

Stratigraphic and Carbon Isotope Evolution of an Ediacaran Mixed
Siliciclastic Deep-Marine Base-of-Slope System, First Isaac Carbonate,
Windermere Supergroup, Canadian Cordillera, British Columbia.

Dylan Cochrane
B.Sc. Brock University, 2015

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

Ottawa-Carleton Geoscience Centre
Faculty of Science
University of Ottawa

© Dylan Cochrane, Ottawa, Canada, 2018

Abstract

The first Isaac carbonate (FIC) is a mixed siliciclastic-carbonate base-of-slope succession in the Neoproterozoic Windermere Supergroup (WSG). Outstanding outcrop exposure at three study areas provided an excellent opportunity to observe the stratigraphic and isotopic evolution of an ancient deepwater mixed turbidite system. Based on lithological and stratal dimensions, the FIC can be subdivided into lower and upper parts suggesting temporal changes in patterns of sediment transport and deposition. $\delta^{13}\text{C}_{\text{carb}}$ also changes from -5.2‰ at the base of the FIC to 2.5‰ in the middle and then decreases to -6.3‰ at the top. Notably, the $\delta^{13}\text{C}_{\text{carb}}$ of primary cement in FIC strata is substantially more positive than most other Neoproterozoic deep-marine sections, suggesting the retention of their original shallow-marine isotopic signature. Nevertheless, this trend potentially correlates with the EN2 excursion in China and therefore the Gaskiers glaciation (~580 Ma), although better age control of WSG is needed to corroborate this correlation.

Résumé

La première séquence de carbonates Isaac (FIC) est un mélange siliciclastique-carbonate déposée au bas d'une pente continentale appartenant au Supergroupe Néoprotérozoïque de Windermere (WSG). L'excellente exposition de trois affleurements a permis d'observer l'évolution stratigraphique et isotopique d'un ancien système de turbidite profonde. Selon la lithologie et les dimensions de strates, le FIC peut être divisé en deux parties, ce qui suggère un changement temporel de transport et déposition. De plus, le $\delta^{13}\text{C}_{\text{carb}}$ varie de -5.2‰, à la base du FIC, à 2.5‰ au centre et diminue à -6.3‰ dans le haut. À noter, le $\delta^{13}\text{C}_{\text{carb}}$ du ciment primaire dans le FIC est nettement plus élevé que la plupart des autres sections Néoprotérozoïques sous-marines, suggérant la préservation de leur signature isotopique peu profonde originale. Néanmoins, cette tendance peut potentiellement être corrélée avec l'excursion EN2 en Chine et la glaciation Gaskiers (~580 Ma). Cependant un meilleur contrôle sur l'âge du WSG est nécessaire pour corroborer cette corrélation.

Extended Abstract

The first Isaac carbonate (FIC) is a mixed siliciclastic-carbonate, base-of-slope succession in the otherwise siliciclastic-dominated deep-marine Windermere Supergroup (WSG). Excellent outcrop exposure at three study areas (Castle Creek, Hill Section and Milk River) provided the opportunity to observe the spatial and temporal stratigraphic and isotopic evolution of an ancient deepwater mixed system. Based on detailed sedimentological analysis seven stratal elements make-up the FIC (channel complexes, proximal levees, distal levees, “123” units, scour dominated units, debrites and slides), but their distribution is not uniform across the study areas. Instead, the FIC can be divided into lower and upper parts with the former characterized by uniformity of stratal units and the latter marked by lithological heterogeneity and discontinuity between the study areas. This suggests that during early FIC development channel belts were mobile and wandered across much of the slope, but then patterns of transport and deposition changed with CC-HS situated closer to the primary transport pathway whereas Milk River was situated on the adjacent distal mud-rich levee. The $\delta^{13}\text{C}_{\text{carb}}$ of seafloor and early burial microspar, which was neomorphosed from primary detrital aragonite, in FIC carbonate strata is substantially more positive than those reported from other similarly aged deep-marine sections, suggesting the retention of the primary shallow-marine isotopic signature even after downslope transport. At the base of the FIC $\delta^{13}\text{C}_{\text{carb}}$ is initially negative and progressively increases to 2.5‰ in the middle of the succession, after which it plateaus for ~ 50 m before declining to a nadir of -6.3‰ at the top. This trend closely resembles the trend reported in the upper June beds-Gametrail Formation of the WSG in the Mackenzie Mountains, but its interpreted association with the Shuram-Wonoka anomaly makes the correlation unlikely. Instead, the FIC excursion potentially correlates with the EN2 excursion in the Doushantuo Formation of South China (and its potential global correlatives), which has been interpreted to be associated with the Gaskiers glaciation (~ 580 Ma). Therefore, the FIC might preserve the isotopic record preceding the Gaskiers, which notably lacks any sedimentological evidence in the WSG, however, better age control of WSG strata is needed to test this correlation.

Acknowledgements

This thesis would not have been possible with the help of numerous individuals along the way and I am grateful to every one of them. The first thank you goes out to my supervisor Bill Arnott for leading me on this journey and introducing me to the world of deep-marine sedimentology. When I started this project I knew very little about carbonates and $\delta^{13}\text{C}_{\text{carb}}$ chemostratigraphy (although I of course nodded my head when you first asked me if I was familiar with it and wanted to do a project on it four days before I left for the field), but it took me down an interesting path and now I could not imagine having studied anything else. You were always there to sit down and handle any questions I threw your way and you allowed me to do the geochemical analyses I deemed necessary, regardless of the cost. Your energy in the field was contagious and your love of sedimentology was evident every time you discussed it. During my first field season I said this, and it is going to stay with me for a very long time: “If I end up doing something I love $\frac{1}{4}$ as much as Bill loves his job, I will be a very happy man.”

I would also like to thank Dr. André Desrochers and Dr. George Dix for taking the time to serve on my committee and providing feedback on how to improve this thesis.

Thank you to the Windermere Consortium for funding the majority of this project: NSERC, Anadarko, Apache, Husky Energy, Nexen and Statoil. Additional funding for this project was provided by a 2016 Ontario Graduate Scholarship and a 2016-2017 NSERC CGS-Masters Scholarship.

We stand upon the shoulders of those who came before us, therefore a huge thank you is owed to Dr. Lilian Navarro, whose PhD research spawned this project. You helped me out countless times: from sending me papers to initially read in the field to get me started on my project, to answering my countless questions and commenting on various theories that I threw around and providing feedback on my presentations and the written portion of this thesis. Also, thank you for doing the XRD analysis on two of my samples after my thesis defense.

Thank you to my two field assistants: Quinn Dabros - you allowed me to start my thesis off on the right foot and pushed me to walk that gruelling hour plus hike to the hill section everyday – and Sean Ludzki – you battled through a broken leg and improved everyday to the point where you became a great geologist in the field. You also provided hours of great conversation about baseball (where you told me of your obsession with Joe Mauer) and Harry Potter theories. Without you two far less data would have been obtained for this thesis.

Thank you to the rest of the Windermere research group: Viktor Terlaky, Nataša Popović, Katrina Angus, Seungmin Leo Lee, Genevieve Huyer, Dave Lowe, Derrick Midwinter, Mike Tilston,

Anika Bergen, Curran Wearmouth, Jag Ningthoujam, Tyler Billington, Gabriela Milczarek, Celeste Cunningham and Nicole Miklovich for helping me along the way. Your questions and comments during Friday meetings were very helpful and pushed this thesis forward. Despite not being part of our research group, I would also like to thank Wilder Greenman from Carleton University. We were both doing projects focusing on Neoproterozoic chemostratigraphy (although our sections differed in age by ~300 Myr) and it was helpful to bounce ideas back and forth and see how our studies progressed.

I would like to thank George Mrazek and Alain Mauviel for preparing my thin sections and Pascale Daoust for helping me stain them. Thank you also Paul Middlestead, Wendy Abdi, Patricia Wickham and everybody else at the G.G. Hatch Stable Isotope Laboratory for putting up with my loud drilling and performing the analyses on my samples. Glenn Poirier for allowing me to do cathodoluminescence microscopy and Dr. Nimal De Silva for analyzing some of my samples for Mn/Sr.

Thank you to the support staff in the Department of Earth Sciences: H  l  ne De Gouffe, Liss-Robin Murphy, Caroline Bisson and Caroline Poirier for answering my many questions and helping me out.

I would also like to thank all of my fellow graduate students. You guys made my time at the University of Ottawa an enjoyable experience.

Merci beaucoup    Emilie Gagnon d'  tre l   pour moi ces deux derni  res ann  es, malgr   qu'on a habit   loin pendant ce temps. I thoroughly enjoy every second that I spend with you and I am so proud of how you've progressed in your career thus far. I couldn't imagine completing this thesis without your support. Je t'aime beaucoup.

Last, but not least, I would like to thank my family. Particularly my parents, Allen Cochrane and Sherry Pepper, for everything they have done for me over the past 26.5 years. Both of you have always supported everything I wanted to do and have taught me a lot about life and I am still learning lessons from you everyday (some I wish I figured out when I was 18). This thesis would not have started without you guys and its completion is as much your accomplishment as it is mine. Thank you also to my Nanny, Mona Pepper, for helping me out whenever I needed it and allowing me stay at her place over the last few months of writing this thesis.

Table of Contents

Abstract	ii
Résumé	iii
Extended Abstract	iv
Acknowledgements	V
List of Figures	ix
List of Tables	xiv
List of Abbreviations	xv
Chapter 1: Thesis Introduction	1
I Thesis Rationale	1
II Windermere Supergroup Overview	3
II.I Location of the Windermere Supergroup	3
II.II Tectonic Setting and Metamorphism	3
II.III Development of the Windermere Supergroup	8
II.III.I Syn-Rift	11
II.III.II Post-Rift	11
II.IV Geochronological Constraints	14
II.V Provenance and Paleocurrents	14
II.VI Regional Markers of the Windermere Supergroup	15
II.VII Study Areas	16
II.VII.I Castle Creek Study Area	17
II.VII.II Milk River Study Area	19
II.VII.III First Isaac Carbonate	21
II.VIII Previous Work	22
III Deep-Marine Sedimentation Processes	23
III.I Classification of Sediment-Gravity Flows	24
III.I.I Frictional Flows	24
III.I.II Cohesive Flows	25
III.II Mass Movement Deposits	26
III.III Deep-Marine Channels and Levees	27
IV Neoproterozoic Climate and $\delta^{13}\text{C}_{\text{carb}}$ Chemostratigraphy	30
IV.I Neoproterozoic Glaciations	31
IV.I.I Proposed Models for Neoproterozoic Glaciations	35
IV.II $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ Overview	38
IV.II.I Neoproterozoic Carbon Isotope Trends	41
IV.III Previous Post-Marinoan Windermere Isotope Studies	45
V Thesis Objectives and Structure	50
Chapter 2: Comparison of Stratal Elements and Their Spatial Distribution in a Neoproterozoic Mixed Carbonate-Siliciclastic Base-of-Slope System, Windermere Supergroup, Canadian Cordillera, British Columbia	53
I Introduction	53
II Geologic Setting.....	56
II.I Windermere Supergroup	56
II.II First Isaac Carbonate	59
II.III Study Areas	61
III Methodology.....	62
IV Results	65
IV.I Facies	65

IV.I.I Facies 1: Structureless, Normally Graded Sandstone	66
IV.I.II Facies 2: Traction-Structured Sandstone.....	71
IV.I.III Facies 3: Mud-Rich Turbidites	74
IV.I.IV Facies 4: Poorly Sorted, Calcilutite-Rich Deposits.....	76
IV.I.V Facies 5: Internally Sheared and Deformed Strata	79
IV.II Stratal Elements	83
IV.II.I Channel Complexes	85
IV.II.II Proximal Levees	91
IV.II.III Distal Levees.....	98
IV.II.IV 123s	101
IV.II.V Scour Dominated Sandstones	111
IV.II.VI Mass Transport Deposits	113
V Discussion	116
Chapter 3: Geochemical Evolution of an Ediacaran Mixed Carbonate-Siliciclastic Continental Slope System, Windermere Supergroup, Canadian Cordillera, British Columbia	123
I Introduction	123
II Geologic Setting.....	126
II.I Windermere Supergroup	126
II.II Study Areas	129
III Lithological and Stratal Make-Up of the FIC	129
III.I Facies	129
III.II Stratal Elements	136
III.II.I Channel Complexes	137
III.II.II Proximal Levees	138
III.II.III Distal Levees.....	141
III.II.IV 123s	143
III.II.V Scour Dominated Sandstones	144
III.II.VI Mass Transport Deposits	145
III.III Sequence Stratigraphy of the FIC	145
IV Geochemistry	146
IV.I Methodology	146
IV.II Results	149
IV.II.I Petrography and Cathodoluminescence Microscopy	149
IV.II.II Geochemistry	156
V Discussion	163
V.I Strength of the Carbon Isotopic Signal	163
V.II Source of the Carbon Isotopic Signal	166
V.III Connection Between Eustasy and $\delta^{13}\text{C}_{\text{carb}}$	168
V.IV Correlation to the Windermere Supergroup in the Mackenzie Mountains?	169
V.V Correlation to Worldwide Ediacaran $\delta^{13}\text{C}_{\text{carb}}$ Datasets	171
Chapter 4: Thesis Conclusions and Areas for Future Research	176
I Conclusions	176
II Areas for Future Research	180
References	182
Appendix A: Samples analyzed for $\delta^{13}\text{C}_{\text{carb}}$, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}_{\text{org}}$	212
Appendix B: Samples analyzed for manganese and strontium contents	224
Appendix C: Stratal element, facies and lithology proportion in each study area	225
Appendix D: Lithology and facies proportions in each study area	228

List of Figures

Figure 1.1: Location of the Windermere Supergroup through Mexico, the United States and Canada. (Redrawn from Ross, 1991 and Terlaky, 2014).	4
Figure 1.2: Schematic cross-section of western North America ca. ~ 600 Ma. (modified from Ross and Arnott, 2007).	5
Figure 1.3: Map of western Canada showing the five morpho-tectonic belts that comprise the Canadian Cordillera, resulting from collisional tectonism along western Laurentia between 185-50 Ma (Modified from Ross et al., 2002 and Smith, 2009).	6
Figure 1.4: Dimensional comparison of the reconstructed Windermere turbidite system and other turbidite basins worldwide (Modified from Ross and Arnott, 2007).	8
Figure 1.5: Neoproterozoic global reconstructions by Li et al. (2013) for four different time intervals: (A) 825 Ma, with Laurentia indicated by red L; (B) 720 Ma; (C) 635 Ma; (D) 580 Ma. Windermere margin indicated by pink diamond in B-D.	9
Figure 1.6: Generalized stratigraphic column of the Windermere Supergroup in the southern Canadian Cordillera. This study focuses on the first Isaac carbonate, which is indicated by the red rectangle (modified from Ross and Arnott, 2007).	10
Figure 1.7: Stratigraphic nomenclature of the Windermere Supergroup in the southern Canadian Cordillera (modified from Smith et al., 2014).	12
Figure 1.8: (A) Satellite image with study areas indicated by red polygons (background image © Province of British Columbia 2017, © CNES/Airbus 2017, © DigitalGlobe 2017, © Google 2017). (B) Cross-section of the regional geology of the Castle Creek Area (modified from Ross and Ferguson 2003).	18
Figure 1.9: (A) Aerial photograph of the Castle Creek and Hill Section study areas and (B) Photograph of the Milk River study area. Blue box represents outcrop exposure of the first Isaac Carbonate.	20
Figure 1.10: Composite stratigraphic section of the Castle Creek study area (modified from Ross & Arnott, 2007). (B) Generalized log through the FIC showing long- (blue; 3 rd order) and short-term (green; 4 th /5 th order) eustatic changes and mineralogical composition of sediment supply into local Windermere basin. (modified from Navarro, 2016).	22
Figure 1.11: Idealized Bouma sequence deposited by a decelerating, dilute turbidity current; (redrawn after Bouma, 1962; Terlaky, 2014; Angus, 2016).	26
Figure 1.12: The downslope transition of sediment gravity flows (modified from Shanmugam et al., 1994; Terlaky, 2014).	27
Figure 1.13: (A) 3D seismic horizon slice of a submarine channel complex in offshore Angola (from Abreu et al., 2003) and (B) seismic profile across channel-levee complex in the Amazon fan (from Piper and Normark, 2001).	29

Figure 1.14: Paleolatitudinal distribution of diamictites associated with the three major Neoproterozoic glaciations (modified after Li et al., 2013).....	32
Figure 1.15: Present day geographical distribution of deposits associated with the (A) Sturtian (B) Marinoan and (C) Gaskiers glaciations. The red circles indicate glacial deposits in the Windermere Supergroup of western Canada. (modified from Halverson, 2006).	34
Figure 1.16: Schematic showing the estimated changes in global mean surface temperature during the development and onset of a snowball Earth event (from Hoffman and Schrag, 2002).	37
Figure 1.17: Composite Neoproterozoic $\delta^{13}\text{C}$ record from marine carbonates (modified after Halverson et al., 2010).	42
Figure 1.18: Post-Marinoan $\delta^{13}\text{C}_{\text{carb}}$ record of the Windermere Supergroup in the (A) southern Canadian Cordillera (Smith, 2009) and (B) Mackenzie Mountains (modified from MacDonald et al., 2013).	48
Figure 1.19: Stratigraphic correlation of Cryogenian to Ediacaran strata of the Windermere Supergroup in the southern Canadian Cordillera, Monkman Pass area and Mackenzie Mountains, which represent different paleogeographic positions along a shelf to basin profile (modified from Smith, 2009).	49
Figure 2.1: (A) Outcrop of the Windermere Supergroup in western North America (redrawn from Ross, 1991 and Terlaky, 2014). (B) Generalized stratigraphic section of the WSG in the southern Canadian Cordillera (modified after Ross and Arnott, 2007). (C) Satellite image with the study areas indicated by red polygons (background image © Province of British Columbia 2017, © CNES/Airbus 2017, © DigitalGlobe 2017, © Google 2017).	58
Figure 2.2: Generalized stratigraphic log through the FIC at CC showing short- (green) and long-term (blue) changes in sea level and composition of sediment supply into the deep-water Windermere basin (modified from Navarro, 2016).	62
Figure 2.3: Aerial photographs of the study areas with major units correlated; measured stratigraphic logs indicated by yellow lines. (A) Castle Creek, (B) Hill Section, (C) Milk River.	64
Figure 2.4: Examples of Facies 1A (A-D) and 1B (E-F) from the FIC.	67
Figure 2.5: Stained photomicrographs of Facies 1 strata.	69
Figure 2.6: Field photographs of facies 2 strata.	73
Figure 2.7: Photographs and stained photomicrographs of facies 3 strata.	76
Figure 2.8: Field photographs of Facies 4 (debrites) in the FIC.	78

Figure 2.9: Outcrop examples of facies 5 strata (slides).	80
Figure 2.10: Correlation of stratal elements identified in CC and HS. White lettering indicates Unit names (see text for details).	82
Figure 2.11: Correlation of stratal elements identified at MR. White lettering indicates unit names (see text for details).	84
Figure 2.12: (A) Stratigraphic section through BST 1 at HS. (B-F) Representative photographs of channel deposits in the FIC. (G) Lithological (left) and facies (right) make-up of FIC channel complexes.	86
Figure 2.13: (A) Stratigraphic section of CT3 at MR-S with sharply-bounded mud-rich and sand-rich (S) turbidite packages. (B-F) Representative photographs of proximal-levee deposits in the FIC. (G) Lithological (left) and facies (right) make-up of FIC proximal levee units.	92
Figure 2.14: Bergen's (2017) schematic model for the deposition of lower and upper levee parts of levee packages.	95
Figure 2.15: Schematic model for the lateral transition from PL 2 (A) to BST 4 (D) at Milk River.	97
Figure 2.16:(A) Stratigraphic section of CT 2 at HS. (B-E) Representative photographs of Siliceous calcilitite-rich distal levee deposits in the FIC. (F) Lithological (left) and facies (right) make-up of siliciclastic mud-rich distal levee deposits in the FIC. (G) Lithological (left) and facies (right) make-up of calcilitite-rich distal levee deposits in the FIC.	99
Figure 2.17: Representative photographs of 123 packages in the FIC. (A-C) 123s at CC-S showing the characteristic repetitive stacking of the three packages (labelled 1, 2, 3). (D-G) Lateral transition of sandstones 7 and 29 at Castle Creek south. (H-J) Lateral transition of sandstone 5 from north to south at Castle Creek south.	102
Figure 2.18: 123 sections logged at Castle Creek south with the sandstone beds correlated across the outcrop indicated by brown coloured areas. Red arrows indicate stratigraphic pinch-outs.	104
Figure 2.19: (A) Histogram showing number of traceable sandstone versus stratigraphic height in the 123s at CC-S. (B) Histogram showing the location of the thickest part of each sandstone traced in relation to the middle of the outcrop exposure in the 123s at CC-S. (C) Plot of vertical (stratigraphic) height of each sandstone bed versus the location of its thickest part relative to the middle of the outcrop at CC-S. (D) Three bed rolling average of the maximum thickness of each sandstone versus stratigraphic height. (E) Schematic of the transition matrix illustrating the vertical succession of facies in the 123 units. (F) Lithological (left) and facies (right) make-up of 123 units in the FIC.	106
Figure 2.20: Transitions observed in the three detailed stratigraphic logs measured through the 123 units. Bold numbers highlight the most probable transitions.	109

Figure 2.21: (A-C) Representative photographs of strata in a scour-dominated unit at MR-N. (D) Lithological (left) and facies (right) make-up of scour dominated units in the FIC. ...	112
Figure 2.22: SE to NW correlation from CC-S to MR-N showing the spatial distribution of stratal elements in the FIC; datum is the base of BST 1; unit names in white.	119
Figure 2.23: Abundance of stratal elements in the study areas. (A) Stratal element percentage relative to total thickness of the FIC. (B) Cumulative stratigraphic thickness of individual stratal elements identified in the FIC. (C) Stratal element percentage relative to total thickness of the FIC below the top of BST 1. (D) Stratal element percentage relative to thickness of the FIC above BST 1.	120
Figure 2.24: Schematic model illustrating the temporal evolution of the FIC.	121
Figure 3.1: Representative photographs showing the facies and depositional elements observed in the FIC.	135
Figure 3.2: SE to NW correlation from CC-S to MR-N showing the spatial distribution of stratal elements in the FIC; datum is the base of BST 1; unit names in white..	140
Figure 3.3: Representative photomicrographs of carbonate cements observed in the FIC.	153
Figure 3.4: Measured X-ray Diffractograms obtained from siliceous calcilutite strata in CT 2 and CT 3 to identify abundances of calcite, chlorite, ferroan dolomite, muscovite and quartz.	154
Figure 3.5: Relative paragenetic timing of FIC carbonate cements.	156
Figure 3.6: Correlation of strata between CC-S, HS and MR-S study areas. Stable isotope data is presented to right of each stratigraphic column. $\delta^{13}\text{C}_{\text{carb}}$, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}_{\text{org}}$ samples are denoted by circles, diamonds and triangles, respectively. Red and black shapes indicate samples collected as part of this study, with the black corresponding to carbonate clasts and red to matrix cements, and green shapes are the unpublished data of G.M. Ross (2003).	157
Figure 3.7: Cross plots of geochemical data from the FIC (A) $\delta^{13}\text{C}_{\text{carb}}$ vs. $\delta^{18}\text{O}$, (B) $\Delta\delta$ ($\delta^{13}\text{C}_{\text{carb-org}}$) vs. total organic carbon, (C) $\delta^{13}\text{C}_{\text{carb}}$ vs. total organic carbon and (D) $\delta^{13}\text{C}_{\text{carb}}$ vs. Mn/Sr.	160
Figure 3.8: Combined FIC $\delta^{13}\text{C}_{\text{carb}}$ from CC and HS with eustatic curve and sequence stratigraphy based on the chemostratigraphy (black) and lithostratigraphy (orange; Navarro, 2016). The light blue region within the sea-level curve corresponds to carbonate production on the shelf platform.	165

Figure 3.9: Comparison of $\delta^{13}\text{C}_{\text{carb}}$ chemostratigraphies for the Ediacaran in nine regions: the SCC in western Canada (Smith, 2009; this study), Mackenzie Mountains in northwestern Canada (MacDonald et al., 2013), South China (Jiang et al., 2007), Death Valley in western U.S.A. (Corsetti and Kaufman, 2003; Kaufman et al., 2007), northern (Halverson et al., 2005) and southern (Saylor et al., 2008) Namibia, Australia (Calver, 2000), and India (Kaufman et al., 2006). Excursions associated with the Shuram-Wonoka anomaly are highlighted in purple, whereas those potentially correlatable to EN-2 in China (Gaskiers glaciation) are highlighted in pink. 173

List of Tables

Table 2.1: Overview of rock classification scheme utilized in this study.	66
Table 2.2: Stratal elements in the FIC and the proportional make-up of their constituent facies.	83
Table 3.1: Summary of deep-water facies (siliciclastic and carbonate) in the first Isaac carbonate.	131
Table 3.2: Carbonate cements identified in FIC strata with their respective thin section and cathodoluminescence descriptions and interpreted timing of cementation.	151
Table 3.3: Isotopic values measured in carbonate clasts and their host matrix. There appears to be no correlation between $\delta^{13}\text{C}_{\text{carb}}$ in clasts compared to the host matrix, with differences ranging from -2.1‰ to 2.9‰.	159
Table 3.4: Result table for an F-test comparing the $\delta^{13}\text{C}_{\text{carb}}$ values within clasts in mass-transport deposits and mixed siliciclastic-carbonate channel complexes to those of their host matrix. The F value is less than F-critical, thus the null hypothesis that the variance is strongly overlapping cannot be rejected.	159

List of Abbreviations

123-1	Lower 123 unit
123-2	Upper 123 unit
$^{86}\text{Sr}/^{87}\text{Sr}$	Strontium isotope ratio
BIF	Banded-iron formation
BST	Bacon sandstone channel complex
BST 1-4	Bacon channel complexes 1-4
C 1-5	Calcite cements 1-5
CC	Castle Creek study area
CC-S	Castle Creek south
CC-N	Castle Creek north
CD	Calciobrite
CD 1-3	Calciobrites 1-3
CL	Cathodoluminescence
CT 1-4	Calcioburidite units 1-4
D 1-3	Dolomite cements 1-3
Dm	Decametre
$\delta^{13}\text{C}_{\text{carb}}$	Carbon isotope composition of carbonate carbon (inorganic)
$\delta^{13}\text{C}_{\text{org}}$	Carbon isotope composition of sedimentary organic matter
$\delta^{18}\text{O}$	Oxygen isotope composition of carbonate
$\delta^{34}\text{S}$	Sulfur isotope composition in sedimentary rocks
$\Delta\delta$	$\delta^{13}\text{C}_{\text{carb}} - \delta^{13}\text{C}_{\text{org}}$
DIC	Dissolved inorganic carbon
DOC	Dissolved organic carbon
Dm	Decametre
EN1	Ediacaran $\delta^{13}\text{C}_{\text{carb}}$ negative excursion 1 (China – Marinoan cap carbonates)
EN2	Ediacaran $\delta^{13}\text{C}_{\text{carb}}$ negative excursion 2 (China - Gaskiers glaciation)
EN3	Ediacaran $\delta^{13}\text{C}_{\text{carb}}$ negative excursion 3 (China – Shuram-Wonoka anomaly)
EN4	Ediacaran $\delta^{13}\text{C}_{\text{carb}}$ negative excursion 4 (China – Precambrian-Cambrian boundary)
F1A	Facies 1A
F1B	Facies 1B
FIC	First/lower Isaac carbonate
FSST	Falling stage systems tract
GBR	Great Barrier Reef
GSM	Geikie Siding Member of the Old Fort Point Formation
HS	Hill Section study area
HST	Highstand systems tract
ICP-MS	Inductively coupled plasma mass spectrometry
IRMS	Isotope ratio mass spectrometer
Ka	Thousand years (datum)
Kbar	Kilobar
LGM	Last glacial maximum
LST	Lowstand systems tract
Ma	Million years (datum)
MR	Milk River study area
MR-S	Milk River south
MR-N	Milk River north

MTC	Mass transport complex
MTD	Mass transport deposits
Myr	Million years (duration)
NCC	Northern Canadian Cordillera
OFP	Old Fort Point Formation
PL 1-3	Proximal levee units 1-3
Re-Os	Rhenium-Osmium
S 1-3	Slide units 1-3
SB	Sequence boundary
SCC	Southern Canadian Cordillera
SD 1-2	Scour-dominated sandstone units 1-2
SIC	Second/upper Isaac carbonate
SRMT	Southern Rocky Mountain Trench
T _a	A division of Bouma sequence: massive to structureless sandstone
T _b	B division of Bouma sequence: planar-laminated sandstone
T _c	C division of Bouma sequence: ripple cross-stratified sandstone
TCD	Thermoconductivity detector
T _d	D division of Bouma sequence: discontinuous interlaminated silt-mud couplets
T _e	E division of Bouma sequence: structureless mudstone
TLM	Temple Lake Member of the Old Fort Point Formation
TOC	Total organic carbon
TST	Transgressive systems tract
U-Pb	Uranium lead dating
VPDB	Vienna Pee Dee Belemnite
WMM	Whitehorn Mountain Member of the Old Fort Point Formation
WSG	Windermere Supergroup

Chapter 1: Thesis Introduction

Part I Thesis Rationale

The Neoproterozoic Era (1000-541 Ma) was a period of extreme change in the Earth system that is unparalleled to anytime during the Phanerozoic. This era saw the assembly and breakup of the supercontinent Rodinia, three glaciations with potentially global extent (Sturtian, Marinoan and Gaskiers) that followed a ~ 1.5 billion-year-long interglacial, the return of banded-iron formations after a ~ 1 billion-year hiatus, globally correlatable carbonates formed in response to rapid climate change during early interglacials, and a shift to more oxic atmospheric conditions that may have stimulated the evolution of metazoan life (Ediacaran biota), which eventually led to the innovation of calcified animals during the Cambrian explosion (Kirshvink, 1992; Hoffman et al., 1998; Xiao and Kaufman, 2006; Halverson, 2006; Fairchild and Kennedy, 2007). Accordingly, many studies have focused on this time in Earth history and investigated the factors that led to these drastic changes. One proxy that has been widely used over the past ~ 25 years is the carbon isotopic compositions of carbonate rocks ($\delta^{13}\text{C}_{\text{carb}}$), whose secular variations have been shown to be repeatable at both regional and global scales, and combined with other stratigraphic tools (e.g. sequence stratigraphy and other marine proxy records) have improved the correlation of Neoproterozoic strata in the absence of biostratigraphic control (e.g. Knoll et al., 1986; Kaufman et al., 1997; Hoffman et al., 1998; Hoffman and Schrag, 2002; Halverson et al., 2005; Halverson et al., 2010; Tahata et al., 2013; MacDonald et al., 2013). Notably the Neoproterozoic was characterized by prolonged periods of elevated $\delta^{13}\text{C}_{\text{carb}}$ with five globally correlatable large-amplitude excursions, three during the Ediacaran Period (635-541 Ma), and two of those associated with major glacial episodes (Halverson et al., 2010).

The first Isaac carbonate (FIC) is a mixed carbonate-siliciclastic deep-marine base-of-slope succession within the Isaac Formation in the otherwise siliciclastic dominated Windermere Supergroup (WSG) turbidite system in the southern Canadian Cordillera. The FIC, therefore, represents a

stratigraphic outlier that occurs ~ 1 km stratigraphically above deposits interpreted to be associated with the transgression that following the Marinoan glaciation (~ 635 Ma; Ross et al., 1995; Smith, 2009; Smith et al., 2011; Smith et al., 2014 a,b) and was first identified as a regionally extensive unit in the Cariboo Mountains by Ross and Murphy (1988). However, in spite of its stratigraphic significance (Ross et al., 1989; Ross et al., 1995) the FIC has been largely unstudied except for the Ph.D thesis research of Lilian Navarro (2016). This study builds on the work of Navarro and has two principal objectives:

- 1) Despite being primary targets of hydrocarbon exploration (i.e. offshore Nova Scotia and West Africa) the understanding of continental slope mixed carbonate-siliciclastic or carbonate systems is dwarfed by their siliciclastic counterparts. Presented here is a detailed account of the stratal units that make-up a mixed siliciclastic system (FIC) in three study areas (Castle Creek, Hill Section and Milk River) that are separated by up to ~ 20 km. Moreover, the spatial and temporal distribution of the component stratal elements illustrate the underlying allogenic and autogenic processes that controlled the stratigraphic evolution;
- 2) Use the temporal changes of the $\delta^{13}\text{C}_{\text{carb}}$ signal to assess paleoenvironmental conditions during deposition of the FIC. These changes are then compared with other Ediacaran sections worldwide, notably shallow-marine strata of the Windermere Supergroup in the Mackenzie Mountains (Yukon) and the Doushantuo Formation in South China. This puts the post-Marinoan part of the Windermere Supergroup in a global context, and therefore the ability to assess development of the Windermere basin and paleo-oceanographic conditions during this time.

Part II Windermere Supergroup Overview

II.I Location of the Windermere Supergroup

The term “Windermere System” was first coined by Walker (1926) to describe sedimentary rocks that crop out in the Windermere Valley of southeastern British Columbia. Later the use of this term expanded to include a series of Neoproterozoic rocks exposed locally from northwestern Mexico and the western United States, continuously throughout most of southwestern Canada through to the Yukon-Alaska border region, a strike length of just over 4000 km (Figure 1.1; Ross and Arnott, 2007). In Mexico and the United States rocks of the Windermere Supergroup (WSG) comprise continental and shallow marine strata (Link et al., 1993), whereas those in the southern Canadian Cordillera consist of superbly exposed successions of deep-marine, siliciclastic-dominated meta-sedimentary rocks (Campbell et al., 1973; Ross et al., 1995). In the Mackenzie Mountains of northwestern Canada WSG rocks preserve extensive carbonate-rich shallow-marine and upper continental slope facies. (Figure 1.2; Ross and Arnott, 2007).

II.II Tectonic Setting and Metamorphism of Windermere Supergroup in the Canadian Cordillera

The present-day configuration of the Canadian Cordillera is divided into five NW-SE trending morphological belts, which from east to west are the Foreland fold and thrust belt, Omineca belt, Intermontane belt, Coast belt and Insular belt (Figure 1.3; Monger et al., 1982; Ross and Arnott, 2007). These belts are the result of Late Jurassic to Early Tertiary collisional tectonics associated with the accretion of two large allochthonous terranes along the western margin of ancestral North America (Laurentia), causing crustal thickening, thrust faulting, folding and metamorphism of autochthonous Proterozoic and Paleozoic sedimentary rocks (Reid et al., 2002). This orogen evolved in three stages, expressed in the current preservation of the aforementioned belts (Angus, 2016), beginning with deposition of the WSG and overlying Paleozoic strata through Neoproterozoic rifting and successive

development of a passive margin along present-day western Laurentia. This was followed by terrane accretion and subsequent orogenesis during the Paleozoic and Mesozoic to Eocene, causing the continent to expand laterally by ~ 500 km (Monger et al., 1982). Late Cretaceous to Early Eocene tectonism, principally strike-slip faults and regional extension, in the Cordillera makes up the final stage.

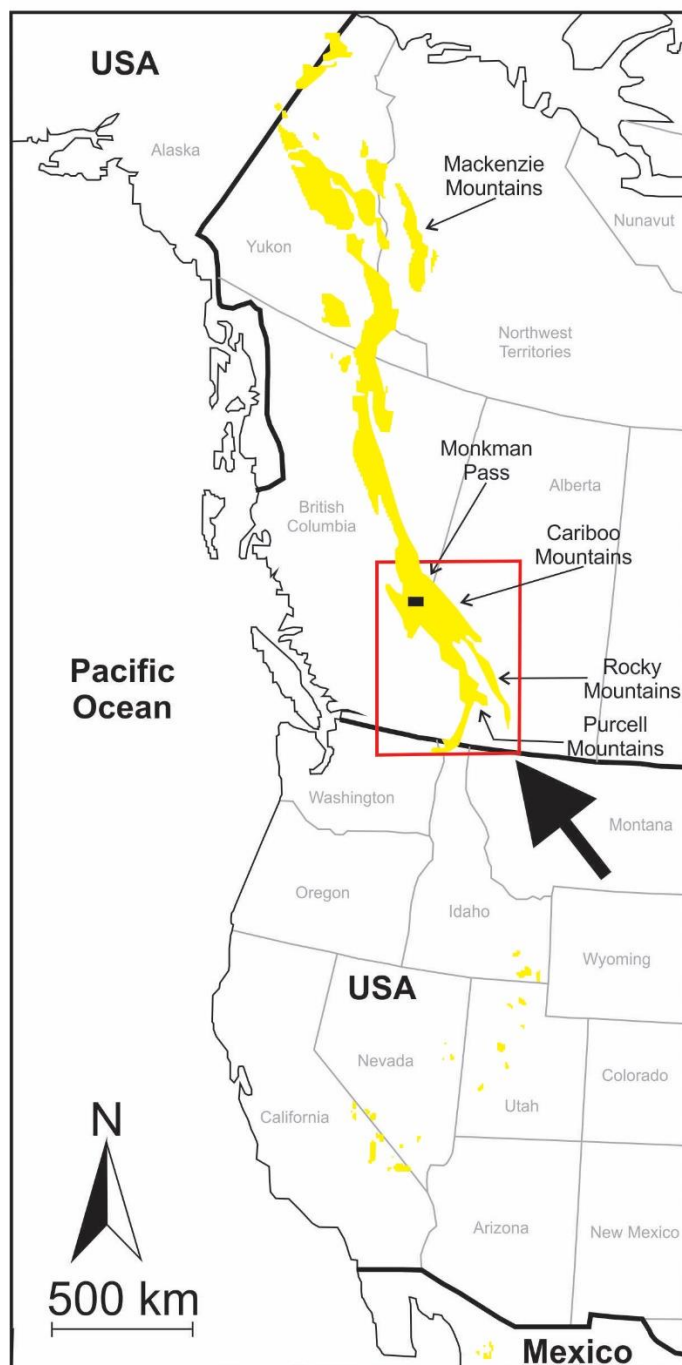


Figure 1.1: Location of the Windermere Supergroup through Mexico, the United States and Canada. The deep marine portion of the WSG is delineated by the red rectangle with black box indicating the study area. The thick, black arrow highlights the primary sediment transport direction into the Windermere turbidite system (Redrawn from Ross, 1991 and Terlaky, 2014).

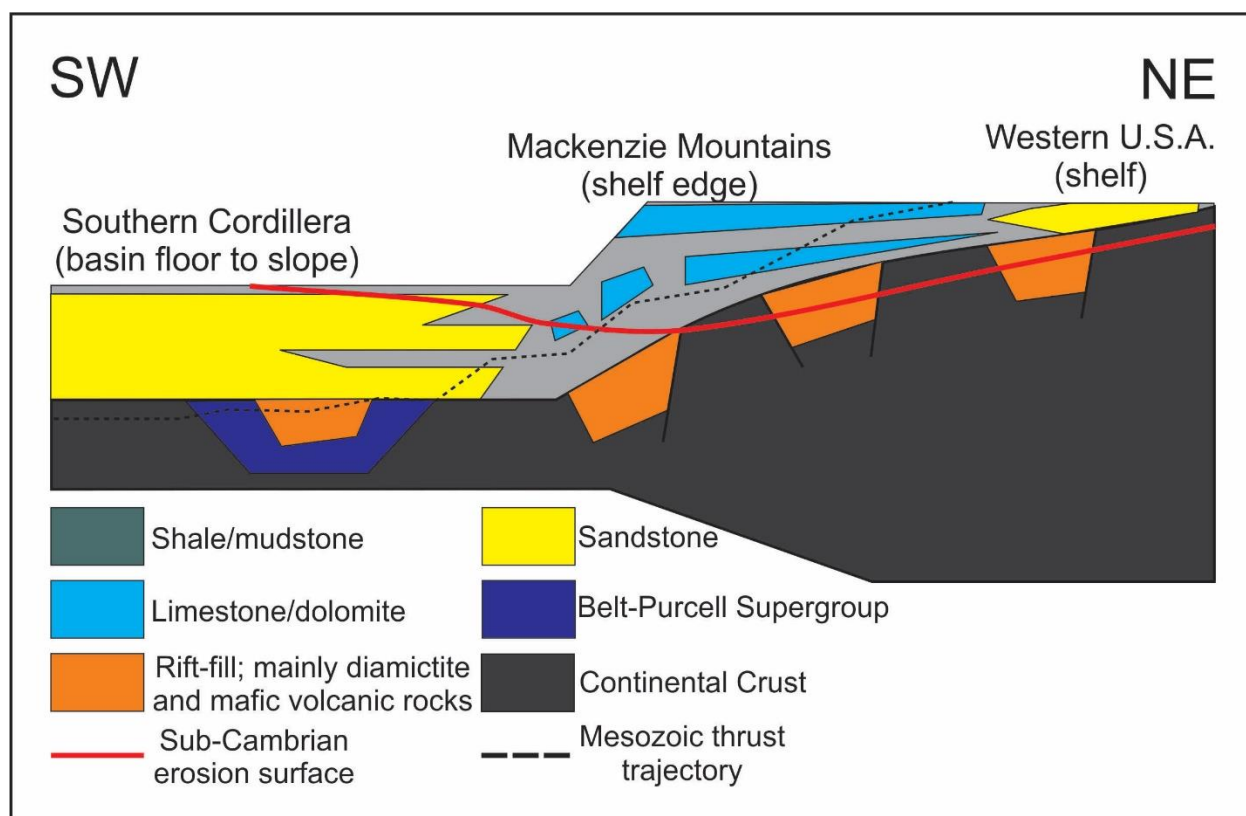


Figure 1.2: Schematic cross-section of western North America ca. ~ 600 Ma. The Windermere in the western United States consists mostly of shelf deposits compared with shelf-edge and upper slope facies in the Mackenzie Mountains, and deep-marine base-of-slope and basin floor settings in the southern Canadian Cordillera. The trace of the sub-Cambrian erosion surface, which has removed shallow-water strata in the southern Cordillera, is indicated by the red line (modified from Ross and Arnott, 2007).

Rocks of the WSG are well exposed in thrust sheets of the Foreland fold and thrust belt and the Omineca belt of the southern Canadian Cordillera; the two belts being separated by the Southern Rocky Mountain Trench (SRMT) (Ross and Arnott, 2007; Smith, 2009). East of the SRMT levels of deformation and metamorphism are low, with metamorphic grade increasing towards the trench, and WSG strata occur in broad structural panels separated by thrust faults. In contrast, metamorphic grade and structural complexity is generally high throughout much of the Omineca belt (up to amphibolite facies), although localized areas of lower grade metamorphism (sub-greenschist and greenschist facies) and tectonic deformation allow for detailed sedimentological study.

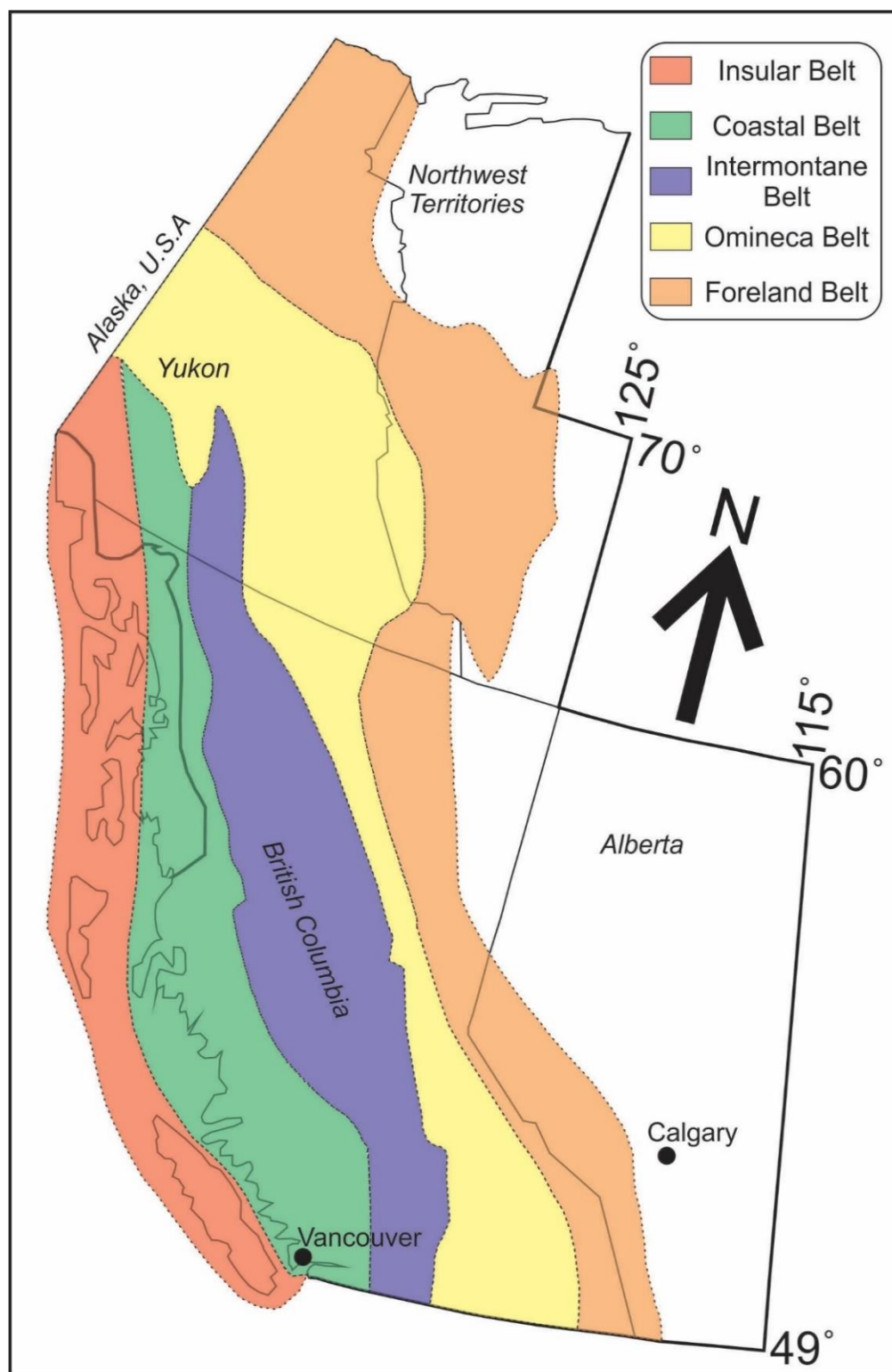


Figure 1.3: Map of western Canada showing the five morpho-tectonic belts that comprise the Canadian Cordillera, and are the result of mountain building events related to collisional tectonism and accretion of allochthonous terranes along western Laurentia between 185-50 Ma. From west to east they are the: Insular Belt, Coastal Belt, Intermontane Belt, Omineca Belt and Foreland Belt (Modified from Ross et al., 2002 and Smith, 2009).

In the McBride map area, south of the SRMT, the Cariboo Mountains have been affected by four phases of deformation (D1-D4) and at least two episodes of low-grade metamorphism (M1 and M2), related to the previously mentioned terrane accretion. (Murphy and Rees, 1983; Murphy, 1987a,b). The first phase (D1) was associated with the obduction of the first terrane onto western North America during the middle Jurassic, producing northeast verging folds and M1, characterized by a low-grade mineral assemblage of muscovite, biotite and chlorite (Murphy, 1987b; Ross & Arnott, 2007). The next stage (D2) produced map-scale southwest verging structures, mainly overturned folds and reverse faults, that form the main structural grain in the region. This phase was associated with significant crustal thickening and M2 onset, which is manifest as metamorphic minerals that overprint D2 structures (Murphy, 1987b). The mid-Jurassic Hobson Lake Pluton, located ~ 75 km south of McBride, cross-cuts D1 and D2 structures indicating that they are older than 174 Ma (Reid et al., 2002; Pigage, 1977; Gerasimoff, 1988). The late- to post-metamorphic D3 cross-cuts and refolds D1 and D2 structures into northwest trending open and upright folds with associated reverse faults (Murphy, 1987b; Reid et al., 2002). The final phase (D4) is post-metamorphic and refolds all previous deformation structures (D1-D3) into metre to decametre northeast-trending, steeply-inclined upright folds with tight inter-limb angles and chevron hinge zones (Campbell, 1970; Campbell et al., 1973; Ferguson, 1994; Reid et al., 1997). Regional metamorphism peaked between D2 and D3, at temperatures estimated between ~ 350-450°C and pressures ranging from 4-9 kbar (Lee, 2016). These stages of deformation and tectonism in the southern Canadian Cordillera have resulted in the deep-marine outcrop belt of the WSG being substantially shortened and now exposed over an area of about 35,000 km² (Ross & Arnott, 2007). If the amount of shortening estimated to be about 30%, then the original size of the Windermere turbidite system was at least 80,000 km², which is similar in area with a number of modern turbidite systems and also makes it one of the world's largest ancient turbidite systems (Figure 1.4; Ross & Arnott, 2007).

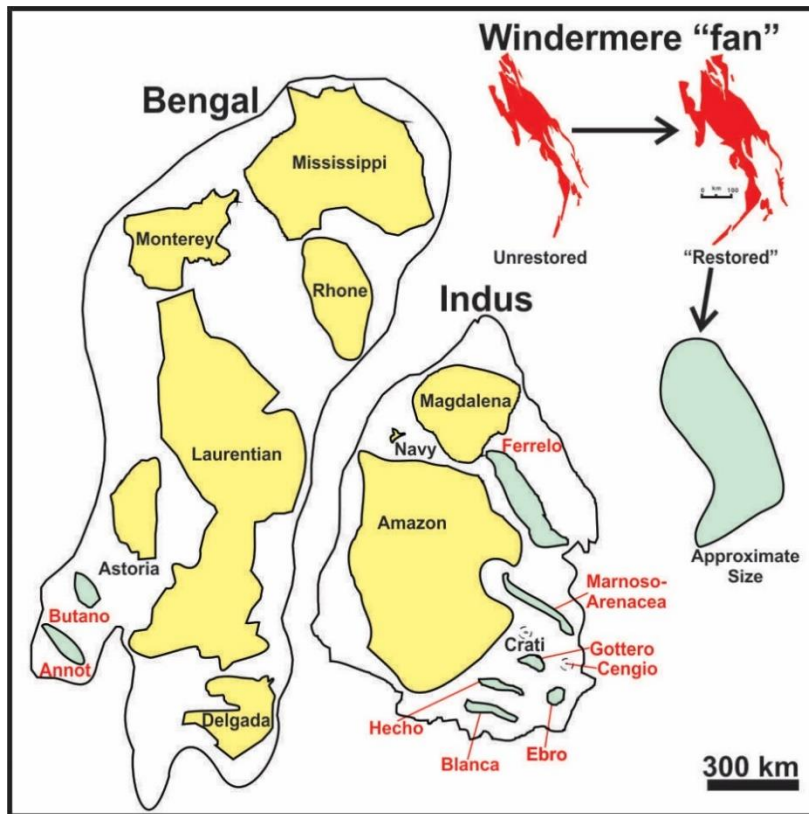


Figure 1.4: Dimensional comparison of the reconstructed Windermere turbidite system and other turbidite basins worldwide, both modern (in black) and ancient (in red) (Modified from Ross and Arnott, 2007).

II.III Development of the Windermere Supergroup

The most significant paleogeographic feature during the Neoproterozoic was the supercontinent Rodinia, which had completely assembled by about 825 Ma, and at that time extended from the northern polar region to the equator, with Laurentia in the centre (Figure 1.5a; Li et al., 2013; Hoffman, 1991; Moores, 1991). At around 780 Ma Rodinia began to break up and this continued throughout much of the remaining Neoproterozoic (Figure 1.5b-d; Li et al., 2013). The onset of supercontinent breakup coincided with the beginning of Windermere Supergroup deposition (Ross, 1991). Throughout this rifting stage Laurentia was situated around the paleoequator with the present-day western margin facing northward, before drifting to the south (30-60°S) near the end of the Neoproterozoic (~ 580 Ma) due to continued opening of the Panthalassa Ocean (paleo-Pacific Ocean) (Li et al., 2013).

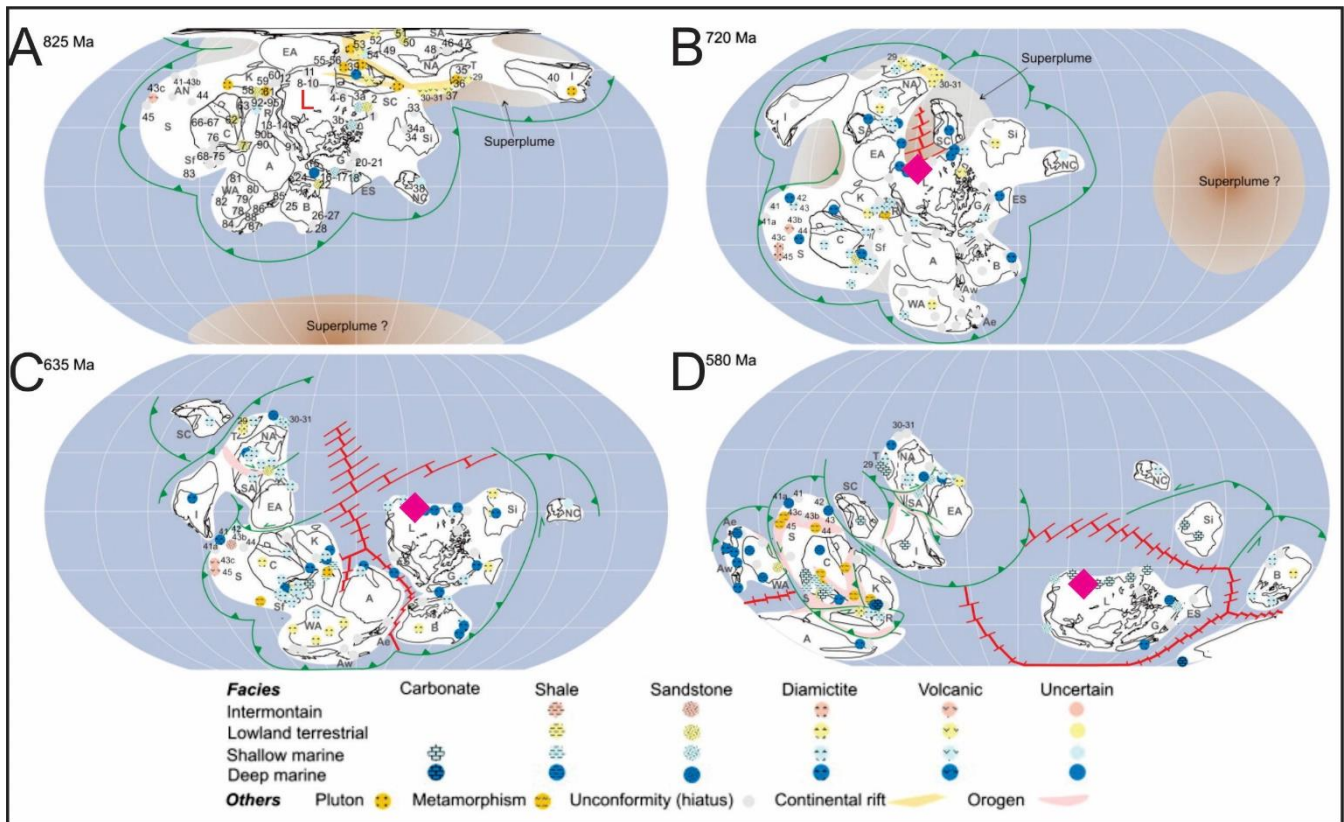


Figure 1.5: Neoproterozoic global reconstructions by Li et al. (2013) for four different time intervals: (A) 825 Ma; during which mid-life Rodinia extended from the paleoequator to the north pole with Laurentia in the middle (indicated by the red L); (B) 720 Ma; onset of Rodinia breakup and global Sturtian glaciation; (C) 635 Ma; continuation of Rodinia rifting and onset of global Marinoan glaciation; (D) 580 Ma; termination of Rodinia rifting and initiation of Gondwana assemblage and Gaskiers glaciation. The approximate location of the Windermere margin is indicated by the pink diamond in B-D.

The WSG is an unconformity bounded sequence that in the southern Canadian Cordillera overlies metasedimentary rocks of the 1.5-1.4 Ga Belt-Purcell Supergroup (southeastern British Columbia and northwestern United States), 2.2-0.73 Ga crystalline basement of the Deserter Gneiss (north-central British Columbia), or the Malton Gneiss (east-central British Columbia) (Evans et al., 2000; Evenchick et al., 1984; Murphy, 1990) and underlies Lower Cambrian strata of the Gog Group (Campbell et al., 1973; Ferguson, 1994). Accumulation of the WSG proceeded in two phases (Stewart, 1972): a lower, syn-rift phase, which in southeastern B.C. consists of laterally discontinuous glacial diamictite (Toby Formation) and volcanics (Irene Formation) deposited in isolated rift basins, and an

upper, post-rift sequence, which in the Cariboo Mountains, comprises laterally continuous sedimentary units of the Kaza and overlying Cariboo groups, which indicate deposition along a passive continental margin developed in response to thermally driven subsidence (Figure 1.6; Ross, 1991; Ross & Arnott, 2007).

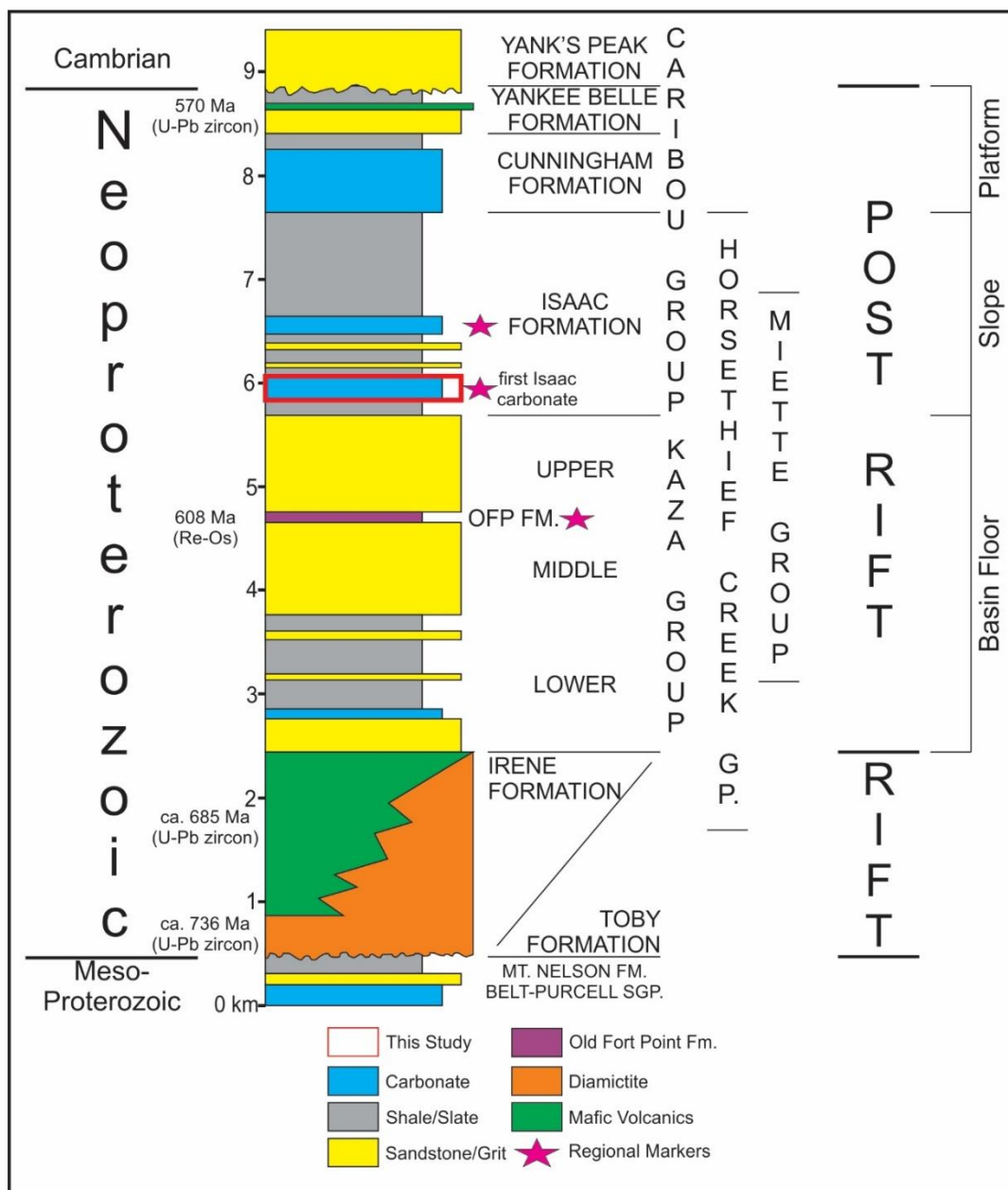


Figure 1.6: Generalized stratigraphic column of the Windermere Supergroup in the southern Canadian Cordillera. The WSG comprises syn-rift (Irene and Toby formations in the Purcell Mountains) and the overlying post-rift succession (Kaza Group and Isaac, Cunningham and Yankee Belle formations in the Cariboo Mountains). This study focuses on the first Isaac carbonate, which is indicated by the red rectangle (modified from Ross and Arnott, 2007).

II.III.I Syn-Rift

The syn-rift phase was marked by deposition in fault controlled isolated rift basins, and the characteristic rapid changes in lithology and thicknesses over short geographical distances (Smith et al., 2011; Ross & Arnott, 2007; Stewart, 1972). The Toby Formation (0-2500 m-thick) was named by Walker (1926) after outcrops along Toby Creek, located west of Invermere, B.C. and is equivalent to the Shedroof and Huckleberry conglomerates of Idaho and Washington (Aalto, 1971; Smith et al., 2011). It consists predominantly of diamictite interbedded with mudstone, sandstone, limestone and conglomerate. Diamictites make up most of the basal part of the unit, but gradually transition upwards to fine, argillaceous sandstone (Smith et al., 2011; Aalto 1971). Based on the occurrence of striated clasts, dropstones and exotic granitic clasts, the Toby is interpreted to be a glaciomarine deposit made up of resedimented glacial deposits associated with the Sturtian glaciation (~ 720-660 Ma (Hoffman et al., 2017; Rooney et al., 2014)). In the Purcell Mountains the Toby Formation is overlain by sedimentary strata of the Horsethief Creek Group, but to the south it is interbedded and conformably overlain by foliated volcanic greenstones and volcanic conglomerates of the Irene Formation (Aalto, 1971). Rocks interpreted to be correlative to the Irene Formation crop out in northern British Columbia and Idaho and have been dated between 726-684 Ma, further suggesting synchronous deposition and igneous activity during the rifting of Rodinia (Devlin and Bond, 1988; Ferri et al., 1999; Lund et al., 2003; Fanning and Link, 2004; Li et al., 2013).

II.III.II Post-Rift

The post-rift succession of the WSG is dominated by sandstone and granule conglomerate turbidites with shale dominated intervals, minor carbonate and one diamictite (Vreeland Formation) and hosts one of the world's best outcrop examples of deep-marine sedimentary rocks (Ross & Arnott, 2007; Ross et al., 1995; Smith, 2009). It represents a 5-7 km succession that shoals upwards from basin floor, to continental slope, to continental shelf in response to long-term basinward progradation of the Laurentian continental margin (Ross & Arnott, 2007). In the Cariboo Mountains (this study's

more details see the Regional Markers (1.II.VI) and Previous Post-Marinoan Windermere Isotope Studies (1.IV.III) sections) (Ross & Murphy, 1988; Smith, 2009; Smith et al., 2014a). In the Monkman Pass area (Figure 1.1) glacial diamictite of the Vreeland Formation (350-2000 m thick) underlies the OFP and changes facies laterally to coarse-grained deep-marine turbidites of the Kaza Group to the west and south (McMechan, 1987; McMechan and Thompson, 1995; Smith et al., 2011). These diamictites are interpreted to result from the Marinoan glaciation (650-635 Ma) and were transported to the deep-marine by the resedimentation of shallow-marine glacial deposits via sediment-gravity flows (McMechan, 2000; Fairchild et al., 2016).

The overlying Cariboo Group is up to 5 km-thick in the Cariboo Mountains and is subdivided into three stratigraphic units. The up to 2.4 km-thick Isaac Formation conformably overlies the Kaza Group and is composed primarily of mudstones (levee deposits) that encase laterally discontinuous, ~ 20-100 m-thick, coarse-grained sandstone units interpreted to represent stacked leveed-channel complexes (Ross and Arnott, 2007; Arnott et al., 2011). A slope setting is interpreted based on the mud-dominated nature of the unit and the commonality of mass movement deposits (slumps, slides and debris flows), suggesting increased gravitational slope instability. Conformably overlying the Isaac Formation are the Cunningham and Yankee Belle formations, which are interpreted as upper-slope to shallow-marine, high-energy shelf deposits (Ross et al., 1995). The Cunningham Formation is up to 550 m thick and is made-up of oolitic intraclastic limestone with minor mudstone, siltstone and shale (Rowe, 2003), while the Yankee Belle is composed of up to 900 m of alternating limestone, siltstone, sandstone and shale (Gabrielse and Campbell, 1991). The top of the Yankee Belle Formation is a regional unconformity that separates the WSG from Lower Cambrian rocks of the Yank's Peak Formation (Aitken, 1969). This is believed to be the result of a second rifting event that caused uplift and as much 4 km of strata to be eroded in the southern Canadian Cordillera. Erosion increases eastward and is responsible for the lack of shallow-marine and continental WSG strata in the eastern part of the basin (Figure 1.1 and 1.2; Ross et al., 1995; Ross & Arnott, 2007).

II.IV Geochronological Constraints

Geochronological data is sparse throughout the WSG in the southern Canadian Cordillera (SCC) due to its mostly siliciclastic composition, lack of volcanic rocks and the absence of fossils. The timing of deposition is constrained by U-Pb zircon dates from rocks that unconformably underlie and overlie the WSG (Figure 1.6). The maximum age of deposition, also coincident with onset of rifting, is constrained from 740 ± 36 to 728 ± 8 Ma from underlying crystalline basement rocks in the Canadian Cordillera (Evenchick et al., 1984; Parrish and Scammell., 1988; McDonough and Parrish, 1991; Li et al., 2013). Timing of the second rifting event at the top of the WSG is well constrained to 569 ± 4.6 Ma from synrift volcanics near the base of the Hamill Group (Colpron et al., 2002). Two dates have been obtained from rocks in the Windermere Supergroup. The first is a ~ 685 Ma U-Pb zircon date from volcanic rocks in Idaho that correlate with rift-related rocks of the Irene Formation (Lund et al., 2003) and the second is a rhenium-osmium (Re-Os) age of 607.8 ± 4.7 Ma from an organic mudstone in the Geikie Siding Member of the OFP (Figure 1.6; Kendall et al., 2004). Based on these constraints the Windermere turbidite system was active for a maximum of ~ 210 myr during which ~ 8000 m of sediment was deposited, ~ 4000 m during the last ~ 40 myr of Windermere history.

II.V Provenance and Paleocurrents

Uranium-lead dating of detrital zircons in the WSG exhibit a pronounced bimodal distribution of ages, > 2.6 Ga (Archean) and 1.9-1.75 Ga (Paleoproterozoic), with a notable absence from 2.6-1.9 Ma (Ross and Bowring, 1990; Ross and Parrish, 1991a, b). This gap in ages constrains the source of sediment to be from the southern Canadian Shield and its continuation into the northwestern United States (Ross and Arnott, 2007). In addition, paleocurrent data, albeit from a rather meagre dataset, is predominantly toward the west-northwest (Ross and Arnott, 2007), which is consistent with an eastern to southeastern sediment source, although variations related to local geomorphological effects have been reported (Figure 1.1; Arnott and Hein, 1986; Schwarz and Arnott, 2007; Khan and Arnott, 2011;

Bergen, 2017). The present-day distribution of WSG strata in the southern Canadian Cordillera shows a progressive northwestward change from submarine canyons to slope to basin floor deposits (Eisbacher, 1985; Ross, 1991), suggesting that it forms a single continental-scale turbidite system fed from the southeast by a single continental (sediment) dispersal system (Terlaky, 2014).

II.VI Regional Markers of the Windermere Supergroup

Throughout the SCC three marker units have been identified: the areally extensive OFP and two less extensive deep-marine carbonate to mixed carbonate-siliciclastic units, informally named the first/lower (FIC, this study) and the second/upper (SIC) Isaac carbonate, respectively (Figure 1.6; Ross and Murphy, 1988; Ross and Arnott, 2007).

The OFP is the stratigraphically lowest of the three and ranges from ~ 50-450 m thick (Ross & Arnott, 2007; Smith et al., 2014a). Glacial deposits of the Marinoan Vreeland Formation locally underlie the OFP, which is made-up of three lithologically distinct members: Temple Lake, Geikie Siding and Whitehorn Mountain members (Smith et al., 2011; Smith et al., 2014a,b). The basal part of the Temple Lake Member is composed of siltstone and mudstone that grades upward into rhythmic limestone-siltstone couplets. Significantly, these strata have been interpreted to be the deep-marine equivalent of the Ravensthorpe cap carbonate in the Mackenzie Mountains, Yukon Territory (Smith, 2009; James et al., 2001). Black/grey organic rich mudstone dominates the middle member with thin-beds of dark-grey limestone and very-fine ripple cross stratified sandstone cropping out locally. The basal contact of the Whitehorn Mountain Member locally scours, and in places completely erodes the lower two members, causing its thickness to vary spatially. Lithologically the Whitehorn Member exhibits a distinctive variety of lithofacies, including mudstone, sandstone, calcareous arenite, arenaceous limestone and diamictite (Smith, 2009). Due to its generally fine-grained nature, which contrasts the sheet-like coarse-grained nature of the surrounding Kaza Group, the OFP is interpreted to represent deposition during a major eustatic sea-level rise related to the melting of the Marinoan ice

sheet that caused the shutdown of coarse siliciclastic sediment input into the Windermere basin (Ross and Arnott, 2007; Smith, 2009; Smith et al., 2011; Smith et al., 2014a).

The FIC and SIC range from ~ 10-260 m-thick, and according to Ross and Ferguson (2003) are lithologically similar. The FIC is characterized by three laterally extensive, decametre-thick units composed of very thin- to thin-bedded carbonate turbidites (hereafter termed calciturbidites) made-up of black to greenish brown siliceous calcilutite and very fine- to fine-grained siliceous calcarenite intercalated locally with up to ~ 20 m-thick granule siliceous calcirudite (conglomerate) to medium-grained siliceous calcarenite. Interbedded with these carbonate units are up to ~ 60 m thick successions of siliciclastic mud dominated, fine-grained, thin-bedded turbidites. As noted above, the SIC is similar in lithology to the FIC, but with local intercalated sheets of quartz arenite (Ross and Ferguson, 2003). Both the FIC and SIC are interpreted to reflect elevated sea-level that flooded the continental shelf and enhanced carbonate sediment production on the platform, which subsequently was resedimented into the deeper-water part of the Windermere basin (Ross et al, 1995; Ross and Arnott, 2007).

II.VII Study Areas

This study focused on the Castle Creek (CC) and Milk River (MR) study areas in the Cariboo Mountains of the SCC (Figure 1.8, 1.1). Geologic work in the McBride region was first conducted by Sutherland Brown (1963) and was subsequently refined in the 1:250,000 geological map and report by Campbell et al. (1973). A more detailed 1:50,000 map, which includes the Castle Creek study area, was later published by Ross and Ferguson (2003), unfortunately Milk River is located just beyond the western edge of the map. Both field areas occur in mid-Jurassic, northwest-trending folds of the Isaac Synclinorium and Premier Anticlinorium associated with the D2 deformation event (Campbell et al., 1973; Murphy, 1987b; Murphy et al., 1995; Reid et al., 2002). Additionally, both are located on the steeply-dipping limbs of two broad anticlines where bedding dips at ~ 89° and ~ 75°, respectively. Rocks have been metamorphosed to lower greenschist facies, but primary sedimentary textures and

structures are still well preserved (Ross & Arnott, 2007). Both areas are situated beside rapidly retreating, north-facing glaciers, resulting in rocks that are glacially polished and free of vegetation. This, then, allows for easy stratigraphic logging and the tracing of beds for hundreds of metres laterally.

II.VII.1 Castle Creek Study Area

At Castle Creek rocks of the Upper Kaza Group and the lower part of the Isaac Formation crop out in an exposure that is 2.5 km-thick (perpendicular to bedding) and 7 km-wide (parallel to bedding) (Figure 1.9a; Meyer, 2004; Ross and Arnott, 2007; Smith, 2009; Angus, 2016). Here the Upper Kaza Group is ~ 800 m-thick and comprises decametre-thick, sheetlike to lobate, coarse-grained sandstone units that grade laterally and intercalate vertically with mud-rich, thin-bedded turbidites (75:25 sandstone:mudstone ratio) (Ross and Arnott, 2007; Terlaky, 2014; Popović, 2016; Angus, 2016). These sandstones are interpreted as depositional lobe elements deposited by unconfined flows on the basin floor, an interpretation supported by the scarcity of mass movement deposits, suggesting negligible gravitational instability (Meyer and Ross, 2007; Terlaky et al, 2015). Near the top of the Upper Kaza Group sandstone filled scours and lesser channels become observed for the first time, marking the superposition of the channel-lobe transition zone above basin floor stratigraphy, and therefore continent-scale progradation of the Windermere system (Rocheleau, 2011; Navarro, 2016; Navarro and Arnott, 2017).

The Isaac Formation is 1600 m-thick at CC and is dominated by fine-grained strata (25:75 sandstone:mudstone ratio) that encase coarse-grained, 10-100 m-thick, few km-wide sandstone units and common mass transport deposits (slumps, slides and debris flows), and a ~ 150-200 m mixed carbonate-siliciclastic unit (see below) (Ross & Arnott, 2007; Davis, 2011; Terlaky, 2014). The sandstone units are typically composed of coarse-grained amalgamated sandstones that fine and thin upward into more interstratified strata and are interpreted to be stacked channel fills bound on their margins by genetically related, fine-grained levees (Navarro, 2006; Arnott, 2007ab; Gammon et al., 2007; Navarro et al., 2007; O'Byrne et al., 2007; Schwarz and Arnott, 2007ab; Khan and Arnott, 2011;

Khan et al., 2011a; Khan, 2012; Dumouchel, 2015; Bergen, 2017). Seven informal channel-levee complexes have been identified in the Isaac Formation of CC.

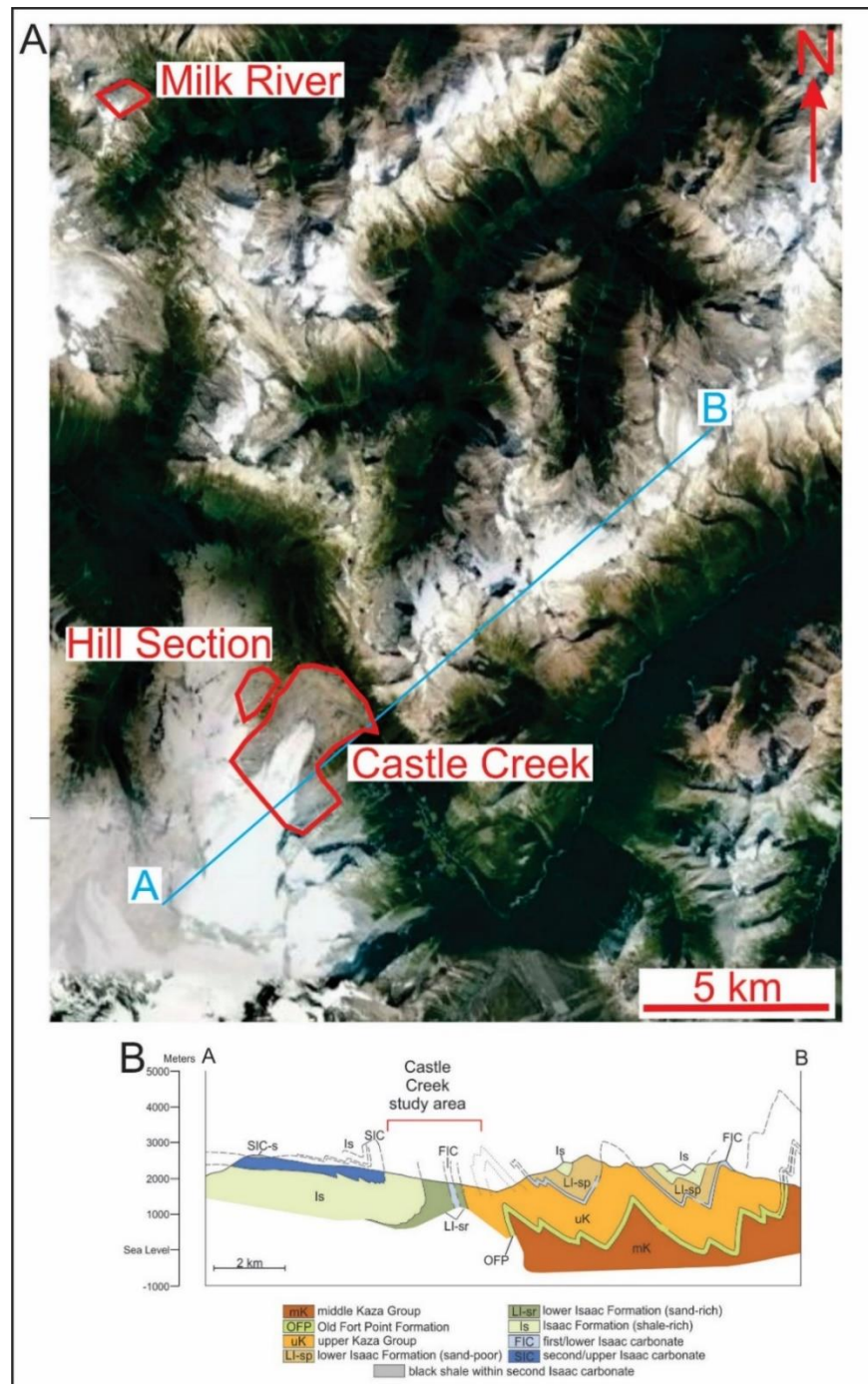


Figure 1.8: (A) Satellite image near McBride, British Columbia with the study areas indicated by red polygons – Castle Creek, Hill Section and Milk River (background image © Province of British Columbia 2017, © CNES/Airbus 2017, © DigitalGlobe 2017, © Google 2017). (B) Cross section of the regional geology of the Castle Creek area. The Castle Creek study area is indicated by the red bracket (excerpt from the Eddy 1:50,000 map area; modified from Ross and Ferguson 2003).

