

**SEDIMENTOLOGIC AND PETROGRAPHIC EVIDENCE OF FLOW
CONFINEMENT IN A PASSIVE CONTINENTAL MARGIN SLOPE
CHANNEL COMPLEX, ISAAC FORMATION, WINDERMERE
SUPERGROUP, BRITISH COLUMBIA, CANADA**

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Abstract

At the Castle Creek study area in east-central British Columbia a well-exposed section about 450 m wide and 30 m thick in the (Neoproterozoic) Isaac Formation was analyzed to document vertical and lateral changes in a succession of distinctively heterolithic strata. Strata are interpreted to have been deposited on a deep-marine levee that was sandwiched between its genetically related channel on one side and an erosional escarpment sculpted by an older (underlying) channel on the other. Flows that overspilled the channel (incident flow) eventually encountered the escarpment, which then set up a return flow oriented more or less opposite to the incident (from the channel) flow. This created an area of complex flow that became manifested in the sedimentary record as a highly tabular succession of intricately interstratified sand and mud overlain by an anomalously thick, plane-parallel interlaminated sand-mud unit capped finally by a claystone.

Résumé

Dans la zone d'étude de Castle Creek, situé au centre-est de la Colombie-Britannique, se trouve une section exposée de 30 m d'épaisseur et de 450 m de large dans la formation Isaac (Neoprotérozoïque). Cette formation a été analysée afin de documenter les changements verticaux et latéraux d'une succession de strates hétérolithiques. Il est suggéré que ces strates se sont déposés lors d'une transgression marine. Les sédiments sont alors en étau entre leur chenal et un escarpement dû à l'érosion, sculpté par un ancien chenal (sous-jacent). L'écoulement dépassant ce chenal atteint éventuellement l'escarpement, ce qui causera un retour de ce courant, en sens inverse de l'incident. La déposition des sédiments sera alors le reflet de la complexité de ce courant, se manifestant sous forme d'une succession de lits de sable et de boue finement interstratifiés, suivi par une unité parallèle au plan de sable-boue interlaminée anormalement épaisse, le tout recouvert par de l'argilite.

Extended Abstract

Along the base of passive continental margins, like the Canadian east coast, extensive deep-water clastic turbidite systems are present. Through the 21st century these systems have been increasingly recognized as prolific petroleum reservoirs and currently represent about 20-30% of global production. In spite of their excellent resource potential, deep-marine turbidite systems also come with many uncertainties and substantial drilling costs. In order to reduce economic risk and improve exploration success, petroleum geologists have turned to the ancient deep-marine sedimentary record to find meaningful analogues to describe passive margin deep-water turbidite systems. These deposits are often classified using the Bouma turbidite model; a generally upward-fining succession composed of five sharply bounded layers, each exhibiting a unique assemblage of predictable sedimentary structures and textures. In the Windermere Supergroup, however, in addition to classical turbidites, another type of strata, which superficially resemble turbidites, are observed. Where fully developed these strata consist of a basal massive or structured sandstone (F1) overlain sharply by structured, interstratified clayey sandstone and mudstone (F2). This layer is then overlain gradationally by distinctively rhythmically interstratified, sandy claystone capped sharply by structureless claystone (F3).

The study area crops out for over 450 metres laterally and can be subdivided into two parts. The first part consists of a basal amalgamated unit overlain by a heterolithic unit capped sharply by a thick succession of fine-grained, thin-bedded classical turbidites. The basal amalgamated unit is 15 metres thick, composed of amalgamated coarse-grained sandstone to pebble conglomerate, and interpreted to be a channel complex. The second part occurs at the same stratigraphic level, but 250 metres to the northwest. Here strata consist of a thick succession of rhythmically interlaminated, well-sorted, very fine-grained sandstone overlain sharply by mudstone that

distinctively lack traction transport structures and are intercalated with thin- to very thin-bedded turbidites (F4). Although the contact between the two parts is covered, such a dramatic lithological change over only a few hundred metres suggests that the channel complex scoured these fine-grained strata (F4) and then became partly filled with the amalgamated unit. These strata grade upward over about one metre into the heterolithic unit that forms a 13 metre-thick succession that comprises 2-5 metre-thick bedsets comprising F1 to F3.

Collectively, the lithological difference between strata in the heterolithic unit and classical turbidites is interpreted to be a consequence of flow confinement. Here two-way confinement was provided by the escarpment that had previously been sculpted by the channel complex in the fine-grained unit to the northwest, and growth of a levee associated with a younger, but out-of-the-plane-of-the-outcrop channel located to the southeast. Flows overspilled the new channel and over the developing levee where it eventually encountered the escarpment resulting in a return flow orientated in the opposite direction. The interaction of the incident (from the channel) and return flow (from the escarpment) created a local area of complex flow and depositional conditions, which here is manifested as the heterolithic unit. With continued deposition, relief of the escarpment was reduced and, in turn the strength of the return flow diminished. Later the development of an even younger channel further to the southeast, and the flattening of the local sea floor, allowed overspill flows to move unobstructed and unconfined and deposit the thin-bedded turbidites that cap the study area.

The ability to recognize the distinctive lithological characteristics of these heterolithic strata, but equally, appreciate and understand their interpretive significance, will aid in modelling these sub-seismic features, which also may be important in predicting reservoir

compartmentalization and fluid flow pathways in hydrocarbon reservoirs hosted in submarine channel complexes.

"If I have seen further it is by standing on the shoulders of Giants."
– Isaac Newton (1675)



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Table of Contents

Chapter 1: Thesis Introduction	1
1.1 Thesis Rationale.....	1
1.2 Windermere Supergroup	4
1.2.1 Regional Structure & Metamorphism.....	5
1.2.2 Regional Distribution.....	9
1.2.3 Regional Markers.....	10
1.2.4 Sediment Provenance & Paleoflow	13
1.2.5 Geochronology.....	14
1.2.6 Tectonic Model	15
1.2.7 Paleoenvironmental Interpretation.....	17
1.2.8 Castle Creek Study Area.....	19
1.2.9 Stratigraphy.....	20
1.2.9.1 Toby Formation	22
1.2.9.2 Irene Formation.....	23
1.2.9.3 Kaza Group	24
1.2.9.4 Isaac Formation.....	25
1.2.9.5 Cunningham & Yankee Belle Formations.....	25
1.3 Methodology	26
1.4 Previous Work	28
1.5 Objectives & Thesis Structure	30
Chapter 2: Deep-Marine Processes	32
2.1 Classification: Subaqueous Sediment-Gravity Processes.....	33
2.2 Subaqueous Mass Movements.....	39
2.3 Sediment-Gravity Flows	40
2.3.1 Cohesive Flows.....	40
2.3.2 Frictional Flows	42
2.3.2.1 Hyperconcentrated Density Flows.....	43
2.3.2.2 Concentrated Density Flows	44
2.3.2.3 Turbidity Currents.....	46
Fine-Grained Turbidity Currents	51
2.4 Bottom Currents.....	55
Chapter 3: Facies Descriptions and Interpretations	56
3.1 Terminology.....	56

3.2 Metamorphic Fabric.....	59
3.3 General Petrographic Characteristics.....	61
3.4 Facies 1: Massive or graded, structureless or planar/wavy-stratified clayey sandstone.....	64
3.4.1 Macroscopic Characteristics	64
3.4.2 Microscopic Characteristics.....	67
3.4.3 Interpretation.....	69
3.5 Facies 2: Graded, plane parallel-, cross- or wavy-stratified clayey sandstone with intercalated structureless sandy claystone	70
3.5.1 Macroscopic Characteristics	70
3.5.2 Microscopic Characteristics.....	74
3.5.3 Interpretation.....	85
3.6 Facies 3: Graded, plane parallel-laminated sandy claystone to massive or graded, structureless claystone	90
3.6.1 Macroscopic Characteristics	90
3.6.2 Microscopic Characteristics.....	91
3.6.3 Interpretation.....	93
3.7 Facies 4: Massive, interstratified, structureless clayey sandstone and claystone	96
3.7.1 Macroscopic Characteristics	96
3.7.2 Microscopic Characteristics.....	98
3.7.3 Interpretation.....	99
Chapter 4: Sedimentologic and Petrographic Evidence of Flow Confinement in a Passive Continental Margin Slope Channel Complex, Isaac Formation, Windermere Supergroup, British Columbia, Canada.....	100
4.1 Regional Geology	100
4.2 Methodology	107
4.3 Facies	110
4.3.1 F1: Massive or graded, structureless or planar/wavy-stratified clayey sandstone.....	111
4.3.2 F2: Graded, plane parallel-, cross- or wavy-stratified clayey sandstone with intercalated structureless sandy claystone	114
4.3.3 F3: Graded, plane parallel-laminated sandy claystone to massive or graded, structureless claystone	125
4.3.4 F4: Massive, interstratified, structureless clayey sandstone and claystone	128
4.4 Architectural Model.....	130
4.5 Stratigraphic Evidence of Flow Confinement.....	132
4.5.1 Tabularity	133
4.5.2 Lithological Characteristics: Vertical Trends	135

Sedimentation Events.....	135
Bedsets	138
Markov Chain Analysis	142
Paleocurrent Reversal	143
4.5.3 Lithological Characteristics: Lateral Trends in Grain Size within Facies	144
4.6 Comparison with Classical (Bouma) Turbidites	144
Chapter 5: Conclusions & Areas for Future Work	148
References	156
Appendix 1: General Petrographic Characteristics	174
Appendix 2: Correlation Panel.....	180

List of Figures

1. Thesis Introduction

Figure 1.1: Location of exposed strata of the Windermere Supergroup through western North America.....	4
Figure 1.2: The five physiographic regions of the western Canadian Cordillera.....	6
Figure 1.3: Regional geology of the Castle Creek study area.....	8
Figure 1.4: Size comparison of the restored Windermere turbidite system with modern and ancient turbidite systems	9
Figure 1.5: Geological map of the southern Canadian Cordillera showing the distribution of the Windermere Supergroup and a map of western Canada showing the location of different mountain ranges in the Canadian Cordillera and important localities associated with the Windermere Supergroup	11
Figure 1.6: Depositional model for the Old Fort Point Formation and the overlying calcareous sandstone	12
Figure 1.7: Map showing exposed rocks of the Windermere Supergroup in western Canada and the United States..	15
Figure 1.8: Schematic representation of the Windermere margin (western North America) during the Neoproterozoic	17
Figure 1.9: Schematic paleogeographic reconstruction of the Windermere turbidite system in the southern Canadian Cordillera.....	19
Figure 1.10: The Castle Creek study area located in the Cariboo Mountains of east-central British Columbia.....	20
Figure 1.11: Stratigraphic nomenclature of Windermere Supergroup (upper Proterozoic/ Neoproterozoic) and Lower Cambrian strata in the southern Canadian Cordillera	21
Figure 1.12: Stratigraphy of the Windermere Supergroup from regional to outcrop scale. A) Correlated schematic stratigraphy of the Windermere Supergroup. B) Composite stratigraphy of the Castle Creek study area in the Cariboo Mountains. C) Geologic map of the stratigraphy in the Castle Creek study area	22
Figure 1.13: Detailed stratigraphic correlation of the lower Isaac Formation in the Windermere turbidite system, Cariboo Mountains, British Columbia, Canada.....	30

2. Deep-Marine Processes

Figure 2.1: Flow chart showing the main flow initiation mechanisms, type of subaqueous sediment-gravity flow generated, and resultant depositional facies model for each process.....	34
Figure 2.2: Schematic of a shelf edge sediment failure and its downslope evolution from a coherent mass to a cohesionless sediment-water dispersion	35
Figure 2.3: Nomenclature of flow type based on volume percent sediment concentration	36
Figure 2.4: The classification of sediment-gravity flows by Mulder and Alexander (2001).....	38
Figure 2.5: Diagram illustrating the changes in flow characteristics of frictional flows across the different subdivisions	43
Figure 2.6: Idealized Lowe sequence for concentrated density currents showing the typical grain size and sedimentary structures for each of the six divisions as well as the primary sedimentation type.....	46
Figure 2.7: A) Laboratory saline turbidity current with a well-developed head and trailing body. B) Diagram illustrating the typical morphology and velocity profile of a turbidity current.....	48
Figure 2.8: Complete Bouma sequence showing grain size and sedimentary structures observed in each division ...	49

Figure 2.9: Comparison of the bedform stability diagram developed from flume experiments with 20 cm flow depth	50
Figure 2.10: Idealized facies model for fine-grained turbidites showing its distinctive component parts and their average thickness	53

3. Facies Descriptions and Interpretations

Figure 3.1: Terminology for the thickness of beds and laminae	57
Figure 3.2: Hierarchy of stratal terminology used here from micro to macro scales	58
Figure 3.3: Folk (1980) sandstone classification of facies 1-4	59
Figure 3.4: Photomicrographs showing recrystallized quartz formed by bulging recrystallization between framework quartz grains	61
Figure 3.5: Thickness variation in Facies 1 strata	65
Figure 3.6: Massive or graded, structureless or planar/wavy-stratified clayey F1 sandstone	66
Figure 3.7: Photomicrographs illustrating the range in grain size of F1 strata	68
Figure 3.8: Facies 2 showing a transition in the style of stratification, which stratigraphically upward consists of planar-stratified (F2.A) to low then moderate amplitude isolated cross-stratified (F2.B) sandstone	71
Figure 3.9: Thickness variation in Facies 2 strata	71
Figure 3.10: Grain size analysis along two vertical transects in a thin-bedded F2 division	73
Figure 3.11: Deposits with varying sand and mud content: lenticular (25:75), wavy (50:50), and flaser (75:25)	74
Figure 3.12: Wavy-stratified (F2.C), very thin-bedded F2 with coarser-grained, carbonate cemented interlaminae each draped with structureless claystone from macroscopic to microscopic scale	75
Figure 3.13: Comparison of F2 interstratified coarser-grained sand layers and finer-grained mud layers in thin section and photomicrograph	77
Figure 3.14: Photomicrographs of the sharp contact between fine-grained mud lamina overlain by a coarser-grained, very thin-bedded sand lamina	77
Figure 3.15: Flow chart of the common styles of stratification observed in matrix-poor, carbonate cemented, or matrix-rich F2 strata, and the corresponding stacking patterns where two or more styles of stratification stack to form a single F2 division	78
Figure 3.16: Examples of matrix-rich and matrix-poor planar-stratified F2.A strata	79
Figure 3.17: Comparison between climbing, low-amplitude, cross- (F2.A) to wavy-stratified F2.C strata and high-angle, cross-stratified climbing ripple (T_c) set in levee deposits, both in the Castle Creek study area	81
Figure 3.18: Photo and accompanying line diagram illustrating F2.B divisions consisting of two stacked styles of stratification	81
Figure 3.19: Examples of matrix-rich and matrix-poor wavy-stratified F2.C strata	83
Figure 3.20: Various styles of cross-stratification in the F2.B division, each truncated and then overlain by an anomalously thick (0.5-2 cm) mud layer where some mud layers separate cross-stratification both migrating and climbing in opposite directions	84
Figure 3.21: Thick (0.5-2 cm), laterally extensive mud layers that extend for at least several metres along strike	85
Figure 3.22: Schematic of the energy cascade illustrating the input of energy into the system at the integral scale of turbulence and the successive nonlinear break up and transfer of energy from large eddies into progressively smaller eddies, which eventually become so small that viscosity dominates and converts turbulent kinetic energy into heat	89

Figure 3.23: Sharp based strata of F3 that exhibit distinctive alternation of planar layers of sandy claystone overlain by claystone that collectively thin and fine upward, and ultimately are capped sharply by a structureless claystone from macroscopic to microscopic scale	90
Figure 3.24: Photo of a thin section showing planar-laminated sandy claystone capped by structureless claystone in F3 strata and SEM images at increasing magnification near the top of a structureless claystone cap in a F3 bed	92
Figure 3.25: Grain size analysis through a very thinly-bedded F3 bed based on point counting of 30 points identified every 0.5 mm along three separate vertical transects.....	93
Figure 3.26: Thin bed of F3 strata comprising multiple faint interlaminated layers of coarser-grained strata in the basal part and overlain by a faintly laminated claystone cap	96
Figure 3.27: Facies 4 from macroscopic to microscopic scale	97
Figure 3.28: Large, elongated chloritoid grains in both the basal coarse-grained and upper fine-grained layer in F4 strata	98

4. Manuscript: Sedimentologic and Petrographic Evidence of Flow Confinement in A Passive Continental Margin Slope Channel Complex, Isaac Formation, Windermere Supergroup, British Columbia, Canada

Figure 4.1: Stratigraphy of the Windermere Supergroup from regional to outcrop scale and an annotated aerial outcrop photo	102
Figure 4.2: Schematic paleogeographic reconstruction of the Windermere turbidite system in the southern Canadian Cordillera	104
Figure 4.3: Map of the five physiographic regions in the Canadian Cordillera also showing exposed rocks of the Windermere Supergroup in western Canada and the United States	105
Figure 4.4: Size comparison of the restored Windermere turbidite system with modern and ancient turbidite systems	107
Figure 4.5: Aerial drone photo of a 100 m lateral segment of the study area where all component facies are delineated	108
Figure 4.6: Hierarchy of stratal terminology used here from micro to macro scales.....	110
Figure 4.7: Photomicrographs illustrating the range of grain size in F1 strata	112
Figure 4.8: Massive or graded, structureless or planar/wavy-stratified clayey F1 sandstone	113
Figure 4.9: Grain size analysis along two stratigraphically-upward transects in a thin-bedded F2 division	115
Figure 4.10: Flow chart of the common styles of stratification observed in matrix-poor, carbonate cemented, or matrix-rich F2 strata, and the corresponding stacking patterns where two or more styles of stratification stack to form a single F2 division.....	116
Figure 4.11: Examples of matrix-rich and matrix-poor planar-stratified F2.A strata	117
Figure 4.12: Photo and accompanying line diagram illustrating F2.B divisions consisting of two stacked styles of stratification.....	119
Figure 4.13: Examples of matrix-rich and matrix-poor wavy-stratified F2.C strata.....	120
Figure 4.14: Various styles of cross-stratification in the F2.B division, each truncated and then overlain by an anomalously thick (0.5-2 cm) mud layer where some mud layers separate cross-stratification both migrating and climbing in opposite directions.....	121
Figure 4.15: Thick (0.5-2 cm), laterally extensive mud layers that extend for at least several metres along strike ..	122
Figure 4.16: : Schematic of the energy cascade illustrating the input of energy into the system at the integral scale of turbulence and the successive nonlinear break up and transfer of energy from large eddies into progressively smaller	

eddies, which eventually become so small that viscosity dominates and converts turbulent kinetic energy into heat	124
Figure 4.17: Sharp based strata of F3 that exhibit distinctive alternation of planar layers of sandy claystone overlain by claystone that collectively thin and fine upward, and ultimately are capped sharply by a structureless claystone from macroscopic to microscopic scale	126
Figure 4.18: Grain size analysis through a very thinly-bedded F3 bed based on point counting of 30 points identified every 0.5 mm along three separate vertical transects.....	126
Figure 4.19: Facies 4 from macroscopic (field) to microscopic (thin section) scale	129
Figure 4.20: Large, elongated chloritoid grains in both the basal coarse-grained and upper fine-grained layer in F4 strata	130
Figure 4.21: Schematic model of the study area showing the spatial and temporal relationship of the four major units that overlie an MTC used here as the stratigraphic datum – representative photos of each unit are provided	132
Figure 4.22: Thickening and then thinning trend of levee deposits away from the channel margin	135
Figure 4.23: A stack of four sedimentation events	136
Figure 4.24: Evolution of a confined flow and its interaction with sea floor topography	138
Figure 4.25: Detailed lateral correlation of bedset 4 from macroscopic to microscopic scale	139
Figure 4.26: Textural and dimensional analysis of four stacked heterolithic bedsets in log 2.....	141
Figure 4.27: Markov chain analysis consisting of 274 facies transitions between facies in log 2 with statistically significant results	142
Figure 4.28: Comparison of facies in this study and their apparent equivalents in a classical (Bouma) turbidite.....	147
Figure 4.29: Comparison between climbing, low-amplitude (6-8° at base to 3-4° near top of F2 layer), cross- to wavy-stratified F2 strata in and high-angle (30-32° lee face inclination), cross-stratified climbing ripple (T _c) set in levee deposits, both in the Castle Creek study area	147

5. Conclusion

Figure 5.1: Comparison between T_{bcd} classical Bouma turbidite and strata described in this study **Error! Bookmark not defined.**

List of Tables

Table 3.1: Summary of facies descriptions and interpretations	56
Table 4.1: Macroscopic and microscopic characteristics of facies F1 to F4.....	111

List of Abbreviations

Analytical Techniques

PPL	Transmitted, plane-polarized light
XPL	Transmitted, cross-polarized light
SEM	Scanning electron microscopy
XRD	X-ray powder diffraction
U-Pb	Uranium-lead dating
AFES	Alizarin red-S with potassium ferricyanide stain for ferroan carbonates

Minerals

Qtz	Quartz
Cal	Calcite
Mus	Muscovite
Dol	Dolomite
Chl	Chlorite

Measurement

Ga	Billion years ago
Ma	Million years ago
km²	Kilometre squared
g/L	grams per litre
km	Kilometre
Dm	Decametre
m	Metre
cm	Centimetre
mm	Millimetre
μm	Micrometre
mD	Millidarcy
2-D	Two dimensional
3-D	Three dimensional
km/h	Kilometres per hour
m/s	metres per second
cm/s	centimetres per second
°C	Degrees Celsius

Bouma Sequence (1962) – Medium-grained Turbidites

T_a	A division – basal massive or normally graded sandstone
T_b	B division – planar-stratified sandstone
T_c	C division – high-angle, small-scale ripple cross-stratified sandstone
T_d	D division – subtly parallel-laminated siltstone
T_e	E division – massive mudstone

Modifications to Bouma Sequence

e^t	Kuenen (1964) modification to Bouma E division – turbiditic mud deposition
e^p	Kuenen (1964) modification to Bouma E division – passive pelagic, hemipelagic, and/or biogenic
T_e	van der Linen (1969) and Hesse (1975) modification to Bouma E division – turbiditic mud deposition
T_f	van der Linen (1969) and Hesse (1975) modification to Bouma E division – passive pelagic, hemipelagic, and/or biogenic
E₁	Piper (1978) modification to Bouma E division – turbiditic basal laminated mud
E₂	Piper (1978) modification to Bouma E division – turbiditic graded mud
E₃	Piper (1978) modification to Bouma E division – turbiditic ungraded mud
F	Piper (1978) modification to Bouma E division – passive pelagic, hemipelagic, and/or biogenic

Lowe Sequence (1982) – Coarse-grained Turbidites

R₁	Basal structureless conglomerate
R₂	Inversely-graded conglomerate
R₃	Normally-graded conglomerate
S₁	Planar- and/or cross-stratified sandstone
S₂	Inversely-graded coarse- to fine-grained sandstone
S₃	Massive or normally graded coarse- to fine-grained sandstone

Stow and Shanmugam (1980) – Fine-grained Turbidites

T₀	Basal normally-graded, planar-laminated to cross-laminated sandstone
T₁	Mudstone with convolute silt laminae
T₂	Low-amplitude, long-wavelength cross-laminated with interstratified siltstone and mudstone
T₃	Planar-laminated with interstratified siltstone and mudstone
T₄	Mudstone with faint discontinuous siltstone laminae
T₅	Mudstone with subtle wispy or convolute silt laminae
T₆	Graded mudstone with dispersed silt lenses
T₇	Structureless mudstone (turbiditic)
T₈	Bioturbated mudstone (pelagic and/or hemipelagic sedimentation)

Facies (this thesis)

F1	Facies 1 – Massive or graded, structureless or planar/wavy-stratified clayey sandstone
F2	Facies 2 – Graded, plane parallel-, cross- or wavy-stratified clayey sandstone with intercalated structureless sandy claystone
F3	Facies 3 – Graded, plane parallel-laminated sandy claystone to massive or graded, structureless claystone
F4	Facies 4 – Massive, interstratified, structureless clayey sandstone and claystone

Misc.

ICC0	Isaac leveed-channel complex zero
OFP	Old Fort Point Formation
SCC	Southern Canadian Cordillera
SRMT	Southern Rocky Mountain Trench
TKE	Turbulent kinetic energy
WSG	Windermere Supergroup

Chapter 1: Thesis Introduction

1.1 Thesis Rationale

For centuries geologists have been on a quest to understand the Earth and the processes that shape it. The early work of James Hutton, often considered the father of modern geology, in the mid 1700s was pivotal in establishing geology as a proper science. He conceptualized deep time and understood that ongoing processes such as sedimentation and erosion might help determine the history, and more specifically the age of the Earth. His theory of uniformitarianism, which states that the same natural processes and laws that operate on Earth now have operated the same in the past, was instrumental in advancing scientific thought at the time and is still integral to scientific understanding today. However, his work was not generally accepted until 1830 when Charles Lyell published *Principles of Geology* where he famously summarized the thoughts of James Hutton in saying, “the present is the key to the past.”

Geologists have also been driven by the exploration and exploitation of natural resources that have aided in the advancement of our modern society. The successful scientifically-based exploration of these natural resources, such as coal, was pioneered by the work of William Smith who recognized that strata in the Somerset coalfield were deposited in a predictable pattern, and an individual stratum could be identified by its unique fossil assemblage. This method of exploration is still used today to explore for a variety of natural resources. For example, petroleum geologists use these scientifically-based methods to populate their generally poorly resolved seismic models with data from outcrop-based ancient analogues.

Deep-water (>125 m) exploration began in the late 1970s, but grew rapidly in the mid 1980s (Pettingill and Weimer, 2002), with continued growth of 25% over the past decade (Manning, 2016). Today, as conventional onshore hydrocarbon reservoirs become depleted, deep-

marine sedimentary rocks represent the last vast frontier for conventional reservoirs. Most of this exploration and production is located along passive margins, downdip from producing Cenozoic deltaic systems (Pettingill and Weimer, 2002). In these areas, sand-rich turbidites can have excellent porosity and permeability (>30% and 1,000s mD respectively), thus forming primary reservoirs that currently account for *c.* 90% of deep-water reservoirs worldwide (Pettingill and Weimer, 2002). This high porosity is often maintained due to two characteristics: low geothermal gradient that hinders extensive diagenesis, and underconsolidation of sediment due to overpressure (Pettingill and Weimer, 2002). However, in spite of excellent resource potential, deep-marine turbidite systems also come with many uncertainties and substantial drilling costs. To reduce economic risk and improve exploration success, petroleum geologists have turned to the ancient deep-marine sedimentary record to find meaningful analogues to describe passive margin deep-marine turbidite systems.

The internal stratigraphic architecture and geometries of passive margin turbidite systems remain poorly understood due to a paucity of examples preserved in the ancient sedimentary rock record. These deposits remain an enigma because of their inaccessibility, unpredictable nature, rarity, and destructive forces of the very processes that build up the modern sedimentary record, namely turbidity currents (Piper and Normark, 2009), and, therefore, the difficulty in linking the present to the past. Thus, the majority of studies describing the behaviour of the flows is inferred from lab-based flume experiments and modelling studies as well as outcrop, core, and 3-D seismic reflection imaging (Piper and Normark, 2009).

Modern turbidite systems have been studied primarily using 2-D and 3-D seismic, geophysical well logs, and core. Recent improvements in seismic resolution, particularly with the development of 3-D seismic techniques, has greatly improved our understanding over the past few

decades of the deep-marine sedimentary system as it provides both cross-sectional and plan-view perspectives (*e.g.* Posamentier and Kolla, 2003). However, at industry depths (1,000s m) these surveys are still only capable of resolving large-scale architectural elements that generally are 15 m or more in thickness, and the individual stratal components that make up these elements remain below detection (Arnott, 2010).

The Windermere Consortium, which includes this thesis, is an industry-, government-, and academia-funded research initiative whose aim is to improve the understanding of deep-marine turbidite systems through field studies of an ancient, well-exposed, passive-margin turbidite system (Windermere turbidite system) in the Neoproterozoic Windermere Supergroup (*Fig. 1.1*), as well as experimental studies. Even though the Windermere Supergroup has no economic potential, it represents a useful analogue for the analysis of other deep-marine turbidite systems, both modern and ancient (Ross and Arnott, 2007). This study is located at Castle Creek in the Cariboo Mountains of the southern Canadian Cordillera (*Fig. 1.1*) where recently-deglaciated, vertically-dipping strata provide an unparalleled opportunity to study the deep-marine sedimentary record on scales ranging from millimetres to kilometres, both vertically and laterally. This allows lithology, stratigraphic relationships, and associations of architectural elements to be examined in detail.

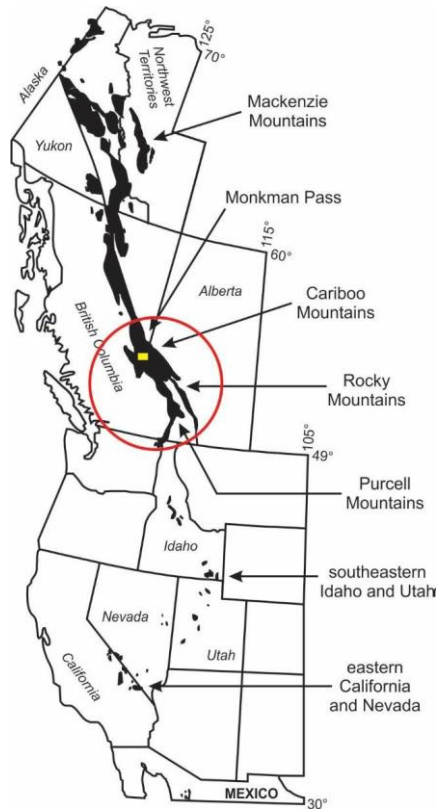


Figure 1.1: Location of exposed strata of the Windermere Supergroup (black) through western North America. The red circle denotes the approximate location of the deep-marine Windermere turbidite system in the southern Canadian Cordillera, and the yellow rectangle indicates the location of the Castle Creek study area (modified from Ross *et al.*, 1995).

1.2 Windermere Supergroup

Walker (1926) originally coined the term Windermere Series to describe primarily sedimentary rocks that crop out in the Windermere Valley of southeastern British Columbia. Later, the United States Geological Survey classified Neoproterozoic rocks in the American Cordillera as “Windermere Group” strata, whereas more completely preserved and better exposed Neoproterozoic rocks in the Canadian Cordillera are referred to as the “Windermere Supergroup” (Link *et al.*, 1993). In this thesis all Neoproterozoic rocks in western North America will be referred to as the Windermere Supergroup.

1.2.1 Regional Structure & Metamorphism

In the southern Canadian Cordillera Windermere Supergroup strata dominate the exposed stratigraphy in the western Main Ranges and Omineca Belt of the southern Canadian Cordillera (Ross *et al.*, 1989). The Omineca Belt is sandwiched between the ancient continental margin of western Laurentia (proto-North America) to the east after accretion that began in the mid-Jurassic, and a large composite of exotic allochthonous terranes to the west (*Fig. 1.2*) (Monger *et al.*, 1982). These strata have locally experienced several phases of structural deformation and metamorphism segregated between different structural panels resulting from the Cordilleran orogeny from 170-55 Ma (late Early Jurassic – Late Paleocene) (Price, 2000). This has imparted a variety of deformational styles in the Cordillera such as thrust faulting, strike-slip faulting and associated drag features, and significant tectonic discontinuities (Stewart, 1972). Nevertheless, within both the Foreland fold and thrust belt and Omineca Belt significant areas exist where strata have been subjected to only minor amounts of structural deformation and low-grade metamorphism (sub-greenschist to greenschist facies), and where reliable sedimentological data can be gathered, including the Jasper and Lake Louise areas, northern Cariboo Mountains (location of Castle Creek study area), and the eastern Purcell Mountains (*Fig. 1.2, 1.5*) (Smith *et al.*, 2011).

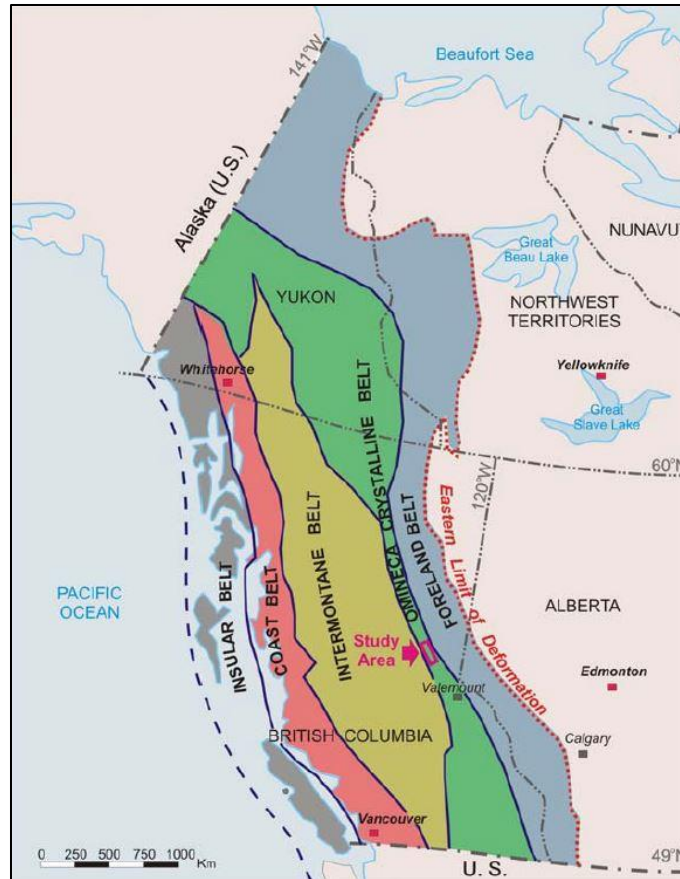


Figure 1.2: The Canadian Cordillera is subdivided into five physiographic regions, which from west to east are the Insular Belt, Coast Belt, Intermontane Belt, Omineca Belt, and Foreland Belt. The pink box denotes the approximate location of the Castle Creek study area (modified from Wheeler and McFeely, 1991; Gabrielse *et al.*, 1991).

The Castle Creek study area in the northern Cariboo Mountains (*Fig. 1.5B*) occurs on the overturned limb of a southwest-verging anticline part of the eastern portion of the Isaac Synclinorium within the Omineca Belt (*Fig. 1.3*) (Ross and Arnott, 2007). This area has undergone multiple phases of structural deformation and at least two episodes of low-grade metamorphism during the Mesozoic and early Cenozoic (Brown *et al.*, 1986; Murphy, 1987). Northeast-verging folds and associated bedding-parallel (locally inclined) cleavage formed during the first phase of deformation (Murphy, 1987). Metamorphism to biotite or lower grade (sub-greenschist to greenschist facies) is suggested to have been synchronous with this deformation (Murphy, 1987). This was followed by a second phase of deformation that produced map-scale southwest-verging structures forming a prominent axial planar crenulation cleavage (Murphy, 1987). The first two

structural events and first metamorphic episode were linked to the obduction of the Windermere turbidite system onto the western margin of proto-North America from 196-165 Ma (Jurassic) (Brown *et al.*, 1986; Murphy, 1987; Reid, 2003). A second episode of metamorphism, up to greenschist facies, overprinted the earlier fabrics (Murphy, 1987). Although polyphase structural deformation is evident on both macroscopic and microscopic scales, primary sedimentological attributes and larger scale stratal elements were little affected (Ross and Arnott, 2007). More recent uplift of the Premier Range in the southeastern Cariboo Mountains tilted the strata into its current northwest-plunging attitude (Murphy, 1987).

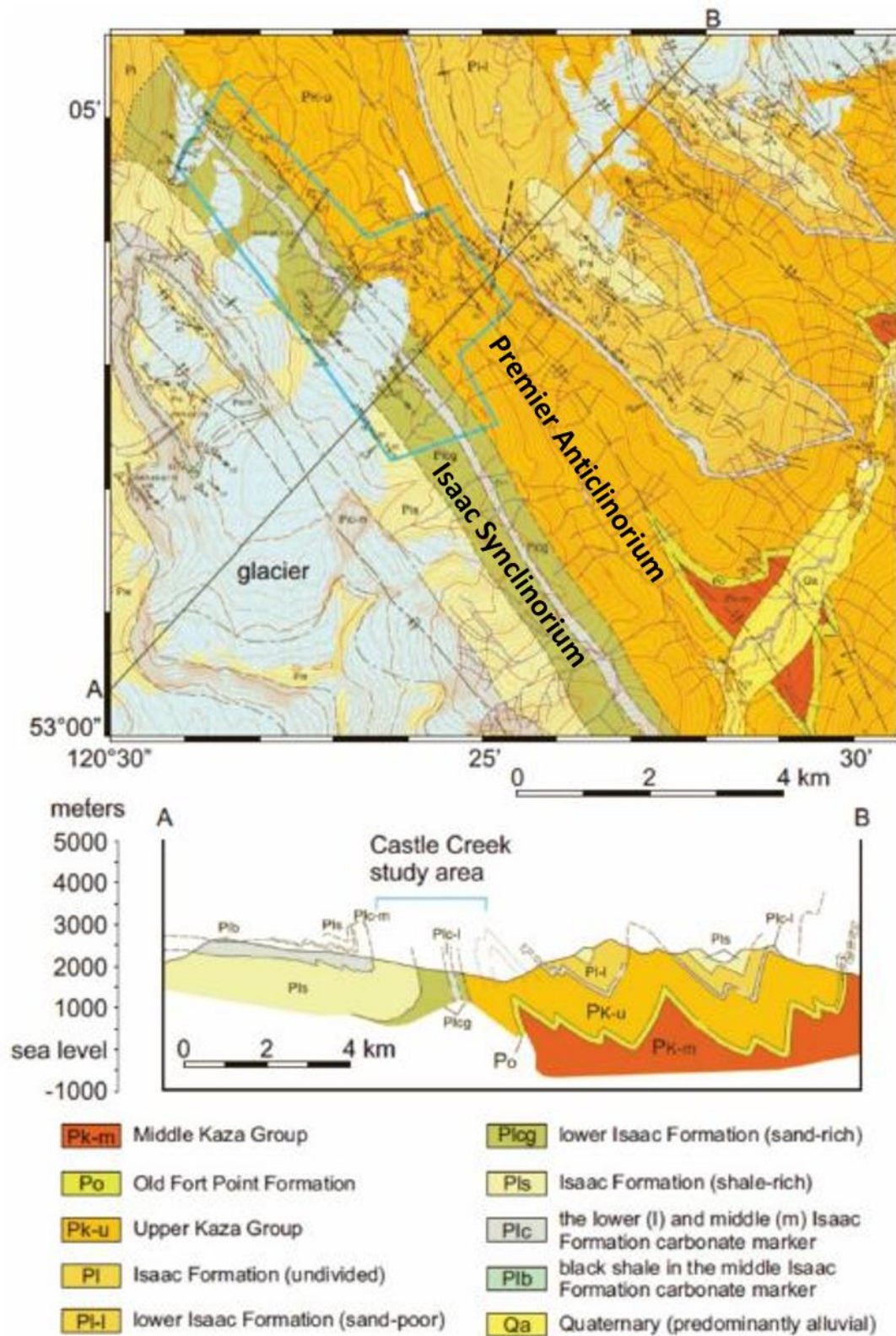


Figure 1.3: Regional geology of the Castle Creek study area (outlined in blue). Excerpt from 1:50,000 Eddy, British Columbia map sheet: Geological Survey of Canada, "A" Series Map 1967A, 2003 (modified from Ross and Ferguson, 2003a).

In the southern Canadian Cordillera, the Windermere Supergroup crops out over an area of about 35,000 km² (Ross *et al.*, 1995; Ross and Arnott, 2007). If the effects of contractional Mesozoic deformation are considered, and a conservative shortening factor of 30% applied (Brown *et al.*, 1986; McDonough and Simony, 1988; Price, 2000; Arnott and Ross, 2007), the exposed part of the deep-water turbidite system is suggested to have been c. 80,000-100,000 km², which would make it comparable in size to modern submarine fans like the Mississippi, Astoria, Magdalena or Monterey (*Fig. 1.4*) (Barnes and Normark, 1985), and also the largest known ancient turbidite system (Ross *et al.*, 1995; Ross and Arnott, 2007).

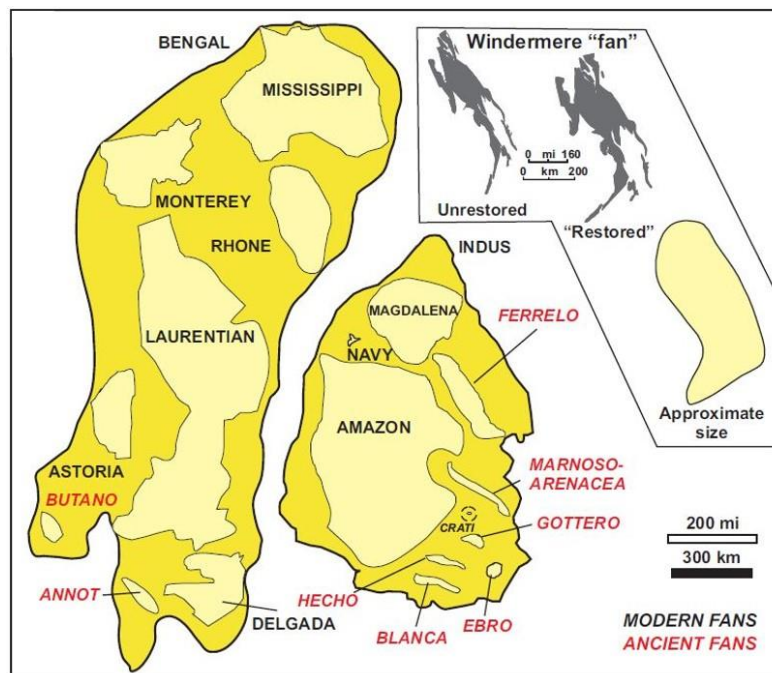


Figure 1.4: Size comparison of the restored Windermere turbidite system (Windermere “fan”) with modern (black text) and ancient (red text) turbidite systems (Ross, 2001 modified from Barnes and Normark, 1985).

1.2.2 Regional Distribution

The Windermere Supergroup refers to Neoproterozoic strata that form a long, semi-continuously exposed arcuate belt in western North America (*Fig. 1.1*) with correlatives from the Sonoran Desert in northwestern Mexico to the Yukon-Alaska border (Ross *et al.*, 1989; Ross,

1991). Outcrops in the United States and Mexico are more discontinuous and fragmented compared to the extensive, continuous belt in the southern Canadian Cordillera. Strata in the United States crop out in structurally expanded portions of the Basin and Range Province and the Sevier Belt (Christie-Blick and Levy, 1989; Ross and Arnott, 2007).

1.2.3 Regional Markers

During regional mapping of the Windermere turbidite system, recognition of up to three regionally extensive lithostratigraphic markers were identified in the Kaza and Cariboo groups of the Cariboo Mountains, Horsethief Group of the Selkirk and Purcell Mountains, and Miette Group of the Main Ranges (from Lake Louise and Jasper west to the Southern Rocky Mountain Trench; SRMT; *Fig. 1.5A*), and Jasper and Lake Louise areas (*Fig. 1.11*) (Ross *et al.*, 1995). These consist of a lower mixed siliciclastic and carbonate marker called the Old Fort Point Formation (OFP) (Arnott and Ross, 2007), which is exposed locally over the entire Canadian Cordillera (Smith *et al.*, 2011), and the upper two are carbonate markers that crop out in the Cariboo Mountains, termed the first and second Isaac carbonate (Ross *et al.*, 1989; Ross and Ferguson, 2003a). The recognition and use of these regional markers has substantially improved the correlation of Windermere Supergroup strata regionally and across structurally complex terrain comprising rocks of various metamorphic grade (Charlesworth *et al.*, 1967; Aitken, 1969; Ross and Murphy, 1988).

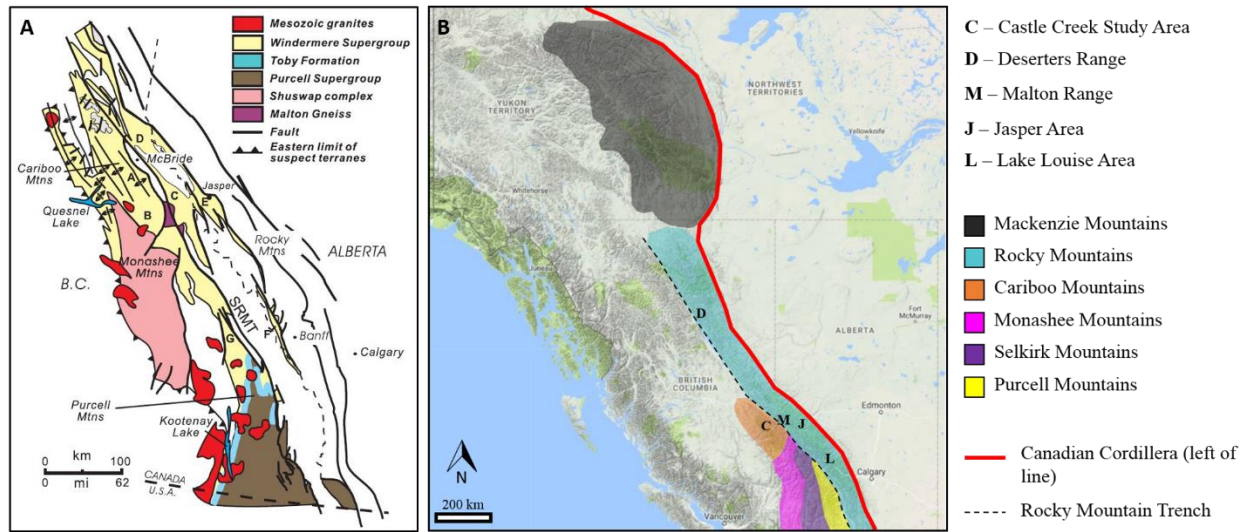


Figure 1.5: A) Geological map of the southern Canadian Cordillera showing the distribution of the Windermere Supergroup. Letters A-G show known outcrop locations of the Old Fort Point Formation. The Shuswap complex comprises high-grade (amphibolite facies) metamorphic rocks in the core of the Omineca Belt that are thought to correlate with the Windermere but are unsuitable for sedimentary analysis due to their elevated metamorphic grade (Ross and Arnott, 2007). B) Map of western Canada showing the location of different mountain ranges in the Canadian Cordillera and important localities associated with the Windermere Supergroup (modified from Google Maps, 2017).

The Old Fort Point Formation ranges from 50-450 m thick and is a geochemically and lithologically distinctive unit in the Windermere sedimentary pile in the southern Canadian Cordillera (Walcott, 1910; Weiner, 1966; Charlesworth *et al.*, 1967; Ross and Murphy, 1988; Ross *et al.*, 1989; Smith *et al.*, 2014). The Old Fort Point Formation conformably overlies and in turn is overlain by tan-colored coarse-grained deep-marine siliciclastic turbidite deposits (Smith *et al.*, 2011, 2014). Recently, Smith (2011) and Smith *et al.* (2014) subdivided the formation into three lithostratigraphic members, which stratigraphically upward are: 1) Temple Lake Member comprising varicolored siltstone-mudstone grading upwards into limestone-siltstone rhythmic couplets; 2) Geikie Siding Member containing organic-rich mudstone; and 3) the sharply bounded, heterolithic Whitehorn Mountain Member comprising conglomerate, breccia, sandstone, siltstone-mudstone, limestone, and diamictite.

The distinctive nature of the Old Fort Point Formation has been interpreted to correlate with an abrupt basin-wide eustatic rise associated with the end of the Marinoan glaciation (Kendall

et al., 2004; McMechan, 2015). This shut off much of the siliciclastic input into the Windermere turbidite system and resulted in an upward-fining and -thinning succession (Ross and Murphy, 1988) dominated by chemogenic sediment. This is made evident by limestone turbidites shed from the highstand carbonate shelf, and the organic-rich black shale indicating increased nutrient supply (Fig. 1.6) (Ross and Murphy, 1988; Smith *et al.*, 2014), and suggests that the Old Fort Point Formation represents a transgressive condensed section (Ross *et al.*, 1995; Ross, 2000).

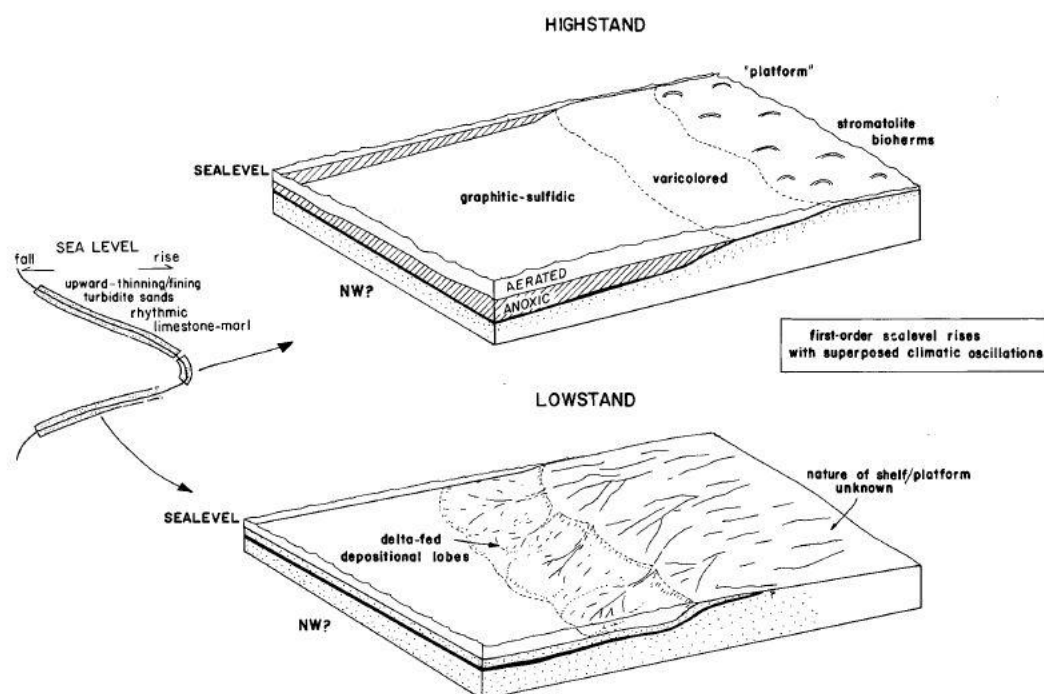


Figure 1.6: Depositional model for the Old Fort Point Formation (highstand) and the overlying calcareous sandstone (lowstand) (Ross *et al.*, 1989). The facies transition is evident in the lowstand conditions shown above from dysaerobic to aerobic varicolored shales in the east to anaerobic, sulfidic black shales in the west. The heterolithic nature to the upper Whitehorn Mountain Member is interpreted to represent an abrupt relative sea-level fall (Smith *et al.*, 2014).

The two other regional markers, informally termed the first and second Isaac carbonate marker units, are deep-marine calciturbidite horizons composed of sediment sourced from the adjacent shallow-marine carbonate platform (Ross *et al.*, 1989; Ross, 1991; Ross and Ferguson, 2003a). Both show dramatic and abrupt changes in thickness laterally, ranging from 10 m to greater than 200 m, as well as lithology and grain size comprising calcilutite, thin-bedded carbonate

turbidites, ooid grainstone, and decametre shelf carbonate slide blocks (Arnott and Ross, 2007). These regional carbonate markers have been interpreted to be deposited during the late transgressive to early falling stage systems tract that caused the continental shelf to flood and carbonate to replace siliciclastic sediment input (B. Arnott, personal communication, 2017).

1.2.4 Sediment Provenance & Paleoflow

The deep-marine Windermere basin in the southern Canadian Cordillera was part of a longitudinal dispersal system that transported sediment from southeast to northwest based on regional facies patterns, sediment provenance, and paleocurrent data (Ressor, 1957; Mountjoy and Aitken, 1963; Seeland, 1968; Arnott and Hein, 1986; Ross and Parrish, 1991a, b; Ross, 2001). Sediment provenance studies were conducted using U-Pb detrital zircon geochronology, suggesting that the primary source of sediment was Archean and Paleoproterozoic rocks to the east-southeast (Ross and Parrish, 1991a, b; Ross, 2001). The absence of zircon grains between 1.9-2.6 Ga rules out a source from the Alberta basement, thus indicating that sediment was derived almost entirely from the southern Canadian shield and its correlatives in the northwestern United States (Ross and Arnott, 2007). The homogeneity of the framework grains (85% quartz, 15% feldspar) and uniformity of detrital zircons imply that the depositional system was regionally integrated with a common source (Ross and Arnott, 2007). This is further supported by consistent paleocurrent direction from ripple and dune cross-stratification, cobble imbrication, and less commonly flutes throughout the Windermere turbidite system in the Canadian Cordillera, which show general transport toward the west-northwest (Arnott and Hein, 1986; Ross and Arnott, 2007), although local variations are noted (Charlesworth *et al.*, 1967; Arnott and Hein, 1986; Ross *et al.*, 1989).

1.2.5 Geochronology

Age constraints within the Neoproterozoic Windermere Supergroup remain poor because of the absence of biostratigraphic control (Ross *et al.*, 1989; Terlaky *et al.*, 2016), the predominantly siliciclastic lithology that lacks suitable material for conventional radiometric dating (Ross *et al.*, 1989; Smith *et al.*, 2011), and a lack of zircons in exposed volcanic rocks (Lund *et al.*, 2003). However, highly precise radiometric ages on rocks that unconformably underlie and overlie the Windermere constrain the maximum and minimum ages of deposition of the turbidite system from c. 740-728 Ma to c. 569 Ma, respectively (*Fig. 1.7*) (Evenchick *et al.*, 1984; Parrish and Scammell, 1988; McDonough and Parrish, 1991; Colpron *et al.*, 2002). Maximum depositional ages were obtained from U-Pb zircon dates in granitic rocks beneath the Windermere Supergroup in the Deserters Range of northeastern British Columbia (*Fig. 1.5B*), which yielded an age of $728 \pm 8/-7$ Ma (Evenchick *et al.*, 1984). This is further supported by U-Pb zircon dates from basement orthogneisses in southeastern British Columbia in the Monashee Mountains dated at 740 ± 36 Ma (Parrish and Scammell, 1988), and in the Malton Range yielding $736 \pm 23/-17$ Ma (McDonough and Parrish, 1991). Within the Windermere Supergroup, rift-related Irene Formation correlatives in northern British Columbia, Idaho, and Utah have been dated using U-Pb zircon geochronology that yielded ages ranging from 709 to 667 Ma (Ferri *et al.*, 1999; Lund *et al.*, 2003; Fanning and Link, 2004; Balgord *et al.*, 2013). The organic-rich mudstone in the middle Geikie Siding Member of the Old Fort Point Formation in the Jasper area was dated using Re-Os at 607.8 ± 4.7 Ma (Kendall *et al.*, 2004). The minimum age of deposition is constrained by a U-Pb zircon date of 569.6 ± 5.3 Ma from rift-related felsic volcanic rocks at the base of the Hamill Group in southeastern British Columbia, which post-date termination of the Windermere turbidite system (Colpron *et al.*, 2002).

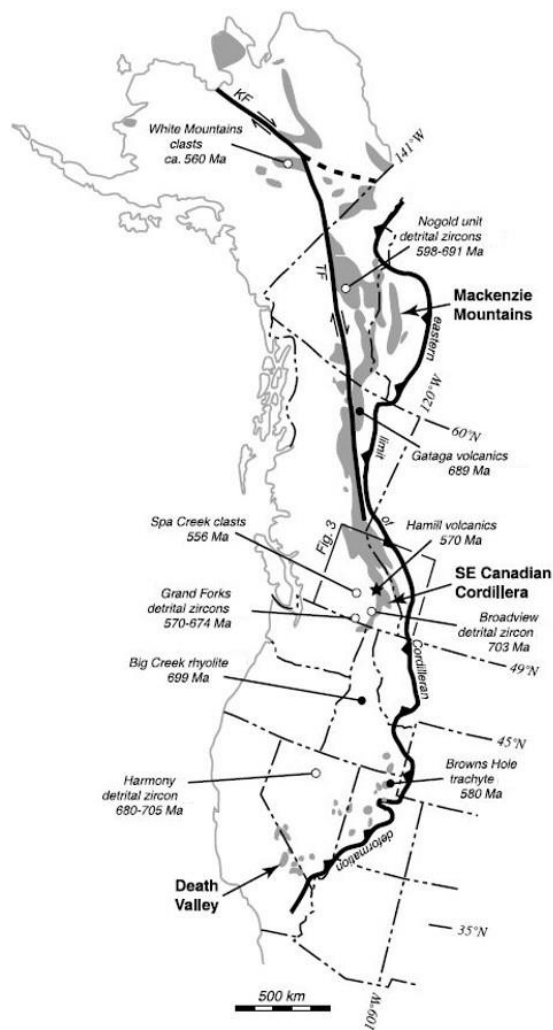


Figure 1.7: Map showing exposed rocks of the Windermere Supergroup (grey) in western Canada and the United States. Neoproterozoic geochronological constraints are indicated by closed circles (igneous rock ages) or open circles (detrital zircon ages) (modified from Colpron *et al.*, 2002).

