Allostratigraphy, Paleogeographic Evolution and Accommodation Controls of the Lower Colorado Allogroup in West-Central Alberta, Canada (Western Canada Foreland Basin)

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Abstract

The Lower Colorado allogroup of the Western Canada Foreland Basin is a mud-dominated rock succession of marine, estuarine, and coastal plain depositional origin. Relative sea-level oscillations produced bounding discontinuities that allow the allogroup to be divided into four alloformations: the Joli Fou, Viking, Westgate, and Fish Scales.

Rocks of the Lower Colorado allogroup within the study area record marine depositional environments represented by 9 facies. The facies represent depositional environments ranging from offshore marine to lower shoreface through to upper shoreface and river mouth environments.

Geometric analysis of the isopach maps of the regional allomembers of the Lower Colorado allogroup has allowed the recognition and differentiation of the tectonic and eustatic contributions to the generation of accommodation patterns within the study area. Accommodation for the Joli Fou alloformation and allomembers VA and VB of the Viking alloformation were determined to be the result of eustatic changes, while allomembers VD, WB, WC, and FB were determined to be the result of tectonism.

Analysis of Viking allomembers VA, VB1, and VB2 revealed multiple internal surfaces that lead to the identification and mapping of sand bodies within the Viking allomembers. This identification led to the creation of a hypothetical sea-level curve that illustrates how sea-level changed during the Viking from the transgression of VE0 to the transgression of VE3.

Keywords
Allostratigraphy, Western Canada Foreland Basin, Lower Colorado Group, Viking Formation, Correlation, Tectonics and Eustasy, Isopachs, Accommodation Controls.
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Chapter 1: Introduction

1.1 Introduction

The Lower Colorado allogroup of the Western Canada Foreland Basin is a muddy-dominated succession deposited in marine, estuarine, and coastal plain settings. Relative sea-level oscillations produced regionally-mappable surfaces that allow the allogroup to be divided into four alloformations: the Joli Fou, Viking, Westgate, and Fish Scales. With the application of an allostratigraphic interpretation to subsurface data, the physical bounding discontinuities, can be traced and correlated in cross-section. The correlation of the discontinuities allows the interpretation of allogenic controls by determining the subsurface geometry of allostratigraphic units.

1.2 Statement of Problem

The Western Canada Foreland Basin has been the subject of many stratigraphic analyses in an effort to determine basin history, with particular emphasis on mapping petroleum source and reservoir rocks. The WCFB is a component of the Western Canada Sedimentary Basin. The entire sedimentary succession in the WCFB contains multiple large petroleum systems that comprise multiple source and reservoir units, including the Viking Formation reservoir. Many field-scale studies (e.g. Downing and Walker, 1988; Boreen and Walker, 1991; Posamentier and Chamberlain, 1991; Walker and Wiseman, 1993; Bartlet, 1994; amongst others) have focused on localized sandbodies in the Lower Cretaceous (U. Albian) Viking Formation that are hydrocarbon rich. Although many of these previous studies used an allostratigraphic method, they were not linked into a coherent regional allostratigraphic framework, such as the regional master surfaces of Roca et al. (2008). A regional allostratigraphic analysis is important because it will allow ‘genetic’ rock units to be mapped, regardless of rock type. Stacking patterns of internal units can be analyzed to determine the history of relative sea-level changes. The ultimate aim of this study is to determine the relative roles of regional tectonic deformation, global (eustatic) sea-level change, and locally-variable sediment supply as controls on deposition and erosion. The relationship between the Viking sandstone and the Pelican
sandstone to the north may also be defined. The project forms part of a basin-scale study focused on recognizing and correlating Cretaceous eustatic events on a global basis.

1.3 Research objectives

The research conducted for this thesis has four main objectives:

1) To establish a set of bounding surfaces that spans all the previous studies in the area so as to allow interpretation in a consistent temporal framework.

2) To determine the geometry of Colorado Group rock bodies (Joli Fou, Viking, Westgate).

3) To understand depositional processes, environments, and paleogeography of the Lower Colorado allogroup.

4) To distinguish tectonic from eustatic controls on sea-level change.

1.4 Study Area

The study area of this project encompasses approximately 26,000 km² of north-central Alberta, Canada (Figure 1.1). It spans Townships 40 to 60 and Ranges 8W5 to 16W4. The western, northern, and southern boundaries of this study area are shared with those of related studies: Roca et al. (2008) to the west, Vannelli (2016) to the north, and Drjlepan (in progress) to the south.

1.5 Database

A total of 1,082 gamma ray and resistivity logs were used to construct a grid of 20 cross-sections that are oriented north-south and west-east. Cross-sections were constructed on every even-numbered Township and Range. This was done to match the spacing of cross-sections used by Vannelli (2016) and Drjlepan (in progress) so correlations could be extended into each study area. Sedimentological analysis was based on detailed logs of 52 cored sections with a cumulative thickness of approximately 1130 m.
Figure 1.1: Map of study area in north-central Alberta. The locations of well logs and cored wells used to develop regional cross-sections are indicated by the small circles. Locations of oil fields have also been included.
1.6 Methods

Borehole geophysical logs, specifically gamma ray and resistivity logs, were used to construct a regional grid of cross-sections, ten cross-sections (Appendix A) running North-South and ten running East-West (Figure 1.2). Gamma ray logs measure the natural radioactivity of rock units. Potassium-40 emits a gamma ray when it spontaneously decays to Argon-40 and the signal is measured by a gamma ray spectrometer as it moves vertically in a borehole. Potassium is a common element found in clay-rich sediments. Therefore, mudstone has a high gamma ray measurement whereas non-feldspathic sandstones result in lower radioactivity readings. Gamma ray logs are calibrated in API units scaled from 0-150. The resistivity log is generated by inverting the signal from the dual induction log that measure the bulk conductivity of the formation. This is of particular importance because a rock unit saturated with hydrocarbons will have a high electrical resistivity whereas a rock unit saturated in brine has a very low resistivity (other factors such as the presence of pore lining clays can also affect the conductivity of a formation). The geophysical logs, supplemented by cores, allowed the recognition and correlation of regional bounding surfaces, and the construction of an allostratigraphic framework (Figure 1.3). Analysis of physical and biogenic structures in core permitted determination of both physical processes and the paleoecology of the infauna, from which depositional conditions (turbidity, salinity, sedimentation rate, depth, etc), could be determined (Chapter 5). Measurements of allomember thicknesses allowed the creation of detailed isopach maps of each allomember (Chapter 6), which revealed evolving patterns of subsidence. Detailed correlation of parasequences within each allomember revealed internal organization and stacking patterns and allowed paleogeographic reconstruction (Chapter 7).

1.7 Overall significance of this study

The application of an allostratigraphic interpretation to the subsurface data will show the temporal relationship between the Joli Fou, Viking, and Westgate alloformations. The information will then allow an interpretation of the factors that controlled the deposition and preservation of each rock body, and in turn, inferences about changes in accommodation that may have been due to tectonic subsidence or uplift or possibly only eustatic sea-level change. Various subsidence and
Figure 1.2: Base map of study area with location of cross-sections. N-S cross-sections are labelled A through E and W-E cross-sections are labelled 1 through 6.
Figure 1.3: Reference well for the allostratigraphic framework used in this study.
erosional mechanisms generate different accommodation patterns and therefore the geometry of
the sediment body can be used to infer the process that controlled subsidence (Plint et al., 2012).

The fact that a foreland basin is loaded on only one side results in the subsidence gradually
decreasing away from the tectonic load. This asymmetry results in the units being deposited with
an overall wedge-shape that is thickest at the loaded side (Plint et al., 2012). Eustatic oscillations,
in contrast, produce stratigraphic units with a tabular geometry that results from the spatially
uniform change in accommodation (Plint et al., 2012).

This study, in conjunction with numerous other studies conducted at the University of
Western Ontario (e.g. Rylaarsdam (2006), Zhang (2006), Roca (2007), Buckley (2009), Vannelli
(2016), and Drljepan (PhD in progress)), will allow a palaeogeographic interpretation of not only
this study area but that of over 300,000 km² of Alberta and British Columbia. The large-scale study
has shown that four regionally-mappable marine transgressive surfaces can be correlated through
Alberta and BC with high confidence (Plint et al., 2016).
Chapter 2: Tectonic Setting and Controls on Sedimentation

2.1 Introduction

This chapter summarizes the tectonic setting in which a foreland basin is formed and the processes that control the sedimentary fill of the basin. This chapter also includes a description of the varying scales of cyclicity in the stratigraphic record and how recognizable stratigraphic surfaces are formed and preserved. A focus will be placed in the Western Canada Foreland Basin.

2.2 Foreland Basins

The use of the term ‘foreland basin’ became commonplace in geology when plate tectonic setting served as the basis for the classification of sedimentary basins (Dickinson, 1974). The introduction of plate tectonics in the 1960s made it much easier to explain the mechanisms of crustal deformation within a sedimentary basin. Peripheral foreland basins and retroarc foreland basins were named to differentiate the two tectonic settings in which they were formed (Dickinson, 1974). The term retroarc foreland basin describes a basin formed behind (cratonwards) the compressional arc on the upper, continental plate, whereas a peripheral foreland basin forms on top of the plate being subducted (Ingersoll, 1988) (Figure 2.1).

Retroarc foreland basin systems form adjacent to Andean-style orogenic belts associated with a rapidly converging oceanic-continental subduction zones (DeCelles, 2012). The thickened crust and increased mechanical load imposed by the accretion of terranes translates cratonwards and causes two flexural deflections of the continental lithosphere: the foredeep and the forebulge (Turcotte and Schubert, 2002). The foredeep is located adjacent to the orogen and is a wedge shaped trough that thins toward the craton, whereas the forebulge is a broad, subtle crustal uplift that has a location dependant on the flexural strength of the lithosphere (DeCelles and Giles, 1996). The distance between orogen and forebulge can vary from 100-400 km from the load, with a stiffer plate producing a more distant forebulge (DeCelles, 2012). As the rate of loading increases from the accretion of exotic terranes, the
Figure 2.1: Retroarc Foreland Basins involve a combination of static loads with an additional dynamic slab load overlying a mantle wedge that is being dragged downwards by the subducting plate from a ‘slab-driven’ effect where the downward circulating mantle wedge material entrained by the subducting plate couples with the base of the continental plate, resulting in subsidence of the upper continental plate (modified from DeCelles and Giles, 1996).
flexure in the foredeep increases; the greater flexure also results in greater uplift in the forebulge.

Retroarc foreland basins can also experience broad regional subsidence due to dynamic subducted slabs. Figure 2.1 (ii) depicts this ‘slab-driven’ effect where the downward circulating mantle material, entrained by the subducting plate, couples with the base of the continental plate to result in subsidence of the upper continental plate (Gurnis, 1992). When a foreland basin formed from the subduction of an oceanic plate beneath a relatively older (and colder) strong continental lithosphere, the load is distributed over a wider area and generates a broader basin with less local subsidence (Watts, 1992). With this rheology, the lithosphere behaves as an elastic solid. Under elastic conditions, the stresses cannot be dissipated and the morphology of the downwarped zone created by the tectonic load does not vary through time (Jordan, 1981). Conversely, a weaker and younger (and hotter) lithosphere will produce a foreland basin that is deeper and narrower (Watts, 1992). Under these conditions, the lithosphere displays a viscoelastic behaviour, which allows the relaxation of tectonic stress and causes the forebulge to migrate towards the orogen through time (Beaumont, 1981).

Ultimately, Quinlan and Beaumont (1984) suggest that the best models of foreland basin generation that reflect the vertical change in temperature of the lithosphere consider the uppermost part of the lithosphere to behave elastically whereas the deeper regions of the lithosphere have a viscoelastic behaviour.

DeCelles and Giles (1996) described a foreland basin as a system composed of four depozones aligned parallel to the orogen (Figure 2.2). Progressing outward from the orogen, the depozones are named the wedge top, foredeep, forebulge, and back-bulge. The depozones are characterized with regards to their proximity to the orogen, subsidence rate of each depozone, and structure of the basin (DeCelles and Giles, 1996). As subduction progressively compresses and thickens the orogen, the deformation front advances, leading to corresponding migration of the thrust belt (DeCelles and Gilles, 1996). Understanding the lateral movement of the thrust belt-foreland basin system allows interpretation of the foreland basin.
Figure 2.2: (i) Schematic view of a representative foreland basin. Vertical line represents cross-section as depicted in ii. (ii) Geometry of foreland basin in transverse cross-section view. This cross-section is represented with much vertical exaggeration. (Modified from DeCelles and Giles, 1996)
stratigraphic record in regards to how tectonics and sedimentary environments evolved over time (Flemings and Jordan, 1989).

In the wedge-top depozone, sediment accumulates on and between active thrust sheets (Figure 2.2 ii). Sediment in this depozone is typically compositionally and texturally immature because of short transport distances and steep gradients (DeCelles and Giles, 1996). Continued advance of the fold-and-thrust belt results in uplift of the wedge-top depozone; resulting in short sediment residence times and the generation of unconformities due to the gradual tectonic cannibalization of the basin fill (Ori and Friend, 1984).

In the foredeep depozone, subsidence is primarily a response to loading by the adjacent orogenic wedge coupled with the isostatic load imposed by the mass of sediment derived from the orogen (Beaumont, 1981). Within a foreland basin, the foredeep is the area with the most rapid subsidence and therefore potentially the greatest amount of sediment accumulation and preservation (the foredeep can accumulate 2-8 km of sediment; DeCelles and Giles, 1996). Located cratonward of the foredeep, the forebulge depozone is an area characterized by low subsidence rate or subtle uplift that can lead to subaerial exposure and erosion (Crampton and Allen, 1995). The uplift of the forebulge is typically less than 5% of the amount of subsidence in the foredeep and the position and elevation of the forebulge is controlled by both the supracrustal load (Figure 2.3) and flexural strength of the lithosphere. The position of the forebulge can be traced by mapping the distribution of the forebulge unconformity through time (Plint et al., 1993). For example, in the Western Canada Foreland Basin, the trace of the forebulge is marked by erosion and offlap/onlap patterns (Plint et al., 1993).

The back-bulge is the most distal depositional setting and is located between the forebulge and the craton. The back-bulge depozone is relatively broad and experiences much less subsidence than the foredeep (DeCelles and Giles, 1996). The sediment in the back-bulge basin may be derived from the orogenic belt and the craton (DeCelles and Giles, 1996).
Figure 2.3: Diagram to illustrate the viscoelastic lithosphere response to tectonic loading. As the lithosphere relaxes, the load subsides more (1-3) and the forebulge undergoes progressive uplift and moves towards to the orogen. (Modified from Beaumont et al., 1993).
2.3 Western Canada Foreland Basin

The tectonic processes described above can be used to explain the history of subsidence of the Western Canada Foreland basin. The evolution of the basin is closely linked to the geological history of the Rocky Mountain thrust belt (Price, 1973). There are two distinct episodes of development of the orogenic wedge: 1. The Columbian Orogeny, which occurred during the Late Jurassic to Early Cretaceous and 2. The Laramide Orogeny, which occurred during the Late Cretaceous to Paleocene (Douglas et al., 1970).

The Western Canada Foreland Basin consists of a foredeep depozone that is thickest adjacent to the Rocky Mountains and thins eastward. The foredeep is filled with sediment supplied from the orogenic wedge. The back-bulge depozone is filled with sediment from the craton, the forbulge, and the orogen (Figure 2.4) (Armstrong, 1968; Porter et al., 1982). The WCFB extends almost 2000 km parallel to the Cordillera and extends up to 900 km perpendicular to the thrust belt (Wright et al., 1994). The foredeep accumulated a maximum thickness of 6 km in the fold-and-thrust belt and the sedimentary wedge onlaps the Precambrian crystalline basement in the north-east (Wright et al., 1994). The wedge-top depozone is very poorly preserved in Canada because of post-orogenic uplift and erosion.

2.4 Cyclicity and Sequences in the Stratigraphic Record

2.4.1 Introduction to Sequence Stratigraphy

Within the stratigraphic record of any particular basin, there are many types of sequences. The duration and periodicity of depositional sequences have been shown to span at least sixteen orders of magnitude ranging from daily tidal cycles to 100 million year continental cycles (Miall, 2010).

The sequence stratigraphic method of interpreting seismic sections developed by Peter Vail (1977) and others at Exxon started a revolution in stratigraphy. The fundamental stratigraphic unit defined by Vail (1977) was the depositional sequence. Vail (1977) defined a
Figure 2.4: Geological sections across the Western Canada Foreland Basin, east of the thrust belt (from Wright, 1984). Approx. 40 times vertical exaggeration
sequence as a, “relatively conformable succession of genetically related strata bounded at its top and base by unconformities”. The unconformities, or bounding discontinuities, form as a result of relative changes in sea level and delineate packages within which sediments are approximately coeval.

The lowest-frequency sequences preserved in the stratigraphic record are the product of large-scale plate tectonic processes and span hundreds of millions of years. The highest-frequency stratigraphic sequences are in the Milankovitch-band, which span of the order of 20-400 kyr. The large range of cycle types reflect the independent operation of at least four types of processes, including regional tectonism and various controls on eustasy (Miall, 2010). In their representation of a global sea level curve (Figure 2.1), Vail et al. (1977) defined three scales of cycles in the stratigraphic record; first-order, second-order, and third-order. First-order cycles range from 200-400 million years in duration and have been interpreted as the global supercontinent cycle. Second-order cycles range from 10-100 million years in duration, and 3rd-order cycles range roughly 1-10 million years in duration. Vail et al. (1977) also hypothesized causes for the variation in cycle duration. Geotectonic mechanisms appeared to be the only factor capable of spanning a timescale long enough for first-order cycles and some second-order cycles whereas glaciation and deglaciation were thought to explain many third-order cycles and some second-order cycles.

