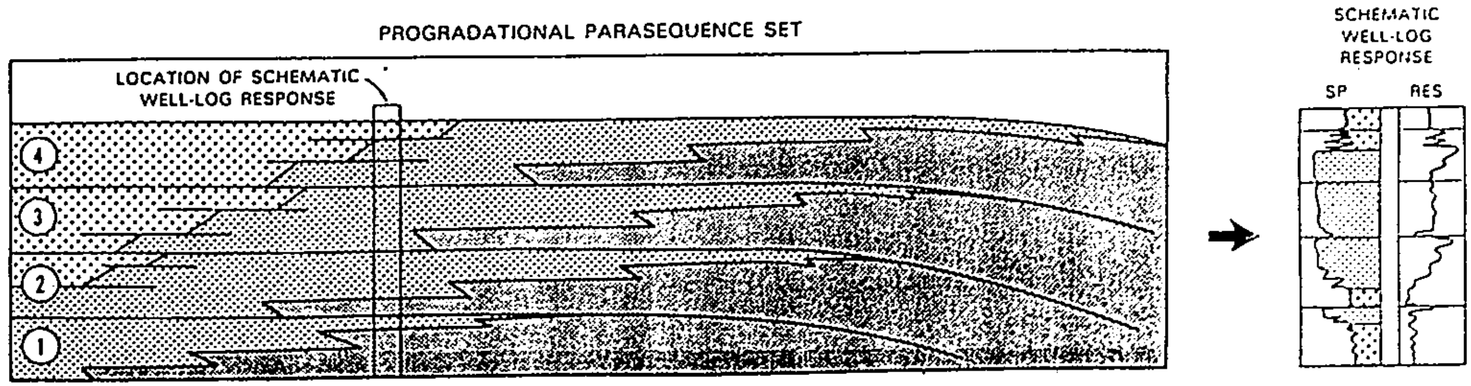
However, certain key aspects of the Lea Park - Belly River cycles do not fit the definition of a parasequence. Van Wagoner et al. (1990, p. 8) state that

Figure 12.7: Stratigraphic section of a theoretical parasequence formed due to progradation of a deltaic shoreline (from Mitchum and Van Wagoner, 1991).



GR
API UNITS
0
150
ENVIRON-MENTS
BEDDING IN CORE
MEDIUM GRAINED
VERY FINE GRAINED
PARASEQUENCE BOUNDARY
100 FEET
INTERVAL
CORED
SH
OSMB
DF
PRO D
WITHIN EACH PARASEQUENCE:
SANDSTONE BEDS OR BEDSETS THICKEN UPWARD
SANDSTONE/MUDSTONE RATIO INCREASES UPWARD
GRAIN SIZE INCREASES UPWARD
LAMINAE GEOMETRY BECOME STEEPER UPWARD
BIOTURBATION INCREASES UPWARD TO THE PARASEQUENCE BOUNDARY
FACIES WITHIN THE PARASEQUENCE SHOAL UPWARD
PARASEQUENCE BOUNDARY MARKED BY:
• ABRUPT CHANGE IN LITHOLOGY FROM SANDSTONE BELOW TO MUDSTONE ABOVE
• ABRUPT DECREASE IN BED THICKNESS
• POSSIBLE SLIGHT TRUNCATION OF UNDERLYING LAMINAE
• HORIZON OF BIOTURBATION; BURROWING INTENSITY DECREASES DOWNWARD
• GLAUCONITE, SHELL HASH, PHOSPHORITE, OR ORGANIC-RICH SHALE
• ABRUPT DEEPENING IN DEPOSITIONAL ENVIRONMENT ACROSS THE BOUNDARY
SANDSTONE
MUDSTONE
TROUGH-CROSS-BEDS
CURRENT-RIPPLE LAMINAE
PLANAR LAMINAE
H
HOMOGENEOUS
TURBIDITE
BURROWS
WAVE-RIPPLED LAMINAE
OSMB = OUTER STEAM-MOUTH BAR, DF = DELTA FRONT, PRO D = PRO DELTA, SH = SHELF

Figure 12.8: Theoretical geometry of a prograding parasequence set (from Mitchum and Van Wagoner, 1991).



PROGRADATIONAL PARASEQUENCE SET
LOCATION OF SCHEMATIC WELL-LOG RESPONSE
4
3
2
1
SCHEMATIC WELL-LOG RESPONSE
SP
RES

"Evidence of an abrupt decrease in water depth, ... has not been observed within parasequences." This is clearly not the case in the Lea Park - Belly River cycles. The basal surface of each cycle is interpreted to be an unconformity representing a drop in sea level, and as such each Lea Park - Belly River cycle, along with the overlying transgressive sediments, could be classified as a sequence. This is probably the more correct classification.