In a recent study on the microstructural deformation at CC, Lee (2016) identified two planar foliation structures that were oriented NNW-SSE, and termed them S1 and S2. S1 is defined by shape-preferred orientation fabrics of shortened quartz grains, muscovite crystals and elongate sericite fields and is associated with D1. S2, on the other hand, comprises undulose muscovite ribbons and micro-crenulations and is attributed to D2. Moreover, lower greenschist metamorphism has caused quartz grains in grain-supported sandstones to display lobate grain boundaries caused by grain boundary bulging recrystallization, and locally even subgrain boundary rotation in more cemented samples (Hirth and Tullis, 1992; Lee, 2016). Samples with higher matrix proportions display lower degrees of micro-deformation as the mud-dominated matrix better accommodates strain and acts as a buffer between the quartz grains “floating within” (Popović, 2016; Lee, 2016). However, metamorphism has resulted in recrystallization of all original clay minerals to muscovite and chlorite, and therefore the mineralogical make up and absolute amount of detrital clay mineral matrix is unknown, but it is estimated to have been at least 5% to more than 70% (Terlaky, 2014; Lee, 2016). Based on the predominance of bulging recrystallization of quartz grains Lee (2016) assumed a deformational temperature range at CC of $\sim 280\text{-}400^{\circ}\text{C}$, which is consistent with greenschist facies mineral assemblages, but slightly lower than the $350\text{-}450^{\circ}\text{C}$ regional temperature. In accordance with regional trends, it was found that there is a slight decrease in strain gradient stratigraphically upward through the Castle Creek section (Lee, 2016).

II.VII.11 Milk River Study Area

Milk River is located ~ 20 km basinward from Castle Creek and is approximately 1000 m thick and 700 m wide (Figure 1.8; 1.9b). Like CC, rocks of the Upper Kaza Group and lower Isaac Formation are superbly exposed and steeply dipping ($\sim 75^{\circ}$). The stratigraphic location of Milk River was confirmed by the presence of the FIC, which is significantly thicker than the equivalent section at Castle Creek (up to ~ 260 m). MR has much less overburden covering the rocks than at CC (i.e. less common moraines), which most probably is related to the steeper inclination of the outcrop. However,

the outcrop quality is diminished due to its smaller area and the slightly higher grade of metamorphism, that generally obscures fine-scale primary sedimentary structures, making detailed sedimentological studies unlikely (e.g. Terlaky and Arnott, 2014; Popović, 2016; Angus, 2016; Wearmouth, 2018).

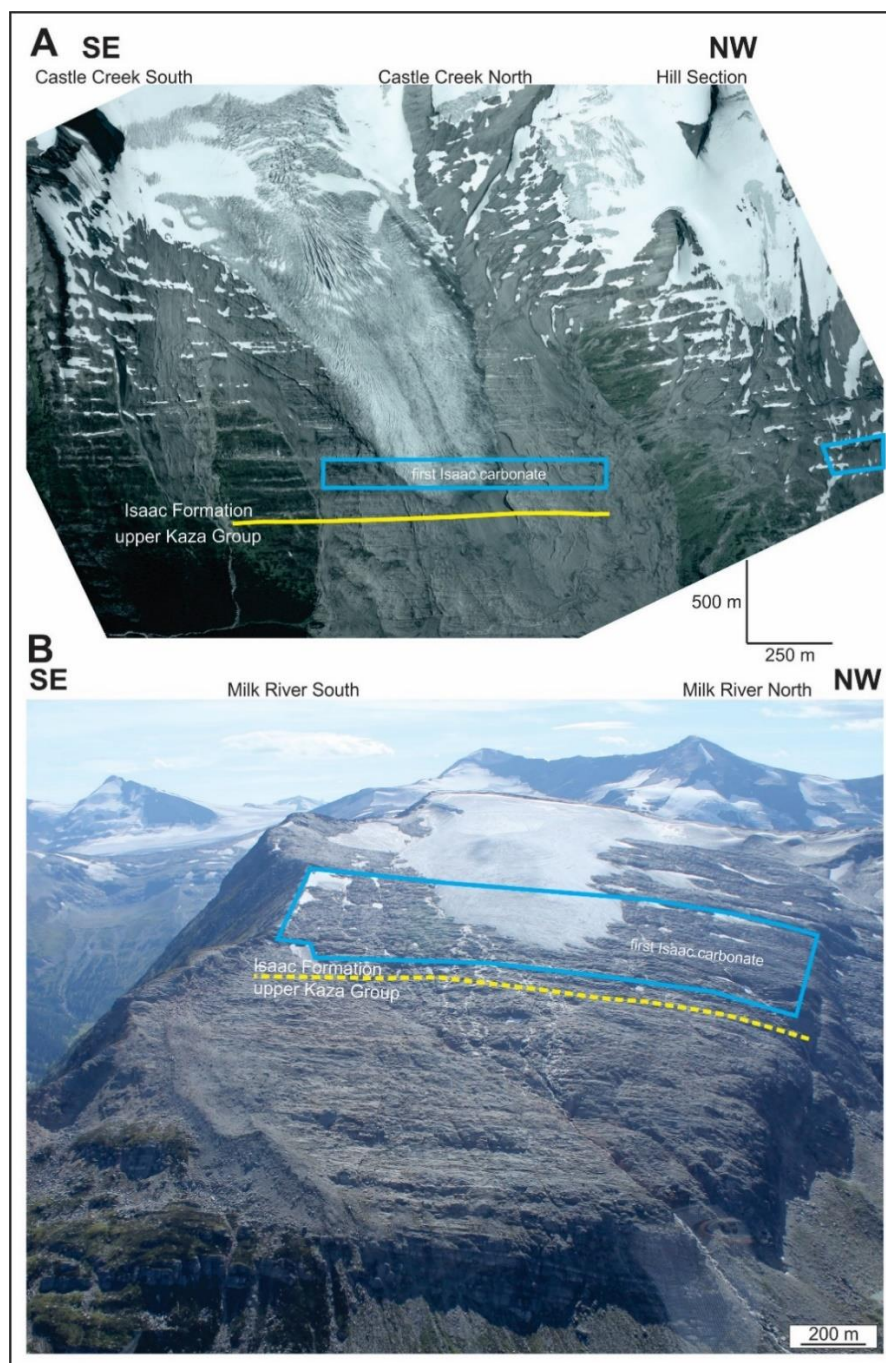


Figure 1.9: (A) Aerial photograph of the Castle Creek and Hill Section study areas and (B) Photograph of the Milk River study area. The contact between proximal basin-floor deposits (Kaza Group) and stacked channel-levee complexes (Isaac Formation) is indicated by the yellow line. Blue box represents outcrop exposure of the first Isaac carbonate.

II.VII.1III **First Isaac Carbonate**

Just above the Kaza-Isaac contact the re-introduction of carbonate sediment into the Windermere turbidite system begins in the form of distinctly red-coloured carbonate-cemented sandstone. Approximately 150 m above this contact, at the base of the FIC, the proportion of carbonate increases dramatically due to the presence of decametre-thick packages of fine-grained, very thin- to medium-bedded calciturbidites. The mixed carbonate-siliciclastic nature of the FIC interrupts an otherwise siliciclastic dominated pile of deep-water sedimentary rocks in the study areas (Fig 1.10a). It is believed that the FIC was deposited during a long-term (3rd order) eustatic rise that led to the initiation of a well-developed carbonate platform and, at times, the shutdown of siliciclastic input into the basin (Figure 1.10b; Navarro, 2016). Carbonates were resedimented from the platform and transported downslope by turbidity currents and debris flows in a process termed highstand shedding, which has been identified in numerous modern and ancient carbonate systems (e.g. Schlager et al., 1994). Superimposed on this long-term trend are shorter-term episodes of sea-level fall that formed decametre-thick channel complexes filled with amalgamated, carbonate-cemented coarse-grained sandstone. Rising sea level was then followed by a dramatic eustatic fall and the deposition of a > 100 m-thick channel complex (Isaac channel 1) that incises the top of the FIC and is devoid of carbonate sediment or cement; the base represents a sequence boundary and attendant shutdown of the carbonate factory. Between the first and second Isaac carbonate units, carbonate reappears sporadically and only as carbonate-cemented sandstone, in particular above Isaac channel 5 and in levee deposits associated with channel 4.5 (Bergen, 2017), or as clasts in debris flow deposits (Terlaky et al., 2016; Bergen and Arnott, 2016; Bergen 2017). Unfortunately, evidence of the shallow-water carbonate platform, or even continental shelf for that matter, have been completely removed by the sub-Cambrian unconformity, but it was likely located southeast of the Cariboo Mountains (Ross, 1991).

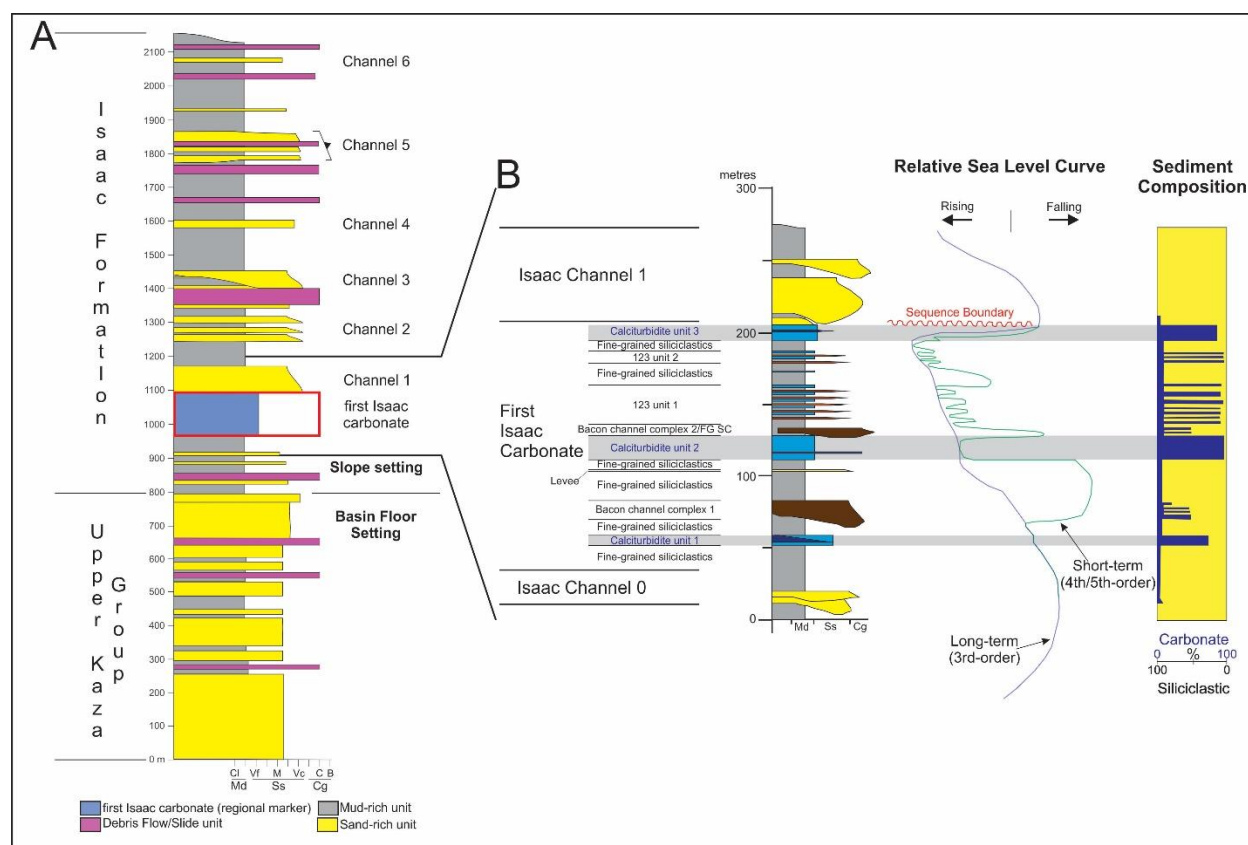


Figure 1.10: Composite stratigraphic section of the Castle Creek study area with the FIC outlined in red (modified from Ross & Arnott, 2007). (B) Generalized log through the FIC showing long- (blue; 3rd order) and short-term (green; 4th/5th order) eustatic changes and mineralogical composition of sediment supply into local Windermere basin. The FIC was initiated during a long-term eustatic rise that was terminated by a major sea level fall marking the top of the FIC and initiation of Isaac channel 1 (modified from Navarro, 2016).

II.VIII Previous Work

Prior to 2001 most studies of the WSG in the Cariboo Mountains focused on regional mapping and structural geology (Sutherland Brown, 1963; Aitken, 1969; Campbell et al., 1973; Murphy and Rees, 1983; Murphy 1987ab; Ross and Murphy, 1988; Gabrielse, 1991; Gabrielse and Campbell, 1991; Hein and McMechan, 1994; Reid et al., 2002; Ross and Ferguson, 2003). The onset of the Windermere Consortium brought new focus into understanding the sedimentology, stratigraphy, architecture and geochemistry of deep-marine strata in the WSG. At Castle Creek base-of-slope deposits of the Isaac Formation have been examined by Wallace (2004), Laurin (2005), Leclair and Arnott (2005), Marion (2005), Navarro (2006), Altosaar (2007), Gammon et al. (2007), Navarro et al. (2007ab), O'Byrne et

al. (2007), Schwarz and Arnott (2007ab), Mussa-Caleca (2008), Buckley et al. (2009), Anthony (2011); Arnott et al. (2011), Davis (2011), Khan et al. (2011a), Khan and Arnott (2011), Dumouchel (2012), Khan (2012), Dumouchel (2015), Angus (2016), Huyer (2016) and Bergen (2017) and geochemistry and stratigraphy of the underlying OFP were studied by Smith (2009); Smith et al. (2011), Smith et al. (2014ab). However, in spite of being long recognized as an important regional marker in the Cariboo Mountains (Ross et al., 1989; Ross et al., 1995; Ross and Ferguson, 2003; Ross and Arnott, 2007), the FIC has been paid little sedimentological attention. G.M. Ross previously measured a stratigraphic log and sampled across the upper portion of the FIC for bulk carbon isotope analysis, data that has been incorporated into this study. Gammon and Arnott (2007) analyzed the petrography and geochemistry of a calcidebrite in the FIC, but only recently was a complete and detailed stratigraphic log measured through the FIC (Navarro, 2016). Additionally, this study is the first to report on rocks at the Milk River study area.

Part III Deep-Marine Sedimentation Processes

The dominant agents for transporting and depositing sediment from the continental shelf to the deep marine are sediment-gravity flows, which of sediment-water mixtures driven downslope by gravity (Arnott, 2010). Sediment-gravity flows commonly initiate beyond the continental shelf-slope break when the downward force of gravity exceeds the shear strength of an accumulation of sediment (Shanmugam, 2006), but they can also result from dense, bottom-hugging suspensions related to hyperpycnal river discharge or local storm events (Piper and Normark, 2009). In these flows sediment is mixed and suspended in the water column, causing the mixture to be denser than the ambient fluid, resulting in downslope movement of sediment as the it displaces the ambient fluid (Terlaky, 2014). Sediment-gravity flows can travel at speeds up to 100 km/h and during a single event can transport more than 100 km³ of sediment hundreds to 1000s of kilometres across the sea-floor (Rothwell et al, 1998; Piper et al., 1999; Talling et al., 2007; Wynn et al., 2002; Frenz et al., 2008). Despite these being

the largest transporters of sediment in the world, they were overlooked for much of early sedimentology (Arnott, 2010). It was only after the increase in sea exploration, principally associated with submarine warfare during World War II that researchers began to focus on deep-marine turbidite systems and their potential for hosting large hydrocarbon reserves (Heezen and Ewing, 1952; Kuenen and Migliorini, 1950; Stow and Mayall, 2000; Arnott, 2010).

III.I Classification of Sediment-Gravity Flows

In-situ documentation of modern sediment-gravity flows is rare because of their destructive and infrequent nature; thus, their classification is based largely on field studies and laboratory experiments. Mulder and Alexander (2001) constructed a scheme based on flow behaviour, grain-size distribution, and grain-support mechanisms in which they proposed two end member types of flow: frictional and cohesive, although a variety of intermediate flow types exist.

III.I.I Frictional Flows

The most common type of frictional flows are turbidity currents, which according to Mulder and Alexander (2001) are sediment suspensions that contain less than 9% sediment concentration by volume and particles are suspended primarily by the upward directed component of fluid turbulence and secondarily by grain-to-grain interactions and buoyancy (Bagnold, 1962; Middleton and Hampton, 1973; Mulder and Alexander, 2001; Arnott, 2010). An idealized turbidity current has a well-developed structure and can be subdivided horizontally into three parts: head, body and tail (Kneller and Buckee, 2000). The head has the highest sediment concentration and is where mixing with the ambient fluid is intense and sweeps sediment-rich fluid toward the body. In laboratory experiments the head moves slower than the body of the flow because it must displace the ambient fluid. These currents typically hug the ocean floor and their low sediment concentration allows for deposition through progressive differential settling of grains that produces a characteristic upward decrease in grain size (Kuenen and Migliorini, 1950; Kuenen, 1964; Bouma, 1962; Terlaky, 2014). Additionally, this fining trend is

accompanied by a distinctive upward change in sedimentary structures that collectively forms the well-known five-part (A-E) Bouma sequence (Figure 1.11; Bouma, 1962). The A division commonly has an erosive base and is composed of massive to coarse-tail graded structureless sandstone and/or conglomerate. This is then sharply overlain by planar-laminated sandstone (B), overlain by ripple cross-stratified sandstone (C), capped by discontinuous interlaminated silt-mud couplets (D) and finally structureless mudstone (E) (Bouma, 1962; Stow, 1979; Stow and Shanmugam, 1980; Kelts and Arthur, 1981; Arnott, 2010). The T_{a-b} divisions are commonly interpreted to form in the upper flow regime while the T_c deposited in the lower flow regime (Walker, 1965). However, the planar laminated T_b division does not necessarily imply upper plane conditions, but more simply spatially uniform bed load transport on a bed that lacked surface defects necessary to form hydrodynamic instabilities, and thus angular bedforms (i.e. ripples) (Arnott, 2012).

III.II.II Cohesive Flows

The most common cohesive flows are debris flows, where the volume concentration of the solid and fluid phases are of the same order of magnitude allowing them to possess a substantial mechanical strength (Mohrig et al., 1998; Mulder and Alexander, 2001). In these flows heavy grains, generally equating to large grain size, are suspended by a cohesive matrix composed of fluid and fine-grained sediment, more specifically clay-sized clay minerals, aided by local high pore pressure, particle-particle interactions and buoyancy (Mulder and Alexander, 2001; Arnott, 2010). This cohesive, fine-grained matrix diminishes the settling velocity of particles in the flow, reducing the rate of deposition. When the shear resistance of the flow exceeds the driving force of gravity it freezes inward and deposits en masse. Debris flows can run-out for distances > 100 km and their deposits, or debrites, typically consist of a disorganized, poorly sorted mixture of variously sized clasts dispersed throughout a fine-grained matrix (Mulder and Alexander, 2001). Their disorganized, poorly sorted nature allows debrites to have a distinct chaotic character in seismic, making identification easy

(Arnott, 2010). As debris flows are transported downslope they can form an overriding into frictional flow from sediment being eroded from the flow and incorporated into a turbulent suspension above.

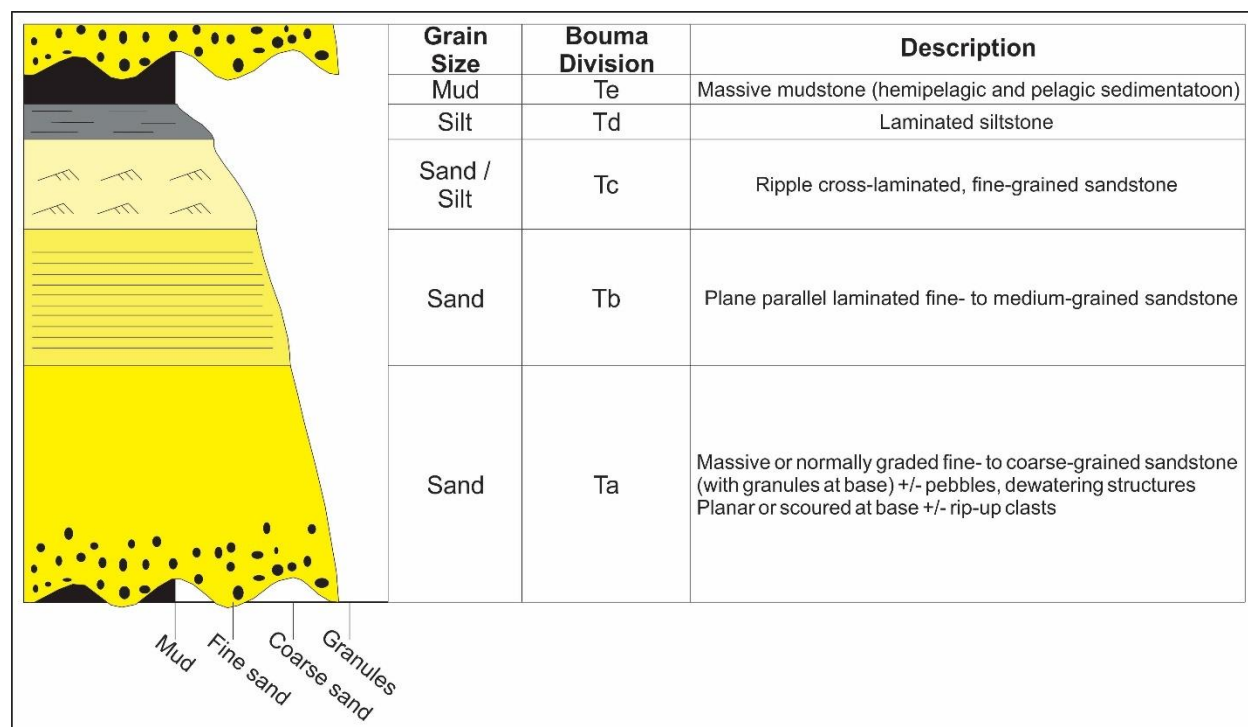


Figure 1.11: Idealized Bouma sequence deposited by a decelerating, dilute turbidity current; showing upward change in grain-size and sedimentary structures (redrawn after Bouma, 1962; Terlaky, 2014; Angus, 2016).

III.II Mass Movement Deposits

In contrast to turbidity currents and debris flows, mass movements (slides and slumps) are gravity induced downslope movements of coherent masses of sediment along discrete failure planes (Arnott, 2010; Angus, 2016). They occur when the tensile strength of the sedimentary pile is exceeded by the driving force of gravity and often are confined to proximal parts of the sedimentary system (i.e. submarine canyon and continental slope) (Shanmugam, 2006; Arnott, 2010). As slides and slumps move downslope internal reorganization may cause the structural integrity of the mass to be lost, resulting in the transition to a debris flow, and if sufficiently diluted, eventually transformed into a turbidity current (Figure 1.12 Shanmugam, 2006). Deposition occurs en masse when the gravitational driving force is exceeded by friction along the base of the moving mass.

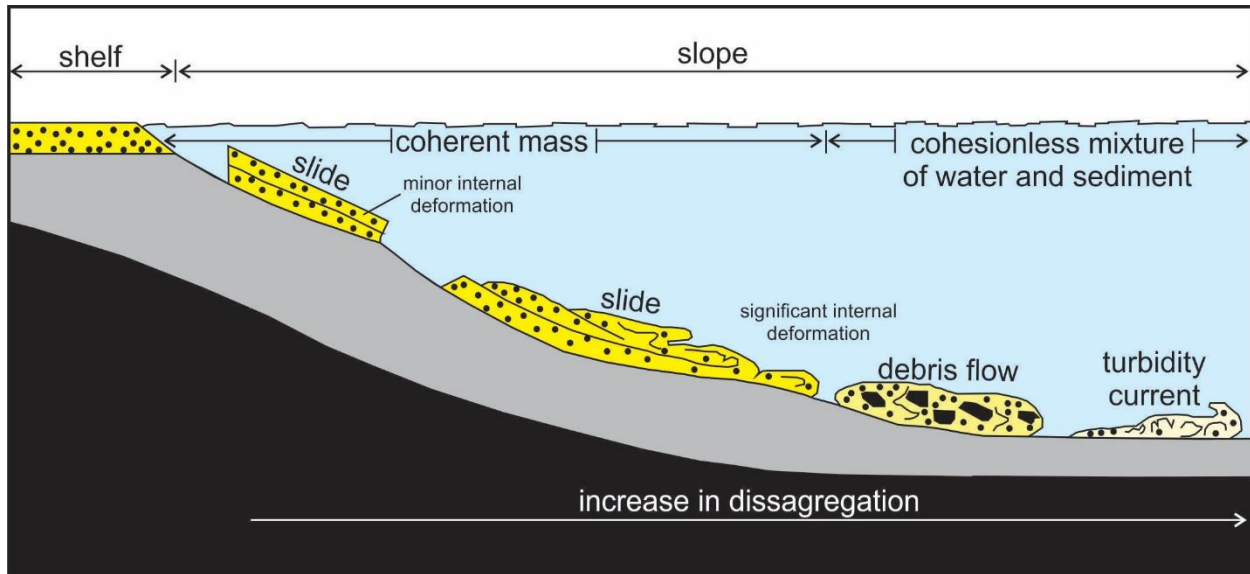


Figure 1.12: The downslope transition of sediment gravity flows. At the shelf edge initial sediment failure forms coherent slumps and slides that transform into debris flows and finally turbidity currents (modified from Shanmugam et al., 1994; Terlaky, 2014).

III.III Deep-Marine Channels and Levees

Deep-marine channels are negative topographic elements that are the primary pathway for transporting sediments from the slope to the basin floor (Arnott, 2010). Slope channels are morphologically similar to their subaerial counterparts as they commonly exhibit sinuous, meandering planforms and are bounded on their margins by levees formed by flow overspill (Figure 1.13a; Posamentier & Kolla, 2003; Babonneau et al., 2002; Khan, 2012). Channels are defined by their degree of confinement, which is a function of the interplay between erosion, deposition and levee development. Generally, channel width and depth decrease downflow in deepwater systems as there is a basinward transition from deeply incised erosional canyons/channels on the upper slope to aggradational levee-bounded channels on the middle and lower slope (Normark et al., 1985; Babonneau et al., 2002; Arnott, 2010; Terlaky, 2014). Poorly- and highly- confined leveed channels are the two end-member types of channel-levee deposits in the rock record (Arnott, 2010). Poorly confined channels range from several meters to a few 10s of m thick and commonly fine and thin upward and laterally. The basal contact is commonly asymmetric with abundant mudstone intraclasts,

and like meandering rivers is more steeply inclined along the outer bend side (“cut bank”) of the channel. Along this margin levee deposits are in erosional contact with the channel, whereas on the opposite side (“point bar”) coarse-grained, amalgamated sandstone channel-fill strata progressively fine, thin and become increasingly interstratified with mudstone as channel deposits grade continuously into the adjacent inner bend levee. In contrast, highly confined channels are generally smaller and occur as disconnected channel fills that tend to cluster laterally and vertically (Arnott, 2010). Channel fills display negligible scouring along their basal contact and change little in grain size upward or laterally. Additionally, coarse-grained, amalgamated sandstones (channel fill) interfinger obliquely upward with thin-bedded, fine-grained turbidites. Relative to the basal contact these strata are inclined at about 7-12° and are interpreted to represent lateral accretion deposits formed on the inner bend of a sinuous deep-marine channel.

Levees form as flows overspill the margins of the channel. In contrast to subaerial conditions, the small density difference between the flow and ambient fluid in submarine systems provides buoyancy and the ability of the overspill flow to extend far from the related channel (up a few tens of kilometres) (Posamentier and Kolla, 2003; Wynn et al, 2007; Khan et al., 2011b; Arnott, 2010). Upon escaping the channel, the flow expands rapidly and collapses, leading to high rates of sedimentation adjacent to the channel, which then decreases laterally, giving levee deposits a their distinctive “gull-wing” shape in seismic images (Figure 1.13b; Arnott, 2010; Khan et al., 2011b). This causes levee deposits to typically fine and thin away from the channel, although this trend is not always the case (Khan, 2012; Bergen, 2017). During the early stages of channel development, the relief between the levee-top and channel base is less, allowing the sand-rich basal part of the channelized flow to overspill, depositing sand-rich strata along the margins (de Leeuw et al., 2016). As the levees aggrade the relief eventually allows only the upper, more dilute portions of the flow to overspill, which is manifest as the typical upward fining and thinning trend in levee deposits (Bergen, 2017). In outcrop

this fine-grained portion is characterized by upper division, T_{cde} , turbidites with multiple ripple sets, indicative of competence-driven deposition in sustained, low-energy flows (Arnott, 2010).

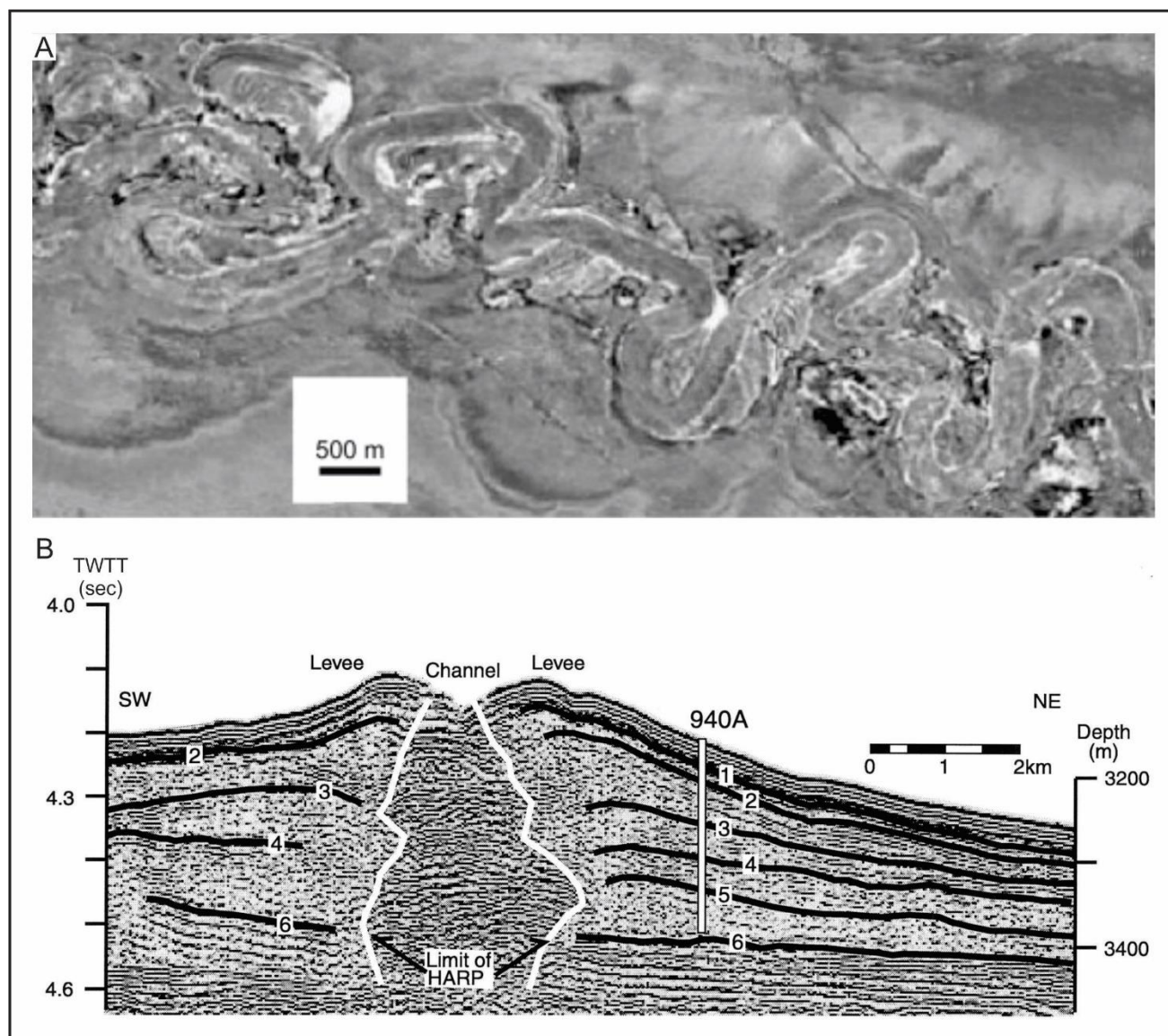


Figure 1.13: (A) 3D seismic horizon slice of the moderate to highly sinuous Green Channel Complex, offshore Angola (from Abreu et al., 2003). (B) Seismic profile across a channel-levee complex in the Amazon fan. White lines delineate the margin of the channel. Note how levee deposits on either side of the channel display the characteristic “gull-wing” morphology away from the channel (from Piper and Normark, 2001).

Part IV Neoproterozoic Climate and $\delta^{13}\text{C}$ Chemostratigraphy

Interest in the Neoproterozoic Era has steadily grown over the past few decades as it represents a period of drastic change in the Earth system that is unparalleled in magnitude to anytime during the Phanerozoic. For example, the Neoproterozoic saw the assembly and breakup of a supercontinent (Rodinia), three worldwide glaciations that followed an ~ 1.5 billion-year-long interglacial period, the return of banded-iron formations (BIFs) after a 1 billion-year hiatus, globally-correlatable carbonates formed during early interglacials in response to rapid climate change, and a transition to a more oxic atmosphere that is believed to have stimulated the evolution of metazoan life (Ediacaran biota), which eventually led to the development of calcified skeletons during the Cambrian explosion (Kirshvink, 1992; Hoffman et al., 1998; Hoffman and Schrag, 2002; Xiao and Kaufman, 2006; Halverson, 2006; Fike et al., 2006; Fairchild and Kennedy, 2007) .

The stratigraphic record has been crucial in reconstructing the Neoproterozoic Era, but unlike the Phanerozoic, regional and global correlation have been hampered by the lack of biostratigraphic control (Halverson, 2006; Fairchild and Kennedy, 2007; Halverson et al., 2010). Accordingly, geologists rely on chemostratigraphy, the study of variations in sediments' chemical makeup, for meaningful correlation (Halverson, 2010). Ideally these data would be supplemented with absolute radiometric ages. However, in many cases appropriate rocks are not present or cannot be accurately dated with current dating techniques, although methods have much improved in recent years (e.g. Rooney et al., 2015). The most widely used geochemical tools for correlation are $\delta^{13}\text{C}$, $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{34}\text{S}$. The $\delta^{13}\text{C}_{\text{carb}}$ (carbon isotopic composition of unaltered carbonates) has shown that Neoproterozoic glaciations are associated with negative anomalies that are reproducible at both basinal and inter-basinal scales (Knoll and Walker, 1992; Halverson, 2006; Halverson et al., 2010). Similarly, the marine strontium isotope ($^{87}\text{Sr}/^{86}\text{Sr}$) record shows a steady increase from < 0.7055 to > 0.7080 during the

Neoproterozoic, with abrupt (~ 0.005) rises during post-glacial episodes and just prior to the dawn of the Cambrian (Shields et al., 1997; Halverson et al., 2010).

IV.I Neoproterozoic Glaciations

Despite earlier periods when much of Rodinia was situated at high latitudes (e.g. 825 Ma), evidence for glaciation in the Neoproterozoic sedimentary record is lacking before about 730 Ma, indicating a stable greenhouse world (Li et al., 2013). Following this, synchronous diamictites have been found on almost every Neoproterozoic continental block preserved today, suggestive of periods of widespread glaciation. This observation, in addition to paleomagnetic data, led Harland (1964) to propose the idea of a worldwide Neoproterozoic glaciation. With better understanding of tectonics and improved paleomagnetic and radiometric dating techniques it has been increasingly shown that there were three widespread glaciations, generally termed the Sturtian, Marinoan and Gaskiers, that are marked by glacial deposits that extend from high latitudes to near the equator (Figure 1.14; Halverson et al., 2005; Li et al., 2013). During the Sturtian and Marinoan glacials much of the Earth's continental landmasses lay close to the equator, which contrasts the configuration during the much more recent Quaternary glaciations (Li et al., 2013). Factors that caused the planet to switch from a greenhouse to icehouse world include: (1) the presence of a stable supercontinent leading to reduced ocean spreading, and, therefore, lowered sea level due to older underlying oceanic crust (Russell, 1968; Valentine and Moores, 1970). Kirschvink (1992) proposed that this fall in sea-level would expose large expanses of reflective continental landmasses, and in turn cool the climate. (2) Enhanced silicate weathering due to the breakup of Rodinia, in addition to increased precipitation rates on the extensive low-latitude continents and (3) extrusion of easily weathered flood basalts in the tropics (i.e. Franklin Large Igneous Province) coincided with the onset of Sturtian glaciation (Worsley et al., 1991; Godd  ris et al., 2003; Donnadieu et al., 2004; Denyszyn et al., 2009; MacDonald et al., 2010). Silicate weathering is a sink

for CO₂, which then reduces atmospheric pCO₂ and, as a consequence, global temperature (Marshall et al., 1988).

Sturtian glacial deposits are found on every continent with a preserved Neoproterozoic stratigraphic record, extending throughout the entire paleolatitude range from 60°N-60°S, with the best exposures in northern Namibia, South Australia and in the Mackenzie Mountains of northwestern Canada (Figure 1.15A; Halverson et al., 2005; Li et al., 2013). The timing and duration of the Sturtian glaciation has been a topic of much debate, with some studies suggesting it was short lived (< 5 million years) and started ~ 740 Ma (Frimmel et al., 1996; Key et al., 2001) while others support a long glaciation that persisted to about 670 Ma (Halverson, 2006). Recent Re-Os dating by Rooney et al. (2015) from strata immediately above and below Sturtian diamictites indicate the glacial epoch began after ~ 727 Ma and ended about 660 Ma (~ 57 myr glacial).

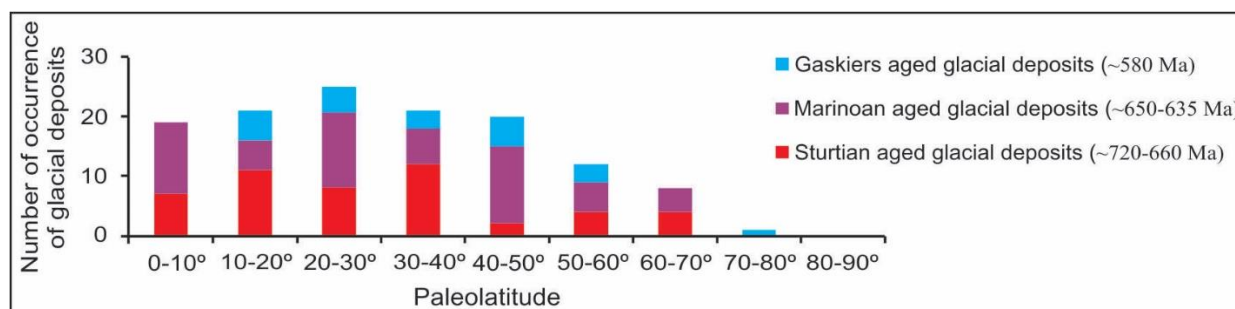


Figure 1.14: Paleolatitudinal distribution of diamictites associated with the three major Neoproterozoic glaciations (modified after Li et al., 2013).

Deposits associated with the Marinoan were deposited after a brief interglacial (~ 10 myr) and are the most geographically widespread of the Neoproterozoic glaciations, with diamictites deposited at all paleolatitudes from > 70° to the equator (Figure 1.15B; Halverson et al., 2005; Li et al., 2013). Radiometric data suggests that the Marinoan glacial lasted for about 5-15 myr and terminated at about 635 Ma (Hoffman et al., 2004; Condon et al., 2005; Calver et al., 2013; Rooney et al., 2015).

Almost all Sturtian and Marinoan glacial deposits are sharply overlain by metre- to decametre-thick dolostones, with local limestones, that persist globally; even in successions that are generally

devoid of carbonates (Kennedy, 1996; Hoffman et al., 1998; James et al., 2001; Hoffman and Schragg, 2002; Fairchild and Kennedy, 2007; Hoffman et al., 2007). The basal contact of these units show little evidence of hiatus and together with their regional to global extent, indicate the rapid and synchronous termination of a worldwide glaciation. During such glaciations, sinks for carbon dioxide, like silicate weathering, would be greatly reduced, thereby allowing it to accumulate in the atmosphere through volcanic outgassing (Hoffman and Schragg, 2002). Once CO₂ levels exceeded a critical threshold, swift deglaciation would ensue, leading to the transfer of atmospheric CO₂ to the ocean and precipitation of carbonate rocks in the warm surface waters found throughout the world (Hoffman et al., 1998). The appearance of these ubiquitous carbonates that were deposited over a relatively short period of geologic time (~ 1 myr) give further evidence that the first two Neoproterozoic glaciations were global in scale and therefore unlike those in the later Phanerozoic.

The Ediacaran Gaskiers glaciation is the most controversial of the Neoproterozoic glacial events due to its diminished distribution relative to the Sturtian and Marinoan (Figure 1. 15C). Deposits associated with this glaciation are reported from several continents covering a wide range of paleolatitudes, but are much less widespread than their predecessors and are not globally overlain by a cap carbonate (Halverson, 2006; Li et al., 2003). In addition, Newfoundland's Gaskiers Formation is dated at 580 Ma and is thought to have been deposited over less than 1 myr (Bowring et al., 2003). The above suggests this glaciation does not fulfil all the criteria for a global glaciation as outlined by Hoffman et al. (1998), but it could potentially represent a bridge between Neoproterozoic- and Phanerozoic-type ice ages (Halverson et al., 2005; Halverson, 2006; Li et al., 2013). The Gaskiers glaciation has long been thought to be associated with the Ediacaran Shuram/Wonoka anomaly, a steep and rapid decrease in $\delta^{13}\text{C}_{\text{carb}}$ to values as low as -12‰ (see section IV.II and IV.III below for more on carbon isotopes and trends throughout the Neoproterozoic), but its timing and association with glaciation remain poorly constrained (Calver, 2000; Corsetti and Kaufman, 2003; Halverson et al., 2005; Le Guerroué et al., 2006a,b). However, more recent work by MacDonald et al. (2013), building

on studies from the Ediacaran record of South China that showed two excursions (Zhou and Xiao, 2007; McFadden et al., 2008), suggests that a small magnitude negative excursion at 580 Ma coincides with the Gaskiers, which precedes the large ~ 560-550 Ma Shuram/Wonoka excursion.

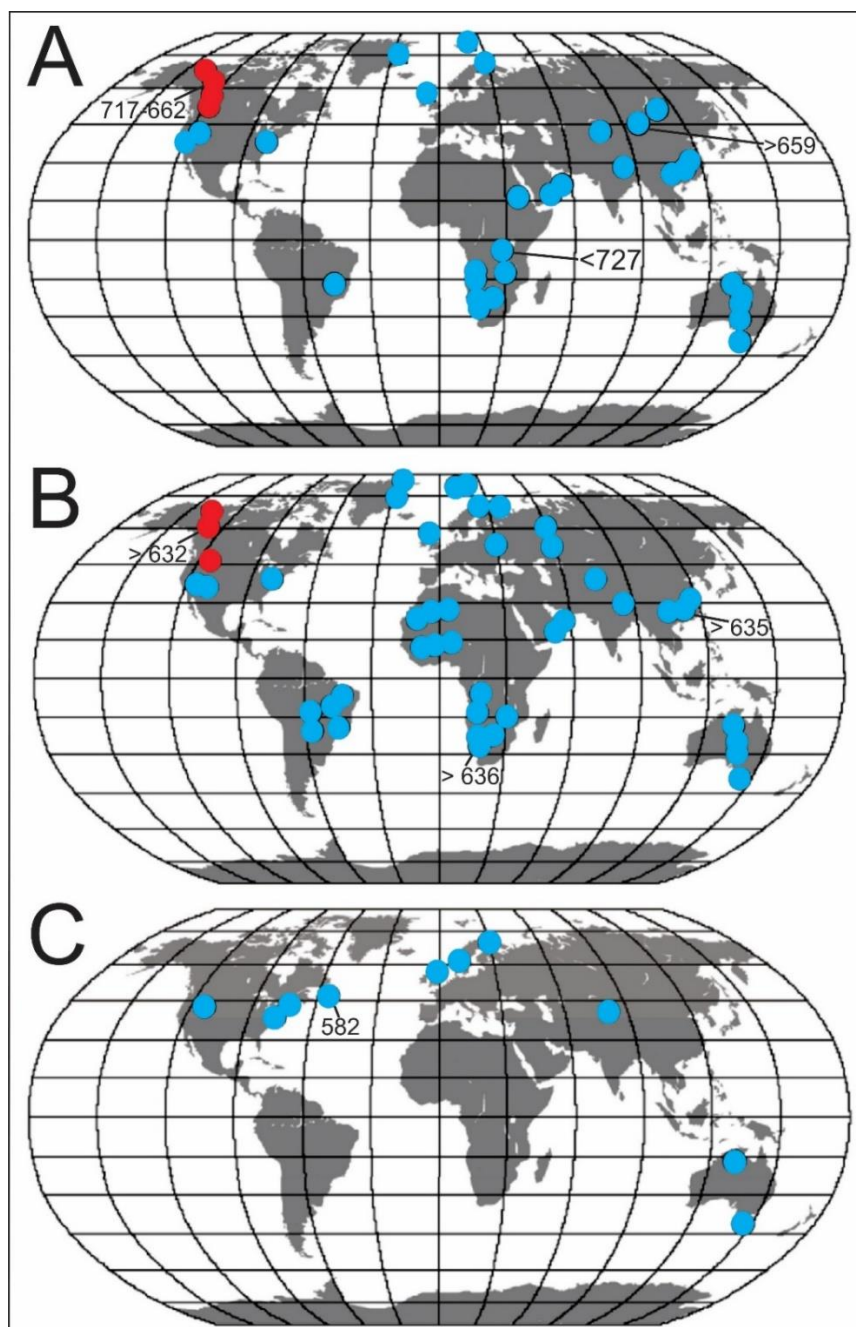


Figure 1.15 Present day geographical distribution of deposits associated with the (A) Sturtian, (B) Marinoan and (C) Gaskiers glaciations. The red circles indicate glacial deposits in the Windermere Supergroup of western Canada. Numbers beside some of the circles indicate some published age constraints on the timing of glaciation (in Ma) from Bowring et al. (2003), Hoffman et al. (2004), Condon et al. (2005), and Rooney et al. (2014; 2015) (modified from Halverson, 2006).

IV.I.I Proposed Models for Neoproterozoic Glaciations

Since Harland's (1964) pioneering theory the mechanisms that led to the extensive glacial deposits associated with the Sturtian, Marinoan and Gaskiers glaciations have been vigorously debated, and most notably include the: High-Tilt Earth, Zipper-Rift Earth, Snowball Earth and Slushball Earth models (Fairchild and Kennedy, 2007). Only the last two models support the premise that glaciations were long-lived, worldwide phenomena.

Williams (1975, 1993, 2000) first proposed the High-Tilt Earth model to explain the low-latitude distribution of Neoproterozoic glacial deposits. In this model an increase in the Earth's obliquity (tilt of its spin axis with respect to its orbital plane) to more than 54° would cause polar regions to receive more solar radiation than the equator. This, and the predominantly low-latitude continental distribution at the time, would have led to equatorial glaciations. The principal objection to this model is that it is difficult to so markedly change the Earth's obliquity over such a short time period, and once the planet achieved low obliquity the gravitational coupling of the Earth and Moon system would then prevent it from increasing again (Laskar et al., 1993; Hoffman and Schrag, 2002; Fairchild and Kennedy, 2007). Also, the change in the distribution of solar energy would cause the areas of primary carbonate production to move from the tropics to the poles, however this is not supported by the Neoproterozoic geological record (Kirschvink, 1992; Hoffman and Schrage, 2002).

The Zipper-Rift Earth model was proposed by Eyles and Januszcak (2004) because they believed that Neoproterozoic diamictites were preferentially preserved rift basins, due to the occurrence of glaciers on uplifted rift-shoulders, and were not the result of global glaciations. They were also skeptical of low-latitude glaciation and the deposition of cap carbonates in response to rapid deglaciation and suggested that the lithological similarity between Neoproterozoic and Phanerozoic diamictites was evidence that similar forcing mechanisms created them. However, more recent work has shown overwhelming evidence of low-latitude glaciation during the Neoproterozoic ice ages and deposition of cap carbonates in response to rapid deglaciation (Hoffman and Schrag, 2002; Li et al.,

2013). This and the lack of evidence for the association between rifting and glaciation put the Zipper-Rift theory in doubt.

Widespread evidence for low-latitude Neoproterozoic glacial deposits led Kirschvink (1992) to hypothesize that they were the result of worldwide glaciations, during which “the Earth would have resembled a highly reflective snowball”. This author hypothesized that the equatorial continental distribution would make the planet more susceptible to glaciation, which began at the poles. These ice sheets would increase Earth’s albedo, enhancing the cooling of the climate and, in turn, promote the growth of ice. Once the ice progressed to a critical latitude ($\sim 30^\circ$) runaway glaciation would ensue (Figure 1.16; Budyko, 1969; Caldeira and Kasting, 1992; Hoffman and Schrag, 2002). This icehouse state would shutdown the carbon cycle, allowing for the buildup of CO_2 in the atmosphere, and once levels exceeded a critical threshold (~ 0.12 kbar) global temperature would swiftly rise, resulting in rapid deglaciation (Kirschvink, 1992; Caldeira and Kasting, 1992; Hoffman et al., 1998; Fairchild and Kennedy, 2007). Hoffman et al. (1998) estimated that it would take between 4-30 myr to raise atmospheric CO_2 to sufficient levels, indicating that snowball glaciations tended to be long-lived. Kirschvink (1992) stated that the snowball hypothesis could be further strengthened if the synchronicity of geographically dispersed glacial units and evidence for rapid climatic fluctuations could be demonstrated. Later work has demonstrated that Neoproterozoic glacial units are bounded by globally correlatable high-magnitude $\delta^{13}\text{C}_{\text{carb}}$ excursions (discussed in more detail in section IV.II.I) thereby demonstrating their synchronicity (Halverson et al., 2010). In addition, the ubiquitous conformable occurrence of superimposed cap carbonates is evidence of rapid climatic shifts from icehouse to greenhouse conditions (Hoffman and Schrag, 2002). Also, the reappearance of BIFs in the sedimentary record can be explained by a completely frozen ocean that might have shutdown global oceanic circulation, leading to stagnation and deep-water anoxia, and ultimately, buildup of dissolved iron in the water column (Kirschvink, 1992). Upon deglaciation, the resumption of oceanic circulation would re-oxygenate the ocean, causing iron to oxidize and BIFs to form. Since being introduced in the

1990s the snowball Earth hypothesis has become the favoured model to explain at least two of the Neoproterozoic glaciations as more data corroborates the theory of long-lived, runaway glaciation (e.g. Li et al., 2013; Rooney et al., 2015).

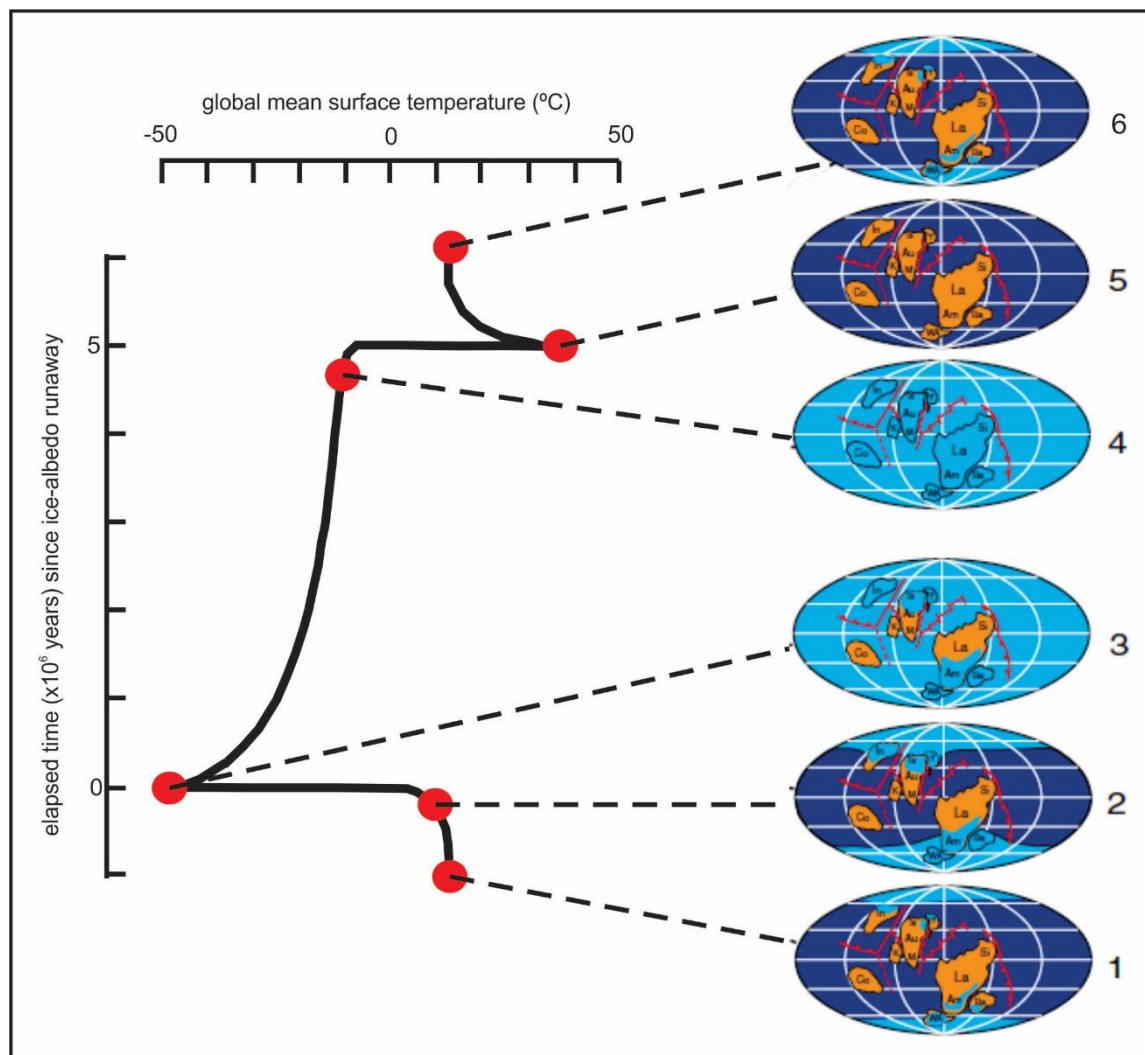


Figure 1.16: Schematic showing the estimated changes in global mean surface temperature during the development and onset of a snowball Earth event. (1) Ice sheets slowly develop at high latitudes. (2-3) Once the ice reaches a critical latitude ($\sim 30^\circ$), rapid worldwide glaciation ensues due to ice albedo feedback. (3-4) Surface temperature increases as CO_2 builds up in the atmosphere. (5) Once levels reach a critical threshold glacial ice swiftly retreats. (6) Increased rates of weathering and return of productivity in the oceans would reduce atmospheric CO_2 and lower surface temperatures (from Hoffman and Schrag, 2002).

The slushball Earth model is similar in many respects to snowball Earth except that open water conditions were sustained in equatorial regions during glaciation (Harland and Rudwick, 1964;

Fairchild and Kennedy, 2007). This is supported by multiple studies that report Neoproterozoic glacial deposits grading into open marine sediments (Eyles and Eyles, 1983; Leather et al., 2002; Smith, 2009) and the study of Arnaud & Eyles (2006) that reported the intercalation of glacial and non-glacial units. Together these studies suggest that the planet was not completely frozen or without a hydrological cycle (Fairchild and Kennedy, 2007).

IV.II $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ Overview

Despite the study of carbon isotopic compositions in carbonate rocks as far back as the 1950s (Craig, 1953), their utilization for illustrating secular variations in Proterozoic strata only began in the late 1970s/early 1980s (i.e. Veizer and Hoefs, 1976; Veizer et al., 1980; Knoll et al., 1986; Magaritz et al., 1986; Tucker, 1986). However, the 1990s saw an increase in these studies continuing to the present day as carbon isotope stratigraphy, or more generally, chemostratigraphy, has proven invaluable for correlating Neoproterozoic strata that otherwise lack the traditional means for stratigraphic correlation (i.e. fossils and radiometric dating).

The isotopic value of an element, expressed by lowercase delta (δ), is the difference in the ratio of the heavy, rarer isotope versus the light, more abundant isotope (e.g. $^{13}\text{C}/^{12}\text{C}$, $^{18}\text{O}/^{16}\text{O}$) compared to an internationally recognized standard (i.e. Vienna Pee Dee Belemnite, VPDB), with values presented in permil (‰) (Anderson and Arthur, 1983). Carbon isotopes are the most widely used to study Neoproterozoic rocks, which can be attributed to their ease of measurement, comparatively low analytical costs, resistance to alteration, and wide variations over short timescales (Halverson et al., 2010). It has been widely accepted that the carbon isotope composition of unaltered carbonate ($\delta^{13}\text{C}_{\text{carb}}$) precipitated in equilibrium with seawater approximates the isotopic composition of the dissolved inorganic carbon (DIC) reservoir in the fluid from which they formed, which typically is seawater (Anderson and Arthur, 1983; Halverson et al., 2010). Thus, it is believed that the history of the biogeochemical cycle is recorded in the $\delta^{13}\text{C}$ record in ancient sedimentary rocks (Des Marais, 2001).

During the autotrophic fixation of CO_2 , ^{12}C is preferentially taken-up over ^{13}C , leaving the DIC reservoir relatively enriched in ^{13}C (Kaufman and Knoll, 1995). Carbon being brought into Earth surface reservoirs (e.g. the ocean, through hydrothermal vents) is balanced by the burial of carbon as organic carbon, which is preferentially enriched in ^{12}C , or as calcium carbonate. Therefore, changes in the isotopic composition of carbonate deposited in the oceans is a function of the amount of organic carbon versus inorganic carbonate burial (Hayes et al., 1999; Schrag et al., 2013). Accordingly, an increase in $\delta^{13}\text{C}_{\text{carb}}$ indicates more organic carbon burial relative to inorganic, which in turn implies periods of elevated productivity, and vice versa for decreases. However, some authors have suggested that fluctuations in the Neoproterozoic carbon isotopic record are due to meteoric (Swart and Kennedy, 2012) or burial diagenesis (Derry, 2010), changes in authigenic carbonate precipitation (Schrag et al., 2013; MacDonald et al., 2013), or bacterial sulphate reduction (Nédélec et al., 2007) (see below for more details).

In the modern oceans, there is a general $\sim 2\text{‰}$ decrease in $\delta^{13}\text{C}_{\text{carb}}$ from the shoreline into deeper water offshore (Kroopnick, 1985). A similar trend has been identified in some Neoproterozoic examples (Jiang et al., 2007; Giddings and Wallace, 2009; Smith, 2009). Kroopnick (1985) attributed this to the operation of a shallow water biological pump. Specifically, in the shallow photic zone there is the preferential uptake of ^{12}C by primary producers that enriches the shallow surface ocean waters in ^{13}C , while at the same time particulate organic matter, depleted in ^{13}C , settles to the deep-ocean, resulting in lower $\delta^{13}\text{C}$ values. Alternatively, Swart (2008) reported that $\delta^{13}\text{C}$ variations in carbonate sediments in platform/ramp systems over the past 10 myr are different than the global pattern of open ocean $\delta^{13}\text{C}$ over the same interval and instead record the relative mixing of $\delta^{13}\text{C}$ enriched platform and depleted pelagic sediments, with lower isotope values indicating greater proportions of deep-water sediments and the opposite for higher values. This led to the suggestion that the $\delta^{13}\text{C}_{\text{carb}}$ measured in the ancient rock record might not record changes in the global carbon cycle or variations in the burial of organic carbon. However, the occurrence of isotopic variations that can be correlated between

different sedimentary basins are indisputable, but the factors that control these variations remain a source of much debate.

When primary productivity is high and the flux of organic sediment exceeds the rate of water column oxidation organic carbon accumulates on and in the sediment pile. (Anderson and Arthur, 1983). The carbon isotope ratios of sedimentary organic matter ($\delta^{13}\text{C}_{\text{org}}$) are comparatively more time-consuming to measure than inorganic carbon because samples must first be treated with acid to remove the inorganic (carbonate) carbon, and are more difficult to acquire than their inorganic counterparts because diagenesis and thermal alteration can distort the isotopic connection between extant total organic carbon (TOC) and original biomass (Halverson et al., 2010). The $\delta^{13}\text{C}_{\text{org}}$ values reflect the isotopic fractionation associated with biological fixation of CO_2 and its modification by various consumers (Des Marais, 2001) — present day values for warm-water marine phytoplankton cluster between -17 to -22‰ (Anderson and Arthur, 1983), but deep-water dissolved organic carbon (DOC) are slightly lower, -20 to -24‰ (Williams and Gordon, 1970; Eadie et al., 1978). Throughout most of earth history $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ have oscillated around 0‰ and -25‰, respectively, with the $\delta^{13}\text{C}$ of carbon entering the ocean closely approximated depleted mantle values of -6 ± 1 ‰ (Schrage et al., 2013).

Oxygen isotopes are typically analyzed in concert with $\delta^{13}\text{C}_{\text{carb}}$, making them relatively inexpensive and easy to acquire. However, $\delta^{18}\text{O}$ values are easily affected by recrystallization and metamorphism (Jacobsen and Kaufman, 1999), therefore, care must be taken to ensure the signal is pristine, especially in Neoproterozoic rocks. Studies have shown that $\delta^{18}\text{O}$ in ancient carbonates are depleted relative to their Cenozoic counterparts (Veizer et al., 1999). The origin of this change is controversial, with two schools of thought. The first proposed that the $\delta^{18}\text{O}$ of seawater has been constant through time, but the isotopic composition of oceanic crust has changed, which buffers the $\delta^{18}\text{O}$ signal (Gregory and Taylor, 1981; Meuhlenbachs, 1998). However, more recent studies have shown that the $\delta^{18}\text{O}$ values of oceanic crust has remained stable throughout geologic time, while $\delta^{18}\text{O}$

in carbonate has increased by $\sim 13\text{‰}$ over the past 3.4 Ga (Jaffrés et al., 2007). This supports the alternate view that $\delta^{18}\text{O}$ of seawater has in fact changed over time (Veizer et al., 1999).

IV.II.I Neoproterozoic Carbon Isotope Trends

The Neoproterozoic carbon isotopic record is characterized by extended periods with elevated $\delta^{13}\text{C}_{\text{carb}}$ with five globally-correlatable, large-amplitude negative excursions, three of which precede worldwide glaciations (Figure 1.17; Knoll et al., 1986; Kaufman and Knoll, 1995; Hoffman et al., 1998; Halverson et al., 2005; Halverson, 2006; Halverson et al., 2010). The two anomalies that lack sedimentological evidence of glaciation are the 800 Ma Bitter Springs stage, which is believed to be related to eustatic changes, and the Precambrian-Cambrian boundary (542 Ma). The negative excursions appear to predate any major fluctuation in sea-level associated with glaciation. Therefore, the onset of glaciation did not cause these precipitous falls, which must have been the result of difference forcing mechanisms (Halverson, 2006). Carbonates that cap Sturtian and Marinoan glacial deposits typically display an initial $\delta^{13}\text{C}_{\text{carb}}$ depletion stratigraphically upward, with values approaching those of the mantle ($-6\pm\text{‰}$). This is followed by an upward increase towards less negative values (Hoffman et al., 1998; James et al., 2001; Hoffman and Halverson, 2011). Hoffman et al. (1998) used these trends to support their snowball Earth hypothesis, suggesting that biological productivity in the surface ocean collapsed for millions of years, which in turn reduced the proportion of organic carbon to total carbon (Hayes et al., 1999). The complete shutdown of productivity would allow the isotopic composition of carbonates precipitated from ocean water following glaciation to approach the $\delta^{13}\text{C}$ values of carbon entering through hydrothermal input at mid-ocean ridges ($-6\pm 1\text{‰}$) (Hoffman et al., 1998; Hayes et al., 1999; Hoffman and Schrag, 2002). Reestablishment of the biological pump following glaciation would slowly drive $\delta^{13}\text{C}_{\text{carb}}$ to more positive values. Alternatively, other authors have ascribed the negative carbon excursions to other mechanisms, including temperature- or alkalinity-controlled fractionations (Higgins and Schrag, 2003), upwelling/mixing of $\delta^{13}\text{C}_{\text{carb}}$ depleted

bottom waters (Grotzinger and Knoll, 1995; Swart, 2008), methane hydrate destabilization (Kennedy et al., 2001), or decreases in the burial of authigenic carbonate (Schrag et al., 2013).

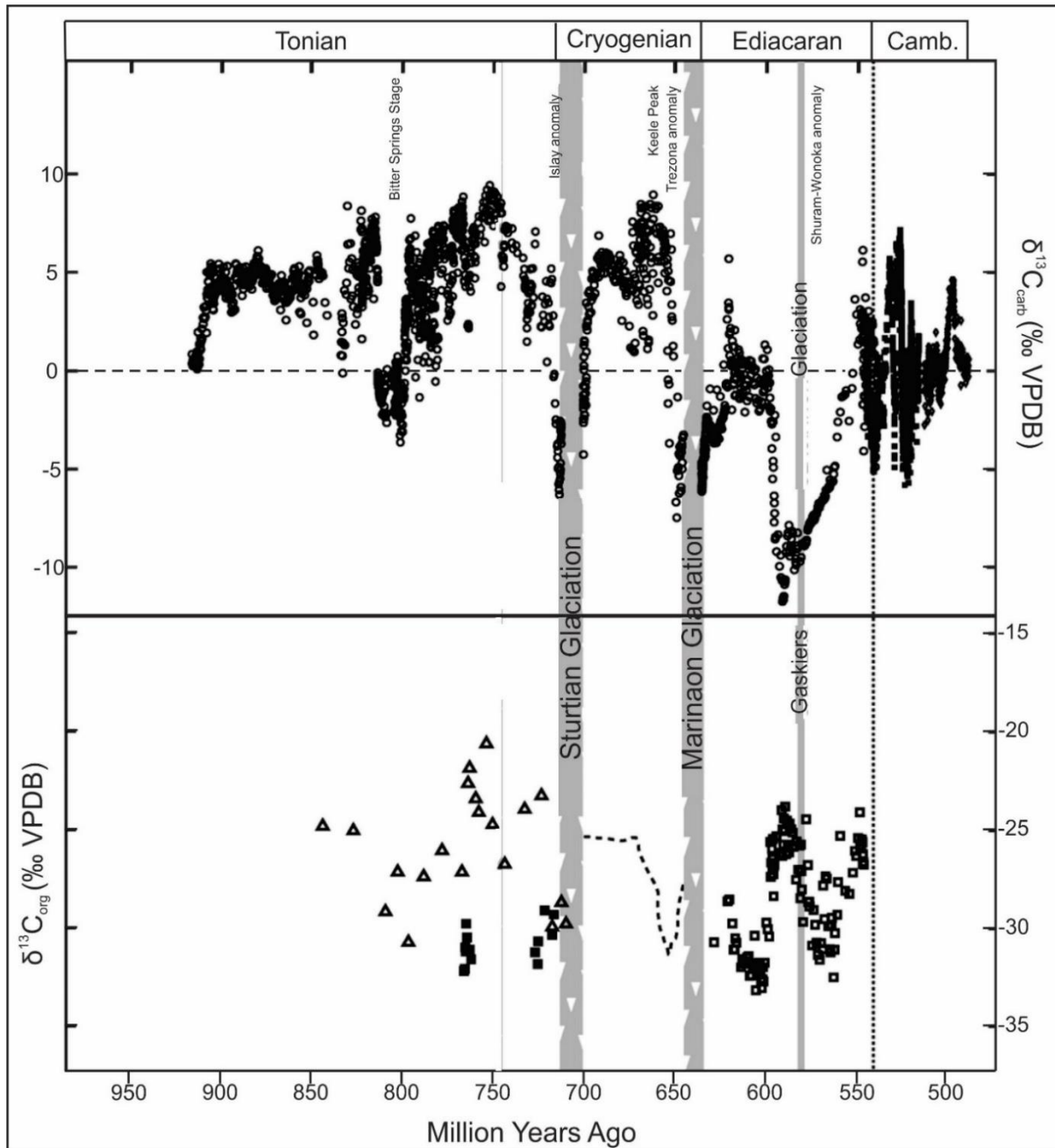


Figure 1.17: Composite Neoproterozoic $\delta^{13}\text{C}$ record from marine carbonates (modified after Halverson et al., 2010). Note the $\delta^{13}\text{C}_{\text{carb}}$ record is mostly positive with negative excursions preceding glaciations, and $\delta^{13}\text{C}_{\text{org}}$ broadly correlates the $\delta^{13}\text{C}_{\text{carb}}$ curve, but typically is $\sim 30\text{‰}$ more negative (e.g. Hayes et al., 1999). Nevertheless, there are periods when the two become decoupled, for example during the Shuram-Wonoka anomaly (associated with the Gaskiers glaciation according to this figure) (Fike et al., 2006). Note: since the making of this compilation the Sturtian glaciation has been expanded to terminate at ~ 660 Ma, while the Gaskiers glaciation is believed to precede the Shuram-Wonoka anomaly and is instead associated with a smaller-magnitude excursion (EN-2; Jiang et al., 2007).

Some of the earliest Neoproterozoic $\delta^{13}\text{C}$ datasets noted a broad covariance between $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ (Fig 1.17), with the difference between the two ($\Delta\delta$) being about 25-30‰ (Knoll et al., 1986; Kaufman and Knoll, 1995). However, more recent datasets from the Ediacaran show sustained intervals when inorganic and organic carbon isotopes were decoupled worldwide, particularly during the Shuram-Wonoka anomaly, which appears to be the largest known perturbation in the global carbon cycle in Earth history and is marked by $\delta^{13}\text{C}_{\text{carb}}$ values below the values associated with the hydrothermal input of carbon (Calver, 2000; Fike et al., 2006; McFadden et al., 2008; Verdel et al., 2011; MacDonald et al., 2013). Authors have attributed the divergence between $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ values and the Shuram-Wonoka excursion to be a response to the oxidation of Ediacaran oceans (Rothman et al., 2003; Fike et al., 2006). This resulted in the oxidation of the large DOC pool, transferring ^{13}C depleted carbon to the DIC, thus lowering its $\delta^{13}\text{C}_{\text{carb}}$, whereas the large size of the DOC reservoir buffered against isotopic change. Therefore, during this time, isotopic values from the DIC were limited by the $\delta^{13}\text{C}$ of the organic reservoir ($\sim 30\text{‰}$) and not the hydrothermal carbon input ($\sim -6\text{‰}$), resulting in highly depleted $\delta^{13}\text{C}_{\text{carb}}$ values. The variability in the magnitude of the excursion across the globe has led others to suggest the excursion is due to the zone of authigenesis expanding onto carbonate platforms during a transgression with disparities in mixing causing the differing excursion magnitudes (Schrag et al., 2013).

Much like late Cenozoic and modern systems (Kroopnick, 1985; Swart, 2008), Neoproterozoic basins display a shallow-to-deep water gradient in $\delta^{13}\text{C}_{\text{carb}}$ ranging from 4-14‰ due to the operation of a shallow water biological pump and/or increased authigenic carbonate development (Calver, 2000; Jiang et al., 2007; Giddings and Wallace, 2009; Smith, 2009; Shen et al., 2011). This gradient is best developed in the Neoproterozoic Doushantuo Formation of South China where shallow deposits are mostly positive and record four negative excursions, while deep-water sections are consistently negative with lower magnitude fluctuations (Jiang et al., 2007). These $\delta^{13}\text{C}_{\text{carb}}$ gradients are interpreted to be proof of long-term deep-ocean anoxia in relation to ocean stratification, (Giddings and Wallace,

2009; Smith, 2009). The preservation of organic matter that cascaded down into the deep-marine would have developed a large deep-water reservoir depleted in $\delta^{13}\text{C}_{\text{carb}}$, which, because of stratification, was not returned to the surface, ultimately causing a drawdown of atmospheric CO_2 that quite possibly resulted in glaciation (Smith, 2009).

A concern with Neoproterozoic $\delta^{13}\text{C}$ data is whether it represents primary seawater conditions or secondary diagenetic overprints (Knauth and Kennedy, 2009). Studies of Miocene to recent carbonate platform sediments have shown isotopic variations that are unrelated to changes in the oceanic carbon cycle, instead likely reflecting eustatic changes associated with glacial-interglacial cycles (Swart, 2008) and widespread $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ correlatability between sections that have undergone meteoric diagenetic alteration (Swart and Kennedy, 2012). However, several authors, for example Halverson et al. (2010), suggest that the worldwide reproducibility of isotopic trends are unlikely to be a result of local diagenesis and instead reflect changes in the global carbon cycle.

Numerous studies have indicated that the covariance of $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}$ is a manifestation of diagenetic alteration (eg. Marshall, 1992; Melim et al., 2004; Knauth and Kennedy, 2009; Derry, 2010). Derry (2010), for example, used covariance to question the validity of isotopic values associated with the Shuram/Wonoka anomaly (~ 560 Ma), hypothesizing that it was the result of diagenetic alteration (i.e. fluid-rock and fluid-fluid interaction) during burial diagenesis, which then was modelled to reproduce other Neoproterozoic excursions. However, Husson et al. (2012) later reported that the observed isotopic covariance was in fact acquired by Wonoka Formation platform rocks prior to being resedimented in offshore canyons, and therefore was a primary signal and not the result of meteoric or burial diagenesis. It has been widely reported that even if diagenetically altered, samples are more likely to preserve their original $\delta^{13}\text{C}$ composition rather than $\delta^{18}\text{O}$ or $^{87}\text{Sr}/^{86}\text{Sr}$ because the amount of dissolved carbon in diagenetic fluids is substantially less than in carbonate rocks, and samples subjected to metamorphic conditions at or below greenschist facies are insufficiently thermally altered to change their carbon isotope composition (Brand and Veizer, 1980; Brand and Veizer, 1981; Banner

and Hanson, 1990; Jacobsen and Kaufman, 1999; Veizer et al., 1999; Des Marais, 2001; Halverson et al., 2005). Despite this, the potential of diagenetic alteration must be evaluated when studying Neoproterozoic sections. This has led to the establishment of a variety of petrographic and geochemical procedures to ensure the near-primary isotopic signal is being recorded (e.g. Kaufman and Knoll, 1995), several of which have been utilized in this study (see results section in Chapter 3)

IV.III Previous Post-Marinoan Windermere Isotope Studies

Due to the dominance of siliciclastic strata in deep-water deposits of the WSG in the SCC, isotopic studies have been sparse. Of note then is the work of Ross et al. (1995) who studied $\delta^{34}\text{S}$ values from large, disseminated pyrite crystals in rocks of the Middle to Upper Kaza Group and Isaac Formation. Results showed a wide range of $\delta^{34}\text{S}$, suggesting that sulphides were sourced from bacterial reduction of seawater sulfate, with a trend toward more positive values stratigraphically upward. This was attributed to the preferential sequestration of ^{32}S along continental margins, leading to oceanic ^{34}S enrichment and potentially the oxidation of large amounts of organic carbon that caused $\delta^{13}\text{C}_{\text{carb}}$ to be depleted. Despite this proposal for profound changes in the oceanic carbon reservoir, and the recognition of several carbonate-rich units in the WSG, particularly in the Isaac and Cunningham formations (Ross and Murphy, 1988; Ross et al., 1995), only the Ph.D. work of Smith (2009) in the OFP has attempted to further these concepts

In the WSG of the northern Canadian Cordillera, calcareous strata are much more prevalent than in the SCC, and, as a consequence, carbon isotope studies are more numerous (e.g. Narbonne et al., 1994; Kaufman et al., 1997; James et al. 2001; Hoffman and Halverson, 2011; MacDonald et al., 2013). In this region Sturtian and Marinoan glacial deposits are represented by the Rapitan and Ice Brook formations, respectively (Aitken, 1991 a,b). Sharply overlying the Ice Brook diamictites are buff- to yellow-weathering, granular to peloidal dolostones of the Ravensthorpe Formation, overlain by dark-grey, thin- to medium-bedded limestones of the Hayhook Formation, containing variously

shaped aragonite fan pseudomorphs (James et al., 2001). Together, the Ravensthorpe and Hayhook make up the post-Marinoan cap carbonate (Eisbacher, 1981; Aitken, 1991b; James et al., 2001). The Hayhook is then overlain by black shales of the Sheepbed Formation, that transition upward to dolostone in the upper hundred metres of the formation (informally termed the Sheepbed carbonate by Aitken 1982; MacDonald et al., 2013). These three formations are interpreted to have been deposited during the post-Marinoan transgression with the maximum flooding surface occurring in the black, fissile shales of the Sheepbed. Upper slope deposits of the June beds unconformably overlie the Sheepbed carbonate and consist of debrites overlain by a thick succession of thin-bedded, lime-rich mudstone and brown siliciclastic siltstone turbidites (MacDonald et al., 2013). The June beds in turn, are overlain by thin-bedded limestone interbedded with debrites of the Gametrail Formation, overlain by siliciclastic mudstone interbedded with siliciclastic sandstone turbidites and ribbon-bedded carbonates of the Blueflower Formation (Narbonne et al., 1994). The uppermost Neoproterozoic in the Mackenzie Mountains consists of crystalline dolostone of the Risky Formation (Narbonne et al., 1994).

Carbonate carbon isotopes in the basal Ravensthorpe are negative (i.e. depleted) and become increasingly negative in the overlying Hayhook. However, this trend reverses in the Sheepbed Formation, where values cross over to positive below the Sheepbed carbonate and then plateau at about 4‰ at the top of the unit (Figure 1.18; Narbonne et al., 1994; Kaufman et al., 1997; James et al., 2001; Hoffman et al., 2011; MacDonald et al., 2013). In the OFP Formation of the southern Cordillera a similar trend is observed, as values in the siltstone-limestone rhythmites of the basal Temple Lake Member (TLM) are as low as -12‰ and progressively increase to near 0‰ at the top of the Whitehorn Mountain Member (WMM) (Figure 1.18- Same as above; Smith, 2009). Utilizing these geochemical trends and lithological similarities, Smith (2009) correlated the OFP with the Ravensthorpe-Hayhook-Sheepbed succession in the Mackenzie Mountains (Figure 1.19 - Smith correlation). Specifically, the basal, green to purple siltstone-rich basal part of the TLM correlated with the Ravensthorpe, and the overlying limestone-siltstone couplets with the Hayhook, suggesting that the TLM is the deep-marine

equivalent of a cap carbonate. Organic-rich shales in the overlying Geikie Siding Member (GSM) were interpreted to indicate maximum flooding conditions and were correlated with the lower, black-shale-rich portion of the Sheepbed Formation. Coarse-grained submarine canyon fills at the base of the WMM contain clasts of bladed-calcite crystals similar to those in the Hayhook Formation and are interpreted to represent lowstand deposition containing eroded fragments of shelf carbonates. Correlation of the upper part of the WMM with strata in the Mackenzie Mountains is more poorly constrained because coarse submarine canyon strata are absent and instead consist of fine-grained mud-rich turbidites (Smith, 2009). Nevertheless, the top of the OFP is marked by a shift in the FeHR/FeT ratio indicated a change from anoxic to oxic conditions (Smith, 2009), which may correlate to a similar shift in the Sheepbed Formation (Shen et al., 2008).

