Characterization of a ~1 billion year old carbonate ramp in
the Brock Inlier, Arctic Canada.

by

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Abstract

The Brock Inlier is an uplifted region east of Darnley Bay, Northwest Territories, composed of mainly carbonate and siliciclastic sedimentary rocks of the early Neoproterozoic (~1000-720 Ma) Shaler Supergroup. Strata were deposited in a large embayment (Amundsen Basin), an epeiric sea within the supercontinent Rodinia. Strata in the Brock Inlier correlate with other inliers of the Shaler Supergroup to the northeast on Victoria and Banks islands, and with the Mackenzie Mountains Supergroup, in the northern Cordillera, to the southwest. The ~900 Ma Boot Inlet Formation, occurs stratigraphically in the middle of the Shaler Supergroup and facies are interpreted to represent deposition along a storm-dominated carbonate ramp. Four transgressive-regressive sequences are preserved in approximately 400 m of section. Stable isotope analysis of 131 carbonate rock samples from the Boot Inlet Formation yield $\delta^{13}\text{C}$ values ranging from -0.2 to 7.6‰, trending positively upsection. Significant covariation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values is not observed, suggesting that diagenesis did not significantly modify the primary marine isotopic signal. $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ data coupled with lithostratigraphy and a transgressive-regressive sequence stratigraphic framework corroborate correlations between the lower Boot Inlet Formation on Victoria Island (~400km to the northeast) and the lower part of the Little Dal Group in the Mackenzie Mountains (~600km to the southwest). This establishes the broad (>1200km wide) regional distribution of a carbonate platform that extended from the Amundsen Basin, to a contemporary platform in the Mackenzie Basin. These early Neoproterozoic sedimentary rocks may record a transitional phase from low to high $\delta^{13}\text{C}$ isotopic

variability reflecting heightened sensitivity to fluxes caused by biogeochemical perturbations.

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1 Chapter: Introduction and Regional Geology

Intracratonic basins deposited within the supercontinent of Rodinia record global climatic, tectonic and paleontological variation through the late Meso- to Neoproterozoic (Halverson, 2006). The late Meso- to mid-Neoproterozoic Amundsen Basin, located in northwestern Canada comprises ~5 km of evaporite, carbonate and fluvio- deltaic siliciclastic rocks of the Shaler Supergroup (Figure 1) that are interpreted to represent deposition within an embayment connected to a much larger epeiric sea (Young, 1981; Rainbird et al, 1996; 1998, Evans, 2009).

The Boot Inlet Formation is exposed in the Brock Inlier, Northwest Territories, and in the Minto Inlier on Victoria Island and this carbonate succession is interpreted to be correlative with the Silverberry and Stone Knife formations in the Mackenzie Mountains Supergroup approximately 400 km to the southwest (Young, 1977; Rainbird et al., 1996; Turner and Long, 2012; Figure 1, 2, 3). Correlations based on chemo-, litho, and sequence stratigraphy have been proposed between Victoria Island and the Mackenzie Mountains, however the less well studied and centrally located Brock Inlier is located between these two field areas and offers an opportunity to test and refine these correlations (Figure 2).

Prior to the breakup of Rodinia, the Boot Inlet Formation was deposited on a storm-dominated carbonate ramp in a shallow, tectonically stable embayment (Amundsen Basin), (Morin and Rainbird, 1993; Narbonne et al., 2000). Deposition of the Boot Inlet

Formation within the Reynolds Point Group (Figure 1) is interpreted to have occurred during a period of transgression and prolific reef growth that interrupted deposition of fluvial-deltaic sediments represented by the underlying and overlying Grassy Bay and Fort Collinson formations, respectively (Young, 1981).

A framework combining litho-, sequence- and chemostratigraphy is presented for the Boot Inlet Formation in the Brock Inlier based on ten stratigraphic sections which tracks the depositional history of the Amundsen Basin during this period and enabling previously established correlations between the Shaler and Mackenzie Mountains Supergroups to be tested (Figure 1, 3). Previous work in northwestern Canada suggest that the terminal Mesoproterozoic leading into the early Neoproterozoic marks a transitional time period where large scale perturbations in the $\delta^{13}\text{C}$ record become more typical (Halverson, 2006). This begs the question of what changes occurred in carbon cycling that allow large isotopic excursions ($>5\text{‰}$) to become more prevalent in the Neoproterozoic $\delta^{13}\text{C}$ record as well as what driving mechanisms are responsible for such variation. Isotopic values average $\sim 0\text{‰}$ in the Mesoproterozoic with relatively low isotopic variation whereas excursions as high $+10\text{‰}$ and as low as -10‰ are globally recognized in the Neoproterozoic (Knauth and Kennedy, 2009; Swanson-Hysell et al., 2012). Given that rocks of the Amundsen and Mackenzie basins track this time period it will be important to assess how long-lasting this transitional phase is and whether or not previously reported isotopic excursions from the Mackenzie Mountains are reproducible. This work, in conjunction with micropaleontological studies of early eukaryotes may shed light on changes in biochemical carbon cycling.

1.1 Local Geological Setting

The Brock Inlier outcrops along the Hornaday River Canyon and its tributaries exposing small vertical sections (30-100 m) relative to the Minto Inlier, where continuous measureable sections of the Shaler Supergroup exceed 500 metres (Morin and Rainbird, 1993). The Boot Inlet Formation is the thickest formation in the Reynolds Point Group (Figure 1), and, in Brock Inlier, rocks are unmetamorphosed and sub-horizontal. The basal measured sections are preserved as limestone, or very fine-grained mimetic crystalline dolostone. Near the base, silicification also replaces allochems, collectively leading to high-quality preservation of the basal Boot Inlet Formation. Moving upsection, the remainder of the Boot Inlet Formation is composed of high proportions of fabric destructive dolostone. The lower contact with the underlying Grassy Bay Formation is exposed at one locality, although it is very recessive. The contact with the overlying Fort Collinson Formation is not exposed in the Brock Inlier, instead the Boot Inlet Formation is disconformably overlain by the Cambrian aged Mount Clark Formation.

A suite of uranium-lead and rhenium-osmium ages from the Shaler Supergroup constrain the depositional history within the late Mesoproterozoic to mid Neoproterozoic Amundsen Basin (Figure 1). The ages specifically constraining the Boot Inlet Formation come from the Minto Inlier. The underlying Nelson Head and directly overlying Fort Collinson formations have maximum depositional ages of 1013 ± 25 Ma and 891 ± 22 Ma, respectively, derived from uranium-lead detrital zircon geochronology (Rayner and Rainbird, 2013). The Boot Inlet Formation has a depositional age of 892 ± 13 Ma as

determined by rhenium-osmium dating on black shales from the Minto Inlier (van Acken et al., 2013), whereas the black shale unit is not observed in the Brock Inlier, in the Minto Inlier it is approximately 100 m from the overlying conformable contact with the Fort Collinson Formation.

1.2 Previous Work

The Shaler Group was originally described by Thorsteinsson and Tozer (1962) on Victoria Island (Minto Inlier), and mapping of the Brock Inlier, which included descriptions of the Shaler Group, was undertaken by Cook and Aitken (1969) and Balkwill and Yorath (1970; Figure 2). The Reynolds Point Formation was described by Young (1974, 1977) but was later elevated to Group status when the Shaler Group was elevated to Supergroup status by Rainbird et al., (1994). The lithostratigraphic assemblages described in the Minto Inlier are the rhythmic, stromatolitic and oolitic facies associations with an observed overall shallowing of facies assemblages interpreted to represent four stages of progradation (Morin and Rainbird, 1993). A detailed description of the spectacular stromatolite reefs from the Boot Inlet Formation on Victoria Island is given by Narbonne et al. (2000). Chemostratigraphy using $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ is described for the Minto Inlier from the Aok to the Wynniatt formations (Thomson et al., 2015a; Jones et al., 2010).

Correlations based on lithostratigraphic similarities between the Shaler and Mackenzie Mountains supergroups and the Wernecke Mountains were proposed by

Young (1977; 1982) Aitken et al., (1978a; 1978b) and Young et al., (1979). The “orange-weathering stromatolite biostome”, which characterizes the Aok Formation in the Shaler Supergroup, is observed in the Mackenzie Mountains and Western Arctic Canada representing an important marker bed in the regional lithostratigraphic framework (Jefferson and Young, 1989). Chemostratigraphic and sequence stratigraphic frameworks have been proposed that build upon this foundation (Long et al., 2008; Macdonald et al., 2012; Thomson et al., 2015a, b; Jones et al., 2010).

1.3 Methods

Ten stratigraphic sections measure the upper Nelson Head, Aok, Grassy Bay and Boot Inlet formations, with a focus on the Boot Inlet Formation. Sections were measured using a Jacobs staff and stable isotope samples were collected at 1-1.5 m intervals from the Aok and Boot Inlet formations as well as representative lithologies from each formation for additional petrographic analysis. Microfossil samples were collected from the upper Nelson Head and Aok formations, and are outside the scope of this project. A correlation of the measured stratigraphic sections from the Brock Inlier is presented in Appendix A.12.

Hand samples, collected for stable isotope analysis, were cut in half with a lapidary saw at Carleton University in order to expose fresh surfaces for facies analysis and for stable isotope sampling. Stable isotope samples were analyzed at the Stable Isotope Laboratory at McGill University, Montreal. 51 unpolished thin sections of

representative lithologies were analyzed in plane and cross-polarised light to characterize lithofacies and assess diagenetic alteration. Thin sections were stained with potassium ferricyanide and alizarin red to help determine their mineralogy.

Figure 1: Generalized stratigraphy of Shaler Supergroup, after Thomson et al., (2014) and Rainbird et al., (1994). Note the vertical black line denoting the exposed portion in the Brock Inlier. Age constraints from Rayner and Rainbird (2013), van Acken et al. (2012), and Heaman et al. (1992).

Figure 2: Amundsen Basin inliers and Mackenzie Mountains, northwestern Canada after Young et al. (1979). Orange stars denote locations with exposures of the Boot Inlet Formation. Thick dotted line represents hypothetical former margin of the Amundsen Basin (Young et al., 1981). White dotted line shows outline of Figure 4.

Mackenzie Basin			Amundsen Basin		
		Formations	Formations		
Mackenzie Mountains Supergroup	Little Dal Group	Ram Head	Wynniatt		Shaler Supergroup
		Snail Spring			
		Ten Stone			
		Gayna	Fort Collinson/ Jago Bay	Reynolds Point Group	
		Stone Knife/ Silverberry	Boot Inlet		
		Dodo Creek	Grassy Bay		
	Katherine Group	Abraham Plains		Rae Group	
		McClure	Aok		
		Shattered Range	Nelson Head		
		Etagochile			
		Grafe River			
		Tawu			
		Eduni			
		Tsezotene			
		Tabasco	Mikkelsen Islands		

Figure 3: Nomenclature and proposed correlation between Mackenzie and Amundsen basins (after Long and Turner 2014; Long et al., 2008; Rainbird et al., 1994).

Figure 4: Brock Inlier field area with locations of measured sections. Formation data after Okulitch, (2000). Stratigraphic sections, their GPS coordinates and basal correlation shown in Appendix A1-A12.

2 Chapter: Lithostratigraphy

The purpose of this chapter is to strip away artifacts of the post depositional history of the Boot Inlet Formation that do not represent primary or early sedimentation, focusing on original processes that took place on the ancient carbonate platform, shortly following lithification. This chapter is divided into part 1, which describes the lithofacies and facies associations; and part 2, which uses part 1 information to describe the stratigraphy, while discussing possible mechanisms causing upsection variation. Stratigraphic sections and basinal correlation are presented in Appendix A. The stratigraphy provides a foundation for sequence stratigraphic and chemostratigraphic analysis and, eventually, regional correlations.