In a recent publication, Mitchum and Van Wagoner (1991) state that recent studies on fourth-order sequences show that they tend to be arranged in a distinct geometric relationships with each other with respect to the third-order sequence in which they are contained. A number of fourth-order sequences grouped into a distinctive progradational, aggradational, or retrogradation pattern is referred to as a *sequence set*. The progradational, eastward shingling pattern of the eight cycles in the Lea Park - Belly River transition observed within the study area can then perhaps be termed a sequence set.

However, there are some discrepancies between the nature of cycles observed in the Lea Park - Belly River transition and the heirarchical system of sequence stratigraphy presently adhered to by Exxon. Mitchum and Van Wagoner (1991, p. 138) state that "Whenever high-frequency (fourth-order) sequences have been observed, they themselves are composed of stacked parasequences that presumably have a fifth-order cyclicity." This relationship is not observed in the Lea Park - Belly River transition. In almost all cases, the internal composition of each cycle, or fourth-order sequence, is a single upwards-shallowing succession, which in Exxon terms is usually called a parasequence. The individual delta lobes observed in some of the cycles might be considered to represent this fifth-order cyclicity, but this is unlikely. The delta lobes are geographically separate from each other, and their is no evidence of fifth-order flooding surfaces within any cycle. There is no evidence in any of the transgressive deposits except perhaps for those between Cycles G and H for transgressive systems tracts containing retrogradational parasequences, or for that matter highstand systems tracts containing progradational parasequences which should be present underlying each sequence boundary according to the Exxon stratal hierarchy. Exxon sequence stratigraphy has mostly been developed for sequences reflecting third-order cyclicity. The basic components of their sequences are parasequences and systems tracts, which are arranged in a predictable geometric pattern according to where relative sea level is in its rise-fall cycle. The absence of these parasequences and systems tracts in higher-frequency sequences calls into question whether or not sequence stratigraphy is similar on all scales. It may also be that many of the parasequences which have previously been interpreted as components of third-order sequences are in fact sequences in their own right, and reflect both rise and fall of relative sea level, rather than just periodic rises, as is required by the definition of a parasequence.

It should also be noted that the most visible surfaces of erosion within the Lea Park - Belly River transition are regressive surfaces of marine erosion.

In Exxon terminology, these surfaces are presently not formally recognized, unless it as the correlative conformity of subaerial erosion surfaces. The importance of such surfaces needs to be addressed if Exxon sequence stratigraphy is to be applicable in more situations.

The points discussed above should not be interpreted as a condemnation of Exxon sequence stratigraphy. The sediments of the Lea Park - Belly River transition in central Alberta can be classified as a fourth-order sequence set, which is superimposed on the falling limb of a third-order sequence. The basic principles involved in this method are very sound, and the success of interpreting basin history that this method allows is reflected in its widespread use in industry and academia. Most of the problems with sequence stratigraphy focus on the rigid terminology which Exxon has defined. On a detailed scale, it appears that these definitions are not always adequate, and that individual systems will produce variations from the model hierarchy.

12.7.2: Genetic Sequence Stratigraphy

This method was recently proposed by Galloway (1989) as an alternative way of classifying deposits in a sequence stratigraphic manner. The principles of how depositional systems will react to relative changes in sea level are basically the same as in Exxon sequence stratigraphy. According to Galloway (1989, p. 125) a genetic stratigraphic sequence can be defined as:

> *"...the sedimentary product of a depositional episode. ... Each sequence consists of the progradational, aggradational, and retrogradational or transgressive facies deposited during a period of regional paleogeographic stability."*

This basically amounts to the same as the definition for an Exxon sequence. The only difference is that instead of using the erosional surface generated by sea level drop as the defining bounding surface of a sequence, genetic sequence stratigraphy uses the surface of maximum flooding created during the transgression. What this amounts to is that each sequence is a half-cycle of sea level fluctuation offset from an Exxon sequence. Galloway (1989) believes that sediments below the maximum flooding surface are derived from the underlying regressive sediments, and should therefore be included in the same genetic sequence as the regressive deposits. He also believes that surfaces of maximum flooding are easier to identify than Exxon sequence boundaries, which rely on subaerial erosion surfaces for identification, and therefore genetic stratigraphic sequence stratigraphy is easier to apply in more situations.

