University of Alberta

Paleogeography and sedimentology of the MacKenzie Basin, Northwest Territories, Canada: An evaluation of Devonian sea-level change, paleoecological controls on Paleozoic reef growth, and early diagenetic conditions.

by

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ABSTRACT

The MacKenzie Basin, located in the District of MacKenzie in the southern part of the Northwest Territories, Canada, includes a thick succession of Middle Devonian strata. This basin, bordered to the east by the Canadian Shield and to the south by the Tathlina Uplift, was directly connected to the open ocean that lay to the northwest. Comprehensive facies analyses of the Chinchaga Formation, Lonely Bay Formation, Horn Plateau Formation, and Horn River Formation, which formed in this basin during the Early and Middle Devonian, shows that sedimentation was largely controlled by eustatic sea level changes. Accordingly, these strata reflect a long period of sea level rise during which shallow water evaporite deposition in the Eifelian was followed by open marine conditions that led to reef growth in the Givetian, and ultimately pelagic shale deposition in the Frasnian.

The Horn Plateau Formation is comprised of numerous isolated reefs that are located along northeast-southwest direction over a distance of 350 km along the MacKenzie Basin ramp. Reefs in the southwest are dominated by stromatoporoids whereas those in the northeast are dominated by corals. Although difficult to prove, it appears that the distribution of the stromatoporoids and corals may have been controlled by nutrients coming from coastal upwelling or runoff from the exposed Canadian Shield.

Effects of early diagenetic processes were evident on the MacKenzie Basin ramp in an intensely bioturbated facies in the Lonely Bay Formation. Burrows from this facies are dolomite-filled further down the ramp and calcite-filled proximal to the Canadian Shield in the east. Anoxic conditions and the presence of sulphate reducing bacteria may have promoted early dolomite formation in the burrows located in deeper water. Burrows further up the ramp were oxygenated and show evidence of input from the exposed Canadian Shield, both of which may have inhibited low-temperature dolomite formation.

ACKNOWLEDGEMENT

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It has been a long road to reach this point in my career and I could not have faced all the challenges associated with completing a PhD without the support of my family and friends. The friendships I have made at the University

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CHAPTER 1: INTRODUCTION

Canada's largest reservoirs for crude oil and natural gas are found in Devonian strata. The Canadian Association of Petroleum Producers (CAPP) published initial estimated reserves for Canada in 2009 that indicated Devonian rocks hold 37.4% of crude oil reserves and 29.6% of raw natural gas reserves (CAPP Report 2010), which is the highest percentage for any geological age. The economic interest in Devonian strata has resulted in a vast amount of high quality data from outcrop and wells (Weissenberger and Potma 2001), particularly in the Western Canadian Sedimentary Basin (WCSB). The Northwest Territories (NWT) is largely underdeveloped with regards to hydrocarbon exploration and does not, therefore, have as large a database as Alberta, Saskatchewan, and Northern British Columbia. Despite the sparse coverage in the southern NWT, drill cores and outcrop data from this area have shown that the Devonian strata are extremely well preserved and largely undeformed throughout the area. Devonian rocks in this area, called the District of MacKenzie, were part of the MacKenzie Basin, sometimes referred to as the Horn River Basin (Figure 1-1; Oldale *et al.* 1994). This basin was bordered by exposed Canadian Shield Rocks in the east and by a Precambrian high to the south, called the Tathlina Arch (Belyea 1972). During the Early and Middle Devonian, the open ocean was encroaching on Western Canada from the northwest and the MacKenzie Basin was directly connected to the encroaching open ocean throughout its exsistence. The excellent preservation of the Devonian strata in the MacKenzie Basin allowed for interpretation (Johnson *et al.* 1985; Corlett and Jones *in press*) of the relative sea level history in the area based on facies associations. Unlike most other Devonian basins in Western Canada, the MacKenzie Basin was continuously connected to the open ocean from the time of its inception in the Eifelian. The MacKenzie Basin strata record an almost continuous sea-level rise throughout the Early and Middle Devonian.

The WCSB did not always have direct access to open ocean waters due to the Tathlina Uplift and the development of a barrier complex at the Alberta and NWT border. As a result, the central and northern parts of the WCSB experienced evaporitic drawdown when thick evaporites accumulated in the basin. (Muskeg Formation and Prarie Evaporite Formation; Oldale and Munday 1994). During the Middle Givetian, tectonic uplift and clastic shedding resulted in a period of exposure and a halt in the carbonate factory (Drees 1993; Wendte and Uyeno 2005). This exposure was seen in several formations in the northern part of the WCSB, such as the Keg River Formation, the Presqu'ile Formation and the Sulphur Point Formation (Drees 1993). Meanwhile, basinal carbonates continued to form in the MacKenzie Basin throughout the Middle Devonian, including reefs in the Horn Plateau Formation.

The Givetian Horn Plateau Formation reefs are aligned along a southwest-northeast trend that stretches over 350 km on a carbonate ramp in the MacKenzie Basin. The isolated reefs are 52-113 m high and the biological composition of the reefs changes along the ramp from stromatoporoid-dominated in the distal part of the ramp to coral-dominated, proximal to the shoreline. These contrasting reefs offer a unique opportunity to evaluate ecological controls on Devonian reef growth such as temperature, oxygen levels, and especially nutrients, an important factor whose significance has not been fully determined (Wood 1993; Kershaw 1998; Hallock 2003; MacNeil 2008). Other reef complexes in the NWT include the Kee Scarp Formation in Norman Wells (Figure 1-1), built mostly by stromatoporoids (Hladil 1989), and the Alexandra Formation reefs near Hay River (Figure 1-1), with one reef complex composed mostly of stromatoporoids and another with a microbial framework (MacNeil 2006). With no established method to directly measure nutrient levels as a possible control on reef growth, MacNeil (2008) used modern analogs to infer that increased nutrient levels and

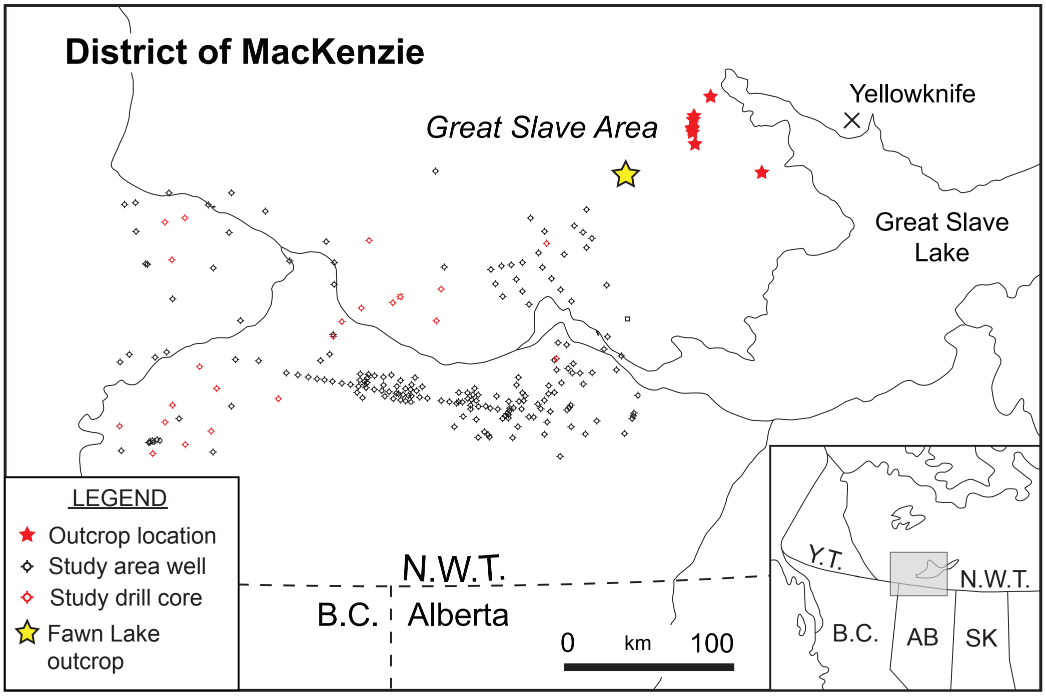

Figure 1-1: Location map showing electronic wells, drill cores, and outcrops examined in thesis study area.

mesotrophic conditions may have caused microbial-stromatoporoid reef growth and oligiotrophic conditions were prevalent during stromatoporoid reef growth. In the MacKenzie Basin, the paleogeographic setting with its connection to the open ocean, ramp geometry, and the exposed Canadian Shield rocks to the east, may have influenced stratification of nutrients, which in turn affected the development of the reefs.

The Lonely Bay Formation, a thickly bedded limestone that underlies the Horn Plateau Formation reefs, contains a very distinctive facies of intensely bioturbated carbonate mud. The burrows on the MacKenzie Basin ramp are filled with dolomite, further out in the basin and calcite in the more proximal position, closer to the exposed Canadian Shield rocks. In early studies (Abed and Scheider 1980; Gunatilaka *et al.* 1987) of dolomitized carbonate burrows it was suggested that the increased permeability and porosity associated with the burrows facilitated fluid flow through the rocks, resulting in dolomitization of burrow material. More recently (Gingras *et al.* 2004; Rameil 2008), studies of dolomitized burrows have recognized some geochemical markers that are associated with decaying marine organic matter, anoxic conditions, and sulphate reducing bacteria. The distribution of the calcite-filled versus dolomite-filled burrows on the MacKenzie Basin ramp suggests that even early diagenetic conditions may have been influenced by paleogeography of the MacKenzie Basin.

With the MacKenzie Basin directly connected to the open ocean, sea level changes in this basin may be representative of Devonian eustasy. The MacKenzie Basin strata represent an almost continuous large-scale sea level rise on a gently dipping Devonian carbonate ramp. The distribution of contrasting reef types and burrowed facies on the ramp present an opportunity to improve our understanding of ecological controls on Paleozoic reef growth and early low-temperature diagenesis in burrowed carbonate rocks.

Study Area and Methods

The MacKenzie Basin is located in the southern Northwest Territories (NWT), Canada between 60º to 64º N and 112º to 120º W (Figure 1-1; Norris 1965; Drees 1993), in an area known as the District of MacKenzie. Although most of these Devonian strata are found in the subsurface, there are exposures in the Great Slave Plains (Figure 1-1; 61º15’ to 62º35’ N and 115º15’ to 117º15’ W), which is located to the west of Great Slave lake. Field work for this project took place in September 2008 when 127 samples were collected from eleven outcrops that were accessed from Highway 93 and one exposure investigated by helicopter (Figures. 1 and 2). Each exposure was described, measured using a Jacob’s staff, and representative samples were collected from every facies.

The subsurface strata of the MacKenzie Basin strata were assessed from drill core (n = 29), thin section (n = 327), and electronic well logs (n = 208). The drill cores used in this study are stored in the core repository at the Geological

Survey of Canada (Calgary), with the exception of five drill cores given to Dr. Brian Jones, which are now held at the University of Alberta. Detailed facies analysis of the drill cores and correlation of electronic well logs was used to establish the regional stratigraphy of the basin and interpretation of depositional environments.

Several types of geochemical analyses were conducted as part of this research including stable isotopes, rare earth elements, and major and minor trace elements. Approximately 170 samples were analyzed for stable isotopes ($\delta^{13}C_{(PDB)}$ and $\delta^{18}O_{(PDB)}$) and 22 samples for trace elements and rare earth elements. All of the analyses were carried out in the stable isotope, electron microprobe, and mass spectrometer laboratories at the University of Alberta.

Previous Research in the District of MacKenzie

Geological research began in the District of MacKenzie with the first exploration of the western arm of Great Slave Lake in 1789 by Alexander MacKenzie (Norris 1965). The first official geological map of the Great Slave Area (Figure 1-1) by Isbister (1855) showed Precambrian basement overlain by Silurian and Devonian sedimentary rocks. Exploration continued in the Great Slave Lake area (Figure 1-1) and a report published by McConnell (1891) documented a traverse from Fort Providence to Rae Point. Cameron (1918) divided the Devonian strata into Pine Point limestones, Presqu'ile dolostones, and Slave Point limestones and completed a comprehensive summary of the Paleozoic geology of the Great Slave Lake area (1922; Norris 1965).

Petroleum exploration of the area began in 1957 when the Geological Survey of Canada instigated "Operation MacKenzie" (Drees 1993). Belyea and McLaren (1962) and Belyea and Norris (1962) described Paleozoic stratigraphy in the subsurface from borehole data. Norris (1965) later provided an overview of

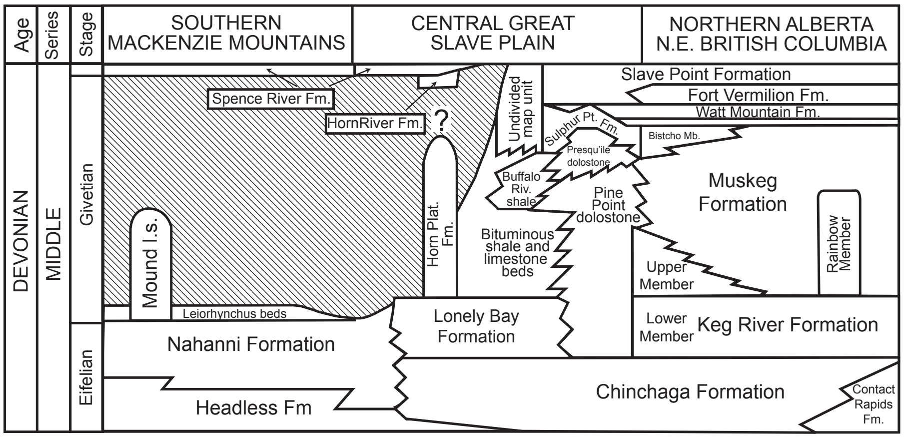

Figure 1-2: Stratigraphy of the Great Slave Area and surrounding areas (adapted from Drees 1993).

this research.

During “Operation MacKenzie”, a helicopter reconnaissance revealed a reefal limestone near Fawn Lake in the Great Slave Lake area (Figure 1-1) that was later described by McLaren and Norris (1964). The exposure at Fawn Lake is the only known exposure of this reefal limestone, which is part of the Horn Plateau Formation. The Pan American Petroleum Corporation drilled five boreholes in the Fawn Lake reef in 1968. Vopni and Lerbeckmo (1969; 1972a; 1972b) subsequently outlined the nature of the Horn Plateau Formation based on the Fawn Lake exposure and the five cores. Eventually it became apparent that there were several of these reefs in the subsurface in the District of MacKenzie (Fuller and Pollock 1972; Drees 1993). Excellent preservation of fossils and lack of dolomite in the Horn Plateau Formation has led to numerous paleontological reports on the tabulate and rugose corals in the reefs (Pedder 1986; Pedder and Babcock 1986; Pedder and Babcock 1986). Interest in possible petroleum resources in the southern Territories peaked again in the 1970s. The Geological Survey of Canada ran another study from 1971 to 1977 to assess the potential of petroleum resources in the Northwest Territories. This study resulted in comprehensive and in-depth reports from Drees (1989; 1993) that incorporated electronic well logs and drill cores from the District of MacKenzie and the Great

Bear Plains (Figure 1-1).

Stratigraphy of the MacKenzie Basin

The Paleozoic stratigraphy in the southern NWT is extremely complex and establishment of high-resolution stratigraphy is complicated by a lack of well control and exposures that are “widely scattered and relatively thin” (Norris 1965). The Paleozoic strata in the Great Slave Plains are part of a wide-ranging homoclinal succession that trends northwest and dips slightly (variable up to a maximum of 10°) to the southwest (Norris 1965). The Precambrian basement in this area underwent pre-Paleozoic extensional tectonic stress, which resulted in horst and graben structures (Morrell *et al*. 1995). The Precambrian highs were onlapped by Paleozoic sediment. Cambrian and Ordovician are generally found as valley-fill in the basement lows.

This study focused on four Devonian formations found in the eastern and central parts of the MacKenzie Basin: the Chinchaga Formation, the Lonely Bay Formation, the Horn Plateau Formation, and the Horn River Formation (Figures 1 and 2). Deposition of the Early to Middle Devonian sediments in this basin was controlled largely by: (1) antecedent topography, (2) progressive isostatic subsidence, and (3) a continuous eustatic rise in sea level (Belyea 1971; House 1975; Drees 1993).

When relative sea level in the area started to rise in the Lower Devonian, evaporites, carbonates and minor clastics were deposited in a proximal setting, fringing the Canadian Shield (Johnson 1971, Morrell *et al.* 1995) and east of the Tathlina Uplift (Figure 1-1). This basal unit in the MacKenzie Basin, which rests directly on the Precambrian Shield, is the Chinchaga Formation (Figure 1-2). It is laterally extensive and composed mostly of evaporite and dolomite facies that are representative of semi-restricted, shallow water environments (Corlett and Jones in press). West of the Tathina Uplift, sediments were deposited in the Root Basin and the Willow-Lake Embayment (Drees 1993). Lower Devonian formations found in the Root Basin include the Tsetso Formation, Camsell Formation, Cadillac Formation, the red beds of the Mirage Point Formation, Fort Norman Formation, and Sombre Formation (Drees 1993).

Sea level rise and carbonate deposition continued into the Late Eifelian (Figure 1-2). The Headless Formation, Hume Formation, Nahanni Formation, and Lonely Bay Formation represent shallow to open marine conditions in the MacKenzie Basin (Drees 1993). The Lonely Bay Formation, which directly overlies the Chinchaga Formation, contains numerous fossils and intensely bioturbated lime mudstones that are consistent with open marine conditions (Corlett and Jones in press).

Increased rates of sea level rise during the Givetian initiated reef growth in the MacKenzie basin (Drees 1993). These isolated reefs, known collectively as the Horn Plateau Formation, overlie the Lonely Bay Formation (Figure 1-2). The Horn Plateau Formation reefs, spread over 350 km of the MacKenzie Basin ramp (Figure 1-1), are composed mainly of stromatoporoids on the distal part of the ramp and corals on the proximal part of the ramp, close to the Canadian Shield (Corlett and Jones in press).

Finally, continued sea level rise during Late Devonian led to deep water conditions and extensive shale accumulation (Gabrielse 1967). The Horn River Formation, Fort Simpson Formation, Canol Formation, and Hare Indian Formation are all composed mainly of organic-rich black shales. The Horn River Formation encased the Horn Plateau Formation reefs. These shales can be correlated to several Late Devonian transgressive cycles (Hallam 1984; House 1975; Johnson *et al*. 1985; Sandberg *et al*. 1988; Ziegler and Sandberg 1990).

Objectives

Due to an apparent lack of petroleum resources in the southern NWT, this region has remained largely unexplored, especially when compared to the age-equivalent strata in Alberta. Existing studies from the District of MacKenzie have either focused on the regional Paleozoic geology of the NWT (Belyea and Norris 1962; Norris 1965; Drees 1989; Drees 1993) or on detailed paleontological studies of the Horn Plateau Formation (Vopni and Lerbeckmo 1969; 1972a; 1972b; Pedder 1989). This is the first basin scale study of the area. Examination of the MacKenzie Basin revealed differences in the paleogeography and the resulting depositional history between this basin and the well-studied WCSB that lies to the south. The paleogeography has greatly influenced the stratigraphy of the MacKenzie basin, the faunal composition of reefs in the basin, and even early diagenetic regimes. With somewhat limited well coverage in the area, conventional sedimentological analyses of the Middle Devonian MacKenzie Basin strata was not sufficient. In order to determine the paleoenvironmental conditions and how they affected carbonate sedimentation and diagenesis, geochemical analyses were combined with traditional sedimentological methods. This approach has managed to shed new light on some persistent questions regarding paleoecological controls on Devonian reef growth and early low-temperature formation of burrow dolomites.

This thesis on the MacKenzie Basin Devonian strata is "paper based". Although each paper has a different focus, they are collectively linked to the paloegeography evolution and depositional history of the MacKenzie Basin. The papers are incorporated into this thesis as Chapters 2 through 4.

Chapter 2 – This chapter introduces the Early to Middle Devonian formations that represent the history of relative sea level change on the MacKenzie Basin ramp, through facies analysis and association. These findings are compared to eustatic sea level curves to demonstrate that global sea level change does not necessarily affect adjacent basins in the same manner.

Chapter 3 – This study examined the paleoecological controls on Devonian stromatoporoid-dominated reef growth versus coral-dominated reef growth in the Horn Plateau Formation.

Chapter 4 – This section of the thesis examines the geochemical and sedimentological differences in early diagenetic controls on some dolomite-filled versus calcite-filled burrows in the Lonely Bay Formation.

Chapter 5 – This chapter summarizes the conclusions of the entire thesis.

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CHAPTER 2: THE INFLUENCE OF PALEOGEOGRAPHY ON RELATIVE SEA LEVEL CHANGE IN DEVONIAN EPICONTINENTAL SEAS[1]

Introduction

Sea level rise during the Early and Middle Devonian led to the formation of many large sedimentary basins throughout the world. In North America, for example, the Western Canadian Sedimentary Basin (WCSB) formed in the west and the Williston Basin, Michigan Basin, Illinois Basin, and Appalachian Basin developed in the east (Bond and Kominz, 1991). Eustatic sea level rise during the Devonian initiated a seaway through the Northwest Territories and into western Canada. This incursion initiated clastic and carbonate deposition throughout the WCSB, a depocenter that today encompasses eastern British Columbia, Alberta, and western Saskatchewan. To the north, the Root Basin, the Willow Lake Embayment (Drees, 1993, 1994; Figure 2-1), and the MacKenzie Basin (Hunt, 1954) developed in the southern part of the Northwest Territories (NWT).

This paper provides an overview of the Devonian succession that developed in the MacKenzie Basin in the Great Slave Lake area (Figure 2-1) and demonstrates that these strata evolved from a eustatic sea-level rise that was intimately linked to the open ocean that lay to the north. This study is framed against the widely used Devonian eustatic sea-level curves that Lenz (1982) and Johnson et al. (1985) developed from their analyses of various successions in North America and Europe (Figure 2-2). In particular, this study determines if global sea-level controlled sedimentation in the MacKenzie Basin in the same manner as in the adjacent Western Canadian Sedimentary Basin (WCSB). This also has implications for worldwide Devonian eustatic changes. A key element of

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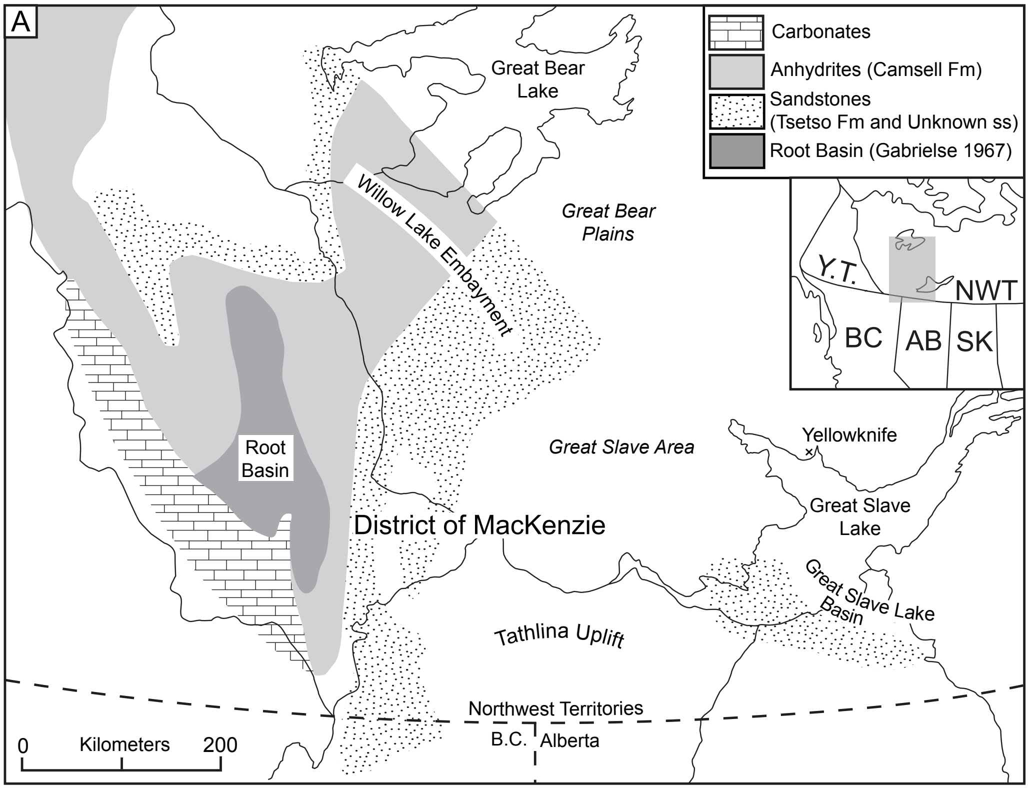

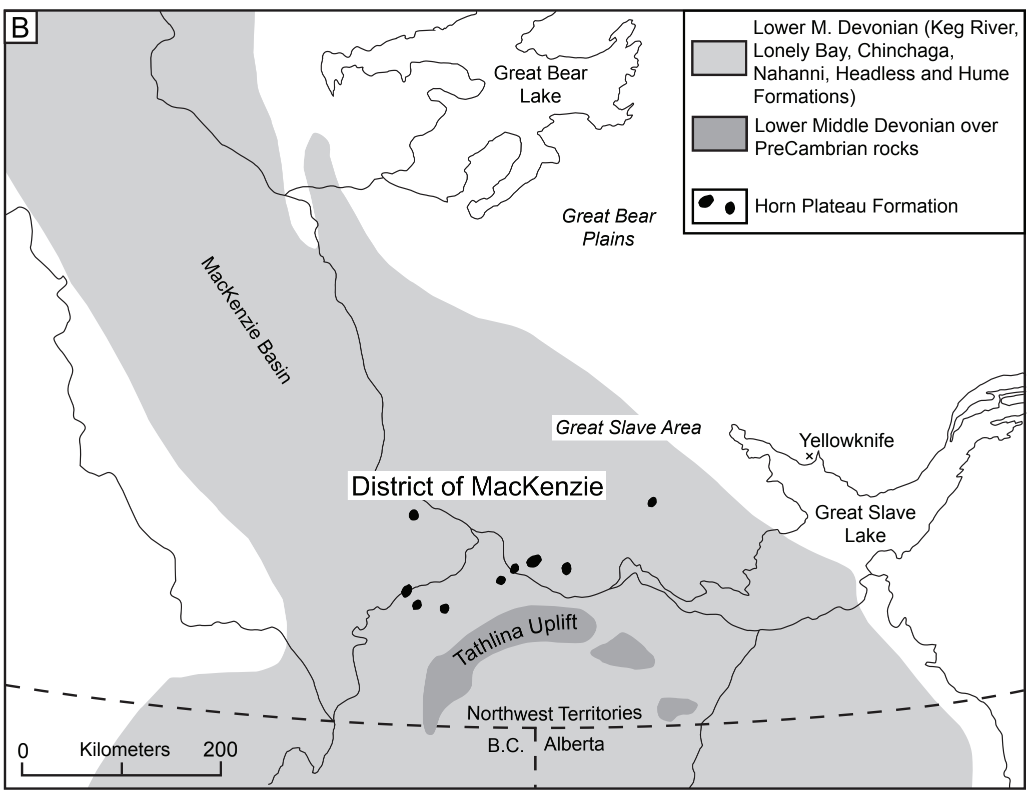

Figure 2-1. Location map of southern NWT, Canada. Labeled areas are referred to in this paper (adapted from Drees 1993). (A) Map showing the distribution of Lower Devonian rocks in the southern NWT. (B) Map of Lower to Middle Devonian strata in the southern NWT.

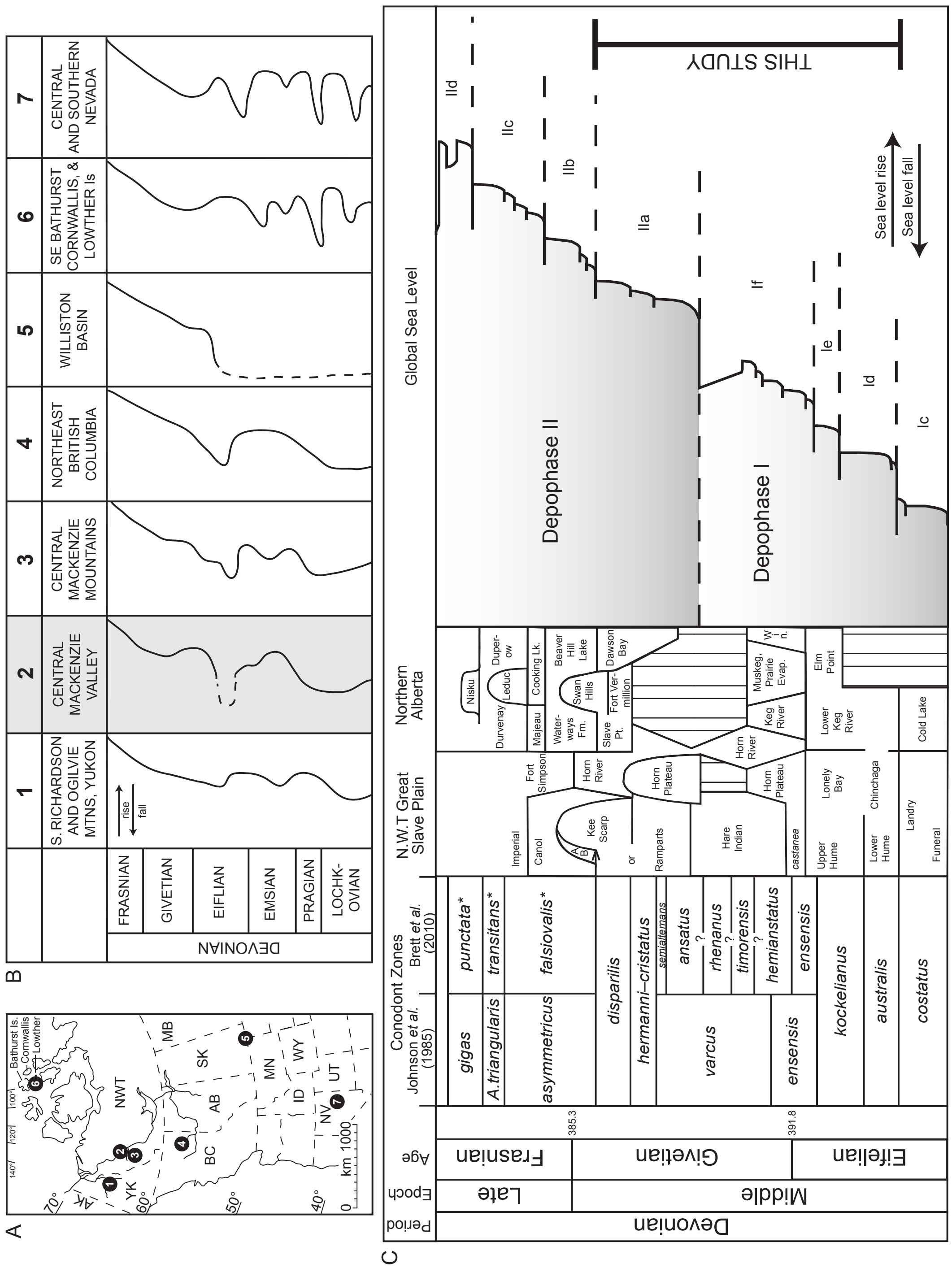

Figure 2-2. Devonian sea level studies (Lenz 1982; Johnson 1984). (A) Location map of sea level curve study (adapted from Lenz 1982). (B) Sea level curves (adapted from Lenz 1982). Note: shading indicates the area that is adjacent and therefore most relevant to this study. (C) Devonian sea level curve and conodont zones. Stratigraphy of the southern Northwest Territories and northern Alberta (adapted from Johnson et al. 1985). Note: time interval of this study is indicated on the right.

both curves is the Mid-Givetian regression (upper part of *T-R cycle If*) that was based largely on successions in the WCSB and the Central MacKenzie Basin. Johnson et al. (1985) noted that this regressive event is not evident in successions in Europe, where deposition appears to have continued without interruption, and suggested that there is a need to look for other sequences in Europe that support a depositional hiatus. This, however, assumes that the regression affected all Devonian basins in the same manner. Herein, evidence is presented that shows that in North America the Mid-Givetian regression is a localized phenomenon experienced only in the interior basins, and not directly related to eustatic sea level changes.

Implicitly, we aim to test the central hypothesis of Johnson et al. (1985) that “Devonian sea-level fluctuations occurred and were of a magnitude to affect sedimentation simultaneously in disjunct regions that had different rates and patterns of subsidence and uplift”. Facies analyses of the strata in the MacKenzie Basin and comparison with relative sea level changes in the WCSB provides an excellent example of how eustatic changes may have different impacts on two adjacent basins; the MacKenzie Basin, which had consistent contact with the open ocean and the semi-restricted Western Canadian Sedimentary Basin.

Nomenclature

Geographic names in the study area are perplexing and confusing. In this study, the following definitions are used.

District of MacKenzie – The area between the MacKenzie Mountains and Great Slave Lake in the southern Northwest Territories, between 60º to 64º latitude and 112º to 120º longitude (Drees, 1993; Figure 2-1).

Great Slave Area – The area west of Great Slave Lake and north of the Tathlina Uplift (Drees, 1993; Figure 2-1).

Great Bear Plains – The area north of Great Slave Lake and south of Great Bear Lake (Drees, 1993; Figure 2-1).

Tathlina Uplift – Located in the southern NWT (Figure 2-1), the Tathlina Uplift is one of the major highs in the Great Slave Area. Subaerially exposed until the Late Devonian, it is onlapped by Devonian strata (Belyea, 1971). The Tathlina Uplift was a barrier between the Western Canadian Sedimentary Basin to the south and the Root Basin and MacKenzie Basin to the north.

Root Basin – The area west of the Tathlina Uplift and east of the MacKenzie mountains (Morrow and Cook, 1987; Drees, 1993; Figure 2-1A).

Willow-Lake Embayment – This embayment was located on the northern flank of the Tathlina Uplift during the Early to Middle Devonian (Drees, 1993; Figure 2-1).

MacKenzie Basin – Hunt (1954) used this name to describe the area of deposition north of the Tathlina High and west of the northern arm of Great Slave Lake. This area has also been referred to as the Horn River Basin (Oldale and Munday, 1994) and the MacKenzie Shelf (Drees, 1993; Nadjiwon et al., 2000).

Methods

Facies analysis of the Middle Devonian Chinchaga Formation, Lonely Bay Formation, Horn Plateau Formation, and Horn River Formation, is based on outcrop, core from 29 wells, and 225 thin sections. Field outcrops of the Chinchaga Formation and Lonely Bay Formation, accessible west of Great Slave Lake (Figure 2-3), were examined in September 2008. Wells and outcrops are located in the study area between 60°05’W-62°45’W and 122°05’-116°00’N (Figure 2-2).