2.1 Lithofacies descriptions

The Boot Inlet Formation, can be subdivided into eight distinct lithofacies summarized in Table 1, using the Dunham classification system (Dunham, 1962) supplemented with the term “calcsiltite” denoting mechanically transported allochems between ~5 - 62.5 μm after Kay (1952).

Table 1: Facies descriptions of the Boot Inlet Formation

Facies	Lithology	Sedimentary Structures	Additional Features	Depositional Environment
1. Stromatolite biostrome	Dolomite	Large domal to tabular morphology tens of meters thick	Laminations with high Fe oxide/organic content interbedded with fine crystalline dolomite (<62um)	Mid-ramp, above storm wave base
2. Stromatolite bioherm	Dolomite	Domal features approximately 2 m across	Medium-coarse dolomite crystals	Inner-ramp, shallow water above fair-weather wave base.
3. Intraclastic grainstone/rudstone	Limestone-Dolomite	Framework supported stromatoclasts	Framework supported. Crystal size of cements differs from that of allochems in recrystallized thin sections	Inner-, mid-ramp, in close association with stromatolites
4. Ooid grainstone	Limestone-Dolomite	Elongate intraclasts present, often cross-bedded. Stylolitized, locally compacted. Herring-bone cross-stratified near top of Boot Inlet Formation	Chert replacement some allochems are ghosts.	Inner-ramp, high-energy environment
5. Laminated mudstone with molar-tooth structure	Limestone-Dolomite	Mudstone with molar-tooth structures (MTS) disrupting laminations. MTS are lighter colour than host, brecciated into elongate molar-tooth intraclasts (>2mm) that form interbeds.	Nodular pyrite	Inner- to mid-ramp, above storm wave base, low-energy environment aside from intermittent storm activity
6. hummocky cross-stratified calci/dolo-siltite	Limestone-Dolomite	Hummocky cross-stratified (HCS) to planar cross-stratified carbonate silt sized carbonate grains	Rare detrital quartz horizons.	Mid-ramp, between fair-weather wave base, shallower than facies 5.
7. Dolo/lime mudstone	Limestone-Dolomite	Laminations that have pinch-and-swell structures, organic rich.	Laminations with horizons rich in organic/Fe oxide content	Outer-ramp; below storm wave base
8. Shale	Mudrock, Claystone	Laminations	Laminations	Outer-ramp; below storm wave base

2.1.1 Facies 1: Stromatolite biostrome

Description. Two thick stromatolite biostromes (section 7 and 8) appear to represent one regionally extensive reef structure (Appendix A7 and 8). Bedding is laterally extensive and tabular with thickness ranging from 0.3-2 m. At section 7, biostromes have an uninterrupted thickness of approximately 85 m (Figure 5A). Stromatolite bioherms underlie and grade into thick, tabular biostromes. Stems are unbranching and resemble the morphology of *Baicalia* to *Jurusania* (Narbonne et al., 2000; Walter, 1972; Figure 5B, C). All observed biostromes in the Brock Inlier have been recrystallized to crystalline dolomite. Petrography reveals that laminations are steeply convex and typically are accentuated by organic matter or Fe-oxides. Minute amounts of detrital quartz (modal percentage < 5) are trapped within stromatolite laminations, but are more commonly observed above them. Horizontal stylolites following microbial laminations, or in the underlying beds, are common.

Interpretation. Stromatolite biostromes are interpreted to have been deposited as low amplitude sheets just below fair-weather wave base during periods of platform stability allowing for complex regional biogenic structures to develop. When stromatolites grew above fair-weather wave base, constant wave agitation eroded their tops, supplying clasts (stromatoclasts) to the ramp.

The initial growth by isolated bioherms (Figure 4E) overlain by continuous stromatolite sheets (biostromes) is similar to what is described from the Boot Inlet

Formation from the Minto Inlier (Narbonne et al. 2000; Figure 4B). Their work likens the transition from mounds to sheets to “catch-up” reefs as described by Neumann and Macintyre (1985). The low-amplitude domal morphology (~5 m synoptic relief) overlain by tabular sheets of columnar stromatolites observed at section 8, resembles the shallowing-upward stromatolite cycles described from the Bitter Springs Formation in Australia by Southgate (1989) but in the Amundsen Basin they are on a much larger scale. These previous interpretations suggest that the change in morphology of microbial communities preserved in these stromatolites represent shallowing. Biostromal reef complexes of the Boot Inlet Formation are potentially correlative between the Minto and Brock inliers, suggesting that this could have represented a basin-scale shallowing characterized by a period of prolific reef growth. Topographic relief of the largest complexes is interpreted by Narbonne et al. (2000) to be no greater than 3-5 m in the Minto Inlier and this is supported in the Brock Inlier where the synoptic relief is no more than 5 m. Collectively, this implies that while these reefs were regionally extensive and thick, they did not appear to form barriers to cause restricted depositional environments.

Figure 5: Facies 1 and 2 (stromatolite biostromes and bioherms). A: Sheets of tabular stromatolite biostromes, person for scale. B: Thickly bedded stromatolite biostrome, hammer for scale. C: Cross section view of stromatolites with morphology similar to *Jurusania* (Walter, 1972). D: Top view of stromatolite (*Baicalia*). E, F: Stromatolite bioherms.

2.1.2 Facies 2: Stromatolite bioherms

Description. Bioherms are ~1-3 m in diameter, appearing as resistant rounded mounds with ~0.3-1 m synoptic relief, composed of columnar stromatolites. They are observed at the base of section 3, the middle of section 7 and 8 and the top of section 4 and 10 (Figure 5E, F; Appendix A3, A4, A7, A8 and A10). Individual stromatolites resemble *Tungussia* as described by Walter (1972), while areas between bioherms are infilled with grainstone/rudstone facies composed of stromatoclasts and ooids.

Interpretation. Bioherms would likely have formed in inner-ramp settings on the basis that they are largely found in conjunction with cross-bedded ooid grainstones. Bioherms appear to be overlain by deeper water facies, and two responses to sea-level change are observed. Observed at the base of section 3, infill by normally graded sediments overlain by clays suggests they were drowned, too quickly for stromatolite growth to continue. The lack of karsting features caused by dissolution indicate that bioherms were never subaerially exposed. Bioherms at section 7 and 8 are observed to be overlain by biostromes likely representing a rise in sea-level, but a gradual one that would facilitate their growth. It therefore appears that the facies overlying bioherms in the Brock Inlier reflect sea-level rise, although the rate of sea-level rise in the Amundsen Basin resulted in two different sedimentation responses.

2.1.3 Facies 3: Intraclast rudstone

Description. Intraclast rudstone facies, best observed at the base of section 3 is characterized by medium- to thickly bedded, commonly dune cross-stratified stromatolclast rudstones that are mixed with ooids and aggregates, most commonly composed of smaller ooids (Figure 6A, B, E). Clasts do not express a preferred orientation. Allochems are framework-supported and isopachous cements are commonly visible where rocks have not been dolomitized; remaining voids are filled with equant, blocky calcite cement. This facies is most commonly associated with stromatolite bioherms.

Interpretation. This poorly sorted facies contains the largest allochems observed in the Boot Inlet Formation, that appear to have been shed from stromatolites. Above the fair-weather wave base (FWWB), bioherms would be subject to constant wave agitation as would biostromes if they aggraded to this zone, resulting in the shedding of stromatoclasts. The fragility of grains, paired with the lack of a consistent orientation suggests that shed material mixed with ooids in a high-energy ephemeral event where reworking was minimal. Deposition of intraclast rudstone facies would have occurred very close to stromatolites, mostly in inner-ramp to the uppermost mid-ramp depositional environments.

Figure 6: Facies 3 and 4 (Intraclast grainstone/rudstone and ooid grainstone). A: Intraclast grainstone, scale bar is 4 cm. **B:** Plan view exposure of intraclast rudstone facies. **C, D:** Ooid grainstone facies. **E:** Thin-section photo of intraclast rudstone. **F:** Ooid grainstone showing “elephant parade” fabric (Scholle and Ulmer-Scholle, 2003). Notches in E and F are each 1 mm.

2.1.4 Facies 4: Ooid grainstone

Description. Ooid grainstone facies, composed of ~1 mm ooids, is best observed at sections 3, 4 and 5 (Fig 6C, D, F), (Appendix A3, A4, A5). Rare stromatoclasts, detrital quartz, and aggregates of small ooids are present. They are generally cross-bedded, with bedding thickness between approximately 5 and 30 cm (Figure 6C). Certain horizons display “elephant parade” fabric (Scholle and Ulmer-Scholle, 2003; Figure 6F). In section 5, ooids are rarely replaced by mimetic chert.

Interpretation. Ooid grainstone facies would have been deposited in an inner-ramp setting above the fair-weather wave base. The formation of ooids relies on wave agitation, in warm seawater highly saturated with calcium carbonate. Rare remineralization of ooids to chert preserves their concentric laminations (Section 5), as does early mimetic dolomitization (Section 9). Silicification, if occurring during early diagenesis, may imply that a range of fluid compositions persisted (Manning-Berg and Kah, 2017). Ooid grainstone beds are less common in the Brock Inlier than in the Minto Inlier (cf. Morin and Rainbird, 1993), as they did not form large shoals, potentially due to unfavourable seawater chemistry or a lack of wave-energy.

2.1.5 Facies 5: Molar-tooth mudstone

Description. Molar-tooth mudstone facies is very common in the Boot Inlet Formation in the Brock Inlier and is observed in section 3, 4, 6, 7, 8, and 10. This facies consists of small ptymatically folded, light-coloured molar-tooth structures (MTS) hosted in laminated lime- or dolomudstone composed of rounded grains (4-15 μm in diameter; Figure 7F) with rare framboidal pyrite. Molar-tooth structures are filled with coarse, white calcite spar or fine (4-15 μm) rounded grains of calcite and are commonly recrystallized to crystalline dolomite. Laminated mudstone layers are interbedded with 5-50 cm thick massive to cross-bedded rudstone beds with erosive basal contacts, composed of elongate, poorly sorted molar-tooth structure clasts (Figure 7A). Grain size in the host rock does not vary greatly, laminations are instead visible due to the presence of dark interstitial material along laminations. Molar-tooth mudstone facies is observed in close association with or are interbedded with thin shaley units (Figure 7C), bioherms and biostromes.

Interpretation. Molar-tooth structure is an enigmatic feature and its origin has been debated for over a century. Two popular interpretations include seismically induced fluid escape structures (Pratt, 1998) and biogenically induced gas expansion cracks (Furniss, 1998). The Boot Inlet Formation was deposited within a very stable intracratonic basin and given that there are no other structures representing seismic activity, the latter hypothesis is favoured. Molar-tooth structure does not appear to form when the host rock's grain size exceeds approximately 15 μm .

The presence of MTS-clast rudstone interbeds with erosive contacts suggest molar-tooth mudstone facies must have formed above storm wave base. Given their close association with shallow and deep-water facies, the depositional environment of molar-tooth mudstone facies appears to range from SWB to FWFB. The timing of MTS formation is interpreted to be prior to lithification of the host muds, as they are preserved as lithified clasts in rudstone interbeds.

Figure 7: Facies 5 (molar-tooth mudstone facies). A: Molar-tooth mudstone interbedded with cross-bedded molar-tooth clast rudstone. B, D: Plan view exposure of molar-tooth structure. C: Molar-tooth mudstone facies overlain by shaley interval. E: Hand sample showing ptymatically folded molar-tooth structure. F: Thin section photo of molar-tooth structure, scale bar is 1 cm.

2.1.6 Facies 6: Hummocky cross-stratified calci/dolo-siltite

Description. This facies is common at sections 3, 4, 5 and 9 and is composed of limestone, mimetic and fabric destructive dolomite expressing ripple, dune and hummocky cross-stratification (HCS), (Figure 8A, B, C). Weathered, dolomitized surfaces have an orange cast and primary structures are difficult to discern. Well-preserved specimens that are composed of calcite or that have been recrystallized by fabric-retentive dolomite show rounding of small, silt- to sand-sized grains. MTS is distinctly absent from this facies. Siltite often contains small ooids (<1mm), forming wackestone and packstone fabrics (Figure 8F).