The Lea Park - Belly River transition, as observed in this study, does not contain identifiable maximum flooding surfaces, and therefore, cannot be classified in a genetic sequence stratigraphic sense. For these sediments, it appears that maximum flooding surfaces are not more easily identified than subaerial erosion surfaces.

12.7.3: Summary

It would appear that of the two proposed sequence stratigraphic methods, Exxon sequence stratigraphy is more applicable to the sediments of the Lea Park - Belly River transition. In some ways, though, Exxon sequence stratigraphy is also unsatisfactory when applied to the sediments of this study. The major problem with both of these methods is that both an Exxon sequence and a genetic sequence are a record of deposition for an entire cycle of sea level rise and fall. Thus, there is always an important sedimentological and allostratigraphic discontinuity within each sequence related to either the rise of sea level (Exxon sequence) or the fall of sea level (genetic sequence). Both of these boundaries can be very important as erosional surfaces or flooding surfaces in an allostratigraphic sense. In either case, the sediments above and below the surface are likely to be genetically unrelated to each other. It makes more sense sedimentologically to classify units according to the definition of allostratigraphy, which does not rank bounding discontinuities related to sea level drop or rise as being more or less important. This was the approach used in defining the Lea Park - Belly River cycles, which are bounded by both erosional surfaces due to relative sea level drop and flooding surface due to relative rise in sea level. One can then use this allostratigraphy to place the sediments into a sequence stratigraphic or chronstratigraphic framework.

CHAPTER 13: CONCLUSIONS AND IMPLICATIONS OF THIS STUDY

13.1: Primary Questions of the Study

In chapter 1, six questions were posed as the main problems that this study would attempt to answer. The preceding text has shown that, to varying degrees of satisfaction, all of these questions have been answered.

Facies analysis has shown that the sediments of the Lea Park - Belly River transition in central Alberta consist of a wide variety of nearshore, shoreline, and non-marine depositional systems. The Lea Park - Belly River transition in this area can be generally interpreted as representing the establishment of prograding coastal plain environments as the Western Canada Sedimentary Basin was filled by deposition of the Belly River wedge of sediment. The shorelines of these coastal plains were dominated by sandy depositional systems, which in most cases tended to be deltaic in nature. Within the non-marine environments to the west, fluvial deposition in meandering streams was dominant in the alluvial plains to the west of the shoreline systems. Sediments deposited in overbank areas within this environment are dominated by fine-grained ponded or lacustrine sediment and contain abundant coals and organic matter, indicating that the coastal plain environment was a wet, marsh-like environment. Fluvial systems farther to the west (more proximal to the mountain source) sometimes show evidence of being higher-energy coarse-grained braided stream environments, but the relationship of these systems to the delta plain and shoreline systems is unknown.

The shoreline systems within the Lea Park - Belly River transition in central Alberta are interpreted to have been characterized by deltas in which fluvial processes and deposition of fine-grained sand were dominant as opposed to wave or tidal processes. The resulting sediments are, however, thought to have been dramatically altered by post-depositional mass movement of delta mouth sediments. The deltaic succession is unlike any previously described within modern deltas or from the ancient record. The deltas are interpreted to have been fine-grained multiple-channel systems, in which a maze of numerous small channels supplied sediment to the basin over a wide length of coastline. The resultant deltaic succession is interpreted to be dominated by turbididty current deposition. The most likely origin of these turbididty currents is that they were created by slumping of highly unstable delta mouth sands which were deposited rapidly and with high pore water contents by fluvial flooding. It is also possible that the turbidity currents are the marine continuation of dense, sediment-laden flood flows from the rivers feeding the delta. These floods are most likely to have been caused by severe storms in the deltaic watershed, rather than being due to annual flooding. Wave-modification of the sediments is evident, but the preserved succession is dominated by the turbidites. The resulting geometry of the deltaic systems consists of proximal elongate, shore-normal lobes of sandstone which are surrounded by broader, more lobate bodies of sandstone. This is a non-actualistic delta model, and these deltaic systems cannot really be classified adequately by the present tripartite classification of deltas as either being fluvial-, wave-, or tide-dominated. The deltas are being interpreted as being fluvial-dominated for the most part, but the geometry and sedimentary succession is very different from other modern or ancient fluvially-dominated deltas. The results of this study indicate that while delta front morphology is important in reflecting the general processes operating within the depositional system, more emphasis on the relationship between the sediments and the depositional processes within deltas is required for more complete understanding and classification.

