SEDIMENTOLOGY AND ALLOSTRATIGRAPHY

OF THE

BASAL BELLY RIVER FORMATION

OF

CENTRAL ALBERTA

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CENTRAL ALBERTA

BY

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Abstract

The nature of the transition from marine Lea Park Formation to continental Belly River Formation has been studied in a 50 m thick section in the Campanian of central Alberta. The sediments are subdivided into three *allomembers (A1 to A3), bounded by erosional discontinuities which formed* as a result of relative sea level changes.

The topmost 18 m of the Lea Park Formation consists of interbedded mudstones, siltstones and sandstones, deposited in a marine offshore environment below fairweather wave base (FWWB).

Allomember 1 consists of 11m thick sandstones with *Macaronichnus* traces which were interpreted as the shoreface of a prograding fluvial- and wave- dominated delta. These shoreface sandstones lie sharply on offshore mudstones of the Lea Park Formation. The lower bounding discontinuity is interpreted as a regressive surface of erosion (RSE) and consequently, the progradation of the shoreface succession of allomember 1 is due to a drop in sea level, termed a *forced regression*.

The sandstones of A1 are truncated by non-marine channel sandstones, mudstones with root traces and coal beds of A2. Channel fills are up to 20 m thick and occur in a channel belt up to 20 km wide. Fine-grained non-marine sediments and coal beds are restricted in the northwest of the study area. These coals are responsible for limiting the depth of erosion during the *ensuing deposition of transgressive and regressive* sediments of A 3.

Transgressive sediments are preserved only in A3 and comprise a 30 cm trangressive lag overlain by offshore mudstones. The extent of the lag to the northwest was mapped and found to coincide with the limit of coal deposition. Regressive sediments of A3 include fluvial and wave dominated lobate delta front sandstones, sharply overlying coal in the NW, but gradationally-based in the SE, overlying offshore mudstones. Gradationally-based shorefaces formed as a result of autocyclic delta lobe switching.

The allomembers define a fourth order cyclicity estimated at about 190,000 ka. The controls on high frequency cycles are not resolved, but two possibilities include regional tectonics and glacio-eustasy.

Acknowledgments

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The "Newf" from St John's who's endless serendipity meant we had the best digs in town is thanked for being a lively house-mate and good friend. I also wish to thank all the new friends I have made in Canada who made my stay more comfortable and enjoyable.

Thanks to Lola Pereira for the daily cups of tea and for the support in the final phase of preparing this thesis.

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Finally 'thank-you' cannot express my gratitude to my mother and family for the love they have always shown.

And The Capital of Kansas is ....

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1 Introduction

1.1 The Geological problem

The Belly River Formation, in the study area, comprises a 260 m thick succession of predominantly continental clastic deposits of Mid-Campanian Age. The Formation is bounded below by thick marine shales of the Lea Park (Pakowki) Formation and above by the marine shales of the Bearpaw Formation. Within the formation, laterally continuous sandstones are limited to the lower 50 m.

McLean (1971), Ogunyomi and Hills (1977), and Doig (1986) described these basal sediments as shallow marine and deltaic, representing a transition from marine shales of the Lea Park Formation to continental deposition in the Belly River Formation. Ogunyomi and Hills (1977) further noted that in Southern Alberta, the basal Belly River showed distinct cyclicity, with five prograding barrier-island depositional cycles being established. The cycles start at the base with offshore marine mudstones which grade into shoreface-foreshore deposits of a barrier island; these deposits are overlain by lagoonal sediments. The lateral extent of each progradational package was, however, not defined and in the subsurface, several important sedimentological questions remain unanswered;

(i) what are the lateral and vertical facies of the basal Belly River Formation in the subsurface?

(ii) are the sediments deltaic; if so, can the delta morphology be defined?

(iii) are the sediments cyclic?

(iv) what is the frequency of the cyclicity, and how can it be defined?

(v) what controls the cyclicity?

This thesis attempts to answer these questions for a particular area of Central Alberta where the core and well log control is good. This will lead to the establishment of a detailed regional stratigraphic correlation for the basal Belly River in central Alberta.

The principal technique employed here for stratigraphic analysis is allostratigraphy. Allostratigraphic units are defined as "mappable stratiform bodies of sedimentary rock defined and identified on the basis of bounding discontinuities" (NACSN, 1983), where the bounding discontinuities are commonly erosional surfaces, generally interpreted as being related to changes in relative sea level. Recognition and mapping of these discontinuity surfaces within the basal Belly River will be critical in interpreting the response of these coastal sediments to changes in relative sea level. An attempt is made to trace the areal extent of the bounding discontinuities to determine the limits of transgressions and regressions in the study area.

In deltaic settings autocyclic controls are also known to produce bounding discontinuities and an attempt will be made to differentiate between autocyclic and allocyclic (relative sea-level) processes and qualify their relative importance.

Finally, the occurrence of thick apparently structureless sandstones within the basal Belly River Formation presents two difficult interpretive problems. Do they represent primary depositional features, and if so, what processes are responsible for their deposition? Alternatively, is the apparently massive nature of these sandstones a result of modification by diagenetic processes? These problems are unresolved in the existing literature.

1.2 Geographical location and data base

The study area covers some 1500 km^2 in the plains of central Alberta, in the vicinity of Red Deer (Township 32 to Township 37, Range 27W4 to Range 2W5, Figure 1.1). The area lies to the east of the Rocky Mountain Deformed Belt, and the sediments are essentially flat lying, with a gentle south-westerly dip of between one and two degrees. There is no faulting in the area and few wells are deviated.

The research involved detailed description of 78 cores and correlation of over 500 electrical well logs (Figure 1.2, Appendix I). Core description was undertaken largely during the summer of 1991 at the Energy Resources Conservation Board (E.R.C.B.) laboratory in Calgary.

The oldest wells were drilled in 1958, and have spontaneous potential (SP) and resistivity well logs. Some of the more modern logs also used include gamma ray and sonic information. Unfortunately very few density or neutron logs were available.

Figure 1.1: Location of the study area.

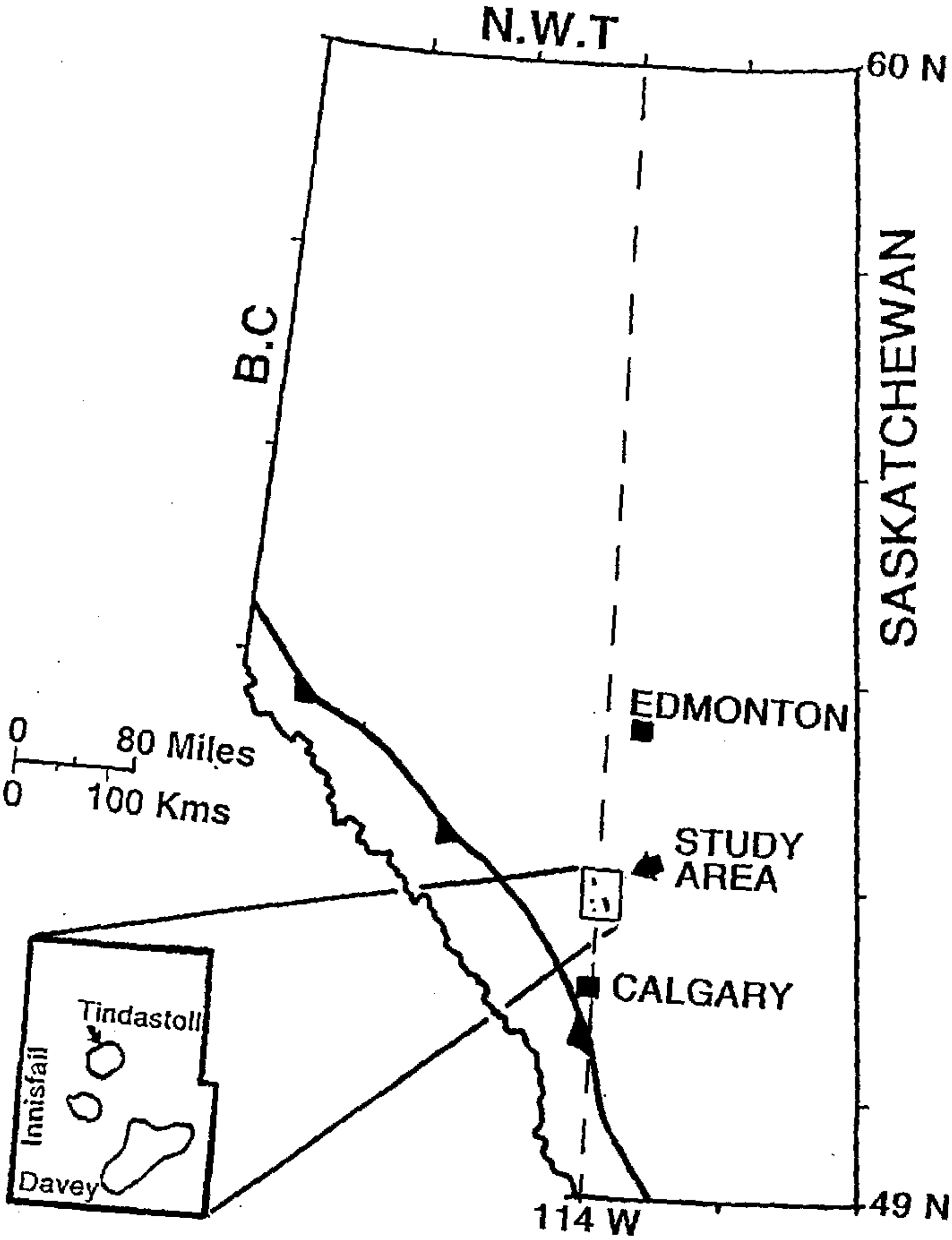
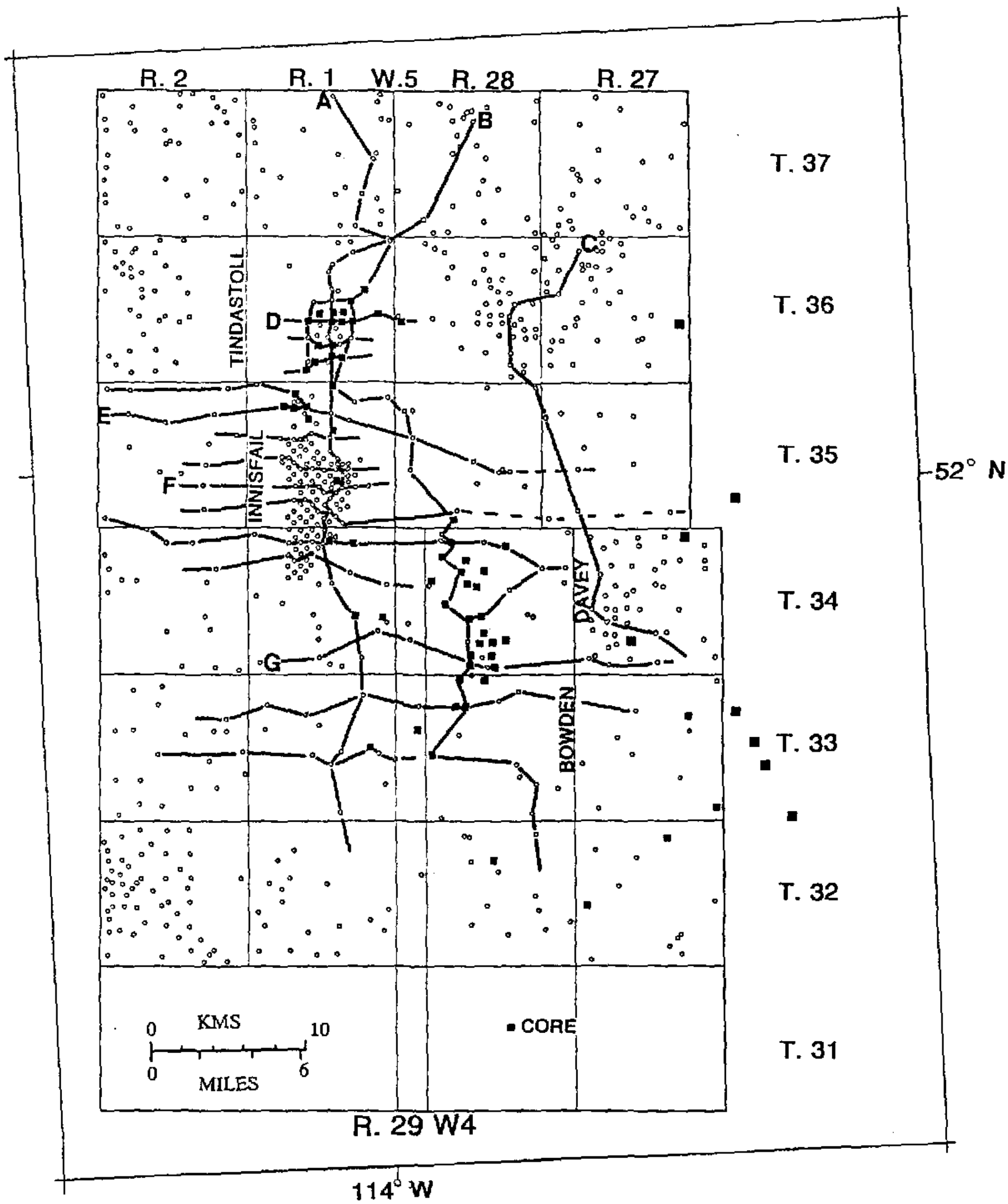

N.W.T
60 N
B.C
SASKATCHEWAN
EDMONTON
0
80 Miles
0
100 Kms
STUDY
AREA
CALGARY
Tindastoll
Innisfail
Davey
114 W
49 N

Figure 1.2: Detailed location map showing the hydrocarbon fields studied, the distribution of wells and the core control in the fields. Fifteen cross-sections were drawn to establish the correlation. Seven cross-sections (A to G) are used in this thesis.


R. 2
R. 1
W.5
R. 28
R. 27
T. 37
T. 36
T. 35
T. 34
T. 33
T. 32
T. 31
52° N
A
B
C
D
E
F
G
TINDASTOLL
INNISFAIL
DAVEY
BOWDEN
0
KMS
10
0
MILES
6
CORE
R. 29 W4
114° W

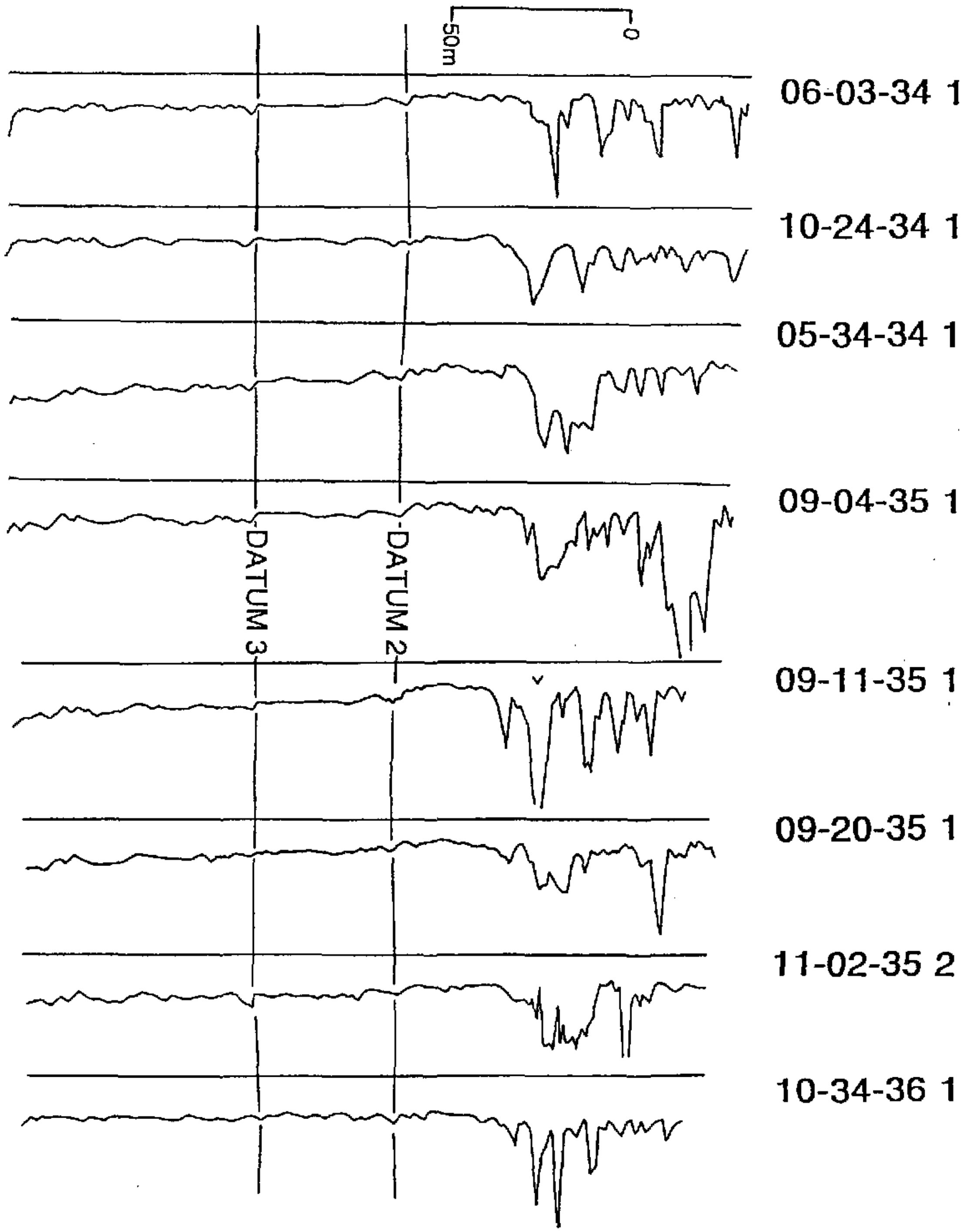
1.2.1 Log correlation

Log correlations were established on the basis of several 'markers'. Where possible, use was made of a prominent regionally extensive marker known to petroleum geologists as the Milk River Shoulder (Figure 1.3, Datum 2). This is a higher resistivity 'kick' present within the Lea Park Formation. Similarly, a second lower datum, of regional extent (Datum 3) was used in conjunction with Datum 2. Few wells penetrate these datums, and hence considerable use was made of local markers, such as thin coal beds within the basal Belly River and/or sand stringers in the topmost Lea Park. No one datum could be established over the whole area, and regional correlation depends on an array of markers.

1.3 Thesis layout

A brief review of the regional stratigraphy, tectonic setting and the nomenclature is given in Chapter 2. This chapter also introduces the allostratigraphy and facies scheme of the basal Belly River defined in the area. Chapter 3 describes the sedimentology of the underlying Lea Park Formation. The basal Belly River is subdivided into three informal allomembers. The allomembers are made up of several facies which are described in some detail in Chapter 4. Chapters 5 and 6 are interpretive sections discussing the vertical facies succession and geometry of individual allomembers, and the lateral relationships between allomembers. Aspects of the structureless sandstones are also considered in Chapter 6. Regional

Figure 1.3: Resistivity logs showing datums in the Lea Park Formation which were used in correlation.



50m
0
06-03-34 1
10-24-34 1
05-34-34 1
09-04-35 1
DATUM 3
DATUM 2
09-11-35 1
09-20-35 1
11-02-35 2
10-34-36 1

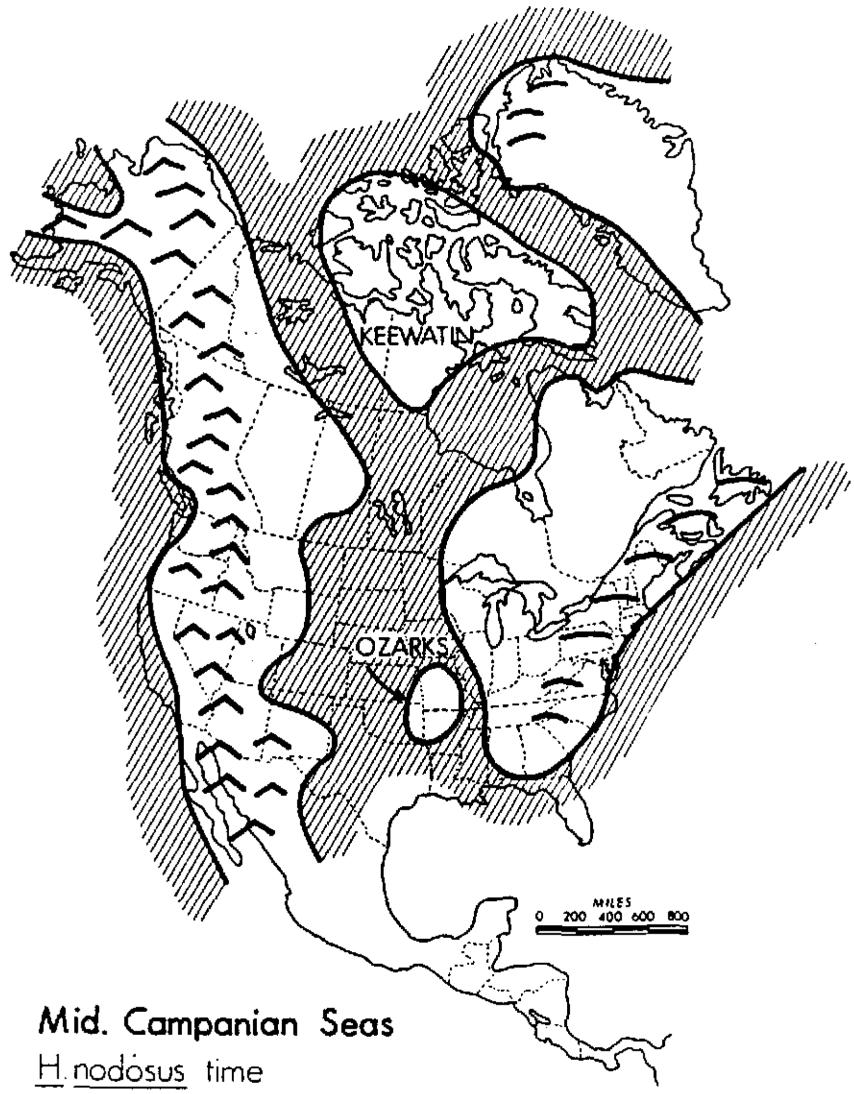
tectonic and eustatic controls on sedimentation are addressed in the following chapter, and finally the conclusions are outlined in Chapter 8.

2. Regional Setting

2.1 Tectonic setting and provenance

The Belly River Formation is an eastward thinning clastic wedge which varies in thickness from a kilometre along the Foothills to 75 m in west-central Saskatchewan (Eberth et al., 1990) and thins out at its depositional edge in south-west Saskatchewan. These sediments were deposited in the Western Canada Sedimentary basin, a foreland basin comprising a thick succession of Upper Jurassic to Paleocene sediments (Cant and Stockmal, 1989). Sedimentation took place along the western margin of a broad inland sea (Figure 2.1) during Mid-Campanian times, largely in response to the rising cordillera associated with the late stages of the Laramide Orogeny. Mountain building due to thrusting to the west resulted in the southward retreat of the Lea Park sea (Stott,1984) and the prevalence of continental deposition as indicated by the sediments of the Belly River Formation. This episode is also thought to be coincident with the third order eustatic sea level fall of Haq et al. (1988) at 80 Ma, and at least part of the R8 regressive cycle of Kauffman (1984). The regressive phase of the Lea Park Sea was slow, and consequently the transition to non-marine sedimentation is highly diachronous, younging south-eastwards (North and Caldwell,1964). The end of Belly River deposition is marked by a major transgression during which the overlying Bearpaw Shales were deposited.

Figure 2.1: Palaegeography of the Western Interior Seaway in the Mid-Campanian (Williams and Stelck, 1975).



Mid. Campanian Seas
H. nodosus time

Measured palaeocurrent directions indicate a strong south-east flow component with minor west to east trends close to the mountain belt (Jerzekiewicz and Labonte, 1991). Studies indicate provenance from the Omineca Crystalline Belt due to episodic thrusting along the western margin of the foreland basin (Mack and Jerzekiewicz, 1989). Composition is dominated by volcanic andesitic rock fragments, with minor low-grade metamorphics and sedimentary clasts. The sandstones of the basal Belly River are lithic arenites (Doig, 1986) except in southern Alberta, where they consist of arkosic arenites (Ogunyomi and Hills, 1977).

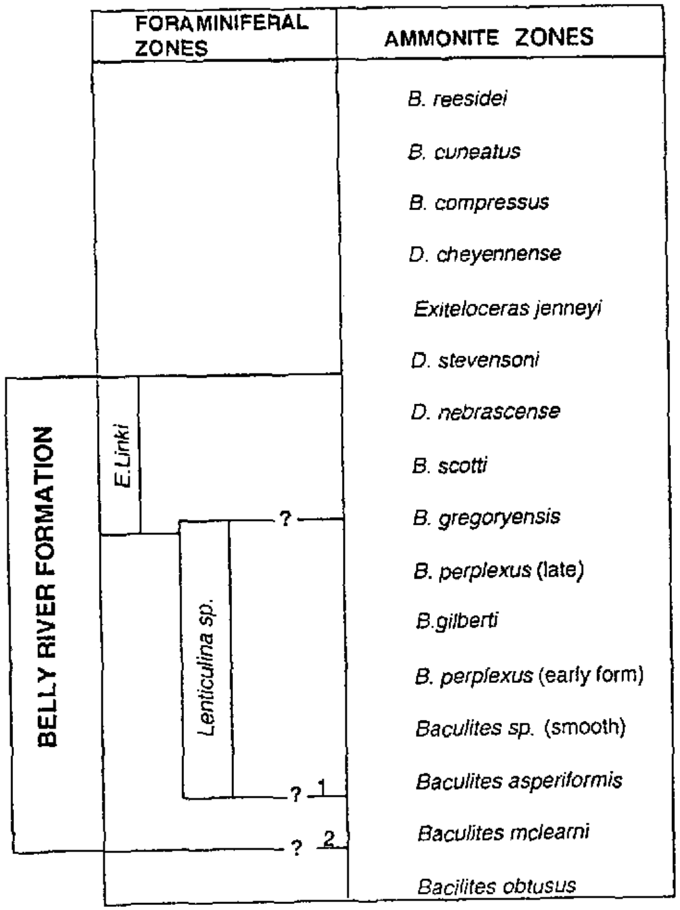
2.2 Stratigraphy

Stratigraphic correlation within the Belly River Formation is difficult due to a scarcity of fauna within the largely non-marine section, and the diachronous nature of the sediments. Furthermore, foraminiferal Zones change from the foothills to the plains and correlation of these Zones into the Foothills is incomplete (Table 2.1). In SW Saskatchewan, the Formation is assigned to the *Eoeponidella linki* foraminiferal Zone which encompasses the uppermost Lea Park, the entire Belly River and the lowermost Bearpaw Formations (Caldwell et al., 1978). This zone correlates with three ammonite zones of Obradovich and Cobban (1975); *Baculites gregoryensi, B. Scotti* and *Didymoceras nebrascense* Zones (Table 2.1). Along the foothills of Central Alberta, the *Lenticulina sp.* Zone is found to correlate with the lower part of the *Eoeponidella linki* Zone. The correlation of the *Lenticulina sp.* Zone to ammonite zones is incomplete; however, the base of this zone extends to at least the *Baculites asperiformis* zone. These zones give a Middle Campanian Age for the entire formation. The overlying Bearpaw Shales are assigned to the *Baculites compressus* ammonite Zone, with maximum flooding taking place at the time of *Baculites reesidei* (Williams and Stelck, 1975).

Foraminiferal samples extracted from the sediments of the study area yielded poor results. The foraminifera were poorly preserved, arenaceous and difficult to identify.

Absolute dates from the uppermost Lea Park Formation are scarce. The topmost Lea Park (Pakowki) Formation in southern Alberta occurs within the *Baculites asperiformis* ammonite zone and is dated at 77 Ma (Lerbekmo, 1989). Mack and Jerzykiewicz (1989) date the topmost marginal marine sediments along the foothills (equivalent to the Pakowki shales in Wyoming) at 79.5 Ma (Table 2.1). In doing so, Mack and Jerzykiewicz (1989) have chosen a radiometric date from Gill and Cobban (1973) which records the start of the Claggett transgression (T8 of Kauffman, 1984) in Wyoming, as the top of the equivalent marginal marine sediments in the foothills. Their topmost Pakowki shales equivalent is included in *Baculites mclearni*, which is one ammonite zone lower than *Baculites asperiformis*. The exact stratigraphic position of the *B.mclearni* Zone within the Pakowki Formation along the foothills is not indicated by Mack and Jerzykiewicz (1989). In northern Montana, Goodwin and Deino (1989) established an age of 78 Ma from two bentonite beds within the Taber coal zone, which marks the boundary between the Foremost and Oldman formations (McLean, 1971). Deposition of the basal Belly

Table 2.1: Mid-Campanian biostratigraphy in Alberta. Correlation of ammonite Zones with foraminiferal Zones is from Caldwell et al. (1978). Estimated absolute dates show the base of the Belly River Formation to be between 77-79.5 Ma. (ammonite Zones from Obradovich and Cobban, 1975).



REFERENCE	LOCATION	SOURCE	AGE (Ma)	AMMONITE ZONES
1. Lerbekmo, 1989	southern Alberta	Obradovich and Cobban (1975)	**77**	*Baculites asperiformis* (based on 0.5 Ma duration of ammonite zones; *B.obtusus* at 77.9 Ma
2. Mack and Jerzykiewicz, 1989	Alberta Foothills	Gill and Cobban, (1973)	**79.5**	*Baculites mclearni*

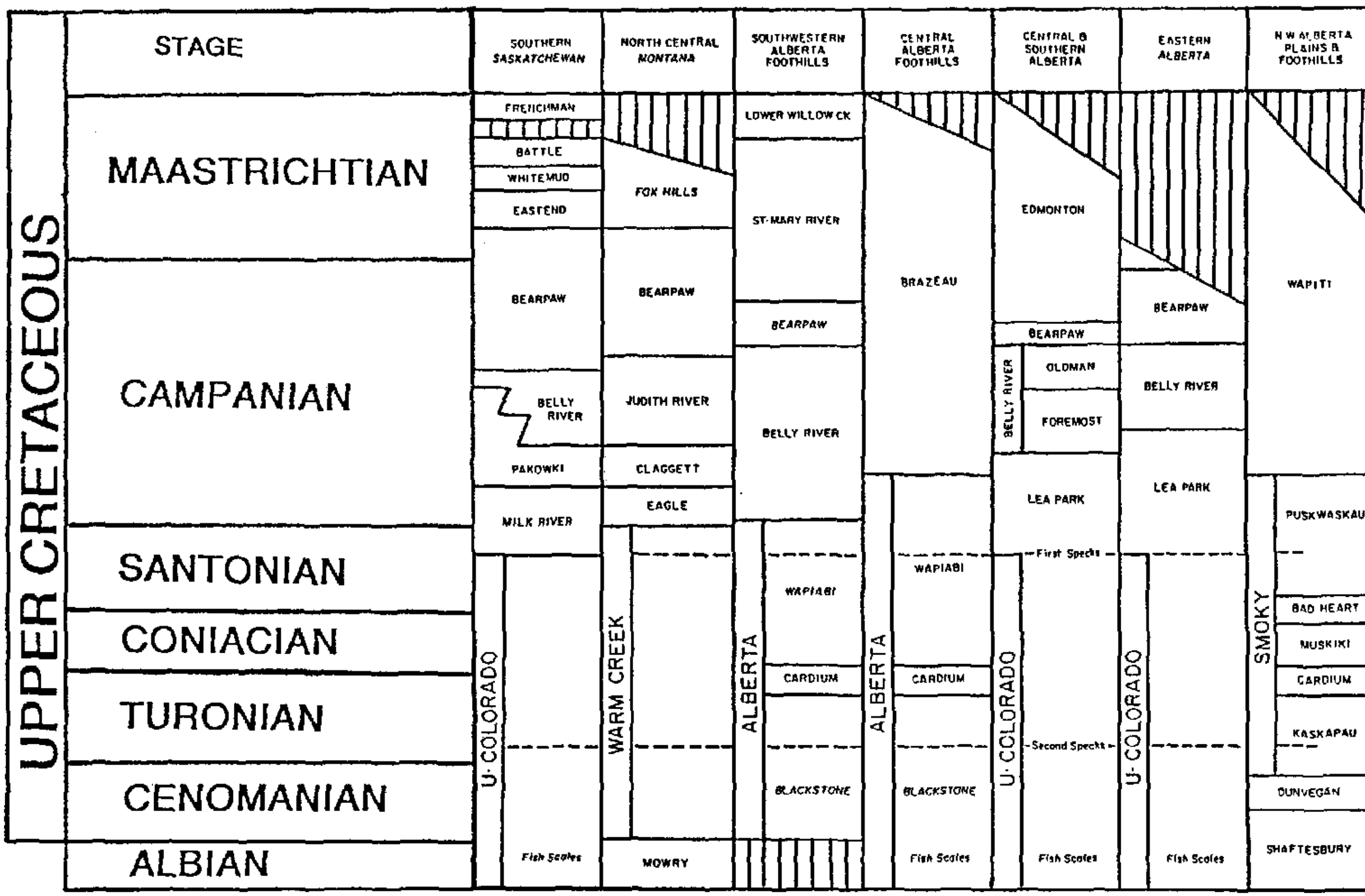
River therefore, spans a period of one million years or less (if the base is taken at 79.5 Ma). Moreover, the dates above suggest that the regression of the Lea Park and onset of basal Belly River deposition may not be coincident with the third order eustatic sea level fall of Haq et al. (1988) which occurs at 80 Ma.

The top of the Judith River (Belly River) Formation is dated at 74-75 Ma from bentonites within the Lethbridge Coal Zone (Thomas et al., 1990).

2.3 Nomenclature

Dawson (1883) used the term Belly River Formation for a succession of predominantly continental deposits found in outcrop in southern Alberta. Tyrell (in Shaw and Harding, 1948) later extended its usage to include east-central Alberta. Since then a confusion of names has arisen for what is essentially the same formation (Table 2.2). The current nomenclature is partly based on locality; in southern Alberta, and in the United States, the Belly River equivalent is known as the Judith River Formation, whereas along the foothills of central Alberta the name Brazeau Group (or Formation) is assigned to a thick clastic wedge that includes the Belly River Formation at its base. In north-west Alberta, the Brazeau Group equivalent is known as the Wapiti Formation. Away from the foothills, discussions are centered on the subdivision of the Belly River Formation. In places, the bBR Formation is upgraded to Group status comprising two formations; the Foremost and Oldman Formations (Table 2.2). McLean (1971) suggested, however that because these formations are not clearly defined, they be

Table 2.2: Lithostratigraphic nomenclature of the Belly River Formation in western North America (after Williams and Burk, 1964).


UPPER CRETACEOUS
STAGE
MAASTRICHTIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
CENOMANIAN
ALBIAN
SOUTHERN SASKATCHEWAN
NORTH CENTRAL MONTANA
SOUTHWESTERN ALBERTA FOOTHILLS
CENTRAL ALBERTA FOOTHILLS
CENTRAL & SOUTHERN ALBERTA
EASTERN ALBERTA
N.W. ALBERTA PLAINS & FOOTHILLS
FRENCHMAN
BATTLE
WHITEMUD
EASTEND
BEARPAW
BELLY RIVER
PAKOWKI
MILK RIVER
U·COLORADO
Fish Scales
FOX HILLS
BEARPAW
JUDITH RIVER
CLAGGETT
EAGLE
WARM CREEK
MOWRY
LOWER WILLOW CK
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
PUSKWASKAU
SMOKY
BAD HEART
MUSKIKI
CARDIUM
KASKAPAU
DUNVEGAN
SHAFTESBURY

regarded as informal members, and that by historical precedence (Hayden, 1871; in McLean, 1971) the name "Judith River Formation" be used regionally. This suggestion has not been generally adopted and to date (1993) a unique nomenclature has not yet been established.

For the purpose of this thesis, the informal term "basal Belly River" (bBR) Formation is applied to the marginal marine sediments of the formation which record the transition from marine to continental sedimentation.

2.4 Allostratigraphy of the basal Belly River Formation

The purpose of this section is to provide a brief definition of the allostratigraphic scheme established and used throughout this thesis. This is accomplished by introducing, at an early stage, one log and two core cross-sections illustrating the stratigraphic relationships between the allomembers.

The only recent stratigraphic work in Alberta is that of Wasser (1988) and B.A.Power (pers. comm., 1992) in the Pembina Field. Both these authors recognised eight prograding deltaic lobes. These lobes offlap and young to the south-east. In addition, the latter author defined individual lobes as allomembers. A correlation of these allomembers into the study area has not yet been established. A.P.Hamblin (pers. comm., 1992) outlined five "regionally-defined cycles" throughout Alberta. These cycles young eastwards and the sediments of this study area fall into Hamblin's cycles 1 and 2. Preliminary studies indicate that the cycles comprise stacked parasequences (A.P.Hamblin, pers. comm., 1992).

The sediments of the basal Belly River Formation in this study are subdivided into three informal allomembers (A) bounded by mappable bounding discontinuities (BD) surfaces. These surfaces represent a break in deposition and can be erosive or non-erosive (Bhattacharya and Walker, 1991b). The bounding discontinuities described in this thesis are regressive (RSE) and transgressive surfaces of erosion (TSE) and marine flooding surfaces, FS, (Van Wagoner et al., 1990). The allomembers are separated at the base by a regressive surface of erosion, and above by a marine flooding surface or a transgressive surface of erosion.

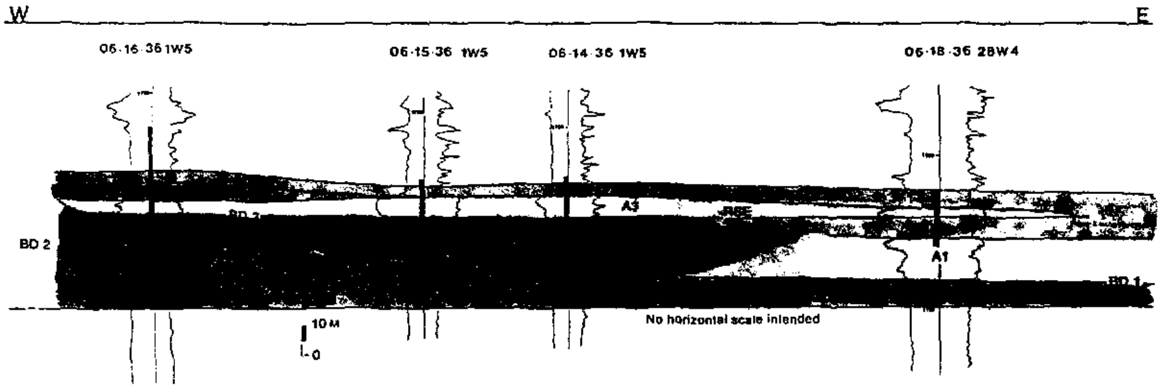
The allostratigraphic scheme was derived by regional correlation of core and well logs. The core and log data were principally "hung" on two regional datums within the Lea Park Formation (Figure 1.3). These two datums are parallel to each other, laterally persistent and easy to identify. This enabled correlation of the allomembers and recognition of the geometries of the various facies. Moreover, the relationship between the allomembers were established by a detailed correlation involving both well logs and cores.

The regional sediments comprise the Lea Park marine shales and mudstones. These are sharply overlain by marine sandstones of allomember 1 (A1). The sandstones will be interpreted later as sharp-based prograding shorefaces. The sharp contact between the Lea Park and Belly River Formations is evident across the study area. Therefore the contact between the Lea Park Formation and allomember 1 is a mappable regressive surface of erosion (RSE) designated in this thesis as BD 1 (Figure 2.2). This surface will be interpreted to be due to a forced regression (A.P.Plint pers. comm., 1988; Posamentier et al., 1990).

Sediments of allomember 2 (A2) consist of non-marine sandstones, siltstones and mudstones. These represent fluvial channel incisions which can be mapped across the study area. Three types of channel fills are present and are discussed in the following chapters. The base of allomember 2 constitutes a second regressive surface of erosion (BD 2, Figure 2.2) represented by channel incision. The extent of the incision into allomember 1 (Figure 2.2) can be shown in core 06-14-36-1W5, where a fluvial channel sits erosively on marine shales of the Lea Park Formation. To the east (core 06-18-36-1W5) however, the channels lie at the same stratigraphic position as the shoreface succession of allomember 1. Throughout the study area, this relationship is evident. Where part of the shoreface of allomember 1 is preserved, BD 2 is delineated by a break in grainsize. Channel sandstones of allomember 2 are coarser grained than the underlying shoreface sandstones. These commonly have a basal lag comprising chert pebbles, clay clasts and coalified plant debris. The extent of bounding discontinuity 2 in the marine section is not observed in the study area.

Allomember 3 (A3) is the highest allomember in the bBR Formation in the study area and comprises transgressive deposits and lobate deltaic sandbodies that overlie allomembers 1 and 2. Transgressive sediments include marine mudstones and thin conglomeratic lag deposits. The base of allomember 3 is a transgressive surface of erosion (BD 3) in the south and

Figure 2.2: Core and log cross-section D-D' showing the three allomembers (A1 to A3) defined in the study area, bounded by erosional discontinuities (BD) which include regressive surfaces of erosion (RSE), transgressive surfaces of erosion (TSE) or marine flooding surfaces (FS).