Correlation of the OFP with the glacioeustatic transgression that followed Marinoan glaciation is made questionable by the Re-Os date of 607.8 ± 4.7 Ma from organic-rich black shale in the GSM at Jasper, Alberta (Kendall et al., 2004). This age was interpreted as a minimum constraint for the timing of the Marinoan glaciation in western Canada. However, recent work has obtained an Re-Os date of 632.3 ± 5.9 Ma for the basal Sheepbed Formation (Rooney et al., 2015), which is consistent with worldwide post-Marinoan U-Pb ages that cluster around 635 Ma (e.g. Hoffman et al., 2004; Condon et al., 2005; Calver et al., 2013). Paradoxically, Kendall et al. (2004) initially obtained a less precise Re-Os date of 634 ± 57 Ma using the inverse aqua regia digestion medium method (rather than the CrO_3 - H_2SO_4 technique in the 607.8 Ma date). Reasons to reconcile this apparent contradiction are: 1) the 607.8 Ma date is accurate and the TLM and basal GSM were deposited over a glacioeustatic sea-level rise that lasted for more than 25 myr, 2) the OFP is unrelated to post-Marinoan transgression, or 3) the date is inaccurate. The first two explanations seem unlikely based on the lithological (e.g. the presence of bladed calcite crystals unique to the Hayhook in all WSG successions) and $\delta^{13}\text{C}_{\text{carb}}$ similarities of the OFP and Ravensthorpe-Hayhook-Sheepbed successions (James et al., 2001; Smith, 2009).

Therefore, the 607.8 Ma date may be the source of the problem, but to date no attempt has been made to re-date the GSM.

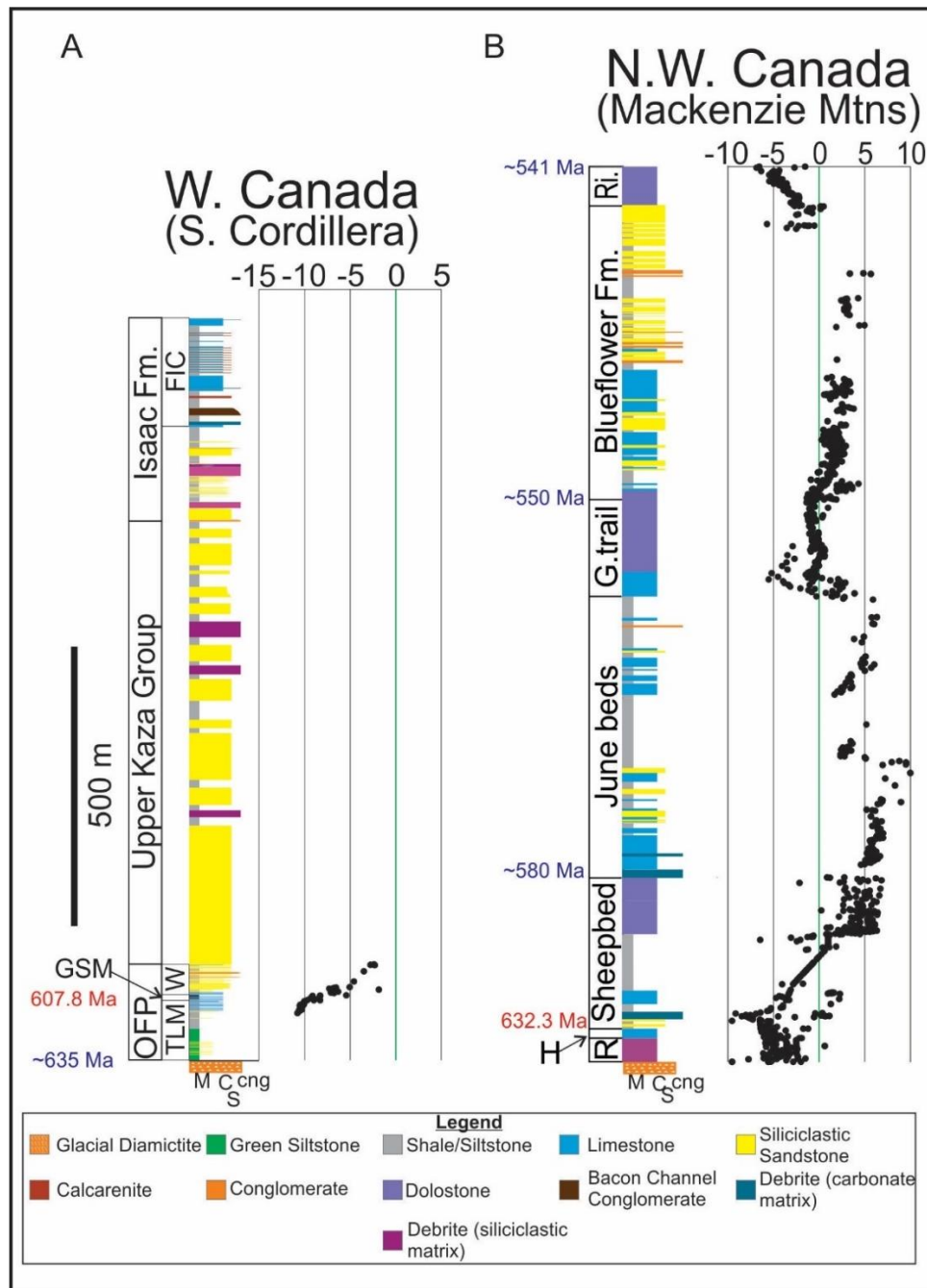


Figure 1.18: Post-Marinoan Marinoan $\delta^{13}\text{C}_{\text{carb}}$ record of the Windermere Supergroup in the (A) southern Canadian Cordillera (Smith, 2009) and (B) Mackenzie Mountains (modified from MacDonald et al., 2013). Red numbers represent Re-Os ages (Kendall et al., 2004; Rooney et al., 2015), blue numbers represent inferred ages based on biostratigraphy or stratigraphic correlation (MacDonald et al., 2013). OFP = Old Fort Point Fm., TLM = Temple Lake Mb., GSM = Geikie Siding Mb., W = Whitehorn Mountain Mb., FIC = first Isaac carbonate, R = Ravensthorpe Fm., H = Hayhook Fm., G.trail = Gametrail Fm., Ri = Risky Fm.

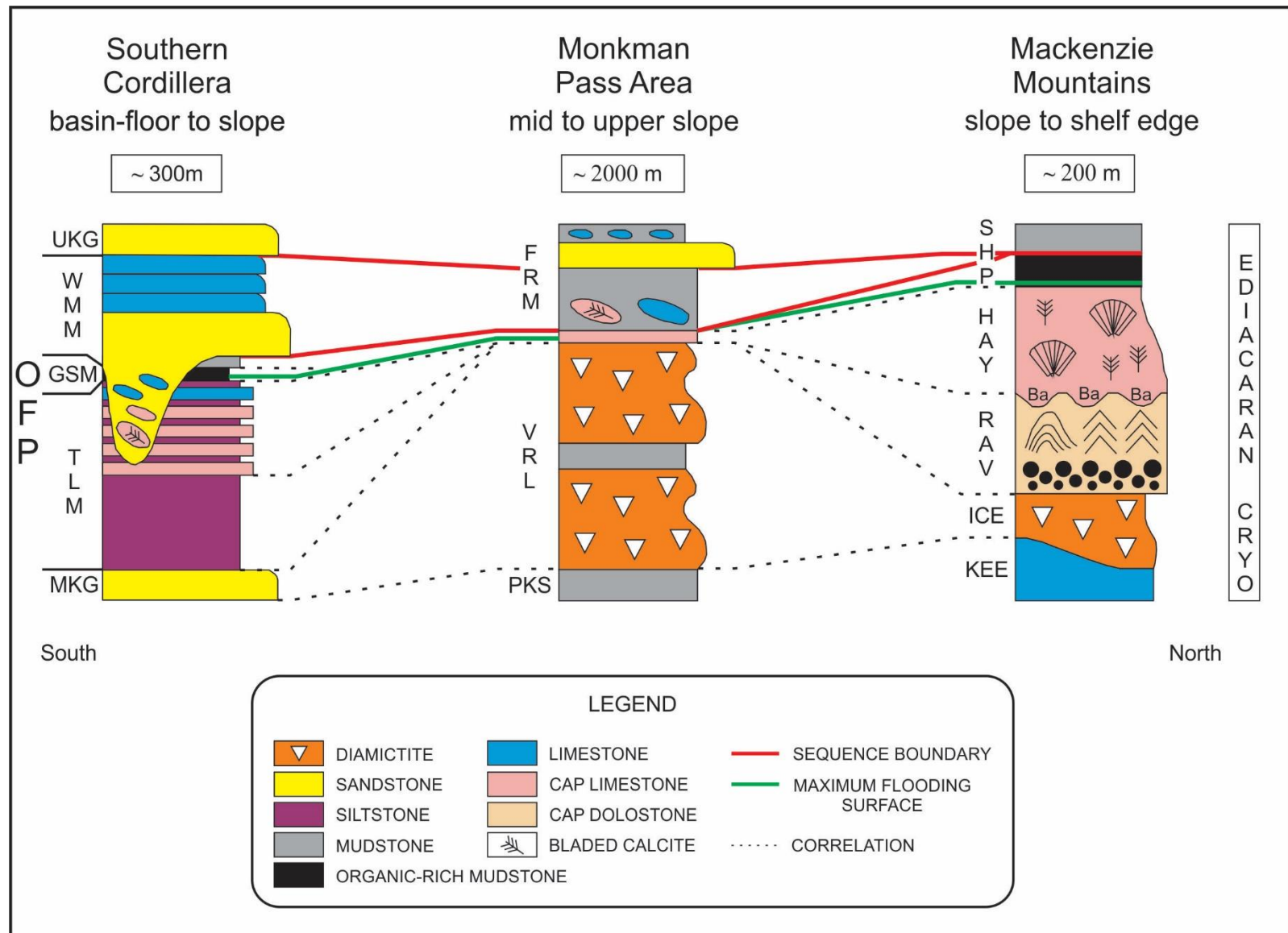


Figure 1.19 (previous page): Stratigraphic correlation of Cryogenian to Ediacaran strata of the Windermere Supergroup in the southern Canadian Cordillera, Monkman Pass area and Mackenzie Mountains, which represent different paleogeographic positions along a shelf to basin profile. Marinoan glaciation is represented by the Middle Kaza Group (MKG) in the southern Canadian Cordillera, Vreeland Fm. (VRL) in the Monkman Pass area and Ice Brook Formation (ICE) in the Mackenzie Mountains. The minimum age of the terminal Marinoan is constrained by a 632.3 ± 5.9 Ma Re-Os date from the basal part of the Sheepbed Formation, whereas a 607.8 ± 4.7 Ma Re-Os date constrains the age of black shale in the Geikie Siding Member (GSM); note that, this date remains a matter of debate (see text; Smith, 2009; Rooney et al., 2015). The Old Fort Point Fm. (southern Canadian Cordillera), Framestead Fm. (Monkman Pass) and Ravensthorpe-Hayhook-Sheepbed formations (Mackenzie Mountains) are interpreted to represent deposition during the transgression associated with the global Marinoan deglaciation (modified from Smith, 2009).

Above the Sheepbed Formation $\delta^{13}\text{C}_{\text{carb}}$ values remain positive throughout the basal June beds, with a small, less positive excursion midway through the unit before recovering to $\sim 5\text{‰}$ (MacDonald et al., 2013). The contact between the June beds and overlying Gametrail Formation is marked by a steep decline ($> 10\text{‰}$ in some sections), before rebounding to positive values in the upper Gametrail/lower Blueflower. Above the basal Blueflower there are fewer carbonates, leaving the $\delta^{13}\text{C}_{\text{carb}}$ record sparse until the Risky Formation, which records a negative excursions interpreted to mark the Precambrian-Cambrian boundary (Narbonne et al., 1994; Kaufman et al., 1997; MacDonald et al., 2013).

Part V Thesis Objectives and Structure

The primary objective of this thesis was to build on the pioneering work of Navarro (2016) by conducting a detailed analysis of the sedimentological and geochemical make-up of the FIC in the Castle Creek and Milk River study areas in order to reconstruct the depositional paleogeography and physiochemical conditions in the Neoproterozoic Panthalassa Ocean, and potentially correlate the FIC with strata of the WSG in the Mackenzie Mountains and other Ediacaran datasets worldwide. Two new study areas have been included: the Hill Section (HS), Castle Creek's most northwestern exposure, and Milk River (Figure 1.8), which allows stratigraphic and geochemical relationships to be analyzed over a 20 km transect. Field work assisted by Quinn Dabros and Sean Ludzki, was conducted over the summers of 2015 and 2016 (~ 12 weeks total) and included the following:

- At the Hill Section six stratigraphic logs (~ 150 m thick) were measured at cm-scale through sandstone and siliciclastic mud-rich intervals and mm-scale through the calciturbidite and 123 units. The entire exposure was mapped at 1:500 scale highlighting the distribution of stratigraphic elements across the study area (Chapter 2).
- At Milk River the entire FIC was mapped at 1:500 scale and two complete stratigraphic logs on the south (MR-S) and north (MR-S) side of the glacier were measured in cm detail (Chapter 2).
- On the south side of the Castle Creek glacier three cm-scale logs were measured through the 123 units, with each sandstone > ~ 5 cm-thick being traced laterally across the ~ 250 m wide outcrop. This was done to better elucidate the stacking pattern of these lithologically distinctive beds and how the sandstone packages change laterally (Chapter 2).
- 208 samples were collected from strata of the FIC at the three study locations using a sample spacing interval of 2-3 m in carbonate-rich intervals and 5 m in siliciclastic-rich intervals. $\delta^{13}\text{C}_{\text{carb}}$ analysis was conducted on the primary cements, which were first identified in thin section and through cathodoluminescence microscopy, to determine oceanographic conditions during development of the FIC and to compare those data with the global post-Marinoan database (Chapter 3).

Work at the University of Ottawa included:

- Analysis of stained thin sections for identification of the carbonate cements in FIC strata at the study areas (Chapter 3).
- Cathodoluminescence microscopy of select samples to aid in the identification of primary marine cements at the Microanalytical Laboratory (Chapter 3).

- $\delta^{13}\text{C}_{\text{carb}}$, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}^{\text{org}}$ analysis at the G.G. Hatch Stable Isotope Laboratory. The samples were prepared and treated by me, with the help of Paul Middlestead and Wendy Abdi. The samples were analyzed by Wendy (Chapter 3).
- Manganese/strontium ratios from 25 siliceous calcilutite and fine-grained siliceous calcarenite samples were measured at the University's Geochemistry Laboratory by Dr. Nimal De Silva (Chapter 3).

Two chapters make up the body of this thesis. Both of which are written in paper-format style for future submission to peer-reviewed journals. Chapter 2 describes the facies identified in the FIC and compares the vertical and lateral distributions of the stratigraphic elements that make up the FIC across the primary study areas to determine its long-term evolution. The third chapter reports on trends in the $\delta^{13}\text{C}_{\text{carb}}$ data to (chemostratigraphically) correlate between the study areas and to better understand the oceanographic and climatic conditions that led to the development of this stratigraphic outlier. These data are then compared with previously published post-Marinoan curves to determine if the changes are global or instead more geographically restricted. First-draft versions of both manuscripts were written by this author with discussions and editorial assistance provided by Dr. Bill Arnott.

Chapter 2: Comparison of Stratal Elements and Their Spatial Distribution in a Neoproterozoic Mixed Carbonate-Siliciclastic Base-of-Slope System, Windermere Supergroup, Canadian Cordillera, British Columbia

Part I Introduction

Compared to their siliciclastic counterparts, which have been thoroughly studied in both modern and ancient settings (e.g. Normark, 1978; Clark and Pickering, 1996; Pasamentier and Kolla, 2003; Posamentier and Walker, 2006; Deptuck et al., 2007; Kolla et al., 2007; Nilsen et al., 2007; Huenneke and Mulder, 2010; McHargue et al., 2011), deep-marine slope sedimentation processes and deposits in carbonate and mixed siliciclastic-carbonate, hereafter termed mixed, have received much less attention and remain more poorly understood, particularly those deposited in tectonically passive margin settings. Today there remains debate in mixed and carbonate systems about when siliciclastic versus carbonate sediments are supplied to the slope due to changes in relative sea-level (e.g. Wilson, 1967; Droxler and Schlager, 1985; Peerdeman et al., 1993; Schlager et al., 1994; Dunbar and Dickens, 2003; Page et al., 2003; Francis et al., 2007; Harper et al., 2015), how they are sourced from the shallow-marine platform and subsequently moved basinward (e.g. Cook et al., 1983; Mullins et al., 1984; Braga et al., 2001; Payros and Pujalte, 2008 and references therein) and details about the morphology of their transport systems (e.g. Payros and Pujalte, 2008; Puga-Bernabéu et al., 2014; Mulder et al., 2014).

Currently there are two prominent models for predicting the relationship between relative sea-level and the export of siliciclastic, mixed or carbonate sediments into deeper waters in carbonate and mixed systems. The first is the long established reciprocal sedimentation model, also known as highstand shedding, in which siliciclastic material is deposited during sea-level lowstands and carbonate sediment is shed from the platform during highstands (Wilson, 1967; Sarg, 1988; Posamentier and Vail, 1988; Dolan, 1988; Handford and Loucks, 1993; Schlager et al., 1994). The

second model (transgressive shedding) is based on observed sedimentation patterns along the Great Barrier Reef (GBR) and the Gulf of Papua following the last glacial maximum (LGM), at ~ 18 Ka. During the LGM lowstand sedimentation rates for both siliciclastics and carbonates were low, but then increased dramatically during the subsequent transgression before being dominated by carbonate sedimentation during the highstand (Peerdeman and Davies, 1993; Dunbar et al., 2000; Page et al., 2003; Francis et al., 2007; Webster et al., 2012; Puga-Bernab   et al., 2014; Harper et al., 2015). Additionally, older stratigraphy along the GBR indicates that sea-level rise occasionally drowned the platform and caused the sediment supply to become predominantly siliciclastic. In these cases carbonate production and downslope resedimentation did not return until the ensuing regression sufficiently reduced water depth (Droxler et al., 1993).

Recent work has shown that deposits in mixed and carbonate systems show similar morphological and sedimentological characteristics to those in siliciclastic dominated basins (e.g. Mulder et al., 2012a; Payros and Pujalte, 2008), but a scaling issue exists, as most mixed systems are up to an order of magnitude smaller than their siliciclastic counterparts. In mixed systems there is also a sensitive interplay between platform development, eustatic change, and the morphology of the shelf-edge and upper slope canyons that influence the type of sediment supplied to the deep-marine. For example, along the Great Barrier Reef siliciclastic and mixed turbidites were deposited in the deep-marine when canyons deeply incised the shelf, whereas strict carbonate sedimentation was associated with times when the heads of canyons terminated basinward of the shelf-slope break (Puga-Bernab   et al., 2014). Moreover, carbonate sediment in carbonate/mixed systems can be line-sourced forming strike-elongated aprons, or like the situation in most siliciclastic systems, point sourced (Vigorito et al., 2006). Therefore, the depositional characteristics of mixed and carbonate systems may not apply directly to models created for their siliciclastic counterparts. However, offshore hydrocarbon exploration has recently begun to target reservoirs located basinward of mixed shelves, for example in offshore Nova Scotia and West Africa (Sanchez et al., 2017), highlighting the need for detailed studies

of deep-marine mixed systems to better understand their sedimentological and stratal architectural make-up in order to better predict reservoir distribution and how it compares to siliciclastic systems (e.g. are channels as laterally continuous as in siliciclastic systems? Are they similarly connected to their associated levee deposits? How is hydrocarbon potential diminished/enhanced by the presence of carbonate in the system?).

This study focuses on the first Isaac carbonate (FIC) in the Neoproterozoic Isaac Formation, Cariboo Mountains, east-central British Columbia. In this area the FIC is a well-exposed, regionally extensive base-of-slope mixed system that forms an easily recognized regional marker in a predominantly siliciclastic pile of deep-marine sedimentary rocks of the Windermere Supergroup (WSG) (Ross et al., 1989; Ross et al., 1995). Despite its stratigraphic significance, little sedimentological attention has been paid to the FIC, excepting the recent PhD work of Navarro (2016) at Castle Creek and Mount Quanstrom. This study builds on the work of Navarro (2016) and includes two new study areas, the Hill Section and Milk River (Figure 2.1C) in order to provide a more complete sedimentological and stratigraphic description of the architecture and component stratal elements that make-up a Neoproterozoic deep-marine mixed slope system. To date, mixed systems have principally been studied using seismic imaging in modern settings, but the resolution is generally too coarse to resolve small-scale stratigraphic and sedimentological detail (e.g. Francis et al., 2008; Mulder et al., 2012a,b; Mulder et al., 2014; Puga-Bernab   et al., 2014), or outcrops that lack sufficient vertical and/or lateral continuity (e.g. George et al., 1997; van Konijnenburg et al., 1999; Braga et al., 2001). Accordingly, few outcrop studies have been able to discern the depositional architecture of the sedimentary bodies that populate deep-marine mixed systems (e.g. Murru et al., 2001; Vigorito et al., 2005; Vigorito et al., 2006). The exceptional preservation and outcrop exposure of the FIC in the study areas allows for the detailed logging and mapping of stratal elements at both sub-seismic and seismic scales, in addition to illustrating how the stratal elements and their distribution changed in both time and space due to changing patterns of transport and deposition. The objective of this study, therefore,

is to provide an in-depth analysis of the facies and stratal elemental make-up of an ancient mixed system and propose a model for how it evolved over-time.

Part II Geologic Setting

II.I Windermere Supergroup

The WSG is a succession of Neoproterozoic rocks exposed locally from northwestern Mexico to the Yukon-Alaska border region, a strike length of just over 4000 km (Figure 2.1A); Ross and Arnott, 2007). Outcrops in Mexico and the USA comprise continental and shallow marine strata (Link et al., 1993), whereas those in the southern Canadian Cordillera (SCC) consist of superbly exposed deep-marine, siliciclastic-dominated metasedimentary rocks (Campbell et al., 1973; Ross et al., 1995). Further northward in the Mackenzie Mountains of the northern Canadian Cordillera (NCC) rocks of the WSG are dominated by carbonate-rich shallow-marine and continental shelf facies (Aitken, 1991a,b; Narbonne et al., 1994).

In the SCC deep-marine rocks of the WSG are well exposed in thrust sheets of the Foreland fold and thrust belt and the Omineca belt; their boundary marked by the Southern Rocky Mountain Trench (SRMT) (Ross and Arnott, 2007; Smith, 2009). These are two of the five NW-SE trending morphological belts that make-up the Late Jurassic to Early Tertiary Western Canadian Cordillera; a physiographic feature related to collisional tectonics along the western margin of Laurentia (Monger et al., 1982; Ross and Arnott, 2007; Smith, 2009). Orogenic deformation resulted in significant shortening of the deep-marine outcrop belt, which currently is exposed over an area of $\sim 35,000 \text{ km}^2$ in the SCC, which if conservatively palinspastically restored may represent a system of more than $80,000 \text{ km}^2$, making it one of the world's largest ancient turbidite systems (Ross and Arnott, 2007).

Accumulation of the WSG proceeded in two phases (Stewart, 1972). The first coincided with rifting and breakup of the Rodinia Supercontinent, which in southeastern British Columbia consists of laterally discontinuous glacial diamictites associated with Sturtian glaciation (Toby Formation) and

extrusive volcanic rocks (Irene Formation) deposited in isolated rift basins. The second episode represents thermally-driven subsidence along the developing Laurentian passive continental margin (ancestral North America) and deposition of a 5-7 km-thick, post-rift sequence, which in the Cariboo Mountains of east-central B.C. comprises laterally continuous sedimentary units of the Kaza and overlying Cariboo groups (Figure 2.1B; Ross, 1991; Ross and Arnott, 2007). Basin floor deposits of the Kaza Group make up the basal 2-4 km of the post-rift succession, which is subdivided into a Lower mudstone dominated part and coarse sandstone-rich Middle and Upper parts, interpreted to represent lobe deposition on a more or less unconfined basin floor (Ross and Murphy, 1988; Ross, 1991; Ross et al., 1995; Meyer and Ross, 2007; Terlaky and Arnott, 2014; Terlaky et al., 2015). The Middle and Upper Kaza groups are separated by the regionally extensive Old Fort Point Formation (OFP), a distinctive lithological and geochemical marker that is interpreted to have been deposited during the eustatic rise that followed the Marinoan glaciation (650-635 Ma) (Ross and Arnott, 2007; Smith, 2009; Smith et al., 2011; Smith et al., 2014a,b).

The overlying up to 5 km-thick Cariboo Group is subdivided into three stratigraphic units. The up to 2.4 km-thick Isaac Formation conformably overlies the Upper Kaza Group and is composed primarily of mudstone (levee deposits) that encase laterally discontinuous, ~ 20-100 m-thick, coarse sandstone filled leveed-channels (Ross and Arnott, 2007; Arnott et al., 2011). The interpreted base-of-slope setting is based on its mud-dominated nature and commonality of mass movement deposits, suggestive of gravitational (slope) instability. In the Isaac Formation are two regionally continuous carbonate-rich units, informally termed the first and second Isaac carbonates (FIC and SIC) (Figure 2.1B; Ross et al., 1989; Ross, 1991; Ross et al., 1995; Ross and Ferguson, 2003). The FIC (this study) comprises three laterally extensive, decametre-thick units of very thin- to thin-bedded carbonate turbidites, hereafter referred to as calciturbidites), and very fine- to fine-grained siliceous calcarenites intercalated locally with up to ~ 20 m-thick successions of granule siliceous calcirudite to medium-grained siliceous calcarenite. These strata are intercalated with up to ~ 60 m-thick units of siliciclastic

mud dominated, fine-grained, thin-bedded turbidites. The SIC is similar to the FIC, but with local intercalated sheets of quartz arenite (Ross and Ferguson, 2003).⁷

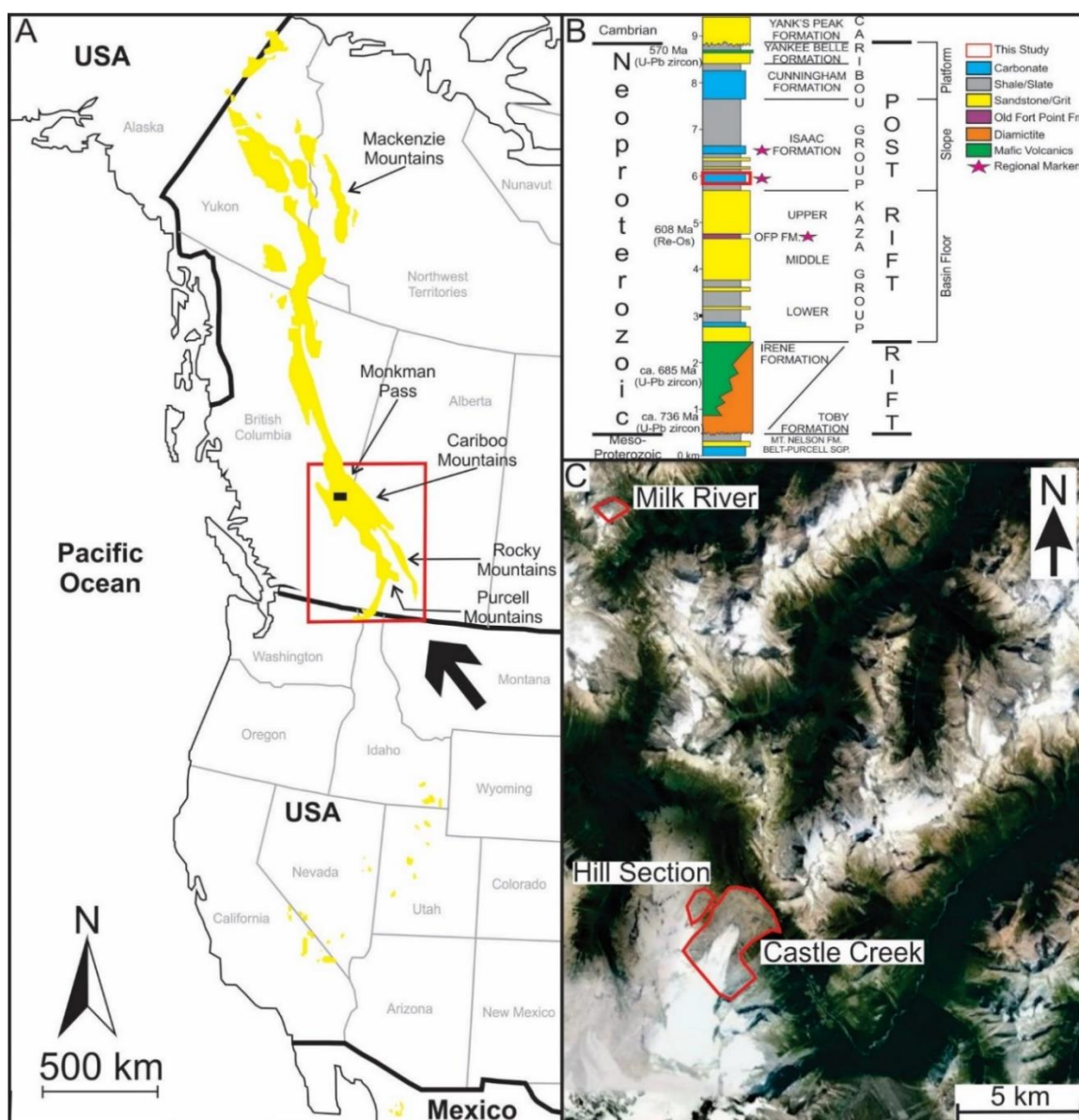


Figure 2.1: (A) Outcrop of the Windermere Supergroup in western North America with the deep-marine portion delineated by the red rectangle and the location of this study is the black box. The black arrow identifies the principle sediment transport direction of the Windermere turbidite system (redrawn from Ross, 1991 and Terlaky, 2014). (B) Generalized stratigraphic section of the WSG in the southern Canadian Cordillera (SCC). The basal ~ 2 km comprises the syn-rift Irene and Toby formations, overlain by a post-rift, upward-shoaling passive margin succession (Kaza Group, Isaac, Cunningham and Yankee Belle formations). This study focuses on the first Isaac carbonate, which is highlighted by a red rectangle (modified after Ross and Arnott, 2007). In ascending stratigraphic order, geochronological dates are from McDonough and Parrish (1991), Lund et al. (2003), Kendall et al. (2004) and Colpron et al. (2002). (C) Satellite image near McBride, British Columbia with the study areas indicated by red polygons – Castle Creek (CC), Hill Section (HS) and Milk River (MR) (background image © Province of British Columbia 2017, © CNES/Airbus 2017, © DigitalGlobe 2017, © Google 2017).

Conformably overlying the Isaac Formation are upper slope to high-energy shelf deposits of the Cunningham (oolitic limestones with minor mudstones) and the Yankee Belle formations (alternating limestone, siltstone, sandstone and shale) (Ross et al., 1995). The top contact of the Yankee Belle is a regional unconformity that separates the top of the WSG from Lower Cambrian rocks of the Yank's Peak Formation and is responsible for the lack of shallow-marine and continental WSG strata in eastern outcrops (Aitken, 1969; Ross et al., 1995; Ross and Arnott, 2007).

The dominance of siliciclastic strata, lack of volcanic rocks (except in the basal rift strata) and the absence of fossils throughout the WSG in the SCC results in sparse geochronological control (Ross and Arnott, 2007). The timing of deposition is constrained between 740 ± 36 to 728 ± 8 Ma (U-Pb) from underlying crystalline basement rocks in the Canadian Cordillera (Evenchick et al., 1984; Parrish and Scammell, 1988; McDonough and Parrish, 1991) and a 569 ± 4.6 Ma U-Pb date from volcanic rocks associated with rifting that resulted in the unconformity that caps the WSG (Colpron et al., 2002). The only date within WSG strata in the SCC is a rhenium-osmium (Re-Os) age of 607.8 ± 4.7 Ma from an organic mudstone in the Geikie Siding Member of the OFP (Kendall et al., 2004).

II.II First Isaac Carbonate

Just above the Kaza-Isaac contact, carbonate sediment was reintroduced into the Windermere turbidite system in the form of distinctively red-coloured, carbonate-cemented sandstone (calcareous silicarenite/siliceous calcarenite), in which carbonate represents the neomorphism of original carbonate grains from aragonite (Hardie, 2003) to calcite along the base of slope during early diagenesis. Approximately 150 m above this contact, at the base of the FIC, the proportion of carbonate increases dramatically with the presence of decametre-thick packages of fine-grained, very thin- to medium-bedded calciturbidites. The mixed carbonate-siliciclastic nature of the FIC interrupts an otherwise siliciclastic dominated succession of deep-water sedimentary rocks in the Cariboo Mountains. Regional mapping of the SCC led to the recognition

of the FIC, which was determined to be regionally extensive and thus recognized as a regional marker (Ross et al., 1989; Ross et al., 1995; Ross and Ferguson, 2003; Ross and Arnott, 2007), and therefore held great promise for understanding the sedimentological, geochemical and eustatic changes that led to the deposition of this stratigraphic outlier. Nevertheless, in spite of its recognition as far back as the late 1980s, the FIC received little sedimentological attention, save for a general stratigraphic log and systematic sampling of the upper FIC for bulk carbon isotope analysis by G.M. Ross, an analysis of the petrography and geochemistry of a calcidebrite in the FIC (Gammon and Arnott, 2007) and a preliminary study of the stratal makeup of the FIC at Castle Creek and Mount Quanstrom (Navarro, 2016).

Ross et al. (1989, 1995) initially suggested that the FIC was deposited during basinward progradation of an active carbonate platform during a eustatic fall, however later work by Navarro (2016) proposed that the base of the FIC coincided with the beginning of a long-term (3rd order) eustatic rise that resulted in the initiation and development of a shallow-water carbonate platform and termination of siliciclastic supply into the local Windermere basin (Figure 2.2). Carbonate sediment was then resedimented downslope by a variety of particle settling and sediment-gravity flow processes – a set of conditions termed highstand shedding (e.g. Schlager et al., 1994), which in this case deposited the basal two calciturbidite horizons (CT 1 and CT 2). Superimposed on this trend were shorter duration (4th/5th-order) episodes of sea-level fall that deposited units dominated by amalgamated, carbonate-cemented, coarse-grained sandstone to granule conglomerate. Nevertheless, the long term eustatic rise eventually drowned the carbonate platform after deposition of CT 2 and it did not develop again until later eustatic fall re-established water depths in the carbonate production window (e.g. Droxler et al., 1993; Harper et al., 2015), which in this study is manifest as the uppermost calciturbidite unit (CT 3). Eventual lowstand conditions, and

development of a sequence boundary, are marked by the base of a > 100 m-thick siliciclastic leveed-channel complex (Isaac channel 1) that truncates the top of the FIC (Navarro, 2005). The > 1 km-thick section between the FIC and SIC is largely devoid of carbonate, except for some < 10 m-thick sections with abundant orange carbonate-cemented sandstones (Bergen, 2017) or as clasts in debris flow and/or channel deposits (Terlaky et al., 2016; Bergen and Arnott, 2016; Bergen, 2017). This suggests that the ecological conditions needed for the growth and maintenance of a robust carbonate platform, such as during the deposition of the FIC, did not redevelop until deposition of the SIC. Additionally, the sub-Cambrian unconformity completely removed all continental shelf and shallower marine strata southeast of the Cariboo Mountains (Ross, 1991).

II.III Study Areas

This study focused on the Castle Creek (CC; divided into Castle Creek south (CC-S) and north (CC-N)), Hill Section (HS) and Milk River (MR; divided into Milk River south (MR-S) and north (MR-N)) study areas in the Cariboo Mountains of the SCC (Figure 2.1C; Figure 2.3). Bedding at all sites is nearly-vertical ($\sim 75\text{-}89^\circ$) due to their location on the upright limb of two broad anticlines in the Isaac Synclinorium and Premier Anticlinorium (Campbell et al., 1973; Murphy, 1987a,b; Murphy et al., 1995; Reid et al., 2002). Rocks have been metamorphosed to lower greenschist facies, but primary sedimentary structures remain well preserved. Furthermore, all areas are situated near rapidly retreating, north-facing glaciers, and as a consequence rocks are glacially polished and free of vegetation, allowing for easy stratigraphic logging and beds to be traced laterally for hundreds of metres (Ross and Arnott, 2007). However, at MR outcrop quality is somewhat poorer compared to CC due to its slightly smaller area and slightly higher metamorphic grade, which generally obscures fine-scale primary sedimentary structures.

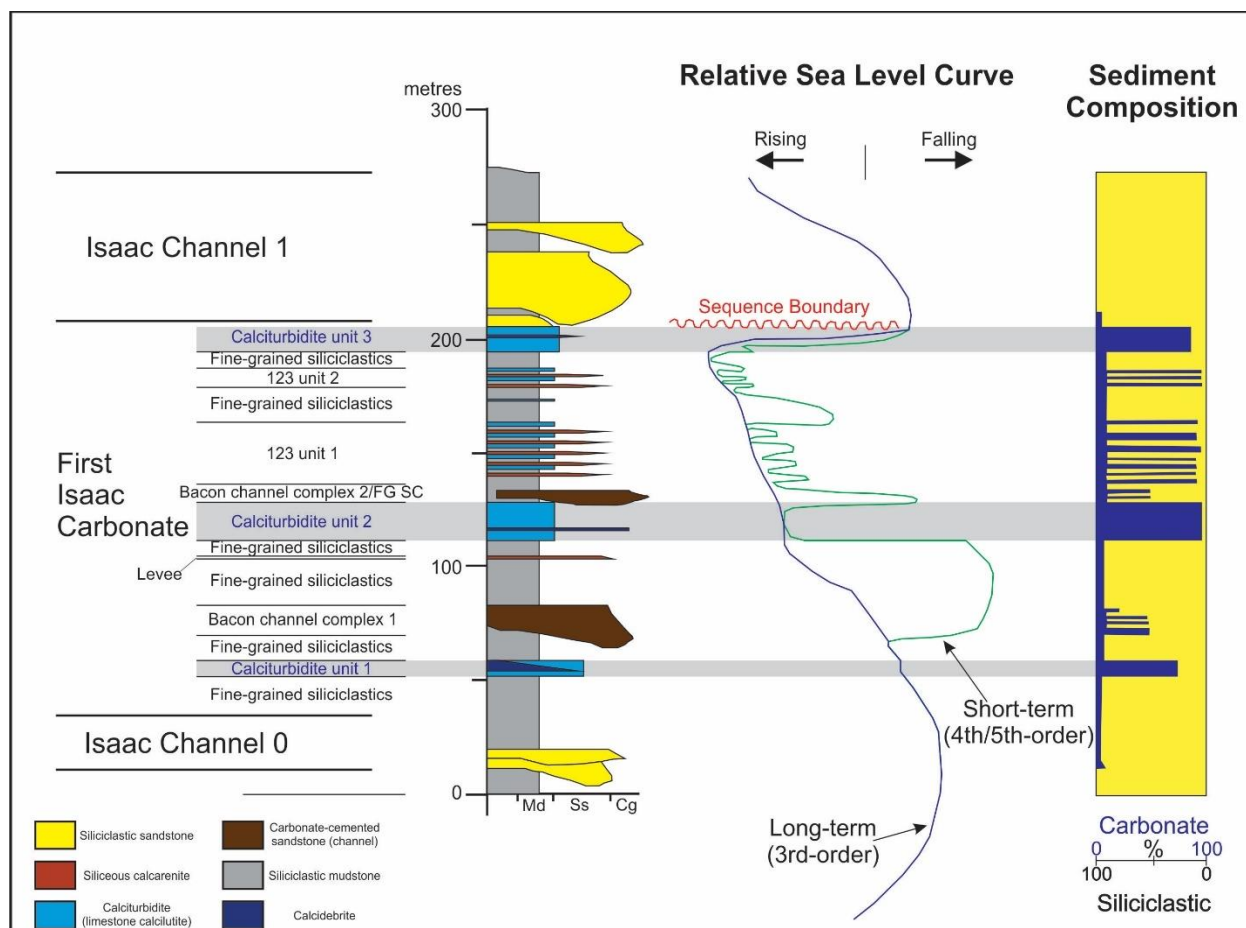
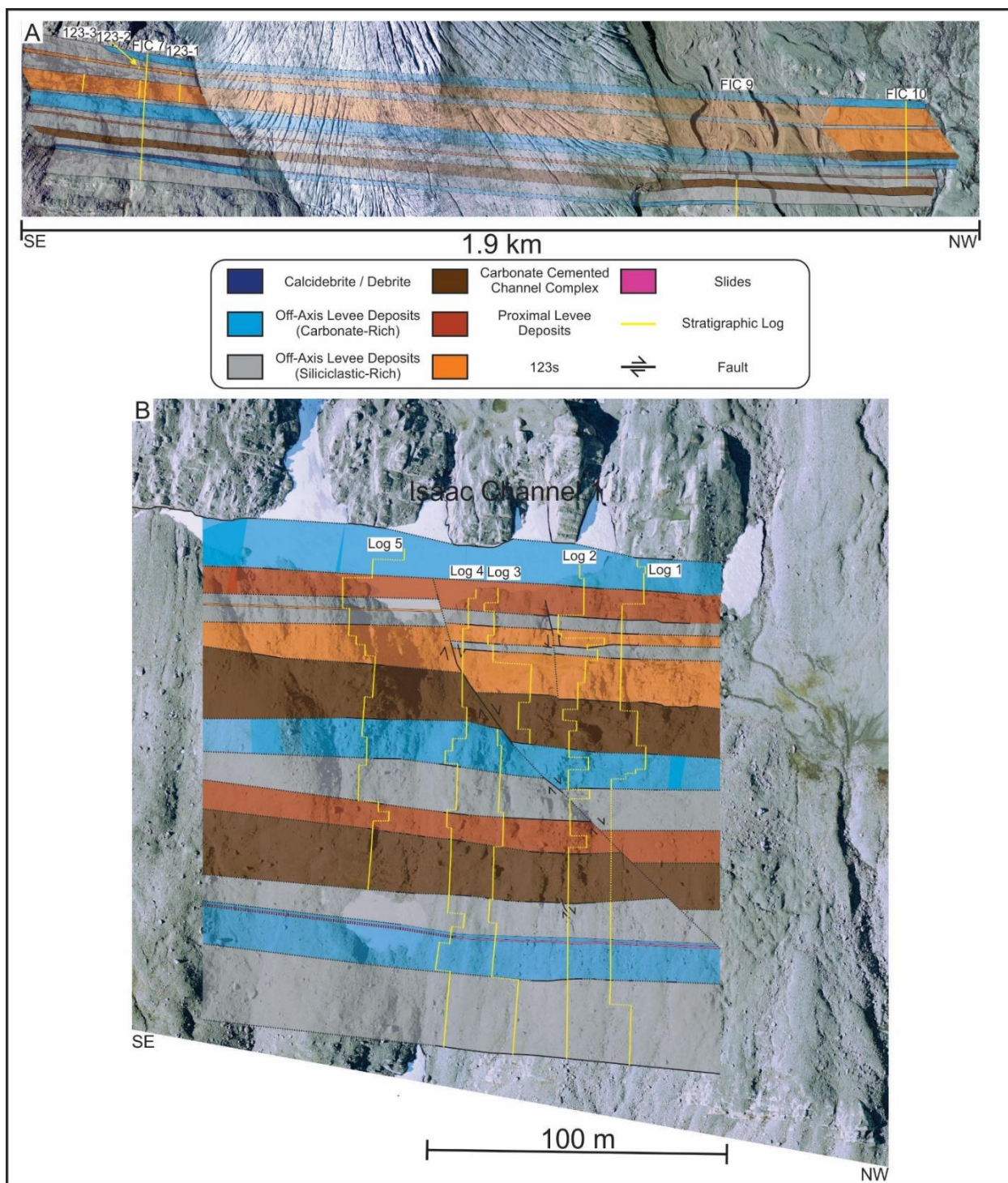


Figure 2.2: Generalized stratigraphic log through the FIC at CC showing short- (green) and long-term (blue) changes in sea level and composition of sediment supply into the deep-water Windermere basin. The FIC was initiated during a long-term sea-level rise (TST and HST) that was terminated by a major fall (sequence boundary) and initiation of Isaac channel 1 (modified from Navarro, 2016).

Part III Methodology

This study utilizes data collected during two field seasons (summer 2015 and 2016) at the CC, HS, and MR study areas. At the HS six sections were logged in mm-scale, bed-by-bed detail and range from 119 to 178 m in thickness (total thickness = 975 m) and spread over a strike length of ~ 225 m. At Milk River a ~ 260 m-thick section was logged in cm-scale, bed-by-bed detail at two locations (MR-S and MR-N) situated at about the same stratigraphic level. Stratal elements identified during logging, were then mapped and correlated using high resolution aerial (1:500) photographs (Figure 2.3). This study also incorporated the stratigraphic logs and stratal element maps of Navarro (2016) in the CC

study area. Logs were then digitized and units correlated across the study areas based on their lithological makeup (e.g. internal facies makeup, abundance of coarse- versus fine-grained strata, bedding contacts, and sedimentary structures) and stratigraphic position. Additionally, three shorter stratigraphic logs (33-51.5 m-thick) were measured through the 123 units at CC-S. In this section any sandstone bed thicker than 5 cm was traced laterally and re-logged every ~ 20 m, or where there was a significant change in thickness and/or sedimentary structures, until the bed either pinched out, became obscured by local tectonic deformation, or was covered by glacial debris. Seventy-one samples were collected and made into thin-sections for grain-size and mineralogical analysis. Most thin sections were stained using potassium ferricyanide and Alizarin Red-S solution (Dickson, 1966) for easier identification of carbonate occurrence and mineralogy. Visual estimates were made to approximate the percentage composition of mineral components in thin section.



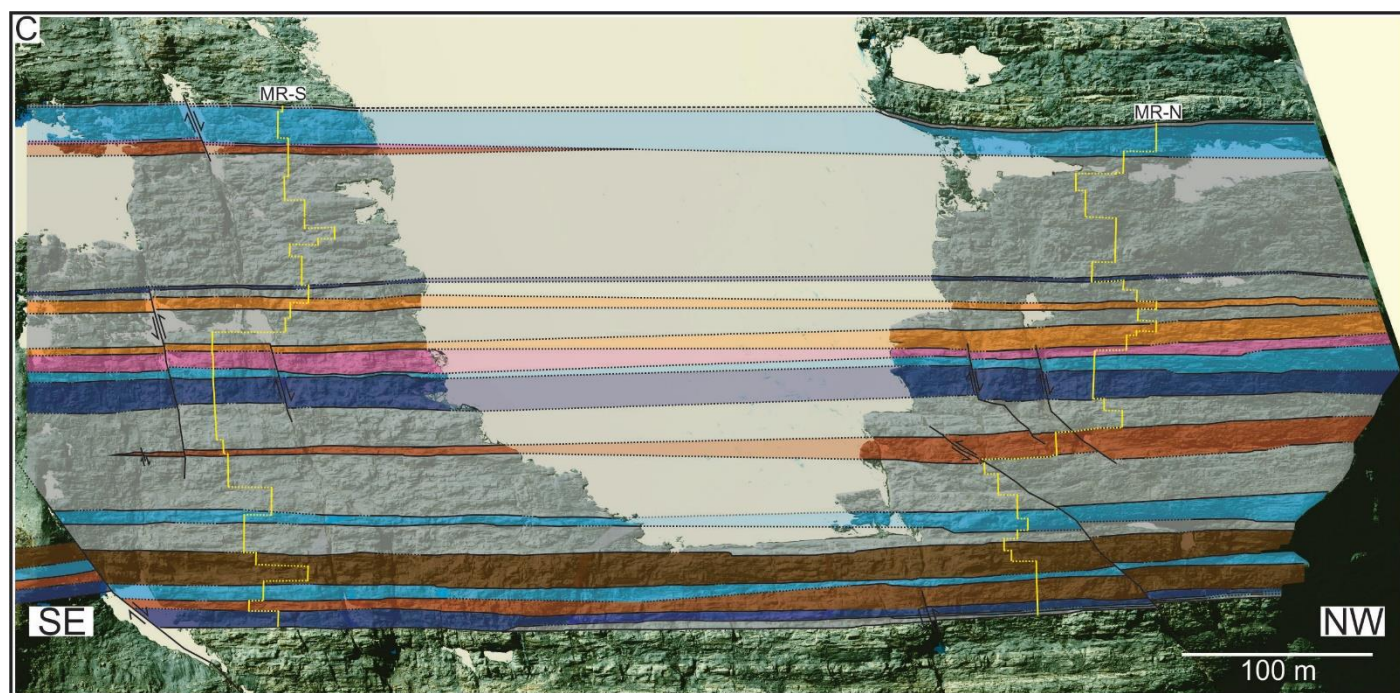


Figure 2.3: Aerial photographs of the study areas with major units correlated; measured stratigraphic logs indicated by yellow lines. (A) Castle Creek, (B) Hill Section, (C) Milk River.

Part IV Results

IV.I Facies

Rocks that make up the FIC comprise five facies: (1) Structureless, normally-graded sandstone, (2) Traction structured sandstone, (3) Fine-grained, mudstone-rich deposits, (4) Poorly-sorted, calcilutite-rich deposits, and (5) Internally sheared and deformed strata. Ingram's (1954) bed thickness classification is used in this study: very thin-bedded (1-3 cm), thin-bedded (3-10 cm), medium-bedded (10-30 cm), thick-bedded (30-100 cm) and very thick-bedded (> 100 cm). Lithological nomenclature for carbonate strata follows the convention of Grubau (1904), in which the adjective corresponds to the lesser mineralogical component and the prefix of the noun refers to the more abundant constituent. The noun refers to the bed's grain size, which utilizes the classification of Folk (1962): lutite (< 0.062 mm), arenite (0.062-1 mm) and rudite (> 1 mm) (Table 2.1). For example, a bed composed predominantly of mud-sized calcite with lesser quartz silt

grains is termed a siliceous calcilutite, whereas a sandstone bed consisting mostly of quartz sand grains with less abundant calcite is a calcareous silicarenite. Siliciclastic grain size classification follows the standard Wentworth scale (Wentworth, 1922), and turbidite classification follows the model of Bouma (1962).

Grain Size (mm)	Lesser Mineralogy	More abundant mineralogy	Adjective	Noun Prefix	Noun	Full name
< 0.062	None	Siliciclastic mud to silt quartz grains	None	Silica-	Lutite	Siliclutite
	Calcite	Siliciclastic mud to silt quartz grains	Calcareous	Silica-	Lutite	Calcareous siliclutite
	None	Calcite	None	Calc-	Lutite	Calcilutite
	Siliciclastic quartz grains/mud	Calcite	Siliceous	Calc-	Lutite	Siliceous calcilutite
0.062-1	None	Siliciclastic quartz sand	None	Silica-	Arenite	Silicarenite
	Calcite	Siliciclastic quartz sand	Calcareous	Silica-	Arenite	Calcareous silicarenite
	None	Calcite	None	Calc-	Arenite	Calcarenite
	Siliciclastic quartz sand	Calcite	Siliceous	Calc-	Arenite	Siliceous calcarenite
> 1 mm	None	Siliciclastic quartz sand, granules and pebbles	None	Silica-	Rudite	Silicrudite
	Calcite	Siliciclastic quartz sand, granules and pebbles	Calcareous	Silica-	Rudite	Calcareous silicrudite
	None	Calcite	None	Calc-	Rudite	Calcirudite
	Siliciclastic quartz sand, granules and pebbles	Calcite	Siliceous	Calc-	Rudite	Siliceous calcirudite

Table 2.1: Overview of rock classification scheme utilized in this study. Lithological nomenclature follows the convention of Grabau (1904) and grain-size classification follows Folk (1962).

IV.I.I Facies 1: Structureless, normally graded sandstone

Facies 1 makes up 27% of the FIC and can be subdivided into two sub-facies, 1A and 1B. Facies 1A is composed of medium- to very thick-bedded (23-240 cm), massive sandstones that are typically amalgamated and coarse-tail graded (T_a) (Figure 2.4A-D). Strata are predominantly dark- to light-brown, reddish-brown/-orange, brownish-grey and greenish-brown siliceous calcarenites (carbonate-cemented) interbedded with rare beige to grey silicarenites (siliciclastic-cemented). On average beds are 90 cm-thick and mudstone and calcirudite intraclasts, which are common at the base of beds, are oriented parallel to bedding. Where not amalgamated, beds fine upwards from medium-grained sandstone to pebble conglomerate at the base, although typically lower coarse-grained sandstone, to traction structured medium-grained sandstone (T_b and T_c) and/or a thin interlaminated siltstone (T_d) or structureless mudstone (T_e). The T_{de} is typically confined to the upper ~ 5-10% of the bed. Bed bases are commonly undulatory with rare flame structures.



Figure 2.4: Examples of Facies 1A (A-D) and 1B (E-F) from the FIC. (A) Granule conglomerate at the base of bacon channel complex 1 at HS with abundant dispersed granule quartz grains (white) and larger calcite (orange-brown (siliceous calcarenite) to yellow-orange (calcilutite)) and siliciclastic mudstone clasts (some indicated by white arrows) in a medium-grained calcarenite matrix. (B) Base of the bacon channel complex 1 at MR. Beige-coloured clasts are limestone (calcilutite) with blue arrows pointing to smaller calcilutite clasts, white arrows indicate siliciclastic mudstone clasts. Pencil for scale (circled in red). (C) Base of channel complex 3 at HS, consisting of a > 2 m-thick, coarse-grained massive orange-brown T_a siliceous calcarenite. White box indicates location of photo (D). (D) Basal portion of the basal, > 2 m-thick siliceous calcarenite bed in (C) with a large mudstone rip-up clast and a dark orange siliceous calcarenite clast (white arrow). (E) Complete Bouma sequence bed (Facies 1B) (F) Facies 1B bed consisting of a massive T_a overlain by planar stratified sandstone (T_b) overlain by a $T_{d/e}$ cap (i.e. the intervening T_c (ripple-cross stratified) unit is absent). White arrows indicate stratigraphic up direction.

In stained thin section dark-coloured, coarse-grained siliceous calcarenite to granule siliceous conglomerate consist of abundant coarse-grained sand to granule quartz and minor feldspar grains (43%; average grain size = 3081 μm , very fine gravel) surrounded by a calcite matrix (50%) with < 3% siliciclastic mud, dolomite and opaque minerals (likely pyrite). Grain boundaries commonly exhibit bulging recrystallization indicative of deformation temperatures of $\sim 280\text{-}400^\circ\text{C}$ (Hirth and Tullis, 1992; Lee, 2016). The calcite matrix consists predominantly of blue to dark-blue iron-rich microspar, interpreted to have resulted from the neomorphism of detrital coarse carbonate grains. These are then followed by cements formed during late burial diagenesis and metamorphism (Figure 2.5A-D) (for more on carbonate cements and on their proposed origin and timing go to section IV.II.I in Chapter 3) The coarse to granule quartz grains were likely delivered to the platform by rivers during periods of lowered sea level and subsequently resedimented downslope by turbidity currents. These flows also incorporated coarse detrital platformal carbonate grains that were hydraulically equivalent to the siliciclastic grains (e.g. Braga et al., 2001). However, after deposition the carbonate grains were recrystallized to iron-rich and iron-poor microspar during early diagenesis (e.g. Scholle and Ulmer-Scholle, 2003; van Der Kooij et al., 2010; James and Jones, 2016).

In contrast to the dark-coloured layers, lighter layers comprise massive, very coarse-grained calcareous silicarenite, composed predominantly of quartz (71%), siliciclastic mud/clay (12%) and significantly less calcite cement (14%), the latter occurring as small patches between quartz grains (Figure 2.5E-F).

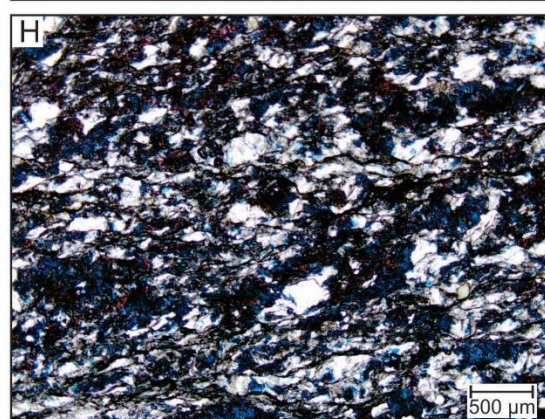
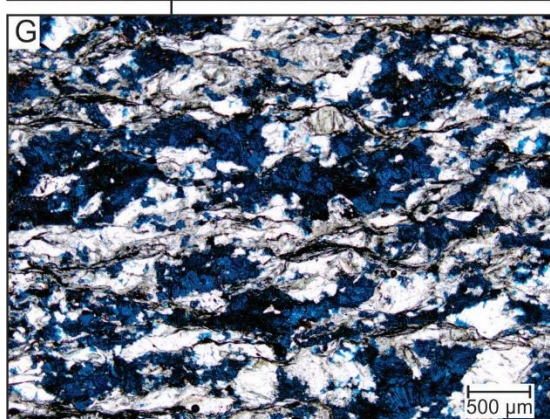
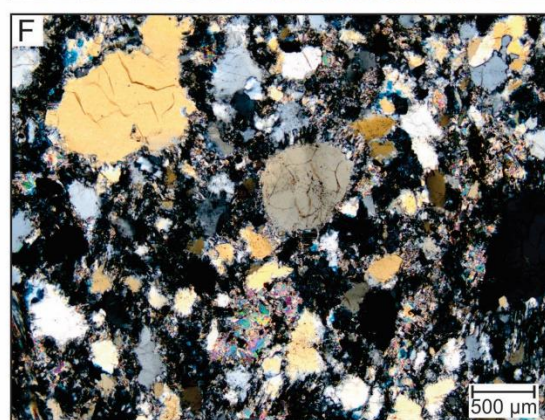
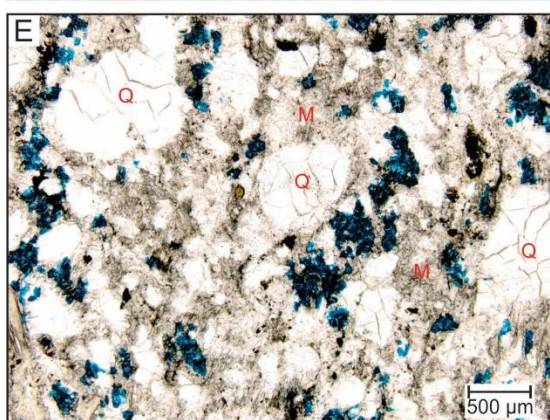
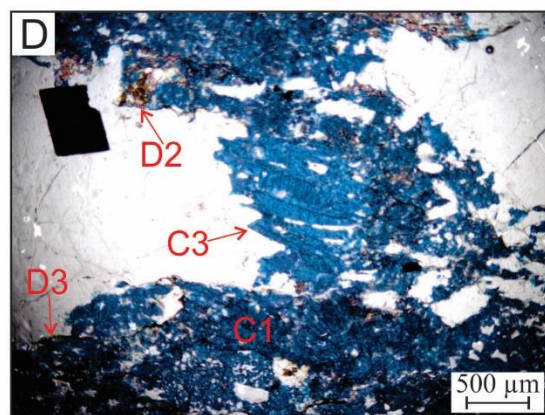
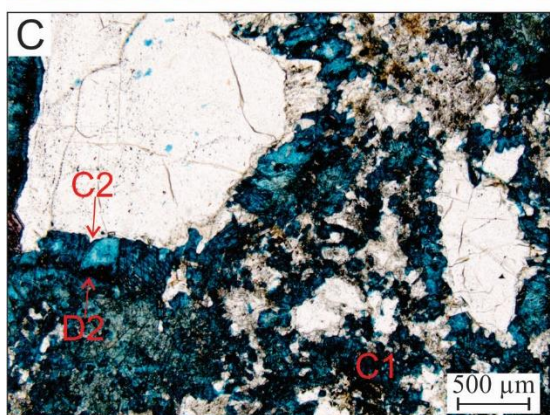
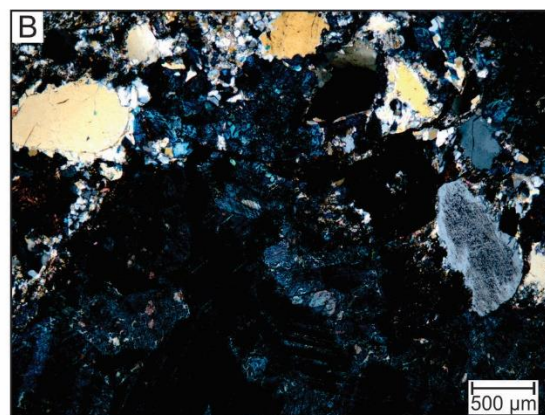
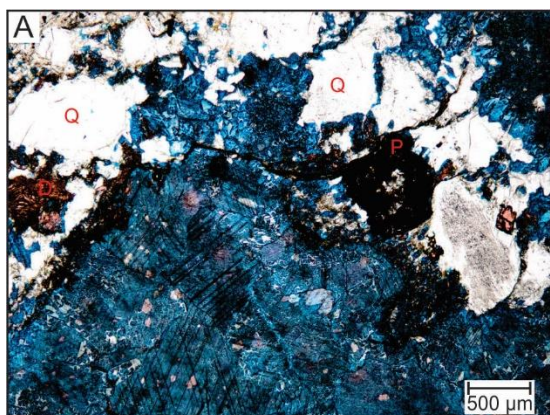


Figure 2.5 (previous page): Stained photomicrographs of Facies 1. (A) Coarse-grained massive sandstone (siliceous calcarenite) from MR. The sample is dominated by dark-blue iron-rich microspar with dispersed smaller, red iron-poor patches. These cements represent early burial diagenetic neomorphism of detrital coarse limestone grains (See Chapter 3: IV.II.I), with some of the original grains still displaying distinctive calcite twinning. White quartz grains (Q) appear to be “floating” in a matrix of this microspar. Brown patches along fractures and the edges of quartz grains are dolomitic (D) relating to late burial diagenesis/metamorphism. (B) Figure (A) under XPL. (C-D) Another dark-brown siliceous calcirudite bed. The primary neomorphosed cement is indicated by “C1”, whereas “C2” and “C3” are cements formed during late burial diagenesis and metamorphism (for more on cements see Chapter 3 IV.II.I). (E) Coarse-grained, light-brown sandstone layer (calcareous silicarenite) within a bacon channel complex. Iron-rich microspar cement is much less abundant (~ 10%) compared to (A) and the matrix is composed mostly of siliciclastic mud (M). Note the more abundant finer quartz sand grains (lower fine to lower medium sand) compared to the darker brown layers. (F) XPL of (C) that accentuates the more abundant fine to medium quartz sand grains. (G-H) Fine- to medium-grained sandstone (calcareous silicarenite) in which quartz grains are surrounded by a microspar to metaspar matrix. The metamorphic fabric is oriented E-W in both photos (thin irregular black lines). Photo F contains more dark-brown dolomite associated with iron-poor calcite formed during late burial diagenesis.

Facies 1B are very thin- to very thick-bedded, fine- to coarse-grained reddish-orange siliceous calcarenites/calcareous silicarenites with uncommon beige to yellow silicarenites. Beds are massive at their base (T_a) and grade upward to fine- to medium-grained traction-structured sandstone commonly overlain by a siltstone/mudstone cap (T_{b-e}) (Figure 2.4E-F). The average bed thickness is 23 cm and basal contacts are usually sharp and planar. The lower massive portion is typically < 50% of the total bed thickness with rare 5-10 cm-thick horizons of mudstone clasts oriented parallel to bedding.

In comparison to Facies 1A strata, the thinner-bedded, finer-grained siliceous calcarenites that make up Facies 1B consist of upper fine to medium quartz grains (35%) surrounded by microspar (58%; Fe-poor and Fe-rich) and lesser stringy dolomite (7%) cement (Figure 2.5G-H). In these samples the quartz grains exhibit less deformation along their boundaries than dark-brown Facies 1A beds, suggesting that Mesozoic deformational stresses were preferentially dissipated by the microspar matrix (e.g. Popović, 2016; Lee, 2016).

IV.I.I.I Interpretation

The coarse-grained, massive sandstones of facies 1A represented rapid suspension fallout from sand- to gravel-rich, high-density turbidity currents (Lowe, 1988; Arnott and Hand, 1989; Mulder and Alexander, 2001), or hyper-concentrated flows (Shanmugam, 1996; 1997). High sediment concentration and rapid deposition inhibited the development of tractional sedimentary structures (Arnott and Hand, 1989). Dispersed calcilutite and siliceous calcarenite clasts, in addition to detrital carbonate sand grains, the latter being recrystallized to cement, were eroded from a shallow-water carbonate platform and resedimented downslope. Mudstone clasts, on the other hand, were most probably sourced by upflow (upslope) erosion of the local muddy seabed. Undulatory basal contacts with common load and rare flame structures in F1A beds suggests high rates of sedimentation (Navarro, 2005). The transition from structureless to better-sorted, traction-structured sandstone (T_a to T_b and/or T_c) in facies 1B, and rarely in 1A, represents deposition in the tailward part of the same turbidity current that reworked previously deposited sediment (the underlying T_a part of the bed) (Arnott and Hand, 1989; Angus, 2016). The rarely preserved thin mudstone (T_e) caps consist of fine sediment that settled from the dilute tail of the flow (Lowe, 1982) or later hemipelagic fallout (Meyer, 2004; Angus, 2016).

IV.I.II Facies 2: Traction-structured sandstones

Making up ~ 13% of FIC stratigraphy Facies 2 is composed of reddish-orange to orange traction-structured siliceous calcarenite and uncommon beige silicarenite. Beds are very thin- to medium-bedded (average thickness = 12.4 cm) and typically fine upward from upper fine to lower medium sand topped by a thin siltstone/mudstone cap. Basal contacts are usually sharp and planar, and strata are either planar stratified overlain sharply by ripple-cross stratification (T_{bde}/T_{bcde}) or are simply ripple cross-stratified (T_{cde}) (Figure 2.6A). Ripple horizons consist mostly of 1-3 non-climbing sets (93%) with rare multisets (6%) or climbing ripple sets (1%) (Figure 2.6B-D). Additionally, four

medium-bedded, high-angle cross stratified sandstone sets are observed (Figure 2.6E-F). Two of which crop-out in the 123-1 unit at CC-S (see stratal elements) and persist for ~ 50 m laterally, before thinning to < 10 cm-thick planar-stratified sandstone.

IV.I.II.I Interpretation

Beds of facies 2 are interpreted to represent incomplete Bouma sequences deposited from waning sand-rich turbidity currents. The occurrence of traction-structured sand above sharp basal contacts suggests flows were immediately depositional with sediment concentrations and/or flow speeds conducive to the development of bed forms. Planar laminae indicate that conditions in the near-bed layer were unfavourable to the establishment and propagation of angular bedforms, possibly because flow speed and/or sediment concentration was initially too high to form the requisite bed defects from which ripples and dunes initiated (e.g. Southard and Boguchwal, 1990; Sumner, 2008; Arnott, 2012; Tilston et al., 2015). These defects change the pattern of fluid flow near the bed, ultimately causing flow separation, defect amplification and the spawning of new defects. In turbidites where the ripple cross-stratified unit (T_c) overlies a flat, planar bed (T_b) the defect must form spontaneously on a flat bed surface (Southard and Dingler, 1971; Venditti et al., 2005; Arnott, 2012). Experimental work by Venditti et al. (2005, 2006) showed that bed defects can develop spontaneously on a flat bed when sediment transport is widespread. The initiation of these defects was proposed to be related to a near-bed hydrodynamic (Kelvin-Helmholtz) instability developed along the interface between the slow moving, bed-load layer and overlying less dense, faster moving near-bed fluid. Shear along the interface eventually overcomes the stabilizing effect of stratification, which then causes the interface to adopt a waveform pattern and a spatially regular pattern of sediment erosion or deposition to develop on the bed that quickly builds the bed defects, and in turn ripples grow (Venditti et al., 2005, 2006; Arnott, 2012; Tilston et al., 2015).

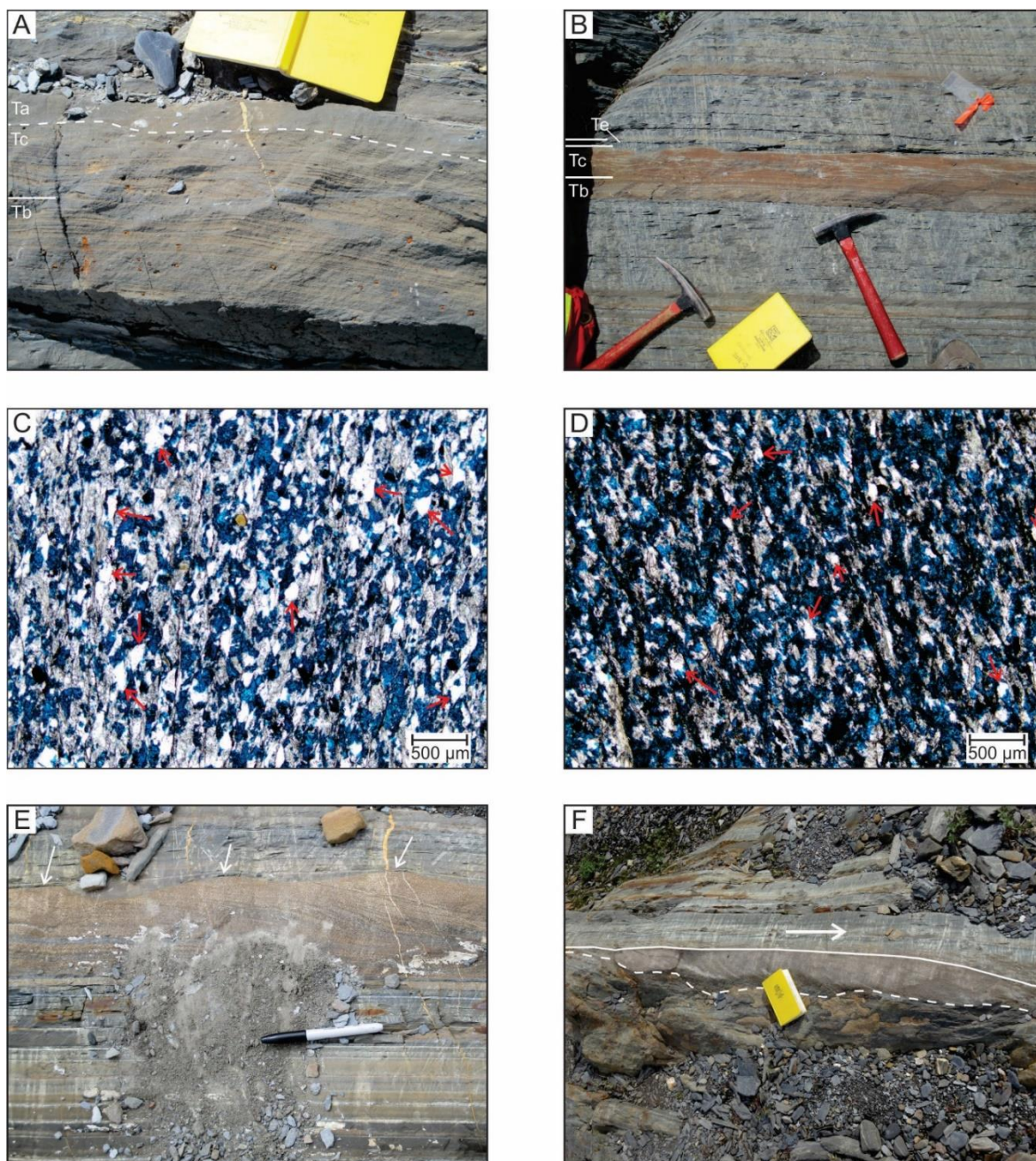


Figure 2.6: Field photographs of facies 2 strata. (A) T_{bc} bed – note the stacked ripple sets in the T_c part. (B) Reddish-orange T_{bcc} siliceous calcarenite with a light grey siliciclastic mudstone cap. Like in (A) the T_c unit contains multiple ripple sets. (C-D) Photomicrographs of Facies 2 strata. Like Facies 1A, Facies 2 strata consist of a pervasive (~ 65%) Fe-rich microspar matrix with lesser globular pyrite and dolomite and dispersed fine quartz sand grains (red arrows). (E) Siliceous calcarenite bed with a basal planar-stratified unit (T_b) overlain by a dune cross-stratified medium-grained siliceous calcarenite capped by ripple formsets along its upper surface (white arrows), in turn overlain by mudstone. (F) Pseudodune that thickens (due to erosion along its base) abruptly from left to right; arrow points to downflow end of scour. Dashed and solid white lines indicate the base and top of the bed, respectively.

In sediment coarser than middle fine sand below a decelerating open-channel flow planar-stratified sand (T_b) should be overlain by dune cross-stratified sand (e.g. Southard and Boguchwal, 1990). However, dunes are rare in the deep-marine, and more typically planar stratification is overlain by ripple cross-stratification (e.g. the T_b to T_c transition in a classical Bouma turbidite), the reason for which remains a source of much debate (Walker, 1965; Allen, 1969; Allen 1970; Lowe, 1988; Arnott, 2012). Recent work suggests that the near-bed density gradient required to create the requisite hydrodynamic instabilities that promote the inception of bed defects does not develop until flow speed is well within the ripple stability field, leading to a scarcity of dunes in the deep-marine (Arnott, 2012; Tilston et al., 2015).

The two high-angle cross-stratified beds in the 123s are examples of rare deep-marine dune cross-stratified sandstones. However, their pinch-and-swell nature and limited lateral extent (< 50 m) suggest that sediment supply was insufficient to develop a field of dunes, instead isolated, or starved, dunes formed.

The other high-angle cross-stratified sandstone beds are significantly less laterally extensive (< 1 m) and the angle of stratification declines downflow, eventually becoming planar stratified (Figure 2.6F). These are interpreted as rare base-of-slope examples of pseudo dunes, which although similar in character to the cross-stratification formed by migrating subaqueous dunes represent the migrating steep depositional front of isolated scours formed in an area of flow expansion akin to a hydraulic jump in open channel flow (Al-Mufti, 2013; Arnott and Al-Mufti, 2017).

Like in facies 1, the laminated siltstone and structureless mudstone caps that overlies the sandstone formed during the late stages of flow or post-flow hemipelagic fallout (Lowe, 1982; Meyer, 2004; Angus, 2016).

IV.I.III Facies 3: Mud-rich turbidites

This is the most common facies in the FIC (51% of stratigraphy) and comprises very thin- to thin-bedded siliceous calcilutite, siltstone and mudstone with occasional very thin siliceous calcarenite

(average thickness = 2 cm). Beds are typically structureless, but T_{de} , T_{ade} , T_{abde} , and T_{cde} (with isolated ripples) are observed. Siliciclastic mudstone is typically light to medium grey intercalated with rare thin dark-bluish and greenish black, organic-rich mudstone (Figure 2.7A). Siliceous calcilutites are composed of brownish-green “silty carbonate” and black to greenish-black “muddy carbonate” (Figure 2.7B). Facies 3 strata form uncommon interbeds in sandstone-rich strata, but more commonly stack to form decametre-thick stratal packages. In most cases these packages are composed of thin-bedded, fine-grained (mud) siliciclastic turbidites, but three, up to 21 m-thick packages composed principally (up to 74%) of siliceous calcilutite occur, which due to their distinctive mineralogy form recognizable marker horizons that can be correlated across the entire study area.

In thin section siliceous calcilutite beds are dominated by iron-poor microspar (80%) with dispersed 10-269 μm -thick (average = 76 μm ; very fine sand) quartz grains (3%) and euhedral pyrite crystals (Figure 2.7C). The very thin- to thin-bedded siliceous calcarenites in siliceous calcilutite-rich successions are similar to the siliceous calcilutites, except for more abundant (34%) upper fine sand to silt grains and less microspar matrix (56%), which makes them lighter coloured in plane-polarized microscopy (Figure 2.7D). Moreover, Facies 3 strata typically show little evidence of tectonic deformation in thin section.

IV.I.III.I Interpretation

Strata of Facies 3 were deposited by fine-grained, low-concentration, low-energy, turbidity currents. Very thin to thin beds of mud (T_e) may also be the result of hemipelagic fallout between successive flow events. Also, thin siliceous calcilutites may also represent sediments line-sourced from an active carbonate platform during episodes of elevated sea-level. These sediments were likely cemented shortly after deposition, which then probably limited subsequent alteration during late burial diagenesis and metamorphism. Thick accumulations of this facies are interpreted to have been deposited on the distal levees of active channel belts, whereas thinner packages in sand-rich

successions (e.g. channels) likely represent a period of quiescence between successive turbidity currents (e.g. Mulder et al., 2014).

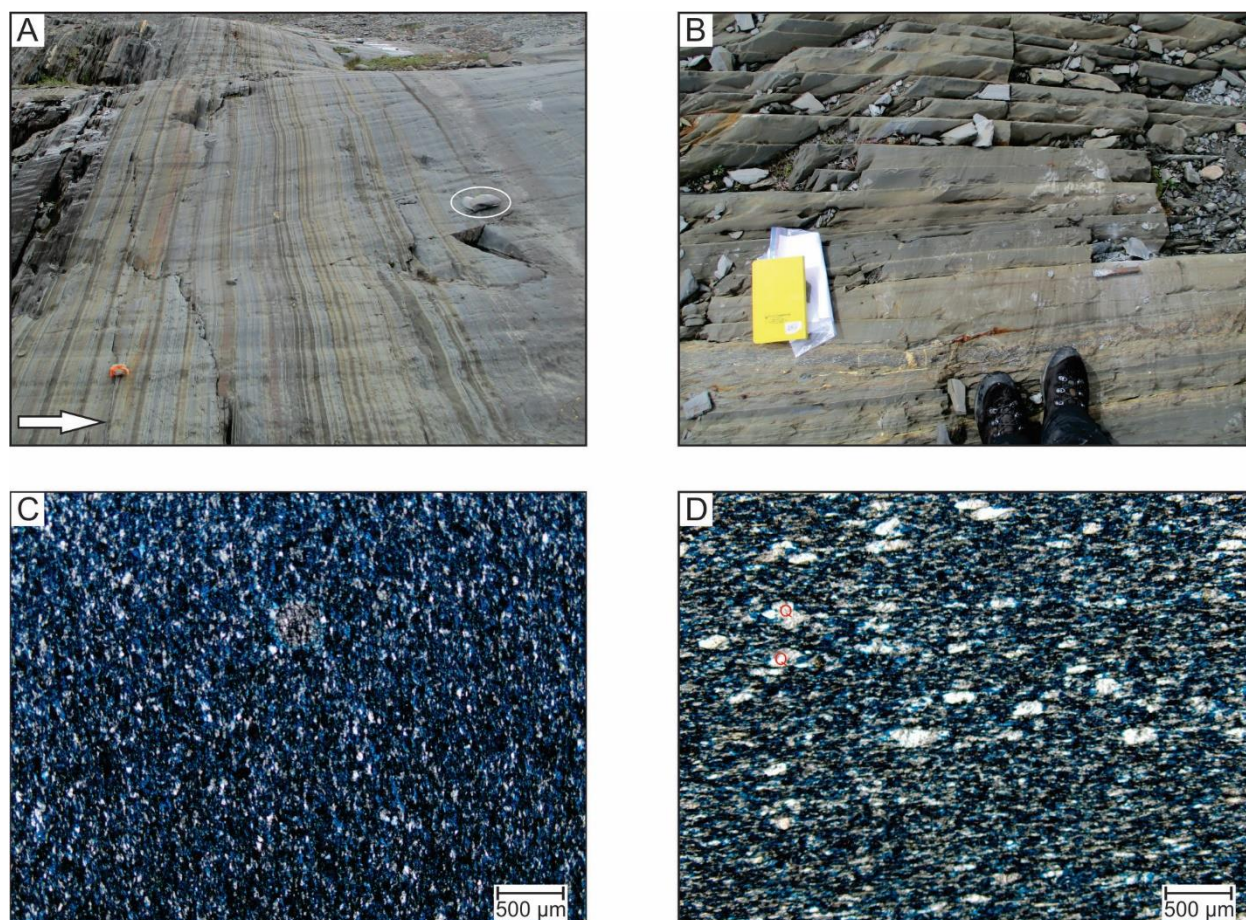


Figure 2.7: Photographs and stained photomicrographs of facies 3 strata. (A) Thin-bedded mud-rich turbidite section consisting of alternating siliceous calcilutite (black, olive-green, brownish-green) and siliciclastic mudstone (grey). Baseball cap for scale (circled in white). White arrow indicates stratigraphic up direction. (B) Part of a calciturbidite-rich section at CC. (C) Photomicrograph of a siliceous calcilutite dominated by iron-rich microspar matrix (dark blue) and lesser iron-poor microspar (grey/dirty grey) and grains of quartz silt. Small black dots are dolomite crystals that formed shortly after deposition. (D) Photomicrograph of a fine-grained siliceous calcarenite. These beds contain more silt to very fine quartz sand grains and dispersed upper fine to medium quartz sand grains than in C, resulting in a noticeably lighter colour in thin section.

