Stratigraphy and Sedimentology of the Late Cretaceous (Coniacian) Muskiki and Marshybank Members, Southern Alberta and Northwestern Montana

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Graduate Program in Geology
A thesis submitted in partial fulfillment of the requirements for the degree in Master of Science
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STRATIGRAPHY AND SEDIMENTOLOGY OF THE LATE CRETACEOUS (CONIACIAN) MUSKIKI AND MARSHYBANK MEMBERS, SOUTHERN ALBERTA AND NORTHWESTERN MONTANA

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Meriem GRIFI

Graduate Program in Geology

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science

The School of Graduate and Postdoctoral Studies
The University of Western Ontario
London, Ontario, Canada

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The thesis by
Meriem Grifi
entitled:

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Abstract

A high-resolution allostratigraphic study of the Coniacian Muskiki and Marshybank members of the Wapiabi Formation in southern Alberta revealed a southwest thickening wedge of mudstone-dominated strata that was deposited on a shallow storm-dominated shelf. Well-log correlations and biostratigraphy show that the Muskiki Member forms the bulk of the succession; the Marshybank Member is thin or absent in the subsurface. The lower and middle units of the Muskiki Member display regional subsidence patterns consistent with thrust sheet loading in the Cordillera. The upper unit comprises a linear trough filled by southeastward-accreting clinothems. The northeast boundary of the clinoform package corresponds with an Archean thrust fault that may have undergone extensional reactivation during the Coniacian, forming a local trough. An isopach map of the basal Santonian strata shows thinning coincident with the Vulcan magnetic anomaly, suggesting differential subsidence across the structure.

Keywords

Cretaceous, Wapiabi Formation, Muskiki, Marshybank, Coniacian, biostratigraphy, Kevin Member, MacGowan concretionary bed, clinoforms, Orion Low, Vulcan Structure, mud floccules, petrography, extensional faulting, foreland basin
For my grandmother,

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CHAPTER 1

INTRODUCTION TO THE MUSKIKI AND MARSHYBANK MEMBERS

1.1 Introduction

Mudstones account for more than 70% of the world’s sedimentary rocks, yet they remain poorly understood (Aplin et al., 1999). Traditionally viewed as sediments that accumulated in deep, quiet water settings (e.g. Potter et al., 2005), recent studies have shown that muds form aggregates that can be transported and deposited in shallower settings and as a result of faster flows, thus behaving hydro-dynamically as sand grains (e.g. Schieber et al., 2007; Schieber and Southard 2009; Scieber and Yawar, 2009; Macquaker et al., 2010). The rock record shows that mudstones have a more complex depositional history than previously assumed (e.g Cattaneo et al., 2007; Liu et al., 2007; Plint et al., 2009).

The Upper Cretaceous Muskiki and Marshybank members of the Wapiabi Formation in southern Alberta consist of marine mudstones and siltstones that were deposited in the Western Interior Seaway that occupied the Western Canada foreland basin. These rocks lack significant hydrocarbon potential, making them of little interest to industry and the subject of few detailed academic studies (e.g. Rosenthal, 1984; Medioli, 1999; and Nielsen, 2003).

In view of the resurgence of interest in the stratigraphy and sedimentology of mudstones, the Muskiki and Marshybank members present an opportunity to gain a better appreciation for rocks that were previously regarded as unimportant to the understanding of tectonic and eustatic processes within foreland basins.

1.2 Purpose of the Study

The Muskiki and Marshybank members of the Wapiabi Formation in southern Alberta contain subtle yet laterally-persistent, coarsening-upward successions that are regionally traceable over an area of approximately 190 by 250 km (Fig. 1.1). The extent of these coarsening-upward successions suggests that autocyclic mechanisms alone are insufficient to explain the patterns of sedimentation within the study interval. A regional
Figure 1.1 Generalized map of North America with close up of Alberta and Montana showing thesis study area.
stratigraphic framework for the Muskiki and Marshybank formations in northwestern Alberta was established by Plint (1990). A single correlation line (Plint, unpublished data) was constructed in a preliminary attempt to link existing allomembers in the north with the new allostratigraphic scheme developed in this study.

The purpose of this study was to carry out a detailed stratigraphic and sedimentological analysis of the Muskiki and Marshybank members in southern Alberta. The main objectives of this project are to:

1) Establish a regional allostratigraphic framework for the Muskiki and Marshybank members in southern Alberta. A preliminary effort will be made to link with the allostratigraphic framework established further north (Plint, 1990, 1991; Plint et al., 1990; Plint and Norris, 1991);

2) Incorporate petrological data from outcrop and core into the regional stratigraphic framework

3) Use the better biostratigraphic control available in the southern part of the study area to extend biostratigraphic zones to the northern part of the basin. This will be done in collaboration with Ireneusz Walaszczyk (University of Warsaw, Poland);

4) Establish the stratigraphic relationship between the Muskiki and Marshybank members of the Wapiabi Formation in southern Alberta with the subsurface lithostratigraphic units of the Niobrara Formation in the Alberta Plains, as well as with parts of the Kevin Member of the Marias River Formation in northwestern Montana;

5) Produce isopach maps to investigate the relative influence of tectonism and eustasy on the geometry and sedimentology of the study interval; and

6) Reconstruct the evolving paleogeography of the study area during Coniacian time.

1.3 Stratigraphic Nomenclature

In southern Alberta, the Muskiki and Marshybank members represent the lowermost units of the Wapiabi Formation within the Upper Cretaceous Alberta Group (Stott, 1961, 1963). The members are named after tributary creeks of the Brazeau River, approximately 300 km northwest of Calgary. In the subsurface, east of the Foothills, the Muskiki and Marshybank members are presently considered to be equivalent to the
Verger Member and lowermost Medicine Hat Member of the Niobrara Formation, within the Upper Cretaceous Colorado Group (Nielsen, 2003; Nielsen et al., 2003).

The Muskiki and Marshybank members, on the basis of macrofauna, are broadly equivalent to the lower and middle Kevin Member of the Marias River Shale Formation in northwestern Montana (Cobban et al., 1976; Nielsen et al., 2003). Based on molluscan assemblages, Plint (1991) suggested that most of the Marshybank could be equivalent with the MacGowan Concretionary Bed and a few metres of overlying strata of the middle Kevin Member in northwestern Montana.

North of the Athabasca River, the Wapiabi Formation forms part of the Upper Cretaceous Smoky Group (Stott, 1967). The Muskiki was assigned formation status, whereas the term Marshybank was not extended northward, with the sandy strata above the Muskiki being assigned to the Bad Heart Formation (Stott, 1967). In a re-interpretation of the stratigraphy, Plint et al. (1990) confined the term Badheart to the mudstone, ooidal ironstone and sandstone deposits of the northwest Alberta Plains, and the Marshybank Formation was used to describe the massive siltstone and sandstone deposits in the central Alberta Foothills. A summary of the various stratigraphic schemes is given in Figure 1.2.

1.4 Biostratigraphy and Age

1.4.1 Macro- and Microfauna of the Western Interior

Stott (1963, 1967), documented the ammonite species *Scaphites preventricosus* in the lower part of the Muskiki, and *Scaphites ventricosus* in the upper part of the Muskiki and the lower part of the Marshybank. *Scaphites depressus* was present through most of the Marshybank Formation. The age assignments of these ammonites have varied over the years, ranging from Late Turonian to Santonian (Jeletzky in Stott, 1963). However, it is currently accepted that *Scaphites preventricosus*, *Scaphites ventricosus*, and *Scaphites depressus* characterize lower, middle and upper Coniacian strata, respectively (Cobban, 1993; Cobban et al., 2005).

In central Montana, Cobban (1951) collected the inoceramid *Inoceramus deformis* and the ammonite *Scaphites preventricosus* from the Niobrara Formation. At the Kevin outcrop in northwestern Montana, which is the type section for the Kevin Member of the
Figure 1.2 Stratigraphic nomenclature for the Muskiki and Marshybank members in southern Alberta and related areas. The southern Alberta Foothills terminology is used in this thesis. Radiometric ages from Siewert et al., 2012.
Marias River Formation, the ammonites *Scaphites preventricosus* (beds 7, 20, and 28) and *Scaphites mariasensis* (bed 20) were collected from the lower unit of the Kevin Member (Fig. 1.1; Cobban et al., 1976). These fossils indicate an early Coniacian age (Cobban et al., 2005), and suggest correlation with the upper part of the Cardium Formation and lower part of the Muskiki Member in southern Alberta. The middle unit of the Kevin Member contains *Inoceramus involutus* (beds 71, 84, and 96), and the ammonite *Scaphites ventricosus* (beds 71, 81, and 88), which are of middle Coniacian age (Cobban et al., 1976; Cobban et al., 2005). Nielsen (2003; Nielsen et al., 2003) correlated the Verger Member to the lower unit and lower part of the middle unit of the Kevin Member. The late Coniacian ammonite *Scaphites depressus* was collected from beds 104 and 106 at Kevin. *Scaphites depressus* was also found in allomembers B to L of the Marshybank Formation in northwestern Alberta, supporting further correlation between these rock units (Plint, 1991). A summary of the ammonite and inoceramid zones is shown in Figure 1.3a.

A microfaunal scheme for the Alberta Group was established by Caldwell et al. (1978). The foraminiferan *Pseudoclavulina sp.* was found to characterize part of the upper Turonian, whereas the *Trochammina sp.* and *Bullopora laevis* zones occurred within strata assigned to the Coniacian, based on correlation with molluscan faunal zones in the Western Interior (Jeletzky, 1971). On the basis of ammonite zonation, and micropaleontology work done in cores of the Verger Member in southern Alberta, Nielsen et al. (2008) established a new subzone: *Marssonella oxicorn*, within the *Pseudoclavulina sp.* zone. The boundaries of the *Pseudoclavulina sp.* zone were also shifted to overlap with late Turonian to early Coniacian ammonite zones (Nielsen et al., 2008). Although Nielsen et al. (2008) shifted the *Trochammina sp.* zone to the late Turonian, they acknowledged that this faunal assemblage was also documented in Coniacian strata of the Muskiki Member in the Foothills (Wall and Germundson, 1963; Wall and Rosene, 1977), which Nielsen (2003) correlated to the lower Verger Member. The disparity in zonation assignment implies that the surfaces in the subsurface may not have been correlated correctly to those in outcrop. This thesis places the *Trochammina sp.* zone within the late Turonian to early Coniacian stages, to follow fauna documented
<table>
<thead>
<tr>
<th>Stage/Substage</th>
<th>a. Macrofaunal Zonal Scheme</th>
<th>b. Foraminiferal Zonal Scheme</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>Ammonite</td>
<td>Inoceramid</td>
</tr>
<tr>
<td>Santonian</td>
<td></td>
<td></td>
</tr>
<tr>
<td>lower</td>
<td>Clioscaphites saxitoniannus</td>
<td>Cladoceramus undulatopicus</td>
</tr>
<tr>
<td>Coniacian</td>
<td></td>
<td></td>
</tr>
<tr>
<td>upper</td>
<td>Scaphites depressus</td>
<td>Megadiceramus crenelatus</td>
</tr>
<tr>
<td>middle</td>
<td>Scaphites ventricosus</td>
<td>Megadiceramus subquadratus</td>
</tr>
<tr>
<td>lower</td>
<td>Scaphites preverticosus</td>
<td>Vorticiceras koeni</td>
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</tbody>
</table>

Figure 1.3 a and b: Biostratigraphic zonal schemes for the Muskiki and Marshybank Formations, based on ammonite and inoceramid zones (a), and foraminiferal zones/sub-zones (b). Arrow indicates Santonian stage continues further. Macrofauna data from Cobban et al., 2005, 2006; Microfauna data from Caldwell et al., (1978), and Nielsen et al., (2003; 2008).
by Wall and Germundson (1963). A summary of the zones and integrated macrofaunal schemes is shown in Figure 1.3b.

1.4.2 Radiometric Dating of Late Cretaceous Strata

Several ages for late Turonian to upper Coniacian rocks have been determined from radiometric dating of bentonites. Sanidine and zircon crystals collected from bentonites interbedded with fossiliferous marine sediments allow relatively precise dating of sedimentary units. Obradovich (1993) dated sanidine crystals from the Scaphites preventricosus zone in the Marias River Shale in Montana using the $^{40}\text{Ar}/^{39}\text{Ar}$ laser fusion technique, and determined an age of $88.34 \pm 0.60$ million years (Ma). The Scaphites preventricosus zone was thought to be indicative of the middle Coniacian at the time, and only one other ammonite zone, Forresteria peruana, was assigned to the lower Coniacian. The average duration of an ammonite zone was assumed by Obradovich (1993) to span about 250,000 years, so he placed the Turonian-Coniacian (T-C) boundary at $88.7 \pm 0.5$ Ma. A second bentonite dated by Obradovich (1993) from the Late Coniacian Scaphites depressus zone gave an age of $86.92 \pm 0.39$ Ma. Obradovich assigned the Coniacian Stage a duration of 2.4 my, and placed the Coniacian-Santonian (C-S) boundary at $86.3 \pm 0.5$ Ma.

Ogg et al. (2004) obtained an age of $89.27 \pm 1$ Ma for the T-C boundary, and an age of $85.8 \pm 0.7$ Ma for the C-S boundary, giving the Coniacian stage a duration of 3.5 my, with an uncertainty of $\pm 1.7$ my (it should be noted here that Ogg et al. (2004) quote an uncertainty of $\pm 0.3$ my in their paper, however, they subtracted, instead of adding, their margin of error). The new date by Ogg et al. (2004) was acquired by fitting and refining the results of Obradovich (1993) to a cubic spline curve - a smooth curve that best fits a set of data points (Agterberg, 1994). Using cyclostratigraphy, Locklair and Sageman (2008) obtain a duration of 3.26 to 3.50 my for the Coniacian, supporting the work of Ogg et al. (2004).

Siewert et al. (2012) combined the $^{40}\text{Ar}/^{39}\text{Ar}$ method, using sanidine crystals, and the $^{238}\text{U}/^{206}\text{Pb}$ method, using zircon crystals, with astrochronology and biostratigraphy. They obtained an age of $89.65 \pm 0.28$ Ma for the T-C boundary and an age of $86.32 \pm$
0.32 Ma for the C-S boundary, suggesting a 3.33 my duration for the Coniacian. The dates obtained by Siewert et al. (2012) will be used in this thesis (Fig. 1.2).

1.5  Regional Cretaceous Paleogeography

Deposition of the Wapiabi Formation in southern Alberta took place at a paleolatitude of approximately 58-60°N (Irving et al., 1993). The Muskiki and Marshybank members in southern Alberta were deposited during the early part of the Niobrara marine cyclothem within the Western Canada sedimentary basin. The rocks represent transgressive-regressive cycles deposited on a low-gradient shallow marine shelf to shoreline setting (Plint and Norris, 1991; Kauffman and Caldwell, 1993). Warm southern Tethyan seawater mixed with cool water from the northern Boreal Sea, expanding the Niobrara Sea to its maximum limit during the early Coniacian (Caldwell et al., 1993). Volcanic activity took place during much of the Cretaceous, with particularly frequent eruptions occurring at the beginning of the Coniacian (McGookey et al., 1972; Roberts and Kirschbaum, 1995).

In terms of sea level rise, the Niobrara cycle was second in magnitude to the preceding Greenhorn cycle, and was interpreted to have experienced a sea level rise of 100 ± 50 m above present day levels (Miller et al., 2005). Unlike the Greenhorn cycle, the Niobrara Sea remained expanded, fluctuating very little, through a series of four major transgressive-regressive cycles that occurred in the Middle Early Coniacian, Late Coniacian, Middle Santonian, and earliest Campanian (Fig. 1.4; Caldwell et al., 1993).

In northwestern Alberta, deposition of the Muskiki Formation is interpreted to have taken place during relative sea level rise, followed by deposition of the Marshybank Formation during relative sea level high-stand and fall (Plint, 1990). The Bad Heart Formation, of the Northwest Alberta plains, is suggested to have been deposited during relative sea level lowstand (Plint, 1990). In the United States portion of the Western Interior, elevated sea levels resulted in deposition of chalks and marls of the Niobrara Formation (Finn and Johnson, 2005; Roberts and Kirschbaum, 1995). A paleogeographic map for the Coniacian Stage is shown in Figure 1.5.
<table>
<thead>
<tr>
<th>STAGES</th>
<th>KAUFFMAN'S GRAPH OF GLOBAL T-R CYCLES</th>
</tr>
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<tbody>
<tr>
<td>MAASTRICHTIAN</td>
<td></td>
</tr>
<tr>
<td>CAMPIANIAN</td>
<td></td>
</tr>
<tr>
<td>SANTONIAN</td>
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<td>CONIACIAN</td>
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<td>TURONIAN</td>
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<tr>
<td>CENOANIANIAN</td>
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<tr>
<td>ALBIAN</td>
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<td>APTIAN</td>
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<td>BARREMIAN</td>
<td></td>
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<tr>
<td>HAUTERIVIAN</td>
<td></td>
</tr>
<tr>
<td>VALANGINIAN</td>
<td></td>
</tr>
<tr>
<td>BERRIASIAN</td>
<td></td>
</tr>
</tbody>
</table>

Figure 1.4 Cretaceous stages with the various transgressive (T) and regressive cycles (R). Note the four transgressive-regressive (a-d) cycles of the Niobrara spanning late Turonian to earliest Campanian time. Modified from Kauffman, 1982.
Figure 1.5 Regional paleogeographic map (modified from Roberts and Kirschbaum, 1995) of the Western Interior during the Coniacian-Santonian stages of the Late Cretaceous. Paleolatitudes from Irving, 1993.
1.6 Previous Studies

1.6.1 Early Investigations

Dawson (1881) described the Cretaceous ‘Upper Shales’ along the Peace River, some 700 km to the north of the present study area. These ‘Upper Shales’ were described as grey to black shales with an abundance of ironstone, and occasional iron-rich septarian nodules. McLearn (1919) applied Dawson’s stratigraphic scheme along the Smoky River and subdivided the ‘Upper Shales’ into three units: a Lower shale, Middle sandstone, and an Upper shale.

The name Wapiabi was first used by Malloch (1911), who established, in ascending order, the Blackstone Shale, Bighorn Sandstone, and Wapiabi Shale successions in the central Alberta Foothills; the latter named after Wapiabi Creek in the Bighorn Range northwest of Calgary. Malloch described the Wapiabi rocks as brown, grey, or dark shales and acknowledged that they were similar to shales of the underlying Blackstone Formation, but different in that the rocks contained concretions.

1.6.2 Lithostratigraphy

Rutherford (1927) described dark to black shales of marine origin lying above the Cardium sandstones along the Bow River, assigned them to the Upper Benton Formation, and suggested that they correlated with the Wapiabi shales described by Malloch (1911) further north. The Upper Benton Formation was renamed the Wapiabi Formation by Webb and Hertlein (1934), who conducted geological surveys between township 5 (Mill Creek area) and township 55 (Berland River area). Webb and Hertlein noted the sharp contact between the dark grey Wapiabi shales with nodular brown ironstone, and the underlying sandstone or conglomerate of the Cardium Formation. The Wapiabi Formation was divided into four units, in ascending order: the Lower Concretionary Shale zone, Platy Shale zone, Upper Concretionary Shale zone, and the Transition zone (Webb and Hertlein, 1934).

Based on field work in southern Alberta conducted between 1954 and 1957, Stott (1963) further subdivided the Wapiabi Formation into seven members: Muskiki, Marshybank, Dowling, Thistle, Hanson, Chungo, and Nomad. The Muskiki and
Marshybank members were previously part of Webb and Hertlein’s Lower Concretionary Shale zone. Stott described the Muskiki Member as alternating beds of rubbly, bioturbated, and flaky shale, with localized siderite concretions and occasional sandstone beds. It is due to these alternating beds that the rocks of the Muskiki Member were previously described as ‘striped’ by Hake et al. (1942). The base of the Marshybank Member is placed where siltstone beds of the Muskiki Member grade upward into more massive siltstone and sandstone beds. The top of the Marshybank Member is typically concretionary, locally conglomeratic, and overlain by shales of the Dowling Member (Stott, 1963). Both the Muskiki and Marshybank members were found to thin eastward and southward in the Foothills, although Stott did not work on the Muskiki and Marshybank members south of Highwood River.

Using subsurface data in the Waterton gas field, south of Mill Creek, Herr (1967) divided the Wapiabi Formation into five informal units: lower shale unit, lower siltstone unit, middle concretionary shale and siltstone unit, upper sandstone, and upper shale unit. Herr noted that not all units were recognizable in the wells, but did note the presence of a pebble bed at the contact between the Cardium Formation and overlying Muskiki Member. This pebble bed was not recorded by Wall and Germundson (1963) on Mill Creek, ~10 km north of the Waterton gas field. More recently, in a subsurface study, the Verger Member of the Niobrara Formation was correlated to the Muskiki Member and the lower part of the Marshybank Member (Nielsen, 2003; Nielsen et al., 2003). The Medicine Hat Member was correlated to the upper part of the Marshybank and Dowling members in the Foothills. The top surface of the Verger Member was described as being covered by a lag of well-rounded pebbles with calcareous ooids, glauconite, fish and bone debris cemented by siderite. Nielsen (2003) documented a second pebble bed above the Verger Member-Medicine Hat boundary, and, together with the pebble bed at the top of the Verger Member, correlated them to pebble beds in outcrops in Kevin, Montana, initially described by Cobban et al. (1976). It remains unclear as to how the upper and lower bounding surfaces of the Verger Member correlate to the established stratigraphy in the Foothills because those studies lack sufficient and convincing subsurface to outcrop correlations (Nielsen 2003; Nielsen et al., 2003; 2008).
1.6.3 Allostratigraphy

To the north of the present study area, Stott (1967) correlated the Bad Heart ooidal sandstone exposed in the Peace River Plains to its apparently stratigraphic equivalent in the Foothills, which is underlain by the Muskiki Formation. Norris (1989), using an allostratigraphic approach, showed through subsurface correlation that the Bad Heart of the Plains was younger than the ‘Bad Heart’ of the Foothills, the type Bad Heart thinning and disappearing westward into the Foothills. The ‘Bad Heart' Formation in the Foothills was renamed the Marshybank Formation, the Marshybank terminology being extended from southern Alberta (Plint et al., 1990). The Bad Heart Formation was constrained to the northwestern Alberta Plains, and was separated from the older Muskiki and Marshybank formations in the Foothills by a regional disconformity, formally documented in Plint et al., (1990). The Bad Heart Formation was interpreted to have been deposited during a relative sea level lowstand in a shallow water setting over the crest of a rising forebulge that may have also been influenced by the Peace River Arch (Donaldson et al., 1998; 1999).

An integrated outcrop to subsurface study of the Muskiki and Marshybank formations between townships 44 and 76 in the northwest Alberta and northeast British Columbia foothills (Fig. 1.6a, Plint, 1991; Plint and Norris, 1991) suggested that the Muskiki Formation was deposited during relative sea level rise. The Formation consists mainly of mudstone to siltstone with interbedded thin sandstones. Two regional log markers representing flooding surfaces, M1 and M2, were followed in the subsurface and correlated to outcrop (Fig. 1.6b). The Marshybank Formation is made up of 12 allostratigraphic units, A-L, most of which represent shoreface sandstones. With the exception of unit G, which is non-marine, the Marshybank succession shows 12 regionally traceable allomembers (A-L) interpreted to represent cycles of eustatic sea level rise and fall. Unit A, capped by pebbles, is interpreted as a lowstand deposit, and also forms the downlap surface for overlying allomembers B-L (Fig. 1.6b). Overall, the Marshybank Formation was interpreted to have been deposited during relative sea level highstand (Plint, 1990, 1991).

Durbano (2009) and Plint (unpublished work), traced several subsurface markers within the Muskiki and Marshybank formations from the northern study area of Plint
Figure 1.6 A- Study area in which an allostratigraphic scheme for the Muskiki and Marshybank formations was first established, in northwestern Alberta; B- The Allostratigraphic framework established by Plint for the Muskiki and Marshybank formations. Note the interpreted systems tract for each Formation. From Plint, 1990.
(1990) to the northwest corner of the present study area, in order to determine whether allomembers in the north could be recognized in the south. Because outcrop sections in the southern Foothills could be correlated with considerable confidence to nearby well logs, it was logical to extend the Muskiki and Marshybank member terminology of Stott (1963) into the subsurface. This was considered preferable to adopting the rather ill-constrained Verger Member terminology of Nielsen et al. (2003).