2.4.2 Supercontinent cycle: Cycles Greater than 100 Million years

The recognition of the supercontinent cycle is arguably one of the most important advances in earth science since the plate tectonic theory (Nance and Murphy, 2013). Earth’s history is seen as having been punctuated by the periodic assembly and breakup of supercontinents that have influenced the rock record more than any other phenomena (Nance and Murphy, 2013). Vail et al. (1977) referred to this cycle type as “first-order cycle” and postulated that geotectonic mechanisms had sufficient duration to explain the length of cycle. First order sequences are now regarded as the product of changes in sea level due to the long-term cycle of supercontinent assembly and dispersal.
Figure 2.5: Original sea Level curve presented by Vail et al. (1977). Shown on the left is the long term first-order cycle and on the right are second- and third-order cycles.
### 2.4.3 Sloss sequences: 10-100 Million year Events

Sequences spanning 10–100 million years can be traced and correlated among several of the Earth’s major continental interiors, including the interior of the United States, Canada, Russia, and Brazil (Miall, 2010). Vail et al. (1977) referred to these sequences as the product of second-order cycles but they are now commonly called *Sloss sequences*, after L. L. Sloss, who in 1963, first identified and described six sequences of North America. Hallam (1963) attributed Sloss sequences to eustatic changes in sea level in response to volume changes of oceanic spreading centres, based on his summary of Late Cretaceous and Cenozoic events. However, Miall (2010) suggested that eustatic changes cannot explain all the features of the major Sloss sequences because they are commonly separated by angular unconformities and many of the sequences contain thick continental deposits. A passive rise in sea level would terminate widespread nonmarine deposition; therefore, additional processes must be involved (Miall, 2010). It is believed that sequences spanning tens of millions of years may be the result of regional tectonism and have hence been named tectostratigraphic sequences (Miall, 2010). These sequences are the product of various tectonic processes including plate rifting, fragmentation, thermal subsidence, as well as global processes including eustatic sea-level change and long-term climate change (Miall, 2010).

### 2.4.4 Million Year Cycles

First described by Vail et al. (1977) as third-order cycles, detailed stratigraphic and facies studies of Phanerozoic strata have generated an abundance of information concerning cyclicity over time scales of 1–10 million years. A wide variety of techniques have been used in an attempt to reconstruct a chronostratigraphic record of this type of cycle. There are the methods of Vail et al. (1977) and Haq and Schutter (2008) who applied their seismic and sequence stratigraphic techniques to delineate million year sequences on a global sea-level curve. Other studies have employed detailed stratigraphic reconstructions emphasizing lithofacies or biofacies data (Miall, 2010).

Million-year sequences are preserved in a wide range of basins and tectonic settings including continental margins, arc-related basins, and foreland basins. The million year cycles
are largely the product of tectonics. Tectonic and eustatic cycles generate unconformities through differing forcing mechanisms and are preserved as sequence boundaries (Figure 2.6) (Jordan and Flemmings, 1991). Subaerial erosion of proximal settings takes place during tectonic inactivity due to flexural rebound (Figure 2.6 i). The opposite occurs during active thrusting when distal settings are uplifted and eroded (Jordan and Flemmings, 1991). Both of these tectonically related unconformity surfaces are diachronous and are separated by a correlative conformity (Jordan and Flemmings, 1991). Eustatic fall, however, generates synchronous proximal and distal subaerial unconformities unless the rate of subsidence exceeds the rate of eustatic fall. Such a case is depicted in Figure 2.6 ii.

The WCFB provides examples, such as the recent Buckley et al. (2016) study, that show million year cycles in the stratigraphic record. The Buckley et al. (2016) study of the Harmon and Cadotte alloformations show that tectonism seemed to control subsidence at the 1-2 myr scale while eustasy produced internal parasequences on a timescale of 500 kyr.

### 2.4.5 <1 Million Year Sequences

High frequency sequences span from $10^5$ to $10^3$ years. These types of sequences can be classified on the basis of their sedimentological character, tectonic, or climatic setting and may be the result of more than one, possibly several, interacting mechanisms (Miall, 2010). Typically high frequency sequences are considered to be the product of climatic processes driven by orbital cycles; these are also referred to as Milankovitch processes. These cycles result from astronomical cycles, such as the change in eccentricity of the Earth’s orbit, the tilt of the Earth’s axis, and the precession of the equinoxes (Bennett, 1990). The duration of these cycles are approximately 400 and 100 ky, 40 ky, and 20 ky respectively. Glacioeustasy is the single most important result of Milankovitch processes, but the importance of “non-glacial Milankovitch cyclicity” is being increasingly recognized throughout the geological record (Miall, 2010). Cycles of loading and unloading due to contractional tectonism, crustal flexure, uplift, and erosion may generate changes in accommodation on a timescale < 1 million years (Miall, 2010). Using the Western Canada Foreland Basin as an example, there is evidence from the Cretaceous stratigraphic record of both climatic and tectonic forcing of accommodation and sedimentation.
Figure 2.6: Sequences developed by tectonism and eustasy in a foreland basin setting. (i) A1 and B1 represent rapid tectonic subsidence that led to vertical aggradation of sequences. A2 and B2 represent a time of no tectonic subsidence and the stacking pattern becomes progradational. Unconformity 1 is formed at end of A2 time due to erosion. Unconformity 2 is formed at B1 due to uplift of forbulge. (ii) An example of constant subsidence and sea level fall. Dotted line represents unconformity formed during sea level fall but it only manifests near forbulge in the area of slow subsidence. Proximal to the orogeny, the subsidence rate is faster than eustatic fall. The sequence boundary is a correlative conformity (Modified from Jordan and Flemings, 1991).
on a high-frequency timescale. The large amount of detailed data from this area provides one of the best areas in the world to study the relationships between tectonics, eustasy, climate forcing, and sedimentation.

Detailed studies on the Cardium Formation (Plint et al., 1986), as well as the Viking Formation (Boreen and Walker, 1991), divided the formations using allostratigraphic methods based on the recognition and mapping of major bounding surfaces. Originally considered controversial, the recognition of this type of stratigraphic method presented by Plint et al. (1986) started a major new approach to understanding the stratigraphy of the foreland basin geology. These ideas are expanded on in section 3.3.2 of this thesis.

2.4.5 System Tracts Model

Posamentier and Vail (1988) built upon the sequence stratigraphic method by modelling the interaction of an invariant tectonic subsidence rate with a sinusoidal eustatic curve. Posamentier and Vail (1988) showed that bounding discontinuities were related to eustatic fall inflection points and would produce type 1 or type 2 unconformities, depending on the rate of relative sea level fall. Stream rejuvenation and fluvial incision, sedimentary bypass of the shelf, and a basinward shift of coastal onlap characterize the type 1 unconformity, whereas the type 2 unconformity is not characterized by stream rejuvenation. Type 2 is characterized by a basinward shift of coastal onlap and slow, widespread subaerial erosion accompanied by gradual degradation of the landscape (Posamentier and Vail, 1988). The occurrence of a type 1 or 2 unconformity would depend on whether the rate of eustatic fall exceeds or is less than the rate of subsidence at the depositional shoreline break. Since its introduction, the original sequence model has evolved and only type 1 sequences are currently recognized (Posamentier and Allen, 1999).

Posamentier and Vail (1998) used a sinuous curve to represent the relative position of sea level. The function of this curve is to illustrate sediment supply (S) and accommodation (A). Sediment can be deposited in an area where there is enough space, or accommodation, available. This space can be added or removed by factors such as a change in sea-level or
isostatic subsidence. When the rate of accommodation being created is greater than the supply of sediment filling the space, there is a landward migration of the shoreline, or transgression. When the rate of sediment supply is greater than the rate at which space is being created, there is a regression and the shoreline migrates basinward. The accommodation made available for the sediment is the sum of tectonic subsidence or uplift and eustatic sea-level change, which combine to control relative sea level (Posamentier et al., 1988). Posamentier and Vail (1988, Figs. 1-6) outline the framework for their systems tract model using a selection of coastal environment diagrams accompanied with a eustasy curve. Progression through the systems tracts is as follows: highstand systems tract (HST), lowstand systems tract (LST)- basin floor fan, LST- lowstand wedge, Transgressive systems tract (TST), HST II, and shelf margin systems tract.

Evolution of the systems tract model modified the scheme to have four components: LST, TST, HST, and FSST (Figure 2.7). The beginning of a sequence takes place with the deposition of the LST when the rate of sediment supply is greater than that of the accommodation being created. There is a normal regression of the shoreline as it migrates into the basin. When the relative sea level rises and the accommodation rate begins to exceed the sedimentation rate, deposition of the TST begins and the shoreline migrates back towards land. The HST occurs when the rate of sediment supply equals and then exceeds the accommodation rate at the shoreline until the point that relative sea level reaches its maximum. The final part of the sequence is the FSST (Plint and Nummedal, 2000), which is deposited between the highest and lowest point of relative sea level. During FSST, as sea level falls, a subaerial unconformity develops bounding the top of the sequence (Figure 2.8).

2.4.6 Important Stratigraphic Surfaces

An important aspect of the research project is to recognize and correlate significant stratigraphic surfaces that allow the mapping of distinct rock units. These surfaces are formed through the course of a sea-level cycle as shown in Figures 2.7 and 2.8. As the sea-level begins to rise, the shoreline stops prograding and begins to retrograde. As the shoreline moves landward, wave scouring produces a significant erosional surface called a ravinement surface.
Figure 2.7: Relative sea-level curve of complete sequence. S: supply; A: accommodation; SB: Sequence Boundary; LST: Lowstand systems tract; TS: Transgressive surface; TST: Transgressive systems tract; MFS: Maximum flooding surface; HST: Highstand systems tract; RSME: Regressive surface of marine erosion; FSST: Falling stage systems tract.
Figure 2.8: The four systems tract sequence model. A) Stratigraphic product of a relative sea level cycle. B) Chronostratigraphic chart projected from the stratigraphic cross-section (from Plint and Nummedal, 2000).
A ravinement surface is often represented in core by a pebble lag. The offshore equivalent of a ravinement surface is a flooding surface, which is a surface that establishes a time of an abrupt increase in water depth (Van Wagoner et al., 1988). In core, there would be a noticeable change in lithology of coarse grained units to finer grained units. Sea level continues to rise until the sedimentation rate equals the accommodation and transgression of the shoreline ceases. The maximum flooding surface demarcates the furthest landward extent of the shoreline and is the highest point that sea-level reaches (Catuneanu, 2006). The sedimentation rate is now greater than the accommodation rate and there is progradation of highstand sediments. As sea level begins to fall, rivers close to shore erode into previously deposited sediments to maintain an equilibrium profile with the sea level, creating valleys (Emery and Myers, 1996). The effects of sea-level change are limited to the downstream portion of a river and diminish upstream (Shanley and McCabe, 1994). Away from the valley, the sequence boundary may be marked by well-developed paleosols. For example, on the ancient Dunvegan coastal plain, nonmarine sequence boundaries manifest as interfluve paleosols that can be quite subtle and difficult to identify (McCarthy and Plint, 1998).

The sequence boundary is therefore a subaerial unconformity that includes both the interfluve surface and the fluvial erosion surface that underlies the valley (Catuneanu, 2006). During relative sea-level fall, wave scour on the inner shelf can produce a surface called the regressive surface of marine erosion (Plint and Nummedal, 2000). The lowest point of relative sea-level, or subaerial unconformity, represents a sequence boundary.
Chapter 3: Stratigraphic Framework

3.1 Introduction

This chapter summarizes the lithostratigraphic framework of the Lower Colorado Group within the WCFB as well as briefly summarize coeval strata in adjacent areas of Alberta where different terminology has been applied to equivalent formations throughout the basin.

3.2 Lithostratigraphy

3.2.1 Introduction

The mid Cretaceous Colorado Group consists predominantly of mudstone interstratified with sandstone and conglomerate units (Leckie et al., 1994). Other lithologies present in this group include bentonite, horizons of fish debris, siderite, and chalky limestones (Leckie et al., 1994). Rocks of the Lower Colorado Group record epicontinental marine and paralic deposition in the Western Interior Seaway. At the time of deposition, global sea level was high, with sea-level maxima occurring specifically in the late Albian, early Turonian, and middle Santonian (Caldwell, 1984; Haq et al., 1987). Tectonism and regional down-flexing of the North American craton was active during deposition at the time of Colorado Group (Lambeck et al., 1987).

3.2.2 The Lower Colorado Group

Rudkin (1964) informally subdivided the Colorado Group into upper and lower subgroups separated by the Base of Fish Scales Zone. The Lower Colorado Group, the main focus of this study, is of lower Cretaceous late Albian age. The stratigraphic and geographic extent of the formal lithostratigraphic units assigned to the Albian rocks in this study, and coeval strata in the closest surrounding areas, are summarized in Figure 3.1. The Colorado Group overlies the Mannville Group in southern Alberta and the Fort St. John Group in northeastern British Columbia and northwestern Alberta. The contact between the Mannville
and the Lower Colorado Group is generally represented by a thin conglomerate (Leckie et al., 1994). In southeastern Alberta there is a thin (< 5 m thick) unit named the Basal
Figure 3.1: Lithostratigraphic chart of the Colorado Group in the regions of Alberta in closest proximity to the study area (Modified from Leckie et al. (1994) and Roca et al. (2008))
Colorado Sandstone that occurs as a northwest-trending body, approximately 100 km wide, and situated in a topographic low (Banerjee, 1989).

The Joli Fou Formation represents the onset of Lower Colorado deposition throughout a large extent of the basin (Stelck, 1958). The lithology of the Joli Fou Formation is dark grey, noncalcareous marine mudstone with small proportion of interbedded very fine- to medium-grained sandstone (Leckie et al., 1994). U/Pb ages from Perovskite crystals in kimberlite erupted during the deposition of the Joli Fou date the formation to 103 ± 1.0 Ma (Zonneveld et al., 2004).

In other areas of Alberta, the Joli Fou has been correlated to be depositionally time equivalent to the Lower Paddy alloformation in the northwestern plains of Alberta (Vannelli et al., 2017) and to pass laterally into the Bow Island Formation in the southern plains of Alberta (Mellon, 1967). The Joli Fou has a type section along the Athabasca River (Wickenden, 1949) but it is not known to occur in outcrop in the Rocky Mountain Foothills.

The Viking Formation was first named by Slipper (1918) who described a gas-rich sandstone encased by two mudstone units. Sandstones of the Viking Formation average thickness of 15 to 30 m in central Alberta but thicken to over 75 m in southern Alberta and thin eastward towards Saskatchewan where the unit eventually pinches out (Reinson et al., 1994). The Viking Formation has been shown be partially time equivalent to the Bow Island Formation in southwestern Alberta, the Pelican Formation in northeastern Alberta, and the Paddy Member of the Peace River Formation in northwestern Alberta (Leckie et al., 1994; Roca et al., 2008). The Viking Formation consists of interbedded mudstone, fine- to coarse-grained marine sandstone and conglomerate that form a series of upward-coarsening sequences (Reinson et al., 1994). Sandstones and conglomerates of the Viking Formation provide reservoirs for significant oil and natural gas resources in Alberta and. A detailed summary of the history of investigations of the Viking Formation and research into the depositional mechanisms of the unit is discussed in section 3.2.2.
The Westgate Formation, originally named by McNeil and Caldwell (1981) in Manitoba for the mudstone unit between the Newcastle and Belle Fourche members, was extended westward by Bloch et al. (1993) to include the shale unit lying stratigraphically above the Viking Formation. In the northwestern plains of Alberta, the Shaftsbury Formation of the Fort St. John Group is a mudstone succession coeval with the Westgate Formation. The Westgate Formation has been interpreted to represent offshore environments of the Mowry Sea (Caldwell, 1984).

The Fish Scales Formation, and specifically the Fish Scales Marker at the base of the Formation, is a highly radioactive unit of rock that contains abundant fish scales and skeletal material that is interbedded with finely laminated, generally nonbioturbated sandstone and siltstone with local zones of pebbles and nodular phosphorites (Leckie et al., 1994). The Fish Scales Formation is the lowest unit of the Upper Colorado Group and the base of the Albian/Cenomanian boundary (Leckie et al., 1994).

Figure 3.2 shows an isopach map of the Lower Colorado Group and approximate equivalents from the Base of Fish Scales to the top of the Mannville throughout northeastern British Columbia, Alberta, and Saskatchewan. The isopach shows two distinct depocentres along the thrust belt. The northern depocenter is located in the Peace River area of northeastern British Columbia and northwestern Alberta and the southern depocenter extends across the southern part of the Alberta and Saskatchewan (Leckie et al., 1994).

