**NEOPROTEROZOIC CARBONATE RAMP DEVELOPMENT IN THE
CANADIAN ARCTIC (SHALER SUPERGROUP): SEDIMENTOLOGY,
SEQUENCE STRATIGRAPHY, AND CHEMOSTRATIGRAPHY**

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FRONTISPIECE

~ Evening river crossing, NW Minto Inlier

ABSTRACT

This study describes Tonian-Cryogenian pre-rift strata of the upper Shaler Supergroup, deposited in the Amundsen Basin (Victoria Island, Canada). The Wynniatt Formation, part of the upper Shaler Supergroup, consists of three carbonate ramp successions separated by regional unconformities. Each cycle represents part of a distally steepened, storm-dominated carbonate ramp. Temporal changes in carbonate facies record depositional environments spanning from supratidal to outer ramp settings. Interpretation of an intertidal mudflat depositional environment for the *Tawuia-Chuaria* assemblage zone provides depositional context for interbasinal biostratigraphic correlation. Correlation between the Amundsen and Fifteenmile (Yukon) basins is complicated by differing rates and regimes of subsidence, with the exception of a regional basin-deepening event. Contrary to previous correlations, it is proposed that the upper Shaler Supergroup and Little Dal Group of the Mackenzie Mountains Supergroup (Mackenzie Basin) are equivalent to the entire Fifteenmile Group.

Five transgressive-regressive (T-R) cycles were identified in the Jago Bay, Minto Inlet, and Wynniatt formations. Three tectonostratigraphic units for the upper Shaler Supergroup record an initial sag basin, followed by early extension and thermal doming, and finally rifting of the Amundsen Basin. Neoproterozoic successions of northwest Laurentia are an example where rift and intracontinental basins are spatially and temporally related. Subsidence possibly was related to multiple cycles of intra-plate extension that complemented coeval fault-controlled subsidence.

An improved $\delta^{13}\text{C}_{\text{carb}}$ curve paired with $\delta^{13}\text{C}_{\text{org}}$ values provides support for correlation of a significant negative $\delta^{13}\text{C}$ excursion that has been linked to the Bitter Springs stage. In the Amundsen Basin, $\delta^{13}\text{C}_{\text{carb}}$ values drop to markedly negative values ($> -14\text{‰}$), which is attributed to local overprints whereby isotopically light carbon in pore waters was precipitated as authigenic carbonate cement fuelled by oxidation of methane or organic matter during sulphate and iron reduction. Redox proxies indicate local basin euxinia and anoxia. Pronounced Mo, V, and U enrichments in euxinic black shales suggest that the Bitter Springs stage was a transitional time, where the global oceans became more oxygenated relative to preceding portions of the mid-Proterozoic, but remained more reducing compared to the Phanerozoic.

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This whole adventure started over beers with Rob at the Black Knight in Yellowknife, an auspicious start to a project in Canada's north, and not to mention a good way to get to know your future supervisor! I would like to thank Rob for his guidance through the past four years; he has been a constant source of wisdom, and is always present to offer his help. One could not ask for a better mentor.

Thank you to George Dix for never seeming to mind when I would pop into his office with a question, and for incredibly fast reviews of drafts. Thank you to Beth and Sheila for all your help over the years. And of course thank you to all the other students at Carleton University that I have met over the years for their help, laughter, and enthusiasm.

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To Andrey and Noah, who may not agree on everything but definitely opened my eyes to the complexities of geochemical interpretations. Thank you to Wendy Abdi and everyone in the G. G. Hatch Lab at Ottawa U for their help with carbon isotope analysis.

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Finally, thank you to my family for all their support throughout the past few years. Lorraine and Mario thank you so much for the delicious home-cooked meals and being there when I needed your help. Mom and Dad – I could not have made the home stretch without you. Much love.

PREFACE

Method of Presentation

This thesis is presented as five chapters: an introduction, conclusion, and three stand-alone manuscripts addressing specific geological problems. While there may be some overlap given the focused nature of this project, redundancies are generally restricted to the introductory portions of the chapters. The order of the manuscripts follows the order of publication given that later manuscripts build upon the framework provided by earlier manuscripts. At the time of this submission, chapter two is published in *Sedimentary Geology* (Thomson, D., Rainbird, R.H., Dix, G., 2014. *Architecture of a Neoproterozoic intracratonic carbonate ramp succession: Wynniatt Formation, Amundsen Basin, Arctic Canada. Sedimentary Geology*, 299: 119-138.), chapter three is re-submitted to *GSA Bulletin* (Thomson, D. Rainbird, R. H., Krapez, B., submitted. *T-R cycles and tectonostratigraphy of the Neoproterozoic (Tonian-Cryogenian) Amundsen Basin prior to supercontinent (Rodinia) breakup*), and chapter four is submitted to *Precambrian Research* (Thomson, D., Rainbird, R. H., Planavsky, N., Lyons, T., Bekker, A. *Chemostratigraphy of the Shaler Supergroup, Victoria Island, NW Canada: a record of ocean composition prior to the Cryogenian glaciations*). I am the primary author on all three contributions.

Original Contribution

The aims of this research were to formally describe the members of the Wynniatt Formation, including the nature of their bounding surfaces, lithostratigraphy, lithofacies, and depositional environments. Sequence (T-R) stratigraphic, chemostratigraphic, and geochemical studies were also undertaken for the upper Shaler Supergroup (including the Wynniatt Formation) to better understand cyclicity within the basin, mechanisms for basin subsidence, improve stratigraphic correlations, and understand seawater chemistry (C, O, S, redox proxies). This required integrating previously published data with newly acquired stratigraphic, sedimentological, and geochemical data. The choice of study areas and research objectives were my own. Fieldwork was carried out over two summers (2010 and 2011), and measurement of stratigraphic sections and sampling was done by myself, with the aid of a field assistant.

Samples collected during the summer were selected by myself for petrography and geochemical analyses. Cut samples were sent to Vancouver Petrographics and made into thin sections. I performed the petrography at Carleton University. Rock powders for C and O isotope analyses were drilled by myself, or drilling locations were selected by myself and drilled by an assistant at Carleton University using a dental drill. Samples for C and O isotope analysis were analyzed at the G. G. Hatch Laboratory of the University

of Ottawa, where I weighed and prepared the samples under the supervision of Wendy Abdi. I performed SEM analyses at Carleton University under the supervision of George Dix. I prepared and ran XRF analyses for major and trace elements at the University of Ottawa under the supervision of Ron Hartree. Samples were sent to the University of Manitoba for organic C and S isotope analysis, performed by Andrey Bekker and Misuk Yun. A subset of these samples was sent to Noah Planavsky for Fe speciation analysis, performed in Tim Lyon's lab at the University of California, Riverside. Samples were sent to ACME Laboratories, Vancouver, for major and trace element analysis.

I drafted all figures and prepared all tables in this thesis. This thesis presents a wealth of new stratigraphic, sedimentological, and geochemical data. The models presented in this thesis represent my interpretation of these data. I prepared initial drafts of all papers, with editorial modifications suggested by co-authors. Specifically, B. Krapez provided technical expertise on intracontinental basins and sequence stratigraphy; N. Planavsky and A. Bekker provided geochemical expertise on stable isotopes and redox proxies (methods, analysis, and interpretation). Finally, I am wholly responsible for any omissions or inaccuracies in this document.

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CHAPTER 1: INTRODUCTION

1.1 General introduction

The Neoproterozoic Era is a critical period for understanding Earth history, and currently garners much attention from researchers. The changes that occurred during the Neoproterozoic set the stage for the explosion of metazoan life and the evolution of earth systems to Phanerozoic conditions. During the Neoproterozoic the supercontinent Rodinia was formed and broke apart (e.g., Hoffman, 1991), there were several extensive, low-latitude ice ages (e.g. Kirschvink, 1992), dramatic fluctuations in seawater chemistry that lead to further oxygenation of the atmosphere (e.g. Des Marais et al., 1992), and unique sedimentary textures, such as molar tooth structure, were preserved in rocks (e.g. James et al., 1998).

Situated in the Canadian Arctic, the Tonian-Cryogenian Shaler Supergroup is a thick succession of sedimentary rocks that were deposited in the Amundsen Basin, a shallow intracratonic basin within an epeiric sea that formed at the core of Rodinia. The rocks are well preserved, and are relatively well exposed on Victoria Island, NWT. The Minto Inlier is a northeast trending belt of Neoproterozoic rock exposed on Victoria Island, and is an ideal field locality to study sedimentary rocks of Neoproterozoic affinity (Fig. 1).

This is a thematic study that was supported by the Geological Survey of Canada's (GSC) Geomapping for Energy and Minerals (GEM) Victoria Island project. Fieldwork was carried out during the summers of 2010 and 2011 on Victoria Island, NWT. The purpose of the GSC's GEM Victoria Island project was to map the southwest portion of

Figure 1. Geology of the Minto Inlier (after Thorsteinsson and Tozer, 1962) and stratigraphy of the Shaler Supergroup, with study locations highlighted by stippled ovals. The NE study area also corresponds to the study area of Jones et al. (2010). Note the black star is the location of drill core GNME 07-04. Inset in top left corner shows Victoria Island in grey, the bounding box is the extent of the map. Re-Os ages from van Acken et al. (2013), U-Pb detrital zircon ages from Rayner and Rainbird (2013), and U-Pb lead ages from Heaman et al. (1992) and Macdonald et al. (2010).

Minto Inlier (Fig. 1), and evaluate the possible economic potential of those rocks. There were several PhD, M.Sc, and B.Sc. honours projects associated with the GSC mapping

project. This project focused on the stratigraphy, sedimentology, and geochemistry of the upper Shaler Supergroup.

Along with detailed outcrop analysis, there is also stratigraphic information preserved in diamond drill cores left in the field by Great Northern Mining Exploration (GNME), who explored for base metals in the Minto Inlier from 2005-2007. Although drilling targeted the Franklin diabase sills, it recovered several hundred meters of sedimentary strata, including almost complete sections of the Jago Bay and Boot Inlet formations, and core through the part of the upper Wynniatt Formation.

1.2 Location and Access

The study area is located in the Minto Inlier, on Victoria Island, NWT, Canada. There were two main areas of study, one located in the southwest that centered around Minto Inlet, the other in the northeast, in the vicinity of Kilian Lake (Fig. 1). A list of coordinates for all stratigraphic sections measured is provided in Appendix 1.

The field area is remote and accessed via helicopter or bush plane equipped with tundra tires. One can get a flight to the hamlet of Ulukhaktok from Yellowknife, and from there a helicopter must be taken to access section localities. The sections in the northeast around Kilian Lake are a long flight from Ulukhaktok and require a mid-way fuel cache for a helicopter.

1.3 Geologic setting

The Tonian-Cryogenian (~1000 – 723 Ma) Shaler Supergroup is exposed in several inliers on the mainland and adjacent islands of Arctic Canada (Minto, Cape Lambton, Duke of York, Brock, and Coppermine inliers) (Rainbird et al. 1994; Rainbird et al. 1996; Thorsteinsson and Tozer 1962; Young 1981). On Victoria Island, the Shaler Supergroup is exposed in the Minto Inlier, which preserves a broad NE-SW trending, shallowly dipping, anticline-syncline pair; the Holman Island Syncline-Walker Bay Anticline. The folded Neoproterozoic strata are unconformably overlain by early Cambrian sandstones to late Ordovician carbonates (Thorsteinsson and Tozer, 1962).

The Shaler Supergroup is more than four kilometers thick and composed of fluvio-deltaic sedimentary rocks, platform carbonate rocks and sulphate evaporite rocks interpreted to have been deposited in the Amundsen Basin, a shallow embayment of a larger epeiric sea, situated in the interior of supercontinent Rodinia (Fig. 2; Evans 2009; Li et al. 2008; Rainbird et al. 1996; Rainbird et al. 1998; Young 1981). This epeiric sea would have been periodically connected to the open ocean via seaways, analogous to the Western Interior Sea of North America during the Cretaceous (Williams and Stelck 1975). Rainbird et al., (1996) interpreted the Amundsen Basin as an intracratonic sag basin, based on the lack of synsedimentary faulting, particularly in the lower Shaler Supergroup; as well as the lateral continuity of stratigraphic units, some for over 1000 kilometers (see also Long et al., 2008).

The maximum age of the Shaler Supergroup is constrained by detrital zircon geochronology at <1151 Ma (Rayner and Rainbird, 2013) and the minimum age at 723 ±4/-2 Ma by U-Pb geochronology of diabase sills that feed overlying flood basalts, part

of the Franklin Large Igneous Province (Heaman et al., 1992). Three depositional ages of

ca. 780 Ma; modified from Evans 2009

Figure 2. Paleogeographic reconstruction of Rodinia, modified from Evans, 2009. Location of Amundsen Basin indicated and epeiric sea with interconnecting seaways overlain. There are several different reconstructions of Rodinia that have been proposed, the paleogeographic fit of the continents is still being debated. I chose to present this version as an example because it highlights the interior sea, where the Amundsen Basin is located. Most workers now agree however that Laurentia was situated at the core of Rodinia, likely with Siberia to the north and Australia to the west.

892 ± 13 Ma (Boot Inlet Formation), 849 ± 48 Ma (Black Shale Member, of the Wynniatt Formation), and 761 ± 41 Ma (Upper Carbonate Member of the Wynniatt Formation),

dated using Re-Os black shale geochronology add time resolution to the stratigraphy (Fig. 1; van Acken et al., 2013).

The Shaler Supergroup is composed of, in ascending stratigraphic order, the Rae Group, comprising the Escape Rapids, Mikkelsen Islands, Nelson Head, and Aok formations, the Reynolds Point Group, comprising the Grassy Bay, Boot Inlet, Fort Collinson, and Jago Bay formations, and the ungrouped Minto Inlet, Wynniatt, Kilian, and Kuujjua formations (Fig. 1). The sedimentary rock units are overlain by the Natkusiak Formation, which is composed of continental flood basalts that are fed by the Franklin gabbro sills and dykes, which intrude the sedimentary succession.

The Minto Inlier is an excellent locality for stratigraphic study because: (1) it hosts the most complete and regionally extensive exposure of Shaler Supergroup strata; (2) strata are relatively undeformed with dips typically less than 10 degrees; and (3) rocks have experienced only low-grade contact metamorphism by diabase sills.

1.4 Previous work

The first descriptions of the geology of Victoria Island come from the arctic expeditions of O'Neill (1924) and Washburn (1947). Thorsteinsson and Tozer (1962) published a geologic map of Victoria Island, and proposed the Shaler Group and descriptions of the original formations within. Detailed descriptions and stratigraphic logs of sedimentary units are given in Young (1981), which incorporate previous studies on the sedimentology and stratigraphy of the Shaler Supergroup by Young and Long (1976, 1977a, 1977b). Updated descriptions and subdivisions of Minto Inlier stratigraphy is

provided in Rainbird et al. (1994), where several new formations were introduced, elevating the Rae and Reynolds Point formations to Group status, and the Shaler Group to Supergroup status.

Young (1979) subdivided the Proterozoic stratigraphy of northern Canada into three, unconformity-bound, supersequences. In that framework, the Shaler Supergroup belongs to Sequence B. Long et al. (2008), further subdivided sequence B into five sub-sequences that correlate across northern Canada. Sequence stratigraphic analysis and sedimentological descriptions of the reefs in the Boot Inlet Formation was preformed by Narbonne and James (2000). Rainbird (1993) interpreted four T-R cycles for the Kilian and Kuujjua formations.

Field observations of the Shaler Supergroup indicate a northwesterly sediment transport direction, and detrital zircon geochronology includes a large proportion of Grenville-age zircons, which led to the interpretation of continental-scale river systems that transported sediments of Grenville affinity across Laurentia, where they were deposited in the Amundsen Basin (Rainbird et al., 1992 & 1997; see also Rayner and Rainbird, 2013).

Previous geochemical studies on the sedimentary rocks of the Shaler Supergroup include Asmeron et al., (1991), who presented Sr isotope data on all the main carbonate units, and Jones et al., (2010), who presented a C and O isotope chemostratigraphic curve for the Shaler Supergroup, and used this to propose a new stratigraphic correlation scheme with the Mackenzie Mountains Supergroup, located in the Mackenzie Mountains of the northern Cordillera.

The *Tawuia-Chuaria* macrofossil assemblage zone was discovered in the Wynniatt Formation by Hofmann and Rainbird (1995), and it was correlated with the upper Little Dal Group, of the Mackenzie Mountains Supergroup. The organic-walled macrofossils are described as carbonaceous compressions, and are unicellular forms of indeterminate affinity. Several different acritarchs, filamentous cyanobacteria, and other microfossils have been described from the Wynniatt Formation (Butterfield and Rainbird, 1998; Butterfield 2005).

1.5 Study objectives

The goals of this research project are:

1) To further refine the stratigraphic members of the Wynniatt Formation, Shaler Supergroup. There is presently much global interest in Neoproterozoic successions, and detailed description of this remote succession adds to the greater body of knowledge of carbonate successions from this time period. Further, the Wynniatt Formation was mapped at the member scale (Rainbird et al., 2013), and therefore these members must be properly characterized and defined.

2) Interpret the depositional environments of the Wynniatt Formation. This involves detailed sedimentological descriptions of units, interpretations of different facies, and a working model of the different depositional environments and how they changed through time.

3) To characterize the evolution of the carbonate ramp during deposition of the Wynniatt Formation. If this formation was deposited in ~100 million years, can one

recognize discrete stages of ramp development? Are there changes in the depocentre location?

4) To provide a sequence stratigraphic framework for the upper Shaler Supergroup. What is driving subsidence patterns that control cyclicity within the basin? Further, recent studies have documented subsidence patterns and tectonostratigraphy of strata deposited elsewhere along the margin of Laurentia. How do these patterns compare to those observed in the intracontinental Amundsen Basin?

5) To document C isotope stratigraphy through the upper Shaler Supergroup, and compare variations observed both locally (NE vs. SW Minto Inlier) and with similar data from the Mackenzie and Ogilvie mountains. Jones et al. (2010) addressed the chemostratigraphy of the Shaler Supergroup in the NE part of the Minto Inlier and recognized a significant negative excursion in the $\delta^{13}\text{C}$ values within the Wynniatt Formation. This anomaly was correlated with the global Bitter Springs Stage, however the samples were collected directly below a ~50 m thick diabase sill, heat from which could have altered the primary isotope values. Detailed sampling through this interval in the SW portion of the inlier, where no diabase is present in the section, tests the validity of this excursion. New data from both outcrop and drill core is obtained to fill in data gaps.

6) To interpret the cause of the large negative excursion in the Wynniatt Formation and investigate why the negative values are amplified in the Amundsen Basin.

7) To interpret the basin redox history using a multi-proxy approach. Redox-sensitive trace metals, pyrite sulfur isotopes, carbonate carbon and organic carbon

isotopes, and Fe speciation data is incorporated to interpret the redox history of the Amundsen Basin and how this relates to contemporaneous global seawater patterns.

1.6 Methodology

A methods section for each chapter is provided that gives the methodology of specific analytical techniques relating to the analyses done for that specific study. See sections 2.4, 3.3, and 4.3 for specific descriptions of methods used. Stratigraphic analysis was performed at the outcrop scale for sections measured throughout Minto Inlier. Available outcrop locations imposed restrictions on stratigraphic analysis as they were predominantly limited to a transect that coincided with depositional strike, limiting the construction of cross-sections perpendicular to the regional strike and dip of Amundsen Basin.

CHAPTER 2: SEDIMENTARY ARCHITECTURE OF THE WYNNIATT FORMATION

Thomson, D., Rainbird, R.H., Dix, G., 2014. Architecture of a Neoproterozoic intracratonic carbonate ramp succession: Wynniatt Formation, Amundsen Basin, Arctic Canada. Sedimentary Geology, 299: 119-138

2.1 Introduction

The geologic record of Neoproterozoic intracratonic basins are now widely distributed on various continents following the breakup of the supercontinent Rodinia (e.g., Li et al., 2008). These types of basins are characterized by: long-lived (>200 m.y.) depocentres with slow subsidence rates that may be coupled with shifting depocentres (e.g. Howell and van der Pluijm, 1990), lack of major syn-sedimentary faulting, a general paucity of deep-water facies, and basin-fill patterns that are dominated by shallow-marine carbonate strata, basin-central evaporite facies, and terrestrial siliciclastic rocks that may form peripheral coastal facies belts (Allen and Allen, 2005).

These attributes characterize a long-lived (50+ m.y.) carbonate succession constituting the Neoproterozoic Wynniatt Formation, Shaler Supergroup, in the Amundsen Basin of the Canadian Arctic (Fig. 1). The formation is exposed in the Minto Inlier (Fig. 1), in which strata are relatively undeformed and the sedimentary rocks have experienced, at most, only low-grade contact metamorphism associated with intrusion of diabase sills (Fig. 3). We present a sedimentological framework illustrating stages of

carbonate ramp development allied with changing subsidence depocentres. These

Figure 3. Selected measured section localities, Wynniatt Formation. Dashed red lines show approximate lines of where the sections were measured. 1) Sections 10DTS3 and 10DTS4 from the SW part of the inlier (see Table 1 for locations). View to the north-northeast across Minto Inlet. Mi (Minto Inlet Formation) overlain by LCM (Lower Carbonate Member), BSM (Black Shale Member), and SCM (Stromatolitic Carbonate Member). 2) Section 10DTS8, showing the contact of the SCM and UCM (Upper Carbonate Member). View to the north, SW Minto Inlier. 3) Section 11DTS9 measured in the NE part of the inlier, where the LCM, BSM, and part of the SCM are exposed. View to the southwest.

improve the understanding of basin and carbonate ramp development for a period likely overlapping with initiation (~830 Ma) of supercontinent (Rodinia) breakup and suggest a mechanism to explain local sedimentation patterns, which has implications for interpretations of regional paleogeography and tectonic evolution of Amundsen Basin. Our work provides sedimentary context for sequence and chemical stratigraphy (Jones et al., 2010) and biostratigraphy (Hofmann and Rainbird, 1995) within the Amundsen Basin, and for improved correlation among coeval Neoproterozoic basins.

2.2 Geologic Setting

The Amundsen Basin is interpreted as an intracratonic sag (Rainbird et al., 1996) based on the lack of synsedimentary faulting, particularly in the lower Shaler Supergroup, as well as the lateral continuity of stratigraphic units, some for over 1000 kilometres (see also Long et al., 2008). The basin split apart during break-up of Rodinia but the location of the conjugate margin is unknown (Young, 1981). The Amundsen Basin is interpreted to represent an embayment along a larger epeiric basin within the supercontinent Rodinia (Young, 1981; Rainbird et al., 1996; Rainbird et al., 1998; Li et al., 2008; Evans, 2009). The epeiric sea was episodically connected to the open ocean via seaways, analogous to the Western Interior Seaway of North America during the Cretaceous (Williams and Stelck, 1975).

The Shaler Supergroup is exposed in several inliers on the mainland and adjacent islands of Arctic Canada (Thorsteinsson and Tozer, 1962; Young, 1981; Rainbird et al., 1994; Rainbird et al., 1996) and is represented by more than four kilometres of sedimentary rocks along with intrusive (diabase sills) and extrusive (continental flood basalts) igneous rocks (Fig. 1). The Wynniatt Formation occurs in the upper Shaler Supergroup (Fig. 1), exposed only in the Minto Inlier of Victoria Island (Fig. 1). It conformably overlies basin-central sulphate evaporite and marine carbonate rocks of the Minto Inlet Formation and is unconformably overlain by sabkha-style evaporite rocks and marine carbonate strata of the Kilian Formation (Fig. 1).

A broadly defined geochronological framework for the Shaler Supergroup includes maximum depositional ages based on U-Pb detrital zircon geochronology (Fig.

1), the Escape Rapids Formation gives a lower age of 1115 ± 13 Ma (Rainbird et al., 1992; Rainbird et al., 1997; Rayner and Rainbird, 2013). Shaler Supergroup strata were intruded by diabase sills dated at $723 \pm 4/-2$ Ma (Heaman et al., 1992) and 716.33 ± 0.54 Ma (Fig. 3; Macdonald et al., 2010). Re-Os geochronology of black shale units representing the lower to middle parts of the Wynniatt Formation identify a depositional age range from 849 ± 48 Ma to 761 ± 41 Ma, respectively (Fig. 3; van Acken et al., 2013). A negative carbon isotope excursion reported from the Wynniatt Formation is correlated to the global Bitter Springs stage carbon isotope anomaly (Halverson et al., 2005; Jones et al., 2010; Thomson et al., 2012), which has been dated at 811.5 Ma (Macdonald et al., 2010).

2.3 Previous Work

The Wynniatt Formation was proposed and first described during reconnaissance geological mapping of Victoria Island by Thorsteinsson and Tozer (1962). There is no type section; a composite section from the northeastern Minto Inlier (Young, 1981) located near the east side of Kilian Lake (Fig. 1) represents the only detailed lithological analysis of the Wynniatt Formation. The Wynniatt Formation contains a diverse assemblage of acritarchs and other organic-walled and filamentous microfossils, and the macrofossils *Chuaria* and *Tawuia* (Hofmann and Rainbird, 1995; Butterfield and Rainbird, 1998; Butterfield, 2005).

2.4 Methods

The Wynniatt Formation is divided here into four, informal, regional members, correlated across the Minto Inlier. A total of 20.8 km of stratigraphic section, including 16 outcrop sections and three drill cores, was measured (Fig. 4; Appendix 1). Owing to the constraints of exposure, some member descriptions rely on composite sections (Fig. 4).

Lithofacies are defined by lithology, texture, and sedimentary structures (Fig. 5). Stromatolites are informally subdivided according to general morphology; in some cases tentative identifications of form-genera are included.

2.5 Results

2.5.1 Stratigraphic Members, Wynniatt Formation

The basal contact of the Wynniatt Formation with the underlying Minto Inlet Formation is inferred to be gradational based on facies stacking patterns (see below). The Lower Carbonate member (Figs. 4 - 6; Table 1) is characterized by lime mudstone with parallel laminae, desiccation cracks, teepee structures, chert nodules and intraclast breccias in the lower part, wavy to nodular bedded, hummocky cross-stratified grainstone interbedded with molar-tooth lime mudstone and minor shale in the middle part, and interbedded stromatolitic dolostone, dolograinstone, and black silty shale in the upper part. A regionally extensive marker bed of distinctive orange-weathering stromatolitic dolostone forms the upper 10 metres of the member. The Lower Carbonate and Black

Shale member contact is gradational. The Black Shale member is a recessive interval of

Figure 4. Sections through the members of the Wynniatt Formation. Coordinates for sections are provided in Table 1. The Black Shale member and Upper Carbonate member sections are composite sections. Note: stratigraphic location of *Tawuia-Chuarina* assemblage shown at the Black Shale and Stromatolitic Carbonate member contact, correlated from Hofmann and Rainbird (1995). Sh – shale, m – mudstone, w – wackestone, p – packstone, g – grainstone, stm – stromatolitic lime/dolostone, r – rudstone.

dark-grey siltstone and silty shale with flute and gutter casts, ball and pillow structures, channel and fill structures, and climbing ripple marks in siltstone. Thin- to medium-thick, discontinuous-to-continuous beds of ripple-topped quartz arenite are dispersed

Legend

Figure 5. Legend for stratigraphic sections and paleoenvironmental reconstruction.

throughout the member but are more commonly preserved near its top. The *Tawuia-Chuaria* zone described by Hofmann and Rainbird (1995) was documented from the upper ~5 m of this member, as well as a diverse suite of acritarchs and other organic-walled and filamentous microfossils (Butterfield and Rainbird, 1998; Butterfield, 2005), from within an interval of wavy-lenticular sandstones interbedded with siltstone and shale. The Black Shale and Stromatolitic Carbonate member contact is gradational. The Stromatolitic Carbonate member contains four main lithologies: stromatolitic dolostone, intraclast grainstone, molar-tooth mudstone, and mud-cracked dololite. A regionally extensive interval of stromatolitic dolostone contains oncolites up to 20 cm in diameter (see Fig. 8). The Stromatolitic Carbonate and Upper Carbonate member contact is sharp. The lower Upper Carbonate member is composed of nodular, black, calcareous shale, succeeded by a thick middle interval of thinly bedded, quartz-sandy, normally graded calcarenite, overlain by a rhythmic successions of metre- to decimeter-thick interbedded stromatolitic dolostone and crossbedded intraclast grainstone. The contact with the overlying Kilian Formation is unconformable.

Member	Thickness	Description	Nature of contacts	Age and correlation	Figure reference
<i>Lower Carbonate Member (LCM)</i>	~130 m in SW; ~85 m in NE (Young, 1981)	Lower LCM (0 ~ 45 m): lime mudstone with minor amounts of black shale; parallel laminae, desiccation cracks, teepee structures, chert nodules, intraclast breccias, carbonate nodules, lenticular beds of quartz arenite, rip-up clasts of mudstone, and shallowly scoured beds in lime mudstone. Middle LCM (~45 – 90 m): wavy to nodular bedded, hummocky cross-stratified grainstone interbedded with lime mudstone and minor shale. Molartooth structure common. Upper (~90 – 125 m) LCM: interbedded stromatolitic dolostone, dolograstone, black silty shale. A regionally extensive marker bed of distinctive orange-weathering stromatolitic dolostone forms the upper ten meters of the subdivision (Fig. 3)	Lower contact with Minto Inlet Fm. is not exposed but interpreted as gradational based on internal facies patterns; upper contact with BSM is sharp where dark grey shale overlies orange-weathering stromatolitic dolostone	<911 Ma, >849±48 Ma; correlative to the Snail Spring Fm. (MMSG) (Turner and Long, 2012) and Reefal Assemblage (FMG) (Halverson et al., 2012)	section 10DTS3; Table 1, Fig. 5
<i>Black Shale Member (BSM)</i>	~200 m in SW; <10 m in NE (Young, 1981),	Recessive; dark grey siltstone and silty shale with flute and gutter casts, ball and pillow structures, channel and fill structures, and climbing ripple marks in siltstone. Decimeter-thick interbeds of stromatolitic dolostone and dololite near the contact with the underlying LCM. Lenticular carbonate nodules and pyrite throughout. Thin- to medium-thick, discontinuous-to-continuous beds of ripple-topped quartz arenite throughout but more commonly preserved near top	Upper contact is gradational and drawn at the first reappearance of resistant-weathering dololites	849±48 Ma, correlative to the Snail Spring Fm. (MMSG) and Reefal Assemblage (FMG)	section 10DTS4; Table 1, Fig. 5
<i>Stromatolitic Carbonate Member (SCM)</i>	160 m in SW; 320 m in NE (Young, 1981)	Four main lithologies: stromatolitic dolostone, intraclast grainstone, molartooth mudstone, and mud-cracked dololite. Stromatolitic dolostone forms buildups that together with onlapping strata, extend laterally from one meter up to ~20 meters, have local topographic relief of several meters. A regionally extensive (>200 km, NE - SW) interval of stromatolitic dolostone contains oncolites up to 20 cm in diameter (see Fig. 7). Interbedded intraclast grainstone and mudstone are wavy-to-nodular bedded, and contain rip-up clasts and intraformational breccia units with scoured lower bounding surfaces. Dololite is variably parallel-laminated or microbially laminated, with mudcracks, teepee structures, and brecciated beds.	Contact with the overlying UCM is sharp, erosive and unconformable. Marks an abrupt change to an interval of black, calcareous shale (Fig. 3)	<849±48 Ma, >761±41 Ma, correlative to Ram Head Fm. (MMSG) (Turner and Long, 2012) and upper Reefal Assemblage (FMG)	10DTS7; Table 1, Fig. 5
<i>Upper Carbonate Member (UCM)</i>	~380 m in SW; ~110 m in NE (Young, 1981)	Lower UCM (0 ~ 160 m): nodular, black calcareous shale, succeeded by a thick middle interval of thinly bedded, quartz-sandy, normally graded calcarenite. Upper UCM (~160 – 980 m): rhythmic successions of meter to decimeter thick interbedded stromatolitic dolostone and cross-bedded intraclast grainstone. Local occurrences of herringbone cross-bedded quartz-arenite and microbially laminated lime mudstone. Chert is common as nodules or tabular layers.	Upper contact unconformable with Kilian Fm.: a three-meter thick black shale forms the basal Kilian Fm., which sharply overlies stromatolitic dolostones of the Wynniatt Fm. The black shale is abruptly overlain by sulphate evaporites.	Lower UCM has a Re-Os age of 761±41 Ma; correlative to upper Reefal Assemblage (FMG); upper UCM correlative with upper Ram Head Fm. (MMSG)	(10DTS8 & 9, 11DTS2-4; Table 1, Fig. 5)

Table 1. Descriptions of members of the Wynniatt Formation. MMSG – Mackenzie Mountains Supergroup, FMG – Fifteenmile Group.

2.5.2 Lithofacies

2.5.2.1 Carbonaceous calcareous black shale and calcarenite facies (F1; Table 2)

Description.--- This facies consists predominantly of laminated carbonaceous calcareous black shale. It is interstratified with laminae of green calcisiltite, thin beds of grey calcareous claystone, and massive to nodular beds (>20 cm thick) of calcarenite composed of rounded lithoclasts of fine to medium sand with erosive or sharp bases, and flattened discoidal lenses (<50 cm in diameter) of similar calcarenite (Fig. 4). Loading and dewatering structures (flame structures) are present in the shale. Disseminated crystalline pyrite occurs throughout. Early seafloor dissolution and cementation is visible along bedding contacts as irregular contacts that are buckled or micro-fractured, and are draped by overlying shale. Beds typically are cemented early by micritic calcite cement. Paleoporosity is filled by blocky, subhedral, cloudy, non-equidimensional calcite cement.

Interpretation.--- This facies is interpreted to represent predominantly suspension sedimentation below storm wave base (SWB) on the outer slope. Calcarenite beds are interpreted as turbidites, and correspond to Bouma Ta subdivisions. Nodules of calcarenite, the same composition as Ta turbidites, were formed *in situ* during early diagenesis, as uneven sediment loading and compaction resulted in differential compaction of shale around the nodules. Early-cemented beds reflect condensed intervals or periods of non-deposition where sedimentation rates were low enough to allow the seafloor to become at least partially lithified. Pyrite is diagenetic.