Other work conducted in Alberta includes more broad sedimentological studies of the Wapiabi Formation in the northern and central Foothills (Ferguson, 1984) and in the southwestern Foothills (Rosenthal, 1984; Rosenthal and Walker, 1987). Medioli (1999) conducted a micropaleontological and sedimentological study of the Blackstone, Cardium, and Wapiabi formations at Mill Creek and proposed the outcrop as a reference section for the southern Foothills. Collom (2001) conducted a regional outcrop study of the Wapiabi Formation, spanning 900 km along the Alberta and B.C. Foothills, and incorporated micropaleontology, lithostratigraphic, and geochemical analysis into the study.

This thesis presents an opportunity to integrate biostratigraphy, sedimentology, and subsurface to outcrop stratigraphy into the regional geological framework at a level of detail that has not previously been attempted.

1.7 Study Area

The study area covers approximately 60,000 km$^2$ in southern Alberta and includes a small area in northwestern Montana (Fig. 1.7). The study area covers an area from ranges 9W5 to 8W4 in Township 30, to ranges 4W5 to 8W4 in Township 1. The area that extends into Montana covers ranges 7W to 3E from Township 37 to Township 35N, approximately 30 km inside of the Canada-US border. Outcrop sections in the deformed Rocky Mountain Foothills were correlated to the closest available well logs.

1.8 Database and Methods

In order to provide a detailed stratigraphic framework for the Upper Cretaceous Muskiki and Marshybank members, approximately 710 gamma ray and induction resistivity well logs were examined. From these, a network of 23 cross sections were
Figure 1.7 Thesis study area showing well, core, and outcrop database, as well as relevant geographic features.
constructed with the township (N-S) and range (E-W) lines spaced about 20 km and 40 km apart, respectively (Fig. 1.7). Sedimentological and petrographic analyses were based on 14 outcrop sections located in the Foothills disturbed belt and Montana, for a total measured thickness of 1180 m, and eight drill cores in the subsurface, for a total measured thickness of 173 m.

1.8.1 Correlation method

The stratigraphy of the Muskiki and Marshybank members was subdivided into smaller units by correlation of flooding surfaces, following an allostratigraphic approach. **Allostratigraphy** is a classification scheme that subdivides sedimentary successions on the basis of bounding surfaces, rather than lithology (NACSN, 2005). Laterally traceable bounding surfaces in this thesis are represented by marine flooding surfaces and erosional unconformities that were traced in the subsurface, and correlated to outcrops and drill cores where possible.

1.8.2 Palinspastic Restoration of Outcrop Sections

All of the outcrop sections within this study, with exception of those located at Deer Creek and Kevin, Montana, are located within the deformed Foothills belt. As a result, the outcrops have undergone varying amounts of displacement from their original Late Cretaceous positions. Based on available published data, outcrops were restored to their original, pre-deformation positions (Table 1.1). This was completed by adding the lengths of individual thrust slices that were located in front of individual outcrop sections, in order to obtain an approximate distance over which an outcrop had been displaced.

1.8.3 Scanning Electron Microscopy

Scanning Electron Microscopy (SEM) images of mudstone samples from core were collected using a Hitachi SU6600 field emission gun- scanning electron microscope (FEG-SEM) located in the Zircon and Accessory Phase Laboratory (ZAP-LAB) in the Department of Earth Sciences at the University of Western Ontario. Samples were carbon coated and placed in a vacuum chamber before images could be obtained. Identification
<table>
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<th>Outcrop</th>
<th>Map Sheet (Topography/Geology)</th>
<th>UTM</th>
<th>Displacement (km)</th>
<th>Displacement Direction (degrees)</th>
<th>Restoration Reference</th>
<th>Restored UTM Latitude</th>
<th>Restored UTM Longitude</th>
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<td>Rottenfusser et al., 1999</td>
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</table>

¹ Five individual sections measured - location denotes most westerly outcrop (stratigraphically lowest)
² Four individual sections measured - location denotes most westerly outcrop (stratigraphically lowest) - sections based on Cobban et al., 1976

Table 1.1 Summary of all the outcrop sections that were measured for this study. Table outlines their geographical location and the topographic map they can be found on. The references used to help with displacement are included, as well as restored UTM data, which was then converted to latitude and longitude format. Many references were found with the gracious assistance of G.S. Stockmal, GSC.
of grains of interest was obtained using X-ray energy dispersive spectrometry (EDS), with an Oxford Instruments X-Max 80 mm² drift detector (Moser et al., 2011).

1.8.4 Petrographic Sections

Petrographic analysis of thin sections was done using a standard optical petrographic microscope. A Nikon SMX1500 stereoscope microscope was also used to obtain images of thin sections and pebble samples. Images were obtained using a Nikon Digital Camera DXM1200 mounted on top of the microscope. Additionally, thin section scans were acquired using a standard flatbed scanner set in ‘back lit’ mode, at 1200 dpi (dots per inch) quality.
CHAPTER 2

FORELAND BASIN EVOLUTION AND SEQUENCE STRATIGRAPHY

2.1 Introduction

The following chapter is divided into two main topics of study. The first section introduces the reader to the regional tectonic history and setting in which the Muskiki and Marshybank members were deposited. The second portion of the chapter is a review of sequence stratigraphy, sea level change, and allostratigraphy. These concepts will be used when reconstructing the depositional history of the Muskiki and Marshybank members in chapters three to five.

2.2 The Western Canada Sedimentary Basin

The Western Canada Sedimentary Basin (WCSB) encompasses a wedge of sedimentary rocks of Proterozoic to Oligocene age, that thickens westward from a zero edge on the Canadian Shield to more than 18 kilometres (km) in the northeastern margin of the Cordilleran fold and thrust belt (Fig. 2.1; Ricketts, 1989; Price, 1994). Within the deformed belt, the sediments that comprise the WCSB can be up to 9 km thick (Wright et al., 1994). The WCSB is underlain by Precambrian crystalline basement rocks.

The evolution of the WCSB can be divided into two main stages: 1) a late Proterozoic to Middle to Late Jurassic passive margin stage, in which the main source of sediment was the north American craton, and 2) a Late Jurassic to Eocene foreland basin stage, in which the main source of sediment was to the southwest, in the emerging Cordilleran mountain belt (Kauffman and Caldwell, 1993; Price, 1994).

During the passive margin stage, repeated marine transgressions resulted in deposition of siliciclastic and carbonate rocks that prograded westward to form a wedge along the eastern rifted margin of the North American craton (Kauffman and Caldwell, 1993). Sediments of Cambrian age indicate that syndepositional block faulting occurred (Kent, 1994). From the Cambrian to the Middle Jurassic, for a period of about 350 million years (my), protracted thermal subsidence developed on margins of the craton,
Figure 2.1 Outline of the Western Canada Sedimentary Basin between the Canada – U.S. border and 60° North. West of the fold and thrust belt are the four accreted terranes that form the Mesozoic western portion of the craton (see Section 2.2 for discussion). From Ricketts, 1989.
and depositional environments included hypersaline basins, reefs, and carbonate ramps (Kent, 1994).

As early as Late Triassic, several terranes accreted against the western margin of North America, collectively forming the Intermontane super terrane (Ricketts, 2008). Westward movement of the North American plate, in response to opening of the Atlantic Ocean, caused subduction of the Farallon plate. The ensuing collision of the Intermontane super terrane with the North American craton was responsible for the Columbian Orogeny, which is represented by the high grade metamorphic rocks of the Omineca Belt (Fig. 2.1; Leckie and Smith, 1992). The accreted Intermontane terrane was thickened by folding and imbrication as it overrode the craton, and was in places backfolded and thrust westward (Kauffman and Caldwell, 1993). The resulting shortening and vertical stacking of the crust emplaced a substantial supracrustal load on the craton margin. Crustal thickening was consequently balanced by isostatic subsidence, which formed a broad flexural foreland basin to the east; this represents the foreland basin stage of the WCSB (Kauffman and Caldwell, 1993).

As a result of subsidence within the foreland basin, the craton was flooded from both the north and south. Arctic waters eventually mixed with warmer waters from the Gulf of Mexico, forming the Western Interior Seaway. Sedimentation during the earliest stage of the foreland basin is represented by the Late Jurassic to Early Cretaceous siliciclastic successions of the Fernie and Kootenai formations (Price, 1994; Miall et al., 2008). By the end of the Early Cretaceous, continued accretion saw collision of another major terrane, the Insular belt, with the North American craton. The suture zone is represented by the Coast Plutonic Belt (Fig. 2.1; Leckie and Smith, 1992).

From the Late Cretaceous to Earliest Eocene, renewed thrusting and horizontal shortening in the orogen resulted in sediment being shed to the east into the subsiding foreland basin (Hayes et al., 1994; Evenchick et al., 2007). Cycles of marine transgression and regression between the Albian and Santonian resulted in widespread shale deposition, punctuated by regressive episodes of shallow marine sandstone and conglomerate deposition (Leckie et al., 1994). The rocks that form the subject of this study, the lower Wapiabi Formation, were deposited during the Late Cretaceous.
Major hydrocarbon source and reservoir rocks were also deposited at this time, including the Late Cretaceous Cardium (reservoir) and Niobrara (mostly source rock) formations (e.g. Hankel et al., 1989; Krause et al., 1994). A schematic representation of the history of the foreland basin from the Jurassic to the Paleocene is shown in Figure 2.2.

Withdrawal of the Western Interior Seaway began during the late Campanian/Maastrichtian as a result of global sea level fall (Miall et al., 2008). Dark brown silty shales of the Bearpaw Formation represent the last marine strata to be deposited in Montana and the Canadian portion of the basin (Dawson et al., 1994; Miall et al., 2008). During the Eocene, the basin underwent post orogenic uplift. This occurred due to plate movement becoming transtensional, which resulted in reduced activity in the fold and thrust belt (McCausland et al., 2006). Consequently, unroofing and erosion of the foreland basin began (McCausland et al., 2006, Plint et al., 2012). It has been suggested that two to three kilometres of Mesozoic and Cenozoic strata have since been removed by post Miocene erosion (Beaumont, 1981; Kalkreuth and McMechan, 1988).

2.3 Foreland Basins

The term foreland basin was first introduced by Dickinson (1974) to describe basins located adjacent to arc terrains. More specifically, a foreland basin is an elongate trough that forms between a linear orogenic belt and a stable craton, whereby its formation is driven mainly by flexural subsidence due to thrust sheet loading in the orogen (Fig. 2.3; Beaumont, 1981; DeCelles and Giles, 1996).

Foreland basins are divided into two main types, based on their tectonic settings: retroarc and peripheral basins (Dickinson, 1974). Peripheral foreland basins develop on the subducting plate adjacent to suture zones in front of thrust belts, and form as a result of continent-continent collision (Beaumont, 1981; Jordan, 1981). Examples include the Indo-Himalaya basin, and the Alpine foreland basin of Central Europe.

A retroarc foreland basin develops on the cratonic side of a foreland thrust belt adjacent to a magmatic arc above a subduction zone (Jordan, 1981; Leckie and Smith, 1992). They are usually the result of oceanic-continental plate collision, and examples include the Subandean foreland basin in South America, and the Western Canada foreland basin.
Figure 2.2 A simplified illustration to explain the evolution of the Western Canada Sedimentary Basin, from the early Jurassic through to the Paleocene. 

a) Early Jurassic- extensional tectonics developed on the North American craton. 
b) Middle Jurassic- accretion of several terranes against the Western margin of North America and subduction of the Farallon plate began. 
c) Early Cretaceous- shortening and vertical stacking of the crust created a supracrustal load on the craton, forming a foreland basin. 
d) Paleocene- continued activity in the fold and thrust belt contributed sediment to the foreland basin. 

2.3.1 Mechanisms of Subsidence

Static loading of an orogen by a thrust belt defines the features of a retroarc foreland basin into a distinctive foredeep, forebulge and backbulge geometry (Fig. 2.3; Catuneanu, 2004b). The accretion of terranes onto the North American margin generated a static supra-lithospheric load on the lithosphere, causing it to flex down and subside. Subduction of the Farallon plate beneath the North American plate also contributed to static loading of the lithosphere (Catuneanu, 2004b). Downwarping of the lithosphere causes an arcuate deflection over several hundred kilometres that is manifested as a foreland basin (Fig. 2.4; Flemings and Jordan, 1989; Decelles and Giles, 1996). The amount of deflection depends on the amount and distribution of loads, as well as the flexural rigidity of the underlying lithosphere (McMechan and Thompson, 1993; Catuneanu, 2004b). Loading onto a thick, cool and rigid lithosphere produces a wider and shallower basin, whereas loading onto a thin, hot and weaker lithosphere generates a narrower and deeper basin (Fig. 2.4; McMechan and Thompson, 1993).

The thickness and width of a foreland basin is too great to be explained by static loading, which produces a foreland basin only several hundred km wide (Plint et al., 2012). Viscous coupling between the subducting plate and the lithosphere generates corner flow above the subducting plate, resulting in dynamic loading (Fig. 2.5; Catuneanu, 2004b). Dynamic (viscous) stresses within the asthenosphere result in regional subsidence that extends beyond the limits of those produced by static loading, over a distance greater than1000 km (Mitrovica et al., 1989; Royden, 1993; Burgess and Moressi, 1999). Although generated independently of each other, static and dynamic loading are the principal mechanisms that generate subsidence within a foreland basin (Fig. 2.5; Catuneanu, 2004b).

Flexure of the lithosphere is predicted by models that assume an elastic behaviour for the plate, which imply that, once the load is removed, the lithosphere should return to its original state (Beaumont, 1981; Jordan, 1981; Flemings and Jordan, 1989). Viscoelastic models on the other hand, predict that over time, the flexural rigidity of the lithosphere diminishes due to the persistent supracrustal load, resulting in narrowing and deepening of the basin (Fig. 2.4; Beaumont, 1981; Beaumont et al., 1993). Although
Figure 2.3 A generalized model for a foreland basin, with four distinct depozones: Wedge-top, foredeep, forebulge, and back-bulge areas. Modified from Decelles and Giles (1996).
Figure 2.4 Schematic illustration of lithosphere flexure during periods of static loading and unloading in the fold and thrust belt. Loads 1, 2, and 3 progressively narrow and deepen the basin, and move the forebulge closer to the fold and thrust belt. Unloading results in a reverse relationship. Forebulge uplift exaggerated for clarity. From Beaumont et al., 1993.
Figure 2.5 Overview of the dynamic loading mechanism: as the subducting slab is pulled down, a drag flow creates corner flow within the asthenosphere, resulting in regional subsidence. Static and dynamic loading work in combination to create subsidence within a foreland basin. From Catuneanu, 2004.
there is continuing debate on which model best predicts foreland basin development, studies from the Western Canada Sedimentary Basin are better explained in terms of an elastic plate model (e.g. Flemings and Jordan, 1990; Sinclair et al., 1991; Plint et al., 2012).

2.3.2 Tectonics, Subsidence, and Sedimentation

Numerical models of foreland basin formation and evolution assume that subsidence involves a homogenous lithosphere (e.g. Beaumont, 1981; Jordan 1981). Realistically, the underlying structure of the basement is quite heterogeneous because it is composed of various terranes that accreted over time. Boundaries between accreted terranes may behave as weak zones, resulting in preferential flexure over those areas. Differential subsidence and uplift of basement blocks that have different flexural rigidity may generate smaller basins and arches within a foreland basin (Catuneanu, 2004a). Such variable activity, together with water and sediment contributing to the basin load, form secondary controls on accommodation (Catuneanu, 2004a). Examples of basement control on sedimentation have been documented for the Cretaceous in Western Canada (eg. Donaldson et al., 1998; Kreitner and Plint, 2006). Variations in the strength of the lithosphere can cause flexure to be concentrated in a forebulge located within a zone of weakness. Advance of the deformation front may cause the forebulge to shift to the next weak zone (Waschbusch and Royden 1992). Evidence of forebulge migration can be seen in upper Cretaceous strata in Western Canada (Plint et al., 2012).

Thrusting within an orogen leads to rapid isostatic subsidence and sediment accumulation so that sediment derived from the uplifted thrust sheet tends to follow the area of subsidence (Heller et al., 1988). The resulting sediment package is wedge-shaped, with the thickest accumulation adjacent to the tectonic front (Fig. 2.6a). Wedge-shaped rock packages typically develop when rates of subsidence are high (Plint et al., 2012). As tectonic activity diminishes, the orogen is eroded and subsequently uplifted by isostatic rebound. This results in a decreasing rate of subsidence in the deepest parts of the foredeep, and newly shed sediment begins to prograde further across the basin. Thus, at times of tectonic quiescence, coarser material is able to prograde further into the basin, a process possibly augmented by sea level fall (Heller et al., 1988; Plint et al., 2012).
Figure 2.6 A- Subsidence during thrust loading tends to lead to a wedge geometry in which coarse material is more commonly found adjacent to the thrust belt. B- Rebound post orogenic thrusting allows coarser sediment to prograde further into the basin. Sediments previously deposited may also get eroded and reworked. From Heller et al., 1988.
Additionally, sediment previously deposited in the most proximal foredeep may also be eroded and reworked across the basin (Fig. 2.6b; Heller et al., 1988). Tectonic unloading generally leads to deposition of sediment in a more sheet-like geometry (Plint et al., 2012). Examples of contrasting wedge and tabular geometries of sedimentary packages are observed in the Upper Cretaceous Kaskapau and Cardium formations of western Canada (Varban and Plint, 2008b).

2.3.3 Foreland Basin Systems

Early foreland basin models focused on sedimentation within the foredeep, and little attention was paid to the forebulge zone and beyond. The concept of a foreland basin system was therefore introduced to explain sedimentation beyond the main area of subsidence (DeCelles and Giles, 1996). The model was expanded to include sediments derived from the orogen that accumulate on the frontal edge of the thrust belt. Consequently, a foreland basin system is made up of four distinct areas: wedge-top, foredeep, forebulge, and back-bulge depozones (Fig. 2.7).

The wedge-top depozone thins onto the orogenic wedge, and may be tens of kilometres in width. The proximity of the wedge-top to the fold and thrust belt leaves it very susceptible to deformation and erosion. The resulting stratigraphy is therefore exemplified by numerous unconformities and deformation features such as folds, faults and progressively rotated cleavages. The sediment deposited therein demonstrates extreme textural and compositional immaturity (DeCelles and Giles, 1996).

The foredeep depozone is the site of the greatest sediment accumulation within a foreland basin, and as such has been the focus of many stratigraphic studies. Foredeep sediment is derived predominantly from the fold and thrust belt, with minor contributions from the forebulge and craton. The fill of the foredeep is typically 200 – 300 km wide and two to eight km thick. The proximal foredeep merges with the wedge-top depozone, where sediment bypass may occur, and numerous unconformities are common (DeCelles and Giles, 1996).

The forebulge depozone is difficult to recognize in the rock record because it is essentially a very small positive feature that can migrate over time. Erosion or nondeposition may occur over the forebulge if it is located in a sufficiently shallow or
Figure 2.7 A summary of the foreland basin system, made up of the wedge-top, foredeep, forebulge, and back-bulge depozones. See section 2.3.2 for details. From Decelles and Giles, 1996.
subaerial environment. In a subaqueous environment, the relief produced by the forebulge may be negligible, and no facies shift may be observed over the feature. Sedimentation over the forebulge may be condensed and chemical and carbonate deposits may form (DeCelles and Giles, 1996). The amount of uplift of the forebulge, which may be 10s of metres, depends on the flexural rigidity of the underlying plate. A narrower, higher forebulge will form over a hotter lithosphere, whereas wider, and more subdued uplift may occur on a cooler plate (Fig. 2.4; Beaumont, 1981).

Within the back-bulge depozone, sediment is still predominantly derived from the orogenic belt, although the cratonic side of the basin plays a greater role in supplying sediment. Carbonate platforms may also form in the back-bulge depozone, generating additional sediment (Decelles and Giles, 1996). Stratigraphic units are much thinner than those within the foredeep, and have a more sheet-like geometry because of limited subsidence (Flemings and Jordan, 1989). The back-bulge depozone remains the most poorly understood part of the basin system.

2.4 Tectonic Elements of the Study Area

As was briefly discussed in the previous section, heterogeneous regions within the Precambrian basement cause zones of weakness in the crust that may be susceptible to later flexure. Various basement blocks may be activated at different times due to different stresses (Catuneanu, 2004a). In southern Alberta, various structural elements have been shown to have influenced the geology of the area. These structural elements range from highly localized features to more regionally extensive ones. This section introduces these tectonic elements and their potential effects on sedimentation.

2.4.1 Basement Structures

The crystalline basement underlying the WCSB has been divided into 23 distinct domains (in Alberta) based on geophysical and geochronological data (Fig. 2.8; Ross et al., 1991; Villeneuve et al., 1993; Pilkington et al., 2000). Dates were obtained using U-Pb geochronology of zircon and monazite crystals from basement drill core samples, which were available from hydrocarbon exploration projects (Ross et al., 1991). Potential field signatures such as magnetic and gravity data were used to distinguish the different domains. Domains were recognized on an aeromagnetic anomaly map based on their
Figure 2.8 Precambrian basement domains of Alberta, as interpreted based on geophysical and geochronological available data. Thesis study area is underlain by parts of the Medicine Hat (Hb), Vulcan (Hv), and Lovernia domains (Hl). The Matzhiwin domain (Hz) forms part of the Lovernia basement structure. From Pilkington et al., 2000.
intensity and textural characteristics (Fig. 2.9). Rocks containing ferromagnetic minerals tend to give highly positive anomalies (shaded red; Fig. 2.9). Most positive curvilinear patterns on the aeromagnetic anomaly map represent magmatic belts (Ross et al., 1994). Regionally extensive negative anomalies (shaded blue; Fig. 2.9) are usually harder to interpret, because even magnetite-bearing rocks can produce a negative signal. However, processes such as regional metamorphism, which oxidizes magnetite to hematite, can produce aeromagnetic lows (Robinson et al., 1985).

2.4.2 Medicine Hat Block

The Medicine Hat Block (MHB) domain forms most of the crystalline basement that underlies the thesis study area (Fig. 2.9). The MHB is of Archean age, 2.6 to 3.3 billion years old, and has a complex tectonic history (Lemieux, 1999; Lemieux et al., 2000). The boundaries of the MHB are delineated as follows: to the west, the MHB is buried beneath overthrust Mesozoic and Cenozoic strata of the Cordillera; to the east, it is bounded by the early Proterozoic Trans-Hudson Orogen; to the north, it is bounded by the early Proterozoic Vulcan structure; and to the south, by the Great Falls tectonic zone (Lemieux et al., 2000).

Within the MHB, prominent linear, north and northwest-trending anomalies exist (Fig. 2.10). These anomalies are characterized by long wavelength magnetic signatures, which are usually attributed to sources deep within the crust and mantle. The crust within the MHB, on the basis of both magnetic and seismic data, is interpreted to have been deformed under a compressional regime. This is based on three main lines of evidence: 1) presence of prominent crustal reflectors, 2) truncation of reflectors along dipping surfaces, and 3) antiform structures developed along the dipping reflectors. During crustal shortening, antiforms may have formed on the hanging wall of thrust sheets (Lemieux et al., 2000). These characteristics form the primary evidence for the presence of faults within the basement (Ross and Eaton, 1999).