W
E
06-16-36 1W5
06-15-36 1W5
06-14-36 1W5
06-18-36 28W4
BD 2
A3
RSE
A1
BD 1
10 M
0
No horizontal scale intended

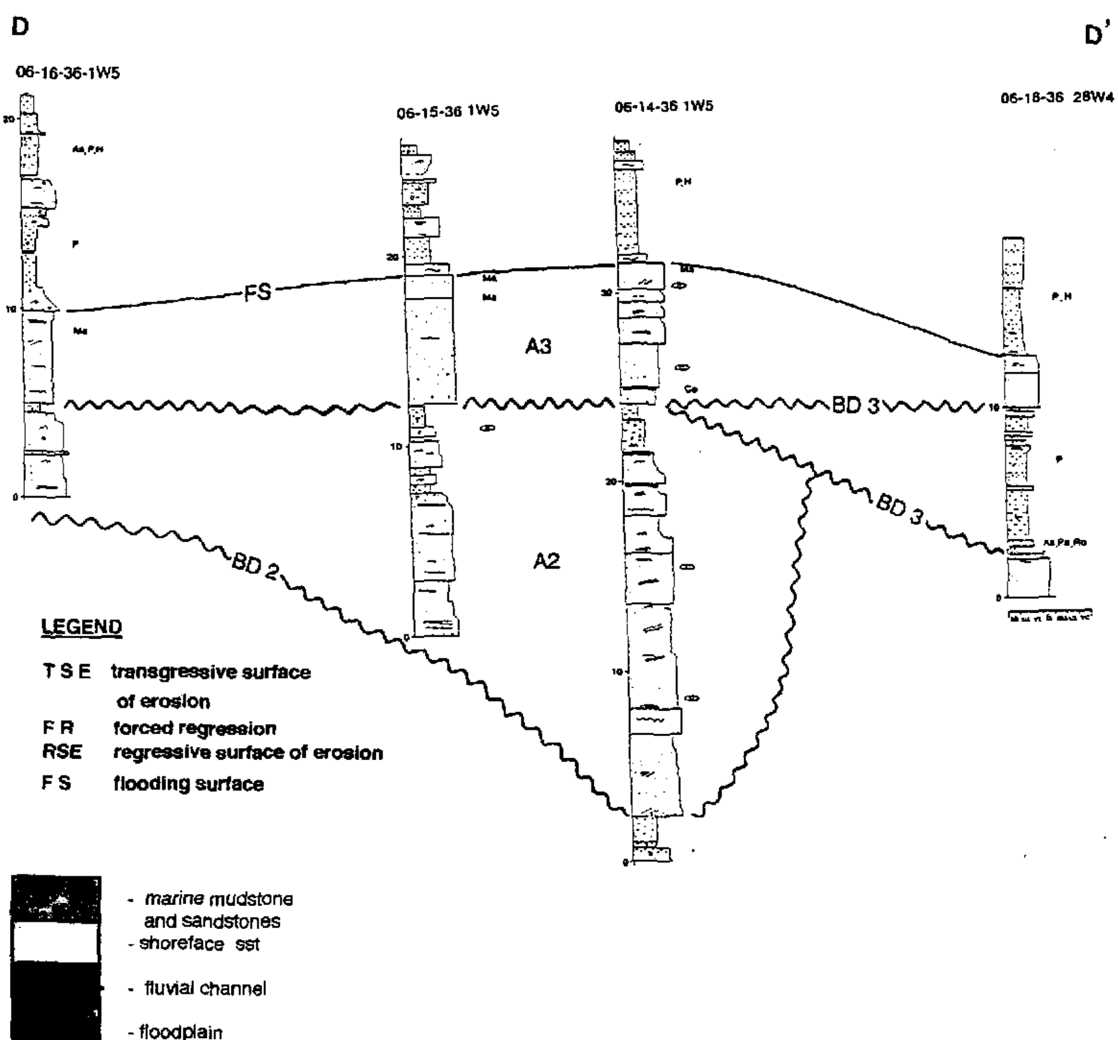
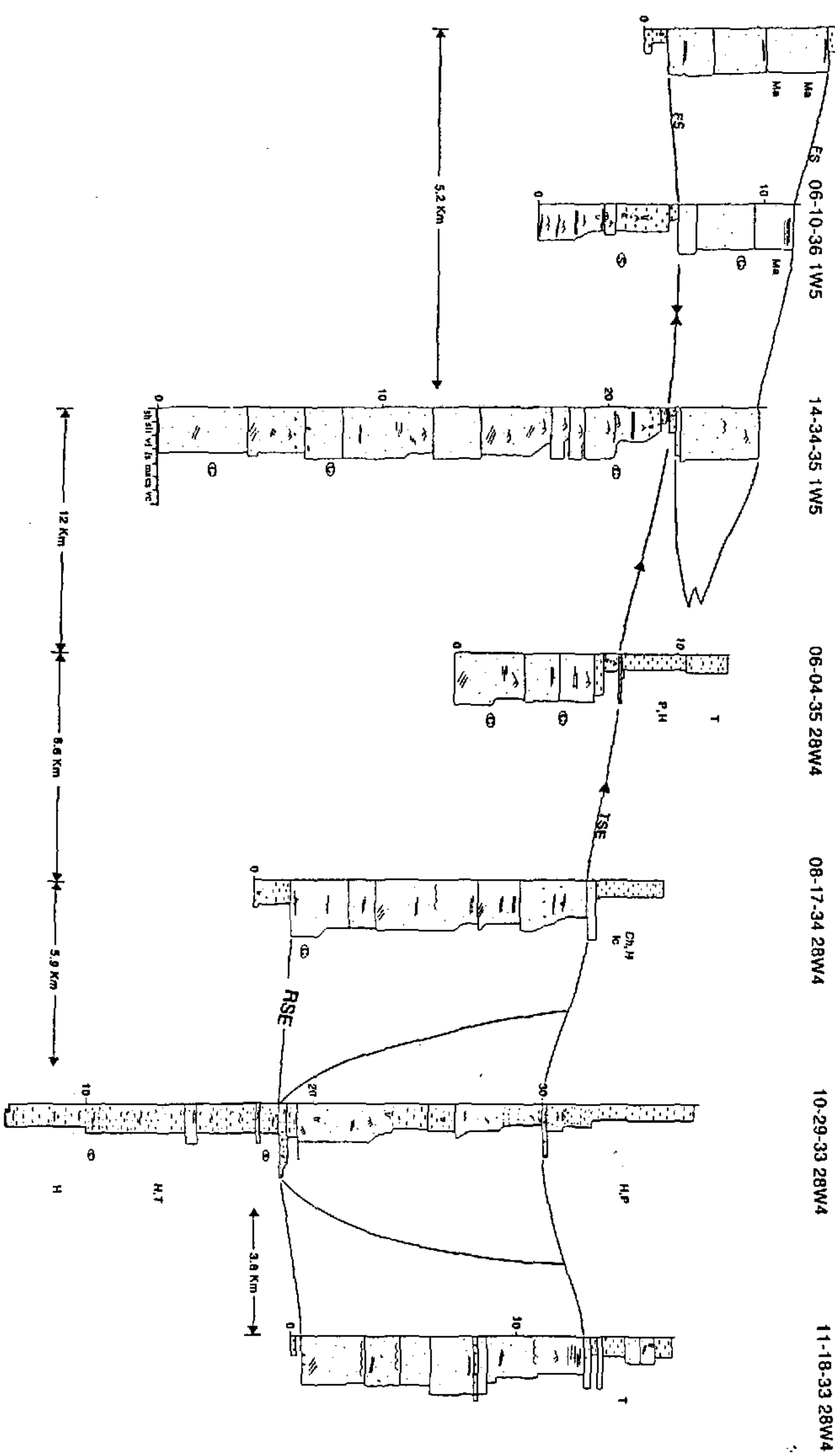

D
D'
06-16-36-1W5
06-15-36 1W5
06-14-36 1W5
06-18-36 28W4
FS
A3
BD 3
BD 3
A2
BD 2
LEGEND
T S E transgressive surface of erosion
F R forced regression
RSE regressive surface of erosion
F S flooding surface
- marine mudstone and sandstones
- shoreface sst
- fluvial channel
- floodplain facies

Figure 2.3: N-S core cross-section outlining the geometry of the allomembers and the lateral relationships of the bounding discontinuities .



14-15-36 1W5
06-10-36 1W5
14-34-35 1W5
06-04-35 28W4
08-17-34 28W4
10-29-33 28W4
11-18-33 28W4
FS
FS
TSE
RSE
5.2 Km
12 Km
8.6 Km
5.9 Km
3.8 Km

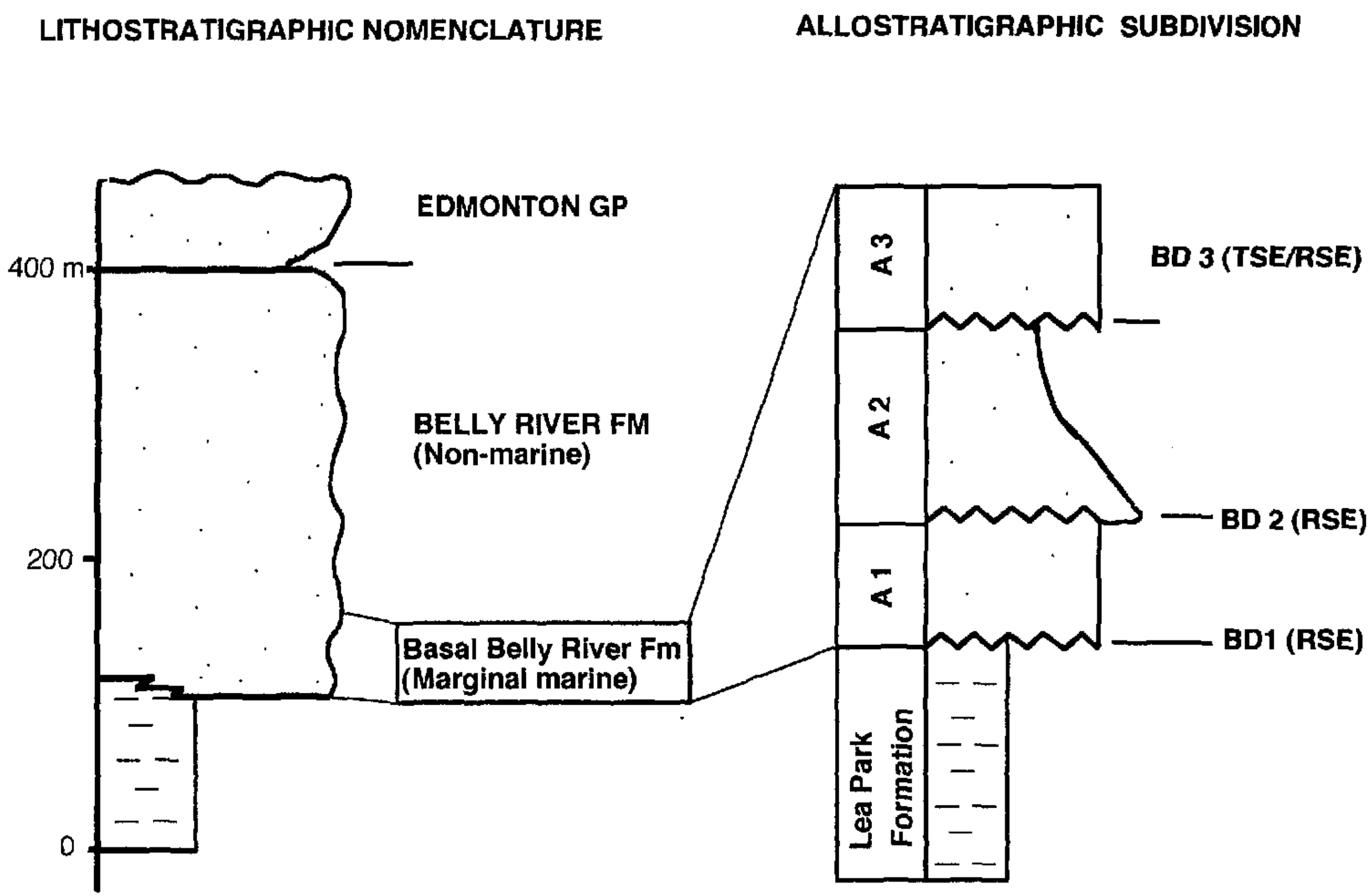
east of the area (Figure 2.3). In the north-west however, BD 3 is overlain by a second sharp-based marine sandstone (Figures 2.3). Consequently, BD 3 is a coplanar surface of erosion formed during transgression and subsequent regression.

In summary, three allomembers are defined, bounded below by discontinuity surfaces (Figure 2.4). This subdivision is compared with the lithostratigraphic subdivision. Allostratigraphic subdivision based on bounding discontinuities makes it easier to study the effects of relative sea level change. Two of these surfaces (BD 1 and BD 2) represent erosional discontuinities as a result of a regressive phase. BD 3 is primarily a transgressive surface of erosion (TSE). The full evidence defining these relationships is presented in the ensuing chapters.

2.5 Facies outline

There has been some disagreement as to the position of the base of the Belly River Formation due to its diachronous nature (McLean, 1971). A lithological boundary is commonly placed at the base of first major sandstone. This scheme is followed in this thesis because the allostratigraphic subdivision employed recognises the base of the sandstone as a regional extensive erosional surface (BD 1). Outside the present study area, the basal sandstone may be missing and carbonaceous shales commonly with coal, overlie the mudstones of the Lea Park Formation (McLean, 1971).

Figure 2.4: Allostratigraphic subdivision, compared with the lithostratigraphic subdivision.


LITHOSTRATIGRAPHIC NOMENCLATURE
ALLOSTRATIGRAPHIC SUBDIVISION
EDMONTON GP
400 m
BELLY RIVER FM
(Non-marine)
200
Basal Belly River Fm
(Marginal marine)
0
A 3
A 2
A 1
Lea Park
Formation
BD 3 (TSE/RSE)
BD 2 (RSE)
BD1 (RSE)

In the study area, six facies are described. These can be further subdivided into several types. The subdivision was based on lithology, bed thickness, and grain size, bioturbation and physical structures. The facies are given interpretive names to help the reader; however, a clear separation of the facts (description) and interpretation is presented in the facies chapters. The common log trends of the facies are shown here (*Figure 2.5*) to familiarise the reader with the different log responses used in establishing detailed correlation of the facies.

The log responses for the facies of A1 *and* A2 *can be difficult to* distinguish where there is no core control. This is partly due to ubiquitous calcite-cemented zones which suppress the SP and give anomalously high resistivity kicks (Iwuagwu and Lerbekmo, 1984). Consequently, the log trend of a particular facies association can be totally altered. Furthermore, log responses of A1 are similar to those of facies 3a (of A2; Figure 2.5b & e), posing some difficulty in distinguishing one facies from the next. These uncertainties do not interfere with the overall interpretation due to good core control in the area of interest.

Facies 1 (Figure 2.5a) comprises sediments of the underlying Lea Park Formation and consists of interbedded marine mudstones and sandstones which are subdivided into two types based *on the trace fossils and sand* content. The basal Belly River comprises two major sandstone facies; shoreface (Figure 2.5b,c) and channel sandstones (Figure 2.5e,f). The shoreface sandstones have blocky or funnel-shaped log responses which delineate sharp-based and gradually coarsening-upward sandstones

Figure 2.5: Sp-Res log responses for the facies of the study area.

a) Offshore mudstones of the Lea Park Formation (Facies 1a and 1b).

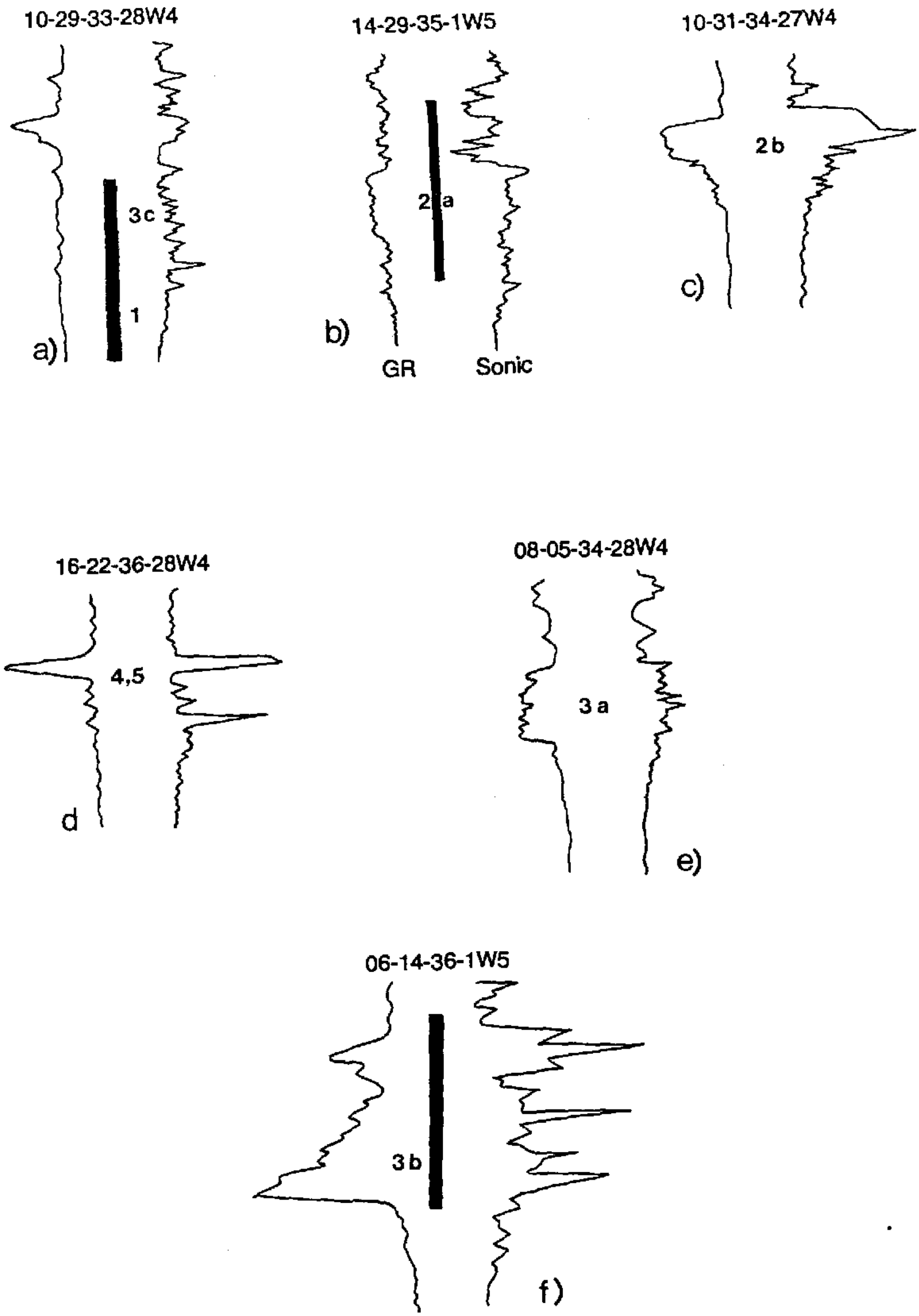
b) Blocky log response of a sharp-based shoreface sandstone.

c) Gradational-based shoreface sandstones.

d) Erratic log responses representing non-marine and brackish water interbedded mudstones and sandstones.

e) Blocky response representing predominantly sandy channel fills.

f) Bell-shaped sp log response of a fining-upward channel fill.


10-29-33-28W4
3 c
1
a)
14-29-35-1W5
2 a
b)
GR
Sonic
10-31-34-27W4
2 b
c)
16-22-36-28W4
4,5
d
08-05-34-28W4
3 a
e)
06-14-36-1W5
3 b
f)

respectively. Channel fill facies are subdivided into three types. The log responses for the channel fill facies are blocky, bell-shaped or serrated with no vertical trend (Figure 2.5e,f). Non-marine and brackish water fine-grained deposits have erratic log signatures commonly of low resistivity with thin higher resistivity kicks (Figure 2.5d). The last two facies are transgressive deposits consisting of conglomeratic lag and mudstones. The facies are described and interpreted in the next two chapters, using predominantly cores and supplemented by log facies.

3. Sediments of the Lea Park Formation

3.1 Lea Park Formation

An understanding of the sediments of the Lea Park Formation is essential to the correct interpretation of the transition from marine to continental deposits, as represented by the Lea Park/Belly River boundary (Chapter 5).

The Lea Park Formation is a 130 m thick section of marine mudstones underlying the Belly River Formation. However, only the topmost part has been cored, with a maximum continuous section of 18 metres (Figure 3.1, 3.2). Most cores only penetrate the top 6 m of the formation.

One facies is present, subdivided into two types on the basis of sand content, bioturbation and physical structures.

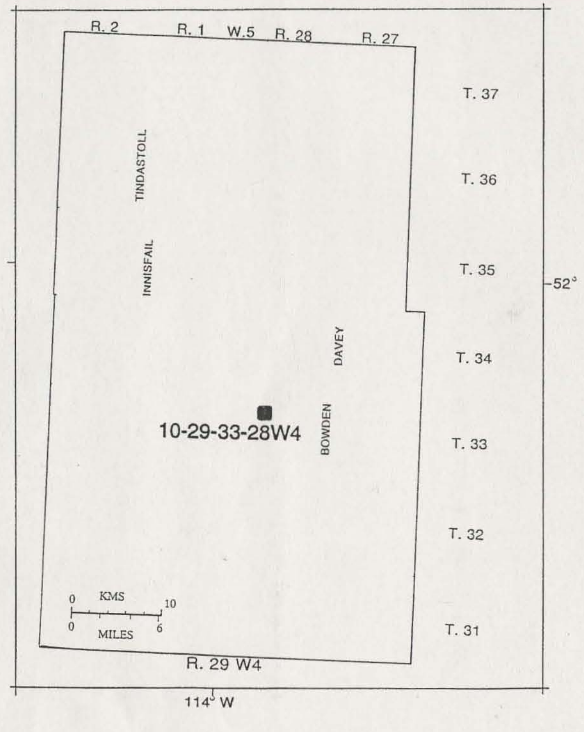
3.2.1 Facies 1a: *Helminthopsis* mudstones

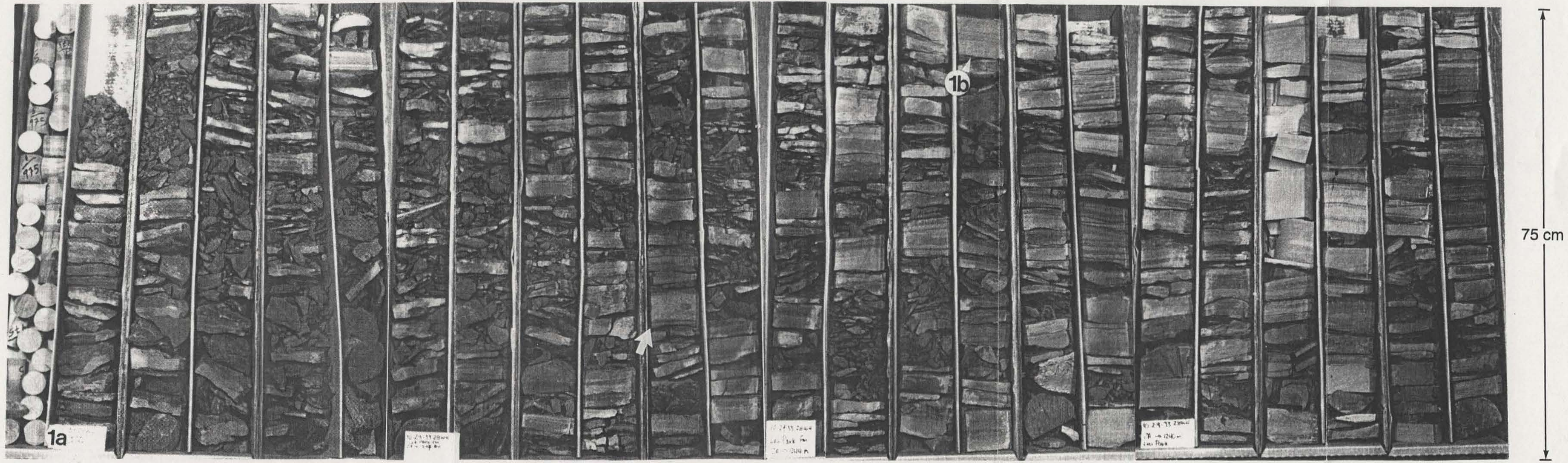
This facies has a minimum thickness of 10 m and consists predominantly of fissile, crumbly mudstones, with thin siltstone and minor very fine-grained sandstone interbeds (Figure 3.1). Sandstone and siltstone interbeds vary from 1-15 cm in thickness, whereas mudstones beds may be up to 50 centimetres. Sandstones are invariably sharp-based. Sedimentary structures within the sandstones include undulatory low angle laminations and few symmetrical ripples. The sandstone beds increase in thickness towards the top. Trace fossils are present within the mudstones and siltstone beds and include *Helminthopsis* (Figure 3.3a) and *Planolites,* with

Figure 3.1: Continuous core through facies 1a and 1b of the Lea Park Formation. Arrow indicates position of facies photography in Figure 3.3a. Note the increase in sandstone bed thickness and presence of soft sediment deformation structures towards the top.

(Well 10-29-33-28W4,1229-1247m).

In all core boxes, the base is at the bottom left and top is to the right. Way-up is from the base to the top of each box.


R. 2
R. 1
W.5
R. 28
R. 27
T. 37
T. 36
T. 35
T. 34
T. 33
T. 32
T. 31
52°
TINDASTOLL
INNISFAIL
DAVEY
BOWDEN
10-29-33-28W4
0 KMS 10
0 MILES 6
R. 29 W4
114° W


1a
1b
75 cm

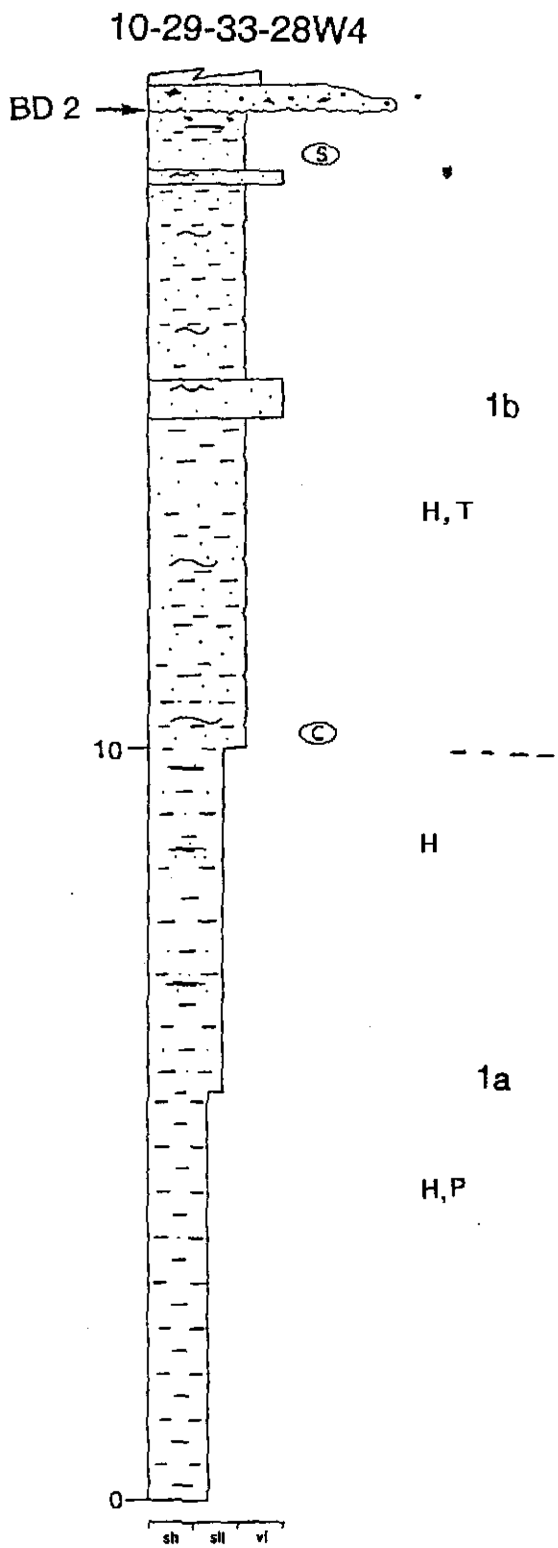


Figure 3.2: Stratigraphic log of facies 1a and 1b (from Figure 3.1).

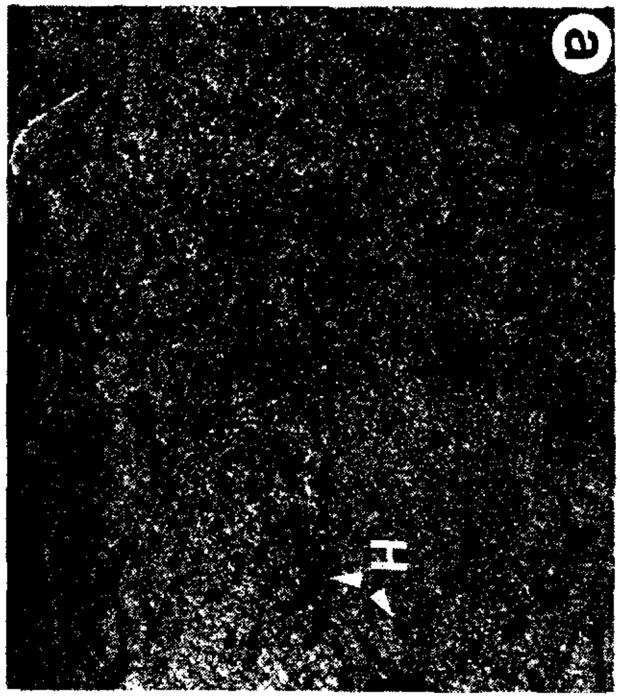
Figure 3.3: Facies 1;

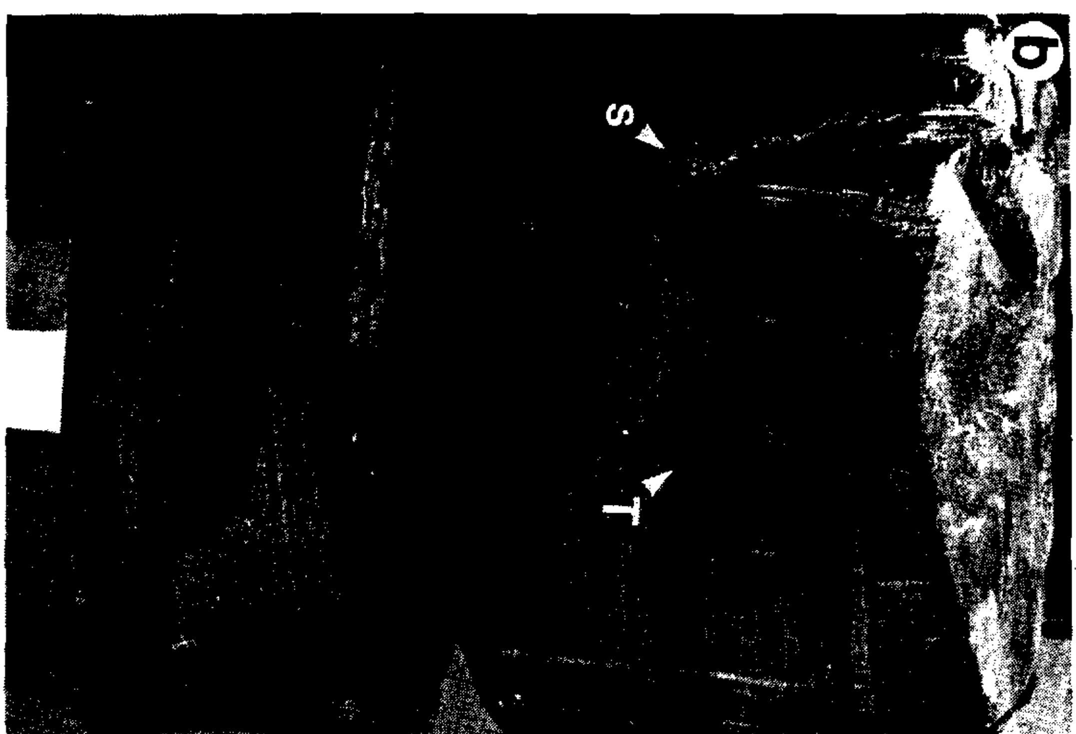
a) Facies 1a, offshore mudstones with *Heminthopsis* (H).

(Well 10-29-33-28W4, at 1244m)

b) Facies 1b, Interbedded sandstones, siltstones and mudstones, with *Teichichnus* burrow (T) and synaerisis cracks (s).

(Well 06-35-34-1W5, at 1238.5m)


a
H


b
S
T

rare burrows of *Terebellina* and rarer *Teichichnus*. These trace fossils constitute a "distal Cruziana" assemblage (MacEachern and Pemberton, 1992).

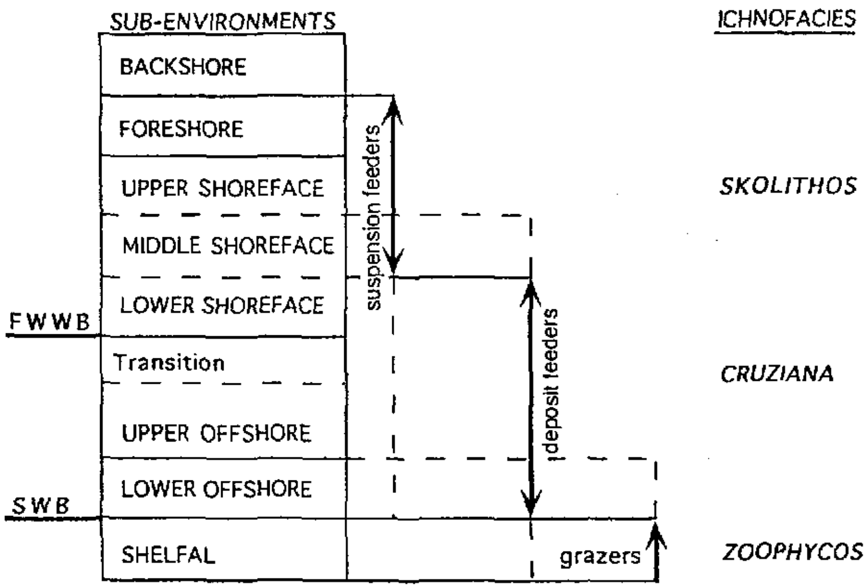
3.2.2 Interpretation

The predominance of mudstones, and the trace fauna indicate a low energy marine environment. The low-angle stratification probably represents hummocky cross-stratification (HCS) and the sandstones are therefore interpreted as storm deposits. The sandstone beds are relatively few and thin in this facies suggesting minimum storm activity. These storm beds and the trace fauna assemblage suggest a lower to upper offshore environment (MacEachern and Pemberton, 1992; Figure 3.4a), above storm wave base. However, in their shoreface model, the offshore sediments are commonly bioturbated, a feature which is not present within the cored interval of the Lea Park Formation. The shoreface model of MacEachern and Pemberton (1992) is used in the interpretation of the environments rather than Walker and Plint (1992; Figure 3.4b) because the trace faunal assemblage are present in sufficient detail to warrant the use this model.

3.2.3 Facies 1b: Banded mudstones

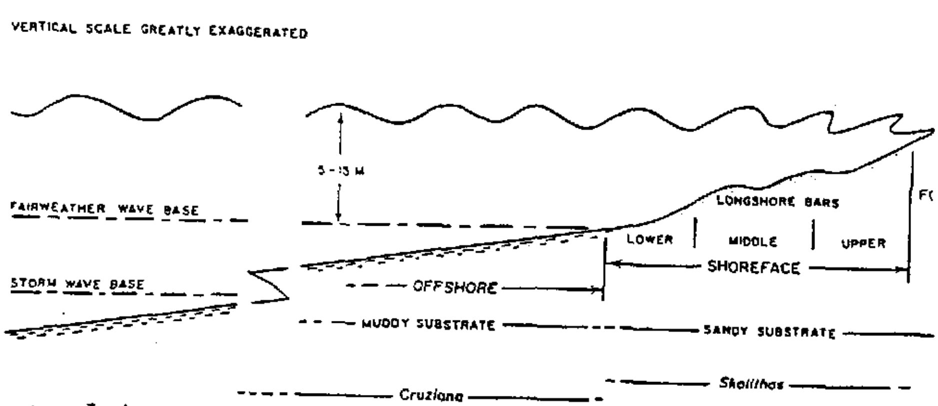
This facies consists of centimetre to decimetre scale mudstone, siltstone and sandstone interbeds with a minimum thickness of 8m (Figure 3.1, top ten boxes, 3.2). Compared with facies 1a, this facies shows an increase in siltstone and sandstone beds, a scarcity of *Helminthopsis* and the presence of synaeresis cracks. Two scales of

Figure 3.4: Shoreface model using ichnofacies, of MacEachern and Pemberton (1992) which is used in the interpretation of the subenvironments. This is compared with the shoreface model of (Walker and Plint 1992; b). Note the subdivision of the upper offshore in a).



SUB-ENVIRONMENTS
ICHNOFACIES
BACKSHORE
FORESHORE
UPPER SHOREFACE
MIDDLE SHOREFACE
LOWER SHOREFACE
FWWB
Transition
UPPER OFFSHORE
LOWER OFFSHORE
SWB
SHELFAL
suspension feeders
deposit feeders
grazers
SKOLITHOS
CRUZIANA
ZOOPHYCOS

a)



VERTICAL SCALE GREATLY EXAGGERATED
FAIRWEATHER WAVE BASE
STORM WAVE BASE
LONGSHORE BARS
LOWER
MIDDLE
UPPER
SHOREFACE
OFFSHORE
MUDDY SUBSTRATE
SANDY SUBSTRATE
Skolithos
Cruziana
Zoophycos

b)

sandstone interbeds were recognised. The first scale of interbedding comprises 5-30 cm thick graded beds of sharp-based lenticular sandstones, siltstones and mudstones (Figure 3.3b). Symmetrical ripples and low angle to parallel lamination are common, although some beds appear structureless. Soft sediment deformation is ubiquitous and includes flame structures and structureless, deformed silty mudstone beds (Figure 3.5a,b). The mudstone and siltstone beds contain an abundance of disseminated carbonaceous material.

The second scale of interbedding consists of clean sharp-based sandstones beds, up to 50 cm thick (Figure 3.1). These commonly appear structureless or with low-angle stratification. The laminae may be gently undulating (Figure 3.6a) and intersecting at low angles. Commonly associated with these sandstones are millimetre thick carbonaceous laminations (Figure 3.6a,b). The laminations can be in pairs or triplets separated by thin sandstone laminae, however, no bundling was observed. The trace fauna includes *Planolites, Asterosoma, Teichichnus*, and rarer burrows of *Rhizocorallium* and *Zoophycos*. These are typical burrows of the *Cruziana* ichnofacies. Some opportunistic and escape burrows were observed within the low-angle stratified sandstone beds (Figure 3.6b).

3.2.4 Interpretation

The presence of sharp-based sandstone beds with low-angle stratification, intercalated with mudstones, provides evidence of intermittent higher energy activity between periods of quieter mud deposition. The low-angle undulating stratification is interpreted to be HCS and its presence with sharp-based sandstone beds indicates sand emplacement during storms.

Figure 3.5. Facies 1b, soft sediment deformation structures.

a) Flame structures formed by rapid loading of sand onto unlithified mudstones.

b) Massively deformed structureless siltstones which may be the result of slumping. (Well 04-20-33-26W4, at 1021m)

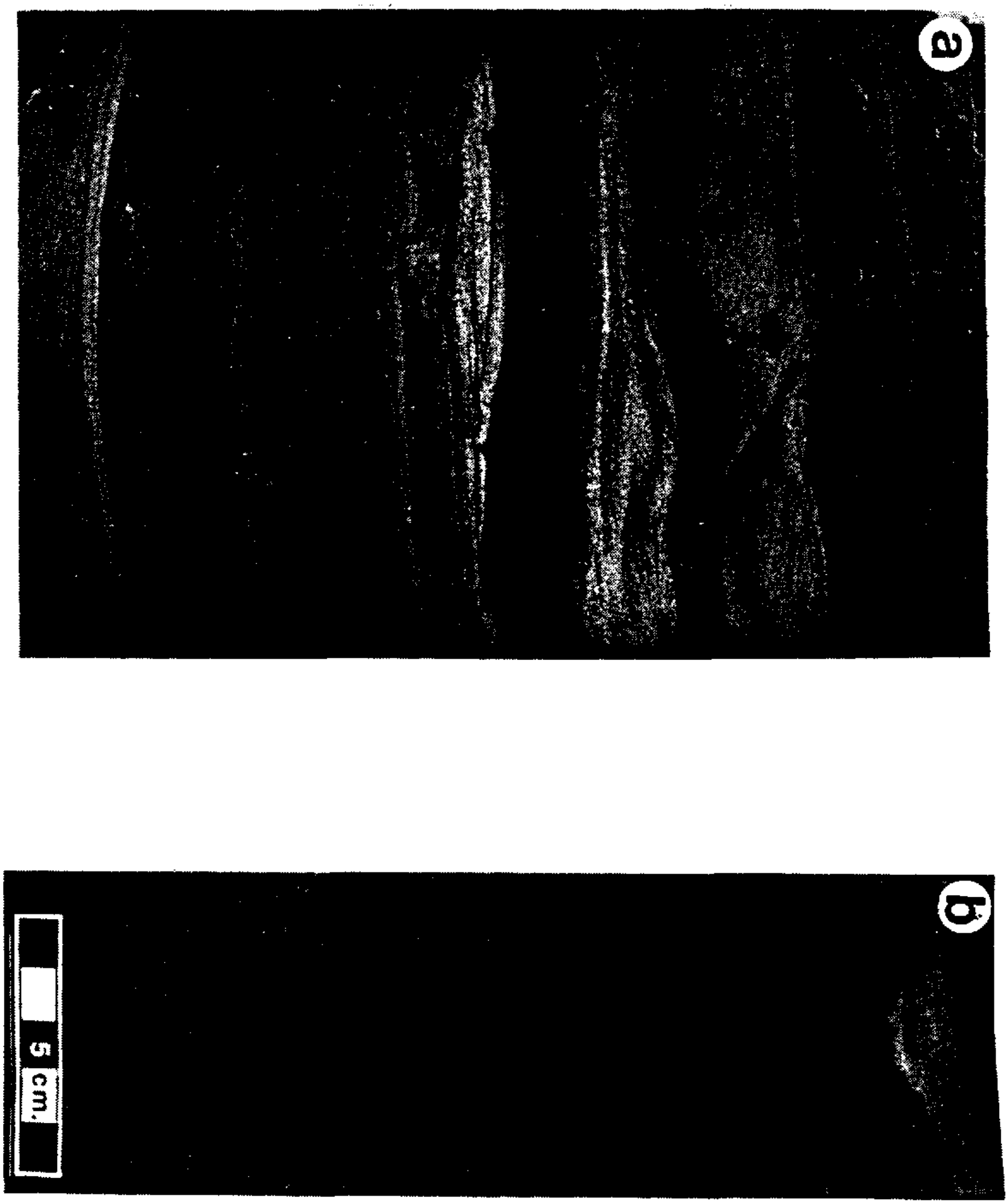
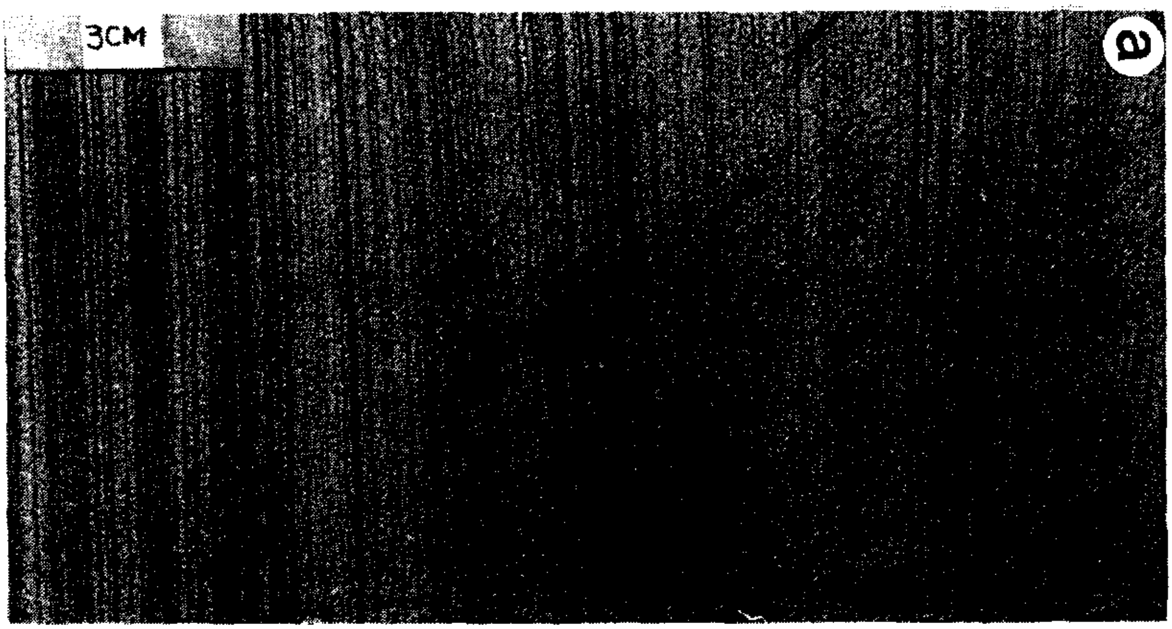
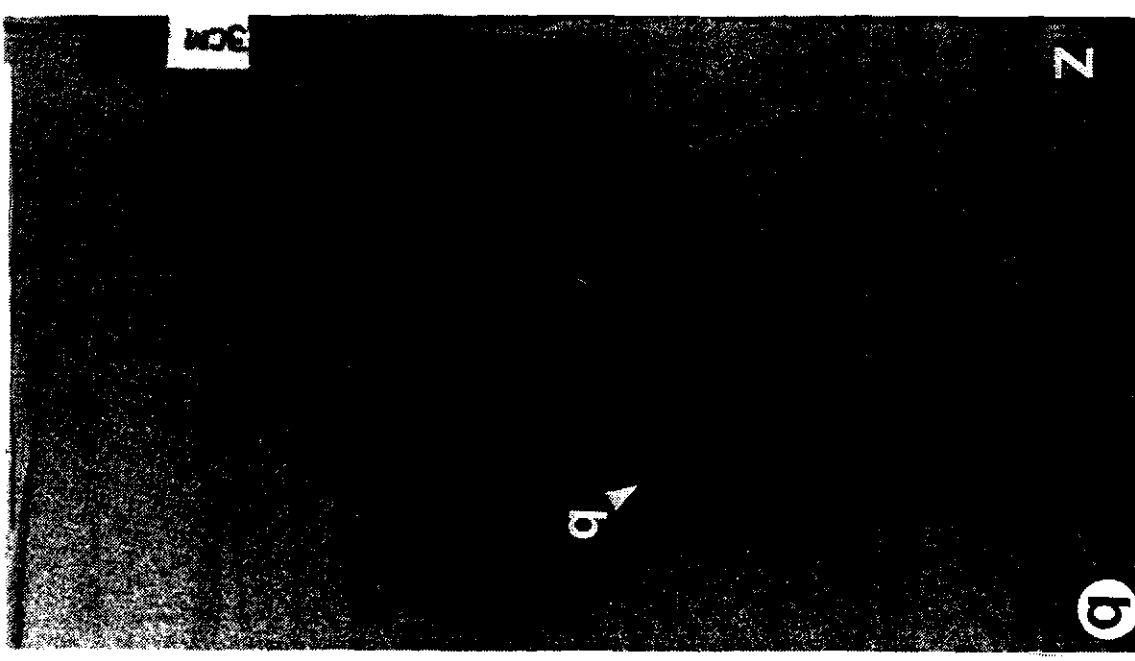

a
b
5 cm.