3.3 Previous Studies

3.3.1 Introduction

The following section provides a brief history of studies of the Viking Formation. Different interpretations of the origin of Viking depositional environments and sedimentary processes are discussed herein. There will also be a brief review of the development of an allostratigraphic scheme of the Viking Formation and how those specific studies led to the research presented here.
Figure 3.2: Isopach map of the Lower Colorado Group and equivalent strata from the Base of Fish Scales to the top of the Mannville Group. The study area for this project is represented by the red boundary line. Note that the study area is in an area of little thickness variation while to the northwest and southeast of the study area there are two distinct depocentres. (Modified from Leckie et al. (1994)
3.3.2 Previous Interpretations of the Viking Formation

The depositional environments in which the Viking Formation was deposited have been debated since the early 1900s. The studies of Beach (1955) and DeWeil (1956), made environmental interpretations based on a lithostratigraphic model by observing sedimentary structures seen in core. Beach (1955) postulated that the sandstones, in particular the chert pebble units, of the Viking Formation were too coarse to have been deposited by pelagic sedimentation and therefore must have been deposited by turbidity currents. This hypothesis was strongly contested by DeWeil (1956) who argued that the hypothetical slope that formed the turbidity currents was not theoretically possible using evidence from the Grand Banks event and those of the Viking sea. DeWeil (1956) determined that Viking sands were, in fact, deposited in a shallow marine setting, which was supported by evidence of upward-coarsening successions and wave-produced sedimentary structures. A storm enhanced tidal current model of deposition was proposed by Evans (1970) who interpreted the Viking sediments as deposits of migrating tidal ridges. This idea was supported many years later by other works such as Leckie (1986) and Bartlett (1994). Beaumont (1984) suggested that deposition of the Viking sand bodies was influenced by relative changes in sea level. Prograding Viking deltas were then reworked into linear shelf sand bodies by subsequent transgressions. The work of Beaumont (1984) was supplemented by results from Hein et al. (1986), Leckie (1986), and Walker and Wiseman (1994) who interpreted the Viking deposits to represent a prograding shoreline formed during relative lowering of sea level, followed by a rise in sea level, which resulted in the deposits being reworked during transgression.

The importance of sea level fluctuations became an important element of environmental interpretations in the 1980s. The study of Plint et al. (1986) on the Cardium Formation, was the first to emphasize the importance of erosional surfaces caused by the rise and fall of sea level. This led to the introduction of an “incised shoreface model” (e.g. Plint, 1988), which was subsequently applied to the Viking Formation by Downing and Walker (1988), who described an erosive-based, 38 km long sandstone body at Joffre field, and interpreted it to have been a shoreface. This form of sequence stratigraphy led Boreen and Walker (1991) to
make the first attempt to establish a sub-regional allostratigraphic scheme based on the
correlation of laterally traceable surfaces, which represent stratigraphic discontinuities (North
American Commission on Stratigraphic Nomenclature, 2005). In the present thesis, the formal
lithostratigraphic definition of the Lower Colorado Group becomes the informal
allostratigraphic definition of the Lower Colorado allogroup.

Other works such as Davies and Walker (1993), Pattison and Walker (1994), and Walker
and Wiseman (1995) used the same allostratigraphic framework outlined by Boreen and Walker
(1991). The framework was developed by correlation of erosional surfaces within the Viking
strata named VE1, VE2, VE3, and VE4 divided the Viking alloformation into five “allomembers”
named A through E (Figure 3.3).

Boreen and Walker (1991) described allomembers A and B, separated by erosional
surface VE1, as collectively comprising three to five regionally-extensive sandier-upward
successions. The number of preserved successions depends on the degree of erosion by basal
surface of the following succession. Allomembers A and B were interpreted by Boreen and
Walker (1991) to represent deposition in an offshore marine realm below fair-weather
wavebase. Allomember C was interpreted to represent several generations of paleovalley fills
of estuarine muds, sands, and conglomerates. The erosional surface VE2 (base of allomember
C) cuts into allomembers A and B and was interpreted to be a surface of subaerial erosion that
was incised during sea level fall. VE3 is a marine transgressive surface that caps allomembers A,
B, and also C. In the Willesden Green field, the majority of the oil and gas producing sand
bodies belong to allomember C (Pattison, 1991; Pattison and Walker, 1994). Allomember D is
bounded below by VE3 and above by VE4. It consists of bioturbated mudstone and wave-
rippled sandstone, which indicate a prograding marine shoreface environment (Boreen and
into nonmarine facies in the scheme of Boreen and Walker (1991). VE4 and the Base of Fish
Scales marker define the bottom and top of allomember E respectively. Allomember E was
deposited during an overall transgressive period with several relative sea level fall and rise
Figure 3.3: Allostratigraphic scheme for the Viking Formation in the Crystal to Caroline area, after Boreen and Walker (1991).
cycles producing repeated episodes of progradation and transgression.

Pattison (1991) and Wiseman (1994) used the framework developed by Boreen and Walker (1991) and extended it into their own project areas. Pattison (1991) set out to determine the sedimentological and stratigraphic relationships between the regional, valley-fill, shoreface, and transgressive types of Viking deposits north of the Willesden Green Area. In addition, his intent was to establish how Viking sedimentation developed on a basinwide scale and to determine the effect of sea level fluctuations on the deposits. The regional deposits consisted of coarsening upward successions deposited in a progradational shoreline controlled by inferred glacio-eustatic fluctuations or differential subsidence. Viking valleys, formed during the VE2 lowstand, were filled by coarse-grained sediments during the ensuing VE3 transgression. These valley-fill deposits are present throughout the Viking areas including Crystal, Sunnybrook, Sundance, and Edson. Lastly, Pattison (1991) recognized several paleoshorelines deposited during the stillstands and regression that were then superimposed on the VE3 transgression. He interpreted that the relative sea-level cycles were controlled by 63,000-year cycles and as a consequence controlled the VE3 shoreline positions.

Wiseman (1994) identified three distinct shoreface sandstone bodies in a study area northeast of Edmonton: Lindbrook, Joarcam, and Beaverhill Lake (all of which are located in the study area of this thesis). Identification and correlation of distinct markers allowed for the reconstruction of the history of sea level changes in that area. Lindbrook was interpreted to have formed first as a lowstand shoreface, which was incised by Joarcam during a pause in ensuing transgression while the shoreline was moving landward (southwest). A second major lowering of sea level moved the shoreline northwest, causing shoreface incision at Beaverhill Lake (Walker and Wiseman, 1995). The surfaces correlated by Walker and Wiseman (1995) did not follow the VE nomenclature but instead used surface labels that designate a system tract (e.g. BHL RSE: Beaverhill Lake Regressive Surface of Erosion). However, these surfaces correlate with significant surfaces used in the present study.

Posamentier and Chamberlain (1991) conducted a sequence-stratigraphic analysis of the Viking Formation at the Joarcam Field. This study used wire-line logs to correlate three units
that contain hydrocarbons: the ‘upper’, ‘main’, and ‘third’ sands. These units correlate approximately with allomembers VD, VB, and VA used in the present study. Posamentier and Chamberlain (1991) hypothesized that the original description of the ‘main’ sand being a prograding beach deposit was incorrect and that, in fact, the ‘main’ sand could be split into two units: a lowstand systems tract and a transgressive systems tract, each with unique reservoir attributes (Figure 3.4). It was determined that the LST comprised upward-coarsening shoreface deposits isolated from and seaward of the underlying highstand shoreline and overlying SB (Posamentier and Chamberlain, 1991). The overlying TST comprises alternating offshore shales and storm deposited sandstones characterized by horizontal bedding, which is different than the shingled geometry of the LST deposit (Posamentier and Chamberlain, 1991).

Additional studies of the Joarcam Field, such as MacEachern et al. (1992), focused on the ichnology and sedimentology of transgressive Viking deposits. It was suggested that the transgressive deposits of the Viking Formation could be differentiated into six distinct facies: A-F. Facies A and B were assigned to the *Glossifungites* ichnofacies, which was commonly developed on the transgressive surfaces of erosion, cross-cutting the original softground trace fossil suit, Facies C and D reflected distal stillstand progradational cycles that occurred within the overall transgression, and Facies E and F correspond to deposition under conditions of rising sea level unrelated to progradational cycles. These Facies descriptions were compared to the facies descriptions of other studies of Viking transgressive deposits (Hein et al., 1986; Leckie, 1986; Downing and Walker, 1988; Boreen and Walker, 1991; Posamentier and Chamberlain, 1991) to establish an encompassing facies framework for the Viking Formation.

Taking the allostratigraphic framework developed by Boreen and Walker (1991) and that of other allostratigraphic interpretations that used the Boreen and Walker (1991) framework, Roca (2007) integrated the previous work into a new expanded framework. Roca (2007) mapped fifteen discontinuities within the Lower Colorado allogroup. Roca (2007) gave each discontinuity a synthetic name composed of three terms: 1. Alloformation (e.g. “V” for Viking, “W” for Westgate), 2. regressive (“E”) or transgressive (“T”), which denotes the origin
Figure 3.4: i) Interpretation of the ‘main sand’ as a northeasterly prograding shoreline resulting in deposition of a sheet sand across the Joarcam Field. ii) Interpretation of the ‘main sand’ as two separate units: the LST and TST. Redrawn from Posamentier and Chamberlain (1991).
of the discontinuity, and 3. the relative sea-level cycle (e.g. “3”). In order to preserve the original nomenclature defined by Boreen and Walker (1991), the areal extent of a composite discontinuity that developed through transgressive ravinement (T) of a pre-existing subaerial unconformity generated during the preceding regression (E) is referred as an “E” surface. Roca (2007) summarized his configuration of the Lower Colorado discontinuities as follows. JE0 was recognized as the lowest surface bounding the entire allogroup and FSU as the highest. The discontinuities from JE0 to WE2 consist of transgressive surfaces of regional extent, which Roca (2007) interpreted to be correlative with approximately equivalent nonmarine surfaces in outcrop. There are five basin-scale discontinuities that were generated during a series of relative sea-level falls. These surfaces are labelled from JE0 to VE4. Roca (2007) divided the Joli Fou alloformation starting with the basal allomember “Ja”, which is constrained to paleotopographic lows, and the top allomember as “JA”. Roca (2007) left the allomembers A through D as previously defined by Boreen and Walker (1991) but renamed them VA through VD to show that they belong to the Viking alloformation. A change was made to Boreen and Walker’s (1991) allomember E by raising it to alloformation status and it was assigned to the Westgate Formation following Bloch et al. (1993). The Westgate was then divided into four allomembers with a newly defined basal paleovalley fill “Wa” marking the bottom surface with three regional allomembers named WA through WC above it. Finally, two allomembers named FA and FB represent the Fish Scales alloformation and also the top of the Lower Colorado allogroup as a whole.
Chapter 4: Regional Allostratigraphy of the Lower Colorado Allogroup

4.1 Introduction

The North American Commission on Stratigraphic Nomenclature (NACSN) (2005) defines allostratigraphy as a method of mapping stratified rocks through the correlation of laterally traceable bounding discontinuities that can be interpreted as approximate timelines (NACSN, 2005). Allostratigraphy does not use lithological criteria to define allostratigraphic units and can therefore overcome possible geographical limitations of lithostratigraphic correlation. These limitations are caused by the lateral changes in facies distribution for a sedimentary depositional system.

The allostratigraphic scheme presented in this chapter is an extension of the framework developed by Roca et al. (2008), with some modifications.

4.2 Allostratigraphic Framework of this Study

The locations of summary cross-sections can be seen on the basemap in Figure 1.2. A total of 11 summary well-log cross-sections (Appendix A, cross-sections 1-6 and A-E), ‘exploded’ allomember cross-sections (Chapter 6) and summary diagrams, and two core cross-sections have been constructed to illustrate the allostratigraphic geometry. These summary cross-sections illustrate the correlation of significant bounding surfaces as well as the stratigraphic relationships of the allomembers throughout the study area. Five cross-sections are oriented in a north to south direction and are labelled A to E and six cross-sections are oriented west to east and are labelled 1 to 6. The cross-sections integrate a total of 185 gamma-ray/resistivity logs (distilled down from 1,082 wells). Bounding discontinuities observed in 29 core intervals have been correlated to nearby wells in various summary cross-sections.

The Base of Fish Scales Marker (Roca et al., 2008) was chosen as the datum for the summary cross-sections because of its regional extent and ease of recognition on well logs. The Fish Scales alloformation produces a large, sharp-based deflection of the gamma-ray log. In the
allostratigraphic framework, this prominent marker is labelled BFSM (Base of Fish Scales Marker).

4.2.1 Bounding Discontinuities

Relative changes in sea-level can generate regional discontinuities that define genetic packages of rock. Marine flooding surfaces, or transgressive surfaces, are considered to represent geologically instantaneous events (Van Wagoner et al., 1988). These surfaces can be correlated regionally to establish an allostratigraphic framework. The geometry of an allostratigraphic unit results from the interaction between several geological processes that control the evolution of accommodation with respect to area and time, the volume and rate of deposition, as well as post depositional erosion (Miall, 1991). Various subsidence and erosional mechanisms generate different accommodation patterns and therefore the geometry of the sediment body can be used to determine the processes that controlled the generation or removal of accommodation (Roca et al., 2008). An analysis of allomember geometry is presented in chapter 6.

The allostratigraphic framework of Roca et al. (2008), previously discussed in section 3.3.2, was used as the basis for the stratigraphic framework used in this thesis, subject to some modifications. Bounding discontinuities WE1 and FE1 are not included here. These surfaces pinch out through onlap in the field area of Roca et al. (2008) due to subsidence and greater amount of accommodation in the foredeep. Geographic limitations control the extent of WE1 and FE1 and therefore they cannot be traced in the study area of this project. Without the WE1 and BFS bounding discontinuities, allomembers WA and FA are not recognizable and therefore are not included in the framework of this project.

The allostratigraphic framework used for this project is shown in Figure 4.1. A total of ten discontinuities have been correlated throughout the entire study area. The lowermost (JE0) and uppermost (FSU) surfaces bound the Lower Colorado allogroup as a whole.

Two new surfaces have been recognized in this project that were not originally part of the framework established by Roca et al. (2008). An erosional surface within the Joli Fou
alloformation, originally mapped by Vannelli et al. (2017), was named JE1. Surface JE1 is a regionally-mappable flooding surface that was also correlated throughout the entire field area of this project. Vannelli et al., (2017) used JE1 to divide the Joli Fou alloformation into two allomembers, JA and JB. A second surface included herein is an erosional surface within allomember VB that separates VB into two distinct units, VB1 and VB2. This surface has been named VEM.

Figure 4.1: Nomenclature of bounding discontinuities, allomembers, and alloformations of the Lower Colorado allogroup. Surfaces WE1 and FE1 have been included to illustrate lap-out patterns determined by Roca et al. (2008)
4.2.2 Allostratigraphic Units

In the present study area, the Lower Colorado allogroup is composed of 9 distinct allomembers (Figures 4.1, 4.2). The four alloformations of the Lower Colorado allogroup are named the Joli Fou, Viking, Westgate, and Fish Scales alloformations, which are subsequently divided into their respective allomembers. The Joli Fou alloformation is divided into two allomembers: JA and JB. Above the Joli Fou, the Viking alloformation is divided into four allomembers: VA, VB1, VB2, and VD. The Westgate alloformation is divided into two allomembers: WB and WC. Lastly, the Fish Scales alloformation is composed of a single unit: FB.

The alloformations are given the same designation as the original lithostratigraphic formations defined by Bloch et al. (1993); however, it is important to note that the boundary of each alloformation does not necessarily correspond with the boundaries of the lithostratigraphically defined formations. For example, the lithostratigraphic top of the Joli Fou Formation is generally defined by the first appearance of a sandstone unit; whereas the base of the overlying Viking alloformation, defined by surface VE0, can occur at a subtle flooding surface within the sandstones of the first sandy unit. These sandstones would normally be assigned to the lithostratigraphic Viking unit. This example can be seen in Reference Well B (Figure 4.2). A second example of the difference between lithostratigraphic and allostratigraphic definition of units is shown with surface VE4. Typically, the top of the Viking Formation is identified by the last coarse-grained sandstone unit; however, in the southern part of this project area, surface VE4 is correlated above the youngest sand unit and is placed in what would lithologically considered to be the Westgate Formation (Figure 4.2 A).

4.2.3 Shared Surfaces and Units Within Previous Viking Studies

This section will briefly describe the similarities and differences of two previous studies (Bartlett, 1994; Walker and Wiseman, 1995) with that of this thesis. It is important to note that both previous studies focused on a geographically smaller area than this study.
Figure 4.2: Reference wells for the allostratigraphic framework used in this study. Well A depicts the disparity between the lithostratigraphic and allostratigraphic classification of the Lower Colorado Group where VE0 and VE4 are not placed within the Viking Formation but are the bounding surfaces of the Viking alloformation.
Bartlett (1994) set out to investigate the processes that distributed coarse sediment that formed gradationally based units offshore of described lowstand Viking shorefaces. Bartlett’s study included a detailed facies analysis, sequence descriptions, log correlations, and analysis of unit geometries. The key difference between Bartlett (1994) and the present study was that correlation was not based on an allostratigraphic scheme but on distinct lithological boundaries observed in core as well as on well-log. Bartlett (1994) placed Viking unit boundaries at points of rapid facies changes, which corresponded to points of rapid change in sand content. However, the use of a purely lithostratigraphic scheme by Bartlett (1994) does not allow the recognition of the temporal relationships between the various lithostratigraphic units and therefore a regional chronostratigraphic framework is necessary to understand a regional scale depositional history.