Figure 6. Examples of outer- and mid-ramp facies. Hammer is 30 cm long. A) *Carbonaceous calcareous black shale and calcarenite facies* (F1), photo from drill core GNME 07-04. B) *Lithoclastic float/rudstone with dololutite matrix facies* (F2). C) *Thin-bedded calcarenite facies* (F3). D) *Conoform stromatolitic dolostone facies* (F7). E) *Dark grey siltstone and silty shale facies* (F5). F) *Molar-tooth mudstone facies* (F8).

2.5.2.2 *Lithoclastic float/rudstone with dololutite matrix facies (F2)*

Description.--- This facies is characterized by beds, up to 75 cm thick, of carbonate rudstone and/or floatstone containing clasts of dololutite that are up to cobble-size, sub-rounded, and tabular (Fig. 6). Lithoclasts may be aligned randomly or imbricated along the bases of beds, which are typically erosional. Clasts are supported in a dololutite matrix. Float/rudstone beds are interbedded with dololutite beds, 5-30 cm thick. Thin shale interbeds are also associated with this facies.

Interpretation.--- This facies is interpreted as debris flows deposited along the outer slope. Debris can be transported for long distances along low-angle slopes (Cook and Mullins 1983), but these flows were probably at least initiated at some break-in-slope, where the angle of repose steepened to generate debris flows. The similar composition of clasts to underlying beds suggests that float/rudstone beds were cannibalized from underlying strata. This facies is interpreted as slope-derived debrites interbedded with dololutites deposited from suspension, below SWB.

2.5.2.3 *Thin-bedded calcarenite facies (F3)*

Description.--- This facies comprises thin, regularly layered and well-defined (2 cm – 10 cm) beds of fine-grained calcarenite that are normally graded to parallel laminated (Fig. 6). Thicker beds contain up to ~25% well-rounded quartz sand grains. Beds have consistent thickness and thin, intervening mud laminae are recessive such that

the sand beds stick out as ribbons; similar rocks have been referred to as ribbon limestones (i.e. Sherman, et al. 2000). Generally, beds of calcarenite have sharp non-erosive to erosive lower contacts that may preserve thin, coarse-grained lags of randomly oriented tabular clasts of dololomite.

Interpretation.--- Beds of calcarenite correspond to Bouma Ta and Tb subdivisions and are interpreted to reflect turbidite deposition. Carbonate sand was derived from higher energy environments that also contained well-rounded quartz sand (e.g. sand shoals), possibly introduced from a nearby coastal setting (e.g. Balazs and Klein 1972), and then redeposited down-slope via turbidity currents possibly triggered by storms. The presence of the Bouma Tb subdivision supports a more distal position for initiation of turbidity flows.

2.5.2.4 Mg-silicate and evaporite facies (F4)

Description.--- This facies is restricted to the Upper Carbonate member. It occurs only at one locality, stratigraphically positioned within sedimentary rocks of F2, a few metres above sedimentary rocks of F1. The lower contact of this facies is erosional, and the upper contact is gradational. It consists of approximately 15 metres of interbedded black, poorly indurated calcareous claystone, shale, nodular lime mudstone, and calcisiltite. Sedimentary structures include slump folds, rip-up clasts, chaotic bedding, flat-chip and edgewise breccias, and wavy-nodular compactional bedding (Fig. 7). Enigmatic textures that resemble Liesegang bands or some diagenetic front occur in lime

mudstone (Fig. 7). Petrography reveals bottom-grown crusts (<1 mm thick) composed of

Figure 7. Photographs of *Mg-silicate and evaporite facies* (F4). Hammer is 30 cm long. A) Nodular bedded calcareous shale and Mg-silicate. B) Bottom-grown crusts of gypsum replaced by dolomite in a ferroan calcite matrix. Thin section has been died using Alizarin Red. C) Mg-silicate clast debris in a calcareous shale matrix. D) Enigmatic diagenetic texture resembling leiseegang banding. E) Doubly-fibrous Mg-silicate (Mg-Si) crystals within a calcium carbonate (CC) matrix. F) Slump folds within lime mudstone.

gypsum bushes that are replaced by dolomite and interlaminated within a ferroan calcite matrix (Fig. 7). Blue-purple, fibrous, doubly radial crystals within a lime mudstone matrix have medium birefringence, yellow-orange interference colours and through SEM analysis are interpreted as a Mg-silicate mineral, most likely talc ($\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2$).

Interpretation.--- This facies is interpreted to have been deposited below SWB due to its association with other deep-water facies, lack of wave-generated sedimentary structures, and fine-grained nature. Sedimentation was gravity-driven, with preserved debris flows, slumping, and suspension-settled shale. Slump folds allude to the possibility of at least some over-steepening of the slope to allow downward creep of sediments. However, this facies also represents a period of basin restriction allowing precipitation of both sulphate evaporites and sedimentary talc (cf. Tosca et al., 2011). The growth of the talc occurred very early because it does not replace the carbonate, but appears to be intergrown. Bottom-grown crusts of an evaporite mineral suggest that seawater was supersaturated with respect to sulphate, and that the depositional environment was low energy allowing for the growth of this delicate fabric. This quiet-water environment, however, was interrupted by debris flows. Sedimentary talc has also been described from the Callison Lake Dolostone in the Ogilvie Mountains, and Akademikerbreen Group of Svalbard (Tosca et al., 2011), where it is interpreted to have precipitated in shallow water.

2.5.2.5 Dark grey siltstone and silty shale facies (F5)

Description.--- The overall grain size varies from silt- to mud-dominated, with minor amounts of fine-grained sand (Fig. 6). Sedimentary structures include flute and gutter casts, ball-and-pillow, channel-and-fill, tool marks, and ripple crosslamination in siltstone. Siltstone and shale layers are laminated to thinly bedded. Rare beds of thin to thickly bedded, fine- to medium-grained, well-rounded, well-sorted quartz arenite are preserved within the siltstone and shale. Quartz arenite beds are sharp-based and laterally discontinuous preserving hummocky cross-stratification (HCS) and asymmetrical ripple marks. Carbonate nodules and pyrite are present throughout. In thin section, imbricated rip-up clasts of siltstone are present as well as carbonate microconcretions whose cores are commonly partially to completely replaced by pyrite.

Interpretation.--- This facies is interpreted to represent influx of siliciclastic detritus to the basin possible on a pro-delta slope between fair-weather and storm-weather wave base. Sandstone beds represent storm events that reworking delta-front sands and re-deposited them on the slope. HCS is widely accepted as deposited by storm-generated flows (e.g., Walker and Plint 1992; Dumas, et al. 2005). Asymmetric ripples are the result of waning flow during deposition of storm-suspended sands.

2.5.2.6 *Black shale facies (F6)*

Description.--- Thin beds of black to grey shale are distributed throughout the Wynniatt Formation, either at the tops of upward-deepening successions or at the bases of upward-shallowing successions. Units range in thickness from paper-thin partings between beds, to beds of medium thickness, but most commonly they occur as thin

interbeds within carbonate rocks. This facies differs from F1 in its lack of carbonate, and from F5 in its lack of sedimentary structures or sandstone interbeds. Also, this facies does not occur in thick successions like those described above.

Interpretation.--- This facies is interpreted as suspension-settled sediment deposited below fair-weather wave base (FWWB), during slack water intervals between storms.

2.5.2.7 *Conoform stromatolitic dolostone facies (F7)*

Description.--- This facies is composed of conical, columnar stromatolites encased in dolomudstone (Fig. 6) in the Lower Carbonate member. Solitary conical stromatolites attain a maximum observed height of ~20 cm and width of ~10 cm, with several centimetres of synoptic relief. Wall-parallel laminae are steeply dipping ($>70^\circ$) with narrow, acute apices. Inter- and intrastromatolite sediment consists predominantly of dolomudstone.

Interpretation.--- Stromatolites fit the form of *Conophyton* sp. that are indicative of relatively deep water (e.g. Hoffman 1976), and are typically associated with transgressions (Kah, et al. 2009). Similar facies have been described from the Mesoproterozoic Atar Group in Mauritania (e.g. Bertrand-Sarfati and Moussine-Pouchkine 1985). There, the lack of coarse-grained interstromatolite sediment and high synoptic relief were interpreted to indicate deposition in quiet water, below FWWB.

2.5.2.8 *Molar-tooth mudstone facies (F8)*

Description.--- Light- to dark-grey limestone and dolostone characterize this facies. Beds are thin- to thick-bedded and generally massive. Strata are commonly organic rich and silty, with molar-tooth structure that is sparsely distributed to completely pervasive (Fig. 6). Ptygmatically folded, sub-vertical cracks are the most common form, but horizontal sheet cracks, simple vertical spindle, and tadpole forms are also preserved. Molar-tooth mudstone facies occurs as a monotonous stacking of beds, interbedded with F10 (see below), or at the tops of graded floatstone or packstone beds. Shale is commonly interbedded with beds of molar-tooth mudstone.

Interpretation.--- Molar-tooth structure is ubiquitous in Proterozoic carbonate rocks and is widely regarded as paleoenvironmentally equivalent to bioturbated, fossiliferous subtidal limestones of the Phanerozoic (James, et al. 1998). There is controversy over the origin of molar-tooth structure, whether it is created through microbially induced gas bubble and expansion cracks (Furniss, et al. 1998), synsedimentary earthquake activity (Pratt 1998), or clathrate destabilization (Marshall and Anglin 2004). Despite the debate on its origin, it is generally agreed that molar-tooth forms within the subtidal depositional regime.

2.5.2.9 *Molar-tooth clast packstone facies (F9)*

Description.--- Molar-tooth clast packstone facies is most commonly preserved sharply overlying beds of molar-tooth mudstone (Fig. 8), and is also interbedded with

F10 (see below). The lower contact between beds is erosive, and the upper contacts may

Figure 8. Examples of mid- to inner ramp facies. Hammer is 30 cm long. A) *Cross-bedded intraclast grainstone facies* (F10). B) *Molar-tooth clast packstone facies* (F9). C) *Lithoclast floatstone with grainstone matrix facies* (F11). D) Polished slab of *oncoide-digitate stromatolite facies* (F12). Note the yellow arrow points to a broken piece of stromatolite that the oncoide is nucleated on, and that digitate stromatolites grew upward (toward top of photo) from the top of the oncoide. E) *Cross-bedded peloid grainstone facies* (F14). Note the apparent bi-directionality of the cross-beds shown in this photo (arrows). F) *Dolomitic siltstone with desiccation cracks facies* (F15).

be gradational or sharp. Beds of molar-tooth clast packstone typically are graded and clasts are imbricated. The matrix consists of fine-grained detrital carbonate and microspar cement.

Interpretation.--- The spatial and repetitive stratigraphic association with molar-tooth mudstone, and the presence of graded texture and imbricated clasts, are taken to identify this facies as subtidal tempestites arising from the erosion and re-working of molar-tooth mudstone. This supports molar-tooth structure as a product of early diagenesis and a process-generated, as opposed to an event-generated, structure (Furniss et al., 1998).

2.5.2.10 Cross-bedded intraclast grainstone facies (F10)

Description.--- Medium-thick beds of HCS, low-angle cross-bedded, trough cross-bedded, and rare swaley cross-stratified (SCS) intraclast grainstone characterize this facies (Fig. 8). Bases of beds commonly are scalloped erosional surfaces. Allochems consist of reworked clasts of molar-tooth microspar, coated grains, peloids, stromatolitic fragments, and rounded lithoclasts of lime mudstone. Grainstones form thick (>1 metre) amalgamated bodies, or thin sheets interbedded with other facies (e.g., facies F8, F9, and F13).

Interpretation.--- Grainstones require constant wave motion for winnowing of mud and reworking of intraclasts. The close association of this facies with molar-tooth mudstones (F8) and tempestites (F9), as well as the preservation of HCS, positions this facies along a hydrodynamic transition from below to above FWFB. This facies is

interpreted as storm-transported, subtidal sediment, deposited above FWFB. Allochem composition varies depending on adjacent and underlying sediment composition.

2.5.2.11 Lithoclast floatstone with grainstone matrix facies (F11)

Description.--- Lithoclast floatstone with grainstone matrix facies is not abundant in the Wynniatt Formation. Beds are medium-thick and laterally continuous with erosive bases (Fig. 8). Lithoclasts are usually tabular, gravel sized, and subrounded with variable lithologies, including calcilutite, calcisiltite and ooid grainstone. Grainstone allochems variously include peloids, pisoids and rare ooids. Lithoclasts may be imbricated or randomly oriented and grading is faintly distinguishable in some beds. Rounded quartz sand grains are present locally.

Interpretation.--- Beds of lithoclast floatstone with grainstone matrix facies are interpreted as storm lags. Increased wave action during storms eroded proximal shallow water sediments, scouring into beds and re-depositing lithoclasts. Ooids are rare throughout the Wynniatt Formation; conditions must not have been conducive to their growth. This facies is interpreted as a shallow, grain-rich, subtidal flat that episodically experienced large storms. This facies is similar to intraclast flat-pebble conglomerate subfacies described by Clough and Goldhammer (2000), from the Neoproterozoic Katakturuk Dolomite in Alaska.

2.5.2.12 Oncoid-digitate stromatolitic dolostone facies (F12)

Description.--- This facies is unique to the Stromatolitic Carbonate member. It forms an up to 40 m-thick buildup and can be traced along strike for hundreds of kilometres. There is a consistent vertical succession within the buildup from an assemblage of giant oncoids and branching columnar stromatolites at the base, to narrower, rarely branching, digitate stromatolites at the top. Oncoids are up to 20 cm in diameter, and are cored by broken pieces of stromatolite (stromaclasts). Some oncoids have digitate stromatolites branching from them (Fig. 8). Intercolumnar space is grainstone composed of stromaclasts, composite pisoids, and ooids.

Interpretation.--- This facies was deposited in a high-energy environment dominated by wave action that liberated coin-sized pieces of stromatolite that became nuclei for the growth of oncoids. Hydrodynamic conditions were energetic enough to keep large oncoids in motion. When oncoids became lodged, they served as a substrate for new stromatolite growth. That stromatolite laminations can be traced around oncoids leads to the inference that the microbial community that composed the oncoids was the same as that of the stromatolites. Comparable facies have yet to be described from the Shaler Supergroup, or from correlative rocks in the northern Cordillera. The depositional facies is interpreted as shallow subtidal, possibly in the wave-break zone.

2.5.2.13 Columnar stromatolite facies (F13)

Description.--- This facies occurs at scales varying from decimetre-thick biostromes to buildups over 10 metres thick. Stromatolite columns generally are 3-10 cm

wide and may be solitary or branching (Fig. 9). Stromatolites typically grow as vertical

Figure 9. Examples of variation in *columnar stromatolite facies* (F13). A) Columnar stromatolites with coarse intercolumnar sediment. B) Aerial view of an outcrop of exhumed stromatolite mounds in the Upper Carbonate member. Field of view is approximately 250 m; object in lower right is a helicopter skid. C) Small, domal bioherm of honeycomb-shaped, digitate, unbranching columnar stromatolites. Growth directions range from horizontal at the base to vertical at the top. D) Elongate stromatolitic reef (unbedded area on right side) in the upper carbonate member with flanking, tabular-bedded grainstones. Person circled for scale is approximately 1.5 m tall. E) Columnar stromatolites in the upper carbonate member, identified as form genus *Tungussia*? F) Columnar stromatolites in the stromatolitic carbonate member.

columns, although columns also grew laterally to sub-horizontally. Synoptic relief of individual layers is commonly a few millimetres up to approximately 1 cm, but stromatolite columns grew up to tens of centimeters high. Stromatolites most closely resemble the form genus *Tungussia*, *Acaciella*, *Baicalia*, *Jurusania* and *Tilemsina* (Walter 1972; Narbonne, et al. 2000; Kah, et al. 2009). Branching stromatolite forms are variably divergent, both vertically and horizontally. Stromatolites exhibit high morphological variability within this facies (Fig. 9). Bioherms range from tight geometric “honey-comb” shaped mounds with unbranching stromatolites that radiate out from a central core, to irregular petaloid branching forms. Buildups occur as metre-scale isolated mounds, patch reefs, and table reefs. Reefs can be very steep sided ($>45^\circ$). Reefs may have internal variation or be remarkably homogenous throughout. The lateral extent varies from local to regional, biostromes form resistant benches and thick (decametre) buildups that can be traced across the inlier for hundreds of kilometres. The intercolumnar material may be composed of calcilutite, but more commonly comprises intraclasts, stromaclasts, coated grains, oncoids and lithoclasts, with micritic to blocky calcite cement (Fig. 9).

Interpretation.--- This facies is interpreted to represent a shallow-water, subtidal, inner ramp environment containing stromatolite buildups. Localities with an upward succession of morphotypes from unbranching to branching columnar stromatolites are interpreted to identify a gradational change from a deeper to shallower water environment.

This facies is spatially associated with crossbedded intraclast grainstone facies (F10), and the coarse interstromatolite sediment suggests constant agitation, indicating

deposition above FWWB. Some buildups were successful in forming hydrodynamic barriers resulting in back-barrier lagoons, but most are encased in grainstones. Similar facies and stromatolite morphotypes have been described from the underlying Boot Inlet Formation (Narbonne et al., 2000) and from the correlative Little Dal Group in the northern Cordillera (Batten, et al. 2004).

2.5.2.14 Cross-bedded peloid grainstone facies (F14)

Description.--- This facies is associated with the domal stromatolite facies (F16 – see below) and is characterized by medium to thick, wedge-planar to tabular-planar cross-bedded peloidal intraclast grainstone. Low-angle crossbedding and planar stratification also are preserved. Adjacent sets of cross-beds have foresets with apparent opposed (bidirectional) transport (Fig. 8). Allochems are predominantly peloids, with subordinate micritic lithoclasts and rare fine-grained quartz sand grains.

Interpretation.--- Transverse sand waves superimposed on tidal bars resulted in wedge-shaped, dune cross-bedded grainstones. Planar stratification as a result of upper-flow-regime storm currents planed the tops of tidal cross-beds. Clough and Goldhammer (2000) described a similar facies from Neoproterozoic carbonate rocks in Alaska. They interpreted their facies as linear tidal sand bars. The wedge-shape geometry and bimodal-bipolar crossbeds supports an interpretation that this facies defines an inner ramp, linear tidal sand bar.

2.5.2.15 Dolomitic siltstone with desiccation cracks facies (F15)

Description.--- This facies occurs exclusively in the Stromatolitic Carbonate member and is located stratigraphically above and adjacent to facies F13. It consists of red, plane-laminated to thin-bedded siltstone. Green redox alteration occurs along fractures. Desiccation cracks are common and are filled with fine-grained, well-sorted, sub- to well-rounded quartz sand (Fig. 8). Thin interbeds of massive quartz arenite are preserved, along with thin mud-clast breccias.

Interpretation.--- The lack of tractional sedimentary structures and presence of regular laminations with desiccation cracks imply a quiet-water, protected environment that was subject to periodic subaerial exposure. Terrigenous sedimentation contributed red mud and silt to the lagoon. Desiccation cracks are filled with sand that was periodically blown in to the lagoon. Eolian sand is also preserved as thin interbeds in siltstone. This facies is interpreted to have been deposited in a protected setting, such as a shallow lagoon that developed behind a stromatolite reef (F12). This back-barrier system can be traced for more than 200 km along Minto Inlier.

2.5.2.16 Domal stromatolite facies (F16)

Description.--- Domal stromatolites occur mainly in the Upper Carbonate member forming laterally linked (Logan, 1964) or isolated mounds with generally low relief (< 1 metre) (Fig. 10). Forms are elongate, bun shaped, or undulatory. Isolated domal stromatolites are encased in crossbedded carbonate sand and parallel-laminated

lime mudstone. Isolated mounds have channels between them.

Figure 10. Examples of inner ramp facies. A) *Domal stromatolite facies* (F16). Bedding surface of laterally linked hemispheroid stromatolites. B) *Cross-bedded quartz-arenite facies* (F17). C) *Flaser-lenticular bedded, quartzose dolosiltite and dolarenite facies* (F18). D) *Laminated lime/dolomudstone facies* (F19). Scale card in B and D is 8 cm wide.

Interpretation.--- Elongation of domal stromatolites in some localities may identify influence from tidal currents and/or prevailing wave directions (Logan, et al. 1974). Domal stromatolites are interpreted as intertidal (Logan et al. 1964).

2.5.2.17 *Cross-bedded quartz-arenite facies (F17)*

Description.--- This facies consists of beige, fine- to medium-grained, well-rounded, well-sorted, quartz arenite. In thin section, grains are cemented by syntaxial quartz overgrowths. Strata are commonly thin- to medium-thick bedded with herringbone crossbedding. Crossbedding is predominantly planar-tabular but trough crossbeds are also preserved (Fig. 10). At one location, five paleocurrents measurements define a bimodal distribution of approximately 354° (n=2) and 102° (n=3).

Interpretation.--- Orthogonal paleocurrent directions are likely the result of competing wave and bottom currents. This facies may represent an interval where sedimentation rates were low and sand could be continually washed back and forth, rounding sand grains and winnowing mud. High textural and compositional maturity suggests either lengthy reworking, or derivation from eolian beach dunes. Herringbone crossbedding identifies reworking by tidal currents. Cross-bedded quartz-arenite facies is interpreted as intertidal to subtidal sand bars.

2.5.2.18 Flaser-lenticular bedded, quartzose dolosiltite and dolarenite (F18)

Description.--- Quartz-sandy dolarenite and dolosiltite are interbedded with dark grey silty shale and preserve flaser-lenticular bedding (Fig. 10). Sandstone beds commonly are topped by symmetric wave-ripple marks. Dolarenite is preserved in shallow (<1 m), flat-topped, concave-based, channels that are up to 10 m in lateral extent. Gutter casts, ripple crosslamination, and trough crossbedding are also preserved in this facies. Paleocurrents measured from crossbedding average 025° (n=2) and 230° (n=3), suggesting a bimodal-bipolar distribution.

Interpretation.--- Flaser-lenticular bedding has long been interpreted as tidal in origin (Reineck and Wunderlich 1968). Sand transported from tidally generated currents is ripple crosslaminated with opposing paleocurrents, and is coupled with mud flasers that settled out in ripple troughs during slack periods between tides. Channel sandstones represent storm deposits re-worked by tidal currents. Bimodal-bipolar paleocurrents and flaser-lenticular bedding support high-energy deposition on an intertidal flat.

2.5.2.19 Laminated lime/dolomudstone facies (F19)

Description.--- This facies consists predominantly of microbially, and less commonly traction-laminated dololutite and calcilutite. Sedimentary features include desiccation cracks, teepee structures, and intraclast breccia (Fig. 10). Desiccation cracks are filled with quartz sandstone, microbially laminated dolostone or mud intraclasts, and microspar cement. Some intervals of dololutite preserve randomly oriented, approximately 2 – 5 mm long, lathe-shaped molds that appear to be of similar shape to gypsum crystals. Black chert in the form of nodules or replacement along laminae is preserved in this facies. Well-rounded, frosted quartz sand grains occur in some beds. Microscopically, microbial mats have dissolution voids that are filled by dolomite cement and later replaced by ferroan calcite.

Interpretation.--- This facies represents deposition along a shallow intertidal to supratidal evaporative mudflat. Sediments were periodically flooded during storms and high tides to wet and maintain growth of microbial mats and transport carbonate mud. However, mud flats were also periodically exposed long enough to dry and form teepees

and desiccation cracks. Intraclast breccias represent storm rip-up deposits. Gypsum formed displacive intrasediment crystals prior to lithification and dissolved during diagenesis. Frosted quartz sand grains were blown onto the tidal flat by wind. Early diagenetic chert is common in Proterozoic tidal flats (Pratt 2010); the chert in this facies is black because it preserves organic material present in the original sediment.

Interpreted Depositional Environment (DE)	Facies associated with DE	Facies description	Composition	Facies Interpretation	Preferred stratigraphic order	Strat. Dist.	Comments	Water depth
DE1: Outer ramp	F1 - carbonaceous calcareous black shale and calcarenite	laminated, carbonate nodules, massive to normally graded, erosive or sharp lower contacts, loading and dewatering structures	carbonaceous calcareous black shale, green calcisiltite, calcareous claystone, thin beds of calcarenite	predominantly suspension settled sedimentation below SWB in outer slope, punctuated by calcarenite turbidites	interbedded F1, F2, F3	Restricted to UCM	Deep marine sedimentation, F4 records basin restriction	below SWB
	F2 - lithoclastic float/rudstone with dololite matrix	cobble-sized, sub-rounded, tabular clasts, imbricated to randomly aligned, erosive bases	dololite clasts, dololite matrix	slope-derived debris deposited along the outer-slope, below SWB turbidites				
	F3 - thin-bedded calcarenite	sharp, erosive bases, normally graded, parallel laminated, thin coarse grained lags, with quartz sand grains, lacks black shale	calcarenite					
	F4 - Mg-silicate and evaporite	slump folds, rip-up clasts, chaotic bedding, flat chip and edgewise breccias, wavy nodular compactional bedding, bottom grown crusts of gypsum replaced by dolomite, Mg-silicate (talc?) present	poorly indurated claystone, mudstone, and calcisiltite	basin restriction, subtidal below SWB				
DE2: Distal pro-delta	F5 - dark grey siltstone and silty shale	flute and gutter casts, ball-and-pillow, channel-and-fill, tool marks, ripple cross-lamination, HCS in quartz arenite interbeds	shale, siltstone, sandstone	pro-delta to delta slope, below FWB but above SWB	n/a	BSM	distal end of deltaic complex; proximal component is not preserved in Minto Inlier (see section 5.2.5)	below FWB, above SWB
DE3: Mid-ramp subtidal	F6 - black shale	paper thin to medium thick, parallel laminated	shale	suspension-settled sediment, below FWB	upward shallowing in the order F6, F8, F9	all members, except F7 is restricted to LCM	Commonly occurs at the base of upward shallowing intervals	below FWB, above SWB
	F7 - coniform stromatolitic dolostone	conical columnar stromatolites	dolomudstone	subtidal, below FWB				
	F8 - molar-tooth mudstone	pygmatically folded, sub-vertical, horizontal, vertical spindle, and tadpole shaped cracks	organic rich, silty, lime or dolomudstone	subtidal				
	F9 - molar-tooth clast packstone	graded with imbricated clasts	packstone, matrix is fine-grained detrital carbonate and microspar cement	subtidal tempestites				

Table 2. Depositional environments and facies of the Wynniatt Formation.

DE4: Inner-ramp subtidal	F6 - black shale	as above	as above	as above	F10 is commonly interbedded with or onlaps F13	all members, except F12 is restricted to the SCM	FA4 sediments were constantly subjected to wave reworking, and are interpreted as proximal to FA3	above FWWB
	F10 - cross-bedded intraclast grainstone	medium-thick beds, HCS, SCS, low-angle and trough crossbedding, scalloped erosional surfaces	as above	allochthems: reworked clasts of molartooth grains, peloids, stromatolitic fragments, and rounded lithoclasts	storm-transported, subtidal, above FWWB			
	F11 - lithoclast floatstone with grainstone matrix	medium thick beds with erosive bases, lithoclasts tabular, gravel-sized, subrounded, imbricated or randomly oriented, faintly graded	as above	floatstone clasts are lithoclasts, grainstone matrix allochthems consist of peloids, pisoids, and rare ooids,	storm lags on a shallow, grain-rich, subtidal flat			
	F12 - oncoïd-digitate stromatolite dolostone	oncoïds and branching columnar stromatolites to narrow, rarely branching digitate stromatolites. Oncoïds 20 cm, nucleated on stromatolites	intercolumnar space is grainstone with stromatolites, composite pisoids, and ooid allochthems	shallow subtidal, possibly in the wave-break zone				
	F13 - columnar stromatolite	decimeter-thick biostromes to reefal buildups, solitary and branching forms, variably divergent both vertically and horizontally; <i>Tungussia</i> , <i>Acaciella</i> , <i>Baicalia</i> , <i>Jurusania</i> , <i>Tilemsina</i>	intercolumnar material: calcilitite, intraclasts, stromatolites, coated grains, oncoïds, and lithoclasts; cement: micritic to blocky calcite	shallow-water, subtidal, inner-ramp stromatolite reefs				
DE5: Peritidal mud flat and lagoon	F14 - cross-bedded peloid grainstone	medium to thick, wedge-planar to tabular-planar, dune cross-bedded, low-angle cross-bedding, planar stratification	as above	allochthems are peloids, micritic lithoclasts, and fine-grained quartz sand grains	inner-ramp linear tidal sand bars	n/a	complex system of intertidal deposits with variable siliclastic input; back-barrier lagoons, tidal flats, tidal channels, sand shoals	above FWWB and periodically exposed
	F15 - dolomitic siltstone with dessication cracks	planar laminated, thin-bedded, red to green, dessication cracks	as above	siltstone	back-barrier shallow lagoon			

Table 2 cont'd.

DE5: Peritidal mud flat and lagoon (cont'd)	F16 - domal stromatolite	laterally linked or isolated mounds, elongate, bun-shaped, or undulatory forms, encased in cross-bedded carbonate sand and parallel laminated mud	carbonate sand and intertidal lime mudstone			
	F17 - cross-bedded quartz-arenite	thin- to medium-thick bedded, herringbone crossbedding, planar-tabular and trough cross-beds	off-white, fine- to medium-grained, well-sorted, quartz arenite	inter- to subtidal sand bars		
	F18 - flaser-lenticular bedded, quartzose dolosiltite and dolarenite	symmetric wave ripples, channels, gutter casts, ripple cross-lamination, trough cross-bedding	quartz-sandy dolarenite and dolosiltite, shale	tidal deposits in an intertidal mudflat		
	F19 - laminated lime/dolomudstone	microbially laminated, dessication cracks, teepee structures, gypsum molds, black chert, frosted quartz	lime/dolomudstone	inner-ramp, shallow inner- to supratidal evaporative mudflats		

Table 2 cont'd.

2.6 Interpretation

2.6.1 *Depositional Framework Interpretation*

The nineteen lithofacies described above collectively record deposition from five depositional environments (DE1 – DE5, Table 2, Fig. 11). These characterize deposition on a storm dominated, distally steepened ramp, which deepened to the northwest (present coordinates) (Rainbird et al., 1996). Ramp divisions discussed below are based on the classification of Burchette and Wright (1992), whereby the division between inner and mid-ramp is defined by fair-weather wave-base (FWWB), and between mid- and outer ramp by storm wave base (SWB) (Fig. 11). Defining criteria that led to the classification of a ramp depositional system for the Wynniatt Formation include the lateral continuity of facies for hundreds of kilometres and the absence of any facies that suggest build-up of an extensive, steep-sided, rimmed platform. In localized areas the columnar stromatolite facies (F13) formed buildups with relatively minor (<15 m) paleotopography (Fig. 10D), however the buildups are neither large, nor extensive enough to have formed a rimmed platform. Despite being an embayment of an epeiric sea, there must have been significant fetch to produce strong, storm-generated currents. The shallow inner ramp was also subject to constant wave agitation, as recorded by sedimentary structures such as wavy-lenticular bedding and symmetrical ripples (e.g. F18).

An outer ramp setting (DE1, Fig. 11) is characterized by facies consisting of re-sedimented coarse- and fine-grained lithologies (Table 2), in the Upper Carbonate member. This is marked by carbonate sediment introduced to the basin via gravity flow

deposits. The intraformational composition of clasts (Fig. 6) suggests that these deposits

Figure 11. Block diagram showing the generalised distribution of facies assemblages from inner to outer ramp. There is considerable heterogeneity in the inner ramp, with lagoon, supra- and intertidal mudflats, tidal creeks, and sand shoal depositional environments. Mid-ramp deposits include stromatolite buildups. The ramp is distally steepened, as indicated by the slope break on the outer ramp. The inset shows DE2, when the development of a delta drowned the carbonate factory with an influx of siliciclastic sediments.

incorporated only slope-derived sediment, supporting not only the presence of an increased gradient in a seaward position, but the initiation of gravity flows below that oceanographic break. Calcarenite turbidites (F3) stratigraphically overlie strata of lithoclastic float/rudstone facies (F2) in an upward-shallowing succession in the Upper Carbonate member.

A pro-delta depositional environment (DE2, Fig. 11) is proposed for the Black Shale member, which is characterized by siliciclastic shale and siltstone facies (Table 2,

F5). This might represent a carbonate ramp bypass system, or a period in which a siliciclastic system replaced the shelf carbonate factory (Fig. 11). During this time, the increased supply of siliciclastic sediment overwhelmed carbonate production. We interpret a river-fed delta environment as opposed to a muddy terrigenous shelf to provide a mechanism of transport of siliciclastic sediment. Sediment accumulated in a mid-ramp setting at water depths below FWFB but above SWB.

Mid-ramp deposits are preserved throughout the Wynniatt Formation in DE3 (Fig. 11), represented by lithofacies F6 to F9 (Table 2, Figs. 7 and 9). They are best preserved at the base of upward-shallowing intervals (Fig. 5) where they are overlain by inner ramp, crossbedded intraclast grainstones (F10). Grainstones require constant wave-motion for sediment winnowing and reworking, and therefore strata of facies F10 are interpreted to straddle the mid-to-inner ramp transition where sediments were consistently above FWFB but still closely associated with mid-ramp facies deposited below FWFB.

A regional inner ramp subtidal environment (DE4), represented by facies F6 and F10 to F13 (Table 2, Figs. 8 and 9) and a peritidal mudflat and lagoon environment (DE5), represented by facies F14 to F19 (Table 2, Figs. 8 and 10) constitute the principal inner ramp deposits of the Wynniatt Formation (Fig. 11). It is in the inner ramp where the regional development of columnar stromatolite buildups (F13) is best preserved (Figs. 9 and 11). The oncoid-bearing facies (F12) also forms a laterally extensive barrier system, and is a useful stratigraphic marker in the Stromatolitic Carbonate member. These facies represent a biophysical response to the offshore-nearshore oceanographic transition that characterizes a ramp because there is no seaward impediment to open ocean energy (James et al., 2010). The above intertidal and subtidal facies illustrate development of an

inner ramp, mixed siliciclastic-carbonate system, with evidence of wave and tidal current influence. The variety of facies illustrates a mosaic of depositional systems including back-barrier lagoons and supratidal to intertidal flats with tidal channels and sand shoals (Fig. 11).