1.2.6 Tectonic Model

The Windermere Supergroup is interpreted to have experienced two discrete episodes of deposition (Stewart, 1972; Ross, 1991). The first records an episode of rifting related to the disassembly of the supercontinent Rodinia, whereas the second episode represents post-rift thermal subsidence. While there is general consensus in the scientific community regarding these two tectonic episodes, the exact tectonic model of the Windermere basin in the southern Canadian Cordillera, and its temporal evolution, remains a source of debate. Some geologists have

interpreted that post-rift sedimentation occurred along a passive margin after the separation of Laurentia and Australia forming the Panthalassic Ocean (proto-Pacific Ocean) during the Neoproterozoic (Gabrielse, 1972; Stewart, 1972; Burchfiel and Davis, 1975; Stewart and Suczek, 1977; Monger and Price, 1979; Eisbacher, 1981; Arnott and Hein, 1986; Ross *et al.*, 1989; Ross, 1991; Ross *et al.*, 1995; Dalrymple and Narbonne, 1996) with a second rifting event responsible for uplift and formation of the Cambro-Ordovician passive margin that resulted in substantial sub-Cambrian erosion (*Fig. 1.8*) (Ross, 1988). Others argue that initially the Windermere margin was a 2,500 km intracratonic rift system that opened to an oceanic basin to the north, suggesting that continental separation and formation of the Panthalassic Ocean and establishment of a passive margin along the western margin of Laurentia did not occur until the Neoproterozoic-early Cambrian (Colpron *et al.*, 2002). The uncertainty and controversy in the southern Canadian Cordillera is due to a multitude of factors, including two intervals of rift-related igneous intrusions in the southern Canadian Cordillera (Bond and Kominz, 1984; Ross 1991; Colpron *et al.*, 2002), diachronous dates for western Laurentian rifting (Lund *et al.*, 2003), and Mesozoic tectonic deformation that hinders the identification of the Windermere margin.

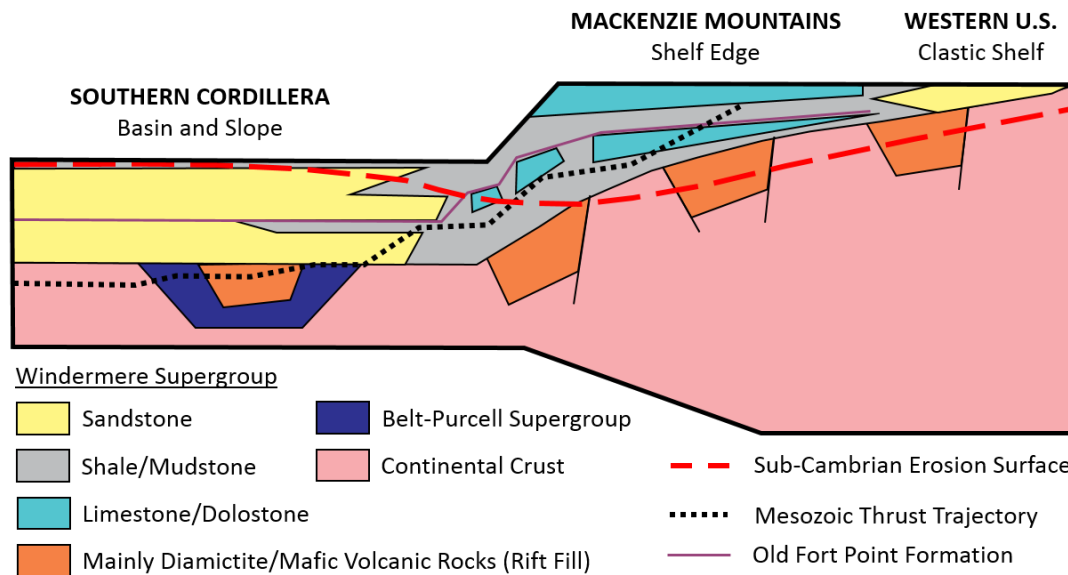


Figure 1.8: Schematic representation of the Windermere margin (western North America) during the Neoproterozoic. The black dotted line indicates the truncation of strata in the footwall due to eastward shallowing thrust faults. The red dashed line represents preservation below the sub-Cambrian unconformity that removed the shallow-marine Windermere strata in the southeastern Cordillera. Orange represents syn-rift (rift) deposits, whereas the overlying deposits are post-rift (modified from Ross and Arnott, 2007).

1.2.7 Paleoenvironmental Interpretation

As discussed above, the Windermere Supergroup has been interpreted as a complete rift-drift passive margin sequence formed during the disassembly of Rodinia (Ross, 1991). The basin experienced diminishing rates of thermal subsidence as the lithosphere cooled in the post-rift stage causing the gradual progradation of the continental margin westward into the basin, which is preserved in the up to about nine-kilometre-thick succession of basin floor overlain by slope strata (Ross, 1991; Ross *et al.*, 1995; Ross and Arnott, 2007). The lower rift succession comprises laterally discontinuous mafic volcanic rocks and glaciogenic diamictites, whereas the upper succession consists of laterally continuous siliciclastics with subordinate carbonate rocks associated with post-rift thermal subsidence (*Fig. 1.8*) (Arnott and Ross, 2007). During this second stage the basin contained an extensive, elongate submarine turbidite system that currently is exposed throughout much of the southern Canadian Cordillera (Ross *et al.*, 1989).

The paleolatitude of the western Laurentian coast has been inferred from paleomagnetic data obtained from correlative Windermere Supergroup strata in the Mackenzie Mountains and suggests deposition in a near-equatorial paleogeographic position (Park, 1997). Based on paleocurrent data and detrital zircon analysis the basin is interpreted to have been located off the coast of Laurentia with turbidity currents flowing down the continental slope and out onto the basin floor (see discussion in 1.2.4) (Ross and Bowring, 1990; Ross and Parrish, 1991a, b). This is supported by evidence of submarine canyon fills in the Lake Louise area that were incised by basinward-flowing turbidity currents (L in *Fig. 1.9*) (Arnott and Hein, 1986). Although fragmentary due to the effects of sub-Cambrian erosion, shallow-marine upper slope and shelf facies are observed in the Mackenzie Mountains (northern Canadian Cordillera) and the upper part of the succession in the southern Canadian Cordillera (Ross *et al.*, 1995), while shallow shelf and terrestrial siliciclastic and subordinate carbonate rocks crop out in Mexico and western United States (Link *et al.*, 1993). Subordinate deep-marine carbonates in the southern Canadian Cordillera are suggested to have been sourced from a shallow-marine shelf (*Fig.1.8*) to the east, which may correlate to incompletely preserved strata in the Mackenzie Mountains in the northern Canadian Cordillera (*Fig. 1.8*) (Ross, 1991).

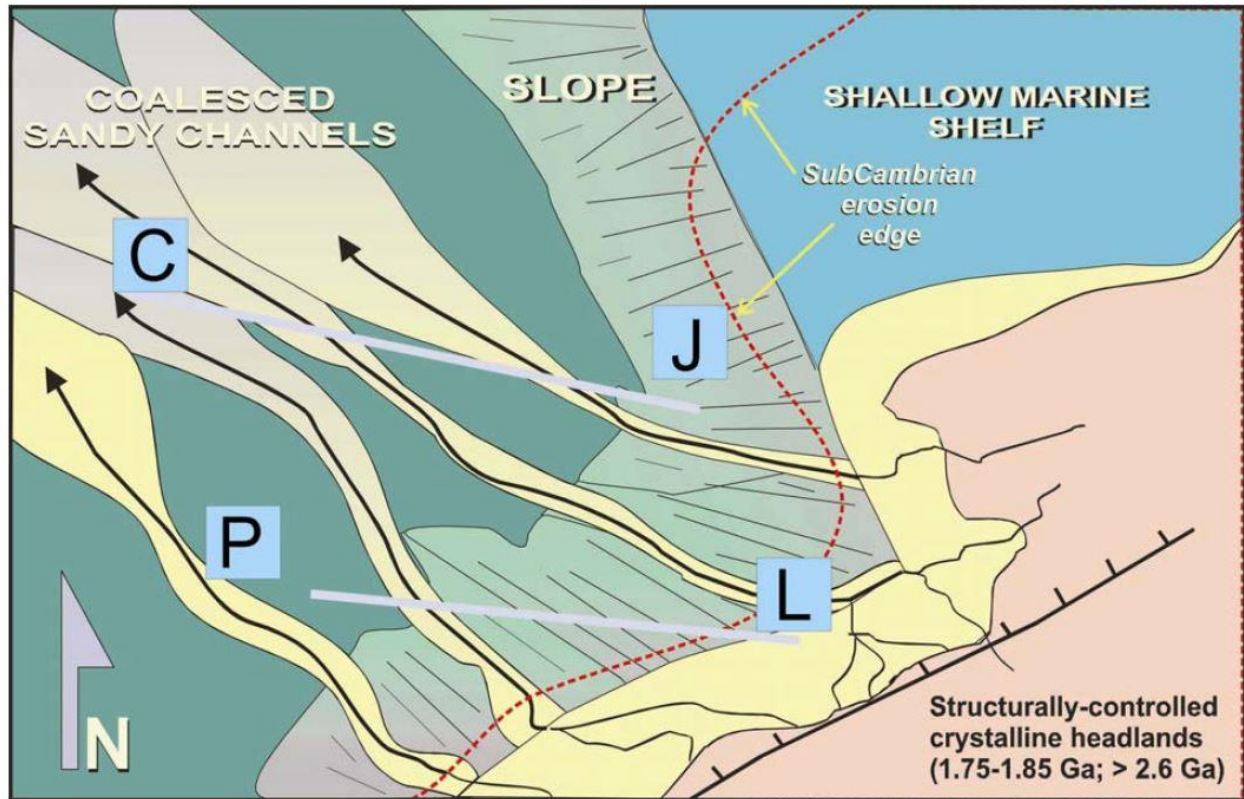


Figure 1.9: Schematic paleogeographic reconstruction of the Windermere turbidite system in the southern Canadian Cordillera. Paleoflow was toward the northwest (present day coordinates) from the upper slope in Lake Louise (L) to the lower slope and base-of-slope in Jasper (J) and Purcell (P) regions, respectively, and then onto the basin floor in the Castle Creek study area (C) (Ross, 2000).

1.2.8 Castle Creek Study Area

The Neoproterozoic Windermere Supergroup in the northern Cariboo Mountains of east-central British Columbia is a world-class example of an ancient passive margin turbidite system comprising a complete succession of deposits from basin floor to slope turbidites and shelf carbonates (Ross, 1991). At the Castle Creek study area (*Fig. 1.10, 1.12*) these strata are exposed in a recently-deglaciated, vegetation free, vertically-dipping section that extends for 8 km parallel to bedding and 2.5 km perpendicular to bedding. However, glacial overburden poses local challenges to lateral stratigraphic correlation. Nevertheless, this study area provides an unparalleled opportunity to study the deep-marine sedimentary record on scales ranging from millimetres (sub-seismic-scale) to kilometres (seismic-scale).

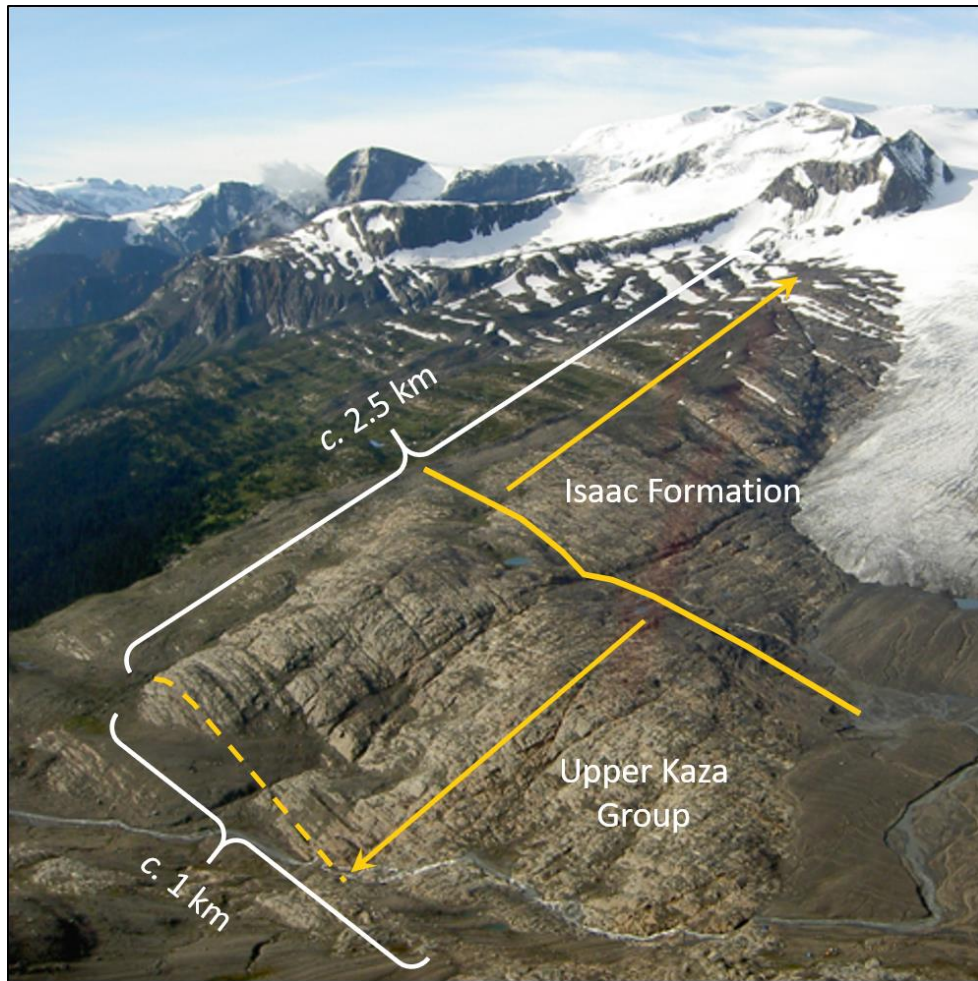


Figure 1.10: The Castle Creek study area located in the Cariboo Mountains of east-central British Columbia. Here, a 2.5-km-thick succession of recently deglaciated, vertically-dipping strata is exposed consisting of proximal basin floor deposits of the Upper Kaza Group (c. 0.8 km thick) conformably overlain by channel-levee complexes of the Isaac Formation (c. 1.7 km thick).

1.2.9 Stratigraphy

Post-rift strata of the Windermere Supergroup are best exposed and most thickly preserved in the Cariboo Mountains, British Columbia where two stratigraphic assemblages, termed the Kaza and Cariboo groups, are recognized, and further subdivided into the Lower, Middle and Upper Kaza Group, and the Isaac, Cunningham and Yankee Belle formations of the Cariboo Group (*Fig. 1.11, 1.12A, B*). Correlation of these strata across western North America is difficult due to the lack of biostratigraphic and geochronological control (Ross *et al.*, 1989), and structural

deformation and metamorphism related to the Mesozoic Cordilleran Orogeny (Smith *et al.*, 2011). This has led to a variety of names assigned to rocks of similar character in different localities (Fig. 1.11) (Gussow, 1957; Charlesworth *et al.*, 1967; Aitken, 1969; Brown *et al.*, 1978; Mansy and Gabrielse, 1978; Eisbacher, 1981; Carey and Simony, 1985; Pell and Simony, 1987; Ross and Murphy, 1988; McDonough, 1989; Kubli, 1990; Ross *et al.*, 1995; Ross and Ferguson, 2003a).

	Rocky Mountains				Cariboo Mountains		Rocky Mountains			Central Purcell Mountains	Southern Selkirk Mountains
	Deserters Range	Mt. Vreeland	Cushing Creek				Yellowhead Pass - Jasper	Lake Louise	Kicking Horse Pass		
	Gog Gp	Gog Group	Gog Group		Cariboo Group		Gog Group	Gog Group	Gog Group		
Lower Cambrian	top of exposure	Hota	Hota		Dome Creek		Hota	Peyto	Peyto	Lardeau Group	Laib
	Mural	Mahto	Mahto		Mural		Mahto		St. Piran	Badshot	
		Mural	Mural		Midas		Mural	Gog Group undivided		Mohican	
		McNaughton	McNaughton		Yanks Peak		McNaughton		Lake Louise		Reno
					Yankee Belle			Fort Mountain		Hamill Group	Quartzite Range
					Cunningham						
							Byng				
							East Twin				
							McKale	Hector	Hector		
							Old Fort Point	Corral Creek	Corral Creek		
Upper Proterozoic							McKale	base not exposed	base not exposed		
							unnamed				
Middle Prot.											

Figure 1.11: Stratigraphic nomenclature of Windermere Supergroup (upper Proterozoic/Neoproterozoic) and Lower Cambrian strata in the southern Canadian Cordillera. The red rectangle indicates strata of this study. Data sources include: Evenchick (1988); Young (1979); Carey and Simony (1985); Bond *et al.* (1985); Teitz and Mountjoy (1985); Steiner (1962); Arnott (1984); Magwood (1988); Palonen (1976); Devlin (1989); Devlin and Bond (1988); Hein (*unpublished field notes*); Wolberg (1986); Aitken (1969); Aitken (*unpublished field notes*); Lindsey and Gaylord (1992); Höy (1980); McMechan (*unpublished field notes*); Charlesworth *et al.* (1967); Rice (1941); Little (1960); Reesor (1958, 1973); Kubli (1990); McMechan (1991) (modified from Hein and McMechan, 1994).

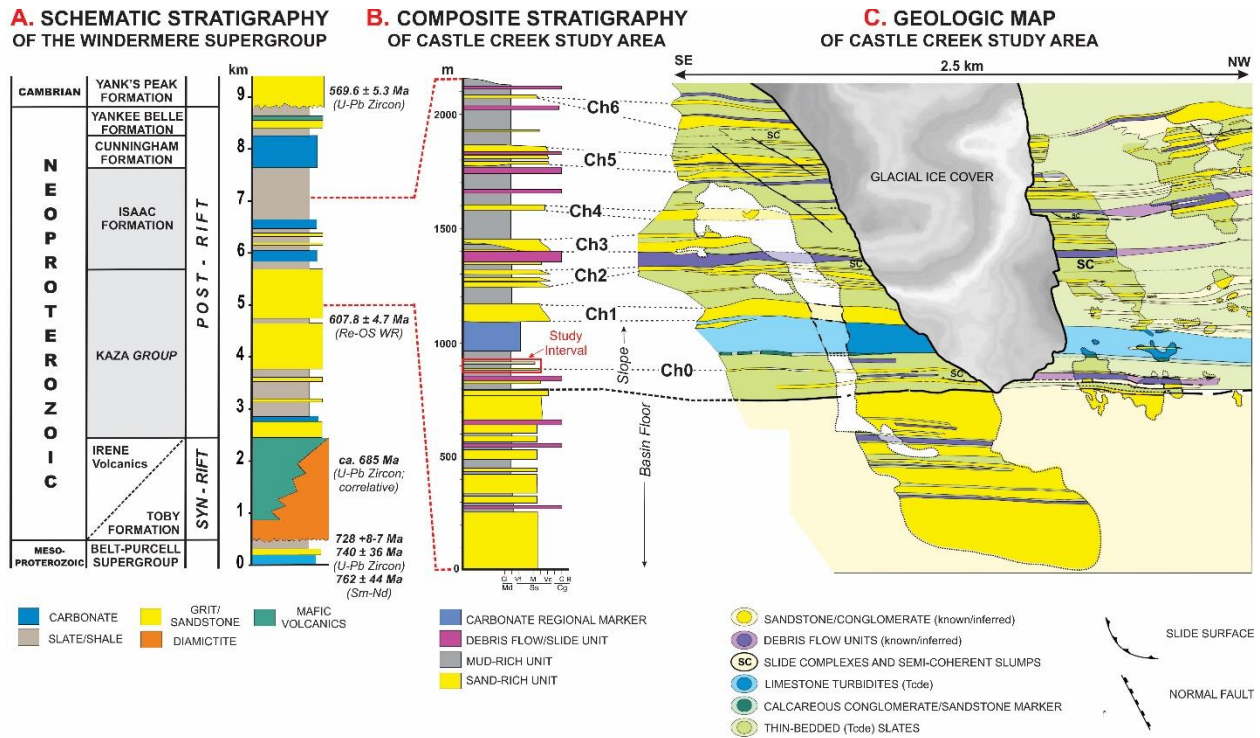


Figure 1.12: Stratigraphy of the Windermere Supergroup from regional to outcrop scale. A) Correlated schematic stratigraphy of the Windermere Supergroup. Age control is provided on the right of the log (see discussion in 1.2.5). B) Composite stratigraphy of the Castle Creek study area in the Cariboo Mountains with the study interval for this thesis outlined in the red rectangle. C) Geologic map of the stratigraphy in the Castle Creek study area (modified from Lee, 2016).

1.2.9.1 Toby Formation

The Toby Formation (*Fig. 1.12A, B*) was named by Walker (1926) for strata exposed along Toby Creek, west of Windermere, British Columbia. The basal Toby Formation is associated with deposition during initial rifting, and although regionally extensive (Stewart, 1972), is discontinuously exposed in southeastern British Columbia to northeastern Washington and northwestern Idaho, and has been correlated to the Shedroof and Huckleberry conglomerates, respectively (Aalto, 1971; Ross and Arnott, 2007), in addition to correlative strata in California and Nevada (Stewart, 1972). In the southern Canadian Cordillera, the Toby Formation unconformably overlies Mesoproterozoic metasedimentary rocks of the Belt-Purcell Supergroup in southeastern British Columbia and the Muskwa assemblage in northeastern British Columbia (Reesor, 1957; Ross *et al.*, 1989; Evans *et al.*, 2000) or the crystalline basement (Crowley, 1999).

The Toby Formation consists mainly of thick-bedded or lenses of diamictite with subordinate interbeds of conglomerate, sandstone, and mudstone with dispersed megaclasts (Aalto, 1971). Most notably, the thickness of the formation varies considerably from a few to 2,500 m (Aalto, 1971). This variation has been suggested to be the result of syn-sedimentary fault-controlled deposition during the active rifting phase (Aalto 1971; Root, 1987). The Toby Formation has been interpreted as a glaciomarine diamictite (*sensu* Eyles, 1985) due to its texture (poorly sorted), composition (dispersed dolomite and quartzite pebble-boulder clasts in sand-mud matrix), lack of regional variation, and stratigraphic associations (Aalto, 1971; Eisbacher, 1981; Root, 1987; Ross *et al.*, 1989) and is thought to be associated with the Neoproterozoic Sturtian glaciation (Ross *et al.*, 1995). A localized, discontinuous carbonate unit at the top of the Toby Formation may be analogous to a cap carbonate that often postdate global glaciations (Hoffman *et al.*, 1998).

1.2.9.2 Irene Formation

The Irene Formation (*Fig. 1.12A, B*) is also associated with active rifting in the Windermere Supergroup (Ross and Arnott, 2007). This formation is restricted to two areas west of the Purcell Mountains (Ross *et al.*, 1989), but may correlate with the Leola and Huckleberry volcanics in northern Washington state (Ross and Arnott, 2007). The Irene Formation conformably overlies or intertongues with the Toby Formation (Stewart, 1972; Ross *et al.*, 1989).

The Irene Formation comprises locally thickly developed and altered and/or metamorphosed tholeiitic basalt with local amygdules and/or pillow structures, breccias, and volcaniclastic deposits (Park and Cannon, 1943; Little, 1960; Aalto, 1971; Leclair, 1982; Root, 1987). The thickness of the formation varies and is thickest near the Canada-United States border where it is up to 1,600 m, thinning from there to the north and south (Little, 1960; Leclair, 1982;

Ross *et al.*, 1989; Ross *et al.*, 1995). Some of these rocks have been identified as subaqueous pillow lavas, which commonly are associated with continental rifting and crustal thinning (*sensu* Dewey, 1969).

1.2.9.3 Kaza Group

The Kaza Group is the lowest part of the post-rift phase of the Windermere Supergroup (*Fig. 1.12A, B*) and is of the order of 2-3 km thick (Ross and Arnott, 2007). As mentioned above, strata of the Kaza Group in the Cariboo Mountains are subdivided into the Lower, Middle, and Upper parts of the Kaza group (Campbell *et al.*, 1973), with the Middle and Upper Kaza Group being separated by the lithologically distinctive Old Fort Point Formation (Smith *et al.*, 2011). Rocks of the Kaza Group correlate with the Horsethief Creek Group in the northern Purcell Mountains, which overlies rift-related diamictite of the Toby Formation or mafic volcanic rocks of the Irene Formation (Aalto, 1971). Although the contact between rift and post-rift rocks is not exposed in the Cariboo Mountains, the Kaza Group is presumed to be underlain by the regionally extensive Toby Formation and/or subordinate Irene Formation. Grain size and bed thickness generally increase from the Lower to Middle Kaza and then changes little into the Upper Kaza (Ross and Murphy, 1988).

At Castle Creek, the Upper Kaza Group consists mostly of sand-rich, sheet-like to lobate deposits that change little in thickness or composition along their extent (Ross and Arnott, 2007). Strata are sand rich and made up of massive or normally graded Bouma sequence T_a beds interstratified with fine-grained, thin-bedded Bouma sequence T_{bc} turbidites (Bouma, 1962; Ross *et al.*, 1995). Sandstone:mudstone ratio is generally high at c. 75:25 (Meyer, 2004). Sandstone beds range from 10-900 cm thick (Ross *et al.*, 1995) that stack to form sheet-like stratal elements

that are up to 40 m thick and interpreted to be basin floor lobes and splays interstratified with uncommon debrites (Ross and Arnott, 2007; Terlaky *et al.*, 2016).

1.2.9.4 Isaac Formation

The Isaac Formation (Cariboo Group) (*Fig. 1.12A, B*) conformably overlies the Upper Kaza Group (Ross and Arnott, 2007) and is composed mostly of mudstone (75%) with subordinate (c. 25%) discontinuous lenses of sandstone and minor conglomerate, in addition to common mass-transport deposits (slide, slump, debrites) (Ross and Arnott, 2007). At Castle Creek, the Isaac Formation is more than 1,600 m thick and comprises six several-decametre-thick slope channel complexes bounded laterally by fine-grained, thinner-bedded deposits (Ross and Arnott, 2007). In addition, several mass-transport complexes (*Fig. 1.12B, C*) (Navarro *et al.*, 2007a; Ross and Arnott, 2007) and two, more than 100 m thick, deep-marine carbonate deposits interbedded with sandstone and mudstone, informally termed the first and second Isaac carbonates, are observed (Ross *et al.*, 1995). Clasts within particularly thick debrites contain oolite and stromatolite fragments, suggesting sourcing from a shallow-marine carbonate platform. Moreover, the thickly-developed Isaac carbonates indicate transgressive to highstand conditions, whereas most of the siliciclastic-filled channel complexes represent conditions approaching and including lowstand (Ross *et al.*, 1989; Ross *et al.*, 1995; Ross and Arnott, 2007).

1.2.9.5 Cunningham & Yankee Belle Formations

The Cunningham and overlying Yankee Belle formations (*Fig. 1.12A, B*) are part of the post-rift phase of the Windermere Supergroup (Ross and Arnott, 2007), and form the upper part of the Cariboo Group (*Fig. 1.11*). The Cunningham Formation conformably overlies the Isaac Formation, and in turn is conformably overlain by the Yankee Belle Formation (Ross and Murphy,

1988). Post-depositional uplift and significant regional sub-Cambrian erosion has removed much of the upper part of the Windermere section in the eastern part of the southern Canadian Cordillera (see sub-Cambrian erosion edge in *Fig. 1.9*) (Devlin and Bond, 1988; Ross and Murphy, 1988). Accordingly, there is a significant unconformity between the Windermere Supergroup and the overlying Lower Cambrian Yank's Peak Formation in the Cariboo Mountains or equivalents in the lower Gog Group of the northern and southern Rocky Mountains, and the Hamill Group in the Purcell Mountains (*Fig. 1.11*) (Hein and McMechan, 1994). In the Cariboo Mountains the Cunningham Formation ranges up to 600 m thick and is composed mostly of oolitic-intraclastic limestone, while the Yankee Belle Formation is composed of an up to 900 m-thick mixed siliciclastic-carbonate succession (Ross *et al.*, 1995). The Cunningham and Yankee Belle formations have been interpreted, respectively, to be upper slope and shelf limestone deposits associated with high-energy shallow-marine platform carbonates in the upper Windermere Supergroup (Campbell *et al.*, 1973; Ross and Murphy, 1988; Ross *et al.*, 1995). More specifically, the Cunningham Formation has been suggested to record the upward transition from a ramp to rimmed platform (Ross *et al.*, 1995).

1.3 Methodology

Fieldwork was conducted at Castle Creek over two summers for 6-7 weeks each (July-August 2016 and 2017). During the first summer emphasis was placed on measuring stratigraphic columns and sampling. During the second summer, these stratigraphic logs were revised based on facies determined from samples collected in the first summer field season, mapping and bed-by-bed correlation in outcrop, and samples were collected.

The study area (*Fig. 1.12*) crops out for over 450 m laterally and comprises two general units in slope deposits of the Isaac Formation: a lower, highly-amalgamated sandstone unit, and

an upper, heterolithic unit composed of sandstone interstratified with siltstone or mudstone. A total of four stratigraphic logs were measured in bed-by-bed (*i.e.* centimetre-scale) detail. Here, logs ranging from 11.4-17.4 m long were measured (*App.* 2). The location of each stratigraphic log was determined after the section was walked out and placed where important lithological changes occurred, or where outcrop conditions were best. In each log, bed thickness, lithology, grain size, basal contact morphology, colour, primary and secondary sedimentary structures, and mud clast presence and features (size, shape, abundance, and orientation) were noted and described. Grain size was determined using a standard grain size comparator and a 10x hand lens, and all thickness measurements were made with a standard tape measure or grain size comparator. Photographs were taken bed-by-bed with a Canon Rebel T3i camera, whereas videos and pictures of larger-scale features were captured with a DJI Phantom 3 Professional drone. Agisoft PhotoScan was used to stitch images together to create panoramas of the section. Excellent lateral continuity of the outcrop due to limited overburden allowed individual beds to be walked out and correlated for 100s of metres with a high degree of confidence. In areas of surface cover, debris was removed by hand to expose certain marker beds that ensured the accuracy of the lateral bed-by-bed correlations.

71 hand samples from the study area were collected and petrographically analysed using an Olympus BX41 transmitted light microscope. The majority of thin sections are oversized (38 x 76 mm/1.5 x 3 in) in order to better analyze sedimentary structures. In this thesis thin sections are denoted as being standard (25 x 46 mm/1 x 1.8 in) or oversized. A selection of 23 thin sections from all facies were stained with a 1:1 solution of Alizarin Red S and potassium ferricyanide, in order to differentiate calcite from dolomite, and to identify ferroan calcite or dolomite (Dickson, 1966). Thin sections were analyzed for grain size (minimum, maximum, and average), sorting, grain contacts, mineralogy, sedimentary structures, and fabric using qualitative description

techniques of estimation. 41 of these thin sections were further point counted using an Olympus mechanical rotating stage where a single vertical transect through each thin section measured 30 points every 0.5 mm to capture trends in grain size and mineralogy.

Since all rocks in the study area have undergone low-grade metamorphism, a variety of metamorphic microstructures are observed in all thin sections. The most important changes caused by metamorphism are: primary clay minerals have been recrystallized to chlorite and micas, and bulging recrystallization and sub-grain rotation occur along some grain boundaries. These metamorphic changes were considered when analyzing thin sections. Small, intergrown crystals of chlorite and mica, as well as silt-sized or larger crystals were counted as matrix in sandstone and original primary clay in siltstone, mudstone or claystone. Matrix was differentiated based on mineralogy when point counting and then grouped together for statistical purposes in further analysis. Scanning electron microscopy (SEM) using a JEOL JSM-6610LV was conducted on six samples to better characterize grain size and micro-sedimentary structures as well as validate mineralogy using energy-dispersive X-ray spectroscopy in fine-grained rocks, especially facies 3 and facies 4. Additionally, in a particularly well-exposed part of the heterolithic unit, a 12.5 m-long section (perpendicular to bedding) was continuously cut with a hand-held rock saw to better resolve millimetre- and centimetre-scale sedimentological features in these fine-grained strata.

1.4 Previous Work

Field mapping in the McBride map sheet (1:250 000 scale), which includes the Castle Creek study area, in addition to comprehensive description of regional stratigraphy, deformation, and metamorphism was first conducted by Campbell *et al.* (1973). This study also included regional correlation of the Proterozoic and Lower Paleozoic rocks with those in the adjacent Rocky Mountains. Later, part of this area was remapped at 1:50 000 scale (*Fig. 1.3*) by Ferguson and

Ross (2003) and Ross and Ferguson (2003a, b). In addition, numerous research reports have been published focussing primarily on regional and structural geology (*e.g.* Murphy and Rees, 1983; Murphy, 1987; Ross and Murphy, 1988; Ross 1991; Gabrielse and Campbell, 1991; Hein and McMechan, 1994; Ross *et al.*, 1995; Ross, 2000; Reid *et al.*, 2002; Reid, 2003). With the inception of The Windermere Consortium, an industry-, government-, and academia-funded research initiative based at the University of Ottawa, the detailed sedimentological, stratigraphic, architectural, and geochemical attributes of the Windermere deep-marine strata became a focus of investigation (Ross and Arnott, 2007). This includes research focusing on a wide range of deep-marine depositional elements such as slope mass-transport deposits (*e.g.* Laurin *et al.*, 2007; Arnott *et al.*, 2011), slope and base-of-slope channel-levee complexes (Navarro, 2006; Arnott, 2007a; Gammon *et al.*, 2007; Navarro *et al.*, 2007a, b; O'Byrne *et al.*, 2007; Schwarz and Arnott, 2007; Mussa-Caleca, 2008; Dumouchel, 2015), basin-floor deposits (Meyer, 2004; Meyer and Ross, 2007; Rocheleau, 2011; Terlaky, 2014; Popović, 2016; Terlaky *et al.*, 2016), and overbank deposits (Arnott, 2007b; Davis, 2011; Terlaky and Arnott, 2014).

In regards to this thesis, Navarro (2016) was the first to describe the stratal architecture and component facies in the part of the lower Isaac Formation that includes this study. Her work focussed on logging and identifying larger-scale stratal elements (Dm to several Dm) and arranging these into the overall architecture. She reported that the succession consisted of four vertically-stacked packages, each about 3-5 m thick and laterally traceable for at least 400 m. The individual packages were composed of multiple thin- to medium-bedded, normally-graded sandstones with anomalously thick mud caps (Navarro, 2016). Also, these beds were interpreted to be enriched in fine-grained sediment, which was thought to be a consequence of partial flow confinement caused by local topography and termed ponded deposits (*Fig. 1.13*). The objective of

this thesis, therefore, was to test this interpretation and develop a more comprehensive understanding of these distinctively mud-rich strata.

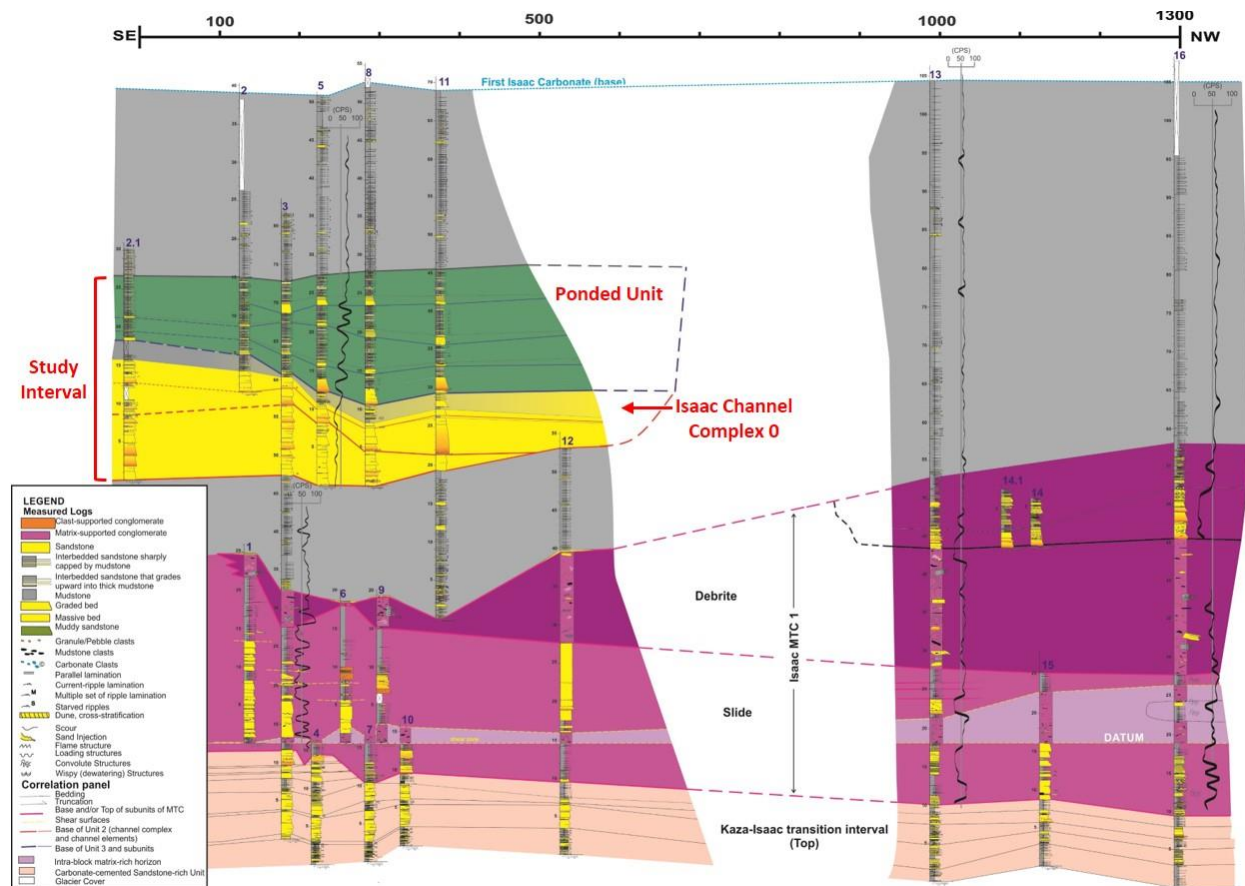


Figure 1.13: Detailed stratigraphic correlation of the lower Isaac Formation in the Windermere turbidite system, Cariboo Mountains, British Columbia, Canada. Note the ponded unit (green) above Isaac channel complex 0 (modified from Navarro, 2016).

1.5 Objectives & Thesis Structure

The primary objective of this thesis is to build on part of the Ph.D. work of Navarro (2016) by conducting a detailed analysis of the sedimentological and petrographic characteristics of Isaac Channel Complex 0 (ICC0), but more specifically the heterolithic deposits that sharply overlie ICC0 in order to better reconstruct the depositional conditions that deposited this unique strata. The primary goals include: 1) detailed description, both macroscopically and microscopically, of

strata in the study area, 2) describe the vertical stacking patterns of facies and their lateral changes, and 3) interpret the depositional conditions of heterolithic strata.

This thesis is written in a combination of paper- and traditional-format, with chapter 4 as a stand-alone manuscript for future submission to a peer-reviewed journal. Chapter 1 provides an overview of the regional and local geology as well as study methods, chapter 2 is a literature review of deep-marine systems with a focus on fine-grained sediment-gravity flows, chapter 3 is a detailed lithological and petrographic description of all facies, chapter 4 is a stand-alone manuscript, and chapter 5 summarizes conclusions and proposes areas for future research.

Chapter 2: Deep-Marine Processes

The deep sea is one of the Earth's most enigmatic environments due to its inaccessibility and vastness. In the early 20th century it was generally thought the deep-marine seafloor was a tranquil environment where only pelagic sediment accumulated from fine-grained sediment fallout (Friedman and Sanders, 1997). However, the first published evidence contrary to this idea was in 1897 when John Milne recognized the occurrence of submarine landslides breaking cables on the ocean floor (Milne, 1897). It was not until several decades later that a name for the mechanism responsible for these breaks, and also for most of the sand and gravel transport in the world's deep oceans was introduced, and termed turbidity currents (Johnson, 1938). The later research of Marie Tharp and Bruce Heezen further suggested the sequential breakup of Transatlantic telegraph cables on the ocean floor downslope of the epicentre of the 7.2 magnitude 1929 Grand Banks earthquake offshore Newfoundland was attributed to turbidity currents triggered by the earthquake (Heezen and Ewing, 1952; Heezen *et al.*, 1964). Still later, it was Philip Kuenen who proposed the term 'turbidite' to refer to the deposit of a turbidity current (Kuenen, 1957), which then was followed by Arnold Bouma who described the idealized vertical facies succession of turbidites (Bouma, 1962). Presently it is well known that the deep sea, at least broadly, is a highly dynamic environment where gravity-controlled mass-wasting and mass-flow processes transport coarse sediment hundreds of kilometres downslope at speeds of 3-19 m/s (*c.* 11-68 km/h) (Talling, 2014), and deep bottom currents (*i.e.* contour currents) related to the global thermohaline circulation system redistribute vast quantities of sediment along the continental slope (Shanmugam, 2000).

2.1 Classification: Subaqueous Sediment-Gravity Processes

The deep sea refers to areas > 125 m deep (bathyal) beyond the continental shelf-slope break (Pettingill and Weimer, 2002). Here the physiographic elements include the continental slope, rise, and basin floor, and where mass transport processes such as slides, slumps, debris flows, and turbidity currents, in addition to bottom currents are the dominant transport and depositional mechanisms of sand and coarse sediment (Shanmugam, 2000). Sediment-gravity flows (a specific kind of density flow) are arguably the most volumetrically important sediment-transporting mechanism on Earth, and in turn build up the Earth's largest geomorphic features termed submarine fans (Piper and Normark, 2009; Talling, 2014). They can be triggered by a variety of mechanisms, including earthquakes/seismic loading (see above), failure of an oversteepened continental slope, plunging of hyperpycnal river floodwater, delta-lip failure, cyclic storm-wave loading, gas charging, failure of rapidly deposited, underconsolidated sediment, transformation of slumps, *etc.* (Normark and Piper, 1991; Locat and Lee, 2002; Piper and Normark, 2009; Talling, 2014). These various processes of flow initiation control the character of the eventual resultant deposit (*Fig. 2.1*). However, natural sediment-gravity flows are rare on observable timescales, difficult to monitor directly, and episodic (Piper and Normark, 2009). Thus much of our current knowledge is based on observations from the ancient sedimentary record and laboratory experiments.

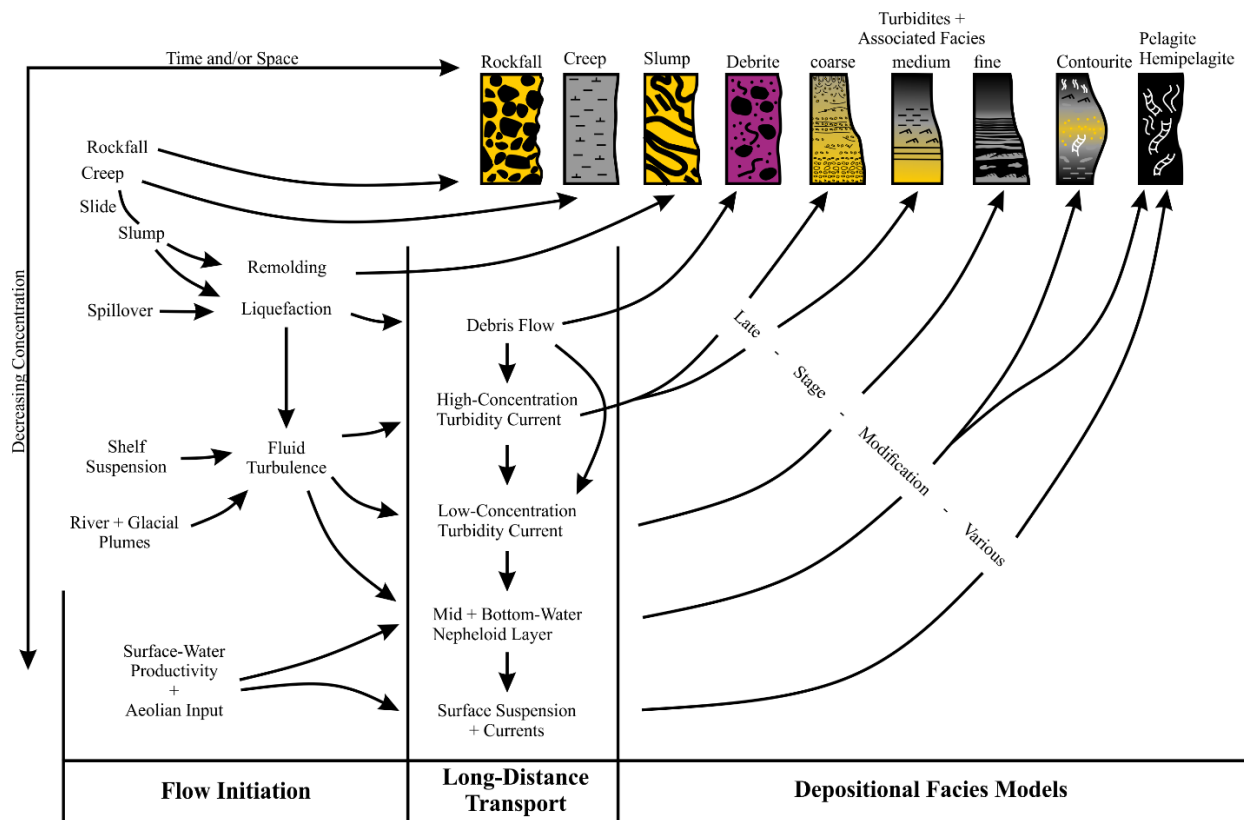


Figure 2.1: Flow chart showing the main flow initiation mechanisms, type of subaqueous sediment-gravity flow generated, and resultant depositional facies model for each process. Coarse-grained turbidites are described by the Lowe classification (1982), medium-grained turbidites by the Bouma classification (1962), and fine-grained turbidites by the Stow and Shanmugam classification (1980). Note that the different kinds of flow initiation, flow types, and depositional facies form a continuum that generally transforms from one type into another (redrawn and modified from Stow and Piper, 1984a).

Different flow types may evolve and change character along the flow path (*Fig. 2.2*) (Fisher, 1983), typically resulting from a reduction in sediment concentration caused by the progressive influx of water during downslope movement and/or sediment deposition (Mulder and Alexander, 2001). Flow transformation, which is profoundly affected by the rate of fluid entrainment, is reliant on factors such as water depth, slope gradient, flow thickness, lateral confinement, and bed roughness (Mulder and Alexander, 2001).

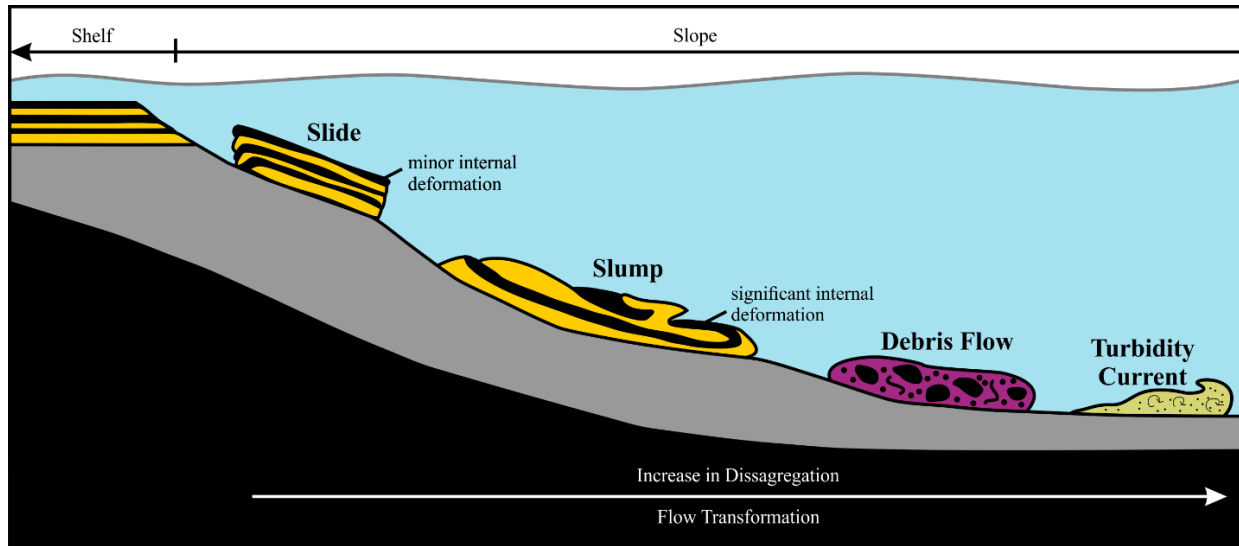


Figure 2.2: Schematic of a shelf edge sediment failure and its downslope evolution from a coherent mass (*e.g.* slide, slump) to a cohesionless sediment-water dispersion (*e.g.* debris flow, turbidity current). Note that this evolution represents only one of several possible styles of flow transformation (redrawn and modified from Shanmugam *et al.*, 1994).

Due to the complexity of sediment-gravity flows, the wide variety of depositional processes and added complexity of deformation and diagenesis makes interpreting the original flow conditions based on depositional characteristics in the sedimentary record difficult (Mulder and Alexander, 2001). This complexity, then, has led to a plethora of nomenclature in the scientific literature that often has caused confusion (*Fig. 2.3*).

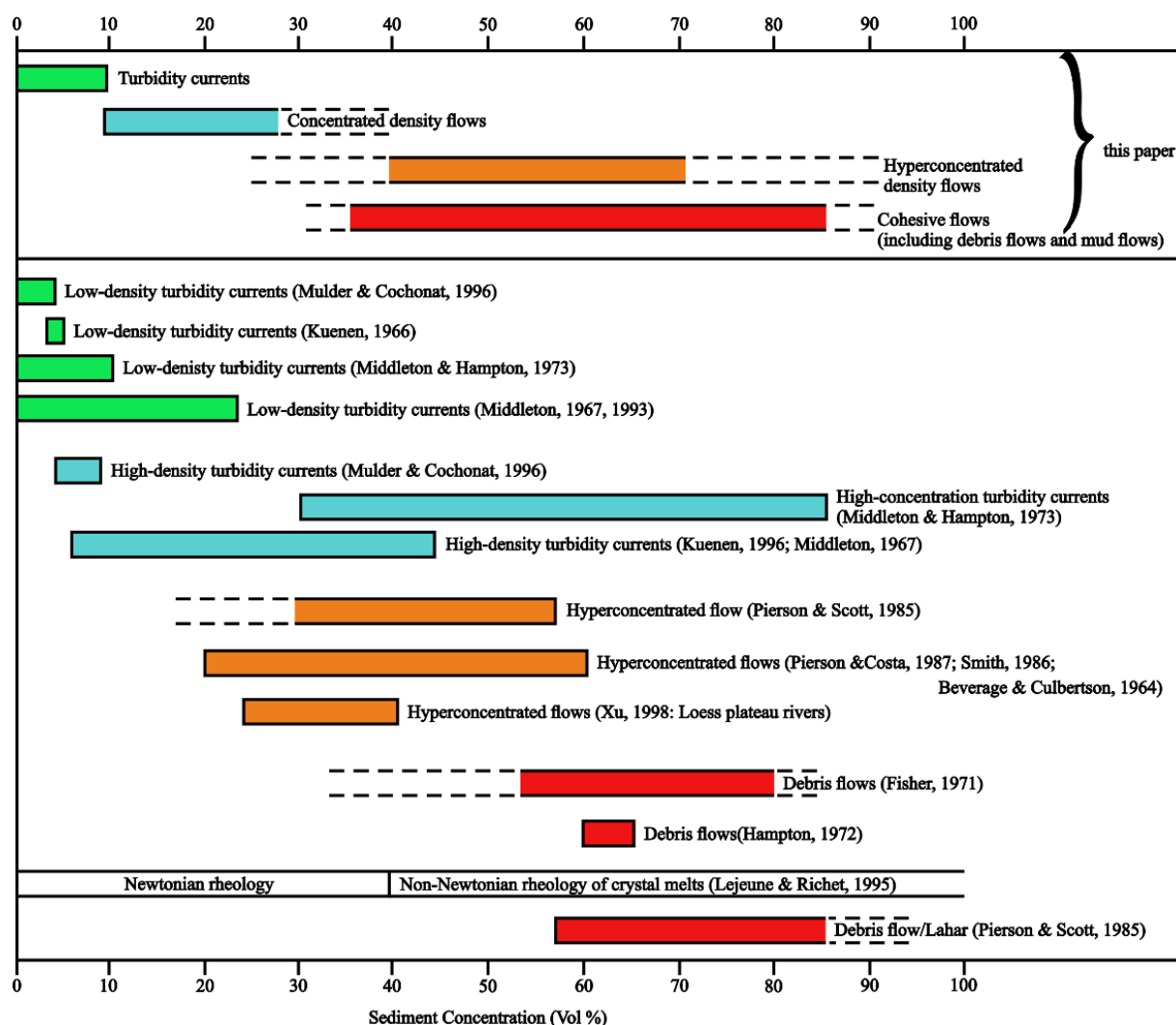


Figure 2.3: Nomenclature of flow type based on volume percent sediment concentration where the dashed lines represent the range of possible conditions for boundaries proposed by different models. Overabundance and frequent misuse of terms has led to common disagreement and controversy in the scientific literature (redrawn and modified from Mulder and Alexander, 2001).

The early classification schemes of Middleton and Hampton (1973, 1976) recognized four end-member kinds of flows based on the dominant sediment supporting mechanisms: 1) debris flows maintained by matrix strength, 2) grain flows sustained by grain-to-grain collisions, 3) fluidized/liquefied flows supported by escaping fluid, and 4) turbidity currents maintained by fluid turbulence. Alternative schemes, for example, classified flows based on fluid rheology and flow state (*e.g.* Dott, 1963), specifically Newtonian flows (*i.e.* flows with no strength) or non-Newtonian flows with plastic rheology (*e.g.* debris flows), whereas others considered both

sediment-support mechanisms and rheology (Lowe, 1979, 1982). More recently Mulder and Alexander (2001) attempted to reconcile these differences with a classification based on sediment-support mechanisms and physical flow properties (*e.g.* flow duration, cohesive properties of particles, sediment concentration) and in which two end-member types of flows were recognized: cohesive flows and non-cohesive or frictional flows (*Fig. 2.4*).

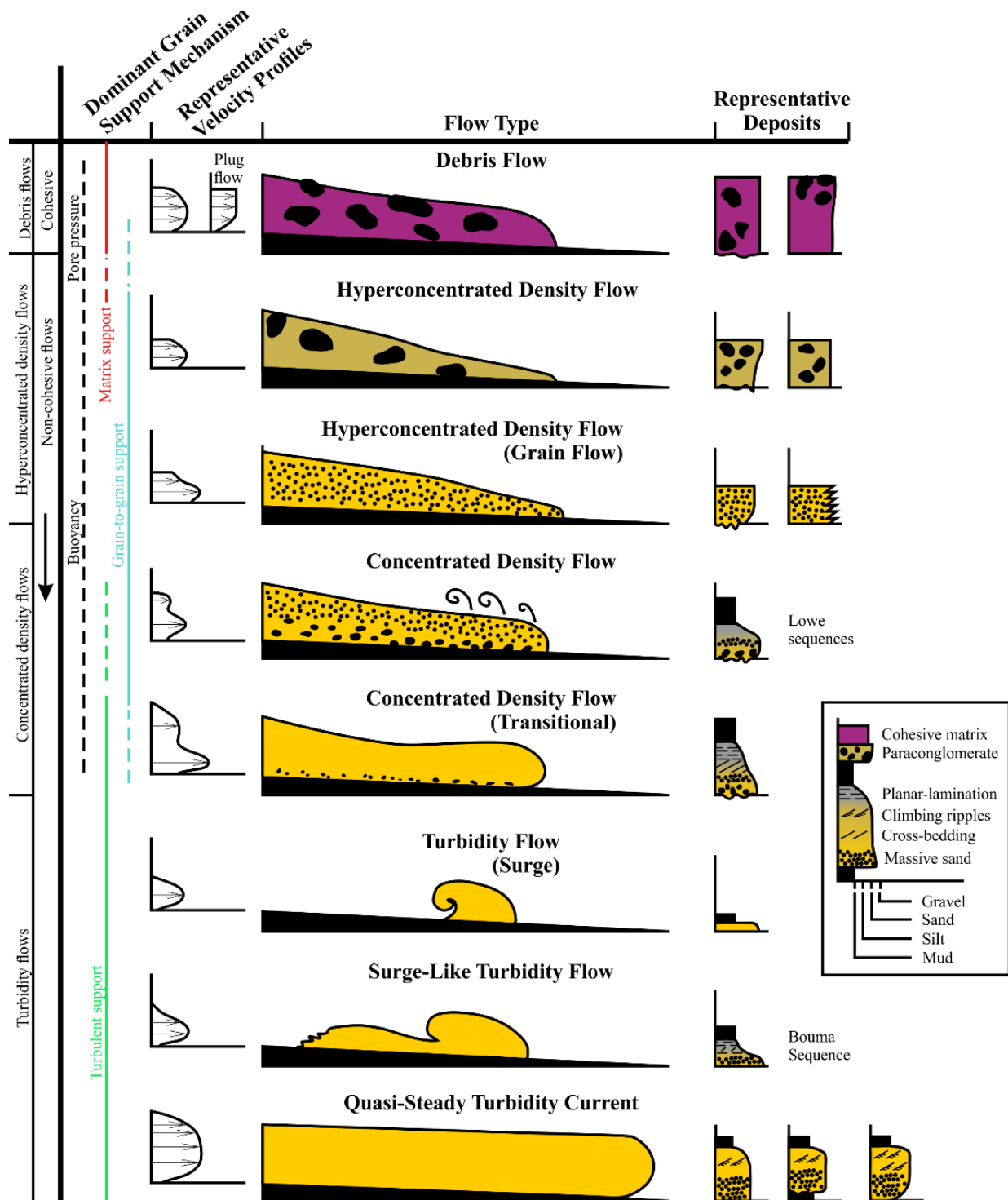


Figure 2.4: The classification of Mulder and Alexander (2001), which is adopted in this thesis, divides subaqueous density flows into cohesive and frictional flows. Debris flows represent cohesive flows, whereas non-cohesive, or similarly frictional flows, are subdivided into hyperconcentrated density flows, concentrated density flows, and turbidity currents that are further differentiated based on flow duration into surge, surge-like flow, and quasi-steady currents (redrawn and modified from Mulder and Alexander, 2001).