Figure 3.6: Sandstones of facies 1b.

a) Fine-grained sandstone, with low-angle, gently undulating laminae interpreted to be HCS. (Well 10-24-33-29W4, at 1261.1m)

b) Escape structures? (b) and *Zoophycos*(z) burrow in HCS sandstone bed. (Well 08-04-34-28W4, at 1215m)


3CM
a


N
b
b

The presence of a *Cruziana* ichnofacies assemblage suggests deposition within a moderate to relatively low energy environment, typically below fairweather wave base (FWWB) but above storm wave base (SWB). The thicker storm beds and a low diversity of trace fossils indicate deposition in shallower water, possibly in the upper offshore to transitional into the lower shoreface (MacEachern and Pemberton, 1992; Figure 3.4). The sharp-based graded beds suggest waning storm-generated flows or density undercurrents. Synaeresis cracks indicate either dewatering of rapidly deposited sediments during compaction or contraction of the mudstones due to changes in salinity. Soft sediment structures could be formed by rapid loading of sand onto unlithified mudstones and the massively deformed siltstones which could be the result of slumping.

The presence of carbonaceous laminations (some paired) defining the HCS sandstone beds presents difficulties in interpreting them as formed by tidal processes. Hunter and Clifton (1982) observed laminae in HCS beds composed of sand with carbonaceous and micaceous sand couplets, and suggested that they may be result of rapid deposition by waning of individual wave orbitals. It is not clear how the couplets observed in the study area were formed.

Facies 1a grades upward into Facies 1b, resulting in an overall coarsening-upward trend (Figure 3.2).

4. Description and Interpretation of Facies of the Basal Belly River Formation

4.1 Introduction

The deposits of the basal Belly River in the study area can be divided into five facies. Each of the facies is further subdivided into several types. Shoreface sandstones are of two types, sharp-based and gradational shoreface sandstones. Channel sandstones are of three types. These are subdivided on the basis of grainsize, internal vertical trends, the overall vertical trend and their lateral continuity. Fine-grained facies comprise non-marine floodplain deposits and brackish water open bay sediments. Finally, transgressive deposits are of two types; conglomeratic lag and marine mudstones. The facies are described and interpreted below.

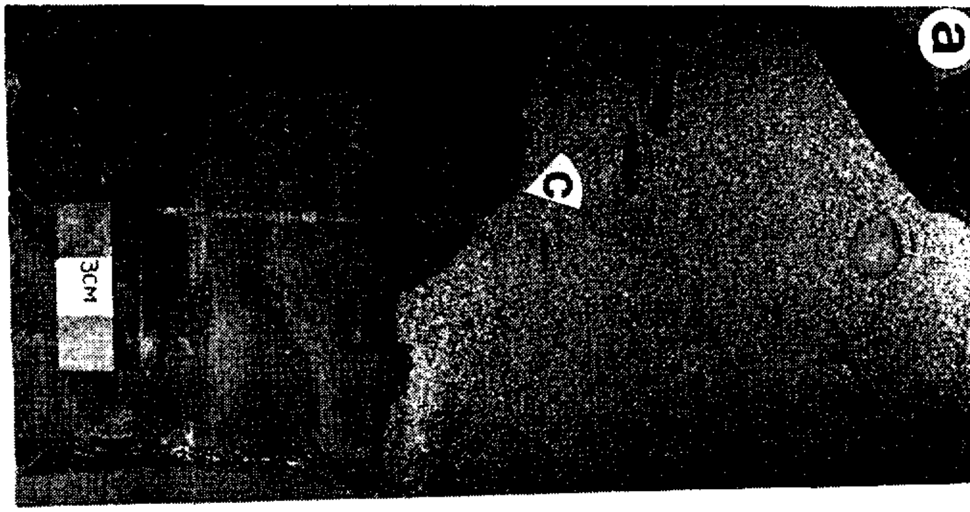
4.2 Facies 2a: Sharp-based prograding shoreface sandstones

The sediments of facies 2a comprise fine- to medium-grained sandstones, up to 12 m thick. The sandstones are sharp-based, overlying facies 1 of the Lea Park Formation (Figure 4.1a) or coal and coaly mudstones of facies 4 (Figure 4.1b). In the latter case, sand filled burrows cut into the underlying coal. Detailed description of the contacts is given when describing individual allomember successions. The basal metre of the sandstones contains sideritised ripped-up clay clasts (Figure 4.2) or coalified

Figure 4.1:

a) Sharp-contact between Lea Park offshore mudstones and shoreface sandstones. Contact is the RSE, BD 1. (Well 14-22-35-1W5, at 1202.8m)

b) Sharp-contact between coal and shoreface sandstones (BD3). Sand-filled burrows (b) within the coal may be characteristic of the *Teredolites* ichnofacies ? (Well 06-04-36-1W5, at 1176.8m)


a
c
3CM

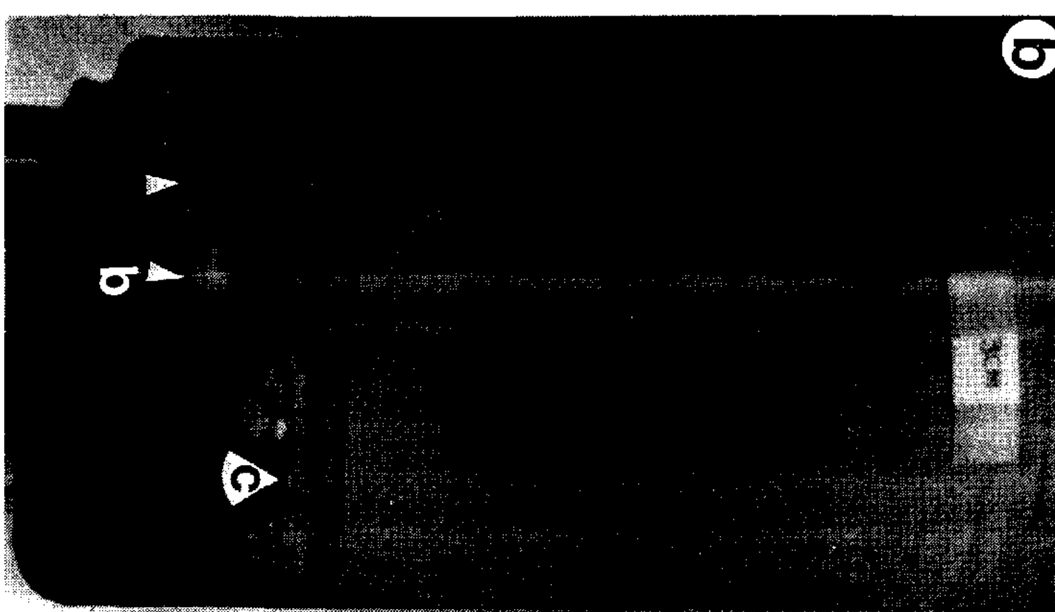
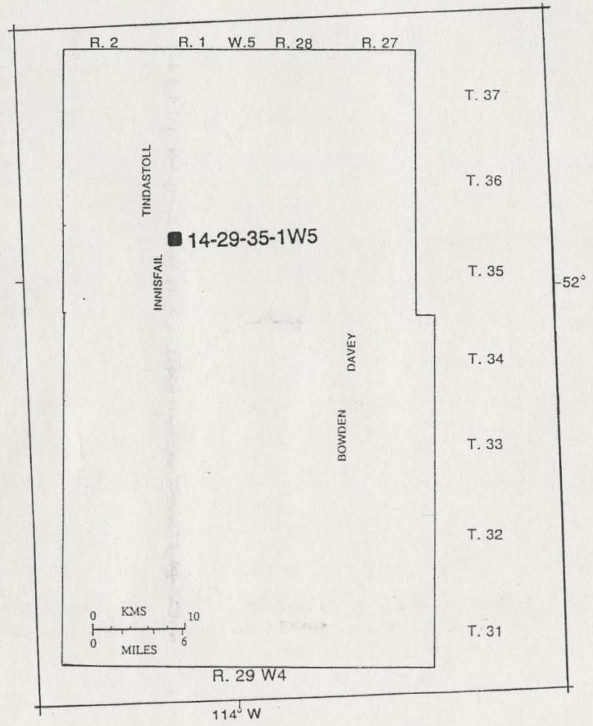
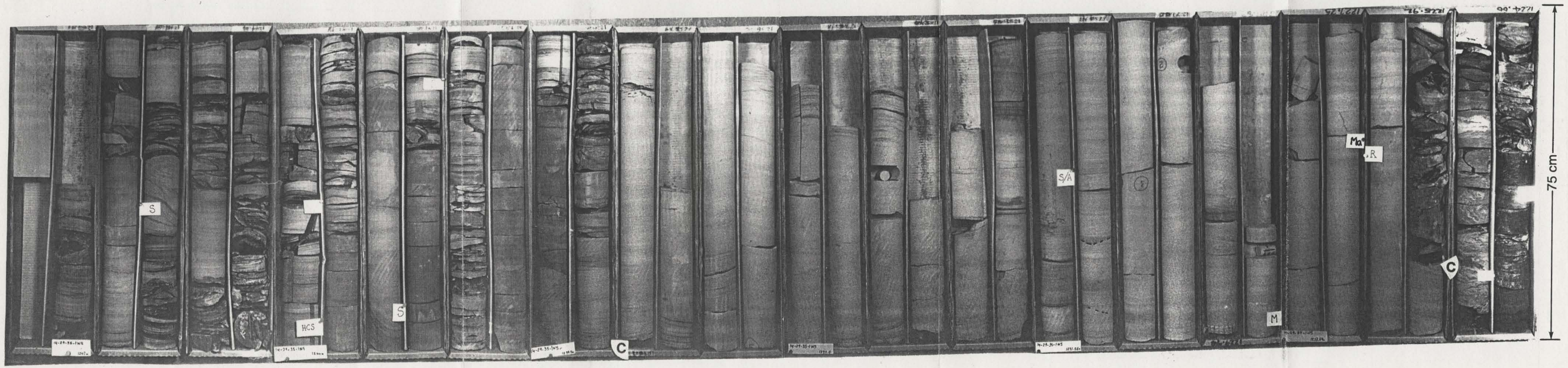

b
b
c

Figure 4.2: Continous core through sharp-based shoreface of allomember 1. Note the diffuse low-angle stratification highlighted by dark organic material, and the infilling of small scale scours in the sandstones beds. *Skolithos* (S), *Asteresoma* (A) and *Macaronichnus* burrows (Ma) are present as well as root traces (R) at the top of the succession.
(14-29-35-1W5, 1224-1247m).


R. 2
R. 1
W.5
R. 28
R. 27
T. 37
T. 36
T. 35
T. 34
T. 33
T. 32
T. 31
52°
TINDASTOLL
INNISFAIL
14-29-35-1W5
DAVEY
BOWDEN
KMS
0
10
MILES
0
6
R. 29 W4
114° W


S
HCS
S
C
S/A
Ma
R
M
C
75 cm

plant debris and disarticulated shell debris (Figure 4.3). Individual sandstone beds, up to 2m thick are commonly homogeneous but some show coarsening-upward trends. Overall the lower part of sandstone is structureless (Figure 4.4a) or with diffuse low-angle stratification (Figure 4.2). Cross-bedding is scarce and is commonly truncated by the next structureless sandstone bed (Figure 4.4b). Results of x-radiographic analysis of these sandstones indicate that some of the sandstones show diffuse stratification. Figure 4.5a shows a core picture of structureless sandstones of the shoreface succession, with very diffuse stratification in places. When x-rayed, the cores show a crude low- to medium-angle stratification interbedded with truly stuctureless sandstone beds. Structureless sandstone may be up to 4 m thick in the shoreface of A3.

Where present, the stratification is defined by centimetre to millimetre thick carbonaceous laminae. The carbonaceous laminae commonly occur as doublets or triplets separated by thin sandstone laminae. No bundling or cyclicity was observed. Thinner, 20-30 cm sandstone beds have scoured bases with current or climbing ripple stratification. The climbing ripples grade upwards into low-angle stratification (Figure 4.6a). This relationship is commonly associated with an abundance of carbonaceous detritus. In some beds however, the opposite is common, with planar stratification grading upwards into current ripple stratification (Figure 4.6b).

In the top metre of the sandstone, clay clasts are partly sideritised, and planar stratification forms the dominant sedimentary structure. Trace fossils include *Asterosoma,Skolithos* (Figure 4.2) and *Conichnus* burrows (Figure 4.7a) in the middle of the succession, with *Macaronichnus*

Figure 4.3: Sharp-based shoreface sandstone, with abundant coalified plant debris. (06-10-36-1W5, 1171.8-1167.6m)

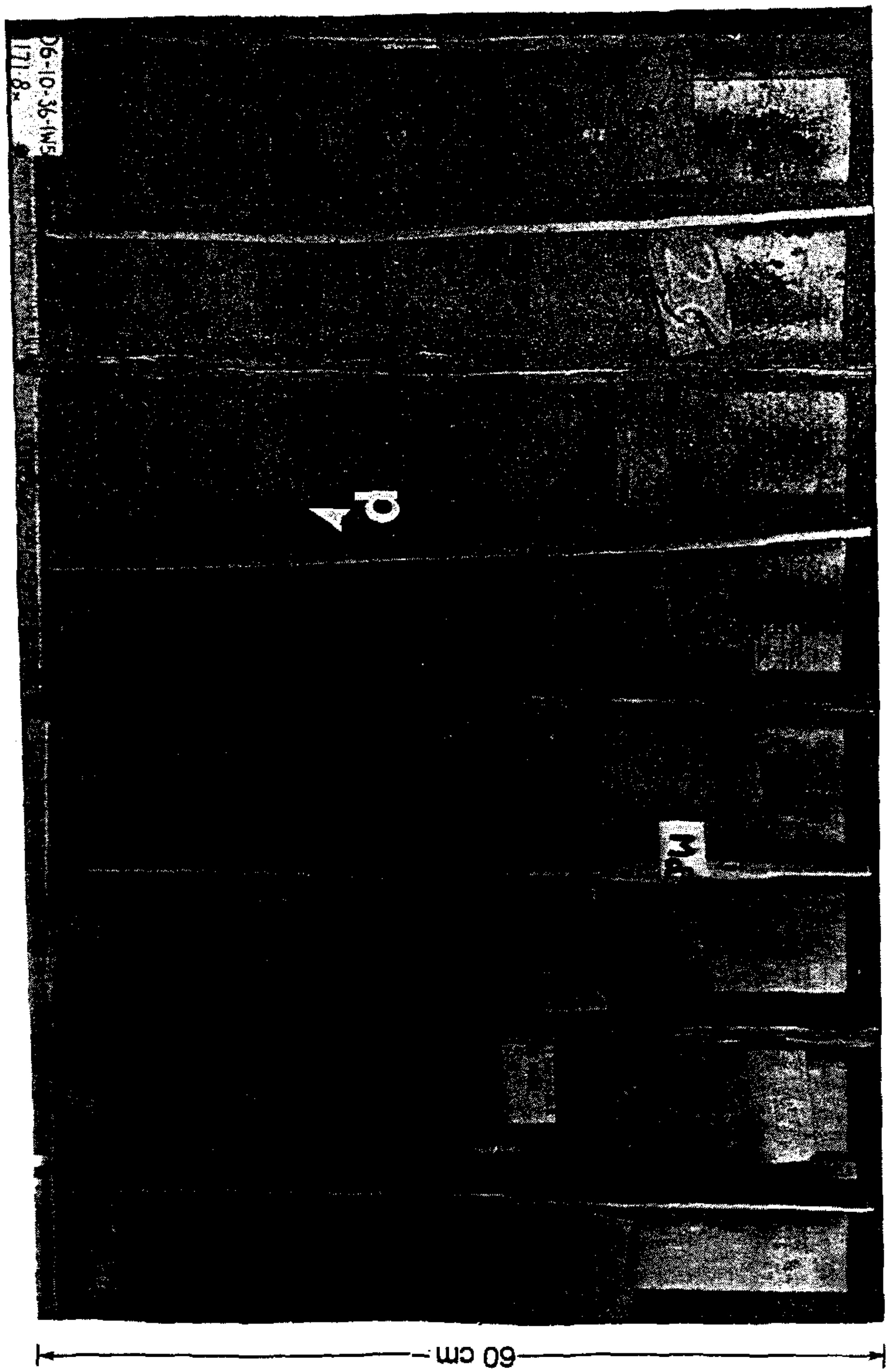
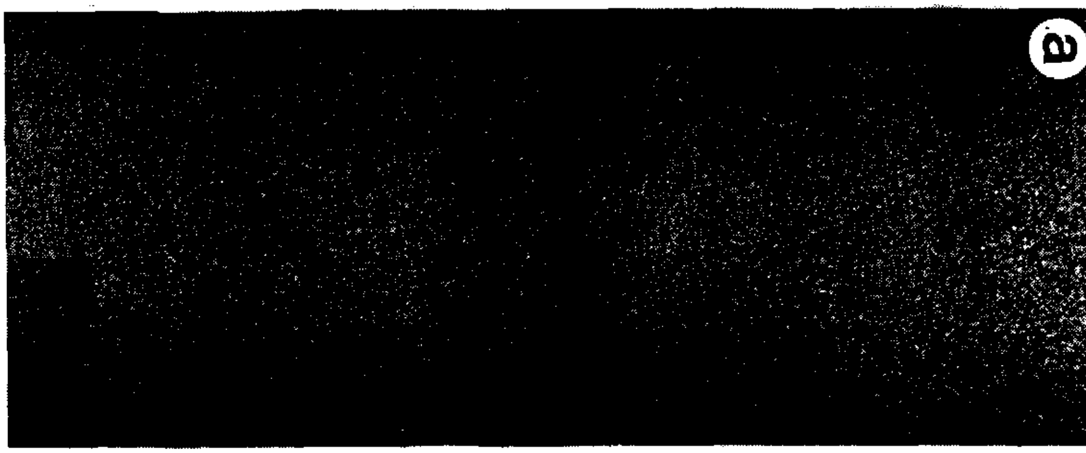

60 cm

Figure 4.4: Facies 2a;

a) Structureless sandstones. (Well 06-15-36-1W5, at 1219m)

b) Cross-bedding truncated by structureless sandstones.

(Well 14-29-35-1W5, at 1234.6m)


a


b

Figure 4.5: X-radiography of structureless sandstones.

a) Crudely stratified shoreface sandstones, with thin structureless beds. Stratification is emphasized when x-rayed.

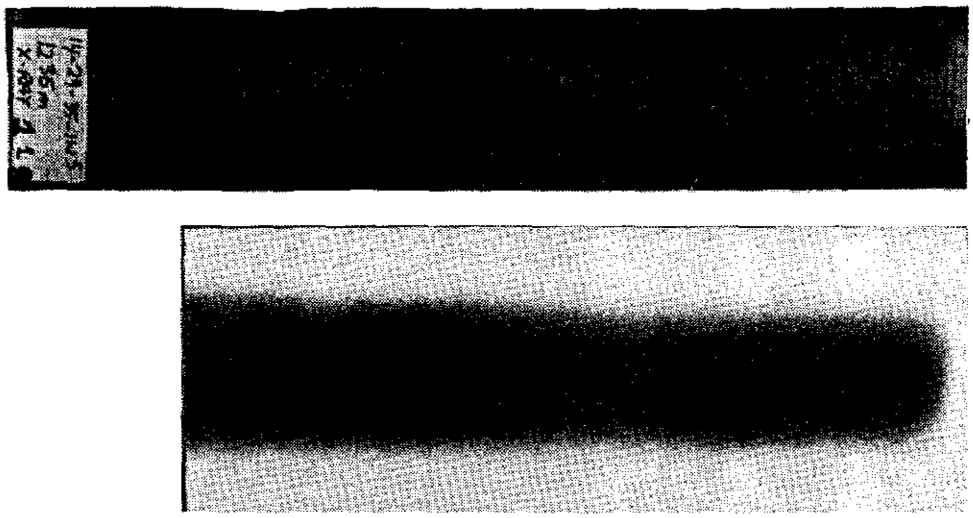
(14-29-35-1W5, 1236m)

b) Calcite-cemented structureless channel sandstones

Sandstones are structureless even when x-rayed.

(05-16-34-28W4, 1213m)

a



b

Figure 4.6: Sedimentary structures within the shoreface sandstones.

a) Climbing ripple stratification which grade upwards into low-angle stratification.

b) Planar stratification grading upwards into ripple stratification.

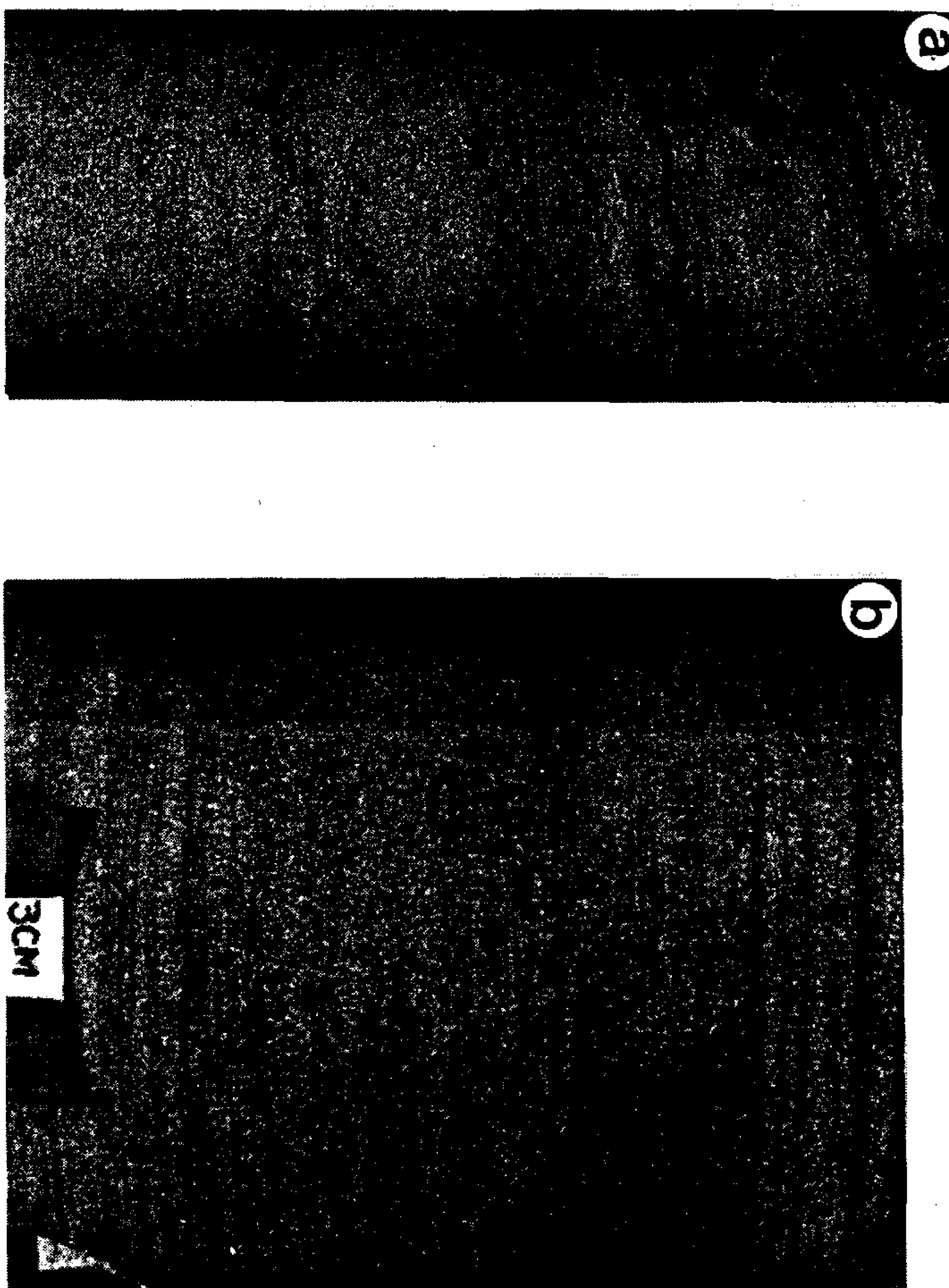

a
b
3CM

Figure 4.7: Trace fossils within the shoreface sandstones.

a) *Conichnus* burrow in the lower part of the shoreface sandstone (Well 06-14-36-1W5, at 1179.4m.

b) *Macaronichnus* burrows, commonly found in the top 3 m of the shoreface sandstones (Well 06-10-36-1W5, at 1169m).


a
Co


b
Ma

(Figure 4.7b) being abundant near the top. In some cored wells, the top of the sandstones shows a purple discolouration, an abundance of carbonaceous material, and root traces.

Sharp-based sandstones are present in A1 and A3. These show some differences which include the nature of their lower bounding contacts, the absence of roots in the shoreface of A3 and the absence of a shell lag and coalified plant debris in A3. *Macaronichnus* trace fossils are abundant in the shoreface of both allomembers.

4.2.1 Interpretation

The trace fauna and the association of this facies with marine mudstones below indicate marine deposition. The presence of *Macaronichnus* (MacEachern and Pemberton,1992) suggest deposition in a highly energetic environment, typically above fairweather wave-base, in the upper shoreface. The planar to ripple stratified sandstone beds reflect deposition of sand during waning storm flows (Walker and Plint, 1992). Preserved climbing ripple stratification highlighted by carbonaceous detritus, suggest rapid and high sedimentation rates. The climbing ripples grade into low-angle stratification and are associated with an abundance of carbonaceous debris which was probably transported in suspension. The transition from ripples and dunes to upper plane bed is attributed to be due to increasing sedimentation rate from suspension (Lowe, 1988). The low-angle stratification present is not flat and may indicate that the transition was not complete.

The abundance of structureless sandstones with crude stratification could indicate rapid deposition of sand out of suspension (Lowe, 1988) or the result of intense biological activity which removed all evidence of stratification. These possibilities are discussed in detail in chapter 6. The uppermost metre shows stratification that is typical of beach deposits, where waves swash and backwash as they break in the surf zone.

This overall sandy succession, lying sharply above marine mudstones, with marine trace fauna and roots at the top suggests a sharp-based prograding shoreface.

4.3 Facies 2b: Coarsening-upward shoreface sandstones

Facies 2b is a log facies (Figure 2.5c) since no cores were available for description. The log trend for this facies suggests a gradational basal contact (Figure 2.5c) from the underlying marine mudstones of the Lea Park Formation or transgressive facies (described below and discussed in chapter 5). The log response is funnel-shaped and shows an increase in sand content from marine mudstones (where cores are available) to a topmost sandstone of 10 m (measured from logs).

4.3.1 Interpretation

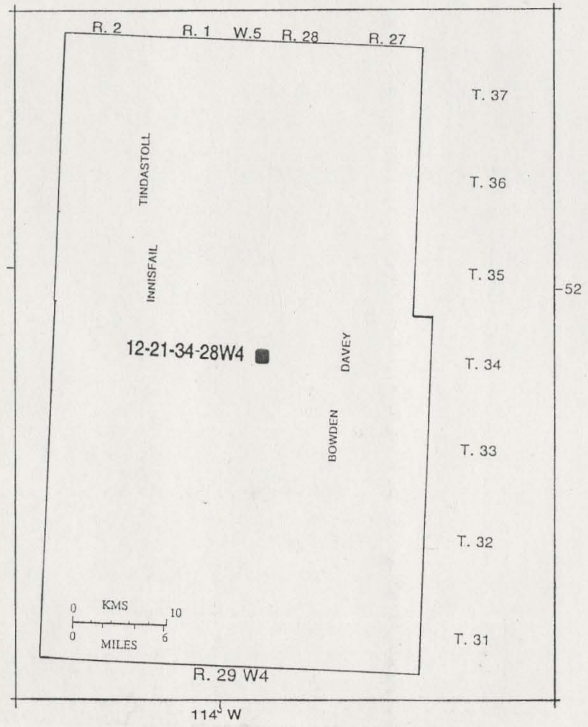
The interpretation of this facies is largely dependent on its association with other facies. The presence of marine mudstones below and the gradual coarsening upward log trend suggest a marine origin. Therefore the coarsening-upward sandstone (gradationally overlying the marine mudstones) suggests decreasing water depth and, may represent a prograding shoreface succession.

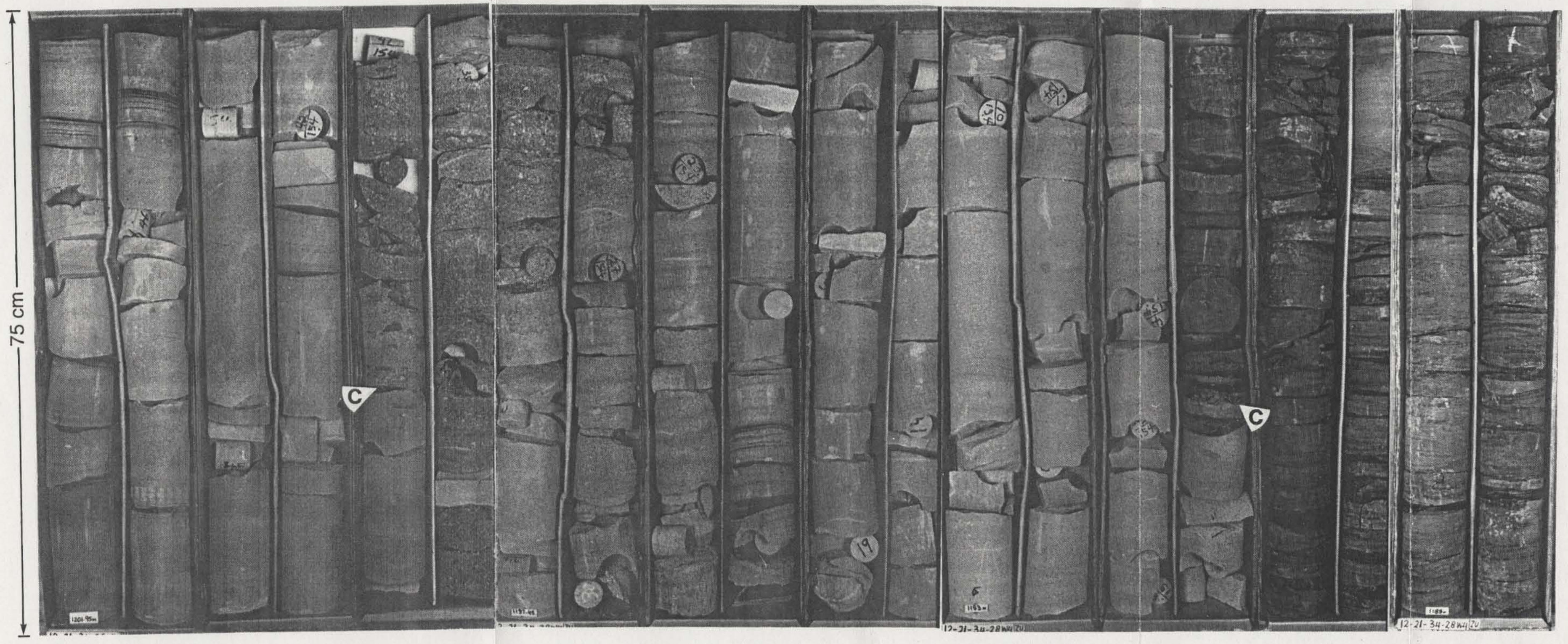
4.4 Facies 3a: Distributary channel fill

Facies 3a consists of a 7-19 m thick stack of sandstones, with individual beds up to 3m thick (Figure 4.8). The base is commonly erosive with a conglomeratic lag of chert and minor clay and coal clasts (Figure 4.9a). In several cores, the conglomeratic lag recurs within the succession. Chert is the most abundant clast in the coarser sand fractions (Iwuagwu and Lerbekmo, 1984; Doig, 1986). In places the pebbles are crudely imbricated. The conglomeratic lag grades into pebbly sandstones and medium- to fine-grained cross-bedded sandstones (Figure 4.9b). Individual cross-bedded sets are 20-60 cm thick (Figure 4.10a). Some beds are structureless or with low-angle stratification. The structureless sandstones, up to 5m thick, are associated with calcite-cementation and when x-rayed, the sandstones are still structureless (Figure 4.5b). Contacts between the sandstone beds are sharp, and are usually defined by an increase in grain size (Figure 4.10b). In several wells, the topmost metre and a half of the sandstone has burrows of *Asterosoma* and mud-lined shafts (Figure 4.10c). Current ripple stratification is only apparent in the top 2 m of the facies.

The overall vertical facies trend can be coarsening-upward, fining upward or with no vertical trends. This facies lacks fine-grained sediments and is dominated by sand. This facies is principally found in the south-

Figure 4.8: Continous core through distributary channel sandstone (facies 3a). The lower contact (c) represents BD2, sitting erosively on shoreface sandstone of allomember 1. Top contact represents BD3, overlain by transgressive lag (facies 6a) and offshore mudstones (facies 6b).
(Well 12-21-34-28W4, 1202- 1187m)


R. 2
R. 1
W.5
R. 28
R. 27
T. 37
T. 36
T. 35
T. 34
T. 33
T. 32
T. 31
52
TINDASTOLL
INNISFAIL
12-21-34-28W4
DAVEY
BOWDEN
0 KMS 10
0 MILES 6
R. 29 W4
114° W


75 cm
C
C

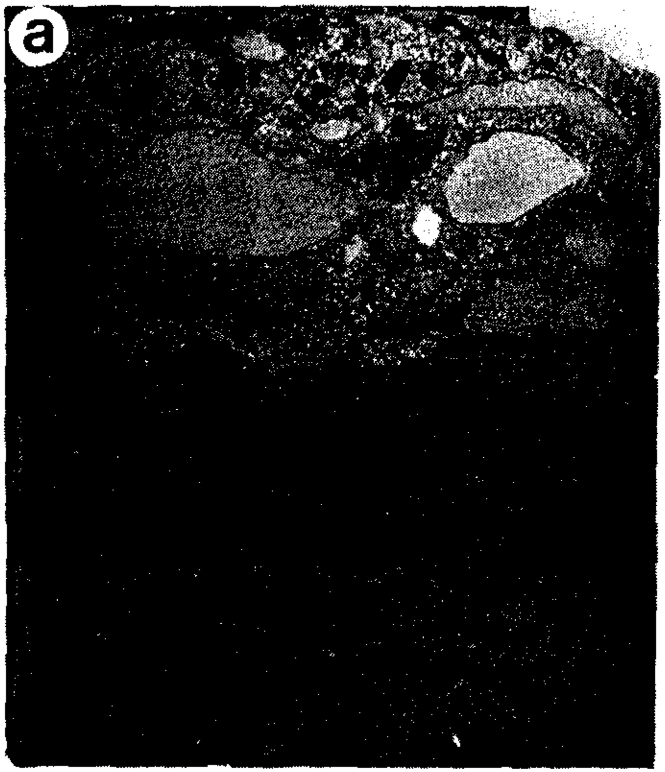
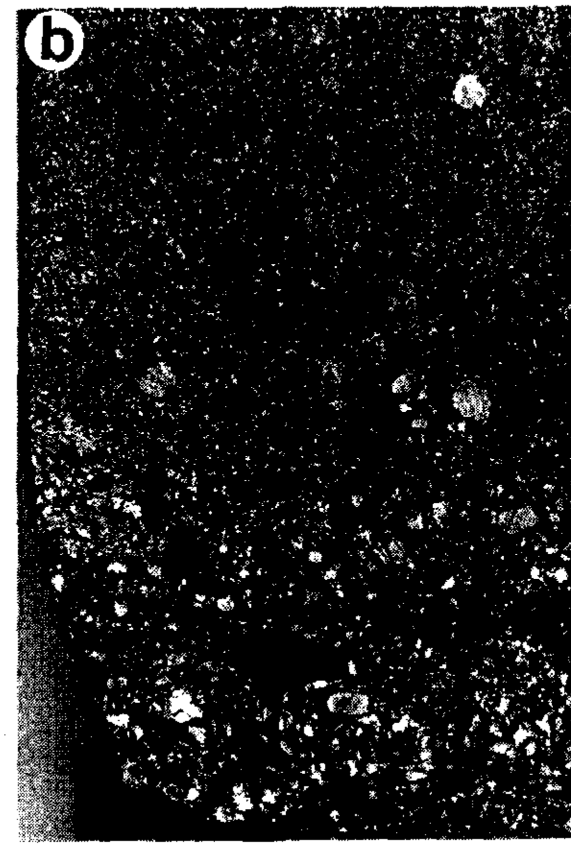


Figure 4.9:

a) Coarse channel lag cutting erosively onto offshore mudstones of the facies 1b (BD2). (Well 14-33-33-28W4, 1186.7m)



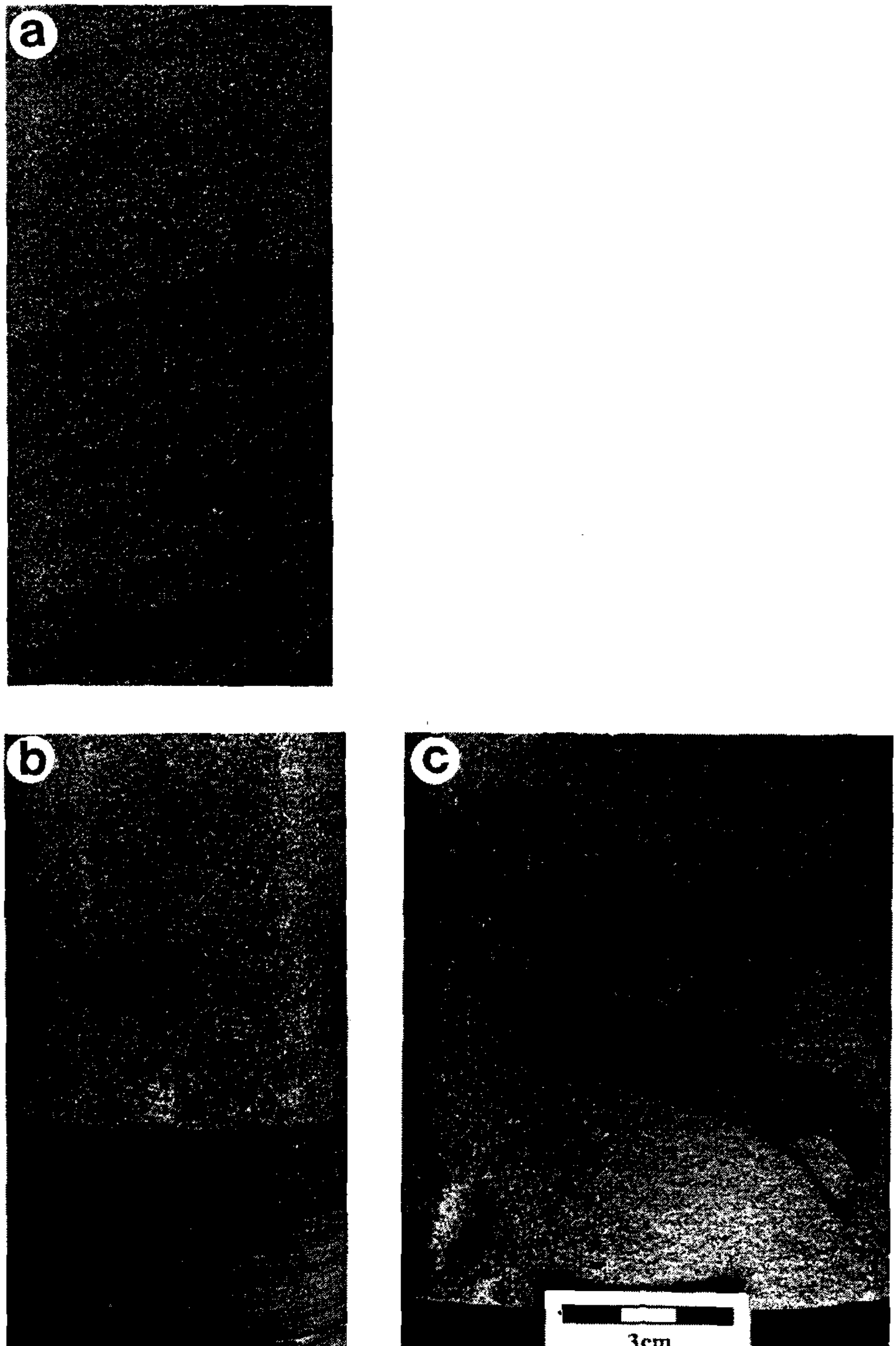
b) Chert-dominated pebbly lag. (Well 14-32-33-28W4, at 1235.7m)

Figure 4.10: Structures within channel sandstones.

a) Cross-bedded channel sandstones (Well 15-30-34-28W4, at 1202.9m)

b) Sharp contact between sandstone beds, defined by grainsize differences. (Well 07-28-32-28W4, at 1209.7m)

c) Robust *Asteresoma* burrow present in structureless sandstones of facies 3a. (Well 11-09-34-28W4, at 1201.4m)


a
b
c
3cm

eastern portion of the study area. Log correlation of this facies shows it is of limited lateral extent and cuts erosively into facies 1 and 2.