IV.I.IV Facies 4: Poorly sorted, calcilutite-rich deposits

Facies 4 makes up ~ 6% of FIC stratigraphy and is characterized by beds with abundant clasts displaying various degrees of deformation and dispersed in a matrix of greenish-brown, reddish-brown, brown, dark grey or black calcilutite or rare siliciclastic mud matrix (Figure 2.8). Beds are typically

lenticular with sharp irregular basal and top contacts that can be traced to the limit of the outcrop (at least ~ 600 m) and typically occur in close stratigraphic association with calciturbidite horizons. Clasts commonly range from mm to m-scale and consist of mudstone fragments, granule- to pebble-sized quartz grains, yellow calcilutites (with rare microbialite laminations (Figure 2.8E)), yellowish-orange oolites (Figure 2.8F), green to black siliceous calcilutites and reddish-orange siliceous calcarenites (some of which are planar stratified). At MR one facies 4 bed's matrix consists of calcilutite that in its upper part contains a ~ 3 m-thick thin-bedded, mud-rich turbidite clast that extends laterally for ~ 100 m and abruptly pinches out laterally in both directions.

IV.IV.I Interpretation

Due to the lack of internal bedding and grading, the sharp and irregular nature of their upper and lower contacts, and the presence of large clasts dispersed in a fine-grained matrix, strata of Facies 4 are interpreted as having been deposited by carbonate-rich cohesive gravity flows termed debris flows (Mulder and Alexander, 2001), hereafter termed calcidebrites (e.g. Navarro, 2016). Based on the abundance of carbonate clasts, especially those containing photic zone growth forms and oolites (Figure 2.8E-F), it is likely that sediment was resedimented from a shallow-water carbonate platform. Clasts, which in some cases ranged up to > 100 m-wide were supported by the cohesive strength and buoyancy provided by the fluid carbonate mud matrix ($\pm$ siliciclastic mud), which because of its high density and low permeability limited fluid dilution during movement (Mohrig et al., 1998; Mulder and Alexander, 2001; Arnott, 2010). Cohesive strength inhibited particle settling and promoted downslope transport. Deposition, however, occurred en masse when the shear resistance of the flow (viscosity and friction) exceeded the gravity (driving) force. Also, the thickness of the deposit (prior to post-depositional compaction) approximated the thickness of the parent flow. (Iverson, 1997; Mulder and Alexander, 2001).



Figure 2.8 (previous page): Field photographs of Facies 4 (debrites) in the FIC. (A) Calcidebrite 1 at MR-S. Note the large yellow limestone clast and many smaller limestone (yellow (calclutite) to orange (calcarenite)) and mudstone (dark grey to black) clasts dispersed throughout the dark green calclutite matrix. Rock hammer circled in red for scale. (B) Close-up of a clast-rich part of the debrite shown in (A). Clasts are composed mostly of limestone calclutite (yellow to dark yellow) or mudstone (grey to black) with lesser calcarenite (orange). (C) Calcidebrite 2 (CD 2) at MR-N with its basal and top contacts indicated by the white lines. In the field this debrite is identified by its lack of bedding, darker colour compared to the mud-rich turbidites below, and randomly dispersed carbonate clasts. (D) Close-up of CD2. Debrite has a weathered brown siliceous calclutite matrix with dispersed calclutite (red arrows) and siliceous calcarenite (yellow arrow) clasts and one large mudstone clast (outlined in white). (E) Yellow, microbially-laminated carbonate cobble at the top of CD2 (MR-N). Head of rock hammer for scale (outlined in white). (F) A yellow-orange oolite clast (white arrow) in CD2 at MR-N. Camera case is 12 cm across. (G) Debrite stratigraphically above CD2 at MR-N; upper and lower contacts are indicated by white lines. The basal reddish-orange siliceous calcarenite clast (white arrow) is planar stratified and oriented at an angle to bedding. Many smaller orange siliceous calcarenite and siliceous calclutite clasts are spread throughout the dark grey-to-brown calclutite matrix. (H) Calcidebrite 3 at CC-S. This thin (~ 65 cm) calcidebrite is characterized by a brown carbonate matrix with abundant white quartz pebbles and numerous mudstone clasts oriented parallel to bedding. White arrows indicate stratigraphic up direction.

IV.I.V **Facies 5: Internally sheared and deformed strata**

The FIC is host to three examples of Facies 5. Beds range from 61-794 cm-thick and like calcidebrites occur in close stratigraphic association with calciturbidite horizons. One bed (S1) is located near the base of the HS in calciturbidite 1 (Figure 2.10) and consists of a ~ 60 cm-thick, internally sheared, dark grey siliciclastic mudstone, underlain by a ~ 45 cm-thick dark brown siliceous calcarenite bed that irregularly thickens and thins laterally (Figure 2.9A). Despite poor exposure this internally sheared mudstone can be traced laterally over > 200 m and at the northernmost HS overlies a > 3 m-thick calcidebrite. Another bed (S2) is up to ~ 8 m-thick and extends across the entire MR outcrop (~ 600 m). The basal part of the bed contains abundant black, lenticular calcidebrite blocks with irregular basal contacts. These debrites abruptly pinch laterally into blocks of thin-bedded, upper-division turbidites (siliceous calclutite- and siliciclastic mud-rich) that are internally deformed (i.e. folding of turbidites within) and are oriented oblique to the regional strike of the FIC (130°) (Figure 2.9B-D). The stratigraphically highest example of Facies 5 (S3) is a 60 cm-thick bed that directly underlies CT 3. It has a greenish-brown siliceous calclutite matrix and the basal half of the bed contains dispersed up to 15 cm-wide (commonly < 1 cm) limestone, siliceous calcarenite and

siliciclastic mudstone clasts. Its upper half contains numerous deformed siliciclastic mud layers and reddish-orange siliceous calcarenites that pinch and swell laterally. This bed can be traced to the most southeastern extent of MR, but in the opposite direction thins to 30 cm-thick moving near the Milk River glacier, and cannot be identified in MR-N, suggesting that it probably pinches out beneath the glacier.

IV.I.V.I Interpretation

The presence of numerous internally deformed blocks oriented at a shallow angle to regional strike, in addition to the irregular upper and basal contacts suggests that strata of Facies 5 represent slides related to the failure of the upslope seafloor and displacement along discrete internal failure planes during movement (Arnott, 2010).

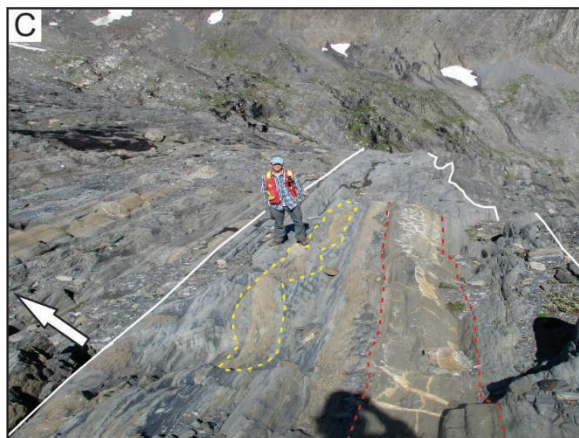
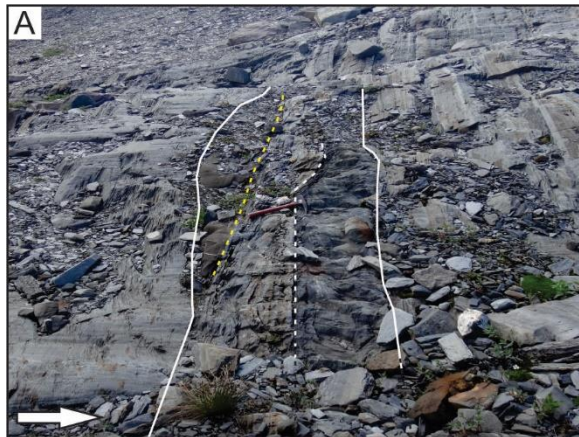


Figure 2.9 (previous page): Outcrop examples of facies 5 strata (slides). (A) Slide sharply bounding the top of calciturbidite 1 at HS. At its base is a dark-brown siliceous calcarenite (base is the leftmost solid white line and top is yellow dashed line) that can be traced across the full width of the outcrop (~ 230 m). Overlying is an internally deformed mudstone (base indicated by dashed white line, top is rightmost solid white line) that can only be traced laterally for ~ 50 m. (B) Base of an ~ 8 m-thick slide at MR-S. Here the base of the slide consists of an ~ 3 m-thick black debrite block that scours the underlying calciturbidites and transitions laterally way to smaller and irregularly-shaped debrite and thin-bedded turbidite (carbonate- and siliciclastic-rich) blocks. Solid white line represents the base of the slide and dashed white line represents the top of the basal debrite block (C) Same slide as (B), but at MR-N; basal and upper contacts indicated by white lines. Yellow and red dashed lines delineate a deformed sandstone and black calcidebrite block. (D) Same as (C), but from the edge of the debrite block outlined in (C). Here at the stratigraphic level of the debrite block is at least three internally deformed thin-bedded, siliciclastic mud-rich turbidite blocks (outlined by black dashed lines), each at an angle to the strike of the underlying calciturbidite beds (black beds). White arrows indicate stratigraphic up direction.

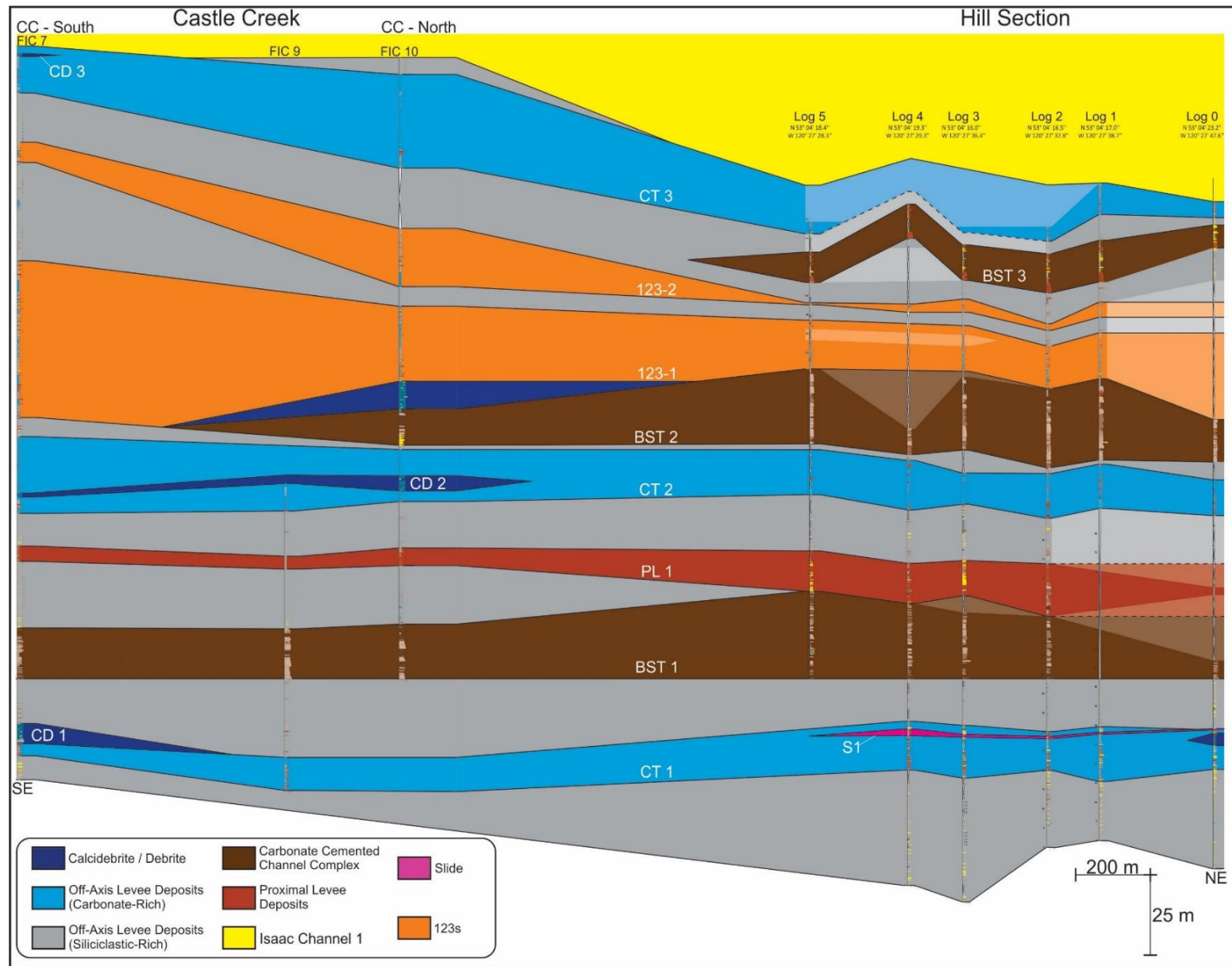


Figure 2.10 (previous page): Correlation of stratal elements identified in CC and HS. The six stratigraphic sections from HS were measured as part of this study whereas the three from CC are from Navarro (2016). White lettering indicates unit names (see text for details).

IV.II Stratal Elements

An architectural element is a distinct three-dimensional sedimentary body bounded above and below by genetically-related surfaces that formed in the same depositional setting (Pickering et al., 1989), whereas a stratal element is its two-dimensional expression in outcrop (Terlaky et al., 2015). Strata described here, therefore, are appropriately termed stratal elements and based on geometry, scale and internal lithological composition form seven distinct elements in the FIC, including: channel complexes, proximal levees, distal levees - further subdivided into siliciclastic- and carbonate-rich, “123” units, scour dominated units, debrites and slides (Table 2.2; Figure 2.10; Figure 2.11). Each is described next.

Stratal Elements	Total Thickness in FIC (m)	Percentage of total FIC thickness (%)	Facies 1 (%)	Facies 2 (%)	Facies 3 (%)	Facies 4 (%)	Facies 5 (%)
Channel Complexes	126.7	11.9	86.5	5.9	7.0	0.0	0.0
Proximal Levees	90.5	8.5	45.7	29.3	25.0	0.0	0.0
Distal Levees							
a) Siliciclastic-rich	487.2	45.8	16.6	15.5	67.5	0.0	0.0
b) Carbonate-rich	145.7	13.7	23.2	14.3	62.3	0.2	0.0
Scour Dominated Units	8.5	0.8	82.6	0.0	17.4	0.0	0.0
"123s"	130.3	12.2	16.5	7.3	76.1	0.0	0.0
Debrites	61.9	5.8	0.0	0.0	0.0	100.0	0.0
Slides	13.3	1.3	0.0	0.0	0.0	0.0	100.0
Total			27.3	12.9	50.8	5.8	1.3

Table 2.2: Stratal elements in the FIC and the proportional make-up of their constituent facies.

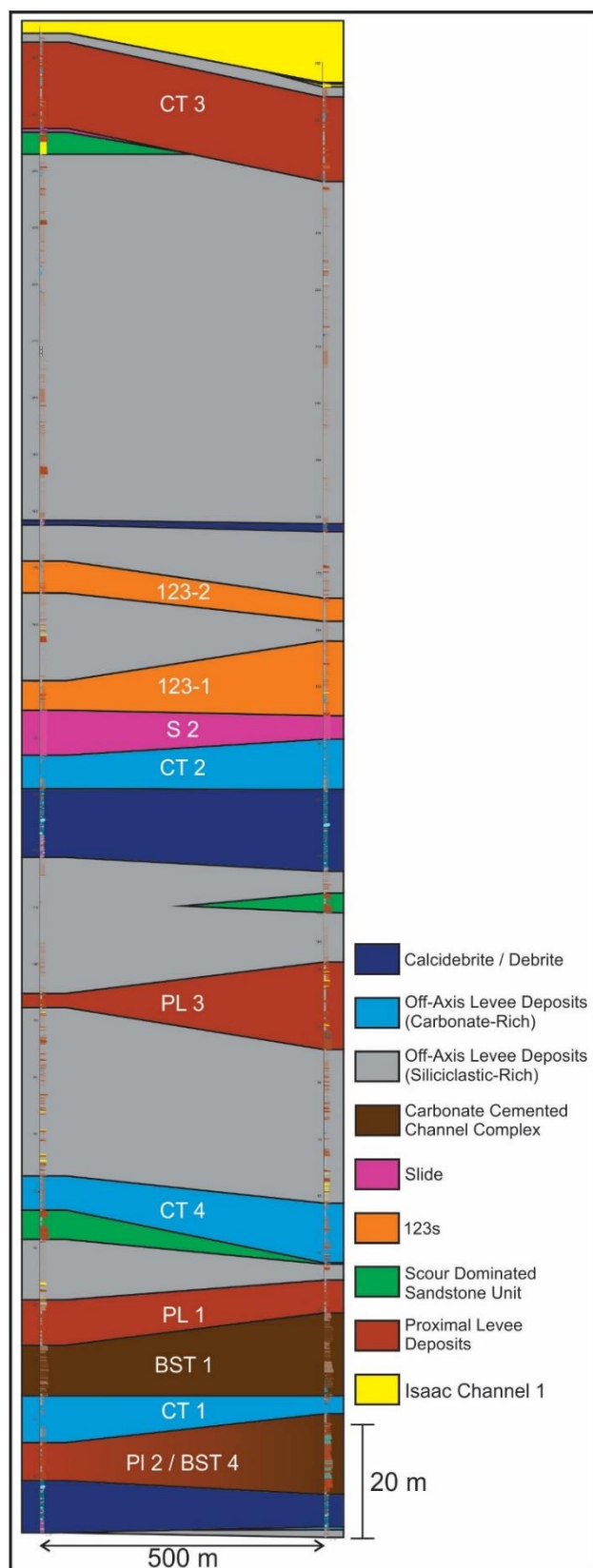


Figure 2.11: Correlation of stratal elements identified at MR. Both stratigraphic sections were measured as part of this study. White lettering indicates unit names (see text for details).

IV.II.I Channel Complexes

IV.II.I.I Description

Channel complexes comprise five Dm-thick units dominated by amalgamated, medium- to very thick-bedded (15-351 cm-thick), medium- to very coarse-grained massive siliceous calcarenite/silicarenite to granule siliceous calcirudite (conglomerate) (T_a; facies 1, 87%) with rare, < 5 cm-thick, mudstone/siltstone caps (T_{de}) (Figure 2.12A-B). The base of each unit scours underlying stratigraphy and is overlain by granule conglomerate/siliceous calcirudite (Figure 2.12C) with abundant dispersed very coarse sand to pebble quartz grains and intraclasts of mudstone, orange siliceous calcarenite and yellow calcilutite (Figure 2.4A-D), some with irregular planar laminae suggestive of microbial activity (e.g. microbialites). Stratigraphically upwards, mudstone and carbonate clasts become progressively less common and sandstone beds generally fine and thin. Additionally, traction-structured sandstone intervals thicken, and similarly their mudstone cap (facies 1 and 2, 6%). These facies stack to form 4-20 m-thick channel complexes comprising 1-3 up to 6 m-thick channel fills (Figure 2.12A; e.g. Clark and Pickering, 1996; Campion et al., 2000; Navarro et al., 2007a,b; Schwarz and Arnott, 2007a; Dumouchel, 2015). Individual channel fills can be traced laterally for 100s of metres in CC and MR with bases marked by an abrupt increase in grain size, bed thickness and abundance of intraclasts. The thickness and lateral extent of individual channel fills is towards the lower end of those observed in carbonate and/or mixed slope deposits worldwide (both ancient and modern), which average 18 m-thick and ~ 1 km wide (George et al., 1997; van Konijnenburg, 1999; Braga et al., 2001; Vigorito et al., 2005, 2006; Payros et al., 2007; Payros and Pujalte, 2008; Mudler et al., 2012a, 2014). However, these dimensions are comparable to channels along the modern Great Bahama Bank (Mulder et al., 2014) and ancient deposits in the Napier Formation of western Australia (George et al., 1997) – both are up to 6 m-thick, and up to 100 m-wide in the ancient example and 500 m-wide in the modern.

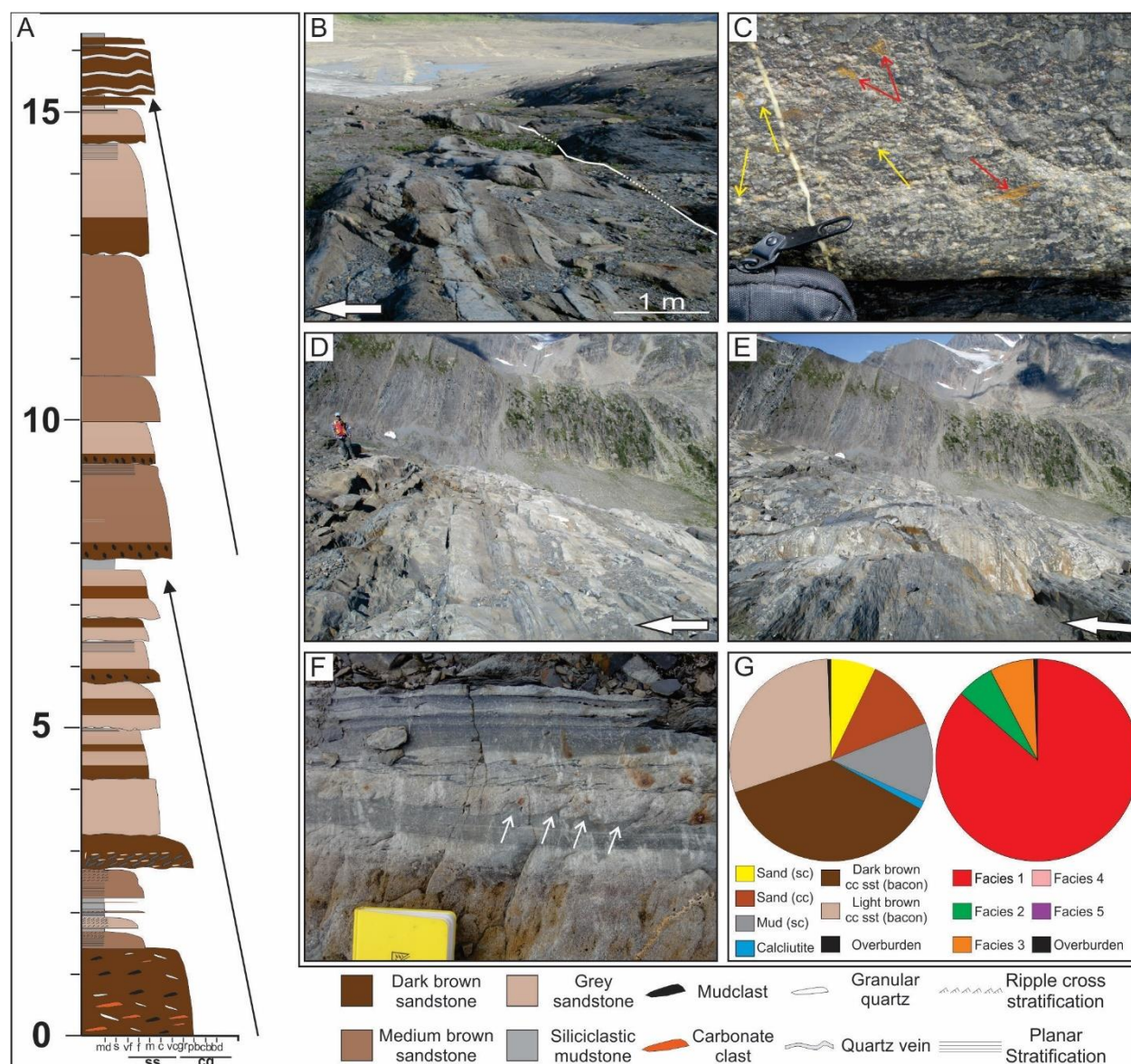


Figure 2.12: (A) Stratigraphic section through BST 1 at HS. The base of the complex consists of very thick-bedded granule siliceous calcirudite (conglomerate) with abundant siliciclastic mudstone and carbonate clasts that then is overlain by an upward fining and thinning succession. The BST 1 at HS comprises two upward fining units interpreted to be the remnants of at least two discrete channel fills. (B-F) Representative photographs of channel deposits in the FIC. (B) Base of BST 1 at CC-S displaying the distinctive light and dark layering of siliceous calcarenite- and calcirudite-rich FIC channel deposits. The white line outlines the sharp, erosive basal contact. (C) Close-up of the basal bed of BST 1 at MR-S. This dark siliceous calcarenite contains numerous quartz grains (up to upper very coarse sand; indicated by yellow arrows) and some orange siliceous calcarenite clasts/grains (red arrows). (D) BST 4 at MR-N, which is a 14 m-thick succession of mostly amalgamated massive brown siliceous calcarenite and grey sandstone. (E) Strata of BST 4 about 100 m SE of (D). Here the succession consists of thin-bedded, siliciclastic mud-rich turbidites with occasional thin- to medium-bedded sandstone. (F) Sandstone bed in BST 3 at HS with evidence of syn-depositional deformation along its basal contact (flame structures, white arrows). (G) Lithological (left) and facies (right) make-up of FIC channel complexes. White arrows indicate stratigraphic up direction.

Channel complexes in the FIC exhibit a distinctive alternation of dark-brown and light-brown to grey layering in outcrop (Figure 2.12B), leading them to be informally termed “bacon sandstones” (BST). This striping is a consequence of varying amounts of microspar cement (Figure 2.5), with the brown layers containing 47% on average (range of 35-65%; calcareous silicarenite to siliceous calcarenite) and lighter layers ~ 10% (siliceous calcarenite). Notably layering does not necessarily parallel bedding and often beds consist of multiple light and dark layers with no apparent change in grain size between successive layers. Dark layers exhibit the greatest number of calcite cement phases, whereas light layers contain only Fe-rich micritic calcite and it is only a minor constituent with the matrix being primarily composed of fine quartz grains and recrystallized siliciclastic mud. The isotopic composition of (inorganic) carbonate carbon ($\delta^{13}\text{C}_{\text{carb}}$) in primary microspar cement in bacon channel sandstones is typically ~ 0‰, which differs from most other similarly aged deep-marine carbonate rocks (e.g. Jiang et al., 2007; Giddings and Wallace, 2009; Smith, 2009) and/or those composed of microbially mediated carbonate mineral precipitates (e.g. Schrag et al., 2013; MacDonald et al., 2013; Al-Mufti, 2013) (see chapter 3).

Overall, the FIC contains five channel complexes, which make up 12% of the stratigraphy. They are most abundant at HS, where three complexes make up 30% of the stratigraphy, and are rarest at MR-S (5%). Only one complex, BST 1, crops out in all the study areas (see below). Due to its areal extent and consistent thickness (~ 15 m) the base of BST 1 was used as the stratigraphic datum in this study. Overlying calciturbidite 2 in CC-N and HS is BST 2, which is > 20 m-thick at HS and comprises two upward-fining and -thinning channel fills, whereas at CC-N it is ~ 9 m-thick and overlain abruptly by a 7.2 m-thick calcidebrite. Notably, BST 2 is absent in CC-S, where calciturbidite 2 is overlain by a thin siliciclastic mud-rich succession and 123-1.

Near the top of the FIC at HS is a 6.0-10.9 m-thick channel that is bounded above and below by siliciclastic mud-rich turbidites. The basal ~ 2 m-thick part of the channel fill consists of a massive, brown-coloured, lower coarse- to upper very coarse-grained siliceous calcarenite bed with abundant

mudstone and rare orange siliceous calcarenite clasts near its base, and can be confidently traced across the entire HS outcrop. Its basal contact is irregular with local flame structures and scours underlying mud-rich thin-bedded turbidites with local flame structures (Figure 2.12F). In thin section this bed appears similar to dark brown siliceous calcarenites from other BST complexes with coarse to very coarse quartz grains surrounded by a fine grained micritic calcite matrix (Fe-poor and Fe-rich) with minor late stage cements (< 5%). However, strata that overlie this calcarenite are thinner (thin- to medium-bedded with only rare thick beds), finer (upper medium grained and finer), and contain thicker siliciclastic mud-rich interbeds (T_{de}/T_e ; > 50%) and more yellow silicarenites than those that overlie the basal beds in BST 1 and 2 (Figure 2.11A,B). An exception is the most northern exposure where most sandstone beds are erosively based and medium- to thick-bedded with some containing abundant mudstone intraclasts.

At Milk River, thick channel complexes are notably absent above BST 1. Below BST 1, and only at MR-N, is an ~ 14 m-thick complex (BST 4) that crops out between calcidebrite 1 and calciturbidite 1. BST 4's channel fill is ~ 11 m-thick with a ~ 9 m-thick interval comprised of 8 thick- to very thick-bedded (52-266 cm; average = 133), massive, coarse- to very coarse-grained siliceous calcarenites (99:1 sandstone:mudstone ratio), overlain by ~ 2 m of medium- to coarse-grained T_{abde}/T_{bde} beds with up to 20 cm-thick siltstone/mudstone caps (75:25 sand:mud) (Figure 2.12D). Like other FIC channel fills, the sandstones exhibit alternating light- and medium-brown layers (carbonate cement-rich) and greenish grey layers (carbonate-cement poor), sometimes in a single bed. However, along the same stratigraphic horizon at MR-S (~ 600 m south) strata are composed mostly of thick- to very thick-bedded T_{abde} , T_{abcde} and T_{bcde} turbidites with thick T_b intervals and multiple ripple-set T_c parts, intercalated with minor thin-bedded, upper division turbidites (65:35 sandstone:mudstone), which suggests deposition in a levee adjacent to the channel. Interestingly, between these two sections strata consist of well-stratified, upper-division, siliciclastic mud-rich turbidites (35:65 sandstone:mudstone) (Figure 2.12E). These dramatic differences in lithology suggest dramatic differences in depositional

conditions, but occurring adjacent to one another along a single stratigraphic horizon (See IV.II.II Proximal Levees below)

IV.II.I.II Interpretation

The irregular, erosive basal contact of channel complexes represents incision by high-energy, high concentration, carbonate- and siliciclastic-rich turbidity currents. Following an initial period of bypass, an amalgamated succession of massive, thick- to very thick-bedded (31-351 cm, average = 105), coarse- to very coarse-grained siliceous calcarenites and siliceous calcirudites, with abundant mudstone and carbonate intraclasts were deposited in the axial part of the channel system (90:10 sandstone to mudstone ratio). With time the magnitude of the transiting flows decreased which is manifest as an upward decrease in bed thickness and grain size and deposition of more complete Bouma turbidites and thicker successions of mud-rich turbidites. The abrupt superposition of amalgamated coarse-grained sandstone represents the base of a younger channel fill (Figure 2.12A). Additionally, BST 2 at the HS comprises two up to 10 m-thick, upward-fining and -thinning sequences, whereas at the same stratigraphic level in CC-N BST 2 is ~ 8 m thick and overlain abruptly by < 1 m of thin-bedded calciturbidites capped by a ~ 7 m-thick calcidebrite that likely plugged the channel at CC-N. However, the channel remained open at the HS and deposited the upper channel fill (e.g. Samuel et al., 2003). In addition to thinning- and fining-upward the channel-filling siliceous calcarenite and siliceous calcirudite (conglomerate) beds also thin laterally (spatially) from amalgamated T_a sandstone to beds that increasingly contain traction sedimentary structures in their upper part (i.e. b and c divisions) and become increasingly interbedded with mud-rich turbidites (T_e or T_{ae} turbidites with very thin-bedded siliceous calcarenites) (50:50 sandstone:mudstone ratio).

As described earlier, the light and dark layering of medium- to coarse-grained sandstone and granule conglomerate layers reflect differences in the mineralogical and textural makeup of the strata. Also, the layering is typically laterally discontinuous and a single bed can contain multiple layers. This suggests that during a single sedimentation event (i.e. deposition of a single bed) the

buildup of sediment on the bed was not texturally uniform, which here is reflected also as differences in mineralogy. These spatial variations are interpreted to indicate that beds of coarse and generally graded sandstone and conglomerate represent the amalgamation of discontinuous patches of sediment that accumulated on the bed during a single transport event. This, in turn, suggests that conditions within the flow similarly varied both spatially and temporally and that the bed is an amalgamated patchwork of sediment deposited by largely bypass flows. Additionally, Navarro (2016) reported that brown layers were more common in the coarser axial part of the channel fills, whereas they become increasingly inter-layered with light grey layers toward the margin. This would also be expected based on the extensive calcite cement in the axial part of the channel fill, which reflects abundant coarse carbonate grains that were transported and deposited closer to the flow axis (dark-brown layers). Lighter coloured layers become increasingly common laterally, reflecting the proportionate increase of fine and medium quartz sand grains along the margins of the flow.

The comparatively heavy isotope values of the primary cement from bacon channel sandstones ($\sim 0\text{‰}$) relative to other Neoproterozoic deep-marine sections (approximately -6‰ ; Jiang et al., 2007) and the presence of carbonate clasts with shallow-water growth forms (e.g. microbialites) suggest that carbonate sediment precipitated and was subsequently resedimented from a coeval shallow-water carbonate platform. The microspar cement surrounding the quartz grains is interpreted to be the recrystallized remnant of detrital carbonate grains that were hydraulically equivalent to coarse quartz sand grains transported in the lower, high-energy part of throughgoing turbidity currents (Lowe, 1988; Arnott and Hand, 1989; Mulder and Alexander, 2001). After deposition, these grains neomorphosed on the seafloor and/or in the shallow subsurface to form the microspar cement that encompasses the quartz grains (e.g. Machel, 1985; Major, 1991; Scholle and Ulmer-Scholle, 2003; Boggs and Krinsley, 2006; van Der Kooij et al., 2010; James and Jones, 2016).

IV.II.II Proximal Levees

Proximal levee units are slightly less common in the FIC than channel complexes and make-up ~ 8% of the stratigraphy. Overall, these units have approximately equal proportions of sandstone (54%; both silicarenite and siliceous calcarenite) and mud-rich deposits (46%; mudstone and siliceous calcilutite) (Figure 2.13G). Sandstones are normally graded and medium- to thick-bedded with abundant traction structures, including thick planar-stratified intervals and multiset ripples (Facies 2 = 29%) (Figure 2.13C-F), however, more complete lower division turbidites also occur (T_{ade} , T_{abde} , T_{ace} ; Facies 1 = 46%). Sandstones typically fine upwards into up to 10 cm-thick mudstone, siltstone, or siliceous calcilutite caps. Also, these beds are sometimes intercalated with up to 3.5 m-thick intervals of mud-dominated very thin- to thin-bedded turbidites (facies 3 = 25%). The FIC in the study areas hosts four proximal levee units (PL 1-3 and CT 3 at MR).

PL 1 overlies BST 1 and crops out in all study areas. It is thickest at the HS (13 m), but is ~ 5 m-thick at all other localities. PL 2 only occurs directly above calcidebrite 1 at MR-S and is ~ 7 m-thick. Both PL 1 and PL 2 are lithologically similar and are comprised of fine- to coarse-grained, thin- to very thick-bedded (3-123 cm-thick) T_{ade} , T_{ab} , T_{abde} , T_{abcde} , T_{bcde} , T_{cde} turbidites (facies 1 and 2; 43% and 41%, respectively) intercalated with minor very thin- to thin-bedded upper division turbidites (Facies 3, 18%). The basal portion of most sandstones are medium-grained and are characterized by thick T_b intervals and thinner T_c units that commonly have multiple ripple sets (46%) and rare climbing ripples (9%). Significantly, at MR-N BST 4 crops out at the same stratigraphic level as PL 2 (~600 m N) (Figure 2.14). Approximately 150 m south of MR-N BST 4 strata transition to well-stratified, thin- to medium-bedded T_{ade} and T_{de} turbidites (35:65 sandstone:mudstone ratio). Here siliceous calcarenites (T_a) are typically < 5 cm thick, but occasionally are greater than 10 cm thick. Further south from this location (~ 250 m) this unit is much more mud-rich (T_{de}/T_e) and siliceous calcarenite interbeds become thinner (< 3 cm-thick; T_{ae}/T_{ade}) and less common. Finally, another 200 m south (at MR-S) is PL 2 (65:35 sandstone:mudstone).

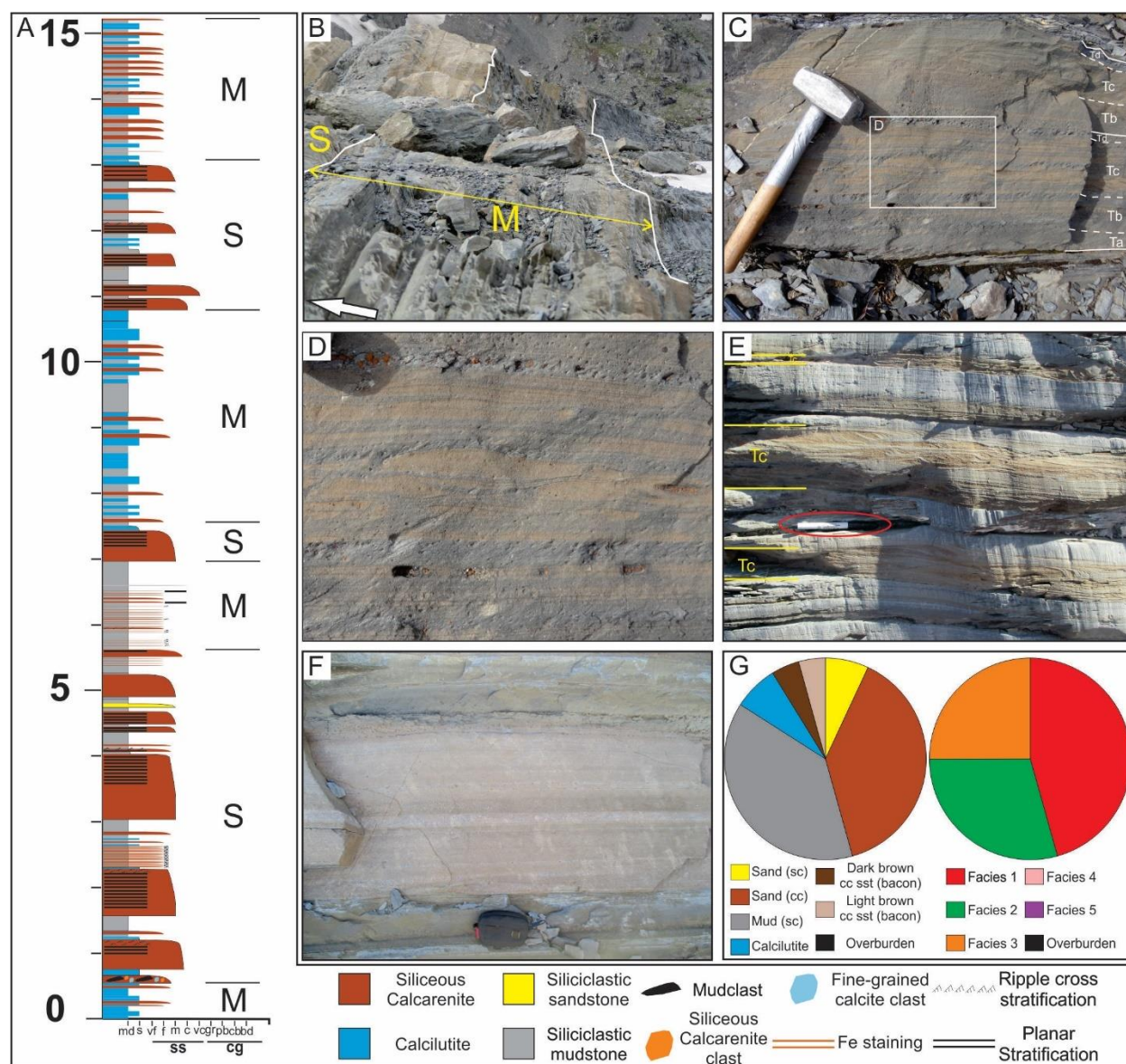


Figure 2.13: (A) Stratigraphic section of CT3 at MR-S with sharply-bounded mud-rich (M; siliciclastic and siliceous calcilutite) and sand-rich (S) turbidite packages. (B-F) Representative photographs of proximal-levee deposits in the FIC. (B) Base of CT3 at MR-N. White lines indicate the basal and top contacts of a siliceous calcilutite-dominated package (M) overlain by a sand-rich package (S). White arrow indicates stratigraphic up direction. (C) T_{abcd} and T_{bcd} beds in PL1 at HS. Both beds contain a thick T_c composed of multiple ripple sets, characteristic of levee deposits. (D) Close-up of photo C showing the well-developed stacked ripple sets. (E) Mud-rich portion of CT3 with three siliceous calcarenite T_{ce} beds containing well-developed multiple ripple sets. Ripple cross-stratification in the middle bed shows a moderate climb (i.e. inclination) to the right. Marker for scale circled in red. (F) Thick T_b bed in CT3. (G) Lithological (left) and facies (right) make-up of FIC proximal levee units.

The make-up of the uppermost calciturbidite horizon (CT 3) in the FIC changes between CC-HS and MR as the proportion of facies 2 sandstone increases from 8% to 50% and siliceous calcilutite

decreases from 64% to 18%. At MR, CT 3 and PL 3 manifest as a series of alternating packages of siliceous calcarenite-rich and mud-rich very thin- to thin-bedded turbidites with sharp contacts between them (Figure 2.13A-B). The sandstone-rich intervals are up to 5 m-thick with individual very thin- to very thick-bedded (2-141 cm-thick, average = 25), upper fine- to lower-coarse-grained (majority are upper-medium) T_{ade} , T_{abde} , T_{abcde} and T_{bcde} turbidites. Most beds are planar-stratified and are uncommonly capped by an up to 5 cm-thick ripple cross-stratified unit composed of one or less commonly multiple sets (Figure 2.13E-F). Overall, sandstone makes-up 58% of the stratigraphy in these intervals, and strata between the sandstones consists of up to 30 cm-thick thin-bedded, mud-rich turbidites with some < 1 cm-thick ripple-cross stratified siliceous calcarenite (single- and multiple-set). The mud-rich packages are up to ~ 3.5 m-thick and are predominantly made up of siliciclastic mudstone (59%) and lesser siliceous calcilutite (27%) and siliceous calcarenite (13%) with those in CT 3 being significantly more siliceous calcilutite rich than PL 3 (33% and 3%, respectively). Most beds are very thin- to thin-bedded T_{de} turbidites with occasional up to 30 cm-thick sandstones (T_{ade} and T_{bde}) and < 2.5 cm-thick T_{ce} turbidites.

IV.II.II.I Interpretation

Despite cropping out adjacent to channel sandstones in only one place, strata described here are interpreted to be sand-rich levee deposited formed close to a coeval channel. This interpretation is based on the intercalation of medium- to very thick-bedded siliciclastic- and carbonate-cemented, sand-rich turbidite and very thin- to thin-bedded turbidite packages that abruptly overlie one another in CT 3 and PL 3 at MR, in addition to abundant multiset ripple-stratified intervals and proportionately thick planar-stratified units (e.g. King et al., 1994; Coleman, 2000; Cronin et al., 2000; Basu and Bouma, 2000; Hickson and Lowe, 2002; Kane et al., 2007; Arnott, 2010). The occurrence of multiset ripple cross-stratification with negligible to low angle of climb suggests competence-driven deposition in sustained, low-energy flows, which is consistent with conditions on channel bounding levees (e.g. Walker, 1985; Khan and Arnott, 2011; Khan et al., 2011; Khan, 2012; Bergen, 2017).

Previous work by Bergen (2017) showed that proximal levee deposits formed 6-21 m-thick stratal packages with distinct upper and lower parts. The lower part consisted of medium- to thick-bedded, upper medium- to coarse-grained sandstone (T_a , T_{ab}) intercalated with very thin- to thin-bedded fine-grained, upper-division turbidites (T_{ces}). It then was sharply overlain by very thin- to thin-bedded upper-division turbidites intercalated with medium-bedded lower-division turbidites. Utilizing previous experimental and theoretical work (de Leeuw et al., 2016; Tilston et al., 2015; Tilston, 2017) and field observations, Bergen (2017) argued that channel inception initially formed ridges (incipient levees) on either side of the axis of flow, which subsequently confined further flows (Figure 2.14). Early in channel formation coarse-grained flows had a plug-like density structure (negligible vertical stratification and high flow efficiency; Tilston et al., 2015; Kneller et al., 2016) and a velocity maximum elevated well above the height of the channel margins that allowed the sand-rich lower part of the flow to spill over the margins and deposit thick sandstones forming the lower part of levee packages. The intercalated thinner-bedded, mud-rich turbidites resulted from smaller magnitude flows in which only the finer-grained upper portion overspilled the channel margins. With time, and ongoing deposition on the margin by successive flows, the height of the now well-developed levees exceeded the height of the velocity maximum (and maximum rate of density change), allowing only the finer, upper part of currents to overspill, depositing packages of thinner-bedded, mud-rich turbidites. The uncommon thick interbedded sandstones, on the other hand, were deposited by exceptionally large flows where the lower part of the flow overspilled. These characteristics resemble those in CT 3 and PL 3 at MR, where the siliceous calcarenite dominated intervals are analogous to the lower packages, which then are sharply overlain by siliceous calcilutite and siliciclastic mud-rich intervals, akin to the upper packages of Bergen (2017).

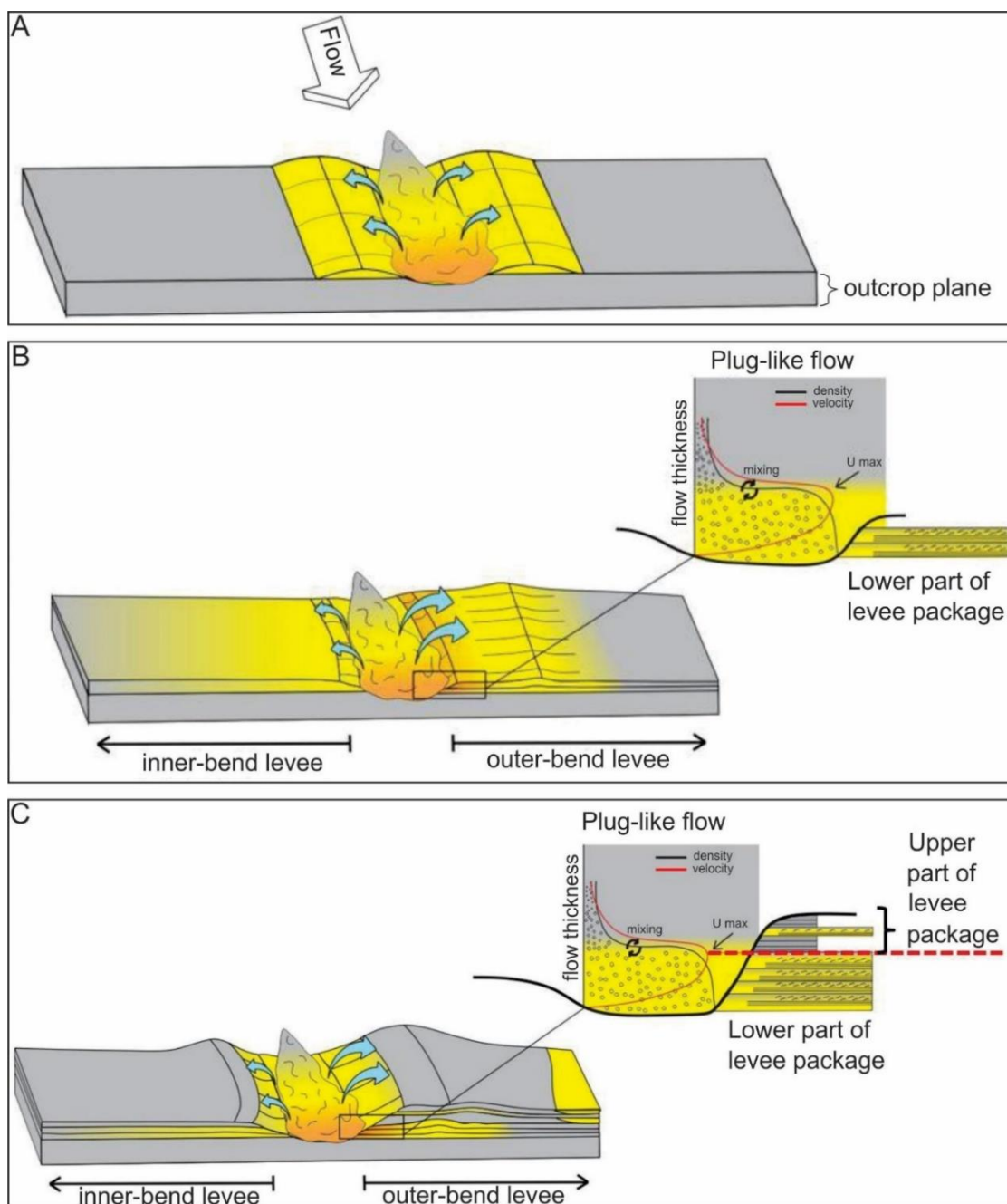


Figure 2.14: Bergen's (2017) schematic model for the deposition of lower and upper levee parts of levee packages. (A) The sharp basal contact of each package overlain by coarse-grained turbidites is interpreted to mark channel inception and the base of incipient levees that confined the axial part of subsequent flows. (B) The predominantly coarse-grained beds in the lower part of packages suggest that channelized flows were similarly coarse-grained and well-sorted. Thoroughgoing flows would have had a plug-like density structure with the height of the velocity maximum above the channel margins, which allowed the lower, coarse-grained dense part of flows to overspill and as a consequence thick-bedded, coarse-grained turbidites in the lower part of packages to be deposited. (C) Continued levee aggradation would have caused the relief between the channel floor and levee crest to exceed the height of the velocity maximum of the average channelized flows. Thus, only the upper, fine-grained parts of currents typically overspilled, depositing the upper markedly finer part of each levee package. Coarse-grained sediment, which is transported in the lower part of the flow, remained confined to the channel.

Based on the abundance of thick- to very thick-bedded, medium- to coarse-grained sand-rich T_{acde} and T_{bcde} turbidites (average sandstone:mudstone ratio = 65:35) with thick T_b intervals, strata in PL 1 and PL 2 are interpreted to represent outer-bend levee deposits (Posamentier, 2003; Arnott, 2010; Khan and Arnott, 2011; Morris et al., 2014) (Figure 2.15). These sand-rich strata were likely deposited by inertial overspill, which allowed some of the lower, medium- to coarse-grained part of the flow to overtop the channel margin while the rest of the flow remained confined to the channel and continued downslope (e.g. Imran et al., 1999; Straub et al., 2008; Khan and Arnott, 2011). Conversely, the thin-bedded, fine-grained, siliciclastic mud-dominated (35:65 sandstone:mudstone) strata between PL 2 and BST 4 at MR are interpreted to be inner-bend levee deposits associated with the same channel (BST 4) (e.g. Khan and Arnott, 2011). Therefore, the unexposed sinuous channel associated with these levee units likely curved away from the exposed outcrop surface and then curved back and intersected the outcrop at MR-N (Figure 2.15). Previous work by Khan and Arnott (2011) reported that inner-bend levee deposits grade laterally from thin-bedded mud-rich turbidites with occasional medium- to thick-bedded sandstone T_{a-d} turbidites to fine, thinly-bedded turbidites over < 150 m, which is consistent with the dimension of the lithological changes and spatial dimensions described here.

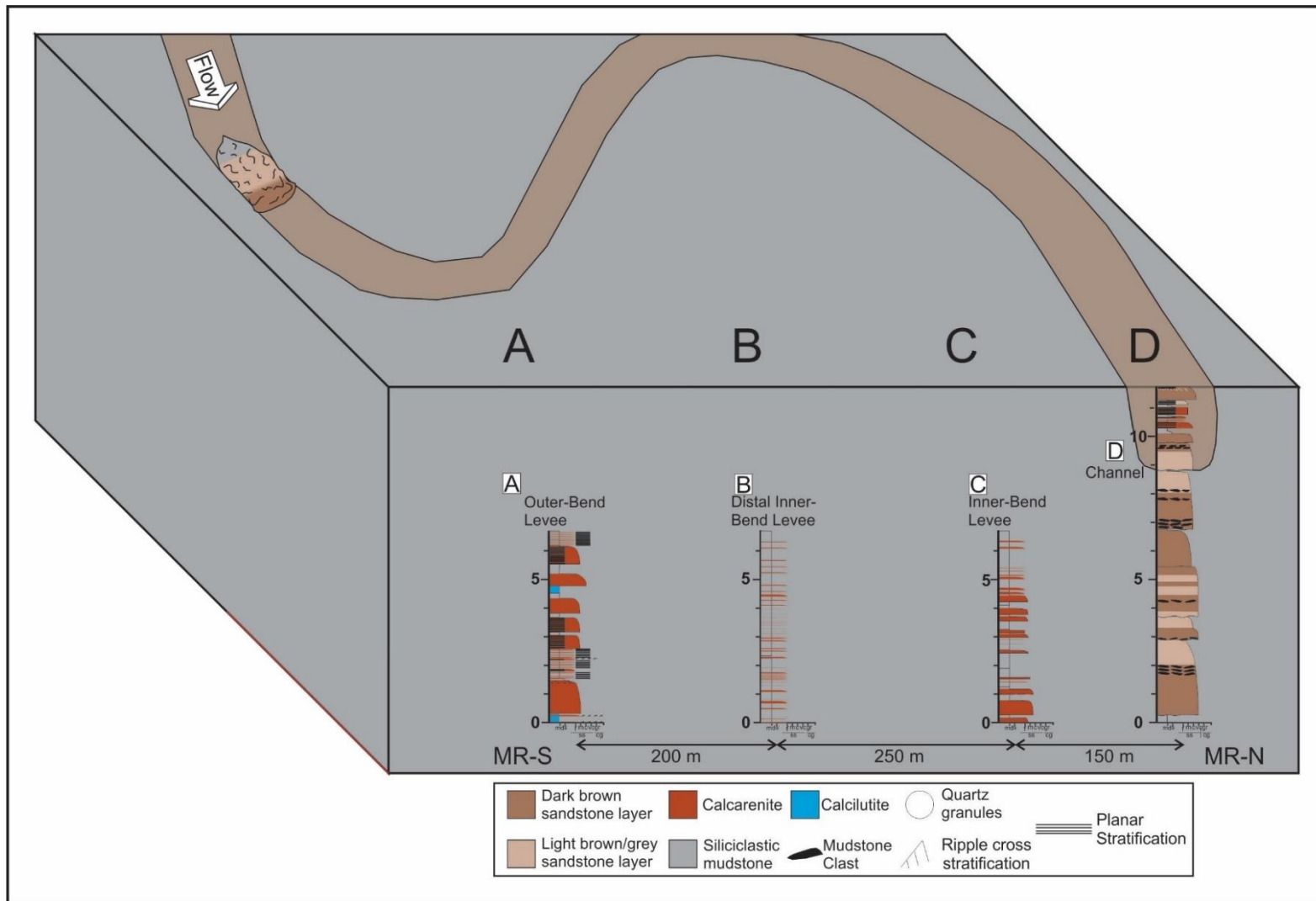


Figure 2.15: Schematic model for the lateral transition from PL 2 (A) to BST 4 (D) at Milk River. (A) and (D) are measured stratigraphic sections, whereas (B) and (C) are schematic logs based on field photographs. PL 2 (location A) is interpreted to have been deposited on the outer-bend levee of the unexposed BST 4 channel, however the channel then curved away from MR (into the outcrop) before bending back and intersecting the MR outcrop at location D. Thinner, finer grained strata in locations B and C represent deposition on the inner-bend levee of the BST 4 channel.

V.II.III Distal Levee Deposits

Distal levee deposits are the most common stratal element in the FIC, making up just under 60% of total FIC stratigraphy. These units are composed primarily of mud-rich, thin-bedded turbidites (Facies 3, 66%) with occasional very thin- to medium-bedded sandstone that fine upward to siltstone/mudstone and/or siliceous calcilutite caps (facies 1 and 2; 18% and 24% respectively). In the FIC these deposits are further divided into two subtypes based on their mineralogy: siliciclastic-rich and carbonate-rich.

IV.II.III.I Carbonate-Rich Distal Levee Deposits

The FIC, and its recognition as a regional marker, is based largely on the presence of three Dm-thick, areally extensive, fine-grained calcareous turbidite, or simply calciturbidite, horizons in an otherwise siliciclastic pile of deep-marine sedimentary rocks (Ross and Murphy, 1988; Ross et al., 1989). The base and top of each unit is sharp and can be traced across the entire study area, however, as stated above, the uppermost calciturbidite unit (CT 3) is substantially more sand rich at MR.

The lowermost distal levee succession (CT 1) occurs ~ 15 m below the base of BST 1 and ranges from 3-9 m-thick. Of the three calciturbidite units at CC-HS CT 1 has the most abundant siliceous calcarenites (35%) and fewest siliceous calcilutites (16%). It is composed of thin- to medium-bedded T_{ade} , T_{abde} , T_{acde} , T_{bcde} , and T_{cde} turbidites (facies 1 and 2; 33% and 25% respectively) with fine- to medium-grained reddish-orange siliceous calcarenites capped by siliciclastic siltstone and/or mudstone caps. Calciturbidite 2 (CT 2) is 9-19 m-thick and underlies BST 2 at CC-N and HS and 123-1 at CC-S (see below). It has a distinct olive-green to dark greenish-grey colour in outcrop due to abundant very thin- to thin-bedded siliceous calcilutites (Facies 3, 73%), which are interbedded with rare very thin- to thick-bedded massive siliceous calcarenites (Facies 1, 16%) and very thin- to thin-beds of siliciclastic mudstone (Facies 3, 11%) (Figure 2.16A-B). CT 3 crops out at the top of the FIC and at CC and HS locally is eroded completely by Isaac channel 1, which limits its stratigraphic thickness in some sections (4-21 m-thick). Its internal make-up is similar to CT 2, but siliceous

calclutites are darker (dark-grey to black) due to more abundant fine-grained pyrite compared to CT 1 or CT 2 (up to 15%) (Figure 2.16C). The increased iron content, and therefore more extensive oxidation, also gives CT 3 siliceous calcarenites a deeper red colour in outcrop.

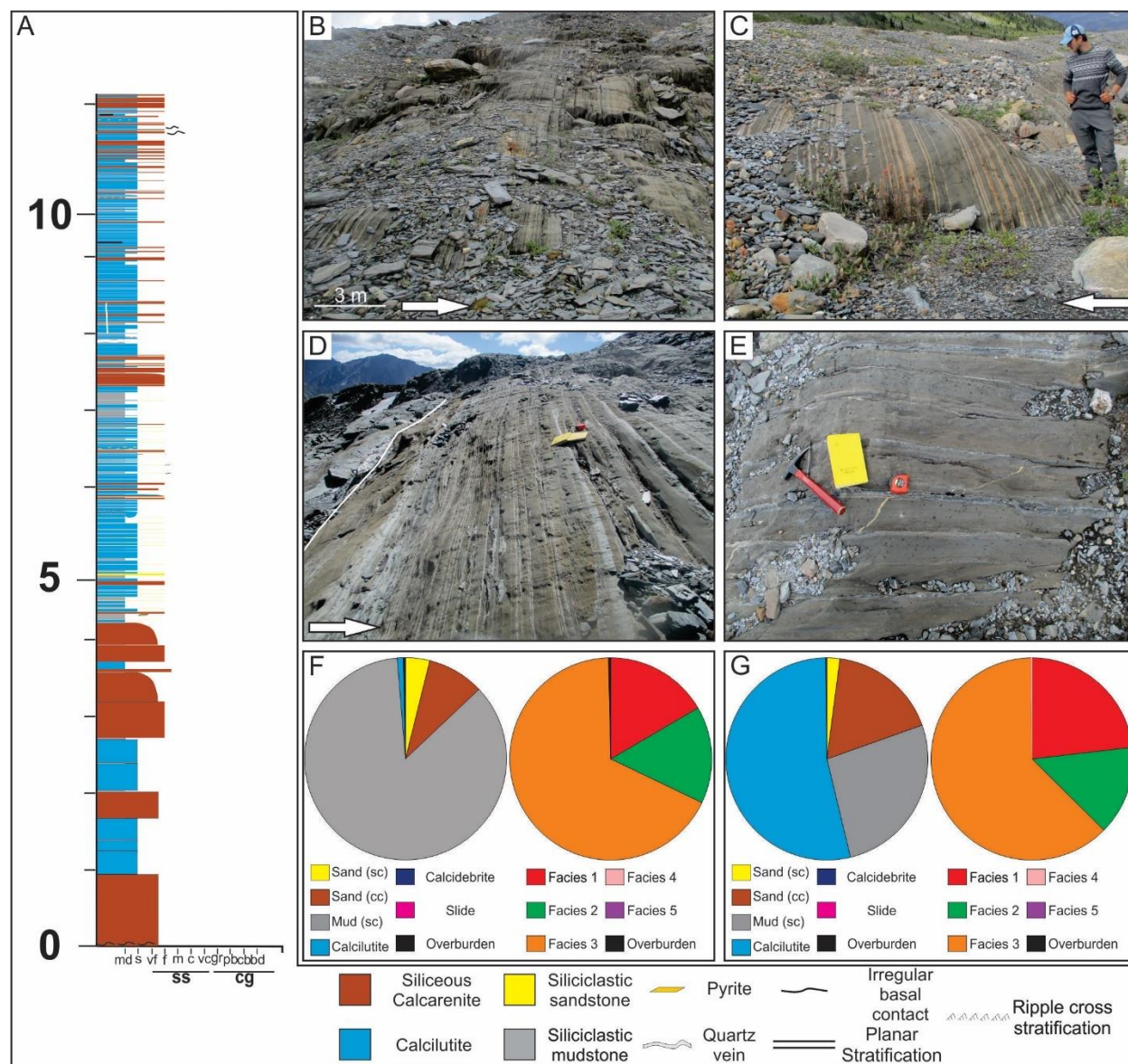


Figure 2.16:(A) Stratigraphic section of CT 2 at HS. (B-E) Representative photographs of siliceous calcilutite-rich distal levee deposits in the FIC. White arrows indicate stratigraphic up direction. (B) CT 2 at HS. At CC and HS CT 2 can be easily identified by its olive green and greenish-grey colour due to abundant siliceous calcilutite beds. (C) CT 3 at CC-N with multiple thin, interbedded red and darker-coloured siliceous calcilutite beds. (D) CT 4 at MR-S. (E) Photograph of multiple brown and dark grey siliceous calcilutites (T_d) in a calciturbidite unit. Each calcilutite is separated by an ~ 1 cm-thick grey, siliciclastic mudstone (T_e). (F) Lithological (left) and facies (right) make-up of siliciclastic mud-rich distal levee deposits in the FIC. (G) Lithological (left) and facies (right) make-up of siliceous calcilutite-rich distal levee deposits in the FIC.

At MR there is an additional ~ 10 m-thick calciturbidite horizon (CT 4) located in the siliciclastic mud-rich unit above PL 1 (Figure 2.16D). CT 1, 2 and 4 all have similar lithological compositions and outcrop appearances at MR. They are dominated by very thin- to thin-bedded, light brown, dark reddish-brown and dark-grey siliceous calcilutites (46%) interbedded with thin- to thick-bedded, fine to coarse-grained reddish-brown siliceous calcarenites (T_{ade} ; 24%; $\pm$ rare silicarenites (< 1%)) and siliciclastic mud-rich, thin-bedded turbidites (T_{de} ; 28%). Siliceous calcarenites are typically massive and fine-grained (T_a ; Facies 1, 34%), although thicker beds tend to be coarser and occasionally planar stratified (T_b ; facies 2, 11%). Only thirteen beds contain ripple cross stratification, making up a minor proportion of calciturbidite units.

IV.II.III.II Siliciclastic-Rich Distal Levee Deposits

Siliciclastic mud-rich, very thin- to thin-bedded turbidite units represent most of FIC stratigraphy (46%) and many individual units occur in all study areas (Figure 2.16F; Figure 2.21). They are composed of 85% grey to dark-grey very thin- to thin-bedded siliciclastic mud-rich turbidites (Facies 3) with isolated thin- to thick-bedded, fine- to coarse-grained sandstones (T_{ae} , T_{be} , and T_{ce} turbidites (facies 1 and 2); predominantly siliceous calcarenite, occasionally silicarenite) and rare, local, < 2 m-thick siliceous calcilutite rich successions (Facies 3). At CC and HS these units range from 1-20 m-thick, but are significantly thicker at MR, two being > 60 cm-thick.

IV.II.III.III Interpretation

Siliciclastic mudstone-rich and siliceous calcilutite-rich units are interpreted to represent deposition on the distal levees of active channel belts. Most beds were deposited by low concentration, fine-grained turbidity currents, although some mudstone (T_e) beds probably represent deposition from hemipelagic fallout (Meyer, 2004; Payros and Pujalte, 2008; Angus, 2016). Rare interbeds of medium- to thick-bedded sandstones are likely related to anomalously dense, high-energy channel flows that allowed sand-size particles to overspill the levees and travel further from the channel margin.

The sharp bounding contacts of calciturbidite horizons indicate that the change from siliciclastic- to carbonate-dominated mineralogy was abrupt and linked to the mineralogical makeup on the coeval continental shelf. Calciturbidites most probably reflect episodes of sustained elevated sea level that allowed for the development and proliferation of a shallow water carbonate platform, and downslope resedimentation of mostly fine-grained carbonate-rich sediment (Droxler and Schlager, 1985; Schlager et al., 1994). Coarse sediment remained confined to the channel and, to a lesser extent, the adjacent proximal levee, with the fine-grained fraction, composed mostly of fine carbonate sediment overspilling and depositing calciturbidite-rich horizons 100s of metres to several kilometres beyond the channel margin. The sharp upper contacts between siliceous calcilutite and siliciclastic mudstone units suggests the abrupt shutdown of the carbonate platform either by drowning (eustatic rise) or subaerial exposure (eustatic fall), and the return of the more typical supply of siliciclastic sediments from the hinterland.

IV.II.IV 123s

IV.II.IV.I Description

These units are characterized by a tri-partite repetition of packages, which form units up to 46 m-thick, typically with a basal fine- to very coarse-grained, thin- to thick-bedded reddish-orange siliceous calcarenite or occasional black to dark brown carbonaceous sandstone (Package 1; facies 1 and 2), overlain by siliciclastic mud-rich turbidites (Facies 3) with occasional thin-bedded, fine- to medium-grained siliceous calcarenites (T_{ade} , T_{bde} , T_{bcde} , T_{cde} , T_{de}) (Package 2) and draped by a package of very thin- to thin-bedded greenish-brown to black siliceous calcilutites (Package 3; Facies 3) (Figure 2.17A-C). The basal sandstone tends to be less laterally persistent compared to facies 2 and 3, and in some successions is absent, resulting in the alternation of 2 and 3 packages. Overall, 123 units are composed predominantly of Facies 3 (76%) with lesser Facies 1 and Facies 2 (17% and 7%, respectively). There are two 123 units, 123-1 (stratigraphically lower) and 123-2 (higher), that crop out in all study areas and make up 12% of FIC's total stratigraphy.

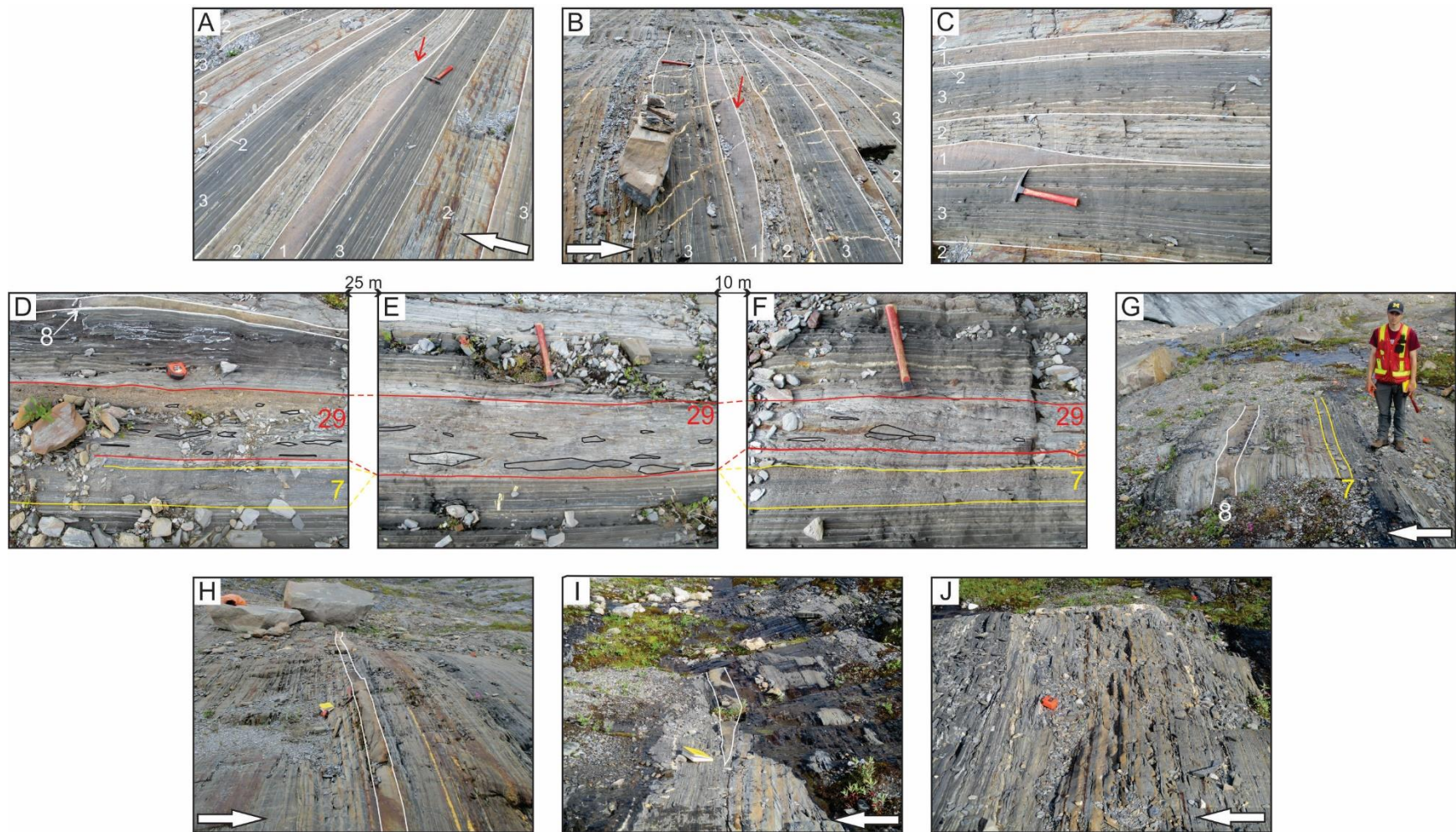


Figure 2.17 (previous page): Representative photographs of 123 packages in the FIC. White arrows indicate stratigraphic up direction. (A) 123s at CC-S showing the characteristic repetitive stacking of the three packages (labelled 1, 2, 3). The base of the central 123 is marked by isolated sets of dune cross-stratified siliceous calcarenite (Sandstone 6 in Figure 2.18) that thin laterally as they pass into planar-stratified sandstone. (B) Same dune cross-stratified (DCS) sandstone in (A; indicated by red arrow), but viewed in the opposite direction. It maintains a similar thickness along its length but then thins rapidly at the bottom of the photograph. (C) Close-up of abrupt thinning of the DCS sandstone in Photo A and B, which then thickens ~ 20 m laterally and then thins. (D) Sandstones 7 and 29 at 123 Log 3.1. Sandstone 7 (outlined in yellow in D-G) is a planar-stratified, carbonaceous sandstone overlain by a 3 cm-thick siliciclastic mudstone. Sandstone 29 (outlined by solid red line in D-F) consists of a carbonaceous basal part and orange siliceous calcarenite upper part with abundant elongate mudclasts oriented parallel to bedding throughout (some outlined in black). Sandstone above 29 is Sandstone 8 (white arrow; outlined in white in D and G). (E) 25 m S of photo D: sandstone 29 has scoured and here completely eroded sandstone 7. (F) 35 m S of photo D: sandstone 29 has thinned and is once again underlain by sandstone 7. (G) ~ 35 m N of photo D, looking downstrike: Sandstone 29 pinches out just beyond the bottom the photograph, but sandstone 7 and 8 remain, separated by ~ 70 cm of siliciclastic mudstone-rich, thin-bedded turbidites. (H) Sandstone 5 just south of log 2.1 (outlined in white in H-I): Note that it is underlain by 6 thin-bedded T_{ae} beds with reddish-orange siliceous calcarenites. (I) 35 m south of photo H: sandstone 5 abruptly pinches out, but the thinner T_{ae} beds below continue. (J) Same stratigraphic level at 123 Log 3: sandstone 5 is noticeably absent. The thinner T_{ae} beds are still present, suggesting the pinch out of sandstone 5 is a stratigraphic feature and not a result of orogenic deformation.

Sandstone beds (Package 1) are predominantly siliceous calcarenite with rare (< 3%) silicarenite. The siliceous calcarenite is composed, on average, of ~ 30% quartz grains surrounded by mostly iron-rich microspar (~ 60%) with < 10% opaque minerals, dolomite and muscovite. Most (73%) of the sandstones in the 123s exhibit traction sedimentary structures, including planar-, ripple- and dune cross-stratification. The remainder are massive. These beds (25%) contain abundant mudclasts and scour underlying strata (see Sandstone 29 description below). Additionally, some of the thickest Package 1 layers (average = 42 cm) tend to have the least lateral extent (average = 30 m)

Excellent exposure of the 123 units at CC-S allowed for thorough analysis of Package 1 sandstone beds within. A detailed (mm-scale) correlation panel was constructed of sandstone beds > 5 cm-thick to identify any lateral lithological or dimensional trends (Figure 2.18). In total 35 sandstones were traced, 27 from 123-1 unit and 8 from 123-2, making up 7% of the 123 stratigraphy, on average. Beds ranged from 4-83 cm at their thickest point (average = 25) and their lateral extents ranged from < 1 m to 162 m (average = 75).

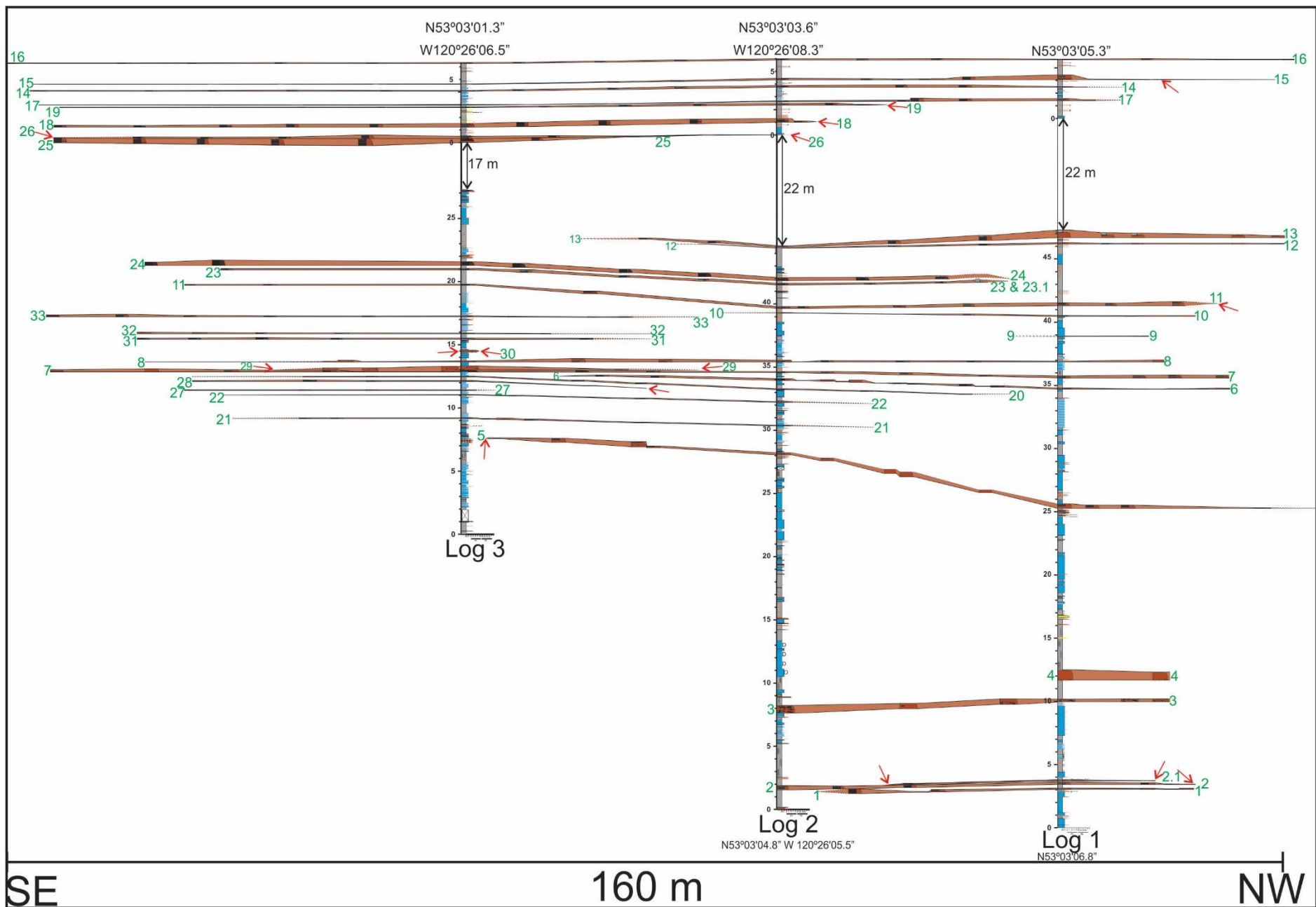


Figure 2.18 (previous page): 123 sections logged at Castle Creek south. The scale on each log is in metres. Sandstones thicker than 5 cm ($n = 35$) were traced laterally across the outcrop and their correlations are indicated by the brown coloured areas. Green numbers correspond to the sandstone beds that were traced. Red arrows indicate stratigraphic pinch-outs.