2.2 Subaqueous Mass Movements

Just as subaerial mass-movements are active across the globe, so too are subaqueous mass-movement or mass-wasting events that occur when gravity, the main driving force, equals or exceeds the tensile strength of the sediment pile resulting in downslope movement (Arnott, 2010). Displacement is often up to hundreds of kilometres along detachment planes (*décollement*) (Arnott, 2010), which based on seismic images, appear to follow stratigraphic horizons (Levesque *et al.*, 2004). Once gravity exceeds the tensile strength movement is initiated and continues until the resisting forces, primarily friction along the basal *décollement*, surpasses gravity and the mass-movement freezes *en masse* (Arnott, 2010).

The deposits of subaqueous mass-movements, termed mass-transport deposits (*i.e.* MTDs), are classified by some geologists to include debris flow deposits (Nardin *et al.*, 1979), whereas in the Mulder and Alexander (2001) classification debris flows are considered cohesive flows. In general, two end-member types of submarine mass-movements are recognized: slides and slumps (Varnes, 1958). Both end-members remain coherent during movement and are differentiated based on the intensity and nature of internal deformation (Arnott, 2010). Slides exhibit minor deformation dominated by brittle fractures (Armitage *et al.*, 2009). Slumps, on the other hand, exhibit rotation and deformation that is more intense and ductile (Arnott, 2010). Mass-transport complexes often consist of an assemblage of complexly arranged, disorganized slide, slump, and debris flow deposits (Arnott, 2010). In some cases, these deposits form enormous features that cover areas up to several 1,000 square-kilometres and are up to 100s of metres thick (Arnott, 2010).

2.3 Sediment-Gravity Flows

Gravity currents occur where a dense fluid displaces a less-dense fluid as it moves (Arnott, 2010). A density current composed of suspended sediment is termed a sediment-gravity flow (Arnott, 2010). Most natural gravity currents are stratified with vertical variation in sediment concentration, sediment type, and/or water density caused by fluid entrainment (*i.e.* seawater) and particle settling (Garcia and Parker, 1993; Garcia, 1994; Altinakar *et al.*, 1996). The character of a flow can also vary in the direction of flow (*i.e.* non-uniform) and/or time (*i.e.* unsteady) (Kneller, 1995). Therefore, the deposit of a single transport event may have formed under a variety of flow types exhibiting different lithological characteristics making it difficult to classify the complex interaction of component flow types prior to deposition (Mulder and Alexander, 2001). Thus, when a deposit cannot be related to a particular flow type, the general term sediment-gravity-flow deposit is used.

2.3.1 Cohesive Flows

Cohesive flows, commonly termed debris flows or mud flows, are flows where particle-particle cohesive forces suspend sediment (typically a silt-clay mixture giving the flow matrix strength) (Nardin *et al.*, 1979; Mulder and Alexander, 2001; Arnott, 2010). Cohesion between adjacent particles reduces the ingress of ambient fluid into the flow, which helps maintain strength and reduce turbulent mixing (Mulder and Alexander, 2001). The solid and liquid phases in cohesive flows are of similar magnitude by volume concentration, but it is the cohesive matrix that gives the flow its pseudoplastic rheology (Mohrig *et al.*, 1998; Mulder and Alexander, 2001) and permits particles ranging from clay to boulders, some up to several metres in diameter, to be suspended (Rodine and Johnson, 1976; Lowe, 1982; Leigh and Hartley, 1992; van Weering *et al.*, 1998). However, it is important to note that the clay fraction can be as low as a few volume percent

and still suspend larger grains (Hampton, 1975). In addition to matrix strength, mechanisms like particle-particle interaction, buoyancy effects, hindered settling, and elevated pore pressure aid in particle suspension (Pierson, 1981; Iverson *et al.*, 1997; Major and Iverson, 1999; Mulder and Alexander, 2001). Debris flows may also transform partially or fully into frictional flows as a result of mixing and erosion at the head of the flow (Hampton, 1972). In some cases, mixing develops a turbulent suspension along the top of the flow, but due to the low permeability of most debris flows, and therefore limited water infiltration, the reworked volume is negligible ($< 1\%$) (Arnott, 2010). However, as the proportion of sediment coarser than medium silt increases, strength decreases causing debris flows to undergo more extensive mixing, and therefore be more prone to transform into a frictional flow (Hampton, 1972, 1975).

Modern debris flows have been shown to runout for distances of up to several hundred kilometres (Gardner and Kidd, 1983; Simm *et al.*, 1991; Gee *et al.*, 1999) with little to no basal erosion (Pickering *et al.*, 1989; Gee *et al.*, 1999). These features have been attributed to hydroplaning (Mohrig *et al.*, 1998); a process whereby a layer of water becomes trapped between the bed and the mud-rich, low permeability flow resulting in a dramatic reduction in basal drag (*i.e.* resistance to flow) (Iverson, 1997; Mohrig *et al.*, 1998). Like slides and slumps, debris flows deposit *en masse* when the gravitational force (the main driving force) equals or exceeds the shear resistance controlled by viscosity and friction at the bed (Lowe, 1982; Postma, 1986). This type of rapid, *en masse* deposition, in addition to limited internal grain mobility, explains the chaotic arrangement and characteristically poorly sorted nature of debris flow deposits, or debrites (Mulder and Alexander, 2001).

Cohesive flows are divided into debris flows and mud flows based on grain size distribution, and mud flows are further subdivided into silty mud flows ($< 25\%$ clay) and clay-rich

mud flows (> 40% clay) (Hawkins and Pinches, 1992; Mulder and Alexander, 2001). These poorly sorted deposits are classified based on the percentage of gravel by volume and the proportion of mud:sand (Folk, 1954; Moncrieff, 1989), where mud flows have less than 5% gravel and a mud:sand ratio more than 1:1, thus transporting negligible coarse sediment (Mulder and Alexander, 2001). Because water content in these flows is low, the thickness of the deposit is similar to the thickness of the parent flow with only minor thinning due to post-depositional consolidation and compaction (Mulder and Alexander, 2001). Despite these effects, examples in the ancient sedimentary rock record range up to more than 200 m thick (Hiscott and James, 1985; Leigh and Hartley, 1992) and have steep margins reflecting matrix strength (Arnott, 2010). Spatially, these deposits can be up to tens of kilometres wide and exhibit a distinctive chaotic reflection appearance in seismic images (Arnott, 2010).

2.3.2 Frictional Flows

Frictional flows (non-cohesive) form a continuum based primarily on sediment concentration, and in the classification of Mulder and Alexander (2001), three main subdivisions based on decreasing sediment concentration are recognized: hyperconcentrated density flows, concentrated density flows, and turbidity currents. Each subdivision is differentiated by the dominant sediment-supporting mechanism, which includes grain-to-grain interaction, buoyancy, turbulent Reynolds stresses, pore pressure, and bed support (*Fig. 2.4, 2.5*) (Mulder and Alexander, 2001). Factors influencing the dominant sediment-supporting mechanism are sediment concentration, flow conditions, sediment type, and grain size distribution (Mulder and Alexander, 2001). The behaviour of frictional flows is directly related to sediment concentration, which during flow run-out can change dramatically due to turbulent mixing and the ingress of ambient fluid (Mulder and Alexander, 2001). The proportion of cohesive and non-cohesive particles that define

the division between cohesive and frictional flow behaviour is poorly understood for subaqueous flows and varies depending on sediment concentration and flow conditions (Fisher, 1971; Iverson, 1997).

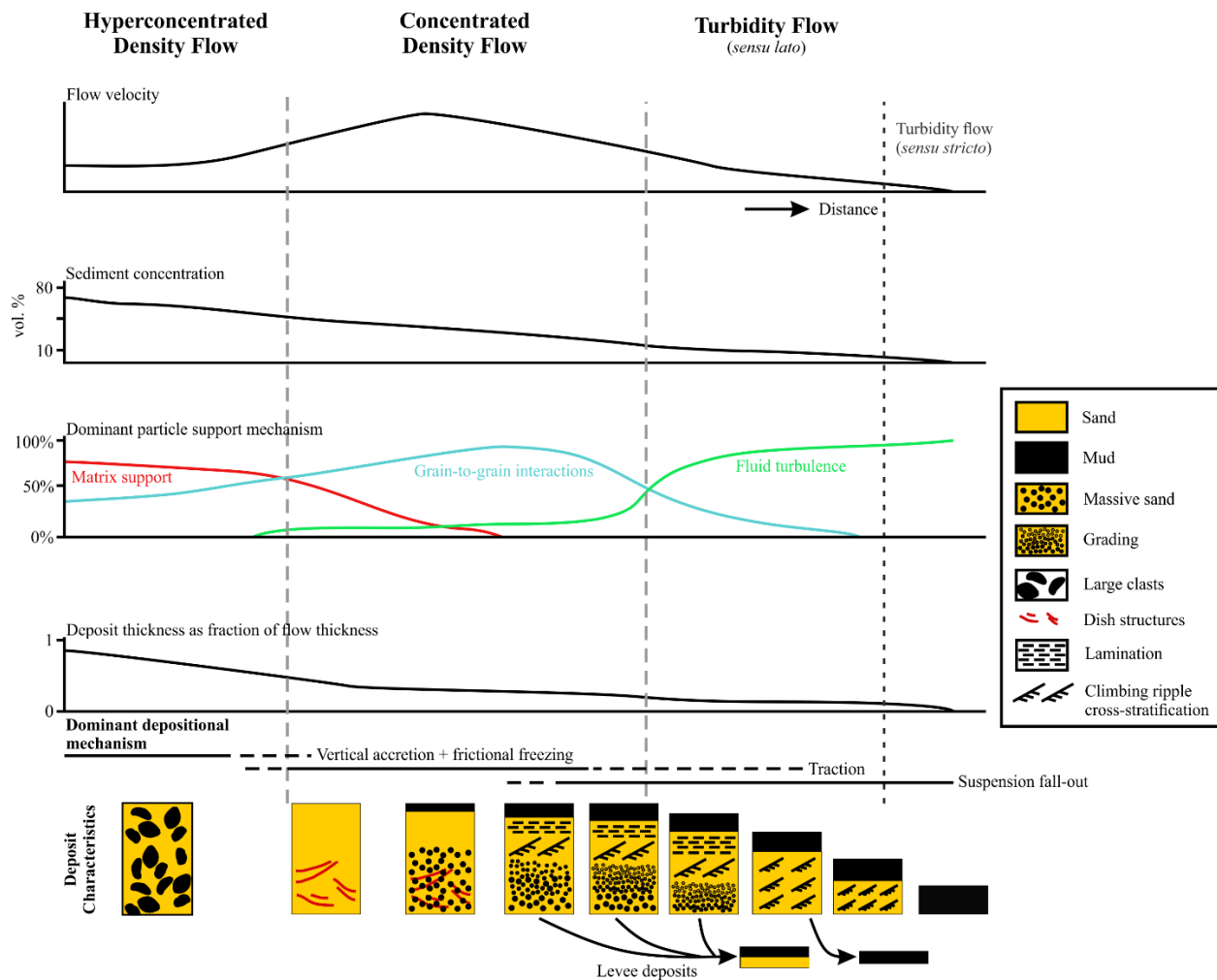


Figure 2.5: Diagram illustrating the changes in flow characteristics of frictional flows across the different subdivisions. Flow types are differentiated based on differences in sediment concentration, which results in differences in flow behaviour and depositional characteristics (redrawn and modified from Mulder and Alexander, 2001).

2.3.2.1 Hyperconcentrated Density Flows

Hyperconcentrated density flows have previously been termed frictional or sandy debris flows (Shanmugam, 1996), inertial grain flows (Takahashi, 1981), and noncohesive/cohesionless debris flows (Falk and Dorsey, 1998). Although similar in sediment-to-fluid content to cohesive

flows, hyperconcentrated density flows are more depleted in cohesive particles and/or are more internally sheared (friction dominated) due to higher flow velocities and possible flow over an irregular surface (Mulder and Alexander, 2001).

Hyperconcentrated density flows consist of more than 25% sediment volume concentration and commonly are highly erosive (Mulder and Alexander, 2001). Flow rheology depends not only on sediment concentration, but also the relative densities of the sediment and fluid, and sediment size (Julien and Lan, 1991). Owing to high internal friction, hyperconcentrated flows require sufficiently steep slopes for movement and therefore tend to have limited runout distances (Mulder and Alexander, 2001). Particle interlocking (*i.e.* frictional freezing) is the primary mechanism causing deposition, which generally forms structureless, ungraded beds (Mulder and Alexander, 2001). Hyperconcentrated flows containing some cohesive particles may undergo local fluidization or liquefaction (*cf.* Nichols, 1995) producing post-depositional fluid escape structures such as dish and pillar structures (Lowe and Guy, 2000). Similar to cohesive flows, large rafted blocks and clasts are common (Mulder and Alexander, 2001).

2.3.2.2 Concentrated Density Flows

Concentrated density flows have lower sediment concentration compared to hyperconcentrated flows and exhibit a change in sediment-support mechanisms from primarily buoyancy and inertial forces to grain-to-grain interactions generating dispersive pressure (Kneller and Branney, 1995). The sediment concentration at which this change occurs depends on a variety of factors, including sediment size, sorting, relative density, and composition, thus making it difficult to determine a single threshold condition for the flows or resultant deposits (Mulder and Alexander, 2001), but likely is close to the transition between non-Newtonian and Newtonian behaviour (Mulder and Alexander, 2001). With reduced sediment concentration grains begin to

move more freely allowing differential settling of larger and/or denser particles, and accordingly coarse-tail normal grading (Mulder and Alexander, 2001). Additionally, lower sediment concentration results in greater entrainment of ambient water causing flow dilution and the onset of Newtonian fluid behaviour (Lowe, 1982). Fluid turbulence is a secondary sediment-support mechanism in concentrated density flows but is the dominant mechanism in the upper part (mixing cloud) and head region of the flow, where because of extensive turbulent entrainment of ambient fluid causes the flow to become stratified (Mulder and Alexander, 2001). Also, being more turbulent allows concentrated density flows to entrain more sediment due to enhanced seabed erosion, which in some cases becomes the primary source of suspended particles (Piper and Savoye, 1993; Mulder *et al.*, 1997) – this increase in density also allows for longer duration. In the literature these flows have also been termed high-density turbidity currents, but because grain-to-grain interaction rather than fluid turbulence is the main particle support mechanism, the more appropriate term is concentrated density flow (Mulder and Alexander, 2001).

In 1982, Donald Lowe developed a six-part (R_1 , R_2 , R_3 , S_1 , S_2 , S_3) classification of concentrated density flow deposits with gravel-rich (R) and sand-rich (S) end-members (*Fig. 2.6*). Gravel-rich R_1 consists of structureless gravel, R_2 is inversely-graded gravel deposited by traction carpet deposition (unique kind of frictional freezing; *sensu* Lowe, 1982), and R_3 comprises normally-graded gravel deposited from suspension (Lowe, 1982). In sand-rich strata, S_1 is formed by bed load transport and is either planar- and/or cross-stratified sandstone, S_2 is deposited by traction carpet deposition as sedimentation rate increases, and S_3 (equivalent to Bouma T_a division) caps the succession where sand is deposited from suspension after the cessation of traction carpet sedimentation (Lowe, 1982). Deposits of concentrated density flows, compared to

hyperconcentrated density flows, exhibit better developed normal grading and sorting, and traction transport sedimentary structures (Mulder and Alexander, 2001).

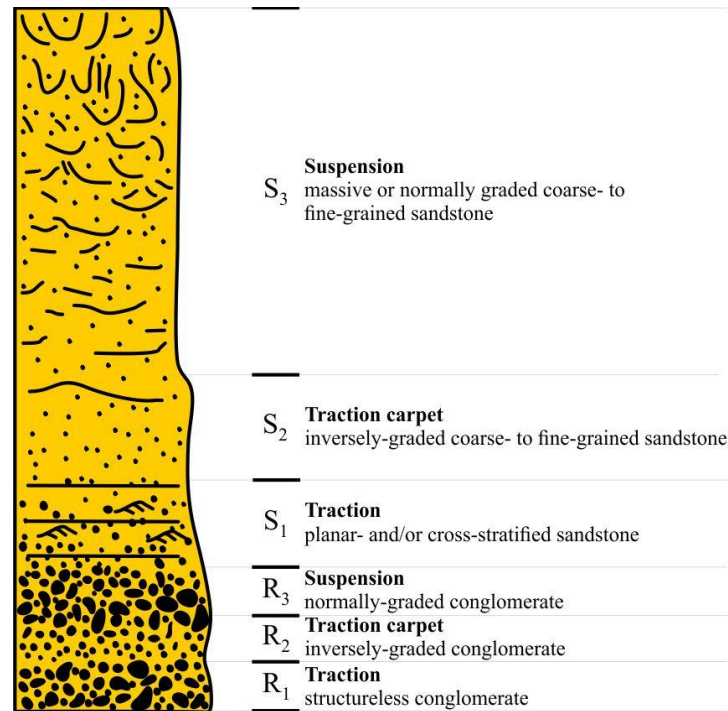


Figure 2.6: Idealized Lowe sequence for concentrated density currents showing the typical grain size and sedimentary structures for each of the six divisions as well as the primary sedimentation type (redrawn and modified from Lowe, 1982).

2.3.2.3 Turbidity Currents

The most common frictional flow is turbidity currents (Newtonian flows) which form the turbidites so ubiquitous in the deep-marine sedimentary record (Arnott, 2010). The term ‘turbidity current’ describes a flow made turbid by suspended sediment (Mulder and Alexander, 2001). However, due to the inaccessibility, destructive and episodic nature of natural turbidity currents, the connection between the formative process and the deposit remained speculative until the late 1800s (see earlier discussion) (Arnott, 2010). It was not until the 1960s that researchers began to conduct laboratory experiments to study the dynamics of turbidity currents and attempt to link it to depositional characteristics in the sedimentary record (Middleton, 1966; Simpson, 1969). Based

on these flume experiments it was observed that turbidity currents exhibit three distinct, but not sharply bounded parts: head, body, and tail (*Fig. 2.7*) (Kneller and Buckee, 2000). The majority of mixing with the surrounding ambient fluid occurs along and beneath the head, which also is the most sediment-rich part of the flow (Arnott, 2010). The head is characterized by a sharp overhanging nose that slopes back upstream due to (fluid) resistance by the ambient fluid and carries sediment-rich fluid back toward the body (Arnott, 2010). In order to be sustained the body of the flow must continually feed sediment into the head, which accordingly moves slower than the body (Arnott, 2010). The head and headward part of the body usually exhibit waxing flow conditions (*i.e.* mean flow velocity increasing), whereas conditions in the rest of the body and tail are waning (*i.e.* mean velocity at a point is decreasing) (Kneller, 1995). Moreover, the three parts of a turbidity current consist of different sediment populations (Arnott, 2010). Due to differences in settling velocity, turbidity currents become longitudinally differentiated by grain size with coarse sediment becoming concentrated in the lower part of the head as finer sediment is displaced up and backwards into the body, and eventually forming a fine-grained low-density sediment dispersion that makes up the tail of the flow (Arnott, 2010).

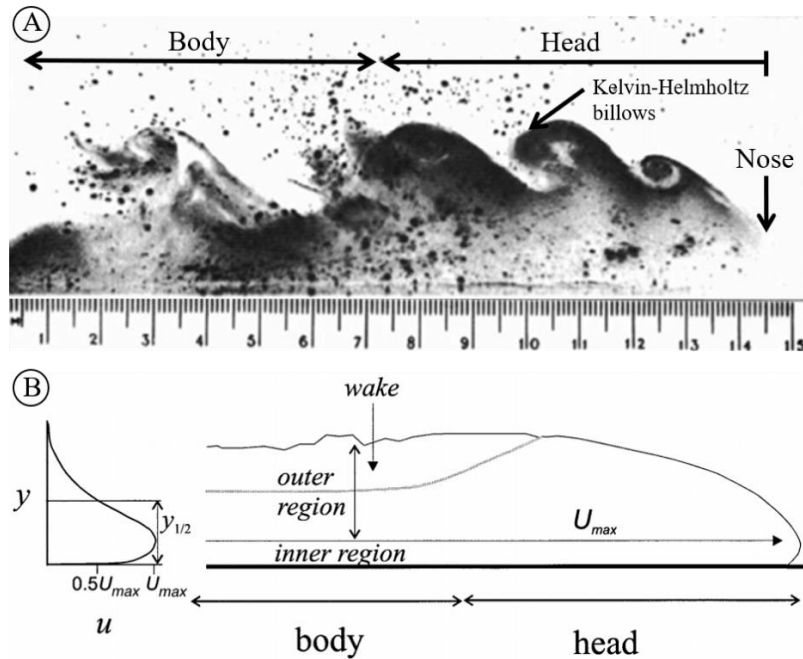


Figure 2.7: A) Laboratory saline turbidity current with a well-developed head and trailing body. Note also the well-developed Kelvin-Helmholtz instability and the associated wavy pattern generated along the top of the current (modified from Simpson, 1969). B) Diagram illustrating the typical morphology (right) and velocity profile (left) of a turbidity current. Note the region of extensive mixing in the wake due to interfacial instability along the top of the flow (Kneller and Buckee, 2000).

The threshold sediment concentration for turbidity currents is reported to be $< 9\%$ volume sediment concentration, which is termed the Bagnold limit (Bagnold, 1962). In such a flow the main particle-support mechanism is fluid turbulence as grain-to-grain interaction becomes less influential due to ambient fluid entrainment and consequent dilution (*Fig. 2.5*) (Middleton and Hampton, 1973; Hallworth *et al.*, 1993). Above the Bagnold limit, particles become so closely spaced that fluid turbulence begins to be dampened and other particle-support mechanisms like dispersive pressure, hindered settling, and buoyancy become increasingly important (Arnott, 2010).

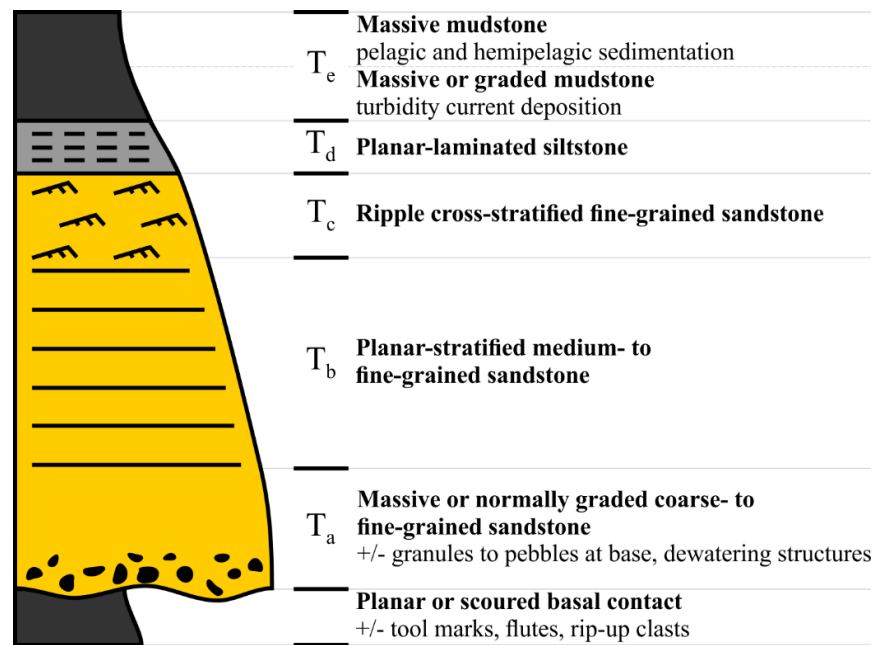


Figure 2.8: Complete Bouma sequence showing grain size and sedimentary structures observed in each division. Note that the full succession is rarely preserved in turbidites (redrawn and modified from Bouma, 1962).

In 1962, Arnold Bouma published what would become the standard vertical facies model for sand-rich turbidites (*Fig. 2.8*). He reported that turbidites, where fully developed, comprise five sharply bounded layers, or divisions: the basal layer consists of massive or normally graded sandstone (T_a) overlain successively by planar-stratified sandstone (T_b), high-angle, small-scale ripple cross-stratified sandstone (T_c), subtly parallel-laminated siltstone (T_d), and finally massive mudstone (T_e). Note that the complete set of divisions is rarely present in any one bed. Nevertheless, beds typically show a characteristic upward fining (normal grading) and the predictable vertical succession of sedimentary structures that indicates deposition from waning, low-density turbidity currents due to direct suspension settling (A- and E-divisions), traction transport (B- and C-divisions), and a combination of both (D division) (Pickering and Hilton, 1998; Shanmugam, 2000). The vertical transition from the traction-transport B- to C-division is often compared with sedimentary structures observed in open-channel flume experiments (*Fig. 2.9*) (Simons, *et al.*, 1965; Southard, 1975; Southard and Boguchwal, 1990). However, the notable

absence of an intervening layer of dune cross-stratification makes this correlation equivocal, if not doubtful (Arnott, 2012).

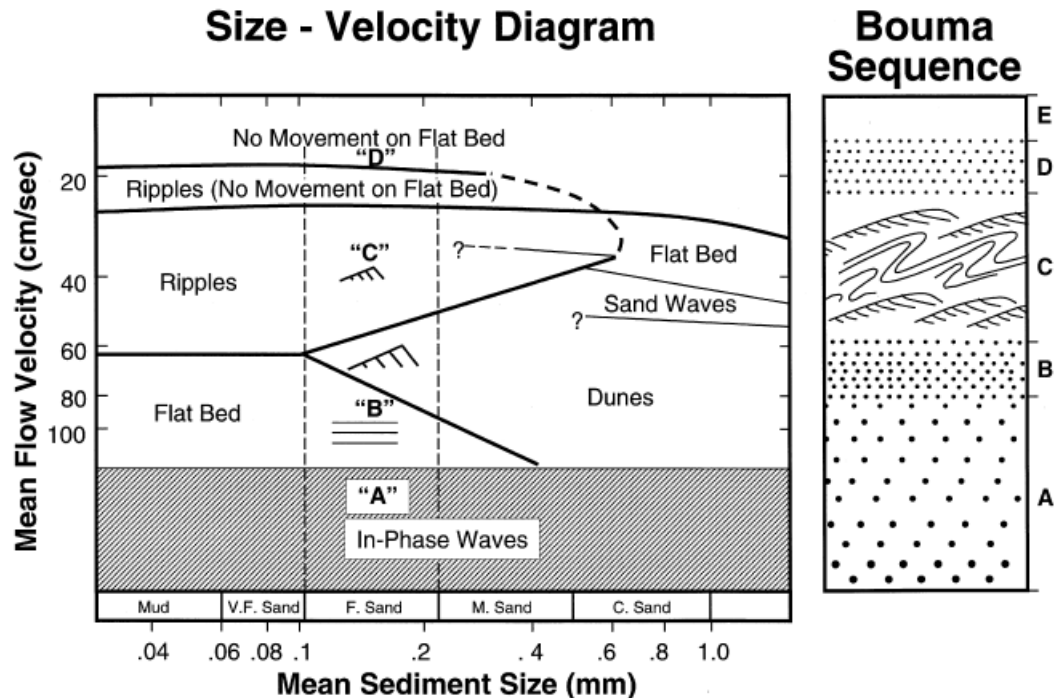


Figure 2.9: Comparison of the bedform stability diagram developed from flume experiments with 20 cm flow depth (Shanmugam, 2000 compiled from many sources, Southard, 1975; Southard and Boguchwal, 1990) with the Bouma sequence (1962). The Bouma divisions are labelled on the size-velocity diagram within their inferred stability field. Note that the dune stability field does not correspond to any Bouma division and the basal normally-graded T_a division of the Bouma sequence is absent in flume experiments.

Turbidity currents are generally unsteady flows with durations ranging from instantaneous surges (minutes-hours), to short-duration surge-like flows (hours-few tens of hours), and lastly prolonged quasi-steady flows (hours to months) (Mulder and Syvitski, 1995, 1996; Mulder and Alexander, 2001). It is useful to subdivide these flows based on flow duration as it influences flow character, and accordingly, the nature of deposition (Mulder and Alexander, 2001). Surge flows are likely uncommon in nature and consist only of an isolated head (*Fig. 2.4*) (Mulder and Alexander, 2001).

Surge-like flows tend to generate sedimentary structures typical of Bouma T_b - T_d division turbidites (Mulder and Alexander, 2001). Quasi-steady flows (*Fig. 2.4*), namely hyperpycnal

turbidity currents, form a more diverse array of vertically stacked sedimentary structures, but typically contain a basal inversely-graded unit that is often separated from the upper normally-graded Bouma-like succession by a sharp contact or erosional surface (Mulder *et al.*, 1998). Climbing ripples are also common and develop thick basal deposits where the flow is steady, depletive (mean velocity decreases in the flow direction), velocity remains low (1-2 cm/s to 1 m/s), and sedimentation rate high (Mulder and Alexander, 2001).

Deposition occurs via unhindered settling of discrete grains in a turbulent suspension with a Newtonian rheology, and is represented in the ancient rock record as normal grading (Dott, 1963; Nardin *et al.*, 1979; Lowe, 1979, 1982). The deposits of turbidity currents and concentrated density flows can be differentiated based on sediment caliber, the proportion of the Bouma T_a division relative to the total bed thickness, and the character of the sedimentary structures (Mulder and Alexander, 2001). Specifically, owing to their greater competence, concentrated density flows tend to be coarse with a thicker basal graded T_a division relative to the total bed thickness (Mulder and Alexander, 2001).

Fine-Grained Turbidity Currents

Fine-grained sediments (*i.e.* over 50% < 63 μm /silt), including fine-very fine sand, silt, and mud are the dominant sedimentary deposit in the world's oceans (silt and clay 2-10x sand in deep-marine) (Stow, 1979). However, research on the processes that transport and deposit these sediments is lacking compared to their coarser-grained counterparts, partly because of their common structurally homogeneous appearance, poor preservation potential, tectonic deformation, and generally poor exposure in the sedimentary record (McCave, 1972; Piper, 1978; Gorsline, 1984; Stow and Shanmugam, 1980; Stow and Piper, 1984a). Even though the approach and techniques used to examine fine-grained turbidites are similar to their coarser-grained

counterparts, a few unique problems arise: sediment is too fine to be properly described in hand sample or core and required more labour intensive and more costly laboratory techniques; clay minerals undergo significant post-depositional alteration that make it challenging to reconstruct conditions during deposition; and textural analysis cannot be as easily used to deduce hydrodynamic processes like in sandy turbidites because clay particles flocculate and change flow dynamics (Stow and Piper, 1984b). Further impeding research on fine-grained turbidites is the generally held assumption that clay particles can only settle out of suspension from dilute, low-velocity turbidity currents, resulting in deposition of a thin layer (Dzulynski *et al.*, 1959). However, clay particle flocculation and the formation of larger floccules, or aggregates, increases settling velocity and permits bed-load transport (Einstein and Krone, 1962; Partheniades, 1965).

Soon after the publication of the Bouma sequence (1962), geologists realized that the upper divisions (T_d - T_e) of the facies model did not adequately capture all the subtle features in these fine-grained strata (Kuenen, 1964; Van der Lingen, 1969; Hesse, 1975; Stow, 1977; Piper, 1978; Stow and Shanmugam, 1980). Kuenen (1964) was the first to propose modifications, most importantly to the T_e division, where a distinction was made for mud deposited by turbidity currents (e^t) or passive pelagic, hemipelagic, and/or biogenic sedimentation (e^p), which were termed T_e and T_f respectively by van der Lingen (1969) and Hesse (1975). Piper (1978) further subdivided the E division into three structural parts: basal laminated mud (E_1), graded mud (E_2), and ungraded mud (E_3).

In 1980, Stow and Shanmugam proposed a comprehensive nine-layer (T_0 - T_8) facies model (Fig. 2.10) for fine-grained turbidites. The thick (c. 1 cm), sharp-based ($\pm$ scoured base, load casts), normally graded basal unit contains the coarsest sediment (very-fine sand – coarse silt) and often comprises three parts: lenticular silt lamina overlain by planar- or cross-laminated middle, and

fading ripples (type C ripples, *sensu* Jopling and Walker, 1968) that are described as very-fine sand and/or silt grading into muddy troughs which form an irregular or lenticular interlamination as the ripples migrate (T₀). This is overlain by a mud layer with convolute silt laminae (T₁), thin, irregular, silt laminae interstratified with mud, generally with low-amplitude and long-wavelength cross-laminated climbing ripples that produce interlaminated silt/mud lamina that are slightly inclined caused by the migration of the ripples (T₂), rhythmic planar silt and mud laminae (T₃), mud with faint discontinuous interstratified silt laminae (T₄) and indistinct wispy or convolute silt laminae (T₅), graded mud with dispersed silt lenses (T₆), structureless mud with rare silt pseudonodules (T₇), and finally bioturbated mud ($\pm$ pelagic and/or hemipelagic sediment) (T₈) (Stow, 1977; Piper, 1978; Stow and Shanmugam, 1980).

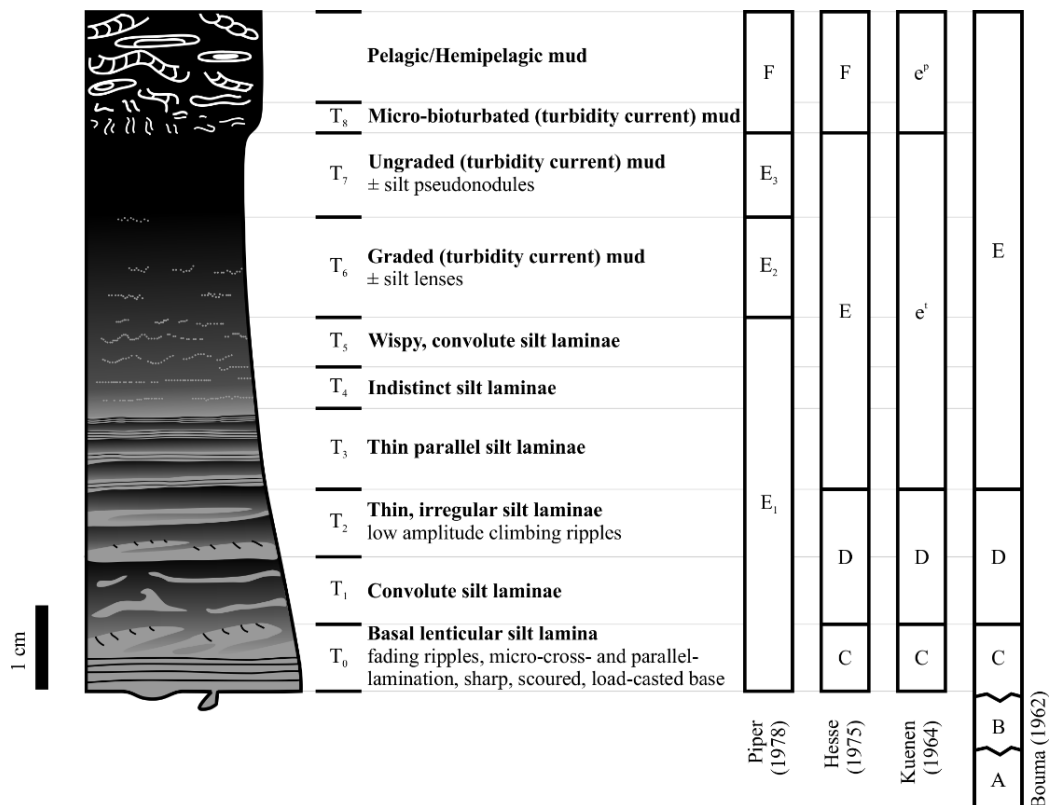


Figure 2.10: Idealized facies model for fine-grained turbidites showing its distinctive component parts and their average thickness. The complete sequence is interpreted to be deposited in a number of hours to a few days from a single waning fine-grained turbidity current and typically is of the order of seven centimetres thick. Note that beds rarely exhibit all divisions (redrawn and modified from Stow and Shanmugam, 1980; Stow and Piper, 1984a).

A characteristic feature of deep-marine fine-grained turbidites is the abundance of sharply bounded silt laminae in the basal part of the bed (*Fig. 2.10*) (Stow and Bowen, 1978). Silt laminae vary in thickness from < 0.5-10 mm (Stow and Bowen, 1980) and in some cases, such as the T₀ and T₂ divisions, are parallel or cross-laminated (Stow, 1979). Some laminae are accentuated by a concentration of heavy minerals, lutite, and/or foraminifera in them (Stow, 1979). The base of laminae is commonly irregular and marked by erosional scours, rip-up clasts, and post-depositional loading features like flame structures and load casts (Stow, 1979).

Fine-grained turbidites are also progressively graded (T₀-T₆) in terms of their modal grain size, thickness of silt laminae, and silt/clay ratio, however changes in the interlaminated mud layers tend to be more subtle (Stow and Bowen, 1980; Stow and Piper, 1984a). Changes in colour (*e.g.* light grey to dark grey/black) often reflect compositional and textural grading and provides the simplest visual method for identifying layering in fine-grained turbidites in the field or core (Stow and Piper, 1984a).

Fine-grained turbidity currents, being composed of fine, easily suspended sediment, are inherently low concentration flows that become more dilute with distance of travel (Stow and Bowen, 1978). Flows are estimated to have low concentrations (1-2.5 g/L), be at least several hundreds of metres thick, and travel downslope at velocities of 10-20 cm/s (0.36-0.72 km/h) (Stow and Bowen, 1980). Due to their cohesive properties, fine-grained turbidity currents often flow further from their origin (see discussion in 2.3.1) compared to noncohesive sediment-gravity flows (Baker *et al.*, 2017).

2.4 Bottom Currents

Thermohaline-induced geostrophic bottom currents (contour currents) in modern oceans tend to flow approximately parallel to bathymetric contours (along continental slopes) below c. 300 m and are important agents in reworking deep-marine sediments on the ocean floor (Heezen *et al.*, 1966; Flood and Hollister, 1974; Stow *et al.*, 1998). The resulting deposit is called a contourite (Hollister, 1967). These currents can transport sediment from mud to coarse sand forming large, elongate (10-100 km long), 10s km wide, and up to > 1 km thick, slope-parallel sediment bodies called sediment drifts (Stow and Piper, 1984b; Arnott, 2010). Sediment deposited in ocean basins by sediment-gravity flows are in places being continually reworked and redeposited by bottom currents (Shanmugam, 1997; Shanmugam, 2000), and makes the differentiation of turbidites and contourites problematic (Stow, 1979; Arnott, 2010).

Contourites can be differentiated from turbidites based on inverse grading exhibited at various scales, principally fine-grained sand and silt, thin-bedded to laminated sand (typically < 5 cm), internal erosional surfaces, sharply bounded units (when interstratified with turbidites), no distinct vertical facies succession, well sorted sediment with low depositional matrix due to winnowing of fines, and rhythmicity of sand and mud layers (Stow, 1979; Shanmugam, 2000; Arnott, 2010). Contourites often exhibit a homogenous texture due to intense bioturbation, which reflects long periods of exposure on the sea floor and low instantaneous sedimentation rate that contrast the conditions of turbidite deposition (Stow, 1979).

Chapter 3: Facies Descriptions and Interpretations

Macro- and microscopic analysis of strata in the lower Isaac Formation identified four lithofacies in the study area. Based on sedimentary structures and grain size, each facies succession (in order from facies 1-3) represents a bed, however, not all beds contain all facies. It is important to note that these facies often grade both vertically and laterally from one to another forming a depositional continuum. Facies 1-3 are observed throughout much of the study area, however, are all rarely (6% of beds) contained within a single bed. Facies 4, on the other hand, is restricted to the northwestern part of the study area and is not the primary focus of this study and therefore is only described briefly.

Table 3.1: Summary of facies descriptions and interpretations.

Facies	Facies Description	Interpretation
F1	Massive or graded, structureless or planar/wavy-stratified clayey sandstone	High rates of suspension fallout overwhelms the development of tractional sedimentary structures depositing structureless sandstone. With decreasing rates of fallout better developed planar-stratification forms.
F2	Graded, plane parallel- (F2.A), cross- (F2.B) or wavy-stratified (F2.C) clayey sandstone with intercalated structureless sandy claystone	Deposition from turbidity currents that experienced a complex history of deceleration and patterns of deposition. Uniform bedload transport (F2.A) until the requisite hydraulic conditions in the near-bed region were able to initiate and amplify bed-surface defects that eventually develop into mature ripples (F2.B). With continued reduction in flow speed, the near-bed hydrodynamic instability (still capable of generating bed defects) was unable to cause sufficient amplification of the defect to sufficiently alter (sediment) transport patterns (F2.C).
F3	Graded, plane parallel-laminated sandy claystone to massive or graded, structureless claystone	Deposition from a dilute, fine-grained suspension. Planar-lamination may relate to the action of internal waves within the channelized turbidity current or alternating rheological changes in the viscous sublayer followed by <i>en masse</i> deposition of a fluid mud layer.
F4	Massive, interstratified, structureless clayey sandstone and claystone	Deposition from suspension fallout and/or background hemipelagic fallout interrupted episodically by deposition from low-concentration turbidity currents.

3.1 Terminology

Macroscopic and microscopic features and depositional interpretations for each facies are outlined next. Macroscopic features can be identified in the field with the naked eye (metre to

millimetre scale) or with the aid of a hand lens (< 0.125 mm), while microscopic characteristics are observed in thin section where grains or crystals as small as $3.9\text{ }\mu\text{m}$ can be discerned. Microscopic analysis was particularly useful in characterizing the grain size for fine-grained samples which helped validate the field-based measurements that were difficult to obtain.

The Udden-Wentworth grain-size scale (Udden, 1914; Wentworth, 1922) is used to describe particle size. Grains were measured along their long axis and classified as silt ($3.9\text{--}62.5\text{ }\mu\text{m}$), sand ($0.0625\text{--}2$ mm), granules ($2\text{--}4$ mm), and pebbles (up to c. 8 cm). Folk's (1965) terminology for carbonate cement crystal size was used to describe any carbonate cement and includes extremely coarse (> 4 mm), very coarse ($1\text{--}4$ mm), coarse ($0.25\text{--}1$ mm), medium ($0.062\text{--}0.25$ mm), fine ($16\text{--}62\text{ }\mu\text{m}$), very finely crystalline ($4\text{--}16\text{ }\mu\text{m}$), and aphanocrystalline ($< 1\text{ }\mu\text{m}$). Bedding is classified according to Ingram's (1954) genetic approach to define laminae and beds (*Fig. 3.1*). For clarification, a bed is characterized as a distinctive sharply bounded layer or stratum. Beds then stack to form an upward fining and thinning succession interpreted to have accumulated during a single sedimentation (*i.e.* depositional) event (*Fig. 3.2*). Laminae are discrete, thin layers that are < 1 cm thick. Interstratification is the rhythmic alternation of two distinct lithologies, each > 1 cm thick, whereas interlamination is similar, but layers are < 1 cm thick.

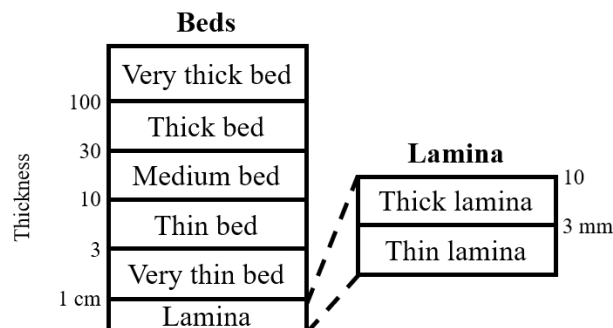


Figure 3.1: Terminology for the thickness of beds and laminae (adapted from Ingram, 1954).

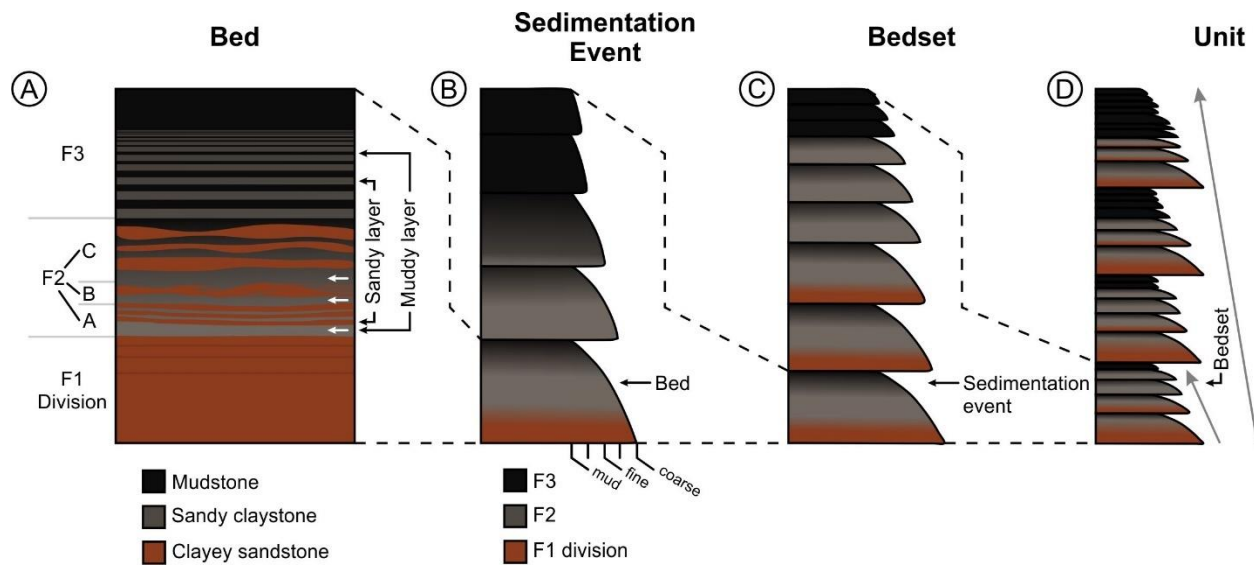


Figure 3.2: Hierarchy of stratal terminology used here from micro to macro scales. A) Idealized facies succession where a complete bed (centimetres to 10s cm thick) comprises an upward fining structureless overlain by planar-stratified F1 sandstone division sharply overlain by a thick mud layer (denoted by white arrows). This, then, is overlain by subfacies divisions F2.A, F2.B, and F2.C. Capping the bed is the F3 division, which is plane-parallel-laminated mudstone in the bottom part sharply overlain by structureless claystone. Note that F2 and F3 divisions are interstratified with sandier and muddy layers (see below). B) Schematic sedimentation event (10s cm to metres thick) comprising multiple beds (distinct basal and top contact) that typically fine and thin stratigraphically upwards and represent deposition during a single sedimentation event. C) Schematic bedset (metres thick), which is composed of multiple sedimentation events that generally fine and thin upwards. D) Schematic unit (10s m thick) that generally fine and thin vertically (indicated by grey arrows), which is made up of numerous bedsets that also fine and thin, and represent the largest stratigraphic element discussed here.

Strata were classified using Folk's classification scheme (Folk *et al.*, 1970; Folk, 1980) based on the relative proportion of quartz, feldspar, and rock fragments (*Fig. 3.3A*). Folk's (1980) textural classification was used to describe the different facies (*Fig. 3.3B*). Point counting in thin section (see discussion in 1.3) was used to quantitatively classify the texture of each facies (30 points were measured vertically in each thin section).

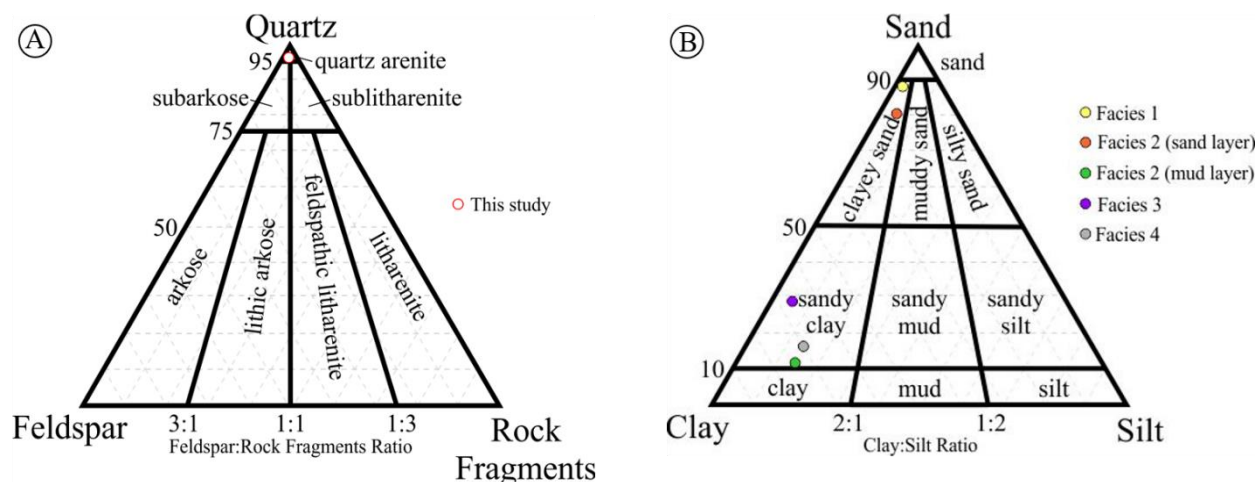


Figure 3.3: A) Sandstone classification based on the relative proportion of quartz (including metaquartzite), feldspar (including granite and gneiss fragments), and all other rock fragments (including chert) (redrawn after Folk *et al.*, 1970; Folk, 1980). Strata of all facies in the study area are mineralogically similar and according to the Folk (1980) classification are termed quartz arenite. B) Textural classification of fine-grained strata based on the relative proportion of sand (0.0625-2 mm), silt (3.9-62.5 μm), and clay (< 3.9 μm) from vertically point counting 30 points in thin section (redrawn after Folk, 1980).

3.2 Metamorphic Fabric

Despite low-grade greenschist facies metamorphism (200-450 °C) primary physical sedimentary structures are well preserved in strata in the study area (Murphy, 1987; Zou *et al.*, 2013), and therefore terminology applied to sedimentary rocks will be used to describe these strata. Metamorphic alteration has transformed detrital clay mineral particles, possibly kaolinite and/or smectite based on common diagenetic clay mineral reaction and transformations (Ulmer-Scholle *et al.*, 2014), and also paleogeographic reconstructions suggesting the Windermere turbidite system accumulated off the equatorial coast of western Laurentia (proto-North America) (Li *et al.*, 2013), to chlorite and mica (muscovite and sericite). Accordingly, this has modified the size and mineralogy of the (primary) detrital grains, and therefore all interstitial material, regardless of crystal size, is considered, along with unaltered silt-sized siliciclastic quartz, to be matrix.

Strata in the Castle Creek study area have undergone dynamic quartz recrystallization, which is a common microstructural transformation that occurs during deformation and metamorphism (Stipp *et al.*, 2002). Three types of dynamic quartz recrystallization have been

described based on increasing temperature of deformation and strain rate: 1) bulging recrystallization, 2) subgrain rotation, and 3) grain-boundary migration (Stipp *et al.*, 2002). Dynamic quartz recrystallization reduces the strain on the crystal lattice and in the case of quartz begins with bulging recrystallization at temperatures from 250-400 °C where grain boundaries bulge (up to 30% of grains) into highly strained areas until the bulge separates from the parent grain forming many recrystallized unstrained silt to fine sand-sized grains along grain boundaries and microfractures (*Fig. 3.4*) (Stipp *et al.*, 2002). Subgrain rotation initiates between 400-500 °C where grains geometrically soften (30-90% of grains) allowing for rotation and the creation of medium to coarse sand with relatively straight grain boundaries and an absence of undulatory extinction (Stipp *et al.*, 2002), which contrast the sutured boundaries and undulose extinction in metamorphic quartz grains. Lastly, grain-boundary migration occurs at temperatures from 400-500 °C where grain boundaries expand (100% of grains) creating very coarse sand to granule quartz grains (Stipp *et al.*, 2002). Owing to the low temperatures of greenschist facies metamorphism (200-450 °C) (Zou *et al.*, 2013), only bulging recrystallization (most common) and subgrain rotation are important in this study. Also, recrystallization tends to be more intense in framework-supported strata with low clay (matrix) content because the strain is accommodated by deformation and recrystallization of the rigid framework grains rather than slip in the fine-grained matrix (Drury and Urai, 1990).

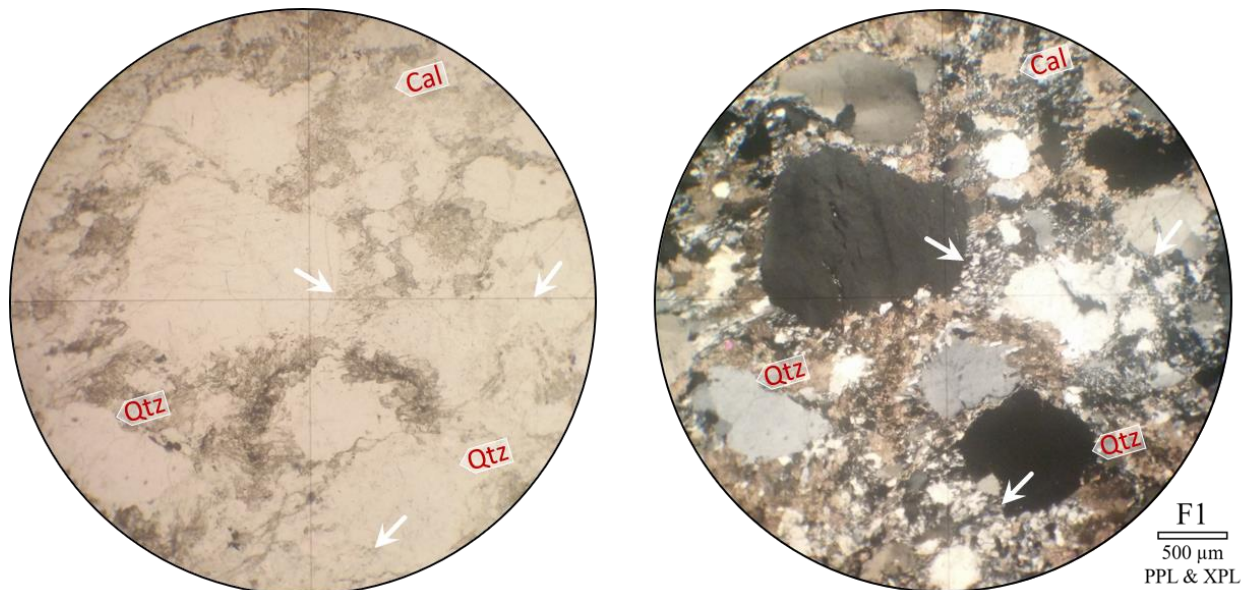


Figure 3.4: Plane light (left) and cross-polar (right) photomicrographs showing recrystallized quartz (indicated by the white arrows) formed by bulging recrystallization between framework (detrital) quartz grains (*sensu* Stipp *et al.*, 2002). Beds that are coarser-grained (> 0.25 mm, medium sand) and framework-supported typically have more recrystallized quartz at grain-to-grain contacts due to locally elevated temperature, strain, and strain rate.

3.3 General Petrographic Characteristics

The mineralogy of the framework grains is almost entirely quartz ($> 99\%$), which range from angular to sub-rounded, equant to elongate, and spherical to non-spherical. However, it is important to note that detrital grain textures (*i.e.* shape, rounding, and sphericity) have been metamorphically altered and therefore may not necessarily reflect original textures. Quartz grains are mostly monocrystalline with minor ($< 5\%$) polycrystalline grains exhibiting preferential elongation of the grain long axis parallel to bedding and undulatory extinction of the individual sub-crystals. Monocrystalline quartz grains also display sweeping undulose extinction unless they have experienced dynamic quartz recrystallization via bulging recrystallization or subgrain rotation (see discussion in 3.2), in which case, parallel extinction is observed. Strata with a higher framework grain content tend to have more recrystallized quartz grains around grain boundaries (Fig. 3.4), in addition to sutured boundaries and overgrowths due to strain-induced pressure dissolution of silica at grain-to-grain contacts. Overgrowths and recrystallization have altered the

grain size of many quartz grains, which is important to take into consideration when analyzing grain size data as grains will be slightly coarser than at the time of deposition. All grains > 0.25 mm (medium sand) have abundant vacuoles (micron-sized fluid-filled inclusions) and uncommon Boehm lamellae (multiple bands of subparallel, planar vacuoles) that tend to be aligned, and therefore suggesting a high-stress regime (Ulmer-Scholle *et al.*, 2014), and probably formed during Mesozoic deformation as the Boehm lamellae in most detrital quartz grains are similarly orientated indicating similar stress fields during formation. The remaining detrital framework grains (< 1%) are composed of accessory minerals such as biotite, muscovite, orthoclase, plagioclase, rutile, and zircon. Detrital micaceous minerals (biotite, muscovite) are foliated and are observed to have undergone ductile mechanical deformation, which is manifest as kinked foliation and varying birefringence when sandwiched between stronger, more rigid grains like quartz. Some chlorite in matrix-poor strata could also be the result of alteration of detrital biotite (Ulmer-Scholle *et al.*, 2014). However, it is difficult to differentiate chlorite formed by alteration or metamorphism, and thus it is likely underestimated. Feldspars (orthoclase, plagioclase) are highly degraded, and in some cases partially altered to sericite or calcite. Some orthoclase grains appear cloudy due to abundant vacuoles related to alteration.