2.6.2 Ramp Evolution of the Wynniatt Formation

Spatial (vertical, lateral) arrangement of facies in each of the stratigraphic members of the Wynniatt Formation demonstrates a punctuated evolution of carbonate ramp development. The lower Lower Carbonate member preserves an upward-shallowing succession of peritidal mudflat and lagoon (DE5) deposits. The upper Lower Carbonate member is an upward-deepening succession of mid-ramp deposits consisting of mid- and inner ramp subtidal (DE3 and DE4) paleoenvironments. The top of the Lower Carbonate member is marked by a biostrome of coniform stromatolitic dolostone facies (F7), signaling drowning of the ramp. Carbonate production was then suppressed by deposition of shale and siltstone facies of the Black Shale member, interpreted to have been deposited on a pro-delta slope (DE2).

The Black Shale member shallows upward, with a return to carbonate-dominated sedimentation that marks the base of the Stromatolitic Carbonate member. A subaerial unconformity is preserved near the base of the Stromatolitic Carbonate member (Fig. 4). It caps an interval of orange-weathered stromatolitic dolostone and dololomite, the surface of which is irregular, undercut and irregularly pitted suggesting there was karstification. Overlying this, a thin bed of quartz arenite is locally distributed and contains a lag of

pebbles and cobbles derived from the underlying stromatolitic dolostone and dololomite. This records a subsequent transgression leading to deposition of overlying inner ramp lagoonal deposits (facies F15, Table 2).

The initial inner ramp deposits are predominantly intertidal with evidence of desiccation (i.e., facies F15 and F18). The overlying and regionally distributed (~30-m-thick) oncoid-stromatolite facies (F12) records initial deepening. Above this marker bed are several cyclic intervals of upward-shallowing, subtidal to peritidal, mid- to inner ramp (DE3, DE4, and DE5) deposits (Figure 11; Table 2). The contact of the Stromatolitic Carbonate member and Upper Carbonate Member is a deeply (~75 cm) scoured unconformity that defines a major shift from inner to outer ramp deposition (DE1).

The lower ~50 m of the Upper Carbonate member consists of deep-water, black calcareous shale facies (F1), followed by an interval of Mg-silicate and evaporite facies (F4), which identifies a period of basin restriction. An overlying ~60 m succession of calcarenite turbidites (F3) signals a return to open ramp circulation and downslope transport. The remaining upper half of the Upper Carbonate member contains cyclical deposits of mid- to inner ramp DE3, DE4, and DE5 (Table 2).

The duration of carbonate ramp development preserved by the Wynniatt Formation is poorly constrained, but represents at least 50 million years (van Acken et al., 2013; Fig. 1). As a result, the two prominent intraformational unconformities probably separate three different ramp successions, where each break represents a significant hiatus and re-arrangement of ramp architecture. The lower ramp succession comprises the Lower Carbonate and Black Shale members. It defines a gentle slope that

preserves a transgression succeeding basin-central evaporites of the underlying Minto Inlet Formation, leading to development of pro-delta siltstone strata of the Black Shale member. The second ramp succession comprises most of the Stromatolitic Carbonate member, and preserves a series of upward-shallowing cycles of subtidal, mid- to inner ramp deposits (facies F8, F10, F13, and F19). The third ramp succession, represented by the Upper Carbonate member, preserves an outer distally steepened ramp that shallows upward to an inner ramp, and preserves the steepest and most variable slope gradient. While still technically a ramp, lithofacies in the outer ramp depositional environment (F2 and F3; Table 2) provide evidence that it is distally steepened. As well, there are steep-sided stromatolitic patch reefs that formed in a subtidal inner ramp setting (Fig. 9), which may be recording the early stages of transition to a rimmed platform.

2.6.3 Ramp Architecture and Basin Subsidence

The Amundsen Basin is envisaged as an embayment separated from the Mackenzie Basin by a promontory, the Great Bear Arch (Fig. 12; Young, 1977; Young, 1981; Rainbird et al., 1996). Paleogeographic data from strata in the Mackenzie Mountains indicate a southwesterly deepening basin marginal to the north-northeasterly curvilinear Ogilvie-Mackenzie platform (Aitken and Long, 1978; Turner and Long, 2008). In contrast, the platform faces north-northwest in the Amundsen Basin (Young, 1977; Rainbird et al., 1992). A northwest-protruding promontory, dividing, but not separating the Ogilvie-Mackenzie and Amundsen platforms, is supported by geophysical data (Aitken and Pugh, 1984; Aspler et al., 2003). This arch was the zero-edge of

deposition along the south-southwest oriented margin of the Amundsen Basin. The

Figure 12. Isopach maps of sediment thickness. Top map shows the paleogeography of northwest Laurentia during deposition of the Wynniatt Formation, modified from Rainbird et al. (1996). The grey area represents land, the blue is water. The estimated zero edge of deposition is shown along the approximate location of the Great Bear Arch; the thickest accumulation of sediment is in the southwest of the Minto Inlier, which thins towards the northeast, forming an interpreted concentric pattern. The bottom three inset maps show migration of the depocentre location during the three different ramp successions of the Wynniatt Formation, where accumulation was thickest in the southwest during deposition of the Lower Carbonate member (LCM) and Black Shale member (BSM) and Upper Carbonate member (UCM), and thickest in the northeast of Minto Inlier during deposition of the Stromatolitic Carbonate member (SCM). All of the dashed lines shown are interpretive, there is no direct field evidence to draw isolines as they are shown, however they are best-guess estimates given the geology and known thicknesses.

greatest accumulation of Wynniatt Formation strata is preserved in southwestern Minto Inlier (Fig. 12), with thinning toward the northeast. Thus, the depositional pattern of the Wynniatt Formation in the Amundsen Basin is interpreted as concentric, based on isopachs of unit thicknesses (Fig. 12), analogous in size and shape to Paleozoic intracratonic basins (Howell and van der Pluijm, 1990).

However, regional variations in the thicknesses of stratigraphic units (Fig. 12) indicate that the depocentre of the basin switched from the southwest to northeast and then back to southwest (Fig. 12). The combined thickness of the Lower Carbonate member and Black Shale member (i.e. lower ramp) in the southwest is 230 m, 2.5 times greater than its thickness in the northeast, where it is only 95 m. Opposing this trend, the Stromatolitic Carbonate member (i.e. middle ramp) doubles in thickness from 160 m in the southwest to 320 m in the NE. A return to a southwestern depocentre is indicated by the 380 m-thick section of Upper Carbonate member (i.e. upper ramp), which is 3.5 times greater than its thickness in the northeast (110 m; Fig. 12).

2.7 Discussion

2.7.1 Subsidence patterns during deposition of Wynniatt Formation

A comparison of subsidence history and ramp development in the Wynniatt Formation with the Paleozoic Michigan Basin (Howell and van der Pluijm, 1990) offers

useful insights for analysis of the Amundsen Basin. There are four distinct episodes of basin subsidence documented for the Michigan Basin that occur on a similar time scale to changes in the Wynnatt Formation, where changes in subsidence are: 1) early extension, 2) basin-centred subsidence, 3) eastward tilting towards the Appalachian orogen, and 4) return to basin central subsidence. Howell and van der Pluijm (1990) propose that these episodic changes in subsidence and basin reactivation in the Michigan Basin are likely related to Appalachian events and the resultant intraplate stress variations.

A similar paleotectonic arrangement places the Amundsen Basin ~3000 km northeast of the Grenville orogen of southeastern Laurentia. However, the youngest contractional activity in the Grenvillian Orogeny in Laurentia occurred at ~1000 - 980 Ma (Hynes and Rivers, 2010), approximately 200 m.y. before deposition of the Wynnatt Formation. Thus, changes in Amundsen Basin subsidence patterns during deposition of the Wynnatt Formation may have been controlled by other far-field stresses that occurred along other contemporaneous, and possibly conjugate margins to Laurentia. For example, during initial rifting of Rodinia, signaled by the emplacement of the Gairdner dyke swarm in Australia at ~827 Ma (Wingate et al., 1998), and the Bikou basalts in South China between 820-810 Ma (Wang et al., 2008). Both events occur on cratons that, in many reconstructions, are adjacent to northwest Laurentia (e.g., Li et al., 2008; Li and Evans, 2011). There is evidence of contemporaneous extension in correlative strata of the Fifteenmile Group (Macdonald et al., 2012) and Mackenzie Mountains Supergroup (Turner and Long, 2008) in the northern cordillera, >1000 km to the southwest (Fig. 12). Although basin subsidence in the Paleozoic Michigan Basin is related to collisional events as opposed to what were probably extensional tectonic events in the

Neoproterozoic Amundsen Basin, there are similar patterns that are repeated in terms of episodic subsidence, migration of depocentres, and ramp dynamics between these different basins.

2.7.2 Comparison of sedimentary patterns

Our work provides the first detailed facies and stratigraphic analysis of the Wynniatt Formation, thereby allowing evaluation of intra- and interbasinal comparisons. In the lower Shaler Supergroup of the Minto Inlier, the Boot Inlet Formation (Fig. 1) is a ~500 m thick carbonate unit, which is also interpreted to have accumulated on a storm-dominated carbonate ramp (Young and Jefferson, 1975; Young and Long, 1977; Morin and Rainbird, 1993; Rainbird et al., 1994; Narbonne et al., 2000). There are, however, two prominent differences between the formations. First, unlike stromatolitic buildups within the Wynniatt Formation, which typically formed tabular (biostromal) bodies that can be correlated across Minto Inlier, the Boot Inlet Formation preserves a variety of more isolated (biohermal) reef morphotypes, which include: patch reefs, table reefs, and large concentric-sheet bioherms that are up to 60 m in diameter (Narbonne et al., 2000). Second, ooids are abundant in the Boot Inlet Formation compared to the scarcity of these allochems in the Wynniatt Formation. There was obviously some important difference in either the paleoclimate and/or seawater chemistry (Lloyd et al., 1987) that favoured ooid production during Boot Inlet time.

Stromatolite types of the Boot Inlet and Wynniatt formations are similar, comprising *Tungussia*, *Baicalia*, and *Jurusania*; however *Acaciella* is present only in the Wynniatt Formation. Interestingly, *Acaciella* has been documented from the correlative Neoproterozoic Amadeus Basin of Australia (Bell and Jefferson, 1987; Rainbird et al., 1996; Hill et al., 2000; Planavsky and Grey, 2008). Specifically, the *Acaciella australica* stromatolite assemblage (Hill et al., 2000) occurs in the Loves Creek Member of the Bitter Springs Formation, coinciding with a negative carbon isotope excursion (Hill and Walter, 2000; Walter et al., 2000; Swanson-Hysell et al., 2010). This negative excursion has also been reported from the Wynniatt Formation (Jones et al., 2010; Thomson et al., 2012) and has been used for global stratigraphic correlation (Halverson et al., 2007a; Halverson et al., 2010). It is possible that the *A. australica* assemblage is temporally restricted in the Shaler Supergroup, although this needs to be substantiated.

The striking similarities documented between the Shaler and Mackenzie Mountains supergroups have made for robust, long-standing litho- and sequence stratigraphic correlations (Young et al., 1979; Rainbird et al., 1996; Long et al., 2008). Within that framework, the Wynniatt Formation is correlative to the Snail Spring and Ram Head formations (Turner and Long, 2012). The Snail Spring Formation is equivalent to the Lower Carbonate and Black Shale members, and like in those members, represents shallow-water deposition and shut-down of the carbonate factory, followed by transgression (Aitken, 1981; Turner and Long, 2008). The Ram Head Formation records upward-shallowing intervals of grainstone, molartooth mudstone, and cryptalgal laminite (Aitken, 1981), analogous to the upward-shallowing successions observed in the Stromatolitic Carbonate member.

Samples containing *Tawuia* and *Chuaria* specimens were identified by Hofmann and Rainbird (1995) from northeastern Minto Inlier, near Kilian Lake (Fig. 1). The stratigraphic interval that contains these fossil occurrences is described as “just below and above the contact zone between a distinctive rusty black shale member and the overlying grey dolostone member of the Wynniatt Formation” (Hofmann and Rainbird, 1995, pg. 723). This is also the same stratigraphic interval from which a diverse suite of acritarchs and other organic-walled and filamentous microfossils was described (Butterfield and Rainbird, 1998; Butterfield, 2005). We correlate this stratigraphic interval to the contact between the Black Shale member and the Stromatolitic Carbonate member (Fig. 4). In the Wynniatt Formation, this occurrence of the *Tawuia-Chuaria* assemblage zone is preserved within a peritidal mudflat and lagoon depositional environment (DE5, Table 2). This supports the original interpretation of Hofmann and Rainbird (1995), that these fossils were preserved in intertidal sedimentary rocks that were periodically subaerially exposed and influenced by storms. The Black Shale member is correlative with the Snail Spring Formation (Rainbird et al., 1996; Turner and Long, 2012), where the *Tawuia-Chuaria* assemblage has also been recovered (Hofmann, 1985). *Tawuia* is also documented from lower in the Mackenzie Mountains Supergroup, in deep-water rhythmites of the Stone Knife Formation (Hofmann and Aitken, 1979; Turner and Long, 2012). Our work improves the depositional and stratigraphic context for this assemblage zone allowing for future greater interbasinal correlations, and improved understanding of paleogeographic distribution of depositional environments.

2.8 Conclusions

This contribution presents new sedimentological descriptions and interpretations for a hitherto undescribed stratigraphic unit. Coupled with spatial organization of depositional environments and their component lithofacies, as well as two prominent, regional unconformities, the Wynniatt Formation represents a succession of three high-energy, distally steepened carbonate ramps. Sedimentation was predominantly controlled by storm-generated processes, with minor reworking by wave and tidal currents. Isopach maps based on sediment thicknesses provide support that the Wynniatt Formation accumulated in a circular basin. Inferred changes in the position of regional depocentres are associated with each stage of ramp development. This is likely the result of far-field extensional stresses, given that the age of the formation overlaps with initiation of breakup of the Rodinia supercontinent, and that extension is manifested as fault-related sedimentation in coeval basins located closer to the paleomargin of Laurentia. Our work provides sedimentary and stratigraphic contexts for intra- and interbasinal correlations; this includes an improved understanding of basin and carbonate-ramp development in the Amundsen Basin, and an improved basis for interbasinal correlation among coeval Neoproterozoic basins.

CHAPTER 3: TRANSGRESSIVE-REGRESSIVE CYCLES AND TECTONOSTRATIGRAPHY OF THE UPPER SHALER SUPERGROUP

Thomson, D., Rainbird, R. H., Krapez, B. T-R cycles and tectonostratigraphy of the Neoproterozoic (Tonian-Cryogenian) Amundsen Basin prior to supercontinent (Rodinia) Breakup. In review, re-submitted to GSA Bulletin

3.1 Introduction

The Neoproterozoic was a dynamic era in Earth history, during which major changes in the isotopic composition of seawater (e.g. Halverson et al., 2007a), the evolution of early animals (e.g. Morris, 1993), low latitude glaciations (e.g. Kirschvink, 1992), and the formation and breakup of the supercontinent Rodinia occurred (e.g. Hoffman, 1991). Recognition of a late Mesoproterozoic supercontinent evolved from the correlation of sedimentary successions between northwestern Canada and Australia (Eisbacher, 1985; Bell and Jefferson, 1987); and reconfiguration of remnant Grenville terranes (1080 – 985 Ma) preserved on several continents (Moores, 1991; Gower and Krogh, 2002; Davidson, 2008). In most reconstructions of Rodinia, Laurentia is interpreted to have formed the core of the supercontinent (e.g. Hoffman 1991; Li et al. 2008). The Amundsen and Mackenzie basins are considered to have been embayments of an extensive epeiric sea that covered northwestern Laurentia during the time between amalgamation and break-up of Rodinia (Fig. 13; Young, 1981; Rainbird et al., 1996). There remains debate about the paleogeographic fit of the continental blocks that surrounded Laurentia, with numerous different models proposed (e.g., Hoffman, 1991;

al., 2012), Mackenzie Mountains, Northwest Territories (Turner and Long, 2008), and the Amundsen Basin, Northwest Territories (Rainbird et al., 1996) invoked different models for pre-720 Ma extension in a northwest-southeast orientation. Turner and Long (2008) attributed extension to simple shear, lower-plate upper-plate detachment along transfer faults for the Mackenzie Mountains Supergroup, whereas Macdonald et al. (2012) interpreted normal faults related to extension and to thermal decay of a mantle plume for the Fifteenmile Group. The interpretation of Macdonald et al. (2012) corroborates the interpretation of an intracontinental sag basin for the Shaler Supergroup (Rainbird et al., 1996).

In this paper, we identify several transgressive-regressive (T-R) cycles, which help to provide a tectonostratigraphic framework for a succession in the Amundsen Basin, Northwest Territories (Canada) that corresponds to deposition ca. 900-780 Ma, prior to the breakup of Rodinia. Stratigraphic cycles record global to intrabasinal controls, and can be used to correlate and discriminate among coeval stratigraphic sections in now geographically disparate basins. It is well established that the Shaler Supergroup of the Amundsen Basin records deposition in terrestrial and rhythmically alternating restricted to open-marine conditions (Young, 1981; Rainbird et al., 1996), but stratigraphic division that would better document the causes and responses of cyclic paleoenvironmental changes has previously been attempted only for the Kilian and Kuujjua formations (Rainbird, 1993) and Boot Inlet Formation (Narbonne et al., 2000; Fig. 1).

Pre-breakup successions of intracontinental sag basins have stratigraphic architectures that differ from those of fault-bound rift basins and of passive margins

where the geometry and architecture of thick prograding sequences can more easily be recognized (Lindsay et al., 1993; Vescei and Durringer, 2003). Sedimentation rates in the Amundsen Basin generally kept pace with, or were briefly outmatched by subsidence, and there is little evidence for large-scale progradation, thereby resulting in a broad layer-cake stratigraphy. Correspondingly, shallow-water facies predominate and were deposited on gently sloping ramps with little, if any, fault-bound compartmentalization. Assignment of subsidence mechanism, or the recognition that such cycles relate to pre-rift extension, is difficult. Despite the Amundsen Basin being a broad (~300 km), shallow-water basin in which most of the sediments were deposited above storm wave-base (SWB), it is possible to identify cycles and interpret subsidence patterns prior to break-up of Rodinia, which has implications for understanding stratigraphic architecture of other intracontinental sag basins.

3.2 Regional Geology

The Neoproterozoic (Tonian-Cryogenian) Shaler Supergroup is over 4 km thick in the Amundsen Basin (Fig. 1), comprising in ascending order the Rae Group, Reynolds Point Group, and the ungrouped Minto Inlet, Wynniatt, Kilian, and Kuujjua formations. The sedimentary succession is intruded by diabase sills and dykes that fed the Natkusiak flood basalts, part of the ~720 Ma Franklin Large Igneous Province (LIP) that was related to the rifting and break-up of Rodinia. Depositional ages of sedimentary units in the Shaler Supergroup are constrained by detrital zircon geochronology of sandstones and

Re-Os geochronology of organic-rich black shale (Fig. 1), but no tuffs have been discovered for precise U-Pb dating.

The Shaler Supergroup is exposed in several erosional windows throughout northwestern Canada; namely the Minto, Cape Lambton, Duke of York, Brock, and Coppermine inliers. Our study is of the upper Shaler Supergroup in the Minto Inlier, which exposes a broad northeast-southwest trending, gently dipping, anticline-syncline pair on northwestern Victoria Island, Northwest Territories, Canada (Fig. 1).

Proterozoic stratigraphic successions of northwestern Canada were divided into three unconformity-bounded sequences, termed A, B, and C by Young et al. (1979). The Shaler Supergroup and correlative rocks in the Mackenzie Mountains (Mackenzie Mountain Supergroup) and Ogilvie Mountains (Fifteenmile Group), constitute Sequence B (Fig. 14), deposition of which postdates the end of the Grenvillian orogeny at ~1000 Ma and predates breakup of Rodinia at ~720 Ma in northwestern Laurentia (Heaman et al., 1992; Macdonald et al., 2010). There are long-standing lithostratigraphic and sequence stratigraphic correlations among Sequence B strata (e.g., Young et al., 1979; Rainbird et al., 1996; Long et al., 2008). In particular, Long et al. (2008) subdivided Sequence B into five sub-sequences (i.e. sB1 to sB5). In that sequence stratigraphic framework, the Jago Bay Formation belongs to sB3 and the Minto Inlet and Wynniatt formations fall within sub-sequence sB4, which correlate to the Little Dal Group of the Mackenzie Mountains and the upper Fifteenmile Group in the Ogilvie Mountains (Macdonald et al. 2012). Macdonald et al. (2012) documented eight sequences within the

Fifteenmile Group (Fig. 14), correlating S5 with the Fort Collinson and Jago Bay

Figure 14. Correlation of Fifteenmile Group, Mackenzie Mountains Supergroup, and Shaler Supergroup stratigraphy, with sequences indicated. References for stratigraphy and U-Pb dates provided on the figure underneath stratigraphic columns. Tripartite Sequence A, B, and C division modified from Young (1981). Sub-sequences sB1-5 after Long et al. (2008). Higher order sequences for Fifteenmile Group after Macdonald et al. (2012); Kilian and Kuujjuua formations after Rainbird (1993). The grey area represents stratigraphy that we consider correlative, with black dashed lines showing approximate correlations of cycles.

formations and the upper three (S6, S7 and S8) with the Minto Inlet and Wynniatt formations on the basis of carbon isotope stratigraphy.

Sequence stratigraphy of the Kilian and Kuujjua formations was documented by Rainbird (1993), in which four transgressive-regressive (emergent-submergent) cycles were interpreted. The Kilian and Kuujjua formations display northeastward stratigraphic thinning, erosional beveling, and evidence for block faulting, consistent with pre-eruptive crustal doming and mantle-plume generated asthenospheric upwelling related to the emplacement of the Franklin LIP.

3.3 Methodology

Fieldwork was undertaken in the summers of 2010 and 2011 with stratigraphic analysis focused in two areas with well-exposed stratigraphic sections (Fig. 1). Twenty outcrop sections and two drill cores were measured totaling 23.3 km (see Appendix 1 for coordinates). Sedimentary descriptions described below incorporate previous work.

3.4 Lithostratigraphy and sedimentology

3.4.1 Jago Bay Formation

The Jago Bay Formation, the uppermost unit of the Reynolds Point Group, conformably overlies sandstones of the Fort Collinson Formation (Fig. 1), which is composed of cross-bedded quartz arenite and quartzose carbonate arenite interpreted to

be of marine-reworked fluvial deposits (Rainbird et al., 1994). The contact between the Fort Collinson and Jago Bay formations is gradational within an upward deepening succession between fluvial to marginal-marine deposits (Fort Collinson Formation) and marine carbonate rocks (Jago Bay Formation), which in turn shallow upward into a gradational contact with the overlying evaporitic Minto Inlet Formation. The thickness of the Jago Bay Formation varies from <65 m in the northeast Minto Inlier to >200 m in the southwest Minto Inlier (Young et al., 1979), consistent with overall thickening of the upper Shaler Supergroup to the southwest (Thomson et al., 2014; Long et al., 2008). Rainbird et al. (1994) described it as an upward-shallowing succession of carbonate-rich sediments, from intertidal-lagoonal to supratidal and restricted intertidal. Where we observed the lower Jago Bay Formation in southwest Minto Inlier, it is composed of carbonaceous calcilutite with black shale interbeds and molar-tooth structure, supporting a lagoonal depositional environment (cf. James et al., 1998).

3.4.2 Minto Inlet Formation

The contact between the Jago Bay and Minto Inlet formations is not exposed, and is intruded by a diabase sill at the formational stratigraphic boundary in drill core, but appears to represent a gradational transition from supratidal and restricted intertidal carbonate rocks of the upper Jago Bay Formation (Rainbird et al., 1994) to supratidal maroon, gypsiferous carbonate mudstones that characterize the lower Minto Inlet

Formation (Fig. 15). The Minto Inlet Formation ranges in thickness between 260 m and

Figure 15. Measured stratigraphic sections of the Jago Bay and Minto Inlet formations. Interpretations of T-R cycles are shown. Inset map shows locations of stratigraphic sections.

400 m (Rainbird et al., 1996), and comprises rhythmically alternating bedded gypsum, gypsiferous shale, and carbonate. An informally defined lower evaporitic unit comprises interbedded red gypsiferous and calcareous siltstones, and thinly bedded gypsum and anhydrite. Ripple cross-lamination, chickenwire texture, satinspar veinlets, selenite crystals, and nodular anhydrite are preserved, consistent with a supratidal mudflat environment (cf. Kendall, 2010). An overlying lower carbonate unit is characterized by stromatolites, hummocky cross-stratified (HCS) calcarenite, and molar-tooth structure in silty, carbonaceous lime mudstone. This lower carbonate unit is interpreted as a basin-freshening event with deposition occurring in a shallow subtidal environment. Bedded gypsum, chickenwire anhydrite, and satinspar veining characterize a middle evaporitic unit, and are interpreted as subaqueous evaporite deposition where gypsum was precipitated at the sediment-air interface producing cumulate gypsite deposits (Kendall, 2010). The upper evaporitic unit contains rhythmically alternating limestone, interbedded limestone-siltstone-gypsum, and nodular gypsum. This upper evaporitic unit is interpreted to record alternating basin freshening and restriction in a shallow subtidal to supratidal depositional environment.

3.4.3 Wynniatt Formation

The contact between the Minto Inlet and Wynniatt formations is not exposed in the Minto Inlier, but is interpreted to be gradational, based on facies stacking patterns.

The thickness of the Wynniatt Formation in Minto Inlier varies from ~1000 meters in the

Figure 16. Measured stratigraphic sections of the Wynniatt Formation and members, with T-R cycles, parasequences, and important stratigraphic surfaces indicated. Refer to Figure 16 for a map of stratigraphic section locations.

SW to ~480 meters in the NE (Young, 1981; Rainbird et al., 2012; Thomson et al., 2014).

The Wynniatt Formation is divided into four informal members (Fig. 16), 1) Lower Carbonate member, an upward-deepening succession of supra- to sub-tidal carbonate rocks; 2) Black Shale member, a recessive interval of dark-grey siltstone and silty shale deposited in a pro-delta setting; 3) Stromatolitic Carbonate member, comprising stacked upward-shallowing cycles of subtidal to supratidal carbonate rocks and, 4) Upper Carbonate member, an upward-shallowing succession from subtidal black calcareous shale to peritidal cross-bedded intraclast grainstone and stromatolitic limestone. The Wynniatt Formation is interpreted to have been deposited along a distally steepened carbonate ramp, dominated by storm-generated depositional processes, with minor reworking by wave- and tidal currents (Thomson et al., 2014).

3.5 T-R cycles

A transgressive -regressive (T-R) cycle is defined as a complete cycle between one deepening event and the start of the next event of the same scale, whereby one T-R cycle may be composed of several smaller cycles or, in turn, may be a part of a larger-scale cycle (Johnson et al., 1985). Best used to document and characterize basin-scale changes in stratigraphy (Embry, 1993), a T-R cycle is recognized by changes in facies stacking patterns that record upward-deepening and upward-shallowing successions, the change in which may be marked by a subaerial unconformity, regressive surface of marine erosion, shoreline ravinement surface, or flooding surface (cf. Embry, 2009). This method is a useful approach to understanding basin-fill patterns for a data set that is

based on outcrop studies, in particular when rocks are poorly constrained in terms of geochronology, as is the case for the Amundsen Basin. Therefore, we follow stratigraphic models of Embry (1993; 2009), and interpret several T-R cycles that are nested within the previously defined sequence stratigraphic framework of Long et al. (2008) and Young et al. (1979), discussed above.

Intracontinental basins typically are shallow-water, long-lived depositories characterized by slow subsidence rates (e.g. Xie and Heller, 2009). Sedimentation rates generally are lower compared to other types of basins, and sequences can be traced laterally for long distances (Vecsei and Düringer, 2003). These shallow-marine sag basins present particular problems for documentation of cycles because base-level changes usually are subtle and their impact on facies tracts have wide lateral extent. Also, with poorly defined or very gentle slopes, changes are difficult to infer. Ramp, rather than shelf-slope geometries dominate, although some sag basins lack obvious seaward paleobathymetric changes. Documentation of accommodation cycles, and interpretation of their causes, is therefore a challenge. A sag basin generally exhibits such subtle stratigraphic changes that it is important to document and recognize the larger significance of what may be seemingly insignificant surfaces, within the context of internal facies stacking patterns.

3.5.1 Sequence sB3

Sequence sB3 in the Amundsen Basin comprises the Boot Inlet, Fort Collinson, and Jago Bay formations (Long et al., 2008). Within sB3, we document one T-R cycle (T-R1), described below.

3.5.1.1 T-R1

The transition from emergent, re-worked fluvial deposits of the Fort Collinson Formation, through submergent peritidal and lagoonal carbonate rocks of the Jago Bay Formation, and back to emergent supratidal mudflat deposits at the base of the Minto Inlet Formation represents a transgressive-regressive cycle that we define as T-R1 (Fig. 15). The contact between the Jago Bay and Minto Inlet formations (that is, the sB3-sB4 boundary), is not exposed, and in drill core is obscured by a diabase intrusion.

3.5.2 Sequence sB4

As defined by Long et al. (2008), sequence sB4 is composed of the Minto Inlet and Wynniatt formations in the Amundsen Basin. Within sB4, we document four T-R cycles (T-R2 to T-R5), described below.

3.5.2.1 T-R2

T-R2 begins at the base of the Minto Inlet Formation, and extends upward through the base of the Lower Carbonate member of the Wynniatt Formation (Fig. 16). The contact between the Minto Inlet and Wynniatt formations is not exposed in the study area, but it appears to be gradational with facies variations recording an upward-shallowing trend. The lower Minto Inlet Formation comprises maroon, gypsiferous lime mudstones that we interpret as shallow-water, marginal-marine deposits. These pass upward into subtidal carbonate rocks that suggest deposition in deeper water, open-marine conditions (Fig. 15). The upper Minto Inlet Formation is composed of basin-central, sulphate crystal cumulate deposits (Kendall, 2010) that shallow upward into

cyclically interbedded carbonate and evaporite rocks. The basal section of the overlying Lower Carbonate member of the Wynniatt Formation comprises shallow-water carbonate rocks with features that include mud-cracks, tepee structures, and halite pseudomorphs, thereby defining emergent conditions.

3.5.2.2 *T-R3*

The boundary between T-R2 and T-R3 is an erosion surface about 45 m above the base of the Lower Carbonate member in the southwest Minto Inlier (Fig. 16). This surface can be traced ~160 km across the Minto Inlier to a section measured in the northeast (Fig. 17), however the surface there is not as well developed. Below that surface, the uppermost carbonate rocks of T-R2 relate to base-level low during which large-parts of the platform must have been periodically exposed and eroded. Above the surface, carbonate floatstone comprising clasts of peloidal floatstone and dololutite in a grainstone matrix (Fig. 17), is interpreted to record subaerial erosion of the platform, and is defined as the basal unit of T-R3. Facies stacking above the floatstone unit is grossly upward-deepening (see below), such that the erosion surface fits the definition of a shoreline ravinement (cf. Embry, 2009), that is, an abrupt, scoured contact whose overlying sediments fine upward as a consequence of water deepening.

The remainder of the Lower Carbonate member, above the shoreline ravinement, and the Black Shale member together preserve an upward-deepening facies trend (Figs. 16 & 17). Individual parasequences are upward deepening, commonly from columnar stromatolites to cross-bedded intraclast grainstones that are overlain by black shale. Collectively, parasequence stacking is also upward-deepening as the proportion of shale

increases up-section. A maximum flooding surface (MFS) is interpreted to occur

Figure 17. (previous page) Southwest to northeast cross-section of the Wynniatt Formation, showing regional T-R cycle correlations. LCM – Lower Carbonate member, BSM – Black Shale member, SCM – Stromatolitic Carbonate member, UCM – Upper Carbonate member.

approximately twenty meters above the base of the Black Shale member, at the position where there is no longer any carbonate and the shale is extremely silt-poor (Fig. 16). The MFS therefore records complete submergence of the platform below fair-weather wave base (FWWB), as hemipelagic sedimentation overwhelmed carbonate production.

Above the MFS in the Black Shale member, sedimentation units coarsen upward and shallow from an interpreted distal pro-delta depositional environment to a peritidal flat and lagoonal depositional environment (Thomson et al., 2014). This upward-shallowing trend defines the onset of the regressive phase of T-R3, as sediment flux offset generation of new accommodation.

3.5.2.3 *T-R4*

The boundary between Sequences T-R3 and T-R4 occurs near the base of the Stromatolitic Carbonate member and is preserved across Minto Inlier, being recognized in all sections where this sequence is exposed (Fig. 17). It is a surface characterized by karstification features such as irregular pitting and undercutting of the underlying orange-weathered stromatolitic dolostone and dololutite (Fig. 18), thereby identifying it as a subaerially cut unconformity. Overlying the unconformity is a thin bed of quartz arenite that locally contains a lag of pebbles and cobbles derived from the underlying dolostone. The quartz arenite is overlain by less than a meter of dolomitic siltstone with desiccation cracks, consistent with a shallow-water marine environment and drowning of the karstic

surface.