In 123-1 the number of traceable sandstones increases stratigraphically upward, with 6 beds in the lower 30 m and 21 in the upper 31-50 m (Figure 2.18; Figure 2.19A). This suggests that during early 123-1 development, the CC-S study area was located far from the primary pathway for throughgoing turbidity currents or experienced longer periods of elevated sea-level that led to the more common deposition of lime mud-rich calciturbidites. Nevertheless, later in 123-1 development sea level fluctuations became more common and/or of high amplitude, leading to an increase in sand-rich flows depositing Package 1 sandstones.

The location where each sandstone was thickest varied from 67 m north and south of the middle of the outcrop exposure, with the most occurring 1-10 m S. A histogram of the thickest location of each sandstone traced resembles a normal distribution with frequency increasing towards the centre of the exposure in both directions, except for rare occurrences from 1-19 m N (Figure 2.19B). A plot of the N-S thickest location of each sandstone versus stratigraphic height shows that sandstones in the basal ~ 10 m of 123-1 are thickest in the northern portion of CC-S, but shift to the south in the upper ~ 30 m, and, finally, in 123-2 is more or less uniform, although possibly slightly more abundant to the north (Figure 2.19C). Accordingly, it appears that the axis of the flows that deposited 123-1 sandstones were preferentially located just to the south of the centre of CC-S exposure.

However, south of Log 2 the basal part of 123-1 is highly (structurally) deformed, making the sandstones impossible to follow with confidence in this direction and three of the five lower sandstones in this part are at their thickest immediately adjacent to the deformed region (Figure 2.18). Thus, it is plausible that these beds are thickest to the south and the clustering of thickest sandstones to the north in the basal part of 123-1 is an artefact of exposure.

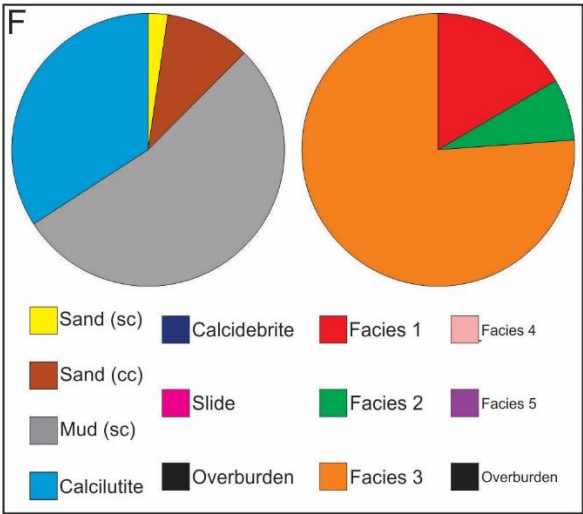
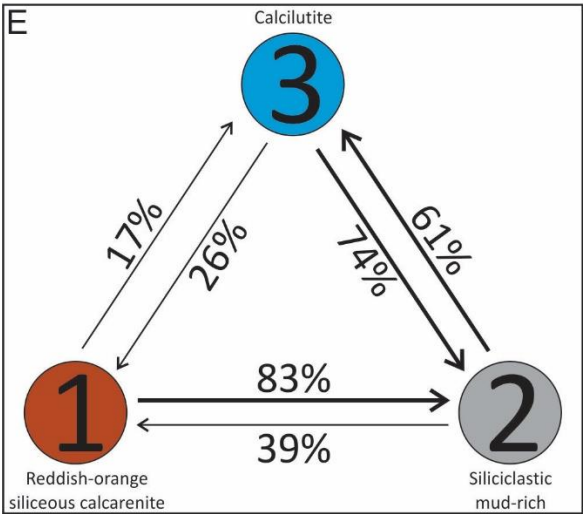
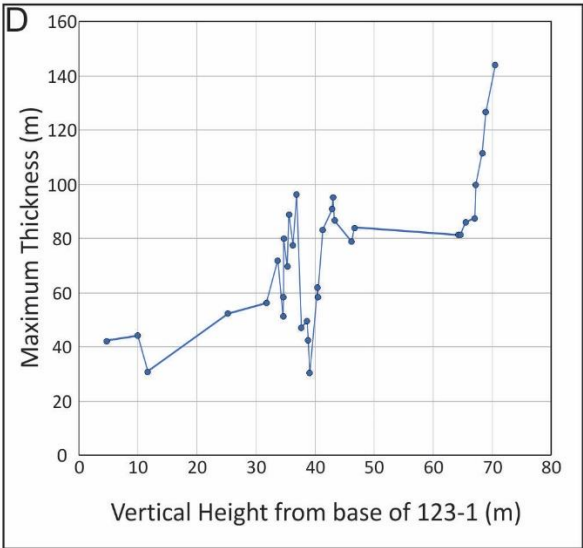
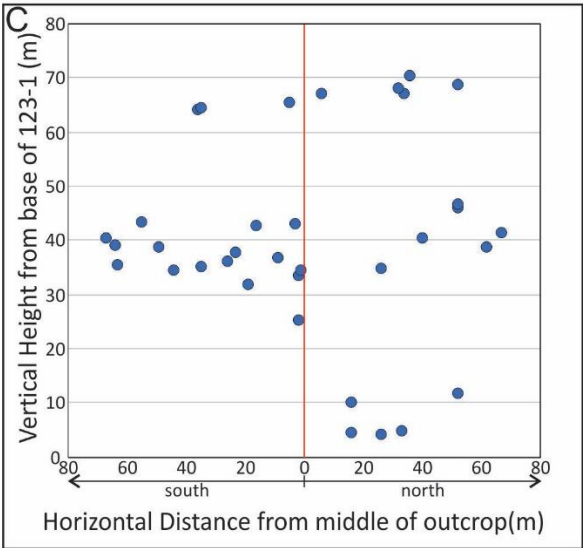
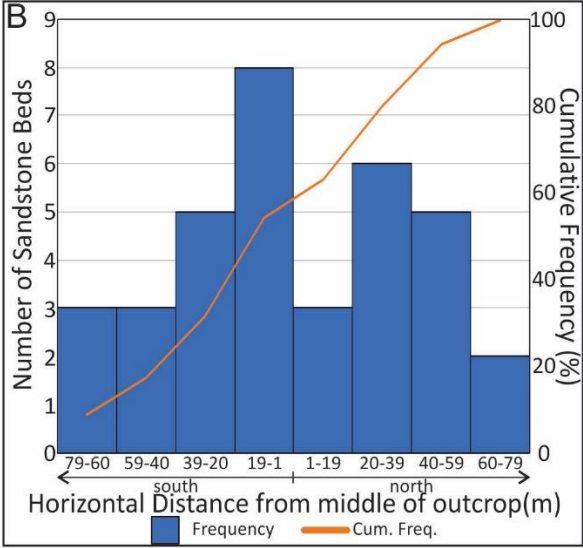
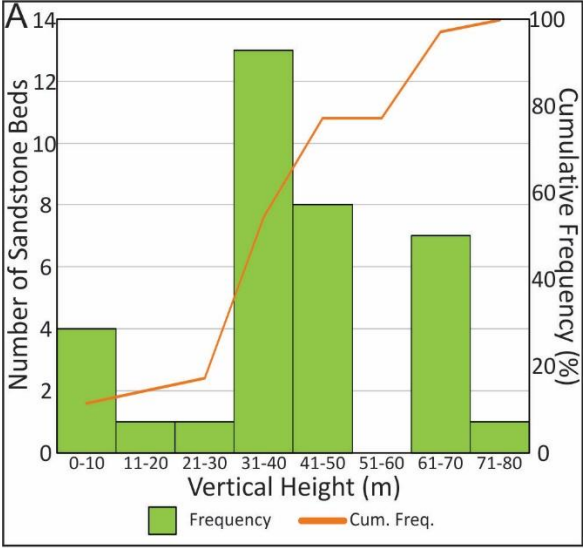


Figure 2.19 (previous page): (A) Histogram showing number of traceable sandstone versus stratigraphic height in the 123s at CC-S. (B) Histogram showing the location of the thickest part of each sandstone traced in relation to the middle of the outcrop exposure in the 123s at CC-S. In (A) and (B) the orange line represents the cumulative frequency of the traceable sandstone beds. (C) Plot of vertical (stratigraphic) height of each sandstone bed (datum is the base of 123-1) versus the location of its thickest part relative to the middle of the outcrop at CC-S. (D) Three bed rolling average of the maximum thickness of each sandstone versus stratigraphic height. (E) Schematic of the transition matrix illustrating the vertical succession of facies in the 123 units. Thicker arrows correspond to the most common transition from each state to the next (i.e. Package 1 is most often vertically overlain by a 2 package (83%) rather than a 3 (17%)). (F) Lithological (left) and facies (right) make-up of 123 units in the FIC.

Additionally, the total lateral extent of Package 1 sandstones increases stratigraphically upwards (Figure 2.19D), although a few caveats exist. First, the limited lateral extent of sandstones in the basal 30 m is an artefact of the deformation south of log 2. However, even if the sandstones that extend into this deformed area are omitted the trend remains as the beds in 123-2 extend further than those in the upper part of 123-1. Secondly, the lack of paleoflow indicators throughout the section makes it difficult to determine the orientation the outcrop relative to flow direction, which if oblique would lengthen the bed in outcrop. However, 10 of the traceable sandstones stratigraphically pinched out lateral to their point of maximum thickness, indicating that at least some of them were cut at an angle close to perpendicular to the flow direction. For example, Sandstone 29 is up to 47 cm-thick and consists of massive coarse-grained siliceous calcarenite with abundant mudstone clasts and locally completely erodes the underlying Sandstone 7 (Figure 2.17D-F). Forty-five metres north from this location 10 cm of siliciclastic mud separated Sandstones 29 and 7, and 29 is a 6 cm-thick, medium-grained T_{be} before abruptly pinching out to the north (Figure 2.17G), while sandstone persists. On the other hand, Sandstone 5 pinches out rapidly to the south: at its thickest it is a medium to coarse-grained 42 cm-thick T_{abcde} and is underlain by six thinner T_{ae} turbidites (Figure 2.17H). Thirty-seven metres south of this location Sandstone 5 abruptly thins to a 7 cm-thick T_{ade} before pinching out (Figure 2.17I), however the underlying thin turbidites remain after this, suggesting that it is a true stratigraphic pinch out and not the result of tectonic deformation (Figure 2.17J).

There are also a few sandstone beds that do not gradually thin away from their maximum thickness, instead they pinch-and-swell multiple times, similar to what has been seen in previous Isaac Formation studies (e.g. Khan, 2012; Bergen, 2017). For example, Sandstone 6 goes from a 14 cm-thick medium-grained T_b to a 1.5 cm-thick, fine-grained T_b just 5 metres north and then thickens once again to a 13 cm-thick dune cross-stratified sandstone (Figure 2.17A-C). This bed thins and thickens once more before it gradually thins until pinching out. A similar trend occurred in the other dune cross-stratified bed (~ 130 cm above Sandstone 6), indicating that the available sediment supply of sediment was insufficient for the development of an expansive field of dunes.

To determine any systematic stacking pattern within the 123 units (i.e. does a “2” commonly overlie a “1” and is in turn overlain by a “3”) the digitized logs were divided into their constituent 1, 2, or 3 packages and a transition matrix tabulated. Although unimportant for the analysis Package 1 typically consists of a single sandstone bed, where “2” and “3” packages commonly comprise numerous beds. A change from one package to another was considered when the overlying bed, or group of beds with the same lithology, was > 2 cm. For example, a < 2 cm-thick siliciclastic mudstone interbed in a > 2 cm-thick package of thin-bedded siliceous calcilutites would not be considered a transition to a new package and instead considered a “3”. In total 611 transitions were identified (Figure 2.19E; Figure 2.20) and showed that Package 1 was overlain by Package 2 and Package 2 by Package 3 83% and 61% of the time, respectively. A notable discrepancy is that Package 3 is more commonly overlain by Package 2 rather than a Package 1 (74% and 26%, respectively). This is likely due to the rarity and comparatively more limited areal extent of Package 1 sandstones compared to strata that make-up package 2 and 3. The transition matrix was then evaluated by a full first-order Markov chain analysis (see Terlaky (2014) and Terlaky and Arnott (2016) for details) which showed the stacking pattern of the various packages making up 123 units is not random, but rather follows a systematic trend with a basal siliceous calcarenite (Package 1) overlain by siliciclastic mud-rich, thin-bedded turbidites (Package 2) and draped by thin-bedded calciturbidites (Package 3).

A					B				
Observed Transition Matrix - Counts					Observed Transition Matrix - Probabilities				
From/To	1	2	3	Row Totals	From/To	1	2	3	Row Totals
1	0	126	26	152	1	0.000	0.829	0.171	1.0
2	103	0	164	267	2	0.386	0.000	0.614	1.0
3	50	142	0	192	3	0.260	0.740	0.000	1.0
Column Totals	153	268	190	611	Column Totals	0.646	1.569	0.785	

Figure 2.20: Transitions observed in the three detailed stratigraphic logs measured through the 123 units. (A) Observed transition count matrix. (B) Probability transition matrix. The bold numbers highlight the most probable transitions.

IV.II.IV.II Interpretation

The repetitive occurrence of siliciclastic (package 1 and 2) overlain by carbonate lithologies (Package 3) that occur at the cm- to m-scale in the 123 units suggests the systematic and most probably, rapid fluctuation in sediment mineralogy, which quite possibly are related to high-frequency (i.e. Milankovitch cycles) eustatic oscillations. Package 1 strata most likely represent deposition during lowstand and resedimentation of coarse siliciclastic sediment and carbonate debris eroded from a deactivated, possibly subaerially exposed, carbonate platform. As sea level rose coarse sediment supply to the slope was terminated and replaced by fine siliciclastic mud (Package 2). Eventually rising sea level terminated siliciclastic input and initiated fine-grained carbonate shedding from a coeval shallow-water carbonate platform (Package 3).

Package 1 sandstones in the 123s were originally interpreted to be broad, shallow channels associated with overbank-splay deposition (Gammon and Arnott, 2007), however the vertical repetition of facies makes this unlikely. Later, Navarro (2016) suggested that they were gullies like those formed along slopes of modern carbonate and mixed systems (for example, the Great Bahamas Bank (Mulder et al., 2012a,b, 2014), Little Bahamas Bank (Schlager and Chernak, 1979; Rankey and Doolittle, 2012), Great Barrier Reef (Feary et al., 1993; Puga-Bernab  u et al., 2011, 2013) and northwest shelf of Australia (Fubara, 2014)). In most cases these modern gullies were identified using seismic and are characterized by an irregular basal contact overlain by high-amplitude layer

(potentially sand) which is draped by a low-amplitude layer (potentially mud) (Principaud et al., 2013). However, the sand-rich basal part is tens of metres thick and hundreds of metres wide, whereas the thickest sandstone in the FIC's 123s is 83 cm and the most laterally extensive example can only be traced for ~ 160 m (although this is constrained by the dimension of the outcrop). This dimensional disparity makes the correlation between slope gullies and Package 1 sandstones in the 123s unlikely.

The basal sandstones (Package 1) were previously interpreted to be the result of either the episodic passage of small-volume siliciclastic-rich turbidity currents or multiple gullies formed along discrete stratigraphic horizons, which subsequently steered later flows (Navarro, 2016). The detailed study of the 123s at CC-S shows no evidence for isolated sandstone beds along any single stratigraphic horizon. Instead it appears that there are ~ 5-8 m-thick intervals where medium- to coarse-grained Package 1 sandstones are more common, potentially indicating periods when sea level fluctuated more rapidly or simply when more coarse-grained turbidity currents were routed through CC-S. Additionally, Package 1 sandstone beds are typically an order of magnitude smaller than modern gullies, and therefore were probably incapable of steering any natural turbidity current. Thus, the medium- to coarse-grained sandstones of Package 1 are interpreted to represent low-relief (shallow) scours that formed near the bottom of much larger throughgoing turbidity currents. Scours were then infilled with medium to coarse sand that subsequently was reworked and/or eroded by later flows. These medium- to coarse-grained beds, therefore, are not the product of a single turbidity current, but rather the deposit of the last flow that moved over the scour. These flows were then succeeded abruptly by low-energy, mud-rich turbidity currents or hemipelagic fallout that deposited the thin-bedded siliciclastic mud-rich turbidite strata of Package 2. This was followed by a eustatic rise that resulted in the re-development of the carbonate platform and the downslope resedimentation of fine-grained carbonate sediment (e.g. Schlager et al., 1994; Wilson and Roberts, 1995) that built up the few m-thick calciturbidites of Package 3. A later eustatic fall terminated carbonate sedimentation and mobilized

coarse palimpsest and relict siliciclastic sediment into the basin, marking the beginning of Package 1 sandstone deposition at the base of the next 123 succession.

IV.II.V Scour Dominated Sandstone Units

IV.II.V.I Description

Two scour-dominated sandstone units occur in the FIC (SD 1 and SD 2), making up 1% of stratigraphy. Both are at MR and are underlain by an up to 10 m succession of siliciclastic mud-rich thin-bedded turbidites. SD 1 is overlain by CT 4, while SD 2 is overlain by more mud-rich deposits. The units range in thickness from 0.2-5.0 m and consist of 1 to 5 medium-grained to granule, medium to very thick-bedded, massive siliceous calcarenites to siliceous calcirudites, (T_a) intercalated with mud-rich, thin-bedded turbidites (Facies 1; Figure 2.21A,D). Mudclasts, and lesser carbonate clasts, are common in the T_a portion of the siliceous calcarenite beds, although their abundance sometimes decreases upwards. Clasts are typically aligned with their long axes parallel to bedding. Occasionally the coarse basal part of the bed fines upwards and is overlain by planar-stratified medium-grained sandstone capped by a 3-10 cm siliciclastic mudstone cap (i.e. a T_{abde} turbidite). Significantly, over tens to hundreds of metres laterally the coarse basal T_a pinches out, whereas the overlying planar-stratified sandstone persists (Figure 2.21B-C). Only one calcarenite (in SD 1) can be traced across the MR study area (~ 500 m).

IV.II.V.II Interpretation

Based on their thinness relative to other units (i.e. distal levee deposits, channel complexes, proximal levees, 123s), their limited lateral dimensions, coarse-grain size and the abundance of mudstone intraclasts, these strata are interpreted to be the fill of decametre-wide, metre deep slope scours formed as throughgoing turbidity currents abraded the siliciclastic mud-rich seabed. Although similar to the origin of Package 1 sandstone strata in the 123 units, the greater abundance of mudstone intraclasts and increased bed thickness may reflect differences in the substrate beneath the throughgoing turbidity currents. For example, the scour dominated units overlie successions of

siliciclastic mud-rich turbidites, whereas Package 1 sandstones typically overlie calciturbidites. These latter strata quite likely became partly to completely lithified shortly after deposition, which rendered them more resistant to scouring by later turbidity currents. Siliciclastic mud-rich sediment, on the other hand, was at best only partly compacted and therefore more susceptible to being eroded by later turbidity currents.

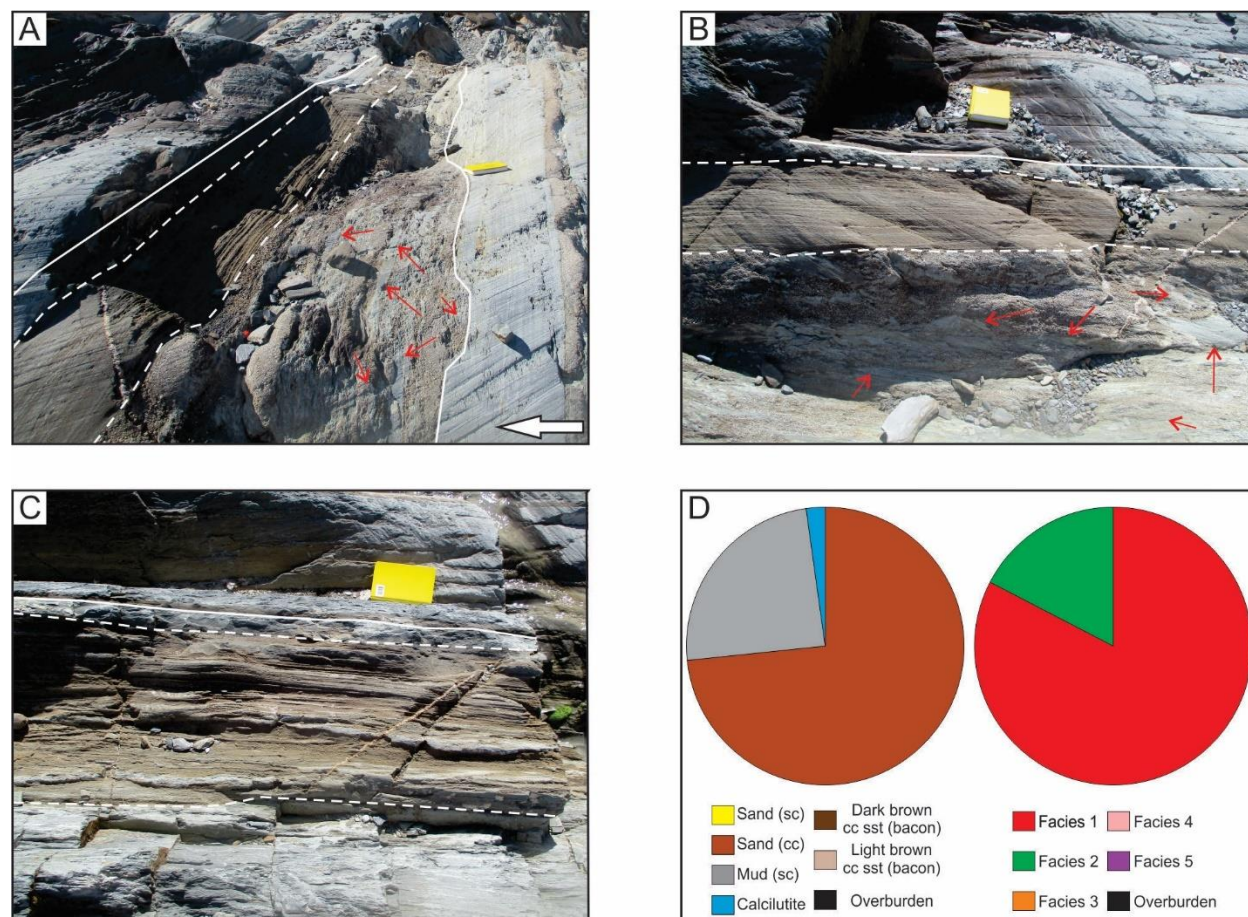


Figure 2.21: (A-C) Representative photographs of strata in a scour-dominated unit at MR-N. (A) Reddish-orange to brown siliceous calcarenite T_{abde} bed with an irregular, scoured basal contact and sharp upper contact (solid white lines). The T_a portion contains abundant mudstone intraclasts oriented parallel to bedding (red arrows). The T_b part is indicated by dashed white lines. White arrow indicates stratigraphic up direction. (B) Close-up photo of A showing basal very coarse-grained T_a sandstone overlain by a darker-brown planar laminated medium-grained sandstone. (C) Same bed in (A) and (B) but ~ 10 north of (B). Note that the basal T_a part is absent, and only the overlying T_{bde} part remains. (D) Lithological (left) and facies (right) make-up of scour dominated units in the FIC.

IV.II.VI **Mass Transport Deposits**

Due to the rarity of slides (facies 5) and, to a lesser extent, debris flows (facies 4) in the FIC, these units have been combined into mass transport deposits (MTDs), which make up 7% of the FIC.

IV.II.VI.I Carbonate-rich debris flows

Carbonate-rich debris flows (calcodebrites; CD) are uncommon in the FIC (~ 6%) and frequently occur in association with calciturbidite horizons. At CC there are four < 7 m-thick calcodebrites (~ 5% of stratigraphy), one in each of CT 1, 2 and 3, with the last directly overlying BST 2 at CC-N. Typically these debrites can only be traced laterally for < 200 m, but the calcodebrite associated with CT 2 (CD 2) is correlatable over > 1 km. These strata consist predominantly of matrix-supported conglomerate with a dark grey, olive-green, or reddish-orange calcilutite matrix with abundant, but dispersed, cm- to m-scale clasts of granule- to pebble-sized white quartz sand, mudstone, yellow to yellow-orange calcilutite and reddish-orange siliceous calcarenite fragments (Figure 2.8). Some of the finer-grained calcilutite clasts contain microbial laminae (Figure 2.8E), suggesting they were sourced from a shallow-marine carbonate platform. Like the calciturbidite horizons, calcodebrites at CC and HS can be differentiated based not only on stratigraphic position, but also on appearance in outcrop. The matrix of CD 1 varies from dark-grey to black siliciclastic mudstone to greenish-brown calcilutite and hosts abundant elongated dark grey siliciclastic mudstone clasts with rare cm-scale calcarenite clasts. Also, there are abundant dispersed quartz grains, which gives this debrite its distinctive “starry night” appearance. CD 2 has an olive-green to dark brown calcilutite matrix with many mm- to cm-scale carbonate and mudstone clasts, rare coarse-grained, light-dark striped siliceous calcarenite clasts (striping similar to bacon sandstone strata) and boulder sized carbonate clasts. CD 3’s matrix is dark-red, and therefore the colour of most siliceous calcarenites in the FIC, and contains abundant granule- to pebble-sized quartz grains, yellowish-grey to yellow carbonate clasts and grey mudstone clasts that are elongated parallel to bedding (2-55 cm along their long-axis) (Figure 2.8H).

At MR calcidebrites are thicker (up to 14.5 m, ~ 8% of FIC stratigraphy), more laterally extensive (all can be confidently correlated over > 600 metres) and contain substantially larger clasts (up to 3 metres wide) than those at CC and HS (Figure 2.8). Their lithological composition is similar to those at CC, but their matrices are typically a darker greenish-brown to brown colour. There is also a greater abundance of microbialite and oolite clasts relative to CC (Figure 2.8E-F). Like at CC the two lowermost CDs at MR occur in close stratigraphic proximity to CT 1 and CT 2, respectively. However, the two at MR underlie the calciturbidite units instead of occurring within them, suggesting that they do not correlate with those at CC (Figure 2.8A-F; Figure 2.10; Figure 2.11). A third, ~140 cm-thick CD crops out exclusively in the siliciclastic mud-rich unit above 123-2 at MR (Figure 2.8G). This debrite is composed of a dark-grey to grey-brown calcilutite matrix with dispersed cm-scale siliceous calcilutite and siliceous calcarenite clasts. However, two larger clasts are present, specifically a 43 cm-thick red T_b siliceous calcarenite at the base of the debrite at MR-N in which planar laminae are at an angle to depositional strike and a 1.5 m-wide yellow limestone clast near the top of the debrite at MR-S.

IV.II.VI.II Slides

Slides are even rarer in the FIC, making up ~ 1% of stratigraphy, and, much like debrites, occur in close stratigraphic association with calciturbidite horizons. At HS there is a < 1 m-thick slide unit in CT 1 consisting of an intensely-sheared, dark grey to black siliciclastic mudstone that can be traced laterally for > 200 m. At MR a > 8 m-thick slide scours CT 2 and extends across the entire outcrop (~600 m). The basal portion of the slide consist of sharply-bounded blocks of black calcidebrite and thin-bedded, upper division turbidites (both calcilutite and siliciclastic mud-rich) with bedding that strikes at an angle to the depositional strike of the underlying CT 2 unit. Fine-grained turbidite blocks commonly show a centre-to-margin increase in the degree of deformation. 123-1 abruptly overlies this slide.

IV.II.VI.I Interpretation

Calcodebrites are interpreted to have been deposited from carbonate-rich debris flows, likely associated with upslope failure. The abundance of siliceous calcarenite and calcilutite clasts, particularly those with microbial laminae or ooids, suggests that they originated from a shallow-marine carbonate platform. Fine-grained mud-rich carbonate sediment that make up the matrix was similarly sourced from the platform, but possibly also the upper slope.

Slides also suggest upslope failure and subsequent downslope movement of previously deposited slope sediments along discrete failure planes (Arnott, 2010). Initial failure could have been caused by myriad factors, including high rates of sedimentation, sediment loading, seismic activity, excess pore pressure, or eustatic change (e.g. Spence and Tucker, 1997; Locat and Lee, 2002; Canals et al., 2004).

The uncommon occurrence of MTDs in the FIC suggests that conditions on the slope were generally gravitationally stable, except during periods of calciturbidite deposition. The calciturbidites are interpreted to have been deposited during times of elevated sea level when the shallow-water carbonate platform was at its greatest extent (Navarro, 2016; Chapter 3). These conditions contrast what is reported from other, mostly siliciclastic deep-marine successions, MTDs are most common during falling sea level (FSST) (e.g. Wallace, 2004; Arnott et al., 2011; Navarro, 2016; Terlaky et al., 2016; Bergen, 2017). The upward growth of the platform, and subsequent shedding and rapid cementation and lithification of carbonate detritus onto the slope would have likely steepened the slope beyond what is typical for siliciclastic slopes causing it to become gravitationally unstable (e.g. Hedberg, 1970; Grammer et al., 1993; Ross et al., 1994). Additionally, MTDs in the FIC, particularly in CC-HS, are substantially thinner and less laterally extensive than those observed in modern carbonate settings, where they are up to a few hundred metres thick and tens of kilometres wide (e.g. Mulder et al., 2012a; Jo et al., 2015; Pincipaud et al., 2015; Reijmer et al., 2015). This disparity,

however, could be the result of proximity to the main conduit for basinward-flowing turbidity currents (see discussion).

Part V Discussion

The correlations of the various stratigraphic units in the FIC across the study areas (Figure 2.22) reveal a number of important observations: 1) many units can be confidently correlated across the ~ 20 km-wide study area, specifically CT 1-3, BST 1, PL 1, both 123 units and two siliciclastic-rich distal levee units; 2) the FIC at MR is ~ 70 m and ~ 100 m thicker compared to CC and HS, respectively, much of the difference being the thickness of siliciclastic-rich distal levee deposits (Figure 2.23A-B); 3) channel complexes are more common at CC-HS; 4) MTDs are more common, thicker and more tabular at MR. Based on these differences in stratal make-up and architecture, the FIC can be divided into lower and upper parts separated by BST 1.

The lower part makes up ~ 20% of FIC stratigraphy (Figure 2.23C) and is characterized by units that, for the most part, correlate across the study areas (i.e. CT 1 and BST 1). This causes the proportion and lithological makeup of the different stratal elements to be about the same and the stratigraphy to be more or less sheetlike. In contrast, stratigraphy in the upper part (upper 80% of FIC) is spatially discontinuous and lithologically more heterogeneous (Figure 2.23D). For example, at CC and HS channel complexes are more common, whereas siliciclastic mud-rich distal levee units make-up most of the section at MR. Two of these siliciclastic mud-rich units at MR are ~ 60 m thick compared with a maximum of only 12 m at CC and HS. Additionally, mass transport deposits are thicker and more abundant at MR (Figure 2.23A-B).

The stratal uniformity in the lower part of the FIC and its comparative absence in the upper part suggests temporal changes in the patterns of sediment transport and deposition. The lateral continuity of units in the lower part, particularly channel complexes, suggests that channel belts were initially mobile and wandered across much of the study area (Figure 2.24A). However, deposition of

BST 1 was followed by a dramatic change in the patterns of sediment transport and deposition, which is manifest as marked discrepancies in stratal architecture and thickness between the study areas. Specifically, CC and HS were located more proximal to the primary transport fairway for high-energy turbidity currents transiting the slope on their way to the basin floor, evinced by two up to 20 m-thick channel complexes (BST 2 and BST 3). In contrast, coeval deposition in MR was marked by the accumulation of thick successions of siliciclastic-rich distal levee deposits (Figure 2.24B). Furthermore, the FIC is thinnest at HS (~ 160 m), slightly thicker at CC (~190 m), but substantially thicker at MR (~ 260 m) which also lacks channel complexes above BST 1. These spatial differences in stratal thickness suggests greater sediment bypass in the more channel prone CC and HS study areas compared with higher rates of net sedimentation on the adjacent levees at MR. Also, the temporal (i.e. stratigraphically upward) change from wide channel belts to more laterally confined transport fairways has been reported in a number of siliciclastic slope systems (e.g. Hubbard et al., 2008; Cross et al., 2009; Labourdette and Bez, 2010; Di Celma et al., 2011; Hubbard, 2017), suggesting that the temporal evolution of channel belts in mixed systems is similar to their siliciclastic counterparts.

Spatial differences in sediment deposition and bypass in the upper part of the FIC can also explain the differences in the abundance, thickness and lateral extent of MTDs between the study areas. The more common occurrence of channel deposits at CC and HS indicates that high-energy turbidity currents were a common occurrence in these areas. Successive flows would have acted to establish an equilibrium slope gradient that would promote sediment bypass and thereby minimize net sedimentation – a process termed autosuspension (Parquet et al., 1986). In contrast, adjacent to the main transport fairway (i.e. Milk River), the paucity of slope modifying flows resulted in higher rates of net sedimentation over the adjacent levees and accordingly accumulation of a comparatively thicker section. This, in turn, increased the local slope which led to enhanced gravitational instability and mass wasting (e.g. Yose and Heller, 1989; Prather, 2000; Leynaud and Vanneste, 2009; Olafiranye et al., 2013; Hubbard, 2017), which at MR is manifest as common mass transport deposits. The comparative

rarity of MTDs at CC and HS is probably a combination of a more shallowly inclined slope along the main transport corridors, but also erosion by the more common throughgoing high-energy turbidity currents.

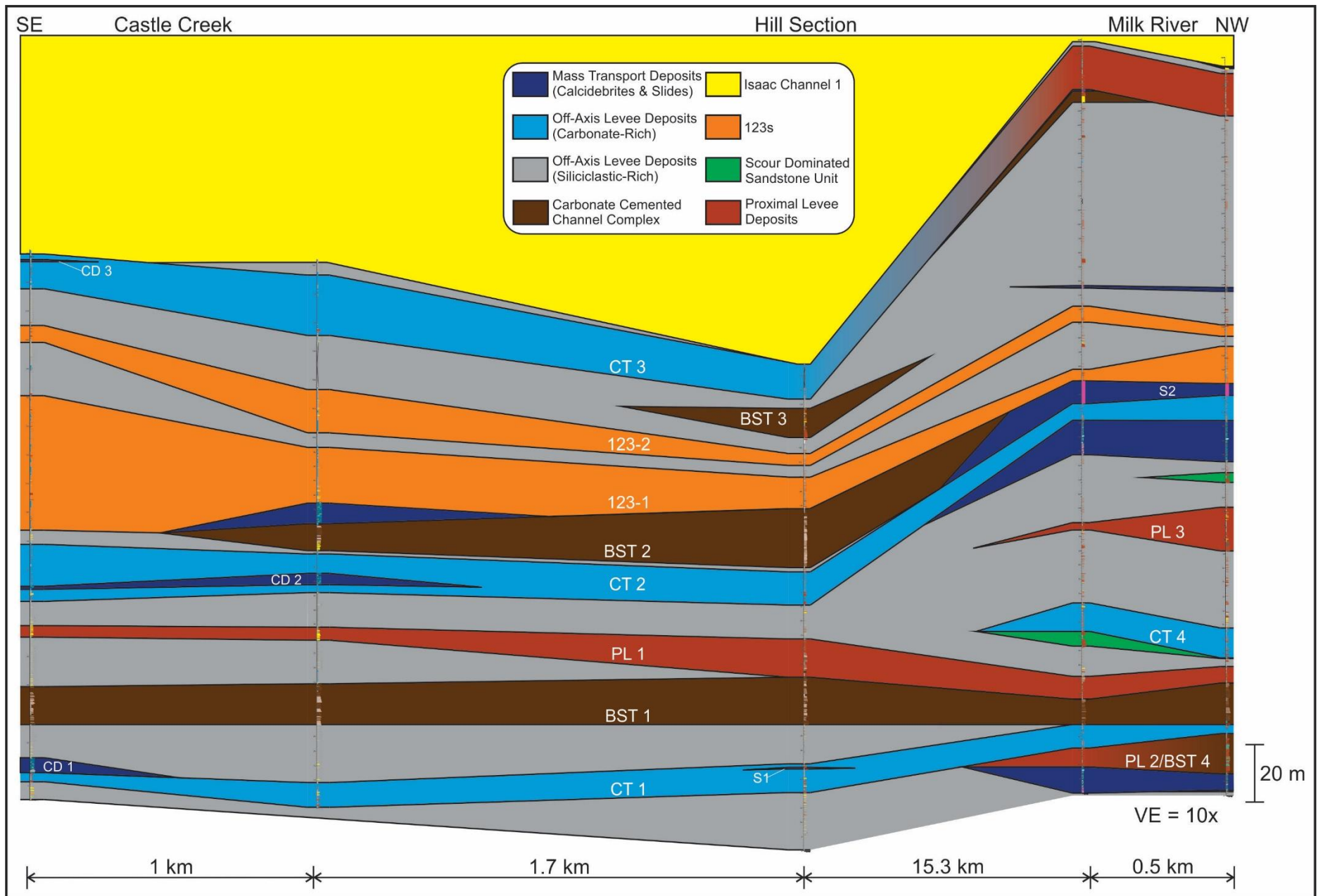


Figure 2.22: SE to NW correlation from CC-S to MR-N showing the spatial distribution of stratal elements in the FIC; datum is the base of BST 1; unit names in white. Several units can be confidently correlated across the study areas based on their distinctive lithological make-up and stratigraphic position – these include three calciturbidite horizons, BST 1 and the two 123 units.

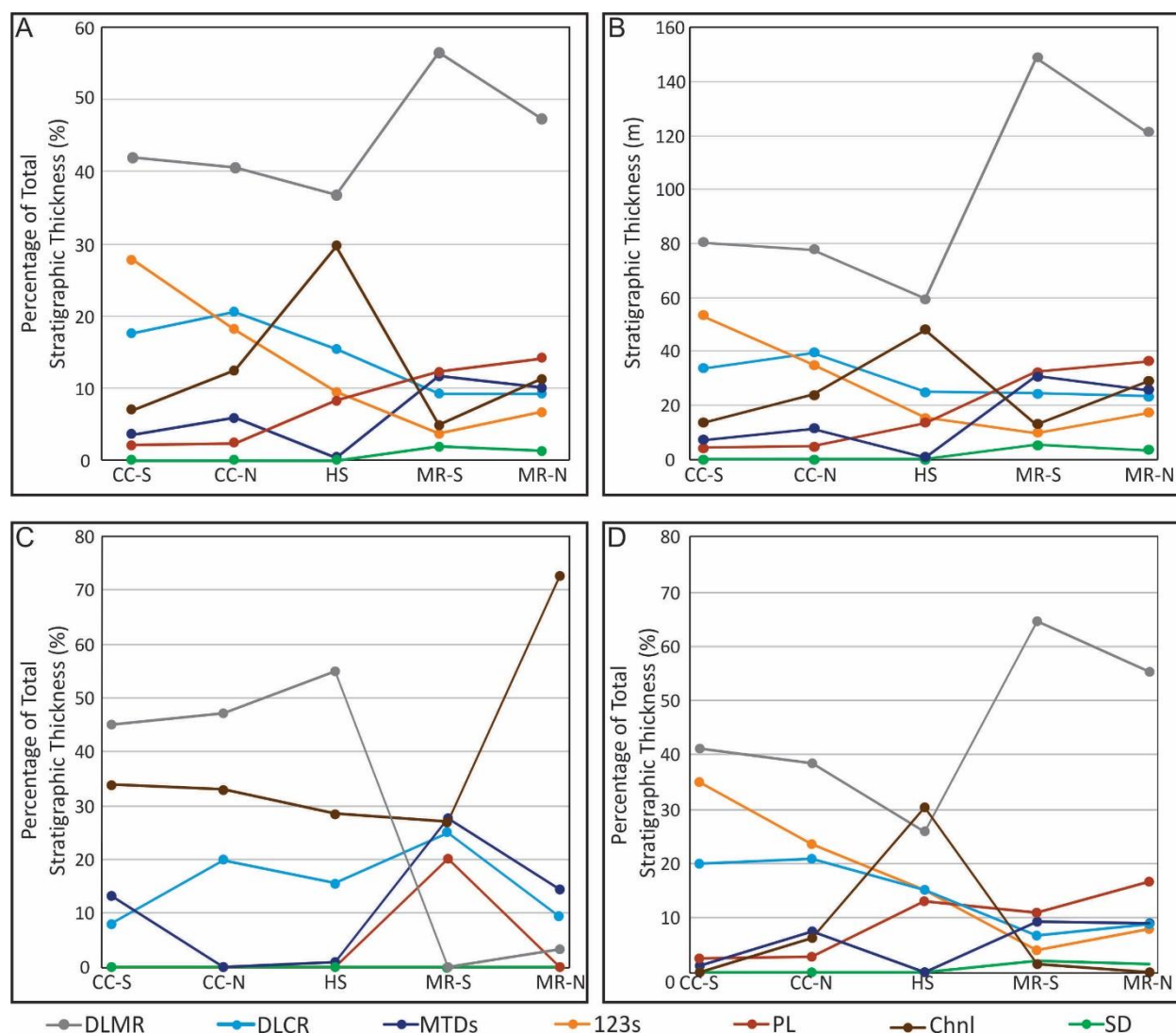


Figure 2.23: Abundance of stratal elements in the study areas. DLMR = siliciclastic mud-rich distal levees; DLCR = siliceous calciturbidite-rich distal levees; MTDs = mass transport deposits; PL = proximal levees; Chnl = channel complexes; SD = scour dominated units. (A) Stratal element percentage relative to total thickness of the FIC. (B) Cumulative stratigraphic thickness of individual stratal elements identified in the FIC. (C) Stratal element percentage relative to total thickness of the FIC below the top of BST 1. (D) Stratal element percentage relative to thickness of the FIC above BST 1. Note here that calciturbidite units appear to be less common at MR, but that is because CT 3 at MR is classified as a proximal rather than a distal levee.

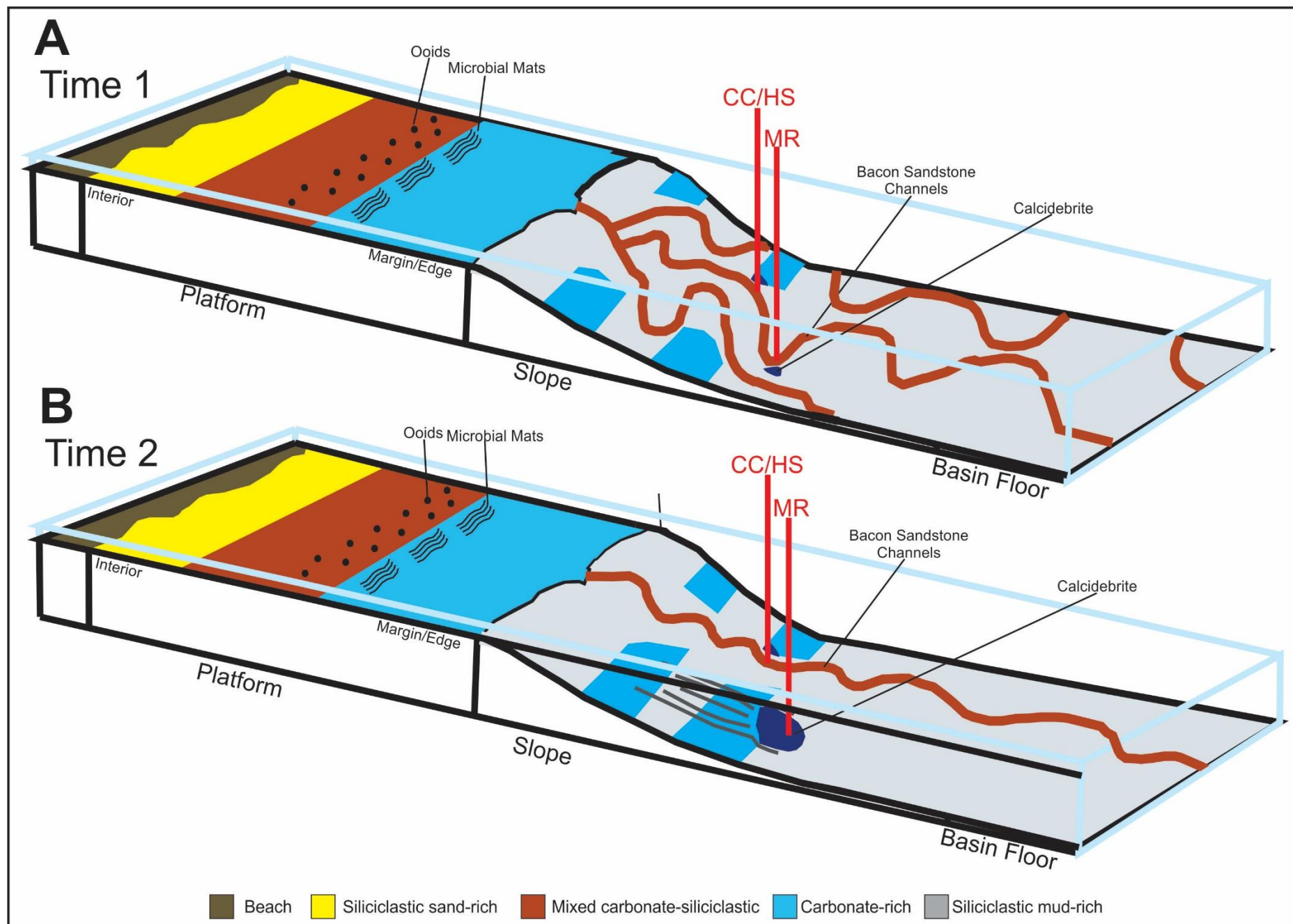


Figure 2.24: Schematic model illustrating the temporal evolution of the FIC. (A) Based on stratal element composition and thickness, channel belts were initially highly mobile and wandered across much of the study area during early FIC deposition (below the top of PL1; ~ 20% of FIC). (B) After deposition of PL1 the spatial patterns of transport and deposition changed with channel complexes becoming confined to the Castle Creek and Hill Section areas (particularly HS). This resulted in significant bypass and deposition of a comparatively thinner stratal pile compared to Milk River. In contrast, the thicker succession at Milk River indicates higher net sedimentation, which in this case was dominated by fine-grained off-axis turbidites and thicker, more common mass transport deposits. The shallow-marine transition from siliciclastic sand-rich to mixed carbonate-siliciclastic to carbonate-rich is based on models near the siliciclastic-dominated end of the spectrum from Schwartz et al. (2018), particularly D'Agostini et al.'s (2015) model for the Abrolhos shelf.

Chapter 3: Geochemical Evolution of an Ediacaran Mixed Carbonate-Siliciclastic Continental Slope System, Canadian Cordillera, British Columbia

Part I Introduction

Interest in the Neoproterozoic Era (1000-542 Ma) has grown steadily over the past few decades as it represents a pivotal period in change of the Earth systems that is unmatched by anything during the Phanerozoic. The Neoproterozoic saw the assembly and breakup of a supercontinent (Rodinia), three potentially global glaciations following an ~ 1.5 billion-year-long interglacial period, the return of banded-iron formations after a 1 billion-year hiatus, globally-correlatable carbonates formed during early interglacials in response to rapid climate change, and a transition to a more oxic atmosphere that may have stimulated the evolution of metazoan life (Ediacaran biota), which eventually led to the development of calcified skeletons that became omnipresent during the Cambrian explosion (Kirshvink, 1992; Narbonne et al., 1994; Hoffman et al., 1998; Hoffman and Schrag, 2002; Xiao and Kaufman, 2006; Halverson, 2006; Fike et al., 2006; Fairchild and Kennedy, 2007).

The stratigraphic record has been central to reconstructing the Neoproterozoic Era, but unlike the Phanerozoic, regional and global correlations are hampered by the lack of biostratigraphic control (Halverson, 2006; Fairchild and Kennedy, 2007; Halverson et al., 2010). To circumvent this, scientists in the late 1970 and early 1980s began correlating Proterozoic strata based on secular changes in chemical components, namely (carbon) isotopes in carbonate rocks ($\delta^{13}\text{C}_{\text{carb}}$) (e.g. Veizer and Hoefs, 1976; Veizer et al., 1980; Knoll et al., 1986; Magaritz et al., 1986; Tucker, 1986), a technique that had been used to study Phanerozoic strata as far back as the 1950s (Craig, 1953).

The Neoproterozoic carbon isotopic record is marked by highly positive $\delta^{13}\text{C}_{\text{carb}}$ values with five globally correlatable, large-amplitude negative excursions, two of which preceded major glaciations (Figure 1.17; Knoll et al., 1986; Kaufman and Knoll, 1995; Hoffman et al., 1998; Halverson et al., 2005; Halverson, 2006; Halverson et al., 2010). Three of these isotopic anomalies occurred

during the ~ 94 myr Ediacaran Period, which spans from the top of Marinoan diamictites (~ 635 Ma) to the Precambrian-Cambrian boundary (~542 Ma), and includes: i) global cap carbonates that overlie Marinoan glacial deposits (e.g. Kennedy, 1996; Hoffman et al., 1998; James et al., 2001; Jiang et al., 2007; Halverson et al., 2005), ii) the Shuram-Wonoka anomaly (< 580 Ma) (e.g. Burns and Matter, 1993; Calver, 2000; Halverson et al., 2005; Le Guerroué et al., 2006ab; Jiang et al. 2007; Husson et al., 2012) and iii) the terminal Proterozoic (Halverson et al., 2010). The timing and duration of the Shuram-Wonoka anomaly has been widely debated with some authors attributing it to the Gaskiers Glaciation, whereas others interpret it to be a short-lived post-Gaskiers event (Jiang et al., 2007; Tahata et al., 2013; MacDonald et al., 2013), or alternatively a 50 myr long interglacial (Le Guerroué et al., 2006ab). Despite this uncertainty, it is generally accepted that this excursion is closely associated with the oxygenation of the deep ocean and Earth's atmosphere (e.g. Narbonne et al., 1994; Fike, 2006) and the subsequent evolution of complex Ediacaran biota with elevated metabolic needs (MacDonald et al., 2013). Furthermore, other studies have reported a fourth negative carbon excursion, known in South China as EN2 (Jiang et al., 2007; McFadden et al., 2008; Tahata et al., 2013), which has potential correlatives in India (Jiang et al., 2002; Kaufman et al., 2006) and in Death Valley, U.S.A. (Corsetti and Kaufman, 2003; Kaufman et al., 2007). However, in spite of being widely recognized, the explanations for these large-amplitude Ediacaran negative carbon excursions and the subsequent oxidation of the Earth's atmosphere are varied and include: methane hydrate destabilization (Kennedy et al., 2001; Jiang et al., 2003, 2006a), upwelling and mixing of $\delta^{13}\text{C}$ depleted deep-marine water during deglaciation (Kaufman et al., 1991; Grotzinger and Knoll, 1995; Knoll et al., 1996; Shields, 2005), remineralization of a large dissolved organic carbon (DOC) reservoir through sulfate reduction during regression (Rothman et al., 2003; Fike et al., 2006; Jiang et al., 2007) and authigenic carbonate precipitation in near-surface sediments (Schrage et al., 2013; MacDonald et al., 2013).

The Neoproterozoic Windermere Supergroup (WSG) in the southern Canadian Cordillera (SCC) represents a ~ 9 km-thick rift-to-drift succession associated with the breakup of Rodinia and

subsequent development of a passive margin, which is dominated by siliciclastic-rich deep-marine strata (Figure 2.1A-B; Ross and Arnott, 2007). Despite its generally siliciclastic composition, the WSG hosts five carbonate-rich intervals, four of which overlie strata correlated to the Marinoan glaciation. Stratigraphically upward, these four intervals are: deep-marine carbonates of the Old Fort Point Formation (OFP), the first and second Isaac carbonates (FIC and SIC), and shallow-water platform carbonates of the Cunningham Formation (Figure 2.1B; Ross and Arnott, 2007). In spite of the large number of $\delta^{13}\text{C}_{\text{carb}}$ studies on Neoproterozoic strata worldwide, these intervals in the WSG in the SCC have been generally understudied, although the work of Smith (2009) on the OFP is a notable exception.

This study focuses on the first Isaac carbonate, a mixed carbonate-siliciclastic base-of-slope succession whose base is located ~ 150 m above the contact between the Isaac Formation and basin floor deposits of the Kaza Group. This succession and the SIC were first recognized in regional mapping of the WSG and subsequently have been used as stratigraphic markers for the lower and upper Isaac Formation, respectively (Ross and Murphy, 1988; Ross and Ferguson, 2003; Ross and Arnott, 2007) and therefore held great promise for future isotope chemostratigraphic analysis (Ross et al., 1989; Ross et al., 1995). Despite its recognition dating back to the late 1980s, the FIC has been paid little sedimentological or geochemical attention, until the PhD work of Navarro (2016), which represented its first detailed sedimentologic and stratigraphic study at Castle Creek and Mount Quanstrom. This study builds on Navarro's work and improves our understanding of the conditions that led to the deposition of this sedimentological outlier in the deep-marine Windermere basin. New and archival data from the thoroughly studied Castle Creek study area (e.g. Ross and Arnott, 2007 and references therein) and new data from the Hill Section and Milk River study areas (Figure 2.1C) were used to study the $\delta^{13}\text{C}_{\text{carb}}$ signal in the FIC in an attempt to correlate the deep-marine WSG to potentially coeval shallow-water Windermere strata in the Mackenzie Mountains and other Ediacaran

basins worldwide, and to better understand the oceanographic and climatic conditions that led to the development of the FIC.

Part II Geologic Setting

II.I Windermere Supergroup

The Neoproterozoic WSG is a succession of mostly sedimentary rocks exposed locally from northwestern Mexico and the western United States and continuously throughout most of southwestern Canada through to the Yukon-Alaska border region; a strike length of just over 4000 km (Figure 2.1A; Ross and Arnott, 2007). Outcrops in Mexico and the USA comprise continental and shallow marine strata (Link et al., 1993), whereas those in the SCC consist of superbly exposed successions of deep-marine, siliciclastic-dominated metasedimentary rocks (Campbell et al., 1973; Ross et al., 1995), and in the Mackenzie Mountains of northwestern Canada preserve extensive carbonate-rich upper continental slope and continental shelf facies (Aitken, 1991a,b; Narbonne et al., 1994).

In the SCC deep-marine rocks of the WSG are well exposed in thrust sheets of the Foreland fold and thrust belt and the Omineca belt, which are separated by the Southern Rocky Mountain Trench (SRMT) (Ross and Arnott, 2007; Smith, 2009) and represent two of the five NW-SE trending morphological belts that make up the Late Jurassic to Early Tertiary Canadian Cordillera (Monger et al., 1982; Ross and Arnott, 2007; Smith, 2009). Deformation associated with the Mesozoic Cordilleran Orogeny caused the deep-marine outcrop belt of the WSG to be substantially shortened, but also exposed over an area of about 35,000 km², which, if palinspastically reconstructed, equates to ~ 80,000 km², making it one of the world's largest ancient turbidite systems (Ross and Arnott, 2007).

Formation of the WSG proceeded in two phases (Stewart, 1972): a lower, syn-rift phase associated with Rodinia breakup, which in southeastern B.C. consists of laterally discontinuous glacial diamictites associated with the Sturtian glaciation (Toby Formation) and volcanics (Irene Formation) deposited in isolated rift basins and an upper, 5-7 km-thick, post-rift to drift sequence that in the

Cariboo Mountains of east-central British Columbia comprises laterally continuous sedimentary units of the Kaza and overlying Cariboo groups, deposited along a passive continental margin developed in response to thermally-driven subsidence (Figure 2.1B; Ross, 1991; Ross and Arnott, 2007). Basin floor deposits of the Kaza Group make up the basal 2-4 km of the post-rift succession, and has been subdivided into a lower mudstone-dominated part overlain by feldspathic, coarse-grained sandstone rich middle and upper parts, interpreted to represent deposition from unconfined flows on the distal to progressively more proximal basin floor (Ross and Murphy, 1988; Ross, 1991; Ross et al., 1995; Meyer and Ross, 2007; Terlaky and Arnott, 2014; Terlaky et al., 2015). The Middle and Upper Kaza groups are separated by the regionally extensive OFP, which represents a distinctive lithological and geochemical marker that aids in regional correlation (Ross and Arnott, 2007; Smith, 2009; Smith et al., 2011; Smith et al., 2014a,b). In the Monkman Pass area (Figure 2.1A) glacial diamictite of the 350-2000 m-thick Vreeland Formation underlie the OFP, changing laterally to coarse-grained turbidites of the Kaza Group to the west and south, and represent the resedimentation of shallow-marine glacial deposits interpreted to be associated with the Marinoan glaciation (650-635 Ma) (McMechan, 1987; McMechan and Thompson, 1995; McMechan, 2000; Smith et al., 2011).

The overlying, up to 5-km thick, Cariboo Group is subdivided into three stratigraphic units. The up to 2.4 km-thick Isaac Formation conformably overlies the Kaza Group and is composed primarily of mudstones (levee deposits) that encase laterally discontinuous, ~ 20-100 m-thick, coarse-grained sandstone units interpreted to represent a series of stacked leveed-channel complexes (Ross and Arnott, 2007; Arnott et al., 2011). A base-of-slope setting is interpreted based on their mud-dominated nature and the commonality of mass movement deposits indicating gravitationally unstable slope conditions. In the Isaac Formation are two regionally continuous, lithologically similar carbonate-rich units, the FIC and SIC (Figure 2.1B; Ross et al., 1989; Ross, 1991; Ross et al., 1995; Ross and Ferguson, 2003). The FIC (this study) is characterized by three laterally extensive, decametre-thick units of very thin- to thin-bedded carbonate-rich turbidites (hereafter termed calciturbidites) and

very fine- to fine-grained siliceous calcarenites intercalated locally with up to ~ 20 m-thick successions of granule conglomerate to medium-grained sandstone, all with a matrix of neomorphosed carbonate; carbonate-cemented sandstones and conglomerates are hereafter termed siliceous calcarenites and siliceous calcirudites (see Part III for an overview of the lithological nomenclature used here), respectively. Interbedded with these carbonate units are up to ~ 60 m-thick accumulations of siliciclastic mud-dominated, fine-grained, thin-bedded turbidites. The SIC has a makeup similar to the FIC, but with local intercalated sheets of quartz silicarenite (Ross and Ferguson, 2003).

Conformably overlying the Isaac Formation are upper-slope to shallow-marine, high-energy shelf deposits of the Cunningham (oolitic limestones with minor mudstones) and Yankee Belle formations (alternating limestone, siltstone, sandstone and shale) (Ross et al., 1995). The top contact of the Yankee Belle is a regional unconformity that separates the top of the WSG from Lower Cambrian rocks of the Yank's Peak Formation and is responsible for the absence of shallow-marine and continental WSG strata in eastern outcrops (Aitken, 1969; Ross et al., 1995; Ross and Arnott, 2007).

The dominance of siliciclastic strata, lack of volcanic rocks, and the absence of fossils throughout the WSG in the SCC results in sparse geochronological control (Ross and Arnott, 2007). Nevertheless, the timing of Windermere deposition can be constrained between 740 ± 36 to 728 ± 8 Ma (U-Pb) from crystalline basement rocks that underlie the Canadian Cordillera (Evenchick et al., 1984; Parrish and Scammell, 1988; McDonough and Parrish, 1991) and a 569 ± 4.6 Ma (U-Pb) date from volcanic rocks associated with rifting that resulted in the unconformity that caps the Yankee Belle Formation (Colpron et al., 2002). The only date from within the WSG stratal pile in the SCC is a rhenium-osmium (Re-Os) age of 607.8 ± 4.7 Ma from organic mudstone in the Geikie Siding Member (GSM) of the OFP (Kendall et al., 2004).

II.II Study Areas

This study focuses on the Castle Creek (CC; divided into Castle Creek south (CC-S) and north (CC-N)), Hill Section (HS) and Milk River (MR; divided into Milk River south (MR-S) and north (MR-N)) study areas in the Cariboo Mountains of the SCC (Figure 2.1C). Bedding in all areas is near-vertical ($\sim 75\text{-}89^\circ$) due to their location on the upright limbs of two regional anticlines in the Isaac Synclinorium and Premier Anticlinorium (Campbell et al., 1973; Murphy 1987a,b; Murphy et al., 1995; Reid et al., 2002). Rocks have been metamorphosed to lower greenschist facies, but primary sedimentary structures remain well preserved. Also, all areas are situated near rapidly retreating, north-facing glaciers, resulting in glacially polished and vegetation free exposures that can be easily measured and beds traced laterally for hundreds of metres (Ross and Arnott, 2007). However, outcrop quality at MR is diminished compared to CC due to its smaller area and slightly higher metamorphic grade, which generally obscures fine-scale primary sedimentary structures.

Part III Lithological and Stratal Make-Up of the FIC

III.I Facies

Rocks that make up the FIC comprise five facies: (1) Structureless, normally-graded sandstone, (2) Traction-structured sandstone, (3) Fine-grained, mud-rich deposits, (4) Poorly-sorted, calcilutite-rich deposits, and (5) Internally sheared and deformed strata. Ingram's (1954) classification for bed thickness is used in this study: very thin-bedded (1-3 cm), thin-bedded (3-10 cm), medium-bedded (10-30 cm), thick-bedded (30-100 cm) and very thick-bedded (> 100 cm). Classification of carbonate strata follows Grabau (1904) with grain-size boundaries from Folk (1962): lutite is < 0.062 mm, arenite ranges from 0.062-1 mm and rudite is > 1 mm. In this classification the first word corresponds to the lesser mineralogical component, while the prefix of

the second word refers to the more component one and the suffix of the second word is based on the grain-size (Table 2.1). For example, a bed composed predominantly of mud-sized calcite crystals with lesser silt-sized quartz grains is termed a siliceous calcilutite. Grain size follows the standard geometric scale of Wentworth (1922), whereas turbidite classification follows the model of Bouma (1962). Table 3.1 summarizes the main stratal attributes of each lithofacies and their interpreted formative processes, while Figure 3.1 contains representative photographs of each. For more detailed description and interpretation of facies and depositional elements see chapter 2.

Facies	% of FIC make-up	Lithology, texture and physical sedimentary structure	Bed thickness, geometry and contacts	Depositional processes	Bouma divisions	Photo(s)
F1: Structureless, normally graded sandstone	27 %	1A - dark- to light-brown, reddish brown, brownish-grey and greenish-brown siliceous calcarenites with rare beige silicarenites. -Massive beds, coarse-tail graded medium-grained sandstone to pebble conglomerate (siliceous calcirudite). -Commonly fine upwards. -Mudstone and calcilutite intraclasts are common in lower portion of beds, 1B: -Reddish-orange siliceous calcarenites, uncommon beige silicarenites -Massive base (<50% of total bed thickness), transition upwards to traction structured sandstones and a siltstone / mudstone cap. -Rare mudstone clasts	1A: -Bed thickness ranges from 20-240 cm (average = 90 m). -Beds commonly amalgamated -Flame structures observed, but are uncommon. 1B: -Very thin- to thick-bedded, average thickness = 20 cm. -Bases typically planar and sharp.	-Massive portion: rapid suspension fallout from sand--rich, high-density turbidity currents (Lowe, 1988; Arnott and Hand, 1989; Mulder and Alexander, 2001). -Tractional sedimentary structures inhibited by high sediment concentration -T_c/T_c : reworking of T_a by tail of turbidity current (Arnott and Hand, 1989) -T_{de}: deposited by dilute tail of flow or hemipelagic fallout (Lowe, 1982; Meyer, 2004)	1A: T_a. -Uncommonly the upper ~ 5-10% of a bed will contain T_b and/or T_c. -Capped by a T_{de} where not eroded by a later flow. 1B: T_a (< 50% of bed), grading upwards to T_b, T_{bde}, T_c, T_{cde}, or T_{de}.	1A: 3.2 A-C

F2: Traction structured sandstone	13%	-Traction structured reddish-orange to orange siliceous calcarenite and uncommon beige silicarenite. -Fine- to medium-grained base, commonly fines upwards -Capped by thin mudstone/siltstone cap -Ripple horizons: typically 1-3 non-climbing sets; rare multiset or climbing ripples. -Four high-angle cross-stratified sandstones observed	-Very thin- to medium-bedded -Average thickness = 10 cm. -Basal contacts generally sharp and planar.	-Deposited from waning sand-rich turbidity currents. -Planar lamination indicative that near-bed conditions were not conducive to the establishment of angular bedforms (e.g. Southard and Boguchwal, 1990; Venditti et al., 2005; Sumner, 2008; Tilston et al., 2015) -Ripples form when hydrodynamic instabilities required for bed-defect development formed and flow was in ripple stability field. -Two high-angle cross-stratified beds are rare examples of deep-marine dunes; other two represent high-angle fill of bed surface scours (e.g. Arnott and Al-Mufti, 2017).	$T_b, T_{bc}, T_{bcde}, T_{be}, T_c, T_{cde}, T_{ce}.$	3.2 D-F
F3: Mud-rich deposits	51%	-Siliceous calcilutite (brownish-green, greenish-black and black), siliciclastic siltstone and mudstone (medium grey) -Typically structureless, occasionally thinly laminated. -Some beds contain a < 1 cm-thick basal siliceous calcarenite.	-Very thin- to thin-bedded (average = 2 cm) -Sharp, planar basal contacts. -Upper contact sometimes undulatory due to local erosion.	-Bed-load deposition beneath low-concentration, low-energy, fine-grained turbidity currents or suspension settling from dilute turbidity current tail. -Very thin structureless sandstone beds indicate flows with low sand and abundant mud/silt content	Predominantly T_e or T_{de} , occasional $T_{ade}, T_{abde}, T_{cde}.$	3.2 G-H

		-Typically stack to form Dm-thick stratal packages (siliciclastic mud-rich or siliceous calcilutite rich) -Less commonly form interbeds in sandstone-rich strata.				
F4: Poorly sorted calcilutite-rich deposits	6%	-Greenish-brown, reddish-brown, brown, dark grey and black structureless calcilutite matrix (locally grey siliciclastic mud) with abundant dispersed clasts -Clasts: mudstone fragments, up to pebble-sized quartz grains, yellow calcilutites ($\pm$ microbial laminations), orange oolites and siliceous calcarenites -Clasts range from mm-scale to > 100 m -These beds/units commonly associated with calciturbidite horizons	-Beds range from 0.61-14.42 m-thick, typically lenticular. -Sharp and irregular contacts	-Deposited from carbonate-rich debris flows initiated on a shallow-marine carbonate platform (evinced by photic zone growth forms and oolites). -Deposits are hereafter termed calcidebrites.		3.2 I-L
F5: Internally sheared and deformed Strata	1%	3 examples (all occur in association with calciturbidites and sometimes calcidebrites): 1) < 1 m-thick intensely sheared siliciclastic mudstone unit.	-Thickness ranges from 0.6-7.9 m and ~ 100 to > 600 m laterally. -Sharp and erosive basal contacts, often evinced by increase in internal ductile deformation of beds. -Irregular and sharp upper contacts.	-Slide deposits produced by upslope failure of the seabed. -Composed of semi-coherent blocks of thin-bedded turbidites and calcidebrites bounded by shear planes.		3.2 M-O

		2) Up to 8 m-thick unit with abundant black calcidebrite blocks truncated by blocks containing internally deformed thin-bedded, upper-division turbidites (T_{ce}). Blocks are oriented oblique to bedding. 3) ~ 60 cm thick greenish-brown siliceous calcilutite. Uncommon carbonate clasts in basal half of bed, upper half contains numerous deformed siliciclastic mud layers and discontinuous siliceous calcarenite beds.				
--	--	--	--	--	--	--

Table 3.1: Summary of deep-water facies (siliciclastic and carbonate) in the first Isaac carbonate



Figure 3.1 (previous page): Representative photographs showing the facies and depositional elements observed in the FIC. (A) The base of the first bacon sandstone channel complex at CC-S, which is made-up of multiple amalgamated, siliceous calcarenite and granular to pebble siliceous calcirudite (facies 1). White line marks the sharp, erosive basal contact. (B) Yellow microbially-laminated carbonate clast in a dark brown, massive siliceous calcirudite (facies 1) in a channel complex at MR-S. Carbonate (calcilutite and siliceous calcarenite (white arrows)) and mudstone clasts (yellow arrows), and gravel (~ 5 mm) quartz grains (red arrows) dispersed in a matrix of carbonate-cemented, upper medium-grained sandstone. (C) Conglomerate with cobble/boulder mudstone and carbonate clasts at the base of bacon channel complex 1 at HS. (D) T_{bde} , T_{cde} , and T_{bcde} (facies 2) turbidites in proximal levee deposits. (E) 123s at CC-S showing the characteristic repetitive stacking of the three packages. The lowermost siliceous calcarenite bed is dune-cross stratified and thickens and thins multiple times laterally before transitioning to planar stratified sandstone in both directions. (F) Close-up of a medium-grained orange-brown siliceous calcarenite in the 123s, underlain by a 3 and 2 package. The basal part of the bed is planar stratified (T_b), overlain by dune-cross stratification and ripple formsets (white arrows point to ripple crests) (Facies 2). (G) CT2 at HS (Facies 3). (H) CT3 at CC-N with multiple thin, interbedded red and darker coloured siliceous calcilutite beds. (I) Calcidebrite 1 (facies 4) at MR-S. Note the large carbonate clast and many smaller carbonate and mudstone clasts dispersed throughout the dark green calcilutite matrix. Rock hammer circled in red for scale. (J) Debrite stratigraphically above calcidebrite 2 at MR-N; upper and lower contacts are indicated by white lines. The basal reddish-orange siliceous calcarenite clast (white arrow) is planar stratified at an angle to bedding. Many smaller orange siliceous calcarenite clasts and finer carbonate clasts are spread throughout the darker matrix. (K) A yellow, microbially-laminated carbonate clast at the top of a calcidebrite at MR-N. (L) A yellowish-orange oolite clast (white arrow) within calcidebrite in (K). Camera case for scale in K and L (circled in red in K). (M) Thin slide (facies 5) at the top of CT1 at HS. The upper, internally deformed mudstone can only be traced laterally for ~ 50 m, but the dark brown siliceous calcarenite at the base can be traced across the full width of the outcrop (~ 230 m). (N) Slide at MR-N with basal and upper contacts indicated by black solid lines. A yellow dashed line indicates a deformed sandstone block and a red dashed line indicates a black calcidebrite block. (O) Photo of slide in (N), but taken from the edge of the debrite block outlined in (N). Here the debrite block is gone, giving way to, at least, three internally deformed siliciclastic mud-rich blocks (outlined by dashed lines) that are inclined to the depositional strike of the calciturbidite unit below (black beds). (P) T_{abde} reddish-orange to brown siliceous calcarenite below Calcidebrite 2 at MR-N. The T_a portion (Facies 1) contains abundant mudstone intraclasts elongated parallel to bedding and it pinches out a few metres north (along-strike) from this location, where only the overlying dark-brown T_b (Facies 2) is present. Based on its short lateral extent and abundance of mudstone clasts within, the T_a is interpreted as a scour.

III.II Stratal Elements

A stratal element is the two-dimensional outcrop expression of a three-dimensional sedimentary body bounded above and below by genetically related surfaces that formed in the same depositional setting (Pickering et al., 1989; Terlaky et al., 2015). The facies described above buildup seven stratal elements that then variously stack to form the FIC. These elements include: channel complexes, proximal levees, distal off-axis levees (which is further divided into siliciclastic- and carbonate-rich), “123” units, scour dominated units, debrites and slides (Table 2.2; Figure 3.2), of

which five of which are described below. The latter two are composed entirely of facies 4 and 5, respectively, and are described above.

III.II.I Channel Complexes

Slope channel complexes make up 12% of the total FIC stratigraphy and are dominated by amalgamated, thick- to very-thick-bedded, massive, medium- to very-coarse-grained carbonate-cemented sandstone (siliceous calcarenite) to granule conglomerate (siliceous calcirudite) (Facies 1; Figure 3.1A). The lowermost beds scour underlying stratigraphy and typically are the coarsest grained with abundant mudstone and carbonate clasts. Carbonate clasts consist of fine-grained yellow calcilutite, occasionally with algal laminations, and brownish-orange to orange siliceous calcarenite. (Figure 3.1B-C). Stratigraphically upward, beds become better graded and generally fine and thin to thin- to medium-bedded, massive or traction structured, siliceous calcarenite with siliciclastic mud caps (facies 1 and 2) and occasional fine-grained, siliciclastic mud-rich, thin-bedded turbidites (Facies 3). These beds stack to form 4-20 m thick channel complexes comprising one to three up to 6 m-thick channel fills, some of which can be correlated for > 3.5 km laterally. Channel complexes in the FIC are informally termed “bacon sandstones” because of their alternating dark-brown and light-brown to grey striping in outcrop (Figure 3.1A). This banding is due to variable amounts of micritic cement, specifically 47% on average in dark layers (siliceous calcarenite) and only ~ 10% in lighter layers (calcareous silicarenite). Importantly, changes in colour do not necessarily represent bed contacts, and in some cases a single bed contains multiple layers.

Four channel complexes were identified in the FIC. Three crop out in the HS and make up 30% of the stratigraphy. At CC channel complexes make up 7-12% of the stratigraphy and the basal complex (bacon sandstone 1 (BST 1)) occurs in each study area. At MR BST 1 extends across the study area, however, there is an additional ~14 m-thick complex beneath this unit that only crops out in MR-N (BST 4). Channel complexes comprise 5% and 11% of the stratigraphy at MR-S and MR-N, respectively (see chapter 2 for details and explanation of the distribution of channel complexes in each

study area). The lowermost complex at CC and HS (BST 1) is present in all study areas and here is used as the datum for stratigraphic correlation (Figure 3.2). All other channel units are more laterally restricted and crop out in only one or two of the study areas.

III.II.II Proximal Levees

Proximal levees make up slightly less than 6% of FIC stratigraphy and consist mostly of medium- to thick-bedded sandstone with abundant traction structures, including multiset ripples, that typically fine upward to a mud/silt cap (facies 1 and 2 = 80%; Figure 3.1D-E), or less commonly mud-dominated (siliciclastic and/or siliceous calcilutite) thin-bedded turbidites (Facies 3; 20%). Only one proximal levee unit, located just above bacon sandstone 1 (BST 1), can be correlated throughout the entire study area. At MR three additional proximal levee units crop out (Figure 3.2), one directly overlying calcidebrite 1 (PL 2) another ~ 15 m below calcidebrite 2 (PL 3) and the third being the uppermost calciturbidite horizon (CT 3). Here CT 3 is more siliceous calcarenite-rich and siliceous calcilutite-poor than at CC-HS, which is interpreted to reflect comparatively more proximal (levee) deposition. Both CT 3 and PL 3 consist of a series of alternating, sharply bounded, siliceous calcarenite- and mud-rich, very thin- to thin-bedded turbidites. The sandstone-rich intervals are up to 5 m-thick with individual very thin- to very thick-bedded, fine- to coarse-grained T_{ade} , T_{abde} , T_{abcde} , and T_{bcde} turbidites. Most beds are planar-stratified with some capped by an up to 5 cm-thick ripple cross-stratified unit (one or multiple sets). The mud-rich packages are up to 3.5 m-thick and are composed predominantly of very thin-bedded to thin-bedded T_{de} turbidites with occasional up to 30 cm-thick sandstones (T_{ade} and T_{bde}) and <2.5 cm-thick T_{ce} turbidites. Mud is typically siliciclastic, but in CT 3 packages are more siliceous calcilutite-rich than PL 3. PL 2 is ~ 6.5 m-thick and comprises six thick- to very thick-bedded T_{bde} and T_{bcde} (56-123 cm-thick) turbidites interbedded with thin-bedded T_{be}/T_{ce} turbidites (65:35 sandstone:mudstone ratio). Cropping out at the same stratigraphic level in MR-N is a ~ 10 m-thick interval of amalgamated, massive sandstone more characteristic of a channel fill (BST 4; 85:15 sandstone:mudstone ratio). However, 150 m laterally to the south (i.e. toward PL 2) strata consist

of interbedded, well-stratified T_{ade} and T_{de} turbidites (35:65 sandstone:mudstone ratio) that 250 m further laterally become siliciclastic mud-rich turbidites interbedded with less common very thin to thin-bedded T_a beds.

Due to their close association with units interpreted to be channel complexes, the sharp intercalation of medium-bedded sandstone and siliciclastic mud-rich, thin-bedded turbidite packages, and the common occurrence of multiset ripple cross-stratification, these units are interpreted to have been deposited on the margins immediately adjacent to the main sediment transport pathway (e.g. Khan and Arnott, 2011; Bergen, 2017). The sandstone-rich packages were deposited early during channel formation. Flows had a plug-like density structure (Tilston et al., 2015; Kneller et al., 2016) with its velocity maximum located above the height of the channel margins, resulting in significant overspill and deposition of medium- to thick-bedded, medium- to coarse-grained, sand-rich turbidites (de Leeuw et al., 2016; Bergen, 2017). With continued deposition the height of the channel-margin levees eventually exceeded the height of the velocity maximum, and as a result only the fine, upper parts of later flows overspilled and deposited very thin- to thin-bedded, fine-grained, mud-rich turbidites.

The abundance of sand-rich strata in PL 1 and PL 2 suggest that they were deposited on the proximal outer-bend levee of a channel where inertial overspill allowed medium- to coarse-grained sand carried in the lower part of the throughgoing flow to overtop the channel margin (Imran et al., 1999; Straub et al., 2008; Arnott, 2010; Khan and Arnott, 2011). In contrast, the generally thinner-bedded, finer-grained, mud-dominated strata between PL 2 and BST 4 at MR are interpreted to be associated with the same channel (BST 4), but were deposited on its inner-bend levee. This implies that in MR-S the unexposed sinuous channel associated with these levee units most probably curved away from the outcrop exposure and then veered back, intersecting the outcrop at MR-N (see Figure 2.15).