The Lea Park - Belly River transition within the study area can be divided into an allostratigraphic framework of eight regressive shoreline cycles, which are bounded at their base by marine regressive surfaces of erosion and at their tops by marine flooding surfaces. Each cycle is separated from the overlying cycle by deeper water "transgressive" sediments. Each of these cycles (and the intervening transgressive units) can be regarded as an informal allostratigraphic member, because each one is bounded by discontinuities. The cycles are arranged in a progradational, shingled geometry, so that each successive cycle is deposited further eastward into the basin than the previous cycle. When traced in a landward direction, cycles are laterally equivalent to fluvial/non-marine sediments which are incised into shelf sediments. The subaerial surface of erosion at the base of the fluvial sediments and the submarine regressive surface of erosion at the base of the laterally equivalent shoreline cycle tend to be at the same stratigraphic level, and indicate that the same relative drop in sea level was probably responsible for both the incision of fluvial sediments into shelf sediments and the creation of the erosional surface at the base of the shoreface succession.

The sharp-based shoreface successions typical of the Lea Park - Belly River cycles are indicative of forced regressions caused by small-scale relative drops in sea level. The lack of a transition interval of interbedded shelf mudstones and lower shoreface sandstones may be a typical feature of coastal successions deposited at least partially in response to a drop in relative sea level as opposed to gradual progradation due to sediment infilling with a stable relative sea level. The presence of sharp-based shoreface successions may be one way of determining whether cyclicity in deltaic successions is allocyclic or autocyclic in nature.

One of the ways commonly cited to distinguish between fluvial incision related to autocyclic switching of distributaries and incision due to allocyclic sea level drop is the amount of incision into the underlying shoreline deposits. It is generally thought that autocyclic switching should produce less incision than even small-scale incision due to an allocyclic drop in relative sea level. Within the coastal sediments of the Lea Park - Belly River transition, this was difficult to determine. The situations where fluvial sediments were sitting erosively on shelf sediments were attributed to allocyclic incision. However, there were many locations where fluvial channels erosively sat on shoreface sediments, and were laterally equivalent to fluvial channels which sat directly on shelf sediments a few kilometres further into the basin. It is not known whether these channels represent autocyclic incision into the shoreline sediments of the coeval cycle as the shoreline system progrades into the basin or whether they represent incision of fluvial channels related to the subsequent allocyclic drop in sea level. Interpretation of channels encased in shoreline/shoreface sediments as being due to autocyclic progradation or switching is therefore rather questionable.

It can be hypothesized that the effects of sea level fluctuation in the fluvial and coastal plain should be recorded by periods of soil development associated with sea level fall and flooding during sea level rise. The data from this study are very inconclusive in this matter. Paleosols are rare within the coastal plain sediments of the lowermost Belly River Formation, and could not be correlated between wells. Therefore no link could be established between soil development and sea level fall. Bioturbated intervals within the coastal plain sediments reflect the periodic incursion of saline waters into the environment. In several cases, these intervals were approximately stratigraphically equivalent to one of the transgressive units which separate the regressive cycles. It is possible that the effects of the relative rise in sea level at its most landward reach would be the development of partially saline ponded areas in the coastal plain. Unfortunately, the number of occurrences of bioturbated intervals associated with transgressive units is small, and a definite cause and effect relationship could not be established. In most locations, the most landward transgressive sediments are interpreted to have been removed by fluvial erosion related to the subsequent drop in relative sea level.

The eight shoreline cycles within the study area are interpreted to be deposited with a fourth-order periodicity of between 100,000-150,000 years. This cyclicity is superimposed on the falling limb of a third-order eustatic drop at 80 Ma associated with the progradation of the Belly River Formation into the basin. The cause of the fourth-order fluctuations in sea level is thought to be due to either glacio-eustatic fluctuations in sea level caused by variations in orbital eccentricity (100,000 year Milankovitch cycle), or to a combination of variations in tectonic subsidence rate due to active thrusting combined with possible variations in the rate of the third-order drop in sea level. A good case can be made for either interpretation, but until definite evidence for large-scale Cretaceous glaciation can be found, the tectonic interpretation would seem to be more plausible.