Interpretation. Facies comprising predominantly hummocky cross-stratified siltite suggests a depositional environment influenced by storm activity. MTS are absent, the formation of which is interpreted to be inhibited by the larger grain size. Allochems are interpreted to have been sourced from stromatolites, and mixed with silt-sized material including ooids that were deposited on the upper-mid to inner-ramp.

Figure 8: Facies 6, 7 and 8 (Hummocky cross-stratified calci/dolosiltite and laminated mudstone facies interbedded with shales. A, B, C: Field exposure of dolosiltite/arenite. D, E: Laminated dolomudstone, E expressing pinch-and-swell structures. F: Ooid wackestone.

2.1.7 Facies 7: Laminated mudstone

Description. Facies 6 is uncommon in the Boot Inlet Formation and it is best observed at the base of section 4, where it is approximately 5 m thick (Figure 8D). It is composed laminated carbonate mudstone (grain size <5 µm), that lack MTS and wave-generated structures. Colour variation between laminations appears to be caused by dark interstitial material in the matrix, rather than a change in grain size. Primary fabrics have been replaced by fine, crystalline dolomite. Pinch-and-swell structures (Figure 8E) occur in this facies and are observed near the bases of section 3 and 4.

Interpretation. Laminated mudstone facies would have been deposited below SWB, in the lowest-energy depositional environment of the Boot Inlet Formation on the outer ramp below storm wave base. This is supported by the fine grain size and the lack of any wave generated structures.

2.1.8 Facies 8: Shale

Description: Shale facies occur as thin laminations that predominantly form thin interbeds, and are best exposed at the base of section 3 and 4. Green, friable claystones observed at Section 3 also compose this facies. This facies is commonly associated with laminated mudstone facies and rarely with molar-tooth mudstone facies. Shale facies weathers very recessively.

Interpretation. The shale facies occur as compressed thin layers between facies 7 and in close association with molar-tooth mudstone facies. Fine terrigenous wash-load sediment is interpreted to have been transported offshore and deposited in the low-energy outer-ramp environment through suspension fallout.

2.2 Lithofacies associations

The basinal setting of the Shaler Supergroup is interpreted to be an embayment of a larger intracontinental sea within the supercontinent, Rodinia (Young, 1981). Facies assemblages reflect deposition around the periphery of the embayment, which was affected by tides, storms and wind-generated waves comprising the inner-ramp (facies association 1); where water was calm aside from periodic storm activity (facies association 2), and the outer-ramp position (facies association 3), that was not influenced by these processes. The depositional model of the Boot Inlet Formation is summarized in Figure 9.

Figure 9: 3D depositional model of the Boot Inlet Formation.

2.2.1 Facies association 1: Inner-ramp

The inner-ramp depositional environment would have been subject to constant wave agitation and comprises the area of the ramp between upper shoreface and fair-weather wave base (Burchette and Wright, 1992). Ooids (facies 4) and stromatoclasts (facies 3) would have represented the dominant grain types. Early-cemented substrates or topographic highs would have been colonized by stromatolite bioherms, which shed stromatoclasts during storm-surges. Some of these clasts would have been lithified, forming facies 3, whereas many were reworked into sand- and silt-sized carbonate clasts and transported to the mid-ramp. Bioherms may have caused local restriction, although inter-bioherm channels are observed, suggesting that water flowed freely between them (Figure 10). The inner-ramp is also interpreted as entirely subtidal, as there is no evidence of subaerial exposure, such as desiccation cracks or evaporites.

Figure 10: Facies association 1: inner-ramp. Stromatolite bioherms (orange outline) surrounded by inter-reef channels (yellow). Hammer for scale.

2.2.2 Facies association 2: Mid-ramp

Stromatolite biostromes (facies 1), hummocky cross-stratified calcisiltite (facies 6), and molar-tooth mudstone (facies 5) comprise facies association 2. The mid-ramp facies association would have been reworked by storms and swells and represents the area on the ramp between fair-weather wave base and storm wave base (Burchette and Wright, 1992).

The stromatolitic portion of the mid-ramp facies association is observed at sections 7 and 8 and is composed of facies 1 representing large, regional biogenic structure (Figure 12A). They are observed stratigraphically above stromatolite bioherms (facies 2), forming table reefs (Narbonne et al., 2000). Their position on the ramp is constrained to below fair-weather wave base as it is interpreted that constant wave agitation would erode the microbial films of biostromes. Given the great thickness (~90 m) and lateral extent of the biostromes, they must represent long periods of relatively calm water and platform stability. These reefs show great affinity to the “type 4” stromatolite boundstones reported in the platformal assemblage of the Little Dal Group from northwestern Canada, interpreted to have formed in quiet water conditions along the mid-ramp (Batten et al., 2004). The increased turbidity and unconsolidated substrates resulting from mud in the water column may have also had detrimental effects on the nucleation and growth of stromatolites which implies that they were most prolific during periods of decreased carbonate production (Sherman et al., 2000). Periods of platform stability would have facilitated the formation of complex microbial structures for long

periods of time and when they aggraded to fair-weather wave base constant wave agitation would plane them (Southgate, 1989). During relative sea-level rise, it appears that the morphology *Jurusania* became dominant. Once stromatolite biostromes developed, it appears that they were capable of keeping pace with sea-level rise.

The distal portion of the mid-ramp facies association is composed of facies 5 (molar-tooth mudstone facies) and facies 6 (hummocky cross-stratified calcisiltite facies). HCS calci-dolo-siltite grade offshore into molar-tooth mudstone facies. Molar-tooth mudstone facies is observed in association with outer-ramp deposits (facies association 3) suggesting that they were deposited close to storm wave base. During storms, semi-consolidated carbonate mud containing molar-tooth structure were reworked into molar-tooth clast rudstone interbeds. Mid-ramp deposits are interpreted to have been sourced from stromatolites, being mechanically transported during storm activity and reworked into facies 5 and 6.

Figure 11: Facies association 2: mid-ramp. A: Molar-tooth mudstone interbedded with molar-tooth clast rudstone. B: HCS calcisiltite overlying ooid grainstone.

2.2.3 Facies association 3: Outer-ramp

Facies association 3 is composed of facies 7 and 8 and is characterized by very fine grain size, thin parallel lamination and no wave or molar-tooth structure (Figure 11b); implying low rates of sedimentation below storm wave base. It is not volumetrically abundant but occurs above facies association 1 (at the base of section 3) and facies association 2 (at the base of section 4). The latter observation represents an overall shallowing from outer- to mid- to inner-ramp deposits. Its occurrence above inner-ramp deposits at section 4 implies a rapid rise in sea-level. The lack of wave structures implies a very quiet-water depositional environment, likely below storm wave base, therefore representing the most off-shore depositional environment represented by strata of the Boot Inlet Formation in the Brock Inlier.

Figure 12: Facies associations 2 and 3 (mid- and outer-ramp). A: Low amplitude stromatolite biostrome, height of broad domes is approximately 5 m. B: thinly laminated dolomudstone.

2.3 Lithostratigraphy

The Boot Inlet Formation can be subdivided into six distinct, informal stratigraphic members that are easily recognizable in the Brock Inlier (Table 2). The Boot Inlet Formation is mainly exposed in cliff sections along the Hornaday, Little Hornaday and Roscoe rivers; plan view exposures are very limited. Thick stromatolites of the second stromatolite member form prominent features throughout the inlier, acting as a useful stratigraphic marker horizon. The stratigraphic record of the Boot Inlet Formation in the Brock Inlier is summarized into an approximately 400-metre-thick composite section with small gaps of estimated thicknesses (Figure 13). The lower contact with the Grassy Bay Formation is gradational, although it is quite recessive; the upper contact with the Fort Collinson Formation is covered. In the Hornaday River canyon, the upper part of the formation is disconformably overlain by the Cambrian Mount Clark Formation (Figure 13).

Table 2: Summary of informal stratigraphic members of the Boot Inlet Formation in the Brock Inlier. See Figure 13 for stratigraphic position and thickness.

Member	Lithology	Basal Contact	Thickness
Mixed molar-tooth mudstone member	Biohermal stromatolites at base and top. Lime/dolomudstone with abundant molar-tooth structure is dominant. Herring-bone cross-stratification and detrital quartz content increase upwards.	Not exposed	~70 m at section 10
Second stromatolite member	Base of unit is composed of stromatolite bioherms that are overlain by molar-tooth mudstone facies. Thickly bedded stromatolite biostromes comprise the upper 70 m.	Conformable	~90 m at section 7
Dolostone member	Predominantly orange, fabric-destructive dolostone. Begins laminated greenish claystones near base. Middle contains 5 m unit of thinly laminated dolomudstone containing pinch-and-swell structures. Upward decrease in shale interbeds. Microbialites, and grainstone becomes more abundant near the top.	Drowning unconformity	~80 m at section 3 and 4
First stromatolite member	Bioherms with inter-reef channels composed of grainstone. Beds of intraclast rudstone, ooid grainstone with abundant cross-stratification.	Not exposed	≤5 m at section 3
Molar-tooth lime mudstone member	Molar-tooth mudstone interbedded with intraclast rudstone/ grainstone facies. Thickly bedded intraclast rudstone at base.	gradational	~40 m at section 6
Calcsiltite member	Hummocky cross-stratified calcsiltite. Base contains quartz silt/arenite interbeds; ooid packstone/grainstone facies interbedded near the top.	Erosional contact with Grassy Bay Formation	~40 m at section 9

Figure 13: Lithostratigraphic members of the Boot Inlet Formation from the Brock Inlier. Composite section shown, basinal correlation shown in appendix A.12. Sandstone at top represents Cambrian aged Mount Clark Formation.

2.3.1 Stratigraphic members of the Boot Inlet Formation

2.3.1.1 Description

The calcisiltite member is a recessive unit approximately 40 m thick that is best exposed at section 9, where it gradationally overlies the Grassy Bay Formation. It begins with a dolomudstone-intraclast conglomerate that was also observed in the Minto Inlier (Young and Long, 1977; Morin and Rainbird 1993). In the Brock Inlier, dolomudstone intraclasts are faintly laminated. This conglomerate is overlain by calcisiltite and dolosiltite (facies 6) interbedded with hummocky cross-stratified, fine grained quartz arenite. Upsection, quartz sand content diminishes and ooid wackestone and grainstone interbeds (facies 4) become more abundant. This transition is observed at section 5, where the calcisiltite member is erosively overlain by molar-tooth-clast rudstone beds characterizing the molar-tooth lime mudstone member.

The base of the molar-tooth lime mudstone member is exposed at section 5, where it is cut by a conformable diabase sill that can be used as a stratigraphic marker to infer that the gap between it and overlying strata at section 6 is likely under 20 m. Section 6 is approximately 40 m thick and comprises molar-tooth lime mudstone cyclically interbedded with molar-tooth-clast rudstone (facies 5).

The first stromatolite member is exposed at the base of section 3, along the Hornaday River (Figure 10), but its contact with the underlying molar-tooth lime

mudstone member is not exposed. The first stromatolite member, a thin unit of approximately 5 m, is composed of laterally-linked metre-scale bioherms composed of resistant stromatolites resembling *Baicalia*. Bioherms are surrounded by ooid grainstone facies with symmetrical wave-rippled tops, and stromatolite-clast rudstone facies. These facies are best-preserved in channels between bioherms, where synoptic relief is greatest. The first stromatolite member is abruptly overlain by the dolostone member.