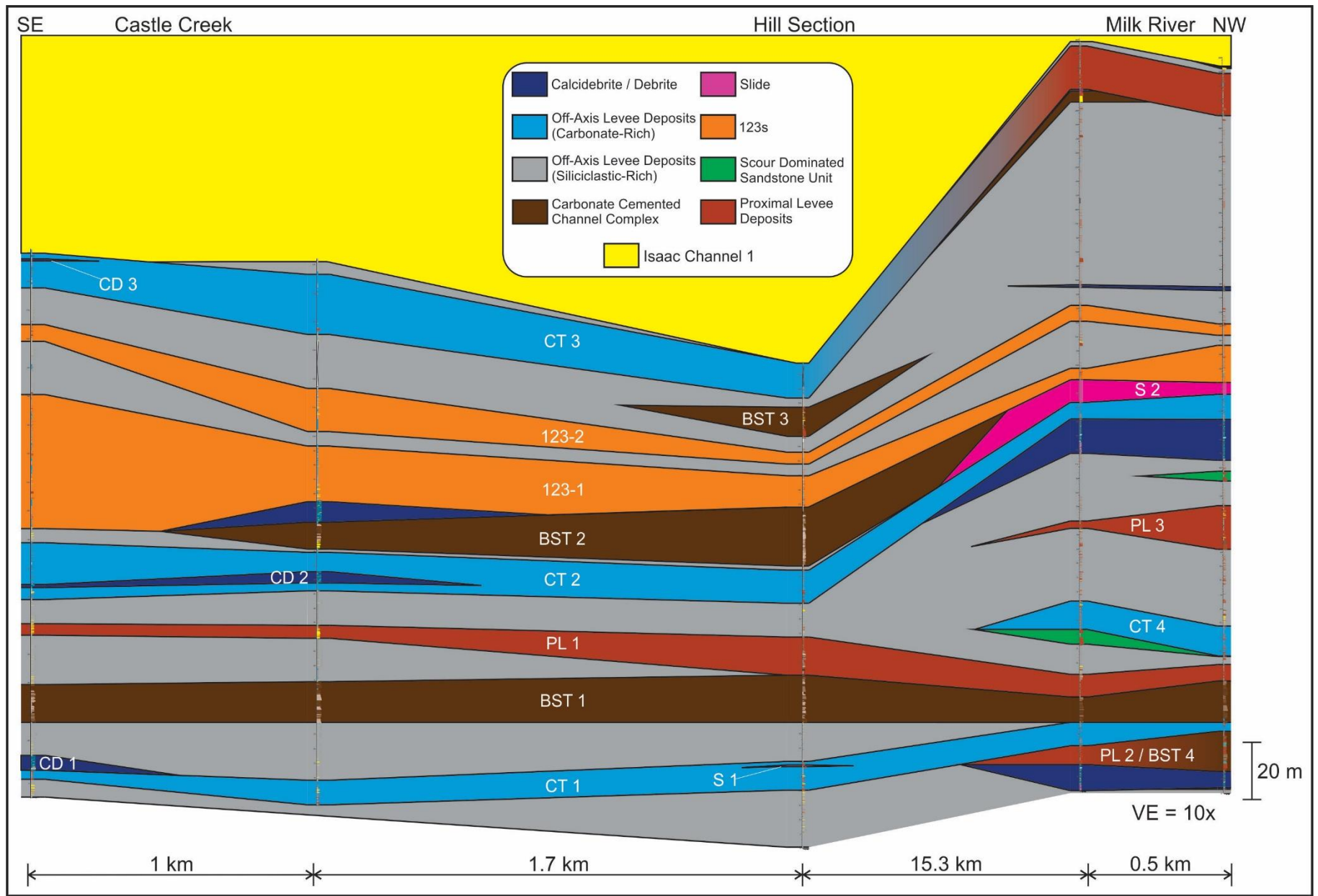


Figure 3.2 (previous page): SE to NW correlation from CC-S to MR-N showing the spatial distribution of stratal elements in the FIC; datum is the base of the lowermost bacon sandstone channel complex. Several units can be confidently correlated across the study areas, based on their distinctive lithological make-up and stratigraphic position - including three calciturbidite horizons, the basal sandstone channel complex and two 123 units.

III.II.III Distal Levee Deposits

Distal levee deposits are the most common stratigraphic elements in the FIC, making up > 62% of the stratigraphy. These units are composed predominantly of thin-bedded, mud-rich turbidites (Facies 3) with uncommon interbeds of medium-bedded sandstone that grade upward to a siltstone and then mudstone cap (facies 1 and 2). In this study, this element has been further divided into two subtypes based on mineralogy: siliciclastic mud-rich and calcilutite-rich.

Siliciclastic-rich units are composed of ~ 85% grey siliciclastic mud (predominantly upper-division turbidites of Facies 3) with local thin- to medium-bedded sandstone (carbonate-cemented and siliciclastic), rare thick-bedded sandstone and siliceous calcilutite. These strata make up ~ 46% of total FIC stratigraphy and most of the thicker units can be correlated all study areas. At CC and HS units range from 1-20 m thick, but in two instances are > 60 m-thick at MR (see Chapter 2 for details).

Carbonate-rich distal levee deposits are less common than their siliciclastic counterparts, comprising 14% of the FIC. These consist of abundant (~ 54%) very thin- to thin-bedded, olive-green to black, fine-grained calcareous turbidites (hereafter calciturbidites), and lesser siliciclastic mud and siliceous calcarenite (~ 27% and ~ 17%, respectively), that stack to form decametre-thick units (3-21 m). Overall, the FIC contains four calciturbidite horizons (CT 1-CT 4), three of which crop out in all study areas (CT 1-CT 3); CT 4 is absent at CC and HS. The lowermost unit (CT 1) is composed principally of medium-bedded, fine- to medium-grained, reddish-orange, traction-structured and massive siliceous calcarenite (facies 1 and 2) and siliciclastic mud-rich, thin-bedded turbidites (Facies 3) with occasional thin-bedded, structureless sandstone (Facies 1, siliceous calcarenite and silicarenite) and very-thin- to thin-bedded siliceous calcilutites (Facies 3). Calciturbidite 2 (CT 2) is composed

primarily of very thin- to thin-bedded siliceous calcilutites, giving it a distinctive olive green to dark-greenish-grey colour in outcrop, with less frequent very-thin- to thick-bedded reddish-orange siliceous calcarenite (facies 1 and 2; Figure 3.1H). Much like CT 2, the uppermost calciturbidite horizon (CT 3) is dominated by siliceous calcilutites with lesser thin-bedded siliceous calcarenites (Figure 3.1I). Calcilutites in CT 3, however, are much darker (dark-grey to black), and thinner beds have a deeper red colour that most probably relates to higher iron content.

At MR, CT 1, CT 2 and CT 4 are lithologically similar and consist of greenish-brown to black siliceous calcilutites and dark reddish-brown siliceous calcarenites. In contrast, CT 3 is significantly more sand-rich than anywhere else (~ 40% compared to ~ 10%) and consists of alternating, sharply bounded packages of calcarenite-rich T_{ade} , T_{abde} , T_{abcde} and T_{bcde} turbidites (up to 5 m-thick) and siliceous calcilutite- and/or siliciclastic mudstone-rich strata (up to 3.5 m-thick) (see Chapter 3: III.II.II Chapter 2: IV.II.II for details).

Both siliciclastic mudstone- and siliceous calcilutite-rich units are interpreted to have been deposited along the distal levees of active channel belts by low-density turbidity currents. Calciturbidite horizons are interpreted to have been sourced from a coeval shallow-water carbonate platform during periods of elevated sea level that shut down siliciclastic sediment supply to the deep basin. Siliciclastic mud-rich units, on the other hand, represent periods when the carbonate platform was inactive, due either to subaerial exposure caused by a sea level fall or drowning during rapid sea level rise. Some siliciclastic mudstone beds might also represent hemipelagic fallout during periods of quiescence (Meyer, 2004; Payros and Pujalte, 2008; Angus, 2016). The rare, intercalated medium- to thick-bedded sandstones in these units probably resulted from anomalously high-energy channelized flows that allowed sand-size particles to overspill the levee margins and travel further away from the channel.

III.II.IV 123s

The 123 units consists of a basal fine- to very-coarse grained, thin- to thick-bedded, reddish-orange siliceous calcarenite and rare carbonaceous sandstone (Package 1), overlain sharply by a succession of siliciclastic mud-rich turbidites with occasional interbeds of very thin-bedded, fine- to medium-grained siliceous calcarenites (T_{ade} , T_{bde} , T_{bcde} , T_{cde} , T_{de}) (Package 2), which then is capped sharply by a succession of very-thin to thin-bedded greenish-brown to black siliceous calciturbidites (Package 3) (Figure 3.1F-G). The base of the basal sandstone layer is locally scoured ($\pm$ mudstone and carbonate clasts) and overlain by planar-stratified (56%), massive (23%), ripple cross-stratified (19%), and rarely dune cross-stratified (2%) sandstone. Most thicker sandstone beds can be traced laterally as far as ~ 170 m, but in most cases irregularly thicken and thin, often over short distances, multiple times before ultimately pinching out. Additionally, the internal make-up of the sandstone beds can vary laterally. For example, one sandstone layer transitions from planar-stratified to dune cross-stratified and then back to planar-stratified sandstone over about 12 m. In many places the basal sandstone layer is absent, leading to a repetition of siliciclastic mud-rich (2) and carbonate mud-rich (3) packages. Also, the strata that make up the 2 and 3 packages show negligible lithological change laterally.

Two 123 units are observed in the study areas. Here they range from 4-47 m-thick and make-up $\sim 12\%$ of the total FIC stratigraphy. Thin-bedded siliciclastic (Package 2) and calcareous (Package 3) turbidites of Facies 3 dominate the stratigraphy (75%), whereas coarser-grained, thicker-bedded layers of facies 1 or 2 (17% and 7%, respectively) comprise the remaining 25%.

The abrupt and repetitive cm- to m-scale alternation of siliciclastic (packages 1 and 2) and carbonate (Package 3) deposition in the 123 units is unique to this part of the FIC and suggests the systematic and rapid fluctuations in sediment mineralogy were quite possibly related to high-frequency eustatic oscillations (i.e. Milankovitch cycles), although this requires further work to substantiate. Recently Navarro (2016) suggested that Package 1 sandstones were the fill of (erosional) gullies based on lithological characteristics observed in seismic images of modern carbonate platforms (i.e. Bahamas

Bank (Rankey and Doolittle, 2012; Mulder et al., 2012a,b, 2014) and Great Barrier Reef (Puga-Bernabéu et al., 2011, 2013)) and were interpreted to have been deposited by: 1) the episodic passage of single siliciclastic-rich turbidity currents; or 2) periods during which multiple, laterally adjacent, coeval high-energy flows were preferentially routed through the gullies. However, based on detailed logging of the 123s at CC-S it does not appear that multiple sandstones were deposited along any single stratigraphic horizon. Instead it appears that Package 1 beds only become more common in a single ~ 5-8 m-thick stratigraphic interval. Also, there is a scaling issue when comparing 123 sandstones to gullies, as the latter are typically tens of metres deep and hundreds of metres wide, which is almost an order of magnitude larger than sandstone packages (1) in the 123s (all < 1 m-thick and < 170 m-wide). The limited relief of Package 1 sandstones suggests that their basal surfaces would not have been sufficiently large to confine later turbidity currents, which then makes the correlation with modern slope gullies improbable. Instead they are interpreted to be shallow scours formed at the base of bypassing turbidity currents that most probably deposited a thin, most likely discontinuous residuum of medium and coarse sand that subsequently was reworked by later flows. The overlying siliciclastic mud-rich strata of Package 2 was then deposited during times of quiescence or by low-energy turbidity currents that draped and blanketed the underlying Package 1. This, then, was followed by a rise of sea level that terminated siliciclastic sedimentation and initiated the growth of a carbonate platform that shed fine-grained carbonate sediment downslope building up Package 3.

III.II.V Scour Dominated Sandstones

Scour dominated sandstone units are minor stratal elements in the FIC, making up < 1% of total stratigraphy, and are only observed at MR. These units range in thickness from ~ 0.2-5 m and consist of one to five medium- to very thick-bedded, medium- to very-coarse grained, massive siliceous calcarenite and granule siliceous calcirudite (T_a) with abundant mudstone clasts (Facies 1). These beds typically have erosive (irregular) bases, and fine upward to fine- to medium-grained, traction-structured sandstone (T_{abde} and T_{acde}). Individual sandstone beds are either amalgamated or separated

by thin-bedded turbidites. Over tens to hundreds of metres laterally, sandstones typically thin and onlap mud-rich turbidites, with only one unit extending across the width of the MR outcrop (~ 600 m) (Figure 3.1P). In all cases these units are underlain and overlain by thin-bedded turbidite units (mostly mud-rich, however one unit is overlain by CT 4).

Due to their thinness, (generally < 3.5 m), lateral discontinuity, abundance of mud intraclasts and typically coarse-grained make-up, strata described here are interpreted to be shallow scours that, like Package 1 strata in the 123s, were filled with a residuum of coarse sediment left behind by erosive throughgoing turbidity currents. Their greater thickness and abundance of mudstone clasts relative to Package 1 sandstones might be a result of differences in composition of the substrate beneath the throughgoing turbidity currents, specifically a moderately compact and therefore comparatively more easily eroded siliciclastic sea bed versus a partly lithified carbonate substrate.

III.II.VI Carbonate-Rich Debrites and Slides (Mass Transport Deposits)

Ranging in thickness from 0.6-14.4 m, calcidebrites make up 6% of the total FIC thickness. Slide deposits are confined to only three occurrences and represent 1% of FIC stratigraphy. These units commonly occur in close stratigraphic association with calciturbidite horizons interpreted to have been deposited during times of elevated sea level and the greatest extent of the carbonate platform. During this time the platform's upward growth would have likely steepened the slope, causing it to become gravitationally unstable and shed sediment basinward in the form of debris flows and slides (e.g. Hedberg, 1970; Grammer et al., 1993; Ross et al., 1994).

III.III Sequence Stratigraphy of the FIC

The following sea level interpretation relating changes of eustasy and sediment supply to stratal characteristics of the FIC was proposed by Navarro (2016). The base of the FIC represents the beginning of a long-term (possibly 3rd-order) eustatic rise that followed the lowstand marked by Isaac channel 0 (Figure 2.2). This eustatic rise resulted in the initiation and development of a shallow water

carbonate platform that became a source of calcareous sediment that was resedimented down the continental slope. At the same time siliciclastic input into the deep-water basin was terminated. This platform was likely located southeast of the Cariboo Mountains which was subsequently eroded by the sub-Cambrian unconformity. The two lowermost calciturbidite horizons (CT 1 and CT 2) were deposited in response to the growth of this platform. However, continued transgression eventually drowned the platform, which did not redevelop until regression lowered sea level back into the carbonate production window (e.g. Droxler et al., 1993; Harper et al., 2015), depositing CT 3. The FIC is then truncated at its top by a > 100 m-thick siliciclastic leveed-channel complex, whose base represents a sequence boundary (SB) related to a eustatic fall (Navarro, 2006) that most probably exposed the shelf and terminated carbonate production. Superimposed on the this long-term (3rd-order) trend were shorter duration (4th/5th-order) sea level fluctuations that formed decametre-thick channel complexes filled with amalgamated, carbonate-cemented, coarse-grained sandstone (bacon channel complexes 1 and 2 (BST 1 and BST 2) during conditions of lowered sea level.

Part IV Geochemistry

IV.I Methodology

During two field seasons eight complete bed-by-bed stratigraphic sections were logged through the FIC at the Milk River and Hill Section field areas, totalling a thickness of 1500 m. The FIC was then sampled at each study area, utilizing a sample interval of ~ 2-3 m in carbonate-rich strata and ~ 5 m in siliciclastic dominated strata (wherever a carbonate bed was located within). In total, 193 samples were selected for $\delta^{13}\text{C}_{\text{carb}}$ analysis, eight of which were from carbonate clasts in debrite and sandstone units. This database was then supplemented with 21 samples bulk analyzed previously for $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}$ by G.M. Ross (unpublished data, 2003).

Since strata of the WSG in the SCC have been subjected to greenschist metamorphism, samples were first screened using a variety of petrographic techniques before analyzing their isotopic

composition to ensure that primary marine signatures were being measured. Polished thin sections were made from a suite of samples (~ 80) representing the various lithologies in the FIC. Additionally, 36 samples from MR and HS were photographed using cathodoluminescence (CL) microscopy and the spectral signatures of the various cements obtained using software at the University of Ottawa's MicroAnalysis Laboratory. Samples were then stained using potassium ferricyanide and Alizarin Red-S solution (Dickson, 1966) and then analyzed petrographically to determine the approximate mineralogical proportions, cement paragenesis (aided with the CL image), with the ultimate goal being to identify the primary depositional and least altered phase.

All isotopic preparation and analyses were done at the University of Ottawa's G.G. Hatch Stable Isotope Laboratory. For $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}$ analysis samples with multiple cements (i.e. siliceous calcarenites and siliceous calcirudites from the bacon sandstone units) were micro-drilled using a Micromill with a 400 μm drill bit, whereas those composed predominantly of micrite to microspar matrix (i.e. siliceous calcilutites) were drilled less precisely with a hand drill and a 1/8" bit. Between 0.5-2.1 mg of powdered sample, depending on the proportion of calcite versus siliciclastic content in the sample, was then weighed into exetainers and treated with 0.1 mL of phosphoric acid (H_3PO_4) and flushed with helium before being placed in a 25°C bath for 24 hours. Measurements were performed on a Thermo Finnigan Delta XP and Gas Bench II (see Application Flash Report G 31 from Thermo Finnigan for more instrument details). Data were normalized using international standards NBS-18, NBS-19 and LVSEC (C only) and isotopic compositions are reported in permil (‰) relative to the PDB standard. Analytical precision (2σ) is $\pm 0.1\text{‰}$ with duplicate $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ analyses differing by an average of 0.06‰.

Thirty-seven samples composed mostly of microspar were selected for $\delta^{13}\text{C}_{\text{org}}$ analysis, 25 were also analyzed to determine their manganese and strontium content. These samples were initially ground into a fine powder using a vibratory pulveriser. For $\delta^{13}\text{C}_{\text{org}}$ analysis ~ 500 mg of powder was placed in 10 mL beakers and then treated with ~ 7 mL of 10% hydrochloric acid (HCl) and left to soak

overnight to eliminate carbonate. Samples were then decanted, washed twice with distilled water to lower the pH, and then placed in an oven for 72 hours to dry. To estimate the amount of sample needed to obtain isotope values, the percent carbon was determined by running 11x5 mm tin capsules filled with ~ 10 mg of sample mixed with 20-25 mg of tungsten oxide (WO_3) in an elemental analyzer. Specifically, the capsules are loaded into the carousel of an autosampler and a sample descends into a column of solid chemicals at 1150°C or 925°C and then is flash combusted at 1800°C with the addition of oxygen. Helium is used to carry the combusted gases through the column of chemicals to obtain N_2 , CO_2 , H_2O and SO_2 . Next, a thermoconductivity detector (TCD) measures the gases as they are released and then Elementar software is then used to process the results from the TCD using the K-factor method. Analytical precision (2σ) is $\pm 0.1\%$.

After determining the total organic carbon (TOC) content, the desired amount of sample was weighed out and placed into tin capsules, before adding WO_3 , and then loaded into an elemental analyzer connected to an isotope ratio mass spectrometer (IRMS). Samples were then run as described above before being sent to the IRMS. The results were normalized using an internal standard (C-55) calibrated to international standards IAEA-CH-6, NBS-22, USGS-40 and USGS-41. The analytical precision using this standard is typically $< 0.2\%$.

Twenty-five samples were crushed to a fine powder using a vibratory pulveriser then submitted to the University of Ottawa's Inductively Coupled Plasma Mass Spectrometry (ICP-MS) Laboratory, where the Mn/Sr ratio was determined. A 50 mg portion of each sample was subjected to a two-step acid digestion, first in concentrated HNO_3 (0.5 mL) and HCl (1.5 mL) and then, after evaporating overnight at 110°C, in HNO_3 (0.5 mL) and deionized water (2 mL) and then centrifuged. Following this the clear portion of the solution was decanted, leaving behind a residue that was treated once with HF and HNO_3 (0.5 mL each) and then 0.5 mL of HNO_3 and 2 mL of water and gravimetrically diluted to 10 g. The solutions were then analyzed using an Agilent 8800 QQ ICP-M utilizing a combination

of “Drift Correction by Interpolation” and “Internal Standardization” to maximize measurement precision. Internal standards used included In, Rh and Y, of which Y proved to be the most useful.

The mineralogical composition of two siliceous calciturbidites from CT 2 and CT 3 were analyzed using x-ray diffraction. For these analyses randomly oriented samples were loaded onto a sample holder and analyzed in a Rigaku Ultima IV X-ray diffractometer system operated at 40 kV, 35 mA, and an incidence angle that spanned from 3° to 85° at 0.03° step interval. This X-ray Powder Diffractometer is a theta-2theta (2Θ) goniometer instrument system with one copper source and one diffracted beam. Data were evaluated and minerals identified using the PDXL2, an integrated X-ray powder diffraction software by Rigaku.

IV.II Results

When analyzing Neoproterozoic carbonates, especially those that have been subjected to low-grade metamorphism, there is always the possibility of post-depositional alteration of the geochemical signal. Therefore, it is imperative that all possible precautions are taken to ensure the measurements reflect primary depositional conditions, and, by extension, the isotopic composition of the Proterozoic seawater from which it was precipitated. With the increasing interest in Neoproterozoic geochemical studies many techniques have been developed to test for diagenetic alteration (e.g. Narbonne et al., 1994; Kaufman and Knoll, 1995). For $\delta^{13}\text{C}_{\text{carb}}$ data these include petrographic screening, cathodoluminescence microscopy, micro-sampling for isotope analyses, $\delta^{13}\text{C}_{\text{org}}$, manganese and strontium ratios, and oxygen isotopes.

IV.II.I Petrography and Cathodoluminescence Microscopy

Petrographic analyses on select samples (n=47) identified seven different carbonate phases in the FIC (Table 3.2), which collectively represent a continuum beginning with early seafloor neomorphism of primary (i.e. detrital) matrix, through recrystallization during early to late burial diagenesis and into metamorphism associated with Mesozoic tectonism. Microscopic crystal size

follows the classification of Folk (1965): micrite ($< 4 \mu\text{m}$), microspar (5-30 μm), pseudospar (30-50 μm), spar/metaspars ($> 50\mu\text{m}$).

Calcite phase 1 (C1) is principally an iron-rich (ferroan; C1A, 95%) microspar (average = 12 μm) and is the most common phase in the FIC. It is ubiquitous in siliceous calcilutite samples, making up $> 90\%$ of the thin section, and enveloping silt-sized quartz grains and C1B patches (Figure 3.3A-B). In siliceous calcarenites and siliceous calcirudites, C1 occurs in the intergranular space between quartz grains, giving the appearance that the grains are floating in a matrix of C1 (poikilitic). C1 crystals are sub-rounded with irregular boundaries that are difficult to identify, particularly in siliceous calcilutites, because of the cement's very-fine and omnipresent nature. In some samples, this microspar phase is iron-poor (C1B), but these portions typically occur as poikilitic patches in a sample otherwise dominated by blue to dark blue stained Fe-rich calcite. The elevated amount of iron causes C1 to typically display a dull dark-brown luminescence (Figure 3.3C), suggesting it crystallized from reduced pore fluids during burial diagenesis (Scholle and Ulmer-Scholle, 2003; James and Jones, 2016). C1's microspar crystal size and its ubiquity in fine-grained samples suggests it represents neomorphism of the primary carbonate phase after deposition on the seafloor and during early burial diagenesis, and therefore likely retains much of the primary shallow-marine isotopic signal (see discussion). During the latter $\sim 350 \text{ Ma}$ of the Neoproterozoic the primary shallow-marine precipitate was likely metastable aragonite (Hardie, 2003), which during early burial diagenesis was transformed to more stable calcite. The rare non-ferroan C1B patches likely record near-surface neomorphism of C1, however its rarity suggests it was a short-lived episode of neomorphism as the carbonate sediment was rapidly buried and pore fluids became anoxic and Fe-rich.

Calcite phase 2 (C2) is the least common phase (in 39% of samples) and consists of prismatic calcite pseudospar to spar (average = 119 μm) that commonly occurs along the edges of quartz grains in siliceous calcarenites and siliceous calcirudites, and occasionally along pyrite crystals. It is typically ferroan (blue), but uncommon non-ferroan (red) examples are observed. In thin section ferroan C2

crystals are stained a lighter shade of blue than C1A (Figure 3.3D). Under cathodoluminescence C2 ranges in colour from dark orange to light brown with a “cloudy” texture and appears to be subtly zoned, with the lightest colours occurring along crystal rims (Figure 3.3E). C2’s prismatic form and occurrence along the edges of quartz grains suggests that it formed during deeper burial diagenesis as compaction dissolved and recrystallized microspar of C1.

Cement	Thin Section Description	Cathodoluminescence Description	Relative Timing of Cementation
C1	Calcite microspar. Commonly dark blue iron-rich (C1A; 95%) with rare pink to red iron-poor patches (C1B; 5%).	Most commonly dark-brown, with occasional light brown to dark orange. Occasionally non-luminescent.	Neomorphism of primary carbonate phase to early burial diagenesis.
C2	Prismatic calcite pseudospar to spar that commonly occurs along the edges of quartz grains or along edges of pyrite grains. Iron-rich examples are slightly lighter blue than C1, whereas iron-poor are clear	Dark orange to light brown, with lightest portion occurring along edge of crystals. Also has a “cloudy” texture.	Burial diagenesis.
C3	Fibrous or elongated calcite typically found along pressure shadows and along/within veins. Predominantly iron-rich with a lighter blue colour than C1, sometimes “dirty” white colour.	Orange to light brown. Lighter than C1.	Late burial diagenesis/ Metamorphism.
C4	Iron-poor microspar, commonly associated with clusters of brown, anhedral D1 crystals and crosscut by D2.	The iron-poor microspar, when alone, is dark brown, but common association with D1 gives it a “dirty” orange appearance.	Late burial diagenesis/ metamorphism.
D1	Dolomicrospar that occurs as single crystals distributed sporadically throughout C1. Typically makes up < 2% of samples, more common in CT3 calcilutites.	Bright orange and yellow spots within dull/non-luminescent C1+C2.	Primary marine cementation to early burial diagenesis.
D2	Anhedral to euhedral, brown or clear dolo-microspar, manganese rich. Sometimes poikilitic. Occasionally associated with rounded pyrite grains.	Bright orange.	Burial diagenesis (after C1, likely around timing of C2).
D3	Brown dolomite microspar filling veins or fractures.	Bright orange.	Metamorphism.

Table 3.2: Carbonate cements identified in FIC strata with their respective thin section and cathodoluminescence descriptions and interpreted timing of cementation.

Calcite phase 3 (C3) is most common in samples with abundant ferroan C1 (e.g. siliceous calcilutites and dark brown bacon channel sandstones) and consists of fibrous to anhedral ferroan calcite pseudospar to metaspar (40-330 μm , average = 152 μm) that occurs along the irregular edges deformed quartz grains, in fractures in quartz grains and in veins (Figure 3.3F-H). This phase can be distinguished from C1 by its larger crystal sizes, lighter blue colour in stained thin section and lighter colour under CL (orange to light brown). The presence of C3 in pressure shadows and in/along veins that cross-cut C1 and C2 suggests it is the result of fluid remobilization and recrystallization of these cements during late burial diagenesis and orogenic deformation/metamorphism.

Dolomite phase 1 (D1) consists of dolomicrospar crystals dispersed throughout neomorphosed C1 matrix. These small brown crystals are difficult to resolve in stained thin sections, but appear as bright orange and yellow patches in CL (average diameter = 14 μm) (Figure 3.3C). Based on visual estimates in thin sections this phase makes up $\sim 2\%$ of FIC carbonates on average, but is slightly more abundant in CT 3 calcilutites ($\sim 3\text{-}5\%$). The rarity of dolomite in siliceous calcilutite samples was corroborated by XRD analyses that showed dolomite content ranged from 2-4% (Figure 3.4). Based on its small crystal size and association with early carbonate phases, D1 is interpreted to have formed during primary seafloor cementation to early burial diagenesis, which is consistent with other Proterozoic studies (e.g. Tucker, 1983; Fairchild et al., 1991).

The final calcite phase (C4) comprises light red, iron-poor microspar ($\sim 20 \mu\text{m}$) that is associated with dolomite phase 2 (D2) in fine- to medium-grained siliceous calcarenites and siliceous calcilutites. The association with D2 typically gives C4 a brownish colour and brighter luminescence (Figure 3.3D-F,I). D2 consists of euhedral to anhedral dolomite crystals whose size cluster around two values. Euhedral crystals are brown to clear and range from 70-130 μm in diameter (mean = 92 μm , dolosparite) and commonly occur close to the edges of quartz grains. Darker brown anhedral crystals, on the other hand, are 15 μm on average (microdolomite) and typically dispersed in C4 and

occasionally associated with rounded/anhydral pyrite grains, particularly in calciturbidite 3. Under CL, D2 crystal clusters are bright orange, indicative of elevated Mn^{2+} contents.

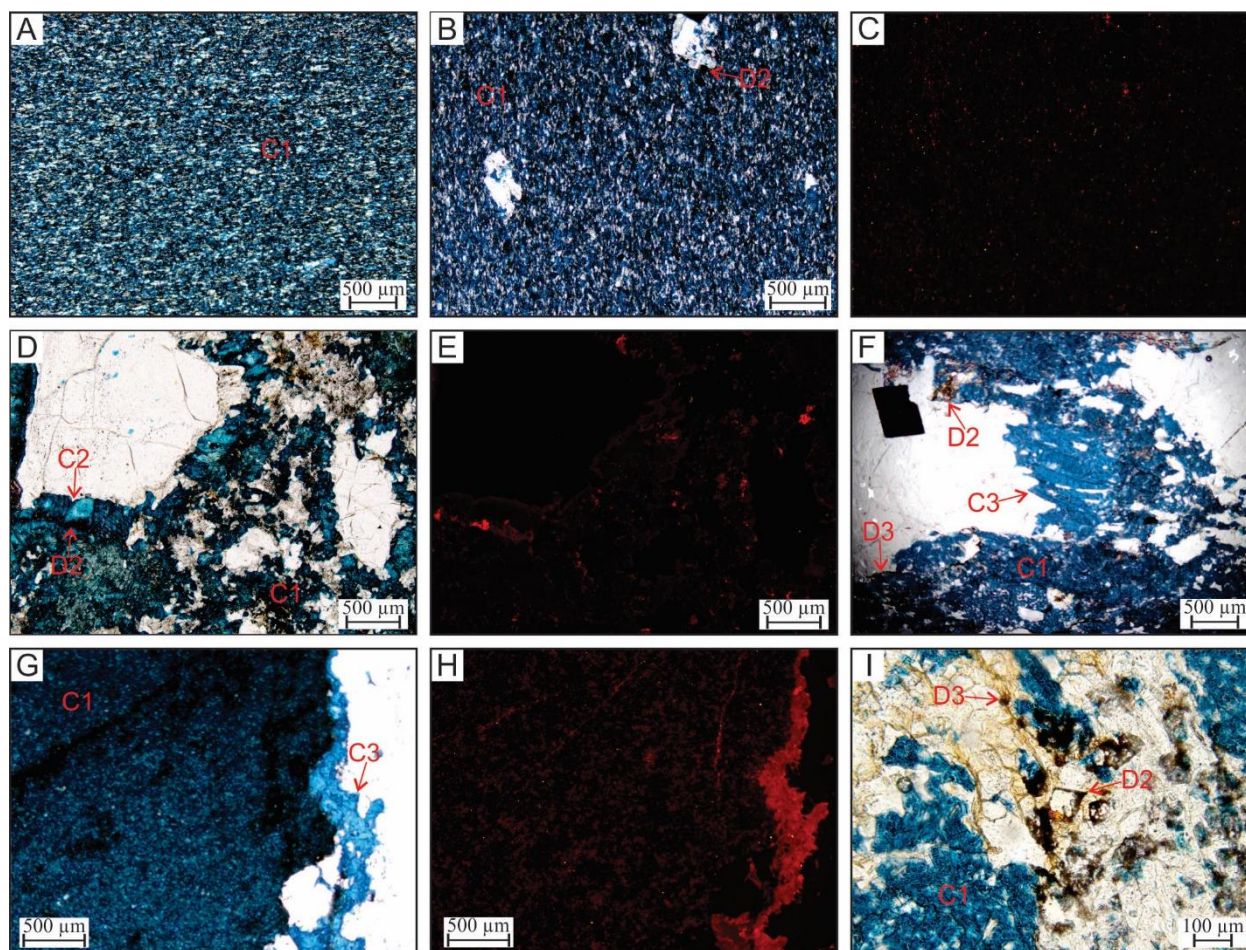


Figure 3.3: Representative photomicrographs of carbonate cements observed in the FIC. (A) Siliceous calcilutite from CT2 at HS with the pervasive blue, poikilitic ferroan calcite microspar cement (C1). (B) Another CT 2 siliceous calcilutite and its corresponding cathodoluminescence (CL) in photomicrograph (C). C1 cement commonly displays dull, dark-brown luminescence with occasional non-luminescence, whereas randomly dispersed bright orange and yellow spots correspond to primary/detrital dolomicropar (D1). (D) Dark-brown siliceous calcirudite from bacon sandstone 1 at HS and its corresponding CL in photomicrograph (E). Blue prismatic ferroan calcite crystals (C2; lighter blue than C1) occur at the edge of both quartz grains (white). Note also the clusters of brown dolomicropar and C1 matrix throughout the image. Under CL C2 is lighter brown than C1 matrix, with the lightest portion occurring along the edge of crystals, and has a “cloudy” texture. (F) Sample from same bed as D/E with fibrous or elongated ferroan calcite crystals (C3) occurring along the edge of the quartz grain. Below the quartz grain is brown dolomicropar filling fractures (D3) and a cluster of anhedral dolomicropar (D2) above. (G) Siliceous calcilutite with dark blue C1 matrix and a quartz vein along the right side of the photomicrograph. Ferroan calcite cement (C3) occurs along the margin of the vein. (H) Under CL this cement is orange, and therefore, distinctly different from dull luminescent C2. (I) Euhedral dolomite crystals (D2) within quartz rich (white-beige coloured) portion of a bacon sandstone bed. C1 is the dominant micritic cement and there are also clusters of D3 in the upper part of the image.

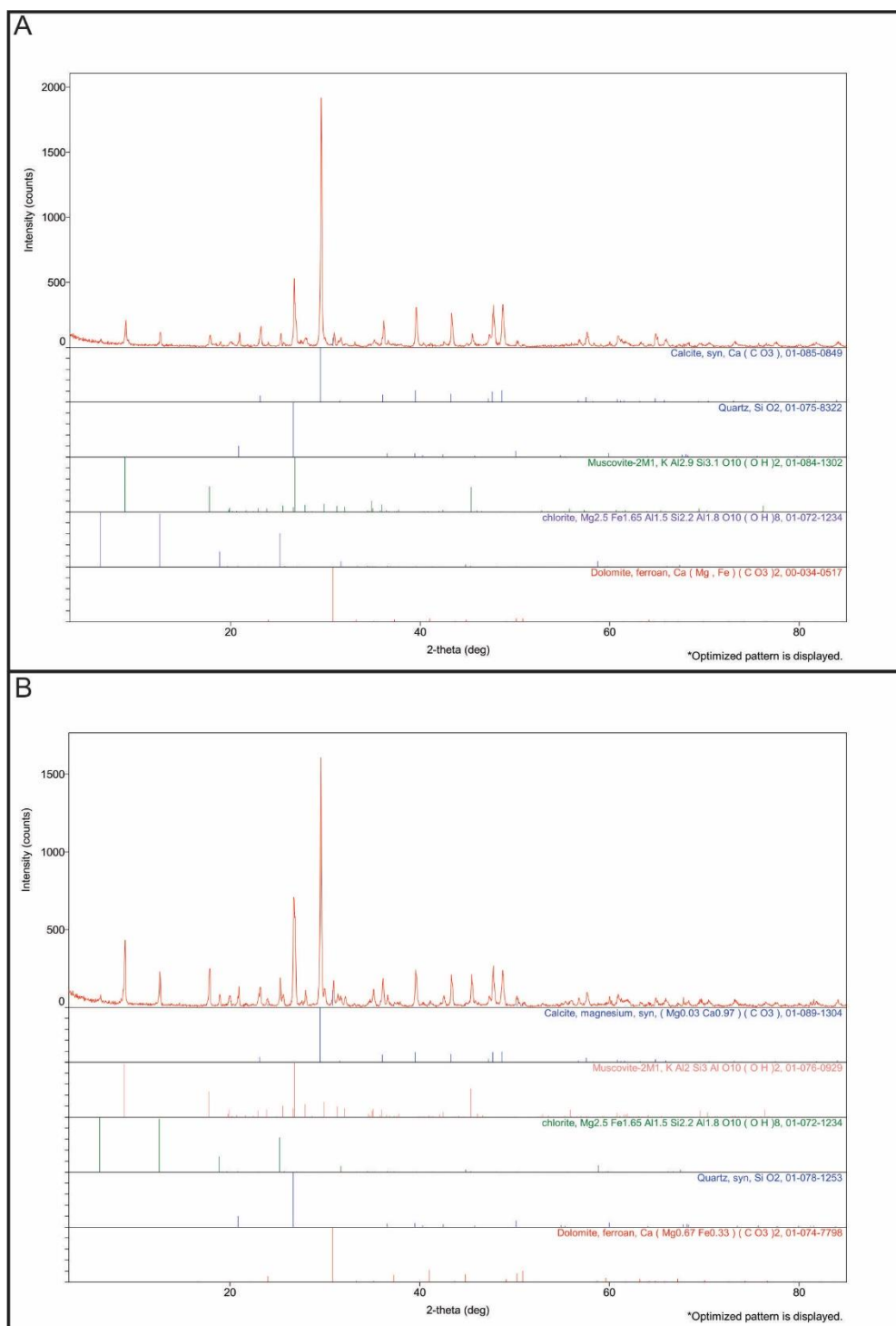


Figure 3.4: Measured X-ray diffractogram pattern obtained from a representative siliceous calcilutite sample from (A) Calciturbidite 2 and (B) Calciturbidite 3, analyzed as random bulk powders. Major intensities or peaks were identified for the following minerals (bottom): calcite, chlorite, ferroan dolomite, muscovite, and/or quartz.

Dolomite phase 3 (D3) is a brown to dark brown microspar that typically occurs along the edges of veins or fractures that cross-cut C1, C2, D1 and D3 phases and commonly parallels the pervasive metamorphic fabric in more deformed samples (Figure 3.3F,I). Like other dolomite phases D3 is bright orange under CL. The cross-cutting relationship with early cements and association with veins and the metamorphic fabric suggest this cement is the result of Mesozoic metamorphism and deformation. The similarity in colour, both in thin section and under CL, to anhedral microspar D1 and D2 crystals might indicate that D3 crystallized from metamorphic fluids that dissolved and reprecipitated D1 and D2 cements perpendicular to the primary direction of orogenic stress (i.e. σ_3).

Based on the previous cement descriptions the proposed paragenetic sequence for FIC carbonate cements is:

C1B → C1A → D1 → C2 → D2 and C3 → C4 and D3 (see Figure 3.5).

Fine-grained samples (e.g. siliceous calcilutites and fine- to medium-grained siliceous calcarenites) consist predominantly of early cements (C1 & C2), whereas coarser samples (e.g. coarse to very-coarse siliceous calcarenites and dark brown siliceous calcirudites in bacon channels) typically contain most of the cements in various proportions. This suggests that finer samples were most probably lithified early, and therefore were less susceptible to further cementation and/or alteration during later burial diagenesis and subsequent metamorphism (e.g. Popovic, 2016; Lee, 2016). In summary, the generations of cement observed in strata of the FIC show a progressive change in environment from seafloor, to shallow subsurface, through burial diagenesis and ultimately metamorphism associated with Mesozoic tectonism.

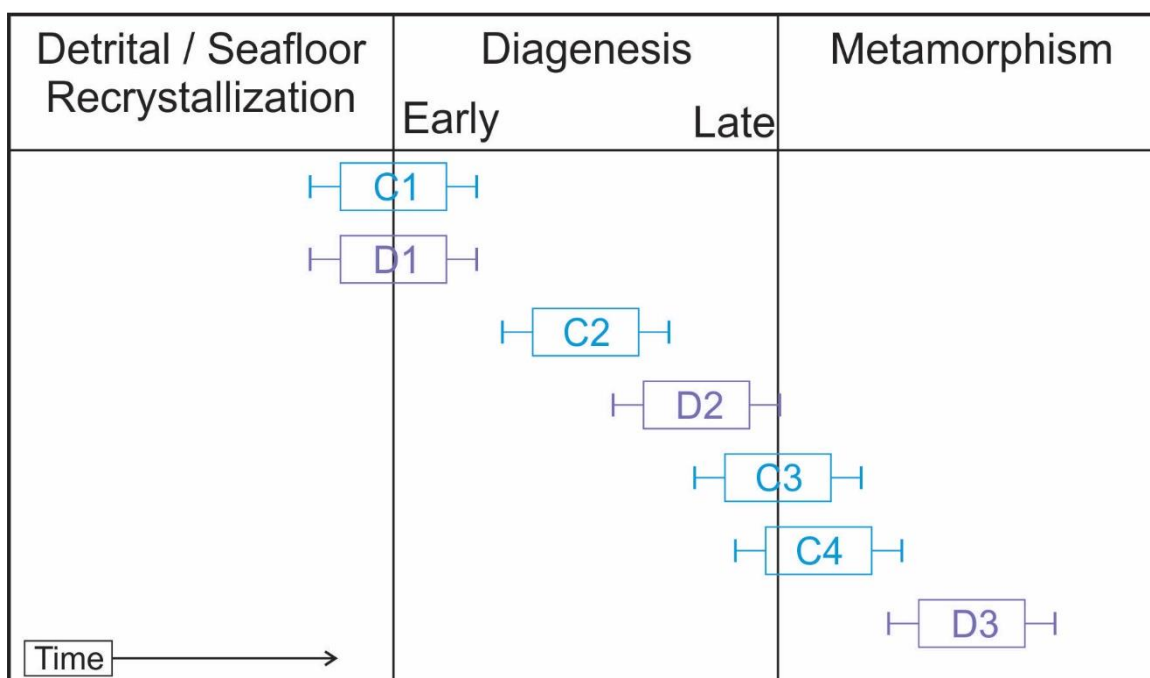


Figure 3.5: Paragenetic history of FIC carbonate phases. C1 is interpreted to represent neomorphism of primary detrital carbonate sediment during seafloor early burial diagenesis. D1 is interpreted to be contemporaneous. C2 and D2 are interpreted to deeper burial phases, whereas C3 and C4 formed during late burial diagenesis and early metamorphism. Dolomitic microspar that occurs along fractures that parallel the metamorphic fabric (D3) is associated with metamorphism during Mesozoic orogenesis.

IV.II.II Geochemistry

Figure 3.6 shows the compilation of $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}$ data obtained from all study areas. These data, plus the unpublished data of G.M. Ross (2003) are present in Appendix A.

At the base of the FIC at Castle Creek $\delta^{13}\text{C}_{\text{carb}}$ values are negative ($\sim -5\text{‰}$), and then increase gradually upsection to $\sim 0\text{‰}$ at the base of the second calciturbidite horizon (CT 2), where it then increases abruptly to 2‰ before plateauing around 2.5‰ for more than 65 metres. In the upper part of 123-1, isotopes values become progressively more negative (lighter) up to the base of CT 3. Approximately 5 m higher the values drop abruptly to a nadir of -6.3‰ , which, in total, represents a decrease of 8.9‰ from the upper part of 123-1. The top of the FIC at Castle Creek South is truncated abruptly by the base of Isaac channel 1. Therefore, it is unclear whether the decreasing trend continues or $\delta^{13}\text{C}_{\text{carb}}$ returns to more positive values.

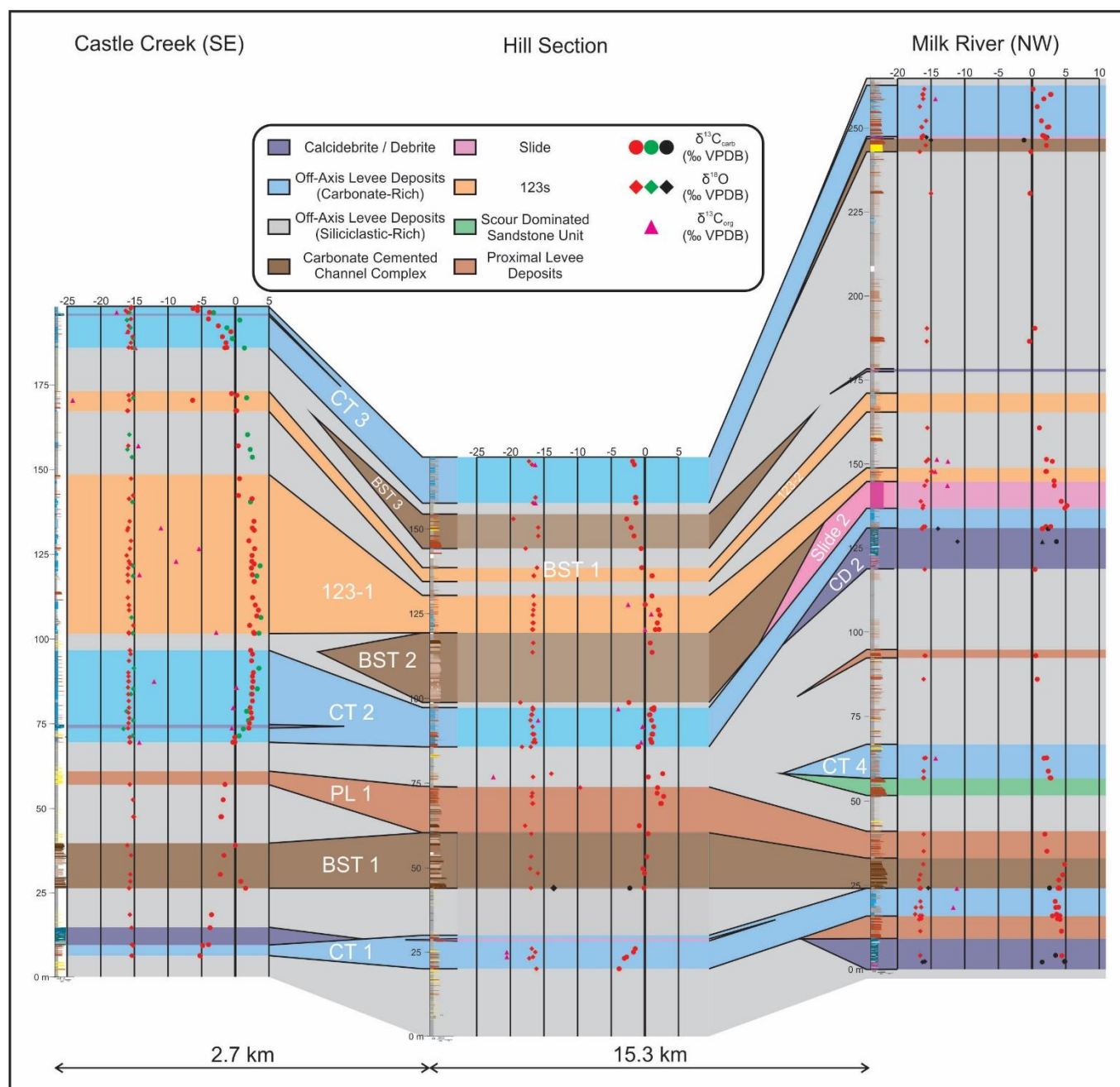


Figure 3.6: Correlation of strata between CC-S, HS and MR-S study areas. Stable isotope data is presented to right of each stratigraphic column. $\delta^{13}\text{C}_{\text{carb}}$, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}_{\text{org}}$ samples are denoted by circles, diamonds and triangles, respectively. Red and black shapes indicate samples collected as part of this study, with the black corresponding to carbonate clasts and red to matrix cements, and green shapes are the unpublished data of G.M. Ross (2003).

Like at CC, $\delta^{13}\text{C}_{\text{carb}}$ is negative, although slightly heavier (-3.8‰), at the base of the Hill Section.

This is then followed upwards by a progressive increase to 0‰ at the base of BST 1 and to $\sim 2.7\text{‰}$

just above the proximal levee unit, before becoming slightly negative at the base of CT 2, before increasing and plateauing at around 2‰ ~ 2 m higher. This plateau terminates at the top of 123-1 and then values become progressively lighter, with a change to negative values in 123-2 before reaching a low of -2.7‰ at the top of BST 3. After this nadir, $\delta^{13}\text{C}_{\text{carb}}$ becomes slightly heavier (-1.3‰) at the base of CT 3 before decreasing slightly again towards the top of the FIC.

Unlike the other two study areas, $\delta^{13}\text{C}_{\text{carb}}$ at the base of the FIC at Milk River is positive (3.5-4.9‰) persisting into BST 1, where after it declines to a minimum of 0.5‰ at the base of CD 2. This, then, is succeeded by a change to progressively higher values at the base of the slide above CT 2 (max of 5‰). Further upwards values gradually decrease and become negative around the top of 123-2. $\delta^{13}\text{C}_{\text{carb}}$ continues to straddle 0‰, but becomes positive near the base of CT 3. From here there is a gradual decline followed by an abrupt positive shift and another decrease to near-zero values at the top of CT 3 immediately below Isaac channel 1.

Samples from carbonate clasts appear to show little correlation with their host strata, with differences ranging from -2.1‰ to 2.9‰ (Table 3.3). Two clasts obtained from BST 1 at MR and HS were 1.4‰ and 2.1‰ lighter, respectively, than the host matrix. Also, two clasts from Calcidebrite 1 at MR were lighter than the matrix (2.9‰ and 0.9‰ less), whereas a third clast was isotopically heavier (0.4‰ more). However, an F-test was performed on the isotopic values from clasts and host matrix at 95% confidence with the null hypothesis that the variance of the datasets is overlapping, and therefore have a common origin. Results (Table 3.4) indicate that the F value is less than F-critical, thus the null hypothesis cannot be rejected, indicating that there is no significant difference in $\delta^{13}\text{C}_{\text{carb}}$ content between clasts and their host strata. Two of the clasts that were sampled contained microbial laminae and occurred in strata that also had oolite clasts, indicating they were likely eroded from a shallow-water carbonate platform and resedimented downslope. Their statistically similar isotopic values to their host strata indicates that the host matrix was probably also precipitated in shallow marine

conditions and retained its primary shallow-water isotopic signature after deposition on the base-of-slope.

Study Area	Unit	$\delta^{13}\text{C}_{\text{carb}}$ Matrix	$\delta^{18}\text{O}$ Matrix	$\delta^{13}\text{C}_{\text{carb}}$ Clast	$\delta^{18}\text{O}$ Clast	$\delta^{13}\text{C}_{\text{carb}}$ Difference	$\delta^{18}\text{O}$ Difference
Milk River	Caldecibrite 1	4.43	-16.63	1.5	-16.28	2.93	-0.35
Milk River	Caldecibrite 1	4.43	-16.63	4.87	-15.99	-0.44	-0.64
Milk River	Caldecibrite 1	4.43	-16.63	3.5	-16.61	0.93	-0.02
Hill Section	Bacon Sandstone 1	-0.13	-16.98	-2.24	-13.55	2.11	-3.43
Milk River	Bacon Sandstone 1	3.96	-16.58	2.6	-15.42	1.36	-1.16
Milk River	Caldecibrite 2	1.54	-16.31	2.41	-13.97	-0.87	-2.34
Milk River	Caldecibrite 2	1.54	-16.31	3.63	-11.08	-2.09	-5.23
Milk River	Slide 3	1.86	-16.39	2.18	-15.68	-0.32	-0.71

Table 3.3: Isotopic values measured in carbonate clasts and their host matrix. There appears to be no correlation between $\delta^{13}\text{C}_{\text{carb}}$ in clasts compared to the host matrix, with differences ranging from -2.1‰ to 2.9‰.

	$\delta^{13}\text{C}_{\text{carb}}$ Matrix	$\delta^{13}\text{C}_{\text{carb}}$ Clast
Mean	2.31	2.76
Variance	4.45	3.14
Observations	8	8
df	7	7
F	1.42	
P(F<=f) one-tail	0.33	
F Critical one-tail	3.79	

Table 3.4: Result table for an F-test comparing the $\delta^{13}\text{C}_{\text{carb}}$ values within clasts in mass-transport deposits and mixed siliciclastic-carbonate channel complexes to those of their host matrix.

Unlike $\delta^{13}\text{C}_{\text{carb}}$, $\delta^{18}\text{O}$ from the FIC is more or less uniform in all samples, rarely deviating from -15‰ and displaying little systematic change with stratigraphic position (Figure 3.6), which is similar to previous geochemical studies in the WSG of the SCC (e.g. Smith, 2009). Smith (2009) suggested that the rather uniform $\delta^{18}\text{O}$ values reflect metamorphic homogenization rather than primary depositional conditions, and as a result are not discussed further.

A simple test for diagenetic alteration is a crossplot of $\delta^{13}\text{C}_{\text{carb}}$ versus $\delta^{18}\text{O}$. If the result forms a straight line with positive slope the covariance is believed to result from alteration due to meteoric diagenesis (Fairchild et al., 1990; Kaufman and Knoll, 1995; Le Guerroué et al., 2006a). Results show a slight negative slope at CC and MR, but a slight positive slope at HS (Figure. 3.7A). Moreover, the R^2 values range from 0.027-0.056, suggesting little correlation between $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}$ values. This suggests that the $\delta^{13}\text{C}_{\text{carb}}$ signal was little affected by meteoric diagenesis, despite the overprint of the oxygen isotopes.

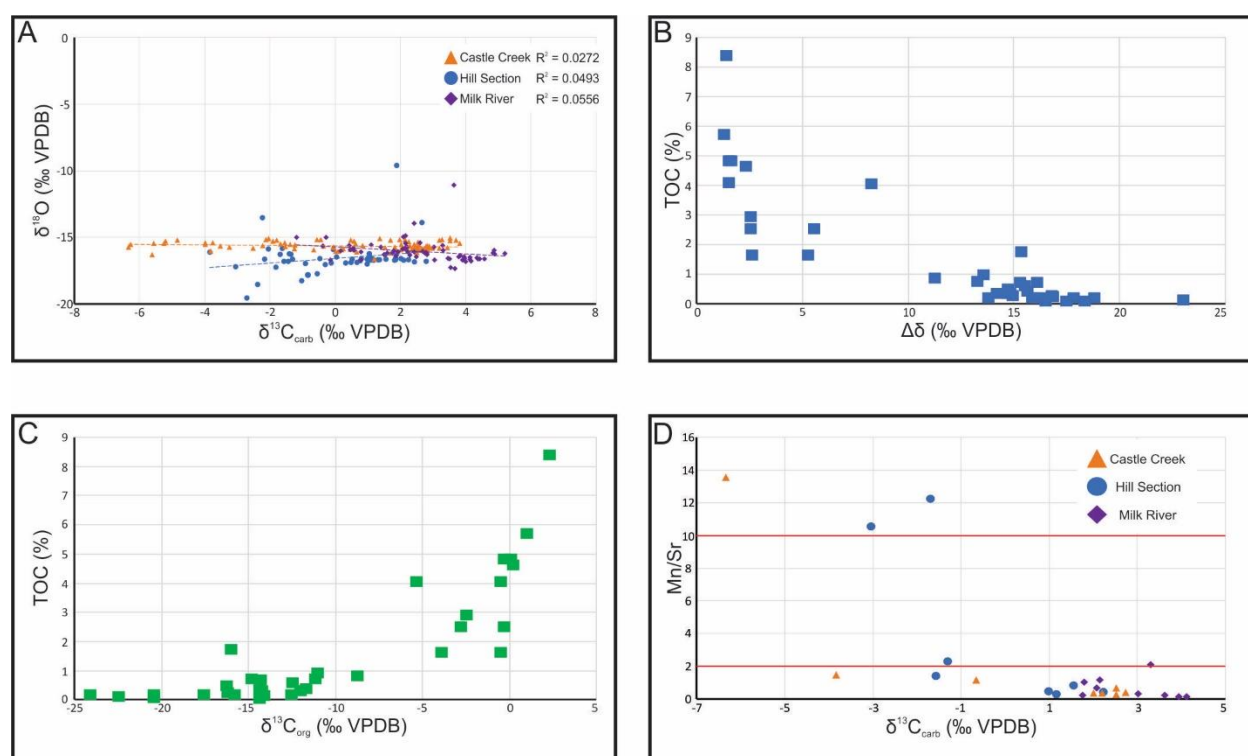


Figure 3.7: Cross plots of geochemical data from the FIC: (A) $\delta^{13}\text{C}_{\text{carb}}$ vs. $\delta^{18}\text{O}$, (B) $\Delta\delta$ ($\delta^{13}\text{C}_{\text{carb-org}}$) vs. total organic carbon, (C) $\delta^{13}\text{C}_{\text{carb}}$ vs. total organic carbon and (D) $\delta^{13}\text{C}_{\text{carb}}$ vs. Mn/Sr.

Knoll et al. (1986) were the first to report that $\delta^{13}\text{C}_{\text{org}}$ and $\delta^{13}\text{C}_{\text{carb}}$ from Neoproterozoic carbonates, although producing similar shaped curves, exhibited isotopic values that consistently differed by a similar value ($\Delta\delta = 26\text{-}30\text{‰}$). This covariance has since been used as a test for alteration as it was widely believed that no diagenetic process can alter the isotopic composition of $\delta^{13}\text{C}_{\text{org}}$ and $\delta^{13}\text{C}_{\text{carb}}$ in the same direction and by the same magnitude (Narbonne et al., 1994; Kaufman and Knoll,

1995; Hayes et al., 1999; Johnson et al., 2012). However, more recent work by Oehlert and Swart (2014) has shown covariance between organic and inorganic carbon isotopes in sections that were diagenetically altered, and accordingly casts doubt on this assumption. However, they determined that sections that underwent marine burial diagenesis were less altered than other diagenetic environments. Accordingly, deep marine samples in this study may be less altered than coeval, shallow-water carbonates.

The $\delta^{13}\text{C}_{\text{org}}$ values from the FIC produced two clusters (Figure 3.7B-C): one ranging from -11 to -24‰ and another around 0‰. The values straddling 0‰ were not expected, as organic carbon is known to be isotopically light. However, the values in the more positive group all occur in samples that have $\text{TOC} > 1\%$, and there is a clear correlation between TOC and $\delta^{13}\text{C}_{\text{org}}$ ($R^2 = 0.68$). TOC values from elsewhere in the Isaac Formation typically range from 0.1-0.8% (Davis, 2011; Bergen, 2017), which correspond to depositional values from 0.3-3.5% when corrected for organic material loss due to low-grade metamorphism (~ 50-75%) (Hayes et al., 1983). Therefore, it is likely that current TOC values in the FIC rarely exceed 1%, let alone 8%, as seen in one of the collected samples. The samples that were selected for $\delta^{13}\text{C}_{\text{org}}$ analysis in this study were all carbonates (siliceous calcarenites and siliceous calcilutites), therefore the likeliest explanation for elevated TOC values is that these samples were insufficiently subjected to treatment with hydrochloric acid and as a result the data reflects a mixture of organic and inorganic carbonate carbon.

To confirm this hypothesis the crushed powder of two samples with high TOC samples (2.7% and 5.5%) were subjected to 10% HCl for a week to allow for more aggressive carbonate carbon dissolution. Following acid treatment, these samples had TOC values of 0.3% and 0.6%, substantially lower than after the first treatment. This suggests that the TOC in the 0‰ $\delta^{13}\text{C}_{\text{org}}$ samples were not treated aggressively enough with HCl, likely skewing the isotopic values closer to the $\delta^{13}\text{C}_{\text{carb}}$ of FIC samples.

The $\delta^{13}\text{C}_{\text{org}}$ values from samples with TOC less than 1% are consistently less negative (average = -15‰) than what has been reported from most Neoproterozoic successions (e.g. Knoll et al., 1986; Kaufman and Knoll, 1995; Halverson et al., 2010), but they are similar to values of siliciclastic-shale samples from the FIC (average = -18‰; Ross, unpublished data, 2003; Milczarek, 2018). This suggests that the low TOC samples from this study were sufficiently treated with hydrochloric acid and that $\delta^{13}\text{C}_{\text{org}}$ was slightly less negative during FIC deposition. Additionally, the difference between organic and carbonate carbon ($\Delta\delta$) from the < 1% TOC samples (average = 15‰) was lower than the 25-32‰ range reported from other Neoproterozoic successions (Narbonne et al., 1994; Kaufman and Knoll, 1995; Hayes et al., 1999; Johnson et al., 2012), but similar to those from the OFP's Whitehorn Mountain Member (average $\Delta\delta$ = 14‰) (Smith, 2009), suggesting this is a common trend in the WSG of the SCC. The heavier $\delta^{13}\text{C}_{\text{org}}$ and lower $\Delta\delta$ could mean that the organic carbon isotopic signal was slightly shifted to less negative values through thermal alteration during burial diagenesis and low-grade metamorphism of the WSG whereas the carbonate carbon signal was little affected (e.g. Valley, 2001; Des Marais, 2001).

Veizer (1983) suggested that increasing meteoric diagenetic alteration caused Sr to be expelled from marine carbonates and replaced by Mn and therefore the Mn/Sr to increase. Subsequent studies proposed that Mn/Sr < 2 indicate near pristine carbon isotope values, whereas those < 10 are passable and anything > 10 are too altered to be reliable (Derry et al., 1992; Kaufman et al., 1993; Narbonne et al., 1994; Kaufman and Knoll, 1995). Of the 25 samples analyzed for their Mn and Sr composition, 20 (80%) had Mn/Sr < 2, 2 (8%) were slightly greater than two and 3 (12%) were > 10 (Figure 3.7D). There was also no strong correlation between Mn/Sr values and study location. Two of the samples with Mn/Sr > 10 were carbonate-cemented sandstones, which commonly contain numerous other cements in addition to the primary Fe-rich microspar. The Mn/Sr analyses were performed on bulk samples; therefore, it is likely that other diagenetic and/or metamorphic cements were incorporated into the final values, leading to the elevated ratio. The other sample with Mn/Sr > 10 had a $\delta^{13}\text{C}_{\text{carb}}$

value that was an outlier with respect to those from samples in close stratigraphic proximity. It was collected from the 123-2 unit at CC and had a $\delta^{13}\text{C}_{\text{carb}}$ composition of -6.3‰, whereas the nearby samples are near 0‰ (this sample is easy to locate in Figure 3.6). Therefore, this is the only sample analyzed for Mn/Sr that suggests the primary cement was post-depositionally altered.

Part V Discussion

V.I Strength of the Carbon Isotopic Signal

The ubiquitous, iron-rich C1 calcite phase represents the neomorphic alteration of primary detrital carbonate sediment during early burial diagenesis and therefore, potentially do not represent seawater values. However, in previous work on the underlying OFP, Smith (2009) compared the $\delta^{13}\text{C}_{\text{carb}}$ isotopic differences between ferroan and non-ferroan early phases and found that values for ferroan phases were 0.5-1‰ lower. If the median of this difference is assumed for ferroan C1 cement (-0.75‰), this amount represents just 8% of the range in FIC $\delta^{13}\text{C}_{\text{carb}}$ values, and therefore does not significantly alter the observed trend. If the $\sim 0.75\text{‰}$ difference between primary detrital carbonate phases and C1 does exist, it suggests the plateau that persists throughout much of the FIC is closer to 3.5-4‰, which is more consistent with general post-Marinoan values observed worldwide, excluding the negative excursions associated with the Gaskiers glaciation and the Shuram-Wonoka anomaly (e.g. Narbonne et al., 1994; Le Guerroué et al., 2006a; Fike et al., 2006; Kaufman et al., 2006; Jiang et al., 2007; McFadden et al., 2008; Tahata et al., 2013; MacDonald et al., 2013). This suggests that the isotopic values obtained from C1 most likely preserve primary seafloor chemical conditions rather than late burial diagenetic or metamorphic phases, notwithstanding the reset of $\delta^{18}\text{O}$ by metamorphism. Previous work has shown that $\delta^{13}\text{C}$ values are typically better preserved than $\delta^{18}\text{O}$ because most metamorphic fluids generally have low carbon content relative to that in carbonate rocks (Banner and Hanson, 1990). The preservation of primary chemical conditions is further corroborated by the isotopic similarity between siliceous calcilutites and very fine- to fine-grained siliceous calcarenites in the FIC,

in which C1 matrix makes up $\sim 90\%$ and $\sim 45\%$ of their composition, respectively. In the plateau between CT 2 and 123-1, $\delta^{13}\text{C}_{\text{carb}}$ values are independent of lithology, as both average $\sim 2.6\text{‰}$. This suggests that early micritic cements substantially reduced porosity and permeability in these strata. This then limited the potential for later alteration, which also is evinced by the smaller number of carbonate phases compared to coarser-grained lithologies (e.g. bacon channel conglomerates).

The occurrence of dispersed dolomicrospar crystals in C1 microspar suggests that the measured $\delta^{13}\text{C}_{\text{carb}}$ values includes some contribution from the dolomite. However, based on visual estimates from numerous siliceous calcilutite and very fine- to fine-grained siliceous calcarenite thin sections, in addition to X-ray diffraction results from two siliceous calcilutites, suggests that dolomicrospar makes up less than 5% of the carbonate content in these strata, and therefore probably had negligible effect on the measured $\delta^{13}\text{C}_{\text{carb}}$ values.

The $\delta^{13}\text{C}_{\text{carb}}$ results from CC and HS display similar trends suggesting isotopic variation in the FIC is repeatable at the km-scale (Figure 3.6), and therefore data can be combined into a single generalized FIC log (Figure 3.8). $\delta^{13}\text{C}_{\text{carb}}$ values at the base of CT 1 are depleted (-3.8‰ to -5.2‰) and gradually increase upsection, crossing over to positive values near the base of CT 2. In the basal two metres of CT 2 $\delta^{13}\text{C}_{\text{carb}}$ increases abruptly to 2‰ before plateauing at about 2.0‰ - 3.0‰ up to 123-1. At the top of 123-1 isotope values fall to $\sim 0\text{‰}$ and remain there until 123-2 where values become negative and gradually decrease upward. Then in CT 3 $\delta^{13}\text{C}_{\text{carb}}$ more rapidly decreases to a nadir at the top of the FIC (-6.3‰ in CC and -2.7‰ in HS) before being truncated by the base of Isaac channel 1. Despite having depleted $\delta^{13}\text{C}_{\text{carb}}$ values at the base and top of the section, carbonates are predominantly positive ($\sim 60\%$ of samples) throughout the FIC, with most between 2‰ - 3‰ , similar to other post-Marinoan interglacial successions (e.g. Narbonne et al., 1994; Saylor et al., 1998; Jiang et al., 2002; Halverson et al., 2005; Kaufman et al., 2006; Le Guerroué et al., 2006a,b; Jiang et al., 2007; Kaufman et al., 2007; Zhou & Xiao, 2007; McFadden et al., 2008; Tahata et al., 2013; MacDonald et al., 2013; Wang et al., 2016). Milk River, on the other hand, displays a $\delta^{13}\text{C}_{\text{carb}}$ trend unlike CC and HS, particularly in the

lower part, with positive values persisting throughout much of the section. This disparity is interpreted to be the result of the higher metamorphic grade at MR that has altered the primary isotopic signal, making it uncorrelatable to the other sections.

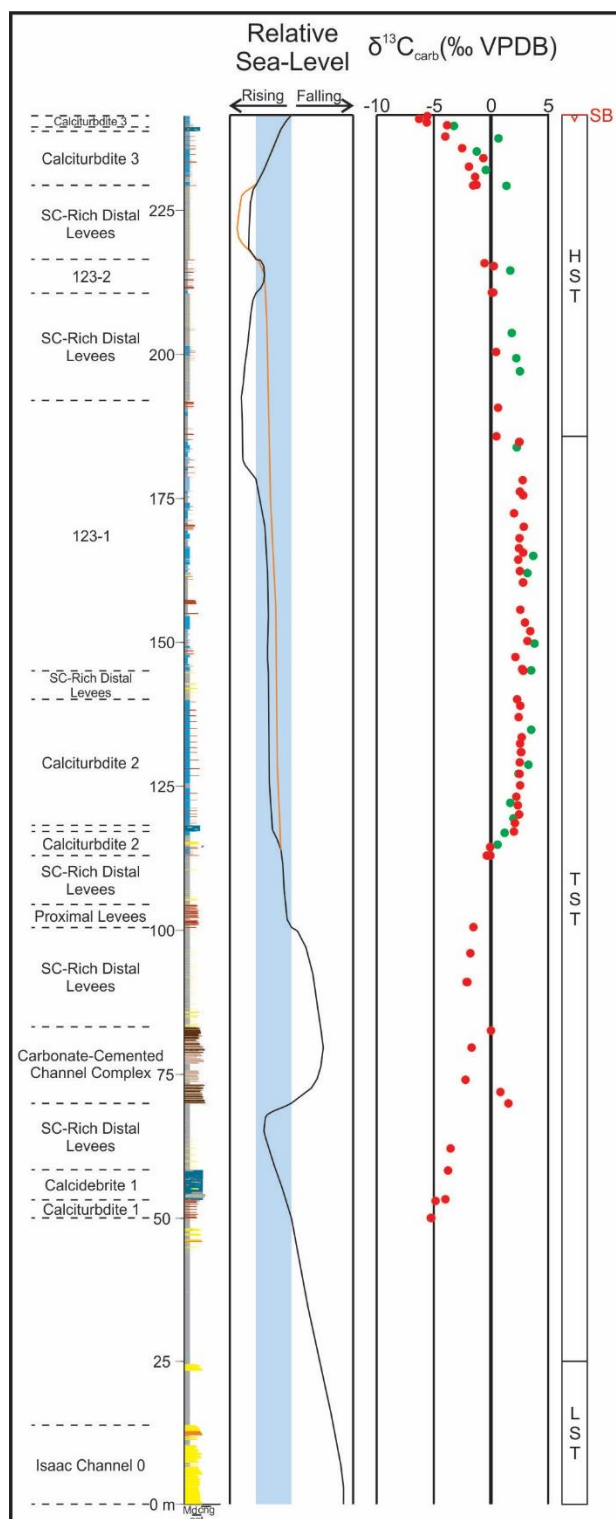


Figure 3.8: Combined FIC $\delta^{13}\text{C}_{\text{carb}}$ from CC and HS with eustatic curve and sequence stratigraphy based on the chemostratigraphy (black) and lithostratigraphy (orange; Navarro, 2016). Red circles represent samples from this study and green are the unpublished data of Ross (2003). The light blue region within the sea-level curve corresponds to carbonate production on the shelf platform. The hypothesized eustatic curve closely resembles Navarro's (2016) interpretation of the FIC based on lithological changes. Sandstone-rich Isaac channel 0, below the FIC, represents lowstand deposition (LST). Stratigraphy upward sea-level and $\delta^{13}\text{C}_{\text{carb}}$ steadily rise through the FIC, except for a short-lived regression associated with bacon channel 1, and then plateaus in the region of CT 2 to 123-1. In the upper part of 123-1 isotope values begin to steadily decline, which is interpreted to coincide with falling sea-level (HST/FSST) and continues to the top of the FIC, which then is truncated by a sequence boundary (SB) at the base of Isaac channel 1.

V.II Source of the Carbonate Isotopic Signal

Many studies of sedimentary basins from the Neoproterozoic to modern display progressively more negative $\delta^{13}\text{C}_{\text{carb}}$ values from shallow to deep water (Kroopnick, 1985; Calver, 2000; Jiang et al., 2007; Swart, 2008; Giddings and Wallace, 2009; Smith, 2009; Shen et al., 2011). During the Neoproterozoic $\delta^{13}\text{C}_{\text{carb}}$ in slope to basinal settings are reported to be up to 14‰ lighter than those in coeval nearshore strata (the difference is much reduced ($\sim 2\%$) in the modern oceans). In a previous study of the deep-marine marine OFP, which is stratigraphically ~ 1 km below the FIC, $\delta^{13}\text{C}$ values in the Temple Lake Member were 7-8‰ more negative than the correlative shallow-water Hayhook Formation in the Mackenzie Mountains (Smith, 2009). In a study of pre-Marinoan strata in South Australia, Giddings and Wallace (2009) hypothesized that the shallow- to deep-water difference was evidence that Neoproterozoic oceans were stratified due to an enhanced biological pump in which particulate organic matter (enriched in ^{12}C) produced in the shallow marine settled into the deep marine and diluted (i.e. depleted) its ^{13}C carbon reservoir. Moreover, stratification would also have helped maintain the gradient by disrupting the transport (upwelling) of ^{13}C back to the surface. Alternatively, Schrag et al. (2013) suggested that the highly depleted offshore values (i.e. as low as -16‰ in the Doushantuo Formation of China (Jiang et al., 2007; Shen et al., 2011)) resulted from enhanced authigenic carbonate precipitation in deeper waters. Authigenic carbonate refers to any carbonate mineral that precipitates at or below the sediment-water interface in association with anaerobic respiration and is typically highly depleted in $\delta^{13}\text{C}$ (MacDonald et al., 2013). Authigenesis is then thought to be more widespread with increasing depth and commensurately $\delta^{13}\text{C}_{\text{carb}}$ values to decrease (Schrag et al., 2013).

In this study of base-of-slope carbonates, $\delta^{13}\text{C}_{\text{carb}}$ values from CC and HS range from -6.3 to 3.4‰, with 60% of the samples being $> 0\%$. If we assume a shallow-to-deep isotopic gradient of 7-8‰, based on Smith (2009), then time-equivalent shallow-marine strata should have $\delta^{13}\text{C}_{\text{carb}}$ ranging

from 0.6-11.4‰, and average about 8‰, which is higher than most post-Marinoan strata studied to date. However, if the difference is of the order of 4‰, then continental shelf strata would range between -2 and 7‰, with most about ~ 4‰, which is more consistent with other Ediacaran isotopic studies (e.g. Jiang et al., 2007; Grotzinger et al., 1995; Saylor et al., 1998; Le Guerroué et al., 2006a,b; MacDonald et al., 2013; Wood et al., 2015). However, the predominance of positive values from the base-of-slope FIC contrasts results reported from many other deep-marine studies (e.g. Jiang et al., 2007; Giddings and Wallace, 2009; Shen et al., 2011). For example, $\delta^{13}\text{C}$ values from interpreted slope settings in the Doushantuo Formation of China are consistently negative, except in the uppermost 10 m of the section (Jiang et al., 2007). In fact, the average $\delta^{13}\text{C}_{\text{carb}}$ value from the FIC is more positive than the 91st percentile from the Doushantuo slope. Also, there is no statistically significant difference between the $\delta^{13}\text{C}_{\text{carb}}$ of carbonate clasts in mass-transport deposits/channel complexes and their host strata. This suggests the Ediacaran Panthalassa Ocean may have been less stratified than the deep-water basin in South China (Li et al., 2013), or conversely, and more probably, FIC carbonates first precipitated on a shallow water platform (suggested also by microbialite and oolite clasts) and retained their isotopic signature following resedimentation downslope.

In a previous study, Al-Mufti (2013) analyzed red-coloured carbonate cements in pseudo-dune cross-stratified beds in the Upper Kaza Group of the WSG. $\delta^{13}\text{C}_{\text{carb}}$ values were highly depleted (average = -18‰), suggesting that the cements were authigenic and related to methanogenic and bacterial sulfate reduction (e.g. Schrag et al., 2013). $\delta^{13}\text{C}_{\text{carb}}$ values in the FIC, however, are significantly more positive (average = 1‰), the lowest being -6.3‰ and therefore similar to $\delta^{13}\text{C}_{\text{carb}}$ from mantle input (-6±1‰), suggesting these values reflect oceanic conditions at the time of deposition/lithification rather than post-depositional authigenic carbonate precipitation.

V.III Connection Between Eustasy and $\delta^{13}\text{C}_{\text{carb}}$

The $\delta^{13}\text{C}_{\text{carb}}$ trend through the FIC at the CC and HS study areas shows a close relationship with Navarro's (2016) interpreted sea level curve based on changes in lithology (Figure 2.2, 3.7). About 100 m below the FIC is a > 35 m-thick mass transport complex (MTC) with abundant siliceous calcarenite and rare limestone clasts. This MTC is interpreted to represent an episode of slope instability related to a eustatic fall (FSST). Strata overlying the MTC are incised by Isaac channel 0, representing lowstand conditions (LST). A long-term transgression followed, and deepening conditions flooded the continental shelf and led to shallow-water carbonate production that sourced CT 1 and CT 2 (TST). This rising trend was interrupted by two short-term regressions (eustatic fall) that temporarily deactivated and eroded the platform and deposited decametre-thick channel complexes (BST 1 and BST 2) above CT 1 and CT 2, respectively (Figure 3.2, 3.4). However, the eustatic rise that followed BST 2 eventually created conditions that were not conducive to carbonate production above 123-2, either because the rate of sea level rise exceeded the rate of carbonate production and/or ecological conditions became unfavourable. Nevertheless, during the ensuing eustatic fall (late HST/FSST), sea level re-entered the carbonate production window with the deposition of CT 3 before being completely terminated with the deposition of coarse siliciclastic sediment associated with Isaac channel 1.

In both CC and HS study areas $\delta^{13}\text{C}_{\text{carb}}$ is depleted (light) at the base of CT 1 (as low as -5.2‰) and progressively increases upward before plateauing in CT 2 (2-2.5‰) (Figure 3.6, 3.8), suggesting that rising sea level increased productivity and, in turn, organic carbon burial, resulting in more positive $\delta^{13}\text{C}_{\text{carb}}$ values (Hayes et al., 1999). A small-magnitude negative $\delta^{13}\text{C}_{\text{carb}}$ excursion briefly interrupts this increasing trend and coincides with the base of BST 1. The plateau in CT 2 continues into the upper part of 123-1 where after $\delta^{13}\text{C}_{\text{carb}}$ declines to ~ -6.3‰ in CT 3, suggesting continuously declining organic productivity.

Combining the lithological and geochemical data suggests that the final regression in the FIC and onset of the highstand systems tract (HST) occurred earlier than originally proposed by Navarro (2016) (Figure 3.8). The continued decrease in $\delta^{13}\text{C}_{\text{carb}}$ above 123-1 suggests regression continued during the growth of a large carbonate platform (CT 3) that subsequently was erosively overlain by lowstand (LST) deposition of Isaac Channel 1, whose base is interpreted to be a (deep-marine) sequence boundary (Navarro, 2005; Navarro, 2016).

The observed relationship between changes in $\delta^{13}\text{C}_{\text{carb}}$ and eustasy have also been identified in many other Neoproterozoic studies (e.g. Nogueira et al., 2007; Giddings and Wallace, 2009; Smith, 2009; Greenman, 2017). These consistent changes, therefore, suggest that eustasy exerts a first order control, not only on the lithological make-up of the stratigraphic record, but also its isotopic composition.

V.IV Correlation to the Windermere Supergroup in the Mackenzie Mountains?

To determine the accuracy of isotope geochemical data it must be demonstrated that the signal is repeatable at multiple scales. In this study the carbonate carbon isotopic and stratigraphic data from this study are repeatable and consistent at kilometre spatial scale, with the next step being to establish relationships with Ediacaran strata from other parts of the Windermere basin, for example upper shelf strata of the WSG in northwestern Canada (Narbonne and Aitken, 1995; Ross and Arnott, 2007; Smith, 2009).