The final question proposed in chapter 1 involved applying the Lea Park - Belly River allostratigraphy proposed in this study to Exxon sequence stratigraphy. The eight progradational cycles along with the intercyclic transgressive sediments can be regarded as a sequence set in Exxon terminology. Each cycle can be considered a sequence, although the scale of the cycles is more appropriate to the definition of a parasequence. Internally, the Lea Park - Belly River cycles do not appear to contain the systems tracts which are interpreted to be the building blocks of Exxon sequences. Much of the reason for this is interpreted to be due to the fact that the hierarchy of Exxon sequence stratigraphy was developed for third-order sequences. The systems tracts and parasequences present within these third-order sequences may not be present within higher-frequency sequences. They certainly are not present in the Lea Park - Belly River transition. At present, it is probably best to classify the Lea Park - Belly River cycles as informal allostratigraphic members. They are bounded at their base by regressive surfaces of erosion and at their tops by marine flooding surfaces. Both of these surfaces are sedimentologically very important, and regarding them as being equally important allows a stratigraphic framework to be developed that more naturally reflects the history of relative sea level fluctuation and its effect on sedimentation within the Lea Park - Belly River transition.

13.2: Broader Implications of this Study

In addition to the primary questions which this study has attempted to answer, there are a few broader implications of this work with regards to our present understanding of coastal depositional systems.

Perhaps the most important of these focuses on how coastal environments are established in foreland basins during times of falling sea level. The present day situation is one of rising global sea level due to deglaciation following the Pleistocene. Therefore almost all of our models based on modern coastal depositional systems are inherently biased towards coastlines which develop under conditions of rising sea level. Individual shorelines may be progradational, but this progradation has had to overcome the effects of sea level. This is undoubtedly reflected in the sedimentary successions which have been deposited during the Holocene. The typical coastal succession is a gradual coarsening upwards succession from shelf to beach, indicative of a gradual decrease in water depth as the sediment supplied to the shoreline slowly fills the accommodation space made available due to rising relative sea level and subsidence. Sediment supply must be greater than the combined effects of sea level rise and subsidence for the coastline to be progradational. This situation of progradation of coastlines during times of rising sea level or sea level highstands has been incorporated heavily into the system of Exxon sequence stratigraphy. The basic building block of systems tracts, a parasequence, is interpreted to be deposited during times of rising sea level or sea level highstand. This is especially true for highstand systems tracts, which is where most progradational coastal successions occur in Exxon sequence stratigraphy.

The Lea Park - Belly River transition in central Alberta reflects the establishment of coastal depositional environments in the Alberta basin under a different set of conditions. Eustatic sea level is thought to be dropping at this time. This is being counteracted by subsidence due to active loading in the Cordillera. The net effect is that the creation of accommodation space within the basin, and the filling of this space by coastal progradation has occurred in a different manner than on modern shorelines. Periods of incision related to relative drop in sea level have allowed the shoreline to "jump" basinward in steps, and deposition of shoreface sediments under conditions of relative sea level fall has allowed the development of distinct sharp-based shoreface successions, rather than the gradual coarsening-upwards succession of modern shoreline systems. Transgressive or retrogradational shoreline systems do not appear to be preserved within the Lea Park - Belly River transition. Nor does there appear to be any evidence of progradational successions deposited in highstand systems tracts. This is interpreted to be a direct effect of the fact that the Lea Park - Belly River shoreline systems were deposited under conditions of overall relative sea level drop, and the accommodation space necessary for the development and preservation of retrogradational or highstand systems tracts was not available.

There are numerous other systems in the ancient record which appear to record similar cyclicity within shoreline systems at the base of a prograding wedge of sediment. Many of these systems have been studied with regards to the sedimentology of the shorelines, but allostratigraphy, especially with regards to fourth-order cyclicities has not yet been applied to these sediments. An example of a system that may be very similar to the Lea Park - Belly River transition is the Campanian-Maastrichtian Bearpaw - Horseshoe Canyon transition in Alberta (Ainsworth, 1991). These deposits also reflect the establishment of coastal depositional environments at the base of the next major wedge of sediment to be deposited in the Western Canada Basin following the Belly River Formation, known as the Edmonton Group. Ainsworth (1991) has also observed the presence of sharp-based shoreface successions within fourth-order sequences. This study is an outcrop study, but the overall geometry of the system may be very similar to the progradational sequence set of the Lea Park - Belly River transition. One major difference between the Horseshoe Canyon sediments and the Lea Park - Belly River transition is that the Horseshoe Canyon contains transgressive deposits, including estuarine sediments. This may reflect a difference in the time and amount of accommodation space created during sea level rise between the two systems. Another system which geometrically resembles the Lea Park - Belly River transition is the transition between the Cretaceous Cody Shale and the Parkman Sandstone in the Powder River Basin of Wyoming (Asquith, 1970). The shingled, progradational nature of the Parkman Sandstone shorelines is very similar to the Lea Park - Belly River transition. Allostratigraphic analysis of these sediments may reveal a similar history of relative sea level rise and fall as interpreted in this study.