Bartlett (1994) identified six units within his study area: Mu, Epsilon, Delta, Gamma, Beta, and Alpha (Bottom to top). Comparison of these units to the allostratigraphic units of this study are as follows; Mu, Epsilon, and Delta approximate what would be considered Viking allomember VA; Gamma is approximately equivalent to VB1; Beta is approximately equivalent to VB2; Alpha is approximately equivalent to VD.

The study of Walker and Wiseman (1995) focused on three long and narrow sandbodies of the Viking Formation: Joarcam, Beaverhill Lake, and Lindbrook. The goal of Walker and Wiseman (1995) was to identify shorefaces formed at relative sea level lowstand, and to compare them with shorefaces incised during minor stillstands in the subsequent transgression. Because their study encompassed a smaller geographic area, Walker and Wiseman (1995) used a local scheme to subdived and understand the shoreface successions. Therefore, instead of using the Boreen and Walker (1991) scheme (VE1, VE3, VE4), Walker and Wiseman (1995) labelled the surfaces with regards to the specific sand body and type of discontinuity (e.g. BHL TSE = Beaverhill Lake transgressive surface of erosion).

Comparison of the surfaces correlated by Walker and Wiseman (1995) to the surfaces used in this study yielded some differing results. Direct comparison of wells used in both studies showed that VE1 of this study was generally equivalent to BHL RSE (regressive surface of
erosion) but also to JCAM ITI (initial transgressive incision). Surface VEM of this study was shown to be equivalent to BHL TSE but also equivalent to JCAM ITI in one well. One possible solution to discrepancy could be due to the fact that Walker and Wiseman (1995) used only 3 dip-oriented and 1 strike oriented cross-sections. There is potential for correlation error if surfaces are not properly looped.

4.3 The Lower Colorado allogroup

4.3.1 Joli Fou alloformation

JE0 is the basal discontinuity of the Lower Colorado Allogroup as well as the base of the Joli Fou alloformation. The JE0 erosional surface marks the boundary from the non-marine rocks of the Mannville Group to the marine mudrocks of the Lower Colorado Group. JE0 was not observed in core; however, in well-logs and cross-sections, the surface is marked by change from a distinct low to high gamma ray and high to low resistivity readings. An internal Joli Fou surface, initially identified by Vannelli et al. (2017) and termed JE1, was mapped throughout the study area. JE1 splits the Joli Fou into two, nearly tabular mudstone dominated units. The top the Joli Fou alloformation is defined by VE0, which is a marine transgressive surface.

Within the study area, the Joli Fou alloformation has a maximum thickness of 23 m and a minimum of 7 m. The alloformation thins steadily from the SE to the NW over 190 km. It was shown by Vannelli et al. (2017) that the Joli Fou continues to thin and laps out towards the NW against a subaerial topographic ridge, informally termed the ‘Smoky River Ridge’.

The Joli Fou alloformation consists of marine mudrocks. Cored intervals of this unit were scarce in the study area, the most complete core being in well 11-12-046-26W4 (Figure 4.3), which consists of approximately 12 m of clay-rich mudstone with rare, thin (<1 cm) beds of very
Figure 4.3: Upper approx. 3m of Joli Fou cored interval (11-12-046-26W4). Dark grey/black clay-rich marine mudstone. White arrows point to thin sandstone beds. Scale is 20cm. Located at 1323-1326m TVD.
fine lower sandstone. The cored interval occurs within the middle of the alloformation and therefore the bounding discontinuities JE0 and VE0 were not sampled.

4.3.2 Viking alloformation

The Viking alloformation overlies the Joli Fou alloformation and is composed of four allomembers: VA, VB1, VB2, and VD. The Viking allomembers are bounded by surfaces VE0 (base), VE1, VEM, VE3, and VE4 (top). VE0 is a regional transgressive surface that defines the boundary between the Joli Fou and Viking alloformations. It generally appears below, or near the base of, the gamma ray and resistivity deflections that mark the base of the lithostratigraphic Viking Formation. Cores that span the VE0 surface were rare. VE0 in well-logs had a distinct log signature, however, in the cored wells that did span this surface, VE0 generally had a very subtle, if any, visible sedimentological expression. The clearest expression of this surface occurs in well 10-22-048-17W4 where a sideritized chert pebble lag is present at the very base of the cored interval (Figure 4.4).

4.3.2.1 Allomember VA

Allomember VA is the basal unit of the Viking alloformation and is bounded by VE0 at the base and VE1 at the top. Allomember VA reaches a maximum thickness of 24 m in the NW corner of the study area, and a minimum of 4 m in the SE. Allomember VA consists of one to three regional, upward-coarsening successions (or parasequences) of shallow marine facies. Sedimentological facies included in VA are dark grey mudstone (facies 2), heterolothic bedding (facies 4), Thoroughly bioturbated silty-sandstone (facies 5), mud-draped sandstone (facies 6), and massive to cross-bedded sandstone (facies 7). The number of parasequences, degree of bioturbation, and percentage of sand in VA varies throughout the field area. A representative core through VA is shown in Figure 4.5.

Surface VE1 marks the top of VA and is a regional marine transgressive surface that was correlated over the entire study area. In local areas, VE1 was difficult to distinguish from other flooding surfaces (Appendix A, cross-section 2, well 15-08-056-01W5). In other examples,
Figure 4.4: Discontinuity VE0 represented by a sideritize-cemented siliclastic lag in well 10-22-048-17W4. See cross-section 4, well 10-16-048-17W4 for core-log/well-log correlation. Located at 823 m TVD.
Figure 4.5: Cored interval of VA from well 04-20-057-26W4. In this core, VE1 is mantled by a lag of poorly sorted granules, pebbles, and sideritized mud-clasts. Scale is 20cm. Depth of cored interval is 897.3 m to 915.5 m TVD.
VE1 is more easily recognized, such as in well 06-20-056-8W5 (Figure 4.5), where VE1 is mantled by a poorly sorted lag of granules, chert pebbles, and sideritized mud-clasts.

### 4.3.2.2 Allomember VB

Allomember VB, previously defined by Roca et al. (2008) is defined by surfaces VE1 below and VE3 above. In the present study, it was recognized that VB could be divided into two separate units where the bulk of the deposition shifts from the northern half of the study area (VB1) to the southern half of the study area (VB2). These units are separated by an erosional surface, here termed VEM, which could be mapped throughout the study area. The total thickness of allomember VB reaches a maximum of 34 m in the central part of the field area of (T48 and R26W4). The minimum thickness occurs in the northwestern (T60) portion where the unit thins to as little as 4 m.

Within allomember VB there is a large degree of facies variability comparable to allomember VA. The facies observed in core through allomember VB depend on location within the study area, and therefore, depositional environment. There is a distinct spatial variability in sand content, sand thickness, and intensity of bioturbation that changes within the allomember. Unlike VA, multiple conglomeratic units logged in cores occur in VB successions.

VE3 is a marine transgressive surface that is has been correlated regionally throughout the entire study area and throughout Alberta (Roca et al., 2008; Buckley et al, 2016; Vannelli et al., 2017, Drljepan, in progress). Surface VE3 marks the top of allomember VB2. VEM is erosionally truncated by VE3 in the northern part of the field area and other internal parasequences within VB2 were shown to be truncated by VE3 as well (see chapter 7). In core, VE3 is most commonly mantled by a transgressive pebble lag and capped by fine-grained marine mudrocks. A cored interval of VB1 and VB2 (Figure 4.6) shows multiple coarsening upwards successions capped by a pebble lag at VE3.
Figure 4.6: Cored through VB1 and VB2. Pictures show multiple upward-coarsening sequences within VB capped by VE3 pebble lag. Detailed core log shows position of discontinuities within the section. Scale bar is 20cm. Cored interval from 1415 m to 1450.5 m TVD.
4.3.2.3 Allomember VD

Allomember VD is the youngest allomember of the Viking alloformation and is defined by VE3 at the base and VE4 at the top. Allomember VD is a thin, approximately sheet-like body ranging from three to nine metres thick. In core, VD is generally represented by a single succession that coarsens upward from marine mudstone into fine to coarse-grained sandstone, which is then mantled by a coarse lag. Lags range in thickness from a granule bed to a >5cm thick pebble lag. Figure 4.7 illustrates a cored interval of allomember VD.

VE4 represents a marine transgressive surface that truncates Allomember VD (see cross-section D, Figure 4.7). VE4 crosses the ‘lithological classification’ boundaries, meaning that in well-logs, VE4 is traced into what would lithologically be classified as the Westgate Formation (Appendix A, cross-section 6)

4.3.3 Westgate Alloformation

4.3.3.1 Allomembers WA, WB, and WC

Allomember WA is bounded by VE4 at the base and WE1 at the top. However, this study area does not contain the WE1 marine transgressive surface because allomember WA pinches out northwest of this study area and therefore will not be discussed herein. The bounding surfaces of WB are surfaces VE4 and WE2 at the bottom and top respectively. Allomember WB reaches a maximum thickness of 26 m in the northwestern section of the study area and thins to a minimum of 14 m in the southeastern section.

Westgate allomember WB is not extensively cored and is generally sampled in the top few meters immediately overlying the Viking formation. The lithology is dominantly clay-rich mudstone with very rare thinly bedded rippled sand laminae (Figure 4.8). Gamma ray and resistivity logs indicate that this unit is composed mainly of mud-dominated rocks represented by high gamma ray and low resistivity readings.
Figure 4.7: Cored interval of VD showing the contact of VE3 overlain by mudstone, coarsening upwards to a sandy siltstone with increased bioturbation, then to a more chert-rich medium sandstone-dominated lithology and capped by the VE4 chert pebble lag. VE4 is overlain by approximately 60 cm of Westgate allomember WB. Scale bar is 20cm. Cored interval from 1361 m to 1376.4 m TVD.
Figure 4.8: Cored interval of Westgate allomember WB. VE4 is marked by a single chert pebble. The lithology of WB is dominantly clay-rich mudstone with rare very fine sandstone laminae. Scale bar is 20 cm. Cored interval is from 1511 m to 1532 m TVD.
Discontinuity WE2 is the upper boundary of allomember WB. It is a marine transgressive surface that has been correlated throughout the entire study area. Although never observed in core, the wire-line log signature produced by this surface was relatively easy to correlate (see Appendix A cross-section B). Cores in the eastern part of the Roca et al. (2008) study area show that surface WE2 is commonly mantled by a thin, fine-grained sandstone or a bed with a high concentration of fish debris indicative of sediment starvation resulting from transgressive reworking.

Allomember WC is defined below and above by the surfaces WE2 and BFSM respectively. Allomember WC ranges from 12 to 36 m in thickness and is thickest in the eastern portion of the study area and thins westward. Like allomember WB, WC is dominated by mudrocks with high gamma ray and low resistivity wire-line log readings. There was no core available from this allomember.

4.3.4 Fish Scales alloformation

4.3.4.1 Allomember FB

Within the study area, discontinuity BFSM defines the base of the Fish Scales alloformation. This surface was used as a datum for all cross-sections because it is the most distinct geophysical signature and is present in all wells studied. The gamma ray measurement for the Fish Scales alloformation has the highest gamma reading because of a high uranium material content that was adsorbed on phosphatic fish debris due to the onset of an anoxic event that produced the Fish Scales marker (Schroder-Adams et al., 1996).

Allomember FA is confined to the fordeep and is truncated by BFSM in the study area of Roca et al. (2008). Allomember FB is the upper most allomember of both the Fish Scales alloformation and of the entire Lower Colorado allogroup. Allomember FB is defined by BFSM at the bottom and FSU at the top respectively. The thickness of this unit remains moderately consistent throughout the study area. There is a slight thickening from 3 m in the east to a maximum of 6 m in the west. The lithology of this unit consists of thinly-bedded, fine-grained sandstone and mudstone with a high concentration of fish scales (Stelck et al., 1958; Stott, 1982).
Discontinuity FSU defines the top of the Lower Colorado allogroup and is a condensed, organic rich package of rock that corresponds to the downlap surface below the Dunvegan alloformation (Plint, 2000).
Chapter 5: Facies Analysis

5.1 Introduction

In the study area of this project, rocks of the Lower Colorado allogroup record deposition in marine environments. In order to determine an environmental interpretation for the rocks observed in core, it is important to divide the rocks into separate facies, which are defined by unique assemblages of sedimentological characteristics. A total of 9 distinct facies have been identified and are described and discussed in this chapter. Sedimentological analysis was based on detailed logs of 52 cored sections with a cumulative thickness of approximately 1130 m.

5.2: Facies Descriptions and Interpretations

5.2.1 Facies 1: Dark Grey Mudstone (Joli Fou Alloformation)

Description: The dominant aspect of this facies is the overall partially laminated to massive appearance and dark grey colour (Figure 5.1). The lithology is clay with very rare silt or VFL (very fine lower) sand laminae. Fine-grained and sharp-based wave-rippled silt beds were observed within this facies; however they were rare. This facies includes the occasional fish scale and bentonite bed (2-3 cm thick). Bands of orange siderite cementation can occur within this facies. In one example, the siderite band contains multiple fragments of inoceramid (bivalve) shells, occurring both in the bedding plane and cross-cutting the bedding plane. Other fauna were not observed. No bioturbation is visible in this facies as the cored sections tend to be fissile and broken in a platy to blocky fashion. Of all cores analyzed, only one was completely Joli Fou mudstone.

Interpretation: The fine grain size of the dark grey Joli Fou mudstones suggest that transport of this sediment was likely to have been as a suspended load that was subsequently deposited in a low energy marine environment (Swift et al., 1987). The source of this mud is interpreted to be hyperpycnal plumes entering a marine setting from river mouths during
Facies 1: Dark Grey Mudstone (Joli Fou)

Figure 5.1: A: section of core showing approximately 4.5 m of fissile mudstone (red box) and the occasional cemented horizon (white box). B: Approximately 3 cm thick bentonite bed found within the mudstones of facies 1 (blue box). Bentonite beds were not common but always had a distinct grey-blue colour. C: sideritized horizon of facies 1 containing inoceramid shell fragments. Due to the cross-cutting nature of the inoceramid, it suggests a semi-infaunal mode of life. This was the only example of macrofauna found. Cored interval approximate depth is 1332 to 1328 m
river floods (Hill et al., 2000). Observations of ancient muddy shelf deposits (e.g., Varban and Plint, 2005) are leading to the recognition that a majority of mud deposition does not just occur as a passive process of the vertical settling of particles, but is a more dynamic process that involves long-distance horizontal transport of mud aggregates with a complex history of depositional, erosional, and biotic colonization events that occurs in water much shallower than typically investigated (Plint, 2010). Mud may be re-suspended from nearshore environments by storm waves. It may then travel offshore as a wave enhanced sediment gravity flow (Macquaker et al., 2010; Plint, 2014), or it may travel as a wave-suspended bottom layer driven offshore by storm-driven geostrophic currents (Varban & Plint 2008; Plint, Macquaker, Varban, 2012). Fine-grained and sharp-based wave-rippled silt and sandstone beds preserved in the mudstone suggest deposition occurred from decelerating storm-generated flows above storm wave base at a relatively shallow water depth (Schieber, 1994).

The orange weathering of certain beds of mudstone suggests that siderite is the predominant cement phase. These cemented horizons are interpreted to be the product of early diagenesis (McKay et al., 1995). This early diagenesis could have occurred from an influx of fresh water which would suggest there was possible subaerial emergence of the shelf. The presence of small granule or pebble beds within siderite horizons would potentially support this hypothesis as it could provide evidence of shoreface ravinement and flooding following emergence (Nummedal and Swift, 1987). The presence of inoceramid shell fragments (Figure 5.1 C) indicate that water conditions, at times, were favourable for benthic fauna; however, the majority of facies 1 shows little to no evidence of bioturbation, which suggests that environmental conditions at the time of deposition were relatively unfavourable to most organisms.

Clay-rich, grey bentonite beds within facies 1 record the product of explosive volcanic eruptions. Due to the fine-grained nature of the volcanic ash, it would suggest that deposition occurred in relatively low energy, relatively deep-water conditions to ensure preservation.
5.2.2 Facies 2: Laminated Dark Grey Mudstone (Viking Alloformation)

**Description:** Laminated dark grey mudstone of facies 2, was deposited as part of the Viking alloformation, and generally overlies coarser grained rocks or an interclastic lag (facies 8) (Figure 5.2). Facies 2 commonly grades up into a coarser-grained unit such as facies 4 or facies 5. The lithology of facies 2 in dominantly clay to silt and may contain thin laminae of very fine-grained sand with scoured bases. The bioturbation index of facies 2 varies from 0 to 2. In the case of no to little bioturbation, the mudrock is very finely laminated with mm- to sub-mm scale bedding. Silt and VFL sand laminae are light grey in colour and can have well-preserved ripple laminations. Trace fossils in facies 2 include *Chondrites*, *Planolities*, and *Zoophycos*.

Several bentonite beds, ranging from 2 to 20 cm in thickness, were observed in facies 2 and are light bluish-grey in colour. Pyrite is present in facies 2 and typically occurs as cement in isolated burrows. Fish scales and debris are also found in facies 2; the fish fragments can be found as solitary scales, groups of scales, or fish bone concentrates.