Here the Ravensthorpe, Hayhook and Sheepbed formations, which were correlated with the OFP in the SCC (Smith, 2009; Smith et al., 2011), represent one of four large-scale depositional sequences in the WSG, and were deposited in response to post-Marinoan glacioeustatic sea level rise (McNaughton et al., 2000; Pyle et al., 2004; MacDonald et al., 2013). These strata are then unconformably overlain by the June beds, which represent the lowstand systems tract (LST) at the base of the next sequence. The basal part of this unit consists of matrix-supported debrites containing oolite and sandstone clasts, representing a period of slope instability associated with lowered sea level. This

mass-wasting-dominated section is succeeded by interbedded shale and limestone, marking the transition to TST deposits, which then is overlain by highstand deposits (HST) of the limestone-rich Gametrail Formation (MacDonald et al., 2013). Inorganic (carbonate) carbon isotopes in the basal June beds are positive (most $> 5\text{‰}$), but ~ 200 m above the base of the unit decrease to just above 0‰ (Figure 3.9). Subsequently, there is a rise to a maximum around 5‰ and plateaus in the overlying TST before dropping to values as low as -8‰ at the base of the Gametrail Formation (HST) (MacDonald et al., 2013). In some sections the magnitude of this excursion is $> 10\text{‰}$, but $\delta^{13}\text{C}_{\text{carb}}$ -values rebound to near 0‰ at the top of the Gametrail.

The lithological and sequence stratigraphic interpretation of the June beds and Gametrail Formation resembles the model of Navarro (2016) for the Kaza-Isaac transition zone and overlying FIC in the SCC. Also, the negative to positive to negative $\delta^{13}\text{C}_{\text{carb}}$ trend in the upper June beds and lower Gametrail (~ 200 m) is similar to that observed in the FIC. The negative excursion in the uppermost FIC is similar in magnitude (a 9‰ decrease in $\delta^{13}\text{C}_{\text{carb}}$) with a nadir of -6.5‰ compared to -8‰ in the Gametrail. Nevertheless, in spite of the consistent isotopic and lithological trends between the two sections, any correlation is equivocal due to the lack of age control in the SCC.

Additionally, MacDonald et al. (2013) interpreted the Sheepbed Formation and overlying June beds to be separated by an unconformity, which based on biostratigraphical data and the absence of Gaskiers glacial deposits, represented a hiatus of about 34 myr (~ 614 to < 580 Ma). Accordingly, these authors proposed that the Gametrail excursion correlates to the global Shuram-Wonoka anomaly (Figure 3.9) observed in several localities worldwide; including Oman (Burns and Matter, 1993; Le Guerroué et al., 2006 a,b; Fike et al., 2006), Australia (Pell et al., 1993; Calver et al., 2000; Husson et al., 2012), India (Jiang et al., 2002; Kaufman et al., 2006), Death Valley, U.S.A. (Corsetti and Kaufman, 2003; Kaufman et al., 2007; Bergmann et al., 2011; Verdel et al., 2011), Norway (Rice et al., 2011). However, the age of this anomaly has been widely contested (i.e. Corsetti and Kaufman, 2003; Le Guerroué et al., 2006 a,b; McFadden et al., 2008; Husson et al., 2012), with the only definitive

date being 551 Ma (U-Pb), which coincides with the excursion's return to positive values in the Doushantuo Formation in China (Condon et al., 2005), and therefore its minimum age. The anomaly in the Mackenzie Mountains was assumed to have occurred after the first appearance of Ediacara-type fossils (~ 580 Ma), but before extensive bioturbation in the overlying Blueflower Formation (~ 547 Ma), with the most likely date being ~ 560 Ma (Narbonne et al., 2012; MacDonald et al., 2013). However, utilizing Ediacaran assemblages as chronostratigraphic markers remains controversial, as their distribution can be influenced by facies and depositional environment (Gehling and Droser, 2013).

In the SCC, the minimum age of the WSG is constrained by a ~ 570 Ma (U-Pb) date in the Hamill Group of southeastern British Columbia, which occurs above the unconformity marking the top of the WSG (Colpron et al., 2002). Therefore, if the Shuram-Wonoka anomaly occurred around 560 Ma, as proposed by MacDonald et al. (2013), it postdates deposition of the uppermost part of the WSG in the SCC. Even if the excursion is closer to the 580 Ma maximum (MacDonald et al., 2013), it seems unlikely that the 2.5 km of strata that occurs between the FIC and base of the Cambrian was deposited in less than 10 myr, which makes the correlation between the lower Isaac-FIC and June beds-Gametrail successions questionable.

V.V Correlation to Worldwide Ediacaran $\delta^{13}\text{C}_{\text{carb}}$ Datasets

An alternative to the correlation presented above is that the FIC was deposited during the hiatus represented by the Sheepbed-June beds unconformity, and therefore represents a negative anomaly not preserved in the WSG of northwestern Canada. Several post-Marinoan studies from the Doushantuo and Dengying formations of China, for example, have reported four negative $\delta^{13}\text{C}_{\text{carb}}$ anomalies (EN 1-4), with the first occurring in a post-Marinoan cap carbonate (~ 635 Ma) and the last marking the Precambrian-Cambrian boundary (542 Ma) (Figure 3.9; Zhou and Xiao, 2007; Jiang et al., 2007; McFadden et al., 2008; Tahata et al., 2013). The top of EN3 coincides with the 551 Ma U-Pb date for the upper part of the Shuram-Wonoka anomaly (Condon et al., 2005), but its base and the older EN2

excursion are poorly constrained chronologically. Approximately 20-25 m below the base of EN2 in South China, $\delta^{13}\text{C}_{\text{carb}}$ values shift from $\sim 0\text{‰}$ to $\sim 6\text{‰}$ before dropping to as low as -10‰ during the excursion (Jiang et al, 2007; McFadden et al., 2008). The shift from near-zero to positive (6‰) values might correspond to the negative to positive trend at the base of the FIC, and much like the negative excursion in the SCC, EN2 occurs near the top of a shallowing upward succession (Jiang et al., 2007; McFadden et al., 2008). However, this excursion is overlain by a major flooding surface in China compared to a sequence boundary in this study. A possible explanation for this discrepancy is that the flooding surface that overlies EN2 in China may in fact be an amalgamated sequence boundary-flooding surface, which is not uncommon in shallow marine strata.

Previous studies have correlated EN2 to negative excursions in the Infra Krol Formation in India (Jiang et al., 2002; Kaufman et al., 2006), Rainstorm Member of the Johnnie Formation in Death Valley (Kaufman et al., 2007) and Karibib Formation in Namibia (Halverson et al., 2005; MacDonald et al., 2013) (Figure 3.9), with the Infra Krol and Johnnie excursions also coinciding with an overlying sequence boundary. The correlation to Death Valley is the least reliable due to the sparsity of carbonate strata in its Ediacaran section, resulting in a lack of isotopic data following the negative values in the Rainstorm Member (Figure 3.9). $\delta^{13}\text{C}_{\text{carb}}$ is depleted at both the top of the Johnnie Formation and in the middle of the overlying Stirling Formation, but the two are separated by ~ 350 m of strata. Thus, it is unknown whether these are part of a single event or multiple excursions. Also, Kaufman et al. (2007) noted that it is unlikely that the Rainstorm excursion corresponds to the Shuram-Wonoka anomaly as it would imply an exceptionally high sedimentation rate (> 250 m/Ma) for the remaining shallow-marine, passive margin Ediacaran succession in Death Valley. Therefore, it is plausible that the negative $\delta^{13}\text{C}_{\text{carb}}$ in the Rainstorm could correlate to the older EN2 excursion.

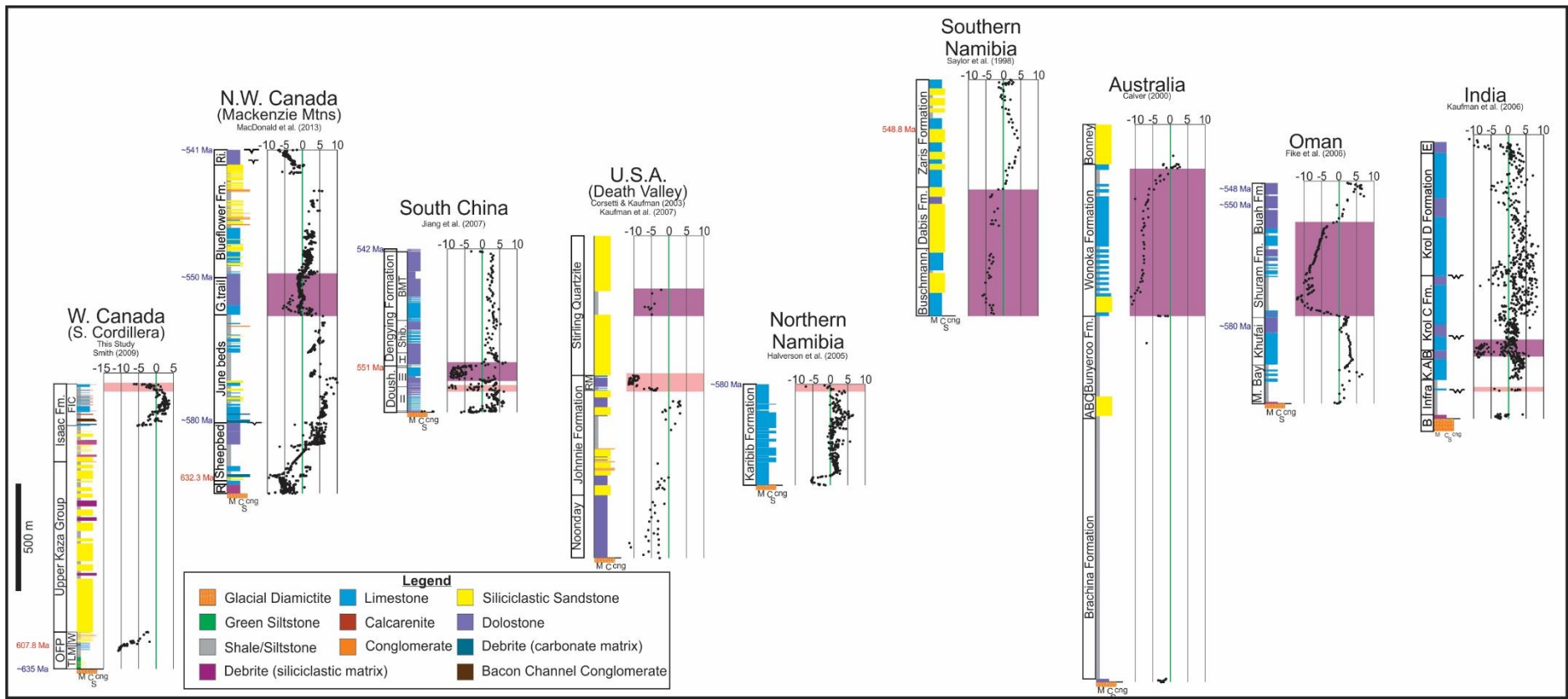


Figure 3.9: Comparison of $\delta^{13}\text{C}_{\text{carb}}$ chemostratigraphies for the Ediacaran in nine regions: the SCC in western Canada (Smith, 2009; this study), Mackenzie Mountains in northwestern Canada (MacDonald et al., 2013), South China (Jiang et al., 2007), Death Valley in western U.S.A. (Corsetti and Kaufman, 2003; Kaufman et al., 2007), northern (Halverson et al., 2005) and southern (Saylor et al., 2008) Namibia, Australia (Calver, 2000), and India (Kaufman et al., 2006). Datum is the base of the EN-2 excursion. Excursions associated with the Shuram-Wonoka anomaly are highlighted in purple whereas those potentially correlatable to EN-2 in China (Gaskiers glaciation) are highlighted in pink. Absolute age constraints (e.g. U-Pb or Re-Os) are in red, and inferred ages (based on biostratigraphy or stratigraphic correlation) are in blue and are from Grotzinger et al. (1995), Kendall et al. (2004), Condon et al. (2005), MacDonald et al. (2013) and Rooney et al. (2015).

A more recent isotopic study in drill core from South China found that negative $\delta^{13}\text{C}_{\text{carb}}$ values associated with EN2 corresponded to elevated $\delta^{18}\text{O}$ values (up to +2‰), suggesting that this excursion was associated with a post-glacial eustatic rise that likely corresponded to the termination of the Gaskiers glaciation (~ 580 Ma) (Tahata et al., 2013). There is also potential evidence of coeval glaciation in the Death Valley Ediacaran section as the Rainstorm Member is scoured by an incised valley whose base might coincide with a glacially controlled eustatic fall (Christie-Blick and Levy, 1989; Abolins et al., 2000). Therefore, if the correlation of the excursion at the top of the FIC to EN2 is correct it could represent an event associated with this late-Ediacaran glaciation.

Currently there are few reliable dates from the WSG, and none from the areas in this study. Current age control is limited to the controversial 607.8 Ma Re-Os date from the underlying OFP (Kendall et al., 2004; see Part IV of Chapter 1 of this thesis and/or Section 5.9.4 of Smith (2009)) and a 570 Ma U-Pb date from rocks that unconformably overlie the WSG (Colpron et al., 2002). Thus, it is difficult to correlate the negative excursions at the top of the FIC with the global Gaskiers event as stratigraphic evidence for post-Marinoan glaciation is lacking in all potentially correlative sections (Figure 3.9); including the Mackenzie Mountains (Narbonne et al., 1994; MacDonald et al., 2013), Namibia (Halverson et al., 2005; Saylor et al., 1998), Death Valley (Corsetti et al., 2003; Kaufman et al., 2007), India (Kaufman et al., 2006) and China (Jiang et al., 2007; Tahata et al., 2013). The unconformity in the WSG in the Mackenzie Mountains (~ 614 to < 580 Ma) may have removed shallow-marine Gaskiers glacial deposits, whereas the deep-marine section of the WSG in the SCC may represent continuous sedimentation and therefore more complete stratigraphic preservation, which might potentially fill this gap in geochemical data. However, there is no lithological evidence of post-Marinoan glaciation in the SCC (i.e. remobilized diamictites, extrabasinal basement clasts, or dropstones (McMechan, 2000; Smith et al., 2011)), making it difficult to definitely attribute the upper FIC excursion to the Gaskiers glaciation.

Given the suite of data available this correlation is tentative and needs to be strengthened with better age control both within and beyond the Cariboo Mountains. A thorough, basin-wide study of the carbonate-rich successions in the WSG in the SCC, for example the second Isaac carbonate and overlying Cunningham Formation would, therefore, be beneficial, and help to expand the $\delta^{13}\text{C}_{\text{carb}}$ database from this study.

Chapter 4: Thesis Conclusions and Areas for Future Research

Part I Conclusions

This study builds on the earlier work of Navarro (2016) at Castle Creek (CC) and presents the results and interpretations from the first detailed analysis of the stratigraphic make-up of the first Isaac carbonate (FIC) at the Hill Section (HS) and Milk River (MR) study areas. It also is the first to analyze the inorganic carbon isotopic ($\delta^{13}\text{C}_{\text{carb}}$) evolution of base-of-slope strata in the study areas and ascertain if FIC carbonate strata retained their shallow-marine source area signatures, or were modified by later diagenetic processes. If shown to be primary, these data could then be compared with other Ediacaran (Neoproterozoic) successions worldwide and thereby potentially place the Windermere Supergroup in the southern Canadian Cordillera (SCC) into a more global context.

Strata of the FIC comprise seven stratal elements, each composed of a unique combination of stratal geometry, scale and lithological composition: channel complexes, proximal levees, distal levees (siliciclastic- and carbonate-rich), “123” units, scour dominated units, debrites and slides. Correlation of these elements between the study areas allows the FIC to be subdivided into two parts separated vertically by bacon sandstone 1 (BST 1). The lower part represents ~ 20% of FIC stratigraphy and is characterized by stratal units that can be correlated across the study areas, specifically BST 1 and calciturbidite 1 (CT 1). Additionally, the unit is more or less isopachous, which collectively suggests spatially uniform patterns of sediment transport and deposition that are manifest stratigraphically as sheetlike stratal geometries composed of similar elements. In contrast, the upper 80% of the FIC sediment pile is decidedly more lithologically diverse and composed of more areally restricted stratal elements – specifically channel complexes are better developed at CC-HS compared with siliciclastic mud-rich distal levees at MR. Additionally, the FIC section at MR is ~ 70-100 m thicker with more abundant mass transport deposits compared to the correlative section at CC-HS. This suggests that during early FIC development (lower part) channel belts were mobile and wandered across much of

the slope. However, following deposition of BST 1 patterns of deposition and transport changed (upper part) with the primary transport fairway for basinward-moving turbidity currents being located near CC-HS, evinced by more common channels and a thinner stratal pile. Meanwhile, MR was situated on the margins of the fairway where higher rates of net sedimentation built up a thick pile of mostly siliciclastic, mud-rich distal levee deposits. Also, the more common occurrence of mass transport deposits at MR is interpreted to reflect spatial differences in slope stability. At CC-HS the greater frequency of high-energy turbidity currents is interpreted to have continuously maintained the slope at a lower, and therefore more gravitationally stable angle compared to that upslope of the MR study area.

Strata in the FIC contain up to eight different cements whose origins range from early seafloor recrystallization to metamorphic alteration. C1 is interpreted to be the primary cement due to its dirty appearance in stained thin-section and non-luminescence. C1 is followed and enveloped by the more voluminous C2, which consists predominantly of iron-poor microspar and represents early burial diagenetic recrystallization. Due to its abundance, most $\delta^{13}\text{C}_{\text{carb}}$ measurements in very early diagenetic cements were made in C2, but, despite being a slightly later phase than C1, likely had little impact on the final results (e.g. Smith, 2009). In siliceous calcilutite and fine-grained siliceous calcarenite samples C2 ($\pm$ C1) represents most of the carbonate cements (> 85 %), whereas medium- to coarse-grained siliceous calcarenite and granule siliceous calcirudite contained a greater proportion of later stage diagenetic and metamorphic cements. This suggests that finer-grained samples were sufficiently impermeable to have limited post-depositional fluid invasion and attendant alteration, and, therefore, preserve primary (seawater) isotopic values.

The $\delta^{13}\text{C}_{\text{carb}}$ results at CC and HS are similar (~ 3 km apart laterally). Near the base of FIC (in CT 1) values are initially depleted (-3.8‰ to -5.2‰) and then gradually increase upsection, crossing over to positive values near the base of CT 2. $\delta^{13}\text{C}_{\text{carb}}$ then rises abruptly rises to 2.0‰ - 3.0‰ and remains more or less constant until the top of 123-1 (~ 50 m). Above 123-1 values steadily decrease and become negative at the top of 123-2. Further upward $\delta^{13}\text{C}$ continues to decrease, but near the base of CT 3

steepens significantly and reaches a minimum of -6.3‰ at the top of the CT 3 section, which then is erosively overlain by the base of Isaac channel 1. A similar pattern of changing $\delta^{13}\text{C}_{\text{carb}}$ is not observed at MR (~20 km downdip from CC-HS), which is interpreted to be a consequence of the higher metamorphic grade at MR that overprinted the $\delta^{13}\text{C}_{\text{carb}}$ signal and rendered it unusable.

FIC $\delta^{13}\text{C}_{\text{carb}}$ values are significantly heavier (more positive) than those reported from other Neoproterozoic deep-marine successions. Instead, these data are more consistent with a shallow-marine origin (i.e. Mackenzie Mountains, South China), suggesting that FIC strata retained their shallow seawater isotopic signature even after being resedimented downslope. Also, these data are inconsistent with a stratified Panthalassa Ocean at this time (e.g. Jiang et al., 2007; Giddings and Wallace, 2009) or post-depositional authigenic carbonate precipitation (e.g. Schrag et al., 2013).

The CC-HS $\delta^{13}\text{C}_{\text{carb}}$ trend closely parallels the interpreted eustatic evolution of the FIC based on changes in lithology (Navarro, 2016). The FIC was interpreted to have been initiated during a long-term (3rd-order) eustatic rise (TST) that periodically resulted in the development of a shallow-water carbonate platform that provided carbonate sediment to CT 1 and CT 2. Continued sea level rise drowned the platform (above 123-2) and not until the ensuing sea level fall (FSST) did water depths return to the carbonate production window with the deposition of CT 3. Continued sea level fall eventually exposed the shelf which terminated carbonate sedimentation and initiated the deposition of Isaac channel 1 during lowstand. Superimposed on this long-term (3rd-order) trend were shorter duration (4th/5th-order) sea level falls that formed decametre-thick, generally coarse-grained siliciclastic-carbonate-filled channel complexes (i.e. BST 1 and BST 2). The only discrepancy between Navarro's interpretation and the geochemical data presented here is that the onset of the HST/FSST, and thus sea level fall, occurred at the top of 123-1 rather than 123-2.

The sequence stratigraphic framework and $\delta^{13}\text{C}_{\text{carb}}$ evolution of the upper Kaza-Isaac transition zone and overlying FIC in the SCC closely resembles the upper June beds-Gametrail Formation succession in the WSG of the Mackenzie Mountains (MM) in the northern Canadian Cordillera

(MacDonald et al., 2013) Here the basal Gametrail negative excursion exhibits a similar range in value ($\sim 8\text{-}9\text{‰}$), with minima of -8‰ and maxima of $\sim 5\text{‰}$, compared with -6.3‰ and 3.5‰ in the SCC. In the NCC this excursion is interpreted to be associated with the Shuram-Wonoka anomaly, which is reported to have begun $\sim 570\text{-}560$ Ma. However, the top of the WSG in the SCC (~ 2.5 km above the FIC) is dated at ~ 570 , suggesting that the correspondence is likely coincidental rather than factual.

Alternatively, deposition of the FIC occurred during the hiatus represented by the Sheepbed-June beds unconformity ($\sim 614 - < 580$ Ma), and therefore represents a negative anomaly not preserved in the WSG in the NCC. An isotopic and sequence stratigraphic history similar to that in the FIC is represented by the EN2 excursion in the Doushantuo Formation of South China (Jiang et al., 2007; McFadden et al., 2008; Tahata et al., 2013) and its interpreted correlatives in India (Kaufman et al., 2006), Death Valley (Corsetti and Kaufman, 2003; Kaufman et al., 2007), and Namibia (Saylor et al., 1998; Halverson et al., 2005), which have been proposed to be associated with the Gaskiers glaciation (~ 580 Ma; Tahata et al., 2013). The Gaskiers, however, is the most controversial of the Neoproterozoic glaciations due to its diminished global distribution relative to the Marinoan and Sturtian glaciations, its reported short duration (< 1 myr), and lack of globally correlative cap carbonates. Notably, sedimentological evidence of the Gaskiers glaciation, for example diamictites, extrabasinal basement clasts and dropstones, is absent in the Windermere Supergroup in both the SCC and NCC. Nevertheless, the deep-marine setting of the FIC in the SCC should have preserved a more complete stratigraphic section compared to the generally shallow-water setting of Neoproterozoic rocks in the NCC where erosion may have removed part of the stratigraphic section. Therefore, in the SCC the FIC could potentially represent the terminus of the interglacial preceding the Gaskiers and the overlying Isaac channel 1 the ensuing glacial lowstand, despite lacking sedimentological evidence of glaciation. If correct this would suggest that the Gaskiers was indeed a more widespread phenomenon, but requires better age control both within and beyond the SCC to correlate the upper FIC excursion with EN2. To aid in this the compilation of a more complete $\delta^{13}\text{C}_{\text{carb}}$ record through the inclusion of other carbonate-

rich intervals in the deep-water WSG (i.e. second Isaac carbonate, Cunningham and Yankee Belle formations) would help to better test and constrain this correlation.

Part II Areas for Future Research

Two broad areas that would advance the work presented here are: a more detailed sedimentological analysis of particular stratal units in the FIC (1-3 below); and a more regional geochemical analysis of the FIC and other carbonate intervals in the SCC to advance our understanding of the oceanographic evolution of $\delta^{13}\text{C}_{\text{carb}}$ during the late Neoproterozoic (4 and 5).

- 1) A comprehensive study of the enigmatic 123 units. The rhythmic and apparent systematic changes in sediment supply, both in terms of granulometry and mineralogy, to the deep-water Windermere basin could potentially indicate eustatic changes on significantly shorter timescales than what has previously been interpreted from the WSG. This was attempted in this study by tracing the basal medium to coarse-grained sandstone (1) packages laterally in 123-1 and 123-2 at CC-S. A similar analysis could be conducted in CC-N, HS, MR and any other FIC exposures where 123 lithologies crop out. This larger database would then be more amenable to statistical analysis (i.e. Markov-chain) and strengthen any interpretation about the natural forcing mechanisms – mechanisms which might include systematic variations in Earth's orbital (i.e. Milankovitch) patterns.
- 2) Further in-depth analysis on the vertical and lateral trends in the bacon sandstone channel complexes and their component channel fills. These data could then be compared to similar elements in siliciclastic systems, for example in other parts of the Isaac Formation and in carbonate or other mixed systems (i.e. The Bahamas).
- 3) A study on the PL 2 to BST 4 transition at MR, which couldn't be properly ascertained in this study due to time constraints. Multiple stratigraphic logs could be measured between

- MR-S and MR-N to more thoroughly document the lateral lithological changes, and therein test the model of deposition on the inner and outer bends of a single sinuous channel.
- 4) Analyze the sedimentological and, more specifically, geochemical make-up of the FIC in other parts of the SCC to better assess its regional correlatability and strengthen/update the global correlations attempted in this study.
 - 5) In addition to expanding the study areally, it could be temporally (i.e. stratigraphically upward) expanded by including carbonate-rich units above the FIC, including the SIC, Cunningham Formation and occasional limestones in the Yankee Belle Formation. This would be similar to the study by MacDonald et al. (2013) in the Mackenzie Mountains and significantly expand the time interval being sampled. Accordingly, this would provide a much longer-term assessment of paleo-oceanographic conditions and allow for more accurate comparison and potential correlation with other Ediacaran intervals globally.

References

- Aalto, K.R. 1971. Glacial marine sedimentation and stratigraphy of the Toby Conglomerate (Upper Proterozoic), southeastern British Columbia, Northwestern Idaho and Northeastern Washington, *Canadian Journal of Earth Sciences*, **8**: 753-787.
- Abolins, M., Oskin, R., Prave, T., Summa, C. and Corsetti, F.A. 2000. Neoproterozoic glacial record in the Death Valley region, California and Nevada. *In* Great Basin and Sierra Nevada: Boulder, Colorado. *Edited by* D.R.P. Lageson, G. Stephen, M.M. Lahren. Geological Society of America Field Guide, **2**: 319-335.
- Abreu, V., Sullivan, M., Pirmex, C., Mohrig, D., and Mutti, E. 2003. Lateral accretion packages (LAPs); an important reservoir element in deep water sinuous channels, *Marine and Petroleum Geology*, **20**: 631-648.
- Aitken, J.D. 1969. Documentation of the sub-Cambrian unconformity, Rocky Mountain Main Ranges, Alberta, *Canadian Journal of Earth Sciences*, **6**: 193-200.
- Aitken, J.D. 1982. Precambrian geology of the Mackenzie Fold Belt – a stratigraphic and tectonic overview. *In* Precambrian Sulphide Deposits. *Edited by* R.W. Hutchinson, C.D. Spence and J.M. Franklin. Geological Association of Canada, Waterloo, Ontario, pp. 150-161.
- Aitken, J.D. 1991a. Two Late Proterozoic glaciations, Mackenzie Mountains, northwestern Canada, *Geology*, **19**: 445-448.
- Aitken, J.D. 1991b. The Ice Brook Formation and post-Rapitan, Late Proterozoic glaciation, Mackenzie Mountains, Northwest Territories, *Geological Survey of Canada Bulletin* 404, 43 p.
- Al-Mufti, O. 2013. Dune or dune-like cross-stratification in deep-marine sandstones of the Neoproterozoic Windermere Supergroup, Cariboo Mountains, British Columbia, Canada. B.Sc. thesis, University of Ottawa, Ottawa, ON.
- Allen, J.R.L. 1969. Some recent advances in the physics of sedimentation, *Proceedings of the Geologists' Association*, **80**: 1-42.
- Allen, J.R.L. 1970. The sequence of sedimentary structures in turbidites, with special reference to dunes, *Scottish Journal of Geology*, **6**: 146-161.
- Altosaar, A. 2007. Stratigraphic architecture and sedimentology of interstratified splay, levee and channel deposits, Isaac Formation (Windermere Supergroup), southern Canadian Cordillera. B.Sc. thesis, University of Ottawa, Ottawa, ON.
- Anderson, T.F., and Arthur, M.A. 1983. Stable isotopes of oxygen and carbon and their application to sedimentologic and paleoenvironmental problems. *In* Stable Isotopes in Sedimentary Geology: SEPM Short Course No. 10. Society for Sedimentary Geology, Tulsa, OK, USA, pp. 1-1-1-151.

- Angus, K. 2016. Lateral facies trends in deep-marine slope and basin floor matrix-rich beds, Neoproterozoic Windermere Supergroup, British Columbia, Canada. M.Sc. thesis, University of Ottawa, Ottawa, ON.
- Anthony, D. 2011. Architecture and depositional evolution of a Neoproterozoic deep marine slope channel complex: Isaac Formation, Windermere Supergroup, Southern Canadian Cordillera. B.Sc. thesis, University of Ottawa, Ottawa, ON.
- Arnaud, E., and Eyles, C.H. 2006. Neoproterozoic environmental change recorded in the Port Askaig Formation, Scotland; climatic vs. tectonic controls, *Sedimentary Geology*, **183**: 99-124.
- Arnott, R.W.C. 2007a. Stratal architecture and origin of lateral accretion deposits (LADs) and conterminuous inner-bank levee deposits in a base-of-slope sinuous channel, lower Isaac Formation (Neoproterozoic), East-Central British Columbia, Canada, *Marine and Petroleum Geology*, **24**: 515-528.
- Arnott, R.W.C. 2007b. Stratigraphic architecture and depositional processes of a proximal crevasse splay and genetically related, sinuous channel fill, Isaac Formation, British Columbia, Canada. *In Atlas of deep-water outcrops: AAPG Studies in Geology 56. Edited by T. H. Nilsen, R. D. Shew, G. S. Steffens, and J. R. J. Studlick*, CD-ROM, 12 p.
- Arnott, R.W.C. 2010. Deep-marine sediments and sedimentary systems. *In Facies Models 4. Edited by N.P. James and R.W. Dalrymple*. Geological Association of Canada, St. John's, NF, Canada, pp. 295-322.
- Arnott, R.W.C. 2012. Turbidites, and the case of the missing dunes, *Journal of Sedimentary Research*, **82**: 379-384.
- Arnott, R.W.C., and Al-Mufti, O. 2017. Deep-marine pseudo dune cross-stratification – Similar, but not completely different, *Journal of Sedimentary Research*, **87**: 312-323.
- Arnott, R.W.C., and Hand, B.M. 1989. Bedforms, primary structures and grain fabric in the presence of suspended sediment rain, *Journal of Sedimentary Petrology*, **59**: 1062-1069.
- Arnott, R.W., and Hein, F.J. 1986. Submarine canyon fills of the Hector Formation, Lake Louise, Alberta: Late Precambrian syn-rift deposits of the proto-Pacific miogeocline, *Bulletin of Canadian Petroleum Geology*, **34**: 395-407.
- Arnott, R.W.C., Wallace, K., and Laurin, J. 2011. Stratal architecture and temporal evolution of a passive margin mass-transport deposit, Neoproterozoic Isaac Formation, Cariboo Mountains, British Columbia. *In Mass-Transport Deposits in Deepwater Settings*, v. SEPM Special Publication. Edited by R.C. Shipp, P. Weimer, and H.W. Posamentier. SEPM, pp. 221-234.
- Babonneau, N., Savoye, B., Cremer, M., and Klein, B. 2002. Morphology and architecture of the present canyon and channel system of the Zaire deep-sea fan, *Marine and Petroleum Geology*, **19**: 445-467.

- Bagnold, R.A. 1962. Auto-suspension of transported sediment: turbidity currents, *Proceedings of the Royal Society of London*, **265**: 315-319.
- Banner, J.L., and Hanson, G.N. 1990. Calculation of simultaneous isotopic and trace element variations during water-rock interaction with applications to carbonate diagenesis, *Geochimica et Cosmochimica Acta*, **54**: 3123-3137.
- Basu, D., and Bouma, A.H. 2000. Thin-bedded turbidites of the Tanqua Karro: Physical and depositional characteristics. *In* *Fine-grained Turbidite Systems. Edited by A.H. Bouma and C.G. Stone*. AAPG Memoir 72/SEPM Special Publication 68, Tulsa, OK, pp. 263-278.
- Bergen, A. 2017. Vertical and lateral facies architectures of slope channels and genetically related levee deposits, Isaac Formation, Neoproterozoic Windermere Supergroup, Cariboo Mountains, B.C. M.Sc. thesis, University of Ottawa, Ottawa, ON.
- Bergen, A., and Arnott, R.W.C. 2016. Lateral and vertical changes in deep marine levee deposits – Where intuition and geological reality diverge: Neoproterozoic Isaac Formation, Windermere Supergroup, B.C., Canada. AAPG Annual Conference and Exhibition, Calgary, Alberta, Canada.
- Bergmann, K.D., Zentmyer, R.A. and Fischer, W.W. 2011. The stratigraphic expression of a large negative carbon excursion from the Ediacaran Johnnie Formation, Death Valley, Precambrian Research, **188**: 45-56.
- Boggs, S, and Krinsley, D. 2006. Application of cathodoluminescence imaging to the study of sedimentary rocks. Cambridge University Press, Cambridge, United Kingdom.
- Bouma, A.H. 1962. Sedimentology of some Flysch deposits; A graphic approach to facies interpretation. Elsevier Publishing Company, New York, U.S.A.
- Bowring, S., Myrow, P., Landing, E., Ramezani, J., and Grotzinger, J. 2003. Geochronological constraints on terminal Neoproterozoic events and the rise of Metazoans, Abstracts from the 2003 EGS-AGU-EUG Joint Assembly, Nice, France.
- Braga, J.C., Martin, J.M., and Wood, J.L. Submarine lobes and feeder channels of redeposited, temperate carbonate and mixed siliciclastic-carbonate platform deposits (Vera Basin, Almería, southern Spain), *Sedimentology*, **48**: 99-116.
- Brand, U., and Veizer, J. 1980. Chemical diagenesis of a multicomponent carbonate system – 1: Trace elements, *Journal of Sedimentary Petrology*, **50**: 1219-1236.
- Brand, U., and Veizer, J. 1981. Chemical diagenesis of a multicomponent carbonate system – 2: Stable isotopes, *Journal of Sedimentary Petrology*, **51**: 987-997.
- Buckley, S.J., Schwarz, E., Terlaky, V., Howell, J.A., and Arnott, R.W.C. 2009. Terrestrial laser scanning combined with photogrammetry for digital outcrop modelling, *ISPRS Congress 2009*, 75-80.

- Budyko, M.I. 1969. The effect of solar radiation variations on the climate of the Earth, *Tellus*, **21**: 611-619.
- Burns, S.J. and Matter, A. 1993. Carbon isotopic record of the latest Proterozoic from Oman, *Eclogae Geologicae Helvetiae*, **86**: 595-607.
- Caldeira, K., and Kasting, J.F. 1992. Susceptibility of the early Earth to irreversible glaciation caused by carbon dioxide clouds, *Nature*, **359**: 226-228.
- Calver, C.R. 2000. Isotope stratigraphy of the Ediacaran (Neoproterozoic III) of the Adelaide Rift Complex, Australia, and the overprint of water column stratification, *Precambrian Research*, **100**, 121-150.
- Calver, C.R., Crowley, J.I., Wingate, M.T.D., Evans, D.A.D., Raub, T.D., and Schmitz, M.D. 2013. Globally synchronous Marinoan deglaciation indicated by U-Pb geochronology of the Cottons Breccia, Tasmania, Australia, *Geology*, **41**: 1127-1130.
- Campbell, R.B. 1970. Structural and metamorphic transitions from infrastructure to suprastructure, Cariboo Mountains, British Columbia; structure of the southern Canadian Cordillera. *In* Geological Association of Canada, Special Paper 6. *Edited by* J.O. Wheeler. Pp. 67-72.
- Campbell, R.B., Mountjoy, E.W., and Young, F.G. 1973. Geology of McBride map-area, British Columbia, Geological Survey of Canada, Paper 72-35, 103 p.
- Campion, K.M., Sprague, A.R., Mohrig, D., Lovell, R.W., Drzewiecki, P.A., Sullivan, M.D., Ardill, J.A., Jensen, G.N., and Sickafoose, D.K. 2000. Outcrop expression of confined channel complexes. *In* Deep-water Reservoirs of the World: Gulf Coast Section. *Edited by* P. Weimer, R.M. Slarr, J. Coleman, N.C. Rosen, H. Nelson, A.H. Bouma, M.J. Styzen and D.T. Lawrence. SEPM Foundation 20th Annual Research Conference, pp. 127-150.
- Canals, M., Lastras, G., Urgeles, R., Casamor, J.L., Mienert, J., Cattaneo, A., De Batist, M., Haflidason, H., Imbo, Y., Laberg, J.S., Locat, J., Long, D., Longiva, O., Masson, D.G., Sultan, N., Trincardi, F., and Bryn, P. 2004. Slope failure dynamics and impacts from seafloor and shallow sub-seafloor geophysical data: Case studies from the COSTA project, *Marine Geology*, **213**: 9-72.
- Charlesworth, H.A.K., Weiner, J.L., Akehurst, A.J., Bielenstein, H.U., Evans, C.R., Griffiths, R.E., Remington, R.E., Stauffer, M.R., and Steiner, J. 1967. Precambrian geology of the Jasper region, Alberta, Research Council of Alberta Bulletin 23, 74.
- Christie-Blick, N. and Levy, M. 1989. Stratigraphic and tectonic framework of upper Proterozoic and Cambrian rocks in the western United States. *In* Late Proterozoic and Cambrian tectonics, sedimentation, and record of Metazoan radiation in the western United States. *Edited BY* N. Christie-Blick, M. Levy, J.F. Mount, P.W. Signor and P.K. Link. American Geophysical Union, 28th International Geological Congress field trip guide series 7-21, p. 113.
- Clark, J.D., and Pickering, K.T. 1996. Architectural elements and growth patterns of submarine channels: Application to hydrocarbon exploration, *American Association of Petroleum Geologists Bulletin*, **80**: 194-221.

- Coleman, J.L., Browne, G.H. Browne, King, P.R., Slatt, R.M., Spang, R.J., Williams, E.T., and Clemenceau, G.R. 2000. The inter-relationships of scales of heterogeneity in subsurface, deep-water E&P projects – Lessons learned from the Mount Messenger Formation (Miocene), Taranaki Basin, New Zealand. . *In* Deep-water Reservoirs of the World: Gulf Coast Section. *Edited by* P. Weimer, R.M. Slarr, J. Coleman, N.C. Rosen, H. Nelson, A.H. Bouma, M.J. Styzen and D.T. Lawrence. SEPM Foundation 20th Annual Research Conference, pp. 263-283.
- Colpron, M., Logan, J.M., and Mortensen, J.K. 2002. U-Pb zircon age constraint for late Neoproterozoic rifting and initiation of the lower Paleozoic passive margin of western Laurentia, *Canadian Journal of Earth Sciences*, **39**: 133-143.
- Condon, D., Zhu, M., Bowring, S., Wang, W., Yang, A., and Jin, Y. 2005. U-Pb ages from the Neoproterozoic Doushantuo Formation, China, *Science*, 308: 95-98.
- Cook, H.E., Hine, A.C., and Mullins, H.T. 1983. Platform margin and deep water carbonates, SEPM Short Course No. 12, pp. 573.
- Corsetti, F.A., and Kaufman, A.J. 2003. Stratigraphic investigations of carbon isotope anomalies and Neoproterozoic ice ages in Death Valley, California, *GSA Bulletin*, 115: 916-932.
- Craig, H. 1953. The geochemistry of the stable carbon isotopes, *Geochimica et Cosmochimica Acta*, 3: 53-92.
- Cronin, B.T., Hurst, A., Celik, H., and Türkmen, I. 2000. An outcrop example of channel and levee complex in an ancient deep-water slope environment, *Sedimentary Geology*, 132: 205-216.
- Cross, N.E., Cunningham, A., Cook, R.J., Taha, A., Esmatie, E., and El Swidan, N. 2009. Three-dimensional seismic geomorphology of a deep-water slope-channel system: The Sequoia field, offshore west Nile Delta, Egypt, *American Association of Petroleum Geologists Bulletin*, 93: 1063-1086.
- D'Agostini, D.P., Bastos, A.C. and Dos Reis, A.T. 2015. The modern mixed carbonate-siliciclastic Abrolhos Shelf: Implications for a mixed depositional model, *Journal of Sedimentary Research*, **85**: 124-139.
- Davis, L. 2011. Architecture of deep-marine interchannel deposits: Isaac Formation, Windermere Supergroup (Neoproterozoic), southern Canadian Cordillera, M.Sc. thesis, University of Ottawa, Ottawa, ON.
- Denyszyn, S.W., Halls, H.C., Davis, D.W., Evans, D.A.D. 2009. Paleomagnetism and U-Pb geochronology of Franklin dykes in High Arctic Canada and Greenland: A revised age and paleomagnetic pole constraining block rotations in the Nares Strait region, *Canadian Journal of Earth Sciences*, **46**: 689-705.
- Deptuck, M.E., Sylvester, Z., Pirmez, C. and O'Byrne, C. 2007. Migration-aggradation history and 3-D seismic geomorphology of submarine channels in the Pleistocene Benin-major Canyon, western Niger Delta slope, *Marine and Petroleum Geology*, **24**: 406-433.

- Derry, L.A. 2010. A burial diagenesis origin for the Ediacaran Shuram-Wonoka carbon isotope anomaly, *Earth and Planetary Science Letters*, 294: 152-162.
- Derry, L.A., Kaufman, A.J., and Jacobsen, S.B. 1992. Sedimentary cycling and environmental change in the Late Proterozoic: Evidence from stable and radiogenic isotopes, *Geochimica et Cosmochimica Acta*, 56: 1317-1329.
- Des Marais, D.J. 2001. Isotopic evolution of the biogeochemical carbon cycle during the Precambrian, *Reviews in Mineralogy and Geochemistry*, 43: 555-578.
- Devlin, W.J., and Bond, G.C. 1988. The initiation of the early Paleozoic Cordilleran miogeocline: evidence from uppermost Proterozoic-Lower Cambrian Hamill Group of southeastern British Columbia, *Canadian Journal of Earth Sciences*, 25: 1-19.
- Di Celma, C.N., Brunt, R.L., Hodgson, D.M., Flint, S.S., and Kavanagh, J.P. 2011. Spatial and temporal evolution of a Permian submarine slope channel-levee system, Karoo Basin, South Africa, *Journal of Sedimentary Research*, **81**: 579-599.
- Dickson, J.A.D. 1966. Carbonate identification and genesis as revealed by staining, *Journal of Sedimentary Research*, **36**: 491-505.
- Dolan, J.F. 1989. Eustatic and tectonic controls on deposition of hybrid siliciclastic/carbonate basinal cycles: Discussion with examples, *American Association of Petroleum Geologists Bulletin*, **73**: 1233-1246.
- Donnadieu, Y., Godd  ris, Y., Ramstein, G., N  d  lec, A., and Meert, J. 2004. A ‘snowball’ Earth climate triggered by continental break-up through changes in run-off, *Nature*, **428**: 303-306.
- Droxler, A.W., and Schalger, W. 1985. Glacial versus interglacial sedimentation rates and turbidite frequency in the Bahamas, *Geology*, **13**: 799-802.
- Droxler, A.W., Haddad, G.A., Kroon, D., Gartner, S., Wei, W., and McNeill, D. 1993. Late Pliocene (2.9 Ma) partial recovery of shallow carbonate banks on the Queensland Plateau: Signal of bank-top re-entry into the photic zone during a lowering in sea level
- Dumouchel, I.G. 2012. Effects of relative sea-level change on the stratigraphic architecture of a passive margin deep-marine channel fill complex, Windermere Supergroup, British Columbia, Canada, B.Sc. thesis, University of Ottawa, Ottawa, ON.
- Dumouchel, I.G. 2015. Stratigraphic architecture and depositional history of laterally- accreted channel fills in the lower Isaac Formation, Windermere Supergroup, British Columbia, Canada, M.Sc. thesis, University of Ottawa, Ottawa, ON.
- Dunbar, G.B., and Dickens, G.R. 2003. Massive siliciclastic discharge to slopes of the Great Barrier Reef Platform during sea-level transgression: constraints from sediment cores between 15°S and 16°S latitude and possible explanations, *Sedimentary Geology*, **162**: 141-158.
- Dunbar, G.B., Dickens, G.R., and Carter, R.M. 2000. Sediment flux across the Great Barrier Reef shelf to the Queensland Trough over the last 300 ky, *Sedimentary Geology*, **133**: 49-92.

- Eadie, B.J., Jeffrey, L.M., and Sackett, W.M. 1978. Some observations on the stable carbon isotope composition of dissolved and particulate organic carbon in the marine environment, *Geochimica et Cosmochimica Acta*, **42**: 1265-1269.
- Eisbacher, G.H. 1981. Sedimentary tectonics and glacial record in the Windermere Supergroup, Mackenzie Mountains, northwestern Canada, *Geological Survey of Canada Paper* 80-27, 40 p.
- Eisbacher, G.H. 1985. Late Proterozoic rifting, glacial sedimentation, and sedimentary cycles in the light of Windermere deposition, western Canada, *Palaeogeography, Palaeoclimatology, Palaeoecology*, **51**: 231-254.
- Evans, K.V., Aleinikoff, J.N., Obradovich, J.D., and Fanning, C.M. 2000. SHRIMP U-Pb geochronology of volcanic rocks, Belt Supergroup, western Montana: evidence for rapid deposition of sedimentary strata, *Canadian Journal of Earth Sciences*, **37**: 1287-1300.
- Evenchick, C.A., Parrish, R.R., and Gabrielse, H. 1984. Precambrian gneiss and Late Proterozoic sedimentation in north-central British Columbia, *Geology*, **12**: 233-237.
- Eyles, C.H., and Eyles, N. 1983. A glaciomarine model for Upper Precambrian diamictites of the Port Askaig Formation, Scotland, *Geology*, **11**: 692-696.
- Eyles, N., and Januszczak, N. 2004. 'Zipper-rift': A tectonic model for Neoproterozoic glaciations during the breakup of Rodinia after 750 Ma., *Earth Science Reviews*, **65**: 1-73.
- Fairchild, I.J., and Kennedy, M.J. 2007. Neoproterozoic glaciation in the Earth System, *Journal of the Geological Society*, **164**: 895-921.
- Fairchild, I.J., Marshall, J.D., and Bertrand-Sarafti, J. 1990. Stratigraphic shifts in carbon isotopes from Proterozoic stromatolitic carbonates: Influences of primary mineralogy and diagenesis, *American Journal of Science*, **290**: 46-79.
- Fairchild, I.J., Knoll, A.H., and Swett, K. 1991. Coastal lithofacies and biofacies associated with syndepositional dolomitization and silicification (Draken Formation, upper Riphean, Svalbard), *Precambrian Research*, **53**: 165-197.
- Fairchild, I.J., Fleming, E.J., Bao, H., Benn, B.I., Boomer, I., Dublyansky, Y.V., Halverson, G.P., Hambrey, M.J., Hendy, C., McMillan, E.A., Spötl, C., Stevenson, C.T.E., and Wynn, P.W. 2016. Continental carbonate facies of a Neoproterozoic panglaciation, north-east Svalbard, *Sedimentology*, **63**: 443-497.
- Fanning, C.M., and Link, P.K. 2004. U-Pb SHRIMP ages of Neoproterozoic (Sturtian) glaciogenic Pocatello Formation, southeastern Idaho, *Geology*, **32**: 881-884.
- Feary, D.A., Symonds, P.A., Davies, P.J., Pigram, C.J., and Jarrard, R.D. 1993. Geometry of Pleistocene facies on the Great Barrier Reef outer shelf and upper slope – seismic stratigraphy of sites 819, 820, and 821, *Proceedings of the Ocean Drilling Program, Scientific Results*, **133**: 327-351.

- Ferguson, C.A. 1994. Structural geology and stratigraphy of the northern Cariboo Mountains between Isaac Lake and Fraser River, British Columbia. Map, University of Calgary, Calgary, AB,. (1997-047810) 157.
- Ferri, F., Höy, T., and Friedman, R.M. 1999. Description, U-Pb age and tectonic setting of the Quesnel Lake Gneiss, east-central British Columbia, British Columbia Geological Survey Branch: Geological Fieldwork 1998, 69-80.
- Fike, D.A., Grotzinger, J.P., Pratt, L.M., and Summons, R.E. 2006. Oxidation of the Ediacaran ocean, *Nature*, **444**: 744-747.
- Folk, R.L. 1962. Spectral subdivision of limestone types. *In* Classification of Carbonate Rocks – a Symposium: American Association of Petroleum Geologists Memoir I. *Edited by* W.E. Ham. Pp 62-84.
- Folk, R.L. 1965. Some aspects of recrystallization in ancient limestones. *In* Dolomitization and Diagenesis: SEPM Special Publication No. 13. *Edited by* L.C. Pray and R.S. Murray. Pp 14-48.
- Francis, J.M., Dunbar, G.B., Dickens, G.R., Sutherland, I.A., and Droxler, A.W. 2007. Siliciclastic sediment across the North Queensland Margin (Australia). A Holocene perspective on reciprocal versus coeval deposition in tropical mixed siliciclastic-carbonate systems, *Journal of Sedimentary Research*, **77**: 572-586.
- Francis, J.M., Daniell, J.J., Droxler, A.W., Dickens, G.R., Bentley, S.J., Peterson, L.C., Opdyke, B.N., Beaufort, L. 2008. Deep water geomorphology of the mixed siliciclastic-carbonate system, Gulf of Papua, *Journal of Geophysical Research: Earth Surface*, **113**: F01S16, 22 p.
- Frenz, M., Wynn, R.B., Georgiopoulou, A., Bender, V.B., Hough, G., Masson, D.G., Talling, P.J., and Cronin, B.T. 2008. Unravelling turbidite basin fill complexity: A case study from the late Quaternary turbidite fill of the Agadir Basin, offshore Atlantic Morocco. *International Journal of Earth Sciences*, **98**: 721-733.
- Frimmel, H.E., Klötzli, U., and Seigfried, P. 1996. New Pb-Pb single zircon age constraints on the timing of Neoproterozoic glaciation and continental break-up in Namibia, *Journal of Geology*, **104**: 459-469.
- Fubara, A.E. 2014. Miocene depositional architecture in the mixed clastic-carbonate Rankin Platform: Implications for clinoform development morphodynamic diversity. AAPG Annual Conference and Exhibition, Houston, TX, U.S.A..
- Gabrielse, H. 1991. Late Paleozoic and Mesozoic terrane interactions in north-central British Columbia, *Canadian Journal of Earth Sciences*, **28**: 947-957.
- Gabrielse, H., and Campbell, R. 1991. Upper Proterozoic Assemblages. *In* Geology of the Cordilleran Orogen in Canada. *Edited by* H. Gabrielse and C.J. Yorath. Geological Society of America, G-2, pp. 125-150.

- Gammon, P.R., and Arnott, R.W.C. 2007. Source control over calciturbidite facies distributions in the Lower Isaac Carbonate, Windermere Supergroup, Canada. AAPG Annual Convention and Exhibition, Long Beach, California, U.S.A.
- Gammon, P.R., Arnott, R.W.C., and Ross, G.M. 2007. Architectural relationship between channels, levees, and debris flows: Isaac Channel 6, Castle Creek South, lower Isaac Formation, Windermere Supergroup, British Columbia, Canada. *In* Atlas of deep-water outcrops: AAPG Studies in Geology 56. *Edited by* T. H. Nilsen, R. D. Shew, G. S. Steffens, and J. R. J. Studlick, CD-ROM, 4 p.
- Gehling, J.G., and Droser, M.L. 2013. How well do fossil assemblages of the Ediacara biota tell time?, *Geology*, **41**: 447-450.
- George, A.D., Playford, P.E., Powell, C.M., and Tornatora, P.M. 1997. Lithofacies and sequence development on an Upper Devonian mixed carbonate-siliciclastic fore-reef slope, Canning Basin, Western Australia, *Sedimentology*, **44**: 843-867.
- Gerasimof, M.D. 1988. The Hobson Lake pluton, Cariboo Mountains, British Columbia, and its significance to Mesozoic and early Cenozoic Cordilleran tectonics, M.Sc. thesis, Queen's University, Kingston, ON.
- Giddings, J.A., and Wallace, M.W. 2009. Facies-dependent $\delta^{13}\text{C}$ variation from a Cryogenian platform margin, South Australia: Evidence for stratified Neoproterozoic oceans?, *Palaeogeography, Palaeoclimatology, Palaeocology*, **271**: 196-241.
- Godd  ris, Y., Donnadieu, Y., N  d  lec, A., Dupr  , B., Dessert, C., Grard, A., Ramstein, G., Francois, L. 2003. The Sturtian 'snowball' glaciation: Fire and ice, *Earth and Planetary Science Letters*, **211**: 1-12.
- Grabau, A.W. 1904. On the classification of sedimentary rocks, *The American Geologist*, **33**: 228-247.
- Grammer, G.M., Ginsburg, R.N., and Harris, P.M. 1993. Timing of deposition, diagenesis, and failure of steep carbonate slopes in response to a high-amplitude/high-frequency fluctuation in sea level, Tongue of the Ocean, Bahamas. *In* Carbonate Sequence Stratigraphy: Recent developments and applications, AAPG Memoir 57. *Edited by* R.G. Loucks and J.F. Sarg. Tulsa, OK.
- Greenman, J.W. 2017. Characterization of a ~1 billion year old carbonate ramp in the Brock Inlier, Arctic Canada, M.Sc. thesis, Carleton University, Ottawa, ON.
- Gregory, R.T., and Taylor, H.P. 1981. An oxygen isotope profile in a section of Cretaceous oceanic crust, Samail Ophiolite, Oman: Evidence for $\delta^{18}\text{O}$ buffering of the oceans by deep (>5 km) seawater-hydrothermal circulation at mid-ocean ridges, *Journal of Geophysical Research*, **84**: 2727-2755.
- Grotzinger, J.P., and Knoll, A.H. 1995. Anomalous carbonate precipitates; is the Precambrian the key to the Permian?, *Palaaios*, **10**: 578-596.

- Grotzinger, J.P., Bowring, S.A., Saylor, B.Z. and Kaufman, A.J. 1995. Biostratigraphic and geochronological constraints on early animal evolution, *Science*, **270**: 598-604.
- Halverson, G.P., 2006. A Neoproterozoic chronology. *In* Neoproterozoic geobiology and paleobiology. *Edited by* S. Xiao and A.J. Kaufman. Springer, Dordrecht, Netherlands, pp. 231-271.
- Halverson, G.P., Hoffman, P.F., Schrag, D.P., Maloof, A.C., Rice, A.H.N. 2005. Toward a Neoproterozoic composite carbon-isotope record, *GSA Bulletin*, **117**: 1181-1207.
- Halverson, G.P., Wade, B.P., Hurtgen, M.T., and Barovich, K.M. 2010. Neoproterozoic chemostratigraphy, *Precambrian Research*, **182**: 337-350.
- Handford, C.R. and Loucks, R.G. 1993. Carbonate depositional sequences and systems tracts: Responses of carbonate platforms to relative sea-level changes. *In* Carbonate Sequence Stratigraphy: Recent Developments and Applications. *Edited by* R.G. Loucks and J.F. Sarg. American Association of Petroleum Geologists Memoir 57, Tulsa, OK, pp. 3-41.
- Hardie, L.A. 2003. Secular variations in Precambrian seawater chemistry and timing of Precambrian aragonite seas and calcite seas, *Geology*, **31**: 785-788.
- Harland, W.B. 1964. Critical evidence for a great Infra-Cambrian glaciation, *Geologische Rundschau*, **54**: 45-61.
- Harland, W.B., and Rudwick, M.J.S. 1964. The great infra-Cambrian ice age, *Scientific American*, **211**: 28-36.
- Harper, B.B., Puga-Bernab  , A., Droxler, A.W., Webster, J.M., Gischler, E., Tiwari, M., Lado-Insua, T., Thomas, A.L., Morgan, S., Jovane, L., and R  hl, U. 2015. Mixed carbonate-siliciclastic sedimentation along the Great Barrier Reef upper slope: A challenge to the reciprocal sedimentation model.
- Hayes, J.M., Kaplan, I.R., and Wedeking, K.W. 1983. Precambrian organic geochemistry: Preservation of the record. *In* Earth's earliest biosphere: Its origin and evolution. *Edited by* J.W. Schopf. Princeton University Press, pp. 93-134.
- Hayes, J.M., Strauss, H., and Kaufman, A.J. 1999. The abundance of C in marine organic carbon and isotopic fractionation in the global biogeochemical cycle of carbon during the past 800 Ma, *Chemical Geology*, **161**: 103-125.
- Hedberg, H.D. 1970. Continental margins from viewpoint of the petroleum geologist, *American Association of Petroleum Geologists Bulletin*, **54**: 3-43.
- Heezen, B.C., and Ewing, M. 1952. Turbidity currents and submarine slumps, and the 1929 Grand Banks earthquake, *American Journal of Science*, **250**: 849-873.
- Hein, F.J., and McMechan, M.E. 1994. Proterozoic and Lower Cambrian strata of the western Canada sedimentary basin. *In* Geological Atlas of the Western Canada Sedimentary Basin (57-68). *Edited by* G.D. Mossop and I. Shetsen. Canadian Society of Petroleum Geologists.

- Hickson, T.A., and Lowe, D.R. 2002. Facies architecture of a submarine fan channel-levee complex: The Juniper Ridge Conglomerate, Coalinga, California, *Sedimentology*, **49**: 35-362.
- Higgins, J.A., and Schrag, D.P. 2003. Aftermath of a snowball Earth, *Geochemistry, Geophysics, Geosystems*, **4**: 1028-1047.
- Hirth, G., and Tullis, J. 1992. Dislocation creep regimes in quartz aggregates, *Journal of Structural Geology*, **14**: 145-159.
- Hoffman, P.F. 1991. Did the breakout of Laurentia turn Godwanaland inside-out?, *Science*, **255**: 1409-1412.
- Hoffman, P.F., and Halverson, G.P. 2011. Neoproterozoic glacial record in the Mackenzie Mountains, northern Canadian Cordillera. *In The Geological Record of Neoproterozoic of Glaciations. Edited by E. Arnaud, G.P. Halverson and Shields-Zhou, G.. Geological Society London Memoir*, pp. 397-411.
- Hoffman, P.F., and Schrag, D.P. 2002. The snowball Earth hypothesis: Testing the limits of global change, *Terra Nova*, **14**: 129-155.
- Hoffman, P.F., Kaufman, A.J., Halverson, G.P., and Schrag, D.P. 1998. A Neoproterozoic snowball Earth, *Science*, **281**: 1342-1346.
- Hoffman, P.F., Condon, D.J., Bowring, S.A., and Crowley, J.L. 2004. U-Pb zircon date from the Neoproterozoic Ghaub Formation, Namibia: Constraints on the Marinoan glaciation, *Geology*, **32**: 817-820.
- Hoffman, P.F., Halverson, G.P., Domack, E.W., Husson, J.M., Higgins, J.A., and Schrag, D.P. 2007. Are basal Ediacaran (635 Ma) post-glacial “cap dolostones” diachronous? *Earth and Planetary Science Letters*, **258**: 114-131.
- Hoffman, P.F., Lamothe, K.G., LoBianco, S.J.C., Hodgskiss, M.S.W., Bellefroid, E.J., Johnson, B.W., Hodgkin, E.B., and Halverson, G.P. 2017. Sedimentary depocenters on Snowball Earth: Case studies from the Sturtian Chuos Formation in northern Namibia, *Geosphere*, **13**: 811-837.
- Hubbard, S.M. 2017. The stratigraphic expression of slope channel processes, GAC-MAC Joint Annual Meeting, Kingston, Ontario, Canada.
- Hubbard, S.M., Romans, B.W., and Graham, S.A. 2008. Deep-water foreland basin deposits of the Cerro Toro Formation, Magallanes basin, Chile: Architectural elements of a sinuous basin axial channel belt, *Sedimentology*, **55**: 1333-1359.
- Hueneke, H., and Mulder, T. 2010. *Deep-Sea Sediments*, Elsevier, Amsterdam, The Netherlands.
- Husson, J.M., Maloof, A.C., Schoene, B. 2012. A syn-depositional age for Earth’s deepest $\delta^{13}\text{C}$ excursion required by isotope conglomerate tests, *Terra Nova*, **24**: 318-325.

- Huyer, G. 2016. Architecture and stratigraphy of a Neoproterozoic deep-marine slope channel-complex set, Windermere Supergroup, Castle Creek, BC, B.Sc. thesis, University of Ottawa, Ottawa, ON.
- Ingram, R.L. 1954. Terminology for the thickness of stratification and parting units in sedimentary rocks, *Geological Society of America Bulletin*, **65**: 130-165.
- Iverson, R.M. 1997. The physics of debris flows, *Reviews of Geophysics*, **35**: 245-296.
- Jacobsen, S.B., and Kaufman, A.J. 1999. The Sr, C and O isotopic evolution of Neoproterozoic seawater, *Chemical Geology*, **161**: 37-57.
- Jaffrés, J.B.D., Shields, G.A., and Wallman, K. 2007. The oxygen isotope evolution of seawater: a critical review of a long-standing controversy and an improved geological water cycle model for the past 3.4 billion years, *Earth-Science Reviews*, **83**: 83-122.
- James, N.P. and Jones, B. 2016. *Origin of carbonate sedimentary rocks*, John Wiley & Sons Ltd., West Sussex, U.K.
- James, N.P., Narbonne, G.M., and Kyser, T.K. 2001. Late Neoproterozoic cap carbonates: Mackenzie Mountains, northwestern Canada: precipitation and global glacial meltdown, *Canadian Journal of Earth Sciences*, **38**: 1229-1262.
- Jiang, G., Christie-Blick, N., Kaufman, A.J., Banerjee, D.M., and Rai, V. 2002. Sequence stratigraphy of the Neoproterozoic Infra Krol Formation and Krol Group, Lesser Himalaya, India, *Journal of Sedimentary Research*, **72**: 524-542.
- Jiang G., Kennedy, M.J., Christie-Blick, N., Wu, H., Zhang, S. 2006a. Stratigraphic, sedimentary structures, and textures of the late Neoproterozoic Doushantuo cap carbonate in South China, *Journal of Sedimentary Research*, **76**: 978-995.
- Jiang, G., Shi, X., Zhang, S. 2006b. Methane seeps, methane hydrate destabilization, and the late Neoproterozoic postglacial cap carbonates, *Chinese Science Bulletin*, **10**: 1152-1173.
- Jiang, G., Kaufman, A.J., Christie-Blick, N., Zhang, S., and Wu, H. 2007. Carbon isotope variability across the Ediacaran Yangtze platform in South China: Implications for a large surface-to-deep ocean $\delta^{13}\text{C}$ gradient, *Earth and Planetary Science Letters*, **261**: 303-320.
- Jo, A. 2013. Carbonate slope morphology and sedimentary processes along southwestern Great Bahama Bank, M.Sc. thesis, University of Miami, Miami, FL.
- Jo, A., Eberli, G.P., and Grasmueck, M. 2015. Margin collapse and slope failure along southwestern Great Bahama Bank, *Sedimentary Geology*, **317**: 43-52.
- Johnson, D.T., MacDonald, F.A., Gill, B.C., Hoffman, P.F., Schrag, D.P. 2012. Uncovering the Neoproterozoic carbon cycle, *Nature*, **483**: 320-324.

- Kane, I.A., Kneller, B.C., Dykstra, M., Kassem, A., and McCaffrey, W.D. 2007. Anatomy of a submarine channel-levee: An example from Upper Cretaceous slope sediments, Rosario Formation, Baja, California, Mexico, *Marine and Petroleum Geology*, **24**: 540-563.
- Kaufman, A.J., and Knoll, A.H. 1995. Neoproterozoic variations in the C-isotopic composition of seawater: stratigraphic and biogeochemical implications, *Precambrian Research*, **73**: 27-49.
- Kaufman, A.J., Hayes, J.M., Knoll, A.H. and Germs, G.J. 1991. Isotopic compositions of carbonates and organic carbon from upper Proterozoic successions in Namibia: Stratigraphic variation and the effects of diagenesis and metamorphism, *Precambrian Research*, **49**: 301-327.
- Kaufman, A.J., Jacobsen, S.B., and Knoll, A.H. 1993. The Vendian record of Sr- and C-isotopic variations in seawater: Implications for tectonics and paleoclimate. *Earth and Planetary Science Letters*, **120**: 409-430.
- Kaufman, A.J., Knoll, A.H., and Narbonne, G.M. 1997. Isotopes, ice ages, and terminal Proterozoic earth history, *Proceedings of the National Academy of Sciences of the United States of America*, **94**: 660-6605.
- Kaufman, A.J. Jiang, G., Christie-Blick, Banerjee, D. and Rai, V. Stable isotope record of the terminal Neoproterozoic Krol platform in the lesser Himalayas of northern India, *Precambrian Research*, **147**: 156-185.
- Kaufman, A.J. Jiang, G., Christie-Blick, Banerjee, D. and Rai, V. 2006. Stable isotope record of the terminal Neoproterozoic Krol platform in the lesser Himalayas of northern India, *Precambrian Research*, **147**: 156-185.
- Kelts, K., and Arthur, M.A. 1981. Turbidites after ten years of deep-sea drilling – Wringing out the mop?, *SEPM Special Publication*, **32**: 91-127.
- Kendall, B.S., Creaser, R.A., Ross, G.M., and Selby, D. 2004. Constraints on the timing of Marinoan "Snowball Earth" glaciation by ^{187}Re - ^{187}Os dating of a Neoproterozoic, post-glacial black shale in Western Canada, *Earth and Planetary Science Letters*, **222**: 729-740.
- Kennedy, M.J. 1996. Stratigraphy, sedimentology and isotope geochemistry of Australian Neoproterozoic postglacial cap dolostones: Deglaciation, $\delta^{13}\text{C}$ excursions, and carbonate precipitation, *Journal of Sedimentary Research*, **66**: 1050-1064.
- Kennedy, M.J., Christie-Blick, N., and Sohl, L.E. 2001. Are Proterozoic cap carbonates and isotopic excursions a record of gas hydrate destabilization following Earth's coldest intervals?, *Geology*, **29**: 443-446.
- Key, R.M., Liyungu, A.K., Njamu, F.M., Somwe, V., Banda, J., Mosley, P.N., and Armstrong, R.A. 2001. The western arm of the Lufilian Arc in NW Zambia and its potential for copper mineralization, *Journal of African Earth Sciences*, **33**: 503-528.
- Khan, Z. 2012. Origin and architecture of deep-water level deposits: Insight from the ancient rock record and experiments, Ph.D. thesis, University of Ottawa, Ottawa, ON.

- Khan, Z.A., and Arnott, R.W.C. 2011. Stratal attributes and evolution of asymmetric inner- and outer-bend levee deposits associated with an ancient deep-water channel-levee complex within the Isaac Formation, southern Canada, *Marine and Petroleum Geology*, **28**: 824-842.
- Khan, Z.A., Arnott, B., and Pugin, A. 2011. An alternative model of producing topography in the crest region of deep-water levees, *AAPG Bulletin*, **95**: 2085-2106.
- King, P.R., Browne, G.H., Slatt, R.M., Weimer, P., Bouma, A.H., and Perkins, B. 1994. Sequence architecture of exposed late Miocene basin floor fan and channel-levee complexes (Mount Messenger Formation), Taranki Basin, New Zealand. *In Submarine Fans and Turbidite Systems: Sequence Stratigraphy, Reservoir Architecture and Production Characteristics: Gulf Coast Section. Edited by P. Weimer, A.H. Bouma, and B.F. Perkins. SEPM Foundation 15th Annual Research Conference*, pp. 177-192.
- Kirschvink, J.L. 1992. Late Proterozoic low-latitude glaciation: the snowball Earth. *In The Proterozoic Biosphere. Edited by J.W. Schopf and C. Klein. Cambridge University Press, Cambridge*, pp. 51-52.
- Knauth, L.P., and Kennedy, M.J. 2009. The late Precambrian greening of the Earth, *Nature*, **460**: 728-732.
- Kneller, B., and Buckee, C. 2000. The structure and fluid mechanics of turbidity currents: a review of some recent studies and their geological implications, *Sedimentology*, **47**: 62-94.
- Kneller, B., Nasr-Azadani, M.M., Radhakrishnan, S., and Meiburg, E. 2016. Long-range sediment transport in the world's oceans by stably stratified turbidity currents, *Journal of Geophysical Research: Oceans*, **121**: 8608-8620.
- Knoll, A.H., and Walter, M.R. 1992. Latest Proterozoic stratigraphy and Earth history, *Nature*, **356**: 673-677.
- Knoll, A.H., Hayes, J.M., and Kaufman, A.J., Swett, K., Lambert, I.B. 1986. Secular variation in carbon isotope ratios from Upper Proterozoic successions of Svalbard and East Greenland, *Nature*, **321**: 832-838.
- Kolla, V., Posamentier, H.W., and Wood, L.J. 2007. Deep-water and fluvial sinuous channels – Characteristics, similarities and dissimilarities, and modes of formation, *Marine and Petroleum Geology*, **24**: 388-405.
- van Konijnenburg, J.H. 1999. Stratigraphic architecture of a Lower Cretaceous-Lower Tertiary carbonate base-of-slope succession: Gran Sasso d'Italia (central Appennines, Italy). *In Advances in carbonate sequence stratigraphy: Application to reservoirs, outcrops and models, SEPM Special Publication no. 63. Edited by P.M. Harris, A.H. Saller, and J.A. Simo. SEPM, Tulsa, OK*, pp. 291-315.
- van der Kooij, B., Immenhauser, A., Steuber, T., Bahamonde Rionda, J.R., Merino Tomé, O. 2010. Controlling factors of volumetrically important marine carbonate cementation in deep slope settings, *Sedimentology*, **57**: 1491-1525.

- Kroopnick, P.M. 1985. The distribution of (super 13) C of sigma CO (sub 2) in the world oceans, Deep-Sea Research Part A: Oceanographic Research Paper V, **32**: 57-84.
- Kubli, T.H. 1990. Geology of the Dogtooth Range, northern Purcell Mountains, British Columbia, Ph.D. thesis, University of Calgary, Calgary, AB.
- Kuenen, P. H. 1964. Deep sea sands and ancient turbidites, *Developments in Sedimentology*, 3: 3-33.
- Kuenen, P.H., and Migliorini, C.I. 1950. Turbidity Currents as a Cause of Graded Bedding, *The Journal of Geology*, 58: 91-127.
- Labourdette, R., and Bez, M.. 2010. Element migration in turbidite systems: Random or systematic depositional processes?, *American Association of Petroleum Geologists Bulletin*, **94**: 345-368.
- Laskar, J., Joutel, F., and Robutel, P. 1993. Stabilization of the Earth's obliquity by the Moon, *Nature*, **361**: 615-617.
- Laurin, J. 2005. Depositional history and sequence stratigraphic context of a Neoproterozoic mass transport deposit, Isaac Formation, east-central B.C., B.Sc. thesis, University of Ottawa, Ottawa, ON.
- Le Guerroué, E., Allen, P.A., and Cozzi, A. 2006a. Chemostratigraphic and sedimentologically framework of the largest negative carbon isotopic excursion in Earth history: The Neoproterozoic Shuram Formation (Nafun Group, Oman), *Precambrian Research*, **146**: 68-92.
- Le Guerroué, E., Allen, P.A., Cozzi, A., Etienne, J.L , and Fanning, M. 2006b. 50 Myr recovery from the largest $\delta^{13}\text{C}$ excursion in the Ediacaran ocean, *Terra Nova*, **18**: 147-153.
- Leather, J., Allen, P.A., Brasier, M.D., and Cozzi, A. 2002. Neoproterozoic snowball Earth under scrutiny: evidence from the Fiq glaciation of Oman, *Geology*, **30**: 891-894.
- Leclair, S.F., and Arnott, R.W.C. 2005. Parallel lamination formed by high-density turbidity currents, *Journal of Sedimentary Research*, **75**: 1-5.
- Lee, S.L. 2016. Microstructural and external characterization of a low-grade metaturbidite succession in the Windermere Supergroup, British Columbia, Canada, B.Sc. thesis, University of Ottawa, Ottawa, ON.
- de Leeuw, J., Eggenhuisen, J.T. and, Cartigny, M.J.B. 2016. Morphodynamics of submarine channel inception revealed by new experimental approach, *Nature Communications*, **7**: 1-7.
- Leynaud, D., Mienert, J., and Vanneste, M. 2009. Submarine mass movements on glaciated and non-glaciated European continental margins: A review of triggering mechanisms and preconditions to failure, *Marine and Petroleum Geology*, **26**: 618-632.

- Li, Z.X., Evans, D.A.D., and Halverson, G.P. 2013. Neoproterozoic glaciations in a revised global palaeogeography from the breakup of Rodinia to the assembly of Gondwanaland, *Sedimentary Geology*, **294**: 219-232.
- Link, P.K., Christie-Blick, N., Devlin, W.J., Elston, D.P., Horodyski, R.J., Levy, M., Miller, J.M.G., Pearson, R.C., Prave, A., Stewart, J.H., Winston, D., Wright, L.A., and Wrucke, C.T. 2003. Middle and late Proterozoic stratified rocks in the western U.S. Cordillera, Colorado Plateau, and Basin and Range Province. *In* Precambrian: Conterminous U.S., *The Geology of North America: Geological Society of America Decade of North American Geology Series*, v. c-3. *Edited by* J.C. Reed, Jr. M.E. Bickford, R.S. Houston, P.K. Link, D.W. Rankin, P.K. Sims, and W.R. Van Schmus, pp. 474-690.
- Locat, J., and Lee, H.J. Submarine landslides: Advances and challenges, *Canadian Geotechnical Journal*, **39**: 193-212.
- Lowe, D.R. 1982. Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents, *Journal of Sedimentary Petrology*, **52**: 279-297.
- Lowe, D.R. 1988. Suspended-load fallout rate as an independent variable in the analysis of current structures, *Sedimentology*, **35**: 765-776.
- Lund, K., Aleinikoff, J.N., Evans, K.V., and Fanning C.M. 2003. SHRIMP U-Pb geochronology of Neoproterozoic Windermere Supergroup, central Idaho: Implications for rifting of western Laurentia and synchronicity of Sturtian glacial deposits, *Geological Society of America Bulletin*, **115**: 349-372.
- MacDonald, F.A., Schmitz, M.D., Crowley, J.L., Roots, C.F., Jones, D.S., Maloof, A.C., Strauss, J.V., Cohen, P.A., Johnston, D.T., and Schrag, D.P. 2010. Calibrating the Cryogenian, *Science*, **327**: 1241-1243.
- MacDonald, F.A., Strauss, J.V., Sperling, E.A., Halverson, G.P., Narbonne, G.M., Johnston, D.T., Kunzmann, M., Schrag, D.P., and Higgins, J.A. 2013. The stratigraphic relationship between the Shuram carbon isotope excursion, the oxygenation of Neoproterozoic oceans, and the first appearance of the Ediacara biota and bilaterian trace fossils in northwestern Canada, *Chemical Geology*, **362**: 250-272.
- Machel, H.G. 1985. Cathodoluminescence in calcite and dolomite and its chemical interpretation, *Geoscience Canada*, **12**: 139-147.
- MacNaughton, R.B., Narbonne, G.M., and Dalrymple, R.W. 2000. Neoproterozoic slope deposits, Mackenzie Mountains, northwestern Canada: Implications for passive-margin development and Ediacaran faunal ecology, *Canadian Journal of Earth Sciences*, **37**: 997-1020.
- Magaritz, M., Holser, W.T., and Kirschvink, J.L. 1986. Carbon-isotope events across the Precambrian/Cambrian boundary on the Siberian Platform, *Nature*, **30**: 258-259.
- Major, R.P. 1991. Cathodoluminescence in post-Miocene carbonates. *In* *Luminescence Microscopy and Spectroscopy: Qualitative and Quantitative Applications: SEPM Short Course Vol. 25*. *Edited by* C. Barker, R. Burruss, O. Kopp, H. Machel, D. Marshall. Pp 149-154.

- Marion, C. 2005. Architecture and sedimentology of a channel and slide complex in the Neoproterozoic Isaac Formation, British Columbia, B.Sc. thesis, University of Ottawa, Ottawa, Ontario.
- Marshall, J.D. 1992. Climatic and oceanographic isotope signals from the carbonate rock record and their preservation, *Geological Magazine*, **129**:143-160.
- Marshall, H.G., Walker, J.C.G., and Kuhn, W.R. 1988. Long-term climate change and the geochemical cycle of carbon, *Journal of Geophysical Research*, **93**: 791-801.
- McDonough, R.M., and Parrish, R.R. 1991. Proterozoic gneisses of the Malton Complex, near Valemount, British Columbia; U-Pb ages and Nd isotopic signatures, *Canadian Journal of Earth Sciences*, **28**: 1202-1216.
- McFadden, K.A., Huang, J., Chu, X., Jiang, G., Kaufman, A.J., Zhou, ÜC., Yuan, X., and Xiao, S. 2008. Pulsed oxidation and biological evolution in the Ediacaran Doushantuo Formation, *Proceedings of the National Academy of Sciences of the United States of America*, **105**: 3197-3202.
- McHargue, T.M., Pyrcz, J., Sullivan, M.D., Clark, J.D., Fildani, A., Romans, B.W., Covaults, J.A., Levy, M., Posamentier, H.W., and Drinkwater, N.J. 2011. Architecture of turbidite channel systems on the continental slope: Patterns and predictions, *Marine and Petroleum Geology*, **28**: 728-743.
- McMechan, M.E. 1987. Stratigraphy and structure of the Mount Selwyn area, Rocky Mountains, northeastern British Columbia, *Geological Survey of Canada Paper* 85-28.
- McMechan, M.E. 2000. Vreeland diamictites; Neoproterozoic glaciogenic slope deposits, Rocky Mountains, Northeast British Columbia, *Bulletin of Canadian Petroleum Geology*, **48**: 246-261.
- McMechan, M.E. 2015. The Neoproterozoic succession of the central Rocky Mountains, Canada, *Bulletin of Canadian Petroleum Geology*, **63**: 243-273.
- McMechan, M.E., and Thompson, R.I. 1995. Geology, Ovington Creek, West of Sixth Meridian, British Columbia, *Geological Survey of Canada Map* 1873A, scale 1:50000, 1 sheet.
- McNaughton, R.B., Narbonne, G.M., Dalrymple, R.W. 2000. Neoproterozoic slope deposits, Mackenzie Mountains, northwestern Canada: implications for passive margin development and Ediacaran faunal ecology, *Canadian Journal of Earth Sciences*, **37**: 997-1020.
- Melim, L.A., Swart, P.K., Eberli, G.P. 2004. Mixing-zone diagenesis in the subsurface of Florida and the Bahamas, *Journal of Sedimentary Research*, **74**: 904-913.
- Meuhlenbachs, K. 1998. The oxygen isotope composition of the oceans, sediments, and seafloor, *Chemical Geology*, **145**: 263-273.