The similarities between the overall geometry of the Lea Park - Belly River transition and that of many progradational parasequences within the literature also brings to mind the possibility that many of these parasequence sets and systems tracts may contain evidence of higher-frequency sea level fluctuations than previously thought. Another look at some examples of progradational parasequence sets may be warranted.

The second implication of this study regards the nature of depositional systems and processes within deltas. Studies of subsurface deltaic systems which include both detailed sedimentological descriptions and sediment body geometries are few. This study adds to this small list, and shows that modern delta classification methods are not completely adequate when describing the deltas of the Lea Park - Belly River transition. The existing tripartite classification scheme is supposedly process-based, but it is apparent that similar sand-body geometries can result from quite different depositional mechanisms. The specific way in which fluvial, tidal, or wave processes affect the depositional mechanisms must be better integrated into a model for delta classification. The possible effects of post-depositional mass movement of shallow water deltaic sediments must also be better integrated into future model development. This study, along with others such as Martinsen (1990) and Pulham (1989), shows that it is necessary to reanalyze how useful the tripartite delta classification system is as the focal point of deltaic facies models. This scheme really only emphasizes the relative importance of fluvial, wave and tidal processes with respect to each other. Many other factors, such as tectonic setting, sea level state, and climate can be extremely important for individual delta systems, and the relative importance of all of these factors undoubtedly varies for each individual system. I do not wish to propose a new rigid classification scheme for deltas, as any I might propose would probably be little improvement on the old system. The tripartite classification system is still very useful as a reference point upon which to base a description of an individual deltaic system. However, each system should be interpreted in a manner that weighs the relative importance of all controlling parameters which affect deposition for the delta system under study. This will lead to a better understanding of how all these factors affect deltaic sedimentation, and how delta systems may evolve over time, a subject that the present classification system does not address.

This study also raises questions about the nature of fluvial-derived depositional currents within deltas. This study, along with other recent studies of ancient deltaic systems may indicate that high-density currents carrying large amounts of sand in suspension are able to travel considerable distances (up to several kilometres) from the river mouth into the basin before depositing their sediment load. This is not the favoured interpretation for the Lea Park - Belly River systems, but it remains a definite possibility. Why such features are not observed in modern delta systems is problematic. There would seem to be no logical reason why such currents would operate during the Cretaceous, but not at the present day. However, the lack of modern analogues for these deposits does no alter the fact that they may be present in several ancient successions.

Further research, perhaps experimental, must be conducted to establish how such shallow water density currents operate and affect sedimentation before the problem of how these deposits were formed can be answered with any real confidence.

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APPENDIX A: MEASURED CORE INTERVALS

Well Location	Cored Interval
Township 39	
8-10-39-2W5	1233-1251 m
Township 40	
6-11-40-1W5	1062-1091 m
10-29-40-8W5	1710-1727.7 m
1-36-40-9W5	1714-1729.25 m
Township 41	
10-32-41-4W5	1275.9-1291.2 m
4-35-41-5W5	1384.1-1398.2 m
6-18-41-6W5	1607.6-1620.8 m
4-6-41-8W5	1687-1699.6 m
4-8-41-8W5	1681.4-1699.7 m
4-17-41-8W5	1708.8-1727.1 m
12-17-41-8W5	1674-1698.6 m
10-20-41-8W5	1665.6-1682.9 m
11-28-41-8W5	1610-1628 m
Township 42	
3-36-42-2W5	1109.8-1128 m
14-29-42-4W5	1285-1293.5 m
8-2-42-5W5	1337-1371 m
10-23-42-5W5	1312-1330 m
10-27-42-5W5	1317.1-1336.3 m
6-16-42-6W5	1573.5-1603.7 m
8-17-42-6W5	1597.3-1608.5 m
8-24-42-7W5	1561.7-1575 m
10-15-42-8W5	1500-1530.2 m
10-26-42-8W5	1523.5-1541.8 m
10-35-42-8W5	1521-1535.4 m
Township 43	
10-3-43-27W4	935.1-953.3 m
8-6-43-27W4	932.5-943.2 m
14-6-43-27W4	935-949.4 m
8-7-43-27W4	952.4-964.2 m
10-8-43-27W4	921-939 m
11-8-43-27W4	933.25-945 m