The dolostone member is approximately 70 m thick, with the lower portion being exposed at section 3, and the upper half at section 4 (Appendix A3 and A4). It begins with thinly laminated green clays interbedded with shales, with abundant pinch-and-swell structures (Figure 8E). This member is characterized by orange-weathering fabric-destructive dolostone with irregular bedding thicknesses, and molar-tooth structure is most evident on fresh surfaces. The base of section 4 begins with 5 m of rhythmically-laminated dolomudstone facies with pinch-and-swell structures (facies 7). This is overlain by interbedded hummocky cross-stratified calcisiltites (facies 6), molar-tooth mudstone (facies 5), microbialites and ooid grainstones (facies 4) characteristic of the inner- to mid-ramp facies association. The top of the dolostone member is conformably overlain by the second stromatolite member.

The second stromatolite member is a resistant, well-exposed unit that serves as a prominent stratigraphic marker in the Brock Inlier. It was measured at sections 7 and 8 but is prominent throughout the inlier (Figure 5A, 12A). At section 8, two metres of stromatolite bioherms (facies 2) are overlain by five metres of molar-tooth mudstone

facies (facies 5) with abundant small-scale cross-bedding in the molar-tooth clast rudstone interbeds (Appendix A8). These are, in turn, overlain by a thick (at least 85 m at section 7) (Appendix A7), domal to tabular-bedded, stromatolitic reef complex (facies 1). Reefal morphologies shift from metre-scale bioherms (facies 2) to larger low amplitude domal structures with widths greater than 500 m (Figure 12a) to flat, tabular biostromes (Figure 4A). Bioherms and domal reefs are composed of *Baicalia* stromatolites that grade upwards into tabular biostromes composed of *Jurusania* (Walter, 1972; Figure 5C). Tabular-bedded biostromes cap the top of section 7 and 8. There is a vertical gap of approximately 100 m between the second stromatolite member in section 7 and the overlying mixed molar-tooth mudstone member in section 10 (Figure 12).

The mixed molar-tooth mudstone member is only exposed at section 10, where it is approximately 70 m thick and is disconformably overlain by the Cambrian aged Mount Clark Formation (Appendix A10). It begins with dolomudstone, overlain by approximately 3 m of poorly preserved stromatolite bioherms (facies 2). An 18 m covered interval separates these stromatolites from the overlying interbedded molar-tooth lime- and dolo-mudstone. Gradually, over 50 m, detrital quartz content increases in the fabric of carbonate facies and herring-bone cross-stratified ooid grainstone becomes more abundant near the top of this member. The last exposed portion of the mixed molar-tooth mudstone member is composed of 2 m of stromatolite bioherms.

2.3.1.2 Interpretation

The basal conglomerate at the base of the calcisiltite member is interpreted as a transgressive lag on the basis that it is regionally extensive and it separates fluvial-deltaic siliciclastic deposits (Grassy Bay Formation) from carbonate rocks representing a deeper-water, mid-ramp depositional environment (Boot Inlet Formation) and because the intraclasts are derived from a deeper-water, outer-ramp setting. The succession from the Grassy Bay Formation to the top of the molar-tooth lime mudstone member marks a shift to a fully marine carbonate ramp. The calcisiltite member reflects an abrupt decrease in siliciclastic sediment flux relative to carbonate sedimentation. The molar-tooth lime mudstone member represents quiet water deposition along the mid-ramp with frequent storm reworking as indicated by the accumulation of molar-tooth clast rudstone facies.

The first stromatolite member represents shallow water, relatively high-energy deposition based on the presence of abundant ooids, stromatolite bioherms and inter-reef channels characteristic of facies association 1. The transition from the molar-tooth lime mudstone member is not exposed but it is interpreted as a progradation of the carbonate platform as molar-tooth mudstone facies aggraded towards sea-level, and facies stacking patterns display a shift from mid- to inner-ramp.

The transition from the first stromatolite member to the dolostone member is interpreted to represent a rapid sea-level rise that drowned stromatolite growth analogous to “give-up” reefs described by Neumann and Macintyre (1985). The lower half of the dolostone member (exposed at section 3) is composed predominantly of facies 5, 6, 7 and

8 representing mid to outer-ramp deposition. At section 4, planar and trough cross-bedding as well as HCS in rudstones become common upsection, as does detrital quartz and the occurrence of microbial laminites. A transition from facies association 3 to 2 to 1 through section 4 represents an overall shallowing as the carbonate platform gradually progrades, eventually facilitating the development of inner-ramp stromatolite bioherms (facies 2) that mark the base of the second stromatolite member.

The second stromatolite member is laterally extensive in the Brock Inlier and represents the most prolific period of reef growth. Table reefs that are observed at section 7 and 8 resemble those which occur near the transition from inner- to mid-ramp from the Boot Inlet Formation in northeastern Minto Inlier (Narbonne et al., 2000). The change in reef morphology resembles stacking patterns observed in the Loves Creek Member of the Bitter Springs Formation in the Amadeus Basin of central Australia, where they are interpreted to represent upward shallowing (Southgate, 1989). Collectively, the shift from metre-scale bioherms (facies association 1), to molar-tooth mudstone facies to broad biostromes (facies association 2), is interpreted to represent an initial rise in sea-level and shift to a mid-ramp depositional environment during a period of platform stability capable of supporting a complex stromatolite reef. Subsequently, the shift from broad domal biostromes to tabular ones is interpreted to represent more rapid sea-level rise resulting in the proliferation of a “catch-up” reef as described by Neumann and Macintyre (1985).

Increased quartz content, herring-bone cross-bedding and stromatolite bioherms comprise the mixed molar-tooth mudstone member and suggest shallow water, higher-energy deposition characteristic of facies association 1. Herring-bone cross-bedded ooid grainstone characteristic of the upper Boot Inlet Formation (Morin and Rainbird, 1993) were observed near the top of section 10, likely indicating that section 10 is very close to its upper contact with the Fort Collinson Formation, which is typified by herring-bone cross-bedded quartz arenite (Rainbird et al., 1994). The Fort Collinson Formation is not well developed in the Brock Inlier.

Lithostratigraphy of the Boot Inlet Formation reveals an initial gradual progradation of the carbonate platform as siliciclastic input from the Grassy Bay Formation diminished during an overall sea-level rise. Facies association stacking patterns indicate that the platform then underwent four, shallowing upward cycles. The platform aggraded towards sea-level in the fourth shallowing upward cycle, siliciclastic input began to increase, resulting in deposition of the Fort Collinson Formation.

3 Chapter: Sequence Stratigraphy of the Boot Inlet Formation

3.1 Introduction

Proterozoic sedimentary rocks of northwestern Laurentia were divided into three first order sequences (A, B, C) by Young et al. (1979). The Shaler Supergroup in the Amundsen Basin comprises Sequence B and the Reynolds Point Group matches the upper portion of sequence B2 (Grassy Bay Formation) and sequence B3 (Boot Inlet, Fort Collinson and Jago Bay formations). Subdivisions of sequence B are outlined by Long et al. (2008). The transition from sequence B2 to B3 is interpreted as a conformable flooding surface (Long et al., 2008).

The sequence stratigraphy for the Reynolds Point Group has not previously been described in the Brock Inlier; however, in the Minto Inlier on Victoria Island, alternating deep and shallow water facies define five, transgressive-regressive sequences (Morin and Rainbird, 1993; Figure 18). In their analysis, the Boot Inlet Formation is divided into transgressive-regressive carbonate sequences that are gradational over several metres with no abrupt change in sedimentation at sequence boundaries which are placed at subaerial unconformities (Morin and Rainbird, 1993). The scale of sequences from the Shaler Supergroup has been compared to that of third-order cycles observed in Phanerozoic successions (Rainbird, 1991), and the Reynolds Point Group is interpreted to record global eustatic sea-level fluctuations within a passive subsidence regime (Morin and Rainbird, 1993; Narbonne et al., 2000).

Transgressive-Regressive (T-R) sequence stratigraphy (e.g. Embry 2002) is used to describe the relationship between sedimentation and accommodation space in the Brock Inlier. These types of sequences were originally defined by Embry and Johannssen (1992) and discussed in greater detail by Embry (1993, 1995, 2002). Sequence boundaries (SBs) are placed at the subaerial unconformity (SU) or co-terminus maximum regression surface (MRS), while the maximum flooding surface (MFS) divides depositional sequences into the lower transgressive systems tract (TST) and upper regressive systems tract (RST) (Embry, 2002). These relationships are characterized by changes in base level, which defines a surface of equilibrium between erosion and deposition; and furthermore does not attempt to separate individual variables such as tectonism and eustasy as key drivers, but rather describes the net effect as it relates to sedimentation (Embry, 2002). Thicknesses of T-R sequences in the Brock Inlier and their relationship to facies trends are illustrated in Figure 16.

3.2 Results:

3.2.1 Sequence 1

The contact between the Grassy Bay and Boot Inlet formations is marked by a regionally extensive transgressive lag deposit, which is overlain by hummocky cross-stratified dolosiltites (facies 6) and then molar-tooth mudstone. This sequence is interpreted to constitute a TST. The MFS and lower portion of the RST are not exposed aside from the very top of the RST which is observed at the base of section 3, where stromatolite bioherm and ooid grainstone facies interpreted as inner-ramp deposits (facies

association 1) are observed. A sequence boundary is placed at the contact between these inner-ramp facies and overlying claystones, with pinch-and-swell structures, interpreted as outer-ramp deposits representing a period of rapid base level rise.

3.2.2 Sequence 2

Claystones overlain by molar-tooth dolomudstone facies are exposed at section 3 and define the lower portion of a TST and the base of Sequence 2. The TST is capped by laminated dolomudstone facies exposed at the base of section 4 and the transition to the overlying strata represents the MFS. The overlying RST comprises dolostones containing ooids, MTS, microbial laminations and an upsection increase in quartz content. Facies shallow up to stromatolite bioherms, the top of which marks the subsequent sequence boundary.

3.2.3 Sequence 3

The TST of sequence 3 comprises molar-tooth mudstone facies, overlain by stromatolite bioherms that grade into biostromes. The reefal morphology changes upsection, beginning with broad domal stromatolites at the base overlain by tabular ones composed of pencil stromatolites (*Jurusania*). This transition in stromatolite morphology is interpreted to represent a shift from rising to dropping base level and consequently the beginning of the RST.

3.2.4 Sequence 4

Sequence 4 is interpreted to occur just below the contact with the Fort Collinson Formation, due to the similarity of its facies to the uppermost Boot Inlet Formation in the Minto Inlier (cf. Morin and Rainbird, 1992). Sequence 4 is observed at section 10 where stromatolite biostromes transition into molar-tooth lime- and dolo-mudstone facies and eventually herring-bone cross-stratified ooid grainstone facies. An upsection increase in detrital quartz content is observed throughout section 10. These facies stacking patterns comprise the RST of sequence 4, the underlying TST of which is not exposed.

3.3 Interpretation

The sequence stratigraphic framework of the Boot Inlet Formation in the Brock Inlier is characterized by three T-R sequences that are each approximately 100-200 m thick and an inferred fourth sequence where just the upper RST is exposed (Figure 17). The initial TST records an initial progradation of the platform as base level rose. Most of the overlying RST is covered but the top is exposed at section 3. Sequence 1 is overlain by two complete sequences and one incomplete sequence with relatively thin TST's followed by thick regressive systems tracts. Exposed sequence boundaries for the Boot Inlet Formation in the Brock Inlier are generally conformable suggesting that changes in base level were gradual, corroborating interpretations from studies in the Minto Inlier (cf. Morin and Rainbird, 1992; Narbonne et al., 2000).