Figure 18. Photographs of sequence boundaries and T-R cycles in outcrop. (A) Shoreline ravinement (SR) and boundary surface between T-R2 and T-R3 in the Lower Carbonate member. (B) Subaerial unconformity (SU) and boundary between T-R3 and T-R4 in Stromatolitic Carbonate member. (C) Maximum surface of regression (MRS) and boundary between T-R4 and T-R5 at the contact between the Stromatolitic Carbonate member and Upper Carbonate member. (D) Cycles composed of sharp-based, orange weathering, thin-bedded dolostone grading upward into laminated black shale interpreted as upward-deepening parasequences (green arrows) near the top of Lower Carbonate member. (E) Maximum flooding surface (MFS) above the T-R4 and T-R5 contact. The boundary between the two is marked by a maximum surface of regression (MRS) that coincides with the Stromatolitic Carbonate member and Upper Carbonate member contact. (F) Excavated contact between the Wynniatt Formation and overlying Kilian Formation, which forms the sB4 – sB5 (cf. Long et al., 2008) boundary.

A set of upward-shallowing parasequences makes up the remainder of T-R4 (Fig.

16). Facies assemblages shallow upward from mid-ramp subtidal deposits, typically

composed of interbedded molar-tooth mudstone and molar-tooth microspar packstone, to inner-ramp subtidal deposits, culminating at the top of the parasequence set with peritidal flat and lagoonal deposits, most commonly preserved as dolomitic siltstone with dessication cracks. The second parasequence identified in T-R4 (Fig. 16) is more than twice as thick (>50 metres thick) as the other parasequences in T-R4, indicating that either there was a change in local subsidence rates or the identification of more parasequences is masked by the nature of the deeper depositional environment.

3.5.2.4 *T-R5*

There is a deeply scoured surface between the Stromatolitic Carbonate member and the overlying Upper Carbonate member. The surface separates, shallow-marine carbonate rocks at the top of an upward-shallowing parasequence set, from overlying, deep-water, black, nodular, calcareous shale at the base of an upward-deepening parasequence set. The Stromatolitic Carbonate member - Upper Carbonate member boundary is therefore interpreted to be a surface of maximum regression (MRS; Fig. 16). As a MRS marks the onset of transgression (Embry, 2009), the surface is also interpreted as the boundary between cycles T-R4 and T-R5.

The base of T-R5 is overlain by thin beds of limestone-pebble conglomerate that grade upward into calcareous shale, which in turn pass upward into black carbonaceous shale (Figs. 16-18). The package is interpreted to represent the transgression, overlying which a MFS marks the change from upward-deepening to upward-shallowing facies, indicating the subsequent regression.

Above the MFS, a thick (~160 m) upward-shallowing succession begins with carbonaceous, calcareous shale that pass upward into thinly bedded calcarenite turbidites. The depositional site is interpreted to have been outer-ramp. Above this succession is a parasequence set that alternates between inner-ramp deposits of interbedded cross-bedded intraclast grainstones, columnar stromatolites, and laminated dolomudstones (Fig. 16).

Stromatolitic carbonate rocks at the top of the Upper Carbonate member are sharply overlain by ~3 m of carbonaceous shale that grade into nodular gypsum; these evaporite rocks mark the base of the Kilian Formation (Fig. 17). A hiatal surface is interpreted on the top of Upper Carbonate member, above the carbonate rocks and at the base of the shale, corresponding to a switch from open-marine carbonate rocks to a restricted, sabkha environment. An unconformity separates the Upper Carbonate member from the overlying Kilian Formation (Figs. 16 & 17), but outcrops are not laterally extensive, and there is no obvious evidence for subaerial exposure or stratal discordance. A hiatus across this boundary is interpreted to account for the abrupt change from open-marine carbonate rocks to sabkha evaporite rocks. Other changes that coincide with this interval include variations in sulfur and carbonate isotopes (Jones et al., 2010; Prince, 2014, Thomson et al., in review), and a switch to a subsidence regime interpreted to have been influenced by thermal doming in the overlying middle Kilian Formation (Rainbird, 1993). Long et al. (2008) placed the sB4-sB5 sequence boundary at the boundary between the Wynniatt and Kilian formations and that interpretation is followed here.

3.6 Correlation with adjacent basins

Neither sequence nor T-R stratigraphy has been documented for the Little Dal Group in the Mackenzie Mountain Supergroup. Sequences defined for the Fifteenmile Group of the Ogilvie Mountains are difficult to correlate with the upper Shaler Supergroup because deposition of the the Fifteenmile Group was strongly influenced by local tectonism (Macdonald et al., 2012). Nevertheless, sequences within the Fifteenmile Group have been correlated across NW Laurentia using carbon isotope stratigraphy (Macdonald et al., 2012). Sequence S6 of the Fifteenmile Group in the Ogilvie Mountains is interpreted to be significantly condensed compared to sequences in the Shaler and Mackenzie Mountains supergroups. S6 is correlated with the combined Minto Inlet Formation and Lower Carbonate member, Black Shale member and Stromatolitic Carbonate member of the Wynniatt Formation in Minto Inlier, and the Ten Stone, Snail Spring, and base of the Ram Head formations in the Mackenzie Mountains Supergroup (Macdonald et al., 2012; Turner and Long 2012; Fig. 14). S6 currently is interpreted to record a steady increase in $\delta^{13}\text{C}$ (VPDB) from +2‰ to +10‰ in the lower half of the sequence in the Fifteenmile Group, whereas in the Mackenzie Mountains the opposite trend is preserved, with a steady decline from +6‰ to 0‰ for the lower half of the interpreted correlative sequence (Halverson, 2006; Macdonald et al., 2012). In the Shaler Supergroup, C isotopes show a steady decline from ~4‰ to -1‰ over the same correlated interval (Jones et al., 2010; Macdonald et al., 2012; Thomson et al., in review). We suggest that the chemostratigraphic correlations are tenuous, and that correlation of sequences from the Fifteenmile Group need to be re-evaluated because of poor chemostratigraphic correlation.

Sequence S7 in the Fifteenmile Basin (Macdonald et al., 2012) correlates well with T-R5 in the Amundsen Basin and the Ramhead Formation in the Mackenzie Basin, and preserves a negative carbon isotope excursion in all three basins (Jones et al., 2010; Halverson et al., 2012; Macdonald et al., 2012, Thomson et al., in review). This sequence represents a basin-deepening event that is present across NW Laurentia, so sea-level rise was likely eustatic; although attendant tectonically driven subsidence in each basin probably exaggerated the sediment thickness, particularly in the Reefal assemblage of the Fifteenmile Group. This is an important correlation because the base of this sequence has been dated at 811.5 Ma from a tuff bed in the Reefal assemblage (U-Pb zircon; Macdonald et al., 2010), and because it contains a pronounced negative carbon isotope excursion that can be correlated with the global Bitter Springs Stage (Jones et al., 2010; Macdonald et al., 2012).

Sedimentation rates for the Fifteenmile Group are interpreted to have been high due to rapid generation of accommodation by extensional faulting that also was responsible for abrupt thickness and facies changes (Macdonald et al., 2012). Gravity-driven slope deposits, such as olistoliths and talus breccias, as well as prograding stromatolite reef complexes, dominate lithofacies of the Reefal assemblage and Craggy dolostone (Macdonald et al., 2012). Extension-controlled accommodation appears to have been rapidly infilled within the Fifteenmile basin. This contrasts with the slow sedimentation rates inferred for the Shaler Supergroup, where deposition occurred on a slowly subsiding carbonate ramp dominated by shallow-marine facies, such as mud-cracked dololomite, molar-tooth mudstone, and biostromal stromatolites that produced a layer-cake stratigraphy (Thomson et al., 2014). We therefore expect that the stratigraphy

of the Shaler Supergroup should be condensed relative to the Fifteenmile Group, rather than the opposite, as was suggested by Macdonald et al. (2012). Unless there is a significant hiatus, which may be recorded in a mudcracked, weakly karstic, subaerial exposure surface reported from the top of the Gibben Formation (Halverson et al., 2012; Macdonald et al., 2012), it appears unlikely that the base of the Fifteenmile Group could be much older than ~900 Ma, given the 811.5 Ma age from an ash layer in the upper the Reefal Assemblage (Macdonald et al., 2010). The Boot Inlet Formation of the Shaler Supergroup has a Re-Os depositional age of 892 ± 13 Ma from drill-core samples of black shale (van Acken et al., 2013), and there is an additional 1.5 km of strata preserved below it. We propose that the Fifteenmile Group does not correlate with the lower Mackenzie Mountains or Shaler supergroups (Fig. 14), but instead only with the upper Mackenzie Mountains Supergroup (Little Dal Group) and the upper Shaler Supergroup (Boot Inlet to Kilian formations).

3.7 Cause of cyclicity

Despite the Amundsen Basin being an intracontinental sag basin prior to deposition of the Kilian and Kuujjua formations (Rainbird, 1993), subsidence does not appear to have been gradual; rather it was rhythmic and generally outpaced sediment flux. Cyclic deposition during this time period was not all symmetrical or harmonious (Figs. 15 and 16), and difficult to correlate with other basins (Fig. 14). The overall depositional trend of the studied stratigraphic interval is transgressive, although there are sections within which parasequence stacking is consistent with regression.

Most cyclicity in the rock record is attributed to eustasy caused by climate change, oceanic ridge pulses, rhythmic basin subsidence, and the expanse or retreat of glaciers (i.e. glacioeustasy). Although the late Neoproterozoic is noted for global scale, low latitude glaciations, the studied formations were deposited between 850 and 750 Ma (van Acken et al., 2013), before the first mid-Cryogenian glaciation (Kendall et al., 2006; Smith, 2009). Therefore, glacioeustasy likely was not the driver for the documented cyclicity, however there may not be a record of contemporaneous glaciation. Other possible drivers for sea-level change are difficult to attribute specifically to the cyclicity documented from the Amundsen Basin. Greater understanding of causal factors may come from correlation with coeval sequences worldwide, for example with the Amadeus Basin of central Australia (Rainbird et al., 2012), however given the current lack of geochronological constraints, assigning causality to global cyclicity is difficult.

3.8 Basin duration, subsidence patterns and tectonostratigraphy

Lacking biostratigraphic controls and dated tuff beds, constraints from detrital zircon geochronology indicate the maximum duration for deposition of the Shaler Supergroup was approximately 400 million years (Rayner and Rainbird, 2013; Fig. 1), equivalent in Phanerozoic time from the base of the Devonian to the present day, although sedimentation was likely not continuous. This is not an unreasonable estimate if it is accepted that the upper half of the succession was deposited in approximately 180 million years (van Acken et al., 2013). Nevertheless, the Amundsen Basin appears to have been long-lived and its fill relatively thick (~4 km) for an intracontinental basin (cf.

Allen and Armitage, 2012). The duration of deposition of the Wynniatt Formation is also poorly constrained, but is estimated to have been 50-100 million years, and is up to 1 km at its thickest, in the southwest Minto Inlier. There are other examples of long-lived Paleozoic intracontinental basins, for example, the Williston Basin in Saskatchewan (~400 m.y.; Xie and Heller, 2009), the Northeast German Basin (~300 m.y.; Xie and Heller, 2009), and the Hudson Bay Basin (Upper Ordovician to present; Pinet et al., 2013).

Long et al. (2008) determined that there are at least five, unconformity-bound, supersequences (i.e. Sloss-style second-order cycles) in the Shaler Supergroup, implying that there were lengthy hiatuses between sequences, and that unconformities may separate discrete sub-basins or tectonic stages (cf. Krapez, 1997). Sequence sB4 (Long et al., 2008), and its component T-R cycles (T-R2 to T-R5), likely were deposited in one of these discrete, unconformity-bounded basins, separated from the underlying succession by a karstic unconformity at the top of the Boot Inlet Formation (Narbonne et al., 2000). This would imply that the surface separates different tectonic stages of the Amundsen Basin (Fig. 19), although it does not imply those tectonic stages were unique to the Amundsen Basin. Separate tectonic stages were described for the Ogilvie and Mackenzie basins (Turner and Long, 2008; Macdonald et al., 2012), but it was not established whether they were of the same rank as the unconformity-bounded sequences recognized in the Amundsen Basin. The episodic pattern of basin subsidence described is consistent with Paleozoic intracontinental basins such as the Michigan Basin, where four subsidence episodes (or tectonic stages) were identified using stratigraphic geometry, backstripping of subsidence rate, and depositional facies architecture (Howell and van der Pluijm,

1990). This suggests there is nothing novel about dividing the Amundsen Basin into unconformity-bound sequences that each relate to a different tectonic stage. The implication is that the stratigraphic succession of the Amundsen Basin is the product of stacked tectonic stages, each with a possibly different tectonic driver and subsidence regime.

We identify three different tectonostratigraphic units for the upper Shaler Supergroup, Stages 1-3 (Fig. 19). Stage 1 comprises T-R cycles 1 to 5 (of sB3 and sB4), which document a sag phase of basin development. This stage probably also includes underlying strata not discussed in this paper. Stage 2 comprises sequence sB5 (Long et al., 2008), which is made up of T-R cycles 1 to 4 of Rainbird (1993), and documents a period of active extension in the Amundsen Basin that is expressed through synsedimentary faulting interpreted to have been caused by plume-related thermal doming (Rainbird et al., 1993). Stage 3 comprises the Kuujjua (fluvial sandstone; Rainbird, 1992) and Natkusiak (flood basalts) formations that record the rifting stage of the Amundsen Basin.

The lack of extensional faulting, lateral continuity of strata, and predominance of shallow-water facies in the Amundsen Basin, supports sagging as the subsidence mechanism during Stage 1 (Figs. 14 and 19). The question is: what did that subsidence relate to? If it related to thermal decay, then it presumes there was a prior period of lithospheric stretching or asthenospheric doming. This sag phase of basin formation in the Amundsen Basin precludes the more active extensional phase of subsidence that accommodated the overlying Kilian, Kuujjua, and Natkusiak formations (Stages 2 and 3)

in rift-controlled basins. The subsequent thermal decay and flexural subsidence phases

Figure 19. Composite lithostratigraphy, T-R stratigraphy, sequence stratigraphy, and tectonostratigraphy of the Jago Bay, Minto Inlet, Wynniatt, Kilian, Kuujua, and Natkusiak formations. Lithostratigraphy and T-R stratigraphy of the Kilian Formation after Rainbird (1993). Supersequence boundaries after Young et al. (1981) and sub-sequence boundaries after Long et al. (2008). MDA – maximum depositional age.

(i.e. drift phase), after ocean opening, is not preserved or exposed in Minto Inlier, but is preserved in the Windermere Supergroup in the Mackenzie Mountains (e.g. Ross, 1991). Therefore, the combined Jago Bay, Minto Inlet and Wynniatt formations (Stage 1, T-R1 to T-R5) relate to subsidence that preceded the breakup of Rodinia along the north-northwest margin of Laurentia.

Macdonald et al. (2012) attributed post-830 Ma subsidence throughout NW Laurentia to extension and subsequent thermal decay associated with Rodinia passing over a mantle plume. That passage seemingly culminated at the emplacement of the 780 Ma Gunbarrel dykes and Tsezotene Sills (Harlan et al., 2003). This interpretation is based on the assumption that the conjugate margin to NW Laurentia was South China (Li et al., 1995). There was therefore a global-tectonic, mechanistic link between basin initiation in China, Australia, and Laurentia at ~830 Ma, and emplacement of the Gairdner and Guibei LIPs between ~830 and 810 Ma (Wingate et al., 1998; Li et al., 1999; Macdonald et al., 2012). The tectonic event correlates stratigraphically to the basin-forming event that accommodated the Reefal assemblage and Craggy dolostone (Fifteenmile Group), Little Dal Group (Mackenzie Mountains Supergroup), and possibly the Fort Collinson, Jago Bay, Minto Inlet, and Wynniatt formations (Shaler Supergroup). The maximum depositional age of the Fort Collinson Formation is 891 ± 22 Ma (Rayner and Rainbird, 2013) and is interpreted as high-energy, marine-reworked fluvial sandstone (Young and Long, 1977; Rainbird et al., 1996). This sandstone may record the initial transgression of this isolated basin event in the Amundsen Basin, supporting the timing of the subsidence model of Macdonald et al. (2012). The Gayna Formation of the Little Dal Group

(Mackenzie Mountains Supergroup) is considered stratigraphically correlative with the Fort Collinson Formation, and is composed of high-energy, cross-bedded oolitic dolograstone (Turner and Long, 2012), and therefore also represents a transgressive unit overlying the Silverberry/Stoneknife formations. It is important to note however, that there is no magmatic evidence for an ~830 Ma plume event in Laurentia, and that the 830 Ma event preceded the Franklin LIP by >100 million years, and the Gunbarrel LIP by 50 million years (Harlan et al., 2003). In fact, there is no evidence of an 830 Ma igneous event in Laurentia.

The tectonostratigraphic model of Macdonald et al. (2012) is nullified if the conjugate margin to northwest Laurentia was not South China. Regardless of the tectonic figuration, the most feasible interpretation of subsidence in the Amundsen Basin between ~900 to 780 Ma is multiple cycles of intra-plate extension that complemented coeval fault-controlled subsidence, sedimentation and volcanism of the same age in the Mackenzie and Ogilvie basins.

3.9 Inferences for the break-up of Rodinia along the northwest margin of Laurentia

Rift basins are by definition associated with extensional faulting, and have stratigraphic architectures pinned to rift-margin faults and fault-bound compartments. Although intracontinental basins lack obvious basin compartments and extensional faults, they typically lie inboard of rifted continental margins and have the same timing, but the causal mechanism for their subsidence has been much debated (e.g. Allen and Armitage,

2012). The Neoproterozoic successions of NW Laurentia present a similar spatial and temporal relationship between rift basins and intracontinental basins.

Sequence B strata that accumulated in NW-facing rift-basin margins are well documented from the Mackenzie Mountain Supergroup and Fifteenmile Group strata in the northern Cordillera (Figs. 1 & 14; Turner and Long, 2008; Macdonald et al., 2012). In contrast, coeval strata in the Shaler Supergroup were not associated with basin-margin faults or intrabasinal fault-bound compartments. Therefore, while faulting and extension were active in the Mackenzie and Ogilvie Mountains, the Amundsen Basin was subject to a passive subsidence regime up to and including deposition of the Wynniatt Formation (Stage 1, Fig. 19). There is evidence for thermal doming and extensional faulting during deposition of the overlying Kilian and Kuujjua formations (~780-720 Ma). Rainbird (1993) reported stratigraphic thinning, erosional beveling, and block faulting in those formations, and interpreted that this was consistent with crustal doming that accompanied mantle plume upwelling before the emplacement of the ~720 Ma Franklin LIP (Stage 2, Fig. 19). Despite the contrasts in stratigraphic architecture, these are coeval responses to regional-scale tectonic events, implying that subsidence drivers for the Amundsen Basin were somehow linked to those of the coeval rift basins in the northern Cordillera (Fig. 14).

That deposition of Sequence B strata south and west of Minto Inlier was already, at least in part, controlled by rift-related extensional faulting, has implications for correlation of higher-order basin events across northwestern Canada. Only deposition that responded to global (e.g. eustatic) controls may be registered in each basin, hence, the correlation of T-R5 in the Amundsen Basin with S7 in the Fifteenmile Basin (Macdonald

et al., 2012). In post-Wynniatt strata (Stages 2 and 3), rift signals extended along the entire northwest margin of Laurentia, supporting multi-stage, diachronous, non-correlative break-up of Rodinia.

3.10 Conclusions

- The Jago Bay Formation is part of Sequence sB3 and Minto Inlet and Wynniatt formations fit with Sequence sB4 of Long et al. (2008), and comprise one T-R cycle. Sequence sB4 has been further subdivided into four T-R cycles. T-R2 is composed of the Minto Inlet Formation and the basal part of the Lower Carbonate member, Wynniatt Formation. T-R3 comprises the remainder of the Lower Carbonate member, Black Shale member, and basal part of the Stromatolitic Carbonate member of the Wynniatt Formation. The boundary between T-R3 and T-R4 is a subaerial unconformity representing a hiatus of unknown duration. T-R4 comprises the remainder of the Stromatolitic Carbonate member, and the boundary between it and T-R5 lies at the base of the Upper Carbonate member of the Wynniatt Formation. T-R5 comprises the Upper Carbonate member, and its upper boundary corresponds to the formational boundary with the Kilian Formation, as well as the base of sequence sB5 (Long et al., 2008).
- T-R cycles are difficult to correlate regionally between basins because accommodation was greatly controlled by local tectonic factors (e.g. in the Fifteenmile basin). The exception appears to be sequence S7 from the Fifteenmile Group (Macdonald et al., 2012) and the coeval T-R5 defined here, which reflect a

significant regional-scale deepening event that may relate to the onset of a global-scale marine transgression.

- We propose that the Fifteenmile Group is only correlative to the upper Shaler Supergroup (Boot Inlet to Kilian formations) and Little Dal Group (Mackenzie Mountains Supergroup).
- We recognize three tectonostratigraphic units in the upper Shaler Supergroup. Stage 1 records subsidence of a sag basin, Stage 2 preserves evidence of syndepositional faulting and thermal doming, and Stage 3 represents the rifting phase with the extrusion of continental flood basalts.
- Subsidence during deposition of the middle to upper Shaler Supergroup in the Amundsen Basin may be related to lithospheric stretching and thinning if northwest Laurentia had passed over a mantle plume at 830 Ma as suggested by Macdonald et al., (2012), or may be related to intra-plate stresses related to some other cause of extension forming fault-bound rift basins in the Mackenzie Mountains and Fifteenmile basins.
- Analysis of pre-rift strata in the Amundsen Basin supports multi-phase, non-correlative break-up of Rodinia along the northwest margin of Laurentia.

CHAPTER 4: CHEMOSTRATIGRAPHY AND GEOCHEMISTRY

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Chemostratigraphy of the Shaler Supergroup, Victoria Island, NW Canada: a record of
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Research.*

4.1 Introduction

The Neoproterozoic Era records dramatic changes in seawater chemistry that eventually led to emergence of Phanerozoic surface conditions. Various isotopic systems have been employed to interpret secular changes in Neoproterozoic ocean chemistry, most commonly carbonate and organic carbon isotopes (e.g. Halverson et al., 2005; Halverson et al., 2010; Swanson-Hysell et al., 2010; Walter et al., 2000), strontium isotopes (e.g. Halverson et al., 2007b; Thomas et al., 2004; Veizer, 1989), and sulphur isotopes (e.g. Kaufman et al., 2007; Loyd et al., 2012). Much of the work has focused on the period of the late Neoproterozoic glaciations and the Ediacaran - Cambrian transition, although attention is now being drawn to older successions and a broader array of redox proxies (e.g. Canfield et al., 2008; Johnston et al., 2010; Sperling et al., 2013).

In contrast to the Phanerozoic, the Neoproterozoic is characterized by intervals with sustained positive $\delta^{13}\text{C}_{\text{carb}}$ values punctuated by abrupt shifts to negative $\delta^{13}\text{C}_{\text{carb}}$ values (e.g. Halverson et al., 2010). The majority of the large, negative perturbations to the Neoproterozoic carbon cycle are associated with glaciations, but there are exceptions. One of them, the Bitter Springs stage, is defined by a negative excursion in the global carbon isotope record reported from many localities, including Australia (Walter et al.,

2000), Zambia (Bull et al., 2011), Svalbard (Halverson et al., 2007a), Scotland (Prave et al., 2009), and northern Canada (e.g. Jones et al., 2010; Macdonald et al., 2010). A U-Pb age from an interbedded tuff preserved below the base of the Bitter Springs stage in the Ogilvie Mountains, Canada, yielded an age of 811.5 +/- 0.25 Ma (Macdonald et al., 2010). Although the Bitter Springs stage is not associated with a glaciation, it is interpreted to represent an interval of global change in the carbon cycle.

It is commonly assumed that the deep ocean became permanently oxygenated in the terminal Neoproterozoic and that this global redox shift was a driving force behind the diversification of multicellular life (e.g. Canfield and Teske, 1996; Des Marais et al., 1992). Many studies have sought to characterize the redox state of the deep ocean and specifically pinpoint when the oceans became fully oxygenated (e.g. Sahoo et al., 2012; Scott et al., 2008; Sperling et al., 2013). However, despite a surge of work over the past decade, there is still debate about Neoproterozoic marine redox evolution. Several independent studies suggest ocean oxygenation around 580 Ma (Canfield et al., 2007; Fike et al., 2006; McFadden et al., 2008). More recent work suggests at least temporary pervasive ocean oxygenation earlier in the Ediacaran (ca. 635–630 Ma), in the immediate aftermath of the Marinoan ‘Snowball Earth’ glaciation (Sahoo et al., 2012).

Sedimentary rocks of the Shaler Supergroup were deposited in the Amundsen Basin and are exposed in the Minto Inlier, Victoria Island, Northwest Territories, Canada (Fig. 1). The upper Shaler Supergroup was deposited between ~900 and 720 Ma and is exceptionally well preserved, providing an ideal target for marine paleoredox studies of the time interval leading up to the Sturtian glaciation. Further, this succession provides an opportunity to further explore $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ trends through this important period of

time. Herein, we present new C isotope, Fe-speciation, trace metal, and pyrite sulphur isotope data to interpret marine redox conditions and carbon cycle dynamics, including their global implications.

4.2 Geologic Setting

The Tonian (1000 – 850 Ma) to Cryogenian (850 – 635 Ma) Shaler Supergroup (~1100 – 723 Ma) is a > 4-km-thick sedimentary succession composed of predominantly shallow-marine carbonate rocks interstratified with mudstones, sandstones, and sulphate evaporites (Fig. 1) deposited in the Amundsen Basin, a shallow embayment within a larger epeiric sea that developed after the assembly of Rodinia (Rainbird et al., 1996; Rainbird et al., 1998; Young, 1981). The maximum age of the Shaler Supergroup is constrained by detrital zircon geochronology at <1151 Ma (Rayner and Rainbird, 2013). The minimum age is 723 $\pm$ 4/-2 Ma based on U-Pb geochronology of diabase sills that intrude the sedimentary succession and feed overlying flood basalts, part of the Franklin Large Igneous Province (Heaman et al., 1992). Three depositional ages of 892 $\pm$ 13 Ma (Boot Inlet Formation), 849 $\pm$ 48 Ma (Black Shale member of the Wynniatt Formation), and 761 $\pm$ 41 Ma (Upper Carbonate member of the Wynniatt Formation) based on Re-Os black shale geochronology provide additional chronostratigraphic constraints (Fig. 1, van Acken et al., 2013). The strata are relatively undeformed, with dips typically less than 10 degrees, and have experienced only low-grade contact metamorphism by the diabase sills (Fig. 3).

Figure 20. Photos of rocks that record the negative carbon isotope excursion in the Upper Carbonate member. A) Outcrop photo of black shale and limestone nodules enveloped by shales showing differential compaction. B) Photo from drill core GNME 07-04. Meter 305 is marked on the core, stratigraphic up is to the right. C) Thin section photo of carbonaceous, calcareous black shale, plane polarized light (PPL). D) Measured section with $\delta^{13}\text{C}_{\text{carb}}$ values overlain. Note that the values suddenly drop over the sharp boundary of the limestone to shale. E) Thin section photo of a calcarenite (turbidite). The bright white is the calcite cement that has infilled pore space surrounding the carbonate sand allochems. PPL.

The Boot Inlet Formation is characterized by large stromatolitic bioherms that formed along a carbonate ramp, with interstratified inner and mid-ramp deposits (Narbonne et al., 2000). An upward-shallowing trend is preserved in the overlying Fort Collinson Formation, an interval of fluvial and marine-reworked sandstones (Rainbird et al., 1994). The overlying Jago Bay Formation is an upward-shallowing succession of

carbonate-rich sedimentary rocks deposited in intertidal/lagoonal to supratidal and restricted intertidal depositional environments (Rainbird et al., 1994). Overlying the Jago Bay Formation, the Minto Inlet Formation consists of sedimentary rocks deposited under alternating restricted and open-marine conditions as reflected in bedded gypsum deposits and shallow subtidal limestones, respectively (Rainbird et al., 1996). The overlying Wynniatt Formation, interpreted as a storm-dominated carbonate ramp, is divided into four informal members: (1) Lower Carbonate member, an upward-deepening succession of supra- to sub-tidal carbonate rocks; (2) Black Shale member, a recessive interval of dark-grey siltstone and silty shale deposited in a pro-delta setting; (3) Stromatolitic Carbonate member, comprising stacked upward-shallowing cycles of subtidal to supratidal carbonate rocks and, (4) Upper Carbonate member, an upward-shallowing succession from subtidal black calcareous shale to peritidal, cross-bedded intraclastic grainstone and stromatolitic limestone (Thomson et al., 2014). The Kilian Formation is characterized by cyclically alternating shallow sub- to intertidal carbonate rocks and sabkha-related evaporite facies (Rainbird et al., 1996).

Asmerom et al. (1991) reported several Sr isotope values for the Shaler Supergroup, however the lack of good age control hindered robust global correlation. Jones et al. (2010) published the first detailed $\delta^{13}\text{C}_{\text{carb}}$ chemostratigraphic curve for the Shaler Supergroup. A negative excursion reported from the Wynniatt Formation was correlated with the global Bitter Springs stage, and a younger negative excursion documented in the Kilian Formation was tentatively correlated to the 'Islay' excursion (Brasier and Shields, 2000; Jones et al., 2010).

4.3 Methods

4.3.1 Geochemical approach to paleoredox interpretation

A multi-proxy approach using iron speciation, redox-sensitive trace metals, and pyrite sulphur isotope ratios is applied to estimate the paleoredox conditions in the Amundsen Basin. The Fe speciation proxy provides a fingerprint for the presence of anoxic water column conditions. It is typically accepted that in average continental and slope deposits, iron that is highly reactive toward hydrogen sulphide on short time scales (Fe_{HR}) accounts for less than ~40% of the total iron (Fe_{T}) pool (Poulton and Raiswell, 2002; Raiswell and Canfield, 1998). In contrast, stratigraphically persistent, highly reactive iron enrichments develop when there is Fe scavenging under an anoxic water column (Poulton and Canfield, 2011). Therefore, typical continental margin sediments with $\text{Fe}_{\text{HR}}/\text{Fe}_{\text{T}}$ ratios greater than 0.4 are interpreted to have been deposited under an anoxic water column. However, it should be noted that sediments deposited close to large anoxic zones can also have highly reactive iron enrichments (Scholz et al., 2014), but these settings are unlikely to leave a laterally and stratigraphically persistent record.

If Fe speciation data suggests that sediments were deposited under anoxic water column conditions, one can explore further whether ferruginous ($\text{Fe}^{2+} > \text{H}_2\text{S}$) or euxinic ($\text{Fe}^{2+} < \text{H}_2\text{S}$) water column conditions prevailed based on the ratio of pyrite (Fe_{Py}) to Fe_{HR} . If the $\text{Fe}_{\text{Py}}/\text{Fe}_{\text{HR}}$ ratio is >0.7 , the anoxic shales can be attributed to iron-limited, sulphide-rich (euxinic) depositional conditions, while a $\text{Fe}_{\text{Py}}/\text{Fe}_{\text{HR}}$ ratio <0.7 indicates ferruginous conditions (Poulton and Canfield, 2011).

The degree of redox-sensitive trace metal enrichment (Mo, U, V) for shales deposited under euxinic conditions, based on the Fe speciation proxies, can be used to make first-order inferences about the global marine redox state (e.g. Algeo and Lyons, 2006; Lyons et al., 2009; Partin et al., 2013; Reinhard et al., 2013; Scott et al., 2008). This idea builds from three key principles. First, because these elements are more soluble under oxic than anoxic conditions, the global ratio of oxic-to-anoxic marine conditions controls the magnitude of the marine burial fluxes and ultimately the size of the dissolved Mo, V, and U reservoirs in seawater. Greater extents of anoxia and euxinia decrease the sizes of the Mo, V, and U reservoirs (Emerson and Huested, 1991). Second, after the permanent rise of atmospheric oxygen during the ‘Great Oxidation Event’ (around 2.32 billion years ago [Ga]; Bekker et al., 2004), the source flux of these elements (e.g. oxidation of continental crust) was decoupled from atmospheric oxygen levels and remained constant (see Partin et al., 2013; Reinhard et al., 2013). This assumption is based on the low levels of atmospheric oxygen needed to oxidize sulphide minerals and to mobilize Mo, V, and U during weathering (e.g. Anbar et al., 2007; Reinhard et al., 2009). Lastly, the sizes of the Mo, V, and U marine reservoirs control the magnitudes of enrichment in euxinic sediments and rocks (e.g., Algeo and Lyons, 2006). Furthermore, because particulate organic matter sequesters Mo, V, and U, there is strong link between particulate organic matter fluxes and Mo, V, and U burial (Algeo and Lyons, 2006). Therefore, ratios of redox-sensitive elements to total organic carbon (TOC), in combination with traditional enrichment factors, are the most effective way of tracking the size of the Mo, V, and U marine reservoirs (Lyons et al., 2009). The utility of this approach is clearly illustrated in euxinic sections that capture the onset, the middle, and

the end of Cretaceous Ocean Anoxic Events (OAEs) (e.g. Hetzel et al., 2009), and it has been widely used to track both Phanerozoic and Precambrian redox evolution (Gill et al., 2011; Sahoo et al., 2012; Scott et al., 2008), and the extent of euxinia in the global ocean (Reinhard et al., 2013).

Light (^{32}S -enriched) sulphur isotope compositions for pyrite ($\delta^{34}\text{S}_{\text{py}}$) indicate marine sulphate concentrations sufficiently high enough to avoid reservoir effects that can mute the ^{32}S preference that characterizes bacterial sulphate reduction (e.g., Leavitt et al., 2013; Scott et al., 2014). Similarly, heavy $\delta^{34}\text{S}_{\text{py}}$ values can be linked to very low marine sulphate concentrations. Low concentrations (<1 mM) are often a consequence of widespread marine anoxia and accompanying high globally integrated rates of marine pyrite burial (Scott et al., 2014 and references therein). Locally, high fluxes of labile particulate organic matter can fuel high rates of microbial sulphate reduction, and this factor can result in small fractionations during sulphate reduction and thus relatively heavy pyrite (see Levitt et al., 2013 and references therein).