The four formations were examined in 29 drill cores (4 to 10 cm in diameter) that are 7 to 88 m long. Facies analysis was carried out with specific attention

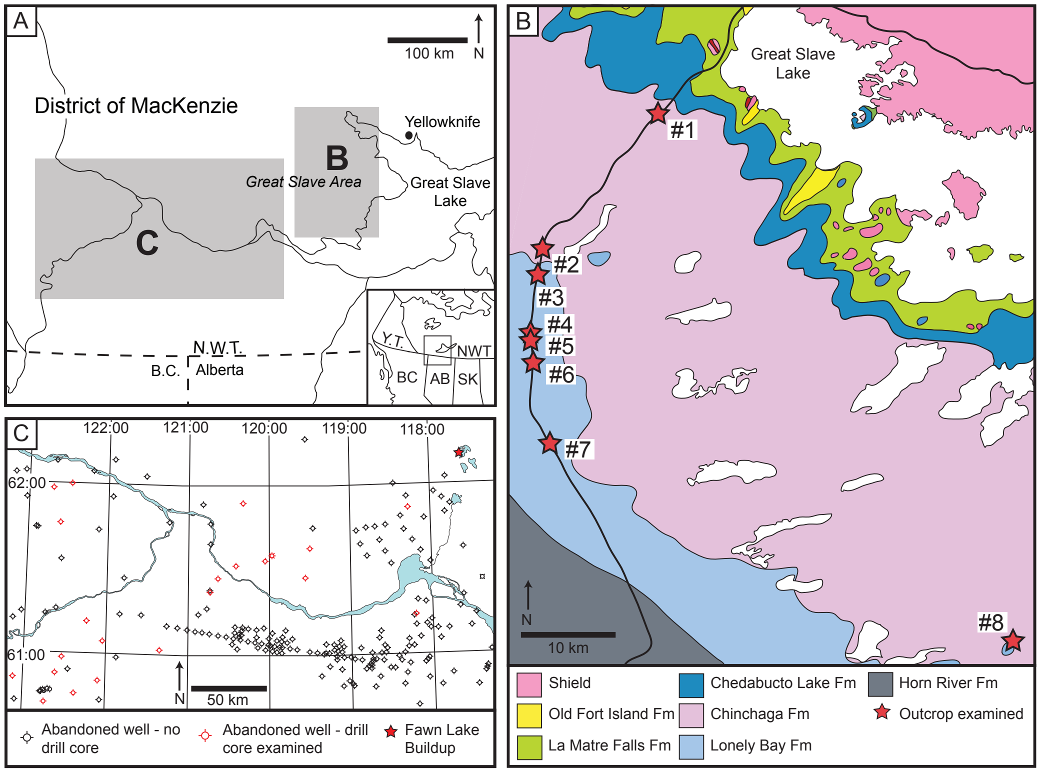

Figure 2-3. Study area. (A) Location map, southern NWT. (B) Outcrops examined in this study. (C) Wells examined in this study. Red wells indicate available drill cores that were examined in this study.

to depositional texture, fossil assemblages, and diagenetic alteration. Fifteen of the 29 drill cores contained the Horn Plateau Formation, including 5 cores (4 cm diameter) that were drilled at Fawn Lake, the only known exposure of the formation. Several (~5) of the other wells were drilled in the early 1970s as part of "Operation Reef", a co-operative exploration project undertaken by Horn River Resources Ltd. The wells were located to penetrate "reefs", rooted on top of the Lonely Bay Formation that had been detected in a series of seismic lines shot in the area. Unfortunately, all attempts to obtain the seismic data proved unsuccessful. The only core that provides full coverage of the Horn Plateau Formation is core #3, drilled by Amoco Canada Ltd. in the Fawn Lake "reef". All other cores come from the upper parts of the "reefs".

Electronic well logs from 208 wells, located throughout the study area were used to correlate the formations and hence determine their lateral extent

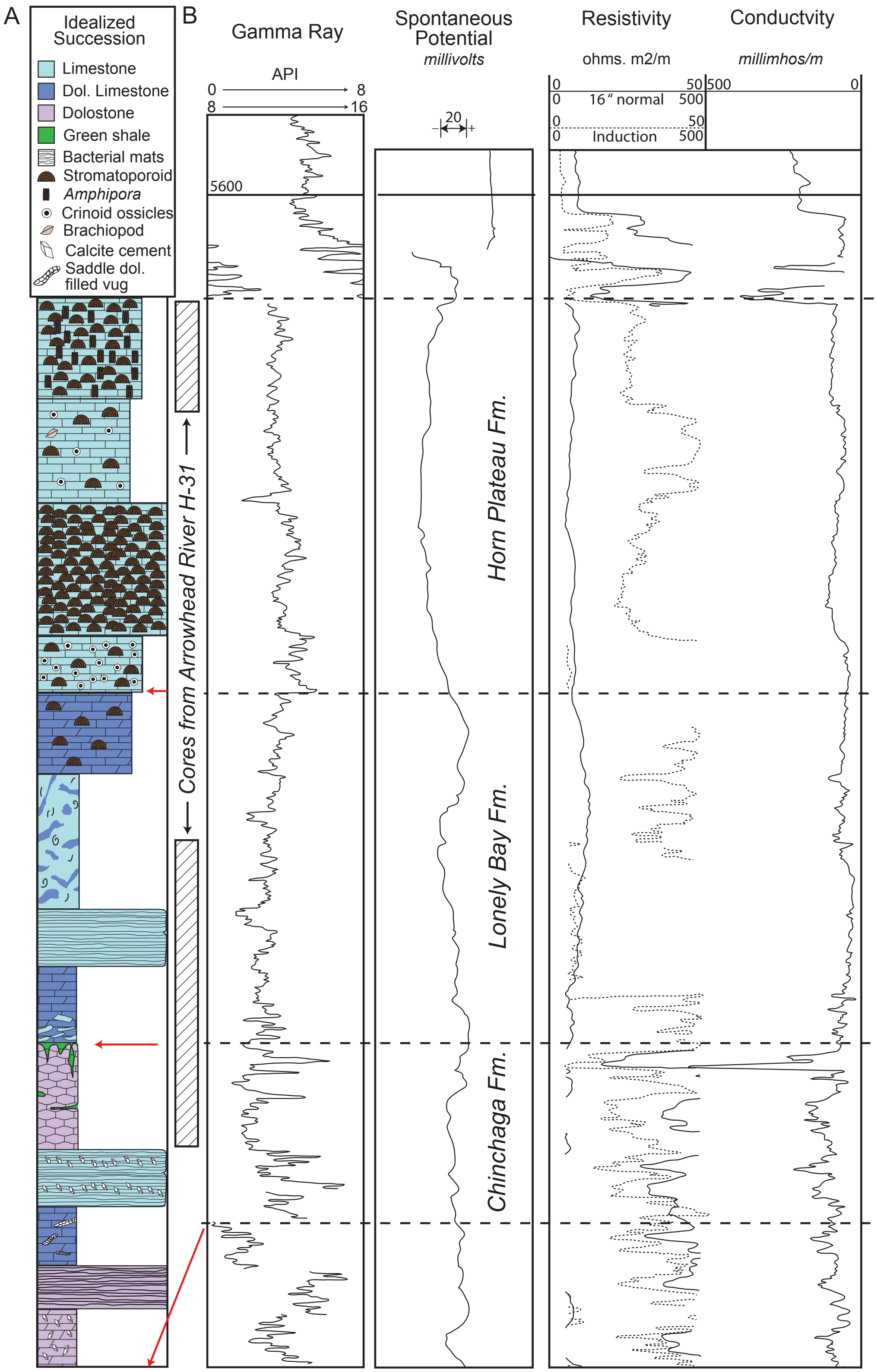

Figure 2-4. Well: Arrowhead River H-31. (A) Idealized stratigraphic column representing facies in drill core. (B) Electronic well logs.

and thicknesses (Figure 2-3). Due to the similarities between the Chinchaga Formation, Lonely Bay Formation, and the Horn Plateau Formation (Figure 2-4), the formation tops were picked using a combination of electronic logs, core descriptions, and descriptions of well cuttings. The geographically extensive Chinchaga Formation is recognized on electronic well logs because it contains evaporites and some shale (Figure 2-4). The uppermost facies in the Chinchaga Formation contains green shale, which helps distinguish it from the overlying Lonely Bay Formation. The Lonely Bay Formation and the Horn Plateau Formation have very similar profiles on the gamma ray log (Figure 2-4). The two formations can be easily distinguished from each other in core, due to a major change in fossil content. The contact between the Horn Plateau Formation and the Horn River Formation is easily recognized because the Horn River Formation is formed largely of shale. The thickness of the Horn River Formation shale was determined using sample descriptions in the well files because its upper contact is with the Fort Simpson Formation shale.

Results - Strata of the MacKenzie Basin

The stratigraphy in the MacKenzie Basin is perplexing and some intervals that contain age equivalent strata are given different names based on when and where they were first described. In this section of the paper each formation is introduced and their constituent facies described. Each facies is given an alphanumeric code (e.g. C-1), where the letter denotes the formation name (Table 2-1). Facies analyses are linked to bathymetry, a method used in many studies of sea-level change (Johnson et al., 1985; Lenz, 1982; Hladil, 1986; Hallam, 1999).

Chinchaga Formation

The Chinchaga Formation, first defined by Law (1955), overlies the

Table 2-1. Summary of major facies from the Chinchaga Fm., the Lonely Bay Fm., the Horn Plateau Fm., and the Horn River Fm. * Depositional environments are interpreted based on lithology, biota, facies succession and are compared with past examples of similar environments.

Facies: Depositional Texture	Description (e.g. allochems, matrix etc.)	Key Features (e.g. biota, porosity etc.)	Depositional Environments*
C-1: Solution Breccia	Dark grey limestone matrix, buff beige/orange angular dolomite clasts	Brecciated clasts found within breccias, multiple collapse features	Intertidal/Tidal Flat, possible periodic exposure (Middleton 1961; Flügel 2004)
C-2: Dolostone with anhydrite	Fine to cryptocrystalline orange/beige dolostone	Coarse dark grey to black anhydrite crystals, minor authigenic quartz	Tidal flat, hypersaline (Flügel 2004)
C-3: Interbedded dolomite and gypsum	Brown/beige dolomite, white slender gypsum crystals	Chickenwire texture	Tidal flat (Scholle 1983; Warren 2006; Flügel 2004)
C-4: Laminated dolomite/limestone	Finely crystalline beige dolomite, fine-grained grey-brown limestone	Millimeter-scale laminations, fenestral porosity	Tidal flat, intertidal (Scholle *et al.* 1983; Walker and James 1992)
C-5: Vuggy limestone/ dolomite	Fine grained beige limestone, finely crystalline beige dolomite	Centimeter-scale vugs lined cm-sized calcite crystals, disturbed bedding (teepee)	Tidalq/mud flat (Assereto and Kendall 1977, Schoelle 1983, Flügel 2004)
C-6: Partially dolomitized mudstone and green shale	Grey partially dolomitized pseudolaminated mudstone	Karst surface, cavities filled in with green shale	Inter/supratidal (Scholle *et al.* 1983)
LB-1: Dolomitized intraclast mudstone	Dark grey matrix, dark grey/ brown clasts	Subrounded clasts, small shelly fossils in clasts	Inter/supratidal (Scholle *et al.* 1983; Flügel 2004)
LB-2: Peloidal partially dolomitized wacke-grainstone	Brown limestone, varying mud peloid content	Fossils and fossil lags: brachiopods, trilobite fragments, ostracods, corals	Inner ramp (Burchette and Wright 1992; Flügel 2004)
LB-3: Brach/Coral/ Stromatoporoid Floatstone	Fine grained mud matrix	Dominant fossils: corals, brachiopods, stromatoporoids; fully marine	Mid-inner ramp (Machel and Hunter 1994, Döring and Kazmierczak 2001)
LB-4: Bioturbated mudstone	Fine-grained brown limestone matrix; black fine-med partially dolomitized infill	Dolomite- and calcite-filled burrows surrounded by fossiliferous wackestone	Middle ramp (Flügel 2004)
HP-1a: Crinoid coral floatstone	Beige crinoid wacke-packstone matrix	Fossils: corals, crinoids, brachiopods, molluscs; fossils not *in situ*	Backreef (Machel and Hunter 1984, Shen and Zhang 1997, Flügel 2004)
HP-2a: Coral bafflestone	Beige mud-wackestone matrix, some small crinoid ossicles in matrix	Fossils: tabulate and rugose corals, minor brachiopods; fossils *in situ*	Reef-front (Machel and Hunter 1984, Hladil 1986, Shen and Zhang 1997)
HP-3a: Crinoid coral rudstone	Beige crinoid wacke-packstone	Fossils: crinoids, tabulate and rugose corals, brachiopods, stromatoporoid fragments	Reef-crest/slope (Machel and Hunter 1984, Shen and Zhang 1997)
HP-4b: Stromatoporoid floatstone	Beige/dark grey matrix	Fossils: Stromatoporoids, brachiopods, crinoids, trilobites, ostracods, rugose corals	Back-reef (Wood 2000; MacNeil and Jones 2006)
HP-5b: Stromatoporoid framestone	Beige/grey mudstone matrix	Mainly stromatoporoids, *in situ*, geopetal cement shows way up	Reef-front (Scholle 1983; MacNeil and Jones 2006)
HP-6b: Coral brachiopod floatstone	Beige/grey crinoid wacke- to packstone	Fossils: brachiopods, rugose and tabulate corals, trilobites, crinoids, stromatoporoids	Lagoon/backreef (Scholle *et al.* 1983, MacNeil and Jones 2006)
HR-1: Black shale	Calcareous and non-calcareous intervals	Minor pyrite and pryrite-replaced shells	Off-ramp (Wendte and Belka 1991)

Chedabucto Lake Formation (Ordovician); the red beds of the Mirage Point Formation (Early Devonian), and locally, the Precambrian basement (Norris, 1965; Drees, 1993). Conodonts recovered from the upper part of the Chinchaga Formation are from the *australis* Zone (Drees, 1993; Figure 2-2). The surface of the underlying Precambrian rocks is highly irregular with relief up to 90 meters (Craig et al., 1967). The Chinchaga Formation lies between these topographic highs. Although the Chinchaga Formation onlaps the Tathlina High (Figure 2-1), it is not found on its crest (Belyea, 1971; Drees, 1993).

The Chinchaga Formation outcrops along a northwest-southeast belt west of Great Slave Lake, following the Canadian Shield on the eastern edge of the basin (Figure 2-3B). The Chinchaga Formation contains evaporites. At many locations, however, the only evidence of the Chinchaga Formation is small patches of white gypsiferous-rich soil. In the subsurface, the Chinchaga Formation is present almost everywhere in the Great Slave Area to a maximum thickness of 109 meters in the central part of the region (Drees, 1993).

The Chinchaga Formation is characterized by rapid lithological changes between breccias, dolostones, limestones, and evaporites (Figure 2-5). Facies C-1 is comprised of breccias found at the base of the Chinchaga Formation (Figure 2-5A and B; Table 2-1). The breccias are chaotic but crude stratification is locally evident. The outcrop-scale breccias (~10 meters thick) contain clasts that are up to one meter long. Some of the larger clasts are also formed of breccia. Most of the angular, beige or orange clasts are composed of dolomite. The surrounding matrix is dark grey vuggy limestone with many vugs filled with calcite. Blocky calcite cement surrounded some of the clasts in the breccia where little or no limestone matrix is present. No fossils were found in the clasts or limestone matrix.

Facies C-2 is an orange-beige, structureless, nonfossiliferous, cryptocrystalline dolostone that contains varying concentrations (10-30%) of

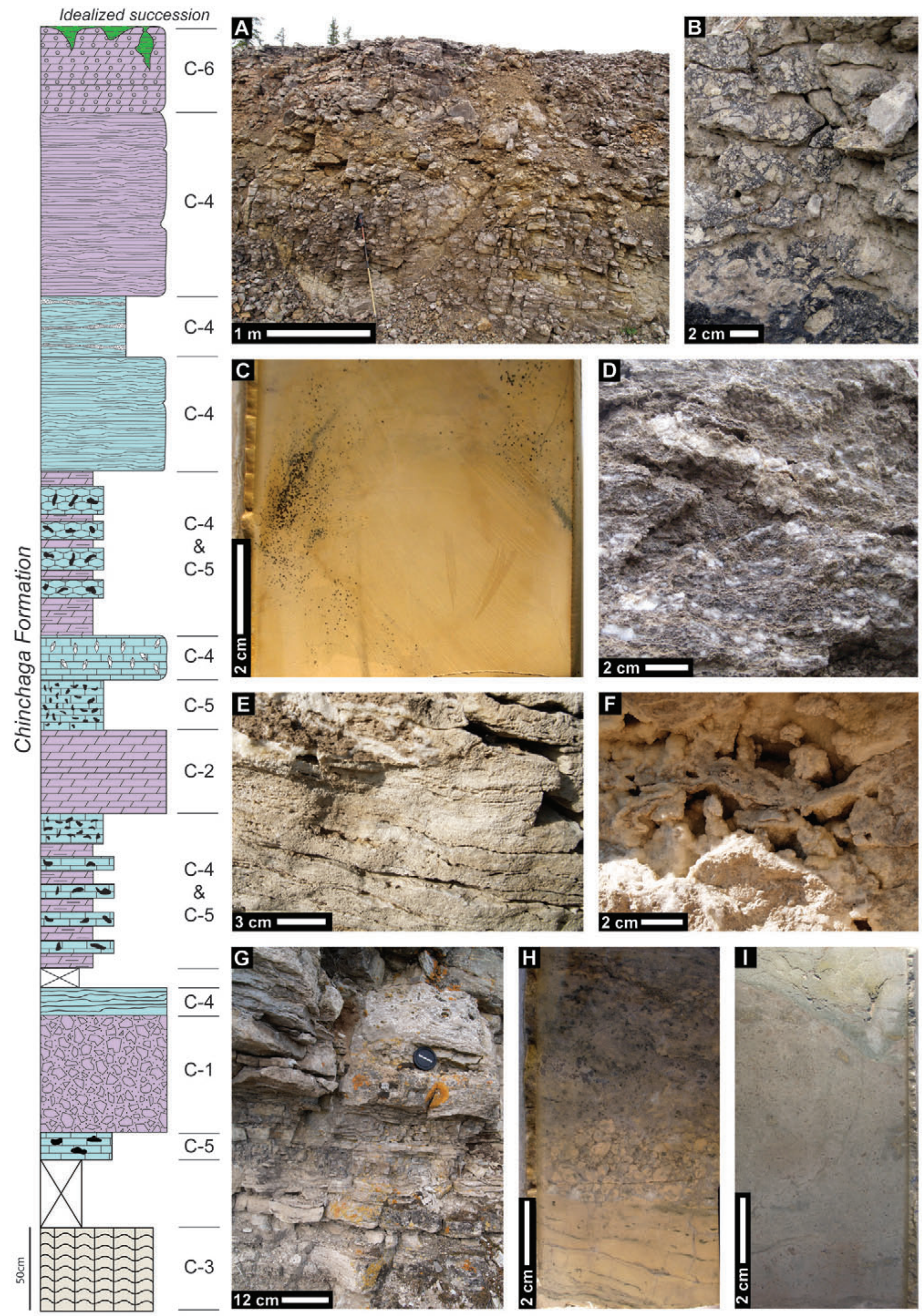

Figure 2-5. Stratigraphic column of the Chinchaga Formation (left) illustrating distribution of major facies. Locations of outcrops are illustrated in Figure 2-3B. (A) Facies C-1: Solution breccia in outcrop (#1). (B) Facies C-1: Solution breccia in outcrop (#1). (C) Facies C-2: Dolostone with large anhydrite crystals from Arrowhead River H-31 drill core. (D) Facies C-3: Interbedded gypsum and dolomite in outcrop (#2). (E) Facies C-4: Laminated dolostone in outcrop (#8). (F) Facies C-5: Porous vuggy dolostone in outcrop (#8). (G) Facies C-4 and C-5: Discontinuous laminated beige dolostone/grey limestone and Facies C-5 (above lens cap) porous vuggy dolostone in outcrop (#8). (H) Facies C-4: Discontinuously laminated beige dolostone/grey limestone from Arrowhead H-31 drill core. (I) Facies C-6: Partially dolomitized mudstone with green shale from Arrowhead River H-31 drill core.

coarse dark grey anhydrite crystals and authigenic quartz (Figure 2-5C; Table 1). There is little to no porosity, no cement, and no fossils in this facies.

Facies C-3 is formed of interbedded gypsum and dolomite (Figure 2-5D; Table 1). This recessive lithology commonly displays a chicken-wire texture. The gypsum occurs as a cm-sized crust in between thin mm-size layers of beige-brown powdery dolomite. Several small white soil patches containing gypsum crystals were seen in the study area, which suggests evaporite units may have been a great deal thicker than the interval measured in this study (~60 cm).

Facies C-4 is formed of beige dolostone (2-10 cm) and dark grey limestone (10-30 cm). The beds are uneven and contain dark grey wavy discontinuous laminations (Figure 2-5E, G, and H; Table 1). Small (mm-sized) vugs and fenestrae are found in the dolostone and limestone beds. Some of the vugs are filled with fine, soft crumbly calcite matrix.

Facies C-5 is a highly porous vuggy limestone/dolostone (Figure 2-5F and G; Table 2-1). This facies displays laminations that appear to “buckle” and form teepee-like structures over large vugs that are lined with euhedral calcite crystals or in some occurrences, saddle dolomite.

Facies C-6 is a grey partially dolomitized mudstone (Figure 2-5I; Table 2-1). Cavities and scour surfaces in this facies have been filled with green shale. Laminations disturbed by clusters of anhydrite crystals give the rock a mottled texture. Fenestral pores, present where the laminae remain relatively undisturbed, are filled with dark anhydrite crystals.

Depositional Setting

The lack of fossils, presence of laminated dolostones, and limestones with fenestrae, evaporites and associated breccias in the Chinchaga Formation collectively indicate an intertidal to supratidal setting. Many of the thin laminae in the limestones and dolostones and vuggy limestone facies are probably indicative of bacterial mats akin to those commonly found in the intertidal zone (Assereto and Kendall, 1977; James, 1984; Jones and Desrochers, 1992; Flügel, 2004). The chicken-wire gypsum and dolomite facies (C-3) are diagnostic of the supratidal zone (Flügel, 2004).

The oligomict clasts, crude stratification, minimal amount of muddy matrix, and the presence of evaporites in the vicinity of the breccias in Facies C-1 indicate an interstratal dissolution breccia (Warren, 2006). The sharp base and irregular upper contact of the breccias also represent interstratal dissolution breccias (Morrow, 1982). These types of breccias are most commonly found in the intertidal or supratidal zone (Flügel, 2004; Gandin and Wright, 2007).

Lonely Bay Formation

The Lonely Bay Formation, defined by Norris (1965), overlies the Chinchaga Formation in most of the Great Slave Area and the Headless Formation along the western edge. Conodont dating places the Lonely Bay Formation in the *ensensis* Zone (Drees, 1993; Figure 2-4). The Lonely Bay Formation is 30-60 m thick. It outcrops northwest of Great Slave Lake and is present in the subsurface throughout the Great Slave Area. The Lonely Bay Formation is absent north of the Great Slave Area (Figure 2-3B) where there is a lateral transition from limestone to shale.

Facies changes are less frequent in the Lonely Bay Formation than in the Chinchaga Formation. Most facies are thick (meter-scale) and laterally continuous. Although formed mostly of limestone there are some thin dolostone beds in the basal part of the formation. Fossils in the Lonely Bay Formation include: *Atrypa* sp., *Emanuella meristoides* (Meek), *Michelinoceros* sp., *Productella* sp., *Spinatrypa* sp. cf. *S. lata* (Warren), *Alveolites* sp., *Aulopora* sp.,

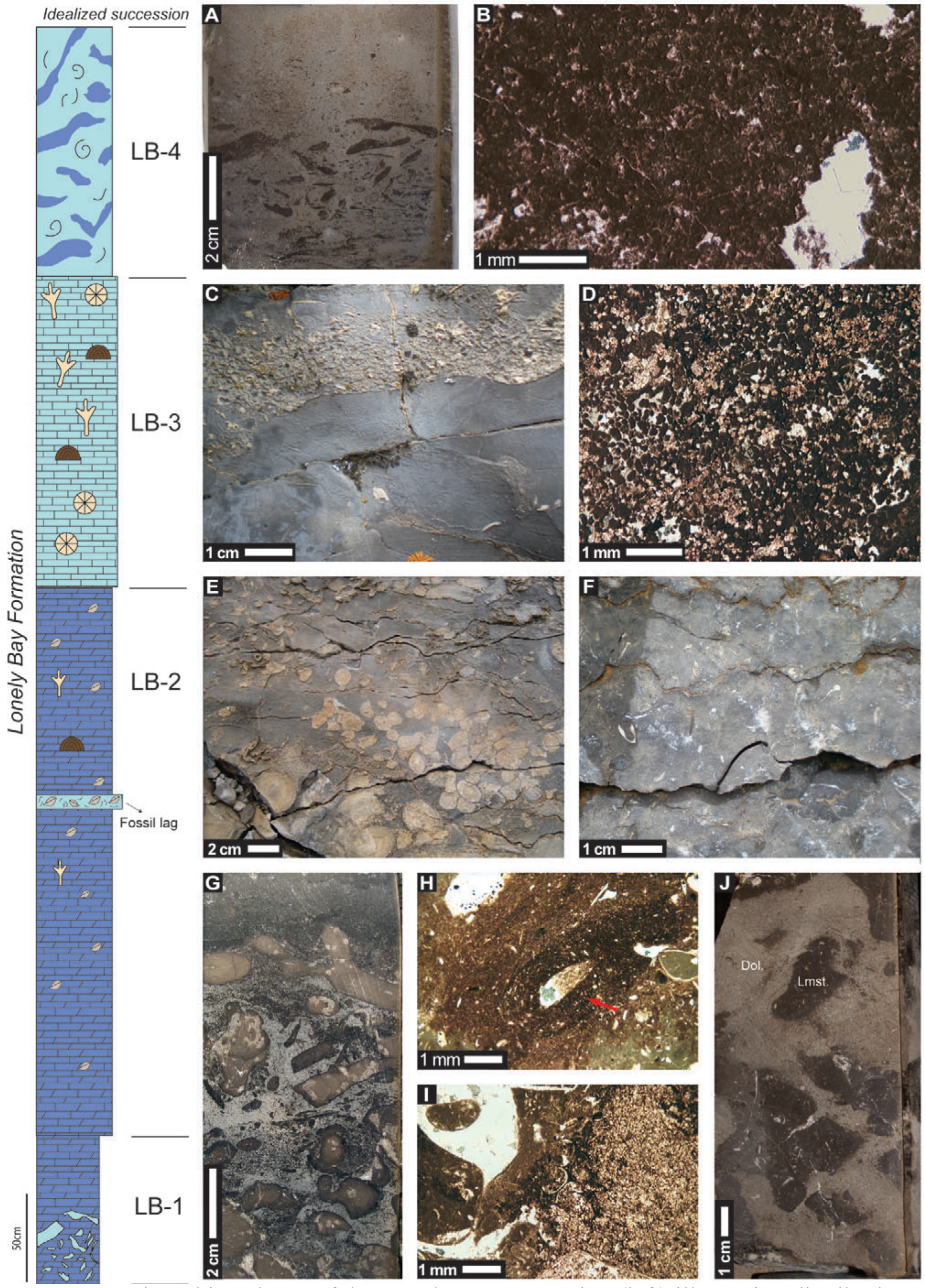

Figure 2-6. Stratigraphic column of the Lonely Bay Formation (left) illustrating distribution of major facies. Locations of outcrops are illustrated in Figure 2-3B. (A) Facies LB-1: Dolomitized intraclast mudstone from Arrowhead River H-31 drill core. (B) Photomicrograph of Facies LB-2: Peloidal partially dolomitized wackestone. (C) Facies LB-2: Peloidal partially dolomitized wacke-grainstone, note fossil lag on scour surface (#3). (D) Photomicrograph of Facies LB-2: Peloidal grainstone. (E) Facies LB-3: Coral flatstone in outcrop (#3). (F) Facies LB-3: Brachiopod floatstone in outcrop (#6). (G) Facies LB-3: Stromatoporoid floatstone from Arrowhead River H-31. (H) Photomicrograph of Facies LB-4: Partially dolomitized strongly bioturbated wackestone, note causative burrow (arrow). (I) Photomicrograph of Facies LB-4: Partially dolomitized strongly bioturbated wackestone. (J) Facies LB-4: Partially dolomitized strongly bioturbated wackestone from Arrowhead River I-46.

Syringopora sp., bulbous stromatoporoids, *Amphipora,* and some unidentified gastropods, ostracods, and crinoid ossicles (Norris, 1965).

Facies LB-1 is a dolomitized intraclast mudstone facies that is common near the base of the Lonely Bay Formation (Figure 2-6A; Table 1). The dark grey or brown angular to subrounded mud clasts (2-5 cm long and 1 cm wide) float in a muddy matrix and decrease in abundance up section. Euhedral dolomite crystals (<1mm) float in the mudstone matrix. There are small unidentifiable fragmented shelly fossils in the matrix.

Facies LB-2 is a peloidal partially dolomitized wacke-grainstone that is grey on weathered surface and brown on fresh surfaces (Figure 2-6B-D; Table 1). Fossils, floating in a dominantly peloidal matrix, include brachiopods, stromatoporoids, trilobites, ostracods, and solitary rugose corals. In some intervals the rocks have a grainstone texture where small (<1mm) blocky calcite cement surrounds the peloids. Large (cm-sized amplitude) stylolites are present throughout this facies. There are also some small intervals (~2cm) of concentrated shells (Figure 2-6C).

Facies LB-3, a brachiopod/coral/stromatoporoid floatstone is a reoccurring facies that consists of various fossil concentrations (Figure 2-6E-G; Table 2-1). The dominant species in the floatstone changes locally. The corals, stromatoporoids, and brachiopods are generally intact and some appear to be *in situ*. The corals in this facies are rugose and tabulate, mostly in branched forms. Although most of the stromatoporoids are bulbous (10-20 cm diameter), there are some tabular and branching forms. Other fossils in this facies include *Amphipora*, gastropods, and trilobite fragments. The fossils float in a greyish-brown mud matrix that has a greyish-brown weathered surface and brown fresh surface. There are a few intervals of concentrated fossils in this facies (5 cm thick) where the muddy matrix is minimal or absent.

LB-4 is a partially dolomitized strongly bioturbated wackestone (Figure 2-6H and I; Table 2-1). The weathered surface is greyish brown whereas the fresh surface is black and brown. Mixing of the black and brown matrices is due to bioturbation. The black mud is composed of calcite, ankerite, and dolomite. Thin (~2 mm) calcite-filled tube-like structures found in the black material are probably the initial “causative” burrows whereas the ragged edged black material is a digenetic halo surrounding the initial burrow (Figure 2-6H). Euhedral to subhedral dolomite is present in the black burrows. Some rhombs have a cloudy centre with a clear rim whereas other dolomite crystals are corroded or completely dissolved (dedolomite). The mud matrix that surrounds the burrows contains gastropods, trilobite fragments, calcispheres, ostracodes, small brachiopod fragments, and foraminifera. Two small intervals (3 cm) of concentrated articulated brachiopod fossils were also seen in this facies.

Depositional Setting

The abundance and nature of fossils in the Lonely Bay Formation indicates a subtidal environment. The presence of *in situ* large bulbous stromatoporoids (10-20 cm) and clusters of branching corals indicate open marine conditions with low to moderate energy levels that would not be capable of toppling over these large organisms (Hladil, 1986; Kershaw, 1990; Konigshof and Kershaw, 2006). The diverse biota indicates favourable environmental conditions, and the presence of *Amphipora* is typical of Devonian lagoons (Klovan, 1964; Hladil, 1986; Sheng and Zang, 1997; MacNeil and Jones, 2006). The floatstones and bioturbated facies in the Lonely Bay Formation contain large quantities of mud, placing them below fair-weather wave base, but fossil lags indicate periodic storm waves. The intraclast mudstone Facies LB-1 also indicates stronger wave action. Intraclasts of this size and shape are common in the intertidal zone (Scholle et al., 1983, Flügel, 2004). This facies was consistently found near the

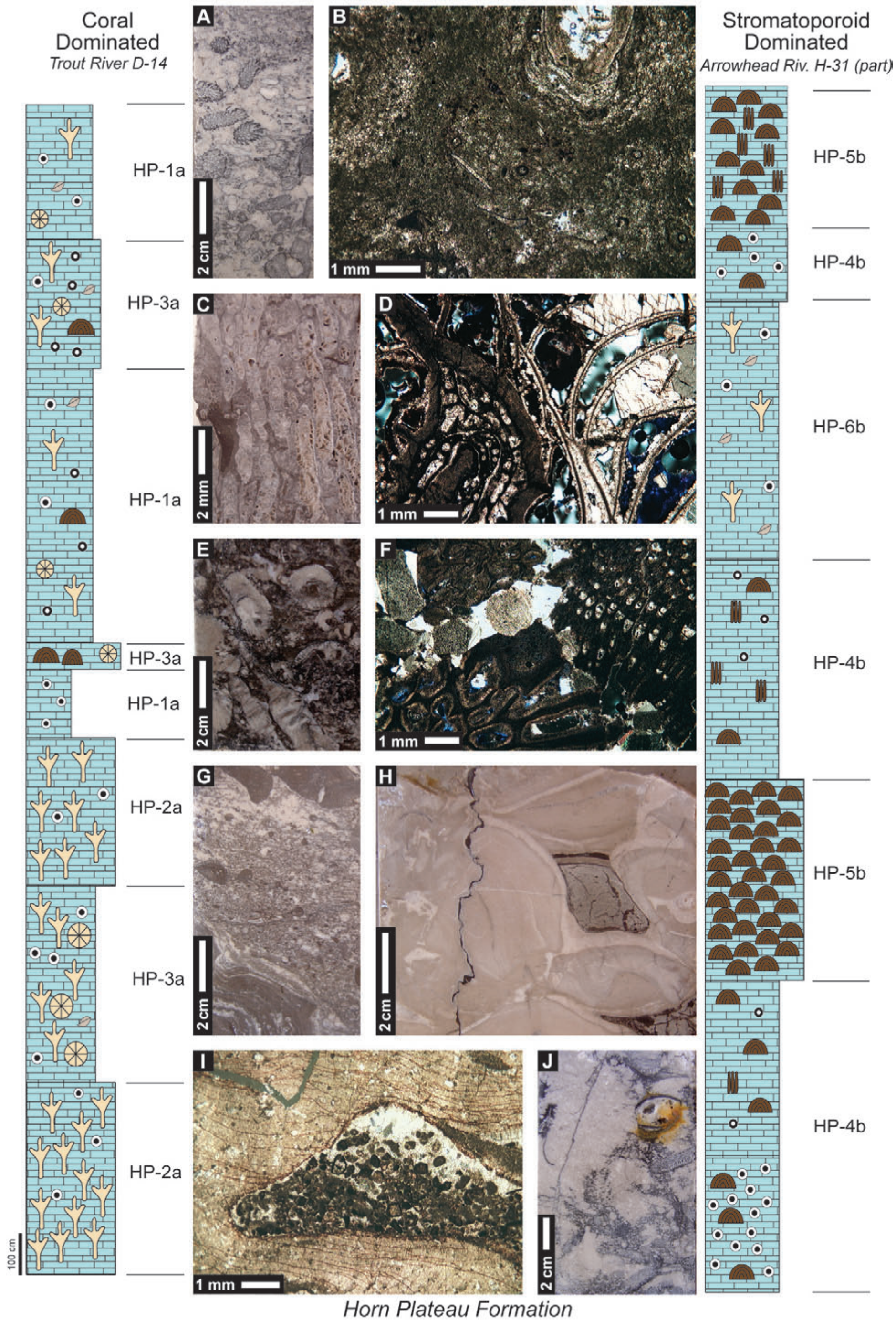

Figure 2-7. Stratigraphic columns of the Horn Plateau Formation showing coral-dominated cores (left) and stromatoporoid-dominated (right) cores. (A) Facies HP-1a: Coral floatstone from Trout River D-66 drill core. (B) Photomicrograph of Facies HP-1a: Coral floatstone. (C) Facies HP-2a: Coral bafflestone from Trout River D-14 drill core. (D) Photomicrograph of Facies HP-2a: Coral bafflestone. (E) Facies HP-3a: Crinoid coral rudstone from Trout River D-14 drill core. (F) Photomicrograph of Facies HP-3a: Crinoid coral rudstone. (G) Facies HP-4b: Stromatoporoid floatstone in Poplar River G-32 drill core. (H) Facies HP-5b: Stromatoporoid framestone from Poplar River G-32 drill core. (I) Photomicrograph of Facies 5b: Stromatoporoid framestone showing sorting of peloids and geopetal cement. (J) Facies HP-6b: Coral brachiopod floatstone from Cormack N-33 drill core.

base of the Lonely Bay Formation indicating shallow water conditions at the initiation of deposition of this succession.