3.3.1 Causes of cyclicity

The assembly of Rodinia is interpreted to have been completed shortly before 1070 Ma (Evans, 2002) roughly coincident with U-Pb maximum depositional age from the Nelson Head Formation (1013 ± 25 Ma). Evidence for the breakup of Rodinia is found in the Kuujua and overlying Natkusiak formations at the top of the Shaler Supergroup (Rainbird, 1993). Diabase dykes and sills feeding the Natkusiak Formation have a U-Pb age of $723 \pm 4/-2$ Ma (Heaman et al., 1992) constraining deposition of the uppermost Shaler Supergroup (Figure 1). The Boot Inlet Formation was deposited around 892 ± 13 Ma (van Acken et al., 2013), roughly in the middle of the ~300-400 million-year lifespan of Rodinia.

The cause of subsidence in the Amundsen Basin is interpreted to be related subsidence drivers of coeval rift basins in the Mackenzie and Ogilvie Mountains (Thomson et al., 2015); although, the Boot Inlet Formation shows no evidence of syndepositional faulting. Another occurrence in which two basins separated by a few hundred kilometres share the same subsidence drivers is observed in Western Australia between the Ordovician aged fill of the Canning and Amadeus Basins (Middleton, 1989). The Canning Basin, predominantly composed of shale and carbonates is interpreted to represent deposition within a shallow epeiric sea (Brown et al., 1984), and comprises all the characteristics of an intracratonic sag basin. The Canning Basin may act as an analogue for assessing the subsidence drivers in the Amundsen Basin (cf. Middleton,

1989); however, a lack of depositional ages in the Shaler Supergroup makes it impossible to accurately calibrate the orders of cyclicity.

Each of the four T-R sequences observed in the Boot Inlet Formation could represent deposition in the order of a few million to tens of millions of years; presenting a problem when interpreting the orders of cyclicity of the observed depositional sequences. These vast windows of time (millions to tens of millions of years) associated with transgressive regressive sequences could be explained by second or third order sequence stratigraphic subdivisions (Krapez, 1996). Second order “supercycles” have been attributed to multiple factors, such as; mid-ocean ridge spreading or the expansion and contraction of oceanic basins (Krapez, 1996). Third-order cycles are likely tied to a combination of causes largely revolving around tectonic and climatic mechanisms. The Amundsen Basin appears to only have undergone low levels of tectonism and therefore eustasy influenced by climatic variations is more likely. Third order cycles have been attributed to glacioeustasy in the Neoproterozoic, although globally recognized Neoproterozoic glaciations are not observed until the mid-Cryogenian (Kendall et al., 2006; Smith, 2009), making it an unlikely mechanism influencing deposition of the Boot Inlet Formation.

4 Chapter: Chemostratigraphy

4.1 Introduction

Chemostratigraphy using stable isotopes of carbon and oxygen has become a powerful tool for correlating Neoproterozoic successions when used in conjunction with other tools such as sequence stratigraphy (Knoll and Walter, 1992). In the Neoproterozoic, biostratigraphic resolution is insufficient for making meaningful correlations and the lack of zircon-bearing volcanic ash beds in older Neoproterozoic intracratonic basins used for absolute dating using U-Pb geochronology makes it difficult to constrain the ages of stratigraphic sections (Halverson, 2006). Analysis of $\delta^{13}\text{C}$ ratios in carbonate minerals is the most robust geochemical tool for correlation of stratigraphic sections because carbonate sample material is abundant, carbon is generally robust with respect to diagenetic overprinting, and large variations have been observed in datasets, which can be useful for correlations (Halverson et al., 2010; Thomson et al., 2015a).

Regional or global correlations rely on the assumption that variations in $\delta^{13}\text{C}$ reflect secular variations in the isotopic composition of seawater (Knoll and Walter, 1992). The trends of isotopic variation (excursions) represent correlatable features, whereas absolute values of stable isotope analyses of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ are not as meaningful for correlations due to intrinsic localized influences (Allan and Matthews, 1982). For instance, the globally recognized “Bitter Springs Stage” (BSS) is a dramatic excursion, first described from the Bitter Springs Formation in the Amadeus Basin

(central Australia), where $\delta^{13}\text{C}$ values shift from +5 to -1 and back to +5 ‰ (Hill et al., 2000; Swanson-Hysell et al., 2008). The BSS is also observed in the Wynniatt Formation of the Amundsen Basin, although simultaneous destabilization of methane clathrates during a sea-level fall is interpreted to have shifted the $\delta^{13}\text{C}_{\text{DIC}}$ pool to anomalously negative values, meaning this correlation depends solely on an isotopic trend, given that intrinsic fluctuations in absolute values are observed (cf. Thomson et al., 2015a).

Previous $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ stratigraphic data from northwestern Canada supported correlations between the Minto Inlier, Mackenzie Mountains and Ogilvie Mountains (Thomson et al., 2015a; Macdonald et al., 2012; Jones et al., 2010). This study offers new $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ data from the Brock Inlier from Boot Inlet and Aok formations. The Boot Inlet Formation was also sampled in more detail to examine the possible relationship between lithology/eustasy and stable isotope composition in the Amundsen Basin. New data are also presented from the Stone Knife Formation in the Mackenzie Mountains Supergroup, the sedimentology and stratigraphy of which is described in (Turner and Long, 2012). Collectively, this dataset strengthens and refines previously established chemostratigraphic correlations across northwestern Canada.

4.2 Sampling Methods

Detailed petrographic studies are necessary to be able to establish that samples are not strongly altered and thus are more likely to preserve a primary depositional signature (Metzger and Fike, 2013). Screening began with a detailed petrographic analysis on 42

thin sections of representative lithologies. Powdered samples were obtained by micro-drilling with a Foredom® Series B drilling tool, with a Pfingst 432 flex shaft, and diamond H 023 dental drilling bit. Material was collected onto 4"x4" sheets of weighing paper and weighed with a Mettler BB240 scale. At least 10mg of powdered material was sampled from each rock specimen. Samples that contained extensive veining or fracturing were omitted. Carbonate mudstone (micrite) was the preferred target and specimens with large molar-tooth structure were sampled both in the host carbonate, and in the structure, if enough of the targeted material could confidently be obtained. Material was also collected directly from ooid cores and stromatolites. An amount of ~80 micrograms samples were taken for the mass spectrometer using a Mettler Toledo XPE Micro-analytical balance and stored in a desiccator until loaded. Isotope samples were processed using a Nu Instruments® Perspective mass spectrometer at McGill University with an error of ± 0.1 ‰ for both $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$. All data manipulation was done using Microsoft Excel.

4.3 Results:

Screening

Covariation plots compare $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ from the individual stratigraphic sections of the Boot Inlet Formation (Figure 15A: section 3-7, 9 and 10) and a composite section of the Aok Formation (Figure 15B: Aok Fm.) to look for evidence of diagenetic overprinting. Section 5 (n=3) and Aok (n=18) plots show the highest R^2 values of 0.538 and 0.440, respectively (Figures 15A, B).

When isotopic values of the Boot Inlet Formation are assigned to their respective facies associations (inner- mid- and outer-ramp - see chapter 2), values generally cluster between 1 and 5 ‰ for $\delta^{13}\text{C}$ and -9 to -6 ‰ for $\delta^{18}\text{O}$ (Figure 14). Values having $\delta^{13}\text{C} > 6$ ‰ come from section 10 and $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values from outer ramp samples have the least variability, clustering tightly around 3 ‰ for $\delta^{13}\text{C}$ and -8 ‰ for $\delta^{18}\text{O}$.

Covariation plots from the Stone Knife Formation are presented in Figure 15C. Values are from two stratigraphic sections here named Stone Knife 1 and 2 with R^2 values of 0.0004 and 0.288, respectively.

Stable isotope stratigraphy

Raw isotopic data are presented from the Brock Inlier (Boot Inlet and Aok formations) and the Mackenzie Mountains Supergroup (Stone Knife Formation) in Appendix B1, B2 and B3.

Figure 14: $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values from the Boot Inlet Formation in the Brock Inlier divided by their interpreted position on the carbonate ramp. Inner-ramp (n=53), mid-ramp (n=69), outer-ramp (n=16).

Figure 15A: Covariation plots from the Brock Inlier through sections 3, 4, 5 and 6.

Figure 15B: Covariation plots from the Brock Inlier through sections 7, 9, 10 and the Aok Formation composite dataset which contains samples from sections 1, 2 and 5.

Figure 15C: Covariation plots from the Stone Knife Formation in the Mackenzie Mountains.

4.3.1 Stable isotope stratigraphy of the Brock Inlier and Mackenzie Mountains

$\delta^{13}\text{C}$ values collected from three stratigraphic sections through the Aok Formation in the Brock Inlier (n=19) have an average value of 1.84 ‰ and a range from -0.1 to 3.4 ‰. $\delta^{18}\text{O}$ values average -6.5 ‰ and range from -8.9 to -5.0 ‰. There are no significant isotopic trends in the composite section of the Aok Formation (~50 m thick; Figure 17).

131 stable isotope samples from seven stratigraphic sections represent the isotopic variation for the Boot Inlet Formation in the Brock Inlier (Figure 17). Values have an average $\delta^{13}\text{C}$ of 3.48 ‰, and a range from -0.2 to 7.6 ‰ and an overall upsection positive trend is observed for $\delta^{13}\text{C}$ in the Boot Inlet Formation (figure 17). A positive excursion from approximately 0 ‰ to 6 ‰ over the basal 100 m of section is observed, through the calcisiltite member to the top of the molar-tooth lime mudstone member. $\delta^{13}\text{C}$ values drop abruptly to 2 ‰ across the gap at the top of the molar-tooth lime mudstone member and increase gradually to 4 ‰ at the top of the dolostone member. The base of the second stromatolite member shows a sharp rise from 2 ‰ to 5 ‰ then and gradual drop to 3 ‰. The uppermost mixed molar-tooth mudstone member displays an initial positive trend from 3 ‰ to 8 ‰, followed by a negative trend back to 4 ‰, just below the inferred contact with the Fort Collinson Formation.

The average $\delta^{18}\text{O}$ value in the Boot Inlet Formation is -7.4 ‰ with a range from -9.7 to -4.1 ‰. Values are scattered in comparison to the $\delta^{13}\text{C}$ data but an overall negative trend is observed through the Boot Inlet Formation in the Brock Inlier (Figure 17). Over the basal 100 m, values gradually increase from -8 ‰ to a maximum of -4 ‰. At section

3, the transition from the first stromatolite member to the dolostone member reveals a sharp negative trend from -8 ‰ to -9 ‰ over 5 metres, with a pronounced positive trend above, gradually shifting to -6 ‰ over the proceeding 100 m throughout the dolostone member. Isotopic values in the second stromatolite member are scattered, and trends over this ~100 m interval are difficult to determine. Data from the mixed molar-tooth mudstone member are also very scattered at the base, but show a positive trend from -9 to -7 ‰ in the upper 40 m.

4.3.2 Stone Knife Formation results

138 values from two stratigraphic logs through the Stone Knife Formation in the Mackenzie Mountains comprise an ~600 metre-thick composite chemostratigraphic section (Figure 19). $\delta^{13}\text{C}$ values from the Stone Knife Formation average 4 ‰ and range from -2 to 10 ‰ with an overall positive trend. The basal 65 metres contain a pronounced positive excursion from 0 to 5 ‰. The proceeding ~150 m has a negative excursion from ~4 to ~2 ‰. Above this ~300 m of section expresses a broad positive trend from ~4 ‰ to 6 ‰ which returns to 4 ‰ in the uppermost 50 m of the section.

$\delta^{18}\text{O}$ values average -7 ‰ and range from -10 ‰ to 0 ‰. The basal 75 m has very scattered $\delta^{18}\text{O}$ values that show a broad negative trend from -8 to -10 ‰. The overlying 100 m contains a positive excursion from -10 to -6 ‰ followed by a negative excursion from -6 to -10 ‰ in the subsequent 200 m of section. The remaining section contains very scattered values showing a broad positive excursion.