4.3.2 *Sampling*

785 samples spanning the stratigraphic interval from the Boot Inlet Formation to the Kilian Formation (Fig. 21) were collected for stable isotope analysis from measured stratigraphic sections and logged exploration drill cores. Altered thermal contact aureoles adjacent to diabase sills were avoided. Carbonate samples were cut to expose fresh surfaces and textures and microdrilled with a diamond dental drill to produce powders. Micritic microfacies were preferentially sampled, avoiding veins. Thin sections were

prepared from all representative lithofacies for petrographic screening and description. Shale samples were also cut to expose fresh surfaces and to avoid veins and visible alteration zones, ultrasonicated with deionized water, oven dried, and pulverized in agate.

4.3.3 Analytical methods

$\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ analyses were performed at the G.G. Hatch Isotope Laboratory at the University of Ottawa. Samples were weighed into exetainers followed by addition of 0.1 mL of H_3PO_4 to the side of the exetainer. The exetainers were then helium-flushed while horizontal. Reaction at 25°C for 24 hrs (calcite) or 50°C for 24 hrs (dolomite) was followed by analysis in continuous flow. The measurements were performed on a Delta XP connected to a Gasbench II, both from Thermo FinniganTM. All $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values are reported in delta notation relative to international standards on the Vienna-Pee Dee Belemnite (VPDB) scale. Analytical precision (2 sigma) for both carbon and oxygen isotope analyses is +/- 0.1permil. Data for C and O were normalized using international standards NBS-18, NBS-19, and LSVEC (C only). Quality control duplicates were analyzed at the beginning, middle, and end of each run.

$\delta^{13}\text{C}_{\text{org}}$ analyses were performed at the Stable Isotopes for Innovative Research (SIFIR) laboratory at the University of Manitoba. The concentration of total organic carbon was determined on an Eltra IR C/S analyzer as the difference between total carbon (determined by combustion) and total inorganic carbon (determined by acidification). The concentration of total sulphur was also determined on an Eltra IR C/S analyzer. For $\delta^{13}\text{C}_{\text{org}}$ analyses, sample powder was weighed in a silver cup, and ultrapure 6N HCl was

added to remove inorganic carbon. To improve the efficiency of sample combustion, temperature in the oxidation column was raised to 1100°C, and a ‘macro’ O₂ injection loop was utilized. Tin cups were used as a catalyst to increase the combustion temperature through exothermic reaction. Analyses were performed using a Costech™ 4010 Elemental Analyzer (EA) coupled to a Thermo Finnigan™ Delta V Plus isotope-ratio mass-spectrometer via an open-split interface (ConFlo III, Thermo Finnigan™). Sample normalization was performed using the two-point calibration described in Coplen et al. (2006). Two international standards (USGS40 and USGS41) were analyzed at the beginning, middle, and end of each run. A calibration line was calculated for each run by least-squares linear regression using the known and measured isotope values of the calibration standards. To monitor the quality of sample preparation and analysis performance, the international standard USGS Green River shale SGR-1b ($\delta^{13}\text{C}_{\text{org}} = -29.3 \pm 0.1\text{‰ V-PDB}$) was treated and analyzed as an unknown. Replicate analyses of SGR-1b standard yielded the $\delta^{13}\text{C}_{\text{org}}$ value of $-29.5 \pm 0.1\text{‰}$ (n=24).

Sulphur isotope analyses were also performed at the SIFIR laboratory at the University of Manitoba. Chromium-reducible sulphur (CRS), dominantly pyrite in our samples, was extracted following the method of Canfield et al. (1986). Approximately 0.3-0.5 mg of extracted Ag₂S was weighed into Sn cups (5x3.5 mm), wrapped, and analyzed for S isotopes on a Costech™ 4010 Elemental Analyzer (EA) coupled to a Thermo Finnigan™ Delta V Plus isotope-ratio mass-spectrometer (IRMS) via an open-split interface (ConFlo IV, Thermo Finnigan™). In the EA, the sample Sn cup was dropped in a helium flow to the 1st reaction column, containing oxidation and reduction chemicals at 1050°C and a ‘macro’ O₂ injection loop was utilized. The produced SO₂

passed through the 2nd reaction column at 890°C, containing quartz chips. This was done to minimize the uncertainty in corrections related to oxygen isotopes (Fry et al., 2002). All S isotope ratios are reported in delta notation relative to international standards on the VCDT scale. Sample normalization was performed using the two-point calibration described in Coplen et al. (2006). Three international standards (IAEA-S1, IAEA-S2, and IAEA-S3) were analysed at the beginning, middle, and end of each run. A calibration line was calculated for each run by least-squares linear regression using the known and measured isotope values of the calibration standards. To monitor the quality of sample preparation and analysis performance, one internal sulphide standard (Maine Light (Ag₂S), $\delta^{34}\text{S} = -30.9 \pm 0.4\text{‰}$ VCDT, n = 24) was treated and analysed as an unknown. Replicate analyses of the quality-check standard yielded the $\delta^{34}\text{S}$ value of $-31.1 \pm 0.3\text{‰}$ (n=7).

The sequential Fe extractions were performed at the Biogeochemistry Lab at the University of California, Riverside, following well-established techniques (e.g. Reinhard et al., 2009). The pyrite Fe concentrations (Fe_{py}) were calculated, assuming a pyrite stoichiometry, using the CRS concentrations determined at the SIFIR laboratory. Approximately 100 mg of fresh sample were used following the Poulton and Canfield (2005) sequential Fe speciation technique. Iron concentrations for the solutions were measured with an Agilent 7500 inductively coupled plasma mass spectrometer (ICP-MS). Aluminum concentrations were also measured for each extract to monitor Fe contribution from less reactive silicates and it was found to be less than 0.1%. ICP-MS precision and accuracy were monitored with multi-element standards and a geostandard (USGS-SDO-1). Drift within a run led to less than 5% error. Duplicate measurements, within the same

or multiple ICP-MS runs, agreed within 7%. The sequential Fe procedure isolates the following operationally defined Fe fractions: (1) Na-acetate Fe representing carbonate-bound Fe, herein referred to as Fe_{carb} ; (2) Na dithionite Fe representing crystalline oxides such as goethite and hematite, herein referred to as Fe_{ox} ; and (3) oxalate Fe representing magnetite, herein referred to as Fe_{mag} . Combined, Fe_{py} , Fe_{carb} , Fe_{ox} , and Fe_{mag} represent the ‘highly reactive’ Fe pool (Fe_{HR}).

Major and minor element concentrations were measured at ACME Analytical Laboratories Ltd using ICP-MS analysis after multi-acid digestion. Accuracy and precision were monitored with duplicate analyses and a blind shale geostandard (USGS SGR1b). ACME Analytical Laboratories uses a quality monitoring system compliant with the International Standards Organization (ISO) 9001 Model for Quality Assurance and ISO/IEC 17025 General Requirements for the Competence of Testing and Calibration Laboratories. Further details are available from ACME Analytical Laboratories, Vancouver, British Columbia.

4.4 Results and Basic Data Interpretations

4.4.1 Carbon isotopes and carbonate sedimentology

4.4.1.1 Carbon isotopes

The carbon isotope curve for the upper Shaler Supergroup (Fig. 21) represents a composite of 19 measured sections and three drill cores logged and sampled from the southwest and northeast portions of the Minto Inlier (Fig. 1); see supplementary data for a complete set of tabulated data. $\delta^{13}\text{C}_{\text{carb}}$ values range from -14.3 to +7.7‰, although

Figure 21. Composite section of the upper Shaler Supergroup. A) $\delta^{13}C_{carb}$, $\delta^{13}C_{org}$, and $\delta^{18}O$ values are shown. Open black circles are data from Jones et al. (2010). B) Mo, V, U (in ppm), Mo, V, and U (normalized to TOC), TOC, and $\delta^{34}S_{py}$ data. C) Cross-plot of $\delta^{13}C_{carb}$ and $\delta^{13}C_{org}$ values. D) Legend. VPDB – Vienna Pee Dee Belemnite, VCDT – Vienna Canyon Diablo Troilite, TOC – total organic carbon.

most samples fall within a range of $\sim 0 - +5\%$, with an average value of $+3.0\%$, which is

a baseline for the Neoproterozoic (e.g. Halverson et al., 2010). Negative deviations occur within the Boot Inlet Formation, the Fort Collinson Formation, the Upper Carbonate member of the Wynniatt Formation, and the upper Kilian Formation (Fig. 21). The most dramatic of these is the negative excursion preserved in the Upper Carbonate member (Wynniatt Formation), where multiple $\delta^{13}\text{C}_{\text{carb}}$ values ranging from ~ -10 to -14‰ are recorded, which are well below the mantle value of -5 to -6‰ .

$\delta^{13}\text{C}_{\text{org}}$ values average -25.9‰ but range from -34.2 to -14.4‰ . $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ values co-vary throughout the sampled intervals (Fig. 21). A cross-plot of $\delta^{13}\text{C}_{\text{carb}}$ versus $\delta^{13}\text{C}_{\text{org}}$ values displays good correlation ($R^2 = 0.66$, Fig. 21), although values cluster more tightly around the line of best fit for the more negative part of the dataset, with a greater spread in data for the samples with the most positive $\delta^{13}\text{C}_{\text{carb}}$ values. In the traditional framework of interpretation, this would indicate that the C_{org} was derived and fractionated from the same contemporaneous dissolved inorganic carbon captured in the carbonate carbon isotope values (e.g., Knoll et al., 1986; Kump et al., 2011; Swanson-Hysell et al., 2010).

4.4.1.2 Sedimentology and stratigraphy of the negative excursion stratigraphic interval

The stratigraphic interval marked by the large negative excursion in the Upper Carbonate member consists predominantly of laminated, carbonaceous calcareous black shale (Fig. 20b and c) representing sedimentation below storm wave-base (SWB) on the outer slope (Thomson et al., 2014). Thin beds of normally graded calcarenite, interpreted as turbidites, are interbedded within the shale. Although the carbonate beds are composed

largely of transported material, grains are cemented in a micritic diagenetic matrix (Fig. 20e). There are also carbonate-rich nodules within the shales, which formed *in situ* during early diagenesis as indicated by enveloping shale, which is warped around the nodules by later differential compaction (Fig. 20a). Early seafloor dissolution and cementation is visible along bedding contacts as irregular surfaces that are buckled or micro-fractured and are draped by overlying shale. Where deformed, pore space is filled by blocky, subhedral, cloudy, non-equidimensional calcite cement. All secondary features, including nodules, fractures, and blocky cement, were avoided during sampling for chemostratigraphy.

An erosional surface separates underlying, shallow-marine carbonate rocks from the interval of deep-water, black, nodular, calcareous shale, and likely represents a sedimentary hiatus (Thomson et al., 2014). The surface is coincident with the boundary between the Stromatolitic Carbonate and Upper Carbonate members (Fig. 21), and the overlying shale represents a major marine transgression. The negative $\delta^{13}\text{C}_{\text{carb}}$ excursion is preserved in this shale (Fig. 20d).

4.4.2 Paleoredox proxies

Most of the data collected from our composite section are too sparse and ambiguous to support confident interpretations about paleoredox conditions. A carbonaceous shale interval in the Upper Carbonate member is the stand-alone exception (Figs. 20 and 22). It is likely though that the majority of the studied interval contains sediments deposited in oxic settings, as the lithologies and sedimentary structures indicate deposition in shallow-marine to emergent conditions (Rainbird et al., 1996; Thomson et al., 2014). The only interval that preserves slightly deeper sedimentation in

the upper Shaler Supergroup is the Black Shale member of the Wynniatt Formation, which was deposited below fair-weather wave-base, but above storm wave-base (Thomson et al., 2014), under oxic to suboxic conditions (Lafond, 2011).

Several redox proxies point to deposition of the carbonaceous shale interval in the lower part of the Upper Carbonate member under euxinic conditions (Fig. 22). $\text{Fe}_{\text{HR}}/\text{Fe}_{\text{T}}$ ratios are well above 0.4, indicating anoxic conditions, and furthermore $\text{Fe}_{\text{Py}}/\text{Fe}_{\text{HR}}$ ratios above 0.7 indicate euxinic conditions (Fig. 22). There are also significant enrichments in

GNME drill core 07-04 (samples 1 RAT-DT095)

Figure 22. Geochemical data from drill core GNME 07-04 spanning the Bitter Springs stage. Interpretation of paleoredox conditions is shown beside the stratigraphic section. The typical continental margin baseline for Fe speciation parameters is indicated by the vertical red lines. Note: the location of this core is indicated on Fig. 1 by a black star, and the V data is presented using a log scale. Re-Os age is from van Acken et al. (2013). EOH – end of hole, sh - shale, are – arenite, rud – rudstone, VPDB – Vienna Pee Dee Belemnite, VCDT – Vienna Canyon Diablo Troilite, PPM – parts per million.

Mo (up to 179 ppm), V (up to 2826 ppm), and U (up to 20.1 ppm), comparative to continental crust averages of 1.1 ppm, 98 ppm, and 1.7 ppm, respectively (Wedepohl, 1995). High Mo enrichments independently corroborate the euxinic conditions suggested by the Fe proxy data (see Scott and Lyons, 2012).

Given that the Upper Carbonate member shales were deposited under euxinic conditions, the Mo, V, and U enrichments can shed light on broader marine redox conditions (cf. Scott et al., 2008). The Mo enrichments in the carbonaceous shale interval of the Upper Carbonate member are elevated relative to the mean value for the mid-Proterozoic (2.0–0.74 Ga) euxinic shale (40.5 ± 22.5 ppm at the 95% confidence level), and are comparable to the mean value for Phanerozoic euxinic shale (170.2 ± 33.4 ppm at the 95% confidence level; Reinhard et al., 2013). The elevated Mo to TOC ratios from the euxinic shale interval in our study are comparable to mid-Proterozoic ratios (<100 ppm/wt %) and lower than Phanerozoic ratios (>100 ppm/wt %; see Sahoo et al., 2012). U enrichments and U/TOC ratios are similar to those recorded in other Mesoproterozoic and early Neoproterozoic successions (Fig. 22; cf. Partin et al., 2013). The V enrichments and V/TOC ratios, in contrast, are well above the typical Proterozoic values and similar to Phanerozoic values (cf. Sahoo et al., 2012). Additionally, the Mo, U, and V enrichments are notably higher than those measured in the younger Walcott member of the Chuar Group (Dahl et al., 2011).

There are light $\delta^{34}\text{S}$ values in the Upper Carbonate member, with the lowest $\delta^{34}\text{S}$ values below -35‰ . Assuming that at no point in the Cryogenian does sulphate $\delta^{34}\text{S}$ ($\delta^{34}\text{S}_{\text{SO}_4}$) drop below $+10\text{‰}$ (Halverson et al., 2010), the $\Delta^{34}\text{S}$ ($\delta^{34}\text{S}_{\text{SO}_4} - \delta^{34}\text{S}_{\text{pyr}}$) values in the Upper Carbonate member are commonly greater than 45‰ , which is unusually high

for Precambrian euxinic shales (Scott et al., 2014). Given euxinic depositional conditions and the absence of macroscopic sulphides, finely-disseminated pyrite in the analysed samples (Fig. 20c) is likely to be depositional (e.g., Lyons, 1997). These highly negative $\delta^{34}\text{S}$ values of pyrite indicate a water column where sulphate was not limiting, allowing for preferential microbial utilization of ^{32}S during dissimilatory sulphate reduction. This relationship suggests mM-level seawater sulphate concentrations (cf. Scott et al., 2014). On a longer time scale, there is a broad trend towards more negative S isotope values from the Boot Inlet Formation to the Jago Bay Formation, followed by a highly variable trend towards more positive S isotope values through the Wynniatt Formation with a notably sharp rise in the lower part of the Upper Carbonate member, after which S isotope values hover at $\sim 0\text{‰}$ (Fig. 21). Given that the pyrite in our samples formed under a range of redox conditions (e.g., in the water column under euxinic conditions and in pore waters beneath oxic bottom waters), it is speculative to interpret the environmental or geochemical significance of this trend.

4.5 Discussion

4.5.1 Testing existing correlations with new chemostratigraphic data

While most of the $\delta^{13}\text{C}_{\text{carb}}$ values through the sampled portion of the Shaler Supergroup are positive and range from ~ 2 to $+5\text{‰}$, within the Boot Inlet Formation, $\delta^{13}\text{C}_{\text{carb}}$ values fall close to 0‰ (Fig. 21). This interval likely correlates with the lower Stone Knife/Silverberry formations of the Mackenzie Mountains Supergroup, Mackenzie Mountains, and the Gibben Formation of the Fifteenmile Group, Ogilvie Mountains (Fig.

23), where similar values were reported. While data from this stratigraphic interval in our study area is limited, this inference is consistent with previous stratigraphic correlations based on litho- and sequence stratigraphy (Long et al., 2008; Macdonald et al., 2012; Rainbird et al., 1996; Young et al., 1979).

There is a drop in $\delta^{13}\text{C}_{\text{carb}}$ values in the Fort Collinson Formation extending into the lower part of the Jago Bay Formation (Fig. 23; highlighted in grey). This decrease occurs within a siliciclastic-rich stratigraphic interval composed of cross-bedded quartz arenites and dolomitic quartz arenites (Rainbird et al., 1994). Values for $\delta^{13}\text{C}_{\text{carb}}$ range from ~ -1 to $+6$ ‰ through the Jago Bay and Minto Inlet formations and the lower part of the Wynniatt Formation. The magnitude of variability in the C isotope curve appears to be enhanced in our study area when compared to the Fifteenmile Group and the Mackenzie Mountains Supergroup (Fig. 23), likely due to lower sedimentation rates in the Amundsen Basin (Thomson et al., in review). The $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ values co-vary through this interval (Fig. 21), which is typically interpreted as a signal of alteration (e.g., Derry, 2010b; for alternative view see Bjerrum and Canfield, 2011). Further, carbon isotope values in dolomitic quartz arenites, as for the Fort Collinson Formation, commonly do not reflect the isotopic composition of the marine DIC reservoir (Knauth and Kennedy, 2009). Given these potential concerns, the $\delta^{13}\text{C}_{\text{carb}}$ trend in this stratigraphic interval cannot be confidently used to make detailed chemostratigraphic correlations with sedimentary successions in other basins.

One of the outstanding questions posed by Jones et al. (2010) was whether the negative excursion present in the Upper Carbonate member of the Wynniatt Formation

was the result of fluid alteration at the contact with a thick diabase sill that intruded the

Figure 23. Potential correlation of the Shaler Supergroup with the Fifteenmile Group and Mackenzie Mountains Supergroup based on lithostratigraphy and $\delta^{13}\text{C}_{\text{carb}}$ data. Data from the Ogilvie Mountains is after Macdonald et al. (2012), data from the Mackenzie Mountains is after Halverson et al. (2006). Inset map shows location of the Amundsen Basin with respect to the Ogilvie and Mackenzie mountains. Two red arrows point to the “double positive peak” that occurs below the Bitter Springs stage, which is highlighted in orange. The green bar highlights the correlation of the Boot Inlet Formation with the Stone Knife, Silverberry and Gibben formations. The grey bar highlights the negative C isotope values of the siliciclastic-dominated Fort Collinson Formation that may reflect early diagenetic alteration. DCF – Dodo Creek Formation, lcm – Lower Carbonate member, bsm – Black Shale member, scm – Stromatolitic Carbonate member, ucm – Upper Carbonate member.

strata at their sampling location. Where we sampled that horizon, both in core and outcrop, 86 and 150 km away, respectively, from the section Jones et al. (2010) sampled (Fig. 1), a similar excursion is preserved. There are no sills in, or adjacent to, our sections, and the presence of this negative excursion in all three localities sampled confirms that negative $\delta^{13}\text{C}_{\text{carb}}$ values in the Upper Carbonate member of the Wynniatt Formation are present throughout the basin. Jones et al. (2010) tentatively linked these negative $\delta^{13}\text{C}_{\text{carb}}$ values to the roughly coeval Bitter Springs isotopic stage.

The Bitter Springs isotopic stage follows two positive peaks in the $\delta^{13}\text{C}_{\text{carb}}$ record that occur prior to the downturn that defines the negative excursion, first recognized and thus named for the Bitter Springs Formation from the Amadeus Basin, Australia (Hill and Walter, 2000). Where this excursion has been reported in other successions, there is an abrupt shift to negative values of ~ -4 ‰, and then the trend gradually recovers and returns to positive values, producing an asymmetric curve (e.g., Halverson, 2006; Halverson et al., 2007b; Macdonald et al., 2010; Swanson-Hysell et al., 2010). $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ values co-vary where they previously were measured through the Bitter Springs excursion (e.g., Swanson-Hysell et al., 2010).

The carbon isotope trend in the Upper Carbonate member of the Wynniatt Formation is likely to represent the Bitter Springs isotopic stage. Foremost, the double positive peak preceding the Bitter Springs isotopic stage is clearly defined in the Wynniatt Formation. However, the highly negative values in the Upper Carbonate member are more difficult to interpret, given that they occur in deeper-water facies, and C isotope values are much lighter than those reported from other localities. Authigenic carbonate phases are likely to have $\delta^{13}\text{C}_{\text{carb}}$ values that reflect pore water compositions

rather than overlying seawater DIC values. However, beds composed largely of transported detrital carbonate still have negative $\delta^{13}\text{C}_{\text{carb}}$ values, which suggests that the negative $\delta^{13}\text{C}_{\text{carb}}$ values are at least partially linked to a drop in surface marine $\delta^{13}\text{C}_{\text{DIC}}$ values, unless they have been completely diagenetically reset. The very negative $\delta^{13}\text{C}_{\text{carb}}$ values in this interval ($< -6 \text{ ‰}$) likely reflect lithologies more strongly affected by authigenic carbonate. We return to a discussion of the origin of the negative $\delta^{13}\text{C}_{\text{carb}}$ values in section 4.5.2.

Assuming that the Upper Carbonate member of the Wynniatt Formation is correlative to the Bitter Springs isotopic stage, there is an apparent discrepancy in the two available ages for this stage. An age of $811.5 \pm 0.25 \text{ Ma}$ was reported from a tuff from the Fifteenmile Group in the Ogilvie Mountains (Macdonald et al., 2010), while our study area has an age of $761 \pm 41 \text{ Ma}$ from Re-Os dating of black shale (Van Acken et al., 2013). This difference can easily be explained however, as the ca. 811 Ma tuff occurs ~50 m stratigraphically below the onset of the decrease in $\delta^{13}\text{C}_{\text{carb}}$ values in the Fifteenmile Group (Macdonald et al., 2012; Macdonald et al., 2010), while the ca. 761 Ma age occurs within the negative excursion in the Wynniatt Formation, specifically, ~25 m above the base of the excursion, near the inflection point where values begin to increase. Also, some of the discrepancy may relate to the observation that the ca. 761 Ma age comes from a stratigraphic interval of black shale deposited on the outer slope where sediment accumulation rates are slow and compaction is great. In addition, at both localities, the Bitter Springs stage occurs in successions directly overlying sequence boundaries (Macdonald et al., 2012; Thomson et al, in review). In the Ogilvie Mountains, it overlies a transgressive-regressive sequence boundary that occurs above the dated tuff

(Macdonald et al., 2012; Fig. 5), and directly overlies an unconformity in the Minto Inlier (Thomson et al., 2014). As the ca. 811 Ma age comes from the strata below an unconformity and below the Bitter Springs stage, it implies a further loss of time between the dated horizon and the strata that host the isotope excursion. One must also consider the large error on the 761 ± 41 Ma Re-Os age (Van Acken et al., 2013), and so the sampled interval may be as old as 802 Ma. Further complicating matters, basalts of the Johnny's Creek Member of the Bitter Springs Formation, considered to belong to the same mantle plume event as the 827 ± 6 Ma Gairdner Dyke Swarm (Wingate et al., 1998) based on similar trace element geochemistry (Zhao et al., 1994), directly overlie the Love's Creek Member, the type section for the Bitter Springs Stage. Clearly, additional work combining precise geochronology and chemostratigraphy is needed to resolve this issue and to test the idea that the Bitter Springs Stage is globally contemporaneous phenomena.

After recovery of $\delta^{13}\text{C}_{\text{carb}}$ values above the Bitter Springs stage, there is a steady decrease in $\delta^{13}\text{C}_{\text{carb}}$ values down to $\sim -5\%$ through the Kilian Formation. This trend was originally reported by Jones et al. (2010), who interpreted the negative excursion to be correlative with the pre-Sturtian, "Islay" excursion (Brasier and Shields, 2000) on the basis of: (1) similar trend and amplitude, (2) the relative age of the Kilian Formation, and (3) because they sampled sedimentary rocks recording the excursion in an outcrop capped by a diamictite. We have not observed this diamictite; however, the age of this excursion is consistent with the notion of equivalence to the Islay excursion, as the age of the Kilian Formation is bracketed by the Re-Os age of the underlying Wynniatt Formation (761 ± 41 Ma; Van Acken et al., 2013) and the $723 \pm 4/-2$ Ma U-Pb age of crosscutting sills

(Heaman et al., 1992). We agree with the correlation of Jones et al. (2010), who correlated the Kilian Formation to the Coates Lake Group, which is also considered correlative with the Callison Lake dolostone of the Mount Harper Group of the Ogilvie Mountains (Macdonald et al., 2012). Importantly, this chemostratigraphy-based correlation supports earlier correlations based on litho- and sequence stratigraphy (Aitken et al., 1978; Rainbird et al., 1996; Young et al., 1979).

4.5.2 Negative carbon isotope anomalies: Evaluation of the different mechanisms and an interpretation for the Amundsen Basin

Determining the significance of highly negative $\delta^{13}\text{C}_{\text{carb}}$ values, below those of the mantle carbon, is essential to our understanding of the evolution of Earth surface processes during the Neoproterozoic. There are several differing interpretations of negative shifts in $\delta^{13}\text{C}$ values: (1) resetting during early diagenesis; for example, during a drop in sea-level whereby meteoric waters are flushed through the carbonate platform, driving C isotope values down (Knauth and Kennedy, 2009; Swart and Kennedy, 2012); (2) resetting during burial diagenesis as a result of rock-fluid interaction (Derry, 2010a, b); (3) that C isotope values record primary seawater chemistry whereby fluctuations were caused by changes in the global carbon cycle through, for example, the oxidation of a large pool of dissolved organic carbon (DOC; Jiang et al., 2007; Schrag et al., 2002; Swanson-Hysell et al., 2010) or methane released by destabilization of methane clathrates (Bjerrum and Canfield, 2011; Dickens et al., 1997; Jiang et al., 2003); and (4) fluctuations were caused by precipitation of authigenic carbonate from sediment pore-fluids during early diagenesis (Schrag et al., 2013). We add to this discussion by

evaluating the applicability of each of these models to the negative $\delta^{13}\text{C}$ anomaly in the Upper Carbonate member of the Wynniatt Formation, which likely records the Bitter Springs isotope stage.

4.5.2.1 Early meteoric diagenesis

The Bitter Springs stage is preserved in carbonaceous, calcareous shales that occur stratigraphically above a sequence boundary within sediments that were deposited during a transgression (see Thomson et al., 2014). If meteoric fluids were indeed responsible for the negative carbon isotope excursion, one would expect, in contrast, to see the negative values preserved in the carbonates below the sequence boundary, reflecting a time when sea-level was at its lowest, and drawdown of meteoric fluids would have flushed fluids with isotopically light carbon through the carbonate platform; not in the transgressive sediments above the sequence boundary as is the case in the Wynniatt Formation.

4.5.2.2 Burial diagenesis

The Bitter Springs stage is reported from stratigraphic sections on several continents, although its age remains poorly constrained. If deep burial diagenesis caused the Bitter Springs anomaly, then it must have affected a number of sections synchronously, which seems highly unlikely. Even in the Amundsen Basin, the negative anomaly is recorded in three sections that form a transect of >150 km in length. Further, the $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ trends are coupled, which would not be expected if the carbonate

C isotope values were diagenetically reset at the burial stage (Jiang et al., 2012). Close co-variation between the $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ values is commonly used as the strongest argument against diagenetic alteration (Sansjofre et al., 2011).

4.5.2.3a Change in primary seawater chemistry by oxidation of organic matter

Considering that the $\delta^{13}\text{C}_{\text{carb}}$ record is primary, one way to explain $\delta^{13}\text{C}_{\text{carb}}$ fluctuations is by inferring change in the rate of burial of organic carbon relative to total carbon (e.g. Hayes et al., 1999), which has already been proposed as a mechanism for sustained positive $\delta^{13}\text{C}_{\text{carb}}$ values through the Neoproterozoic (e.g. Holland, 1984; Knoll et al., 1986). In a deviation from this traditional view, some authors have attributed the markedly negative excursions in C isotope values of the Neoproterozoic carbonates to the oxidation of a large DOC pool (e.g. Fike et al., 2006; Jiang et al., 2007; Rothman et al., 2003; Schrag et al., 2002; Swanson-Hysell et al., 2010). However, this possibility has been challenged on the basis that DOC oxidation would rapidly consume all oxidants in the atmosphere-ocean system and would dramatically increase atmospheric CO_2 (Bristow and Kennedy, 2008; Schrag et al., 2013). Although our data do not bear directly on this interpretation, based on the arguments presented above oxidation of DOC seems unlikely to be the sole explanation for the markedly negative $\delta^{13}\text{C}_{\text{carb}}$ excursion.

4.5.2.3b Release of methane clathrates

Dissipation of methane clathrate reservoirs has been inferred to explain negative carbon isotope excursions in seawater composition, for example in association with

‘Snowball Earth’ glaciations (Schrag et al., 2002), and fluctuations in the Neoproterozoic and younger (e.g. Paleogene) carbon cycle (Bjerrum and Canfield, 2011; Dickens, 2003). Sustained positive $\delta^{13}\text{C}_{\text{carb}}$ values in the underlying carbonate rocks of the Shaler Supergroup (Fig. 21) and in the carbon isotope record leading to the Bitter Springs stage, in general indicate high relative burial rates of organic carbon globally (cf. Hayes et al., 1999 and references therein), a condition favourable to methane production during diagenesis. Some of this methane likely escaped to the ocean and atmosphere, where it was oxidized, but some may have trapped in the sediments forming methane clathrates. A subsequent dissipation of the methane clathrate reservoir could result in water column deoxygenation, as well as the source of isotopically light carbon (methane $\delta^{13}\text{C} \sim -60\text{‰}$; Kvenvolden, 1993) to produce a negative carbon isotope excursion.

4.5.2.4 Precipitation of authigenic carbonate

Carbonate can be precipitated from sedimentary pore fluids during very early diagenesis in association with reduction of iron and sulphate, primarily through the oxidation of methane or organic carbon, and provides a sink for isotopically light carbon in the carbon cycle (Schrag et al., 2013). Bacterially mediated oxidation of organic carbon produces $\delta^{13}\text{C}$ values of approximately -25‰ , which is then incorporated into carbonate (see Irwin et al., 1977). In Proterozoic carbonates it can be difficult to identify authigenic carbonate petrographically since almost all of the carbonate has been neomorphosed to microspar, and dolomite replacement is common. Carbonate in the stratigraphic interval that records the highly negative carbon isotope values contains calcarenite layers that are interpreted to have been transported to the outer slope by

turbidity currents (Thomson et al., 2014; Fig. 2). The allochems are encased in carbonate cement, precipitated *in situ* during early diagenesis (Fig. 20e). We were unable to analyze the allochems and cement separately, but assuming that the carbonate sand was sourced from shallower settings, it likely records the global seawater signature, while the cement precipitated *in situ* likely records local pore-water chemistry.

4.5.2.5 Synthesis: Interpretation for the Amundsen Basin

The negative carbon isotope anomaly in the Upper Carbonate member of the Wynniatt Formation far exceeds that in other sections recording the Bitter Springs stage, and therefore requires a local mechanism for amplification. We interpret the negative C

Figure 24. Schematic diagram representing our interpretation of events that led to the negative carbon isotope excursion in the Wynniatt Formation, and how they relate to global tectonics, sea-level changes, and basin-scale to global seawater composition. Before the Bitter Springs stage waters were oxygenated, there was a large seawater sulphate reservoir, and high relative rates of organic carbon burial resulted in a positive $\delta^{13}\text{C}_{\text{carb}}$ excursion and buildup of methane clathrates in the sediments. During the Bitter Springs stage, a transgression occurred in the Amundsen Basin as Laurentia rotated, positioning the basin at equatorial latitudes and increasing local productivity. At the same time, basin restriction resulted in local anoxic to euxinic conditions. Methane clathrates reacted with the seawater oxidants, and contributed light carbon to DIC pool. This signal was further amplified in pore waters, where methane and organic matter were oxidized driving precipitation of authigenic carbonate cement with highly negative carbon isotope values. After the Bitter Springs stage, settings returned to previous conditions. TWP – true polar wander, py – pyrite, VPDB – Vienna Pee Dee Belemnite.

isotope excursion as follows (see Fig. 24). During deposition of the Upper Carbonate member, the Amundsen Basin became partially restricted from the open ocean (Thomson et al., 2014). However, the well-mixed, oxygenated surface waters likely were still in open communication with the ocean, and sediments precipitated in shallow-water environments would have recorded the C isotope composition of the global DIC pool. High relative burial rates of organic carbon that led to the sustained positive $\delta^{13}\text{C}$ values of seawater recorded in the underlying carbonate rocks of the Shaler Supergroup would also have facilitated build-up of methane in sediments. A proposed true polar wander event at ca. 800 Ma (Maloof et al., 2006) rotated Laurentia closer to the equator, resulting in a transgression as the continent moved through the equatorial bulge, and increased productivity in the Amundsen Basin. Continents that were transported to high latitudes experienced destabilization of methane clathrates along their margins in association with relative sea-level fall (Fig. 24). The released methane would have been oxidized by a large seawater sulphate reservoir accumulated by that time (Prince, 2014), shifting marine $\delta^{13}\text{C}_{\text{DIC}}$ composition to negative values, and causing the significant reduction in ocean sulphate concentration that occurred between deposition of the Minto Inlet and Kilian formations (Prince, 2014). In deep-water environments (below storm wave-base), isotopically light carbon in pore waters was derived from either: reaction of methane with sulphate and iron oxyhydroxides; and/or bacterially mediated oxidation of organic matter, precipitated during early diagenesis as authigenic carbonate cement with highly negative $\delta^{13}\text{C}_{\text{carb}}$ values. The authigenic cement amplified the already light $\delta^{13}\text{C}_{\text{carb}}$ signature preserved in the carbonate sand, resulting in the highly negative $\delta^{13}\text{C}_{\text{carb}}$ values observed. Further supporting this argument is the contrast between $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ values.