Horn Plateau Formation

The Horn Plateau Formation, first described by Norris (1965), encompasses “reefs” that are rooted on top of the Lonely Bay Formation in the south part of the MacKenzie Basin. The term “reef” has been used (e.g. Vopni and Lerbeckmo, 1969, 1972a, b; Fuller and Pollock, 1972) because of the numerous corals and stromatoporoids that are evident in cores from this succession (Figure 2-7). Herein, they are referred as buildups because there is insufficient data to precisely delineate their size, geographic configuration, and internal architecture. Shales belonging to the Horn River Formation encase the buildups. Conodont dating places the initiation of these buildups in the *varcus* Zone and growth of the buildups ceased in the *asymmetricus* Zone (Vopni and Lerbeckmo, 1972; Fuller and Pollock, 1972; Figure 2-2).

The buildups in the Horn Plateau Formation, 52 to 116 m (170-380 ft) high, are aligned along a southwest to northeast trend that stretches for ~350 km (Figure 2-8A and C). Determining the areal extent and geographic configuration of these buildups is difficult because the wells are 13 to 86 km apart (Figure 2-8A and C). The reef exposed at Fawn Lake is ~ 1.3 km in diameter (Vopni and Lerbeckmo, 1969) and wells drilled by Amoco Canada Petroleum Ltd. showed that it was up to 105 m (344 ft) thick. Although there is a high probability that the wells to the southwest also penetrated isolated reefs, the sparse well coverage and lack of seismic information precludes accurate resolution of this issue.

Cores from the buildups in the Horn Plateau Formation are formed entirely of limestone containing well-preserved fossils. Cores from the buildups in the southwest are stromatoporoid-dominated whereas those from the northeast

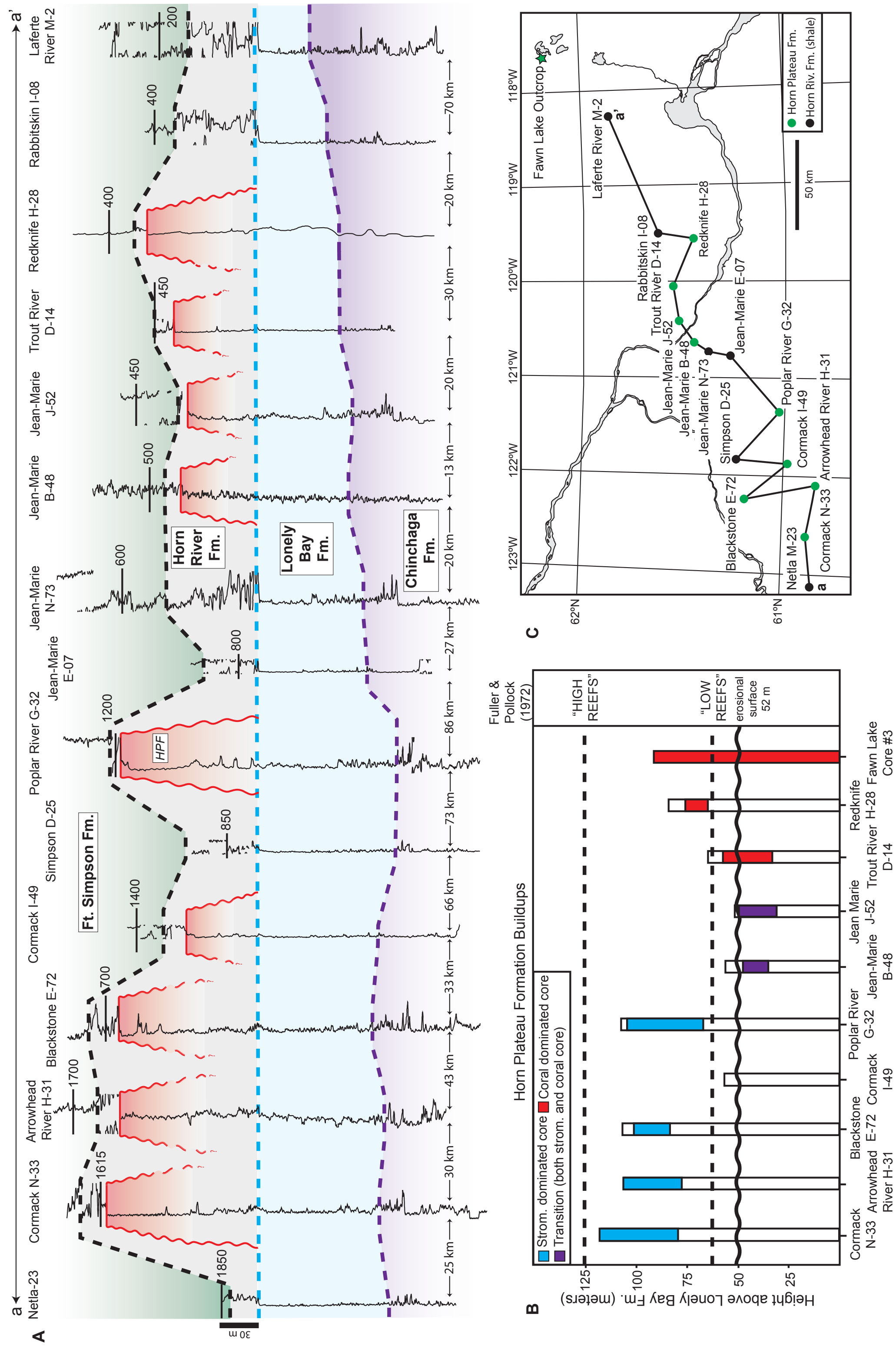

Figure 2-8. Horn Plateau Formation. (A) Cross-section a-a' showing the correlation between wells in the District of MacKenzie. (B) Graph showing drill cores that penetrated the Horn Plateau Formation. Information from Fuller and Pollock (1972) is displayed in the graph for comparison. (C) Location map showing the location of wells included in cross-section a-a'.

buildups are coral-dominated. Cores from the central part of the transect contain a mixed coral-stromatoporoid biota (Figure 2-8B). Core #3 from Fawn Lake, the only core that encompasses the entire formation, is coral-dominated throughout. All of the other cores came from the upper portion of the buildups; thus, it is impossible to determine if the biota evident in the core is representative of the entire formation.

The depositional textures evident in the stromatoporoid- and coral-dominated limestones are characterized by different facies that herein reflect the fossil content, morphology of the constituent organisms, and the amount and type of matrix and cements.

Coral Buildups

Facies HP-1a, the most common facies in the coral buildups is the crinoid coral floatstone facies (Figure 2-7A and B; Table 2-1). Corals in these buildups include: *Favosites* sp., *Siphonophrentis?* sp., *Disphyllum salicis* n. sp., *Cylindrophyllum gruensis* n. sp., *Grypophyllum cornus* n. sp., *Neostringophyllum craigi* n. sp., *Austalophyllum*? cf. *A.? thomasae* (Hill and Jones), *Heliophyllum borealis* n. sp., *Cyathophyllum (Peripaedium) greteneri* n. sp., *Sinospongophyllum (Sociophyllum) redactum* n. sp., *Lekanophyllum* cf. *L. punctatum* Wedekind, *Cystiphylloides spinosum* n. sp., *Atelophyllum nebracis* n. sp. (MacLaren and Norris, 1964). Other fossils include brachiopods, crinoids, and molluscs. The light beige wacke- to packstone matrix contains variable concentrations of small (2-5 mm) crinoid ossicles. Bladed isopachous calcite lines intragranular pore space in the corals and blocky cement filling the remaining intergranular pore space.

Facies HP-2a is a coral bafflestone composed mainly of branching tabulate corals held in a beige mud- to wackestone matrix that is mostly mud with some floating crinoid ossicles and broken shelly fossils (Figure 2-7C and D; Table 2-1) and scattered disarticulated brachiopods. The branching corals in this facies are all oriented in the same direction. Intragranular porosity in this facies is high and most pores are lined with small euhedral calcite.

Facies HP-3a is crinoid coral rudstone containing a diverse biota of crinoids (up to cm-scale), tabulate corals, rugose corals, stromatoporoid fragments, and brachiopods (Figure 2-7E and F; Table 2-1). There are some articulated crinoid ossicles in this facies. There is little to no mud matrix, and where present, the matrix is a crinoid wackestone to packstone. Calcite cements in this facies line pores of the coral and brachiopod intragranular porosity.

Stromatoporoid buildups

Facies HP-4b, a stromatoporoid floatstone, is the most common facies in the stromatoporoid-dominated buildups (Figure 2-8G). This facies includes bulbous and branching stromatoporoids, commonly fragmented, that are floating in a muddy matrix. The mudstone to crinoid wackestone matrix is dark grey and beige. Fossils in this facies include: *Amphipora,* brachiopods, crinoid ossicles, trilobite fragments, ostracodes, and scattered rugose corals. Some fossils are coated with a fine rind of fibrous calcite cement on the outside or inside of the skeleton or shell.

Facies HP-5b, the organic core of these buildups, is a stromatoporoid framestone. This facies is composed almost entirely of large (10-25 cm diameter) bulbous stromatoporoids that appear to be *in situ* (Figure 2-7H and I; Table 2-1). Other allochems in this facies include peloids. Geopetal structures in this facies are filled with pelloidal matrix that appears to have undergone minor sorting. Fibrous calcite cement lines the outside of some of the fossils and intragranular calcite cement is found in the galleries of the stromatoporoids.

Facies HP-6b is a coral brachiopod floatstone facies formed largely of tabulate and rugose corals held in a grey or beige crinoid wacke- to packstone matrix (Figure 2-7J; Table 2-1). Other fossils include brachiopods, rugose corals, tabulate corals, trilobite fragments, crinoid ossicles (2-5 mm diameter), stromatoporoids (bulbous, *Stachyodes* and *Amphipora*) and large cephalopods. Although not in growth position, the fossils are well preserved. Cements line the inside of many of the brachiopods.

Depositional Setting

Fuller and Pollock (1972), who examined cores from five wells (Cormack N-33, Blackstone E-72, Poplar River G-32, Jean Marie B-48, Trout River D-14) and the cores from the Fawn Lake reef, divided the buildups in the Horn Plateau Formation into two “low” and four “high” reefs and used that to suggest that there were two stages of reef growth in the area (Figure 2-8B). They noted an exposure surface in the Fawn Lake Horn Plateau Formation buildup at ~ 51.8 m (~170 ft) “above the platform” (herein assumed to be the top of the Lonely Bay Formation). From their descriptions it is not clear if that surface is related, in any respect, to the “low” or “high” reefs.

Vopni and Lerbeckmo (1969, 1972a, b) made no mention of exposure surfaces in the Fawn Lake buildup. They divided this reef into the reef flat facies, reef flank facies, organic reef facies, and two deep-water facies that had a high mud content. In this study, the floatstone facies (HP-1a and HP-6a) with 40-70% mud is akin to the deep-water facies; facies HP-2a and HP-4b of the stromatoporoid and coral dominated buildups are equivalent to the organic reef facies; and the rudstone facies equate to the reef flank facies. Cores 2, 4, and 5, which were drilled around the outer edges of the reef, penetrate these flank deposits.

The buildups, dominated by stromatoporoids in the southwest and corals in the northeast, are 52 to 116 m high. Fuller and Pollock (1972), in dividing the buildups into “high” and “low” groups implicitly assumed that the well had penetrated through the thickest part of the reef rather than through the reef flank. The difference in height between the reefs may also be the direct result of available accomodation space. The underlying Lonely Bay Formation thickens down-ramp from the northeast to the southwest (Figure 2-8). Reefs in the Horn Plateau Formation also appear to be thicker overall in the southwest.

The Horn River Formation

The Horn River Formation (HR-1), defined by Whittaker (1922), overlies the Lonely Bay Formation and encases the Horn Plateau Formation. The Horn River Formation is overlain by the Late Devonian Fort Simpson Formation and the Hay River Formation (Norris 1965). Conodont dating places the Horn River Formation in the Lower *asymmetricus* Zone (Drees 1993; Figures 2-2 and 2-4). The Horn River Formation is discontinuously exposed along the banks of the Horn River, with the thickest section being ~10 m thick (Douglas and Norris, 1960). In the subsurface it is 45 to 300 meters (Norris, 1965). The Horn River Formation is composed of dark grey to black, locally calcareous shales. There are a few burrowed intervals. Other fossil remains include scattered pyritized brachiopods.

Depositional Setting

These shales represent a subtidal environment. The calcareous intervals may represent pulses of calcareous sediment that has been transported into the deeper water environment.

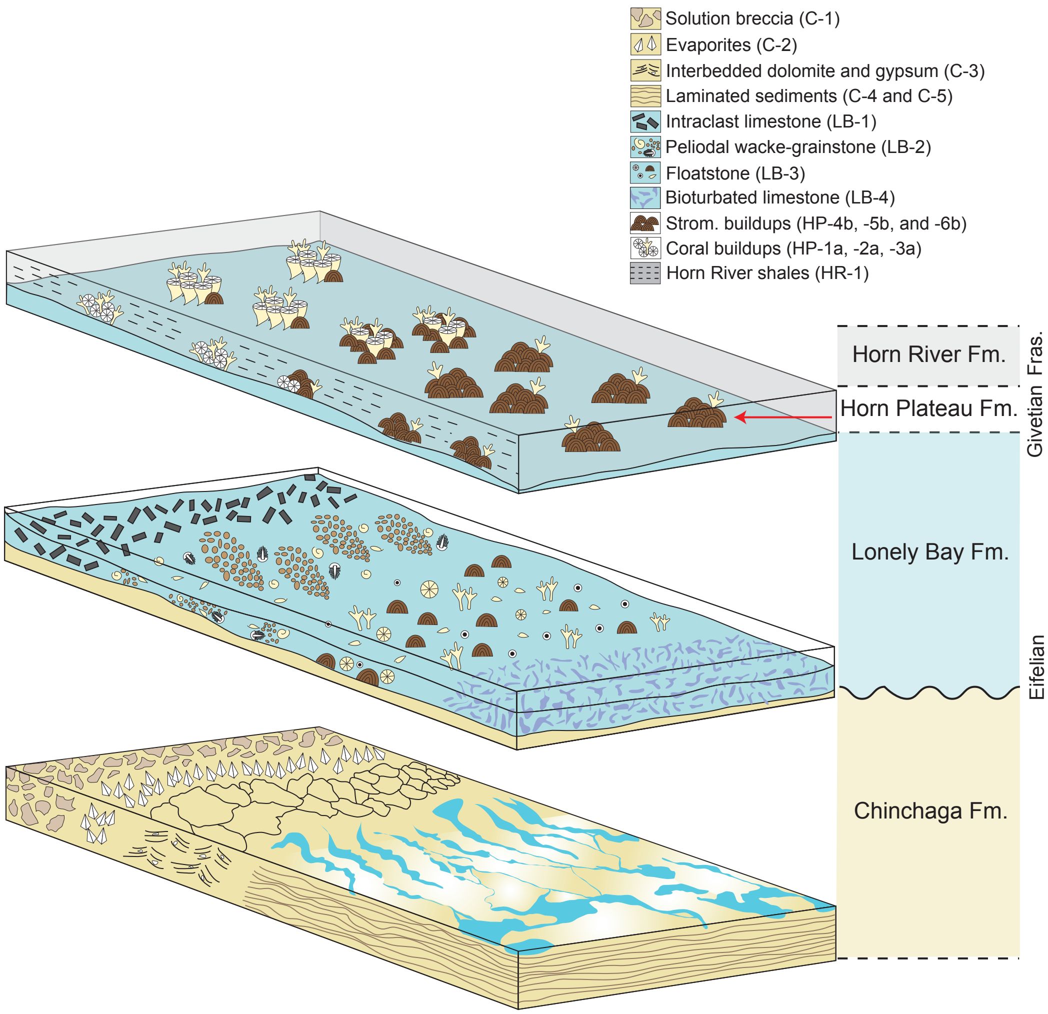

Figure 2-9. Schematic block diagrams and stratigraphy of the MacKenzie Basin. (A) The MacKenzie Basin ramp through time, showing major facies from the Chinchaga Formation, the Lonely Bay Formation, the Horn Plateau Formation, and the Horn River Formation shown in depositional model and simplified stratigraphic column of the MacKenzie Basin.

Depositional Regimes

The Chinchaga Formation, Lonely Bay Formation, Horn Plateau Formation, and Horn River Formation include facies that developed on a shallow dipping (~10°) carbonate ramp (Figure 2-9). There is no evidence of any break in slope.

Facies in the Chinchaga Formation formed in a shallow supratidal to intertidal, inner ramp setting (Figures 2-5 and 2-9). Evidence for evaporative, shallow water conditions typically found in the supratidal is apparent in facies

C-1, C-2, C-3, and C-6 where evaporites are either present or inferred (i.e. solution breccias). The intertidal zone is represented by facies C-4 and C-5, where microbial mats imply periodically submerged conditions and teepee structures indicate periodic exposure and desiccation (cf. Middleton, 1961; Assereto and Kendall, 1977; Ahr, 1985).

The Lonely Bay Formation contains facies that represent a transitional inner ramp to mid-ramp setting (Figures 2-6 and 2-9). Rip-up clasts in a mud matrix in facies LB-1 formed as a result of fluctuating energy levels. Facies LB-2 is above fair-weather wave base, where mud was winnowed from between allochems and grains underwent sorting. The high mud content in facies LB-3 and LB-4 point to quiet-water conditions, probably below fair weather wave base. The presence of fossil lags, however, indicates water depths above storm wave base. Facies LB-4, characterized by extensive burrowing, probably formed at or below storm wave base.

The buildups in the Horn Plateau Formation, which are encased by shales of the Horn River Formation (Figures 2-8 and 2-9), are akin to those commonly found on the outer to mid ramp of interior cratonic basins (e.g. Burchette and Wright, 1992). Facies HP-3a and HP-5b represent the organic framework of these buildups, where the highest wave energies were absorbed. The other facies are moderate to quiet water facies. Facies HP-2a, the coral bafflestone, is typical of a moderate energy environment. The corals in the bafflestone are surrounded by mud and fragments of other biota carried in by wave action and deposited between the branching corals. Facies HP-1a and HP-4b are quiet water facies that contain talus from the core organic facies floating in a mud matrix. The high mud content suggests a lagoonal type environment or a deeper water facies of the reef.

The Horn River Formation shales that surround the Horn Plateau Formation represent a deeper water environment. There is little to no fossil material present in the Horn River Formation shales.

Discussion

Overall, Devonian strata in the MacKenzie Basin record a gradual sea level rise, beginning with evaporitic tidal flat conditions in the Chinchaga Formation and ending with the deep-water shales of the Horn River Formation. To evaluate the relevance of these findings on a North American and global scale, these strata are viewed in terms of the sea level curves of Lenz (1982) and Johnson et al. (1985; Figure 2-2).

Lenz (1982) and Johnson et al. (1985) developed sea level curves by examining strata from various parts in the world. Lenz (1982) compared paleobathymetric curves from seven North American locations and highlighted the “major” and “secondary” transgressions and regressions that were common to all study areas (Figure 2-2A and B). He gave no official designations to these transgressive and regressive cycles. One of the areas, the “Central MacKenzie Valley”, located in the Great Bear Plains (Figure 2-1), lies ~100 km north of the MacKenzie Basin. It has stratigraphically equivalent formations and therefore provides an ideal comparison curve for the MacKenzie Basin strata examined in this study.

One of the most well known and utilized sea-level curves for the Devonian by Johnson et al. (1985), compared rocks in western Canada (MacKenzie Basin and WCSB), the western United States (Idaho and Nevada), New York, and Europe (Belgium and Germany). By combining sedimentological and conodont data from all of these sites, they produced a eustatic curve for the Devonian that displayed numerous transgressive-regressive (T-R) cycles. They recognized two major cycles or “depophases” designated by Roman numerals, that were further divided into cycles, designated by lower-case letters (e.g. *T-R cycle Ia, Ib* or *IIb*; Figure 2-2). This curve was selected for comparison with the MacKenzie Basin strata because Johnson et al. (1985) included the District of MacKenzie as one of the five study areas in their paper.

Although the eustatic curve of Johnson et al. (1985) displays higher resolution sea-level changes than the Lenz (1982) curve for the Central MacKenzie Valley, both show an overall transgression in the Devonian until the end of the Frasnian. A notable difference between the two curves is in the Middle Eiflian, where Lenz (1982) implied a regression that was correlated to a depositional hiatus (Figure 2-2). Both the Lenz (1982) and Johnson et al. (1985) curves show a regression in the Early-Middle Givetian.

Starting in the middle Eiflian, a transgression that corresponds to a secondary transgression as defined by Lenz (1982) and *T-R cycle Id* (Johnson et al., 1985), produced the shallow water facies of the Chinchaga Formation (Figure 2-2). The evaporites, presence of bacterial mats, and teepee structures all point toward a shallow restricted environment (Figures 2-5 and 2-9). This Middle Devonian transgression marks the beginning of widespread carbonate deposition in Western Canada.

The second transgressive event that influenced the MacKenzie Basin took place in the late Eiflian *ensensis* Zone (Figure 2-2). Shown in Lenz (1982) and in Johnson et al. (1985) as the *T-R cycle Ie*, this deepening corresponds to open marine conditions represented by the Lonely Bay Formation. The diverse, *in situ* fossils and muddy substrate in the Lonely Bay Formation are consistent with relatively quiet, fully marine conditions.

At the Eiflian-Givetian Boundary, the sea-level curve in Lenz (1982) indicates no change in sea-level whereas the Johnson et al. (1985) curve shows a transgression at the beginning of *T-R cycle If* (Figure 2-2). Growth of the Givetian-aged Horn Plateau Formation buildups, which were rooted on the Lonely Bay Formation, began at that time. If all of the Horn Plateau Formation buildups are akin to the pinnacle-like reef at Fawn Lake, it seems probably that their preferential vertical growth was achieved as they "kept-up" with the constantly rising sea level associated with the ongoing transgression.

In the early to Mid-Givetian, a regression is indicated on both sea level curves (Lenz, 1982; Johnson et al., 1985; Figure 2-2). Lenz (1982) based this regression on one restricted ostracod unit in the "Central MacKenzie Valley" area at the base of the Hume Formation (Figure 2-4B). The two units overlying this, however, were richly fossiliferous and consistent with open marine conditions (Lenz, 1982). In Johnson et al. (1985), the evidence for a regression in *T-R cycle If* in Western Canada was based on an (1) an unconformity associated with the Watt Mountain Formation in the WCSB and (2) exposure of the Horn Plateau Formation buildups.

The Watt Mountain Formation is a sandy shale found in the Western Canadian Sedimentary Basin. There is an unconformity at the base of the Watt Mountain Formation that has led to some debate as to whether the Watt Mountain Formation belongs in the top of the Elk Point Group (Belyea and Norris, 1962; Drees, 1993) or the base of the Beaverhill Lake Group (Basset and Stout, 1968; Braun et al., 1988; Oldale and Munday, 1994; Potma et al., 2001, Wendte and Uyeno, 2005). Regional uplift, clastic shedding and deposition (Wendte and Uyeno, 2005) led to relative sea level fall and exposure in the WCSB. The sub-aerial exposure and unconformity at the base of the Watt Mountain Formation is cited as evidence for the Givetian regression in *T-R cycle If* regression (Johnson et al., 1985). Stratigraphically, the sub-Watt Mountain unconformity is equivalent to the late stage of the Horn Plateau Formation Buildups; however, the Watt Mountain and its basal unconformity were not observed in any of the drill core examined in this study of the MacKenzie Basin. The farthest north that the Watt Mountain Formation has been clearly recognized is just east of the Hay River

Bank margin (Oldale and Munday, 1994). The lack of Watt Mountain Formation siltstones and shales and its associated unconformity in the MacKenzie Basin, north of the Presqu-ile Barrier indicates that the Mid-Givetian regression was limited to the restricted Elk Point Basin, just south of the District of MacKenzie (Figure 2-1).

The second piece of evidence for the Mid-Givetian regression in *T-R cycle If* is the "exposure surface" in the Horn Plateau Formation buildups (Johnson et al., 1985), which was based on information provided by Fuller and Pollock (1972). Based on data from five wells (Cormack N-33, Blackstone E-72, Poplar River G-32, Jean-Marie B-48, Trout River D-14) and the Fawn Lake reef, Fuller and Pollock (1972) argued that exposure of the buildups was evident from the (1) "low" and "high" stages of reef growth in the Horn Plateau Formation, (2) exposure surface in the core from the Fawn Lake area (Figure 2-8), (3) filled "fissures" that characterized some of these buildups, and (4) conodont dates obtained from the Horn Plateau Formation. A detailed examination of all of the core from the wells that penetrated these buildups and the five cores from the Fawn Lake area during this study failed to support these basic conclusions. Given their importance of these features, each feature is dealt with below.

- The notion of "low" and "high" reefs is not supported by the regional data. Fuller and Pollock (1972) found two "low" reefs and four "high" reefs in their study of the Horn Plateau Formation (Figure 2-8B). The 11 buildups examined in this study are 52 to 116 m high (Figure 2-8A). The thicknesses of the buildups are similar on the northeastern and southwestern parts of the cross-section (91-116 m), but there are lower (~52 m) reefs in between. The underlying Lonely Bay Formation and Chinchaga Formation thicken to the west, further away from the Canadian Shield that was emergent during throughout the Devonian (Figures 2-1 and 2-2). This pattern of thickening to the west is not evident in the Horn Plateau Formation. Likewise there does not seem to be any correlation between the height of the buildups and the dominant framework-building organism seen in the cores (Figure 2-8B). There is no way to be certain that each buildup was drilled through the thickest part of the reef. Some of the wells could have been drilled through the thinner flank deposits.

- Fuller and Pollock (1972) regarded the "exposure surface" in Fawn Lake reef (51.8 m above base of reef) as further evidence for two stages of reef growth. Assessment of this exposure surface is hampered by the fact that most cores from the Horn Plateau Formation buildups come from the upper part of the formation (Figure 2-8B) and therefore do not reach the depth where the exposure surface is suppose to exist. Only Core #3 from Fawn Lake reef includes all of the Horn River Formation and the upper part of the Lonely Bay Formation. Fuller and Pollock (1972), however, did not specify the exact position of the erosion surface in core #3 – instead they argued that they saw the "exposure surface" 52 m above the top of the "platform" in Core #1. Core #1, however, does not penetrate the entire formation; thus, the exact location of the "exposure surface" is difficult to establish, especially since they did not provide any photographs to support their contention. Their argument seems to rest entirely on the presence of coral pebbles in lime mudstone that are "…not found higher than about 52 m…" (Fuller and Pollock, 1972). Coral rubble however, is very common in reefs that are dominated by coral growth (James and Bourque, 1992; Flügel, 2004). There are no obvious erosion surfaces in any of the cores from the Fawn Lake area and the lack of cements

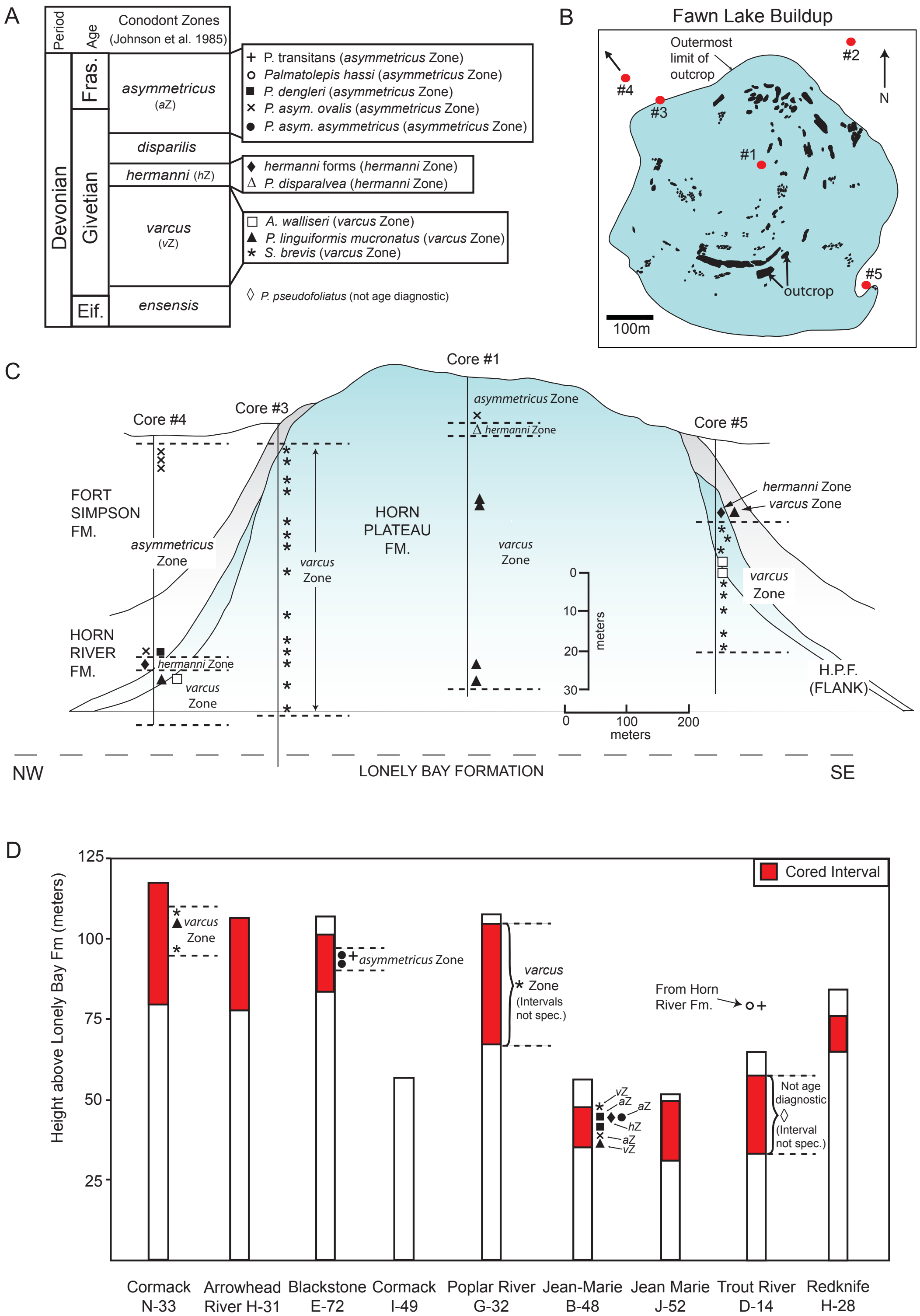

Figure 2-10. Summary of conodont data from Fuller and Pollock (1972). (A) Middle Devonian time scale and relevant conodont Zones. (B) Map of Horn Plateau Formation exposures (adapted from Vopni and Lerbeckmo 1972a; 1972b). (C) Cross-section view of the Fawn Lake buildup showing the conodonts that were recovered from the Horn Plateau Formation (adapted from Fuller and Pollock 1972). (D) Graphs of the Horn Plateau Formation subsurface buildups and the conodonts recovered from the available drill core.

other than marine phreatic cements further supports this assessment. Furthermore, no erosional surfaces were evident in any of the cores from the Horn Plateau Formation. There are small excursions in the gamma ray logs from some wells, possibly representing shale stringers, but none of these can be correlated from well to well.