Seismic reflection profiles along lines 29, 30, and 31 of the SALT Lithoprobe project have allowed the division of the MHB into five distinct zones (1-5), which show strong correspondence to aeromagnetic data (Fig. 2.10). Within zone three, which shows listric geometry in the lower crust, are subtle southwest dipping reflectors that align with
Figure 2.9 Aeromagnetic map for Alberta, measured in nanotesla units. Thesis study area is outlined in red (Fig. 1.2). MB- Medicine Hat Block, VL- Vulcan Low, LB- Loverna Block. Long dashed line represents limit of deformation front; short dashed line shows Phanerozoic edge. From Pilkington et al., 2000.
Figure 2.10 A block diagram showing the interpreted five zones of the Medicine Hat Block and their relationship to aeromagnetic data. The MHB is located in the basement underneath the southern half of the thesis study area. Crustal surfaces CS1 and CS2 correspond to north trending anomalies, and crustal surfaces CS3 and CS4 correspond to northwest trending anomalies. No VE- No vertical exaggeration. From Lemieux et al., 2000.
north and northwest-trending fabrics on the aeromagnetic map. The crust within the MHB is interpreted to have undergone two main stages of crustal deformation. The first stage involved continental collision of two Archean domains, the MHB and the Loverna Block (LB), along a crustal-scale ramp in the late Archean (Fig. 2.11a and b). As a result, portions of the Loverna block were incorporated into the MHB crust. The second stage involved early Proterozoic crustal shortening that was associated with the formation of the Vulcan structure. Juxtaposition of the MHB crust over delaminated Loverna block crust may have occurred along the south dipping subduction (zone five), resulting in the seismic reflectors observed today (Fig. 2.11c and d; Lemieux et al., 2000).

The complex structure of the MHB leaves zones of weakness within the basement that could become reactivated, with deformation propagating upward into overlying sedimentary strata. Recurrent movement of basement fault blocks may occur due to changes in the local stress field, and less frequently, the regional stress field (Shurr and Rice, 1986). A fault block may be elevated at one time and depressed at another, depending on the change of both magnitude and orientation of the stress field (Anna, 1986). Shurr and Rice (1986) attribute patterns of sedimentation within the Cretaceous Niobrara, Telegraph Creek, and Eagle Sandstone formations in central Montana to lineament-bounded fault blocks that were active beneath the sedimentary basin. Landsat images provide further evidence, independent of stratigraphic data, that basement lineaments are expressed in the surficial topography (Shurr and Rice, 1986). Figure 2.12 is an illustration of how that may be expressed. Ugalde et al. (2008) incorporated aeromagnetic with topographic data, to show that many geographical features in southeastern Alberta, such as streams and rivers, demonstrate a strong correspondence between aeromagnetic and gravity anomalies in the basement.

2.4.3 Vulcan Structure

The Vulcan Structure, is a prominent east-trending aeromagnetic anomaly that is 40-70 km wide and extends for over 350 km across southern Alberta (Fig. 2.13; Eaton et al., 1999). The Vulcan Structure is made up of the negative aeromagnetic Vulcan Low, and the positive Matzhiwin High (Hope and Eaton, 2002). Together, the two structures
Figure 2.11 Stages of crustal deformation interpreted for the Medicine Hat Block in Southern Alberta. Panels A and B represent the first stage of crustal deformation, in which continental collision of the MHB and the Lovernia block occurred along a crustal scale ramp. Panels C and D represent the second stage of crustal deformation, in which crustal shortening occurs in association with the Vulcan structure’s formation. The resulting morphology is made up of seismic reflectors observed today. From Lemieux et al., 2000.
Figure 2.12 A simple model to show propagation of fault bounded basement blocks upward through the stratigraphy. On the earth's surface, faults may be represented as linear features. From Shurr and Rice, 1986.
Figure 2.13 Aeromagnetic map showing portions of the Medicine Hat Block (MHB) and Vulcan Low (VL). Solid line delineates Vulcan Low to incorporate positive features assigned to the Matzhiwin High by Villeneuve et al., 1993, which is shown by dashed line. Thin line in the northeast portion of the map delineates granitic rock (not discussed). LB- Loverna Block. From Eaton et al., 1999.
form a gravity low (Hope and Eaton, 2002). The Vulcan Structure separates the Medicine Hat Block (MHB) from the Lovernia Block (LB), is interpreted to represent a continental collision, and is probably the suture zone between the Wyoming Province to the south, and the Hearne Province to the north (Eaton et al., 1999; 2000; Hope and Eaton, 2002). The formation of the Vulcan Structure was responsible for the second stage of crustal deformation within the MHB, discussed in 2.4.3, and figures 2.10 and 2.11c. Previous interpretations of the Vulcan Structure include a failed rift (Kanasewich et al., 1969), the principal suture zone between the Wyoming and Hearne provinces (Hoffman, 1988), and an internal component of the Hearne Province (Hoffman, 1990).

Subtle yet persistent sedimentation trends that coincide with the Vulcan Structure have been documented in the Lower Cretaceous Basal Quartz Member of the Lower Mannville Formation in southern Alberta (Zaitlin et al., 2002). Depositional trends show that the Vulcan Low represents a paleodrainage divide between sedimentary packages. Narrow, shallow tributary systems are observed in units south of the Vulcan Low, in contrast with deeper ones to the north (Zaitlin et al., 2002). In a high resolution stratigraphic study of the Cardium Formation in southern Alberta, upper sand packages were shown to thicken and thin over the Vulcan structure, suggesting episodic reactivation of the suture zone (Shank and Plint, 2011). The overlying Muskiki and Marshybank members may also have been affected by the Vulcan structure.

2.4.4 Sweetgrass Arch

The Sweetgrass arch in southern Alberta and northwestern Montana encompasses a large antiformal structure that is primarily located in northwestern Montana, but extends into southern Alberta. In Alberta, the Sweetgrass Arch extends beneath the Plains and meets the south-plunging North Battleford Arch that extends from the Canadian Shield (Kent and Christopher, 1994). The Sweetgrass Arch is made up of three main components: the northwest-plunging South arch, the northwest-striking Kevin-Sunburst Dome, and the northeast-plunging Bow Island Arch; the latter separates the Alberta foreland basin from the intracratonic Williston basin (Fig. 2.14; Podruski, 1988). Numerous tight, northwest-trending and north-plunging faulted folds further complicate
Figure 2.14 Elements of the Sweetgrass Arch in Southern Alberta and Northwestern Montana: Kevin-Sunburst Dome, South arch, and the Bow Island Arch in Alberta. Outline of study area is also shown. Contours are in feet above sea level, from base of the Colorado Shale. Modified from Grifi et al., in revision.
the structure (Tovell, 1958). Initially thought to have originated together, structural contours and mapping have shown that the Kevin Sunburst Dome is distinct from the Bow Island Arch (Tovell, 1958). The South Arch is separated from the Kevin-Sunburst Dome by the Marias River Saddle (or Pendroy Fault Zone).

Herbaly (1974) provided detailed examples in which evidence for uplift and erosion over the Sweetgrass arch was found to occur sporadically throughout the Paleozoic. More examples that showed evidence of uplift were also summarized by Lorenz (1982). Evidence of active uplift of the Sweetgrass Arch during the Cretaceous has been observed in the upper Aptian- lower Cenomanian Bootlegger Member in Montana (Arnott et al., 1995). In southern Alberta, evidence for uplift and subsidence of the Bow Island Arch is documented in the Santonian Medicine Hat sandstones (Hankel et al., 1989; Nielsen et al., 2003), the Santonian Sweetgrass Member (Schroder-Adams et al., 1997; 1998), and the Santonian-Campanian Milk River Formation (Payenberg, 2002). Ammonoid faunas are also shown to differ between the western and eastern flanks of the arch in the Late Campanian Bearpaw Formation, suggesting presence of a topographic high at the time (Tsujita and Westermann, 1998). No evidence for active uplift of the Sweetgrass Arch has been shown for Coniacian strata, due to the lack of detailed studies of the interval.

Tovell (1958) suggested that movement of the Sweetgrass arch was caused by compressional forces associated with mountain building processes. Lorenz (1982) argued that horizontal compression forces were insufficient to initiate flexure and uplift of the arch, but could have been a factor once the Sweetgrass arch had been elevated due to an underlying brittle-elastic-ductile lithosphere. The weakened crust would then have been more susceptible to flexure, despite changing thrust loads (Lorenz, 1982). The episodic uplift of the arch has also been suggested to be attributed to movement of individual blocks within the basement (Payenberg, 2002), although direct evidence is still lacking.

2.4.5 Extensional Faulting in southern Alberta

In the Southern Foothills of Alberta, the triangle zone, a thin-skinned tectonic wedge, usually marks the eastern limit of Cordilleran deformation within the basin (e.g. Wright et al., 1994; Stockmal et al., 2001). However, deformation has been observed as
far as 65 km east of the triangle zone, for example in the Pincher Creek area (Hiebert and Spratt, 1996). Lemieux (1999) used seismic reflection data from the Southern Alberta Lithospheric Transect (SALT) project to show that extensional faulting occurred as far as 140 km east of the triangle zone.

Using three seismic profiles that run from the Pincher Creek area to south of Medicine Hat, Lemieux (1999) identified five northwest-trending extensional faults that propagate from the near basement surface upward into Devonian and Upper Cretaceous strata. Some areas show folding of upper Cretaceous strata over these faults, which are consistent with extensional forced folds. Extensional forced folding of sediment may occur over buried normal faults, which laboratory experiments have shown to occur if the sedimentary cover overlies more brittle basement (Withjack et al., 1990). The mechanism of extensional fault formation was interpreted to be part of the flexural subsidence that was generated by thrust loading along the continental margin of North America during the Late Cretaceous (Lemieux, 1999).

The extensional faulting and subsequent deformation of Upper Cretaceous strata observed in seismic profiles was not detected specifically for rocks of Coniacian age. However, the high resolution correlation of surfaces within Coniacian rocks has the potential to identify areas of abrupt thickness change, which may suggest that syn-depositional faulting has occurred.

2.5 Sea Level Change and Cyclicity

*Eustasy*, or global sea level, is the motion of the sea surface relative to a fixed datum, such as the center of the earth (Posamentier et al., 1988). Evidence for sea level change within a sedimentary basin does not necessarily indicate eustatic change, so the term ‘relative sea level’ is used instead. *Relative sea level* refers to movement of the sea surface in relation to a local datum, such as a buried stratigraphic surface, or the sea floor (Fig. 2.15; Posamentier et al., 1988). The relative change of sea level within a sedimentary basin depends on the creation or removal of accommodation within the basin, which is in turn controlled by tectonic and eustatic mechanisms (Galloway, 1989; Plint et al., 1992).
Figure 2.15 Relationship between eustatic sea level and relative sea level. From Allen and Allen, 2005.
Sedimentary processes tend to be cyclical in nature, with depositional patterns recurring on many temporal and spatial scales (Helland-Hansen and Gjelberg, 1994). Some of these depositional patterns are observed to occur on a global scale, and suggest eustasy as the main driving mechanism (e.g. Haq et al., 1988). Eustatic change is primarily caused by changes in the volume of water in ocean basins (Miller et al., 2005). Factors controlling water volume variations are attributed to geological processes which occur on many time scales (Fig. 2.16).

Cycles of sea level change were originally categorized in terms of their magnitude. Vail et al (1977), on the basis of seismic data, initially defined three orders of eustatic cycles, with fourth and fifth order cycles later recognized from well logs, cores and outcrops (e.g. Mitchum and Van Wagoner, 1990). More recently however, it has been shown that these ‘orders’ are often arbitrary and highly variable. Instead, sedimentary processes are now described in terms of their episodicity (Miall, 2010). This episodicity is summarized below:

1) Sedimentary cycles that occur on time scales of 400-500 million years are referred to as global supercontinent cycles. A supercontinent cycle is generated as a result of the assembly of continents and their subsequent rifting and dispersal. The assembly of a supercontinent inhibits radiogenic heat loss from the core and mantle, creating a thermal blanket underneath the supercontinent and causing epeirogenic uplift over time. This may result in relative sea level fall on continental margins. There is still however, uncertainty in the long term mechanism behind the cycle (Miall, 2010).

2) Sedimentary cycles that span 10-100 million years are caused by oceanic spreading centers and plate movement (Plint et al., 1992; Miall, 2010). The changing volume of oceanic ridges is affected by the change in spreading rates that are controlled by mantle processes. Plate kinematics cause subtle yet noticeable movement in basement structures due to thermal changes in the crust and mantle, sediment loading, and/or crustal thickening and thinning (Miall, 2010).

3) Sedimentary cycles that operate on time scales of 0.1 to 10 million years are controlled by regional to local basement movement. The movements are
Figure 2.16 Timing and amplitude relationships for different geological mechanisms and their effect on eustatic change. SF- Sea Floor; Cont- Continental. From Miller et al., 2005.
controlled by regional plate kinematics, including changes in the intraplate stress regime. Such stress changes could lead to movement of crustal structures, such as the Vulcan Low and Medicine Hat Block described in section 2.4 (Miall, 2010).

4) Eustatic cycles that operate on timescales of 0.01 to 0.5 million years are recorded by sedimentary successions deposited in both nearshore and pelagic environments, and suggest several generating mechanisms. Many of these cycles are climatic in origin and are caused by changes in the amount of solar radiation and its global distribution across Earth. These cycles occur regularly, are known as Milankovitch cycles, and are affected by variations in three main components of Earth’s orbital rotation (Fig. 2.17):

i) Eccentricity: the shape of Earth’s orbit around the sun, which operates on time scales of 400,000 years (yrs) and 100,000 yrs.

ii) Obliquity: changes in the tilt of the earth, which operates on a time scale of 41,000 yrs.

iii) Precession: changes in the ‘wobble’ of Earth’s orbit, which operates on a time scale of 21,000 yrs.

Orbitally-driven climate change on Milankovitch time scales drives the growth and decay of continental ice sheets, which is the principal mechanism responsible for high-frequency eustatic cycles (Plint et al., 1992; Miall, 2010). It is important to note however, that nonglacial sedimentary cyclicity is also possible. As an example, on an alternating time scale of 10,000 years, the interaction between the eccentricity and precession components of the Milankovitch band cause contrasting weather on Earth’s hemispheres (Miall, 2010). One hemisphere experiences short, hot summers and long, cold winters, while the other undergoes a period of long, cool summers, and short, mild winters. Such changes have a strong influence on oceanic circulation, wind patterns, and surface air-water temperature distributions (Miall, 2010). This leaves potential for effects on sedimentation (e.g. climate change) that are unrelated to eustatic change.

High frequency sequences have been recognized in the Cretaceous (e.g. Fischer, 1986; Plint, 1991; Sageman et al., 1997; Plint and Kreitner, 2007), but direct evidence for
Figure 2.17 The three components of earth’s orbital cycle that cause Milankovitch cycles: eccentricity, obliquity, and precession. See section 2.5 for details From Miall, 2010.
their origin due to glacioeustacy is still a matter of debate. Nevertheless, recognition of high frequency sedimentary sequences occurring over long distances and across different tectonic and climatic regimes strongly suggests a Milankovitch control (Miall et al., 2008; Varban and Plint, 2008b).

2.6 Stratigraphic Terminology

Several types of stratigraphic terminology can be used to subdivide the rock record. Examples include biostratigraphy, subdivision of rock units based on fossil content; and lithostratigraphy, subdivision of rock units based on lithic characteristics and stratigraphic position (NACSN, 2005). This thesis applies an allostratigraphic approach to subdivision of the Muskiki and Marshybank members, where rock units are subdivided based on their bounding surfaces, rather than lithology (NACSN, 2005). The bounding surfaces can include subaerial unconformities, marine flooding surfaces, and maximum flooding surfaces.

An unconformity is a surface that separates younger from older strata, with a hiatus in deposition. Posamentier et al. (1988) used unconformities to define depositional sequences, the boundaries of which embodied evidence of subaerial exposure. However, it was later recognized that evidence for subaerial exposure may not always be preserved, and so the evidence may be inferred (Posamentier in Walker, 1992). A conformity is a surface that separates younger from older strata, in which although no evidence of erosion or significant hiatus is indicated (Van Wagoner et al., 1988). A conformity includes surfaces onto which deposition is slow, which translates to very thin deposits over long periods of geological time.

Early studies that investigated the relationship between sedimentation, sea level, and unconformities led to the idea of craton-wide depositional sequences, bounded by major unconformities (e.g. Sloss et al., 1949; Wheeler, 1964). The assumption then was that eustasy was the main driving force behind sequence formation in the stratigraphic record (Vail et al., 1977). The incorporation of well and outcrop data into seismic stratigraphy led to the concept of sequence stratigraphy (Posamentier and Vail, 1988; Van Wagoner et al., 1988). Posamentier et al. (1988) combined tectonics and eustasy as
the main driving forces behind changes in relative sea level, which are still widely accepted today.

2.6.1 Sequence Stratigraphy

Sequence stratigraphy is defined as the “study of rock relationships within a chronostratigraphic framework of repetitive, genetically related strata bounded by surfaces of erosion or nondeposition, or their correlative conformities” (Van Wagoner et al., 1988). The principal unit of sequence stratigraphy is the sequence, a sedimentary package that is bounded by unconformities and their correlative conformities (Van Wagoner et al., 1988). Within a sequence, parasequences usually develop, and are made up of relatively conformable successions of genetically related beds bounded by marine flooding surfaces and their correlative conformities (Van Wagoner, 1985). A marine flooding surface separates younger from older strata, and shows evidence of an increase in water depth (Van Wagoner, 1985). Minor erosion may be indicated at this surface however, no basinward shift in facies, or rock type, is observed (Van Wagoner et al., 1988).

Sequences can be divided into systems tracts, which are coeval depositional systems defined by stratal geometry at bounding surfaces, position within the sequence, and internal parasequence stacking patterns (Posamentier et al., 1988). Systems tracts reflect the interaction of accommodation and sediment supply for each part of the relative sea level curve. Posamentier and Vail (1988) recognized three systems tracts: highstand, lowstand, and transgressive. A fourth systems tract, forced regression, was first introduced by Plint (1991), and later supported by Hunt and Tucker (1992), and is used to account for siliciclastic and carbonate sediments deposited between the onset of relative sea level fall and lowstand. This is more commonly known as the falling stage systems tract, and the recognition of four systems tracts is now widely accepted (Fig. 2.18; Plint and Nummedal, 2000; Catuneanu, 2006).

2.6.2 Highstand Systems Tract

The highstand system tract (HST) begins when sediment supply exceeds accommodation rate, and the shoreline begins to prograde into the basin (Fig. 2.18a). Sediments deposited during the HST will downlap onto the maximum flooding surface
Figure 2.18 The four systems tracts defined on the basis of their stratal geometries and positions on the relative sea level curve. l-FR: late Forced Regression, e-FR- early Forced Regression, e-T- early Transgression, l-T- late Transgression. See text for details on each systems tract. From Catuneanu, 2006.
(MFS). The MFS develops when marine sediments reach their most landward position and forms the basal sequence boundary of the HST (Van Wagoner et al., 1988). Depositional trends are represented by a combination of aggrading and prograding packages, although occurring at a relatively low rate (Catuneanu, 2006). At the top, the HST is bound by the **regressive surface of marine erosion** (RSME), which erodes previously deposited nearshore marine sediment as relative sea level begins to fall (Plint and Nummedal, 2000; Catuneanu, 2006).

### 2.6.3 Falling Stage Systems Tract

The falling stage systems tract (FSST) begins at the onset of relative sea level fall at the shoreline and is characterized by a basinward shift in facies (Fig. 2.18b; Plint and Nummedal, 2000). Relative sea level fall leads to forced regression of the shoreline, and the FSST is characterized by offlap of strata. However, recognition of this systems tract is usually difficult, due to later subaerial erosion or transgressive ravinement of sediment (Plint and Nummedal, 2000). The FSST is bounded at the base by the RSME, which forms as the shoreface migrates seaward. As relative sea level fall continues, a subaerial erosion surface forms on top of the FSST, forming the sequence boundary that separates the FSST from the lowstand systems tract (LST). In updip areas, the FSST is characterized by sediment bypass, progressive valley incision, and formation of paleosols on the interfluves (e.g., Plint et al., 2001).

### 2.6.4 Lowstand Systems Tract

The LST is bounded at the base by the subaerial unconformity and its marine correlative conformity (Walker, 1992). The LST encompasses time between maximum regression and the beginning of relative sea level rise. During this period, sediment supply is greater than accommodation, leading to shoreline progradation, and normal regression of the shoreline (Fig. 2.18c). Although relative sea level rises, sediment supply is sufficient to allow for progradation of the shoreline. Once accommodation exceeds sediment supply rate, the LST ends and the first major flooding event signifies the beginning of the overlying transgressive system tract (TST; Posamentier and Vail, 1988).
2.6.5 Transgressive Systems Tract

The base of the TST is the transgressive surface that marks the top of the LST (Fig. 2.18d; Van Wagoner et al., 1988). The TST begins when the accommodation rate exceeds the sediment supply in the basin, and the shoreline begins to retreat. The TST is characterized by a series of flooding events that result in retrogradational packages being deposited (Posamentier and Vail, 1988). As the shoreline retreats, the basal surface of the TST is reworked by transgressive ravinement, and represents an erosion surface of a formerly subaerial environment (Walker, 1992). As relative sea level rise continues, entrapment of clastic sediment within fluvial and coastal systems results in greatly diminished sediment supply to the marine shelf. The maximum transgression may be recorded in the basin as a condensed section that represents slow deposition. The marine surface that corresponds to maximum shoreline transgression and maximum clastic starvation of the shelf is the MFS, which forms the top of the TST and sequence boundary (Walker, 1992).
CHAPTER 3

SHALLOW MARINE SEDIMENT TRANSPORT AND FACIES

3.1 Introduction

The following chapter provides a summary of the current understanding of sediment transport processes across marine shelves; particular attention will be paid to recent advances in the understanding of mud transport and deposition. Descriptions and interpretations of the various rock units identified within this study will then be discussed in terms of the sediment transport mechanisms outlined below. A firm grasp on modern sediment transport processes in shallow marine environments is an important prerequisite to the understanding and appreciation of the depositional environments in which ancient sediments accumulated.

3.2 Sediment Transport on a Shelf

In a generalized view of a shelf to shoreline profile, sedimentation occurs within the foreshore, shoreface, and offshore environments (Fig. 3.1). The foreshore, or beach environment, is made up of sediments deposited above the low tide line, and includes sediment deposited as a result of overwash during storms. The shoreface environment forms the innermost part of the shelf, where sediment transport is controlled mainly by waves, resulting in sandy sediment dominating (Plint, 2010). The shoreface is typically 5-10 m in height and has a steeper gradient, ~1:200, than the shelf, which is ~1:1000 (Plint, 2010). Sand dominates the middle to upper shoreface and commonly contains hummocky and swaley cross bedding. In the transition zone between the lower shoreface and the offshore environment, interstratified mudstone and sandstone beds are common, and sedimentary structures include hummocky cross stratification (HCS) and rippled sand beds. Bioturbation within the transition zone is commonly intense as a result of good oxygenation, high nutrient supply and low to moderate sedimentation and salinity stresses, all of which create a favorable environment for benthic organisms (MacEachern et al., 2010). Even further away from the shore and the effect of waves, sediment is predominantly made up of mud that is transported by geostrophic flows and long term
Figure 3.1 A generalized shallow marine profile: Wave base is approximately equal to half the wave-length of a surface wave (L/2), and a typical wave base is 5-15m. Expected sedimentary features for offshore, shoreface, and foreshore environments are shown. Waves entering shallow water become asymmetric and contribute to longshore water circulation (see text and Figure 3.2 for details). Modified from Walker and Plint, 1992; Plint, 2010.
oceanic currents. In this offshore setting, the rate of clastic sediment supply can be low, and pelagic sediment may accumulate.

Marine shelves are typified by water depths of less than 100 m, and the sea floor has a low topographic gradient (Plint, 2010). As a result, energy derived from waves can have a significant impact on sediment transport across the seafloor, a process that is accentuated during storm conditions. Waves entering the nearshore zone result in longshore and rip currents (Fig. 3.2). Although capable of transporting coarse sediment such as sand and gravel, rip currents quickly lose energy and die out no more than a few hundred metres offshore. Longshore currents on the other hand are more efficient at transporting sediment along the coast. Away from the shoreface, sediment transport to deeper water is driven primarily by storm-generated combined flows.

3.2.1 Sediment transport during storms

Storms dominate shelf processes on about 80% of continental margins, and wave-formed sedimentary structures and erosional features are common in the rock record (Swift and Thorne, 1991). Although storm wave base is highly variable and depends on the intensity of an individual storm, the generally accepted depth for maximum storm wave base is ~200 m (Walker and Plint, 1992).