4.4 Interpretation

Screening interpretation

Covariation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ is argued to be caused by diagenetic overprinting making it a useful proxy screening isotopic data (Jones et al., 2010). Covariation ($R^2=0.440$) in the Aok Formation ($n=18$) is not localized to a single section, instead it is shared by samples from section 1, 2 and 5 (Appendix 1A, 2A and 5A). This moderate covariation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ implies that diagenesis may have altered the primary marine isotopic signature. In the Boot Inlet Formation, covariation is observed in section 5; however, this is likely a result of the low sample number ($n=3$) making it unlikely that this can be attributed to diagenetic overprinting with any certainty.

Isotopic values split into their respective facies associations from the Boot Inlet Formation generally overlap, indicating that there is little to no systematic variation in $\delta^{13}\text{C}$ with changing lithologies thereby suggesting that isotopic trends were not influenced by the position on the ramp (Klaebe et al., 2017; Figure 14). One exception to this is section 10 where all $\delta^{13}\text{C}$ values are greater than 3.5 ‰. Section 10 comprises strata just below the Fort Collinson Formation (Appendix A10); these predominantly inner-ramp deposits have anomalously enriched $\delta^{13}\text{C}$ values, which most likely represent deposition in a restricted environment, resulting in a localized isotopic signal (Knauth and Kennedy, 2009). The assumption that all other isotopic variation in the Brock Inlier represents secular variation in seawater is corroborated by lack of covariation between

carbon and oxygen isotopes and the reproducibility of isotopic excursions across greater than 1200 km between the Mackenzie Mountains and Minto and Brock inliers (this study and Thomson et al., 2015a; Macdonald et al., 2012).

There is no covariation found in the first section of the Stone Knife Formation (Stone Knife 1), increasing the confidence that these samples are recording primary marine signatures for the lower portion. $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ from the upper section (Stone Knife 2) only weakly covary ($R^2=0.288$), suggesting that isotopic values still represent primary values.

Analysis of major elements, such as Fe, Mg, Mn, Ca and Sr, is often paired with chemostratigraphic data to look for influences of diagenesis on isotopic composition (Kaufman et al., 1991). Given the reproducibility of isotopic excursions across northwestern Canada (Thomson et al., 2015a; Macdonald et al., 2012; Jones et al., 2010), it is unlikely that a diagenetic mechanism would be capable of simultaneously affecting such a large area, thereby suggesting that the observed variation is likely caused by primary shifts in seawater chemistry and making an analysis of major elements unnecessary.

Pre-850 Myr old marine calcite and dolomite have $\delta^{13}\text{C}$ values that predominantly cluster between $0\text{‰} \pm 2\text{‰}$ (Knauth and Kennedy, 2009). Values from the Brock Inlier, Minto Inlier and Mackenzie Mountains are generally more enriched in ^{13}C relative to global averages from the pre-850 Myr old Proterozoic (Figure 15). High $\delta^{13}\text{C}$ values in

the Shaler Supergroup have been interpreted to be caused by high rates of organic carbon burial (Thomson et al., 2015a). Data from this study in the Brock Inlier together with data from the Minto Inlier and Mackenzie Mountains establish that this phenomenon is regional, spanning over 1200 km.

Figure 16: Pre-850 Myr old stable isotope values compared with values from the Brock and Minto Inlier and Mackenzie Mountains. Global data from Knauth and Kennedy (2009).

5 Chapter: Discussion

Reconstructions of past climatic, tectonic and biological histories through the Neoproterozoic rely heavily on the analysis of sedimentary basins, acting as pieces of a much broader puzzle. Piecing these together can shed light on whether interpreted processes were localized within these basins, regional across much larger areas, or even potentially globally recognized. The development of a lithostratigraphic, sequence stratigraphic and chemostratigraphic framework from the Boot Inlet Formation in the Brock Inlier (Figure 17) allows previously established correlations in northwestern Canada to be tested and refined. New high resolution $\delta^{13}\text{C}$ data from the Brock Inlier and Mackenzie Mountains improves the chemostratigraphic dataset for Neoproterozoic rocks exposed in northwestern Canada, and bolsters previously established regional correlations between the Mackenzie and Amundsen basins.

5.1.1 Relationship between chemo- and sequence stratigraphy

Three positive $\delta^{13}\text{C}$ excursions in the Boot Inlet Formation appear to be associated with transgressive systems tracts (Figure 17). The two uppermost regressive systems tracts appear to be associated with negative $\delta^{13}\text{C}$ excursions. This suggests that deposition in deeper waters are isotopically heavier with respect to $\delta^{13}\text{C}$ and lighter in shallower environments.

Ocean stratification during the Neoproterozoic can cause $\delta^{13}\text{C}$ gradients with variation up to 8 ‰ to 11‰ (Giddings and Wallace, 2009). While the $\delta^{13}\text{C}$ variations

between deeper and shallower water deposition are not as large as others from the Neoproterozoic, they are noticeable and suggest a correlation. The Boot Inlet Formation represents deposition on a shallow, storm-influenced carbonate ramp, where ocean circulation was high and deep-water basinal facies were absent. Allochems moved freely across the platform, given that stromatolitic bioherms are discontinuous, and biostromes did not appear to have significant topographic relief, resulting in a general lack of water circulation barriers.

Figure 13 shows that outer-ramp deposits are not isotopically heavier than mid- or inner-ramp ones, and the transgressive systems tracts on which these positive excursions are observed occur in subtle transitions from the inner- to mid-ramp. Carbonate muds on the ramp are interpreted to be mechanically derived from shallower carbonate depositional environments and the variation in their isotopic signature may only differ in their interaction with pore waters during dissolution and recrystallization (Land, 1986). Studies from the Cenozoic have shown that coastal environments tend to be isotopically lighter with respect to the stable isotopes of carbon and oxygen, a relationship caused by a flourishing biomass that incorporates depleted carbon into meteoric fluids that will inherently contain isotopically light oxygen (Knauth and Kennedy, 2009). This may explain the correlative pattern between T-R sequences and $\delta^{13}\text{C}$ excursions from the Boot Inlet Formation as isotopes trend positively during deeper-water sedimentation offshore and negatively during shallower-water sedimentation. This relationship may be the result of shifts in pore water chemistry, as inner-ramp sediments would interact more with meteoric fluids whereas the marine/mixed meteoric fluid ratio in pore waters interacting

with mid-ramp sediments would be much higher. Regardless of the mechanisms responsible for the relationship, isotopic excursions observed in the Boot Inlet Formation may be somewhat dependent on changes in sea-level that are represented by T-R sequences. However, these trends should still represent non-restricted phenomena that are applicable for the purposes of correlating the Brock Inlier regionally, but it might be misguided to attribute these excursions to secular variations in seawater chemistry if they are not observed on a more regional scale.

Figure 17: Brock Inlier dataset. First column contains lithostratigraphic members. Transgressive regressive sequences in blue. For stable isotope data: O18 data in orange, C13 data in black. Boot Inlet Formation is disconformably overlain by Mount Clark Formation (Cambrian).

5.1.2 Regional Correlation

5.1.2.1 Correlation of the Boot Inlet Formation between the Brock and Minto Inlier

New data from the Brock Inlier allow a refined, high resolution correlation between the Aok, Grassy Bay and Boot Inlet formations in the Minto and Brock inliers to be proposed (summarized in Figure 18). Isotopic values from the Minto Inlier are generally more enriched in C¹³ than in the Brock Inlier, except for the uppermost Boot Inlet Formation, where a localized mechanism just below the contact with the Fort Collinson Formation is interpreted to have amplified the isotopic signal. T-R sequences 2 and 3 (Morin and Rainbird, 1993) from the Minto Inlier align well with the two lowermost T-R sequences from the Boot Inlet Formation in the Brock Inlier as does the first half of the third T-R sequence, above which there is a gap in the stratigraphic section.

In the Brock Inlier, an isotopic excursion from 0 ‰ to 5 ‰ is observed in the Boot Inlet Formation, which is potentially correlative with an excursion at 200 m in the composite Minto Inlier dataset (Figure 18). In the Minto Inlier, the contact of the Boot Inlet Formation with the underlying Grassy Bay Formation is placed stratigraphically higher (Morin and Rainbird, 1993), above the transgressive lag representing the interpreted contact in the Brock Inlier. This could explain why the excursion is stratigraphically higher in the Minto Inlier, assuming that they are each represent results of the same primary variation in seawater chemistry.

Lithostratigraphic members for the Boot Inlet Formation are not well correlated between the two inliers. For example, outer-ramp “concentric sheet bioherms”, described by Narbonne et al., (2000) from the northern Minto Inlier were not observed in the Brock Inlier. Additionally, the basal stromatolite member from the Minto Inlier is not present in the Brock Inlier, and the thick second stromatolite member in the Brock Inlier does not have an equivalently thick unit in the Minto Inlier (Figure 18). These two units appear equivalent based on sedimentological similarities, but chemostratigraphic and sequence stratigraphic attributes, from which the proposed correlation is based, offer a more robust correlation. Additionally, if the two thick stromatolite units were correlative, it would refute the correlation with the Stone Knife Formation in the Mackenzie Basin, proposed below. Therefore, these two thick reef complexes are interpreted to be diachronous.

In summary, the Brock Inlier is interpreted to contain a higher proportion of calm-water, mid-ramp deposits than the Minto Inlier. During deposition of the Boot Inlet Formation, the Amundsen Basin was rimmed by a carbonate ramp dominated by shallow-water facies, suggesting depositional settings that rarely deepened below storm wave base. Four T-R sequences are recognized in the stratigraphic section and composite sections comprising the Boot Inlet Formation have similar thicknesses in both inliers, loosely implying similar subsidence rates. The Grassy Bay and Fort Collinson formations are also better developed in the Minto Inlier. The relatively poor expression of siliciclastic sedimentary rocks in the Reynolds Point Group indicates that terrigenous influx was reduced in the region now occupied by the Brock Inlier. Complex mid and

inner-ramp stromatolite complexes were abundant and form thick, laterally continuous stratigraphic units. A diverse suite of stromatolites grew in the Amundsen Basin, supplying detrital carbonate sediments to the ramp, especially during storm activity. They certainly formed immense tabular regional biogenic structures that extend for tens of kilometres in this part of the basin.

Figure 18: Proposed correlation of Boot Inlet, Grassy Bay and Aok formations between the composite sections of the Brock and Minto inliers utilizing stable carbon isotope chemostratigraphy, transgressive-regressive sequence stratigraphy and lithostratigraphy. Minto Inlier stable isotope data comes from Thomson et al., 2015a (triangles) and Jones et al., 2010 (circles). Transgressive-regressive sequences from Morin and Rainbird (1993). Stratigraphic section after Narbonne et al., (2000).

5.1.2.2 Correlation between the Boot Inlet and Stone Knife Formation

The proposed correlation between the Stone Knife Formation in the Mackenzie Mountains and the Boot Inlet Formation in the Brock Inlier is summarized in Figure 19. Isotopic data show sufficient variation to establish chemostratigraphic correlations, and a high degree of similarity is observed between the $\delta^{13}\text{C}$ datasets both in the isotopic excursions and in the magnitude of $\delta^{13}\text{C}$ values forming the foundation for the proposed correlation. The four T-R sequences interpreted in the Brock Inlier match well with the four Little Dal sequences identified by Turner et al. (1997); corroborating the proposed correlations based on chemostratigraphy.

The basal 75 m of the Stone Knife Formation and the basal 100 m of the Boot Inlet Formation share a distinct positive $\delta^{13}\text{C}$ excursion from 0 to 5 ‰ that coincides with a correlatable transgressive-regressive sequence. This interval is observed directly above the contact with the underlying Dodo Creek and Grassy Bay formations, respectively, and both intervals are capped with potentially correlative stromatolitic facies. Collectively, this represents a directly correlatable interval between the Amundsen and Mackenzie basins that is supported by sequence stratigraphic, chemostratigraphic and lithostratigraphic evidence. This interval matches well with the basal part (0-130 m in Figure 19) of the $\delta^{13}\text{C}$ dataset from the Stone Knife Formation from Halverson (2006) in which a positive $\delta^{13}\text{C}$ excursion, from 0 ‰ to 5 ‰, also occurs.