- Fuller and Pollock (1972) claimed that fissure fills, evident in the core from the Fawn Lake reef supported the notion of an exposure surface. They argued that these “fissures” formed in “low” and “high” reefs when the summits of the reefs were 90 m above sea level. It is difficult to substantiate their presence and significance because Fuller and Pollock (1972) did not include precise locations for these fills and provided no photographs of the “fissures”. They described them in the Fawn Lake cores as “fissures” that contain crinoidal debris and geopetal structures, both of which are not unusual in reef settings. Given that, the Fawn Lake cores are only 4 cm in diameter, it is difficult to establish the presence of fissures. Vopni and Lerbeckmo (1970, 1972a, b), who examined the Horn Plateau Formation exposed at Fawn Lake and the five cores from the area made no mention of the “fissures” in their study. Similarly, detailed examination of every core through the Horn Plateau Formation in this study failed to locate any evidence of fissures.
- Fuller and Pollock (1972) used conodonts from cores at Fawn Lake and the other five wells in their study (Cormack N-33, Poplar River B-32, Blackstone E-72, Jean-Marie B-48, and Trout River D-14) to date the buildups in the Horn Plateau Formation. With five cores drilled at Fawn Lake, it is the only location that provides the resolution necessary to map out the timeline of reef growth

using available conodont data. Two cores that penetrate the main "organic" facies (Vopni and Lerbeckmo, 1969) of the reef are Cores #1 and #3. Conodonts recovered from these cores revealed conodont zones corresponding to the Early to Mid Givetian *varcus* Zone from the base to the middle of the reef. Two subsequent ages in the *hermanni* and Lower *asymmetricus* Zone were found at the top of the reef. The conodonts are in stratigraphic order, which means that reef growth began in the Early to Middle Givetian and continued through to the Late Givetian. The flank deposits, which were seen in Cores #2, #4, and #5, contain conodonts from the *varcus, hermanni, and asymmetricus* zones, except these conodonts are not found in stratigraphic order (Figure 10). This is not unusual in flank deposits that are composed of off-reef debris. Conodonts recovered from Horn River Formation shales directly overlying the Fawn Lake buildup are from the *asymmetricus* Zone, which is the youngest conodont zone that was found in the center and at the top of the buildup in Core #1. In the remaining five wells that were used in Fuller and Pollock (1972), it is difficult to accurately date each buildup because there is no way of determining if the well penetrates the center of the buildup or the flank deposits. Using the distribution of the conodonts in the Fawn Lake cores as a guide it could be surmised that Cormack N-33 and Poplar River G-32, which contain conodonts belonging to the *varcus* Zone, are through the center "organic reef facies" (Vopni and Lerbeckmo 1969, 1972a, b) of the buildups. The conodonts recovered from Jean-Marie B-48 (*varcus* Zone to Lower *asymmetricus* Zone) could, in turn, be interpreted as part of the flank deposits (Figure 2-10). Blackstone E-72 contains conodonts from the *asymmetricus* Zone in the upper part

> of the core, which could mean that this well penetrated either the flank or organic reef facies, since these conodonts were seen in both the upper flank and the organic reef facies in the Fawn Lake buildup. The uncertainty of well placement with respect to the thickest, central part of the buildups makes age determination and correlation impossible.

In their detailed study of the Fawn Lake reef, Vopni and Lerbeckmo (1970, 1972a, b) made no mention of the features listed in Fuller and Pollock (1972). Similarly, our examination of all the cores through the Horn Plateau Formation failed to substantiate any of the features that Fuller and Pollock (1972) used to support their notion of an erosion surface and two stages of “low” and “high” reef growth.

From the available data (electronic wells, drill cores, and thin sections) on the Horn Plateau Formation in the District of MacKenzie (Figure 2-1), there is no conclusive evidence to support two stages of “low and high reef growth”, or a long period of exposure of the buildups. Growth of the Horn Plateau Formation buildups first started in the Mid-Givetian and the lower two thirds of the buildup at Fawn Lake are placed in the *varcus* Zone. Two intervals at the top and in the center (Core #1) of the buildup indicate that reef growth continued into the *hermanni* Zone. The flanking facies, which contain fossils from all three conodont zones, represent the reef talus and debris. The Horn River Formation shales, containing conodonts belonging to the *asymmetricus* Zone, started accumulating around these buildups during the last phase of reef growth (Figure 2-10). The accumulation of these shales co-incides with *T-R cycle IIa*, which is at the beginning of the second major depophase in Johnson et al. (1985).

In summary, the Mid-Givetian regression (Figure 2-2) is not evident in the MacKenzie Basin. There is evidence of a relative sea level drop further south where the Tathlina Arch and surrounding barrier reefs rooted in the Lower Keg

River Platform limited access to the open ocean (Skall, 1975; Kent, 1994; Fu *et al.*, 2006), but the MacKenzie Basin was directly connected to the open ocean. The mid-Givetian regression is not included in the other locations in North America that were examined by Lenz (1982). Johnson et al. (1985) noted that deposition is commonly shown as being uninterrupted in Europe during the *T-R cycle If* regression. There is also no compelling evidence for a break in carbonate deposition in the Canning Basin, Australia (Becker and House, 1997).

Two areas, that share similar paleographic settings to the MacKenzie Basin during the Middle Devonian, are the southwestern United States (Nevada) and the southwestern margin of the Siberian continent (Altai-Salair belt). There is no clear evidence for the *T-R cycle If* regression in Nevada or the Altai-Salair belt. Both of these areas experienced carbonate deposition directly on basement rocks that were gently inclined seaward and connected to ocean waters (Yolkin et al., 1997). They also share a similar history of relative sea level rise and fall throughout the Givetian. The transgression at the beginning of the Givetian, which corresponds to the beginning of *T-R cycle If* resulted in the initiation of reef growth (Johnson and Murphy, 1984; Yolkin et al., 1997). These two areas that were connected to the open ocean, do not show evidence of a major regression in the Mid-Givetian. Another example that mirrors the differences in relative sea level change between the MacKenzie Basin and the Western Canadian Sedimentary Basin are the Moravia (Czech Republic) and Ardennes (Belgium), two Middle Devonian Basins that were located in the Rheic Ocean. The proximity of the Ardennes Basin to the mainland resulted in increased detrital inputs and progradation of the shoreline and Moravia, which is disconnected from the mainland shows a pattern of retrogradation and relative sea level rise (Boulvain et al., 2010). These two basins, despite their proximity, did not respond in the same manner despite eustatic sea level changes.

In comparing the depositional history of the MacKenzie Basin to the timing of Devonian sea level changes by Lenz (1982; Figure 2-2 – Central Valley Location) and Johnson et al. (1985), it is apparent that the major regression shown on both curves is not evident in the MacKenzie Basin. Other basins that were connected to the open ocean, such as Nevada, the Altai-Salair belt, and Moravia in Belgium also show a continuous sea-level rise throughout the Givetian (Yolkin et al., 1997; Boulvain et al. 2010). The restricted basins, considered to be inland basins, do show evidence for a regression. The Western Canadian Sedimentary Basin, which is adjacent to the MacKenzie Basin, experienced a prolonged period of evaporite deposition and clastic shedding from the highlands that were exposed at the time. The differences between the MacKenzie Basin and the WCSB indicate that deposition in the inland seas was not directly controlled by eustasy. Tectonic changes were the dominant controls on deposition in interior basins during the overall sea-level rise in the Middle Devonian.

Conclusions

This study of the MacKenzie Basin has revealed the importance of paleogeographic settings of an area when evaluating the effect of eustatic sea-level changes. The detailed study of the Chinchaga Formation, Lonely Bay Formation, Horn Plateau Formation, and Horn River Formation has led to the following conclusions:

1) Strata of the MacKenzie Basin reflect deposition on a large, shallowly dipping carbonate ramp
2) An almost consistent sea level rise was the major control on deposition in the MacKenzie Basin. The facies of the formations in the MacKenzie Basin reflect a steady Devonian eustatic sea level rise, with punctuated periods of deepening.

3) The Horn Plateau Formation buildups grew continuously until the Late Givetian.
4) Carbonate deposition was continuous in the Givetian in basins that were directly connected to the open ocean. Restricted basins, such as the Western Canadian Sedimentary Basin, display evidence of relative sea-level drop, which can be attributed to a tectonic influence.
5) The MacKenzie Basin is ideal type of basin for evaluating changes in global sea level because it was continuously connected to the open ocean. The strata examined are near the basin edge where minor changes are reflected in the sedimentological record, and the deepening events that were recorded can be correlated on a global scale.

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CHAPTER 3: ECOLOGICAL CONTROLS ON DEVONIAN STROMATOPOROID-DOMINATED AND CORAL-DOMINATED REEF GROWTH[1]

Introduction

The Devonian period saw the most extensive reef development on the planet (Copper 1994). Today, these reefs are important in Western Canada because they host vast petroleum resources. The reefs in the Keg River Formation, Swan Hills Formation, Leduc Formation, and Nisku Formation, for example, are some of Alberta's largest hydrocarbon reservoirs (Moore 2001). Despite their importance, the ecological requirements for Paleozoic coral and stromatoporoid reef growth such as, light, depth, and especially nutrients are largely unknown. Many of the Devonian reefs in the Western Canadian Sedimentary Basin (WCSB) were dominated by stromatoporoids. Tabulate and rugose corals are usually a minor component of Devonian buildups but they were not typically the dominant reef-builders. The scarcity of coral-dominated reefs has caused difficulty in assessing the ecological controls that support coral versus stromatoporoid reef growth during the Devonian. The Horn Plateau Formation, found in the MacKenzie Basin in the Northwest Territories, Canada (Figure 3-1), is composed of isolated reef buildups that are aligned along a northeast-southwest trend that stretches over 300 km. In the northeastern part of the trend, the reefs are coral-dominated whereas reefs in the southwestern part of the trend are stromatoporoid-dominated. This provides an ideal opportunity for comparing both reef types with specific attention being paid to the ecological conditions that influenced their growth.

The general setting of the reefs in the Horn Plateau Formation is

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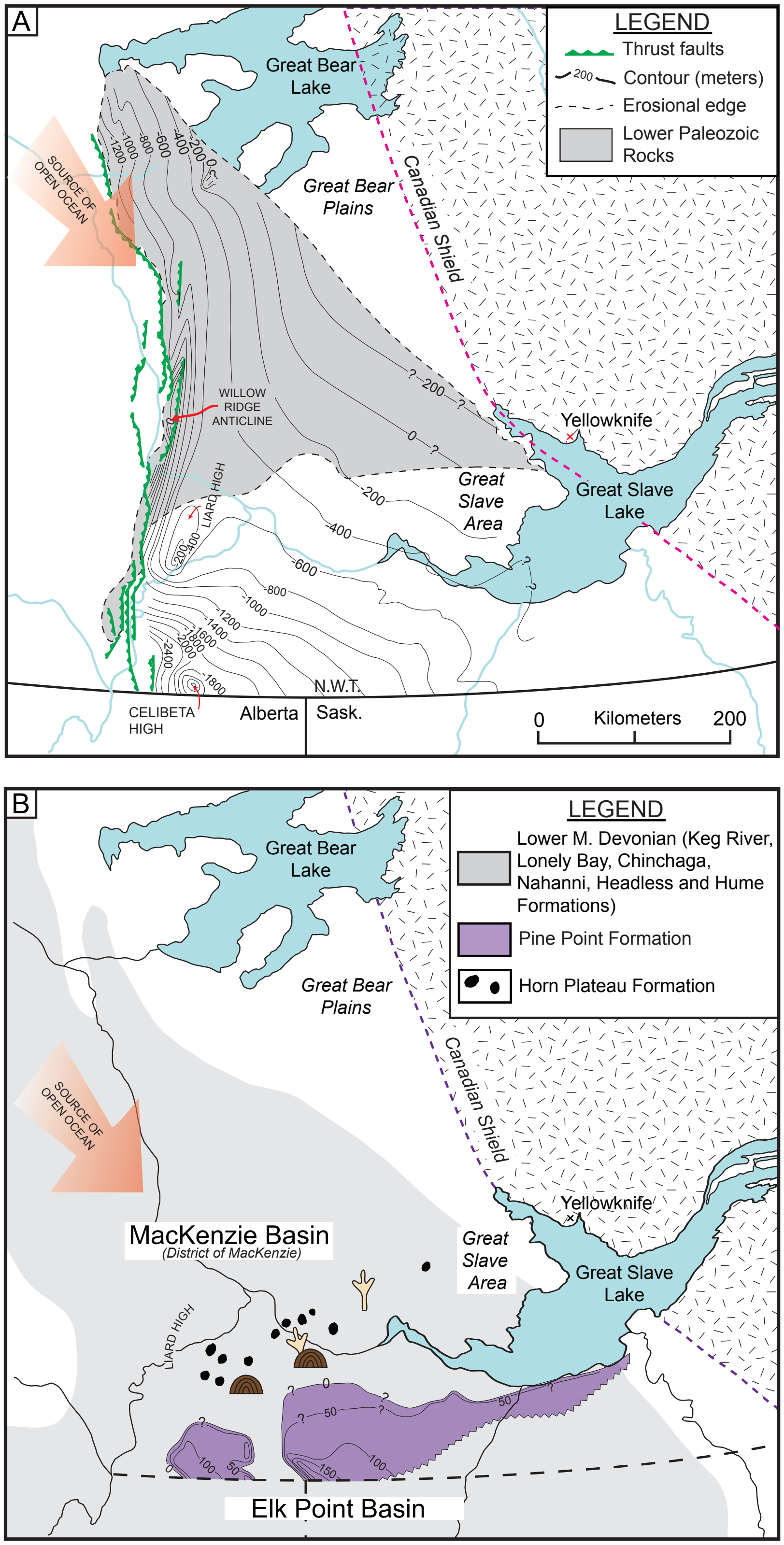

Figure 3-1. Location map (adapted from Drees 1993). (A) Structural map of the pre-Devonian surface in the District of MacKenzie. (B) Locations of the Horn Plateau Formation reefs and boundary of the MacKenzie Basin.

established from detailed stratigraphic and facies analyses that are based on cores that came from numerous wells throughout the area of study. Evaluation of other parameters such as salinity, oxygen levels and particularly, nutrient levels are more difficult because there are no reliable and proven proxies for their determination. Previous research assessing the nutrient requirements of Paleozoic reef building organisms has revealed the need for more than one proxy (Mutti and Hallock 2003) to evaluate nutrient levels and that integration of a stratigraphic framework with geochemical data could lead to a better understanding of the relationship between nutrients and ancient reef growth (MacNeil 2008).

In this study, the possibility of using isotopes in combination with rare earth elements (REE) to resolve paleoecological constraints is tested by examining their distributions and variations relative to the paleography determined from the detailed stratigraphic and facies analyses. This approach suggests that nutrient levels that were related to their positions on the ramp and may have contributed to the distribution of the stromatoporoid and coral reefs.

Methods

The database for this study was compiled from core, thin section analysis, stable isotopes ($\delta^{13}C_{(PDB)}$ and $\delta^{18}O_{(PDB)}$), and trace element data.

Drill core and petrography

Over 300 well files were examined in this study to determine the distribution of the Horn Plateau Formation reefs in the District of MacKenzie. Electronic logs revealed 15 wells that contained the Horn Plateau Formation and 14 of these had drill core through the reefs, including five cores that were drilled at the only known exposure of the formation near Fawn Lake, NWT (Figure 3-2). Amoco Canada Petroleum Company Limited drilled the Fawn Lake reef in 1968 and several of the other wells (~5) were drilled in the early 1970s as part

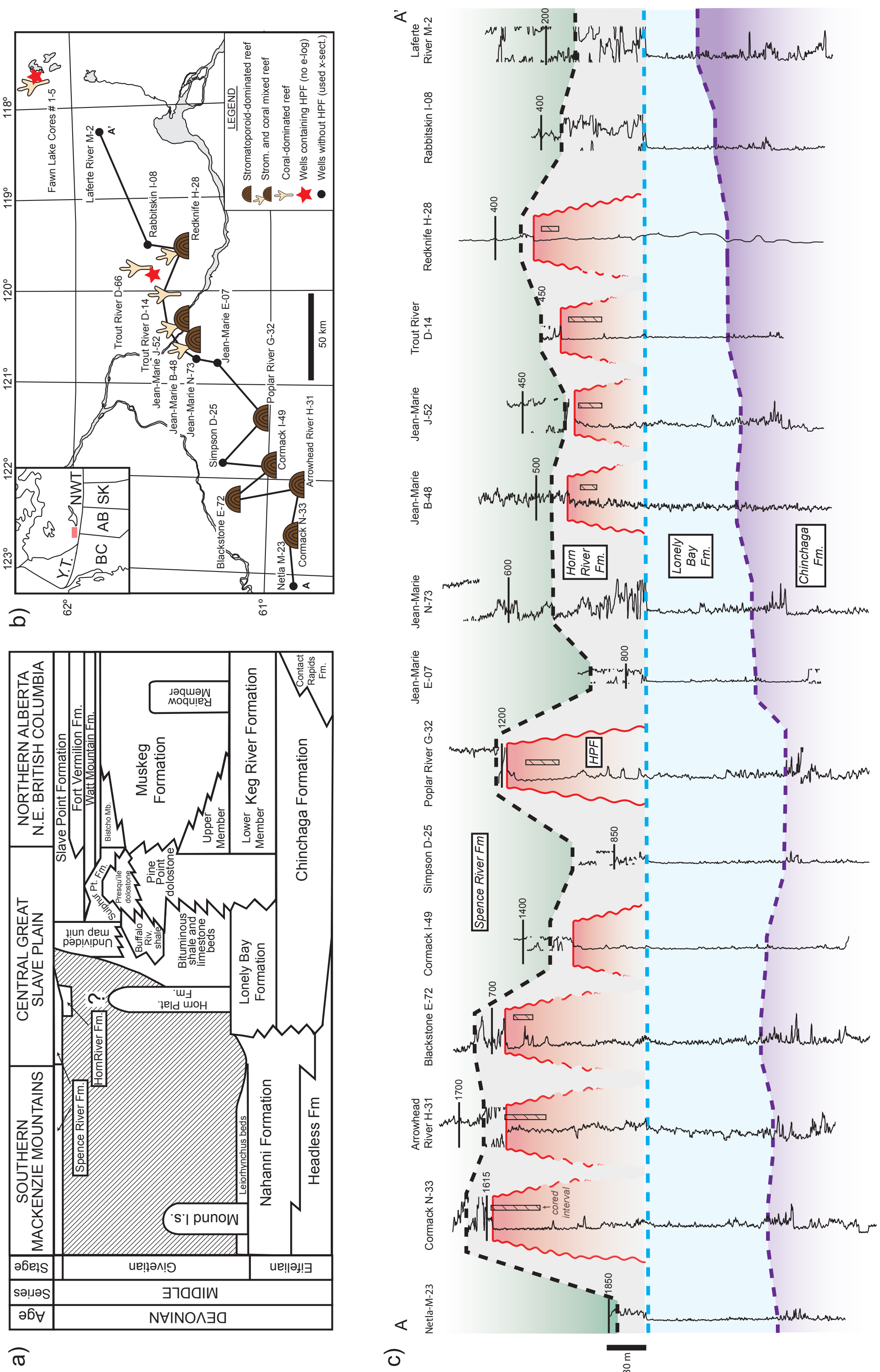

Figure 3-2. Stratigraphy and cross-section of the MacKenzie Basin. (A) Stratigraphy chart of the Great Slave Plain and adjacent areas (adapted from Drees 1993). (B) Map showing wells used in Horn Plateau Formation cross-section A-A'. (C) Cross-section A-A'.

of “Operation Reef”, a co-operative exploration project undertaken by Horn River Resources Ltd. The cores examined in this study are 4 to 10 cm in diameter and 7 to 88 m long. All available cores and 225 thin sections were examined with specific attention given to depositional textures and fossil content in order to determine if environment surroundings controlled the bcomposition of the stromatoporoid-dominated reefs versus coral-dominated reefs.

Sampling – Stable isotope and Rare Earth Element Analyses

Samples used for stable isotope ($\delta^{18}O_{(PDB)}$ and $\delta^{13}C_{(PDB)}$) and trace element analyses were collected from five drill cores from the Fawn Lake reef, housed at the University of Alberta, and nine additional cores from the study area, held at the Geological Survey of Canada, Calgary. Samples for isotope and REE analyses were obtained using a Dremel drill, and carbide drill bits less than or equal to 1 mm in diameter. Samples were taken from each of the major groups of allochems (i.e., corals, stromatoporoids, crinoids, brachiopods), mud matrix, fibrous marine phreatic and blocky, void-filling calcite cements. In all, 150 samples were obtained for stable isotope analysis. Twenty of these powders, from the mud matrix, fossils, and cements analyzed for stable isotopes were also used for trace element analysis.

Data Collection

Stable isotope analysis ($\delta^{18}O_{(PDB)}$ and $\delta^{13}C_{(PDB)}$) was carried out following the method outlined by McCrea (1950). 10 to 50 mg of each sample were placed into glass reaction vessels, which were then placed on a vacuum line so that all gases could be evacuated. Each sample was reacted with 3 ml of H_3PO_4 for at least one hour in a 25º C water bath. The gas was collected on an extraction line and analyzed using a Mat 252 Mass Spectrometer. Error margins were calculated at 0.6‰ for $\delta^{18}O_{(PDB)}$ and less than 0.1‰ for $\delta^{13}C_{(PDB)}$.

Twenty powdered samples (each ~100 mg), ten from the coral-dominated reefs and ten from the stromatoporoid-dominated reefs, were analyzed for 60 trace elements using a NuPlasma Multi-Collector ICP Mass Spectrometer in the University of Alberta's Radiogenic Isotope Facility. The samples were selected from different fossil groups (i.e. stromatoporoids, corals, crinoids), the mud matrix, and blocky calcite cement, for the two different reef types. For comparison to other REE seawater proxy studies, all REE + Y data was normalized to Post-Archean Australian Shales (PAAS) according to McLennan (1989). The REE concentrations for each fossil group were averaged and plotted using a logarithmic scale.

Geological Setting

From the Middle Ordovician to the Late Devonian, much of the Laurentian continent (70-80%) was covered by shallow epeiric seas (Edinger 2002). In North America, several inland basins were created during the Devonian, including the Michigan Basin, Appalachian Basin, Williston Basin, and the Western Canadian Sedimentary Basin (Johnson et al. 1985). In the Early Devonian, the sea encroached onto Western Canada from the northwest (Figure 3-1) and an inland seaway stretched from the Northwest Territories to North Dakota, USA (Fu et al. 2006). The MacKenzie Basin was formed in the Northwest Territories in the area known today as the District of MacKenzie (Figure 3-1). Carbonate deposition began in the Early to Middle Devonian, directly onto the Precambrian basement. Initially, the MacKenzie Basin was bordered by the Canadian Shield to the East, the emergent Tathlina Arch to the south, and the Laurentian Highland to the southwest. Sea level rose in this area almost continuously until the Middle Frasnian. Growth of the reefs, which now form the Horn Plateau Formation, began in the Early Givetian and terminated in the Late Givetian/Early Frasnian

(Corlett and Jones *in press*).

Stratigraphy – MacKenzie Basin

In the District of MacKenzie, shallow, evaporative conditions were prevalent through most of the Early Devonian. The Tsetoe Formation, Mirage Point Formation, Fort Norman Formation, Sombre Formation, and Chinchaga Formation represent the shallow, evaporative conditions that prevailed during that time (Figure 3-2A). These formations are composed largely of dolomitic limestones and anhydrites that formed in supra- and intertidal environments (Corlett and Jones in press). Sea level continued to rise into the Late Eiflian and the MacKenzie Basin ramp began to experience shallow to open marine conditions. The Headless Formation, Hume Formation, Nahanni Formation, and Lonely Bay Formation overlie the older intertidal sediments that were first deposited on the MacKenzie Basin ramp (Drees 1993; Figure 3-2A). Dolomitic limestones of the Lonely Bay Formation covered the southeastern District of MacKenzie. Argillaceous limestones belonging to the Headless Formation, the Hume Formation, and the Nahanni Formation are found west and north of the Lonely Bay Formation (Figure 3-1). These Late Eiflian formations, composed primarily of limestone contain a variety of fossils (Corlett and Jones in press). They record the change from intertidal evaporative conditions to shallow, and eventually open marine conditions that existed on the MacKenzie Basin ramp in the Middle Devonian.

Reefs, belonging to the Horn Plateau Formation, which began their growth in the Early Givetian (Fuller and Pollock 1972; Corlett and Jones in press), are rooted on the top of the Lonely Bay Formation (Figure 3-2C). The reefs, 52-113 m high, are aligned along a northeast-southwest trend that stretches over 350 km. The Horn Plateau Formation reefs are encased in shale of the Frasnian Horn

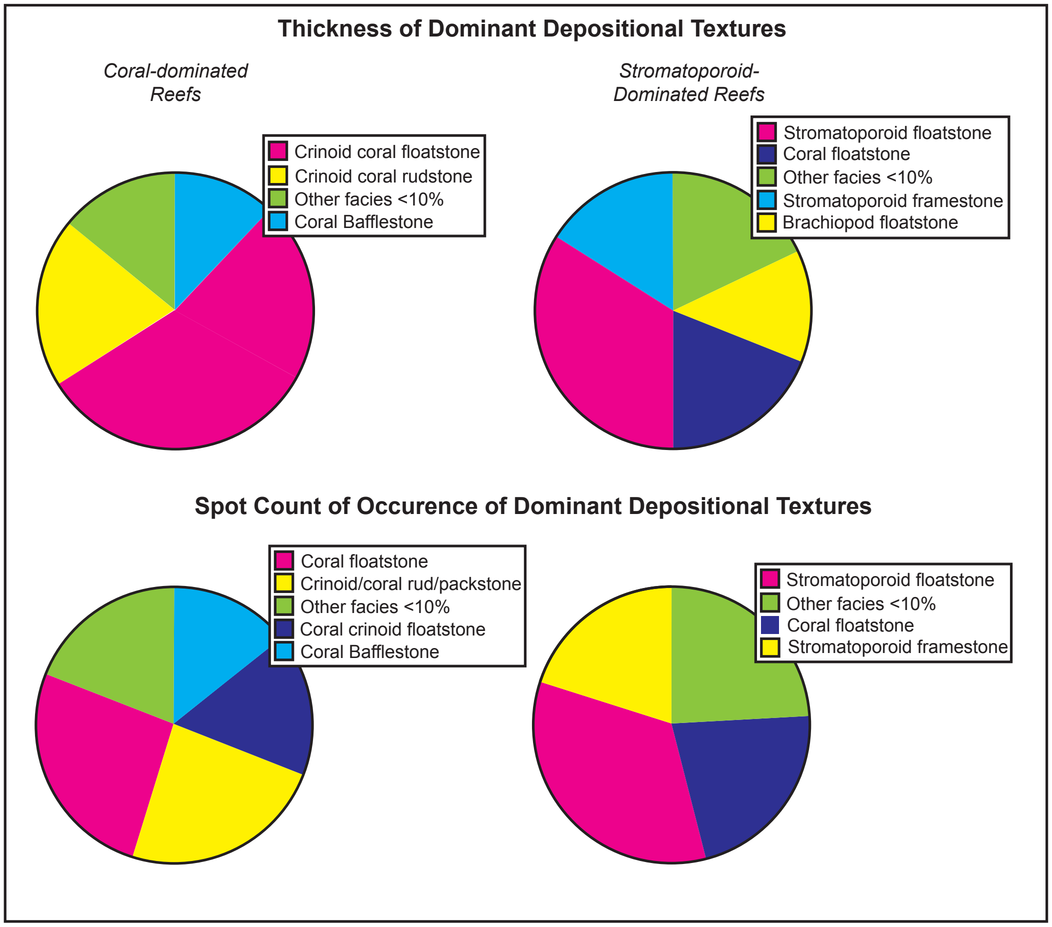

Figure 3-3. Thickness and relative distribution of Horn Plateau Formation facies.

River Formation (Figure 3-2A and C) that is, in turn, overlain by the Spence River Formation shales. Deposition of pelagic shales continued into the Late Devonian in the MacKenzie Basin.

Results

Reefs in the Horn Plateau Formation are dominated by stromatoporoids in the southwest and corals in the northeast (Figure 3-2b). Both reef types which contain diverse fossil assemblages, are formed of several reoccurring facies (Figure 3-3).

Stromatoporoid reefs – Stromatoporoid floatstone facies

This facies, the most common facies in the stromatoporoid-dominated reefs

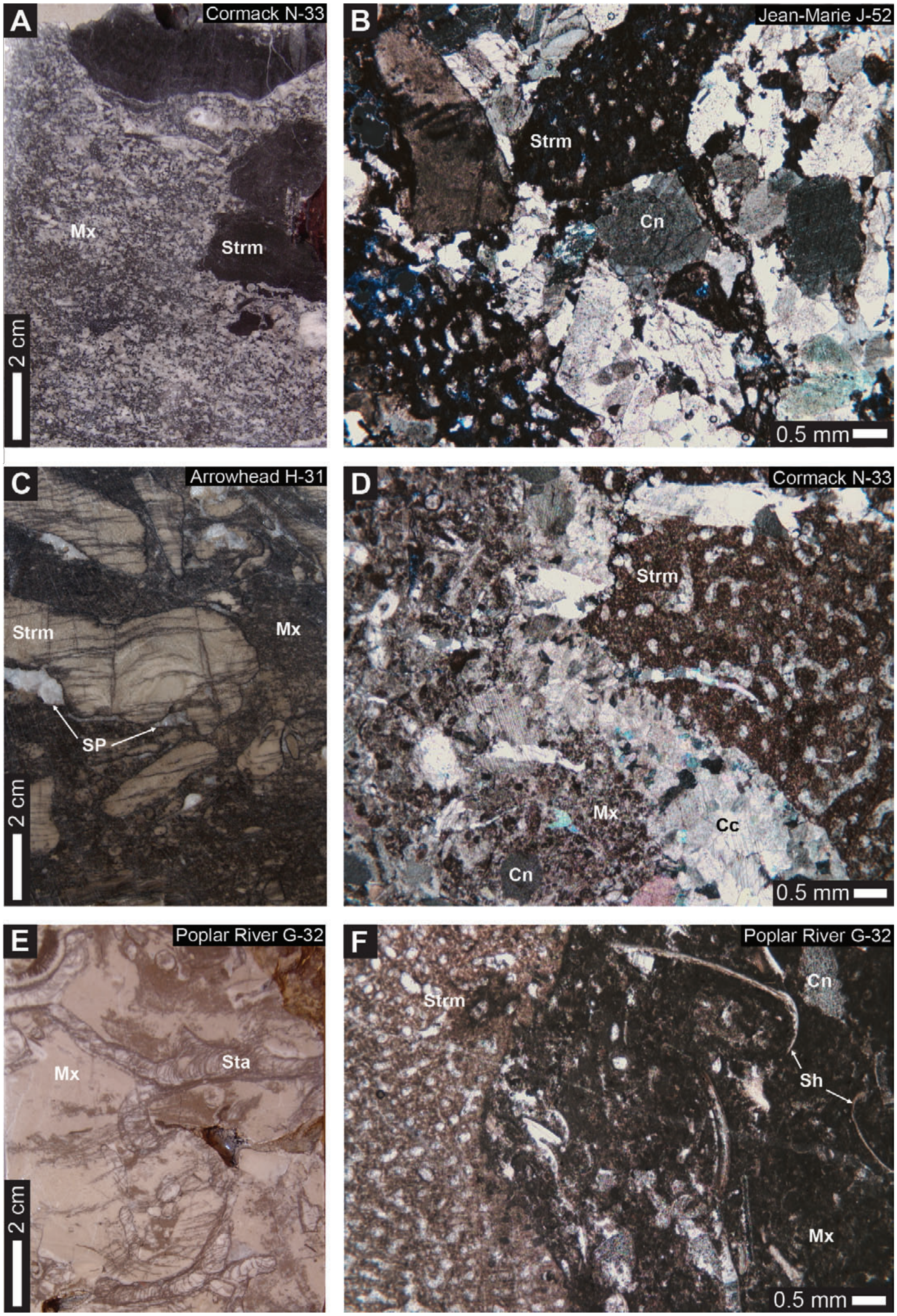

Figure 3-4. Stromatoporoid floatstone facies from the stromatoporoid-dominated reefs in the Horn Plateau Formation. (a), (c), and (e) drill core photographs. (b), (d), and (f) thin section photomicrographs. Mx = crinoid wackestone matrix, Strm = stromatoporoid, SP = shelter porosity, Cc = calcite cement, Cn = Crinoid, Sh = shell fragments, Sta = *Stachyodes*.

(Figures 3-3 and 3-4), contains a diverse assemblage of fossils that are held in a dark grey to brown lime mud matrix. In decreasing abundance these fossils are: stromatoporoids, brachiopods, crinoids, rugose corals, *Amphipora*, trilobite fragments, and cephalopods. Some of the bulbous and branched stromatoporoids are in situ whereas others are fragmented and no longer in growth position. The brachiopods are largely articulated and intact whereas the crinoids and trilobites are fragmented and disarticulated. Locally, the matrix is dominated by small (5 mm diameter) crinoid ossicles.

This facies is mud-supported, which suggests a quiet environment. The fragmentation and disarticulation of the some of the fossils, however, may indicate high-energy intervals. Larger allochems, such as the bulbous stromatoporoids and the rugose horn corals, must have been transported during these episodes of higher energy levels. Overall, the stromatoporoid floatstone facies represents a time when reef growth had slowed or paused and quiet conditions allowed for settling of mud and fine materials. These intervals of slow sedimentation were, however interrupted by brief high-energy events.