During a storm, onshore wind stress generates a coastal set-up where the sea surface near the shore is elevated, sometimes by up to a few metres (Fig. 3.3). The resulting pressure gradient moves water near the seabed in an offshore direction, in response to the hydraulic head. The bottom flow, initially moving perpendicular to the shoreline, is progressively deflected to the right due to Coriolis force. In the southern hemisphere this deflection occurs to the left. Currents that are sustained by a balance between pressure gradients and Coriolis force are called geostrophic currents (Swift et al., 1986).

Geostrophic currents, along with regular wave-induced oscillatory flows are very important in transporting sediment during storms, and result in combined flows. Sediment that is thrown into suspension by oscillatory wave-generated flows is then carried obliquely across the shelf by more slowly-moving geostrophic currents.
Figure 3.2 Water circulation patterns in a nearshore area. As waves enter shallow water above wave base, orbital motion of waves become asymmetric such that landward strokes are short and strong while seaward motion is longer and weaker. As a result water piles up against the beach and water is carried back seaward in narrow rip-current systems, carving channels in the upper shoreface. Waves approaching the beach at an oblique angle tend to cause more unidirectional movement of the longshore currents. From Plint, 2010.
Figure 3.3 Characteristics of a storm driven shelf system. Onshore winds generate a coastal set-up that creates a pressure gradient within the system (A). As a result lower layers of water are carried seaward as relaxation flows. Coriolis force deflects the flow progressively to the right generating coast parallel to coast oblique geostrophic currents (B). A combined flow is generated when oscillating current driven by waves interacts with the geostrophic current. Sediment suspended by waves is therefore able to migrate further offshore and along shore. From Plint, 2010.
In order to understand circulation patterns within the Western Interior Seaway (WIS), various authors in the 1990s developed models that attempted to mimic the environmental conditions affecting water circulation in the Late Cretaceous (e.g. Ericksen and Slingerland, 1990; Slingerland et al., 1996). Model parameters such as paleobathymetry, temperature, and salinity were based on stratigraphic and isotopic data available at the time, and model runs were repeated using various numerical combinations of the parameters. Resulting models suggested that circulation within the WIS was largely storm dominated, with dominant winds from the north and northeast that induced longshore currents to the south. This prediction was supported by paleoflow data from storm influenced deposits that spanned the Middle Albian to Early Campanian (Ericksen and Slingerland, 1990). Models also suggested that fresh-water runoff entering the western part of the WIS exited the seaway in a southerly directed coastal current, (Slingerland et al., 1996).

Studies of Cretaceous sediments in the WIS, using paleocurrent indicators such as combined flow ripples and gutter casts, support modelling results, and indicate that geostrophic currents led to a net southerly transport of sediment (e.g. Hart et al., 1990; Slingerland and Keen, 1999; Varban and Plint, 2008 a,b; Hu and Plint, 2009).

3.2.2 Mud Transport and Deposition on Continental Shelves

Major rivers supply terrigenous mud to the sea. However, most of that mud is trapped on the inner shelf and only a small percentage of terrigenous mud may escape to the deep sea (Plint, 2010). Mud accumulation on continental shelves is dependent on factors such as the amount of suspended matter and hydrodynamic energy that controls transport and deposition of suspended mud particles (McCave, 1972). Although some mud is deposited at or near river mouths, resuspension of mud by waves during storms acts to redistribute sediment further along and across shelves by geostrophic flows and permanent along-shelf currents (Cattaneo et al., 2007). Such is the case of modern subaqueous muddy deltas, such as the Amazon Shelf (e.g. Nittrouer et al., 1996), Adriatic shelf (e.g. Cattaneo et al., 2007), and East China Sea (e.g. Liu et al., 2007; 2009). A wedge-shaped body of mud, partitioned internally by seaward-dipping clinoforms develops as a result of storm wave resuspension moving mud across the top of the
subaqueous delta. The wedge-shaped mud body may extend for hundreds of kilometres along the shelf beyond the river mouth, with clinoforms dipping seaward at up to 1° (Fig. 3.4; Plint, 2010).

Transport mechanisms of suspended mud are now better understood as a result of flume studies in which the amount of sediment, flow velocity and water depth can be varied (e.g. Schieber et al., 2007, 2010; Schieber, 2011). Experiments reveal that resuspended mud forms aggregates of silt and sand sized particles, commonly known as ‘floccules’. Flocculation can occur by three main processes: electrostatic forces attracting particles together; organisms concentrating mud particles into fecal pellets; and bonding of dissolved mineral particles by dissolved organic material (Plint, 2010). Once formed, floccules behave hydraulically as silt- or even sand- sized particles, and are capable of forming ripples (Schieber et al., 2007). Although mud deposition as a result of suspension settling occurs, more recent research based on textural analysis, reveals that a significant amount of mud is transported and deposited as a result of wave energy and unidirectional current energy (Aplin and Macquaker, 2011). Additionally, storm waves are capable of forming a dense fluid mud layer within the wave boundary layer which, under the influence of gravity, can flow downslope as a gravity driven flow (Traykovski et al. 2000). Such transport mechanisms can only persist however, as long as wave energy is vigorous enough to reach the sea floor and keep mud in suspension (e.g. Traykovski et al., 2000). Work by Macquaker et al. (2010) shows that wave-enhanced sediment-gravity flows (WESGF) are capable of transporting large volumes of fluid mud across low gradient shelves. WESGF deposits were found to be common in both modern and ancient shallow marine settings.

Mapping of muddy sediments from the Western Interior shows that mud packages can extend for hundreds of kilometres offshore (e.g. Varban and Plint, 2008 a, b; Hu and Plint, 2009). Applying the new knowledge of mud transport to ancient mudstone rocks of the Western Interior leads to the conclusion that mud should no longer be thought of as sediment deposited by low energy and suspension processes, but in fact records deposition in dynamic sedimentary environments.
Figure 3.4 Model showing how mud can accumulate down-drift from a major river mouth. Storms continually re-suspend mud that is then carried along the shelf by geostrophic currents, as well as permanent along-shelf currents. Mud deposited by permanent along-shelf currents may extend for over 100 km downwind. Seaward, lateral accretion of mud forms clinothems that downlap onto a sediment starved surface, such as a maximum flooding surface. This model is used to explain mud transport on mud rich shelves such as the Adriatic and East China Sea. From Plint, 2010.
3.3 Sedimentary Facies Descriptions

Sediments of the Muskiki Member can be described in terms of facies— a body of rock with specific characteristics (Reading, 2001). The term facies was originally introduced by Gressly (1838), to describe distinctive bodies of rock that formed under certain physical and biological conditions of sedimentation, and which reflect a particular depositional environment. A facies, which may be a single bed but more commonly a group of beds, may be defined on the basis of colour, texture, composition, fossil content, and sedimentary structures observed (Reading, 2001). Examining a single bed or facies is, however, an insufficient basis for interpreting an ancient depositional environment. An understanding of many examples of facies, coupled with documentation of vertical and lateral facies distributions, is necessary to allow a full characterization of a depositional environment.

A progressive vertical or lateral change in the character of a rock unit, such as grain size, sand content, and sedimentary structures, leads to interpretation of a facies succession (Dalrymple, 2010). The gradational aspect to a facies succession is important, because it implies a vertical, in addition to a lateral transition between two facies. Known as Walther’s Law, the vertical gradational transition between two facies suggests the two facies represent environments that were at one time laterally adjacent (Dalrymple, 2010). In contrast, sharp or erosional contacts between facies suggest important surfaces that could be unconformities that may separate stratigraphic sequences.

3.3.1 Previous Work and Methods

A detailed facies analysis for the Muskiki and Marshybank formations in northwestern Alberta and northeastern British Columbia was conducted by Plint and Norris (1991) on the basis of outcrop data combined with subsurface well control. Their study recognized 14 facies that include marine and non-marine strata (Table 3.1). Within a study area that overlaps with this study, Nielsen et al. (2008) recognized six facies within the Verger Member, which is broadly equivalent to the Muskiki Member in the Foothills. The present study in southern Alberta and northwestern Montana uses data collected from outcrop and core, combined with macrofossils to define five facies that represent the various depositional environments of the Muskiki Member. The facies
<table>
<thead>
<tr>
<th>Plint and Norris (1991) Northwestern Alberta and B.C.</th>
<th>Nielsen et al., 2008 (Southern Alberta Plains)</th>
<th>This Study (Southwestern Alberta Foothills and Plains)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Marine facies</strong></td>
<td><strong>Marine facies</strong></td>
<td><strong>Marine facies</strong></td>
</tr>
<tr>
<td>1. Non-sideritic laminated siltstone</td>
<td>1. Fine laminated calcareous shale</td>
<td>1. Dark laminated mudstone (A)</td>
</tr>
<tr>
<td>2. Thin bedded siltstone and sandstone</td>
<td>2. Bioturbated non-calcareous shale</td>
<td>2. Sideritic bioturbated silty mudstone (B)</td>
</tr>
<tr>
<td>3. Blocky sideritic siltstone</td>
<td>2. Bioturbated non-calcareous shale</td>
<td></td>
</tr>
<tr>
<td>5. Bioturbated sandy siltstone</td>
<td>4. Bioturbated fine-grained shaly sandstone</td>
<td>3. Heterolithic Facies (C)</td>
</tr>
<tr>
<td>6. Weakly bioturbated sandy siltstone</td>
<td></td>
<td></td>
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<tr>
<td>7. Hummocky cross-stratified sandstone</td>
<td></td>
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<tr>
<td>8. Chert pebble conglomerate and pebbly mudstone</td>
<td>5. Lag and related facies</td>
<td>4. Conglomerate and lag deposits (D)</td>
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<td></td>
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<td></td>
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<tr>
<td>9. Swaley cross-stratified sandstone</td>
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<td>5. Ooidal ironstone (E)</td>
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<td>10. Parallel laminated sandstone</td>
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<tr>
<td>11. Parallel bedded sandstone</td>
<td></td>
<td>Not present in study area</td>
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<tr>
<td>12. Crudely stratified pebbly sandstone</td>
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<tr>
<td><strong>Non-marine facies</strong></td>
<td><strong>Non-marine facies</strong></td>
<td><strong>Non-marine facies</strong></td>
</tr>
<tr>
<td>13. Cross-bedded sandstone</td>
<td>6. Paleosol</td>
<td>Not present in study area</td>
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<tr>
<td>14. Rooted carbonaceous siltstone and sandstone</td>
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Table 3.1 A comparison of the different facies observed for the Muskiki and Marshybank formations in northwestern Alberta and B.C., with that of southern Alberta Plains, and the Muskiki and Marshybank members within this study. This study did not recognize non-marine facies within the study area. Middle to upper-shore faced deposits (facies 9-12 of Plint and Norris, 1991) were also not observed.
recognized by Nielsen et al. (2008) are similar to those presented herein, although their study did not recognize the presence of a thin but significant ooidal ironstone facies (Table 3.1). Fewer facies are recognized in the present study than those defined by Plint and Norris (1991) because the Marshybank Formation, which comprises nearshore and terrestrial facies in northwestern Alberta (Fig. 1.6), is thin or absent within the present study area.

3.3.2 Facies A: Dark, laminated mudstone

**Description:** Facies A is well developed above major flooding surfaces, such as above the Cardium Formation at Burnt Timber Creek (Fig. 3.5), and in distal cores 10-30-18-15W4 and 13-20-17-7W4 (Fig. 3.6). Facies A consists of dark grey to black, laminated mudstone with occasional mm-scale siltstone to very fine-grained sandstone laminae. Very fine-grained sandstones sometimes preserve low-amplitude ripples. Thin sections reveal that the mudstone is commonly made up of silt-sized aggregates of clay, and very fine-scale ripple laminations (Fig. 3.7). Laminated mudstone of facies A varies from a few centimetres to several metres in thickness, and usually grades upward into bioturbated mudrock of facies B, or heterolithic bedding (facies C). Facies A commonly unconformably overlies coarser-grained rocks such as conglomerates (facies D).

Abundant mm- to cm-scale bentonites are commonly preserved within facies A.

Bioturbation is weak or absent (bioturbation index of 2; MacEachern et al., 2010), and is dominated by *Planolites* and *Gordia*. Molluscan shell debris is common, and shells are sometimes preserved as lags in distal cores (Fig. 3.8A). Biogenic material such as wood, phosphatic nodules, and foraminifera are rare. Portions of facies A in core have a strong calcite cement and exhibit abundant white specks (Fig. 3.8B). Scanning Electron Microscopy (SEM) imaging of calcareous cement reveals abundant whole and broken coccoliths (Fig. 3.9), fish scales and organic debris (Fig. 3.10A-D), as well as framboidal pyrite (FeS$_2$; Fig. 3.10E-F).

Facies A occasionally exhibits a rusty appearance in outcrop due to weathering of finely disseminated pyrite. Septarian siderite concretions, usually 5-10 cm in diameter, can be found sporadically throughout the facies, and are commonly nucleated around inoceramid bivalves and ammonites.
Figure 3.5 Dark, laminated mudstone of facies A. A) Contact between Cardium Formation and Muskiki Member at Burnt Timber Creek (dashed white line). B) Laminated and unbioturbated mudstones of facies A grade upward into bioturbated deposits of facies B at Drywood River. Prominent grey and pale yellow bentonites commonly found within facies A and B, and are denoted by --v--. FA- Facies A; FB- Facies B. Drywood River Section A” 0-10 m.
Figure 3.6 Core box photos showing laminated fabric of facies A. A) Overview of facies in core 10-30-18-15W4, ~556-563 m; scale bar= 20 cm. B) Fine silty laminae (arrows)- note ripple-cross laminae shown by lower two arrows, indicating current transport; core 10-30-18-15W4, ~561m; black and white scale bars are 1 cm each. C) Overview of closely-spaced bentonites (red arrows) interbedded with mudstones of facies A; core 13-20-17-7W4, ~537-540.6 m; Scale bar= 20 centimetres.
Figure 3.7 Microstructures within facies A. A) Small-scale current ripple laminations in core 13-20-17-7W4, ~543 m. Arrows indicate silt and clay aggregate (CA) dominated laminae draping a ripple formed of siliceous silt. B) Close up view of clay aggregates in A. C) Clay aggregates (CA), foraminifera (F), and organic matter (OM) in core 13-20-17-7W4, ~539 m.
Figure 3.8 Sedimentological features within facies A, represented in cross-sectional cuts of core. A) Shell lag in core 13-20-17-7W4, 543.4 m. B) Abundant white specks observed in cemented calcareous mud in core 8-36-24-13W4, ~547 m. White specks were identified by SEM as fecal pellets made up of coccoliths. Black and white scale bars for A and B are 1 cm each.
Figure 3.9 Scanning Electron Microscope images of calcareous coccoliths and micrite present in the Muskiki Member, facies A, core 13-20-17-7W4, ~539 m. A) Overview of coccolith fragments and texture in sample. Arrows indicate less fragmented coccoliths. B-E) Different views of coccoliths showing textural variety. F) ‘White speck’ fecal pellet made up of coccoliths, enclosed in a clay rich (CM) matrix. Dashed line shows contact between clay matter and fecal pellet.
Figure 3.10 Scanning Electron Microscope images of fish scales (Fs) and organic matter (OM) of facies A in core 13-20-17-7W4, ~539 m. A) Fish scale. B) Close-up of A, showing growth lines on the fish scale, a characteristic feature of phosphatized fish debris C) Phosphatized fish scale surrounded by a clay matrix (CM). D) Organic matter, probably plant material, surrounded by clay matrix. Organic and phosphatic material was identified using elemental mapping. E) Framboidal pyrite surrounded by clay matrix F) Detailed view of frambooidal pyrite, showing individual pyrite crystal structure.
**Interpretation:** The fine grain size of facies A is indicative of a distal marine setting that was subject to short-lived episodes of high energy, which punctuated a generally quiet depositional environment. Preservation of thin siltstone and sandstone laminae indicate that deposition took place between fair weather and storm wave base. Both oscillatory wave and geostrophic currents, which formed combined flows, were efficient in transporting sediment offshore during storms (Swift et al., 1987; Varban and Plint, 2008a; Plint, 2010). Regional correlations and core show that mudstone within the Muskiki Member was transported at least ~250 km away from the shoreline. Transport of mud with as much as 85% water content has been shown to be capable of being eroded, thrown into suspension, and redeposited by bottom currents (Schieber et al., 2010). Silt sized aggregates of clay particles within the Muskiki Member indicate reworking of consolidated mud by storms, which transported the aggregate grains as bedload (Schieber and Southard, 2009).

The presence of coccoliths in core 13-20-17-7W4 suggests that, although most sediment was land-derived and transported offshore by storm-driven flows, pelagic sediment was also being deposited. The well-cemented calcareous component of facies A is most likely the product of slow sediment accumulation in the distal offshore setting (Schieber, 1999).

The low level of bioturbation suggests dysaerobic conditions were present at the sediment-water interface, which would have prevented many benthic organisms from thriving.

Framboidal pyrite is interpreted to have precipitated from iron-dominated porewaters during early diagenesis of sediment. High sulfide production rates caused porewaters to reach supersaturation with respect to FeS$_2$, which resulted in framboidal pyrite precipitation (Taylor and Macquaker, 2000). Taylor and Macquaker (2000) proposed that framboidal pyrite in Lower Jurassic clay-rich mudstones in England may have formed as a result of high bacterial sulfate reduction, as a result of low oxygen content within bottom waters and high organic matter content. Nielsen et al. (2008) reported an average total organic carbon (TOC) value of 1-2% for the Verger Member in southern Alberta, but higher TOC values of 3-4% was observed in core 13-20-17-7W4, where the framboidal pyrite is present. The higher organic carbon supplied by rain-out of
pelagic sediment may have promoted increased bacterial sulfate reduction, and precipitation of framboidal pyrite.

3.3.3 Facies B: Sideritic bioturbated silty mudstone

**Description:** Facies B consists of bioturbated silty mudstone to coarse siltstone and very fine-grained sandstone. In outcrop, the facies has a massive to rubbly appearance with primary bedding rarely preserved (Fig. 3.11). Bioturbated mud and siltstone units range from a few decimetres to several metres in thickness. Mud and siltstone units are intensely bioturbated, and give the facies a structureless appearance. Thin sections show that, like facies A, clay forms silt-sized clay aggregates (Fig. 3.12). Facies B gradationally overlies facies A, and grades upward into facies C. An upward increase in silt and very fine-grained sand is common in facies B, with stratigraphically higher beds containing as much as 40% sand.

Pervasive bioturbation has removed much of the evidence for primary stratification. Ichnofossils include *Chondrites, Gordia, Helminthopsis, Planolites, Schaubcylindrichnus*, and *Skolithos*, giving the facies a bioturbation index between 3 and 4 (MacEachern et al., 2010; Fig. 3.13). Lingulid brachiopods are also present, as well as fish and bivalve shell debris.

In distal examples of facies B, pore-filling glauconite appears as an authigenic component within bioturbated silty sandstone intervals that immediately underlie ooidal ironstone of facies E (Fig. 3.14). Rare fecal pellets can be recognized, and small amounts of pyrite are present within some burrows, or in the form of mm- to cm- scale nodules (Fig. 3.13A).

Siderite nodules form a distinctive component of facies B. They occur in outcrop as continuous bands or as isolated nodules averaging from 5-10 cm in diameter, although some 40-50 cm nodules are also present. Siderite nodule bands occur more commonly toward the top of upward coarsening successions, although many nodules are also distributed throughout the facies, regardless of stratigraphic position.

**Interpretation:** Remnants of lamination preserved in facies B suggest that deposition took place under episodic storm conditions above storm wave base, but below fair weather base. Homogenization of sediment due to intense bioturbation suggests that
Figure 3.11 Outcrop expressions of Facies B. A) Rubbly bioturbated siltstone at Horseshoe Dam, ~1-4 m. Arrow shows a coarsening up succession, overlain by finer-grained rocks. Orange bands are thin, micaceous bentonites. B) Overview of bioturbated siltstone at Lynx Creek, 1-4 m. Note the large sideritized burrows and nodules.
Figure 3.12 Thin section views of Facies B. A) Thin section scan showing bioturbated mud and siltstone in core 10-30-18-15W4, ~550.8 m. Arrow shows stratigraphic top. B) Extensive bioturbation and Planolites in core 10-30-18-15W4, ~552 m; Note preservation of few fine scale laminae. Red circle shows probable Rhizocorallium. Arrow shows stratigraphic top. C) Thin section view of scanned section in A, showing silt and sand-sized clay aggregates (arrows), similar to those present in facies A. Clear grains are quartz silt. Thin section showing interbedded mudstone and siltstone beds. Note the presence of small scale erosional scours (white arrows), as well as pervasive bioturbation (red arrows). D) Thin section scan showing pervasive bioturbation (red arrows), as well as an erosional base to a storm bed (black arrows) in core 13-20-17-7W4, ~543 m. White arrow shows stratigraphic top.
Figure 3.13 Core expression of Facies B. A) Thoroughly bioturbated muddy sandstone showing pyritized burrow (dashed) and pyrite nodules (arrows) in core 8-36-24-13W4, ~544 m; B) *Schaubcylindrichnus* (Sc) and *Planolites* (Pl) in core 16-4-22-15W4, ~526 m. A combined flow ripple can be seen toward the top of the image (arrow). C) Pyritized *Gordia* (arrows) in core 8-13-22-16W4, ~559.6 m. D) Silt filled burrow (B) and *Chondrites* (Ch) in 16-4-22-15W4, ~567.2 m. All scale bars are in centimetres.
Figure 3.14 Bioturbated muddy sandstone of facies B. 
A) Glauconite rich bioturbated sand in core 2/10-30-18-15W4, ~550 m. B) Fecal pellets in glauconitic sand in core 8-13-22-16W4, ~565 m. C) Thin section micrograph showing Planolites burrow and sub-angular to sub-rounded glauconite grains (arrows) in core 6-3-24-13W4, ~520 m. Area in red box enlarged in D. 
D) Detailed thin section view of red box in C, showing clay rich matrix and sand grains.
the sea floor was well oxygenated, and traces represent the *Cruziana* ichnofacies (MacEachern et al., 2010). The coarser grain size indicates that sediments were deposited closer to the shoreline, relative to facies A. Small scale erosional features observed in thin section are also consistent with a storm influenced depositional environment (e.g. Schieber, 1998). As with facies A, the presence of silt-sized clay aggregates suggests mud was transported offshore as reworked aggregate grains that travelled as bedload by storm-induced geostrophic currents.

Authigenic precipitation of glauconite typically occurs on the outer margins of modern continental shelves, in areas of low sediment input (Stonecipher, 1999). Glauconite precipitation occurs just below the sediment-water interface and requires slow sedimentation rates in a semi-restricted environment (Amorosi, 1997). As a result of slow sedimentation rates, glauconite concentrates as relatively thin deposits that are characterized by non-selective spatial distribution of glauconite grains (Amorosi, 1997). Glauconite present in facies B is interpreted to have precipitated in pores below the sediment-water interface, where low permeability provided the reducing environment necessary for growth (Stonecipher, 1999). High-energy storm conditions would have intermittently reworked the seafloor, resulting in the concentration of glauconite grains.

Pyritized burrows and pyrite nodules observed in core are interpreted to have precipitated during early diagenesis. Pyrite precipitation probably followed glauconite formation, and is interpreted to have occurred in small amounts due to lack of sufficient organic matter, which is necessary in promoting extensive sulphate reduction (e.g. McKay et al., 1995). The large quantities of siderite (FeCO$_3$) also suggest that sulphide activities within porewaters were low, which in turn allowed for significant amounts of siderite to form (Macquaker and Taylor, 1996). The preferential occurrence of siderite bands near the top of upward coarsening successions is interpreted to be a result of formation under slightly reducing conditions taking place during diagenesis, where coarser sediment may have provided greater permeability for iron-rich porewaters (McKay et al., 1995; Stonecipher, 1999).

The presence of glauconite, pyrite and abundant siderite in bioturbated siltstone of facies B closely resembles the mineral assemblage that was present in an aerobic offshore shelf environment represented by bioturbated siltstone of the Marshybank Formation in
northwestern Alberta (McKay et al., 1995). The paragenetic sequence glauconite, pyrite, and siderite was suggested for the Marshybank Formation, and therefore may provide a reasonable explanation for the diagenetic mineral phases observed in facies B of the present study (McKay et al., 1995).