A negative excursion, from +4 ‰ to -3 ‰, occurs in an interval from 100-200 m in the Stone Knife Formation (Figure 19) and is interpreted to represent deposition below the redoxcline in a bathymetric low generated by syndepositional faulting (Turner and Long 2008). This is interpreted to be a localized phenomenon that would not correlate with the more tectonically stable shallow-water environments in the Amundsen Basin. The regional correlation in the isotopic dataset is re-established above ~200 m in the Mackenzie Mountains stratigraphic section.

The lithostratigraphic correlation between the Stone Knife Formation and the Boot Inlet Formation is strengthened by the data from this study (Figure 3, 19). The Dodo Creek Formation (underlying the Stone Knife/Silverberry formations) has been previously correlated with the lower Boot Inlet Formation (Rainbird et al., 1996; Long et al., 2008) but appears to be more directly correlatable with the Grassy Bay Formation based on lithological and stratigraphic affinities (cf. Turner and Long, 2012; Morin and Rainbird, 1993). The underlying Abraham Plains Formation does not have a clear lithostratigraphic equivalent in the Brock Inlier, because of the lack of terrigenous influx described above, but shows the highest affinity to the lower Grassy Bay Formation from the Minto Inlier (cf. Morin and Rainbird, 1993; Young and Long, 1977). The underlying Aok Formation correlates with the McClure Formation, below the Abraham Plains Formation which underlies the Dodo Creek Formation (Long and Turner, 2014; Aitken et al., 1978 a, b; Jefferson and Young, 1989; Long et al., 2008).

Figure 19: Proposed correlation between the Boot Inlet Formation in the Brock Inlier and the Stone Knife Formation in the Mackenzie Mountains utilizing stable carbon isotope chemostratigraphy, transgressive-regressive sequence stratigraphy and lithostratigraphy. Stratigraphic section from Stone Knife Formation from Turner and Long (2012).

Correlation Summary

This contribution adds new chemostratigraphic, sequence stratigraphic and lithostratigraphic data from the Brock Inlier as well as chemostratigraphic data from the Mackenzie Mountains. These data strengthen the regional correlation proposed by Thomson et al. (2015), which relies more heavily on lithostratigraphy and sequence stratigraphy than those which calibrate the stratigraphic section by employing chemostratigraphy (cf. Jones et al., 2010). The basal interval of the Boot Inlet and Stone Knife formations, marked by an excursion from 0 ‰ to 5 ‰, was originally reported by Halverson (2006) and is now established in chemostratigraphic datasets across northwestern Canada. This potentially also extends to the Gibben Formation from the Fifteenmile Group in the Ogilvie Mountains (cf. Macdonald et al., 2012). A proposed correlation highlighting this tightly correlatable interval across northwestern Canada is summarized in Figure 20.

Figure 20: Regional correlation between Mackenzie Mountains and Shaler supergroups using C13 chemostratigraphy and transgressive-regressive sequence stratigraphy. Note the change in scale for $\delta^{13}\text{C}$ between Amundsen and Mackenzie Mountains Basin. Blue polygon emphasizes isotopic excursion from 0‰ to 5‰.

Potential transitional phase in Neoproterozoic carbon cycling

It has been speculated that the rise in $\delta^{13}\text{C}$ from 0‰ to 5‰ in the basal interval of Stone Knife formations, now established in the Boot Inlet Formation, records an important transition from the late Mesoproterozoic (avg $\delta^{13}\text{C} \sim 3\text{‰}$) to more enriched and more variable $\delta^{13}\text{C}$ values in the Neoproterozoic (Halverson, 2006; Bartley et al., 2001). Data from the underlying Aok Formation in the Amundsen Basin average +2‰, possibly recording an initial decrease from late Mesoproterozoic averages, prior to higher more variable values above. This is challenged by the observation of moderate covariation between carbon and oxygen isotopes in the Aok Formation possibly suggesting diagenetic overprinting, so these data should be interpreted cautiously.

Stable isotope analyses have shown that there appears to have been a gradual decrease in the dissolved inorganic carbon (DIC) reservoir throughout the Proterozoic (Bartley and Kah, 2004), and this has implications for the increased variability of $\delta^{13}\text{C}$ during the Neoproterozoic. A progressively smaller DIC reservoir will be less effective at attenuating isotopic variation caused by biogeochemical fluxes. Consequently, the lower marine carbonate saturation with respect to DIC during the Neoproterozoic led to higher sensitivity to small-scale perturbations such as shifts in C_{org} productivity. The early Neoproterozoic is also marked as a node for the evolutionary expansion of eukaryotes, which would potentially facilitate non-steady state biogeochemical perturbations in $\delta^{13}\text{C}$ (Knoll, 1994).

The $\delta^{13}\text{C}$ excursion from 0‰ to 5‰ that is observed in Neoproterozoic sequences across northwestern Canada (Figure 20) may represent a transitional period in the early Neoproterozoic in which high isotopic variation and prolonged positive values becomes the norm following more stable values of the late Mesoproterozoic. The ability to record these short-term shifts may be the result of the gradual shift to a smaller DIC reservoir through the Proterozoic. The correlation between T-R sequences and isotopic excursions observed in the Brock and Minto inliers and Mackenzie Mountains may represent a relationship that influences isotopic variation which could be misinterpreted as primary variations in seawater chemistry. Chemostratigraphic data is not reported elsewhere across this interval outside of northwestern Canada, as stratigraphic sections do not cover this young Neoproterozoic interval (Halverson 2005; Swanson-Hysell et al., 2010; Swanson-Hysell, 2012).

6 Chapter: Conclusions

The Boot Inlet Formation, deposited in the Amundsen Basin ~900 Mya, may record an important transitional period in the evolution of Earth's carbon cycle. A ~400 metre-thick composite stratigraphic section represents mainly carbonate deposition along a storm-influenced ramp on which spectacular stromatolite reefs thrived, growing to uninterrupted thicknesses exceeding 85 m. These reefs seemingly sourced and distributed carbonate detritus to the ramp as there was no observed obstruction to basin circulation. New chemostratigraphic, sequence stratigraphic and lithostratigraphic data from the Brock Inlier strengthen correlations between early Neoproterozoic successions across northwestern Canada, confirming a link between the Amundsen and Mackenzie basins and supporting the great lateral extent (>1200 km) of directly correlatable intervals.

C¹³ variations in the Boot Inlet Formation do not show a relationship with interpreted lithofacies, but may be partly dependent on sea-level changes represented by four transgressive-regressive sequences, possibly indicating another mechanism responsible for isotopic variation along with secular variations in seawater chemistry. Regardless of the causal mechanisms of Neoproterozoic isotopic variation, a gradual decrease in the size of the DIC reservoir throughout the Proterozoic may have resulted in basin sedimentation that was more sensitive to changes in seawater chemistry, thus producing better-defined and higher frequency C¹³ excursions that are useful for stratigraphic correlation.

A depositional age of 892 ± 13 Ma from shale samples (Re-Os) from the Boot Inlet Formation approximately 100 m below the contact with the Fort Collinson Formation in the Minto Inlier (van Acken et al., 2013), constrains the timing of these isotopic trends. A depositional age lower in the Shaler Supergroup would better establish rates of sedimentation and the order of cyclicity of the proposed T-R sequences within the Reynolds Point Group.

Given that the Aok and Boot Inlet formations potentially record a transitional period in carbon cycling it would be beneficial to develop a chemostratigraphic curve for the underlying Mikkelsen Islands Formation. Investigating this older carbonate succession would help identify where this early Neoproterozoic transition in carbon cycling occurs stratigraphically.

Appendices

Appendix A

Legend

Lithologies

stomatolite

mudstone

rudstone/grainstone

mud/shale

diabase sill

sandstone

Sedimentary structures

biostromal stromatolite

biohermal stromatolite

intraclasts

molar-tooth structure

microbial laminations

cross bedding

quartz sand

ooids

chert replacement

pinch and swell structure

hummocky cross-stratification

microfossil sample site

herring-bone cross-stratification

Carbonate classification

M	Mudstone
W	Wackestone
P	Packstone
G	Grainstone
R	Rudstone
F	Floatstone
B	Boundstone

Siliciclastic classification

M	Mud	C Clay
		S Silt
		F Fine
S	Sand	M Medium
		C Coarse
G	Gravel

A.1 Stratigraphic Section 1

A.2 Stratigraphic Section 2

A.3 Stratigraphic Section 3

A.5 Stratigraphic Section 5

A.7 Stratigraphic Section 7

A.8 Stratigraphic Section 8

A.9 Stratigraphic Section 9

A.10 Stratigraphic Section 10

A.11: GPS coordinates of stratigraphic sections

Section	GSC* ID number	Latitude	Longitude	Formations measured
1	15RAT02	69°0'18.80"N	122°39'47.05"W	Aok Nelson Head
2	15RAT028	69°0'22.12"N	122°40'14.13"W	Aok
3	15RAT026	69°00'53.88"N	122°40'14.69"W	Boot Inlet
4	15RAT027	69°01'56.84"N	122°41'15.94"W	Boot Inlet
5	15RAT012	68°54'11.00"N	121°44'20.20"W	Boot Inlet Grassy Bay Aok
6	15RAT013	68°56'40"N	121°3'34"W	Boot Inlet
7	15RAT017	68°21'50.5"8N	121°20'58.57"W	Boot Inlet
8	15RAT019	69°05'52.99"N	122°44'54.48"W	Boot Inlet
9	15RAT021	68°55'41.92"N	121°44'14.20"W	Boot Inlet Grassy Bay
10	15RAT024	68°24'08.72"N	120°58'07.07"W	Boot Inlet

*Geological Survey of Canada

A.12: Correlation of measured stratigraphic sections from the Brock Inlier.