Stromatoporoid reefs – Coral floatstone with crinoid packstone

The coral floatstone facies is formed of corals held in a beige to brown limestone matrix that is composed largely of crinoid debris (Figure 3-5). The fossils in this facies, in decreasing order of abundance, are: rugose corals, stromatoporoids (bulbous and S*tachyodes*), *Amphipora*, crinoids, brachiopods, tabulate corals, trilobite fragments, and cephalopods. Although the fossils in this facies are not in situ and are usually fragmented, some intervals do contain well-preserved fossils (Figure 3-5C). In most intervals of this facies, the allochems are supported by a crinoid sand. With so many large, fragmented allochems, crinoid debris, and lack of mud, this represents a reef rubble facies deposited during

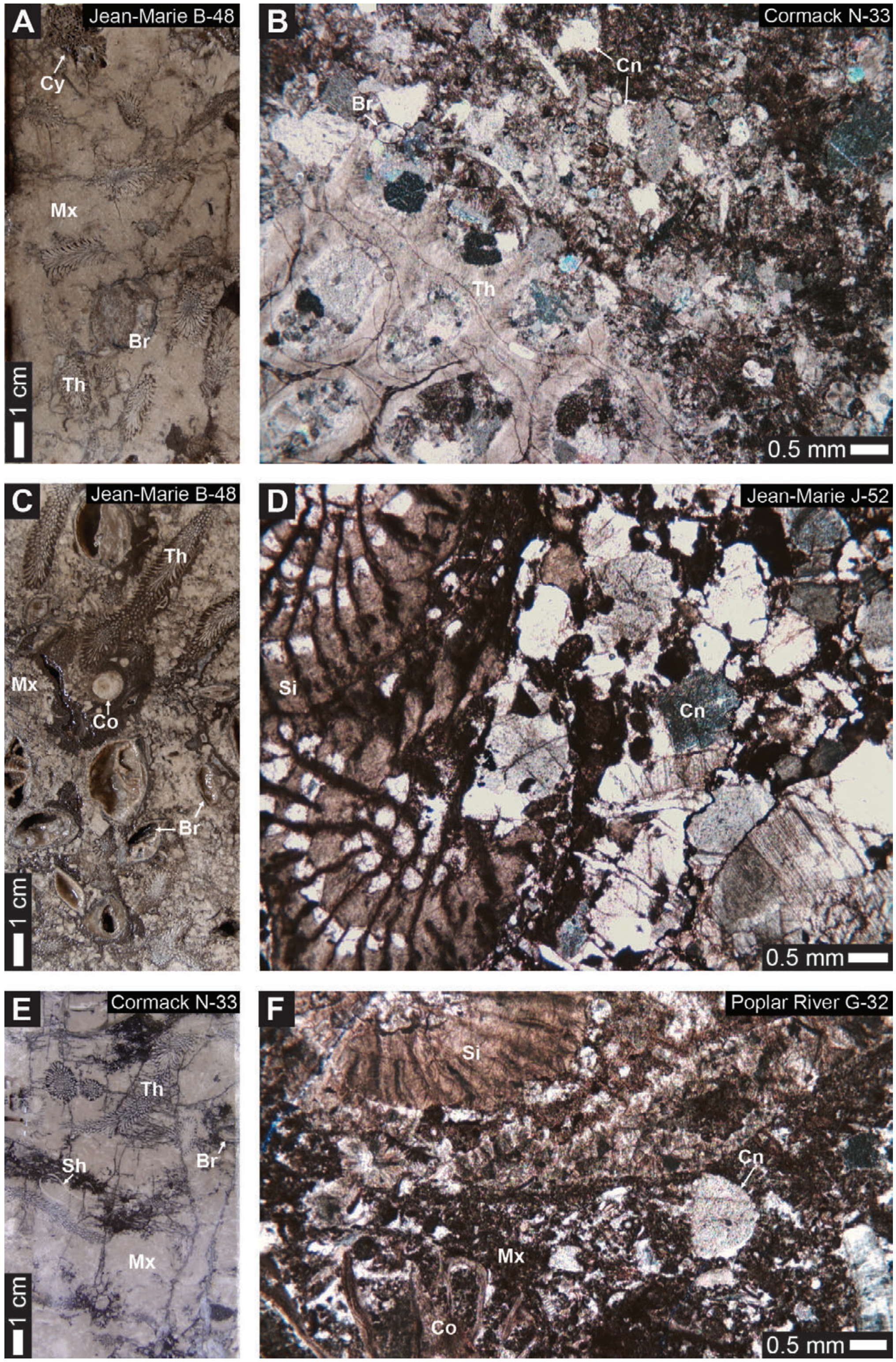

Figure 3-5. Coral floatstone facies from the stromatoporoid-dominated reefs in the Horn Plateau Formation. (a), (c), and (e) drill core photographs. (b), (d), and (f) thin section photomicrographs. Mx = crinoid wacke to grainstone matrix, Strm = stromatoporoid, Cn = crinoid, Br = brachiopod, Sh = shell fragments, Cy = Cystiphylloides, Th = *Thamnopora*, Si = *Siphonophrentis*, Co = unknown coral.

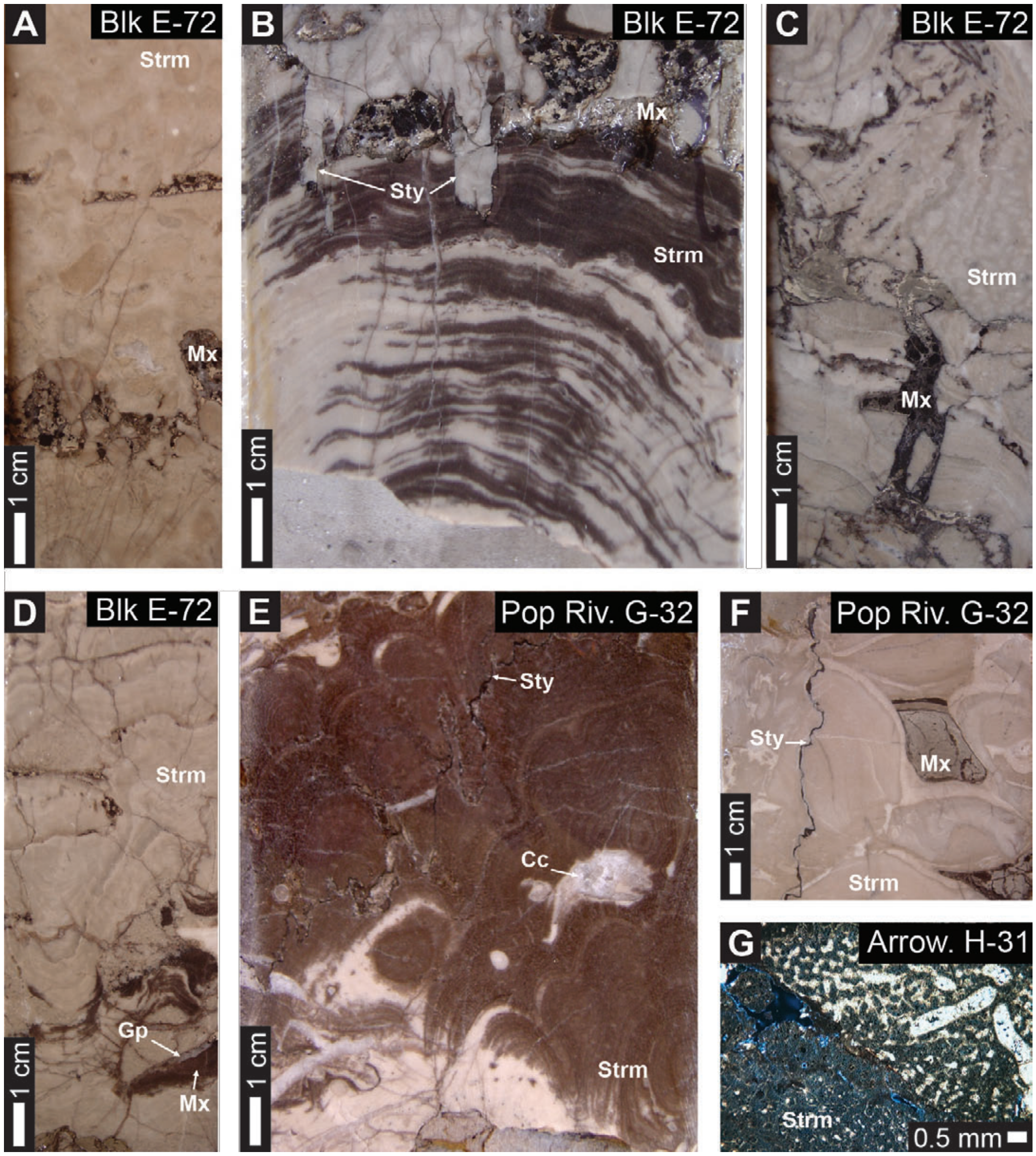

Figure 3-6. Stromatoporoid framestone facies from the stromatoporoid-dominated reefs in the Horn Plateau Formation. (a) through (f) drill core photographs. (g) thin section photomicrograph. Strm = stromatoporoid, Mx = mud matrix, Cc = calcite cement, Sty = stylolites, Gp = geopetal cement.

intervals of very high energy, possibly stormy conditions.

Stromatoporoid reefs – Stromatoporoid framestone

This facies is formed largely of bulbous stromatoporoids that have coalesced into a framework, along with scattered tabular and encrusting stromatoporoid forms (Figure 3-6). The framework porosity is filled with grey or beige lime mud matrix and geopetal cement. Fragmented fossils in the matrix include: *Stachyodes*, rugose corals, *Amphipora*, and crinoids. Some intervals in this facies contain fragmented stromatoporoids held in a mud matrix.

The framework, built by the stromatoporoids, is indicative of a reef environment. Brecciation within the framework indicates a high-energy environment. This facies, composed of large bulbous stromatoporoids, represents the organic core of the stromatoporoid-dominated reefs.

Coral Reefs – Coral floatstone with wacke-packstone matrix

The most common facies in the coral-dominated reefs in the Horn Plateau Formation is coral floatstone with a crinoid wackestone to packstone matrix (Fig 3-3 and 3-7). This matrix-supported facies is usually light grey or beige. Allochems are 1 mm to 5 cm long and include, in decreasing order of abundance: tabulate and rugose corals, crinoid ossicles (up to 3 cm diameter), brachiopods, stromatoporoids (branching *Stachyodes* and bulbous), *Amphipora*, calcispheres, and bryozoans. The diverse coral assemblage in this facies includes: *Thamnapora*, *Hexagonaria*, *Disphyllum*, *Stringophyllum redactum* McLaren, and *Syringopora*. Generally, the fossils are intact, and locally in situ.

This matrix-supported facies, with most fossils in situ indicates intervals of quiescence during the growth of the coral-dominated reefs.

Coral Reefs – Coral rudstone

The coral rudstone facies is grain-supported and contains a diverse array of fossils (Figure 3-8). The matrix, if present, is usually dark grey or beige and is locally composed entirely of coral debris. The allochems are 1 mm to 10 cm long and include, in decreasing order of abundance: corals (*Thamnapora*, *Cyathophyllum*, *Alelophyllum nebracis* McLaren), crinoids (up to 3 cm diameter),

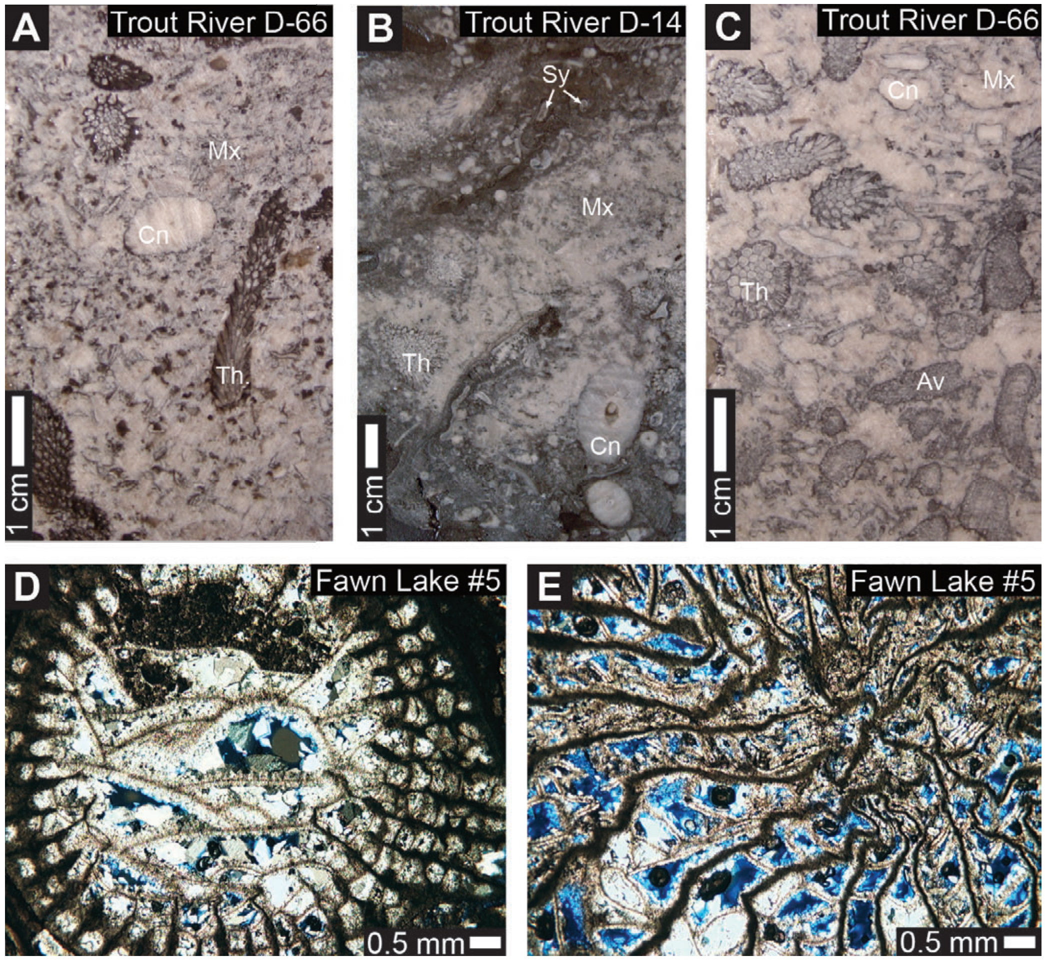

Figure 3-7. Coral floatstone facies from the coral-dominated reefs in the Horn Plateau Formation. (a), (b), and (c) drill core photographs. Th = *Thamnopora*; Av = *Alveolites*; Cn = crinoid; Mx = lime mud-packstone matrix. (d) thin section photomicrograph, *?Cylindrophyllum* (e) *?Heliophyllum*.

brachiopods, stromatoporoid pieces, bryozoans, trilobite fragments, ostracodes, and foraminifera. The fossils are fragmented and disarticulated with none in growth position.

The nature of preservation in this facies indicates a high-energy environment. This facies is interpreted as reef rubble deposited storm events.

Coral Reef – Coral bafflestone

The coral bafflestone facies has the least diverse fossil assemblage of the major facies (Figure 3-9). The branching tabulate coral *Stringophyllum redactum* McLaren are held in by a grey and beige limestone matrix that is mainly

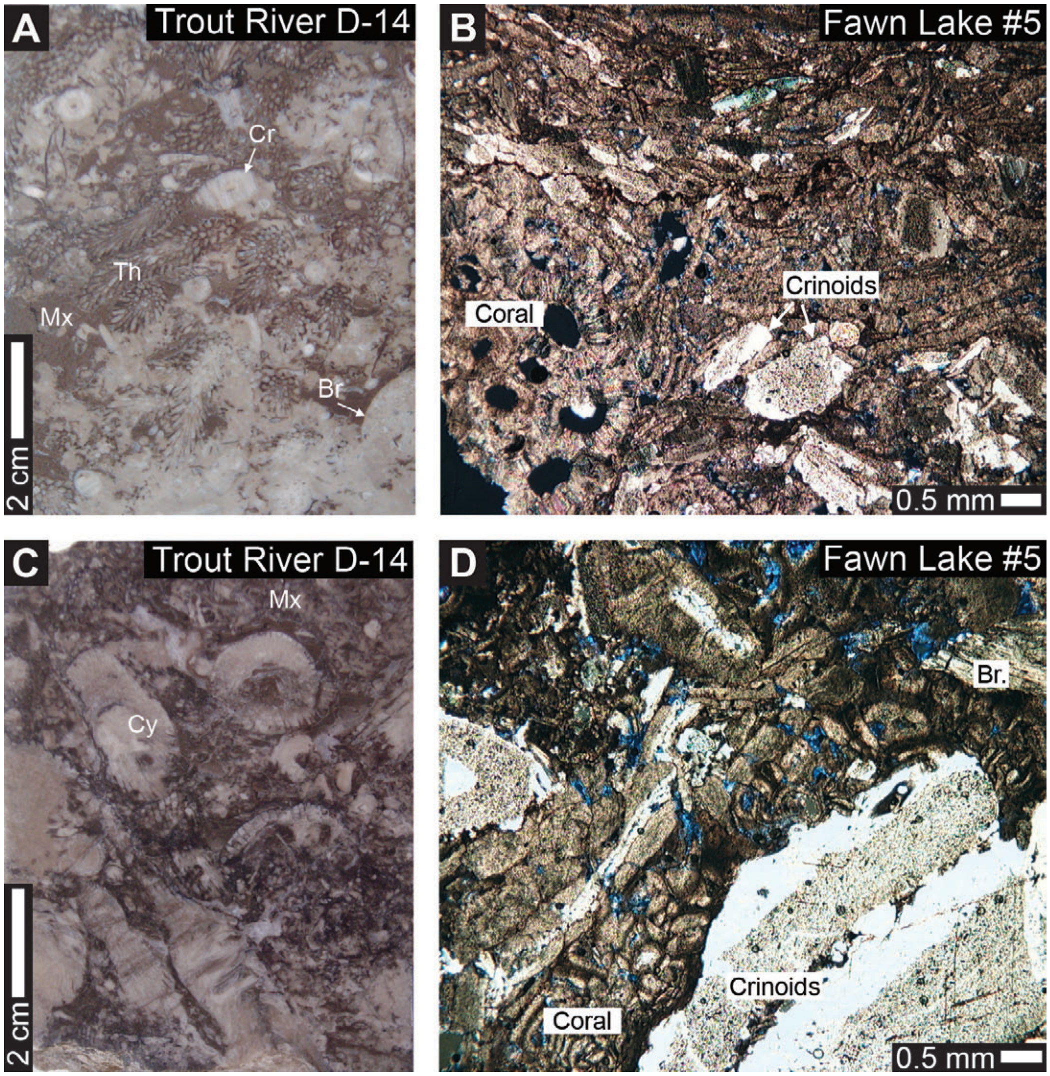

Figure 3-8. Coral rudstone facies from the coral-dominated reefs in the Horn Plateau Formation (*a*) and (*c*) Drill core photographs. (*b*) and (*d*) Thin section photomicrographs. Th = *Thamnopora*, Cy = *Cylindrophyllum*; Br = brachiopod; Cr = crinoid; Mx = crinoid pack grainstone matrix.

composed of lime mud and a few biofragments (Figure 3-9). Small pieces of stromatoporoids are locally present in the matrix between the corals.

The branching corals that dominate this facies would not be capable of withstanding high energy levels. The matrix between the corals contains fossils, which require some energy to transport. A medium level energy regime is required for this type of facies. The branching tabulate corals likely grew as the mud gradually accumulated and surrounded them, since they did not have an effective attachment mechanism.

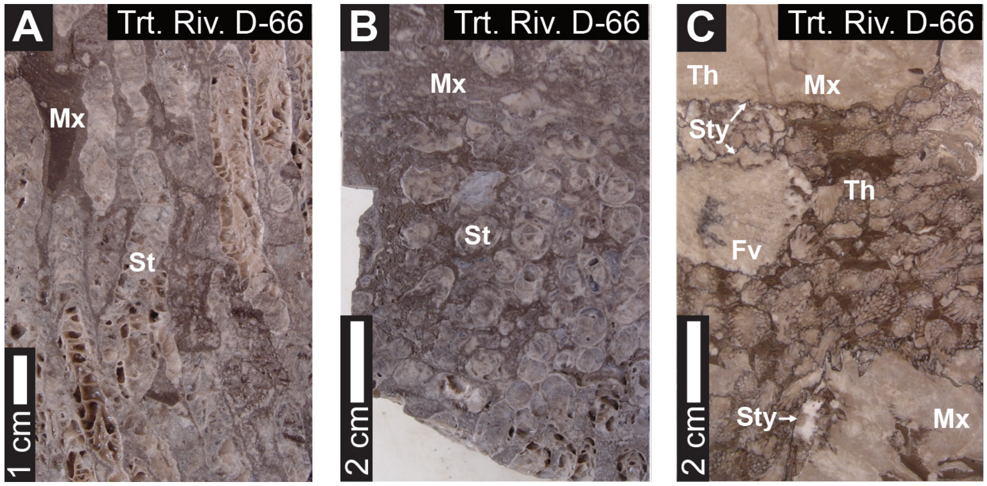

Figure 3-9. Coral bafflestone facies from the coral-dominated reefs in the Horn Plateau Formation. (a), (b), and (c) drill core photographs. St = *Stringophyllum*; Th = *Thamnopora*; Fv = *Favosites*; Mx = lime mud-wackestone matrix, Sty = stylolites.

Stable isotope data

Comparison of the stable isotope data from the coral- and stromatoporoid-dominated reefs shows a slight separation in $\delta^{13}C_{(PDB)}$ and $\delta^{18}O_{(PDB)}$ values (Figure 3-10) of the biogenic components and matrix. The $\delta^{13}C_{(PDB)}$ values are the most divergent with the coral reefs having a range of 0.8 to 6.2‰ and an average value of 2.8‰ whereas the stromatoporoid reefs range from -0.5 to 2.6‰ with an average of 1.6‰ (Figure 3-10). The $\delta^{18}O_{(PDB)}$ values for the coral reefs ranged from -13.5 to -2.3‰ and had an average of -7.2‰ whereas the stromatoporoid reefs had $\delta^{18}O_{(PDB)}$ values between -10.6 to -0.6‰ and an average of -7.8‰ (Figure 3-10).

Cements from the coral reefs range in $\delta^{18}O_{(PDB)}$ from -18.2 to -9.1‰ and in $\delta^{13}C_{(PDB)}$ from 2.1 to 3.4‰. The cements from the stromatoporoid reefs ranged in $\delta^{18}O_{(PDB)}$ from -14.9 to -10.2‰ and in $\delta^{13}C_{(PDB)}$ from -0.8 to 1.1‰ (Figure 3-10). The $\delta^{18}O_{(PDB)}$ values for the cement are very low and significantly lower than the fossil samples, which means that they were not precipitated in seawater.

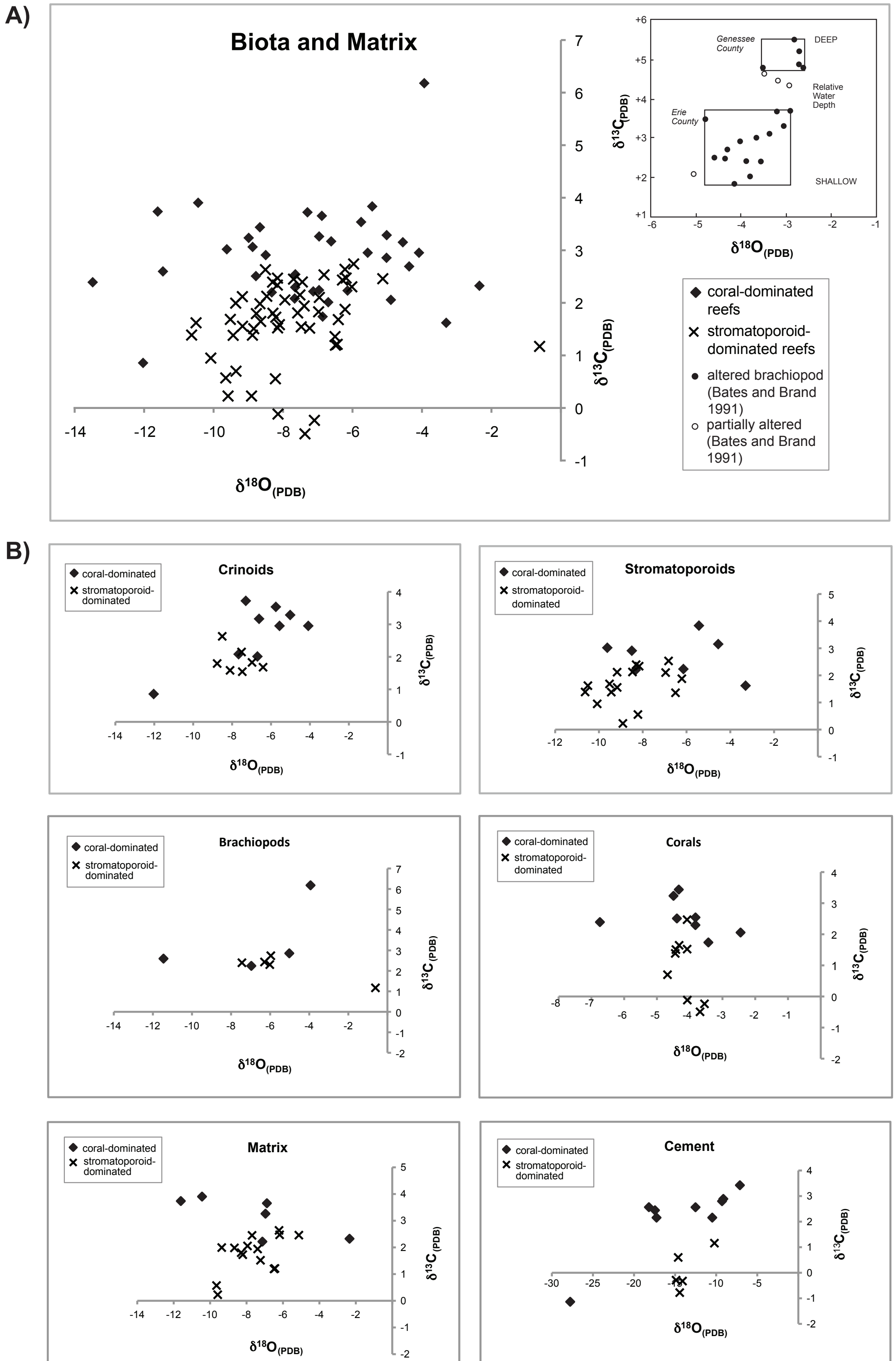

Figure 3-10. Stable isotope data. (A) All fossil and matrix samples. Insert: Devonian reefs from the Appalachian and Michigan Basin (Bates and Brand 1991) (B) Individual plots for each type of sample analyzed.

Rare Earth Elements

All of but one of the biogenic samples analyzed in the coral- and stromatoporoid dominated reefs display patterns that are typical of modern seawater REE +Y patterns (Figure 3-11). Skeletal material from different types of organisms, regardless of being precipitated from the same seawater, will likely have slight variations in their REE pattern due to the partitioning behaviour of different organisms (Webb and Kamber 2000; Northdruft et al. 2004). Overall, these fossil samples are relatively enriched in the heavy REEs (HREEs), which is characteristic of seawater plots. The plots (Figure 3-11) also show strong negative Ce anomalies and slight positive La anomalies that are typically associated with modern seawater plots. The relative enrichment in HREEs, negative Ce anomaly, and slightly postive La anomaly support established parameters for appropriate seawater proxy (Northdruft et al. 2004). One coral sample had a linear REE + Y pattern, which is most likely the result of sample contamination from diagenetic pore cement.

The fossil samples also show a clear split between the coral-dominated and stromatoporoid-dominated reef data sets. Although both reef types have a seawater pattern, the samples from the stromatoporoid-dominated reefs contain higher concentrations of REEs than the coral-dominated reefs (Figure 3-11).

Facies distribution and architecture of the Fawn Lake reef

The only known exposure of the Horn Plateau Formation is near Fawn Lake (Figure 3-2B). The Fawn Lake reef is circular in shape, approximately one kilometer in diameter at its base, and is up to 105 meters high (Vopni 1969). Five boreholes were drilled through the Fawn Lake reef in 1968 by Amoco Canada Petroleum Company Limited (Vopni 1969) to determine its stratigraphic setting and the vertical and lateral extent of the reef.

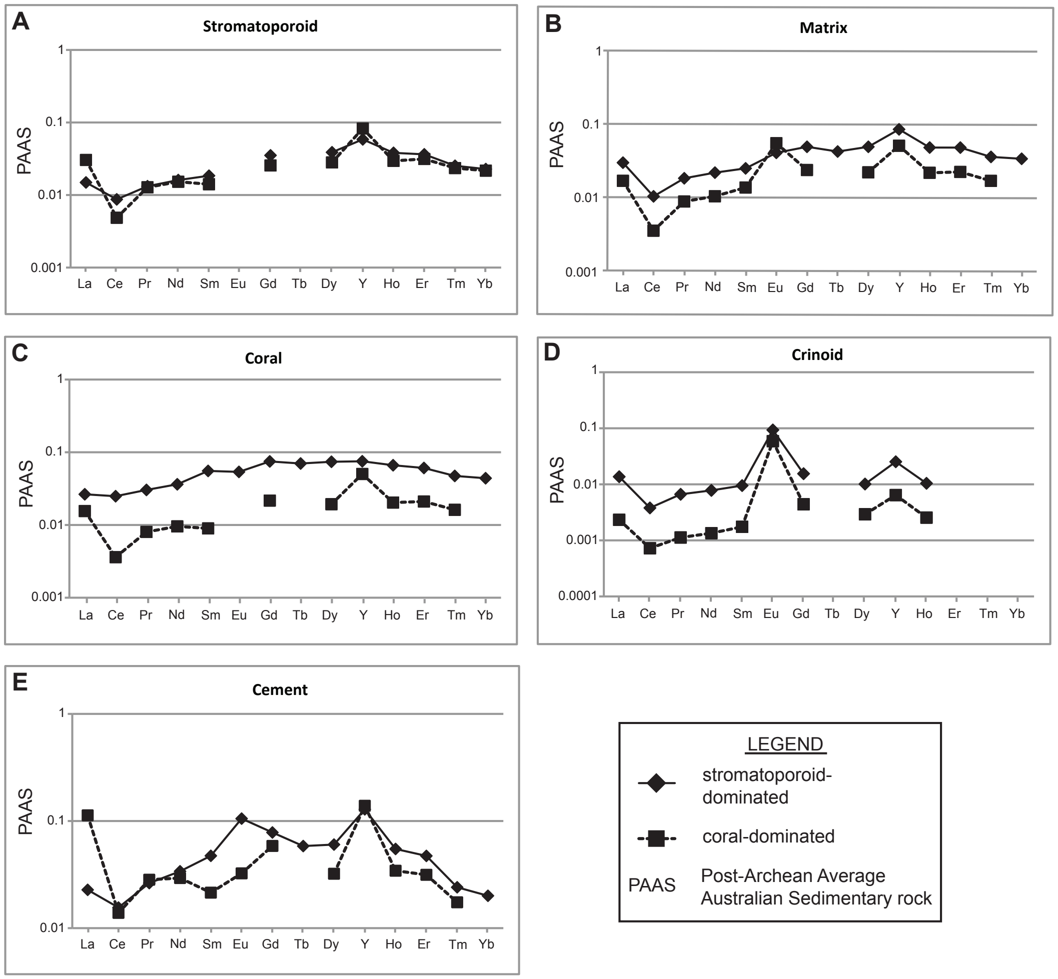

Figure 3-11. Rare earth element plots for each type of sample analyzed. Results shown here are average values for ten samples from the coral-dominated reefs and ten from the stromatoporoid-dominated reefs. (A) Stromatoporoid. (B) Matrix. (C) Coral. (D) Crinoid. (E) Cement.

The three main facies in the Fawn Lake reef are the coral floatstone, coral bafflestone, and the coral rudstone. The high-energy rudstone facies that is composed mostly of large allochems and little mud matrix dominates the northwestern, basinward side of the reef (Figure 3-12). The flank deposits are thickest on the southeast side of the leeward side of the reef (Figure 3-12). The coral bafflestone facies is found in three different intervals in the center and leeward side of the reef. There are two intervals in the Fawn Lake reef that contain numerous stromatoporoids (Figure 3-12). These coral stromatoporoid rudstones to boundstones are on the southeast leeward side of the reef and become thicker near the top of the reef (Figure 3-12).

The coral floatstone with wackestone to packstone matrix is interbedded with intervals of the coral rudstone and bafflestone facies in the Fawn Lake reef. There are three intervals of coral rudstones or bafflestones that are interbedded with the coral floatstone facies (Figure 3-12). A similar pattern of deposition is also evident in coral-dominated Devonian reefs from New York. Isolated reefs in the Eifelian Edgecliff Member of the Onandaga Formation in the Appalachian Basin are composed of branching colonial rugose corals surrounded by and interbedded with crinoidal grainstones and packstones that contain domal favostid corals (Wolosz 1997). The two paleocommunities in the "successional mounds" of the Edgecliff Member are cyclically interbedded and eventually produce "mound/bank reefs" (Wolosz 1992a). In the Fawn Lake reef the coral floatstones with crinoid wacke-packstone matrix are interbedded with the coral bafflestone facies. These coral-dominated reefs may then also be considered to be "mound/bank reefs" caused by successional growth of the two facies. In fact it may have been necessary for coral-dominated reefs in the Paleozoic to have this type of successional growth. Most tabulate and rugose corals lacked an effective attachment mechanism, which made aggradation of the substrate an important factor in Paleozoic coral reef growth (Scrutton 1999).

The Fawn Lake reef is the only one in the Horn Plateau Formation penetrated by more than one drill core and the only one for which core is available for the entire formation. Each of the other reefs are represented by one drill core that typically comes from the upper part of the reef (Figure 3-2C).

Substrate requirements for coral versus stromatoporoid reef growth

Considering the worldwide importance of Paleozoic reefs as petroleum reservoirs, little is known about the ecological controls that influenced the growth of these massive structures. Most information regarding stromatoporoid and coral

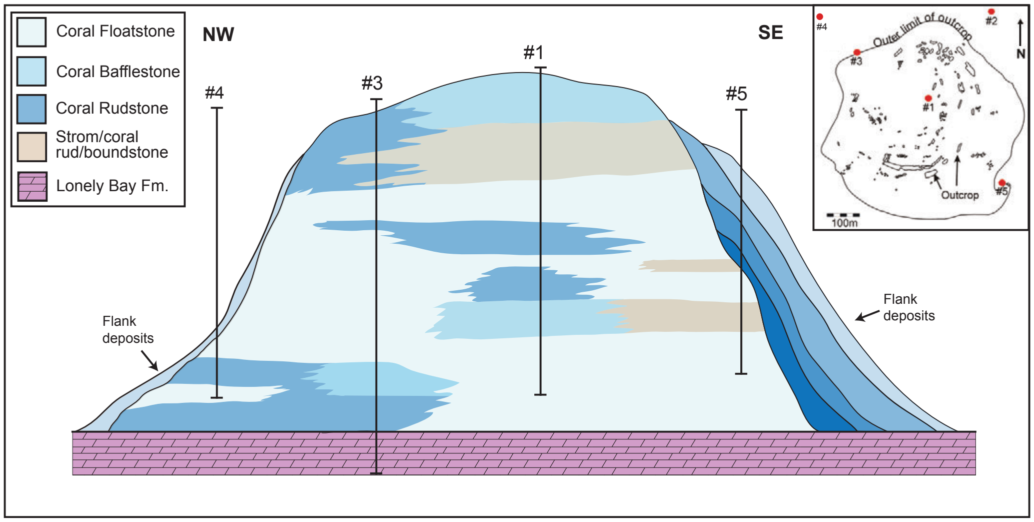

Figure 3-12. Distribution of facies within the coral-dominated Fawn Lake reef.

growth comes from interpretations of the rock record that are based largely on comparison with today's modern reef environment. Tabulate and Rugose corals are commonly compared to Scleractinian corals, but there are many differences between ancient and modern corals especially in their choice of substrate, attachment mechanisms, and their role in reef construction (Scrutton 1999).