The mineralogy of the matrix is slightly more diverse than the framework grains where detrital clays have been altered to chlorite, muscovite, and sericite. Silt-sized (< 62.5 μm) grains of quartz (9% of total normalized point counted matrix) are also observed and display parallel extinction. This, then, makes it challenging to differentiate detrital silt from (bulging) recrystallized quartz grains, and therefore is probably underestimated in point count data. SEM showed that silt-sized quartz is pervasive throughout all strata in the study area, even at the very top of structureless claystone caps (see discussion in 3.6.2; *Fig. 3.24*). Silt grains are angular to

subangular, elongate to bladed, and non-spherical. Metamorphic chlorite (17% of total normalized point counted matrix) recrystallized from detrital clay forming small ($< 62.5 \mu\text{m}$) crystals in matrix-poor strata and increasingly larger porphyroblasts (up to coarse sand, 0.5-1 mm) in strata with higher matrix content. Mechanical compaction has resulted in bedding-parallel realignment of quartz and recrystallized chlorite and mica grains in matrix-rich strata, whereas recrystallized matrix in matrix-poor strata shows no preferred fabric (*e.g. Fig. 3.13*). Additionally, chemical compaction in matrix-rich strata has produced numerous bedding-parallel, thin ($< 150 \mu\text{m}$), dark, wispy, and elongate solution seams (microstylolites) that contain elevated amounts of insoluble materials like detrital clays, iron oxides and/or sulphides. Lastly, metamorphic micas (74% of total normalized point counted matrix) comprise muscovite and sericite forming small ($< 3.9 \mu\text{m}$), tabular crystals.

Thin sections were stained with Alizarin Red S and potassium ferricyanide to help differentiate calcite from dolomite and ferroan calcite or dolomite, respectively (see discussion in 1.3) (Dickson, 1966). Carbonate cements make up 15% of the total normalized point counted constituents and can be as high as 34% in some coarse-grained, matrix-poor strata. Two different carbonate cements were identified: calcite (13%) and dolomite (2%); both were determined to be ferroan based on chemical staining and XRD analysis (see discussion in 1.3). However, where dolomite is present, there is little calcite cement and vice versa (*e.g. Fig. 3.7A vs. 3.7B*). Medium to coarse crystalline sparry calcite cement is prominent (up to 27%) in coarse-grained ($> 0.125 \text{ mm}$, fine sand) strata, while minor ($< 5\%$) very fine to fine crystalline micrite is observed in matrix-poor finer-grained ($< 0.125 \text{ mm}$) rocks. Sparry calcite cement is often equant and translucent with varying levels of brown inclusions under plane-polarized light. Locally, some coarser ($> 0.25 \text{ mm}$) crystals exhibit well-developed twinning and poikilotopic fabric (*Fig. 3.7A*). Sparry calcite cement

is so abundant in some coarse-grained rocks that it fills the intergranular space between framework grains that appear to float in the cement. Micrite tends to fill intergranular pore space forming small, brown-coloured patches of randomly orientated acicular calcite. Less commonly, dolomite forms small (< 0.5 mm) clusters of dark brown, rhombohedral crystals and rare dispersed medium rhombohedral crystals. Most dolomite crystals appear to have undergone dedolomitization where the dissolution of dolomite leaves a relic outline (ghost boundary) that then is filled with calcite.

Lastly, rare diagenetic and metamorphic pyrite forms < 1% of total normalized point counted minerals. Pyrite tends to be concentrated in finer-grained rocks where it makes up to 11% of the mineralogy in a sample. Pyrite forms small, opaque framboids or large (0.1 to > 1 mm), cubic to subhedral crystals. These larger crystals are often mantled by fibrous intergrown quartz that form pressure shadows. See Appendix 1 for more detailed annotated photomicrographs and descriptions of some of the features discussed in this section.

3.4 Facies 1: Massive or graded, structureless or planar/wavy-stratified clayey sandstone

3.4.1 Macroscopic Characteristics

Facies 1 (F1) comprises structureless or planar- or wavy-stratified clayey sandstone (*Fig. 3.6*) and is observed in 12% of beds. Strata are characteristically carbonate cemented, which imparts a distinctive rusty orange colour in the field (see discussion in 3.4.2). Strata range in thickness from 2-112 cm (average 18 cm, mode 13 cm) (*Fig. 3.5*) and where present makes up 16-100% of the bed thickness (average 57%). Where present, F1 is always the basal division of a composite bed and only rarely (4%) makes up an entire bed. F1 is overlain abruptly by Facies 2 (54%) or Facies 3 (42%). However, the contact between F1 and F2 or F3 becomes more gradational as the grain size and thickness of F1 decreases and becomes more similar to the overlying strata

(Fig. 3.5B). Rarely, thick (12-77 cm) beds are composed of only F1. The basal contact is usually sharp and planar (58%), although uncommonly (21%) it undulates (1-2 m spacing and 1-2 cm height) or is loaded (21%) with load casts and flame structures.

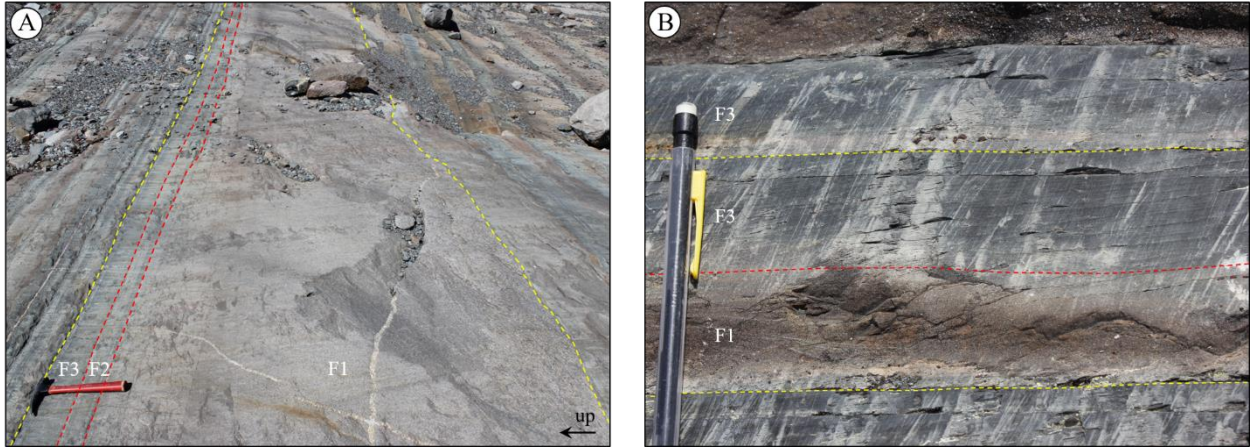


Figure 3.5: Facies 1 sandwiched between yellow and red dashed lines. A) Very thickly-bedded (112 cm), structureless sandstone (hammer handle at the top of F1 portion of the bed). B) Thin-bedded (3.5 cm), structureless sandstone (below red dashed line). Dashed yellow lines bound beds at the top and bottom, whereas the red lines separate the different facies that make up a bed.

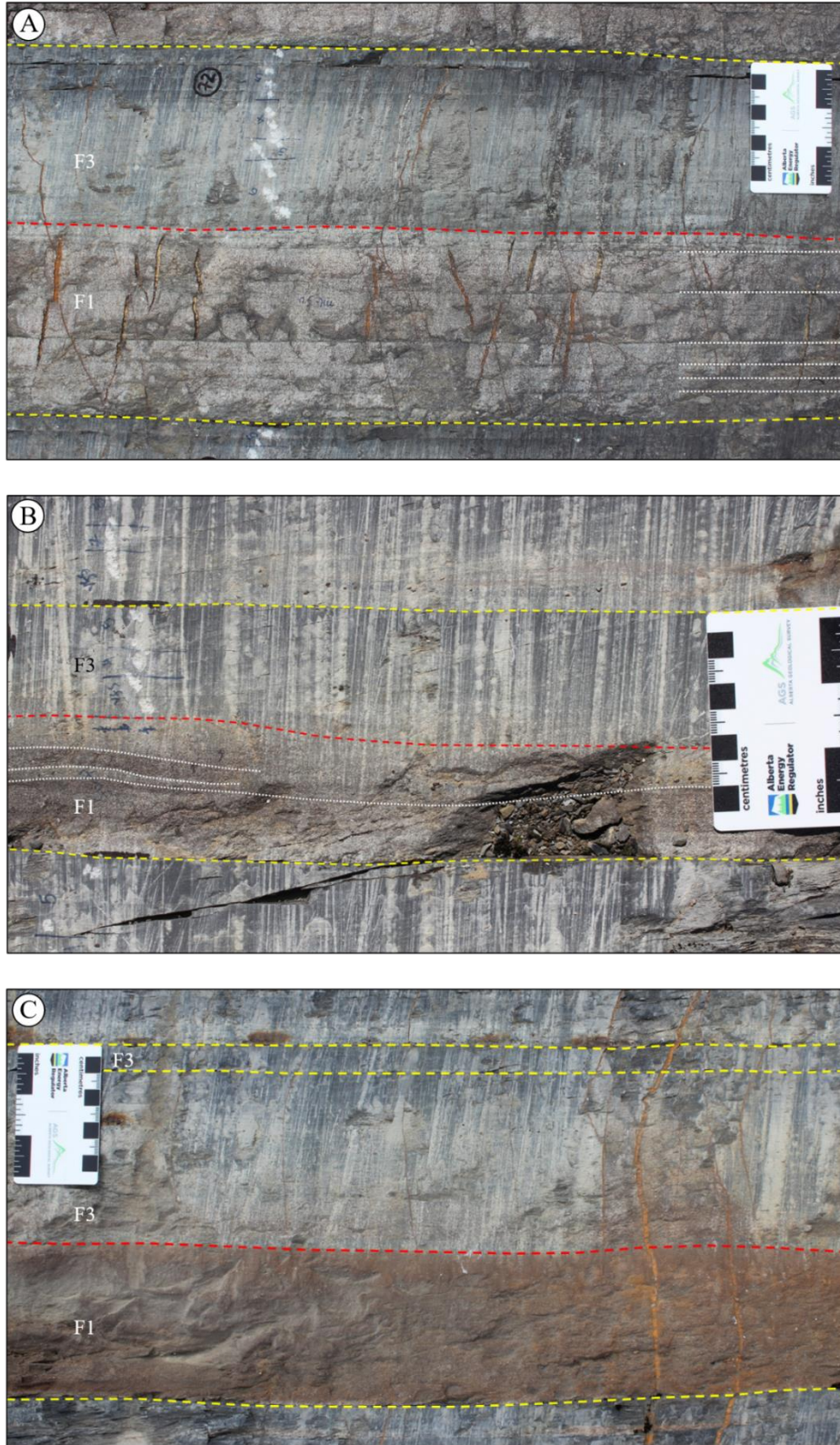


Figure 3.6: Structureless, graded or massive F1 sandstone (top of F1 indicated by dashed red line; base of individual bed indicated by dashed yellow lines). A) Planar-stratified sandstone (stratification highlighted by white dotted lines on right side of image). B) Wavy-stratified sandstone. Note the white dotted lines that highlight the wavy-stratification. C) Structureless, coarse-tail graded sandstone.

F1 strata are mostly massive (76%) and the rest (24%) are normally graded. Additionally, beds are structureless (45%), planar (52%), or wavy (3%) stratified (*Fig. 3.6*). Normally graded beds range in grain size from very coarse to very fine sand and most commonly grade from medium to fine sand (33%), followed by coarse to fine sand (24%), coarse to medium (10%), medium to very fine (10%), fine to very fine (10%), and very coarse to fine (5%). Dispersed anomalously coarse grains are observed near the base of 14% of beds, specifically coarse sand is common (60%), followed by very coarse sand (20%), medium sand (13%), and granules (7%).

3.4.2 Microscopic Characteristics

In the Folk (1980) classification F1 is classified as a clayey sandstone based on point counts in 11 samples after normalizing the sand, silt, and clay percentages. On average this facies comprises approximately 58% framework grains, 8% matrix (muscovite, chlorite, sericite, silt-sized quartz), but always < 10%, and 34% carbonate cement (calcite, dolomite). The coarser-grained (> 0.25 mm, medium sand) bed divisions are typically poorly sorted, while finer-grained (< 0.25 mm) beds are moderately sorted (*Fig. 3.7*). Framework grains are usually sub-angular to sub-rounded and elongate to equant. Where cement content is low (< 10-15%), strata tend to be framework supported. The average grain size of the framework grains is commonly medium sand (32%) followed by coarse (22%), fine (21%), very coarse (18%), and very fine sand (7%). At the base of some beds, sand grains shallowly (*c.* 1 mm) penetrate the fine-grained cap of the underlying bed.

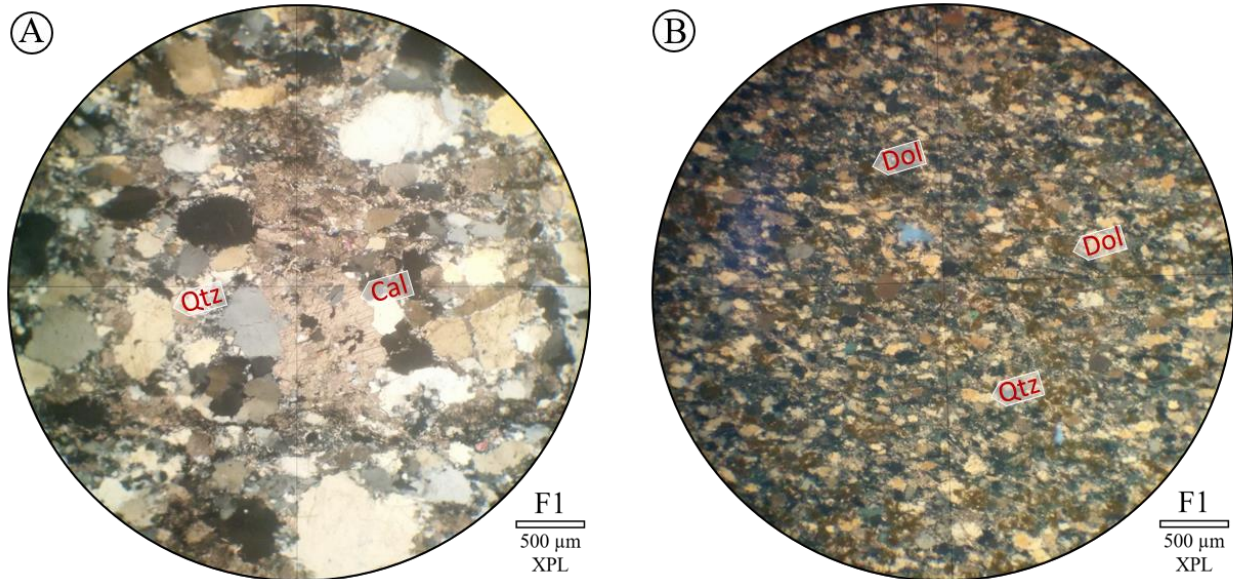


Figure 3.7: Photomicrographs illustrating the range in grain size of F1 strata. Both samples have about the same amount of carbonate cement (c. 25%). In (A) the cement occurs as large calcite (Cal) crystals that locally exhibit a poikilotopic fabric, whereas in (B) it occurs as smaller dispersed patches of brown rhombohedral dolomite (Dol) and minor (< 5%) calcite. A) Very coarse- to medium-grained clayey sandstone with negligible (< 2%) matrix (recrystallized chlorite and micas). B) Fine-grained sandstone with about 5-10% matrix content.

Quartz dominates the mineralogy of the framework grains (98%) that display undulatory extinction under cross-polarized light. Detrital biotite, muscovite, orthoclase, plagioclase, rutile, and zircon make up the remaining framework constituents (each < 1%). Fine-grained (< 62.5 μm) recrystallized quartz grains tend to form along the edges of larger (> 0.25 mm) quartz grains and display parallel extinction (*Fig. 3.4*). Micas form small intergranular patches or are rarely grain-rimming where they are sandwiched between coarser (> 0.25 mm) sand grains or enveloped by calcite cement.

The intergranular space is filled primarily with carbonate cement (81%), specifically sparry calcite (85%) and ferroan dolomite (15%). The matrix comprises 90% mica (muscovite and sericite), 5% silt-sized quartz, and 5% chlorite that typically form sand-sized porphyroblasts.

3.4.3 Interpretation

Facies 1 bed divisions are interpreted to reflect deposition under high sedimentation rates from rapidly decelerating high-density, sand-rich turbidity currents (Lowe, 1982) or concentrated to hyper-concentrated density flows (*sensu* Mulder and Alexander, 2001). Strata of F1 are equivalent to the Bouma T_a or T_b divisions. A combination of fluid turbulence, grain-to-grain interactions, and buoyancy effects are the principal particle-support mechanisms operating in these flows. The general lack of tractional structures (45% F1 beds structureless) suggests high rates of suspended sediment deposition (see also Arnott and Hand, 1989; Leclair and Arnott, 2003). The poorly sorted nature of F1 further supports rapid deposition from flows with a high sedimentation rate and near-bed concentration that hinders settling impeding the segregation of grains (Banerjee, 1977; Sumner *et al.*, 2008). However, if the aggradation rate is reduced, planar stratification develops as observed in some F1 beds (Arnott and Hand, 1989, Leclair and Arnott, 2005). This planar stratification is slightly better sorted with subtle grain size change across individual lamina, although the upper and lower bounding surfaces are diffuse (*sensu* Sumner *et al.*, 2008). Based on experiments, planar-stratification, and even slightly wavy-stratification, develops from crudely sorted laminar shear layers in decelerating flows (Sumner *et al.*, 2008).

The basal contact is commonly sharp and planar indicating near-bed turbulence suspension. Uncommonly, the basal contact is locally undulatory with 1-2 cm relief indicating localized sea floor scours resulting from erosion by turbulent eddies, or alternatively, post-depositional (gravitational) soft sediment deformation, namely loading (*i.e.* density inversion), which indicates a short time interval between successive depositional events.

F1 is abruptly overlain by finer-grained, traction-structured deposits (*i.e.* F2 and F3) suggesting a period of non-deposition and sediment bypass. As discussed in 2.3.2.3, turbidity

currents are longitudinally stratified with coarser sediment concentrated in the head of the flow and deposits the basal part of the bed, whereas finer-grained sediment is displaced backward towards the tail and deposits the upper part of the bed (Arnott, 2010).

Burial diagenesis includes carbonate cementation and pyrite recrystallization. Calcite cement was likely recrystallized from detrital carbonate shed from the flooded shelf during transgressive or highstand conditions and transported to the deep sea by sediment-gravity flows, while the rhombic ferroan dolomite likely formed during diagenesis and metamorphism as it appears to replace minerals. The presence of sulphide minerals, namely pyrite, indicates shallow burial bacterial reduction under anoxic/euxinic basin conditions (Ross *et al.*, 1995). These processes of carbonate cementation and pyrite recrystallization likely occurred in similar strata in the study area (F2-F4) where carbonate cement and/or pyrite is also present.

3.5 Facies 2: Graded, plane parallel-, cross- or wavy-stratified clayey sandstone with intercalated structureless sandy claystone

3.5.1 Macroscopic Characteristics

Facies 2 (F2) is composed of rhythmically interstratified sand-mud layers where planar- (F2.A), cross- (F2.B) or wavy-stratified (F2.C; subfacies described below) clayey sandstone is abruptly overlain by structureless sandy claystone (*i.e.* mud drapes) (*Fig. 3.8*). F2 is observed in 26% of beds. These strata range from 0.5-34.5 cm thick (average 5.8 cm, mode 2.8 cm) (*Fig. 3.9*) and where present make up 4-88% of the bed thickness (average 44%). F2 either sharply overlies F1 (24%) or the basal contact of the bed (76%). In the latter, basal contacts are usually sharp and planar (87%), although uncommonly it locally undulates (1-2 m laterally) with 1-2 cm of relief (10%) or are loaded with load casts and flame structures (3%). F2 is typically overlain by F3 (99%) or rarely (1%) forms the top of the bed.

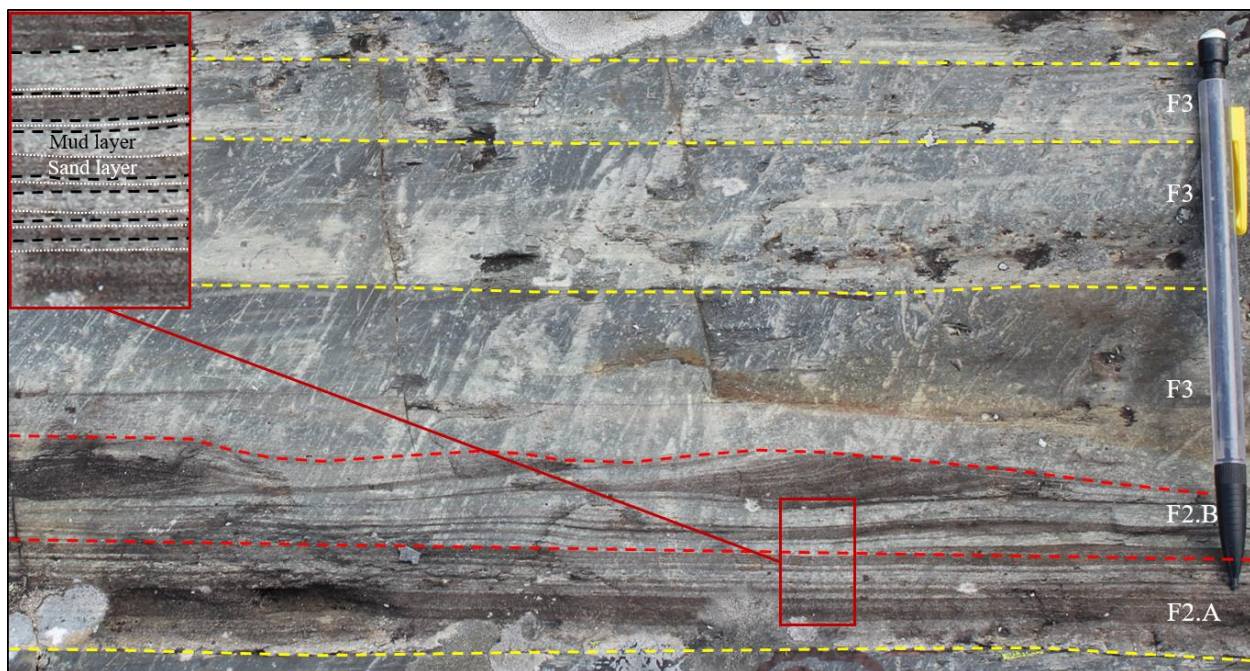


Figure 3.8: Facies 2 showing a transition in the style of stratification, which stratigraphically upward consists of planar-stratified (F2.A) to low then moderate amplitude isolated cross-stratified (F2.B) sandstone. All forms of stratification are draped by structureless sandy claystone (light grey strata) forming intercalated sand-mud layers (indicated in insert in top left corner where sand-mud layers are separated by black dashed lines and the contact between the two components is denoted by the white dotted lines; the whole succession represents rhythmically stacked sand-mud layers). Yellow dashed lines indicate bed contacts and the red dashed lines indicate the contact between different facies or subfacies.

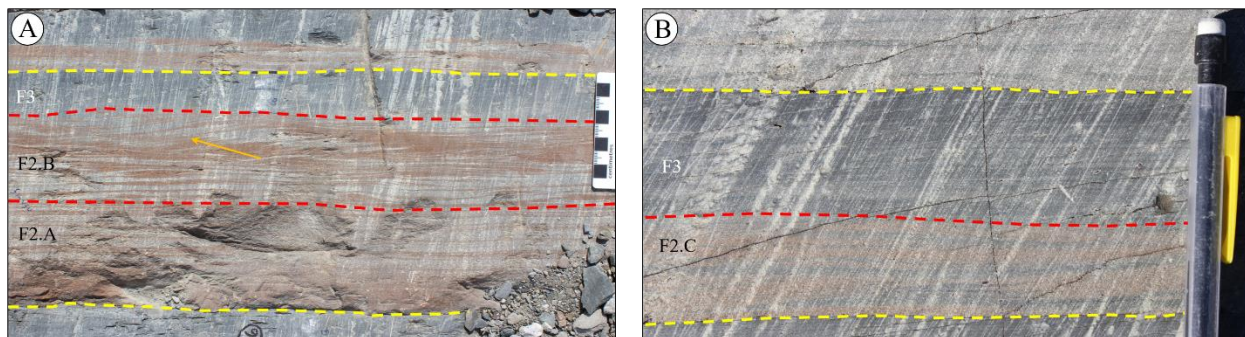


Figure 3.9: Thickness variation in Facies 2 strata (between red and yellow dashed lines). A) Medium bedded (15 cm thick), carbonate cemented (sand layers here are a rusty orange colour), planar- to cross-stratified with mud troughs overlain by climbing ripple cross-stratification (indicated by orange arrow) in the F2.B division of the bed. Note that the sand-rich crest of ripple foresets grade obliquely-downward into mud in the bottomset, forming discontinuous mud layers commonly termed mud flasers. B) Very thin bedded (2 cm thick), carbonated cemented, wavy-laminated Facies 2 interstratified with mud flasers. Bed contacts are indicated by yellow dashed lines and red dashed lines separate different facies.

F2 is commonly normally graded throughout its thickness (*Fig. 3.10*). Normally graded beds range from medium to very fine sand but most commonly grade from fine to very fine sand

(79%), or medium to fine sand (11%), or medium to very fine sand (10%) (see discussion in 3.5.2 for grain size distribution in graded beds). Grading is easily observed in the field in coarser-grained sand layers, and more subtly microscopically in dispersed sediment ($> 62.5 \mu\text{m}$) within fine-grained mud layers. Where F2 has appreciable amounts ($> 10\%$) of calcite cement it is easily recognized in the field by the rusty orange colour in sandier strata related to the oxidation of ferroan carbonate cement (*Fig. 3.8, 3.9*) (observed in 10 of 11 point counted samples). Most characteristically, F2 is dominated by traction-transport sedimentary structures, namely wavy-stratified (55% of beds) (sinuous ripple lamination *sensu* Jopling and Walker (1968) or lenticular bedding with thick or flat lenses to wavy flaser bedding (*Fig. 3.11*) *sensu* Reineck and Wunderlich (1968) in tidally-influenced deposition), planar-stratified (34%) or cross-stratified (11%) clayey sandstone (*i.e.* sand layer) overlain abruptly by massive sandy claystone (*i.e.* mud layer) (*Fig. 3.8*).

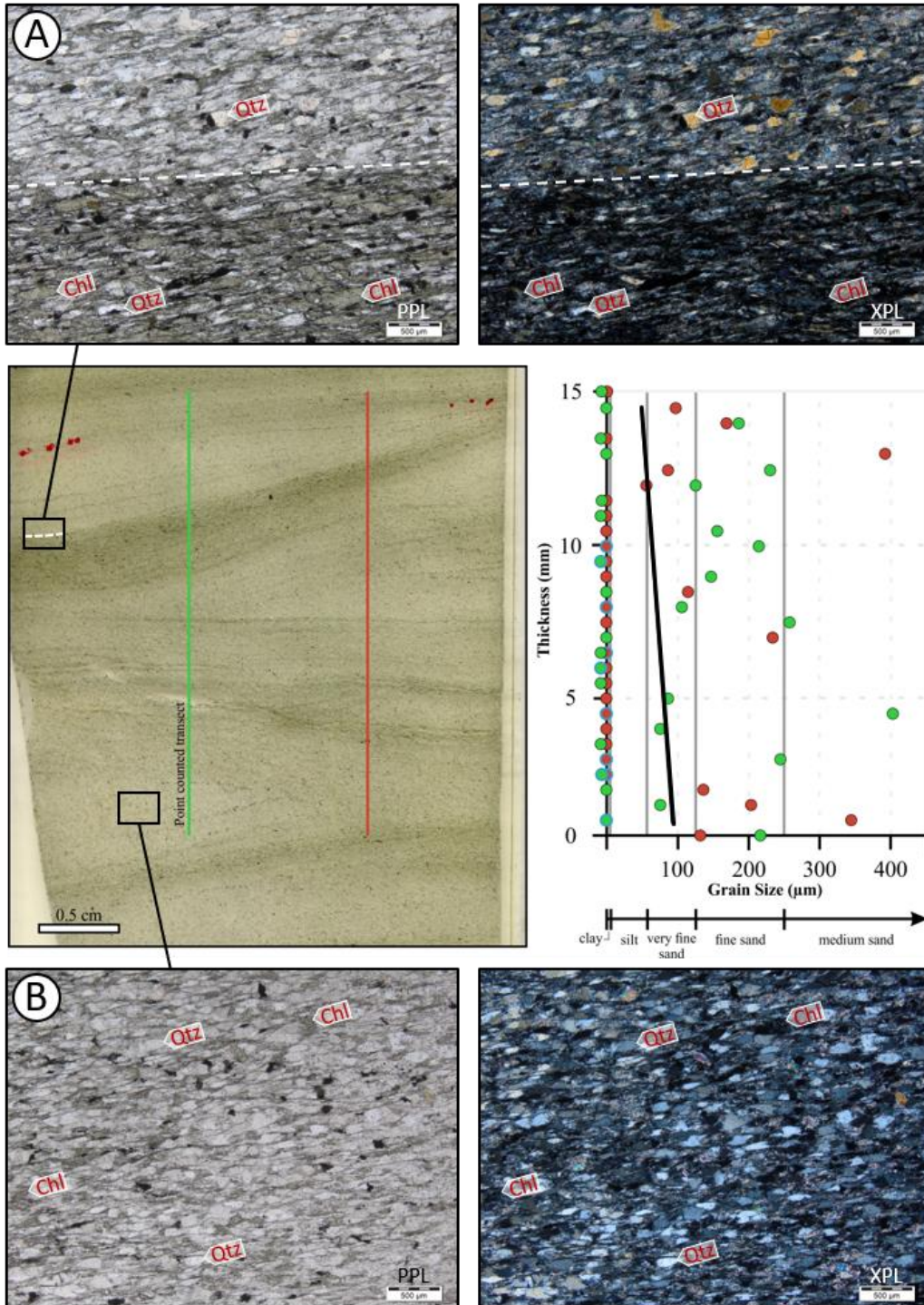


Figure 3.10: Grain size analysis (middle right) along two vertical transects (green and red lines) in a thin-bedded F2 (F2.B) division. Grain size (green and red dots) was measured every 0.5 mm over a vertical length of 15 mm (total of 30 points). Although somewhat muted due to the poorly sorted nature of the strata, grain size does show a subtle upward fining (indicated by the black trendline). Note that data points lying on the vertical axis represent recrystallized clay or carbonate cement and are not considered framework grains. The bottom photomicrographs (B) are taken in a sand layer, and top photomicrographs (A) depict the contact (separated by the white dashed line) between a mud layer overlain sharply by a sand layer.

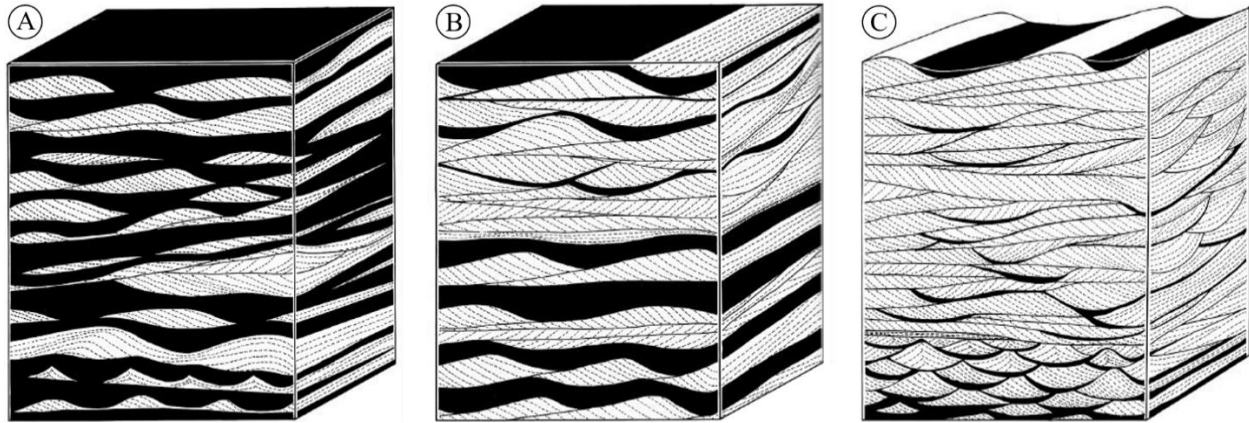


Figure 3.11: Deposits with varying sand and mud content, respectively (A) lenticular (25:75), (B) wavy (50:50), and (C) flaser (75:25). A) Lenticular bedding is composed of discrete sand ripples encased in mud. B) Wavy bedding, an intermediate form, is made up of roughly equal proportions of sand:mud where ripple formsets typically form a single set and are encased in mud. C) Flaser bedding is characterized by thin, isolated mud drapes, typically in the troughs of ripples, interstratified with multi-set ripple cross-stratified sand (modified from Holland, 2018, adapted from Reineck and Wunderlich, 1968).

3.5.2 Microscopic Characteristics

Facies 2 is a clayey sandstone (point counted in 11 samples after normalizing the sand, silt, and clay content) (Folk, 1980). This facies is typically 44% framework grains, 36% matrix (muscovite, chlorite, sericite, silt-sized quartz), although matrix can range from 10-60%, and 20% carbonate cement (calcite, dolomite). Framework grains are usually sub-angular to sub-rounded and equant to elongate (*Fig. 3.12*). Strata with more framework grains tend to exhibit more recrystallization around the margins of quartz grains. The average grain size of the framework grains is fine sand (59%) followed by very fine (27%) and medium sand (14%). At the base of some beds sand grains slightly (*c.* 1 mm) penetrate into the fine-grained cap of the underlying bed.

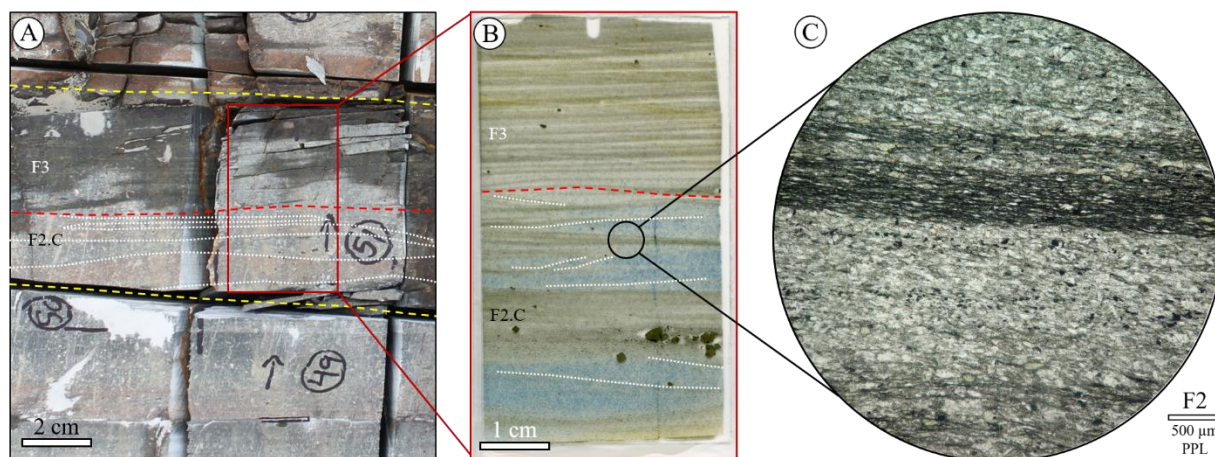


Figure 3.12: A) Wavy-stratified (F2.C), very thin-bedded F2 with coarser-grained, carbonate cemented (rusty orange strata) interlaminae each draped with structureless claystone (dark grey strata). Note that white dotted lines separate the sand and mud layers. F2 is then overlain gradationally by F3 (above red dashed line). B) F2 (below the red dashed line) with low-angle cross-lamination in coarser-grained sand layers that have been stained blue with Alizarin Red S and potassium ferricyanide (see discussion in 3.3). The basal part is very thin bedded (> 1 cm) and coarser grained (> 0.125 cm, fine sand) but then overlain by laminae that gradually fine and thin upward until they transition into very thin (pinstripe) lamina (< 3 mm) in F3 (above red dashed line). C) Photomicrograph of a single interlaminated clayey sandstone (sand layer) draped abruptly with sandy claystone (mud layer). Note that the size of grains dispersed in the mud layer is a mixture of the modal grain size in the underlying sand layer as well as silt-sized grains.

Undulatory quartz dominates the mineralogy of the framework grains (100%). Fine-grained ($< 62.5 \mu\text{m}$) recrystallized quartz grains tend to form along the edges of larger (> 0.25 mm) quartz grains and display parallel extinction. Grain-rimming micas surround coarse (> 0.25 mm) sand grains and are surrounded by calcite cement, while also forming small intergranular patches. The matrix consists mostly of mica (muscovite and sericite) (average 76%), chlorite (20%), pyrite (3%), and minor silt-sized quartz (1%). The chlorite typically forms sand-sized porphyroblasts. The cement comprises sparry calcite (98%) with rare rhombohedral ferroan dolomite (2%) and fills the intergranular space in sandier layers.

F2 is composed of intricately interstratified millimetre- to centimetre-thick, sand-rich (sandy layer) and sand-poor (muddy layer) layers that show no systematic change in thickness upward. The sandier layers are poorly sorted, very thinly bedded to thinly-laminated and consists of traction-structured clayey sandstone whereas the muddier layers are thick- to thinly-laminated and made of structureless sandy claystone (Fig. 3.12-3.14). From bed to bed sandy layers vary the

most in their matrix content (10-60%), which within a F2 division in a bed tends to remain consistent both laterally and vertically and are subdivided into two end-members: matrix-poor (10-30% matrix) and matrix-rich (30-60% matrix). However, rarely (2%) matrix-poor strata are overlain sharply by matrix-rich strata (*e.g. Fig. 3.16C* below). Additionally, matrix-poor strata contain 15-35% ferroan calcite cement that oxidizes to a distinctive rusty orange colour in the field. In contrast, muddy layers lack carbonate cement and are grey. Sand layers uncommonly exhibit subtle normal grading and generally consist of fine-grained sandstone commonly with dispersed medium sand grains (up to coarse sand). In some sand layers, quartz grains gradually become better aligned with bedding as it changes upward into the muddier layers. Muddy layers are massive, moderate to well sorted, and composed primarily of recrystallized clay and (quartz) silt grains with 15-30% dispersed medium to very fine sand. Chlorite porphyroblasts, in addition to the other recrystallized matrix (micas), are preferentially aligned parallel to bedding in muddier layers (*Fig. 3.13*). Significantly, the range of grain sizes in a sandy and overlying muddy layer is similar, and stratigraphically upward show an overall upward fining (*Fig. 3.10*).

F2 is further subdivided into three subfacies based on the type of sedimentary structure(s) deposited: planar-stratified (F2.A), cross-stratified (F2.B), and wavy-stratified (F2.C). Most commonly (81%) the F2 division consists of a single style of stratification (F2.A, F2.B, or F2.C). However, beds with two, and rarely (< 1%) three styles of stratification are observed; specifically, cross-stratified (F2.B) overlain by wavy-stratified (F2.C); plane-parallel stratified (F2.A) overlain by wavy-stratified (F2.C); or plane-parallel stratified (F2.A) overlain by cross-stratified (F2.B) overlain by wavy-stratified (F2.C) (see also *Fig. 3.15*). Where F2 overlies a planar-stratified F1 it too is planar-stratified (F2.A), whereas where F1 is absent and F2 strata overlie the basal contact, F2 strata are typically wavy-stratified.

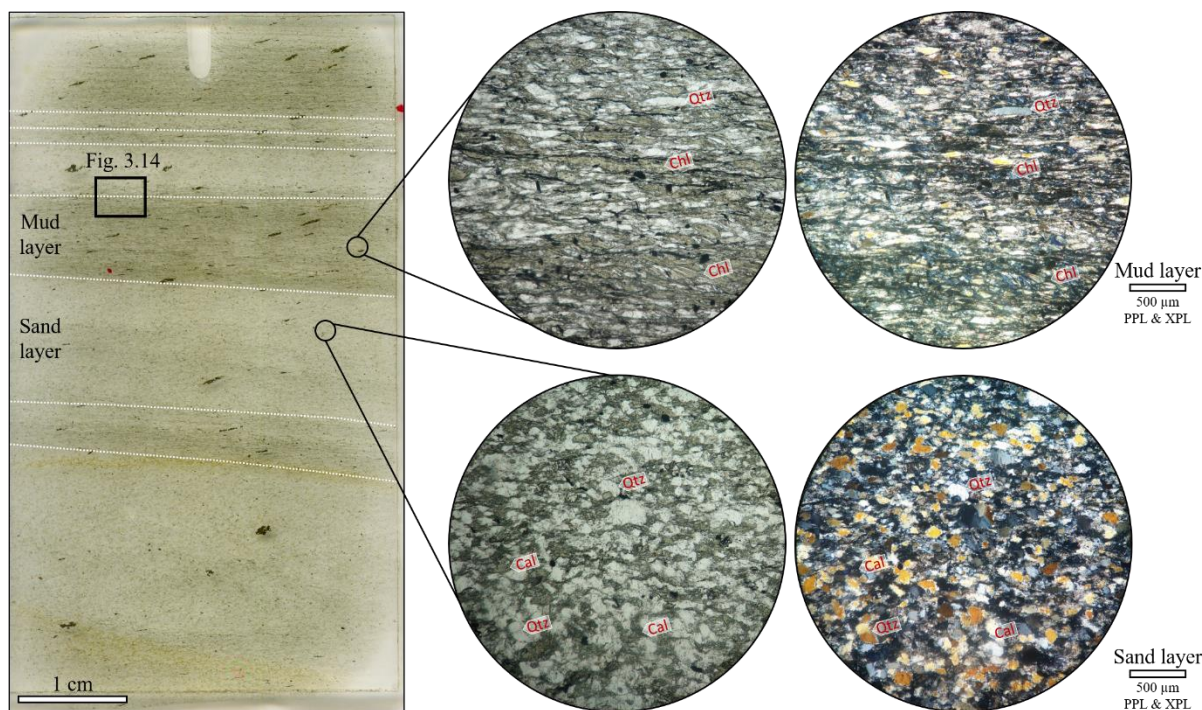


Figure 3.13: Comparison of F2 interstratified coarser-grained sand layers and finer-grained mud layers (darker laminae) in thin section and photomicrograph. Note the difference in abundance and fabric of the chlorite (Chl) porphyroblasts (proxy for recrystallized detrital clay) in the sand versus mud layers. In mud layers intergranular chlorite and micas are more abundant and generally larger; chlorite crystals ($c. > 100 \mu\text{m}$) display sigmoidal crenulation cleavage (product of metamorphism and successive deformation) and are preferentially aligned parallel to bedding. In sand layers intergranular chlorite and mica crystals are rare (in this example $< 1\%$), much smaller, and randomly aligned. Quartz (Qtz) grains are also preferentially aligned in finer-grained mud layers and randomly orientated in coarser-grained sand layers. There is also a compositional difference as sand layers have much more carbonate cement ($> 20\%$ on average; Cal), whereas adjacent mud layers contain significant matrix ($c. > 80\%$) and negligible carbonate cement.

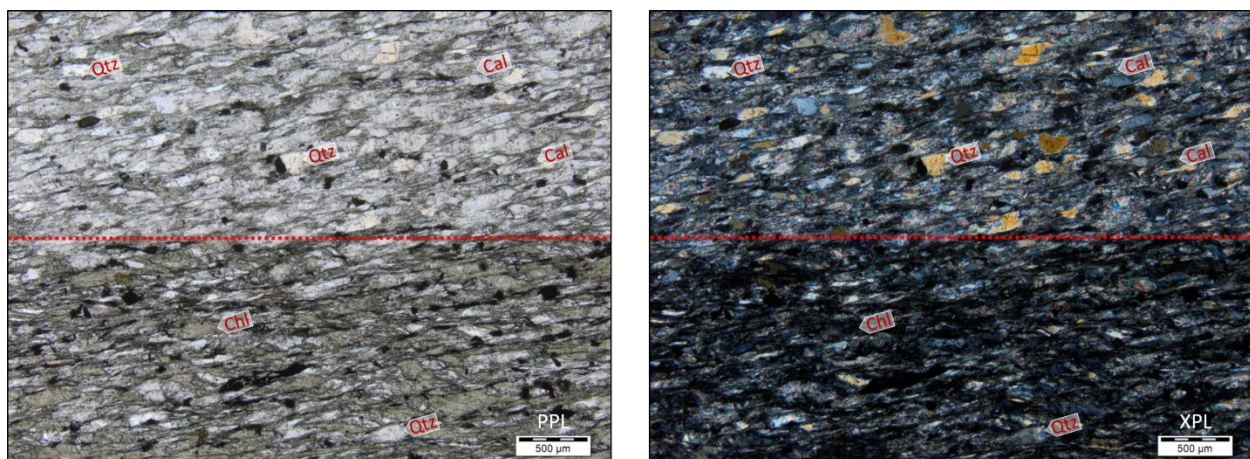


Figure 3.14: Photomicrographs (plane- and cross-polarized light, respectively) of the sharp contact (red dotted line) between fine-grained mud lamina overlain by a coarser-grained, very thin-bedded sand lamina (location of photomicrograph outlined in Fig. 3.13). Note the more abundant chlorite (Chl) and the parallel alignment of quartz (Qtz) grains and chlorite (Chl) porphyroblasts in the mud layer, whereas the sand layer has more calcite (Cal) cement.

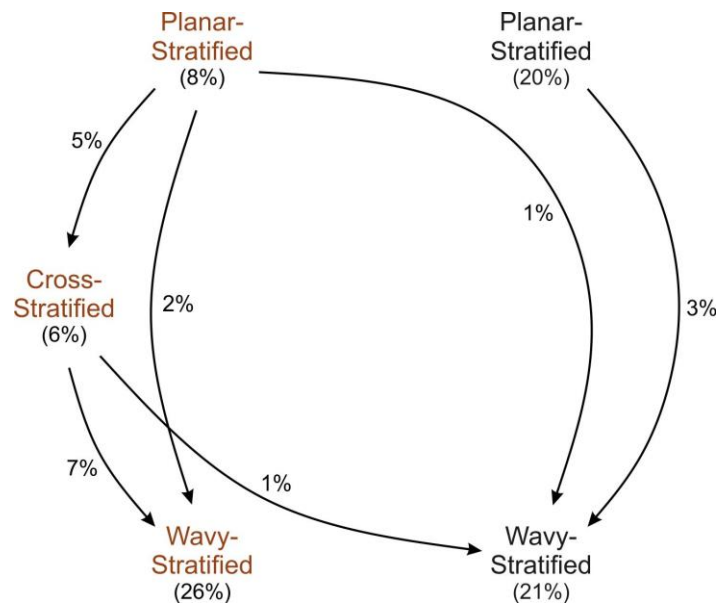


Figure 3.15: Flow chart of the common styles of stratification observed in matrix-poor, carbonate cemented (orange text), or matrix-rich (black text) F2 strata, and the corresponding stacking patterns (indicated by arrows) where two or more styles of stratification stack to form a single F2 division.

Plane-parallel F2.A strata are typically rich in matrix (62% of F2 divisions) and the sand layers are grey, whereas less common matrix-poor (38%) sand layers are calcite cemented and are distinctively orange colour (*e.g.* Fig. 3.9, 3.16). Matrix-rich F2.A strata tend to be thinner (average 7.8 cm) and laminated with sharper, more distinct boundaries separating the interstratified sand-mud layers that range in thickness from 0.1-1.5 cm, but more commonly are < 1 cm. Matrix-poor strata, on the other hand, tend to be thicker (average 9.2 cm) and sand and mud interlayers are thicker (0.2-3 cm, commonly > 1 cm) and bounded by more diffuse boundaries. Average grain size for F2.A is upper fine sand, and generally is coarser-grained in matrix-poor strata compared to matrix-rich strata.

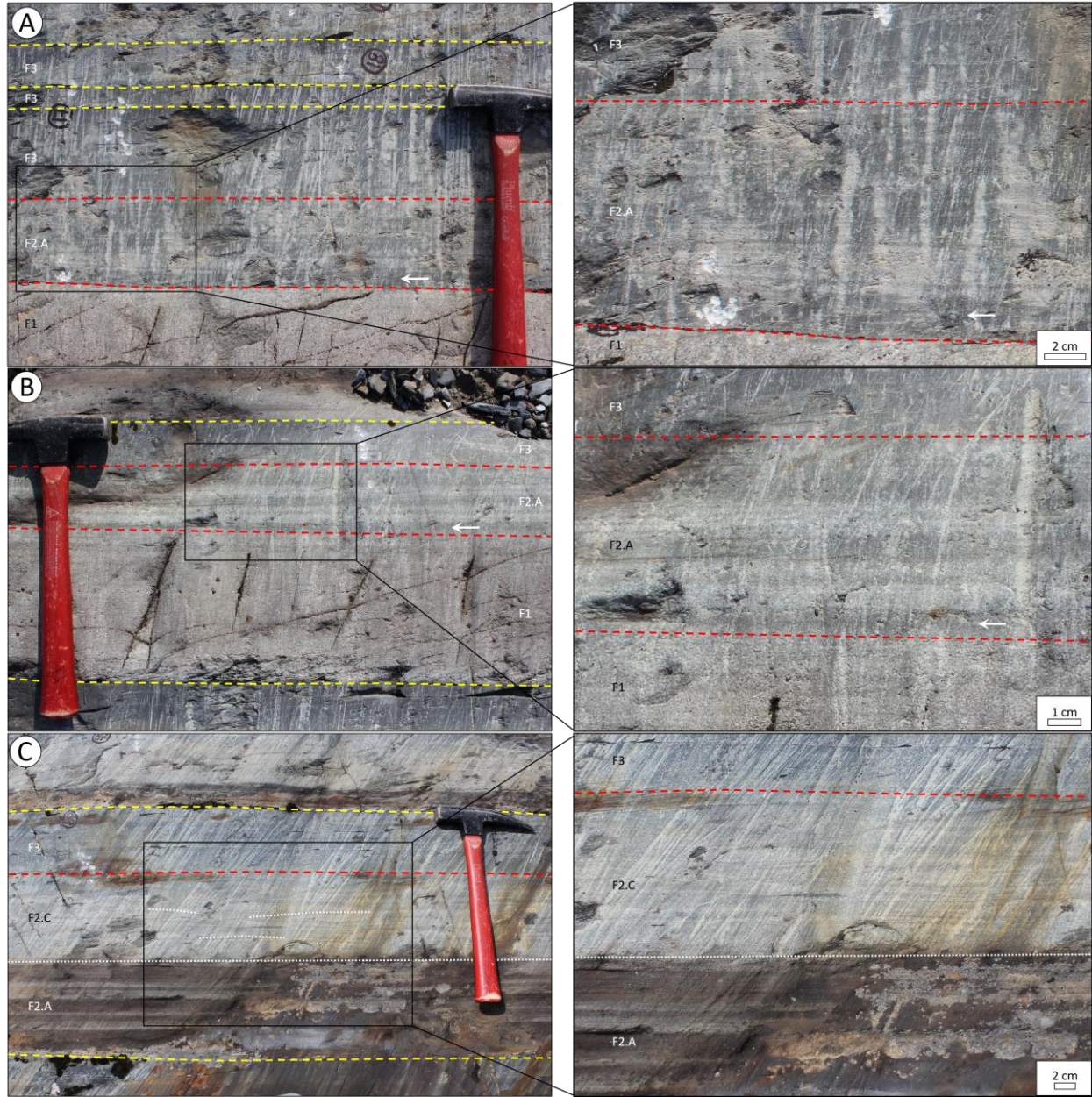


Figure 3.16: Examples of matrix-rich (A, B) and matrix-poor (C) planar-stratified F2.A. Photos are presented as pairs, with the photo on the right being an expanded view of those on the left in order to help highlight the stratification. Matrix-poor, carbonate cemented F2.A strata tend to be more thickly stratified than similarly thick matrix-rich F2.A strata. A) Matrix-rich, planar-stratified F2.A division sharply overlying F1 and gradationally overlain by F3. Note the thick mud layer separating F1 from F2.A (indicated by the white arrow). B) Matrix-rich, planar-stratified F2.A with thickly stratified sand and mud layers that overlie an anomalously thick mud layer that sharply caps F1 (indicated by white arrow). C) Matrix-poor, carbonate cemented planar-stratified clayey sandstone (F2.A) at the base of the bed overlain sharply by matrix-rich, wavy-stratification (F2.C). However, subfacies transitions from matrix-poor, carbonate cemented strata to matrix-rich strata lacking carbonate cement within F2 divisions, as shown here, is uncommon (2% of beds containing F2).

Where F2.A is an intermediate division in a bed, commonly (36%) the underlying F1 division is similarly planar-stratified. Planar-stratification typically forms the entire F2 division

and is either matrix-rich (71%) or matrix-poor and carbonate cemented (29%). If more than one style of stratification (*i.e.* subfacies) is deposited within an F2 division (19% of F2 beds), the initial type of stratification is typically planar-stratified where matrix-poor, carbonate cemented F2.A divisions commonly changes vertically to similarly matrix-poor cross-stratification (46%; F2.B) or wavy-stratification (18%; F2.C). Alternatively, matrix-rich F2.A strata changes vertically to similarly matrix-rich F2.C (27%). Uncommonly (9%), matrix-poor, carbonated cemented F2.A strata changes vertically to finer-grained matrix-rich wavy-stratification (F2.C) (*e.g.* *Fig. 3.16C*).

Cross-stratified F2.B strata have an average grain size of lower fine sand, and dip between 4-18° (commonly > 10°), which is low compared to typical Tc divisions elsewhere in the Castle Creek study area (*e.g.* *Fig. 3.17*) (Khan, 2012; Bergen, 2017). Additionally, unlike the wide range in matrix content (10-60%) in the sand layers in subfacies F2.A and F2.C (*i.e.* matrix-poor and matrix-rich end-members; see discussion above), sand layers in F2.B are always matrix-poor (10-30% matrix) and carbonate cemented with ferroan calcite (up to 34% in thin section), which form a spectrum of cross-stratified styles, often with abundant internal truncation surfaces. These styles include isolated cross-stratified sets encased in mud, cross-stratified sets that are sand-rich in their upper part and feather into mud in their (downflow) troughs, or climbing ripple-stratification that feathers into mud in up- and downflow directions (*Fig. 3.18*). Interstratified sand and mud layers are laterally discontinuous and often extend along strike for less than 1 metre.

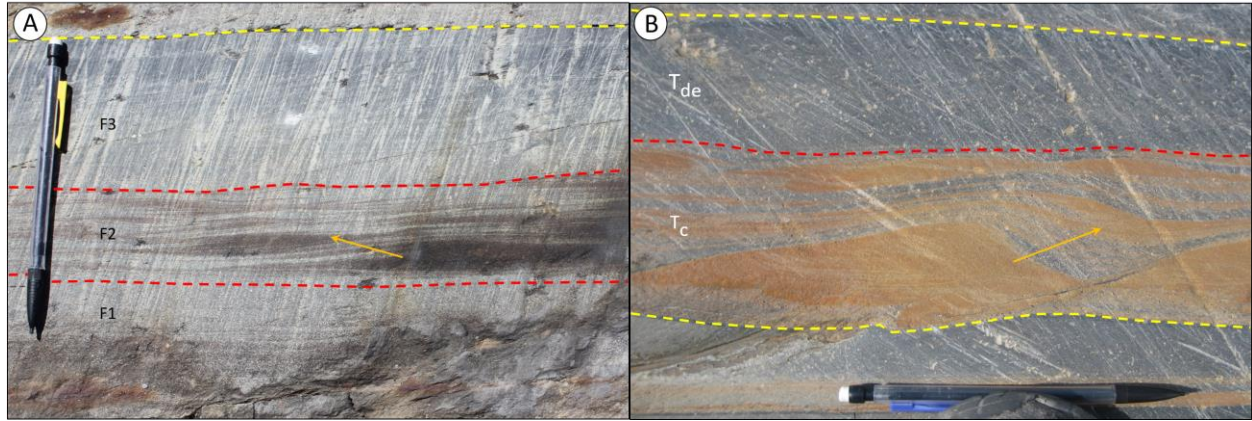


Figure 3.17: Comparison between climbing (indicated by the orange arrows), low-amplitude ($6-8^\circ$ at base to $3-4^\circ$ near top of F2 division), cross- (F2.A) to wavy-stratified F2.C strata (bounded between red dashed lines) in (A) and high-angle ($30-32^\circ$ lee face inclination), cross-stratified climbing ripple (T_c) set in levee deposits (rusty orange strata below red dashed line in part B) (Bergen, 2017), both in the Castle Creek study area. Note the consistently lower angle of cross-stratification and the lateral discontinuity of the sand-rich (dark brown) strata in the F2 division. Bed contacts are indicated by yellow dashed lines, while red dashed lines separate different facies or Bouma divisions.