Appendix B

Stable Isotope Data

B.1 Stable isotope data from the Aok Formation in the Brock Inlier

Section number	Stratigraphic height (m)	$\delta^{13}\text{C}$ (‰, V-PDB)	$\delta^{18}\text{O}$ (‰, V-PDB)	Carbonate mineralogy
S1	9	-0.1	-7.8	dl
S1	10	1.1	-7.8	dl
S1	11	2.0	-7.3	dl
S1	13	2.3	-5.8	dl
S1	14	2.0	-6.1	dl
S1	17	2.6	-5.0	dl
S1	18	3.0	-5.3	dl
S1	20	3.4	-6.4	dl
S1	21	2.4	-5.0	dl
S1	22	2.8	-6.7	dl
S1	28	3.0	-6.1	dl
S2	19	0.5	-6.4	dl
S2	20	0.5	-6.4	dl
S2	21	1.5	-7.9	dl
S2	22	1.4	-6.7	dl
S2	23	1.0	-7.5	dl
S2	24	1.2	-6.4	dl
S5	37	0.6	-8.6	dl
S5	40	0.4	-8.9	dl

B.2 Stable isotope data from the Boot Inlet Formation in the Brock Inlier

Section number	Stratigraphic height (m)	$\delta^{13}\text{C}$ (‰, V-PDB)	$\delta^{18}\text{O}$ (‰, V-PDB)	Carbonate mineralogy
S9	76	0.0	-7.9	dl
S9	88	1.5	-7.6	dl
S9	89	1.8	-6.3	dl
S9	96	2.0	-5.5	dl
S9	98	1.4	-5.9	dl
S9	110	2.1	-8.0	ls
S9	111	2.3	-6.5	ls
S5	113	3.0	-6.2	ls
S5	116	3.5	-7.6	ls
S5	117	3.3	-6.1	ls
S6	141	2.2	-6.8	ls
S6	144	2.7	-4.4	ls
S6	145	3.4	-5.0	ls
S6	146	3.4	-4.6	ls
S6	147	1.8	-5.9	ls
S6	148	3.1	-4.4	ls
S6	150	3.4	-5.2	ls
S6	151	4.2	-4.3	ls
S6	152	3.1	-6.7	ls
S6	155	5.0	-4.4	ls
S6	156	3.6	-5.7	ls
S6	158	4.2	-8.1	ls
S6	161	4.9	-5.5	ls
S6	162	5.0	-5.2	ls
S6	163	6.0	-5.0	ls
S6	169	6.1	-6.4	ls
S6	169	5.1	-5.1	ls
S6	170	4.8	-4.5	ls
S6	172	5.3	-7.0	ls
S6	174	5.3	-7.6	ls
S3	186	2.2	-7.6	dl
S3	189	2.2	-7.3	dl
S3	190	2.3	-9.3	dl
S3	191	2.0	-9.4	dl

S3	192	2.4	-9.2	dl
S3	193	3.2	-9.4	dl
S3	193	2.5	-9.7	dl
S3	194	3.2	-9.0	dl
S3	195	3.0	-8.6	dl
S3	196	2.2	-8.7	dl
S3	197	2.9	-7.8	dl
S3	198	2.5	-6.7	dl
S3	199	2.0	-7.6	dl
S3	200	3.2	-7.1	dl
S3	200	2.3	-8.9	dl
S3	201	3.6	-7.2	dl
S3	202	2.3	-7.1	dl
S3	203	2.8	-7.8	dl
S3	204	2.9	-8.0	dl
S3	206	0.9	-7.4	dl
S3	207	-0.2	-8.3	dl
S3	208	2.8	-8.3	dl
S3	209	2.3	-8.1	dl
S3	210	2.8	-6.9	dl
S3	211	2.7	-6.9	dl
S3	214	2.4	-7.4	dl
S3	215	2.8	-8.1	dl
S3	216	2.4	-7.6	dl
S3	217	2.5	-7.3	dl
S3	218	3.7	-7.7	dl
S3	221	3.2	-6.8	dl
S4	236	2.9	-7.7	dl
S4	237	3.3	-7.5	dl
S4	238	2.7	-7.2	dl
S4	239	2.7	-7.3	dl
S4	240	2.9	-7.2	dl
S4	241	2.9	-7.5	dl
S4	242	2.9	-7.3	dl
S4	244	2.8	-7.5	dl
S4	245	3.3	-7.7	dl
S4	246	2.7	-7.5	dl

S4	247	3.3	-7.4	dl
S4	248	2.7	-7.4	dl
S4	249	2.9	-7.4	dl
S4	250	2.9	-7.7	dl
S4	252	3.4	-7.8	dl
S4	253	2.2	-7.4	dl
S4	254	2.5	-7.1	dl
S4	255	2.7	-7.1	dl
S4	256	2.1	-7.5	dl
S4	260	2.1	-7.5	dl
S4	262	1.9	-6.7	dl
S4	274	2.7	-6.6	dl
S4	275	3.0	-6.8	dl
S4	276	2.9	-6.9	dl
S4	277	4.1	-6.5	dl
S4	278	2.7	-9.5	dl
S4	279	3.4	-6.4	dl
S4	280	3.8	-6.9	dl
S7	306	2.3	-8.1	dl
S7	307	2.9	-8.5	dl
S7	308	2.8	-8.1	dl
S7	309	3.9	-7.9	dl
S7	311	2.7	-8.8	dl
S7	315	5.0	-8.6	dl
S7	316	5.3	-7.5	dl
S7	321	4.4	-8.8	dl
S7	322	4.2	-9.4	dl
S7	325	4.3	-7.6	dl
S7	326	3.9	-8.0	dl
S7	327	4.1	-8.2	dl
S7	328	4.1	-8.2	dl
S7	331	4.1	-5.8	dl
S7	333	4.1	-7.6	dl
S7	334	4.1	-6.4	dl
S7	335	4.0	-7.6	dl
S7	337	4.1	-9.2	dl
S7	340	3.9	-8.9	dl

S7	351	3.5	-8.6	dl
S7	352	3.4	-8.9	dl
S7	353	3.4	-9.0	dl
S7	361	3.1	-9.3	dl
S7	367	3.4	-8.7	dl
S7	372	3.2	-8.3	dl
S10	401	3.5	-6.9	dl
S10	402	4.5	-9.5	dl
S10	421	7.1	-4.1	ls
S10	423	7.2	-7.2	ls
S10	424	7.5	-8.6	dl
S10	426	7.6	-9.3	dl
S10	431	7.1	-5.9	dl
S10	436	7.2	-8.8	dl
S10	438	6.7	-8.9	dl
S10	449	6.8	-8.9	dl
S10	453	5.8	-9.0	dl
S10	455	5.2	-8.5	dl
S10	456	4.8	-8.2	dl
S10	462	4.7	-8.3	dl
S10	465	4.5	-8.1	dl
S10	467	4.8	-6.5	dl
S10	468	4.4	-7.1	dl

B.3 Stable isotope data from the Stone Knife Formation in the Mackenzie Mountains

Section number*	Stratigraphic height (m)	$\delta^{13}\text{C}$ (‰, V-PDB)	$\delta^{18}\text{O}$ (‰, V-PDB)	Carbonate mineralogy
SK	0.1	1.4	-8.7	dl
SK	0.2	1.6	-9.4	dl
SK	2	1.6	-9.0	dl
SK	3	1.7	-9.3	dl
SK	4	1.5	-9.8	dl
SK	5	-0.3	-6.1	dl
SK	6	0.3	-5.6	dl
SK	7	1.4	-8.2	ls
SK	8	0.2	-6.9	ls
SK	8.01	1.4	0.0	dl
SK	8.6	0.7	-10.9	dl
SK	13.1	1.4	-6.4	ls
SK	15.3	1.6	-7.4	ls
SK	15.3	1.0	-9.9	ls
SK	16.9	1.0	-9.7	dl
SK	19.7	1.7	-10.1	dl
SK	20	1.8	-10.2	ls
SK	21.8	2.1	-8.2	dl
SK	21.8	2.0	-8.0	dl
SK	22.5	1.9	-7.9	dl
SK	24.6	2.0	-7.1	dl
SK	24.6	2.0	-7.9	dl
SK	26	2.0	-6.7	dl
SK	26	2.1	-8.3	dl
SK	27.6	3.1	-8.6	dl
SK	29.7	1.4	-9.5	ls
SK	37.7	4.3	-8.4	ls
SK	37.7	4.4	-8.8	ls
SK	39	3.6	-8.7	ls
SK	40.2	4.4	-8.6	ls
SK	40.2	4.7	-6.8	ls
SK	41	4.4	-8.4	ls
SK	42.3	4.3	-7.6	ls
SK	43	4.8	-8.8	ls

SK	43.5	5.0	-8.7	ls
SK	45.3	4.4	-8.6	ls
SK	45.3a	4.5	-8.7	ls
SK	46	4.0	-7.6	ls
SK	47.3	3.7	-8.4	ls
SK	47.9	3.5	-8.6	ls
SK	50	3.5	-8.7	ls
SK	53	3.5	-8.4	ls
SK	54	3.5	-8.8	ls
SK	54.9	3.7	-8.6	ls
SK	56	3.4	-8.9	ls
SK	57.5	3.8	-8.9	ls
SK	59	3.7	-9.1	ls
SK	62.8	1.9	-9.8	ls
SK	63.5	2.4	-9.7	ls
SK	64.2	2.1	-10.0	ls
SK	65	3.2	-10.4	ls
SK	72.5	-0.5	-10.4	ls
SK	74.5	-0.6	-9.9	ls
SK	75	-1.5	-10.5	ls
JS	100.1	-1.9	-8.7	dl
JS	124	2.2	-10.9	ls
JS	131.5	3.2	-10.3	ls
JS	133.5	3.4	-7.4	ls
JS	135.5	4.2	-7.5	ls
JS	141.5	4.5	-7.0	ls
JS	150.5	4.3	-6.0	ls
JS	155.5	4.3	-5.7	ls
JS	158.5	4.3	-5.9	ls
JS	177.4	3.7	-6.2	ls
JS	180.7	4.0	-6.7	ls
JS	183.7	4.1	-6.7	ls
JS	192.7	4.3	-6.4	ls
JS	195.7	4.6	-6.4	ls
JS	197.7	4.1	-6.3	ls
JS	205.7	4.1	-6.5	ls
JS	205.7	4.4	-6.4	ls

JS	196.1	4.0	-7.0	ls
JS	199.4	4.3	-7.2	ls
JS	205.1	5.1	-2.4	dl
JS	212.6	3.7	-5.6	dl
JS	217.1	4.0	-7.1	ls
JS	225	4.4	-6.9	ls
JS	229.2	4.3	-6.7	ls
JS	235.2	3.4	-6.9	ls
JS	239.7	3.1	-6.5	ls
JS	242.7	3.3	-6.3	ls
JS	247.7	3.4	-5.5	ls
JS	259.2	1.9	-7.6	ls
JS	265.1	1.9	-7.4	ls
JS	271.1	2.5	-7.5	ls
JS	274.9	3.2	-7.5	ls
JS	283.7	-1.8	-8.7	ls
JS	286.9	4.5	-8.5	ls
JS	290.4	4.9	-8.5	ls
JS	297.9	4.9	-8.4	ls
JS	303.9	4.9	-8.5	ls
JS	322.6	3.6	-8.6	ls
JS	327.1	5.4	-9.1	ls
JS	332.1	5.5	-9.0	ls
JS	337.1	4.5	-9.1	ls
JS	343.1	3.6	-9.2	ls
JS	350.1	4.0	-9.5	ls
JS	376.1	2.5	-9.1	ls
JS	382.9	4.7	-9.1	ls
JS	386.9	4.9	-8.1	ls
JS	386.9	4.5	-7.3	ls
JS	400.9	5.3	-8.4	ls
JS	418.9	4.9	-8.5	ls
JS	426.2	5.1	-7.8	ls
JS	428.9	5.4	-7.2	ls
JS	428.9	5.2	-7.2	ls
JS	433.4	5.5	-7.5	ls
JS	433.4	9.9	-5.7	ls

JS	433.4	5.4	-7.1	ls
JS	433.4	7.2	-5.0	ls
JS	441.9	6.6	-2.3	ls
JS	441.9	5.0	-6.4	ls
JS	444.8	5.7	-3.8	dl
JS	449.3	6.1	-4.3	dl
JS	452.3	6.4	-4.1	dl
JS	452.3	6.4	-3.7	dl
JS	456.8	6.1	-4.0	dl
JS	459.8	6.0	-3.6	dl
JS	462.8	7.6	-2.7	dl
JS	466.3	6.3	-5.3	dl
JS	472.5	6.9	-8.0	dl
JS	480	6.8	-3.6	dl
JS	485	6.1	-4.0	dl
JS	485	5.8	-5.0	ls
JS	492	6.4	-5.1	dl
JS	492	5.7	-7.7	dl
JS	496.7	6.2	-3.9	dl
JS	496.7	3.2	-11.3	dl
JS	502	6.3	-5.1	dl
JS	502	6.1	-7.0	dl
JS	508.5	6.8	-3.4	dl
JS	508.5	6.4	-4.5	dl
JS	511.6	5.8	-3.9	dl
JS	518.7	5.0	-6.2	ls
JS	527.7	4.7	-6.8	ls
JS	527.7	5.2	-7.8	ls
JS	556.9	4.3	-6.7	dl
JS	558.7	4.4	-8.7	dl

*SK and JS represent Stone Knife 1 and 2 respectively. Reference sections from Turner and Long (2012).

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