Most Paleozoic corals lack an efficient attachment mechanism. Two of the three dominant facies in the Horn Plateau Formation coral-dominated reefs are formed largely of mud. The soft substrate, indicated by the high mud content in the coral-dominated reefs, provided support for the fasciculate (branched) tabulate corals common to the Horn Plateau coral-dominated reefs (cf. Shen and Zhang 1997, Woloscz 1997). These types of corals could cope with fast sedimentation rates and it seems that these corals grew with only a small portion of the branched corallites above the sediment water interface (Shen and Shang 1993; Wood 1993; Wolocz 1997; Scrutton 1999). The spaces between the branched corallites were filled with sediment and anchored the corals in the substrate. This kept the corals from toppling over or becoming dislodged during episodes of high-energy (e.g., wave energy or storm events). The layered or successional pattern of growth in the Fawn Lake reef (Figure 3-12), interbedded coral bafflestone and matrix-dominated coral floatstone, may imply that there were times when the sedimentation rates overcame coral growth. In contrast, the stromatoporoid-dominated reefs contained little mud. The stromatoporoid framestone facies, considered the core organic reef facies, contained bulbous stromatoporoids that encrusted and grew on top of one another. There is only a small amount of mud in the framework porosity, and there is no indication that this mud was necessary to support the stromatoporoid reef growth.

Evidence from drill core and petrographic analysis indicates that the coral and stromatoporoid reefs may have grown in areas where there were different substrates and different sedimentation rates.

Discussion

A map of the pre-Devonian structure of the MacKenzie Basin (Figure 3-1A) shows that the Canadian Shield was exposed in the east-northeast with the basin floor sloping to the southwest (Corlett and Jones in press). The Horn Plateau Formation reefs are aligned northeast to southwest (Figure 3-1B) with the coral-dominated reefs proximal to the shoreline and the stromatoporoid-dominated reefs located lower on the ramp, in a more basinward location. The paleogeography, inferred from detailed stratigraphic and facies analyses of the succession indicates that the stromatoporoid reefs in grew in deeper water than the coral-dominated reefs (Figure 3-13).

The difference in the $\delta^{18}O_{(PDB)}$ and $\delta^{13}C_{(PDB)}$isotope data (Figure 3-10) from the two reef types may be a reflection of their paleogeographic settings. It is, however, dangerous to base such interpretations solely on these isotopes because they may reflect many different factors including vital effects, water temperature, and diagenetic alteration (e.g., Knauth and Epstein, 1976; Muehlenbachs and Clayton, 1976; Land 1986; Karhu and Epstien, 1986; Veizer et al., 1999).

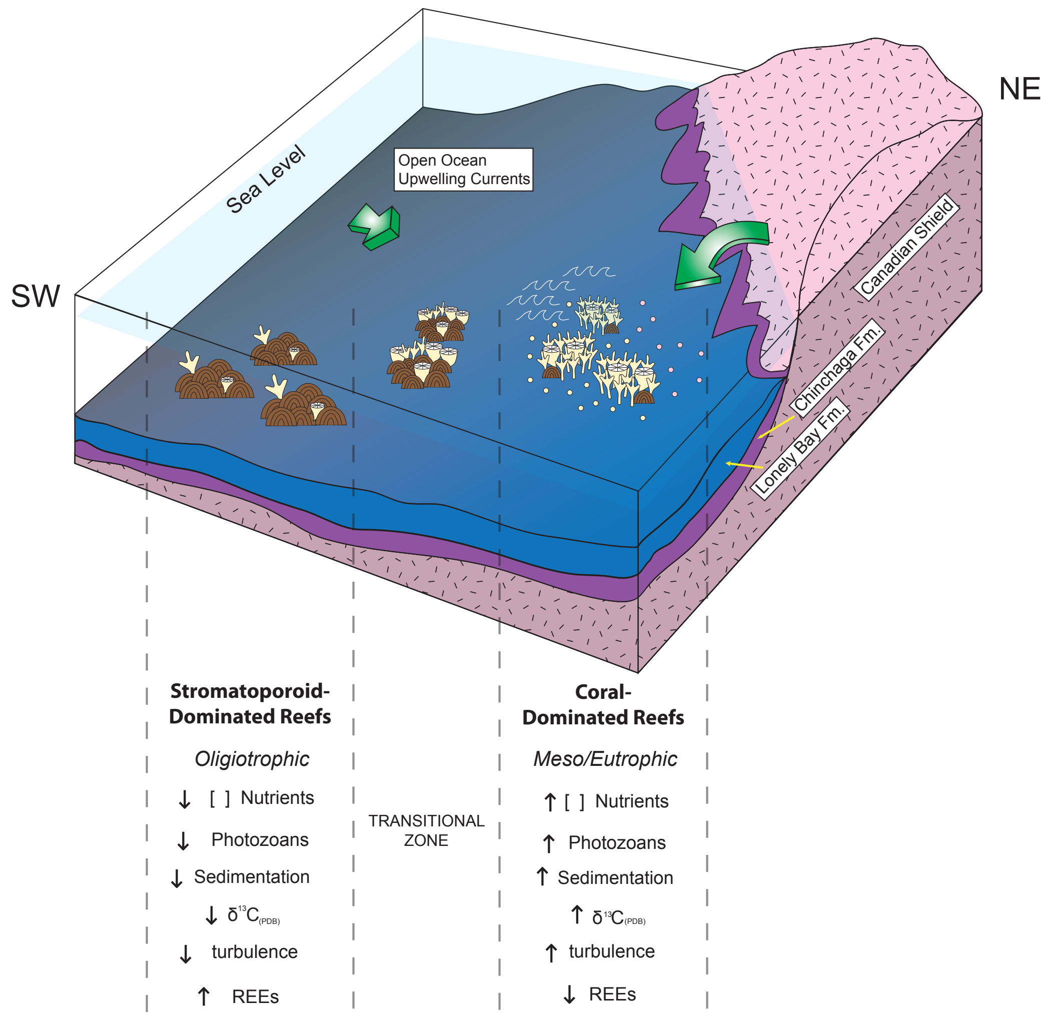

Figure 3-13. Schematic diagram of the MacKenzie Basin ramp illustrating the position of the coral- versus stromatoporoid-dominated reefs and associated paleoenvironmental conditions.

Nevertheless, these data can be used in combination with rare earth element data that were obtained from the stromatoporoid- and coral-dominated reefs. This approach is feasible because it has recently been demonstrated that rare earth elements (plus Yttrium; REE + Y) in limestones can be used to assess ancient seawater and paleoceanographic reconstruction (Shaw and Wasserburg 1985; Webb and Kamber 2000; Northdruft et al. 2004; Tanaka et al., 2003). The REEs are ideal because the trivalent elements (all but Ce and Eu, dependant on oxygen levels) demonstrate uniform behaviour in seawater (Northdruft et al. 2004). Past studies using REEs as a seawater proxy have found that fossil material, mud matrix, and marine cements all display a modern-day seawater REE +Y patterns with a positive La^{3+} anomaly, a negative Ce^{3+} anomaly, and depleted light rare earth elements (LREE) relative to the heavy rare earth elements (HREE; Webb and Kamber 2000; Northdruft et al. 2004). It now appears that the input of salinity, water depth, and oxygenation levels can be interpreted from REE concentrations in seawater and limestone (Elderfield 1988; Pipegrass and Jacobson 1992; Betram and Elderfield 1993; Greaves et al. 1999; Northdruft et al. 2004). Diagenesis seems to have little effect on the REE patterns of marine carbonates (Grandjean et al. 1987; Banner et al. 1988; Northdruft et al. 2004). The REE can, therefore, be used in combination with stable isotope data to determine certain physiochemical conditions of past environments.

REE concentrations in fossils from the coral- and stromatoporoid-dominated reefs have patterns that are consistent with modern seawater with normal salinity and oxygen levels. The concentrations of trivalent REEs are higher in the stromatoporoid-dominated reefs than in the coral-dominated reefs (Figure 3-11). Past studies examining the behaviour of the REEs in seawater and limestones indicate that all of the REEs (except Ce and Eu, dependent on oxidation levels) increase in concentration with water depth (Grandjean et al. 1987; Elderfield et al. 1990; Holser 1997; Kamber and Webb 2001). This REE enrichment occurs within even first hundred meters of seawater (McLennan 1989). Since the samples collected from stromatoporoid-dominated reefs contained higher REE concentrations than the coral-dominated reefs, a distal, deeper water depositional environment is indicated for the stromatoporoid-dominated reefs. This is consistent with the local paleogeography and with this depositonal framework (Figure 3-13).

The $\delta^{18}O_{(PDB)}$ values from the coral-dominated reefs (-7.2‰) are slightly heavier than those from the stromatoporoid reefs (-7.8‰; Figure 3-10). The notion that this may indicate a cooler, basinward or deeper environment for the coral reefs (cf. Emiliani 1955; Bates and Brand 1991) is converse to the known paleogeographic framework. The higher $\delta^{18}O_{(PDB)}$ values in the coral reefs might, however, reflect coastal upwelling whereby shallow coastal waters moves basinward and are replaced by deeper, upwelling nutrient-rich waters (Stanton 2006). Under this scenario, the slightly heavier $\delta^{18}O_{(PDB)}$ values from the coral-dominated reefs would be attributed to cooler waters brought into the coastal areas by upwelling (Emiliani 1955; Bates and Brand 1991).

The $\delta^{13}C_{(PDB)}$ values for the coral-dominated reefs (2.8‰) in the Horn Plateau Formation are heavier than the stromatoporoid-dominated reefs (1.6‰; Figure 3-10). Heavier $\delta^{13}C_{(PDB)}$ values generally imply surface productivity (Surge et al. 1997; Edinger 2002). During photosynthesis, the photoautotrophs (e.g. phytoplankton) preferentially remove the lighter C^{12} isotope from the CO_2 in seawater, thereby producing positive $\delta^{13}C_{(PDB)}$ values (Surge et al. 1997; Peeters et al. 2002). The coral-dominated reefs, that are located closer to the shoreline than the stromatoporoid-dominated reefs reflect the increased $\delta^{13}C_{(PDB)}$ values because the organisms in the reef precipitated their skeletons from ambient sea water that has been relatively increased with respect to C^{13}. The $\delta^{13}C_{(PDB)}$ values in the samples from coral-dominated reefs could be a reflection of increased nutrients, since the photoautotrophs responsible for the relative increase in $\delta^{13}C_{(PDB)}$ (Stanton 2006) utilize nutrients (e.g. nitrogen, phosphorous, and iron) for reproduction and protein production (Mutti and Hallock 2003; Jones 2010).

Stable isotope data from coral-dominated reefs in the Onandaga Formation in the Appalachian Basin and stromatoporoid-dominated reefs from the Amherstberg Formation in the Michigan Basin (Bates and Brand 1991) are characterized by patterns that are similar to those from the Horn Plateau

Formation reefs in the MacKenzie Basin (Figure 3-10A - insert). Reefs in the Onandaga Formation are composed mostly of phaceloid (branching) and dendroid corals whereas reefs in the Amherstburg Formation (Endinger 2002) are dominated by tabular and laminar stromatoporoids. The coral-dominated reefs in the Appalachian Basin (Onandaga Formation) have higher $\delta^{13}C_{(PDB)}$ (+ 5.01‰) and $\delta^{18}O_{(PDB)}$ (- 2.85‰) values than the stromatoporoid-dominated reefs from the Michigan Basin (Amherstburg Formation ; $\delta^{13}C_{(PDB)}$: + 2.79‰ and $\delta^{18}O_{(PDB)}$: - 3.83‰; Figure 3-10A – insert; Bates and Brand 1991). The differences in values between the two reef types were attributed to a cooler, deep-water environment for the coral-dominated reefs (Bates and Brand 1991; Endinger 2002). This interpretation, however, did not agree with the paleogeography of the study areas as both reef types grew in shallow water environments (Wolosz 1997). The Appalachian Basin was connected to the open Rheic Ocean to the southwest, which allowed for upwelling of deeper nutrient-rich water into shallow proximal environments, where the coral-dominated reefs were growing (termed "quasi-estuarine circulation" by Endinger 2002). Coastal upwelling resulted in apparent cooler water conditions, reflected in higher $\delta^{18}O_{(PDB)}$ values for the coral-dominated reefs in the Edgecliff Member. High $\delta^{13}C_{(PDB)}$ in the coral-dominated reefs in this formation were attributed to primary productivity as a result of the increased nutrients brought up by coastal upwelling (Endinger 2002).

At present, there is no direct proxy for determining past nutrient levels in carbonate rocks because oceanic biogenic particulate matter does not incorporate many of the trace elements associated with nutrients in seawater in direct proportion in their skeletons (Chester 1990). There is also no established method of determining the source of nutrients in carbonate rocks. Nutrients are sourced from coastal upwelling, terrestrial runoff, dissolution of coastal carbonates, or storm disturbance of offshore oceanographic systems (Jones 2010). In this study, higher nutrient levels are implied by higher $\delta^{13}C_{(PDB)}$ values for the coral-dominated reefs that were located closer to the mainland. Being closer to the exposed Canadian Shield, the increased nutrients and photoautotroph production in the area where the coral-dominated reefs were growing may have been sourced from terrestrial runoff or groundwater seepage on the ramp (Figure 3-13). Both of these are viable mechanisms for nutrient increase (Stanton 2006); however, areas that are experiencing a large amount of river runoff or ground water seepage are characterized by lower salinities resulting in low carbonate accumulation and, in the case of seepage, dissolution (Mutti and Hallock 2003; MacNeil 2008). The REE pattern for the coral-dominated reefs reflect normal seawater and would have a much flatter pattern if there were a large freshwater input in the area where the coral reefs grew. If there had been significant groundwater seepage, the reefs or the platform they grew on (Lonely Bay Formation) would likely display vuggy porosity (cf. MacNeil). It is unlikely that there either of these methods of delivering nutrients was prolific on the MacKenzie Basin ramp; however, it is impossible to completely discount their contribution to higher nutrient levels.

Increased nutrients in more proximal areas of the MacKenzie Basin could have also been introduced through coastal upwelling, as Endinger (2002) suggested for the coral-dominated reefs in the Appalachian Basin. This process is mainly controlled by local geography and the interaction of ocean currents with the seafloor (Stanton 2006). Due to their connection to the open ocean, and geometry of the large shallowly dipping ramp, it seems likely that nutrients in the coastal areas of the Appalachian Basin and MacKenzie Basin might have been sourced in this manner. Without a method to directly measure ancient nutrient levels from limestone, the nutrient requirements for Paleozoic corals versus stromatoporoids and the source(s) for the nutrient input remains uncertain. If the nutrients were sourced through coastal upwelling processes, this would suggest that coastal upwelling and increased nutrients might have also affected the stromatoporoid-reefs, further down the ramp. Yet, the stromatoporoiod-dominated reefs have lower $\delta^{13}C_{(PDB)}$ values, indicating less photoautotrophic productivity further out in the basin and possibly lower nutrient levels. MacNeil (2008) demonstrated that Devonian stromatoporoid-dominated reefs from a carbonate ramp setting in the Alexandra Formation, NWT, were not tolerant of nutrients and grew in oligiotrophic conditions. Input of nutrients from groundwater seepage, sporadic runoff, storm events, and seasonally forced deepenings on the carbonate ramp in the Alexandra Formation resulted in a change from stromatoporoid-dominated reefs to microbial-carbonate reefs that were tolerant of enriched nutrient conditions (MacNeil 2008). Assuming that stromatoporoids were sensitive to nutrient enrichment, the MacKenzie Basin ramp might have been stratified with higher nutrient levels in the more proximal position on the ramp where the coral-dominated reefs grew, possibly due to a locally derived additional source of nutrients from terrestrial runoff (Figure 3-13).

As the dominant modern warm-water reef builders, scleractinian corals require low nutrient (oligiophic) conditions (Hallock 1987). This is because many of the corals in modern reefs require light for their photosynthetic symbionts. With high nutrient (meso or eutrophic) levels, the increase in photoautotrophs and particulate organic matter in the water column results in a decrease in the depth of light penetration (Mutti and Hallock 2003; Jones 2010). MacNeil (2008) observed that stromatoporoids, tabulate corals, and rugose corals likely lived in a much wider range of nutrients than what has traditionally been considered. It has also been suggested that Paleozoic reef building organisms may have actually been dependent on nutrients (Wood 1993; Stanton 2006; Surge et al. 1997; Scrutton 1999) and that photosymbiosis became necessary when nutrient levels were too low. Many more case studies of Paleozoic reefs in various paleogeographical settings, using an integrated approach of sedimentological and geochemical data are needed to test this hypothesis.

Conclusions

Observations from drill core and the petrographic, stable isotope, and rare earth element data collected from the Horn Plateau Formation reefs revealed the following conclusions:

1) the Horn Plateau Formation is stromatoporoid-dominated in the western side of the MacKenzie Basin and coral-dominated in the eastern side,
2) the coral-dominated reefs required muddy substrate and increased rates of sediment accumulation to support the corals, since they do not have adequate attachment mechanisms,
3) geochemical data indicate that the two reef-types grew under different physiochemical conditions and that the coral-dominated reefs were in a more proximal position on the ramp, which was closer to the exposed Canadian Shield,
4) nutrient levels on the the MacKenzie Basin ramp appear to be stratified, which is likely the result of coastal upwelling bringing nutrients into the more proximal area of the ramp, and possibly an additional source of nutrients input from the Canadian Shield.
5) geochemical evidence suggests that nutrients may have been stratified on the ramp, and the stromatoporoid-dominated reefs grew on the middle to lower ramp where they may have experienced lower nutrient levels,
6) Paleozoic corals do not seem to be limited by nutrients as many photosymbiotic scleractinian corals are in modern oceans.

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CHAPTER 4: PETROGRAPHIC AND GEOCHEMICAL CONTRASTS BETWEEN CALCITE- AND DOLOMITE-FILLED BURROWS: IMPLICATIONS FOR DOLOMITE FORMATION IN PALEOZOIC BURROWS[1]

Introduction

Carbonates are subjected to diagenetic changes from the moment they are deposited. There are two early main processes at work within the first few centimeters of carbonate deposits at the sediment-water interface: 1) dissolution and precipitation, and 2) modification of pore-water chemistry (Morse and MacKenzie 1990). For these processes to begin, there must be a mechanism to facilitate fluid flow through the sediments. Carbonate muds in particular tend to be impermeable, but when they are mixed through bioturbation, diagenetic processes affect the burrows and material adjacent to the burrows (Gingras *et al.* 2004).

The Lonely Bay Formation (Middle Devonian), found in the District of MacKenzie N.W.T. (Figure 4-1), contains a distinctive bioturbated lime mudstone facies that forms beds up to 1.5 meters thick. The burrows and their "diagenetic halo" (Gingras *et al.* 2004) have been partially to fully dolomitized. Similar fabrics have been recognized in several localities around the world (e.g. Abed and Scheider 1980; Chow and Longstaffe 1995; Gingras *et al.* 2004; Ramiel 2008). Nevertheless, there are still numerous questions surrounding their diagenetic histories, as the timing and cause of dolomitization in the burrows are still debatable.

Low temperature dolomitization remains somewhat of a mystery and there are only a few examples where low temperature dolomite or protodolomite has been precipitated in the laboratory (Vasconcelos *et al.* 1995; Wright 1999;

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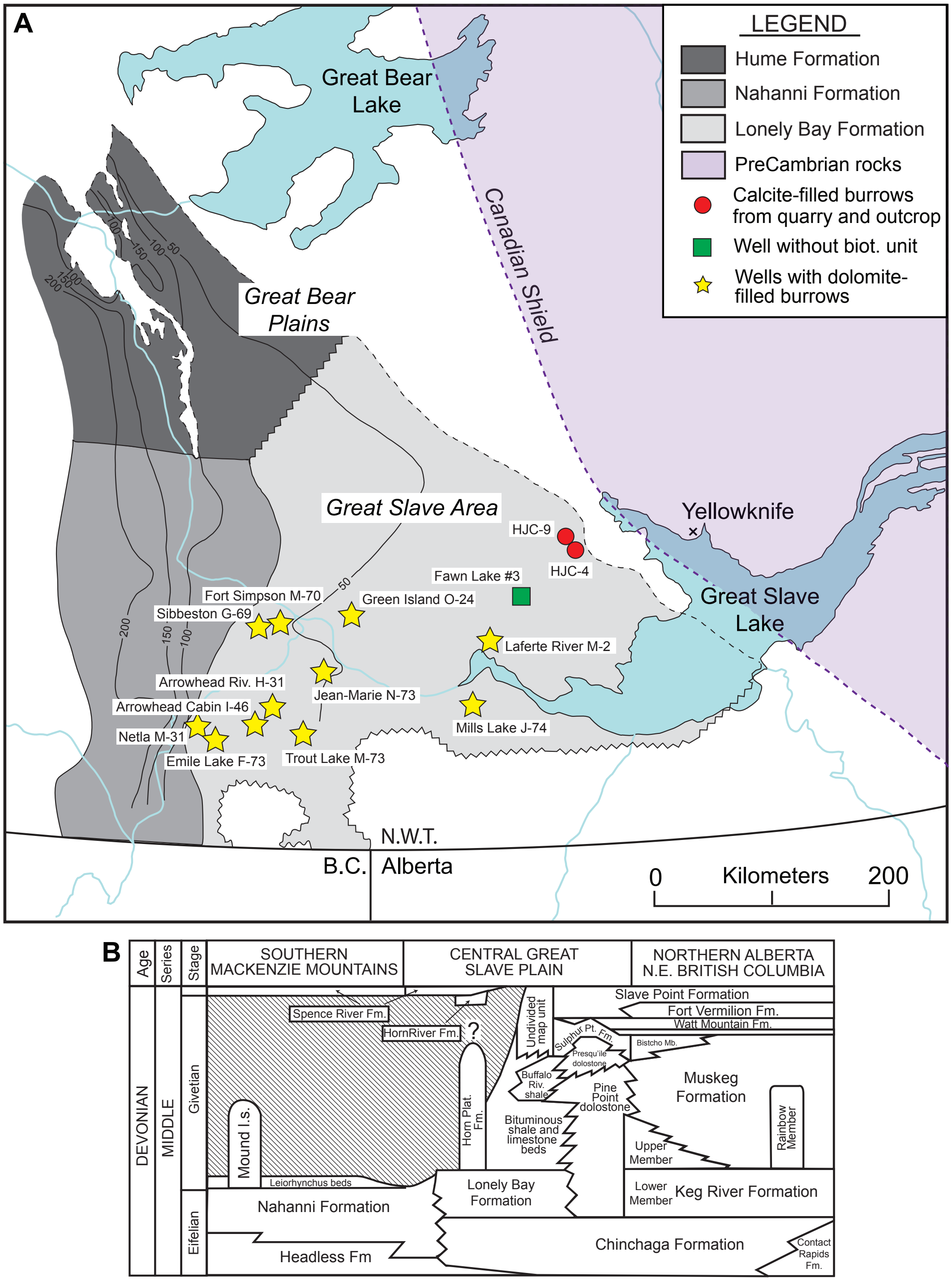

Figure 4-1. Location map and stratigraphy of study area (adapted from Drees 1993). A) Map showing well and outcrop locations Great Slave Area and Great Bear Plains Area with Middle Devonian strata. B) Middle Devonian stratigraphy.

Warthmann *et al.* 2000; Van Lith 2003). Such studies have provided valuable insight into the role of biology and the physiochemical parameters necessary for early dolomitization in carbonate sediments. Integration of petrographic information, isotope analyses, and REE analyses from the calcite-filled and dolomite-filled burrows provide the basis for explaining why dolomitization affected some of the burrows whereas others underwent little diagenetic modification.

Geological Setting

The Lonely Bay Formation, defined by Norris (1965), is a Middle Devonian sequence found in the Great Slave Lake Area of the District of MacKenzie, N.W.T., Canada (Figure 4-1). Mostly found in the subsurface, the Lonely Bay Formation is 30-60 m thick and is exposed just west of Great Slave Lake (Figure 4-1A). The Lonely Bay Formation is part of a carbonate ramp complex that developed in the MacKenzie Basin (Corlett and Jones, in press) that is located just north of the Tathlina High (Figure 4-1A), a Precambrian high that separated it from the Elk Point Basin to the south (Belyea 1972). The Lonely Bay Formation overlies the shallow-water, Chinchaga Formation and is overlain by the Horn Plateau Formation buildups and the Horn River Formation shales that encase the Horn Plateau buildups (Figure 4-1B; Corlett and Jones, in press). The Lonely Bay Formation is comprised of four main facies: a dolomitized intraclast mudstone facies, a peloidal partially dolomitized wacke-grainstone facies, a brachiopod/coral/stromatoporoid floatstone facies, and the focus of this paper, a partially dolomitized bioturbated wackestone facies. These facies formed in a mid- to outer ramp depositional setting.

Methods

This study is based on the examination of twelve drill cores, nine outcrops, and 55 thin sections. Drill core containing the Lonely Bay Formation was examined at the Geological Survey of Canada, Calgary. Field work in the area west of Great Slave Lake (Figure 4-1) was done in the fall of 2008. The partially dolomitized strongly bioturbated wackestone facies was seen in almost every core and at two locations where the Lonely Bay Formation is exposed (Figure 4-1). Most calcite-filled burrows were sampled from a quarry (HJC4; Figure 4-1) accessed during fieldwork and the dolomite-filled burrows were sampled from drill core. Thin sections were examined under a polarized microscope and cathololuminescence to discern the diagenetic histories of the burrows.

All of the geochemical analyses were performed at the Department of Earth and Atmospheric Sciences at the University of Alberta. Two polished thin sections were analyzed using a JEOL 8900 microprobe. The thin sections were used to create elemental maps (i.e., Ca, Mg, Fe, P, and Si) along 1 cm transects across one dolomite-filled burrow and one calcite-filled burrow. Further spot analyses were then performed on the microprobe to determine major element values (i.e., Ca, Mg, Fe, P, and Si) of the matrix and burrows in each sample. Values obtained from spot analysis were used for calibration during an *in situ* laser ablation transect across the same two burrow types. The laser ablation transect was performed using a New Wave Research UP-213 laser ablation system on the NuPlasma Multi-Collector ICP Mass Spectrometer. Each laser spot in the transect through two types of burrows analyzed for 67 different elements, including the rare earth elements (REEs). 1 to 1.5 cm long transects across each burrow type and the surrounding matrix were performed to obtain trace and REE concentrations. Rare earth element concentrations were normalized to the Post Archean Australian Shales (PAAS) for comparison with previous studies using REE element seawater proxies (Nance and Taylor 1976; Hayley et al. 2004; Northdruft 2004). REEs are useful seawater proxies because their concentrations are usually retained in limestones, even after dolomitization, providing the diagenetic fluids and rocks have similar REE distributions (Banner *et al.* 1988; Northdruft 2004). If the dolomitizing fluids were from seawater, the seawater signature would be reflected in the REEs.

Stable isotope analyses ($\delta^{18}O_{(PDB)}$ and $\delta^{13}C_{(PDB)}$) of the burrow fill and the surrounding matrices were also performed. The burrows and matrix were sampled from drill core (Arrowhead River H-31; Figure 4-1) and from samples taken from the quarry (HJC4; Figure 4-1) using a Dremel drill with a 1 mm size drill bit. Twenty samples were analyzed for $\delta^{18}O_{(PDB)}$ and $\delta^{13}C_{(PDB)}$ using the method outlined by McCrea (1950). Approximately 10 mg of powder from each fossil and matrix type were dissolved in 3 ml of phosphoric acid. The calcite samples were left to react for 1 hour, and the dolomite samples for 24 hours, in vacuum-sealed reactions vessels in a 25°C water bath. The CO_2 samples were then collected on a vacuum line and analyzed on a Finnigan Mat 252 Mass Spectrometer. Error margins for these samples were calculated at 0.02‰ for $\delta^{13}C_{(PDB)}$ and 0.03‰ for $\delta^{18}O_{(PDB)}$.

Results

The bioturbated facies in the Lonely Bay Formation has a distinct texture that has been referred to as a "nodular" limestone fabric (Figure 4-2; Norris 1965; Meijer Drees 1993). The "nodules" are actually burrows that are surrounded by an irregular diagenetic halo. The matrix between the burrows is composed mostly of brown lime mud that contains peloids and numerous small (1-5 mm) biofragments (Figures 4-3, 4-4) derived from trilobites, brachiopods, gastropods, and crinoids. Most of the allochems in the matrix are disarticulated and fragmented. Some of the larger gastropods and brachiopods have remained intact.

In outcrop and hand sample the burrows are 1 to 2 cm in diameter and

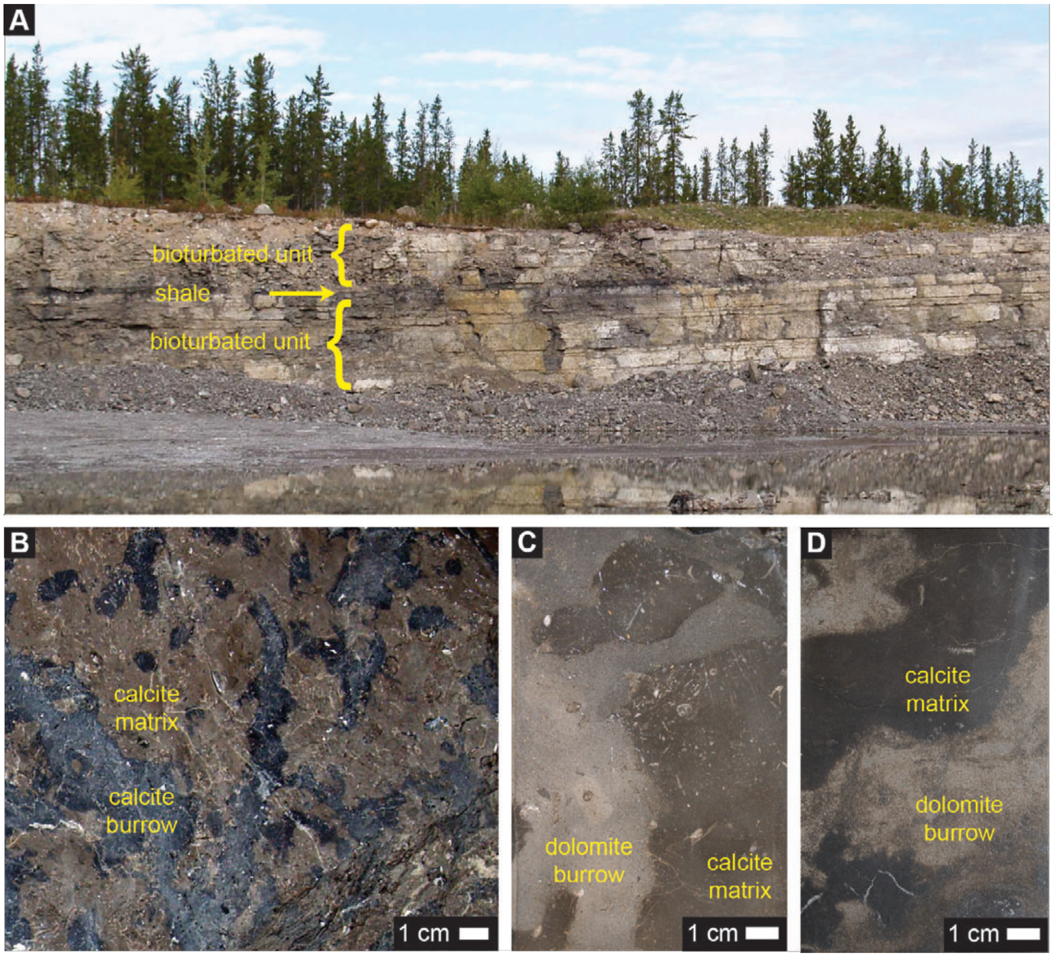

Figure 4-2. Photographs of quarry where calcite-filled burrows were collected and dolomite-filled burrows in drill core. A) 20 m section in quarry (HJC4; Figure 4-1) showing meter-sized beds of bioturbated unit in Lonely Bay Formation. B) Close-up of fresh surface on quarry wall. Calcite-filled burrows (black) surrounded by lime mud-wackestone matrix (brown). C) Dolomite-filled burrows (grey) surrounded by lime mud-wackstone matrix (brown; 1790 m – Arrowhead Cabin I-46; Figure 4-1). D) Dolomite-filled burrows (grey) surrounded by lime mudstone matrix (brown; 2022 m – Arrowhead Cabin I-46; Figure 4-1).

have a three-dimensional maze pattern. In thin section however, it is apparent that the burrows are < 5 mm in diameter with a ragged-edged diagenetic halo surrounding them. Both the calcite- and dolomite filled burrows show a pattern that is consistent with deposit feeding behaviour (Gingras pers comm.). For the purpose of this study, they are divided into the calcite-filled and dolomite-filled, based on their mineralogical and textural characteristics that largely reflect varying degrees of diagenetic alteration and dolomitized. The internal structure of the burrows is visible in the calcite burrows, but has been completely destroyed in the dolomite-filled burrows.

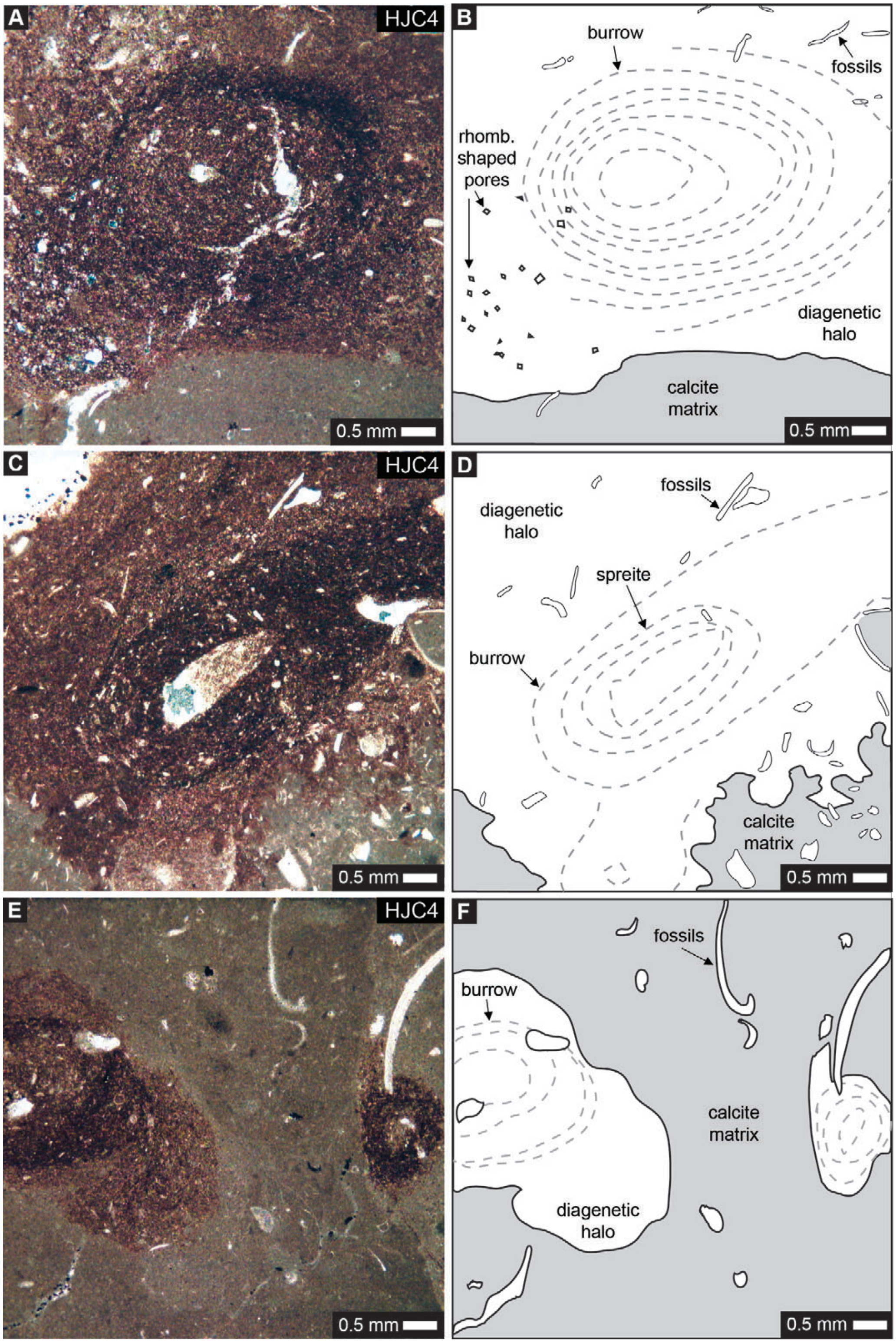

Figure 4-3. Thin-section photomicrographs (A, C, E) and corresponding schematic line-digrams (B, D, F) of calcite-filled burrows from Lonely Bay Formation found in quarry HJC4 (Figure 4-1) showing main components of calcite-filled burrows.