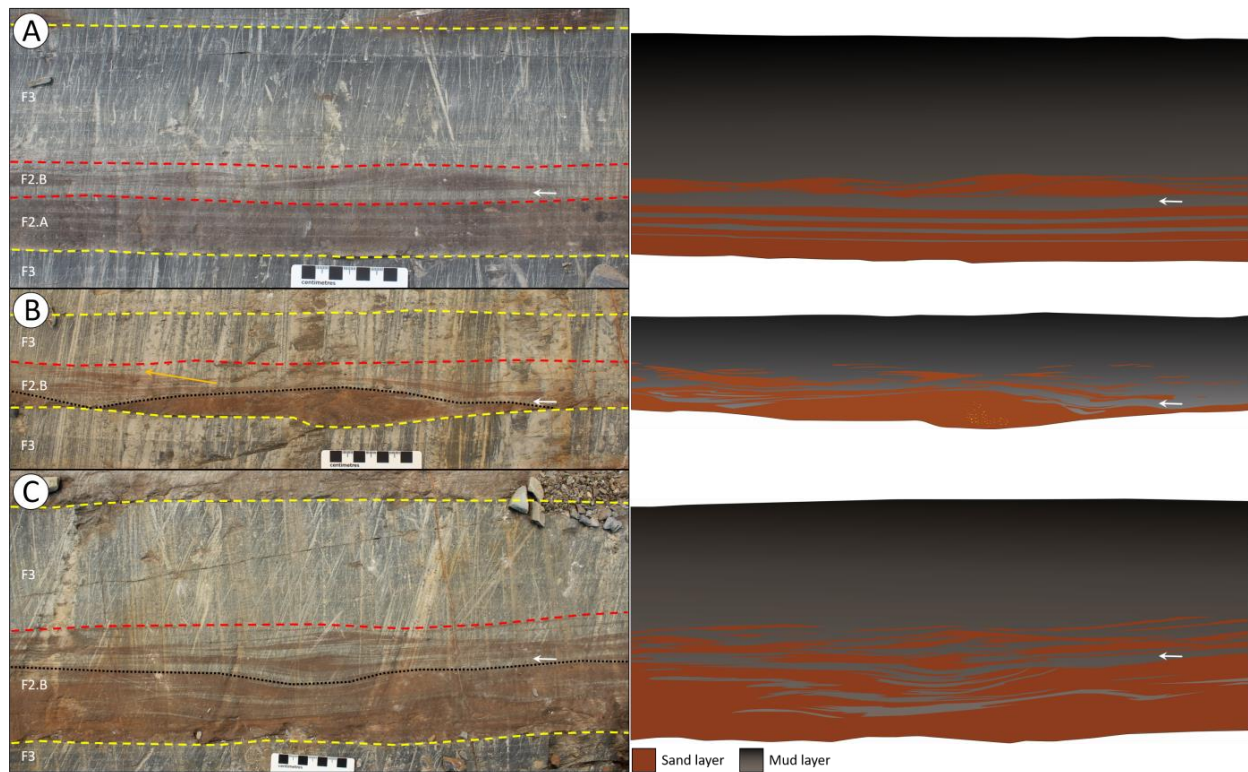


Figure 3.18: Photo pairs (photo and accompanying line diagram) illustrating F2.B divisions consisting of two stacked styles of stratification; contact indicated by black dotted line. Note that the different styles are separated by a thick laterally continuous mud layer (indicated by white arrows). Line drawings on the right help to highlight the intricate and often subtle interstratification of the sand and mud layers. A) Basal planar-stratified, carbonate cemented F2.A strata overlain abruptly by a thick (1.5 cm) laterally continuous mud layer (indicated by white arrows), which then is overlain by isolated sand-rich mounds draped with mud layers, ultimately capped by F3 strata. In the F2.B division paleoflow is very generally toward the left (SE). More significantly, the F2.B division consists of discrete sand-rich mounds separated by continuous mud-rich laminae, and is interpreted to be a single ripple set that experienced recurring changes in local sediment supply and ripple behaviour. B) Isolated ripple filling a scour; paleoflow is very

generally towards the right (NW). Note the abundant, dispersed coarse sand grains concentrated in the ripple trough. This, then, is overlain by a thick (up to 6 mm) mud layer (base delineated by black dotted line) that drapes the ripple formset and separates it from low-angle climbing ripple cross-stratification (orange arrow indicates angle of climb) that feathers into mud in both up- and downflow directions. Very generally paleoflow is toward the left (SE), which is more or less opposite to the underlying ripple formset. C) Thick (up to 1 cm), laterally continuous mud layer (base denoted by black dotted line) that abruptly separates higher-angle cross-stratified sets with muddy troughs at the base of the bed from more widely spaced, more isolated cross-stratified sets with sandy stoss and crestal areas that then feather into thicker muddier troughs. In all photos bed contacts are denoted by yellow dashed lines whereas red dashed lines separate different facies or subfacies.

Where F2.B makes up the entire F2 division the type of cross-stratification (*i.e.* isolated, climbing, *etc.*) remains similar throughout its entire thickness. More commonly where more than one subfacies of F2 is deposited in a bed, cross-stratification commonly transitions vertically to either matrix-poor, carbonate cemented (88%), or matrix-rich (12%) wavy-stratification. Paleocurrent reversals have also been identified in some F2 divisions where two or more types of cross-stratification are observed (*e.g.* *Fig. 3.18B*; see discussion below).

Wavy-stratification (F2.C) is the most common (55%) style of stratification in F2 divisions, and is also the most common type of F2 stratification where the F2 division overlies the basal contact. F2.C comprises about sub-equal occurrences (see *Fig. 3.15*) of matrix-poor, calcite-cemented (*Fig. 3.19A*), and matrix-rich (*Fig. 3.19B*) end-members. Where the F2 division comprises a single subfacies, it most commonly is F2.C (77%), or F2.C forms the top of the division (23%) where underlain by other F2 subfacies.

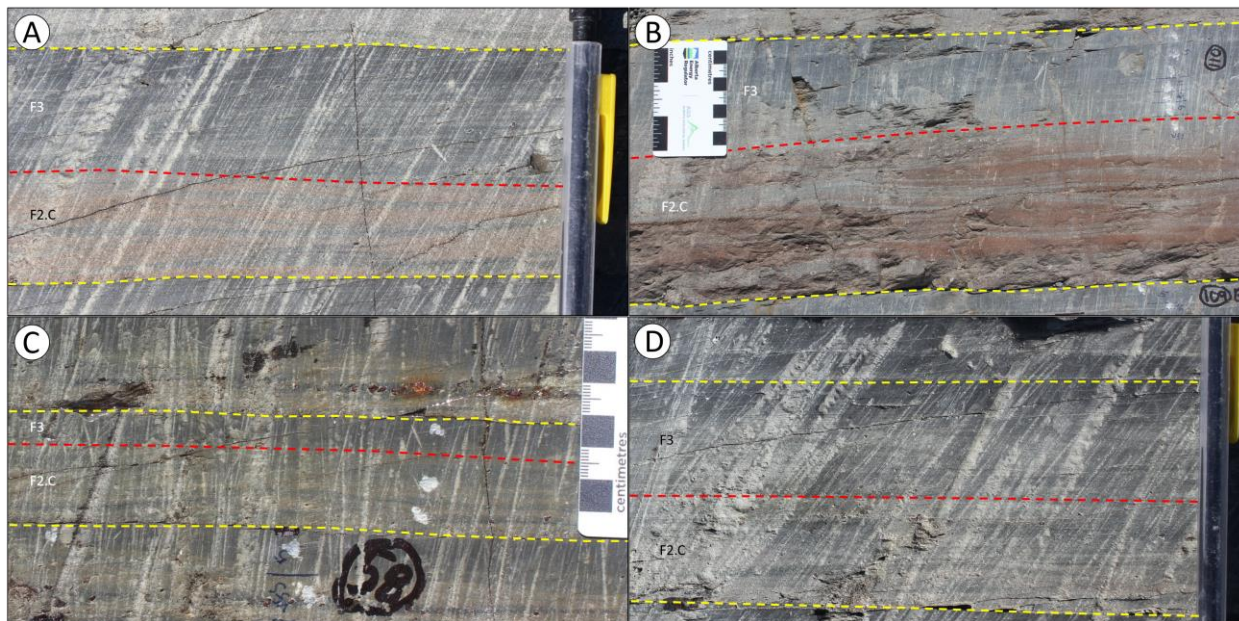


Figure 3.19: Wavy-stratified strata of F2.C. Significantly, and in comparison to F2.B, mud layers in strata of F2.C can commonly be traced for metres along strike. A) Thin, matrix-poor, carbonate cemented, wavy-stratification overlain gradationally by F3. B) Anomalously thick (11 cm) matrix-poor, carbonate cemented, wavy-stratified division. C) Thin, matrix-rich, wavy-stratified F2.C division. D) Matrix-rich, subtly wavy-stratified F2.C division gradationally overlain by F3 strata. In all photos basal bed contacts are denoted by the yellow dashed lines whereas red dashed lines separate different facies or subfacies in a bed.

Compared to F2.B, cross-stratification in F2.C is always much lower angle ranging from 1-8° (typically *c.* 4°), and mud layers tend to be more laterally continuous (> 1 m) rather than filling depressions on one or both sides of the raised part of the ripple formset in F2.B. Sand and mud layers tend to be thinner and more intricately interlaminated. Spacing and height of the undulation is of the order of 15-30 cm and a few millimetres, respectively. On average grain size in F2.C is lower fine sand, where the range of grain size is similar between matrix-poor and matrix-rich end-members and the difference is in their proportions of matrix.

Where F2 overlies F1 (*e.g.* *Fig. 3.16A, B, 3.17A*), or comprises more than one type of stratification (*e.g.* *Fig. 3.18B, C, 3.20*), the contact often exhibits evidence of erosion; which is especially evident where the underlying subfacies is cross-stratified (see *Fig. 3.20* below) and then is overlain by an anomalously thick (0.5-2 cm) mud layer (indicated by white arrows in *Fig. 3.16, 3.18, 3.20, 3.21*). Compositionally these mud layers are similar to the thinner intercalated mud

layers, being composed primarily of recrystallized clay and silt grains with 15-30% dispersed medium to very fine sand, but differ only in thickness and along-strike continuity (extend for 10s metres and change little in thickness along their length; *Fig. 3.21*). Additionally, in some cross-stratified F2.B units, internal truncation surfaces are overlain not only by a thick mud layer but also coincide with a reversal in paleocurrent direction (*Fig. 3.18B, 3.20A*).

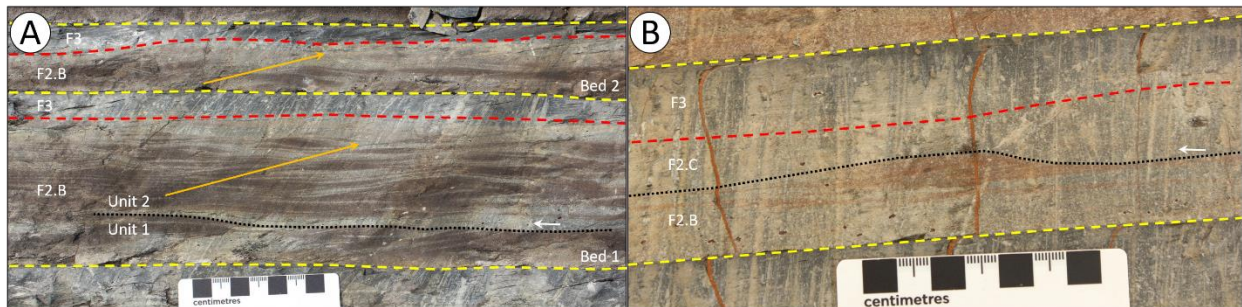


Figure 3.20: A) Two beds composed of an F2.B unit overlain gradationally (red dashed lines) by F3; bed bases indicated by yellow dashed lines. Two stacked F2.B units separated by an anomalously thick (0.5-2 cm) mud layer (indicated by white arrows; base delineated by black dotted lines). Note that the thick mud layer overlies an erosional surface that truncates cross-stratification (highlighted by white dotted lines) at the top of the lower F2.B unit. Unlike the lower unit, cross-stratification in the upper unit is both migrating and climbing (orange arrows) toward the upper right of the photo, which is opposite to the direction of cross-stratification in the lower unit. Note also the transition of the sand-rich crestal parts of ripples into mud-rich deposits in the downflow trough. B) Isolated ripple that in the downflow direction (left) interfingers with more mud-rich strata, and in the upflow direction is truncated and then sharply overlain by a thick mud layer (white arrow).

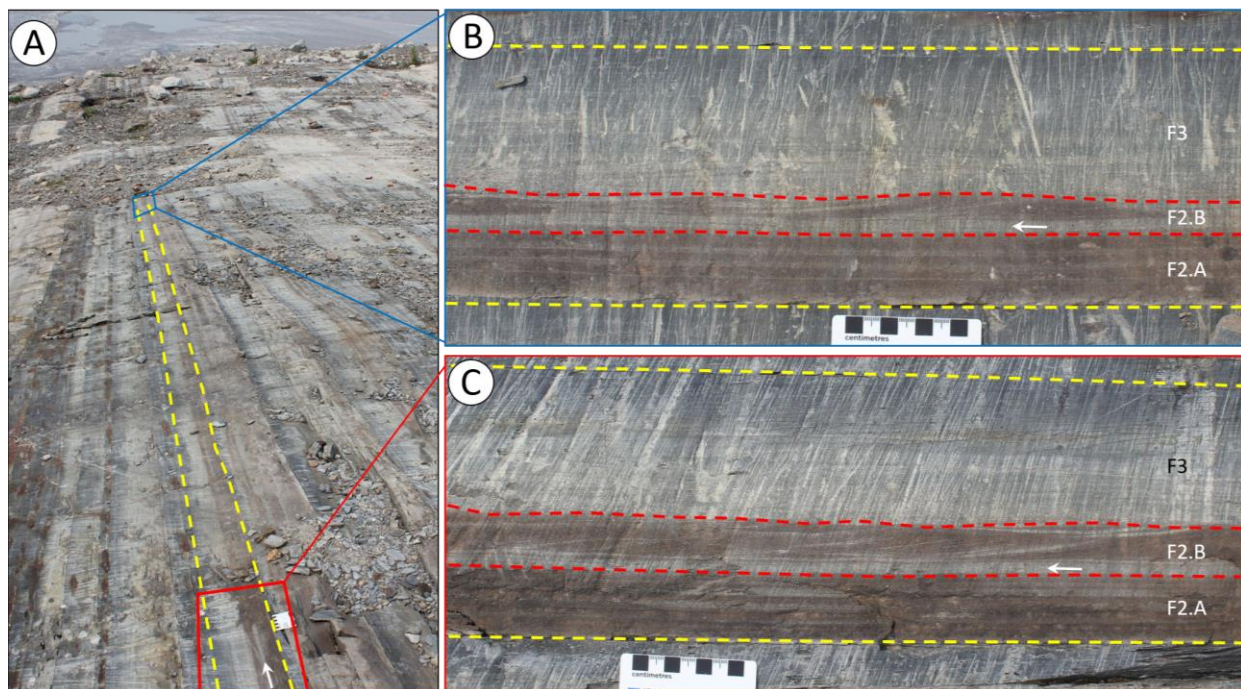


Figure 3.21: A) Thick (0.5-2 cm), laterally extensive mud layers that extend for at least several metres along strike (indicated by white arrow; B and C are separated by about 5 m along strike). Almost always thick mud layers overlie a scour surface and separate different styles of stratification in F2 strata. Bed bases are indicated by yellow dashed lines. Note that stratigraphic top is to the left; black squares on scale card are 1 cm (left of card). B and C) Closeups of (A) showing a basal matrix-poor, carbonate cemented, planar-stratified F2.A overlain sharply by a thick mud layer and then by a single cross-stratified set consisting of isolated sand-rich regions that transition downflow (left/SE) into more interstratified sand-mud cross-lamination – F2.B is then capped gradationally by F3 strata. Note that in both (B) and (C) subfacies F2.A and F2.B are separated by the same thick mud layer, which is indicated by the arrow in (A).

3.5.3 Interpretation

Heterolithic successions containing interstratified layers of sand and mud, like F2, are commonly interpreted to be associated with tidal processes in shallow-marine environments where suspended sediment passively settles from the water column (*sensu* Reineck and Wunderlich, 1968; Smith, 1988; Räsänen *et al.*, 1995). In part this is because criteria for identifying dynamically deposited mudstone layers from a single decelerating sediment-gravity flow with high near-bed suspended sediment concentration (*i.e.* > 10 g/L) are not well established. However recent laboratory experiments have shown that mud can accumulate much more rapidly, compared to passive sedimentation, from dense (> 10 g/L), near-bed suspensions even where current velocities exceed the threshold for mud erosion (*e.g.* Baas and Best, 2002; Baas *et al.*, 2009; Baas *et al.*,

2013), suggesting that the deposition of up to 1-2 cm-thick (compacted) mud layers in F2 may only require minutes to hours (Mehta *et al.*, 1989). In the field, geologists have identified a suite of lithological characteristics suggestive of deposition from energetic fluid mud layers (Bhattacharya and MacEachern, 2009; Ichaso and Dalrymple, 2009; Mackay and Dalrymple, 2011), which principally are the rhythmic intercalation of sand- and mud-rich layers, and the absence of grading within the mud layers indicating hindered settling conditions in the near-bed region (*cf.* Mehta, 1991; Winterwerp, 2002).

It is clear that interstratified mud-rich F2 strata were not formed by turbidity currents that deposit classical turbidites consisting of a normally-graded succession with a consistent superimposed suite of sedimentary structures (Bouma, 1962) (see discussion in 2.3.2.3). Instead, F2 strata represent deposition from turbidity currents that experienced a complex history of deceleration and patterns of deposition. It is important to note that fluctuating flow velocity is not a requisite condition to produce interstratified sand and mud deposits (Baas *et al.*, 2015), like previously interpreted for the formation of flaser, wavy, or lenticular bedding (*sensu* Reineck and Wunderlich, 1968), and interstratified sand and mud layers in deep-marine sediment-gravity flow deposits (Kneller, 1995). In general, most flows, at least initially, were sufficiently competent to transport medium, and in some cases, coarse sand (*e.g.* Fig. 3.18B). With time flows waned and strata became progressively overlain by fine and then very fine sand with abundant interstitial mud, which also is rhythmically intercalated with mud interlayers. Considering that F2 deposits have considerable amounts of mud (36% matrix on average), it can be inferred that the flows that deposited these strata had even higher concentrations of mud, which exerted a major control on flow dynamics, and in turn, the resultant deposit (Baas *et al.*, 2011, 2013, 2015). In addition to these textural changes, sedimentary structures also show a consistent upward change from plane-

parallel-, to cross-, to wavy-stratification (*Fig. 3.15*). Although, the full suite of stratification styles is only rarely observed in a single bed, the more common occurrence of partial suites is always in the same order.

Plane-parallel-stratification (F2.A) indicates uniform bedload transport on the bed where flow speeds were either too fast for the development of ripples (equivalent to upper-stage plane bed), or the requisite hydraulic conditions in the near-bed region were unable to initiate and amplify bed-surface defects that eventually would develop into mature ripples. As a result, sediment transport on the bed remained spatially uniform and resulted in deposition of planar-stratified sediment, possibly at flow speeds that otherwise would have formed ripples in a clear-water flow (Tilston *et al.*, 2015). With further flow deceleration sediment continued to fall from suspension into the near-bed region and more fully stratified the flow, which in turn promoted the development and strengthening of a near-bed Kelvin-Helmholtz instability. This, then, coupled with sediment transport along the bed and resulted in spatial changes in the pattern of sediment transport and deposition, and with the inception and growth of bed surface defects that ultimately formed downstream-migrating ripples (F2.B) (Arnott, 2012; Tilston *et al.*, 2015), which sequestered sand in their topographically elevated part (crest) and allowed mud to accumulate in the adjacent lower-energy trough (*cf.* Baas *et al.*, 2011; Becker *et al.*, 2013). These processes formed strata of F2.B. Uncommonly (< 3% of F2 beds), sediment fallout rate was sufficiently high that the bed aggraded fast enough that the ripples were forced to climb (Allen, 1971; Ashley *et al.*, 1982). With continued reduction in flow speed, and the reduced availability of (coarse) particles to settle into the near-bed region, the near-bed hydrodynamic instability, although still capable of generating bed defects, was unable to cause enough amplification of the defect to sufficiently alter (sediment) transport patterns on the up- and downflow sides of defects. As a consequence, a wavy

bed surface (F2.C) remained the stable bed state. In terms of angular bedforms, cohesive particles significantly slow the rate of their growth (Baas *et al.*, 2013), and make the substrate more difficult to erode compared to pure sand (Mitchener and Torfs, 1996). However, subtle changes in other parameters like flow velocity, clay type, concentration and size, initial turbulent structure, and concentration of non-cohesive sediment can also dramatically alter flow conditions and accordingly bed-form morphology (Baas and Best, 2002; Baker *et al.*, 2017). However, at present the effect of these parameters on bedform development in the near-bed region of the mud-rich sediment gravity flows and their resultant deposits remain poorly understood and points to the need for continued experimental- and field-based research.

Where F2 overlies F1 strata or is composed of two or more different types of stratification the contact is marked by an anomalously thick, laterally continuous mud layer up to 2 cm thick. Additionally, where F1 is overlain by F2 there is a sharp upward increase in mud content. Significantly, these thick mud layers overlie an erosion surface that is interpreted to have formed by low frequency, high energy turbulent eddies that scoured the sea bed and increased the sediment load in the lower part of the flow. This new sediment would have substantially reduced the turbulent kinetic energy (TKE) in the flow as energy was consumed by accelerating and maintaining these particles in suspension. Note that although the incorporation of new suspended sediment into the basal part of the flow may have decreased TKE, it had little effect on the most energetic scales of fluid motion (*i.e.* integral scales; *Fig. 3.22*) (Bennett *et al.*, 2014). Nevertheless, scouring of the bed would have substantially increased the near-bed sediment load and, especially in this case, the volume of clay mineral particles. The abundance of these particles would have led to particle aggregation and ultimately gelling followed by *en masse* deposition of thick, laterally continuous, poorly-sorted, massive mud layers (McAnally *et al.*, 2007). Following deposition of

these fine-grained layers, and depending on the hydraulic and sediment conditions in the near-bed region, deposition of F2.A, F2.B, or F2.C strata ensued.

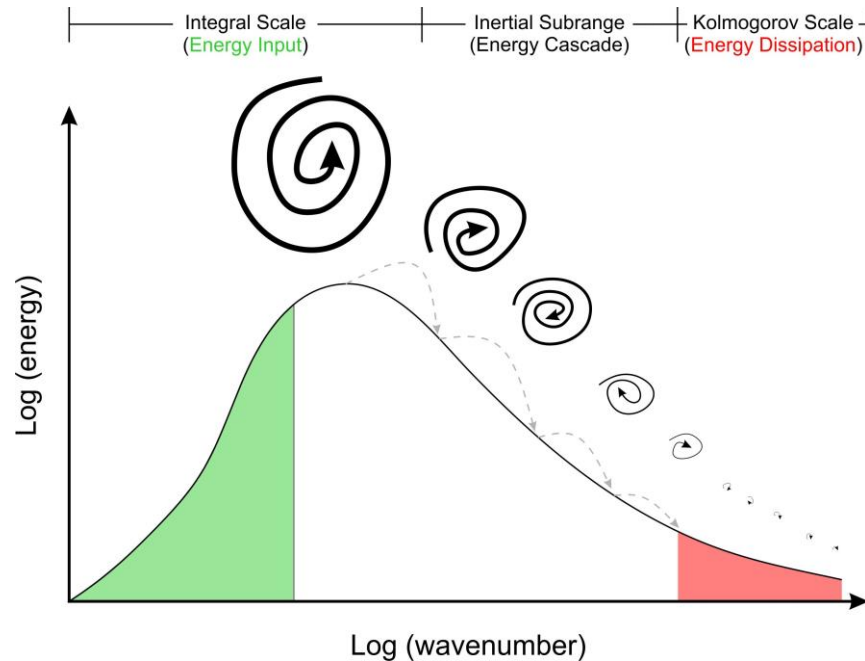


Figure 3.22: Schematic of the energy cascade illustrating the input of energy into the system at the integral scale of turbulence (green area) and the successive nonlinear break up and transfer of energy from large eddies into progressively smaller eddies (white area), which eventually become so small (red area) that viscosity dominates and converts turbulent kinetic energy (TKE) into heat. As eddies become progressively smaller, they also become more abundant (adapted from Davidson, 2015).

Where F2 lies at the base of a bed, the basal contact is commonly sharp and planar suggesting turbulence suspension near the bed. Uncommonly, the basal contact is locally sharp and undulatory (1-2 cm relief) indicating localized sea bed scours due to erosion by turbulent eddies, or instead rare post-depositional soft sediment deformation where structures such as flames are observed suggesting short temporal intervals between depositional events.

F2 strata commonly grade into F3 where there is a considerable decrease in the sand fraction, especially medium sand grain sizes or larger, possibly due to a lack of available sand, or flow velocity was too slow to transport grains larger than medium sand.

3.6 Facies 3: Graded, plane parallel-laminated sandy claystone to massive or graded, structureless claystone

3.6.1 Macroscopic Characteristics

Facies 3 (F3) is a bed division composed of plane-parallel-laminated sandy claystone capped by massive or graded, structureless claystone (described in more detail in 3.6.2). Planar lamination, especially its coarser basal portion (*Fig. 3.23B*), can be correlated for up to 10s of metres along strike. Beds composed of only F3 commonly stack to form sedimentation units up to 17 beds thick (average 3 beds). However, without first sawing the outcrop surface it is difficult to discern stratification in outcrop due to weathering (see discussion in 3.6.2). This facies is easily recognized in the field due to its distinctive colour change from light grey (laminated basal portion) to dark grey (structureless portion) (*e.g. Fig. 3.23A*).

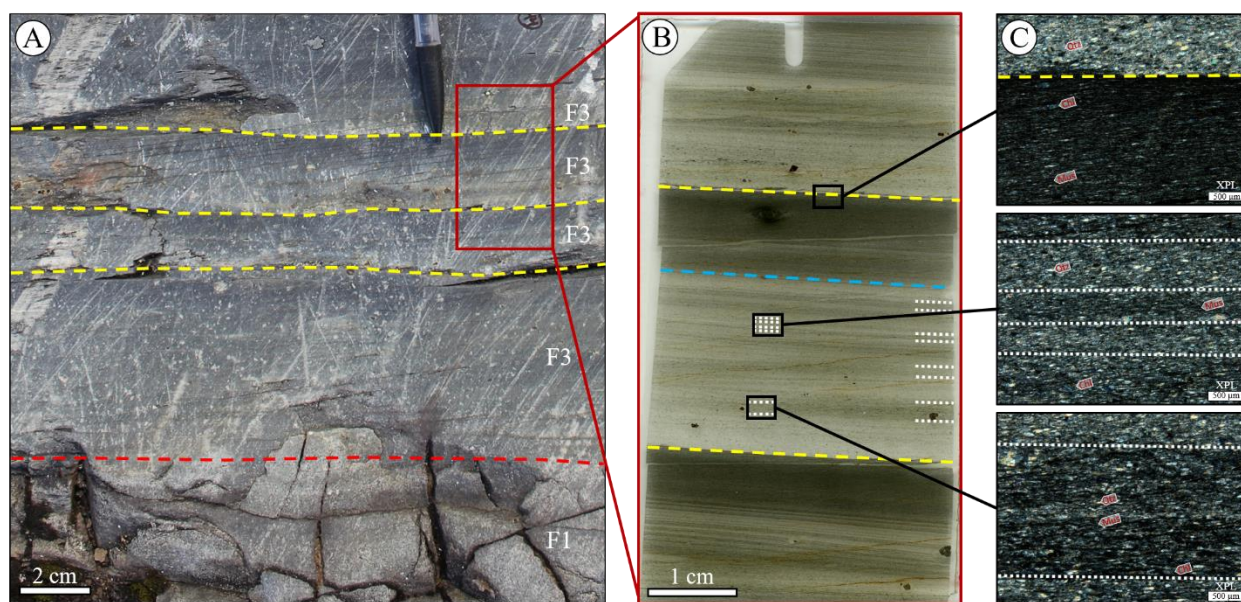


Figure 3.23: A) Sharp based (yellow dashed lines) strata of F3. Although (parallel) lamination is not apparent in the lower, light-coloured part of the bed in the field, the gradual upward colour change from light grey to dark grey at the top (structureless upper part) can be easily discerned and used to differentiate individual F3 beds. B) In thin section strata of F3 exhibit a distinctive alternation of planar layers of sandy claystone (lighter laminae) overlain by claystone (lighter laminae; separated by white dotted lines) that collectively thin and fine upward, and ultimately are capped sharply by a structureless claystone (above blue dashed line). C) Photomicrographs of thinly laminated (2 mm thick; bottom) sandy claystone with claystone interlaminae (separated by white dotted lines) that thin to very thin laminae (0.5 mm thick; middle) before grading into structureless claystone at the top of the bed. Common minerals: quartz (Qtz), muscovite (Mus), and chlorite (Chl) are indicated.

F3 is observed in almost all (99%) of the measured beds and makes up about 3-100% of the bed thickness (average 81%). This facies ranges in thickness from 0.5-15 cm (average 3.2 cm, mode 2 cm). However, thick occurrences (> 10 cm) likely consist of multiple thinner F3 beds, but bed contacts in this fine-grained strata are difficult to discern without a freshly sawed surface. Where F3 makes up the whole bed (68%) the basal contact is generally sharp and planar (96%), or uncommonly undulates (up to 1 m laterally) with 0.5-1 cm of relief (4%), or are loaded with load casts and flame structures (< 1%). Where not at the base of the bed, F3 abruptly or gradationally overlies F1 (5%) or F2 (27%), respectively. However, the contact between F1 and F3 becomes more gradational as the grain size and thickness of F1 decrease and becomes more similar to F3.

3.6.2 Microscopic Characteristics

According to the Folk (1980) classification, Facies 3 is classified a clayey sandstone based on the relative percentage of sand grains (quartz) and matrix (silt and clay) point counted in 20 samples after normalizing the sand, silt, and clay. This facies typically comprises 30% grains, 70% matrix (muscovite, chlorite, sericite, and silt-sized quartz), and < 1% carbonate cement (calcite and dolomite). Grains are usually angular to sub-angular and elongate to equant. Average grain size is typically very fine (46%) or fine (45%) sand, with lesser (9%) medium sand which are often dispersed amongst smaller grains. Textural grading is manifest microscopically as a progressive decrease in the maximum, mean and range of grain sizes vertically through successive laminae, as well as a decrease in the sand:clay ratio in the lower planar-laminated part.

Grains are dominantly undulatory quartz (99%). Rutile makes up the remaining granular constituent (< 1%). The matrix, on the other hand, consists of mica (muscovite and sericite) (79%), chlorite (17%), and minor silt-sized quartz (3%). Note that because of the difficulty in

differentiating detrital silt from (bulging) recrystallized quartz grains the silt fraction is likely underestimated in point count data as silt was observed in SEM analysis even at the tops of structureless claystone caps in F3 (*Fig. 3.24*). The chlorite typically forms sand-sized porphyroblasts, and along with quartz grains and recrystallized matrix (finer chlorite and micas), tend to be aligned parallel to bedding, especially in the mud layers (*e.g. Fig. 3.23C*). Cement is rare (< 1%) and is composed exclusively of small (< 0.125 mm) rhombohedral ferroan dolomite crystals.

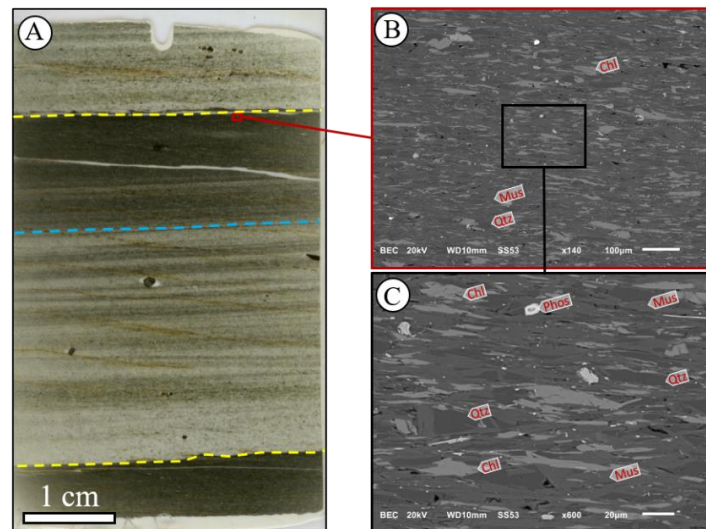


Figure 3.24: A) Photo of a thin section showing planar-laminated sandy claystone (below blue dashed line) capped by structureless claystone in F3 strata (bounded between yellow dashed lines). B and C) SEM images at increasing magnification (100 µm and 20 µm scale bar in (B) and (C), respectively) near the top of a structureless claystone cap in a F3 bed. Note that silt-sized quartz grains (darkest grains) are dispersed throughout the claystone, even up to the top indicating that deposition was not just from quiescent suspension fallout but instead form fine-grained fluid muds (Sumner *et al.*, 2008; Baas *et al.*, 2015). Note that the quartz grains are generally aligned parallel to bedding indicating possible grain rotation and elongation due to compaction and/or multiple stages of deformation in the study area. Common minerals, including quartz (Qtz, dark grey), muscovite (Mus, grey), chlorite (Chl, light grey), and phosphate (Phos, white) have been indicated.

Facies 3 is made up of strata comprising two parts: a plane-parallel laminated basal sandy claystone (*Fig. 3.25*) overlain sharply by a graded or ungraded claystone (*Fig. 3.23B*). The basal part is always planar-laminated and composed internally of rhythmically interstratified poorly sorted sand (sandy claystone) laminae and moderately to poorly sorted mud (claystone) laminae.

Sand and mudstone laminae thin vertically from 2-5 mm at the base to 1 mm-hairline laminae at the top. Boundaries between laminae become better defined stratigraphically upwards as the range of grain sizes between the sand and mud layers converge. Each sand lamina fines upwards as do the dispersed sand grains in the superjacent mud lamina (*Fig. 3.25*). The coarsest (fine-grained sand) basal sand laminae of F3 have dispersed, up to medium-grained sand grains, which then fine to silt vertically. The abundance of dispersed grains is directly proportional to the thickness of the basal laminated part. As the recrystallized clay content increases upwards, the abundance and size of chlorite porphyroblasts in the sand and mud lamina increase (up to coarse sand size).

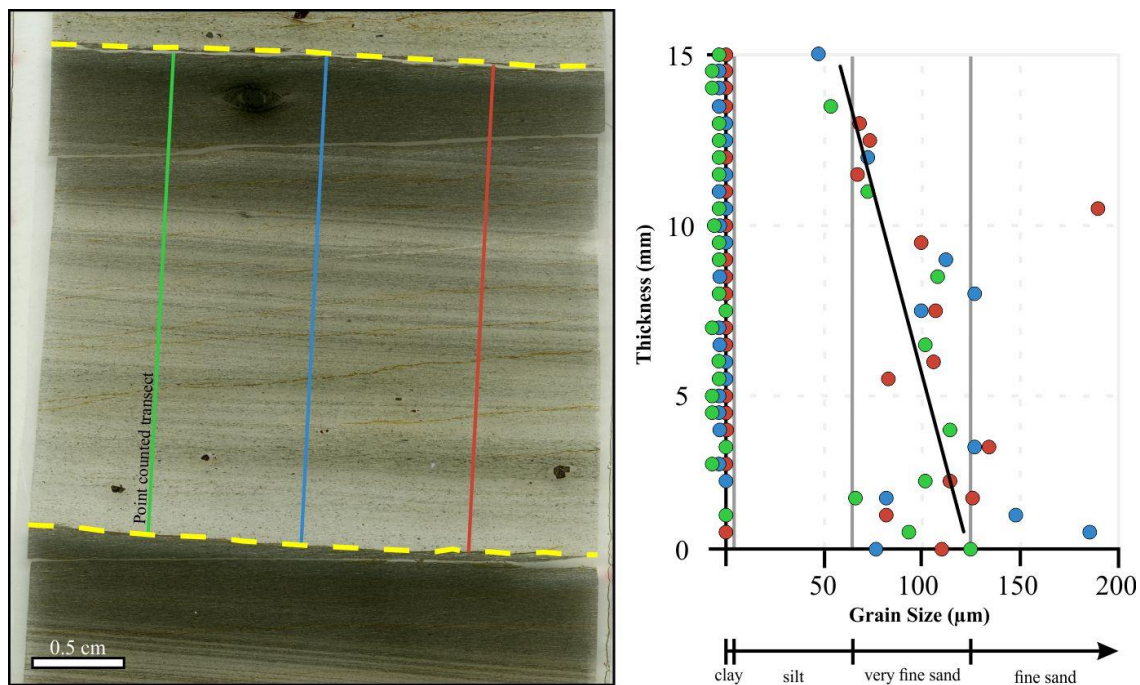


Figure 3.25: Grain size analysis through a very thinly-bedded F3 bed (bounded between yellow dashed lines) based on point counting of 30 points identified every 0.5 mm along three separate vertical transects (*i.e.* 90 points total). Note that data points lying on the vertical axis represent recrystallized clay (*i.e.* chlorite, muscovite, and sericite). Although dominated by matrix (points on vertical axis), the coarser grained fraction shows a clear upward fining (highlighted by the black linear trendline). Significantly, the range of grain sizes in adjacent sandy and muddy layers (grey layers) is similar, and stratigraphically upward show a similar upward fining.

3.6.3 Interpretation

F3 is interpreted to be similar to a T_{de} turbidite or upper T_{de} part of a more complete turbidite with deposition from a dilute, fine-grained suspension (Stow and Shanmugam, 1980;

Stow and Piper, 1984a). The gradual upward fining from mostly very fine-grained sand to silt suggests a progressive reduction in flow competence during a single sedimentation event. Pervasive planar lamination composed of rhythmically interlaminated sand (sandy claystone) and mud (claystone) layers may relate to the action of internal waves within the channelized turbidity current (Khan and Arnott, 2011). Alternatively, Stow and Bowen (1978, 1980) proposed that the millimetric interlamination of comparatively coarse- and fine-grained layers was the result of alternating rheological changes in the viscous sublayer of a low-energy turbidity flow, specifically episodes of individual coarse particle settling followed by near-bed gelling and deposition of a finer-grained clay layer. However, sand and mud lamina in F3 layers are more poorly sorted, and the coarse sand layer has a wider range of grain sizes than described in the depositional sorting model proposed by Stow and Bowen (1978, 1980).

Sand laminae commonly have > 60% matrix and dispersed grains up to medium sand, whereas mud laminae contain up to 15% dispersed silt to fine-grained sand. Therefore, the process of segregating silt and clay into separate, alternating lamina is likely more inefficient than suggested by Stow and Bowen (1978, 1980) as sand, silt, and recrystallized clay are deposited in different proportions in both the sand and mud layers in F3. The poorly sorted nature of the sand and mud lamina suggests that shear sorting in the boundary layer is ineffective at fully sorting the coarse sediment from the fines. Dispersed grains in the mud lamina could also be a result of clay flocculating around individual silt to fine sand-sized grains before deposition. The massive or graded claystone cap is deposited via suspension fallout in the final stages of the flow and later hemipelagic fallout. Also, other authors have proposed that this type of planar-lamination might be the result of velocity fluctuations during a single turbidity current (Lombard, 1963), reflection of a turbidity current off the walls of a small basin which then reworks the bed (Kuenen, 1964; van

Andel and Komar, 1969), quasi-cyclic bursting in the near-bed part of a turbulent boundary layer (Hesse and Chough, 1980), or a series of small, distinct flows or suspension clouds (Dzulynski and Radomski, 1955).

Lastly, structureless, poorly- to moderately-sorted, sharp-based, massive or graded claystone caps overlying the lower planar-laminated part are interpreted to have been deposited rapidly *en masse* as a fluid mud layer (McAnally *et al.*, 2007). Grading is likely related to post-depositional settling of very fine sand and silt grains in the fluid mud (Talling *et al.*, 2012). Upon deposition, these fluid mud caps in F3 strata are interpreted to be less viscous than similar claystone interpreted as being deposited as fluid mud in F2 strata indicated by thinner claystone lamina in F3 strata (*i.e.* lower suspended sediment concentrations in flows that deposited F3 strata *sensu* Mackay and Dalrymple (2011)). In thickly (*c.* 1-2 cm) developed claystone caps, subtle planar lamination is observed in graded claystone caps (*Fig. 3.26*) suggesting the stacking of discrete fluid mud layers, possibly due to the lack of noncohesive sediment in the flow and/or increased cohesive sediment in the near-bed region allowing for the deposition of thicker, repeated fluid mud layers. Conversely, massive claystone strata with dispersed silt grains suggests *en masse* deposition of a fluid mud layer with a more viscous gel-like rheology resulting from high near-bed clay concentrations (McAnally *et al.*, 2007). Alternatively, where caps grade upward from sandy claystone to claystone, particles may have segregated by their different settling velocities as suspension settling.

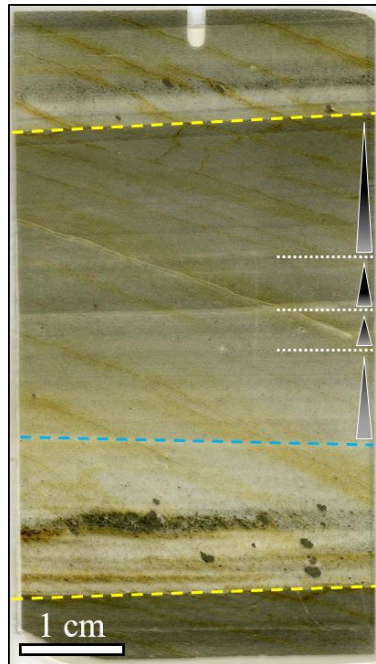


Figure 3.26: Thin bed of F3 strata (bounded between yellow dashed lines) comprising multiple faint interlaminated layers of coarser-grained (fine sand) strata in the basal part (below blue dashed line) and overlain by a faintly laminated claystone cap (above blue dashed line where contacts are indicated by white dotted lines). This faint lamination is interpreted to be multiple stacked, *en masse* deposited fluid mud layers in which post-depositional settling allowed noncohesive grains to accumulate at the base of each fluid mud layer (grading is indicated by the triangles which change from grey, representing accumulated noncohesive grains at the base, to black, representing predominantly recrystallized clay with dispersed silt near the top of each).

3.7 Facies 4: Massive, interstratified, structureless clayey sandstone and claystone

Facies 4 crops out in the northwest corner of the study area, which is outside of the primary focus area. Thus, field analysis was conducted at a reconnaissance level and only a small number of samples were collected and analyzed.

3.7.1 Macroscopic Characteristics

Facies 4 (F4) comprises > 1 cm massive fine to very fine-grained sandstone (30-40% matrix) overlain sharply by massive mudstone (with up to 10-15% dispersed silt) (Fig. 3.27). Interbeds of thin-bedded, fine-grained, carbonate cemented T_{cde}/T_{ce} sandstone turbidites are common and generally consist of isolated, single set ripple cross-stratified formsets. Strata ranges in thickness from c. 1 to > 20 cm thick and stack to form bedsets that range up to 10s m thick

showing no intercalation with other lithologies. Due to the fine-grained nature of these strata it is generally difficult to discern bedding contacts without first sawing the outcrop surface. On a cut surface, multiple bedding-parallel, closely spaced (< 1 cm), thin (< 150 μm), dark, wispy, and elongate solution seams (microstylolites) with elevated levels of insoluble material (detrital clays, iron oxides and/or sulphides) become visible (*e.g.* Fig. 3.27). The basal contacts of F4 strata is typically sharp and planar, except where coarser (> 0.25 mm), thicker (> 3 cm) sediment overlies a muddier bed resulting in a locally undulating (1-2 m spacing) basal contact.

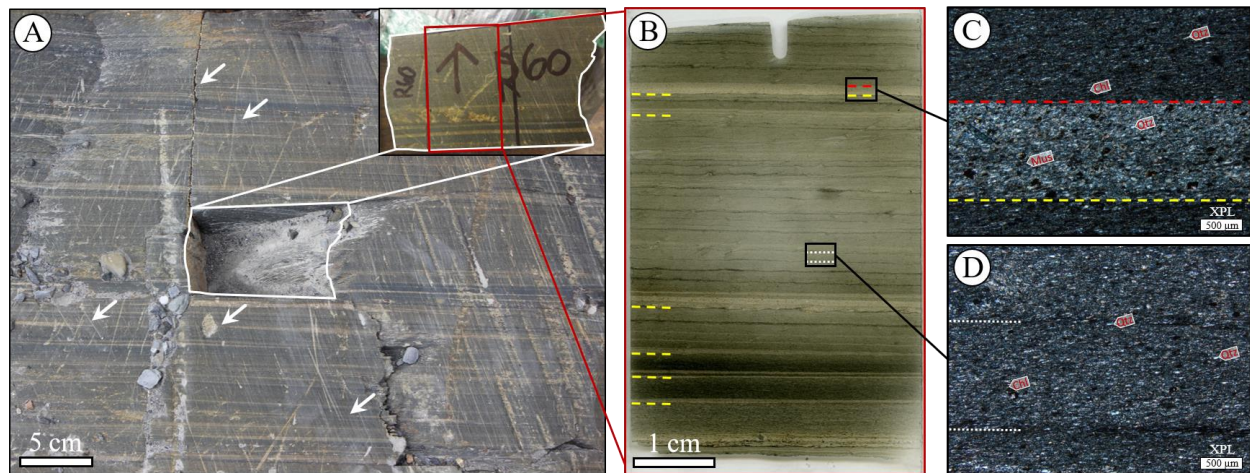


Figure 3.27: Facies 4 from macroscopic (field) to microscopic (thin section) scale. A) Outcrop photo of thin- to very thinly-bedded, fine-grained F4 beds. Bedding contacts are difficult to discern due to the fine-grained nature of this facies. The base of some units is marked by up to fine, but generally very fine sand laminae that often pinch and swell over 1-2 m along strike (possible starved ripples), indicated by white arrows, whereas others are planar-laminated or planar but structureless. B) Multiple thinly-bedded F4 strata (beds bounded between yellow dashed lines) in thin section. Even though the basal coarser-grained part appears structureless or planar-laminated here, as shown in (A), these layers often pinch out over short distances laterally (1-2 m). The contact between the coarser-grained basal layer and fine-grained cap is almost always sharp and planar. Note also the abundant thin, closely spaced (< 1 cm), bedding-parallel, dark, wispy solution seams (microstylolites) throughout the fine-grained layers (indicated by white dotted lines). C) Closeup of layering in F4 strata, specifically a structureless sand-rich layer (bounded between the red and yellow dashed line) overlain sharply by structureless sandy claystone. Note the dramatic decrease in grain size between the basal and upper parts (indicated by a red dashed line). Significantly, both parts have similar ranges in grain size, however, sand grains in the finer-grained upper part are less abundant and are floating in a matrix of mud. D) Photomicrograph of fine-grained strata showing solution seams (highlighted by white dotted lines) that are continuous for a few to tens of metres along strike in outcrop. Some common minerals like quartz (Qtz), muscovite (Mus), and chlorite (Chl) are indicated.

3.7.2 Microscopic Characteristics

Facies 4 can be classified according to the classification scheme of Folk (1974) as sandy claystone based on the relative percentage of sand grains (quartz) and matrix (silt and clay) point counted from multiple beds in one thin section after normalizing the sand, silt, and clay. F4 consists primarily (*c.* > 70%) of fine-grained (< 62.5 μm) strata. Even the coarser-grained (up to *c.* medium sand) basal part has high proportions of matrix (up to *c.* 50%) (*Fig. 3.27C*).

Grains are commonly quartz with rare plagioclase and rutile grains and range in size from fine to very fine sand. Also, rare anomalously large (medium to coarse sand size) chloritoid grains with rounded and corroded edges are observed in both sandstone and mudstone layers (*Fig. 3.28*). Grains (> 62.5 μm) are usually sub-angular to sub-rounded and elongate to equant. The matrix is composed of angular and elongate quartz silt, in addition to recrystallized (detrital clay) chlorite (typically forming sand-sized porphyroblasts), muscovite, and sericite. However, more research is needed to properly characterize these strata both macro- and microscopically.

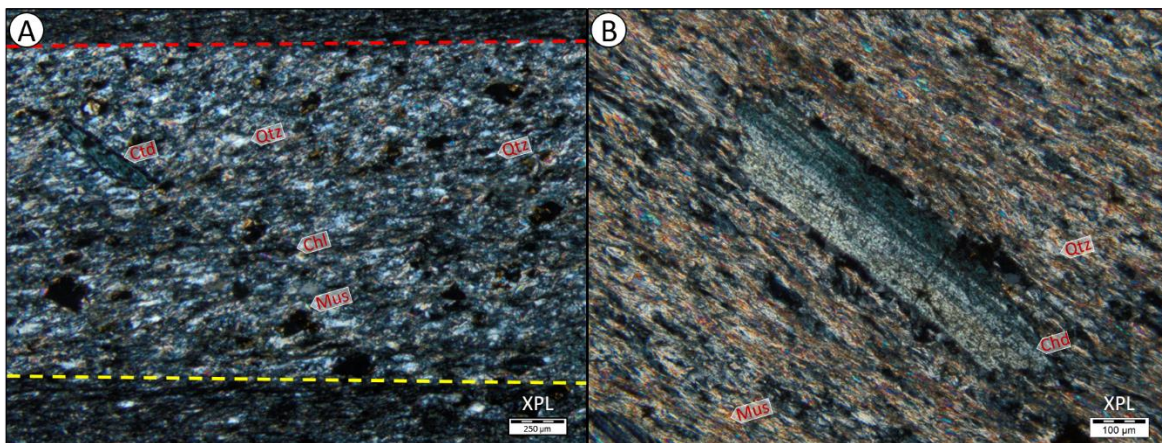


Figure 3.28: Large, elongated chloritoid (Chd; compound mineral percentages from EDS: Mg – 2.33%, Si – 25.46%, Fe – 30.48%, Al – 41.73%) porphyroblast in both the basal coarse-grained (A) and upper fine-grained layer (B) in F4 strata. A) Large (*c.* 400 μm) chloritoid porphyroblast in a coarse-grained F4 layer (bounded by yellow and red dashed lines). B) Photomicrograph of a large chloritoid porphyroblast in the fine-grained part encased in recrystallized clay (primarily micas here) with dispersed silt sized quartz and very fine-grained sand. Note the polysynthetic twinning. Note also other minerals identified in photomicrographs are quartz (Qtz), muscovite (Mus), and chlorite (Chl).

3.7.3 Interpretation

The ungraded and structureless character of both sand and mud-rich layers suggests deposition from suspension fallout and/or background hemipelagic fallout. These conditions were interrupted episodically by the emplacement of single set (starved) ripple formsets indicating traction transport and deposition from low-concentration turbidity currents.

Chapter 4: Sedimentologic and Petrographic Evidence of Flow Confinement in a Passive Continental Margin Slope Channel Complex, Isaac Formation, Windermere Supergroup, British Columbia, Canada

4.1 Regional Geology

In 1926 Walker coined the term Windermere Series to describe Proterozoic rocks that crop out in the Windermere Valley of southeastern British Columbia. Later, the United States Geological Survey classified Neoproterozoic rocks in the American Cordillera as “Windermere Group” strata, whereas better exposed and more completely preserved Neoproterozoic rocks in the Canadian Cordillera are referred to as the “Windermere Supergroup” (Link *et al.*, 1993). In this paper all Neoproterozoic rocks (728 – 570 Ma; Evenchick *et al.*, 1984; Colpron *et al.*, 2002) will be referred to as the Windermere Supergroup, or more simply WSG.

In the southern Canadian Cordillera (SCC) the WSG is interpreted to have experienced two discrete episodes of deposition (Stewart, 1972; Ross, 1991). The first records an episode of rifting related to the disassembly of the supercontinent Rodinia and consists of glaciomarine diamictite of the Toby Formation overlain locally by mafic volcanics of the Irene Formation forming a succession more than 1 km thick (*Fig. 4.1A*) (Aalto, 1971). The second episode represents post-rift thermal subsidence and the passive margin phase of WSG deposition and consists of a thick (5-7 km) succession of mostly deep-marine siliciclastic rocks and subordinate carbonates rocks (Ross and Arnott, 2007). The Kaza Group is the lowest part of the post-rift phase of the WSG and represents basin floor deposits (*Fig. 4.1B*) comprising primarily sand-rich (sandstone:mudstone ratio *c.* 75:25), sheet-like deposits composed mostly of massive or normally graded (T_a) sandstone intercalated with units made up of thin-bedded, fine-grained (T_{bc}) turbidites (Bouma, 1962; Ross *et al.*, 1995). These deposits are interpreted to be parts of basin floor depositional lobes (Ross and

Arnott, 2007; Terlaky *et al.*, 2016). The Isaac Formation conformably overlies the Kaza Group and represents slope deposits composed mostly of mudstone (75%) with subordinate (c. 25%) discontinuous lenses of sandstone and minor conglomerate. These deposits are interpreted to be leveed-slope-channel deposits with common intercalated mass-transport deposits (Ross and Arnott, 2007). In addition, carbonate deposits interbedded with sandstone and mudstone, informally termed the first and second Isaac carbonates, are observed (Ross *et al.*, 1995). Clasts within particularly thick debrites contain oolites and stromatolite fragments, suggesting sourcing from a shallow-marine carbonate platform. The Cunningham Formation conformably overlies the Isaac Formation, and in turn is conformably overlain by the Yankee Belle Formation (Ross and Murphy, 1988) forming a succession of up to > 1 km thick (Ross *et al.*, 1995). Strata comprise mostly oolitic-intraclastic limestone and mixed siliciclastic-carbonate rocks, respectively. The Cunningham and Yankee Belle formations have been interpreted, respectively, to be upper slope and shelf limestone deposits associated with high-energy shallow-marine platformal carbonates in the upper Windermere Supergroup (Campbell *et al.*, 1973; Ross and Murphy, 1988; Ross *et al.*, 1995).

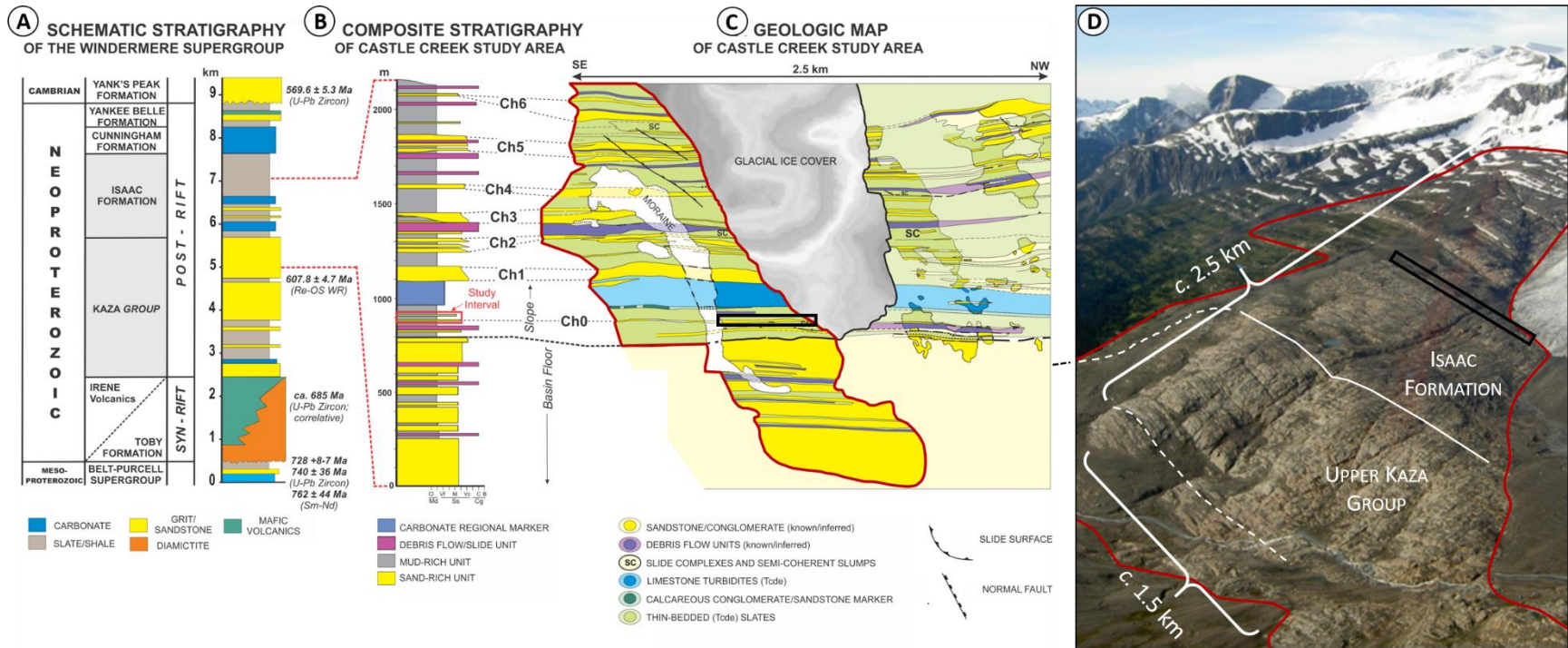


Figure 4.1: Stratigraphy of the Windermere Supergroup from regional to outcrop scale. A) Schematic regional stratigraphy of the Windermere Supergroup. Age constraints are denoted to the right of the log. B) Composite stratigraphy of the Windermere Supergroup located at the Castle Creek study area in the Cariboo Mountains. The study interval for this thesis is outlined in the red rectangle. C) Geologic map of the stratigraphy in the Castle Creek study area. The approximate location of the study area is outlined in the black rectangle and the exposed outcrop outlined in red corresponds partially to the red outline in (D). Note that channel complexes (Ch) are represented to the left (modified from Lee, 2016). D) The Castle Creek study area located in the Cariboo Mountains of east-central British Columbia. Here, a 2.5-km-thick succession of recently deglaciated, vertically-dipping strata is exposed consisting of proximal basin floor deposits of the Upper Kaza Group (c. 0.8 km thick) conformably overlain by channel-levee complexes of the Isaac Formation (c. 1.7 km thick).