3.3.4 Facies C: Heterolithic Facies

**Description:** Heterolithic facies consist of interstratified siltstone and sandstone beds, with rare interbedded mudstone (Fig. 3.15). Sandstone beds are usually fine-grained and laminated, 2-4 cm thick, and show an upward increase in abundance, ranging from 20 to 80% sand content. Beds are usually rippled, and may show combined flow ripple lamination. Hummocky cross stratification (HCS) is rare, and is only observed in thicker, laterally discontinuous sandstone beds that are 10-20 cm thick (Fig. 3.16). Sandstone beds are usually interstratified with moderately bioturbated siltstone, similar to that of facies B. Gutter casts are also sometimes preserved, although never more than a few cm deep and 5-10 cm in width. Facies C typically overlies facies B and is sometimes overlain by conglomerate (facies D), but more typically by flooding surfaces and finer-grained sediment (facies A and B).

Ammonites and inoceramid bivalves are scattered throughout facies C. *Planolites*, *Skolithos*, and *Thalassinoides* dominate the ichnofauna. Rare sideritized logs containing *Teredolites* were observed at Sullivan Creek, within bioturbated silty sandstone (Fig. 3.17).

**Interpretation:** Facies C is most reasonably interpreted to represent an offshore to lower shoreface transitional setting deposited under alternating high and low energy conditions (Varban and Plint, 2008a; Plint, 2010). Interbedded siltstone and mudstone represent sedimentation below fair-weather base, but above storm wave-base, and the overall sandier-upward succession reflects progradation of the shoreline. Combined flow ripples indicate deposition under the influence of storm waves and geostrophic flows which were responsible for transporting sand from a shoreline to the southwest (e.g. Arnott and Southard, 1990; Duke et al., 1991). Rare HCS may record larger storms.

The presence of abundant bioturbation indicates a well-oxygenated sea floor that was episodically reworked by wave energy. Ichnofossils present within facies C further
Figure 3.15 A) Interbedded sand- and silt-stone of facies C at Burnt Timber Creek, 34 m above the base of the Muskiki Member.
B) Bioturbated siltstone with ribby sands at Sheep River, 32 m above the base of the Muskiki Member. Image shows the top of one coarsening up succession (arrow), and the base of the next.
Figure 3.16 Heterolithic character of facies C at Highwood River West, 55 m above the base of the Muskiki Member. A) overview of stratigraphy and HCS lens interstratified with sandy ribs and siltstone. B) Close-up view of gutter cast, showing hummocky cross stratified fine sand.
Figure 3.17 Ichnofossils in facies D. A) *Teredolites* in wood at Sullivan Creek, 42.2 m above the base of the Muskiki Member. B) *Thalassinoides* on top of a rippled fine upper sand bed at Sullivan Creek, 43.6 m above the base of the Muskiki Member.
support deposition within a fully marine setting, in the *Cruziana* ichnofacies (MacEachern et al., 2010). Siderite nodules are interpreted to have formed during early suboxic diagenesis, similar to siderite found in facies B (McKay et al., 1995).

### 3.3.6 Facies D: Conglomerate and lag deposits

**Description:** Conglomerates are present on regionally traceable erosion surfaces on the E7 surface at the base of the Muskiki Member, at the top of the member, on the DE1 surface, and as well as a few conglomerates within the member (Figs. 3.18, 3.19). Conglomerate consists mainly of chert and quartz that was originally derived from Paleozoic rocks in the evolving fold and thrust belt of the Cordillera. Rarely, locally derived phosphatic and carbonate clasts may form pebble beds. Conglomerates are generally thin, varying from one pebble veneers to a few decimeters (10-40 cm). One exception is a 3 m thick conglomerate present at Millarville. Stratification within conglomerates is not readily visible. Conglomerates are usually poorly sorted, and consist of sub-rounded to well-rounded clasts 0.4-4 cm in diameter. Although most conglomerates are siliceous in origin, phosphatic and mud intraclasts are sometimes preserved within lags as well (Figs. 3.20, 3.21).

Conglomerates that occur at the top of the Muskiki Member are interbedded with bioturbated granular silty sandstone, and cm-scale gravel scours are sometimes preserved within the bioturbated facies (Fig. 3.22). Beds immediately below lags usually have preserved *Thalassinoides* and *Rhizocorallium*, and gravel-filled ?*Skolithos* are present at the base of one conglomerate at Highwood River (Fig. 3.23). Pervasive siderite cementation of conglomerate is observed at several outcrops, including Millarville, Mill Creek, and Highwood River (Fig. 3.19).

A lag of phosphatic pebbles is present at the Kevin outcrop in Montana (Bed 108 of Cobban et al., 1976), where phosphatic intraclasts lie on a subtle, yet sharp erosion surface with a lag of inoceramid shell debris. Pebbles are grey or green in colour, sub to well-rounded, and 2-4 cm in diameter (Fig. 3.24). White, mm scale *Chondrites* can be seen on some surfaces of phosphate pebbles, and glauconite grains are observed on cracked surfaces and in thin sections of many pebbles. Some intraclasts are phosphatised steinkerns of *Baculites* ammonites and inoceramid shells.
Figure 3.18 Conglomerates of facies D. A) a 20 cm thick conglomerate at Burnt Timber Creek, 8 m above the base of the Muskiki Member. Scale bar is 20 centimetres long. B) Gravel conglomerate at Mill Creek, at the top of allomember ME5 (129 m). C) Close-up of B, showing pervasive siderite cementation of conglomerate. Handlens ~3 centimetres across.
Figure 3.19 Conglomerate deposits of facies D. A) Series decimetre scale beds of conglomerate (thick arrows) at Sullivan Creek. Small arrow indicates continuous bed of clean cemented sand; ~54-58 m above the base of the Muskiki member; B) Overview of conglomerate at the top of the Muskiki Member, Highwood River, 80 m above the base of the Muskiki Member; stratigraphy younging to the right. C) Close-up view of conglomerate in B at Highwood River. Note sideritized top surface.
Figure 3.20 A) Plan view of a siliciclastic pebble lag at Lynx Creek, 3.5 m above the base of the Muskiki Member. B) Belemnite fragment surrounded by chert pebbles on same lag surface as A. Scale bar subdivisions are 1 cm each.
Figure 3.21 A) Pits (arrows) where mud intraclasts have been eroded can be seen at Sullivan Creek, associated with gravel conglomerate and lag; 57 m above the base of the Muskiki Member. Scale bar is 20 centimetres. B) Phosphatized *Baculites* fragment in chert pebble lag at Lynx Creek, 3.4 m above the base of the Muskiki Member. Scale bar subdivisions are 1 cm each.
Figure 3.22 A) Conglomerate at the top of the Muskiki Member at Highwood River West, 77 m above the base of the Muskiki Member. Siltstone above is bioturbated and contains granules interspersed throughout, as well as gravel pods. B) Close-up view of siltstone above conglomerate in A, showing gravel lenses (arrow). C) Gravel filled Thalassinoides (arrows) in conglomerate at Sullivan Creek, 57 m above the base of the Muskiki Member. Scale bar is 20 centimetres.
Figure 3.23 A) Horizontal burrows (dashed lines) on lag surface at Lynx Creek, 3.4 m above the base of the Muskiki Member. B) *Rhizocorallium* (dashed line) in bioturbated sand just below lag surface in A. Scale bar subdivisions are 1 cm each.
Figure 3.24 Phosphatic pebble beds in facies D. A) A lag of phosphatic nodules (arrows) in core 8-36-24-13W4, ~545 m. The nodules are in mudstone of facies A, just below transition into facies B; black and white scale bars are 1 cm each. B) Phosphatic pebbles from bed 108 at Kevin, Montana, 89 m. Bivalve shell (Bv) and Baculites fragment (Bc) moulds preserved. C) Close-up view of Chondrites and surface cracks in nodule from bed 108 in B. D) Cracked phosphatic pebble from bed 108 in B, showing glauconite (green) and siderite (brown) grains.
**Interpretation:** Siliceous conglomerate beds are interpreted as transgressive lags that formed following landward movement of the shoreline. Gravel was originally transported to the shore by rivers, followed by possible subaerial exposure (Plint and Norris, 1991). Subsequent transgression and reworking of shoreline deposits by waves redistributed the gravel as a veneer on the seafloor (e.g. Bergman and Walker, 1987; Plint, 1988; Hart and Plint, 1995; Hu and Plint, 2009). Gravel-filled scours are interpreted to have formed in a nearshore environment influenced by longshore and rip-currents (Clifton, 2003; Plint, 2010).

The presence of abundant *Rhizocorallium* and *Skolithos* below pebble-veneered transgressive surfaces, as well as locally-derived mud clasts suggests the seafloor was a semi-consolidated firmground, within the *Glossifungites* ichnofacies, prior to deposition of some of the conglomerates (MacEachern et al., 2010).

The thick conglomerate at Millarville may originally have been deposited in a shoreface setting close to a gravel-bearing river mouth. However, the interbedded conglomerate and mud suggests complete reworking during transgression so that the succession visible today must represent a shallow marine depositional environment.

Siderite cement within conglomerates is interpreted to have formed during early diagenesis. A similar interpretation has been presumed for sideritized conglomerates that occur at the top of the Marshybank Formation further north (McKay et al., 1995).

Phosphatic pebble beds are associated with condensed sections, depositional breaks, and/or unconformities (Loutit et al., 1989; Flügel, 2010). Phosphorite-rich sediment can form at the sediment-water interface or during early diagenesis through physio- or biochemical precipitation (Flügel, 2010). Phosphatic nodules are interpreted to have formed during shallow burial, by precipitation and cementation of apatite around nuclei made up of pre-existing baculites shells, as well as inoceramid prisms and any detrital sediment grains that were available on the seafloor (e.g. Soudry and Lewy, 1988; Lewy, 1990; Nathan et al., 1990). Periodic exhumation of the nodules on the seafloor, by waves and currents, concentrated the nodules as a lag of clasts, and are today observed as pebbles. The presence of glauconite within phosphatic nodules in bed 108 at Kevin reinforces the interpretation of a surface that represents a major depositional break. The phosphatic bed is significant because it lies on a major unconformity, the DE1 surface,
and represents a significant depositional hiatus within the Muskiki and Marshybank
stratigraphy. The surface merges northward in the subsurface with ooidal ironstone of
facies E (equivalent to the MacGowan Concretionary Bed).

3.3.7 Facies E: Ooidal ironstone

**Description:** Ooidal ironstone is usually developed immediately above the
unconformity that defines the top of the Muskiki and/or Marshybank members. The
exception is located at Castle River, where ooidal ironstone is also present on the
flooding surface that defines the base of the Muskiki Member (ie. E7 surface). Facies E
is observed in core over the north-central part of the study area, and in outcrop at Kevin,
Montana (Fig 3.25). The ooidal ironstone in the subsurface appears to correlate with the
MacGowan Concretionary Bed at Kevin (bed 100 of Cobban et al., 1976). The
MacGowan Concretionary Bed is rich in chert and quartz pebbles and granules that are
cemented by siderite (Fig. 3.26). Facies E is usually intensely bioturbated, and heavily
sideritized and cemented. Well-developed *Thalassinoides* are common at the base of the
ooloidal bed in the subsurface, and extend a few cm into the underlying sediment. Most
ooloids within the ironstone appear to have been dissolved, leaving voids (Fig. 3.27).
Where ooids are intact however, concentric growth rings are preserved, and ooids
resemble goethite ooids found in the Bad Heart Formation of northwestern Alberta
(Donaldson, 1997). The original composition of ooids within the present study however,
is not known.

**Interpretation:** The ooidal ironstone facies lies on a regional unconformity, the
DE1 surface, and is interpreted to have formed during marine transgression that initiated
deposition of the overlying Dowling Member. The presence of extrabasinal chert pebbles
and granules within facies E implies a previous sea level fall that led to the introduction
of chert clasts, presumably by rivers. The lowstand deposits were then reworked during
transgressive ravinement. The ooidal facies is generally interpreted as evidence of very
slow clastic sedimentation (Macquaker and Taylor, 1996). Extensive *Thalassinoides* at
the base of facies E imply that the sea floor was a *Glossifungites* firmground, consistent
with erosion of compacted sediments (MacEachern et al., 2010). The ooidal ironstone
facies in this study is similar to that of the Upper Coniacian Bad Heart Formation.
Figure 3.25 Representative core photos of ooidal ironstone facies E. A) Ooidal ironstone at the top of the Muskiki Member in core 16-4-22-15W4, ~540 M. B) Close-up view of box in B, showing ooidal ironstone contact and nodular fabric. C) Close-up of C, showing ooid rich zone (box) within ironstone facies. Scale bar is 20 centimetres.
Figure 3.26 A) Representative thin sections of the MacGowan concretionary bed (Bed 100) at Kevin, Montana, 85 m above the base of the Muskiki Member. Note the poor sorting and wide range in clast shape. Each section is ~4.5 X 2.5 cm. B) Plain polarized light micrograph of red box in A (approximate), showing siderite cement (Sid) surrounding limestone (Lm) and quartz (Q) clasts. C) Cross polarized view of B.
Figure 3.27 Petrographic details of facies G. A) Thin section scan showing abundant ooids surrounded by nodular siderite cement and calcite filled fractures in core 16-04-22-15W4, ~552 m. Section is ~ 7.5 X 2.5 centimetres. B) Close-up view of Ooid rich cluster (box) in A. Note the predominance of dissolved ooid voids, with very few intact ooids (arrows). C) Thin section showing ferruginous ooid and concentric layering from ooidal ironstone in A.
(Donaldson et al., 1999) and the Upper Albian Peace River Formation (Taylor et al., 2002), both located in north central Alberta. Both ironstone deposits are interpreted to have formed during times of low sediment accumulation in well oxygenated conditions.

Ferruginous ooid formation is interpreted to have occurred in a suboxic environment just below the sediment-water interface, where iron-rich clays such as berthierine initially started to precipitate around detrital nuclei (e.g. Taylor and Macquaker, 2000). Periodic exhumation and agitation by storms would then oxidize the iron rich ooids to goethite. Repeated exhumation and burial, coupled with agitation on the seafloor, resulted in growth of concentric layers of clay and iron oxides around the nuclei (e.g. Taylor and Curtis, 1995; Heikoop et al., 1996). Where original ooids are not preserved, subsequent reworking of sediment is thought to have dissolved away the original internal structures.

In modern shallow shelf environments where free oxygen is not abundant in porewaters, bacterially-mediated sulphate reduction is the predominant method of oxidizing organic matter, which leads to preferential precipitation of pyrite. In the absence of sulphate reduction however, iron rich minerals such as siderite can precipitate (Macquaker and Taylor, 1996). In Lower Jurassic ooidal ironstone deposits in eastern England, low sedimentation rates are thought to have promoted ironstone formation (Taylor and Macquaker, 2000). Low sedimentation rates resulted in low organic matter preservation, and therefore negligible sulphate reduction took place. Iron reduction processes took place instead, leading to precipitation of siderite (Taylor and Macquaker, 2000).

### 3.4 Facies Succession

The facies present in the study area all represent marine environments, and are grouped into an offshore to lower shoreface succession (Fig. 3.28). A typical facies succession may be several metres thick and is made up of a gradational vertical change in facies that represents lateral movement of the shoreline. The marine facies succession begins with deposition of dark grey to black laminated mudstone of facies A above a flooding surface. A good example is usually observed at the Cardium Formation-Muskiki Member contact. Laminated mudstones of facies A grade up into more bioturbated
Typical marine facies succession for the Muskiki and Marshybank Members: Offshore to lower shoreface facies succession

Figure 3.28 A representative marine facies succession for the study. Facies are labelled to the right of the stratigraphic column. The legend on the far right shows components of the stratigraphy for the whole study area, and can be used as a reference for cross-sections in Chapter 4 as well. FA- Facies A, FB- Facies B, FC- Facies C.
deposits of Facies B. Numerous bentonites are characteristic of facies A, and, to a lesser extent, facies B. The facies succession consists of a progressively coarsening up succession, depicting an overall shallowing marine environment that is progressively more prone to fair and storm wave base sedimentation.

The normal progression of the facies succession within the study interval is sometimes separated by erosional surfaces that may or may not contain siliciclastic pebble lags (Facies D). These erosional unconformities represent remnants of a seafloor that was subject to reworking, a process interpreted to be a result of sea level fluctuations. In outcrop, thicker conglomerates are sometimes preserved at the top of the study interval. These conglomerates probably represent the transgressive reworking of a nearby shoreline. Above facies D usually lie muddier sediments of facies A and B, indicating relative sea level rise and an increase in accommodation.

In the subsurface over most of the study area, ooidal ironstone (facies E) lies erosively on a regional unconformity, the DE1 surface. The ooidal ironstone typically overlies bioturbated silty sandstone of facies C. The ooidal ironstone is interpreted to have been deposited during marine transgression that resulted in deposition of the overlying Dowling Member, and is therefore not included in the above facies succession.

### 3.5 Bentonites in the Muskiki Member

The abundance of bentonites in the Muskiki Member within the study area warrants discussion. Bentonites form by the subaqueous devitrification and chemical alteration of a glassy volcanic tuff or ash (Ross and Shannon, 1926; Altaner et al., 1984). They are the product of diagenesis, usually dominated by smectite clays, and are commonly formed in marine and lacustrine settings. Due to their fine-grained nature, preservation potential is greatly increased under low energy conditions (Straeten, 2004). The majority of bentonites recorded in this study occur in the lower part of the Muskiki Member, above the Turonian-Coniacian boundary, currently dated at 89.65 ± 0.28 Ma (Siewert et al., 2012).

Bentonites within this study vary in thickness, from a few millimetres to 40 cm (Fig. 3.5; 3.6). Fewer bentonites occur in the upper part of the Muskiki Member and the Marshybank Member. Those bentonites are usually a few millimetres to one to two
centimetres thick. One exception is a 12 centimetre pale grey bentonite interbedded within granular sand observed at Sullivan Creek, and is located just below the top of the study interval (Fig. 3.19). The majority of bentonites can be described as either creamy grey/white or rusty orange/yellow in colour. A few of the thinner bentonites are rusty and micaceous. Coarse-grained bentonites are also sometimes observed, and contain biotite crystals readily visible to the naked eye. In coarser facies, due to reworking, bentonites tend to be thinner and may be contaminated with silt and sand.

3.5.1 Possible sources for Muskiki bentonites

The greatest intensity of Cordilleran magmatism occurred between 95-90 Ma, with widespread plutonism recorded in the Coast plutonic and Omineca tectonic belts of the Canadian Cordillera (Fig. 2.1; Armstrong, 1988; Armstrong and Ward, 1993). While volcanoes that occur in association with their respective plutons are not very well known, the vast numbers of bentonites that occur in the Late Cretaceous attest to the presence of volcanic eruptions. This study did not involve any analysis of the composition of bentonites found in the Muskiki and Marshybank members. However, the stratigraphic occurrence, and variety in texture and colour of bentonites suggests different sources for the volcanic ash. The occurrence of these bentonites coincides with episodes of extensive magmatism in the Cordillera. Possible source regions for the bentonites are outlined below.

The Sierra-Nevada batholith, located in California, was magmatically active between 210-85 Ma, and is suggested as a possible source for the bentonites (Bateman, 1992). Magmatism from the batholith is thought to have been present periodically throughout the Mesozoic, but culminated with the highest volume production between 85-100 Ma (Saleeby et al., 2008). Overlapping in time with the Sierra-Nevada batholith is the Idaho batholith, with rocks that date 217-38 Ma, most of which are concentrated in the Cretaceous (Armstrong, 1988). The Cassiar batholith, located in southern Yukon Territory and northern B.C., is the largest plutonic body in the Canadian Cordillera, and was emplaced ~100 Ma (Driver et al., 2000). The batholith is thought to have contributed to widespread magmatism that spanned the middle Cretaceous to Eocene, and is also suggested as a possible course for volcanic ash.
Elder (1988), based on correlations of macrofaunal zones, suggested a northwestern volcanic source for a prominent bentonite bed that covers much of Arizona, Colorado, Kansas, and Utah. The bed is located near the Cenomanian-Turonian Stage boundary. The ash bed is correlated in Canada as the Bighorn River, or “Red” bentonite located within the underlying Blackstone Formation (Tyagi et al., 2007). The source of volcanic activity was attributed to be somewhere near the Canada-U.S. border in Alberta. This was based on bentonite dispersal patterns, although a more specific source was not determined. It is not unreasonable to infer that a volcanic source for this Cenomanian bentonite persisted through the early to middle Coniacian, and perhaps to a lesser degree in the Middle to Late Coniacian. A volcanic source to the west of the study area has been suggested by Fuentes et al. (2011), based on the abundant tuff and plagioclase rich sandstone units deposited in the Late Cretaceous.

By no means do the above ideas offer a final answer to the source of bentonites found throughout the study. At present, there is no hard evidence to advocate one suggestion over the other. It is clear however, that volcanism was active throughout deposition of most of the Muskiki and Marshybank members.
CHAPTER 4

ALLOSTRATIGRAPHY OF THE LOWER WAPIABI FORMATION

4.1 Introduction

An allostratigraphic approach (Section 2.6) was used to examine the internal stratigraphic architecture of the Muskiki Member of the Wapiabi Formation. Correlations were based on tracing marine flooding surfaces, represented in well logs. Flooding surfaces were also correlated between outcrop and drill-core localities, where available. Flooding surfaces were regionally traceable throughout the study area, except for where they lapped out or were erosionally truncated. In this manner, the Muskiki Member can be informally divided into 19 allomembers (Fig. 4.1). One allomember in the overlying Dowling Member is also named. The horizontal datum used in this study is the DE1 surface that marks the base of the Dowling Member. Dividing the geology into smaller units allows localized controls on deposition to be distinguished from those of regional scale, such as eustasy and large-scale plate flexure.

This chapter describes the new allostratigraphic scheme for the Lower Wapiabi Formation. Summary cross-sections, located at the back of the thesis, illustrate the allostratigraphic correlations that form the basis for the interpretations of depositional history. New biostratigraphic results, outlined at the end of this chapter, are integrated with the allostratigraphic framework, and allow for some biostratigraphic horizons to be traced regionally.

4.2 Well Logs: Resistivity and Gamma Ray

710 gamma-ray and resistivity well logs were used to delineate the subsurface stratigraphy of the Muskiki Formation (Fig. 1.7). A gamma ray log records the natural radioactivity of a rock generated by the decay of naturally-occurring uranium, thorium, and potassium (Gluyas and Swarbrick, 2004). Potassium is the most common radioactive element that occurs in sedimentary rocks, while uranium and thorium tend to be adsorbed onto clay minerals, and incorporated into mudrocks (Gluyas and Swarbrick, 2004). The result is that shales tend to give higher gamma ray readings than sandstones. The
Figure 4.1 Summary allostratigraphic diagram of part of the lower Wapiabi Formation in southern Alberta. Allomembers have been assigned letters whereas bounding surfaces have been assigned numbers. Allomembers MCA-MCL have been assigned C letters to denote that they have a clinothem geometry. Dowling allomember DA has also been mapped throughout the study area.
intensity of radiation measured by a log is calibrated in American Petroleum Institute (API) units.

A typical induction well-logging tool measures the conductivity of a rock formation. The inverse of conductivity, or resistivity, is the property usually displayed in a well log, and is measured in units of ohms-m\(^2\)/m or ohm-m for short (Gluyas and Swarbrick, 2004). The resistivity of a rock depends on the nature of its porosity, as well as the type of fluid that may fill the pore spaces. For example, hydrocarbon bearing sandstones will have higher resistivity values than those filled with saline water. As a result of the higher porosity, sandstones tend to have higher resistivity than shales, although this is not always true. High resistivity values in the Muskiki Member, for example, are interpreted to reflect low porosity due to pore-filling calcite cement, which is observed in core and outcrop. Alternatively, the high porosity values may also be explained by fresh water saturating rock units. Gamma ray and resistivity well logs were used together in this study to trace flooding surfaces and identify facies changes across the basin (Fig. 4.2). Both sets of logs were also used qualitatively to make an approximate determination of lithology.