Well Location	Cored Interval
6-18-43-27W4	977-990.5 m
6-19-43-27W4	944-960.5 m
7-20-43-27W4	922.9-940.5 m
6-1-43-28W4	943.5-963.6 m
16-1-43-28W4	940-958 m
8-12-43-28W4	953-965.7 m
16-12-43-28W4	980.5-992.25 m
6-13-43-28W4	1017-1035 m
14-13-43-28W4	1009-1027 m
16-13-43-28W4	979-997 m
3-14-43-28W4	1009-1029 m
6-15-43-28W4	950-968 m
14-15-43-28W4	946.5-964.5 m
8-21-43-28W4	953-971.4 m
8-22-43-28W4	977-995 m
6-23-43-28W4	1023-1041 m
8-23-43-28W4	1002-1020.2 m
14-23-43-28W4	1018-1036 m
8-24-43-28W4	956-974.4 m
16-24-43-28W4	960-978 m
6-26-43-28W4	1012-1027 m
14-26-43-28W4	967-985 m
4-20-43-2W5	1094.2-1109.5 m
14-8-43-4W5	1264-1282.25 m
16-9-43-4W5	1261.3-1279.5 m
6-10-43-4W5	1265-1283 m
6-19-43-4W5	1287-1300.2 m
12-14-43-8W5	1481-1499 m
3-21-43-8W5	1529-1547 m
Township 44	
6-7-44-27W4	896.3-914.6 m
6-30-44-27W4	889-904 m
8-3-44-28W4	993-1011 m
16-3-44-28W4	941-959.25 m
Township 45	
10-3-45-28W4	910.4-923.5 m
8-10-45-28W4	914-923 m
8-19-45-1W5	1003-1016.8 m
14-32-45-1W5	925-950.5 m
16-18-45-6W5	1299-1325 m

Well Location	Cored Interval
6-26-45-7W5	1348-1378.7 m
6-11-45-8W5	1364.9-1383.2 m
Township 46	
10-9-46-1W5	970.7-987.8 m
4-14-46-1W5	927.4-945.7 m
6-16-46-1W5	971-1011.2 m
6-29-46-1W5	940-958.3 m
6-30-46-1W5	959-977 m
8-30-46-1W5	934-952 m
6-1-46-2W5	1036-1046 m
6-23-46-2W5	979.6-1003.5 m
14-24-46-2W5	1002-1023.5 m
8-26-46-2W5	1000-1018.25 m
10-26-46-2W5	1001.5-1015 m
11-26-46-2W5	1005-1023 m
11-35-46-2W5	995-1014 m
12-36-46-2W5	974-992 m
3-25-46-3W5	1024.7-1037.8 m
10-30-46-3W5	1088.4-1111 m
2-32-46-3W5	1061-1079.3 m
4-34-46-3W5	1051.8-1070.1 m
4-35-46-3W5	1027.7-1045.4 m
11-33-46-4W5	1073.8-1092 m
16-36-46-4W5	1067-1085.25 m
12-20-46-5W5	1161.6-1179.9 m
14-26-46-5W5	1153.7-1168.6 m
4-36-46-5W5	1179.3-1197.6 m
12-20-46-6W5	1227.5-1245.75 m
8-11-46-7W5	1262-1273 m
7-23-46-7W5	1224.1-1239.3 m
14-28-46-7W5	1240.9-1259.[illegible] m
Township 47	
12-19-47-1W5	1239-1256 m
2-2-47-2W5	962.6-980.6 m
10-2-47-2W5	968-982 m
10-9-47-2W5	997.6-1012.8 m
7-15-47-2W5	942.1-960.4 m
2-17-47-2W5	947-965.2 m
8-21-47-2W5	943.9-980.4 m
6-23-47-2W5	932.9-958.8 m