Based on paleomagnetic data from WSG strata in the Mackenzie Mountains of the northern Canadian Cordillera the paleolatitude of western Laurentia has been inferred to be in a near-equatorial location (Park, 1997). Based on regional facies patterns, sediment provenance, and paleocurrent data the basin is interpreted to have been located off the coast of Laurentia with turbidity currents flowing toward the west-northwest (present-day geographical coordinates) down the continental slope and out onto the basin floor (*Fig. 4.2*) (Ressor, 1957; Mountjoy and Aitken, 1963; Arnott and Hein, 1986; Ross and Bowring, 1990; Ross and Parrish, 1991a, b). This is supported by evidence of submarine canyon fills in the Lake Louise area that were incised by basinward-flowing turbidity currents (L in *Fig. 4.2*) (Arnott and Hein, 1986). Although, fragmentary because of sub-Cambrian erosion, shallow-marine upper slope and shelf facies are observed in the Mackenzie Mountains (northern Canadian Cordillera) and the upper part of the succession in the southern Canadian Cordillera (Ross *et al.*, 1995), while shallow shelf and terrestrial siliciclastic and subordinate carbonate rocks crop out in Mexico and western United States (*Fig. 4.3*) (Link *et al.*, 1993). Subordinate carbonates in deep-marine deposits of the southern Canadian Cordillera are suggested to have been sourced from the shallow-marine shelf to the east (*Fig. 4.2*), which may correlate to strata in the Mackenzie Mountains in the northern Canadian Cordillera (*Fig. 4.3*) (Ross, 1991).

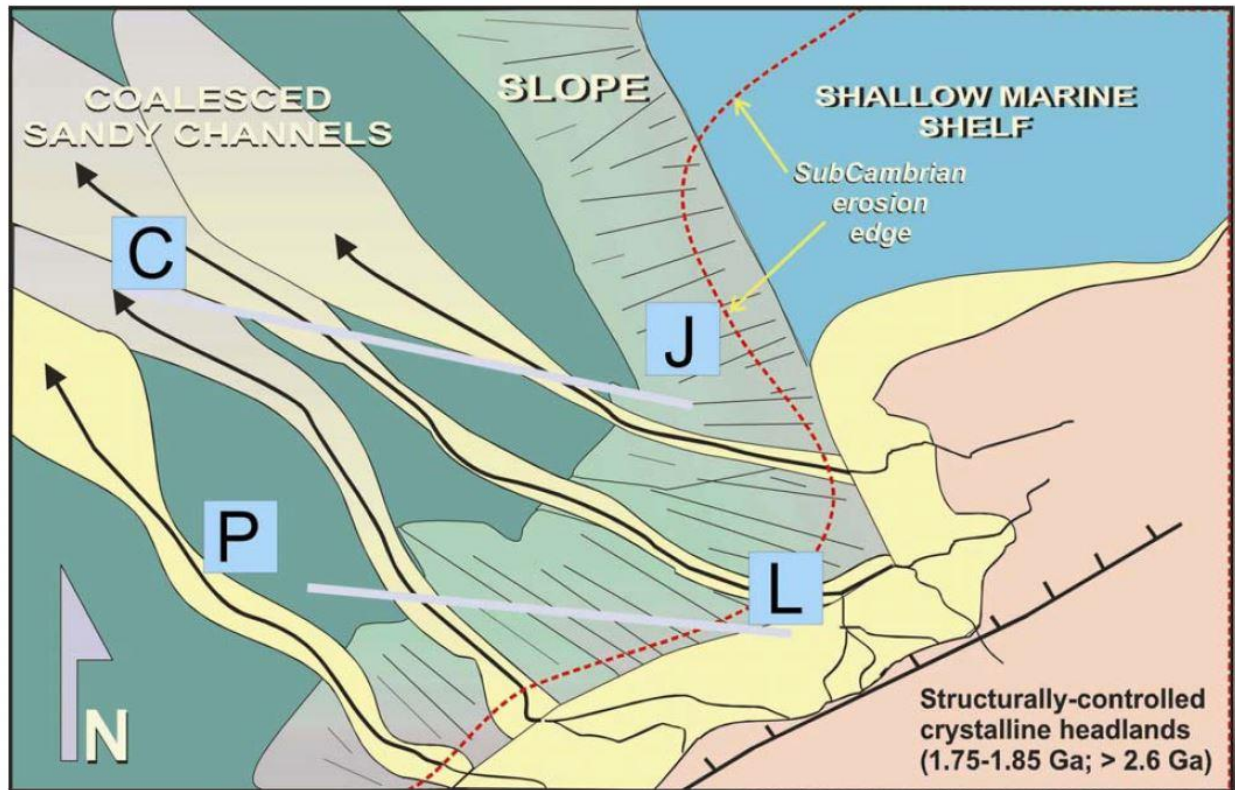


Figure 4.2: Schematic paleogeographic reconstruction of the Windermere turbidite system in the southern Canadian Cordillera. Paleoflow was toward the northwest (present day coordinates) in Lake Louise (L) down the lower slope and base-of-slope in Jasper (J) and Purcell (P) regions, respectively, and onto the basin floor in the Castle Creek study area (C) (Ross, 2000).

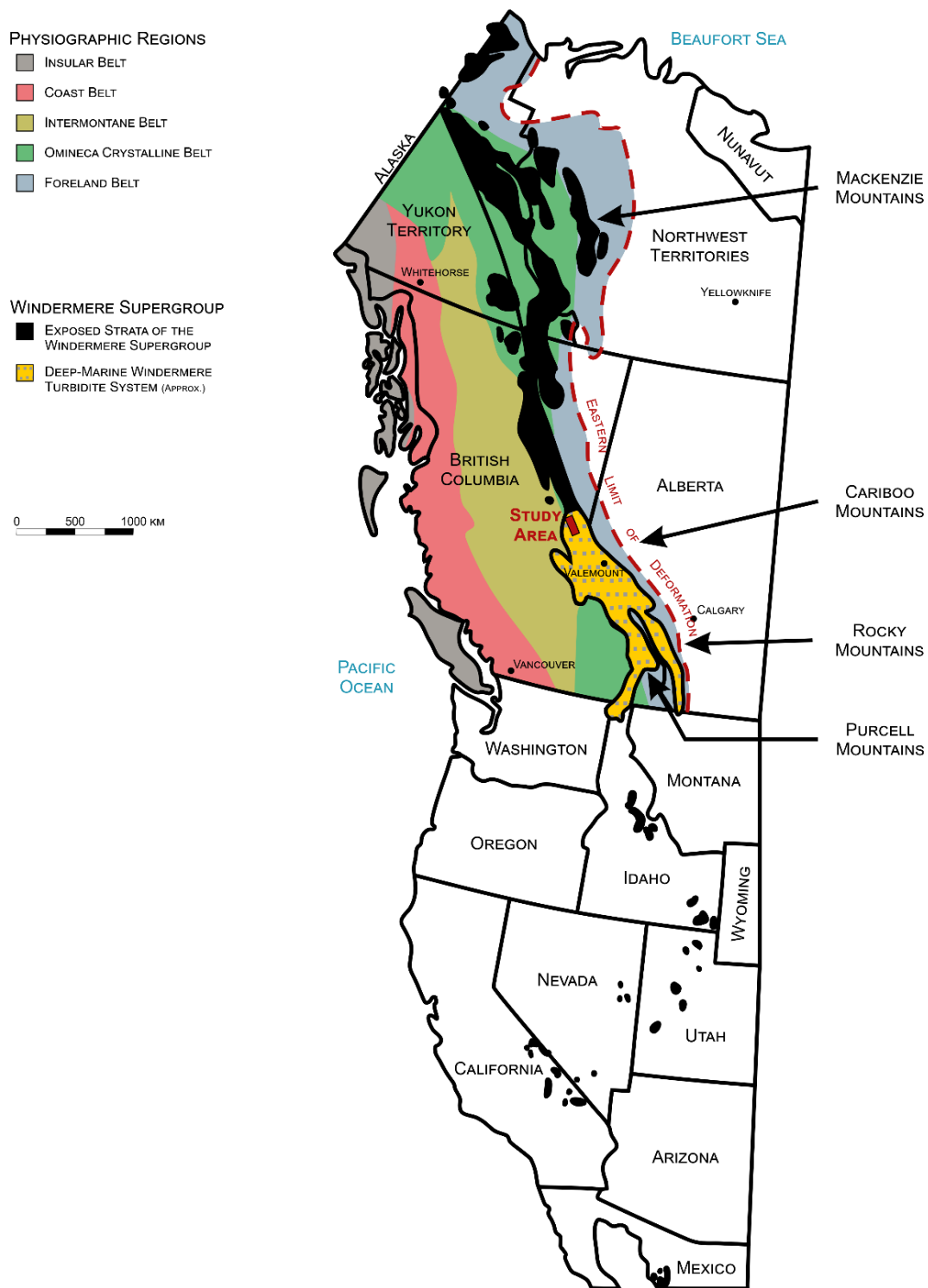


Figure 4.3: The Windermere Supergroup refers to Neoproterozoic strata that form a long, semi-continuously exposed arcuate belt in western North America with correlatives from the Sonoran Desert in northwestern Mexico to the Yukon-Alaska border (Ross *et al.*, 1989; Ross, 1991). Outcrops in the United States and Mexico are more discontinuous and fragmented compared to the extensive, continuous belt in the southern Canadian Cordillera. Strata in the United States crop out in structurally expanded portions of the Basin and Range Province and the Sevier Belt (Christie-Blick and Levy, 1989; Ross and Arnott, 2007) (redrawn and modified from Wheeler and McFeely, 1991; Gabrielse *et al.*, 1991; Ross *et al.*, 1995).

WSG strata dominate the exposed stratigraphy in the western Foreland Belt and Omineca Belt of the SCC (*Fig. 4.3*) (Ross *et al.*, 1989). The Omineca Belt was sandwiched between the ancient continental margin of western Laurentia (proto-North America) to the east after accretion that began in the mid-Jurassic, and a large composite of exotic allochthonous terranes to the west (Monger *et al.*, 1982). These strata have locally experienced several phases of structural deformation and metamorphism segregated between different structural panels resulting from the Cordilleran orogeny from 170-55 Ma (late Early Jurassic – Late Paleocene) (Price, 2000). Nevertheless, within both the Foreland Belt and Omineca Belt significant areas exist where strata have been subjected to only minor amounts of structural deformation and low-grade metamorphism (sub-greenschist to greenschist facies), and where reliable sedimentological data can be gathered, including the Jasper and Lake Louise areas, northern Cariboo Mountains (location of Castle Creek study area), and the eastern Purcell Mountains (*Fig. 4.3*) (Smith *et al.*, 2011). In the SCC, strata of the WSG strata crop out over an area of about 35,000 km² (Ross *et al.*, 1995; Ross and Arnott, 2007). If the effects of contractional Mesozoic deformation are considered, and a conservative shortening factor of 30% applied (Brown *et al.*, 1986; McDonough and Simony, 1988; Price, 2000; Arnott and Ross, 2007), the exposed part of the deep-water turbidite system is suggested to have been c. 80,000-100,000 km², which would make it comparable in size to modern submarine fans like the Mississippi, Astoria, Magdalena or Monterey (*Fig. 4.4*) (Barnes and Normark, 1985), and also the largest known ancient turbidite system (Ross *et al.*, 1995; Ross and Arnott, 2007).

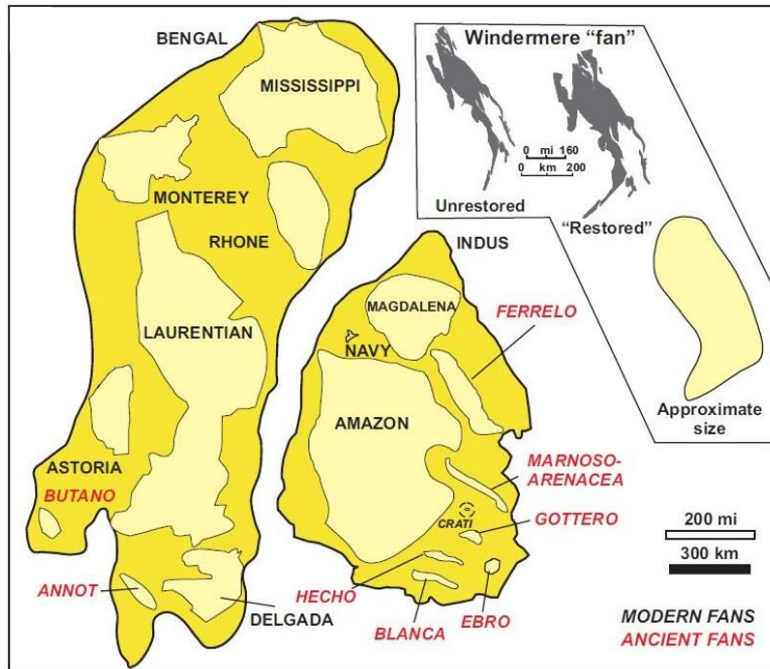


Figure 4.4: Size comparison of the restored Windermere turbidite system (Windermere “fan”) with modern (black text) and ancient (red text) turbidite systems (Ross, 2001 modified from Barnes and Normark, 1985).

The deep-marine portion of WSG strata are best exposed and most thickly preserved in the northern Cariboo Mountains of east-central British Columbia where a world class example of an ancient passive margin turbidite system comprising a complete succession of deposits from basin floor to continental shelf (Ross, 1991). At the Castle Creek study area (*Fig. 4.1C, D*) these strata are exposed in a recently-deglaciated, vegetation free, vertically-dipping section that extends for 8 km parallel to bedding and 2.5 km perpendicular to bedding. However, glacial overburden poses local challenges to lateral stratigraphic correlation. Nevertheless, this study area provides an unparalleled opportunity to study the deep-marine sedimentary record on scales ranging from millimetres (sub-seismic-scale) to kilometres (seismic-scale).

4.2 Methodology

In this study slope deposits of the Isaac Formation (*Fig. 4.1C, D*) crop out for over 450 m laterally and 11-13 m vertically and comprises two general units (*Fig. 4.5*): a lower, highly-

amalgamated sandstone unit, and an upper, heterolithic unit composed of sandstone interstratified with siltstone or mudstone, and is the primary focus here. A total of four stratigraphic logs ranging from 11.21-13.35 m were measured in bed-by-bed (*i.e.* centimetre-scale) detail. In each log, bed thickness, lithology, grain size, basal contact morphology, colour, primary and secondary sedimentary structures, and size, shape, abundance, and orientation of mud intraclasts were noted and described. Beds were then walked out for *c.* 400 m along strike and mapped on high resolution aerial photos.

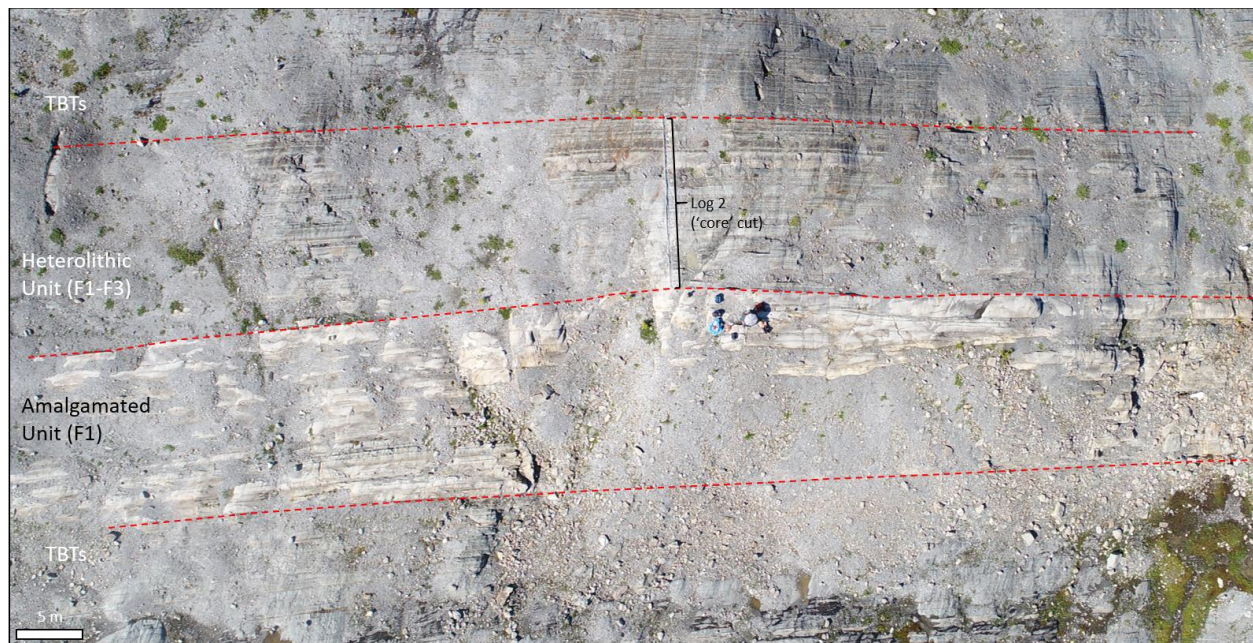


Figure 4.5: Aerial drone photo of a 100 m lateral segment of the study area where all component facies are delineated. The different units (separated by red dashed lines) can be differentiated based on their distinct colouring as the thin-bedded turbidites (TBTs) are black/dark grey, the amalgamated unit is tan, and the heterolithic unit is striped with a variety of tans, grey, and rusty orange colours. This is located over log 2 which was continuously cut for 12.5 m vertically to expose an unweathered surface that was logged in detail in order to better define bed contacts, especially in rhythmic successions of fine-grained F3 beds, thus providing a more accurate vertical representation of the stratigraphy. The roughly 30 m thick strata of the study area are sandwiched between thin-bedded turbidites and comprises a basal amalgamated unit overlain by a finer-grained heterolithic unit, which is the primary focus of this research. Note that for correlation purposes, glacial debris covering the outcrop was removed, such as the vertical transect in the upper left under TBTs, in order to more confidently correlate beds by walking then out between logs.

71 hand samples were collected and petrographically analyzed. Additionally, 23 thin sections from all facies were stained with a 1:1 solution of Alizarin Red S and potassium

ferricyanide, in order to differentiate, respectively, calcite from dolomite and ferroan calcite from dolomite (Dickson, 1966). Thin sections were analyzed for grain size (minimum, maximum, and average), sorting, grain contacts, mineralogy, sedimentary structures, and fabric using qualitative descriptors. 41 thin sections were then point counted, 30 points every 0.5 mm, along a vertical transect through each thin section to capture vertical trends in grain size. Despite low-grade greenschist facies metamorphism (200–450 °C) (Murphy, 1987; Zou *et al.*, 2013) primary physical sedimentary structures are well preserved in strata, and therefore classification and terminology applied to sedimentary rocks will be used. Metamorphic alteration has transformed detrital clay mineral particles to chlorite and mica (muscovite and sericite). Accordingly, this has modified the size and mineralogy fine detrital grains, and therefore all interstitial material, regardless of crystal size, is considered, in addition to unaltered quartz silt, to be matrix. Another important alteration caused by metamorphism is dynamic quartz recrystallization which includes bulging recrystallization and sub-grain rotation (Stipp *et al.*, 2002). Dynamic quartz recrystallization tends to be more intense in framework-supported strata with low clay (matrix) content because strain is accommodated by deformation and recrystallization of the rigid framework grains rather than slip in the fine-grained matrix (Drury and Urai, 1990). These changes, especially development of sub-grains, were considered when point counting thin sections. Scanning electron microscopy (SEM) was conducted on six samples to better characterize grain size and micro-sedimentary structures, as well as validate mineralogy in fine-grained rocks, especially facies 3 and facies 4. Additionally, a continuous 12.5 m section was cut with a hand-held rock saw in the upper heterolithic part (‘core cut’ in *Fig. 4.5*) to better resolve the millimetre- and centimetre-scale characteristics in these fine-grained strata.

4.3 Facies

Macro- and microscopic analysis of strata in the lower Isaac Formation identified four lithofacies in the study area. Based on changes in physical sedimentary structures and upward trends in grain size, facies 1-3 stack to form an idealized facies succession forming a bed (*Fig. 4.6*). Note that not all beds contain all facies; in fact, only 6% of beds contain all three facies, and where present comprises Facies 1 overlain by Facies 2 overlain by Facies 3. In the northwestern part of the study area Facies 4 crops out exclusively. Refer to *Table 4.1* for a summary of macroscopic and microscopic details of each facies and also Billington (2019) for more detailed facies descriptions.

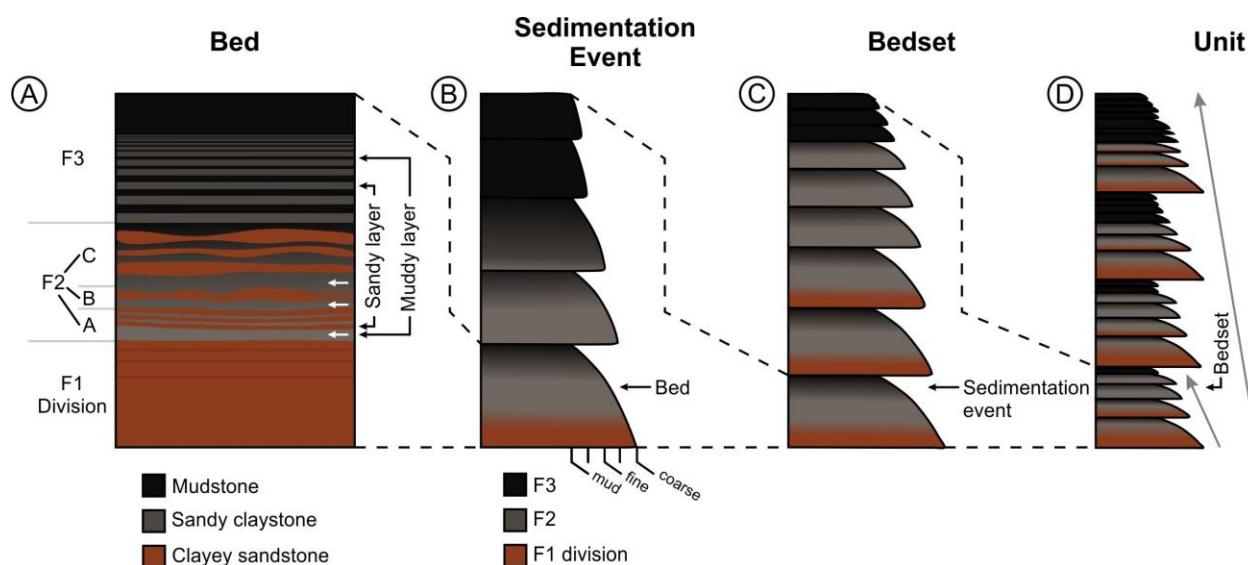


Figure 4.6: Hierarchy of stratal terminology used here from micro to macro scales. A) Idealized facies succession where a complete bed (centimetres to 10s cm thick) comprises an upward fining structureless overlain by planar-stratified F1 sandstone division sharply overlain by a thick mud layer (denoted by white arrows). This, then, is overlain by subfacies divisions F2.A, F2.B, and F2.C. Capping the bed is the F3 division, which is plane-parallel-laminated mudstone in the bottom part sharply overlain by structureless claystone. Note that F2 and F3 divisions are interstratified with sandier and muddy layers (see below). B) Schematic sedimentation event (10s cm to metres thick) comprising multiple beds (distinct basal and top contact) that typically fine and thin stratigraphically upwards and represent deposition during a single sedimentation event. C) Schematic bedset (metres thick), which is composed of multiple sedimentation events that generally fine and thin upwards. D) Schematic unit (10s m thick) that generally fine and thin vertically (indicated by grey arrows), which is made up of numerous bedsets that also fine and thin, and represent the largest stratigraphic element discussed here.

Table 4.1: Macroscopic and microscopic characteristics of facies F1 to F4. Facies 1 to 3 are observed in the heterolithic unit whereas F4 is exclusive to the fine-grained unit in the northwest part of the study area (Fig. 4.21).

		Heterolithic Unit			Fine-Grained Unit
		F1	F2	F3	F4
	Description	Massive or graded, structureless or planar/wavy-stratified clayey sandstone	Graded, plane parallel-, cross- or wavy-stratified clayey sandstone with intercalated structureless sandy claystone	Graded, plane parallel-laminated sandy claystone to massive or graded, structureless claystone	Massive, interstratified, structureless clayey sandstone and claystone
	% Beds observed in	12%	26%	99%	n/a
Basal Contact	Sharp and planar	58%	87%	96%	yes
	Undulatory	21%	10%	4%	yes
	Loaded	21%	3%	< 1%	no
Sedimentary Structures	Structureless	45%	0%	0%	yes
	Planar-stratified	52%	34%	100% (laminated)	yes
	Wavy-stratified	3%	55%	0%	no
	Cross-stratified	0%	11%	0%	yes
Thickness	Minimum	2 cm	0.5 cm	0.5 cm	1 cm
	Maximum	112 cm	34.5 cm	15 cm	> 20 cm
	Average	18 cm	5.8 cm	3.2 cm	n/a
	Mode	13 cm	2.8 cm	2 cm	n/a
	% Bed Thickness Minimum	16%	4%	3%	n/a
	% Bed Thickness Maximum	100%	88%	100%	n/a
	% Bed Thickness Average	57%	44%	81%	n/a
Components	Framework Grains (> 62.5 μm)	58%	44%	30%	< 30%
	Matrix (> 62.5 μm ; silt and recrystallized clay)	8%	36%	70%	> 70%
	Carbonate Cement	34%	20%	< 1%	< 1%
Grain Size Proportion (Framework)	Very coarse (1,000-2,000 μm)	18%	0%	0%	n/a
	Coarse (500-1,000 μm)	22%	0%	0%	n/a
	Medium (250-500 μm)	32%	14%	9%	n/a
	Fine (125-250 μm)	21%	59%	45%	n/a
	Very fine (62.5-125 μm)	7%	27%	46%	n/a
Facies Transitions	Overlying Facies within Bed	F2 (54%) F3 (42%) none (top facies) (4%)	F3 (99%) none (1%)	none (top facies)	n/a
	Underlying Facies within Bed	none (basal facies)	F1 (24%) none (76%)	F1 (5%) F2 (27%) none (68%)	n/a

4.3.1 F1: Massive or graded, structureless or planar/wavy-stratified clayey sandstone

Description

Facies 1 is observed in 12% of beds and where observed typically overlies a sharp planar contact at the base of the bed. F1 ranges from 1-112 cm thick but averages 18 cm (mode 13 cm)

and makes up 16-100% (average 57%) of the bed thickness. It preferentially thickens and coarsens to the NW and consists of massive (76% of beds), medium-grained carbonate cemented sandstone (matrix content averages 8%, but always is < 10%) (*Fig. 4.7*). In an individual bed carbonate cement content changes little along strike in a bed but decreases stratigraphically upwards in each stratigraphic log measured (inversely related to %mud in the bed). Strata of F1 are planar-stratified (52%), structureless (45%), or rarely wavy-stratified (4%) (*Fig. 4.8*) and are overlain sharply by F2 (54%), F3 (42%), or rarely makes up the entire bed (4%). Thicker (average 20.9 cm), coarser-grained (medium sand) F1 layers are overlain by similarly thicker (average 9.6 cm) F2 layers and thicker (average 7 cm) F3 layers. Conversely, thinner (average 8.4 cm), finer-grained (fine sand) F1 layers are overlain by thinner (average 5.5 cm), finer-grained (silt) F3 layers, and the F2 layer is absent.

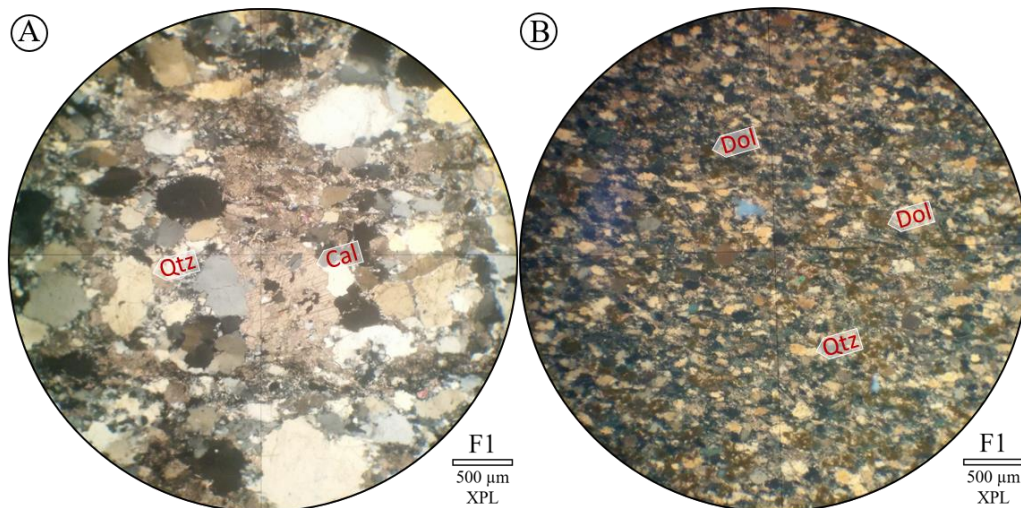


Figure 4.7: Photomicrographs illustrating the range of grain size in F1 strata. Both samples have about the same amount of carbonate cement (*c.* 25%). A) Very coarse- to medium-grained clayey sandstone with negligible (< 2%) matrix (recrystallized chlorite and micas); cement occurs as large calcite (Cal) crystals that locally exhibit a poikilotopic fabric. B) Fine-grained sandstone with about 5-10% matrix content; cement occurs as small dispersed patches of brown rhombohedral dolomite (Dol) and minor (< 5%) calcite.



Figure 4.8: Structureless, graded or massive F1 sandstone overlain abruptly (red dashed line) by strata of F3. Base of individual beds indicated by dashed yellow lines. A) Planar-stratified sandstone (stratification highlighted by white dotted lines on right side of image). B) Wavy-stratified sandstone. Note the white dotted lines that highlight the wavy-stratification. C) Structureless, coarse-tail graded sandstone.

Interpretation

F1 represents deposition from sand-rich turbidity currents. The comparatively low (< 10%) matrix content in F1 strata indicates deposition from turbidity currents enriched in sand in the near-bed region. Structureless and massive strata indicate rapid and direct deposition from suspension whereas diffuse to well-developed planar stratified units indicate reduced rates of fallout and episodic deposition from laminar shear layers (*sensu* Vrolijk and Southard, 1997; see also Sumner *et al.*, 2008).

4.3.2 F2: Graded, plane parallel-, cross- or wavy-stratified clayey sandstone with intercalated structureless sandy claystone

Description

Facies 2 occurs in 26% of beds and most commonly overlies a planar base, or sharply overlies F1, and virtually always (99%) is overlain by F3. Beds generally thicken toward the NW and range from 0.5-34.5 cm (average 5.8 cm, mode 2.8 cm) and make up 4-88% (average 44%) of the bed thickness.

F2 is composed of intricately interstratified millimetre- to centimetre-thick, sand-rich (sandy layer) and sand-poor (muddy layer) layers that show no systematic change in thickness upward. From bed to bed sandy layers vary the most in their matrix content (10-60%), which within a F2 division in a bed tends to remain consistent both laterally and vertically and are subdivided into two end-members: matrix-poor (10-30% matrix) and matrix-rich (30-60% matrix). However, rarely (2%) matrix-poor strata are overlain sharply by matrix-rich strata (*e.g.* Fig. 4.11C). Additionally, matrix-poor layers contain 15-35% ferroan calcite cement that oxidizes to a distinctive rusty orange colour in the field. In contrast, muddy layers lack carbonate cement and are grey. Sand layers uncommonly exhibit subtle normal grading and generally consist of fine-

grained sandstone commonly with dispersed medium sand grains (up to coarse sand). Muddy layers are massive, moderate to well sorted, and composed primarily of recrystallized clay and (quartz) silt grains with 15-30% dispersed medium to very fine sand. Significantly, the range of grain sizes in a sandy and overlying muddy layer is similar, and stratigraphically upward show an overall upward fining (*Fig. 4.9*).

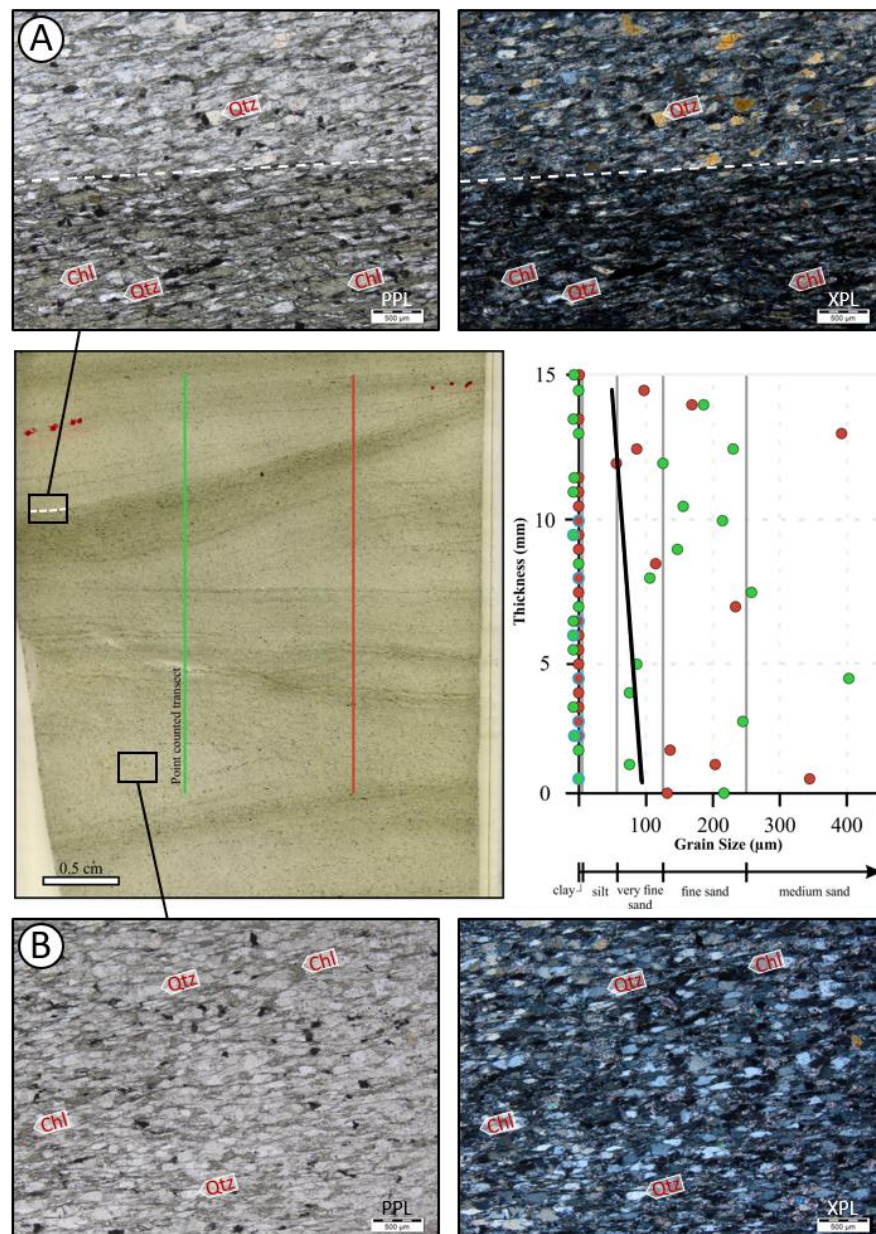


Figure 4.9: Grain size analysis (middle right) along two stratigraphically-upward transects (green and red lines) in a thin-bedded F2 (F2.B) division. Grain size (green and red dots) was measured every 0.5 mm over a vertical length of

15 mm (total of 30 points). Although somewhat muted due to the poorly sorted nature of the strata, grain size does show a minor upward fining (indicated by the black linear trendline). Note that data points lying on the vertical axis represent recrystallized clay or carbonate cement. The bottom photomicrographs are taken through a sand layer, and top photomicrographs depict the contact (separated by the white dashed line) between a mud layer overlain sharply by a sand layer.

Most commonly (81%) the F2 part of a bed consists of a single style of stratification (F2.A, F2.B, or F2.C). However, beds with two, and rarely three, styles of stratification are observed; specifically, cross-stratified (F2.B) overlain by wavy-stratified (F2.C), plane-parallel stratified (F2.A) overlain by wavy-stratified (F2.C), or plane-parallel stratified (F2.A) overlain by cross-stratified (F2.B) overlain by wavy-stratified (F2.C) (see also *Fig. 4.10*). Where F2 overlies a planar-stratified F1 it too is planar-stratified (F2.A), whereas where F1 is absent and F2 strata overlie the basal contact, F2 strata are typically wavy-stratified.

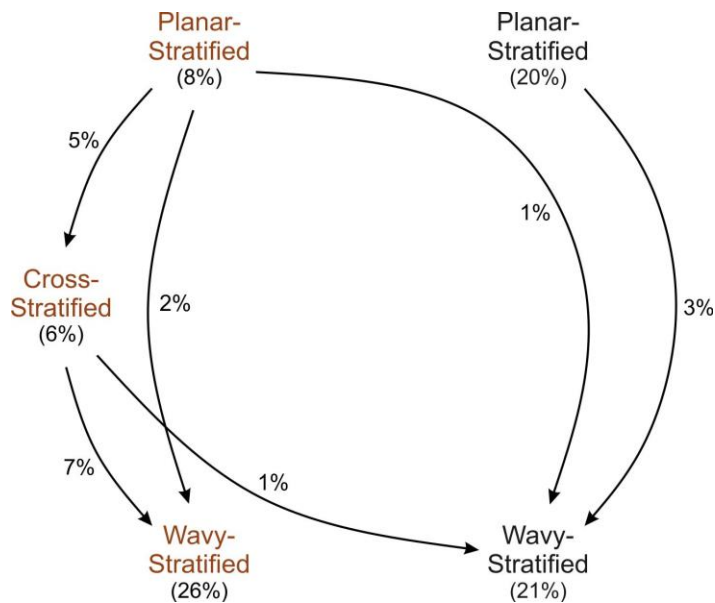


Figure 4.10: Flow chart of the common styles of stratification observed in matrix-poor, carbonate cemented (orange text), or matrix-rich (black text) F2 strata, and the corresponding stacking patterns (indicated by arrows) where two or more styles of stratification stack to form a single F2 division.

Plane-parallel F2.A strata are typically rich in matrix and the sand layers are grey, whereas less common matrix-poor sand layers are calcite cemented and are distinctively orange colour (*Fig. 4.11*). Matrix-rich F2.A strata tend to be thinner and laminated with sharper, more distinct

boundaries separating the interstratified sand-mud layers. Matrix-poor strata, on the other hand, tend to be thicker and sand and mud interlayers are thicker (commonly > 1 cm) and bounded by more diffuse boundaries. Average grain size for F2.A is upper fine sand, and generally is coarser-grained in matrix-poor strata compared to matrix-rich strata.

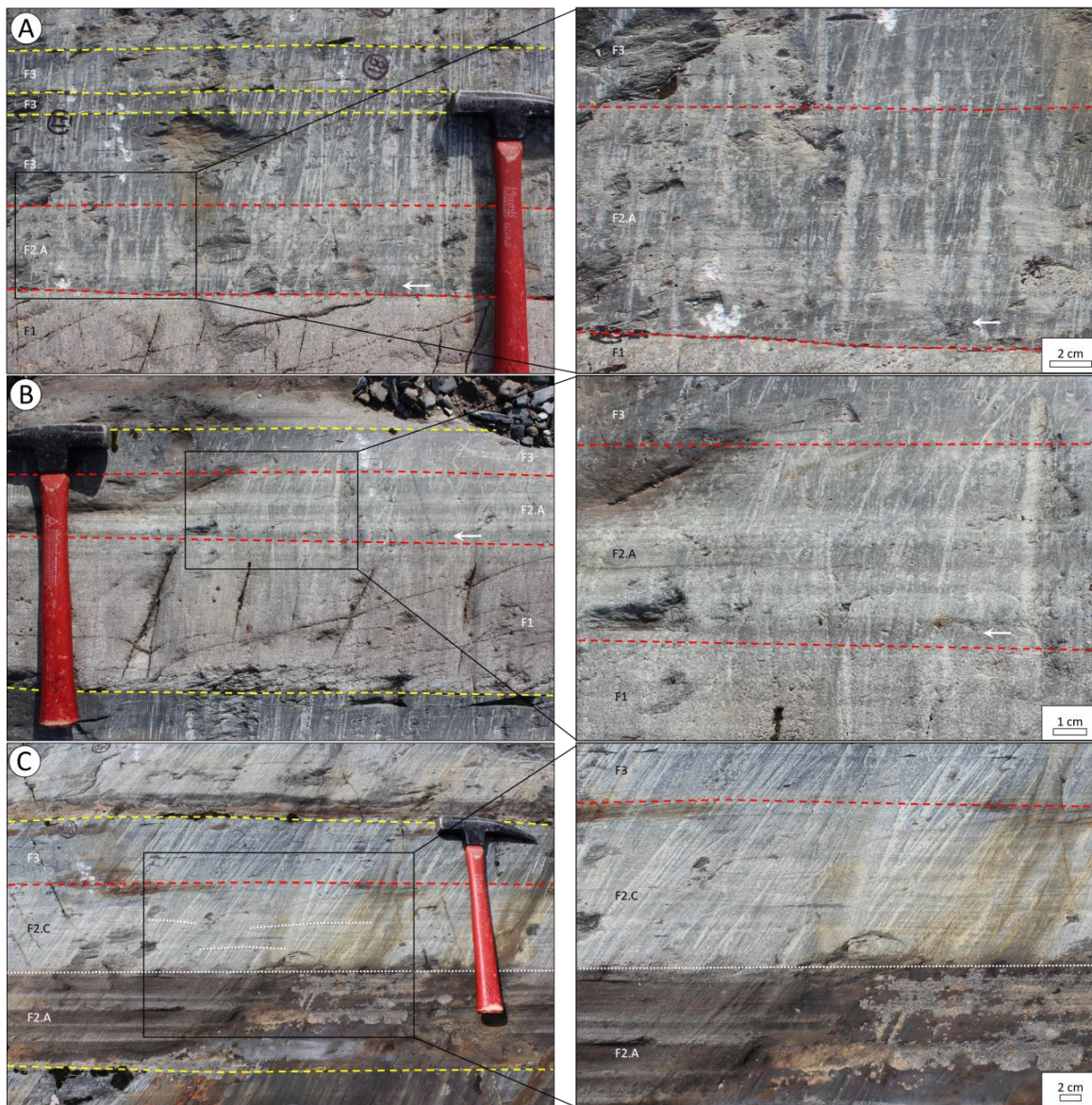


Figure 4.11: Examples of matrix-rich (A, B) and matrix-poor (C) planar-stratified F2. Photos are presented as pairs, with the photo on the right being an expanded view of those on the left in order to help highlight the stratification. Matrix-poor, carbonate cemented F2 strata tend to be more thickly stratified than similarly thick matrix-rich F2 layers. A) Matrix-rich, planar-stratified F2.A division sharply overlying F1 and gradationally overlain by F3. Note the thick mud layer separating F1 from F2.A (indicated by the white arrow). B) Matrix-rich, planar-stratified F2.A with thickly stratified sand and mud layers that overlies an anomalously thick mud layer that sharply overlies F1. C) Matrix-poor,

carbonate cemented planar-stratified clayey sandstone (F2.A) at the base of the bed overlain sharply by matrix-rich, wavy-stratification (F2.C).

Cross-stratified F2.B strata have an average grain size of lower fine sand, and dip at angles between 4-18°, and on average about 10°. F2.B strata are always matrix-poor and calcite cemented with abundant ferroan calcite (up to 34% in thin section) that form a spectrum of cross-stratified styles, often with abundant internal truncation surfaces. These styles include isolated cross-stratified sets encased in mud, cross-stratified sets that are sand-rich in their upper part and feather into mud in their (downflow) troughs, or climbing ripple-stratification that feathers into mud in up- and downflow directions (*Fig. 4.12*).

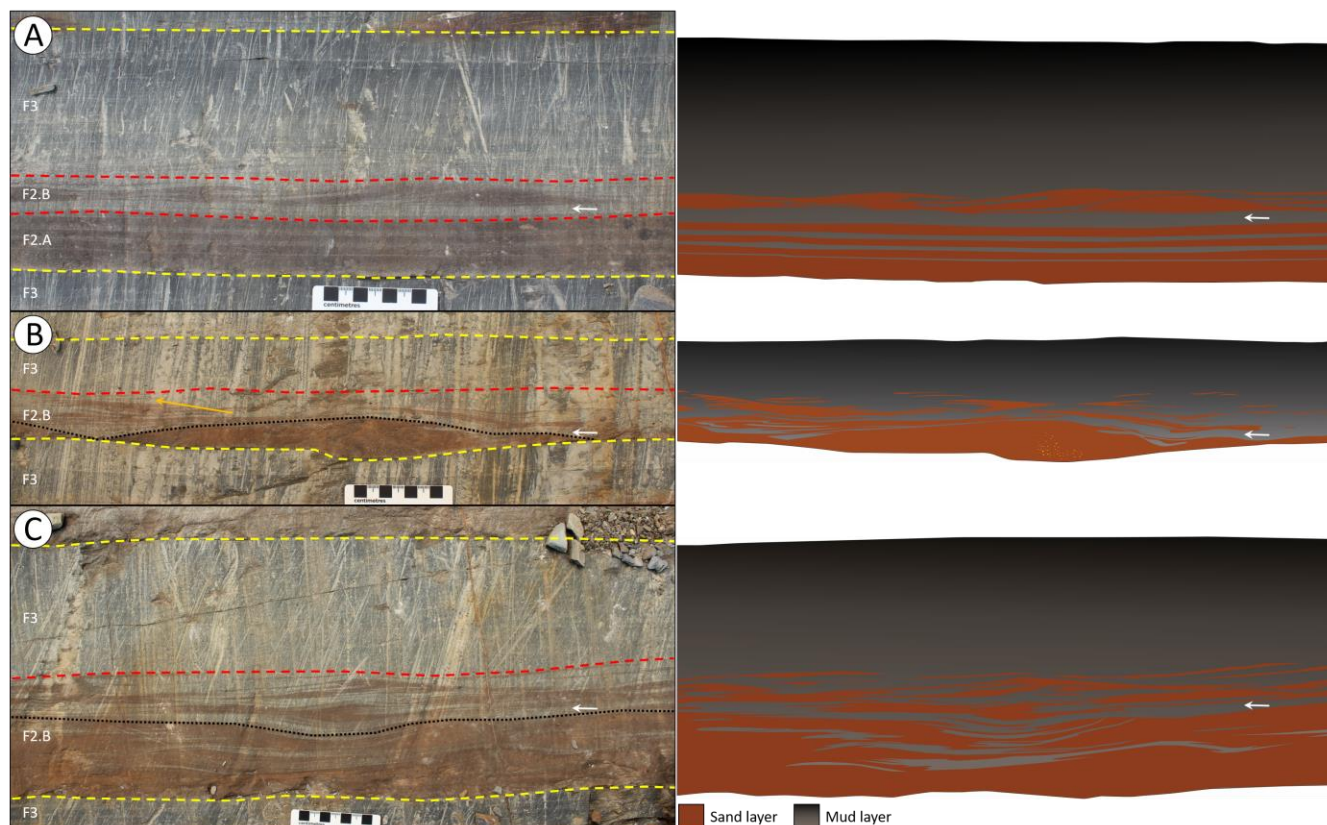


Figure 4.12: Photo pairs (photo and accompanying line diagram) illustrating F2.B divisions consisting of two stacked styles of stratification; contact indicated by black dotted line. Note that the different styles are separated by a thick laterally continuous mud layer (indicated by white arrows). Line drawings on the right help to highlight the intricate and often subtle interstratification of the sand and mud layers. A) Basal planar-stratified, carbonate cemented F2.A strata overlain abruptly by a thick (1.5 cm) laterally continuous mud layer (indicated by white arrows), which then is overlain by isolated sand-rich mounds draped with mud layers, ultimately capped by F3 strata. In the F2.B division paleoflow is very generally toward the left (SE). More significantly, the F2.B division consists of discrete sand-rich mounds separated by continuous mud-rich laminae, and is interpreted to be a single ripple set that experienced recurring changes in local sediment supply and ripple behaviour. B) Isolated ripple filling a scour; paleoflow is very generally towards the right (NW). Note the abundant, dispersed coarse sand grains concentrated in the ripple trough. This, then, is overlain by a thick (up to 6 mm) mud layer (base delineated by black dotted line) that drapes the ripple formset and separates it from low-angle climbing ripple cross-stratification (orange arrow indicates angle of climb) that feathers into mud in both up- and downflow directions. Very generally paleoflow is toward the left (SE), which is more or less opposite to the underlying ripple formset. C) Thick (up to 1 cm), laterally continuous mud layer (base denoted by black dotted line) that abruptly separates higher-angle cross-stratified sets with muddy troughs at the base of the bed from more widely spaced, more isolated cross-stratified sets with sandy stoss and crestal areas that then feather into thicker muddier troughs. In all photos bed contacts are denoted by yellow dashed lines whereas red dashed lines separate different facies or subfacies.

Wavy-stratification (F2.C) is the most common style of stratification in F2 divisions. F2.C comprises about sub-equal occurrences (see *Fig. 4.10*) of matrix-poor, calcite cemented, and matrix-rich end-members (*Fig. 4.13*). Compared to F2.B, cross-stratification in F2.C is always much lower angle (typically c. 4°), and mud layers tend to be more laterally continuous (> 1 m) rather than filling depressions on one or both sides of the raised part of the ripple formset in F2.B. Sand and mud layers tend to be thinner and more intricately interlaminated. Spacing and height of the undulation is of the order of 15-30 cm and a few millimetres, respectively. On average grain size in F2.C is lower fine sand, where the range of grain size is similar between matrix-poor and matrix-rich end-members and the difference is in their proportions of matrix.

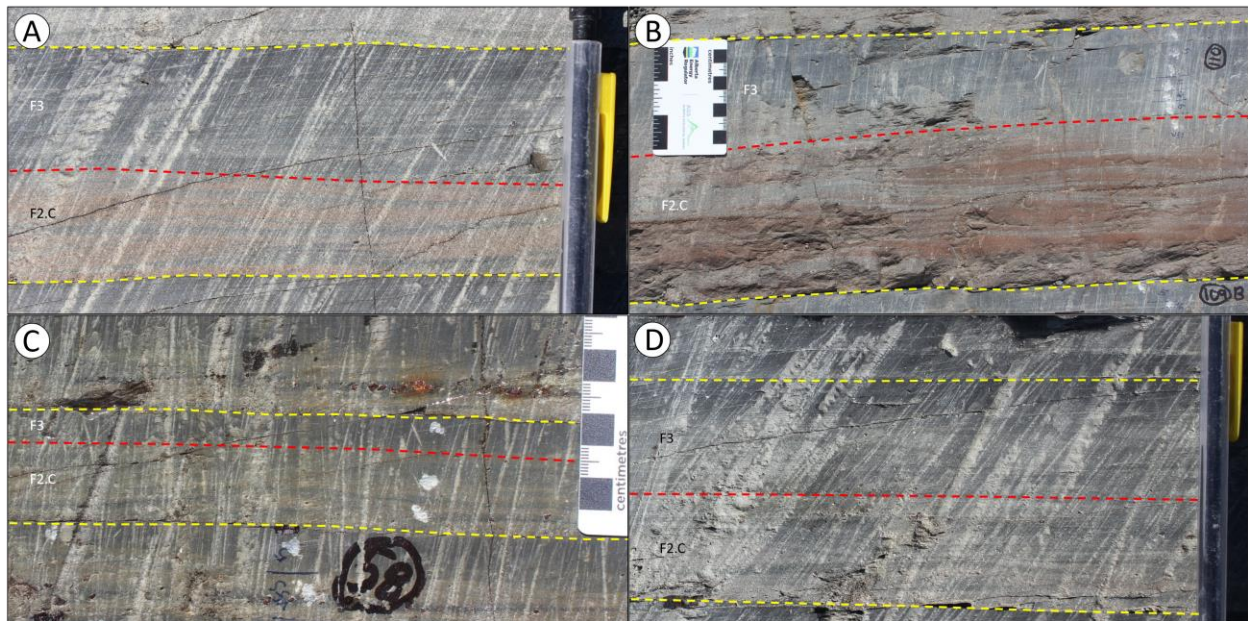


Figure 4.13: Wavy-stratified strata of F2.C. Significantly, and in comparison to F2.B, mud layers in strata of F2.C can commonly be traced for metres along strike. A) Thin, matrix-poor, carbonate cemented, wavy-stratification overlain gradationally by F3. B) Anomalously thick (11 cm) matrix-poor, carbonate cemented, wavy-stratified division. C) Thin, matrix-rich, wavy-stratified F2.C division. D) Matrix-rich, subtly wavy-stratified F2.C layer gradationally overlain by F3 strata. In all photos basal bed contacts are denoted by the yellow dashed lines whereas red dashed lines separate different facies or subfacies in a bed.

Where F2 overlies F1 (*e.g. Fig. 4.11A, B*), or comprises more than one type of stratification (*Fig. 4.12B, C*) the contact often exhibits evidence of erosion (*Fig. 4.14*) and then is overlain by

an anomalously thick (0.5-2 cm) mud layer. Compositionally these mud layers are similar to the intercalated mud layers, being composed primarily of recrystallized clay and silt grains with 15-30% dispersed medium to very fine sand, but differ only in thickness and along-strike continuity (extend for 10s metres and change little in thickness along their length; *Fig. 4.15*). Additionally, in some cross-stratified F2.B units, internal truncation surfaces are overlain not only by a thick mud layer but also coincide with a reversal in paleocurrent direction (*Fig. 4.12B, 4.14A*).

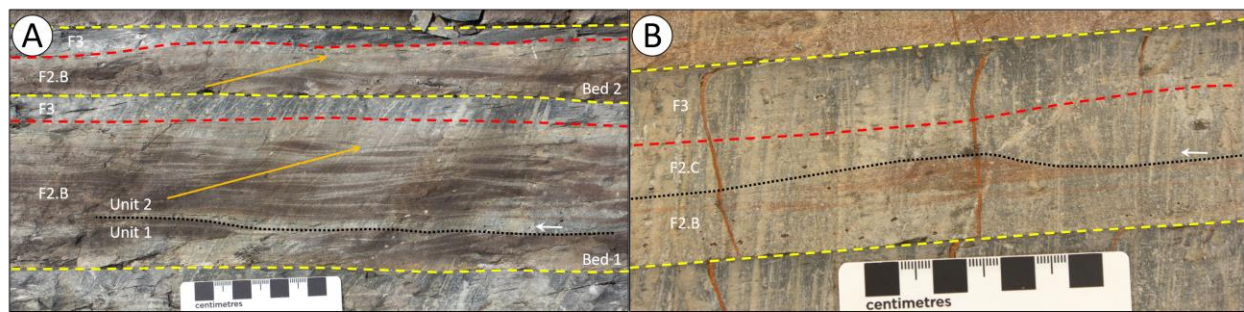


Figure 4.14: A) Two beds composed of an F2.B unit overlain gradationally (red dashed lines) by F3; bed bases indicated by yellow dashed lines. Two stacked F2.B units separated by an anomalously thick (0.5-2 cm) mud layer (indicated by white arrows; base delineated by black dotted lines). Note that the thick mud layer overlies an erosional surface that truncates cross-stratification (highlighted by white dotted lines) at the top of the lower F2.B unit. Unlike the lower unit, cross-stratification in the upper unit is both migrating and climbing (orange arrows) toward the upper right of the photo, which is opposite to the direction of cross-stratification in the lower unit. Note also the transition of the sand-rich crestal parts of ripples into mud-rich deposits in the downflow trough. B) Isolated ripple that in the downflow direction (left) interfingers with more mud-rich strata, and in the upflow direction is truncated and then sharply overlain by a thick mud layer (white arrow) and overlain by F2.C strata.

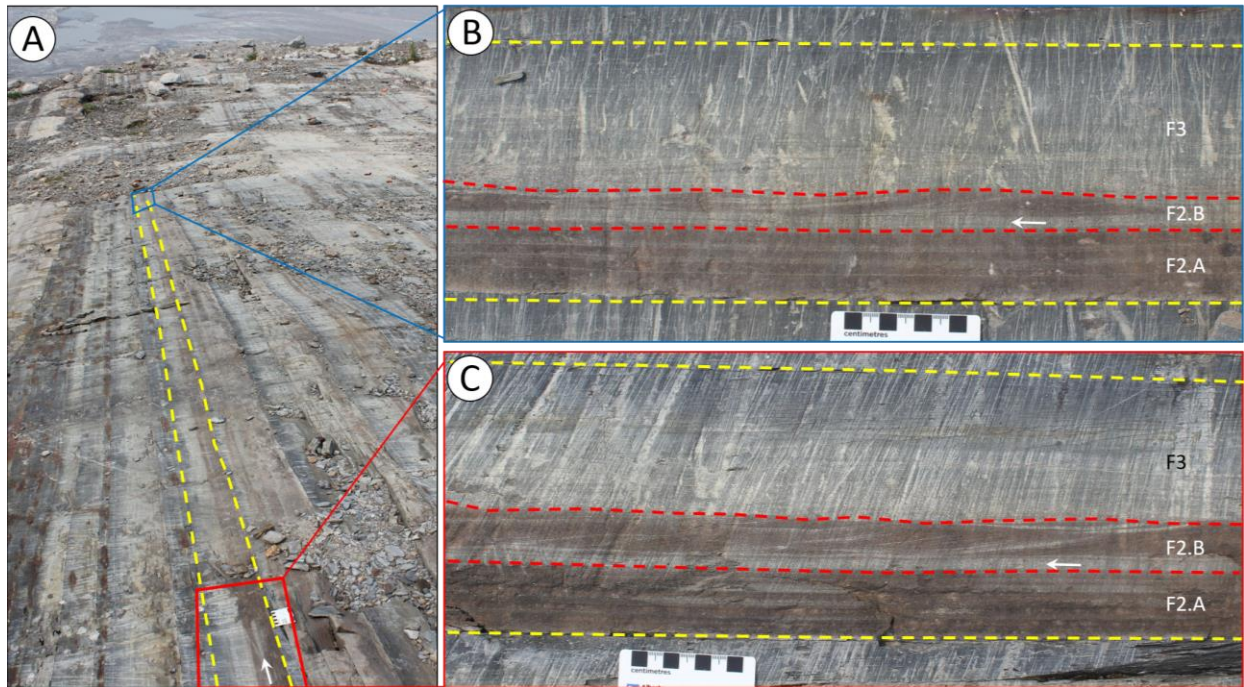


Figure 4.15: A) Thick (0.5-2 cm), laterally extensive mud layers that extend for at least several metres along strike (indicated by white arrow; B and C are separated by about 5 m along strike). Almost always thick mud layers overlie a scour surface and separate different styles of stratification in F2 strata. Bed bases are indicated by yellow dashed lines. Note that stratigraphic top is to the left; black squares on scale card are 1 cm (left of card). B and C) Closeups of (A) showing a basal matrix-poor, carbonate cemented, planar-stratified F2.A overlain sharply by a thick mud layer and then by a single cross-stratified set consisting of isolated sand-rich regions that transition downflow (left/SE) into more interstratified sand-mud cross-lamination – F2.B is then capped gradationally by F3 strata. Note that in both (B) and (C) subfacies F2.A and F2.B are separated by the same thick mud layer, which is indicated by the arrow in (A).