This facies is sharply overlain by transgressive marine sediments, commonly a thin conglomeratic sandstone of facies 6.

4.4.1 Interpretation

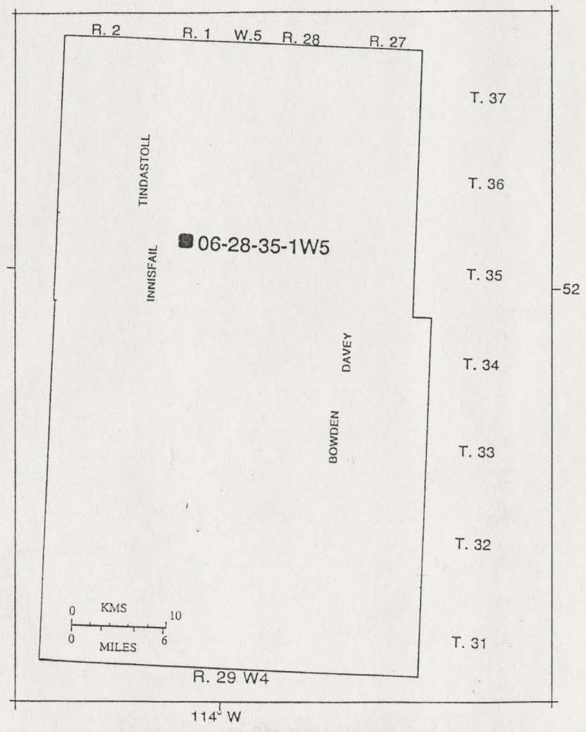
The presence of a basal lag, the dominance of cross-bedding, the absence of fine-grained sediments, and the limited lateral continuity of this facies is consistent with deposits left by frequently switching channels. Internally, the sandstone beds probably represent composite channel fills, evidenced by the stacked sandstone beds with recurring conglomeratic lag in the middle of the succession. Alternatively, the lag could reflect sediment fluctuations in the source area.

The marine fauna present within the top metre of the succession is probably not related to overlying transgressive sediments. The burrows are mud-lined, commonly horizontal and are found up to 1.5 m below the contact with facies 6a. Consequently, marine inundation during low-water stages or in a temporarily abandoned channel allowed colonization by marine organisms and thus offer an explanation for their presence in the channel sandstones.

4.5 Facies 3b: Upward-fining channel fill

Facies 3b is finer-grained than facies 3a, comprising 8 to 19 m thick sandstone succession with an overall fining-upward trend. Individual beds are two to four metres thick (Figure 4.11) and also show fining-upward

Figure 4.11: Continous core through fining-upward channel sandstone (facies 3b). Note the metre thick basal lag of partly sideritised clay clasts. (Well 06-28-35-1W5, 1198-1214m)


R. 2
R. 1
W.5
R. 28
R. 27
T. 37
T. 36
T. 35
T. 34
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52
TINDASTOLL
INNISFAIL
06-28-35-1W5
DAVEY
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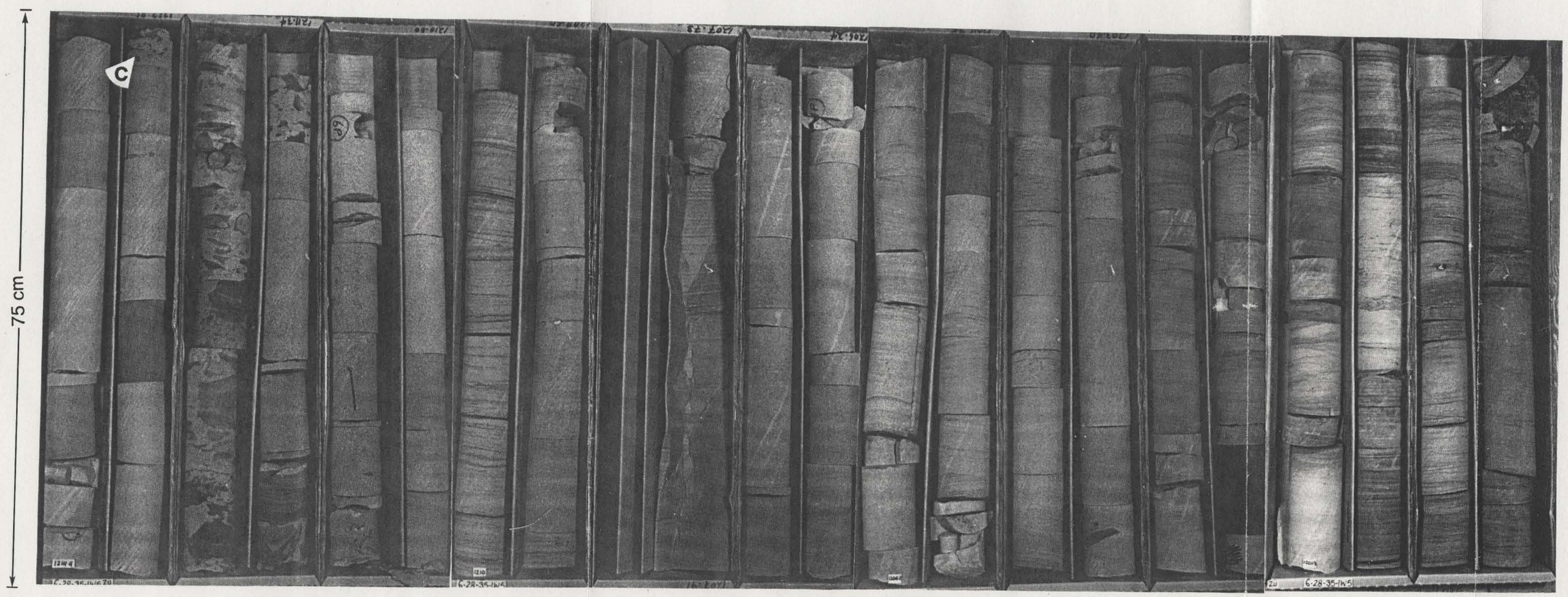
0 KMS 10
0 MILES 6
R. 29 W4
114° W


75 cm
C

trends from medium- to fine-grained sandstones. A basal lag, up to a metre thick in places, cuts erosively into the facies 1b of the Lea Park Formation (Figure 2.2) or into shoreface sandstones of allomember 1 (Figure 4.11). The lag consists of partly sideritised ripped-up clay clasts up to 8 cm in diameter. These clasts are of three types (massive black mudstones, burrowed and slumped mudstones, and finely laminated mudstone), and occur in a fine to medium-grained sandstone matrix. The basal lag grades into medium- to very fine-grained sandstone beds. Internally, beds show two stratification trends, which occur repeatedly throughout;

(i) structureless sandstones grading upwards into low angle discontinuous laminations.

(ii)structureless sandstones grading upwards into cross-bedded sandstone, and up into current- and climbing ripple stratification.

Carbonaceous detritus is common and usually accentuates the stratification. In places, the carbonaceous laminations occur in pairs or triplets in sandstone beds up to 30 cm thick. However, no bundling was observed. The topmost 2 to 6 m of this facies is dominated by very fine-grained sandstones where current-, climbing- and wave-ripple stratification are observed (Figure 4.12a). In turn, these grade into thin mudstone/siltstone interbeds and flaser beds, with root traces (Figure 4.12b) and small unidentified burrows. This succession is overlain by floodplain deposits of Facies 4.

4.5.1 Interpretation

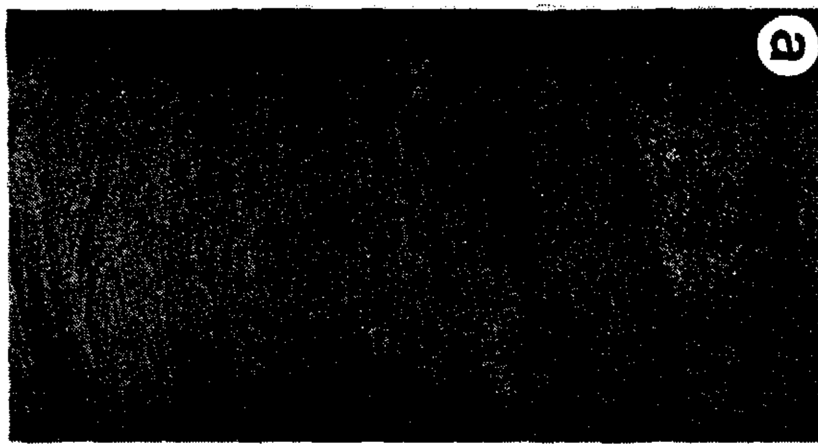
A fluvial channel-fill origin is suggested by the presence of a basal lag, the upward-fining trend, root traces and the lack of marine fauna. The

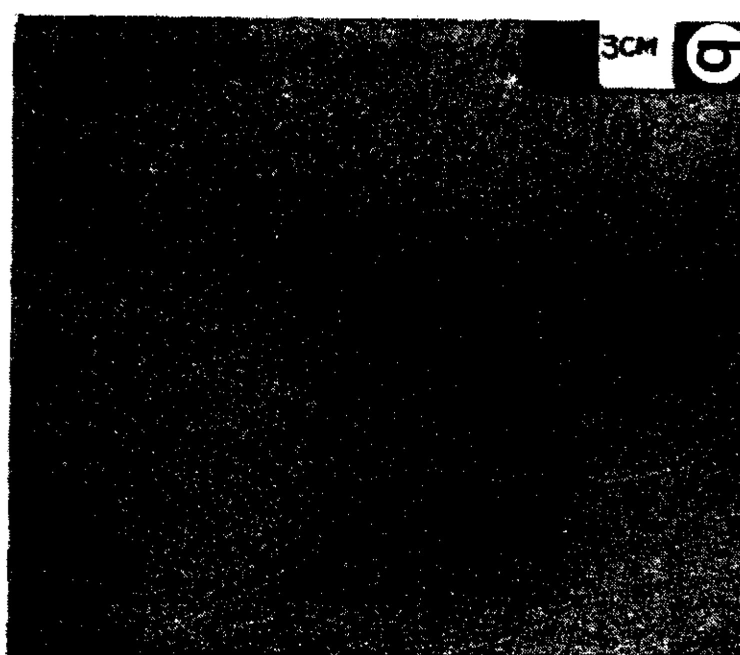
Figure 4.12: Sedimentary structures in facies 3b.

a) Climbing-ripple stratification. (Well 16-14-33-1W5, at 1271m)

b) Root traces in the topmost metre.

(Well 14-34-35-1W5, at 1166.7m)


a


3CM
b

overall vertical fining-upward trend, with a dominance of current and wave ripples at the top of the channel, and the stratification trends within individual sandstone beds, suggest waning flow. The preservation of interbedded mudstones and siltstone with root traces at the top of the succession implies gradual channel abandonment.

Low-angle stratification in thick marine sandstone successions is often interpreted as swaley cross-stratification. Nevertheless, in core, low-angle stratification in medium sandstone could also represent the bottomsets of large trough cross-bedding. Based on the non-marine nature of the deposits, the latter interpretation is preferred to explain the low angle stratification observed.

4.6 Facies 3c: Abandoned channel fill

This facies was only observed in one core, 10-29-33-28W4, and consists of 11 m of mudstones, siltstones and sandstones. The basal contact lies sharply on marine mudstones, and consists of a lag of partly sideritised clay clasts in a medium-grained sandstone matrix. The bulk of this facies consists of fine-grained mudstones and siltstones. Two sandstone beds, up to 3 m thick, fine upwards into siltstones and mudstones with interbedded sandstones. Stratification is dominated by current, climbing and wave ripples. The interbedded sandstones, siltstones and mudstones up to 40 cm thick show evidence of soft sediment deformation, which includes massively deformed mudstones and flame structures. Small unidentified burrows are present in the top 3 m.

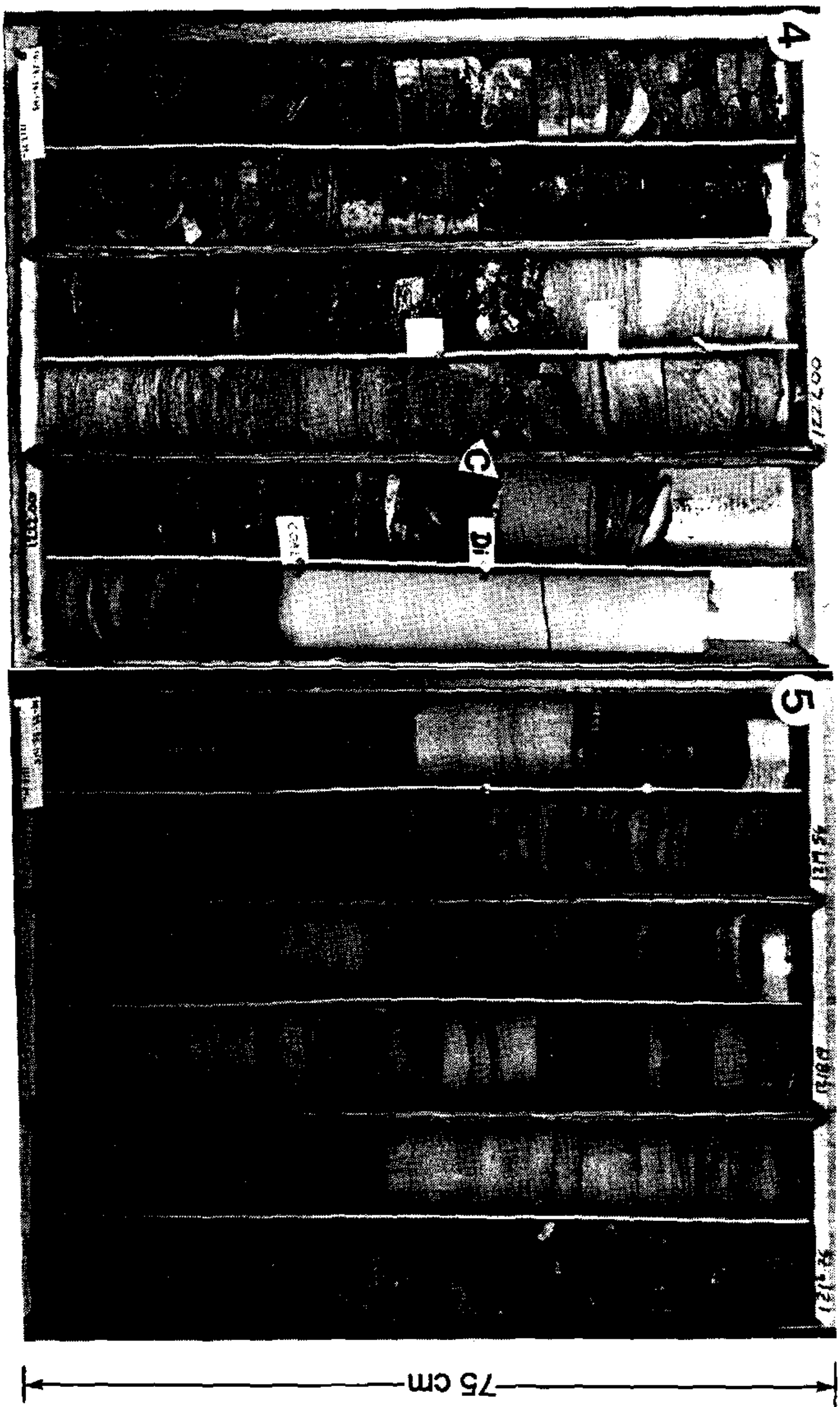
4.6.1 Interpretation

The basal lag, erosively overlying the Lea Park Formation, and log and core correlations (Figure 2.3), indicate that this facies has a channel geometry. The predominance of mudstones and siltstones suggest gradual channel abandonment fill. The presence of sandstone beds suggests at least part of the channel fill was active. The thin sandstone interbeds indicate periodic channel flooding and the introduction of coarser sand fractions during high discharge, in between periods of quieter mudstone deposition. The lack of marine trace fossils indicates that the sediments are non-marine or brackish.

4.7 Facies 4: Interdistributary deposits

This facies is up to 7 m thick and occurs predominantly in the west and northwest portions of the study area. It consists of several facies types but these are rarely all present in any one core. The most common facies is dark, highly organic mudstone (coaly mudstone) which always occurs below and, occasionally also above, a thin (less than 25 cm) coal bed. In places this coaly mudstone is up to 2.5 m thick (Figure 4.13). Commonly, it is overlain by a pale, bleached siltstone, which has abundant roots and shows evidence of soft sediment deformation. Thin, typically fining upwards sandstone beds (commonly less than 2 m) also occur within this facies. They can be totally bioturbated with *Arenicolites,* or show a wide variety of sedimentary structures that include mm-scale parallel carbonaceous laminations, current-, wave- and climbing ripple stratification. The final

Figure 4.13: Continous core boxes of mudstones, thin coal beds, and sandstones of facies 4 and brackish water deposits of facies 5. Facies 5 with HCS beds at the base become progressively non-marine at the top. The contact (c) separates coaly mudstones of facies 4 from open bay deposits of facies 5. *Diplocraterion* (Di) trace burrowed down into the coal. This contact is a marine flooding surface (FS) which is correlated with the TSE (BD3). (Well 14-29-35-1W5, 1215-1223.4m)


4
5
75 cm

facies consists of millimetre to centimetre interbeds of mudstone, siltstones and ripple cross-laminated sandstones. Sediment loading is ubiquitous, and small burrows of *Planolites* are common.

4.7.1 **Interpretation**

Sediments of facies 4 represent interdistributary floodplain deposits. Overbank flooding and crevassing processes are responsible for the deposition of the thin sandstone beds. The mm-scale interbedded mudstone, siltstone and sandstones represent deposition in ponded areas of the floodplain. Roots and palaeosol indicate repeated emergence of bay(s) and ponded areas. The abundance of coaly mudstones with thin coal beds suggests a high clastic input to the swamps. Thin bioturbated beds of *Arenicolites* indicate brackish influence.

4.8 <u>Facies 5: Open bay deposits</u>

The succession comprises 4.6 m of mudstones with thin sharp-based, lenticular siltstone and sandstone interbeds (Figure 4.13). This facies is present in Innisfail and the western part of Tindastoll and closely resembles facies 1b of the Lea' Park Formation, but lacks a well developed marine trace-fauna. A few robust *Teichichnus* and *Planolites* burrows are present. Symmetrical ripples, HCS and graded beds are the dominant sedimentary structures. HCS sandstone beds are 30-50 cm thick with sharp bases, and occur only in the basal 3 m. This facies is gradationally overlain by 5 m of highly carbonaceous rooted mudstones and siltstones of facies 4. The association between facies 4 and facies 5 is crucial to the interpretation of this facies as open bay (below).

4.8.1 **Interpretation**

The trace fossils suggest deposition in a brackish to marine environment. However, the lack of a well-developed marine ichnofacies suggests brackish water rather than a fully marine environment. The mudstone beds indicate quiescent periods of deposition from suspension. Thin graded beds and HCS sandstone beds were probably deposited periodically during storms and floods. This succession is therefore interpreted as deposits of an open bay environment. Further justification for this interpretation is provided by the presence of non-marine facies (4) gradationally above, which suggests gradual infilling and abandonment of the bay.

4.9 Facies 6: Transgressive deposits

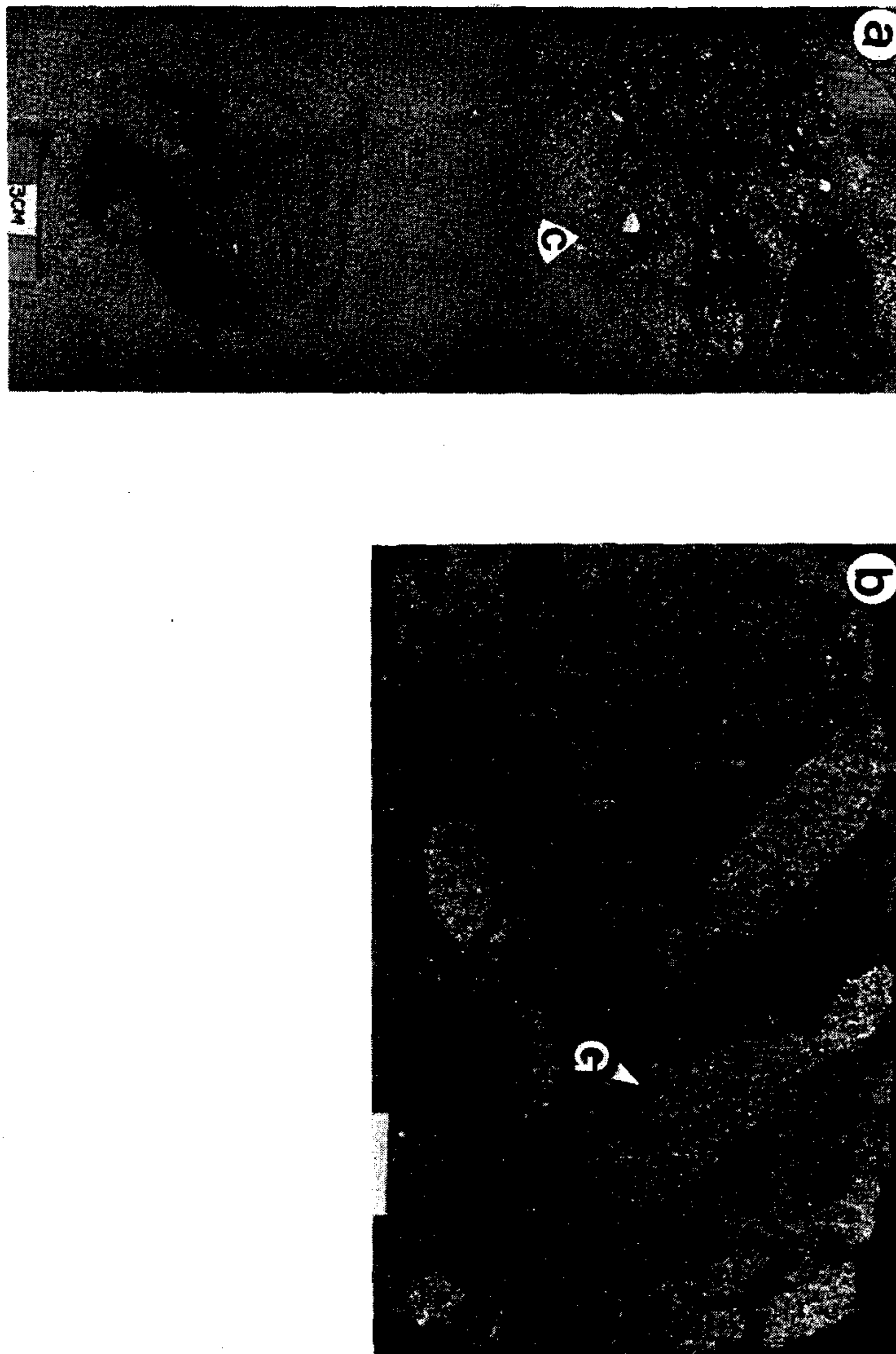
Sediments of facies 6 are present in the southeast and east of the study area and consist of two types. The first facies (6a) consists of a 20-50 cm thick lag of coarse sand or conglomeratic sandstone, with clasts of black chert, partly sideritised mudstones, and plutonic rock fragments (Figure 4.14a). The lag overlies non-marine channel sandstones of facies 3a. Vertical to sub-vertical, unlined and passively filled burrows of *Diplocraterion* and *Thalassinoides* (Figure 4.14b) are present. These trace fossils commonly burrow down into the underlying facies (facies 3a). In

Figure 4.14: Transgressive deposits

a) Sharp contact (c) between trangressive lag of coarse sand, chert pebbles and clay clasts (facies 6a) and channel sandstones of facies 3a. This contact is a TSE (BD3). Burrow within the channel sandstone may be *Teichichnus.* (Well 12-29-33-28W4, 1231.5m)

b) Unlined, passively filled burrows of the *Glossifungites* ichnofacies (G).

(Well 08-17-34-28W4, 1210.3m)


a
3CM
G
b
G

core, this facies is only present in Davey Field (T33-34, R28-27W4) and part of Bowden Field (T34, R1W5).

The second facies (6b) comprises a minimum of 18 m of highly fissile, burrowed black shaly mudstones, with rare silty sandstone beds (Figure 4.8). Commonly, the basal 50 cm of the mudstones contain scattered grains of coarse sandstone (Figure 4.15a). In places, a fining-upward trend is evident, with a thin fine-grained sandstone at the base, which then grades into the mudstones. *Helminthopsis* and *Chondrites* (Figure 4.15b) are the common trace fauna. *Inoceramus* shells (Figure 4.15c) and fish scales have also been identified.

4.9.1 Interpretation

The lag lies above a transgressive surface of erosion (TSE) which was cut across a subaerially exposed surface (MacEachern et al, 1992), of facies 3a. The composition of clasts is similar to the underlying facies, indicating reworking of the underlying non-marine sandstones by waves to produce a thin transgressive lag. The burrows present are typical of the substrate-controlled *Glossifungites* ichnofacies (Pemberton et al, 1992).

The presence of *Chondrites* and *Helminthopsis* and the scarcity of sandstone beds indicate that facies 6b was deposited in an offshore marine environment where wave activity was minimal. The coarser grains within the mudstones at the base suggest reworking of facies 6a. The fining-upward trend observed in some wells also supports increasing water depth.

Figure 4.15: Transgressive deposits

a) Gritty mudstones (Well 16-05-34-28W4, 1215m)

b) *Chondrites* and *Helminthopsis* burrows in offshore mudstones of facies 6b. (Well 11-18-33-28W4, 1240m)

c) *Inoceramus* shell fragments.(Well 08-17-34-28W4, 1207.7m)


c
lc
a
b
Ch

5. Depositional History of Allomembers 1 and 2

5.1 Introduction

This chapter is an interpretive section discussing the lateral and vertical relationships of the sediments of allomembers 1 and 2. This is accomplished by detailed subsurface correlation of core and wireline logs, in addition to the facies descriptions and brief interpretations outlined in the previous chapters. In-field correlation was done using local markers, commonly thin coal beds within the basal Belly River Formation. Between fields, datums within the Lea Park (Figure 1.3) were used. The legend to symbols for the stratigraphic logs is given in Table 5.1.

5.2 Allomember 1

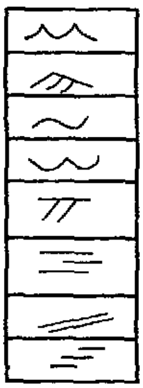
5.2.1 Sediments of allomember 1

Three facies successions are present in allomember 1. However only one (facies succession A, Figure 5.1) was completely cored and the other two successions are log facies identified during detailed log correlation (section 5.2.2). Facies succession A of allomember 1 comprises blocky sandstones (facies 2a), lying sharply on offshore marine mudstones (facies 1b), and the transitional facies of a gradual coarsening-upward facies succession are missing. This sandstone facies, with marine trace fossils,

Table 5.1: Legend to stratigraphic logs and cross-sections for Figures 5.1 to 6.6.

LEGEND

STRUCTURES



Wave ripples

Current ripples

HCS

Trough cross-bedding

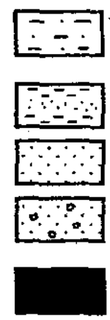
Planar Tabular cross-bedding

Low angle cross-stratification

Stratification

Climbing ripples

LITHOLOGIES



- Siltstones/ mudstones
- Interbedded ssts/ mdsts units
- Sandstones
- Pebbly sandstones
- Coal

GRAINSIZES

Shl - shale
Slt - silt
vf - very fine sand
fs - fine sand
ms - medium sand
cs - coarse sand
vc - very coarse /pebbly sand

- marine mudstone and sandstones
- shoreface sst
- fluvial channel
- floodplain facies

SYMBOLS

- rootlets
- mud clasts
- plant debris
- comminuted plant debris
- coalified plant debris
- synaerisis cracks

C - calcite cementation

S - siderite cementation

TRACE FAUNA

Ar - *Arenicolites*
As - *Asterosoma*
Co - *Conichnus*
Ch - *Chondrites*
Diplo - *Diplocraterion*
G - *Glossifungites*
H - *Helminthopsis*
Ma - *Macaronichnus*
Pa - *Palaeophycus*
P - *Planolites*
Rz - *Rhizocorallium*
Ro - *Rosselia*
Sk - *Skolithos*
T - *Teichichnus*
Te - *Teredolites*
ϴ - burrows

Macrofauna

- *Ostrea*

Ic - *Inoceramus*

TSE transgressive surface of erosion
FR forced regression
RSE regressive surface of erosion
FS flooding surface

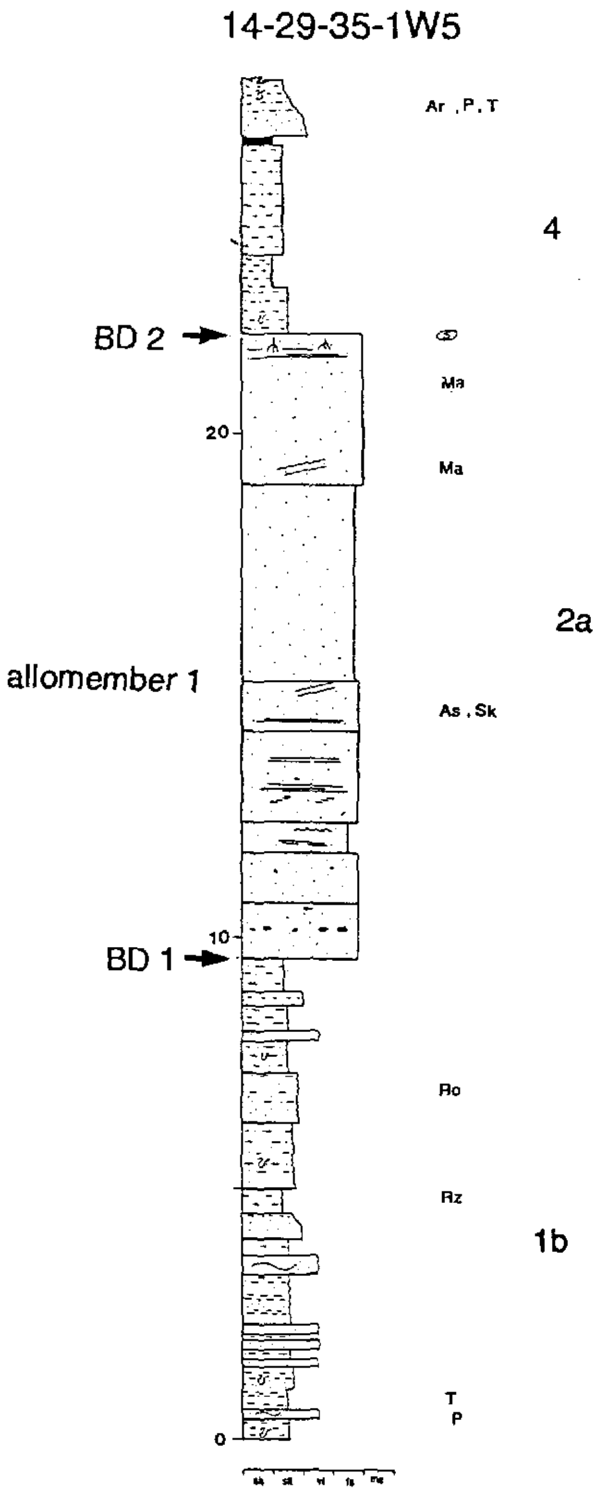


Figure 5.1: Facies succession A of A1.

and root traces in the top of the succession was interpreted in the previous chapter as the deposit of a prograding shoreface.

The nature of the shoreface (i.e. the dominant processes acting at the shoreline) can be interpreted from a detailed examination of the internal sedimentary structures within the facies succession, and the sandbody geometry to be described in section 5.2.2.

The shoreface sandstones of facies 2a preserve climbing ripple stratification (Figure 4.7b) which is associated with abundant carbonaceous detritus. Climbing ripple stratification indicates high sedimentation rates, and the abundant carbonaceous detritus suggests that the shoreface sandstone was deposited close to a delta mouth. In delta fronts, high sedimentation rates, especially during floods, coupled with rapid burial is likely to favour the preservation of climbing ripples. These structures are observed in modern delta front sediments of the Mississippi (Frazier, 1967; Coleman and Wright, 1975; Elliott, 1986) and are also observed in ancient river-dominated delta front facies (Bhattarcharya and Walker, 1991a).

However, to interpret the shoreface as deltaic, good three dimensional control is required. The exact geometry of the shoreface sediments of allomember 1 is not clear (as will be discussed below) but the sedimentary structures suggest that the sandstone was deposited in close proximity to a fluvial input, supplying large amounts of carbonaceous material. With high discharge during floods, the sedimentary structures were quickly buried, preventing local reworking by waves.

Macaronichnus traces occur in the top of the shoreface succession suggesting local high energy conditions due to waves breaking in the nearshore (MacEachern and Pemberton, 1992).

Sedimentary structures of the underlying Lea Park Formation also display evidence of fluvial input. The mudstone and sandstone interbeds of facies 1b have abundant carbonaceous material and soft sediment deformation structures. Storm activity is suggested by the HCS sandstone beds. *The facies succession of allomember 1 therefore show a dominance of both wave and fluvial processes acting on the delta front.*

5.2.2 Lateral facies variation and facies geometry

The well logs of allomember 1 show three different responses (Figure 2.5b,c,d). The most common log response is the blocky coarsening-upward log trend which characterises facies succession A and was discussed above. This facies is found predominantly in Innisfail (T35, R.1W5) and Davey (T33-34, R28W4). Two other log facies (facies 2b and 5) are present. Facies 2b was interpreted in chapter 4 as a gradual coarsening-upward shoreface sandstone. The third log facies represents open bay deposits (Figure 5.2, well 16-14-36-28W4). *Within allomember 1, these two facies are concentrated in the eastern and southeastern parts of the study area.* These two log facies are essential to the interpretation of the geometry and are discussed further with the cross-sections below.

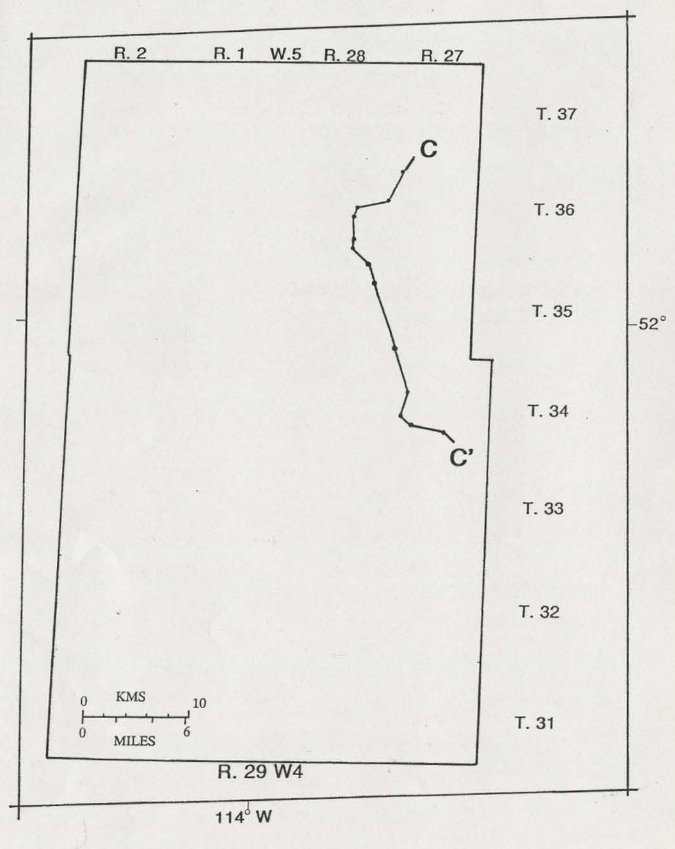
The geometry of allomember 1 is difficult to define in the study area, mainly because sediments of allomember 1 are cut-out by sediments of allomember 2. However, in the eastern portion of the field where one north-south cross-section was constructed, there was minimum incision, and log correlation can be used to establish the geometry of allomember 1.

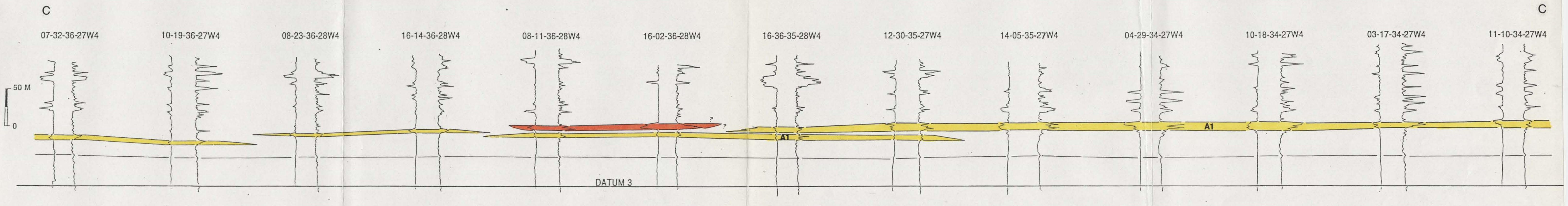
In cross-section C-C' (Figure 5.2), the shoreface of allomember 1 is represented by a series of discontinuous sandbodies, some with overlapping relationships. The overlapping geometry is illustrated in wells 16-36-35-28-W4 and 12-30-35-27-W4 where two shoreface successions are present. These sandstones are up to 10 m thick and thin out to the northwest and southeast. The first sandstone extends from 08-11-36-28W4 to 12-30-35-27-W4, a distance of 5 km. The second sandstone is separated from the first by a thin mudstone (as indicated by the lower resisitivity between the higher resisitivity sandstones), and extends for a longer distance than the first sandstone, a minimum of 20 km. It is not clear whether the second sandstone between 08-11-36-28W4 and 16-02-36-28W4 is a continuation of the second shoreface or represents a channel sandstone. The resistivity log response shows a crude fining-upward trend, despite the blocky SP, and is interpreted to be a channel sandstone.

The northernmost sandstone pinches out between 10-19-36-27W4 and 08-23-36-28W4, and is at a lower stratigraphic horizon with respect to the datums. This suggests that it is older than the two shorefaces farther south.

Between 08-23-36-28W4 and 08-11-36-28W4, the sandstone is thinner and discontinuous and could represent another thinner shoreface succession. However, the presence of thin higher resistivity beds in between these shoreface sandstones suggests that this log response represents facies 5, open bay sediments, and the sandstones between 08-

Figure 5.2: Log cross-section C-C' showing the discontinuous sandbody geometry of the shoreface of A1. The thin sandstone between 08-23-36-28W4 and 16-14-36-28W4 represents interlobate areas between the prograding shoreface sandstones. (see Table 5.1 for legend to colours used in the logs)


R. 2
R. 1
W.5
R. 28
R. 27
T. 37
T. 36
T. 35
T. 34
T. 33
T. 32
T. 31
52°
C
C'
0 KMS 10
0 MILES 6
R. 29 W4
114° W


C
C
07-32-36-27W4
10-19-36-27W4
08-23-36-28W4
16-14-36-28W4
08-11-36-28W4
16-02-36-28W4
16-36-35-28W4
12-30-35-27W4
14-05-35-27W4
04-29-34-27W4
10-18-34-27W4
03-17-34-27W4
11-10-34-27W4
50 M
0
A1
A1
DATUM 3

23-36-28W4 and 16-14-36-28W4 could represent localised coarsening-upward facies, infilling bay areas.

East-west sections also show the sandstones to be laterally discontinous, further emphasising the limited lateral extent of the shoreface sandstones of allomember 1. In section E-E' (Figure 5.3), the cored well 14-29-35-1W5 comprises an 11.5 m thick shoreface sandstone. A second E-W cross-section parallel to E-E' (section F-F'; Figure 5.4), shows a thinner shoreface sandstone (2 m thick), suggesting either localised infilling of an open bay area or the southeast thinning of the shoreface sandstone present in section E-E'.