Well Location	Cored Interval
16-29-47-2W5	1004.6-1017.4 m
2-34-47-2W5	995-1009 m
10-34-47-2W5	985-995 m
6-35-47-2W5	990.9-1004.6 m
12-2-47-3W5	980.2-993.3 m
8-3-47-3W5	969.5-985.4 m
10-7-47-3W5	1013-1031.2 m
6-9-47-3W5	968-986.2 m
4-10-47-3W5	963.4-981.7 m
4-12-47-3W5	1030-1044 m
2-14-47-3W5	1030.5-1047.6 m
14-14-47-3W5	1043-1061 m
4-16-47-3W5	975-989.5 m
4-18-47-3W5	1029-1046 m
2-21-47-3W5	1022.9-1041.2
4-22-47-3W5	1034.1-1052.4 m
6-23-47-3W5	1034.8-1058.2 m
10-29-47-3W5	1022-1039.9 m
10-30-47-3W5	1000.6-1015.9 m
16-32-47-3W5	971-986.9 m
16-34-47-3W5	975.9-991.2 m
6-36-47-3W5	1009.1-1024.4 m
12-2-47-4W5	1043.3-1058.8 m
4-10-47-4W5	1017.1-1037.5 m
10-12-47-4W5	1042.7-1074.4 m
10-14-47-4W5	1030.5-1057.9 m
2-25-47-4W5	1013.7-1032 m
10-27-47-4W5	971-989 m
4-36-47-4W5	980.9-1008.2 m
15-1-47-5W5	1132.3-1148.2 m
10-4-47-5W5	1182.9-1201.2 m
6-11-47-5W5	1116.2-1131.4 m
14-15-47-7W5	1242-1260 m
12-24-47-7W5	1216.5-1234.8 m
14-34-47-7W5	1141-1173 m
16-30-47-8W5	1155-1173 m
16-12-47-9W5	1185.4-1207.6 m
12-14-47-9W5	1263.1-1293.6 m
14-25-47-9W5	1269.5-1284.4 m
Township 48	
10-7-48-1W5	867.4-882.6 m

Well Location	Cored Interval
10-28-48-1W5	781.4-799.7 m
14-33-48-1W5	769-787 m
8-2-48-2W5	942.5-953.5 m
14-2-48-2W5	904-920 m
6-3-48-2W5	923.8-939.9 m
8-3-48-2W5	931.4-948.2 m
8-4-48-2W5	931.4-947.6 m
6-6-48-2W5	972.6-987.8 m
4-11-48-2W5	902-916 m
11-19-48-2W5	916.75-926.5 m
10-28-48-2W5	867.2-882.4 m
3-29-48-2W5	917-935 m
6-4-48-3W5	949.7-964.9 m
6-6-48-3W5	984.5-1000.6 m
8-8-48-3W5	925.2-942.1 m
8-10-48-3W5	954-970.7 m
14-11-48-3W5	943.6-961.6 m
14-14-48-3W5	924.4-938.1 m
10-16-48-3W5	948.2-966.5 m
2-26-48-3W5	906.1-921.3 m
14-1-48-4W5	993.9-1009.5 m
8-11-48-4W5	993.9-1009.1 m
8-12-48-4W5	998.5-1013.7 m
16-27-48-4W5	956.7-975 m
16-28-48-4W5	978.7-993.9 m
4-30-48-4W5	970.7-989 m
12-30-48-4W5	961.9-980.2 m
8-4-48-5W5	1051.2-1069.5 m
6-22-48-5W5	1006.1-1024.4 m
10-25-48-5W5	966.8-983.2 m
2-14-48-6W5	1075.6-1091.2 m
2-15-48-6W5	1075.6-1091.2 m
16-19-48-6W5	1096-1116.8 m
10-23-48-6W5	1057.3-1072.6 m
16-27-48-6W5	1036.3-1051.5 m
8-28-48-6W5	1060.7-1080 m
14-28-48-6W5	1066.5-1100 m
2-29-48-6W5	1092.7-1109.1 m
6-31-48-6W5	1070.1-1090.2 m
16-31-48-6W5	1071-1094.8 m
6-33-48-6W5	1060.7-1079.3 m
8-33-48-6W5	1049.7-1077.4 m

Well Location	Cored Interval
8-34-48-6W5	998.5-1016.2 m
8-9-48-7W5	1149.4-1164.9 m
6-25-48-7W5	1080.8-1099.1 m
14-25-48-7W5	1068-1083.2 m
4-27-48-7W5	1020.7-1043.6 m
14-30-48-7W5	1138-1162 m
6-35-48-7W5	1064-1081 m
16-10-48-8W5	1234.1-1249.4 m
4-36-48-8W5	1181.1-1204.9 m
10-15-48-9W5	1228.4-1243.6 m
10-16-48-9W5	1274.7-1289.9 m
Township 49	
14-5-49-6W5	1057-1075.2 m
16-6-49-6W5	1054-1072.3 m
16-7-49-6W5	1022-1047 m
10-1-49-7W5	1047.3-1070.1 m
14-13-49-7W5	960-978 m
8-22-49-7W5	1038-1056 m
6-29-49-7W5	1075-1093 m
14-8-49-9W5	1294.8-1312.8 m
15-17-49-9W5	1285.1-1316.5 m
2-21-49-9W5	1221-1239.3 m