Interpretation

F2 strata represents deposition from turbidity currents that experienced a complex history of deceleration and patterns of deposition. In general, most flows, at least initially, were sufficiently competent to transport medium, and in some cases, coarse sand (*Fig. 4.12B*). With time flows waned and strata became progressively overlain by fine and then very fine sand with abundant interstitial mud, which also is rhythmically intercalated with mud interlayers. In addition to these textural changes, sedimentary structures also show a consistent upward change from plane-parallel-, to cross-, to wavy-stratification (*Fig. 4.10*). Although, the full suite of stratification styles is only rarely observed in a single bed, the more common occurrence is partial suites, but always in the same order.

Plane-parallel-stratification (F2.A) indicates uniform bedload transport on the bed where flow speeds were either too fast for the development of ripples (equivalent to upper-stage plane bed), or the requisite hydraulic conditions in the near-bed region were unable to initiate and amplify bed-surface defects that eventually would develop into mature ripples. As a result, sediment transport on the bed remained spatially uniform and resulted in deposition of planar-stratified sediment, possibly at flow speeds that otherwise would have formed ripples in a clear-water flow (Tilston *et al.*, 2015). With further flow deceleration sediment continued to fall from suspension into the near-bed region and more fully stratified the flow, which in turn promoted the development and strengthening of a near-bed Kelvin-Helmholtz instability. This, then, coupled with sediment transport along the bed and resulted in spatial changes in the pattern of sediment transport and deposition, and with the inception and growth of bed surface defects that ultimately formed downstream-migrating ripples (F2.B) (Arnott, 2012; Tilston *et al.*, 2015), which sequestered sand in their topographically elevated part (crest) and allowed mud to accumulate in the adjacent lower-energy trough. These processes formed strata of F2.B. Uncommonly (< 3% of F2 beds), sediment fallout rate was sufficiently high that the bed aggraded fast enough that the ripples were forced to climb (Allen, 1971; Ashley *et al.*, 1982). With continued reduction in flow speed, and the reduced availability of (coarse) particles to settle into the near-bed region, the near-bed hydrodynamic instability, although still capable of generating bed defects, was unable to cause sufficient amplification of the defect to sufficiently alter (sediment) transport patterns on the up- and downflow sides of defects. As a consequence, a wavy bed surface (F2.C) remained the stable bed state.

Where F2 overlies F1 strata or is composed of two or more different types of stratification the contact is marked by an anomalously thick, laterally continuous mud layer up to 2 cm thick.

Additionally, where F1 is overlain by F2 there is a sharp upward increase in mud content. Significantly, these thick mud layers overlie an erosion surface that is interpreted to have formed by low frequency, high energy turbulent eddies that scoured the sea bed and increased the sediment load in the lower part of the flow. This new sediment would have substantially reduced the turbulent kinetic energy (TKE) in the flow as energy was consumed by accelerating and maintaining these particles in suspension. Note that although the incorporation of new suspended sediment into the basal part of the flow may have decreased TKE, it had little effect on the most energetic scales of fluid motion (*i.e.* integral scales; *Fig. 4.16*) (Bennett *et al.*, 2014). Nevertheless, scouring of the bed would have substantially increased the near-bed sediment load, and especially in this case, the volume of clay mineral particles. The abundance of these particles would have led to particle aggregation and ultimately gelling followed by *en masse* deposition of thick, laterally continuous, poorly-sorted, massive mud layers (McAnally *et al.*, 2007). Following deposition of these fine-grained layers, and depending on the hydraulic and sediment conditions in the near-bed region, deposition of F2.A, F2.B, or F2.C strata ensued.

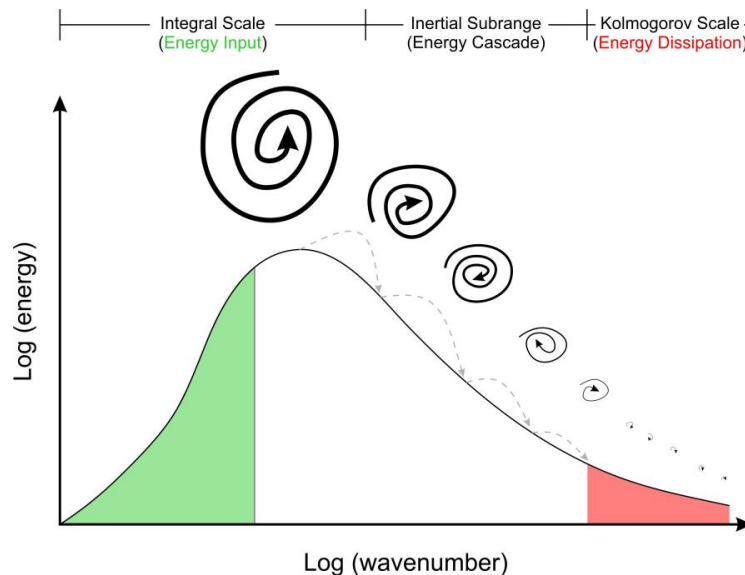


Figure 4.16: Schematic of the energy cascade illustrating the input of energy into the system at the integral scale of turbulence (green area) and the successive nonlinear break up and transfer of energy from large eddies into

progressively smaller eddies (white area), which eventually become so small (red area) that viscosity dominates and converts turbulent kinetic energy (TKE) into heat. As eddies become progressively smaller, they also become more abundant (adapted from Davidson, 2015).

4.3.3 F3: Graded, plane parallel-laminated sandy claystone to massive or graded, structureless claystone

Description

Facies 3 is made up of strata comprising two parts: a plane-parallel laminated basal part overlain sharply by a graded or ungraded claystone (*Fig. 4.17*). The basal part is always planar-laminated and composed internally of rhythmically interstratified sand (sandy claystone) and mud (claystone) laminae. Sand and mud laminae thin vertically from 2-5 mm at the base to 1 mm-hairline laminae at the top. Boundaries between laminae become better defined stratigraphically upwards. Each sand lamina fines upwards as do dispersed sand grains in superjacent mud couplets (*Fig. 4.18*). The coarsest (fine-grained sand) basal sand layers of F3 have dispersed, up to medium-grained sand grains, which then fine to silt vertically. The abundance of dispersed grains is directly proportional to the thickness of the basal laminated part. As the recrystallized clay content increases upwards, the abundance and size of chlorite porphyroblasts in the sand and mud lamina increase (up to coarse sand size).

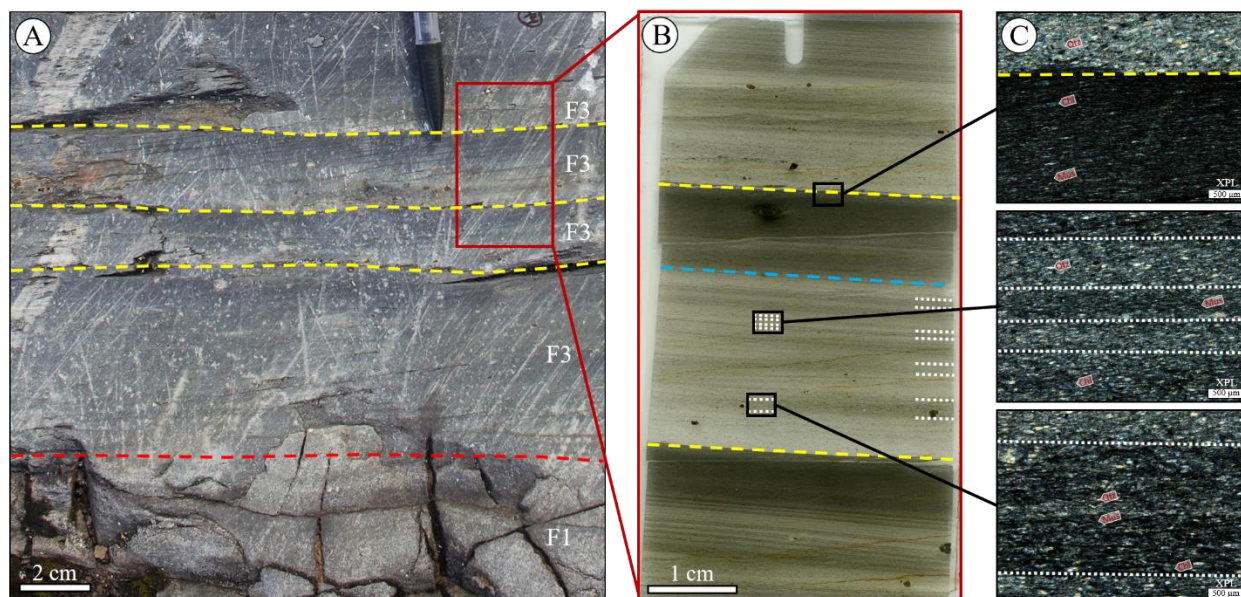


Figure 4.17: A) Sharply based (yellow dashed lines) strata of F3. Although (parallel) lamination is not apparent in the lower, light-coloured part of the bed in the field, the gradual upward colour change from light grey to dark grey at the top (structureless upper part) can be easily discerned and used to differentiate individual F3 beds. B) In thin section strata of F3 exhibit a distinctive alternation of planar layers of sandy claystone (lighter layers) overlain sharply by claystone layers (lighter layers; separated by white dotted lines) that collectively thin and fine upward, and ultimately is capped sharply by a structureless claystone (above blue dashed line). C) Photomicrographs of thinly laminated (2 mm thick; bottom) sandy claystone with claystone interlaminae (separated by white dotted lines) that thin to very thin laminae (0.5 mm thick; middle) before grading into structureless claystone at the top of the bed. Common minerals like quartz (Qtz), muscovite (Mus), and chlorite (Chl) have been indicated.

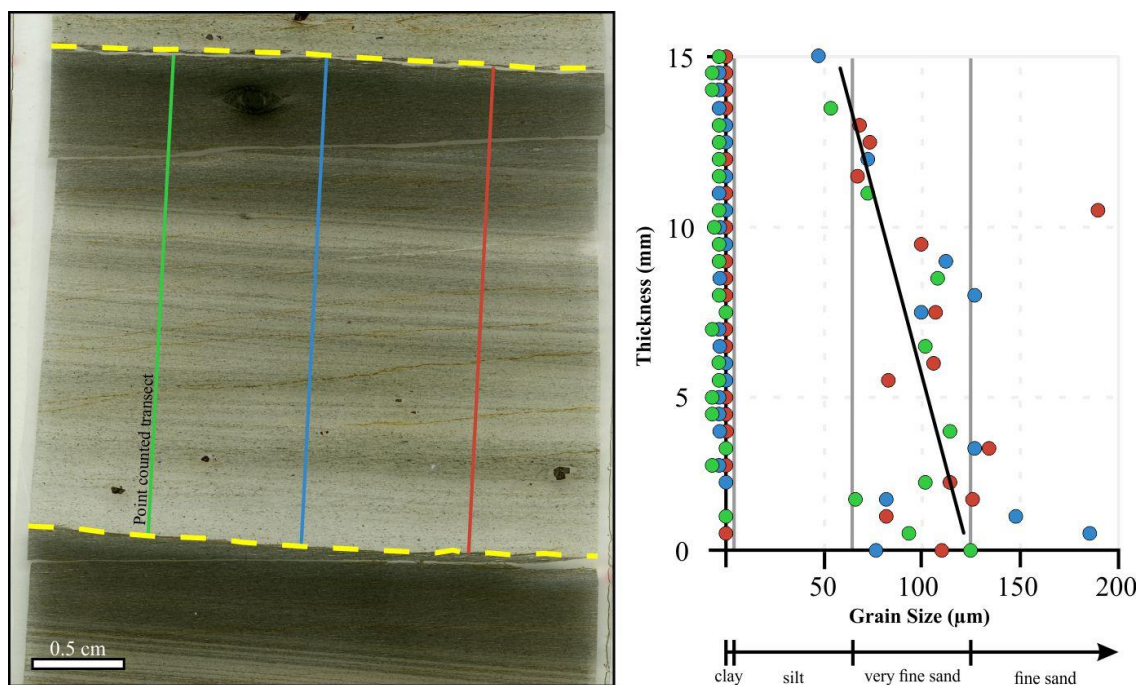


Figure 4.18: Grain size analysis through a very thinly-bedded F3 bed (bounded between yellow dashed lines) based on point counting of 30 points identified every 0.5 mm along three separate vertical transects (*i.e.* 90 points total). Note that data points lying on the vertical axis represent recrystallized clay (*i.e.* chlorite, muscovite, and sericite).

Although dominated by matrix (points on vertical axis), the coarser grained fraction shows a clear upward fining (highlighted by the black linear trendline). Significantly, the range of grain sizes in adjacent sandy and muddy layers (grey layers) is similar, and stratigraphically upward show a similar upward fining.

Facies 3 is deposited almost everywhere across the study area (99% of beds), and very commonly (68%) forms the entire bed where the basal contact is sharp and planar. F3 strata that form the whole bed commonly consist of packages up to 17 beds thick, but packages are more commonly 2-5 beds thick. Alternatively, where F1 or F2 are present, it overlies a F2 layer gradationally (27%) or rarely (5%) a F1 layer sharply. Where F1 underlies F3, F3 tends to be coarser-grained compared to where the bed transitions vertically from F1 to F2 to F3. Facies thickness ranges from 0.5-15 cm (average 3.2 cm, mode 2 cm) thick and makes up 3-100% (average 81%) of bed thickness. F3 preferentially thickens to the NW, whereas grain size coarsens to the SE. Strata are always graded and matrix-rich (average 70% recrystallized clay and silt sized quartz based on point counting), very fine sand with rare (< 1%) carbonate cement.

Interpretation

F3 is interpreted to be similar to a T_{de} turbidite or parts of a more complete turbidite with deposition from a dilute, fine-grained suspension. The gradual upward fining from mostly very fine-grained sand to silt suggests a progressive reduction in flow competence during a single sedimentation event. Pervasive planar-lamination composed of rhythmically interlaminated sand (sandy claystone) and mud (claystone) layers may relate to the action of internal waves within the channelized turbidity current (Khan and Arnott, 2010). Alternatively, Stow and Bowen (1980) proposed that the millimetric interlamination of comparatively coarse- and fine-grained layers was the result of alternating rheological changes in the viscous sublayer of a low-energy turbidity flow, specifically episodes of individual coarse particle settling followed by near-bed gelling and deposition of a finer-grained clay layer. However, sand and mud lamina in F3 layers are more

poorly sorted, and the coarse sand layer has a wider range of grain sizes than described in the depositional sorting model proposed by Stow and Bowen (1980). Sand laminae commonly have > 60% matrix and dispersed grains up to medium sand, whereas mud laminae contain up to 15% dispersed silt to fine-grained sand. Therefore, the process of segregating silt and clay into separate, alternating lamina is likely more inefficient than suggested by Stow and Bowen (1980) as sand, silt, and recrystallized clay are deposited in different proportions in both the sand and mud layers in F3. The poorly sorted nature of the sand and mud lamina suggests that shear sorting in the boundary layer is ineffective at fully sorting the coarse sediment from the fines. Dispersed grains in the mud lamina could also be a result of clay flocculating around individual silt to fine sand-sized grains before deposition. The massive or graded claystone cap is deposited via suspension fallout in the final stages of the flow and later hemipelagic fallout.

4.3.4 F4: Massive, interstratified, structureless clayey sandstone and claystone Description

Facies 4 is restricted to the northwest corner of the study area. Analysis was restricted to reconnaissance level observation and only a small number of samples were collected and analyzed.

F4 beds have sharp, planar basal contacts and consist of > 1 cm massive fine to very fine-grained sandstone (30-40% matrix) overlain sharply by massive mudstone (with up to 10-15% dispersed silt) (*Fig. 4.19*). Rare anomalously large (medium to coarse sand size) chloritoid grains are observed in both sandstone and mudstone layers. Additionally, abundant bedding-parallel, thin (< 150 μm), dark, wispy, and elongate seams (microstylolites) with elevated amounts of clay, iron oxides and/or sulphides (*e.g.* pyrite) are observed. Interbeds of thin-bedded, fine-grained, carbonate cemented $T_{\text{cde}}/T_{\text{ce}}$ sandstone turbidites are common and generally consist of isolated, single set ripple cross-stratified formsets.

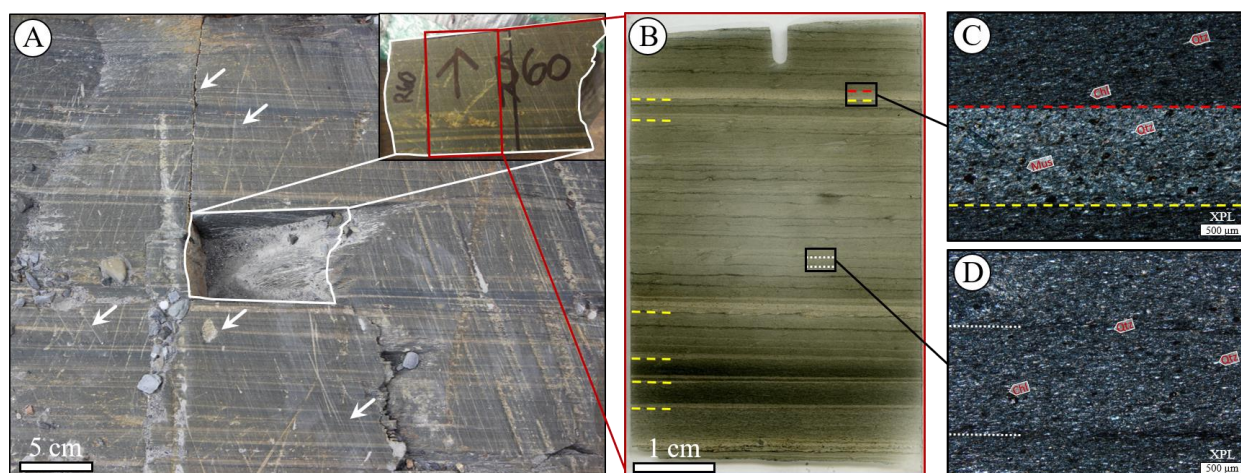


Figure 4.19: Facies 4 from macroscopic (field) to microscopic (thin section) scale. A) Outcrop photo of thin- to very thinly-bedded, fine-grained F4 beds. Bedding contacts are difficult to discern due to the fine-grained nature of this facies. The base of some units is marked by up to fine, but generally very fine sand laminae that often pinch and swell along strike (possible starved ripples), indicated by white arrows, whereas others are planar-laminated or planar but structureless. B) Multiple thinly-bedded F4 strata (beds bounded between yellow dashed lines) in thin section. Even though the basal coarser-grained part appears structureless or planar-laminated here, as shown in (A), these layers often pinch out over short lateral distances (1-2 m). The contact between the coarser-grained basal layer and fine-grained cap is almost always sharp and planar. Note also the abundant thin, closely spaced (< 1 cm), bedding-parallel, dark, wispy solution seams (microstylolites) throughout the fine-grained layers (indicated by white dotted lines). C) Closeup of layering in F4 strata, specifically a structureless sand-rich layer (bounded between the red and yellow dashed line) overlain sharply by structureless sandy claystone. Note the dramatic decrease in grain size between the basal and upper parts (indicated by a red dashed line). Significantly, both parts have similar ranges in grain size, however, sand grains in the finer-grained upper part are less abundant and are floating in a matrix of mud. D) Photomicrograph of fine-grained strata showing solution seams (highlighted by white dotted lines) that are continuous for a few to tens of metres along strike in outcrop. Some common minerals like quartz (Qtz), muscovite (Mus), and chlorite (Chl) are indicated.

Interpretation

The ungraded and structureless character of both sand and mud-rich layers suggests deposition from suspension fallout. These conditions were interrupted episodically by the emplacement of single set (starved) ripple formsets indicating traction transport and deposition from low-concentration turbidity currents.

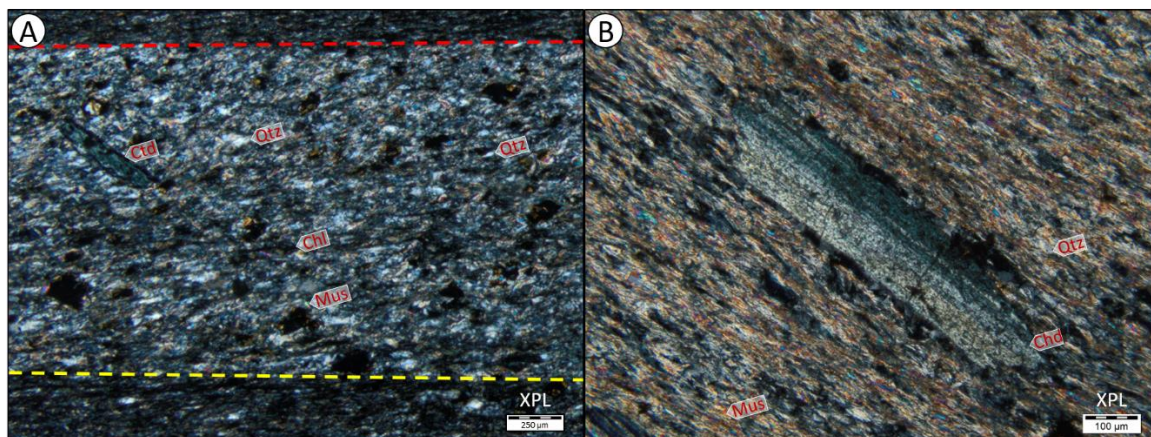


Figure 4.20: Large, elongated chloritoid (Chd; compound mineral percentages from EDS: Mg – 2.33%, Si – 25.46%, Fe – 30.48%, Al – 41.73%) porphyroblast in both the basal coarse-grained (A) and upper fine-grained layer (B) in F4 strata. A) Large (c. 400 µm) chloritoid grain in a coarse-grained F4 layer (bounded by yellow and red dashed lines). B) Photomicrograph of a large chloritoid porphyroblast in the fine-grained part encased in recrystallized clay (primarily micas here) with dispersed silt sized quartz and very fine-grained sand. Note the polysynthetic twinning, but more importantly the rounded and corroded texture of the grain boundaries, suggesting the grain is detrital rather than diagenetic or metamorphic. Note also other minerals identified in photomicrographs are quartz (Qtz), muscovite (Mus), and chlorite (Chl).

4.4 Architectural Model

Strata in the study area consist of an amalgamated sandstone and conglomerate unit overlain abruptly by a heterolithic unit capped sharply by a thick succession of fine-grained, thin-bedded classical turbidites (*Fig. 4.21*). The heterolithic unit is the primary focus here. The basal amalgamated unit is 15 m thick and composed of amalgamated coarse-grained sandstone to pebble conglomerate interpreted to be part of a slope channel complex, informally termed Isaac Channel Complex 0 (ICC0). At the same stratigraphic level, but 250 m to the northwest, strata consist of a thick (> 40 m) succession of rhythmically interstratified, thin- to very thin-bedded, very fine-grained sandstone overlain sharply by mudstone – all strata distinctively lack traction transport structures (F4). Although the contact between the two units is covered, the dramatic lithological change over only a few hundred metres suggests the coarse, amalgamated strata of ICC0 represents an erosionally bounded channel that partly infills a deep scour incised into fine-grained F4 strata. Heterolithic strata that abruptly overlies ICC0 are interpreted to represent levee deposits associated

with a younger (unexposed) channel whose axis is now located some distance to the SE (*Fig. 4.21*). These strata, like levee deposits everywhere, were deposited from flows that overspilled the margins of the (new) channel. However, in this case, the outward movement of the flows, at least on one side of the channel, was restricted by the erosional levee sculpted previously by ICC0 into the fine-grained (F4) unit. This two-way flow confinement, therefore, resulted in a set of unique and localized sedimentological conditions that became manifest as heterolithic strata rather than more typical classical turbidites. More specifically, flows that overspilled the new channel to the SE and flowed down the developing levee eventually encountered the earlier part of the same flow that had reversed off the erosional levee (escarpment), termed here an escarpment, causing an upstream-moving bore. As flows continued to overspill the channel, the incident flow (from the channel) and return flow (from the escarpment) began to continuously interact and resulted in the trapping and deposition of the mud-rich parts of heterolithic strata, specifically, Facies 2 and 3 within each event bed. With continued deposition, the relief of the escarpment became reduced, and as a consequence, the return flow diminished. Additionally, this second channel was abruptly succeeded by a third channel located even further to the southeast, and therefore further from the study area. Collectively, the diminished relief of the escarpment and the more distal location on the levee resulted in deposition of thin-bedded turbidites that more closely resemble turbidites deposited on modern and ancient deep-marine levees (*e.g.* Khan and Arnott, 2011).

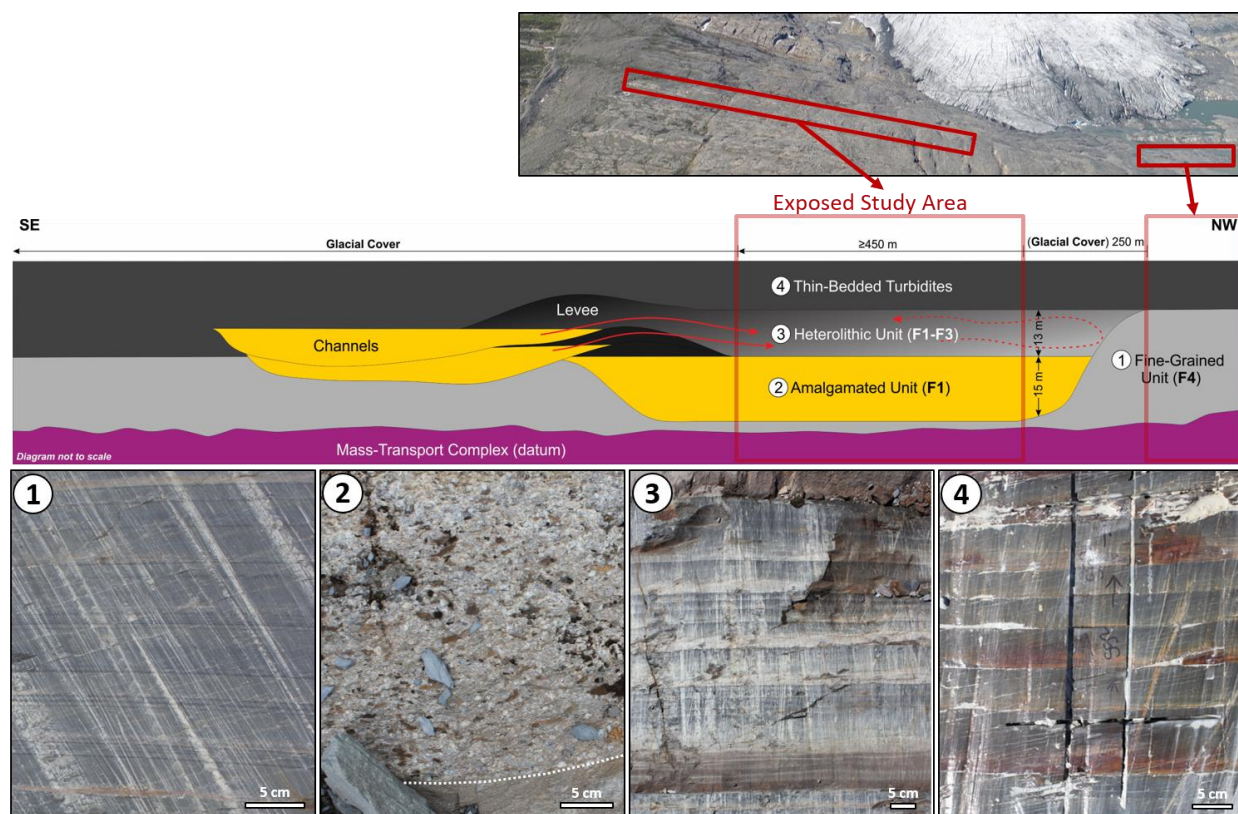


Figure 4.21: Schematic model of the study area showing the spatial and temporal relationship of the four major units that overlie an MTC used here as the stratigraphic datum – representative photos of each unit are provided at the bottom (note that in the photos the subvertical/vertical white lines are glacial striations). The channel complex is interpreted to consist of multiple channels (yellow) that progressively aggraded and migrated to the SE with levees (black to grey) that built up as flows overspilled the channel margins. Flow exiting the channels along their northwest-facing levees interacted with the same, but earlier part of the flow that had reversed off the erosional levee (*i.e.* escarpment; light grey) further to the northwest causing an upstream-moving bore. This resulted in a complex flow field in this semi-confined basin that in the Windermere sedimentary record is manifest by the rhythmic stacking of fining and thinning successions composed of beds exhibiting a distinctive vertical succession of facies: F1 (matrix poor) overlain by F2 (36% matrix on average) and capped by F3 (70% matrix). Note that strata observed in the study area is outlined in red boxes whereas the rest of the study area is covered in glacial deposits and are inferred.

4.5 Stratigraphic Evidence of Flow Confinement

Flow confinement in a deep-marine system occurs where turbidity currents encounter intraslope or basin floor topography that modifies the flow path. Typically, the term flow confinement, or more commonly simply *ponding*, is used to imply full containment of the flow within a bathymetric low and as a consequence entraps sediment, particularly the fine-grained fraction, which more typically is transported further basinward (van Andel and Komar, 1969). However, the more general term *confinement* will be used here to describe turbidity currents

affected significantly by seafloor topography but not fully contained (*sensu* Lomas and Joseph, 2004). Such partially confined flows are commonly reported in tectonically active deep-marine settings, such as the Castagnola Basin, Italy (Marini *et al.*, 2016), above remobilized salt formations (*e.g.* Winker, 1996; Wang *et al.*, 2017), and less commonly on passive-margin continental slopes with local intraslope relief (this study). Previous authors have reported stratigraphic evidence suggestive of partial or complete flow confinement in both ancient and modern systems, including bed tabularity, paleoflow reversals, abundance of mud, onlap relationships, and distinctive facies changes (*e.g.* Ricci Lucchi and Valmori, 1980; Pickering and Hiscott, 1985; Drinkwater and Pickering, 2001; McCaffrey and Kneller, 2001; Sinclair and Tomasso, 2002; Spychala *et al.*, 2015; Marini *et al.*, 2016; Liu *et al.*, 2018). Although onlap relationships were not observed in this study, most probably due to exposure conditions (covered), the following compares these other attributes, in addition to a suite of additional features, to features observed in this high-resolution study.

4.5.1 Tabularity

One of the most commonly reported depositional manifestations of flow confinement is bed tabularity as unconfined systems tend to be more lobate and stack compensationally (Terlaky *et al.*, 2016). Here we assess bed tabularity using an equation adapted from Liu *et al.* (2018):

$$T_{Bed} = \sum_{i=1}^n \left(\frac{B_i - B_{i+1}}{D_i n} \right)$$

where T_{Bed} represents the along-strike thinning/thickening rate (cm/km) of individual beds, B_i is the total bed thickness (cm) at a specific location, D_i is the along-strike distance (km) between adjacent logs, and n is the total number of measurements in a single along-strike transect. Here we use data from beds consisting of F1 to F3, where positive or negative T_{Bed} indicates, respectively,

along-strike thinning or thickening to the NW (*i.e.* away from the channel). Beds that extended across the entire study area (> 300 m) and showed no evidence of bed amalgamation or scour surfaces were selected and analyzed at four along-strike logging locations. In this analysis 50 beds (roughly a quarter of all beds in a single vertical stratigraphic log) were included in the analysis. Note that this analysis only captures general changes in bed thickness, thus neglecting subtle bed thickness variation between logs. Beds were then grouped into three categories based on bed thickness using Ingram's (1954) classification for bed thickness: 1) very thin to thin beds (< 10 cm thick), 2) medium beds (10-30 cm thick), and 3) thick to very thick beds (> 30 cm thick). This was done to reduce the bias between the absolute thinning/thickening rates between thin, medium, and very thick beds as thicker beds will intrinsically have a higher rate of thinning/thickening compared to thinner beds with the same relative rate (*e.g.* average thinning/thickening rate) (Tókéš and Patacci, 2018). In general, beds thicken towards the NW, with higher rates of thickening related to thicker beds, specifically: thick to very thick beds – 0.7 m/km, medium beds – 0.1 m/km, and very thin to thin beds – 0.008 m/km. The higher rate of thickening of thicker beds, although not surprising, still only represents a 0.2% change in bed thickness across the width of the study area. Therefore, irrespective of thickness, all beds exhibit negligible change in bed thickness for kilometres along strike (See *Fig. 4.25*), and therein are consistent with Pickering and Hiscott's (2015) criteria for tabular beds, as well as being classified as a highly confined system according to Liu and colleagues (2018) analysis. Additionally, in beds that do thicken toward the NW, most of the thickening occurs in F1 strata at the base of the bed, which also are generally the thickest and coarsest grained facies making up a bed.

Based on general lithological attributes and stratigraphic position and association, strata of the heterolithic unit are interpreted to be levee deposits, however their highly tabular nature

contrasts the geometry more commonly reported from unconfined levee deposits in the Castle Creek study area. Specifically, levees initially thicken (up to 10 m) in the first 200-600 m away from the channel margin and then thin by about the same amount over the next 200-300 m (*Fig. 4.22*) (Khan and Arnott, 2011; Khan, 2012; Bergen, 2017). Beds in this study, however, show negligible along-strike change in thickness – 10s cm per kilometre compared to metres per kilometre.

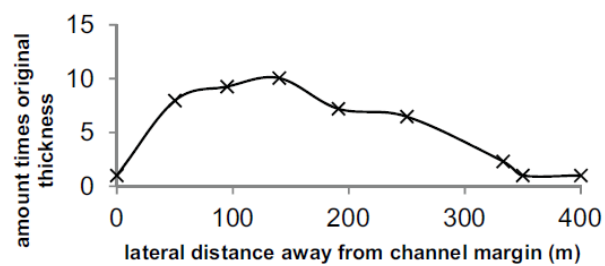


Figure 4.22: Thickening and then thinning trend of levee deposits away from the channel margin. Note that the vertical axis represents the average change in thickness of 25 thick- and medium-bedded sandstones from their original thickness adjacent to the channel margin (Khan, 2012).

4.5.2 Lithological Characteristics: Vertical Trends

Sedimentation Events

In the deep-marine sedimentary record it is generally assumed that a single bed indicates a single sedimentation event (*i.e.* deposit from a single turbidity current). Here, however, multiple beds, although separated by distinct basal contacts, form a consistent and well-defined overall upward fining and thinning trend suggesting that they form part of a single sedimentation event (*Fig. 4.23*). Each sedimentation event begins with deposition of F1 or F2 strata, or uncommonly F1 overlain by F2, which then is overlain by rhythmically stacked F3 strata (self-similar stacking) that generally fine and thin upwards and are made up of 1-17 F3 beds thick, but more commonly are 2-5 beds thick. Additionally, within the 11.21-13.35 m-thick heterolithic unit, the number of

F3 beds in each sedimentation event increases upwards whereas the number of F1 divisions decreases (see *Fig. 4.26*).

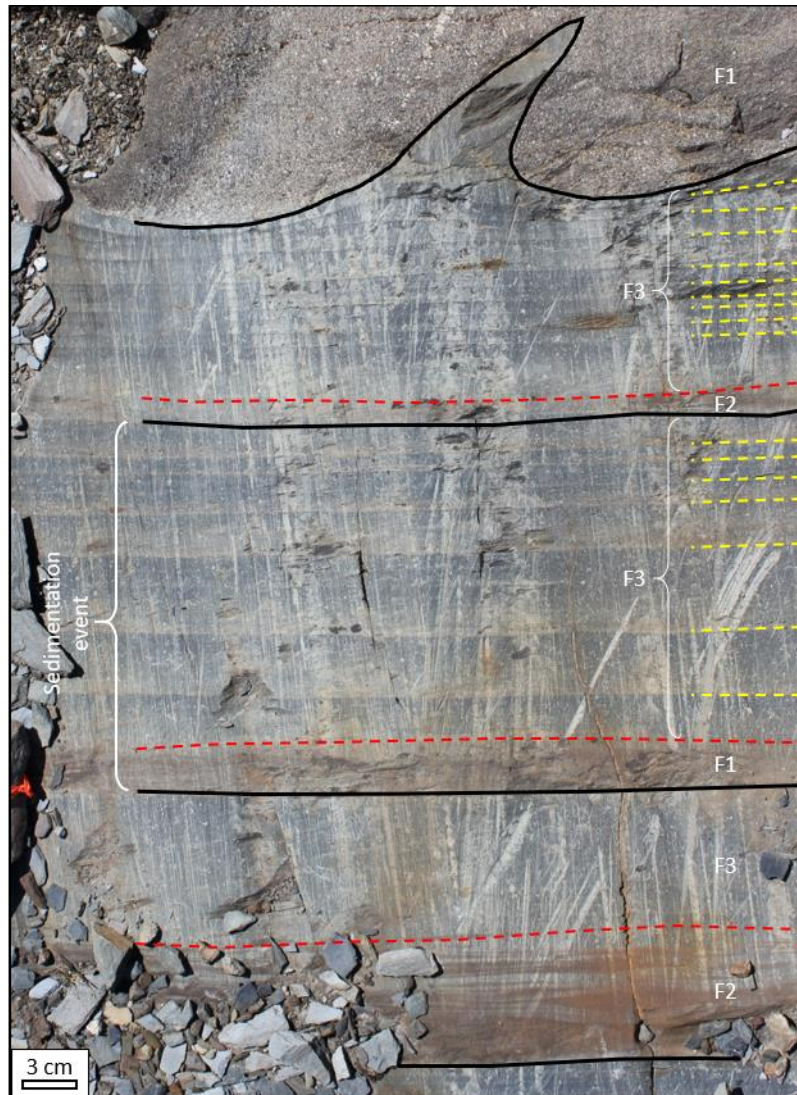


Figure 4.23: A stack of four sedimentation events; the base of each indicated by a black line). Within each sedimentation event the dashed red lines separate the component facies and dashed yellow lines indicate bed contacts in a stack of similar facies. Beds within a sedimentation event generally fine and thin upwards, as well as the F1 or F2 divisions at the base of each sedimentation event in a bedset. Note the well-developed flame structure at the base of the uppermost sedimentation event.

Many geologists have reported a basal sandstone overlain by an anomalously thick mud cap related to a sand-mud suspension trapped within a mini-basin (Pickering and Hiscott, 1985; Haughton, 1994), whereas the fine, upper part of these and larger flows would mostly overtop the

confining topography (*e.g.* Marini *et al.*, 2015). However, in many of these studies the vertical and horizontal resolution of fine-grained deposits is generally poor; in the ancient because of their recessive weathering character and common vegetative cover, and in the modern because of widely spaced core control. As a result many researchers have grouped all fine-grained strata in a single category and not attempted to differentiate thin mud turbidites from the mud caps of thicker beds (Liu *et al.*, 2018), or have disregarded the mud caps all together and instead focused on the sandstone part of beds (Tőkés and Patacci, 2018). Here, vertically-dipping, glacially polished fine-grained strata with only local surface cover provide an unparalleled opportunity to study the millimetre-scale detail of fine-grained F3 beds. Additionally, Liu and colleagues (2018) reported that of the four systems they analyzed, their most confined system had the highest mud content, ranging from 34-76%. In this study, mud content in the heterolithic part of each sedimentation event ranges from 0-91% (average 51%) indicating efficient entrapment of fine-grained sediment in a highly confined system.

The abrupt superposition of F2 above F1, the very occurrence of F2 strata, and the rhythmically stacked F3 beds at the top of most sedimentation events is interpreted to be a result of the interaction of a more dense flow, in this case representing the incident flow spilling out of an adjacent channel, with a bore flowing away from a confining slope (*Fig. 4.24*). More specifically, the bore represents an inflated sediment suspension composed of sediment sourced from the incoming incident flow as it rises along the confining slope converting its kinetic to potential energy, which then is converted to kinetic as the suspension collapses and forms a bore moving down the confining slope and into the incident flow (van Andel and Komar, 1969; Kneller and McCaffrey, 1999; Toniolo *et al.*, 2006; Patacci *et al.*, 2015; Tőkés and Patacci, 2018).

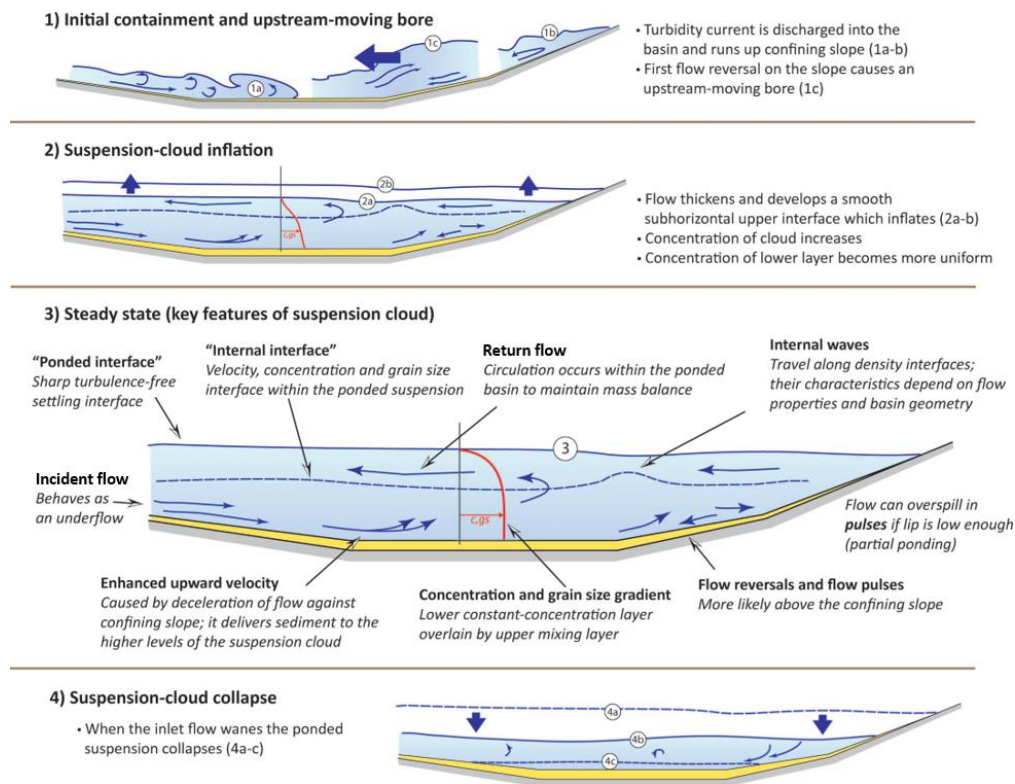


Figure 4.24: Evolution of a confined flow and its interaction with sea floor topography (modified from Patacci *et al.*, 2015).

Bedsets

Bedsets progressively fine and thin upwards and range from 2.01-4.12 m thick and comprise multiple sedimentation events (10-24 made up of 41-59 beds). Also, the number of sedimentation events per bedset decreases upwards. The base of each bedset is marked by an anomalously thick (up to 1.12 m; see *Fig. 3.5*), coarse-grained (up to very coarse sand) F1 sandstone (*Fig. 4.25A, 4.26*). Across the study area bedset thickness varies by up to two metres, which in large part depends on changes in the thickness of the lowermost F1 sandstone. This, then, is overlain by sedimentation events with thinner F1 and/or F2 divisions overlain by rhythmically stacked F3 beds until the deposition of the next thick, coarser-grained F1 sandstone at the base of the next bedset. On a larger scale, successive bedsets exhibit a general upward fining and thinning where the maximum grain size and range of grain sizes decreases stratigraphically upwards.

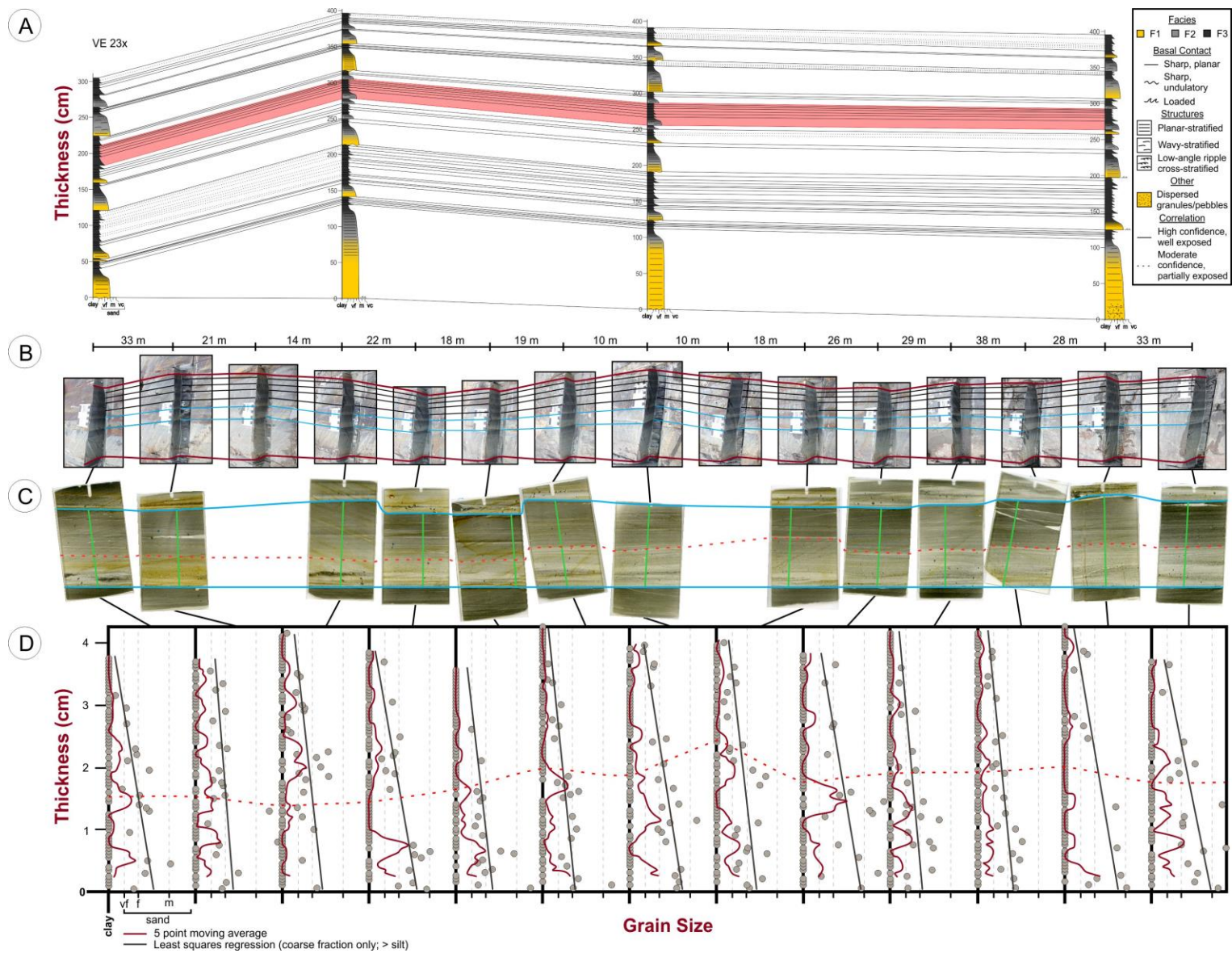


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Figure 4.25: Detailed lateral correlation of bedset 4 from macroscopic to microscopic scale. A) Correlation panel of bedset 4 showing tabular, laterally continuous beds; note that vertical exaggeration is 23x. The red band represents the beds included in samples cut for detailed analysis and shown in (B). B) 15 samples cut about every 20 m through a succession of seven upward fining and thinning beds (red band in (A)). The bed bounded by blue lines was sampled for thin section analysis shown in (C). Scale card is 8.5 cm long. C) In 13 oversized (38 x 76 mm) thin sections grain size was microscopically point counted every 0.5 mm along a single vertical transect (green lines); data plotted in (D)). The red dotted line separates coarser-grained, planar-laminated F3 basal strata from more mud-rich, more subtly laminated or structureless strata. The upward fining and increase in mud-content is also manifest as an upward colour change from light grey to dark grey/black. Due to the fine-grained nature of these strata two samples were not suitable for thin section preparation and therefore are missing from the microscopic analysis. D) Point counted grain size data (green line in (C)). Note that data points on the vertical axis represent recrystallized clay (*i.e.* chlorite, muscovite, and sericite). Based on a five-point moving average (red line), and a least squares regression (black line) that only accounts the coarser grained fraction ($> 62.5 \mu\text{m/silt}$), grain size consistently fines upward in all F3 strata here

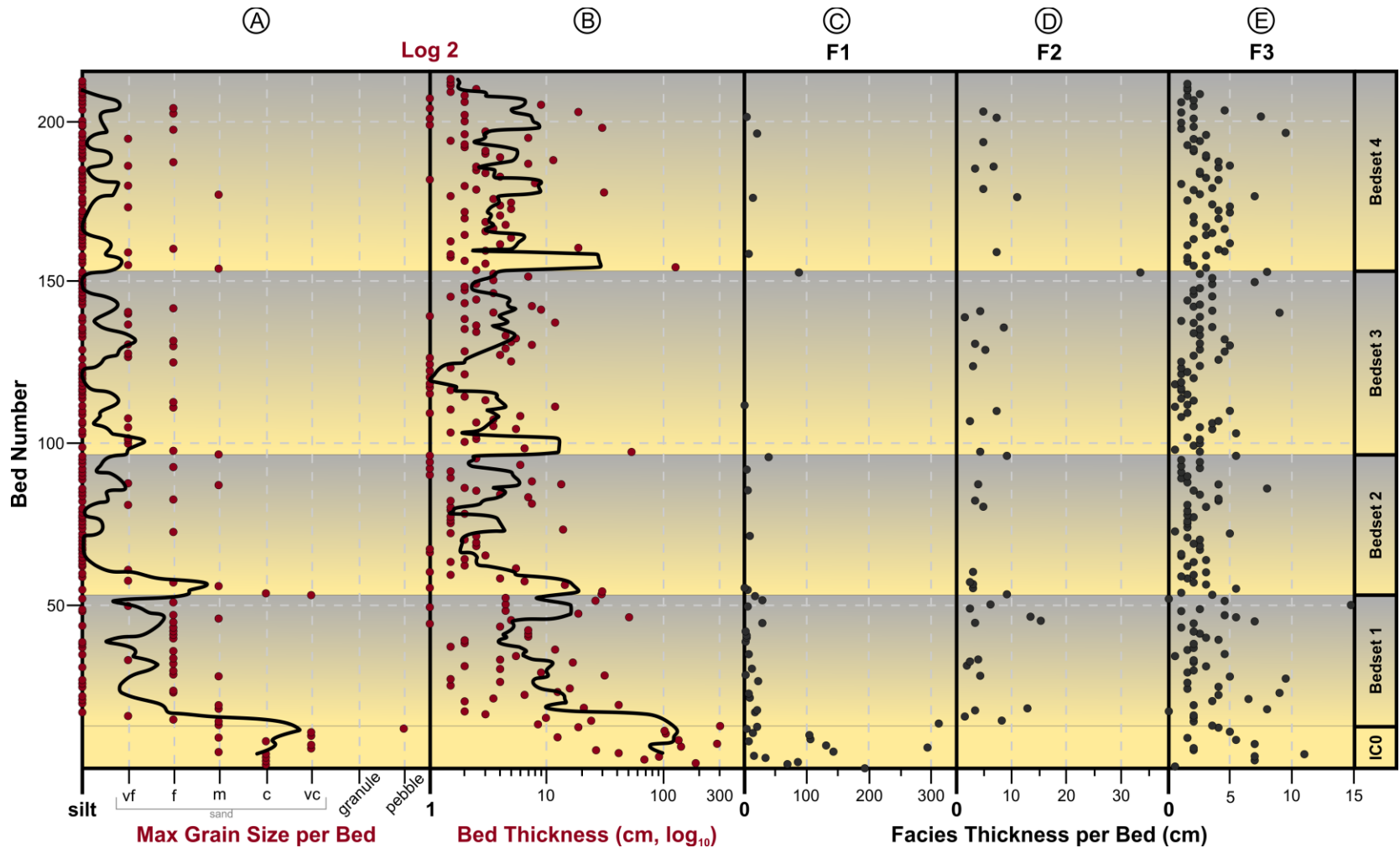


Figure 4.26: Textural and dimensional analysis of four stacked heterolithic bedsets in log 2. A) Maximum grain size; black line represents the five-point moving average. Note the marked increase in grain size at the base of each bedset followed by an abrupt fining. Note also the overall upward fining across the four bedsets, which is somewhat more muted within individual bedsets. B) Bed thickness; note that the horizontal scale is logarithmic in order to capture subtle changes in beds thinner than 10 cm (as they make up most of the heterolithic unit). Black line is the five-point moving average and shows that beds are thickest at the base of each bedset, decreases abruptly and then varies irregularly thereafter. C-E) Thickness of the component facies (F1-F3) that make up each bedset. In (C) note that coarser-grained F1 strata become slightly finer but less abundant upwards, whereas F2 (D) and F3 (E) show little to no upward trend.

Markov Chain Analysis

In order to identify any statistically significant pattern in stratal stacking a first-order Markov chain analysis was performed (summarized in *Fig. 4.27*). A Markov chain is a mathematical model describing sequential changes in a system composed of discrete, finite states (Schwarzacher, 1975; Everitt and Skrondal, 2010). The analysis was performed on log 2 where a continuous 12.5 m-thick exposure was made with a rock saw (see *Fig. 4.5*). The polished, unweathered surface significantly facilitated the identification of component facies and bed contacts.

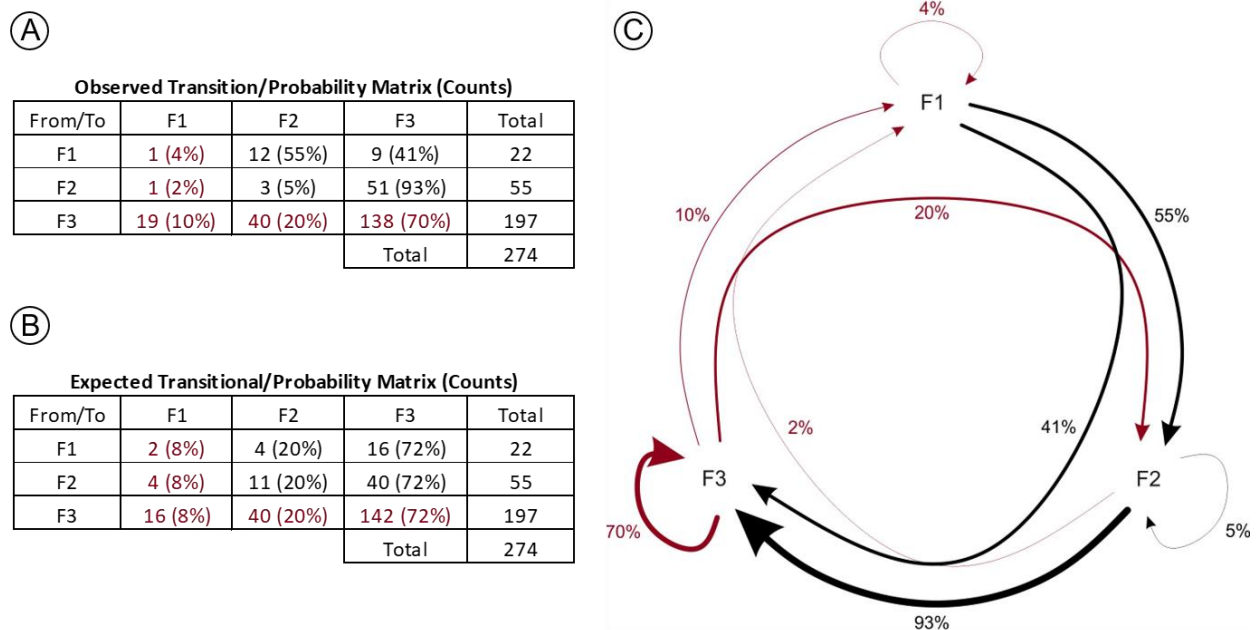


Figure 4.27: Markov chain analysis consisting of 274 facies transitions between facies in log 2 with statistically significant results. Note that the black text/arrows indicate facies transitions within a bed whereas red text/arrows indicate facies transitions between beds. A) Observed transitional matrix (probabilities in brackets) consisting of F1 (massive or graded, structureless or planar/wavy-stratified clayey sandstone), F2 (graded, plane parallel-, cross- or wavy-stratified clayey sandstone with intercalated structureless sandy claystone), and F3 (graded, plane parallel-laminated sandy claystone to massive or graded, structureless claystone) states. B) Expected transitional matrix based on a random distribution. C) Summary transitional state diagram; higher transition probabilities are indicated by thicker arrows.