In shallow marine nearshore environments, discontinous sand bodies may represent either barrier islands or delta front sandstones. The lateral discontinuity of the shoreface sandstones, with overlapping geometries suggest that these represent delta front bars rather than barriers. Barriers are commonly formed during transgressions, where delta front sandstones are reworked during shoreface retreat (Penland et al., 1988). Barrier islands are also associated with tidal inlets, and the sandstones are extensively bioturbated. There is no evidence to indicate any of these features within the sediments of allomember 1. The preserved sedimentary structures and the presence of *Macaronichnus* (which require high energy conditions), imply deposition in an actively prograding shoreface rather than reworking of an abandoned delta front succession. Therefore, the discontinous sandbody geometry may be formed by progradation of individual deltaic lobes. The overlapping geometry is of limited areal extent, and may be due to lateral shifting of the lobes rather than to changes in relative sea level.

Figure 5.3: East-West cross-section (E-E') showing the thickness and limited lateral extent of the shoreface sandstone of allomember 1. (see Table 5.1 for legend to colours)

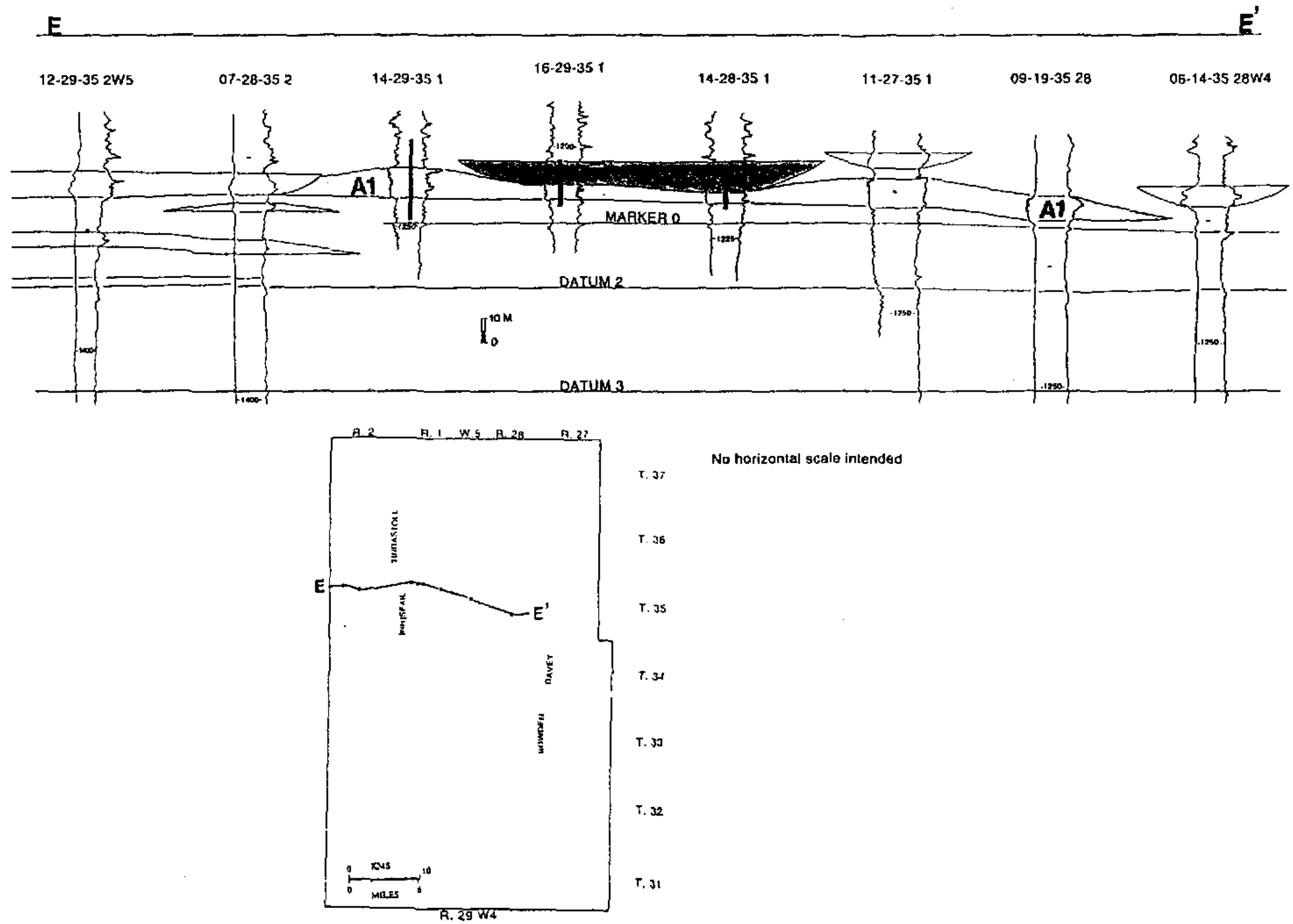
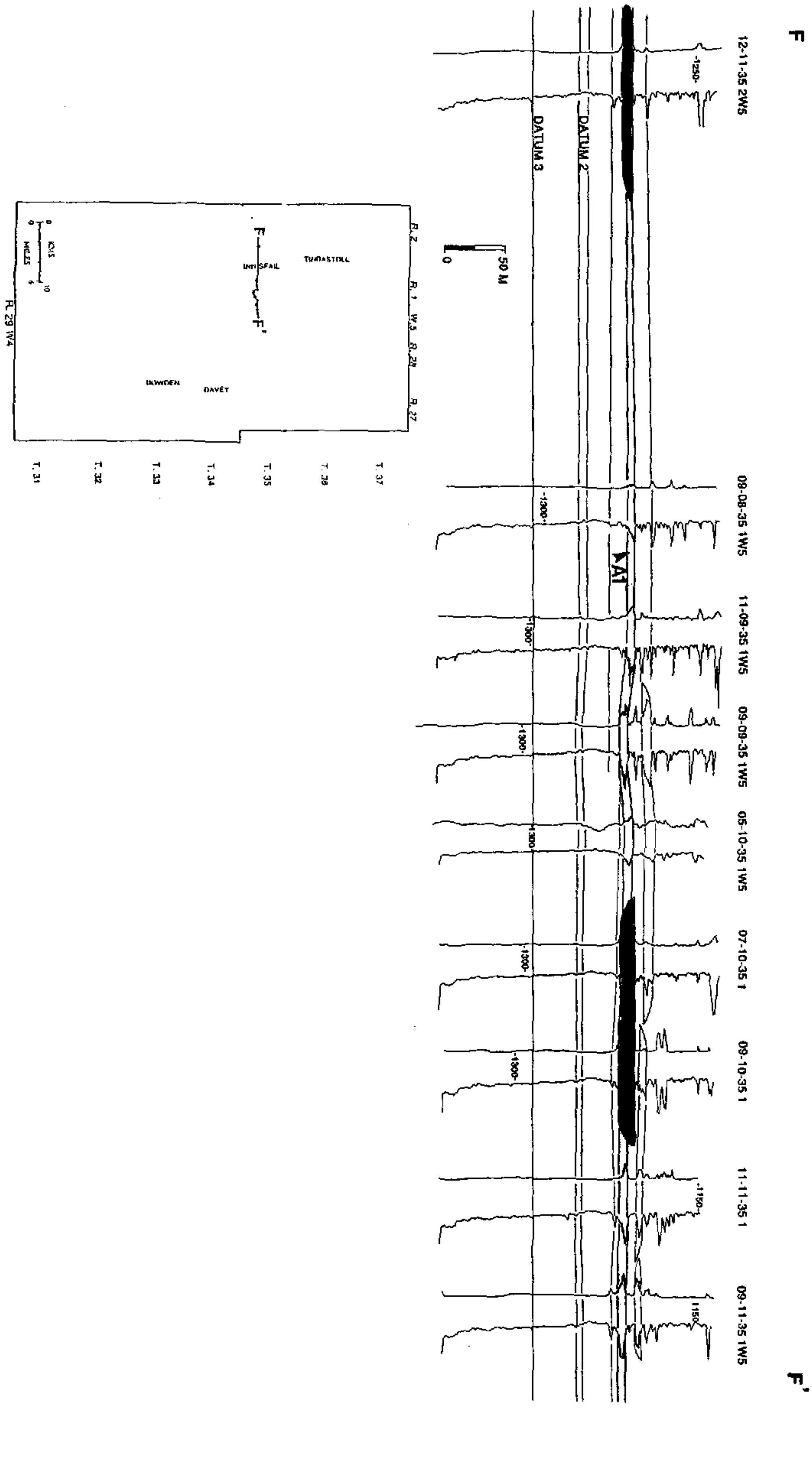

E
E'
12-29-35 2W5
07-28-35 2
14-29-35 1
16-29-35 1
14-28-35 1
11-27-35 1
09-19-35 28
06-14-35 28W4
A1
A1
MARKER 0
DATUM 2
DATUM 3
10 M
0
No horizontal scale intended
R. 2
R. 1
W 5
R. 28
R. 27
T. 37
T. 36
T. 35
T. 34
T. 33
T. 32
T. 31
E
E'
DAVEY
KMS
MILES
R. 29 W4

Figure 5.4: East-West cross-section (F-F') parallel to section E-E' where the shoreface sandstone is thinner than in section E-E'. (see Table 5.1 for legend to colours)



F
F'
12-11-35 2W5
09-08-35 1W5
11-09-35 1W5
09-09-35 1W5
05-10-35 1W5
07-10-35 1
09-10-35 1
11-11-35 1
09-11-35 1W5
DATUM 2
DATUM 3
A1
50 M
0
-1250-
-1300-
-1150-
INNISFAIL
TROCHU
BOWDEN
DAVEY
R.2
R.1 W.5
R.28
R.27
R.29 W4
T.31
T.32
T.33
T.34
T.35
T.36
T.37
KMS
MILES

Lobe switching may be due to upstream channel avulsion, in response to a gradient advantage. The open bay areas between the sandstone lobes can be described as interlobate; they progressively become non-marine upwards. These lobes may be similar to shingles of Bhattacharya and Walker (1991b) and parasequences of Van Wagoner et al. (1990).

Although the 3D geometry of the sandstones cannot be better defined, the presence of laterally discontinous sandstones associated with open bay sediments, and the fluvial-dominated sedimentary structures such as climbing ripples, loading features and an abundance of carbonaceous material in the marine succession favour a deltaic depositional environment.

5.2.3 Bounding discontinuity 1

Offshore interbedded mudstones and sandstones of the Lea Park Formation are abruptly overlain by thick, upper shoreface sandstones (with abundant intraclasts at the base of the sandstone) of allomember 1. This sharp contact between the sandstones and mudstones is therefore mapped as a regionally extensive RSE and defines the lower bounding surface (BD1) of allomember 1 (Figure 4.1, 5.1).

5.2.4 Shoreline response to relative sea level fall

The delta front sandstones of allomember 1 prograded over a sharp regressive surface of erosion (RSE), formed during a lowering of sea level. Thus lowering of fair-weather wave base resulted in wave scouring of offshore mudstones to form an erosional surface. Progradation of the shoreface was therefore stimulated by a drop of relative sea level drop rather than sediment flux variations; this progradation is termed a *forced regression* (Plint, 1988; Posamentier et al., 1990; 1992; Walker and Plint, 1992; Figure 5.5). Criteria for recognising falling stage forced regressions are outlined by Plint (1988) and Posamentier et al. (1992) based on examples from the Cardium and Viking Formations in the Western Interior. The distinguishing feature of forced regressions is the sharp-based shoreface lying on offshore mudstones. In the regressive Kakwa allomember of the Cardium Formation, the shoreface sandstones alternate between sharp-based and gradational across the entire width of the sandbody. However, in the Viking Formation at Joarcam, the sharp-based shoreface passes distally (seaward of the lowstand incision) into a normal prograding shoreface. This relationship is implied to be due to the finer grained nature of the Viking deposits and lower wave energy farther seaward (Posamentier et al.,1992). However, these distal shoreface sandstones may be gradationally based because they prograded into deeper water. Shoreface sandstones of the basal Belly River show similar features, with both sharp and gradational bases, but there are no apparent proximal-distal trends as in the Viking Formation at Joarcam.

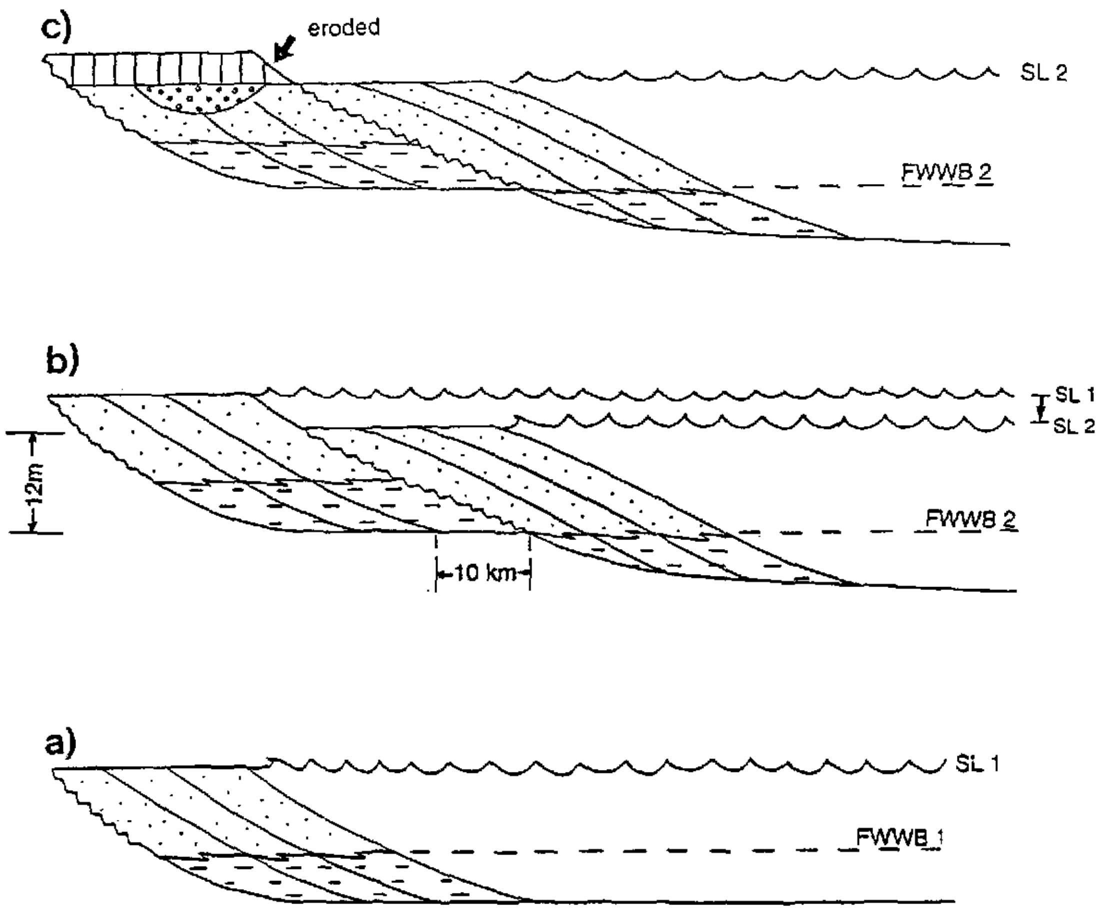
The geometry of the lowstand shoreface sandstones is also different from normal prograding shorefaces (Figure 5.5a). In the diagram, sharp-based shoreface successions show erosion of the offshore mudstones, and shaling-out and reappearance of thick sandstone successions, at a lower stratigraphic position basinwards (Figure 5.5b). Consequently, a wedge-shaped seaward stepping geometry can be produced (Tesson et al., 1990; Posamentier et al,1992). This seaward stepping geometry of the shoreface is due to successive lowering of sea level.

Figure 5.5 : Formation of a sharp-based shoreface (After Plint, 1988)

a) A normal prograding shoreface, with gradationally coarsening-upward from offshore mudstones to shoreface sandstones.

b) Formation of a sharp-based shoreface. Erosion of offshore mudstones due to lowering of relative sea level (from SL1 to SL 2) resulted in progradation of a sharp-based shoreface onto offshore marine mudstones. The resulting sandbody geometry steps seaward and is at a lower stratigraphic position than the gradational-based shoreface.

c) As progradation continues, channel incision into older and coeval shoreface successions may occur. Channel incision may be due to autocyclic controls or due to a further drop in relative sea level.


c)
eroded
SL 2
FWWB 2
b)
SL 1
SL 2
12m
FWWB 2
10 km
a)
SL 1
FWWB 1

The width of the sharp-based sandbody is generally related to the duration of stillstand after maximum sea level drop (Posamentier et al., 1992). In the examples studied by Posamentier et al. (1992), the lowstand shoreface deposits occur in a narrow belt, less than 12 km wide. Both sharp-based and gradational coarsening-upward shoreface successions are present in allomember 1 of Belly River Formation. The sharp-based shoreface sandstones extend downdip for about 10 km, but the gradational coarsening-upward shoreface succession extends downdip much farther. This suggests that progradation in the study area, after the sea level drop, was much farther and perhaps of longer duration, than observed in Joarcam Field. The different progradation rates may also be due to differences in sedimentation rates between the two formations. Calculated rates of sedimentation during the deposition of the Viking Formation are less than those of the Campanian Belly River Formation (Chamberlain et al., 1989).

In contrast, other examples in the Western Interior show sharp-based shorefaces that have prograded much farther basinward. The Santonian-Campanian Virgelle Member of the Milk River Formation (McCrory and Walker, 1986) and the laterally equivalent Chungo Member of the Wapiabi Formation (Rosenthal and Walker, 1987) prograde as sharp-based shorefaces for distances of at least 100 km. The Kakwa allomember of the Cardium Formation prograded up to 200 km (Plint, 1988).

Updip, the shoreface sandstones may be subjected to fluvial channel incision (Figure 5.5c), which may be coeval with the falling-stage prograding shoreface (Tesson et al., 1990) or they may be incising into older shorefaces. In the Late Quaternary lowstand deltaic wedges on the Rhone continental shelf (Tesson et al., 1990), distributary channel incision is localised and cuts the seaward stepping shoreface sandstones. The nature of fluvial incision into shoreface sandstones of allomember 1 are discussed in section 5.3 below.

Although the geometry of the shoreface of allomember 1 is not well-defined, sandstones shale out and reappear at slightly different stratigraphic levels. The sandbodies are localised, with overlapping relationships (Figure 5.2). Their small areal extent is more likely to be the result of autocyclic processes of delta lobe switching rather than seaward stepping shorefaces formed as a result of relative sea level drop. Therefore the gradual coarsening-upward profiles seen to the east and south-east of the study area reflect autocyclic processes rather than the lateral equivalent seaward sediments of the lowstand incision. Autocyclic processes such as delta lobe switching due to abandonment of a distributary channel could result in progradation of a new lobe with a normal gradual coarsening-upward succession. In this study area, fluvial incision is extensive and may be attributed to the overall regressive nature of the sedimentation during Belly River time. The major differences between other well studied examples of the Western Interior and the basal Belly River Formation lies in the facies relationships. The shoreface sandstones of the Cardium and Viking Formations are encased in offshore shales (except for the Kakwa allomember; Plint, 1988) and lack fluvial incision. Moreover, the shoreface successions of Cardium Formation and some of the Viking Formation are transgressive, unlike the shoreface of this study area, which is regressive.

5.2.5 Comparison with other studies

In outcrop (T2 R7-8 W4), Ogunyomi and Hills (1977) interpreted the transition from Lea Park to Belly River Formations as a series of five depositional cycles, each cycle showing a coarsening-upward succession from offshore to barrier island to lagoonal sediments (which are rarely preserved). Along the foothills, the basal Belly River sandstone is storm-dominated with swaley cross-stratification (Rosenthal and Walker, 1987).

The base of the Belly River Formation is sharp in other localities both in outcrop and in the subsurface of Alberta (B.A.Power, pers. comm. 1992; Rosenthal and Walker, 1987; Ogunyomi and Hills, 1977). However, Doig (1986), suggested that the contact was gradational in this study area. Iwuagwu and Lerbekmo (1984) found the contact between the Lea Park and basal Belly River sandstone to be gradational in the Ferrier area (T40, R8-9W5). However, Iwuagwu and Lerbekmo (1984) worked with limited subsurface data and on a much more regional study of the Ferrier area, B.A.Power (pers. comm., 1992) found that the contact to be sharp-based.

Some workers have implied a deltaic model, but also lack sufficient information regarding the 3D sandbody geometry to prove their case (Doig, 1986; Iwuagwu and Lerbekmo, 1984). McLean (1971) worked on a regional scale from central Alberta to Saskatchewan. He noted that individual deltas geometries were difficult to recognise "due to overlap, erosion, and truncation of older strata...".

In summary, the evolution of allomember 1 is associated with a relative sea level fall, represented by sediments of a falling stage shoreface prograding over a regressive surface of erosion (BD1). Although sediments of allomember 1 are cut out by sediments of allomember 2 and the 3D facies geometry cannot be determined, the evidence presented above suggests that the sediments of allomember 1 were deposited near a fluvial input and the shoreface sandstones represent delta front sandstones. The discontinous sandstone geometry with overlapping patterns is indicative of autocyclic processes of channel avulsion and delta lobe switching.

Comparison with other workers shows that the transition from marine to non-marine, as illustrated by the Belly River Formation, is represented by different environments in different localities. However, the erosive nature of the lower contact (RSE) in most localities suggests that shoreface progradation was due to a drop in relative sea level. This regressive surface of erosion is not a regionally isochronous surface.

5.3 Allomember 2

5.3.1 Facies successions

Sediments of allomember 2 consist predominantly of non-marine, channel sandstones of facies 3 (chapter 4). Three main facies successions are described (Figure 5.6);

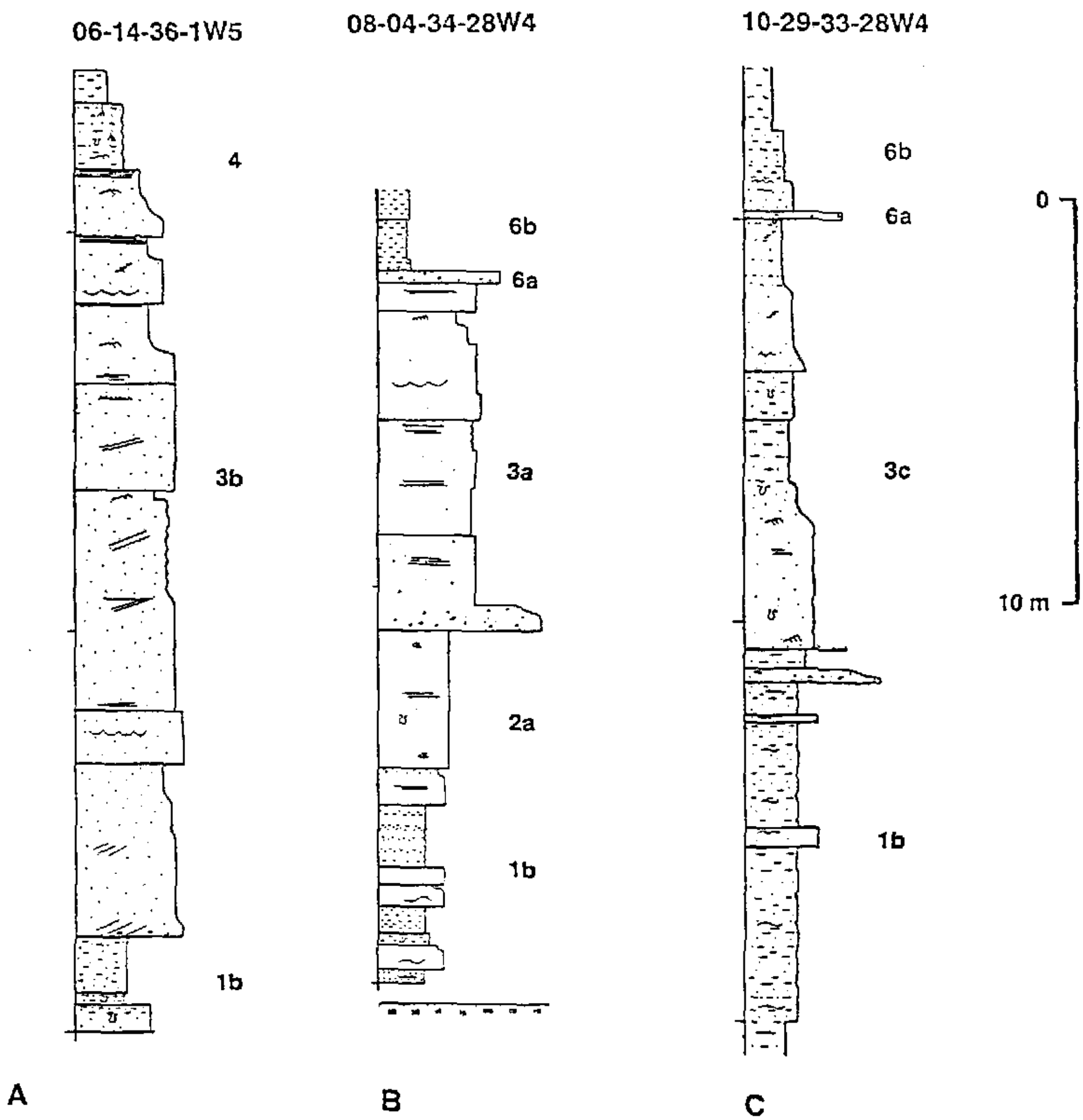
A) **1b : 3b : 4 : 2a**

B) **1b : 2a : 3a : 6a**

C) **1b : 3c : 6a**

Figure 5.6: Facies successions of allomember 2.

(see Table 5.1 for legend to symbols)


06-14-36-1W5
08-04-34-28W4
10-29-33-28W4
4
3b
1b
6b
6a
3a
2a
1b
6b
6a
3c
1b
0
10 m
A
B
C

The basal lag, the overall vertical fining-upward trend, the scarcity of marine fossils, an abundance of cross-stratification, and the sandbody geometry (discussed in detail below) suggest that these sediments are channel fills (chapter 4). Facies succession A (Figure 5.6) consists of fining-upward channel sandstones (Facies 3b), overlain by coal and fine grained deposits of the interdistributary areas (facies 4). This succession is predominantly found in the northwest of the study area, in Innisfail and Tindastoll Fields (T35-36, R1W5). The top of the channel fill in this succession has thin mudstone beds rarely more than a metre thick which represent gradual channel abandonment. These are overlain by overbank mudstones and thin coal beds (20-30 cm) of the interdistributary area (facies 4). Although laterally impersistent across fields, the coals provide excellent local markers for infield correlation.

In contrast, facies succession B occurs in the southeast, in Davey Field (T33-34, R27-28W4). This succession is characterised by the coarser grained channel fill of facies 3b (Figure 4.8) and a lack of fine grained deposits. The lack of floodplain fines in succession B could imply their subsequent removal during the ensuing transgression (allomember 3). However, the preservation of thick floodplain sediments in the north-west, and the scarcity of clay clasts within the transgressive lag suggest minor floodplain deposition in the southeast.

In general, the sedimentary structures of the channel sandstones of facies 3a and 3b indicate a predominance of unidirectional bedload transport. In a continuous core, there was no evidence to indicate reversing currents, such as tidal bundles or herringbone cross-stratification. Facies 3a displays repeated waning flow structures as sandstone beds grade upwards from massive into trough cross-bedded and ripple cross-stratified. The scattered beds of massive sandstone therefore indicate periods of strong current activity. The thickness of the channel sediments is comparable to the depth of most sandy-dominated distributary channels in modern deltas. Distributary channels in the Niger and Rhone deltas can be up to 25 m deep, and some channel sediments are multistorey indicating repeated cut and fill within the channel or reflecting minor fluctuations of the channel position (Elliott, 1986).

A third channel fill (facies succession C) is encountered in one well (Figure 5.6). This is predominantly mud-filled and represents an abandoned channel fill. The presence of a transgressive lag above the channel and the absence of marine trace fossils within the channel indicates that the channel was filled prior to the ensuing transgression that deposited sediments of allomember 3 (Chapter 6). This channel could not be mapped out in detail due to the lack of well control.

These three channels successions are interpreted as distributary channels because of the following lines of evidence;

- They contain marine to brackish water trace fossils, especially *Asterosoma*, which indicates periodic inundation by marine waters during times of low river stage.
- The delta-plain sediments laterally equivalent to the channels show evidence of brackish water influence, with thin, bioturbated (fine-grained) sandstones with *Arenicolites*.

5.3.2 Channel abandonment facies

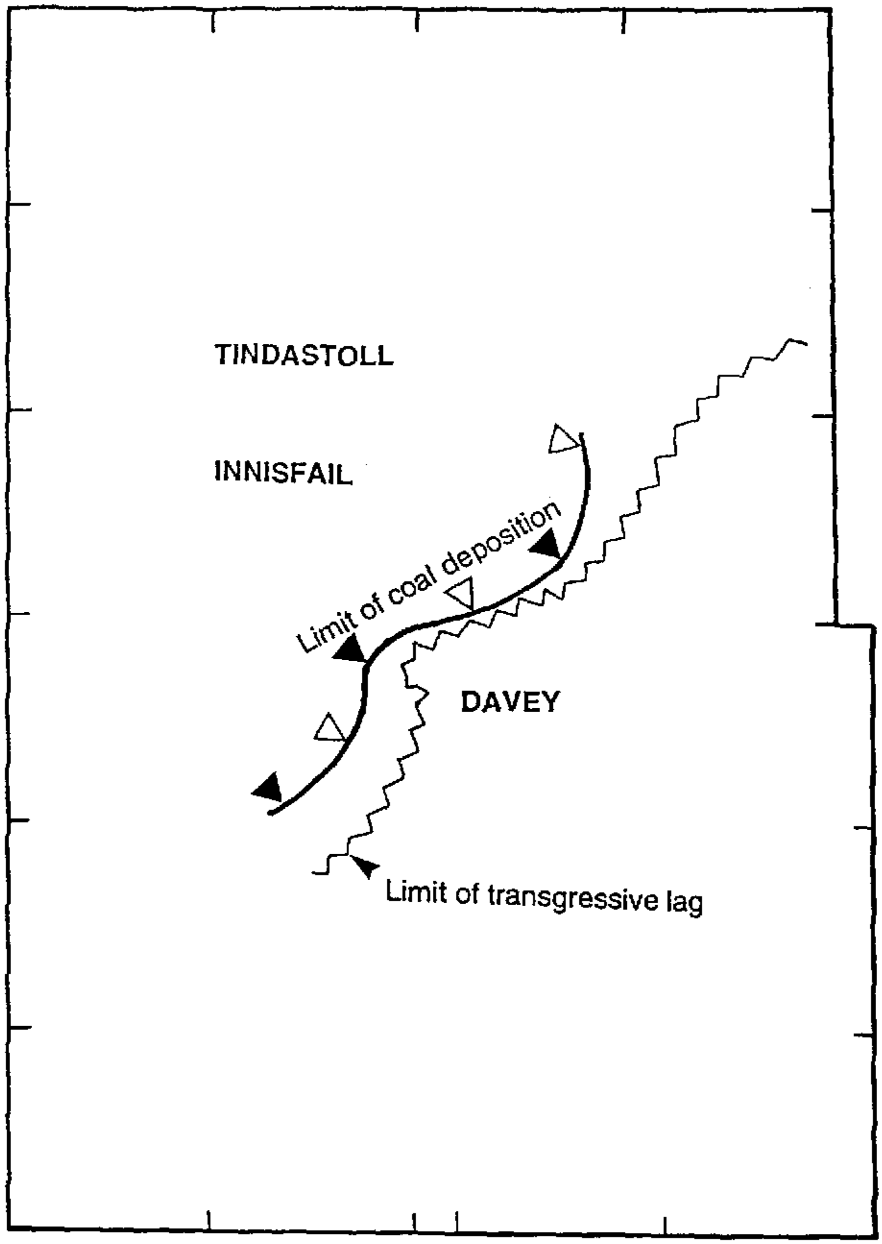
The abandonment facies of the channel types also varies. Facies 3a and 3c are sharply overlain by a thin transgressive lag (facies 6a). The composition of the lag is dominated by sand fractions with clay clasts but lacks plant debris, and indicates reworking of the underlying channel sediments. This lag is correlative with a thin, 20-30 cm coal bed in the north-west (Figure 2.3). Mapping of the limit of floodplain sediments and coal shows that the southern limit of floodplain deposits is coincident with the northern limit of the transgressive lag (Figure 5.7). The peat, which occurs as resistant flexible mats of intertwined plant material (McCabe, 1984), probably resisted erosion during the ensuing transgression. This further strengthens the suggestion that no interdistributary sediments (of comparable thickness to those in the northwest) were deposited in the southeast.

This relationship between floodplain and lag deposits has been described from the Yoredale Series by Elliott (1975), in order to determine the proximal and distal portions of a deltaic lobe. In this example, the distal abandonment facies is a thin localised fossiliferous horizon instead of a coarse pebbly lag, and coal forms the proximal facies. Therefore, the coal beds form the proximal abandonment facies of allomember 2 in the study area and indicate a palaeolandward direction to the northwest and west (Tindastoll Field).

5.3.3 Facies geometry

The geometry of sediments of allomember 2 can be described in two ways. Core and log cross-sections were constructed to determine the

Figure 5.7: The southern limit of coal and floodplain deposits is coincident with the northern limit of the transgressive lag overlying sediments of allomember 2 in the southeast.


TINDASTOLL
INNISFAIL
Limit of coal deposition
DAVEY
Limit of transgressive lag

extent of the channel incision and to establish the width of the channel belt. Cross-sections D-D' (Figure 2.2) and G-G' (Figure 5.8) are east-west cross-sections and section A-A' (Figure 5.9) is a N-S section. In the northwest (Figure 2.2), channels are isolated or form a belt which is less than 3 km wide. Channels are more laterally continuous in a N-S direction (section A-A', Figures 5.9) and extend for 10-15 km. In Davey Field, the width of the channel belt is greater, up to 10 km. To the east of T34, R28W4, complete shoreface successions are preserved as shown above (section C-C', Figure 5.2). Therefore channel incision within the study area is concentrated in a major NW-SE belt which crosses into Davey Field (Figure 5.10). Minor E-W fluvial trends are also evident but these could only be mapped as isolated single channels.

The lateral extent of individual channels is difficult to define in the subsurface. Doig (1986) correlated calcite-cemented zones within the sandstone of facies 3a for a lateral distance of 250 m suggesting channels are laterally continous for that distance.

5.3.4 Empirical methods of palaeochannel reconstruction

Determining individual channel dimensions using logs and cores is not realistic. The only independent variable that can be used in calculating palaeochannel parameters is the maximum channel depth, measured from the thickness of the in-channel facies. This can be used to estimate bankfull width, width-depth ratio, and sinuosity of the palaeofluvial system. The problem arises in that there is only one method (in the subsurface) that can be used to estimate width from depth; Leeder's (1973) regression equation.

Figure 5.8: East-West cross-section G-G' illustrating the depth of channel and extent of channel incision in allomember 2. The width of the channel belt in the south-east is about 10-20 km. The channels cut into shoreface sandstones of allomember 1 and into the Lea Park Formation.

(see Table 5.1 for legend to colours)

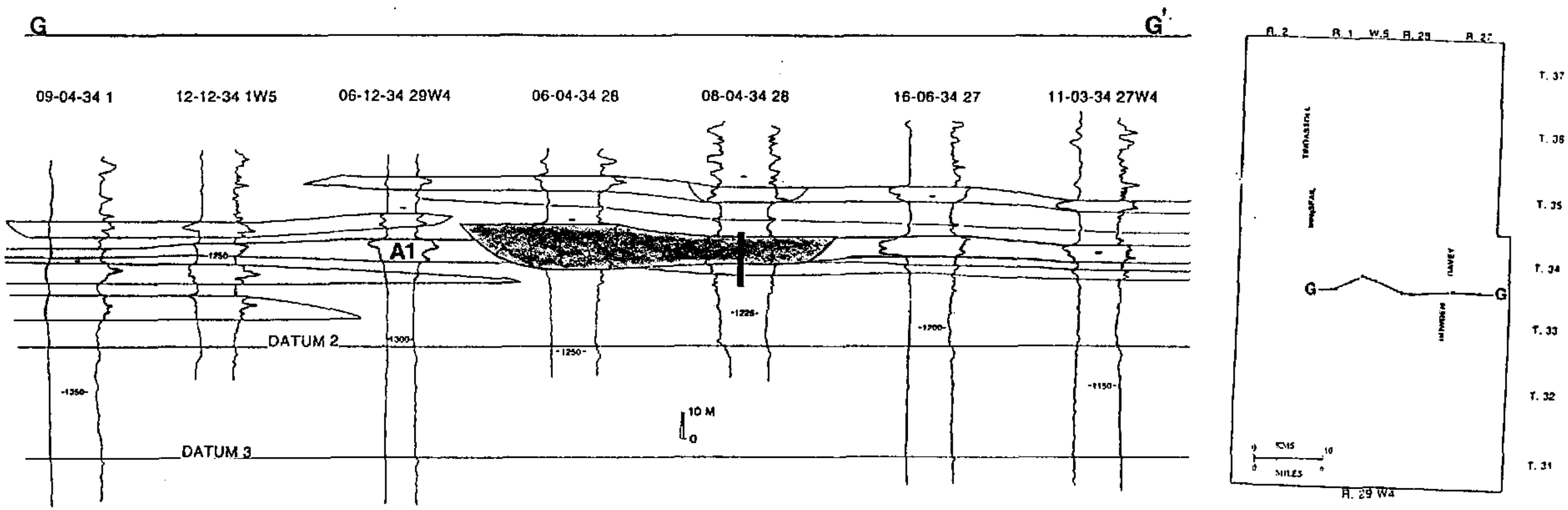
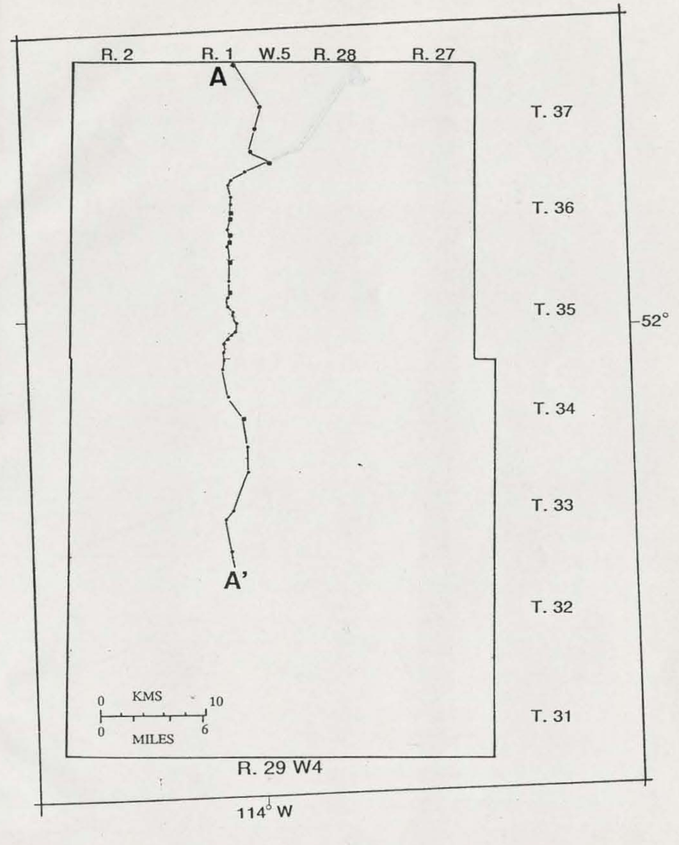
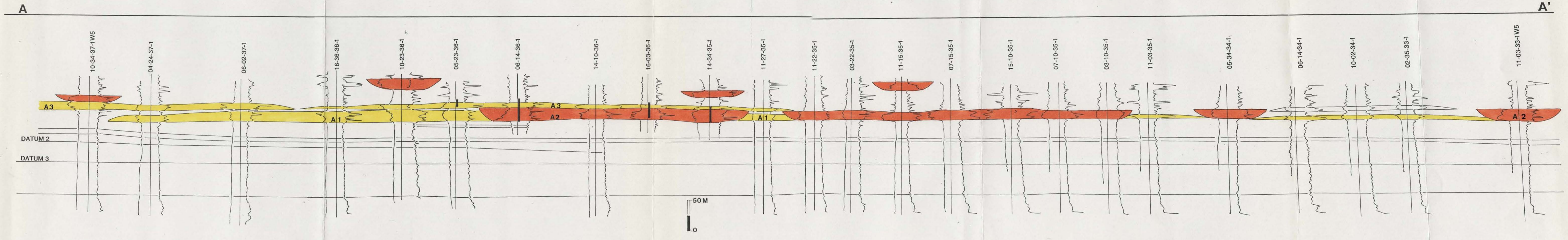

G
G'
09-04-34 1
12-12-34 1W5
06-12-34 29W4
06-04-34 28
08-04-34 28
16-06-34 27
11-03-34 27W4
A1
DATUM 2
DATUM 3
10 M
0
R. 2
R. 1
W.5
R. 29
T. 37
T. 36
T. 35
T. 34
T. 33
T. 32
T. 31
R. 29 W4
G
G

Figure 5.9: N-S log cross-section A-A', showing the regional extent of the channel incision.In this cross-section, the incision has removed all the shoreface succession of allomember 1 and the base of allomember 2 erosively on offshore mudstones of the Lea Park Formation. (see Table 5.1 for legend to colours)


R. 2
R. 1
W.5
R. 28
R. 27
A
T. 37
T. 36
T. 35
52°
T. 34
T. 33
A'
T. 32
0 KMS 10
0 MILES 6
T. 31
R. 29 W4
114° W


A
A'
10-34-37-1W5
04-24-37-1
06-02-37-1
16-36-36-1
10-23-36-1
05-23-36-1
06-14-36-1
14-10-36-1
16-03-36-1
14-34-35-1
11-27-35-1
11-22-35-1
03-22-35-1
11-15-35-1
07-15-35-1
15-10-35-1
07-10-35-1
03-10-35-1
11-03-35-1
05-34-34-1
06-14-34-1
10-02-34-1
02-35-33-1
11-03-33-1W5
A3
A1
A3
A2
A1
A2
DATUM 2
DATUM 3
50 M
0

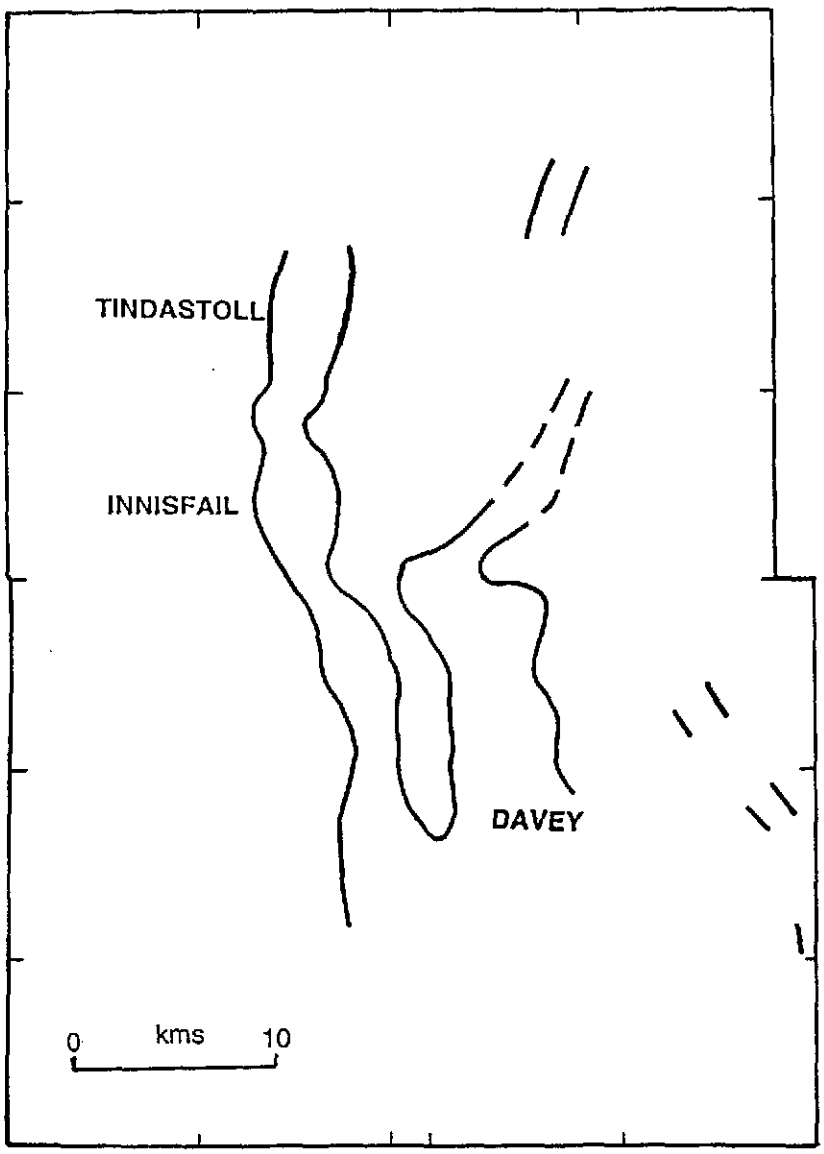


Figure 5.10: N-S trend of the channel belt. The channel belt in the northwest is narrower than in the southeast.