- Meyer, L. 2004. Internal architecture of an ancient deep-water, passive margin, basin-floor fan system, Upper Kaza Group, Windermere Supergroup, Castle Creek, British Columbia, M.Sc. thesis, University of Calgary, Calgary, AB.
- Meyer, L., and Ross, G.M. 2007. Channelized lobe and sheet sandstones of the upper Kaza Group basin-floor turbidite system (Windermere Supergroup), British Columbia, Canada. *In* Atlas of deep-water outcrops: AAPG Studies in Geology 56. *Edited by* T. H. Nilsen, R. D. Shew, G. S. Steffens, and J. R. J. Studlick, CD-ROM, 4 p.
- Middleton and Hampton 1973. Sediment gravity flows: mechanics of flow and deposition. *In* Turbidity and Deep Water Sedimentation. *Edited by* G.V. Middleton and A.H. Bouma, SEPM, Pacific Section, Short Course Lecture Notes, 1-38.
- Milczarek, G. 2018. Chemostratigraphic analysis of mudrocks and its implication on provenance and sequence stratigraphy in the Isaac Formation, Windermere turbidite system, Castle Creek, B.C., Canada, B.Sc. Thesis, University of Ottawa, Ottawa, ON.
- Mohrig, D., Ellis, C., Parker, G., Whipple, K.X., Hondzo, M. 1998. Hydroplaning of subaqueous debris flows, Geological Society of America Bulletin, **11**: 387-394.
- Monger, J.W.H., Price, R.A., Temelman-Kluit, D.J. 1982. Tectonic accretion and the origin of the two major metamorphic and plutonic belts in the Canadian Cordillera, Geology, **10**:70-75.
- Moore, E.M. 1991. Southwest U.S.-East Antarctic (SWEAT) connection: A hypothesis, Geology, **19**: 425-428.
- Morris, E.A., Hodgson, D.M., Brunt, R.L., and Flint, S.S. 2014. Origin, evolution and anatomy of silt-prone submarine external levees, Sedimentology, **61**: 1734-1763.
- Mountjoy, E.W. 1980. Mount Robson, Alberta-British Columbia; Geological Survey of Canada Map 1499A, scale 1:250000, 1 sheet.
- Mulder, T., and Alexander, J. 2001. The physical character of subaqueous density flows and their deposits, Sedimentology, **48**: 269-299.
- Mulder, T., Ducassou, E., Eberli, G.P., Gonthier, E., Kindler, P., Principaud, M., Fournier, F., Léonide, P., Billeaud, I., Marsset, B., Reijmer, J.J.G., Bondu, C., Joussiaume, R., and Pakiades, M. 2012a. New insights into the morphology and sedimentary processes along the western slope of Great Bahama Bank, Geology, **40**: 603-606.
- Mulder, T., Ducassou, E., Gillet, H., Hanquiez, V., Tournadour, E., Combes, J., Eberli, G.P., Kindler, P., Gonthier, E., Conesa, G., Robin, C., Sianipar, R., Reijmer, J.J.G., and François, A. 2012b. Canyon morphology on a modern carbonate slope of the Bahamas: Evidence of regional tectonic tilting, Geology, **40**: 771-774.
- Mulder, T., Ducassou, E., Gillet, H., Hanquiez, V., Principaud, M., Chabaud, L., Eberli, G.P., Kindler, P., Billeaud, I., Gonthier, E., Fournier, F., Léonide, P., Borgomano, J. 2014. First discovery of channel-levee complexes in a modern deep-water carbonate slope environment, Journal of Sedimentary Research, **84**: 1139-1146.

- Mullins, H.T., Heath, K.C., Van Buren, H.M., and Newton, C.R. 1984. Anatomy of a modern open-ocean carbonate slope: Northern Little Bahama Bank, *Sedimentology*, **31**: 252-168.
- Murphy, D.C. 1987a. Kaza Group, eastern wells gray park, British Columbia, Geological Survey of Canada Paper 87-1A, pp. 735-742.
- Murphy, D.C., 1987b. Suprastructure/infrastructure transition, east-central Cariboo Mountains, British Columbia; geometry, kinematics and tectonic implications, *Journal of Structural Geology*, **9**: 13-29.
- Murphy, D.C. 1990. Stratigraphy and structure, southern Rocky Mountain Trench to the headwaters of the North Thompson River, Cariboo Mountains, British Columbia *In* Current Research Part E, Cordillera and Pacific Margin, Geological Survey of Canada Paper 90-1E, 71-79.
- Murphy, D.C., and Rees, C.J. 1983. Structural transition and stratigraphy in the Cariboo Mountains, British Columbia. *In* Current Research Part A, Geological Survey of Canada Paper 83-1A, 245-252.
- Murphy, D.C., van der Heyden, P., Parrish, R.R., Klepacki, D.W., Struik, L.C., and Gabites, J. 1995. New geochronological constraints on Jurassic deformation of the western edge of North America, southeastern Canadian Cordillera, Geological Society of America Special Papers, **299**: 159-172.
- Murru, M., Simone, L., Vigorito, M. 2001. Carbonate channel network in the Miocene syn-rift Sardinia basins, *Géologie Méditerranéenne*, **28**: 133-137.
- Mussa-Caleca, M. 2008. Architecture and depositional history of a Neoproterozoic deep-water slope channel complex in a passive margin setting: Isaac Formation, Windermere Supergroup, southern Canadian Cordillera, M.Sc. thesis, University of Ottawa, Ottawa, ON.
- Narbonne, G.M. and Aitken, J.D. 1995. Neoproterozoic of the Mackenzie Mountains, northwestern Canada, *Precambrian Research*, **73**: 101-121.
- Narbonne, G.M., Kaufman, A.J., and Knoll, A.H. 1994. Integrated chemostratigraphy and biostratigraphy of the Windermere Supergroup, northwestern Canada: Implications for Neoproterozoic correlations and the early evolution of animals, *Geological Society of America Bulletin*, **106**: 1281-1292.
- Narbonne, G.M., Xiao, S., and Shields-Zhou, G. 2012. Ediacaran period. *In* Geological Timescale 2012. *Edited by* F. Gradstein, J. Ogg, M.D. Schmitz and G. Ogg. Elsevier, Oxford, pp. 427-449.
- Navarro, L. 2005. Depositional architecture and evolution of deep-water base-of-slope and slope channel complexes in a passive margin setting: Isaac Formation, Windermere Supergroup (Neoproterozoic) southern Canadian Cordillera, M.Sc. thesis, University of Ottawa, Ottawa, ON.

- Navarro, L. 2016. Stratigraphic architecture depositional processes and reservoir implications of the basin floor to slope transition, Neoproterozoic Windermere turbidite system, Canada, Ph.D. thesis, University of Ottawa, Ottawa, ON.
- Navarro, L. and Arnott, B. 2016. Stratigraphic evolution of an ancient channel-lobe transition zone (CLTZ) in the Windermere turbidite system and implications for reservoir development. AAPG Annual Conference and Exhibition, Houston, Texas, U.S.A.
- Navarro, L., Khan, Z., and Arnott, R.W.C. 2007a. Architecture of a deep-water channel-levee complex: Channel 3, Castle Creek South, Isaac Formation, Windermere Supergroup, British Columbia, Canada. *In Atlas of deep-water outcrops: AAPG Studies in Geology 56. Edited by T. H. Nilsen, R. D. Shew, G. S. Steffens, and J. R. J. Studlick*, CD-ROM, 4 p.
- Navarro, L., Khan, Z., and Arnott, R.W.C. 2007b. Depositional architecture and evolution of a deep-marine channel-levee complex: Isaac Formation (Windermere Supergroup), Southern Canadian Cordillera., *In Atlas of deep-water outcrops: AAPG Studies in Geology 56. Edited by T. H. Nilsen, R. D. Shew, G. S. Steffens, and J. R. J. Studlick*, CD-ROM, 22 p.
- Nédélec, A., Affaton, P., France-Lanord, C., Charrière, A., and Alvaro, J, 2007. Sedimentology and chemostratigraphy of the Bwipe Neoproterozoic cap dolostones (Ghana, Volta Basin): A record of microbial activity in a peritidal environment, *Comptes Rendus Geoscience*, **339**: 223-239.
- Nilsen, T.H., Shew, R.D., Steffens, G.S., Studlick, J.R.J. 2007. *Atlas of Deep-Water Outcrops*, American Association of Petroleum Geologists, Tulsa, OK.
- Nogueira, A.C.R., Riccomini, C., Sial, A.N., Moura, C.A.V., Trindade, R.I.F., Fairchild, T.R. 2007. Carbon and strontium isotope fluctuations and paleoceanographic changes in the late Neoproterozoic Araras carbonate platform, southern Amazon craton, Brazil, *Chemical Geology*, **237**: 168-190.
- Normark, W.R. 1978. Fan valleys, channels, and depositional lobes on modern submarine fans: Characters for recognition of sandy turbidite environments, *American Association of Petroleum Geologists Bulletin*, **62**: 912-931.
- Normark, W.R., Barnes, N.E., and Coumes, F. 1985. Rhone Fan, Mediterranean. *In Submarine Fans and Related Turbidite Systems. Edited by A.H. Bouma, W.R. Normark, and N.E. Barnes*. Springer, New York, U.S.A, pp. 151-156.
- O'Byrne, C.J.M., Barton, M.D., Steffens, G.S., Primez, C., and Buergisser, H. 2007. Architecture of a laterally migrating channel complex: Channel 4, Isaac Formation, Windermere Supergroup, Castle Creek north, British Columbia, Canada. *In Atlas of deep-water outcrops: AAPG Studies in Geology 56. Edited by T. H. Nilsen, R. D. Shew, G. S. Steffens, and J. R. J. Studlick*, CD-ROM, 4 p.
- Oehlert, A.M., and Swart, P.K. 2014. Interpreting carbonate and organic isotope covariance in the sedimentary record, *Nature Communications*, **5**: 1-7.

- Olafiranye, K., Jackson, C.A.L., and Hodgson, D.M. 2013. The role of tectonics and mass-transport complex emplacement on upper slope stratigraphic evolution: A 3D seismic case study from offshore Angola, *Marine and Petroleum Geology*, **44**: 196-216.
- Page, M.C., Dickens, G.R., and Dunbar, G.R. 2003. Tropical view of Quaternary sequence stratigraphy: Siliciclastic accumulation on slopes east of the Great Barrier Reef since the Last Glacial Maximum, *Geology*, **31**: 1013-1016.
- Parker, G., Fukushima, Y., and Pantin, H.M. 1986. Self-accelerating turbidity currents, *Journal of Fluid Mechanics*, **171**: 145-181.
- Parrish, R.R., and Scammell, R.J. 1988. The age of the Mount Copeland syenite gneiss and its metamorphic zircons, Monashee Complex, southeastern British Columbia. *In* Radiogenic age and isotopic studies: Report 2, Geological Survey of Canada Paper 88-2, 21-28.
- Payros, A., and Pujalte, V. 2008. Calciclastic submarine fans: An integrated overview, *Earth Science Reviews*, **86**: 203-246.
- Payros, A., Pujalte, V., and Orue-Etxebarria, X. 2007. A point-sourced calciclastic submarine fan complex (Eocene Anotz Formation, western Pyrenees): Facies architecture, evolution and controlling factors, *Sedimentology*, **54**: 137-168.
- Pell, S.D., McKirdy, D.M., Jansyn, J. and Jenkins, R.J.F. 1993. Ediacaran carbon isotope stratigraphy of South Australia – An initial study, *Transactions of the Royal Society of South Australia*, **117**: 153-161.
- Peerdeman, F.M., and Davies, P.J. 1993. Sedimentological response of an outer-shelf, upper-slope sequence to rapid changes in Pleistocene eustatic sea level: Hole 820A, northeastern Australian margin, *Proceedings of the Ocean Drilling Program, Scientific Results*, **133**: 303-313.
- Peerdeman, F.M., Davies, P.J., and Chivas, A.R. 1993. The stable oxygen isotope signal in shallow-water, upper slope sediments off the Great Barrier Reef (Hole 820A), *Proceedings of the Ocean Drilling Program, Scientific Results*, **133**: 163-173.
- Pickering, K.T., Hiscott, R., and Hein, F.J. 1989. *Deep-marine Environments: Clastic Sedimentation and Tectonics*. Unwin Hyman, London.
- Pigage, L.C. 1977. Rb–Sr dates for granodiorite intrusions on the northeast margin of the Shuswap Metamorphic Complex, Cariboo Mountains, *Canadian Journal of Earth Sciences*, **14**: 1690-1695.
- Piper, D.J.W., and Normark, W.R. 2001. Sandy fans – from Amazon to Hueneme and beyond, *American Association of Petroleum Geologists Bulletin*, **85**: 1407-1438.
- Piper, D.J.W., and Normark, W.R. 2009. Processes that initiate turbidity currents and their influence on turbidites: A marine geology perspective, *Journal of Sedimentary Research*, **79**: 347-362.

- Piper, D.J.W., Cochonat, P., and Morrison, M.L. 1999. The sequence of events around the epicenter of the 1929 Grand Banks earthquake: Initiation of debris flows and turbidity current inferred from sidescan sonar, *Sedimentology*, **46**: 79-97.
- Popović, N. 2016. Sedimentology and stratigraphy of a matrix-poor to matrix-rich depositional continuum in proximal basin floor strata, upper Kaza Group, Windermere Supergroup, B.C., Canada, M.Sc. thesis, University of Ottawa, Ottawa, ON.
- Posamentier, H.W. 2003. Depositional elements associated with a basin floor channel-levee system: case study from the Gulf of Mexico, *Marine and Petroleum Geology*, **20**: 677-690.
- Posamentier, H.W., and Kolla, V. 2003. Seismic geomorphology and stratigraphy of depositional elements in deep-water settings, *Journal of Sedimentary Research*, **79**: 367-388.
- Posamentier, H.W. and Vail, P.R. 1988. Eustatic control on clastic deposition II – sequence and system tracts models. *In* Sea-level Changes, an Integrated Approach. *Edited by* C.G. Wilgus, B.S. Hastings, C.A. Ross, H.W. Posamentier, J.C. Van Wagoner and C.G. Kendall. SEPM Special Publication 42, Tulsa, OK, pp. 125-154.
- Posamentier, H.W., and Walker, R.G. 2006. Deep-water turbidites and submarine fans, SEPM Special Publication, **84**: 397-520.
- Prather, B.E. 2000. Calibration and visualization of depositional process models for above-grade slopes: A case study from the Gulf of Mexico, *Marine and Petroleum Geology*, **17**: 619-638.
- Principaud, M., Mulder, T., Borgomano, J., Ducassou, E., Hanquiez, V., Gillet, H., Marieu, V., and Sorriaux, P. 2013. Carbonate slope gully system on the Westside Great Bahama Bank, EGU General Assembly, Vienna, Austria.
- Principaud, M., Mulder, T., Gillet, H., and Borgomano, J. 2015. Large-scale carbonate submarine mass-wasting along the northwestern slope of the Great Bahama Bank (Bahamas): Morphology, architecture and mechanisms, *Sedimentary Geology*, **317**: 27-42.
- Puga-Bernabéu, A., Webster, J.M., Beaman, R.J., and Guilbaud, V. 2011. Morphology and controls on the evolution of a mixed carbonate-siliciclastic submarine canyon system, Great Barrier Reef margin, north-eastern Australia, *Marine Geology*, **289**: 100-116.
- Puga-Bernabéu, A., Webster, J.M., Beaman, R.J., and Guilbaud, V. 2013. Variation in canyon morphology on the Great Barrier Reef margin, north-eastern Australia: The influence of slope and barrier reefs, *Geomorphology*, **191**: 35-50.
- Puga-Bernabéu, A., Webster, J.M., Beaman, R.J., Reimer, P.J., Renema, W. 2014. Filling the gap: A 60 ky record of mixed carbonate-siliciclastic turbidite deposition from the Great Barrier Reef., *Marine and Petroleum Geology*, **50**: 40-50.
- Pyle, L.J., Narbonne, G.M., James, N.P., Dalrymple, R.W., and Kaufman, A.J. 2004. Integrated Ediacaran chronostratigraphy, Wernecke Mountains, northwestern Canada, *Precambrian Research*, **132**: 1-27.

- Rankey, E.C., and Doolittle, D.F. 2012. Geomorphology of carbonate platform-marginal uppermost slopes: Insights from a Holocene analogue, Little Bahama Bank, Bahamas, *Sedimentology*, **59**: 2146-2171.
- Reesor, J.E. 1973. Geology of the Lardeau Map-area, east-half, British Columbia, Geological Survey of Canada Memoir 369, 129 p.
- Reid, L.F., Cook, F.A., and Erdmer, P. 1997. Structural correlation across the suprastructure-infrastructure transition (Cariboo Mountains, British Columbia); possible, impossible or work in progress?, lithoprobe slave Northern cordillera lithospheric evolution (SNORCLE) and cordilleran tectonics workshop; report of the 1997 combined meeting. *Lithoprobe Report*, 56, 181-182.
- Reid, L.F., Simony, P.S., and Ross, G.M. 2002. Dextral strike-slip faulting in the Cariboo Mountains, British Columbia: a natural example of wrench tectonics in relation to Cordilleran tectonics, *Canadian Journal of Earth Sciences*, **39**: 953-970.
- Reijmer, J.J.G., Palmieri, P., Groen, and Floquet, M. 2015. Calciturbidites and calcidebrites: Sea-level variations or tectonic processes, *Sedimentary Geology*, 317: 53-70.
- Rice, A.H.N., Edwards, M.B., Hansen, T.A., Arnaud, E. and Halverson, G.P. 2011. Glaciogenic rocks of the Neoproterozoic Smallfjord and Mortensnes formations, Vestertana Group, E, Finnmark, Norway. *In The Geological Record of Neoproterozoic Glaciations. Edited by E. Arnaud, G.P. Halverson and G. Shields-Zhou*, pp. 593-602.
- Rocheleau, J. 2011. Depositional architecture of a near-slope turbidite succession: Upper Kaza Group, Windermere Supergroup, Castle Creek, British Columbia, Canada, M.Sc. thesis, University of Ottawa, Ottawa, ON.
- Rooney, A.D., MacDonald, F.A., Strauss, J.V., Dudás, F, Ö, Hallmann, C., and Selby, D. 2014. Re-Os geochronology and coupled Os-Sr isotope constraints on the Sturtian snowball Earth, *Proceedings of the National Academy of Sciences of the United States of America*, 111: 51-56.
- Rooney, A.D., Strauss, J.V., Brandon, A.D., MacDonald, F.A. 2015. A Cryogenian chronology: Two long-lasting synchronous Neoproterozoic glaciations, *Geology*, **43**: 459-462.
- Ross, G.M. 1991. Tectonic setting of the Windermere Supergroup revisited, *Geology*, **19**: 1125-1128.
- Ross, G.M., and Arnott, R.W.C. 2007. Regional Geology of the Windermere Supergroup, Southern Canadian Cordillera and Stratigraphic Setting of the Castle Creek Study Area, Canada. *In Atlas of deep-water outcrops: AAPG Studies in Geology 56. Edited by T. H. Nilsen, R. D. Shew, G. S. Steffens, and J. R. J. Studlick*, CD-ROM, 16 p.
- Ross, G.M., and Bowring, S.A. 1990. Detrital zircon geochronology of the Windermere Supergroup and the tectonic assembly of the southern Canadian Cordillera, *The Journal of Geology*, **98**: 879-893.

- Ross, G.M., and Ferguson, C.A. 2003. Geology and structure cross-sections, Eddy, British Columbia; Geological Survey of Canada map 1967A scale 1:50000, 1 sheet.
- Ross, G.M., and Murphy, D.C. 1988. Transgressive stratigraphy, anoxia, and regional correlations within the late Precambrian Windermere grit of the southern Canadian Cordillera, *Geology*, **16**: 139-143.
- Ross, G.M., and Parrish, R.R., 1991a. Detrital zircon geochronology of metasedimentary rocks in the southern Omineca Belt, Canadian Cordillera, *Canadian Journal of Earth Sciences*, **28**: 1254-1270.
- Ross, G.M., and Parrish, R.R. 1991b. Single grain U-Pb and Sm-Nd provenance of the Middle Proterozoic Belt Supergroup (northwestern United States): implications for pre- Panthalassa plate reconstructions. *Geological Society of America Abstracts with Programs*, 23, A210.
- Ross, G., McMechan, M.E., and Hein, F.J. 1989. Chapter 6: Proterozoic history: The birth of the Miogeocline. *In Western Canada Sedimentary Basin: A Case History. Edited by B.D. Ricketts*. Canadian Society of Petroleum Geologists, Calgary, Canada, pp. 79-105
- Ross, W.C., Halliwell, B.A., May, J.A., Watts, D.E., and Syvitski, J.P.M. 1994. Slope readjustment: A new model for the development of submarine fans and aprons, *Geology*, **22**: 511-514
- Ross, G.M., Bloch, J.D., and Krouse, H.R. 1995. Neoproterozoic strata of the southern Canadian Cordillera and the isotopic evolution of seawater sulfate, *Precambrian Research*, **73**: 71-99.
- Rothman, D.H., Hayes, J.M., and Summons, R.E. 2003. Dynamics of the Neoproterozoic carbon cycle, *Proceedings of the National Academy of Sciences of the United States of America*, **100**: 8124-8129.
- Rothwell, R.G., Thomson, J., and Kähler, G. 1998. Low-sea-level emplacement of a very large Late Pleistocene 'megaturbidite' in the western Mediterranean Sea, *Nature*, **392**: 377-380.
- Rowe, C.E. 2003. The Cunningham formation: a ramp to platform carbonate in the terminal Proterozoic of the Cariboo Mountains, British Columbia, M.Sc. thesis, University of Calgary, Calgary, AB.
- Russell, K.L. 1968. Oceanic ridges and eustatic changes in sea level, *Nature*, **218**: 861-862.
- Samuel, A., Kneller, B., Raslan, S., Sharp, A., Parsons, C. 2003. Prolific deep-marine slope channels of the Nile Delta, Egypt, *American Association of Petroleum Geologists Bulletin*, **87**: 541-560.
- Sanchez, C., Covault, J.A., Janson, X., Dunlap, D. 2017. Mixed carbonate-siliciclastic systems : Dynamics and controls, Eocene-Oligocene Browse basin. AAPG Annual Conference and Exhibition, Houston, Texas, U.S.A.
- Sarg, J.F. 1988. Carbonate sequence stratigraphy. *In Sea-level Changes, an Integrated Approach. Edited by C.G. Wilgus, B.S. Hastings, C.A. Ross, H.W. Posamentier, J.C. Van Wagoner and C.G.St.C. Kendall*. SEPM Special Publication 42, Tulsa, OK, pp. 155-181.

- Saylor, B.Z., Kaufman, A.J., Grotzinger, J.P., and Urban, F. 1998. A composite reference section for terminal Proterozoic strata of southern India, *Journal of Sedimentary Research*, **68**: 1223-1235.
- Schlager, W., and Chermak, A. 1979. Modern sediment facies of platform-basin transition, Tongue of the Ocean, Bahamas. *In* *Geology of Continental Slopes*, SEPM Special Publication no. 27. *Edited by* O.H. Pilkey, and L.J. Doyle. SEPM, Tulsa, OK.
- Schlager, W., Reijmer, J.J., and Droxler, A. 1994. Highstand Shedding of Carbonate Platforms, *Journal of Sedimentary Research*, **64B**: 270-281.
- Scholle, P.A. and Ulmer-Scholle, D.S. 2003. A color guide to the petrography of carbonate rocks: Grains, textures, porosity, diagenesis: AAPG Memoir 77. The American Association of Petroleum Geologists, Tulsa, OK, U.S.A.
- Schrag, D.P., Higgins, J.A., MacDonald, F.A., and Johnston, D.T. 2013. Authigenic carbonate and the history of the global carbon cycle, *Science*, **339**: 540-543.
- Schwarz, E., and Arnott, R.W.C. 2007a. Anatomy and evolution of a slope channel-complex set (Neoproterozoic Isaac Formation, Windermere Supergroup, southern Canadian Cordillera): implications for reservoir characterization, *Journal of Sedimentary Research*, **77**: 89-109.
- Schwarz, E., and Arnott, R.W.C. 2007b. Outcrop characterization of a passive- margin channel-complex set: Isaac Channel 5, Neoproterozoic Isaac Formation, British Columbia, Canada. . *In* *Atlas of deep-water outcrops: AAPG Studies in Geology 56*. *Edited by* T. H. Nilsen, R. D. Shew, G. S. Steffens, and J. R. J. Studlick, CD-ROM, 15 p.
- Schwarz, E., Veiga, G.D., Alvarez Trentini, G., Isla, M.F., Spalletti, L.A. 2018. Expanding the spectrum of shallow-marine, mixed carbonate-siliciclastic systems: Processes, facies distribution and depositional controls of a siliciclastic-dominated example, *Sedimentology*, doi: 10.1111/sed12438.
- Shanmugam, G. 1996. High-density turbidity currents: Are they sandy debris flows?, *Journal of Sedimentary Research*, **66**: 2-10.
- Shanmugam, G. 1997. Deepwater exploration; Conceptual models and their uncertainties, *The Bulletin of the Houston Geological Society*, **39**: 13.
- Shanmugam, G. 2006. Deep-water processes and facies models: Implications for sandstone petroleum reservoirs. Elsevier, New York, U.S.A.
- Shen, B., Xiao, S., Huiming, B., Kaufman, A.J., Zhou, C., Yuan, X. 2011. Carbon, sulfur and oxygen isotope evidence for a strong depth gradient and oceanic oxidation after the Ediacaran Hankalchough glaciation, *Geochimica et Cosmochimica Acta*, **75**: 1357-1373.
- Shen, Y., Zhang, T., and Hoffman, P.F. 2008. On the coevolution of Ediacaran oceans and animals, *Proceedings of the National Academy of Sciences of the United States of America*, **105**: 7376-7381.

- Shields, G.A. 2005. Neoproterozoic cap carbonates: A critical appraisal of existing models and the *plumeworld* hypothesis, *Terra Nova*, **17**: 299-310.
- Shields, G., Stille, P., Brasier, M.D., Atudorei, N.V. 1997. Stratified oceans and oxygenation of the late Precambrian environments: A post glacial geochemical record from the Neoproterozoic of W. Mongolia, *Terra Nova*, **9**: 218-222.
- Smith, M.D. 2009. Stratigraphic and geochemical evolution of the Old Fort Point Fm., southern Canadian Cordillera: The deep-marine perspective of Ediacaran post-glacial environmental change, Ph.D. thesis, University of Ottawa, Ottawa, ON.
- Smith, M.D., Arnaud, E., and Arnott, R.W.C. 2011. The record of Neoproterozoic glaciations in the Windermere Supergroup, southern Canadian Cordillera, Geological Society, London, *Memoirs*, **36**: 413-424.
- Smith, M.D., Arnott, R.W.C., and Ross, G.M. 2014a. The Old Fort Point Formation: redefinition and formal subdivision of a distinctive stratigraphic marker in the Neoproterozoic Windermere Supergroup, southern Canadian Cordillera, *Bulletin of Canadian Petroleum Geology*, **62**: 1-13.
- Smith, M.D., Arnott, R.W.C., and Ross, G.M. 2014b. Physical and geochemical controls on sedimentation along an ancient continental margin: The deep-marine Old Fort Point Formation (Ediacaran), southern Canadian Cordillera, *Bulletin of Canadian Petroleum Geology*, **62**: 14-36.
- Southard, J.B., and Boguchwal, L.A. 1990. Bed configurations in steady unidirectional water flows. Part 2. Synthesis of flume data, *Journal of Sedimentary Research*, **60**: 658-679.
- Southard, J.B., and Dingler, J.R. 1971. Flume study of ripple propagation behind mounds on flat sand beds, *Sedimentology*, **16**: 251-263.
- Spence, G.H., and Tucker, M.E. 1997. Genesis of limestone megabreccias and their significance in carbonate sequence stratigraphic models: A review, *Sedimentary Geology*, **112**: 163-193.
- Stewart, J.H. 1972. Initial deposits in the cordilleran geosyncline; evidence of a late Precambrian (<850 m.y.) continental separation, *Geological Society of America Bulletin*, **83**: 1345-1360.
- Stow, D.A.V. 1979. Distinguishing between fine-grained turbidites and contourites on the Nova Scotian deep water margin, *Sedimentology*, **26**: 371-387.
- Stow, D.A.V. and Mayall, M. 2000. Deep-water sedimentary systems: New models for the 21st century, *Marine and Petroleum Geology*, **17**: 125-135.
- Stow, D.A.V., and Shanmugam, G. 1980. Sequence of structures in fine-grained turbidites: Comparison of recent deep-sea and ancient flysch sediments, *Sedimentary Geology*, **25**: 23-42.
- Sumner, E.J., Amy, L.A. and Tailing, P.J. 2008. Deposit structure and processes of sand depositing from decelerating sediment suspensions, *Journal of Sedimentary Research*, **78**: 529-547.

- Swart, P.K. 2008. Global synchronous changes in the carbon isotopic composition of carbonate sediments unrelated to changes in the global carbon cycle, *Proceedings of the National Academy of Sciences in the United States of America*, **105**: 13741-13745.
- Swart, P.K., and Kennedy, M.J. 2012. Does the global stratigraphic reproducibility of $\delta^{13}\text{C}$ in Neoproterozoic carbonates require a marine origin? A Pliocene-Pleistocene comparison, *Geology*, **40**: 87-90.
- Sutherland Brown, A. 1963. Geology of the Cariboo River Area, British Columbia; British Columbia Department of Mines and Petroleum Resources Bulletin 47, 68 p.
- Tahata, M., Ueno, Y., Ishikawa, T., Sawaki, Y., Murakami, K., Han, J., Shu, D., Li, Y., Yoshida, N. and Komiya, T. 2013. Carbon and oxygen isotope chemostratigraphies of the Yangtze Platform, South China: Decoding temperature and environmental changes through the Ediacaran, *Geochimica et Cosmochimica Acta*, **23**: 333-353.
- Talling, P.J., Wynn, R.B., Masson, D.G., Frenz, M., Cronin, B.T., Schiebel, R., Akhmetzhanov, A.M., Dallmeier-Tiessen, S., Benetti, S., Weaver, P.P.E., Georgiopoulou, A., Zühlendorf, C. and Amy, L.A. 2007. Onset of submarine debris flow deposition far from original giant landslide, *Nature*, **450**: 541-544.
- Terlaky, V. 2014. Sedimentology, stratigraphy, architecture and origin of deep-water, basin-floor deposits: Middle and upper Kaza Group, Windermere Supergroup, B.C., Canada, Ph.D. thesis, University of Ottawa, Ottawa, ON.
- Terlaky, V., and Arnott, R.W.C. 2014. Matrix-rich and associated matrix-poor sandstones: Avulsion splays in slope and basin floor strata, *Sedimentology*, **61**: 1175-1197.
- Terlaky, V., and Arnott, R.W.C. 2016. The control of terminal-splay sedimentation on depositional patterns and stratigraphic evolution in avulsion-dominated, unconfined, deep-marine basin-floor systems, *Journal of Sedimentary Research*, **86**: 786-799.
- Terlaky, V., Rocheleau, J., and Arnott, R.W.C. 2015. Stratal composition and stratigraphic organization of stratal elements in an ancient deep-marine basin-floor succession, Neoproterozoic Windermere Supergroup, British Columbia, Canada, *Sedimentology*, **63**: 136-175.
- Terlaky, V., Bergen, A., and Arnott, R.W.C. 2016. Characteristics of falling stage systems tract slope channels: An example from the Isaac Formation, Windermere Supergroup, Canada. AAPG Annual Conference and Exhibition, Calgary, Alberta, Canada.
- Tilston, M. 2016. The role of particle size and concentration in defining the flow structure of turbidity currents and the morphology of their deposits: Insights from computed tomography, Ph.D. thesis, University of Ottawa, Ottawa, ON.
- Tilston, M., Arnott, R.W.C., Rennie, C.D., and Long, B. 2015. The influence of grain size on the velocity and sediment concentration profiles and depositional record of turbidity currents, *Geology*, **43**: 839-842.

- Tucker, M.E. 1983. Diagenesis, geochemistry, and origin of a Precambrian dolomite: The Beck Spring Dolomite, eastern California, *Journal of Sedimentary Petrology*, **53**: 1097-1119.
- Tucker, M.E. 1986. Carbon isotope excursions in Precambrian/Cambrian boundary beds, Morocco, *Nature*, **319**: 48-50.
- Valentine, J.W, and Moores, E.M. 1970. Plate-tectonic regulation of faunal diversity and sea-level: A model, *Nature*, **228**: 657-659.
- Valley, J.W. 2001. Stable isotope thermometry at high temperatures, *Reviews in Mineralogy and Geochemistry*, **43**: 365-413.
- Veizer, J. 1983. Chemical diagenesis of carbonates: Theory and application. *In* Stable isotopes in Sedimentary Geology: SEPM Short Course No. 10. Society for Sedimentary Geology, Tulsa, OK, USA, pp. 3-1-3-100.
- Veizer, J., and Hoefs, J. 1976. The nature of O^{18}/O^{16} and C^{13}/C^{12} secular trends in sedimentary carbonate rocks, *Geochimica et Cosmochimica Acta*, **40**: 1387-1395.
- Veizer, J., Holser, W.T., and Wilgus, C.K. 1980. Correlation of $^{13}C^{12}C$ and $^{24}S^{32}S$ secular variations, *Geochimica et Cosmochimica Acta*, **44**: 579-897.
- Veizer, J., Ala, D., Azmy, K., Bruckschen, P., Buhl, D., Bruhn, F., Garden, G.A.F., Diener, A., Ebner, S., Godderis, Y., Jasper, T., Korte, C., Pawallek, F., Podlaha, O.G., Strauss, H. 1999. $^{87}Sr/^{86}Sr$, $\delta^{13}C$ and $\delta^{18}O$ evolution of Phanerozoic seawater, *Chemical Geology*, **161**: 59-88.
- Venditti, J.G., Church, M., and Bennett, S.J. 2005. Bed form initiation from a flat sand bed, *Journal of Geophysical Research*, **110**: F01009, 1-19.
- Venditti, J.G., Church, M., and Bennett, S.J. 2006. On interfacial instability as a cause of transverse subcritical bed forms, *Water Resources Research*, **42**: W07423.
- Verdel, C., Wernicke, B.P., and Bowring, S.A. 2011. The Shuram and subsequent Ediacaran carbon isotope excursions from southwest Laurentia, and implications for environmental stability during the metazoan radiation, *Geological Society of America Bulletin*, **123**: 1539-1559.
- Vigorito, M., Murru, M., and Simone, L. 2005. Anatomy of a submarine channel system and related fan in a foramol/rhodalg carbonate sedimentary setting: A case history from the Miocene syn-rift Sardinia Basin, Italy, *Sedimentary Geology*, **174**: 1-30.
- Vigorito, M., Murru, M., and Simone, L. 2006. Architectural patterns in a multistory mixed carbonate-siliciclastic submarine channel, Porto Torres Basin, Miocene, Sardinia, Italy, *Sedimentary Geology*, **186**: 213-236.
- Walker, J.F. 1926. Geology and mineral deposits of Windermere map-area, British Columbia, *Geological Society of Canada Memoir* 148, 78 p.

- Walker, R.F. 1965. The origin and significance of the internal sedimentary structures of turbidites, *Proceedings of the Yorkshire Geological Society*, 35: 1-32.
- Walker, R.G. 1985. Mudstones and thin-bedded turbidites with the Upper Cretaceous Wheeler Gorge conglomerates, California: A possible channel-levee complex, *Journal of Sedimentary Petrology*, 55: 279-290.
- Wallace, K. 2004. The architecture and sedimentology of two submarine debris flow deposits in the Neoproterozoic Isaac Formation, east-central British Columbia. B.Sc. thesis, University of Ottawa, Ottawa, ON.
- Wang, X., Jiang, G., Shi, X., and Xiao, S. 2016. Paired carbonate and organic carbon isotope variations of the Ediacaran Doushantuo Formation from an upper slope section at Siduping, South China, *Precambrian Research*, 273: 53-66.
- Watts, K.F. 1988. Triassic carbonate submarine fans along the Arabian platform, Sumeini Group, Oman, *Sedimentology*, 35: 43-71.
- Webster, J.M., Beaman, R.J., Puga-Bernab  a, A., Ludman, D., Renema, W., Wust, R.A.J., George, N.P.J., Reimer, P.J., Jacobsen, G.E., and Moss, P. 2012. Late Pleistocene history of turbidite sedimentation in a submarine canyon off the northern Great Barrier Reef, Australia, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 331-332: 75-89.
- Wearmouth, C. 2018. Spatial, temporal, and physical origin of matrix-poor to matrix-rich sandstones, Neoproterozoic Windermere Supergroup, British Columbia, Canada, M.Sc. Thesis, University of Ottawa, Ottawa, ON.
- Weimer, P. 1990. Sequence stratigraphy, facies geometries, and depositional history of the Mississippi fan, Gulf of Mexico, *American Association of Petroleum Geologists Bulletin*, 74: 425-453.
- Wentworth, C.K. 1922. A scale of grade and class terms for clastic sediments, *The Journal of Geology*, 30: 377-392.
- Williams, G.E. 1975. Late Precambrian glacial climate and the Earth's obliquity, *Geological magazine*, 112: 441-465.
- Williams, G.E. 1993. History of the Earth's Obliquity, *Earth Science Reviews*, 34: 1-45.
- Williams, G.E. 2000. Geological constraints on the Precambrian history of Earth's rotation and the Moon's orbit, *Reviews of Geophysics*, 38: 37-59.
- Williams, P.M., and Gordon, L.I. 1970. Carbon-13: carbon-12 ratios in dissolved and particulate organic matter in the sea, *Deep Sea Research and Oceanographic Abstracts*, 17: 19-27.
- Wilson, J.L. 1967. Cyclic and reciprocal sedimentation in Virgilian strata of southern New Mexico, *Geological Society of America Bulletin*, 78: 805-818.

- Wilson, P.A., and Roberts, H.H. 1992. Carbonate-periplatform sedimentation by density flows: A mechanism for rapid off-bank and vertical transport of shallow-water fines, *Geology*, **20**: 713-716.
- Wood, R.A., Poulton, S.W., Prave, A.R., Hoffman, K.H., Clarkson, M.O., Guilbaud, R., Lyne, J.W., Tostevin, R., Bowyer, F., Penny, A.M., Curtis, A. and Kasemann, S.A. 2015. Dynamic redox conditions control late Ediacaran metazoan ecosystems in the Nama Group, Namibia, *Precambrian Research*, **261**: 252-271.
- Worsley, T.R., and Kidder, D.L. 1991. First-order coupling of paleogeography and CO₂ with global surface temperature and its latitudinal contrast, *Geology*, **19**: 1161-1164.
- Wynn, R.B., Cronin, B.T., and Peakall, J. 2007. Sinuous deep-water channels: Genesis, geometry, architecture, *Marine and Petroleum Geology*, **24**: 341-387.
- Xiao, S., and Kaufman, A.J. 2006. Neoproterozoic geobiology and paleobiology. Springer. Dordrecht, Netherlands.
- Yose, L.A., and Heller, P.L. 1989. Sea-level control of mixed-carbonate-siliciclastic, gravity-flow deposition: Lower part of the Keeler Canyon Formation (Pennsylvanian), southeastern California, *Geological Society of America Bulletin*, **101**: 427-439.
- Zhou, C., and Xiao, S. 2007. Ediacaran $\delta^{13}\text{C}$ chemostratigraphy of South China, *Chemical Geology*, **237**: 107-126.

Appendix A: Samples analyzed for $\delta^{13}\text{C}_{\text{carb}}$, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}_{\text{org}}$

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-HS-2	20.07	CT1	Tbd (calcarenite)	-3.84	-16.12				N53°04'20.5" W120°27'32.7"
16-HS-3	23.18	CT1	Tabd (calcarenite)	-3.05	-17.19	0.09	-20.51	17.46	N53°04'19.9" W120°27'32.7"
15-HS-3	23.49	CT1	UF (calcarenite)	-2.17	-16.63	0.09			
16-HS-4	25	CT1	Tbcd (calcarenite)	-1.69	-16.29	0.18	-20.52	18.83	N53°04'20.0" W120°27'32.9"
16-HS-5	25.81	CT1	Tabd (calcarenite)	-1.44	-16.81				N53°04'19.6" W120°27'32.7"
16-HS-6	30	CT1	Tbcd (calcarenite)	Not analyzed					N53°04'19.5" W120°27'33.1"
16-HS-50	44	BST1	Massive yellow/pale orange calcite clast	-2.24	-13.55				N53°04'18.9" W120°27'32.9"
16-HS-10	44	BST 1	Dark brown Carbonate-cemented granular conglomerate	-0.13	-16.93				N53°04'18.9" W120°27'32.9"
16-HS-11	47.61	BST 1	Light grey Tab calcarenite	Not analyzed					N53°04'18.9" W120°27'33.2"
15-HS-7	48.5	BST1	Light grey Ta calcarenite	-0.06	-15.92				
16-HS-12	49.8	BST 1	Dark brown LVC calcarenite / granular conglomerate	-0.3	-17.05				N53°04'18.9" W120°27'33.7"
16-HS-13	53.35	BST 1	Dark brown UM Tab (calcarenite)	0.31	-16.93				N53°04'18.8" W120°27'33.6"
16-HS-14	56.86	BST 1	LM calcarenite	Not analyzed					N53°04'18.7" W120°27'33.6"
16-HS-15	60.4	BST 1	LC dark brown calcarenite	0.48	-16.89				N53°04'18.1" W120°27'33.3"
16-HS-16	62.46	Proximal Levees	Light brown Tbcd (calcarenite)	-0.84	-17.81				N53°04'18.2" W120°27'33.9"
16-HS-17	66.49	Proximal Levees	Tbd (calcarenite)	Not analyzed					N53°04'17.6" W120°27'33.1"

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-HS-18	69	Proximal Levees	Tbcd (calcarenite)	2.43	-16.86				N53°04'18.4" W120°27'34.0"
16-HS-19	71.37	Proximal Levees	Te calcilutite	2.79	-16.82				N53°04'18.3" W120°27'34.3"
16-HS-20	72	Proximal Levees	Tbc (calcarenite)	1.82	-16.73				N53°04'18.3" W120°27'34.1"
15-HS-10	73.89	Proximal Levees	Td calcilutite	1.89	-9.62				
16-HS-21	76.84	SC Mud dominated	LM calcarenite	0.51	-16.63	0.13	-22.52	23.03	N53°04'18.3" W120°27'34.2"
15-HS-11	77.8	SC Mud dominated	Calcarenite	2.65	-13.9				
16-HS-24	85.74	CT2	LF calcarenite	-1.03	-18.27				N53°04'18.3" W120°27'34.7"
15-HS-12	85.74	CT2	LF calcarenite	-0.91	-16.97				
15-HS-13	87	CT2	Te calcilutite	1.05	-16.25				
16-HS-23	87.07	CT2	Td calcilutite	0.98	-16.63	4.08	-0.54	1.52	
16-HS-25	89.46	CT2	Te calcilutite	1.17	-16.7				N53°04'18.5" W120°27'34.8"
				1.15	-16.62				
16-HS-26	91.68	CT2	Te calcilutite	1.3	-16.69	4.84	-0.35	1.65	N53°04'18.2" W120°27'35.1"
16-HS-27	93.57	CT2	Td calcilutite	0.98	-17.01	0.21	-15.87	16.85	N53°04'18.4" W120°27'35.3"
16-HS-28	95.2	CT2	UF calcarenite	0.67	-16.89				N53°04'18.4" W120°27'35.0"
				0.76	-16.76				
16-HS-29	96.67	CT2	Te calcilutite	1.33	-16.66	1.64	-3.95	5.28	
15-HS-16	96.89	CT2	Td calcilutite	1.05	-16.25				
16-HS-30	97.29	CT2	Te calcilutite	1.35	-16.68				N53°04'18.5" W120°27'36.2"
16-HS-31	98.82	BST 2	UM Tabd (calcarenite)	-2.38	-18.55				N53°04'17.6" W120°27'34.6"

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-HS-32	101.67	BST 2	Greenish grey sandstone	Not analyzed					N53°04'18.1" W120°27'35.6"
16-HS-33	104.88	BST 2	UM sandstone	Not analyzed					N53°04'18.4" W120°27'36.4"
16-HS-34	113.61	BST 2	UC-LM Tab (calcarenite)	1.07	-16.7				N53°04'18.1" W120°27'36.6"
16-HS-35	116.36	BST 2	Tab (calcarenite)	0.79	-16.6				N53°04'18.1" W120°27'36.7"
16-HS-36	120.36	123s-1	Te calcilutite	1.56	-16.63	4.84	0.03	1.53	N53°04'17.5" W120°27'36.7"
15-HS-21	120.36	123s-1	Te calcilutite	2.1	-16.74				
16-HS-37	122.28	123s-1	Te calcilutite	1.85	-16.67				N53°04'17.7" W120°27'36.6"
				1.85	-16.55				
16-HS-38	124.61	123s-1	Te calcilutite	2.24	-16.66	5.71	0.96	1.28	N53°04'17.6" W120°27'36.7"
15-HS-20	126.08	123s-1	Te calcilutite	2.01	-16.58				
16-HS-40	127.68	123s-1	Te calcilutite	0.04	-16.49	2.92	-2.48	2.52	N53°04'17.6" W120°27'36.1"
16-HS-41	130.4	123s-1	Dark brown calcarenite	1.05	-16.57	2.01			N53°04'17.3" W120°27'36.9"
16-HS-43	136.16	123s-2	Te calcilutite	1.09	-16.54				N53°04'17.0" W120°27'36.4"
				1.05	-16.48				
15-HS-24	138.6	123s-2	Tde calcilutite	-0.48	-16.61				
16-HS-45	144.23	Upper SST	LC Tabde (calcarenite)	-0.56	-17.73				N53°04'16.8" W120°27'37.1"
16-HS-46	147.92	Upper SST	UM Tad (calcarenite)	-1.62	-15.85				N53°04'17.2" W120°27'38.0"
16-HS-47	150.44	Upper SST	UC Tad (calcarenite)	-2.05	-15.86				N53°04'17.5" W120°27'38.5"
16-HS-48	154	Upper SST	UM-LM Tbd (calcarenite)	-2.71	-19.55				N53°04'16.9" W120°27'37.9"

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-HS-49	158	CT3	Td calcilutite	-1.31	-16.62	0.26	-16.26	14.95	N53°04'16.6" W120°27'37.7"
15-HS-27	159.27	CT3	UF calcarenite	-1.36	-16.28				
		CT3		-1.41	-16.25				
16-HS-51	169	CT3	Tde calcilutite	-1.57	-16.79	0.5	-16.3	14.73	
15-HS-28	170	CT3	Td calcilutite	-1.83	-17.23				
16-CC-62	6.3	CT1	Fine-grained calcarenite	-5.21	-15.3				N53°03'11.2" W120°26'05.9"
15-CCS-1	6.3	CT1		-5.26	-15.44				
16-CC-63	9.27	CT1	Fine-grained calcarenite	-4.84	-15.2				N53°03'11.0" W120°26'05.6"
16-CC-61	9.56	CD1	Calcdebrite	-3.98	-15.39				
16-CC-60	14.71	CD1	Calcdebrite	-3.75	-15.42				
16-CC-64	18.45	SC Mud dominated	Calcarenite	-3.53	-15.67				N53°03'01.4" W120°26'03.8"
16-CC-1	26.32	BST1	Dark brown Carbonate-cemented granular conglomerate	1.53	-15.79				N53°03'07.2" W120°26'03.8"
16-CC-3	27.77	BST1	LVC grey sandstone	Not analyzed					
16-CC-2	28.33	BST1	Dark brown Carbonate-cemented granular conglomerate	0.83	-15.61				N53°03'05.8" W120°26'03.1"
16-CC-4	30.45	BST1	UC dark brown calcarenite	-2.22	-15.61				N53°03'03.2" W120°26'00.0"
16-CC-5	33.51	BST1	Dark brown Ta calcarenite	Not analyzed					N53°03'03.2" W120°25'58.8"
16-CC-6	36.08	BST1	UC calcarenite	-1.69	-15.49				N53°03'01.4" W120°25'58.6"
16-CC-7	39.08	BST1	UC dark brown calcarenite, some UVC grains	0.01	-16.05				N53°03'01.4" W120°25'58.4"

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-CC-9	47.54	SC Mud dominated	LM-UF Tcd calcarenite	-2.13	-15.15				N53°03'02.8" W120°26'00.5"
				-2.05	-15.07				
16-CC-10	52.55	SC Mud Dominated	UF Tbc calcarenite	-1.8	-15.2				N53°03'02.6" W120°26'00.9"
16-CC-11	57	Proximal Levees	Calcarenite	-1.53	-15.69				N53°03'02.4" W120°26'01.0"
16-CC-13	69.53	CT2	Td calcilutite	-0.05	-15.74	0.33	-14.27	14.22	N53°03'01.6" W120°26'01.2"
15-CCS-5	69.53	CT2	Td calcilutite	-0.36	-15.66				
16-CC-14	71	CT2	Td calcilutite	-0.07	-15.72				N53°03'01.5" W120°26'01.2"
16-CC-15	73.69	CT2	Td calcilutite	2.01	-15.93	1.63	-0.57	2.58	N53°03'01.2" W120°26'01.2"
16-CC-16	75.18	CT2	Td calcilutite	2.1	-15.87				N53°03'01.2" W120°26'01.2"
16-CC-17	76.64	CT2	Td calcilutite	2.47	-15.75				N53°03'01.1" W120°26'01.3"
16-CC-18	78.24	CT2	Td calcilutite	2.35	-15.88				N53°03'01.2" W120°26'01.3"
16-CC-20	79.73	CT2	Te calcilutite	2.21	-15.8	0.36	-0.33	2.54	
16-CC-21	81.73	CT2	Td calcilutite	2.55	-15.83				N53°03'01.0" W120°26'01.4"
16-CC-19	83.73	CT2	Td calcilutite	2.5	-15.9				N53°03'01.0" W120°26'01.5"
16-CC-22	85.73	CT2	Td calcilutite	2.52	-15.94	0.61	0.19	2.33	N53°03'01.0" W120°26'01.5"
16-CC-23	87.53	CT2	Te calcilutite	2.63	-15.89	0.32	-12.07	14.7	N53°03'00.8" W120°26'01.4"
				2.65	-15.81	0.32	-12.07	14.72	
16-CC-24	89.03	CT2	Td calcilutite	2.55	-15.87				N53°03'00.8" W120°26'01.5"

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-CC-25	90.14	CT2	Td calcilutite	2.69	-15.89				
16-CC-26	93.64	CT2	Td calcilutite	2.43	-15.8				N53°03'00.6" W120°26'01.8"
16-CC-27	95.64	CT2	Td calcilutite	2.57	-15.55				N53°03'01.0" W120°26'02.4"
16-CC-28	96.76	CT2	Td calcilutite	2.29	-15.66				N53°03'01.0" W120°26'02.5"
15-CCS-7	101.74	123-1	Te calcilutite	2.81	-15.8				
				2.85	-15.79				
16-CC-38	102.02	123-1	Te calcilutite	2.73	-15.84	2.52	-2.84	5.57	N53°03'05.9" W120°26'09.5"
16-CC-39	104.12	123-1	UM Ta calcarenite	2.14	-15.15				N53°03'06.6" W120°26'07.4"
16-CC-40	106.92	123-1	Td calcilutite	3.19	-15.8				N53°03'06.4" W120°26'07.3"
16-CC-42	108.65	123-1	UF Tade (calcarenite, calcilutite)	3.44	-15.78				N53°03'06.6" W120°26'07.7"
16-CC-43	110.15	123-1	UVC Ta calcarenite	2.98	-15.81				
15-CCS-14	112.39	123-1		2.57	-15.97				
16-CC-44	117.15	123-1	Td calcilutite	2.82	-15.89				N53°03'06.6" W120°26'08.3"
		123-1		2.81	-15.93				
16-CC-45	119.15	123-1	VF Tade (calcarenite, calcilutite)	2.53	-15.98	0.26	-14.25	16.78	N53°03'06.6 W120°26'
16-CC-46	121.15	123-1	Td calcilutite	2.39	-15.91				
15-CCS-13	122.37	123-1	Te calcilutite	2.82	-15.62				
16-CC-47	123.15	123-1	Te calcilutite	2.46	-15.82	0.84	-8.8	11.26	
16-CC-48	124.9	123-1	LM Tab calcilutite	2.5	-16.15				N53°03'06.5" W120°26'08.7"

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-CC-49	126.9	123-1	Te calcilutite	2.87	-15.67	4.06	-5.39	8.26	N53°03'06.3" W120°26'08.7"
16-CC-50	129.25	123-1	Fine-grained calcarenite	2.02	-15.51				
15-CCS-12	132.38	123-1		2.82	-16.04				
16-CC-51	133.05	123-1	Td calcilutite	2.53	-15.93	0.95	-11.06	13.59	N53°03'06.2" W120°26'08.4"
16-CC-52	135.05	123-1	Td calcilutite	2.78	-15.7				
16-CC-53	141.72	123-1	Td calcilutite	2.5	-15.79				
		123-1		2.47	-15.79				
16-CC-54	142.71	123-1	LM Tabcd (calcarenite)	0.48	-15.19				N53°03'05.9" W120°26'09.4"
16-CC-55	147.71	123-1	LM calcarenite	0.62	-15.48				N53°03'06.2" W120°26'10.0"
16-CC-56	156.2	SC Mud Dominated between 123s	Te calcilutite	0.45	-15.9	0.35	-14.41	14.86	N53°03'06.0" W120°26'10.0"
16-CC-57	166.47	123-2	Te calcilutite	0.08	-16.06				N53°03'05.6" W120°26'10.4"
				0.2	-15.94				
16-CC-58	169.72	123-2	Td calcilutite	-6.34	-15.73	0.2	-24.18	17.84	N53°03'05.5" W120°26'10.6"
15-CCs-11	171.21	123-2	Te calcilutite	0.25	-15.85				
16-CC-59	171.72	123-2	LM calcarenite	-0.56	-15.17				N53°03'04.04" W120°26'09.9"
16-CC-29	185.33	CT3	Tbd (calcarenite)	-1.55	-15.41	0.74	-14.85	13.3	N53°03'03.3" W120°26'09.2"
15-CCS-10	185.41	CT3	Td calcilutite	-1.26	-15.85				
16-CC-30	186.81	CT3	LF-UVF calcarenite	-1.38	-15.51				N53°03'03.4" W120°26'09.5"
16-CC-31	188.59	CT3	UF calcarenite	-1.92	-15.29				N53°03'03.6" W120°26'09.6"
16-CC-32	190.09	CT3	Te calcilutite	-0.66	-15.92	1.75	-16.04	15.38	N53°03'03.6" W120°26'09.7"

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-CC-33	191.83	CT3	LF calcarenite	-2.52	-15.83				N53°03'04.1" W120°26'10.7"
16-CC-34	193.68	CT3	UF calcarenite	-3.99	-15.52				
16-CC-37	195.75	CT3	Red carbonaceous mud	-3.83	-16.01	0.19	-17.63	13.8	
16-CC-36	196.33	CT3	Te calcilutite	-5.61	-16.29				N53°03'04.2" W120°26'10.7"
16-CC-35	196.93	CT3	Calcarenite	-6.28	-15.53				
15-CCS-9	197.49	CT3	Td calcilutite	-5.57	-15.46				
16-CC-Vein				2.28	-15.97				
16-MR-CD1 Clast 3	2.2	CD1	Yellow calcite clast with algal laminations	1.5	-16.28				N53°12'05.0" W120°30'06.5"
16-MR-CD1 Clast 1	2.35	CD1	Yellow calcite clast with algal laminations	4.87	-15.99				N53°12'07.3" W120°30'10.5"
16-MR-1	4.1	CD1	Calcidebrite	4.43	-16.63				N53°12'04.4" W120°30'05.0"
16-MR-CD1 Clast 2	4.2	CD1	Orange calcite clast	3.5	-16.61				N53°12'05.6" W120°30'07.6"
16-MR-2	11.39	Proximal Levees	Calcarenite	4.38	-16.6				N53°12'05.0" W120°30'07.0"
15-MR-6	14.85	Proximal Levees	Calcarenite	4.16	-16.78				
16-MR-3	15.09	CT1	UF/LM calcarenite	3.78	-16.45				N53°12'05.1" W120°30'07.3"
15-MR-2	15.91	CT1	Calcarenite	3.04	-16.68				
16-MR-4	16	CT1	Tde calcilutite	3.84	-16.36				N53°12'05.0" W120°30'07.4"
15-MR-G1	16	CT1	Te calcilutite	4.13	-16.65				
15-MR-3	16.39	CT1	Te calcilutite	3.66	-17.36				
15-MR-5	18.42	CT1	Td calcilutite	3.53	-17.28				

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-MR-5	18.5	CT1	Td calcilutite	3.95	-16.47	0.41	-11.7	15.65	N53°12'04.9" W120°30'07.3"
16-MR-6	20.22	CT1	Te calcilutite	3.45	-16.59				N53°12'04.9" W120°30'07.5"
16-MR-7	24.1	CT1	Te calcilutite	4.13	-16.5	0.72	-11.18	15.31	N53°12'04.5" W120°30'07.1"
15-MR-G2	24.1	CT1	Te calcilutite	3.9	-16.72				
16-MR-BST Clast	24.21	BST 1	Yellow calcite clast with algal laminations	2.6	-15.42				N53°12'04.0" W120°30'06.2"
16-MR-8	24.29	BST 1	Ta (calcarenite)	3.96	-16.58				N53°12'04.8" W120°30'07.7"
15-MR-8	26.61	BST 1		3.99	-16.82				
16-MR-9	28.16	BST 1	Dark brown calcarenite	4.58	-16.6				N53°12'04.3" W120°30'07.1"
16-MR-10	31.25	BST 1	Coarse-grained calcarenite	4.84	-16.18				N53°12'04.2" W120°30'07.3"
16-MR-11	35.13	Proximal Levees	Calcarenite	2.22	-16.11				N53°12'04.2" W120°30'07.5"
16-MR-12	40.2	Proximal Levees	Medium-grained calcarenite	1.92	-16.11				N53°12'04.3" W120°30'08.2"
15-MR-G4	56.81	CTs above BST	Te calcilutite	2.8	-16.16				
16-MR-17	56.84	CTs above BST	Td calcilutite	2.68	-16.07				N53°12'04.1" W120°30'09.1"
16-MR-18	58.84	CTs above BST	Calcarenite	2.42	-16.11				N53°12'03.8" W120°30'08.6"
16-MR-19	62.63	CTs above BST	Td calcilutite	1.79	-16.02	0.18	-14.32	16.11	N53°12'03.7" W120°30'08.9"
15-MR-G5	62.76	CTs above BST	Te calcilutite	2.18	-15.79				
16-MR-23	85.97	FG SC Dominated	Light reddish brown calcarenite	0.78	-16.11				N53°12'03.1" W120°30'09.6"
16-MR-24	92.96	Proximal Levees	Td calcilutite	0.57	-15.97				N53°12'03.0" W120°30'09.8"

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
16-MR-24	92.96	Proximal Levees	Td calcilutite	0.55	-15.83				
16-MR-25	95.74	FG SC Dominated	Reddish brown calcarenite	Not analyzed					N53°12'03.0" W120°30'09.9"
15-MR-G6	118.64	FG SC Dominated	UF calcarenite	0.45	-15.95				
16-MR-28	118.86	CD2	Debrite (siliciclastic mud matrix)	Not analyzed					N53°12'02.4" W120°30'11.0"
16-MR-CD2 Clast	126.91	CD2	Orangish yellow calcite clast	3.63	-11.08	8.4	2.25	1.38	N53°12'01.5" W120°30'09.9"
16-MR-29	130.73	CD2	Calceibrite	1.54	-16.31				N53°12'02.0" W120°30'11.3"
16-MR-N	130.73	CD2	Yellow calcite clast with algal laminations	2.41	-13.97				
16-MR-30	131.39	CT2	Td calcilutite	2.07	-15.87				N53°12'02.1" W120°30'11.4"
15-MR-G7	131.39	CT2	Te calcilutite	2.78	-16.13				
16-MR-31	136.89	Slide	Calceibrite within slide	4.89	-16.25				N53°12'01.9" W120°30'11.3"
15-MR-10	137.62	Slide	Calceibrite within slide	5.2	-16.22				
16-MR-32	138.89	Slide	Calceibrite within slide	4.35	-16.53				N53°12'01.9" W120°30'11.3"
16-MR-33	143.57	Slide	Td calcilutite	3.31	-15.96	0.19	-12.57	15.88	N53°12'02.0" W120°30'11.8"
15-MR-G8	145	123s - 1	Td calcilutite	3.29	-15.63				
16-MR-35	147.77	123s -1	Td calcilutite	2.08	-14.96	0.07	-14.4	16.48	N53°12'01.3" W120°30'10.9"
				2.15	-14.9				
16-MR-36	150.78	FG SC Dominated	Te calcilutite	3.02	-15.74	0.6	-12.5	15.52	N53°12'01.3" W120°30'11.3"
16-MR-37	151.37	FG SC Dominated	Te calcilutite	2.15	-15.45	0.17	-14.14	16.29	N53°12'01.1" W120°30'11.2"
15-MR-G9	160.75	FG SC Dominated	UF calcarenite	1.10	-15.55				

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
15-MR-11	186.51	FG SC Dominated	Calcarenite	-0.38	-15.65				
16-MR-39	190.37	FG SC Dominated	Coarse-grained calcarenite	0.43	-15.62				N53°12'00.2" W120°30'12.2"
16-MR-41	230.52	FG SC Dominated	Reddish brown calcarenite	-0.29	-15.02				N53°11'59.4" W120°30'14.3"
16-MR-43	242.92	Channel	Reddish brown calcarenite	-0.15	-16.68				N53°11'59.0" W120°30'14.4"
16-MR-44	244.81	Channel	Coarse-grained Tb (calcarenite)	2.15	-15.81				N53°11'58.6" W120°30'13.9"
16-MR-Clast below CD3	246.36	Channel	Orange carbonate clast	-1.19	-15.02				N53°11'59.9" W120°30'16.4"
16-MR-45	246.95	Channel	Red brown calcarenite	Not analyzed					N53°11'58.6" W120°30'13.9"
16-MR-CD3 Clast	247.23	Calci-Slide	Brown with hint of orange carbonate clast	2.18	-15.68				N53°12'00.6" W120°30'17.5"
16-MR-46	247.47	CT3	Td calcilutite	1.91	-16.11				N53°11'25.6" W120°30'14.1"
15-MR-G12	247.49	CT3	Te calcilutite	1.86	-16.39				
		CT3		1.88	-16.33				
16-MR-52	247.78	CT3	Td calcilutite	1.59	-16.16				
15-MR-G13	250.21	CT3	Td calcilutite	2.28	-16.35				
16-MR-47	250.3	CT3	Reddish brown calcarenite	2.49	-16.44				N53°11'58.4" W120°30'14.1"
16-MR-48	252.13	CT3	Reddish brown calcarenite	1.4	-15.75				N53°11'58.3" W120°30'14.3"
16-MR-49	256.34	CT3	Reddish brown calcarenite	0.78	-16.69				
16-MR-50	258.66	CT3	Te calcilutite	1.76	-16.18	0.7	-14.36	16.12	
16-MR-51	260	CT3	Te calcilutite	2.82	-16.3				
		CT3		2.74	-16.16				

Sample	Stratigraphic Height (m)	Element	Lithology	$\delta^{13}\text{C}_{\text{carb}}$	$\delta^{18}\text{O}$	%C	$\delta^{13}\text{C}_{\text{org}}$	$\Delta\delta$	GPS Coordinates
15-MR-14	261.57	CT3	Calcarenite	0.16	-16.03				
RAR-R071-B	71.43	CT2		0.58	-15.27				
RAR-R071-D	73.5	CT2		1.20	-16.63				
RAR-R071-E	76.01	CT2		1.97	-15.16				
RAR-R071-F	78.73	CT2		1.69	-16.17				
RAR-R071-H	83.82	CT2		2.41	-15.51				
RAR-R071-I	85.4	CT2		3.28	-15.15				
RAR-R071-J	91.52	CT2		3.52	-15.06				
RAR-R071-L	101.9	123-1		3.52	-15.26				
RAR-R071-O	106.58	123-1		3.81	-15.43				
RAR-R071-Q	118.91	123-1		3.19	-15.21				
RAR-R071-R	121.92	123-1		3.70	-15.22				
RAR-R071-T	140.9	123-1		2.26	-15.32				
RAR-R071-U	154.2	SC Mud Dominated between 123s		2.54	-15.36				
RAR-R071-W	156.49	SC Mud Dominated between 123s		2.22	-16.04				
RAR-R071-X	160.9	SC Mud Dominated between 123s		1.82	-15.75				
RAR-R071-Z	171.83	123-2		1.69	-15.20				
RAR-R071-DD	186.72	CT3		1.36	-15.10				
RAR-R071-EE	189.5	CT3		-0.45	-15.18				
RAR-R071-FF	192.73	CT3		-1.24	-15.58				
RAR-R071-GG	195.02	CT3		0.66	-16.10				
RAR-R071-HH	197.2	CT3		-3.24	-15.73				

Appendix B: Samples analyzed for Mn and Sr

Sample	Stratigraphic Height (m)	Element	Lithology	Mn (ppm)	Sr (ppm)	Mn/Sr
16-HS-3	23.18	CT1	Tabd (calcarenite)	1947.3	184.1	10.58
16-HS-4	25	CT1	Tbcd (calcarenite)	2317.5	189	12.26
16-HS-25	89.46	CT2	Te calcilutite	429.5	1443.5	0.30
16-HS-27	93.57	CT2	Td calcilutite	380.6	821	0.46
16-HS-36	120.36	123s-1	Te calcilutite	966.3	1145.6	0.84
16-HS-38	124.61	123s-1	Te calcilutite	544.9	1302.7	0.42
16-HS-49	158	CT3	Td calcilutite	354.1	153.2	2.31
16-HS-51	169	CT3	Tde calcilutite	690.4	482.5	1.43
16-CC-15	73.69	CT2	Td calcilutite	307.7	875	0.35
16-CC-20	79.73	CT2	Te calcilutite	340.5	892.9	0.38
16-CC-22	85.73	CT2	Td calcilutite	276.6	1057.4	0.26
16-CC-38	102.02	123-1	Te calcilutite	400.3	991.8	0.40
16-CC-45	119.15	123-1	VF Tade (calcarenite, calcilutite)	504.3	715.6	0.70
16-CC-58	169.72	123-2	Td calcilutite	1523.3	112.1	13.59
16-CC-32	190.09	CT3	Te calcilutite	623.3	537.4	1.16
16-CC-37	195.75	CT3	Red carbonaceous mud	288.5	192	1.50
16-MR-5	18.5	CT1	Td calcilutite	244.4	1631.8	0.15
16-MR-7	24.1	CT1	Te calcilutite	336.6	2273.8	0.15
16-MR-19	62.63	CTs above BST	Td calcilutite	683.4	664.4	1.03
16-MR-CD2 Clast	126.91	CD2	Orangish yellow calcite clast	225.6	915.6	0.25
16-MR-33	143.57	Slide	Td calcilutite	1264.4	602.8	2.10
16-MR-35	147.77	123s -1	Td calcilutite	193.1	274.5	0.70
16-MR-36	150.78	FG SC Dominated	Te calcilutite	329.1	1061.7	0.31
16-MR-37	151.37	FG SC Dominated	Te calcilutite	250.2	216.5	1.16
16-MR-50	258.66	CT3	Te calcilutite	268.3	1058.4	0.25

Appendix C: Stratal element, facies and lithology proportion in each study area

Study Area	Total Thickness (m)	Distal Levees - Mud Rich		Distal Levees - Carbonate Rich		123s		Proximal Levees	
		Thickness	%	Thickness	%	Thickness	%	Thickness	%
Castle Creek South	191.61	80.34	41.9	33.63	17.6	53.22	27.8	4.04	2.1
Castle Creek North	191.65	77.65	40.5	39.5	20.6	34.8	18.2	4.56	2.4
Hill Section	161.41	59.34	36.8	24.87	15.4	15.36	9.5	13.32	8.3
Milk River South	263.57	148.69	56.4	24.32	9.2	9.71	3.7	32.34	12.3
Milk River North	255.81	121.2	47.4	23.34	9.1	17.17	6.7	36.22	14.2
Total	1064.05	487.22		145.66		130.29		90.48	

Study Area	Total Thickness (m)	Channels		Debrites		Slides		Scour Dominated	
		Thickness	%	Thickness	%	Thickness	%	Thickness	%
Castle Creek South	191.61	13.4	7.0	7.0	3.6	0	0.0	0.00	0.0
Castle Creek North	191.65	23.85	12.4	11.26	5.9	0	0.0	0	0.0
Hill Section	161.41	47.89	29.7	0.0	0.0	0.62	0.4	0.00	0.0
Milk River South	263.57	12.8	4.9	22.1	8.4	8.54	3.2	5.07	1.9
Milk River North	255.81	28.74	11.2	21.54	8.4	4.15	1.6	3.43	1.3
Total	1064.05	126.68		61.88		13.31		8.5	

Table C.1: Stratal element proportion in each study area.

Study Area	Total Thickness	Facies 1		Facies 2		Facies 3		Facies 4		Facies 5	
		Thickness	%	Thickness	%	Thickness	%	Thickness	%	Thickness	%
Castle Creek South	191.61	27.87	14.5	39.15	20.4	117.6	61.4	6.98	3.6	0	0.0
Castle Creek North	191.65	33.29	17.4	20.06	10.5	111.24	58.0	11.26	5.9	0	0.0
Hill Section	161.41	62.86	38.9	15.61	9.7	80.09	49.6	0	0.0	0.62	0.4
Milk River South	263.57	82.43	31.3	33.51	12.7	115.31	43.7	22.1	8.4	8.54	3.2
Milk River North	255.81	84.58	33.1	29.06	11.4	116.21	45.4	21.82	8.5	4.14	1.6
	1064.05	291.03		137.39		540.45		62.16		13.30	

Table C.2: Facies proportions in each study area.

Section	Total Thickness	Sand (SC)		Sand (CC)		Mud (SC)		Calclutite	
		Thickness	%	Thickness	%	Thickness	%	Thickness	%
Castle Creek South	191.61	11.74	6.1	9.87	5.2	109.5	57.1	42.36	22.1
Castle Creek North	191.65	11.75	6.1	2.74	1.4	89.91	46.9	41.01	21.4
Hill Section	161.41	5.87	3.6	21.02	13.0	80.68	50.0	12.59	7.8
Milk River South	263.57	4.05	1.5	50.39	19.1	146.89	55.7	19.15	7.3
Milk River North	255.81	6.17	2.4	53.68	21.0	137.2	53.6	20.23	7.9
Total	1064.05	39.58		137.7		564.18		135.34	

Section	Dark Brown CC SST		Light Brown CC SST		Slide		Debrite		Overburden	
	Thickness	%	Thickness	%	Thickness	%	Thickness	%	Thickness	%
Castle Creek South	4.93	2.6	6.23	3.3	0.00	0.0	6.98	3.6	0.00	0.0
Castle Creek North	11.43	6.0	7.74	4.0	0.00	0.0	11.26	5.9	15.8	8.2
Hill Section	18.67	11.6	19.74	12.2	0.62	0.4	0.00	0.0	2.23	1.4
Milk River South	7.44	2.8	3.33	1.3	8.54	3.2	22.10	8.4	1.68	0.6
Milk River North	8.27	3.2	4.29	1.7	4.15	1.6	21.81	8.5	0.00	0.0
Total	50.74		41.33		13.31		62.15		19.71	

Table C.3: Lithological proportions in each study area.

Appendix D: Lithology and facies proportions in each study area

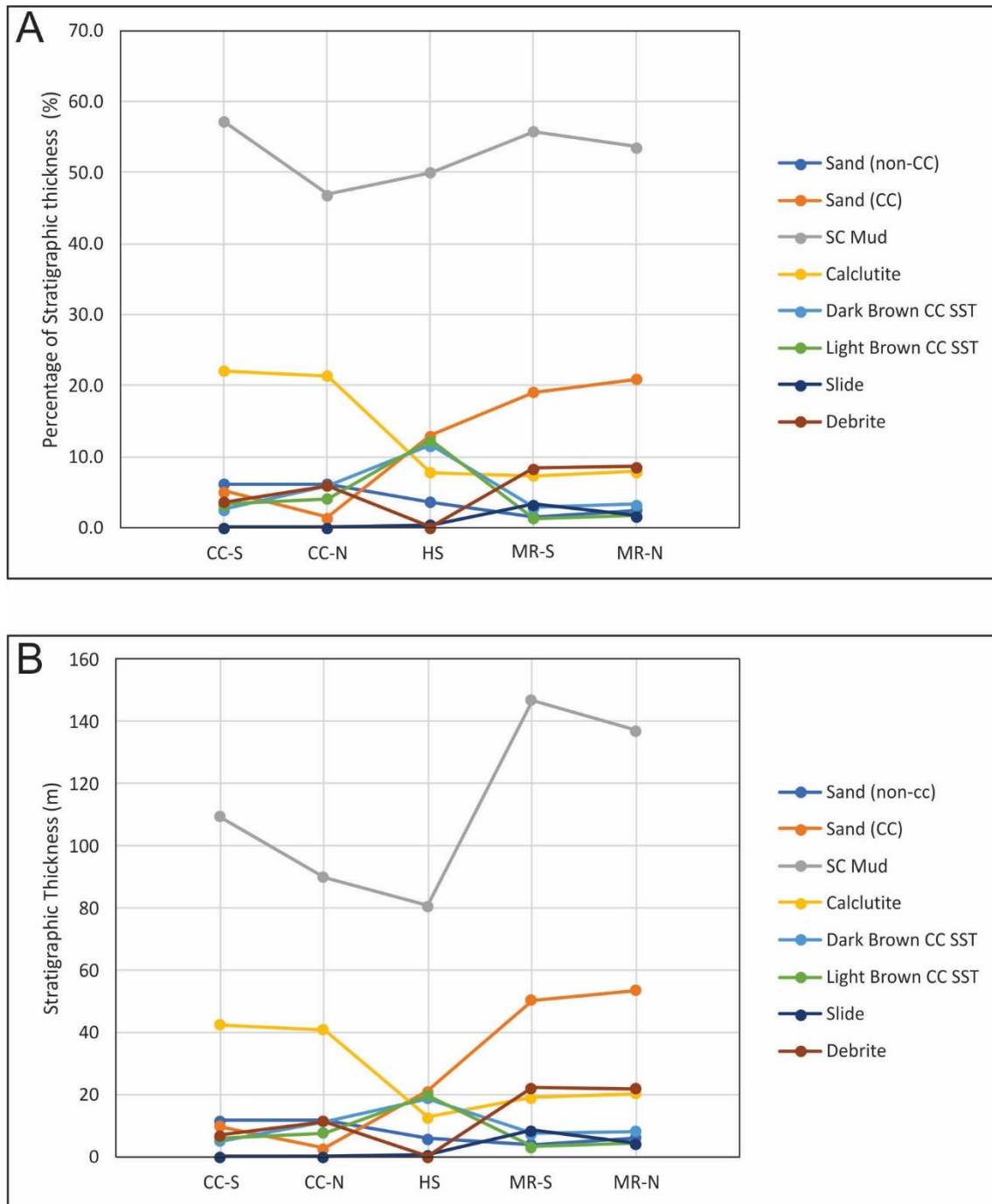


Figure D.1: Lithological proportions in each study area based on (A) percentage of stratigraphic thickness and (B) total stratigraphic thickness.

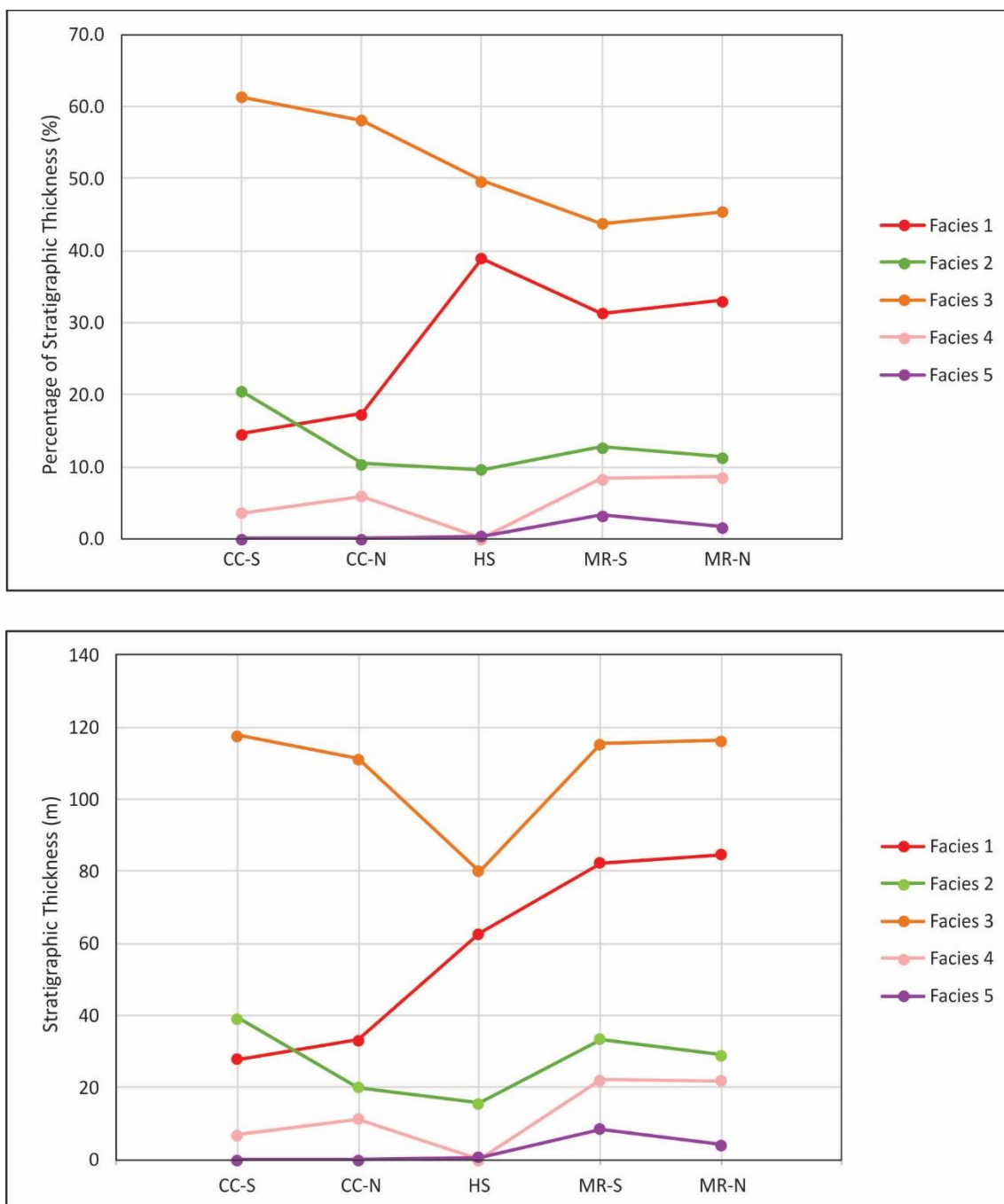


Figure D.2: Facies proportions in each study area based on (A) percentage of stratigraphic thickness and (B) total stratigraphic thickness.