**Interpretation:** Like facies 1, the fine-grained nature of facies 2 suggests deposition in a relatively low energy environment. Facies 2 has extremely well preserved fine laminations of mud and silt showing evidence of small wave ripples. The ripples indicate that the sediment was transported, probably by storm waves that had the energy to move sediment below fair-weather wave-base. Mudstones that overlie flooding surfaces (Figure 5.2 Ai and Aii) are interpreted to record slow sedimentation during transgressive events, whereas the siltier laminated mudrocks are inferred to have been deposited during progradation of the marine coastal system (Schutter, 1998). Typically, facies 2 coarsens upwards into a more silt or very-fine sand-rich facies such as facies 4.

Compared to the relatively ubioturbated facies 1, facies 2 contains a greater abundance of trace fossils. A possible reason for the presence of trace fossils could be due to a time period of increased oxygen levels at the sea-floor, making the environment more hospitable for benthic fauna. Oxygenation can occur as the result of storm events if the sea floor is above storm wave base (Dashtgard and MacEachern, 2016). It could therefore be assumed that the bioturbated mudrocks of facies 2 were deposited at a shallower depth than
Facies 2: Laminated Dark Grey Mudstone (Viking)

Figure 5.2: A)i: Laminated dark grey mudstones of facies 2 deposited sharply on a coarse granule layer (lag lying on a ravinement surface). This particular section is very finely laminated with little to no visible bioturbation. A)ii: sharp contact of chert granule lag to mudstone. B: 8 cm section of facies 2 showing a wave-ripple laminated VFL sandstone bed (blue arrow) and a sharp, erosional-based storm bed (red arrow). Some bioturbation of primary bedding is visible in this sample (green arrow). Cored interval at approximately 838 to 842m TVD.
the unbioturbated mudrocks of facies 1.

Figure 5.3 C depicts a concentrated fish bone bed which is interpreted to record a low rate of siliclastic sedimentation and protracted physical winnowing which allowed the accumulation of the organic debris (Schroder-Adams et al., 2001).

The greyish-blue bentonite beds found in facies 2 suggests that deposition occurred in relatively low energy, relatively deep-water conditions. The thickest bentonite bed found in facies 2 was located in well 08-14-048-04W5 (Figure 5.3 A), which had a thickness of approximately 20 cm. A bentonite of this thickness would have been produced by a particularly large volcanic eruption or the area was in the direct path of the ash plume.

5.2.3 Facies 3: Dark Laminated Mudstone (Westgate Alloformation)

Description: Facies 3 is composed of laminated black mudstones of the Westgate alloformation, which has a sharp lower contact with a coarse pebble lag (facies 8) or, in the absence of a lag, a coarse-grained sandstone unit (Figure 5.4). The lithology is dominantly clay and silt and appear much darker in colour than facies 1 and 2. Whispy VFL sharp-based sandstone with visible ripple laminations beds is occasionally preserved.

Bioturbation is rare in facies 3 but fish scales are very common. Both solitary scales as wells as large concentrations of scales were found.

Interpretation: Like facies 1, facies 3 is dominantly a very fine grained mudstone unit that suggests that sediment transport probably took places as a suspended load that was subsequently deposited in a low energy marine environment. Mud probably originated from rivers during a flood, which is then transported by a hyperpycnal plume (Hill et al., 2000). The appearance of very fine rippled sand laminae indicates that storm waves impinged on the sea floor which therefore must have lain above storm wave-base. Fissile mudstones overlying flooding surfaces and pebble lags (facies 8) are interpreted to record slow sedimentation during transgressive and highstand events (Schutter, 1998). Primary mudrock lamination appears to be very well preserved as there is no visible bioturbation within facies 3.
Figure 5.3: A: A thick (approx. 20 cm) bentonite bed within facies 2 (orange box). This was the thickest bentonite found in the study area. This section of facies 2 shows a slightly greater degree of bioturbation because some of the primary bedding of individual storm-beds has been disrupted.

Approximately 721 to 725 m TVD B: Facies 2 sharply overlying massive sandstone (facies 6). Red dotted line represents surface VE1. There is also a pyritized burrow visible (yellow box) (1570 m TVD) C: A highly concentrated bed of fish bones. Blue arrow highlights a relatively long skeletal fragment. The debris has a blueish colour due to a high phosphate content.
Figure 5.4: A)i: Approximately 3 m of dark fissile mudstones of the Westgate alloformation with the occasional siltstone storm bed (A ii). Little to no bioturbation was visible within facies 3. B) A concentrated bed of fish scales (green box) within the mudstones. Both solitary scales and groups of scales were commonly found in facies 3. C)i: Illustrating the sharp contact between the dark mudstones of facies 3 and a pebble lag (facies 8). Red box on stratigraphic log shows the position of the contact within the entire section. Silty storm beds are also visible as the colour contrast between the dark (possibly organic rich) mudstones and the lighter grey colour of the silt beds. C)ii: pebble lag sharply overlain by facies 3 mudstones.
The lack of bioturbation suggests unsuitable environmental conditions for benthic fauna. Other factors that could have affected the presence of bioturbation in facies 3 could be the lack of food, salinity, and substrate consistency (MacEachern et al., 2010).

5.2.4 Facies 4: Centimetre-scale Heterolithic Bedding

Facies 4 consists of varying proportions of interbedded mudstone and VF to F sandstone (Figure 5.5). The proportion of sandstone tends to increase upwards. Facies 4 commonly records the transition from dark grey laminated mudstone (facies 2) to a coarser-grained facies such as thoroughly bioturbated silty-sandstone (facies 5), massive sandstone (facies 6), or mud-draped sandstone (facies 7). Mudstone bed lithology is dominantly clay, and the sandstone interbeds tend to have a grain size ranging from VFL to FU; however, there are a small number of examples where the sand interbeds reach grain sizes of medium to coarse sand. Individual beds of sandstone and mudstone range from 1 cm to typically no more than 5 cm thick. The sandstone beds in facies 4 can show laminations with evidence of wave ripples. The base of the sandstone beds appear sharp and are subtly erosive. The tops of the beds can be sharp or can show normal grading into a mudstone.

Bioturbation of facies 4 ranges from a BI of 0 to 3 but is typically localised to distinct beds. Common types of trace fossils include Chondrites, Planolites, Teichichnus, overall indicating a Cruziana ichnofacies (MacEachern et al., 2010).

Interpretation: Due to the increase in sand content compared to facies 2, facies 4 is interpreted to have been deposited more proximal to the shore where storm and wave-related processes had a more significant effect on the sea-bed. The evidence for storm-wave interaction with the sediment is the nature of the base of the sandstone beds being sharp based and scoured. Sedimentary structures, such as oscillation ripples, were found in fine sandstone units of facies 4. The interstratified mudstones of facies 4 are likely to be the result of the deposition of mud clouds suspended by storms (Scheiber, 1998) or from hypopycnal plumes issuing from river mouths (Hill et al., 2000).
Figure 5.5: A: Heterolithic bedding (facies 4) showing increased silt/VF sand content compared to facies 2. B: An enlarged view of a small section that better exemplifies the scale of interstratification. Primary bedding of storm beds is visible. Other examples of facies 4 occasionally contain increased bioturbation. 834 m TVD.
The degree of bioturbation in facies 4 ranges from none (BI= 0), to moderate (BI= 3). Packages of mudrock with no visible bioturbation may indicate that environmental conditions were not favourable for benthic fauna on the sea floor. Occasional pyrite cementation found within facies 4 suggests that the environment during the deposition of some units of mudrock were anaerobic at least a few mm below the sediment-water interface (Macquaker et al., 1998). Moderate bioturbation of primary lamination suggests that oxygen levels increased enough to allow for benthic fauna to colonize the sea-floor.

Studies by Bhattacharya and MacEachern (2009) on delta systems of the Ferron Sandstone in Utah and the Dunvegan Formation in Alberta attribute this type of heterolithic facies to prodeltaic muddy “hyperpycnites”. Bhattacharya and MacEachern (2009) showed that the mudstones show diffusely bedded, centimeter-thick, normally to inversely graded siltstone and very fine-grained sandstone beds, with internal scours, suggesting deposition during waxing as well as waning hyperpycnal flows.

5.2.5 Facies 5: Thoroughly Bioturbated Silty-Sandstone/Sandy-Siltstone

**Description:** Facies 5 is characterized by a sequence of thoroughly bioturbated VF to F sand and silty sandstone (Figure 5.6). Facies 5 tends to coarsen upwards as the silt fraction decreases and sand increases. Facies 5 typically occurs above a fine grained facies such as facies 2 or 4 and is followed by the deposition of a coarser sandstone unit such as massive or cross-bedded sandstone (facies 6).

The most distinctive feature of facies 5 is the degree of bioturbation. With a bioturbation index between 4 and 6, very little original bedding is preserved and the units appear quite massive and relatively poorly sorted. Typical macroscopic trace fossils include; *Diplocraterion, Rhizocorallium, Rosselia, Skolithos, Teichichnus, and Zoophycos*. This suit of trace fossils would indicate a *Skolithos* ichnofacies (MacEachern et al., 2010). Large *Skolithos* burrows on the scale of 20-30 cm in length are not uncommon in this facies.

**Interpretation:** The distinguishing feature of facies 5 is the degree of bioturbation. Facies 5 is a more intensely bioturbated version of facies 4 (and sometimes facies 6). An
Facies 5: Thoroughly Bioturbated Silty-Sandstone/Sandy-Siltstone

Figure 5.6: A-C showing the extent of bioturbation in a cored section of well 08-06-044-26W4. 1447 to 1448 m TVD
increase in sand content and grainsize, as well as the variety of the trace fossils present, suggests that facies 5 was deposited more proximal to the shore; probably within the lower shoreface (MacEachern and Pemberton, 1992; MacEachern and Bann, 2008). This is a higher energy environment with a sufficient nutrient supply to support a high abundance of organisms. In most examples of facies 5, all primary stratification is obliterated (BI= 5-6).

5.2.6 Facies 6: Massive to Cross-bedded Sandstone

**Description:** Massive sandstones range in grainsize from FL to VCU. Primary sedimentary structures are absent and burrows are rare. The sandstone units can be 3-4 m thick, but can be interstratified with thin mud drapes. Poorly sorted granules and pebbles, sideritized mud-clasts, and woody plant material are common. The granules and chert pebbles found within the massive sandstone units appear to be ovoid and well rounded. Some of the sandstone units contain thin, wispy mud-draps; however, not enough mudstone is present to be considered facies 7.

Cross-bedded sandstone beds are common within the thick sandstone units; the grainsize of the cross-bedded units ranges from FU to VCU. Some beds show normal grading from VC at the base to ML-MU at the top (Figure 5.7). The occasional sideritized mud-clast and coal fragment can be found within the cross-bedded units. In some sandstone packages, planar parallel beds can be seen where the strata appear to have no dip angle.

Facies 6 is typically found at the gradational top of a coarsening-upwards succession but can also be sharp based, where facies 6 was deposited on a fine-grained facies such as facies 1 or 2. Facies 6 can be sharply overlain by a fine-grained facies (Figure 5.8) but can also be overlain by a coarser conglomeratic unit or pebble lag (Figure 5.9).

**Interpretation:** Alone, massive sandstones cannot easily be interpreted due to the lack of information provided. Scarcity of sedimentary structures within massive sandstone has been attributed to be the result of biological modification effected by microscopic organisms (meiofauna) acting on individual grains, leading to the thorough homogenization of the sandstone body without generating and trace structures (Bromley, 1996). Cross-stratification is
Facies 6: Massive to Cross-bedded Sandstone

Figure 5.7: A: an approximately 3 m section showing multiple sets of cross-bedding. The magnified section shows normal grading and the well-preserved contact between two cross-sets. Some thin mud-drapes are also visible (1399 to 1396 m TVD). B: Cross-bedding with a steeper dip angle (903 m TVD). C: Sideritized mud-clasts (blue arrow) within cross-bedded sandstone. Primary lamination with mud-clast is still visible (1320 m TVD).
Facies 6: Massive to Cross-bedded Sandstone

Figure 5.8: A: Section of core showing the transition from thoroughly bioturbated sandstone (Facies 5) to a massive sandstone with little bioturbation and the occasional thin mud-drape. Blue arrow indicates approximate facies transition from facies 5 to 6. B: Close-up view of flooding surface and sharp contact between facies 6 and facies 2 (1570 m TVD). C: Section of massive sandstone with very thin mud drapes.
Facies 6: Massive to Cross-bedded Sandstone

Figure 5.9: A: section of core showing a coarse grained sandstone unit with both massive (Ai) and cross-bedded units (Aii) that is overlain by a poorly sorted conglomeratic unit (facies 9). A)i: Massive sandstone with large Skolithos (blue arrow) that has been filled with small pebbles. The burrows appear to be a darker brown/orange colour and indicate possible siderite cementation. A)ii: Cross-bedded unit with granules and small pebbles poorly sorted throughout. Cored interval at 1165 to 1161 m TVD.
interpreted to record deposition in subaqueous dunes produced by longshore and shore-normal currents in the upper shoreface (Clifton et al., 1971).

The basal contact of facies 6 (gradational or sharp-based) can be attributed to the trend of relative sea-level change during the deposition of the shoreface. Where the base is gradational, it would suggest that there was a conformable progradation of the shoreface, allowing the preservation of underlying and laterally contiguous environments (Plint, 1988). When massive to cross-bedded sandstone is deposited unconformably on top of fine-grained offshore facies, it indicates deposition took place during a relative sea-level fall and forced regression (Plint, 1988; Plint and Norris, 1991; Posamentier et al., 1992).

Evidence that deposition of facies 6 occurred close to river mouths includes the appearance of coal and woody fragments as well as sideritized mud-clasts. The wood and plant material would have been transported down river, probably during a flood, and deposited at the river mouth. Sideritized mud clasts are believed to be evidence of a delta distributary channel.

5.2.7 Facies 7: Trimodal Mud-draped Sandstone

**Description:** Facies 7, like facies 4, is characterized by distinct heterolithic bedding; however, facies 7 is much more trimodal in terms of grain size (Figure 5.10). Facies 7 is characterized by coarse to very coarse-grained sand that has been draped by mud and silt. The thickness of both the sandstone and mudstone layers varies but is typically 2-3 mm, or as large as 1-2 cm. In the sandstone layers, sedimentary structures are difficult to determine; however, faint cross-stratification is locally visible. Occasional silt or VFL sand laminae within mudstone layers show evidence of unidirectional ripple lamination.

Facies 7 is most commonly deposited at the top of an upwards-coarsening succession of massive or cross-bedded sandstone (facies 6). Well 04-20-057-26W4 (see stratigraphic log in Appendix A, cross-section C) illustrates a complete upwards-coarsening sequence from facies 2, into facies 4, 5, 6, 7, and ends with a poorly sorted conglomerate bed that is over 2 m thick.
Figure 5.10: Examples of trimodal grainsize distribution found in facies 7. A: Alternating beds of coarse sandstone, very fine sandstone, and black mud (blue box). Each bed is approximately 2-3 mm thick (1010 m TVD). B: Approximately 4 m of facies 7. Pervasive siderite bands or nodules (orange box) suggesting fresh water input was common during early diagenesis and multiple burrows through mudstone layers (green box) are common in this section. C: Section of core showing the pronounced heterogeneity of facies 7. Homogeneous mud alternating with very coarse sand (yellow box).
Sideritization of coarse-grained sand layers and the inclusion of sideritized mud clasts is common in facies 7 (Figure 5.10 B). Bioturbation (BI 2-3) of mudstone layers is also common in facies 7.

**Interpretation:** Facies 7 displays a trimodal grain-size distribution combining mudstone, fine-, and coarse-grained sandstone beds and laminae. The coarser grain sizes of the sandstone bed suggest that this facies was deposited much closer to the source.

Facies 7 has the appearance of flaser or wavy tidal bedding; however, while these types of deposits are common in a tidal setting, to establish a tidal origin for these deposits, one must look for diurnal inequality or the presence of bipolar ripple laminations (Dalrymple, 2010). Strong evidence of tidal action was not observed in the cores examined in this study.

The siderite nodules present in Figure 5.10 B may have been produced by early diagenesis, which commonly requires an influx of fresh water to reduce sulphate activity and usually occurs where salt water and fresh water mix, such as in bays, lagoons or estuaries (Bhattacharya, 2006).

**5.2.8 Facies 8: Pebble Lag**

**Description:** The pebble lags are composed of poorly sorted chert and quartzite clasts that range in size from granules to pebbles; with pebbles being the most common (Figure 5.11). The pebble lags vary in thickness and range from a small group of isolated pebbles to thin, poorly-sorted conglomeratic units up to 15 cm thick. In the Eastern half of the study area, the lags are thinner, typically not much larger than granules 2-3mm in diameter.

Facies 8 always sharply overlies an erosion surface and caps an upward-coarsening, upward-shallowing succession. The conglomerate of facies 8 is generally sharply overlain by mudstone of facies 2 or 3. In some wells, the flooding surface was directly overlain by a few centimeters of mud before the pebble lag was deposited. In a few rare instances, two pebble beds, separated by 10-20 cm of mud, were observed.
Facies 8: Pebble Lag

Figure 5.11: A variety of pebble lags observed throughout the study area. Lags were commonly seen as cm thick units (A-C) or as thin bands of granules or pebbles (D). Lags were poorly sorted and commonly composed of chert pebbles.
No bioturbation was observed within facies 8. Each lag appears to be isolated from the facies below and above.