There is a 10‰ shift in $\delta^{13}\text{C}_{\text{org}}$ values through the Bitter Springs stage, from -25‰ to -35‰ (Fig. 21), which is the same amplitude of change observed globally in the $\delta^{13}\text{C}_{\text{carb}}$ values (values drop from $\sim +5\text{‰}$ to $\sim -5\text{‰}$), and therefore what one would expect to observe in the $\delta^{13}\text{C}_{\text{org}}$ values as well if they are fractionated from global DIC. However, $\delta^{13}\text{C}_{\text{carb}}$ values record a 24‰ shift (+5 to -14, Fig. 21) in the Amundsen Basin, supporting the interpretation that $\delta^{13}\text{C}_{\text{carb}}$ values record local basin effects, where authigenic carbonate is precipitated during early diagenesis.

4.5.3 Euxinia and redox-sensitive trace metal enrichments in the Amundsen Basin

Set in an intracontinental setting, the Amundsen Basin was intermittently restricted from the open ocean, notably during deposition of the Minto Inlet and Kilian formations (Fig. 1). There is physical evidence for restriction in the Upper Carbonate member of the Wynniatt Formation, implied by the preservation of lathe-shaped gypsum crystals, dolomite pseudomorphs of evaporative crusts, and the presence of a Mg-silicate (likely talc) mineral (Thomson et al., 2014). However, despite this evidence for restricted (evaporative) conditions, the overall depositional environment of the Upper Carbonate member of the Wynniatt Formation is interpreted as deep-water and below storm-wave base (Thomson et al., 2014). Strong evidence from Fe proxies for euxinic conditions is not surprising given the partially restricted setting; euxinia would have been favoured by water mass isolation, potentially promoting estuarine-style circulation.

The Mo, V, and U enrichments in the Upper Carbonate member of the Wynniatt Formation (Figs. 21 and 22) reflect paleoredox conditions in the Amundsen Basin as well

as the global ocean redox state. As noted above, Mo, V, and U enrichments well above crustal levels also provide evidence for locally euxinic conditions. The magnitude of Mo enrichment in euxinic facies is controlled largely by the global extent of euxinic conditions (see Reinhard et al., 2013, and reference within). Therefore, Mo/TOC values comparable to or higher than those typical of the mid-Proterozoic rule out the possibility that there was more extensive euxinia during the pre-glacial Cryogenian compared to the rest of the Proterozoic; a scenario that was previously favoured based on Mo concentrations and isotope data from the roughly coeval (~800 - 750 Ma) rift deposits of the Chuar Group (Dahl et al., 2011; Reinhard et al., 2013), containing euxinic shales with lower Mo enrichments (<12 ppm). Indeed, our Mo/TOC ratios fall between mean values for the mid-Proterozoic and Phanerozoic, suggesting that the time interval leading up to the onset of widespread Neoproterozoic glaciations was a transitional period in Earth's redox evolution, during which oceans were likely less reducing than during the preceding billion years. The possibility that the large Mo enrichments have local rather than global significance and are due to basin restriction can be ruled out, as it is well established that basin restriction reduces Mo enrichment (Algeo and Lyons, 2006; Lyons et al., 2009). However, Mo drawdown in a restricted, isolated rift basin can explain the contrast between the Shaler Supergroup and the Chuar Group Mo records.

The U and V enrichments (Fig. 21) provide additional support for the idea that the pre-glacial Cryogenian represents a transitional period in Earth's redox evolution. Low U enrichments and U/TOC ratios relative to the Phanerozoic anoxic shale mean (Partin et al., 2013), but high V and V/TOC ratios relative to the Phanerozoic anoxic shale mean (Sahoo et al., 2012) are unusual. These elements respond similarly to anoxic marine

conditions, but the modern marine residence and response times for U are much longer than for V (see mass balance modelling in Sahoo et al., 2012, and Partin et al., 2013). Therefore, we would expect that in a semi-restricted basin, like the Amundsen Basin, V reservoir drawdown would precede U drawdown, resulting in substantial U enrichment relative to V in anoxic shales, the opposite of what we observe in the Upper Carbonate member of the Wynniatt Formation. A potential explanation for these U-V enrichment trends may lie in the fact that suboxic marine sediments can be a source of V to oceans (Hastings et al., 1996). Suboxic sediments, for this case, are defined as those where oxygen is depleted in the uppermost 1 cm of the substrate and there is no surface solid phase Mn maximum. These sediments are likely overlain by a weakly oxic water column. The V in these sediments is sourced from particulates, primarily metal oxides (see Hastings et al., 1996 for discussion). In a weakly oxidizing ocean, this could have led to a global-scale build-up of the dissolved V reservoir, and thus the potential for large sedimentary V enrichment. Building from this framework, we propose that there may have been a shift in the pre-glacial Cryogenian to an ocean with an expansive, weakly oxic water column.

Persistently light $\delta^{34}\text{S}_{\text{pyrite}}$ values in the euxinic section of the Upper Carbonate member of the Wynniatt Formation suggest that the pre-glacial Cryogenian was marked by a growth in marine sulphate reservoir. Although there is a large dataset of sulphur isotope values for Proterozoic sedimentary pyrite (e.g. Canfield and Raiswell, 1999), euxinic shales do not typically show large ranges of $\Delta^{34}\text{S}$ values ($\Delta^{34}\text{S} < 30\text{‰}$) prior to 635 Ma (Sahoo et al., 2012; Scott et al., 2014). Scott et al. (2014) linked this shift in shale sulphur isotope systematics to growth of the marine sulphur reservoir. Thick sulphate

evaporites in the Minto Inlet Formation and elsewhere globally, for example the Ten Stone Formation, Mackenzie Mountains (Turner and Long, 2012), and the Gillen Member of the Bitter Springs Formation, Amadeus Basin (Lindsay, 1987), occur immediately below sediments recording the Bitter Springs Stage, adding support to the notion that there was a significant seawater sulphate reservoir at that time (Prince, 2014). This is consistent with lower pyrite burial rates under generally more oxidizing conditions in the global ocean.

4.6 Conclusions

- A composite carbon isotope curve with paired $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ values is presented for the upper Shaler Supergroup. $\delta^{13}\text{C}_{\text{carb}}$ values range from -14.3 to +7.7‰, with the average of +3.0‰, while $\delta^{13}\text{C}_{\text{org}}$ values range from -34.2 to -14.4‰, with the average of -25.9‰.
- Basin-scale variations in $\delta^{13}\text{C}_{\text{carb}}$ values for the upper Shaler Supergroup (~900 – 720 Ma) make detailed chemostratigraphic correlation with stratigraphic sections in the northern Cordillera difficult. The Boot Inlet Formation may correlate with the lower Stone Knife/Silverberry formations of the Mackenzie Mountains Supergroup and the Gibben Formation of the Fifteenmile Group, supporting previous litho- and sequence stratigraphic correlations. Our data also supports previous correlations of the Kilian Formation with the Coates Lake Group, also considered correlative to the Callison Lake dolostone. The C isotope curve appears to be compressed in our study area when compared to those for the

Fifteenmile Group and Mackenzie Mountains Supergroup, likely due to lower sedimentation rates in the Amundsen Basin.

- The Bitter Springs stage is present in the Wynniatt Formation of the Shaler Supergroup; however, $\delta^{13}\text{C}$ values are much lighter than those described from coeval sedimentary successions. Therefore, these values require a local amplification, perhaps reflecting precipitation of isotopically light carbon in pore waters derived from reactions of methane and organic carbon with sulphate and ferric iron, resulting in precipitation of authigenic carbonate cement with highly negative $\delta^{13}\text{C}_{\text{carb}}$ values.
- The majority of the studied interval includes sediments likely to have been deposited in oxic settings; however, redox proxies of a carbonaceous shale interval at the base of the Upper Carbonate member of the Wynniatt Formation indicate deposition under both euxinic and anoxic (ferruginous) conditions, with high concentrations of Mo, V, and U. In these settings, Mo and V were sequestered and buried, but the global ocean must have been oxygenated before the Bitter Springs stage to allow for buildup of the large dissolved Mo and V reservoirs. This new data suggests that a shift to more oxygenated marine conditions preceded the oldest Cryogenian glaciation.

CHAPTER 5: CONCLUSIONS

5.1 Conclusions

The main conclusions of this study are:

1. The Wynniatt Formation can be subdivided into four new members, which are recognized and described.
2. The Wynniatt Formation is interpreted to represent a succession of three high-energy, distally steepened carbonate ramps. Sedimentation was predominantly controlled by storm-generated processes, with minor reworking by wave and tidal currents. Isopach maps provide support that the Wynniatt Formation accumulated in a circular basin. Inferred changes in the position of regional depocentres are associated with each stage of ramp development.
3. Five unconformity-bounded, transgressive-regressive (T-R) cycles are recognized through a succession that includes, in ascending stratigraphic order, the Jago Bay, Minto Inlet, and Wynniatt formations. T-R cycles are difficult to correlate regionally because accommodation was affected by local tectonically controlled subsidence (e.g. Fifteenmile basin). The exception is sequence S7 from the Fifteenmile Group and the coeval T-R5 defined here from the Shaler Supergroup, which reflects a significant regional basin-deepening event that may relate to the onset of a global marine transgression.

4. The upper Shaler Supergroup (Boot Inlet to Kilian formations) and Little Dal Group (Mackenzie Mountains Supergroup) are correlative with the Fifteenmile Group in its entirety.
5. Three tectonostratigraphic units are recognized in the upper Shaler Supergroup. Stage 1 records subsidence of a sag basin, Stage 2 preserves evidence of syndepositional faulting and thermal doming, and Stage 3 represents the rifting phase that concluded with extrusion of a thick succession of continental flood basalts.
6. Subsidence during deposition of the middle to upper Shaler Supergroup in the Amundsen Basin may have been influenced by lithospheric stretching and thinning, if northwest Laurentia had passed over a mantle plume at 830 Ma, or by intra-plate stresses related to some other cause of extension, which formed fault-bound rift basins in the Mackenzie Mountains and Fifteenmile basins. Inferred changes in subsidence patterns in the Amundsen Basin are likely the result of far-field extensional stresses, which were related to more pronounced extension that manifested as fault-related sedimentation in coeval basins located closer to the paleomargin of Laurentia.
7. $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ composite secular curves for the upper Shaler Supergroup (~900 – 720 Ma) are presented. Detailed chemostratigraphic correlation with adjacent successions is tenuous, complicated by lower sedimentation rates in the Amundsen Basin, resulting in condensation of the C isotope curve.
8. The global Bitter Springs stage negative $\delta^{13}\text{C}$ anomaly is recorded in the Wynniatt Formation; however, $\delta^{13}\text{C}$ values are much lighter than those

measured from coeval sedimentary successions. The more negative values in the Wynniatt Formation are interpreted to reflect precipitation of isotopically light carbon in pore waters derived from reaction of methane and organic carbon with sulphate and ferric Fe. This resulted in precipitation of authigenic carbonate cement with highly negative $\delta^{13}\text{C}_{\text{carb}}$ values, locally amplifying negative values.

9. The majority of the upper Shaler Supergroup includes sediments likely to have been deposited in oxic settings; however, redox proxies in the Upper Carbonate member of the Wynniatt Formation indicate deposition under both euxinic and anoxic (ferruginous) conditions, with high concentrations of Mo, V, and U. The global ocean must have been oxygenated in phase with the Bitter Springs stage to allow for build-up of the large dissolved Mo and V reservoirs, suggesting there was a shift to more oxygenated marine conditions in the interval preceding the oldest Cryogenian glaciation.

5.2 Future Research

There are many questions that could still be addressed, including:

- Characterization of sequence or T-R stratigraphy of the correlative Little Dal Group in the Mackenzie Mountains, to see if sequences there can be correlated with the ones defined in this study.
- If subsidence in the upper Shaler Supergroup is related to lithospheric thinning signaling the onset of Rodinia breakup, what are the subsidence

mechanisms that created accommodation for deposition of the lower Shaler Supergroup?

- Additional C-isotope study of the turbidite layers within the shale that preserves the negative carbon excursion in the Upper Carbonate member of the Wynniatt Formation. This study predicts that the isotopic values of the cement compared to the allochems would be significantly different.
- Further micropaleontological studies are suggested. There are other known localities of scale and vase-shaped microfossils in successions of the same age globally, however the previous study of microfossils from the Shaler Supergroup was limited to the uppermost Black Shale member. This study predicts that there is good potential for diversification of life at this time, if it is linked to oxygenation of the oceans, as is observed here.

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Appendixes

<i>Section</i>	<i>Latitude</i>	<i>Longitude</i>	<i>Notes</i>	<i>GSC station number</i>
Outcrop Sections				
10DTS1	71.322	-115.5826	Minto Inlet Fm.	10RAT-DT022
10DTS2	71.2602	-116.0748	Minto Inlet Fm.	10RAT-DT024
10DTS3	71.4418	-116.0118		10RAT-DT027
			lower carbonate member	
10DTS4	71.4219	-116.0693	black shale member	10RAT-DT030
10DTS5	71.4761	-115.5157	stromatolitic carbonate member	10RAT-DT055
10DTS6	71.3795	-115.1603		10RAT-DT057
			upper carbonate member	
10DTS7	71.4453	-115.2464	stromatolitic carbonate member	10RAT-DT058
10DTS8	71.4374	-115.2262		10RAT-DT060
			upper carbonate member, west side of river	
10DTS9	71.4265	-115.1988		10RAT-DT062
			upper carbonate member, east side of river	
10DTS10	72.1982	-111.7301	stromatolitic carbonate member	10RAT-DT074
10DTS11	72.2209	-111.6861	lower carbonate and black shale members	10RAT-DT076
10DTS12	71.3772	-115.1588		10RAT-DT105
			upper carbonate member	
11DTS2	71.3431	-115.5731	upper carbonate member, SW side of hill	11RAT-DT029
11DTS3	71.3376	-115.5552	upper carbonate member, SE side of hill	11RAT-DT034
11DTS4	71.3376	-115.5552		11RAT-DT034
			upper carbonate member, south side of hill	
11DTS5	71.1725	-117.1078		11RAT-DT102
			upper carbonate member	
11DTS6	71.4515	-117.1434	Fort Collinson Fm.	11RAT-DT127
11DTS7	71.3241	-116.4923		11RAT-DT129
			upper carbonate member	
11DTS8	71.4521	-117.1236	Jago Bay Fm.	11RAT-DT147
11DTS9	72.1222	-112.2748	lower carbonate, black shale, and stromatolitic carbonate members	11RAT-DT168
Drill core				
GNME 06-03	71.6282	-115.8962	Minto Inlet Fm. and lower carbonate member, Wynniatt Fm.	10RAT-DT111
GNME 07-04	71.7286	-113.6019		11RAT-DT095
			upper carbonate Member	

Appendix 1. Locations and descriptions of measured outcrop sections and drill core.

Appendix 2. Stable isotope data

Sample- A series	Sample -B series	Fm/Mb	m (measured)	m (base Boot Inlet)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT022A1		Minto	1	472.7					
10RAT-DT022A2		Minto	19	490.7					
10RAT-DT022A3		Minto	20	491.7					
10RAT-DT022A4		Minto	22	493.7	3.94	-4.52			
10RAT-DT022A5		Minto	26.5	498.2	3.90	-9.02			
10RAT-DT022A6		Minto	66	537.7	3.23	-8.50			
10RAT-DT022A7		Minto	67.2	538.9	3.40	-8.71			
10RAT-DT024A1		Minto	2.8	599.5					
10RAT-DT024A2		Minto	5	601.7					
10RAT-DT024A3		Minto	17.5	614.2					
10RAT-DT024A4		Minto	21.1	617.8	0.45	-8.18			
10RAT-DT024A5		Minto	26.4	623.1	1.67	-9.13			
10RAT-DT024A6		Minto	27	623.7					
10RAT-DT024A7		Minto	27.4	624.1	2.32	-8.08			
10RAT-DT024A8		Minto	28	624.7	0.72	-8.74			
10RAT-DT024A9		Minto	30.4	627.1	4.14	-6.66			
10RAT-DT024A10		Minto	31.9	628.6	3.48	-6.05			
10RAT-DT024A11		Minto	32.9	629.6	2.93	-7.73			
10RAT-DT024A12		Minto	34	630.7	3.22	-6.27			
10RAT-DT024A13		Minto	34.9	631.6	3.49	-4.67			
10RAT-DT024A14		Minto	36.1	632.8	3.53	-7.45			
10RAT-DT024A15		Minto	36.7	633.4	3.14	-8.53			
10RAT-DT024A16		Minto	37.7	634.4	1.31	-8.67			
10RAT-DT024A17		Minto	38.6	635.3	1.28	-8.13			
10RAT-DT024A18		Minto	40.9	637.6					
10RAT-DT024A19		Minto	43	639.7	-1.33	-11.30			
10RAT-DT027A1		lcm	1	657.7	4.51	-8.89			
10RAT-DT027A2		lcm	7.4	664.1	4.94	-6.01	-24.40	0.20	
10RAT-DT027A3		lcm	16	672.7	3.27	-3.97			
10RAT-DT027A4		lcm	17	673.7	2.66	-4.28			
10RAT-DT027A5		lcm	20	676.7	4.62	-6.65			
10RAT-DT027A6		lcm	25.3	682	3.72	-5.60			
10RAT-DT027A7		lcm	37.5	694.2	3.23	-5.53			
10RAT-DT027A8		lcm	41	697.7	3.24	-5.53			
10RAT-DT027A9		lcm	45.5	702.2	3.99	-8.00			
10RAT-DT027A10		lcm	46.5	703.2	3.38	-3.36			
10RAT-DT027A11		lcm	53	709.7	4.48	-8.93	-25.00	0.43	-4.00
10RAT-DT027A12		lcm	57.1	713.8	4.94	-8.61	-22.70	0.13	
10RAT-DT027A13		lcm	61.1	717.8	5.12	-4.77	-24.70	0.22	
10RAT-DT027A14		lcm	66.1	722.8	5.33	-6.47			
10RAT-DT027A15		lcm	67.8	724.5	4.66	-6.93	-24.70	0.27	
10RAT-DT027A16		lcm	72.7	729.4	4.84	-4.68			
10RAT-DT027A17		lcm	79.4	736.1	4.84	-3.65			
10RAT-DT027A19		lcm	91.6	748.3	4.40	-9.41	-23.20	1.02	
10RAT-DT027A20		lcm	99.15	755.85	5.34	-7.37			
10RAT-DT027A21		lcm	107.5	764.2	3.14	-7.43			
10RAT-DT027A22		lcm	113.2	769.9	2.98	-8.13			
10RAT-DT027A23		lcm	118	774.7	4.11	-9.56			
10RAT-DT027A24		lcm	124	780.7	4.62	-8.23			
10RAT-DT030A1		bsm	0.5	781.2	4.29	-9.60			

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT030A4		bsm	8.8	789.5	2.65	-8.72			
10RAT-DT030A5		bsm	11.4	792.1	0.84	-9.82			
10RAT-DT030A6		bsm	12.4	793.1	0.99	-8.89	-24.60	0.15	
10RAT-DT030A7		bsm	18	798.7	-0.01	-7.71	-25.90	0.14	13.20
10RAT-DT030A8		bsm	23	803.7	0.12	-8.63			
10RAT-DT030A9		bsm	26.1	806.8	-0.56	-8.89			
10RAT-DT030A10		bsm	54	834.7					
10RAT-DT030A11		bsm	62.5	843.2	-3.48	-9.69			
10RAT-DT030A12		bsm	73.5	854.2			-21.70	0.11	10.60
10RAT-DT030A13		bsm	89	869.7					
10RAT-DT030A14		bsm	93	873.7					
10RAT-DT052A1		bsm	5.7	932.4					
10RAT-DT052A2		bsm	40.2	966.9	2.64	-9.22			
10RAT-DT052A3		bsm	43.6	970.3	3.25	-4.54			
10RAT-DT052A4		bsm	45.9	972.6					
10RAT-DT052A5		bsm	46.8	973.5	2.56	-6.90			
10RAT-DT055A1		bsm	1	908.4					
10RAT-DT055A2		bsm	53	960.4					
10RAT-DT055A3		bsm	68	975.4	2.18	-8.78			
10RAT-DT055A4		bsm	67.5	974.9					
10RAT-DT055A5		scm	71.8	979.2	3.72	-5.63			
10RAT-DT055A6		scm	73.5	980.9	4.34	-4.97			
10RAT-DT055A7		scm	75.2	982.6	4.03	-4.58			
10RAT-DT055A8		scm	75.3	982.7					
10RAT-DT055A9		scm	77	984.4					
10RAT-DT055A10		scm	80	987.4	4.29	-3.49			
10RAT-DT055A11		scm	85.4	992.8	4.13	-8.98			
10RAT-DT055A12		scm	87.2	994.6					
10RAT-DT055A13		scm	87.7	995.1	4.60	-8.33			
10RAT-DT055A14		scm	90.9	998.3	2.54	-8.88			
10RAT-DT055A15		scm	92.5	999.9	6.97	-7.60			
10RAT-DT055A16		scm	97.1	1004.5	4.77	-8.65			
10RAT-DT057A1		ucm	0.5	1375.4					
10RAT-DT057A2		ucm	1.6	1376.5	4.40	-5.02			
10RAT-DT057A3		ucm	7.8	1382.7	5.52	-5.26	-23.94	0.67	
10RAT-DT057A4		ucm	9.8	1384.7	5.13	-5.32			
10RAT-DT057A5		ucm	16.6	1391.5	4.28	-6.98			
10RAT-DT057A6		ucm	17	1391.9					
10RAT-DT057A8		ucm	25.6	1400.5	3.96	-6.15			
10RAT-DT057A9		ucm	25.75	1400.65	2.41	-6.09			
10RAT-DT057A10		ucm	27.9	1402.8	4.07	-7.70	-26.29	0.19	
10RAT-DT057A11		ucm	29.5	1404.4					
10RAT-DT057A12		ucm	30.6	1405.5	3.25	-5.02			
10RAT-DT057A13		ucm	35.9	1410.8	3.00	-6.22			
10RAT-DT057A14		ucm	41.8	1416.7	2.83	-7.50			
10RAT-DT058A1		scm	1	975.9	2.38	-9.30			
10RAT-DT058A2		scm	10.6	985.5	4.04	-4.73			
10RAT-DT058A3		scm	15.1	990					
10RAT-DT058A4		scm	18.3	993.2	5.60	-8.83			
10RAT-DT058A5		scm	23	997.9	6.36	-7.31			
10RAT-DT058A6		scm	26	1000.9	5.64	-7.33			
10RAT-DT058A7		scm	37.6	1012.5	7.67	-8.03			
10RAT-DT058A8		scm	44.1	1019	7.16	-8.71			
10RAT-DT058A9		scm	47.2	1022.1	6.49	-8.10	-24.27	0.27	-12.60
10RAT-DT058A10		scm	55	1029.9	4.51	-7.69			
10RAT-DT058A11		scm	57.8	1032.7	4.99	-7.97			

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT058A13		scm	70.9	1045.8	4.70	-8.59			
10RAT-DT058A14		scm	80.9	1055.8	5.75	-7.27			
10RAT-DT058A15		scm	99.9	1074.8	4.37	-8.76			
10RAT-DT060A1		scm	1.3	1081.2	4.42	-8.14			
10RAT-DT060A2		scm	7.5	1087.4	5.01	-8.88			
10RAT-DT060A3		scm	13	1092.9	5.10	-7.60			
10RAT-DT060A4		scm	17	1096.9	4.98	-8.06	-25.85	0.53	
10RAT-DT060A5		scm	25.8	1105.7	5.86	-8.26			
10RAT-DT060A6		scm	30	1109.9	6.44	-8.22			
10RAT-DT060A7		scm	35.9	1115.8	6.34	-8.00			
10RAT-DT060A8		scm	40	1119.9	6.29	-8.28			
10RAT-DT060A9		scm	43.5	1123.4	5.37	-7.10			
10RAT-DT060A10		scm	49.5	1129.4	5.92	-2.74			
10RAT-DT060A11		scm	52.6	1132.5	5.20	-5.06			
10RAT-DT060A19		scm	53.1	1133	4.85	-3.99			
10RAT-DT060A20		scm	53.6	1133.5	5.12	-4.16			
10RAT-DT060A21		scm	54.1	1134	4.98	-4.46			
10RAT-DT060A22		scm	54.6	1134.5	4.39	-5.38			
10RAT-DT060A23		scm	55.1	1135	4.62	-5.40	-25.38	0.75	-18.40
10RAT-DT060A24		scm	55.6	1135.5	3.94	-4.35			
10RAT-DT060A25		scm	56.1	1136	1.10	-5.13			
10RAT-DT060A26		ucm	56.3	1136.2	0.77	-7.37			
10RAT-DT060A27		ucm	56.6	1136.5	1.29	-7.97			
10RAT-DT060A12		ucm	57	1136.9	-0.12	-5.14			
10RAT-DT060A28		ucm	57.3	1137.2	-0.18	-4.23			
10RAT-DT060A29		ucm	57.6	1137.5	-0.16	-5.78			
10RAT-DT060A30		ucm	58	1137.9	-1.19	-6.71			
10RAT-DT060A31		ucm	58.2	1138.1					
10RAT-DT060A33		ucm	60.2	1140.1	-2.85	-7.94			
10RAT-DT060A34		ucm	62.2	1142.1	-2.06	-9.16			
10RAT-DT060A13		ucm	65.7	1145.60	-5.50	-9.38			
10RAT-DT060A14		ucm	79	1158.9					
10RAT-DT060A15		ucm	80	1159.9					
10RAT-DT060A16		ucm	85	1164.9	-5.56	-8.32			
10RAT-DT060A17		ucm	97.1	1177					
10RAT-DT060A18		ucm	97.1	1177					
10RAT-DT062A1		ucm	3	1217.9	3.62	-7.66			
10RAT-DT062A2		ucm	10.8	1225.7					
10RAT-DT062A3		ucm	11.35	1226.25					
10RAT-DT062A4		ucm	10.2	1225.1					
10RAT-DT062A5		ucm	12.4	1227.3	2.30	-6.50			
10RAT-DT062A6		ucm	18	1232.9	1.94	-7.03			
10RAT-DT062A7		ucm	23	1237.9	5.71	-6.12	-24.12	0.37	
10RAT-DT062A8		ucm	28.5	1243.4	5.51	-5.92	-22.21	0.74	12.5
10RAT-DT062A9		ucm	42.5	1257.4	5.14	-8.39			
10RAT-DT062A10		ucm	60	1274.9	6.22	-6.33			
10RAT-DT062A11		ucm	68	1282.9	5.83	-7.48			
10RAT-DT062A12		ucm	82	1296.9	6.21	-6.84			
10RAT-DT062A13		ucm	93	1307.9	5.72	-8.35			
10RAT-DT074A1		scm	9	973.6					
10RAT-DT074A2		scm	10.3	974.9	0.58	-9.28			
10RAT-DT074A3		scm	15	979.6					
10RAT-DT074A4		scm	17.6	982.2					
10RAT-DT074A5		scm	22	986.6	6.76	-6.29			
10RAT-DT074A6		scm	26.4	991	5.98	-6.31	-25.99	0.42	-13.5
10RAT-DT074A7		scm	32.1	996.7	5.75	-5.38	-23.69	0.16	

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT074A9		scm	51.5	1016.1	6.36	-6.03			
10RAT-DT074A10		scm	58	1022.6	6.48	-5.62			
10RAT-DT074A11		scm	63	1027.6	5.08	-7.42			
10RAT-DT074A12		scm	68	1032.6	5.14	-6.98	-25.43	0.20	
10RAT-DT074A13		scm	72.8	1037.4	5.55	-6.86	-24.57	0.87	-13.5
10RAT-DT074A14		scm	77.8	1042.4	4.84	-6.80	-25.31	0.14	
10RAT-DT074A15		scm	85.9	1050.5	4.64	-8.73			
10RAT-DT074A16		scm	87.8	1052.4	4.46	-9.75			
10RAT-DT074A17		scm	90.3	1054.9	5.08	-8.16	-26.03	0.22	
10RAT-DT074A18		scm	94.8	1059.4	5.16	-7.88			
10RAT-DT074A19		scm	101	1065.6	5.62	-6.33			
10RAT-DT076A1		lcm	0.3	740	5.55	-5.87			
10RAT-DT076A2		lcm	1.1	740.8	4.94	-6.93			
10RAT-DT076A3		lcm	5	744.7	4.64	-6.13			
10RAT-DT076A4		lcm	10.5	750.2	4.90	-6.98			
10RAT-DT076A5		lcm	15	754.7	4.82	-7.86	-25.33	0.07	
10RAT-DT076A6		lcm	20	759.7	3.71	-5.73	-23.88	0.06	
10RAT-DT076A7		lcm	36.4	776.1	3.86	-7.60			
10RAT-DT076A9		lcm	44.7	784.4	2.58	-9.17			
10RAT-DT105A1		ucm	0.3	1417	2.31	-5.83			
10RAT-DT105A2		ucm	4.7	1421.4	3.71	-8.04			
10RAT-DT105A3		ucm	6.2	1422.9					
10RAT-DT105A4		ucm	11.4	1428.1	4.40	-7.25			
10RAT-DT105A5		ucm	17	1433.7	4.37	-6.51			
10RAT-DT105A6		ucm	18.8	1435.5	3.55	-6.07			
10RAT-DT111A1		Minto	15.8	683.4	5.14	-8.44			
10RAT-DT111A2		Minto	17.9	681.3	4.86	-9.41			
10RAT-DT111A3		Minto	20.1	679.1	4.96	-9.61			
10RAT-DT111A4		Minto	22.1	677.1	5.01	-8.52			
10RAT-DT111A5	10RAT-DT111B1	Minto	22.5	676.7	4.94	-8.93	-21.90	0.13	
10RAT-DT111A6	10RAT-DT111B2	Minto	23.2	676	4.37	-6.89	-22.10	0.01	
10RAT-DT111A7		Minto	24.1	675.1	4.51	-7.58			
10RAT-DT111A8		Minto	25.8	673.4	2.96	-10.80			
10RAT-DT111A9		Minto	34.6	664.6	3.04	-7.26			
10RAT-DT111A10		Minto	36.4	662.8	1.51	-10.70			
10RAT-DT111A11	10RAT-DT111B3	Minto	36.6	662.6	1.31	-11.33	-23.60	0.72	-9.9
10RAT-DT111A12		Minto	37	662.2	1.59	-11.39			
10RAT-DT111A13		Minto	38.9	660.3	2.04	-6.90			
10RAT-DT111A14		Minto	40.9	658.3	1.60	-8.97			
10RAT-DT111A15		Minto	46.5	652.7	1.70	-10.55			
10RAT-DT111A16		Minto	48.5	650.7	2.53	-7.30			
10RAT-DT111A17		Minto	50.4	648.8	3.34	-5.56			
10RAT-DT111A18		Minto	52.6	646.6	2.55	-9.33			
10RAT-DT111A19		Minto	54.4	644.8	3.28	-7.29			
10RAT-DT111A20		Minto	56.3	642.9	3.92	-6.94			
10RAT-DT111A21		Minto	58.5	640.7	3.18	-11.52			
10RAT-DT111A22		Minto	60.6	638.6	3.30	-10.72			
10RAT-DT111A23		Minto	62.4	636.8	3.85	-6.90			
10RAT-DT111A24		Minto	64.3	634.9	3.62	-7.60			
10RAT-DT111A25		Minto	67.5	631.7	3.29	-8.47			
10RAT-DT111A26		Minto	69.5	629.7	4.06	-6.49			
10RAT-DT111A27		Minto	70.9	628.3	3.57	-6.58			
10RAT-DT111A28		Minto	72.9	626.3	3.54	-6.25			
10RAT-DT111A29		Minto	75	624.2	3.41	-6.29			
10RAT-DT111A30		Minto	77.1	622.1	2.00	-11.39			
10RAT-DT111A31		Minto	79.1	620.1	2.35	-11.57			