This use of this method assumes that the channel a) was meandering; b) had a sinuosity of 1.7 or more; and c) the maximum depth can be measured (Williams, 1988). Therefore, a prerequisite of this method requires an understanding of the channel type which may not be very obvious in the subsurface. Leeder's method cannot be applied in the basal Belly River because an assumption of sinuosity of 1.7 is needed, thereby defeating the whole purpose of the calculations which is to determine the sinuosity. All other methods of palaeowidth calculations (outlined by Ethridge and Schumm, 1978 Williams, 1984; 1988) require initial parameters (lateral accretion thickness, meander wavelength) that cannot be determined from the subsurface.

5.3.5 Fluvial systems of allomember 2

In the study area, the identification of fluvial systems as meandering or braiding is difficult because of the lack of three dimensional control necessary to identify the channel type. The limitations of quantitative methods of palaeochannel reconstruction have been described above. Therefore, qualitative methods are summarised (Table 5.2), and the channels are classified as either low or high sinuosity, avoiding using interpretative terms such as meandering or braided. A sinuosity greater than 1.7 separates high sinuosity streams from low sinuosity streams (Leeder, 1973). These are distinguished on the nature of the log response, the lateral extent of the channel belt and the amount of fine-grained in-channel and overbank deposits (Table 5.2).

Table 5.2: Summary description of the channels styles in the Tindastoll and Davey Fields.

Locality	Tindastoll and Innisfail Fields (NW)	Davey Field (SE)
Log response	*Fining-upward trend*	Blocky response
In-channel features	Fining-upwards; facies succession A (Figure 5.6)	Fining-upwards, coarsening upwards or blocky sandstones; facies succession B.
In-channel mudstones	Present at the top (up to 2 m thick) of channel.	not available
Overbank fines	up to 4 m thick, with thin coal beds	not available
Channel belt width	3 km or isolated single channels	10 km

a) fining-upward, high sinuosity channels

The general fining-upward nature of the channel fills is probably due either to lateral accretion of point bar deposition, or to vertical accretion during gradual channel abandonment. If the channel succession A (Figure 5.6) represents point bar deposits, these are very sandy and do not conform to the classic fining-upward point bar deposits formed by lateral accretion on the inside of a meander bend (Miall, 1992a, Figure 9). The best studied modern example of sandy point bar deposits is the Wabash River (Jackson, 1976).

In a meander loop, Jackson recognised three gradational zones which vary as functions of channel curvature, channel width and water stage. The three zones (transitional, fully developed and intermediate) form from the upstream of any meander bend to the downstream end. The classical point bar is the fully developed facies, which occurs after flow has changed its sense of rotation from one meander bend to the next. The thick sandy succession of facies succession A (Figure 5.6A) shows that these are more in line with Jackson's (1976) transitional depositional facies. The transitional facies of Jackson show less organised structures, and can show coarsening-upward trends. It occurs at the entry to a new meander bend, and forms under flow conditions inherited from the previous upstream bend. The facies of the transitional zone are therefore sandier than the fully developed facies, with a thinner succession of vertical aggrading fines. They are comparable with facies succession A of allomember 2.

b) low-sinuosity channels

The channels in the southeast are more laterally extensive than those to the northwest. The presence of coarse grained clasts recurring in the channel sandstones of facies succession B offers two possible scenarios. The lag may either represent repeated cut and fill within the channel, or indicate a variation in sediment load delivered within the channel perhaps during flood periods. The lateral extent of the channel belt in Davey would favour repeated cut and fill as channels migrate freely on the delta plain. It seems therefore that the channels in Davey Field were probably of lower sinuosity than those of Tindastoll Field to the northwest. These low sinuosity channels would favour lateral migration on the delta plain because no inter-channel deposits have been observed that may have confined the channels.

A second difference exists between the channels in the southeast and the northwest. In Davey, *Asterosoma* traces occur within the channels, but they were not observed in the northwest. This suggests that in the southeast, channels were inundated by marine water, perhaps during low river stage and suggests closer proximity to the river mouth than in the northwest. Therefore, the presence of coal in the inter-channel areas of the northwest, and the marine trace fossils in the channels of the southeast indicate proximal-distal relationships (with respect to a shoreline) between the two channel types. Furthermore, the blocky well log trends observed in some of the sandstones could be the result of introduction of sand during storms from the sea, which tends to suppress fining-upwards trends in distributary channels close to the delta mouth (Elliott, 1986).

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The reasons for channel switching in fluvial-dominated deltas can be due to gradient advantages as channels create shorter routes to sea, or as a result of an increase in sediment load and the sediment type. An example where gradient advantages may cause channel avulsion can be seen in the Gulf Coast. Today, the Atachafalaya River is diverting flow from the modern Mississippi River but is prevented from avulsion by man-made controls (Elliott, 1986).

The presence of a sandy sediment load with little fine-grained sediments would also favour laterally migrating channels instead of fixed channels. Such variations in distributary channel load can be observed in the Mississippi delta plain. The Modern Mississippi (Balize) delta is birdfoot in shape, and the position of the main distributary channel has been fixed for a relatively long time (Frazier, 1967, Bhattacharya and Walker, 1992). This partly due to the finer-grained nature of the bedload (carrying only 2% fine sand), the fine-grained nature of the underlying substrate, and the presence of vegetation. In contrast, the pre-Modern Lafourche delta was characterised by sandier bedload, prograding into shallow water. Channel switching was more rapid, resulting in lobate mouth bars and laterally, more continuous channel sandstones.

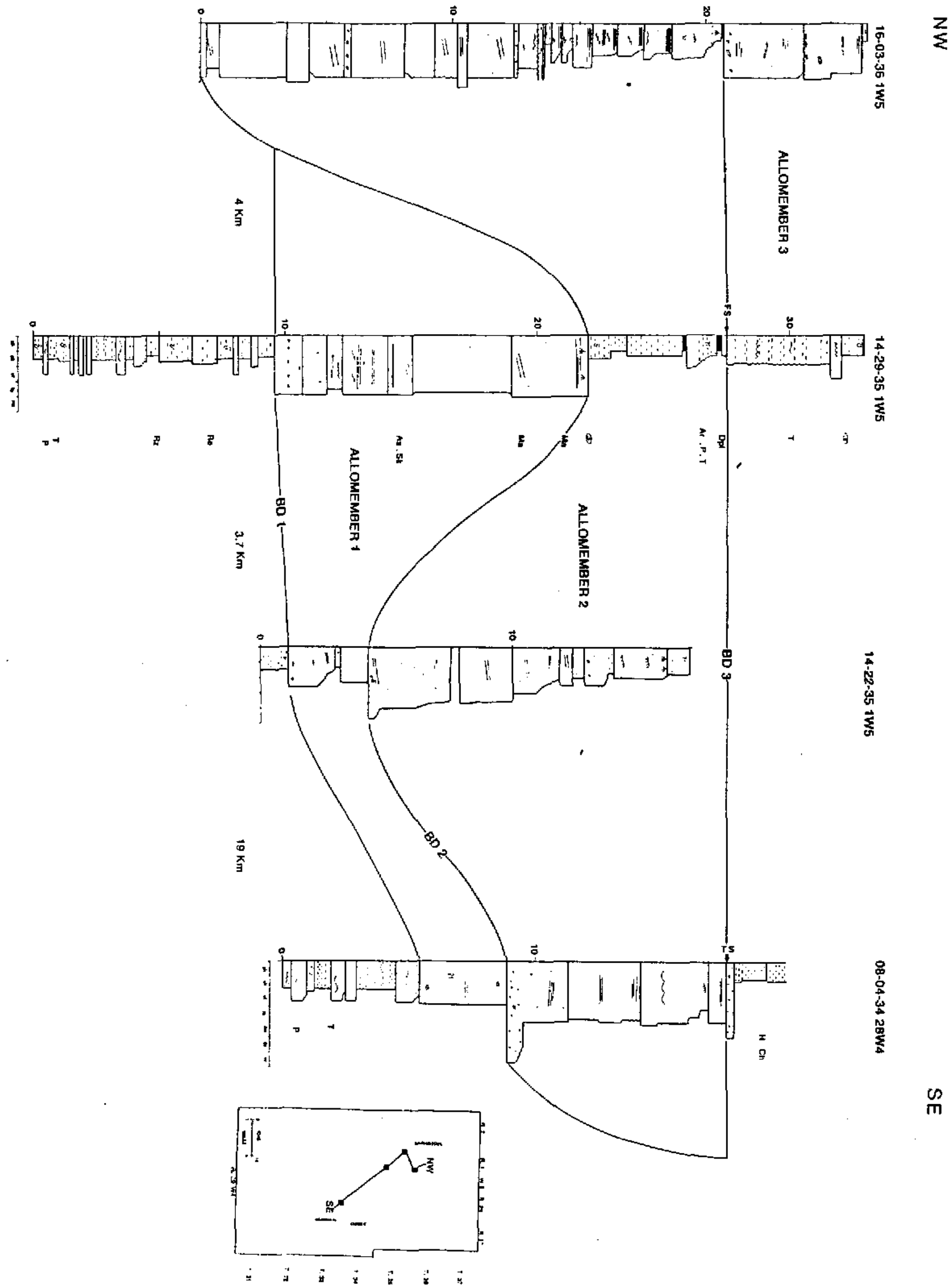
Some differences in channel styles due to differences in sediment type are also present in the basal Belly River Formation but are not of the extent of the Mississippi channels. Although both facies successions are sand dominated, the width of the channel belt in the west and northwest is not as extensive as in the southeast (Figure 5.10; Table 5). The lack of fine-grained facies in succession B (in the south) would favour the lateral migration of channels. In contrast, in the northwest where fine-grained sediments reach a thickness of up to 4 m and with peat deposition, the lateral switching of channels was probably restricted. The importance of peat in limiting the depth of erosion, and in this case, preventing the lateral migration of channels, is emphasised by McCabe (1984).

5.3.6 Origin of channel incision of allomember 2

The base of allomember 2 represents a channel incision which cuts down into the underlying facies of allomember 1 and Lea Park Formation and defines a second regressive surface of erosion (BD 2). This relationship is shown by log cross-sections (Figures 2.2, 5.8, 5.9), and one core cross-section (Figure 5.11). Channel incision removes all or part of the shoreface succession. Where no shoreface sandstone is preserved, the channels lie erosively on offshore mudstones (facies 1b) of the Lea Park Formation. In some wells, 2-4 m of the shoreface succession of allomember 1 may be preserved.

The regional extent of the incision is seen in section A-A' (Figure 5.9). The facies successions of allomember 1 consist of delta front sandstones with interlobe areas. It follows that in some wells (without core control), the channels of allomember 2 might have eroded not into sandstones, but into open bay deposits of the interlobe areas. Thus, depending on the depth of the incision, the channels may be underlain by brackish to non-marine sediments. The core cross-section (Figure 5.11) through Innisfail (T35, R1W5) shows an upward-fining channel (facies 3b; Well 06-28-35-1W5) at the same stratigraphic horizon as the shoreface succession of allomember 1 (well 14-29-35-1W5). Where the shoreface

Figure 5.11: Core cross-section across the three field illustrating the lateral relationship between channels of A2 and shoreface succession of A1. In 14-29-35-1W5, BD2 is the contact between the top of the shoreface sandstone and non-marine mudstones.


NW
SE
16-03-36 1W5
14-29-35 1W5
14-22-35 1W5
08-04-34 28W4
ALLOMEMBER 3
ALLOMEMBER 2
ALLOMEMBER 1
BD 1
BD 2
BD 3
FS
TS
4 Km
3.7 Km
19 Km

succession is preserved, interdistributary deposits above allomember 1 are coeval with the adjacent channel sediments. Also, bounding discontinuity 2 occurs at the contact between the top of the shoreface sandstone and the coaly mudstones of facies 4 (Figure 5.11).

a) channel incision and incised valleys

Channel incision may be due to autocyclic switching of the channel onto its coeval shoreface deposits (as in the Modern day Petit Rhone Delta, Elliott, 1986), or by allocyclic mechanisms associated with a drop in base level. The channel incision in the study area show;

- a basinward shift in facies suggests where channel sandstones lie erosively on offshore marine mudstones of the Lea Park Formation.

- the channel fill is largely non-marine, or show evidence of brackish water influence at the top of the facies succession.

A basinward shift in facies suggests that the incision may be due to a relative sea level fall. Posamentier et al. (1992) have argued that the nature of the fill or the degree of downcutting may not be due to relative sea level fall but that the juxtaposition of fluvial sandstones above offshore mudstones is most likely to be due to a relative sea level fall. These channels are not incised valleys in the sense of Van Wagoner et al., (1990) because incised valleys are commonly associated with two phases, an incision phase during relative sea level fall and a deposition phase in response to a relative rise in sea level, during the late lowstand or transgressive systems tract (Van Wagoner et al., 1990). The lack of core control of basinward facies coeval to the channels of the study area makes it difficult to determine whether the fill was deposited during transgression.

In an incised valley, during lowstand, sediments bypass the valley and are deposited as a lowstand wedge in the basin, and the subsequent valley fill is therefore commonly transgressive (Pattison, 1992). Consequently, most examples of incised valley fill comprise estuarine or marine deposits (van Wagoner et al., 1992; Pattison, 1992). The channel fill of allomember 2 is largely non-marine which would indicate the proximal part of the valley. In the study area, there is insufficient evidence to suggest that the channels represent incised valleys in the sense of van Wagoner et al. (1990).

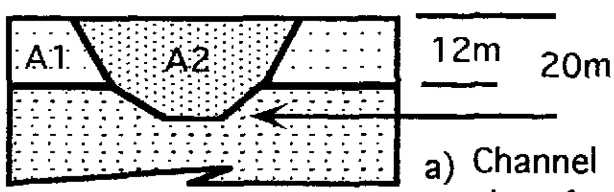
The extent of the channel incision outside the study area is not known. However, cores in T33 R26W4 (to the east of the study area) still show similar relationships to those in the study area.

Channel incision into underlying older shoreface successions is shown by Power (1989) in Pembina Field and was attributed to a relative sea level fall. What is the relationship between the channels of allomember 2 with the shoreface of allomember 1? Are the channels incising into older shoreface or do they incise into their coeval shoreface. Several possibilities exist which are discussed below.

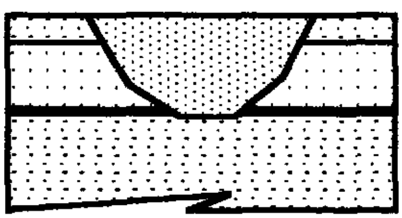
b) channel incision models

The channels of allomember 2 are up to 20 m thick, but the complete shoreface succession preserved in the study area is only 11.5 m thick. Because the channels cannot aggrade significantly above the top of the shoreface, the channels must therefore have incised 9.5 m below the shoreface into the Lea Park Formation, if the shoreface is coeval (Figure 5.12a). It is not possible to determine the depth of the incision into the Lea

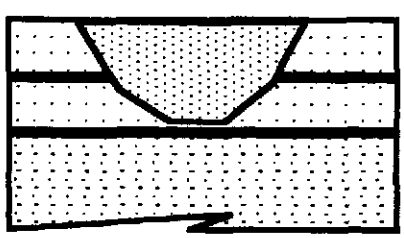
Figure 5.12: Models showing possible relationships between shoreface sandstone of allomember 1 and channel sandstones of allomember 2.



a) Channel incision into coeval shoreface with considerable erosion of the Lea Park Formation



b) floodplain aggradation before channel incision



c) incision into coeval and older shoreface successions

Park Formation because the channels do not intersect any of the markers used. Moreover, the channels with the greatest thickness are in Tindastoll Field and the well logs terminate a few metres below the sandstone, making it impossible to determine the extent of the incision into the Lea Park. However, in all the cores described, the channels did not incise deeply enough to intersect facies 1a which is present at least 8 m below facies 1b. Therefore, the incision must have eroded at least part of facies 1b.

If incision into the Lea Park Formation was minimum, then the shoreface must have aggraded up to 9.5 m of sediments prior to the incision (Figure 5.12b). The floodplain sediments above the shoreface sandstones are only 4 m thick, and appear to be laterally equivalent to the inter-channel sediments of allomember 2 (Figure 5.11). No break was observed in the floodplain facies that could indicate two different phases of sedimentation, one associated with shoreface and the other representing floodplain sediments of allomember 2. In contrast, the break in deposition occurs at the top of the shoreface, which is taken to be BD2.

These channels may possibly have cut into successive older shorefaces (Figure 5.12c). The channels may have eroded through more than one shoreface and now lie erosively on an older shoreface succession, having eroded all of their coeval shoreface succession. Alternative, the channels may be cutting into older shoreface and feeding a coeval shoreface further to the south. The incision would therefore be due to a further drop in relative sea level not associated with the shoreface that it is eroding into. Older shoreface successions are observed along the western edge of the study area. These shorefaces however, pinch out in the study area.

Therefore, there is no evidence to suggest incision into successively older shorefaces.

To summarise, the evolution of allomembers 1 and 2 is attributed to a two-fold drop in relative sea level. Sediments of allomember 1 represent a lowstand shoreface deposited during a forced regression. The depth of incision of the channels of allomember 2 suggest that the incision was probably related to a second phase of relative sea level fall. Thus the fluvial channels prograded seaward, eroding into an older shoreface succession and part of the Lea Park Formation, and were probably feeding a lowstand shoreface further south (out of the study area).

6. Sediments of Allomember 3

6.1 Introduction

This purpose of this chapter is three-fold. The first section (6.2.1 to 6.2.3) is primarily descriptive, outlining facies successions of allomember 3, their lateral facies relationships, the sandbody geometry and the bounding discontinuities. The lateral and vertical stacking of the facies succession can be interpreted as deposits of a deltaic system. The second part discusses the facies relationships and the sandbody geometry in view of existing deltaic models and by comparison with some modern deltas. The final section discusses the structureless sandstones commonly present in sediments of the basal Belly River Formation.

6.2 Allomember 3

6.2.1 Facies successions

Four facies successions form the sediments of allomember 3;

A. **2a : 5 : 4**

B(i). **5 : 2a: 4: 5 : 4**

B(ii) **5 : 2a: 5**

C. **5: 4**

D. **6a : 6b : 2b**

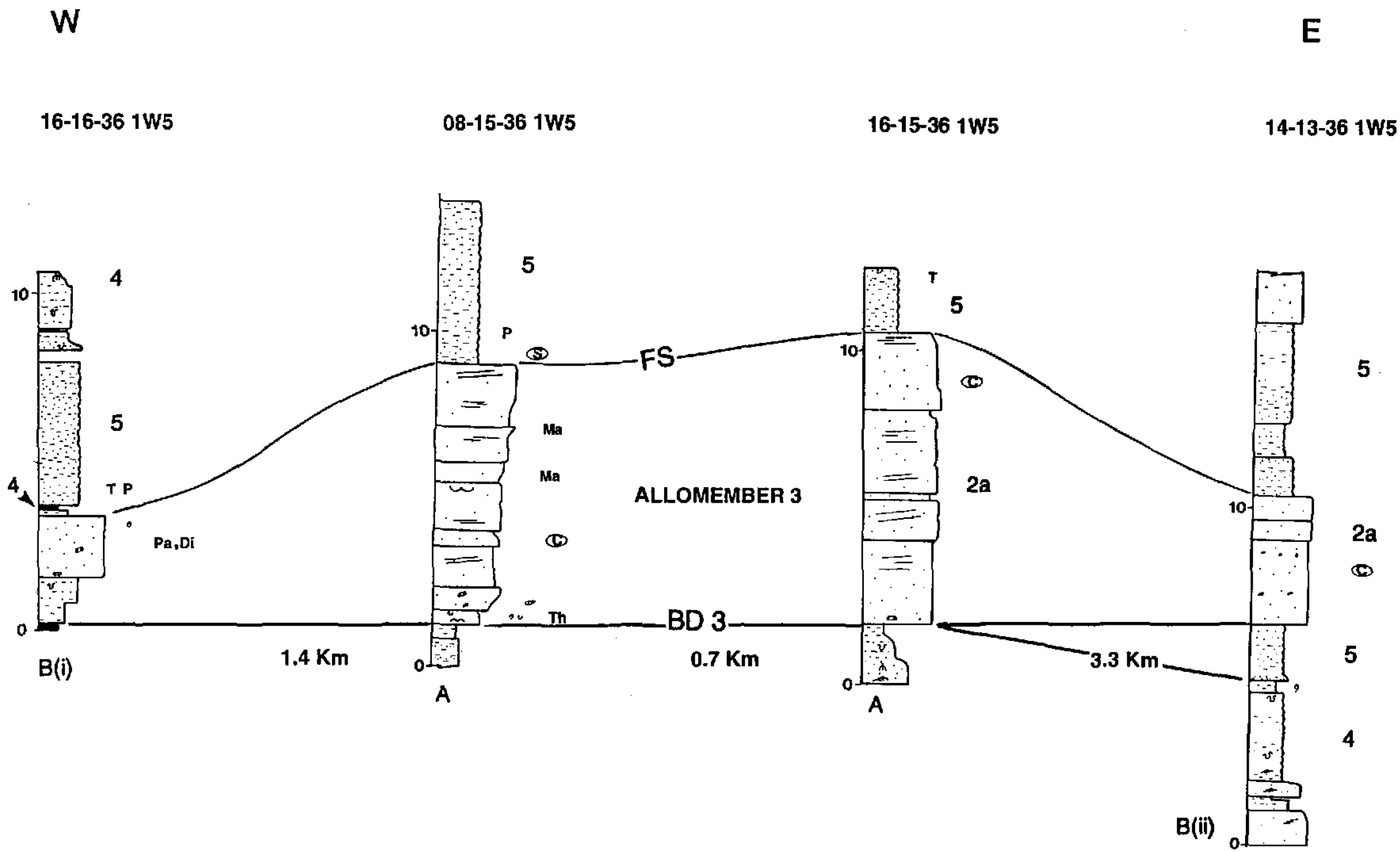
Facies succession A comprises shoreface sandstones of facies 2a sharply overlying coal or coaly mudstones (facies 4) of allomember 2 (Figure 6.1). The shoreface sandstone varies from 2 to 11 m thick. Abruptly overlying facies 2a are open bay sediments (Facies 5; Figure 6.1, for example, well 08-15-36-1W5).

Facies succession B is a variation of succession A and occurs along the western and eastern edges of Tindastoll Field (Figure 6.1; wells 16-16-36-1W5 and 14-13-36-1W5). Both facies successions (B(i) and B(ii)) have sharp contacts with the underlying sediments (of allomember 2), and the open bay sediments of 16-16-36-1W5 contain abundant disarticulated oyster shells. The difference between B(i) and B(ii) is the presence of facies 4 above the shoreface sandstone in B(i), observed along the western edge of the field.

The sharp-based shoreface has features which are not observed in the shoreface of allomember 1. The base of the shoreface of allomember 3 not only has mudstone clasts, but coalified plant fragments are abundant and *Macaronichnus* traces occur up to 4 m below the top of the sandstone. No roots were observed at the top of the shoreface sandstone, but in two wells, facies 4 (succession B(i),Figure 2.4) was deposited before deposition of facies 5. Within the shoreface sandstones, diffuse low-angle stratified and structureless beds are dominant, although some high angle cross-beds have been observed. Sandstone beds show both coarsening-upward and fining-upward trends.

Facies succession C is present in Innisfail (Figure 6.2a). It consists of

Figure 6.1: E-W core cross-section of facies successions of allomember 3 in the NW. Shoreface sandstones lie erosively (BD 3) on coal or coaly mudstone suggesting a *forced regression*. The sandstone thins out to the east and west. The *shoreface succession* is overlain by open bay sediments (*facies 5*), FS is a flooding surface, formed due to *autocyclic processes*.
(see Table 5.1 for legend to symbols)


W
E
16-16-36 1W5
08-15-36 1W5
16-15-36 1W5
14-13-36 1W5
FS
ALLOMEMBER 3
BD 3
1.4 Km
0.7 Km
3.3 Km
B(i)
A
A
B(ii)
T P
Pa,Di
Ma
Ma
P
Th
T
2a
2a

facies 5 sharply overlying facies 4 of allomember 2 and overlain gradational by facies 4. No shoreface sandstone was developed here.

Facies succession D consists of facies 6a overlain by facies 6b and by the gradual coarsening-upward shoreface succession of log facies 2b (Figure 6.2b).

The facies successions are dependent on location. Facies successions A and B occur in Tindastoll Field (T36, R1W5). In Innisfail Field and surrounding areas (T35 R28W4-1W5), facies succession C (Figure 6.2a) is present, and facies succession D (Figure 6.2b) occurs in the southeast, in Davey Field (T33-34 R28-27W4).

6.2.2 Sandbody geometry and lateral facies relationships

The geometry and lateral facies relationships are described using log and core correlations, and by isopach mapping of the shoreface sandstones. The 3D geometry of the shoreface sandstones (facies 2a and 2b) is best illustrated in an isopach map (Figure 6.3) of the total sand thickness. This map shows an irregular, lobe-shaped sandbody with two north-south trending thicks, one in Tindastoll (T36 R1W5) and one in Davey (T34 R28W4) Fields. These form individual sandbodies with elongate geometries of limited extent. The sandbody in Tindastoll is 6 km in wide and 12 km long. Erratic (less than 2 m) sandstones extend further eastwards forming the irregular lobate shape. In between the two fields, sandstone beds are thin and discontinous and mudstones and siltstone beds of facies 5 are predominant.

Figure 6.2: Facies succession of allomember 3 in Innisfail (C) and Davey (D) fields. In Innisfail, no shoreface developed, but open bay deposits are present. BD 3 is therefore a flooding surface (FS). In Davey, trangressive deposits are preserved and BD 3 is a TSE. The transgressive lag is overlain by offshore mudstones and gradational based shoreface succession (not cored).
These relationships and the sandbody geometry suggest that sediments of allomember 3 were deposited in a deltaic environment.

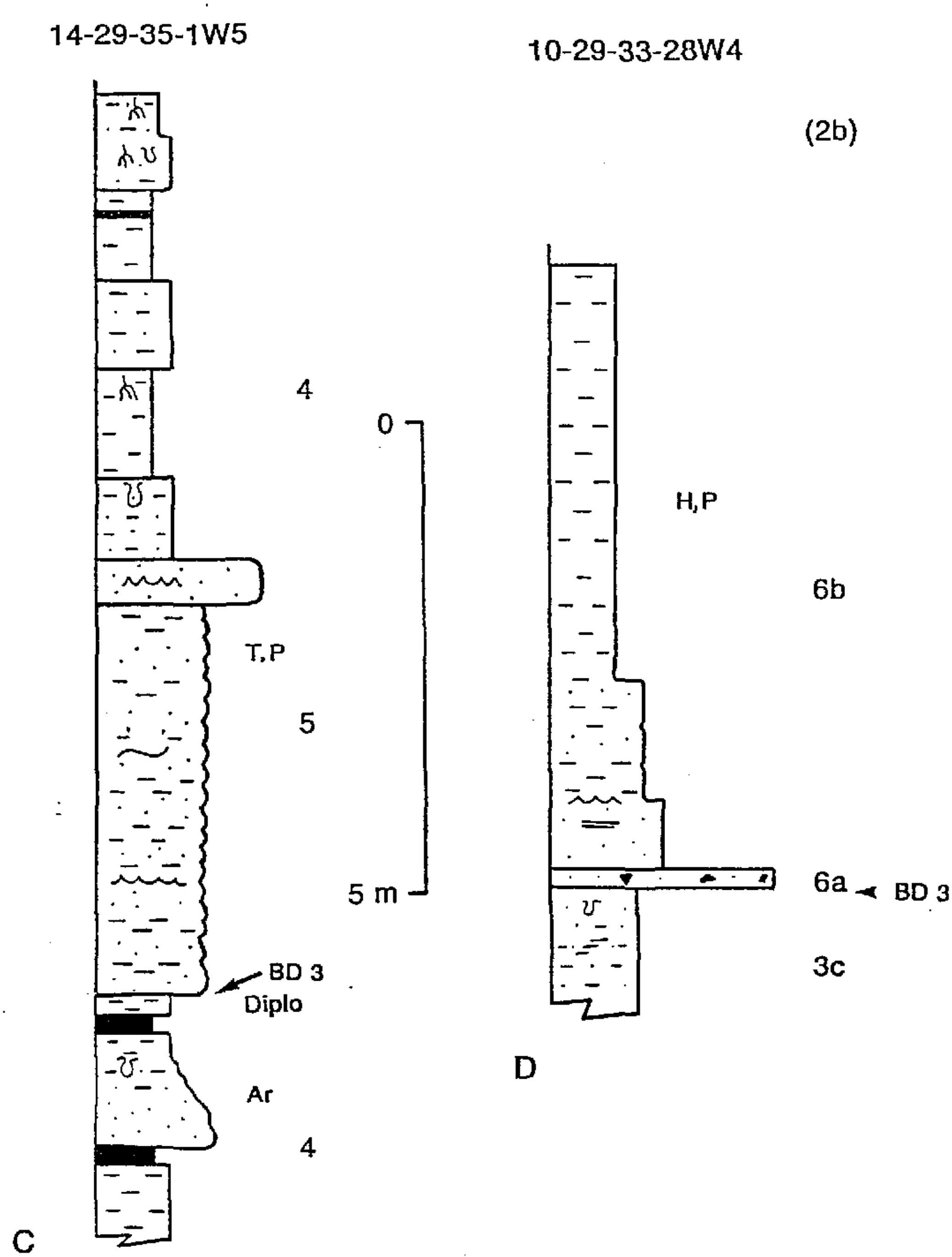
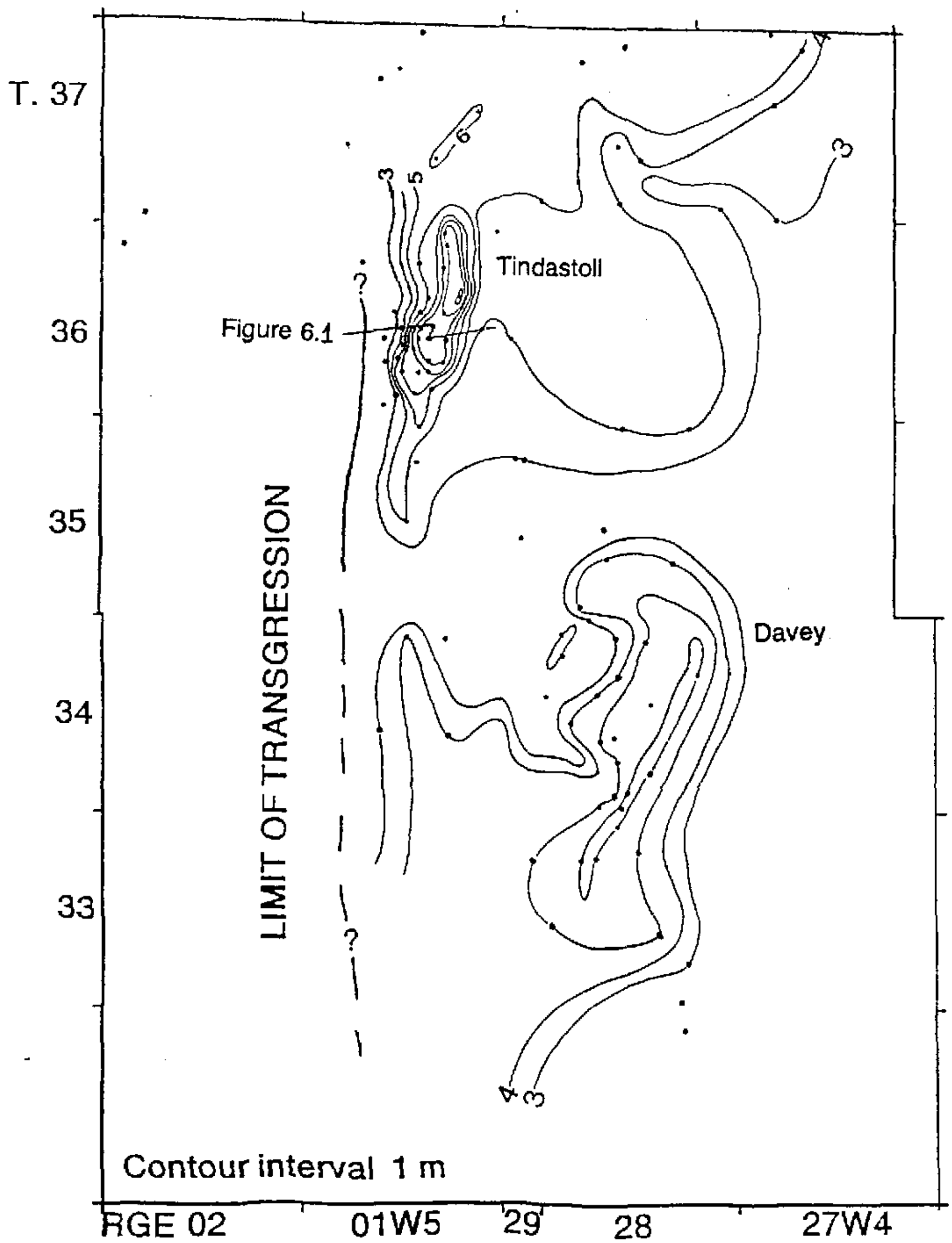

14-29-35-1W5
4
T,P
5
BD 3
Diplo
Ar
4
C
0
5 m
10-29-33-28W4
(2b)
H,P
6b
6a
BD 3
3c
D

Figure 6.3: Isopach of shoreface sandstone thickness. Two elongate sandstone thicks are mapped, separated by thin lobate sandbodies. The shoreface in the SE is at a slightly higher stratigraphic position than in the NW and progradation of the shoreface occured following the abandonment of the shoreface in the northwest. The geometry suggests a river- and wave-dominated deltaic system.



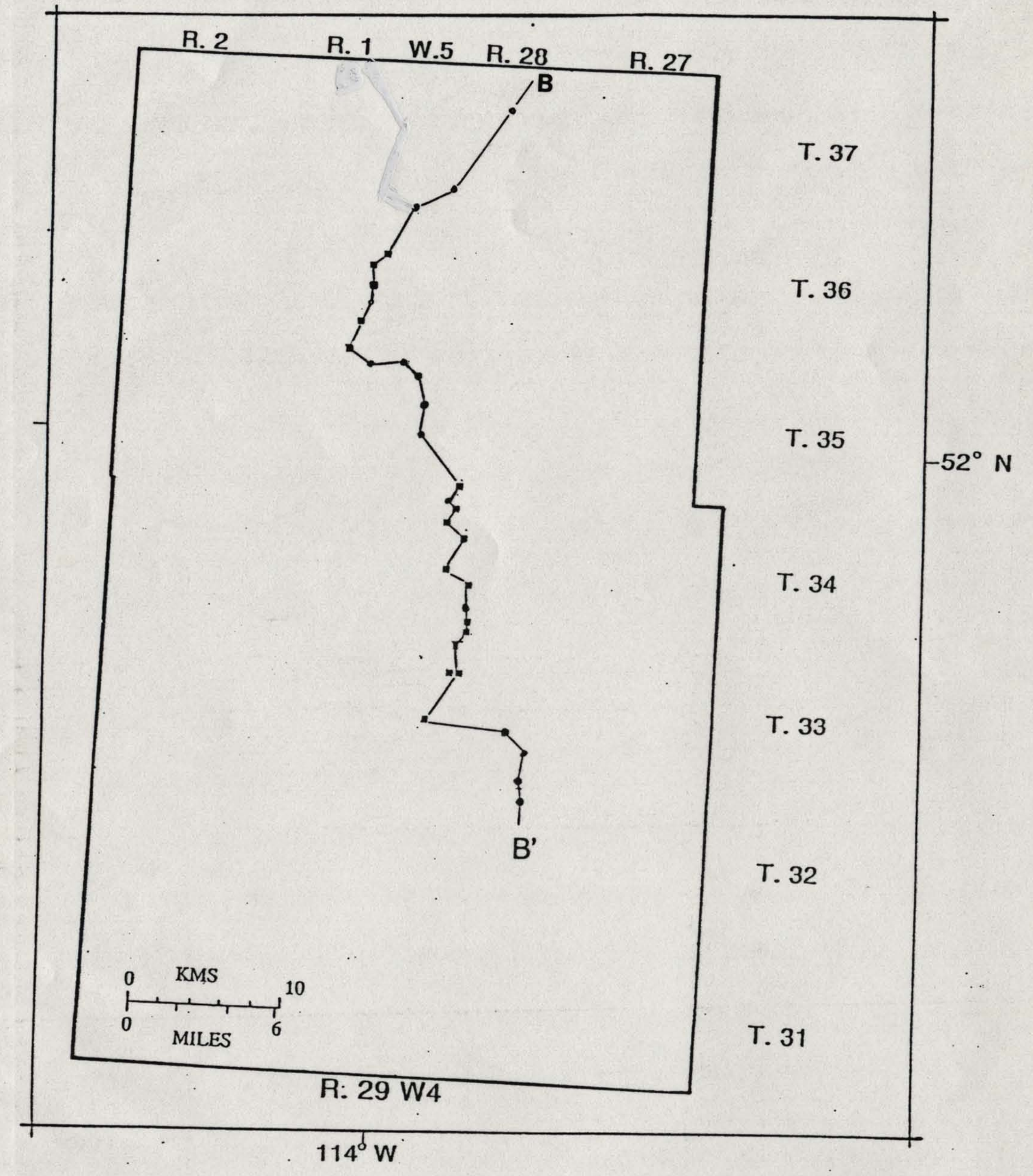
T. 37
36
35
34
33
Tindastoll
Davey
Figure 6.1
LIMIT OF TRANSGRESSION
Contour interval 1 m
RGE 02
01W5
29
28
27W4

The cross-sectional geometry can be established from the log and core correlations. In Figure 6.1 (core section), the core cross-section shows facies succession A in the north-west of the field (08-15-36-1W5) sharply overlying coal. This sandstone pinches out to the west and although the contact is not seen in 16-16-36-1W5 (Figure 6.1), the shoreface is only 2 m thick. The thickest portion, in the middle of the sandbody (Figure 6.1), is 11 m thick. Figure 2.2 is a west-east log and core cross-section which illustrates the width and pod-like cross-sectional geometry of the sandbody, with thinning eastwards. It also shows the extent of the transgressive open bay sediments before the progradation of the shoreface sandstone. Westward, facies 4 overlies the shoreface before being inundated by marine to brackish water sediments of facies 5.

The NW-SE core cross-section and corresponding log correlation (Figures 2.3 and 6.4 respectively) show the lateral relationships of facies 2a and 2b across the study area. In Figure 2.3, the shoreface sandstones of succession A are laterally equivalent to marine mudstones of facies succession D in the south-east. The mudstones are overlain by a gradual coarsening-upward sandstone (facies 2b) which is stratigraphically higher than the shoreface in the north-west. Therefore, the two lateral discontinous sandstones may represent two different prograding deltaic lobes of slightly different age. Their overall evolution is interpreted below. Thicker sandstones are present in the topmost portion of the study area and could represent another lobe.

In between Davey and Tindastoll (where the two shoreface sandstones are present), no shoreface sandstone was deposited; instead,

Figure 6.4: N-S log cross-section showing the two shoreface sandstones, at slightly different stratigraphic position and the extent of the sharp-based and gradational-based shoreface sandstones, and the trangressive mudstones. (see Table 5.1 for legend to colours)


R. 2
R. 1
W.5
R. 28
R. 27
B
B'
T. 37
T. 36
T. 35
T. 34
T. 33
T. 32
T. 31
52° N
0 KMS 10
0 MILES 6
R: 29 W4
114° W


B
B'
14-27-37-28W4
11-05-37-28
16-36-36-1W5
06-27-36-1
06-22-36-1
14-15-36-1
06-10-36-1
14-34-35-1
14-30-35-28W4
06-04-35-28
07-31-34-28
08-17-34-28
14-32-33-28
10-29-33-28
11-18-33-28
07-15-33-28
06-11-33-28
06-02-33-23
06-35-32-28W4
A1
A2
A3
MARKER 0
DATUM 2
20 M
0
No horizontal scale intended

the shoreface sandstones thin out and are replaced by open bay deposits (facies succession C; Innisfail Field). Therefore, these bay deposits may represent the interlobe areas in between two deltaic lobes.