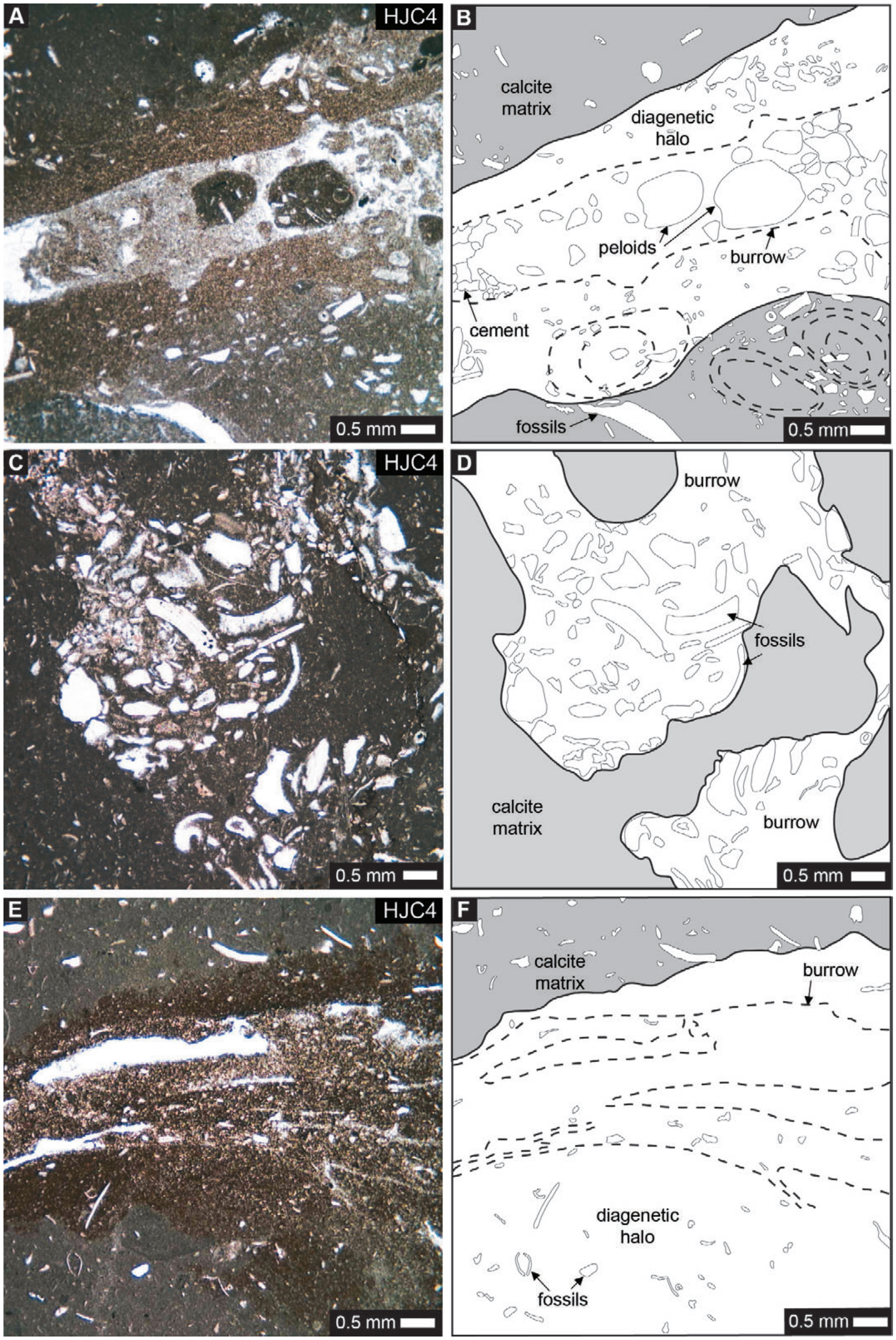

Figure 4-4. Thin-section photomicrographs (A, C, E) and corresponding schematic line-diagrams (B, D, F) of calcite-filled burrows from Lonely Bay Formation found in quarry HJC4 (Figure 4-1) showing main components of calcite-filled burrows.

Calcite-filled burrows

The calcite-filled burrows are branched, usually 1 to 3 mm in diameter, and surrounded by a diagenetic halo that is up to 15 mm wide (Figures 4-3, 4-4). They are backfilled with broken and fragmented fossils (Figures 4-3B, 4-4A, and B), dark brown calcite mudstone matrix, and peloids. Some of the calcite-filled burrows show alignment of shell material, revealing concentric burrow linings, or spreite, whereas others do not contain aligned fossils but are largely filled with peloids held in a mud matrix.

The diagenetic halo, surrounding the caustive burrows, is composed of dark brown calcite. The boundary between the diagenetic halo and the surrounding mud matrix is irregular. Rarely, the diagenetic halo contains scattered dolomite rhombs and rhombohedral-shaped pores (Figure 4-3A). When present, the dolomite rhombs fluoresce bright orange under cathodoluminescence and are not zoned. Some of the burrows are filled with blocky calcite cement (Figures 4-4A, E). Dissolution of the dolomite rhombs in the diagenetic halos associated with this burrow type most likely occurred after the blocky calcite cement filled the central part of the burrow, otherwise the rhombohedral-shaped pores would have been filled with calcite cement.

Dolomitized burrows

The dolomitized burrows, which are 1 to 3 cm in diameter, lack any internal structure and, it is therefore impossible to distinguish the causative burrow from the diagenetic halo (Figures. 4-5, 4-6). The halos and burrows have undergone pervasive dolomitization with only scattered brachiopod fragments being apparent in some of the burrows (Figures 4-5A, C and 4-6C). The burrow-filling dolomite is subhedral (0.05 mm – 0.3 mm) with some interlocking crystals.

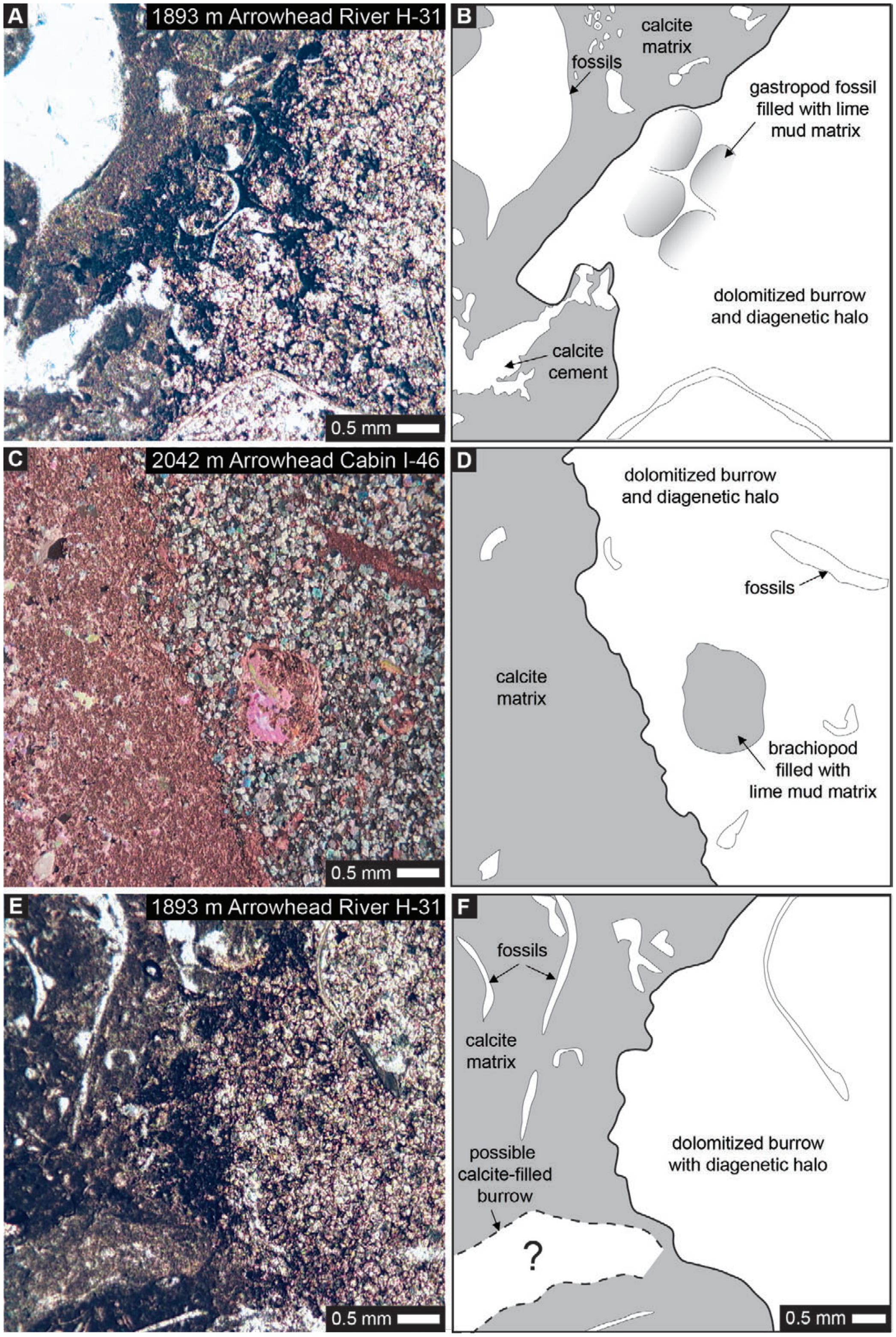

Figure 4-5. Thin-section photomicrographs (A, C, E) and corresponding schematic line-diagrams (B, D, F) of dolomite-filled burrows surrounded by calcite matrices from drill cores (Figure 4-1). The well ID of each sample is in the top right hand corner of the photomicrographs. Red colour in (C) is due to Alizarin red stain.

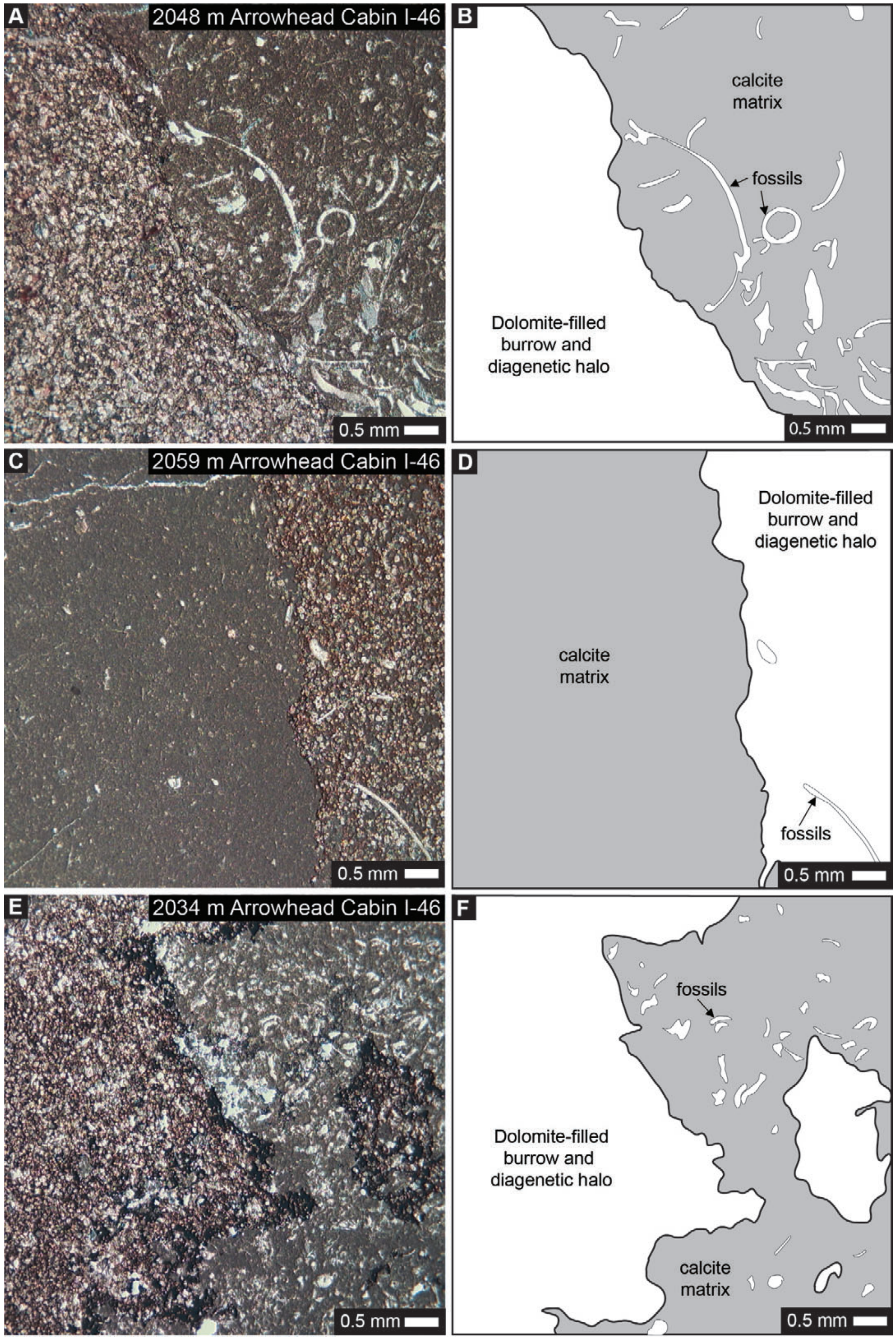

Figure 4-6. Thin-section photomicrographs (A, C, E) and corresponding schematic line-diagrams (B, D, F) of dolomite-filled burrows surrounded by calcite matrices from drill cores (Figure 4-1). The well ID of each sample is in the top right hand corner of the photomicrographs.

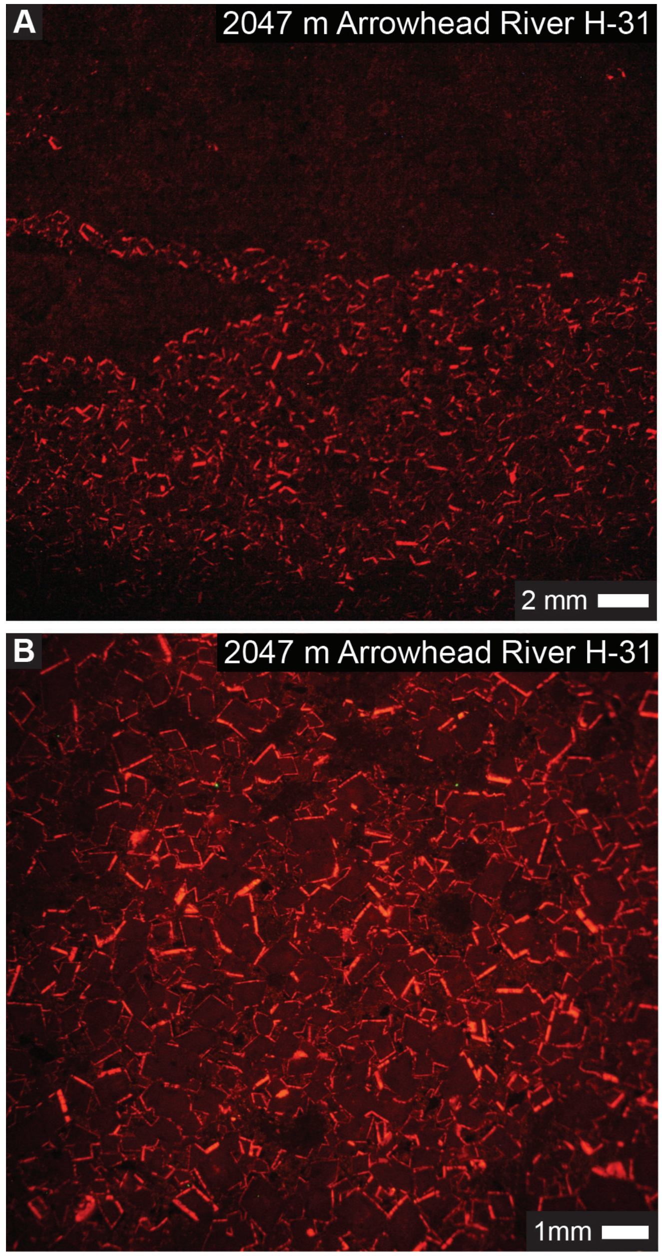

Figure 4-7. Dolomite-filled burrows (2047 m in Arrowhead River H-31) under cathodoluminescence. Note bright red rims of dolomite in burrows. The surrounding lime mud-wackestone matrix is non-luminescent.

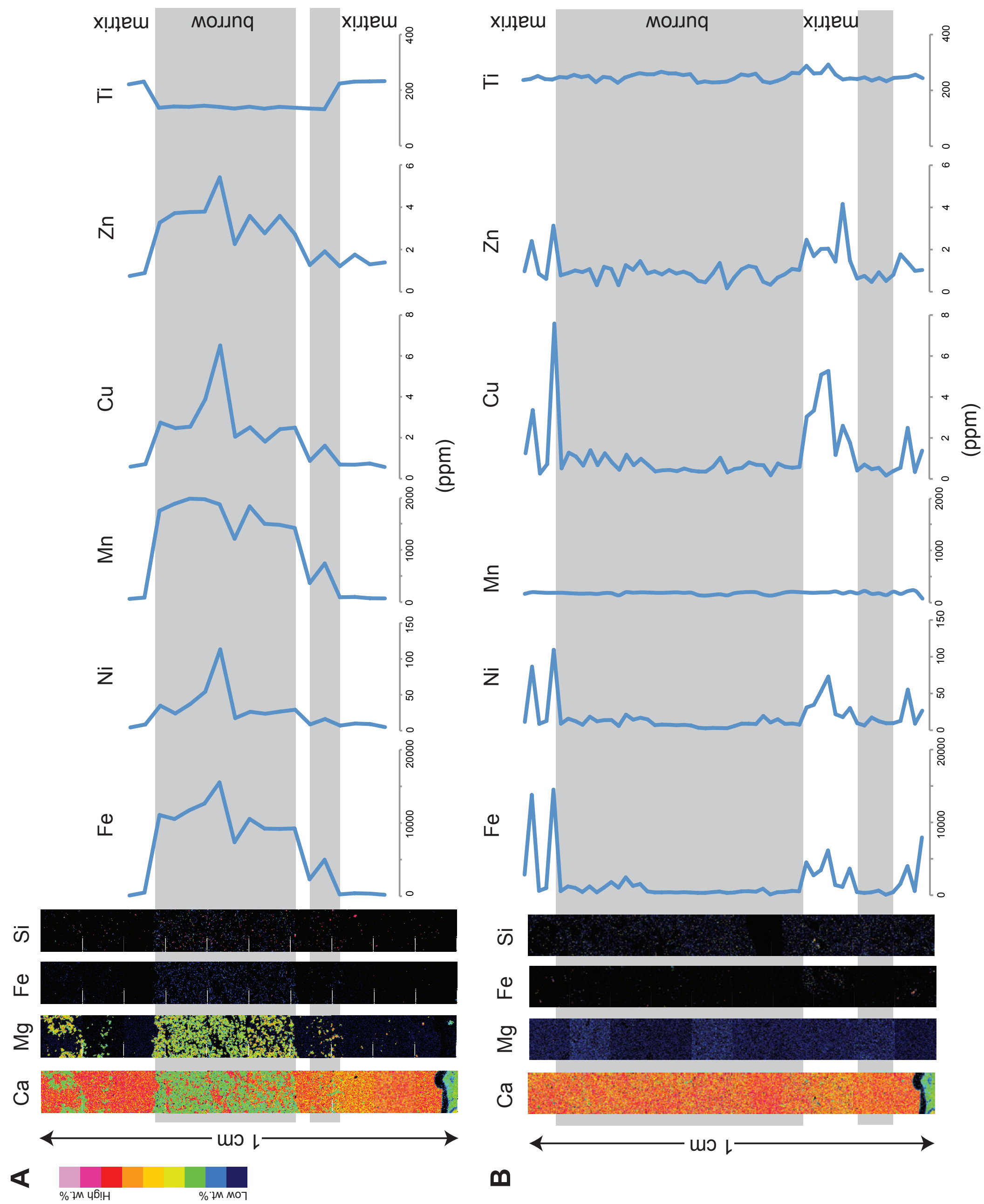

Figure 4-8. Elemental maps and trace element concentrations (ppm) of A) dolomite-filled burrows (2047 m – Arrowhead River H-31) and B) calcite-filled burrows (quarry – HJC4).

These dolomites show two zones under cathodoluminescence; the inner zone does not luminesce whereas the outer zone luminesces bright red (Figure 4-7).

Distribution of Dolomite- and Calcite-filled Burrows

The Lonely Bay Formation stretches over 350 km from west of the Great Slave Lake to where the MacKenzie Mountains are located today (Figure 4-1A). In the eastern part of the study area, where the Canadian Shield borders the MacKenzie Basin, the burrows are calcite-filled whereas the burrows found in the western part of the area are dolomitized (Figure 4-1A).

Major and Trace Elements

In the dolomitized burrows, backscatter electron images highlight a distinct boundary between the burrow with its diagenetic halo fill and the surrounding matrix (Figure 4-8). The Mg, Fe, and Si contents in the dolomitized burrow (Mg: 15.7 wt%, Fe: 4.1 wt%, Si: 2.5 wt%) are elevated relative to the matrix (Mg: 0.3 wt%, Fe: 0.1 wt %, Si: 0.1 wt%). In the calcite burrows, the Fe and Mg contents are negligible (Mg: 0.1 wt%, Fe: 0.8 wt%), and the Si and Ca content are similar in the burrow (Si: 1.0 wt%, Ca: 55.4) and the surrounding matrix (Si: 0.04 wt %, 57.5 wt %).

Trace element concentrations reveal some significant differences between the calcite and dolomite burrows (Figure 4-8). In the dolomitized burrow, transition metal cation concentrations (Fe, Cu, Ni, Mn, and Zn) are elevated in the burrow relative to the surrounding matrix (Figure 4-8; Table 4-1). In contrast, metal cation concentrations in the calcite-filled burrow are lower in the burrow than in the surrounding matrix (Figure 4-8) and to the dolomite-filled burrows (Figure 4-9; Table 4-2).

The REE pattern from the calcite-filled burrow shows a “linear” pattern (Haley et al. 2004) with a positive La anomaly, a negative Ce anomaly, and LREE

TABLE 4-1. *Trace element data (ppm) from the calcite-filled burrows. Theses data are represented diagrammatically in Figure 8. Shaded areas represent burrows.*

Sample No.	Fe	Ni	Mn	Cu	Zn	Ti	Sr	Mg	Si
LBC-1	8181.0	5.4	70.3	1.4	1.0	243.8	1365.4	3903.6	2417.0
LBC-2	861.2	1.8	215.8	0.3	1.0	256.4	346.4	7697.8	9019.9
LBC-3	4253.9	11.2	208.7	2.5	1.4	248.3	354.2	7275.8	7337.9
LBC-4	1804.8	2.6	157.0	0.5	1.8	246.3	620.5	6507.9	5611.8
LBC-5	683.1	2.0	197.7	0.4	0.8	244.5	443.4	7333.3	4139.1
LBC-6	349.2	1.9	135.2	0.2	0.5	232.7	513.2	6000.1	652.6
LBC-7	909.2	2.5	168.0	0.5	0.9	244.2	521.3	6504.6	5340.6
LBC-8	671.5	3.5	157.7	0.5	0.5	235.0	469.3	6734.3	1046.5
LBC-9	587.3	1.3	211.3	0.7	0.8	246.8	336.4	7238.8	5828.8
LBC-10	711.4	1.9	166.4	0.4	0.6	240.5	466.8	5851.5	4826.9
LBC-11	3909.8	6.1	195.1	1.8	1.5	242.8	376.0	6448.2	7694.1
LBC-12	1416.3	3.6	164.4	2.6	4.2	239.2	288.7	21204.6	163657.1
LBC-13	1665.3	4.4	202.4	1.2	1.4	256.5	368.3	7129.3	11832.8
LBC-14	6395.7	14.8	184.4	5.3	2.0	292.3	378.7	6578.2	14410.8
LBC-15	3707.2	10.8	184.1	5.1	2.0	261.9	381.3	7484.6	19698.0
LBC-16	2997.5	7.0	176.8	3.3	1.7	260.6	342.0	6524.7	13302.1
LBC-17	4759.0	6.3	181.7	3.0	2.5	287.7	334.1	8503.1	32518.5
LBC-18	811.5	1.5	189.3	0.6	1.0	261.3	307.2	5683.9	13899.7
LBC-19	863.0	1.9	196.3	0.5	1.1	262.6	293.0	5717.3	15672.0
LBC-20	722.6	1.7	184.7	0.6	0.8	244.3	319.0	5494.1	12561.4
LBC-21	689.1	3.0	150.1	0.8	0.7	234.3	405.6	6446.1	7027.0
LBC-22	375.0	2.1	125.7	0.2	0.3	227.0	454.6	6369.7	152.2
LBC-23	1160.1	3.9	143.7	0.7	0.5	231.9	417.9	5758.7	1749.4
LBC-24	751.6	1.7	187.3	0.7	1.2	259.9	317.3	5376.6	12828.8
LBC-25	830.0	1.8	192.2	0.8	1.2	253.0	300.4	5913.0	17584.3
LBC-26	801.2	1.8	185.4	0.5	1.1	257.1	281.5	6044.8	20839.2
LBC-27	645.7	1.1	170.7	0.5	0.7	241.5	306.4	4952.8	10203.9
LBC-28	587.1	0.6	130.2	0.3	0.2	231.4	292.5	4191.5	5093.6
LBC-29	776.6	0.6	149.9	1.0	1.4	229.3	273.5	9702.3	54107.1
LBC-30	692.0	0.6	138.5	0.6	0.9	228.3	300.5	8646.6	40796.8
LBC-31	595.5	0.5	125.7	0.4	0.4	232.1	321.0	5120.0	12539.5
LBC-32	592.6	0.7	134.9	0.4	0.5	227.1	280.3	4651.8	9805.9
LBC-33	646.9	1.3	180.5	0.4	0.8	258.2	281.7	4868.1	10881.6
LBC-34	692.4	1.5	176.3	0.5	1.0	254.7	293.9	4977.7	13195.0
LBC-35	636.1	1.4	186.0	0.4	0.9	260.8	285.8	5132.2	13048.8
LBC-36	688.7	1.5	181.0	0.4	1.0	260.6	310.2	5212.4	13940.9
LBC-37	661.6	1.5	176.3	0.4	0.8	266.7	293.1	5046.1	12061.1
LBC-38	660.1	1.4	178.7	0.4	1.0	257.8	280.9	5217.0	15638.3
LBC-39	802.9	3.0	183.1	0.7	0.9	257.6	289.4	5075.5	11977.8
LBC-40	1805.7	3.4	185.8	1.0	1.5	261.6	308.8	5252.6	16135.8
LBC-41	1551.0	2.9	179.3	0.7	1.0	254.4	310.9	5087.0	11067.9
LBC-42	2721.2	4.3	189.0	1.2	1.3	245.9	319.3	4959.4	10104.2
LBC-43	1313.7	1.2	130.4	0.5	0.3	227.2	449.1	5133.5	2591.0
LBC-44	2067.8	2.8	170.1	0.8	1.1	245.2	320.7	4666.1	9507.7
LBC-45	1331.4	2.8	173.5	1.3	1.2	248.3	334.5	5135.9	8744.8
LBC-46	665.3	2.5	155.7	0.7	0.3	229.4	350.5	5383.6	673.3
LBC-47	1481.1	3.7	164.7	1.4	1.1	252.0	299.1	5429.0	12942.7
LBC-48	721.0	1.5	160.5	0.7	0.9	247.6	316.5	5104.5	11718.2
LBC-49	1281.4	2.5	165.0	1.1	1.0	255.4	343.8	5572.3	12209.2
LBC-50	1481.7	3.2	171.3	1.3	0.9	245.7	337.6	5824.3	6996.9
LBC-51	839.3	1.8	178.2	0.5	0.8	248.0	362.6	6033.2	8310.2
LBC-52	14716.5	22.2	177.6	7.6	3.1	238.8	311.3	4061.3	8529.2
LBC-53	1256.5	2.6	177.1	0.7	0.6	240.5	311.7	6026.5	6220.0
LBC-54	873.5	1.8	183.6	0.3	0.9	251.5	303.3	6601.4	10584.1
LBC-55	14003.1	17.6	189.1	3.4	2.4	241.0	339.7	5900.9	5029.3
LBC-56	3095.3	2.3	158.6	1.3	1.0	237.0	387.1	6322.9	21767.3

TABLE 4-2. *Trace element data (ppm) from the dolomite-filled burrows. These data are represented diagrammatically in Figure 8. Shaded areas represents burrows.*

Sample No.	Fe	Ni	Mn	Cu	Zn	Ti	Sr	Mg	Si
LBD-1	213.7	0.8	61.8	0.6	0.8	220.8	272.6	2960.3	5394.7
LBD-2	633.1	1.6	86.3	0.7	0.9	230.7	305.4	4125.2	3506.1
LBD-3	11829.0	6.9	1748.3	2.7	3.3	136.6	174.7	87856.8	6514.5
LBD-4	11249.5	4.7	1879.8	2.5	3.7	141.2	147.6	97648.5	8699.3
LBD-5	12530.4	7.4	1977.6	2.5	3.8	139.9	161.0	78697.3	7245.2
LBD-6	13487.9	10.9	1968.1	3.9	3.8	144.0	188.8	76659.6	8337.4
LBD-7	16536.0	22.8	1870.7	6.5	5.4	139.7	202.8	83739.8	7970.6
LBD-8	7911.7	3.4	1211.2	2.1	2.3	133.8	160.6	56610.0	4508.8
LBD-9	11266.8	5.2	1830.7	2.5	3.6	140.4	160.7	74707.0	5887.1
LBD-10	9873.0	4.7	1496.5	1.8	2.8	133.5	166.5	55161.8	5963.4
LBD-11	9823.3	5.3	1477.9	2.4	3.6	139.9	165.8	62761.4	6902.6
LBD-12	9889.0	5.8	1417.6	2.5	2.7	136.8	189.4	49001.9	7278.7
LBD-13	2550.3	1.6	366.9	0.9	1.3	133.9	164.4	16182.1	3128.5
LBD-14	5370.3	3.2	739.3	1.6	1.9	131.5	167.4	30139.8	6189.9
LBD-15	374.7	1.3	93.3	0.7	1.2	223.8	280.1	3443.8	5796.3
LBD-16	537.4	1.9	98.7	0.7	1.8	230.6	301.1	4640.7	4198.7
LBD-17	490.1	1.7	73.8	0.7	1.3	231.4	311.1	2724.5	2418.6
LBD-18	314.7	0.9	70.7	0.6	1.4	232.3	314.7	2869.4	3846.0

and middle rare earths (MREEs) that are slightly higher that the heavy rare earths (HREEs; Figure 4-10; Table 4-3). The REE patterns are quite different for the dolomite- and calcite-filled burrows.

The REE pattern from dolomite-filled burrow lacks a significant La anomaly, lacks a negative Ce anomaly, and is enriched in the MREEs (Figure 4-10; Table 4-4). This MREE enrichment in a REE pattern is referred to as an “MREE” bulge (Johannesson and Zhou, 1999; Haley and Klinkhammer 2003; Haley et al. 2004).

Stable Isotope Geochemistry

The stable isotope analysis ($\delta^{13}C_{(PDB)}$ and $\delta^{18}O_{(PDB)}$) revealed significant differences between the matrix surrounding the calcite burrows and matrix around the dolomite burrows (Figure 4-11). The matrix surrounding dolomitized burrows has $\delta^{13}C_{(PDB)}$ values between -1.17 to -1.39‰ whereas the matrix surrounding the calcite burrows showed values of 0.25 to -0.45‰. The $\delta^{18}O_{(PDB)}$ values for the matrix around the dolomite burrows (-9.23 to -9.75‰) are also lower than those from the matrix around the calcite burrows (-5.79 to -7.10‰; Figure 4-11).