4.3 Method

In order to correlate and map allomembers across the study area, well logs were used to build a grid of 23 working cross-sections. Fifteen working cross-sections were oriented from west to east, as dip lines, and seven cross-sections were oriented north to south, to approximately follow the strike of the basin. Working lines intersected at common wells, allowing for consistency and confidence in correlation. Township lines (E-W) were spaced approximately two townships apart (~20 km), and range lines (N-S) were spaced approximately four ranges apart (~40 km). Difficulties in correlation were encountered in the northeast and southern portions of the study area. Within these areas, additional wells were used to help constrain correlations. Nine cores and 14 outcrop sections were used to calibrate the lithological character of the well log signals (Fig. 1.7). As most of the measured outcrops lie within the Foothills deformed belt, only a few unthrusted wells are available directly adjacent to most outcrops. These unthrusted wells
Figure 4.2 Gamma ray and resistivity well log curves to show the typical well log character observed in the subsurface for the Muskiki Member. Well is located 55 km east of the restored Mill Creek outcrop. Dashed red line shows a qualitative method used to define silty shale-sandy siltstone packages when making sandy-silt maps (Chapter 5). Gamma curve that fell outside the red line was used to provide a measure of the thickness of a package represented by sandier rocks. Mudstones tend to have high gamma ray and low resistivity values, whereas sandstones tend to have lower gamma ray and high resistivity values. L- low, H- high. The curves for all the wells in the summary cross-sections in the thesis are presented in this manner.
were used wherever possible to correlate directly to outcrop, and provide a better link to the subsurface.

### 4.3.1 Summary cross-sections

The locations of the 11 well-log cross-sections that summarize the stratigraphy within the study area are shown in Figure 4.3. The cross-sections, Figures 4.4 through 4.14, provide an approximately chronostratigraphic framework for the Muskiki Member, and demonstrate the variability of the stratigraphy throughout the study area. Seven cross-sections oriented west to east are labelled 1 through 7 (Figs. 4.4-4.10), and four cross-sections oriented north to south are labelled A through D (Figs. 4.11-4.14). Where possible, outcrop sections and all core sections were incorporated directly into the summary cross-sections. These 11 cross-sections are located at the back of the thesis (pocket). In addition, four cross-sections linking Foothills outcrops with the subsurface grid (Fig. 4.11) were constructed, and are summarized in Figures 4.15 through 4.18. Each summary cross-section includes a color-coded, scaled summary diagram that illustrates the subsurface geology in a compressed form, which gives a more dramatic impression of the stratal geometry (Figs. 4.4-4.14).

### 4.4 Allostratigraphy of the Muskiki and Marshybank Formations

A subsurface to outcrop allostratigraphic framework for the Muskiki and Marshybank formations in northwestern Alberta and British Columbia was established by Plint (1990). The southern boundary of that study ended at Township 44. Plint (1990), recognized two regional markers (flooding surfaces) within the Muskiki Formation, M1 and M2, and 12 allomembers, termed A to L in ascending order, within the Marshybank Formation (Fig. 1.6). Stratal markers within the Muskiki Formation, plus allomember A of the Marshybank Formation, were found to be essentially parallel from west to east, with subtle divergence of markers from north to south (Plint, 1990). In contrast, allomembers B to L (with exception of non-marine unit G) comprise upward coarsening successions that downlap onto allomember A, from southwest to northeast (Plint, 1990).

To link the allostratigraphy of Plint (1990) with the present study, a working log correlation line was constructed to follow the stratigraphy from township 44 to the
Figure 4.3 Map of study area showing location of summary well log cross-sections. Present day locations of outcrops shown in blue, palinspastically-restored outcrops in black. Where necessary, cores were projected into existing cross-sections.
Figure 4.15 Well-log cross-section demonstrating tie to Highwood River and Sullivan Creek outcrops. See text for details.
Figure 4.16 Well-log cross-section demonstrating tie to Sheep River outcrop. See text for details.
Figure 4.17 Well-log cross-section demonstrating tie to Millarville outcrop. See text for details.
Figure 4.18 Well-log cross-section demonstrating tie to Oldfort Creek and Horseshoe Dam outcrops. See text for details.
northwestern edge of the present study area in township 27 (Plint, unpublished work). The Marshybank Formation was found to thin southward such that siltstone- and sandstone-dominated successions that are typical of the formation in the north are very thin or absent. Correlations showed that allomembers of the Marshybank Formation were progressively removed southward, from the stratigraphy established in the northwest (Plint, unpublished work). In contrast, the Muskiki Formation’s equivalent to the south, the Muskiki Member, thickens southward. The top of the Muskiki Member in the study area is marked by a regional unconformity that has subtle erosional relief in the subsurface that becomes more obvious westward in the Foothills. Correlation suggests that the top of Unit A of the Marshybank Formation in the north is correlative with the erosion surface that forms the top of the study interval in southern Alberta (the DE1 surface).

In the Foothills of southern Alberta, Stott (1963) designated the Muskiki and Marshybank as members of the Wapiabi Formation. Although Nielsen et al. (2003), through subsurface to outcrop correlations, demonstrate how the Verger and Medicine Hat members are equivalent to stratigraphy in outcrop at Kevin, Montana, they do not provide physical or biostratigraphic evidence to explain how the subsurface stratigraphy was correlated with the Muskiki and Marshybank geology in the Foothills of southern Alberta. The physical stratigraphic relationship between the members therefore remained ambiguous.

The new stratigraphic framework established in the present study is an attempt to resolve the uncertainty concerning correlation between the Foothills and the subsurface of the southern Alberta Plains. Because this correlation is established, the Muskiki and Marshybank terminology is extended into the southern Alberta Plains, and the Verger and Medicine Hat Member terminology is not used.

This study recognizes 19 marine flooding surfaces that were correlated in the subsurface. The correlations are based on surfaces that were followed from northwestern Alberta (Plint, 1990) to the northern edge of this study (Plint, unpublished data). The flooding surfaces define 19 allomembers within the study, and have been assigned informal names, ME1 to ME18, from oldest to youngest (Fig. 4.1). The base of the Muskiki Member, represented by the E7 surface, was also traced throughout the
The top of the Muskiki Member is marked by erosion surface DE1, which marks the base of the Dowling Member. Surface DE2 was also traced throughout the basin, and marks the top of Dowling allomember DA.

The base of the Muskiki Member is marked by a major erosional surface, E7 that marks the top of the Cardium Formation. The E7 surface has been delineated in the study area by J. Shank (Shank and Plint, 2012), based on southward extension of previously established subsurface stratigraphy (e.g. Plint et al., 1986; Plint et al., 1988). The E7 erosion surface is readily identifiable throughout most of the study area, because the top of the Cardium Formation is locally marked by a lowstand deposit, and E7 is usually a burrowed surface mantled by well-rounded extra-formational chert pebbles (e.g. Plint and Walker, 1987). At Castle River, the E7 surface is marked by chert pebbles in an ooidal ironstone that marks the marine transgression of the Muskiki Member.

In the eastern part of the study area, the E7 surface is marked by a mm-scale shell lag with fish teeth and wood debris (Fig. 4.19). Subsurface correlation to Deer Creek, Montana, shows that the E7 surface rests on top of a massive siltstone unit overlain by dark silty mudstone with abundant siderite concretions. This subtle marker separates the Ferdig Member (Cardium Formation equivalent) from the Kevin Member, which in part, is stratigraphically equivalent to the Muskiki Member.

The top of the Muskiki Member is defined by the southward extension of the erosion surface that defines the top of the Marshybank Formation in the north. In the south, this is the DE1 surface that separates overlying siltstone of the Dowling Member from the bioturbated sandy siltstone of the Muskiki Member. Extensive *Thalassinoides* burrows mark the boundary, and ferruginous ooids cemented by a pervasive siderite cement form a thin ooidal ironstone that marks the surface throughout most of the subsurface (Figs. 3.25, 3.26). Well log correlations show that north of Kevin, Montana, the DE1 surface cuts down to merge with the MacGowan Concretionary Bed (bed 100 of Cobban et al., 1976). In Foothills outcrop sections (e.g. Highwood River, Millarville, Mill Creek), the top of the Muskiki Member is represented by a sideritized conglomerate up to 3 metres thick (Figs. 4.15, 4.17).

In cross-section, the Muskiki Member has an overall wedge-shaped geometry that thins from 125 m in the southwest to 25 m in the northeast. Dividing the Muskiki
Figure 4.19 Core image showing the contact between the underlying Cardium Formation and Muskiki Member in the subsurface (white triangle on left). A thin mm-scale shell lag with fish teeth and wood debris is present on the E7 surface (white box). Core 01-24-16-05W4, 483 m. Scale bar subdivisions are 1 cm each.
Member into smaller packages allows for a higher resolution stratigraphic framework to be established, which can then be used to observe and interpret subsidence patterns through time.

During the correlation process, it was noted that markers E7, ME1, ME7, and DE1 were subtle regional bevelling surfaces within the study area, and were used to divide the Muskiki Member into three distinct stratal packages: Lower Unit (E7-ME1), Middle Unit (ME1-ME7), and Upper Unit (ME7-DE1).

4.5 Lower Unit of the Muskiki Member

The Lower Unit of the Muskiki Member is bounded by markers E7 below and ME1 above, and forms a southeast-thinning wedge of mudstone. The package thins from 27 m in the NW to a zero edge 260 km to the SE, where ME1 merges with the E7 surface (Fig. 4.4, 4.5, 4.13, 4.14). The base of the Lower Unit forms the base of the Wapiabi Formation, and the contact is placed where mudstone of the Muskiki Member unconformably overlies the generally sandier or siltier Cardium Formation.

Over most of the study area, ME1 is a regional flooding surface. In the northwest part of the study area however, at Sheep River and Horseshoe Dam, ME1 is correlated to a cm-scale, rippled coarse sandstone bed with rare granules (Fig. 4.16, 4.18). At Highwood River East and Sullivan Creek, ME1 is correlated to cm scale granular and pebbly conglomerates. A bentonite sometimes overlies the ME1 flooding surface in outcrop (e.g. Mill Creek, Fig. 4.5), and cores 14-18-19-18W4 and 06-34-30-08W4 (Fig. 4.8, 4.14).

4.5.1 Internal Stratigraphy

The Lower Unit is composed of allomember MA. In outcrop, allomember MA consists of bioturbated siltstone of facies B, and several, 2-15 cm thick bentonites (e.g. Mill Creek, Fig. 4.5). A prominent erosion surface, mantled by chert and quartz pebbles, consistently appears 3-9 m above the E7 surface at several outcrop localities (e.g. Lynx Creek, Sullivan Creek). At Burnt Timber Creek, four distinct coarsening up successions are present in the basal 9 m of the Lower Unit, each of which is capped by a 2-20 cm thick granular conglomerate (Fig. 4.10).
The allomember can be separated into at least three unnamed stratigraphic units. The top of each unit is defined by a prominent flooding surface that can be traced in well logs with relative ease. The un-named bounding surfaces, represented by dashed lines on cross-sections, are sub-parallel and are progressively truncated towards the southeast by the regional ME1 erosion surface, which merges with surface E7 in the southeast (e.g. Figs. 4.5, 4.7, 4.9).

4.5.2 Preliminary interpretation

The wedge shape of the Lower Unit is probably the result of tectonic loading of the basin immediately northwest of the study area. The relatively parallel geometry of markers within allomember MA suggests that the rates of accommodation and sediment supply were approximately equal. Accommodation was probably filled with sediment as fast as it was created, resulting in an aggradational style of deposition for allomember MA (e.g. Pint, 1990; Varban and Plint, 2008a; Hu and Plint, 2009). Sediment was efficiently distributed across the basin by storm processes, as evidenced by clay aggregates and subtle wave-form structures in mud and silt-stone facies (Chapter 3). Deposition of allomember MA probably took place above storm wave base, across a low relief ramp in water probably no more than ~50-70 m deep (e.g. Plint et al., 2012; Plint, submitted).

4.6 Middle Unit of the Muskiki Member

The Middle Unit of the Muskiki Member is bounded by erosion surfaces ME1 at the base, and ME7 at the top. The Middle Unit forms the thickest package within the study, forming a northeast thinning wedge of sediment that thins from 100 m in the SW to ~10 m in the northeast, over a distance of 300 km (e.g. Fig. 4.4, 4.5, 4.13, and 4.14).

The base of the Middle Unit is defined by the transgressive surface ME1 although, in the southeast, the base is defined by the composite E7/ME1 surface, where ME1 merges with the regional E7 surface. The top of the Middle Unit is represented by the ME7 surface, which is correlated to the top of a coarsening up siltstone unit, overlain by siderite nodules (e.g. Fig. 4.10).
4.6.1 Internal stratigraphy

The middle Muskiki Member is divided into six allomembers, MB to MG, each of which is bounded by a marine flooding surface, ME2 to ME7. The allomembers consist of at least six upward-coarsening packages. However, it is evident that in the southwestern part of the study area, more than six upward-coarsening packages exist.

All surfaces bounding allomembers within the Middle Unit are truncated to the east and south by surface ME7, except where surface ME3 merges with the underlying ME2 surface (e.g. Figs. 4.6, 4.7, 4.13). The ME7 surface, which forms the top of the Middle Unit, has up to 35 m erosional relief (Fig. 4.14). Overall, each allomember is thickest in the southwest, and typically thin to a zero erosional or depositional edge to the northeast.

4.6.2 Preliminary interpretation

Like the Lower Unit of the Muskiki Member, the Middle Unit forms a wedge shape. The southwest thickening of the wedge is probably the result of tectonic loading to the southwest of the study area. This is in contrast to the underlying E7-ME1 package, which shows thickening to the northwest. The shift in sediment accumulation, from the northwest to the southwest, suggests that by middle Muskiki time the region of active loading in the thrust belt had shifted to the southwest.

Similarly to the underlying E7-ME1 package, the lack of clinoform geometry and evidence for deposition above storm wave base suggests that the Middle Unit was also deposited in relatively shallow water (e.g. Plint et al., 2009).

4.7 Upper Unit of the Muskiki Member

The Upper Unit of the Muskiki Member is bounded at the base by marker ME7, and at the top by surface DE1. This package a very different geometry compared to the underlying units. The Upper Unit forms a ‘trough’ shaped package of mudstone that thickens from ~5 to 50 metres from northwest to southeast. The ‘trough’ is bounded to the southwest by a relatively tabular ‘platform’ area that varies from 5 to 20 metres in thickness. The subtle thinning of allomembers across the ‘platform’ is demonstrated well
in Fig. 4.7, between wells 16-23-14-26W4 and 06-06-14-21W4, in Fig. 4.7, between wells 06-07-23-26W4 and 08-30-22-22W4, and in Fig. 4.8, between wells 03-09-18-29W4 and 03-18-18-27W4. West of the platform, and like the underlying Middle Unit, the Upper Unit shows minor thickening to the southwest.

In cross sections, the thickening of the trough becomes apparent where surface ME7 bevels underlying markers towards the eastern and southern portions of the study area. In the following order, figures 4.9, 4.7, and 4.5 show the increase in erosional relief of the ME7 surface towards the south. Figure 4.12 provides an excellent axial view across the trough, and also shows that stratigraphic surfaces within the trough toplap against the DE1 surface, which forms the top of the Muskiki Member.

4.7.1 Internal stratigraphy

The Upper Unit is divided into 12 allomembers, MCA to MCL, each of which is separated by a marine flooding surface. The ‘C’ in the allomember name indicates that the allomember is a clinothem package. As the erosional ME7 surface progressively cuts deeper into underlying sediment toward the southeast, clinothem packages appear to shingle off the top of the DE1 surface (e.g. Fig. 4.6, 4.7, 4.13, 4.14).

Allomember MCA forms the lowermost package of sediment within the Upper Unit. It is bound at the base by the regional ME7 unconformity, which is overlain by a bentonite in places (e.g. Core 06-34-30-08W4 in Fig. 4.14). Unit MCA, and the overlying MCB allomember, are the only two units within the Upper Unit that can be shown to downlap onto the ME7 surface within the study area.

Allomember MCC occurs above allomember MCB, and is also the last package within the Muskiki Member to have substantial thickness in the western Foothills. Within allomember MCC, it becomes apparent that sediment begins to thicken within the ‘trough’ in a northwest to southeast direction, as the package is thickest in the southeast (e.g. Figs. 4.4, 4.13).

Following deposition of allomember MCC, allomembers MCD to MCL begin to shingle off the DE1 surface that marks the top of the Muskiki Member. Allomembers MCD through MCL are thin, 5-10 metre packages of sediment that fill the northwest-southeast trending trough from the northwest to the southeast. Each package
progressively appears further to the southeast, so that the uppermost package, MCL, blankets only the very southeast portion of the study area. Strike cross-sections along ranges 16, 12, and 8 West of 4 illustrate the complex geometry (Figs. 4.12, 4.13, and 4.14). The clinothems appear to fill the trough in an east and south accreting direction (Figures 4.9, 4.7, and 4.5).

4.7.2 Preliminary interpretation

The Upper Unit has a geometry that cannot be attributed to flexure driven by a Cordilleran load. The northwest-southeast wedge of sediment suggests localised linear subsidence that might have occurred above a deeply buried fault or faults. The sheet of thinner sediment that borders the western edge of the trough implies the presence of a hinge line across which flexure has created additional accommodation in the east. This idea is discussed further in Chapter 5.

4.8 Dowling Allomember A

One regional flooding surface in the basal part of the Dowling Member was correlated throughout the study area. Dowling allomember DA lies above the Muskiki Member (and Marshybank Member where present) and forms a wedge thinning to the southeast. Isopachs within the Muskiki Member generally thin toward the northeast, whereas allomember DA thins in a southeasterly direction (e.g. Figs. 4.7, 4.10, 4.14). Dowling allomember DA thins from 12 metres in the northwest to a zero depositional edge ~120 km to the southeast.

4.9 Summary of stratigraphy for the Muskiki and Marshybank members

The overall sedimentary package that makes up the Muskiki and Marshybank members has a wedge-shaped geometry that is thickest in the southwest. Internally, the stratigraphy is compartmentalized into three distinct units that are separated by regional erosion surfaces. The Lower Unit, made up of allomember MA, forms a northwest thickening wedge of sediment, and probably reflects tectonic loading of the lithosphere to the northwest of the study area. The Middle Unit, represented by allomembers MB to MG, is made up of a southwest thickening wedge of sediment, and probably reflects the southward shift in the locus of tectonic activity. The Upper Unit in the Muskiki Member
forms a northwest-southeast trending trough of sediment that is filled by clinothem allomembers MCA to MCL. The surfaces of these clinothem packages progressively appear to shingle off the top of the DE1 surface. The overall geometry of the Upper Unit suggests localized linear subsidence and a more complex depositional history (Chapter 5). A three-dimensional block diagram that illustrates the overall stratigraphy for the Muskiki and Marshybank members is shown in Fig. 4.20.

4.10 Biostratigraphy

A great number of well-preserved Coniacian ammonite and inoceramid fauna were collected from various outcrop sections in the study area. Areas of significant contributions to the new biostratigraphic scheme include outcrops at Sheep River, Lynx Creek, Alberta, and Kevin and Deer Creek, Montana. Better biostratigraphic control available in the south allows for a biostratigraphic zonation to be integrated into the stratigraphy and extended to the northern portions of the basin, where macrofossils are not as abundant. A summary of positively identified samples, along with their locations and relative ages, is shown in Table 4.1. Identification of ammonites and inoceramids was done in collaboration with Dr. Ireneusz Walaszczyk (University of Warsaw, Poland) and Dr. William Cobban (USGS, Denver, USA).

4.10.1 Significance of new results

The consistent appearance of *Cremnoceramus crassus crassus* specimens above the E7 surface, but not below, suggests that the lower part of the Muskiki Member can be assigned to the upper part of the Lower Coniacian (Fig. 1.4; Cobban et al., 2005, 2006). Physical well-log correlations to Deer Creek, Montana, show that the E7 surface lies at the top of a massive siltstone unit. Specimens from siderite nodules 1.3 and 3.3 metres above this massive siltstone reveal specimens belonging to the *Cremnoceramus crassus* lineage, further supporting the Lower Coniacian assignment.

At Kevin, Montana, the middle Coniacian inoceramids *Volviceramus koeneni*, and *Volviceramus involutus* appear between 3 and 28 metres below the MacGowan bed (bed 100 of Cobban et al., 1976; Fig. 4.12). There were no macrofossils found between the highest lower Coniacian fossil collected (*Cremnoceramus crassus*) and the lowest middle Coniacian fossil (*Volviceramus koeneni*), leaving an 18 metre gap where the lower-
Figure 4.20 Summary block diagram for the Muskiki and Marshybank members. The left panel represents the most southern west-east cross-section within the study area (Fig. 4.4) and the right panels represents the most eastern north-south cross-section within the study area (Fig. 4.14). The top panel shows clinoform surfaces ME10-ME18 that toplap against the DE1 surface, which forms the top of the Muskiki and Marshybank members. Overall, the study interval is thickest in the southwest, and thins to the northeast.
<table>
<thead>
<tr>
<th>Outcrop Locality</th>
<th>Macrofossil collected</th>
<th>Height above E7 surface (m)</th>
<th>Relative Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Burnt Timber Creek</td>
<td><em>Scaphites ventricosus</em></td>
<td>64</td>
<td>Middle Coniacian</td>
</tr>
<tr>
<td></td>
<td><em>Volviceramus tenuistratus</em></td>
<td>81</td>
<td>Middle Coniacian</td>
</tr>
<tr>
<td>Deer Creek, MT</td>
<td><em>Cremnoceramus crassus lineage</em></td>
<td>1.3, 3.3</td>
<td>Early Coniacian</td>
</tr>
<tr>
<td>Kevin, MT</td>
<td><em>Cremnoceramus crassus crassus</em></td>
<td>30</td>
<td>Late Early Coniacian</td>
</tr>
<tr>
<td></td>
<td><em>Inoceramus kleini</em></td>
<td>5</td>
<td>Late Early/ Early Middle Coniacian</td>
</tr>
<tr>
<td></td>
<td><em>Volviceramus koeneni</em></td>
<td>77</td>
<td>Early Middle Coniacian</td>
</tr>
<tr>
<td></td>
<td><em>Volviceramus involutus</em></td>
<td>52</td>
<td>Early Middle Coniacian</td>
</tr>
<tr>
<td>Lynx Creek</td>
<td><em>Cremnoceramus crassus deformis</em></td>
<td>11</td>
<td>Early Early Coniacian</td>
</tr>
<tr>
<td></td>
<td><em>Tethyoceramus sp.</em></td>
<td>11</td>
<td>Early Coniacian</td>
</tr>
<tr>
<td>Mill Creek</td>
<td><em>Inoceramus ernsti</em></td>
<td>78</td>
<td>Early Coniacian</td>
</tr>
<tr>
<td>Oldfort Creek</td>
<td><em>Tethyoceramus sp.</em></td>
<td>45</td>
<td>Early Coniacian</td>
</tr>
<tr>
<td></td>
<td><em>Scaphites preventricosus</em></td>
<td>35</td>
<td>Early Coniacian</td>
</tr>
<tr>
<td>Sheep River</td>
<td><em>Sphenoceramus</em></td>
<td>64</td>
<td>Late Coniacian</td>
</tr>
<tr>
<td></td>
<td><em>Volviceramus involutus</em></td>
<td>64-66</td>
<td>Late Middle Coniacian</td>
</tr>
<tr>
<td></td>
<td><em>Scaphites depressus</em></td>
<td>69</td>
<td>Late Coniacian</td>
</tr>
</tbody>
</table>

Table 4.1 Summary of the macrofossils that were collected and positively identified within the study area. Samples were identified by I. Walaszczyk, (USGS; University of Warsaw).
middle Coniacian boundary might be. A lower Coniacian inoceramid was collected at Oldfort Creek, *Tethyoceramus* sp. Projecting the biostratigraphy into the subsurface stratigraphy, the inoceramid lies just above the ME3 surface at Oldfort Creek. Following this surface to Kevin, Montana, ME3 lies 5 metres above the highest Early Coniacian fossil collected there. Based on the subsurface extension of this surface to Kevin, it seems reasonable to reduce the 18 m gap with no fossils down to just 13 m, leaving the lower-middle Coniacian boundary somewhere in that interval.