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT111A33		Minto	84.5	614.7	2.29	-11.98			
10RAT-DT111A34		Minto	86.4	612.8					
10RAT-DT111A35		Minto	88.4	610.8	2.10	-12.36			
10RAT-DT111A36		Minto	90.4	608.8					
10RAT-DT111A37		Minto	92.3	606.9	1.68	-12.51			
10RAT-DT111A38		Minto	94.1	605.1	2.26	-13.74			
10RAT-DT111A39		Minto	95.85	603.35	3.66	-12.09			
10RAT-DT111A40		Minto	195.4	503.8	3.71	-11.97			
10RAT-DT111A42		Minto	201.1	498.1	2.78	-12.24			
10RAT-DT111A43	10RAT-DT111B4	Minto	201.4	497.8	3.50	-11.55	-24.22	0.10	
10RAT-DT111A44		Minto	203	496.2	3.46	-11.98			
10RAT-DT111A45		Minto	204.85	494.35	3.23	-12.36			
10RAT-DT111A46		Minto	206.9	492.3	3.74	-10.01			
10RAT-DT111A47		Minto	209	490.2	3.19	-8.21			
10RAT-DT111A48		Minto	211	488.2	3.62	-7.97			
10RAT-DT111A49		Minto	212.9	486.3	2.97	-9.77			
10RAT-DT111A50		Minto	215.2	484	4.11	-8.26			
10RAT-DT111A51		Minto	217.1	482.1	3.64	-8.03			
10RAT-DT111A52		Minto	218.9	480.3	4.04	-6.73			
10RAT-DT111A53		Minto	221	478.2	2.66	-8.04			
10RAT-DT111A54		Minto	223.1	476.1	1.75	-8.62			
10RAT-DT111A55		Minto	224.9	474.3	2.91	-10.77			
10RAT-DT111A56		Minto	226.8	472.4	3.78	-8.64			
10RAT-DT111A57		Jago	256.8	442.4	3.46	-10.49			
10RAT-DT111A58		Jago	258.7	440.5	3.14	-10.13			
10RAT-DT111A59		Jago	261	438.2	3.06	-12.88			
10RAT-DT111A60		Jago	263	436.2	2.72	-12.79			
10RAT-DT111A61		Jago	265.1	434.1	3.14	-12.83			
10RAT-DT111A62		Jago	267.1	432.1	2.75	-10.32			
10RAT-DT111A63		Jago	269.1	430.1	3.00	-12.71			
10RAT-DT111A64		Jago	271.2	428	3.17	-7.69			
10RAT-DT111A65		Jago	272.6	426.6	4.25	-8.22			
10RAT-DT111A66		Jago	273.7	425.5	4.44	-8.27			
10RAT-DT111A67		Jago	274.9	424.3	3.96	-6.64			
10RAT-DT111A68		Jago	276.9	422.3	3.76	-8.61			
10RAT-DT111A69		Jago	277.7	421.5	4.02	-7.25	-26.50	0.12	
10RAT-DT111A70		Jago	278.2	421	4.78	-7.94			
10RAT-DT111A71		Jago	278.8	420.4	5.00	-7.84			
10RAT-DT111A72		Jago	282.1	417.1	4.41	-5.58			
10RAT-DT111A73	10RAT-DT111B5	Jago	281.1	418.1	4.37	-6.94			
10RAT-DT111A74		Jago	281.1	418.1	4.06	-6.92			
10RAT-DT111A75		Jago	283.7	415.5	4.44	-7.28			
10RAT-DT111A76		Jago	285.7	413.5	4.12	-7.42			
10RAT-DT111A77		Jago	287.4	411.8	3.96	-5.98			
10RAT-DT111A78		Jago	289.3	409.9			-25.00	0.23	
10RAT-DT111A79		Jago	291.4	407.8	3.92	-6.34			
10RAT-DT111A80	10RAT-DT111B6	Jago	291.8	407.4	4.33	-7.78	-24.71	0.17	
10RAT-DT111A81	10RAT-DT111B7	Jago	294.2	405	4.47	-7.97	-24.24	-0.03	
10RAT-DT111A82		Jago	293.3	405.9	4.26	-7.92			
10RAT-DT111A83		Jago	295.4	403.8	4.68	-7.61			
10RAT-DT111A84		Jago	296.4	402.8	4.33	-7.12			
10RAT-DT111A85	10RAT-DT111B8	Jago	298.2	401	4.27	-6.72	-24.58	0.36	-16
10RAT-DT111A86		Jago	299.2	400	4.19	-6.36			
10RAT-DT111A87	10RAT-DT111B9	Jago	300.3	398.9	4.47	-7.93	-22.73	-0.04	
10RAT-DT111A88		Jago	301.3	397.9			-26.66	0.21	-32.5

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT111A91		Jago	304.4	394.8	3.79	-7.97			
10RAT-DT111A92		Jago	305.8	393.4	3.77	-7.58			
10RAT-DT111A93		Jago	307.7	391.5	3.78	-7.91			
10RAT-DT111A94		Jago	309.5	389.7	5.73	-8.46			
10RAT-DT111A95	10RAT-DT111B11	Jago	310.5	388.7	5.45	-8.13	-23.32	0.11	
10RAT-DT111A96		Jago	311.5	387.7	4.80	-8.46			
10RAT-DT111A97	10RAT-DT111B12	Jago	312.8	386.4	4.70	-8.17	-23.54	-0.02	-29
10RAT-DT111A98		Jago	313.5	385.7	5.94	-8.22			
10RAT-DT111A99		Jago	315.5	383.7	4.98	-7.58			
10RAT-DT111A100	10RAT-DT111B13	Jago	316.1	383.1	5.23	-5.29	-28.12	0.23	
10RAT-DT111A101		Jago	316.5	382.7	5.54	-7.91			
10RAT-DT111A102		Jago	317.5	381.7	5.54	-7.99			
10RAT-DT111A103	10RAT-DT111B14	Jago	319.3	379.9	5.20	-4.95	-23.34	0.11	
10RAT-DT111A104		Jago	319.5	379.7	4.94	-7.06			
10RAT-DT111A105	10RAT-DT111B15	Jago	320.6	378.6	4.64	-7.66	-25.09	-0.01	
10RAT-DT111A106		Jago	321.4	377.8	4.73	-8.42			
10RAT-DT111A107	10RAT-DT111B16	Jago	321.8	377.4	4.22	-4.60	-25.09	0.19	-33.6
10RAT-DT111A108		Jago	323.6	375.6	3.48	-8.79			
10RAT-DT111A109	10RAT-DT111B17	Jago	324.6	374.6	2.61	-8.85	-23.35	0.23	
10RAT-DT111A110		Jago	328.1	371.1	3.24	-9.09			
10RAT-DT111A111		Jago	330.3	368.9	2.93	-7.03			
10RAT-DT111A112		Jago	332.2	367	3.02	-8.41			
10RAT-DT111A113	10RAT-DT111B18	Jago	332.8	366.4	3.65	-8.59	-26.76	0.18	-31.9
10RAT-DT111A114		Jago	334.2	365	3.79	-8.02			
10RAT-DT111A115	10RAT-DT111B19	Jago	334.9	364.3	3.97	-9.02	-22.19	0.28	
10RAT-DT111A116		Jago	336.3	362.9	3.77	-8.60			
10RAT-DT111A117		Jago	338.2	361	2.71	-8.85			
10RAT-DT111A118		Jago	339.3	359.9	2.82	-8.21			
10RAT-DT111A119	10RAT-DT111B20	Jago	340.1	359.1	3.25	-8.27	-24.97	-0.02	
10RAT-DT111A120		Jago	341.2	358	3.30	-8.62			
10RAT-DT111A121		Jago	342.4	356.8	2.74	-7.68			
10RAT-DT111A122		Jago	343.7	355.5	3.74	-8.90			
10RAT-DT111A123	10RAT-DT111B21	Jago	344.4	354.8	3.15	-6.63	-24.72	0.48	-1.8
10RAT-DT111A124		Jago	345.4	353.8	3.92	-9.13			
10RAT-DT111A125		Jago	346.2	353	4.27	-9.02			
10RAT-DT111A126	10RAT-DT111B22	Jago	348.4	350.8	5.04	-8.63	-26.10	0.01	
10RAT-DT111A127		Jago	348.7	350.5	5.36	-8.05			
10RAT-DT111A128	10RAT-DT111B23	Jago	349.3	349.9	5.57	-7.55	-27.28	0.12	
10RAT-DT111A129		Jago	350.5	348.7	4.71	-7.35			
10RAT-DT111A130		Jago	352.4	346.8	5.61	-7.56			
10RAT-DT111A131	10RAT-DT111B24	Jago	352.7	346.5	5.17	-7.61	-25.38	2.81	-1.2
10RAT-DT111A132	10RAT-DT111B25	Jago	354.2	345	5.32	-7.82	-27.42	0.31	
10RAT-DT111A133		Jago	354.5	344.7	5.89	-7.50			
10RAT-DT111A134	10RAT-DT111B26	Jago	355.7	343.5	5.94	-7.46	-26.99	0.26	-11.3
10RAT-DT111A135		Jago	356.8	342.4	5.75	-7.87			
10RAT-DT111A136	10RAT-DT111B27	Jago	358.1	341.1	5.63	-7.69	-26.09	0.14	
10RAT-DT111A138	10RAT-DT111B28	Jago	360	339.2	5.66	-7.67	-25.44	0.50	-19.2
10RAT-DT111A139		Jago	361	338.2	5.76	-7.76			
10RAT-DT111A140	10RAT-DT111B29	Jago	362	337.2	5.80	-7.87			
10RAT-DT111A141		Jago	363.2	336					
10RAT-DT111A142	10RAT-DT111B30	Jago	364.2	335			-25.33	0.06	
10RAT-DT111A143		Jago	365.2	334					
10RAT-DT111A144	10RAT-DT111B31	Jago	366.2	333			-26.08	0.13	
10RAT-DT111A145		Jago	367.3	331.9					
10RAT-DT111A146	10RAT-DT111B32	Jago	367.9	331.3			-25.28	0.34	-31.5
10RAT-DT111A147	10RAT-DT111B33	Jago	368.9	330.3			-24.83	0.39	

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT111A149	10RAT-DT111B34	Jago	371	328.2			-25.49	0.16	-29.6
10RAT-DT111A150		Jago	366.9	332.3					
10RAT-DT111A151		Jago	371.5	327.7					
10RAT-DT111A152		Jago	372.6	326.6					
10RAT-DT111A153		Jago	373.6	325.6					
10RAT-DT111A154		Jago	375	324.2					
10RAT-DT111A155		Jago	376.2	323					
10RAT-DT111A156	10RAT-DT111B35	Jago	377.4	321.8			-25.59	0.23	
10RAT-DT111A157		Jago	378.2	321					
10RAT-DT111A158		Jago	379.4	319.8					
10RAT-DT111A159	10RAT-DT111B36	Jago	380.1	319.1			-24.89	0.21	-30.7
10RAT-DT111A160		Jago	381.3	317.9					
10RAT-DT111A161	10RAT-DT111B37	Jago	382.1	317.1			-25.45	0.23	
10RAT-DT111A162		Jago	382.6	316.6					
10RAT-DT111A163		Jago	384.1	315.1					
10RAT-DT111A164	10RAT-DT111B38	Jago	385.2	314			-26.48	0.43	0.7
10RAT-DT111A165		Jago	386.2	313					
10RAT-DT111A166		Jago	387.2	312					
10RAT-DT111A167		Jago	388.4	310.8					
10RAT-DT111A168		Jago	389.9	309.3					
10RAT-DT111A169		Jago	391.2	308					
10RAT-DT111A170		Jago	392.5	306.7					
10RAT-DT111A171		Jago	393.3	305.9					
10RAT-DT111A172		Jago	394.8	304.4					
10RAT-DT111A173	10RAT-DT111B39	Jago	395.6	303.6			-23.36	0.35	
10RAT-DT111A174		Jago	396.4	302.8					
10RAT-DT111A175		Jago	397.9	301.3					
10RAT-DT111A176	10RAT-DT111B40	Jago	398.4	300.8			-24.42	0.26	6.1
10RAT-DT111A177		Jago	398.6	300.6					
10RAT-DT111A178		Jago	399.9	299.3					
10RAT-DT111A179	10RAT-DT111B41	Jago	400.6	298.6			-25.26	0.53	12.1
10RAT-DT111A180		Jago	401.03	298.17					
11RAT-DT029A01		ucm	1.5	1417.8	4.08	-5.60			
11RAT-DT029A02		ucm	8.7	1425	1.12	-6.21			
11RAT-DT029A03		ucm	10.4	1426.7	3.46	-6.40			
11RAT-DT029A04		ucm	13.8	1430.1	4.85	-6.17			
11RAT-DT029A06		ucm	98	1514.3	5.20	-7.25			
11RAT-DT029A07		ucm	104.5	1520.8	3.20	-6.05			
11RAT-DT031A01		ucm	0.7	1363.6	5.19	-5.80			
11RAT-DT031A02		ucm	6.3	1369.2	5.77	-5.17			
11RAT-DT031A03		ucm	7.8	1370.7	6.09	-4.92			
11RAT-DT031A04		ucm	11.1	1374	5.37	-5.87			
11RAT-DT031A05		ucm	18.3	1381.2	5.90	-5.50			
11RAT-DT031A06		ucm	25.6	1388.5	3.87	-6.32			
11RAT-DT031A07		ucm	33.5	1396.4	4.18	-5.18			
11RAT-DT031A08		ucm	53.4	1416.3	3.98	-5.89			
11RAT-DT034A01		ucm	0.1	1520.9	4.96	-7.04			
11RAT-DT034A02		ucm	10	1530.8	3.00	-5.94			
11RAT-DT034A03		ucm	28.5	1549.3	4.03	-6.56			
11RAT-DT034A04		ucm	37.5	1558.3	3.83	-6.38			
11RAT-DT034A05		ucm	77.9	1598.7	4.17	-6.97			
11RAT-DT034A06		ucm	79.3	1600.1	4.78	-6.67			
11RAT-DT095A01		ucm	311	1138					
11RAT-DT095A02		ucm	312.1	1137.3					
11RAT-DT095A03		ucm	311.2	1138.2	-1.46	-7.87			
11RAT-DT095A05		ucm	309.2	1140.2	-6.84	-9.57			

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
11RAT-DT095A07		ucm	306.4	1143	-7.78	-9.02			
11RAT-DT095A08		ucm	305.2	1144.2	-8.59	-9.36			
11RAT-DT095A09		ucm	304.2	1145.2					
11RAT-DT095A10		ucm	303.3	1146.1	-9.64	-9.11			
11RAT-DT095A11		ucm	302.2	1147.2					
11RAT-DT095A12		ucm	300.8	1148.6	-8.76	-8.86			
11RAT-DT095A13		ucm	299.8	1149.6	-10.82	-8.89			
11RAT-DT095A14		ucm	298.5	1150.9	-8.56	-9.25			
11RAT-DT095A15		ucm	297.5	1151.9	-10.17	-9.37			
11RAT-DT095A16		ucm	296.8	1152.6	-9.55	-9.47			
11RAT-DT095A17		ucm	295.6	1153.8	-4.50	-8.66			
11RAT-DT095A18		ucm	294.6	1154.8	-5.47	-8.82			
11RAT-DT095A19		ucm	293	1156.4	-14.31	-8.89			
11RAT-DT095A20		ucm	292.1	1157.3	-6.99	-9.02			
11RAT-DT095A21		ucm	290.5	1158.9	-8.47	-9.45			
11RAT-DT095A22		ucm	289.5	1159.9	-9.15	-9.23			
11RAT-DT095A23		ucm	288.5	1160.9	-7.52	-8.85			
11RAT-DT095A24		ucm	287.3	1162.1	-7.28	-8.91			
11RAT-DT095A25		ucm	285.9	1163.5	-8.03	-9.04			
11RAT-DT095A26		ucm	285.5	1163.9	-8.35	-9.02			
11RAT-DT095A27		ucm	284.2	1165.2	-6.75	-9.37			
11RAT-DT095A28		ucm	282.9	1166.5	-7.05	-9.44			
11RAT-DT095A29		ucm	282	1167.4	-6.38	-10.14			
11RAT-DT095A30		ucm	281	1168.4					
11RAT-DT095A31		ucm	280	1169.4	-6.88	-9.17			
11RAT-DT095A33		ucm	278.1	1171.3	-4.49	-9.02			
11RAT-DT095A34		ucm	276.9	1172.5	-5.73	-9.88			
11RAT-DT095A35		ucm	275.6	1173.8	-5.05	-9.03			
11RAT-DT095A36		ucm	274.1	1175.3					
11RAT-DT095A37		ucm	272.7	1176.7	-5.12	-8.51			
11RAT-DT095A38		ucm	271.5	1177.9	-4.80	-12.35			
11RAT-DT095A39		ucm	270.6	1178.8					
11RAT-DT095A40		ucm	269.5	1179.9					
11RAT-DT095A41		ucm	267.5	1181.9					
	11RAT-DT095B01	ucm	311.9	1138.9			-31.57	0.89	7.6
	11RAT-DT095B02	ucm	310.4	1140.4			-31.01	0.63	
	11RAT-DT095B03	ucm	309.6	1141.2			-32.70	0.52	
	11RAT-DT095B04	ucm	308.4	1142.4			-34.11	8.48	-23.82
	11RAT-DT095B05	ucm	307.2	1143.6			-33.72	6.96	
	11RAT-DT095B06	ucm	305.7	1145.1			-33.52	6.87	-31.10
	11RAT-DT095B07	ucm	304.4	1146.4			-33.60	5.12	-25.75
	11RAT-DT095B08	ucm	303.7	1147.1			-33.76	5.53	-27.57
	11RAT-DT095B09	ucm	302.1	1148.7			-34.11	9.95	-25.09
	11RAT-DT095B10	ucm	301	1149.8			-33.88	3.67	-24.34
	11RAT-DT095B11	ucm	300	1150.8			-34.15	4.21	-20.08
	11RAT-DT095B12	ucm	299.1	1151.7			-34.21	6.55	-22.51
	11RAT-DT095B13	ucm	297.5	1153.3			-34.06	3.49	-21.25
	11RAT-DT095B14	ucm	296.5	1154.3			-33.36	12.53	-19.69
	11RAT-DT095B15	ucm	295.3	1155.5			-32.05	1.34	-22.03
	11RAT-DT095B16	ucm	294.2	1156.6			-33.33	2.62	-18.16
	11RAT-DT095B17	ucm	293.2	1157.6			-32.93	7.15	-19.39
	11RAT-DT095B18	ucm	292.1	1158.7			-33.10	4.55	-17.25
	11RAT-DT095B19	ucm	290.6	1160.2			-33.08	5.87	-12.45
	11RAT-DT095B20	ucm	289.5	1161.3			-32.44	8.00	-15.70
	11RAT-DT095B21	ucm	288.4	1162.4			-32.31	11.83	-17.66
	11RAT-DT095B22	ucm	287.5	1163.3			-32.52	4.01	-17.70

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
	11RAT-DT095B24	ucm	285.5	1165.3			-31.94	2.30	-14.91
	11RAT-DT095B25	ucm	284.4	1166.4			-32.03	1.59	-7.18
	11RAT-DT095B26	ucm	283.4	1167.4			-31.32	9.15	-12.95
	11RAT-DT095B27	ucm	282.4	1168.4			-30.87	4.12	-16.49
	11RAT-DT095B28	ucm	281.4	1169.4			-30.99	8.38	-13.66
	11RAT-DT095B29	ucm	280.3	1170.5			-31.18	8.32	-4.87
	11RAT-DT095B30	ucm	278.8	1172			-30.98	6.69	-3.57
	11RAT-DT095B31	ucm	278.1	1172.7			-30.35	8.40	-0.59
	11RAT-DT095B32	ucm	276.8	1174			-30.05	6.47	-5.47
	11RAT-DT095B33	ucm	275.8	1175			-29.58	3.81	-9.46
	11RAT-DT095B34	ucm	274.2	1176.6			-29.34	6.36	-6.32
	11RAT-DT095B35	ucm	272.6	1178.2			-29.20	2.29	-9.40
	11RAT-DT095B36	ucm	271.4	1179.4			-29.01	8.12	-10.35
	11RAT-DT095B37	ucm	270.5	1180.3			-28.99	7.64	-7.88
	11RAT-DT095B38	ucm	269.4	1181.4			-28.64	0.20	
11RAT-DT127A02	Jago		10	260.7	2.26	-11.96			
11RAT-DT127A03	Jago		14.5	264.2	3.38	-10.23			
11RAT-DT130A01	ucm		0	1600.1			-24.71	0.57	-3.3
11RAT-DT130A02	ucm		0.1	1600.2			-24.51	0.24	
11RAT-DT130A03	ucm		0.3	1600.4			-24.85	-0.04	
11RAT-DT130A04	ucm		0.45	1600.55			-25.11	0.40	
11RAT-DT130A05	ucm		0.55	1600.65			-25.51	0.33	
11RAT-DT130A06	ucm		0.7	1600.8			-24.92	0.18	
11RAT-DT130A07	ucm		0.8	1600.9			-25.12	0.08	
11RAT-DT130A08	ucm		0.9	1601			-26.18	0.45	0
11RAT-DT130A09	ucm		0.9	1601			-25.33	0.07	
11RAT-DT130A10	ucm		1	1601.1			-25.68	0.25	
11RAT-DT130A11	ucm		1.05	1601.15			-26.00	0.50	4.3
11RAT-DT130A12	ucm		1.25	1601.35			-26.07	1.20	8.1
11RAT-DT130A13	ucm		1.3	1601.4			-23.51	0.22	
11RAT-DT130A14	ucm		1.4	1601.5			-23.95	0.28	
11RAT-DT130A15	ucm		1.5	1601.6			-23.87	0.14	
11RAT-DT130A16	ucm		1.7	1601.8			-22.99	0.08	
11RAT-DT130A17	ucm		1.75	1601.85					
11RAT-DT130A18	ucm		2	1602.1					
11RAT-DT130A19	ucm		2.15	1602.25					
11RAT-DT130A20	ucm		0	1600.1					
11RAT-DT130A21	ucm		0	1600.1					
11RAT-DT130A22	ucm		1.25	1601.35					
11RAT-DT147A01	Jago		1.5	270.7	4.15	-9.19			
11RAT-DT147A02	Jago		6.7	275.9	2.50	-9.03			
11RAT-DT147A03	Jago		22.1	291.3	2.68	-9.14			
11RAT-DT147A04	Jago		23.8	293	2.54	-9.07			
11RAT-DT147A05	Jago		26	295.2	2.60	-8.55			
11RAT-DT147A06	Jago		27.5	296.7	3.90	-8.39			
11RAT-DT147A07	Jago		28.6	297.8	3.75	-8.46			
11RAT-DT147A08	Jago		30.6	299.8	4.54	-3.68			
11RAT-DT147A09	Jago		32.1	301.3	4.72	-3.21			
11RAT-DT147A10	Jago		33.8	303	4.53	-6.09			
11RAT-DT147A11	Jago		35.1	304.3	4.13	-3.10			
11RAT-DT147A12	Jago		36.1	305.3	3.65	-3.93			
11RAT-DT147A13	Jago		38.1	307.3	4.03	-5.87			
11RAT-DT147A14	Jago		39.6	308.8	4.35	-7.65			
11RAT-DT147A15	Jago		42.2	311.4	3.34	-7.40			
11RAT-DT147A16	Jago		44	313.2	4.56	-2.95			
11RAT-DT147A17	Jago		45.8	315	4.78	-3.64			

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
11RAT-DT147A19		Jago	52.5	321.7	4.82	-3.69			
11RAT-DT147A20		Jago	54.1	323.3	4.66	-5.62			
11RAT-DT147A21		Jago	55.6	324.8	4.46	-1.08			
11RAT-DT147A22		Jago	57.1	326.3	4.49	-5.76			
11RAT-DT147A23		Jago	60.7	329.9	5.22	-3.49			
11RAT-DT147A25		Jago	64.6	333.8	4.99	-4.08			
11RAT-DT147A26		Jago	67.6	336.8	5.04	-5.71			
11RAT-DT147A27		Jago	69.1	338.3					
11RAT-DT147A28		Jago	72.1	341.3	5.19	-5.61			
10RAT-GL028A1		bsm	2	782.7			-24.40		
10RAT-GL028A2		bsm	4	784.7			-25.70		
10RAT-GL028A3		bsm	7.8	788.5			-25.70		
10RAT-GL028A4		bsm	8.7	789.4			-25.30		
10RAT-GL028A5		bsm	10.3	791			-25.10		
10RAT-GL028A6		bsm	11.9	792.6			-25.50		
10RAT-GL028A7		bsm	14	794.7			-26.00		
10RAT-GL028A8		bsm	16.5	797.2			-25.80		
10RAT-GL028A9		bsm	18.4	799.1			-24.50		
10RAT-GL028A10		bsm	21	801.7			-24.50		
10RAT-GL028A11		bsm	22	802.7			-24.00		
10RAT-GL028A12		bsm	24	804.7			-24.50		
10RAT-GL028A13		bsm	25.7	806.4			-22.60		
10RAT-GL028A14		bsm	28	808.7			-22.50		
10RAT-GL028A15		bsm	30	810.7			-23.60		
10RAT-GL028A16		bsm	32	812.7			-24.40		
10RAT-GL028A17		bsm	33	813.7			-23.30		
10RAT-GL028A18		bsm	34.4	815.1			-22.20		
10RAT-GL028A19		bsm	36.2	816.9			-23.90		
10RAT-GL028A20		bsm	38	818.7			-23.30		
10RAT-GL028A21		bsm	40	820.7			-23.20		
10RAT-GL028A22		bsm	41.6	822.3			-25.40		
10RAT-GL028A23		bsm	42	822.7			-22.40		
10RAT-GL028A24		bsm	44	824.7			-20.70		
10RAT-GL028A25		bsm	46	826.7			-23.90		
10RAT-GL028A26		bsm	48	828.7			-22.50		
10RAT-GL028A27		bsm	50	830.7			-23.40		
10RAT-GL028A28		bsm	52	832.7			-24.20		
10RAT-GL028A29		bsm	56	836.7			-21.50		
10RAT-GL028A30		bsm	58	838.7			-23.40		
10RAT-GL028A31		bsm	60	840.7			-23.20		
10RAT-GL028A32		bsm	62	842.7			-21.60		
10RAT-GL028A33		bsm	64	844.7			-20.00		
10RAT-GL028A34		bsm	66	846.7			-24.10		
10RAT-GL028A35		bsm	68	848.7			-22.50		
10RAT-GL028A36		bsm	70	850.7			-22.70		
10RAT-GL028A37		bsm	72	852.7			-21.60		
10RAT-GL028A38		bsm	74	854.7			-21.00		
10RAT-GL028A39		bsm	76	856.7			-20.50		
10RAT-GL028A40		bsm	78	858.7			-20.30		
10RAT-GL028A41		bsm	80	860.7			-20.70		
10RAT-GL028A42		bsm	82.4	863.1			-21.80		
10RAT-GL028A44		bsm	86	866.7			-21.80		
10RAT-GL028A45		bsm	88	868.7			-22.90		
10RAT-GL028A47		bsm	92	872.7			-24.10		
10RAT-GL028A48		bsm	96	876.7			-25.30		
10RAT-GL028A49		bsm	98	878.7			-26.70		

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-GL028A51		bsm	102	882.7			-25.40		
10RAT-GL028A52		bsm	104	884.7			-25.90		
10RAT-GL028A53		bsm	106	886.7			-22.50		
10RAT-GL028A54		bsm	108	888.7			-21.50		
10RAT-GL028A55		bsm	110	890.7			-22.10		
10RAT-GL028A56		bsm	112	892.7			-22.60		
10RAT-GL028A57		bsm	116	896.7			-20.00		
10RAT-GL028A58		bsm	120	900.7			-21.30		
10RAT-GL028A59		bsm	125	905.7			-21.90		
10RAT-GL028A60		bsm	130	910.7			-23.10		
10RAT-GL044A1		bsm	155	935.7			-27.20		
10RAT-GL044A2		bsm	158	938.7			-25.80		
10RAT-GL044A3		bsm	162	942.7			-26.60		
10RAT-GL044A4		bsm	166	946.7			-26.30		
10RAT-GL044A5		bsm	168	948.7			-23.80		
10RAT-GL044A6		bsm	170	950.7			-26.10		
10RAT-GL044A7		bsm	172	952.7			-26.30		
10RAT-GL044A8		bsm	174	954.7			-25.60		
10RAT-GL044A9		bsm	176	956.7			-23.80		
10RAT-GL044A10		bsm	178	958.7			-25.50		
10RAT-GL044A11		bsm	180	960.7			-20.40		
10RAT-GL044A12		bsm	182	962.7			-23.80		
10RAT-GL044A13		bsm	184	964.7			-26.30		
10RAT-DT110-B01		Ft. Collinson	151.5	173.7			-28.36	0.31	-1.5
10RAT-DT110-B02		Ft. Collinson	154.5	170.7			-29.50	0.63	17.70
10RAT-DT110-B03		Ft. Collinson	155.7	169.5	0.45	-13.86	-28.94	0.24	
10RAT-DT110-B04		Ft. Collinson	157.5	167.7	-0.25	-14.73	-28.23	0.07	-15.9
10RAT-DT110-B05		Ft. Collinson	162.4	162.8	-0.04	-14.33	-29.30	0.20	
10RAT-DT110-B06		Ft. Collinson	163.7	161.5	-0.07	-13.00	-30.69	0.21	
10RAT-DT110-B07		Ft. Collinson	167	158.2	-0.52	-14.13	-29.19	0.30	-2.4
10RAT-DT110-B07		Ft. Collinson	169.6	155.6	-0.35	-14.34	-29.94	0.42	
10RAT-DT110-B09		Ft. Collinson	140.4	184.8	0.87	-15.50	-28.66	0.48	
10RAT-DT110-B10		Ft. Collinson	142.9	182.3	-0.31	-13.56	-28.41	0.96	-9
10RAT-DT110-B11		Ft. Collinson	148.6	176.6	1.70	-11.58	-28.64	0.13	
10RAT-DT110-B12		Ft. Collinson	149.9	175.3	3.29	-11.05	-27.73	0.25	
10RAT-DT110-B13		Ft. Collinson	195.1	130.1	1.45	-15.14	-28.83	0.89	9.7
10RAT-DT110-B14		Ft. Collinson	203.1	122.1	5.31	-9.38	-25.86	0.66	7.9
10RAT-DT110-B15		Boot Inlet	211.3	113.9	5.05	-8.58	-28.36	0.50	
10RAT-DT110-B16		Boot Inlet	218.5	106.7	3.66	-4.00	-24.58	0.03	
10RAT-DT110-B18		Boot Inlet	234.2	91.0	4.99	-6.07	-25.77	0.38	
10RAT-DT110-B19		Boot Inlet	235.5	89.7	4.79	-6.14	-25.10	0.24	
10RAT-DT110-B20		Boot Inlet	240.2	85.0	5.42	-5.96	-26.53	0.13	
10RAT-DT110-B21		Boot Inlet	246.9	78.3	4.86	-7.50	-26.08	0.42	
10RAT-DT110-B22		Boot Inlet	249.8	75.4	4.61	-8.95	-26.15	0.17	
10RAT-DT110-B23		Boot Inlet	254.6	70.6	4.62	-9.00	-26.16	0.69	-6.2
10RAT-DT110-B24		Boot Inlet	259.9	65.3	6.14	-6.78	-25.75	0.21	
10RAT-DT110-B25		Boot Inlet	262.1	63.1	5.13	-9.67	-26.33	0.20	
10RAT-DT110-B26		Boot Inlet	264.4	60.8	4.62	-9.80	-25.93	0.26	
10RAT-DT110-B27		Boot Inlet	269.2	56.0	4.49	-10.84	-26.23	0.01	
10RAT-DT110-B28		Boot Inlet	274.8	50.4	4.99	-8.84	-26.67	0.34	
10RAT-DT110-B29		Boot Inlet	278.1	47.1	5.19	-10.02	-26.90	0.41	7.50
10RAT-DT110-B30		Boot Inlet	283.3	41.9	5.64	-7.91	-25.89	0.14	
10RAT-DT110-B31		Boot Inlet	284.5	40.7	4.79	-8.77	-26.26	0.44	9.30
10RAT-DT110-B32		Boot Inlet	288.4	36.8	4.87	-9.14	-26.48	0.50	
10RAT-DT110-B33		Boot Inlet	291.6	33.6	4.65	-8.79	-26.30	0.73	14.2
10RAT-DT110-B34		Boot Inlet	254.5	70.7	4.66	-8.54	-26.30	0.53	

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT110-B36		Boot Inlet	298.4	26.8	4.50	-8.90	-26.73	0.39	11.5
10RAT-DT110-B37		Boot Inlet	303.8	21.4	4.34	-9.68	-26.31	0.05	
10RAT-DT110-B38		Boot Inlet	305.7	19.5	4.26	-9.49	-26.22	0.11	
10RAT-DT110-B39		Boot Inlet	313.2	12	2.52	-11.03	-27.41	0.51	2.8
10RAT-DT110-B40		Boot Inlet	313.4	11.8	4.54	-9.30	-25.27	0.14	
10RAT-DT110A01		Jago	45.2	280.0	0.31	-8.64			
10RAT-DT110A02		Jago	47.2	278.0	0.31	-9.24			
10RAT-DT110A03		Jago	48.9	276.3	0.91	-10.81			
10RAT-DT110A04		Jago	50.2	275.0	0.73	-11.17			
10RAT-DT110A05		Jago	51.3	273.9	0.76	-10.83			
10RAT-DT110A06		Jago	52.3	272.9	0.60	-11.38			
10RAT-DT110A07		Jago	54.8	270.4	2.99	-12.07			
10RAT-DT110A08		Jago	57.2	268.0	4.67	-8.68			
10RAT-DT110A09		Jago	58.8	266.4	4.74	-9.09			
10RAT-DT110A10		Ft. Collinson	70.1	255.1	0.83	-8.28			
10RAT-DT110A11		Ft. Collinson	71.4	253.8	2.17	-8.30			
10RAT-DT110A12		Ft. Collinson	75.3	249.9	3.09	-11.21			
10RAT-DT110A13		Ft. Collinson	78.4	246.8	2.57	-8.30			
10RAT-DT110A14		Ft. Collinson	80.4	244.8	3.58	-10.95			
10RAT-DT110A15		Ft. Collinson	82.1	243.1	3.66	-9.55			
10RAT-DT110A16		Ft. Collinson	84.3	240.9	3.36	-9.56			
10RAT-DT110A17		Ft. Collinson	86.8	238.4	2.92	-9.92			
10RAT-DT110A18		Ft. Collinson	88.9	236.3	3.87	-10.58			
10RAT-DT110A19		Ft. Collinson	91.2	234.0	0.77	-11.66			
10RAT-DT110A20		Ft. Collinson	93.3	231.9	1.58	-11.20			
10RAT-DT110A21		Ft. Collinson	95.2	230.0	2.68	-12.63			
10RAT-DT110A22		Ft. Collinson	97.1	228.1	0.66	-14.13			
10RAT-DT110A23		Ft. Collinson	99.2	226.0	2.01	-15.59			
10RAT-DT110A24		Ft. Collinson	101.1	224.1	2.36	-15.83			
10RAT-DT110A26		Ft. Collinson	136.3	188.9	0.02	-14.14			
10RAT-DT110A27		Ft. Collinson	138.3	186.9	1.44	-14.46			
10RAT-DT110A28		Ft. Collinson	140.2	185.0	2.06	-13.23			
10RAT-DT110A29		Ft. Collinson	142.1	183.1	-0.52	-14.76			
10RAT-DT110A30		Ft. Collinson	144.5	180.7	-0.15	-13.36			
10RAT-DT110A31		Ft. Collinson	146.4	178.8	-0.14	-13.10			
10RAT-DT110A32		Ft. Collinson	148.3	176.9					
10RAT-DT110A33		Ft. Collinson	167	158.2	1.93	-11.53			
10RAT-DT110A34		Ft. Collinson	169.6	155.6	3.07	-10.96			
10RAT-DT110A35		Ft. Collinson	170.9	154.3	3.34	-11.06			
10RAT-DT110A36		Ft. Collinson	173.9	151.3	2.17	-10.50			
10RAT-DT110A37		Ft. Collinson	177.6	147.6	3.13	-12.96			
10RAT-DT110A38		Ft. Collinson	180	145.2	4.31	-12.01			
10RAT-DT110A39		Ft. Collinson	182.2	143.0	4.38	-11.76			
10RAT-DT110A40		Ft. Collinson	183.4	141.8	3.78	-12.08			
10RAT-DT110A41		Ft. Collinson	198.2	127.0	3.38	-11.16			
10RAT-DT110A42		Ft. Collinson	200.1	125.1	5.34	-9.98			
10RAT-DT110A43		Ft. Collinson	201.4	123.8	5.47	-9.03			
10RAT-DT110A44		Ft. Collinson	203	122.2	4.62	-10.09			
10RAT-DT110A45		Ft. Collinson	206.3	118.9	5.39	-9.43			
10RAT-DT110A46		Ft. Collinson	208.3	116.9	4.78	-11.42			
10RAT-DT110A47		Boot Inlet	213.5	111.7	4.84	-8.13			
10RAT-DT110A48		Boot Inlet	215.5	109.7	4.08	-9.26			
10RAT-DT110A49		Boot Inlet	217.3	107.9	3.74	-7.50			
10RAT-DT110A50		Boot Inlet	219.1	106.1	3.57	-8.43			
10RAT-DT110A51		Boot Inlet	221.4	103.8	3.62	-4.21			
10RAT-DT110A52		Boot Inlet	223.5	101.7	3.94	-8.14			