**Interpretation:** Coarse pebble lags are interpreted to be an erosional product of the landward migration of a shoreface (Nummedal and Swift, 1987) rather than an individual facies; however, pebble lags have considerable stratigraphic importance and are therefore are distinguished as a facies here. The presence of a pebble lag directly above a flooding surface implies a succession of events: 1) transport of gravel to the shore by a fluvial system, 2) subsequent subaerial exposure of the depositional site, prior to marine transgression, and 3) wave-reworking of the progradational gravel shoreface by the transgressing shoreline (Plint et al., 1986; Plint, 1988). The lag, therefore, represents the size fraction of the eroded sediment that could not be transported by the wave energy of the transgressive shoreface.

**5.2.9 Facies 9: Poorly-Sorted Conglomerate**

**Description:** Poorly sorted conglomeratic units were observed in multiple wells in the northwestern and north-central portion of the study area (Figure 5.12). One example of facies 9 reached a maximum of 4 m in thickness; however, two examples of conglomerate were located at the bottom and top of cored intervals respectively so the true thickness of those units could only be determined from well-log. It could be difficult to distinguish a coarse pebble lag (facies 8) from a poorly sorted conglomerate (facies 9). In some cases, lithologies of these units appear quite similar with only thickness of the unit being the distinguishing factor. It was decided that if the unit in question exceeded 15 cm in thickness, it would be considered facies 9.

Facies 9 ranges from clast supported to moderately matrix supported. The clasts range from granules (2-4 mm) to pebbles (<1 cm). The matrix is composed of medium to coarse-grained sand. The lithologies of the majority of the clasts appear to be chert and quartzite with a mixture of lithic fragments throughout. Sideritization of the conglomerate was seen in multiple examples of facies 9 together with multiple siderite clasts ranging from 2-8 cm in length. Within the siderite units, *Skolithos* may be present.
Facies 9: Poorly Sorted Conglomerate

Figure 5.12: Two types of conglomerate unit were observed in core; the difference between the varieties was distinguished by siderite cementation. A: A clast-supported chert conglomerate that shows no evidence of sideritization. Pebbles are well rounded. B: A poorly-sorted conglomerate with a greater variation in grain sizes. Siderite horizons and cementation is pervasive throughout this conglomerate. A sharp contact into facies 2 is also observed (B iii). Interval at 861 to 857 m TVD.
The basal contact of facies 9 is rarely observed, and where visible, is sharp. The facies observed below facies 9 include facies 6 and 7. The upper contact of facies 9 is also sharp and is commonly overlain by facies 2, 3, or 4.

**Interpretation:** The energy required to transport and rework the poorly sorted sediments of facies 9 must have been substantial. Environments where this could occur include delta fronts, river mouths, and gravelly shorefaces. Sedimentary structures were difficult to identify within the conglomerate units found in this study area. Unfortunately, no single criterion can be used to unequivocally differentiate between fluvial and shoreface conglomerates (Hart and Plint, 2003). Studies of the Kaskapau and Cardium Formations (Hart and Plint, 2003; Varban, 2004; Varban and Plint, 2005) have shown that marine conglomerates were deposited in close proximity to the mouth of gravel-bearing rivers.

The erosive nature of the basal surface of the conglomerate units is interpreted to be the result of erosion of the sea bed during forced regression (Plint and Nummedal, 2000). The visible upper contacts of the conglomerate units seen in core also appear to be erosive. This is believed to the result of ravinement during transgression.

Figure 5.12 B shows that the some of the conglomerate units were subject to siderite cementation in early diagenesis.

5.2.9.1 Conglomerate Core Cross-sections

Evidence of a potential deltaic system is provided by the presence of conglomerate units. A cross-section of cores is provided in Figure 5.13 and 5.14. The location of the conglomerates indicates the furthest extent of the river into the basin as it would be expected that they were deposited close to a river mouth. Figure 5.13 represents the conglomerate units that are located within allomember VA and 5.14 represents conglomerates found within allomember VB1. VA conglomerates are considerably thick. The conglomerate units belong to VA sandstone unit γ (see Chapter 7 for more information). Both wells 11-17-060-02W5 and 06-21-059-01W5 are sharply overlain by laminated mudstone of facies 2.
Figure 5.13: Core cross-section of conglomerate units in allomember VA. Location of core is indicated by the red dots on the field map. Conglomerates are indicators of a high energy environment. A fluvial river mouth is the probable depositional environment for these conglomerates. The location of these conglomerate units on the field map indicates how far the deltaic system prograded into the basin.
Figure 5.14: Core-log cross-section of conglomerate units in allomember VB1. Location of core is indicated by the red dots on the field map.
5.3 Depositional Environment Overview

Cored intervals within the study area could be divided into 9 distinct facies. These facies vary from fine-grained (off-shore) to coarse-grained (near-shore) units. Sedimentary and biogenic structures suggest the depositional environment was a marine shoreface to shelf system proximal to a river mouth due the presence of deltaic features such as trimodal heterolithic bedding and conglomerate units. A majority of the sedimentary features found within the cored intervals were formed by wave-generated processes. Figure 5.13 illustrates a hypothetical environment in which these facies were deposited. A more detailed paleo environmental interpretation is presented in Chapter 7.

Figure 5.13: Hypothetical depositional environment and the spatial location of each facies. The overall system is a wave-dominated delta with a strandplain shoreface. Green colour represents non-marine rocks (not observed in this field area but necessary to show), yellow represents a foreshore environment, pale yellow represents the shoreface, orange represents conglomeratic facies, grey is offshore marine, and brown is a mud plume out of a river. The red arrow indicates the hypothetical direction of Coriolos deflection (to the right). Facies 1 and 3 are mudstone facies that are just within storm wave-base; facies 2 is slightly more proximal to shore as it contained more abundant storm beds; facies 4 is more proximal to shore and contains a grainsize distribution that could have resulted from being close to a deltaic system; facies 5 is a lower shoreface deposit; facies 6 is composed of upper shoreface deposits; facies 7 contains a trimodal grainsize distribution ranging from mudstone to coarse sand indicating that facies 7 was deposited closer to the source than facies 4; facies 8 can be seen as a transgressive lag produced during relative sea-level rise; and facies 9 is a conglomerate unit deposited close to a river mouth.
Chapter 6: Controls on Accommodation

6.1 Introduction

The geometry of an allostratigraphic unit results from the interaction between several geological processes that control the evolution of accommodation with respect to area and time, the amount and rate of deposition, as well as post depositional erosion (Miall, 1991). Various subsidence and erosional mechanisms generate different accommodation patterns and therefore the geometry of the sediment body can be used to determine the process that controlled the generation of accommodation; including tectonics and eustasy (Plint et al., 2012).

In a foreland basin, the generation or removal of accommodation results from the load imposed by tectonically-thickened crust, the subsidence induced by the sediment load, and fluctuations in sea-level (Jordan and Flemmings, 1991). A foreland basin, because it is tectonically loaded only on one side, experiences subsidence that gradually decreases away from the tectonic load. This asymmetrical subsidence results in units being deposited with an overall wedge-shape that is thickest at the loaded side (Plint et al., 2012). Eustatic oscillations, in contrast, produce units with a tabular geometry that results from the uniform change in accommodation (Roca et al., 2008; Plint et al., 2012).

The aim of this chapter is to illustrate the geometry of successive Lower Colorado allostratigraphic units by means of isopach maps. The isopach maps show the total thickness of each allomember and were created using Surfer 8.0. The thickness of each allomember was measured to the nearest metre for each well.

6.2 Isopach and Isolith maps

6.2.1 Joli Fou Alloformation

The Joli Fou alloformation is the oldest allostratigraphic unit of the Lower Colorado allogroup. This mudstone-dominated unit thickens slightly to the southeast (Figure 6.1), which
suggests differential subsidence in that direction. The average thickness of the alloformation is approximately 15m. Deposition of the Joli Fou mudstones followed the Late Albian Skull Creek transgression, which resulted in the first connection of the northern Boreal Sea to the Tethyan Ocean during the early late Albian (Williams and Stelck, 1975). In combination, the two allomembers, JA and JB, of the Joli Fou alloformation have a near-tabular geometry with some local thickness variations. The tabular geometry of allomembers JA+JB is most reasonably interpreted as the result of eustatic rise during the Skull Creek transgression. The local thickness variation could potentially be attributed to underlying compactional variation of the Mannville Formation.

When mapped northward into the study area of Vannelli (Vannelli, 2016; Vannelli et al., 2017), the Joli Fou alloformation gradually thins and onlaps north-westward onto a topographic feature termed the 'Smoky River Ridge', which separates the Joli Fou depocentre in the east from the Paddy depocentre in the west. Vannelli et al. (2017) determined that Paddy allomembers A to F and the Joli Fou alloformation are time equivalent and both onlap and terminate onto the ‘Smoky River Ridge’ from opposite directions. Active flexural subsidence resulted in the creation of abundant accommodation close to the deformation front. The Paddy alloformation sediment that fills the newly created space had a wedge-shape geometry, reflecting differential subsidence across the foredeep, thinning to the east.

**6.2.2 Viking Allomember VA**

Allomember VA is the first sandstone-dominated unit of the Lower Colorado allogroup. Allomember VA comprises stacked, upward-coarsening successions of shallow marine deposits. The isopach map of allomember VA (Figure 6.2) shows that the unit thins from 24 m in the northwest of the study area to 4 m in the southeast. Allomember VA is effectively a ‘mirror image’ of the Joli Fou alloformation where the thickest part of the Joli Fou underlies the thinnest part of allomember VA. It therefore appears that the area with the most accommodation for the Joli Fou became the area with the least amount of accommodation for VA.
Figure 6.1: Isopach map of the Joli Fou alloformation (JE0 to VE0) illustrating the thickening of the allomember from the northwest to the southeast. Contour interval is 2 m.
Figure 6.2: isopach map of allomember VA (VE0 to VE1). VA thins to the southeast. Contour interval is 2 m.
Most coarser-grained VA sediment was deposited in the northwest of the study area. Allomember VA correlates with Pelican allomembers PeA and PeB (Vannelli et al., 2017) to the north. PeA and PeB display a broadly tabular geometry throughout the Vannelli et al. (2017) study area. Revised Isopach maps of VA in the study area of Roca (2007) show that the allomember has a tabular geometry that onlaps and pinches out to the west (Plint et al., 2016). A similar geometry is observed here, with some variation. The isopach map of VA displays a progressive thickness change from the southeast to the northwest and therefore cannot be considered completely tabular. This pattern is not readily explained as a consequence of tectonic loading because the thick zone is of rather localized distribution, which is not the typical ‘wedge-shaped’ geometry. A possible reason for this trend could be due to isostatic subsidence that resulted from the mass of the sediment preferentially deposited in the northwest. The near-tabular geometry of allomember VA in other study areas (Roca, 2007; Vannelli, 2016) suggests that eustatic rise was primarily responsible for the accommodation generated for this unit. A complex history of sea-level change had an effect on the stratigraphic stacking pattern and spatial distribution of VA sandstones is discussed in chapter 7. Orogenic activity was minimal at VA time with no flexural subsidence, but there was likely isostatic uplift in the west, which explains the onlap of Viking onto older rocks. Mellon (1967) determined that all of the Upper Albian is missing further west in the central Foothills, which implies uplift during that time.

6.2.3 Allomember VB

The isopach pattern of allomember VB differs from the underlying units. VB thins radially from a maximum of 33 m at a point in the central part of the study area (Figure 6.3). Towards the northeast, VB thins to as little as 4 m along a NW-SE trending linear feature that can be traced into the study area of Vannelli (2016). The thinning of allomember VB towards the northeast and the thinning of time-equivalent Pelican allomembers PeC and PeD towards the southwest suggests the linear region of thinning marked a region of very low subsidence rate, or even a topographic high during deposition. The four Pelican allomembers form south-facing sandbodies composed of quartz arenites that were most likely sourced from the
Figure 6.3: Isopach map of allomember VB (VE1 to VE3) showing the maximum thickness of the allomember is at approximately T48 and R26W4. The black arrow indicates the location of the region of low subsidence that continues to the northwest into the study area of Vannelli (2016) and Vannelli et al. (2017). Contour interval is 3 m.
Canadian Shield (Vannelli et al., 2017). The fact that both the Viking and Pelican sand bodies do not form distinct wedge-shape bodies that thicken towards the Cordillera, it is hypothesized that the original accommodation pattern was probably controlled primarily by sea level changes. This conclusion is also supported by Roca et al. (2008) who mapped a VB unit that thickened away from the thrust belt that suggested that flexural subsidence was not the primary control on accommodation.

6.2.4 Allomember VD

Viking allomember VD is bounded by surfaces VE3 and VE4. In this study area, VD is represented by a relatively thin sheet of mudstone-dominated rock with an average thickness of approximately 5 m. The allomember does appear to thicken slightly to the southeast (Figure 6.4). In isolation, the apparent tabular geometry could be interpreted as evidence that accommodation was controlled by eustasy; however, northwest of the study area, VD thickens to over 130 m in a flexural depocentre adjacent the Cordillera (Rylaarsdam, 2006; Buckley (2009); Angiel (2013); Plint et al., (2016). There is a second depocentre in southwest Alberta where VD reaches a maximum thickness of 25 m (Roca et al., 2008). This observation provides evidence for renewed flexural subsidence in the northwest and southwest that took place during upper Viking (VD) time, implying renewed tectonic loading (Plint et al., 2016).

6.2.5 Allomember WB

The Westgate Alloformation onlaps and thins from NW to SE across the whole basin. In this study area, allomember WA has already onlapped onto VE4, so only WB and WC are observed (Roca et al., 2008). Allomember WB is bounded by VE4 and WE2. Allomember WB thickens steadily from approximately 14 m in the southeast to 24 m in the northwest (Figure 6.5). When allomember WB is traced to the northwest into the study areas of Roca (2007), Angiel (2013), and Zhang (2006), the unit continues to thicken towards the thrust belt forming a prominent wedge-shape with a maximum thickness of approximately 70 m. South, into the study areas of Driljepan (In progress) and Sisulak (2007), allomember WB thins and tapers out and is not present south of about 52° N and west of 114°E. The wedge-shape geometry of
Figure 6.4: Isopach map of allomember VD (VE3 to VE4). Although the thickness change is subtle, the isopach shows a distinct thickening on opposite sides of a SW-NE trending ridge that is potentially the axis of the forebulge. Contour interval is 2 m.
Figure 6.5: Isopach map of allomember WB. WB thickens to the north. Contour interval is 2 m.
allomember WB shows that accommodation was controlled primarily by tectonic loading in northeastern British Columbia. Roca (2007) also observed that allomember WB tapered southward more gradually than the underlying allomember WA, leading to the conclusion that accommodation of WB was also, in part, controlled by eustatic rise.

6.2.6 Allomember WC

Allomember WC is the uppermost allomember of the Westgate alloformation in the study area. Allomember WC is bounded by surfaces WE2 and BFSM. WC thickens to the east from a minimum of 14 m to a maximum of 34 m within the study area (Figure 6.6). A key factor in determining the control on accommodation in WC time is to determine the shape of the alloformation throughout Alberta. WC was mapped by Roca (2007) who showed that the thinning observed within his study area was actually a broad arch (oriented southwest-northeast) that separates a southeast-facing wedge from a northwest-facing wedge, the latter thickening to > 100 m in the study areas of Angiel (2013) and Zhang (2006). The southeast-thickening wedge continues to thicken into the study areas of Drljepan (in progress) and Sisulak (2007) to approximately 50 m. The arch that splits the two wedges is interpreted as a flexural arch (a forebulge) related to loading and subsidence of the northern depocenter. Accommodation of WC is therefore inferred to have been strongly controlled by flexural subsidence; however, deposition of mudstone over the flexural arch suggests that accommodation was also generated by eustatic rise (Roca, 2007).

6.2.7 Allomember FB

Allomember FB is the upper allomember of the Fish Scales alloformation, and also the uppermost allomember of the Lower Colorado allogroup. It is bounded by surfaces BFSM and FSU. In the study area, allomember FB thickens from 3 m in the southeast to 6 m in the northwest (Figure 6.7). FB appears to have a tabular geometry in the study area; however, the unit thickens significantly in the northwest of Roca’s (2007) field area to form a very prominent wedge. It is likely that while sea-level rise did generate some space for sediment, flexural subsidence was the main control of accommodation in the proximal foredeep.
Figure 6.6: Isopach of allomember WC. WC shows a distinct thinning to the west. Contour interval is 2 m.
Figure 6.7: Isopach of allomember FB. FB shows a gradual thickening to the west. Contour interval is 1 m.
6.3 Influence of Precambrian Basement Structure

The Precambrian basement of Alberta is composed of accreted terranes that are separated by narrow zones of rheological contrast. This contrast may have acted to localize deformation when the plate was under stress, such as when it was loaded by the growing Cordillera (Ross et al., 1994). Precambrian terrane boundaries have been mapped throughout Alberta using aeromagnetic data, supplemented by drilling to basement in select wells (Ross et al., 1994). The potential influence of Precambrian terranes on accommodation for Cretaceous rocks can be determined by comparing an aeromagnetic anomaly map of the basement structures to the location and orientation of regional isopach trends. Figure 6.8 shows the aeromagnetic anomaly map for this study area. Comparing this map to the isopach maps of each allomember, there does not appear to be any distinct relationship between terrane boundaries and regional isopach trends.
Figure 6.8: Tectonic domains (Wabamun (WA), Thorsby (TH), Rimbey (RY), Lacombe (L), and Loverna Block (HI)) of the basement structures beneath the WSCB. There appears to be little correlation between the basement structure and the thickness patterns seen in allomember isopach maps. (modified from Ross et al., 1994)
Chapter 7: Paleogeographic Evolution of the Lower Colorado Allogroup

7.1 Introduction

This chapter summarizes the evolution of the Lower Colorado allogroup within the study area and discusses the results and conclusions determined in adjacent study areas. An interpretation of the depositional history and paleogeographic evolution of the Joli Fou, Viking, and Westgate alloformations is presented. Particular emphasis is placed on the interpretation of the Viking alloformation. The main tectonic and eustatic events that are interpreted to have taken place from the early late Albian through the late late Albian in the Western Interior Basin of Canada is also discussed.