Upper Coniacian fossils within the study area were only observed at Sheep River in the western Foothills. The Upper Coniacian inoceramid *Sphenoceramus* and the ammonite *Scaphites depressus* were collected 22 and 18 metres, respectively, below the top of the Marshybank Member. This suggests the Upper Coniacian at Sheep River is only represented by 22 metres. At Kevin, Montana, documented for the study area is at the MacGowan Concretionary Bed at the Kevin outcrop, where Cobban et al., (2005) describe the upper Coniacian ammonite *Protexanites*. Below the MacGowan concretionary bed, the inoceramid that marks the base of the Upper Coniacian (*Megadiceramus subquadartus*) is not observed, unlike in localities further south in the Western Interior (Walaszczyk and Cobban, 2006). Above this surface, upper Coniacian *Scaphites depressus* ammonites were collected from siderite nodules in the overlying 3.8 m of mudstone. At 3.8 m above the MacGowan bed, well rounded phosphatic pebbles mark a subtle erosion surface that makes up bed 108 of Cobban et al., 1976. The first Santonian fossil observed is located 6.8 m above bed 108, and appears to be an advanced form of the lower Santonian ammonite *Clioscaphtes saxitonianus*, suggesting an early, but not earliest Santonian age. Based on the erosion surface present at bed 108, and a later form of the Santonian ammonite, it is proposed that the Coniacian-Santonian boundary be placed at bed 108 (Grifi et al., *in revision*). This suggests the Upper Coniacian at Kevin is represented by only 6.8 m of mudstone. Correlation of the surface represented by bed 108 reveals that it cuts down to merge with the MacGowan Concretionary Bed in the subsurface to the north. Between the subsurface and outcrop at Sheep River, the stratigraphy is interpreted to expand such that more Upper Coniacian strata is preserved (Fig. 4.16).
CHAPTER 5
CONTROLS ON SEDIMENTATION AND PALEOGEOGRAPHY OF THE LOWER WAPIABI FORMATION

5.1 Introduction

The following chapter presents sediment isopach maps for allomembers of the Muskiki and Marshybank (where present) members, as well as for the lowermost Dowling Member. The effects of local tectonic events, regional flexural subsidence and eustatic sea-level changes on the geometry of allomembers will be discussed. A paleogeographic interpretation is also provided at the end of this chapter.

The geometry of a chronostratigraphic unit, which in this study is bounded by marine flooding surfaces, is the result of the interaction between geological process that control accommodation through space and time, sediment volume and dispersal rates, and syn- and post- depositional erosion (Allen and Allen, 2005; Miall, 2010). Various subsidence mechanisms generate different and discrete patterns of accommodation. Therefore, the geometry of different packages of sediment can be used to interpret the processes that generated those subsidence patterns.

5.1.1 Isopach and Isolith Maps

Isopach maps for each allomember were generated using the Golden Software program Surfer 8.0, using the default kriging method. The thickness of each allomember for each well was input into a database, which was then used in the Surfer program to generate individual isopach maps for each allomember. In addition, isolith maps were generated for several allomembers that display a ‘dirty’ sandstone character. The latter packages are made up of very fine sand mixed with silt, and therefore contribute to a ‘sandier’ signal in well logs. The isolith maps were generated on a qualitative basis using the gamma ray log (Fig. 4.2). A total of 22 sediment isopach maps and 2 ‘dirty’ sand isolith maps were generated for the Muskiki Member. One isopach map was generated for the lowermost unit in the overlying Dowling Member.

Stratigraphic and sedimentological evidence from this study (Chapter 3) indicates that the Lower and Middle units of the Muskiki Members were deposited across a low relief ramp in water no more than ~50-70 m deep. The near tabular geometry of these
units indicates deposition by storm-driven currents. The stratal geometry therefore suggests that sediment filled available accommodation as soon as it was created.

Stratigraphic and sedimentological evidence from this study (Chapters 3 and 4) suggests that the Upper Unit of the Muskiki Member was also deposited above storm wave base, at least throughout the western portion of the study area. A clinoform geometry however, appears to have developed in the eastern and southern portions of the study area. Sediment dispersal in this part of the basin is more complex, and suggests that accommodation was greater than sediment supply (Plint et al., 2009).

Isopach maps allow recognition of both regional flexure and localized subsidence patterns. The isopach maps presented on the following pages will be discussed in terms of regional subsidence patterns, and localized, Precambrian basement-related structures. Isopach and isolith maps (where present) are discussed for each of the Lower, Middle, and Upper units of the Muskiki Member, as well as the lowermost allomember of the overlying Dowling Member. The data control points from which isopach and isolith maps were constructed are shown on the total isopach maps for the lower Wapiabi Formation (Fig. 5.1).

5.2 Regional Flexural Subsidence

Regional flexural subsidence patterns for the Muskiki Member are best demonstrated at the unit scale. Each of the lower, middle, and upper units of the Muskiki Member provides isopach trends that are unique, and as such, are discussed separately below. The individual allomember isopach maps for the middle unit are also discussed.

5.2.1 Lower Unit

The Lower Unit of the Muskiki Member, represented by allomember MA, forms a wedge-shaped body of mudstone that thins from 27 m in the northwest to a 0 edge to the southeast, over a lateral distance of 260 km (Fig. 5.2). The thickest sediment accumulation in the northwest is divided into two main ‘lobes’, suggesting that generation of accommodation for sediment was not uniform. Isopleths are parallel to an inferred tectonic load in the northwest, except between townships 18 and 22, where they turn abruptly westward, resulting in two areas of thicker sediment accumulation.
Figure 5.1 Isopach map of total thickness of the Muskiki and Marshybank (where present) members. Relevant geography and location of outcrops is also shown. Contours are in 5 m intervals.

Figure 5.2 Isopach map of total thickness of the Lower Unit of the Muskiki Member-allomember MA. Yellow dashed lines show approximate west-east trend of isopleths, which show thinning of sediment over this area. Contours are in 2 m intervals.
5.2.2 Middle Unit

The Middle Unit of the Muskiki Member, represented by allomembers MB to MG, forms a prismatic wedge that thins from 100 m in the southwest to ~10 m in the northeast, over a distance of 300 km (Fig. 5.3). A 90° rotation of the isopleths, from northwest to southwest, is observed between the lower and middle unit. Isopleths on the middle Muskiki isopach map are also inferred to be broadly parallel to the trend of the tectonic load, and show thinning, over an interpreted forebulge approximately 250 km away.

5.2.3 Allomember isopachs in the Middle Unit

Allomembers MB and MC show isopach patterns in which both units thicken to the southwest, in the direction of the tectonic load (Figs. 5.4, 5.5). Allomember MB also shows thicker sediment accumulation toward the east, along the southern part of the study area. The isolith maps of allomembers MB and MC show that silty sandstone accumulated where subsidence was greatest (Figs. 5.6, 5.7). Sedimentological and stratigraphic evidence from the middle Muskiki Member suggests that sand-rich sediment was supplied from the southwest corner of the study area, at least throughout deposition of allomembers MB and MC.

Stratigraphic correlations show subtle thinning of allomembers MD, ME, MF, and MG (Figs. 5.8-5.11) toward the northeast, indicating a loss of accommodation in that direction. However, allomembers are also progressively truncated by overlying surface ME7 toward the northeast and southeast. The isopach maps therefore show a combination of depositional thinning and post-depositional erosion of sediment packages.

5.2.4 Interpretation of the Lower and Middle units

The Lower and Middle units within the Muskiki Member display regional subsidence patterns that can be readily interpreted as a result of static loading by Cordilleran thrust sheets. The 90° rotation in isopleth trend between the Lower and Middle units suggests a shift in the position of active loading within the fold and thrust belt (Fig. 5.12). Deviations from a simple flexural deflection within the lower unit suggest that crustal inhomogeneities also modified the subsidence pattern (Section 5.4.1).
Figure 5.3 Isopach map of total thickness for the Middle Unit of the Muskiki Member. Note the northwest-southeast orientation of isopleths, which are parallel to the interpreted tectonic load to the southwest. Contours are in 5 m intervals.
Figure 5.4 Isopach map of total thickness for Middle Unit allomember MB. Note that the thickest sediment is located in the southwest, however thick sediment is also present in a west-east trend along the southern part of the study area. Contours are in 4 m intervals.

Figure 5.5 Isopach map of total thickness for Middle Unit allomember MC. Note that the thickest sediment occurs in the southern part of the study area, and thins to a zero erosional edge to the east. Contours are in 2 m intervals.
Figure 5.6 Silty sandstone isolith map for Middle Unit allomember MB. Isolith map suggests a sand-rich sediment source to the southwest of the study area. Contours are in 2 m intervals.

Figure 5.7 Silty sandstone isolith map for Middle Unit allomember MC. Isolith map suggests a sand-rich sediment source to the southwest of the study area. Contours are in 2 m intervals.
Figure 5.8 Isopach map of Middle Unit allomember MD. Sediment thins to a zero erosional edge to the east and southeast of the study area. Contours are in 2 m intervals.

Figure 5.9 Isopach map of Middle Unit allomember ME. Sediment thins to a zero erosional edge to the northeast and the southeast. Contours are in 2 m intervals.
Figure 5.10 Isopach map of Middle Unit allomember MF. Sediment is eroded over a small area in the northeast, and over a larger area to the southeast. Contours are in 2 m intervals.

Figure 5.11 Isopach map of Middle Unit allomember MG. Sediment is eroded to a zero edge over a north-south trend in the eastern part of the study area. Contours are in 2 m intervals.
Figure 5.12 Summary block diagrams for the Lower and Middle units of the Muskiki Member. From Grifi et al., in review.

A) The Lower Unit displays a regional subsidence pattern interpreted to be a result of static loading of the lithosphere some distance northwest of the study area. The margin of the flexural moat is inferred in the northern part of the block diagram.

B) The Middle Unit displays a regional subsidence pattern interpreted to be a result of static loading of the lithosphere some distance southwest of the study area. An inferred hinge line is drawn where sediment thins more rapidly (discussed in more detail in Section 5.4.2)
5.2.5 Upper Unit

The Upper Unit of the Muskiki Member occupies a northwest-southeast trending ‘trough’ that thickens from ~5 m in the northwest to 50 m in the southeast (Fig. 5.13). The isopach map for the Upper Unit shows a thin ‘platform’ to the west that bounds the trough, and minor thickening in the southwest of the study area. The trough is filled by mudstone clinothem packages comprising allomembers MCA to MCL, of which allomembers MCC to MCL progressively toplap against the DE1 surface that marks the top of the Muskiki Member (Figs. 5.14-5.24). A subcrop map of surfaces ME10 to ME18 against surface DE1 show broad, irregular arcs concave to the east, suggesting accretion of clinothems in that direction (Fig. 5.13).

5.2.6 Interpretation of the Upper Unit

A relatively minor thickening southwest of the ‘platform area’ in the Upper Unit suggests there was limited Cordilleran activity and associated subsidence southwest of the study area. This however does not explain the northwest-southwest thickening wedge of sediment east of the ‘platform’ area on the isopach map. Interpretation of this pattern is best explained in terms of Precambrian basement control (Section 5.4.3).

5.3 Precambrian basement control on sedimentation

The Precambrian basement in Alberta consists of a mosaic of welded terranes that are composed of different rock types, and therefore produce variable signals on an aeromagnetic anomaly map (Chapter 2). The regional aeromagnetic map of Alberta highlights heterogeneous portions of the underlying Precambrian basement which may embody zones of weakness in the crust that may be susceptible to differential flexure. In order to assess the influence of Precambrian basement structures on subsidence patterns within the Muskiki Member, the regional aeromagnetic map was overlain on most isopach maps. The study area lies above parts of the Medicine Hat Block (MHB) and Loverna Block (LB) terranes, which are separated by the Vulcan Structure. The Vulcan Structure, made up of a high (Matzhiwin)-low (Vulcan) magnetic anomaly and gravity low, forms the suture zone between the MHB and LB (Hope and Eaton, 2002). The Vulcan Structure was found in previous studies to have influenced sedimentation patterns within Early to Late Cretaceous strata (e.g. Zaitlin et al., 2002; Shank and Plint, 2011;
Figure 5.13 Isopach map of total thickness for the Upper Unit of the Muskiki Member. Note the presence of a ‘platform’ that trends northwest-southeast and separates thicker sediment in the southwest from a northwest-southeast trending ‘trough’ to the east of the ‘platform’. Dashed white lines show the toplap of clinothem surfaces ME10-ME18 against the DE1 surface that forms the top of the study interval. Contours are in 4 m intervals.
Figure 5.14 Isopach map of Upper Unit allomember MCA. Sediment thins to a zero limit where allomember MCA is interpreted to downlap on surface ME7. Contours are in 2 m intervals.

Figure 5.15 Isopach map of Upper Unit allomember MCB. Note the thicker sediment accumulation in the northeast and southeast part of the study area. This probably reflects the onset of subsidence of the ‘trough’ and increased accommodation across the area. Contours are in 2 m intervals.
Figure 5.16 Isopach map of Upper Unit allomember MCC. Note the overall thicker sediment accumulation within the northwest-southeast ‘trough’. Contours are in 2 m intervals.

Figure 5.17 Isopach map of Upper Unit allomember MCD. Allomember MCD and succeeding cliniothems only appear in the eastern portion of the study area, filling a northwest-southeast trending ‘trough’. Contours are in 2 m intervals.
Figure 5.18 Isopach map of Upper Unit allomember MCE. Contours are in 2 m intervals.

Figure 5.19 Isopach map of Upper Unit allomember MCF. Contours are in 2 m intervals.
Figure 5.20 Isopach map of Upper Unit allomember MCG. Contours are in 2 m intervals.

Figure 5.21 Isopach map of Upper Unit allomember MCH. Contours are in 2 m intervals.
Figure 5.22 Isopach map of Upper Unit allomember MCI. Contours are in 2 m intervals.

Figure 5.23 Isopach map of Upper Unit allomember MCJ. Contours are in 2 m intervals.
Figure 5.24 Isopach map of Upper Unit allomembers MCK and MCL. Contours are in 2 m intervals.
Chapter 2). Another prominent aeromagnetic low, trending northwest to southeast, is present to the southeast of the Vulcan Structure within the MHB, and has been named the Orion Low (Grifi et al., *in revision*). The Orion Low becomes an important feature in the discussion of the Middle and Upper units of the Muskiki Member isopach maps (Sections 5.4.2 and 5.4.3).

Seismic reflection profiles 29, 30, and 31 of the Southern Alberta Lithospheric Transect (SALT) were also overlain on the regional aeromagnetic map. Along these lines, Lemieux (1999) identified five extensional faults (1-5). Faults 1-4 offset Phanerozoic strata as young as Devonian, and extensional forced folding was observed in overlying as young as Campanian. Fault 5 was found to offset strata as young as the Upper Cretaceous Fish Scales horizon (below the Cardium Formation). The faults identified in the SALT profiles have a regional northwest-southeast strike, which follows the Precambrian basement fabric trend on the aeromagnetic map. Four reflectors (CS1-4) within the basement were interpreted to be major thrust slices (Fig. 2.10; Lemieux et al., 2000). The lower boundary of these thrust slices was interpreted to be a crustal scale ramp that truncated sub-horizontal reflectors to the east (Fig. 2.10; Lemieux et al., 2000). This suggests that structures within the basement influenced the position of these extensional faults at least until the Late Cretaceous. The influence of basement structures and extensional faults on deposition of the Muskiki Member is discussed in sections 5.4.2 and 5.4.3.

### 5.3.1 Lower Unit

Comparison of the Lower Unit isopach map (i.e. allomember MA) with the aeromagnetic anomaly map shows correspondence to the aeromagnetic high that forms part of the Vulcan Structure (Fig. 5.25). North of the negative aeromagnetic zone of the Vulcan Structure, between townships 18 and 22, isopleths curve westward before turning northeastward again. The change in isopleth direction occurs over the positive anomaly of the Vulcan Structure, formerly known as the Matzhiwin high (Hope and Eaton, 2002). The thinning of sediment over this region suggests that subtle flexure of the Vulcan Structure during deposition of the lower unit could have resulted in a slight topographic high over the area, resulting in thinner sediment accumulation, or erosion of already deposited sediment.
Figure 5.25 Isopach of the Lower Unit (allomember MA) of the Muskiki Member overlain on the regional aeromagnetic map. Relevant aeromagnetic structures are labeled on map. Note the thinning of sediment over the Matzhiwin High (deflection of isopleths), shown by dashed red lines. Seismic lines 29, 30, and 31 of the SALT project, as well as faults 1 - 5 interpreted by Lemieux (1999; Lemieux et al., 2000) are also shown. Aeromagnetic values are measured in nanotesla units. Aeromagnetic map courtesy of Mark Pilkington, GSC.
5.3.2 Middle Unit

Comparison of the isopach map of the Middle Muskiki Unit with the regional aeromagnetic map shows that overall, the northwest to southeast trending isopleths correspond well with similar trending anomalies in the southern portion of the area (Fig. 5.26). Thinning of the Middle Unit is observed near the southwest margin of the Orion low, close to interpreted thrust slice CS4 of Lemieux et al. (2000). A change from a positive to a negative magnetic intensity over this area strongly suggests a genetic link between deep crustal structure and sedimentation of the Middle Unit. This southwest margin is interpreted to represent a lithologic contrast in the Precambrian basement, which functioned as a hinge over a weak zone in the basement (Fig. 5.12B; Grifi et al., in revision). It is possible that the 90° rotation in tectonic load from the northwest of the underlying lower unit, to the southwest of the middle unit, exploited the weak zone in the Precambrian basement, and thus promoted flexure across that boundary (c.f. Waschbusch and Royden; Chapter 2). Although subsidence is interpreted to have been a result of tectonic loading in the Cordillera, the presence of weak zones in the Precambrian basement is inferred to have promoted flexure across these boundaries when the tectonic load was oriented such that it exploited the weak zones (Fig. 5.27).

Unlike the overall isopach map for the Middle Unit, allomember scale isopach patterns do not show good correspondence of sediment thickness trends with the regional aeromagnetic map (Fig. 5.28A-C). This suggests that the effects of Precambrian basement structures on overlying stratigraphy are better observed in thicker stratigraphic intervals, and ultimately at larger temporal scales. However, allomembers MB (Fig. 5.28A) and MF (Fig. 5.28C) have closely spaced isopleths that are oriented northwest-southeast, close to similarly striking extensional faults identified by Lemieux (1999). Allomember MB displays rapid thinning of sediment close to faults 1, 2, 3 of Lemieux (1999). Allomember MF displays rapid thinning of sediment east of fault 5. These faults were not interpreted by Lemieux (1999) to have affected stratigraphy during the Coniacian, however it is possible that there may have been subtle displacement of sediment that occurred below the 30 m resolution of the SALT data. It is therefore difficult to prove the reactivation of Precambrian extensional faults during the Coniacian, at least during deposition of the Middle Unit.
Figure 5.26 Isopach of the Middle Unit (allomembers MB to MG) of the Muskiki Member overlain on the regional aeromagnetic map. Relevant aeromagnetic structures labeled on map. Note the closely spaced northwest-southeast trending isopleths west of the Orion Low, indicating rapid thinning of sediment. The thinning of sediment also occurs immediately west of lineament CS4, which is interpreted as a thrust slice in the Precambrian basement (Lemieux et al., 2000). Interpreted crustal thrust ramp of Lemieux et al. (2000), lines 29, 30, and 31 of the SALT project, as well as faults 1 - 5 are also shown (from Lemieux 1999; Lemieux et al., 2000). Aeromagnetic values are measured in nanotesla units.
Figure 5.27 A simple illustration to show preferential flexure over weak zones in the Precambrian basement. A) Static loading of the lithosphere in the northwest results in subsidence being greatest in that direction. Parallel to the northwest direction of loading, the Vulcan Structure (represented by Vulcan Low in above illustration) served as a weak zone in the basement, and promoted bending of the lithosphere in the direction of the tectonic load. The Orion Low is interpreted to have been inactive because the tectonic load was not oriented in a direction to promote flexure across the weak zone.

B) Static loading of the lithosphere in the southwest results in subsidence being greatest in that direction. Parallel to the southwest direction of loading, the Orion Low served as a weak zone in the Precambrian basement, and promoted preferential flexure in the direction of the tectonic load. The Vulcan Low is interpreted to have been inactive because the tectonic load was not oriented in a direction to promote flexure across the weak zone.
Figure 5.28A Isopach maps of Middle Unit allomembers MB and MC overlain on the regional aeromagnetic map. Relevant structures are labelled on map. Aeromagnetic values are measured in nanotesla units. Lines 29, 30, and 31 of the SALT project, as well as faults 1-5 interpreted by Lemieux (1999; Lemieux et al., 2000) are also shown. In allomember MB, thinning of sediment coincides with the location of faults 1-3, suggesting a possible influence of extensional faulting on Muskiki stratigraphy.
Figure 5.28A Isopach maps of Middle Unit allomembers MD and ME overlain on the regional aeromagnetic map. Relevant structures are labelled on map. Aeromagnetic values are measured in nanotesla units. Lines 29, 30, and 31 of the SALT project, as well as faults 1-5 interpreted by Lemieux (1999; Lemieux et al., 2000) are also shown. There is no apparent correspondence between structures and in the Precambrian basement and isopach trends of the allomembers.
Figure 5.28C Isopach maps of Middle Unit MF and MG overlain on the regional aeromagnetic map. Relevant structures are labelled on map. Aeromagnetic values are measured in nanotesla units. Lines 29, 30, and 31 of the SALT project, as well as faults 1 - 5 interpreted by Lemieux (1999; Lemieux et al., 2000) are also shown. In allomember MF, thinning of sediment occurs just east of fault 5, suggesting a possible influence of extensional faulting on Muskiki stratigraphy.
5.3.3 Upper Unit

Comparison of the Upper Unit of the Muskiki Member with the aeromagnetic map shows a strong spatial correspondence between the clinothem-enclosing ‘trough’ and the Orion Low (Fig. 5.29). The initiation of subsidence within the NW-SE ‘trough’ occurs above the Orion Low, and is located east of the inferred hinge zone for the Middle Unit. The Upper Unit has a subsidence pattern that is opposite to that seen in the Middle Unit, with sediment thickest east of the inferred hinge zone (compare Figs. 5.3 and 5.13).

The evidence presented by Lemieux (1999) for Late Cretaceous reactivation of north to northwest-trending faults in the Precambrian basement suggests that extensional faulting could have been active during deposition of the Upper Unit. The crustal scale ramp east of the Orion Low, interpreted by Lemieux et al. (2000), may have been reactivated during the Coniacian, resulting in forced-folding of overlying strata (Grifi et al., in revision), causing subsidence of a linear trough, in which clinothem packages accumulated (Fig. 5.30; Lemieux 1999).

5.3.4 Dowling Allomember DA

The isopach map for the lowermost Dowling allomember DA (Fig. 5.31) shows maximum subsidence in the northwest, resembling the subsidence pattern of the Lower Unit in the Muskiki Member. Allomember DA thins from 12 m in the northwest to a zero depositional wedge 120 km to the southeast. The pattern of subsidence suggests that renewed orogenic activity occurred in the Cordillera to the north-northwest, leading to differential subsidence in that direction. Comparison of the isopach map to the regional aeromagnetic map suggests a spatial correspondence between thinning of sediment and the Vulcan Low (Fig. 5.32). Isopleths appear to thin more rapidly along the trend of the Vulcan Low. Thinning of sediment in this region suggests that the weak zone within the Precambrian basement (represented by the Vulcan Low), may have been reactivated during deposition of the lowermost Dowling allomember.