The western limit of the transgression in the study area was mapped from log responses and using cores, and was established primarily on the extent of brackish or marine sediments lying above allomember 2 in cores along the western portion of the study area. In 16-14-33-1W5, the thin transgressive lag of facies 6a is overlain by 2m of open bay brackish-water deposits which are overlain by non-marine sediments. In the west, isolated channels form laterally equivalent facies. However, no core was available and the exact nature of the fill (although with a sandy lithology), is unknown. Therefore, this limit is an approximate one based on the above observations and it is possible that it may change once more work is done outside the study area.

6.2.3 Bounding discontinuities

The nature of bounding discontinuity 3 varies across the three fields. In Davey, a thin, 20-30 cm transgressive lag lies above channel sandstones of allomember 2. The contact between the two is interpreted as a transgressive surface of erosion (TSE), which forms the lower bounding discontinuity of allomember 3 in the southeast. The NW limit of this lag was mapped earlier and found to coincide with the south-eastern limit of coal deposition (Figure 5.7).

In Tindastoll, the bounding surface is the contact between the top of the coal bed and the overlying shoreface sandstones. Sandstone-filled burrows are present within the coal (Figure 4.1a). However, the burrows could not be identified as part of the *Teredolites* ichnofacies because no cross-cutting relationships were evident. This may be due to limitations of the core. The lower bounding discontinuity surface in Tindastoll is therefore a coplanar erosional surface of marine transgression and subsequent regression.

In Innisfail (between the progradational shoreface sandstones), fine-grained mudstones and interbedded sandstones (interpreted as interlobe deposits) lie above the coal. These are laterally equivalent to the shoreface sandstones above the coal bed in Tindastoll and the transgressive lag and offshore mudstones in Davey. The base of the open bay sediments is also sharp and a burrow of *Diplocraterion*, characteristic of the *Glossifungites* ichnofacies, cuts down into the coaly mudstones of facies 4 (Figure 6.2; succession C). Therefore, this contact can be correlated with confidence to the transgressive surface farther south and the open bay deposits can be interpreted as transgressive.

The presence of the coal bed in the north-west, which is laterally correlative with the transgressive lag found in the east and southeast, may imply that the transgression ended where the coal and lag meet (Figure 5.7). If that were the case, non-marine facies must then overlie the coal bed. However, shoreface sandstones sharply overlie the coal bed and marine inundation must have extended farther northwards, beyond the limit of the transgressive lag. Therefore, in the northwest, the peat probably prevented erosion of the underlying sediments during transgression. The shoreface sandstones include abundant coalified plant debris and clay clasts at the base suggesting that the transgression extended into Tindastoll Field.

A second bounding discontinuity is recognised in Tindastoll Field, occuring at the contact between the top of the shoreface sandstones and open bay sediments. This contact marks a marine flooding surface, which is defined as a depositional discontinuity (Pemberton et al, 1992b). This surface is of limited areal extent and is interpreted as a localised discontinuity surface formed by autocyclic processes and is discussed in detail below.

6.2.4 Interpretation

The lobate geometry, with elongate (discontinous) sandstones, at slightly different stratigraphic positions; the lateral juxtaposition of open bay deposits and shoreface sandstones; and the vertical facies successions suggest that sediments of A3 were deposited in a deltaic environment.

These sediments of A3 were deposited in a trangressive and regressive setting. The transgressive component is preserved in Davey and Innisfail and is represented by facies 5 and 6. In the southeast, the lag (facies 6a) overlying channel sandstones of allomember 2 was formed by wave reworking of the underlying sediments during the ensuing transgression. The lag is overlain by offshore mudstones which indicate a gradual deepening following trangression. Initially, waves (probably during storms) were able to erode some of the lag deposits and in the basal metre of facies 6b, thin layers of coarse sandstones are present within the mudstones (Gritty mudstones, Figure 4.15a). These mudstones, up to 18 m thick, are overlain gradationally by log facies 2b (representing the regressive component). In deltaic environments, the transgression may be autocyclically or allocyclically controlled. The regional extent of facies 6a suggest that this surface is not associated with the channel sandstones underneath and is unlikely to be formed by autocyclic processes. Therefore, the transgression is due to a relative sea level rise which may be due to climatic and/or tectonic changes in the hinterland, or eustasy.

Laterally equivalent facies successions in the northwest are different. In Innisfail, marine to brackish water sediments were deposited. These become increasingly non-marine upwards, representing gradual infilling of interlobe areas of the delta. In Tindastoll, no lag or marine mudstones were observed. Instead, a shoreface sandstone lies erosively on a thin coal bed of allomember 2 (except along the eastern margin, Figure 6.1). This sharp-based shoreface probably prograded as a result of *forced regression*. The shoreface lies sharply above non-marine delta plain deposits indicating a landward translation of the shoreline. This shoreface prograded eastwards a distance of 6 km producing an elongate sandstone geometry. In Davey, offshore marine mudstones were being deposited. As progradation rates declined, wave reworking modified the geometry, making it more lobe-shaped. These changes in geometry as progradation continues are similar to examples from the Mississippi delta and in the Lower Wilcox Group (Early Eocene) of the Gulf Coast (Fisher et al., 1969; Galloway, 1975).

The processes acting on the delta front deposits of the basal Belly River Formation can be interpreted from the sedimentary structures and trace fossils present within the sandstone. The presence of *Macaronichnus* in the middle of the shoreface sandstone indicates localised wave action.

The abundance of carbonaceous material also suggests fluvial input into the delta front. The sandstone is dominated by low-angle stratification with abundant carbonaceous material and structureless sandstones, although some ripple stratification has been observed. The sedimentary structures and trace fossils indicate the dominance of both fluvial and wave processes, similar to sediments of A1.

Isopachs of sand thickness (Figure 6.3) are tightly spaced in the northwest and spread to the south, suggesting gradual sandstone thinning to the south, perhaps due to southerly transport by longshore drift. This transport direction is in agreement with measured palaeocurrent directions in outcrops of the basal Belly River Formation (Taylor and Walker, 1984). The structureless sandstones are discussed in detail below.

In Tindastoll, the shoreface sandstones are sharply overlain by facies 5 and are interpreted as the abandonment facies of the deltaic lobe. These are marine at the base but get progressively non-marine upward (with root traces and the development of thin coal beds), suggesting gradual infilling of the bay following abandonment of the lobe.

Along the western egdes of the shoreface, facies 5 overlies facies 4. These non-marine sediments pinch-out eastwards (Figure 2.2) and comprise a thin coal bed and coarsening-upward sandstones. These represent infilling of isolated swampy areas of the lower delta plain. Fisher et al. (1969) have noted that progradation is commonly faster than the abandonment phase of a delta. Therefore, if certain areas of the delta plain are above base level and sufficiently removed from basinal processes, then depending on the rate of subsidence, different parts of the delta plain will be inundated by marine waters at different times. Hence in the raised areas of the delta plain, non-marine sediments would continue to aggrade, with input from fluvial processes such as overbank flooding. Work done on the Fraser Delta (Williams and Roberts, 1988) indicate that delta plain sediments can be 1-2 m above sea level and peat bogs are raised 3-4 m above sea level. Using peat accumulation rates of 1.6 and 2.3 mm/yr for the Mississippi and raised swamps in S.E.Asia respectively, and assuming a peat to coal compaction ratio of 10 to 1 (McCabe, 1984) , the peat required to form the 20 cm thick coal bed would take between 900-1250 years to accumulate. This places a constraint on how long it would have taken to completely inundate the delta lobe assuming the that the peat were being deposited in a raised area.

The causes of lobe abandonment can be allocyclic or autocyclic. Abandonment of deltaic lobes may be due to a decrease in sediment supply in the source area (hinterland). Therefore, sediment supply cannot keep pace with subsidence and localised transgressions occur. There is no evidence to suggest a decrease in supply in the source area. Furthermore, the localised extent of the lobe and the abandonment facies (open bay deposits), and the gradual coarsening-upward nature of the second shoreface rule out allocyclic controls.

Lobe abandonment is more likely to be due to upstream channel avulsion. After an avulsion, lobe progradation slowed and subsidence resulted in localised transgressions and the sediments of facies 5 were deposited. This resulted in the formation of a marine flooding surface above the delta front sands. This marine flooding surface separating facies 2 (Figure 6.1) and 5 represents the abandonment of the lobe in the northwest and progradation of the second shoreface in Davey (Figure 2.3; 6.3). This shoreface is therefore slightly younger than the shoreface in Tindastoll and the gradual coarsening-upward from transgressive mudstones of facies 6b suggests progradation of the shoreface associated with autocyclic lobe switching due to upstream channel avulsion.

6.3 Discussion

The description and interpretation of deposits of allomember 3 indicate that they are deltaic. This section will review the existing delta models and discuss whether any of the models can be applied to the deltaic sediments of the basal Belly River. The sediments are compared with the Quaternary and Modern deltas.

6.3.1 Delta models

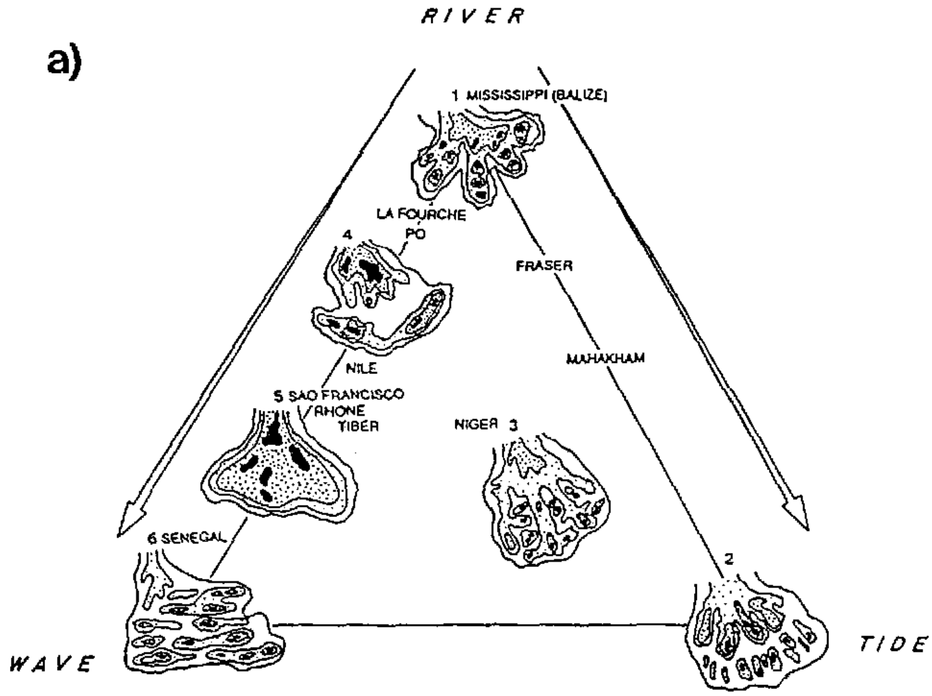
Deltaic sediments form as a result of a complex interplay between basinal processes, basin subsidence, sediment supply, climate and tectonics in the hinterland (Wright and Coleman, 1975). However, existing deltaic models are process-related, based on the morphology of the delta front as a function of the relative importance of fluvial and basinal processes. Galloway (1975) established a tripartite subdivision of deltas based on the ratio between basinal processes (tidal and wave processes) and fluvial processes (Figure 6.5a). Fisher et al. (1969) recognised two end-members of delta types; the fluvial dominated, highly constructive delta and the high destructive delta dominated by wave action. These authors noted that bedload type affected the delta type, with fine grained bedload favouring

Figure 6.5: Delta Models

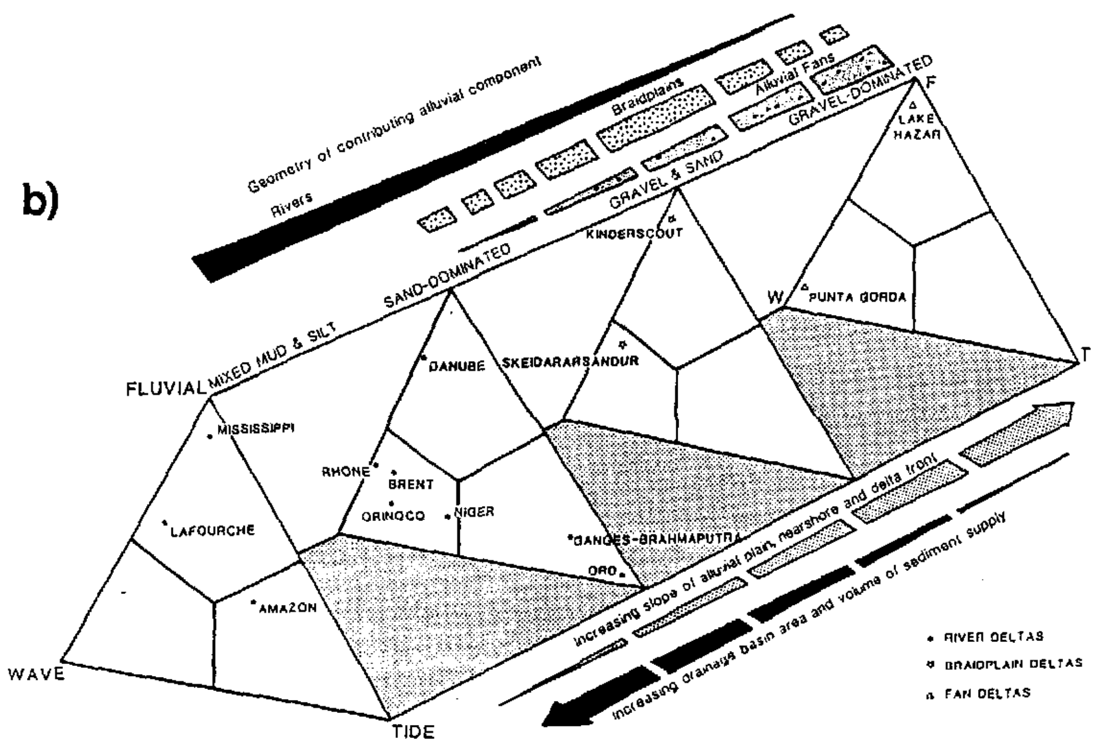
a) Tripartite classification of deltas based on the dominance of basinal and riverine processes (from Walker, 1992).

b) Addition of grainsize in the tripartite subdivision of Galloway (1975) suggesting the importance of sediment type in determining delta front morphology. The sediments of the study area fall into the sand-dominated fluvial and wave influenced delta. (modified from Reading and Orton, 1991).

a)


RIVER
1 MISSISSIPPI (BALIZE)
LA FOURCHE
4
PO
FRASER
NILE
MAHAKHAM
5 SAO FRANCISCO
RHONE
TIBER
NIGER 3
6 SENEGAL
2
WAVE
TIDE

b)


Geometry of contributing alluvial component
Rivers
Braidplains
Alluvial Fans
MIXED MUD & SILT
SAND-DOMINATED
GRAVEL & SAND
GRAVEL-DOMINATED
F
LAKE HAZAR
KINDERSCOUT
W
PUNTA GORDA
T
FLUVIAL
DANUBE
SKEIDARARSANDUR
MISSISSIPPI
RHONE
BRENT
ORINOCO
NIGER
LAFOURCHE
GANGES-BRAHMAPUTRA
ORD
AMAZON
WAVE
TIDE
increasing slope of alluvial plain, nearshore and delta front
increasing drainage basin area and volume of sediment supply
RIVER DELTAS
BRAIDPLAIN DELTAS
FAN DELTAS

highly constructive deltas with elongate sandbody geometries. More importantly, they recognised (both in recent and ancient sediments) that processes (hence delta geometry) vary as progradation continues. Commonly wave reworking after abandonment would change the delta front geometry. This is now well established in the Quaternary sediments of the Gulf Coast (Suter and Berryhill, 1985 Penland et al., 1988; Boyd et al., 1989). Using 34 modern deltas and emphasing river mouth processes, Coleman and Wright (1975) defined six sandbody geometries characterising major deltas today. However, like Galloway (1975), they did not take into account all the major parameters involved in deltaic sedimentation.

Recently, Elliott (1989) has pointed out that no single parameter can be used to define delta models and emphasized the importance of tectonics and sediment load in defining delta morphology. The importance of sediment load is also argued strongly by Reading and Orton (1991). The type of sediment load controls the sandbody geometry of the delta plain and the delta front gradients, which will in turn control the processes at the mouth, and the resulting sandbody geometry. The effectiveness of basinal processes will therefore depend on the slope of the delta front which is determined by the grain size (Reading and Orton, 1991). Commonly, fine grained sediments have gentle gradients (but are unstable due to abundant mass flow processes) and wave processes are generally dissipative. Using grainsize as a fourth component in the classification, Reading and Orton (1991) extended Galloway's (1975) ternary diagram, to include sediment load. The deposits of the basal Belly River Formation are sand-dominated, and show evidence for both fluvial and wave processes. The sandbody geometry is elongate to lobate. These features suggest that the sediments were deposited in a fluvial- and wave- dominated delta. These can be plotted in the uppermost section of the triangle (Figure 6.5b), with morphologically comparable deltas being the Danube and the Po deltas.

Work on tidal deltas resulted in the suggestion to remove the tidal portion of Galloway's (1975) triangle, because tide-dominated deltas bear no resemblance to other deltas types (Walker, 1992). Tide-dominated "deltas" or estuaries have sediment sourced from the basin rather than being supplied directly from the river (Bhattacharya and Walker, 1992).

The existence of deltas in the ancient record which are different from modern deltas (non-actualistic models) is suggested by Elliott (1989) and this idea is used by B.A.Power (pers. comm., 1992) to describe the deltaic sediments of the basal Belly River Formation in Pembina and surrounding fields. Martinsen (1990) has shown that different processes at the mouth can produce the same sandbody geometry, and like Elliott (1989) further emphasizes the need to revise existing deltaic models.

On a larger scale, incorporation of the effects of relative sea level changes to deltaic sedimentation led to the recognition of two delta types which depend on their position on the shelf in which they prograded. Shelf margin deltas, such as those in the northwest Gulf of Mexico formed during sea level lowstands are fixed at the shelf edge and are characterised by extensive growth faulting which exerts a major control on sandstone thickness (Suter and Berryhill, 1985; Elliott, 1989). The deltaic lobes of the pre-modern Mississippi Delta prograded extensively across a shallow water shelf and are known as shelf deltas (Suter and Berryhill, 1985). This distinction between shelf-edge and shelf deltas brings together the interaction between delta sedimentation and relative sea level change which can be analysed using sequence stratigraphic methods (Boyd et al., 1989).

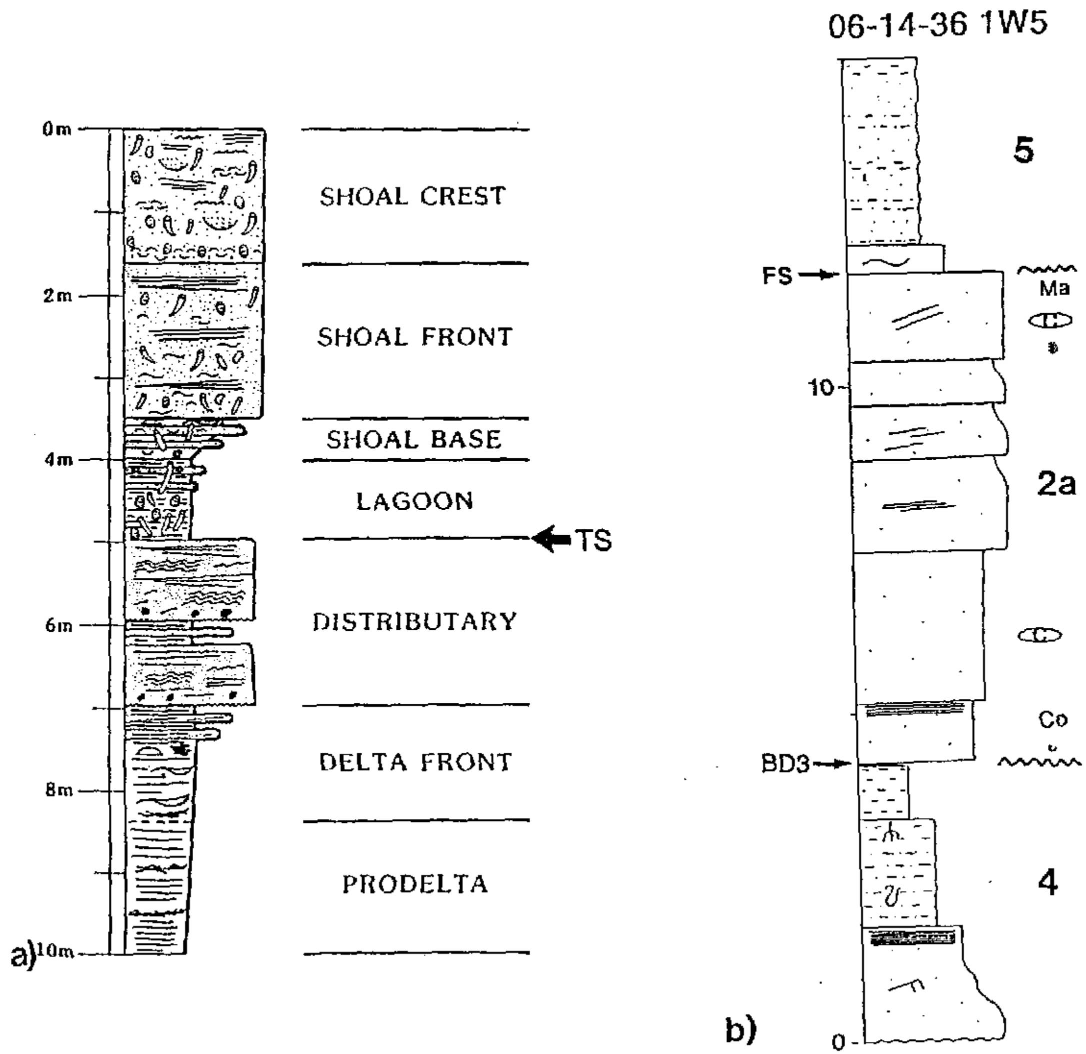
6.3.2 Comparison with the Modern deltas

The Modern Mississippi (Balize) Delta is a fine-grained delta, located within 30 km of the present day shelf edge. Due to its fine-grained nature, the channels are at a fixed position. Therefore, delta progradation is localised in a small area characterised by prograding bar fingers, oriented normal to the shoreline, and distributaries play a major role in delta progradation (Elliott, 1986) . The modern Balize delta is not a good analogue for the basal Belly River Formation because the sediment load is different, and the delta is similar to shelf margin deltas prograding in relatively deep water. However, the elongate geometry of the sandbody of A3 is similar morphologically to the bars of the Mississippi. Other modern deltas such as the Orinoco delta (van Andel, 1967) are also fine-grained, however these have distributaries over the entire delta front and the delta front is dominated by longshore currents and greater tides than in the Mississippi. In the Rhone, delta growth is by accretion of beach ridges. Progradation on the Niger delta is by tidal flat deposition and progradation of fluvial sediments. The Po river delta is sand-dominated, fed by a single distributary channel. However, sediment dispersal mechanisms are dominated by a well-developed salt wedge system (Nelson, 1970), despite having a high fluvial discharge. Therefore, processes in the Po delta front cannot form structureless sandstones as observed in the bBR (section 6.4). The delta front sediments of the bBR Formation are lobate, with elongate thicks. The sandstones were deposited in close proximity to the shoreline, and the sedimentary structures indicate a dominance of both fluvial and wave processes. The nature of the feeder system in A3 is not well defined but channels in the northwest, coeval with the shoreface sandstones are isolated and the non-marine succession is dominated by fine-grained sediments.

The geometries of the sandy pre-Modern Mississippi lobes have been modified by transgression and can not be compared to the bBR formation. However, the nature of the contacts between the transgressive and regressive sediments of the Quaternary Mississippi delta are comparable to the contacts in A3 (Figure 6.6)

A typical succession shows that sediments of the Mississippi delta are divided into two; a regressive component and a transgressive component (Figure 6.6a). The regressive succession is well documented and is represented by a coarsening-upward delta front succession, and associated delta plain facies. As individual lobes become abandoned, reworking of delta front sands occurs with continued and transgression forming beach ridge complexes which translate landwards (Frazier, 1967; Penland et al., 1988). If the rate of subsidence is fast, then the barrier islands and inner shoals form by a three-fold process of transgressive submergence (Penland et al.,1988). This results in a reworked, bioturbated sandstone lying abruptly on delta plain non-marine sediments (e.g. Ship Shoal), similar to facies succession A and B (Figure 6.6b).

Figure 6.6: Comparison of the sharp-based shoreface succession (b) with the *transgressive shoal of the* Mississippi (Boyd et al., 1989; a).


0m
2m
4m
6m
8m
a)10m
SHOAL CREST
SHOAL FRONT
SHOAL BASE
LAGOON
TS
DISTRIBUTARY
DELTA FRONT
PRODELTA
06-14-36 1W5
5
FS
Ma
10
2a
Co
BD3
4
b)
0

Therefore, in deltaic settings, sharp-based shoreface sandstone may not only be related to a drop in relative sea level but could also be due to reworking of delta front sandstones during transgression. Consequently, are the sharp-based shoreface sandstones of allomember 3, lying sharply above non-marine sediments (in Tindastoll) a result of reworking during the transgression or due to a forced regression? Although the sharp-based shoreface shows similar relationships to the shoals of the pre-Modern Mississippi, the following reasons indicate that the shoreface sandstones of A3 formed in response to a forced regression rather than due to reworking with continued transgression of A3.

(i) The internal sedimentary structures within the two sandstones are different. Sandstones of facies succession A and B show no evidence of reworking and extensive bioturbation. Furthermore, the type of trace fossils found (*Macaronichnus*) are not indicative of barrier island deposits. *Macaronichnus* requires high energy conditions as it feeds on bacteria and organic matter on the surface of sand grains (Clifton and Thompson, 1978) and depends on the constant churning of sediments by wave activity.

(ii) No laterally equivalent facies relationships (washover terraces, tidal inlets, flood tidal deltas) which are characteristic of shoal evolution are evident within the study area. The sediments are therefore unlikely to be transgressive.

(iii) The shoreface shows evidence of emergence, above the shoreface sandstones and above the abandoned open bay sediments. This is not indicative of continued transgression.

6.4 Structureless sandstones: A discussion.

Structureless sandstones are 7-8 m thick in Pembina (B.A.Power, pers. comm.), and extend 10s kms offshore. In this study area, the sandstones are commonly about 1.5 m thick but they can be up to 5 m, and are found both in marine and non-marine sandstones.

X-rayed cores were of three different types; structureless sandstones with no calcite cementation, those with calcite cementation and crudely stratified sandstones with no calcite cementation. The results of some of the x-rays of shoreface and channel sandstones are shown in Figure 4.5. In general, where present, stratification was accentuated when x-rayed. In facies 2a (shoreface sandstone), the x-rays of crudely stratified sandstones alternating with structureless beds showed similar features (Figure 4.5a). Some beds, with no calcite cementation were still structureless. However, structureless calcite-cemented non-marine sandstones (Figure 4.5b) showed no stratification when x-rayed. This suggests that the sandstones are truly structureless.

The structureless sandstones are interpreted to result from original depositional conditions and were deposited by hyperpycnal flows; alternatively, they could be due to modification by diagenesis or bioturbation which obliterates the stratification. The presence of structureless sandstones in marine deposits can be explained by looking at processes at the river mouth. However, within the distributary sandstones, it is difficult to explain what processes are responsible for producing 5m thick structureless sandstones.

6.4.1. River mouth processes (Hyperpycnal flows)

Processes at the river mouth are found to be important in determining how sediments are transported and distributed in the basin, and hence on influencing the sandbody geometry of delta front successions (Wright and Coleman, 1974). In river dominated deltas, the depositional patterns would depend on the outflow inertia, turbulent bed friction in the basin, and outflow buoyancy (Wright, 1977).

The density difference between sea water and river discharge results in three types of outflows (Bates, 1953; Wright, 1977). These are homopycnal flows, with negligible density contrasts; hypopycnal flows, which result in buoyant plumes as freshwater and seawater are well stratified, and hyperpycnal flows where the density of freshwater is greater than seawater (inertial processes are dominant) and the discharge flows as a density undercurrent. One method of forming structureless sandstones is by hyperpycnal flows. However, data on hyperpycnal flows in rivers entering the sea are rare (except for deltas in glacial settings), and commonly, models concentrate on the flow being a buoyant plume (such as in Mississippi and Po deltas).

Recent studies of the Huanghe (Yellow River) delta show hyperpycnal plumes off the delta front. These form as a result of a high sediment concentration which varies from 26g/l to 220 g/l during floods (Bornhold et al., 1986; Prior et al., 1986; Wright et al., 1986, 1988). These values are much higher than concentrations in buoyant plumes such as those of the Mississippi delta where the maximum suspended sediment concentration is only 0.2-0.4 g/l. In the Po delta, concentrations of up to 0.5g/l are recorded during floods, carrying 23% fine sand in suspension (Nelson, 1970). However, no underflows develop.

The hyperpycnal plumes off the Huanghe delta are 1-4 m thick, and extend up to 10 km offshore into the Gulf of Bohai (Figure 6.7a). The flows originate from a localised point source, and coupled with the high rates of sediment and very low delta gradients (0.1-0.6 degrees), result in rapid progradation of the delta front as a single lobe, 15 m thick. Although the sediment load is predominantly silt, fine sands are concentrated within 5 km from the delta front (Figure 6.7b).

Farther offshore, these flows are commonly channelised, and they maintain flow because of the unstable delta slope which results in periodic slumping during storms (Wright et al., 1986).

Hyperpycnal flows are one possible mechanism responsible for the formation of structureless sandstones found 10s of kms offshore of their equivalent shoreline in the basal Belly River Formation (B.A.Power, pers. comm., 1992). Structureless sandstones within the basal Belly River Formation in Pembina occur as discrete shore-normal tongues, 5-15 km wide and 10-30 km long. B.A.Power (pers. comm.) suggested deposition by sediment laden suspension currents which deposited sands very rapidly. The sandstones can show very crude low-angle stratification, which Power uses to further strengthen his argument for hyperpycnal flows. Lowe (1988) suggested that increased rates of suspension load fall-out suppress the ability of the bed to form sedimentary structures and massive sandstones are produced, some with diffuse low-angle stratification. Lowe (1988) however, gave no scale of the thickness of these flows. Arnott and

Hand (1988) showed experimentally that ripple stratification is suppressed if rates exceed 4 cm/minute. Elliott (1989) has suggested that the lowering of basin salinities, with a river load having a high suspended sediment concentration could decrease the density contrast between the two effluents and allow undercurrents to extend further seaward. The Western Interior Seaway had reduced salinities due to freshwater input except at times of maximum transgression (Kauffmann, 1984), therefore reducing buoyancy effects. The structureless sandstones of the shoreface succession show diffuse low-angle stratification and may be formed by episodic hyperpycnal flows associated with flooding. These are up to a metre thick but show clear evidence for waning flow. In the study area, the hyperpycnal flows formed in close proximity to the delta mouth and at times of low discharge, sandstones were reworked by waves producing the lobate-shaped.

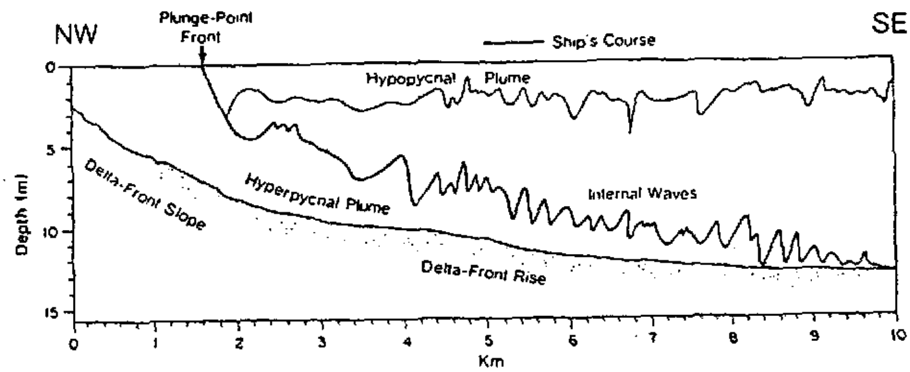
However, as pointed out by Walker (pers. comm., 1992) the thickness of some of the sandstones (especially those in Pembina Field) makes it difficult to visualise a single flow capable of forming 7-8 m structureless sandstones with no breaks in deposition. In the Huanghe delta, sand deposition occurs very close to the mouth of the delta and only silt and clay are carried offshore in suspension (Figure 6.7b). Furthermore, cores from the delta front of the Huanghe river show sandstones with both coarsening and fining-upwards trends. They also have thinly bedded stratified beds, but structureless sandstones were not described (Bornhold et al., 1986). This is probably due to the limitations of data available.

Figure 6.7: Sediment dispersal in the Huanghe river delta.

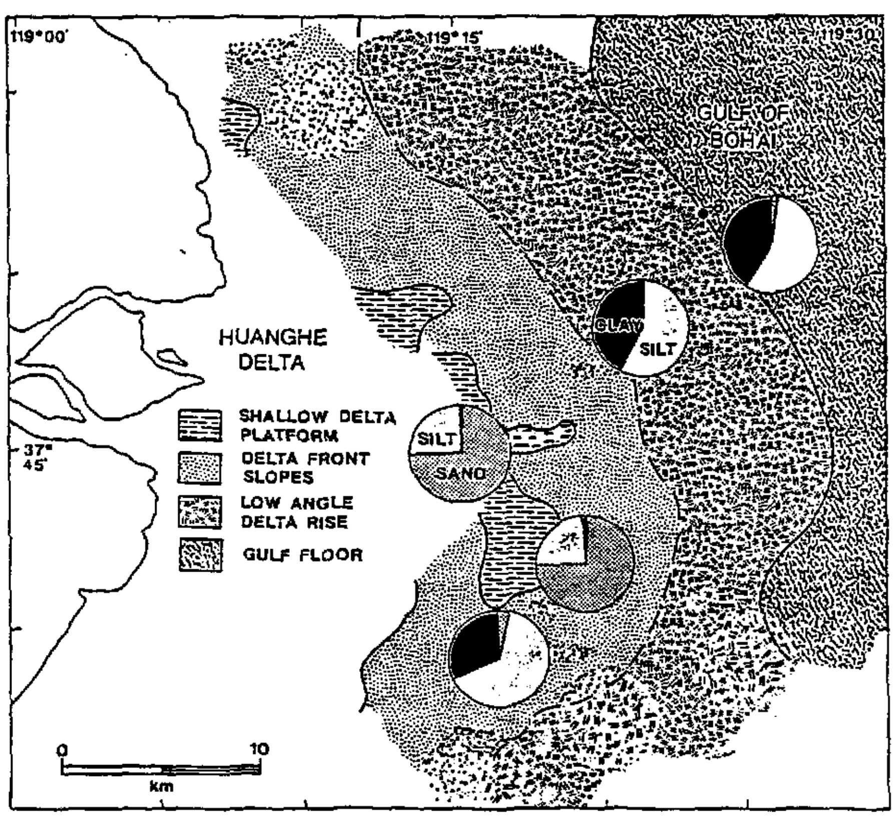
a) Profiles of hyperpycnal and hypopcynal flows off the Huanghe delta front. Hyperpycnal flows extend to about 10 km offshore. (after Wright et al., 1988)

b) Textural characteristics in the Gulf of Bohai . Sand is concentrated close to mouth of delta, in the shallow delta platform.

(after Bornhold et al,1986)


NW
Plunge-Point Front
Ship's Course
SE
Hypopycnal Plume
Hyperpycnal Plume
Internal Waves
Delta-Front Slope
Delta-Front Rise
Depth (m)
0
5
10
15
0
1
2
3
4
5
6
7
8
9
10
Km

a)


119°00'
119°15'
119°30'
GULF OF BOHAI
HUANGHE DELTA
CLAY
SILT
SAND
37°
45'
SHALLOW DELTA PLATFORM
DELTA FRONT SLOPES
LOW ANGLE DELTA RISE
GULF FLOOR
0
10
km

b)

Ancient examples of hyperpycnal flows in river deltas are scarce. In the inertially-dominated deltas of the Carboniferous of northern England (Martinsen, 1990), massive sandstones were believed to have formed from hyperpycnal flows. These flows developed during high discharge periods associated with floods (Martinsen, 1990), which resulted in deposition of sand lobes beyond the river mouth in the prodelta area. The sand was transported to the delta front area through subaqueous channels incised on the delta front. The apparently massive channel sandstones described by Martinsen (1990) are thought to have formed by subaqueous flows at the delta front during floods.

6.4.2 Diagenetic processes

Structureless sandstones are commonly but not exclusively associated with calcite cementation in the sediments of the study area. In several other deltaic formations structureless sandstones are present, associated with calcite cementation (e.g.. Rannoch Formation of the Brent Delta in the Jurassic North Sea; Daws et al., 1992). The calcite cemented zones tend to suppress the SP and result in high resistivity readings (Shouldice, 1979; Iwuagwu and Lerbekmo, 1982; Doig, 1986). Calcite zones up to 2 m thick, are widely distributed in the channel sandstones and occur with less frequency within the shoreface succession. Doig (1986) observed up to five zones in one channel sandstone, assumed to be laterally continuous for 250 m to several kilometres. These zones are lens-shaped, being thickest at the margins, and are due to late diagenesis (Doig, 1986). Their origin was implied to be due to explusion of porewaters within the underlying mudstone beds. However, in Pembina Field Longstaffe (1986) used geochemical data to suggest precipitation of calcite from groundwater. Two thin sections examined in this study showed early mechanical compaction within the coarser channel sediments which have an abundance of volcanic rock fragments, and late calcite diagenesis (after quartz overgrowth), with up to 20% calcite cementation within the finer grained sandstones. In these samples, however, stratification is very prominent. This indicates that the late calcite cement observed is not responsible for the lack of stratification seen within the channel sandstones. Iwuagwu and Lerbekmo (1982) have observed up to 40% of calcite cement with no quartz overgrowths suggesting early calcite cementation. Similarly, Longstaffe (1986) observed early calcite forming instead of chlorite. Calcite cements in the basal Belly River sandstones are therefore the result of both early and late diagenesis. Early calcite precipitation is observed in other structureless sandstones, such as the Rannoch Formation in the Jurassic North Sea. The cementation there is interpreted as being due to meteoric and marine porewaters flushing the sandstones during early burial (Daws et al., 1992). Since even the late cements are associated with high calcite cementation values, showing complete to partial quartz replacement, it is likely that there may be two phases of cementation associated with calcite. Therefore, a tentative explanation for the structureless sandstones would be to suggest that early calcite cement could be responsible for obliterating the stratification. This would at least explain the apparently massive nature of the channel sandstones. The source of calcite could be from shells or microbial degradation of organic matter (I.Hutcheon, pers. comm., 1992). More detailed work is obviously required to verify this suggestion and isotopic studies of the calcite would be able to distinguish between early and late calcite cementation and the source.

The last process capable of modifying the stratification is bioturbation. However, the results of the x-radiography displayed no evidence of extensive bioturbation of the sandstones. Furthermore, the channel sandstones are predominantly non-marine and are unlikely to be bioturbated.

7. Depositional History and Controls on Sea Level Changes

7.1 Introduction

This chapter is subdivided into two major sections. The first section reviews the overall evolution of the basal Belly River Formation in the study area. The sedimentation rates and rates and magnitude of relative sea level changes are outlined. The second section looks at the possible mechanism controlling sea level changes of the scale observed within the sediments of the basal Belly River sedimentation.

7.2 Evolution of the basal Belly River Formation

The Belly River Formation is underlain by a thick succession of marine mudstones (of the Lea Park Formation). The nature of the transition from marine to continental deposition is depicted in the lower sediments of the Belly River Formation. In the study area, the basal Belly River Formation is subdivided into three allomembers bounded by erosional surfaces which were interpreted to have formed by changes in relative sea level. The sediments preserved within the allomembers are largely progradational and were deposited in a deltaic environment.

The contact between the Lea Park Formation and the basal Belly River Formation is diachronous, younging eastwards. This contact is sharp, where offshore marine mudstones of the Lea Park Formation are overlain by shoreface sandstones of allomember 1 (Figure 7.1a). This surface (BD 1) formed because of a drop in sea level, and the progradation of the shoreface succession of allomember 1 is interpreted to be the response of the

Figure 7.1: Evolution of the basal Belly River Formation in the study area.

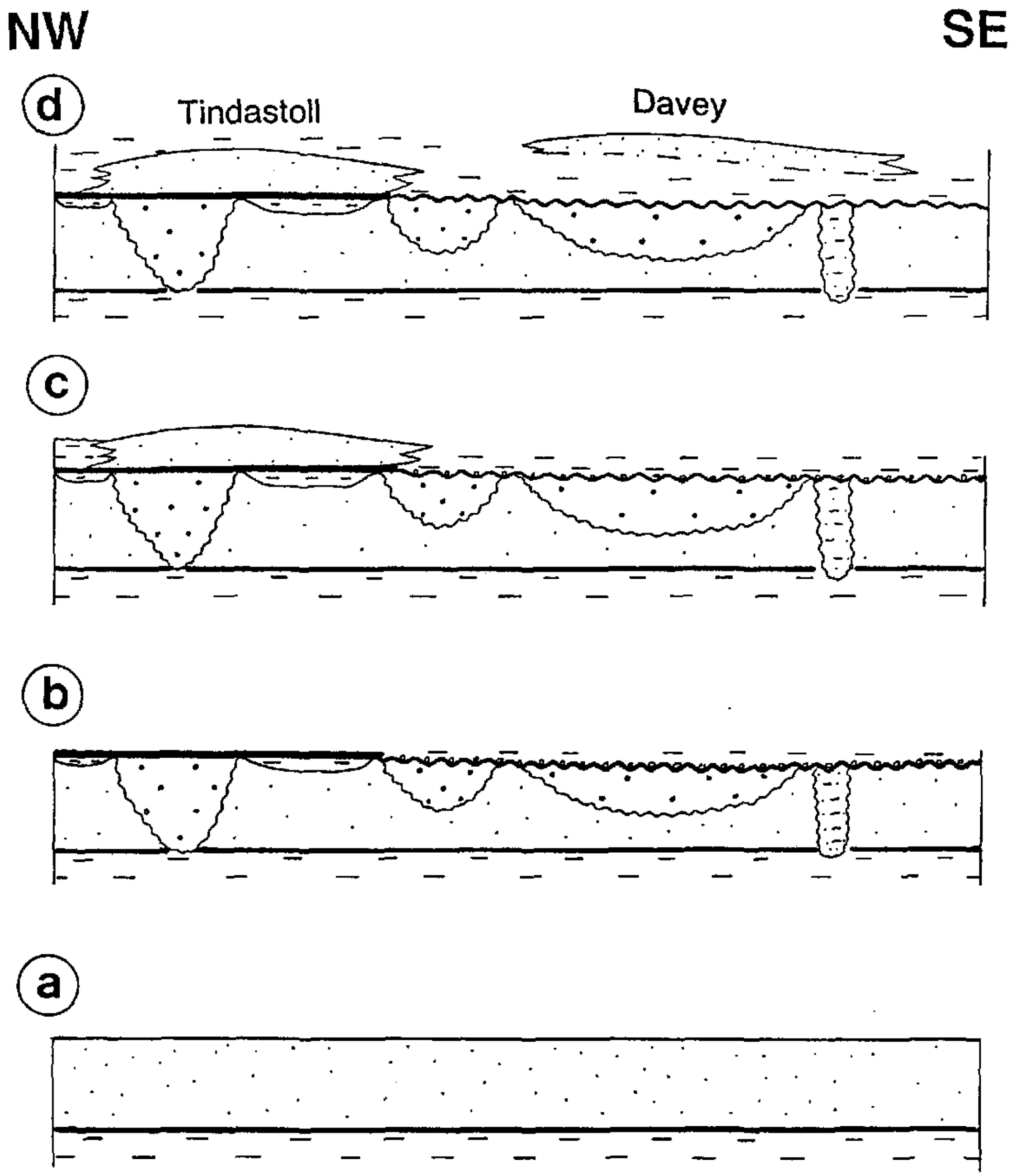
a) Formation of sharp-based shoreface prograding onto offshore mudstones of the Lea Park Formation, due to a drop in relative sea level.

b) Channel incision represented by A2 onto older shoreface succession (A1) in response to a relative sea level fall.