Results are portrayed in the form of sandstone isolith maps, sandstone distribution maps, and fence diagrams. Isolith maps are a quantitative measure of sandstone distribution and provide a gross representation of the area occupied by nearshore depositional environments. A value of approximately 85 API units was used to discriminate relatively sandstone-rich from mudstone-rich facies on gamma-ray logs.

Because isolith maps show only net sandstone thickness and distribution, they do not provide a detailed picture of the organization of the individual sandbodies that can be recognized within each allomember. In order to illustrate the stacking pattern of all the component sandstone bodies, fence diagrams, drawn to scale from well-log cross-sections, were used to map the small-scale stratal architecture within each allomember.

7.2 The Joli Fou-Skull Creek Seaway (Joli Fou Alloformation)

Approximately 102 to 104 Ma, during the early late Albian (Early Cretaceous), eustatic sea-level rise resulted in the Polar and Tethyan oceans flooding the Western Interior Basin of North America from the Polar Ocean to the Gulf of Mexico, to form the Joli Fou-Skull Creek Seaway (Williams and Stelck, 1975) (Figure 7.1). In the United States, mudstone units equivalent to the Joli Fou Formation include the Skull Creek Shale of southern Manitoba and Montana, the Thermopolis Shale of central Wyoming, and the Kiowa Shale of eastern Colorado (McGookey et al., 1972). The mudstones that were deposited in the Skull Creek-Joli Fou Sea
Figure 7.1: Representation of the early late Albian Joli-Fou/Skull Creek Seaway. During this time (approx. 102 to 104 Ma), the seaway formed by the merging of the Polar (from the north) and Tethyan (from the south) oceans. Map from Blakey (cpgeosystems.com).
contain a mixed assemblage of both fully marine Boreal and Tethyan fauna, confirming the connection of the two oceans during the early late Albian (Walaszczyk and Cobban, 2016).

Within the study area, the Joli Fou alloformation forms a sheet-like body of mudstone deposited in an open marine setting. The Joli Fou alloformation consists of at least two subtle siltier-upward successions, separated by surface JE1, first recognized by Vannelli (2016) in the adjacent study area to the north. Surface JE1 has also been traced over the present study area, showing that deposition of Joli Fou mudstone took place during at least two cycles of relative sea-level change.

To the north of the study area, Vannelli et al. (2017) show that the Joli Fou alloformation gradually thins and onlaps northwestward onto a topographic feature termed the 'Smoky River Ridge', which separates the Joli Fou depocentre in the east from the Paddy depocentre in the west. Roca (2007) and Roca et al. (2008) interpreted both the Joli Fou and Viking alloformations to pass laterally westward into terrestrial sediments. However, the palynological evidence presented by Mellon (1967) showed that late Albian rocks are absent in the central Foothills. Therefore the Joli Fou and Viking strata must pinch out to the west against underlying early Albian rocks of the Gates Formation.

7.3 Middle Albian Regression (Viking Alloformation)

Eustatic fall during the middle late Albian resulted in partial retreat of the Joli Fou Seaway and closure of the southern connection to the Gulf of Mexico (Dolson and Muller, 1994) (Figure 7.2). The fall in sea-level resulted in the progradation of coastal depositional systems that formed multiple, geographically isolated shallow marine sandstone bodies encased in offshore mudstone (Roca et al., 2008). These sandstone bodies include those of the Viking alloformation. In the study area, the Viking alloformation is composed of four allomembers: VA, VB1, VB2, and VD. Allomembers VA and VB consist of multiple upward-coarsening successions bounded by marine flooding surfaces, each of which suggest repeated transgressive and regressive events (Boreen and Walker, 1991). The detailed history of transgressive-regressive events, expressed as stacked parasequences and bounding erosion surfaces, is summarized below.
Figure 7.2: Late Albian regression. During this time, progradation of Viking and coeval deltaic systems into the basin resulted in a major narrowing of the seaway. Eventually, the seaway closed to the south, but remained open in the north. However, Vannelli et al. (2017) concluded that the Pelican deltaic system prograded from the eastern margin of the seaway and connected with the Paddy and Viking systems along the western margin of the seaway. Red arrow shows approximately where the Pelican delta was located. Map from Blakey (cpgeosystems.com).
### 7.3.1 Allomember VA

Allomember VA, the oldest allomember of the Viking alloformation, is bounded by discontinuities VE0 and VE1. Although the isopach map of VA (Figure 6.3) gave a broad overview of allomember geometry, it did not reveal the complex internal stratigraphic architecture and facies distribution. Sandstone isoliths, exploded cross-sections, and 3-D fence diagrams are here used to illustrate the internal stacking pattern of sandstone units. The net sandstone isolith map of allomember VA (Figure 7.3) shows that most sandstone was deposited in the northwestern part of the study area. However, the net sandstone isolith map does not reveal the distribution of individual sandstone bodies within VA. To illustrate the internal stacking pattern of sandstones within allomember VA, a 3-D fence diagram was constructed (Figure 7.4) combining multiple ‘exploded’ cross-sections. Sandstone units were distinguished by correlation of internal markers within allomember VA. Analysis of the fence diagram reveals at least four distinct sandstone units in allomember VA: these are here termed VA α, VA β, VA γ, and VA δ. The spatial distribution of the four sandstone units in VA is given in Figure 7.5.

Sandstone VA α is the oldest sandstone unit and also has the most limited areal distribution, being confined to the northwestern side of the study area. The shape of the sand body is relatively straight rather than lobate, which suggest wave action was probably an important influence on sediment dispersal during the deposition of VA α. Core and well-logs reveal that VA α is sharp based (See Fig. 7.5 for core log and Cross-section 2 (Appendix A) for well-log signature) and therefore it is likely that the unit was deposited during a relative fall in sea level that followed the VE0 transgression (Plint and Nummedal, 2000). The western margin of the sandstone unit is not currently known because it extends outside the study area.

Sandstone VA β is also restricted to the northwest corner of the study area (Figure 7.5). A more detailed view of stratal architecture in Township 56 and Range 4W5 is shown in Figure 7.6, which shows that VA β laps out and is partially truncated in Range 4W5 by VA γ; therefore, the original southern extent of VA β is not known. The resulting shape of the VA β sand body is moderately elongated compared to VA α. Analysis of core- and well-logs show that VA β is also sharp based (See cross-section 2 (Appendix A); wells 06-20-056-08W5, 11-15-056-05W5, and 11-14-056-04W5; Cross-section C; 11-09-055-26W4) where it can be seen that massive to
Figure 7.3: Net sandstone isolith map of allomember VA showing the collective thickness of sandstone in all component parasequences. The sandstone reaches a maximum thickness of 23 m in the northwest of the study area. Contour interval is 2 m.
Figure 7.4: Fence diagram for allomember VA illustrating the internal complexity and stacking pattern of sandstones within the allomember. Three sandstone units, VA $\alpha$, VA $\beta$, and VA $\gamma$ form a stacked succession in the northwestern part of the study area and VA $\delta$ is in a basinally-isolated position in the southeast.
Figure 7.5: Distribution of the four sandstone bodies mapped within allomember VA. Core logs representative of each sandstone are illustrated and their position within the study area is indicated by the coloured circle. Bounding surfaces VE0 and VE1 are indicated on core-logs. A-A’ and B-B’ represent the core cross-section presented in section 7.4.
Figure 7.6: Exploded cross-sections of Township 56, Township 52, and Range 4W5. The dotted lines indicate the intersection point between the three cross-sections. This diagram is used to illustrate the internal stacking patterns of the VA sandstone bodies.
cross-bedded sandstone (facies 6) overlies Viking mudstone (facies 2). The coarse grained nature and sedimentary structures indicate that this unit was probably deposited close to a river mouth. The sharp-based sandstone of VA β indicates that progradation of the deltaic shoreface probably took place during relative sea-level fall (e.g. Plint & Nummedal, 2000).

Both the landward and seaward limits of sandstone VA γ are mapped within the study area (Figure 7.5). VA γ is the largest and thickest sandstone unit found in allomember VA. The unit is a relatively lobate, southeasterly prograding sand body (see Figure 7.5, Township 52 and Range 26 cross-sections). The change in the shape of the sand bodies from VA α to VA γ may indicate a possible change of interacting controls such as the effect of waves, rivers, tides, or delta lobe switching. Wave dominated deltas tend to have a more arcuate shape parallel to the coastline whereas river dominated deltas tend to be more lobate; however, most deltas are likely to be mixed influenced and show multiple characteristics of interacting controls (Bhattacharya, 2010). The more lobate shape of VA γ could suggest that the river feeding the sediment into the system had a greater influence on delta shape rather than wave action. The seaward shift of sand deposition could have been the result of a change in location of the river supplying the sediment, or perhaps sea-level did not rise as much during transgression compared to the previous rise in sea-level.

The youngest sandstone unit is VA δ. The geographic location of VA δ represents a significant basinward shift of the system. Core logs and well-logs (see cross-section 4, wells 10-16-048-17W4, 03-20-048-20W4; cross-section C, wells 16-02-044-26W4, 06-12-043-26W4) show that VA δ is a sharp-based sandstone unit, which suggests deposition during a relative fall in sea level that occurred following the deposition of VA γ.

Major transgressive surface VE1 follows the deposition of VA δ and is the top bounding surface of allomember VA. The relative rise in sea-level of VE1 caused a landward shift of the system back to the west where the first sandstone of VB1 was deposited.

An interpreted relative sea-level curve (Figure 7.7) has been drawn to illustrate the changes in sea level that are interpreted to have occurred during Viking VA, VB1, and VB2 time.
Figure 7.7: Relative sea-level curve for VA, VB1, and VB2. Displayed on the curve are the proposed internal sea-level changes for the allomembers inferred from the stacking pattern and distribution of sandstones in allomembers. The major transgressive/ravinement surfaces are highlighted in red.
7.3.2 Allomember VB1

Bounding discontinuity VEM splits allomember VB into two distinct units. VB1 is the older of the two and it is bounded by VE1 and VEM. Figure 7.8 shows that most sandstone within VB1 was deposited in the mid- to north-central part of the study area. A 3-D fence diagram of VB1 was constructed to highlight the internal stratal architecture of the allomember (Figure 7.9). The stacking patterns within VB1 are complex and are interpreted to contain at least 3 separate sandstone units: VB1 α, VB1 β, and VB1 γ. A map of the three sandstone bodies in VB1 is given in Figure 7.10.

Sandstone VB1 α is the oldest sandstone unit and was deposited following the VE1 transgression that moved the Viking shoreline landward. The shape of VB1 α is irregular but is mostly lobate indicating some degree of river influence. Analysis of core- and well-log (see Appendix A, cross-section 4, well 08-24-048-04W4) data reveals that VB α is locally sharp based with cross-bedded sandstone deposited directly onto Viking mudstones (facies 2) (Figure 7.10 core log 08-14-048-04W5). Other well-logs of VB1 α (see cross-section 4, well 06-24-048-25W4) that are located geographically further seaward, show that the base of the sandstone unit is gradational. It is possible that initial deposition of VB1 α occurred during a fall in sea level (depositing the sharp based sandstones on mud) but subsequently sea-level ceased to fall. Enough sediment was then supplied to the shoreline that the system continued to prograde eastward depositing gradational upward-coarsening succlusions as no new accommodation was being created, which is supported by the core-logs of VB1 β (Fig. 7.10, well 10-22-048-17W4) being gradationally-based and upward-coarsening.

VB1 β is the second sandstone body within allomember VB1. It is hypothesized that there was a small transgression following the deposition of VB1 α that led to the deposition of marine mudstone (see Township 48 cross-section in Figure 7.9). Sea-level then fell to its lowest point relative to other VB1 sandbodies. VB1 β was then deposited during this lowstand and the sandstone unit prograded southeastward to an unknown extent outside the study area. The shape of VB1 β is not fully determined and therefore the relative importance of wave or river dominance is not known.
Figure 7.8: Net sandstone isolith map of allomember VB1. VB1 reaches a maximum thickness of 28 m. Contour interval is 4 m.
Figure 7.9: Fence diagram of allomember VB1 illustrating the internal stacking patterns of VB1 sandstone units. VB1 contains three units: VB1 α, VB1 β, and VB1 γ. VB1 β displays a marked basinward shift relative to VB1 α and VB1 γ is a large backstep following β. A similar trend was observed in allomember VA.
Figure 7.10: Distribution of the three sandstone bodies within allomember VB1. Core logs representative of each sandstone is illustrated and their position within the study area is indicated by the coloured circle. Bounding surfaces VE1 and VEM are indicated on core-logs to show the boundaries of the allomember. This map illustrates the dramatic basinward shift of the location of sandstone deposition from $\alpha$ to $\beta$ and the landward shift from $\beta$ to $\gamma$. 
The youngest sandstone unit in allomember VB1 is sandstone VB1 γ. Following the deposition of VB1 β at maximum lowstand, sea-level rose, causing a significant landward shift of the system. When sea-level rise halted, VB1 γ prograded east/southeastward into the basin. A major river mouth that supplied sediment to the VB1 γ system may be inferred from the conglomerate unit observed in core 01-36-054-01W5 (Figure 7.10). The overall lobate shape of the sand body indicates that the river that fed the conglomerate to the shore had a large effect on the shape of the sand body.

### 7.3.3 Allomember VB2

Allomember VB2 is bounded by surfaces VEM and VE3. An isolith map of net sandstone within VB2 is shown in Figure 7.11. Internally, VB2 contains a thick stacked succession of four sandstone bodies that are distinct and mappable. Figure 7.12 illustrates a 3-D fence diagram of VB2 showing the cross-section of Township 48 contains at least three separate sandbodies that are distinguishable: VB2 α, VB2 β, and VB2 γ. The fourth sandstone, VB2 δ, is located in the northwest of the study area. The geographical distribution of the sandstone bodies within VB2 is illustrated in Figure 7.13.

VB2 α was deposited following the transgression represented by surface VEM. VB2 α is a southeastward prograding unit that is relatively limited in lateral extent compared to the other sandstone bodies in VB2. VB2 α reached a thickness of approximately 15 m in Range 26W4 (see Figure 7.13; well log 05-22-048-26W4).

VB2 β is the second sandstone body within allomember VB2. Well-logs in cross-sections 4 (Township 48) and C (Range 26W4) show that VB2 α and β are separated by a prominent flooding surface, indicating that there was probably a relative rise in sea-level before VB2 β was deposited. The shape of the sand body is rounded and lobate suggesting significant river influence. Wells 06-03-048-21W4 and 16-26-042-26W4 show that VB2 β was dominated by massive to cross-bedded sandstones with relatively sharp bases. The coarse-grained sand indicates that deposition was closer to the source (river mouth) and the sharp lower contact suggests that sea level was falling as the system was prograding eastward.
Figure 7.11: Net sandstone isolith of allomember VB2. The locus of sandstone deposition has shifted southward relative to VB1. VB2 reaches a maximum thickness of 18 m at approximately T48 and R4W5. Contour interval is 2 m.
Figure 7.12: Fence diagram of allomember VB2 illustrating the internal stacking patterns of the sandstone units. VB2 α, β, and γ are restricted mainly to the southern to southeastern section of the study area whereas VB2 δ is located primarily in the northwest.
Figure 7.13: Distribution of the four sandstone bodies that form allomember VB2. Core logs representative of each Sandstone are illustrated and their position within the study area is indicated by the coloured circle. Bounding surfaces VEM and VE3 are indicated on core-logs.
Sandstone VB2 \( \gamma \) is the third unit of the stacked sandstones present in allomember VB2. VB2 \( \gamma \) represents a seaward shift of the depositional system. Cross-section 4, wells 01-10-048-24W4 and 08-12-048-23W4 show that VB2 \( \gamma \) is relatively thin and sharp-based.

The core-log of well 16-26-042-26W4 illustrates that VB2 \( \gamma \) is composed of mostly massive to cross-bedded sandstone and is separated from VB2 \( \beta \) by a distinctly coarser-grained bed, indicating a change in sea level (ravinement) between the deposition of the two sandstone bodies. A relative rise in sea-level halted the progradation of VB2 and caused a landward shift of the depositional system. VB2 \( \delta \), the uppermost VB2 sand unit, was deposited in the northwestern section of the study area following this transgression. VB2 \( \delta \) is composed of small eastward-prograding sandstone units.