5.4 Influence of the Sweetgrass Arch

The northwest-striking Kevin-Sunburst Dome and the northeast-plunging Bow Island arch form components of the Sweetgrass Arch in southern Alberta, and transect parts of the study area in the southeast (Fig. 2.14). Previous studies of Cretaceous rocks
Figure 5.29 Isopach of the Upper Unit (allomembers MCA to MCL) of the Muskiki Member overlain on the regional aeromagnetic map. Relevant aeromagnetic structures labeled on map. Note the closely spaced northwest-southeast trending isopleths west of the Orion Low, which delineate a northwest-southeast trending ‘trough’. Sediment thickens considerably just west of lineament CS4, interpreted as a thrust slice in the Precambrian basement (Lemieux et al., 2000). Interpreted crustal thrust ramp of Lemieux et al. (2000), lines 29, 30, and 31 of the SALT project, as well as faults 1 - 5 are also shown (from Lemieux 1999; Lemieux et al., 2000). Aeromagnetic values are measured in nanotesla units.
Figure 5.30 Summary block diagrams for the Upper Unit of the Muskiki Member. An Archean crustal thrust ramp (Lemieux et al., 2000) is interpreted to have been reactivated in the Coniacian, resulting in regional extension of fault blocks within the basement. This is interpreted to have been propagated up into the overlying stratigraphy through forced folding of strata. The clinothem-enclosing ‘trough’ in the Upper Unit is interpreted to have subsided as a direct result of forced folding, creating accommodation for the mudstone clinothems. From Grifi et al., in review.
Figure 5.31 Isopach map of Dowling allomember DA. Contours are in 2 m intervals.

Figure 5.32 Isopach map of Dowling allomember DA overlain on the regional aeromagnetic map. Relevant aeromagnetic structures labeled on map. Note the strong spatial correspondence between thinning of sediment and the Vulcan Low, shown by dashed red lines. Aeromagnetic values are measured in nanotesla units.
(e.g. Schroder-Adams et al., 1997; Payenberg, 2002; Nielsen et al., 2003) have attributed depositional and paleogeographic patterns to subtle uplift of these two structures. The isopach maps presented in this study do not appear to show trends that correspond to either the Kevin-Sunburst Dome or the Bow Island Arch, and therefore these structures appear to have had no influence on Muskiki stratigraphy. In the Middle Unit of the Muskiki Member however, allomembers ME, MF, and MG are eroded by the overlying ME7 surface toward the northeast and southeast of the study area. Although some of the erosion occurs over an area covered by the Bow Island Arch, the erosion of these allomembers does not follow the trend of the northeast-plunging arch, which would be expected if the structure was influencing sedimentation. Therefore, the influence of the Bow Island Arch on the deposition of these allomembers cannot be shown.

5.5 Sedimentation processes in the Upper Unit

The clinothem geometry in the upper Muskiki Member is interpreted to have developed when accommodation rate exceeded sediment supply, which resulted in water depth increase (Plint et al., 2009). The subsidence of the ‘trough’, in which clinothems accumulated, is interpreted to have promoted an increase in accommodation, which would have resulted in newly generated space for sediment to accumulate in. Muddy sediment within the Muskiki Member is interpreted to have been transported by advection across the seafloor, and filled the ‘trough’ once it reached the bathymetric break on the seafloor. Transport of mud by bottom currents is now understood to be capable of depositing resuspended sediment far from the shoreline, as long as wave energy is capable of reaching the sea floor (Schieber and Southard, 2009; Schieber et al., 2020). In this study, clinothem packages MCA and MCB (in part) were observed to lap out onto the ME7 surface. This lap out is observed to occur approximately 250 km seaward of the restored location of the Mill Creek outcrop. Within the study area, the Mill Creek outcrop provides the best evidence of a lower shoreface environment within the Muskiki Member, and is therefore the closest representation of a nearshore environment. The lap out of allomembers MCA and MCB shows that mud was transported for at least 250 km offshore before the seafloor became too deep for wave energy to maintain silt and clay aggregates in suspension.
In contrast to allomembers MCA and MCB, clinothem packages MCC to MCL progressively fill the ‘trough’ in a southeast direction, with packages MCD to MCL only preserved east of the ‘trough’. The mud that formed the clinothem packages was transported across a sediment bypass surface, in a process comparable to that of modern subaqueous deltas, such as the Adriatic (e.g. Cattaneo et al., 2007). Using the modern Adriatic subaqueous delta as an example, mud within the upper Muskiki package is assumed to have been transported across the top of a subaqueous Muskiki delta as suspended bedload. Once sediment reached the bathymetric break on the sea floor (in this study, the western margin of the ‘trough’), sediment moved down the clinoform surface and settled (Fig. 5.33; e.g. Plint, submitted).

The top surface of the Adriatic subaqueous clinoform package ranges from 8 m to 30 m deep and is dependent on wave energy (Cattaneo et al., 2007). The re-suspension of mud by waves prevents mud from building up to sea level, which makes an accurate interpretation of water depth for ancient deposits difficult. The thickness of an ancient clinoform package however, does provide a measure of depositional water depth. In a sediment transport study of the Dunvegan delta system, Plint (submitted) estimated that effective wave base for mud was ~70 m. Below this depth, silt and clay aggregates could not be suspended by waves. The compacted thickness of the upper Muskiki clinoform package, which is up to 50 m thick in the southeast, provides some measure of the water depth in which the clinothems were deposited. Although the water depth for the bypass surface is impossible to identify, a 20 m estimate is suggested, which would give an approximate water depth of 70 m. This number would match the estimated effective wave base for mud presented for the Dunvegan delta, as well as other Cretaceous sediment packages (e.g. Plint et al., 2012; Plint, submitted).

5.6 Anomalously thin sections in outcrop

Three outcrop sections in this study were found to be anomalously thin with respect to the subsurface data and other outcrops (Figs. 4.15, 4.17). Complete outcrop sections (from E7 to DE1) at Millarville, Sullivan Creek, and Highwood River (West) measured 60 to 70 m in thickness, and were thinner than the 120 m thick average at other outcrop localities (e.g. Sheep River, Mill Creek), and the subsurface to the east. These
Figure 5.33 Cartoon illustration to show the lateral transport of mud across the shallow western ‘platform’ and into the deeper, eastern ‘trough’.
It should be noted that the slope of the ‘trough’ is greatly exaggerated to illustrate this idea.
A) The accommodation created by the subsiding ‘trough’ in the upper unit of the Muskiki Member allows mudstone to be deposited in the newly created space.
B) Mud aggregates are transported laterally across a sediment bypass surface. Once mud reaches the bathymetric break on the seafloor, aggregates are deposited downslope into the ‘trough’.
C) Succeeding clinothem packages progressive fill the ‘trough’, and clinoform surfaces toplap against the DE1 surface, which forms the top of the Muskiki Member.
thin outcrop sections are interpreted to have been affected by syn- or immediately post
Coniacian faulting in the area that resulted in erosion of the upper part of the Muskiki
succession. In the Middle and Upper units of the Muskiki Member, allomembers ME to
MCB are interpreted to be absent at Highwood River and Sullivan Creek (Fig. 4.15), and
allomembers MCA and MCB are interpreted to be absent at Millarville (Fig. 4.17). The
allomembers at these outcrops are interpreted to have been deposited on a thrust ramp
that was uplifted and eroded post-deposition of the Muskiki succession (Fig. 5.34). A
more complete Muskiki Member section is present at Sheep River, which is located
further to the West (Fig. 5.1). The thicker and more complete section at Sheep River,
which includes molluscan fossils indicative of the Upper Coniacian, suggests that the
erosion at Highwood River, Millarville, and Sullivan Creek was localized. This is further
supported by the presence of gravel conglomerate that marks the top of the anomalously
thin sections. The conglomerates may represent a nearby shoreline and suggest relative
lowering of sea level in the area that would have promoted reworking of the seafloor
sediment. Faulting that occurred during the Upper Coniacian may therefore have been
responsible for both the anomalously thin stratigraphic sections and also the localization
of a lowstand gravelly shoreline on the up-thrust block.

5.7 Timescale of depositional cyclicity

The Muskiki and Marshybank formations to the northwest of the study area were
interpreted to represent a third order depositional sequence (Plint, 1991). Third order
cycles, better described today in terms of their episodicity, occur on times scales between
0.1 and 10 million years (Miall, 2010; Chapter 2). Regional to local basement movement
is cited as the cause of these cycles however, Milankovitch cycles also operate on time
scales of up to 400,000 years (Miall, 2010).

The regional unconformities at the base and top of the Muskiki Member (and
Marshybank where present), represented by the E7 and DE1 surfaces respectively,
coincide with two major episodes of sea level fall, followed by a sea level rise that
occurred during the Niobrara cyclothem (Caldwell et al., 1993). The Niobrara
transgression that occurred in the middle of the Early Coniacian, represented by the base
of the Muskiki Member, is attributed to eustatic (global sea level) change (Kauffman and
Caldwell, 1993). This suggests that eustasy played a role during deposition of the
Figure 5.34 Series of simple cartoons to illustrate a faulting mechanism that could explain the anomalously thin outcrop sections observed at Highwood River, Millarville, and Sullivan Creek.

1) A site of future thrusting is identified over an area covered by Muskiki stratigraphy.
2a) Uplift of thrust sheet exposes sediment on the tip of the thrust, which is eroded as it becomes elevated.
2b) Continued uplift results in progressively more sediment being eroded.
3) Resulting stratigraphy is such that thin outcrop sections lie on an uplifted, and consequently eroded, thrust sheet.

The Sheep River outcrop, which is interpreted to be stratigraphically more complete, is interpreted to lie on a part of the thrust sheet that was unaffected by erosion, or on a different thrust sheet altogether. The more complete subsurface stratigraphy in well logs are also interpreted to have been unaffected by thrusting.
Muskiki and Marshybank members. Many of the surfaces within this study can be traced along strike for over 350 km and basinwards for 100 to 300 km, suggesting that autocyclic processes alone are insufficient to explain changes in the sedimentation rate. The stratigraphy within the Muskiki Member suggests interplay between subsidence of the foreland basin to the west and eustasy. Flooding surfaces bound upward coarsening successions within the Muskiki Member, and indicate an increase in accommodation in the basin. Upward-coarsening packages suggest small scale cycles of sea level rise and fall, and probably reflect back and forth movement of the shoreline.

5.7.1 Depositional successions within the Muskiki Member

Although 19 allomembers were identified in this study, some allomembers comprise more than one upward-coarsening succession. Allomember MA that makes up the lower Muskiki Member for example, consists of at least two upward-coarsening successions (e.g. Fig. 4.10, Burnt Timber Creek outcrop and well 04-04-32-08W5). Within the middle Muskiki member, at least four upward-coarsening packages can be recognized within allomember MB, and two within allomember MC (e.g. Fig. 4.6, Mill Creek outcrop and well 10-13-05-1W5). Including the allomembers above with allomembers MD to MCL, this study recognized at least 24 upward-coarsening successions. To calculate the approximate duration of each succession, the time in which the whole study interval was deposited must be known. The geochronological control is poor however, due to the scarcity of radiometric age dates.

5.7.2 Biostratigraphy and Geochronology

From a biostratigraphic point of view, the entire Muskiki succession lies between the base of the *Cremnoceramus crassus crassus* zone (late Early Coniacian), and the top of the *Scaphites depressus* zone (Late Coniacian; Chapters 1, 4). Over much of the study area, the *Scaphites depressus* zone is very thin or missing. This indicates that the whole Muskiki Member represents part of the Lower Coniacian, all of the Middle Coniacian, and little to none of the Upper Coniacian. A bentonite dated by Nielsen et al. (2003), taken 1.8 m above the E7 surface in core 13-20-17-7W4, gave an age of 89.4 ± 0.31 myr. A slightly older age of 89.5 myr is therefore tentatively assigned to the E7 surface in an attempt to constrain the age for the base of the Muskiki Member (Grifi et al., *in revision*).
The top of the Coniacian is dated at 86.35 ± 0.11 (Siewert et al., 2012). Based on orbital cycles, Locklair and Sageman (2008) assigned the Upper Coniacian *Scaphites depressus* zone a duration of 1.6 myr. Because most of the Upper Coniacian strata are missing within this study, the duration of the Upper Coniacian is not considered in calculations of the duration of the Muskiki Member. An age of 87.9 is therefore assigned to the top of the Middle Coniacian. This very tentative geochronology suggests that all of the Muskiki Member in the study area was deposited over ~1.6 myr, from about 89.5 Ma at the E7 surface (upper Lower Coniacian) to about 87.9 Ma at the top of the Muskiki Member (top of Middle Coniacian).

Although the duration of each upward-coarsening succession is not very well constrained this study assumes that each succession represents an approximately equal period of time. Assuming a duration of 1.6 myr for the Muskiki Member and 24 upward coarsening packages, an average duration of ~67,000 years is tentatively assigned to each package. The duration of each package falls within the Milankovitch band. Glacioeustacy is the principal mechanism responsible for generating eustatic cycles at the Milankovitch scale (Miall, 2010). Although the Cretaceous period is traditionally viewed as ice-free, Miller et al. (2005) proposed that ice-sheets must have existed for at least geologically short intervals to explain rapid changes in relative sea level observed for the Cretaceous. Although evidence for ice-sheets has only been shown for the Early Cretaceous period (e.g. Frakes and Francis, 1988; Alley and Francis, 2003), it seems possible that ice sheets may have existed during the Coniacian.

### 5.8 Paleogeography

The interpreted paleogeography of the lower Wapiabi Formation is summarized in Figure 5.35. Sedimentological and stratigraphic data presented in this thesis shows that most of the sediment deposited in the study area is made up of mud deposited in an offshore setting. In the southwest portion of the study area, well log signatures for a few allomembers indicate a higher sand content, suggesting that a sand-rich sediment source lay to the southwest.

The presence of coccoliths in core from the northeast part of the study area shows that coccolithophore algae also flourished. Preservation potential for pelagic sediment is much higher away from land, where sediment can dilute pelagic sediment (e.g. HüNeke
Figure 5.35 Paleogeographic map for the Muskiki and Marshybank members in southern Alberta. Rose diagrams showing ripple or gutter cast measurements are also shown beside their corresponding outcrop section.

GC- Gutter Cast, RC- Ripple Crest trend, CFR- preferred migration direction of Combined Flow Ripples.
and Mulder, 2011). The landward limit of this pelagic sedimentation cannot be precisely delineated in this study due to lack of core. However, the boundary of the coccolith-rich sediment is tentatively drawn to coincide with the landward limit of core data, to follow a trend that approximately parallels the strike of the basin (Fig. 5.35).

Paleocurrent data from gutter cast and ripple crest measurements indicate that sediment transport occurred mainly shore-parallel (Fig. 5.35). Combined flow ripples measured at Sheep River and Sullivan Creek indicate preferential sand transport occurred to the northeast, indicating paleocurrents flowed shore perpendicular. At Burnt Timber Creek, the combined flow ripples indicate preferential sand transport was to the southeast, and indicate a shore-parallel flow of paleocurrents.
6.1 Conclusions

The results of this study of the Muskiki and Marshybank members of the Wapiabi Formation can be summarized as follows:

1. The study incorporated 710 well logs, 14 outcrops, and 8 cores to establish a high-resolution subsurface to outcrop allostratigraphic framework for the Muskiki and Marshybank members of the Wapiabi Formation in southern Alberta. The Muskiki and Marshybank members form a wedge-shaped body of strata that thins from ~150 metres in the southwest to less than 30 m to the northeast. Subsurface correlations, supported by biostratigraphic data, suggest that the Marshybank Member is very thin, if not absent over most of the subsurface in the study area. The stratigraphic interval is therefore dominated by rocks of the Muskiki Member.

2. The Muskiki Member is divided into Lower, Middle, and Upper Muskiki units, separated by four regional erosion surfaces. The two lower surfaces, E7 and ME1, as well as the upper surface, DE1, are mantled by extra-basinal pebbles that indicate subaerial exposure of the sea floor. Surface ME7 is not mantled by a siliciclastic lag, but does have erosional relief of up to 30 m in the subsurface.

3. Nineteen allomembers can be recognized and correlated within the Muskiki Member. The Lower Unit, bounded by surfaces E7 and ME1 at the base and top, is made up of allomember MA. The Middle Unit, bound by surfaces ME1 and ME7 at the base and top, comprises six allomembers, MB to MG. The Upper Unit, bound by surfaces ME7 and DE1 at the base and top, is made up of 12 clinothem allomembers, MCA to MCL.

4. Five facies are recognized within this study interval, all of which represent shallow marine wave-influenced environments. The five facies recognized are: dark laminated mudstone, sideritic bioturbated silty mudstone, heterolithic facies, conglomerate and lag deposits, and ooidal ironstone. The interpreted marine facies succession depicts and overall sandier upward succession that was more
prone to fair- and storm- wave base sedimentation. Petrographic analysis of mudstone shows that clay formed silt-size aggregates that were transported across a low-relief ramp as bedload.

5. The Lower and Middle units of the Muskiki Member form wedges that thicken to the northwest and southwest, respectively, and record flexural subsidence that occurred in response to Cordilleran thrust sheet loading. The $90^\circ$ anticlockwise rotation of the pattern of subsidence reflects a southward shift in the locus of tectonic activity in the orogen.

6. The Upper Unit of the Muskiki Member consists of a northwest to southeast trending ‘trough’ that thickens to at least 50 m in the southeast. The trough encloses eastward-accreting mudstone clinothems that resemble a subaqueous delta. The clinothem packages toplap against the DE1 surface, which forms the top of the Muskiki member.

7. Isopach maps of various components of the Muskiki Member, when compared to the regional aeromagnetic map, reveal subsidence patterns that correspond spatially with aeromagnetic anomalies in the Precambrian basement. The Lower Unit within the Muskiki Member shows slight thickening of sediment above the Matzhiwin High, which has been reassigned to the Vulcan Structure (Hope and Eaton, 2002). The Middle Unit within the Muskiki Member shows rapid thinning of sediment over the southwest margin of the Orion Low. These magnetic anomalies represented by the Vulcan Structure and the Orion Low are interpreted to represent weak zones in the Precambrian basement that behaved as flexural hinges. These flexural hinges are interpreted to have been exploited when tectonic loading within the Cordillera occurred parallel to the zone of weakness, promoting flexure across these boundaries.

8. The trough that encloses the clinothems forming the Upper Unit corresponds spatially with the Orion Low, which is bounded to the northeast by an interpreted Archean crustal thrust ramp. The ‘trough’ is interpreted to have developed when Precambrian basement thrust faults were reactivated as normal faults during the Coniacian, which resulted in forced folding of overlying strata, forming a linear northwest-southeast trending ‘trough’. The bathymetric space created was filled
by mud (mainly in the form of aggregates) that was transported laterally by storm-driven geostrophic currents. The mud accreted southeastward from the eastern margin of the trough, forming large clinothems.

9. Although the study area embraces both the Kevin-Sunburst Dome and the Bow Island arch, neither of these features appear to have had an influence on sedimentation patterns within Coniacian Muskiki strata.

10. Subsurface correlations to outcrop at Kevin, Montana, show that the Muskiki Member is equivalent to beds 1 to 108 of Cobban et al. (1976) of the Kevin Member of the Marias River Shale Formation in northwestern Montana. This establishes a firm relationship between the two units, in geographic and stratigraphic contexts.

11. New macrofossil collections, identified through collaboration with Dr. I. Walaszczyk (USGS; University of Warsaw) and Dr. W.A. Cobban (USGS), provide a better understanding of the timescale represented by Muskiki stratigraphy. Collections of inoceramid bivalves and ammonites show that the Muskiki Member embraces part of the Lower Coniacian, all of the Middle Coniacian, and little, to none of the Upper Coniacian. The distribution of inoceramid species suggests that the Marshybank Member is represented by less than 7 metres of rock at Kevin Montana. The studied interval thins northward into the subsurface of southern Alberta, leaving the Marshybank Member virtually nonexistent. Only at Sheep River, in the far western Foothills, is a significant thickness (22 m) of Upper Coniacian strata preserved.

12. A duration of ~1.6 my is tentatively assigned to the Muskiki Member. This incorporates data from available geochronology and biostratigraphy. An average ~67,000 year duration is calculated for each allomember within the Muskiki Member; this periodicity falls within the Milankovitch band. The regional extent of allomembers across the study area coupled with a high-frequency depositional cyclicity suggests that the allomember boundaries record eustatic changes.

13. A paleogeographic interpretation of the study area shows that the Muskiki Member is dominated by marine mudstone that was deposited on a storm influenced shallow marine ramp. A significant proportion of coccolith carbonate
is present ~200 km seaward of an interpreted shoreface environment. In the southwest sandier facies suggest the presence of a sediment source to the southwest.

6.2 Suggestions for future work

1. The subsurface data control for this thesis ends at Township 27, just north of Calgary. The subsurface and outcrop work of Plint (1990) ends at Township 44. The area between townships 27 and 44 is currently under investigation by E. Hooper (M.Sc in progress). An integrated study that will incorporate subsurface stratigraphic correlations from all three studies will aid in a basin-wide reconstruction of the depositional history of the Muskiki and Marshybank members. The results of this integration will enable a complete subsurface to outcrop study that covers an area from northwestern Montana to the northeastern Foothills of B.C.

2. In this study, only clinothem packages MCA and MCB were observed to downlap on the underlying ME7 surface, and clinothem packages MCC to MCL appeared to continue further to the east. Additional subsurface correlations to the east of the study area are needed in order to map the lateral and stratigraphic extent of these clinothem packages.

3. Bentonite samples were collected from outcrops at Mill Creek, Drywood Creek, and Kevin, Montana. Many of the bentonites can be traced into the subsurface stratigraphy with relative confidence. A few samples from this study, taken at Kevin, Montana, are part of a collaborative geochronology project run by Dr. Brad Singer at the University of Wisconsin. Results from this collaborative project will allow better constraints on the duration of some allomembers, and thus enable more accurate interpretation of the depositional cyclicity of these packages.

4. A compositional study on bentonites collected in this study was not done. The variability in texture and colour of samples collected suggests multiple ash sources. There are currently no specific volcanic sources identified for the Coniacian. A geochemical analysis of bentonites collected, similar to studies done
for the Marias River Formation (e.g. Altaner et al., 1984) may aid in future interpretation and identification of possible volcanic sources that were present during deposition of the Muskiki Member.

5. Preliminary research suggests the presence of Late Cretaceous outcrops equivalent to the Muskiki Member in the disturbed belt of northwestern Montana (e.g. Rice and Cobban, 1977). The integration of such outcrops with the subsurface and outcrop stratigraphy established to the east (Kevin and Deer Creek localities), and to the north in southern Alberta, will result in a better paleogeographic reconstruction. Although there was no evidence for a Muskiki shoreline in the study area, well logs in southern Alberta and Montana suggest the presence of a sandy sediment source to the southwest. The outcrops located in the fold and thrust belt of northwestern Montana may provide further information on the sedimentology of the Muskiki Member.

6. Preliminary work incorporating airborne and ground magnetic data with stratigraphy from this study was done by Dr. Bill Morris at McMaster University. The preliminary maps show rectilinear patterns within the study area that represent probable lineaments (and faults) within the Precambrian basement. Further application of this method and comparison with isopach maps generated for the Muskiki Member may help identify areas of faulting that may explain some of the more complex sedimentation patterns observed in subsurface.
REFERENCES


Figure 4.4 Summary west-east cross-section across Township 1. A scale representation of the stratigraphy is included to the right. See text for details.
Figure 4.5 Summary west-east cross-section across Township 6. A scale representation of the stratigraphy is included to the right. See Text for details.
Figure 4.6 Summary west-east cross-section across Township 10. A scale representation of the stratigraphy is included to the right. See Text for details.
Figure 4.7 Summary west-east cross-section across Township 14. A scale representation of the stratigraphy is included to the right. See Text for details.
Figure 4.8 Summary west-east cross-section across Township 18. A scale representation of the stratigraphy is included to the right. See Text for details.
Figure 4.9 Summary west-east cross-section across Township 22. A scale representation of the stratigraphy is included to the right. See Text for details.
Figure 4.10 Summary west-east cross-section across Township 26.
A scale representation of the stratigraphy is included to the right.
Cross-section turns northwards after 03-35-26-29W4.
See Text for details.
Figure 4.11 Summary north-south cross-section between Ranges 28 and 24. A scale representation of the stratigraphy is included below. See text for details.
Figure 4.12 Summary north-south cross-section across Range 16. A scale representation of the stratigraphy is also included. See text for details.
Figure 4.13 Summary north-south cross-section across Range 12. A scale representation of the stratigraphy is also included. See text for details.

North

South

Cardium Formation

Figure 4.13 Summary north-south cross-section across Range 12. A scale representation of the stratigraphy is also included. See text for details.
Figure 4.14 Summary North-South cross-section across Range 8. A scale representation of the stratigraphy is included to the left. See text for details.
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