The ensuing marine transgression, reworked underlying sediments and a transgressive lag was deposited in the S.E of the study area. In the NW, peat deposition prevented erosion of the underlying sediments. Transgressive marine mudstones are only preserved in the SE. The following progradational succession (of A3) is characterised by a *forced regression* (c)

c) Shoreface progradation due to a drop in relative sea level. In the SE, marine mudstones were deposited. The geometry of the shoreface is lobate, with elongate thicks indicating a fluvial- and wave- dominated deltaic depositional system.

d) autocyclic lobe switching due to subsidence and/or a decrease in sediment supply.


NW
SE
d
Tindastoll
Davey
c
b
a

shoreline to this base level change. Progradation of the shoreface resulted in seaward stepping of the shoreline, hence offshore mudstones are sharply and erosively overlain by shoreface sandstones. The erosive nature of BD 1 is due to lowering of wave base. Consequently, mudstone intraclasts are abundant at the base of the shoreface.

The sandy nature of the shoreface succession and the lobe-shaped geometry of the sandbody resembles fluvial- and wave-dominated sandy delta front deposits.

Allomember 1 is extensively cut-out by sediments of allomember 2 (Figure 7.1b). The deposits of allomember 2 comprise low-sinuosity non-marine to brackish, sandy channels. These channels, up to 20m deep occur, as isolated individual channels or form a channel belt which is up to 10 km across. The channel incision is concentrated in a N-S trending belt through Davey Field.

Differences in channel fill are present in the various fields. In the north-west (Innisfail and Tindastoll Fields), the channel sandstones fine-upwards into vertical-accreting in-channel mudstones and siltstones which represent gradual channel abandonment facies. These channels are also associated with an abundance of floodplain fines that include thin coal beds. In contrast, in the south-east (Davey Field), the channels are largely sand-dominated, with no floodplain deposits. Any vertical trends present are due to changes in sand grain size.

A further drop in relative sea level is infered from the regional extent of the incision. The interpretation of these channels as incised valleys (van Wagoner et al., 1990) is not possible in this area because the coeval shoreline sediments are out of the study area. Incised valleys are commonly filled during the late lowstand or early phase of the transgression (Pattison, 1992). The fill may be non-marine at the base and more marine at the top. In the bBR case, the fill is entirely non-marine (with some brackish water trace fossils present in the channels of Davey field). In an incised valley, a non-marine fill usually forms the proximal part of the valley. Therefore, if the channels represent an incised valley then the sediments are proximal or were infilled during lowstand. The response of fluvial channels to a change in base level suggests that channels prograde in the same direction as the movement of its coeval shoreline. Examples from the Quaternary deposits of the Louisiana continental shelf (Suter and Berryhill, 1985) and the Rhone Delta (Tesson et al., 1990) show that channels can prograde up to the shelf edge (similar to the Modern Mississippi), cutting out coeval or older shoreface successions to lie erosively on offshore mudstones. The depth of incision of allomember 2 suggests that these channels were eroding into an older shoreface succession, feeding a coeval shoreface farther south (out of the study area).

Following the deposition of allomember 2, the study area was inundated by marine waters. During the initial stages of the transgression, wave erosion scoured underlying sediments, and redeposited them as a thin transgressive lag. The extent of the lag was mapped and found to coincide with the limit of floodplain deposition (in particular coal). Therefore peat prevented erosion of the underlying sediments in the north-western part of the study area.

Detailed correlation shows that the nature of the lower bounding discontinuity of allomember 3 (BD3) varies in the three fields. In Davey, a lag was deposited above a trangressive surface of erosion. Gradual deepening resulted in deposition of offshore *Helminthopsis* mudstones above the lag. In Tindastoll field, transgressive sediments are thin or missing, and sharp-based shoreface sandstones lie erosively on coal.

A second forced regression resulted in progradation of a second sharp-based shoreface in Tindastoll (Figure 7.1c). In Davey Field, *Helminthopsis* mudstones continued to be deposited. Therefore, these mudstones are transgressive and regressive. Progradation of a second shoreface (above the offshore mudstones) in Davey is due to autocyclic lobe switching (Figure 7.1d), perhaps due to channel avulsion upstream. This shoreface is gradational with the underlying offshore mudstones.

As a result of autocyclic lobe switching, marine flooding surfaces of limited lateral extent occur above the sharp-based shoreface sandstone of Tindastoll. This surface separates the open bay facies of the abandonment phase from the progradational phase. The abandonment facies get progressively non-marine upwards suggesting emergence of the first delta lobe in the northwest.

The nature of the shoreline during deposition of allomember 3 was interpreted from isopach of the sand thickness (Chapter 5). The 3-D geometry shows that sediments of allomember 3 were deposited in a deltaic environment. The sandbody geometry is lobate with elongate thicks close to the river mouth.

These sediments can be compared with modern deltaic systems. The sandy nature of the delta lobes, the sedimentary structures and the trace fossils indicate a river- and wave-dominated deltaic system. Initially, river processes were dominant, but as progradation continued, wave influence becomes increasing important. These processes are similarly observed in the sandy lobes of pre-modern Mississippi deltaic system. The major difference is the temporal and spatial scale of the two systems. The sandy delta lobes of the pre-modern Mississippi deltaic system are an order of magnitude larger than the deltaic system of the bBR, and lobe switching occurs every 500 years.

7.2.1 Rates of sedimentation

The rates of sedimentation can be estimated by two methods (Table 7.1). The first method uses published values derived from Chamberlain et al. (1989), based on rates of subsidence for the Western Canada basin. The average sedimentation rate is estimated to be 97 m/Ma in the study area. The thickness of sediments studied is 55 m. These would therefore have been deposited in a period of less than 1 Ma. Posamentier and Vail (1988) indicated that base level changes due to relative sea level changes affect the whole drainage system and that the river shifts seaward on a similar time scale. Therefore, although allomember 2 consists entirely of channel sandstones, they can be assigned an equal duration to the shoreface succession of allomember 1. By assigning an equal age to the three allomembers, deposition of each allomember would take place in 189,000 year cyclicity. This falls into the high frequency fourth-order cycles of van Wagoner and Mitchum (1991). The frequency of the cyclicity is similar to that estimated for the cycles of the basal Belly River Formation in Pembina Field (B.A.Power, pers. comm., 1992).

The second method is from the correlation of biostratigraphy and absolute age determinations. However, due to the problems of establishing both a biostratigraphic and absolute age of the base of the formation, the method is not feasible. Lerbekmo (1989) used radiometric dates from southern Alberta to estimate sedimentation rates for Upper Campanian and Maastrichtian of 60 m/Ma. The rates would also result in fourth order cycles but the estimated duration of each cycle is nearly twice that estimated by using the rates estimated by Chamberlain et al. (1989). Therefore, the duration of each of the fourth-order cycles is estimated to be between 189 to 305 ka.

7.3. Mechanisms controlling relative sea level changes

The bounding discontinuities that were outlined earlier were interpreted to have formed as a result of relative sea level changes. The controls on relative sea level changes are subdivided into two parts; allocyclic controls which include eustasy and regional tectonics, and autocyclic controls. This section describes briefly the various mechanisms which control sea level fluctuations and those mechanisms which may be applied to the sediments of the basal Belly River Formation.

RATES OF SEDIMENTATION

USING CALCULATED VALUES (CHAMBERLAIN et al., 1989)

At 52° N, 113-114 ° W; sedimentation rates are 73-77 m/Ma.

At 50° N, 113-114 ° W; sedimentation rates are 121 m/Ma.

Average sedimentation rates for 51° N are 97 m/Ma.

Study area lies between 51° N and 52° N, 113-114 ° W.

Sedimentation rates in study area varies from 77-97 m/ma.

Thickness of basal Belly River Formation in study area is about 55m.

Deposition of basal Belly River Formation in study area in less than one million years (0.56-0.7 Ma)

Table 7.1: Estimated sedimentation rates for the basal Belly River Formation in the study area.

7.3.1 Eustasy

Global changes in sea level are controlled by the changes in volume of ocean water (glacio-eustatic controls) or by changes in the volume of ocean basins (tectono-eustatic controls; Donovan and Jones, 1979; Hallam, 1984).

Tectono-eustatic controls can produce sea level changes by changing the volume of oceanic spreading ridges. Rapid rates of spreading (Hays and Pitman, 1973) or changes in the length of the ocean ridges (Donovan and Jones, 1979; Hallam, 1984) can cause changes in sea level of up to 300 m. Another method of increasing the volume of ocean basins (and subsequently raising sea level) is by lithospheric heating and mid-plate volcanism (Schlanger et al., 1981). Schlanger et al. (1981) associate Late Cretaceous sea level rise due to mid-plate volcanism which resulted in swells on the Pacific and Farallon plates. Sea level fall is attributed to thermal subsidence of the swells. This mechanism cannot explain the cyclicity seen as no periodicity is implied.

The rates of sea level changes by tectono-eustatic controls (1 cm/1000 years) are too slow for the frequency of sea level changes observed in the bBR Formation. Tectono-eustatic controls are generally reflected in second-order cycles which span 10-100 Ma (Plint et al., 1992).

Glacio-eustatic changes of sea level are controlled by the volume of terrestial ice and can cause up to 150 m of sea level change. Calculations of the volume of present day terrestial ice indicate a sea level rise of 40-50 m would result if the ice completely melts (Donovan and Jones, 1979). A sea level drop of about 100 m is estimated during the peak of the

Pleistocene glaciation (Donovan and Jones, 1979). Ice sheets grow and decay over a shorter time period and are a feasible mechanism for explaining high frequency cycles (third and fourth order). Rates of sea level change of 1m/1000 years due to glacio-eustasy are an order of magnitude faster than tectono-eustasy. Therefore, glaciation is capable of relative changes of sea level of the scale seen within the deposits of the bBR Formation.

However, several authors have pointed out that in the Cretaceous, sea level changes had little or no glacial component (Hallam, 1984; Chamberlain et al., 1989; Christie-Blick, 1990). Futhermore, due to isostatic rebound, glacio-eustastic controls cannot produce global sea level changes of uniform magnitude (Cloetingh, 1990). Frakes and Francis (1988) have observed ice-rafted deposits in the Lower Cretaceous of Australia, but this implies only the presence of river or coastal ice and does not provide sufficient evidence to suggest extensive low-altitude glaciation in the Cretaceous. In an orogenic belt, the presence of high altitude (alpine) glaciers in the Cretaceous cannot be ruled out. However, melting of present day alpine glaciers would only cause a sea level rise of less than a metre (Plint et al., 1992). Hence, although land ice may have been present, the magnitude of sea level change and the frequency could not be explained by alpine glaciation alone.

There is also no evidence for glaciation in carbonate sediments of equivalent age. In Alabama, the Campanian marl-limestone cycles within the Arcola Limestone Member were probably controlled by changes in sea level due to intraplate deformation or short-term eustatic sea level changes and are not glacially-induced (King, 1990).

In the Western Canada Basin, high-frequency cycles are observed in many formations (Ainsworth, 1991; Plint et al., 1992). Due to the absence of tectonic models to explain the rate and magnitude of sea level changes, glacio-eustasy is commonly cited as mechanism which best explains the cyclicity observed (Cloetingh and Kooi, 1989).

Deposition of the bBR Formation in a foreland basin, during the Laramide orogeny suggests that regional tectonics may have a greater control on relative sea level changes than the effects of eustasy. These tectonic controls are discussed in section 7.3.3.

7.3.2 Correlation with eustatic sea level curves

Exxon's sea level cycle chart (Haq et al., 1988), shows a rapid third-order sea level drop of about 65 m at 80 Ma. Plint et al. (1992) associated this fall with the onset of deposition of the basal Belly River Formation. Therefore, BD 1 may be correlated to the falling limb of a third-order cycle in Haq et al., (1988) eustatic curve. However, two major problems are apparent in establishing such a correlation.

Firstly, due to the diachronous nature of the base of the Belly River Formation, the absolute date for onset of deposition must vary considerably. The problem of dating this contact is evident from the various absolute dates established by various authors (Chapter 2). Lerbekmo (1989) in southern Alberta used Obradovich and Cobban (1975) dates and established the basal Belly River Formation to be 77 Ma, and correlating with the *Baculites asperiformis* Zone. On the other hand, Mack and Jerzykiewicz

(1989) use the dates of Gill and Cobban (1973) to establish the base of the formation in the Foothills. Their date of 79.5 Ma is correlated with *B. mclearni* Zone, which is one ammonite zone lower than *Baculites asperiformis*. Therefore, by using a 0.5 Ma equal duration for each ammonite zone, the variation in absolute dates between the two areas is about 2 Ma. The sediments of the basal Belly River in the study area were deposited in less than a million years hence the biostratigarphic resolution at the moment is not satisfactory to enable correlation with eustatic sea level curves.

Secondly, problems with the precision of the Exxon cycle chart itself and the problems of correlation between different curves to that of Haq et al. (1988) are apparent. Correlation of transgressive-regressive curves of the western interior of USA, (Hancock and Kauffmann 1979; Weimer;1988) with eustatic curves of Europe indicate some variations in the position of the peaks within the Campanian, suggesting a possible tectonic component. In northern Europe, the Middle Campanian records a steady rise in sea level, culminating in the Late Campanian. In the western interior of the USA, there are several transgressive-regressive peaks of lower amplitude. These differences are explained by the effects of the Laramide orogeny. In Canada, Stott (1984) was not able to correlate T-R peaks with the eustatic sea level curve.

The problems of correlating third-order sequences to the Haq et al. (1988) global chart and the methods used to derive the curve have been critically reviewed by several authors (Miall, 1992b; Walker, 1990; Christie-Blick et al., 1990). Miall (1992b) pointed out that the precision of third-order cycles is far greater than the available precision from existing chronostratigraphic tools. The problems of infering amplitude from coastal aggradation is noted by Christie-Blick et al. (1990). Amplitudes are not realistically defined due to the effects of local subsidence and isostatic adjustment due to sediment load (Christie-Blick, 1990).

Sahagian and Holland (1991) showed the estimate of the amplitude of sea level change varies if a stable frame of reference is used. Preliminary results from the stable Russian platform indicate differences in the magnitude of sea level change between Haq et al. (1988) and Sahagian and Holland (1991). Haq et al. (1988) long term curve shows a sea level rise of 200 m for the Bajocian to Turonian. Sahagian and Holland (1991) found for the same period, a rise of about 120 m. Differences are related to the differences in subsidence between passive margins and stable platforms, cumulate sequence thickness and compaction errors.

In summary, eustatic controls on sea level changes are probably not the main mechanisms responsible for the cyclicity observed. Tectono-eustatic rates are two slow for the sedimentation rates (97m/Ma) estimated. High frequency cycles can be produced by glaciation, but there is no evidence to support low-altitude glaciation in the Campanian.

Correlation with Haq et al. (1988) sea level curve is not feasible due to i) problems with the chronostratigraphic resolution of the basal Belly River Formation and ii) problems with the precision of the global cycle chart.

7.3.3 Regional tectonics

Regional tectonics result in lithospheric loading and subsidence in foreland basins. Consequently vertical motions within the basin can give rise to changes in relative sea level and generate the pattern of sedimentation observed in foreland basins. These mechanisms are discussed below.

i) presence of a subducting plate

Regional crustal subsidence and high rates of sedimentation have been correlated with collision of allochthonous terranes, thrusting and lithospheric flexure due to underplating or subduction (Cant and Stockmal, 1989; Cross and Pilger, 1978; Chamberlain et al., 1989).

Low angle subducting plate is attributed to the great thickness of Cretaceous sediments in western interior of USA.

In the western interior of USA, regional subsidence during the Campanian is attributed to subcrustal loading due to shallow subduction of an eastward dipping plate (Cross and Pilger, 1978). In this area, evidence suggests that during the Campanian the Farallon plate was subducting at a shallow angle beneath the central part of western USA. Rapid convergence and active overriding of the Farallon plate by North America increased the shallow angle of subduction. Two mechanisms associated with the subducting plate are responsible for the subsidence observed: isostatic subsidence by subcrustal loading and subcrustal cooling. Density contrasts between the asthenosphere and the subducting oceanic lithosphere results in early isostatic subsidence (up to 3 km) by crustal loading. Cooling of the base of the continental crust due to subduction of a cold lithosphere (by conduction and dehydration of the sinking slab) induced a long-term subsidence trend by increasing the density of the continental crust. Subcrustal cooling can produce up to 1-2 km of subsidence. No periodicities are implied and there is no evidence for an eastward dipping subducting plate in the western Canada basin in the Late Cretaceous (Chamberlain et al., 1989) making this mechanism not applicable to the study area.

ii) thrusting

Regional crustal subsidence can be due to tectonic loading by emplacement of thrust sheets. Jordan (1981) and Beaumont (1981) have shown that lithospheric flexure due to loading results in an asymmetric basin with a peripheral bulge that varies depending on the flexural rigidity of the lithosphere. The relationship between the peripheral bulge with respect to loading of the lithosphere has been modeled by various authors to determine the evolution of foreland basins. Both viscoelastic and elastic lithospheric models have been used (Beaumont, 1981 and Jordan, 1981 respectively). These early models associated deformation (thrust events) with rapid progradation of coarse clastics and fine-grained sediments were associated with tectonic quiescence or sea level rise.

Recent models suggest the opposite; thrusting has been shown to induce rapid subsidence, trapping coarse sediments at the edge of the basin and depositing fine-grained sediments over the rest of the basin. During tectonic quiescence, coarse sediments prograde farther out into the basin. Therefore, the thickness of strata does not mean onset of thrusting except in proximal areas, close to the load.

Quinlan and Beaumont's (1984) model predicts the formation of regional unconformities by the relaxation of a visco-elastic lithosphere.

Flemings and Jordan's (1990) model on foreland basin evolution is one of erosion and redistribution of sediments on an elastic lithosphere.

Flemings and Jordan (1990) incorporate into their elastic model the interplay between erosion and sedimentation to explain the pattern of sedimentation in foreland basins. They modeled sedimentation rates as a function of thrusting and quiescence events on 2 Ma intervals. During thrusting, a wedge-shaped stratal geometry forms which thins basinward. The ensuing quiescent period produces a lens-shaped stratal geometry which progrades farther basinward. This is because during thrusting, subsidence is greatest close to the thrust zone due to crustal thickening in the foreland basin. Therefore, during thrust events, the forebulge migrates towards the thrust, narrowing the basin whereas, during quiescence the basin widens. The stratigraphic record of thrust events therefore shows facies progradation in the direction of the thrust movement, which is punctuated by retrogradation at the onset of thrust cycles. In contrast, in the distal areas of the thrust, an erosional unconformity will mark out the onset of thrusting. Episodic thrusting alternating with quiescence therefore results in a stair stepped geometry bounded by unconformities.

The average rates of thrust movement in foreland basins are taken as 6m/1000 years for individual thrusts and 10m/1000 years for thrust belts (Jordan et al., 1988). These authors also estimated load thickness, with average values of 0.05 to 1 m/1000 years. Hence, rate of subsidence due to thrusting in the foothills can be estimated. Subsidence rates are dependent on the flexural rigidity and will be a fraction of the thickening rate. Jordan et al. (1988) estimate proximal basin subsidence to be 20-40%

of the thickening rate. Therefore, subsidence rates due to thrusting (estimated from thickening rates) vary between 0.01 m/1000 yrs. to 0.4 m/1000 yrs. The rate of subsidence observed in the basal Belly River Formation (0.1 m/1000 years) is within the range of values estimated if subsidence were continuous. It has been suggested that thrusting is episodic (Jordan et al., 1988) and rates can be much higher, as indicated in the subandean thrust belt where rates of up to 30 m/1000 years are estimated. Therefore, thrusting is capable of generating the rates of subsidence seen and produce third-order cycles. Furthermore, Jordan et al. (1988) found that uplift and denudation rates are of the same order of magnitude, suggesting that depending on the lithology, climate and topography, short term rates of erosion may be equal to or greater than uplift rates.

However, the sediments of the bBR Formation have a fourth-order periodicity which is not fully explained by these models.

Mack and Jerzykiewicz (1989) suggested that the initial progradation of the Belly River clastic wedge was due to thrusting in the Rocky Mountain Belt and contemporaneous volcanic activity. Mack and Jerzykiewicz (1989) used vertical trends in composition and related them to specific tectonic events. In the southern Foothills, the sediments are characterised by a high concentration of volcanic detritus which is absent in the underlying Alberta Group. If volcanism were contemporaneous, then thermal subsidence due to contraction of the magmatic material may be important. Thermal subsidence is however slow, decaying at an exponential rate following the initial thermal event (i.e. crustal heating and uplift; Cross and Pilger, 1978).

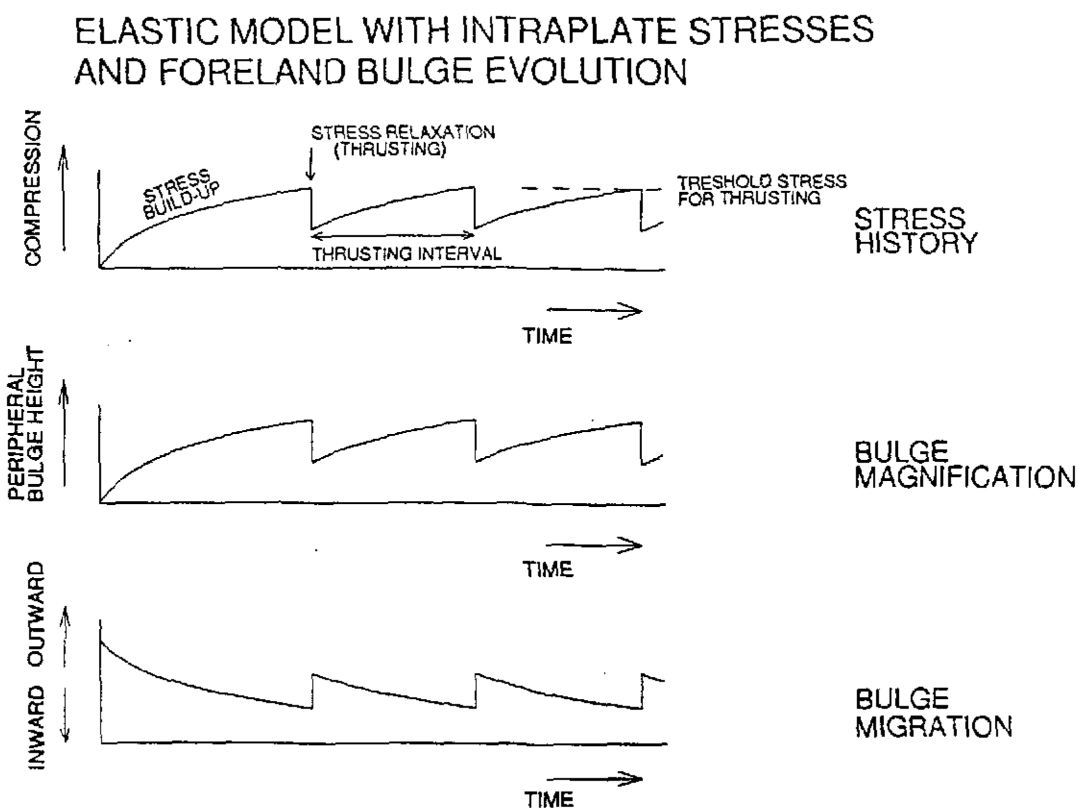
Furthermore, correlating thrusting with coarse clastic progradation would imply sediment supply exceeded subsidence due to thrusting in the basin or that their interpretation is based on the proximal area of the basin, close to the load.

iii) intraplate stresses

Short-term changes in relative sea level can be produced by rapid, stress-induced vertical motions of the lithosphere (Cloetingh and Kooi, 1989; and Cloetingh, 1990) and can significantly contribute to the total tectonic subsidence. Theoretical modeling of horizontal intraplate stresses within the lithosphere showed that stress fields generated at plate boundaries (due to plate reorganizations) can propagate into the plates where they can induce vertical motions within sedimentary basins. Changes in the magnitude and orientation of the stresses can operate on a time-scale of a few million years thereby causing relative sea level fluctuations of the scale of a third-order cycle.

Cloetingh and Kooi (1989) and Cloetingh (1990) have shown that a stratigraphic succession similar to that generated by Quinlan and Beaumont's (1984) visco-elastic model may be generated by the influence of tectonic compressional stress on an elastic lithosphere. The results of their model (Figure 7.2) indicates that tectonic stresses can modify the peripheral bulge resulting in changes in the pattern of sedimentation within a foreland basin. Accumulation of stresses, released by thrust events causes uplift and cratonward migration of the peripheral bulge during quiescence (Figure 7.2). This alternating sequence of stress build-up and thrusting can

Figure 7.2: The effects of stress fluctuations and thrusting on foreland bulge migration and basin subsidence. Build-up of compressional stresses, released during thrusting results in vertical and lateral motions of the bulge, affecting sedimentation patterns in the basin.
(after Cloetingh, 1991)


ELASTIC MODEL WITH INTRAPLATE STRESSES
AND FORELAND BULGE EVOLUTION
COMPRESSION
STRESS BUILD-UP
STRESS RELAXATION (THRUSTING)
TRESHOLD STRESS FOR THRUSTING
THRUSTING INTERVAL
STRESS HISTORY
TIME
PERIPHERAL BULGE HEIGHT
BULGE MAGNIFICATION
TIME
INWARD OUTWARD
BULGE MIGRATION
TIME

produce changes in sea level, capable of generating third-order transgressive and regressive cycles. Therefore the advocates of intraplate stresses model see short-term deviations from long term trends (thermal) in basin subsidence to be due to "stress-induced vertical motions of the lithosphere" rather than due to eustasy. Again, intraplate stress fluctuations alone cannot explain fourth-order cycles.

Recently Peper et al. (1992) using an elastic lithosphere, modeled the combined effects of thrusting and intraplate stress fluctuations to produce an explanation for both short and long term sea level changes. They concluded that short-term ($10\text{-}10^4$ yr.) vertical motions of the basin are due to thrust events while long-term subsidence may be due to large scale deformation of the accretionary wedge.

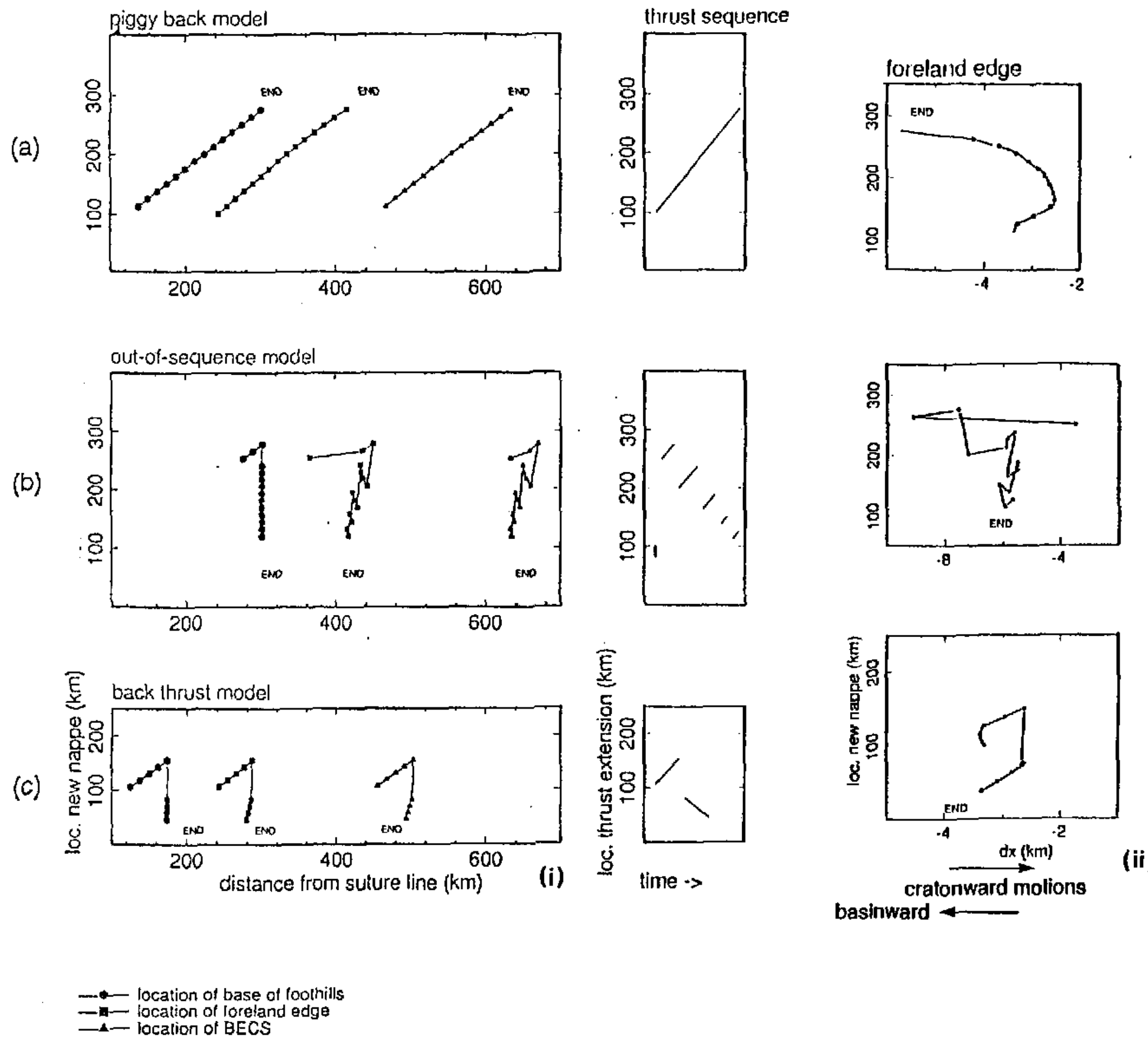
Fault geometries influence the duration of the magnitude of stress fluctuations by changing the volume of the lithosphere. In previous models, the effects of thrusting are analysed by prolonged progradation of a single wedge. Peper et al. (1992) indicate that the wedge is composed of thrust nappes which move independently of each other. Thrusting, due to stress relaxation, results in redistribution of the mass which exerts a major control on the basin shape, as the location of new thrust loading changes with time. The effect of three types of thrusting on basin shape are shown on Figure 7.3. Piggy-back thrusting causes migration and deepening of the foothill area (Figure 7.3a) whilst out of sequence thrusting (OOST) and back thrusting both result in basinward migration of the foreland edge and tectonic subsidence of the foothills (Figure 7.3b,c). As a result, different thrust styles can produce a variety of changes in the basin.

Figure 7.3: Changes of basin shape (i) and the lateral migrations (ii) of the foreland edge due to compression on different thrust styles.

a) Piggy-back thrusting which results in continuous deepening of the foothill area as the basin edge migrates towards the orogenic wedge.

b) & c) Out of sequence thrusting (OOST) and back thrusting. Both show initial migration of the basin edge towards the wedge, followed by basinward migration hence resulting in simultaneous basinward migration of the foreland edge and subsidence in the foothills.

(modified from Peper et al., 1992)



piggy back model
thrust sequence
foreland edge
(a)
(b)
(c)
END
out-of-sequence model
back thrust model
loc. new nappe (km)
loc. thrust extension (km)
distance from suture line (km)
(i)
time ->
dx (km)
(ii)
cratonward motions
basinward
location of base of foothills
location of foreland edge
location of BECS

Combining the effects of intraplate stress fluctuations and different thrust styles, Peper et al. (1992) produced models that predict both short-term and long-term changes. The effects of shortening on the stress field indicate that stress fluctuations in the lithosphere are dependent on the scale of thrusting, the geometry of the thrust and the rheological properties. These models showed that curved faults can lead to inhomogeneous distribution of stresses which can lead to short-term fluctuations in the underthrusted lithosphere. Local stress fluctuations in the order of 50 MPa can occur in 40,000 years.

The consequences of these models in foreland basin sedimentation indicate that the effects of fluctuating stresses and thrusting styles can produce a wide age spectrum of uplift of the basin with time, thereby generating sequences of similar time-scale as the stress fluctuations. Small scale thrust events are capable of producing short-term vertical motions in the basin, while long-term changes (capable of producing third-order cycles) may reflect long term stress fluctuations resulting from major reorganisation in the orogenic wedge.

The makeup of the vertical motions changes with orogenic maturity. The impact of stress fluctuations on tectonic mature accretionary wedges is greater whilst thrusting is important in younger basins. Facies transitions due to thrusting in mature orogens become less important and flexural changes due to stress increase become significant.

The models suggest that stress fluctuations and thrusting are controlled by the presence of and size of the peripheral basin, the thickness of the elastic lithosphere and the volume and shape of the overriding nappe.

The models rely on the interaction of thrust and rift flexure from a peripheral basin (similar to Appalachian foreland basin) but such a basin is not associated with the Alberta foreland basin. Furthermore, their models do not incorporate the effects of erosion on the lithosphere.

7.3.4 Autocyclic controls of delta lobe switching

Autocyclic controls may be related to early compaction of underlying mudstones. The underlying Lea Park Formation consists of 130 m of marine mudstones. Therefore, early compaction of the mudstones may produce the subsidence required for sedimentation of the allomembers. However, the sharp-based nature of the progradation packages imply an allocyclic control rather than gradual subsidence due to compaction.

Autocyclic controls are important with the allomembers. Sediments of allomembers 1 and 3 consist of delta front succesions The initial progradation of the delta front sediments is due to a relative sea level fall. However, within individual allomembers progradation packages are separated by localised marine flooding surfaces. These surfaces were probably formed due to autocyclic delta lobe switching. Lobe switching may be due to changes in sediment supply in the hinterland, or commonly, due to channel avulsion in the upstream part of the delta.

7.3.5 Summary: controls on sea level changes

Two possible controls on sea level changes which are capable of producing fourth-order cycles are glacio-eustasy and regional tectonics (thrusting and intraplate stresses). However, both mechanisms are not well demonstrated in the existing literature. As yet, glacial deposits have not been identified in the Cretaceous to suggest glacio-eustatic controls.

Sea level curves of western interior seaway cannot be correlated with the global chart. Transgressive-regressive peaks in the Mid-Campanian of the western interior of USA suggest that regional tectonic controls may be important, overprinting any eustatic controls. However, tectonic models have yet to explain the cyclicity observed.

Channel incision of allomember 2 represents a response to sea level fall. This incision cuts into older shoreface succession and the Lea Park Formation, and feeds a coeval shoreline which is out of the study area. Channels fills are sandy, with the exception of one muddy channel. The fill is predominantly non-marine, being to brackish only in the southeast of the study area.

Two types of channel fill are recognised. Fining-upward channels are present in the northwest. These are associated with overbank mudstones and siltstones and thin coal beds. In contrast, in the southeast, no overbank or in-channel fine-grained deposits are present. These represent low sinuosity channels formed as a result of frequent channel switching.

Transgressive sediments are thin or missing and are represented by thin pebbly lag with *Glossifungites* ichnofacies and offshore mudstones. The northwestern extent of the lag was found to coincide with the southeastern limit of coal beds. The presence of peat therefore prevented erosion of the underlying sediments during transgression.

The duration of each allomember is estimated to be 190,000 years, indicating a fourth-order periodicity.

(v) Two mechanisms capable of generating fourth-order cycles are glaciation and regional tectonics. The cyclicity is probably not eustatic-driven. As yet there is no evidence to suggest extensive glaciation in the Cretaceous. Tectonic controls in the foreland basin may be more important, however, tectonic models do not adequately explain the cyclicity observed.

Autocyclic delta lobe switching is an important mechanism within the allomembers. Gradual coarsening-upward shoreface successions are due to delta lobe switching.

Autocyclic controls result in marine flooding surfaces of limited area extent.

(vi) Structureless sandstones are present in the channel succession as well as in the delta front. The structureless shoreface sandstones probably represent original depositional features, formed by hyperpycnal flows in the delta front. Structureless sandstones in the channel succession are associated with calcite cementation and may represent the result of modification by early calcite diagenesis.

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Location	Top (metres)	Base (metres)	Recov.d (metres)	Location	Top (metres)	Base (metres)	Recov.d (metres)
06-22-31 28W4	1185.0	1203.0	18.0	01-08-35 26W4	1124.0	1142.0	18.0
08-32-32 26W4	1003.0	1017.0	14.0	01-09-35 26W4	1060.7	1066.8	6.0
06-18-32 27W4	1120.0	1131.0	10.6	06-09-35 26W4	1084.2	1094.8	10.7
07-34-32 27W4	1063.8	1082.0	15.0	06-04-35 28W4	1170.0	1181.9	11.9
07-28-32 28W4	1197.9	1213.7	15.0	06-18-36 28W4	1112.0	1130.0	18.0
10-04-33 26W4	993.0	1002.0	8.4	07-13-32 01W5	1260.3	1281.7	21.4
07-17-33 26W4	1002.0	1016.5	9.4	16-14-33 01W5	1266.0	1284.3	18.1
04-20-33 26W4	1004.0	1022.0	15.0	06-15-33 01W5	1318.7	1338.0	19.0
07-30-33 26W4	1019.5	1028.4	9.0	06-13-34 01W5	1207.0	1217.0	10.0
09-01-33 27W4	1046.8	1055.8	9.0	06-14-34 01W5	1206.0	1222.0	16.0
07-26-33 27W4	1082.0	1100.0	18.0	06-34-34 01W5	1286.0	1301.0	15.0
11-18-33 28W4	1237.7	1255.7	17.7	06-35-34 01W5	1228.0	1250.5	18.0
10-29-33 28W4	1215.0	1251.3	36.6	14-10-35 01W5	1201.0	1218.3	17.3
12-29-33 28W4	1226.0	1252.0	26.3	14-17-35 01W5	1140.0	1154.0	4.9
14-32-33 28W4	1225.0	1243.0	18.0	14-22-35 01W5	1187.0	1204.0	16.9
14-33-33 28W4	1175.0	1192.8	17.0	06-28-35 01W5	1199.0	1217.6	17.9
10-24-33 29W4	1248.0	1265.0	17.0	14-28-35 01W5	1195.4	1213.6	17.6
10-16-34 27W4	1104.9	1123.2	18.2	14-29-35 01W5	1213.0	1248.4	36.0
07-29-34 27W4	1133.9	1152.1	18.3	16-29-35 01W5	1164.0	1172.0	8.0
11-35-34 27W4	1059.8	1078.0	18.3	16-29-35 01W5	1207.0	1224.7	17.7
08-04-34 28W4	1196.0	1217.0	18.5	08-32-35 01W5	1194.0	1217.0	23.3
08-05-34 28W4	1199.0	1216.0	13.4	04-33-35 01W5	1205.0	1222.0	17.0
16-05-34 28W4	1212.0	1230.0	18.0	14-34-35 01W5	1162.0	1186.8	24.7
04-09-34 28W4	1209.0	1222.5	13.9	14-03-36 01W5	1162.0	1192.0	22.0
07-09-34 28W4	1198.5	1223.5	24.0	16-03-36 01W5	1163.0	1189.7	26.7
11-09-34 28W4	1207.0	1225.0	18.0	06-04-36 01W5	1172.0	1198.0	26.0

Appendix I : Core Listing

Location	Top	Base	Recov.d	Location	Top	Base	Recov.d
	(metres)	(metres)	(metres)		(metres)	(metres)	(metres)
05-10-34 28W4	1207.0	1222.0	14.5	16-04-36 01W5	1164.0	1191.5	25.7
05-16-34 28W4	1207.0	1225.0	17.6	08-09-36 01W5	1163.6	1174.7	10.2
08-17-34 28W4	1207.0	1225.2	18.0	06-10-36 01W5	1168.0	1178.5	10.5
16-18-34 28W4	1199.0	1217.0	17.5	14-13-36 01W5	1139.0	1157.0	18.0
13-19-34 28W4	1204.7	1222.3	17.5	06-14-36 01W5	1166.0	1203.0	36.0
10-20-34 28W4	1200.0	1215.5	15.5	06-15-36 01W5	1170.0	1198.8	28.8
12-21-34 28W4	1187.5	1206.0	18.6	08-15-36 01W5	1172.0	1203.0	30.0
03-28-34 28W4	1187.0	1204.5	11.0	14-15-36 01W5	1173.0	1187.5	14.5
03-29-34 28W4	1191.0	1208.5	17.5	16-15-36 01W5	1175.0	1186.7	11.7
10-29-34 28W4	1188.0	1210.0	21.5	06-16-36 01W5	1164.0	1197.0	33.0
05-32-34 28W4	1191.0	1209.3	18.3	16-16-36 01W5	1174.0	1192.0	18.0
15-30-34 28W4	1188.0	1208.4	20.4	05-23-36 01W5	1163.0	1178.0	12.6
04-34-34 28W4	1157.0	1173.0	16.3	16-23-36 01W5	1119.0	1137.2	18.2

Appendix I : Core Listing

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