Sample- A series	Sample -B series	Fm/Mb	m	m (base)	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}_{\text{org}}$	TOC	$\delta^{34}\text{S}$
10RAT-DT110A54		Boot Inlet	227.5	97.7	4.49	-8.06			
10RAT-DT110A55		Boot Inlet	229.5	95.7	4.75	-7.55			
10RAT-DT110A56		Boot Inlet	231.4	93.8	4.48	-7.63			
10RAT-DT110A57		Boot Inlet	233.4	91.8	4.63	-7.83			
10RAT-DT110A58		Boot Inlet	235.6	89.6	4.27	-7.53			
10RAT-DT110A59		Boot Inlet	237.5	87.7	4.30	-7.57			
10RAT-DT110A60		Boot Inlet	238.5	86.7					
10RAT-DT110A61		Boot Inlet	239.5	85.7	4.86	-6.32			
10RAT-DT110A62		Boot Inlet	241.6	83.6	5.23	-5.92			
10RAT-DT110A63		Boot Inlet	243.6	81.6	3.53	-8.11			
10RAT-DT110A64		Boot Inlet	245.5	79.7	4.79	-7.57			
10RAT-DT110A65		Boot Inlet	247.8	77.4	4.94	-7.89			
10RAT-DT110A66		Boot Inlet	249.9	75.3	4.77	-7.30			
10RAT-DT110A67		Boot Inlet	252.2	73.0	4.58	-8.78			
10RAT-DT110A68		Boot Inlet	254.3	70.9	4.86	-7.83			
10RAT-DT110A69		Boot Inlet	256.3	68.9	5.07	-8.26			
10RAT-DT110A70		Boot Inlet	258	67.2	4.99	-8.21			
10RAT-DT110A71		Boot Inlet	260.3	64.9	5.04	-7.93			
10RAT-DT110A74		Boot Inlet	266.1	59.1	5.22	-8.32			
10RAT-DT110A75		Boot Inlet	268	57.2	5.42	-8.35			
10RAT-DT110A76		Boot Inlet	270.1	55.1	5.76	-8.34			
10RAT-DT110A77		Boot Inlet	272	53.2	5.85	-8.61			
10RAT-DT110A78		Boot Inlet	273.5	51.7	6.03	-8.47			
10RAT-DT110A79		Boot Inlet	275.2	50.0	4.89	-8.52			
10RAT-DT110A80		Boot Inlet	279.1	46.1	5.18	-8.16			
10RAT-DT110A81		Boot Inlet	281.1	44.1	5.56	-8.40			
10RAT-DT110A82		Boot Inlet	283.2	42.0	5.00	-8.10			
10RAT-DT110A83		Boot Inlet	285.1	40.1	4.52	-8.83			
10RAT-DT110A84		Boot Inlet	287.1	38.1	4.88	-8.74			
10RAT-DT110A85		Boot Inlet	289.1	36.1	4.51	-8.68			
10RAT-DT110A86		Boot Inlet	291.2	34.0	5.05	-8.59			
10RAT-DT110A87		Boot Inlet	293.1	32.1	4.20	-8.76			
10RAT-DT110A88		Boot Inlet	295.1	30.1	4.75	-8.35			
10RAT-DT110A89		Boot Inlet	297.2	28.0	4.60	-8.73			
10RAT-DT110A90		Boot Inlet	299.2	26.0	4.87	-8.33			
10RAT-DT110A91		Boot Inlet	301.2	24.0	4.34	-8.36			
10RAT-DT110A92		Boot Inlet	303.3	21.9	4.95	-8.41			
10RAT-DT110A93		Boot Inlet	305.2	20.0	4.24	-9.32			
10RAT-DT110A94		Boot Inlet	309.2	16.0	4.50	-7.57			
10RAT-DT110A95		Boot Inlet	311.3	13.9	3.78	-9.56			
10RAT-DT110A96		Boot Inlet	313.4	11.8	3.33	-9.98			
10RAT-DT110A97		Boot Inlet	315.3	9.9	3.52	-9.74			
10RAT-DT110A98		Boot Inlet	316.4	8.8	3.58	-9.83			
10RAT-DT110A99		Boot Inlet	318.3	6.9	3.65	-9.83			
10RAT-DT110A100		Boot Inlet	320.3	4.9	3.52	-10.15			
10RAT-DT110A101		Boot Inlet	322.2	3.0	3.44	-11.07			
10RAT-DT110A102		Boot Inlet	324.3	0.9	3.44	-10.74			
10RAT-DT110A103		Boot Inlet	325.2	0	3.36	-10.80			
10RAT-DT110A104		Boot Inlet	325.2	0	3.44	-9.76			

Appendix 3. Trace element data. MDL – maximum detection limit.

Sample- A series	Fm/Mb	m (measu red)	m Boot Inlet)	Unit MDL	Mo PPM 0.05	V PPM 1	U PPM 0.1	Mn PPM 2	Fe % 0.0	Mg % 0	Al % 0
10RAT-DT027A11	lcm	53	709.7		0.1	4	0.2	832	1.0	11	0.3
10RAT-DT027A19	lcm	91.6	748.3		0.1	7	0.2	771	0.8	12	0.3
10RAT-DT030A7	bsm	18	798.7		0.1	69	1.6	920	4.8	4.2	5.6
10RAT-DT030A12	bsm	73.5	854.2		6.8	60	1.4	657	15.8	0.6	3.5
10RAT-DT057A3	ucm	7.8	1382.7		0.1	3	2.1	48	0.1	0.3	0.3
10RAT-DT058A9	scm	47.2	1022.1		0.1	12	0.9	480	1.1	10	1.7
10RAT-DT060A4	scm	17	1096.9		0.1	2	0.2	285	0.3	12	0.2
10RAT-DT060A23	scm	55.1	1135		0.4	20	0.8	287	1.0	10	0.7
10RAT-DT062A8	ucm	28.5	1243.4		0.08	4	0.2	91	0.4	1.1	0.4
10RAT-DT074A6	scm	26.4	991		0.08	10	0.3	215	0.6	1	1.1
10RAT-DT074A13	scm	72.8	1037.4		0.05	5	0.4	107	0.3	2.7	0.6
10RAT-DT111A11	Minto	36.6	662.6		0.26	71	3.9	89	1.1	5.9	6.1
10RAT-DT111A85	Jago	298.2	401		<0.05	15	1.6	206	0.7	3.3	1.8
10RAT-DT111A97	Jago	312.8	386.4		0.15	31	2.6	208	1.4	2.7	2.9
10RAT-DT111A107	Jago	321.8	377.4		0.2	22	1.7	483	1.7	2	2.3
10RAT-DT111A113	Jago	332.8	366.4		0.3	67	5.7	295	2.2	3.9	6.1
10RAT-DT111A123	Jago	344.4	354.8		0.09	17	2.2	311	0.8	2.8	1.5
10RAT-DT111A131	Jago	352.7	346.5		0.16	17	2.4	310	0.8	2.5	1.4
10RAT-DT111A134	Jago	355.7	343.5		0.3	67	3.3	258	3.1	2.9	5.8
10RAT-DT111A138	Jago	360	339.2		0.19	52	2.3	349	2.5	3	5.4
10RAT-DT111A146	Jago	367.9	331.3		0.23	53	2.9	298	2.1	2.6	4.9
10RAT-DT111A149	Jago	371	328.2		0.1	15	1.9	298	0.7	2.6	1.3
10RAT-DT111A150	Jago	366.9	332.3		sample I.S.	I.S.	I.S.	I.S.	I.S.	I.S.	I.S.
10RAT-DT111A159	Jago	380.1	319.1		0.3	76	3	303	2.9	2.8	6.9
10RAT-DT111A164	Jago	385.2	314		<0.05	8	1.2	198	0.2	0.6	0.5
10RAT-DT111A176	Jago	398.4	300.8		0.08	20	2.6	227	0.9	1.4	1.9
10RAT-DT111A179	Jago	400.6	298.6		0.17	72	5.5	186	1.8	2.5	6.9
11RAT-DT095B01	ucm	311.9	1138.9		0.2	70	0.4	353	3.0	12	2.6
11RAT-DT095B04	ucm	308.4	1142.4		47.6	1732	16	49	1.7	1.4	5.7
11RAT-DT095B05	ucm	307.2	1143.6		24.7	514	9.9	69	1.3	1.4	6.3
11RAT-DT095B06	ucm	305.7	1145.1		14.0	673	6.5	127	0.9	1.1	5.1
11RAT-DT095B07	ucm	304.4	1146.4		15.2	1084	8.7	63	1.7	1.4	8.2
11RAT-DT095B08	ucm	303.7	1147.1		61.3	1732	12	65	2.0	1.4	6.9
11RAT-DT095B09	ucm	302.1	1148.7		43.6	1511	11	42	1.7	1.4	5.6
11RAT-DT095B10	ucm	301	1149.8		41.8	1368	14	111	2.2	4.1	6.8
11RAT-DT095B11	ucm	300	1150.8		48.5	1991	18	123	2.0	2.5	7
11RAT-DT095B12	ucm	299.1	1151.7		92.5	2407	20	99	1.8	2.7	6.7
11RAT-DT095B13	ucm	297.5	1153.3		61.6	2087	7.7	193	1.9	3.5	5.9
11RAT-DT095B14	ucm	296.5	1154.3		51.0	2826	18	67	3.5	1.2	5.8
11RAT-DT095B15	ucm	295.3	1155.5		18.0	730	15	141	2.1	4.4	6.4
11RAT-DT095B16	ucm	294.2	1156.6		64.7	964	12	172	1.0	1.7	4.6
11RAT-DT095B17	ucm	293.2	1157.6		179.2	2578	14	89	1.8	3.1	6.9
11RAT-DT095B18	ucm	292.1	1158.7		27.7	346	16	142	1.1	1	5.1
11RAT-DT095B19	ucm	290.6	1160.2		63.9	695	9.9	88	1.1	1.2	5.8
11RAT-DT095B20	ucm	289.5	1161.3		76.5	1515	11	114	1.2	1.5	5.3
11RAT-DT095B21	ucm	288.4	1162.4		129.2	1572	18	46	1.2	1.1	5.4
11RAT-DT095B22	ucm	287.5	1163.3		54.1	1004	8.8	131	1.3	2.9	6.4
11RAT-DT095B23	ucm	286.5	1164.3		155.7	1272	14	102	2.4	1.9	6.1
11RAT-DT095B24	ucm	285.5	1165.3		29.5	491	5	165	1.1	3.5	5.3
11RAT-DT095B25	ucm	284.4	1166.4		8.9	335	5.2	239	1.8	4.5	4.5
11RAT-DT095B26	ucm	283.4	1167.4		21.7	799	10	69	1.4	1.5	6
11RAT-DT095B27	ucm	282.4	1168.4		27.1	701	12	143	1.2	4.1	6.5
11RAT-DT095B28	ucm	281.4	1169.4		22.3	873	14	109	1.1	2.1	6.3
11RAT-DT095B29	ucm	280.3	1170.5		33.2	496	9.4	124	0.9	1.4	5.4
11RAT-DT095B30	ucm	278.8	1172		11.9	189	8.8	211	1.5	3.1	5.6

Sample- A series	Fm/Mb	m	m	Mo	V	U	Mn	Fe	Mg	Al
11RAT-DT095B32	ucm	276.8	1174	11.2	177	15	182	2.8	3	6
11RAT-DT095B33	ucm	275.8	1175	5.2	97	9.7	169	1.2	2.3	4.2
11RAT-DT095B34	ucm	274.2	1176.6	7.7	443	10	100	1.9	4.4	6.4
11RAT-DT095B35	ucm	272.6	1178.2	2.6	123	6	220	1.5	7.2	3
11RAT-DT095B36	ucm	271.4	1179.4	9.7	267	13	154	1.5	5.7	3.9
11RAT-DT095B37	ucm	270.5	1180.3	5.9	275	12	167	2.8	7.2	5.9
11RAT-DT095B38	ucm	269.4	1181.4	1.29	56	1.9	281	2.0	8.1	2.6
11RAT-DT095B39	ucm	267.6	1183.2	0.31	318	9.2	97	2.2	6.4	6.7
11RAT-DT130A01	ucm	0	1600.1	0.1	<1	<0.1	128	0.2	0.6	0.1
11RAT-DT130A08	ucm	0.9	1601	0.3	35	1.4	108	1.1	11	2.6
11RAT-DT130A11	ucm	1.05	1601.15	0.2	39	1.5	201	1.0	13	2.3
11RAT-DT130A12	ucm	1.25	1601.35	0.1	9	0.6	206	0.6	11	0.8
10RAT-DT110-B01	Ft. Collinson	151.5	173.7	0.9	109	3.2	69	2.9	1.8	7.3
10RAT-DT110-B02	Ft. Collinson	154.5	170.7	0.1	25	1.2	114	0.5	0.7	2.4
10RAT-DT110-B04	Ft. Collinson	157.5	167.7	0.2	26	0.8	120	3.4	1.5	4.9
10RAT-DT110-B07	Ft. Collinson	167	158.2	0.3	49	2.1	134	1.9	1.3	5.8
10RAT-DT110-B10	Ft. Collinson	142.9	182.3	0.7	118	7.8	150	2.2	1.5	8.1
10RAT-DT110-B13	Ft. Collinson	195.1	130.1	0.3	61	3.1	150	2.9	0.9	5.5
10RAT-DT110-B14	Ft. Collinson	203.1	122.1	0.3	74	2.8	83	1.7	1.5	7
10RAT-DT110-B23	Boot Inlet	254.6	70.6	0.3	63	2.3	138	1.8	1.8	6
10RAT-DT110-B29	Boot Inlet	278.1	47.1	0.1	47	1.7	130	1.4	1.3	4.4
10RAT-DT110-B31	Boot Inlet	284.5	40.7	0.3	75	3	131	1.6	1.4	6.6
10RAT-DT110-B33	Boot Inlet	291.6	33.6	0.3	75	3.1	133	2.1	1.9	6.9
10RAT-DT110-B36	Boot Inlet	298.4	26.8	0.2	45	1.7	101	1.7	1	5.1
10RAT-DT110-B39	Boot Inlet	313.2	12	0.14	29	1.4	159	0.9	1	3
GL028-01	bsm	2.0	523.0	0.23	82	2.0				
GL028-02	bsm	6.9	527.9	0.28	121	2.2				
GL028-03	bsm	7.8	528.8	0.28	149	2.3				
GL028-04	bsm	8.7	529.7	0.11	116	1.8				
GL028-06	bsm	11.9	532.9	0.19	114	2.2				
GL028-07	bsm	14.0	535.0	0.39	148	2.2				
GL028-08	bsm	16.5	537.5	0.33	133	2.9				
GL028-09	bsm	18.0	539.0	0.18	73	2.1				
GL028-10	bsm	18.4	539.4	0.23	96	2.2				
GL028-11	bsm	21.0	542.0	0.15	143	2.3				
GL028-13	bsm	24.0	545.0	0.39	90	2.2				
GL028-14	bsm	25.7	546.7	1.66	134	4.1				
GL028-15	bsm	28.0	549.0	0.14	95	1.9				
GL028-18	bsm	33.0	554.0	1.98	110	3.3				
GL028-19	bsm	34.4	555.4	0.32	126	2.4				
GL028-20B	bsm	36.9	557.9	0.53	147	2.4				
GL028-21	bsm	38.0	559.0	0.37	145	2.6				
GL028-22F	bsm	40.0	561.0	0.25	141	2.8				
GL028-24	bsm	42.0	563.0	0.45	83	1.3				
GL028-25	bsm	44.0	565.0	1.92	135	6.9				
GL028-26	bsm	46.0	567.0	0.49	126	3.6				
GL028-28	bsm	50.0	571.0	1.87	132	3.7				
GL028-29	bsm	52.0	573.0	0.16	84	2.3				
GL028-31	bsm	56.0	577.0	0.39	142	3.0				
GL028-34	bsm	62.0	583.0	1.67	114	5.6				
GL028-36	bsm	66.0	587.0	0.23	92	##				
GL028-38	bsm	70.0	591.0	0.29	113	4.0				
GL028-39	bsm	72.0	593.0	0.99	97	3.6				
GL028-41	bsm	76.0	597.0	1.76	119	5.6				
GL028-44F	bsm	82.4	603.4	0.20	95	3.7				
GL028-45	bsm	84.0	605.0	0.61	139	5.8				
GL028-46	bsm	86.0	607.0	0.41	54	2.3				
GL028-47	bsm	88.0	609.0	0.90	147	5.7				

Sample- A series	Fm/Mb	m	m	Mo	V	U	Mn	Fe	Mg	Al
GL028-49	bsm	92.0	613.0	0.11	83	2.8				
GL028-50	bsm	94.0	615.0	0.13	97	2.9				
GL028-51	bsm	96.0	617.0	0.19	68	2.4				
GL028-52	bsm	98.0	619.0	0.19	74	1.8				
GL028-53	bsm	100.0	621.0	0.20	103	2.4				
GL028-54	bsm	102.0	623.0	0.38	134	3.2				
GL028-55	bsm	104.0	625.0	1.22	121	2.5				
GL028-56	bsm	106.0	627.0	0.50	133	2.9				
GL028-57	bsm	108.0	629.0	0.59	119	3.3				
GL028-58	bsm	110.0	631.0	0.81	123	4.5				
GL028-59	bsm	112.0	633.0	0.56	82	2.4				
GL028-60	bsm	116.0	637.0	0.85	92	2.6				
GL028-61	bsm	120.0	641.0	0.61	96	2.4				
GL028-62	bsm	125.0	646.0	0.71	119	3.0				
GL044-01	bsm	155.0	676.0	0.31	101	2.1				
GL044-03	bsm	163.0	684.0	0.85	105	2.4				
GL044-05	bsm	169.0	690.0	0.62	107	2.2				
GL044-06	bsm	171.0	692.0	0.40	88	2.2				
GL044-07	bsm	173.0	694.0	0.13	86	2.1				
GL044-08	bsm	175.0	696.0	0.22	77	1.5				
GL044-09	bsm	177.0	698.0	1.02	86	1.8				
GL044-10	bsm	179.0	700.0	1.73	86	1.7				
GL044-11	bsm	181.0	702.0	1.16	113	2.8				
GL044-13	bsm	185.0	706.0	0.50	41	0.4				

Appendix 4. Fe speciation data

Sample- A series	Fm/Mb	m (measured)	m (base Boot Inlet)	Pyrite Fe %	Oxalate Fe (%)	Dithionite Fe(%)	Na Acetate Fe (%)	Na Acetate Ca (%)	Na Acetate Mg (%)	Na Acetate Mn (%)	Fe(HR)/ Fe(T)	Fe(Py)/ Fe(HR)	Fe/Al
10RAT-DT027A11	lcm	53	709.7		0.04	0.06	0.69	14.99	8.09	0.06			
10RAT-DT027A19	lcm	91.6	748.3		0.08	0.04	0.46	11.08	5.98	0.05			
10RAT-DT030A7	bsm	18	798.7		0.15	0.16	1.86	6.70	2.95	0.09			
10RAT-DT058A9	scm	47.2	1022.1		0.06	0.07	0.61	12.65	6.72	0.03			1.25
10RAT-DT060A4	scm	17	1096.9		0.02 B.D.		0.24	16.47	9.26	0.03			1.30
10RAT-DT060A23	scm	55.1	1135										0.86
10RAT-DT062A8	ucm	28.5	1243.4		B.D.	0.02	0.05	36.32	0.36	0.01			0.51
10RAT-DT074A6	scm	26.4	991										0.48
10RAT-DT074A13	scm	72.8	1037.4										
10RAT-DT11A6	Minto	23.2	676	0.27	0.05	0.03	0.40	9.02	1.57	0.02			
10RAT-DT11A11	Minto	36.6	662.6	0.48	0.02	0.02	0.07	4.44	0.59	0.01	0.53	0.81	0.19
10RAT-DT11A85	Jago	298.2	401	0.23 B.D.		0.02	0.25	23.86	2.97	0.02	0.69	0.46	0.41
10RAT-DT11A88	Jago	301.3	397.9	0.28	0.02	0.03	0.23	7.90	1.29	0.02			
10RAT-DT11A97	Jago	312.8	386.4	0.27									0.48
10RAT-DT11A107	Jago	321.8	377.4	0.51									0.73
10RAT-DT11A113	Jago	332.8	366.4	0.69									0.36
10RAT-DT11A123	Jago	344.4	354.8	0.20									0.55
10RAT-DT11A131	Jago	352.7	346.5	0.23									0.55
10RAT-DT11A134	Jago	355.7	343.5	0.28									0.53
10RAT-DT11A138	Jago	360	339.2		0.04	0.03	0.46	10.66	1.53	0.03	0.21	0.00	0.45
10RAT-DT11A144	Jago	366.2	333	0.14									
10RAT-DT11A146	Jago	367.9	331.3	0.32									0.43
10RAT-DT11A149	Jago	371	328.2	0.63	0.00	0.02	0.23	28.11	1.79	0.02 -	0.72	0.58	
10RAT-DT11A159	Jago	380.1	319.1	0.17	0.02	0.05	0.71	3.69	1.43	0.02	0.33	0.18	0.42
10RAT-DT11A164	Jago	385.2	314	0.20									0.43
10RAT-DT11A176	Jago	398.4	300.8	0.28									0.47
10RAT-DT11A179	Jago	400.6	298.6	0.40	0.01	0.02	0.32	7.51	1.23	0.01	0.43	0.53	0.25
11RAT-DT095B01	ucm	311.9	1138.9	0.13 B.D.		0.07	0.57	10.88	5.98	0.03	0.25	0.17	1.16
11RAT-DT095B04	ucm	308.4	1142.4	0.69	0.03	0.04	0.09	0.14	0.08	0.00	0.49	0.81	0.31
11RAT-DT095B05	ucm	307.2	1143.6	0.00									0.21
11RAT-DT095B06	ucm	305.7	1145.1	0.24									0.18
11RAT-DT095B08	ucm	303.7	1147.1	0.37	0.06	0.03	0.13	1.47	0.13	0.00	0.30	0.62	0.29
11RAT-DT095B09	ucm	302.1	1148.7	0.50	0.02	0.04	0.09	0.72	0.04	0.00	0.39	0.78	0.30
11RAT-DT095B10	ucm	301	1149.8	0.60									0.32
11RAT-DT095B11	ucm	300	1150.8	0.60	0.04	0.03	0.08	3.88	0.09	0.01	0.37	0.81	0.28
11RAT-DT095B12	ucm	299.1	1151.7	0.27									0.27
11RAT-DT095B13	ucm	297.5	1153.3	0.58									0.32

Sample-A series	Fm/Mb	m (measured)	m (base)	Pyrite Fe %	Oxalate Fe (%)	Dithionite Fe (%)	Na Acetate Fe (%)	Na Acetate Ca (%)	Na Acetate Mg (%)	Na Acetate Mn (%)	Fe(HR)/Fe(T)	Fe(Py)/Fe(HR)	Fe/Al
11RAT-DT095B16	ucm	294.2	1156.6	0.39									0.22
11RAT-DT095B17	ucm	293.2	1157.6	0.59	0.04	0.03	0.12	1.88	0.14	0.01	0.44	0.76	0.26
11RAT-DT095B18	ucm	292.1	1158.7	0.52									0.22
11RAT-DT095B19	ucm	290.6	1160.2	0.61									0.19
11RAT-DT095B20	ucm	289.5	1161.3	0.43	0.02	0.02	0.06	4.57	0.06	0.01	0.46	0.81	0.22
11RAT-DT095B21	ucm	288.4	1162.4	0.47	0.03	0.02	0.05	0.74	0.10	0.00	0.50	0.82	0.21
11RAT-DT095B22	ucm	287.5	1163.3	0.29									0.20
11RAT-DT095B23	ucm	286.5	1164.3	1.49	0.04 B.D.		0.07	4.24	0.17	0.01	0.66	0.93	0.40
11RAT-DT095B24	ucm	285.5	1165.3	0.12									0.20
11RAT-DT095B25	ucm	284.4	1166.4	0.80									0.40
11RAT-DT095B26	ucm	283.4	1167.4	0.63	0.01	0.02	0.05	1.47	0.05	0.00	0.52	0.89	0.23
11RAT-DT095B27	ucm	282.4	1168.4	0.23									0.19
11RAT-DT095B28	ucm	281.4	1169.4	0.42									0.17
11RAT-DT095B29	ucm	280.3	1170.5	0.38									0.17
11RAT-DT095B30	ucm	278.8	1172	0.50	0.17	0.03	0.09	10.08	0.10	0.02	0.54	0.63	0.26
11RAT-DT095B31	ucm	278.1	1172.7	0.87	0.12 B.D.		0.07	12.00	0.13	0.02			0.33
11RAT-DT095B32	ucm	276.8	1174	0.90	0.01	0.03	0.06	6.58	0.07	0.01	0.36	0.90	0.46
11RAT-DT095B33	ucm	275.8	1175	0.40									0.29
11RAT-DT095B34	ucm	274.2	1176.6	0.91	0.04	0.03	0.06	1.46	0.19	0.00	0.54	0.87	0.30
11RAT-DT095B35	ucm	272.6	1178.2	0.49									0.50
11RAT-DT095B36	ucm	271.4	1179.4	0.31	0.03	0.05	0.18	4.69	2.40	0.01	0.39	0.55	0.38
11RAT-DT095B37	ucm	270.5	1180.3	0.25	0.12	0.11	0.13	0.13	0.17	0.00	0.21	0.41	0.48
11RAT-DT095B38	ucm	269.4	1181.4	0.02									0.78
11RAT-DT130A01	ucm	0	1600.1	0.04									
11RAT-DT130A08	ucm	0.9	1601	0.07	0.01	0.00	0.09	9.83	1.02	0.01			
11RAT-DT130A11	ucm	1.05	1601.15	0.06	0.01	0.11	0.33	12.04	4.96	0.02			
11RAT-DT130A12	ucm	1.25	1601.35	0.11	0.01	0.04	0.28	15.64	6.00	0.02			
10RAT-DT110-B01	Ft. Collinson	151.5	173.7	0.70	0.05	0.09	0.11	0.17	0.07	0.00			
10RAT-DT110-B02	Ft. Collinson	154.5	170.7	0.17 B.D.		0.02	0.06	28.81	0.26	0.01			
10RAT-DT110-B04	Ft. Collinson	157.5	167.7	1.45	0.02	0.04	0.09	3.73	0.11	0.01			
10RAT-DT110-B07	Ft. Collinson	167	158.2		0.04	0.04	0.10	3.04	0.13	0.01			
10RAT-DT110-B07	Ft. Collinson	169.6	155.6	0.64									
10RAT-DT110-B10	Ft. Collinson	142.9	182.3	0.63	0.04	0.16	0.09	0.20	0.07	0.00			
10RAT-DT110-B13	Ft. Collinson	195.1	130.1	1.47	0.06	0.17	0.08	3.26	0.05	0.01			
10RAT-DT110-B14	Ft. Collinson	203.1	122.1	0.69	0.01	0.03	0.09	7.65	0.30	0.01			
10RAT-DT110-B18	Boot Inlet	234.2	91.0		0.02	0.03	0.31	7.41	1.86	0.03			
10RAT-DT110-B23	Boot Inlet	254.6	70.6	0.67	0.01	0.04	0.20	11.65	0.56	0.01			
10RAT-DT110-B29	Boot Inlet	278.1	47.1	0.59	0.01	0.03	0.17	17.65	0.45	0.01			

Sample-A series	Fm/Mb	m (measured)	m (base)	Pyrite Fe %	Oxalate Fe (%)	Dithionite Fe(%)	Na			Na			Na			Fe(HR)/ Fe(Py)/ Fe(T)	Fe (HR)	Fe/Al
							Fe (%)	Acetate Fe (%)	Ca (%)	Acetate Mg (%)	Mn (%)	Acetate Mn (%)	Fe (%)	Ca (%)	Mg (%)			
10RAT-DT110-B33	Boot Inlet	291.6	33.6	0.64	0.02	0.04	0.03	0.03	31.36	0.20	0.01							
10RAT-DT110-B36	Boot Inlet	298.4	26.8	0.74	0.01	0.02	0.10	13.59	0.19	0.01								
10RAT-DT110-B39	Boot Inlet	313.2	12	0.31														

● 10RAT-DT055A10 C-isotope sample	▨ Satinspar veins		Dolostone		Siltstone
▲ 10RAT-DT055A09 Thin section sample	⤿ Ripple cross lamination		Calcareous mudstone		Sandstone
△ 10RAT-DT022A02 Representative sample	⬢ Selenite crystals		Gypsiferous lime mudstone		Cross-bedded intraclast dolograinstone
— Whipshy shale	⤿ Lenticular fine grained sand		Diabase sill		Cross-bedded intraclast grainstone
≡ Normal graded bed	⤿ Rip-up clasts		Chickenwire texture		Dolomitic sandstone
▨ Calcite veining	⤿ Wave ripples		Bedded gypsum		Intraclast packstone
⊕ Evaporite nodules	⊙ Hydrothermal alteration (vugs/veining/breccia)		Ooid dolograinstone		Stromatolitic dolostone
⊙ Ooids	⤿ HCS		Stromatolitic limestone		
⤿ Water escape structures	⤿ Nodular bedded				
⤿ Low angle cross-bedding	⤿ Wavy bedded				
⤿ Flooding surface	≡ Parallel laminated				
⤿ Compactional bedding	⤿ Intracast breccia				
▨ Tempestite	⬢ Carbonate nodules				
□ Halite casts	⤿ Teepee Structures				
⤿ Scours	⬢ Carbonaceous				
⤿ Exposure surface	▨ Quartz sand				
⬢ Thrombotic	⤿ Flute/gutter casts				
⤿ Columnar stromatolites	⤿ Molartooth structures				
⤿ Digitate stromatolites	⤿ Dessication/mud cracks				
⤿ Conophyton stromatolites	⤿ Upward shallowing				
⤿ Hemispheric stromatolites	⬢ Chert nodules				
⬢ Oncoids	⤿ Wavy laminated				
⬢ Evaporite crusts	Py Pyrite				
⬢ Evaporite casts/molds	⬢ Ball and pillow structures				
⬢ Synsedimentary fissures	⤿ Load structures				
★★ Mg-Si mineral	⤿ Channel and fill				
	⤿ Stylolites				
	⤿ Herringbone cross-bedding				
	⤿ Planar-tabular cross-bedding				
	⤿ Trough cross-bedding/laminae				
	LLH Laterally-linked hemispheric stromatolites				
	⬢ Rudstone				
	⤿ Soft-sediment deformation/slumping				

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Appendix 5 Legend for measured stratigraphic sections

Appendix 6 10DTS1, Minto Inlet Formation

Appendix 7 10DTS2, Minto Inlet Formation

Appendix 8 10DTS3, Lower Carbonate member

Appendix 9 10DTS4a, Black Shale member

10DTS4b - Wynniatt, black shale member

Appendix 10 10DTS4b, Black Shale member

Appendix 11 10DTS5, Black Shale and Stromatolitic Carbonate members

Appendix 12 10DTS6 and 10DTS12, Upper Carbonate member

Appendix 13 10DTS7, Stromatolitic Carbonate member

Appendix 14 10DTS8, Stromatolitic Carbonate and Upper Carbonate members

Appendix 15 10DTS9, Upper Carbonate member

Appendix 16 10DTS10, Stromatolitic Carbonate member

Appendix 17 10DTS11, Lower Carbonate and Black Shale members

Appendix 18 11DTS2, Upper Carbonate member

Appendix 19 10DTS3, Upper Carbonate member

Appendix 20 11DTS4, Upper Carbonate member

Appendix 21 11DTS5, Upper Carbonate member

Appendix 22 11DTS6, Fort Collinson Formation

Appendix 23 11DTS7, Upper Carbonate member

Appendix 24 11DTS8, Jago Bay Formation

11DTS09 - Wynniatt
Formation, lower carbonate
to stromatolitic carbonate members

Appendix 25 11DTS9; Lower Carbonate, Black Shale, and Stromatolitic Carbonate members

GNME drill core 2006-03 Jago Bay and Minto Inlet fms

Appendix 26 GNME drill core 06-03, Jago Bay and Minto Inlet formations

Appendix 27 Section location map