TABLE4-3 . *Rare earth element data (ppm) from the calcite-filled burrows. These data are represented graphically in Figure 10.*

Sample No.	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Y	Ho	Er	Tm	Yb	Lu
LBC-1	2.40	2.47	0.39	1.85	0.27	0.05	0.28	0.04	0.21	2.00	0.04	0.20	--	0.15	--
LBC-2	5.96	7.77	1.06	4.80	0.67	0.17	0.63	0.09	0.61	5.00	0.10	0.28	0.03	0.23	0.04
LBC-3	6.01	7.79	1.10	4.52	0.63	0.18	0.84	0.09	0.72	5.00	0.12	0.33	0.04	0.26	0.05
LBC-4	5.60	5.88	0.87	4.02	0.63	0.17	0.78	0.08	0.60	5.60	0.12	0.31	0.06	0.38	0.04
LBC-5	5.85	7.07	1.08	4.72	0.71	0.19	0.75	0.11	0.61	5.60	0.14	0.24	0.05	0.33	0.05
LBC-6	5.68	5.06	0.91	3.95	0.66	0.16	0.80	0.10	0.63	6.08	0.13	0.36	0.04	0.20	0.03
LBC-7	5.46	6.41	0.93	4.18	0.62	0.18	0.64	0.08	0.60	5.09	0.13	0.31	0.03	0.28	0.04
LBC-8	5.58	5.17	0.84	4.00	0.60	0.16	0.77	0.08	0.56	5.09	0.12	0.36	0.03	0.27	0.04
LBC-9	4.91	5.80	0.90	3.95	0.62	0.13	0.63	0.07	0.47	4.65	0.12	0.27	0.05	0.24	0.03
LBC-10	6.81	8.95	1.37	5.65	0.86	0.22	0.86	0.13	0.67	6.18	0.14	0.45	0.05	0.36	0.04
LBC-11	6.65	8.33	1.15	5.55	0.77	0.19	0.92	0.12	0.73	5.98	0.14	0.42	0.07	0.32	0.05
LBC-12	7.64	9.36	1.41	6.33	0.65	0.39	1.12	0.15	0.95	8.51	0.19	0.55	0.09	--	0.07
LBC-13	7.43	9.71	1.40	6.14	0.83	0.26	1.02	0.12	0.90	7.57	0.18	0.50	0.07	0.46	0.07
LBC-14	9.49	11.70	1.68	7.46	1.10	0.25	1.18	0.16	0.95	10.59	0.23	0.71	0.08	0.55	0.07
LBC-15	9.91	11.73	1.62	7.33	1.09	0.31	1.12	0.14	0.92	8.17	0.20	0.57	0.07	0.48	0.05
LBC-16	9.72	11.34	1.59	7.43	1.08	0.27	1.18	0.17	0.95	8.98	0.22	0.62	0.09	0.51	0.07
LBC-17	9.73	11.68	1.64	7.64	1.12	0.30	1.17	0.17	1.09	9.44	0.20	0.55	0.08	0.52	0.06
LBC-18	6.56	8.38	1.22	5.65	0.89	0.22	1.00	0.12	0.86	8.07	0.16	0.49	0.06	0.35	0.06
LBC-19	8.35	8.72	1.28	5.89	0.99	0.22	0.84	0.12	0.77	7.05	0.16	0.47	0.07	0.29	0.05
LBC-20	7.09	9.05	1.29	5.62	0.88	0.24	0.93	0.11	0.70	7.18	0.16	0.45	0.05	0.37	0.03
LBC-21	8.65	8.98	1.36	6.69	0.82	0.19	0.85	0.10	0.63	6.33	0.12	0.37	0.03	0.34	0.04
LBC-22	3.98	3.11	0.54	2.50	0.38	0.09	0.43	0.05	0.31	3.84	0.07	0.19	0.02	0.12	0.02
LBC-23	7.75	6.86	1.27	5.27	0.84	0.22	0.87	0.10	0.65	6.56	0.14	0.37	0.05	0.30	0.04
LBC-24	6.63	8.74	1.26	5.87	0.85	0.23	0.90	0.12	0.81	7.33	0.16	0.42	0.05	0.34	0.06
LBC-25	7.05	9.84	1.31	6.31	0.89	0.28	1.01	0.12	0.80	7.68	0.17	0.51	0.07	0.41	0.06
LBC-26	6.63	8.60	1.22	5.97	0.81	0.23	1.05	0.14	0.86	7.55	0.18	0.56	0.08	0.35	0.06
LBC-27	5.93	7.87	1.15	5.25	0.79	0.21	0.69	0.11	0.57	6.06	0.14	0.42	0.05	0.28	0.04
LBC-28	3.62	5.05	0.73	3.14	0.47	0.12	0.37	0.06	0.33	2.99	0.06	0.17	0.03	0.13	0.02
LBC-29	2.20	3.90	0.49	2.34	0.35	0.13	0.34	0.07	0.27	1.80	--	0.22	0.04	0.10	0.03
LBC-30	5.66	8.46	1.20	5.65	0.72	0.19	0.60	0.08	0.40	3.98	0.09	0.25	0.04	0.26	0.04
LBC-31	2.87	3.49	0.56	2.51	0.38	0.08	0.34	0.03	0.36	2.63	0.06	0.16	0.01	0.10	--
LBC-32	7.06	8.85	1.40	5.85	0.73	0.17	0.75	0.11	0.56	4.68	0.15	0.31	0.04	0.20	0.04
LBC-33	6.19	7.58	1.14	5.53	0.77	0.24	0.93	0.13	0.94	9.36	0.19	0.63	0.10	0.56	0.09
LBC-34	6.69	8.14	1.21	5.35	0.85	0.23	0.85	0.11	0.89	8.17	0.18	0.54	0.08	0.45	0.05
LBC-35	6.06	7.78	1.14	5.24	0.75	0.22	1.01	0.13	0.77	7.49	0.13	0.52	0.08	0.37	0.03
LBC-36	6.94	8.23	1.22	6.01	0.85	0.25	0.95	0.14	1.03	8.74	0.21	0.62	0.08	0.50	0.08
LBC-37	7.88	9.47	1.31	6.41	0.81	0.24	0.91	0.12	0.83	8.05	0.16	0.47	0.07	0.46	0.06
LBC-38	6.56	8.00	1.17	5.53	0.69	0.25	1.12	0.13	0.79	7.93	0.17	0.48	0.07	0.39	0.05
LBC-39	6.32	8.32	1.23	5.40	0.86	0.25	0.86	0.14	0.83	7.82	0.16	0.50	0.06	0.47	0.06
LBC-40	7.17	9.10	1.33	6.10	1.02	0.23	1.02	0.11	0.91	8.83	0.21	0.52	0.07	0.46	0.05
LBC-41	7.63	9.05	1.35	6.55	0.96	0.27	1.17	0.17	0.93	8.81	0.20	0.48	0.07	0.40	0.06
LBC-42	7.60	8.87	1.39	5.98	1.06	0.26	1.07	0.14	0.96	10.06	0.21	0.57	0.07	0.58	0.06
LBC-43	5.89	5.16	0.90	3.96	0.71	0.17	0.87	0.11	0.71	7.92	0.14	0.43	0.06	0.27	0.05
LBC-44	6.36	7.77	1.20	5.43	0.89	0.22	1.01	0.14	0.84	7.85	0.17	0.41	0.07	0.35	0.07
LBC-45	7.86	8.39	1.39	6.29	0.98	0.26	1.07	0.16	0.86	8.64	0.20	0.50	0.06	0.40	0.07
LBC-46	9.59	6.86	1.25	5.83	0.97	0.26	1.11	0.14	1.00	8.42	0.20	0.48	0.07	0.44	0.06
LBC-47	7.96	9.37	1.42	6.86	0.96	0.27	1.29	0.15	1.06	9.09	0.20	0.59	0.07	0.45	0.07
LBC-48	8.15	10.27	1.65	7.54	1.13	0.30	1.11	0.16	0.82	9.02	0.24	0.54	0.07	0.34	0.06
LBC-49	8.47	10.65	1.62	7.42	1.18	0.29	1.03	0.14	0.84	8.91	0.21	0.57	0.08	0.58	0.06
LBC-50	7.00	7.67	1.26	5.66	0.89	0.23	0.97	0.11	0.94	7.84	0.17	0.63	0.08	0.43	0.05
LBC-51	6.80	8.06	1.21	5.62	0.80	0.20	0.85	0.11	0.74	7.41	0.16	0.45	0.06	0.30	0.05
LBC-52	10.56	15.42	2.29	10.55	1.38	0.35	1.59	0.18	1.35	9.03	0.20	0.68	0.06	0.48	0.04
LBC-53	6.19	6.94	1.13	5.02	0.78	0.21	0.77	0.11	0.70	6.64	0.14	0.39	0.05	0.28	0.05
LBC-54	5.68	6.70	1.04	4.56	0.67	0.17	0.74	0.09	0.62	5.87	0.14	0.27	0.05	0.52	0.04
LBC-55	5.16	5.68	0.93	4.42	0.63	0.19	0.80	0.11	0.70	6.46	0.15	0.42	0.05	0.42	0.05
LBC-56	14.10	19.92	3.00	14.11	2.09	0.48	2.29	0.28	1.58	11.79	0.31	0.80	0.09	0.57	0.07

TABLE4-4. *Rare earth element data (ppm) from the dolomite-filled burrows. These data are represented graphically in Figure 10.*

Sample No.	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Y	Ho	Er	Tm	Yb	Lu
LBD-1	0.73	1.13	0.15	0.80	0.10	0.04	0.13	0.01	0.09	0.44	0.02	0.04	0.01	--	--
LBD-2	0.66	1.17	0.15	0.70	0.12	--	0.17	0.02	0.11	0.40	0.01	0.03	--	--	--
LBD-3	0.43	0.90	0.10	0.59	0.09	0.04	0.10	0.02	0.05	0.46	0.02	0.04	0.01	0.05	0.01
LBD-4	0.52	0.97	0.10	0.53	0.13	0.05	0.15	0.02	0.12	0.51	0.02	0.04	--	--	--
LBD-5	0.44	0.83	0.13	0.57	0.09	0.04	--	0.01	0.09	0.51	0.02	0.04	--	--	--
LBD-6	0.49	0.92	0.12	0.56	0.12	0.04	0.22	0.01	0.10	0.47	0.02	0.06	--	--	--
LBD-7	0.39	0.70	0.12	0.47	0.12	0.04	0.15	0.02	0.08	0.48	0.01	0.05	--	--	--
LBD-8	0.44	0.83	0.10	0.56	0.10	0.04	0.13	0.04	0.06	0.45	0.01	0.03	--	--	--
LBD-9	0.45	0.82	0.12	0.55	0.10	0.04	0.08	0.02	0.10	0.53	--	--	0.01	--	--
LBD-10	0.50	0.86	0.09	0.53	0.12	0.03	0.12	0.02	0.05	0.40	--	0.06	--	--	--
LBD-11	0.49	0.88	0.12	0.49	0.12	0.04	0.12	--	0.08	0.54	--	0.04	0.01	0.07	0.01
LBD-12	0.32	0.69	0.09	0.50	0.06	0.04	0.13	0.01	0.09	0.48	0.02	--	0.01	--	--
LBD-13	0.44	0.81	0.10	0.53	0.10	0.04	0.13	0.01	0.05	0.32	0.01	0.03	--	--	--
LBD-14	0.44	0.82	0.12	0.53	0.11	0.04	0.13	0.01	0.09	0.39	--	0.04	--	0.05	--
LBD-15	0.76	1.25	0.16	0.82	0.13	0.04	0.20	0.02	0.08	0.52	0.01	0.04	--	0.08	--
LBD-16	0.92	1.44	0.21	0.95	0.20	0.05	0.16	0.01	0.06	0.52	0.02	--	0.01	0.04	0.01
LBD-17	0.79	1.32	0.18	0.84	0.18	0.06	--	0.01	0.08	0.43	0.02	0.05	0.01	0.05	--
LBD-18	0.84	1.27	0.17	0.96	0.14	0.05	0.10	0.01	0.07	0.47	0.01	0.03	--	0.05	--

The calcite- and dolomite-filled burrows had similar $\delta^{13}C_{(PDB)}$ and $\delta^{18}O_{(PDB)}$ values (Figure 4-11). The dolomitized burrows have $\delta^{13}C_{(PDB)}$ values between -0.08 to -0.53‰ and $\delta^{18}O_{(PDB)}$ values range from -7.85 to -9.45‰. The calcite burrows show a slightly larger range with $\delta^{13}C_{(PDB}$ values between -0.09 to -0.26‰ and $\delta^{18}O_{(PDB)}$ values between -7.56 to -11.37‰.

Discussion

Low-temperature dolomite precipitation in burrows has been associated with anoxic environments (Baker and Burns 1985), sulphate-reducing bacteria (Brown and Farrow 1978; Gunatilaka et al. 1987; Wright 1999; Van Lith 2003), marine-derived organic matter (Wright 1999; Slaughter and Hill 1991), and fluid flow and source of magnesium (Morrow 1978).

Haley et al. (2004) measured REEs in pore waters in the upper 25 cm of modern day near shore sediment with terrigenous and organic material from the Californian margin and found that patterns in REE plots represent changes in sediment depth and various intervals of reducing environments. REEs from the calcite-filled burrows in the Lonely Bay Formation show a positive La and

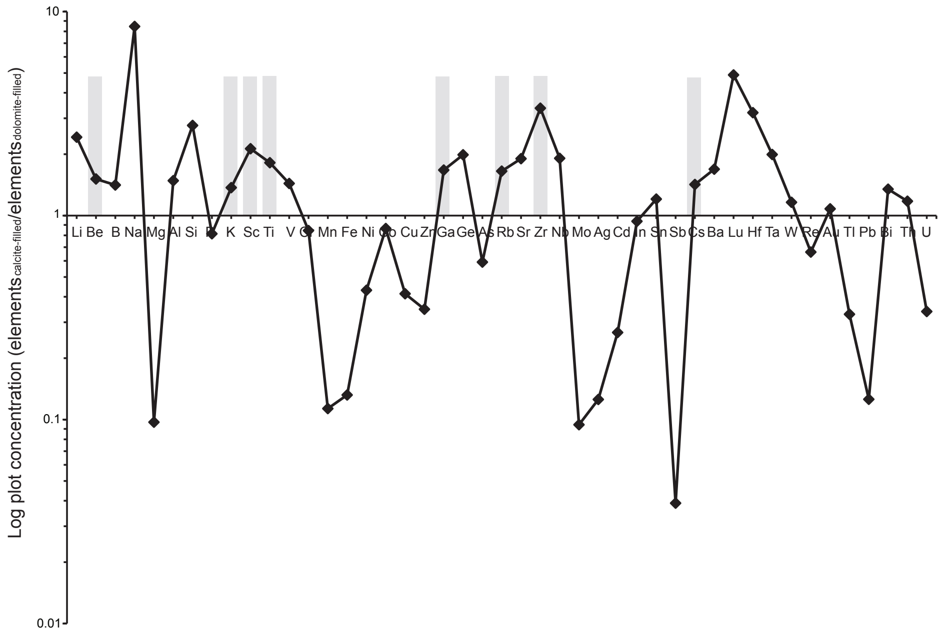

Figure 4-9. Concentration of average trace element values in calcite burrow fill (quarry – HJC4) relative to dolomite burrow fill (2047 m – Arrowhead River H-31).

negative Ce anomaly as well as a LREE and MREE concentrations that are slightly higher than the HREEs (Figure 4-10). This “linear”-type pattern shows a slight enrichment in the LREEs (Haley *et al.*, 2004), which represents oxic to suboxic conditions at the sediment-water interface where LREEs associated with surfaces of particulate organic matter in the water column are released (Sholkovitz *et al.* 1994). This indicates that conditions in the calcite-filled burrows were not fully anoxic.

The REE pattern from the dolomite-burrow shows MREE enrichment relative to the LREEs and HREEs. Referred to as an “MREE-bulge”, this results from fully anoxic conditions (Haley *et al.* 2004) where Fe-oxides become reduced. In the water column Fe-oxides scavenge REEs, specifically, the MREEs (Johannesson and Zhou 1994). Below the sediment-water interface, first the Mn-oxides degrade under reducing conditions (no affect on the REEs) and when full anoxia is reached, Fe-oxides are reduced and the MREEs are released (Haley et al. 2004). The dolomite-filled burrows in the Lonely Bay Formation show MREE enrichment, which represents anoxic conditions.

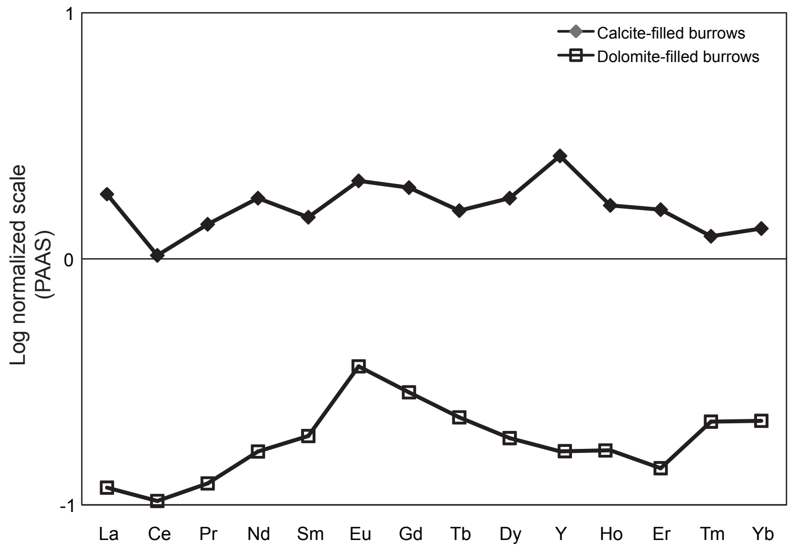

Figure 4-10. Rare earth element patterns for calcite- and dolomite filled burrows (quarry – HJC4; 2047 m – Arrowhead River H-31). Values are plotted on logarithmic scale and relative to PAAS (Post Archean Australian Shales).

Another indication of reducing or anoxic conditions in the dolomite-filled burrows (cf. Finney et al. 1988; Shaw et al. 1990) is the concentration of transition metals (e.g., Fe, Mn, Cu, Ni), which are much higher in the dolomite burrows than in the calcite burrows (Figure 4-8). Metals are enriched in burrow linings (Over 1990; Gingras et al. 2004) but these metals are mobile and constantly travel back and forth between oxic sediments and anoxic sediments, from areas of high metal concentrations to areas of low concentrations. Sulphate reducing bacteria (SRBs) influence the cycling of these metals back and forth across the oxic/anoxic boundary (Johnson 1998).

The SRBs promote metal immobilization and dolomitization by degrading proteins in particulate organic matter thereby releasing CO_2 and ammonia (Slaughter and Hill 1991; Wright 2000). Ammonia acts to raise the pH and alkalinity of ambient seawater, which also increases the activity of the carbonate ion (Berner 1980; Slaughter and Hill, 1991; Wright 2000):

$$NH_3 + H_2O \rightarrow NH^{4+} + OH^-$$

$$OH^- + HCO_3^- \rightarrow H_2O + CO_3^{2-}.$$

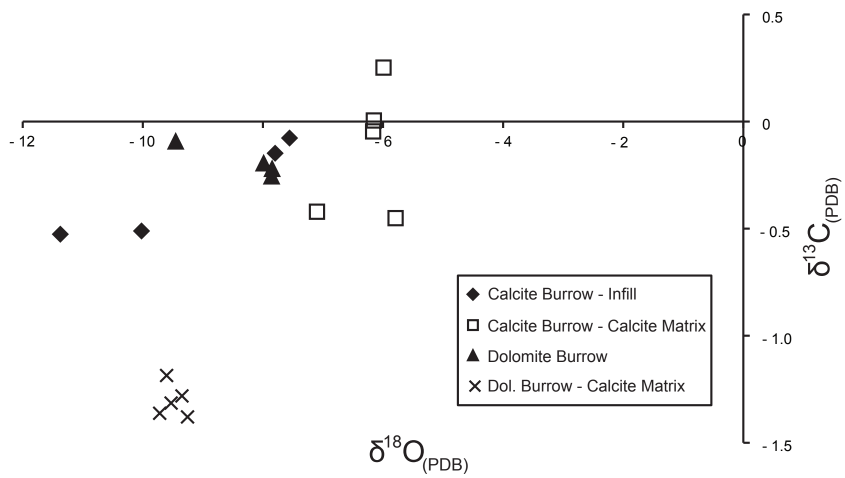

Figure 4-11. Plot of $\delta^{13}C$ vs. $\delta^{18}O$ for samples taken from calcite- and dolomite-filled burrows and their respective lime mud-wackestone matrices (quarry – HJC4; 2047 m – Arrowhead River H-31).

By raising the pH in interstitial pore water SRBs reduce the solubility of the metals in the burrow (Johnson 1998) resulting in increased metal concentration like that found in the dolomite-filled burrows. Sulphate reduction in the burrows can also lead to an increase in Mg^{2+} by releasing the ions from neutral ion pairs (Slaughter and Hill, 1991; Wright 2005):

$$2\ CH_2O + MgSO_4^0 \rightarrow 2\ HCO_3^- + H_2S + Mg^{2+}.$$

The high concentration of these metals in the dolomite burrows compared to the calcite burrows, and compared to the matrix surrounding the dolomite-filled burrows (Figures 4-8, 4-9; Tables 4-1, 4-2), indicates the presence of SRBs and anoxic conditions during dolomite formation in the burrows. The release of the carbonate ion that occurs as a result of this process also promotes dolomitization in the burrow as does the increase in Mg^{2+}.

Marine organic matter (e.g., phytoplankton, zooplankton) contains a high percentage of proteins that are necessary for the production of ammonia, CO_2, and increase in pH and alkalinity needed for the formation of early dolomite (Wright 2000; Wright 2005). Burrows that are proximal to the shoreline, like the calcite-filled burrows in the Lonely Bay Formation, may contain some terrestrial organic matter. Terrestrial organic matter contains little protein and as a result, diagenetic changes only occur at depths greater than one kilometer, whereas protein-rich marine-derived organic matter is diagenetically altered at depths of less than one kilometer (Slaughter and Hill 1991). A trace metal indicator of marine-derived organic matter is nickel (Ni) because like ammonia, it is also derived from proteinaceous material that is present in marine organic matter. Terrestrial organic matter contains high concentrations of lignin with low amounts of proteins, resulting in low concentrations of Ni (Slaughter and Hill 1991). In the Lonely Bay Formation the Ni content in the calcite-filled burrows (8.7 ppm) is very low compared with the dolomite-filled burrows (87.6 ppm; Figures 4-8, 4-9), which points toward a higher concentration of terrestrial-derived organic matter in the calcite-filled burrows.

Stable isotope data (Figure 4-11) from the calcite- and dolomite-filled burrows along with each burrow's surrounding matrix also indicates that a higher concentration of marine organic matter was in the dolomite-filled burrows. The $\delta^{13}C_{(PDB)}$ values from the matrices surrounding the calcite-filled burrows are enriched (+0.25 to -0.45‰) relative to the $\delta^{13}C_{(PDB)}$ values for the matrix surrounding the dolomite-burrows (-1.17 to -1.39‰; Figure 4-11). Carbonates that have precipitated in waters containing more marine organic matter typically have lighter $\delta^{13}C_{(PDB)}$ values, which is seen here in the differences between the matrix surrounding the calcite-filled burrows.

The dolomite from inside the dolomite-filled burrows, with higher concentrations of Ni indicating that they contained marine organic material, had $\delta^{13}C_{(PDB)}$ values that are only slightly negative (Figure 4-11). With only a slight separation from the material from the calcite-filled burrow $\delta^{13}C_{(PDB)}$ values, the stable isotopes do not appear to agree with the trace element data. However, these slightly negative $\delta^{13}C_{(PDB)}$ values are common in Paleozoic dolomites that have been microbially influenced because the source of carbon for the dolomites comes not only from dissolved organic matter, but also from the HCO_3^- that forms in response to the increased alkalinity caused by sulphate reduction in the burrows (Wright 1997). The combination of the two carbon sources in the burrow raised the $\delta^{13}C_{(PDB)}$ values slightly higher than the $\delta^{13}C_{(PDB)}$ of the surrounding matrix.

The stable isotope data for the dolomite-filled burrows and their surrounding matrices shows a 1‰ separation between $\delta^{13}C_{(PDB)}$ values and similar $\delta^{18}O_{(PDB)}$ values (Figure 4-11). Similar $\delta^{13}C_{(PDB)}$ enrichment was seen in dolomite-filled burrows in the Tyndall Limestone (Gingras *et al.* 2004) and may reflect biologically mediated dolomite. The $\delta^{13}C_{(PDB)}$ enrichment between burrow and matrix was not seen in the calcite-filled burrows. The calcite from inside the calcite-filled burrows has similar $\delta^{13}C_{(PDB)}$ values but $\delta^{18}O_{(PDB)}$ values are heavier in the calcite matrix, possibly reflecting minor diagenetic alteration in the calcite-filled burrows.

One of the most important requirements for dolomite formation in any medium is that there has to be a conduit for dolomitization fluids (Morrow 1978; Gingras et al. 2004; Rameil 2008). Carbonate muds usually have particularly low porosities and permeability. Burrows in the Lonely Bay Formation have been backfilled with large fragments of fossil material (Figures 4-3, 4-4, 4-5, 4-6) and, in the dolomite burrows, silica sand (Figure 4-8), seen in the backscatter images (Figure 4-8A). The high concentration of burrows in the Lonely Bay Formation and textural differences (fossil material and silica) in the backfilled burrows has provided the necessary conduit for dolomitizing fluids. Seawater (Van Lith et al. 2003; Rameil 2008), cation enrichment associated with burrow linings (Over 1990; Gingras et al. 2004), and cation exchange between oxic and anoxic environments, provide a source of Mg for dolomitization.

The calcite-filled burrows that dominate the eastern portion of the Lonely

Bay Formation on the more proximal MacKenzie Basin ramp were missing many of the geochemical markers associated with the conditions necessary for dolomite formation. The REE pattern for the calcite-filled burrows are “linear” (Figure 4-10), which indicates oxic to anoxic conditions in interstitial pore waters (Sholkovitz et al. 1994; Haley et al. 2004). The trace metal concentrations in the calcite-filled burrows are very low (Figures 4-8, 4-9; Table 4-1) which is another indication of oxic conditions. In oxic conditions, any transition metals (e.g., Ni, Cu, Mn, Zn) that may have initially been present in burrow linings (Over 1990) are remobilized and bonded in the water column (Kristensen 2000). Oxic conditions may, therefore, explain the lack of trace metals and dolomite in the calcite-filled burrows. The lack of transition metals in the calcite-filled burrows (Figures 4-8, 4-9) suggests that the SRBs responsible for metal immobilization and dolomitization were not present.

The calcite-filled burrows have high trace elements concentrations that are typically associated with aluminosilicates (e.g., K, Rb, Cs, Be, Ti; Figure 4-9; Table 4-1). Clays and silicates containing aluminum have been found to inhibit sulphate reduction by 70-90% (Wong et al. 2003). This may also have inhibited dolomite formation within these burrows. There is also evidence indicating that the organic matter in the calcite-filled burrows may be of terrestrial or mixed terrestrial/marine origin. Ni concentrations, typically influenced by the presence of proteins associated with marine-derived organic matter, are very low in the calcite-filled burrows. The matrix surrounding the calcite-filled burrows is also heavier in $\delta^{13}C_{(PDB)}$, which reflects carbonates formed in waters that was depleted in dissolved inorganic matter (cf. Slaughter and Hill 1991). The proximity of these burrows to the Canadian Shield rocks may have, therefore, contributed to several factors that inhibit dolomite formation.

Conclusions

Burrows from the intensely bioturbated facies found in the Middle Devonian Lonely Bay Formation in the MacKenzie Basin, N.W.T. are filled with calcite in the area nearest the exposed Canadian Shield and filled with dolomite further down the ramp and into the basin (Figure 4-1). Geochemical characterization and comparison of calcite- and dolomite-filled burrows highlighted subtle but important differences in the diagenetic regimes that led to the development of dolomite in some burrows but not in others. The dolomite-filled burrows have an REE pattern that is enriched in the MREEs, which indicated development in an anoxic setting. The sulphate-reducing bacteria (SRB) immobilize transition metals and promoted dolomitization through release of the carbonate ion and Mg^{2+} enrichment. High Ni^{2+} concentration and ligher $\delta^{13}C_{(PDB)}$ values in the matrix surrounding the dolomite-filled burrows implied a higher concentration of marine organic matter in environment where the dolomite-filled burrows formed.

The calcite-filled burrows have a “linear” REE pattern with slight LREE and MREE enrichment, which suggests precipitation in suboxic to oxic conditions. In addition to an oxic environment, the calcite-filled burrows did not have high concentrations of transition metals that are usually found in burrows containing SRBs. Finally, the location of the calcite-filled burrows resulted in an increase in aluminosilicate content (Figure 4-9) likely sourced from the adjacent Canadian Shield (Figure 4-1). This terrigenous material may have also hindered the formation of dolomite in the calcite-filled burrows of the Lonely Bay Formation.

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CHAPTER 5: CONCLUSIONS

The Devonian inland seas were widespread and had diverse antecedent topographies, contrasting local tectonic histories, and variable access to open ocean waters. This resulted in basins with differential relative sea level histories and large reef systems with varied architecture and biological makeup. This study examined four Devonian formations from the central and eastern parts of the MacKenzie Basin in the Northwest Territories (NWT). The Chinchaga Formation, the Lonely Bay Formation, the Horn Plateau Formation, and the Horn River Formation record a history of consistent sea-level rise and subsequent reef development on a large Devonian carbonate ramp north of the well-studied Western Canadian Sedimentary Basin (WCSB). The MacKenzie Basin was directly connected to the open ocean, and did not experience significant tectonic uplift, which resulted in a history of relative sea-level change that differs from the adjacent WCSB. Some important findings regarding Devonian sea level change and the resulting stratigraphy in the MacKenzie Basin include:

1) The ocean first encroached onto northwestern Canada in the Early Devonian and carbonate deposition on the MacKenzie Basin ramp began in the Eifelian.
2) Sea level continued to rise into the Late Eifelian/Early Givetian resulting in open marine conditions and eventually isolated reef growth (Horn Plateau Formation) on the MacKenzie Basin ramp. During this time tectonic uplift in the WCSB led to the development of a platform margin. Reefs grew along the platform edge and amalgamated to form the Presqu’ile Barrier, isolating the WCSB from much of the encroaching ocean to the north.
3) Growth of 11 isolated Horn Plateau Formation reefs on the MacKenzie

Basin ramp was continuous into the Middle to Late Givetian. There is no compelling evidence of an exposure surface in the isolated reef mounds. During this time, in the adjacent WCSB, evaporites of the Muskeg Formation and Prairie Evaporite Formation formed in central part of the basin, and semi-restricted marine conditions existed along the basin's northern edge. Eventually, semi-restricted strata in the WCSB were exposed as tectonic uplift and clastic shedding continued.

4) In the NWT, during the Early Frasnian, pelagic shales were deposited and eventually formed the Horn River Formation that surrounded the Horn Plateau Formation reefs. The Horn River Formation shales were eventually overlain by the Spence River Formation shales.
5) The WCSB and the MacKenzie Basin reflect global sea-level changes differently due to the local geometry and tectonic history of the two adjacent basins.

The architecture of the MacKenzie Basin may have also contributed to the contrasting biological makeup of Horn Plateau Formation reefs along the MacKenzie Basin ramp. The reefs further out in the basin are stromatoporoid-dominated and more proximal to the paleo-shoreline are successional buildups composed mostly of corals. This presented an ideal opportunity evaluate paleoecological controls on Paleozoic reef growth. Another reef complex, in the Alexandra Formation in the southern NWT, contained one stromatoporoid dominated reef and another reef that had a microbial framework. MacNeil (2008) examined the reefs in detail and used established stratigraphy of the area and a modern analogue to conclude that a change in nutrient levels was responsible for the different biological makeup of the reefs. The control of nutrients on Paleozoic reefs is still largely unknown but Mutti and Hallock (2003) and MacNeil (2008) suggested that a multidisciplinary approach might be necessary to elucidate the control of nutrients on reef building organisms. With such excellent preservation in the Horn Plateau Formation and its two contrasting reef types, the following conclusions regarding the use of certain seawater proxies and coral versus stromatoporoid Devonian reef growth were reached:

1) Rare earth elements were successfully used to establish that both reef types grew in normal open marine conditions and confirmed the position of the coral- and stromatoporoid-dominated reefs on the MacKenzie Basin ramp.
2) With the relative depths of the reefs confirmed, $\delta^{13}C_{(PDB)}$ values could be interpreted as being a function of primary productivity, which also reflects nutrients levels on the ramp.
3) Reefs that are formed of a stromatoporoid framework require a hard substrate and reefs composed mostly of branching tabulate and rugose corals require higher carbonate sedimentation rates and a soft substrate.
4) Nutrients appear to be stratified on the MacKenzie Basin ramp with mesotrophic conditions in a more proximal position, due to coastal upwelling and possibly an additional input from a terrestrial source.
5) Stromatoporoid-dominated reefs may not have thrived in areas with higher nutrient levels, whereas Paleozoic corals might have required some nutrients.
6) Modern day Scleractinian corals may have developed a relationship with zooanthallae symbionts as a coping mechanism, during a time when nutrients were not available.

Dolomite-filled carbonate burrows have been found in other Devonian formations in the WCSB such as the Wabuman Formation in Alberta and the Alexandra Formation in the NWT. In both of these formations, the burrows have been pervasively dolomitized. The burrowed facies in the Lonely Bay Formation from the MacKenzie Basin, with some calcite- and some dolomite-filled burrows, allowed for examination and identification of conditions that may have contributed or limited the formation of early low-temperature dolomite. This study has contributed the following to our understanding of early diagenetic changes in carbonates:

1) Rare earth element (REE) patterns from the dolomite in the burrows showed higher concentrations of the middle REEs (MREEs), implying that the dolomite formed under anoxic conditions.
2) REEs revealed that the calcite-filled burrows were in a suboxic or oxic environment.
3) High Ni^{2+} concentrations in the dolomite-filled burrows are the result of protein breakdown in marine organic matter. Lighter $\delta^{13}C_{(PDB)}$ in the matrix that surrounds the dolomite-filled burrows also implies the breakdown of marine organic matter.
4) Sulphate reducing bacteria (SRB) immobilized metals in the dolomite-filled burrows and also contributed to the formation of dolomite.
5) Formation of dolomite in carbonate burrows may have been inhibited by oxic conditions and input of aluminosilicate or terrestrial organic material from the exposed Canadian Shield rocks.

With no major targets for petroleum exploration, the MacKenzie Basin Devonian strata have not been examined in great detail. In this study it became apparent that the paleogeographic setting of the MacKenzie Basin contrasts many of the inland seas and carbonate platforms that existed throughout much of the Devonian. The four formations examined in this thesis, the Chinchaga Formation, the Lonely Bay Formation, the Horn Plateau Formation, and the Horn River Formation, have revealed new information regarding Devonian sea level change and its impact in different basin styles. The distinct differences in biological makeup and mineralogy of certain facies on the MacKenzie Basin ramp have provided an opportunity to elucidate some of the ecological controls on Paleozoic reef growth and conditions of early dolomite formation in carbonate burrows.

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