Bounding discontinuity VE3 is a transgressive erosion surface that defines the top of allomember VB2. Following the VE3 transgression, delivery of sand to the study area diminished greatly, and no major sandstone bodies are present in the overlying allomember VD.

7.3.4 Allomember VD

Viking allomember VD, bounded by surfaces VE3 and VE4, indicates the beginning of a major transgression of the Western Interior Seaway (Figure 7.14). In the study area, VD comprises a sheet of mostly marine mudstone and thin sheets of sandstone located in the northwest and southeast. A sandstone isolith map (Figure 7.15) and 3-D fence diagram (Figure 7.16) have been created to illustrate the distribution of the thin sandstone units within the study area.

Allomember VD represents a time of renewed tectonic loading in the Cordillera. While unit VD is a relatively thin unit in this study area, the allomember thickens into two depocentres located in the northwest and southwest along the thrust belt (Rylaarsdam, (2006); Buckley (2009); Angiel (2013); Plint et al., 2016). The relatively constant thickness of VD in this study area indicates that it was probably situated on the forbulge between the two depocentres, where minimal subsidence greatly limited deposition. Contemporaneous shoreface sandstones in VD are restricted to SW Alberta (Roca et al. 2008).
Figure 7.14: A paleogeographic representation of the Western Interior Seaway broadly representing Late Albian Viking (VD) time. Sea level has risen and a majority of the interior is flooded by the sea. Map from Blakey (cpgeosystems.com).
Figure 7.15: Sandstone Isolith map of allomember VD. Sandstone units are thin and dispersed throughout the study area. In the southeast, the larger sandstone unit extends into the study area of Drljepan (In progress) where it becomes much thicker. Contour interval is 1 m.
Figure 7.16: Fence diagram of Viking allomember VD. This allomember is dominated by mudstone and has a relatively constant thickness throughout the study area. Thin and localised sandstone bodies can be seen in the northwest and southeast portions of the study area.
VE4 marks the top of allomember VD and also the Viking alloformation. VE4 represents a marine transgressive surface that caused a significant landward shift of the shoreline that was followed by deposition of the Westgate alloformation.

### 7.3.5 Westgate Alloformation

Transgression of the sea during the late late Albian led to a major flooding of the Western Interior of Canada (Figure 7.17). The shoreline migrated to the west of the study area; Angiel (2013) mapped nearshore deposits in the vicinity of the foothills to the west of Chetwynd, BC. The Westgate alloformation in the present study area is dominated by mudstone that is interpreted to represent offshore environments of the Mowry Sea, which was a southern incursion of the Polar Ocean during the late late Albian (Kauffman and Caldwell, 1993). The mudstones of the Westgate Formation are in part equivalent to the Shaftesbury Formation in northwestern Alberta and the Mowry Shale throughout the extent of the basin in the U.S.

### 7.4 Core Cross-sections

#### 7.4.1 Introduction

Figures 7.5, 7.10, and 7.13 displayed individual core logs of Viking sandstone bodies. Core log cross-sections have been created to illustrate the spatial variation of the sand bodies throughout the study area. These core logs and cross-sections provide further evidence to support the interpreted sea-level history. Gradational or sharp facies transitions reflect sea-level rise or fall and normal or forced regression.

The core-logs in the cross-sections were chosen in order to display as many facies as possible as well as show ample representation of each individual sandstone body. This was difficult because a large number of cored intervals were relatively short and did not span the entire Viking unit; however, the longest cores have been included in the cross-sections together with multiple smaller cores to supplement the others.
Figure 7.17: Late Late Albian transgression of the Seaway (Mowry Sea) that occurred during Westgate time (approx. 98 Ma). The seaway extends into the US but does not connect with the southern ocean. Map from Blakey (cpgeosystems.com).
7.4.2 Core Cross-section A-A’

7.4.2.1 VA Sand Bodies

Core core-section A-A’ spans the northern part of the study area and best displays the spatial variation of sandstone bodies within VA (Figure 7.18). The solitary core that contained VA α was in well 07-20-056-06W5. This well shows that VA α was a relatively thin sandstone unit (approx. 2 m). This sandstone was sharp-based (sharp contact from mudstone to sandstone) and was capped by a transgressive lag. VA α was dominated by cross-bedded sandstone (facies 6) and contained some plant debris. The sharp-based nature of this sandstone suggests that the body was deposited during a relative fall in sea-level.

Sandstone VA β is present in three core-logs in cross-section A-A’ which shows that VA β becomes less sandstone dominated as the delta prograded eastward. One cored example of VA β demonstrates a relatively gradational base (well 7-20-056-08W5), which suggests that deposition of this sandstone occurred when the supply of sediment was greater than the accommodation being created; however, other examples of VA β, including the core of well 04-20-057-26W4 and well-logs from cross-section 2 (Appendix A) wells 11-15-056-05W5, 11-14-056-04W5 and cross-section C well 11-09-055-26W4 show VA β to be sharp-based. The sharp base of some of the sandstone units indicates the possibility of deposition of VA β during relative sea-level fall.

Sandstone VA γ is present in two cored wells along A-A’. The sandstone thins eastward from 5.5 m to 0.5 m in 17 km. The difference in VA γ thickness between the two wells is thought to be due to proximity to the sediment source. Well 04-20-057-26W4 is the thicker VA γ sand and is composed of poorly sorted fine sand with granules. The sandstone also contains various mud drapes and cross-bedding. The mud-drapes could indicate a strong fluvial influence as described by Bhattacharya and MacEachern (2009). The VA γ sandstone in well 13-03-056-24W4 is well-sorted fine sand but also contains mud drapes and current ripples, indicating there is still an influence of fluvial processes. Both VA γ sandstones in cross-section A-A’ are sharp-based indicating that deposition likely occurred during a relative fall in sea level.
Figure 7.18: Core cross-section A-A' composed of core-logs from the northern half of the study area.
The eastern core-logs displayed in cross-section A-A’ show that sand was not transported that far east during VA time. East of Range 20, the sediment is dominantly mud and silt. VA δ, the youngest VA sand body is not present in cross-section A-A’ because it is located in the southern half of the study area.

7.4.2.2 VB1 Sand Bodies

VB1 sand bodies α and γ are represented in core cross-section A-A’. In well 06-21-059-1W5, VB1 α is an approximately 3 m thick, poorly sorted conglomerate, likely deposited very close to a river mouth and the sediment source for this sand body. Well 04-20-057-26W4, located 26 km away from the previous well, is shown to have been deposited sharply on VE1 and is composed of cross-bedded coarse grained sand. The location of this second core is to the southeast of 059-01W5 and is therefore further from the sediment source; however, due to the coarse grained nature and sedimentary structures present in core 04-20-057-26W4, it is possible that a second river also fed sediment into the system. This second river system then became the dominant river system by VB1 γ time because in core 04-20-057-26W4, a 2 m thick conglomerate unit was deposited presumable indicating proximity to a river mouth. It is important to note that there is a significant amount of time missing in core 04-20-057-26W4 between α and γ. VB1 β (represented in cross-section B-B’) was deposited in the southeast part of the study area at the maximum lowstand for VB1 time. Sea level rose following the deposition of VB1 β and the shoreline moved landward to the point where the VB1 γ sand body could build out into the basin. The VB1 γ sand body in wells 04-20-057-26W4 and 13-03-056-24W4 is sharp-based and are therefore interpreted to have been deposited during a relative fall in sea level.

7.4.2.3 VB2 Sand Bodies

VB2 δ is the lone sand body from VB2 represented in core cross-section A-A’. VB2 δ is a combination of small eastward-prograding sandstone units that are mostly disconnected. VB2 δ is represented by a thin conglomerate in well 06-21-059-01W5 and by medium to coarse cross-bedded sand in wells 07-20-056-08W5 and 13-03-056-24W4. The discontinuity represented by VB2 δ could be the result of more extensive erosion from the major transgression of VE3.
7.4.3 Core Cross-section B-B’

7.4.3.1 VA Sand Bodies

VA δ is a large sandstone unit that was deposited in the southeast part of the study area and is the only VA sand body represented in core cross-section B-B’ (Figure 7.19). Following the deposition of VA γ, there was a fall in sea level and the shoreline reached a maximum lowstand and furthest extent into the basin. VA δ was deposited during this lowstand. Three core-logs in cross-section B-B’ contain sandstone VA δ. The sandstone ranged from very-fine to medium-grained sand. The medium-grained sand is present in well 10-22-048-17W4 and is therefore likely closer to the source of the sediment. All three core-logs show that the VA δ sandstone has fairly thorough bioturbation (facies 5), especially well 08-06-044-26W4 where the bioturbation index is 6. Core logs from cross-section B-B’ and well-logs (see cross-section 4, wells 10-16-048-17W4, 03-20-048-20W4; cross-section C, wells 16-02-044-26W4, 06-12-043-26W4) show that VA δ is a sharp-based sandstone unit, which indicates deposition during a relative fall in sea level that occurred following the deposition of VA γ.

7.4.3.2 VB1 Sand Bodies

Similar to VA δ, VB β is hypothesized to have been deposited during a maximum lowstand when the shoreline shifted basinward. VB β is depicted in multiple core-logs in cross-section B-B’. The thickest sandstone appears in well 08-27-047-022W4. It is an 11 m thick, extensively bioturbated package of fine sandstone geographically located close to the hypothesized landward limit of VB β. To the southwest (58 km), in well 08-06-044-26W4, VB β is much thinner (1 m) and a finer grainsize yet still contains bioturbation. To the east (58 km), in well 10-22-048-17W4, VB β is 9 m thick and composed of fine to medium-grained sandstone with extensive bioturbation. Grain size increases at the top of the sandstone where there is evidence of cross-bedding. It is hypothesized that the main sediment feeder for this system was likely located close to well 10-22-048-17W4.
Figure 7.19: Core cross-section B-B’ composed of core-logs from the southeastern part of the study area.
7.4.3.3 VB2 Sand Bodies

VB2 sand bodies β and γ are well represented in core cross-section B-B’. The younger of the two, β, is shown in four core-logs within the cross-section showing that VB2 β is composed fine to coarse-grained sand. The coarsest sand occurs in well 06-25-046-20W4 where there is also evidence of cross-bedding and localized bioturbation. The sandstone unit in well 06-25-046-20W4, 10-22-048-17W4, and 08-27-047-22W4 is sharp-based, which indicates that VB2 β was likely deposited during a relative fall in sea-level. More examples of the sharp-based nature of this sand body can be seen in cross-sections D, 4, and 5 (Appendix A). The same sharp-based lower contact can be seen in the core-logs of VB2 γ. The sandstones of VB2 γ are represented by very fine- to medium-grained sand with the presence of moderate bioturbation, wave-ripples, cross-bedding (06-25-046-20W4), and mud drapes (08-27-047-22W4).

VB2 α was not observed in core; however, it is hypothesized that between the deposition of VB2 α and β, and β and γ, that there were cycles of sea level rise and fall. VB2 α, β, and γ were each deposited during a fall in sea-level following a previous transgression.

7.5 Discussion of Viking Sea-level changes

The internal organization of sandstone bodies within Viking allomembers VA, VB1 and VB2 allow both the evolving geography of the basin to be reconstructed, and also the succession of relative changes in sea-level to be inferred. The sandstone units within allomembers are separated by traceable flooding surfaces that are attributed to allogenic processes, specifically changes in sea-level. Coarse sandstone beds or granule lags mantle key surfaces that record marine transgression and significant shoreline backstep. A representative sea-level curve (Fig. 7.7) was produced to illustrate the interpreted changes in sea-level throughout allomembers VA, VB1, and VB2. With the aid of cross-section well-logs and core-logs, the nature of the timing of deposition (during sea-level fall, lowstand, etc.) for each sand body within the allomembers was determined. A sandstone body with a gradational-base suggests that it was deposited during 'normal regression' where supply of sediment was greater than the accommodation being created, whereas sharp-based sandstone is evidence for deposition of the sand body during relative sea level fall, or ‘forced regression’. Analysis of the
sand bodies mapped in this thesis shows evidence for both types of regression; however, forced regression appears to have been the most common.
Chapter 8: Conclusions

8.1 Conclusions

The findings of this thesis can be summarized in 6 conclusions:

1. The Lower Colorado Group is a mudstone-dominated package of rock that was deposited into the Western Interior Seaway during the Late Albian. The study area for this thesis is located in north-central Alberta (Edmonton area) spanning approximately 26,000 km². The Lower Colorado Group only occurs in subsurface within this study area and the average thickness of the group is 75 m. The Lower Colorado Group can be lithostratigraphically subdivided into four formations: the Joli Fou (oldest), Viking, Westgate, and Fish Scales (youngest) formations.

2. The rocks of the Lower Colorado Group can be divided on allostratigraphic principles, constituting the Lower Colorado allogroup, composed of the Joli Fou, Viking, Westgate, and Fish Scales alloformations. Ten bounding discontinuities have been traced throughout the study area and also into other regions of Alberta to establish an allostratigraphic framework throughout the basin.

3. Rocks of the Lower Colorado allogroup within the study area record marine depositional environments represented by nine facies. The facies represent depositional environments ranging from offshore marine to lower shoreface through to upper shoreface and river mouth environments.

4. Isopach maps of the Joli Fou, Viking, Westgate, and Fish Scales alloformations has allowed the recognition of both tectonic and eustatic controls on accommodation. The Joli Fou alloformation has a near-tabular geometry that is most reasonably interpreted as the result of eustatic rise during the Skull Creek transgression. Viking allomember VA displays a progressive thickness change from the southeast to the northwest. This pattern is not readily explained as a consequence of tectonic loading because the thick zone is of rather localized distribution, which is not the typical ‘wedge-shaped’ geometry. Local thickening may most reasonably be attributed to localized
sedimentation around a river mouth. When compared to other areas, the allomember is broadly tabular, suggesting eustatic control.

Viking allomember VB thins towards the northwest whereas coeval Pelican allomembers PeC and PeD thin towards the southwest against a linear region that underwent slow subsidence or even uplift. Neither the Viking or Pelican sand bodies form wedge-shape bodies that thicken towards the cordillera, suggesting that the original accommodation was probably controlled primarily by sea level change.

Within the study area, Viking allomember VD has a tabular geometry suggesting that accommodation was controlled primarily by eustatic sea level change; however, studies in adjacent areas have shown that VD thickens to the northwest and southwest into two distinct depocentres indicative of renewed flexural subsidence in both areas.

Westgate allomember WB is a northwest-thickening wedge, indicating that accommodation was controlled primarily by tectonic loading in northeastern British Columbia. Westgate allomember WC thickens to the east and the accommodation is inferred to have been strongly controlled by flexural subsidence.

Fish Scales allomember FB produced an isopach with a gradual thickening to the west, which continues to thicken outside of the study area. Flexural subsidence was the main control of accommodation for this allomember; however, it is likely that sea-level rise generated some space for sediment and produced near tabular geometry of FB in this study area.

5. Viking allomembers VA, VB1 and VB2 can be divided into multiple parasequences separated by flooding surfaces. The stacking pattern of sandstones within parasequences allowed a detailed relative sea-level curve to be constructed.

Allomember VA contains four distinct internal sand bodies that were deposited during successive sea level regressions. VA sand bodies α, β, and γ become progressively more lobate as a result of the river system feeding the sediment into the system developed a greater influence on delta shape. VA δ represents a significant basinward shift of the system and the sand body was deposited at lowstand evident by the sharp-based nature of the lower contacts.
Allomember VB1 contains three internal sand bodies that were deposited during separate sea-level cycles. Following VE1 there was sea-level regression which resulted in the deposition of VB1 α. There was then a relatively small sea-level rise followed by a fall to lowstand and basinward shift of the system as VB1 β is deposited. Transgression then occurred and shifted the system. Sea-level dropped again and deposition of VB1 γ occurred. Major transgression VEM then marks the end of allomember VB1.

Allomember VB2 contains four internal sand bodies. Analysis shows that VB2 α, β, and γ get progressively larger after each small sea-level cycle. The lobate shape of these three sand bodies indicates a strong influence from the river system/systems that fed sediment into the basin. Following the deposition of VB2 γ, there was a transgression and the shoreline shifted landward. VB2 δ was deposited in the northwestern section of the study area. VB2 δ is a combination of small eastward-prograding sandstone units. Bounding discontinuity VE3 is a transgressive erosion surface that defines the top of allomember VB2. Following the VE3 transgression, delivery of sand to the study area diminished greatly, and no major sandstone bodies are present in the overlying allomember VD.

6. The allostratigraphic framework makes it possible to show that: The Joli Fou is time equivalent to the Lower Paddy alloformation in the northwestern Alberta and adjacent British Columbia and to pass laterally into the Bow Island Formation in the southern plains of Alberta. The Viking Formation is partially time equivalent to the Bow Island Formation in Southwestern Alberta, the Pelican Formation in northeastern Alberta, and the upper Paddy alloformation of the Peace River Formation in northwestern Alberta.

8.2 Suggestions for future work

1. Extend the regional allostratigraphic framework eastward from this study area. The seaward limit of lowstand deposits VA δ and VB1 β could possibly be determined.
2. More research into the Westgate alloformation could prove beneficial. There are multiple surfaces within the alloformation that could be correlated regionally to give a more detailed understanding of the unit.
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Publications:
Appendix A

The following pages contain cross-sections 1-6, oriented E to W and cross-sections A-E, oriented N to S.