The statistically significant results show that where present, F1, which occurs at the base of the bed, is overlain equally by either F2 or F3 and only rarely forms the top of a bed. F2 is most

commonly overlain by F3 strata (93%), or another style of F2 stratification (*i.e.* subfacies of F2 stack 5% of the time), or uncommonly (2%) forms the top of the bed. F3 almost always (99%) forms the top of the bed and commonly stacks with other F3 beds (70%) or is overlain by F1 (10%) or F2 (20%) at the base of the next bed.

Paleocurrent Reversal

In some F2.B beds composed of two or more styles of cross-stratification, the base of each is marked by a thick mud layer, and the paleocurrent direction is commonly observed to reverse above the mud layer (*Fig. 4.12B, 4.14A*). Since the reversal(s) occurs in the deposit of a single sedimentation event it suggests that during a single turbidity current local paleoflow was at times directed away (northwest) or toward (southeast) the channel. Flow directed away from the channel represents typical channel overspill, which here is termed the incident flow. Flows directed toward the channel, on the other hand, are related to the interaction of the incident flow with a lateral barrier, in this case the erosional escarpment carved into F4 strata by ICC0 (see above). Specifically, the superelevation of the incident flow along the escarpment, and hence buildup of potential energy is eventually relaxed by a flow directed away from the escarpment, here termed a return flow or bore, as potential energy is converted to kinetic energy. The thick mud layer at the base of the paleocurrent reversal marks the initial interaction of the return flow and the continually flowing incident flow, specifically an internal hydraulic jump and an episode of intense sediment fallout. Rapidly the return flow comes to dominate the flow field and causes sediment transport to be directed toward the channel. Eventually the return flow weakens and the incident flow comes to re-establish itself as the dominant transport direction.

4.5.3 Lithological Characteristics: Lateral Trends in Grain Size within Facies

50 laterally continuous beds were analyzed to assess lateral changes in grain size within each of the stratal divisions in a bed. F1 preferentially coarsens to the NW, most commonly from fine- to medium-grained sand; F2 fines to the NW typically from medium to fine sand or fine to very fine sand; F3 changes little. F1 strata are interpreted to have been deposited by the incident flow as it overspilled the channel margin and before it interacted with the escarpment to the NW. The slight coarsening and thickening of strata are similar to trends reported previously in levee deposits of the Isaac Formation by Khan (2012) and Bergen (2016) and interpreted to be a consequence of flow inertia and the lateral displacement of maximum rate of sedimentation outboard of the channel margin, which in this study may have been accentuated by an adverse slope very near the escarpment. F2, on the other hand, coarsens to the SE, which probably reflects the now well-developed return flow and its interference on the underlying and oppositely orientated incident flow that promotes preferential deposition of coarser grained sediment closer to the channel. F3, like similarly fine-grained strata deposited in small basins (*e.g.* Tókéš and Patacci, 2018), tends to be spatially more equigranular. Finally, the general upward fining and thinning exhibited by virtually all sedimentation event beds reflects the long-term waning of the channel-bound turbidity current, and accordingly, a reduction in the grain size and sediment concentration of the overspilling flow.

4.6 Comparison with Classical (Bouma) Turbidites

It is clear that strata consisting of alternating sand- and mud-rich layers, which in this study make up F2 and F3, were deposited by turbidity currents that differ from those that deposit classical turbidites (Bouma, 1962). An idealized classical (Bouma) turbidite comprises a thickly developed (commonly 10s cm) lower graded or massive (T_a) sandstone overlain by planar stratified (T_b)

sandstone overlain by a thinner (few centimetres), generally finer-grained ripple cross-stratified sandstone and then discontinuously planar-stratified sandstone/siltstone grading upward to mudstone (c-, d- and e-divisions, respectively) (for a detailed granulometric analysis of a classical turbidite see Komar (1985)). This succession, at least superficially, would appear to resemble an idealized heterolithic bed made up of F1 to F2 to F3. F1 consists of structureless overlain by planar-stratified sandstone, which apparently would equate to the T_a and T_b parts of a classical turbidite, and here is interpreted to indicate similar formative processes. F1 is then sharply overlain by F2 overlain by F3 which resemble the b-, c-, and d-/e-divisions, respectively (*Fig. 4.28*). In terms of thickness and grain size (fine to very fine sand) F2 strata resemble the cross-stratified T_c division in a classical turbidite (see Komar, 1985), however, a number of important differences are noted. Firstly, and most notably, is the abundance and rhythmic intercalation of mud in F2 – average mud content in F2 is c. 36% whereas b- and c-divisions in classical turbidites are generally well to very well sorted with negligible mud content (Komar, 1985). Secondly, F2 cross-stratification consists of interstratified sand and mud layers that exhibit a consistently low angle dip, commonly c. 10°, which is much lower than reported from T_c divisions described by Komar (1985) or elsewhere in the Castle Creek study area (typically range from 20-30°) (*Fig. 4.29*) (Khan, 2012; Bergen, 2017). Thirdly, wavy-stratification, which occurs in about 55% of all beds containing F2, is notably absent in classical turbidites. Comparative differences in sedimentary textures and component sedimentary structures between F2 and F3 strata and classical turbidites are interpreted to be a consequence of a higher clay concentration in the near-bed region of flows that deposited F2 strata – even ‘sand-rich’ F2 strata contain up to 20-30% recrystallized clay. Clay, even at low concentrations (c. 2-4 volume %), has been shown to impart cohesive properties to laboratory flows, which substantially influences their fluid dynamic properties and damps turbulence (Baas

and Best, 2002; Baas *et al.*, 2015), alters the dynamic behaviour of sediment in suspension (Winterwerp and van Kesteren, 2004; Schieber *et al.*, 2007), and influences the shape and dimension of bedforms (Basaniak and Verhoeven 2008; Baas *et al.*, 2009; Baas *et al.*, 2013; Schindler *et al.*, 2015). In terms of angular bedforms, cohesive particles significantly slow the rate of bedform growth (Baas *et al.*, 2013), and also strengthens the substrate and makes it more difficult to erode (Mitchener and Torfs, 1996). However, it is important to note that changes in other parameters like flow velocity, clay type, concentration and size, initial turbulent structure, and concentration of non-cohesive sediment can also dramatically alter flow conditions and accordingly bedform morphology (Baas and Best, 2002; Baker *et al.*, 2017). Presently, however, the effect of these parameters on bedform development in the near-bed region of mud-rich sediment-gravity flows, and their resultant deposits, remain poorly understood and points to the need for continued experimental- and field-based research. Gradationally overlying F2 strata is the F3 division made up of two sharply bounded parts: a stack of plane-parallel, rhythmically interlaminated sand and mud couplets that superficially resemble the d-division overlain by structureless claystone that resembles the e-division in classical turbidites. Unlike the typically discontinuous, thin, diffusely bounded, silt-rich laminae in the T_d division of most classical turbidites (Stow and Bowen, 1978), the plane-parallel-interlaminae in the lower part of F3 strata are coarser (up to medium sand in lower laminae), thicker (up to 5 mm at base), more sharply defined, and can be traced along strike for at least several metres. Additionally, the structureless claystone at the top of the F3 division is much thicker (some > 5 cm) than the T_e division in most classical turbidites.

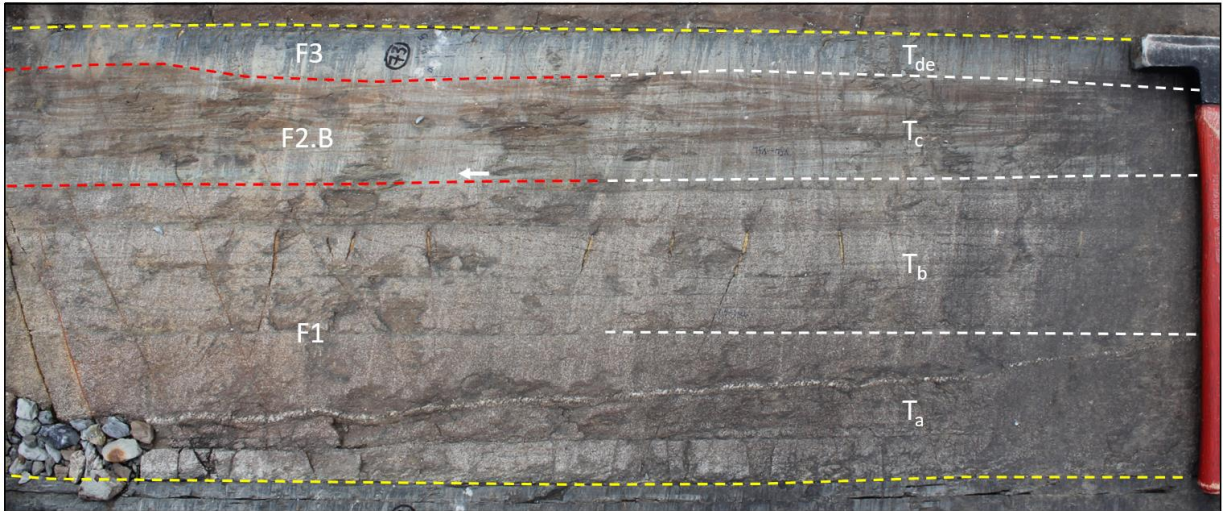


Figure 4.28: Comparison of facies in this study and their apparent equivalents in a classical (Bouma) turbidite. Note that even though this bed exhibits the most complete succession of F1 to F3 in the study area, the planar-stratified (F2.A) and wavy-laminated (F2.C) layers are absent in the F2 division. The bed is 32 cm thick and at its base comprises an F1 unit consisting of structureless overlain by planar-stratified sandstone (separated by white dashed line), which respectively, equate to the a- and b-divisions in a classical turbidite. This, then, is overlain by a thick (1.5 cm), laterally continuous mud layer (indicated by white arrow) overlain by cross-stratified strata (F2.B). Superficially this division appears analogous to the T_c division, but notably overlies a thick mudstone layer, the cross-stratification is consistently low angle, and the strata are composed of rhythmically interstratified sand-rich and mud-rich laminae. This, then, is then overlain gradationally by fine-grained F3 strata, that at least in outcrop (macroscopic), appears to be similar to d- and e-divisions of a classical turbidite, but microscopic analysis indicates important differences (see text for details). Bed bases indicated by the dashed yellow lines; dashed red lines separate facies within the bed.

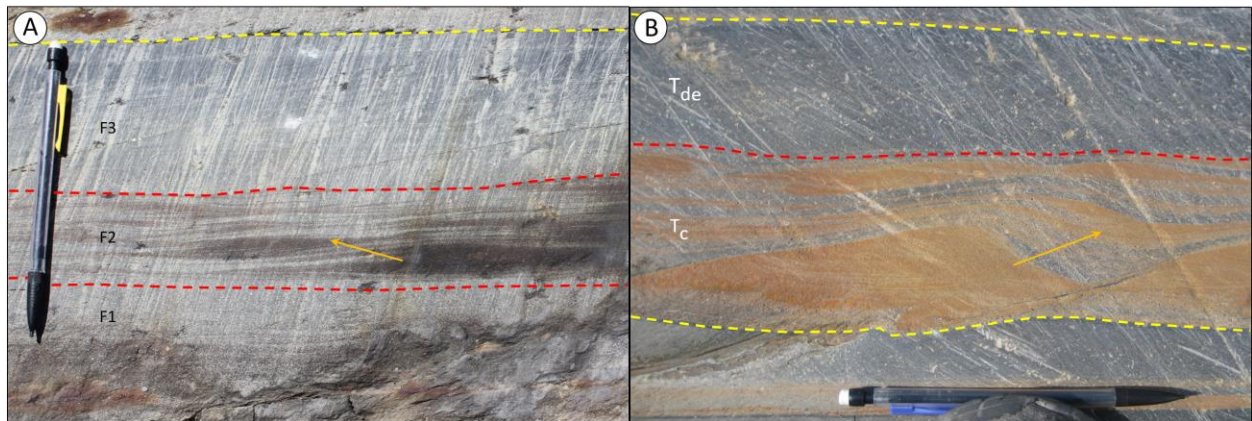


Figure 4.29: Comparison between climbing (indicated by the orange arrows), low-amplitude ($6-8^\circ$ at base to $3-4^\circ$ near top of F2 layer), cross- to wavy-stratified F2 strata (bounded between red dashed lines) in (A) and high-angle ($30-32^\circ$ lee face inclination), cross-stratified climbing ripple (T_c) set in levee deposits (rusty orange strata below red dashed line in part B) (Bergen, 2017), both in the Castle Creek study area. Note the consistently lower angle of cross-stratification and the lateral discontinuity of the sand-rich (dark brown) strata in F2. Bed contacts are indicated by yellow dashed lines, while red dashed lines separate different facies or Bouma divisions.

Chapter 5: Conclusions & Areas for Future Work

In the deep sea the primary sediment-transporting agent is a turbidity current whose depositional product is a classical (Bouma) turbidite. Where fully developed these deposits comprise five sharply bounded, upward-fining divisions where the base consists of massive or normally graded sandstone (T_a) overlain by planar-stratified sandstone (T_b), high-angle, ripple cross-stratified sandstone (T_c), subtly parallel-laminated siltstone (T_d) and capped by massive mudstone (T_e). Classical turbidites have been reported extensively around the globe in both the modern and ancient deep-marine sedimentary records, commonly deposited from unconfined turbidity currents. However, only recently has seafloor topography and its effect on the movement and patterns of deposition from turbidity currents become a topic of interest. Seafloor topography affects turbidity currents to varying degrees depending on the amount of flow confinement, where ponded turbidite systems are the most affected due to complete entrapment of the flow. In general terms such confinement becomes manifest in the sedimentary record by tabular stratal geometries and, as discovered in this work, the pervasive and intricate intercalation of fine and coarser sediment. Presently, however, ponded turbidite systems have only rarely been reported from the ancient deep-marine sedimentary record. Partly this is because their representative lithological characteristics have been poorly documented due to limited spatial scales, both laterally and vertically, of most outcrops, and their typical fine-grained nature that causes them to be easily weathered and vegetated.

This study investigates deposits of an ancient deep-marine slope leveed-channel complex and illustrates how the associated levee deposits exhibit many macroscopic and microscopic features that differ from those exhibited by classical Bouma turbidites. The channel complex is interpreted to consist of multiple channels that progressively aggraded and migrated to the

southeast and were bounded by levees formed from sediment that overspilled the channel margins. Over northwest-facing levees overspill, flows moved away from the channel (incident flow) but then encountered a topographical barrier (erosional escarpment) that set-up a return flow (*i.e.* bore) which then interacted with the incident flow. This interaction resulted in a highly complex flow field in this semi-confined basin, which in the Windermere sedimentary record is manifest by an intricately intercalated sand-mud succession termed the heterolithic unit. Distinctively, the heterolithic unit comprises beds exhibiting a vertical succession of facies that superficially resemble classical turbidites, but with a number of important differences. Specifically, an ideal bed consists of structureless overlain by planar-stratified sandstone (F1) overlain sharply by plane-parallel- then cross-, and finally wavy-stratified intercalated sand-mud strata of F2. This, then, is capped by plane-parallel-laminated and then structureless claystone of F3. However, compared to classical turbidites (*Fig. 5.1*), and in particular turbidites in most modern and ancient levee deposits, strata are ubiquitously planar stratified, which becomes better developed in the finer-grained F3 division at the top of beds. In addition, any cross-stratification is consistently low angle or even wavy, even though the grain size could easily support the development of current ripples forming high-angle cross-stratification. Notably also, the upward change from F1 to F2, or a change in the style of stratification (*i.e.* subfacies) within F2, is marked by a sharply bounded anomalously thick (1-2 cm), laterally continuous (10s metres along strike) sandy mud layer. Collectively all characteristics suggest that a suite of depositional conditions that differ from those that deposit more common classical turbidites, and is interpreted to be a consequence of topographical flow confinement.

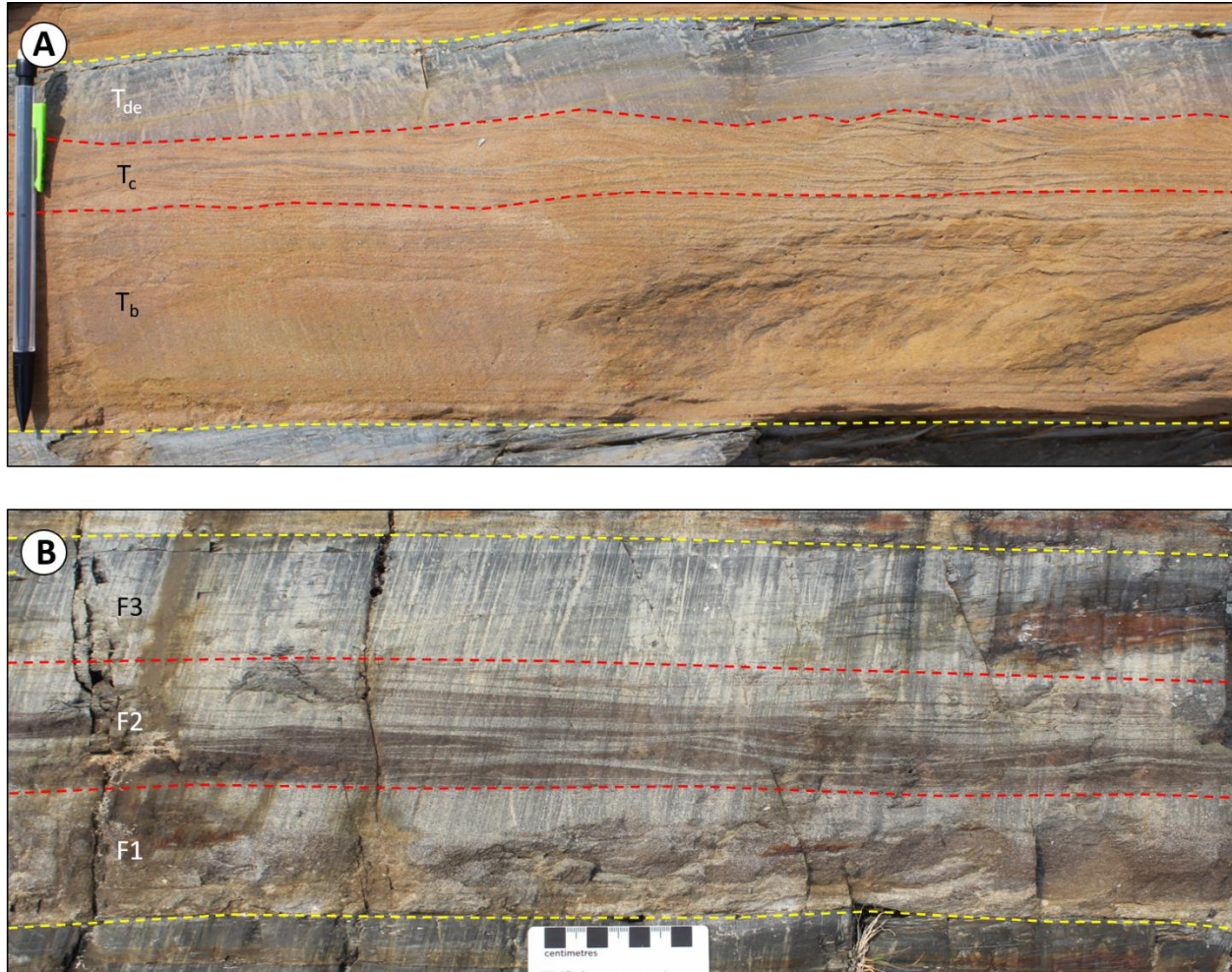


Figure 5.1: Comparison between T_{bcde} classical Bouma turbidite (A) and strata described in this study (B). A) Medium-grained, planar-laminated sandstone (T_b) overlain sharply by high-angle cross-stratified sandstone (T_c) overlain sharply by diffusely planar-laminated siltstone capped by structureless mudstone (T_{de}). B) Medium-grained, structureless sandstone (F1) overlain sharply by graded, low-angle cross- to wavy-stratified clayey sandstone interstratified with structureless sandy claystone (F2) overlain gradationally by graded, plane-parallel-laminated sandy claystone overlain sharply by graded structureless claystone (F3).

As noted above, an idealized bed consists of three major parts, here termed F1, F2 and F3. F1 is the thickest component (average 18 cm) and generally consists of medium-grained, carbonate-cemented sandstone with an average matrix content of 8%, but is always less than 10%. F1 is commonly planar-stratified but uncommonly structureless or wavy-stratified and is interpreted to represent deposition from sand-rich turbidity currents where structureless strata indicate rapid deposition from suspension whereas diffuse planar stratification, and even slightly wavy stratification, indicates reduced rates of sediment fallout and intermittent deposition from

crudely sorted laminar shear layers. This, in turn, is overlain by an anomalously thick sandy mud layer that is compositionally similar to other F2 mud layers but is notably thicker (commonly 1-2 cm) and more laterally extensive (10s metres along strike). Significantly, these thick sandy mud layers overlie an erosion surface that is interpreted to have formed by low frequency, high energy turbulent eddies that scoured the sea bed and increased the sediment load in the lower part of the flow. This, in turn, substantially reduced the turbulent kinetic energy as energy was consumed by accelerating and maintaining these newly scoured particles in suspension. The abundance of these particles, particularly the clay fraction, would have led to particle aggregation (*i.e.* flocculation) and ultimately gelling followed by *en masse* deposition of thick, laterally continuous, poorly-sorted, massive sandy mud layers. Following deposition of these mud layers, and depending on the hydraulic and sediment conditions in the near-bed region of the flow, deposition of plane planar- (F2.A), cross- (F2.B), or wavy-stratification (F2.C) ensued in generally fine-grained F2 strata; notably most F2 divisions (81%) consist of a single style of stratification (*i.e.* F2.A, F2.B, or F2.C). Significantly, F2 is composed of intricately interstratified mm- to cm-thick, sand-rich (sandy) and sand-poor (muddy) layers that show no systematic change in thickness upward, while the coarser sand fraction in sand layers and sediment dispersed in mud layers fines upward. However, from bed to bed sand layers vary considerably in their matrix content ranging from 10-60% and are subdivided into two end-members: matrix poor (10-30% matrix) and matrix rich (30-60% matrix).

Plane-parallel F2.A strata are typically matrix rich with a distinctive grey colour, and are interpreted to represent spatially uniform bed-surface (*i.e.* bedload) transport of mixed sand and mud sediment. Cross-stratified F2.B strata, on the other hand, are always matrix-poor and cemented with abundant ferroan calcite (up to 34% in thin section) that upon being weathered

exhibits a distinctive orange colour in the field. F2.B divisions consist of low-angle cross-stratified sets (average 10°) that are separated by thick, continuous sandy mud layers when two or more styles of stratification are deposited, and uncommonly climb at angles up to 15° . F2.B was deposited under lower energy conditions compared to F2.A, but more importantly in flows whose near-bed density structure allowed the development of a Kelvin-Helmholtz instability that coupled with bed-surface sediment transport and caused the inception and growth of bed surface defects that ultimately evolved into downstream-migrating ripples; sand became sequestered in the ripple crest whereas mud accumulated in the adjacent lower-energy trough.

Significantly, coincident with the thick sandy mud layer that separates cross-stratified sets, is the uncommon reversal in paleocurrent direction in the overlying set. Since the reversal(s) occurs during a single sedimentation event it suggests that during a single turbidity current local paleoflow was at times directed away (northwest) or toward (southeast) the channel, indicating transport by the incident or return flow, respectively. F2.B is then overlain by wavy-stratification (F2.C), which is the most common style of stratification in F2 divisions, and comprises sub-equal occurrences of matrix-poor, calcite cemented, and matrix-rich, calcite-cement poor end-members. Compared to F2.B, cross-stratification in F2.C is always much lower angle (typically *c.* 4°), and both the intercalated sand and mud layers tend to be more laterally continuous (> 1 m). F2.C formed as the flow speed continued to wane from F2.B, but where the near-bed hydrodynamic instability (although still capable of generating bed defects) was unable to cause sufficient amplification of the defect to adequately alter sediment transport patterns on the up- and downflow side of the defects. Accordingly, the bed surface remained wavy and simply aggraded.

Capping all beds is F3 strata that consist of two parts, a plane-parallel-laminated basal part made up of rhythmically interstratified sand (sandy claystone) and mud (claystone) laminae that

systematically fine and thin upward, and then are overlain sharply by a graded or ungraded claystone. Additionally, F3 strata that form the whole bed commonly stack forming packages up to 17 beds thick (more commonly 2-5 beds thick). F3 is interpreted to have been deposited from a dilute, fine-grained suspensions where flow competence during a single sedimentation event progressively reduced causing gradual upward fining. Pervasive planar interlamination of sand and mud laminae in the basal part of F3 strata may relate to the action of internal waves within the channelized turbidity current; or a result of alternating rheological changes in the viscous sublayer of a low-energy turbidity flow, specifically episodes of relatively coarse particle settling followed by near-bed gelling and deposition of a finer-grained clay layer. The massive or graded claystone cap is then deposited via suspension fallout in the final stages of the flow and later hemipelagic fallout.

Based on detailed cm-scale stratigraphic logging, mapping and bed-by-bed correlation in outcrop, and petrographic facies analysis, studies like the research described here will help to better constrain the lithological and architectural evidence suggestive of partial or complete flow confinement – details that most certainly are below seismic resolution but fully captured in core. These features include bed tabularity, anomalously thick, stratified mud caps, interstratification of sand-mud layers on a variety of thickness and lateral scales, and paleoflow reversals. Incredible tabularity is observed in all heterolithic strata, whether they be millimetres to metres in thickness, but consistently exhibit negligible (up to 10s cm/km) changes in thickness over large (at least 100s m) spatial scales. Thick mud caps appear structureless on weathered outcrop surfaces, but on inspection of freshly cut surfaces and orientated thin sections, are shown to be rhythmically stratified and superimposed on a larger scale fining- and thinning-upwards trend associated with an individual sedimentation event. This contrasts the structureless nature of thick mud caps

commonly reported by previous researchers who interpret deposition under similarly confined flow conditions but by simple suspension fallout. Here the intricate interstratification of sand and mud layers and paleoflow reversals are interpreted to be not only a manifestation of flow confinement (*i.e.* F2 and F3), but also the complex interaction of an incident and return flow.

The ability to recognize the distinctive sub-seismic lithological characteristics of heterolithic strata, and to appreciate and understand their interpretive significance, will aid in hydrocarbon reservoir modelling as it will give insight about the predictable vertical and lateral changes in lithology that affect reservoir compartmentalization, quality, and performance. Moreover, the generally fine-grained nature and poor sorting of these strata will most likely have an important influence on fluid flow pathways in deep-marine hydrocarbon reservoirs and as a result, understanding their stratal architecture and spatial distribution is essential.

Areas for future work include:

- More experimental work on clay-rich sediment-gravity flows at different clay concentrations and clay types/mixtures and compare with characteristics of the resultant deposit.
- Closer examination of the petrographic characteristics, especially the recrystallized clay content, in the different kinds of F2 stratification, and how these characteristics change vertically in a single F2 division that comprises more than one F2 subfacies.
- A more exhaustive comparison of sedimentary textures and structures in F2.B strata and the cross-stratified T_c division observed in classical (Bouma) turbidites elsewhere at Castle Creek and described in the scientific literature.
- Collection of more paleocurrent data, especially in the heterolithic unit. Due to the two-dimensionality of the outcrop surface (a consequence of glacial erosion and polishing) and

the general lack of cross-stratification reliable paleocurrent direction was difficult to assess. However, with the use of a rock saw multiple cuts can be made into the strata at different angles to expose a different perspective of targeted cross-stratified layers, which would improve the accuracy of paleocurrent data and strengthen the depositional model proposed here.

- More detailed analysis of planar-stratified F3 strata, particularly assessing the lateral extent of the intercalated sand and mud layers. This could be accomplished by cutting along strike within an F3 division and comparing this to the T_d division observed in classical turbidites.

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Appendix 1: General Petrographic Characteristics

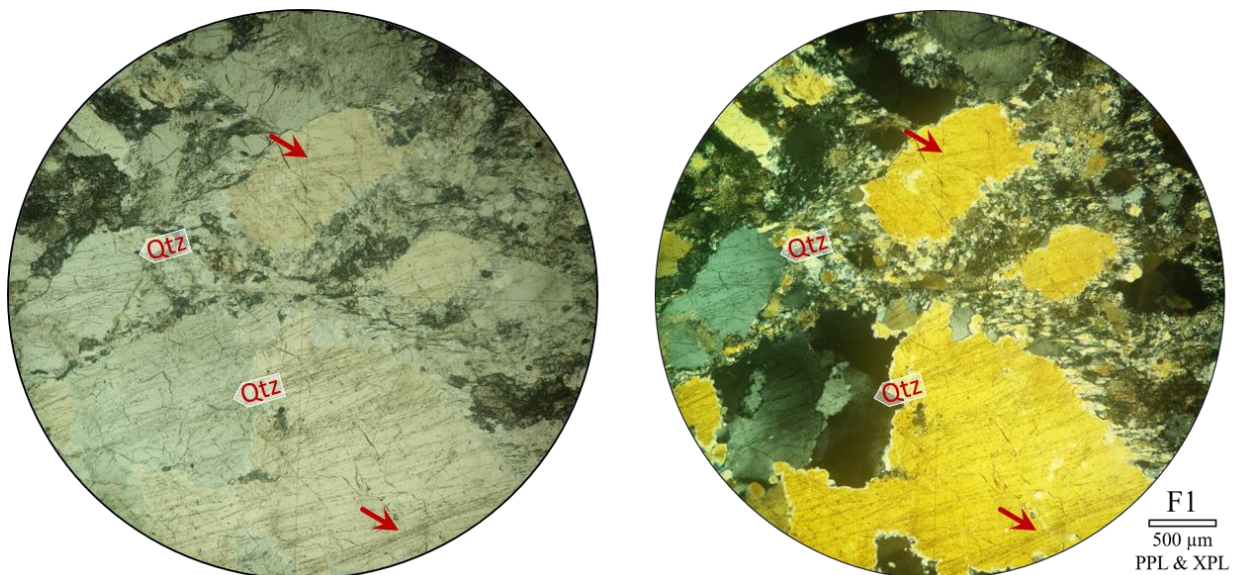


Figure A1.1: Boehm lamellae are observed in various detrital quartz (Qtz) grains (both mono- and polycrystalline here) with thin subparallel lines of micron- to submicron-sized vacuoles (trails of dusty bubble/inclusions in quartz indicated by red arrows) that may be relicts of deformation lamellae formed under intense strain. These lamellae probably formed *in situ* since they appear in most detrital quartz grains and are similarly oriented. Note that this thin section was prepared too thick resulting in higher than normal yellow interference colours of quartz in XPL.

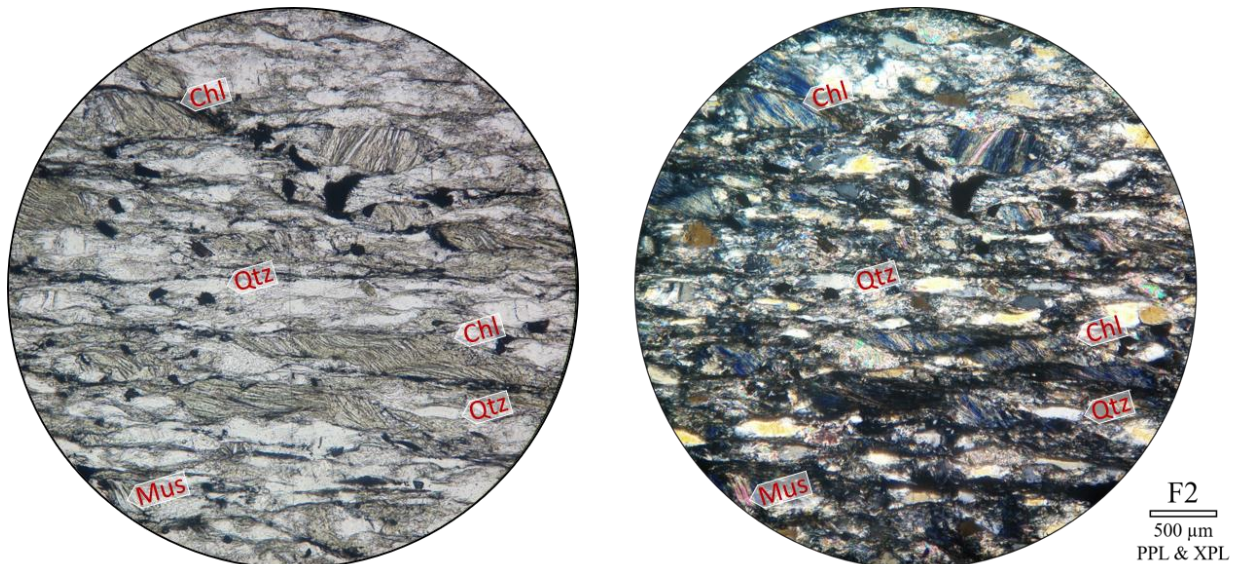


Figure A2.2: Abundant chlorite (Chl) porphyroblasts (light green pleochroism in PPL and anomalous birefringence colours such as deep blue in XPL) which are the product of detrital clay recrystallization. Large chlorite crystals tend to exhibit well developed crenulation (sigmoidal) cleavage with rare mica (Mus) inclusions oriented parallel to bedding.

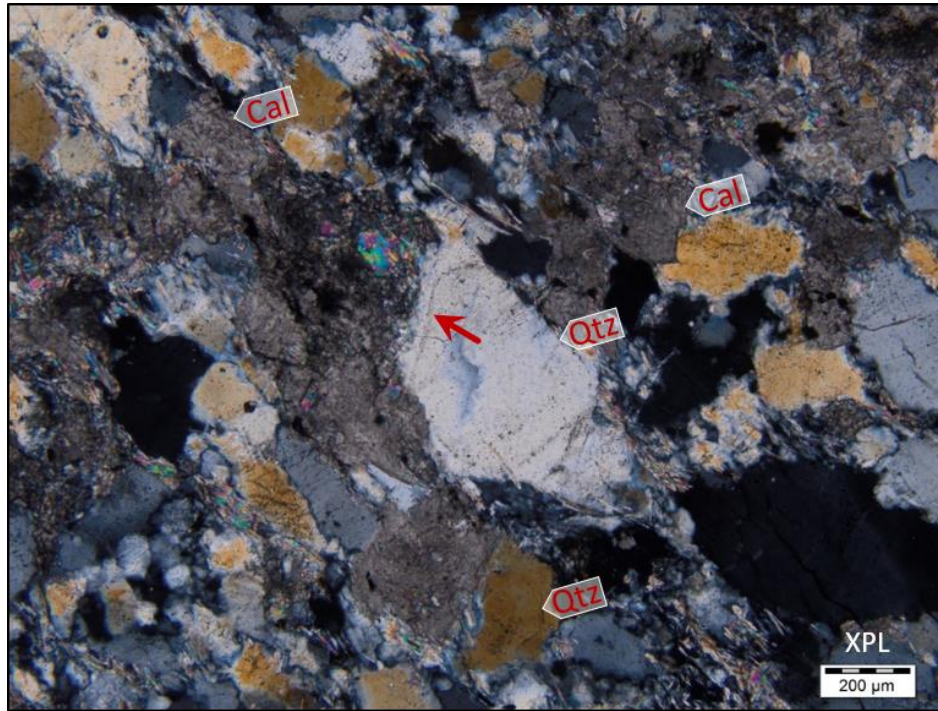


Figure A1.3: Quartz overgrowth (indicated by red arrow) partially surrounds a detrital quartz (Qtz) grain. Due to intense compaction and deformation, framework grains (most commonly quartz in this study) in matrix-poor strata with abundant grain-to-grain contacts display sutured boundaries (not observed here) due to the dissolution of silica. This silica then precipitates around detrital grains filling pore space as authigenic syntaxial quartz overgrowths separated from the detrital grain by a thin dark layer(s) of vacuoles (observed here), clay, and/or iron oxide coatings (‘dust’ rims) (Ulmer-Scholle *et al.*, 2014).

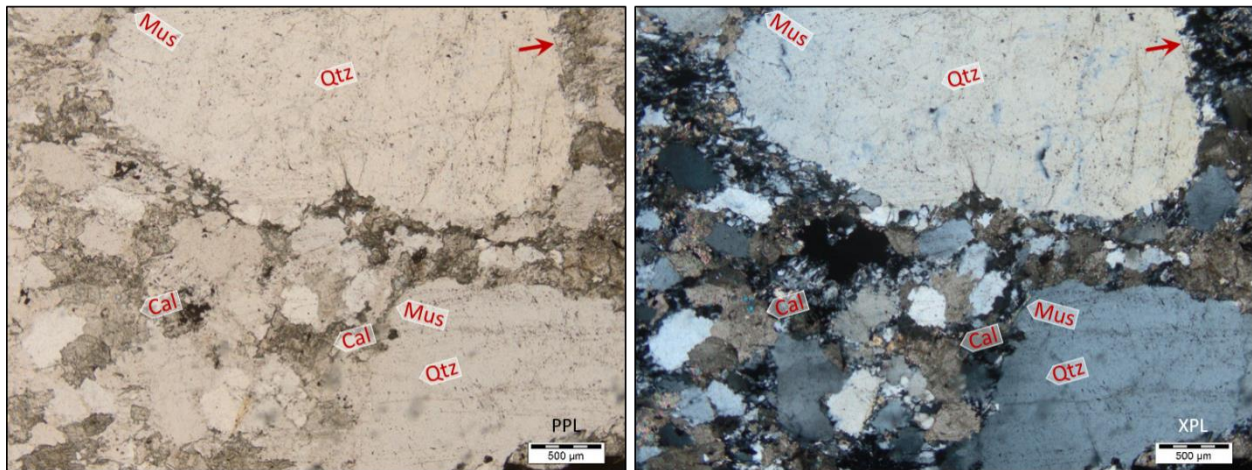


Figure A1.4: Grain rimming detrital mica (Mus; uncommon here) around quartz (Qtz) grains. Also note the quartz overgrowth (red arrows) partially surrounding the quartz grain which has been further altered due to chemical compaction causing the sutured grain boundaries, and bulging recrystallization due to dynamic quartz recrystallization (see discussion in 3.2) due to low-grade metamorphism. Bulging recrystallization forms multiple recrystallized unstrained silt to fine sand-

sized grains along grain boundaries as observed around larger quartz grains in this photomicrograph.

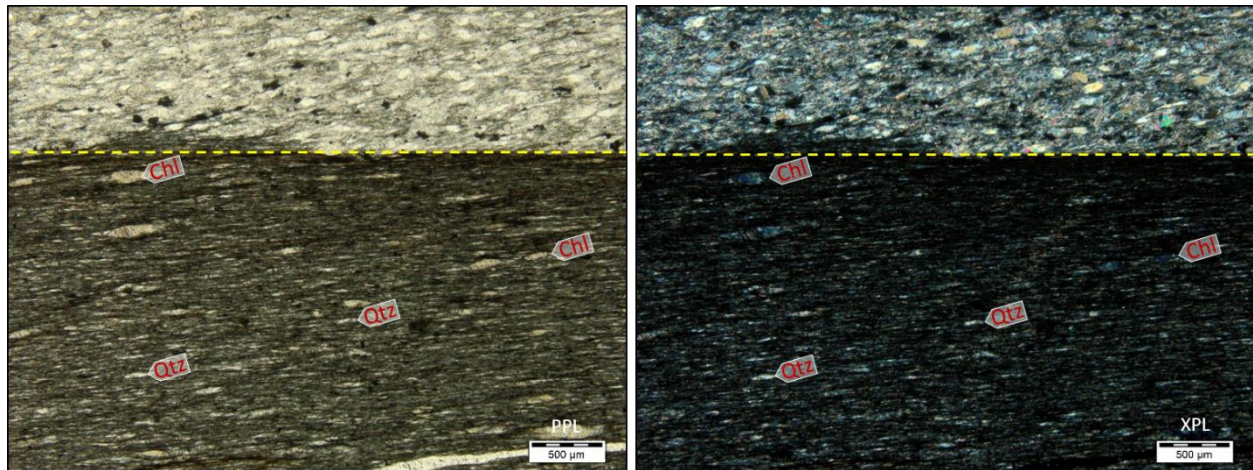


Figure A1.5: Sand-sized chlorite (Chl) porphyroblasts scattered in recrystallized mica and silt (Qtz) in a claystone cap (top of bed indicated by yellow dashed line). Note their subparallel and inclined orientation that parallels (tectonic) cleavage. Chlorite porphyroblasts are common in matrix-rich strata, especially in the claystone cap (F3) of almost every bed and makes determining grain size assessment in the field (using one's teeth) difficult due to the inability to distinguish chlorite from sand-sized framework grains, and indicates the need for microscopic analysis.

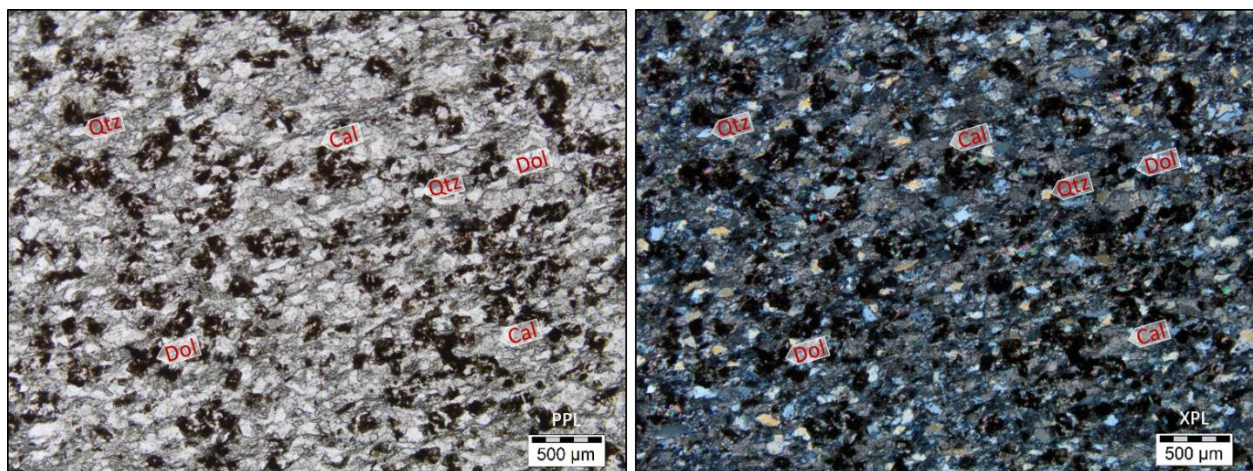


Figure A1.6: Dark brown clusters and rare individual crystals of partially developed or altered interstitial rhombohedral dolomite (Dol) and pervasive pore-filling calcite cement (high interference colors in XPL). Some dolomite crystals have undergone dedolomitization with relic dolomite (ghost boundary) filled with calcite.

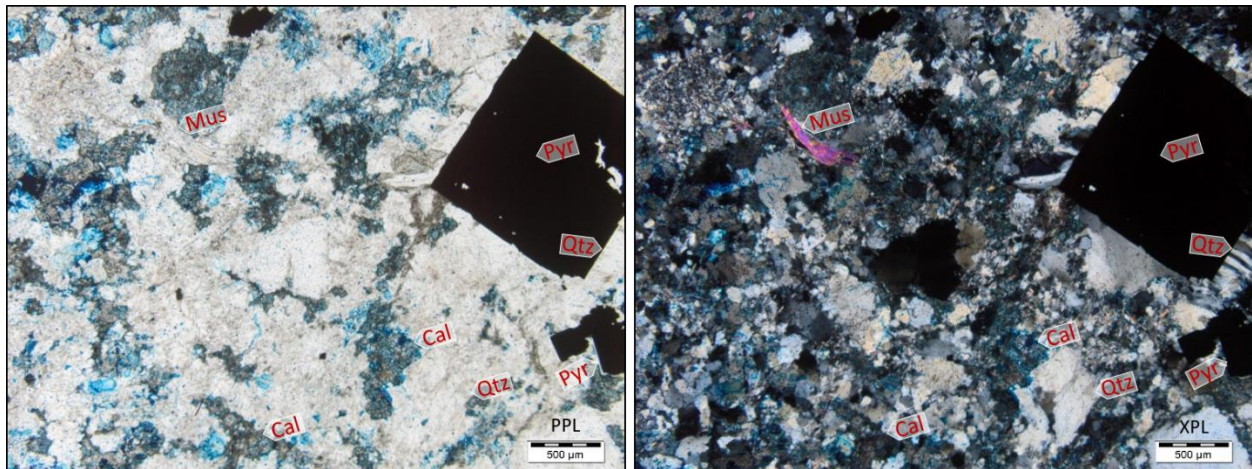


Figure A1.7: Large, isolated cubic (metamorphic) pyrite (Pyr) crystal (black, opaque) with fibrous quartz radiating away from one of the crystal boundaries (pressure shadow). Smaller pyrite crystals tend to form clusters (just below the larger pyrite crystal). Also, note the mechanically (ductile) deformed detrital muscovite grain (Mus) that has been partially compacted between quartz grains. This is indicated by the kinked centre and expanded bottom part of the grain (a commonly feature in highly foliated phyllosilicates). Thin section has been strained with AFeS to identify ferroan carbonates.

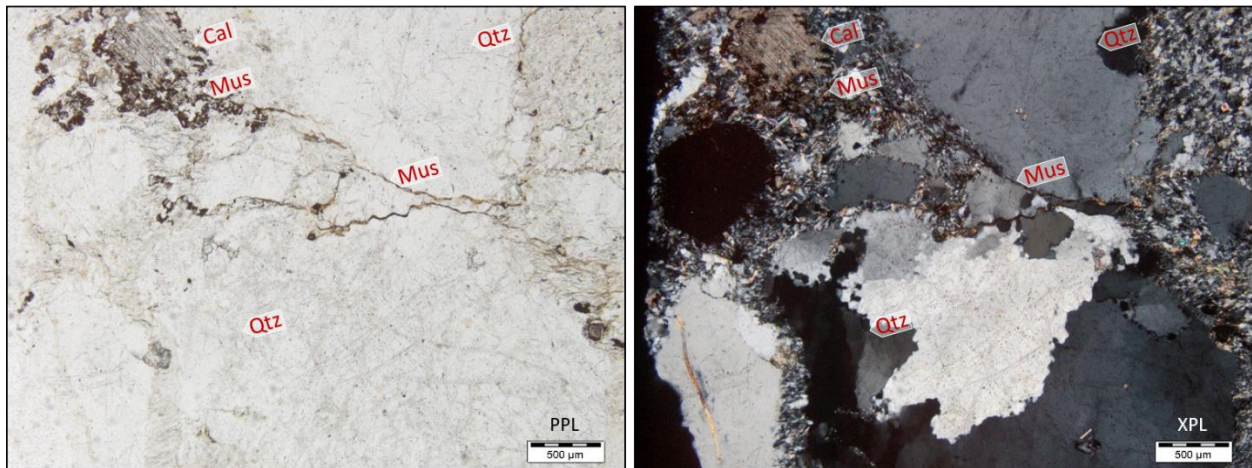


Figure A1.8: A large (granule) polycrystalline quartz (Qtz) grain with undulose extinction. Polycrystalline texture of quartz grains are rare in this strata (most common in F1 strata with grain sizes > 500 µm), but where observed, often exhibit undulatory extinction and elongation of individual subcrystals with sutured boundaries. Also notice that thin, fibrous mica crystals (Mus) have been included in some calcite (Cal) cement and rim some quartz grains.

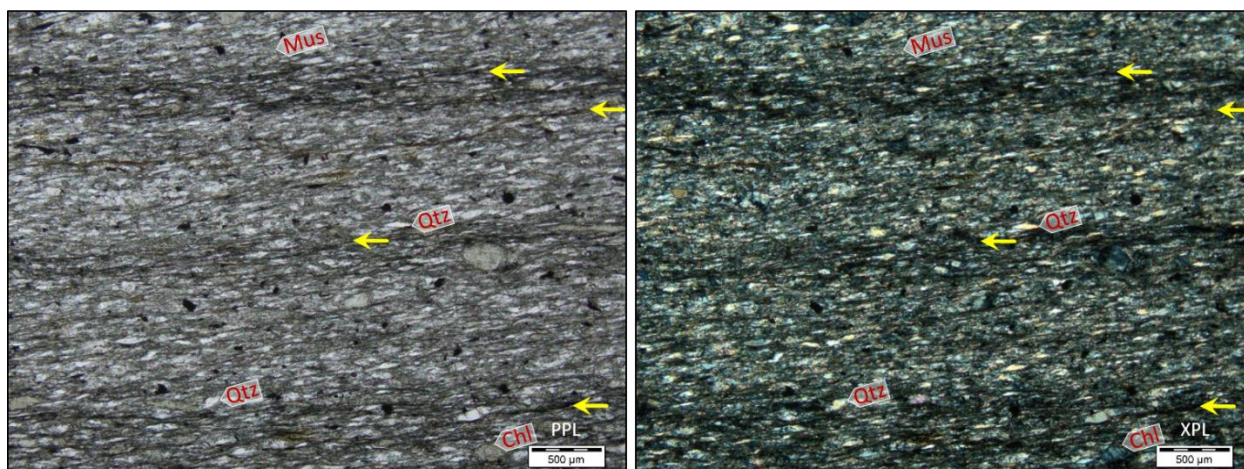


Figure A1.9: Chemical compaction of matrix-rich strata has produced numerous bedding-parallel, thin ($< 150 \mu\text{m}$), dark, wispy, and elongate solution seams or microstylolites (indicated by yellow arrows) containing elevated amounts of insoluble detrital clays (recrystallized to mica and chlorite here), organic matter, iron oxides and/or sulphides (pyrite mainly). These features are common in matrix-rich strata such as F2-F4.

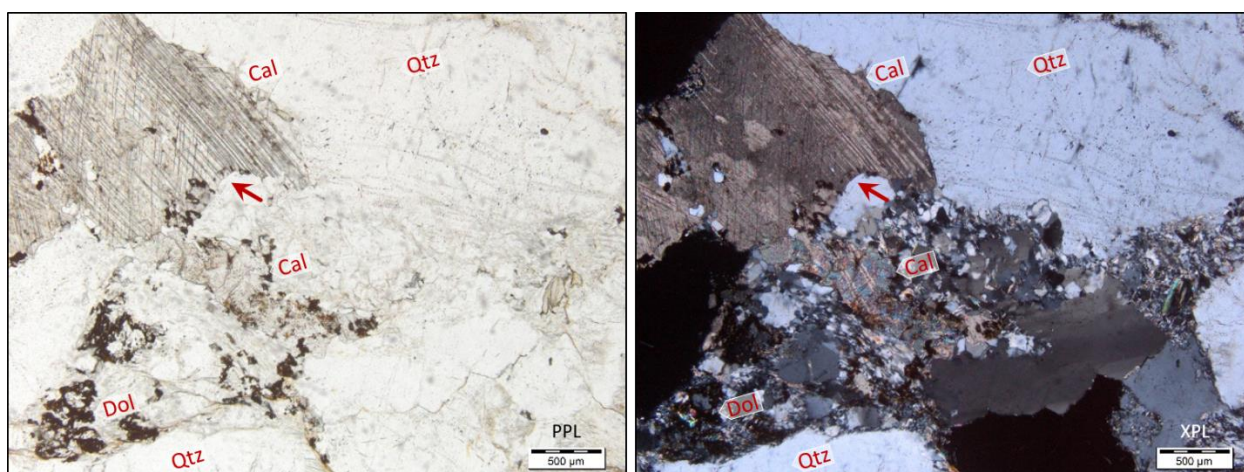


Figure A1.10: High relief, very coarse, twinned calcite crystal, which likely formed as a secondary cement over an initial silica cement, indicated by the subtle 'dust' rim (red arrow) separating the grain below from the quartz (overgrowth) cement on an adjacent grain. The well-developed calcite twins are diagnostic of low stress-induced mechanical deformation.

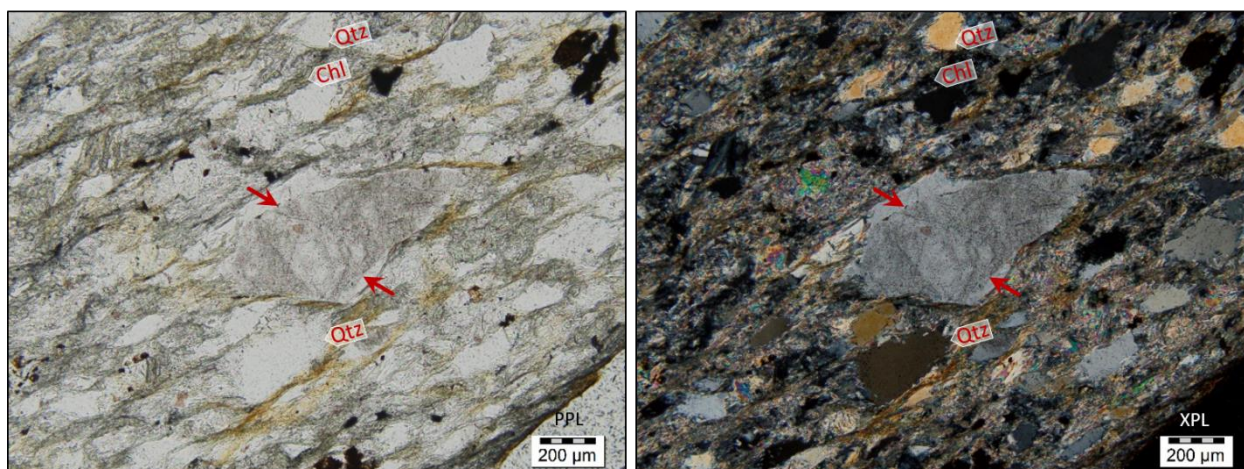


Figure A1.11: A rare altered orthoclase grain (between red arrows) with feldspar overgrowth. The feldspar grain has a cloudy appearance due to abundant vacuoles formed during alteration, thus leaving the later overgrown cement relatively clear with fewer inclusions.

Appendix 2: Correlation Panel

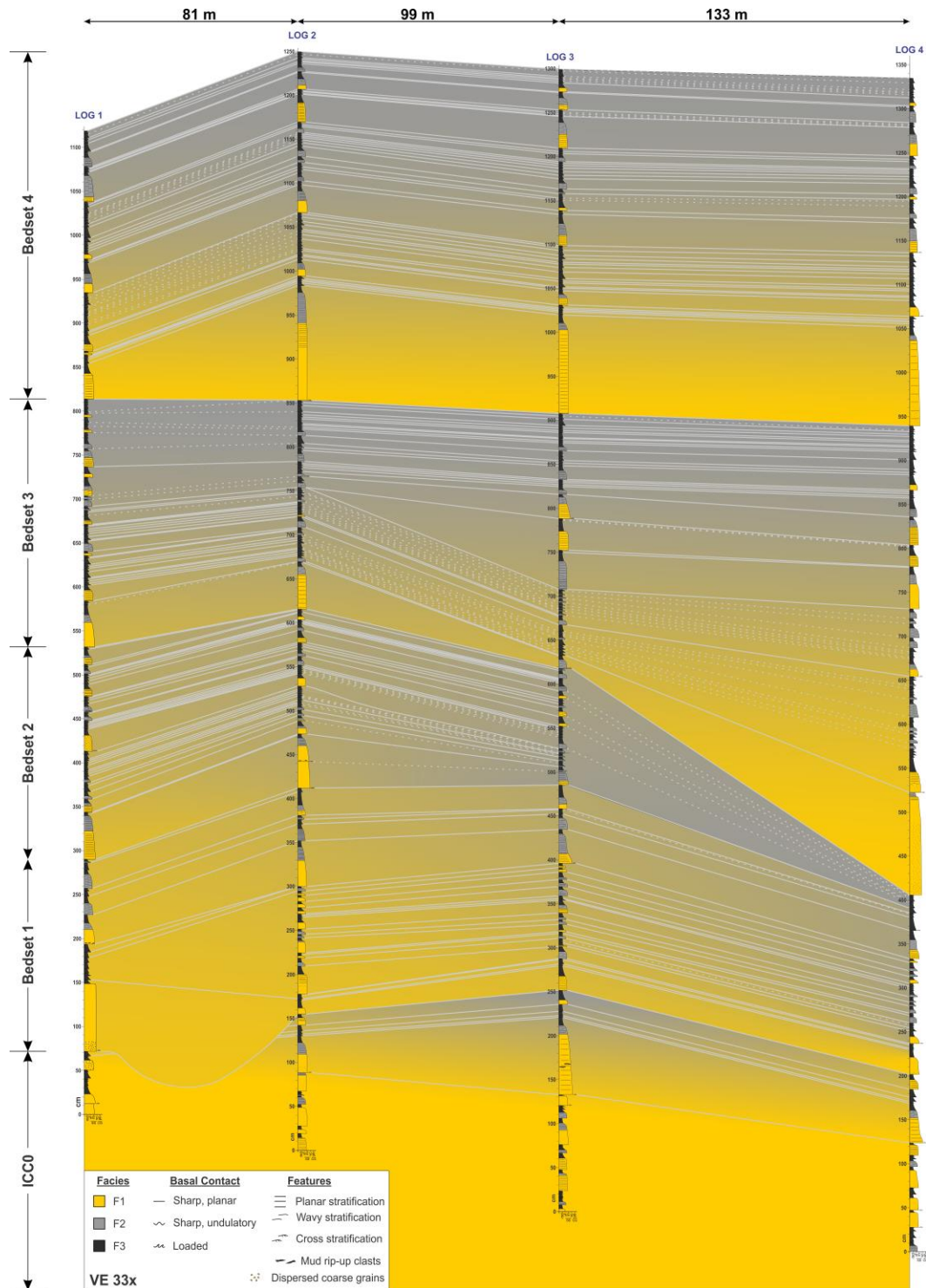


Figure A.2: Correlation panel of the heterolithic unit across the study area. Four stratigraphic logs that range in thickness from 11.2 to 12.5 m were measured in bed-by-bed detail and physically traced across the entire area. Note the large vertical exaggeration (33x) which allows subtle thickening and thinning trends to be detected in apparently tabular strata.